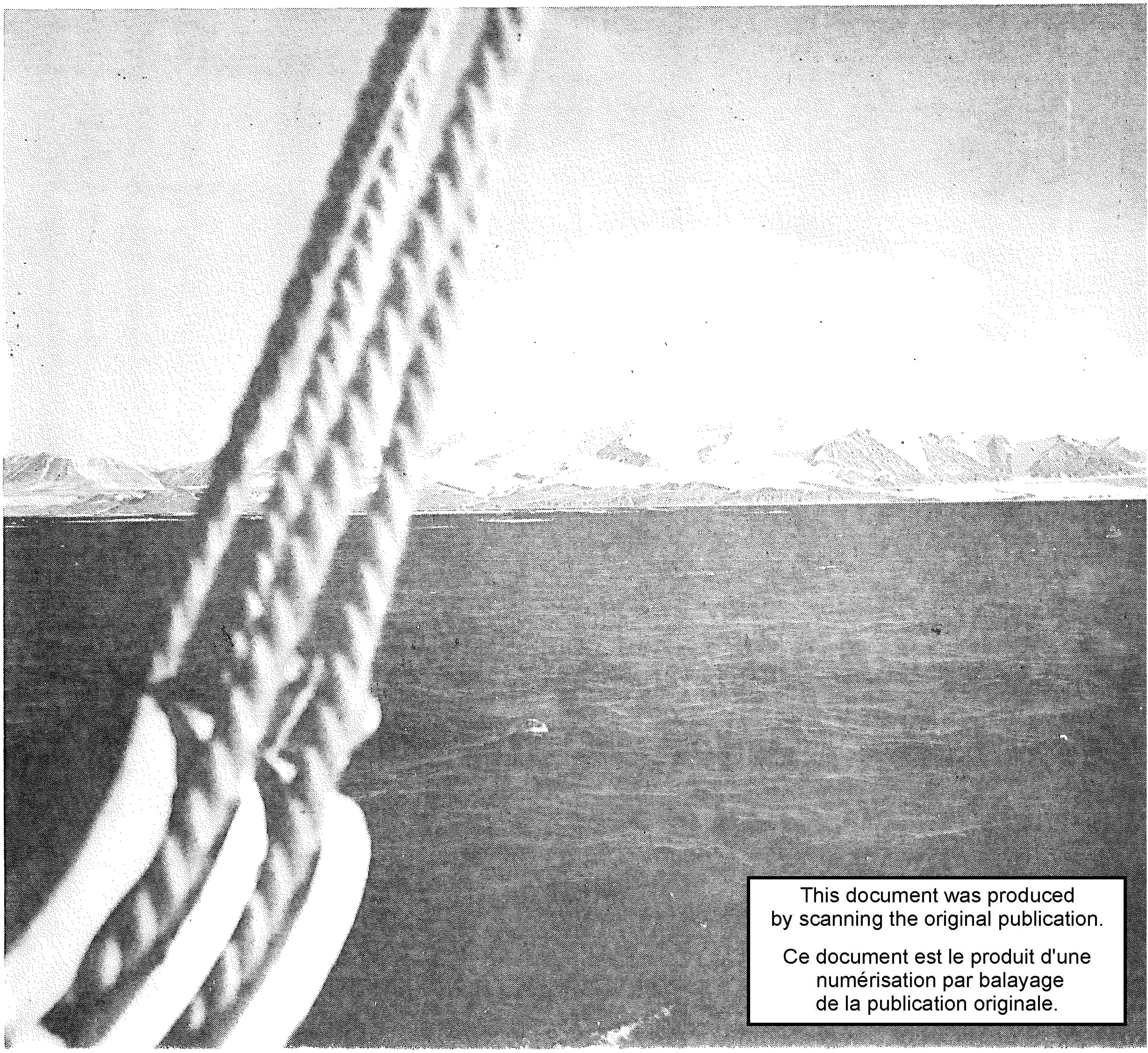
14TH ARCTIC WORKSHOP

NOVEMBER 6-8, 1985

ARCTIC LAND-SEA INTERACTION

GLACIERS, SEDIMENT AND SEA LEVEL, NORTHERN BAY OF FUNDY, NOVA SCOTIA



14th ARCTIC WORKSHOP, NOVA SCOTIA, 1985
FIELD TRIP B
GLACIERS, SEDIMENT AND SEA LEVEL; NORTHERN BAY OF FUNDY, N.S.

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Abstract

The Bay of Fundy coastal area of northern Nova Scotia presents a variety of features which document the changing relationships of land and sea, and the processes which acted upon them during lateglacial and postglacial time. Tills and ice-flow indicators record four glacial advances from different dispersal centres. Buried organic beds give evidence of the intervening nonglacial periods of which the youngest relates to a sharp cooling after 11 ka. Palynology of lake sediment reveals the postglacial climatic evolution.

This excursion (Fig. 1) focuses primarily on features related to the last deglaciation when, as ice was retreating from the north side of Minas Basin, the depressed coast permitted a marine invasion (De Geer Sea) against the ice front where moraines and ice-marginal deltas mark its culmination. The present elevation of the marine facies has been tilted eastward from a maximum 37 m above present tide level to zero over a distance of 90 km. The subsequent glacio-isostatic regression caused truncation and dissection of the marine member, while a fluvial blanket was deposited initially as outwash fans, and later as postglacial plains and terraces. Relative sea level dropped to as much as -40 m from which position a renewed late Holocene submergence caused tidal marsh muds to aggrade.

Dating of tree stumps beneath the tidal mud, and of peats within the marsh sequence show that, while tide level has risen an average of 30 cm per century during the last 4000 years, there were several minor stillstands or regressions. Part of the drowning is due to increasing tidal range — a hypothesis that has been supported by mathematical modelling. The submergence is however a regional phenomenon that is basically due to broad crustal subsidence, perhaps because of a collapsing glacier forebulge. This factor too has been simulated by rheological modelling of crustal movements based on postglacial relative sea-level change. The modern transgression constitutes the final land/sea interaction resulting from glaciation. It has profound coastal consequences including intensified local siltation, extensive erosion whereby barriers grow and migrate, and cliffs and platforms are cut in all rock types. The impact on biologic systems, including man's works, is significant.

OUTLINE OF ROUTE AND STOPS

STOP NO.	
	Halifax/Dartmouth to Truro
	Introduction to general Quaternary lithostratigraphic sequence
	red clay (Lawrencetown Till) of northern provenance over greywacke
	stacked tills and organic beds
	Truro-Glenholme
1	Debert **Paleo-Indian** site (10.5 ka); climatic deterioration?
2	overview of dykelands
	outwash fans dissected and onlapped by salt marsh mud
3	submarine bedforms in a macrotidal estuary
4	submerged forests (3 ^{14}C dates) at Lyon Head, Highland Village
	Saints' Rest (Bass River)
5	**submerged forest** (8 ka) overlain by pond clay and by tidal marsh (1 ka)
6	glaciofluvial terrace truncated by regression
	Upper Economy
7	Peat bed (10.1 ka) under gravel and sand (=lateglacial **climatic reversal?**
8	Economy Point
	Five Islands
9, 10	glaciomarine silt/clay rhythmites (over ice-contact gravel?)
	overlain by outwash up to 14 m above tide level
	= **type locality of Five Islands Formation** -- the lateglacial fluviomarine unit comprising a lower marine Advocate Harbour Member, and upper fluvial Saints' Rest Member
	Moose River Woods
	Cobequid fault (neotectonic activity?)
	Parrsboro
11-19	end moraines, kames, kettles (= **major ice-marginal stand**)
	Gilbert Lake/Leak Lake **palynology** 15 ka BP to present
OVERNIGHT	outwash fan over kettled glacial delta, degradational terraces, resubmergence
	Diligent River
20	kettled outwash fan
	Fox River - Wards Brook
21	**ice-proximal fan-deltas** with kettles, terraces
	Spencers Island
22	Gilbert-type delta with **fossiliferous** clay-silt bottomset beds
	Advocate Harbour
23	emerged **beach plain** with spits like modern barriers
24	sedimentologic estimate of paleotides
25	possible interglacial rock platform
	Cape Chignecto - Squally Point
	rock platform and beach at marine limit (37 m)
	Apple River
	major **ice-marginal zone** with moraines, meltwater channels, kames
	Clam Cove
26	raised beach at marine limit (23 m)
	Joggins
27	**three superposed tills**
	Lower Cove
28	sand over **peat 11.8-11.1 ka** (=Younger Dryas cooling event?)
	Amherst
	extensive dykelands
29	tidal marsh invading living forest
30	deep drilling; Holocene perimarine stratigraphy
	Fort Beauséjour
31	**drowned forest** at -12 m beneath red and grey tidal mud with peat layers
	Folly Lake - Wentworth Valley
	gigantic morainal plug in wind gap

INTRODUCTION

Bedrock Geology and Physiography

Northern Nova Scotia has essentially two terranes (Fig. 2). A central elongate flat-topped upland (ca. 200-300 m), called **Cobequid Mountain** is composed of Silurian to Carboniferous igneous and metamorphic rocks. It is deeply dissected by steep gorges and has two wind gaps holding important morainal deposits. The upland is surrounded by low plains; that to north is called the **Cumberland Lowlands** and is underlain mainly by Carboniferous red and grey sandstones, while to the south along Minas Basin the **Annapolis Lowland** is composed of Triassic sandstone and shale, and locally basalt. At times during glacial advance the Cobequid massif must have deflected the ice flow; during deglacial thinning and retreat the lowlands localized major ice bodies, while the adjoining submarine basins probably promoted marine incursion by calving. The contrasting rock types in these terranes provide indicators for assessing the changing provenance of successive till sheets, and for determining the direction of glacial advance and retreat independently of striations.

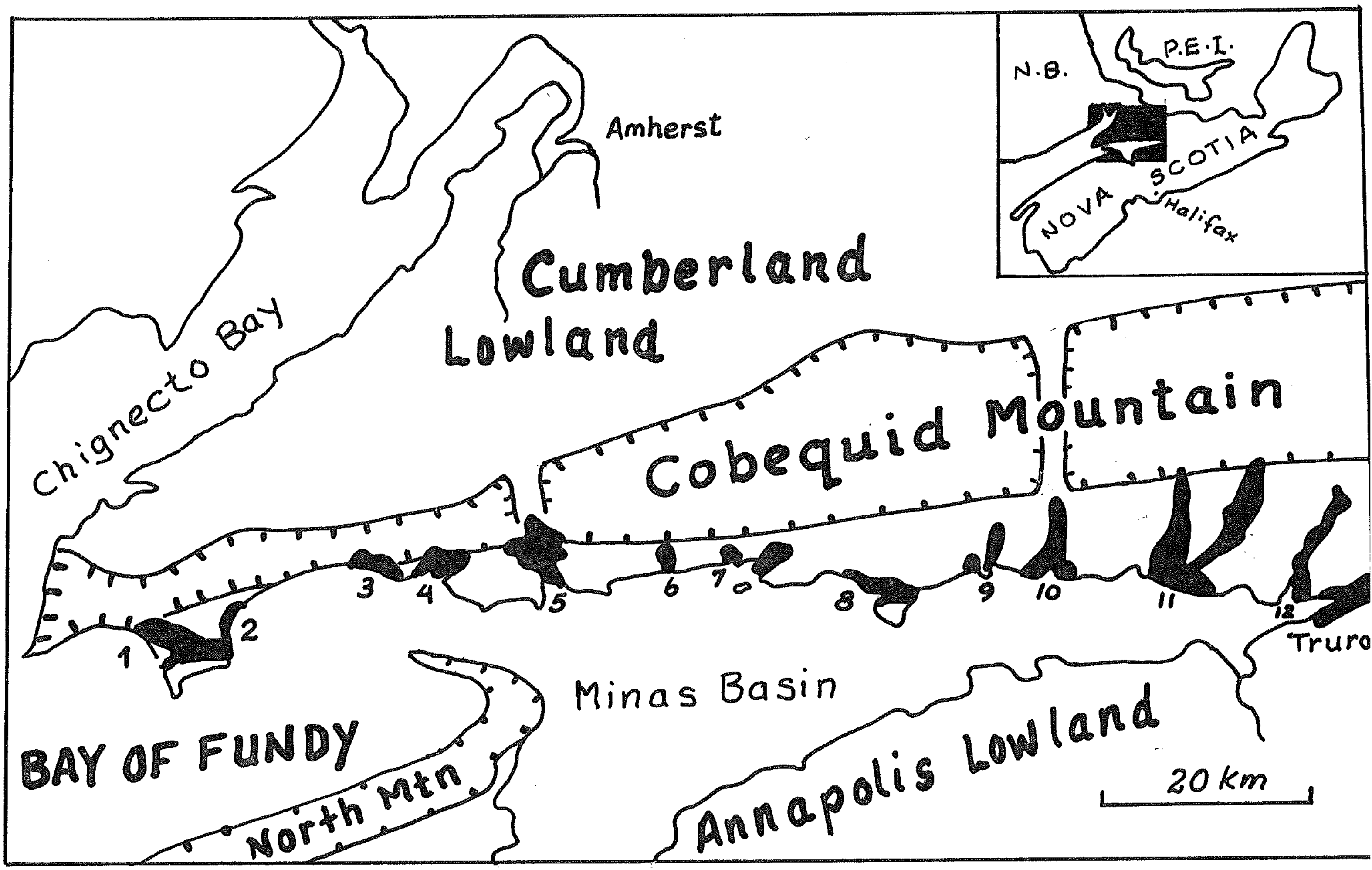

Fig. 2 Bedrock geology and physiography of Minas Basn area showing location of major outwash fan-deltas at: (1) Advocate Harbour; (2) Spencers Island; (3) Wards Brook/Port Greville/Fox R.; (4) Diligent River; (5) Parrsboro; (6) Moose River; (7) Five Islands; (8) Economy; (9) Bass River; (10) Portapique; (11) Debert River; (12) Salmon River. (after Wightman, 1980, p. 3).

Quaternary Geology

The Quaternary geology of the area has been studied for more than 100 years, notably by Dawson (1855), Chalmers (1895), Goldthwait (1924) and Wickenden (1941). Some highlights of the area were described in a guidebook for the 24th International Geological Congress (Prest et al., 1972). However, details of the glacial sequence and sedimentary succession have come into focus only in relatively recent years, mainly through systematic mapping and stratigraphic studies by R.R. Stea of the Nova Scotia Department of Mines and Energy. His work provides the essential basis for understanding the Late Quaternary history of northern Nova Scotia and this guidebook draws heavily on his findings.

Much of the area is mantled by **till** that varies in thickness and texture mainly according to the grain size and hardness of the parent bedrock. Thick muddy tills thus tend to occur on the lowlands, and thin stony drifts on the uplands. Their superposition and lithology, linked to numerous striation measurements has served to elucidate a four-fold sequence of major ice movements. Small end moraines, eskers and sidehill meltwater channels occur locally which corroborate retreat directions inferred from ice flow trends. Built during retreat of the ice mass that lay on and north of Cobequid Mountain ("Flow Pattern 4") was the Minas Basin **fluviomarine outwash terrace** that comprises a series of fans and ice-marginal deltas (Figure 2) commonly joined to **ice-contact kames** and morainic deposits. The abundant Cobequid stones in the outwash shows that meltwater issued from ice receding northward. In the area below high tide level, and now largely reclaimed from the sea by dykes, are extensive areas of **salt marsh mud** consisting of finely laminated clay-silt with organic horizons. Its contact with the substrate, often marked by a zone of terrestrial vegetation, can be viewed along tidal gullies throughout the full range of the tides which here reach a world maximum of 16 m. The Holocene rise of sea level is documented by dates on organics in and under the tidal mud sequence.

The narrower topic of relative sea-level changes following the last glaciation lends interesting and unique aspects to the overall history. Firstly, on the subject of deglacial submergence, the first isobase maps of "marine limit" were by De Geer (1892), Fairchild (1918) and Flint (1940). Modern interpretations by Wightman and Cooke (1978) (Figure 3) and Grant (1980) support the earliest reconstruction and further illustrate that inner Bay of Fundy straddles a strongly warped zone at the limit of net emergence. Lougee (1953) termed this deeper postglacial phase of Bay of Fundy the **De Geer Sea.** Although it was Goldthwait (1924) who first outlined the general features of the Minas Basin outwash terrace, detailed work on the sedimentology began with Borns (1965, 1966) and Swift and Borns (1967) who defined it and the associated fluvial outwash member as the **Five Islands Formation.** Wightman (1976) interpreted the lateglacial tidal range from raised beaches at Advocate. Wightman (1980) documented the sedimentology of the Five Islands Formation along the northern Minas Basin coast and used the upper limit of the delta remnants to define marine limit (Fig. 4). His maps and diagrams are used extensively in this guide.

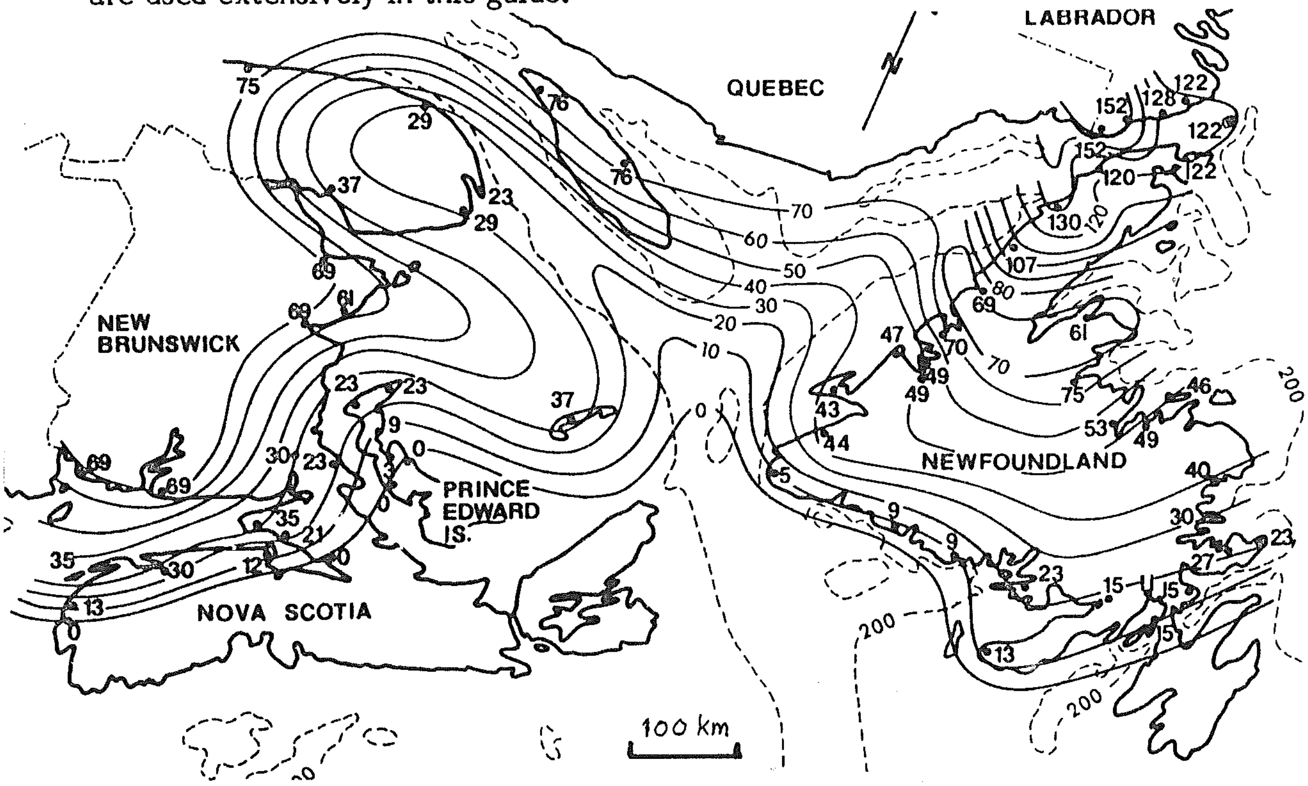

Fig. 3 Isopleths of net emergence in Atlantic Canada (elevations in metres above mean sea level; contour interval 10 m) (from Wightman and Cooke, 1980)

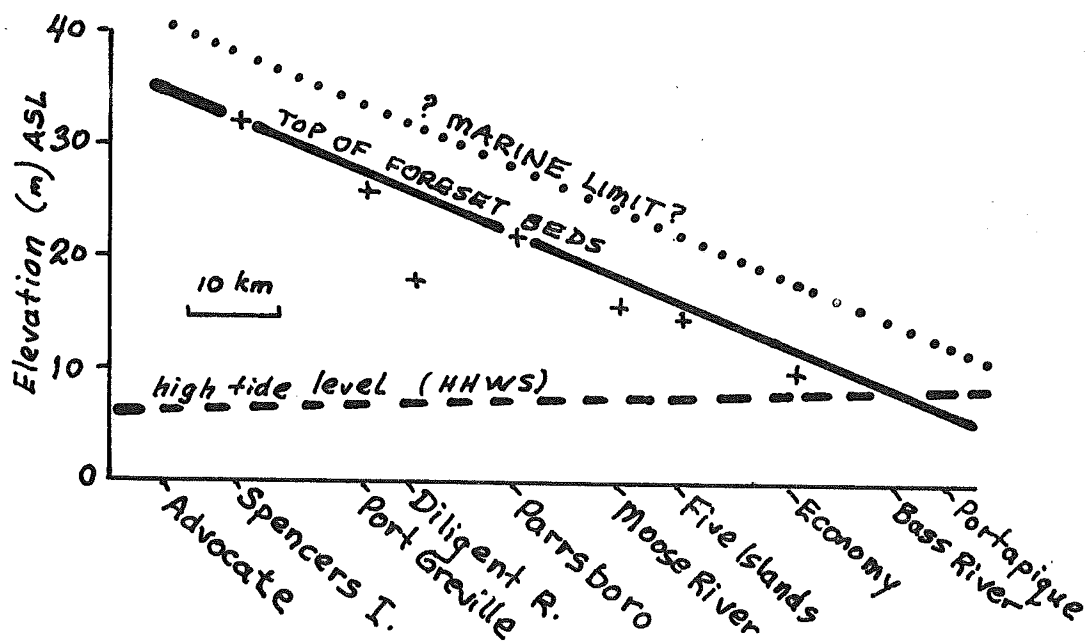

Fig. 4 Tilted lateglacial paleoshore, as defined by upper limit of truncated delta foreset beds, Minas Basin north shore; dotted line = maximum marine limit; dashed line = present high tide level. (after Wightman, 1980, p. 345).

On the subject of the modern re-submergence that is now in progress, Dawson (1856, 1868) first noted forest beds beneath coastal muds 10 m below tide level. Chalmers (1895, p. 125m) showed a drill log that reported terrestrial peat 22 m below high tide. Modern borings intersect the forest bed at depths as great as 40 m deep at head of Bay of Fundy. Dating of these submerged forests to trace the rise of tide level was first reported by Harrison and Lyon (1963). Measurements were extended over a large part of the Bay by Grant (1970) (Fig. 5) who inferred that tides in the Bay began with a nearly tideless De Geer Sea and increased nonlinearly between 4000 and 1000 y BP. Most studies yield average rates of relative rise of 30 cm/100 y, although local differences as well as fluctuations and/or stillstands are possible.

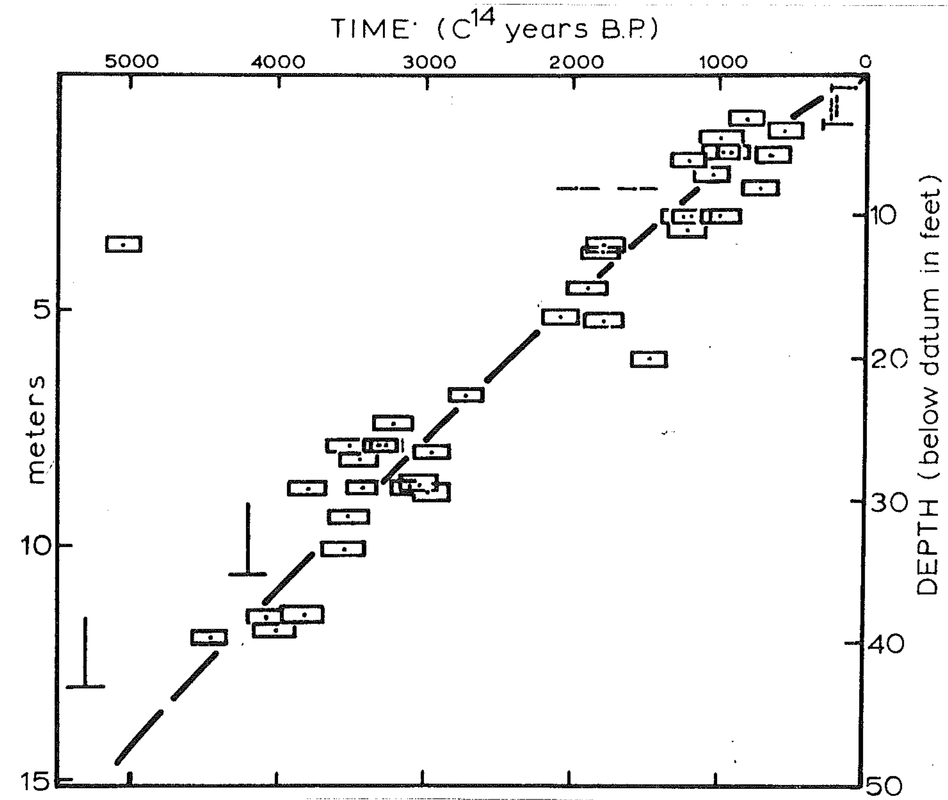

Fig. 5 Generalised trend of rising high tide level in Bay of Fundy area in Late Holocene time based mainly on ages of submerged tree-stumps and basal humus layer beneath tidal marsh mud. Reference datum is present maximum annual tide level. (from Grant, 1975).

The Bay of Fundy area therefore has a two-part sea-level history -- an early deglacial emergence and a modern re-submergence (Fig. 6). The interplay of crustal movement and sea-level recovery is thus transitional between the continuous submergence that characterises the marginal glaciated areas farther to the south, and the continuous emergence that characterises northern areas covered by the Laurentide Ice Sheet.

The most recent development in the field of shorelevel research has been the application of mathematical models to paleosealevel data using computer simulation to reconstruct paleoenvironmental conditions. Using total postglacial relative sea level change at numerous localities in the region, Quinlan and

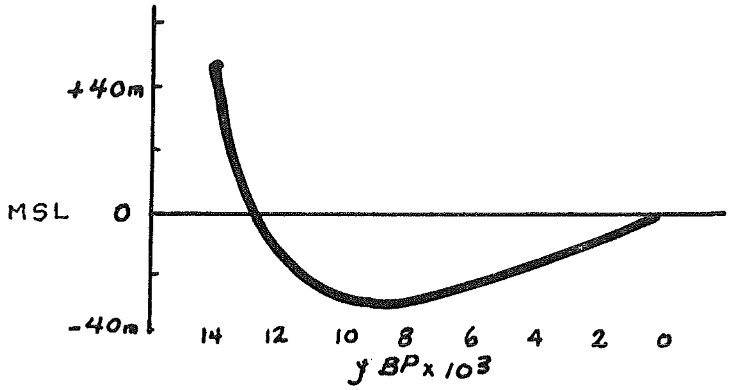

Fig. 6 Total hypothetical postglacial relative sea-level change in northern Bay of Fundy area based on various sources.

Beaumont (1982) modelled the former glacier distribution and found that ice had been localized mainly over Newfoundland and the Maritimes. The results resembled earlier "minimalist" reconstructions based on terrestrial geomorphic evidence, for example by De Geer (1892, p. 473) and Grant (1977). Testing earlier speculations by Swift and Borns (1967), and by Grant (1970) as to the history of changing tides, Scott and Greenberg (1983) modelled Holocene tidal ranges using paleodepths for the entire basin (shallower near the mouth; deeper near the head) and found that amplification had indeed occurred, but probably mostly between 7000 and 4000 y BP. Thus the power of the computer was coupled with the geological value of paleoshores, and first applied in Bay of Fundy region to shed light on the most important problems of Late Quaternary environmental evolution, namely the changing distribution of land, sea and glaciers.

This brief visit to the Bay of Fundy coast introduces the visitor to some of its geological highlights, and illustrates how these have been variously interpreted and used to reconstruct land/sea relations. Only generalized notes are given for each stop, not all of which may be visited because of uncertain scheduling and weather. The features will speak for themselves; much additional commentary is expected to come from the participants!

Acknowledgments

The author is grateful to R.B. Taylor, Atlantic Geoscience Centre, for the opportunity to propose this excursion. To the other prospective co-leaders who were unable to participate, the author acknowledges R.R. Stea, Nova Scotia Department of Mines and Energy, for sharing his ideas on the glacial history, and D.M. Wightman, Alberta Research Council, for discussions of the sedimentology. R.W. Dalrymple and B.A. Zaitlin, Queen's University, kindly contributed the explanation of Bay of Fundy intertidal bedforms and current structures. At the Geological Survey of Canada, J.S. Scott approved the necessary resources, R.J. Mott assisted with the palynological aspects, V.K. Prest contributed his unpublished observations and J. Grainger and M. Slegr processed the words.

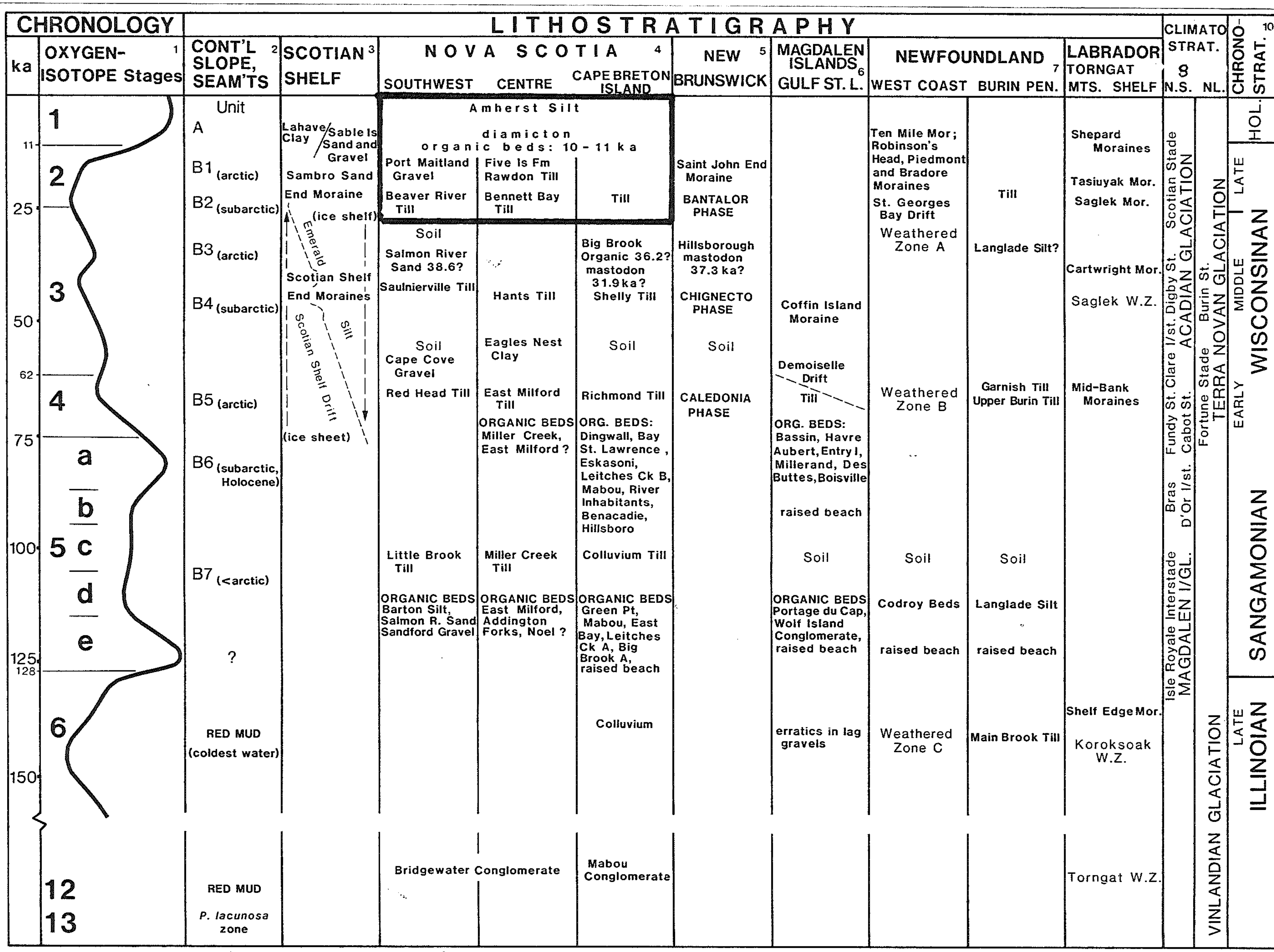

Fig. 8 Regional correlation of Quaternary stratigraphic units (from Grant and King, 1984).

DETAILED DESCRIPTION OF ROUTE AND STOPS

Halifax-Truro

On the first leg, Highway 102 crosses the province from Atlantic Ocean to Bay of Fundy. This area has yielded definitive stratigraphic information which Stea (1982) has used to erect a provisional depositional sequence for the central

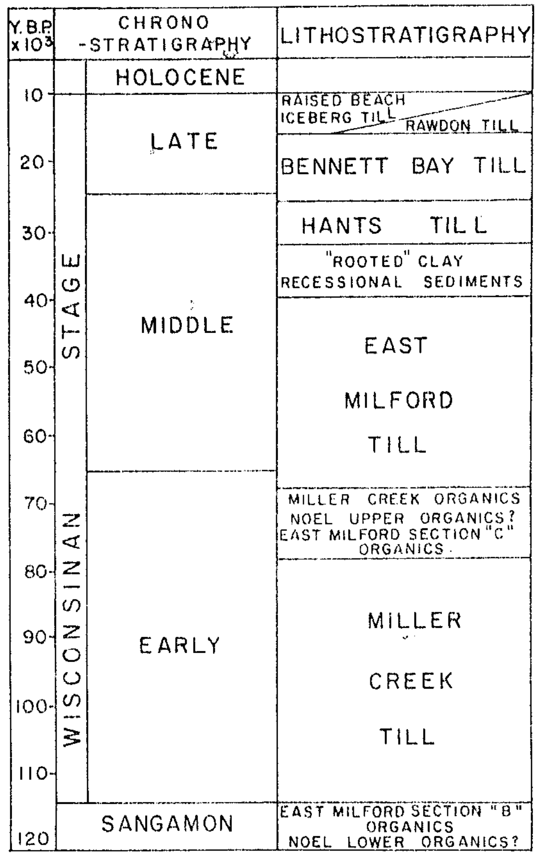

Fig. 7 Lithostratigraphic column for central Nova Scotia (from Stea, 1982).

part of the province (Fig. 7). It is part of the basis for a regional (stratigraphic scheme (Figure 8) (Grant and King, 1984). The sequence features a till from oxygen isotope Stage 6, several Sangamonian (Stage 5) organic beds, and an overlying stack of till sheets representing separate pulses of different ice-dispersal centres during the Wisconsinan. Figure 9 (R.R. Stea, in press) illustrates schematically how the tills and intercalated nonglacial beds relate laterally and vertically from north to south across central Nova Scotia.

A feature of the Atlantic coastal area, as may be seen in road cuts between Dartmouth and the international airport, is a bright red clayey drift, the "Lawrencetown Till". Ranging in thickness from huge drumlins to a thin veneer, its fine texture, bright colour and content of stones derived from northern sources, notably Cobequid Mountain, contrasts with the underlying greywacke and black slate, and demonstrates a powerful southward ice flow across the province. It represents wholesale transport of red material from Carboniferous terranes as

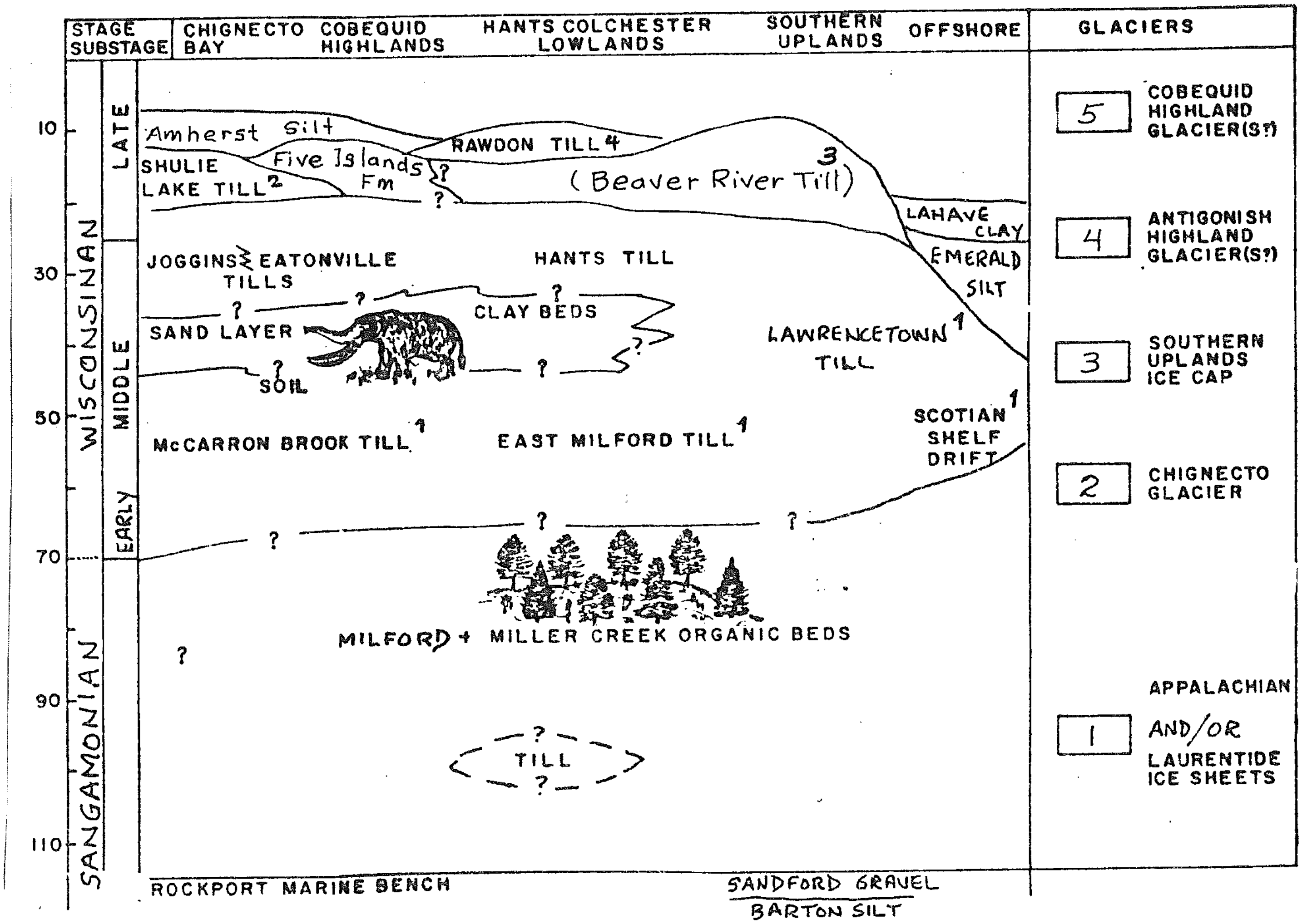

Fig. 9 Schematic time-space diagram of Quaternary sedimentary units across Nova Scotia from Northumberland Strait to Scotian Shelf (after Stea, in press).

much as 100 km distant. This flow occurred early in the Wisconsinan Stage (as well as later) and may have reached the edge of Scotian Shelf, although data is lacking offshore.

Four glacial advances in central Nova Scotia (Fig. 10) are recorded by cross cutting striations and by stacked dissimilar tills (Stea and Finck, 1984). The last phase involved two ice masses: one retreating northward to a centre offshore in Gulf of St. Lawrence; one retreating eastward to a source in central Nova Scotia. The latter evidently cleared the north shore of Minas Basin coast first in order that the sea could invade the depressed basin, and northern ice could debouch outwash freely into it.

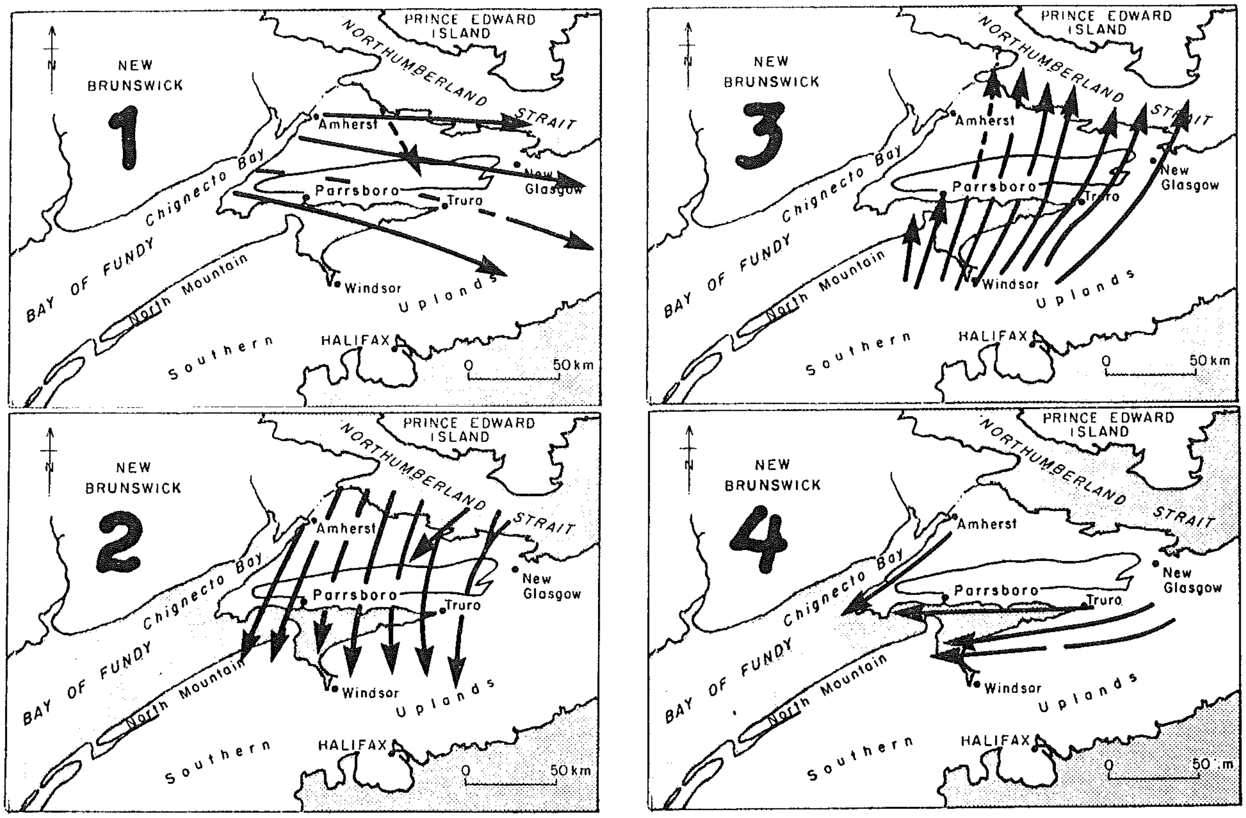

Fig. 10 Sequence of glacier movements in north-central Nova Scotia. Flow 1 may be Early Wisconsinan; Flows 3 and 4 are probably Late Wisconsinan. (from Stea and Finck, 1984).

Truro to Glenholme

STOP 1

DEBERT military base is the site of an ancient Indian encampment discovered in 1951 during construction excavations in a sand plain. More than 4500 stone tools were recovered. The site was used for many decades by hunters of woodland caribou, then suddenly abandoned perhaps because of a shifted migration route. Thirteen charcoal samples passing chi-square test have a mean age of 10,903 ± 48 y BP (Stuckenrath, 1966). The site therefore represents the **earliest evidence of postglacial man in eastern North America.** Belonging to the Maritime archaic culture, but informally referred to as paleo-Indian, the remains are assignable to the Llano Complex which is characterised by Clovis-type fluted points (Byers, 1966). As to the paleogeography, Borns (1966) believed that the site was occupied while the climate was periglacial: permafrost was present in the immediate area and an ice cap persisted in central Nova Scotia. While these contentions remain unsupported by direct evidence, it is interesting that Mott (1985) reports a sharp cooling 11-10 ka BP which he correlates with the glacial-degree shift of the oceanic Polar Front ca. 10.4-10.2 ka BP detected by Ruddiman and McIntyre (1973). Perhaps the site was abandoned because of the intense cold.

STOP 2

Along the **Masstown shore road,** the broad dykelands (reclaimed salt marsh meadows) overlap the rolling till surface. One can visualise how the tidal muds can gently bury and thus preserve soils and vegetation. At Lyon Head a tree stump under marsh at -1.8 m dated 1210 ± 140 y BP (GSC-973). The **Debert River** is

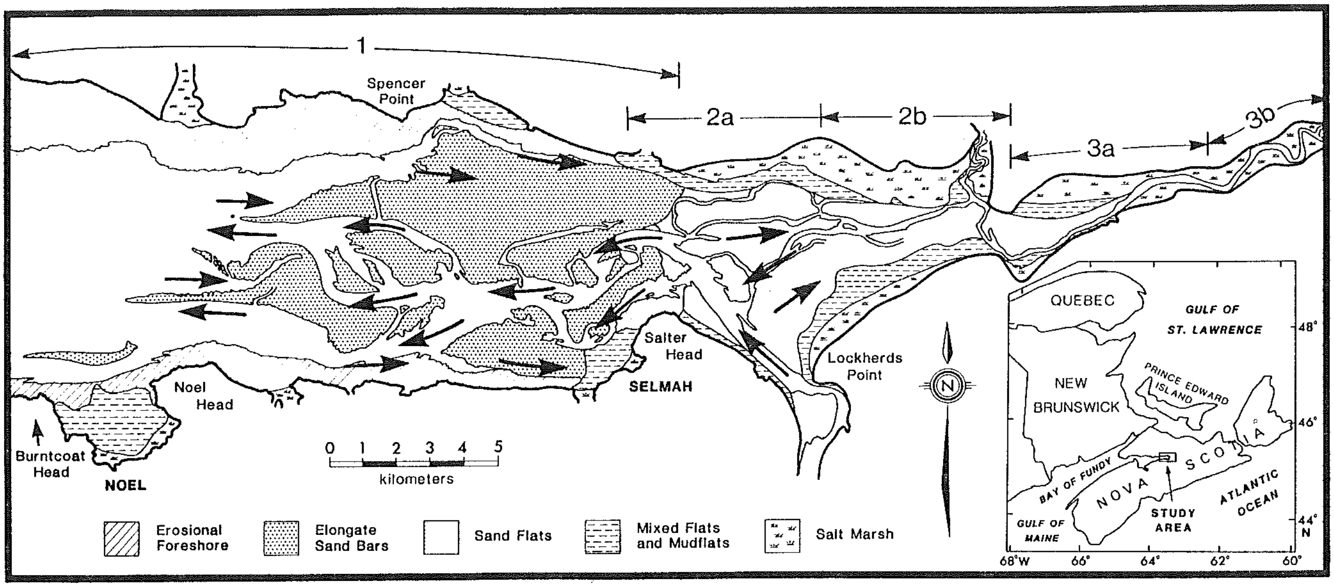

Figure 11a Location map of the study area (inset), and facies zonation within the Cobequid Bay-Salmon River Estuary. Truro is located along the south margin of zone 3b. The arrows give the residual sediment-transport directions.

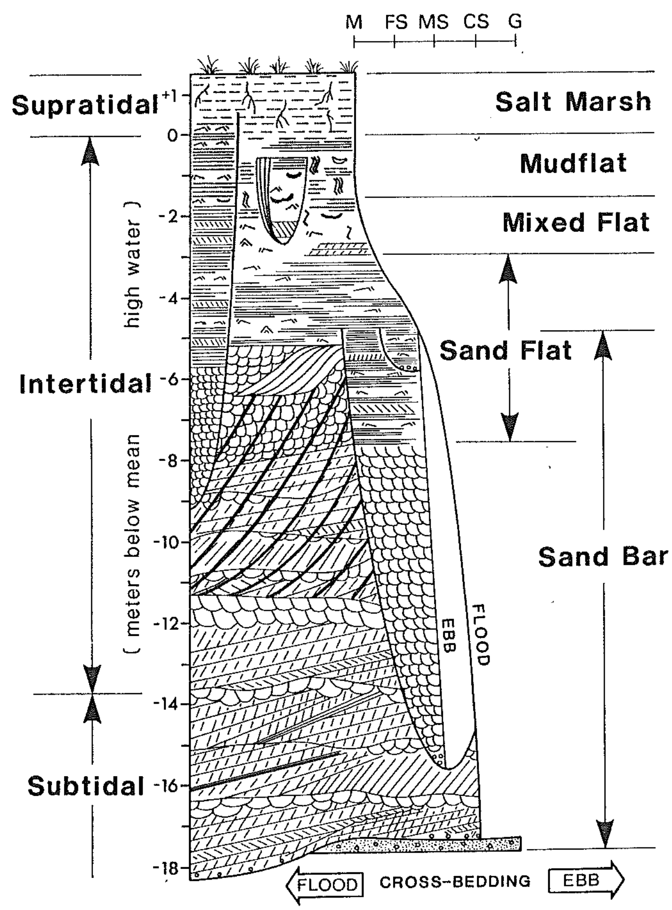

Figure 11b Composite, idealized, progradational/aggradational facies model for the sandy, macrotidal, Cobequid Bay-Salmon River Estuary. See Appendix 1 for explanation. Grain sizes are indicated by the horizontal scale at the top.

Fig. 11 Composite idealised facies model of sediments in the macrotidal Cobequid Bay estuary, with map of bedform types (R. Dalrymple and B. Zaitlin, pers. comm., 1985).

cut in a rock gorge that is backfilled by tidal marsh; this situation indicates baselevel was once at a lower level and has since risen to its present position.

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Leaving Route 4 in **Glenholme;** a lane leads to the shore. On the left (east) salt marsh fills a rock gorge estuary that is flanked by gravel terrace which here is the distal part of a outwash fan built by Folly River when a glacial lobe was melting in Wentworth valley. Kames and kettles (such as Mackay Lake) locally interrupt the graded gravel surface, indicating that dead ice was buried in the alluvium. A pit shows planar-bedded gravel overlain by crossbedded, scour-and-fill gravel.

STOP 3

From a vantage point on the shore at the **migratory bird display,** it may be possible depending on the tide to see huge sand waves and other current bedforms ornamenting the sea floor. These spectacular features and the easy access have made the Fundy tide flats a natural laboratory for the study of current structures produced by a range of velocities operating on different textures. For the record, R.W. Dalrymple and B.A. Zaitlin have contributed a sedimentological analysis (Appendix 1) of the bedform sequence shown in Figure 11.

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STOP 4

East of the hamlet of **Highland Village** a low bank of salt-marsh peat near high tide rests on a thin humus and peat layer over till (Fig. 12). The ages of two tree stumps in the submerged terrestrial horizon are believed to date the first arrival of salt water, and thus provide control on the rise of tide level in upper Minas Basin.

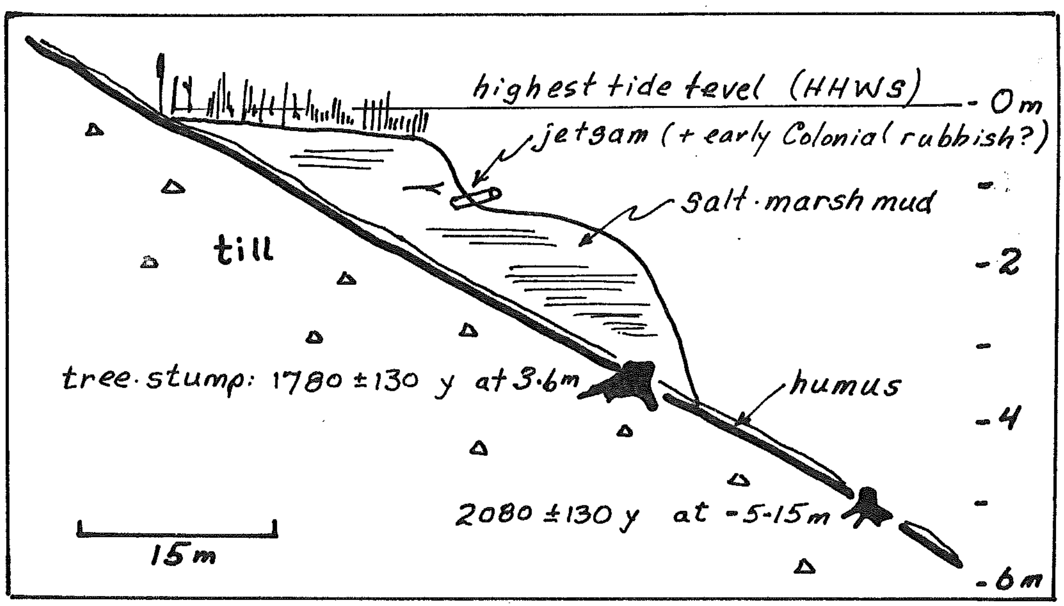

Fig. 12 Submerged forest beneath tidal marsh mud, east of Highland Village (after Grant, 1970, p. 90).

At the mouth of **Portapique River,** another large outwash fan (Fig. 13) about 6 km wide represents major meltwater discharge from northward retreating ice mass. It seems to have been built partly during ice-marginal stands within the deposit and during later recession through the drainage basin.

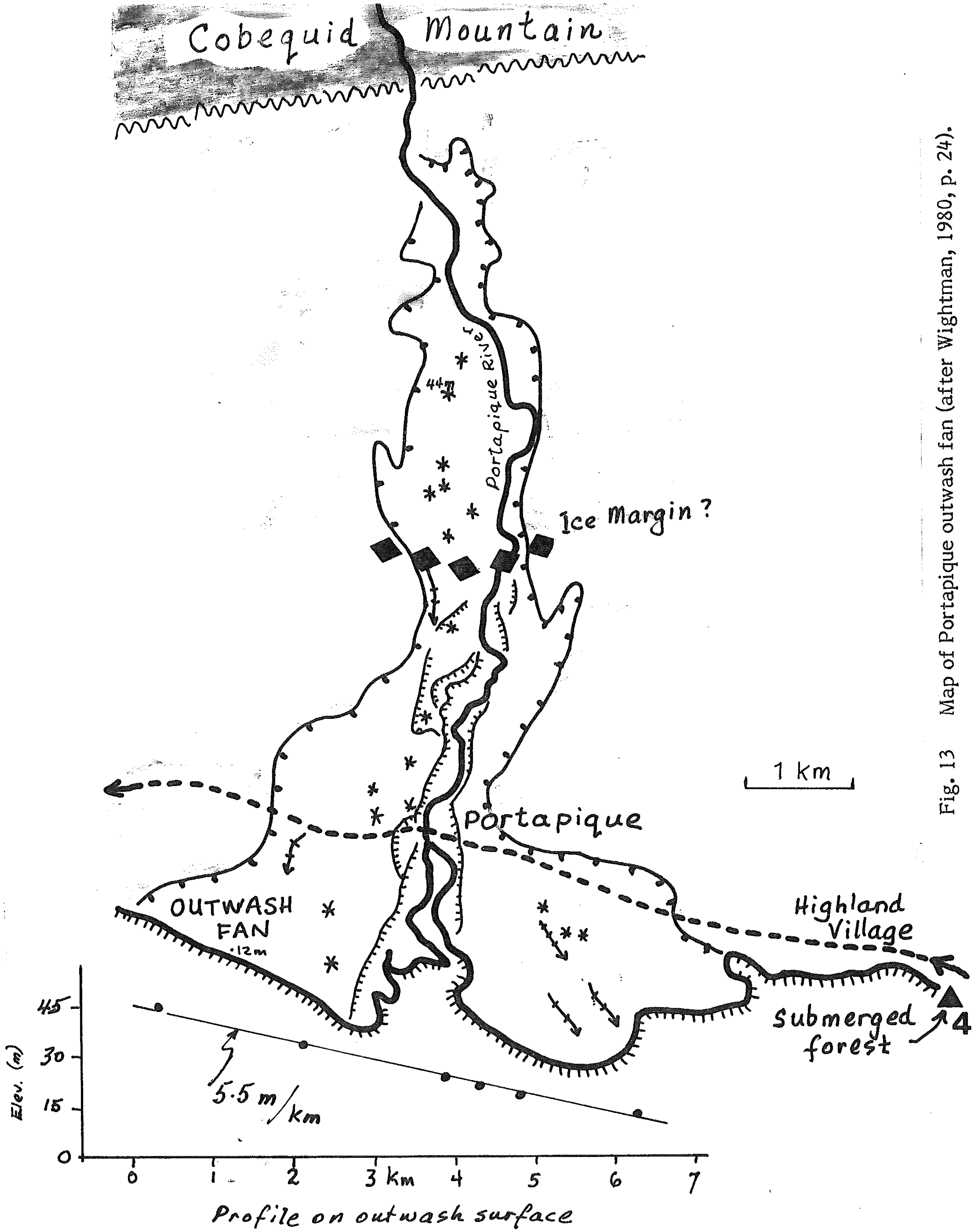

Fig. 13 Map of Portapique outwash fan (after Wightman, 1980, p. 24).

SAINTS' REST is a good place to begin the debate about changing base levels. Here, Swift and Borns (1967, p. 708) recognized both deltaic and fluvial lithosomes; the contact was in the beach face just below high tide level. Hence they chose this as **the type locality for the glaciofluvial Saints' Rest Member of the Five Islands Formation.** For reasons given later, this is the easternmost exposure of deglacial marine sediments. What is seen now along the shore, is a distal remnant of an extensive outwash terrace of lateglacial Bass River (Fig. 14). The

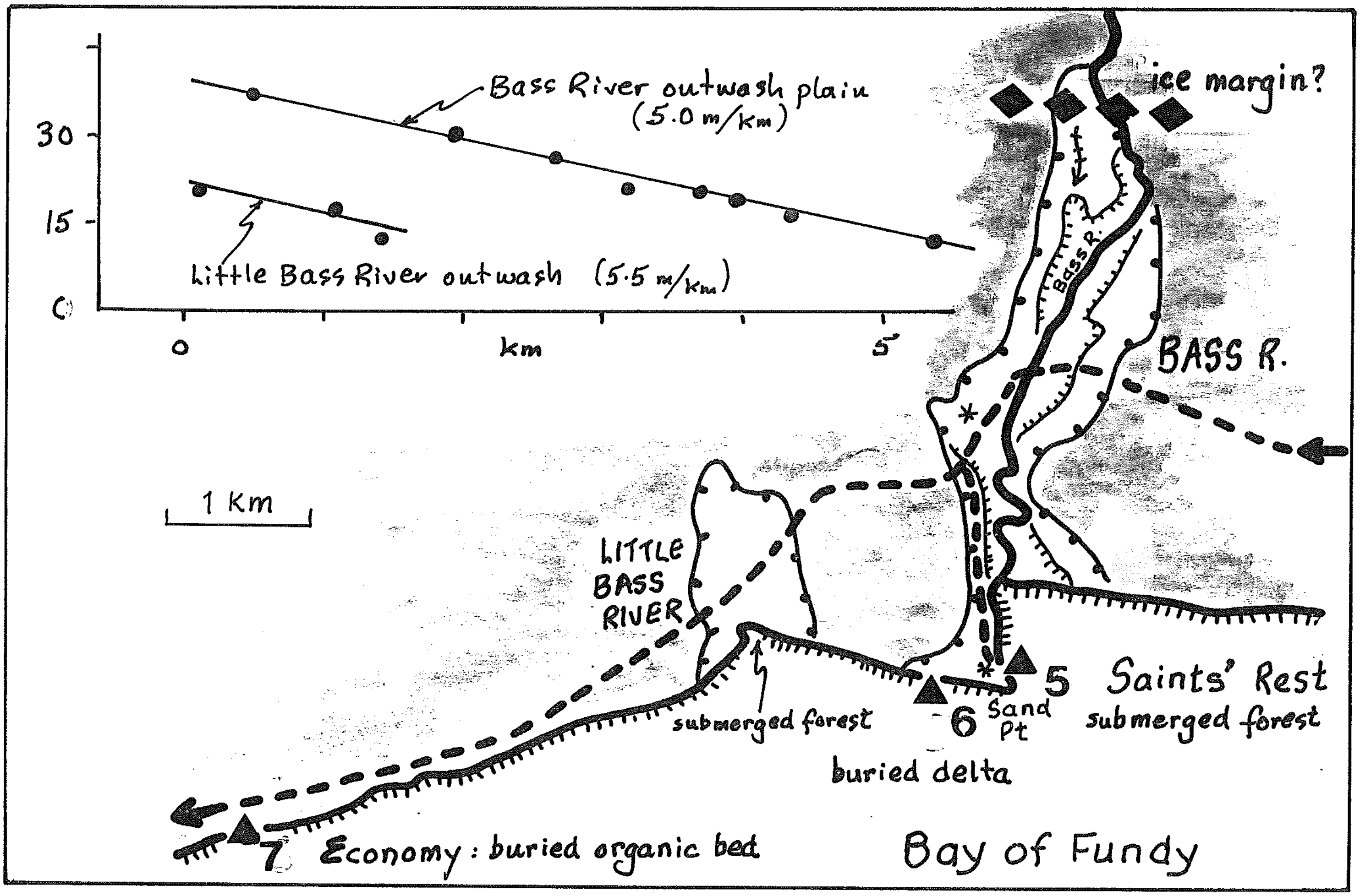

Fig. 14 Map of outwash fan-deltas at Bass River, showing gradient of fluvial surface (after Wightman, 1980, p. 45-57).

bluffs north and west of Sand Point expose gravels of the outwash terrace which slopes inland at 5 m/km like the Portapique terrace. The gravel is closedwork (sand matrix), horizontally stratified, with imbrication of shingle pointing seaward. Wightman (1980, p. 51) interprets the sediments here and at nearby Little Bass River as braided-stream deposits, with tabular crossbeds indicating migration of transverse bars and infilling of steep-sided channels during fluctuating discharge. Kettles, as near the lighthouse, attest to burial of glacier ice, but whether it belonged to the northern or to the eastern glacier is unknown.

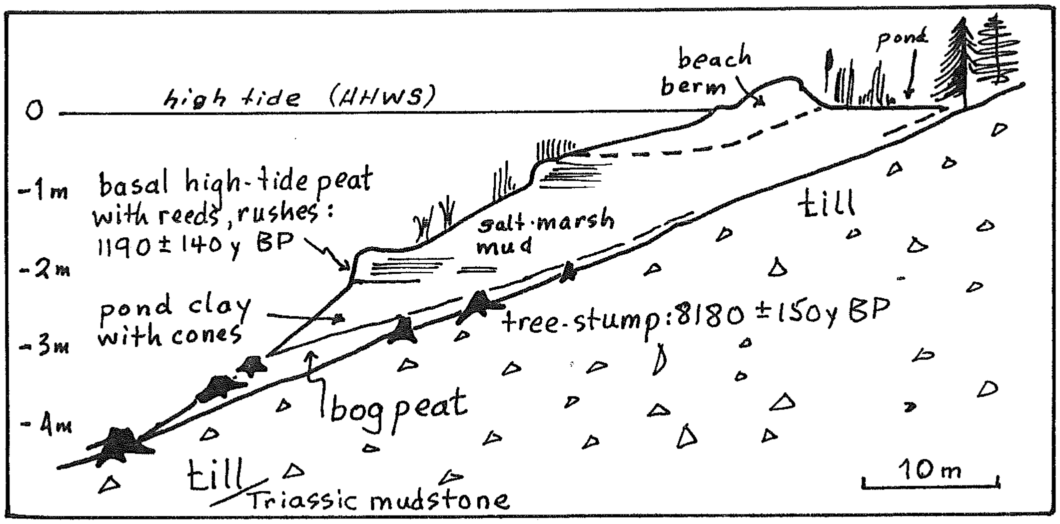

Fig. 15 Diagrammatic geological cross-section of submerged forest and bog peat beneath tidal marsh at Saints' Rest (after Grant, 1970, p. 91)

At its foot **in the beach face is an outcrop of organic beds** resting on till. (Fig. 15). Rooted in the till are large tree stumps one of which at -2.4 m dated 8180 ± 150 (GSC-757) showing that deglacial regression had lowered relative sea level below that position before then. The forest layer is overlain by freshwater bog peat and grey pond clay containing pine cones, suggesting that rising groundwater table inundated the forest. The cause is evidently linked to rising tide level for the freshwater beds grade upward to salt marsh peat, the basal layers of which contain the distinctive black corms of the brackish-water bulrush **Scirpus** and the fine hairy roots of the high tide grass **Spartina patens.** At 3 m below high tide, these yielded a date of 1260 ± 140 y BP (GSC-922). The interval between the dates is a large erosional hiatus which illustrates the hazard of indiscriminate sampling of basal terrestrial organics.

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STOP 7

In **Upper Economy,** the rapidly eroding sea cliff exposes an organic bed buried by a thin layer of gravel believed to be fluvial. The peat dates 10,100 ± 130 y BP (GSC-3963) (R.R. Stea, pers. comm., 1985). Palynologically the peat is dissimilar to other Late Wisconsinan buried organic deposits (R.J. Mott, pers. comm., 1985). Together with the younger age and the cover of fluvial material, it is unlikely that the occurrence signifies the same climatic determination which Mott (1985) documented for the period 11-10 ka BP.

STOP 8

The mouth of **ECONOMY RIVER** cuts into another large outwash fan (Fig. 16) which has scattered kettles and scour channels, and a gently conical slope in three directions. The surface gravel is clearly fluvial but supposed delta foreset beds of sand have been seen up to elevation 5 m AHT in the cutbank of a meander along the Economy Point road, in Morrisons gravel pit up to elevation 10 m ASL in Economy village (Wightman, 1980, p. 60) and in the intertidal zone along the shore westward to Carr's Brook (Swift and Borns, 1967, p.708). If the beds are indeed deltaic, their elevation would accord with the general position of marine limit projected into this area from more definitive features farther west (Fig. 4). Notwithstanding the general agreement, the fluvial/deltaic contacts at this locality clearly demonstrate the common problem of determining the inner and upper reach of the sea ("marine limit") using the tops of delta foreset beds where they have been truncated, dissected and buried by later fluviatile sediment. A better criterion, one that is in common use elsewhere, is the limit of beaches and wave trimlines on till surfaces on hills beyond the reach of fluvial sedimentation. Such features have not been systematically sought on this coast except for one spot check on Gerrish (Economy) Mountain where gravel lag on till pinches out at 16 m AHT. This compares with a marine limit determination of 9 m by Wightman (1980, p. 345). ********************

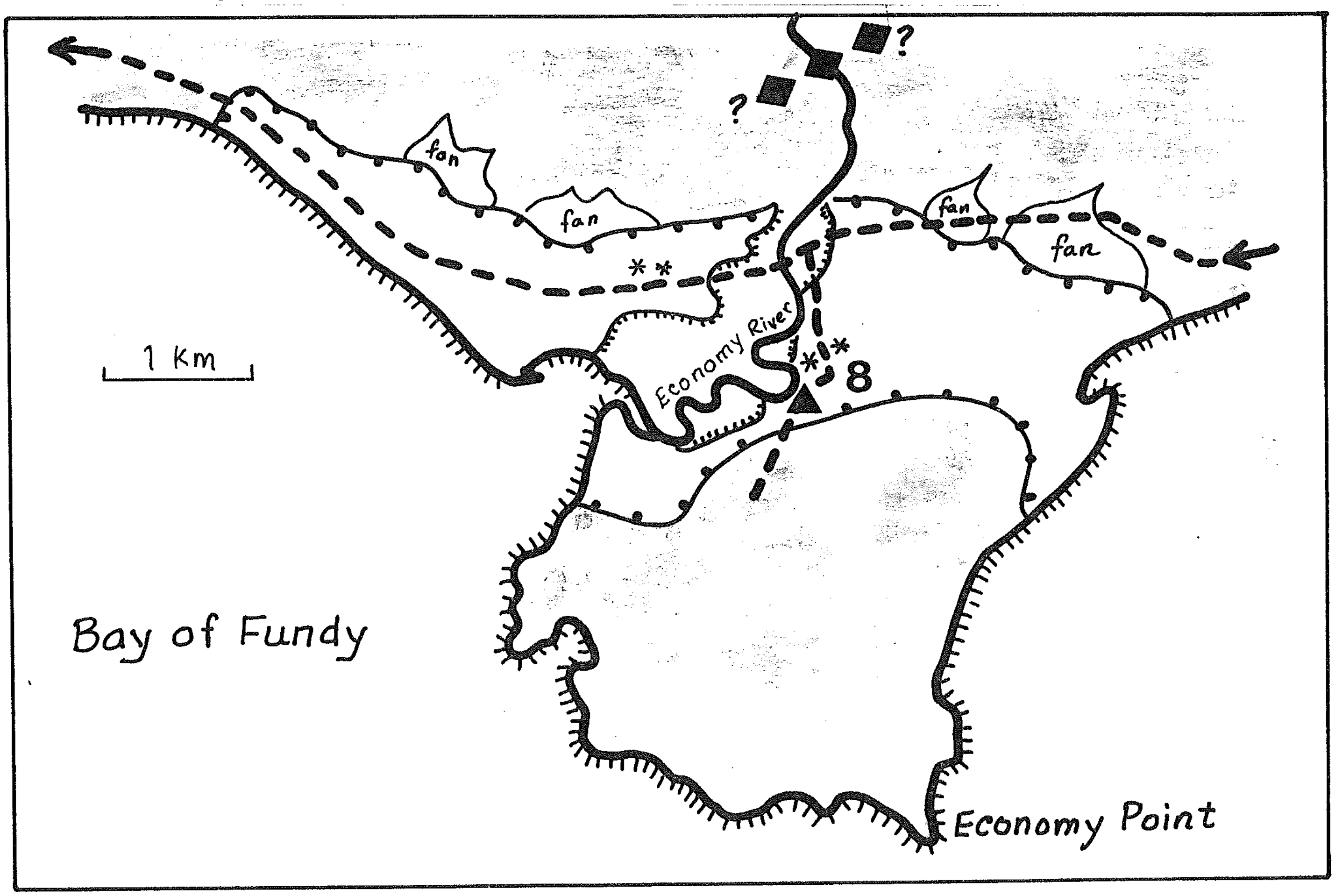

Fig. 16 Map of outwash fan-delta at Economy (after Wightman, 1980, p. 59).

FIVE ISLANDS area (Fig. 17) contains **the type section of the Minas Basin fluviomarine terrace,** as defined by Swift and Borns (1967). The exposures provide the definitive point of reference for assessing the origin and paleogeographic significance of the entire terrace complex.

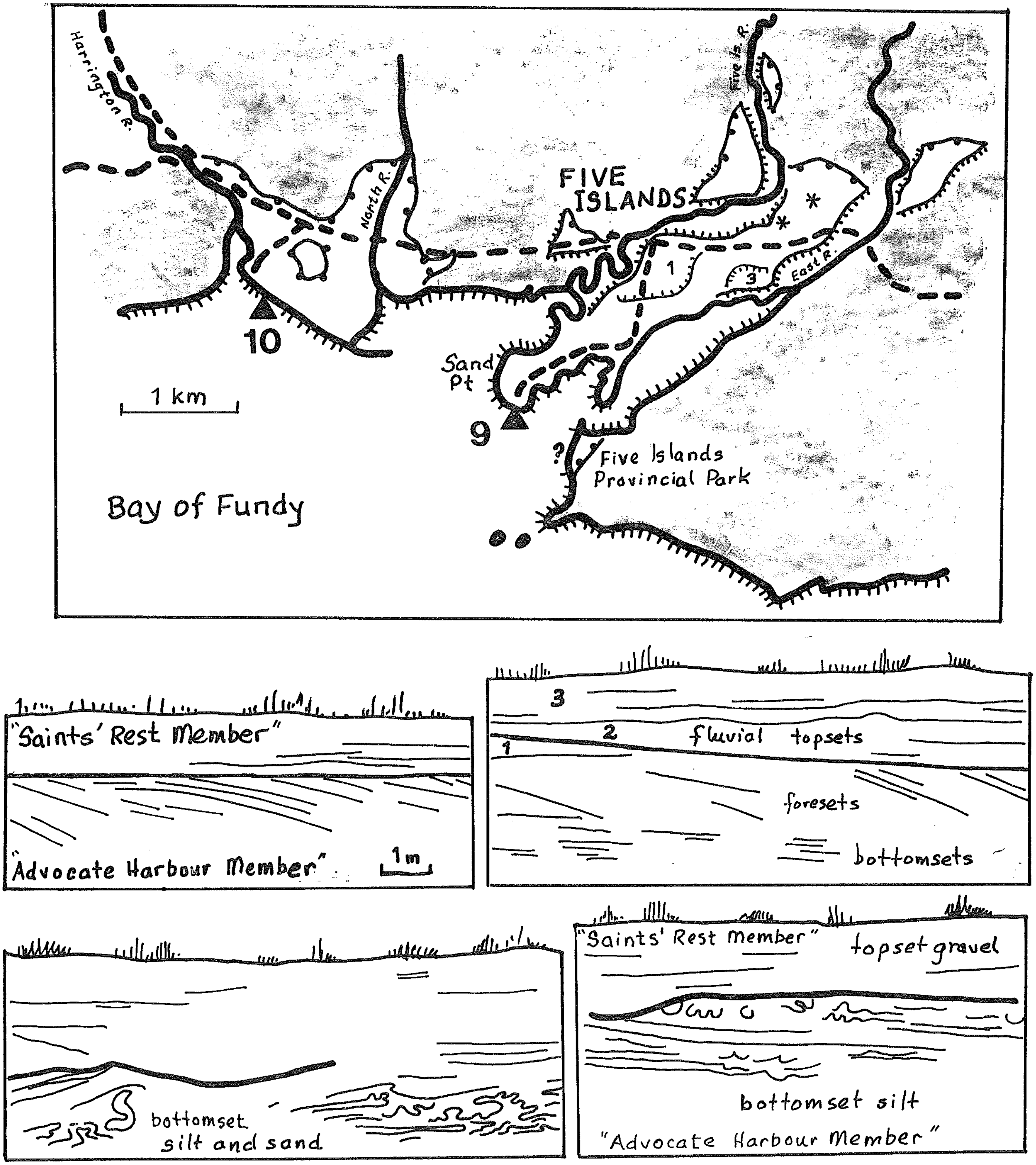

Fig. 17 Map of outwash fan-deltas at Five Islands with photologs of exposures at type locality (STOP 10) (after Wightman, 1980, p. 64, 73, 88, 91).

STOP 9

Sand Point is the terminus of a large gravel outwash plain that slopes seaward at 6.0 m/km from Five Islands Village. Fluvial topsets overlie foreset beds inclined 27°SW.

STOP 10

The main exposure and type section, however, is between the mouths of Harrington and North Rivers (Fig. 17) where it comprises a lower unit of reddish-brown sand/clay rhythmites, at one spot overlying, chaotic-bedded (ice-contact?) boulder gravel. This mud (the Advocate Harbour Member) is contorted near the top by sediment loading, and rises to 7 m AHT. Molds and casts of **Portlandia arctica** were reported by Swift and Borns (1967, p. 695). This sediment is therefore judged to be marine. It grades upward into gravelly foreset beds inclined 20°-30° seaward. Thus, a deltaic couplet is represented. The delta beds are truncated by a few metres of fluvial gravel that forms an outwash plain sloping 6.5 m/km seaward. The outwash contains more Cobequid Mountain crystalline rocks toward the top - a condition which is explained by Swift and Borns (1967) and by Wightman (1980, p. 76) as the result of ice recession northward across the upland. Although this aspect may have other explanations, the essential deltaic origin of the basal sediments seems clear. The minimum elevation of marine limit as shown by the height of the truncated topset beds, is 15 m ASL (8 m AHT) although the upper limit of marine action was probably several metres higher.

Two large sediment-filled structures cut the top and foreset beds. Stratification is vertical and there is no deformation of the adjacent beds. The features are therefore probably not ice-wedge casts as Swift and Borns (1967) supposed but vents or pipes produced by dewatering of the underlying underconsolidated silt beds during emplacement of the superincumbent gravel load.

****

PARRSBORO area presents **the largest outwash fan-delta** (Fig. 18) and is the most illustrative of the ice-marginal setting for the Minas Basin fluviomarine terrace. In essence, there is a narrow pass through Cobequid Mountain in which a lobe of northern ice lodged during the last deglaciation, built small end moraines, large kame fields, and an extensive outwash fan when sea level was at lesta 16 m above present tide level (22 m ASL). Subsequent regression and melting of buried ice blocks has dissected and pitted the terrace. Ponds in the depressions (as well as offshore sediments) yield minimum dates on the beginning of deglaciation; the terrace may be 14-15 ka old.

STOP 11

Leaving Highway 2 about 5 km NE of Parrsboro, and proceeding along a branch road at the foot of Cobequid faultline scarp, an exposure **(KERNAHAN'S PIT)** situated north of East Parrsboro River, is cut into the summit outwash plain at elevation 30 m ASL (=24 m AHT) and shows that at least the top 6 m of gravel is

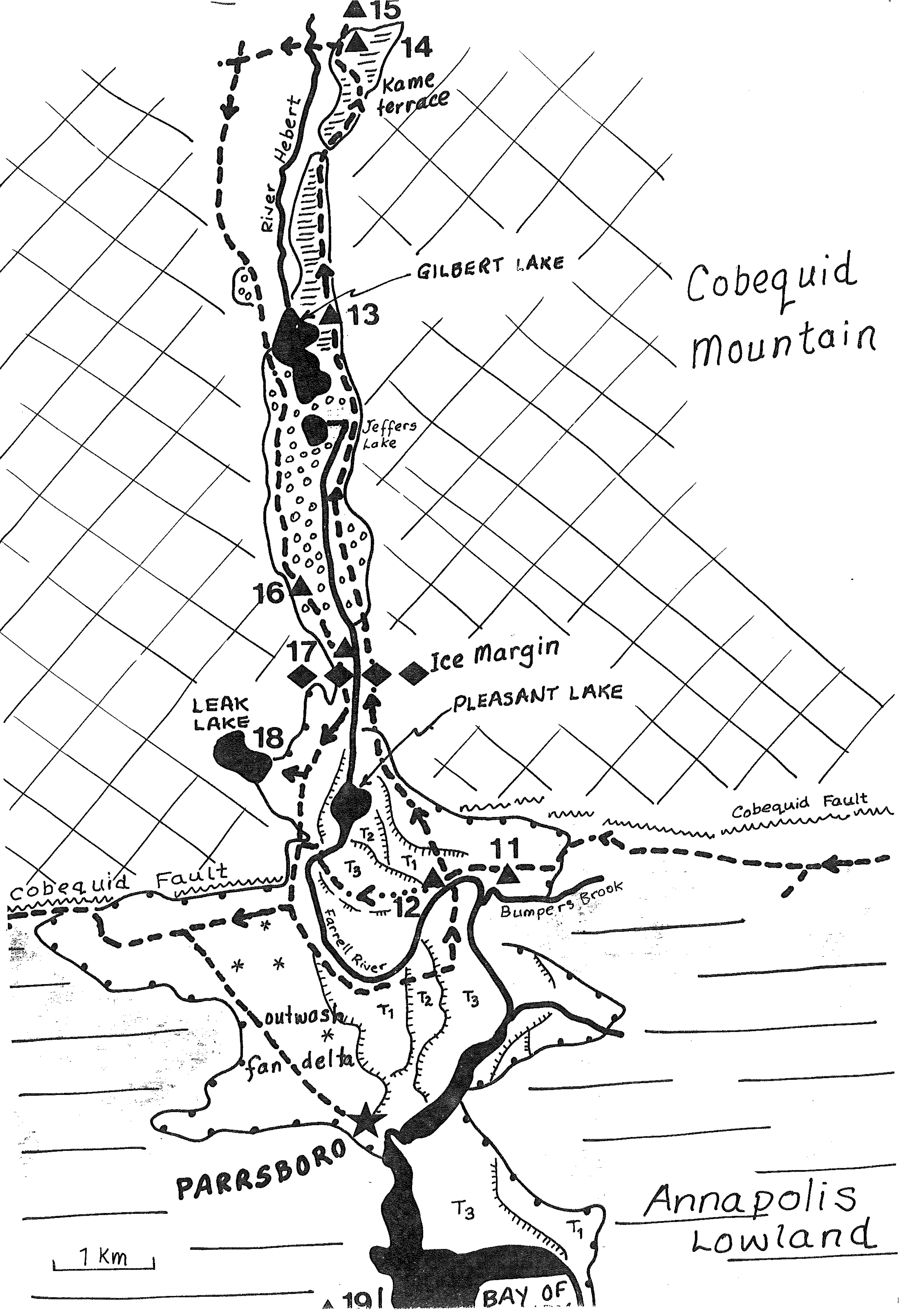

Fig. 18 Map of outwash fan-delta at Parrsboro including ice-marginal zone of end moraines and kame terraces, and gradients of outwash surface and three later terraces (after Wightman, 1980, Map D).

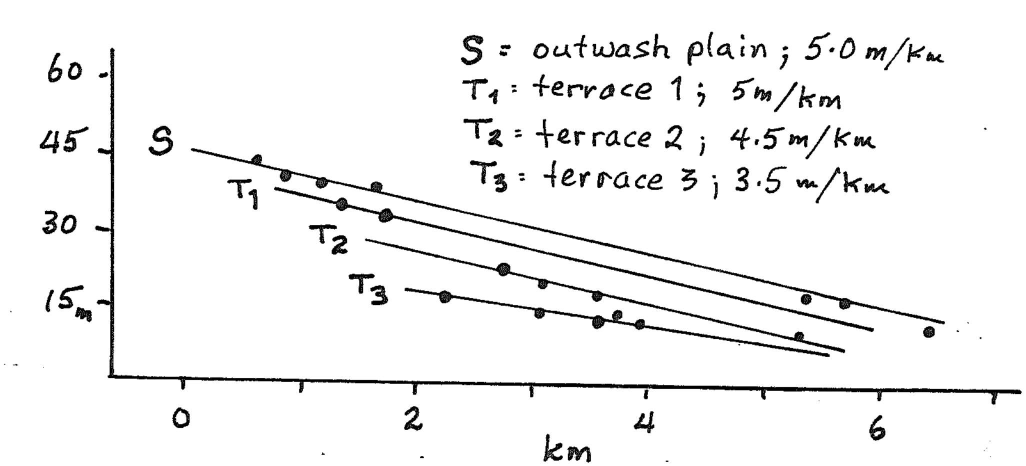

Fig. 21 Gradient of alluvial surfaces, Parrsboro outwash fan-delta

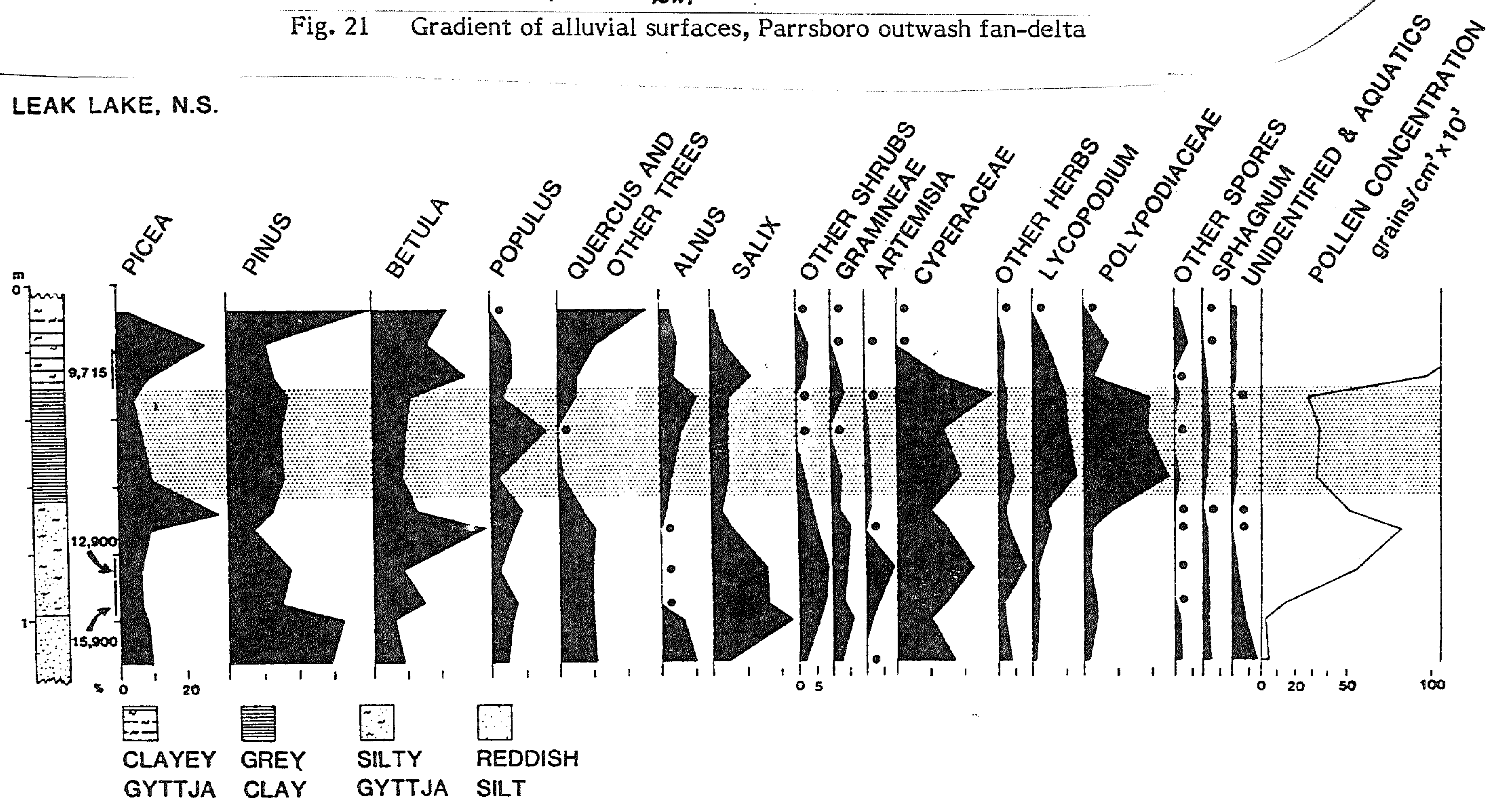

Fig. 19 Pollen diagram for basal part of Leak Lake sediment core, showing early shrub -tundra phase, first spruce maximum, reversion to open shrub and herb vegetation ca. 11-10 ka BP (from Mott, 1985).

fluvial. One kilometre west, beyond the road junction is Durants Pit **(STOP 12)** where fluvial topsets truncate delta foreset beds at 22 m ASL. This gives a minimum estimate for local marine limit (cf. Fig. 4).

****

Going northward from here past **Pleasant Lake** (a kettle in the modern floodplain!) the road gradually ascends the fan surface to 44-46 m ASL where it ends abruptly at a complex of ice-contact gravel knolls and small moraines which plugs Parrsboro valley for the next 5 km.

STOP 13

On a knoll east of **GILBERT LAKE,** a large area of morainic ice-disintegration topography provides a good spot to consider the ice mass which was responsible for the Parrsboro 'fan-delta'.

In addition, **lake-sediment palynology** from Gilbert and Leak Lake kettles can be discused in terms of the record of postglacial climatic evolution (Wightman, 1980; R.J. Mott unpublished data). The sequence (Fig. 19) begins with herbaceous tundra evolving into a spruce woodland that is interrupted by a climatic deterioration ca. 11-10 ka. The latter event is widely represented in the region; it is correlated with the Younger Dryas cooling in Europe, and attributed to a southward shift of the Polar Front (Mott, 1985). Thereafter spruce forest resumed and evolved in a normal way into the more thermophilous modern forest. A basal date of 15 900 ± 1200 y BP (GSC-2880) is rejected because of possible old carbon (coal) contamination (R.J. Mott, pers. comm.), but it may still be a valid minimum for deglaciation here. Indeed Amos (1978) reports dates of 14 180 ± 710 (GX-4514) and greater than 37 000 y (GX-4522) from glacial marine sediment under Holocene mud in Minas Basin. If reliable, these dates suggest that glaciers may have terminated north of Parrsboro during Late Wisconsinan time as depicted by Grant (1977).

With regard to postglacial time, the current belief is that the Parrsboro lake record is sufficiently similar to other well documented postglacial pollen sequences, which are considered to be more reliably dated, that the beginning of postglacial sedimentation in the kettle is placed between 12 and 13 ka BP. However, to deduce when the glacier actually departed from the morainal zone depends on how long it took for the lakes to come into existence by melting of the buried ice, and how long a lag there was before sufficiently rich organic material could accumulate. Ice retreat in the Parrsboro area is thus poorly constrained by chronometry, but an AMS date (in progress) on mollusc periostraca in basal delta beds at nearby Spencer's Island (see below) will soon provide the first direct indication.

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Optional Diversion Northward along River Hebert

From the east side of Gilbert Lake the road travels over a large kame terrace at elevation 62 m ASL which appears coplanar with other kame terraces in the morainal zone, and to be the up-valley extension of the graded fluvial surface represented by the fan. If so, meltwater flowed over buried ice blocks from Newville Lake south to Pleasant Lake. At present, the drainage divide is the small moraine which dams Gilbert Lake.

STOP 14

At the crossroads southeast of Newville Lake is a large kame terrace with deltaic foreset beds reaching 30 m. Wightman (1980, p. 287) referred the deposit to local deposition a small ice-marginal pond, but it is equally possible that the kame formed a continuous surface across the Gilbert Lake divide, and thence to the lower degradational surface (Terrace 3) in the dissected Parrsboro fan-delta.

This is a good place to consider the continuity of the glacier margin when the Parrsboro ice-contact deposits were formed. That ice of northern derivation moved onto and over most of Cobequid Mountain is shown by ice-flow features, till lithology and the great River Hebert esker (Stea, 1983). The crucial question is whether ice crossed Minas Basin to merge with South Mountain ice cap. For this area, Grant (1977) depicted the two as separate ice masses. He linked the Parrsboro ice-marginal deposits to other large and small moraines, kames and meltwater chanels along the north slope of Cobequid Mountain west to Apple River on Chignecto Bay. Because of great differences in terrain maturity and marine limits north and south of this morainal zone, he postulated that it marked the Late Wisconsinan stadial limit of northern ice. That hypothesis is not shared by Wightman (1980) or Stea (1983) who feel that Late Wisconsinan ice flowed completely across Cobequid Mountain. Pedologic and palynologic studies are underway to provide data on the extent and age of glacial ice.

About postglacial events there is more agreement. During final ice disintegration, Newville valley held a small postglacial lake ("Potter Lake"; Wightman 1980, p. 294).

STOP 15

On the old lake bottom, a cutbank of West Brook exposes outwash gravel covered by peat which at its base dates 13 365 ± 420 y BP (DAL-300) (Fig. 20). Pollen indicates open or sparsely treed tundra (R.J. Mott, GSC Palynological Report 78-11). Above this, the spruce pollen maximum dates 9830 ± 100 y BP (GSC-2772). (The older date is considered too old by most authors because of possible contamination by Carboniferous coal despite the fact that the sample was not small and no coal was confirmed. The peat is covered by greenish clay above which is wood debris (3745 ± 120 y BP (DAL-265)) covered by alluvium. Burial of the vegetated surface by fluvial aggradation is taken to indicate rising baselevel as a result of relative sea-level rise in Bay of Fundy (Wightman, 1980).

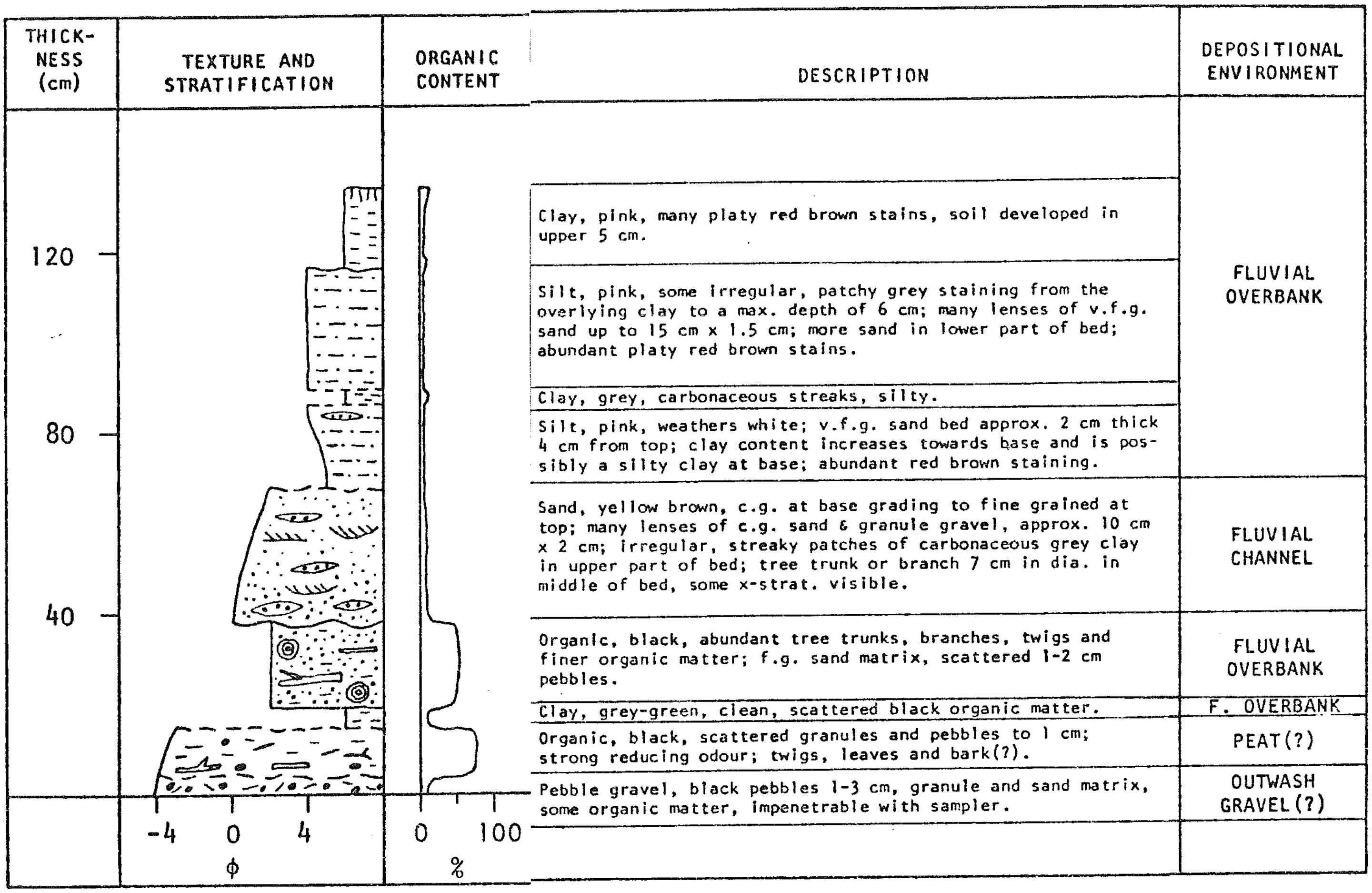

Fig. 20 Exposure of peat buried by alluvium (from Wightman, 1980, p. 299)

Crossing to the west valley wall the excursion route rejoins Highway 2 which undulates over several broad moraines and follows a nest of esker segments.

STOP 16

Two kilometres south of Gilbert Lake an exposure shows the internal constitution of one of the typical small, sharp-crested **end moraines:** chaotic bedding, contortion, folding, faulting, thrusting, and a range of textures. Glacial thrusting and overriding of ice-marginal sediment seems evident.

****

STOP 17

One-half kilometre farther south a huge **kame moraine** consists of muddy, poorly stratified gravel that is faulted and deformed by ice thrusting and melting. This ice-contact deposit marks the limit of the active glacier at the time the Parrsboro fluviomarine terrace was aggraded to its maximum level. The outwash fan begins at this point and slopes seaward from 46 m ASL to 12 m at the Fundy shore in Parrsboro Harbour where in two places it overlies reddish-brown clay-silt rhythmites believed to be delta bottomset beds (STOP 19).

****

STOP 18

Leak Lake, occupies a kettle hole near the apex of the outwash fan. Here was taken the controversial sediment core which yielded the spurious dates (Fig. 19) discussed at Stop 13..

****

By meandering along sideroads north of town it is possible to see other kettles and channels on the upper outwash and to descend the **three major regressional terraces** that were cut into it as Parrsboro River was lowered by sea-level fall (Fig. 21). The terraces are not paired and thus have no sea-level significance (Wightman, 1980).

West of Parrsboro fan, the highway follows the foot of **Cobequid faultline scarp,** and glossy slickensided surfaces are common. It is not known whether this fault may have been reactivated by the large differential stresses of postglacial crustal rebound as the ice stood on or behind the mountain. Postglacial displacement is not impossible, and would depend on maximum ice extent and rate of retreat. Repeated failure of Fox River delta (see below) might have been triggered by faulting but certainly no surface expression of movement on Cobequid Fault has ever been detected, such as lineaments crossing the late glacial and postglacial sediments that are bedded across the fault trace. The question remains hypothetical. (In this connection may be noted the 14-metre postglacial displacement on Aspy Fault -- an extension of Cobequid Fault in Cape Breton Island: Grant, 1975). The highway crosses several young alluvial fans and cones at the mouths of gorges that indent the scarp. Note that each fan has been a preferred site for settlement; a farm house sits on each apex.

STOP 20

DILIGENT RIVER are presents another classic ice-frontal fan-delta (Fig. 22). From an apex of 41 m ASL it slopes with a gentle conical form to 21 m at the present Fundy cliffs where it overlies red clay bottomset beds west of the river mouth. The topset/foreset contact is at only 18 m -- surely far below its true maximum level for this location.

Several large kettles indent the fan surface. A line of four depressions trends eastward and may be seen along the backroads through the blueberry fields south of the village. These may mark a buried glacier margin, prior to retreat and stabilisation of the ice front at the fan's apex. In one **kettle pond,** 11 m deep, **(Stop 20)** sediment from the basal 7 cm of 5.5 m of gyttja gave an age of 11 555 $\pm$ 230 y BP (DAL-301) while in a second core the basal 17 cm dated 10 120 $\pm$ 200 y BP (QU-764). The dates are considered reliable because they come from the spruce pollen maximum (R.J. Mott, GSC Palynological Report 79-15). Thus the Diligent River dates lend support to the notion that a Late Wisconsinan ice front stood in this area not more than a few thousand years earlier (allowing time for ice-block melting and accumulation of organic debris on a pervious substrate).

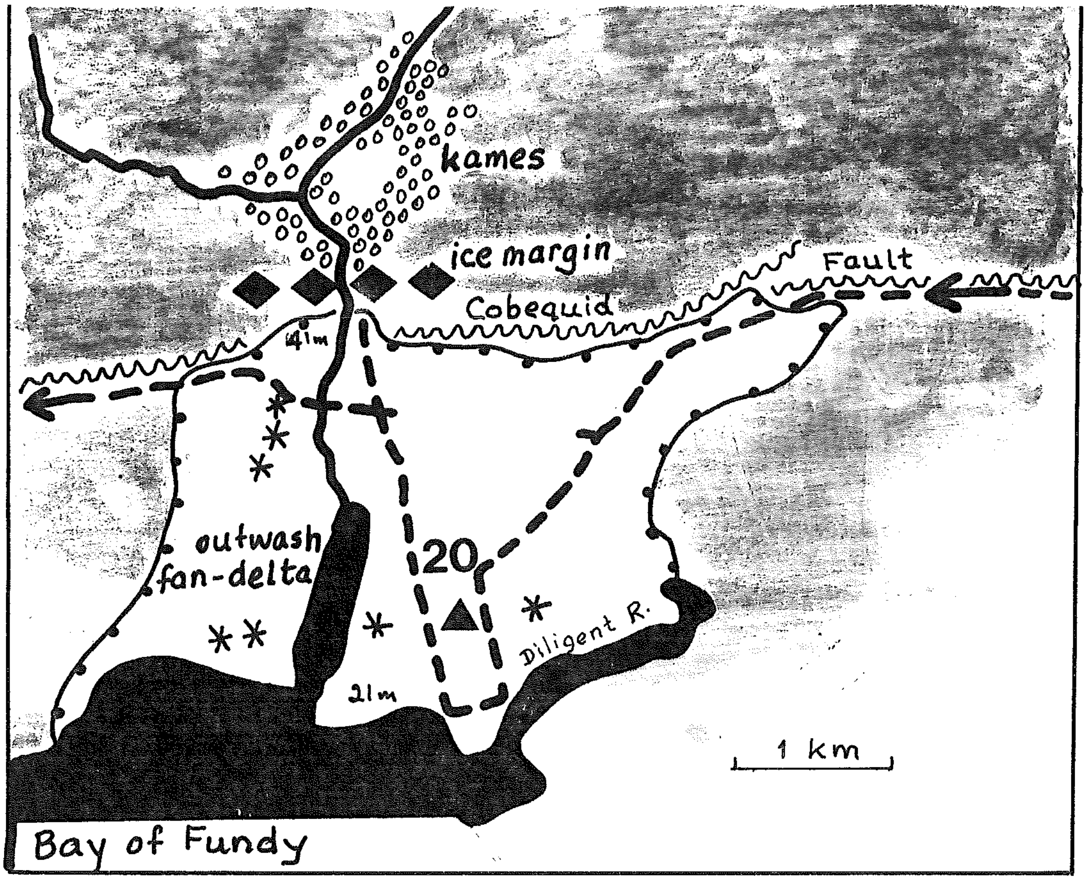

Fig. 22 Map of outwash fan-delta at Diligent River (after Wightman, 1980, p. 135).

The outwash fan ends abruptly at a steep ice-contact face and the valleys inland from these contain gravel hummocks and kame and postglacial terraces that are disturbed by kettles, showing that buried ice persisted after meltwater ceased. These kames show that the fan-delta was built by an ice lobe which lay with its margin localized at the constricted part of the valley, where it breaches the Cobequid fault line scarp.

****

Four kilometres west, **Fox River** presents an identical and even beter ice-contact delta (Fig. 23). From an apex at 34 m ASL, where there is a spectacular 10 m deep kettle, the well-graded conical fan stretches 1.5 km seaward from Cobequid scarp. From that point inland, the valley is almost empty except for patchy ice-contact gravel deposits reaching up to 30-40 m ASL. A stable ice margin of a valley glacier debouching into a higher sea level is again demonstrated. Thus, it may be postulated that, if the northern ice had once extended across Minas Basin and relative sea level was higher during its recession, the ice front might have been calved back out of the deep water area to a stable position along the Cobequid scarp where it could become firmly grounded and anchored in the deep gorges with their narrow mouths. The ice margin would thus follow the high-level paleoshore, as the deposits show.

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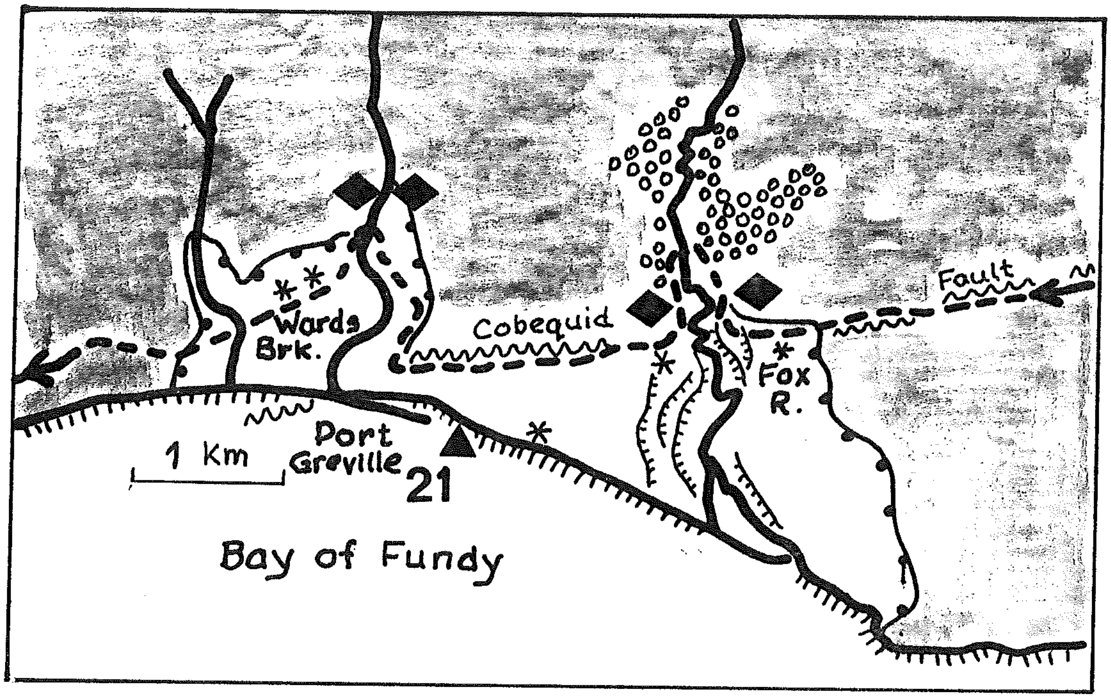

Fig. 23 Map of outwash fan-delta at Fox River, Port Greville and Wards Brook (after wightman, 1980, p. 146).

STOP 21

An abutting delta on the west at **Port Greville** exposes its internal structure in the bluffs east of the harbour. Clearly seen are foreset beds reaching a maximum elevation of 20 m. These have a diverging orientation showing that the delta had a lobate front. More important is the unique occurrence here of three stacked foreset units separated by horizontal (topset?) beds and locally by clay bottomsets. The cycles are thought to represent pulses of aggradation and progradation following episodes of subsidence of the substrate, presumably due to expulsion of water from fine-graded bottomset sediment.

The collapsed foreset beds and the inferred flowage of bottomset beds leads to the question of why only this delta has suffered such disturbance. While it may be due to its great thickness and the steeply sloping substrate as Wightman (1980) believed, perhaps also its location on the Cobequid Fault is significant. Even small movements along this major break in the earth's crust, triggered by deglacial rebound, would be sufficient to disturb the weak sediment pile.

****

From a vantage point high on the delta terrace at Wards Brook a good view is had of Cape Split on the south side of Minas Basin. Famous as a semi-precious, mineral-collecting area, the curved cape is the cuesta edge of a synclinal sheet of resistant Triassic basalt surrounded by weak sandstone and shale.

Continuing west, the so-called **"Parrsboro Shore"** includes the communities of Brookville, Fraserville and Allenville. These hamlets are nestled in deep gorges that incise the flank of Cobequid Mountain. No ice-contact deposits or deltas occur, although there is a terrace at the latter place. Cobequid erratics are absent or rare and locally-derived tills abound. If Cobequid glaciers covered this area there is little evidence.

STOP 22

The village of **Spencers Island** (Fig. 24) has a small raised delta that is the most important in the Five Islands Formation. It is a classic Gilbert-type delta with 1-1.5 m of horizontally stratified closedwork topset outwash gravel deposited by braided meltwater streams. The basal contact, at 32 m ASL (local minimum marine limit) truncates foreset beds which are 15 m thick and inclined 25°-34° to the south away from the source stream, Mahoney Brook. Groups of foresets separated by unconformities are attributed to slumping due to failure of weak bottomset sediment.

About 5 m of bottomset facies are occasionally exposed above and below tide level. They interfinger with the foreset beds, being the basal extension of the more muddy beds. They consist of coarsening-upward couplets of reddish-brown clay/silt and fine sand with dropstones. The rhythmic sedimentation in a glaciomarine environment is explained by Wightman (1980, p. 235) as seasonal variations of grain size; thus the couplets are varves. The beds are locally faulted, slumped and convoluted but the disturbance does not penetrate the foresets indicating that it occurred only as progradation began. Burrows and molds of pelecypods **(Portlandia arctica, Nuculana pernula, Macoma, Mya)** in life position are an important feature. An accelerator date (courtesy R.R. Stea) on the chitinous periostracum that lines the cavities is expected to give the first direct measure of the inception of deglacial marine submergence in Minas Basin .

****

STOPS 23, 24, 25

At the western end of Minas Basin, **Advocate Harbour** area presents extensive raised marine deposits (Fig.24) which are unique in that they include a beach facies that is unaccountably absent elsewhere in the marine deltaic formation. From these comes the first definitive sedimentologic information on **paleotides** in Bay of Fundy during early deglacial time.

The surficial deposits feature a **terrace** at 47 m ASL along McRitchie Brook in West Advocate which Wightman (1980, p. 265) calls a **delta** despite its three times greater gradient (17 m/km) than other Minas Basin deltas. Still, the deposit is only 4 m higher than the upper limit of the gravel beach ridges against till in East Advocate. Perhaps the feature is a remnant of the fluvial fan just slightly inland of the true limit of marine action.

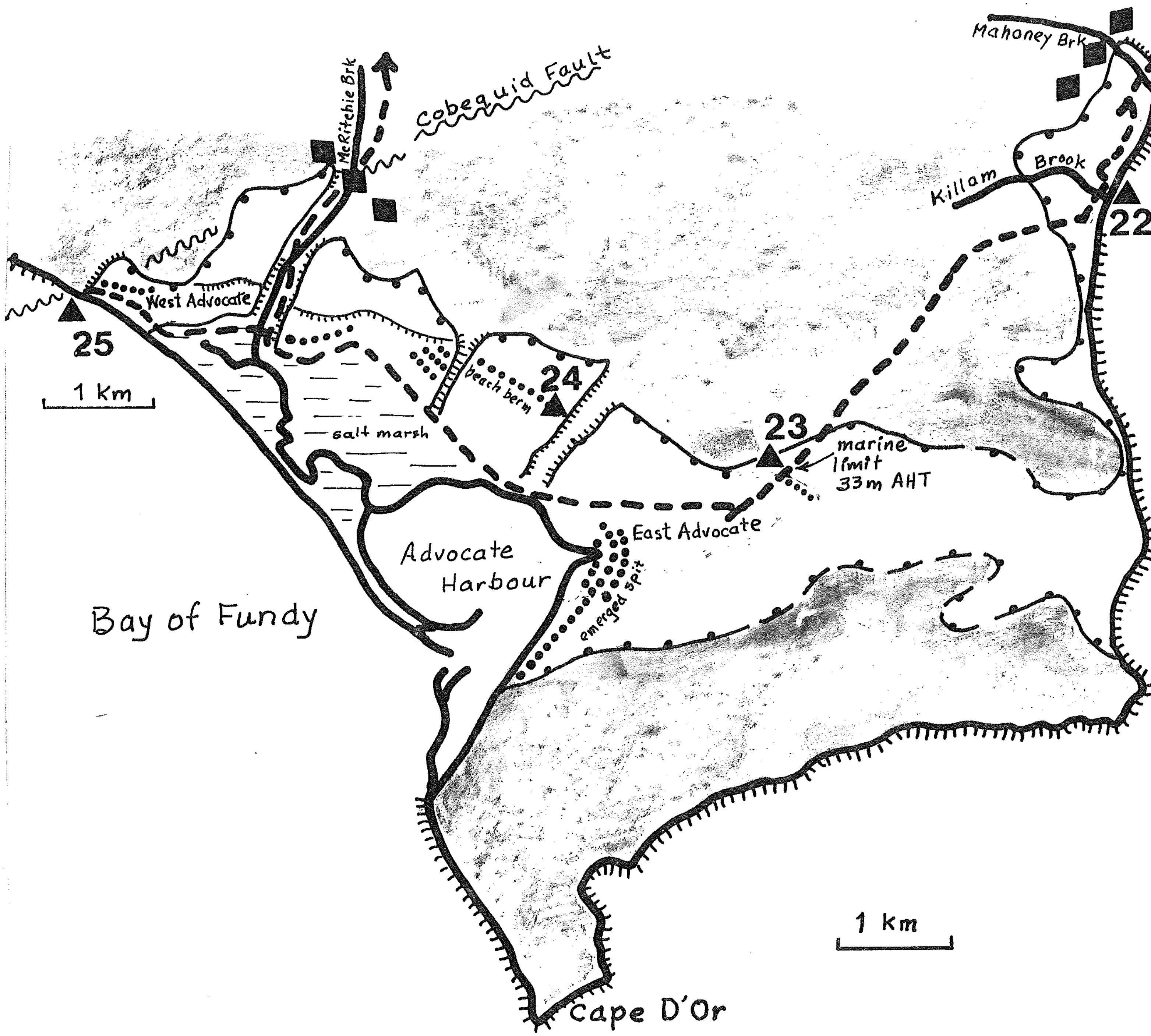

Fig. 24 Map of outwash fan-deltas at Spencers Island and Advocate (after Wightman, 1980, p. 261.

Most of the area is a plain of littoral gravel that slopes from 33 m **(STOP 23)** to present tide level where it is onlapped by salt marsh. It is ornamented with a few beach berms, small scarps, and three minor terraces at 25 m, 18-20 m, and 15 m. Wightman (1976) studied the texture and structure of the berms in a large gravel pit during its enlargement **(STOP 24)**. From numerous exposures, the sedimentary structures could be used to recognize the facies of two littoral zones. The lower extended from subtidal, through foreshore (beach face), to supratidal backshore (Fig. 25). Because the sequence is complete from subtidal

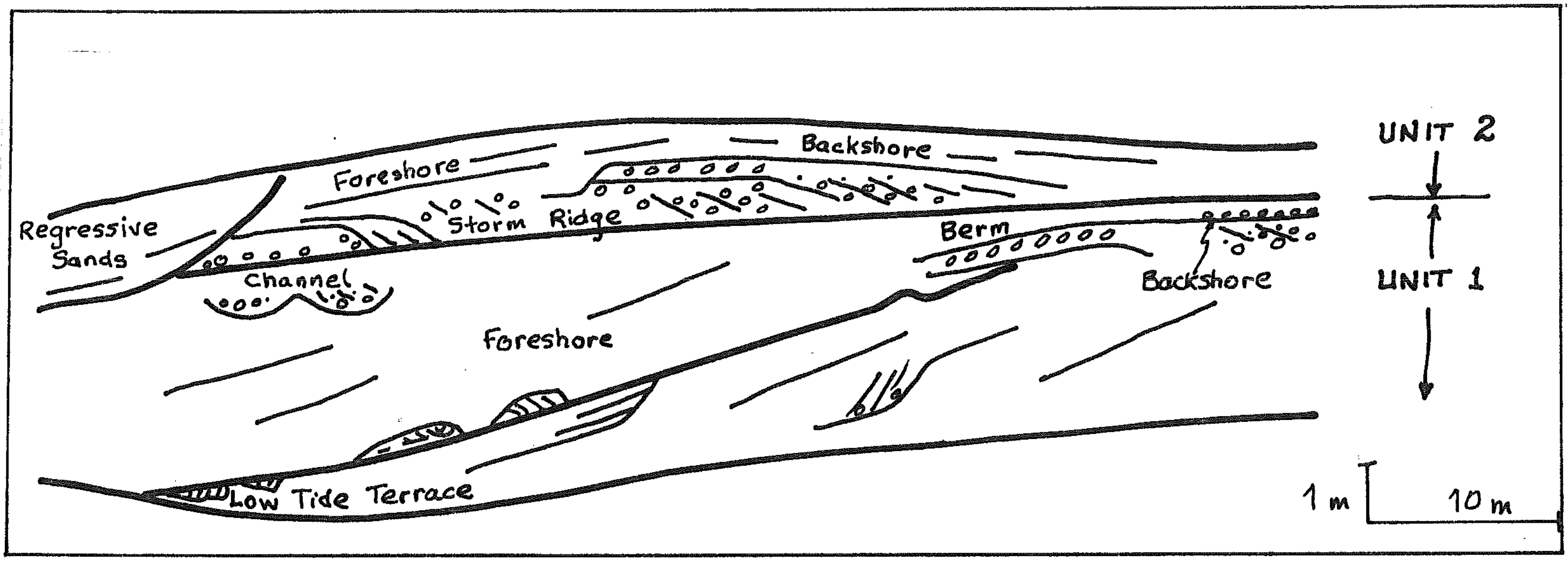

Fig. 25 Composite idealised stratigraphy and facies of a beach berm at Advocate used to measure paleotidal range (from Wightman, 1976, p. 20)

to supratidal, its thickness, **3.4 m, was the maximum paleotidal range.** The upper unit is a coarse washover gravel deposited in the backshore zone of the beach as a storm ridge thrown up during a small 1.7 m transgression.

Mean sea level during formation of the highest raised beach was 23.9 m above present mean sea level. Littoral gravel without beach ridges extends to 33 m ASL (27 m AHT) which Wightman (1980, p. 270) regards as a very early deglacial shoreline and may be considered as the local marine limit in the absence of topset/foreset contacts. As a check, unwashed till surfaces were found at 31 m ASL on the road over Cape D'Or mountain, and good pebble beaches at 26 and 27 m. Thus, for Advocate Harbour area, **marine limit ranges 31-33 m ASL** (25-27 m AHT).

In summary, the Advocate littoral plain was constructed of outwash debouched primarily from McRitchie Brook, presumably from an ice front abutting the north flank of Cobequid Mountain along a zone from Eatonville to New Salem where large masses of ice-contact gravels extend up to the col where McRitchie Brook begins. The outwash was spread eastward by longshore currents to form the beach ridges, much as the modern barrier has been built.

****

STOP 25

A final point in the overall sea-level history of this area is a **supratidal platform** at the mouth of Dewis Brook, West Advocate. At 3 m above its well-developed modern counterpart, the surface truncates Triassic sandstone and is covered by 5 m locally-derived red till and a veneer of marine gravel. The feature may correlate with an old intertidal abrasion surface that is widely seen elsewhere on the Atlantic and Gulf coasts of Nova Scotia at 4-8 m AHT (Grant, 1980). It is assigned to the last interglacial sea-level maximum because of its stratigraphic position relative to Sangamonian organic beds and Wisconsinan tills. Although it has not been sought in upper Fundy embayment, it may occur at Rockport, NB where it is a minimum 1.5 m AHT. The Advocate feature should spur efforts to extend this important stratigraphic and geodynamic datum into Fundy region.

****

From Advocate the route goes northward along Chignecto Bay and thus traces the retreat of a northern ice lobe, and and its associated declining marine limit. Between **New Salem** and **Apple River** a wide belt of thick ice-contact stratified drift with sidehill meltwater channels marks a major ice-marginal stand of the northern ice mass when it stood against the north side of Cobequid Mountain. This position may correlate with the southern edge of the Shulie Lake drift sheet which is the surface till on the Cumberland lowlands and which Stea (in press) interprets as Late Wisconsinan because it overlies a till with a paleosol. (Stacked tills and paleosols are to be seen at Joggins). The Apple River morainal zone is here tentatively correlated with the ice margin that formed the deltas at Fox River and Parrsboro, among others, whether or not it represents the limit of Late Wisconsinan ice.

In keeping with the southward ice flow, and inferred northward retreat of the Chignecto lobe, marine limit declines from a maximum of 37 m ASL on **Squally Point** to 28 m at **Apple River**.

STOP 26

At **Clam Cove** a marine gravel beach with shoreface sand pinches out against unwashed till at 25 m (Stea, 1983, p. 200).

STOP 27

Joggins is famous, not only for its in situ Carboniferous tree trunks, but also for a magnificent 20 m exposure of three tills (Fig.26). First described by Wickenden (1941), and further commented on by Prest et al. (1972, p. 37), the sequence has been studied in detail by Stea (1983, p. 149). He found that the lower till was emplaced by ice moving southeastward from New Brunswick, possibly during Early Wisconsinan time because a supposedly correlative movement buried Sangamonian interglacial organic beds in central Nova Scotia. Vague signs of subaerial alteration at the top of the lower till suggest that the area was deglaciated before emplacement of the middle till which is attributed to south-southwestward movement by an ice mass lying north of Nova Scotia. Alteration at the top of the middle till signifies another paleosol, and hence a second intra-Wisconsinan deglaciation. The surface till has a southwestward fabric due to movement of the last ice lobe that moved down Chignecto Bay. The Joggins sequence is thus one of several Nova Scotia reference sections that preserve a sedimentary record of the major Wisconsinan ice-flow events hitherto discerned mainly by cross-cutting striations. The concordance of depositional and erosional records described by Stea (1982), among others, gives the Nova Scotian sequence prime regional significance despite the paucity of chronometric control.

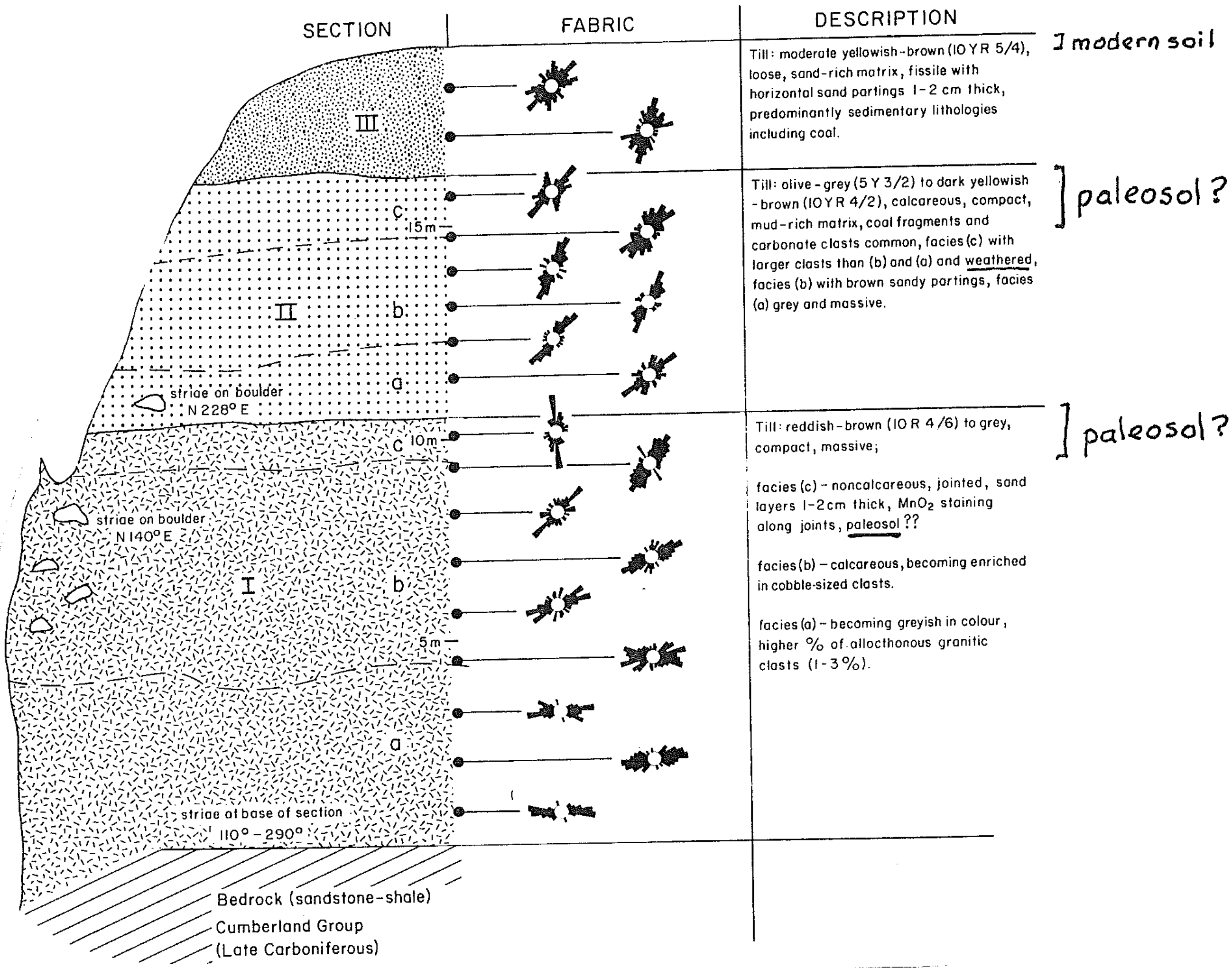

Fig. 26 Analysis of the 3-till section at Joggins (from Stea, 1983).

STOP 28

Just north of Joggins, at **LOWER COVE** (Fig. 28) the sea cliff exposes a 15 cm thick peat layer between surface sand and a reddish-brown muddy diamicton. Similar in age to 6 other near-surface buried organic beds in Nova Scotia, (Fig. 27), the date on the base is 11 800 ± 110 (GSC-3915) and on the top 11 100 ± 120 (GSC-3924). Like the others, it records an abrupt cessation of vegetal accumulation. Whether the cause was a climatic deterioration that caused slope mobility and sedimentation remains to be tested by analyses of the sediments. Nonetheless it is tempting to correlate this occurrence with others which Mott (1985) has ascribed to a short sharp cooling related to southward shift of the oceanic Polar Front. In any case, it lies below local marine limit and thus provides a minimum date for the time that postglacial marine overlap had regressed below a level 6-7 m ASL.

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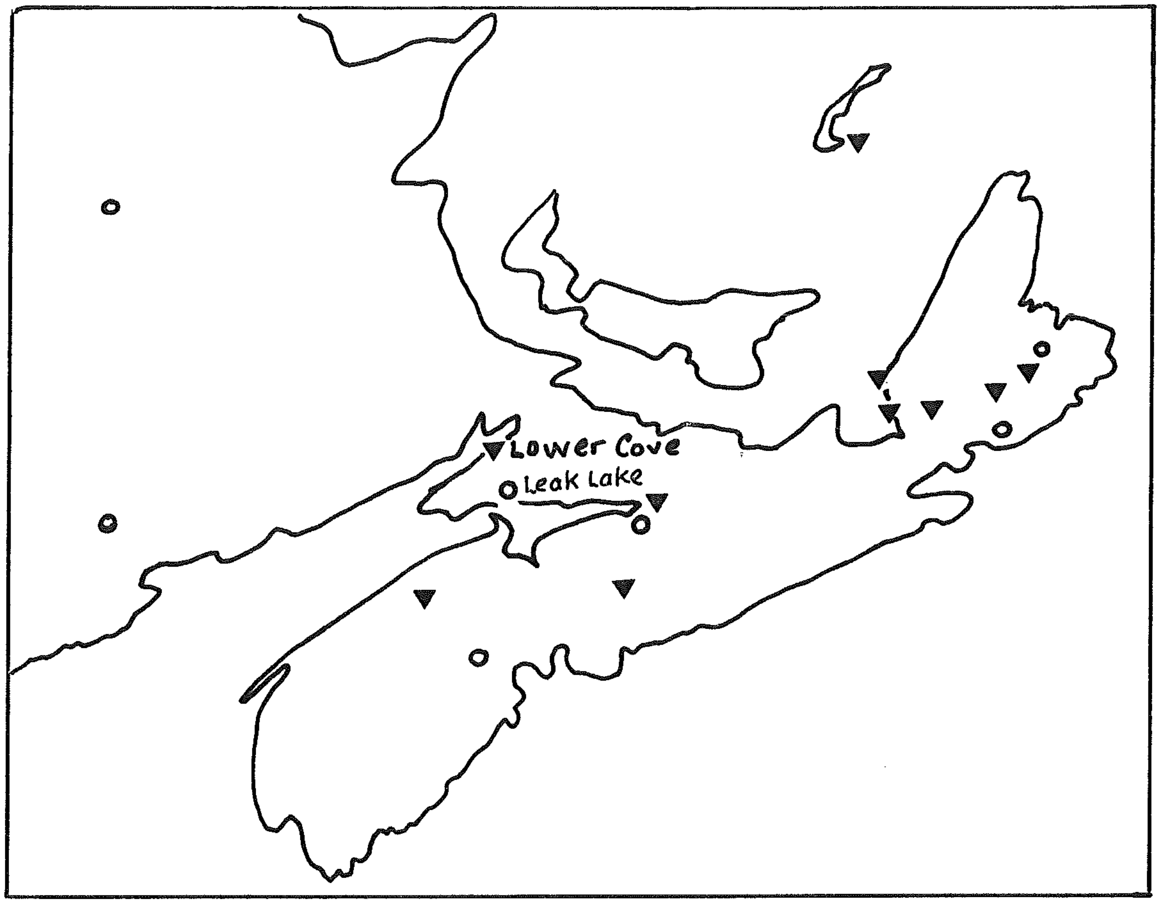

Fig. 27 Map of sites in the Maritime Provinces showing evidence of a climatic deterioration ca. 1-10 ka BP. Circles denote lake-sediment records; triangles are buried organic beds (after Mott, 1985).

AMHERST AREA brings this Quaternary historical review to a conclusion with its record of Late Holocene sea-level fluctuations. The upper Fundy embayment features extensive tracts of tidal sediment (Fig. 28) much of which has been reclaimed from the sea by dykes. On the firm ground of these meadows many borings have been put down for a variety of engineering and scientific reasons, either to probe the depth and composition of the substrate or to study the tidal sequence. About 40 m of marsh mud over a basal forest horizon have been sounded in this way. Detailed information on the upper 15 m comes from numerous hand borings, from study of exposures along tidal gullies, and from the cutbank outside the dykes where the sediment is being eroded by tidal currents.

STOP 29

AMHERST MARSH illustrates the invasion of the living forest by the advancing tidal marsh as relative sea level rises. Saltwater inundation has killed a fringe of trees and the precise limit of marine invasion can be seen as the inner edge of mud and salt-tolerant plants. The innermost tidal limit corresponds to the outer edge of living trees and salt sensitive plants, which are zoned by species according to the duration and frequency of tidal submergence.

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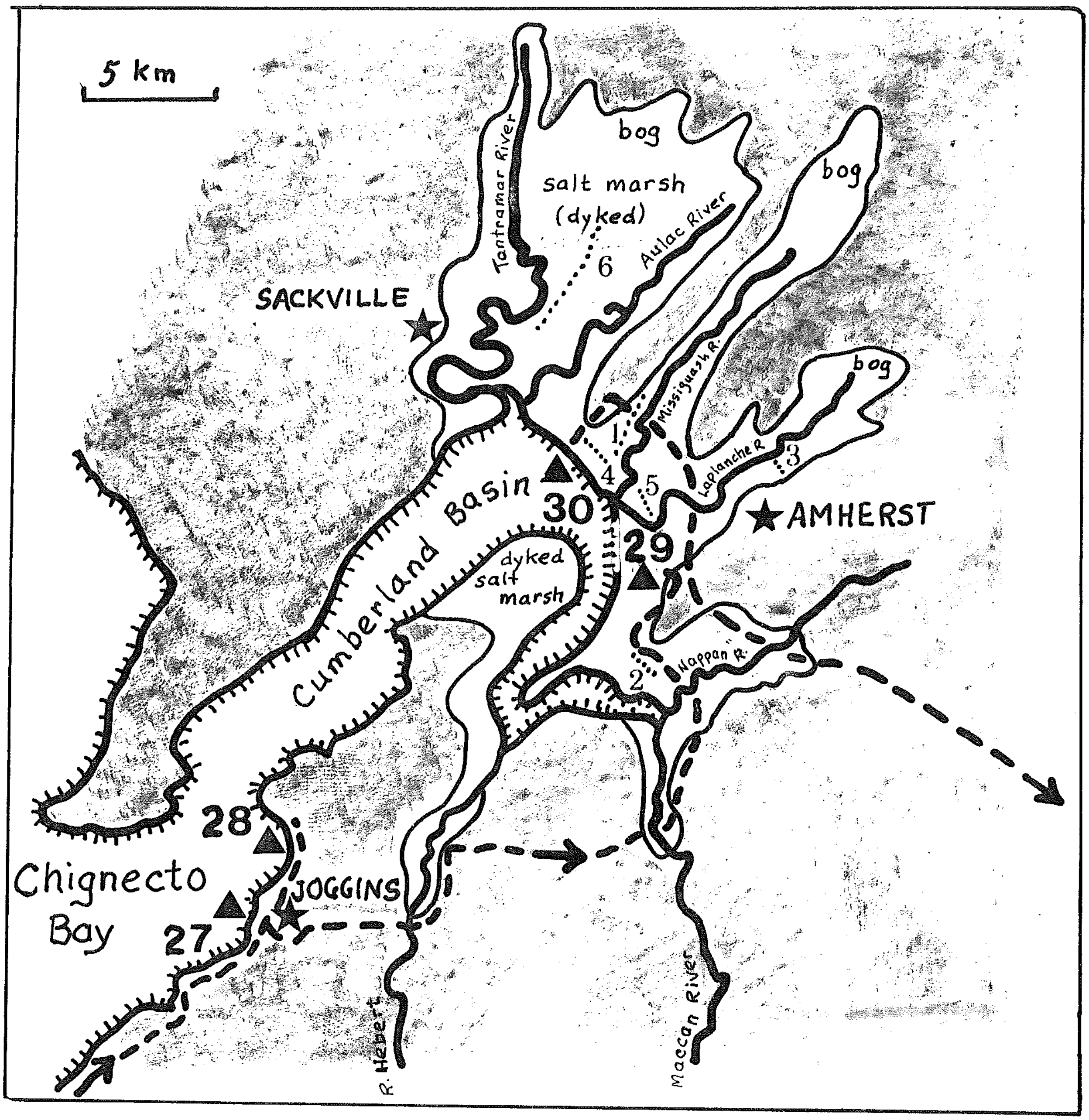

Fig. 28 Map of Holocene tidal marsh deposits in Amherst-Sackville area, showing location of stops, and of boring transects shown in Fig. 25.

STOP 30

FORT BEAUSÉJOUR, just across the border in New Brunswick, gives a good view of the buried submerged forest and the character of the overlying tidal muds (Fig. 29). Here, several large tree stumps, rooted in the solid till substrate, date about 4000 y BP and lie 12-13 m below tide level (Harrison and Lyon, 1963; Grant, 1970). If the tide is fully out, it is possible to see karstic depressions in the till, and giant grooves on the sediment caused by winter ice floes.

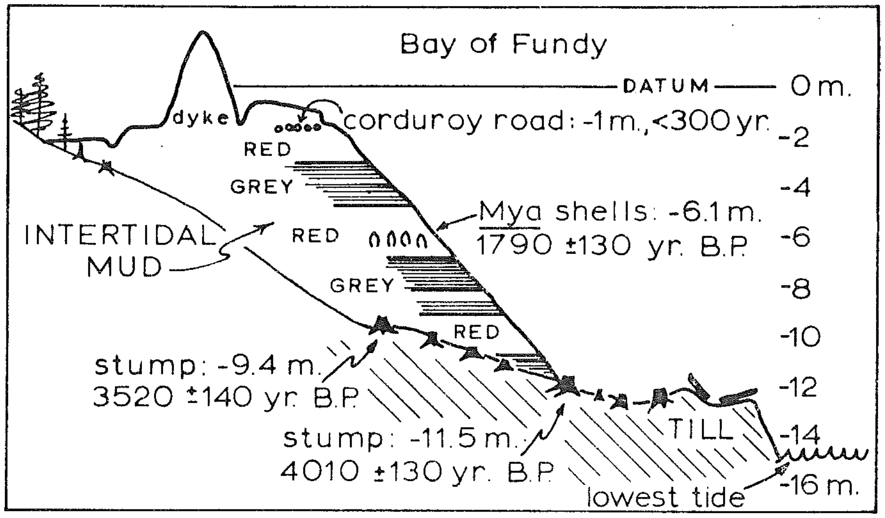

Fig. 29 Cross-section through tidal deposits near Fort Beauséjour NB, showing basal layer of submerged tree stumps rooted in till (from Grant, 1975).

Near the base of the tidal mud sequence, layers of **fresh and brackish-water peat** signify minor regressions of unknown origin. Throughout the sequence, the mud alternates between massive grey and laminated red. The variations are thought to be due to repeated shallowing such that stratification and grass content changed. In the top metre are remains of ditches, drains and roads installed by the Acadian settlers in the early 1600's. Their age and depth agrees with the long term submergence rate determined from submerged trees, as well as with the modern rate of water level rise measured by tide gauges.

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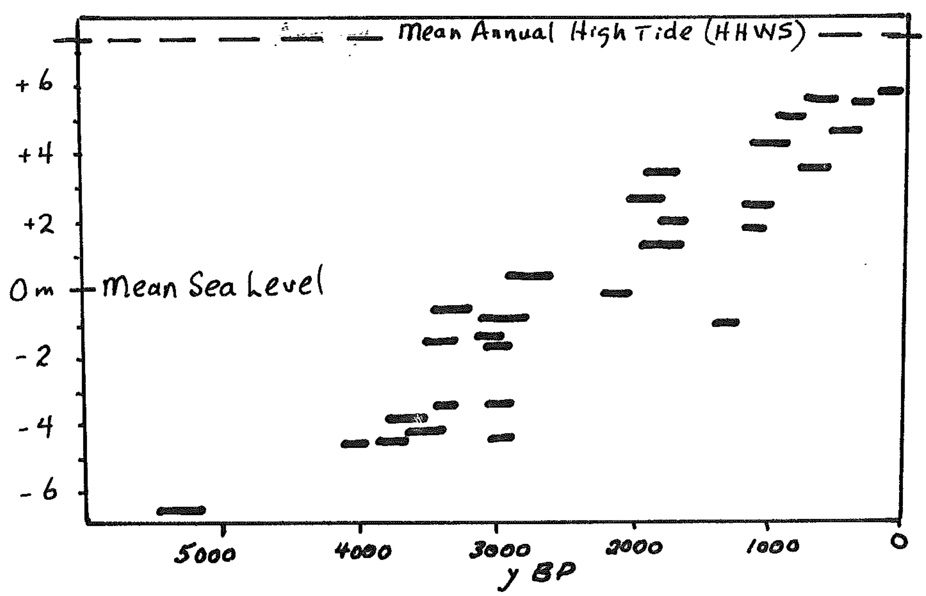

Fig. 31 Summary of ^{14}C dates on submerged freshwater horizons buried by tidal-marsh deposits in the Amherst-Sackville area (from Noordijk and Pronk, 1981). Bars represent error range on dates; no estimate is made of uncertainty in elevation or position of sea level at time of formation.

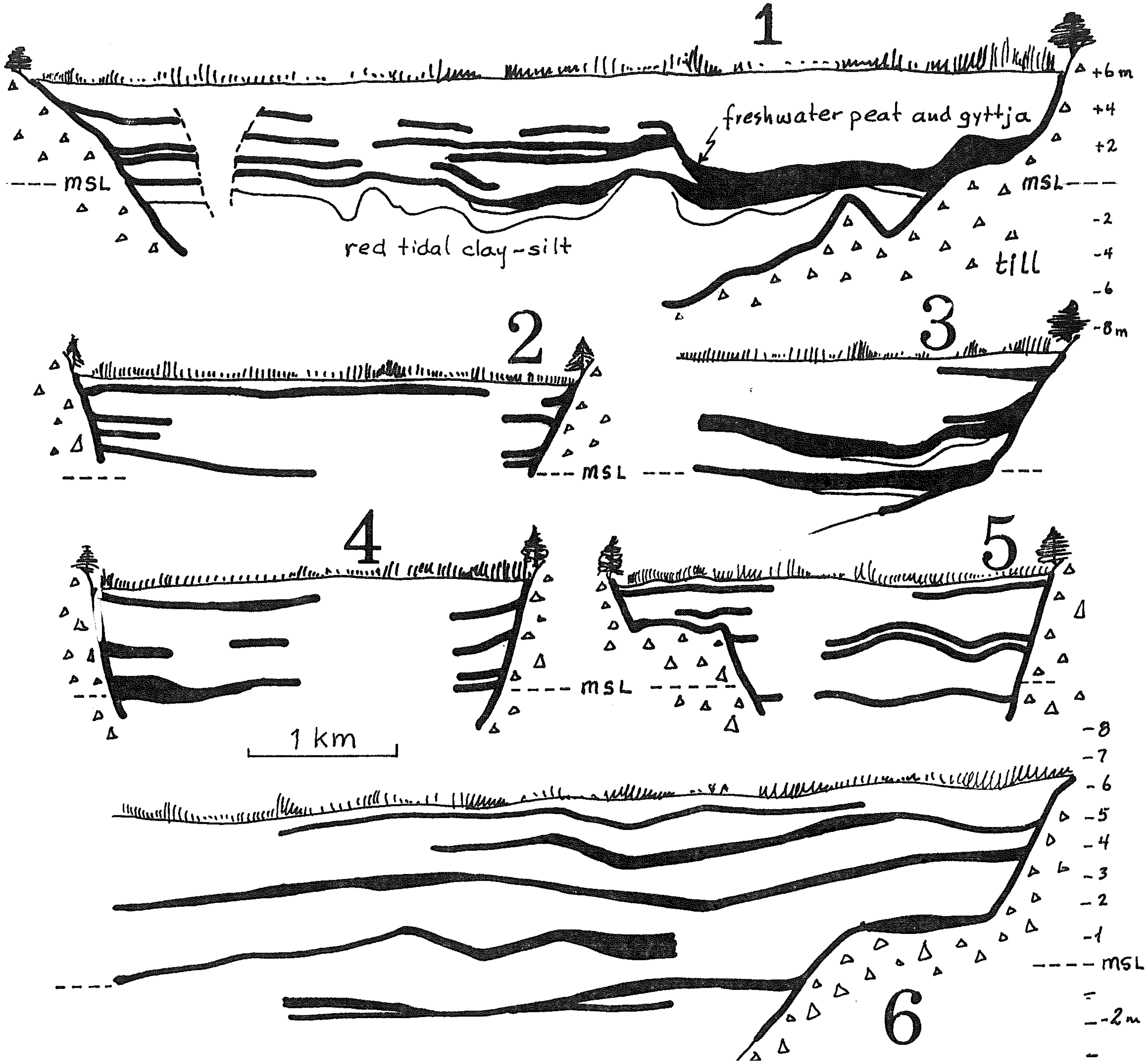

Fig. 30 Selected cross-sections through tidal marsh, Chignecto Isthmus, based on detailed borings, showing four (regressional?) freshwater peat and humus layers (1-5 after Noordijk and Pronk, 1981; 6 after Lammers and De Haan, 1980).

Details of the tidal-marsh stratigraphy have been worked out by Noordijk and Pronk (1981) who made dozens of borings up to 12 m deep along 17 transects across the marshes along the Missiguash, LaPlanche and Nappan Rivers, north and south of Amherst. Lammers and De Haan (1980) conducted a parallel study of the Tantramar marshes in New Brunswick east of Sackville. Six representative cross-sections from their work are shown in Fig. 30. Four extensive freshwater horizons of bog peat and/or humic clay are intercalated in the marine tidal-mud sequence. Dating of the entire sequence (Fig. 31) shows that high tide level has risen about 14 m in the last 4000 years. In the last 3000 years the general transgression has been interrupted at least four times such that the saltwater margin retreated seaward many kilometres and was replaced by freshwater

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APPENDIX 1

SEDIMENTATION IN THE MACROTIDAL, COBEQUID BAY-SALMON RIVER ESTUARY

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Introduction

Estuaries are one of the most obvious sites of land-sea interaction, being, by definition (Pritchard, 1967), locations where marine and fluvial processes meet. Partly as a result of this, estuaries are major sites of sediment accumulation. As geologists, we are interested in the characteristics and three-dimensional organization of these sediments, because such information allows us to recognize similar ancient deposits and to develop predictive models for use in resource exploitation. In order to do this, one must understand the nature of the marine-fluvial interactions, for these control the character and geometry of the deposits.

Nova Scotia, with its lengthy coastline, has many estuaries of various sizes and types. Here we will briefly describe one of the largest and most dynamic estuaries in Nova Scotia, the Cobequid Bay-Salmon River Estuary, a macrotidal estuary with the world's largest tidal range (16.3 m maximum). We will also discuss, in a preliminary fashion, the nature of the tidal-fluvial interactions, and attempt to show how the sedimentary facies are controlled by them.

General Setting

The Cobequid Bay-Salmon River Estuary occupies the headward part of the eastern arm of the Bay of Fundy system (Fig. 1, inset). The estuary is up to 10km wide, and 40 km long from the limit of tidal influence at Truro to the outer end of modern sediment accumulation at Burntcoat Head (Fig. 1). The average tidal range in Cobequid Bay is 11.9m, and maximum tidal current speeds vary from 1.5-2.25 m/s in the outer part of the Bay, increasing to values greater than 2.5 m/s headward of Salter Head (Fig. 1) before decreasing progressively up the river to zero at the limit of tidal influence. Waves are small throughout the area due to the restricted fetch, and reach significant, monthly mean heights of only 0.15-0.6m at the seaward end of the Bay (Amos and Long, 1980), becoming even smaller within the estuary. Impressive amounts of ice cover the Bay from January to April, but its sedimentological significance is minimal (Knight and Dalrymple, 1976). Consequently, sedimentation is tidally dominated throughout most of the estuary; fluvial domination exists only in the 5km stretch below the tidal limit, although some fluvial influence is felt almost to the seaward end of the estuary.

Progressive headward shoaling and tidal wave deformation cause the duration of the flood tide to become increasingly shortened up the Bay. As a result, flood current speeds generally exceed those of the ebb, and most areas experience headward sediment transport (Fig. 1). The most significant sediment source lies seaward of the Bay, and consists of shoreline cliffs of easily-eroded, Triassic sandstone and Pleistocene till and outwash. Because of the sediment influx, net aggradation is occurring (Amos and Long, 1980), despite the fact that the mean high water level is rising at a rate of approximately 30 cm/century (Grant, 1970), due to the combined influence of sea level rise and tidal-range amplification at a rate of 30 cm/century (Amos, 1978). Because of the sandy nature of the source, most of the sediment entering the system is sand, and muddy deposits constitute less than approximately 25% of the Holocene record.

Facies Organization and Descriptions

The sandy deposits in the estuary are divisible into three, broad, depositional zones (Fig. 1): zone 1- the outer, elongate sand bars; zone 2- the inner, "high-energy" sand flats; and zone 3- the tidally-influenced fluvial channel. Sand size decreases headward through these zones, from medium and coarse sand in zone 1, to fine and very fine sand in zone 3. The sands of zone 2 and 3 are bordered by varying amounts of mixed flats, mudflats and salt marsh (Fig. 1). In zone 1, wave action is sufficient to cause coastal erosion, and the foreshores are wave-cut platforms that have only a veneer of mud and/or sand overlying a gravel lag (Knight and Dalrymple, 1975, 1976). This gravel lag is continuous with the gravel that mantles the floor of Cobequid Bay west of the sand bar complex (Amos and Long, 1980).

Zone 1: The elongate sand bars occupy the 20 km-long outer part of Cobequid Bay (Fig. 1). In this zone, both channel bottoms and bar crests rise eastward; channel bottoms reach -25m (relative to mean high water) in the west and rise to -10m at Salter Head, while bar crest elevations range from -15m opposite Noel to -5m at the east end of the complex (Knight, 1980). Within the complex, three main channels exist: a shallow, central, ebb-dominated channel; and two, deeper, marginal, flood-dominated channels. The flood and ebb channels are linked by smaller (1.5-5m relief) channels (swatchways) which migrate headwards at rates of approximately 100m/year. The two marginal channels occupy fixed positions and both terminate headward where bar crestlines attach to the foreshores. The central channel occupies a stable position north of Salter Head, but further seaward its location switches periodically as swatchways grow and capture the main ebb flow. Despite these variations, the bar system retains the chracteristics of a large, ebb-tidal delta (Knight, 1980).

Strong grain size segregations occur between the ebb and flood portions of the bar complex. The flood areas consist of medium to coarse sand derived from the cliffs to the west, whereas the ebb channel contains primarily fine sand derived from the inner sand flats. The swatchways are places of mixing, with coarser sand in their bottoms, and finer sand higher on the depositional, western flank. Bedform distributions closely follow the grain size patterns. In general, sandwaves (Dalrymple, Knight and Lambiase, 1978) cover most of the flood-dominated areas, producing complex, headward-inclined cross-stratification (Dalrymple, 1984) in which herringbone structures are common. Within the finer sands of the ebb zones, three-dimensional, type 2 megaripples (Dalrymple et al., 1978) predominate. These generate nearly unimodal, ebb-oriented, trough cross-bedding. Swatchway migration will generate up to 5 m thick, fining-upward sequences that may contain headward-dipping, epsilon cross-beds with dips of 3-15°.
Cross-bedding at the base of each sequence will be oriented at an oblique angle to the estuary axis, whereas that at the top will be aligned more or less parallel to the length of the Bay. Biogenic structures are absent from most parts of this zone.

Zone 2: This zone can be subdivided into an outer, sand bar-sand flat transition region (zone 2a), and an inner sand flat, or "braid bar", area (zone 2b) (Fig. 1). The transition zone has characteristics of both zones 1 and 2b, and contains intermixed patches of megaripples and/or sandwaves (as in zone 1), and upper-flow-regime plane bed (as in zone 2b). (These differences in bed configuration are due to variations in both current speed and grain size, with grain size exerting the primary control; upper plane bed occurs in fine and very fine sand at the same current speeds as megaripples and sandwaves are stable in medium and coarse sand.) This transition subzone has not been studied in detail, and is not discussed further here.

Subzone 2b, the inner sand flat, is between 4 and 6 km long, and is characterized by multiple, migrating, shallow (0.5-2.0m deep), ebb-dominant channels that bifurcate around flood-asymmetric "braid bars". Individual bars range from 0.25-1.2 km in length, and migrate headwards at rates of up to 2.5m/tidal cycle. The channels range from 5-200m in width, and migrate at rates that are inversely proportional to channel size; the large channel which drains the Salmon River alternates its position between the north and south shores of the Bay on a time scale of several decades, while the smallest channels migrate at the same rate as the braid bars.

Maximum, U_{100} current speeds reach 2.6 m/s on the flood, and 2.1 m/s on the ebb, in water depths of 1-5m, so that upper flow regime conditions prevail over large areas; megaripples are rare as a result, except at the edges of the sand flats, adjacent to the mixed flats. An enigmatic bedform, here termed a sandsheet, is abundant in zone 2b, however. This feature is characterized by very low amplitudes (5-10 cm) and long wavelengths (> 5m), and contains upper-flow-regime horizontal and low-angle stratification. Vertical aggradation of successive sandsheets produces sequences up to 3m thick composed of 5-10 cm-thick sandsheet units separated by heavy mineral lags. Horizontally-spiralled burrows constructed by *Paraonis* and simple *Skolithos*-like burrows are the only biogenic structures present.

Zone 3: This innermost zone is restricted to that portion of the estuary where a single, confined channel occurs. Throughout this zone, the radius of curvature of the channel bends decreases progessively headward, permitting the recognition of 2 subzones: 3a- a "straight" reach with alternating, bank-attached sand flats; and 3b- a tidally-influenced, meandering reach.

Subzone 3a is approximately 3-4km long and is characterized by a meandering thalweg in a gently sinuous channel. The alternating, bank-attached sand flats in the channel have two terraces, the lower one consisting of sand, and the upper one of mixed sand and mud. Maximum, U_{100} values for the flood and ebb on the lower terrace are 2.6 m/s and 1.7 m/s respectively, and upper flow regime conditions prevail; consequently, parallel lamination is abundant. By contrast, flaser and lenticular bedding predominates in the upper terrace. Evidence of vertical accretion of mudflats and salt marsh along both sides of the channel suggest that the channel occupies a relatively stable position.

Subzone 3b is 4-5km in length, and contains headward-tightening, meander loops that are characterized by triple-terraced point bars. Each point bar is symmetrical about a median line, with a flood-dominant, seaward portion and an ebb-dominant, headward segment. U_{100} values are lower than in zone 3a, reaching only 2.0 m/s and 0.8 m/s during the flood and ebb respectively; therefore, each point bar is composed predominantly of tidal bedding, with very fine sands occurring near the channel thalweg. Erosional, epsilon cross-stratification surfaces truncate the tidal bedding at intervals, attesting to the lateral migration of the point bars, which has occurred at an average rate of 3.4 m/year in one documented case. Migration appears to be episodic, however, and is separated by periods when the point bars aggrade vertically. Bioturbation in this zone is minor, with rare surface grazing trails and occasional simple, Skolithos-like burrows in the muds.

Fringing Muddy Facies: The mixed flats, mudflats and salt marshes which border zones 2 and 3 are similar in their characteristics to those described from the North Sea coasts (Reineck, 1972). The mixed flats change landward from ripple cross-laminated sands, through flaser-bedded sand, to interbedded sands and muds. Lenticular bedding predominates in the mudflats, while crude horizontal bedding and massive, root-mottled units characterize the salt marsh. Adjacent to zone 2, U-shaped burrows created by Macoma balthica and Corophium volutator are locally abundant, as are Skolithos-like burrows and other horizontal feeding traces. Further headward, the diversity and density of burrowing decreases rapidly, so that almost nothing lives in the muddy sediments adjacent to zone 3a. Locally in the zone 2 mudflats, 10-20 cm-thick beds are present which pass upwards from non-bioturbated to extensively bioturbated mud. Ice-push (?) deformation structures and ice-rafted pebbles cap these beds in some places, suggesting that each bed represents an annual sedimentation unit. Meandering tidal gulleys are also present through the salt marsh and mudflats, and terminate near the mixed flat-sand flat junction. These gulleys have very stable locations, so that lateral-accretion bedding is not extensive.

Facies Model

The rising high-tide levels have caused the facies zones to trangress headwards erosionally; as a result, the sediments of all facies zones lie directly on a basal gravel lag which has been produced by current and wave erosion of underlying units, and/or by fluvial processes prior to marine transgression. In only a few locations are transgressive sequences observed. One of these occurs immediately to the west of Salter Head where the sands of Selmah Bar are migrating eastward over mudflats (Fig. 1).

Despite the transgressive conditions, the system has a remarkable ability to aggrade/prograde rapidly. Some of the salt marshes north of zone 2 may have been initiated several thousand years ago at a lower sea level, and have grown upwards at the same approximate rate as the rise of high tide level. Over the past 120 years, the sand bars of zone 1 have also experienced net aggradation at a rate that is greater than that of the rise of the high tide level (Amos and Long, 1980). The progradation which has occurred along the south side of zone 2 (Fig. 1) is even more spectacular, however. In 1959, the main channel of the Salmon River hugged the south shore, but then, some time prior to about 1970, it switched to a location close to the north shore. Since then, mudflats and salt marsh have prograded seaward more than one kilometre, producing a sediment wedge up to 8m thick. It is behaviour such as this which leads us to believe that the bulk of the estuary fill will be regressive (progradational and/or aggradational) in character.

The idealized, regressive facies model (Fig. 2), which has been constructed by projecting the facies zones horizontally, consists of an overall fining-upward sequence 20-25m thick. The lower one-half to two-thirds will be composed primarily of medium to coarse sand containing the headward-inclined, complex cross-bedding produced by the flood sandwaves, together with the fining-upward sequences generated by the migrating swatchways. This flood-dominated package will be erosionally overlain in some places by the finer, ebb channel sands containing seaward-directed, trough cross-bedding. The 2-4m thick unit of fine to very fine sand of the overlying "high-energy" sand flats (zones 2b and 3a) will consist predominantly of parallel lamination, with scattered low-angle inclined bedding, ripple cross-lamination, and mud drapes. High-angle cross-bedding will occur rarely, particularly immediatly overlying the parallel-laminated zone. The top 4-5m of the sequence will pass upwards from flaser-bedded sands to massive salt marsh muds. Minor amounts of longitudinal cross-bedding will also be present, and larger fluvial channels will cut through the upper portion of the sequence. The deposits within these channels will generally be similar to those of sand flats, mixed flats, mudflats and salt marsh, but will contain epsilon cross-stratification surfaces and fewer biogenic structures. The extensive parallel lamination of the high-energy sand flats (Fig. 2) is, perhaps, the most diagnostic feature of the macrotidal environment.

Depositional Controls

The order and length of the sandy zones present within the Cobequid Bay-Salmon River estuary are controlled by four, first-order variables: 1) fluvial influence; 2) tidal influence; 3) the depositional gradient; and 4) the basin width. The fluvial and tidal influences increase (or decrease) in opposite directions, and together control the funnel shape of the estuarine system (variable 4). The depositional gradient, which is a function of the coastal zone topography, controls the lateral extent over which the fluvial and tidal influences act. This in turn controls the length of each depositional zone. Within this conceptual framework, we suggest that the boundaries between the depositional zones in the head of the estuary may be defined as follows: the division between zones 2 and 3 is placed at the critical width within the estuary, relative to the size of the fluvial system, where the channel pattern changes from single (zone 3) to multiple (zone 2); and within zone 3, the inner, meandering reach is an area where, to a first approximation, the fluvial influence is greater than the tidal influence. Because of the very large tidal range in the Bay of Fundy system, all of the tidally-dominated zones have greater lengths than they would have under smaller ranges. The small discharge of the Salmon River (264,000 dam^3; = low fluvial influence) contributes to this. On the other hand, the depositional gradient in this area is steep, relative to a coastal-plain setting, and the zones are correspondingly shorter. We would also suggest that this set of four factors is applicable to all estuarine systems, and that it may be possible to develop a unified model for the type and extent of estuarine facies by quantifying these parameters.

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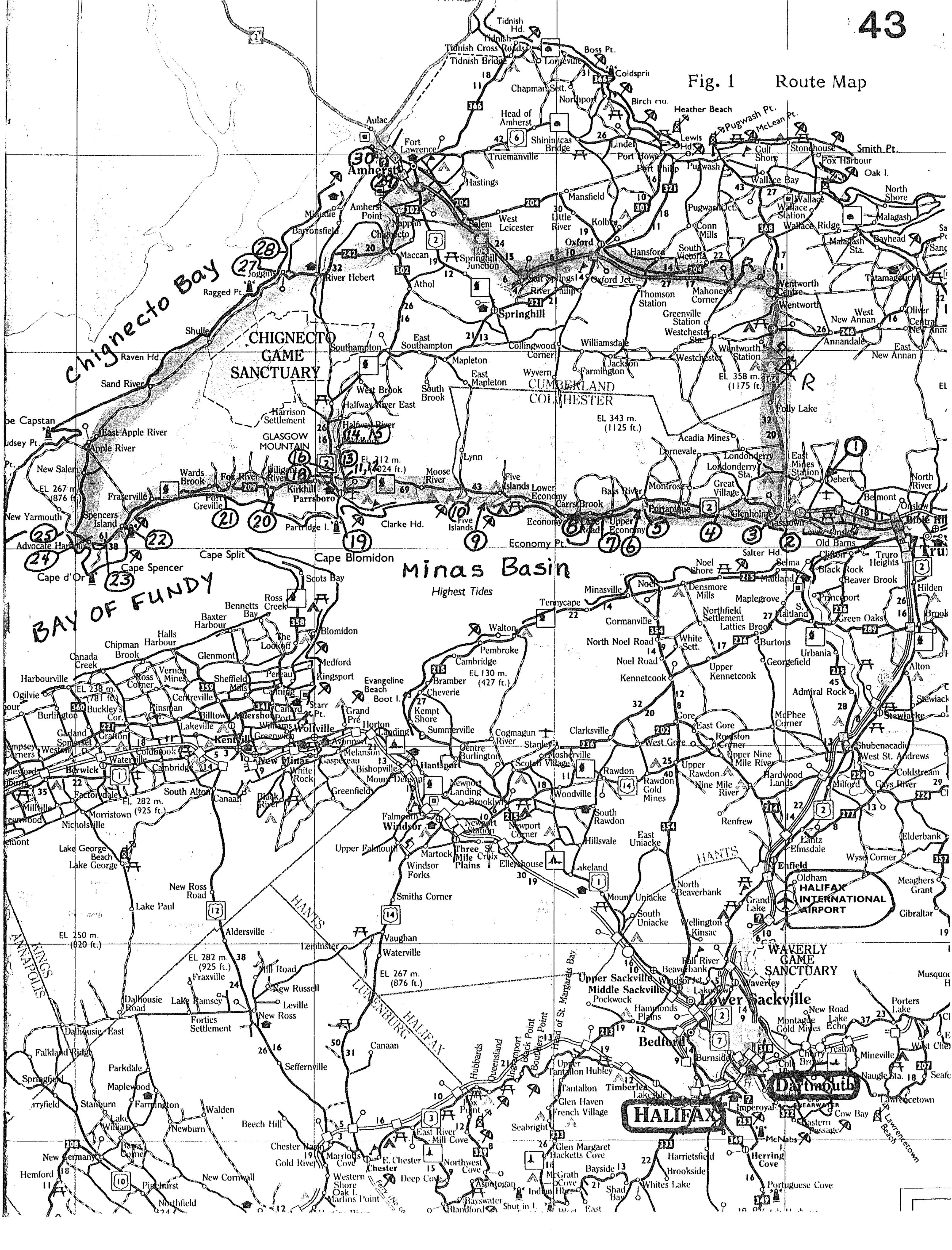

Fig. 1 Route Map