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# QUATERNARY GEOLOGY, CAPE BRETON ISLAND, NOVA SCOTIA

Douglas R. Grant

1994

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**Cover description**

Aspy faultline scarp in northern Cape Breton Island separates a highlands plateau that represents a Tertiary peneplain developed on resistant crystalline rocks, and separates it from a younger lowland peneplain cut across sedimentary rocks. The fault has been upthrown a further 15 m on the south side since the last interglaciation. GSC 204024-T

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# Preface

Cape Breton Island is an important economic and cultural unit of the Maritime Provinces region. Its unique landscape and natural resources make it the heart of Nova Scotia's mineral industry and its prime tourist region.

In recent decades the island has attracted and sustained increased economic development and industrial activity, focused on deepwater trans-shipment facilities, mineral exploration and exploitation, large-scale mechanized forestry operations, and tourism. Given the proven potential of natural and human resources, growth is likely to accelerate and this will place further demands on effective use of the land by competing interests. A major consideration will be the optimal use of scenic wilderness, so as to achieve a balance between development and preservation of environment.

This report describes and explains the unconsolidated surface materials and landforms of Cape Breton Island. It traces the history of geological processes that have shaped the area during late Cenozoic time, with emphasis on the previous interglacial period and the last glacial period. The information is of use for mineral exploration in drift-covered areas, groundwater extraction, waste disposal, foundation siting in areas of complex stratigraphy, inventory of granular resources, and the assessment of the potential for large-scale reforestation. The historical account enhances appreciation of landscape values and provides an insight into the evolution of vegetation, climate, and coastal change up to the present time.

Elkanah A. Babcock
Assistant Deputy Minister
Geological Survey of Canada

# Préface

L'île du Cap-Breton est une importante entité économique et culturelle de la région des provinces Maritimes. Ses paysages uniques et ses richesses naturelles en font le centre de l'industrie minérale de la Nouvelle-Écosse et sa principale région touristique.

Depuis quelques décennies, l'île attire et stimule le développement économique et industriel, axé sur les installations de transbordement dans les secteurs d'eau profonde, la prospection minérale et l'exploitation des minéraux, les opérations à grande échelle de foresterie mécanisée et le tourisme. Étant donné le potentiel prouvé des ressources naturelles et humaines, cette croissance tendra probablement à s'accélérer et augmentera les exigences quant à l'utilisation efficace des terres par des intérêts concurrents. Un important détail sera l'utilisation optimale des réserves naturelles de valeur esthétique, pour que l'on puisse atteindre un équilibre entre le développement et la conservation de l'environnement.

Dans le présent rapport, l'auteur décrit et donne une explication détaillée des matériaux de surface non consolidés et de la topographie de l'île du Cap-Breton. On décrit l'évolution des processus géologiques qui ont façonné la région au cours du Cénozoïque tardif, notamment pendant la période interglaciaire précédente et pendant le dernière période glaciaire. L'information sert à la prospection minérale dans les régions couvertes de sédiments glaciaires, à l'extraction des eaux souterraines, à l'élimination des déchets, au choix de l'emplacement des fondations dans les régions de stratigraphie complexe, à l'inventaire des ressources en granulats et à l'évaluation des possibilités de reforestation à grande échelle. Le compte rendu historique permet de mieux apprécier la valeur des paysages et nous donne un aperçu de l'évolution de la végétation, du climat et des modifications du littoral jusqu'à l'époque actuelle.

Elkanah A. Babcock
Sous-ministre adjoint
Commission géologique du Canada

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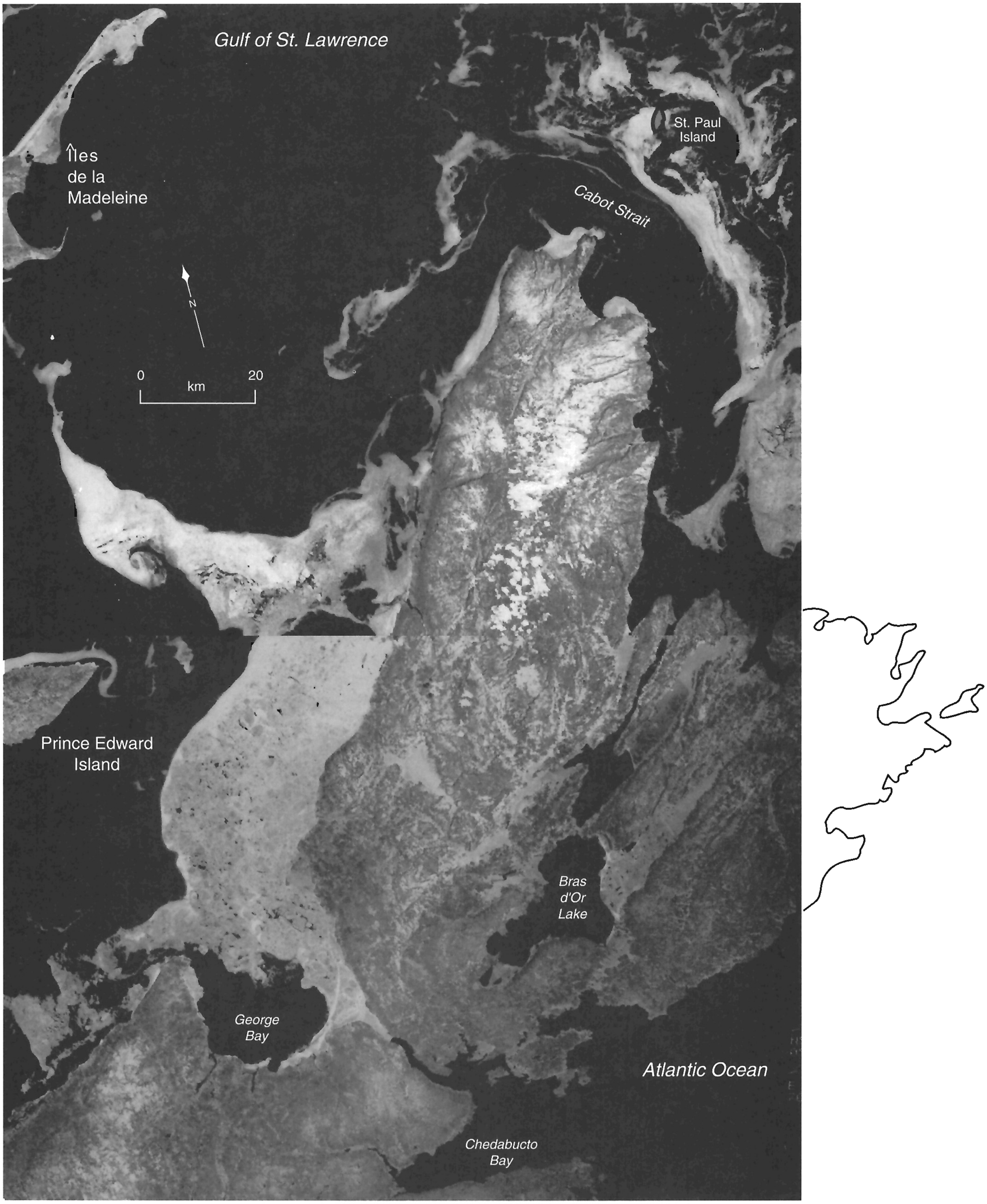



***Frontispiece.*** *Satellite view of northern Cape Breton Island and the Îles de la Madeleine imaged 13 April 1974 showing typical late-lying snow fields on highlands plateau and winter pack ice exiting Gulf of St. Lawrence. Major visible surficial geological features are high-relief morainal tracts on the southern lowlands and the fracture-lineated bedrock plain of the northern highlands. (ERTS 1629-14195)*

# QUATERNARY GEOLOGY, CAPE BRETON ISLAND, NOVA SCOTIA


### *Abstract*

*Tertiary planation and uplift exhumed two unconformities (paleoplains) and produced two peneplains into which Quaternary glaciation excavated submarine basins and channels. The last interglaciation deposited organic beds and cut an intertidal rock bench and littoral gravel, now emerged 2-7 m. The nonglacial period spanned 126-47 ka BP according to 30 Th/U and $^{14}C$ wood ages at 16 sites. Two temperate cycles climaxed 125 and 86 ka BP and are correlated to oxygen isotope substages 5e and 5a; the first had hardwood forest reflecting a warmer than present climate; the second had boreal forest like today. Early-Middle Wisconsinan time is recorded by three tundra/boreal forest alternations, reflecting climatic deterioration until Wisconsinan glaciers arrived 62-47 ka BP. Several ice flow phases left crosscutting striations, superimposed drumlins and fluting, and three main till sheets of differing lithology, colour, texture, and provenance. An ice cap began on the highlands, then an ice sheet from mainland Nova Scotia advanced eastward and deposited a red foreign till on the lowlands. South-flowing regional ice (Laurentide?) crossed the island; it was cold-based on the highlands and preserved preglacial soil and landforms. Weathered striations beneath fresh till suggest an ice free interval. A northward flow from Scotian Shelf which introduced shell fragments may have signalled an ice rise, a flow reversal due to drawdown, or a new ice dome on the emergent shelf. The ice divide shifted onshore forming a Bras d'Or Lake centre. Gulf of St. Lawrence ice on the west coast dammed lakes that drained northward. Climatic cooling 11-10 ka caused glacial readvance and renewed solifluction. Highland ice lasted until 8-9 ka and Neoglacial cirque glaciers may have formed. Neotectonics tilted the bench southward and displaced it 15 m at Aspy Fault. Submerged features, geodetic relevelling, and tide gauging show a relative sea level rise of 20-70 cm/100 a due glacial forebulge collapse. Unstable slopes have landslides and sagging cliffs.*

### *Résumé*

*La pénéplanation et le soulèvement survenus au Tertiaire ont mis en évidence deux discordances (paléoplaines) et produit deux pénéplaines dans lesquelles les glaciations du Quaternaire ont creusé des bassins et chenaux sous-marins. Pendant le dernier interglaciaire se sont déposés des lits organiques et ont été entaillés une banquette rocheuse intertidale et des graviers littoraux, qui émergent actuellement de 2 à 7 m. La période non glaciaire a couvert l'intervalle d'il y a 47 à 126 ka, selon 30 datations par la méthode U/Th et par la méthode $^{14}C$ appliquée au bois à 16 sites. Deux cycles tempérés ont atteint leur apogée il y a 125 ka and 86 ka et ont été corrélés avec les sous-étages 5e et 5a de la stratigraphie isotopique de l'oxygène; le premier était caractérisé par une forêt de feuillus qui indiquait un climat plus chaud que celui de nos jours, le second, par une forêt boréale semblable à la forêt boréale actuelle. Le Wisconsinien précoce-moyen se traduit par trois alternances de la toundra et de la forêt boréale, indiquant une détérioration du climat jusqu'à l'arrivée des glaciers du Wisconsinien il y a entre 47 et 62 ka. Plusieurs phases d'écoulement glaciaire ont laissé des stries glaciaires entrecroisées, des drumlins et cannelures superposés et trois grandes nappes de till de lithologie, couleur, texture et provenance différentes. Une calotte glaciaire a commencé à se former sur les hautes terres, puis un inlandsis en provenance de la région continentale de la Nouvelle-Écosse a progressé vers l'est et a déposé un till rouge exotique sur les basses terres. Des glaces régionales s'écoulant vers le sud (glaces laurentidiennes?) ont traversé l'île; elles avaient une base froide sur les hautes terres et ont conservé les sols et la topographie préglaciaires. La présence de stries altérées sous du till non altéré donne à penser qu'il y a eu un intervalle caractérisé par l'absence de glaces. L'écoulement des glaces vers le nord à partir de la plate-forme néo-écossaise, qui a laissé des fragments de coquillages, a pu indiquer la montée des glaces, l'inversion de l'écoulement due à l'ablation des glaces, ou l'apparition d'un nouveau dôme de glace sur la plate-forme émergente. La ligne de partage glaciaire s'est déplacée vers le littoral en formant un centre du lac Bras d'Or. Sur la côte ouest, les glaces du golfe du Saint-Laurent ont endigué des lacs dont les eaux se vidaient vers le nord. Le refroidissement survenu il y a 10 ou 11 ka a provoqué une réavancée des glaces et une nouvelle phase de solifluxion. Les glaces des hautes terres ont duré jusqu'à il y a 8 ou 9 ka, et il est possible que des glaciers de cirque néoglaciaires se soient formés. Des déformations néotectoniques ont basculé la banquette vers le sud et l'ont déplacée de 15 m à l'emplacement de la faille d'Aspy. Les structures submergées, les nouvelles mesures de nivellement de précision et les mesures des marées montrent que le niveau de la mer a monté de 20 à 70 cm/siècle par suite de l'enfoncement périphérique glaciaire. Les pentes instables ont subi des glissements de terrain et portent des versants affaissés.*

## SUMMARY

This report presents the results of a combined geomorphological, stratigraphic, and glacial geological study of the island's landforms and surficial materials. It shows that Cape Breton Island is important from a Quaternary geological point of view for two main reasons. The complex physiographic development illustrates Cenozoic tectonics and its stratigraphy records a glacial and interglacial sedimentary sequence that is unparalleled in the region for its completeness. The island is the heart of Nova Scotia's mineral industry and its prime tourist region, so this information has a variety of practical applications. It can be adapted to the search for mineral deposits by the practice of drift prospecting and it helps meet the popular demand for geo-historical interpretation of changing landscape, climate, and vegetation. The data is of further value in groundwater extraction, waste disposal, foundation siting, granular resources, forest capability, and real estate acquisition.

Physiographically, the island has strongly differentiated topographic relief which features several ancient planation surfaces that range in age from Paleozoic to Cenozoic. The older surfaces are paleoplains or exhumed unconformities which extend offshore and are dated from the rocks which overlie them. One such paleoplain is of Carboniferous age and forms a facet on the northeast highlands; another of Jurassic/Cretaceous age forms the gently shelving south coast. Three younger surfaces are peneplains of Cenozoic age. The oldest is mid- to late Tertiary age and is developed mainly on resistant crystalline rocks. Represented by the flat-topped hills and highlands of the Atlantic Uplands, it was originally at or near sea level, but it is now tilted up to the northwest so that its present elevation ranges from 100 m in the south to over 500 m in the north. Locally, it preserves intensely weathered saprolite and deeply disintegrated grus with corestones that are considered to be Tertiary soil remnants. Its regional elevation variation is slightly domed over Cabot Strait area, perhaps suggesting upwarp in response to erosional unloading. Two notable tectono-geomorphological features are depressions in granite adjacent to lowlands underlain by gypsum which may indicate that the highlands are a thin overthrust slab that is collapsing into underlying karstic rocks. Inset into the uplands is a second peneplain developed on intervening nonresistant sedimentary rocks during a later planation period. It forms a lowland complex belonging to the Maritime Plain. A third erosion level is represented by flat-floored submarine channels cut to about -400 m during lower sea levels in late Tertiary to early Quaternary time. Except for the 200 m deep basins of Bras d'Or Lake (and the continental shelf) which have been glacially excavated in semiconsolidated rocks, Quaternary geomorphic modification has been relatively minor, being limited to fluvial incision and local glacial scouring of upland areas and voluminous till deposition in lowland areas.

## SOMMAIRE

Dans le présent rapport, on présente les résultats d'une étude combinée de la géomorphologie, de la stratigraphie et de la géologie glaciaire des formes de relief et des matériaux de surface de l'île. Les résultats démontrent que l'île du Cap-Breton est importante du point de vue de la géologie du Quaternaire pour deux raisons principales. Le développement physiographique complexe illustre la tectonique du Cénozoïque et la stratigraphie présente une séquence sédimentaire glaciaire et interglaciaire qui est unique dans la région du point de vue de son intégralité. L'île est le coeur de l'industrie minérale de la Nouvelle-Écosse et sa plus importante région touristique; par conséquent, l'information recueillie lors des travaux a diverses applications pratiques. Elle se prête à l'exploration de gîtes minéraux par la prospection glacio-sédimentaire et aide à répondre à la demande populaire d'interprétation géo-historique de l'évolution des paysages, du climat et de la végétation. Les données sont également importantes pour ce qui est de l'extraction des eaux souterraines, de l'élimination des déchets, du choix de l'emplacement des fondations, des ressources en granulats, des possibilités en foresterie et de l'acquisition de biens immobiliers.

Du point de vue de la physiographie, l'île présente un relief fortement différencié, caractérisé par plusieurs anciennes surfaces d'aplanissement qui s'échelonnent du Paléozoïque au Cénozoïque. Les surfaces les plus anciennes sont des paléoplaines ou des discordances exhumées qui se prolongent vers le large et qui sont datées d'après les roches sus-jacentes. Une paléoplaine de ce type date du Carbonifère et constitue une facette sur les hautes terres du nord-est; une autre, d'âge jurassique/ crétacé, forme la côte sud en pente douce. Trois surfaces plus récentes sont des pénéplaines d'âge cénozoïque. La plus ancienne date du Tertiaire moyen à tardif et s'est constituée principalement sur des roches cristallines résistantes. Représentée par les collines et hautes terres à sommet plat des hautes terres de l'Atlantique, elle se trouvait à l'origine au niveau de la mer ou presque, mais elle a été basculée vers le haut vers le nord-ouest, de sorte que son altitude actuelle passe de 100 m, au sud, à plus de 500 m, au nord. Localement, elle conserve une saprolite fortement altérée et une arène profondément désagrégée contenant des boules de granite considérées comme des vestiges de sols tertiaires. À l'échelle régionale, elle montre un léger bombement en forme de dôme au-dessus de la région du détroit de Cabot, qui pourrait être attribué au relèvement du terrain par suite de la décharge causée par l'érosion. Deux éléments tectono-géomorphologiques notables sont les dépressions présentes dans le granite à proximité des basses terres dont le sous-sol est constitué de gypse, ce qui pourrait indiquer que les hautes terres sont un mince lambeau de charriage qui s'enfonce dans les roches karstiques sous-jacentes. Une deuxième pénéplaine est emboîtée dans les hautes terres et s'est formée sur des roches sédimentaires intermédiaires non résistantes au cours d'une période ultérieure d'aplanissement. Elle forme un complexe de basses terres appartenant à la plaine des Maritimes. Un troisième niveau d'érosion est représenté par des chenaux sous-marins à fond plat qui ont été entaillés jusqu'à environ -400 m lorsque les niveaux marins étaient plus bas au Tertiaire tardif-Quaternaire précoce. À l'exception des bassins de plus de 200 m de profondeur du lac Bras d'Or (et de la plate-forme continentale), qui ont été creusés par les glaces dans des roches semi-consolidées, les modifications géomorphologiques ont été relativement mineures au Quaternaire, se limitant à l'érosion fluviatile et au rabotage glaciaire local des hautes terres et à l'accumulation de quantités volumineuses de till dans les basses terres.

Surficial materials vary greatly in lithology, texture, and distribution depending on elevation, relief, and rock type. Sixteen map units are differentiated. Exposed bedrock and forested bedrock areas (map units Ra and Rb) cover about 30 per cent of the island and are the product of areal glacial scouring and occur mainly on upland summits and in some southern coastal areas. Residuum or regolith (map unit 1) covers about 10 per cent of the island and comprises disintegrated bedrock (grus) with corestones; one occurrence of intensely rotted bedrock (saprolite) was found. The main areas occur on upland summits, notably Mabou Highlands and a wide swath along the axis of the Cape Breton Highlands, where the pre-Quaternary surface form is preserved because of a virtual lack of glacial erosional and depositional effects. The residuum is believed to be a relict of weathering associated with a pre-Quaternary land surface that has been preserved because of limited glacial action due to possible cold-based ice conditions.

Quaternary deposits which predate the last glaciation comprise a suite of littoral and organic beds generally occupying gypsum karst and generally overlain by one or more tills. They are exposed along the coast, in buried valleys, and in drillholes. The littoral beds, mainly well-rounded sand and gravel, overlie and are laterally associated with an emerged rock bench that is identical in every respect to the modern intertidal bench. Its elevation, measured from the upper limit of the modern intertidal bench (i.e., mean high tide level), shows a distinct tilt from 2 m in the south to 7 m in the north. The bench is a paleoshoreline that marks the culmination of a marine transgression believed on stratigraphic grounds to be associated with the warmest episode of the last interglaciation (substage 5e). The age assignment is given by the fact that it (or its associated gravel) is locally overlain in about a dozen places by organic beds which are beyond the limit of $^{14}C$ dating, and which yield uranium/thorium age estimates of 126 ka to 47 ka, and can thus be reliably assigned to the last interglaciation (Sangamonian time) and possibly also to early parts of the last glaciation (Wisconsinan time). Pollen and macrofossil analyses reveal several vegetation cycles representing a fluctuating climatic deterioration lasting from the Sangamonian climatic optimum to the early part of the last glacial period. In the interglacial group (75 ka), the first cycle, with thermophilous hardwood forest indicating a warmer climate than today, climaxed 125 ka and is correlated with oxygen isotope substage 5e; a dubious second cycle (possibly represented at only one or two sites) with boreal forest/tundra/periglacial vegetation indicating a cooler-than-present climate is tentatively correlated with substage 5c (105 ka); a third with boreal forest like today dates about 86 ka and is referred to substage 5a. The rarity of reliable substage 5c beds may indicate that the area was glacier covered or the climate too rigorous for organic accumulation. Several subtill organic beds dating K ka (Wisconsinan time) record

La lithologie, la texture et la distribution des matériaux de surface varient sensiblement selon l'altitude, le relief et le type de roche. On a identifié seize unités cartographiques. Les secteurs de substratum rocheux exposé et les secteurs boisés (unités cartographiques Ra et Rb) couvrent environ 30 % de la superficie de l'île et sont le produit de l'affouillement glaciaire; ils se rencontrent principalement sur les sommets des hautes terres et dans certaines régions côtières méridionales. Des éluvions ou du régolite (unité cartographique 1) couvrent environ 10 % de la superficie de l'île et se composent de roches désagrégées du substratum rocheux (arène) avec des boules de granite; on a reconnu un exemple de roche très fortement décomposée (saprolite). Les principaux secteurs se rencontrent au sommet des hautes terres, notamment sur les hautes terres de Mabou et le long d'une vaste bande de terrain qui suit l'axe des hautes terres du Cap-Breton, où la topographie préquaternaire a été conservée en raison de l'absence virtuelle des effets de l'érosion et de la sédimentation glaciaires. On estime que les éluvions sont des vestiges de l'altération associée à une surface terrestre préquaternaire qui a été conservée en raison de l'action glaciaire limitée, sans doute à cause de la présence de glaces à base froide.

Les dépôts du Quaternaire qui sont antérieurs à la dernière glaciation comprennent une suite de lits littoraux et organiques qui reposent généralement sur du karst gypseux et que recouvrent un ou plusieurs tills. Ils sont visibles sur les côtes, dans les vallées enfouies et dans les trous de sondage. Les lits littoraux, qui se composent surtout de sables et graviers bien arrondis, recouvrent une banquette rocheuse émergée et y sont latéralement associés; la banquette est identique à tous égards à la banquette intertidale actuelle. Son altitude, mesurée à partir de la limite supérieure de la banquette intertidale moderne (c'est-à-dire la ligne normale des hautes eaux), passe d'environ 2 m, au sud, à plus de 7 m, au nord. La banquette est une ancienne ligne de rivage marquant l'apogée d'une transgression marine qui, pour des raisons de stratigraphie, semblerait associée à l'épisode le plus chaud du dernier interglaciaire (sous-étage 5e). L'âge est attribué en fonction du fait que cette banquette (ou le gravier qui lui est associé) est recouvert localement en une douzaine d'endroits environ par des lits organiques qui se situent au-delà de la limite de datation par le $^{14}C$, et dont la datation par U/Th fournit des estimations de l'âge allant de 47 ka à 126 ka et qui, par conséquent, peuvent être attribués avec fiabilité au dernier interglaciaire (Sangamonien) et peut-être aussi à des portions initiales de la dernière glaciation (Wisconsinien). Les analyses palynologiques et les macrofossiles indiquent plusieurs cycles de végétation qui représentent une détérioration fluctuante du climat ayant duré de l'optimum climatique du Sangamonien jusqu'à la portion initiale de la dernière période glaciaire. Dans le groupe interglaciaire (75 ka), le premier cycle, avec une forêt thermophile de feuillus indiquant un climat plus chaud que celui de nos jours, a atteint son point culminant il y a environ 125 ka et se laisse corréler avec le sous-étage 5e de la stratigraphie isotopique de l'oxygène; un deuxième cycle douteux (peut-être représenté en un ou deux sites seulement) avec une végétation de forêt boréale/toundra/région périglaciaire indiquant un climat plus frais que celui de nos jours, est provisoirement corrélé avec le sous-étage 5c (105 ka); un troisième cycle avec forêt boréale de même type que celui de nos jours, daté d'environ 86 ka, appartient au sous-étage 5a. La rareté de lits appartenant avec certitude au sous-étage 5c pourrait indiquer que la région a été couverte par des glaciers ou que le climat était trop rigoureux pour toute accumulation organique. Plusieurs lits organiques sous le till, qui datent de three tundra/ boreal forest alternations which reflect a climate that gradually deteriorated to tundric periglacial character until glaciers arrived on the island 62 ka BP, and locally as late as 47 ka BP. The intervening periods are represented only by mineral sediment, some of which is colluvial and may therefore indicate that conditions were unsuitable for organic production and preservation.

The last glaciation is recorded by erosional and depositional effects. Erosional markings on bedrock include several crosscutting sets of striations and intersecting glacially stossed facets; the older of these are stained with iron and manganese hydroxides, whereas the younger ones are not. More than 600 striation measurements help document the glacial sequence. The depositional record consists of stacks of lithologically dissimilar tills which cover about 60 per cent of the area and were produced by a sequence of several different glacier movements from quite different directions, and by glaciofluvial deposits that record the general direction of ice retreat. For the purposes of mapping, the tills are grouped together and classified according to thickness and extent. Till veneer (map unit 2b) and discontinuous till (map unit 2c) veneer are sandy stony tills, generally locally derived mainly from the more resistant crystalline rocks, on the highlands and lowlands alike. Till blanket (map unit 2a) which is defined as mantle more than 5 m thick, is relatively fine grained, having developed largely from the fine clastic rocks and evaporites that underlie the lowlands, and occurs in two main phases. One is a single sheet of till, chiefly composed of comminuted Carboniferous redbeds; the other comprises layers of different tills, produced by successive advances of the last glaciation. Two notable features of the blanket are drumlins along the East Bay and Gulf of St. Lawrence coasts carved from preexisting stratified sediments and a series of giant till ridges ranging up to 50 m high, 1 km wide, and 30 km long. The ridges were produced by an early movement and were successively modified by later flows. The succession of ice movements has produced complex superposition of drumlins and fluting, the cross-cutting relationships of which, together with the striation evidence, allow the glacial sequence to be reconstructed. Degree of glacial erosion of bedrock varies from nil over the preglacial residual terrain to extreme in glacial troughs and cirques. The provenance of the tills varies from locally derived island rocks to distantly derived from northern and marine assemblages.

Glaciofluvial deposits (map units 3a and 3b) are relatively restricted, covering only 3 per cent of the island. They include ice-contact complexes on the lowlands, the four largest being near Sydney, Mira, Inverness, and Aspy Bay. Extensive outwash is largely restricted to the major highland valleys, chiefly those of the Middle and Margaree rivers. Generally well sorted and composed of stable crystalline rocks, these deposits serve to help outline the pattern of final glacier retreat.

K ka (Wisconsinien), témoignent de trois alternances de toundra/forêt boréale reflétant un climat qui s'est graduellement détérioré en prenant un caractère périglaciaire de toundra jusqu'à l'arrivée des glaciers dans l'île il y a 62 ka, et localement à une date aussi récente que 47 ka. Les périodes intermédiaires ne sont représentées que par des sédiments minéraux, dont certains sont de type colluvial et peuvent par conséquent indiquer que les conditions ne convenaient pas à la production et à la conservation de la matière organique.

Des phénomènes d'érosion et de sédimentation témoignent de la dernière glaciation. Les marques d'érosion que l'on trouve sur le substratum rocheux comprennent plusieurs ensembles entrecroisés de stries et de facettes sculptées par la glace qui se recoupent; les plus anciennes sont tachées d'hydroxydes de fer et de manganèse, contrairement aux plus récentes. On a mesuré plus de 600 stries pour documenter la séquence glaciaire. La colonne sédimentaire comporte des empilements de tills lithologiquement dissemblables qui couvrent environ 60 % de la région et qui sont le produit d'une succession de plusieurs mouvements de glaciers différents provenant de directions très différentes, et de dépôts fluvioglaciaires qui témoignent de la direction générale du retrait des glaces. Aux fins de l'établissement des cartes, on a regroupé les tills et on les a classés selon leur épaisseur et leur étendue. Le placage de till (unité cartographique 2b) et le placage de till discontinu (unité cartographique 2c) représentent des tills sablo-pierreux qui, en général, sont dérivés localement, principalement des roches cristallines plus résistantes, sur les hautes terres comme sur les basses terres. La couverture de till (unité cartographique 2a) qui est définie comme constituant un manteau de plus de 5 m d'épaisseur, a une granulométrie relativement fine et s'est constituée largement à partir des évaporites et des roches clastiques fines qui constituent le sous-sol des basses terres; elle se présente en deux grandes phases : l'une est une simple nappe de till, composée principalement de couches rouges fortement fragmentées du Carbonifère; l'autre englobe des couches de divers tills, produites par des avancées successives de la dernière glaciation. Deux détails notables de la couverture de till sont des drumlins en bordure de la baie East et du golfe du Saint-Laurent, qui sont sculptés dans des sédiments stratifiés préexistants, et une série de crêtes de till géantes pouvant atteindre 50 m de hauteur, 1 km de largeur et 30 km de longueur. Ces crêtes ont été produites par un mouvement initial et modifiées par la suite par des écoulements subséquents. La succession des mouvements des glaces a donné lieu à une superposition complexe de drumlins et de cannelures, dont la nature entrecroisée de même que les indices fournis par les stries permettent de reconstituer la séquence glaciaire. Le degré d'érosion glaciaire du substratum rocheux varie de zéro sur le terrain résiduel préglaciaire à extrême dans les cirques et auges glaciaires. Les tills ont une provenance variée et peuvent être dérivés localement d'îlots rocheux ou bien provenir d'assemblages septentrionaux et marins distants.

Les dépôts fluvioglaciaires (unités cartographiques 3a et 3b) sont relativement limités et ne couvrent que 3 % de l'île. Ils comprennent dans les basses terres des complexes juxtaglaciaires dont les quatre plus vastes se trouvent près de Sydney, de Mira, d'Inverness et d'Aspy Bay. Les vastes épandages fluvioglaciaires se limitent largement aux grandes vallées des hautes terres, notamment celles des rivières Middle et Margaree. Ces dépôts sont généralement bien granoclassés et composés de roches cristallines stables, et ils aident à délimiter la configuration du retrait final des glaces.

Glaciolacustrine deposits, as mapped (map unit 4), are limited to small bodies of muddy, poorly sorted gravel deposited as deltas in a network of late glacial ice-dammed lakes, chiefly in Inhabitants and River Denys lowlands. Small areas of lake-bottom silt (not mapped) are associated with these deltas. A problematical patchy surficial layer of red stony silt occurs on the western coastal lowlands which might represent an outburst event from this lake system. Also included in this group are subsurface lake sediments in River Denys and East Bay lowlands, relating to analogous proglacial lakes dating to the initial Wisconsin advance. The total area of these deposits is less than one per cent of the island.

Les dépôts glaciolacustres, tels qu'ils sont cartographiés (unité cartographique 4), comprennent des masses peu étendues de graviers boueux et mal triés qui se sont déposés sous la forme de deltas dans un réseau de lacs de barrage glaciaire tardiglaciaires, principalement dans les basses terres d'Inhabitants et de River Denys. De petites étendues de silts lacustres (non cartographiées) sont associées à ces deltas. Une couche superficielle sporadique problématique de silt pierreux rouge se rencontre dans les basses terres côtières occidentales et pourrait correspondre à une débâcle survenue dans ce réseau de lacs. Ce groupe comprend aussi des sédiments lacustres de subsurface dans les basses terres de River Denys et de la baie East, qui proviennent de lacs proglaciaires analogues remontant à l'avancée initiale des glaces du Wisconsinien. La superficie totale de ces dépôts représente moins de 1 % de la superficie de l'île.

Fluvial deposits are those sediments deposited after glacier ice left a watershed. It thus includes all of the alluvium deposited up to the present and is divided into two main groups. Paraglacial and postglacial fans and terraces (map unit 5a) lie at elevations intermediate between outwash terraces and above modern floodplains. The paraglacial sediment is generally a muddy, angular gravel with only fairly developed stratification, believed to relate to the earliest postglacial time when surfaces were poorly vegetated and the climate severe. Notable paraglacial fans occur in River Denys lowland at the foot of Creignish Hills. Postglacial terrace sediment is mud free, better sorted, and stratified. Modern alluvium (map unit 5b) includes present day floodplains, active fans, and marine and lake deltas. The sediment is generally clean, well stratified, sandy, fine gravel.

Les dépôts fluviatiles sont les sédiments qui se sont déposés après que les glaces de glacier aient quitté un bassin versant. Ils comprennent ainsi toutes les alluvions qui se sont déposées jusqu'à l'époque actuelle et se laissent subdiviser en deux grands groupes. Les cônes et terrasses paraglaciaires et postglaciaires (unité cartographique 5a) se trouvent à des altitudes entre les terrasses d'épandage fluvio-glaciaire et au-dessus des plaines d'inondation modernes. Les sédiments paraglaciaires sont généralement des graviers boueux anguleux qui ne présentent qu'une stratification faiblement développée, sans doute associés au tout début de l'époque postglaciaire lorsque les surfaces ne portaient qu'une maigre couverture végétale et le climat était rigoureux. On rencontre des cônes paraglaciaires notables dans les basses terres de River Denys au pied des collines Creignish. Les sédiments des terrasses postglaciaires sont stratifiés, libres de boues et mieux granoclassés. Les alluvions modernes (unité cartographique 5b) comprennent les plaines d'inondation actuelles, les cônes alluviaux actifs et les deltas marins et lacustres. Les sédiments sont généralement des graviers sableux fins et bien stratifiés.

Marine deposits, as mapped (map unit 6), cover less than one per cent of the island and comprise only modern littoral deposits, including beaches, barriers, spits, and tombolos. Virtually all of the littoral deposits are derived by erosion of till and glacial alluvium; few are due to direct erosion of bedrock. The beach complexes typically consist of multiple parallel storm ridges, of which the older ones are partly submerged as a consequence of rising sea level.

Les dépôts marins, tels que cartographiés (unité cartographique 6), couvrent moins de 1 % de l'île et ne comprennent que des dépôts littoraux modernes, notamment des plages, des barrières, des flèches et des tombolos. Pratiquement la totalité des dépôts littoraux proviennent de l'érosion des tills et des alluvions glaciaires; quelques-uns viennent de l'érosion directe du substratum rocheux. Les complexes de plage se composent typiquement de multiples crêtes parallèles de tempête dont les plus anciennes sont partiellement submergées par suite de la montée du niveau de la mer.

Colluvial deposits, which cover about 4 per cent of the island, comprise material that has accumulated on the lower parts of slopes as a result of gravity transfer, either by sudden mass movement or slow creep. Three groups are recognized. Talus (map unit 7a) is blocky to cobbly rubble generated by frost-induced failure of bedrock cliffs which has accumulated in steeply inclined beds by the processes of rockfall, landslide, and avalanche in fans and aprons, chiefly on the slopes of the myriad deep gorges that dissect the upland and highland margins. Solifluction valley-fillings (map unit 7b) are low-angle aprons of soliflucted till and interbedded rubble on the lower parts of some slopes which are poorly understood, but are believed to have been shed by slumping and solifluction of unstable till masses and by rockfall from outcrops on the upper parts of the

Les dépôts colluviaux, qui couvrent environ 4 % de l'île, comprennent des matériaux qui se sont accumulés sur les portions inférieures des versants par suite de glissements par gravité causés soit par de soudains mouvements de masse, soit par un lent mouvement de reptation. Trois groupes sont reconnus. Les talus d'éboulis (unité cartographique 7a) se composent de débris blocailleux à caillouteux produits par la rupture de falaises rocheuses due au gel et accumulés dans des lits fortement inclinés à la suite d'écroulements de pierres, de glissements de terrain et d'avalanches dans les cônes et les glacis d'épandage, principalement sur les versants des myriades de gorges profondes qui ont disséqué les marges des hautes terres. Les remblais de vallées produits par solifluxion (unité cartographique 7b) sont des glacis d'épandage faiblement inclinés de till et de blocailles interstratifiées soliflués, entraînés par solifluxion, qui se rencontrent sur les portions inférieures de certains versants; leur existence s'explique mal, mais ils ont vraisemblablement été slope. Sheetwash aprons (map unit 7c) are a category of gently inclined, poorly stratified, sandy rubble, chiefly found on the northern highland margins and overlying the interglacial marine bench, which is believed to have been deposited mainly by torrential streams delivering a reworked mixture of till and talus. Other notable features of gravity failure are landslides and sagging slopes. In addition to innumerable small debris slides in highland gorges, one rockfall, seemingly the largest in the region, occurs on Aspy scarp. It is 80-90 years old by dendrochronology and may have occurred as a result of the 1904 earthquakes. There are two examples of massive failure of bedrock slopes by deepseated creep called sagging. Measuring several kilometres long by hundreds of metres thick, one is at Cap Rouge in steeply inclined slopes that are being undercut by active marine erosion. The other, in massive crystalline rocks at High Capes, is failing possibly because the crystalline terrane is a thrust sheet overlying weaker sedimentary strata.

Organic deposits (map unit 8) cover about 3 per cent of the island and consist of vegetal matter that has accumulated in bogs, fens, swamps, and coastal salt marshes. Bogs typically occupy depressions in till and rock and are composed of sphagnum peat up to 10 m thick. Fens are 1-2 m thick meadows of sedge grasses localized on seepage slopes, usually on the highlands. Swamps are stagnant water areas of reeds and rushes; the largest are along Inhabitants River. Salt marshes are located in the sheltered inner reaches of estuaries and consist of salt-tolerant vegetation that has accumulated in the intertidal zone as a consequence of rising sea level. They are encroaching on the living forest and commonly have fossil tree stumps projecting from them. Eeolian deposits (not mapped) consist of small areas of active and forested dunes composed of sand blown ashore from the larger beaches. Too small to show on the map, they are nonetheless economically important because of their scenic and recreational attributes.

The Cenozoic history is reconstructed from the geomorphological, stratigraphic, and chronological data. It involves a sequence of pre- and early Quaternary regional crustal movements that resulted in periods of erosional planation, a detailed account of changing climate and marine transgression during the last interglaciation, a complex succession of glacier movements during the Wisconsin glaciation, a unique late glacial climatic reversal, and a strong sea level rise during the Holocene. Pre- and Early Quaternary time is represented only by rare paleosols. Otherwise, the record is largely erosional: two inset peneplains and a network of submarine lowlands which represent periods of planation during crustal stability, punctuated by

déposés par l'affaissement et la solifluxion de masses instables de till et par des écroulements de pierres provenant d'affleurements situés sur les parties supérieures des versants. Les glacis d'épandage produits par le ruissellement en nappe (unité cartographique 7c) représentent une catégorie de débris rocheux sableux, faiblement inclinés et mal stratifiés, que l'on rencontre surtout sur les marges des hautes terres du nord et sur la banquette marine interglaciaire; ils ont vraisemblablement été déposés surtout par des cours d'eau torrentiels apportant un mélange remanié de till et de talus d'éboulis. D'autres éléments notables des ruptures par gravité sont les glissements de terrain et les versants affaissés. Outre les innombrables petits glissements de débris dans les gorges des hautes terres, un écroulement de pierres, vraisemblablement le plus vaste de la région, a eu lieu sur l'escarpement d'Aspy. Il date de 80 à 90 ans selon la dendrochronologie et est peut-être un résultat des séismes de 1904. Il existe deux exemples de rupture massive de versants rocheux par reptation profonde, donc par affaissement. L'une d'elles, de plusieurs kilomètres de longueur et de plusieurs centaines de mètres d'épaisseur, se trouve à Cap Rouge dans des versants fortement inclinés qui subissent actuellement une érosion marine active. L'autre, dans des roches cristallines massives à High Capes, subit peut-être une rupture parce que le terrane cristallin est une nappe de charriage qui repose sur des strates sédimentaires moins résistantes.

Des dépôts organiques (unité cartographique 8) couvrent environ 3 % de l'île et se composent de matière végétale qui s'est accumulée dans des tourbières oligotrophes, des tourbières minérotrophes, des marécages et des marais salants. Typiquement, les tourbières oligotrophes occupent des dépressions dans les tills et la roche et se composent de tourbe de sphaignes dont l'épaisseur peut atteindre jusqu'à 10 m. Les tourbières minérotrophes sont des prairies de 1 à 2 m d'épaisseur, composées de carex qui poussent sur les versants de suintement, habituellement sur les hautes terres. Les marécages sont des zones d'eau stagnante peuplées de roseaux et de joncs; les plus vastes se situent le long de la rivière Inhabitants. Les marais salants se situent dans les zones intérieures protégées des estuaires et comportent une végétation halophile qui s'est accumulée dans la zone intertidale par suite de la montée du niveau de la mer. Ils empiètent sur la forêt vivante et souvent des souches d'arbres fossilisés émergent. Les dépôts éoliens (non cartographiés) constituent de petites étendues de dunes actives et de dunes boisées composées de sable amené par le vent à partir de plages plus vastes. Bien que trop petits pour apparaître sur les cartes, ils sont néanmoins d'importance économique en raison de leur valeur esthétique et récréative.

L'histoire du Cénozoïque est reconstituée à partir de données géomorphologiques, stratigraphiques et chronologiques. Elle comporte une séquence de mouvements crustaux remontant au Préquaternaire et au Quaternaire précoce qui ont produit des périodes d'aplanissement, un compte rendu détaillé des variations climatiques et des transgressions marines survenues pendant le dernier interglaciaire, une succession complexe de mouvements de glaciers au cours de la Glaciation du Wisconsinien, une inversion climatique tardiglaciaire exceptionnelle et une forte montée du niveau de la mer pendant l'Holocène. Le Préquaternaire et le Quaternaire précoce ne sont représentés que par de rares paléosols. Autrement, cette période de temps est largement représentée par des structures d'érosion : deux pénéplaines emboîtées et un réseau de basses terres sous-marines qui correspondent à des périodes d'aplanissement pendant que la differential uplift. The movements are linked to the overall evolution of the continental margin when the shelf was created by subsidence offshore and erosion of the hinterland. Some warping and faulting of these surfaces may be evident in the elevational variations and discontinuities with respect to adjacent mainland Nova Scotia.

Modification of the peneplains and submarine lowlands during Quaternary time varies greatly. Glacial erosion and deposition has been almost nil on parts of the highlands plateau where a Tertiary soil remains, but major on the lowlands where thick till mantles the preglacial topography and a 200+ m deep basin has been excavated under Bras d'Or Lake. Estimated total rock removed during glacial time by ice is roughly equivalent to that eroded during nonglacial time by fluvial incision and marine planation during successive interglacial transgressions: glacial basins amount to 90 $km^3$, upland and highland gorges represent 100 $km^3$, and marine planation 50 $km^3$. The similarity may reflect the general equality of glacial and nonglacial time in this area.

Last interglacial or Sangamonian time saw the climate ameliorate to maximal warmth, producing thermophilous hardwood and white pine forests over much of the island. The deposits of this period are dated to about 125 ka and therefore correlate with substage 5e, the altithermal of the last interglaciation. During or slightly after the thermal maximum, the sea rose to at least 7 m above its present level and cut an intertidal rock bench which, during subsequent regression, was mantled with beach gravel. A minor cold period ensued when only mineral sediment was deposited; glaciers may have formed on the highlands. Thereafter, the climate recovered only slightly in that the vegetation became boreal forest and forest tundra. This is dated to 105 ka and referred to substage 5c. A second cooling was followed by a greater degree of warming that supported mixed hardwood and coniferous forests. This phase, which was similar in character to present conditions, is dated to about 85 ka and correlated with substage 5a. A second marine transgression may have cut a minor bench at about 2 m which in one place is followed by cool climate beds reflecting the onset of Wisconsinan glaciation.

Wisconsinan time was dominantly glacial. An ice cap probably formed on the highlands plateau well before ice invaded the lowlands; the latter event may have been as late as 62 ka. The first major glacial event was a strong west-to-east movement across the southern lowlands which introduced massive quantities of foreign red till from the west and formed the giant till ridges. If correctly correlated with a comparable mainland Nova Scotian event, it probably stemmed from far to the west in the Appalachians. Next, ice overlapped southward onto the eastern part of the island from the direction of the Laurentian Channel; this movement may have been of

croûte restait stable, ponctuées par un soulèvement différentiel. Les mouvements sont liés à l'évolution globale de la marge continentale à une époque où la plate-forme continentale a été créée par la subsidence au large de la côte et l'érosion de l'arrière-pays. Un certain degré de gauchissement et de fracture par failles de ces surfaces est mis en évidence par les variations d'altitude et les discontinuités par rapport aux régions adjacentes de la portion continentale de la Nouvelle-Écosse.

La modification des pénéplaines et des basses terres sous-marines a varié fortement au cours du Quaternaire. L'érosion et la sédimentation glaciaires ont été presque nulles dans des parties des hautes terres où subsiste un sol d'âge tertiaire, mais elles ont été importantes dans les basses terres aux endroits où un épais manteau de till recouvre la topographie préglaciaire et où un bassin de 200+ m de profondeur a été creusé sous le lac Bras d'Or. Selon les estimations, le volume total de roche enlevé par les glaces pendant l'époque glaciaire est approximativement égal au volume enlevé pendant l'époque non glaciaire par l'érosion fluviatile et la pénéplanation marine au cours des transgressions interglaciaires successives : les bassins glaciaires représentent environ 90 $km^3$, les gorges des hautes terres, environ 100 $km^3$, et la pénéplanation marine, environ 50 $km^3$. La similarité peut refléter l'égalité générale des époques glaciaires et non glaciaires dans cette région.

Pendant le dernier interglaciaire ou Sangamonien, le climat s'est amélioré jusqu'à atteindre des températures maximales, permettant ainsi l'établissement de forêts thermophiles de feuillus et de pins blancs dans une grande partie de l'île. Les dépôts de cette période remontent à environ 125 ka et, par conséquent, se laissent corréler avec le sous-étage 5e, l'altithermal du dernier interglaciaire. Pendant ou peu après le maximum thermique, le niveau de la mer est monté jusqu'à au moins 7 m au-dessus de son niveau actuel et a entaillé une banquette rocheuse intertidale qui, pendant la régression ultérieure, était recouverte d'un manteau de graviers de plage. Une période mineure de refroidissement a suivi pendant laquelle seuls des sédiments minéraux se sont accumulés; des glaciers se sont peut-être formés sur les hautes terres. Par la suite, le climat ne s'est adouci que légèrement, et la végétation a fait place à la forêt boréale et à la toundra boisée. Cet événement remonte à 105 ka et est corrélé avec le sous-étage 5c. Une deuxième période de refroidissement a été suivie d'un plus fort degré de réchauffement qui a permis l'établissement d'une forêt mixte de feuillus et de conifères. Cette phase, qui rappelle les conditions actuelles, remonte à environ 85 ka et a été corrélée avec le sous-étage 5a. Une deuxième transgression marine a peut-être entaillé une banquette mineure à un niveau d'environ 2 m qui, en un endroit donné, est suivi de lits indicateurs d'un climat frais qui annoncent le commencement de la Glaciation du Wisconsinien.

Le Wisconsinien a surtout été une période glaciaire. Une calotte glaciaire s'est probablement formée sur les hauts plateaux bien avant que les glaces aient envahi les basses terres. Ce dernier événement a pu se produire il y a 62 ka. Le premier événement glaciaire d'importance a été un fort déplacement ouest-est à travers les basses terres méridionales, qui a déposé des quantités massives de till rouge exotique en provenance de l'ouest et formé les crêtes de till géantes. En le corrélant correctement avec un événement comparable survenu dans la portion continentale de la Nouvelle-Écosse, on constate qu'il a probablement eu son origine loin à l'ouest dans les Appalaches. Ensuite, en progressant vers le sud, les glaces ont partiellement recouvert la portion est de l'île à partir de

Laurentide origin, although no Shield indicator erratics are recognized. Iron staining of the glaciated surfaces related to these first two events, if correctly interpreted as subaerial weathering, indicates that a period of ice retreat ensued. This was followed by a strong flow northward from the Scotian Shelf which reshaped outcrops and introduced marine shells into the till. The source may have been a separate ice cap that grew on the emergent shelf, or if there was no prior retreat, it may have been the result either of flow reversal in the regional ice sheet due to drawdown by calving in Gulf of St. Lawrence or an ice dome that formed in the regional ice sheet as a result of marginal nourishment due to string cyclogenesis. Thereafter, ice flow reorganized so that it radiated from the central part of the southern lowland. If correctly correlated with events on the mainland, this occurred during Late Wisconsinan time and the Cape Breton dome was probably a continuation of the Scotian Ice Divide. At a late stage, ice impinged on the west coast, possibly as the outer limit of the Escuminac Ice Centre situated on Magdalen Shallows. It blocked west-draining intermontane valleys, formed moraines and, together with Cape Breton ice, created a series of proglacial lakes that drained northward over Chéticamp lowland, cutting the large paleochannel and depositing the stony red silt over previously deglaciated ground. Lowland ice had shrunk into Bras d'Or basin, marking its retreat with small end moraines and sidehill meltwater channels. A sudden severe climatic deterioration 11-10 ka attributed to discharge of meltwater from the glacial great lakes caused ice to stabilize or readvance locally, vegetation to thin out, and solifluction to intensify, so that moraines were formed in some highland valleys, and on the lowlands till, colluvium, and other mineral sediment was deposited over late glacial organic beds. Highland ice may have persisted until after 9 ka and formed a series of small moraines inland of Ingonish, again possibly because of a second major meltwater discharge 10-8 ka BP. The last ice remnant was a small cap about 5 km in diameter at 480 m elevation in the central part of the national park area. Small névé fields may have re-formed on the highest ground during the Little Ice Age because, according to pollen evidence, climate then was as cool as during early late glacial time.

Crustal movement and sea level change is poorly known, although some major events are clear. The attitude of the last-interglacial littoral bench gives some indication of a long-term and possibly permanent change in the relative level of land and sea over the last 125 ka. Instead of being parallel to, if not coincident with, its modern counterpart as it theoretically should be, the bench elevation ranges from 7 m in the north to as low as 2 m in the south. This inclination may have

la direction du chenal Laurentien; ces glaces ont pu avoir une origine laurentidienne, bien que l'on ne puisse reconnaître aucun bloc erratique indiquant une provenance du Bouclier. La rouille sur les surfaces englacées associées à ces deux premiers événements, si elle est effectivement due à l'altération subaérienne, indiquerait qu'il y a eu ensuite une période de retrait glaciaire. Vint ensuite un fort écoulement des glaces vers le nord à partir de la plate-forme néo-écossaise, qui a remodelé les affleurements et déposé des coquilles marines dans le till. La source était peut-être une calotte glaciaire distincte qui s'est développée sur la plate-forme émergente; autrement, s'il n'y avait pas eu de retrait antérieur, l'écoulement glaciaire aurait peut-être été le résultat soit d'une inversion de l'écoulement dans l'inlandsis régional par suite d'un abaissement de la nappe glaciaire par vêlage des glaces dans le golfe du Saint-Laurent, soit de la formation d'un dôme de glace dans l'inlandsis régional par suite d'une alimentation glaciaire marginale due à une cyclogénèse en série. Par la suite, l'écoulement des glaces s'est modifié de telle sorte qu'il rayonnait à partir de la portion centrale des basses terres méridionales. Si la corrélation avec les événements survenus sur la partie continentale était correcte, cette réorganisation remonterait au Wisconsinien tardif, et le dôme du Cap-Breton serait probablement le prolongement de la ligne de partage glaciaire de la Nouvelle-Écosse. À une étape plus tardive, les glaces ont empiété sur la côte ouest, formant peut-être ainsi la limite extérieure du Centre glaciaire d'Escuminac situé sur les bancs de la Madeleine. Elles ont bloqué les vallées intermontagnardes dans lesquelles l'écoulement s'effectuait vers l'ouest, formé des moraines et, en même temps que les glaces du Cap-Breton, créé une série de lacs proglaciaires dont les eaux s'écoulaient vers le nord sur les basses terres de Chéticamp, entaillé le vaste paléochenal et déposé le silt rouge pierreux sur un terrain antérieurement déglacé. Les glaces des basses terres s'étaient retirées dans le basin de Bras d'Or, laissant de petites moraines frontales et des chenaux d'eaux de fonte à flanc de coteau. Une dégradation climatique soudaine et marquée survenue il y a 10 ou 11 ka, attribuée à l'écoulement d'eaux de fonte des grands lacs glaciaires, a causé la stabilisation ou la réavancée locale des glaces, le démaigrissement de la végétation et l'intensification de la solifluxion, de sorte que des moraines se sont formées dans certaines vallées des hautes terres et que du till, des colluvions et d'autres sédiments minéraux se sont déposés sur des lits organiques tardiglaciaires dans les basses terres. Les glaces des hautes terres ont peut-être persisté jusqu'après 9 ka et ont formé une série de petites moraines à l'intérieur des terres relativement à Ingonish, possiblement encore en raison d'un deuxième important écoulement d'eaux de fonte il y a 8 à 10 ka. Une petite calotte glaciaire d'environ 5 km de diamètre, à 480 m d'altitude dans la portion centrale du parc national, constituait le dernier vestige de glace. De petits champs de névé ont pu se reconstituer sur les terres les plus hautes pendant le Petit âge glaciaire, car, selon les indices palynologiques, le climat était alors aussi frais que pendant le début de la période tardiglaciaire.

Les mouvements crustaux et les variations du niveau de la mer sont méconnus, mais quelques-uns des principaux événements sont évidents. La disposition de la banquette littorale du dernier interglaciaire indique dans une certaine mesure une variation de longue durée et peut-être même permanente du niveau relatif des terres et de la mer depuis 125 ka. Au lieu d'être parallèle à son équivalent moderne, sinon coïncident avec lui, comme ce devrait théoriquement être le cas, le niveau de la banquette se situe entre 7 m au nord et aussi peu que 2 m au sud. Cette inclinaison pourrait s'expliquer two explanations. It may be a real crustal tilting that could have been accomplished by the long term 5-10 cm/ka rate of regional crustal tilting. Alternatively, there could have been a change in the gravitational level of the sea (i.e., the geoid) due to changes of subcrustal mass. An additional puzzle is that the bench is apparently upthrust 15 m where is crosses Aspy Fault. Rather than a true tectonic effect, the offset is more likely to be due to deglacial release of regional crustal stress that was stored during glacial cover. Changes of relative sea level position during Wisconsinan time are largely unknown; shorelines were not registered on the island due to ice cover and the sea is inferred to have been lower than present based on offshore data. As ice retreated, sea level rose from about -110 m, transgressed the inner shelf, and apparently left trimlines at -80 m off Isle Madame and -50 m off Sydney. For the past several thousand years, sea level has been rising, drowning estuaries, causing salt marshes to encroach on the forest, and beach complexes to build outward in a series of ridges. The rate of rise has averaged 30 cm/100 a and this is borne out both by the depth of submerged features at Fortress Louisbourg, built 250 years ago, and the trend of mean sea level measured by tide gauges. The reason for the relative rise is evidently that the crust is subsiding, as documented by recent geodetic releveling which yields negative crustal movement rates of up to 80 cm per century in the area.

Economic applications of the information are varied. For mineral exploration, which is hampered by the thick and extensive till cover, knowledge of the sequence and direction of successive drift transport events is essential in designing and interpreting the results of prospecting programs based on geochemistry of surficial sediments. Reforestation of major clearcut areas, particularly on the highlands where conditions are optimal, will be necessary to regenerate stocks and surficial sediment distribution is expected to guide the planning of this effort. Reserves of good-quality sorted aggregate are minimal and much is unavailable for exploitation because of higher-value development. Planning the optimal disposition of existing stocks will be based on the deposits mapped in this study. Problems and hazards are few in the benign Cape Breton landscape; this report identifies two major concerns – coastal erosion and slope failures. Coastal erosion annually causes significant loss of real estate, along with the current and potential wealth that is generated on that land, and necessitates expensive remedial and defensive measures. This report explains the underlying cause and evaluates the relative susceptibility of different areas; this information can help rationalize investment decisions and construction plans. Slope failures, mainly in the form of debris slides in the highland gorges can impact on development and water quality; relationships of their spatial and temporal occurrence

de deux façons. Il pourrait s'agir d'un réel basculement de la croûte dû au basculement régional à long terme de la croûte qui se fait à la vitesse de 5 à 10 cm/ka. Autrement, il y a peut-être eu une modification du niveau gravitationnel de la mer (c'est-à-dire du géoïde) par suite de modifications de la masse subcrustale. Un autre problème est que la banquette a subi un soulèvement apparent de 15 m à son point de recoupement avec la faille d'Aspy. Au lieu d'être une effet tectonique véritable, ce décalage est sans doute davantage dû à une réduction des contraintes crustales régionales par suite de la déglaciation, contraintes qui s'étaient accumulées sous la couverture glaciaire. Les changements de la position relative du niveau de la mer au cours du Wisconsinien sont généralement peu connus; les lignes de rivage ne sont pas visibles dans l'île en raison de la couverture glaciaire, et le niveau de la mer était vraisemblablement plus bas qu'actuellement selon des données obtenues dans les régions extracôtières. Au fur et à mesure du retrait des glaces, le niveau de la mer est monté à partir d'environ -110 m, a envahi la plate-forme interne et a vraisemblablement laissé des lignes de contact à -80 m au large de l'île Madame et à -50 m au large de Sydney. Depuis plusieurs milliers d'années, le niveau de la mer monte, noyant les estuaires, faisant empiéter les marais salants sur les forêts et favorisant la formation vers l'extérieur de complexes de plage en une série de crêtes. La vitesse de montée de la mer a été en moyenne d'environ 30 cm/100 ans, et c'est ce que confirment à la fois la profondeur des structures submergées à l'emplacement de la forteresse de Louisbourg, édifiée il y a 250 ans, et la tendance du niveau moyen de la mer telle que mesurée avec des maréomètres. La raison de la montée relative du niveau de la mer est de toute évidence la subsidence de la croûte, telle que documentée par de récentes et nouvelles mesures de nivellement de précision qui indiquent des mouvements crustaux négatifs atteignant jusqu'à 80 cm par siècle dans la région.

Les applications économiques de cette information sont diverses. En ce qui concerne la prospection minérale, qui est rendue difficile par l'épaisse et vaste couverture de till, il est essentiel de connaître la séquence et la direction des épisodes successifs de transport des sédiments glaciaires avant de concevoir et d'interpréter les résultats de programmes de prospection basés sur la géochimie des sédiments superficiels. La reforestation d'importantes zones défrichées, notamment sur les hautes terres où les conditions sont optimales, sera nécessaire pour régénérer des réserves, et l'on s'attend à ce que la distribution des sédiments superficiels nous guide dans la planification de cet effort. Les réserves de granulat trié de bonne qualité sont minimales et en grande partie ne se prêtent pas à l'exploitation en raison d'efforts de mise en valeur plus rentables. Pour planifier la disposition optimale des réserves existantes, il faudra tenir compte des dépôts cartographiés dans la présente étude. Les problèmes et risques sont rares dans les paysages tranquilles du Cap-Breton; dans le présent rapport, on identifie deux grands problèmes – l'érosion des côtes et les ruptures de pentes. L'érosion des côtes cause chaque année des pertes notables de biens immobiliers, ainsi que des richesses existantes et potentielles que génèrent ces terres, et des mesures correctives et défensives coûteuses s'imposent. Dans le présent rapport, on explique la cause fondamentale de cette situation et l'on évalue la susceptibilité des différentes régions; cette information aide à rationaliser les décisions en matière d'investissement et les plans de construction. Les ruptures de pente, surtout sous forme de coulées de débris dans les gorges des hautes terres, peuvent avoir une incidence sur la mise en valeur des ressources hydrauliques et sur to hydrological changes need to be investigated. Water quality is not yet a problem, but increasingly there will be a need to develop new reserves in the subsurface and this effort will benefit from the information presented here concerning sediment sequences, particularly the existing of buried valleys and the presence of stratified sediments beneath till. Safe disposal of wastes is becoming increasingly problematical; it is suggested herein that a viable, practicable, and inexpensive solution would be to store the material in rock basins lined with clay till and taking advantage of the vast resource of unused sphagnum peat. Perhaps the greatest potential economic application of the surficial geological information generated by this study is the recognition of several wilderness areas of outstanding scenic value and biological diversity which could be set aside as park reserves. They are chosen to enhance the existing national park system, as well as to form the nucleus of a renewed provincial park initiative. Such parks could greatly increase the revenue generated by the burgeoning hospitality industry and would moreover encourage exploration of hitherto overlooked areas.

la qualité de l'eau; les relations entre d'une part les épisodes spatiaux et temporels des ruptures de pentes, et d'autre part les modifications du régime hydrologique, doivent être examinées. La qualité de l'eau n'est pas encore problématique, mais de plus en plus, il faudra mettre en valeur de nouvelles réserves dans la subsurface, et cet effort bénéficiera de l'information présentée ici sur les séquences sédimentaires, notamment sur l'existence de vallées enfouies et la présence de sédiments stratifiés au-dessous du till. L'élimination des déchets dans des conditions sûres devient de plus en plus problématique; on suggère ici une solution viable, pratique et peu coûteuse permettant de stocker les produits dans des bassins rocheux dont les parois sont garnies de till argileux et de tirer profit des vastes ressources de tourbe de sphaignes non exploitée. Il est possible que la meilleure application potentielle de l'information géologique de surface obtenue grâce à cette étude soit l'identification de plusieurs réserves naturelles de valeur esthétique et de diversité biologique précieuse, qui pourraient être préservées sous forme de parcs naturels. On les choisit pour augmenter le réseau existant de parcs nationaux, et aussi pour fonder de nouvelles initiatives sur les parcs provinciaux. Ces derniers aideraient grandement à accroître les revenus de l'industrie hôtelière naissante et encourageraient de plus l'exploration de secteurs jusqu'ici oubliés.

# INTRODUCTION

## *General statement*

From Canada's beginnings, Cape Breton Island has figured prominently for various reasons. Historically, it occupied a pivotal position in the New World because of its strategic location (Fig. 1, inset). Economically, coal deposits ensured its early entry into the industrial era. Cape Breton is well known as one of the eastern region's most popular tourist destinations because of its culture and scenic attributes. The varied landscapes derive from a diverse geological makeup and a complex geomorphic history. Geologically, Cape Breton Island gained early prominence because of its complex structure and exotic disjunct fossils, now explained by plate tectonics. Now, as detailed in this report, there is growing awareness that its Quaternary sediments and features tell much about the more recent geological history of the region.

## *Background*

The project was launched in 1969 and work began in 1970 as part of a broader program of geological mapping undertaken by the author in the Atlantic Provinces region. Interest in the island's mineral potential was being spurred by exploration for salt and gypsum, by renewed interest in existing base metal prospects, and by an initiative to foster industrial investment which led to the creation the Cape Breton Development Corporation (DEVCO). The prospect of enhanced industrial activity thus created a need for comprehensive, general-purpose data on terrain conditions and surface materials for a variety of engineering and resource-development applications. A particular requirement was for specialized information on glacial dispersal which could be applied to the discovery of buried mineral deposits by geochemical surveying and by tracing of mineralized debris.

Scientific interest in the Quaternary geology of Cape Breton Island was sparked by previous studies, some dating from a century ago, which suggested that the island held promise for helping unravel the glacial history of the whole province. Previous authors showed that Cape Breton Island had a complex overprint of glacial features (Goldthwait, 1924; Prest, 1957, 1970; Wilson et al., 1958) and Weeks (1954) drew attention to very thick and extensive drift sheets. Grant (1963, Table 1) drew attention to a red foreign till sandwiched between two locally derived tills at Capelin Cove on the south coast of the island. Airphoto reconnaissance of drift forms by the author for the 1968 Glacial Map of Canada (Prest et al., 1968) revealed that the pattern was the most complex in the region. Cape Breton Island also appeared interesting from a glaciological point of view in that Flint (1951, p. 316) implied that the highlands had been an ice-dispersal centre. Indeed, Prest and Grant (1969) concluded that the ice-flow evidence pointed mainly to the action of local glaciers, with apparently little effect by Laurentide ice. Moreover, a suite of organic deposits had been found which predated the last glaciation (Mott and Prest, 1967) and thus indicated that the Quaternary sedimentary sequence on Cape Breton Island spanned the last interglacial/glacial cycle.

## *Stratigraphic terminology*

The stratigraphic terms and age connotations used in this report (Fig. 2) generally follow current usage, with departures noted below. It is modified and adapted from that prepared by Fulton (1989) for the summary volume on the "Quaternary Geology of Canada and Greenland". Chronostratigraphic boundaries are taken from Shackleton and Opdyke (1973); those for stage 5 subdivisions are from Ruddiman and McIntyre (1981). The time span covered by Figure 2 is variably represented by the stratigraphic sequences in Cape Breton Island. Tertiary time is represented only by a saprolite paleosol and associated grus and regolith. Apart from inferences about erosional events, there is apparently no depositional record of Early and Middle Quaternary time on Cape Breton Island, except for one questionable occurrence of till-like material beneath last interglacial deposits. The Sangamonian period (sensu lato), also referred to as the "last interglaciation", is equated with all of stage 5 which consists of three warmer periods (5a, 5c, 5e) and two cool interludes (5b, 5d). It is represented by scattered subsurface occurrences, many of which are richly organic. The Wisconsinan time period comprises three cold events, termed oxygen isotope stages 4, 3, 2, and is represented almost wholly by tills, the most areally extensive units in Cape Breton Island. Holocene time saw the continuous deposition of nonglacial deposits.

## *Scope and fieldwork*

This report and the accompanying Map 1631A complete a systematic survey of the Quaternary geology of Cape Breton Island. They deal with the eastern island portion of Nova Scotia, an area of 10 295 km$^2$ or about one-quarter of the province, centred roughly between latitudes 45°30′ and 47°00′ north and longitudes 60°00′ and 61°30′ west (Fig. 1). With the exception of Margaree Island, all offshore islands were visited, including St. Paul Island in Cabot Strait.

Documentation of the surficial geological map units was by ground observation. The initial phase of the field work spanned four months in the summers of 1970 and 1971. Ground observations were made by a two-person team, mainly by means of vehicular traverse of all passable roads (see Fig. 6, below) which, because of intensive widening and other improvements underway those years, fortunately provided clean and nearly continuous exposures that have not been equalled since. Most of the coast (including Bras d'Or Lake) was studied on foot; the rest was followed by boat. Additional incidental observations, local stratigraphic studies, and traverse of new highland forestry access roads were made on an opportunity basis in 1974, 1976, 1982, 1993, and 1994. Three brief helicopter flights afforded opportunities to view less accessible areas: the northern cliffs from Aspy Bay to Chéticamp, the interior of Cape Breton Highlands National Park, and St. Paul Island in Cabot Strait.

Delineation of map units was completed by interpretation of airphotos at various scales. Major physiographic features and glacial patterns were examined in the planning stages of the project using high-altitude, 1:85 000 scale airphotos. An interesting historical sidelight is that the high-altitude

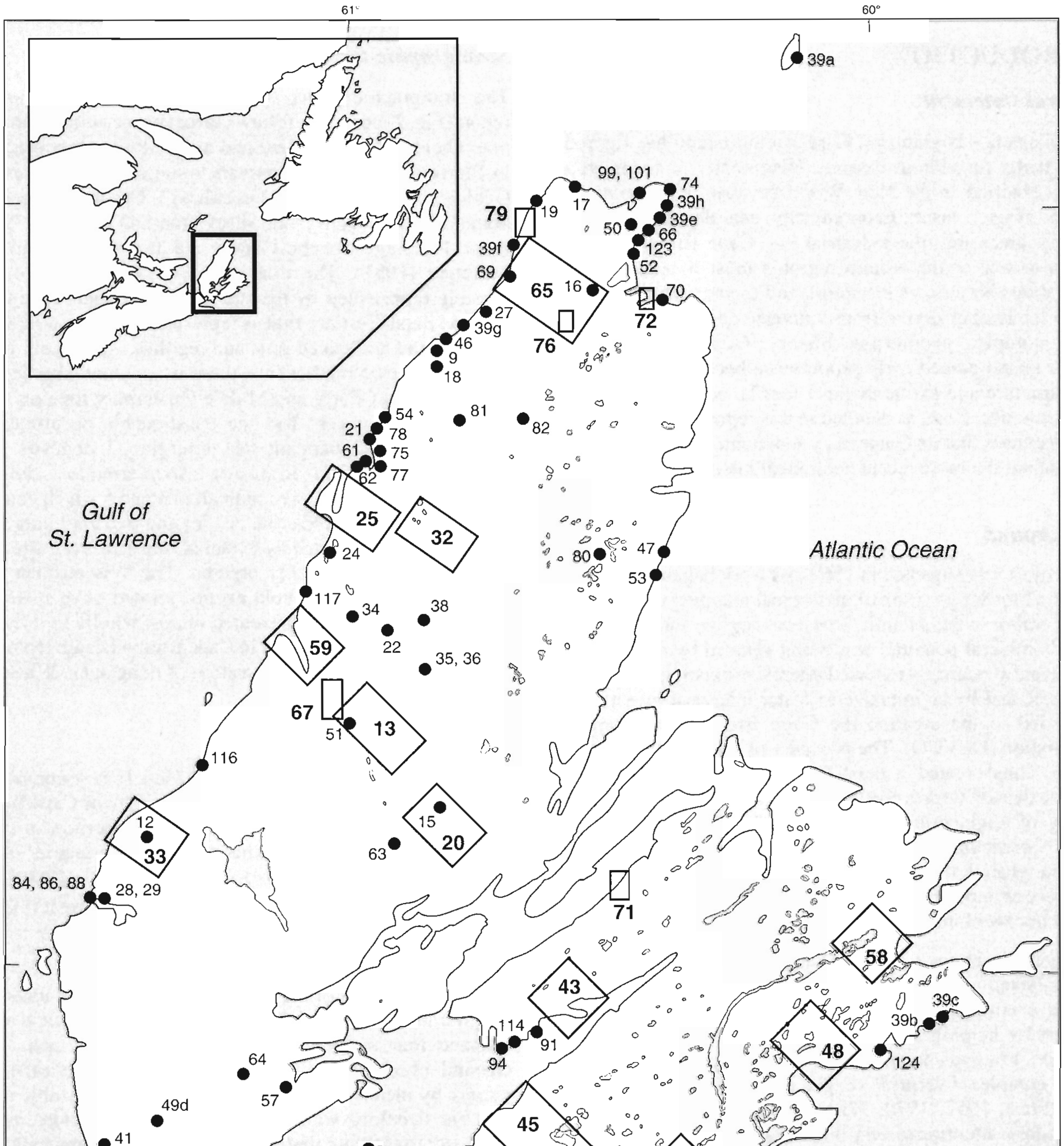



*Figure 1. Location map showing position of photo illustrations.*

photography, a special project arranged by H.L. Cameron and originally planned to be at 1:100 000 scale by military reconnaissance plane U2 of the United States Air Force, flying at 50 000 feet altitude, just before it was lost in a military action over USSR, was acquired by V3, its counterpart in the Royal Air Force under test project "Maple Scan" in 1961. Most of the interpretation, however, was done on 1:68 000 scale black-and-white coverage provided by the National Air Photo Library in 1971 especially for this project. Additional detailed work was done on 1:15 840 colour imagery intended for forest capability mapping. Its resolution enabled very small features to be detected, such as the layering and even the texture of vertical exposures.

The fieldwork brought to light numerous complex sedimentary sequences which provided the stratigraphic basis for the interpretations. Of these, 92 are considered to be key sites and their main attributes are summarized in Table 1. In this report, the site name is accompanied by the site number in square brackets (for example, Dingwall [4]) to facilitate locating them on Map 1631A (in pocket). Elevations of contacts and other features, including the emerged bench, were measured with a tape measure or aneroid altimeter, using local mean annual high tide as datum. (This tidal level is approximated within a few decimetres by the lower limit of flowering plants, and the upper limit of barnacles and fresh jetsam). To convert the measurements to mean sea level add about 1 m (the tidal range is 2-2.5 m around the island). The stratigraphy of several of the larger exposures was traced from photographs taken from a boat or helicopter. Stratigraphic study of the sequences was supplemented by paleoecological analyses mainly by R.J. Mott who was following up his earlier reconnaissance investigations (Mott and Prest, 1967). Later, the stratigraphic study was joined by collaborating teams from the Université du Québec à Montréal and the Université de Montréal who added detailed sedimentological, palynological, and chronometric analyses. This co-operative work established the age of the beds as Sangamonian and Wisconsinan and outlined the major climatic trends and the associated vegetation and glacial changes for the last 130 000 years.

## *Analytical methods*

### Chronometry

Age estimates by several methods have been applied to many surface and subsurface organic deposits and to a few scattered fossils. Surface organics have been dated by conventional $^{14}C$ methods and have yielded ages less than 14 ka. Many of the buried organic beds yielded "greater than" $^{14}C$ ages and so most of them, usually wood, were re-dated in a large counter (5 litres), at high pressure (4 atmospheres), and for periods up to 3 days. In addition, these and other wood samples have also been dated by the thorium/uranium disequilibrium method. One shell sample has been dated by the accelerator mass spectrometer (AMS) method. In total, almost 100 age determinations (Table 1, see p. 136) have been made on 39 sites, of which 34 are shown as stratigraphic sites on Map 1631A; the other five were added or studied after the map was published. The age estimates range back to 126 ka and many should be considered minimum ages. Prior to the availability of Th/U age estimates, amino-acid racemization ratios have also been determined on most of the older wood samples (Fig. 3). This varied chronometric data permit stratigraphic assignment of the deposits to various intervals of Late Pleistocene (130-10 ka) and Holocene (10-0 ka) time, according to the following convention outlined by Fulton (1984; 1989, p. 3). The Sangamonian Stage (the last interglaciation) is equated with oxygen isotope stage 5 (130-80 ka). The Wisconsinan Stage spanned the period 80-10 ka, with the Early/Middle boundary at 65 ka and the Middle/Late boundary at 23 ka. The Holocene Stage, the present interglaciation, began at 10 ka.

#### *$^{14}C$ age estimates*

With the exception of one estimate by the accelerator/mass spectrometer (AMS) method carried out at University of Toronto, most radiocarbon measurements are by the conventional method (1 litre and 2 litre counters at 2 atmospheres) and an additional few at "high pressure" using a large (5 litre) counter at high pressure (4 atmospheres) for long counts of 2 to 5 days. For these few, "greater than" $^{14}C$ age estimates have thus been extended back to >53 000 BP. Most of the age determinations related to the Quaternary geology of Cape Breton Island were initiated by this study and most were made at the Geological Survey of Canada. Most have been previously documented and interpreted by the author in the annual lists of Geological Survey of Canada radiocarbon age determinations. Additional $^{14}C$ dates were made at the University of Québec at Montréal in connection with associated studies by S. Occhietti.

Ages quoted as "finite" are in two groups, 0-14 ka and 20-44 ka. Those quoted as "nonfinite" or "greater than" are beyond the range of the method used at the Geological Survey of Canada (depending on counting time, pressure, volume, and background). Many of the nonfinite ages have been dated more than once as the method improved, and as further studies provided higher quality material in a better understood pollen-stratigraphic context.

For various reasons, a few of the ages are considered anomalous or unreliable; these are underlined in Table 1. Some of the finite ages that are near the limit of the method are considered minimal because of the likelihood of contamination by a small amount of young carbon: wood at Bay St. Lawrence [1] and Dingwall [4], shells at Grantville [47], organic silt at Big Brook quarry [49], and bone at Middle River [15]. Five of the younger finite dates in the range 12.0-24.9 ka are considered anomalous owing to contamination by old carbon (South Aspy River [5]) and by carbonates to produce the so-called "hardwater effect" (Gillis Lake [40].

#### *Age estimates by the thorium/uranium disequilibrium method*

A suite of Th/U age estimates on wood from subtill organic deposits was carried out during the developmental stages of the method by the University of Québec at Montréal, first to calibrate the method by comparing results with $^{14}C$ ages on the same samples, and then, once good correspondence was

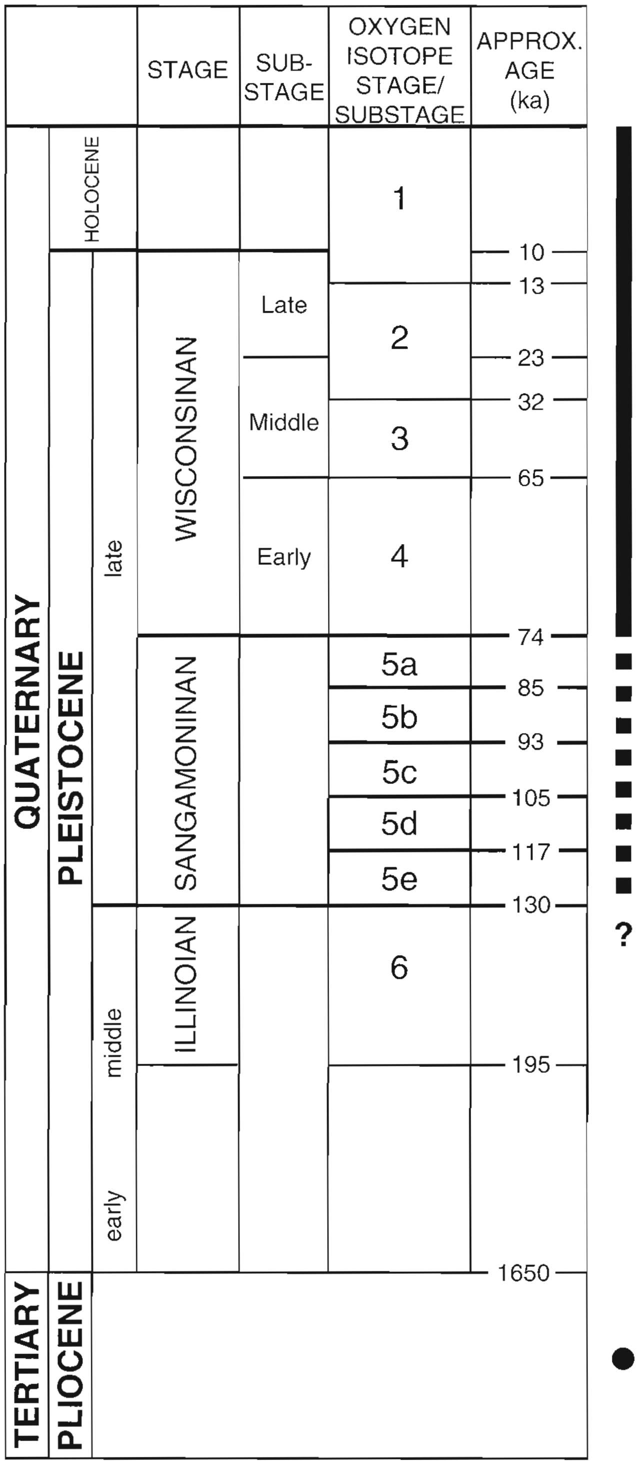



*Figure 2.* *Stratigraphic terminology and age connotations used in this report (the solid, broken and dotted line indicates the degree to which the time interval is represented by sedimentary sequences).*

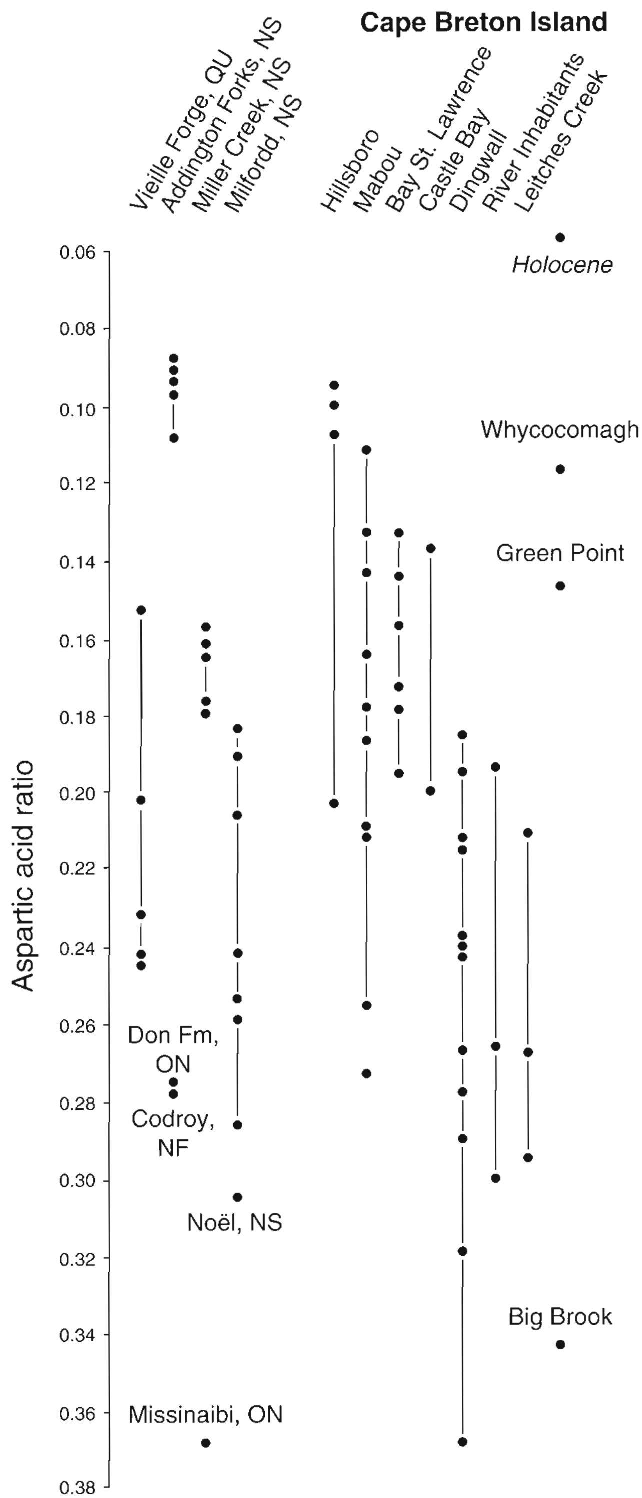



*Figure 3.* *Aspartic amino-acid racemization ratios on wood from subtill buried organic deposits in Cape Breton Island, with a selected few from analogous deposits in other regions for comparison.*

achieved, to support the pollen-stratigraphic studies. The ages range from 126 ka to 47 ka and, although quoted as finite, should be considered minimum estimates. The estimates appear reasonably consistent, apart from some minor anomalies, and only a few conflict with the chronostratigraphic assignments based on pollen analyses; however Stea et al. (1992b) have recently raised a serious challenge to the validity of these estimates based on the assumption that the geochemical/isotopic system is closed. For that reason and pending further analysis of the problem, the Th/U estimates are used in this report in a provisional way to lend support to the tentative stratigraphic assignment for the older part of the lithostratigraphic and pollen stratigraphic sequences. They provide a chronometric basis for the history of climatic fluctuations inferred from pollen analysis for the last interglaciation and the onset of the Wisconsin glaciation in Cape Breton Island.

*Amino-acid racemization ratios*

Most of the wood from the older buried organic deposits which yielded "greater than" $^{14}C$ ages was also analyzed for amino-acid ratios at the University of Alberta, courtesy of N.W. Rutter (Fig. 3). The work was undertaken to help develop the use of wood in amino-acid geochronology and was carried out before the application of Th/U dating to Cape Breton Island samples. The approach was firstly to see how the sequence of wood amino-acid ratios compared with the age differences inferred from gross stratigraphy and pollen correlation and secondly, how the Cape Breton Island results compared with those on pre-Late Wisconsinan, that is to say last interglacial, material from better understood sequences elsewhere in Canada: primarily the group of Sangamonian deposits in mainland Nova Scotia, as well as last-interglacial deposits farther afield at Codroy, Newfoundland, Vieille Forge in southern Quebec, and the Missinaibi and Don formations in Ontario. This wide areal sample also allowed a test of the sensitivity of amino-acid racemization to temperature history. Assays were thus made on known Holocene and Sangamonian wood from the Atlantic region, as well as from Quebec and Ontario where the temperature regime would be expected to have been generally higher over the same time interval. Despite some anomalies and discrepancies, the wood ratios generally parallel the stratigraphic order inferred from pollen and superposition and were in broad agreement with ages estimated by the Th/U method. Additional analyses by Occhietti and Rutter (1982) compared Cape Breton Island wood with Quebec material. Both studies suggest that all the older buried organic deposits on Cape Breton Island belong to one nonglacial period prior to Wisconsin glaciation.

### Palynology

Pollen analysis, done entirely by other workers, some of whom were allied with this project, has been performed on all but a very few of the organic beds in the many multiple-layer sequences. These are complemented by identification of animal and plant microfossils and macrofossils, including wood. In addition, a number of postglacial lake-sediment cores and a few bog sequences have been analyzed. Early pollen work focused on bog and lake sediments (Auer, 1933; Hall, 1949; Killam, 1951; Livingstone and Livingstone, 1958; Schofield and Robinson, 1960; Terasmae, 1974). Then, attention shifted to the buried organic beds (Livingstone, 1968; Mott and Prest, 1967; J.B. Railton, pers. comm., 1970; Mott, 1971; Mott and Grant, 1985; de Vernal et al., 1986; de Vernal and Mott, 1986). That work documented the vegetation succession and inferred climatic trends during the last interglaciation. Of late, palynologists are looking again at detailed postglacial climatic changes (for example, Mott et al., 1986).

## *Settlement, industry, and access*

About 250 years ago the population of indigenous Micmac was swelled by a great influx of European immigrants, mainly French, who cleared and settled chiefly on the deeper, finer-textured soils along the south and west coasts, and have largely remained in those areas. Later, Scottish crofters occupied the remaining generally thin and rocky soils of the interior and higher areas. The spread of settlement by agrarian development is reflected in the varying density of farming villages and the associated road network; the coastal villages mark safe anchorages for fishing.

The island's total population of 160 000 (Statistics Canada, 1991) is located almost entirely in the lowland parts of the island in hundreds of communities, ranging from tiny hamlets to small cities. The largest settlements are now industrially based and are centred in two areas. In the east, the 1991 census (Statistics Canada, pers. comm., 1994) shows that three-quarters of the island's population reside in the Sydney agglomeration (116 100) which ranks twenty-seventh in Canada. Based on a coal and steel industry that began a century ago, it is a group of eight towns, of which the main two are Sydney (26 063), the third largest city in Nova Scotia, and Glace Bay (19 501). North Sydney (7260) is a railhead and the terminal for the two ferry links to Newfoundland. Along Strait of Canso in the west, the towns of Port Hawkesbury (3991) and Port Hastings (362) expanded when the island was linked to the mainland in 1955 by the world's deepest causeway; the area now has heavy industry and a superport because of the deepwater access. Important tourist and service centres are Inverness (1935) and Chéticamp (979) on the west coast, Ingonish (607) on the east coast, Louisbourg (1261) on the south coast, and Baddeck (1064) on interior Bras d'Or Lake.

Cape Breton Island has a mixed economy based on natural resources, manufacturing, and service industries (chiefly tourism) which relate in a number of ways to the surficial and environmental geological themes in this report. The rural one-quarter of the population is occupied mainly with fishing and forestry. From all coastal areas a variety of marine products are harvested and most is processed locally. A fish and shellfish aquaculture has begun. Pulpwood harvesting, mostly highly mechanized, is important in the interior. Cape Breton Island's mineral industry began in 1766 with coal mining in the Sydney area where it still continues. Gypsum quarrying began with small operations at Dingwall, Chéticamp, Mabou, and Baddeck and is now concentrated at

Big Brook quarry and Little Narrows. Celestite has been worked at Enon and brineable reserves of salt and potash have been outlined in Canso Strait area. Base metal deposits occur widely and a few have been worked.

Manufacturing and service industries are concentrated in industrial complexes in two areas. Activity in the Sydney area centres on steelmaking because of local coal and fluxing limestone. At Strait of Canso, deep water and proximity to offshore oil and gas reserves have attracted North America's largest superport, crude oil trans-shipment facilities, oil refineries, a thermal power plant, a pulpwood mill, a gypsum loading terminal, a heavy water plant, and a host of service businesses.

The hospitality industry is a major economic component serving tourists and summer residents who are drawn to scenic attractions and camping opportunities along the famous Cabot Trail and a dozen other scenic byways. Major destinations include Cape Breton Highlands National Park, Alexander Graham Bell National Historic Park, and Fortress Louisbourg National Historic Park, Canada's largest historical reconstruction project, a fortified townsite of several thousand residents 1720-1760. The information contained in this report helps satisfy visitors' requirement for information on the origin and development of the landscape and thus has direct application to the tourist industry.

The Trans-Canada Highway bisects the island, and a network of mainly unpaved roads crisscrosses the lowland areas (Fig. 6) and follows virtually the entire coastline, except for precipitous segments in the northern highlands. Recently, forest haulage roads have penetrated most upland and highland areas, except for the northern third of the highlands plateau, including the national park, which remains a largely trackless wilderness.

### *Climate, vegetation, and soils*

Present-day climate provides insight into the style of past glaciation and explains the island's diverse vegetation, both now and in earlier periods. Cape Breton Island, like the rest of Atlantic Canada, has a predominantly continental climate because it lies astride the converging tracks of most major eastward-moving storms in eastern North America (Hare and Thomas, 1974). However, its maritime setting tempers the hot and cold extremes of continental air arriving from the west. It thus has a moderated seasonality; the surrounding cold ocean produces cool, short summers, long but moderate winters, and abundant and frequent precipitation. In summer, storm tracks pass more to the north and thus bring warm moist Maritime Tropical (MT) air from the south; in winter they pass to the south and draw in cold continental air from the west and north. Fine weather is associated with colder continental air from the west, whereas precipitation arrives mainly on easterly winds from the Atlantic Ocean. Elevation of the highlands is such that the northern plateau is often above cloud line.

Local climate largely reflects the proximity of the relatively cool seawater around the island. Thus, there is a strong gradient inland and monthly isotherms follow the coastline. Based on records from nine stations (Fig. 4) temperature shows a larger range than occurs in a true maritime climate (Environment Canada, 1975a, b). For example, there is an average difference of two Celsius degrees in mean daily temperature between the coast and the interior. In the interior mean daily maxima of +23°C in high summer are 5° higher than coastal sites. Conversely, in mid-winter mean daily minima in the interior drop to -23°C, more than 6° below those at coastal sites.

Local and seasonal differences are strongly influenced by sea-surface temperature, since most parts are within 30 km of the coast. In summer, the west and northeast coasts are relatively warm and dry because of 18°C-20°C water exiting Gulf of St. Lawrence compared to the east and south coasts which are chilled by a branch of the cold (10°C-15°C) Labrador Current with cold, often foggy air from offshore regularly being drawn ashore by inland heating. In winter, however, the west is colder than the east because pack ice covers the Gulf of St. Lawrence whereas the Atlantic Ocean is ice free. There are 158-175 days with frost.

Precipitation likewise varies with coastal proximity and elevation, ranging from 1600 mm on the highlands and uplands to 1200 mm on the southwestern Gulf of St. Lawrence coast. About three-quarters of this falls as rain, mostly in October and November. Snow occurs on about 100 days and exceeds 400 cm on the highlands (more on the northwestern coast), compared to 150-200 cm along the south coast. It is present for four months from early December to late April and commonly persists into June on the highlands. Fog is common along the Atlantic coast, especially in spring and early summer.

The vegetation of Cape Breton Island spans a considerable range of species from hardwood forest to heath barrens. It may be divided into five main zones (Fig. 5) which further reflect the control of elevation and coastal proximity on climate. Like the rest of the Maritime Provinces, Cape Breton Island belongs to the Acadian Forest Region which has coniferous elements of the boreal forest and deciduous species of the Great Lakes-St. Lawrence forest (Loucks, 1962; Rowe, 1972). The distribution of vegetation is determined mainly by potential evapotranspiration, with wind being a secondary factor. In connection with interpretations of past climate based on fossil vegetal deposits, the present day variations of vegetation with geographic position helps explain some of the variance in the pollen content of deposits believed to be of the same general age.

The coastal spruce-fir-bog zone is a belt 10-30 km wide bordering the Atlantic shore. It has dense, low stands of balsam fir and black spruce, interspersed with areas of heath barren, sphagnum bog with tamarack, and lichen-covered bedrock that reflect the excessive moisture and cool, foggy conditions. Blow-down and sculpted thickets of white spruce (krummholz or "tuckamore") illustrate wind stress.

An interior sugar maple-hemlock-pine zone occurs around Bras d'Or Lake and up to 150 m elevation along the east and west coasts. In this area, conifer stands are densest and broadleafed trees more abundant because temperatures are higher, precipitation is lower, and soils deeper than elsewhere on the island. Conifers (mainly white spruce and

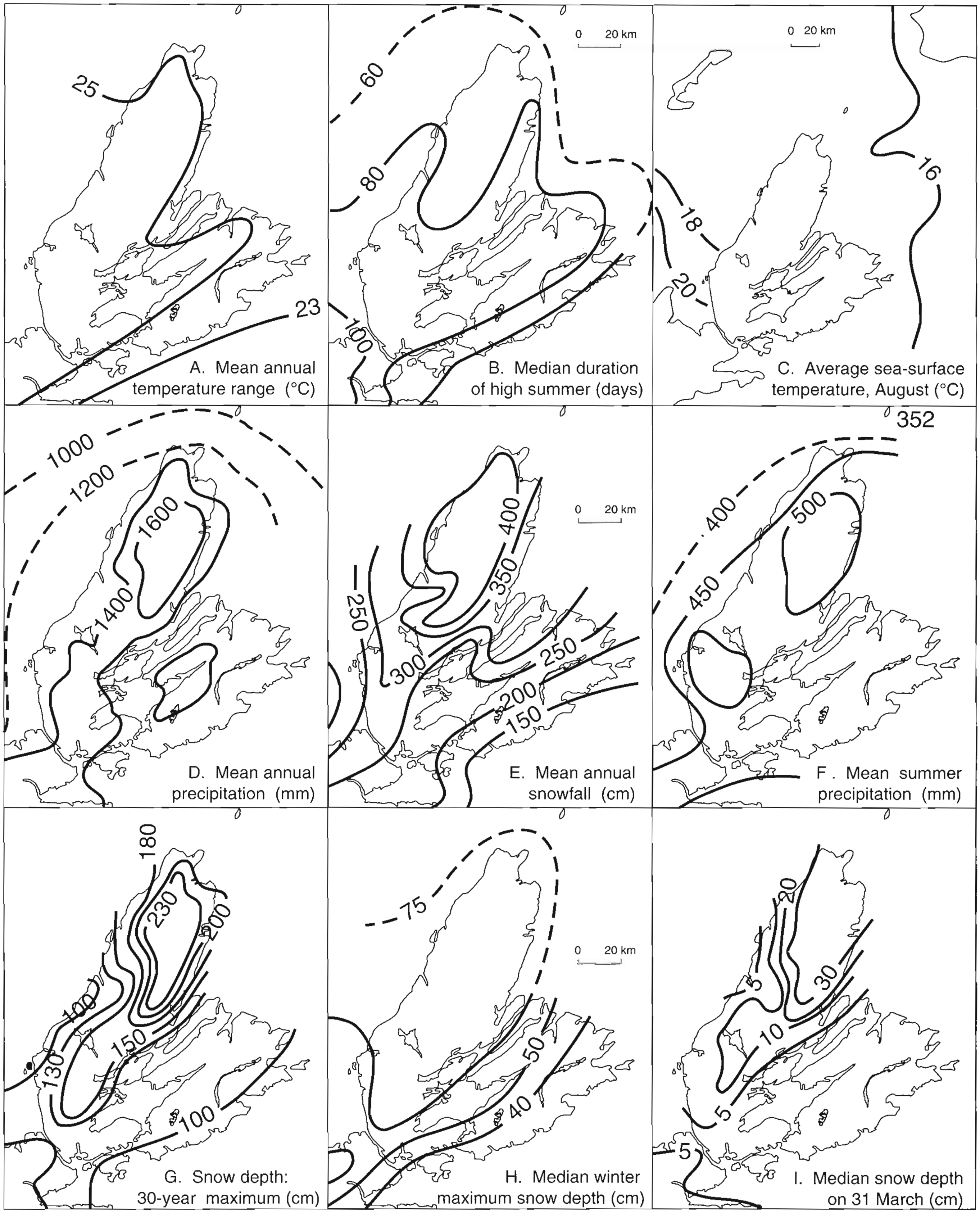



***Figure 4.*** *Selected climatic parameters illustrating general eastward gradient on which local orographic and marine influences are superimposed (adapted from Gates, 1975).*

balsam fir) dominate on heavy-textured, undulating till surfaces, on valley bottoms and lower slopes, and in ravines. White pine is associated with glacial gravel; black spruce with terraces and poor drainage.

An upland sugar maple-yellow birch-fir zone is found on hills above 150 m. On slopes and ridges, conifers give way to tolerant hardwoods such as maple, beech, and birch.

On highland slopes at 250-350 m, depending on aspect and exposure, the tolerant hardwoods are replaced by a forest with more conifers, the plateau fir-pine-birch zone. Above 400 m on the plateau, balsam fir is dominant, along with white and black spruce. Jack pine occurs on granitic rocks northwest of Ingonish where fire is prevalent. Sphagnum bogs and sedge fens occupy depressions and seepage slopes.

On the summital area above 450 m, a taiga-spruce zone indicates a climate too severe for closed forest. Stunted thickets of black spruce, deformed by wind stress, give way to heath-shrub barrens on ridges.

Soils mapping has been carried out for the whole island (Cann et al., 1963) and the classification was based primarily on parent material, thus reflecting the general distribution of the main types of Quaternary deposits. Over most of the lower

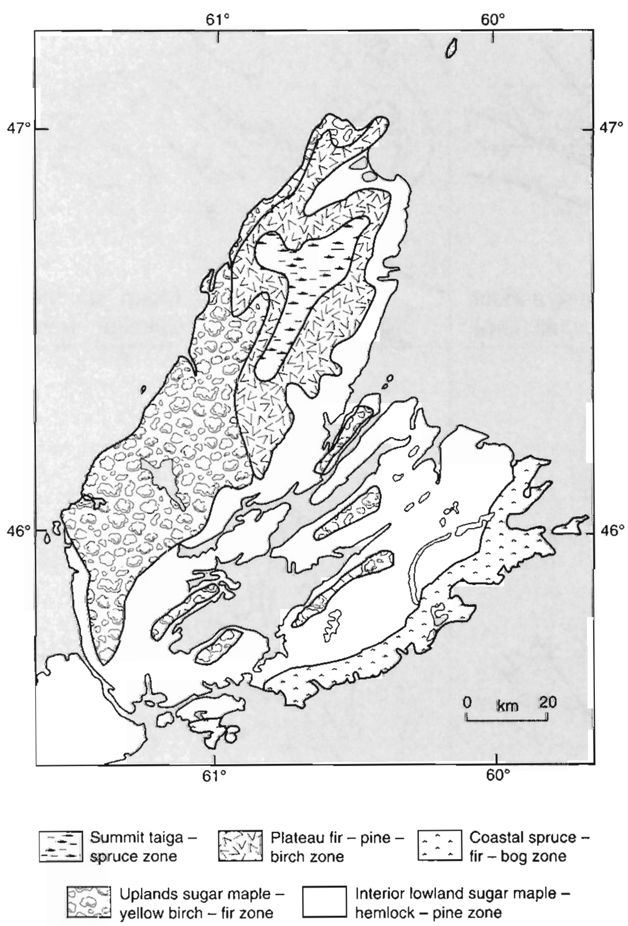



*Figure 5. Vegetation zones (adapted from Loucks, 1962; Rowe, 1972).*

parts of the island, podzols have developed in the extensive areas of till which cover much of the area and are typically shallow with strong horizonation. Locally the B zone has a cemented iron horizon, called "hardpan", which limits forest development and promotes bogs. In contrast, most of the highlands plateau has regosols which are developed in the extensive areas of variably disintegrated bedrock (residuum); small areas of till have ferro-humic podzols and gleysols; bogs and shrub-covered bedrock constitute azonal soils.

## *Previous work*

Prior to this study there were few substantive publications on any aspect of the surficial deposits and Quaternary history of Cape Breton Island. While this report was in preparation, however, interpretations have appeared for adjacent areas which bear on the history presented here. Many incidental observations and interpretations on a variety of aspects of the Quaternary geology had been reported over the past 100 years which raised important questions, posed problems, and generally suggested that Cape Breton Island had a complex history deserving systematic study.

Initial bedrock mapping a century ago provided incidental observations on glacial features and assembled striation data pointing to a great variety of ice-flow directions. Despite little fieldwork in the 1950s and growing evidence of local glaciers elsewhere in Nova Scotia (e.g. MacNeill and Purdy, 1951), overviews espoused a simple hypothesis of general southeastward flow by the Laurentide Ice Sheet which covered all but a few of the highest summits and extended onto the continental shelf (e.g. Flint, 1971, p. 318-320; Mayewski et al., 1981). The present phase of intensive stratigraphic study by several workers in various fields is developing a comprehensive paleogeographic reconstruction of changes in climate, vegetation, glaciers, and sea level. The main themes of previous work have been: geomorphological development, glacial deposits, ice-flow phases and ice extent, relative sea level changes, Quaternary stratigraphy, and the pollen record of postglacial climatic evolution.

### Previous geomorphological interpretations

Physiographic interpretation appears to have begun with Robb (1876) who attributed the elongate Bras d'Or basins to glacial gouging. Drawing attention to the upland and lowland levels, Daly (1901) regarded the summits of hard-rock uplands as remnants of a tilted peneplain and the lower levels on softer rocks as the product of later planation events. Goldthwait (1924) advanced the planation concept and described the landscape of the island in detail. He noted wind gaps as the remnants of ancient valleys consequent on the upland peneplain, inferred a Cretaceous age for the lowlands, and attributed the shallow thresholds of the Bras d'Or Lake basins to morainal dams. Norman (1935) inferred a major buried valley at Inverness. Johnson (1925) classified the entire coastline, suggesting that some segments were unconformities exhumed from beneath Carboniferous and Cretaceous cover rocks. Weeks (1954) referred to the southern coastal area as a partly submerged, warped plateau. Bird

(1972) and Brookes (1972) postulated regional correlations of the tilted degradation levels. Mathews (1975) deduced that the summit peneplain was uplifted in middle to late Tertiary time because the volume of rock eroded below the peneplain level was similar to the volume of deep sea sediment deposited since that time. McKeague et al. (1983) interpreted a mature saprolite on the peneplain as support for the preservation of an unmodified Tertiary landscape. Sanford and Grant (1976) depicted the regional physiography and proposed new divisions. Roland (1982) gave a well-illustrated popular account of landscape development. Grant (1987, 1989) outlined the regional geomorphic evolution.

**Previous glacial-geological studies**

Observations on glacial deposits and of ice-flow directions based on striations and erratics dispersal were initially part of detailed bedrock mapping surveys. Fletcher (1877, 1878, 1881) noted the northward and northeastward ice flow across the southern part of the island and drew attention to the important nonglacial sequences on the East Bay shore. Robb (1876), however, inferred a southwestward flow from the direction of Newfoundland which appears to have been based on the erroneous assumption that crystalline erratics in the Sydney area were derived from the Shield, rather than locally. Gilpin (1886) recorded diverse striae, mentioned a mastodon fossil and peat under till, and inferred that Bras d'Or Lake was glacially excavated. In the first paper devoted to the glacial geology of the island, Honeyman (1890) noted anomalous till thicknesses and large gravel complexes, but viewed all of the striations in the context of a single glaciation by Laurentide ice moving southeastward from Gulf of St. Lawrence. Wilson (1906) extrapolated Robb's (1876) suggestion and proposed that an ice-dispersal centre on Newfoundland had expanded southwestward over Cape Breton Island and as far as Cape Cod – a revolutionary view for which evidence has never been found. Goldthwait (1924), on the other hand, adhered to the Laurentide school of thought, citing the common southeastward trend of striations and postulated that a glacier had spread southward across the southern Gulf of St. Lawrence (which he called the "Acadian Bay Lobe") and crossed Cape Breton Island to terminate at the edge of the continental shelf. Significantly, he noted onshore-directed striations on the south coast and considered, but rejected, a shelf-based ice cap; he explained the trend as topographic divergence of an ice lobe from New Brunswick. Mather (1926) felt that the absence of erratics on Smoky Mountain (northern highlands) meant that it had not been overridden by glaciers. Norman (1935) was the first to report evidence that the island was first invaded by foreign glaciers, then later supported the theory of a local ice cap.

The question of the extent and elevation of glaciers was first posed by Fernald (1925) who found isolated (disjunct) assemblages of arctic/alpine plant species on Cape Breton Island summits (among others in the region) and postulated that they had survived in ice-free enclaves during the last glaciation. K.F. Mather (1926) suggested that a portion of the northern highlands plateau escaped glaciation. Wynne-Edwards (1937) termed this the "nunatak hypothesis". Flint (1940, 1951), on the other hand, postulated that Cape Breton Island was completely glaciated during the latest phase; his view went unchallenged and untested for almost 40 years. Cameron (1961) prepared the first surface materials and till lithology map of the island, largely by recasting the information gained by detailed soils mapping by Cann et al. (1963). His recorded glacial data and interpretations, known to be voluminous, were not published and virtually all manuscript material was discarded after his death, except for two summary maps (copies of which were provided to the author by the Nova Scotia Research Foundation) which he prepared for the 1958 Glacial Map of Canada (Wilson et al., 1958). One shows his inferred limit of the last interglacial high sea stand; the other shows moraines, proglacial lakes, and ice-flow data and seems to indicate that he may have held to Wilson's (1906) idea of southwestward Newfoundland ice flow across the southern coastal area. Prest and Grant (1969) and Prest et al., 1972) presented the first map of Cape Breton Island showing the great variety of ice-flow directions, most of which were gained in connection with the 1968 Glacial Map of Canada (Prest et al., 1968). They postulated a series of local ice caps, variously affected by drawdown into marine calving bays, with Laurentide ice restricted to the west coast. An outline of the Quaternary history since the last interglaciation was given by Prest (1970).

Study of Quaternary sedimentary sequences and on Cape Breton Island began with E.H. Muller who, in unpublished notes dated 1964, concluded that Laurentide ice overrode the northern highlands in Late Wisconsinan time. Grant (1963, Table 1) drew attention to a red foreign till sandwiched between two locally derived tills at Capelin Cove on the south coast of the island. For the southern and northern highlands area, a three-fold ice-flow sequence, involving local ice flows before and after the main event, was deduced from the lithology and fabric of superposed tills (Tang, 1970; Newman, 1971). These were incorporated into a five-fold sequence from superposed striations over the whole island (Grant, 1971b). Interim progress reports on this study (Grant, 1971a, b, 1972, 1974a, 1975a) gave further information on the glacial sequence. Crosscutting ice-flow indicators recorded five ice advances across the island from various directions. One of the flow events, a widespread onshore northward movement, posed glaciological problems because it seemed to require an ice cap situated on the eustatically emergent continental shelf (Grant, 1972). A distantly derived till was reported and a late ice cap was postulated for southern lowlands (Grant in Scott, 1976). Stratigraphic evidence led Grant (1976, 1977) to propose an Early Wisconsinan age for the major englaciation, and to adopt a "minimalist" ice model for Late Wisconsinan time by postulating two ice caps, on the highlands and the lowlands, which did not cover the whole island.

The concept of restricted Late Wisconsinan ice in Cape Breton Island and the Atlantic region in general has been challenged by numerous workers. Mayewski et al. (1981) discounted the ice-limit evidence and adopted Flint's (1940) "maximalist" idea. The two models were being actively debated during and after the field part of this study, and later work has still not resolved the issue. Although the "minimum concept" is still supported by geomorphological results in Newfoundland (Grant, 1992, 1993), ice modelling from sea level data points to an intermediate-sized reconstruction (Quinlan and Beaumont, 1982) and several authors, particularly those working offshore, have presented evidence pointing to a much larger Late Wisconsinan ice complex in the Appalachian region – one extending nearly, if not wholly, to the edge of the continental shelf (Wightman, 1980; Stea and Finck, 1984; Stea et al., 1986; Stea and Wightman, 1987; Gipp and Piper, 1989; King and Fader, 1990; Stea and Mott, 1990).

Attempts were made to summarize the Cape Breton Island findings and to place them in the context of global temperature variations. The three till-forming phases and two nonglacial intervals were correlated with the 23 000 year rhythm of temperature variations recorded by deep sea oxygen isotope variations (Grant and King, 1984). Based on the areal inventory of surface materials (Map 1631A, in pocket), on which the succession of flow phases was depicted, an interim summary of the glacial history of the island was presented (Grant, 1987, 1989).

Reconnaissance and detailed geochemical surveys of surficial materials have been carried out during, and subsequent to, the completion of this report. Fortescue and Hornbrook (1969) made biogeochemical assays of plant material around the Yava lead deposit in Salmon River valley; Rogers (1983) conducted regional stream sediment surveys; and Rogers and MacDonald (1984, 1986) reported on lake sediment sampling. Sangster (1988) linked surface geochemistry in the Blue Mountain area near Marion Bridge to glacial dispersal patterns. MacDonald et al. (1991) and MacDonald and Boner (1993) studied the relation between the geochemistry of the Yava [94] lead showing and the variations of minerals dispersed from it into the vegetation, humus, A-horizon soil, and the three tills comprising a composite drumlin. Detailed till provenance studies were conducted in the Gabarus area (McClenaghan and DiLabio, 1992; McClenaghan and DiLabio, in press; McClenaghan et al., 1992). These geochemical studies lend support to the sequence of glacier movements and the corresponding tills presented in this report.

During the course of this study, successive ice-flow phases have been reconstructed in neighbouring parts of the Maritime Provinces, namely for New Brunswick (Rampton et al., 1984) and Nova Scotia mainland areas (Stea and Finck, 1984; Stea et al., 1986; Stea et al., 1987; Stea and Wightman, 1987; Stea and Brown, 1989; Stea and Mott, 1990). They provide the basis for linking the Cape Breton Island findings presented in this report to the wider region.

### Previous work on relative sea level change

Paleoshorelines have attracted much attention, both in their own right, and as indicators of glacier distribution and crustal movement. For the first time in North America, the regional pattern of glacial-age shorelines was interpreted in terms of ice extent (De Geer, 1892). He inferred that the main glacier masses had lain over Newfoundland and New Brunswick-Nova Scotia and that ice was thin or absent over Gulf of St. Lawrence and much of Nova Scotia, including Cape Breton Island. Fairchild (1918) and Daly (1921) strengthened the concept of limited ice using shoreline data, adding that Nova Scotia was presently submerging because of crustal subsidence. As evidence, McIntosh (1908) cited the drowning and erosion of Fortress Louisbourg. McIntosh (1923) argued that Port Hood Island had been isolated since 1781 because of recent rapid submergence. Flint (1940) depicted a very different pattern, showing raised shorelines in Cape Breton Island (although no source was given), as support for his belief in a major expansion of Laurentide ice over the area. Recent isobase maps by Wightman and Cooke (1978) and by Grant (1980) resemble the original map by De Geer (1892) and contradict Flint (1940).

A shoreline predating the last glaciation was first noted as a "25-foot bench" on an unpublished sketch map by H.L. Cameron dated 1957. It is probably the same rock bench which Neale (1963, 1964) reported overlain by till and locally faulted at Money Point. Grant (1980) reported many additional long segments of the bench around the island, locally overlain by warm-climate organic beds and by as many as three tills. He inferred that the bench was cut at the marine transgressive maximum of the last interglaciation, and proposed it as a stratigraphic marker horizon.

Postglacial relative sea level rise and crustal subsidence was shown by submerged features and relevelling. From dating of forests being onlapped by salt marshes, Grant (1970, 1975b) found evidence of an anomalously rapid recent rise of tide level which he attributed mainly to crustal subsidence as the effect of a migrating collapsing glacier-marginal bulge. Kranck (1972) mapped submerged shorelines in George Bay and inferred a differential subsidence of Cape Breton Island. From geodetic relevelling, Vanícek (1975) gave corroborative evidence of rapid modern sinking of the western part of Cape Breton Island. From sea level data, Quinlan and Beaumont (1982) modelled a former glacier cover that was intermediate in extent between the "minimum concept" and the "maximum concept". De Vernal et al. (1985) and de Vernal and Jetté (1987) determined the time sea water overflowed the sills of Bay St. Lawrence and Bras d'Or Lake and extrapolated submerged rates that are similar to adjacent areas. Sea level changes in the Cape Breton Island region have been outlined by Grant (1987, 1989).

### Earlier studies of subtill nonglacial deposits

Quaternary stratigraphic studies perhaps had their beginning in the first treatise on the geology of (Lower) Canada by Dawson (1855; 1868, p. 63) who mentioned a peat bed beneath "drift" (till). Though not relocated until the present study began, it was the first discovered of many buried organic beds which have since served as markers for subdividing the Quaternary glacial sequence. Others were found in River Denys Lowland by Fletcher (1881). Finds of mastodon fossils – a molar near Baddeck, and a femur and tusk from Middle River area – were reported by Piers (1915) and later became significant for the problem of dating the nonglacial intervals. Ries and Keele (1911) catalogued clay deposits, some of which underlie the surface till. Guernsey (1927, 1928) mentioned a shell-bearing gravel, and additional occurrences of wood and clay beneath till in River Denys Lowland – two observations significant for the present interpretation.

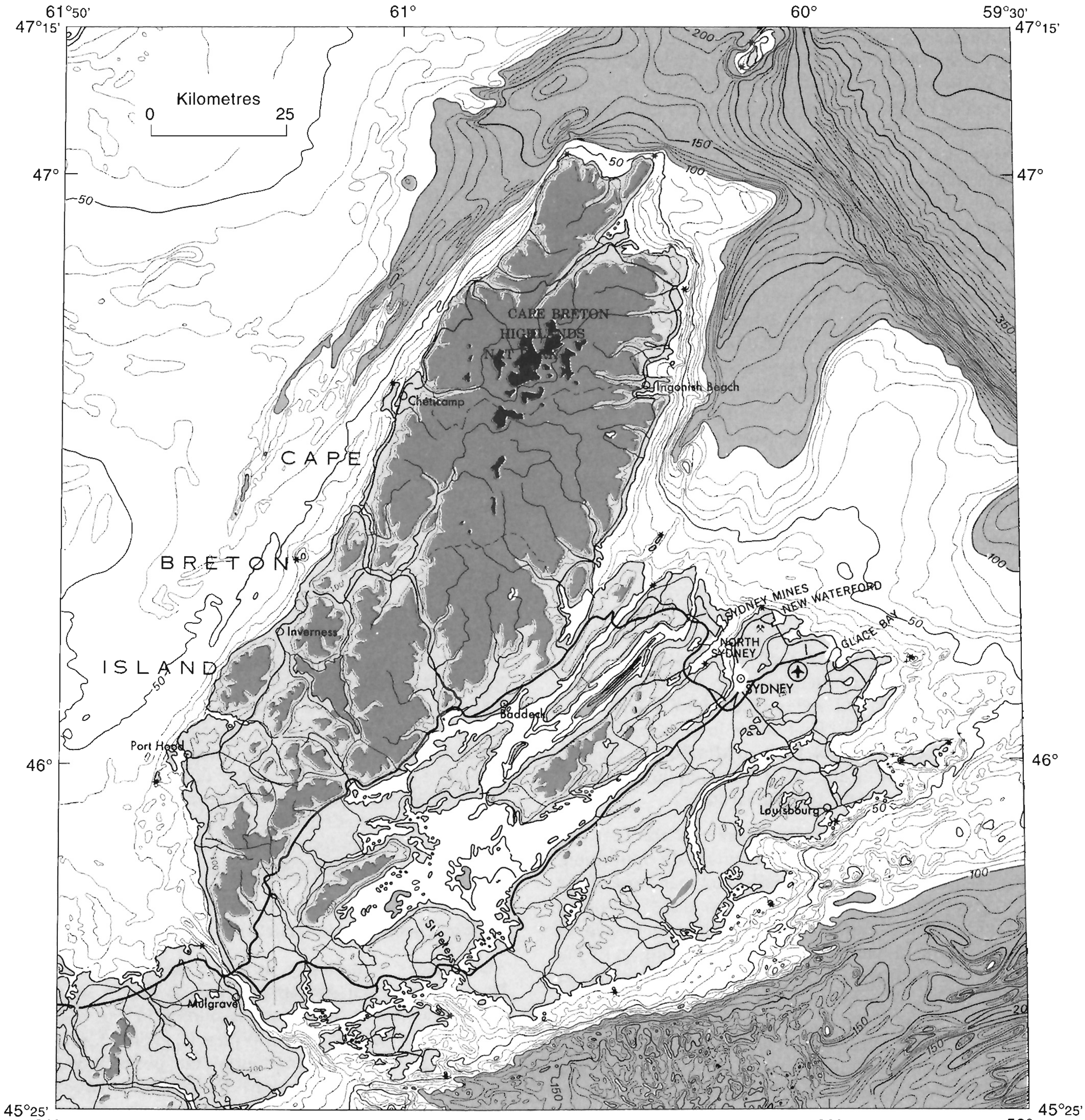



*Figure 6.* *Topography (generalized from International Map of the World sheet NK/NL-20 and Canadian Hydrographic Service Chart 801-A).*

Modern work began with the relocation of the peat bed discovered by Dawson (1868) and the discovery of several other organic beds beneath tills. They gave ages beyond the range of the radiocarbon dating method, and thus, as evidence of pre-last glacial climate, provided pollen-stratigraphic studies with a new focus. In the 1950s, the Hillsboro site (until recently spelled Hillsborough) was discovered by L.D. Wilson, Bay St. Lawrence site by E.R.W. Neale, Whycocomagh by D.G. Kelley, and the Benacadie shore 3 and East Bay sites by L.J. Weeks (Prest, 1957, 1970). Livingstone (1968) reported that the Hillsboro pollen assemblage resembled that of early Postglacial time, while Mott and Prest (1967) found that the others reflected climates both warmer and cooler than the present; these they characterized as "interglacial" and "interstadial", respectively. J.B. Railton (pers. comm., 1970) and Mott (1971) assigned the River Inhabitants organics to a tundra phase prior to the onset of the last glaciation. It was the subtill organic beds in the complex sequences along East Bay (Grant, 1971a, 1972) that subsequently became the definitive components in the stratigraphic basis for the Quaternary history of the region. Amino-acid racemization ratios on wood from these beds, provided by N.W. Rutter, were the basis for placing them all in the interglacial interval corresponding to oxygen isotope stage 5 (Grant, 1983).

In 1982, while this report was being prepared, other workers were invited to collaborate in stratigraphic study of the fourteen subtill organic beds which had been discovered. Further amino-acid work on the wood fraction led Occhietti and Rutter (1982) to assign them to the Early Wisconsinan St. Pierre interstadial interval. A shell bed in colluvium at Bay St. Lawrence was interpreted as deep water sediment, glacially transported and redeposited into a terrestrial sequence (Guilbault, 1982). Peat layers in the same deposit were inferred to represent cool, Early to Middle Wisconsinan conditions (de Vernal et al., 1983), thus implying ice-free conditions during part of the last glacial stage.

On the basis of further pollen work and the first age estimates on wood by the thorium/uranium disequilibrium method, Mott and Grant (1985) reaffirmed that most of the subtill organic beds probably formed in oxygen isotope stage 5 (128-75 ka), but that at least three belonged to Wisconsinan time in that they gave ages as young as 47 ka, also by the $^{14}C$ method, and thus record ice-free conditions as recent as Middle Wisconsinan time. Grant and King (1984) correlated the broad glacial/nonglacial succession of Cape Breton Island, and of other parts of the Atlantic region, to the 23 000 year climate cycle recorded in North Atlantic deep sea sediment. De Vernal and Mott (1986) and de Vernal et al. (1986) refined the palynostratigraphy and reaffirmed that most of the subtill organic beds belonged to the three warm periods of the last interglaciation. These findings have been summarized elsewhere (Grant, 1987, 1989; Mott, 1989).

### Early pollen study of postglacial organics

Quaternary palynology on Cape Breton Island was pioneered by Auer (1930) who inferred changes of vegetation and climate, including a hypsithermal interval, using cores from a peat bog near Strait of Canso. Hall (1949) and Killam (1951) deduced a similar record from four bogs in the northern part of the island.

Some sites record a late glacial climatic reversal and glacial readvance about 10-11 ka. D.A. Livingstone obtained pollen records of postglacial climate since 12 ka by coring lake sediment at five localities on the highlands and lowlands. One site, Gillis lake [40], recorded a sharp cooling and a glacial readvance which he correlated with the Valders (Younger Dryas) episode 10-11 ka (Deevey, 1958; Livingstone, 1963, 1968; Livingstone and Estes, 1967; Livingstone and Livingstone, 1958; Schofield and Robinson, 1960). Three peat beds, dating to the same interval and covered by till, were reported by MacNeill (1969), Grant (1972), and Terasmae (1974). Evidence of a climatic reversal about 11-10 ka was found at 10 sites on Cape Breton Island by Mott et al. (1986) who correlated it with the Allerød/Younger Dryas oscillation in Europe and attributed it to a southward migration of the oceanic Polar Front. Stea and Mott (1989) reported till overlying a 10.9 ka peat bed in adjacent mainland Nova Scotia and postulated a Younger Dryas age readvance of an ice mass in southern Cape Breton Island.

## *Bedrock geology*

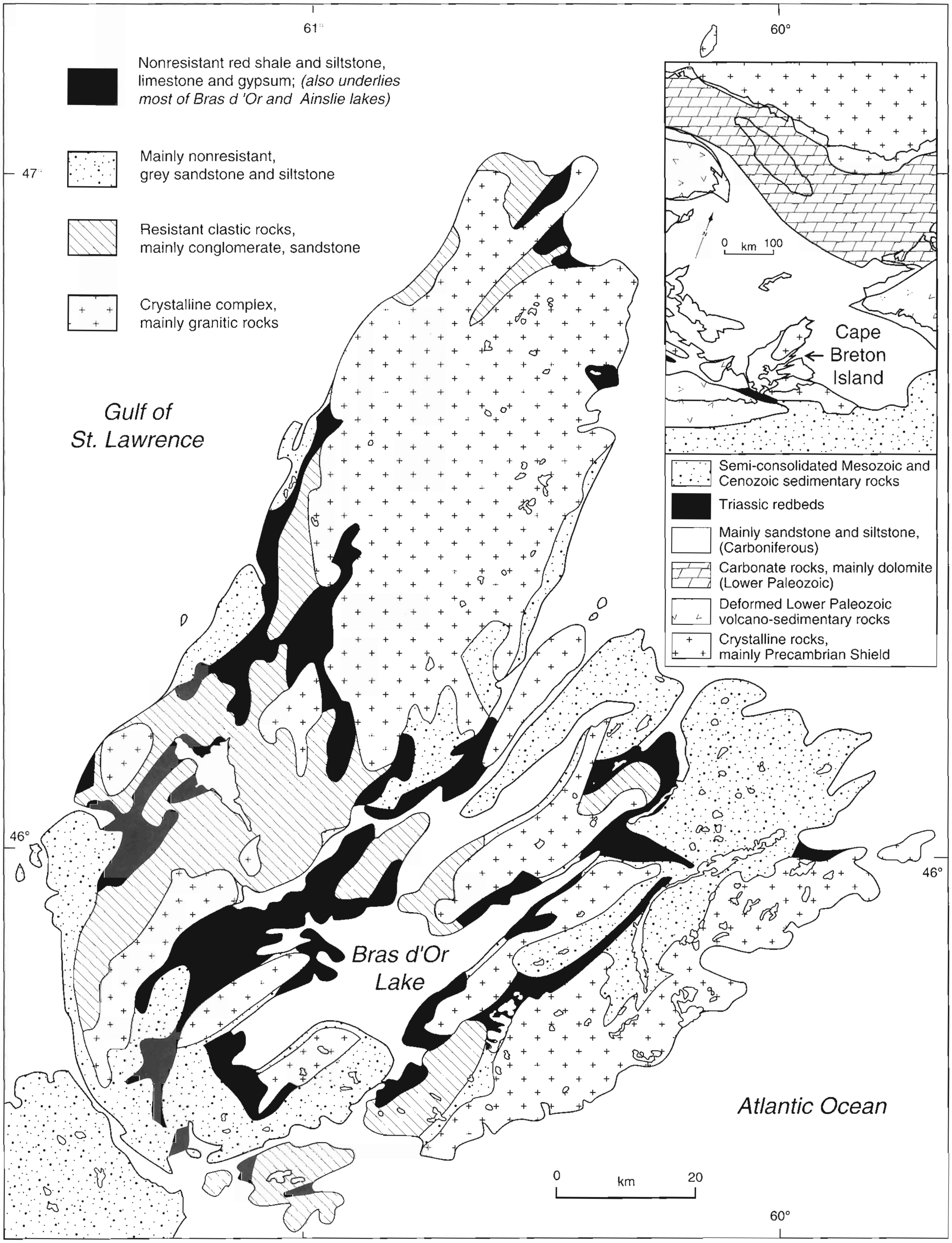
### General character

The landscape of Cape Breton Island is one of the most interesting in the Maritime Provinces region because of the juxtaposition of various contrasting landform groups – flat-topped steep-sided uplands, strongly undulating lowlands, and deep submarine channels and basins (Fig. 6). The island owes its existence mainly to relatively more resistant Precambrian crystalline rocks and folded Paleozoic sedimentary rocks which remain after differential erosion of surrounding weaker Paleozoic sedimentary rocks. The relief is largely the product of Cenozoic planation, uplift, and dissection; pre-Quaternary landforms have been locally accentuated or modified by Quaternary glacial erosion and deposition.

### Bedrock geology and structural evolution

The bedrock stratigraphy, structure, and tectonics of Cape Breton Island and surrounding area is drawn from recent regional syntheses (Keen and Beaumont, 1990; Keen et al., 1990; Gradstein et al., 1990; Williams, in press). As with the rest of the Appalachian region, the bedrock geological makeup and structural history is the product of two cycles of continental collision and rifting involving the American and Eurafrican crustal plates. Three of the region's four main bedrock terranes make up Cape Breton Island and the surrounding seafloor (Fig. 7, inset) (Bujak and Donahoe, 1980).

***Figure 7.*** *Bedrock geology of Cape Breton Island (modified from Bujak and Donahoe, 1980) and of the surrounding region (inset; adapted from Geological Survey of Canada, 1979), expressed in terms of the units which influenced the physiography and the distribution and lithology of Quaternary deposits.*


61°
60°
47°
46°
Nonresistant red shale and siltstone, limestone and gypsum; (also underlies most of Bras d 'Or and Ainslie lakes)
Mainly nonresistant, grey sandstone and siltstone
Resistant clastic rocks, mainly conglomerate, sandstone
Crystalline complex, mainly granitic rocks
Gulf of St. Lawrence
Bras d'Or Lake
Atlantic Ocean
0 km 20
0 km 100
Cape Breton Island
Semi-consolidated Mesozoic and Cenozoic sedimentary rocks
Triassic redbeds
Mainly sandstone and siltstone, (Carboniferous)
Carbonate rocks, mainly dolomite (Lower Paleozoic)
Deformed Lower Paleozoic volcano-sedimentary rocks
Crystalline rocks, mainly Precambrian Shield

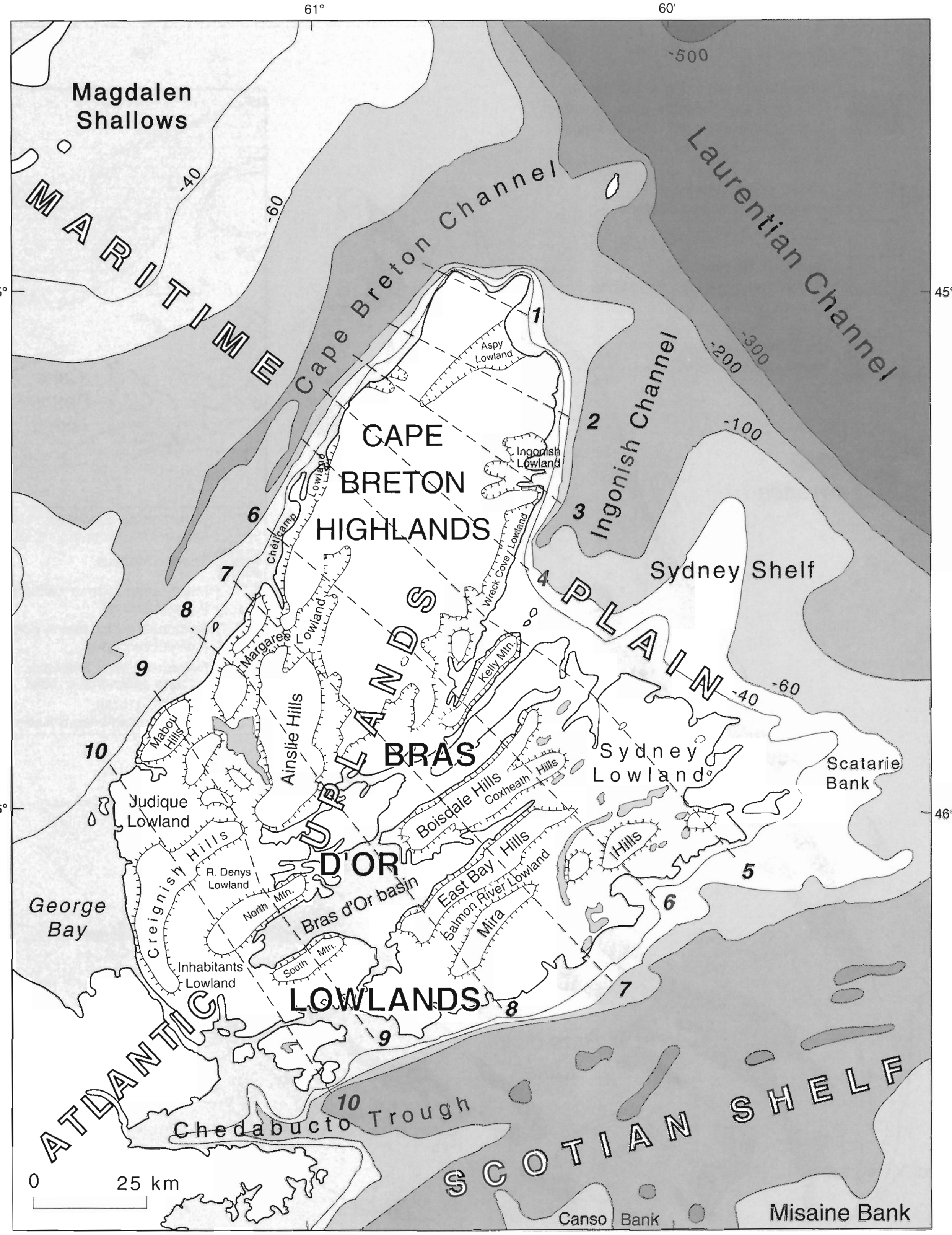



***Figure 8.*** *Physiographic divisions (adapted from Bostock, 1970 and Sanford and Grant, 1976), showing also the locations (1-10) of the topographic profiles depicted in Figure 10.*

The lithotectonic assemblage is the product of basin infilling, folding of sediment, and shelf progradation associated with basement warping and block faulting. Resistant Precambrian crystalline basement blocks, upthrown in Paleozoic time, form the south coastal area and steep-sided hills and highlands. Permo-Carboniferous sedimentary rocks that accumulated in fault-bounded basins underlie the surrounding lowlands, including Gulf of St. Lawrence and Cabot Strait area. A wedge of Mesozoic/Cenozoic sedimentary rock laps onto these older terranes and forms the present continental shelf. Within these larger groupings are the main rock units (Fig. 7) which have determined the shape and surface character of physiographic features, as well as the thickness, texture, and lithology of Quaternary sediments.

The structural evolution of Cape Breton Island began when separation of the American and Eurafrican plates in Cambro-Ordovician time created the proto-Atlantic (Iapetus) ocean. On the American margin of this ocean, shelf deposits on Grenville basement are represented by sedimentary and volcaniclastic cover rocks in Mira valley area. The spreading eventually ceased, the movement reversed, and the plates began to approach one another, thus closing the ocean to create the Pangea supercontinent. They eventually collided in Middle Devonian time and the collision zone involved the Cape Breton Island region. The resulting Acadian Orogeny produced a fold belt with granitic intrusions, called the Appalachian Orogen. The uplands, highlands, and part of the southeastern coast are fragments of Precambrian basement composed of igneous and metamorphic complexes upthrown by the collision.

The plates began to separate again in Carboniferous time, causing the basement to sag and form a basin under Gulf of St. Lawrence where terrestrial sandstone redbeds accumulated. These rocks form a relatively undeformed terrane called the St. Lawrence Platform (Williams, in press). The sea invaded adjoining rift basins in Cape Breton Island and deposited shale redbeds, salt, and gypsum. Renewed uplift in Late Carboniferous time ended marine deposition and created enclosed, fault-bounded grabens in which grey sandstone, conglomerate, and coal accumulated under reducing conditions. Renewed rifting and block faulting in the Triassic produced a graben along the south coast where desertic redbeds and volcanic rocks were deposited.

Eventually, in Jurassic time, Pangea began to break up as the plates separated. A seaway flooded the rift and the present Atlantic basin thus came into being. Post-rift cooling of the trailing continental margin caused the crust to subside, thus providing a locus for deposition. A 4-6 km thick wedge of Mesozoic/Cenozoic sedimentary rock which has accumulated over the last 160 Ma to form the present continental shelf thus represents an average subsidence in the offshore of about 50-100 m/Ma since the Jurassic. At Diogenes Brook in the Creignish Hills, a small outlier of Cretaceous sediment is infaulted into the crystalline basement, indicating that basement rifting continued into the Mesozoic. This may explain the discordant attitude relationships of the unroofed portions of these unconformities (discussed below under Exhumed unconformities (paleoplains)). None of the pre-Quaternary Neogene rocks underlying the shelf extend onshore, which may suggest that sea level is presently at its Mesozoic/Cenozoic highest level. In any case, their absence prevents a direct reconstruction of events on land during this period. However, it may be inferred that, in order to produce the vast quantity of Neogene sediment which accumulated offshore, the hinterland, of which Cape Breton Island is a part, was being eroded and presumably also being generally uplifted during most of post-Jurassic time. Thus, with the offshore subsiding and the hinterland uplifting, the region underwent a broad tilting, which established the tectonic setting for the development of Late Neogene features through widespread erosion. That erosional history is reconstructed by means of the following physiographic analysis.

### *Physiography and geomorphological history*

Cape Breton Island's major and minor relief features are the product of successive episodes of erosional planation, crustal uplift, fluvial incision, partial resubmergence, and finally, widespread glacial deposition and local overdeepening. The island is characterized by strong landscape contrasts involving isolated hills and highlands standing in sharp relief above an extensive complex of lowlands (Fig. 8). The hills and highlands are steep-sided, more or less flat-topped, and owe their relief to their composition of resistant igneous and metamorphic rocks. The intervening lowlands are developed on nonresistant sedimentary rocks (for example, cover photo, Fig. 9) on which most of the Quaternary deposits are localized. Incised onto the lowlands are glacial basins, the largest being Bras d'Or Lake, and the offshore channels, which locally have broad basins representing local glacial overdeepening. The hills and highlands belong to the Atlantic Uplands physiographic region, while the lowlands are part of the Maritime Plain (Bostock, 1970; Sanford and Grant, 1976). The highlands and lowlands are generally considered to be relict erosion levels or degradation surfaces representing successive pre-Quaternary peneplains. The offshore channels (and the basins later cut into them) have been tentatively referred to a third erosion level (Grant, 1989), but are not

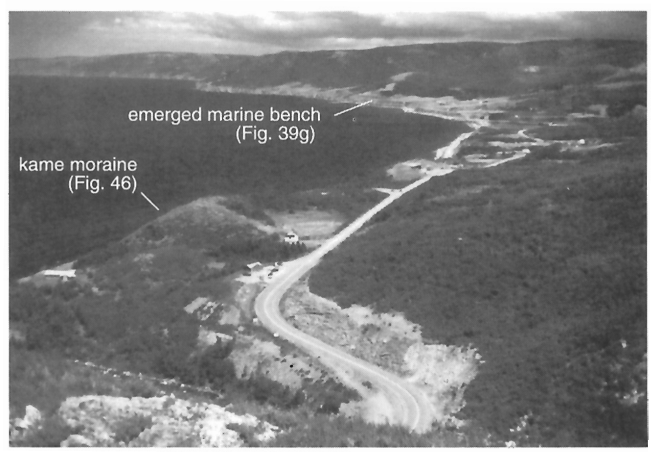



*__Figure 9.__ Coastal lowland at Pleasant Bay, as seen from Cabot Trail on Mackenzie Mountain, with key stratigraphic features indicated. GSC 120178*

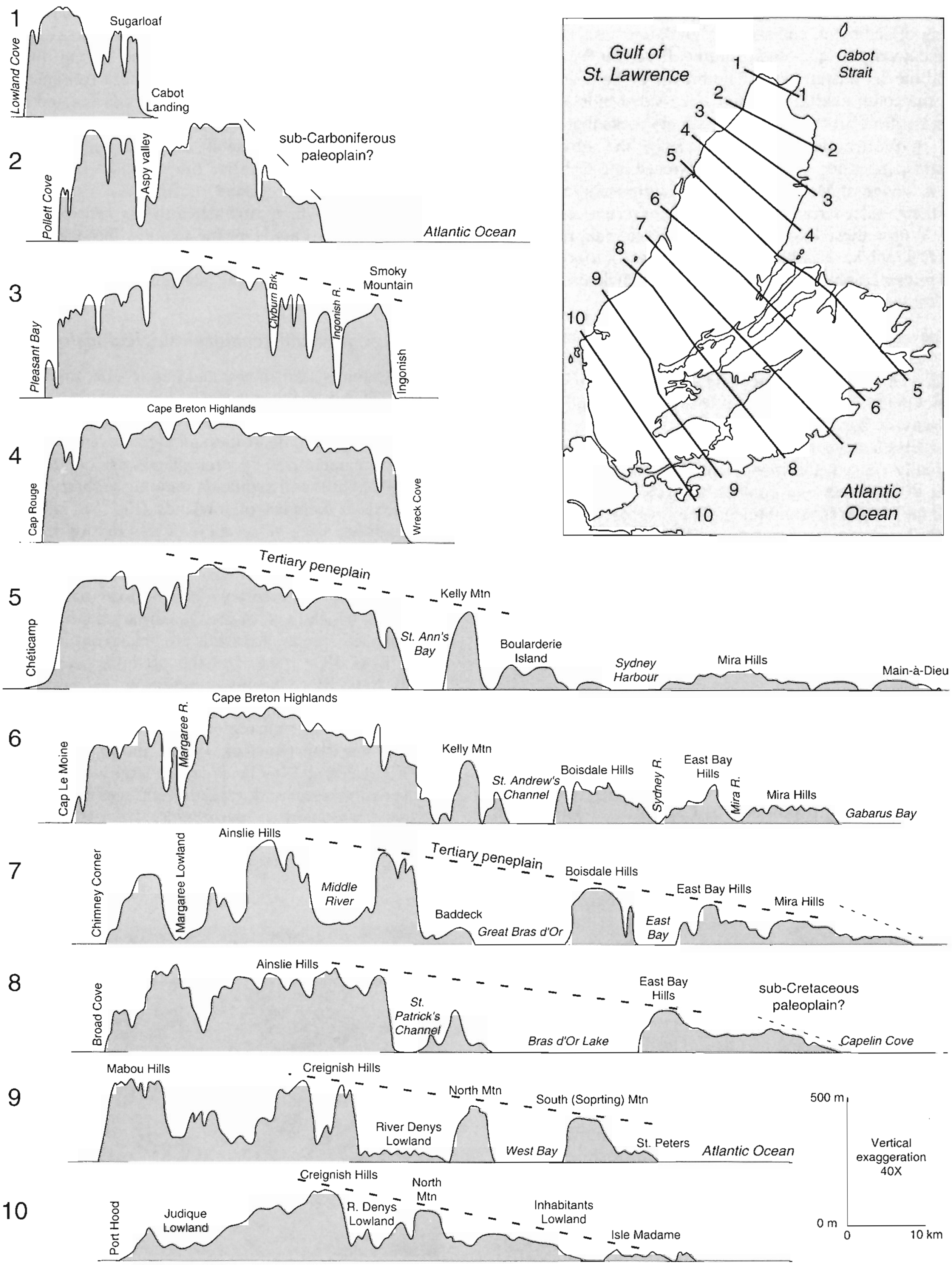



***Figure 10.*** *Representative topographic profiles, oriented across the main physiographic units (locations in Fig. 8).*

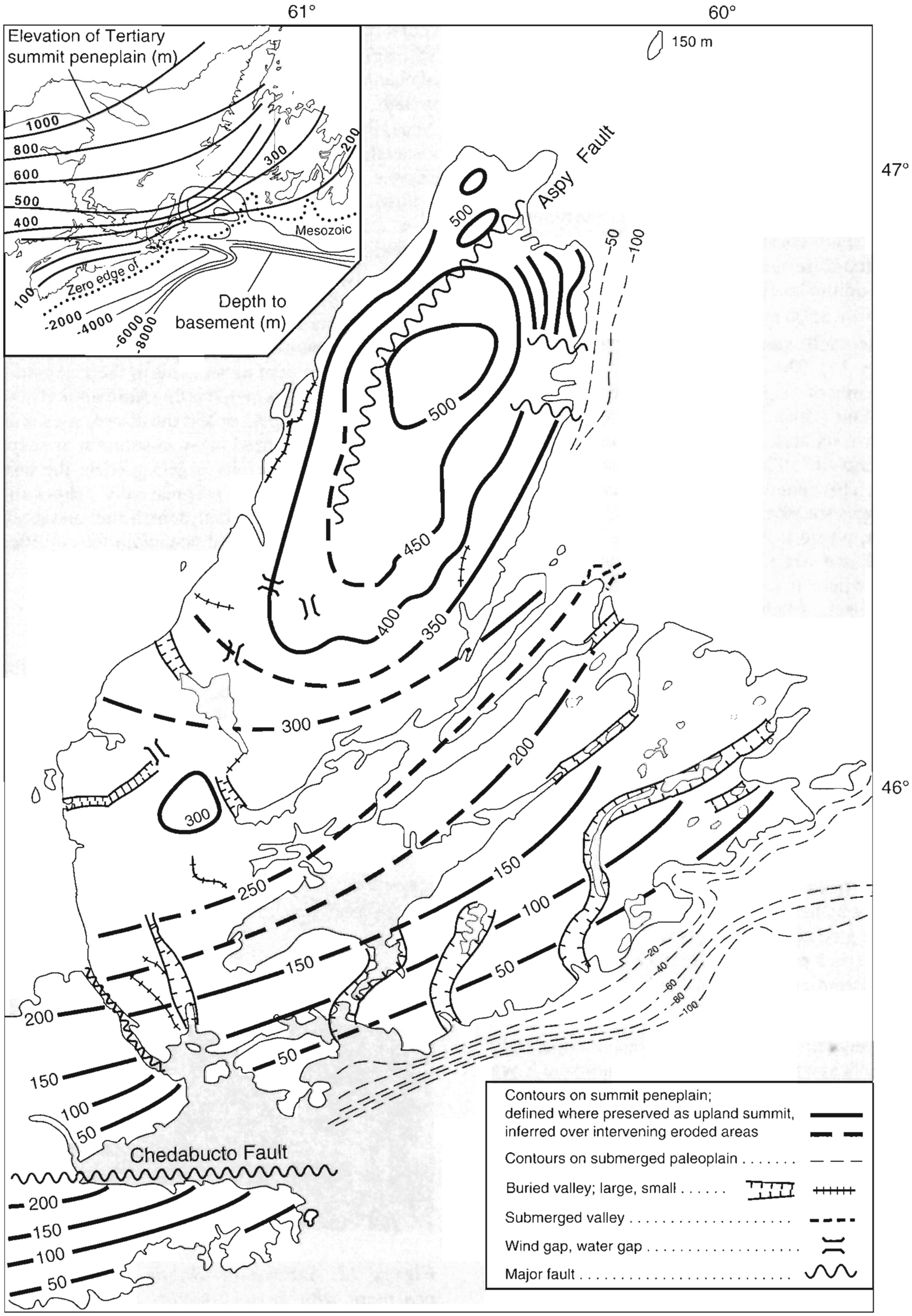



***Figure 11.*** *Attitude of summit peneplain in Cape Breton Island and beyond (adapted from Grant, 1989) in relation to basement contours on the continental shelf (from A.C. Grant, 1988), plus other major physiographic features.*

discussed here. To illustrate the following description and interpretation, ten topographic profiles (Fig. 10), show the strong topographic contrasts, emphasize the depth of incision of the higher areas and reveal the continuity of certain levels.

## Uplands – remnants of a tilted, dissected peneplain

The uplands in Nova Scotia are implicitly defined as discrete blocks standing at various elevations above a distinct lowland level. The general summit level of each upland block appear to conform to a single surface that slopes southeastward and maintains its identity as a geomorphic feature across the entire island. Along the south coast, the upland level merges with the lowlands and thus loses its appellation as an upland in that area. The upland summit level is a generally flat surface which slopes at ≈1:200 from >500 m elevation on the northern coast, to ≈150 m on the south side of Bras d'Or Lake (Fig. 10, profiles 3-10; Fig. 11). The isolated summits are generally regarded as remnants of single regionally extensive erosion surface or peneplain which has been set in relief by later erosion of weaker-rock areas, as Daly (1901) originally discerned and Goldthwait (1924) further elaborated for Nova Scotia as a whole. This denudation surface occurs throughout the region and slopes southeastward from its highest portions in Newfoundland, where it is called the Long Range Peneplain (Twenhofel and MacClintock, 1940), and Gaspésie-New Brunswick where it forms the Fundy Surface (Bird, 1972) (Fig. 11, inset). Mathews (1975) suggested a mid-Tertiary (Miocene?) or younger age for the uplift and incision of the peneplain, by comparing the volume of rock eroded over the intervening lowlands with the volume of abyssal sediment sequences on the Atlantic Ocean deep sea floor.

The surface of the upland peneplain remnants varies in detail from featureless to rough. The Mabou Hills (Fig. 12) and most of Cape Breton Highlands (Fig. 13) have remarkably smooth surfaces. Gradients are very gentle, interfluves are smoothly rounded, bedrock is rarely exposed except in valleys, and the surface is devoid of local undulations, as might be expected from glacial action. In contrast, all other upland summits, whether bedrock or till, have hummocky surfaces typical of glaciated terrain. As with other such areas in the region, there is a profound difference in the degree to which the pre-Quaternary peneplain has been modified by glaciation (Fig. 14).

The upland peneplain of Cape Breton Island is apparently not coplanar with its assumed continuation in mainland Nova Scotia and there are local sharp changes in slope and direction (Fig. 11). The peneplain on the mainland is about 50 m higher on the west side of Cabot Strait and about 200 m higher on the upthrown south side of Chedabucto Fault. On the island, the contours show a sharp warp in the Mira area, the slope swings to northward in Bras d'Or region, and the highlands portion is essentially a broad dome. While the dome may be an original feature, the other discontinuities and irregularities may suggest postplanation warping and faulting, perhaps induced by erosional unloading.

### *Upland valleys*

The uplands are incised by a remarkable dendritic system of deep valleys. Most are V-shaped gorges with interlocking spurs, the hallmark of normal fluvial erosion (Fig. 15, 16, 17, 18). Relatively few have U-shaped cross profiles indicative of channeled glacier erosion such as typify the larger northern valleys, notably the Chéticamp, Ingonish, Dundas, Grande Anse, Blair, and some smaller valleys (Fig. 19). Valleys are excavated in solid bedrock of the plateau surface with varying degrees of integration. The area of the highland valleys is estimated at 500-1000 $km^2$, or more than one-half the plateau area. Local relief ranges from about 10-20 m where the headwater streams begin in shallow swales on the uplands plateau, to chasms that are 200-500 m deep (depending on plateau height) where the streams debouch onto the adjoining lowlands. The volume of removed rock represented by the valleys is immense (60-300 $km^3$), compared to the volume of postglacial alluvium now resting at their mouths on the adjacent lowlands (2-4 $km^3$). It is therefore unlikely that they were cut in postglacial time, unless the discrepancy represents rock which has been carried away in solution or suspension. The upland valleys therefore largely predate the last glaciation. and were probably cut progressively deeper in each interglacial period, and thus represent the net dissection which has occurred since the upland peneplain was uplifted and set in relief since the Tertiary.

### *Exhumed unconformities (paleoplains)*

Two extensive areas of crystalline basement, forming tilted bevels or facets on portions of the upland and lowland surface are almost as smooth as the summit peneplain, but have much steeper gradients (1:25 and 1:55, compared to 1:200). Stratigraphy shows them to have been exhumed from beneath cover rocks and thus to be much older surfaces. One of these extensive bevels forms the southern coast (Fig. 10, profiles

***Figure 12.*** *Summit of Mabou Hills, a pre-Quaternary peneplain with gently rounded slopes smoothed by mass wasting, then dissected following uplift. GSC 203504-S*


Map units: 1. residuum (felsenmeer, rubble, grus, saprolite); 2a. till blanket; 2c. discontinuous till veneer; 3a. ice-contact stratified drift; 3b. outwash; 7. colluvium (slope rubble); 8. bog, fen

*Figure 13. Stereogram showing smooth highland peneplain surface that is deeply dissected around margins; headwaters of Middle River. NAPL A21535-65, 66.*

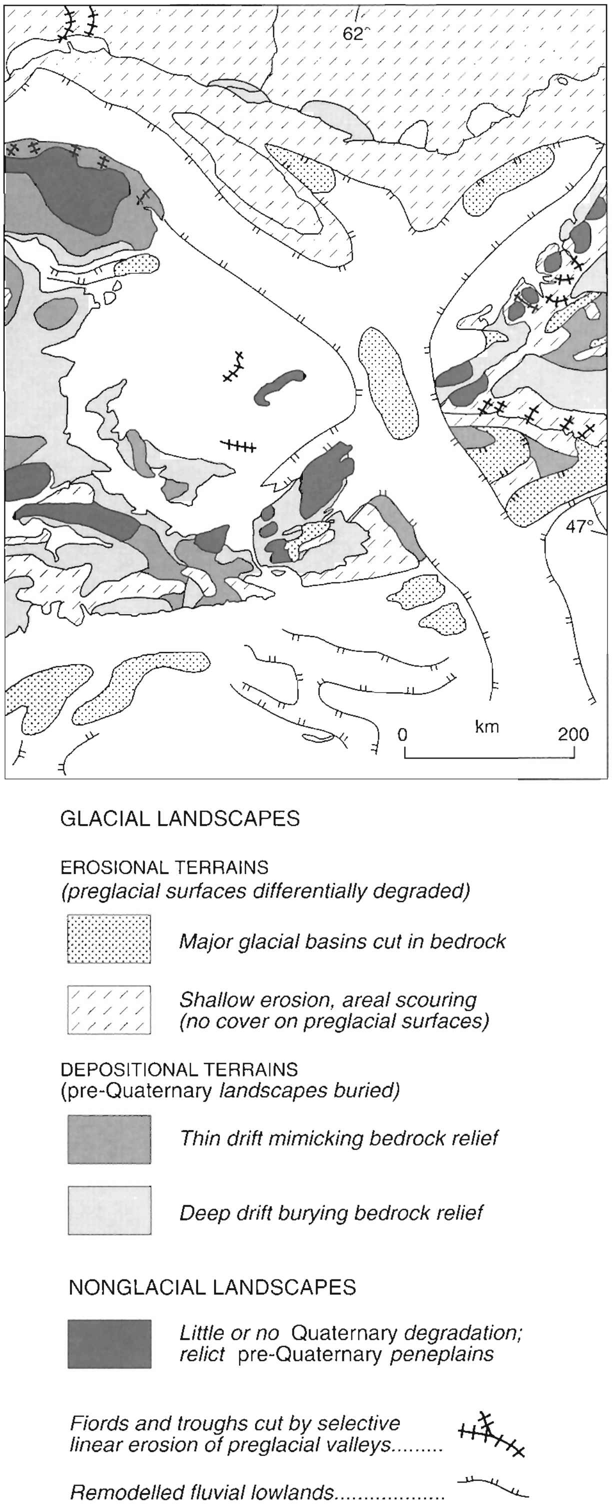



***Figure 14.*** *Degrees of glacial modification of pre-Quaternary landforms (after Grant, 1989).*

7 and 8). It continues offshore where subbottom seismic profiles show that is the basement unconformity beneath continental shelf strata ranging in age from Jurassic (McIver, 1972) to Lower Cretaceous (King et al., 1974). A Cretaceous or older age may explain why this surface appears to be upthrown to higher elevations on the south side of Chedabucto Fault, which was active in Late Cretaceous and Early Tertiary time (King and MacLean, 1970). From there this gently tilted upland surface extends southward to form the mainland Nova Scotia Atlantic coast. That segment has been correlated with the "Fall Zone Peneplain" of

***Figure 15.*** *Aerial view of Cape Breton Highlands plateau margin dissected by 300 m deep gorge, Leonard McLeod Brook area. GSC 204024-Z*

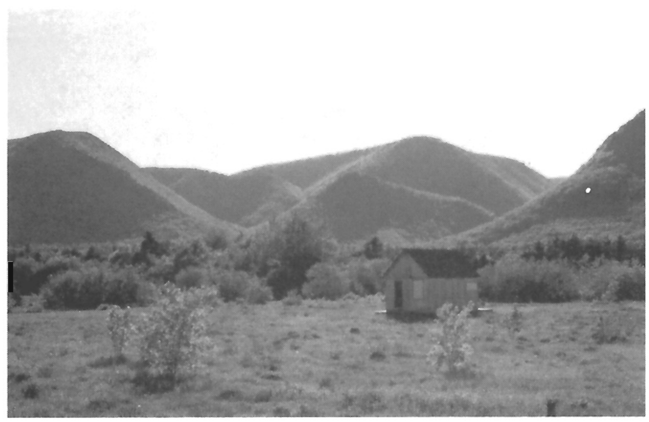


***Figure 16.*** *View of deeply dissected Aspy scarp, mouth of Gray Glen Brook. GSC 204024-Y*

New England (Flint, 1963) which borders the continental margin from Massachusetts to Georgia and slopes seaward beneath the Mesozoic shelf sequence.

The other anomalous coastal facet, first noted by Goldthwait (1924, p. 36,) but not explained, is a ramp between Ingonish and Aspy Bay descending from 450 m to sea level with a regular gradient of about 1:30 (Fig. 10, profile 2; Fig. 11) that lies between the almost flat highlands plateau and the markedly steep highlands margin. This coastal ramp does not correspond to any particular basement lithology or structure. It is abutted on the north and south by grabens of Carboniferous rocks which appear to overlie it just offshore. This facet is therefore tentatively considered to be the unroofed unconformity beneath Carboniferous rocks.

These exhumed unconformities (or resurrected peneplains, as Johnson (1925) called them) have areas of 200-2000 $km^2$ and thus are sufficiently extensive to qualify as "paleoplains" (Ambrose, 1964).

*Depressions (sinkholes?) in granite*

Some small depressions (Fig. 20) on the southern highlands plateau may be significant for understanding local tectonic structure. The pits resemble sinkholes, except that the bedrock is granite gneiss. They are not morainic hollows, as no till is present. Currie (1977) suggested that the Precambrian massif may be a 300 m thick slab thrust over the Carboniferous suite. Tectonic superposition is suggested by the fact that the contact generally follows the foot of the highlands such that each valley is floored with Carboniferous rock which is this area is predominantly gypsum. It is therefore suggested that the depressions are karst pits where the granite has collapsed into underlying solution cavities.

### Bras d'Or Lowlands

Surrounding the uplands is a network of largely interconnected and partly isolated lowlands, part of the Maritime Plain physiographic region, which are here collectively and informally termed the Bras d'Or Lowlands (Fig. 8). Some of the recognizably separate constituent parts are given informal names for convenience of discussion. Their contact with the uplands is generally a sharp break in slope at the foot of the bordering escarpments. The lowlands range from 10 m to 200 m elevation and are developed mainly on Carboniferous sedimentary rock. Their elevation demonstrates a close relationship between denudation depth and geological structure and lithology. The lowest levels are developed on the weakest rocks (Mississippian shale, mudstone, and gypsum), such as along the coast (Fig. 21) and in inland valleys such as those of the Middle and Margaree rivers (Fig. 22). The few areas underlain by resistant Pennsylvanian sandstone stand somewhat higher. The broadest part is the Sydney Lowland; the narrowest is the lakes O'Law wind gap.

The greater part of the Bras d'Or Lowland lies in the 50-80 m elevation range (Fig. 10, profiles 5-10). Although the regularity of the surface has been obscured by later incision and by glacial erosion and deposition, the narrow elevational range suggests that it represents a distinct denudation level or peneplain that is inset into the older summit peneplain. This

*Figure 18. Mackenzie River valley as seen from Cabot Trail is deeply incised in plateau; spurs are interlocking but glacially rounded. GSC 120182*

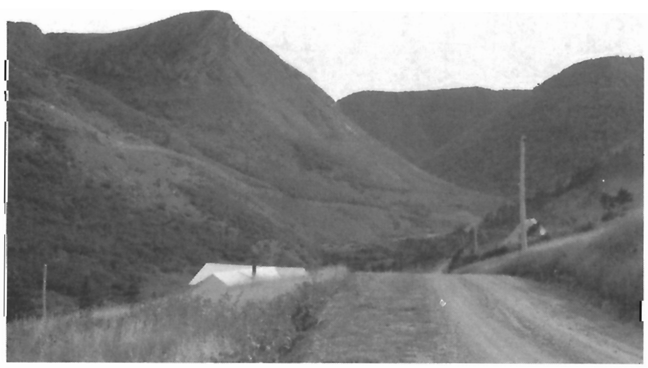
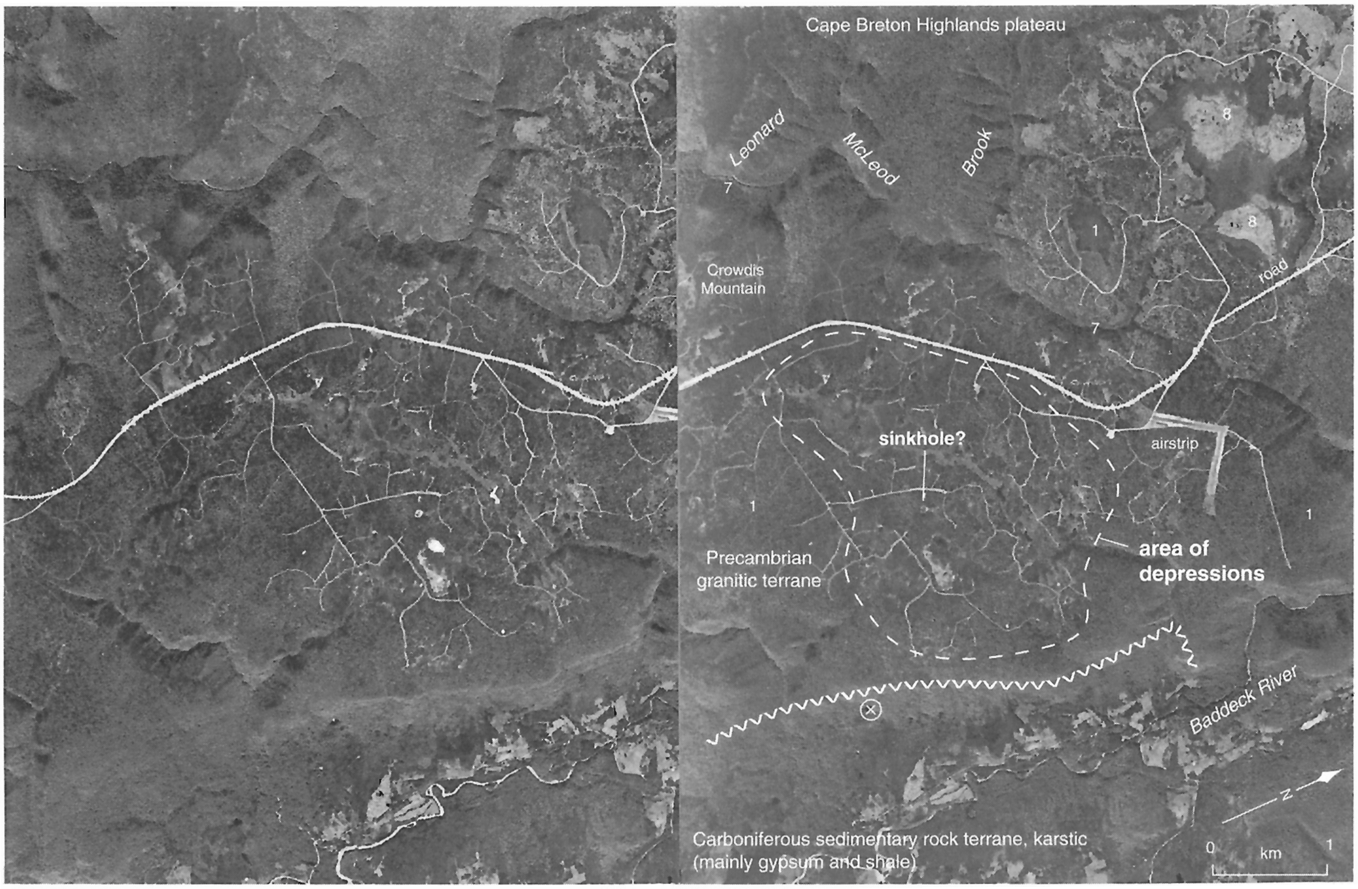


*Figure 17. Gorge cut into northern plateau at Meat Cove. GSC 202024-W*


*Figure 19. Aerial view of highland gorge of Sailor Brook, northwest coast; note slight glacial modification and emerged marine bench (B) at mouth. GSC 203504-W*




Map units: 1. residuum (felsenmeer, grus, saprolite); 7. colluvium (slope rubble); 8. bog, fen ⊗ karst depression in gypsum VVVVVV thrust? fault

***Figure 20.*** *Stereogram of possible karst depressions in granite, Leonard McLeod Brook, Crowdis Mountain. NAPL A21576-73, 74.*

***Figure 21.*** *Presqu'ile, a cuesta showing differential erosion of Carboniferous sedimentary strata, as seen from Cabot Trail. The lagoon may be an extension of Chéticamp paleochannel. GSC 203504-X*

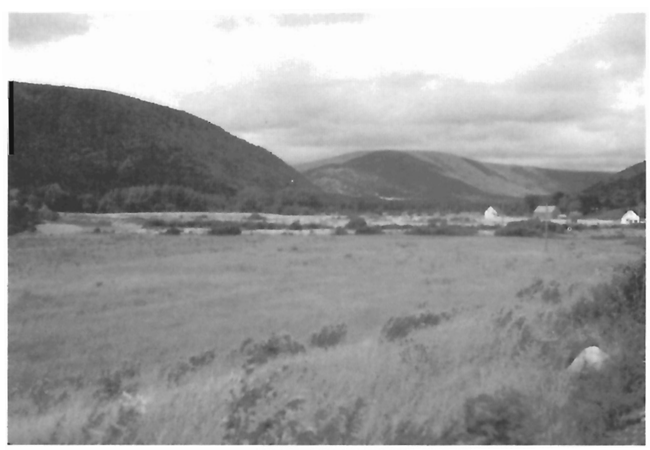


***Figure 22.*** *A small semi-enclosed alluvial flat, or "intervale", thinly covering a relatively flat bedrock floor belonging to an interior lowland that represents a Tertiary peneplain developed on nonresistant sedimentary rocks; upper Margaree River valley at Kingross. GSC 204024-X*

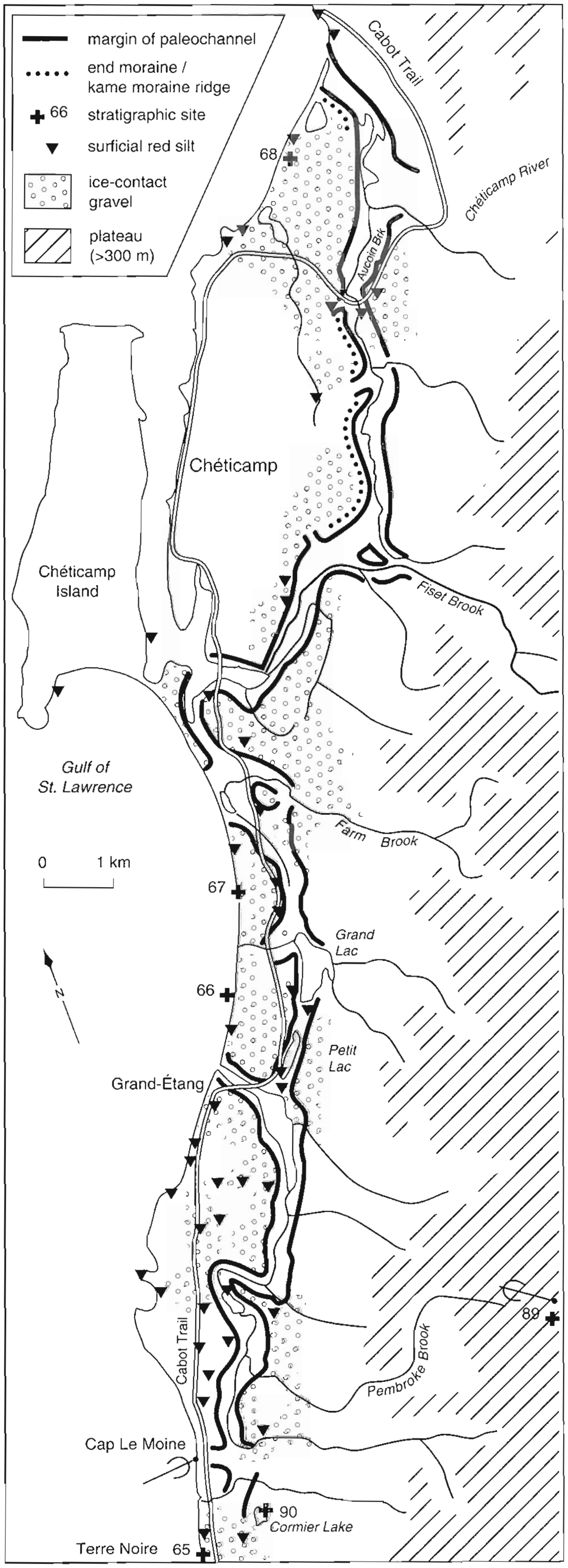



**Figure 23.** *Diagram of abandoned paleochannel on Chéticamp Lowland, showing relation to ice-contact stratified drift and surficial red silt.*

erosion surface probably correlates with a similar level also developed on weak rocks elsewhere Nova Scotia and southern New Brunswick which Bird (1972) termed the Dorchester Surface. The lowland surface appears to be coplanar with an incision level in the Cenozoic shelf sequence which King et al. (1974) have tentatively assigned at Pliocene age.

*Chéticamp paleochannel*

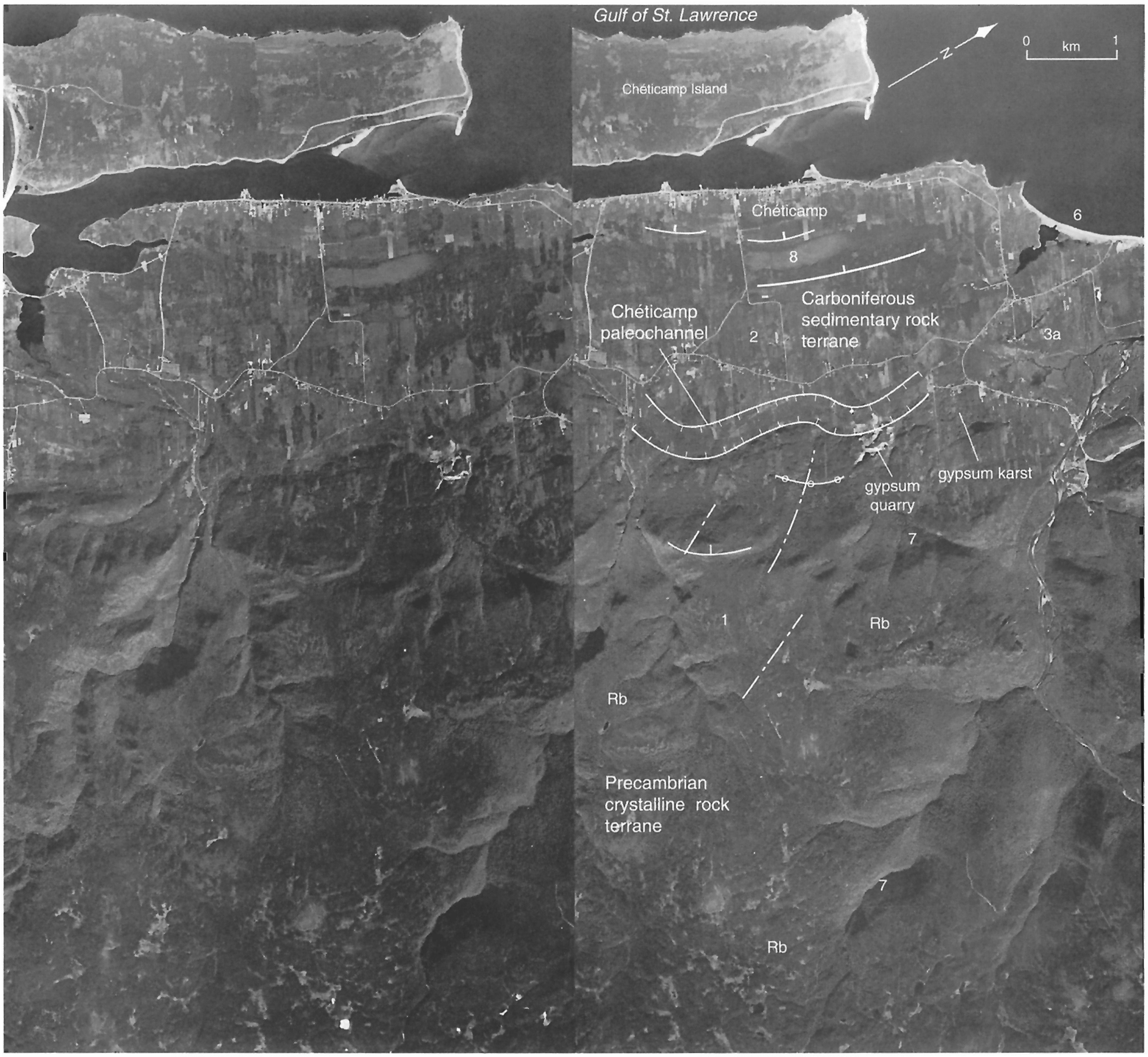
A prominent feature of the Chéticamp coastal lowland, but one of uncertain origin and age, is a sinuous channel which meanders parallel to the coast for more than 30 km from Margaree Harbour to Chéticamp River, and possibly to Presqu'ile (Fig. 11, 21, 23, 24, 25). It ranges from 200 to 800 m wide, 10 to 30 m deep, and slopes northward from +30 m to below sea level, with a breach to the sea at Grand Étang. Cut mostly into nonresistant shale and gypsum in a steeply inclined Carboniferous sandstone sequence, it generally has walls that are steep and sharply defined, although in places they are rounded and one or both are missing. The floor is not continuously graded, but instead is gently undulating with bog- and lake-filled depressions. The channel also cuts through most of the Quaternary sequence, the till and interbedded gravels, but there are scattered mounds of till and gravel kames and on its walls and floor. The west channel wall in the Chéticamp area is an end moraine or kame moraine, and outwash plains end at the channel (for example, those of Chéticamp River, Fiset Brook, Farm Brook, and Pembroke Brook). The channel captures most valleys that drain the plateau; their lower courses are abandoned or buried. While parts of the channel may have originated in pre-Quaternary time (Cameron, 1948) when the lowland erosion surface was being incised during lower base levels, the

**Figure 24.** *View of Chéticamp paleochannel containing moraine-dammed lakes; Petit Lac in foreground; Grand Étang pond in distance. GSC 120218*

incomplete channel morphology and its close relationship to deglacial gravels suggests that it was cut while partly ice-walled (i.e., subglacial) during the final stages of glaciation. If so, the channel would be termed a tunnel valley. Indeed, it is comparable in form and size to the type examples in Netherlands and north Germany where they are called rinnenseen and rinnentäler (Fairbridge, 1968, p. 463). As suggested below (section Cenozoic history), the channel represents a period when meltwater was forced to flow parallel to the mountain slope perhaps because it was confined by a glacier in Gulf of St. Lawrence.

*Buried valleys*

Bedrock is deeply drift covered on the southern lowland half of the island and the present drainage is largely consequent on morainal topography. Nonetheless, several major buried valleys (Fig. 11) may be inferred from drainage anomalies, lake distribution, coastal shape, and present topography. The generally disorganized drainage has meandering shallow streams with abrupt changes of direction and frequent short, rocky reaches. As first alluded to by Weeks (1954), lines of lakes in morainal tracts suggest underlying linear depressions. Large coastal bays reflecting major topographic lows




*Map units:* Rb. forested, glacier-scoured bedrock; 1. residuum; 2. till (undifferentiated); 3a. ice-contact stratified drift; 6. coastal beach; 7. colluvium; 8. bog, fen

fracture lineament; sedimentary rock strike ridge; line of sinkholes

***Figure 25.** Stereogram of Chéticamp paleochannel and adjacent highlands. NAPL A21576-9, 10*

have no equivalent onshore expression and have only unfit tributary streams. These features suggest large bedrock channels in the subsurface.

Several major buried channels are inferred on the Atlantic drainage slope. The largest two are inferred from the anomalous topographic setting of Mira River (actually a lake). It presently exits through a narrow bedrock gorge to Mira Bay, but formerly it may have debouched into Cow Bay through a wide channel now largely obscured by ice-contact gravel. Alternatively, it may have exited southward, probably to Framboise Cove, along a channel seemingly marked by a line of lakes. Another buried valley is possibly represented by the drift-dammed Loch Lomond group of lakes and the present shallow Grand River. St. Peters Inlet, with depths to -30 m below mean sea level, is clogged with drumlins >30 m thick, suggesting a major buried valley, but which may continue across the inner shelf via a filled channel at -30 m (as noted above). River Inhabitants is inferred to flow along the axis of a filled valley because drilling and hammer-seismic profiling reveals several tens of metres of till (R.M. Gagné, GSC, pers. comm., 1971).

The Gulf of St. Lawrence drainage slope also has several major confirmed and suspected buried valleys. The largest and deepest paleochannel extends from Loch Ban to Inverness and has been probed by drilling. Presumably this channel was the original northwestern outlet of Lake Ainslie watershed before it was plugged with till and ice-contact stratified drift by an offshore glacier which thus backed up the flow, created the lake, and forced a new channel to be cut across the northeastern divide to the Margaree River system. Hammer-seismic profiling across the mouths of Judique Brook, Rory Brook, and Graham River has revealed buried channels filled with drift up to 15 m thick and extending to about 15 m below sea level. In the same area, a buried channel is inferred to underlie Mull River on geomorphic grounds; this channel and those of Mabou River tributaries may have once debouched westward to the position of present Little Judique Brook. The lower reaches of River Denys overlie a till-filled valley extending to at least 20 m below sea level. On the Chéticamp Lowland, several valleys are drift filled and diverted by the large paleochannel.

A number of buried channels may be present on the Cabot Strait drainage slope. Sydney Harbour connects to East Bay of Bras d'Or Lake via a valley that is deeply mantled with till, ice-contact stratified drift, and outwash. Possibly the bedrock floor lies at -20 m (or deeper) thus connecting the floor of East Bay to the floor of Sydney Harbour. St. Andrews Channel, which now has an anomalously small connection to the Atlantic Ocean via a narrow rock gorge (called Little Bras d'Or), may once have exited through a seaway that is now plugged with till at Little Pond. The large coastal indentations of Glace Bay and Little Bay have underfit tributary streams; possibly larger drift-filled ancestors underlie the hinterlands. Examples of valleys, now below sea level, that are partly obscured by drift might include the sills and thresholds along the northern arms of Bras d'Or Lake, notably Great Bras d'Or and St. Andrews Channel. Seismic profiling shows small buried channels on Sydney Shelf (Manchester, 1965, p. 28). Off Point Aconi, colour airphotos show several sinuous channels, cut into the bare-rock seafloor. The larger two appear to be the submerged extensions of Aconi Brook and Little Bras d'Or (Fig. 11). In summary, the number of buried valleys which can be seen and reasonably inferred in the heavily drift- covered areas of Cape Breton Island point to the likelihood of others being found.

*Bras d'Or Lake basin*

Bras d'Or Lake is a landlocked sea consisting of several elongate basins which are localized almost entirely on Mississippian shale and gypsum, the weakest rocks on the island, which occupy fault-bounded grabens. The submarine depths are almost as great as the heights of the adjacent hills (Fig. 10, profiles 6-9). The seafloor generally lies below 50 m below sea level, with the maximum depth of 256 m at the junction of St. Andrews Channel and Great Bras d'Or. The rock floor underlies an average of 10-25 m of sediment, including till and outwash (Manchester, 1965), but considerably greater thicknesses, comprising several till sheets with stratified interbeds, are locally present (A. de Vernal, pers. comm., 1990). The present outlet through Great Bras d'Or is over a bedrock sill at 8 m below sea level (de Vernal et Jetté, 1987) and the several basins are largely hemmed in by bedrock. Hence, the closure on the basin system as a whole is at least 250 m, not counting sediment thickness. With an area of $\approx$700 km$^2$ and assuming an average depth of 125 m, the volume of the Bras d'Or depression is $\approx$90 km$^3$, excluding the sediment fill.

The bottom topography varies from generally smooth in the northern arms and basins where postglacial infilling has evidently masked any older relief to quite irregular in the centre and in West Bay where the local relief averages 50 m and has the same form as the numerous till ridges and drumlins which border this part of the lake. In the centre, an irregular area, which appears to be inset within the morainal topography and has steep sides down to a flattish floor below -100 m, may represent a dead-ice depression. The origin of these features will become clear when seismic profiling records are analyzed (A. de Vernal, pers. comm., 1990).

The origin of these large connected depressions is unclear and poses the same problem as basins of similar size and shape cut in weak Cenozoic rocks on eastern Scotian Shelf (Fig. 6). The Cape Breton Island basins may be drift-dammed fluvial channels (as Goldthwait (1924, p. 58) suggested) cut off from the valley systems on the Scotian Shelf which were cut to depths now 400 m below sea level during Tertiary sea level lowerings of more than 300 m (King et al., 1974; King and MacLean, 1976). Indeed, several drift-filled bedrock channels are inferred for the southern lowlands (Fig. 11), but neither of the two which could possibly connect Bras d'Or Lake to the shelf (St. Peters Inlet and East Bay-Sydney Harbour) is lower than 30 m below sea level where it crosses the inner shelf (Manchester, 1965; R.B. Taylor, pers. comm., 1990). On the other hand, if the basins are true depressions entirely hemmed in by bedrock, they may be the result of either tectonic downdropping or glacial excavation. The suggestion that the basins are unfilled tectonic depressions or grabens (J. Clague, pers. comm., 1988) would mean they


*Map units:* Rc. bedrock with till patches; 2a. till blanket; 2b. till veneer; 2c. discontinuous till veneer; 6. coastal beach; 8. bog, fen

strike ridge of sedimentary rock; crestline of giant till ridge; drumlin, fluting; crag-and-tail hill; minor moraine ridge

***Figure 26.*** *Stereogram showing older till ridges from which drumlins have been carved by later ice movement; note also minor moraines and tombolos; southern Isle Madame. NAPL A21536-2, 3*

postdate the supposed Tertiary lowland denudation event, yet there is no evidence of continuing graben formation since Carboniferous time. The only plausible explanation for the Bras d'Or depressions is that they are glacial basins. If so, they indicate selective and localized erosion on a scale unmatched in the region.

## Quaternary geomorphic modification

Pre-existing landscapes have been modified by Quaternary erosion and deposition to three main degrees (Fig. 14) which vary according to elevation and rock type. The summit Tertiary peneplain remnants, which are preserved as the upland plateaus on resistant crystalline rocks, are essentially unmodified except for small till mounds, scoured rock areas, and the valleys cut into their flanks. In contrast, the lowland Tertiary peneplain, which is developed on nonresistant sedimentary rock, has been modified by glaciation. The peneplain surface and the valleys cut into it are widely buried under a till blanket of drumlins and morainic ridges 10-50 m thick (Fig. 26); it is locally glacially excavated along the weak rock belts. Intermontane valleys which were formerly narrow rocky gorges have been transformed by deposition of glacial and postglacial gravels into wide alluvial plains ("intervales" in local parlance) (Fig. 22, 27).

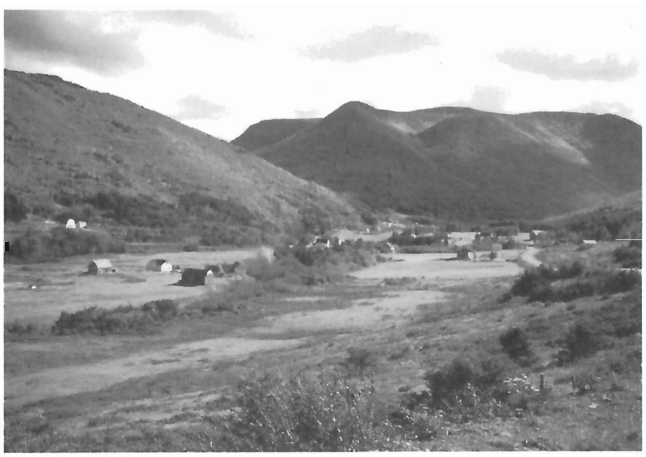


*Figure 27. Shallow glacial outwash plain overlying interglacial marine rock bench, Red River valley. GSC 204024-U*

Widespread dissolution of gypsum bedrock has profoundly modified bedrock topography and, in many areas, also the surface of overlying deposits. Collapse depressions range in size from individual sinkholes, to small gorges (Fig. 28), to large valleys like that of Big Brook quarry. Many valleys are former caverns or cave systems, perhaps unroofed by glaciation. Old karst covered by thick till results in a pseudo-morainic topography, as around River Denys, Little Narrows, Middle River, Chéticamp, and Dingwall. Similarly, some of the glacial gravel in Aspy valley, which has a hummocky surface and is therefore mapped as ice-contact sediment, may actually be outwash that has collapsed into gypsum karst.

Collapse continues today wherever gypsum is in the subsurface. Sinkholes regularly appear in till areas and even on modern floodplains, notably lower Middle River. Collapsed caverns at present sea level, such as on the coast near Wreck Cove head, Ingonish, and Mabou (Fig. 29), probably developed during lower postglacial relative sea levels when the water table was consequently lower and subterranean hydraulic gradients correspondingly steeper.

## *Acknowledgments*

This report is dedicated to the memory of H.L. Cameron, professor of geology at Acadia University, Wolfville, Nova Scotia who pioneered modern Nova Scotian geomorphology, photogeology, and glacial geology. Cameron introduced the author to Cape Breton Island in 1953 as a member of a Nova Scotia Research Foundation underwater archaeological expedition to Fortress Louisbourg and his military connections produced the "Maple Scan" photo coverage. V.K. Prest was the driving force in that he sponsored the project proposal, contributed unpublished notes and maps by E.H. Muller and H.L. Cameron, and reviewed an early draft of this report.

*Figure 28. Uvala or karst valley originating by collapse of cavern in gypsum, Mabou Inlet. GSC 204024-R*

*Figure 29. Former karst valley in gypsum, now partly filled with intertidal salt marsh, Mabou Inlet. GSC 204024-H*

R.H. MacNeill, W.F. Take, W.A. Newman, and W.H. Mathews are thanked for their generous sharing of ideas and unpublished observations. K. Howells of the Nova Scotia Research Foundation loaned some of Cameron's manuscript maps from NSRF files. Resourceful and congenial field assistance was rendered by D.G. Vanderveer in 1970 and 1971 and by C.M. Tucker and P.E. Miller in 1974 and 1975. R.J. Mott provided much of the paleoecological interpretation based mainly on his numerous pollen and macrofossil analyses, and wood identifications, with some also by J.V. Matthews, Jr., L. Wilson, and H. Jetté. Some of the organic sites were discovered by, and others restudied by, A. de Vernal, P. Richard, G. St.-Jean, and S. Occhietti who were instrumental in helping establish the stratigraphic framework based on organic deposits from the last interglaciation. W. Blake, Jr. and R. McNeely expedited radiocarbon age determinations, C. Causse and C. Hillaire-Marcel provided thorium/uranium ages, and N.W. Rutter kindly donated amino-acid racemization ratios. R.B. Taylor brought the Dingwall subtill organic bed to my attention and F. Baechler provided important observations on till stratigraphy in the Sydney and St. Rose areas. R.M. Gagné conducted hammer-seismic profiling of buried valleys.

Helicopter support was provided by the Canadian Coast Guard Service (courtesy L. Hatt) and by the Canadian Forestry Service (courtesy J. Bouzane). Ed Gilkie, boat master, of Pleasant Bay ensured safe passage along the treacherous northern cliffs. Officials of Nova Scotia Power Corporation, Parks Canada, Nova Scotia Forest Industries, Georgia-Pacific Gypsum Corporation, and Kaiser Celestite provided facilities and access to restricted areas. Invaluable discussion in the field was contributed by B.G. Craig, R.J. Fulton, L.A. Dredge, L.H. King, R.R. Stea, and C. Wang and by the participants of 1987 INQUA Congress excursions A-3 and C-3. I sincerely appreciate the hospitality of countless residents who welcomed my visits and shared their knowledge of local features. J.A. Clague, R.J. Fulton, and R.R. Stea are thanked for their reviews of the manuscript. Rachelle Lacroix scanned the airphotos and output all the final illustrations.

## SURFACE MATERIALS AND GEOMORPHIC FEATURES

Eleven primary genetic categories of surficial geological materials are recognized and each has characteristic textural and compositional attributes. The nine with sufficient areal extent are shown on Map 1631A; the other three are either too small to be shown or are in the subsurface. Some of the nine mapped units are further subdivided, as appropriate, either according to landform so as to convey important aspects of process and depositional environment, or according to thickness where this reflects useful textural and stratigraphic properties. The composition, genesis, and stratigraphic order of surface materials was determined by direct observation of natural and artificial exposures. Their extent was mapped by airphoto interpretation. The stratigraphic order and chronology of events are based on 92 key localities which are located on Figure 30. Localities 1-82 are also shown on Map 1631A; the others were found or studied after the map was published. Reference to these localities in the text is by name and site number in square brackets: for example, Green Point [57].

### *Introduction*

#### Identification

The composition of surface materials was determined by ground observation of numerous exposures. Along the seacoast, exposure is virtually continuous because of erosion due to vigorous wave action and rising sea level. The Bras d'Or Lake shore, on the other hand, is relatively sheltered and nearly tideless, so the shore cliffs are normally obscured by dense forest; exposures are small and shortlived, except on exposed headlands. During the field phase of the work, inland exposures were relatively common and fresh along the dense road network which was being upgraded in settled areas and greatly extended in several highland areas.

#### Classification

After ground identification, the extent of the genetic/materials map units was traced on airphotos (despite the almost complete vegetation cover) by means of distinctive patterns of relief, vegetation, and drainage. In rough stratigraphic order, based on vertical succession, the eleven primary units and their main characteristics are: R - bedrock (obvious structural lineaments where exposed as well as where vegetation covered); unit 1 - residuum (degraded terrain devoid of either structural or glacial features); older marine and organic beds (largely in the subsurface; not shown on Map 1631A); unit 2 - till (typical morainic and ice-moulded landforms); unit 3 - glaciofluvial deposits (kame-and-kettle topography, high-level alluvial terraces); unit 4 - glaciolacustrine deltas (small terraces perched at stream mouths); unit 5 - fluvial deposits (flat, sloping, channelled surfaces); unit 6 - marine deposits (modern lake and sea beaches); unit 7 - colluvium (gravity-graded slope deposits); eolian deposits (modern sand dunes too small to show on Map 1631A); unit 8 - organic deposits (vegetal accumulations of bogs, fens, swamps, and marshes). These units belong to four broad age groups: 1) pre-Quaternary (bedrock), 2) pre-Quaternary and/or Quaternary (residuum), 3) Sangamonian and Wisconsinan (older marine and organic beds), and 4) Wisconsin and Holocene (all others).

#### Landforms

Landforms or geomorphic features are recognizable elements on the Earth's surface which have characteristic shapes and are the result of natural processes. Form and implied origin provide keys to the likely composition of surface materials. Map 1631A shows by means of symbols those landforms which were used to help delineate and interpret the surface materials. Landforms in bedrock include stratification ridges, fracture lineaments, and karst topography. Constructional glacial landforms include drumlins and drumlinoids, crag-and-tail hills, and moraine ridges. Erosional glacial features include cirques, roches moutonnées and stoss and lee

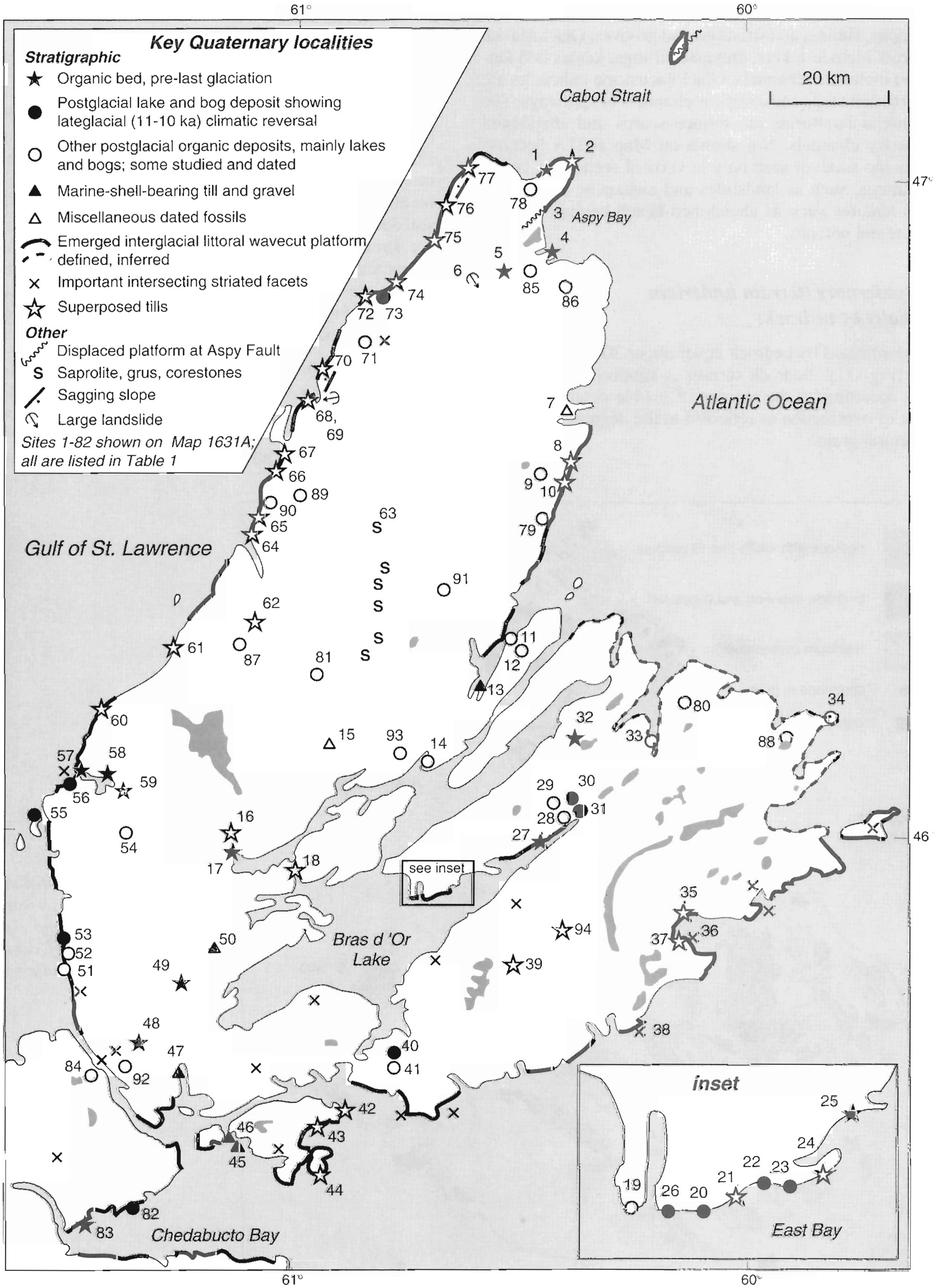



*Figure 30. Important stratigraphic localities; numbers refer to entries in Table 1; 1-82 are also shown on Map 1631A.*

topography, fluting, and striations and grooves. Glaciofluvial landforms include eskers, crevasse fillings, kames and kettles, and meltwater channels. Glaciolacustrine indicators are limited to delta tops and overflow channels or spillways. The only fluvial landforms are terrace scarps and abandoned distributary channels. Not shown on Map 1631A because they are too small or seen only in vertical section are colluvial features, such as landslides and avalanche tracks, and marine features such as abandoned beach berms, wavecut benches, and notches.

## *Pre-Quaternary (terrain underlain essentially by bedrock)*

Areas dominated by bedrock cover about 30 per cent of the island (Fig. 31). Bedrock terrain is subdivided into three classes according to the amount of visible outcrop and the amount of overburden as reflected in the degree of masking of structural grain.

### Exposed bedrock (unit Ra)

On airphotos, exposed rock is recognized by the presence of structural lineaments, namely stratification and foliation ridges, and narrow linear depressions following fracture lines (Fig. 25). Structure lineaments are included on the map to illustrate the structural fabric, especially in areas where they might be mistaken for drift features. Barren bedrock is due mainly to glacial erosion; locally slope processes and loss of soil due to fire have caused further denudation. Rock outcrops are streamlined into basins and knobs with stoss-and-lee asymmetry, and erratic boulders of all sizes litter the surface. On the ground, ice-scoured rock surfaces commonly display a variety of intersecting ice-flow markings such as striations, grooves, miniature crag-and-tail features, and beveled facets.

Bare-rock terrain occupies about two per cent of the island. It occurs throughout the area on all bedrock lithologies, but is generally found in higher areas and near the coast. The largest expanses are on granite north of Chéticamp and Ingonish where repeated fire has allowed the thin drift to

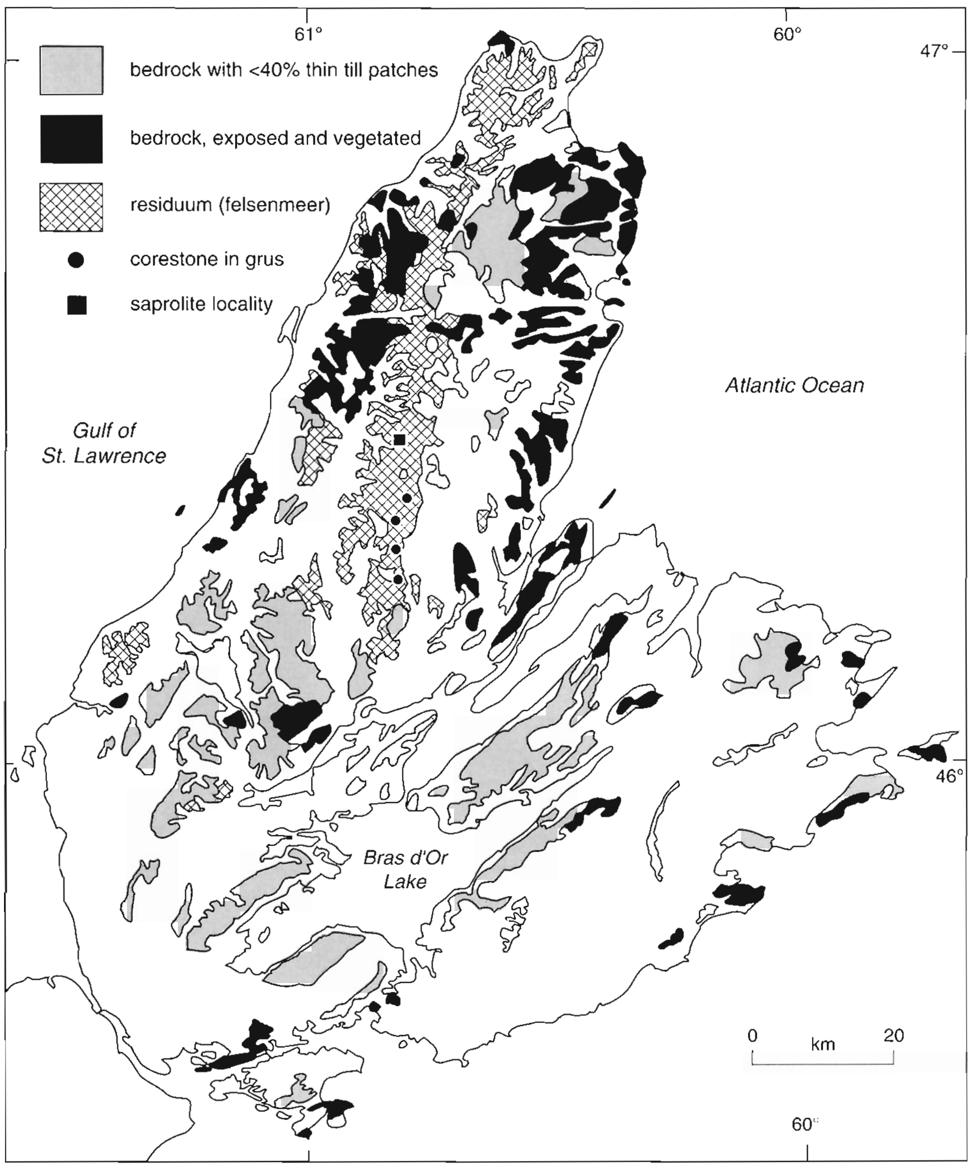



***Figure 31.***

*Map showing distribution of rock-dominated terrain, rock with minor till, and residuum with locations of corestones and saprolite; (adapted from Map 1631A).*

be washed away. Modern wave action and slope washing explains the bare cliffs on the northern highlands coast (Fig. 19) and the coastal outcrops around Louisbourg, Gabarus, and on Isle Madame. Elsewhere, small barren areas near Wreck Cove, Baddeck Lakes, and Lake Ainslie correspond to expanses of resistant siliceous rocks where no till was produced; those on the summits of Kelly Mountain, Coxheath Hills, and East Bay Hills are denuded by glacial scour.

### Bedrock obscured by forest (unit Rb)

Bedrock terrain covered by forest and other vegetation is recognized by small scattered outcrops and by structural relief patterns that are almost as clearly visible as in areas of exposed bedrock. Surficial sediment may be present in thin patches and in depressions. Practical considerations justify the distinction of this unit when it is necessary to know where bedrock would be exposed if vegetation were not present. Bedrock mappers and mineral explorationists will know that, while few outcrops may be visible on airphotos, they will likely be common in the forest and that bedrock will be found at shallow depths. Road builders will be aware that little, if any, overburden will be available for grading and levelling. Those engaged in reforestation will know that, although overburden is thin, it is sufficient to support relatively good forest.

Forested rock terrain occupies 10 per cent of the area and is found at all elevations, usually marginal to bare-rock areas (Fig. 31). Large expanses occur on crystalline rocks around

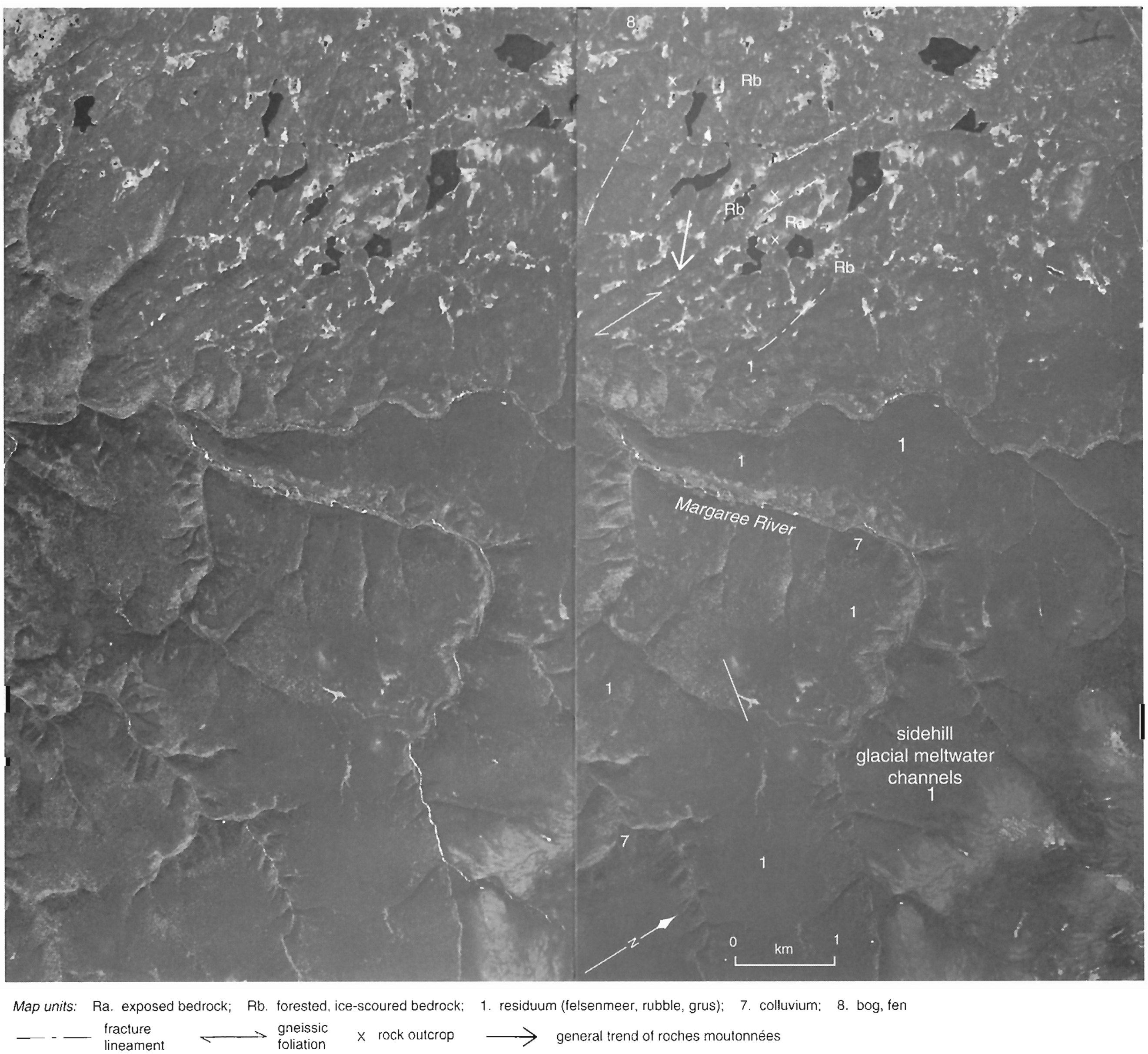



*Figure 32. Stereogram showing contrast and sharp contact between ice-scoured bedrock terrain and residuum terrain, locally with small sidehill meltwater channels; highlands east of Chéticamp. NAPL A21576-31, 32*

the outer parts of the highlands plateau (Fig. 32), and on other upland summits such as Ainslie Hills, Kelly Mountain, Boisdale and Coxheath hills, and in East Bay Hills. On lower ground, obscured bedrock occurs on sandstone in the Sydney and Lennox Passage areas, among others, on siliceous and volcanic rocks of the Fourchu Group on Scatarie Island, in the Gabarus-Fourchu coastal area, and on Isle Madame. As with bare-rock areas, this terrain has been produced by glacial erosion over resistant rocks and over higher ground where there was minimal local till production and no influx of foreign drift.

### Bedrock with undifferentiated patches of thin till (unit Rc)

About 18 per cent of Cape Breton Island is bedrock-dominated terrain with a patchy cover of till (Fig. 31). This unit constitutes the transition between bare (Ra) or vegetated bedrock (Rb) and complete till cover; its contacts with these units are therefore usually gradational. Till comprises up to 40 per cent of the map-unit area, occurring locally as discrete morainic patches and ice-moulded forms such as small flutes, drumlins, and crag-and-tail hills whose distribution varies locally in relation to rock type, relief, and direction of glacier flow. It occurs throughout the island, but is typically located on harder rocks and high areas. On the highlands, large areas are transitional between terrain dominated by bedrock and those with essentially complete till cover. Tracts occur on resistant sandstone ridges near Sydney and St. Peters, and on other resistant sedimentary rock knolls north of Bras d'Or Lake and east of George Bay. A belt borders the coast between Gabarus and Main-à-Dieu. In general, this terrain is attributed to settings where glacial erosion prevailed over production and lodgment of till.

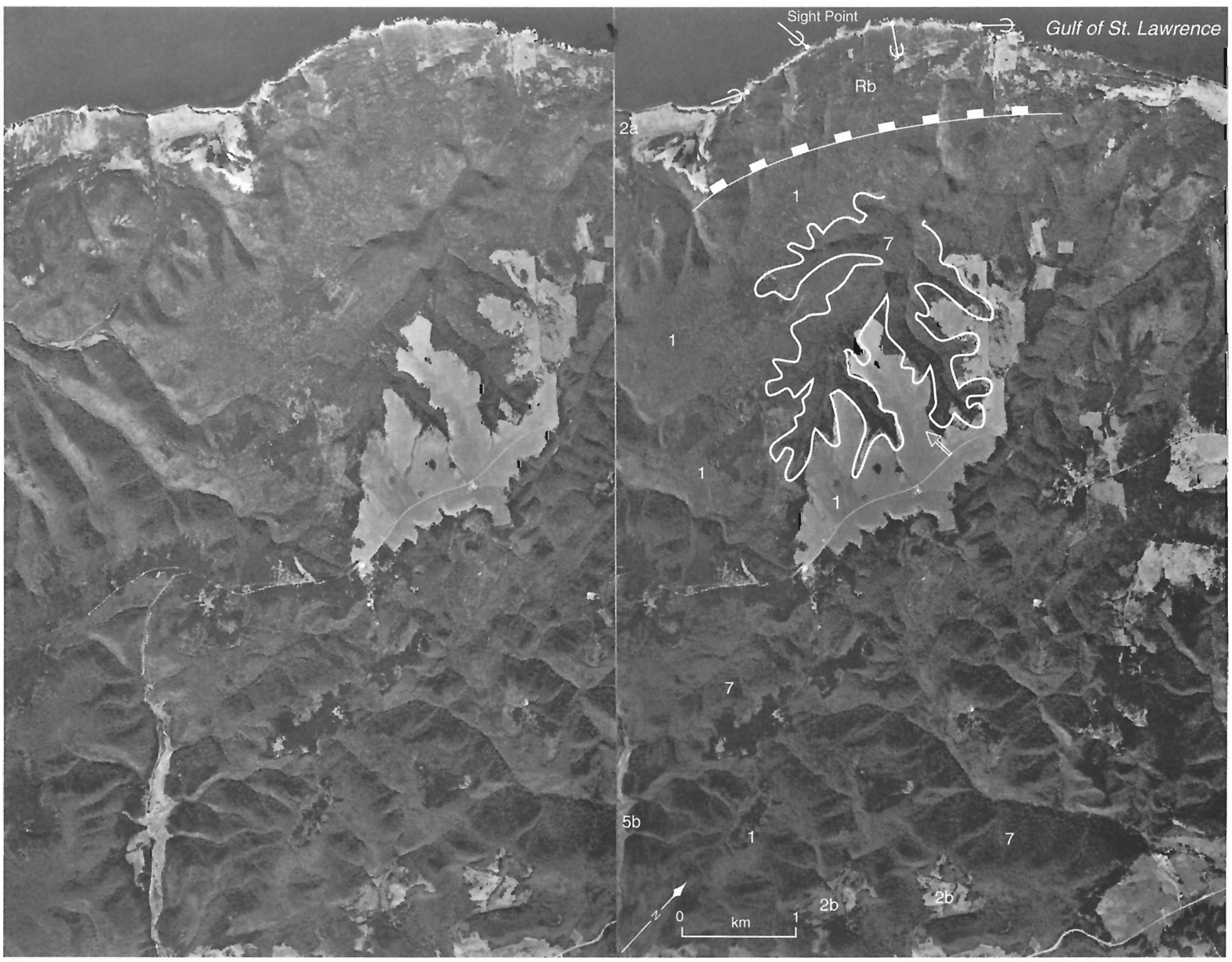



*Figure 33. Stereogram of Mabou Highlands, a dissected, residuum-covered pre-Quaternary peneplain with glacial trimline. NAPL A21536-84, 85*

## *Pre-Quaternary and/or Quaternary (residuum and regolith; unit 1)*

Broken and rotted bedrock, called residuum or regolith, covers about 10 per cent of the island (Fig. 31), chiefly the central third of the highlands plateau, the northern part of Creignish Hills, and the summit of Mabou Hills. The surface of such residual areas is flat to very smoothly rounded (Fig. 12, 33) and devoid of bedrock structural grain and glacial landforms such as mounds and basins. The surface is composed essentially of disaggregated, disintegrated, and altered rock mainly in the form of rubble ("felsenmeer"), with smaller areas of sandy debris ("grus") and one occurrence of maturely weathered paleosol ("saprolite"). The extensive tracts of residuum have rare patches of till and occasional erratics, all of which apparently are dark crystalline rock types from the western part of the highlands; no erratics of sedimentary rock (for example, limestone, gypsum, red sandstone) from the Gulf of St. Lawrence lowlands were seen.

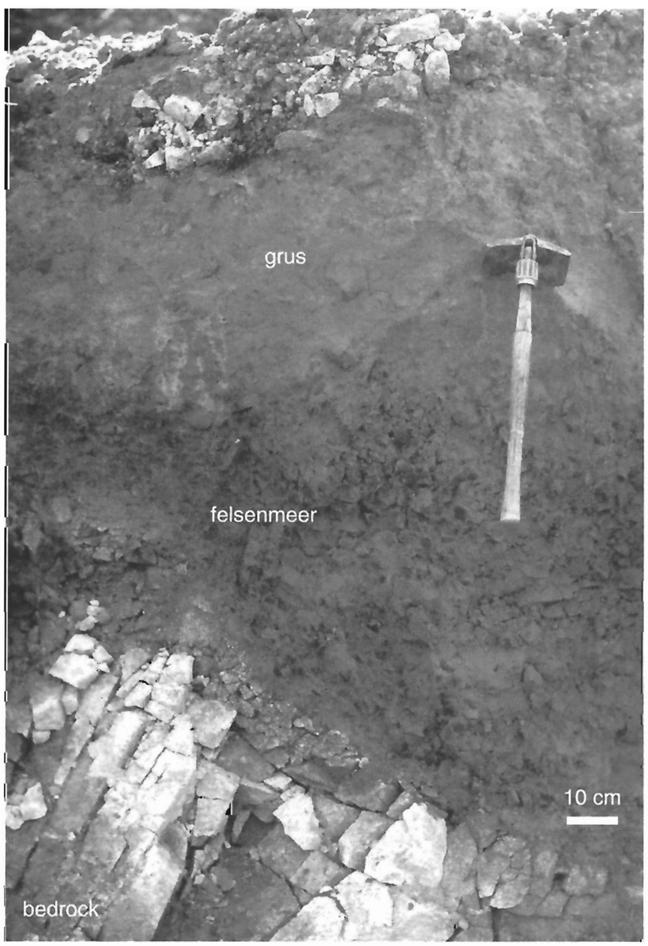



*Figure 34. Exposure of typical residuum composed of disaggregated fractured bedrock (felsenmeer) with interstitial red iron-stained sand (grus); roadcut on highlands east of Belle Côte. GSC 203752-U*

### Rock rubble or felsenmeer

A mantle of rock fragments derived from the underlying solid bedrock covers most of the surface to an average depth of 2-3 m. Without forest cover, the surface would be a vast blockfield or felsenmeer. Typically, the rubble consists of angular joint-bounded fracture blocks that are essentially unaltered except for kaolinization on the surfaces of some feldspars. Near the surface, the fragments are set in a loose sandy matrix or "grus" consisting of finely disintegrated rock that has been reddened or "rubefied" by accumulation of iron compounds in interstices. The rubble grades downward into progressively less fractured, more coherent rock (Fig. 34). Presumably the primary disaggregation is the result of frost action on fractured bedrock, and the iron and secondary clays are the product of pedogenesis.

Contacts with the adjacent till-covered or glacially denuded areas vary from gradational to sharp. On the one hand, a belt about 5 km wide has till outliers in the felsenmeer, and grus shows through windows in the till; on the other, the felsenmeer is laterally transitional to denuded bedrock or till over less than 500 m. In a few places where contact relations can be observed, the felsenmeer is worked into till in down-ice directions, and is overlain by till in up-ice directions, thus proving that the residuum predates the last glaciation of the

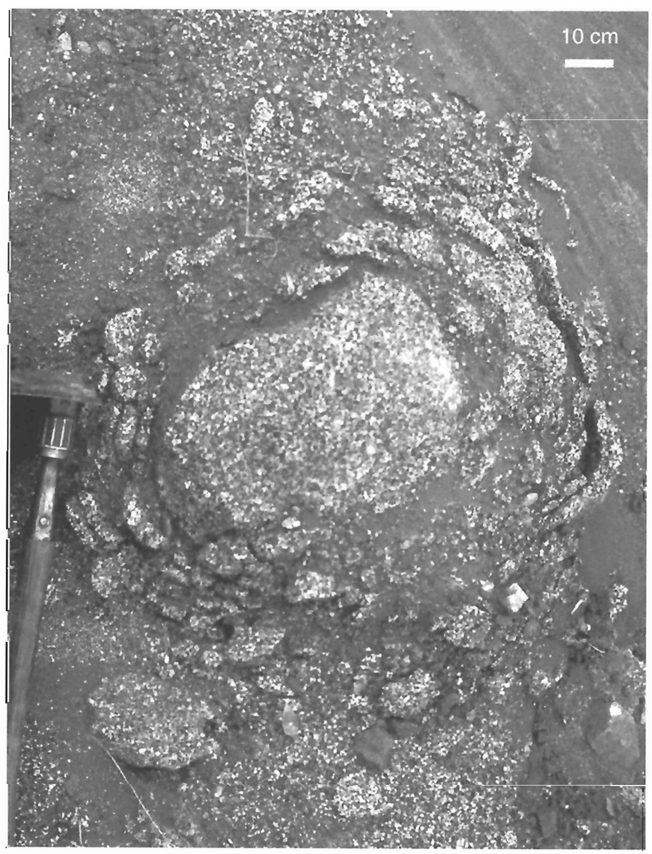



*Figure 35. Cross-section through granite corestone in borrow pit; note exfoliation and enclosing red grus: Highland Road, northern plateau. GSC 202188-W*

highlands. As further discussed below, the fact that residuum is preserved on the higher, central part of the plateau suggests that glacial erosion was infrequent and/or ineffective, perhaps because the ice was cold-based and hence frozen to the bed, rather than sliding over it.


*Figure 36. Exfoliating granite corestones; these boulders are commonly mistaken for glacial erratics; Highland Road on northern plateau. GSC 202188-U*

### Grus and corestones

Locally, massive granites and granite gneisses have disintegrated by grain release to the point where only scattered, metre-sized, subspherical masses of coherent rock or corestones remain in a groundmass of iron-stained gravel or grus composed of individual mineral grains (Fig. 31, 35). Borrow pits reveal that the disaggregation locally exceeds 5 m depth. The corestones are concentrically fractured (Fig. 36), have a rough pitted surface, and feldspars are kaolinized.

### Saprolite

One exposure of advanced chemical weathering of bedrock was seen in a shallow roadcut on Second Fork Brook Road South [63]. There, more than 3 m of intensely altered biotite gneiss is overlain by sand (redeposited grus) and till (Fig. 37). The rock mass is essentially in place, as shown by quartz veins, biotite foliae, and pegmatite masses which are intact and fresh, except for a slight downhill bend probably due to creep (Fig. 38). The gneiss has been rotted to a "saprolite" with cheesy consistency that can be easily shaved with a knife. It contains abundant and relatively mature clay minerals, chiefly kaolinite and gibbsite, which reflect an advanced state of alteration of feldspars (McKeague et al., 1983). The saprolite is overlain sharply by an overlying stratified bed of sand-sized grus and saprolite particles, possibly deposited by sheetwash or rillwash, and in one place, a large volcanic erratic is imbedded in the sand. Saprolite and overlying sand are truncated by 1-2 m of reddish-brown sandy till with striated stones including a few fragments of red siltstone from the Carboniferous lowlands to the west. Till has been forced into the sand and saprolite in an uphill direction so as to form a nose that is oriented as if emplaced in a direction toward

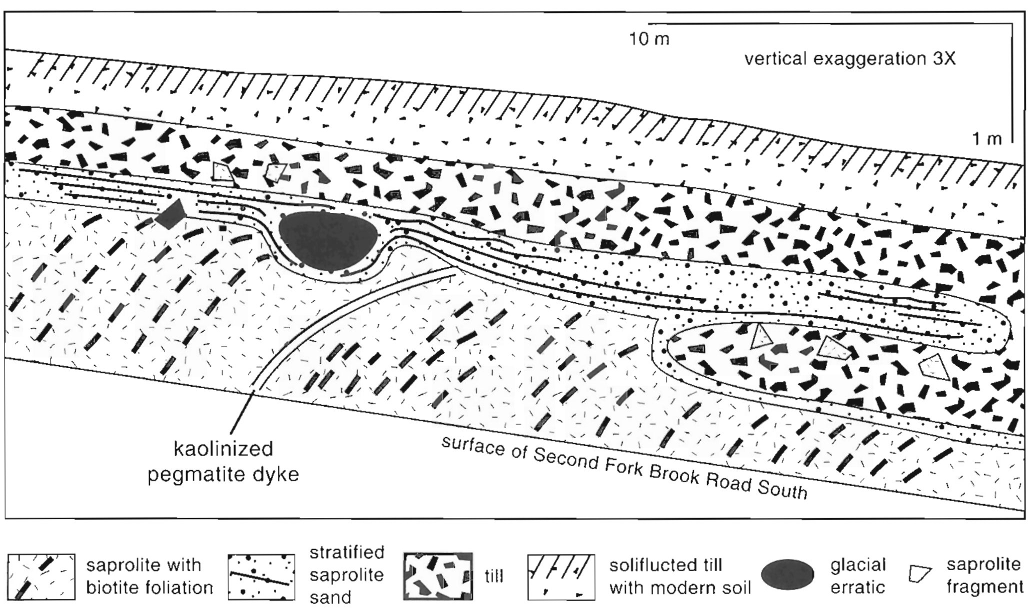



*Figure 37. Geological cross-section of saprolite [63]; note distortion of gneissic foliae, glacial erratics pressed into saprolite, and till nose on right with axis oriented 105°; Second Fork Brook Road South, northern highlands (from Grant, 1987, p. 39).*

105°. Presumably the saprolite is part of a preglacial soil, but it is not known how much has been removed or how deep it extends.

The age and origin of such intensely weathered rock cannot be determined directly but this material can perhaps be compared to saprolites occurring beyond the limit of glaciation in the southeastern United States where the climate has been generally humid and subtropical throughout Cenozoic time. The Cape Breton Island saprolite and associated residual materials may therefore be relics of warmer pre-Quaternary climates. Support for the hypothesis that they are remnant soils of an ancient landscape is the fact that they are situated on a pre-Quaternary degradation surface considered to be an essentially intact Tertiary peneplain. Similar regoliths have been reported elsewhere in the Atlantic Provinces region, notably northern New Brunswick (Wang et al., 1982) and eastern Quebec (LaSalle et al., 1985) and are inferred to be pre-Quaternary warm-climate soil remnants. Preservation of both the residual materials and the peneplain of which they seem to be an integral part is significant to the interpretation of Quaternary processes, especially the vigour and frequency of glaciation, as discussed in the Cenozoic history section.

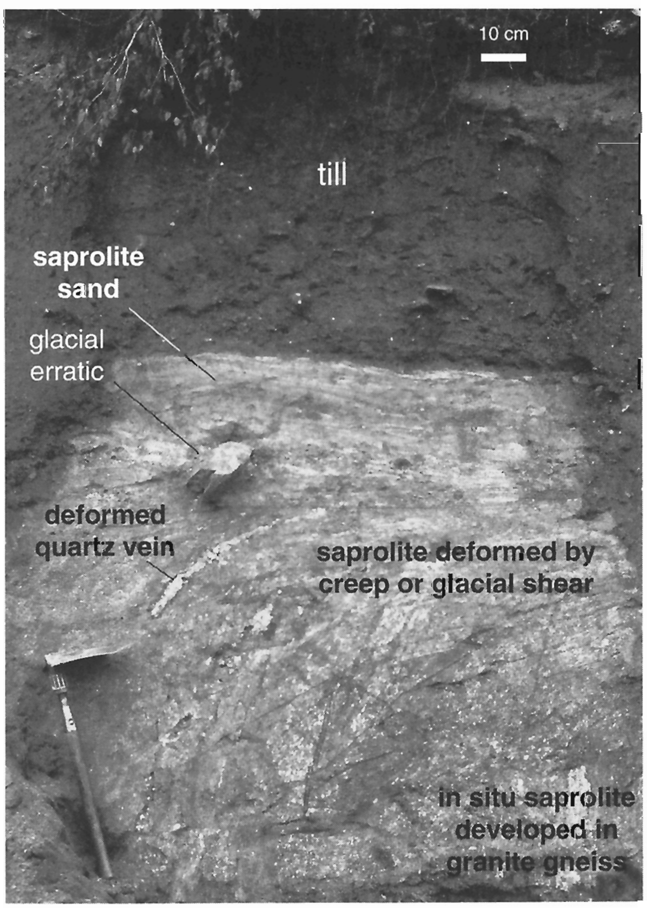



***Figure 38.*** *Photo of till overlying saprolite, showing downhill creep of gneissic foliae; Second Fork Brook Road South. GSC 202188-R*

## *Sangamonian and Early/Middle Wisconsinan deposits*

Locally underlying the surface till(s) is a complex of various organic and organic-rich sediments, colluvium, beach gravel, and an emerged intertidal wave-cut rock bench. The organic beds have been dated by the $^{14}C$ and Th/U methods to the period 130-47 ka. Most fall in the range 125-75 ka and therefore belong to the last or Sangamonian interglaciation (sensu lato) which is equated with oxygen isotope stage 5, the boundaries of which were defined at 127 and 75 ka by Broecker et al. (1968). The younger members of the group, if reliably dated, might be referred to the earlier part of Wisconsinan time.

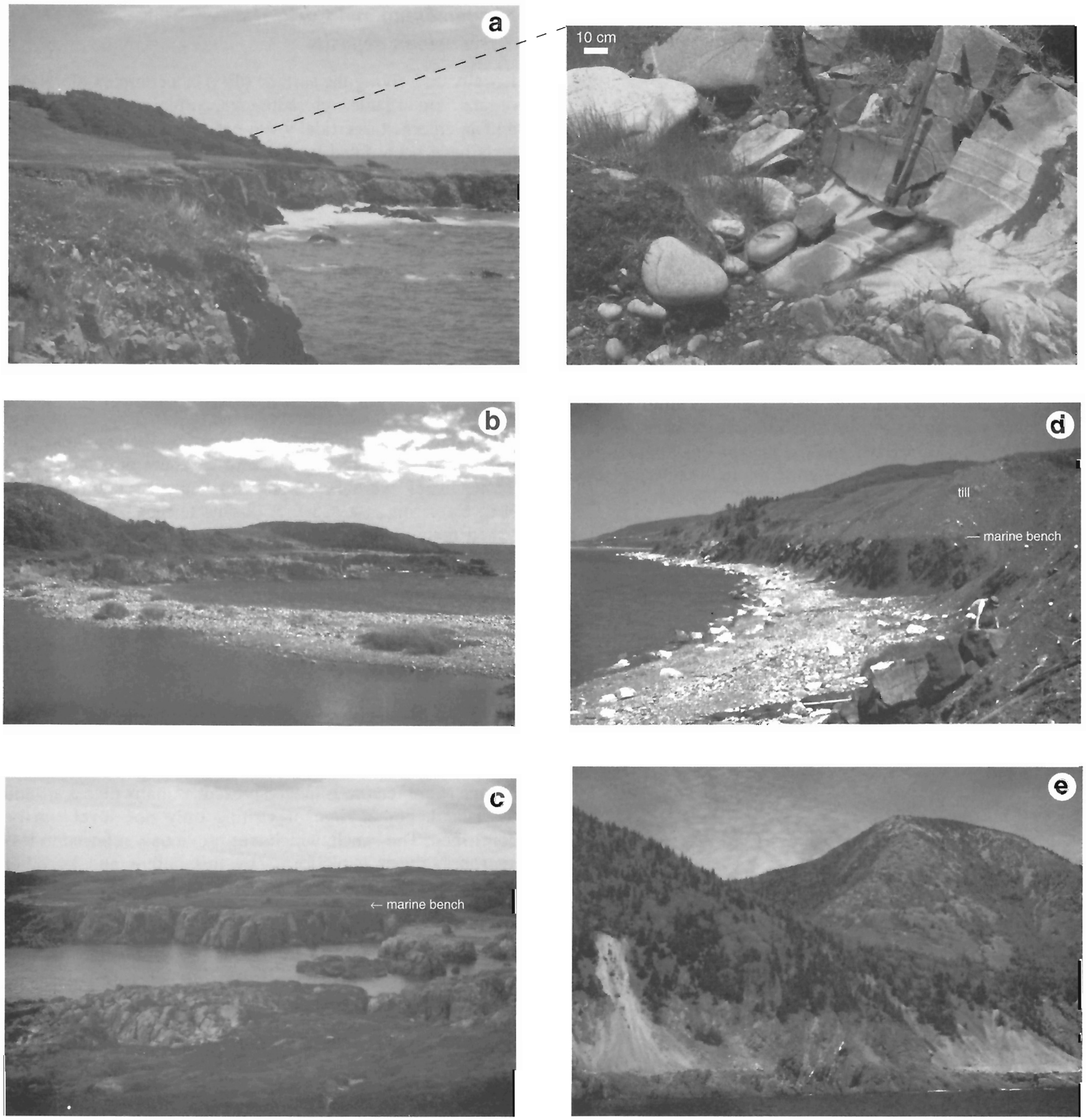
### Emerged littoral wave-abrasion rock bench

MacNeill (1952) was apparently the first to draw attention to till-covered, raised marine benches a few metres above tide level on the George Bay coast of adjacent mainland Nova Scotia. They were first noted on Cape Breton Island by H.L. Cameron as a "25-foot marine terrace" on an unpublished manuscript map dated 1957, but locations were not given. The first definitive published interpretation of emerged marine shore features was in marginal notes on bedrock maps of the Aspy Bay area by Neale (1963, 1964):

'Tilted wave-cut rock benches, formed during one or several periods of relatively higher sea level, are overlain by stratified fluviatile deposits. The benches range in elevation from 3 to 35 feet above sea level, possibly because of post-glacial faulting of an original bench at about 20 feet.'

The present work has verified that a prominent raised marine shore feature is indeed present at many places around the island, but at time of writing only one level can be confirmed. The bench, which was previously referred to less appropriately as a platform by this author and by other authors, is locally overlain by well-rounded beach gravel and occurs along lengthy segments of the entire Cape Breton Island seacoast (Fig. 30), and parts of Bras d'Or Lake, a marine embayment. It may also be present along the low sandstone coast between Great Bras d'Or and Main-à-Dieu. The rock there is truncated to a gently undulating till-covered plain generally less than 10 m above tide level. Lengthy segments lie at 5-6 m or lower but these have no clear limit against the higher parts, so if the bench exists along this part of the coast, its extent cannot yet be defined.

Where considered to be defined, the bench appears as a single, obvious, planar truncation of bedrock up to a few hundred metres wide with a clear inner notch or steep declivity against the surrounding terrain (Fig. 39a). Where not obscured by later deposits, it is glacially rounded to varying degrees (Fig. 39b). Rarely, there may be other vague levels at slightly higher elevations (Fig. 39c), but they may be more apparent than real. In any case, their origin as marine benches cannot be affirmed. At present, therefore, only a single marine bench has been established. In a few valleys, it can be traced a short distance inland as the rock floor or thalweg of a stream, such as Morrison Brook [8] and Red River [74]. Typically the bench is glacially striated and overlain by till(s) (Fig. 39d)




*a.* *Bench with paleocliff glacially rounded and partly infilled with till; east side of St. Paul Island, Cabot Strait; GSC 203752; Inset shows detail of inner notch, GSC 204042-I;*

*b.* *Glacially abraded bench at Gooseberry Cove near Louisbourg where 5.5 m elevation determined (hill in rear, which was island during higher sea level, now has till lodged on northern lee side). GSC 204043*

*c.* *Prominent 5 m emerged bench and possible second higher level; west of Little Lorraine. GSC 202188-K*

*d.* *Measuring striations on emerged bench where overlain by tills; near Creignish, George Bay. GSC 204024-K*

*e.* *Inactive talus cones and aprons overlying emerged 7 m bench cut across granite and greenstone; Aspy scarp near Money Point. GSC 204042-K*

*Figure 39. Representative views of the emerged Sangamonian intertidal rock bench.*

showing that it predates the last glaciation. Locally, it is buried only by muddy colluvium or talus (Fig. 39e). In rare instances it is overlain by organic beds which, by virtue of their dateability, provide minimum age estimates for the bench. The surface is typically smooth or mammillated (Fig. 39f) and is overlain by a few metres of sand with well-rounded cobbles and boulders (Fig. 39g), some percussion marked. Exposures perpendicular to the coast, as at Aspy Fault [3] and Morrison Brook [8], show that the bench has a distinct seaward gradient of about 1:5, much like its modern counterpart. The inner edge, where rarely visible, is a smooth, rounded notch (Fig. 39h) cut into a paleocliff. The emerged bench has all the features which characterize the modern intertidal abrasion surface, so it is believed to be the paleoshoreline of an earlier, relatively higher sea level.

There is a considerable variation in the present elevation of the emerged bench, measured with respect to mean annual high tide, also the upper limit of its modern counterpart. The elevational pattern is also discordant with other emerged shorelines in the region. As seen in section beneath till, the bench edge ranges in elevation from about high tide level in the south to just over 7 m in the north, but much of this apparent range is skewed toward lower values because many exposures are obviously positioned at lower points along its seaward slope. The maximum local elevation was measurable only at those three locations where the inner notch was visible: 7.2 m at Money Point (Fig. 39h) and 6.7 m at Delaney Brook (Fig. 39f) in the northern part of the island, and ≈5 m at Gooseberry Cove (Fig. 39b) on the southeastern coast. It seems to be at least 4 m in the Gabarus Bay area, but numerous segments from Framboise to Cabot Strait were not higher than 2 m (although the inner limit was not seen). It is at 3.7 m or slightly higher at Moose Point [83] on Chedabucto Bay, mainland Nova Scotia. The bench thus slopes very slightly down to the south with a gradient of 3-5 m over this 175 km north-south distance.

This relatively gentle tilt of the emerged bench contrasts with the gradient of late glacial and postglacial shorelines in the region which are isostatically tilted about 75 m over the same distance (Grant, 1989). It is therefore probably not of glacial age, otherwise it would be much more strongly warped. Because the bench is nearly parallel to present interglacial tide level (sea level), it likely represents a sea level position during a nonglacial period comparable to the present (in geodynamic terms). In other words, it is probably a former interglacial shoreline. According to Shackleton (1987), of the 30 or so interglacial sea level high stands during the last 2.3 Ma, only a few reached higher than present interglacial sea level. However, in this area, because the crust has been subsiding >5 m/Ma (Jansa and Wade, 1975), all of those older interglacial shorelines, except that of the last interglaciation, would have been lowered below present sea level. The sea level culmination dating from the warmest part of the last or Sangamon interglaciation (oxygen isotope substage 5e, ≈125 000 BP) is well documented throughout the globe. Its elevation ranges 2-9 m above sea level in stable areas, such as eastern United States, Caribbean, Mediterranean, Africa, Australia, and many Pacific atolls (Marshall and Thom, 1976;




*f.* *Typical wave-abrasion potholes on emerged bench; Delaney Brook, northwest coast. GSC 204024-I*

*g.* *Emerged rock bench overlain by beach gravel, outwash, and colluvium; Red River [74]. GSC 203504*

*h.* *Smooth and polished notch (arrowed) marking inner margin of emerged bench at its highest elevation (7.2 m) at Money Point. GSC 203752-W*

***Figure 39.*** *(cont.)*

Moore, 1982; CLIMAP Project Members, 1984). On this basis, the raised intertidal wavecut rock bench, which ranges up to 7 m elevation in Cape Breton Island, has therefore been correlated with the highest sea level of last interglaciation (Grant, 1980, 1990) which was attained during oxygen isotope substage 5e.

Attention has been drawn to the value of this feature, and interglacial shorelines in general, as geodynamic datum planes (Grant, 1980, p. 203) for assessing the present degree of crustal equilibrium. Since this shoreline marks the interglacial transgressive maximum, it represents a condition when sea level is at its highest and global temperature presumably at its warmest. Both conditions thus reflect minimal global ice volume, and consequently, the greatest degree of glacio-isostatic crustal equilibrium in an area. Like any tilted shoreline, the slight nonparallelism of the interglacial bench to present sea level indicates that crustal and/or geoidal changes have occurred in the interim, or that isostatic equilibrium has been not been re-established. Further, it may be noted that at present rates of relative sea level rise, the bench will be re-occupied by the sea in about 2000 years.

Further support for the last interglacial age assignment of the bench, apart from the fact that it is glacially striated and overlain by tills ascribed to the last glaciation (for example, Fig. 39d), comes from overlying organic sediments most of which have been dated isotopically to the Sangamonian Stage, as detailed in the Stratigraphy section. At present, the five best dates forming the basis for assigning the bench and its associated overlying beach gravel to the Sangamonian sea level high stand are: the bench is associated with peat yielding a minimum Th/U age of 117 ka at Green Point [57]; the associated gravel overlies peat dating ≥126 ka (Th/U) at East Bay [27]; the associated gravel is overlain by organic mud assigned an age of 62 ka on palynological grounds at Castle Bay [25]; the bench is overlain by peat which has a $^{14}C$ age of 47 ka at Bay St. Lawrence [1]; and a probably correlative beach gravel is mixed in a sinkhole with peat with a $^{14}C$ age of >39 ka at Dingwall [4].

### Organic beds underlying the glacial deposits

At thirteen localities, organic and other stratified nonglacial sediments underlie one to three tills and other glacial and nonglacial sediments. The sites are described and illustrated in the Stratigraphy section; the salient features are summarized here, based on studies by Mott and Prest (1967), Mott (1971), Occhietti and Rutter (1982), Guilbault (1982), de Vernal et al. (1983), Mott and Grant (1985), de Vernal and Mott (1986), and de Vernal et al. (1986).

The organic material includes woody peat, gyttja, and lacustrine and marine silt. (Till-like material was found in two of these organic sequences, but it could not be demonstrated to be in situ and of glacial origin, so concrete evidence of pre- or intra-Sangamonian glaciation remains lacking). Ages on the organic material are estimated from Th/U ratios on included wood and range back to 126.4 ka. Ten of the organic beds, which date 82-126 ka, have temperate-climate pollen assemblages indicating interglacial conditions, whereas three of the organic beds, which yield ages of 47-62 ka, have cooler-climate assemblages. Most therefore appear to belong to the last or Sangamonian interglacial Stage (oxygen isotope stage 5; 127-75 ka). Pollen analysis of individual beds reveals vegetation assemblages ranging from thermophilous hardwood forest to shrub and grass tundra. The pollen stratigraphic studies cited above are unanimous in concluding that the sediments record six major distinctive partial and complete cycles of warming and cooling. In the temperate group of beds, there are three warm culminations assigned to oxygen isotope substages 5a, 5c, and 5e of the Sangamonian Stage partly on the basis of age estimates, and, in the cool-climate beds, three alternations between boreal forest and tundra forest, each cooler than its predecessor, which give ages <80 ka and are assigned to Wisconsinan time.

These Sangamonian and Wisconsinan organic beds, together with the underlying littoral bench and gravel, are overlain by the lowest till in the multi-till sequence and the bench is inscribed by the oldest set of striations in the succession of ice-flow phases. These beds thus place a lower limit on the beginning of the last or Wisconsin glaciation in Cape Breton Island and, by extension, help constrain the glacial history for the surrounding region, including offshore areas.

## *Wisconsinan and Holocene deposits and features*

### Glacial deposits (till)

Till, which is glacial sediment deposited directly beneath or in front of a glacier, covers about 48 per cent of the area (Fig. 40a). Its thickness, extent, texture, and lithology vary widely, but there are general relationships with topographic setting and rock type (Fig. 40b). On the resistant crystalline rock uplands, till is thin, patchy, relatively coarse, and with a sandy and silty sand texture (Fig. 40b, inset). It is usually present as a single sheet and locally derived. In contrast, till on the easily erodible, sedimentary rock lowlands, particularly shale and gypsum areas, is generally thicker, more extensive, finer textured (silty clay) and with few or no cobbles and boulders. It is also more lithologically diverse and is commonly stacked in lithologically dissimilar sheets. Till-covered areas are geomorphically youthful with steep, 30° angle-of-repose slopes on morainic topography and bedrock surfaces are unweathered, freshly striated, and polished in the minor included areas of outcrop. Till areas have characteristic surface forms (symbolized on Map 1631A): streamlined ice-moulded features (drumlins, flutings, crag-and-tail hills) produced by active ice flow; hummocky to rolling morainic topography produced by stagnation and disintegration; and ridge-form end moraines deposited at stable ice fronts. An exception is a small till area near Creignish (Fig. 41) which has long, anomalously smooth and gentle (<10°) slopes with catenary profiles, as if graded by solifluction. The till sequence, together with various crosscutting glacial erosional features, documents a succession of several local and regional ice-flows from different directions.

Lithology and texture of the surface till broadly corresponds to bedrock type (Fig. 40b; Fig. 40b, inset), but because of the variety of ice-flow direction, there are wide lateral and vertical variations, considerable admixture of pebbles and cobbles, as well as matrix components, from other source

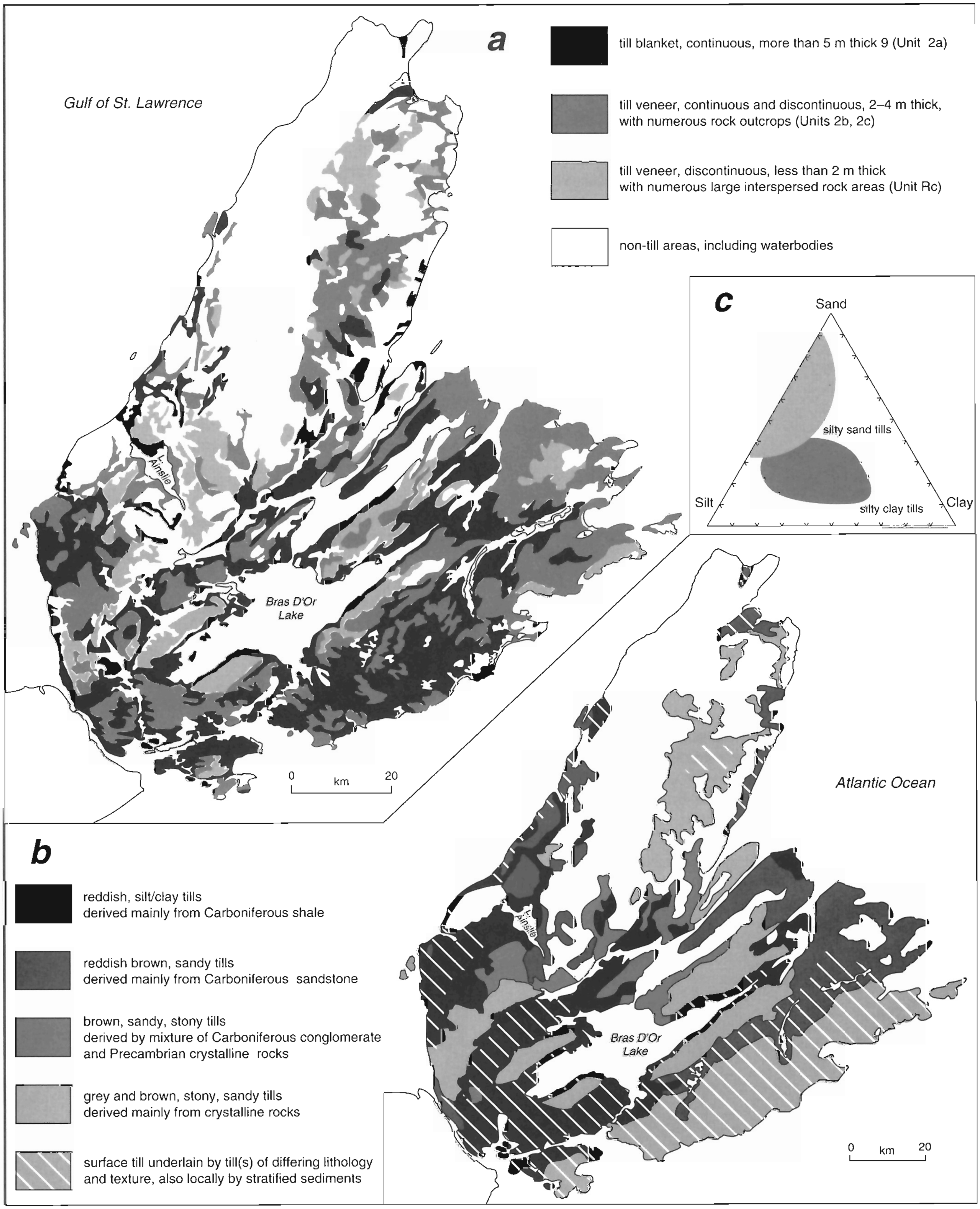



*Figure 40. Generalized thickness, continuity, lithology, and texture of till. **a.** Till areas differentiated by thickness and continuity (adapted from Map 1631A); **b.** Till areas differentiated by gross colour, texture, and lithology (adapted in part from Cann et al., 1963). Note that large areas are underlain by superposed dissimilar tills; **c.** Main textural groupings of the four general lithological types.*

terranes, and a locally extensive overlap of till sheets onto unrelated terranes. Each of the score or more main bedrock types has thus given rise to a lithologically distinct till, but the scale of mapping precluded differentiating each of the numerous lithological and textural phases. In general, however, a suite of loose sandy, locally-derived tills characterize the crystalline rock uplands, whereas various cohesive silty tills with a high proportion of transported stones occur on the sedimentary rock lowlands. Most till species generally have several textural phases occasioned by the degree of introduced material from adjacent terranes. Examples of tills overlapping onto adjacent source terranes are: the fine-textured tills derived from the weak-rock sedimentary lowlands which have been spread onto the crystalline uplands, notably from north to south onto higher ground around Bras d'Or Lake and the stony, sandy tills from the crystalline rock uplands which overlap the lowlands bordering on the south. Thick till accumulations commonly comprise more than one till sheet, usually of quite dissimilar composition. Sequences of multiple tills are common over a broad area of the Atlantic coastal zone from Isle Madame to Louisbourg, along the coast of East Bay, over the River Denys and Mabou lowlands, and along the highlands coast; the major exposures serve as reference stratigraphic sections (Fig. 30).

Till-covered areas are subdivided on the basis of thickness and areal continuity relative to the amount of bedrock outcrop, as inferred both from the size of the till surface forms and from the degree to which they mask the generally rugged 10-20 m local relief which has been glacially etched into the bedrock surface almost everywhere. It may be noted that, on the otherwise smooth residual surfaces of the upland plateaus, the presence of patches of till is revealed on airphotos by telltale undulations of typical morainic topography. Till thickness was chosen as a mapping parameter for Cape Breton Island because it conveys information on texture as well as subsurface stratigraphy because the thicker accumulations tend to be finer grained and to consist of several dissimilar superposed sheets. The three categories of till deposit are: till blanket, continuous till veneer, and discontinuous till veneer.

*Till blanket (silty; unit 2a)*

Till blanket is so named because its thickness is sufficient (5-50 m) to mask the erosional structural relief of the underlying bedrock. On airphotos, areas of till blanket stand noticeably higher than adjacent terrain where bedrock relief is visible. The surface is typically either undulating with disorganized or ridged morainic landforms or streamlined with elongated ice-moulded forms such as drumlins, flutes, and crag-and-tail hills; chaotic and streamlined till terrains are commonly interspersed as a mosaic. Till blanket covers about 25 per cent of the island, mainly in southern lowland areas. It is mainly composed of fine grained clay/silt tills which, on the lowlands, are largely derived from the nonresistant Carboniferous shale, siltstone, mudstone, and gypsum and, on the highlands, are derived from volcanic rocks. In many lowland areas the till is a hybrid of two major source terranes: the fine grained matrix derives from local Carboniferous redbeds, whereas the large included blocks are crystalline erratics from the highlands and uplands (Fig. 42).

The relatively great thicknesses of till where it forms blankets compared to its thickness in adjacent drift areas stems mainly from either the voluminous accumulation of a single till sheet because of local factors such as bedrock erodibility or topography (for example, the infilling of buried valleys; Fig. 11) or the superposition of several dissimilar drift sheets, commonly interbedded with nonglacial sediments, including organic beds. Multiple till sequences are a major portion of the drift cover of Cape Breton Island.

*Till blanket composed of one drift sheet* – Areas of thick till consisting of only one drift sheet are relatively few and, in most, the till commonly has gradational vertical variations of colour, pebble/cobble content, lithology, or fineness of matrix which presumably reflect changing regimen and provenance, given the fact that ice-flow indicators demonstrate that all areas have been overrun from at least two different directions. Deep and extensive exposures

***Figure 41.*** *Anomalously smooth concave till slope, possibly graded by periglacial solifluction; George Bay coast near Creignish. GSC 204024-S*

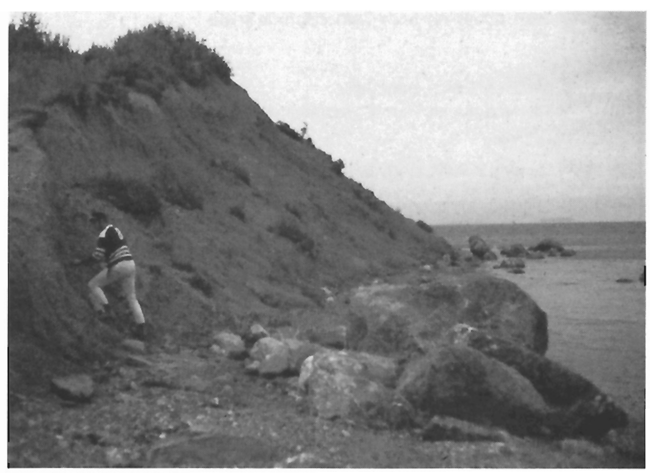


***Figure 42.*** *Sampling typical shoreline exposure of drumlin composed of red clay till with large upland erratics; MacLeod Point, West Bay, Bras d'Or Lake. GSC 204024-M*

indicate thick blankets of a single till sheet in the Mabou/ Judique area, along St. Patricks Channel, and around the West Bay coast of Bras d'Or Lake. The till blankets shown on Boularderie Island, the Margaree Lowlands, and on the highlands around Baddeck Lakes are mapped more on the basis of morphology than outcrop.

*Till blanket consisting of superposed units* – A major portion of the till-covered areas on Cape Breton Island are till blankets consisting of up to four superposed drift sheets, commonly including interbedded stratified sediments, notably organic beds. Geomorphic expression ranges from flat to ridged to drumlinoid. There is considerable complexity in the more than fifty most important exposures of stacked drift sheets which are documented in the section on Stratigraphy, but there is sufficient commonality to derive some general relationships. Composite till blankets consisting of stacked drifts occur throughout the island, mainly on the lowlands. Exposures of definitive sequences are documented in the Stratigraphy section; only the main areas are outlined below. Apart from small areas bordering the northern highlands, till blankets are found mainly on the southern lowlands. The two main areas are a belt of peculiar drumlinoid forms with buried organic beds on the north side of East Bay (Bras d'Or Lake) and a broad swath along the south coast with three morphological expressions – ridges in the Framboise area, drumlins in the Salmon River area, and fluted ground moraine in the Mira Lake area.

The East Bay composite till blanket, as documented by exposures [20]-[27] in the section on Stratigraphy, occurs as northeastward-trending drumlinoid hills and islands (Fig. 43). The streamlined hills are veneered by a thin reddish-brown till which mimics the landform, but are mainly composed of a complex sequence of dissimilar tills and nonglacial beds, including organic layers. The contacts of the lower beds, which are more or less horizontal, are truncated by the surface till, showing that these drumlinoid forms are carved from pre-existing sediments, rather than built up by basal accretion of a single till. This process may apply to other drumlins in Cape Breton Island and elsewhere, particularly those which consist mainly of earlier-deposited till emplaced from a different direction than the trend of their long axis.

The south coast composite till blanket is the largest area of stacked drifts and extends from Strait of Canso to Gabarus Bay in the form of steep-sided, flat-topped, east-trending, slightly curved ridges up to 50 m high (Fig. 44, 45) and >30 km long. The ridges have been variously reformed by subsequent ice movements which have produced complex and interpenetrative patterns of glacial landforms, but there is nonetheless a fairly clear relationship to ice-flow patterns


*Map units:* Rc. bedrock with till patches; 2a. till blanket; 5b. modern alluvium; 6. coastal beach; 7. colluvium (talus); 8. bog, fen —○— drumlin

***Figure 43.*** *Stereogram of drumlins carved from thick, preglacial, mainly waterlain sediments; Eskasoni, East Bay, Bras d'Or Lake. NAPL A21535-125, 126*

deduced from striations. The ridges have small drumlins, and flutings are superimposed and they tail off into drumlin fields at either end. As first noted by Grant (1963, Table 1), the ridges consist of two, and locally three, distinct but dissimilar tills, as typified by numerous exposures in the Strait of Canso to Fourchu area [35, 36, 39, 42, 43, 44]. The lower till, which makes up the bulk of the ridges, is compact, reddish to reddish brown with a sandy/silt texture, and up to 30 m thick. The composition of its pebbles and cobbles and the colour of its matrix show that it is mainly derived from Carboniferous terranes to the west. Pebble fabrics (McClenaghan and DiLabio, 1992; McClenaghan et al., 1992) show that the red till was indeed emplaced by east-flowing ice. The red till overlies an early-formed set of east-trending striations (070°-090°) which have the same general direction as the ridge orientation, and it overlies organic beds in two places [32, 48]. It contains reworked marine shells scattered over a wide area [43, 44, 45, 46, 48]. The shells and the red matrix were presumably derived, at least in part, from the seafloor of Chedabucto Trough which is underlain by Triassic redbeds

*Figure 45. Stereogram of three superimposed generations of till lineaments; Loch Lomond area. Early east-west ridges are re-formed into northeast drumlins and both are superficially ornamented with south-southeast fluting; note thick drift in valley, little on East Bay Hills. NAPL A21536-9, 10, 11.*

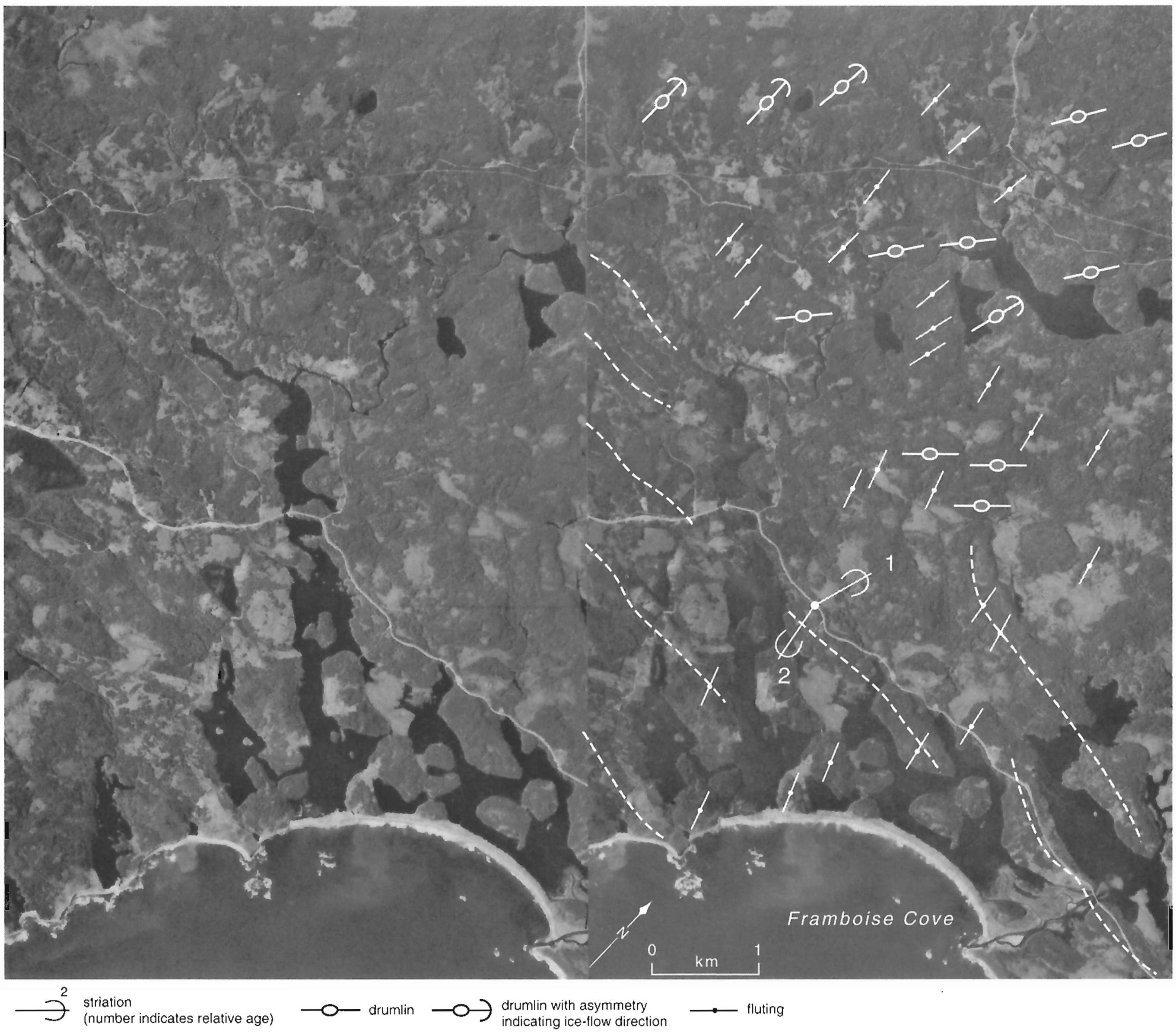
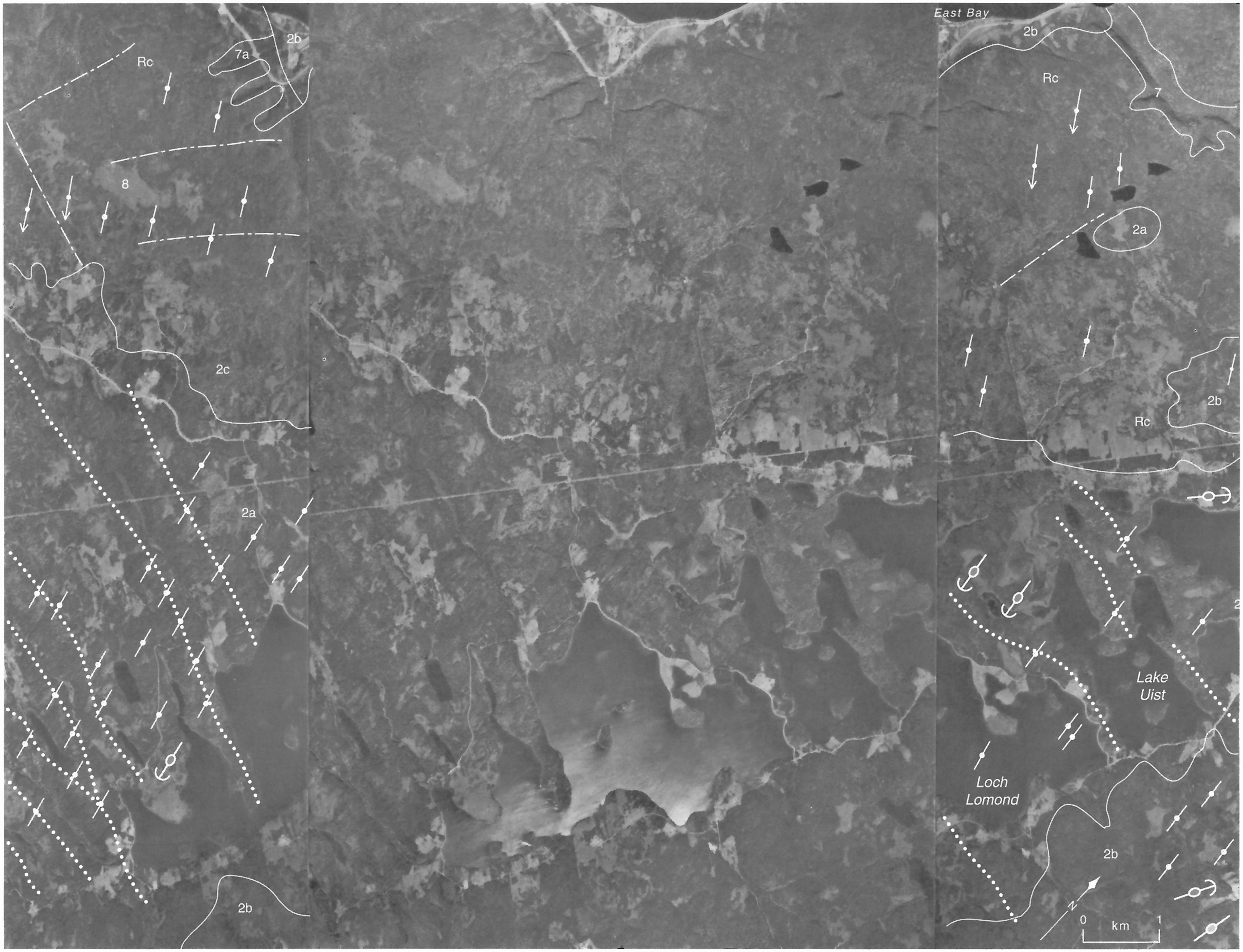



*Figure 44. Stereogram showing older east-west till ridges partly transformed into isolated north- and northeast-trending drumlins; both with light south-southeastward fluting superimposed; Framboise Bay area. (Striation site shown in Fig. 49b has the same ice flow sequence). NAPL A21536-57, 58*




*Map units:* Rc. bedrock with till patches; 2a. till blanket; 2b. till veneer; 2c. discontinuous tilll veneer; 7. colluvium; 8. bog

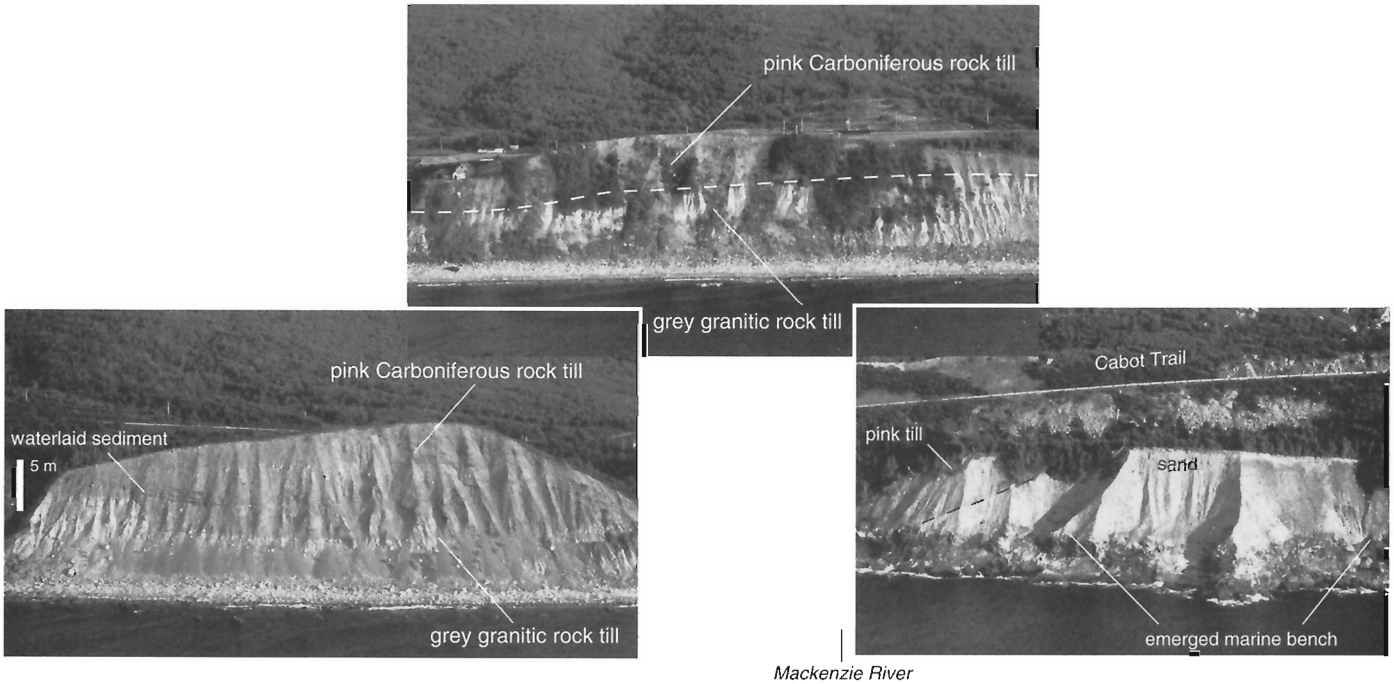



***Figure 46.*** *Aerial view of three segments of kame moraine composed of dissimilar tills and gravel deposited at the confluence of two valley glaciers in Mackenzie and Grand Anse river valleys. GSC 204042-J*

(although no Triassic or other offshore indicator stones were noted). In only three places, one at Capelin Cove (Grant, 1963), and two near English Cove, the red till overlies a grey silty locally-derived till, also with an east-pointing pebble fabric (McClenaghan and DiLabio, 1992; McClenaghan et al., 1992). The red till is almost completely overlain, except for sporadic windows, by a surface veneer 1-5 m thick of grey, loose, sandy and stony till. The grey till is locally derived from the underlying grey, silicic, Fourchu Group volcaniclastic rocks. On top of the red till ridges, the grey surface till forms small drumlins and flutings and in intervening swales, small drumlinoids. These small drumlins and flutings trend southeastward in the western part of the area of large ridges, parallel to the last-formed set of striations (160°), whereas they trend northeastward in the eastern part. There, they have a northeastward-directed fabric (McClenaghan and DiLabio, 1992; McClenaghan et al., 1992).

The eastern and western extremities of the ridge tracts are transitional to longitudinal streamlined till terrains (Fig. 44, 45); drumlins in the Strait of Canso area trend northwestward and drumlins in the Salmon River area and fluted till terrain in Mira River area trend northeastward. Both of the streamlined tracts are parallel to the youngest set of striations in the area. The Mira River fluted till blanket is remarkable in that, as exposed along the highway to Gabarus, adjacent flutes, attenuated drumlins, and the intervening ground moraine are composed of different tills. Most commonly, ribbons of shale till alternate with sandstone till, but one narrow drumlinoid is composed of pinkish granitic porphyry drift which trails northeastward from its source in Gillis Mountain and continues as an erratics train in the fluted till plain west of Sydney. Some of the till ribbons are derived from nearby sources, but a large proportion of the accessory pebbles and cobbles are crystalline rocks from distant East Bay Hills, as Goldthwait (1924) first noted. McClenaghan et al. (1992) document a corresponding strong northeastward transport of material in these drumlins. Bordering the thick drumlinized till, isolated small ovoid drumlins composed of reddish Carboniferous till transported from the south and west are surrounded by till veneer composed of grey, sandy debris derived from the underlying rocks. Near Leitches Creek [32] a 60 m borehole penetrated two till sheets and two peat beds, demonstrating a considerable infilling of the bedrock relief. In terms of glacial history,

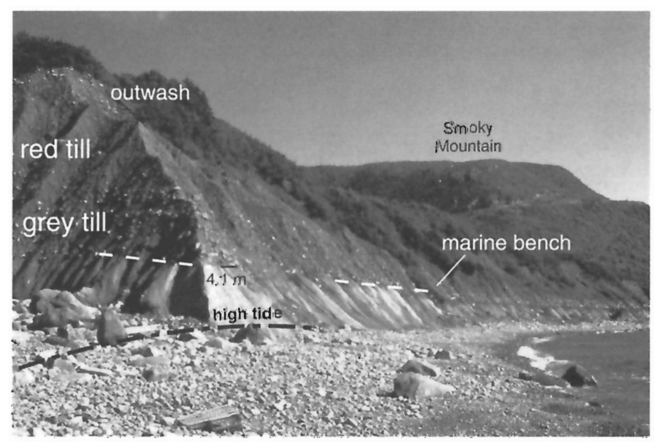



***Figure 47.*** *Emerged marine bench cut across Carboniferous sandstone overlain by two tills and outwash; Morrison Brook. GSC 203752-F*

the till blanket of the Sydney area seems to be mainly the product of glacial transport northeastward and inland from the Atlantic Ocean to Sydney bight.

In addition to the East Bay and south coast composite till blankets, five other areas of composite till blankets may be mentioned. In the River Denys Lowland, Fletcher (1881) annotated his bedrock map in a few places to the effect that wood and peat underlay the surface till and gravel. During this study one such occurrence was seen at Big Brook quarry [49] where a lower till, emplaced from the southeast, overlies organic beds, and an upper till, emplaced from the northwest, overlies laminated lacustrine(?) sand and silt. Near Portage Creek [18], in the same area, brownish till with crystalline stones emplaced from the west is overlain by a reddish silty till composed of redbed materials emplaced from the south. Mineral exploration trenches near the south end of Lake Ainslie at Mullach Brook [16] uncovered three lithologically distinct tills with different fabric directions (Tang, 1970). On the lowland bordering Gulf of St. Lawrence, numerous coastal exposures in the Chéticamp area [60-70] reveal a complex of sands and gravels interbedded with different tills, a lower one derived from offshore, an upper from onshore. Farther north along Pleasant Bay [73], mounds of kame gravel and a grey granitic till are overlain by red Carboniferous till (from offshore?) which presumably were interlayered by confluent valley glaciers (Fig. 46). On the Cabot Strait coast, the Aspy Bay-Dingwall shore shows superposed tills [3, 4]


*Map units:* Rc. bedrock with till patches; 2a. till blanket; 2b. till veneer; 2c. discontinuous till veneer; 6. coastal beach; 8. bog, fen

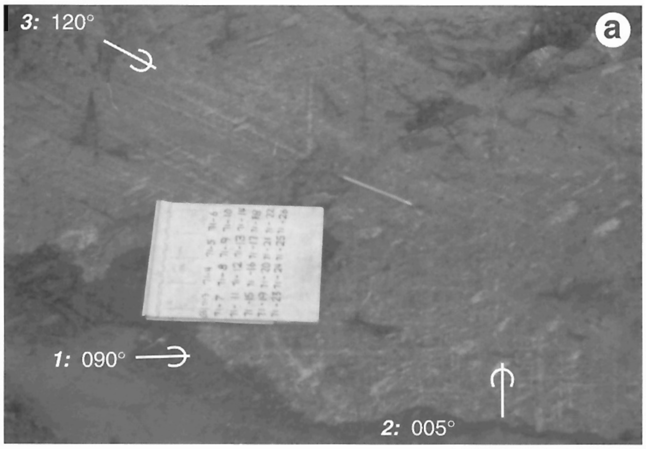
fracture lineament; striation; drumlin; minor moraine

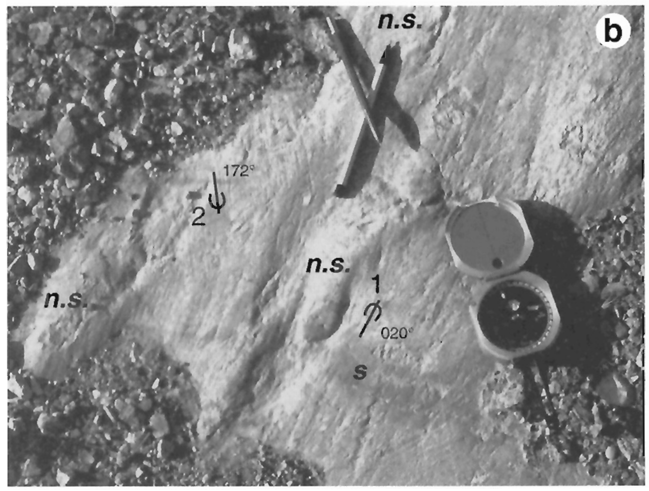
***Figure 48.*** *Stereogram of area north of Gabarus Bay showing drumlins composed of two and three till sheets, separated by till veneer area with minor moraines; adjacent to rocky area with typical structural*

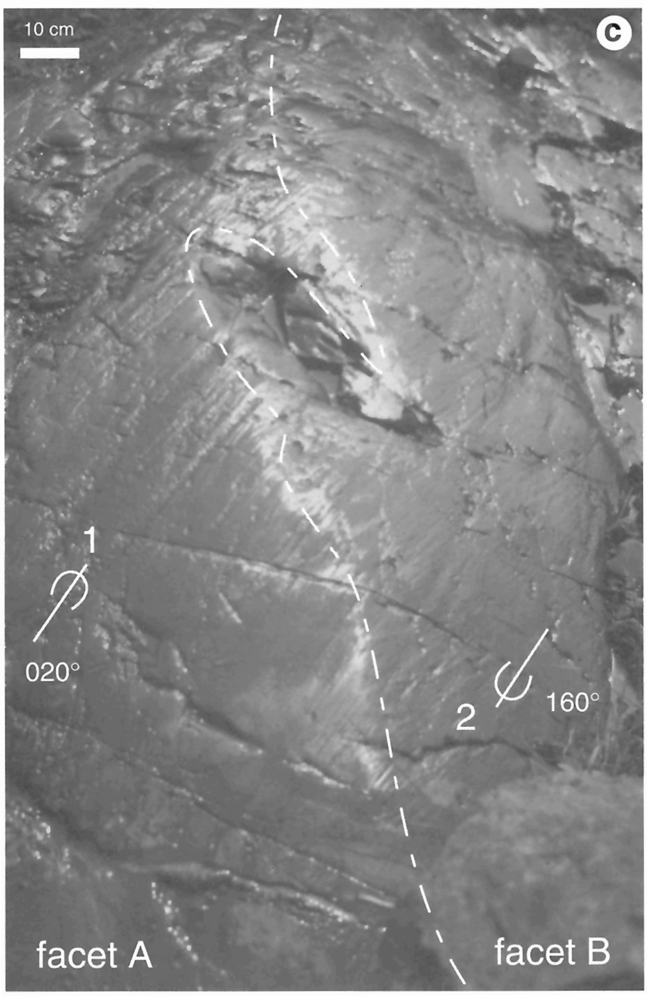
lodged over karstic gypsum, which may suggest that the extensive gypsum karst of the rest of Aspy valley may also have thick composite till sequences. Farther south along the eastern highlands piedmont, the Sangamonian marine bench and its cover of beach gravel is overlain by two tills, the lower derived from the highlands, the upper derived the southern lowlands, as for example at Wreck Cove head [10] and Morrison Brook [8] (Fig. 47). Lastly, on the highlands, two-till sequences were observed in hydroelectric excavations for the Wreck Cove Reservoir (R.R. Stea, pers. comm., 1985).

*Continuous till veneer (sandy; unit 2b)*

Till veneer is a 2-4 m thick mantle that is sufficient to form diagnostic morainic features, yet thin enough to reveal most of the structural relief of the underlying bedrock relief. It is




*a.* *Two intersecting facets on volcanic Fourchu agglomerate with three sets of striations and miniature crag-and-tail; Red Point, Atlantic coast. GSC 202188-X*

*b.* *Glacial pavement on Fourchu Group agglomerate shows miniature crag-and-tail crossed by later striations; ice-free period of weathering between the two sets is inferred from iron stain (s) on early set which has been superficially scraped away (n.s.) by second advance; highway near Framboise. GSC 202188-Y*

*c.* *Outcrop showing two intersecting stoss faces with striations from two almost diametrically opposed directions; MacDonald Mountain, south of Dundee. GSC 1994-411*

*d.* *Two intersecting stossed facets: the first has pedogenic iron stain, on the second it has been glacially removed; Creignish Hills. GSC 1994-412*

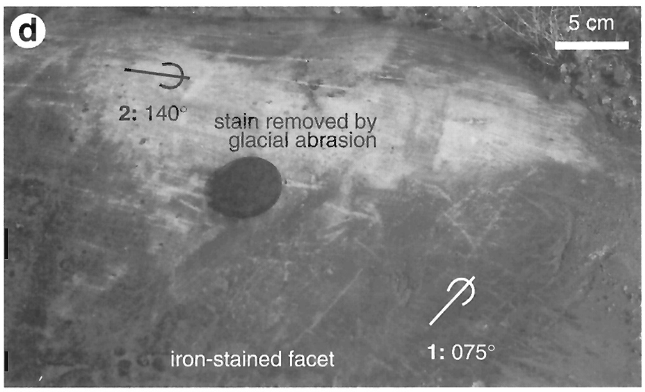
*e.* *Three intersecting striated surfaces showing stoss-and-lee relationship, Strait of Canso shore south of Mulgrave. GSC 204042*

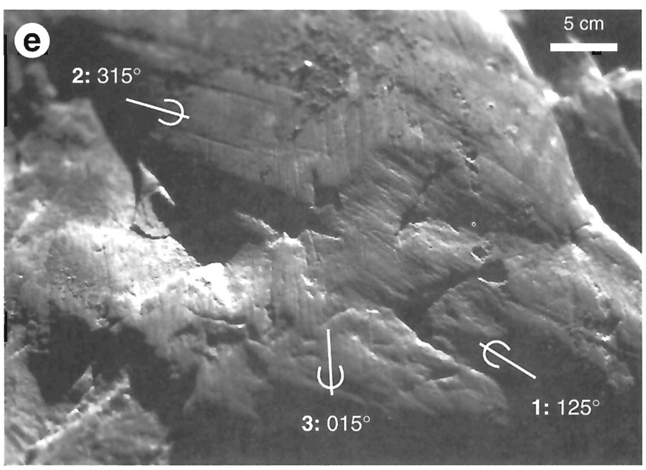
*f.* *Outcrop showing three intersecting stoss-and-lee faces; on highway near Point Michaud. GSC 203797-F*

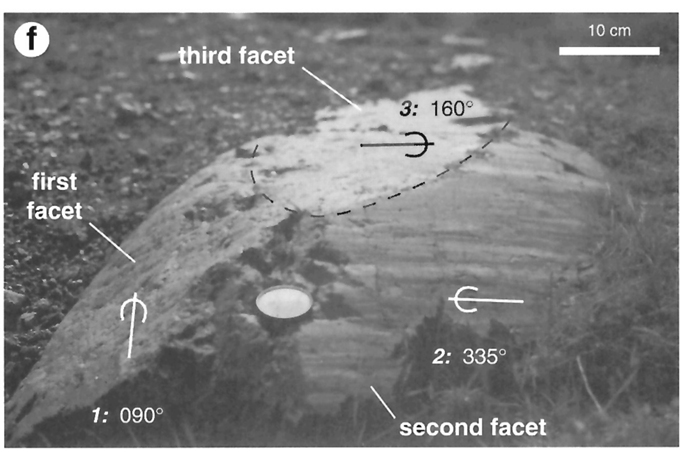
***Figure 49.*** *Crosscutting striation sets and intersecting glacially abraded facets.*

the subdued, but nonetheless visible bedrock relief which allows till veneer to be recognized and delineated on airphotos. The bedrock structural ridges and depressional lineaments mimicked by the till surface are more throughgoing and thus cut across generally shorter and smaller glacial constructional landforms, such as moraines, flutings, drumlins, and crag-and-tail hills (Fig. 48).

Till veneer covers about 15 per cent of the island and occurs mainly on resistant clastic and crystalline bedrock not otherwise covered by thicker, finer grained, foreign tills (Fig. 40a, b). Till veneer is characteristically sandy, is derived from the underlying bedrock, and is typically developed as




the transitional phase between areas of rock and areas of till blanket. In the southern area, till veneer occupies the swales between the drumlins and large till ridges and has sharp contacts with them.

*Discontinuous till veneer (stony; unit 2c)*

This category of thin till denotes areas where rocky areas and numerous smaller outcrops are interspersed in a shallow mantle of till. It is essentially an intermediate phase between continuous till veneer and rocky terrain with some till, with which it forms a mosaic. The term connotes that the proportions of rock outcrop and till cover are about equal and/or not clear enough to enable the terrain to be classified as either dominantly rocky or dominantly till-covered. Till is less than 2 m thick, typically sandy and stony, and is locally derived and immature, meaning that the included stones are mainly angular and show little wear and transport. Such terrain makes up about 7 per cent of the island, chiefly hilltops (Fig. 40a).

### Ice-flow indicators

A sequence of different glacier movement directions is recorded by erosional features and by a succession of till units derived from different source areas. The erosional evidence is judged to be more definitive than the depositional record because the features are widespread, commonly present in multiple sets, and are more completely preserved. Erosional features range from small-scale crosscutting markings inscribed in bedrock to large-size longitudinal till forms that are overlaid on one another. The depositional record of the succession of glacial events comprises stacks of lithologically distinctive till beds, but that record is incomplete because of: 1) partial deposition, 2) erosion by later movements, and 3) a given flow event has produced tills of locally varying lithology which are not readily correlated. Supplementary evidence of ice-flow trends is provided by transverse morainal landforms, by retreatal ice-marginal features, and by erratics dispersal.

*Erosional markings inscribed on bedrock*

Striations and grooves are ubiquitous on all rock types and two or more superimposed sets with different directions are not uncommon on a single outcrop. Of the various directional markings, the most abundant are miniature crag-and-tail forms (also called pressure shadows or rat-tails). This is because a variety of rocks, such as porphyritic gneisses, crystal schists, and conglomeratic sandstones, contain isolated resistant fragments and crystals which can be set in relief by differential erosion. Miniature crag-and-tail is especially well developed on the pyroclastics of the Fourchu Group. Various other features, namely crescentic gouges, chattermarks, and other arcuate fracture features are locally abundant on dense silicic rock types, such as quartzite and acid tuff, and some carbonate rocks such as marble.

The directional sequence is interpreted from a variety of rock-inscribed markings (e.g., striations, miniature crag-and-tail, stoss-and-lee) which commonly occur as differently oriented sets superposed on a single surface or as separate sets on mutually intersecting abraded facets (Fig. 49a-f). One example was noted of glaciotectonic drag of bedrock in a col inland of Bay St. Lawrence (Fig. 50) which shows southeastward flow onshore from Laurentian Channel. Most of the mapped striations are shown on the inset on Map 1631A. Outstanding examples of multiple superimposed striation sets and of intersecting stossed facets occur at Green Point, Creignish shore, Inhabitants Bay, southern Isle Madame, and at many other places along the south coast. Examples of three intersecting facets and striation sets occur at Mulgrave (Fig. 49e) and especially on the siliceous Fourchu Group rocks, as at Red Point (Fig. 49a), near Framboise Cove (Fig. 49b), near Point Michaud (Fig. 49f), at Gabarus [36], Gabarus Cove [37], Fourchu [38], and Scatari Island. The relative age of the facets was deduced from their stoss-and-lee relationships, even where they have almost exactly the same trend but opposite sense, as on Scatari Island and along the Dundee-Grande Anse road (Fig. 49c). A further indication of relative age of such crosscutting striations and multi-faceted outcrops is that one or more of the earlier bevels and striation sets is iron stained, whereas the younger one(s) is fresh and clean (Fig. 49b, c). Moreover, where such outcrops are till-covered, the stained facet is overlain by till which also contains secondary iron and manganese, whereas the fresh facet is overlain by a till apparently devoid of secondary mineral accumulation. The iron/manganese stain appears identical to that which accumulates subaerially on rock outcrops and/or which accumulates on rock surfaces that subcrop in the Bfe soil horizon (although this has yet to be demonstrated by chemical analysis and SEM). By this hypothesis, the iron/manganese stain is presumed to represent subaerial weathering during a former nonglacial period. After deposition of the stain, subsequent glacial abrasion removed the stain from part of the surface before another till was deposited.

The more than 600 measurements of such ice-flow indicators can be grouped in several major patterns, whose relative age is reasonably well documented by the widespread crosscutting relationships. All of the flow phases appear to be Wisconsinan because the striations and associated tills are superimposed on the raised marine rock bench (e.g., Fig. 47)

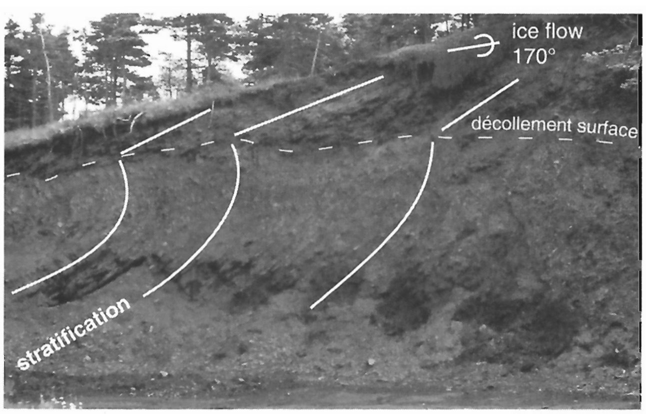



*Figure 50. Shale bedrock with upper part of foliae overturned almost 90° by drag of glacier that advanced uphill from Cabot Strait; borrow pit, Bay Road valley. GSC 202188-Z*

as well as on the overlying organic beds which date to the last interglaciation. If this age assignment is correct, the supposed subaerially-produced iron stain which coats some early striations but not later crosscutting ones, represents a period of glacier withdrawal in the middle of the sequence. This sequence of crosscutting striations and intersecting glacial facets may be referred to as an "erosional stratigraphy" (The term was coined by the author in the early days of this study and has since been adopted by other authors (e.g., Stea et al., 1987, 1992; Stea and Mott, 1990)), although it may be a contradiction in terms. This erosional sequence is corroborated by the less complete depositional succession and thus is used as the framework for the reconstructed glacial sequence.

The above-described sequence of glacier-erosional markings on bedrock serve to outline an ice-flow sequence to which depositional evidence can be compared. The flow sequence is essentially similar to that summarized on Map 1631A (inset), except that some of the patterns have been subdivided such that eight discrete flow phases are recognized in this report. They involve phases of both local ice dispersal and invasion by external ice sheets. A sequence of flow phases (designated by letters rather than number as some may be coeval) with their range of azimuths and inferred sources is summarized as follows: a) radial expansion of a highlands plateau ice cap; b) eastward (070°-110°) from mainland Nova Scotia (and/or beyond) across the southern lowlands; c) southward lobation (160°-175°) from Cape Breton Shelf (Laurentian Channel?) onto Aspy and Sydney lowlands and eventually southeastward (±135°) across the highlands plateau; (A period of ice retreat then possibly occurred, as inferred from an iron stain on these earlier two striation sets as if by subaerial weathering, prior to being crosscut by the following younger striations which are unstained); d) northward (335°-035°) over the southern lowlands from a Scotian Shelf ice cap; e) radial flow of two island-centred ice masses, one in the Bras d'Or basin and a later one on the northern plateau; f) eastward (045°-090°) and northeastward onto the George Bay and northwest coasts by foreign ice; g) radial retreat of Bras d'Or Lowlands ice, with a readvance 11-10 ka; h) radial retreat of a highlands plateau ice cap (and possibly Neoglacial re-glacierization). (As will be further discussed under Cenozoic history, Wisconsinan (oxygen isotope stages 4, 3, 2), this sequence has several points in common with the sequence developed in adjacent mainland Nova Scotia (e.g., Stea and Brown, 1989; Stea and Mott, 1990). These trends, based on erosional markings, have associated with them suites of longitudinal and transverse drift forms and dispersal trends which corroborate their direction and sequence, as described below.

### *Longitudinal, ice-moulded till forms*

The surface of till-covered areas in Cape Breton Island is commonly ornamented with large and small streamlined till forms, mainly crag-and-tail hills, flutes and grooves, drumlinoids, and drumlins (Map 1631A). It may be noted that the Cape Breton Island drumlins are the most complex of the several Nova Scotia drumlin fields which are cited as one of the four largest in the southern zone of the North American ice sheet (Flint, 1971, p. 100). There are five major areas of longitudinally moulded till terrain in Cape Breton Island, four on the lowlands surrounding Bras d'Or Lake and one on the highlands, each with differently oriented sets of features. The largest covers a wide area around Mira Lake and features northeast-trending drumlins and fluting (Fig. 44, 45, 48). A second area, characterized by northwestward-trending drumlins and crag-and-tail hills, extends from the Mira field west to Strait of Canso (Fig. 26) and north to Mabou. The drumlins are aligned at right angles to the Mira trend and merge with it on the west. A third set, comprising much smaller drumlins and associated flutes and grooves with a south-southeast trend, overlies, crosscuts, and locally reforms the Canso drumlins and those in the southern part of the Mira field (Fig. 44). Abutting and at right angles to the south-southeast trend, are southwest-trending drumlins and moulded till terrain in the nearby western bays of Bras d'Or Lake and adjacent lowlands. The fifth area has southeastward-oriented crag-and-tail hills and small drumlins on the highlands plateau between Baddeck and Wreck Cove Reservoir. Except for the West Bay group, the orientation of streamlined till forms in each of these areas is parallel to a set of glacial markings inscribed on underlying and interspersed bedrock outcrops.

*Transverse morainal features*

Various linear glacial deposits are oriented generally at right angles to the mapped direction of ice-flow expressed by striations and longitudinal streamlined till forms. These transverse features include ice-marginal deposits and certain subglacial bedforms. Those composed of stratified drift are the largest, but they are classified as glaciofluvial deposits and are discussed elsewhere; those composed of till are discussed below. They include large and small end moraines and small till ridges called minor moraines and ribbed moraine. Also included here are certain irregularly shaped and anomalously large till ridges of uncertain origin which are discordant in overall trend and pattern to the mapped ice-flow trends.

*End moraines* – End moraines in Cape Breton Island are relatively small, being generally less than a few kilometres long and less than a few tens of metres high. Most are composed of till, though locally contain stratified material. (Glacier-marginal deposits composed largely of stratified drift and termed kame moraines, though having the same paleoenvironmental meaning as end moraines, are treated later in the section on Glaciofluvial deposits and meltwater flow-direction indicators). One group of end moraines relates to the series of ice flows recorded on the southern lowland half of the island. The largest of this group is a mass of drift 30-50 m thick covering ≈4 $km^2$ which almost blocks the outlet of Bras d'Or Lake at Great Bras d'Or. A belt of small moraines between Sydney River and Mira consist of a series of approximately 10 subparallel ridges in short, discontinuous segments; they mark the distal edge of the Mira drumlin field. Another group of small end-moraine ridges, with associated ice-contact stratified drift and northward-directed meltwater channels, lies on the north slope of Creignish Hills east of Judique (where perhaps significantly there is evidence of a Late Wisconsinan readvance (R.R. Stea, pers. comm., 1992)). These small moraines probably relate to the southeastward onshore striation set (Phase F), but may be associated with the earlier northwestward set (Phase D). Numerous short, small, parallel end moraines, developed in thick till on the northern plateau behind Ingonish, are perpendicular to the final eastward flow of a plateau glacier (Phase H). In the col between Middle River and Margaree River intermontane valley at lakes O'Law, a small end moraine with a south-facing proximal slope and northward-directed outwash (Fig. 51) probably relates to a northward-flowing valley glacier fed either by highland or lowland ice.

A second group of end moraines, also relatively small, occur in the lower reaches of highland valleys. Examples on the eastern highland margin are those at Aspy Fault [3] (Fig. 52), Rocky Bay, Indian Brook, Dundas Brook, and Baddeck River. Together with the kame moraines (see below under Glaciofluvial deposits and meltwater flow-direction indicators), these small end moraines were evidently built by ice lobes deploying from a late stage ice cap on the highlands plateau.

***Figure 51.*** *Small hummocky end moraine north of lakes O'Law, deposited by north-flowing outlet glacier from highlands ice cap. GSC 204042-C*

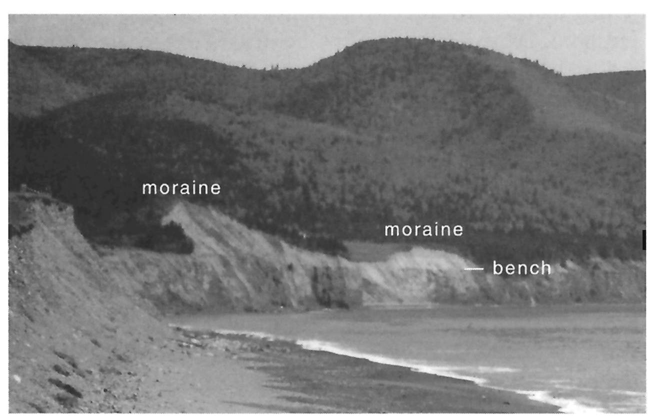



***Figure 52.*** *Two small end moraines deposited on emerged interglacial rock bench by late glacier in Aspy valley; view from Cabot Landing, Aspy Bay. GSC 203752-G*

*Transverse till ridges mainly formed subglacially* – Transverse till ridges, other than end moraines, include large areas of ribbed moraine and lesser areas of nondescript, anastomosing, hummocky ridges of variable width (see Map 1631A). Some of the latter may be frontal moraines, except that they lack the diagnostic ice-contact slope. In ribbed moraine, the individual ridges are a few metres high, a few tens of metres wide, and a few hundred metres long. They are sinuous and typically have stoss-and-lee asymmetry. The tops are generally flat and commonly have superimposed grooving. The grooving or fluting is at right angles to the ridge and both have the same trend as the last-formed set of striations in a given area. Areas of ribbed moraine occur: 1) west of Mira Bay, where they are associated with the end moraines at Sydney River; 2) north of Gabarus Bay (perpendicular to the northeast-trending Mira drumlin field) (Fig. 48); 3) on Isle Madame and near St. Peters (perpendicular to the secondary south-southeastward fluting) (Fig. 26); 4) on the north slope of Creignish Hills in association with small end moraines; and 5) on the highland plateau behind Ingonish where they are related to eastward retreatal ice flow.

*Giant till ridges* – The largest and most extensive linear till deposits on the island is a series of about 50 east-trending ridges which extend 100 km eastward over the Atlantic coastal part of the Bras d'Or Lowland from Strait of Canso to Fourchu Bay and 40 km south from Lake Uist to the Heath Head-Red Head [44] area of Isle Madame (Fig. 26, 44, 45). They also appear to extend a further 20 km to the Guysborough area of mainland Nova Scotia. As Weeks (1954) first noted, the individual ridges account for most of the local relief in this area. They range in height from 20-50 m, in width from 300-1000 m, and in length from a few kilometres to 15 km; collinear segments can be traced up to 30 km. They are generally straight, but those in the Framboise area are broadly concave to the north. Their orientation swings from due east on Isle Madame (Fig. 26) to east-southeast across Lake Uist (Fig. 45). The ridges are mostly composed of a distinctive till with a fine grained reddish-brown matrix that is derived from Carboniferous redbeds, probably those outcropping in the Bras d'Or Lowland, if not also in Chedabucto Trough and elsewhere. The red till in the Fourchu-Gabarus area have strong a eastward-directed pebble fabric (McClenaghan and DiLabio, 1992). In the English Cove area of Gabarus Bay the red till is underlain in two places by a grey silty till, also with an eastward-pointing fabric (McClenaghan and DiLabio, 1992). The ridges are, however, superficially grooved and remolded by later ice flows (Fig. 26, 44, 45) and are generally veneered with younger drifts.

The orientation of the ridges relative to the various mapped ice flows is ambiguous and their time of formation uncertain because they overlie bedrock with three different ice-flow directions inscribed (eastward, northward, southward; phases B, D, E). The simplest explanation would be that the ridges were formed during the youngest flow events in that area (phases E or G; 160°-170°). This would make them transverse ridges which were formed either at the ice margin as end moraines or subglacially as rogen moraines (as H.L. Cameron first implied in Wilson et al., 1958), however, the ridges are ten times larger than typical rogen moraine, they do not look like ice-frontal features with typical scalloped margins and cross-axial meltwater channels, and have, in any case, been completely overrun by the youngest flows which superficially fluted and remolded them into small drumlinoids. Alternatively, the giant till ridges were formed during one of the earlier flows and have been modified and slightly shifted by subsequent ice movement. If so, they could be either end moraines transverse to the northward (≈030°) flow (Phase D) or longitudinal streamlined forms produced by the eastward (070°-110°) flow (Phase B). The last suggestion is preferred for three reasons: the ridges trend generally parallel to the early eastward ice flow, they are composed largely of the till assigned to that ice flow, and they bear the imprint of the later movements in the form of secondary remolding and superposed tills. Further support for this suggestion is that eastward-directed till fabrics were measured in the red till in the Fourchu-Gabarus area (McClenaghan and DiLabio, 1992) and in the lower part of the till that appears to form similar ridges on the north side of Chedabucto Bay in mainland Nova Scotia (Stea and Mott, 1990, p. 74). Hence, the giant till ridges can be regarded as very large drumlinoid features related to the earliest major advance of ice in southern Cape Breton Island.

*Erratics dispersal and till transport*

Qualitative evidence of glacial transport, complementary to directional ice-flow measurements, includes the dispersal of heavy minerals and individual distinctive stones (indicators) and the wholesale overlap of till of one provenance onto an adjacent source terrane. Despite two main factors which limit the number of examples, namely that certain indicator species occur in more than one source area, and that reworking during successive discordant movements has blurred the original dispersal fans and trains and produced a complex mixture of drifts and indicators from various areas, all of the major flow phases mapped independently by striations have a comparable expression in terms of dispersal phenomena. Examples of various scales of drift transport corresponding to the major successive flows to the east, northwest, northeast, and southeast are described below. Several notable examples were reported by early authors, all except one of which were verified in the present study. Attention is drawn to a puzzling apparent lack of expected transport in certain areas.

*Eastward* – The most extensive drift displacement occurred in the southern lowlands along the outer Atlantic coast where red till was shifted eastward and southward onto crystalline rock areas from Carboniferous redbed terranes located in the Bras d'Or basins, if not also in Chedabucto Trough and the Nova Scotia mainland west of Cape Breton Island. This has been borne out by detailed provenance studies of the red till in the Fourchu-Gabarus area (McClenaghan and DiLabio, 1992; McClenaghan et al., 1992). This displacement indicates that the eastward (Phase B) and possibly also the southward ice-flows (phases C, E, G) in this area were vigorous and/or long sustained.

*Northward* – There are several cases of major northward dispersal which attest to the northward ice flow (Phase D) documented by erosional indicators across the entire western half of the island. The most obvious is the northward spread of marine shell fragments in tills (Fig. 30), beginning in southern Isle Madame [43-48] and extending at least 50 km inland to Big Brook quarry [49], Campbell Brook [50], and the East Bay coast [19-26]. Further evidence of northward transport is the spread of igneous and metamorphic erratic stones >50 km from Marble Mountain and Creignish Hills (notably George River Group quartzite and marble) to Mabou and Inverness on the Gulf of St. Lawrence coast. Another example is the northwestward dispersal of granodiorites from near Whycocomagh to Inverness 20-30 km away (Norman, 1935). Masses of upland crystalline drift trail a few kilometres off the north flank of Creignish Hills onto the adjacent sandstone terrane. Finally, small but conspicuous gypsum trains trend northward inland from Carleton Head (Inhabitants Bay) and along George Bay coast and coal from the Inverness-St. Rose fields has been dragged northeastward along the Margaree coast (Bell, 1943).

*Northeastward* – The northeastward flow in the eastern part of the island (phases D, E), evidenced by striations, is also manifest by voluminous drift transport in three major areas. Distinctive volcanic agglomerates from the crystalline Boisdale Hills are displaced many kilometres northeastward onto the grey sandstone terrane in the Sydney area (Goldthwait, 1924, p. 86) and are part of a wholesale northeastward displacement of red southern till from the redbed terrane surrounding Boisdale Hills. This flow also produced streamers of slate, sandstone, and granite from East Bay Hills in the drumlinized till sheet along the highway south of Marion Bridge, and fans of white Cambrian quartzite from Mira River area. Sangster (1988) found a distinct northeastward dispersal fan in the Blue Mountain area, near Marion Bridge. Northeastward transport during deposition of the surface till in the Gabarus and Yava [94] areas have been clearly documented by geochemical assays (McClenaghan and DiLabio, 1992; McClenaghan et al., 1992; MacDonald and Boner, 1993). That this flow extended offshore and moved northeastward along the eastern highland margin is shown by the upper grey till at Morrison Brook [8] (Fig. 47) and Wreck Cove head [10] (Fig. 53) containing diorite from the eastern highland margin to the south.

*Southeastward* – Evidence of southeastward drift transport is present in several highland and lowland areas where lowland till is displaced onto highland (crystalline) terranes and highland-derived tills are displaced downstream onto adjacent lowlands. This flow comes from the general direction of the Gulf of St. Lawrence and, in the first appraisal of the area, all southeastward transport was thought to be the same age and correlated on Map 1631A (inset). However, in the process of reconstructing the glacial paleogeography, closer inspection of the relationship between southeastward and other flow phases, it has been concluded that this correlation is unlikely. In this report the southeastward flow patterns are considered to belong to two separate phases of the last glaciation.

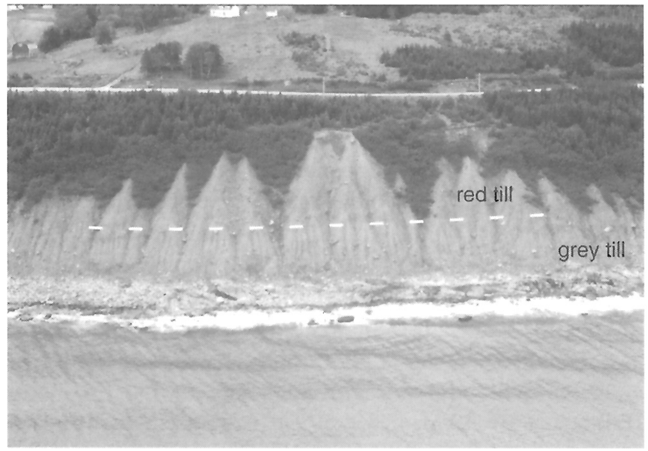



*Figure 53. Aerial view of composite till sequence with lower grey granitic till deposited by eastward ice flow off highlands, and upper reddish till with intermediate rocks deposited by glacier flowing northeastward along the coast; Wreck Cove head [10]. GSC 204042-N*

A strong southeastward ice flow (Phase C) is imprinted on the highlands plateau and there are three examples of southeastward transport. Red till with a matrix composed largely of red lowland Carboniferous material overlies the saprolite at Second Fork Brook Road South [63] (Fig. 38). A train of carbonate boulders, which could only have come from the west coast area, trends eastward for 18 km inland of Ingonish (MacLaren, 1956). Lowland sedimentary rock erratics were reported by Neale (1963) on the plateau east of Bay St. Lawrence (although the present study found none despite good exposures along the Money Point road).

A bright red clay/silt diamicton, evidently derived from the extensive redbeds offshore under Gulf of St. Lawrence has been displaced southeastward onto the west coast from Green Point to Lowland Cove [57-77]. The red foreign material overlies nonglacial materials, such as gravel and organic beds, and grey sandstone and crystalline bedrock. At most places, the red material is a stony, massive, compact till that rests on a stossed surface with southeastward striations [e.g., 57, 60, 70, 76, 77] (Fig. 54). Locally, however, this red foreign till appears to grade laterally into a similarly coloured material which contains few or no coarse stones and is slightly stratified. At other places, the silt appears to overlie the red till [58, 61, 64, 65, 66, 67, 68, 72, 74, 75]. As discussed below under Cenozoic history, Wisconsinan (oxygen isotope stages 4, 3, 2), this material is tentatively related to drainage of proglacial lakes dammed in western intermontane valleys by a late eastward flow of Gulf of St. Lawrence ice (Phase F).

A continuation of the general southeastward movement across the highlands and onto the adjacent eastern lowlands is shown by highland-derived anorthosite erratics and garnet sand in Aspy Lowland tills (Newman, 1971) and by the lower reddish granitic rock till at Morrison Brook [8] (Fig. 47) and Wreck Cove head [10] (Fig. 53) which is derived from a pluton directly inland.

*Southward* – A veneer of locally derived till that overlies the blanket of older, red-brown, distantly derived till is evidence of a late southward ice flow across the southern as in the area between Strait of Canso and Fourchu. It is the latest glacial event in that area and is traceable from the south coast to the shore of Bras d'Or Lake, but not farther north. It therefore appears to represent the movement of a late remnant ice mass over Bras d'Or basin.

notable exceptions to the expected effects of the mapped ice-flow trends which deserve mention. Firstly, the northward flow, which apparently stemmed from a source offshore on the Scotian Shelf, was able to introduce marine shells, but unaccountably did not introduce any of the distinctive Triassic redbeds and other grey Tertiary rocks which underlie its supposed source area. Secondly, the southeastward flow which advanced out of the Gulf of St. Lawrence and crossed the highlands did not deposit on the plateau any of the distinctive bright red drift which it laid down on the west coast (except for a single occurrence at Second Fork Brook Road South [63]). Even more puzzling is the situation on the crystalline rock of St. Paul Island: till is rare, but what was seen consisted only of local crystalline material – there was no sign of any sedimentary rock drift which might have been deposited by the glacier that flowed southeastward down Laurentian Channel and over the island.

*Summary sequence of mapped ice-flow indicators*

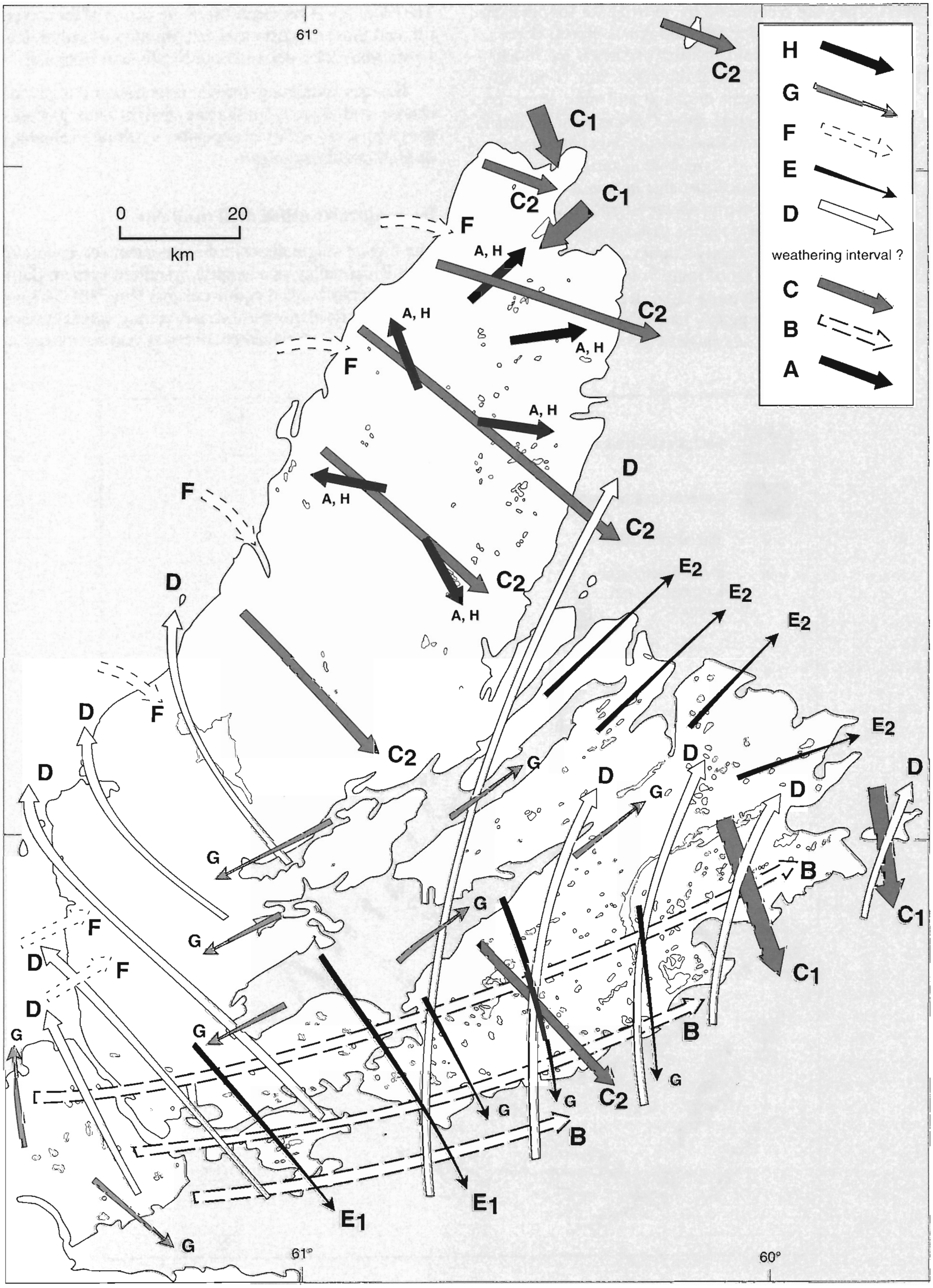
The above data on crossing striation and superimposed till lineaments provides sufficient information to summarize the major mapped ice-flow patterns and their relative ages (Fig. 55) (It resembles the general sequence shown on Map 1631A (inset), except that some of the trends have been subdivided). The patterns (called "Phases") are lettered A-H, rather than numbered, because the sequence may not be complete and because not each is a separate event; some are known or strongly suspected to be coeval. All of the mapped flow patterns and till sequences apparently belong to the Wisconsinan glaciation because they are inscribed upon or overlie the raised interglacial littoral bench and associated interglacial beds. (The two occurrences of till-like material suggesting glaciation during or prior to the last interglaciation are of doubtful origin and stratigraphic context, lie in the subsurface, and do not appear to correlate with any of the main till sheets linked to the mapped ice flow sequence. Evidence of glaciation is therefore restricted to post-last interglacial or Wisconsinan time. The first glacial movement (Phase A) is considered to be the northeastward flow that deposited lower till in Aspy valley. It is tentatively regarded as an outflow of ice from the nearby highlands, which can reasonably be expected to be the site of earliest glacierization. The first main glacial movement recorded over the lowlands (Phase B) was eastward (070°-090°) across much of the southern part of the island. Next a major southeastward flow (110°-160°) moved across much of the island (Phase C), as represented by fluted till plains on the northern highlands and by the flutings on the tops of the till ridges and striations on the southern lowlands; this movement, however, is apparently not recorded in Bras d'Or Lake area. An iron/manganese stain, assumed to be the result of subaerial weathering, has been deposited on these first two sets of striations, but not on sets superimposed on them, so an ice-free period may have intervened before the next glacial phase. Next, there was a northward flow (340°-030°) from Scotian Shelf onto the southern lowlands (Phase D) where it is also manifested as drumlins. Subsequent ice-flow patterns include: a radial dispersal from a centre in Bras d'Or Lake basin (Phase E), a probable penecontemporaneous northeastward (050°) movement onto the George Bay and northwest coasts (Phase F), a final radial flow out of Bras d'Or basin, mainly expressed as a superficial south-southeastward (160°) fluting between Canso and Gabarus (Phase G) during concentric retreat including a local readvance 11-10 ka, and lastly, radial flow during retreat of highlands ice (Phase H). The interpretation of these ice-flow patterns, in terms of discrete glacial events, together with associated tills and landforms, is discussed under Cenozoic history.


***Figure 54.** Pebbly sandstone outcrop with stoss-and-lee asymmetry and east-pointing miniature crag-and-tail overlain by till derived from redbeds offshore; Cap Rouge [70]. GSC 204024-V*

## *Glaciofluvial deposits and meltwater flow-direction indicators*

Glaciofluvial deposits cover about three per cent of the island, mainly in intermontane valleys and some of the larger southern lowland valleys (Fig. 56). Glaciofluvial deposits are sediments transported and laid down by glacial meltwater, either beneath, within, or in front of the glacier as ice-contact stratified drift, or beyond the glacier as proglacial outwash. The ice-contact group of deposits, which makes up about half

***Figure 55.** Summary map of successive glacier flow patterns. Length and position of arrow indicates extent over which ice-flow trend is recorded. Relative age is indicated by overlap and by letter and number at downstream end of arrow; sequence is general, rather than absolute, as some are known or likely to be coeval.*


H
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D
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G
B

the total, includes kames, kame moraines, crevasse fillings, and eskers. They are recognized by an irregular topography with conical hummocks (kames) and small closed depressions (kettles) resulting from melting of buried ice blocks. Mainly the product of subglacial drainage, they commonly have little relation to present drainage and may occur on hilltops, valley sides, or beneath lakes. Outwash, on the other hand, having been deposited subaerially, occurs largely along present watercourses, and is recognized as terraces, marked commonly by shallow braided dry channels and locally by kettles. Outwash terraces have steeper gradients than modern floodplains, and commonly decline downstream from ice-contact deposits, or begin abruptly at relict ice-contact scarps. Glacial alluvium, the product of more voluminous sediment load, greater discharge, and more rapid deposition than prevails in modern streams, typically has a coarse gravelly texture with chaotic bedding and large-scale current structures (Fig. 57). The matrix is commonly poorly sorted and muddy. The lithology is generally the some as that of the surrounding till, and thus, in most cases is a mixture of sedimentary and crystalline rocks, derived both locally and from afar.

Besides indicating former directions of meltwater discharge and thereby providing insight into glacial paleogeography, glaciofluvial deposits constitute a valuable source of high-quality aggregate.

### Ice-contact stratified drift (unit 3a)

The largest single deposit of ice-contact sediment occurs in Mira River valley as a lengthy system of kettled, flat-topped gravel kames banked against eskers (Fig. 58). The complex was evidently deposited on and among stagnant glacier ice blocks after a subglacial channelway had been exposed. The

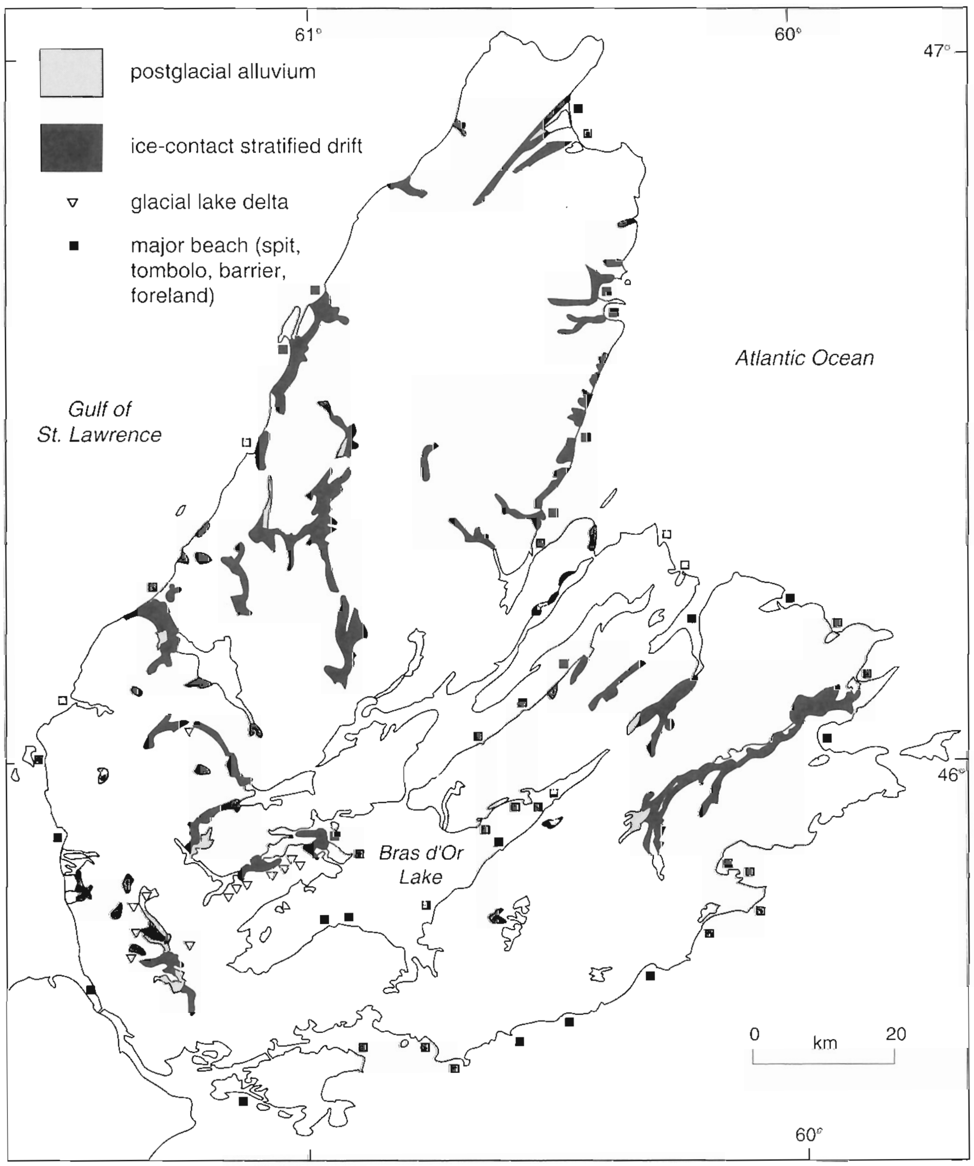



*Figure 56. Map showing distribution of stratified deposits (adapted from Map 1631A).*

*Figure 57. Gravel pit showing northward current-bedded structure in esker in ice-contact complex near Monroe Bridge, River Denys Lowland. GSC 203504-Q.*

seaward end near Cow Bay is a complex of bouldery and sandy eskers and crevasse fillings which grade laterally to minor moraines. Gradients and current structures indicate that deposition took place initially under the ice as subglacial meltwater escaped northward from an ice cap centred over the south coast. The distal portion of crevasse fillings is considered to be a kame moraine that was deposited at a stagnant ice margin which extended westward to Sydney. Later, as the ice front retreated to the southern end of Mira River (actually a lake), proglacial outwash aggraded downstream. Analogous deposits, related to the same northward ice-marginal drainage occur to the west at the mouth of Great Bras d'Or, in the valleys of Sydney River, on the east side of St. Andrews Channel at Frenchvale Brook.

Smaller ice-contact complexes, mainly kame moraines, occur widely in the western part of the southern lowlands. The kame fields, kame moraines, and kame terraces in eastern River Denys Lowland (Fig. 57) were built against the flanks of North Mountain and Creignish Hills by a westward-flowing ice lobe. In adjacent Inhabitants River valley, an

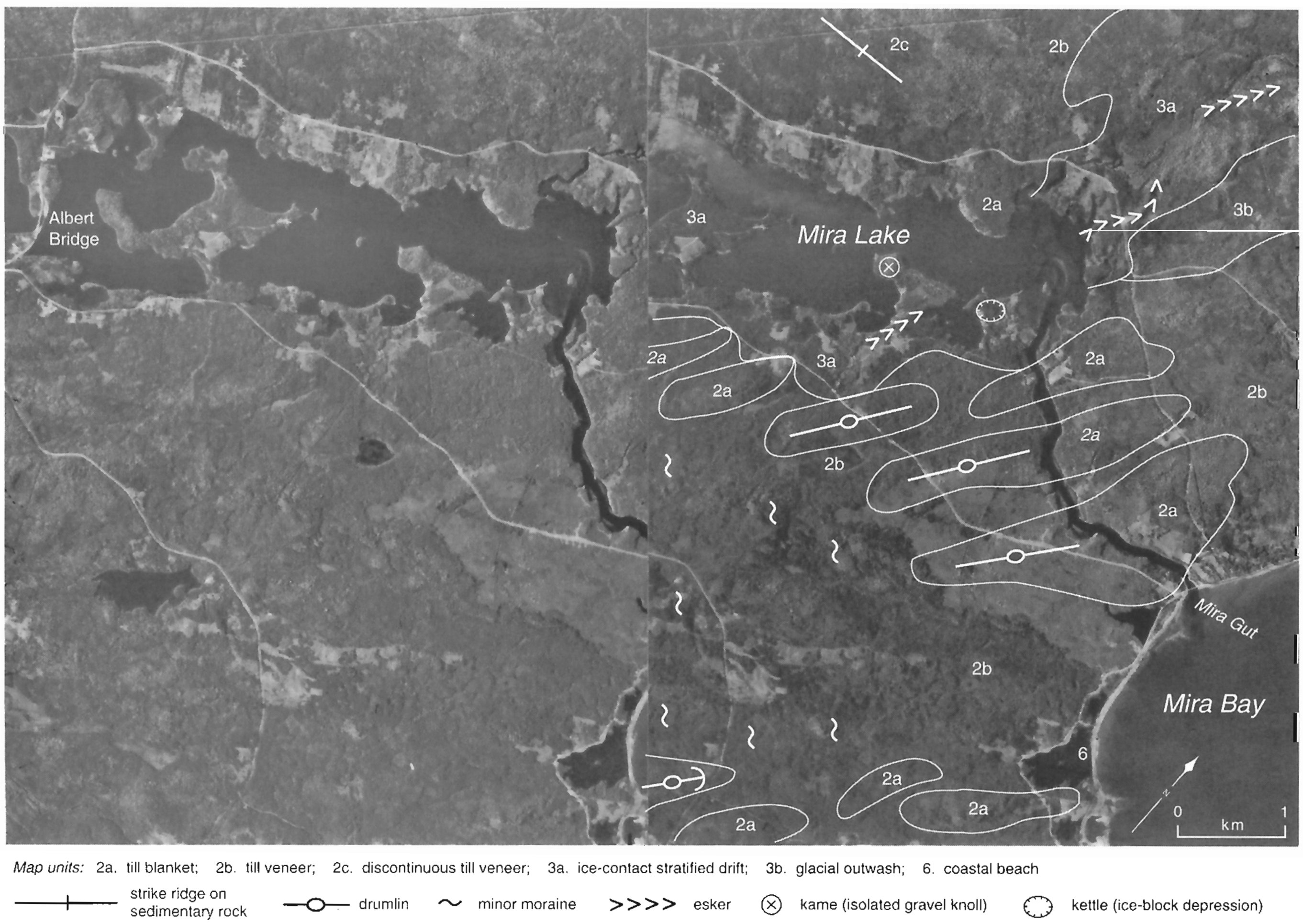
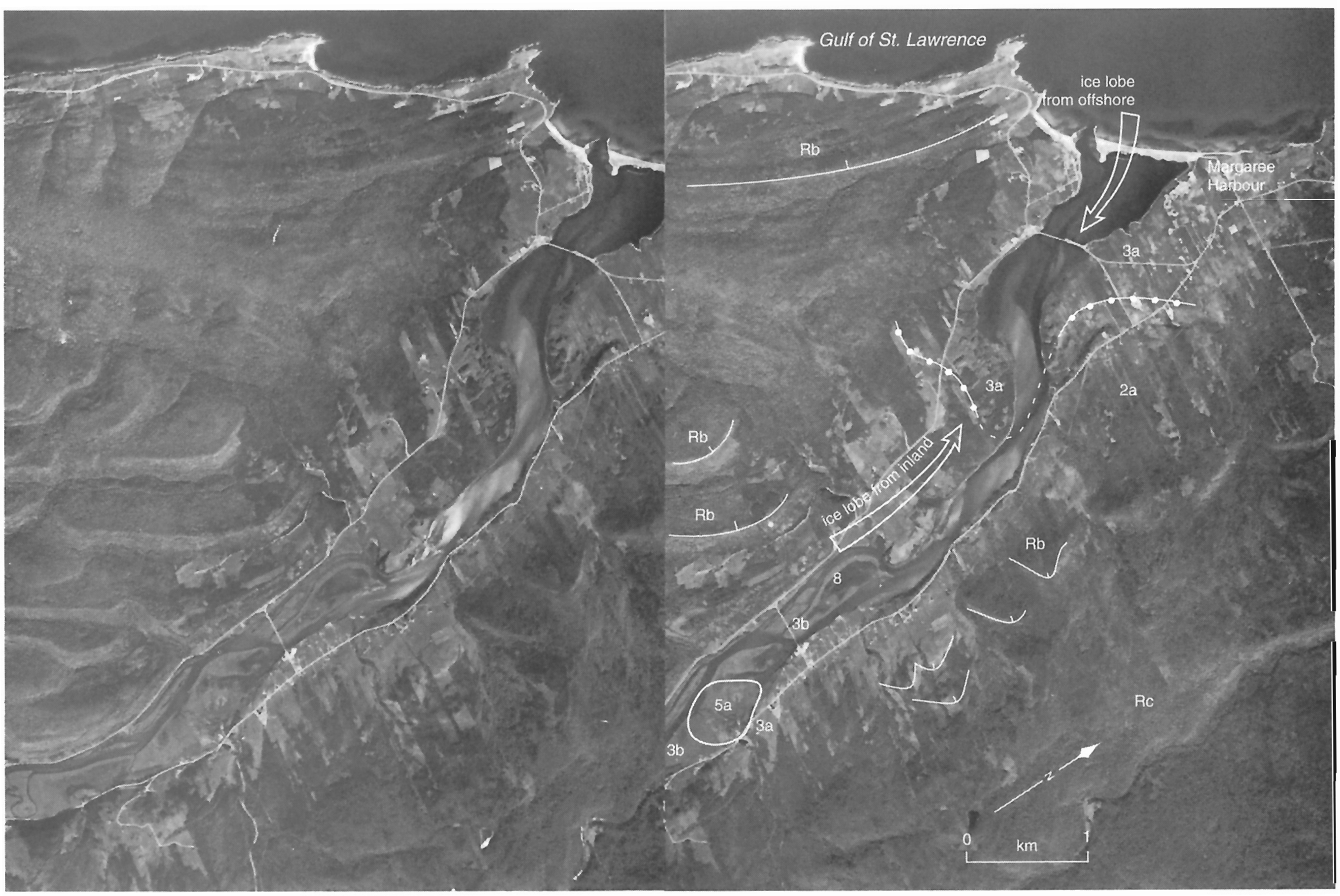



*Figure 58. Stereogram showing kettled ice-contact stratified drift (kame moraine and eskers), and large northeast-trending drumlins blocking inferred former outlet of Mira River (lake). NAPL A21536-84, 85.*

esker with northward-imbricated structures (containing reworked marine shells at Grantville [47]) and a kame moraine across the valley relate to a northward-flowing ice lobe, probably part of the same ice mass that also created a small proglacial lake in the valley. Similarly, kame moraines and sidehill kame terraces on the north flank of Creignish Hills relate to recession of a northward-flowing ice lobe in George Bay. The large kame complex at Inverness and the smaller ones at the mouth of Margaree River (Fig. 59) are probably kame moraines since they apparently represent the margin of an ice mass which impinged on the coast from Gulf of St. Lawrence and created lakes in the intermontane valleys. On a smaller scale, Skye Glen valley has numerous small kames and kame terraces, and is blocked at its southern end by a large kame moraine at Churchview – all related to an ice lobe that retreated southward to Whycocomagh.

Ice-contact stratified drift is common in most of the highlands valleys, presumably because the high relief promoted ice stagnation. Kames and small kame moraines, produced by retreating outlet glaciers from a plateau ice cap, choke northern valleys, notably those of the Ingonish, Clyburn, Dundas, Aspy, Blair (Fig. 60), Red, Grande Anse, Mackenzie (Fig. 46), and Chéticamp (Fig. 61) rivers. On the west coast in the Inverness area, a major sandy kame moraine complex fills a buried valley, thereby blocking the preglacial drainage and creating the present Lake Ainslie. One of its major parts is a large sandy ridge (first noted by Honeyman, 1890) which borders Lake Ainslie at Kenloch. Other major kames and kame moraines occur in the two intermontane valleys which drain about one-quarter of the highlands: in Middle River valley at lakes O'Law and Ulva and in Margaree valley at Margaree Forks and Margaree Harbour. All eastward-draining highlands valleys also contain significant kame deposits, and the adjacent coastal lowland is similarly blanketed with kames and kame moraines. Lastly, the coast between Margaree Harbour and Chéticamp is blanketed by extensive kame fields composed of crystalline gravel from the highlands which is inferred to have been deposited against an ice mass lying offshore. The gravel is incised by the so-called Chéticamp paleochannel which is probably a tunnel valley that carried subglacial meltwater in the inferred contact zone of the onshore and offshore ice masses. Overlying the kame gravel (and present also in the channel) are thin but extensive




Map units: Rb. forested bedrock; Rc. bedrock with thin till patches; 2a. till blamket; 3a. ice-contact stratified drift (kame moraine); 3b. outwash plain; 5a. paraglacial fan; 5b. postglacial alluvial terrace; 6. coastal beach; 8. salt marsh

sandstone hogback (tick on dip slope) — interlobate moraine

*Figure 59. Stereogram of lower reach of Margaree River valley showing hogback ridges, interlobate moraine, and outwash plain being onlapped by fan, modern delta, and salt marsh. NAPL A21576-5, 6*

patches of a red stony silt (Fig. 62) that is discussed below in the section on Glaciolacustrine deposits and lake-level indicators.

### Outwash plains and fans (unit 3b)

Outwash is usually associated with ice-contact deposits because it is the product of proglacial alluviation by meltwater streams, and because it becomes inset into ice-contact deposits during ice retreat when subglacial gravels become dissected and reworked. It is commonly deposited on and among buried ice blocks which, during later melting, produce kettles and collapse structures (Fig. 63). In the absence of collapse structures, outwash terraces are locally difficult to separate from postglacial degradational terraces (unit 5a). Outwash has a composition similar to kame material, but its sorting is better and it has a greater proportion of resistant rock types. The Margaree and Middle river valleys have the most extensive outwash plains; all of the other valleys mentioned above have outwash, as do a number of smaller watercourses.

*Figure 60. Hummocky kame moraine deposited by late valley glacier in Blair River valley. (Site [75] is cutbank at stream mouth); Pollett Cove, northwest coast. GSC 204042-D*

*Figure 61. Interlobate kame moraine ridge composed largely of ice-contact gravel deposited at confluence of offshore ice sheet and outlet glacier issuing from Chéticamp River valley (on left). GSC 120200*


*Figure 62. Outwash with imbricated pebbles, overlain by red stony silt with columnar jointing; mouth of Chéticamp River [68]. GSC 203752-V*

*Figure 63. Muddy outwash gravel with collapse structure due to melting of buried glacier ice. (Similar structures, but with surface pits, occur in postglacial and modern alluvium due to karst development in underlying gypsum); Middle River. GSC 202188-O*

### Meltwater flow-direction indicators

The trend of eskers, the gradient of meltwater channels, and the surface slope of outwash reflects the pattern of meltwater flow. From this, three main glacier masses can be discerned on Cape Breton Island and their retreatal pattern inferred. The convergence of outwash systems over most of the southern lowlands shows that the largest ice mass was situated over Bras d'Or basin and that it retreated more or less concentrically, although with a highly lobate margin because of the major arms of the lake. A smaller, and largely independent ice cap was situated on the highlands plateau. Its early-stage outlet glaciers left deposits in the trunk valleys and its final shrinkage onto the plateau cut flights of tiny proglacial and sidehill meltwater channels into the residuum (Fig. 13). It retreated more or less concentrically to the highest ground; the last remnant occupied the divide between Ingonish and Chéticamp rivers. A third ice mass, evidenced by deposits and features in River Inhabitants and Judique areas, was evidently part of that situated on the eastern uplands of Nova Scotia mainland. It lapped onto the island in the Strait of Canso area and, at an early stage, was probably contiguous with the Bras d'Or ice mass.

## *Glaciolacustrine deposits and lake-level indicators*

Glaciolacustrine deposits are sediments laid down in lakes dammed by former glaciers. Glacial lake sediments are rare in Cape Breton Island and of negligible area; they belong to two groups. One group comprises scattered hanging deltas with accordant tops and associated overflow channels which indicate they are the product of small impoundments dating from the last ice retreat. They are shown on Map 1631A and on Figure 56, but the associated fine grained muddy deposits on the former lake floors are rare and too small in areal extent to be shown. No fossil beaches or trimlines from the inferred lakes were recognized, but no detailed search was conducted on the generally forested slopes where they should occur. Perhaps there was no wave action because the lakes were perennially frozen over. The other group of sediments (not shown on Map 1631A) lies in the subsurface and is only rarely exposed. It is only assumed to be glaciolacustrine on the basis of texture and geographic setting.

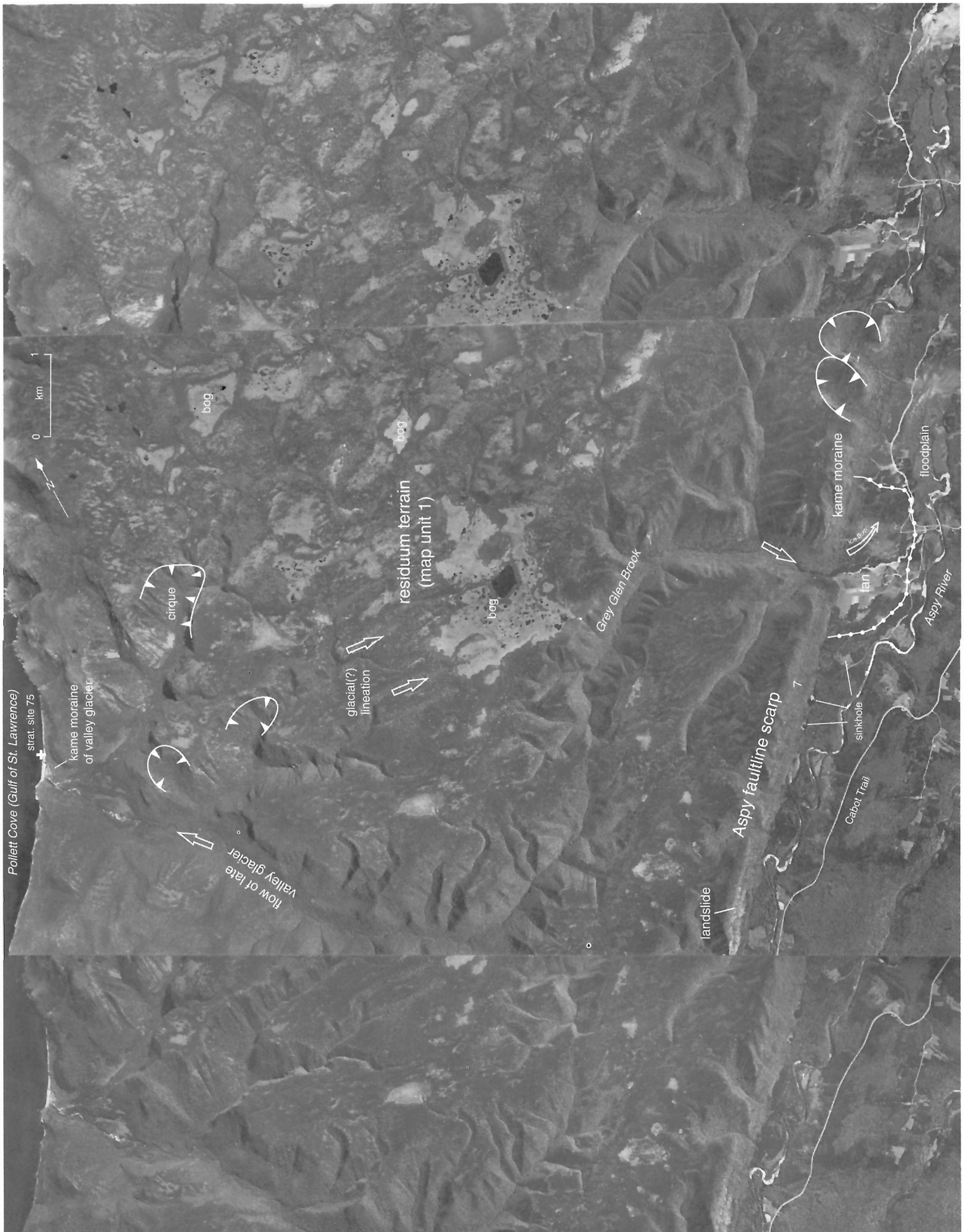
### Subsurface glaciolacustrine(?) deposits

Supposed glaciolacustrine deposits lying in the subsurface consist of a few scattered occurrences of compacted laminated silt and clay beneath till and other sediments. The oldest and perhaps most important group of subsurface deposits of possible glaciolacustrine origin are exposed in coastal bluffs along the East Bay coast. A stony, locally laminated, slightly organic mud underlies surface tills, notably at Castle Bay [25] where it attains an elevation of 20 m. The mud contains a mixture of freshwater, brackish-water, and marine diatoms whose proportions vary irregularly through the section. On that basis, de Vernal and Mott (1986) postulated a marine inundation with a fluctuating fluvial input of freshwater microfossils. They estimated the submergence lasted several tens of thousands of years based on pollen concentration calculations. However, there is no independent evidence for sea levels of that height. The suggested alternative explanation is that the problematical sediment is proglacial lake mud combined with redeposited marine sediment. The postulated scenario is that, as the early eastward-flowing ice crossed eastern Cape Breton Island, it advanced up East Bay converting the marine basin to a proglacial lake. The previously deposited marine mud was plowed up and debouched into the lake where it was mixed in temporally varying proportions with freshwater material. The duration of the lake would have been only a few thousand years, given the short length of the bay and average glacier advance rates. This hypothetical proglacial impoundment is provisionally named "glacial lake Cameron", because just such a lake was figured (but not named) by H.L. Cameron on one of his unpublished map interpretations presumably because he observed the same sediment and was aware of the eastward ice flow.

A younger group of subsurface lake sediments is found in two areas. At Mira Bay on the east end of the island, Ries and Keele (1911) mentioned subsurface lacustrine silt and clay, but did not report the stratigraphy. The deposits were not seen during the present study, so they may simply represent local deposition in kettled kame moraine, or they may signify a lake phase prior to a readvance that produced the ice-contact complex. The other group of younger deposits, of which most were first reported by Fletcher (1881) occurs in the River Denys Lowland west of Bras d'Or Lake and consists of many small, widely scattered exposures of highly compacted, fractured, laminated red silt beneath a thin patchy surface till (Fig. 64). The sediment is barren of diatoms, so its freshwater origin can only be assumed from the fact that this lowland also held a lake during the final retreat of the ice, as evidenced


*Figure 64. Compacted, fissile glaciolacustrine silt and sand overlain by till; highway ditch near Valley Mills, River Denys Lowland. GSC 203504-M*

*Figure 65. Stereogram of Aspy scarp showing V-shaped gorges, some with amphitheatre-shaped heads; also alluvial fans, and large recent landslide. NAPL A21576-24, 25, 26.*


Pollett Cove (Gulf of St. Lawrence)
strat. site 75
kame moraine
of valley glacier
flow of late
valley glacier
cirque
glacial(?)
lineation
residuum terrain
(map unit 1)
bog
bog
bog
N
0 km 1
Grey Glen Brook
Aspy faultline scarp
landslide
sinkhole
Cabot Trail
kame moraine
ice flow
fan
floodplain
Aspy River

by surficial deposits of a similar laminated silt (see below). Nearby at Big Brook quarry [49], organics dating 11 000 ± 90 BP (GSC-3398) are overlain by a silt, presumed to be the same as that elsewhere in the valley, together with an overlying till, so the lake was evidently a late glacial feature which formed prior to a local readvance.

**Surficial glaciolacustrine deposits and features (unit 4)**

Twelve small gravel deltas are perched at an elevation of about 60-70 m in the small gorges draining the north flank of North Mountain bordering River Denys Lowland. They show typical Gilbert-type structure with planar inclined foreset beds and subhorizontal topset beds. Their distribution and accordant elevation indicate that a water body was impounded in River Denys Lowland by an ice mass blocking the lowland on the east. The other side of the lowland has no such perched deltas (or beaches or trimlines), only several large gravelly fans with apexes at ≈75 m, in Diogenes Brook and Blues Brook valleys for instance. They are mapped as alluvium, but perhaps they were actually deposited subaquatically in the same lake as the deltas. Patches of red, plastic silt, only a few decimetres thick and too small to be shown on Map 1631A, occur in the lower parts of the lowland and are presumed to be the associated lake bottom sediment. The thinness and patchiness suggests the lake was shortlived. The deposits and features were presumably noted by Cameron (1957) who showed a lake on the 1958 Glacial Map of Canada (Wilson et al., 1958) which he labelled "Glacial Lake Bell" (hereafter referred to as glacial lake Bell) on an unpublished manuscript sketch map of the area.

In the upper reaches of River Inhabitants valley, across a low divide from River Denys Lowland, there are five similar, though larger, Gilbert-type deltas at 75-90 m elevation. These are inferred to have been built into a proglacial lake impounded by ice lying to the south and anchored on the ends of Creignish Hills and North Mountain. This lake (unnamed) evidently overflowed into the River Denys Lowland at elevation ≈100 m via a large dry channel across the divide north of Glendale and built a small delta at elevation 60-70 m in the River Denys basin proglacial lake.

Smaller ice-dammed lakes in intermontane settings are indicated by deltas (too small to be shown Map 1631A) in Bridgend and Skye Glen valleys north of Whycocomagh. The inferred lake in Bridgend was impounded by northward-advancing Bras d'Or ice which stood at the valley mouth at Churchview where it built a large kame delta with northward-inclined foreset beds with silt laminae on the lake side. Water from the Bridgend impoundment spilled through a cross-axial channel at elevation 160 m to the other ice-dammed lake in neighbouring Skye Glen valley where it built a delta. The Skye Glen lake was also held up by the Churchview dam, while its northern end was dammed by ice that which stood near Nevada Valley settlement and deposited kames. Skye Glen lake overflowed to Lake Ainslie basin via a second cross-axial channel at 120 m. That channel is higher than the present divide to Mabou River and Gulf of St. Lawrence which means that ice from offshore must have plugged Mabou valley during the intermontane lake phase.

Another intermontane glacial lake possibly existed in nearby Margaree Valley. It was first depicted by H.L. Cameron on the 1958 Glacial Map of Canada (Wilson et al., 1958) and named "Glacial Lake Margaree" (hereafter referred to as glacial lake Margaree) on an unpublished sketch map. Later, E.H. Muller (unpub. typescript report, 1964) reported laminated silt and clay in kame gravel at Margaree Forks [62] and presumed it to be the sediment of the lake. Although the silt may represent only local deposition in a kettle pond and no other evidence of a lake in that area was found, such as varved silts on the valley floor and over the ice-contact gravels. The valley of Margaree River is tentatively inferred to have held a proglacial lake because the valley could have been blocked at its mouth by the east-flowing glacier which impinged on the west coast and held up the Skye Glen proglacial lake.

## *Fluvial deposits*

The deposits which are mapped as fluvial (alluvium) are river and stream sediments, mainly sand and gravel, that were laid down in channels and on floodplains not directly associated with glacier ice. They represent all the stages of alluviation and downcutting between the time glacier ice left a watershed and the formation of the present floodplain. Their combined distribution, which amounts to about two per cent, is shown in Figure 56. This category embraces two broad age groups of deposits which can be recognized by their distinctive geomorphic features and are shown on Map 1631A because of their economic and paleoenvironmental significance. One group, representing former higher floodplains, is termed paraglacial and postglacial deposits. It includes all abandoned terraces and fans situated above present floodplains and consists of sediments which were laid down during and after glacier retreat but does not include modern deposits. The earlier phases of this group were deposited while glacier ice was still present in the area, so they represent climatic conditions considerably different than today. The other group is the deposits of modern floodplains and fans.

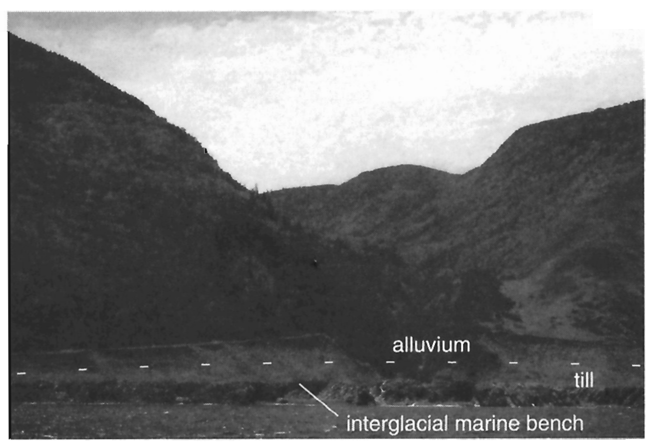
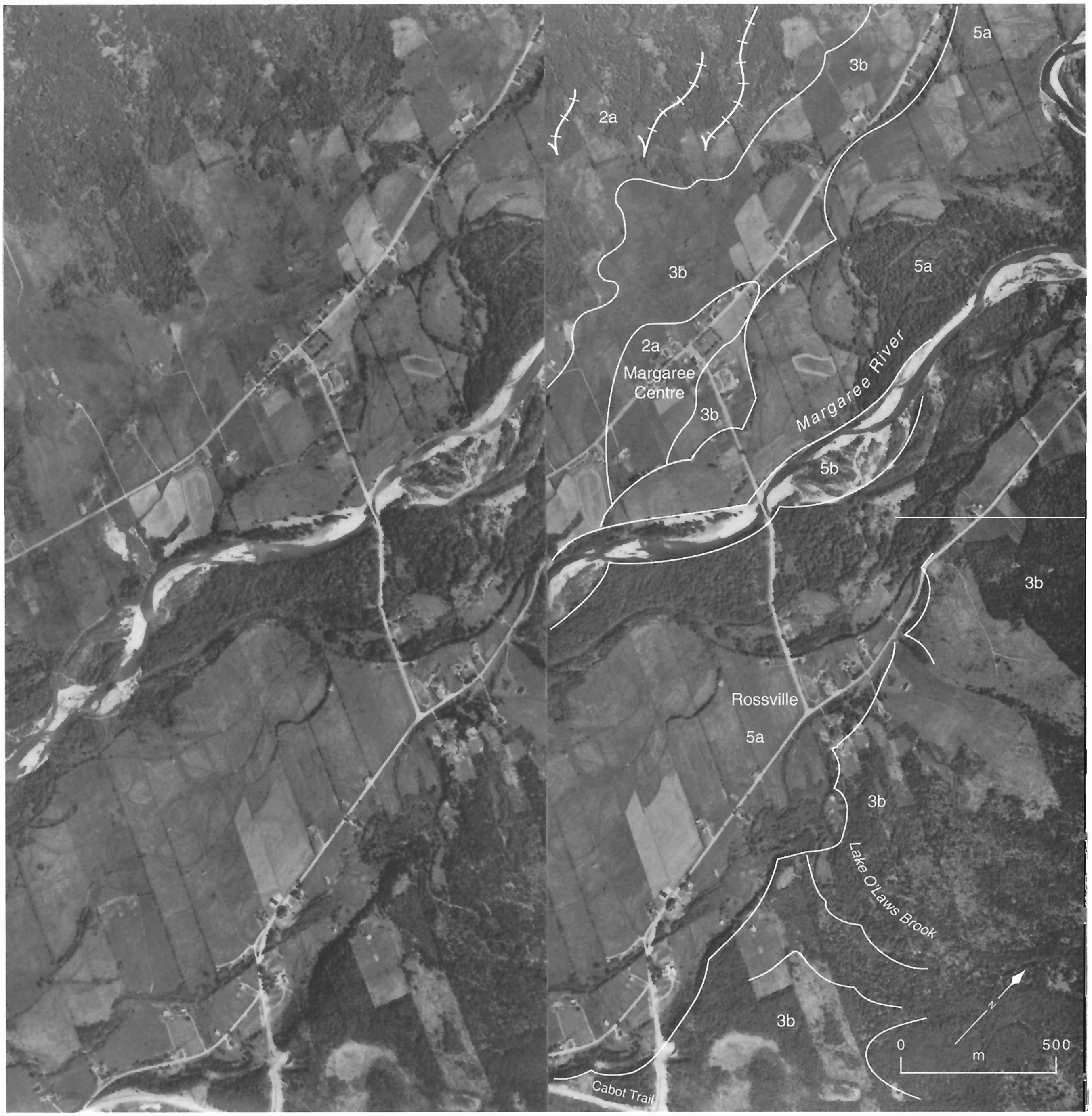



***Figure 66.*** *Inactive paraglacial/postglacial alluvial fan overlying till on emerged rock bench along Aspy scarp, Aspy Bay; note horizontality of bench and amount of shoreline retreat since fan was formed. GSC 203504-K*

## Paraglacial and postglacial alluvial terraces and fans (unit 5a)

This category of alluvium includes terraces lying below the level of outwash plains and above the reach of modern streams, as well as piedmont fans at the mouths of small gorges. Some of the earliest-formed deposits in this category were laid down during the deglacial phase while glacier ice was still in the area. The deposits formed under those conditions are termed paraglacial. The climate at that time was periglacial and vegetation was sparse or tundra-like. Unstable, partly vegetated, till-covered slopes shed muddy debris which accumulated as fans or as the highest postglacial terraces. Paraglacial sediment is therefore typically muddy, poorly stratified, and more angular than the younger, or




*Map units:* 2a. till blanket; 3b. glacial outwash; 5a. postglacial terrace alluvium; 5b. modern floodplain alluvium. terrace scarp meltwater channel

***Figure 67.*** *Stereogram of terraced outwash plain; upper Margaree River valley at Rossville. NAPL A30206-123, 124*

postglacial, deposits in this category. As climate improved and conditions were no longer periglacial, forests took hold, thus limiting the input of mud from till-covered slopes.

Paraglacial fans occur at the mouths of gorges draining steep slopes. Good examples are found in River Denys Lowland, at the margins of North Mountain and Creignish Hills, in Skye Glen, around Lake Ainslie, East Bay, and in Margaree valley. Exposures in borrow pits show the gravel to be angular, muddy, and with crude, surface-parallel stratification. Fossil fans and associated talus aprons along the Aspy scarp (Fig. 65) are truncated by coastal erosion and consist of blocky rubble with a muddy matrix (Fig. 66). They are clearly inactive, as they are incised by modern streams and are heavily forested.

Postglacial terraces are the remnants of former higher floodplains abandoned as streams cut down through earlier deposited alluvium. They are found in most valleys where glaciofluvial material was reworked as sediment supply diminished in response to revegetation of slopes. The gravelly sediment is finer grained, contains fewer mud layers, and is better sorted and stratified compared to the parent glacial and paraglacial gravels. This maturation reflects the generally declining discharge, decreased sediment throughput and greater time for reworking. The largest terrace systems occur in valleys with extensive and thick glaciofluvial deposits, such as along Margaree River (Fig. 67), Middle River, Aspy River, and Baddeck River. At the coast, Holocene terraces disappear beneath tide level because they are graded to lower sea levels and are now being overlapped by salt marsh and modern deltaic deposits (Fig. 59).

### Modern floodplains (unit 5b)

Active alluvial flats, which include floodplains, modern deltas, and undifferentiated alluvial fans, occupy 0.6% of the island (Fig. 56). Floodplains and deltas lie at or just above modern stream level, and are thus subject to periodic inundation. They are recognized by their unvegetated anastomosing channels and immature vegetation cover of grass, fen, or swamp. Four main groups of floodplains are distinguished. Headwater streams on the highland plateaus flow in shallow, gently sloping valleys in residuum. They are fen-covered with a grassy vegetation overlying shallow muddy sediment. These streams become torrents as they descend the escarpments through steep tortuous gorges; their floodplains are represented by narrow bouldery sediments, interrupted by rocky stretches. Floodplains on most lowland valley bottoms are the latest stage of reworking of older gravelly alluvium, and are thus composed of finer, cleaner, and better stratified gravel. (It may be noted in passing that, where these floodplains overlie gypsum, for example along Margaree and Middle rivers, sinkholes are developing indicating active dissolution in the subsurface.) Lastly, some sluggish, low-gradient streams, such as those reworking fine grained tills in areas of deranged drainage, like River Inhabitants, have wide peat-covered silty floodplains. Streams emptying into Bras d'Or Lake typically produce digitate or "bird's foot" deltas (Fig. 68) because of the negligible tidal range, whereas those on the coast have anastomosing channels and islands, grading to broad intertidal flats. Coastal deltas are presently aggrading and transgressing landward over floodplains and older alluvium, owing to current sea level rise.

Modern alluvial fans, like their paraglacial counterparts, are located where small streams debouch from short steep gorges at the foot of highland and upland escarpments. Typical examples occur along the northern coast, along Aspy scarp (Fig. 65), and along steeper slopes of some southern uplands such as North Mountain and Boisdale Hills. They are typically composed of rounded pebble/cobble gravel.

### *Marine and lacustrine deposits (unit 6)*

Marine deposits shown on Map 1631A amount to less than one per cent of the island's area (Fig. 56) and are limited to the larger barriers, bars, and spits which have accumulated by the action of waves and currents on the present seacoast, including Bras d'Or Lake. Not shown on Map 1631A because it is too narrow, is the swath of unconsolidated material that wave attack, erosion, and reworking in the present regime of rising sea level have produced along virtually the entire shoreline, except for some precipitous northern highland segments. The emerged Sangamonian rock bench and associated beach gravel are also not delineated on Map 1631A because they are generally too small and covered by other deposits, but their distribution is shown as a bold line on Figure 30.

Littoral deposits are common on the exposed Atlantic Ocean and Gulf of St. Lawrence coasts where vigorous wave action erodes rock and till and the resulting sediment transported by longshore drift sediment to form spits, tombolos, barriers, and bars. The longest series of barriers and tombolos in Nova Scotia occurs between Gabarus Bay and St. Peters Bay. The largest unit in the series is a festoon of gravelly cordons anchored on rock knolls at Framboise (Fig. 44). The material in these beaches has been derived entirely by destruction of drumlins; the silicic bedrock is virtually indestructible to waves (Wang and Piper, 1982). Smaller barriers and tombolos occur on Isle Madame (Fig. 26, 69). At Aspy Bay, beach

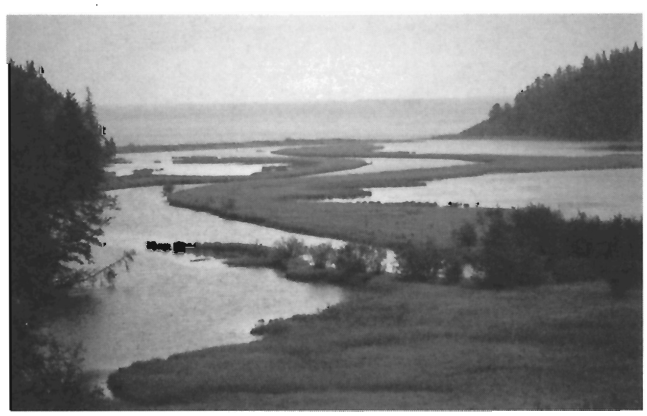


***Figure 68.*** *Digitate "bird's foot" delta, typical of those in microtidal Bras d'Or Lake; Scott Brook, West Bay. GSC 203752-Y*

formation on an area of weak, till covered bedrock has produced large sandy baymouth barriers that have straightened the shore to a perfect curve and created large freshwater lagoons (Fig. 70). Similarly, sandy shingly baymouth barriers and midbay bars typify the weak sedimentary rock coast between Mira Bay and Sydney. Here, the soft sandstone is being destroyed to form a wide intertidal rock bench. The 3 km long Englishtown barrier at the mouth of St. Ann's Bay, is actually a compound recurved spit, the largest on the island, which consists of a dozen progressively more lichen-covered cobbly ridges the oldest more than 2 m lower than the modern one. On the Gulf of St. Lawrence coast, modern transgression has cut back till and nonresistant sedimentary rock to form a series of remarkably gentle curves; the material thus eroded has drifted north and south alongshore to form sandy spits across the mouths of most estuaries, notably at Judique, Mabou, Inverness, Margaree, and Chéticamp. The symmetrical cuspate foreland at Port Hood is the remnant of a tombolo which, according to Desbarre's 1791 chart (McIntosh, 1923), connected Port Hood Island to the mainland when sea level was lower.

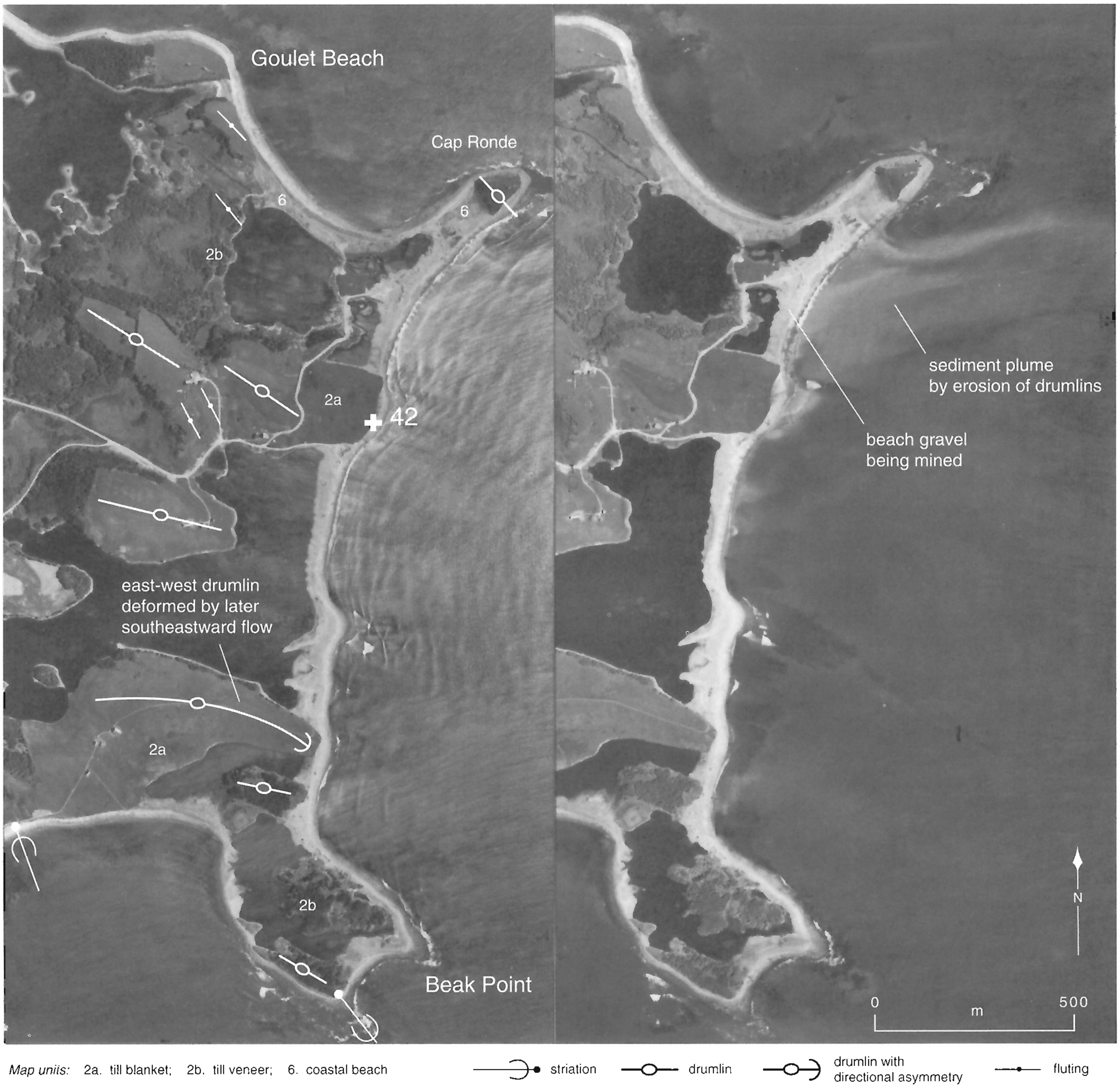



***Figure 69.*** *Stereogram of tombolo beaches resulting from erosion of drumlins at Cap Ronde, Isle Madame; note sediment plume caused by rapid erosion due mining of beach gravel. NAPL A30204-8, 9*

On the more sheltered, less active Bras d'Or Lake shore, beaches are correspondingly fewer and smaller. Those large enough to show on Map 1631A are almost entirely related to the erosion of till, primarily in the form of drumlins. Because the till bordering Bras d'Or Lake is relatively fine grained and with minimal stone content, even the relatively weak wave action is sufficient to truncate and carry away the entire section above water level, leaving only a bouldery carapace in the nearshore. The boulder shoals in West Bay are typical examples of such truncated drumlins and till mounds. Tombolos and spits are more common in East Bay, where longshore drift is appreciable and there are thick deposits of till and interbedded stratified material along the shore, as at Castle Bay site [25]. The numerous remarkably symmetrical cuspate spits and forelands along St. Andrews Channel and Great Bras d'Or (Fig. 71) were suggested to be due to current oscillations or seiches (Tarr, 1898; Woodman, 1899; Johnson, 1925, p. 359). The theory of converging longshore currents is still extant (e.g., Snead, 1985), but the general feeling is that these symmetrical spits can be attributed to edge waves (e.g., Holman, 1983).

The Cape Breton Island spits and barriers typically consist of successive parallel ridges, the crests of which rise in elevation from older to younger. The older ridges are commonly partly submerged or buried by salt marsh (Fig. 72) and the emergent ridges are progressively more lichen covered, showing that they have been added over a long time. This climb of berm height in all settings irrespective of aspect, wave energy, or tidal range suggests progradation in a rising sea level. To estimate the span and frequency at Englishtown: ≈20 emergent ridges climb 1.9 m which, if sea level is rising 16-20 cm/100 a (see Recent relative sea level rise and crustal subsidence), represents ≈1000 years, or approximately 1 ridge per 50 years, like the 45-year Double Hale solar cycle (Fairbridge and Hillaire-Marcel, 1977). That the modern berm truncates the older berms indicates cessation of accretion and a change of coastal regime, possibly because sea level rise has recently accelerated and sediment supply is insufficient to offset it (Grant, 1975b, 1987).

## *Colluvial deposits and slope processes*

Colluvial deposits, as mapped, cover about 4% of the island (Fig. 73) and are defined as sediments that have accumulated as a direct result of gravity-induced movement by falling, sliding, flowing, or creeping, usually assisted to some degree by the action of water. Colluviation thus encompasses such rapid and sudden occurrences as rockfall, landslide, avalanche, and debris flow, as well as the slower, more prolonged movements due to creep and solifluction. Colluvial deposits of sufficient extent and thickness to qualify as map units occur as mantles on steep valley walls and other sloping surfaces, and as fans and aprons at the base of some slopes. They are subdivided into two categories according to topographic setting and mode of emplacement. The most important group (unit 7a) is found throughout the island and comprises steep sheets of rubble or talus on steep slopes, locally surmounted by rock cliffs (Fig. 74, 75). A secondary group (unit 7b) comprises valley-bottom fillings of slumped and solifluted debris interbedded with fluvial sediment which occur locally in southern sedimentary rock uplands. A more localized group (unit 7c) consists of thick, gently sloping aprons of rubble and solifluction/sheetwash debris, which occur at the base of slopes in the northern highlands. Individual colluvial features not qualifying as map units are the rare landslides and certain segments of steep slopes that are failing by creeping or sagging.

### Rubble or talus on steep slopes (unit 7a)

The steep slopes which form the margins of the highlands and uplands, whether along the sea or in the dendritic valley systems, are composed in their lower two-thirds of rubble lying at its angle of repose (≈30°) which has accumulated by frost-induced failure of the upper bedrock part of the slope. The rubble or talus forms gravity-graded slopes that are recognized by their smooth, slightly concave profile and downslope-decreasing gradient (Fig. 74, 75). The deposits are wedge-shaped in cross-section, with a basal width of up to 200 m and thicknesses, as seen in coastal exposures and inferred from topography, which range up to 50 m. The rubble is angular and crudely stratified parallel to the surface slope. Coarse blocky layers alternate with muddy lenses, reflecting the interplay of rockfall and debris flow.

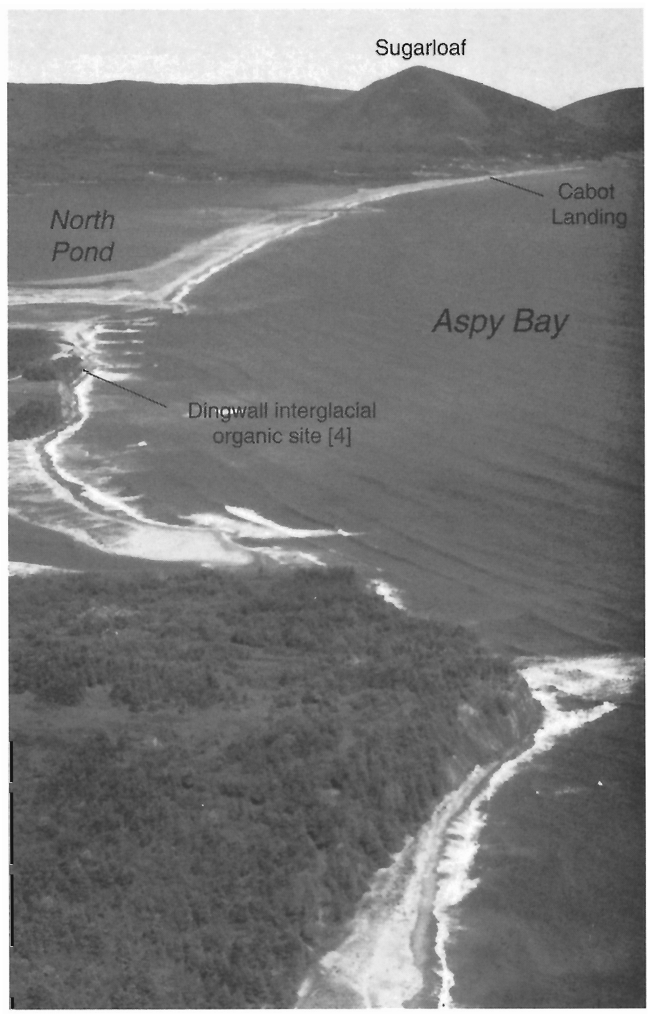



***Figure 70.*** *Aerial view of baymouth barrier beach enclosing lagoons; Aspy Bay. GSC 203752-R*

### Composite solifluction/talus valley-filling (unit 7b)

Certain valley sides and bottoms have received slope-borne material by the combined processes of rockfall and slumping and solifluction of till. The deposits form hummocky or irregular aprons up to about 20 m thick. Rare exposures show beds of rubble interlayered with lenses of loose diamicton which, from its varied and glacially abraded stones, is interpreted to be till sloughed from the upper part of the slope. Examples of such composite slope deposits are seen along Aspy scarp and in Ainslie Hills along Matheson Glen Brook.

### Talus and sheetwash aprons overlying other deposits (unit 7c)

Coastal exposures along the cliffs at the northern tip of the island are seen to be composed of a thick wedge of colluvial sediment overlying noncolluvial sediments. As examples, the coastal segment of Aspy scarp is blanketed with talus cones and aprons over a till bed that lies on the gravel-covered Sangamonian marine bench (Fig. 39e). Similarly, the coast of Bay St. Lawrence [1] is fronted by up to 30 m of colluvium overlying peat and gravel which in turn rest on the emerged bench. The colluvium is a loose, sandy, rubbly diamicton with crude stratification shown by sorted lenses which incline smoothly up to the 400 m high rock wall on the east. The debris was evidently shed from the rock slopes by sheetwash or debris flow. Similar sequences occur on the western highland margin and Mabou Hills, but at a scale too small to be shown on Map 1631A. This colluvium began accumulating as long ago as >49 000 BP, as shown by the ages on the underlying peat at Bay St. Lawrence [1], and continues to the present in most of the talus slopes.

*Landslides*

Innumerable small failures of bedrock cliffs and of till masses locally present on the plateau edge are evidenced by narrow deforested rock chutes and tracks 10-50 m wide on the talus.

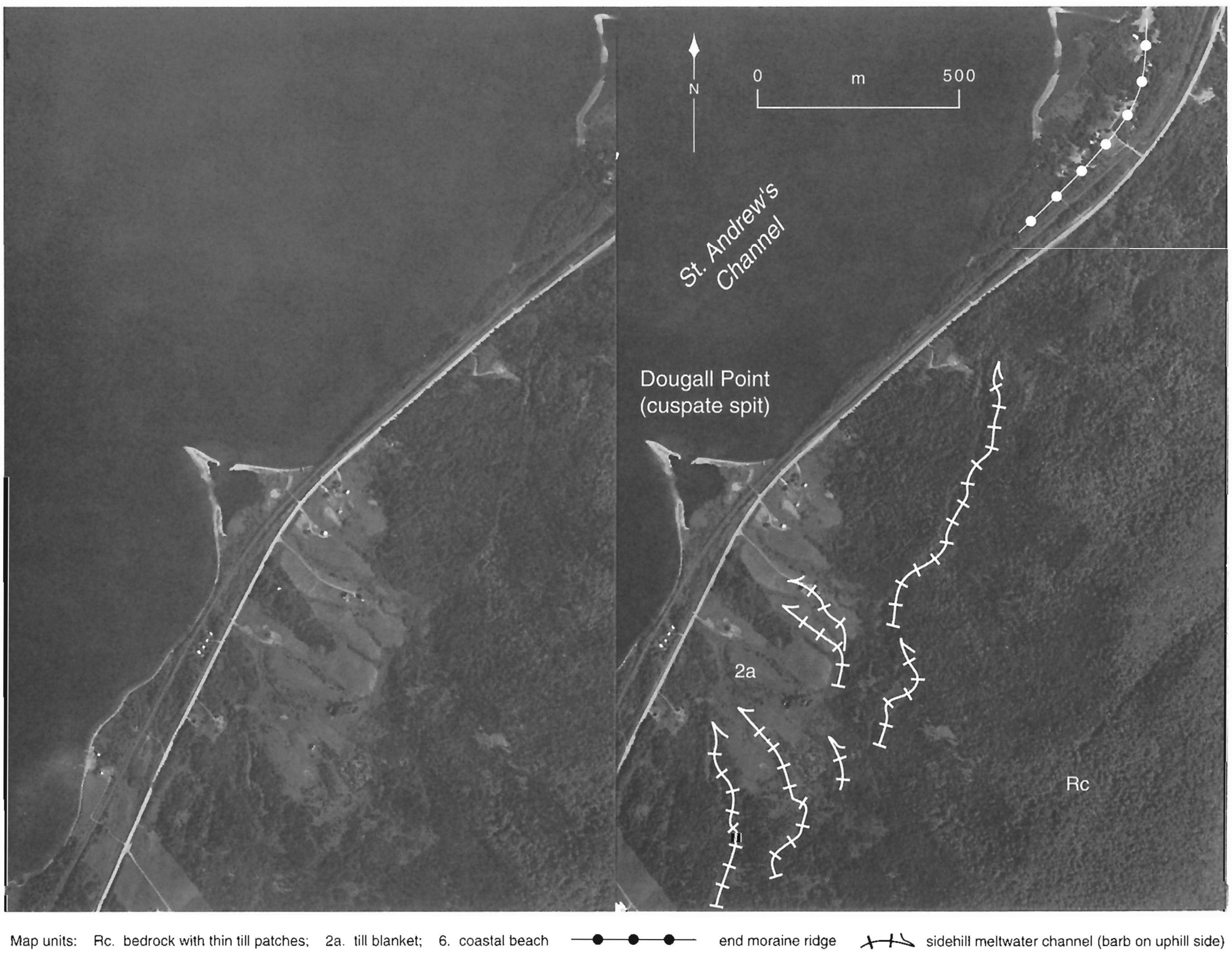
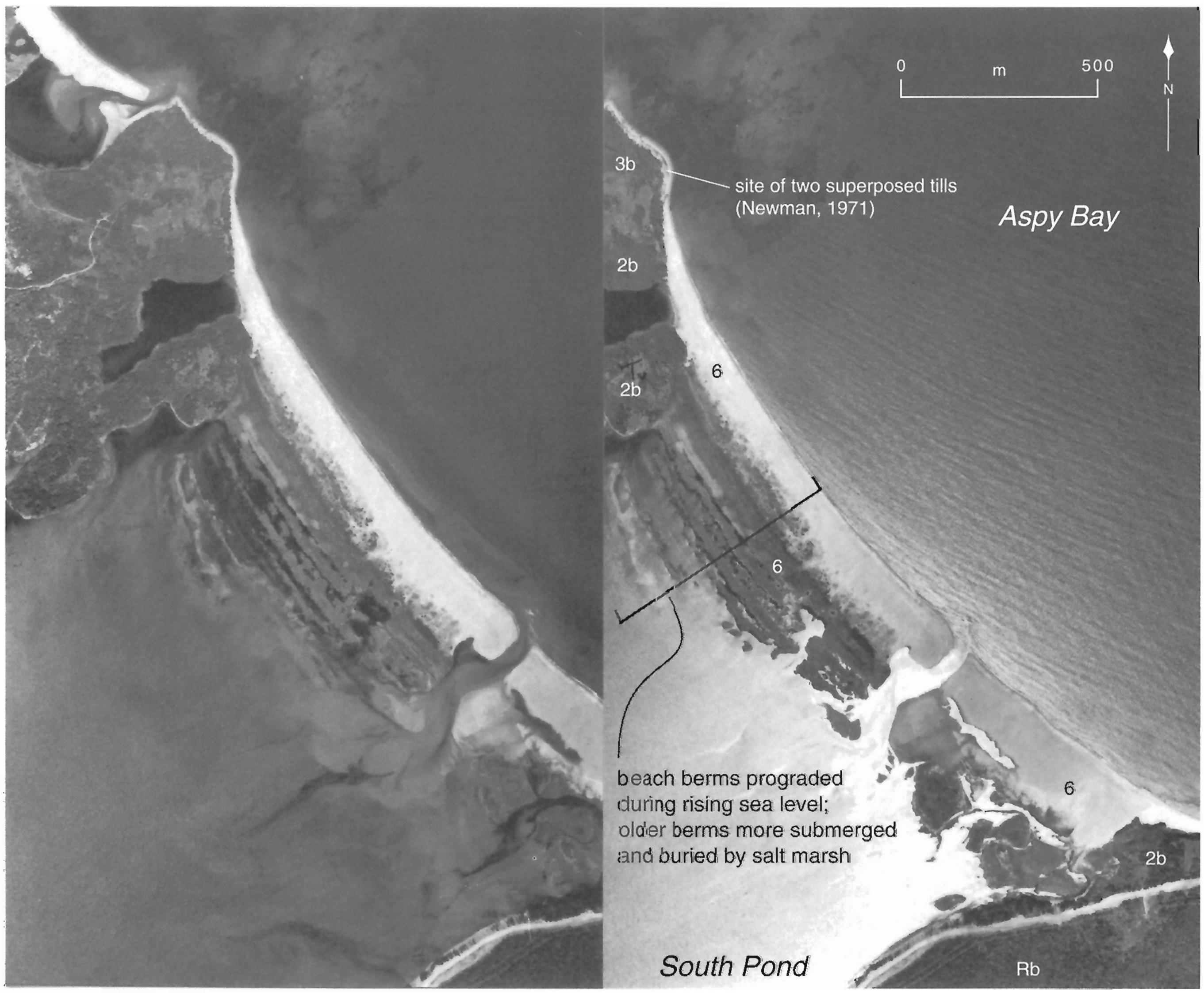



***Figure 71.*** *Stereogram of sidehill meltwater channels and small symmetrical cuspate spit near Boisdale, St. Andrews Channel. NAPL A30220-111, 112*

The slides are clearly the process by which the slopes are backwasted and the talus aprons have accumulated in post-glacial time in response to the widening and deepening of the stream channel and the consequent oversteepening of the slope. Exposures of the accumulated debris in borrow pits, at Rigwash valley, Wreck Cove, Ingonish, and Bay Road Valley, suggest that each fall adds about a metre of debris to the talus pile. If the aprons are about 100-200 m wide at the base and have accumulated in about 10 000 years, there would have been about one fall every 1000 years at any given point. Based on anecdotal evidence, many have occurred in late spring and are associated with snow avalanching. This would accord with the finding by Jordan and Toews (1993) for the Kootenay region of British Columbia that most of this kind of debris landslides occur "in mid to late spring during rain-on-snow or peak snowmelt periods when the ground is saturated and partly snow free". Hydrology is clearly the operative process. It is unclear whether the process is steady and unpredictable, but the impression gained by comparing airphotos taken in the 1960s with those taken about 20 years later was that more avalanche and debris tracks are visible on the later series, particularly in the highlands area. While this needs to be quantified by detailed counting on all of the several successive airphotos coverages since 1936, if correct, the apparent increase may signal a change in groundwater conditions perhaps associated with a climatic shift or perhaps to the forest clearcutting which has mainly taken place during this period (Jordan and Toews, 1993). Some recent work has been done on the temporal and spatial distribution of debris avalanching in the southern half of the highlands (Finck, 1992).




Map units: Rb. forested bedrock; 2a. till blanket; 2b. till veneer; 3b. glacial outwash; 6. coastal beach

***Figure 72.*** *Stereogram of beach complex composed of multiple berms that are progressively more submerged and overlain by salt marsh sediment in the lagoon; South Pond, Aspy Bay. NAPL A30199-49, 50*

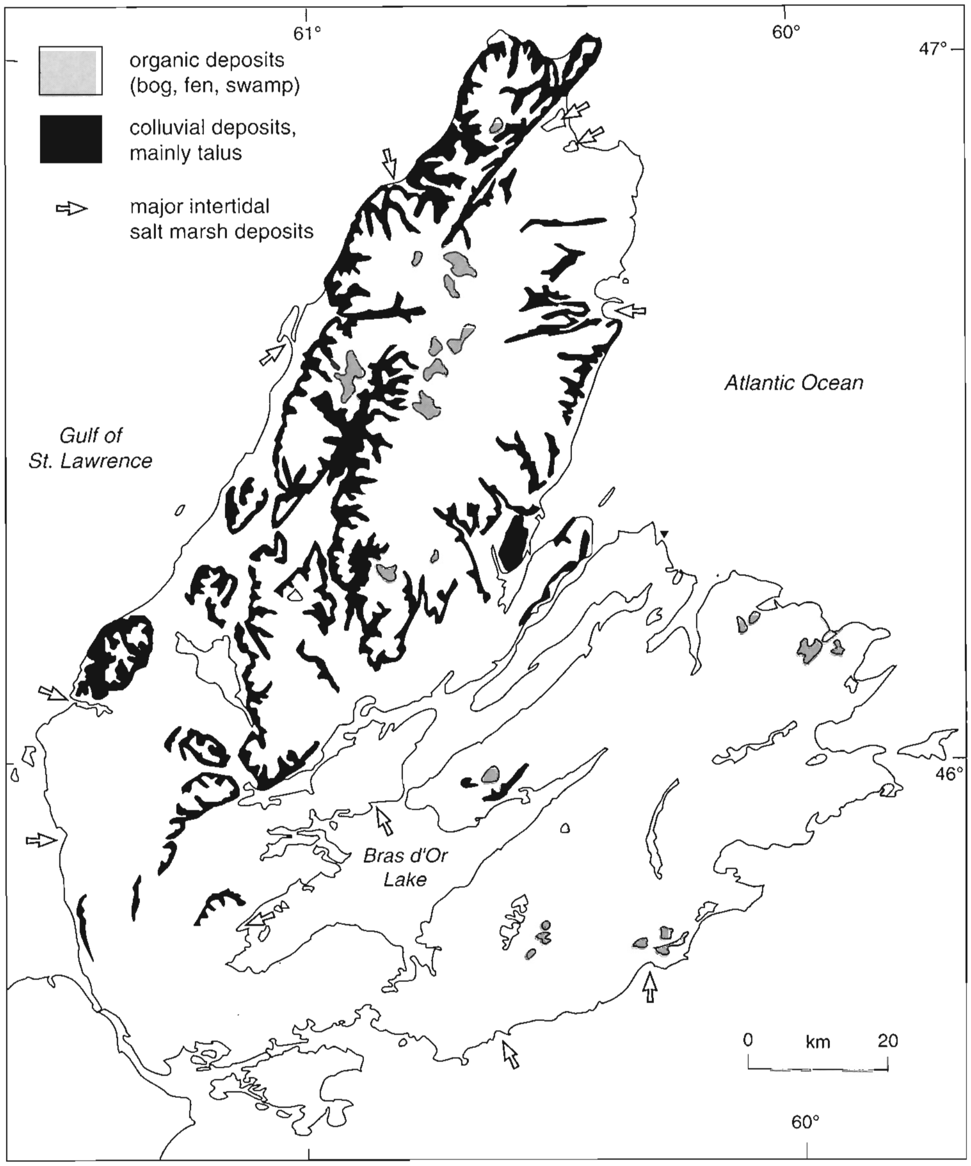



*Figure 73. Map showing distribution of colluvium, terrestrial organic deposits, and major intertidal salt marshes (adapted from Map 1631A).*

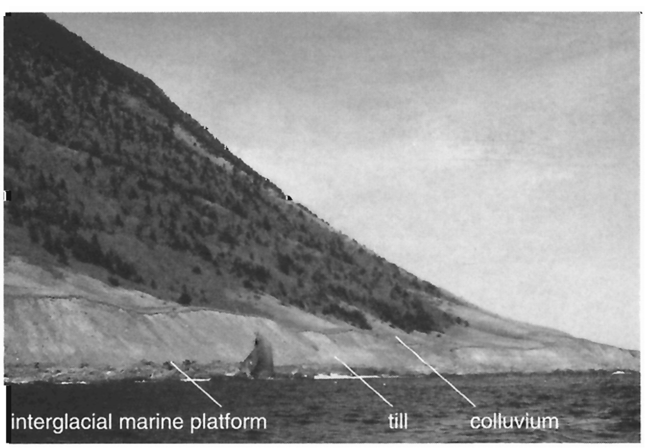



***Figure 74.** Rock rubble or talus composing colluvial slope with typical downward decreasing catenary profile, encroaching over undulating till deposits; Money Point. GSC 202188-P*

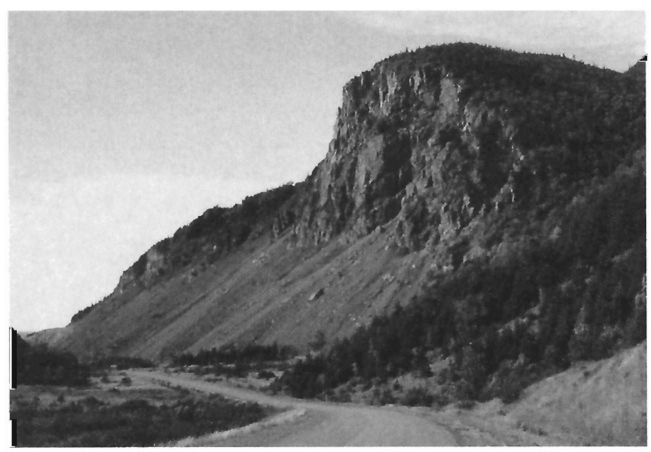
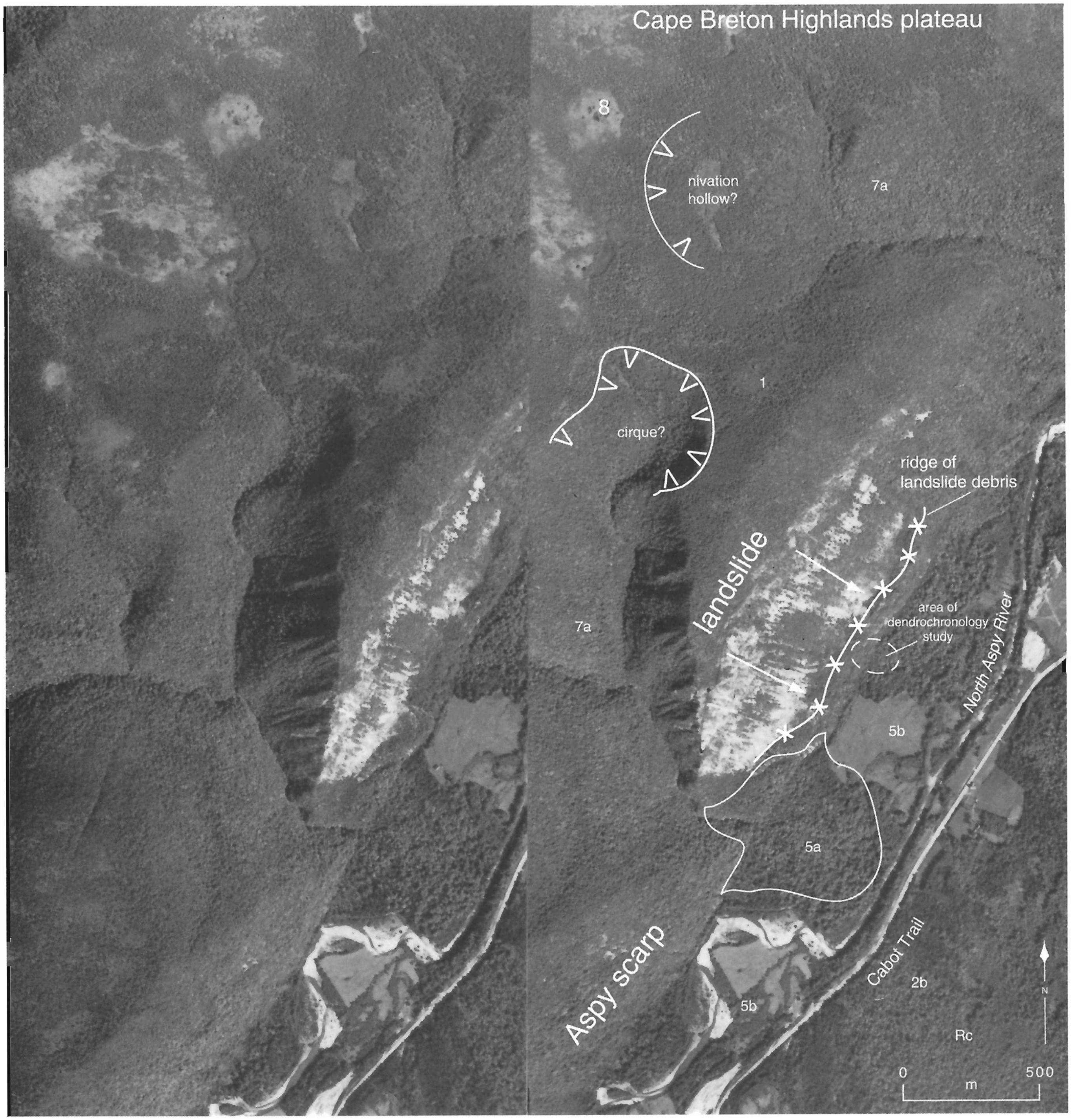


***Figure 75.** Active talus accumulating below failing rock cliffs; Rigwash valley, Cape Breton Highlands National Park. GSC 1994-413*

Two landslides are so anomalously large, however, as to deserve special mention. The largest is on Aspy scarp [6] (Fig. 76) adjacent to the Cabot Trail. The slumped material forms a hummocky linear festoon-like ridge with multiple lobes about 1500 m long, 200 m wide, and 50 m thick which slid onto a Holocene terrace of North Aspy River. Airphotos taken in 1936 show much less vegetation cover than the 1967 photos used in Figure 76. Unless the sparse cover was the result of fire, it seems likely that part or all of the slide occurred relatively recently, perhaps within the last century, although no local report or recollection of the event has yet been found. Based on a special dendrochronological study of trees on and beyond the slide (requested by the author) the slide is estimated to have occurred at least 80, and probably




Map units: 1. residuum (felsenmeer, etc.); Rc. bedrock with thin till patches; 2b. till veneer; 5a. paraglacial fan; 5b. postglacial alluvial terrace; 7a. colluvium talus); 8. bog, fen

***Figure 76.*** *Stereogram of large recent landslide on Aspy faultline scarp. NAPL A30223-50, 51*

closer to 90 years ago (R. Power, unpublished manuscript to J. Bridgland, Canadian Parks Service; pers, comm., 1992). The slide was probably a random event, but it may not be fortuitous that it apparently dates from a period when Cape Breton Island was shaken by two major earthquakes. That of 21 March 1904 occurred in southwestern New Brunswick (Leblanc and Burke, 1985; Ruffman and Peterson, 1988). It was felt in the Sydney area (Woodman, 1908) and therefore may also have shaken northern Cape Breton Island, which is no farther from the epicentre. A more likely candidate is the quake of 20 December 1909 which was centred on Cape Breton Island in the Inverness/Whycocomagh area (McIntosh, 1913) and had an estimated magnitude of 4.0 (Smith, 1962). Although reported not to have been felt in Ingonish (Ruffman and Peterson, 1988), lack of positive or negative reports in the more distant Aspy Bay area, which had no direct communication with the outside world at that time, leave open the possibility that the slide area was also shaken by the 1909 earthquake.

The other large landslide almost completely blocks Rigwash valley in Cape Breton Highlands National Park near the mouth of Chéticamp River (Fig. 77). It, too, is a hummocky linear festoon-like ridge with multiple lobes and also lies below steep, barren rock cliffs which contrast with the adjacent forested slopes. It has the appearance of a relatively young feature and efforts are being made to test its age by dendrochronology of trees on and beneath the toe.

*Slopes failing by deepseated, gravitational creep (sags)*

Two areas of steep mountain slope are failing by what appears to be slow, deep-seated, gravitational creep. A segment of the highlands margin between Presqu'ile and Cap Rouge is sagging downhill en masse (Fig. 78), disrupting the integrity of the Cabot Trail. It apparently is an example of deepseated gravitational rock creep, or "sagging". As detailed by Grant (1974a), the failure is localized on a steep slope composed of Precambrian phyllites inclined steeply seaward that are fractured and dissected by zones of incoherent crushed rock trending parallel and obliquely to the shore. A series of three airphoto coverages show the recent progress of the failure. Those taken in 1935 show the slope is intact and the Cabot Trail above the failure zone on the stable shoulder of the mountain. With the advent of motor traffic, the highway was relocated between 1937 and 1940 to the foot of the escarpment to a more easily negotiable route over a level strip just above tide level. That section of the highway was soon after washed away, presumably because there is a deepwater lead offshore, rather than the usual wide, energy absorbing, intertidal abrasion surface (Fig. 78). The highway was then rerouted to its present location on the sagging middle portion of the slope. Airphotos, taken in 1945 and 1947 when there was more cleared land, clearly show that the slope had begun to fail, producing several, large, displaced rock slices. It is concluded that the failure is the response of an inherently weak rock body oversteepened by accentuated marine erosion. The slope continues to sag and is being monitored. A similar failure has begun nearby at Corney Brook (R. Pereira, Public Works Canada, pers. comm., 1988) in the same structural setting of steep bedrock slices.

Another occurrence of sagging slopes is on the northwestern coast between Delaney and Sailor brooks known as High Capes (Fig. 79). In contrast to the generally vegetation-covered condition of stable slope, the slopes of High Capes are rubbly and marked by a steeped profile. The cliffs are failing in long, narrow slices, the tops of which are backward-tilted steps separated by scarps a few tens of metres high. The steps are being further broken up by expanding fissures, such that their outer parts have disaggregated into rubble. The 6+ m raised marine bench, which is almost continuous on the western highland margin north and south of this locality, is conspicuously absent along the sagging segment, so presumably it has been displaced below sea level by the downward movement of the sagging slope.

## *Organic deposits (unit 8)*

Organic deposits shown on Map 1631A cover about 3% of the island (Fig. 73) and consist of vegetal matter, mainly the remains of mosses, sedges, grasses, shrubs, and trees which


***Figure** 77. Large landslide with hummocky topography nearly blocking Cabot Trail in Rigwash valley, Cape Breton Highlands National Park. GSC 120198*


***Figure 78.** View of sagging bedrock slopes failing by deep seated gravitational creep at Cap Rouge (foreground) and at High Capes (far left), with emerged bench in middle distance. GSC 202188-T*

have accumulated in bogs, fens, swamps, and coastal salt marshes. Their distribution and rate of deposition are related to topography, precipitation, and water level, so these deposits provide a record of changing climate and sea level. Five different types of organic terrain occur in the area, although they have not been subdivided on Map 1631A. Basin bogs, which can exceed 10 m in thickness, generally occupy small, closed depressions in morainic and rocky terrains and have developed by infilling of ponds (Fig. 44, 48). Blanket bogs, while generally less than 2 m thick, are more extensive because they typically occupy the broad flat surfaces which characterize residuum on the highlands (Fig. 13, 20) and certain large till areas. Mostly composed of sphagnum moss, it is interesting that many have built up in areas formerly covered by forest (Fig. 80). The bogs have developed by excess precipitation on surfaces that have become waterlogged because of low permeability due to compactness or cementation. Some blanket bogs on the highest summits have a dry shrub and herb cover that is pimpled with tiny mounds a few metres across (Fig. 81), suggesting palsa or ice-cored swellings. Fens are meadows of sedge grasses, typically with ribs and elongate pools oriented across the slope (Fig. 82). Common on the highlands, and with organic matter usually less than 3 m thick, they have developed by seepage of mineral-rich groundwater. Swamps are stagnant-water areas of reeds and rushes along sluggish streams, such as River Inhabitants. Salt marshes, generally located in sheltered estuaries, are intertidal meadows of salt-tolerant vegetation, mainly grasses, which trap suspended mineral sediment

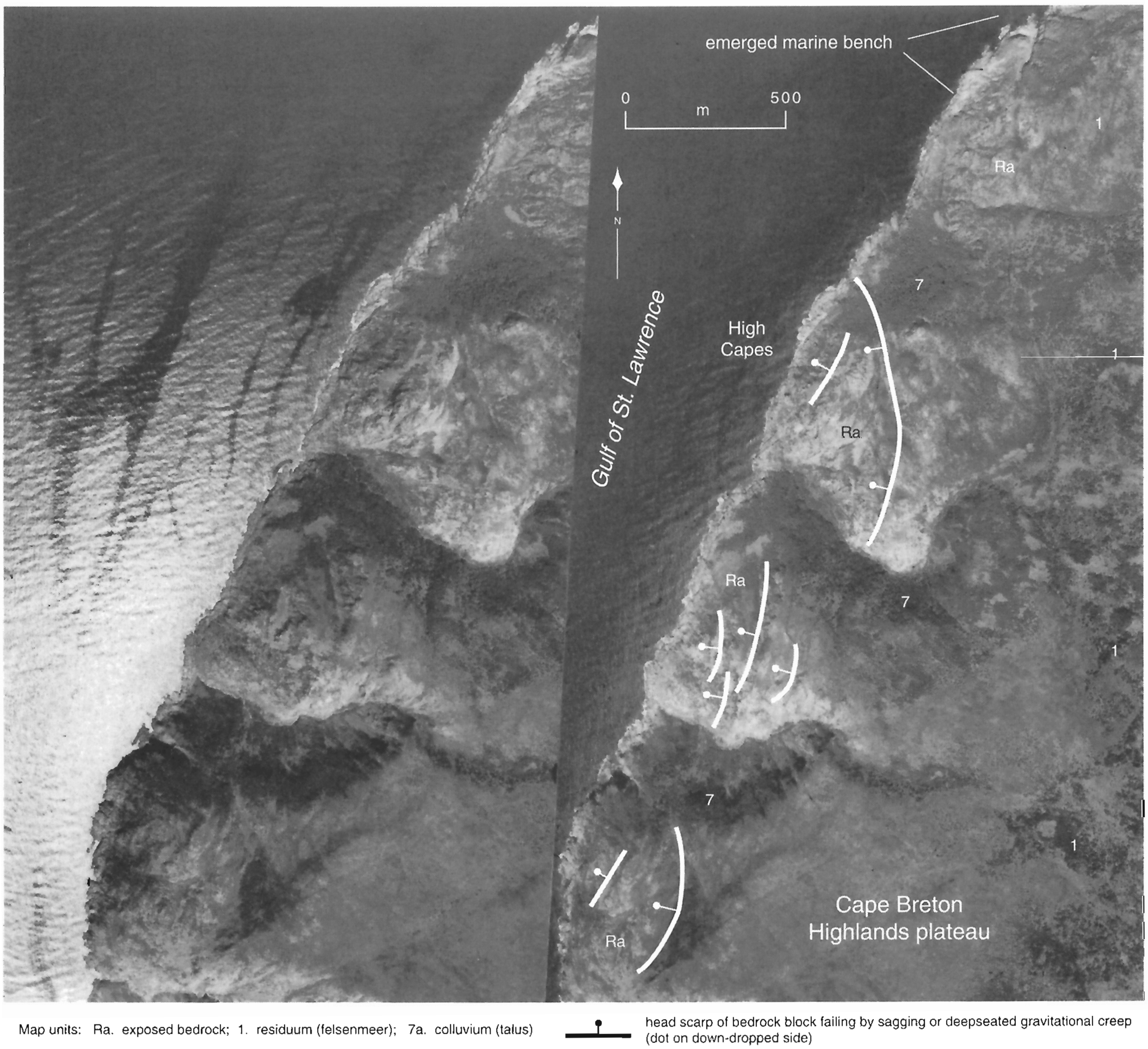



***Figure 79.*** *Stereogram of sagging bedrock slope at High Capes near Delaney Brook. NAPL A30199-26, 27*


*Figure 80. Roadcut through shallow blanket bog showing typical sphagnum peat overlying basal forest horizon rooted in till; near Wreck Cove Reservoir. GSC 203504-D*

during submersion. With rising sea level, they have aggraded to thicknesses of 10 m and have expanded landward over terrestrial humus and forest (Fig. 83).

### Eolian deposits

Eolian deposits are sediments transported and laid down by wind. They take the form of active and forested sand dunes which have developed in modern and subrecent time. Although too small to show on Map 1631A, the dunes are nonetheless important because of the tourist activity they have attracted. They have developed in association with sandy beaches at Point Michaud, Mabou Inlet, Inverness, Chéticamp, and Aspy Bay. The sand has been blown ashore from intertidal flats and reworked from beach berms. As shown by successive humus layers exposed in blowouts,

*Figure 82. Aerial view of slope fen with ladder pattern of elongate pools oriented across the slope; Cape Breton Highlands National Park. GSC 204042-X*

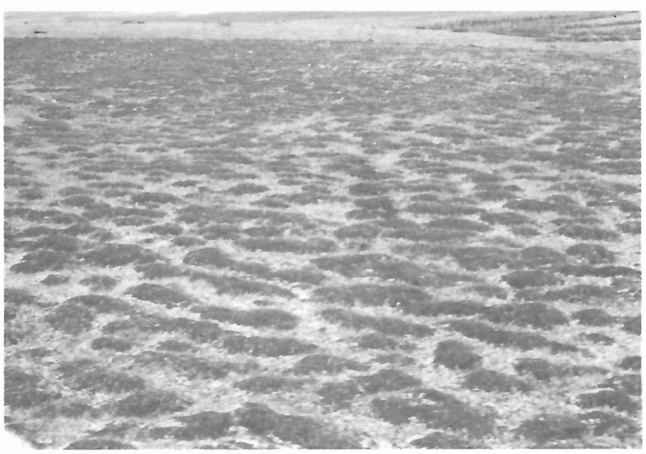


*Figure 81. Aerial view of peat bog with small mounds which may have been formed by ice growth (palsa?); Cape Breton Highlands National Park. GSC 204042-B*

*Figure 83. Salt marsh invading living forest, and showing submerged tree stumps; head of West Bay, Bras d'Or Lake. GSC 202188-S*

periods of sand accumulation and dune formation have alternated with periods of stability when vegetation fixed the surface. Judging by the generally active nature of most of these sand systems, the present climate is conducive to eolian activity.

## STRATIGRAPHY

Vertical successions of beds, together with the distribution and juxtaposition of surface materials and landforms, record the sequence of depositional events, which in turn can be put into a tentative geochronological framework with the help of age estimates by $^{14}C$ and thorium/uranium (Table 1) and amino-acid racemization ratios (Fig. 3). Sedimentary sequences outcrop extensively along natural and artificial exposures, of which about 100 key stratigraphic localities (Fig. 30) serve as the basis of the historical geology interpretation. The sequences vary in age, time span, environmental range, and complexity from preglacial to modern, and from single beds to lengthy composite nonglacial and glacial successions. The sequences span various parts of one lengthy glaciation and the last and the present interglaciations; for the purposes of the following description, they are grouped and organized according to the age of the lowest member of the sequence. Four broad age intervals are thus differentiated: preglacial sequences (predating Quaternary glaciation); Sangamonian and/or Early and Middle Wisconsinan nonglacial sequences; Wisconsinan glacial sequences; and postglacial sequences.

### *Preglacial sequences*

#### Saprolite and associated regolith

The only deposit found which may be preglacial is the saprolite which occurs on Second Fork Brook Road South [63] (Fig. 30, 31) on the highlands plateau in the residuum map unit described above in Surface materials and geomorphic features. Discovered by the author in 1974, when it had been exposed in a new roadcut, the saprolite is significant for both geomorphological and glacial history. McKeague et al. (1983) showed that the saprolite was the result of subaerial weathering and was thus the lower part of a paleosol. They presumed that it dated from the Tertiary, a period of generally warmer climate in the area, because of its degree of alteration and occurrence on a peneplain which Mathews (1975) suggested is of middle to late Tertiary age. The saprolite and small areas of the surrounding residuum are overlain by only one unweathered till (Fig. 38) and the peneplain surface of which it is evidently an integral part is virtually unmodified by glacial erosion. These relationships would appear to suggest that glacial action on this part of the Cape Breton Highlands plateau has been weak or infrequent throughout the Quaternary. A similar conclusion was reached by LaSalle et al. (1985) in connection with saprolites on the same peneplain in northern Gaspésie.

The relation of the saprolite to the surrounding glacial features raises difficult questions about the nature and timing of glaciation of the plateau. The saprolite is overlain by one till which is red by virtue of its matrix of comminuted Carboniferous redbed debris from the western lowlands. The red till is assumed to have been deposited by the same southeastward ice flow (Phase C) that swept across the entire plateau (Fig. 55) and produced the fluted till which covers part of the highlands. In the area of the saprolite and other forms of weathered rock, there is no sign of a till related to the late recessional movements of a local ice cap (Phase H) which elsewhere overlies the older till (as in the area of thick drift near Wreck Cove Reservoir, (R.R. Stea, pers. comm., 1992)). The older southeast flow is widely registered over the highlands plateau and it evidently stemmed from Gulf of St. Lawrence area. As to the age of this older plateau till relative to the tills on the lowlands and to the other mapped ice-flow events, it seems to overlie the tills deposited by northward-flowing ice (Phase D) (which are Wisconsinan because they overlie last interglacial deposits) and it is as unweathered as any of the surface tills. The southeastward plateau ice-flow phase is therefore Wisconsinan and probably one of the later phases.

It is puzzling why the southeastward flow deposited till and produced erosional effects over only about one-third of the plateau, leaving the rest of the plateau a virtually unmodified surface of residuum. It is also difficult to understand why earlier events left no trace. The residuum terrain in which the saprolite occurs encloses patches of till and is surrounded by typical ice-scoured and drumlin-lineated terrain with ice-flow trends which clearly show that glacier ice must have passed across the residuum terrain from one side of the plateau to the other. The only apparent explanation for the juxtaposition of unmodified residuum and ice-scoured till-covered areas is that the ice was cold-based and frozen to its bed over the residuum and warm-based and eroding over the till terrain. Even so, it is still difficult to imagine how all previous glaciations of the plateau could have been similarly cold-based and if they were not, why they left no trace. These unanswered questions render the reconstruction of glacial events on the plateau tentative.

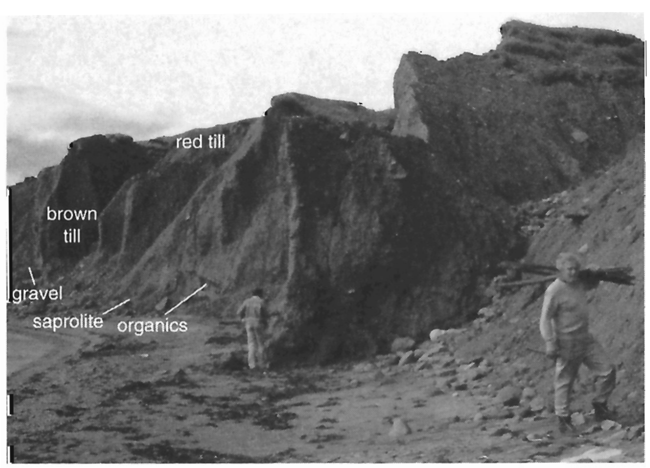



***Figure 84.*** *View of buried interglacial organic site at Green Point [57] overlain by gravel and tills. GSC 204042-L*

### *Sangamonian and/or Early and Middle Wisconsinan nonglacial sequences*

This group of depositional sequences includes those in which the lowest unit is a deposit or feature that is assigned on the basis of dates or stratigraphic position to the last or interglaciation sensu lato, which is broadly equated with oxygen isotope stage 5 (127-75 ka BP) as defined by Broecker et al. (1968), or to later periods, some as young as 47 ka.

#### Green Point [57]

This sequence on the Gulf of St. Lawrence coast at the mouth of Mabou Harbour (Fig. 30) was discovered by the author in 1981 when it first became exposed by cliff retreat (Fig. 84). The 10 m high bluff shows regolith developed in Carboniferous conglomerate bedrock of the Grantmire Formation, overlain successively by a woody organic bed, gravel, and two tills (Fig. 85). As seen in the lower part of the intertidal zone and where it forms Green Point, the conglomerate bedrock grades upward over a 2 m interval from a hard and fresh condition to a decomposed mass of leached and kaolinized stones in a friable matrix of hematitic sand. This rotted rock or regolith extends from midtide to about 1 m above tide level. It is sharply overlain and cleanly truncated on the north by a mass of well-rounded cobbles that are unweathered except for an oxidation rind 1 cm thick, and some secondary yellow iron hydroxide which cements the sandy matrix. (This gravel apparently also occurs under a buried organic bed at nearby Northeast Mabou River [58]; see below)

In one place, over a 30 m distance, a highly compressed organic bed 1-40 cm thick overlies the regolith with an undulating interfolded contact and is interlayered with the gravel along shear planes that slope up to the southeast (Fig. 86). It is composed of organic silt with peat laminae and wood fragments which range in size up to logs 20 cm in diameter and >2 m long. The wood is identified as hickory, ash, and juniper. The largest specimen, juniper, yielded a radiocarbon age of >53 000 BP (GSC-3320) and a Th/U age of 117 400 ± 10 000 BP (UQT-181). The latter age is judged to be a minimum measurement by de Vernal et al. (1986) and believed to be closer to 125 ka. According to R.J. Mott (pers. comm., 1987), pollen assemblages through the organic unit (Fig. 87) are dominated by thermophilous hardwood taxa, particularly *Quercus* (oak), *Tilia* (basswood), *Carya* (hickory), *Carpinus ostrya* (blue beech/ironwood), and *Pinus strobus* (white pine). Polypodiaceae (fern) spores and Gramineae (grass) pollen are abundant. The pollen and macrofossils indicate that closed forests of mixed temperate hardwoods and white pine occupied the surrounding hills. Ferns were abundant in the understory or in suitable locations bordering small ponds and marshy areas near the shore. The climate was significantly warmer and drier than is recorded palynologically at any time during the Holocene, including the Hypsithermal. On the basis of the fossils and supported by the age estimates, Mott and Grant (1985) assigned the organic bed to the thermal maximum of the last interglaciation, specifically to oxygen isotope substage 5e which culminated ≈125 ka.

The origin of the gravel is not clear, but there is reason to believe it may be a marine littoral deposit. The Green Point gravel may be fluvial sediment, but, while it lacks the planar bedding of typical beach gravel, its mature rounding strongly resembles that of the supposed littoral gravel which, elsewhere along the Gulf of St. Lawrence coast, overlies the horizontally truncated bedrock surface at about 4-6 m that is

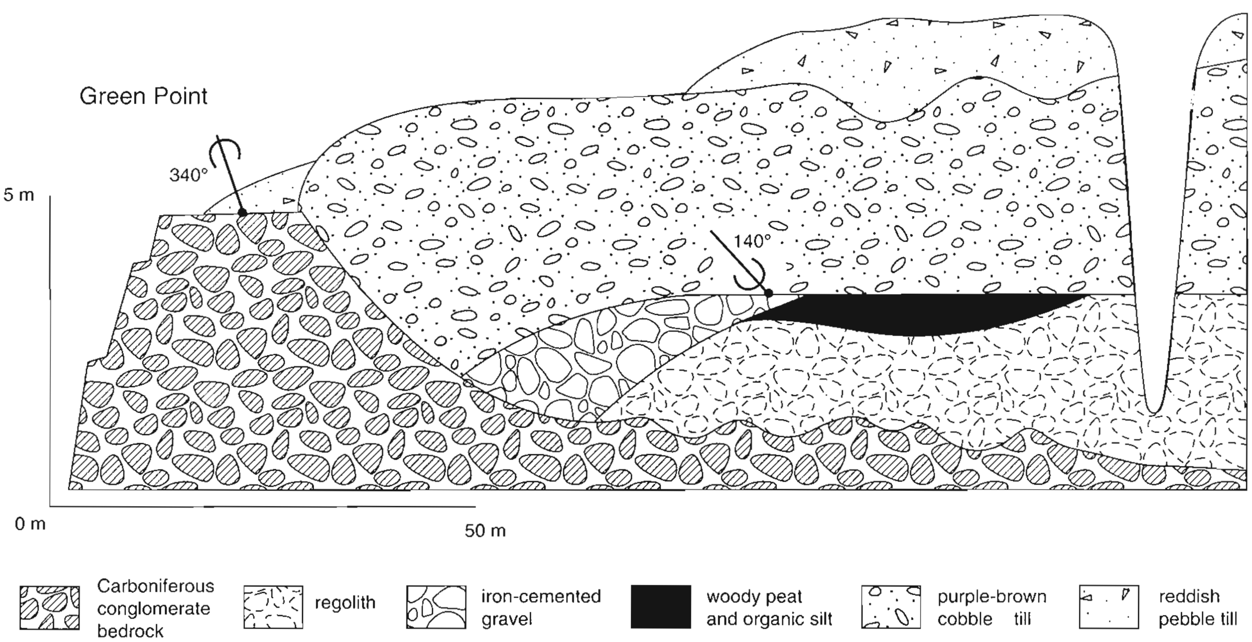



*Figure 85. Diagrammatic geological cross-section of stratigraphy at Green Point interglacial buried organic site [57] (after Grant, 1987, p. 32).*

interpreted as an emerged wavecut bench. The gravel is therefore tentatively regarded as a beach deposit graded to the level of the littoral platform represented by Green Point itself. If so, the beach gravel overlying the organic bed indicates that a marine transgression followed the thermal optimum of the last interglaciation. (The sea level maximum of the present interglaciation has also lagged the thermal optimum.) Alternatively, if the gravel is fluvial and the bench is represented by the truncated top of the regolith under the organic bed, then the organic bed was deposited after the marine maximum, proably as the surface became exposed during the subsequent regression.

The two overlying tills are different in texture, composition, and provenance; the upper till is embayed into the lower (Fig. 88). The lower till, which is several metres thick and


*Figure 86. Detail view of Green Point interglacial organic bed between regolith and till; note platy foliation induced by glacial shear. GSC 204042-S*

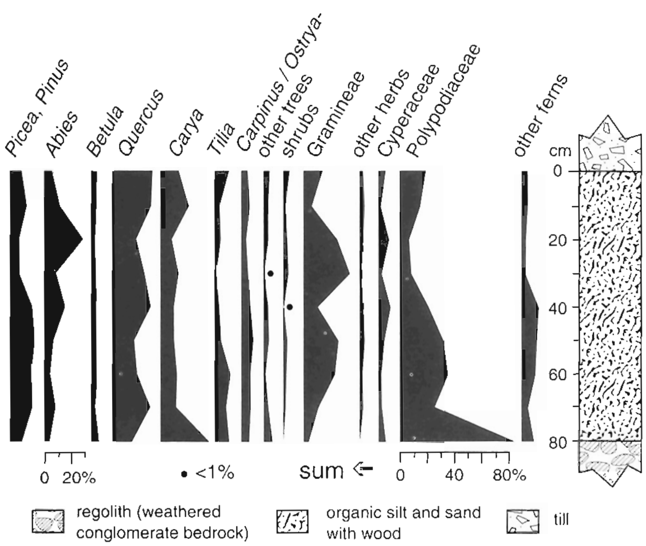



*Figure 87. Simplified pollen diagram for Green Point interglacial organic bed; note high content of thermophilous hardwoods (courtesy R.J. Mott, from Grant, 1987, p. 32).*

forms most of the section, is compact, purplish brown, and contains fragments of both upland crystalline rocks and lowland sedimentary rocks. It owes its dense red colour to a sandy/clay matrix composed mainly of comminuted redbeds that underlie the lowlands and Gulf of St. Lawrence. The direction of emplacement is taken to be 140° (the trend of striations inscribed on several large cobbles in the lower part and on a glacial facet truncating the top of one boulder in the underlying gravel). The same general trend is shown by a fabric based on a cursory count of the orientation of ≈100 elongate stones. The lower till is thus ascribed to an early southeastward onshore advance of ice from Gulf of St. Lawrence (Phase C; Fig. 55).

The upper till contrasts with the lower in that it has far fewer pebbles and cobbles, most of which are from sedimentary rock, and the reddish, sandy silt matrix contains more light coloured mineral components, chiefly from the inland crystalline terranes. The surface till is much thinner and has an undulating erosional contact with the lower till. It was emplaced by ice flowing 340°, judging by the excellent miniature crag-and-tail features on the underlying conglomerate bedrock on Green Point. The upper till is therefore attributed to the northwestward ice flow event which is recorded over much of the southwest quarter of the island (Phase D; Fig. 55).

Green Point thus provides key evidence of a temperate phase that is assigned to the Sangamonian thermal optimum, followed by two ice advances, one southward onto Cape Breton Island from Gulf of St. Lawrence, the other northward off the island.

### East Bay [27]

One of the most complete, albeit complex and disturbed, records in the Atlantic region of the duration and climate of the last interglaciation occurs on the north coast at the inner end of East Bay, Bras d'Or Lake, in a 200 m long bluff just east of the Church of St. Mary of Missions. The East Bay


*Figure 88. Brown sand till with undulating top underlying red silt till with solifluction-graded slope. Green Point [57]. GSC 204024-N*

deposits were evidently first noted by L.J. Weeks in the early 1950s and described as 'wood and leaves overlain by till-like material' (Prest, 1957, p. 447). The deposits unfortunately were not visible during the fieldwork phase of the present study, being obscured by forest which presumably grew up during a quiescent period. They were rediscovered, along with other exposures, by S. Occhietti in 1982 after a series of storms and high water levels cut back lengthy segments of the East Bay coast. The sequence is localized in and over karst depressions in Carboniferous gypsum which are infilled with various organic sediments and capped by two tills. Like all other exposures along the East Bay coast [20-27], the surface of the East Bay sequence is molded into northeast-trending drumlins, thus further demonstrating that the landforms in that area have been carved from mainly stratified sediments (cf. Fig. 43) and are only veneered with till. Chronostratigraphically, East Bay sequence appears to represent portions of a lengthy nonglacial period, perhaps as long as 60 000 years. It features three climatic warming cycles; the earlier two are considered to belong to the last interglaciation (stage 5) and the latter to Early and Middle Wisconsinan time (Mott and Grant, 1985; de Vernal and Mott, 1986; de Vernal et al., 1986; Grant, 1987). The salient features of this site are summarized here, together with observations not reported in

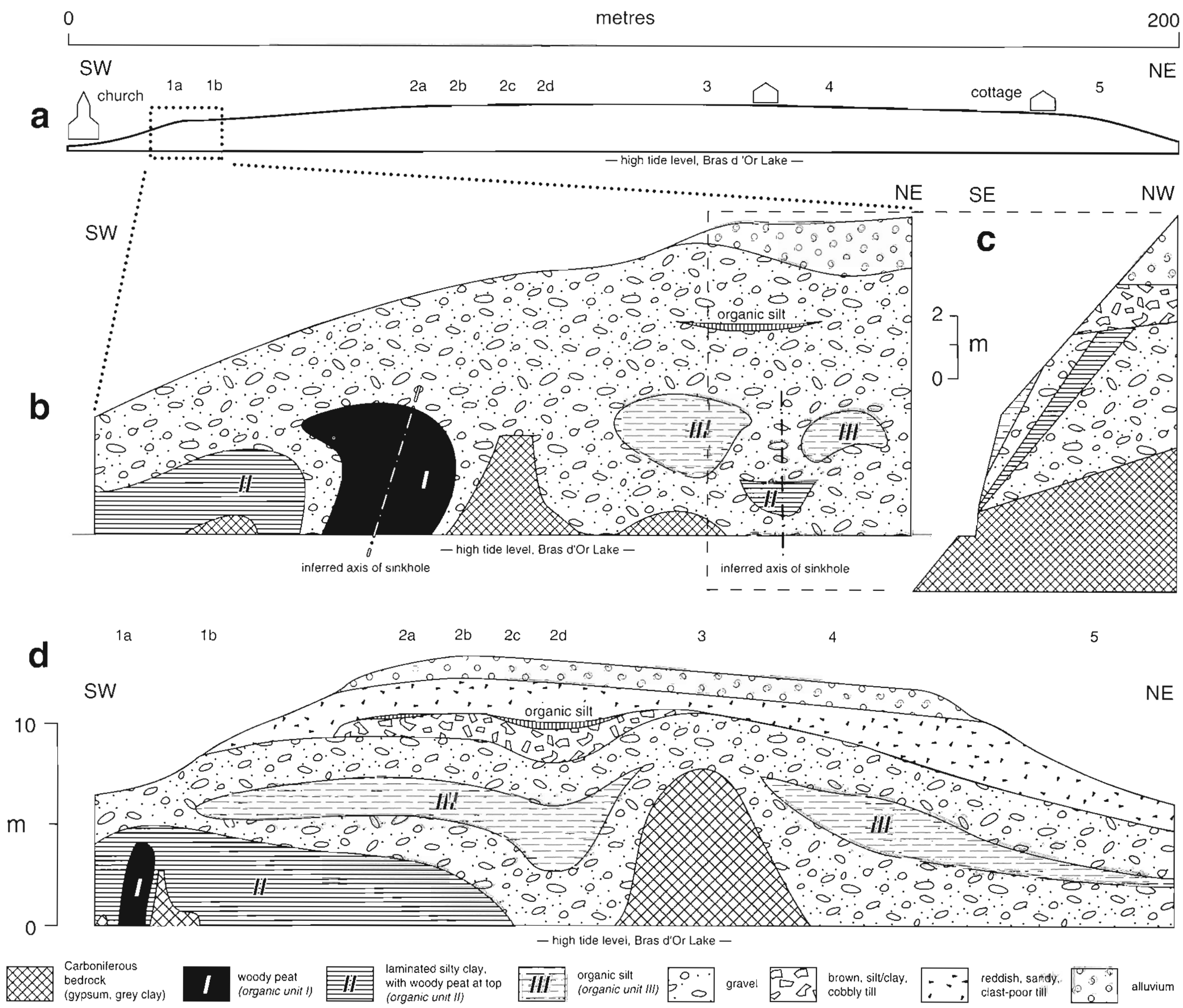



*Figure 89. Lithostratigraphy of the East Bay [27] interglacial organic site. **a.** Location of pollen profiles 1a-5 studied by de Vernal and Mott (1986) and de Vernal et al. (1986); **b.** Detail view of portion depicted by de Vernal et al. (1986) showing complexity due to karstification; **c.** Inferred structure perpendicular to coast at exposure 1b; **d.** Composite lithostratigraphic cross-section based on successive observations 1982-93 showing inferred continuity of various units (partly from Grant, 1987; with additional information from S. Occhietti).*

earlier accounts, because they serve to corroborate the interpretations arrived at earlier in this study and because they help establish a chronological framework for the island as a whole.

*General sequence*

As exposed along the 200 m long lake bluff (Fig. 89a), the stratification of the various sedimentary units is complex, discordant, locally greatly deformed (Fig. 89b), and changes yearly with erosion. In places, the lower organic horizons appear to be completely surrounded by gravel (Fig. 89b) where the pollen studies were mainly done, but stratification attitudes seen in gullies normal to the bluff (Fig. 89c) show that the members are conical-shaped with steep inclinations at various angles to the sloping face of the exposure and that the axes of these masses and the thickest accumulations are located about midway between steep gypsum knobs. It is thus inferred that the sediments accumulated in developing sinkholes and were further deformed by differential compaction from continuing karstic development and/or from subsequent glacial loading. Despite the deformation, seven fair to good exposures make it possible to outline and trace the continuity of three major organic members and to establish their stratigraphic relations to the other units (Fig. 89d). The deposits rest on a highly irregular surface of shale and gypsum bedrock which locally appears to be smoothly truncated as if by wave action 3-4 m above high tide level. Most of the sequence is a body of coarse, angular, muddy gravel composed mainly of crystalline rocks from the adjacent Boisdale Hills. It is vaguely stratified; most of the original structure was obliterated by settlement into the karst. It is blue-grey in the lower two thirds owing to the reducing effect of the enclosed organics. The top third is oxidized yellow-brown and retains a good stratification that is picked out by interbedded silt lenses. Crosscutting relationships of the gravel and included organics in the sinkholes shows that the infilling was interrupted by two main periods of collapse. The origin of the gravel is unclear but its generally conformable bedding with all three organic units and general parallelism and proximity to present sea level suggests a floodplain or beach plain environment. If it is littoral, the sequence shows that a marine transgression followed each organic unit.

The gravel/organic complex is overlain by two tills, locally separated by a thin silt layer; the lower is compact and has abundant cobble-sized crystalline rock stones in a dark reddish-brown silt/clay matrix; the upper is less compact and has only a few pebble-sized stones in a bright reddish-brown sandy matrix. The lower till locally contains much wood reworked from the peaty layers and both tills contain marine shell fragments. The tills have a northeastward fabric which is visibly indicated by the alignment of stones exposed by the floors of gullies cut into the till (although no pebble counts were made to fix the trend). The sequence is capped by postglacial fluvial gravel which crosscuts all the layers.

*Organic beds*

The East Bay sequence contains three organic units: woody peat (unit I), laminated silty clay (unit II), and organic silt (unit III), each of which has a distinctive pollen assemblage or zone (Fig. 90) which represents different vegetation/climate associations. Thorium/Uranium age estimates have been made on each unit. Even if the dates are reliable – and this depends on the basic assumption that the isotopic system remained closed (de Vernal and Mott, 1986; de Vernal et al., 1986) – they are minimum ages. On this basis, the dated beds seem to belong to three widely separated intervals, an inference consistent with the pollen results. It must be noted, however, that Stea et al. (1992) have challenged the basic assumption and rejected the dates. For the purposes of this paper, however, the ages are accepted as given because they seem to fulfill the requirements of stratigraphy and do not violate the palynological findings. Unit I is a compact, woody peat about 4 m thick with silt seams that occurs only on the side of one solution cavity that extends below low tide level. Its bedding has been tilted to vertical as the mass gradually settled into the deepening sinkhole. The peat contains beetle fragments, cones of *Pinus* strobus, and wood that gave a Th/U age of 126 400 ± 15 000 BP (UQT-175). The pollen assemblage is dominated by *Quercus*, *Pinus* strobus and thermophilous hardwoods, indicating a closed forest and a warmer and drier climate than the present. Most of the peat is therefore assigned to oxygen isotope substage 5e, when the warmest conditions of the last interglaciation prevailed about 120-130 ka, although it may well have extended to substage 5c (Mott, 1989).

Unit II is mainly laminated silty clay ranging up to 3 m thick, with interbedded sand and gravel. It grades upward to alternating silty clay and peat with woody layers (at one place including stumps and logs up to 40 cm thick and 3 m long). Wood of *Tsuga canadensis* in the upper part of silty clay gave a Th/U age of 86 900 ± 12 000 BP (UQT-109). Pollen assemblages at the base of this unit are dominated by *Abies balsamea* with some thermophilous hardwoods and *Tsuga*, representing a mixed forest and a cool wet climate possibly comparable to the present climate. Toward the top of this unit the mixed-forest assemblages give way to coniferous

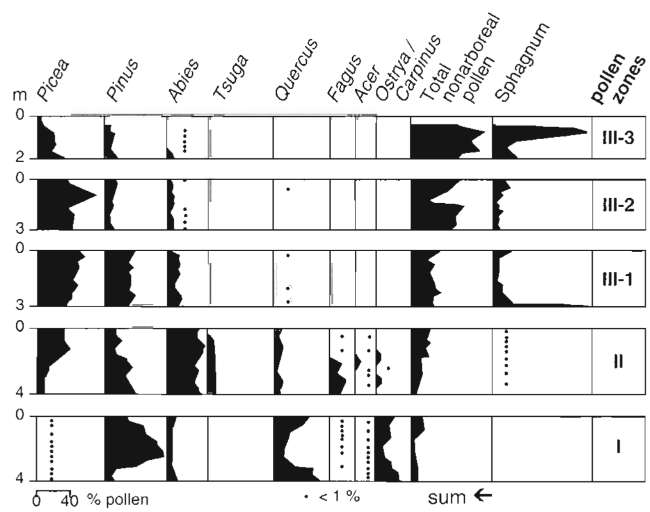



***Figure 90.*** *Pollen diagrams for the three major organic members at East Bay site [27] (from de Vernal et al., 1986, simplified by Mott, 1989).*

assemblages dominated by *Picea* which indicate an impoverishment of the forest and a cooling trend. On the basis of the date and the inferred climate, de Vernal et al. (1986) assigned this middle organic bed to oxygen isotope substage 5a, the last warm episode of the Sangamonian Stage.

Unit III is a thinly laminated silt 1-2 m thick in the upper part of the stratified gravel. Pollen is dominated by *Picea*, *Pinus*, and nonarboreal taxa. Variation in their proportions indicate alternation between boreal forest, open forest, and tundra. They reflect small amplitude changes of a wet climate, cooler than at present. It contains both freshwater and brackish water diatoms and dinoflagellates indicating aquatic sedimentation. Wood of *Picea sp.* yielded a Th/U age of 62 100 ± 5000 BP (UQT-177). De Vernal et al. (1986) accepted the age as given and, noting that the pollen spectra were similar to that of the Bay St. Lawrence peat which dated 44 200 ± 800 BP by $^{14}C$, they assigned unit III to the Middle Wisconsinan, specifically to oxygen isotope stage 3 (65-32 ka). Alternatively, if the age is minimal, the deposit may be more closely related in time to the underlying unit II, that is to say, to late stage 5 and early stage 4, the cooling phase during the onset of Wisconsinan glaciation.

*Depositional scenario*

The following sequence of events is suggested to explain the East Bay deposits. Organic units I and II, together with the enclosing gravels, were laid down adjacent to East Bay on a floodplain, delta, or beach plain that accumulated over an active karst area. Interludes of higher sea level may have occurred after each was laid down. The upper organic bed (unit III) was definitely laid down when the site became inundated. De Vernal and Mott (1986) postulated that the water body was a lake or large pond that experienced brackish-water input. If so, the elevated position of the waterbody would require that some external agent raise the level of East Bay and change it to freshwater. The present author therefore suggests that the inundation was that of a proglacial lake dammed by ice advancing up East Bay, since the upper bed is said to date from the cooling phase of the Wisconsin glaciation (stage 4, if not stage 3). The inclusion of marine microfossils in the sediments of that lake is explained as redeposited marine sediment ploughed up from the former sea floor. Such an impoundment would have been the natural consequence of the initial eastward glacial advance over the southern lowlands recorded by the striation sequence (Phase B). One problem with the proglacial lake scenario is that such a lake would be relatively shortlived, given the length of the bay and average glacier advance rates, whereas the pollen concentration calculations, if valid, suggest that aquatic conditions lasted several tens of thousands of years. Whether the deposits are marine or freshwater, their age, if reliable, provides an minimum estimate on the time of the first advance onto the Bras d'Or Lowlands from mainland Nova Scotia: 62 ka, if the age estimate on unit III is accepted as given; 87 ka if it is regarded as minimal and estimated to be closer in time to the age of unit II. Of the two tills overlying the organic beds, the lower may have been laid down either during the early eastward flow (Phase B) which, in this area, may have taken on a northeastward trend parallel to the orientation of East Bay, or by the later northeastward flow from the Scotian Shelf (Phase D). If so, the silt bed between it and the surface till may represent the nonglacial interval represented elsewhere by the subaerial weathering on the early striations (Phase C-D interval). The surface till would then be the product of the final radial flow of the remnant ice cap over Bras d'Or Lowlands, which in this area was northeastward toward the Sydney Shelf (phases E, G).

## Castle Bay [25]

Located on the north side of East Bay, Bras d'Or Lake (Fig. 30), this sequence forms the shore of Castle Bay, which is named for the earth pillars that have developed in the deposits (Fig. 91). First noted by the author in 1971 (Grant, 1972), the exposure, which cuts obliquely across the long axis of a 1500 m long drumlin (Fig. 92a), measures 1300 m long by up to 32 m above tide level. The sequence consists of the following, mainly stratified, deposits (Fig. 92b): basal angular pebble gravel, a thin gyttja horizon at the base of a thick organic silt member which encloses a lens of sand and gravel, two tills, and a thin organic silt at the top under till or colluvium (Fig. 92b). Like the other sequences in the area, the Castle Bay exposure shows that the northeast-trending drumlins along the East Bay coast (Fig. 43) are carved from a mainly stratified sequence of deposits and are only thinly veneered with till.

*Basal angular pebble gravel*

At the base is a bed of poorly sorted, horizontally-bedded, angular, pebble-sized rock fragments that is generally fairly clean, but locally has a matrix of muddy grit. The angular fragments resemble frost-broken debris and consist almost wholly of crystalline rock debris from the adjacent Boisdale Hills, chiefly a greenish Cambrian granite with partly kaolinized feldspars. The material is oxidized and the feldspars are altered. It is not indurated, but so compact and tightly packed as to support almost vertical cliffs, perhaps because of glacial overconsolidation and the interstitial fines. The top of the gravel is generally smooth and planar where it is in contact with the overlying organic member, but is locally deformed where overlain by till at the east end of the section. The contact slopes up to the east from below tide level to an almost level plane at 3 m elevation in the middle part of the section. It is inferred that the basal gravel relates to a former higher level of Bras d'Or Lake because of its planar stratification, subhorizontal upper level and gradational contact with waterlaid sediment, and similar elevation where it outcrops in a road cut on the north side of the drumlin (G in Fig. 92a) and elsewhere beneath similar sequences along East Bay shore. The gravel may therefore be a littoral deposit, or perhaps the distal part of a piedmont fluvial apron that was deposited marginal to, and partly in, a standing body of water. The top of the gravel is nearly accordant with the rock bench which outcrops at about 4 m above tide level on nearby Benacadie Point [19] and the true maximum elevation of both is undoubtedly higher as the inner limit is not exposed. Thus, if the gravel and inferred associated bench in East Bay can be correlated on the basis of their horizontality and nearly identical elevation with the emerged coastal bench that is assigned to the Sangamonian interglacial sea level maximum (oxygen isotope substage 5e), then the rubbly gravel underlying the organic beds along East Bay is a deposit of that elevated interglacial sea level. All of the overlying deposits thus postdate the marine transgressive maximum of the last interglaciation.

*Sand and gravel body with enclosing organic sediments*

The middle portion of the sequence consists of a large lenticular body of sand and gravel up to 10 m thick enclosed in organic sediment that thickens from 2 m over the crest of the sand/gravel body to 15 m thick on its flanks. The sand, which ranges in thickness from a few decimetres to 8 m thick, is composed of clean, subrounded quartz and feldspar in well stratified, crosscutting sets of long, planar beds. As seen in a large gully which provides a three-dimensional view at right angles to the bluff, the sand beds dip south toward the lake, decreasing from a maximum of ≈40° at their upper end to horizontal where they merge asymptotically with the underlying gyttja. The sand is sharply overlain by a 2-3 m bed of pebble/cobble gravel that has good horizontal to gently inclined stratification and is composed of local crystalline rocks. The form and structure of the sand and gravel body is typical of a delta.

The enclosing organic sediment, which ranges up to 15 m thick, has two main parts. The lower part, a stratified, gritty gyttja 1-3 m thick, lies beneath the sand and extends laterally over the underlying gravel. The gyttja yielded an age of >42 000 BP (GSC-1577). The upper part, which is draped over the sand and gravel and merges laterally with the underlying gyttja, is a compact, stony, slightly organic mud. It is massive, except for rare sandy partings, and has columnar jointing with iron-manganese stain. A small wood fragment yielded an age of >52 000 BP (GSC-1619). Where overlain by till at the east end of the section, the silt is locally interfolded with the underlying gravel. The silt was originally thought to be till (Grant, 1972) because it contains striated stones and appeared massive. Subsequently, however, its waterlaid origin was demonstrated by Lortie et al. (1984), who noticed vague layers which contained varying proportions of freshwater, brackish-water, and marine diatoms. De Vernal and Mott (1986) later documented a matching variation in pollen assemblages, as outlined below.

The complete 15 m thickness of the organic unit lateral to the sand/gravel body, as well as the part beneath and on top, was studied for pollen and other microfossils. Four vegetation assemblages were revealed (de Vernal and Mott, 1986) (Fig. 93). The first, represented by the sandy gyttja at the base, has a macrofossil content typical of lake-bottom mud and a pollen content dominated by *Picea* and *Pinus*, indicative of boreal forest conditions. The overlying silt which forms the bulk of the organic part of the section, has three intervals dominated by nonarboreal pollen indicative of tundra conditions, separated by two intervals of coniferous pollen representing boreal forest vegetation.

The nature of the water body in which the silt accumulated is unknown, yet crucial to the paleogeographic reconstruction. From the 20-30 m height of the sand/gravel/laminated silt sequence and the vertical variation of microfossil salinity indicators, Lortie et al. (1984) and de Vernal and Mott (1986) postulated that glacio-isostatic depression by an approaching ice sheet caused a marine inundation in which there was a fluctuating fluvial input. They further estimated that the submergence lasted 14 000-52 000 years, based on the thickness, pollen concentration, and number of vegetation/climate cycles. This scenario presents two main problems which call for an alternative explanation. Firstly, although marine deposits do occur up to 22 m elevation in neighbouring

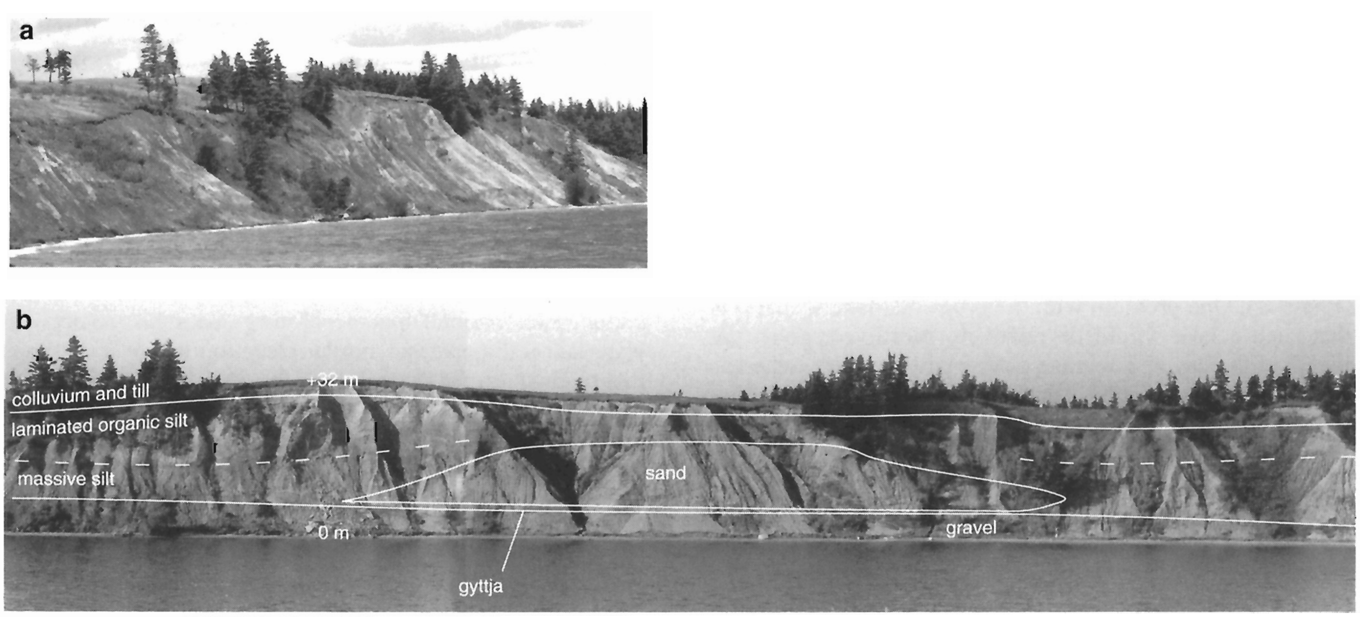



***Figure 91.*** *Views of the interglacial organic site at Castle Bay [25] with contacts of major visible units indicated. GSC 204042-O and GSC 1994-414A, B, C*

Newfoundland (Brookes et al., 1982) and les Îles de la Madeleine (Dredge et al., 1992), there is no independent evidence anywhere in Cape Breton Island of deposits or shorelines indicating that sea level rose to an elevation of 20-30 m during interglacial time (or any subsequent time for that matter). Secondly, the isostatic depression mechanism is inapplicable because crustal deflection normally lags the load application by several thousand years with the result that, during advance, sea level is usually low relative to present level, rather than high. A local inundation, as by a glacial lake would avoid the need for abnormal isostasy and would explain all the features of the deposits. It is therefore suggested that the stratified silt and associated delta beds were laid down in an ice-dammed lake created when a glacier blocked the mouth of the bay. As it advanced, the marine water freshened and became a lake and the underlying submarine mud was ploughed up and redeposited. The vagaries of the process could account for the varying proportions of marine and freshwater microfossils. Such a lake would be the natural consequence of any one of the three main ice-flow trends which moved toward the head of the bay: the early eastward advance (Phase B), the intermediate-age north-northeastward advance (Phase D), and the final northeastward advance (Phase G). The first of the three is the most likely as it would explain the shells in the lowest till and would allow the overlying tills to be assigned to the later phases. Still, as at East Bay, the proglacial lake scenario presents a problem in that such a lake would be relatively shortlived, whereas the pollen work suggests that aquatic sedimentation lasted tens of thousands of years.

*Overlying tills and diamicton*

Two tills and an unexplained surficial diamicton overlie the sand/gravel and organic beds. The tills are thin on the western or stoss flank of the drumlinoid mass, patchy to absent on the crest, and several metres thick on the eastern or lee flank. The lower till is compact, dark purplish brown, and silty; its pebbles and cobbles are a mixture of local sedimentary rocks

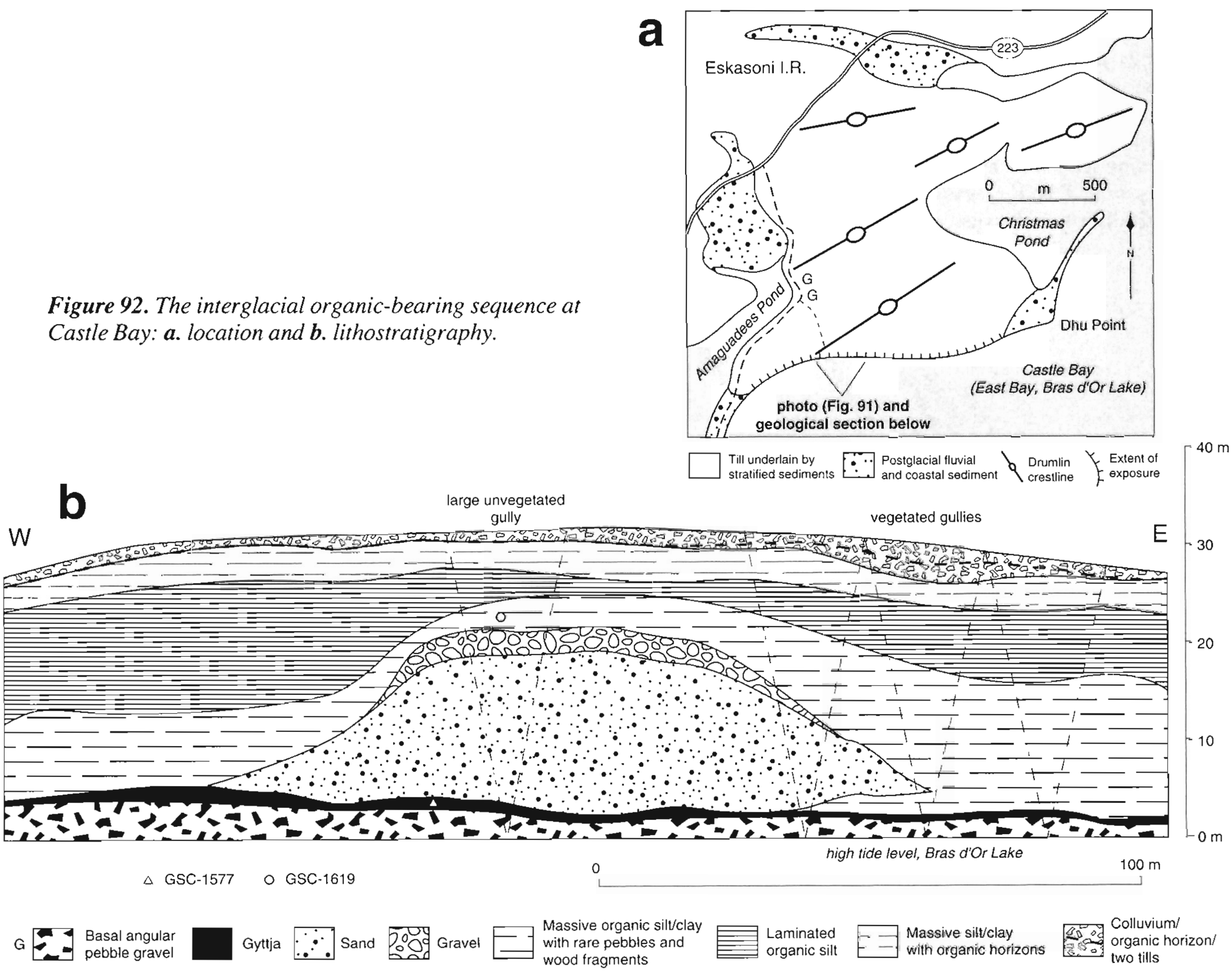



***Figure 92.*** *The interglacial organic-bearing sequence at Castle Bay:* ***a.*** *location and* ***b.*** *lithostratigraphy.*

from the surrounding lowlands, together with crystalline rocks from the neighbouring uplands. The upper till is bright reddish brown, silty, and derived almost wholly from local Carboniferous redbeds which underlie Bras d'Or basin and much of the adjacent lowlands. Locally, the surficial part of the upper till has been reworked by solifluction into a colluvial mantle which is recognized by its platy structure, unconsolidation, and darker, more iron-enriched colour from oxidation. In one place, in a slight depression near the top of the drumlin, the soliflucted till (colluvium) overlies a thin bed of organic silt which yielded ages of 13 030 ± 1270 BP (UQ-246) and 11 100 ± 170 BP (UQ-831) (S. Occhietti, pers. comm., 1987).

*Summary depositional scenario*

The following sequence of events is based on the stratigraphy at this locality and takes account of the findings at other nearby localities. The flat top of the basal gravel and its similar elevation to that of the emerged rock bench suggests that the gravel may also relate to the last interglacial sea level highstand. Possibly its weathered condition was acquired when it later became subaerially exposed, probably during the regression as global sea level fell because of growing Wisconsinan ice sheets. The overlying organic beds and associated delta were then laid down as the level of Bras d'Or Lake rose, either because of isostatic depression due to an approaching ice sheet, as postulated by de Vernal and Mott (1986) or when East Bay was converted into a proglacial lake by the initial eastward invasion of ice onto Cape Breton Island

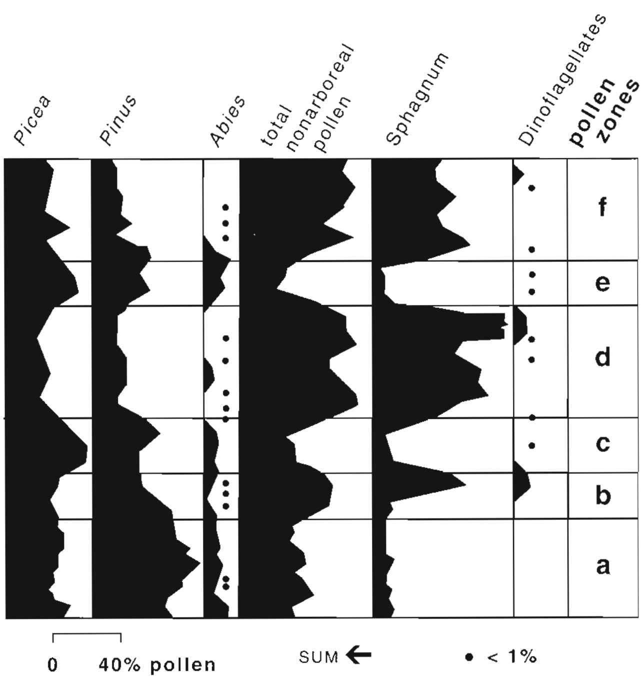



*Figure 93. Pollen diagram for Castle Bay organic sequence based on several profiles from west of the lenticular sand body shown in Figures 91 and 92; (simplified by Mott, 1989 from detailed profiles in de Vernal et al., 1986)*

(Phase B), as favoured in this report. The overlying organic beds and tills are thus considered to date from the Wisconsinan Stage because of the cool-climate pollen assemblages. De Vernal and Mott (1986) estimated that the organic sequence spans an interval of 14 000-52 000 years based on its thickness, the pollen concentration, and the number of vegetation/climate cycles. They correlate it with East Bay unit III which gave an age of 62 ka. If so, the Castle Bay sequence begins in Sangamonian time (sensu lato) and the organic portion may span most of oxygen isotope stage 4 and much of stage 3. The overlying tills are therefore largely limited to oxygen isotope stage 2 and are therefore Late Wisconsinan in age. By this scenario, the nonglacial conditions which began in Sangamonian time lasted well into Wisconsinan time, and glaciers did not enter lowland Cape Breton Island until Middle Wisconsinan time or later. On the other hand, if the submergence phase was a shortlived proglacial lake related to the lower of the two overlying tills, then the nonglacial period represented by the 20 m of organic silt was much shorter and was therefore probably confined to early stage 4. If so, the first major glaciation of lowland Cape Breton Island (Phase B) was Early Wisconsinan. This alternative would accord better with the evidence from the weathered striations which suggest there was an ice-free interval after the first two major glacial phases and before the last major overriding of the island.

**Benacadie shore 3 [21]**

Near the mouth of East Bay, 2 km east of Benacadie Harbour (Fig. 30), shore bluffs expose the first reported occurrence of subtill silt in the East Bay area (Mott and Prest, 1967; Fig. 94). (The sequence should not be confused with nearby occurrences on this shore [20], [22], [26] of subtill organics that are of Late Wisconsinan age). The same oxidized, muddy, angular, granitic pebble gravel as at Castle Bay [25] extends to 3-4 m above tide level and has a horizontal top over a few hundred metres. The gravel is sharply overlain by 1-3 m of grey organic silt with a pollen assemblage resembling that in the Early (and Middle?) Wisconsinan part of the Castle Bay sequence. No macrofossils indicative of aquatic conditions were observed. The silt is overlain by dark olive-brown, cobbly clay till with rare small fragments of temperate water marine shells (e.g., *Mercenaria mercenaria*) like those occurring in the area today. The surface till, on the other hand, contains no shells, is stony, and has a bright brownish red clay/silt matrix. The sequence thus contains four of the main units present at Castle Bay [25] and East Bay [27] sites and helps demonstrate their wide extent.

Given the stratigraphic similarity to other East Bay exposures, depositional events probably followed the same inferred sequence. The rubble and conformably overlying waterlaid silt evidently represent an aquatic episode which reached to about 6 m above present lake level, or almost to the same level as the emerged littoral bench. As there are no microfossils to judge whether that water body may have been marine or freshwater, it therefore may have been either the last interglacial sea or an ice-dammed lake related to deposition of the lower of the two overlying tills. The lower till forms east-trending drumlins and was therefore probably laid down by the early eastward flow (Phase B) which advanced up East Bay and incorporated marine fossils from the sea floor. The upper till may represent any one of the following flow phases, but most likely the radial flow of a remnant lowland ice cap (phases E, G).

It may be noted in connection with this locality that the same two-till sequence occurs 2 km to the east at the site informally designated Amaguadees head [24]. There, the lower till forms an east-northeastward-trending drumlin and the red surface till forms a thin mantle which mimics the form. It is unknown whether the drumlin was formed when the lower till was deposited (Phase B) or was carved into it by a later flow (phases E, G).

### Hillsboro [59]

On the western lowlands, on the inner end of Mabou River estuary (Fig. 30), a roadcut on Highway 252 cuts through high-relief gypsum karst and exposes a 10 m thick filling of a sinkhole consisting of a complex and locally variable sequence which features woody organic sediments beneath till. Discovered in 1952 by L.R. Wilson, while touring the area with R.F. Flint, Yale University, the Hillsboro (former official spelling: Hillsborough) deposit is notable in that it was apparently the second subtill organic site to be found in the Atlantic Provinces region (Prest, 1957, p. 446). As such, it attracted the attention of two major radiocarbon labs, then in their developmental stages, and was thus among the first of such deposits to yield ages beyond the limit of the $^{14}C$ method: >21 000 BP (Y-232) and >38 000 BP (W-157) (Flint, 1956; Flint and Rubin, 1955). The first pollen analysis of the section was by Mott and Prest (1967) and, on the basis of several large temporary artificial excavations by S. Occhietti and R.J. Mott in 1984, an updated pollen profile by R.J. Mott was reported by Grant (1987). However, study is still incomplete because the stratigraphic succession is uncertain owing to deformation by karst development and because proximity to the highway precludes the necessary excavation. The general sequence as presently understood seems to be: ≈1 m gravel and sand with yellow and orange oxidation, overlain by ≈2 m of laminated but contorted grey silt and clay with a layer of coarse wood fragments and minor carbonaceous seams. The organic member is overlain by sand and a thin colluvial layer of slumped till and is topped by a red clayey till of mixed upland and lowland lithologies which resembles the lower till overlying the organics at nearby Green Point [57].

The largest wood samples were *Abies balsamea* and *Picea sp.*; the former yielded an age of >51 000 BP (GSC-370). The pollen sequence (Fig. 95) begins with spectra dominated by *Abies balsamea*, *Alnus rugosa*, *Betula*, and Polypodiacea and changes to spectra dominated by *Picea*, Cyperaceae, and *Sphagnum*. This represents a shift from mixed boreal forest with abundant balsam fir to a spruce forest and indicates a climatic cooling. The pollen sequence resembles other Nova Scotian deposits, particularly the upper part of the East

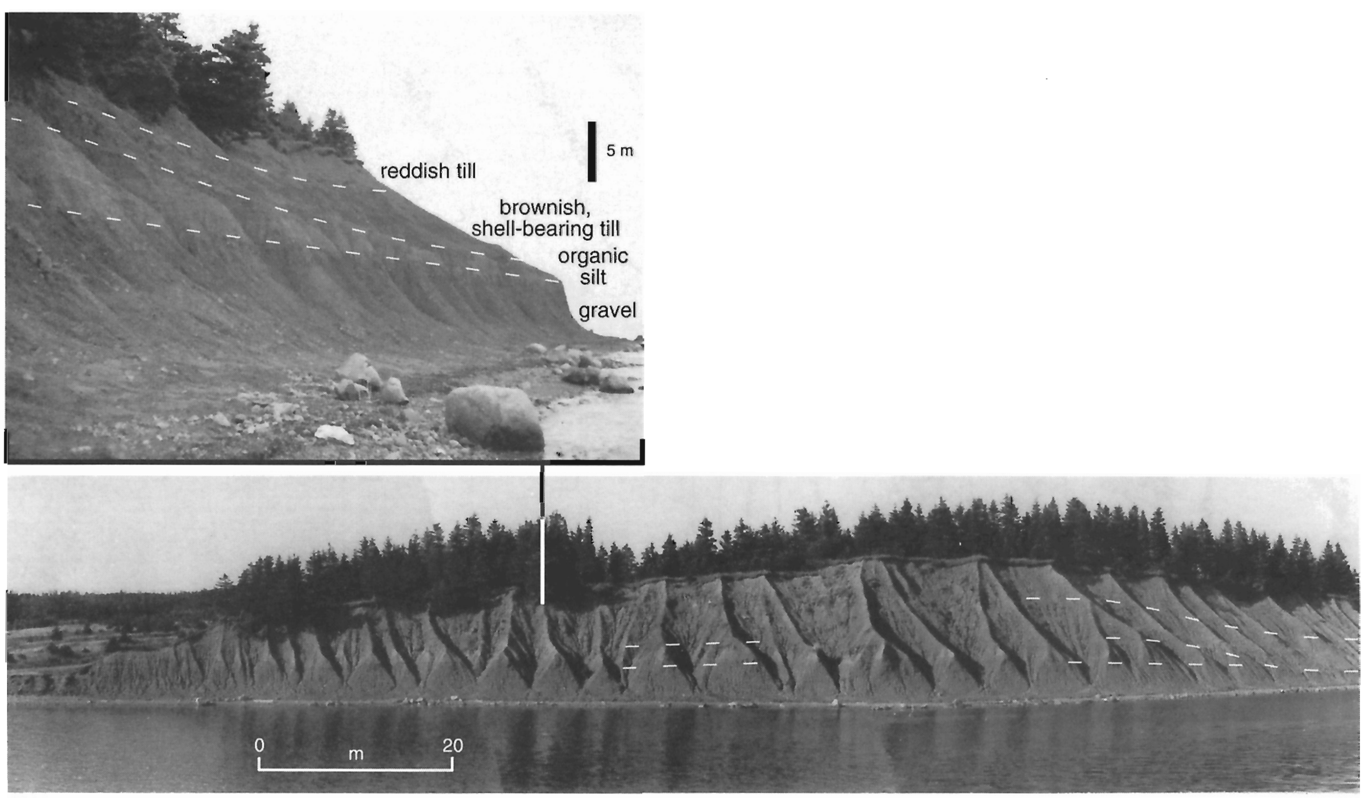



***Figure 94.*** *Views of Benacadie shore interglacial organic site [21], East Bay; onshore view shows texture of units; GSC 204042-E; composite view from offshore with main contacts indicated; GSC 1994-415A, B, C*

Milford deposit (Mott et al., 1982), so the Hillsboro organic member is assigned to the late Sangamonian, specifically to the cooler parts of oxygen isotope substage 5a (R.J. Mott, pers. comm., 1987).

## Big Brook quarry [49]

In the upper reaches of Big Brook (Fig. 30), a tributary of River Denys, quarrying operations of the Georgia-Pacific Gypsum Company have produced good exposures over a broad area of several tills covering high-relief gypsum karst. In 1980 the author discovered three organic layers of different age and climatic affiliation that are interbedded with the tills. Glacial thrusting has complicated the succession, however, necessitating repeated examination over several years to determine the following general succession, which still must be considered provisional.

*Lower organic bed*

The gypsum surface varies from smoothly undulating to highly karstic with deep steep-sided pits. Solution cavities in one such place contain grey organic silt that yielded a finite age of 36 200 ± 1280 BP (GSC-3206). The date, however, is rejected as spurious and minimal because the organic content was low, the possibility of carbonate contamination high, and the age was near the limit of the method. Pollen analysis, courtesy of R.J. Mott (Fig. 96, pers. comm., 1987), provides a more definitive indication of the likely age; his findings and interpretation are outlined below. The spectrum contains entirely cool-climate boreal assemblages, dominated in the lowermost part by *Salix* and Cyperaceae, and then throughout most of the section by abundant *Alnus*, *Picea*, *Pinus*, and Polypodiaceae. The most complete sequence shows a climatic deterioration from forest to forest tundra. The pollen assemblage correlates with other sites which have supplied Th/U age estimates suggesting they are late Sangamonian (oxygen isotope substage 5a).

*Lower till and rubble*

Overlying the organic silt and the gypsum bedrock is a widespread grey calcareous till, 2-4 m thick, with a typical mixed upland and lowland lithology in a matrix of comminuted gypsum. The upper 23 m is light yellow due to accumulation of interstitial iron hydroxides. If this alteration is the result of subaerial weathering, it must have occurred prior to emplacement of the overlying unaltered deposits. In one place, several metres of compact, grey, muddy gypsum rubble truncate the lower till along a steep contact which merges with a steep slope on the underlying gypsum. The rubble has a faint stratification parallel to this contact, so the deposit is taken to represent a sinkhole filling, but whether the material is slumped residuum or till is not clear.

*Middle organic bed*

A second organic silt beds intervenes between the rubble and the overlying three unweathered tills. It thickens toward the inferred centre of the sinkhole and pinches out at its margin,

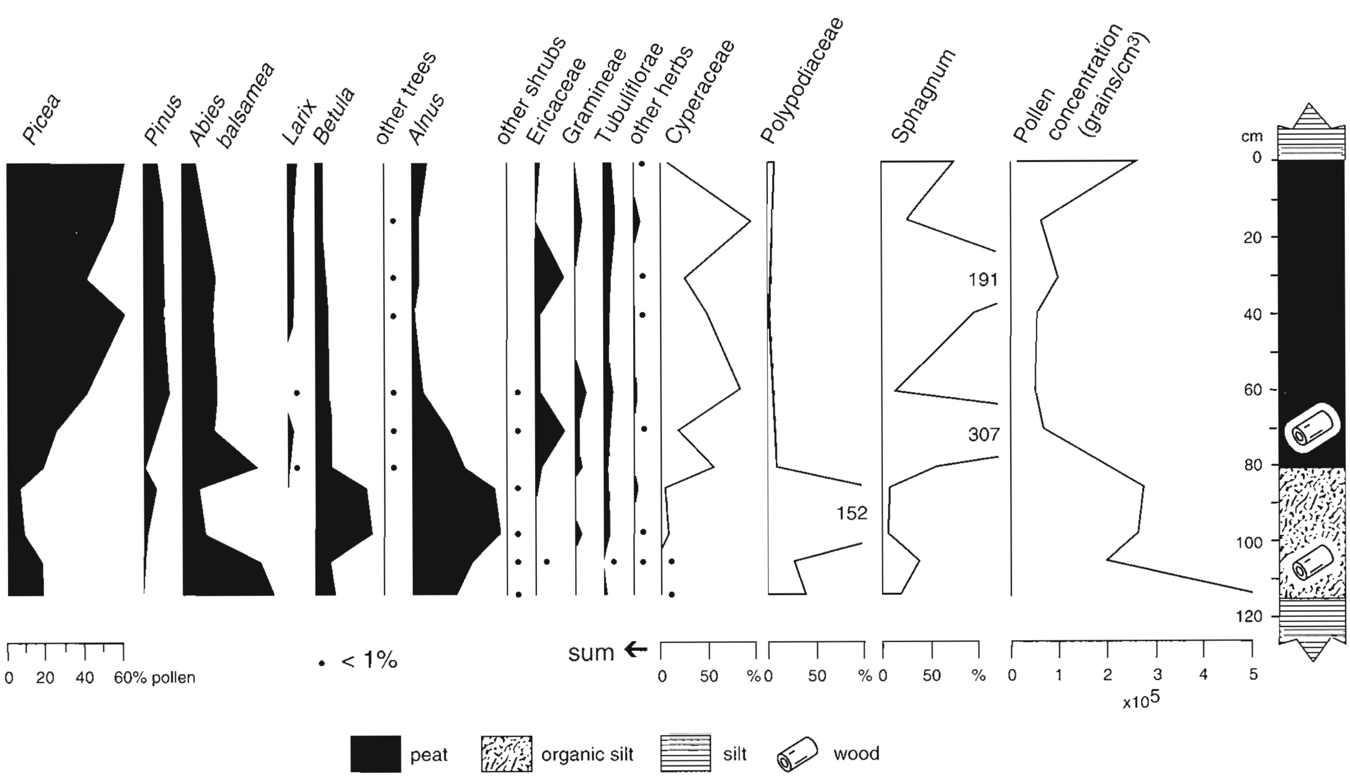



*Figure 95. Simplified pollen diagram for Hillsboro interglacial organic site [59] (courtesy, R.J. Mott, in Grant, 1987, p. 31)*

and thus appears to reflect continual deepening of the depression. The material was unsuitable for dating, but pollen analysis shows that it differs markedly from the lower silt in having about half as much *Pinus* and three times as much *Picea*. It thus also represents boreal forest conditions, but under a distinctly cooler climate. R.J. Mott (pers. comm., 1984) noted its resemblance to the Dingwall [4] subtill organic bed which, in turn, he correlates with the latter part of the East Bay sequence, which has been Th/U dated to the period ranging from substage 5a to possibly as young as oxygen isotope stage 3. The middle organic bed could thus represent either the waning stages of the last interglaciation as the climate cooled during the onset of glaciation, or an intra-Wisconsinan nonglacial period after deposition of the lower till and before the emplacement of the overlying tills. The latter alternative is favoured because the lower till has been weathered, presumably during the same nonglacial period.

*Middle three tills*

Discounting the repetitions of members caused by glacial thrusting, the middle organic bed and residuum are overlain by three tills of different fabric (R.R. Stea, pers. comm., 1981), colour, and texture whose combined thickness ranges from 5-20 m. All have generally the same lithology – a mixture of the main local sandstone and shale sedimentary rocks, together with a great variety of crystalline stones from the surrounding uplands; the differences are in the colour and texture of the matrix. The lowermost is dusky purplish brown, clayey, with an eastward fabric. It contains masses of peat and wood fragments which have been sheared into it, probably from a source related to one of the underlying organic beds. Two such wood samples yielded ages of >49 000 BP (GSC-3289) and >52 000 BP (GSC-3880). It is sharply overlain by a middle till that is greyish red and silty, with a northward fabric. The middle till grades into an overlying greyish brown surface till with a southeastward fabric. The sequence of till emplacement directions agrees with the east/north/south ice-flow sequence (phases B, D, E) determined from crosscutting striations.

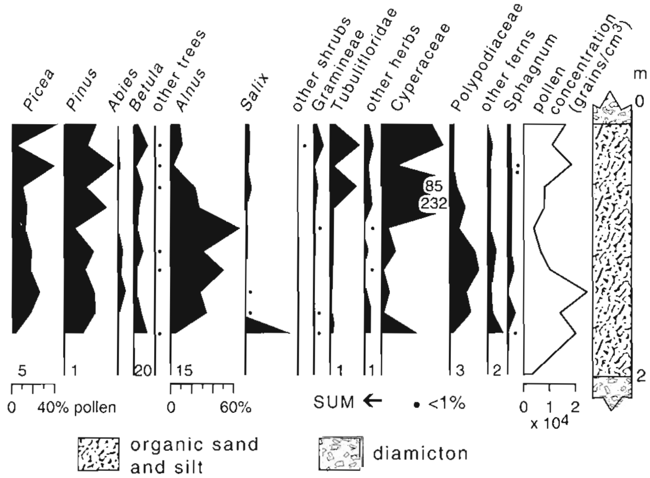



***Figure 96.** Simplified pollen diagram for Big Brook quarry organic site [49] (courtesy, R.J. Mott, in Grant, 1987, p. 30)*

*Upper organic bed and surficial stony silt*

The top two units overlying the upper till are a 1 m surficial veneer of laminated silt with sand partings which has at its base in one place a 20 cm organic seam that yielded an age of 11 000 ± 90 BP (GSC-3778). According to Mott et al. (1986), the pollen spectrum of this organic unit is dominated by sedge, ferns, and club moss, with some shrub birch, willow, and a variety of herbs which represent tundra conditions. On the basis of its similar age and pollen signature, they correlated it with numerous deposits throughout the region and attributed the group to a sharp climatic reversal after the postglacial climatic warming began. Although no specific microfossil indicators were found in it, the surficial stony, laminated silt is evidently an aquatic deposit. It is likely the product of the small proglacial lake which is inferred to have occupied much of the River Denys Lowland, on the basis of the hanging deltas described in the section on Glaciolacustrine deposits and lake-level indicators. That proglacial lake deposits overlie 11 000 year old organics indicates that a sizable glacier remnant persisted in Bras d'Or Lowlands during the Younger Dryas cooling episode. Its supposed re-expansion (Phase G) is correlated with the small readvance in nearby mainland Nova Scotia (Stea and Mott, 1989), which they attributed to the severe climatic deterioration that affected the region (Mott et al., 1986).

**Leitches Creek [32]**

In July 1966, a 60 m deep mineral exploration boring by New Jersey Zinc Corporation near Leitches Creek, west of Sydney (Fig. 30) penetrated two organic layers in a till sequence. According to the company geologist (E.A. Goranson, pers. comm., 1966) the hole was situated at 60 m elevation and penetrated 15 m of "overburden" (the surface till in the area), 3.66 m of highly compressed grey-brown silt with lamination dipping 40° and beds of well-preserved wood fragments, a further 9 m of material said to be similar to the surface till, and at 43 m depth, another bed of well-preserved wood. The bottom 7 m was generally a sand/gravel mixture, with intervals of grey clay and scattered boulders of granite, limestone, and gypsum. (It may be noted that the elevation, texture and lithology of the lower units resembles that at East Bay [27], although the similarity may be fortuitous). The cored intervals of organic matter were donated to GSC where they remain in storage. Because this was a rare case of two organic beds separated by till and because the original core was in poor condition and showed internal disturbance, a second drill hole was put down in the 1980s as close as possible to the same site through a co-operative agreement with R.R. Stea (Nova Scotia Department of Natural Resources). The same sequence was found, although the character of the organic horizons was slightly different and their depths somewhat less.

The organic intervals in both cores were analyzed for pollen by R.J. Mott (pers. comm., 1971, 1987). The lower organic unit contains mainly pollen of *Picea*, *Pinus*, *Abies*, *Betula*, and *Alnus*, plus significant amounts of other deciduous pollen genera, such as *Carpinus/Ostrya* type, *Tilia*, *Acer*, and *Corylus*. These indicate a distinctly warmer climate than present. The upper organic interval in the more recent core is seen to consist of distorted and highly compressed silty peat, associated organic silt, and sand with some pebbles. The peat yielded an age of >52 000 BP (GSC-2678). Its pollen content contrasts with that in the lower organic silt unit in that it is dominated by *Picea*, *Pinus*, *Abies*, *Betula*, and *Alnus*, representative of typical boreal forest conditions. Comparing these results with better dated deposits elsewhere on Cape Breton Island, Mott likened the lower organic unit to the early warm interval of the East Bay [27] interglacial sequence and the upper to the cooler-climate unit of the late interglacial sequence at East Bay [27], Bay St. Lawrence [1], Whycocomagh [17], River Inhabitants [48], and Hillsboro [59]. Both organic layers at Leitches Creek are thus assigned to the Sangamonian and, if correct, the supposed till between the two organic intervals would represent a glacial event during interglacial time, a unique occurrence in Cape Breton Island; however, the diamicton between the two organic units is less like the compact surface till and more like the loose gravelly material in the lower part of the core, so it may not be till at all. Further, considering that the organics are highly deformed raises the possibility of repetition of units by glaciotectonic stacking (as at Big Brook quarry [49]), so the included diamicton may not be in sequence. Further work on the cores was suspended because of this uncertainty about the stratigraphic integrity, so the question of possible glacial action within the last interglaciation could not be resolved during the course of this study.

### Northeast Mabou River [58]

In 1981 the author, in the company R.J. Mott and V.K. Prest, discovered coarse woody peat overlain by gravel exposed in a shallow roadside ditch ≈3 m above tide level near the mouth of Northeast Mabou River (Fig. 30). In 1984, R.J. Mott obtained a more complete section for study by excavating a 4 m deep trench in the floor of an adjacent gravel pit. The excavated part of the sequence consists mainly of iron-cemented muddy gravel near the base (like that at nearby Green Point [57]; see above), overlain by ≈2 m grey silty clay, which contains marine diatoms, and a bed of coarse woody peat near the top. The gravel and the peat interfinger with the clay and decrease in thickness toward the river. The many large wood fragments and tree trunks in the peat included *Picea sp.*, *Larix*, and *Abies balsamea*, the last of which dated >53 000 BP (GSC-3317). A pollen profile by R.J. Mott (pers. comm., 1987) through the 4 m section revealed an assemblage typical of a cool boreal climate, with *Abies*, *Betula*, and Cyperaceae increasing upward. With reference to other dated subtill organic beds on the island, notably at East Bay [27] unit II, Mott (1989) correlated the Mabou River deposit with the cooling phase of oxygen isotope stage 5c. If so, it is the only deposit specifically referred to that period. From the fossil content, the deposit may represent a tidal salt marsh deposit with some fluvial input of wood detritus. If so, it is the only example of last-interglacial, fossiliferous marine sediment on Cape Breton Island.

Abruptly overlying the organic part of the sequence (i.e., above the level of the road) is 1015 m of poorly stratified, muddy gravel composed almost entirely of upland crystalline rocks. The gravel forms an outwash terrace that extends ≈1 km inland and slopes up from 11 m at the coast to 51 m where it ends at a kettled ice-proximal scarp, which evidently marks the terminus of an ice lobe emanating from an inland source, probably related to glacial flow Phase E. Clearly, if the organics are last interglacial in age and the outwash is the latest glacial deposit, there is a large depositional hiatus between the two.

A bed of massive red stony silt about a half metre thick occurs at a depth of 11.5 m in the gravel and its top and bottom contacts are planar and parallel to the gravel stratification. From this it would appear to be waterlaid and one of the latest of the late glacial deposits on the west coast. It appears identical in colour, texture, structure, and stratigraphic position to similarly thin beds of a problematical massive stony silt that occurs on top of or in the upper layers of various other gravel deposits on the west coast of the island. The material is considered to be a mudflow deposit related to proglacial lake discharge when offshore (Gulf of St. Lawrence) ice impinged on the west coast (Phase F), as discussed in the section on Glaciolacustrine deposits and lake-level indicators.

### Dingwall [4]

First noted by R.B. Taylor during a coastal study in 1980, a 10 m high coastal cliff on the shore of Aspy Bay between North Pond and South Pond (Fig. 30, 70) exposes three tills and various nonglacial sediments, some richly organic, which accumulated in a sinkhole in Carboniferous shale and gypsum bedrock (Fig. 97) (Grant, 1987). The sinkhole continued to deepen while the sediments were accumulating, as shown by the downward-increasing concavity of stratification of all units and by the depression which has formed on the surface; however, the deepening has caused the organic units which accumulated in the sinkhole to move downward relative to the other units, so that they now have a penetrative relationship, and the apparent stratigraphic positions may not be as they originally were.

At the base, a brown sandy till is overlain by grey pebbly organic silt with interbedded gyttja and peat which contains fragments of spruce and/or tamarack and willow wood. Pollen spectra, as determined by R.J. Mott (pers. comm., 1987) are dominated by *Picea* and *Alnus crispa*; *Abies balsamea* is a lesser component, and a variety of shrubs and herbs occur in small amounts (Fig. 98). The silt first gave an age of 32 700 ± 560 BP (GSC-3381), but a Picea fragment later dated >39 000 BP (GSC-3417). The pollen indicates coniferous forests and a cool boreal climate. It compares with other organic beds on Cape Breton Island which are Th/U-dated to the interval from oxygen isotope substage 5a through oxygen isotope stage 3. The underlying till may therefore be Illinoian, Sangamonian or Early Wisconsinan, unless there has been an inversion of the sequence due to slumping.

The gravelly silt is overlain by clean well-rounded gravel of two lithologies. The lower is mainly diorite and other dark, intermediate volcanic rocks, whereas the upper consists almost entirely of granite cobbles. The upper contains reworked wood fragments, one of which dated >48 000 BP

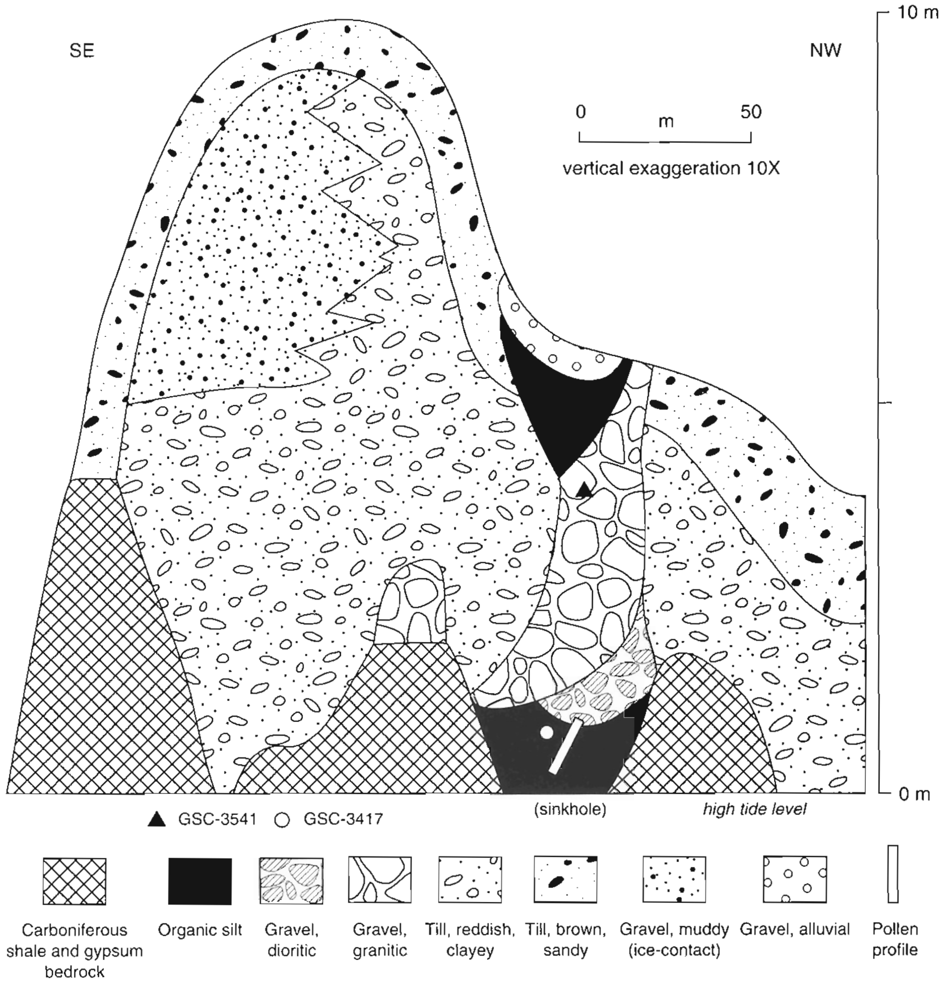



***Figure 97.*** *Geological cross-section of subtill interglacial organic bed at Dingwall [4] (after Grant, 1987, p. 35).*

(GSC-3541). The gravels may be fluvial or glaciofluvial and related to the tills, or, because in one place they overlie a smoothly truncated bedrock surface that resembles the wave-abraded bench seen at other sites, they may be littoral and correlative with the rock bench and associated gravel which underlie other subtill organic occurrences, such as nearby Bay St. Lawrence [1].

A brownish-red, clayey till with a northeastward fabric (Newman, 1971) borders the organic gravel laterally and forms a drumlinoid mound. Interbedded with the till is a westward-thickening wedge of muddy gravel which is assumed to be associated ice-contact sediment. The stratigraphic position of the till relative to the above-described units is unclear. If the organic package has dropped through it, as would seem to be the case, the till would actually be stratigraphically lower and therefore pre-Sangamonian. If there has been no penetrative inversion of order, the till is Wisconsinan. The surface is composed of brown sandy till with southeastward fabric as if emplaced by ice moving off Aspy scarp (Newman, 1971), except over the sinkhole where

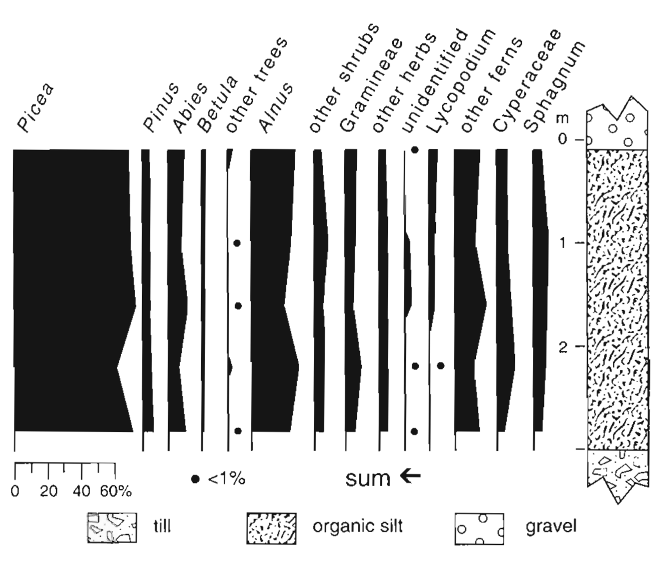



***Figure 98.*** *Simplified pollen diagram for the interglacial organic bed at Dingwall [4] (courtesy R.J. Mott, from Grant, 1987, p. 35).*

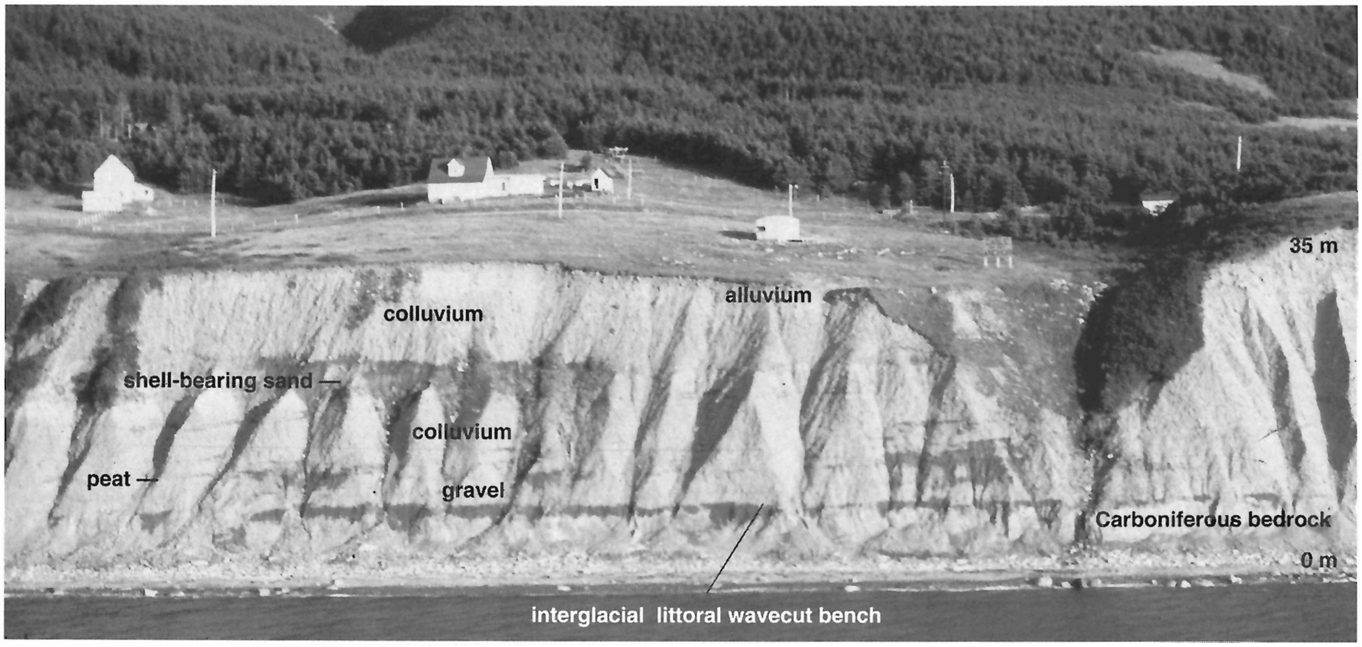



*Figure 99. Aerial view of sequence at Bay St. Lawrence [1] with lithostratigraphic annotation. GSC-203504-J*

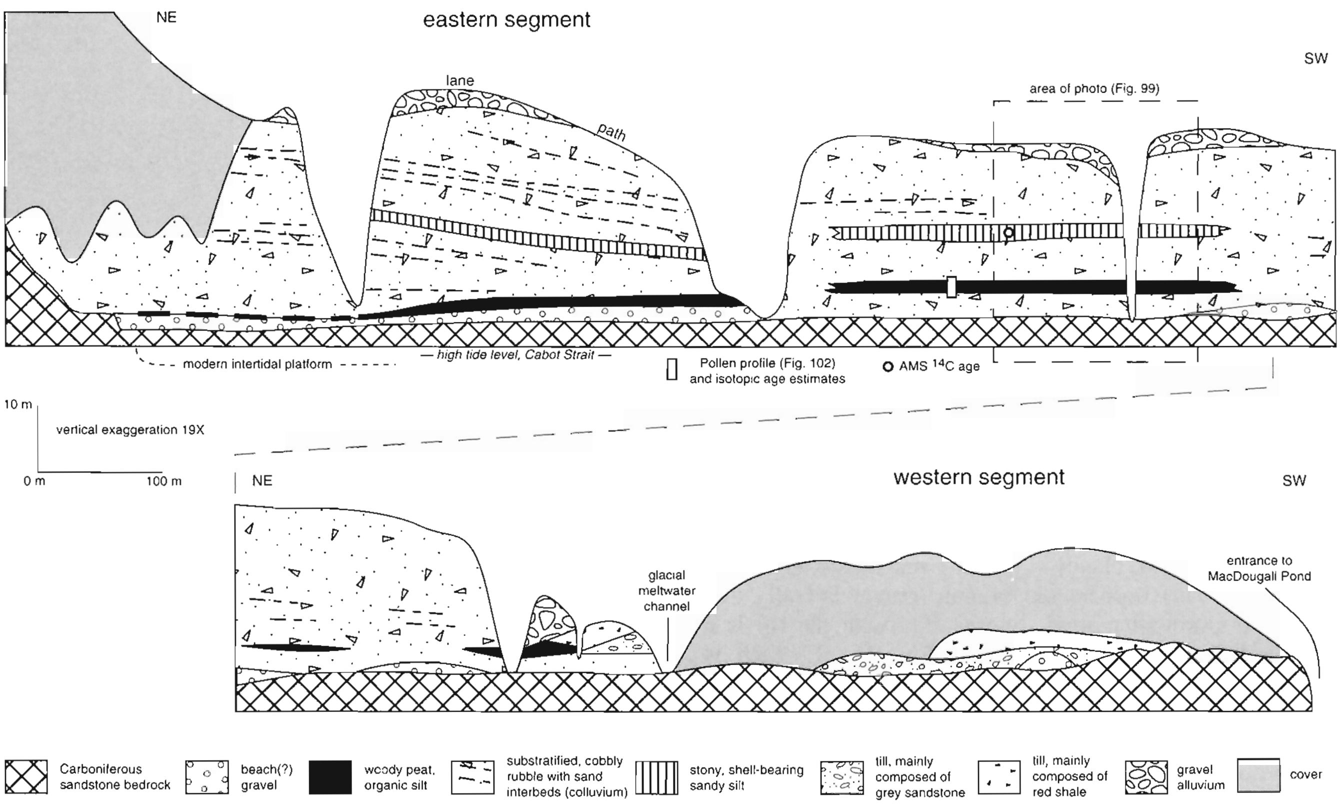



*Figure 100. Geological cross-section of sequence at Bay St. Lawrence [1] (traced from photographs taken offshore).*

there is a second plug of grey organic silt (not analyzed), which appears to truncate the reddish till, and a patch of gravel which is part of an outwash plain in adjacent Aspy valley.

In summary, the sequence seems to record: a possible glaciation before or during the Sangamonian; nonglacial organic deposition during the Sangamonian and possibly also Early Wisconsinan; glaciation by an ice lobe moving northeastward down Aspy valley probably from a highlands ice cap (Phase A); deglaciation and a second period of organic deposition; southeastward ice flow across the valley as external ice invaded the area or lobed onshore from Laurentian Channel (Phase C); and deposition of outwash during final retreat of highlands ice (Phase H).

**Bay St. Lawrence [1]**

This sequence, first noted in 1954 by Neale (1964), extends along the coastal bluffs of Cabot Strait between the cliffs of Cape North and the entrance to MacDougall Pond at the village of Bay St. Lawrence (Fig. 30). The deposits comprise gravel, peat, sandy rubble (colluvium), shell-bearing mud, and till resting on an emerged rock bench (Fig. 99). They form a 10-35 m thick apron over a narrow lowland on Carboniferous sedimentary bedrock between the 430 m high crystalline plateau and the 500 m deep Laurentian Channel. Mott and Prest (1967) focused on the peat and termed it a cool-climate "interstadial" deposit; Newman (1971) interpreted the enclosing rubble as Late Wisconsin till. Later work has narrowed the age of the peat but the origin and age of much of the rest of the sequence is still disputed. For the purposes of the following description, the 1800 m long exposure (Fig. 100) is divided into two sedimentary complexes, a western segment mainly of till and an eastern segment, mainly rubble with two organic beds, which are separated by three deep gullies that hamper correlation.

Under the 600 m long western segment, the rock bench undulates gently between 5-7 m above tide level, but nearer the inlet to MacDougal Pond it rises relatively abruptly to a rolling glacial pavement at 10 m elevation. Over most of this distance it is overlain by grey till mainly composed of the underlying sandstone and, in one place, has striations pointing 190°. The basal part of the grey till contains a few well-rounded boulders evidently incorporated from nearby gravel. From the lithology and striations it is concluded that the lower till was emplaced by ice lobing directly onshore, presumably during the early part of the invasion of the area by foreign ice (Phase B). The grey till is overlain by red till with fragments of local red shale and crystalline rock from the highlands at Cape St. Lawrence to the west. It was evidently deposited by an offshore glacier moving parallel to the coast, probably during the maximal period of Phase B when ice was moving parallel to the axis of Laurentian Channel. At the east end of this segment, most of the beds are cut out by two gullies which are the ends of sidehill meltwater channels that continue inland and have been overdeepened near the coast by modern streams. The channels appear to mark the western limit of the two till sheets. In a small remnant between the gullies, the tills overlie a dark grey, slightly organic silt horizon which thickens to become a major peat bed in the eastern segment of the exposure.

The 1200 m long eastern segment of the section is up to 35 m thick. It rests on Carboniferous sandstone which has a smoothly truncated surface that undulates gently down to the east from 10 m to about 1 m above tide level. On the bedrock cliffs at the eastern end of the section, a wave abrasion notch emerges at 6.9 m elevation and extends completely around the northern cliffs to Money Point where it stands at 7.2 m and is glacially striated (150°). At the lighthouse and farther south, the notch expands to a 10-50 m wide bench that is overlain by till composed of grey Carboniferous Horton sandstone rocks which lie to the west. Given that the bench is seen at other localities to have a distinct seaward slope, that part of the buried truncated bedrock at Bay St. Lawrence which lies below 7 m elevation is considered to be the lower seaward portion of a fossil marine bench or abrasion surface. This bench thus constitutes the starting point of the sedimentary succession.

The lowest unit overlying the bench in the eastern part of the section has been reported by some authors to be till (for example, Newman, 1971; Stea et al., 1992) and this would not be inconsistent with the fact that the bench is overlain by till at Money Point. This study, however, saw no till in this part of the sequence, only a gravelly material (Fig. 101) composed of smooth, but nonstriated, pebbles of crystalline rock from the surrounding highlands. Although many are rounded, most are angular, suggesting frost-broken debris shed by slopewash from the rocky cliffs just behind the exposure. Conspicuous at the base of the gravel are smooth, well-rounded boulders up to a metre in diameter composed of crystalline rock, mostly pink granite from the adajcent highlands. Some have percussion marks typical of those formed on boulders on the present beach by impact of wave-thrown cobbles. These boulders, and the fact that the gravel is horizontally well-stratified parallel to the underlying rock bench and has its top at the same level as the 7 m notch suggests that it is slopewash material that has been slightly reworked in the littoral zone.

The pebble gravel grades up through a silty transition zone to a compact peat (Fig. 101) with greatly flattened large wood fragments, including branches, small trunks, and root crowns. The wood has given $^{14}C$ ages of >38 270 BP (GSC-283) and 44 200 ± 820 BP (GSC-3636), and a Th/U age of 47 000 ± 4700 BP (UQT-178). Pollen analysis (Fig. 102) shows a continuous vegetational succession from open tundra, to coniferous forest containing some hardwood trees, and back to tundra (de Vernal et al., 1983), thus representing a climatic oscillation during which the maximum temperature was lower than at present. Therefore, the peat could date from any period between substage 5c and stage 3. Grant (1977) suggested that the gravel to peat succession indicated marine offlap followed by paludification when the sea regressed from the upper or lower littoral benches during the eustatic sea level fall associated with the onset of Wisconsinan climatic cooling and glacier buildup in oxygen isotope stage 4 (≤75 ka). Occhietti and Rutter (1982)


*Figure 101. Closeup view of organic bed at Bay St. Lawrence [1]. GSC 204042-R*

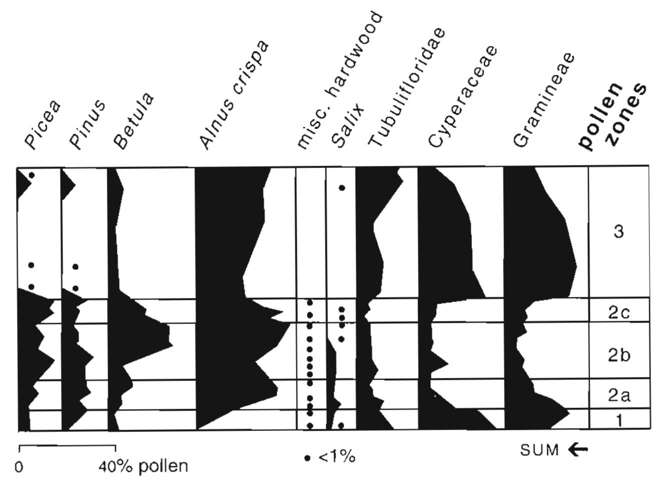



*Figure 102. Pollen diagram for the Bay St. Lawrence [1] woody organic bed (simplified by Mott, 1989 from de Vernal et al., 1983) Profile at location of $^{14}C$ age estimate on Figure 100.*

also suggested an Early Wisconsinan age for the peat based on amino-acid racemization ratios in the wood. De Vernal et al. (1983, 1986), on the other hand, accepted the quoted ages as finite (not minimal) and opted for a Middle Wisconsinan (stage 3) age for the peat.

Crucial to the age of the peat and its relation to interglacial marine fluctuations is whether the peat grades down into a marine deposit and whether it rests on the lower of two possible benches. If there is only one bench, which represents the last-interglacial sea level maximum about 125 ka, and the gravel is nonmarine, then the peat can date from any later time, whether ≥75 ka, as inferred from stratigraphy, or 47 ka as dated by $^{14}C$. That scenario, however, poses the problem of how the meagre thickness of unweathered gravel between bench and peat can represent a 50-70 ka period of subaerial exposure, unless there is a major sedimentary, if not also erosional, hiatus between the two. If, on the other hand, there is a second lower bench at about 2 m that was cut during the second main interglacial high sea level about 85 ka (substage 5a), and if the peat grades into its associated regressional littoral gravel, then the peat would have formed shortly after the regression and its suggested age of about 75 ka would be more reasonable. The latter scenario is tentatively adopted in this report.

The peat is overlain by 20 m of compact, sandy, substratified subangular, cobble-sized rubble composed entirely of crystalline rocks arranged in planar beds (Fig. 99, 101) which incline with a gentle concavity laterally and inland up toward the adjacent rock slopes. Tabular fragments are present in many beds and are nested in an imbricate fashion, especially in the more sandy and better stratified beds. This is seen as indicating deposition by sheet or channel flow downdip or generally seaward. Nearer the top, the rubble has minor lenticular interbeds of loose, reddish-brown, muddy diamicton, one of which near the top contained a 2 m diameter, striated gneiss boulder. The lithology of the diamicton resembles that of the tills farther west, but its unconsolidation, lack of fines, and inclusion with the colluvium suggests that it is probably not a true till in place, but more likely resedimented till material derived from the slopes behind the section and emplaced by mud flows. Newman (1971) interpreted all of the rubble as till, calling the imbrication a till fabric and inferring deposition by ice flowing downslope from the plateau. Stea et al. (1992) also interpreted the material as till, apparently solely because it contained schist material from the neighbouring highlands. As the rubble lacks the striated and faceted stones, clay skins, and compactness that are typical of most tills, this report follows the original interpretation of Grant (1977) that the material is frost-broken debris, shed from the adjacent steep rocky slopes and deposited by sheetwash to form a seaward-dipping apron, and it is shown thus on Map 1631A. De Vernal et al. (1983) concurred, calling it rillwash and torrential sediment in which they found pollen assemblages which varied through the colluvial sequence. From that, they inferred alternations between tundra and forest tundra and suggested that the deposit reflected severely cold conditions as might be associated with glacier ice covering other parts of Cape Breton Island. If the material is till, these varying proportions have no climatic meaning.

In the middle of the rubble bed at 23 m elevation is a bed of substratified, sandy grey silt with gravel lenses and rare fragments of marine shells. It pinches in thickness from 3.5 m in the middle of the section to a thin seam near the meltwater gullies. The shells are those of pelecypods and gastropods, namely *Mercenaria mercenaria*, *Nuculanidae*, *Dentalium occidentale*, *Tachyrhynchus erosus*, *Retusa obtusa*, and *Oenopota cf. turricula*. According to Wagner (1977), most are represented among the modern day fauna down to ≈25 m and thus may be regarded as interglacial in character. A fragment of *Serripes groenlandicus* gave an AMS $^{14}C$ age of 21 920 ± 150 BP (TO-246), but amino-acid ratios suggest the shells are more likely >50 ka and more probably closer to 100 ka (Stea et al., 1992). According to Guilbault (1982), the shell-bearing silt also contains marine microfossils, mainly foraminifera, that comprises a mixed assemblage, varying from those species that inhabit the present nearshore to those that inhabit muds below 100 m depth in Laurentian Channel. In an interpretation arrived at jointly with the present author, Guilbault concluded that the fossils had been transported by glacier from the deep offshore and redeposited in an ice-marginal pond dammed against the slope. Pollen and dinoflagellate assemblages, together with the presence of freshwater *Pediastrum*, support this hypothesis, suggesting that material from former warmer-water beds offshore was redeposited in a pond in a tundra environment (de Vernal et al., 1983). The present author envisages that the silt bed was laid down on an aggrading colluvial apron in an ice-marginal pond in front of the glacier which presumably deposited the tills at the west end of the section. Whatever the origin of the thick stratified rubbly beds, the shell bed signifies that the last glacier thick enough to ground in Laurentian Channel and impinge on northern Cape Breton Island was post-Sangamonian and probably Wisconsinan. In terms of the glacial sequence of this report that event would probably correlate with Phase C.

The colluvium grades upward to cleaner and better sorted fluvial gravel which forms alluvial fans on the surface. Now entrenched by modern streams, the fans possibly reflect the gradual decrease of sediment as slopes became forested during the postglacial climatic amelioration.

In summary, the deposits at Bay St. Lawrence are crucial to the dating of the last major ice sheet in northern Cape Breton Island and Cabot Strait area. A basal marine bench records the last-interglacial sea level maximum and a possible lower bench may represent a second marine transgression, possibly related to the substage 5c warm interval. A prominent peat bed was deposited, probably shortly thereafter (about 75 ka) as climate cooled and sea level regressed from the site. One further glacial event is recorded, that of a major invasion by a glacier grounded in Laurentian Channel.

**River Inhabitants [48]**

The first mention of organics under till in Cape Breton Island, and indeed, in the whole of eastern Canada, was by Dawson (1855; 1868, p. 63) who reported 'hardened peat with roots and branches under 20 feet of till' on "Inhabitants River". The deposit was later referred to as "tender coal" of Carboniferous age by Fletcher (1881) but, because the name of the river had been changed, it could not be located until the author noticed on 1:15 000 scale colour airphotos what appeared to be the same section on Northwest Arm Brook (Fig. 103), a tributary of River Inhabitants (Grant, 1971a). There, the river cuts a 10 m gorge across Carboniferous argillite bedrock with a steep slope that is overlain by a sedimentary sequence that

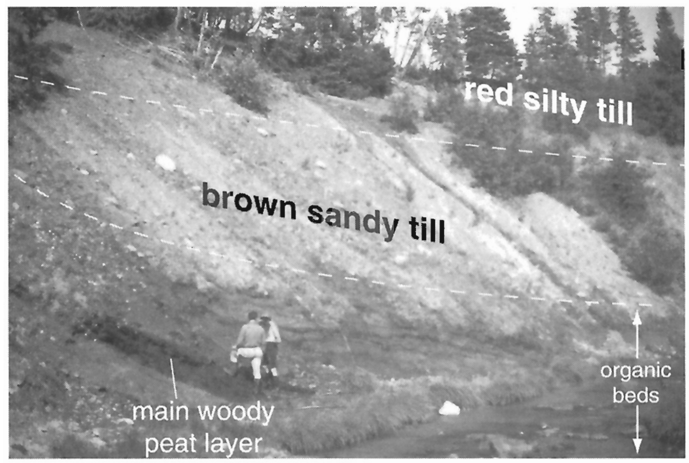



*Figure 103. View of River Inhabitants [48] buried organic sequence. GSC 203504-E*

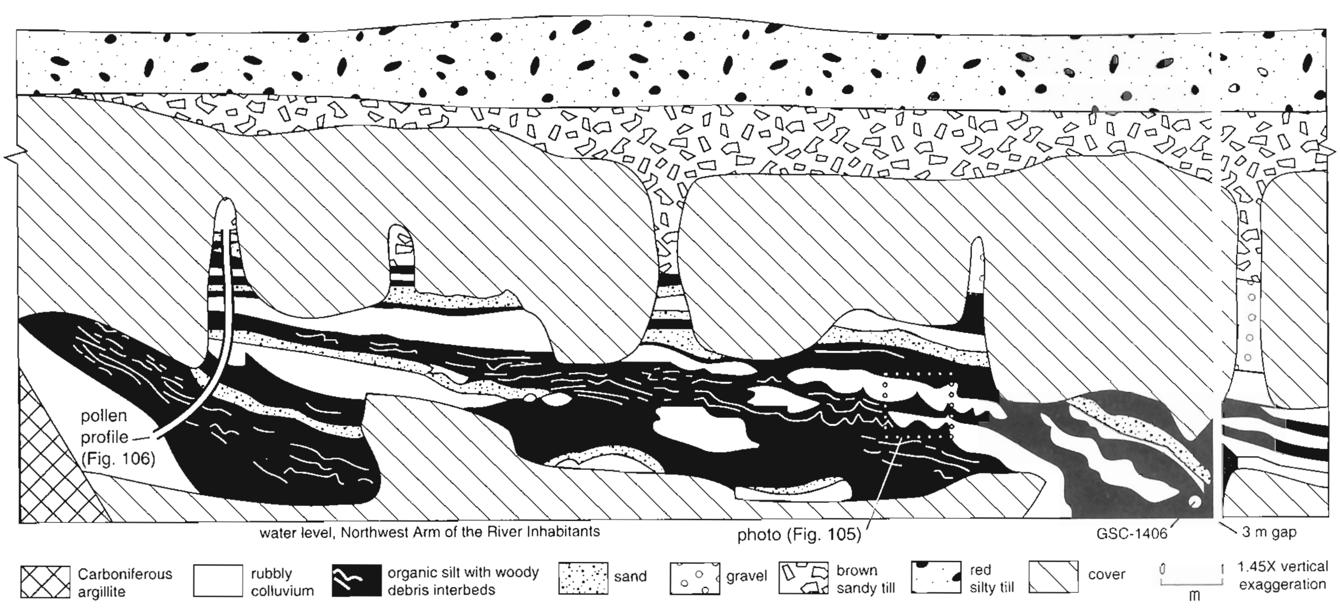



*Figure 104. Lithostratigraphic cross-section of River Inhabitants [48] buried organic beds (traced from photographs).*

appears to fill a buried valley. When first seen, it was apparent that some organic material at water level was overlain by gravel and till. The detailed stratigraphy (Fig. 104) was revealed using a fire-fighting pump to excavate the entire lower part of the section and by cutting a few vertical channels to establish the continuity of the overlying units. The basal ≈3 m thick organic member, which laps onto the steep bedrock slope, consists of seven alternating layers of organic silt and peat, the lowest of which exposed contains abundant wood fragments, interbedded with rock debris. The peat is compact, with a platy structure and the wood is flattened to about half its original diameter. The peat layers are bowed downward, stretched, boudinaged, and involuted (Fig. 105) as if by differential compaction due to glacier loading presumably when the tills were deposited. The organics are overlain by ≤1.5 m yellow iron-stained cobble gravel and by two tills. The lower till is 3 m thick, sandy, brownish, compact, and with a distinct admixture of foreign crystalline rocks; it probably relates either to Phase B or to Phase C. The upper till is 2.5 m thick, silty, salmon-red, and, because of its content of shell fragments, some of which are *Mercenaria mercenaria* (a temperate-water species typical of present coastal conditions), is referred to glacial Phase D which advanced northward from the shelf and introduced shell material to the southern lowlands.

It was originally assumed that the peat/rubble alternations represented colluvial cycles caused by climatic variation, but J.B. Railton (pers. comm., 1970) found no evidence of this in a pollen sequence through the middle four units. Pollen assemblages documented by Mott (1971) are dominated by abundant spruce and alder, and relatively abundant nonarboreal taxa (Fig. 106). Wood fragments composed of spruce, alder, and balsam fir is present; one piece dated >39 000 BP (GSC-1406). Mott (1971) concluded that the assemblages represent wet, northern-boreal climatic conditions, cooler than at present. He termed the deposit "interstadial" and assigned it to the Early Wisconsinan because the narrow definition of interglacial climates prevailing then did not include depressed temperatures. However, the River Inhabitants organics were found to resemble portions of the Castle Bay sequence which had been tentatively correlated with oxygen isotope substage 5a, stage 4, or even early stage 3 (Mott and Grant, 1985). The River Inhabitants sequence thus records part of the transition from interglacial to glacial

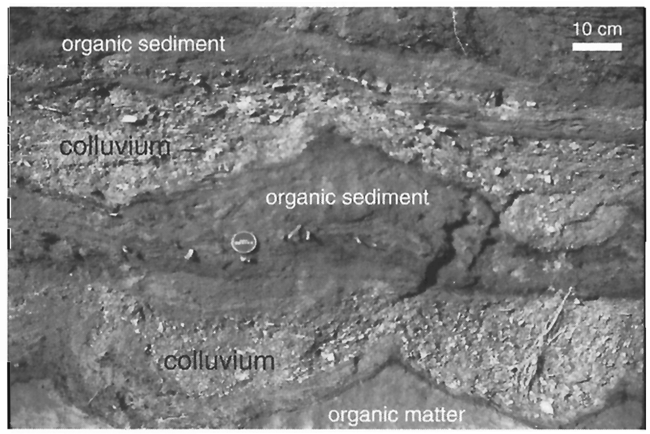



*Figure 105. Detail view of interbedded organic material and colluvium showing glaciotectonic deformation. GSC 203504-N*

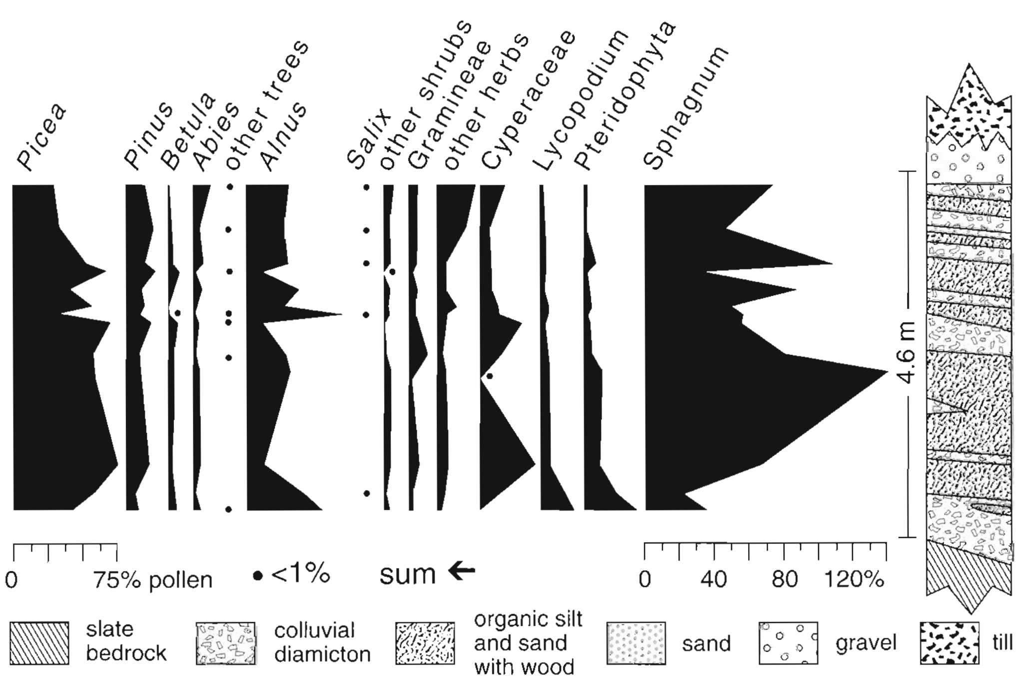



*Figure 106. Simplified pollen diagram of peat/colluvium sequence at River Inhabitants site (redrawn by R.J. Mott from Mott, 1971 for Grant, 1987, p. 30)*

conditions. It further suggests that the northward ice flow (Phase D) which deposited the red shell-bearing till probably predates 49 ka.

### Whycocomagh [17]

Discovered by D.G. Kelley (GSC) in 1955 while mapping the bedrock, the deposits are located at about 20 m elevation in a roadcut on the north side of the Trans-Canada Highway on the northern outskirts of the village of Whycocomagh (Fig. 30) about 1 km east of the junction of Highway 252. The highway cuts obliquely through the crest of a north-trending drumlin-like landform and exposes the top third of the sediment pile. The sequence features a thin organic bed between two tills (Prest, 1957, p. 443), but could not be observed during the present study because the exposure was sodded over. The pollen and spore content of the intertill organic bed was first studied by J. Terasmae (GSC) and reported in its stratigraphic context by Prest (1957, p. 446). Further palynological work by R.J. Mott led to a preliminary paleoecological and age interpretation by Mott and Prest (1967) who described the sequence as 20 cm of peat, locally with logs up to 20 cm diameter, interlaminated with organic clay/silt and interbedded between two ≈3 m thick gravelly tills. They assigned the organics to an Early Wisconsin interstade on the basis of its boreal-forest pollen assemblages and a wood date of >44 000 BP (GSC-290). Later, on the basis of similar pollen assemblages, Mott and Grant (1985) provisionally correlated the Whycocomagh organics with unit III at East Bay which, according to de Vernal et al. (1986), is Middle Wisconsinan.

Although the enclosing tills could not be studied during the present study, they were seen in 1969 by Tang (1970), who described the lower as crimson red, silt/clay with few pebbles and cobbles mostly of the local red Carboniferous siltstone and the upper till as light greyish-brown and sandy with abundant cobble-sized crystalline rocks and a weak northwest/southeast fabric. From this it is provisionally concluded that the lower till was emplaced by the early eastward flow (Phase B) which traversed mainly Carboniferous terranes before reaching the site, and the upper till was deposited by the subsequent northward and onshore movement (Phase D) which crossed two crystalline rock uplands before reaching the site. Ice-flow phases C and E are apparently not recorded at the site. The period of organic sedimentation that evidently occurred between the two till-forming events is provisionally correlated with the period of subaerial weathering between these ice-flow events inferred from differentially iron-stained striations of Phase B (and older) elsewhere.

### Moose Point [83] (mainland Nova Scotia)

This sequence is located outside the map area on the north shore of Chedabucto Bay (Fig. 30), but is included here because it affords further support for the interpretations made in this report. Discovered by R.R. Stea in 1986 (Stea and Mott, 1990), the sequence rests on an emerged rock bench, the exposed edge of which is a minimum of 3.7 m above tide level (and it probably rises higher inland under the overlying sediments). The bench has a thin regolith that is overlain sharply by ≤2 m of silt with layers of peat, containing large wood fragments, one of which dated >49 000 BP (GSC-4419). As determined by R.J. Mott, the pollen assemblages represent northern boreal forest vegetation. In the peat bed, pollen of *Picea sp.* and *Pinus banksiana/resinosa* type increases slightly above the base, then decrease as herbs increase; *Alnus* dominates at the top of the bed. The changes reflect a cooling trend which Stea and Mott (1990) assign to the late last interglacial climatic deterioration during oxygen isotope substage 5a.

The organic bed is overlain by two tills and an interbed of waterlain sediment. The lower till was emplaced by eastward ice flow as evidenced by an eastward fabric (115°-125°), striations pointing 095° where it rests on the rock bench, and distinctive appinite erratics from the northwest. It also contains shell fragments thought to have been scooped out of George Bay, although Chedabucto Bay would be an equally plausible source. Its direction of emplacement suggests that this till probably relates to Phase B. The waterlain sediment consists of a few decimetres of gravel, sand, and silty sand. The upper till has a northwestward-pointing fabric that is parallel to the trend of the drumlins and drumlinoids into which the terrain has been shaped.

The Moose Point section helps link the Cape Breton Island stratigraphic sequence developed in this report to that in eastern mainland Nova Scotia by Stea and Mott (1990). It extends the range of the emerged wavecut bench and confirms that it predates substage 5a, although not necessarily substantially. It also shows that the first Wisconsinan till deposited in this area was by ice flowing eastward (Phase B) and its shell content may suggest that the giant eastward-trending ridges in southern Cape Breton Island were formed as drumlinoids during the initial invasion. The surface till, however, poses a problem in that its inferred northwestward direction of emplacement is exactly opposite in sense to that for the surface till in adjacent Cape Breton Island. The conflict could be resolved by supposing either that its direction of emplacement is wrongly inferred or the surface till at Moose Point correlates with the intermediate-age, northward-emplaced till in Cape Breton Island (Phase D) and that the surface till in Cape Breton Island, formed by final southward flow (phases E, G), was not deposited at Moose Point.

## *Wisconsinan glacial and nonglacial sequences*

The sequences in this group are those which are inferred to begin during Wisconsinan time (75-10 ka BP) based either on direct dating or correlation with deposits and features which overlie and postdate inferred Sangamonian deposits. This group includes several multiple-till sequences and a few organic beds. In addition, some of the best and most telling examples of multiple crosscutting striations are documented to demonstrate that the erosional sequence corresponds to the depositional stratigraphy. These and additional sites are outlined in point form in Table 1 and are located both on Figure 30 and Map 1631A. They are arranged roughly in order of importance and complexity.

## South Aspy River [5]

Discovered in 1966 by W.A. Newman, this exposure is located on the north bank of South Aspy River about 3 km inland of Cabot Trail (Fig. 30). The river has cut down into a 3 m thick outwash plain and exposed 15 m of the underlying sediments (not seen during the present study because of slumping). According to Newman (1971), the outwash is underlain by two dissimilar tills, both of which contain masses of organic silt. The lower till is compact, massive, yellowish brown, and silty, with a west-pointing fabric. It contains a few conspicuously well-rounded and polished stones and some blocks of silt (not analyzed). He attributed this till to a glacier that lobed onshore into Aspy valley and reworked waterlain material, probably fluvial or marine. Relating this till to the ice-flow sequence, it is suggested that this till may have been deposited during the early stages of Phase B.

The upper till is described as dark brown, cobbly, with an upland crystalline lithology and a bimodal south- and southeastward-pointing fabric (Newman, 1971), thus suggesting that it probably relates to glacial Phase C. In it, he found angular blocks of grey silt, of which the largest was 1.5 m in longest dimension and contained "crushed and oxidized vegetation zones". No macrofossil results are reported, but the pollen content of the vegetal layers was analyzed by D.A. Livingstone (pers. comm., 1970) and showed mainly *Betula*, and lesser but equal amounts of Alnus and Coniferae (plus "considerable amount of reworked Carboniferous microfossils") which he interpreted to indicate northern boreal forest, rather than tundra. As the organic matter yielded radiocarbon ages of 20 300 ± 400 BP (I-2438) and 24 900 ± 700 (I-3413), the deposits were taken to indicate the timing of Late Wisconsinan glaciation, however, the ages are suspect for two reasons. On the one hand, the high content of pre-Quaternary material could mean the true ages are younger (and the enclosing tills thus younger still). On the other hand, if the organic concentration was too low (<5%) to ensure a reliable count, the true ages may be much older. The latter possibility is considered more likely in view of the finding by R.J. Mott (pers. comm., 1987) that the pollen assemblage resembles that of the grey silt at nearby Dingwall [4] (which is pollen-correlated to pre-Late Wisconsinan; oxygen isotope substages 5a through 3). The silt blocks are therefore tentatively regarded as reworked from Sangamonian interglacial or Wisconsinan interstadial beds in the area. Notwithstanding this uncertainty, the enclosing tills provide a record of Wisconsinan glacier movements in the Aspy Bay area.

The upper till is overlain by 2 m of planarstratified sand and interbedded grey sandy silt with virtually no pollen and a negligible organic matter content. Above this, the top 2 m of the sequence is composed of horizontally bedded gravel with a muddy sand matrix. Although Newman (1971) called this surficial material till, it seems more likely to be outwash because its top is the surface of an outwash plain that borders the river.

The South Aspy River section is important because it shows that after an interglacial period of organic deposition, a Wisconsin glacier moved onshore incorporating waterlaid sediment (early Phase B); this was followed by a glacier movement that was of sufficient magnitude to overtop the highlands (maximal Phase B). Ultimate glacial retreat was evidently up-valley to an ice cap on the highlands plateau (Phase H), as represented by the outwash gravel.

## Morrison Brook [8] and Wreck Cove head [10]

Reference has been made in the description of thick composite till blankets (unit 2a), to the two-till sequences at Morrison Brook (Fig. 47) and nearby Wreck Cove head (Fig. 53). The first records the period of Sangamonian bench cutting, as represented by the horizontally truncated bedrock surface ranging from at least 4.1 m to at least 6.9 m above tide level. Both exposures show that this was followed by ice flowing eastward off the plateau (Phase A and/or B), which deposited a granitic till of inland derivation, then by ice flowing along the coast in a general northeastward direction (Phase D and/or E) to deposit a till composed mainly of diorite from a pluton to the southwest, admixed with lowland Carboniferous rocks.

## Gabarus Bay-Fourchu area [35-38]

Erosional evidence of the succession of the major glacial phases that affected the island is deduced from crosscutting erosional markings on bedrock. Some outcrops have up to three superposed sets of striations. Many examples of the crosscutting striations are inscribed on the presumed Sangamonian bench, so if its age is correctly inferred, all of the flow phases are Wisconsinan. Outcrops with the greatest number of superimposed striation sets are found mainly in the southern lowlands. Good examples of three superposed sets occur at Red Point (Fig. 49a), near Mulgrave (Fig. 49e), and in southern Isle Madame. The best exposures of multiple striation sets are on the resistant silicic Fourchu Group rocks in the Fourchu-Louisbourg area. The earlier sets on some of these have a weathered stain, whereas the later ones are fresh (Fig. 49b). The Fe/Mn stain is tentatively interpreted as representing accumulation in a soil B-horizon during a period of subaerial exposure between the two glacial advances.

For example, at Gabarus [36] a facet with miniature crag-and-tail pointing 110° and an iron/manganese stain is in the lee of a second stossed facet with striations pointing 030° that is fresh and unstained. There are also three beveled facets at Fourchu [38]. The earlier two facets have stossing and striations trending 070°-090° and 030° and are stained, whereas the latest facet with stossing at 170° is unstained and fresh. The outcrop is overlain by red till derived from Carboniferous and Triassic rocks to the east and south. These relations are interpreted to mean that, after a period of onshore ice flow from a centre on the shelf (Phase D), a nonglacial period of subaerial weathering intervened before the period of offshore flow from a source on the island (phases E, G).

At two sites on Gabarus Bay, one on the south side at Gabarus Cove [37] and the other on the north side near Deep Cove [35], the emerged bench is seen to cut across Carboniferous breccia composed of silicic tuff in a hematitic matrix. The breccia fragments have been etched in relief by glacial scour and show eastward-directed crag-and-tail that has been modified by later northward stossing (Phase D). Two superposed tills nearby corroborate the erosional sequence – a red, foreign drift emplaced by eastward flow (Phase B) is overlain by grey, local drift emplaced by northward flow (Phase D).

At Lighthouse Point in Fortress Louisbourg National Historic Park, the emerged bench has early stossed facets with 050° striations and miniature crag-and-tail (Phase B), that are truncated by later beveled facets with 010° striations (Phase D), and overlain by a reddish till rich in red Carboniferous sandstone and shale indicator erratics from the west.

### Isle Madame area [42-44]

This southernmost extremity of Cape Breton Island is notable for its many exposures of thick drift composed of two distinctly different tills. The best exposures are on the north shore of Bay of Rocks [42, 43] and on Heath Head-Red Head [44]. Rock pavements with eastward striations crossed by northward ones are overlain by a lower reddish, silty, shell-bearing till that is the product of one or more of the early ice flows (phases B, C). It is veneered discontinuously by a sandy grey till composed mainly of the local sandstone bedrock. Locally, the upper till rests on bedrock that is striated 160°, so it is presumed to the product of the final radial dispersal of an ice mass centred on Bras d'Or Lake basin (phases E, G).

### Shell-bearing glacial deposits [13, 19-27, 43-50]

Fragments of marine mollusc shells are widely dispersed in the tills on southern Cape Breton Island between Port Hawkesbury and the head of East Bay (Fig. 30). Those that can be identified are all temperate species, for example *Mercenaria*, *Macoma*, and *Clinocardium*, which occur in inner shelf parts of the area today. The shell material is most abundant in the earlier-deposited till(s), that is to say, in those tills which were deposited during the initial eastward and northward movements. Shells are most abundant in the till on Isle Madame [43-46] and generally decline in abundance inland and northwards, as would be expected if they were brought onshore from the Scotian Shelf during the northward advance (Phase D). Another area of shell-bearing tills occurs along the north side of East Bay [19-25]. These were probably dredged up from Bras d'Or itself, as no shell-bearing till is found south of East Bay inland of the Atlantic coast. Shell fragments in the till have locally been further reworked and recycled into younger waterlaid deposits, such as an esker at Grantville [47], a proglacial delta at Campbell Brook [50] (the most inland occurrence), and Holocene marine sediment in Canso Strait (Wagner, 1975, 1977). The last occurrence contains *Emargulina sp.*, *Chione cancellata*, *Rimula frenulata*, and *Spiratella lesueri* – species which today do not occur north of North Carolina.

The age of the reworked shells cannot be defined precisely, but is probably last interglacial. Stratigraphically, the shells occur in the first Wisconsinan till and so are likely to be pre-Wisconsinan and therefore probably Sangamonian. The age as indicated by radiocarbon dating is generally inconclusive. Shells in till on Janvrin Island [45] dated >34 000 BP (GSC-1639), whereas those recycled into an esker at Grantville [47] gave a finite age of 32 100 ± 900 BP (GSC-1408). The latter is suspect, being on carbonate and near the limit of the method, so the true age is assumed to be much greater.

The paleoecology of the faunal assemblages is more telling of a possible interglacial age. All are temperate species reflecting marine conditions as warm as or warmer than at present and are therefore characteristic of interglacial conditions (Wagner, 1975, 1977). Accordingly, a Sangamonian age assignment is tentatively given to the shell material. Moreover, the presence of shells indicating warmer than present conditions for the last interglaciation is supported by the palynology of the last interglacial beds which indicates that climate became warmer during the last interglaciation than during any period of the present interglaciation (Mott and Grant, 1985).

### Mullach Brook [16]

A till sequence documented by Tang (1970) features three drift sheets that were exposed in an 8 m deep trench opened to test for gold-bearing preglacial alluvium on the south side of Mullach Brook, about 3 km north of Whycocomagh on the north side of Bras d'Or Lake (Fig. 30). The bottom metre was cut in Carboniferous siltstone bedrock and showed successively: 1 m of regolith; 1 m of red, pebbly, silt/clay till derived from the underlying siltstone bedrock; 2-6 m of light grey-brown, pebbly sand till composed mainly of local sedimentary rocks, plus a few crystalline erratics from the uplands to the east; and a 2-6 m thick upper till with a red silt/clay matrix and pebble- to boulder-sized fragments composed almost entirely of local sedimentary rocks. According to Tang (1970), fabrics, heavy minerals, and stone provenance suggested that the lower till was deposited by east-flowing ice (probably Phase B), the middle till by northwest-moving ice (probably Phase D), and upper till by west-moving ice (probably Phase E). The Mullach Brook till sequence thus accords with the succession of ice flows recorded in that area by crosscutting striations and other glacier-erosional indicators.

### Sight Point [60] and Mabou Hills

This western coastal promontory of Mabou Hills (Fig. 30) retains several well-preserved segments of the emerged wavecut bench at 4-6 m above tide level on which two striation sets are inscribed and two diamictons deposited. The bench looks exactly like the modern intertidal bench at the foot of the exposure in that it has smooth potholes and a veneer of well-rounded beach cobbles. Locally, the emerged bench is overlain by a compact till with a dusky reddish brown, silty matrix evidently derived by comminution of the Carboniferous redbeds that lie offshore, plus a few pebbles of the local crystalline rocks. This till overlies two striation sets, an early set oriented 110°-135° and a later set directed to 045°. In terms of lithology and emplacement direction, it resembles the lower till overlying the interglacial organics at nearby Green Point [57] and for that reason is attributed to the same onshore ice-flow event (Phase B). Overlying the till in thin patches is a loose, silty, almost stone-free, reddish-brown diamicton which, apart from the lack of compactness, resembles the upper till at Green Point, which was emplaced by ice flowing against and along the coast (Phase F). Sight Point thus appears to record two of the four Wisconsinan ice-flow events which affected nearby areas. It evidently lay beyond the limit of both the early eastward advance (Phase B) and the late glacial Bras d'Or ice remnant (phases E, G).

Further indications of limited glacier action and/or extent can be inferred from the gross geomorphology and surficial deposits of the Mabou Highlands in general (Fig. 12, 33). Up to 250 m elevation the coastal interfluves have glacially scoured bedrock and rare patches of reddish lowland till and erratics, whereas above that elevation and on the extensive planar summits, the surface is smoothly graded and rubble-covered by mass wasting; no till was seen nor any erratics of the distinctive red Carboniferous rocks underlying the adjacent lowlands. Since glaciers did overtop the highest highland summits to the north, it is assumed that Mabou Highlands were also overridden. The upper limit of ice-scoured terrain may therefore reflect the boundary between warm- and cold-based ice. The sharpness of the contact, however, suggests that it more likely marks the upper limit of a later and relatively thin glacier which advanced from the west and deployed around the summits (Phase F).

**Chéticamp Lowland [64-70]**

The western coastal lowland between the mouths of Margaree and Chéticamp rivers has a complex and locally variable sequence consisting of one main till, which in places is overlain by two gravel members and two silty units, the upper of which is a massive discontinuous veneer of undetermined origin. The lithology of the deposits is a variable mixture of both offshore sedimentary rocks and onshore crystalline rocks, so the sequence evidently records the interplay of foreign and local glaciers. A typical example is the coastal exposure near the mouth of Grand Étang [66] which displays most of the depositional units (Fig. 107). At the base, a compact, greyish-purple till (which locally overlies the emerged bench), contains lowland and highland rocks. It resembles the till at Cap Rouge [70] which was emplaced by ice flowing onshore toward 110° (Fig. 54) (Phase C). The greyish-purple till is overlain by poorly sorted, subrounded cobble gravel which has a similar lithology, and so may be the associated ice-contact stratified drift or outwash. The gravel is overlain by 4 m of soft, brick-red silt that is generally massive but locally well stratified; it is laterally discontinuous and may be a local proglacial pond deposit. Above the silt is 10 m of unstratified, muddy, angular gravel, also of mixed lithology. This upper gravel forms an extensive hummocky blanket of ice-contact stratified drift which appears to represent the disintegration of the last glacier on the Chéticamp Lowland. Holocene fluvial channel gravels are locally inset into the older units.

*Surficial red stony silt of undetermined origin*

A thin red silt of uncertain origin discontinuously overlies all older units. It occurs sporadically all along the west coast from Sight Point in the south to Cape St. Lawrence in the north, but is most common on the Chéticamp Lowland. Beyond those limits, it is conspicuously absent. North of the Chéticamp Lowland, it occurs overlying the emerged rock bench at Lowland Cove [77] and Delaney Brook [76], a valley end moraine at Pollett Cove [75], and a low alluvial terrace at the mouth of Chéticamp River [68] (Fig. 62). South of Chéticamp Lowland, it occurs at Sight Point [60], Broad Cove Marsh [61], and in patches in the vicinity of Inverness and Chimney Corner. The red silt is generally massive, but may locally have thin sand laminae, has columnar jointing, and contains scattered, rounded, matrix supported pebbles of various rock types. Its thickness ranges from 0.3 m to 2 m, but is generally about 1 m. Apart from the admixture of foreign crystalline pebbles, the silt is essentially what would be produced by comminution of red siltstone and sandstone of Carboniferous age. It usually has a conformable contact with the gravels, but locally it is gradational (Fig. 62) and interbedded as a thin layer in the top 1-2 m of the gravel, as in the outwash at Northeast Mabou River [58] and Red River [74]. The elevation generally seems to decline northwards from about 60-70 m near Inverness to about 8 m at Lowland Cove [77]. On the Chéticamp Lowland, where its distribution is more readily mapped because of the gentle slope and good access, its elevation ranges up to about 40 m, though many

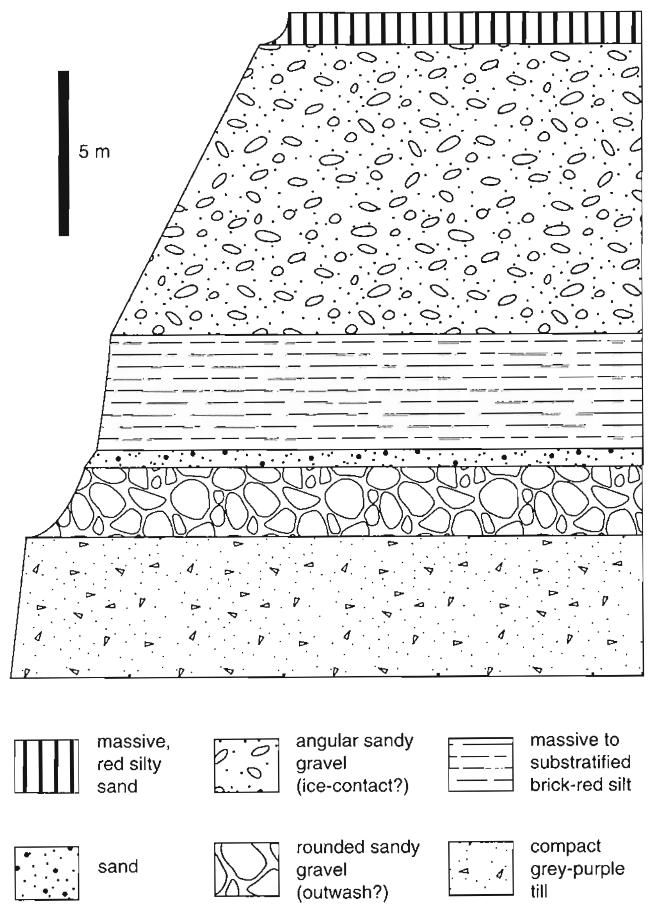



***Figure 107.*** *Diagrammatic geological cross-section of sequence at Grand Étang [66] (with information courtesy S. Occhietti).*

lower areas, notably Holocene alluvium, have none. Where pits provide exposure through gravel hummocks it is seen to pinch out on the flanks of gravel knolls, although at variable elevation and without a morphological break, and small patches occur on the gravel hilltops. Where it overlies hummocky ice-contact gravels, the kame and kettle topography seems more subdued, as if by water washing, than where it is absent. The silt is found as patches on hillocks and in Chéticamp paleochannel, an overfit valley that meanders along the lowland (Fig. 23). No macrofossils and microfossils were found to give an indication of depositional environment.

The origin of the red silt is undetermined because there are no definitive features and several contradictory aspects. Any suggested origin must accommodate the fine texture, suspended stones (dropstones?), variable structure, conformable contacts, absence of fossils, relatively sharp lateral and vertical limits, and apparent postglacial age. A glacial origin seems ruled out by the conformable contacts and more so because it would require a glacier to readvance at a very late stage from a deepwater part of Gulf of St. Lawrence onto a coast already deglaciated by land ice. An eolian origin would best explain the silt-dominated texture, but is rejected because of the included stones, the sharp geographic and elevational limits, and the patchy distribution. Subaquatic deposition could account for suspended stones and the sharp upper limits (if the water plane was above the highest occurrence), and the wide, but defined geographical restriction. A glacio-isostatic marine submergence could theoretically have overlapped the northwest coast, but only during late glacial time, not after Holocene floodplains were in place. A related possibility is that an ice shelf, rich in red debris from the Gulf of St. Lawrence sea floor, impinged on that deepwater segment of the coast and deposited the silt as glaciomarine drift. This suggestion is made on the basis that a similar brick-red stony silt occurs in two other settings nearby: as a patchy veneer over gravel on nearby Les Iles de la Madeleine (Grant, 1987, p. 19) and as two 20-30 cm layers in the upper part of the typically grey, clayey, late glacial marine sequence offshore in Laurentian Channel and interpreted as meltout debris from an ice shelf (Conolly et al., 1967). The problem with both of these marine hypotheses is that they would require a higher than present sea level for which there is no supporting evidence whatsoever. Indeed, this area lies 200 km beyond the zero isobase of late glacial marine overlap, as documented by shoreline features and fossiliferous sediment. A proglacial lake could conceivably have flooded the western lowland, given that a series of ice-dammed lakes occupied intermontane valleys in the same area at a late stage, but it would have required an unusual configuration of glaciers lying offshore at a very late stage after land ice had retreated. In any case, both the marine and lake scenarios are judged untenable because they do not explain why the sediment is patchy and commonly absent at the lowest levels and why the associated shoreline features such as fossil beaches, trimlines, and deltas are completely absent.

The most likely possibility is that the red mud is related to the network of proglacial lakes in the western part of the island which generally drained northward to, if not also along, the Chéticamp Lowland where the silt is most abundant. This relationship suggests that the red stony silt may have been deposited against and under the ice mass lying offshore in Gulf of St. Lawrence which was confining the lake water over the western coastal lowlands (Phase F). By this scenario, the lake discharged northward in the marginal zone of the Gulf of St. Lawrence glacier, variously by sheet flow beneath the ice and by channelized flow in tunnels and between stagnant ice blocks during, and after deposition of the ice-contact gravels. This scenario explains: the occurrence in the same general area into which the lakes are draining; the close association and interbedding with late glacial deposits; the vague and northward declining upper limit; the sharp contacts; and the nonsystematic patchiness. Until detailed sedimentological and geochemical analyses are made, however, the data in hand do not indicate a likely origin for the surficial red stony silt in western Cape Breton Island.

### Middle River [15] (fossils of Pleistocene mastodons and mammoths)

Several scattered mastodon and mammoth fossils have been found in the Nova Scotia area; three of the mastodon fossils are from Cape Breton Island (Dawson, 1868; Piers, 1915). A molar (Fig. 108) was reported to have been found near Baddeck [14], but no further information was given or has since come to light. The most significant find is a femur (Fig. 109) and a tusk that were plowed up in 1834 on the lower postglacial terrace of Middle River [15] on the farm of Alexander MacRae. The femur yielded ages of 32 000 ± 630 BP (GSC-1220) on apatite and 31 300 ± 500 BP (GSC-1220-2) on collagen. While it may be tempting to accept these as finite and postulate a Middle Wisconsinan ice-free interstadial period to explain the fossils, they are considered minimal ages because they are near the limit of the method and because bone is frequently subject to contamination. Pollen in cemented sediment in cavities in the Middle River femur was

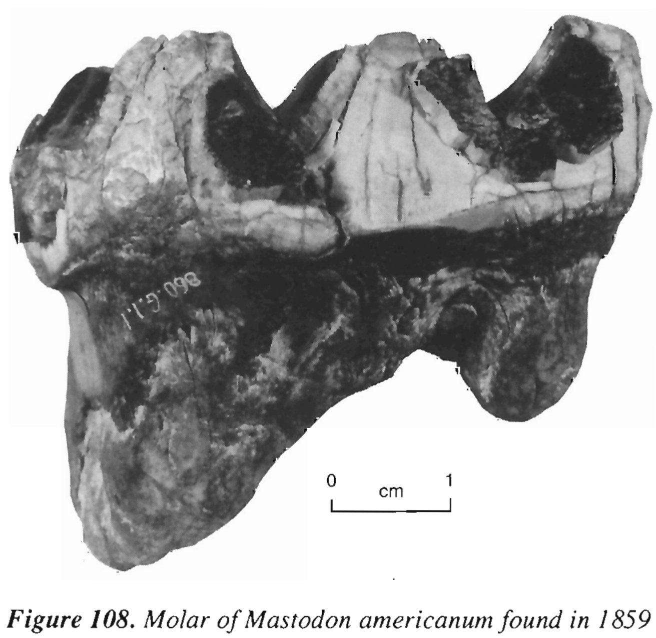



*Figure 108. Molar of Mastodon americanum found in 1859 near Baddeck. GSC 200970-A*

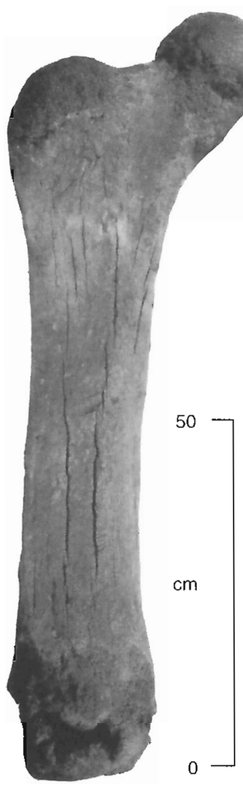



*Figure 109. Femur of Mastodon americanum found in 1834 on Holocene terrace of Middle River [15]. GSC 200987*

analyzed R.J. Mott (as reported in Harington et al., 1993) and found to contain a high proportion of thermophilous hardwood taxa. The assemblage contrasts with those in Holocene sequences and in suspected Middle Wisconsinan deposits, but it correlates well with those in Sangamonian (stage 5) deposits. The Middle River mastodon fossils (and the Baddeck molar for lack of contrary evidence) have therefore been referred to a late stage of the last interglaciation during the cooling phase associated with the onset of Wisconsin glaciation (Harington et al., 1993). In support of this age assignment is the recent find of another nearly complete mastodon skeleton (Globe and Mail, 1992) in a confirmed Sangamonian (substage 5e) deposit at Milford, Nova Scotia (Mott et al., 1982) which has a pollen signature similar to that at Middle River.

### *Postglacial and Holocene sequences*

Natural exposures and cores of bog and lake sediments have been studied to extract a record of vegetation succession which reveals the evolution of late glacial and postglacial climate of Cape Breton Island. The area has provided the first and most definitive evidence that north Atlantic oceanic circulation is the prime controlling factor on climatic conditions in Atlantic Canada and that finding has in turn led to a better understanding of the timing and style of glaciation in this maritime area. Against the background of regional paleoecological trends summarized by Anderson (1985) and Anderson et al. (1989), the Cape Breton Island sites are discussed in groups according to their lithostratigraphic and

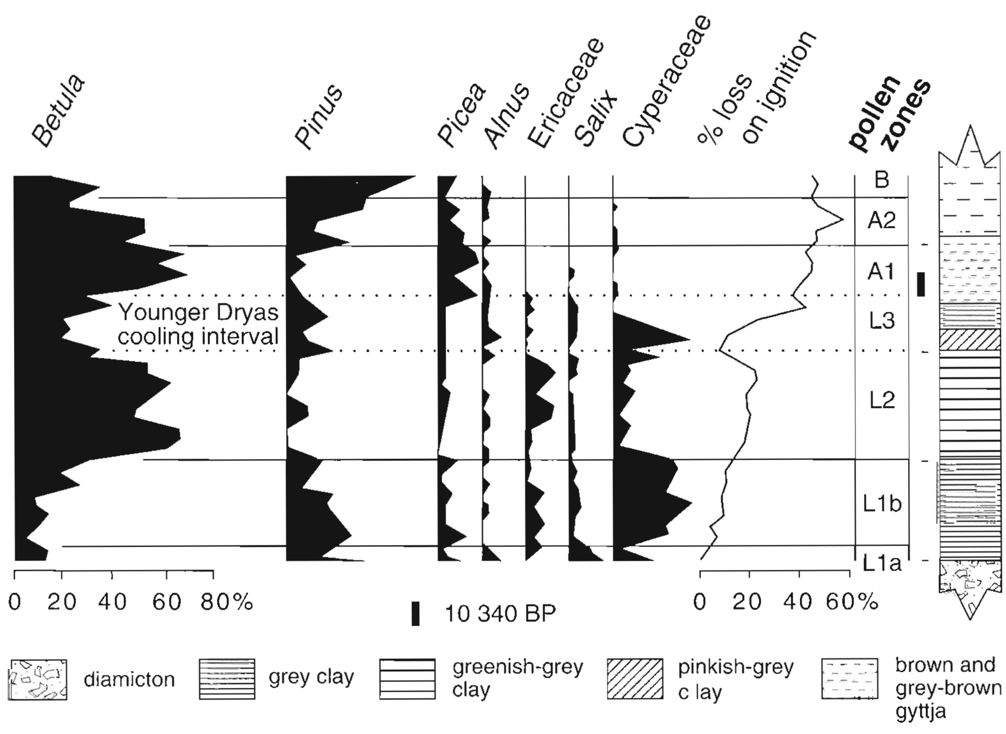



*Figure 110. Simplified pollen diagram for lower part of lake sediment core at Gillis Lake [40] (L3 pollen zone corresponds to Younger Dryas cold period) (adapted by Mott et al., 1986 from Livingstone, 1968).*

glacial geological significance in order to trace how the present understanding developed. Most of the sites on Cape Breton Island are shown on Map 1631A, except for site numbers 84-92 which were discovered or were studied after the map was submitted. All core locations are shown on Figure 30 and the main aspects are summarized in Table 1.

The first studies, conducted before the advent of radiocarbon dating, were of peat bogs situated near Mulgrave [84], and Glasgow Brook bog [85] near Aspy Bay by Hall (1949); and Trout Brook bog [86] inland of Neils Harbour, a Margaree Forks bog [87], and Cunningham bog [88] near Glace Bay by Killam (1951). The bog stratigraphy showed a vegetation succession from early postglacial tundra, to boreal forest, to a mixed hardwood/coniferous forest during a mid-Holocene Hypsithermal, and a slight reversion to more coniferous forest in recent millennia. (Further studies, still in progress, have been done of bogs at Donkin [34] and Kilkenny Lake [80] by J.G. Ogden, III (pers. comm., 1986) and of lake sediments at MacDougall Pond [78], "MacInnis lake" [79], Third Lake O'Law [81], Pembroke Lake [89], Cormier Lake [90], Timber Lake [91], and Hector Lake [92] by R.J. Mott (pers. comm., 1992). The available age determinations from the lakes studied by Mott are included in Table 1.

The first lake-sediment cores in Cape Breton Island to be radiocarbon dated were by D.A. Livingstone of Dalhousie University. The dates showed that the timing of major late glacial events on the highlands was different than on the lowlands. The northern plateau was evidently not deglaciated until after 9.0 ka, as inferred from the age of the basal tundra zone in Wreck Cove pond [9] (now submerged under a hydroelectric reservoir) (Livingstone and Estes, 1967). In comparison, the basal date from Gillis Lake [40] on the southern lowland indicates deglaciation by 13.9 ka followed by tundra conditions which lasted until 10.1 ka (Fig. 110) (Deevey, 1958; Livingstone and Livingstone, 1958; Schofield and Robinson, 1960). Further interpretation of these core led Livingstone (1963) to speculate that glacier ice persisted locally on the lowlands during "Valders" time, the period of renewed cold 11-10 ka BP. Further coring by Livingstone (1968) on the southern lowlands at Salmon River Lake [41], McDougall Lake [29], and Upper Gillies Lake [30] revealed that the normal postglacial progression from tundra to forest which accompanied the general warming was interrupted by a marked reversion back to tundra, indicating a sharp climatic deterioration, before afforestation resumed. In the Upper Gillies Lake core, the reversal coincides with a lithological change to inorganic varves which Livingstone (1968) inferred meant "glacier ice had made a close approach to the site" during the sharp climatic cooling. This was the first indication of how late glacier ice had persisted in lowland Cape Breton Island and that it responded to the climatic cooling cycle.

### Late glacial climatic cooling cycle

Discovery in the 1950s of organics beneath postglacial alluvium in the Benacadie Point area by L.J. Weeks (Prest, 1957, p. 447) was the first discovery of evidence of a major Late Wisconsinan climatic reversal during the overall postglacial climatic improvement. MacNeill (1969) reported direct evidence of a late glacial readvance over organic beds nearby at Benacadie Point [19] (Fig. 30). There, peat with an age of 11 670 ± 170 BP (I-3234) is overlain by what he assumed was till. During the present study, the author found what was probably the same peat bed; it gave a virtually identical age of 11 300 ± 90 BP (GSC-2146) (D.R. Grant in Lowdon and Blake, 1976); however, because the site is at the foot of a steep slope, it could not be ascertained whether the overlying diamicton was in fact till, rather than colluvium. Further evidence of a possible late glacial glacier readvance in Cape Breton Island was given by Grant (1972) who found woody peat with an age of 10 300 ± 150 BP (GSC-1578) between two compact and lithologically dissimilar tills at Blacketts Lake [31] (only 3 km from the Upper Gillies Lake varved sequence). On Port Hood Island [55], Terasmae (1974) found till overlying peat which gave an age of 11 000 ± 170 BP (GSC-540) and had a pollen sequence showing a deteriorating climate to the top of the organic sequence (Fig. 111) but, because the sequence occupies a still-developing gypsum sinkhole, there is a likelihood that the overlying till may have slumped in rather than being emplaced by glacier ice. This possibility led Grant (1977) to downgrade the upper till at Blacketts Lake to a solifluction deposit, but the original interpretation is retained in this report. Also on the west coast of the island, a few kilometres south of Port Hood at Campbell [53], the author found outwash overlying peat that gave an age of 11 200 ± 110 BP (GSC-2212) (Grant in Blake, 1984).

These finds prompted R.J. Mott to conduct further pollen work and dating to define the nature and age of the event. Detailed analysis of the peat on Benacadie Point [19] showed a shrub tundra/birch/tundra cycle (Fig. 112) and the ages of the bottom and top of the layer were narrowed to 12 100 ± 100 BP (GSC-3900) and 10 900 ± 100 BP (GSC-3912), respectively. Analysis of the Campbell [53] site revealed a similar herb tundra/shrub tundra/herb tundra oscillation

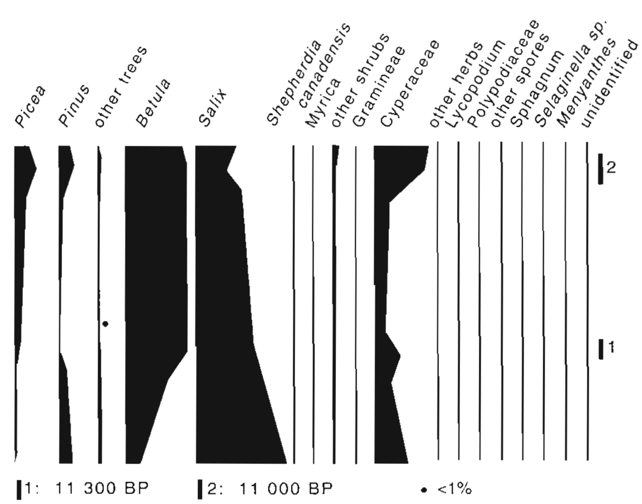



*Figure 111. Pollen diagram of Port Hood Island [55] buried peat (from Terasmae, 1974).*

(Fig. 113) and the bottom and top of the beds was dated to 11 300 ± 100 BP (GSC-3781) and 10 800 ± 100 BP (GSC-3982), respectively.

This led to a wider search for other localities where the surface till, solifluction material, or other detritus overlay organic material. One was found by R.J. Mott and the author in the interior at Big Brook quarry [49] and several more were found by S. Occhietti (pers. comm., 1987) on the East Bay coast at Benacadie shore [20, 22, 26], Amaguadees [24] (Fig. 114), and Castle Bay [25], and on the Gulf of St. Lawrence coast at Mabou beach [56] and Pleasant Bay [73]. All of these late glacial buried organic beds have comparable age ranges and pollen sequences.

These occurrences demonstrated that a late glacial climatic reversal was widely represented in the Cape Breton Island area, so R.J. Mott initiated a lake-sediment coring program to obtain the best record and to test its regional extent. The results of the more than 20 such sites found over a wide area (most of which are in Cape Breton Island; Fig. 115a), were summarized by Mott et al. (1986) (Fig. 115b). All show a comparable sequence and similar ages for the beginning and ending of the major changes. As postglacial climate improved, there was the usual increase in organic production and proliferation of vegetation. Gyttja was deposited in lakes and peat accumulated on land. The pollen sequence shows a normal vegetation succession from shrub and herb tundra, though forest tundra to boreal forest until about 11 ka, when the organic sequences are abruptly overlain by mineral layers in which the pollen shows that vegetation reverted back to tundra. The cool episode lasted for

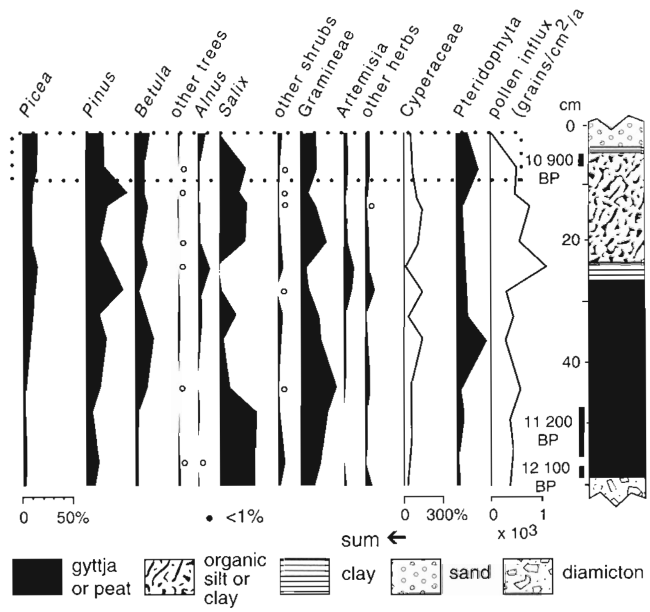



*Figure 112. Simplified pollen diagram of Benacadie Point [19] subtill organic bed (dotted box indicates Younger Dryas cold period) (from Mott et al., 1986).*

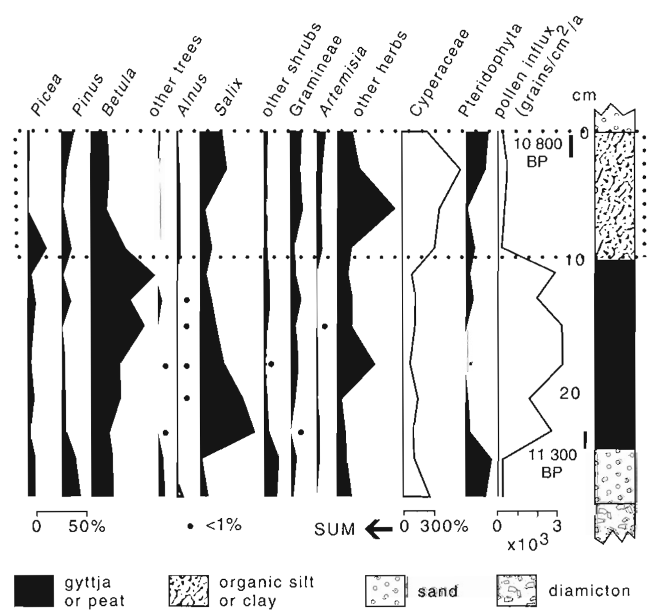



*Figure 113. Simplified pollen diagram of peat bed in sand outwash(?) near Campbell [53], George Bay coast (dotted box indicates Younger Dryas cold period) (by R.J. Mott, in Grant, 1987).*


*Figure 114. View of late glacial organic beds buried by till near Amaguadees site [23]. GSC 202188-N*

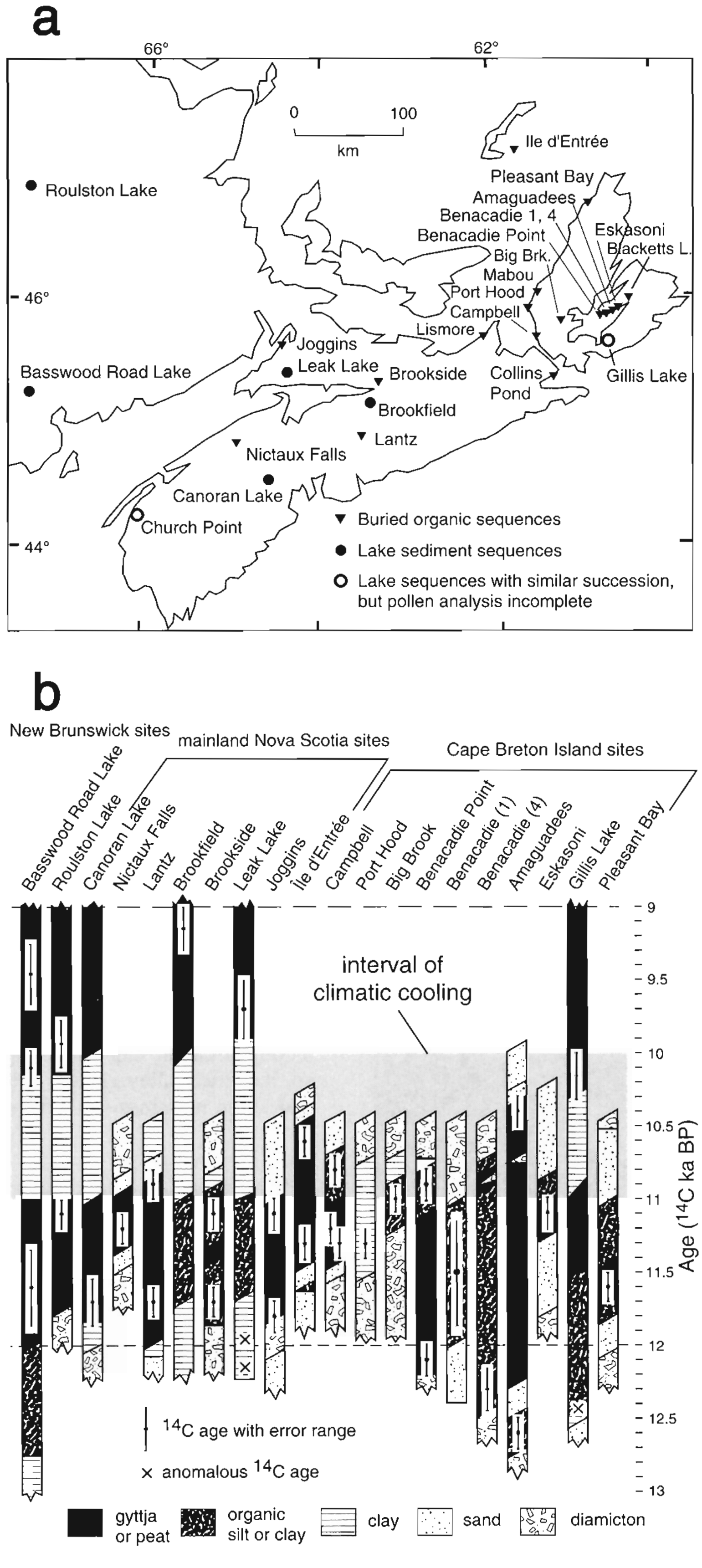



*Figure 115a.* map and *b.* summary lithostratigraphic correlation diagram of organic sequences which record the climatic reversal that occurred 11-10 ka BP during Younger Dryas cold period (adapted from Mott et al., 1986).

500-1000 years until ≈10 ka when normal climatic recovery and afforestation resumed. The similarity of ages and climatic trends in these widely spaced occurrences points to a sharp deterioration of the Cape Breton Island climate during the period 11-10 ka which caused lakes to accumulate more detritus than organic and organic terrains to be buried by colluvium and alluvium. Mott et al. (1986) did not affirm, however, that any possible remnant glaciers in Cape Breton Island readvanced or that new glaciers formed.

To explain the sedimentation changes, Mott et al. (1986) suggested that the climatic cooling caused vegetation to thin out, thus exposing soils which liberated sediment that in turn caused floodplains to aggrade, a mineral layer to accumulate in lakes, and a windblown layer to be deposited locally. Frost activity may also have increased, because colluvium evidently moved downhill, burying soils and organic terrains. Organic layers overlain by till and outwash show that some remnant glacier ice was still present in other areas and that the cold period caused it to readvance. When the climate ameliorated, forest returned, slopes became stabilized by increased vegetation cover, and organic sedimentation resumed. Mott et al. (1986) correlate the cycle with the Aller-d/Younger Dryas climatic reversal in northern Europe and suggested that it was caused by a southward shift of the Polar Front in the north Atlantic Ocean.

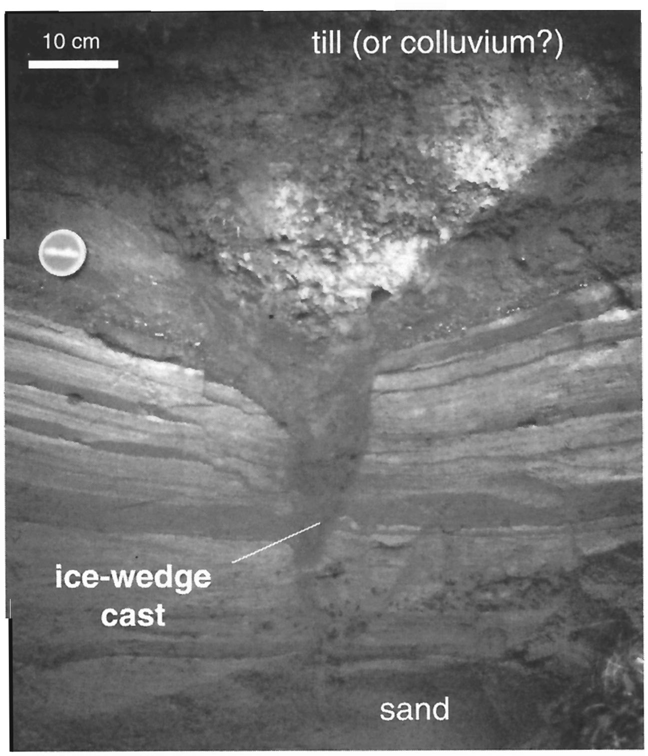



*Figure 116. Ice-wedge cast in sand, overlain by till or colluvium near Broad Cove Marsh [61]. GSC 203504-T*

The extent of remnant glaciers during the 11-10 ka cold period of the Late Wisconsinan can be inferred from various direct and indirect evidence. Ice probably filled most, if not all, of Bras d'Or Lake basin. It covered the eastern part of the basin because it deposited till over peat near Blacketts Lake [31] where it deposited till over peat and laid down varved sediment in Upper Gillies Lake [30], but it probably did not reach beyond Sydney because the Kilkenny Lake core [80] has a basal pollen assemblage dominated by shrubs and herbs similar to those dating ≈12 ka elsewhere on the lowlands (J.G. Ogden III, pers. comm., 1992). Its northern flank did not extend far into the Middle River valley because basal sediment at Third Lake O'Law [81] gives a 12.4 ka age (R.J. Mott, pers. comm., 1992). It was evidently present in the western part of the basin because it dammed a proglacial lake at Big Brook quarry [49] and in Inhabitants Lowland. Conclusive evidence has been presented (Stea and Mott, 1989, 1990) that lowland ice remained on Cape Breton Island during late glacial time and that its western flank readvanced westward onto mainland Nova Scotia at least to Collins Pond [82] because R.R. Stea and co-workers found till overlying a peat bed that gave an age of 11.8 ka. Probably this same ice also readvanced northward into George Bay at least as far as Judique where late glacial peat has been beneath till (R.R. Stea, pers. comm., 1992). The southward extent of this late glacial Bras d'Or ice mass is unclear, but it evidently reached the Atlantic coast over a wide area because sedimentation at Gillis Lake [40], Salmon River Lake [41], and Hector Lake [92] apparently did not begin until about 10 ka. Farther north beyond the lowlands, conditions during the cold period probably involved frost action, rather than glacier overriding, as suggested by an ice-wedge cast (Fig. 116) and a possible frost boil in bedrock (Fig. 117). Palynology and age estimates in progress on a core in Cormier Lake [90] might clarify this point. Given the size of the glacier which the cold period evidently sustained on the lowlands, it is assumed there was a companion ice cap on the highlands plateau (Phase H). If so, its outlet valley glaciers probably built the end moraines and kame moraines at Chéticamp River (Fig. 61), Pleasant


*Figure 117. Superficial fan fold developed in sandstone bedrock either by subglacial shear or by ground-ice growth, near Terre Noire. GSC 120219*

Bay [73] (Fig. 46), Pollett Cove [75] (Fig. 60), Aspy Fault [3] (Fig. 52), Aspy scarp (Fig. 65), Rocky Bay, Indian Brook, Dundas Brook, and Baddeck River. Indirect support for late ice on the highland plateau is the 9.0 ka basal date at Wreck Cove pond [9]. These indications are integrated into a reconstruction of hypothetical glacier distribution that is given in the section on Cenozoic history.

The many late glacial organic sequences demonstrate a cause and effect relationship between climatic change and geological processes which has an important application to understanding the longer term glacial/nonglacial succession in the Cape Breton Island region. Specifically, the change from organic to inorganic sedimentation which occurred 11-10 ka, as documented by Anderson (1985) and Mott et al. (1986), underscores the profound effect that sea-surface conditions in the north Atlantic Ocean have on climate in the Nova Scotia area and the Atlantic Provinces region in general. This finding is significant in the broader context of seeking parallels between the long-term record of average global temperature and the sequence of glacial events in Cape Breton Island, because it is necessary to bridge gaps in the fragmentary geological record and to estimate their relative importance. In this connection, it is significant that several authors have noted that the eastern Canadian Maritime region is particularly subject to a rapid and vigorous glacial response to conditions in the north Atlantic Ocean (Johnson and McClure, 1976; Ruddiman, 1980; Ruddiman and McIntyre, 1979, 1981). The Allerød-Younger Dryas temperature oscillation in the Atlantic Canada region is estimated at 1.5°C-2°C (Walker et al., 1991), yet was evidently sufficient to cause glaciers to readvance and vegetation to revert from forest to tundra. It therefore seems reasonable to suppose that the much larger temperature swings of about 6°C-8°C which occurred during the last glaciation, as recorded by marine oxygen isotopes, can be used as a guide to the possible major alternations of glacial and nonglacial conditions in the Atlantic Provinces region. The record of Late Cenozoic temperatures in the North Atlantic is therefore used to help reconstruct the Quaternary paleogeography of the Cape Breton Island area as outlined in the section below.

## CENOZOIC HISTORY

From the evidence presented in this report, a succession of events is traced from the earlier part of Cenozoic time up to the present. The history thus reconstructed deals primarily with crustal movement, erosional episodes, glacial and interglacial periods, climatic variation, and sea level changes. The recorded events fall into three main time periods – pre-Quaternary and early Quaternary, the last interglaciation, and the last glaciation – and each had a different suite of events. The early history can be only broadly sketched and is limited to the erosional development of the major landscape features, uplands and lowlands, which are interpreted to belong to a regional suite of degradation surfaces or peneplains. An attempt is made to link this terrestrial erosional history to the largely depositional regime which have built the adjacent continental shelf. The latter part of the record, spanning Late Quaternary time, is based on the stratigraphy of the complex of mapped surficial deposits. Climate and sea level of the last or Sangamonian interglaciation, with its wide climatic variations and associated sea level changes, is documented by a suite of scattered, Th/U dated organic beds and associated nonglacial sediments. The last or Wisconsinan glaciation involved several major ice-flow phases that are recorded by a stack of three or more till sheets, and associated glacier erosional features. Postglacial time is represented by sporadic cover of fluvial, colluvial, and organic deposits. Associated sea level changes are inferred mainly from surrounding areas.

### *Pre-Quaternary and earlier Quaternary events*

Lacking a depositional record of events prior to the last interglaciation, the earlier Quaternary history can only be a matter of general speculation based on the major erosional features of the landscape – the paleoplains and peneplains described in the section on Physiography and geomorphological history. These can be viewed in the context of the long term sequence of crustal movements which created the continental margin, and which are recorded in the development of the thick sedimentary wedge that forms the continental shelf adjacent to Cape Breton Island. Because the shelf was constructed from detritus eroded from the hinterland, of which Cape Breton Island is the nearest part, the better-documented record of offshore sedimentation and crustal warping provides valuable clues and forms the only available framework for interpreting onshore physiographic evolution. Information on offshore events, used in the following interpretation of Cape Breton Island physiography, is extrapolated primarily from offshore structural data and paleogeographic reconstructions by King (1970, 1976), King et al. (1974), Jansa and Wade (1975), Gradstein et al. (1990), and Keen and Beaumont (1990). The continental margin came into being when the American and Eurasian crustal plates separated in Jurassic time. Since then, because of cooling, crustal stretching, isostasy, and sediment loading, the margin has been generally tilting seaward by means of subsidence of shelf areas and uplift of land areas. Over the last 210 Ma, the shelf has subsided at rates of 3-10 cm/ka (average ≈5 cm/ka) to accommodate the several kilometres of Mesozoic and Cenozoic rocks that underlie the Scotian Shelf. The subsidence rate has generally decreased exponentially, except for the last few million years of Cenozoic time, when there was a rapid increase to ≥6 cm/ka, perhaps because of tectonism and/or sediment loading partly attributable to Quaternary erosion (Gradstein et al., 1990). Meanwhile, the hinterland was uplifted as much as 1 km (Keen and Beaumont, 1990) causing erosion which supplied much of the necessary sediment. With the increasing shelf sediment load, the isostatic downwarp expanded landward, the basement unconformity developed hingelines, and the shelf sequence onlapped the land. As crustal movement varied with subsidence, tectonism, and eustasy, unconformities developed offshore and planation levels (peneplains) were cut onshore; both became tilted seaward (Fig. 10, 11).

The major planation levels (paleoplains and peneplains) which are recognized in Cape Breton Island can possibly be correlated with some of erosion levels mapped and dated in the offshore sequence. Paleoplains of Carboniferous age in the Ingonish area and the paleoplain on the south coast which appears to be an extension of the Jurassic basement unconformity have been mentioned in Physiography and geomorphological history. The three inset and progressively less tilted planation surface in the Cape Breton Island area alluded to in this report, the upland and highland plateau, the lowlands, and the submarine channels and basins, are considered to represent periods of stability which resulted in widespread erosion, followed by uplift and downcutting which released sediment until the next period of stability occurred. The peneplains on land are therefore tentatively correlated with the younger offshore unconformities mapped by King et al. (1974), whereas the periods of terrestrial downcutting prior to formation of the next planation level correlate with the sedimentary units deposited upon the offshore unconformities. The oldest and highest peneplain, which forms the upland and highland summits, is considered to have been uplifted and dissected in mid- to late Tertiary time (Mathews, 1975) so it must date from an earlier period of relative stability, perhaps the one when a major Tertiary erosional surface was cut across Miocene beds on Scotian Shelf ("event 2" of Keen and Beaumont, 1990, p. 426). The two lower planation levels, the Bras d'Or Lowlands and submarine channels, must be younger and therefore probably correlate with the two major unconformities in the Pliocene part of the offshore sequence which King et al. (1974) place in the period from Late Tertiary to Early Pleistocene. The major landscape regions of Cape Breton Island had therefore largely assumed their form in pre-Quaternary time and were able to influence the course of Quaternary glaciation and marine transgression.

The impact of Quaternary processes on the Cape Breton Island landscape can be judged in terms of the extent to which pre-Quaternary landforms have been modified. Only the net or cumulative effect can be identified; at present there is no data to permit apportioning the effect to any particular time intervals. The main difference between Tertiary and Quaternary time in terms of the operative effects on landscapes was the sudden switch to a greatly fluctuating climatic regime marked, at this latitude, by repeated glaciation. Assuming that glaciation was the major geomorphic agent, it is useful to consider the number of possible glacial/interglacial cycles during Quaternary time. The record of global average temperature over the last 3 Ma is a close approximation of ice volume and thereby also of sea level position. The record presented by Ruddiman and Raymo (1988) shows that prior to 2.4 Ma temperature was higher, more equable, and less variable; this was the regime in which the peneplains were cut and the deep saprolite developed. After 2.4 Ma, there were 30-50 temperature fluctuations corresponding to continental-scale ice-sheet growth ($\approx$40 with a 41 ka period to 0.7 Ma, then 7 with a $\approx$100 ka period). Each of these major glacial periods had several major advances and retreats. Given that Cape Breton Island was completely ice covered and experienced several major advances during the last glacial period, it is assumed that it was similarly affected by the 30-50 previous glacial/interglacial cycles during which times it was alternately attacked by ice, running water, and marine erosion.

The volumes of glacial, fluvial, and marine erosion resulting indirectly from the dramatic swings of Quaternary climate, as estimated in the section Physiography and geomorphological history, are found to be roughly comparable, although very unequally distributed over the landscape. The paleoplains and the summit peneplain, being developed mainly on resistant crystalline rocks, have remained largely intact, with few or no glacial basins having been excavated and little till and other sediment deposited, thus attesting to the generally minimal effect of Quaternary erosion and deposition, at least upon resistant rocks at higher elevations. Masswasting has also been relatively ineffectual on summit areas, presumably because of the already mature slopes, but has participated actively in slope recession along river valleys and coastal cliffs. Ignoring the innumerable smaller glacial basins on the island, the main erosional effect in the Cape Breton Island area of all the glacial periods combined has been the creation of Bras d'Or Lake basin which amounts to at least 90 km$^3$, not counting the volume occupied by sediment. The major effect of fluvial action, on the other hand, has been the creation of the dendritic system of gorges which dissect the upland margins; their volume is estimated to be $\geq$100 km$^3$. With regard to marine action, the main effect has been the production of a very low relief littoral abrasion surface extending seaward to about 100 m depth, the average 100 m glacio-eustatic range. This planation surface evidently represents the cumulative erosional effect of successive post-glacial/interglacial marine transgressions which, coupled with long term crustal subsidence, ensured that each transgressive maximum reached farther inland than previous ones. The net amount of this littoral planation is substantial. A minimum of 10 m of rock (the present average cliff height) has been removed over a 10 km wide swath around the $\approx$500 km perimeter of the island, a volume that amounts to about 50 km$^3$. While admittedly crude and approximate, these estimates seem to suggest that erosion during glacial times has been roughly equal to erosion during interglacial times (combined fluvial and marine). This finding may be regarded as an indirect consequence of Porter's (1988) concept of "average Quaternary conditions" which proposes that, if conditions at 11 ka were the Quaternary average, then glacial and interglacial intervals were of approximately equal total duration.

These suggestions for the local Cape Breton Island area are broadly supported by findings from deep offshore sedimentary sequence over a much wider area. They show that sediment delivery to the deep sea varied little over the long term, except for rare spikes during the three or four greatest glaciations (Alam and Piper, 1977; Alam et al., 1983). So, without evidence to the contrary, it is therefore presumed that, prior to the last interglacial and glacial periods for which there is a stratigraphic record, the work accomplished by the various erosional agents throughout Quaternary time was steady and cumulative and that the topographic relief thus created developed slowly and had progressively greater influence on the course of glaciation and sea level change. The work of previous glaciations and interglaciations prepared the landscape and thus helped influence the course of the events discussed next.

### Pre-Sangamonian glaciation

There is surprisingly little evidence of the Quaternary prior to the last interglaciation. No prior interglacial deposits are recognized and only at Dingwall [4] is there till which may possibly underlie last-interglacial deposits; however, the true stratigraphic position of the till cannot be affirmed because of disturbance due to karst development. There is thus no reliable evidence in Cape Breton Island of pre-Sangamonian glaciations (and interglaciations).

## *Late Quaternary events*

### Sangamonian interglaciation (oxygen isotope stage 5)

*Marine transgression*

A shoreline related to the marine transgressive maximum of the last interglaciation is represented by a fossil intertidal rock bench and its associated littoral sediment, now several metres above present tide level. The bench is a smooth, planar, and almost level notch with a width ranging from 1 m to several kilometres which rings the island. It is identical in every respect to the modern intertidal rock bench, and on those morphological grounds it is interpreted as an emerged littoral feature. The associated sediment has the roundness, sorting, and good horizontal stratification that is typical of littoral sediment. Percussion marks on included boulders is a further indication of the beach environment.

The emerged littoral bench can be referred to the earlier part of the last interglaciation by its stratigraphic setting beneath, or in association with, organic deposits that are assigned chronometrically or palynologically to the last interglaciation (sensu lato), notably at Green Point [57], East Bay [27], and Bay St. Lawrence [1]. The only fossiliferous marine sediment of possible interglacial age which may be associated with the bench is the silt with marine diatoms and wood debris at Northeast Mabou River [58].

Two other reasons for confidence in the assignment of the bench to the last interglacial marine transgressive maximum is its attitude with respect to present (interglacial) sea level, compared to päst (glacial) shorelevels, and its elevation compared to known last interglacial shoreline elsewhere in the world. First, it is obvious that the bench's slope of <5 m over 150 km is about a tenth of that of late glacial shorelines, which rise northward across the region 200-250 m over 600 km. Secondly, the bench elevation falls within the 2-9 m (average 6 m) elevational range of oxygen isotope substage 5e (125 ka) coral reef levels in essentially stable areas worldwide (Marshall and Thom, 1976; CLIMAP Project Members, 1984).

The inland upper limit of the Sangamonian transgression throughout Cape Breton Island may be sketched (Fig. 118) by extrapolating between the few places where its inner notch his visible, as along the present shore of Louisbourg area (Fig. 39b, c), St. Paul Island (Fig. 39a), the northwest coast (Fig. 39f), and Money Point (Fig. 39h). In the intervening segments, a very rough approximation of its position is made with reference to topography, bedrock elevation, and the thickness of till cover. From these indications, the Sangamonian transgressive maximum was evidently more or less coincident with the present shoreline where it lies along the upland and highland margins, but reached up to several kilometres farther inland in lowland areas, particularly where underlain by non-resistant rocks and where covered by thick till. Cape Breton Island may thus have been divided into several islands during the highest last interglacial sea level. The interglacial paleoshore probably lay at the foot of the highland escarpments. Lake Ainslie basin, Margaree valley, and possibly also Mabou estuary were probably arms of the sea which may have connected to Bras d'Or basin. Aspy, Ingonish, Judique, and Inhabitants lowlands likely were extensively submerged, and Isle Madame was much reduced in size. The Bras d'Or basin was much larger by inclusion of most of its bordering lowlands, making islands of adjacent uplands such as Boisdale Hills, and it probably had two additional marine connections at St. Peters and Sydney. The southern lowland was extensively submerged; Loch Lomond was probably a bay and Mira Lake a seaway.

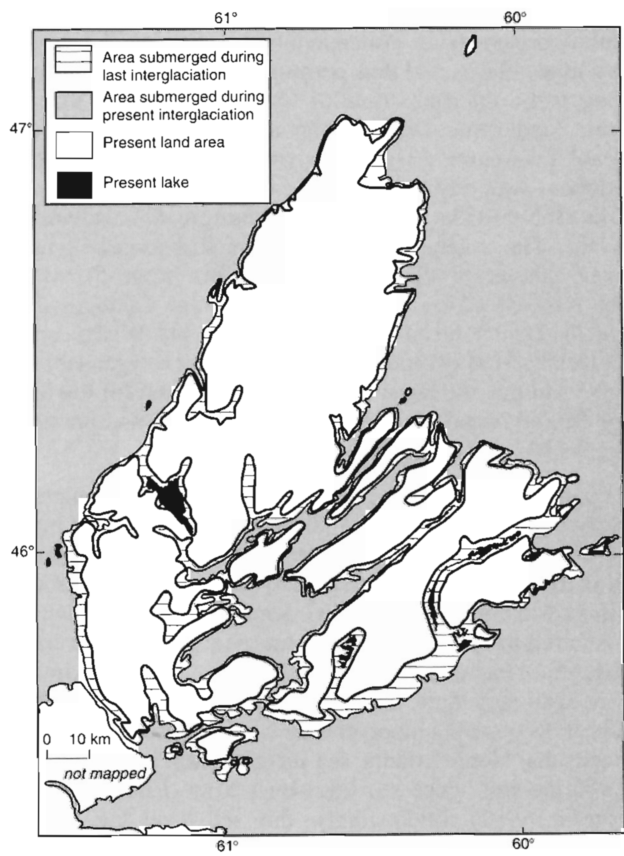



***Figure 118.*** *Inferred position of coastline during highest sea level of last interglacial period (based on an unpublished map by H.L. Cameron).*

It is useful to compare the highest marine shoreline position attained during last interglaciation with that which may likely be reached during the present interglaciation. The last interglacial shore may be several metres above present, but at the rate of 18-70 cm-per-century that sea level has been rising for the last few millennia (see section below on Recent relative sea level rise and crustal subsidence) it will reach the level of the bench within 1-3 ka and thus replicate its previous interglacial position (Grant, 1980). (It may also be noted that this several thousand year lag of the sea level maximum relative to the thermal maximum mirrors the lag that also occurred during the last interglaciation.)

As average global temperature decreased during the transition from oxygen isotope substage 5e to 5d, water was transferred from the world's oceans and stored as ice on land areas. In this way, ocean levels fell, the sea regressed from its maximum stand, and the Cape Breton Island intertidal bench emerged. That relative sea level did indeed fall, rather than rise such as by glacio-isostatic crustal depression due to possible glacier growth in the region, is shown by the red ferruginous weathering (rubefication) on the bench surface and in the overlying beach gravels in a few places.

*Climatic variation during oxygen isotope stage 5*

The record of terrestrial conditions during and after formation of the emerged rock bench is contained in eleven occurrences of subtill organic beds. Pollen in those deposits reveals vegetation assemblages and thus permits a range of climatic conditions to be inferred, some of which were cooler than at present, and some warmer than at present or during the thermal maximum ("Hypsithermal") of the present interglaciation. Age estimates of these units range from at least 126 ka to about 62 ka, and perhaps as young as 47 ka (although this latter date is suspect), thus indicating that some belong to various phases of the last interglaciation (stage 5), while others may extend into the Wisconsinan (stage 4 and possibly stage 3). Their total range shows that the pre-Wisconsinan nonglacial period spanned the full duration of oxygen isotope stage 5 and that the latest episodes of glaciation (of lowland Cape Breton Island) did not begin until stage 4 and possibly stage 3 (Wisconsinan time).

Four of the several interglacial organic beds have a relationship to the bench that allows the three thermal maxima, and particularly the warmest (substage 5e), to be bracketed chronologically with respect to the sea level maximum of the last interglaciation. The Green Point [57] beds, which are referred to the thermal optimum (5e), appear to be truncated by the bench, thus indicating that the sea level maximum followed or lagged the thermal maximum. This relationship offers stratigraphic support for the finding based on oxygen isotopes that North Atlantic sea surface temperatures lagged ice volume and hence sea level by 2-5 ka (Ruddiman and McIntyre, 1981). Interestingly, this sea level lag is also occurring during the present interglaciation; witness the fact that the sea is still rising ≥6 ka after the Holocene thermal maximum ("Hypsithermal") and is actually eroding coastal organic deposits of that period. The last-interglacial beds which overlie the bench or its associated littoral sediments (Bay St. Lawrence [1], Castle Bay [25], Moose Point [83]) belong to the two later warming periods (substage 5c and/or 5a). They confirm that the bench is older than about 85 ka and further show that, after the warmest climate and highest sea level of the last interglaciation, climate remained cooler and sea level lower than at present.

During the last interglaciation, climate on Cape Breton Island went through two, and possibly three, major warming cycles. The sedimentary record of these cycles is not complete; they are inferred from separate and distinct organic units with different ages. The temperature conditions represented by the vegetation recorded in the pollen is linked to form a series of three fluctuations which correspond to the major intervals of the stage 5 oxygen isotope record. The first and warmest phase of the Sangamonian Interglaciation on Cape Breton Island, represented notably by Green Point and by East Bay unit I, had thermophilous hardwood and white pine forest vegetation indicating a climate that was warmer than the Holocene thermal maximum ("Hypsithermal") period and is correlated to oxygen isotope substage 5e about 125 ka. The second warming episode, not dated and possibly represented only at Northeast Mabou River, had boreal forest and forest tundra vegetation and is tentatively correlated to oxygen isotope substage 5c (about 105 ka). Indeed, the virtual absence of any organic beds which can be dated and/or related to substage 5c is circumstantial evidence that conditions were not conducive to organic deposition and preservation. Perhaps the climate was too cool or the area was glacier-covered. The third warming episode, best recorded by East Bay unit II, gives ages of 84.2 ka and 86.9 ka and is thus reasonably reliably correlated with substage 5a. It produced mixed hardwood and coniferous forests similar to the present vegetation and thus indicates that climate during the final warm episode of the last interglaciation evolved to conditions comparable to the present. The relative warmth of these three possible oxygen isotope stage 5 climatic cycles (the first being warmer than present, the last being comparable to the present, and the middle period being much cooler or even glacial) broadly matches the temperature record of this period provided by oxygen isotope ratios (e.g., Shackleton, 1987).

Whether glaciers may have formed in the area during the coolest periods of the last interglaciation (oxygen isotope stage 5) cannot be definitively answered from stratigraphic evidence, although the till-like sediments in the Dingwall [4] and Leitches Creek [32] organic sequences may provide evidence. Certainly the lack of demonstrable substage 5c deposits may be an indirect argument for glacier cover. The deep sea record of global average temperature offers theoretical evidence that glaciation may well have occurred in this area during oxygen isotope stage 5. Numerous authors (e.g., Ruddiman and McIntyre, 1976) have shown that sea-surface temperatures in the northwest Atlantic Ocean during the two main cool periods of the last interglaciation (substages 5b and 5d) were at least as cool as conditions during the late glacial period about 13-11 ka BP when, as will be shown below, glaciers covered much of Cape Breton Island. To the extent that north Atlantic Ocean sea surface temperatures are reflected in the Cape Breton Island area, theoretically, there could have been glacier growth in the area during the cooler parts of the last interglaciation (stage 5).

**Wisconsinan (oxygen isotope stages 4, 3, 2)**

The climatic deterioration that began in late Sangamonian time (sensu lato) continued with fluctuations into Wisconsinan time. The record at East Bay and Castle Bay sites, if correctly dated, shows that, in southern lowland Cape Breton Island, ice-free conditions probably lasted until as late as 62 ka (stage 4). Moreover, vegetation had become tundric, indicative of periglacial conditions leading up to the ultimate arrival of Wisconsin ice which deposited the tills over the organic beds. The Bay St. Lawrence peat age of 47 ka is considered to be minimal, not only, as Stea et al. (1992b) point out, because of possible postdepositional uptake of uranium from pegmatitic debris in the enclosing sediments, but because its pollen signature matches that of some oxygen isotope stage 5 beds and because it bears a close relationship to the raised interglacial marine bench. For the purposes of this report therefore, Wisconsin glaciers are considered to have reached the island sometime in stage 4, although an ice growth probably began earlier on the highlands plateau, and glaciers may not have filled up the deep submarine channels and the adjacent coastal areas until much later.

Several Wisconsinan glacial events (phases A-H) can be inferred from the various patterns of glacial markings (Fig. 55) and from the sequence of superposed drift sheets of differing provenance. Significant additional support for this sequence of ice flows and their corresponding till sheets in the Yava [94] and Fourchu-Gabarus areas is provided by recent till provenance and till geochemistry studies (McClenaghan and DiLabio, 1992; McClenaghan et al., 1992; MacDonald and Boner, 1993). This "erosional" and depositional glacial sequence begins after emergence of the Sangamonian shore bench and deposition of the organic beds. An approximate sequence of glacial advance and retreat phases can be reasonably deduced from superposition and crosscutting relationships, as well as from the pattern of recessional features. Where age relations are ambiguous, the sequence is decided on simple glaciological grounds, for example, ice should accumulate first on higher areas and be topographically controlled during early and late stages of advance. Eight ice-flow phases (and one possible deglacial interval) are recognized and most can be fairly readily integrated with the sequence presented for mainland Nova Scotia and surrounding region (e.g., Rampton et al., 1984; Stea et al, 1992b). Glacier cover was probably continuous throughout the whole sequence; however, there may have been a period of ice retreat and subaerial weathering between Phase C and Phase D. The changing trajectories probably reflect periodic steady-state conditions within a spasmodically changing continuum. Most phases are considered to be individual events, although several were probably coeval. For this reason, and in part for convenience, more than one flow phase are shown on each part of Figure 119. The phases are as follows:

A) radial outflow from a probable early highlands ice cap (Fig. 119a);

B) west to east (070°-115°) flow of a major external ice sheet over the southern Bras d'Or Lowlands (Fig. 119a);

C) southward (160°) flow onto the Gulf of St. Lawrence, Aspy Bay, and Louisbourg coasts, and eventually southeastward (120°) over the highlands plateau (Fig. 119b);

- recession and subaerial weathering(?) -

D) strong northward spreading flow (340°-030°) over the southern lowlands from a source on Scotian Shelf (Fig. 119c);

E) late radial ice dispersal from a centre over Bras d'Or basin (Fig. 119d);

F) eastward (050°) advance onto the west coast by Gulf of St. Lawrence (Fig. 119d);

G) radial flow of lowland ice during concentric retreat, with a late readvance 1110 ka (Fig. 119d); and

H) radial flow of a highlands ice cap during final retreat (Fig. 119d).

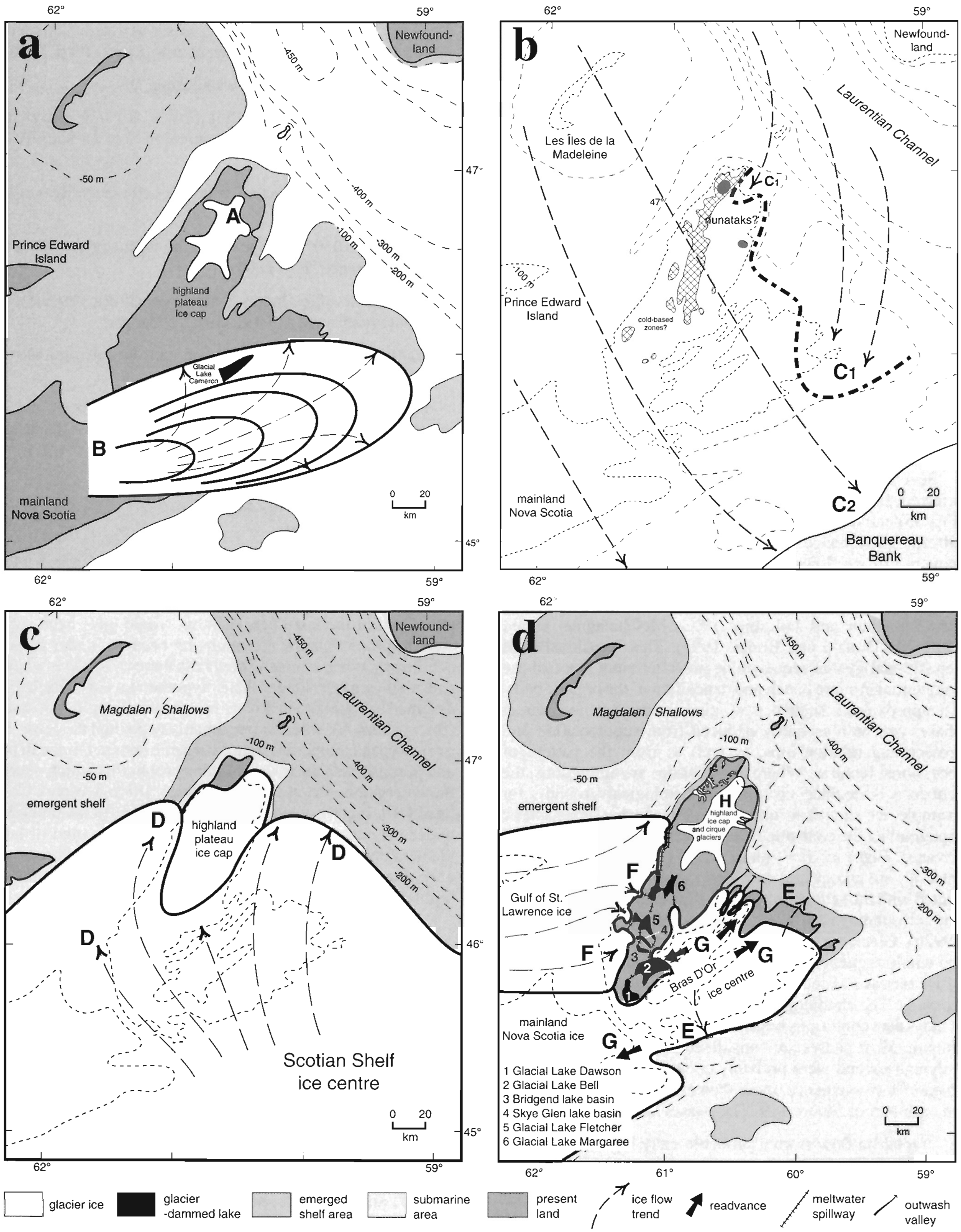
Details of these glacial flow phases, their respective tills, and their associated paleogeographic features are discussed below.

*Phase A – inception and radial spread of a highlands ice cap*

As the climate gradually cooled at the end of the last interglaciation, it is assumed that the first accumulation of glacier ice was on the northern plateau (Fig. 119a). A minor cooling at the end of the last interglaciation would have been sufficient to promote snow accumulation because, under present conditions, twice as much snow falls there as on the lowlands and it stays much longer. This hypothetical early ice cap is assumed to have spread downslope onto the fringing lowlands (flow pattern A), where it presumably deposited the lower till over oxygen isotope substage 5a organic beds at Dingwall [4], and possibly also the lower (highland rock) tills which overlie the emerged bench at Morrison Brook [8] and Wreck Cove head [10]. It is assumed that cold-based conditions prevailed under the central portions of this early ice cap (and of later glaciers that covered this area) because no glacial erosional and depositional features were formed in the central highlands area.

*Phase B – foreign ice advances eastward over the southern lowlands*

The first main Wisconsinan glacial event recorded on the southern lowlands was an eastward flow evidently from a source to the west of Cape Breton Island (Fig. 119a). This movement is recorded by the first or eastward-trending set of striations in the crosscutting sequence and by the till overlying the organic beds in the lowland area which is derived from western and offshore sources. The distribution of these indicators shows that this ice covered almost all of the southern lowlands, but apparently did not override the northern uplands and highlands, perhaps because a local ice cap had already developed over that area. Geographic variation in the modal direction of this movement suggests that the ice seems to have followed the trend of the coast from Canso Strait to


a
Newfound-land
-450 m
-50 m
-400 m
-100 m
-300 m
-200 m
A
Prince Edward Island
highland plateau ice cap
Glacial Lake Cameron
B
mainland Nova Scotia
0 20 km
b
Newfound-land
Laurentian Channel
Les Îles de la Madeleine
C1
nunataks?
-100 m
Prince Edward Island
cold-based zones?
C1
mainland Nova Scotia
C2
Banquereau Bank
c
Newfound-land
Magdalen Shallows
Laurentian Channel
-450 m
-100 m
-400 m
-50 m
emergent shelf
D
highland plateau ice cap
-300 m
-200 m
D
D
Scotian Shelf ice centre
d
Magdalen Shallows
Laurentian Channel
-450 m
-100 m
-400 m
-50 m
emergent shelf
H
highland ice cap and cirque glaciers
F
6
Gulf of St. Lawrence ice
5
4
E
F
3
G
G
2
1
Bras D'Or ice centre
mainland Nova Scotia ice
E
G
1 Glacial Lake Dawson
2 Glacial Lake Bell
3 Bridgend lake basin
4 Skye Glen lake basin
5 Glacial Lake Fletcher
6 Glacial Lake Margaree
62°
61°
60°
59°
47°
46°
45°
glacier ice
glacier -dammed lake
emerged shelf area
submarine area
present land
ice flow trend
readvance
meltwater spillway
outwash valley

Sydney area with a gently curving trajectory. Moreover, from the crosscutting relationships of striations, movement at first was east-southeastward (≈115°), then changed to east-northwestward (070°). These facts may suggest that the ice at first followed the eastward trend of Chedabucto Trough, then as it thickened on the shelf, expanded northward over the island.

Sea level during the advance of this ice into the area is placed at or below -100 m, approximately at its supposed local Wisconsinan sea level minimum (Fader et al., 1982; King and Fader, 1986). That this paleoshoreline is at about the same level as the global eustatic sea level minimum 120 m below present sea level in nonglacial area perhaps reflects the fact that little glacier load was present regionally and that the crustal lag time for full isostatic response is about 5000 years.

The eastward flow had several important effects. Firstly, because the flow was generally parallel to the structural elongation of the weak-rock lowlands in the Bras d'Or area, it probably helped excavate the lake basins. Indeed, if this flow pattern was repeated often during the Quaternary, it may have been the main cause of the basins. Secondly, in erosional terms it was evidently powerful and/or long-continued because it carved the bedrock into roches moutonnées with large grooves and, in depositional terms, it produced the thickest deposit, a reddish silty till derived from Carboniferous redbeds to the west. (The grey, silty locally-derived till with an eastward pebble fabric that underlies the red till in two places (McClenaghan and DiLabio, 1992) was probably deposited as the ice first moved over the landscape and was able to generate a till directly from the local bedrock, before it eventually brought in the masses of red foreign drift). Thirdly, this ice flow produced the large east-west till ridges as giant drumlinoid forms, as shown by the fact that its early 100°-115° trend is parallel to their long axes and by strong eastward-directed pebble fabrics (McClenaghan and DiLabio, 1992). Fourthly, as this ice advanced into East Bay, it would have blocked the drainage, ploughed up marine mud from the sea floor, and transformed the saltwater arm into a proglacial lake into which deltas would have been built and the mud redeposited up to 20+ m elevation, as at Castle Bay [25]. This hypothetical proglacial lake in East Bay is here provisionally named glacial lake Cameron (Fig. 119a) in honour of H.L. Cameron who first postulated ice-dammed lakes in other parts of Cape Breton Island. If the Castle Bay silts are correctly pollen-correlated to the dated East Bay sequence, glacial lake Cameron was in existence by 62 ka BP, thereby providing an approximate maximal age for the beginning of major ice sheet glaciation in Cape Breton Island.

This early eastward flow over southern Cape Breton Island evidently was a continuation of the early generally eastward flow first mapped in southern New Brunswick and northern Nova Scotia by Chalmers (1895) and recently integrated into a single regional flow event that affected all of mainland Nova Scotia and southern New Brunswick (Rampton et al., 1984; Ice Flow Phase 1A of Stea and Brown (1989) and Stea et al., 1992). It apparently originated west of New Brunswick, but as it did not introduce Laurentide erratics, it must have emanated from a burgeoning Appalachian ice cap centred in northern New England. The direction of advance into the area is greatly discordant with most later flows, so this flow phase may be viewed as a temporary transition to a more stable ice sheet configuration. The extension of this flow virtually to the edge of Laurentian Channel apparently indicates that Gulf of St. Lawrence was not filled with Laurentide Ice Sheet. This in turn suggests that glaciers became well established in the Appalachian region well before glaciers from the Shield invaded the area. The peculiar tangential flow parallel to the regional slope remains a glaciological problem.

The timing of this major initial glacial phase cannot be established from available isotopic age estimates and stratigraphy, but the hypothesized relationships of oceanic conditions to ice growth (Johnson and McClure, 1976; Ruddiman and McIntyre, 1979; Ruddiman et al., 1980) would suggest that flow Phase B occurred during isotope stage 4, when sea-surface temperatures were their coldest and the Polar Front was at its farthest southward position. This phase is therefore referred to the advent of stage 4, sometime after 75 ka BP.

*Phase C – a regional ice sheet moves southeastward over the island*

The eastward invasion was followed by a major flow from the north (Fig. 119b). It apparently occurred in two stages: initially southwestward and southward (≥160°) onto parts of the eastern lowlands (Phase $C_1$), then later southeastward (≈140°) over the whole island, including the highlands (Phase $C_2$). It is assumed that sea level remained below -100 m, thus limiting the calving effects and keeping large areas of the shelf exposed for possible in situ build-up of glacier ice. The continuity of this ice-flow event is not clear because it is recorded only in noncontiguous parts of the island. Lacking evidence to the contrary, all are assumed to be the product of the same event. During this event, ice evidently filled

***a.*** *Glacial flow phases A and B. Flow phase A: formation of ice cap on northern highlands plateau; Flow phase B: eastward advance of ice from mainland Nova Scotia over southern lowlands.*

***b.*** *Glacial flow phase C: ice from Laurentian Channel lobing onto Sydney Shelf ($C_1$), thickening to regional ice sheet over Gulf of St. Lawrence which overrode highlands ($C_2$), except possibly for small nunataks.*

***c.*** *Glacial flow phase D: ice from Scotian Shelf advanced northward over the southern part of the island.*

***d.*** *Glacial flow phases E-H: establishment of a Bras d'Or ice dispersal centre (E, G), while Gulf of St. Lawrence ice abuts on the west coast (F), impounding proglacial lakes (Dawson, Bell, Fletcher, and Margaree) in western intermontane valleys, while a plateau ice cap persisted (H).*

***Figure 119.*** *Paleogeographic reconstruction of the main glacial configurations and their constituent ice-flow patterns.*

Laurentian Channel and covered St. Paul Island, leaving till over the interglacial bench and a strongly scoured terrain with roches moutonnées and striations oriented 120° (although later striations there range eastward to 070°). A glacier, evidently also from the Laurentian Channel, lobed onshore from the direction of the Sydney Shelf and crossed Scatari Island and the Louisbourg area in the easternmost part of the island. This Laurentian Channel glacier is assumed to have been the one that advanced onto northernmost part of Cape Breton Island and deposited the tills and the shell-bearing bed at Bay St. Lawrence [1] which, Stea et al. (1992b) referred to the period before 50 ka on the basis of amino-acid ratios. At this time, if not later during the maximal period of this phase, this ice probably also deposited on the western coastal lowlands the distinctive red and fine grained till derived from the Carboniferous redbeds underlying Gulf of St. Lawrence.

Eventually, this early southward-moving ice mass thickened and extended to overtop virtually all of the highest summits, including the northern plateau, and reached the Atlantic coastal area. During this process, its flow direction changed to southeastward and the ice moved without regard for topography. Although some areas bear no evidence of this ice-flow event, its parallel pattern of southeastward-directed scouring and till lineation is widely enough registered over the island, that it can be assumed virtually the entire island was covered during this phase. This major flow event was assumed because of its magnitude and extent to have climatostratigraphic significance and was informally termed the "Cabot Stade" because it apparently marked the only time during the last glaciation when Cabot Strait was "bank full" (Grant in Dreimanis et al., 1981; Grant and King, 1984). Although, the term has occasionally been used subsequently (e.g., King and Fader, 1986), it should probably be discontinued pending definition of properly dated lithostratigraphic equivalents.

Absence of evidence of this flow phase in the central parts of the northern highlands plateau gives important indications of possible ice behaviour during this period. Rare patches of lowland till and few lowland erratics were introduced. Perhaps the reason was that the lower debris-laden levels of the glacier remained below the steep escarpment, while the clean upper levels of the ice continued over the highlands. A more likely explanation of the widespread terrain on the highlands plateau (Fig. 119b) which lacks glacier erosional and depositional effects is that the ice was cold-based and frozen to its bed over this area. Cold-based conditions may have existed because the ice was thinner, because it overrode permafrost, and/or because it slid over the top of the pre-existing ice cap that is assumed to have been cold-based. Thus, the preglacial landscape with its regolith and saprolite soil was left essentially intact and only small patches of till were deposited. A similar suggestion has been made for analogous sharp discontinuities between freshly eroded glacial terrains and degraded, weathered glacial terrains in western Newfoundland (Gosse et al., 1993).

Small areas of the eastern plateau margin at Smoky Mountain (Fig. 120) and near Money Point above 360 m may have projected through the ice sheet as nunataks. This is inferred from their crumbling bedrock, lack of erratics, and sharp contact with adjacent glaciated terrain. These and other assumed nunataks that are inferred from glacier scouring limits at 300-400 m on the other side of Laurentian Channel in Newfoundland (Grant, 1987, p. 44), may thus serve to define the upper limit of the regional ice sheet in the Cabot Strait area during the Wisconsin maximum. Independent support for hypothetical ice-free areas on the highlands, if not also on the continental shelf beyond the ice sheet, during the last glacial maximum comes from evidence of disjunct plant and insect assemblages pointing to ice-age refugia (Belland, 1987; Hamilton and Langor, 1987). Ice thickness in the Cabot Strait area, 300 km from the maximum possible margin at the shelf edge, was thus 400 m (counting the 50 m local water depth) or, 800-900 m thick in nearby Laurentian Channel. This is much thinner than the theoretical 2000 m thickness of a temperate terrestrial ice sheet 300 km from its margin (Paterson, 1981); however, ice sheets moving on deformable beds of soft sediment have much lower basal shear stress values and hence thinner ice. For example, the northwest part of the Laurentide Ice Sheet was ≈300 m thick at the same distance from the margin (Beget, 1987). Moreover, that most of the ice sheet in the Atlantic region was grounded below sea level is an added factor contributing to lowered basal shear stress and hence a flatter profile. There is thus some glaciological support for supposing that ice surface elevations in the soft sediment submarine area of Cabot Strait could have been much lower than they would be where the ice was founded on resistant bedrock above sea level.

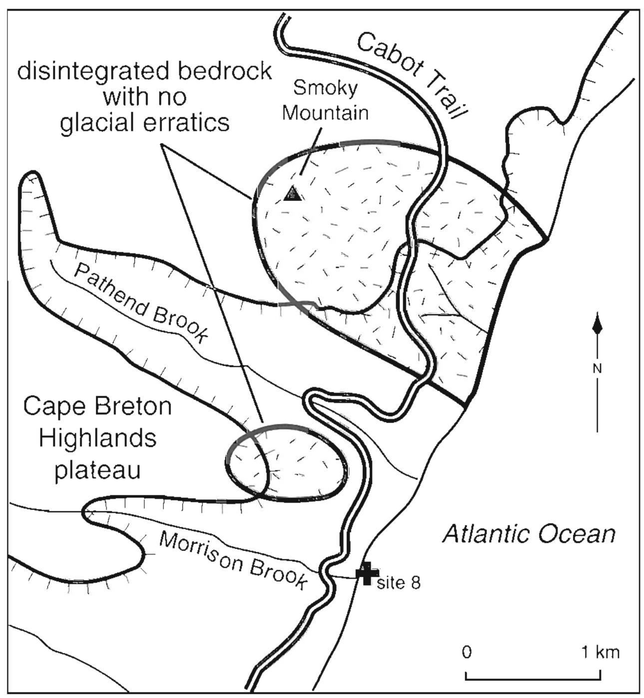



*Figure 120. Speculative ice-free areas (nunataks) in the Smoky Mountain area which, according to Belland (1987), served as biological refugia during the maximal glaciation (from Grant, 1987, p. 37).*

Over the southern or lowland half of the island the effects of the southeastward movement are varied. No trace of its passage is evident over the Bras d'Or Lake area. A possible explanation is that, after it moved off the plateau, it overrode a pre-existing ice mass which had already lobed into the area during the initial stages of the advance. Farther south, over the heavily drift-covered terrain along the Atlantic coastal area, the southeastward-moving ice sheet touched bedrock only locally and slightly rearranged the early east-trending drumlinoid till ridges, giving them an irregular morainic form, resembling ribbed moraine or rogen moraine.

This regional movement correlates best with Flow Phase 1B of Stea et al. (1992b), shown as sweeping more or less unidirectionally across Prince Edward Island and Nova Scotia and referred to as a combination of Appalachian and Laurentide ice masses. Alternatively, it may correlate with their later Phase 2 which emanated from the Escuminac Centre over western Prince Edward Island (Rampton et al., 1984) and had the same trend over Cape Breton Island as their Ice Flow Phase 1B. This latter is the movement first recognized by Honeyman (1890) as emanating from the southern Gulf of St. Lawrence and named the ("Acadian Bay Lobe") by Goldthwait (1924, p. 73).

The unidirectional southeastward trend of this ice flow without regard for the highland and lowland relief shows that it was part of a regional ice sheet coming from a source situated northwest of Cape Breton Island. This flow is therefore tentatively correlated with the Laurentide invasion which dispersed lower Paleozoic carbonate and Shield debris southward over Les Iles de la Madeleine (Dredge et al., 1992) and across Magdalen Shallows almost to Prince Edward Island (Loring and Nota, 1973). If Laurentide ice, however, did indeed move southeastward over Cape Breton Island, its diagnostic Shield erratics have not yet been recognized among the abundant local crystalline rocks of similar lithology. It thus seems likely that Laurentide ice merged with the formerly east-flowing Appalachian ice and changed its trajectory to south(east)ward, but why Laurentide ice did not spread Shield debris completely to its margin on the shelf may be explained by the fact that Appalachian ice had already occupied this outer area during Phase B and also perhaps that the Laurentide phase did not last long enough to accomplish full transport. This early southeastward "ice flood" which extended offshore south of Cape Breton Island onto the eustatically emergent shelf probably deposited the Scotian Shelf Drift, the only till recognized in that area (King and MacLean, 1976). This major regional flow, long recognized as the maximal phase of the last glaciation was the basis for the so-called "maximum model" of ice extent (Prest, 1984) (Fig. 121a). It is regarded as the culmination of the initial eastward advance (Phase B).

Various ages have been proposed for this event, but oxygen isotope stage 4 is likely the most reasonable estimate. It was initially generally assumed to have occurred during the Late Wisconsinan (e.g., Prest and Grant, 1969), but that estimate was later revised to Early Wisconsinan (Grant, 1977; Grant and King, 1984; Fader et al., 1982; King and Fader, 1986) when offshore studies (Alam and Piper, 1977; Alam et al., 1983) showed that to be the time of greatest glaciomarine sedimentation in the adjacent deep sea basins. The inferred tunnel valleys on outer Scotian Shelf, thought to have been cut by sudden discharge of basal meltwater from beneath a regional ice sheet in oxygen isotope stage 4 (Boyd et al., 1988), may be further evidence that the maximal glacier extent occurred in the Early Wisconsinan. The deep sea oxygen isotope record in north Atlantic Ocean (Ruddiman and McIntyre, 1976) also shows stage 4 to be one of the two periods of greatest accumulation of ice-rafted detritus and hence most severe glacial conditions in the eastern Canadian region.

*Possible recessional phase*

There is some scattered and indirect evidence of nonglacial conditions showing that ice withdrew from at least the southern lowland portion of Cape Breton Island after the eastward (Phase B) and southeastward (Phase C) flow events. However, the reality of the event depends mainly on two pieces of evidence: whether the intertill silts, for example at Castle Bay [25] and Big Brook quarry [49], are correctly placed in the stratigraphic sequence and whether the iron stain found locally on the eastward and southeastward striated surfaces (e.g. sites [36] and [38]) represents subaerial weathering. Otherwise, there are no sediments with a distinctive Middle Wisconsinan pollen signature, nor are there any reliable finite Middle Wisconsinan age determinations from the lowlands; all are considered suspect and minimal: the 32 ka Middle River mastodon femur [15], the 32.7 ka wood at Dingwall [4], the 36.2 ka middle silt at Big Brook quarry [49], and the 32 ka marine shells in the Grantville esker [47]. The 44.2 and 47 ka age estimates on wood at Bay St. Lawrence [1], may be valid, but they can only signify that ice retreated from the deepest part of Laurentian Channel, as Stea et al. (1992b) have also suggested. Otherwise, there is a total lack elsewhere in the region of dated sedimentary evidence of an intra-Wisconsin interlude in the glacial history.

Oceanic conditions in the north Atlantic region at this time offer some support for the notion of glacier reduction in the Maritime Provinces area. There is evidence in the global temperature and sea level records that global ice volume during stage 3 decreased significantly relative to the full glacial conditions represented by stages 4 and 2 (Mix and Ruddiman, 1984; Shackleton, 1987). If, as hypothesized above, these oceanic conditions are a reasonable proxy of ice extent in the Cape Breton Island region, there may have been a major retreat of glaciers corresponding to the globally reduced glaciation of oxygen isotope stage 3.

This ice retreat interval, as originally hypothesized, was given interstadial rank and informally termed the "Bras d'Or Interstade" (Grant in Dreimanis et al., 1981) in reference to the seeming evidence that the lowlands were ice free. The term should probably be discontinued until stratigraphic evidence proves that the alteration resulted from weathering during significant deglaciation in the area at that time.

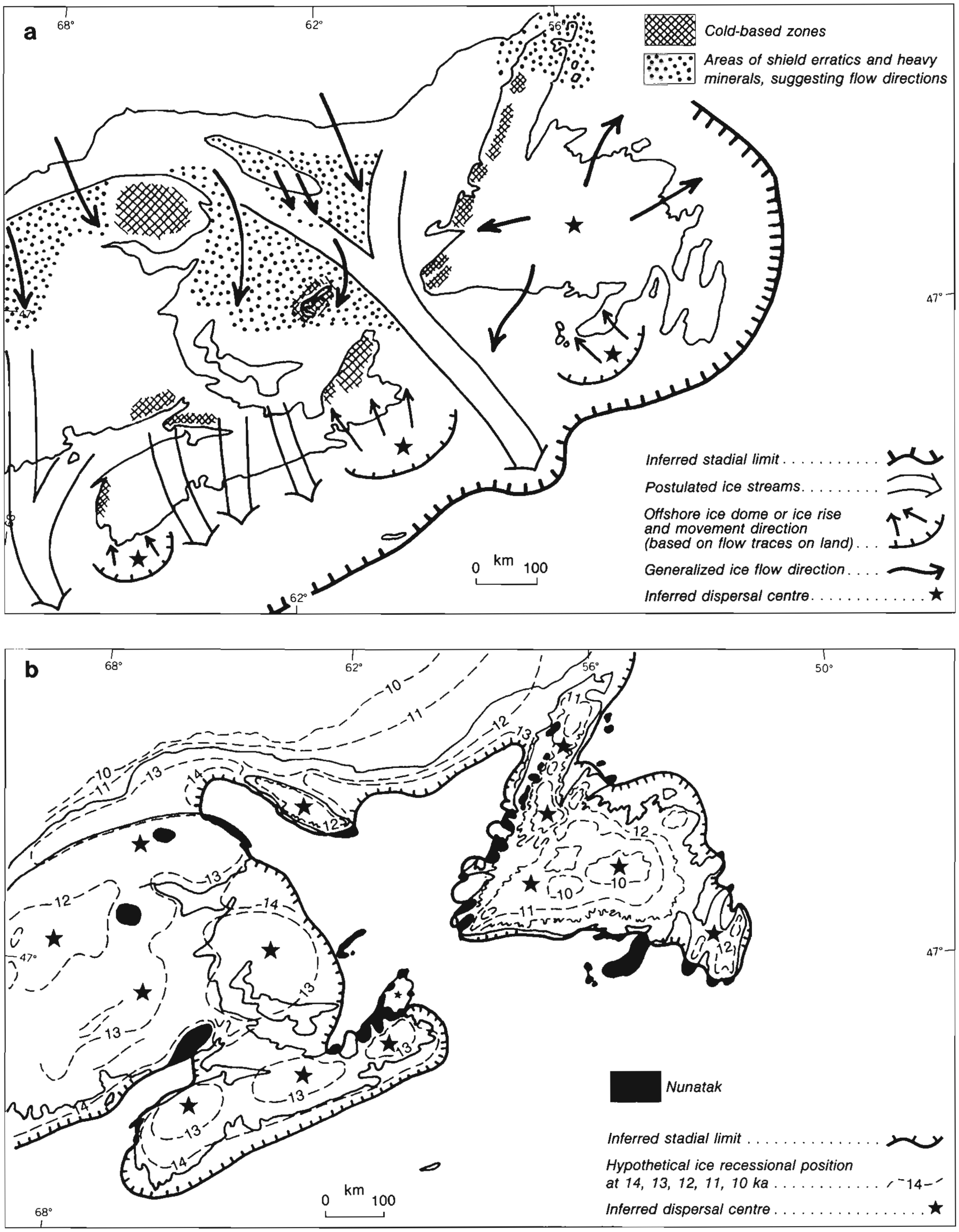



***Figure 121a.*** *Hypothetical maximum extent of glaciers in the Atlantic region during Wisconsinan time (from Grant, 1987, p. 13).* ***b.*** *Hypothetical extent of glaciers in the Atlantic region during the last major Wisconsinan expansion (from Grant, 1987, p. 13).*

*Phase D – ice flows northward from the continental shelf*

After the eastward and southeastward movements, and possibly after an ice-free interval, glacier ice fanned out generally northwards over the southern lowlands (flow pattern D) from a centre that must have been located south of the island (Fig. 119c). The ice-flow direction ranged from northwestward (340°) in the western part of the island, to northward (000°) in the central part, to northeastward (030°) in the eastern part. As this flow pattern is recorded along the entire south coast of the island, the source of the ice must have been situated on Scotian Shelf (Grant, 1977). This glacier movement brought redbed debris and marine shell fragments ashore to form a distinctive, bright red, fossiliferous till. It imparted a broadly sinuous form to the presumably linear axes of the early east/west till drumlinoids and locally carved the ends into drumlins. The northeastward flow over the Gabarus Bay area carved the giant drumlinoids into smaller drumlins and deposited on top of them a locally-derived grey till that it generated from bedrock in the intervening swales. The provenance of this surface till clearly shows the strong northeastward transport (McClenaghan and DiLabio, 1992; McClenaghan et al., 1992). The northward-bedded esker in River Inhabitants valley was probably formed at this time in response to northward hydraulic gradient of this onshore-moving ice mass. The northward gradient is further evidenced by the fact that this movement surmounted the 300 m Creignish Hills, but farther north skirted around the Mabou Highlands below 250 m and below 100 m elevation on the western highlands margin. These features indicate that the thick ice originated from a centre located on the Scotian Shelf south of Cape Breton Island, as previously postulated (Grant, 1971a, b; Grant in Prest et al., 1972).

At its greatest northward extent, this Scotian Shelf ice was evidently in contact with the east and west flanks of a contemporaneous ice cap spreading off the highlands plateau. Scotian Shelf ice wrapped around the highlands ice mass, thus moving parallel to the coast of the fringing lowland and spreading northeastward onto Sydney Shelf. The confluence of the two ice masses is reflected by the mixture of lowland- and highland-margin tills, as at Morrison Brook [8] and Wreck Cove head [10]. Similarly, Scotian Shelf ice extended along the western highlands margin, as shown by striations oriented parallel to the coast as far as north as Margaree Harbour, where it converged with highlands ice along the Chéticamp Lowland and deposited tills and hummocky ice-contact deposits of mixed lowland and highland derivation. The northward gradient of Scotian Shelf ice along the western highlands margin, and its persistence relative to highland ice, is perhaps further evidenced hydrologically by the series of sidehill meltwater channels at 70-200 m elevation which were cut into the highland flank.

The shape of the Scotian Shelf ice dispersal area can only be approximated. The generally northward and uphill flow and the northward-declining surface proves that the ice mass was centred offshore to the south. The location of the southern margin can be estimated using the Nye formula (height of an ice cap at any point is 4.7 times the square root of the distance from the centre to the margin minus the distance of that point to the margin). Assuming that the Scotian Shelf ice mass was symmetrical about a divide that was located just south of Isle Madame, and given that the ice overtopped 300 m summits ≥100 km north of this divide, the southern margin of the shelf glacier may be placed approximately at the north edge of Banquereau Bank. If so, it is interesting to note that the belt of deep basins which are excavated in the soft shelf bedrock in that area may represent erosion by this hypothetical ice cap. Similarly, the north margin of Scotian Shelf ice is placed near the east end of Prince Edward Island where, perhaps coincidentally, there is a large subsea deposit of chaotically bedded sand and gravel, with no obvious source, which may be an ice-marginal deposit (Kranck, 1971).

This apparent shift in the ice-dispersal centre from a position north of Nova Scotia southward toward the oceanic area may indicate an actual shift in the locus of snow accumulation seaward. If so, it may thus be further evidence of low glaciation levels and advection from a strong thermal gradient just offshore (Johnson and McClure, 1976; Ruddiman and McIntyre, 1979; Ruddiman et al., 1980). Two conditions, low sea level and low glaciation level, suggest that Scotian Shelf could have become a ice dispersal centre. If, as suggested above, the preceding regional ice sheet had retreated (during oxygen isotope stage 3), most of the eastern part of the Scotian Shelf would have been emergent and thus available as a site for build-up of glacier ice because sea level was ≈120 m lower during the greater part of the last glaciation (Fader et al., 1982; King and Fader, 1986). Further indication that glaciation level in the area was quite low are the cirques around the Newfoundland coast that are now submerged ≥40 m (Grant, 1992). It is therefore concluded that an ice cap could have grown on the eastern Scotian Shelf.

Alternatively, there are three additional mechanisms to explain an apparent Scotian Shelf ice dispersal centre even if there was no general deglaciation of Scotian Shelf during Wisconsinan time and the regional south-flowing ice sheet remained in place. Ice flow could have been reversed from southward over the shelf to northward over Cape Breton Island in the following ways. Firstly, by the strong landward advection referred to above, the margin of the larger ice sheet on the outer shelf could have developed a subsidiary marginal dome with radial flow that would have had a northward component onto Cape Breton Island. Secondly, if the ice sheet went afloat as an ice shelf in the Chedabucto Trough and associated basins, then regrounded farther out on Canso and Misaine banks, it would have developed an ice dome or ice rise at its margin. The ice rise could have been further nourished, thereby developing strong northward flow toward Cape Breton Island. Thirdly, there may have been calving of the regional ice sheet in the deepwater areas of Gulf of St. Lawrence, perhaps because interior portions were being starved by the development of offshore accumulation centres and/or because glacio-isostatic loading was causing relative sea level to rise and decouple the ice sheet from its bed. Such calving in the ice sheet north of Cape Breton Island would have caused drawdown to the north and a consequent reversal of the former southward flow direction. From the available data, it is not possible to say which scenario is the more likely; perhaps the factors worked in combination.

Whatever the mechanism, the postulated ice dispersal area offshore from Cape Breton Island evidently was not unique. Evidence has been found of glaciers moving landward into the Yarmouth area from the southern Scotian Shelf (Gravenor, 1974) and onto Burin Peninsula of Newfoundland from the Grand Banks (Grant, 1977). Although offshore workers find no direct positive or negative marine geological evidence for these shelf-centred ice caps (e.g. King and Fader, 1986), the explanation may be that proposed by Stea et al. (1992b) who show the ice dispersal was an extension onto the shelf of the Scotian Ice Divide (their Ice Flow Phase 3) which elsewhere remained on shore. A possible reason for its ability to be sustained offshore is that there are more shallow banks on eastern Scotian Shelf to stabilize it.

The timing of the offshore ice-dome phase can be bracketed only approximately. It followed two major flow events which began after 86.9 ka and possibly not until 51 ka if the Th/U dates at East Bay [27] are accurate. If the weathered striations and 32 ka shells at Grantville indicate that ice in the area disappeared before the Scotian Shelf ice-cap phase, then it may date from the Middle Wisconsinan or probably later. The younger end of the Scotian Shelf glacier phase age range is constrained by two later Wisconsinan flow events, the older of which had ended by 11-12 ka. If this offshore glacier reached to the shelf edge then it was Late Wisconsinan, as Mayewski et al. (1981) have argued. King and Fader (1990) concluded that, if moraines and glaciomarine sediments on the shelf are correctly dated, a glacier covering Laurentian Channel and Scotian Shelf culminated at the shelf margin at 26-21 ka. If so, the ice dome could have been a part of that larger ice mass.

This glacial regime, apparently signifying an offshore ice cap which seemed unique in the glacial history of Nova Scotia, was considered to have sufficient climato-stratigraphic significance to warrant designation as a stade and accordingly was informally termed the "Isle Royale Stade" after the original name for Cape Breton Island (Grant in Dreimanis et al., 1981; Grant and King, 1984). As there is no evidence the event was climatic, and some reason to believe it was a mechanical phenomenon, the term should probably be abandoned.

*Phase E – establishment of an island-centred lowland ice cap*

The flow from the offshore centre ended with the inception of an ice-flow regime centred on the southern part of Cape Breton Island. The main manifestation of the new regime was a flow reversal which superposed southeastward (160°) striations and drift lineations on earlier trends (flow pattern $E_1$ of Fig. 55) over the area south of Bras d'Or Lake between Gabarus and Strait of Canso. Its main effect there was to lightly striate any exposed bedrock and to superimpose on the pre-existing till masses a patchy veneer of locally derived, stony till which is ornamented by grooves and fine flutings. The till was derived mainly from the high-standing crystalline rock uplands, such as East Bay Hills, North Mountain, and South (Sporting) Mountain. The northern flank of the ice mass is tentatively placed along the highlands escarpment, such that ice escaped laterally northeastward onto Scotian Shelf.

This hypothetical island-centred ice mass is inferred to have lain over the Bras d'Or basin (Fig. 119d) because its southward ice-flow trends are found only south of the basin and because earlier ice-flow trends north of the basin are unmodified. The shift in the location of ice-dispersal centre from offshore to onshore signalled a complete reorganization of the ice flow structure and is presumed to have been accomplished by a progressive northward migration of the ice divide. The change from northward to southward flow and the inferred shift of the ice divide was probably relatively rapid because there are no intermediate ice-flow trends. Perhaps the change occurred because glacial conditions were gradually lessening, thereby causing the shelf ice cap to shrink preferentially landward. The suddenness of the change, however, suggests that a different mechanical force came into play. Sea level was rising at this time because of both global eustatic recovery and local isostatic crustal lowering, so that calving of the marine-based portions on Scotian Shelf and in Gulf of St. Lawrence may have been a contributing factor in the shrinkage.

The regional context of this island-centred ice-flow regime can be judged from interpretations for mainland Nova Scotia (Stea et al., 1992b). Since there is no evidence that Cape Breton Island ice had separated from mainland ice at this stage, Phase E probably should be correlated with the latter part of the (Stea et al., 1992b) Ice Flow Phase 3 which is characterized by ice dispersal from the topographic axis (Scotian Ice Divide). Stea et al.'s (1992b) estimate of 15-13 ka for this Nova Scotia-centred flow regime is therefore tentatively adopted for glacial Phase E on Cape Breton Island.

*Phase F – Gulf of St. Lawrence ice impinges on the west coast*

When the Bras d'Or ice mass was at its maximum northward extent, or at least when it eventually began to shrink toward the interior, it was evidently met on its northwestern flank by foreign ice that moved eastward and northeastward (050°) from Gulf of St. Lawrence onto the western coastal lowlands (flow pattern F, Fig. 119d). The eastern and upper limit of Gulf of St. Lawrence ice against the island is considered to be marked by the following features from south to north: the moraines banked against the western slope Creignish Hills at ≈250 m; the upper limit of scoured bedrock on Mabou Hills; and the sidehill channels which range up to 200 m at Chéticamp. Other, perhaps somewhat later effects are the minor moraines on north flank of Creignish Hills and the large paleochannel and blanket of hummocky ice-contact stratified drift in the Chéticamp area.

*Intermontane proglacial lakes* – The most important effect of this Gulf of St. Lawrence ice pressing in from the west was to block the drainage of meltwater on the west side of the island. Gulf ice evidently remained against the west coast while inland ice shrank to Bras d'Or Lake area because it dammed the western ends of several intermontane valleys, built kame moraines, and impounded between the two ice masses a declining series of four interconnected proglacial lakes in Bridgend, Skye Glen, Ainslie, and Margaree valleys. They cascaded from one to another in a general northeastward direction (Fig. 119d), presumably because of the gradient of Gulf of St. Lawrence ice. The presence of the lakes proves that glacier ice disappeared over a wide area of western Cape Breton Island, yet they require that westward drainage to the coast be blocked by glacier ice in Gulf of St. Lawrence. Hence, Gulf ice must have persisted while lowland Cape Breton Island ice retreated into the confines of Bras d'Or basin. Further evidence is the fact that the original drainage of Ainslie basin westward to the coast at Inverness became permanently blocked with till and ice-contact stratified drift and was redirected north via a circuitous route over the divide to the Margaree basin.

In the southernmost part of the re-entrant between Gulf of St. Lawrence ice and Bras d'Or ice, proglacial lakes stood at elevations ≈160 m and ≈120 m in Bridgend and Skye Glen valleys, respectively. They were first depicted (albeit as a single water body) and named "glacial lake Fletcher" on a manuscript map by H.L. Cameron. Glacial lake Fletcher was impounded between Gulf ice standing at the Nevada Valley moraine and Bras d'Or ice at the Churchview moraine. It overflowed through a spillway at 120 m elevation into the Ainslie basin where another ice-dammed lake at ≈90 m elevation was created by Gulf of St. Lawrence ice standing at the Strathlorne/Kenloch kame moraine/delta whose summit is ≈105 m elevation. A southern arm of the Ainslie lake may have extended to the Churchview moraine. Ainslie basin water, in turn, spilled over the divide along the Southwest Margaree River valley and thence into another ice-dammed lake in the Margaree basin. It was called "glacial lake Margaree" by H.L. Cameron on an unpublished manuscript and shown on the 1958 Glacial Map of Canada (Wilson et al., 1958). Glacial lake Margaree was impounded by Gulf ice standing at the 45 m high Margaree Harbour kame moraine and was fed also by drainage from interior ice standing at the end moraine just north of lakes O'Law. This last lake in the series drained into the Gulf of St. Lawrence either at the Margaree Harbour moraine or, if Gulf ice extended farther north along the coast at that stage, elsewhere along the Chéticamp Lowland, possibly by way of the Chéticamp paleochannel which descends from ≈30 m elevation near Terre Noire to below present sea level at Chéticamp.

It was during the contact and separation process of Gulf of St. Lawrence and Cape Breton Island ice masses that the unexplained surficial red silt is inferred to have been deposited on the west coast. As discussed in the Stratigraphy section under Grand Étang site [66], it is suggested to be the product of discharge of muddy proglacial lake water which escaped along and under the disintegrating margin of the Gulf ice mass.

The regional context of this late movement of Gulf of St. Lawrence ice which apparently came from the direction of the Magdalen Shallows and crossed George Bay onto the northwest coast is not altogether clear. It would seem to have been an obvious extension of the eastward flow which was the last movement in nearby eastern Prince Edward Island (Prest, 1973). As such, it may have been the eastern part of the Escuminac Ice Centre, a large, radially spreading glacier which was located on the exposed shelf and flowed south and eastward over Prince Edward Island (Rampton et al., 1984). They and Stea et al. (1987) suggest that it persisted until ≈13 ka, but perhaps it remained active until almost 10 ka, judging by the lack of postglacial lake and bog sediments older than that age. Stea et al. (1987) do not, however, depict such an advance onto western Cape Breton Island after the interior parts of the island had been deglaciated. The advance does, however, appear to have predated their final late flow of land ice in a seaward direction into George Bay (Ice Flow Phase 4) which they estimate at 10-11 ka.

### *Phase G – concentric retreatal flow and local readvance of lowland ice*

Gulf of St. Lawrence ice eventually withdrew from the west coast and Bras d'Or ice continued its retreat farther into the confines of the basin. It shrank more or less concentrically, as shown by scattered ice-marginal features such as moraines and side-hill meltwater channels; however, these are discontinuous and ice-flow trends are discordant, thus precluding any meaningful reconstruction of retreatal positions and further showing that the various sectors of the ice mass retreated differently. Lack of dating control rules out isochrons. Nonetheless, it is clear that general retreat was punctuated by moraine-building stillstands and there was one readvance during the period 11-10 ka (although it is areally too small to differentiate from the lowland ice margin depicted in Fig. 119d). Ice-flow trends assigned to this stage diverge from an east-west line through Gabarus Bay and St. Patricks Channel, which is therefore considered to be the ice divide. The trend of this late stage ice divide and its associated ice flow patterns correlate with those in adjacent mainland Nova Scotia and suggest that the Bras d'Or ice mass at this stage remained contiguous with an elongate glacier complex called the Scotian Ice Divide that lay over the southern uplands of Nova Scotia during the latter part of the last glacial maximum (Stea, 1987, p. 11; Stea and Mott, 1989, p. 183). During this stage, the extension of Cape Breton Island ice onto the shelf ultimately retreated and became restricted to the Bras d'Or basin, at which time it became separated from ice on mainland Nova Scotia. As discussed below, the north and south flanks of the Bras d'Or ice mass produced different features probably because they lay over very different kinds of terrain.

On the north side of Bras d'Or basin, the ice mass left moraines only at Churchview and Nyanza and was possibly in contact with highland ice. In the three main eastern Bras d'Or channels, the lowland glacier had a lobe originating in St. Ann's Bay which left small end moraines and interlobate ice-contact gravels along the Wreck Cove Lowland as it separated from highland outlet glaciers. At a slightly later stage, this lobe built the large underwater ridge across St. Ann's Bay at Englishtown while the adjacent lobe in Great Bras d'Or channel built the large moraine at New Campbellton. The lobe in St. Andrews Channel spilled meltwater over the bedrock divide at its seaward end and cut deep channels which continue offshore several kilometres to ≈40-50 m depth (cf. Fig. 11). Significantly, this depth is also the level at which there seems to be a marked truncation of seafloor relief at many places around the island (see bathymetry, Map 1631A), as if wave action was prolonged at this position. These features are considered to reflect the relative position of sea level which, accordingly, is placed at a depth of 50 m below present sea level during the latter stages of the Bras d'Or ice cap. The eastern part of the lowland ice mass during this period probably extended beyond the present coast.

The southern flank of the Bras d'Or ice mass retreated northward into the basin, leaving a few, discontinuous, minor moraines on the generally even terrain, but continuing to flow actively so that it fluted the older till forms. Eventually the southern part of the ice mass shrank within the confines of the lake basin and, as it thinned, it lobed around the uplands of North Mountain, South (Sporting) Mountain, and Creignish Hills. Ice flow in the bays was directed westward against the drainage. Thus, for a time, an ice-dammed lake existed in River Denys Lowland, as witnessed by the compacted laminated silts (which were later overridden and covered with till during the readvance described below).

A readvance of Bras d'Or ice occurred in the interval 11-10 ka. Tills were deposited over organics which had accumulated on the newly deglaciated terrain and lakes were re-impounded. The lobe in East Bay readvanced and deposited till over peat at Blacketts Lake [31] after 10 300 BP. It probably formed the moraine belt which extends eastward from Sydney River and was also responsible for the varved sediments in Upper Gillies Lake [30]. The adjacent lobe of Bras d'Or ice in St. Andrew's Channel probably also readvanced and its limit may be represented by the concentration of sidehill meltwater gullies at Boisdale.

Lobes of Bras d'Or ice in the western bays and lowlands evidently readvanced also because new proglacial lakes were created and late glacial organics were overridden. The ice front again blocked the mouth of River Denys Lowland, perhaps at the Malagawatch ice-contact deposit and created a lake, termed "glacial lake Bell" for reasons given above under Glaciolacustrine deposits and lake-level indicators. It is marked by a series of hanging deltas ranging up to 70 m in which silts were deposited over organics dating 11.0 ka at Big Brook quarry [49]. The western ice margin probably wrapped around North Mountain and deposited the Glenora ice-contact deposit across the mouth of River Inhabitants valley, thus creating a another proglacial lake which may be named "glacial lake Dawson" in recognition of J.W. Dawson's (1855) pioneering observations on the Quaternary stratigraphy there. Hanging deltas and laminated silt deposits at the northern end show that the lake ranged up to about 100 m. Glacial lake Dawson overflowed into neighbouring glacial lake Bell through a spillway which crosses the divide at ≤100 m near Glendale and built a delta at 70 m at its lower end. Farther west, the Bras d'Or ice readvanced northward into George Bay beyond Judique and deposited outwash over organics dating 10.8 ka at Campbell [53]. Its eastern margin on the flank of Creignish Hills probably deposited the series of north-sloping kame terraces and small moraines which decline in elevation northward from ≈100 m near Creignish to ≈30 m near Judique.

The approximate time of the readvance (and of the consequent ice-dammed lakes, can be roughly bracketed to the period 11-10 ka by ages on organic materials which predate and postdate the event. As detailed above, organics which yield ages of 11.0, 10.8, and 10.3 ka are overlain by till and other mineral sediment produced by the event. Ages of 9-10 ka are obtained on surface organics in the same area. From the similar ages on comparable sequences throughout the region, the readvance has been linked to the climatic deterioration which affected the area in Younger Dryas time (Mott et al., 1986) and caused remnant glaciers in other areas to readvance, for example, in northern Newfoundland (Grant, 1969).

The approximate time when this lowland ice mass finally disappeared from the Bras d'Or area can be estimated from the ages of surface organics, although there are too few sites to allow the ice margin to be configured. The northern part of the hypothetical late lowland ice cap had retreated from lower Middle River valley by 12.4 ka, the basal age at Third Lake O'Law [81] and the eastern part had withdrawn from the mouth of East Bay at least by 9.5 ka, the basal age of a peat deposit at Benacadie shore 2 [20]. On the south, the ice margin was well inland of the coast by at least ≈11 ka, if the basal date at Gillis Lake [40] is corrected from 13.9 ka for the hardwater effect according to R.J. Mott (in McNeely and McCuaig, 1991). From one core at the head of East Bay, de Vernal and Jetté (1987) found that the pollen sequence began after the Younger Dryas climatic oscillation and, on that basis, estimated the beginning of lacustrine sedimentation at 10-9 ka. Together, these would limit the ice margin to within the confines of the present lake basin by about 10 ka.

### *Phase H – radial flow and concentric retreat of the highlands remnant ice cap*

Geomorphic features indicate that an independent ice cap occupied the highlands plateau and that, after being in contact with lowland ice for a time, it continued to flow actively while retreating more or less concentrically to the highest part of the plateau. Ages on surface organics suggest that its final remnants disappeared somewhat later than lowland ice, as might be expected from the fact that present day seasonal snow cover lasts about twice as long as on the lowlands. How far highland ice extended onto the surrounding lowlands cannot be defined, but the distance appears to be minimal because there are no places where pure highland crystalline rock till is displaced more than about one kilometre onto the adjacent sedimentary rock terrane. For paleogeographic purposes, three distinctively different retreatal configurations can be outlined on the basis of geomorphic features: 1) highland and lowland ice caps contiguous (Fig. 119c); 2) highland ice a separate cap with several outlet glaciers (Fig. 119d); and 3) a late remnant (or a Neoglacial ice field re-formed) on the highest area (too small to show on Fig. 119d). None of the configurations is closely dated yet, but ages and some correlations can be suggested from indirect evidence.

The first phase was when highland ice was in contact with lowland ice at its greatest northern extension along the east and west highland margins (Fig. 119c). That configuration is considered to have been responsible for the hybrid tills and ice-contact stratified drift that were deposited in the convergence zone. The great volume of these sediments suggest that the convergence lasted a not inconsiderable period. If so, perhaps it dates to the period 14-13 ka when glaciers withdrew from most of the Gulf of St. Lawrence and became localized on land with their margins stabilized in coastal areas (Grant, 1989).

The second major definable phase of highland ice occurred after it separated from lowland ice and stabilized with a highly lobate margin in two main forms (Fig. 119d). One was an ice mass on the highest part of the plateau which sent outlet glaciers down the major valleys, such as that of the Aspy on the north, Dundas, Clyburn, Ingonish, and others on the east, Middle and Baddeck valleys on the south, and Margaree and Chéticamp on the west. The other was a number of small cirque-fed valley glaciers on the northernmost part of the plateau in the Grande Anse, Red, Blair, Delaney, Sailor, Wilkie, and Gray Glen valleys, among others. All of these valleys became sculpted to varying degrees into U-shaped troughs. End moraines were constructed in the lower reaches of some valleys and in others, terminated beyond the present coast. There is no direct information on the age of the moraines, except for a single age determination of 11.6 ka on organics buried by gravel in the kame moraine of the Grande Anse valley glacier at Pleasant Bay [73] (S. Occhietti, pers. comm., 1987). In addition, unpublished ages (R.J. Mott, pers. comm., 1992) of basal sediment in lakes which lie just beyond these features, namely Cormier Lake [90], Third Lake O'Law [81], and "MacInnis lake" [79] indicate that the outlet glacier phase postdates 12 ka. It is therefore reasonable to suppose that, during the resurgence of Bras d'Or Lowlands ice which occurred during the 11-10 ka cold period, highland ice experienced a comparable expansion.

During its ultimate stages of retreat the ice cap was localized entirely on the plateau. When it was about 10 km in diameter, it produced a belt of small end moraines and marginal meltwater channels in the thick till in the headwaters of Ingonish River. The age of the moraine-building phase can be no younger than 8-9 ka, the age of the basal lake sediment at Wreck Cove pond [9] which is situated on the outer edge of the moraine belt. Perhaps the moraines were a response to the climatic cooling, which is represented in Gillis lake [40], Salmon River Lake [41], and Wreck Cove pond [9], which Anderson and Lewis (1992) attributed to a second major discharge of glacial meltwater into Gulf of St. Lawrence from the glacial Great Lakes, much as an earlier major discharge into Gulf of St. Lawrence caused the severe climatic reversal and glacial readvance at 11-10 ka in the Atlantic provinces region (Mott et. al., 1986). Thereafter, flights of small sidehill meltwater channels cut in the residuum (Fig. 32) show that the ice cap shrank to a small carapace about 5 km in diameter situated on the highest ground at ≈480 m elevation in the vicinity of Chéticamp Reservoir in the national park.

It is possible that during the Neoglacial period there was a sufficient cooling to devegetate and destabilize slopes and possibly to cause re-accumulation of glacier ice. Pollen evidence from Newfoundland suggests that climate in the region a few centuries ago during the Little Ice Age reverted to conditions as extreme as during the early postglacial 8-9 ka (T.W. Anderson, pers. comm., 1992). If so, it may have caused intensified solifluction and thus may explain the accumulation of slope debris over 2000 year old peat at Fishing Cove River barrens [71]. Further, if there was glacier ice on the plateau during the earlier cold period, then, theoretically, comparable climatic conditions during the Little Ice Age could have promoted the development of small icefields and cirque glaciers in northern Cape Breton Island during the last millennium.

## Crustal movement and sea level change

Actual and apparent changes in the relative level of land and sea over different spans of time are generally revealed by the changed elevation of paleoshorelines. The only such fossil shoreline in Cape Breton Island is the emerged interglacial littoral bench. Its general tilt and local displacements could be due to any one of a number of possible changes since it was formed 125 ka ago. Relative sea level change in Cape Breton Island due to isostatic crustal recovery since deglaciation cannot be directly evaluated as there are no raised shorelines to provide local information; however, general inferences about how Cape Breton Island behaved can be drawn from the emergence/submergence pattern in adjacent areas (Grant, 1977, 1989), including the surrounding shelf where there are submerged shorelines down to about 120 m below sea level (King and Fader, 1986). Events in the past several millennia involve relative sea level rise due mainly to crustal subsidence which is causing widespread coastal erosion and deposition of intertidal deposits that provide a few dates (de Vernal et al., 1985). Movements during the last few decades are measured by tide gauges and geodetic relevelling (Vanícek, 1975; Vanícek and Nagy, 1981).

### *Tilting of the emerged interglacial marine bench*

Relative to its modern (interglacial) counterpart, the Sangamonian interglacial shore bench in Cape Breton Island slopes down to the south from 7.2 m at the northern tip, to ≈5 m on the southeast coast, to possibly as low as 2 m on the southern tip. This elevational range is not inconsistent with the elevation of interglacial shorelines in stable areas elsewhere in the world, which range from 2-9 m and average 6 m (Moore, 1982; CLIMAP Project Members, 1984). The present elevated position of last-interglacial shorelines respect to present sea level is generally accepted to reflect the fact that last-interglacial global temperature was warmer than during the present interglaciation and hence less water was tied up in glacier ice; however, shorelines representing the highest level of interglacial seas should be parallel to one another and to present sea level, unless there has been differential movement of the crust since the feature was formed, or a change in the shape of the geoid, the equipotential gravitational surface to which sea level approximates. Instead of being level and parallel to present sea level, the Cape Breton Island feature has a tilt of at least 3 m and possibly 5 m which requires explanation.

Considering first the possibility that the geoid has changed its shape (elevation), note that this surface, which is approximated by sea level, presently has a relief of ≈25 m across the region and has a ≈2 m southward height difference across Cape Breton Island (Fig. 122). Geoid elevation mainly reflects subcrustal mass and its large-scale height variations have been linked to mantle convection and core/mantle topography (Cazenave and Dominh, 1985; Cazenave et al., 1989). Geoid elevation also combines lesser oceanographic (steric) factors, namely temperature, salinity, and currents, which determine the dynamic height of the water column. The geoid (and hence also sea level) must therefore be a dynamic and ever-changing horizon, at least over time spans of tens of thousands of years over which mass redistributions take place, as Mörner (1976, 1980) first suggested. A change in any one or more of these factors over the last 125 ka could have reconfigured present sea level such that it is not parallel to its previous interglacial shape. Indeed, Nunn (1986) has suggested that the present height differences of last-interglacial shorelines around the world may indicate geoidal shifts. If so, the present "tilt" of a few metres of the Sangamonian shore bench across Cape Breton Island could reflect a minor change in subcrustal mass.

On the other hand, the 3-5 m height change since 125 ka could be due to true crustal tilting by any one or all of the following three mechanisms. If the long-term Cenozoic subsidence of 5-10 cm/ka which has lowered the shelf (Jansa

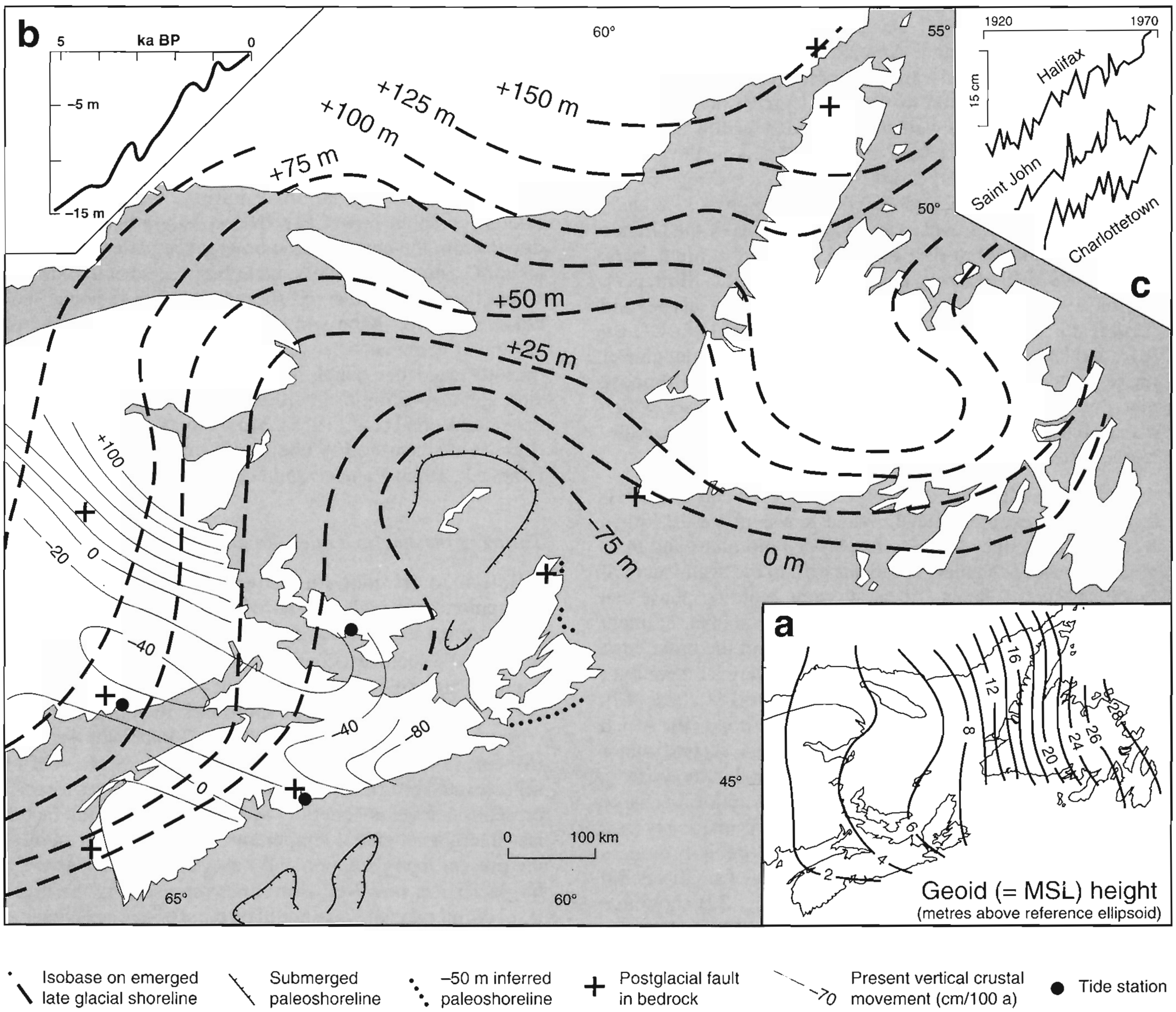



***Figure 122.*** *Summary of evidence of crustal movement and sea level change in the region (after Grant 1989, p. 428). Insets:* ***a.*** *present elevation of the geoid (≈ mean sea level) above the reference ellipsoid;* ***b.*** *general trend of relative sea level rise along the Atlantic coast of Nova Scotia during the last few thousand years (Grant, 1975b);* ***c.*** *rise of mean sea level over the past few decades based on tide gauges (Grant, 1975b).*

and Wade, 1975; Gradstein et al., 1990) has continued to the present and has a hinge line near the present coast, there would have been a concomitant uplift of the landward side, such that northern Cape Breton Island could have been tilted up 5 m relative to the south during the last 125 ka. Alternatively, the tilt could be related to the general ≈0.3 m/km up-to-the-north slope of tilted late glacial shorelines (Fig. 122) (Grant, 1989, p. 428), except that the bench tilt is an order of magnitude less, or, the tilt may be an artifact of incomplete postglacial isostatic crustal recovery. The southern Gulf of St. Lawrence region, including Cape Breton Island, has been subsiding at 20-40 cm/100 a during the last several millennia (Grant, 1970, 1989) and this subsidence is generally attributed to the collapse of a northward-migrating glacial crustal forebulge (for example, Pardi and Newman, 1987). Indeed, the ongoing subsidence in the region shows that isostatic crustal recovery is still in the process of returning the crust to it previous position relative to sea level. Hence, areas to the south (which were deglaciated earlier) have been subsiding longer and at a generally faster rate than areas to the north. This means that the northern part of any paleoshore has yet to be lowered as much as its southern end. It is not possible to say whether the present pattern of subsidence will exactly correct the tilt, but it is interesting to note that, if present rates continue, sea level during the present interglaciation will return to its Sangamonian level in Cape Breton Island within 1-3.5 ka, not counting any enhancement of rate or magnitude caused by eustatic sea level rise related to the Greenhouse Effect.

*Faulting of the emerged interglacial marine bench*

In one place, the bench is visibly upfaulted several metres (Fig. 123). The bench was first reported to be faulted in the vicinity of Money Point and on St. Paul Island (Neale, 1964), but he did not specify exact locations. No faulting of the bench on St. Paul Island was observed by the present writer on a brief visit, despite the fact that the feature can be traced almost continuously and is virtually intact (Fig. 39a), with a distinct notch at ≈8 m (slightly higher than on northernmost Cape Breton Island). A large displacement of the bench was, however, found by the writer where Aspy Fault [3] crosses the Aspy Bay coast (Grant, 1990). On Money Point, the bench is well developed and is glacially scoured in a southeastward direction and overlain by patches of till (Fig. 74). It continues southward along the Aspy Bay coast as a well preserved bench, largely stripped of its gravel and talus cover by wave action (Fig. 39e). In places, a veneer of the original well rounded beach gravel outcrops under the talus. The bench has a good, clean notch that is 7.2 ± 0.5 m above its modern counterpart (Fig. 39h). At a small branch of Aspy Fault which intersects the shore about midway along the scarp, the bench is abruptly upthrown to about 22 m (Fig. 123) [3]. The upthrown segment, overlain by its beach gravel veneer and two tills, continues south for several hundred metres along the highland margin. Where the coast curves seaward away from the highlands, the bench surface slopes away and descends beneath tide level near Cabot Landing; this slope is not a tilt, but the original slope of the intertidal surface. It is concluded from the abrupt offset that there has been a 15 m upthrow of the south side of Aspy Fault sometime in the last 125 000 years. This occurrence, though not unique because a few other postglacial faults occur in the region (Fig. 122), is exceptional because of its seemingly great magnitude.

The age and reason for the faulting is unknown, but some possible mechanisms have been suggested (Grant, 1990). One explanation may be that the lowlands, which have been eroded so much more that the adjacent highlands, have

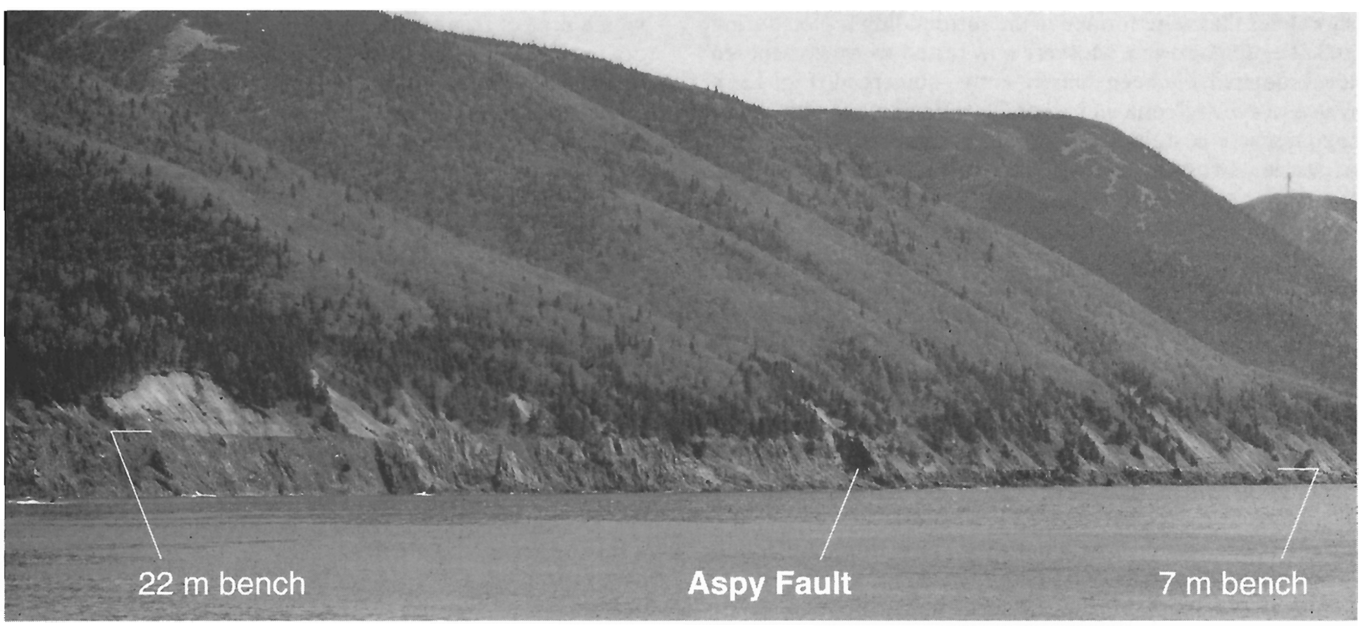



***Figure 123.*** *Telephoto view of 7 m emerged rock bench, upthrown by 15 m to an elevation of 22 m at Aspy Fault, as seen from Cabot Landing, Aspy Bay. GSC 203504-R*

rebounded isostatically along the fault in response to the unloading. A second possibility is that, if the highland crystalline block is an allochthon overlying weak Carboniferous sedimentary rocks (Currie, 1977), glacial loading may have pressed the block down into the underlying weak rock, displacing it laterally and upward at the highland margin along the fault. Or thirdly, postglacial rebound may have been concentrated along the fault, a major line of weakness, especially if the glacier retreating in Laurentian Channel paused in the vicinity, thereby induced a large stress gradient. Fourthly, if, as suggested by Bostrom (1984), the crust is also extended during glacial loading and undergoes thrust faulting, perhaps the Aspy Fault movement occurred as the Laurentide Ice Sheet was expanding over the region. A fifth explanation, most favoured by the author, is based on the fact that the region is being compressed from the northeast (Shih et al., 1988) because of subcrustal drag as a result of plate drift (Plumb and Cox, 1987). The ambient crustal stress is the main reason for modern seismicity causing surface thrust-faulting (Basham and Adams, 1984) and most modern earthquakes in the region cause displacement on old fault zones (Burke, 1984). Further, if crustal stress is stored while large ice sheets cover an area (Johnson, 1987), then the Aspy Fault disruption may have been a locus of such seismo-tectonic stress release at some time during a period of major deglaciation, perhaps the last.

*Postglacial differential isostatic changes of level*

There are no emerged glacial or postglacial shorelines on Cape Breton Island, a fact early recognized by De Geer (1892) (although Flint (1940) erroneously postulated marine limits up to 60 m based on straight-line extrapolation of Newfoundland raised shorelines). Despite the lack of onshore evidence in the area, the general nature of postglacial differential rebound can still be inferred from the relative changes in elevation of the shorelines that were formed in the surrounding region during and after deglaciation. Most are now raised above present sea level; others have been subsequently submerged (Fig. 122). Much of Nova Scotia and all of Cape Breton Island is in the region where postglacial shorelines have become submerged as sea level rose to its present position. North of this area, there has been a net fall of sea level because glacio-isostatic crustal rebound has been greater than eustatic sea level rise. The line of no net change (the zero isobase), where the highest postglacial shoreline is still at present sea level, makes a large re-entrant into Gulf of St. Lawrence, more or less along the axis of Laurentian Channel. Assuming a generally regular gradient in the overall deglacial shoreline delevelling (Fig. 122), it would be expected that the deglacial paleoshore in the Cape Breton Island area would lie about 90-100 m below present sea level, (that is to say, it would be intermediate in depth between the -115 m to -120 m submerged terrace on eastern Scotian Shelf (King and MacLean, 1976) and the submerged terraces at -79 m in George Bay-Northumberland Strait (Kranck, 1972) and -62 m to -72 m on Magdalen Shallows (Loring and Nota, 1973)). Evidence of this hypothetical paleoshore position may be the prominent submarine cliff with its base at -80 m to -90 m, visible on the bathymetric contours of Map 1631A off Isle Madame and Fourchu. As discussed above under deglaciation, the younger shoreline at -50 m, which seen as wave-planed till hummocks on Scotian Shelf (Wang and Piper, 1982) and the truncation of Bird Islands ridge on Sydney Shelf (see bathymetric contours on map 1631A), is thought to have been formed after retreat of late ice in the coastal area. The combined data on emerged and submerged paleoshorelines seems to show that deglacial crustal upwarp extends northwards from the edge of the shelf and has been progressively greater toward the north in the direction of greatest ice thickness. The fact that deglacial and early postglacial shorelines around Cape Breton Island have been submerged rather than raised above sea level supports the hypothesis that there was relatively thin glacier ice over the area, a hypothesis which was developed from ice-flow patterns, nunataks, and cold-based conditions. The pattern further suggests that, during the Late Wisconsinan, glacier thickness in troughs such as Laurentian Channel was relatively thin (presumably because of calving) rather than being thicker because of the greater relief. As expressed by the suite of relative sea level curves assembled for the region (e.g. Grant, 1989), it is clear that emergence ended about 8-10 ka, signalling the end of the uplift phase of postglacial isostatic crustal recovery. Since then, the crust has been subsiding, causing a relative rise of sea level, as discussed next.

*Recent relative sea level rise and crustal subsidence*

There is various qualitative evidence that sea level has risen in late postglacial time and continues to rise at present. That there has been a net rise of sea level since deglaciation is shown by postglacial paleoshores which are now submerged. That the relative rise has continued up to present time is shown by the rapid cutting of coastal cliffs in rock and sediment alike (e.g., Fig. 66), by the progressive rise in elevation of the crests of compound barriers (Fig. 72), and by the overlap of salt marshes onto the living forest, remnants of which project from the tidal grass (Fig. 83); however, it has not been possible to quantify the rate of recent submergence by the usual method of dating submerged terrestrial organics, as elsewhere (e.g., Grant, 1970) because the tidal range is too small to expose the material for direct sampling and because the salt marshes have proved too sandy to permit hand coring below one metre. Nonetheless, it is suggested that the rate of relative sea level rise for Cape Breton Island can be approximated from results in other areas within the same isobase zone, such as the Nova Scotia Atlantic coast and Scotian Shelf. In those areas, dated submerged tree stumps and humus layers beneath aggrading salt marshes show that sea level has risen ≈30 cm/100 a for the past 6 ka (Fig. 122, inset b; Grant, 1970, 1975b). Mathematical modeling of the entire postglacial relative sea level history has arrived at essentially the same rate for eastern mainland Nova Scotia and southern Cape Breton Island (Quinlan and Beaumont, 1982). Seawater is estimated to have flooded in over the -8 m threshold of Bras d'Or Lake, changing it from fresh to salt, at 5-4 ka, from which a submergence rate of 16-20 cm/100 a was estimated (de Vernal et Jetté, 1987). A comparable rate of 18 cm/100 a was found for the accumulation of nearshore marine sediments in MacDougall Pond [78] in northernmost Cape Breton

Island (de Vernal et al., 1985). Both are somewhat slower than the rates for mainland Nova Scotia, presumably because Cape Breton Island is farther north in the isobase pattern.

The rise has continued unabated up to the present, as shown by the subtidal position of early Colonial structures, by rise of mean sea level measured by tide gauges, and by sinking of the crust determined from geodetic relevelling. At Fortress Louisbourg, built 1717-1737, the engineering specifications of drains and harbour facilities were adjusted to mean sea level, as measured shortly before construction began. Now, the drains are submerged daily by the tides, the quay is covered by the modern beach, high tide level reaches through the gates in the seawall (Fig. 124a), and mooring rings are now 37 cm below high tide level (Fig. 124b). These relationships clearly show that high tide level has risen considerably over the last 250 years. The reason is not because tidal range has increased, because it is within 2 cm of the range measured by Chabert (1753). Hence, sea level in southern Cape Breton Island has risen about 30 cm/100 a in the last two centuries. Tide gauging over the last 50-75 years throughout the region, on either side of Cape Breton Island, from Saint John, New Brunswick, to St. John's, Newfoundland, show that sea level has continued to rise at rates of 25-45 cm/100 a (Fig. 122, inset c), thus confirming the long-term rate from geological evidence. The only other likely cause for which there is some evidence is that there has been subsidence of the Earth's crust relative to sea level. Indeed, repeated geodetic leveling in the Maritime Provinces region west of Cape Breton Island since 1920 suggests that the western part of the island at least may be sinking as much as 80 cm/100 a (Fig. 122) (Vanícek, 1975; Lambert and Vanícek, 1979; Vanícek and Nagy, 1981). In summary, the main impact of this relative rise of sea level on the coast of Cape Breton Island is an expansion of estuarine conditions into river valleys and vigorous erosion of the coast. Because of rising sea level, the shoreline is retreating landward at 13 m/a, especially in areas underlain by unconsolidated deposits and weak sedimentary rocks.

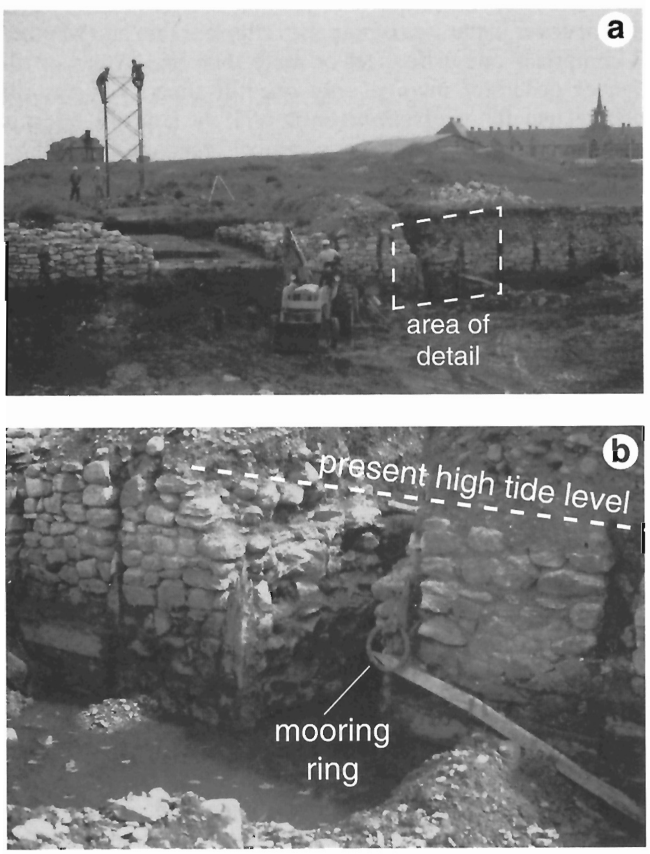



***Figure 124a.*** *View of submerged quay and seawall with mooring ring, being excavated from beneath beach gravel; Fortress Louisbourg National Historic Park. GSC 204024-P.* ***b.*** *Close-up photo and diagram of quayside mooring ring showing depth below present high tide level (modified from Grant, 1970). GSC 204024-J*

## ECONOMIC GEOLOGY

Knowledge and understanding of Quaternary deposits and processes have wide application to most economic activities, namely industrial development, resource exploitation, hazard assessment, and land-use considerations. The state of the natural environment and the changes it undergoes have economic impacts in that they may either hinder or help the creation of wealth and the enhancement of well-being. Certain terrain aspects are liabilities, but most others can be regarded as assets and opportunities which can be more fully developed without sacrificing sustainability. Terrain data and understanding of processes can thus be used to mitigate negative aspects and optimize beneficial features.

### *Quaternary deposits as substrates for economic activity*

#### Application of glacial dispersal trends to mineral exploration

Mineral exploration based on prospecting of bedrock exposures will be hampered to varying degrees depending on the extent and thickness of surficial deposits. On this basis, the mineral potential of the generally rocky upland summit areas and plateaus can be fairly readily evaluated by direct observation of bedrock exposures. Similarly, the belt of residuum terrain on the highlands can be treated as essentially bedrock because it has developed mainly by in situ disintegration and, in some small areas, by chemical weathering. The original lithological identity of these mechanically disaggregated phases will usually be readily apparent and, because their chemical integrity is intact, their mineral potential can be tested directly. The strongly weathered bedrock phases on the other hand, while retaining a direct geochemical relationship to the parent bedrock, should be expected to be differentially depleted or enriched in base metals, depending on the position of the present surface in relation to the paleosol profile. Hence, geochemical prospecting in areas of strongly weathered bedrock paleosols should probably also make use of serial geochemical samples in vertical profiles in order to evaluate any metal anomalies. Fortunately, strongly weathered bedrock is apparently rare and restricted to granitic terranes. Prospecting in areas of bedrock outcrop and residuum can therefore be pursued by normal methods, subject to the qualifications mentioned above.

The greater part (60%) of Cape Breton Island's surface is not bedrock and residuum, but is covered by till (48%) and other glacial and nonglacial sediments (12%), locally of considerable thickness, which can pose a significant impediment to effective and efficient mineral exploration. The following information is offered with a view to reducing the difficulties and uncertainties in pursuing a mineral exploration program in areas of thick and extensive overburden. The following application of surficial geological information to geochemical prospecting deals with the usual case of a program based on the sampling of till.

The extent and thickness of till varies widely over the island and these variations are major limitations on the effectiveness of a geochemical exploration program. Most of the till cover is a thick blanket amounting to 25% of the area. It is commonly composed of lithologically dissimilar layers of wide-ranging provenance, so there are large lateral and vertical variations in geochemical signature within the blanket. Other kinds of till cover are continuous veneer (15%) and discontinuous veneer (7%); these are more closely related to the local bedrock lithology. Surficial deposits other than till, which also effectively prevent direct mineral exploration of bedrock, are waterlaid sediment, colluvium, and organics. Excluding the organic deposits, these deposits are in themselves secondary and tertiary derivatives of till and therefore contain a mixture of bedrock elements derived from a larger area. They also present good targets for mineral evaluation and, in that sense, may aid the process of discovering bedrock sources. For this reason, geochemical assays of till and other surficial sediments have become a prime means of evaluating the mineral potential of the underlying bedrock. Before discussing the optimal techniques of till prospecting as applied to conditions in Cape Breton Island, it is worth noting that there are some advantages in utilizing surficial sediments other than till, such as stream sediment (alluvium), lake sediment, or colluvium, particularly in a reconnaissance program which seeks effective coverage of a large area. These postglacial sediments are largely derived from till and thus are essentially a convenient pre-sorted fine fraction which has been collected and concentrated naturally from a large catchment area; they are a sample of a larger tract of bedrock and they offer a larger target. On the other hand, they are farther removed from the original source and require detailed follow-up sampling to pinpoint the source. Metal anomalies in these sediments will first require that their till sources be determined before applying the following principles. In most cases the parent till is that which is upslope in the catchment area. Once the till source is located, the next step is to locate its bedrock source according to the following general procedures.

Several considerations should be borne in mind when pursuing a till sampling program in Cape Breton Island (these are common to most areas). First, the usual assumption that till chemistry will reflect the underlying bedrock is valid on a regional basis. Locally, however, because glacier erosion is not uniform, some areas of bedrock may be protected, so the absence of metal anomalies in surficial sediment may not necessarily indicate the absence of economic deposits in the underlying bedrock. Bedrock slopes with a lee or downstream position with respect to any ice flow do not contribute as much material to the till as stoss or ice-glacier slopes. Crag-and-tail hills are a prime example of this. The tail side of the bedrock obstacle, where till has been preferentially lodged, is clearly an area where bedrock erosion has been relatively light. Second, each subsequent glaciation must contend with deposits of earlier ones. Not only do these cover bedrock and make it unavailable for further erosion, but they provide to the glacier materials which are not related to the underlying bedrock; hence, in layered sequences the compositional relationship between till and underlying bedrock becomes progressively more remote, the higher in the stratigraphic sequence they lie. Drift prospecting is therefore difficult in areas with multiple till layers, such as the Cape Breton Island lowlands. In contrast, drift prospecting in areas of one thin till will be easier to conduct and interpret than in areas with thick and multiple drift sheets, whether the samples are from the surface or from the base of boreholes.

An effective geochemical till sampling program in Cape Breton Island could proceed in the following way. First, reference should be made to Figure 40a to determine whether the till cover in the area of interest is thick or thin and whether it comprises one drift sheet or more than one. Areas of till veneer generally involve only one till sheet, whereas till blanket usually involves two or more. If the area of interest is till veneer, the source of geochemical anomalies should be first sought along the last ice-flow direction to affect the area, (see Fig. 55). More confidence may be attached to the source direction if dispersal trends in the geochemical pattern are elongated in the direction of the last ice flow. Other than the general pointers, it is not possible to provide a uniform code of practice that is uniformly applicable to the whole island because of the highly variable till thickness, the lithologically dissimilar stacked tills, and the divergence of successive glacial dispersal trends.

Areas of till blanket, such as that over the areas of base metal interest in the southern lowlands, present special problems for geochemical prospecting. Ideally, samples should be taken from the base of the sequence, especially if there is more than one till sheet, in order to obtain material that is most directly related to the source bedrock. The lower till on the southern lowlands, however, is the most distantly derived of the three that occur there, so it will generally not provide the best mineral dispersal patterns for an area. Moreover, sampling of the basal horizon generally requires drilling, as exposures are few. As this is a relatively expensive option, the surface till offers a reasonable alternative sampling medium, providing that the practitioner can establish the relationship between that till and the local bedrock. For example, in applying these techniques to the southern lowlands, the surface till is a good sampling medium because it is locally derived and was produced by the last or next-to-last ice flow; however, it does not form a continuous sheet and the intervening areas are windows where the lower, reddish, fine grained, distantly derived till is at the surface. In summary, when drift prospecting in areas of stacked tills, care must be taken to ensure that all samples come from a single, lithologically distinct till sheet and that it be the one that has the least far-travelled components. Once geochemical anomalies in a given till sheet have been revealed, tracing them to their ultimate bedrock source will begin by applying the appropriate transport direction for that till sheet, as outlined in Figure 55 and detailed in the sections on Ice-flow indicators and Wisconsinan (oxygen isotope stages 4, 3, 2).

Detailed and intensive till sampling and analysis has been carried out by others aimed at providing more quantitative information on the efficacy of the method. Some have been conducted on large till tracts in mainland Nova Scotia in order to develop general principles (Stea and Fowler, 1979; MacEachern et al., 1984; Stea et al., 1986). Other till geochemistry studies have been focused in southeastern Cape Breton Island itself, specifically in the area of complexly-layered thick foreign and local tills in order to evaluate the trends and amounts of dispersal from known mineral occurrences (e.g., McClenaghan and DiLabio, 1992; McClenaghan et al., 1992; McClenaghan and DiLabio, in press; MacDonald and Boner, 1993). Those studies have made significant advances in techniques of sampling, analysis, and interpretation that are applicable to a variety of till settings. The reader is therefore directed to those reports and the further references they contain for essential information on the some of the most effective ways to conduct a mineral exploration program utilizing the till cover as a sampling medium.

### Foundation conditions and trafficability

The complex layering of surficial sediments with different properties, particularly over the till-covered lowlands, will be of concern for various land-use operations. The sandy and stony till phases, which typically occur as till veneer (map units 2b, 2c) (Fig. 40a), would generally present no difficulties for construction because they are unconsolidated enough to permit easy ripping and excavating, yet are compact enough to carry heavy loads without settlement and maintain stable slopes at relatively high angles.

Attention should be paid, however, to the silty, fine grained tills which are relatively weak compared to the stony and sandy tills and have a tendency to flow when saturated, thus requiring that cut slopes to be graded to lower angles or otherwise stabilized with modern methods of surface treatment using stone and "geotextile". Recognizable by their dark red colour, these silty tills occur widely, mainly over the lowlands. They are particularly thick and extensive over their source areas of Carboniferous siltstone and gypsum in the central and western parts of the island (Fig. 7) and thus form the bulk of the surficial cover over the lowlands and are present in all three mapped units of till thickness and extent (Fig. 40a). These silty tills also occur in abundance in the southern and eastern parts of the island far beyond their source areas, having been glacially transported south and east onto crystalline rocks and grey sandstone and then buried beneath dissimilar tills. In those areas, the weak silty tills not only make up most of the overburden shown as till blanket (map unit 2a), but are also present in lesser thickness (till veneer; map units 2b, 2c; Fig. 40a). The presence of these silty tills may be hidden by a cover of coarser till of different colour, as along the Atlantic coast, or by gravel, as over the lowlands bordering Gulf of St. Lawrence. Their presence at depth will be readily apparent in gravel pits, and in coast and river exposures. It will thus be prudent for construction planning to drill or trench in these till blanket areas because the surface composition is generally not indicative of conditions at depth.

### Land-use planning

The composition, thickness, and texture of surface materials provide a basis for assessing the optimal use of the land surface. The lowland half of the island is devoted to a mix of agriculture and diverse economic activities that are unlikely to change in the future. The highland and upland portions, on the other hand, present a greater challenge and opportunity for planning optimal land use now that most of the original-growth forest has been removed and vast areas are available for reforestation. Uncut areas show what combinations of topography and substrate yield the best forest; the surficial geology map can be used as a primary planning tool to evaluate the forest land potential. It is clear that the best conditions for forest growth, as well as for the logistics of forest operations, are the least glaciated areas, that is to say, those with the gentlest slopes because of minimal glacial erosion and deposition. On this basis, the least suitable areas for large-scale forest plantations are the rugged bare-rock areas along the eastern and western fringes of the Cape Breton Highlands (map units Ra, Rb, Rc). Marginally better are the southern uplands around Bras d'Or Lake with their rolling rocky terrain, thin till, and boggy depressions (generally identified by map unit Rc). These include Boisdale Hills, and Kelly, North, and South (locally called Sporting) mountains. The best conditions for reforestation are provided by Creignish and Ainslie hills and the greater part of the Cape Breton Highlands plateau. These areas are flat to gently undulating, well drained, and underlain either by friable sedimentary rock or rotten crystalline bedrock. The uniformity of slope and substrate (and climate) reduce the variables on forest growth and permit very large scale plantations and mechanized operations. For reforestation, application of the surficial geology map can help maximize the potential of proposed programs.

## *Quaternary materials as resources*

### Granular deposits

Sorted aggregates of gravel and sand are essential ingredients for economic activity involving construction. Cape Breton Island has a considerable number of high-quality deposits, both at the surface and in the subsurface beneath other deposits, but relatively little can be counted as exploitable reserves because most are effectively unavailable for exploitation owing to environmental considerations and land-use conflicts. For example, the highest quality, most easily recoverable deposits of gravel and sand are the coastal beaches and the low terraces of Holocene alluvium bordering modern streams. These have been the most favoured sources in the past, but it is now clear, as reflected by protective statutes, that beaches should be preserved as much for the economics of their tourism potential, as for minimizing shoreline erosion. Disturbance of floodplain deposits is generally considered imprudent because of the unacceptable degradation of surface water quality and its consequent negative impact on biotic resources.

Excluding beaches and alluvium, the only deposits actually available for exploitation are the large glaciofluvial complexes and the much smaller glaciolacustrine deltas. The latter are of good grade and lithology, but are scattered and, with a few notable exceptions, are not located near the high-demand areas. The glaciofluvial material is in good proximity to demand areas but most deposits are already devoted to other high-value uses, such as farm land, urban development, and cottage property, very little of which would likely be abandoned in favour of the underlying gravel. For example, the greater part of the large glaciofluvial deposits, some of which serve the Sydney urban region, are effectively excluded from exploitation: those along Sydney River because they are urbanized and those along the Mira River because they constitute prime recreational property. The two main remaining undeveloped deposits in the Sydney area, although nonurbanized, have additional negative aspects: those around Broughton which are presently unworked are of rather poor quality and the remaining deposits on open land upstream along Trout Brook would damage fish habitats unless carefully controlled. The same situation prevails along the built-up west coast area; glaciofluvial deposits around Inverness are fairly densely settled, as are those along the Chéticamp Lowland. The latter, however, offer greater potential because they are extensive in the subsurface beneath a thin red silty diamicton which can be easily removed in environmentally nonsensitive areas. Deposits are particularly lacking the in Canso Strait area, necessitating the use of the more expensive crushed-rock alternative. To meet demand in this area, there is a definite potential in the small, but widely scattered glaciofluvial deposits in the Judique area and the glaciolacustrine deltaic deposits in River Inhabitants valley. Of the remaining deposits on the island, the largest and highest quality are the outwash terraces along Margaree and Middle rivers. Impediments to their development are the long haulage distances and the value of the salmon fishery along these watercourses which could easily be negatively impacted by discharge of muddy waste water and alterations of ground water. The granular deposits mapped and described in this report provide a basis for optimal development of those that would qualify as exploitable reserves.

It should not be overlooked that onshore reserves could possibly be supplemented by sources in adjacent areas and these alternative supplies may in fact be less expensive. Large high quality, direct-shipping deposits occur at tidewater at the heads of fiords in southern Newfoundland (Grant, 1974b). Additionally, extensive gravel and sand deposits exist on the seafloor on certain offshore banks of the Scotian Shelf south of Cape Breton Island. Some of these banks have relatively limited fishery potential and, using existing technology and practice in use elsewhere in United States and Europe, their surficial deposits might be recovered at competitive costs without the negative impact on shorelines that currently results from removal of sand and gravel from coastal beaches.

### Groundwater extraction

Developing subsurface water should take into account the commonplace layering of tills with quite different hydrological properties. In some areas, dense, less pervious tills underlie porous surface tills, such as over the southern coastal lowland, so there is less groundwater at depth than would be normally expected. Conversely, in some areas, such as along the East Bay coast and in the Inverness-Chéticamp area, there are gravels and other pervious sediments beneath fine grained tills, so there may be a greater potential for water withdrawal than would be expected from the surface material. In particular, the potential of buried valleys should not be overlooked in prospecting for major undeveloped aquifers. It is likely that some of the original alluvium is preserved beneath the till which presently fills the valley. If so, artesian conditions are a possibility. At present, water supplies come mostly from surface sources and these are adequate in quantity and quality, but as these are reduced by demand and pollution, attention to groundwater reserves will increase and there will be a need to take into account the large local differences in surficial geological setting in order to discover and develop these effectively.

### Waste disposal

The disposition of waste material (household garbage, industrial debris, sewage sludge, toxic chemicals, etc.) has become an increasingly urgent problem as it has grown in volume and toxicity. Until means are developed to dispose of wastes safely and permanently, the interim solution has been to store these on or in the landscape, euphemistically called landfills. The general practice is to dump the material in exhausted gravel pits and, to a lesser extent, into the sea. In both cases, movement of water through the waste causes harmful fluids to be leached out which then becomes part of the water system. To mitigate the hazard of contamination, consideration could be given to a more judicious choice of dumpsites. If gravel pits are to remain the dumpsite of choice, leachate could be effectively prevented from entering the groundwater system by applying a layer of impervious clay-rich till to the floor so as to form a sealed basin. Thick clay tills, perhaps beneficiated by addition of the clay that is widely associated with Carboniferous gypsum, would be a better choice of dump site, providing it was established by drilling that there were no pervious layers that could allow the leachate to escape. Clay tills are a practicable alternative to gravel pits because they are widely available over the island and generally underlie areas of modest productive value. The best storage site would be a natural peat-filled depression, first because groundwater throughput is minimal and secondly because peat is an adsorbant (e.g., Coupal, 1985). Peat bogs occur throughout the island and would be readily available because they are not presently used for other purposes. In all cases, it would be necessary to conduct a complete three-dimensional study of the surficial geological setting with particular attention paid to the extent and permeability of all units in the setting. The best setting is one in which both groundwater and leachate movement are slow. In that sense, the best way to store wastes would be to interlayer them with peat in a clay-lined natural basin in little-fractured rock. All of these requisites can be found in the surficial geological makeup of Cape Breton Island.

## *Quaternary processes as hazards and problems*

The Cape Breton Island landscape is essentially benign and without significant hazards. However, two presently active processes, coastal erosion and slope failure, pose problems in that they impact on environmental quality, public safety, and property values. Capable of destroying anything in their path, these processes are impossible to stop or to modulate to any significant extent. Therefore, in order to minimize their impact, it is important to try to understand the factors determining their distribution and occurrence. Management and possible mitigation of their effects will be mainly through avoidance; prevention is effectively impossible and at best can offer only temporary relief.

### Coastal erosion

The steady and insidious loss of land by coastal erosion may be the terrain process with the greatest, albeit not yet quantified, negative economic impact. Rising sea level is driving the shoreline inland at rates depending on the erodibility of the shoreline material, the general gradient of the coast, the depth of water, and wave exposure. Although not a life-threatening hazard, coastal erosion has an impact greater than the other hazards because of the economic loss. It has a significant negative impact on property values, investment, and production. Moreover, coastal erosion locally necessitates expensive remedial and defensive structures. The rates of retreat vary around the island as evident in the general configuration of the shoreline, which in turn can be related to its composition: shorelines composed of Quaternary sediment are generally concave because they are nonresistant; shorelines composed of crystalline rock are convex because they are resistant. Shorelines composed of bedrock that is both nonresistant and of low elevation are generally straight because they are retreating at intermediate rates. In the absence of direct measurement, the degree of concavity is thus a general indication of susceptibility to erosion and can be a guide to property valuation. Unfortunately the most vulnerable shorelines and those which are suffering the greatest damage, that is those which are composed of sediment and weak rock, are those which have attracted greatest economic investment, as around Canso Strait and Sydney area, or are the most preferred for vacation homes, as around Bras d'Or Lake.

The retreat of the shorefront, or in other words the loss of property depth, is to be expected based on simple geometric relationships and it can be estimated arithmetically. Rising sea level deepens the water; the extra depth is effectively a transitory void which tends to be filled with material eroded from the shore in order to maintain a stable offshore bottom profile. Thus, according to the Bruun Rule, coastal retreat is a function of submergence rate (Schwartz, 1967). If sea level is currently rising about 2570 cm/100 a, this will theoretically cause the shoreline to move landward at 1-3 metres per year. In support of this, airphotos reveal that the coastline has moved inland several tens of metres over the last 40-50 years. Over the longer term, Wang and Piper (1982) derive an average rate of coastal retreat for the Gabarus area of 0.5 m/a over the last 8000 years. In practical terms, these erosion rates suggest that a shoreline property 30 m deep could disappear in one generation; conversely, 30 m would be the prudent setback for a home with a 25 year mortgage. These very approximate observations are in the interest of providing insight into the main process affecting the coast in order to permit investment decisions regarding the life span of shoreline property.

### Slope failure

As discussed in the section Colluvial deposits and slope processes, slope failure in Cape Breton Island is largely restricted to the steep upland and highland slopes mapped as colluvial (unit 7) and involves the slow or sudden descent of till and rock debris into the sea and watercourses. The slow creep of massive sections of the western highland margin are simple failures evidently related to the unstable structural setting. Those at High Capes are presently in a wilderness area and so pose no immediate problem, apart from the remote possibility that a large enough mass might fall suddenly into the sea and cause a destructive sea wave. Those at Cap Rouge have affected the integrity of the Cabot Trail and have therefore necessitated extensive remedial measures (slope strengthening, highway realignment). Should the slope deteriorate unacceptably, consideration should be given to avoidance as an alternative strategy for coping. The highway could be repositioned higher up the slope (closer to one of its earlier routes). This option would provide a higher vantage point and thus further enhance the visual appeal of the route. On the other hand, it could be argued that, instead of being treated as an expensive liability, the problem could be transformed into a park interpretation feature, a curiosity of interest to passersby. Based on statistics showing that the 300 000 annual visitors spend about $70.00 per day during an average 3.7 day visit (Parks Canada, pers. comm., 1988), a roadside display that attracts a 15 minute stop generates nearly one million dollars per year in indirect spinoff benefit to the local economy. Interesting geological features clearly have a not inconsiderable economic value.

Sudden slope failures by avalanching, landsliding, and rockfall have quite different socio-economic consequences. Relatively small, but nonetheless lethal, falls of rock, till, and debris are occurring in great, but as yet uncounted, numbers each year on the steep slopes of the thousands of deep gorges that incise the highland and upland margins. The failures are simply the process by which the gorges are widened as the watercourses attempt to deepen and widen their channels. The failures are a natural consequence of the downcutting and are the means by which the slopes are backwasted and the material delivered to the streams below. Although the average frequency has been estimated in the section Colluvial deposits and slope processes at about one per 1000 years at any given point over the 10 000 year span of postglacial time, the determining factors are not understood, so prediction is impossible. That being the case, avoidance of these slopes is the only means of managing the hazard. Buildings should not be located on or at the bottom of such slopes and roads should not traverse them unless absolutely necessary. Trails in the national park should avoid these areas and visitors be made aware of the potential danger. Any undercutting of the natural talus slope, either naturally or artificially, will increase the slope instability. If excess water at the cliff break is the factor that increases their frequency and size, then the failure process will be exacerbated by deforestation and/or greater precipitation and spring weather conditions. Given that the biophysical environment has been in equilibrium with this process over the long term, loss of vegetation may be expected to increase their frequency which in turn will increase suspended matter in watercourses and thereby alter water quality for aquatic life. The slopes most susceptible to changes in porewater pressure are those topped with till on the eastern highlands.

Earthquakes are important factors in promoting slope failure. Although this is not to say that earthquakes can create landslides that otherwise would not have occurred, they can be expected to have the potential of advancing the date of failure of metastable slopes because the ground accelerations increase the effective load and lower the shear stress. While no slide has yet been linked to any earthquake in historic time, shock-induced failure of weak slopes or loose slope material, particularly when saturated, remains possible. Again, the slopes most likely to fail by seismic shaking, particularly when saturated, are those of the eastern gorges which are cut in thick till along much of their upper and inner reaches.

## *Quaternary terrains as potential parks*

Cape Breton Island is one of Nova Scotia's prime tourist areas because its outstanding scenery is enhanced by an equable summer climate. Demand for this vital and dwindling resource as a source of cottage property has increased dramatically in recent years, primarily on the shores of Bras d'Or Lake shore and southern Gulf of St. Lawrence coast. This section focuses on the remaining undeveloped or wilderness tracts all of which are in the coastal belt (Fig. 125) and identifies those which have such obvious recreational development potential that it may now be prudent to consider whether some should be preserved as parks. All of these coastal wildernesses have the necessary combination of natural attributes, are large enough to be viable, are near or readily accessible from main roads, and have proximity to an infrastructure that could serve travellers' needs. The coastal wilderness areas considered to having outstanding park potential are (in no particular order): the Mabou Highlands, northernmost Cape Breton Highlands, North and South ponds at Aspy Bay, the Cape Gabarus area, three southern parts of Isle Madame, Big and Little basins of Inhabitants Harbour, and Grand Anse of Lennox Passage. Their major natural features are outlined below; socio-economic evaluation is beyond the scope of this report.

Mabou Highlands can be considered a microcosm of Cape Breton Island's natural features as it combines most of the landscape types found in Nova Scotia. The highlands are elliptical plateau (≈100 km²) with a strikingly flat, prairie-like summit (Fig. 12), incised by a radial system of dendritic gorges clothed in almost virgin forest (Fig. 33). The eastern slopes open onto riverine lowlands with gentle meandering streams, whereas the western flank is fringed by a narrow coastal bench with pocket beaches. It is amenable to campgrounds and a network of trails to access its special geomorphological features. The main west-coast highway skirts the eastern margin and the towns of Inverness and Mabou lie at the north and south ends. The bedrock and Quaternary features record most of the island's geological history. As such, the Mabou Highlands would seem to be a prime candidate as Nova Scotia's largest provincial park.

The northernmost prong of the highlands (≈300 km²), extending to Cape North and Cape St. Lawrence, is perhaps the most wild and dramatic area of Nova Scotia. The gently rolling plateau is incised by 400 m deep gorges (Fig. 19, 65). The western or Gulf of St. Lawrence flank of the plateau block between Pleasant Bay and Pollett Cove has small lowlands corresponding to areas of Carboniferous sedimentary rock (Fig. 60). The narrow emerged fossil tidal bench can be followed almost continuously from Chéticamp to Cape

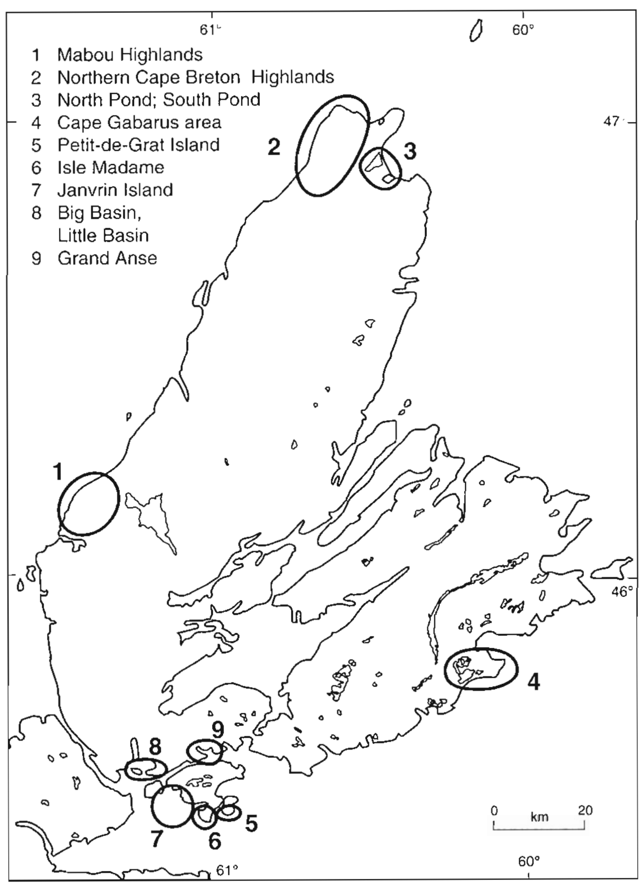



*Figure 125. Suggested potential park reserves.*

St. Lawrence. Aspy faultline scarp forms the eastern or Atlantic margin of the plateau block. Geomorphological features unique in Nova Scotia are the multicoloured sagging slopes (Fig. 79), the huge Aspy landslide (Fig. 76), and the cirques and glacial troughs (Fig. 65). The northwest coast is notable as Nova Scotia's main whale-watching area and both the Les Iles de la Madeleine in Gulf of St. Lawrence and the Long Range Mountains of Newfoundland, 120 km distant, can be seen from the plateau. The entire area is in an almost pristine state, except for small overgrown pioneer settlement areas at Pollett and Lowland coves. Visitors' needs can be served by commercial establishments in Pleasant Bay, Aspy River valley, and Bay St. Lawrence. Apart from the intrinsic value of preserving these unique aspects, the area abuts Cape Breton Highlands National Park which presently has 300 000-400 000 visitors annually, and would therefore be a logical extension to meet its future needs.

North Pond and South Pond areas (≈15 $km^2$) of Aspy Bay integrate most of the features desirable for water sports, in additional to general outdoor activities. Enclosed by sandy sea beaches several kilometres long with tidal inlets, the ponds are broad, generally placid lagoons of temperate water with fringing salt meadows and an intricate and interesting shoreline (Fig. 70, 72). They thus possess most of the features expected in a maritime setting. The great barrier beaches which created these special enclaves have unquestioned park potential. Parks in this area would benefit from the existing infrastructure in nearby villages of Dingwall and Cape North and would be well positioned to serve those tourists circumnavigating Cabot Trail who leave the main route to visit Cabot Landing and take whale-watching boat rides. The parks would offer facilities to sustain longer visits in the area.

The Cape Gabarus area on the Atlantic coast is a moderate-sized (300 $km^2$), but exceptional mosaic of varied terrain and vegetation. It has barren rocky areas, meadow- and forest-covered drumlins, bog and heath, marshes, an inland chain of lakes, and the island's largest coastal beach barrier of sand and gravel. With marine exposure on three sides, it possesses most of the natural features which draw most visitors to the province. The location offers benefits to the nearby Sydney conurbation and its proximity to Fortress Louisbourg National Historic Park, a major destination, would help diversify the present radial tourist route to a more circumferential route that would include the less well travelled south coast with its distinctive Acadian culture.

Three parts of Isle Madame area offer extensive tracts of interesting, varied, and virtually undisturbed terrain. In the bay between Janvrin Island and Isle Madame, and presently already traversed by the highway between the two, is a 5 $km^2$ tract of drumlins with numerous spits and connecting beaches (Fig. 26). The island-tombolo complex provides an extensive lee area of shallow water that is ideal for water sports, especially boating. The shallow water would permit easy construction of additional causeways, thus creating a unique island complex. In contrast, the two southernmost peninsulas are wide-open rugged headlands that offer rugged to rolling terrain with broad vistas and great scenic variety. Cape Hogan area (15 $km^2$) is generally rocky and heaped with heath-covered morainal ridges and mounds but with sheltered forested groves and several good pocket beaches (Fig. 26). The Heath Head-Red Head promontory [44] (10 $km^2$), forming the south end of Petit-de-Grat Island has a low flat rock substrate holding several lakes and ponds that are sheltered from the sea by one of the large east-west till ridges. The vegetation is mainly grassy meadows interspersed with forest groves. All three wilderness areas are conveniently served by the well-developed infrastructure in the several nearby communities. Establishing parks in this area as additional destinations for visitors would help raise the tourism profile of the southern part of Cape Breton Island and would thereby result in more frequent and longer visits to the island as a whole.

The Hawkesbury-St. Peters area has two potential park areas, judged on the basis of their scenic attributes and location. Inhabitants Harbour area (≈20 $km^2$) is an elliptical bay sheltered from the open ocean by drumlin islands and long sandstone ridges mounded with till hills. The shallow and relatively warm water, combined with both open expanses and an intricate shoreline, is ideally suited for all water sports, especially boating and sailboarding. Its proximity to the Strait of Canso urban area gives it added value. Grande Anse Bay (≈5 $km^2$), near St. Peters, the main yachting entrance to Bras d'Or Lake, offers similar features, but in a smaller area with deeper water and steeper surrounding till slopes. The proximity of both these wilderness areas to the island's gateway, to service centres, and to the main southern autoroute (Highway 104) easily qualify them as potential park sites which would, in addition, offer visitors an immediate recreational option within minutes of arriving at the island.

**Table 1.** Key stratigraphic sequences, including radiometric age estimates.

| Site No.[1] | Locality[2] | Main elements of stratigraphic sequence[3] (cited in stratigraphic order) | Material dated[4] | Species[5] | Age[6] | error[7] |
|---|---|---|---|---|---|---|
| 1 | Bay St. Lawrence | fluvial gravel; till?; colluvium; | | *Serripes* | | |
| | | **shelly silt** | shell | *groenlandicus* | 21 920 | ± 150 |
| | | | wood | *Picea sp.* | >49 000 | |
| | | **woody peat** | wood | *Picea sp.* | >46 000 | |
| | | gravel and rubble | wood | *Juniperus sp.* | 47 000 | ± 4700 |
| | | emerged wavecut rock platform | peat | | 44 200 | ± 820 |
| | | cut in siltstone bedrock | wood | *Picea or larix* | >38 270 | |
| 2 | Money Point | alluvial fan gravel | | | | |
| | | till with sedimentary-rock clasts; | | | | |
| | | rounded gravel (littoral?) | | | | |
| | | emerged wavecut rock platform | | | | |
| | | cut in granite bedrock | | | | |
| 3 | Aspy Fault | till(s) | | | | |
| | | gravel (littoral) | | | | |
| | | emerged wavecut platform 22 m | | | | |
| | | on Precambrian greenstone | | | | |
| | | —upfaulted 15 m | | | | |
| 4 | Dingwall | gravel | | | | |
| | | **organic silt** | | | | |
| | | brown till | | | | |
| | | red till | | | | |
| | | gravel (granitic clasts) with wood | wood | | >48 000 | |
| | | gravel (volcanic rock clasts) | wood | *Picea sp.* | 23 700 | ± 560 |
| | | **woody organic silt** | wood | *Picea sp.* | 32 700 | ± 560 |
| | | brown sandy till | wood | *Picea* or *Larix* | >39 000 | |
| | | sinkhole in gypsum bedrock | | | | |
| 5 | South Aspy River | alluvial (outwash) gravel | | | | |
| | | sand with grey silt | | | | |
| | | till (gravel?) with **organic** clasts | organic silt | | 24 900 | ± 700 |
| | | till (or gravel?) | | | 20 300 | ± 400 |

1 Refers to locations on Figure 30 and to numbered symbols, e.g., $+^{17}$, on Map 1631A, except sites marked with an asterisk (83*-94* which were added or studied after the map was published. The sites are numbered in general clockwise order around the island starting at the northern extremity.
2 Names refer to the nearest locality on published topographic maps; uncapitalized names are informal.
3 Abbreviated description only; for details, refer to text. Except as noted in column 3, all are natural exposures.
4 **shell**= tests of marine pelecypods and gastropods; **gyttja** = organic lake sediment; **peat** = plant debris
5 Species named in unpublished paleoecological reports (see note 12) and marine shell identifications by F.J. E. Wagner, Atlantic Geoscience Centre, and M.I. Smith, National Museum of Natural Sciences.
6 All ages are by the $^{14}C$ method using a half life of 5568 ± 30 years, except for those in **boldface** which are by the thorium/uranium disequilibrium method. GSC dates are corrected for $^{12}C/^{13}C$ fractionation relative to the PDB standard by normalizing to 0.0‰ for marine carbonate and to -25.0‰ for other organic matter, such as marine carbonate (shells). GSC shell age determinations are therefore 410 years younger than would be quoted by other labs. Underlined ages are considered anomalous (too old or too young) because of contamination, as discussed in text.
7 For finite GSC (Geological Survey of Canada) dates, the error term is based on a 2σ criterion (that is, the age is accurate to within two standard deviations) and for "greater than" dates it is 4σ; most other labs quote to 1σ.

| Lab no.[8] | Collector[9] | elevation (of base and top of section)[10] | Latitude[11] | Longitude[11] | Unpublished analytical report[12] | References[13] |
|---|---|---|---|---|---|---|
| | | **0-37** | 47°00.75' | 60°26.83' | WR/87-33 | Prest, 1957, p. 447; |
| TO-246 | DRG | | | | WR/82-40 | Prest, 1970, p. 679; |
| | | | | | WR/87-33 | Mott and Prest, 1967; |
| GSC-4487HP | RJM | | | | WR/71-07 | Newman, 1971; |
| GSC-3864HP | RJM | | | | WR/83-01 | Grant in Prest et al., 1972; |
| UQT-178 | | | | | | Guilbault 1982; |
| GSC-3636 | RJM | | | | | de Vernal et al. 1983; |
| GSC-283 | VKP | | | | | Mott and Grant, 1985; |
| | | | | | | Grant, 1987, p. 36; |
| | | | | | | McNeely and McCuaig, 1991 |
| | DRG | 0-10 | 47°01.5' | 60°23.5' | | Neale, 1963 |
| | DRG | **0-35** | 46°57.8' | 60°26.8' | | Neale, 1963; |
| | | | | | | Grant, 1975a, 1987, 1990 |
| | DRG | **0-12** | 46°54.05' | 60°27.02' | PR/81-1 | Newman, 1971; |
| | DRG | | | | WR/81-06 | Grant, 1987, p. 36 |
| | DRG | | | | WR/82-42 | |
| | DRG | | | | WR/82-43 | |
| GSC-3541-2 | DRG | | | | WR/82-44 | |
| GSC-3381-1 | DRG | | | | | |
| GSC-3381-2 | RJM | | | | | |
| GSC-3417 | RJM | | | | | |
| | | **5-23** | 46°52.41' | 60°30.06' | | Newman, 1971 |
| I-3414 | WMN | | | | | |
| I-2438 | WMN | | | | | |

8 GSC = Geological Survey of Canada; I = Isotopes Inc.; TO = University of Toronto accelerator / mass spectrometer facility; UQ = University of Québec at Montréal; UQT = University of Québec at Montréal Thorium/Uranium facility; W = United States Geological Survey; Y (now Ya) = Yale University. **HP** indicates a high pressure count (5 atmospheres).

9 Initials refer to D.R. Grant, F. Baechler, E.A. Goranson, K.N. Greenidge, D.A. Livingstone, R.H. MacNeill, A. MacRae, R.J. Mott, S. Occhietti, V.K. Prest, F. Seymour, G. Saint-Jean, R.R. Stea, Jaan Terasmae, Anne de Vernal, New Jersey Zinc Corp.

10 **Datum is present mean annual high tide level.** (Mean Sea Level is about one metre lower). Elevations in boldface were mainly determined by tape measure, some by aneroid altimeter; all others interpolated and converted from topographic maps with 50 foot (≈15 m) contours.

11 Note: coordinates are to 0.01 minute, not seconds.

12 These are the file numbers for unpublished analytical reports that are archived in the Geological Survey of Canada, Paleoecological Database. PR = Palynological report (by J. Terasmae, R.J. Mott); WR = wood identification report (by R.J. Mott, L.D. Wilson, H. Jetté); PM = plant macrofossil report (by J.V. Matthews Jr.); BR = bryological (fossil moss) report (by M. Kuc); FA = fossil arthropod report (by J.V. Matthews Jr.); DR = Diatom report (by S. Lichti-Federovich)

13 Refers to previous publications containing any part of the data and/or any interpretation of it.

Table 1. (cont.)

| Site No.[1] | Locality[2] | Main elements of stratigraphic sequence[3] (cited in stratigraphic order) | Material dated[4] | Species[5] | Age[6] | error[7] |
|---|---|---|---|---|---|---|
| 6 | Aspy scarp | massive landslide of recent age;<br>talus (rock-glacierized?) | | | | |
| 7 | Ingonish | reindeer antler in ice-contact? gravel | bone | *Rangifer tarandus* | 0 | ± 220 |
| 8 | Morrison Brook | outwash gravel<br>grey till, volcanic clasts (from south)<br>red till, granite clasts (from west)<br>emerged wavecut rock platform<br>Carboniferous arkose | | | | |
| 9 | Wreck Cove pond | 4 m core of **Holocene lake sediment** | gyttja | | 9 030 | ± 170 |
| 10 | Wreck Cove head | outwash gravel<br>red granitic rock till<br>brown sedimentary rock till<br>granite bedrock stossed 068° | | | | |
| 11 | Wilhausen Point | till<br>**sand with organic lenses**<br>gravel<br>siltstone bedrock | | | | |
| 12 | Oyster Pond | gravel (outwash?)<br>**organic clay/silt** | | | | |
| 13 | McLeod Point | red clay till with **marine shells**<br>gypsum bedrock | | | | |
| 14 | Baddeck | colluvium/till<br>**peat and humus**<br>sandstone bedrock? | | | <5 000<br>(palynological estimate) | |
| 15 | Middle River | Holocene gravel alluvium with **femur** of *Mammut americanum* | bone<br>bone | *collagen*<br>*apatite* | 31 300<br>32 000 | ± 500<br>± 630 |
| 16 | Mullach Brook | local red silt/clay till<br>polymictic grey-brown sand till<br>local red silt/clay till<br>regolith<br>Carboniferous siltstone bedrock | | | | |
| 17 | Whycoco-magh | ice-contact gravel?<br>polymictic grey-brown sand till<br>sand; organic clay/silt<br>**organic sediment with wood**<br>clay/silt<br>local red silt/clay till | wood | *Picea sp.* | >44 000 | |
| 18 | Portage Creek | red sandy till, sedimentary clasts<br>brown silty till, crystalline clasts<br>gypsum | | | | |

| Lab no.[8] | Collector[9] | elevation (of base and top of section)[10] | Latitude[11] | Longitude[11] | Unpublished analytical report[12] | References[13] |
|---|---|---|---|---|---|---|
| | | 150-300 | 46°51.5' | 60°35.5' | | |
| GSC-1395 | FS | **20** | 46°39.2' | 60°23.7' | PL/70-1 | |
| | DRG | **0-25** | 46°34.3' | 60°23.5' | | Grant, 1987, p. 38 |
| GSC-335 | DAL | 335 | 46°32.37' | 60°26.75' | | Livingstone and Estes, 1967 |
| | DRG | **0-30** | 46°32.1' | 60°24.9' | | |
| | DRG | **0-14** | 46°18.03' | 60°31.56' | PR/74-3;<br>BR-286 | |
| | DRG | **3-5** | 46°17.83' | 60°31.35' | | |
| | DRG | 0-15 | 46°13.42' | 60°35.83' | | |
| | DRG | 43-47 | 46°07.28' | 60°41.95' | PR/75-14<br>PM/1-75 | |
| GSC-1220-2<br>GSC-1220 | AM | 17 | 46°08.1' | 60°55.2' | | Dawson, 1868;<br>Piers, 1915;<br>Grant, 1987, p. 39 |
| | | **74-75** | 46°00.3' | 61°08.4' | | Tang, 1970 |
| GSC-290 | VKP | **5-10** | 45°58.5' | 61°07.0' | WR/83-02 | Prest, 1957, p. 446;<br>Prest, 1970, p. 680;<br>Mott and Prest, 1967 |
| | DRG<br>DRG | **0-5** | 45°57.0' | 60°59.5' | | |

**Table 1.** (cont.)

| Site No. [1] | Locality[2] | Main elements of stratigraphic sequence[3] (cited in stratigraphic order) | Material dated[4] | Species[5] | Age[6] | error[7] |
|---|---|---|---|---|---|---|
| 19 | Benacadie Point | till? with shell fragments | | | | |
| | | silt and clay | peat | | 10 900 | ± 100 |
| | | **organic sediment with wood** | peat | | 11 300 | ± 90 |
| | | sand | peat | | 11 670 | ± 170 |
| | | emerged wavecut rock platform | peat | | 12 100 | ± 100 |
| | | conglomerate bedrock | | | | |
| 20 | Benacadie shore 2 | pink-brown till with shells | | | | |
| | | **peat** | peat | | 9 560 | ± 235 |
| | | sand | | | | |
| | | dark purplish-brown till | | | | |
| 21 | Benacadie shore 3 | sandy till | | | | |
| | | red silt till with **shell** fragments | | | | |
| | | **organic silt** | | | | |
| | | oxidized granitic rock gravel | | | | |
| 22 | Benacadie shore 4 | red till (soliflucted?)with shells | | | | |
| | | gravel | | | | |
| | | **organic silty sand with wood** | wood | | 12 310 | ± 150 |
| | | cobble gravel | | | | |
| | | purplish-brown till with **shells** | | | | |
| | | oxidized granitic rock gravel | | | | |
| | | in depression (gypsum sinkhole?) | | | | |
| 23 | Amaguadees | sand | | | | |
| | | stony colluvium (soliflucted till) | | | 10 400 | ± 120 |
| | | **organic** silt; **peat** with **twigs** | peat | | 10 850 | ± 700 |
| | | sand with slumped till tongue | | | | |
| | | | peat | | 11 250 | ± 200 |
| | | **peat** | peat | | 11 700 | ± 100 |
| | | silt and sand | peat | | 12 600 | ± 120 |
| | | red till with **shell** fragments | peat | | 12 200 | ± 140 |
| | | depression (sinkhole in gypsum?) | peat | | 11 700 | ± 130 |
| 24 | Amaguadees head | red till with **shell** fragments | | | | |
| | | purplish-brown till | | | | |
| 25 | Castle Bay | till, reddish with shell fragments | | | 11 100 | ± 170 |
| | | **organic silt** | silt | | 13 030 | ± 1270 |
| | | brown till | | | | |
| | | stony silt with **wood** fragments | wood | *Picea or Larix* | >52 000 | |
| | | sand (delta foreset beds?) | | | | |
| | | **organic silt** | silt | | >42 000 | |
| | | granitic gravel (littoral?) | | | | |
| 26 | Benacadie shore 1 | pebbly colluvium (ablation till?) | | | | |
| | | pink-brown till with shells | | | | |
| | | laminated **organic silt** | silt | | 11 370 | ± 290 |
| | | cobble gravel | | | 11 530 | ± 390 |
| | | dark purple-brown till | | | | |

| Lab no.[8] | Collector[9] | elevation (of base and top of section)[10] | Latitude[11] | Longitude[11] | Unpublished analytical report[12] | References[13] |
|---|---|---|---|---|---|---|
| GSC-3912 | | **0-2** | 45°54.16' | 60°43.68' | BR-317 | MacNeill, 1969; |
| GSC-2146 | RJM | | | | | Grant, 1975a; |
| I-3234 | DRG | | | | | Mott et al., 1986 |
| GSC-3900 | RHM | | | | | McNeely and McCuaig, 1991 |
| | SO | | | | | |
| | SO | **0-10** | 45°54.02' | 60°42.46' | | |
| UQ-236 | | | | | | |
| | | **0-13** | 45°54.08' | 60°42.02' | | Grant, 1987, p. 42 |
| | DRG | | | | | |
| | DRG | | | | | |
| | | **0-14** | 45°54.26' | 60°41.30' | | Mott et al., 1986 |
| UQ-394 | SO | | | | | |
| | DRG | | | | | |
| | DRG | | | | | |
| | | **0-9** | 45°54.53' | 60°40.22' | | McNeely and Jorgersen, 1992; |
| GSC-4063 | RJM | | | | | S. Occhietti, pers. comm., 1986; |
| UQ-985 | SO | | | | | McNeely and Jorgersen, 1992 |
| UQ-970 | SO | | | | | |
| UQ-676 | SO | | | | | |
| GSC-4062 | RJM | | | | | |
| GSC-4973 | RJM | | | | | |
| GSC-4974 | RJM | | | | | |
| | | **0-17** | 45°54.47' | 60°39.78' | | R.J. Mott, pers. comm., 1985 |
| UQ-831 | SO | **0-32** | 45°55.22' | 60°38.75' | WR/71-62 | Grant, 1972; |
| UQ-246 | SO | | | | | Mott and Grant, 1985; |
| | | | | | | de Vernal and Mott, 1986; |
| GSC-1619 | DRG | | | | | de Vernal et al., 1986 |
| GSC-1577 | DRG | | | | | |
| | DRG | | | | | |
| | DRG | **5-7** | 45°53.97' | 60°42.68' | PR/83-29 | Mott et al., 1986 |
| | DRG | | | | | |
| UQ-404 | SO | | | | | |
| UQ-395 | SO | | | | | |
| | DRG | | | | | |

Table 1. (cont.)

| Site No. [1] | Locality[2] | Main elements of stratigraphic sequence[3] (cited in stratigraphic order) | Material dated[4] | Species[5] | Age[6] | error[7] |
|---|---|---|---|---|---|---|
| 27 | East Bay | gravel | | | | |
| | | red till with rare shell fragments | | | | |
| | | **organic silt** | wood | | 51 000 | ± 3000 |
| | | brown till with rare **shells** | wood | *Picea sp.* | 62 100 | ± 5000 |
| | | gravel/sand; **organic lenses** | wood | *Picea sp.* | >50 000 | |
| | | **organic silt; peat with wood** | wood | *Tsuga canadensis* | >49 000 | |
| | | rock rubble, oxidized | wood | *Juniperus sp.* | >50 000 | |
| | | sinkhole in gypsum bedrock | wood | *Tsuga canadensis* | 86 900 | ± 6000 |
| | | | wood | | 65 900 | ± 3200 |
| | | | wood | *Pinus strobus* | 101 900 | ± 5400 |
| | | | wood | | 123 400 | ± 30 000 |
| | | | wood | | 126 400 | ± 15 000 |
| 28 | McAdams Lake | fluvial gravel (outwash?)<br>till with **clasts of woody peat** | wood | *Betula sp.* | 1 560 | ± 70 |
| 29 | McDougall Lake | 6.5 m core in **Holocene lake sediment** | | | | |
| 30 | Upper Gillies Lake | 12 m core in **Holocene lake sediment** | | | | |
| 31 | Blacketts Lake | reddish silty till | | | | |
| | | **peat** | peat | | 10 300 | ± 150 |
| | | greyish sandy till | | | | |
| 32 | Leitches Creek | 50 m drill core containing:<br>till, 15 m | | | | |
| | | **org. silt, woody peat,** 3.6 m | peat | | >52 000 | |
| | | till, 9 m<br>**woody peat**<br>gravelly sand, boulders | | | | |
| 33 | Amelia Point | sand (alluvium?) with **organics** | | | | |
| 34 | Donkin | **peat bog** | peat | | 9 590 | ± 160 |
| 35 | Deep Cove | gravel and red silt<br>red till<br>striations 055° on glacial facet on emerged wavecut platform cut in hematitic breccia (Quaternary?)<br>Paleozoic bedrock | | | | |
| 36 | Gabarus lighthouse | grey till<br>red till<br>bedrock with two striated glacial facets: 110° (with stain) crosscut by 030° (fresh) | | | | |

| Lab no.[8] | Collector[9] | elevation (of base and top of section)[10] | Latitude[11] | Longitude[11] | Unpublished analytical report[12] | References[13] |
|---|---|---|---|---|---|---|
| | DRG | **0-12** | 45°59.10' | 60°28.82' | WR/84-42 | Mott and Grant, 1985; |
| | DRG | | | | WR/83-24 | de Vernal et Mott, 1986; |
| UQT-188 | SO | | | | WR/83-23 | de Vernal et al., 1986 |
| UQT-177 | SO | | | | | McNeely and McCuaig, 1991 |
| GSC-3878HP | RJM | | | | | |
| GSC-3871HP | RJM | | | | | |
| GSC-3861HP | RJM | | | | | |
| UQT-109 | SO | | | | | |
| UQT-119 | SO | | | | | |
| UQT-108 | SO | | | | | |
| UQT-176 | SO | | | | | |
| UQT-175 | SO | | | | | |
| GSC-2058 | DRG | 76 | 46°01.50' | 60°24.76' | | |
| | DAL | 150 | 46°03.1' | 60°25.9' | | Livingstone, 1968 |
| | DAL | 80 | 46°04.05' | 60°23.64' | | Livingstone, 1968 |
| | DRG | 50 | 46°02.66' | 60°22.50' | BR-135 | Grant, 1972, 1975a |
| GSC-1578 | | | | | | |
| | | -20 +40 | 46°09.20' | 60°22.87' | PR/71-10 | Prest, 1970, p. 680; Mott and Prest, 1967; |
| GSC-2678 | EAG (NJZ) | | | | | R.J. Mott, in Blake, 1984 |
| | FB | 7 | 46°08.95' | 60°13.22' | PR/85-7 | |
| I-2477 | KNG | **0-15** | 46°10.88' | 59°49.25' | | |
| | DRG | **0-10** | 46°52.93' | 60°08.05' | | Williams, in press |
| | DRG | **0-5** | 45°50.65' | 60°08.96' | | Prest et al., 1972, p. 45; Grant, 1987, p. 43 |

Table 1. (cont.)

| Site No.[1] | Locality[2] | Main elements of stratigraphic sequence[3] (cited in stratigraphic order) | Material dated[4] | Species[5] | Age[6] | error[7] |
|---|---|---|---|---|---|---|
| 37 | Gabarus Cove | red till<br>two striated glacial facets: 090° and 000° on emerged wavecut rock platform cut in hematitic breccia (Quaternary?)<br>bedrock | | | | |
| 38 | Fourchu | red till<br>bedrock with three crosscutting glacial facets: 080° (stained) / 170° (stained) / 030° (fresh) | | | | |
| 39 | Enon | reddish-brown sandy till<br>greyish-brown silty till<br>celestite bedrock | | | | |
| 40 | Gillis Lake | 6 m core in Holocene lake sediment | gyttja<br>gyttja<br>gyttja<br>gyttja | | 12 000<br>10 160<br>13 450<br>13 900 | ± 130<br>± 160<br>± 260<br>± 160 |
| 41 | Salmon River lake | 4.5 m core in Holocene lake sediment; 260—280 cm interval | gyttja<br>gyttja | | 5 540<br>8 770 | ± 140<br>± 150 |
| 42 | Cap La Ronde | grey sandstone till<br>red mudstone till with rare **shell** | | | | |
| 43 | Rocky Bay | grey till<br>red **shell**-bearing till over bedrock stossed 110°, 340°, 160° | | | | |
| 44 | Heath Head / Red Head | grey till<br>red **shell**-bearing till over bedrock stossed 070° and 170° | | | | |
| 45 | Janvrin Island | red clay till (northward fabric) with fragments of temperate-water marine **fossils** | shell | | >34 000 | |
| 46 | Haddock Harbour | red clay till (northwest fabric) with fragments of temperate-water fossil marine **shells** | | | | |
| 47 | Grantville | esker gravel with fragments of temperate-water marine **shells** | shell | *Mercenaria sp.?* | 32 100 | ± 900 |
| 48 | River Inhabitants | **shell**-bearing red till<br>gravel<br>**peat with wood**<br>rubble colluvium<br>siltstone bedrock | wood | *Picea sp.* | >49 000 | |

| Lab no.[8] | Collector[9] | elevation (of base and top of section)[10] | Latitude[11] | Longitude[11] | Unpublished analytical report[12] | References[13] |
|---|---|---|---|---|---|---|
| | DRG | **0-7** | 45°50.28' | 60°09.22' | | Williams, in press |
| | DRG | **0-12** | 45°42.8' | 60°15.0' | | Prest et al., 1972, p. 45;<br>Grant, 1987, p. 43 |
| | DRG | 60 | 45°48.3' | 60°32.5' | | |
| GSC-4230<br>Y-524<br>Y-525<br>GSC-4246 | RJM<br>DAL<br>DAL<br>RJM | 58 | 45°39.8' | 60°46.5' | | Livingstone and Livingstone, 1958;<br>Deevey, 1958;<br>McNeely and McCuaig, 1991 |
| GSC-791<br>GSC-336 | DAL<br>DAL | 38 | 45°38.7' | 60°46.5' | | Livingstone, 1968 |
| | DRG | **0-17** | 45°34.32' | 60°53.68' | | |
| | DRG | **0-15** | 45°34' | 60°54' | | Grant, 1971a |
| | DRG | **0-30** | 45°29' | 60°56' | | Grant, 1971a |
| GSC-1639 | DRG | **0-3** | 45°31.31' | 60°06.75' | | Grant, 1971a, 1987 |
| | DRG | **0-4** | 45°31.78' | 61°08.20' | | Grant, 1971a |
| GSC-1408 | DRG | 10-20 | 45°38.50' | 61°14.20' | | Grant, 1971a; 1987, p. 31 |
| GSC-1406-2 | DRG | **53-65** | 45°40.57' | 61°19.58' | WR/71-13<br>WR/71-14<br>WR/83-31<br>WR/83-26 | Dawson, 1855;<br>Prest, 1957, p. 446;<br>Prest, 1970, p. 680;<br>Grant, 1971a; 1987, p. 29;<br>Mott, 1971 |

**Table 1.** (cont.)

| Site No.[1] | Locality[2] | Main elements of stratigraphic sequence[3] (cited in stratigraphic order) | Material dated[4] | Species[5] | Age[6] | error[7] |
|---|---|---|---|---|---|---|
| 49 | Big Brook quarry | stony diamicton (till, colluvium) | | | | |
| | | **organic silt** (lacustrine?) | bulk silt | | 11 000 | ± 90 |
| | | grey brown clayey till | | | | |
| | | grey-red clayey till with rare shells | wood | *Picea sp.* | >52 000 | |
| | | brown till with **wood** fragments | wood | *Picea sp.* | >49 000 | |
| | | grey **organic** silty sand | | | | |
| | | gypsum rubble | | | | |
| | | yellow, oxidized till (paleosol?) | | | | |
| | | grey **organic** sandy silt | silt | | 36 200 | ± 1280 |
| | | sinkhole in gypsum bedrock | | | | |
| 50 | Campbell Brook | gravel (proglacial lake delta) with temperate-water marine **shell** fragments | shell | *Mercenaria mercenaria* | | |
| 51 | Long Point | **peat** layers in sand in kettle in ice-contact sand | | | | |
| 52 | Chisholm Brook | Holocene fluvial gravel | | | | |
| | | **organic debris** | | | | |
| 53 | Campbell | sand | moss | | 10 800 | ± 100 |
| | | **peat** | peat | | 11 200 | ± 110 |
| | | red silty till | peat | | 11 300 | ± 100 |
| 54 | Mull River | **black organic silt** | silt | | | |
| | | sand and gravel outwash | | | | |
| 55 | Port Hood Island | sand | | | 11 000 | ± 170 |
| | | diamicton (soliflucted till) | wood | *Salix sp.* | 11 300 | ± 160 |
| | | **organic sediment** (incl. wood) | wood | *Abies balsamea* | 10 710 | ± 240 |
| | | sinkhole in gypsum bedrock | wood | | | |
| 56 | Mabou beach ("Hogs Back") | red-brown till | | | | |
| | | peat with wood | wood | | 10 550 | ± 100 |
| | | pinkish brown till | | | | |
| 57 | Green Point | red silty till | | | | |
| | | brown clayey till | wood | *Juniperus sp.* | >53 000 | |
| | | **woody peat** | wood | *Juniperus sp.* | 117 400 | ± 10 000 |
| | | gravel (weathered) | | | | |
| | | emerged wavecut rock platform | | | | |
| | | conglomerate bedrock | | | | |
| 58 | Northeast Mabou River | red diamicton (till?) | | | | |
| | | gravel outwash | | | | |
| | | silt (oxidized) | | | | |
| | | **woody peat** | wood | *Abies balsamea* | >53 000 | |
| | | grey clay with gravel lens | | | | |

| Lab no.[8] | Collector[9] | elevation (of base and top of section)[10] | Latitude[11] | Longitude[11] | Unpublished analytical report[12] | References[13] |
|---|---|---|---|---|---|---|
| GSC-3378<br>GSC-3880HP<br>GSC-3289<br><br><br><br>GSC-3206 | RJM<br>SO<br>RRS<br>DRG | **30-40** | 45°48.4' | 60°12.9' | PR/80-12<br>PR/81-6<br>PR/83-26<br>PR/84-3<br>WR/84-29<br>WR/81-15<br>WR/84-21<br>WR/83-25 | Mott et al., 1986;<br>Grant, 1987, p. 30;<br>McNeely and McCuaig, 1991 |
| | DRG | 30 | 45°49.28' | 61°10.30' | | Grant, 1971a, 1987, p. 31 |
| | DRG | **0-5** | 45°18.25' | 61°29.80' | PR/75-13<br>FA/75-2 | |
| | DRG | **0-5** | 45°48.64' | 61°29.64' | | |
| GSC-3892<br>GSC-2212<br>GSC-3781 | RJM<br>DRG<br>RJM | **6-7** | 45°50.25' | 61°29.70' | PR/75-12<br>FA/75-1 | Grant <u>in</u> Blake, 1984;<br>Mott et al., 1986;<br>Grant, 1987, p. 30;<br>McNeely and McCuaig, 1991 |
| | DRG | 48 | 46°00.02' | 61°22.18' | PR/75-11 | |
| GSC-540<br>GSC-541<br>Y-762 | JT<br>JT<br>JT | **18** | 46°01.18' | 61°34.39' | WR/75-83<br>WR/73-12 | Terasmae, 1974;<br>Grant, 1975a |
| UQ-906 | GSJ | 0-25 | 46°04.64' | 61°29.10' | | S. Occhietti, pers. comm., 1990 |
| GSC-3320<br>UQT-181 | DRG<br>SO | **0-15** | 46°05.37' | 61°28.58' | WR/83-28<br>WR/83-29 | Mott and Grant, 1985;<br>de Vernal et al., 1986;<br>Causse and Hillaire-Marcel, 1986;<br>Grant, 1987, p. 32 |
| GSC-3317 | DRG | **0-15** | 46°05.17' | 61°24.52' | WR/85-10 | Grant, 1987, p. 31 |

Table 1. (cont.)

| Site No. [1] | Locality[2] | Main elements of stratigraphic sequence[3] (cited in stratigraphic order) | Material dated[4] | Species[5] | Age[6] | error[7] |
|---|---|---|---|---|---|---|
| 59 | Hillsboro (formerly Hillsborough) | red clayey till<br>silt<br>**peat with wood**<br>sand<br>gravel | <br>wood<br>wood<br>wood | | <br>>21 000<br>>38 000<br>>51 000 | |
| 60 | Sight Point | red stony silt (till?)<br>rounded (littoral?) gravel<br>striations 040°, 090°, 145° on emerged wavecut platform cut in volcanic bedrock | | | | |
| 61 | Broad Cove Marsh | red silty diamicton (till?)<br>gravel with ice-wedge cast<br>brown sandy till<br>sandstone bedrock | | | | |
| 62 | Margaree Forks | kame gravel<br>red proglacial lake silt<br>till | | | | |
| 63 | Second Fork Brook Road South | colluvium<br>red-brown granitic till<br>saprolite sand<br>saprolite developed in granitic bedrock | | | | |
| 64 | Belle Côte | red stony diamicton (till?)<br>kame gravel<br>red sandy till<br>red sandstone bedrock | | | | |
| 65 | Terre Noire | grey sandstone-clast till<br>red sandy silt<br>gravel (kame?)<br>red stony silt (till?)<br>grey sandstone bedrock | | | | |
| 66 | Grand Étang | fluvial gravel<br>red silty (loess?)<br>gravel (ice-contact?)<br>red silt (glaciolacustrine?)<br>grey-purple till, mixed lithology<br>sandstone bedrock | | | | |
| 67 | Ruisseau du Lac | red stony silt<br>kame gravel<br>brown granitic rock till<br>grey sandstone bedrock | | | | |
| 68 | Chéticamp River | Holocene **peat bog**<br>red silt (loess?)<br>gravel outwash | | | | |

| Lab no.[8] | Collector[9] | elevation (of base and top of section)[10] | Latitude[11] | Longitude[11] | Unpublished analytical report[12] | References[13] |
|---|---|---|---|---|---|---|
| Y-232<br>W-157<br>GSC-370 | DAL<br>DAL<br>VKP | **2-7** | 46°04.37' | 61°22.06' | WR/83-03 | Prest, 1957, p. 446;<br>Prest, 1970, p. 680;<br>Mott and Prest, 1967;<br>Livingstone, 1968;<br>Grant, 1987, p. 31 |
| | DRG | **3-15** | 46°11.1' | 61°25.6' | | |
| | DRG | **15-17** | 46°17.1' | 61°16.0' | | |
| | DRG | 5-30 | 46°19.9' | 61°05.7' | | E.H. Muller, unpublished reports, 1964, 1965 |
| | DRG | 450-453 | 46°28.30' | 60°49.20' | | McKeague et al., 1983;<br>Grant, 1987, p. 38 |
| | DRG | **10-15** | 46°27.94' | 61°05.68' | | |
| | DRG | **0-15** | 46°29.15' | 61°04.90' | | |
| | DRG | **0-17** | 46°33.49' | 61°02.38' | DR/77-115 | Grant, 1987, p. 33 |
| | DRG | **0-15** | 46°34.13' | 61°02.00' | | |
| | DRG | **0-10** | 46°39.68' | 60°58.56' | DR/77-116 | Grant, 1987, p. 34 |

Table 1. (cont.)

| Site No.[1] | Locality[2] | Main elements of stratigraphic sequence[3] (cited in stratigraphic order) | Material dated[4] | Species[5] | Age[6] | error[7] |
|---|---|---|---|---|---|---|
| 69 | Jerome Brook | grey till<br>pink till<br>gravel (littoral?)<br>emerged wavecut platform<br>bedrock | | | | |
| 70 | Cap Rouge | fluvial fan gravel<br>red silty till<br>rounded gravel (littoral?)<br>sandstone bedrock stossed 110° | | | | |
| 71 | Fishing Cove River barrens | peat<br>wood<br>cobbles<br>gyttja<br>till<br>bedrock | gyttja | | 2 050 | ± 130 |
| 72 | Mackenzie River | red silt<br>fluvial gravel (outwash?)<br>emerged wavecut platform<br>(weathered) schist bedrock | | | | |
| 73 | Pleasant Bay | gravel<br>organic sediment in kettle<br>ice-contact sandy gravel<br>granitic-rock till<br>kame gravel | organic sediment | | 11 600 | ± 100 |
| 74 | Red River | fluvial fan gravel (outwash?)<br>red silt (till?)<br>rounded gravel (littoral?)<br>emerged wavecut rock platform | | | | |
| 75 | Pollett Cove | fluvial gravel (outwash)<br>red silt (till?)<br>kame moraine of valley glacier | | | | |
| 76 | Delaney Brook | red stony silt<br>emerged wavecut rock platform | | | | |
| 77 | Lowland Cove | red stony silt (till?)<br>emerged wavecut rock platform stossed 120° | | | | |
| 78 | McDougall Pond | Holocene lacustrine and marine sediment in coastal tidal pond | gyttja<br>veg. debris<br>veg. debris<br>organic mud | | 2 080<br>2 680<br>2 910<br>2 710 | ± 100<br>± 160<br>± 75<br>± 250 |

| Lab no.[8] | Collector[9] | elevation (of base and top of section)[10] | Latitude[11] | Longitude[11] | Unpublished analytical report[12] | References[13] |
|---|---|---|---|---|---|---|
| | DRG | **0-20** | 46°40.7' | 60°57.7' | | |
| | DRG | **0-10** | 46°42.7' | 60°56.4' | | Grant, 1987, p. 34 |
| GSC-281 | VKP | 410 | 46°45.5' | 60°50.9' | | Dyck et al., 1966 |
| | | **0-15** | 46°49.46' | 60°50.30' | | Grant, 1987, p. 34 |
| UQ-971 | DRG<br>SO | **0-15** | 46°49.51' | 60°49.55' | | S. Occhietti, pers. comm., 1987;<br>Grant, 1987, p. 34 |
| | DRG | **0-8** | 46°50.4' | 60°47.2' | | Grant, 1987, p. 34 |
| | DRG | **0-7** | 46°54.9' | 60°41.6' | | Grant, 1987, p. 34 |
| | DRG | **5-10** | 46°57.8' | 60°40.5' | | |
| | DRG | **7** | 47°01.0' | 60°37.5' | | Neale, 1963 |
| UQ-768<br>UQ-616<br>UQ-417<br>QC-1307 | ADV | **0- -5.5** | 47°00.03' | 60°27.94' | | de Vernal et al., 1985 |

**Table 1.** (cont.)

| Site No. [1] | Locality[2] | Main elements of stratigraphic sequence[3] (cited in stratigraphic order) | Material dated[4] | Species[5] | Age[6] | error[7] |
|---|---|---|---|---|---|---|
| 79 | "MacInnis lake" | core in Holocene lake sediment in kettle (sinkhole?) in outwash | gyttja | | 10 900 | ± 110 |
| 80 | Kilkenny Lake | ≈4 m core in Holocene lake sediment in depression in till | | | | |
| 81 | Third Lake O'Law | core in Holocene lake sediment (basal 15 cm of 1165 cm core) | gyttja | | 12 400 | ± 130 |
| 82 | Collins Pond (St. Francis Harbour) | till<br>**peat**<br>till | peat<br>peat<br>peat | | 10 900<br>11 800<br>12 700 | ± 100<br>± 100<br>± 130 |
| 83* | Moose Point | till with northwestward fabric<br>till with eastward fabric<br>**peat with wood**<br>emerged wavecut rock platform<br>Carboniferous bedrock | wood | *Picea sp.* | >49 000 | |
| 84* | Mulgrave bog | | | | | |
| 85* | Glasgow Brook bog | | | | | |
| 86* | Trout Brook ("highland") bog | 1.2 m of peat | | | | |
| 87* | Margaree Forks bog | 4.9 m of surface peat over pond sediment in sinkhole in gypsum | | | | |
| 88* | Cunningham bog | 6.4 m of surface peat in depression in till | | | | |
| 89* | Pembroke Lake | core in Holocene lake sediment | gyttja | | 10 700 | ± 190 |
| 90* | Cormier Lake | core in Holocene lake sediment; 584-586 cm interval | gyttja<br>gyttja | | 10 700<br>9 970 | ± 170<br>± 80 |
| 91* | Timber Lake | core in Holocene lake sediment; 245—248 cm interval | gyttja | | 11 200 | ± 200 |
| 92* | Hector Lake | core in Holocene lake sediment | gyttja | | 13 400<br>9 810<br>11 910 | ± 170<br>± 90<br>± 90 |
| 93* | Baddeck bog | basal 4 cm of organic silt in 610 cm core in peat bog | silt | | 9 050 | ± 100 |
| 94* | Yava (Pb deposit) | three superposed tills studied for fabric and geochemistry; deposited by NE- and NNE-flowing ice. | | | | |

| Lab no.[8] | Collector[9] | elevation (of base and top of section)[10] | Latitude[11] | Longitude[11] | Unpublished analytical report[12] | References[13] |
|---|---|---|---|---|---|---|
| GSC-4656 | RJM | ≈16 | 46°28.97' | 60°26.88' | MS-86-21 | McNeely and Jorgersen, 1992 |
| | JGO | ≈30 | 46°12.5' | 60°08.6' | | J.G. Ogden III, pers. comm., 1992 |
| GSC-4414 | RJM | ≈96 | 46°15.04' | 60°57.12' | | McNeely and Jorgersen, 1993 |
| GSC-4475<br>GSC-4367<br>GSC-4474 | RRS<br>RRS<br>RRS | **0-17** | 45°25.97' | 61°19.90' | | Grant, 1987, p. 28;<br>Stea and Mott, 1989;<br>McNeely and McCuaig, 1991 |
| GSC-4419HP | RJM | **0-15** | 45°24.1' | 61°27.0' | WR/87-15 | Stea and Mott, 1990;<br>Grant, 1987, p. 28;<br>McNeely and McCuaig, 1991 |
| | | ≈150 | 45°36' | 61°25' | | Auer, 1933 |
| | | ≈40 | 46°52.1' | 63°28.9' | | Hall, 1949 |
| | | ≈243 | 46°51' | 60°25' | | Killam, 1951 |
| | | ≈30 | 46°18.6' | 61°07.0' | | Killam, 1951 |
| | | ≈35 | 46°09.0' | 60°55.5' | | Killam, 1951 |
| GSC-5185 | RJM | ≈365 | 46°30.1' | 61° 00.0' | | R.J. Mott, pers. comm., 1992 |
| GSC-5275<br>Beta-61401 | RJM | ≈50 | 46°29.53' | 61°04.00' | | R.J. Mott, pers. comm., 1992 |
| GSC-5259 | RJM | ≈350 | 46°22.75' | 60°40.15' | | R.J. Mott, pers. comm., 1992 |
| GSC-5283<br>TO-3974<br>TO-3975 | RJM | ≈100 | 45°39.1' | 61°21.7' | | R.J. Mott, pers. comm., 1994 |
| GSC-4865 | RJM | ≈60 | 46°06.75' | 60°4 6.33' | | McNeely and Jorgensen, 1993 |
| | | ≈45 | 45°52.1' | 63°23.3' | | MacDonald and Boner, 1993 |

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