Maritime Sediments, Vol. 8, No. 3, December 1972, pp. 88-100.

Reports

Quaternary History and Sedimentation: A Summary and Select Bibliography*

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Introduction

The Pleistocene record in the Maritime Provinces is largely confined to the last glaciation (Classical Wisconsin); the older glacial and non-glacial intervals are represented by fragmentary evidence. Some features, however, do appear: numerous estuaries for example represent the drowned parts of former river systems that may have been operative during times of lower sea-level in Quaternary time, however low sea level almost certainly corresponds with recovery in the Maritime Provinces. Interglacial and interstadial deposits together with associated glacial deposits that pre-date the last glaciation are present at eight localities on Cape Breton Island (Mott and Prest, 1967). Several radio carbon datings on organic remains yielded ages of >40,000 years in the southwestern part of the Island.

Pre-Pleistocene morphology has been treated by several authors. R.H. MacNeill (noted in Prest, 1970) believes that many of the streams in Nova Scotia are re-excavated pre-glacial channels whilst till-mantled coastal terraces were probably cut during pre-glacial times. In Northumberland Strait, Kranck (1972a) identified a drainage network which post-dates the suggested Tertiary peneplain in Eastern Cape Breton (Goldthwait, 1924) and clearly pre-dates the last glaciation. King and MacLean (1970) suggest a Late Cretaceous or Early Tertiary age for an erosional sub-bottom profile cut in Carboniferous and Cretaceous bedrock underlying the Laurentian Channel but believe the channel to have an erosional history dating back possibly as far as the Jurassic period. Swift and Lyall (1968) consider the Bay of Fundy to have a fluvial origin of equal antiquity with glacial action only modifying the form.

Present marine morphology and sedimentation represent continuing adjustment of Pleistocene features to modern processes as changes in the hydrodynamic regimes have taken place through eustatic and isostatic changes since the last glaciation.

Classical Wisconsin

The last continental ice sheet in Canada had three major components, one of which - the Laurentide Ice Sheet - dominated the Maritime Region. It is clear, however, that the Maritime region came under the influence of large independent ice caps (the Appalachian Glacier Complex) both in early Wisconsin times and in the deglaciation process (Prest, 1970; Prest and Grant, 1969), although at the glacial maximum the Maritime ice and Laurentide ice were confluent, if not across the Gulf of St. Lawrence then at least in northern New Brunswick and Newfoundland. It is probable that the Appalachian Glacier Complex consisted of two main parts, one over Newfoundland and the other over the Provinces of Prince Edward Island, New Brunswick and Nova Scotia. The discrete nature of the ice build up in the Appalachian Complex may well have effectively barred Laurentide ice from some parts of the region, because little definitive evidence of Laurentide glaciation is known.

On the basis of crystalline erratics (not Shield rocks however) the terminal positions of one or more ice sheets off the Nova Scotian coast have been thought to lie close to the edge of the continental shelf and to pass through or near Sable Island, about 160 km from the shore. King (1969) reported that the entire Scotian Shelf has geen glaciated and an end-moraine system runs from northeast of Halifax to southeast of Yarmouth at approximately 30 to 40 km offshore and in water depths ranging from 70 to 200 m. The maximum age of the moraine is inferred from a date of 18,800 years BP, on amorphous organic matter in postglacial marine sediments overlying the moraine. From this King concluded that the end-moraine system represents a major late-Wisconsin glacier limit - either a recessional stand during the retreat of the ice from the edge of the continental shelf or the terminus of the last ice sheet.

Regional Ice Flow Patterns:

A generally held concept of regional radial flow of Laurentide ice (Labradorean sector) across the Maritime Provinces and on to the Scotian shelf has been based on several factors:

(i) regional southeast trend of eskers and other ice-flow features in southwestern New Brunswick and southern Nova Scotia coincident with those in the adjoining state of Maine.

* Manuscript received March 1, 1973.

(ii) presence in Nova Scotia of glacial erratics considered to be New Brunswick 'indicators' (Goldthwait, 1924).

Prest and Grant (1969), however, cite several factors at variance with this concept (see Fig. 1):

(i) absence of Precambrian erratics derived from Quebec in Nova Scotia.

(ii) absence over most of the area (northern Nova Scotia, Cape Breton Island, Prince Edward Island and eastern New Brunswick) of southeast ice-flow trends that might be expected from a major flow of Laurentide ice.

(iii) the east southeast direction of the main ice-flow indicated over Prince Edward Island.

(iv) little or no evidence of southeasterly trending ice-flow lineations on Cape Breton Island.

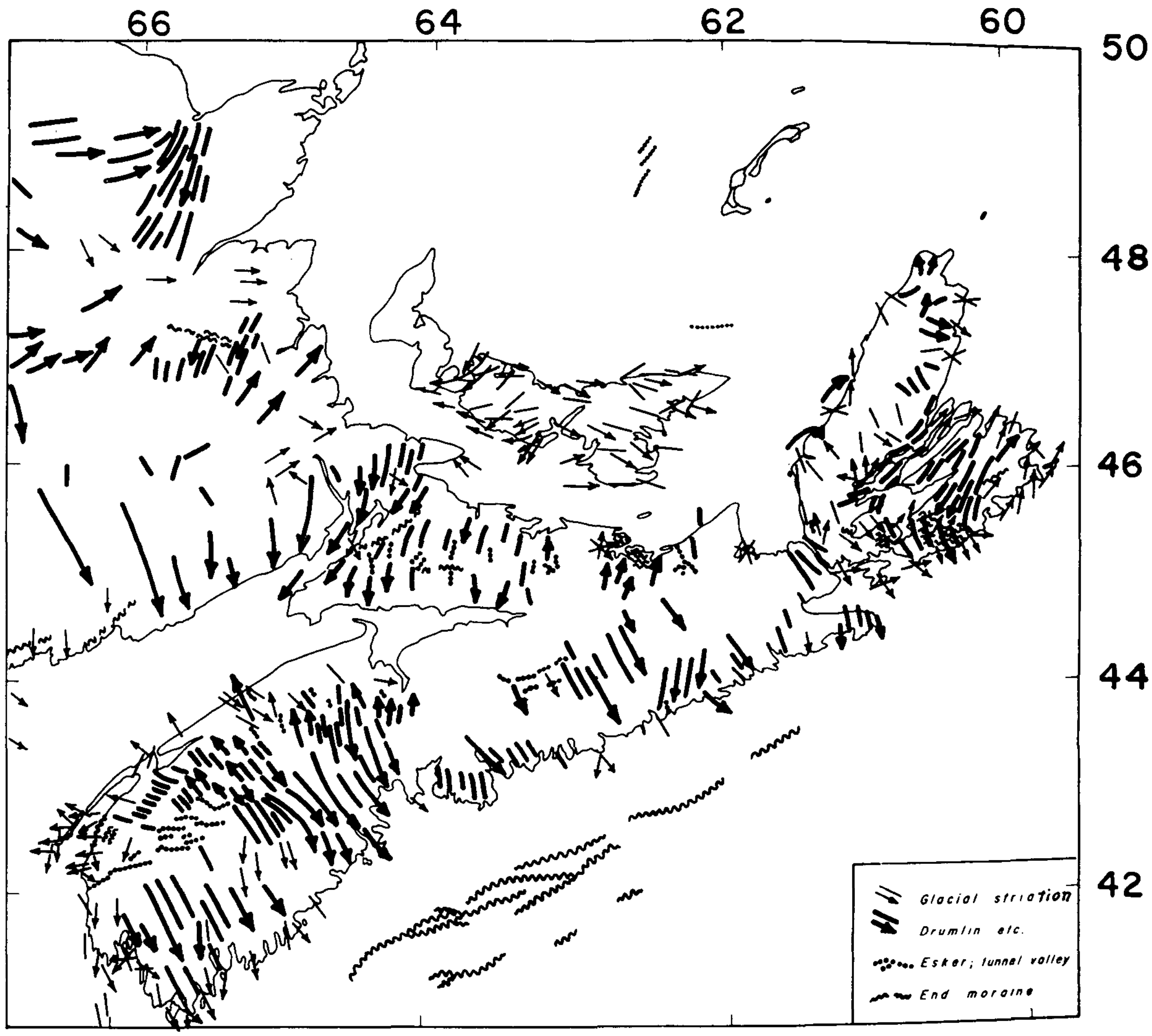

Figure 1 - Glacial features indicative of ice flow patterns in New Brunswick, Prince Edward Island and Nova Scotia (from Prest and Grant, 1969).

Recent mapping by Grant (1971a, b, c; 1972) has further emphasized the more complex nature of ice-flow in parts of Nova Scotia. The general consensus is that during the Wisconsin maximum the ice shed lay over New Brunswick and the Gulf of St. Lawrence. For the period during deglaciation Prest and Grant (1969) suggest localized, more or less radial outflow from certain lowland as well as upland areas, and that many ice flow features and end moraines relate to a rising sea-level over the period from about 18,000 to 11,000 BP. The Maritime climatic regime is thought to have allowed various parts of the Appalachian ice complex to remain active as the sea encroached on the depressed land masses. The deeper-water parts of the submerged coast served as 'leads' into the ice fronts, with consequent development of calving bays and ice draw down. An hypothetical pattern of deglaciation in response to topography, climatic change and sea-level rise has been proposed by Prest and Grant (1969; also Prest, 1969, Map 1257A) and is shown in Figure 2.

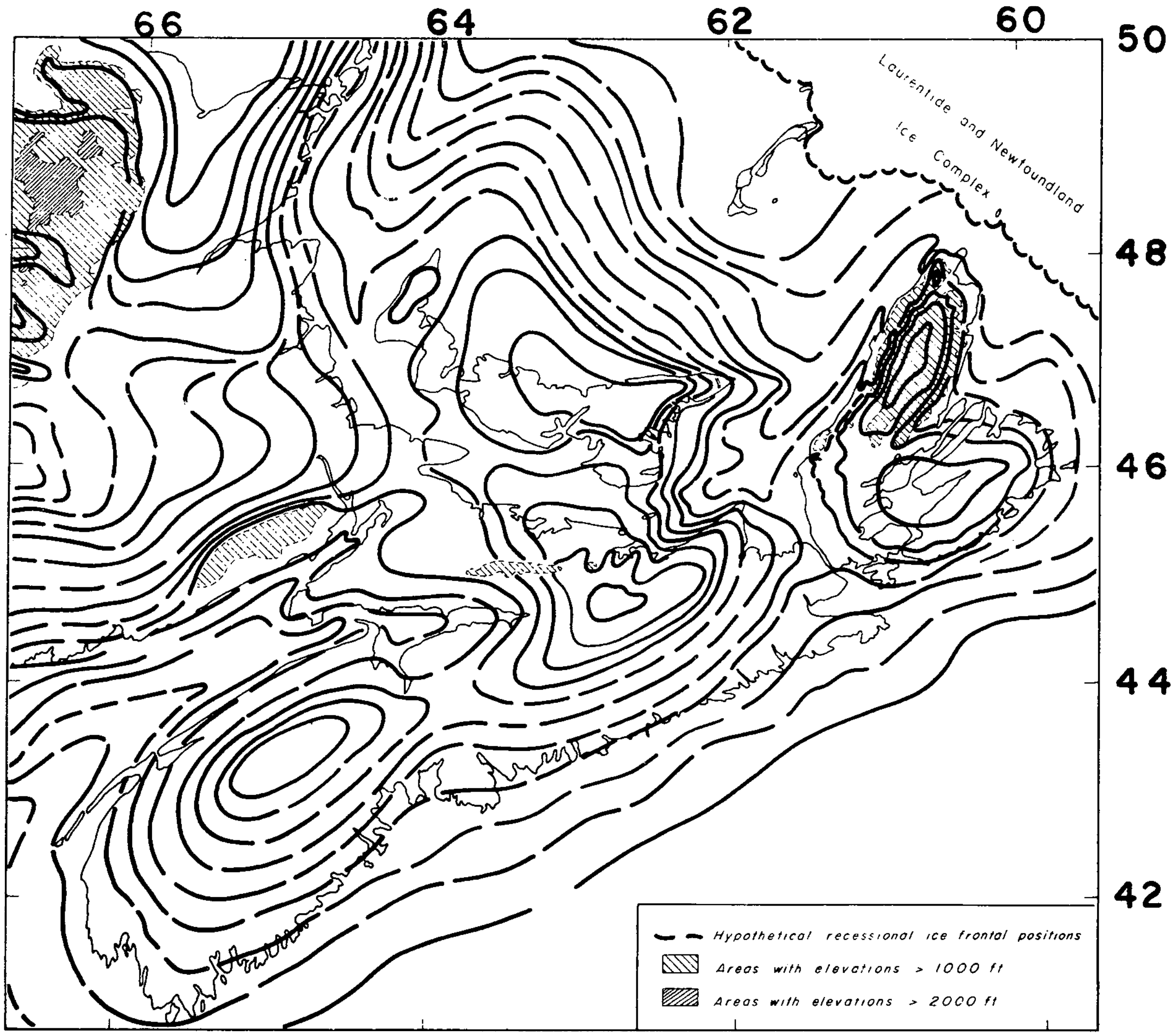

Figure 2 - Hypothetical ice-marginal positions during the recession of Wisconsin ice (from Prest and Grant, 1969).

Post-Wisconsin Marine Events:

The Maritime coastal region has experienced both eustatic and isostatic adjustments subsequent to deglaciation, but a detailed chronology of sea level changes for the whole area is still incomplete and considerable differences of opinion exist on the altitude of the marine limit in various areas.

Bay of Fundy:

The Fundy embayment is the only area in Nova Scotia affording evidence of emergence whereas the whole province is presently involved in submergence (Grant, 1970). A marine limit at the mouth of the bay has been dated at 14,000 ± 200 years BP in southwest Nova Scotia by Grant (1971c), whilst Swift and Borns (1967) have suggested a date of around 13,000 BP for the deglaciation of the Bay.

Late Wisconsin sea levels are (by actual measurement) higher than present on the south and north coasts at the mouth. Chalmers (1890) noted marine terraces along the Fundy shore both southwest and northeast of Saint John. In the former area the deposits had a maximum altitude of 68 m and contained shells; in the latter area the highest terrace deposits were only 38 m and no shells were seen. Lee (1959) found estuarine deposits at 38 m in the St. John River valley at Fredericton. It is possible that the lower part of the Saint John River valley and the New Brunswick side of the Bay were covered by late ice while the high-level terraces were being formed southwest of Saint John.

Subsequently sea-level fell, crossing present sea-level about 11,000 to 12,000 years BP and, according to Grant (1971c), continued falling to about 20 to 30 m below present level, a value estimated on the basis of the thickness of intertidal sediment presently filling estuaries. The sea approached its present position only within the last 3,000 years (Harrison and Lyon, 1963), although the record of rising sea level is at least 5,000 to 6,000 years long.

Determination of post-glacial coastal emergence and submergence in the Bay of Fundy is complicated by the fact that the tidal regime has changed considerably since deglaciation (Grant, 1970). The extreme modern tidal range, (varying from 3 m at the mouth to >17 m in the Minas Basin) has been suggested as a consequence of the dimensions of the Bay which satisfy the conditions for resonant amplification of the tidal wave's semi-diurnal component (King, 1962; Swift and Borns, 1967; Pelletier and McMullen, 1972). Calculations made by Swift and Borns suggest that actual and resonance dimensions have been 95 per cent coincident only since 7,500 BP using Curray's (1965) curve for Holocene sea-level rise. Harleman (1966) and Rao (1968), however, seriously question the concept of resonant amplification and conclude that tidal amplification primarily results from a high tidal range at the Bay entrance, being confined by the narrowing and shallowing cross-section. Whatever the cause of the modern high tide it is still likely that variation in range would have occurred because the depth of the bay has changed markedly since deglaciation.

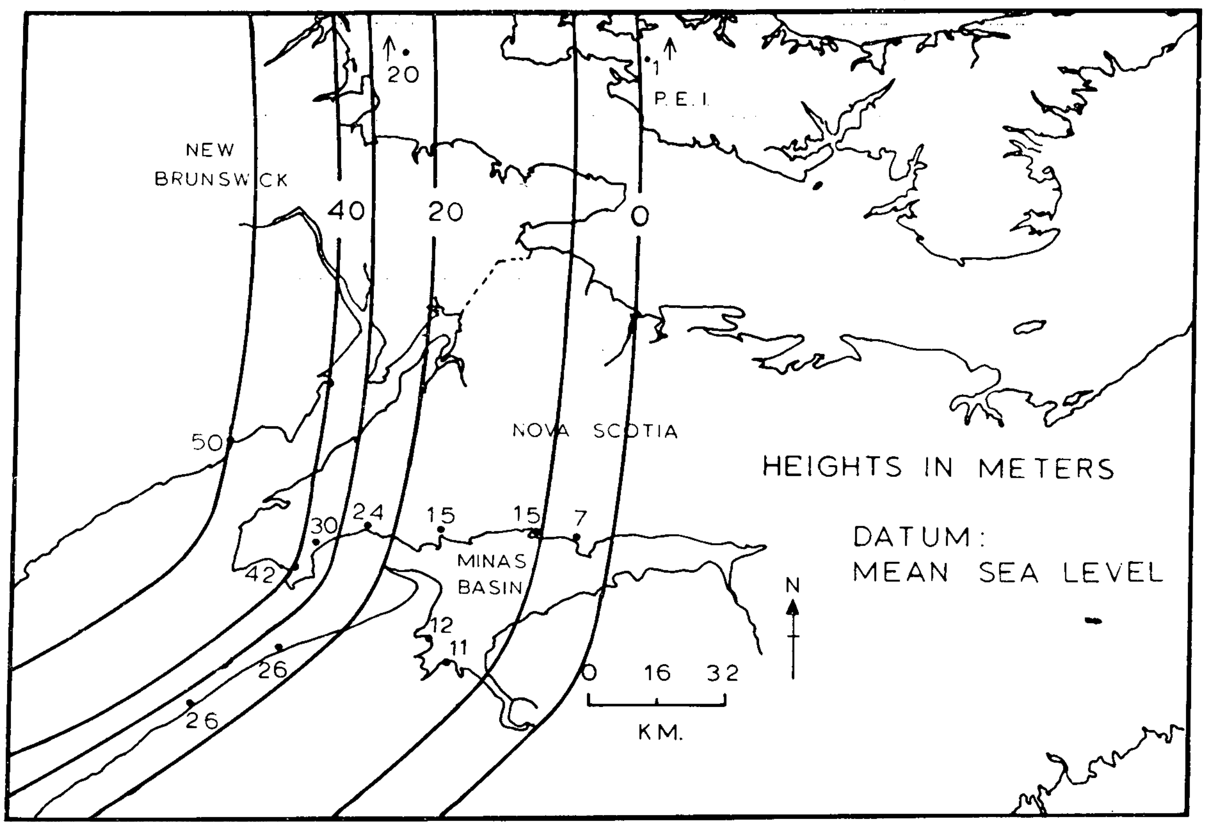

Figure 3 - Isobases of post-glacial emergence (from Swift and Borns, 1967).

Swift and Borns (1967) present a map of minimum postglacial emergence (Fig. 3), based on elevation above mean sea level of the junction between a marine and a fluvial lithosome on the north shore of the Minas Basin, together with data from Hickox (1962) for the Annapolis Valley and Prest (1962, 1964) for Prince Edward Island. In southwest Nova Scotia, isolines on the height of emerged features trend northeasterly, roughly parallel to the Fundy shore, with the zero isoline just north of Yarmouth: uplift is 40 m at Digby and 50 m on Long Island, southwest of Digby (Grant, 1971c).

The north shore of the Minas Basin was clearly affected by a complex series of sea-level movements: the retreating ice front and the eustatically rising sea-level appear to have reached the north shore at about the same time, as indicated by the deposition of marine outwash. Rebound then transpired, resulting in initial dissection of the terrace, and its subsequent burial under fluvial outwash (Swift and Borns, 1967). The isoline trends undergo a marked change in this area and Prest (1970) has suggested this could have resulted from changes in the amount and rate of uplift owing to:

(i) early opening of Bay of Fundy as a calving bay in the ice front;

(ii) to presence of late ice north of the Minas Basin;

(iii) structural influence of the Cobequid Fault; and

(iv) different rates of retreat.

There are clearly serious problems associated with construction of isolines of the marine limit at this time (Swift and Borns, 1967, p. 708, Prest, 1970, Welsted, 1971, 1972). However, it does appear that values for emergence increase in a northwesterly direction (Welsted, 1971).

Southern Gulf of St. Lawrence:

About 14,000 years BP, as Wisconsin ice waned and the sea encroached into the Gulf of St. Lawrence, floating ice fringed the east and south shores of the Magdalen Islands, and marine sand and gravel together with thin layers of till were deposited along the coasts up to an elevation of about 36 m. The sea invaded the northwest part of Prince Edward Island prior to 12,410 years BP and had encroached some 400 km up the St. Lawrence River by 12,720 ± 170 years BP (Prest and Grant, 1969; in Lee, 1963). In all cases there is evidence of glacier ice close by. Ice is thought to have remained over Prince Edward Island and on the floor of the Gulf of St. Lawrence to the north as the sea penetrated up deep 'leads' into Northumberland Strait, although it was not until about 5,000 BP that rising sea levels completed the inundation (Kranck, 1972a). However, sea level has continued rising up to the present.

McRoberts (1968), studying foraminiferal concentrations in marine sediments in the eastern section of Northumberland Strait, suggested that immediately after deglaciation sea level was lower by approximately 71 m. At this time the Strait would be completely exposed to subaerial influences with two estuaries, separated by an isthmus at Abbegweith Strait, and draining the channel to the west and east (Kranck, 1972a).

Marine overlap in western Prince Edward Island reaches a maximum of 23 or 24 m along the west coast. The marine limit in the northwestern part of Malpeque Bay is about 9 m and in the southeastern corner, about 3 m. The zero isobase appears to be near Borden and probably trends south across Northumberland Strait to Cape Tormentine, New Brunswick (Kranck, 1972a). There is no evidence of marine overlap in central or eastern Prince Edward Island.

The transgression in Northumberland Strait is shown by Kranck (1972a) to have produced five marine terraces (60 to 79 m, 45 to 55 m, 35 to 40 m, 24 to 30 m and 18 to 23 m depths) with spatial characteristics reflecting eustatic and isostatic controls (Fig. 4). Greater isostatic rise of land in the west caused regression of the sea between 13,000 and 7,000 BP, whilst in the east eustatic rise in sea level was always greater than rebound. The oldest marine terrace at 60 to 79 m in the eastern end of the Strait is correlated with raised beaches dated at approximately 12,500 years BP on Prince Edward Island by Prest (1962).

Along the east coast of New Brunswick large tracts of low-lying ground formerly covered by the sea show little or no evidence of marine action; there were apparently no significant halts as the land rose and, in general, marine sediments appear to have been removed by erosion or incorporated into the soil profile. Local areas of marine sand, gravel, and poorly washed sediments, and the deposition of ice-rafted boulders do provide some evidence of former marine action (Prest, 1970). The marine limit varies from zero at the eastern end of Cape Tormentine to 30 m at Moncton. It is 48 m about 15 miles west of Richibucto and 68 m at Newcastle and Bathurst. Farther west along Chaleur Bay the marine limit is well marked by prominent terraces, and declines to about 42 m (see Prest, 1969, Map 1253A) as a result of the withdrawal up the bay during the period of maximum overlap along the east coast. The isobases of differential uplift trend northeasterly

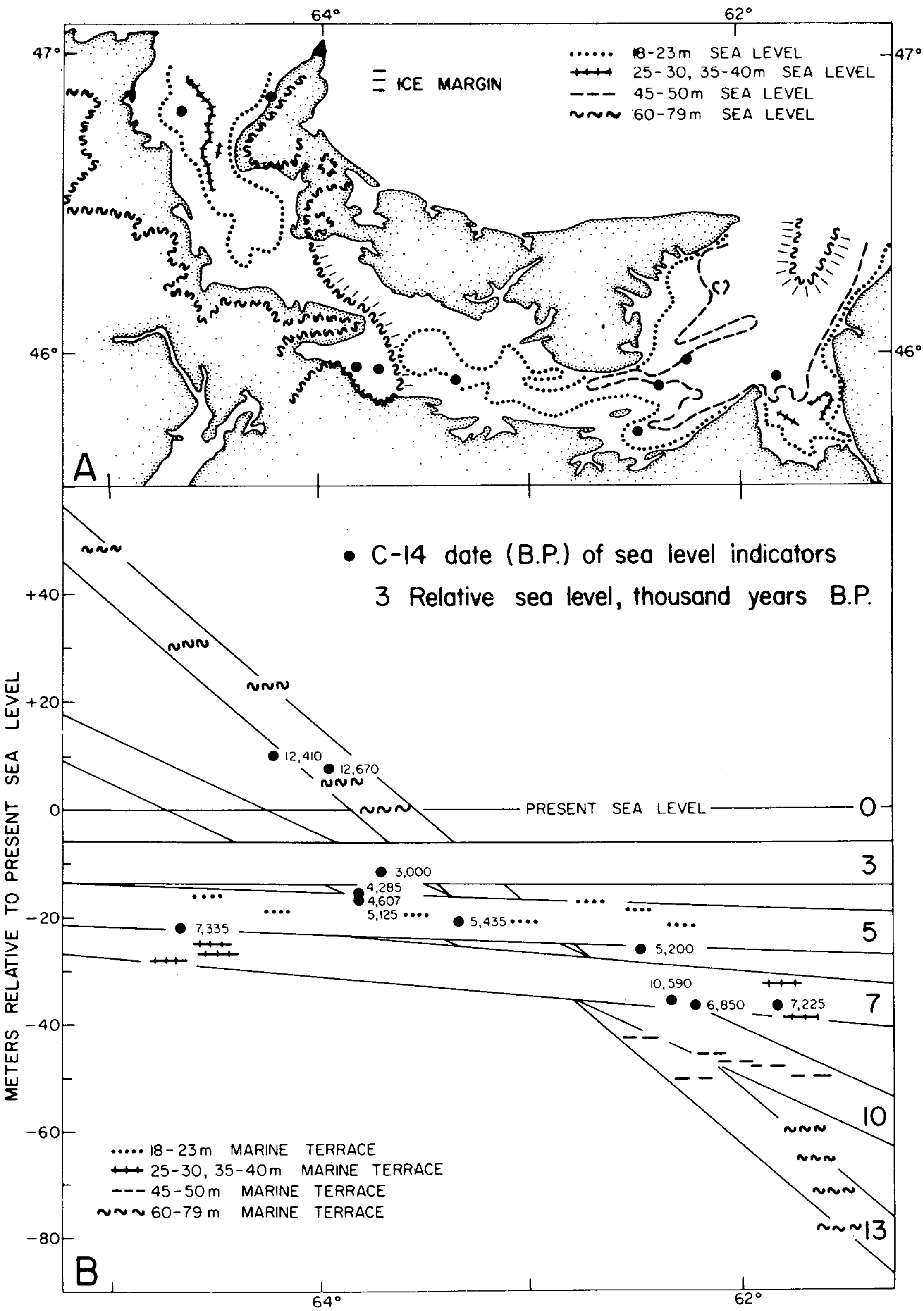

Figure 4 - Ancient sea levels of Northumberland Strait (from Kranck, 1972a). A. Carbon - 14 dates and geographical location of shorelines based on marine terraces; the marine limit (prest et al. 1968) is included synchrous with the 60-79 m level. B. Information from A projected onto a vertical E-W plane; diagonal lines represent the approximate position and age of the marine limits.

along the Bay of Fundy shore, northerly through the Moncton area, and northeasterly again along the eastern shore of the province.

There is no evidence of marine overlap along most of the Northumberland shore of Nova Scotia. A rock cut bench mantled with gravel on the northeasterly trending coast near Arisaig is a questionable record of a former sea level stand some 1.5 to 3 m above the present, and comparable sea level stands have been recorded elsewhere along this and the George Bay coast; all these are believed to be pre-Wisconsin (Prest, 1970). In the extreme northwestern part of the province, however, the zero isobase of postglacial uplift is believed to pass through Northport, with maximum marine overlap increasing westward to perhaps 15 m at the provincial boundary. The evidence is, however, rather inconclusive (Prest, 1970). No unequivocable evidence of marine overlap has been found on Cape Breton Island and no evidence of emergence has been recognized along the southeastern coast of Nova Scotia.

Recent Submergence

Grant (1970) citing hydrographic, archeologic and geologic evidence indicates that submergence of the Maritime Provinces over the last 4000 years has been going on at a rate of about 30 cm/100 years, i.e., three to fives times faster than the 6 cm/100 years rate of eustatic rise of sea level (Shepard and Curray, 1967). The Bay of Fundy shows an anomalous submergence of 24 cm/100 years, after correcting for eustatic change, of which 15 cm/100 years, is probably due to rise of high tide, or increase of tidal range, beginning 4000 to 6000 years ago. Submergence of the Atlantic coast of Nova Scotia exceeds the eustatic rate by 9 cm/100 years. Grant interprets this reduction as a result of isostatic subsidence of the crust as the sea deepened eustatically over the continental shelf (Bloom, 1965; Higgins, 1965).

However, neither tidal amplifications nor water-load subsidence account fully for the anomalous rate at which the Maritime Provinces have been submerging. Grant suggests that (atmospheric and oceanographic elevation of sea level reduction of barometric pressure, increase in water temperature, wind "set-up" effects, although deemed negligible, together with) subsidence as a result of epeirogenic downwarping and collapse of bulges peripheral to the last icesheet, may be the cause of the large relative rise of sea level in the area.

Evidence for continuing contemporary submergence is provided by secular changes in the height of sea level as revealed in tidal records. Records for Halifax, Nova Scotia, Saint John, New Brunswick and Charlottetown, Prince Edward Island show that mean tide level has risen steadily over the past few decades at 41, 46 and 26 cm/100 years respectively. Coastal erosion in the Gulf of St. Lawrence (Forward, 1960), the continued upward growth of salt marshes and the transgressive barrier beaches in the region all point towards continuing submergence.

Modern Marine Morphology and Sedimentation

Southern Gulf of St. Lawrence:

This area comprises a broad almost circular, shallow shelf (the Magdalen Shelf, Prince Edward Island and Northumberland Strait) bordered to the north by the deep submarine trough - the Laurentian Channel - off Cabot Strait, which reaches a depth of more than 200 fathoms (265 m) (Fig. 5). The shelf has an irregular relief composed of elevated banks and trough-shaped valleys. The main positive features are the circular platform crowned by the Magdalen Islands, Bradelle Bank to the west of the islands, Orphan Bank in the northwest corner of the shelf, Prince Edward Island and smaller banks along the northern edge of the shelf. Along its western and eastern margins, the shelf is incised by a number of well-defined trough-shaped valleys 20 to 90 fathoms (37 to 165 m) in depth (Chaleur Trough, Shediac Valley, Cape Breton Trough) and by shallower and broader ones in the central part (Western and Eastern Bradelle Valleys) as shown in Chart 811-A. The whole area, underlain by Paleozoic red sandstone, was modified by glaciation and most of the features of the troughs, which were part of a pre-Pleistocene drainage system, are a result of this erosive modification.

Loring et al (1970) show that a relict Pleistocene sedimentary environment in various stages of preservation exists on the Magdalen shelf. Deposits composed of large areas of uncovered bedrock and unsorted sand and gravel exist north of Prince Edward Island and in the nearshore areas of the Magdalen Islands. In other areas, in contrast, recent depositional processes have produced major changes in the microtopography and texture of relict sediments on offshore banks to produce lag gravel deposits and rippled sand fields. Accumulation of fine grained sediments in lows and troughs has buried the relict sediment cover and modified the morphology in these areas (Fig. 5).

Nota and Loring (1964), studying the bathymetry and sedimentation of the Gulf, concluded that Pleistocene glaciation largely determined the present shape of the submarine trough although the valley is claimed to be tectonically controlled, being the faulted boundary between the Archean crystalline rocks of the Canadian Shield and the Palaeozoic sedimentary rocks of the northern

Appalachians (contrast King and MacLean, 1970). Sediment in the trough is fairly regular and characterized by poorly sorted, coarse grained deposits near shore and an extensive area of pelitic material in the deeper parts. The nearshore areas bordering the submarine trough exhibit reworking and redistribution of glacial drift, while fresh coarse debris is being added by ice-rafting Postglacial deposition in the deeper parts is thought to have occurred at a rate of 22 cm/1000 years.

Kranck (1967, 1971), in a study of the bedrock and sediments of Northumberland Strait, shows areas of topographic highs to be composed of bedrock and gravel whereas the topographic depressions consist chiefly of sands and muds. In many areas the bedrock is coated with till and the broad, nearly horizontal, marine benches mentioned earlier have been cut across rock and glacial sediments alike.

The gravel sediments here are again thought to be relict glacial material mroe or less reworked by wave and current action but the sand deposits interpreted on the basis of sediment parameters, relation to topography and association with present-day river mouths appear to be former river deposits. Modern sand and mud are being transported by asymmetrical tidal currents and associated residual drift (Kranck, 1972b); net sediment transport is towards the western and eastern ends under the higher velocity ebb flows.

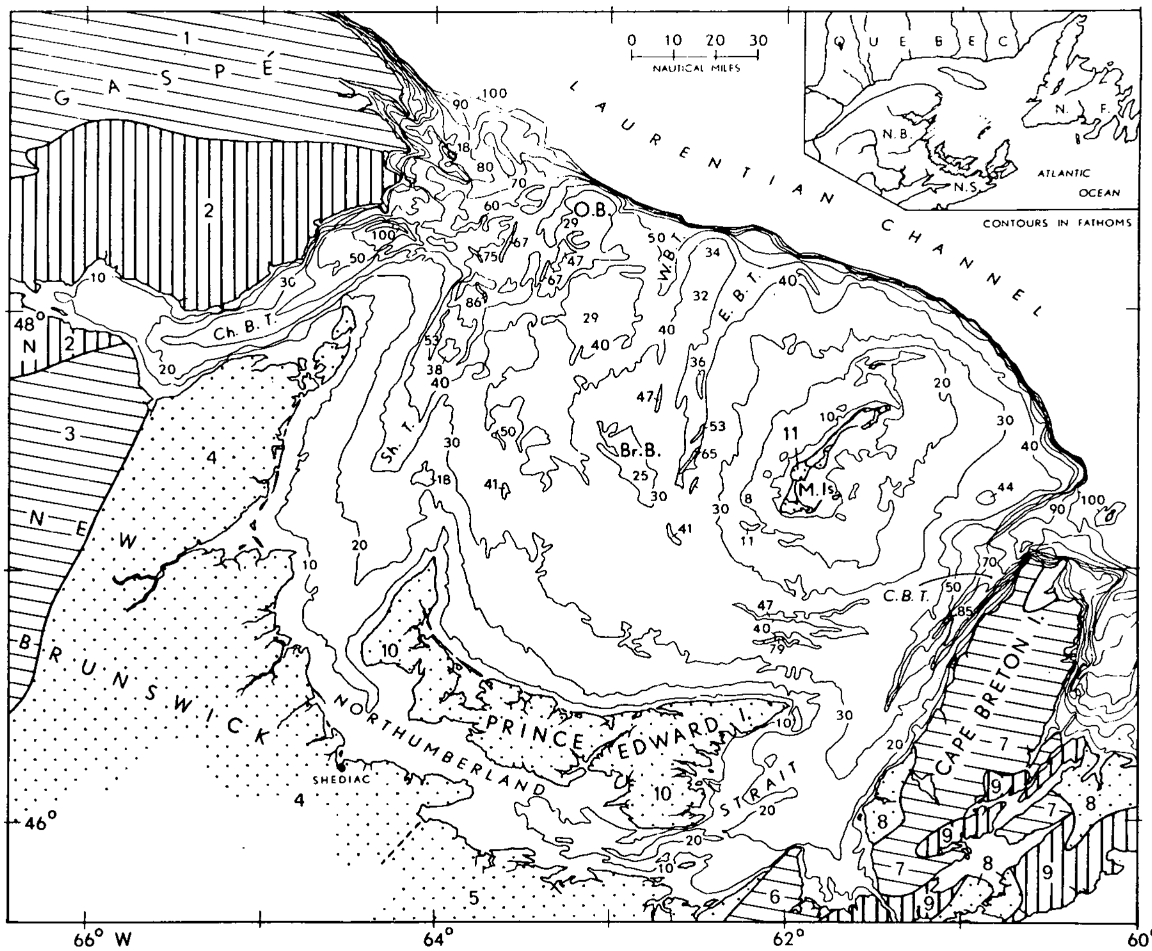

Figure 5 - Generalized geomorphological map of the S. Gulf region (from Loring and Nota, 1966). Ch. B.T. - Chaleur Bay Trough; Sh. T. - Shediac Trough; W.B.T. - Western Bradelle Trough; E.B.T. - Eastern Bradelle Trough; C.P.T. - Cape Breton Trough; M. Is. - Magdalen Islands; Br. B. - Bradelle Bank; O.B. - Orphan Bank; 1 - Notre Dame Mts; 2 - Chaleur Uplands; 3 - New Brunswick Highlands; 4 - New Brunswick Lowlands; 5 - Cumberland Lowlands; 6 - Antigonish Highlands; 7 - Cape Breton Highlands; 8 - Cape Breton Lowlands; 9 - Atlantic Uplands of Nova Scotia; 10 - Prince Edward Island Lowlands; 11- Gulf of St. Lawrence Plain.

Bay of Fundy:

The Bay of Fundy is a funnel-shaped body of water 144 km long, 100 km wide at the base and averages 75 m deep. Considerable debate has ensued on whether the Bay is derived from glacial or fluvial erosion of the underlying Acadian Triassic basin, which consists of friable red clastics and tholeitic basalts set in a half graben (Swift and Lyall, 1968). Fluvial erosion along the so-called Fundian Border Fault was suggested by Johnson (1925) and Koons (1941, 1942) but this was seriously questioned by Shepard (1930, 1942) who concluded that glacial erosion during the Pleistocene created the Bay and its scarps. Swift and Lyall (1968) made a detailed study of the bedrock bathymetry and revealed that the south side of the Bay is a dip-slope shore and the west shore, and portions of the south shore of the Chignecto Peninsula, are fault-line, scarp shorelines. Klein in 1961 rejected the idea of a Fundian scarp but stressed only transverse glaciation of the Bay: examining unpublished work by Take, he suggested a fluvial origin dating back to early Cretaceous times and that possibly the Bay is Jurassic in origin.

Fundy's rock floor is exceedingly flat, with gradients of 2 m/km or less. Swift and Lyall (1968) identify till materials overlain by outwash forming surficial layers commonly up to 10 m in thickness. Sub-bottom profiles reveal 0 to 3 m of outwash and till, and up to 100 m of finer sediment (Swift et al, 1971 & 1973). Bedrock troughs oriented longitudinally in Chignecto Bay and Minas Passage and the large trough at the mouth of the Bay of Fundy which are still blocked with glacial drift clearly owe their present configuration to ice action suggesting that at least at some time the Bay underwent longitudinal glaciation. However these troughs are initially consequent upon structure and therefore are probably only lately modified by ice. Also, flow may not necessarily be parallel to the length of the depression as cases of transverse troughs on continental shelves exhibit evidence that ice moves across them and erodes with lateral components.

The bulk of Fundy's sediment cover was emplaced under subaerial or non-marine conditions during the Pleistocene low stands of the sea, and subsequent encroachment by the modern high-energy tidal regime has generated additional sediment by means of coastal erosion. Much of the shoreline consists of cliffs backing a wide intertidal shore platform, the former receding at a rate of 2 m/yr in places (Churchill, 1924). Thin veneers of coarse, local detritus occupy the platform in exposed areas, and extensive tidal mud flats backed by marshes and large intertidal sand bars occur in protected coves and estuaries (Middleton this issue).

Distinct textural provinces have been identified on the floor of the Bay by Pelletier and McMullen (1972: a) gravel deposits lying generally below the 40 m depth contour beyond the influence of wave-action, especially in the central and southeastern areas; b) sands at shallower depths again in the central and eastern areas; c) muds along the northwestern side. The sands in planoconvex lenses transgress the muds in many areas (Forgeron, 1962), and are clearly tide-maintained. This together with alternating coarse/fine bands of sediment oriented transverse to the strike of the Bay is thought to reflect successive reworking of Pleistocene sediments by the standing wave tidal regime, which may have migrated in periodic intervals during the post-Wisconsin marine transgression (Pelletier and McMullen, 1972). However, only in the mud provinces are large thicknesses of Holocene sediment found. A counterclockwise residual circulation in the Bay together with stronger tidal currents on the south than north shore are clearly influential in the textural zonation at the bed.

The suspended sediment dispersal system has the following components, Swift et al (1971, 1973): 1) an oscillating turbid water mass, with an average turbidity of 6.6 mg/l; 2) bay floor and margin provinces undergoing winnowing, which serve as fine sediment sources; 3) bay floor and margin provinces undergoing fine sediment accretion, which serve as fine sediment reservoirs; 4) minor fresh, turbid water input; and 5) minor salt turbid water output into the Gulf of Maine. Three different rates of sediment transfer may be detected. Exchange between the mud provinces and the overlying water masses occurs during each semi-diurnal tidal cycle. Some sediment undergoes long-term storage of fine sediment in reservoir provinces, with residence times of the order of several millenia. Ultimately fine sediment either escapes at the mouth of the bay (on the order of magnitude of 1.6×10^6 metric tons per year) or undergoes permanent burial on the bay floor. It appears possible to resolve the system into two sub-systems, the main system of bay floor provinces and associated turbid water masses, and a marginal system of intertidal mud flats and associated water mass. The latter is volumetrically less important, but its rates of sediment transfer are much higher.

Atlantic Seaboard:

King (1967a, b) established a facies map for the Scotian Shelf based on echo-sounding, textural analysis of grab samples and topographic characteristics. He suggests that glaciation left the whole area covered by glacial drift which is still preserved in places and probably underlies most of the silt and clay deposits in the deep basins. During a low stand of sea level, indicated by a marine terrace at 110 to 115 m the drift sediments were reworked by wave and current processes with the clays and silts being transported offshore and deposited in the basins. Some of these now mask the end moraine complex previously described. The marine terraces and the shallower areas are covered by deposits of sands and gravels. Drapeau (1970) has reported the development of sand wave fields in areas where there is abundant sand. The coastline is characterized by deep indentations, which are produced by post-Wisconsin submergence of a glacially scoured landscape although these indentations are not essentially produced by glacial scour.

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