43. **GLACIAL GEOLOGY OF NORTHERN BOOTHIA PENINSULA, DISTRICT OF FRANKLIN**

Project 750071

Arthur S. Dyke
Terrain Sciences Division

Dyke, Arthur S., Glacial geology of northern Boothia Peninsula, District of Franklin; in *Current Research, Part B, Geological Survey of Canada, Paper 79-1B, p. 385-394, 1979.*

Abstract

Mapping of glacial and marine features and radiocarbon dating of fossil marine molluscs have provided much new information on the glacial geology of northern Boothia Peninsula. At the late Wisconsin maximum ice flowed eastward to northeastward over the area. This flow is recorded by large fields of rock drumlins and other ice moulded bedrock forms. During recession, lateral meltwater channels, ice dammed lakes, and moraines were formed. These features document westward recession of the ice mass, whose margins and last flow pattern were topographically controlled. The marine limit near the east coast lies at 215 m a.s.l. and was formed 9230 ± 130 years B.P. The marine limit declines westward in the direction of ice recession to 155 m a.s.l. and less in Wrottesley Valley, where it was formed about 9040 ± 100 years ago. The ice margin retreated about 300 m/year, and the ice surface lowered about 5.5 m/year between 9230 and 9040 years B.P. The initial coastal emergence rate was more than 30 m/100 years. Shells dated at >23 300 years, from deltaic sediment at 195 m a.s.l. on the northeast coast, probably represent a pre-late Wisconsin marine incursion caused by ice recession. Organics, dated at >30 000 years, from sands below the Holocene marine limit in Wrottesley Valley, are probably detritus in Holocene marine sediments.

Introduction

The glacial geology of the northern half of Boothia Peninsula (Fig. 43.1, inset) was investigated during field work and airphoto interpretation in 1978. Craig (1964) dealt briefly with the area, and Boydell et al. (1975a, b) mapped the general distribution of surficial materials. This study is part of a more detailed study of the south-central Canadian Arctic Archipelago and adjacent Arctic mainland.

Northern Boothia Peninsula contains an abundance of ice flow and ice marginal features that were previously unmapped. In addition, the marine limit has been mapped and dated at two critical points. This allows establishment of an early Holocene deglacial chronology. Previously dated pre-late Wisconsin materials and the possible correlation of these with "old" materials on Somerset Island are discussed in light of these findings.

Acknowledgments

This study was made possible largely by the logistical support provided by the Polar Continental Shelf Project of Energy, Mines and Resources through their field operations base at Resolute Bay, managed by Mr. Fred Alt. I am thankful to Mr. Robert Hélie and Mr. Steven Black for assistance in the field, and to Dr. R.J. Fulton for his critical reading which helped to clarify several points.

Late Wisconsin Glaciation and Deglaciation

Ice Flow Features

Features, other than striae, shown in Figure 43.1 are easily recognized on 1:60 000 scale airphotos, an indication of their size. Forty striae measured throughout the area are parallel to these features, which confirms that the features were formed by glacial erosion. The most widespread ice flow features are ice moulded bedrock forms which are concentrated in the north and southeast part of the study area. The features include large roches moutonnées, rock drumlins, and crag-and-tail forms, with the latter two being more common.

The bedrock is most strongly ice moulded in the extreme northern part of the peninsula. Ice flowed eastward, normal to the compositional banding in the granitic gneiss (Fig. 43.2) and normal to one of the major bedrock fracture systems. Hence, the moulded features do not appear to be structurally controlled. Also in this area are deep, east-west oriented, U-shaped troughs, occupied by long lakes or arms of the sea. Their forms are partly due to glacial erosion, although they may have originated as grabens in Tertiary time (Kerr and de Vries, 1977). However, the southernmost of these, occupied by Amituryouak Lake, is crossed obliquely by a strongly developed set of ice moulded bedrock features. This crosscutting indicates that formation of the trough preceded the last major ice advance.

Strongly ice moulded bedrock is also seen southwest of Wrottesley Inlet and west of Abernethy Bay; in both areas the ice flowed northeastward. The rest of the study area is devoid of large ice flow features. In most places, however, the bedrock is moulded into smaller forms, mostly roches moutonnées, on many of which striae and grooves are preserved.

The preferential development of large ice flow features in the north and southeast parts of the area probably reflects the locations of vigorously flowing streams of ice within the ice sheet. These streams did not arise from topographic channelling of the ice, because most of the intervening area is no higher than much of the streamlined terrain. Furthermore, the ice defied topographic channelling in areas of considerable (200 to 400 m) local relief, as around Amituryouak Lake and Wrottesley Valley, a wide and deep graben oriented nearly perpendicular to ice flow (Fig. 43.1).

The uniform nature of the ice flow pattern and the lack of topographic control suggest that the features were imprinted on the terrain by thick ice during glacial maximal conditions. The direction of flow indicates that the dispersal centre of the ice mass lay west of Boothia Peninsula. It is undoubtedly the same centre as that which produced the eastward flow over southern and western Somerset Island immediately north of the present study area (Dyke, 1978a).

Ice Marginal Features

Both erosional and depositional ice marginal features are common, particularly in areas devoid of the large glacier bedforms discussed above (cf. Fig. 43.1, 43.3). Figure 43.3 shows the distribution of ice marginal (lateral) meltwater channels, moraines, and former ice dammed lakes. These are discussed below.

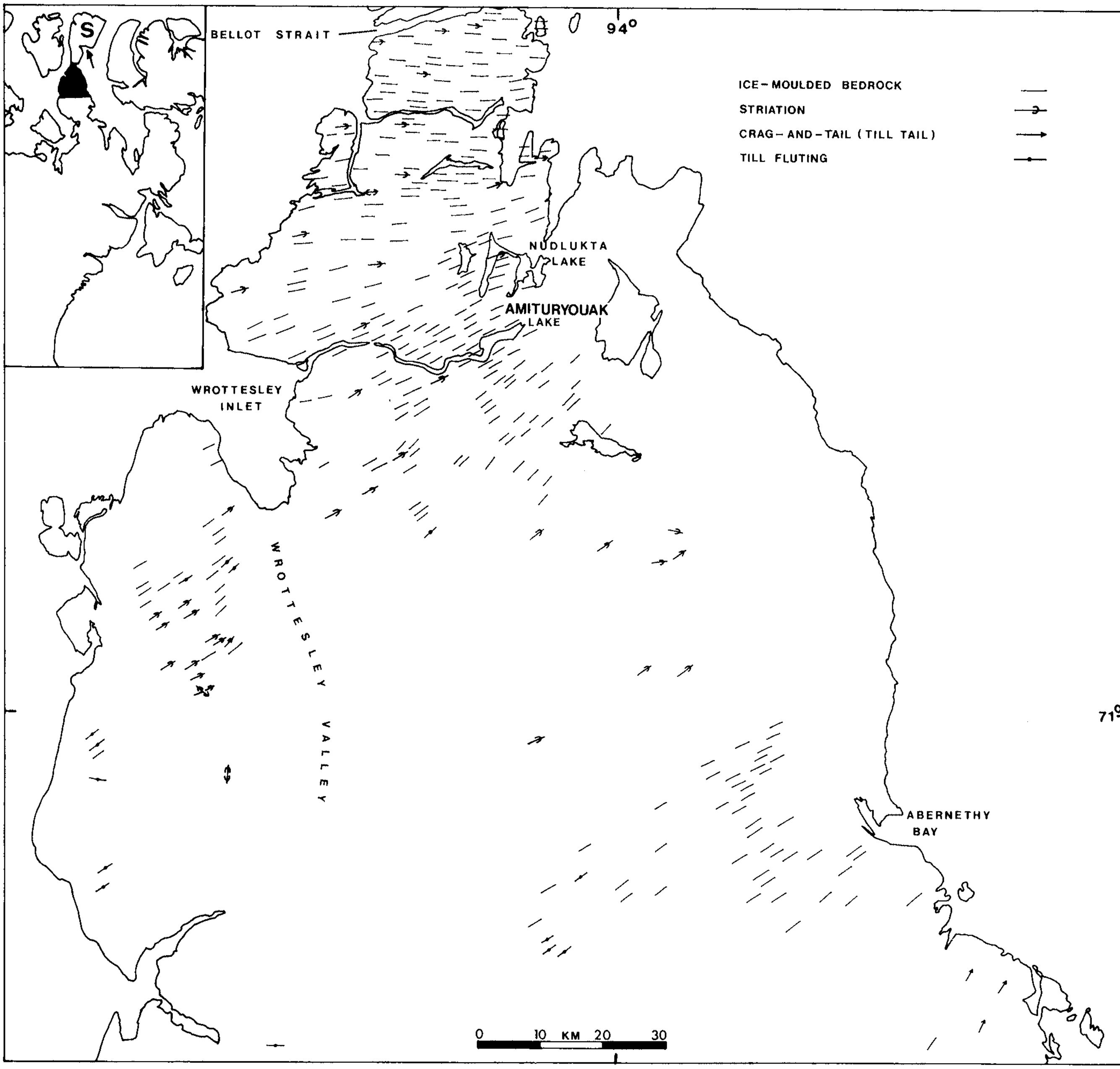

Figure 43.1. Ice flow lineaments on northern Boothia Peninsula. Somerset Island (S) is immediately north of the study area; arrow points to Creswell Bay.

The most common ice marginal features are lateral meltwater channels, but they are confined to areas above the marine limit. These channels run obliquely along hillsides in such a manner that the streams that cut them must have been confined on the downslope side by ice. They are generally incised 1 to 5 m into bedrock and form nested sets of features along many valleysides. Subglacial and proglacial meltwater channels and deposits are also common but are not shown in Figure 43.3 because they are of little use in deciphering ice marginal configurations.

Moraines are the next common ice marginal features, and three general types occur (Fig. 43.3): ridges of till that accumulated at ice fronts on dry land, generally in topographically unconfined areas; ridges of sand and gravel, commonly with distinct ice contact escarpments (kame moraines), that accumulated where ice fronts impinged on steep slopes above marine limit; and till and boulder ridges of various heights that accumulated at the ice front as it retreated across the former seabed in areas below marine limit on the east coast.

The submarine moraines form three morphologic groups, each occupying a separate area. The northern group occurs between Brentford Bay to just north of Abernethy Bay. They consist of closely spaced parallel lines, 1 to 2 m wide, of boulders which overlie till on a broad lowland plain (Fig. 43.4). These are typical of features commonly referred to as De Geer moraines, which are thought to be annual ice marginal accumulations (Hoppe, 1959; Prest, 1968). The central group, which occurs inland from the head of Abernethy Bay, consists of larger parallel ridges of till, about 5 m high (Fig. 43.5). They probably represent longer stands of the ice front than do the De Geer moraines. The southern

Figure 43.2A. *Oblique aerial view of rock drumlins immediately south of Bellot Strait. Ice flowed from right to left, across the compositional banding in the gneiss, clearly visible in the middle foreground. The light-coloured sediment in the foreground is a carbonate-rich till. (GSC 203359-X)*

Figure 43.2B. *Oblique aerial view of ice moulded bedrock forms near Nudlukta Lake. Ice flowed from left to right. Forms vary from symmetrical rock drumlins to crag-and-tail features. The crags are formed by a resistant vertical band in the gneiss which trends across the ice flow features and extends from the lake shore in the middle foreground to the right skyline. (GSC 203359-Y)*

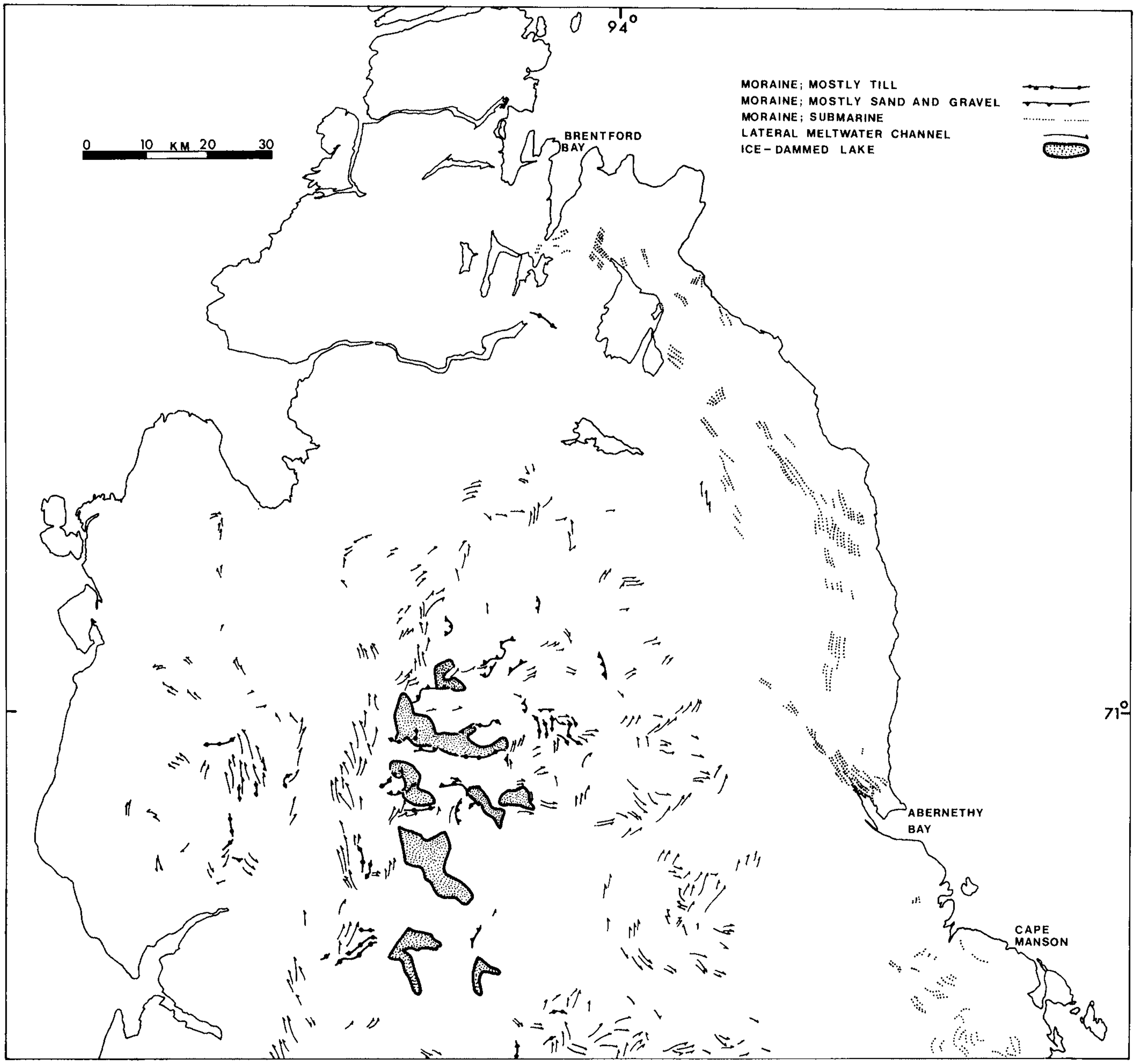

Figure 43.3. *Ice marginal features on northern Boothia Peninsula.*

group, which occurs inland from Cape Manson, is made up of 1 to 2 m high, subparallel, closely spaced, sinuous ridges, best developed in U-shaped valleys. They closely resemble in morphology the "cross-valley" moraines of Baffin Island (Andrews, 1963).

The third important set of ice marginal features are former glacier dammed lakes (Fig. 43.3). These left weakly developed strandlines and thin veneers of sediment, which indicates that they were short lived. They occupied the headwaters of eight westward draining valleys, indicating a westward recession of the ice margin.

Ice Recession

The ice marginal features (Fig. 43.3) were correlated throughout the region to produce a map of ice recession (Fig. 43.6). Correlations were based on orientations and relative positions of features, extrapolation of ice-surface gradients defined by the features, and topography. The oldest ice margins, below marine limit on the east coast, were simple broad arcs, subparallel to the modern coastline. Here the ice retreated in the sea, probably quite rapidly, as ablation probably was aided by calving. Above the marine limit, the configurations of the ice margins, and hence the patterns of ice flow, during retreat were controlled largely by topography. Generally, the ice mass retreated westward; interfluves appeared first through the ice and large tongues and lobes remained in the valleys. The largest lobe, that in Wrottesley Valley, however, retreated in the sea. Generally, the ice remained a coherent unit during retreat, and remnant ice caps were left at only two places: on plateaus northeast and southwest of Wrottesley Valley.

Figure 43.4. *Oblique aerial view of De Geer moraines about 30 km north of Abernethy Bay. The features are composed entirely of limestone boulders and stones forming 2 m-wide stripes on the surface of a fine grained till. The till supports a nearly complete vegetation cover, whereas the moraines are bare. Seven distinct moraines appear in the foreground and several others in the background. (GSC 203359-T)*

Figure 43.5. *Oblique aerial view of moraine ridges, about 5 m high, composed of till on a bedrock lowland near the head of Abernethy Bay. Three distinct ridges trend from lower left to upper right. The moraines were modified during the Holocene marine regression when the upper part of the till was winnowed and sorted by wave action to form gravel beach ridges, particularly obvious just left of the middle foreground. (GSC 203359-U)*

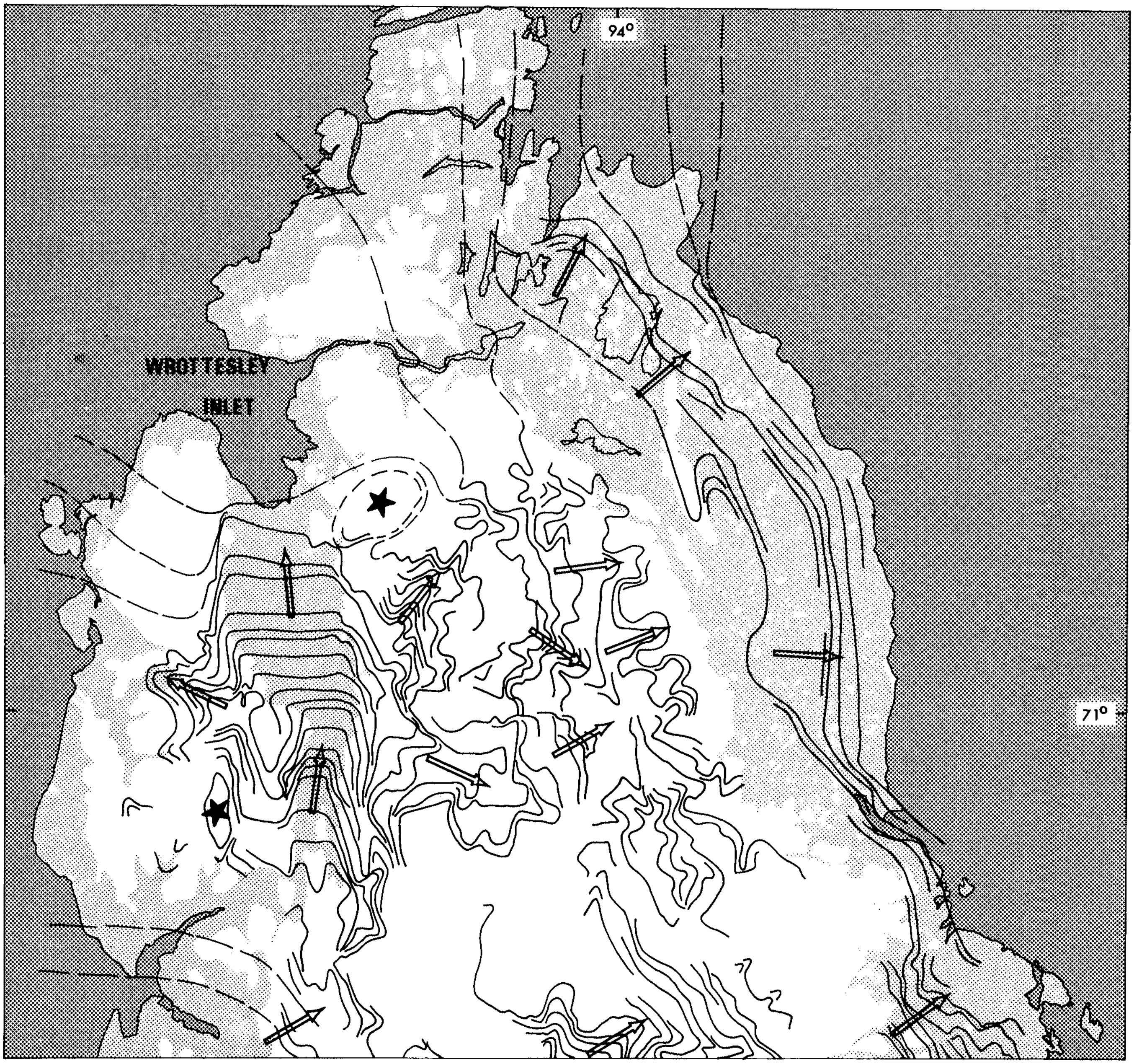

Figure 43.6. *Ice margins (shown by black lines) and flow patterns (arrows) during recession of the late Wisconsin ice sheet on northern Boothia Peninsula. Lightly shaded areas were covered by the early Holocene sea, where it is likely that the ice retreated largely by calving. Two remnant ice caps are marked by stars.*

Marine Limit and Emergence

Marine Limit and Date of Ice Recession

In areas behind the glacial limit, such as northern Boothia Peninsula, the marine limit is formed at the instant of deglaciation of any land which rises high enough to intersect the former water plane. The marine limit on the northeast coast and in Wrottesley Valley is best marked by the uppermost perched deltas, and in a few places by high beaches, pockets of marine sediment, and washing limits on till. On the west coast, adjacent to Wrottesley Valley, the limit is clearly marked by the uppermost raised beaches, above which are till-covered slopes showing no signs of modification by wave action.

Marine deltas deposited by glacial meltwaters are now perched 215 m above sea level on the northeast coast (Fig. 43.7). Fossils of the marine bivalves, **Mya truncata** and **Hiatella arctica**, the common assemblage in early Holocene marine sediments in the Canadian Arctic, were collected from the foreset sands of one of these marine limit deltas, and a single valve of **Mya truncata** yielded a radiocarbon age of 9230 ± 130 years (GSC-2720). The ice margin stood no more than a few kilometres west of the delta at the time of its deposition.

Ice thickness, and therefore crustal depression, increases in the up-ice direction. However, the marine limit may decline in that direction because much rebound can occur as the ice thins before the margin retreats. This is the

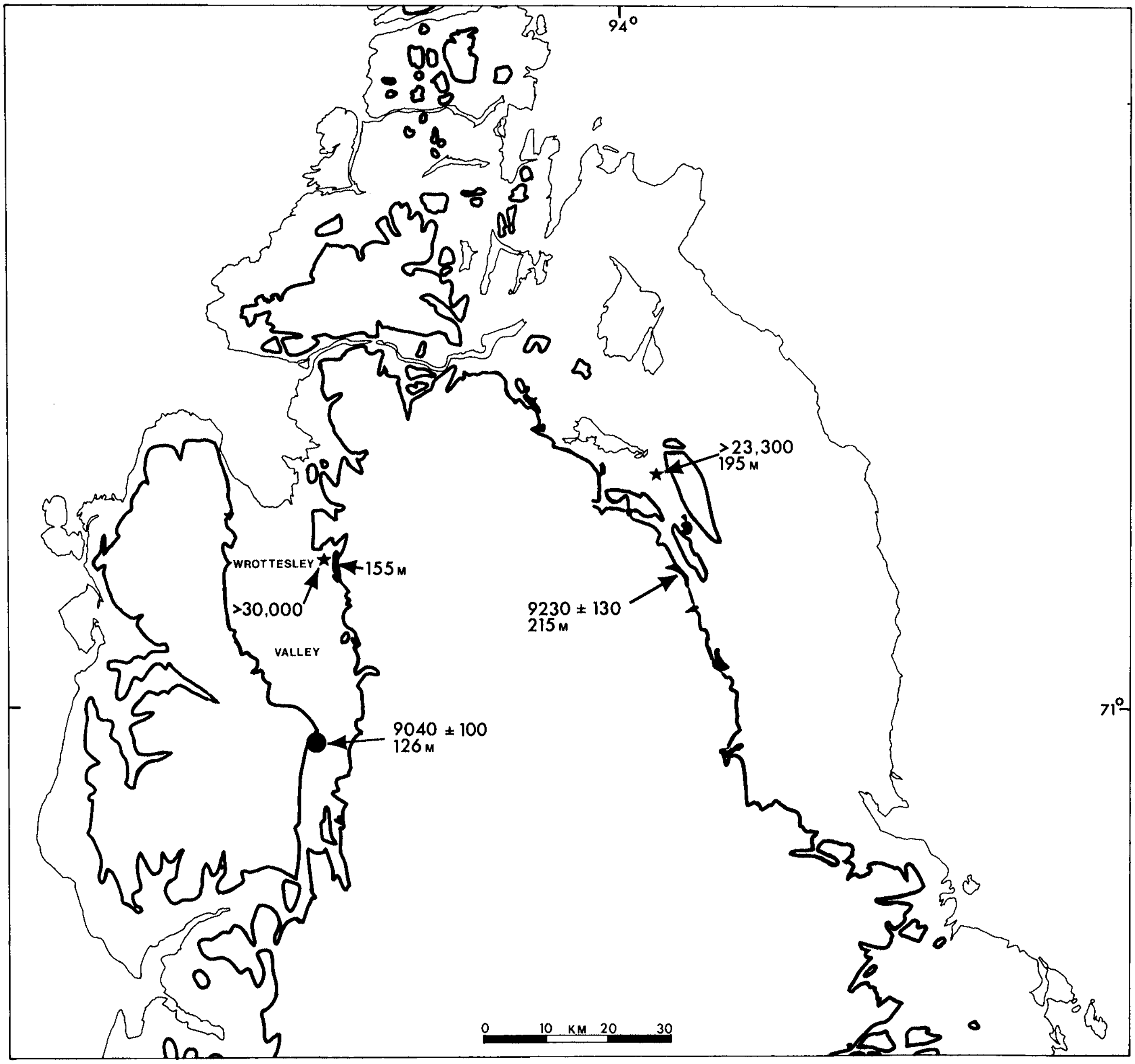

Figure 43.7. Holocene marine limit (heavy line) and marine limit deltas (black areas) and dates on northern Boothia Peninsula, along with locations of samples (shown by stars) of pre-late Wisconsin age.

case on northern Boothia Peninsula, for the marine limit at the mouth of Wrottesley Valley is marked by glaciomarine deltas at only 155 m a.s.l. No marine fossils were found in these deltaic sediments. However, 35 km farther up-valley (southward), where marine limit appears to be at the same elevation or slightly lower, paired valves of **Mya truncata, Hiatella arctica,** and **Macoma calcarea** were collected from a small deposit of silt and sand. The silt and sand fill a depression in till at 126 m a.s.l. Sea level was higher than that during deposition of the sediment, which was probably produced by the melting of the glacier, and may have stood at the marine limit. Therefore, a radiocarbon age of 9040 ± 100 years (GSC-2722), which was obtained on the **Mya truncata** valves, is a reasonable approximation of the date of deglaciation of the site.

These two radiocarbon dates, from opposite sides of the peninsula, show that much of the area was deglaciated in a period of about 200 years, or at most, using maximum quoted error terms (± 2 sigma) on both dates, 400 years. Ignoring the error terms on the dates, the average rate of recession of the ice margin between the two fossil sites was 300 m/year (a recession of 60 km in 200 years). If the ice retreated at the same rate across the area east of the eastern fossil site, then the present coastline was deglaciated about 9350 years ago.

Another perspective on the early Holocene ablation rate can be provided by considering the rate of thinning of the ice sheet, in other words, the rate of lowering of the ice surface. Figure 43.8 shows profiles of the ice sheet along a line that passes through the two fossil sites discussed above, and that fortuitously is parallel to the flow lines in the ice sheet

(Fig. 43.1). These profiles are based on the formula given by Nye (1952) and are reasonable reconstructions as they closely approximate the measured profile of the northern Greenland Ice Sheet. Between 9230 ± 130 and 9040 ± 100 years B.P. the ice surface over the Wrottesley Valley fossil site lowered by 1100 m, at an average rate of 5.5 m/year. Between about 9350 and 9040 years ago, 1450 m of ice melted above the site, at an average rate of 4.8 m/year. The total lowering from the time of the stadial maximum, however, cannot be calculated because the glacial limit off eastern Boothia Peninsula has not been mapped.

Early Holocene Emergence

It is also possible to calculate the rate of coastal emergence, a result of the unloading of ice from the crust, immediately following deglaciation. The Wrottesley Valley date of 9040 ± 100 years B.P. relates to a former sea level more than 126 m a.s.l., but no more than 155 m a.s.l., the height of the local marine limit. The 9040 year old shoreline on the east coast is, therefore, 60 m or more below the 215 m marine limit dated at 9230 ± 130 years. Because the ice thickened westward, it is likely that the 9040 year old shoreline dips eastward (Fig. 43.9) as do the early Holocene shorelines on Somerset Island (Dyke, 1978b). This gives an emergence rate of more than 30 m/100 years.

This rate is much higher than early Holocene emergence rates previously reported for Arctic Canada. For example, Andrews (1970, p. 114) derived a rate of about 3.5 m/100 years for uplift occurring 1000 years after deglaciation for sites in Arctic Canada (emergence rates would be less than uplift rates as Andrews used a eustatic sea level correction), and Blake (1975) gave an average rate of 7 m/100 years for emergence between 9000 and 8000 years ago at Cape Storm, southern Ellesmere Island. The latter area was deglaciated about the same time as northern Boothia Peninsula. Perhaps the explanation of the apparent discrepancy is that the higher rate of emergence reported here for northern Boothia Peninsula applies to only the 200 years immediately following deglaciation, that is, the interval when sea level lowering due to reduced gravitational attraction of the sea by the ice sheet and elastic rebound of the crust are significant factors in determining the emergence rate (Clark, 1976; Farrell and Clark, 1976), especially in areas of rapid ice recession. I am not aware of any other calculation of emergence rates for the first few centuries after deglaciation for any other area.

Evidence of Pre-Late Wisconsin Events

Pre-Late Wisconsin Dates

Because dates on the marine limit in areas behind the glacial limit are also dates on deglaciation, the last ice sheet to cover northern Boothia Peninsula was of late Wisconsin age, and it is likely that the strong eastward to northeastward flow pattern was imprinted during the late Wisconsin stadial maximum. Two samples of pre-late Wisconsin age from northern Boothia Peninsula were dated prior to this study; these require explanation in light of the results presented above.

On the northeast coast (Fig. 43.7), B.G. Craig collected shells, mostly fragments, but some with periostracum still attached, from crossbedded deltaic sands forming terraces up to 195 m a.s.l. The sample, which was dated >23 300 years old (GSC-135; Craig, 1964), contained **Yoldia arctica,** Clinocardium **ciliatum, Astarte** sp., and **Hiatella arctica.** The first three species are uncommon at such elevations; in Craig's extensive fossil collection from the central Canadian Arctic, **Yoldia** occurs no higher than 52 m a.s.l., **Clinocardium** no higher than 113 m a.s.l., and **Astarte** no higher than 91 m a.s.l. (see appendix and table by F.J.E. Wagner in Craig, 1964).

The site was revisited in 1974 by mappers from Terrain Sciences Division, and they noted a veneer of till over the crossbedded sands near the contact of the deposit with the adjacent till-covered hillside (A.N. Boydell et al., 1974, unpublished field notes). Because of its position, the till could either have been deposited directly over the sands by glacier ice or have been redeposited by solifluction from the adjacent hillslope. Likewise, the shells could have been redeposited after transportation by glacier ice, but the presence of periostracum on them argues against this. It is most likely therefore that the shells died during deposition of the deltaic sands and that the overlying till represents the late Wisconsin ice advance. If so, the older marine incursion attained heights similar to those reached by the early Holocene sea. This implies that the pre-late Wisconsin and the late Wisconsin ice loads were of similar size. The original fossil collection was small and entirely consumed to make the age determination, so a re-collection is needed to determine their age more precisely.

On the northwest coast in lower Wrottesley Valley, K.A. Drabinsky in 1974 collected plant remains from bedded sands, which he considered glaciofluvial or marine in origin (Fig. 43.7). A single fragment of **Salix** sp. was dated at >30 000 years old (GSC-2180; Drabinsky in Lowdon and Blake, 1979). M. Kuc (GSC unpublished Bryological Report No. 315) identified the sample as an in situ meso-xeric sod, despite the fact that the material came from bedded sands which had been interpreted as water-laid deposits. The area was re-examined during the course of this study, and it is concluded that the sands from

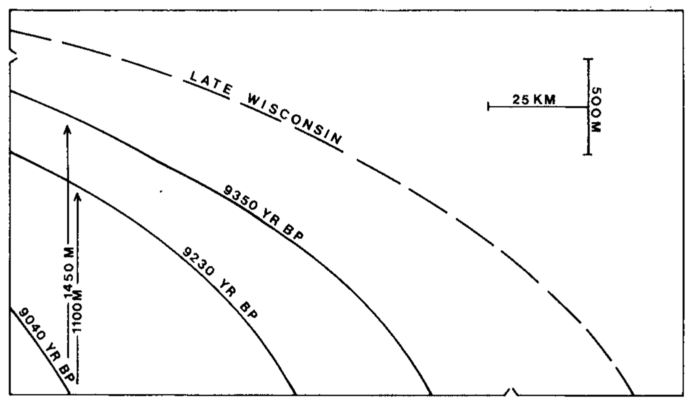

Figure 43.8. *Profiles of the retreating ice sheet along a line passing through the two Holocene fossil sites (cf. Fig. 43.7), showing the amount of ice-surface lowering between 9350 and 9040 years B.P.*

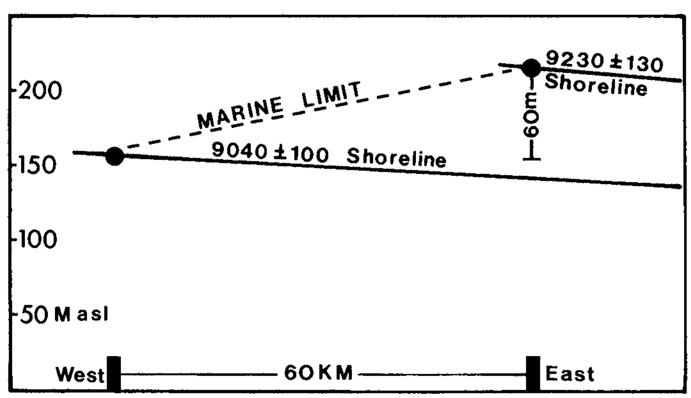

Figure 43.9. *Approximate profiles (cf. Fig. 43.7) of the early Holocene shorelines and marine limit; these allow calculation of the rate of initial coastal emergence.*

which the plant remains were collected are prodeltaic sediments because they occur immediately downslope from a glaciomarine delta terrace marking the marine limit at 155 m a.s.l. (Fig. 43.7). The marine limit was formed during the early Holocene (see above), so the plant material incorporated in the prodeltaic sands must be detrital.

Similar detrital plant remains were collected by the 1974 mapping party from a conical kame composed of sand in the central part of the peninsula (A.N. Boydell et al., 1974, unpublished field notes). Because the organic material occurs in ice-contact sediments, it is likely that it was brought to the site by glacier ice. Therefore, if the absolute age of the material were measured, it probably would give the date of advance of the last ice sheet across the peninsula.

Correlations

Dyke (1978a, b) presented evidence that the eastward flowing ice on northwestern Somerset Island terminated about 20 km inland from the west coast and that the rest of the island, with the exception of two nunataks, was covered by a radially flowing local ice cap. An ice marginal delta formed during recession of the local ice cap in the Creswell Bay area contained driftwood which dated >38 000 years old, and beaches just below the delta contained well preserved shells estimated to be >40 000 years old, but probably younger than Sangamon, on the basis of the extent of amino acid racemization. Hence, it seems that the local ice cap retreated from at least part of the coastal area during middle or early Wisconsin time. It is possible that the deltaic sediments dated at >23 300 years on northeastern Boothia Peninsula are correlatives of the pre-late Wisconsin sediments in Creswell Bay and, therefore, that the main ice sheet also retreated somewhat at that time. On the other hand, it is also possible that the Somerset Island sediments were deposited during a response to a relatively small, short-lived warming, while the main ice sheet remained stationary because of its greater bulk and, hence, longer response time. If this was the case, there need be no correlation between the two deposits.

Conclusions

During the height of the late Wisconsin Stade, an ice mass flowed eastward and northeastward across northern Boothia Peninsula. The ice was coeval with previously recognized eastward flowing ice on southern and western Somerset Island (Dyke, 1978a, b). The main products of this ice flow were hundreds of square kilometres of severely ice moulded bedrock, mainly rock drumlins and crag-and-tail features, concentrated in two areas which were probably overlain by vigorously flowing streams of ice. Striae measurements confirm the interpretation of the lineaments as glacial features. During recession an abundance of ice marginal features formed, including lateral meltwater channels, former ice dammed lakes, and a variety of morainic forms ranging from De Geer moraines, composed only of lines of boulders deposited on the former seabed, to kame moraines, bulky ridges of sand and gravel with ice contact escarpments. These features form the basis for a map of ice retreat. The ice sheet margin retreated westward and its configuration was topographically controlled. Across most of the area, it retreated as a coherent mass, leaving behind only two small remnant ice caps. As the ice retreated, the sea flooded the isostatically depressed land and reached 215 m above its present level on the east coast. Shells from a marine limit delta deposited by glacial meltwater on the east coast have been dated at 9230 ± 130 years. As the ice retreated the land rebounded, and by the time the sea reached Wrottesley Valley, about 9040 ± 100 years ago, it stood only 155 m above its present level. The two dates bracket deglaciation of most of the study area, and allow calculation of average rates of retreat (300 m/year), ice surface lowering (5.5 m/year), and initial coastal emergence (more than 30 m/100 years).

Shells dated >23 300 years old had been collected previously from deltaic sediments at 195 m a.s.l. near the northeast coast (Craig, 1964). These sediments may represent a pre-late Wisconsin ice recession and marine incursion and may correlate with pre-late Wisconsin marine sediments in the Creswell Bay area on Somerset Island. Old plant remains in sand near the northwest coast are interpreted as detritus, redeposited in an early Holocene prodeltaic environment.

References

Andrews, J.T.
1963: Cross-valley moraines of the Rimrock and Isortoq River valleys, Baffin Island, N.W.T.: a descriptive analysis; Geographical Bulletin, no. 19, p. 49-77.

1970: A geomorphological study of postglacial uplift with particular reference to Arctic Canada; Institute of British Geographers, Special Publication No. 2, 156 p.

Blake, W., Jr.
1975: Radiocarbon age determinations and postglacial emergence at Cape Storm, southern Ellesmere Island, Arctic Canada; Geografiska Annaler, Ser. A, v. 57, p. 1-71.

Boydell, A.N., Drabinsky, K.A., and Netterville, J.A.
1975a: Terrain inventory and land classification, Boothia Peninsula and northern Keewatin; in Report of Activities, Part A, Geological Survey of Canada, Paper 75-1A, p. 393-396.

1975b: Surficial geology and geomorphology, Boothia Peninsula, N.W.T.; Geological Survey of Canada, Open File 285, scale 1:125 000.

Clark, J.A.
1976: Greenland's rapid postglacial emergence: a result of ice-water gravitational attraction; Geology, v. 4, p. 310-312.

Craig, B.G.
1964: Surficial geology of Boothia Peninsula and Somerset, King William, and Prince of Wales Islands, District of Franklin; Geological Survey of Canada, Paper 63-44, 10 p.

Dyke, A.S.
1978a: Glacial history of and marine limits on southern Somerset Island, District of Franklin; in Current Research, Part B, Geological Survey of Canada, Paper 78-1B, p. 218-224.

1978b: Glacial and marine limits on Somerset Island, Northwest Territories; Geological Society of America, Abstracts with Programs, v. 10, p. 394.

Farrell, W.E. and Clark, J.A.
1976: On postglacial sea level; Geophysical Journal of the Royal Astronomical Society, v. 46, p. 647-667.

Hoppe, G.
1959: Glacial morphology and inland ice recession in northern Sweden; Geografiska Annaler, v. XLI, p. 193-212.

Kerr, J. Wm. and de Vries, C.D.S.
1977: Structural geology of Somerset Island and Boothia Peninsula, District of Franklin; in Report of Activities, Part A, Geological Survey of Canada, Paper 77-1A, p. 107-111.

Lowdon, J.A. and Blake, W., Jr.
1979: Geological Survey of Canada radiocarbon dates XVIII; Geological Survey of Canada, Paper 78-7.

Nye, F.J.
1952: The mechanics of glacier flow; Journal of Glaciology, v. 2, no. 12, p. 81-93.

Prest, V.K.
1968: Nomenclature of moraines and ice flow features as applied to the Glacial Map of Canada; Geological Survey of Canada, Paper 67-57, 32 p.