Geoscience Canada

Geomorphological Changes and Permafrost Dynamics: Key Factors in Changing Arctic Ecosystems. An Example from Bylot Island, Nunavut, Canada

Michel Allard

Volume 23, Number 4, December 1996

URI: https://id.erudit.org/iderudit/geocan23_4art04

See table of contents

Publisher(s)

The Geological Association of Canada

ISSN

0315-0941 (print)
1911-4850 (digital)

Explore this journal

Cite this article

Allard, M. (1996). Geomorphological Changes and Permafrost Dynamics: Key Factors in Changing Arctic Ecosystems. An Example from Bylot Island, Nunavut, Canada. *Geoscience Canada*, *23*(4), 205–212.

Article abstract

Geomorphological changes related to erosion and sedimentation, periglacial processes, and the thermal regime of permafrost can provide important information on the impact of climatic changes on the terrain in arctic regions. In some areas, organic- (peat-) rich stratigraphie sequences can lead to the interpretation of past changes related to the variable intensity of geomorphological processes.In the Arctic, the dual accumulation of peat and sand is closely related to syn-geneic aggradation of permafrost andto frost cracking, leading to the forma-tion of tundra polygons. Eolian activity may increase in colder and drier periods. Observations of surface changes (geomorphology and vegetation) in a regional network of sensitive sites, along with climate and permafrost temperature monitoring, constitute a comprehensive method to assess the impact of climatic changes on land systems in the Arctic. One such site, on Bylot Island, provides a good example.

Geomorphological Changes and Permafrost Dynamics: Key Factors in Changing Arctic Ecosystems. An Example from Bylot Island, Nunavut, Canada

Michel Allard
Centre d'études nordiques
Université Laval
Sainte-Foy, Québec G1K 7P4
Tel: 418-656-5416
Fax: 418-656-2978
E-mail: michel.allard@cen.ulaval.ca

SUMMARY
Geomorphological changes related to erosion and sedimentation, periglacial processes, and the thermal regime of permafrost can provide important information on the impact of climatic changes on the terrain in arctic regions. In some areas, organic- (peat-) rich stratigraphic sequences can lead to the interpretation of past changes related to the variable intensity of geomorphological processes. In the Arctic, the dual accumulation of peat and sand is closely related to syngenetic aggradation of permafrost and to frost cracking, leading to the formation of tundra polygons. Eolian activity may increase in colder and drier periods. Observations of surface changes (geomorphology and vegetation) in a regional network of sensitive sites, along with climate and permafrost temperature monitoring, constitute a comprehensive method to assess the impact of climatic changes on land systems in the Arctic. One such site, on Bylot Island, provides a good example.

RÉSUMÉ
Les changements géomorphologiques associés à l'érosion et la sédimentation, aux processus périglaciaires et au régime thermique du pergélisol sont porteurs d'information sur l'impact des changements climatiques dans les territoires arctiques. Dans certaines régions, des séquences stratigraphiques riches en tourbe peuvent permettre l'interprétation de changements environnementaux passés, lesquels sont liés a l'intensité variable des processus géomorphologiques. Dans l'Arctique, l'accumulation combinée de tourbe et de sable est associée à l'aggradation syngénétique du pergélisol, lui-même affecté par la fissuration au gel qui conduit à la formation de polygones de toundra. L'activité éolienne augmente vraisemblablement durant les périodes froides et sèches. L'observation des changements écologiques (géomorphologie et végétation) selon un réseau régional de sites sensibles, en concertation avec un suivi du climat et des températures du pergélisol, constitue une méthode intégrée pour évaluer l'impact des changements climatiques dans l'Arctique. Un de ces sites, sur l'île Bylot, constitue un bon exemple.

INTRODUCTION

Arctic ecosystems evolve on landforms and soils that endured changes during the Holocene. The processes responsible for that evolution are still active today. Although the whole landscape may be affected by processes associated with climate and ecological changes, some areas are particularly sensitive and may evolve faster because they have a concentration of fragile permafrost landforms and are the location of a large number of intensive processes associated with rivers, winds, periglacial conditions, and vegetation dynamics. Those more dynamic areas usually have a Holocene and recent history that can be reconstructed through stratigraphic analysis of sediment layers and paleosols, buried vegetation layers, and recent landform patterns (*e.g.*, dunes, pingos, tundra polygons, active layer failures) superimposed over older ones. Areas such as Hot Weather Creek on Ellesmere Island (Gajewski *et al.*, 1995; Garneau, 1992; Lewkowicz, 1990), Bracebridge Inlet on Bathurst Island (Blake, 1974), and the eolian landforms associated with tundra polygons on Banks Island (Pissart *et al.*, 1977) are good examples. In the northern Baffin region, Qunguliqtut (an unofficial name, used according to information provided at the NEAT workshop by people from Pond Inlet), a valley on Bylot Island (Fig.1), is also an excellent example.

In order to illustrate the continuity from Holocene through to present evolution of the land surface, the history of the valley is presented, starting from deglaciation in the Early Holocene. Thereafter, the actual dynamics of permafrost landforms and soils are described and

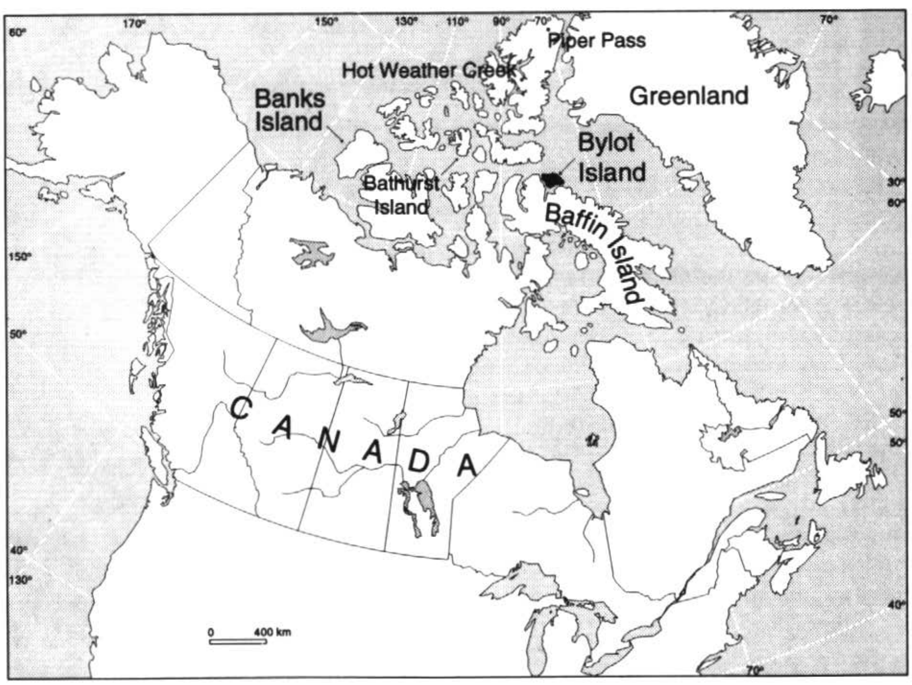

Figure 1 *Location of Bylot Island, and Arctic sites of sand and peat sections referred to in the text.*

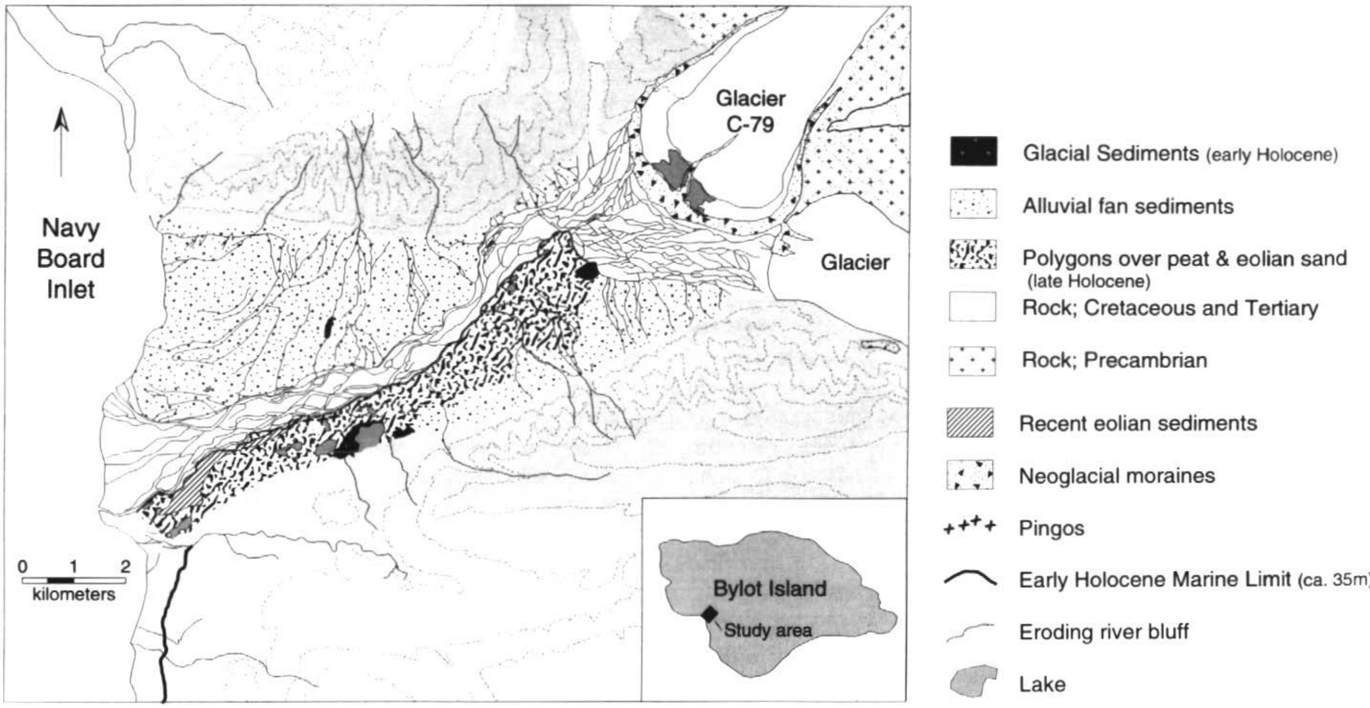

Figure 2 *Surficial geology and landforms in "Qunguliqtut."*

discussed, with some consideration for the interaction between soils, vegetation and the abundant bird population. Finally, several recommendations are presented on how a simple monitoring program could be set up in Qunquliqtut and similar arctic environments.

HOLOCENE HISTORY OF THE QUNGULIQTUT VALLEY

Located on the southwestern side of Bylot Island, the valley is approximately 15 km long and extends from the Precambrian core of the island to the shore of Navy Board Inlet (Fig. 2). It cuts across the Cretaceous and Tertiary formations of the arctic lowlands (Bostock, 1970; Jackson and Sangster, 1987). Two glacier snouts flow into the head of the valley, originating from the intramontane ice cap that covers the major part of the island. A glacio-fluvial outwash, or sandur, is fed by the glaciers and terminates in Navy Board Inlet at a delta (Fig. 2).

Early Holocene: Glacial Retreat and Marine Submergence

Mounds of ice-contact deposits occur on the floor of the valley about midway between the sea shore and the actual glacier front (Fig. 2, 3 and 4). Another isolated mound also exists upstream, about 2 km from the glacier front. Sediments of the ice-contact deposits consist of diffusively stratified coarse sands and gravels with numerous stones. The lithological composition includes material from both Cretaceous–Tertiary and Precambrian sources, strongly suggesting that the deposits were laid by an inland-based glacier, although secondary or tertiary transport of boulders carried from Baffin Island during an earlier glaciation cannot be totally ruled out (Klassen, 1985, 1993). These ice-contact sediments are interpreted to mark past positions of the glacier front in the valley (Fig. 5). At the site of the ice-contact deposit mound closest to the present shoreline, shelly, fine-grained marine sediments outcrop below the glacial drift, on the bank of a lake. Shells collected at this site, all of the *Mya truncata* species, gave a radiocarbon date of 9860±140 BP (before present) (UL-1028, Table 1). As the sampling site is about 15 metres above sea level (masl), it can be concluded that the glacier front was tidal and advanced into

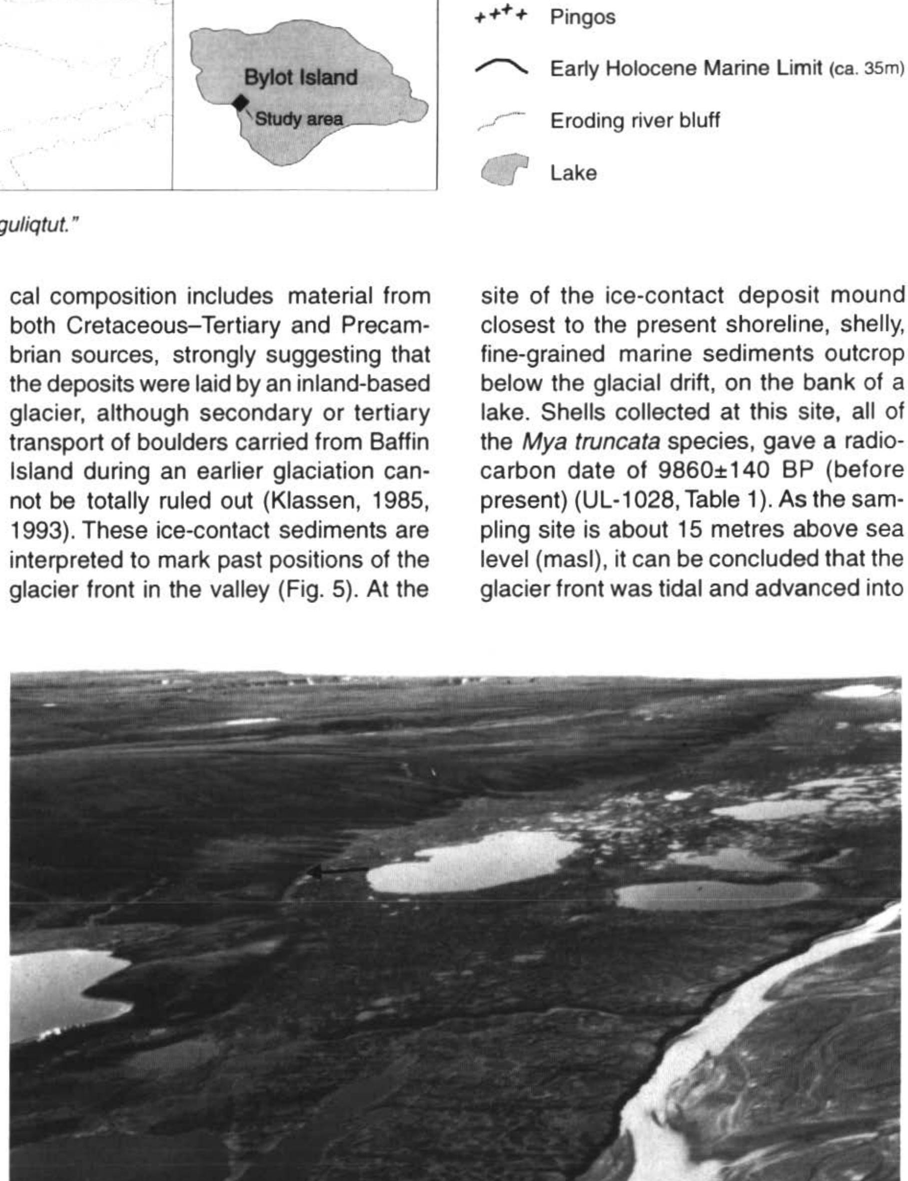

Figure 3 *Oblique aerial view of the left bank terrace. The camp is on the ice-contact sediment mound. A marine shoreline about 12 masl runs along the foot of the mound (arrow). Note the eroding river bluff and thaw lakes. View southwards.*

shallow marine waters. This frontal position at that time is the only known evidence on Bylot Island of an island glacier advance at the end of the Wisconsinan or during the Early Holocene. It is approximately correlative with a glaciofluvial delta dated at 9530±180 (GSC-3318, Table 1) (75 masl) at Cape Hatt, which lies at the northernmost extension of the Wisconsinan ice sheet on Baffin Island, south of Navy Board Inlet (Klassen, 1993). The isostatic depression responsible for the marine flooding of the Qunguliqtut valley under 15-20 m of water can be explained by the load of the Wisconsin ice on Baffin Island and the location of southern Bylot Island at the margin of the peripheral depression of Baffin ice. In fact, the Early Holocene marine limit at about 35 masl is clearly visible on air photographs and in the field along the shore of Navy Board Inlet (outside the valley) (Fig. 2). Shell samples from sediments in the cliff along that shore gave dates between 8710±120 BP (UL-1032, Table 1) and 6100±70 BP (GSC-2948, Table 1).

Downstream from the Early Holocene (or late Wisconsin?) ice-front position, marine sediments outcrop in two eroded pingos on the river bank. Elsewhere along the river, only late Holocene sediments appear (see below), but the growth of the ice cores in the pingos upheaved the underlying strata, exposing sediments from deeper than the present river bed level (Figs. 6 and 7). Despite cryogenic and hydrostatic deformation of strata in the pingo, the diapiric appearance of the silty clay body in the pingo core and the sideways dip of glaciofluvial sands and gravels strongly suggest that the glaciofluvial sediments originally overlaid the marine sediments. As the glacier ice-front receded up-valley, outwash sediments were deposited over the marine clays. This retreat took place after 6000 B.P., as radiocarbon dates on shells from the clays in the two pingos gave ages of 6430±110 (UL-1347, Table 1) and 6020±80 (UL-1027, Table 1). As the next glacier frontal position upstream (Fig. 5) is not dated, it is not known whether it belongs to a halt during the mid-Holocene recession or to a possible readvance that could have occurred more recently.

Figure 4 *Ice-contact sediment mound across the valley, marking the 9.8 ka frontal position (see Fig. 5).*

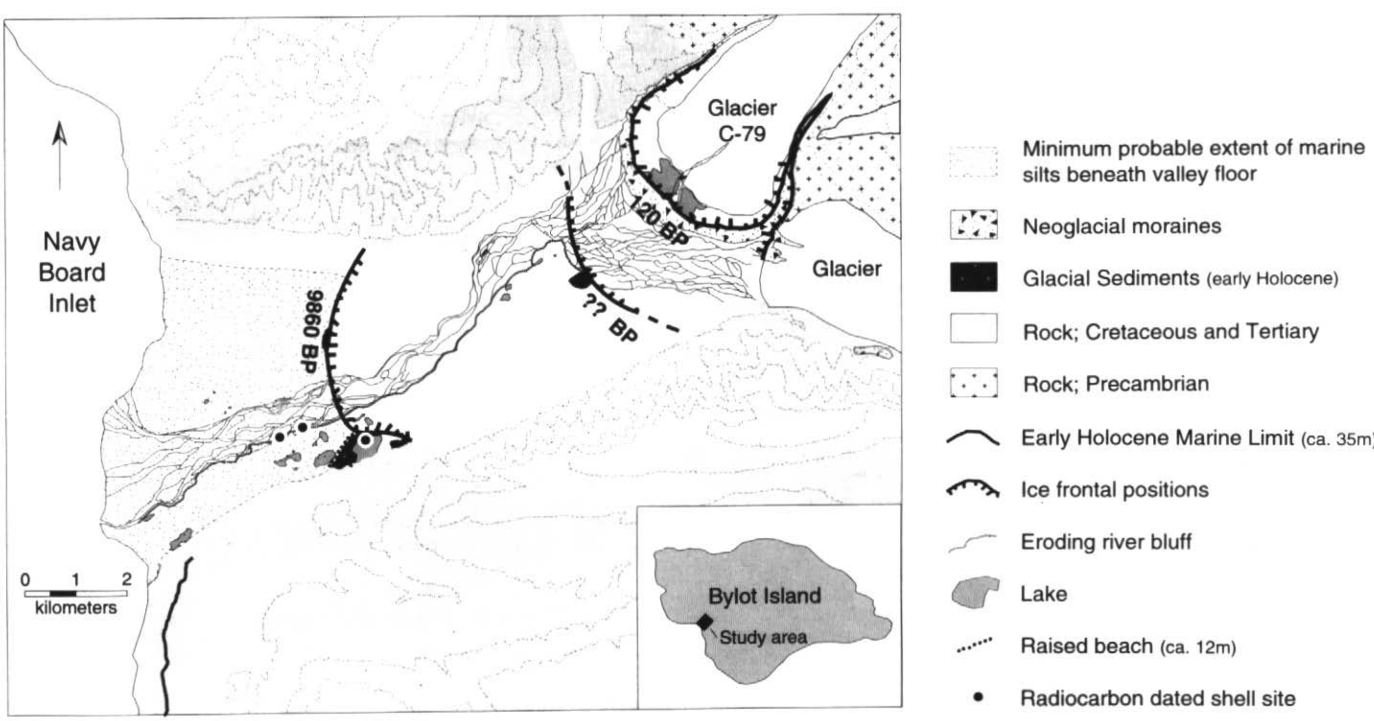

Figure 5 *Holocene glacial frontal positions in the valley, marine sediments and landforms, shell radiocarbon date sites.*

Late Holocene Events

During the Late Holocene (roughly the last 3000 years), glaciers readvanced, eolian and organic sediments accumulated over the valley floor, and permafrost landforms evolved into their present configuration.

Neoglacial Advance

With a few exceptions, most of the fronts of glacier tongues on Bylot Island have recently receded from fresh-looking frontal moraines. Such is the case for glacier C-79 in Qunquliqtut (Fig. 2). Klassen (1993, table 1.1, p. 53) reported a radiocarbon date of 120±80 (GSC-3227) on "twigs of Salix collected from buried soil within a thrust plate forming part of neoglacial moraine of glacier C-79. Maximum extent of ice during neoglacial time occurred subsequent to growth of twigs". The glacier ice-front retreat from the frontal moraine, a distance of about 1 km, took place between 1958 and 1982, as can be determined by comparing air photographs. Although "Neoglacial" refers to a global cooling beginning 3500-3000 years ago, the very recent age of this moraine and of the glacier ice-front retreat suggests that the readvance was initiated during the Little Ice Age which lasted from *ca.* 1550 AD to *ca.* 1880 AD, possibly the coldest period of the entire Holocene (Bradley, 1990).

Sedimentation and Erosion on the Qunguliqtut Valley Floor

The valley floor is covered predominantly by alluvial fan sediments (Fig. 2). On the north bank of the river, torrents flow into deep gullies that incise the poorly consolidated Cretaceous–Tertiary sandstones and shales. The transported sediments, dominantly sands, are spread by multiple channels down to the bank along the river outwash. This continuing sedimentation on the north side of the valley forces the river to migrate laterally to the south, where it is eroding its bank.

The flat terrace on the south bank of the river is built of a sequence of interlayered peat and fine sand deposits. At low-water stages, a peat-rich layer is exposed along almost the whole length of the valley. The layer consists of abundant herb roots and plant remains in stratified sand. Its total thickness is difficult to assess because the base is below river bed level, and because the number and thickness of interstratified sand layers vary along the length of the valley. It is at least 1.2 m thick.

Table 1 Radiocarbon dates. See text for the historical significance of the dates.

Lab. no	Age BP	Material	Cal. date and range*
UL-1338	Modern	Peat	1950-1993 AD
UL-1337	50±100	Peat	1810-1925 AD or after 1950 AD
UL-1336	190±110	Sandy Peat	1801 AD 1718-1820 AD
UL-1047	690±90	Peat	1292 AD 1261-1332 AD
UL-1342	710±190	Sandy Peat	1288 AD 1219-1331 AD
UL-1033	1050±90	Peat	997 AD 936-1058 AD
UL-1048	2210±120	Peat	349 BC 390-157 BC
UL-1034	2510±90	Peat	614 BC 660-518 BC
UL-1035	2600±90	Peat	791 BC 831-753 BC
UL-1025	2900±90	Peat	1064 BC 1132-970 BC
UL-1027	6020±80	Mixed shells	4467 BC 4563-4373 BC
GSC-2948	6100±70	Mixed shells (Navy Board Inlet Klassen, 1993)	4556 BC 4657-4490 BC
UL-1347	6430±110	Mixed shells	4933 BC 5064-4808 BC
GSC-3062	7880±70	Mixed shells (Navy Board Inlet Klassen, 1993)	6360 BC 6411-6261 BC
UL-1032	8710±120	Mixed shells (Navy Board Inlet)	7412 BC 7508-7244 BC
UL-1028	9860±140	Shells (*Mya truncata*)	8796 BC 8930-8525 BC

*Peat dates were calibrated with the dendrochronological calibration curve of Stuiver and Becker (1993); shell dates calibrated with marine calibration model of Stuiver and Braziunas (1993).

Figure 6 *Eroded pingo along the riverbank. The dark soil in centre is marine clay (see Fig. 7). Pack gives the scale.*

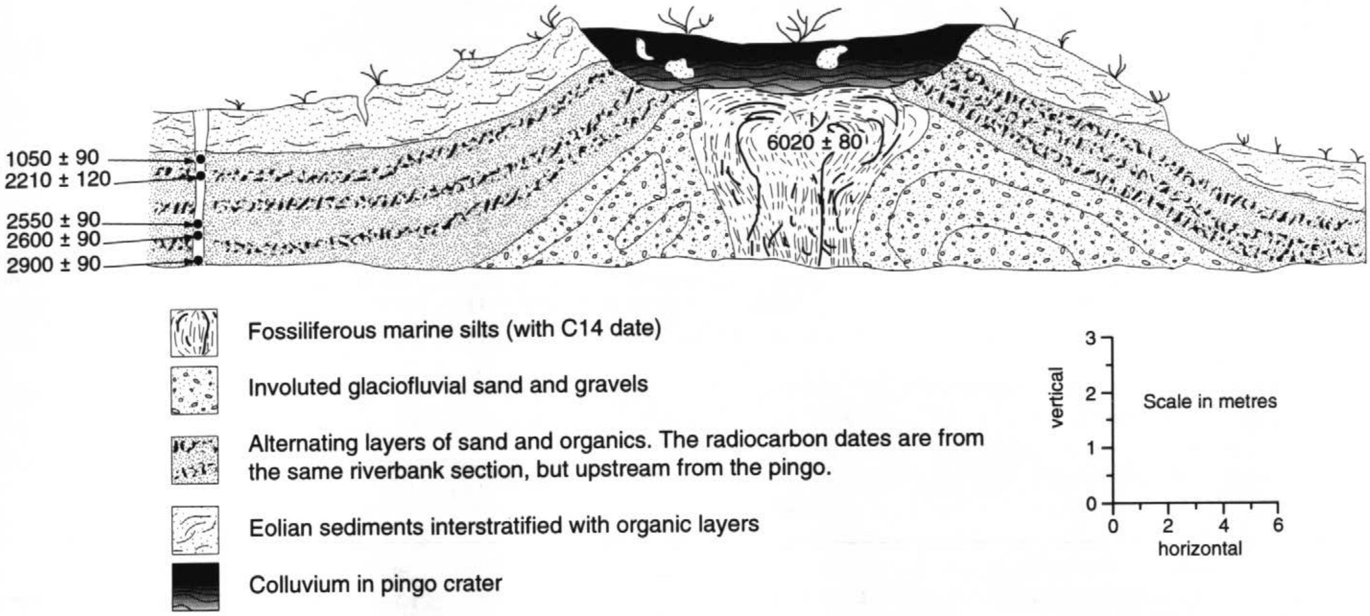

Figure 7 *Sketch of the stratigraphy and structure observed in an eroded pingo.*

The lowermost organic sample recovered from this peat-rich layer, 3.2 m below bank top, gave a radiocarbon age of 2900±90 B.P. (UL-1025, Table 1, Fig. 7). In another section upstream, a sample at a depth of 2.5 m gave an age of 2600±90 (UL-1035, Table 1), and another one at 2.28 m was dated at 2510±90 B.P. (UL-1034, Table 1). Above the peat-rich layer, the sequence is made up of medium and fine sand layers including dispersed vegetation debris and, occasionaly, thin layers of organics. An organic layer, sampled at 1.17 m below the cliff top, was dated at 1050±90 B.P. (UL-1033, Table 1). From that depth to the present surface, the soil consists of fine sand and silt interstratified with thin organic-poor layers. Plotting the calibrated radiocarbon dates from the south bank cliff (Table 1) against depth provides an estimate of variations in accumulation rates and build-up of the terrace (Fig. 8). The plot clearly shows a faster initial build-up at an average rate of 0.25 cm.a^{-1}, followed by a slower accumulation after 2200 B.P. (350 BC), corresponding with a general change from an organic-rich to a sand-rich facies.

The valley floor had completely emerged from the sea 3000 years ago. Sand, deposited probably from alternating and mixed fluvial and eolian events, is present at the river bed level and to some depth below. An organic-rich soil was supporting a wet tundra covering the central portion of the valley 2900-2500 years ago. This layer suggests a wetter, and maybe a warmer, environment than at present. Thereafter, the eolian component in the terrace sediments increases, suggesting drier, and perhaps cooler, conditions. Over the last 1050 years, approximately, eolian sedimentation played an important part in the accretion process over the south bank terrace.

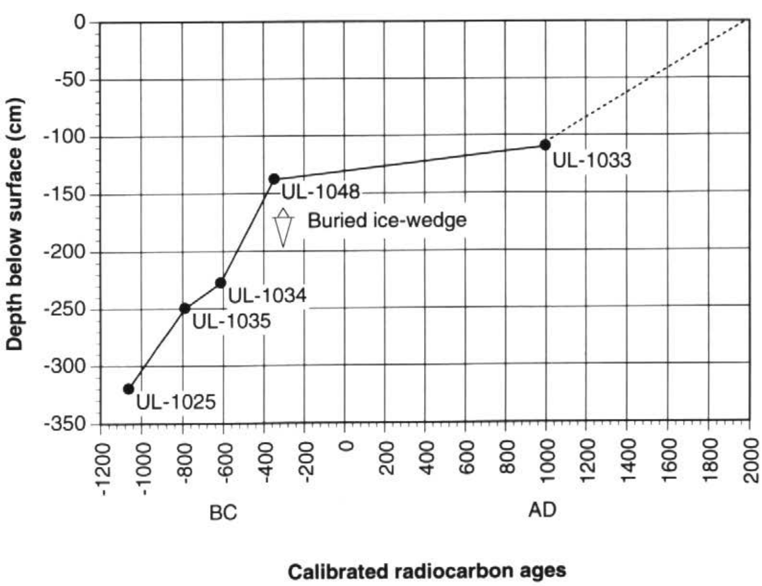

Figure 8 *Age-depth curve for peat and eolian sediments of the south bank terrace. The three lower dates come from a section downstream from the pingos (Fig. 2), the two uppermost ones from other sections upstream.*

In comparison, the north bank of the river valley is low and, apparently, aggrades over the outwash; the braided river system slowly migrates laterally. The outwash surface is actually the main source of eolian sand during low-water stages in dry and cold summer weeks. Large amounts of blown sand are carried by the northwesterly winds and sparsely scattered over the wetlands. Near the delta, where the fetch is larger, a veneer of fresh eolian sediments covers the tundra and masks the polygonal terrain (Figs. 2 and 9).

Ground Ice and Permafrost Landforms

Three principal landform types make up the valley bottom: tundra polygons, thaw lakes, and pingos. The tundra polygons result from repeated frost cracking of the ground in winter due to thermal contraction of the soil at low sub-freezing temperatures; surface water flows into the open cracks during snow melt and forms ice veins which grow thicker with time (centuries and millenia), forming ice wedges (Lachenbruch, 1962). The wedges form a network of polygons 5 m to 40 m in diameter. As the wedges grow thicker, they displace the enclosing soil, forming ridges that ultimately enclose low-centre polygons. In section, the top of the ice wedges coincides with the permafrost table.

In the river bank section, both active and fossil (buried) ice wedges can be observed. One of the fossil wedges was overtopped by a smaller outgrowth feature, indicating an elevation of either the soil surface or the permafrost table (Fig. 10). It was buried under cold climatic conditions, favourable for wedge growth. But this took place during a period when eolian and organic accretion was rapid enough to overtake active layer elevation; at a rate of 0.33 cm.a-1, it is seemingly the period of fastest accumulation (Fig. 8) (see also Lewkowicz, 1994, for examples of buried ice wedges in the eastern high Arctic). Other features, still active today, have characteristics of syngenetic wedges (Mackay, 1990), indicating the upward movement of the active layer at the pace of sediment accretion. The ice wedges and the polygon patterns had a complex evolution over the last 3000 years, with some polygons being fossilized, others changing shape as ice wedges along their sides became either deactivated or reactivated, and as new ones formed under newly deposited soils.

Low-centre polygons gather surface water and humidity. Some have small ponds. Water also accumulates in furrows along the sides of polygons when ice wedges melt. These water bodies coalesce and make up larger lakes under which permafrost thaws; lake shorelines also recess by thermal erosion. Such a series of processes has been described from similar environments elsewhere, for example in northern Alaska (Billings and Peterson, 1980). Eventually, as the river bank retreats, some lakes are catastrophically drained. New polygon networks are reinitiated in those drained lake flats (Mackay, 1986). In some cases, pingos grow in the thaw lake basins (Mackay, 1979). Eleven pingos are scattered over the south bank terrace (Fig. 11). As mentioned above, several of them are actually being eroded by river bank retreat.

Recent Changes in Polygonal Patterns and Inception of Pingos

On the south bank terrace, some areas have evolved into fields of high-centre polygons owing to melting of their ice wedges by running water. The transition from low-centre to high-centre forms induces a major change in surface drainage conditions, as depressed wetlands are transformed into small plateau surfaces. The living mosses die out and are covered by gray lichens such as *Stereocolon* sp. As the age of the dead peat just beneath the lichen surface provides a maximum date for the form change, four attempts at dating were made. The dates obtained were 710±110 BP (UL-1342, Table 1), 190±110 BP (UL-1336, Table 1), 50±100 BP (UL-1337, Table 1), and modern (UL-1338,Table 1). They suggest that for several areas on the terrace, change in polygonal form from depressed to high centre is a process that was active for the several last centuries, and that it continues today.

Similarly, a buried vegetation layer beneath 25 cm of drained lake sediments beside a pingo 8 m in height gave an age of 690±90 BP (UL-1047, Table 1). Since enough time must be allowed for lake flooding over the terrestrial organic matter, talik formation beneath the lake, and drainage by riverbank erosion, the pingo must be younger than that age by many decades.

In summary, the south bank terrace supports a very dynamic and sensitive assemblage of polygons, thaw lakes and pingos. These landforms, and the vegetation patterns on them, have changed continuously over the last 3000 years. The soils were formed by the deposition of eolian fine sand blown off the glaciofluvial outwash by the dominant winds during dry (and colder?) spells, and the accumulation of organic layers dominated by sedges and mosses in wetter (warmer?) spells.

The wet tundra, particularly the areas of low-centre polygons and thaw lakes, at present is grazed intensively by large populations of birds dominated by snow geese. The massive consumption of herbs by these birds and the recycling of nutrients, particularly nitrogen, severely influences the vegetation cover. With time, this process is very likely to affect the polygonal terrain; grazing may lead to wind erosion in some places after destruction of the vegetation cover. In other places, polygons and permafrost thermal integrity may be favoured by increased moss growth related to changes in nutrient cycling (Gauthier *et al.*, 1996). It appears likely that the increase in animal populations will have an impact on the tundra features; this impact needs to be better assessed.

CONSIDERATIONS FOR A MONITORING PROGRAM

The information presented here needs to be better developed in order to base any monitoring of future changes on as complete an understanding of past and present conditions as can be obtained. More detailed work should then be based on precise mapping of polygon and vegetation types in the valley, with empha-

Figure 9 *Oblique aerial view of eolian sand covering tundra polygons near the deltaic section of the glaciofluvial outwash (see Fig. 2).*

sis on the south bank terrace. This mapping should consider polygon types and states of evolution between types, as well as modern peat cover thickness and composition. At the same time, coring across the terrace, feasible now with portable shallow drilling equipment, can provide a better spatial definition of the Holocene dynamics of the peat–eolian sand sequences. Paleoecological interpretations can be based on analysis of vegetal remains and on diatoms (since many layers of pond and lake sediments must be incorporated in the stratigraphy; see Vincent and Pienitz, 1996) and can be supported by abundant ^{14}C dating.

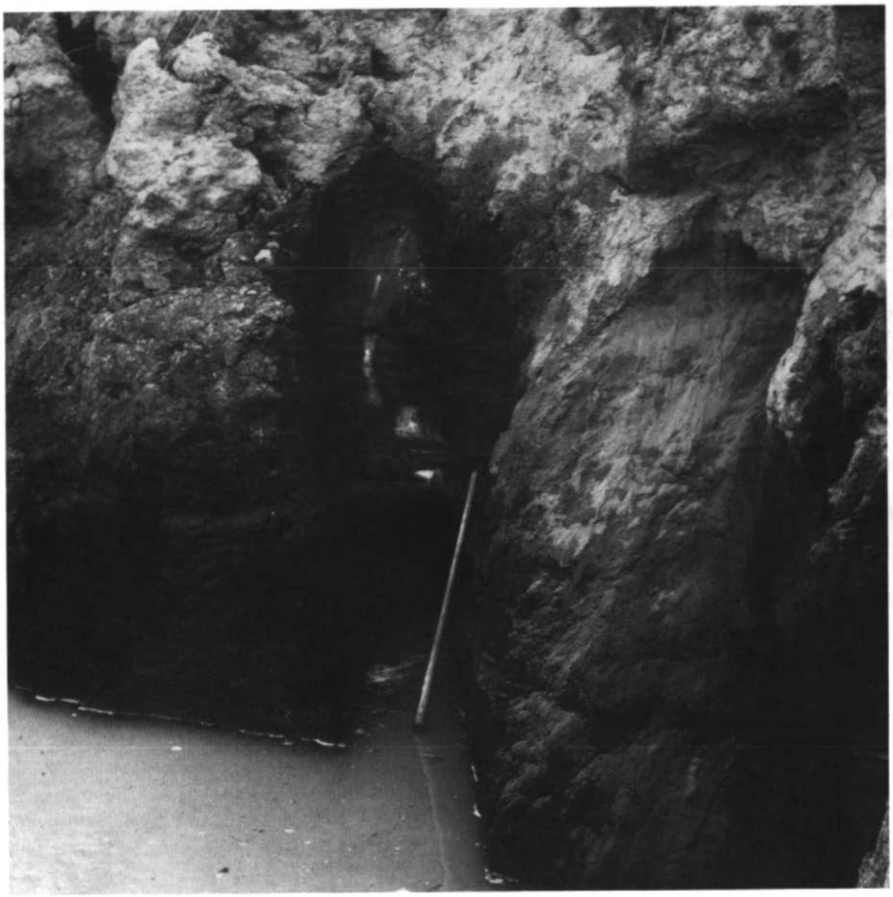

Figure 10 *Buried ice wedge exposed in the riverbank section. The basal "mother" wedge is about 50 cm wide (receded back in the cave by thermo-erosion). Syngenetic wedges (vertical ice bodies) outgrew above it for some time following sediment accretion. Shovel handle is about 1.25 m long.*

Figure 11 *A pingo and polygons on the south bank terrace.*

One or two deep (>100 m) ground thermal profiles could also substantiate the Holocene geological history. Glacial advances, marine inundation, and major geomorphological changes must have forced a very different ground thermal history in the valley as compared with other areas of the island. In fact, the unglaciated plains of Bylot island experienced much less surficial change throughout the Holocene. Taylor *et al.* (1996) have shown the potential use of ground thermal history to understand Holocene geological events in the Mackenzie delta. This approach resembles a similar proposal by Moorman *et al.* (1996) for permafrost thermal monitoring.

A series of field sites, or quadrats, should be selected within which topographical, geomorphological, and vegetation changes can be surveyed repeatedly over many years. The sites should represent the whole array of patterns of polygons, thaw lakes, and pingos. New aerial photographs should be obtained in order to analyse spatially and quantitatively recent changes in the valley.

Finally, several (3-4) thermistor strings should be installed in holes drilled to depths of approximately 20 m. Near-surface permafrost temperature variations, above the depth of zero annual amplitude, record a filtered and integrated signal that is proportional to surface temperature changes, and reflect climatic tendencies. This is a very important aspect of any monitoring program since the eastern Canadian Arctic is affected by climate anomalies related to the atmospheric and oceanic circulation over the North Atlantic and Baffin Bay, and is under-equipped in ground thermal monitoring sites. The anomalies apparently counteract the forced global warming trend observed elsewhere, as documented in the Hudson Strait region (Allard *et al.*, 1995). Documenting ground temperatures and supporting a number of automatic meteorological stations in the eastern Arctic is necessary to understand regional trends (Jacobs, 1996). Also, these thermal measurements would provide an important means to tie together climatic changes, permafrost thermal changes, and surface changes in geomorphology and vegetation. Such a geomorphology-based approach would complement glacier mass balance monitoring in the region (Bell and Jacobs, 1996), in order to assess the impact of climate changes on land systems.

According to the surficial geology map of Bylot Island and adjacent areas (Klassen, 1993), Qunguliqtut is unique in the region for its fine sands and its concentration of permafrost features, an observation confirmed during a reconnaissance flight over the island and the northern coast of Baffin Island. These basic geological characteristics underlie the development in the area of the valley of the richest ecosystem in the region. Other sensitive landscapes, either with similar or different geology and processes, should be identified across the eastern Arctic, in the Baffin region, and the Hudson Strait region in order to organize a geographically representative network for monitoring climatic and ecological changes. In particular, a regional effort for detecting wide-ranging Holocene changes through correlations of environmental reconstructions from peat sections should be undertaken. Although the observed changes at each site tend to be dominated by local geomorphological and ecological processes, some general trends should be revealed through a joint regional effort (Bradley, 1990; Gajewski *et al.*, 1995 (Hot Weather Creek); Lafarge-England *et al.*, 1991 (Piper Pass, Fig. 1); Ovenden, 1988).

ACKNOWLEDGEMENTS

The author is grateful for access to Gilles Gauthier and Austin Reed's field camp and the friendly support of their staff during two visits in 1993 and 1995. Agreement from the Hunters and Trappers Organization of Pond Inlet (Mittimatallik) in a meeting in 1995 to conduct this initial reseach work is very much appreciated. Helicopter support from the Polar Continental Shelf Project of Natural Resources Canada was essential, and much appreciated. This research was supported by NSERC. The author thanks Drs. Chris Burn and Trevor Bell for their useful review of the manuscript. The organizers of the Nunavut Environment Assessment Transect (NEAT) workshop provided an excellent occasion for presentation of the results to the community and to colleagues.

REFERENCES

Allard, M., Wang, B. and Pilon, J.A., 1995, Recent cooling along the southern shore of Hudson Strait, Quebec, Canada, documented from permafrost temperature measurements: Arctic and Alpine Research, v.27, p.157-166.

Bell, T., and Jacobs, J.D., 1996, Glacier monitoring for climate change detection in Nunavut: Geoscience Canada, v. 23, n. 4, p. 195-204.

Billings, W.D. and Peterson, K.M., 1980, Vegetational change and ice-wedge polygons through the thaw-lake cycle in Arctic Alaska: Arctic and Alpine Research, v. 12, p. 413-432.

Blake, W., Jr., 1974, Periglacial features and landscape evolution, Central Bathurst Island, District of Franklin: Geological Survey of Canada, Paper 74-1, Part B, p. 235-244.

Bostock, H.S., 1970, Physiographic subdivisions of Canada, *in* Douglas, R.J.W., ed., Geology and Economic Minerals of Canada: Geological Survey of Canada, Economic Geology, Report n. 1, p.9-30.

Bradley, R.S., 1990, Holocene paleoclimatology of the Queen Elizabeth Islands, Canadian High Arctic: Quaternary Science Reviews, v. 9, p. 356-384.

Gajewski, K., Garneau, M. and Bourgeois, J.C., 1995, Paleoenvironments of the Canadian High Arctic derived from pollen and plant macrofossils: problems and potentials: Quaternary Science Reviews, v.14, p.609-629

Garneau, M., 1992, Analyses macrofossiles d'un dépôt de tourbe dans la région de Hot Weather Creek, péninsule de Fosheim, île d'Ellesmere, Territoires du Nord-Ouest: Géographie physique et Quaternaire, v. 46, p. 285-294.

Gauthier, G. 1996, The exploitation of wetland ecosystems by herbivores on Bylot Island: Geoscience Canada, v. 23, n. 4, p. 253-259.

Jackson, G.D. and Sangster, D.F., 1987, Geology and resource potential of a proposed national park, Bylot Island and northwest Baffin Island, Northwest Territories: Geological Survey of Canada, Paper 87-17, 31 p.

Jacobs, J.D. 1996, Climate anomalies in Northeastern Canada–Reflections from Baffin Island: Centre d'études nordiques, Abstract of a talk, unpublished typed document, 3 p.

Klassen, R.A., 1985, An outline of glacial history of Bylot Island, District of Franklin, N.W.T., *in* Andrews, J.T., ed., Quaternary Environments, Eastern Canadian Arctic and Western Greenland: Allen and Unwin, Boston, p. 428-460.

Klassen, R.A., 1993, Quaternary geology and glacial history of Bylot Island, Northwest Territories: Geological Survey of Canada, Memoir 429, 93 p. [with map 1686A, Surficial geology, Bylot Island and adjacent areas, scale 1:250 000].

Lachenbruch, A.H., 1962, Mechanics of thermal contraction cracks and ice-wedge polygons in permafrost: Geological Society of America, Special Paper 70, 69 p.

Lafarge-England, C., Vitt, D.H. and England, J., 1991, Holocene soligenous fens on a high Arctic fault block, Northern Ellesmere Island (82ON), NWT., Canada: Arctic and Alpine Research, v. 23, p. 80-98.

Lewkowicz, A.G., 1990, Morphology, frequency and magnitude of active layer detachment slides, Fosheim Peninsula, Ellesmere Island, N.W.T., *in* Permafrost-Canada, Proceedings of the Fifth Canadian Permafrost Conference: Université Laval, Nordicana n. 56, p.111-118.

Lewkowicz, A.G., 1994, Ice-wedge rejuvenation, Fosheim Peninsula, Ellesmere Island, Canada: Permafrost and Periglacial Processes, v. 5, p. 251-268.

Mackay, J.R., 1979, Pingos of the Tuktoyaktuk area: Géographie physique et Quaternaire, v. 33, p. 3-62.

Mackay, J.R., 1986, The first 7 years (1978-1985) of ice-wedge growth, Illisarvik experimental drained lake site, western Arctic coast: Canadian Journal of Earth Sciences, v. 23, p. 1782-1795.

Mackay, J.R., 1990, Some observations on the growth and deformation of epigenetic, syngenetic and anti-syngenetic ice wedges: Permafrost and Periglacial Processes, v. 1, p. 15-29.

Moorman, B.J., Michel, F.A. and Drimmie, R.J., 1996, The climate record preserved by glaciers: Geoscience Canada, v. 23, n. 4, p. 189-194.

Ovenden, L., 1988, Holocene proxy-climate data from the Canadian Arctic: Geological Survey of Canada, Paper 88-22, 11 p.

Pissart, A., Vincent, J.S. and Edlund, S.A., 1977, Dépôts et phénomènes éoliens sur l'île de Banks, Territoire du Nord-Ouest, Canad: Canadian Journal of Earth Sciences, v. 14, p. 2462-2480.

Stuiver, M. and Becker, B., 1993, High-precision decadal calibration of the radiocarbon time scale: Radiocarbon, v. 35, p. 35-65.

Stuiver, M. and Braziunas, T.F., 1993, Modelling atmospheric ^{14}C influences and ^{14}C ages of marine samples to 10,000 BC: Radiocarbon, v. 35., p. 137-189.

Taylor, A., Dallimore, S. and Judge, A.S., 1996, Late Quaternary history of the Mackenzie-Beaufort region, Arctic Canada, from modelling of permafrost temperatures. 2. The Mackenzie delta–Tuktoyaktuk Coastlands: Canadian Journal of Earth Sciences, v. 33, p. 62-71.

Vincent, W. and Pienitz, R., 1996, Sensitivity of high latitude freshwater ecosystems to global change: Temperature and solar ultraviolet radiation: Geoscience Canada, v. 23, n. 4, p. 231-236.

Accepted as revised 15 December 1996.