The Nastapoka drift belt, eastern Hudson Bay: implications of a stillstand of the Quebec–Labrador ice margin in the Tyrrell Sea at 8 ka BP

Patrick Lajeunesse and Michel Allard

Abstract: During deglaciation of eastern Hudson Bay, the western margin of the Quebec-Labrador sector of the Laurentide Ice Sheet came to a stillstand about 8 ^{14}C ka BP along the Nastapoka Hills, a series of topographic highs along the bay. These hills are the northward continuation of the eastern Hudson Bay cuesta system. It left a drift belt consisting of ice-contact submarine fans along the western slopes of the hills, small frontal moraines on hilltops, and grounding-line deposits on sills between the hills. Geomorphological, sedimentary, and radiometric evidence suggest that the stillstand responsible for deposition of the Nastapoka drift belt was either entirely or partly synchronous with the deposition of the Sakami moraine farther south. There was a period when these two morainic systems marked a continuous ice margin. These stillstands occurred due to reduction of ablation at the ice margin. In the Nastapoka Hills, ablation slowed down when the ice margin was anchored on higher relief and stood at a regional break of slope that grounded the ice margin and reduced water depth at the ice terminus, therefore, putting an end to intensive calving. In eastern James Bay and southeastern Hudson Bay, stabilization of the ice margin was caused by a reequilibrium of the ice terminus after a rapid drop of water level due to the drainage of Glacial Lake Ojibway. The new data improves the resolution of the position ice margin in eastern Hudson Bay at 8 ka BP.

Résumé : Pendant la déglaciation de l'est de la Baie d'Hudson, la marge ouest du secteur Québec–Labrador de la calotte glaciaire laurentidienne s'est stabilisée vers 8 ^{14}C ka BP le long des collines Nastapoka, une série de crêtes rocheuses longeant la baie. Cette stabilisation a laissé des éventails sous-marins de contact glaciaire le long des versants ouest des collines, des petites moraines frontales sur les sommets ainsi que des dépôts de ligne d'échouage glaciaire sur les seuils entre les collines. Les données géomorphologiques, sédimentaires et radiométriques suggèrent que la pause glaciaire responsable de la mise en place de la ceinture de drift de Nastapoka était entièrement ou partiellement synchrone à la mise en place de la moraine de Sakami au sud. Il y a eu une période où ces deux systèmes morainiques constituaient une marge glaciaire continue. Ces phases de stabilisation se sont produites à cause de la réduction de l'ablation au front glaciaire. Le long des collines Nastapoka, l'ablation a été ralentie lorsque la marge glaciaire s'est accrochée à une topographie plus élevée et est restée ancré sur une rupture de pente régionale, réduisant ainsi l'ablation de la glace en diminuant la profondeur de l'eau au front glaciaire et mettant fin à la phase de vêlage intense dans la Baie d'Hudson. À l'est de la Baie de James et au sud-est de la Baie d'Hudson, la stabilisation de la marge glaciaire a été produite par le ré-équilibre du front glaciaire après une chute rapide du niveau de l'eau causée par le drainage du Lac glaciaire Ojibway. Ces nouvelles données permettent de préciser la reconstitution de la marge glaciaire et des étapes de la déglaciation à l'est de la Baie d'Hudson aux environs de 8 ka BP.

Introduction

Catastrophic deglaciation of Hudson Bay between 8.4 and 8 ^{14}C ka BP resulted in the drainage of Glacial Lake Ojibway by breaching the Laurentide Ice Sheet (LIS) into two domes over Quebec–Labrador and Keewatin (Dyke and Prest 1987). This event may have triggered the cooling detected in the Greenland Ice Sheet Project 2 (GISP2) ice core and elsewhere after 7.7 ka BP (8470 calendar years BP) through the sudden outflow of large volumes of freshwater

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P. Lajeunesse[1,2] and M. Allard. Centre d'études nordiques and Département de géographie, Université Laval, Québec, QC G1K 7P4, Canada.

[1]Corresponding author (e-mail address: patrick_lajeunesse@uqar.qc.ca).
[2]Present address: Institut des sciences de la mer de Rimouski (ISMER), Université du Québec à Rimouski, 310, allée des Ursulines, Rimouski, QC G5L 3A1, Canada.

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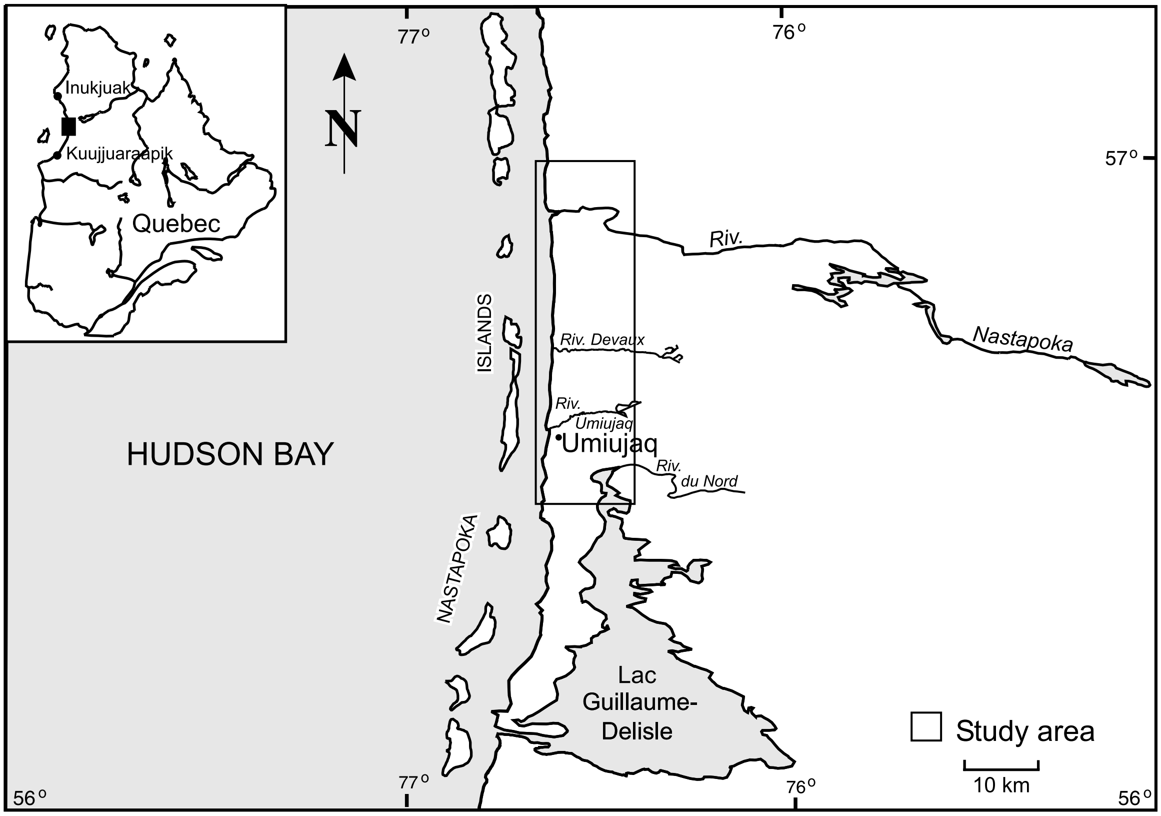
Fig. 1. Location of the study area, eastern Hudson Bay, Quebec.



into the North Atlantic (Barber et al. 1999). When the ice sheet was breached, the Tyrrell Sea inundated much of the isostatically depressed Hudson Bay basin. The abrupt change in water level at the ice margin in Lake Ojibway from near 500 m above sea level (a.s.l.) to about 280 m in the Tyrrell Sea engendered a stillstand of the ice front between Lac Mistassini and Grande-Rivière-de-la-Baleine, for the duration necessary for the reequilibration of the ice sheet profile (Hillaire-Marcel et al. 1981). Deposition at the glacier margin during this stillstand formed the Sakami moraine in eastern James Bay and southeastern Hudson Bay (Lee et al. 1960; Vincent 1974; Hardy 1977; Hillaire-Marcel et al. 1981). Up to now, except for marine geophysical surveys (Zevenhuizen 1996), little is known about deglaciation of Hudson Bay north of Grande-Rivière-de-la-Baleine and no correlative sediments and landforms were reported to support an improved reconstruction of deglaciation phases.

This paper presents results from fieldwork between Lac Guillaume-Delisle and Rivière Nastapoka (Fig. 1). Its builds upon recent sedimentary and stratigraphic investigations of raised glaciomarine deposits (Figs. 2, 3) along the Nastapoka Hills (Lajeunesse and Allard 2002) that indicate a stillstand of the ice margin of the Quebec–Labrador sector of the LIS in the Tyrrell Sea soon after the breakup of the ice sheet over Hudson Bay. The dating of this event improves the resolution of the ice-margin geometry of this sector of Hudson Bay around 8 ka BP (Dyke and Prest 1987).

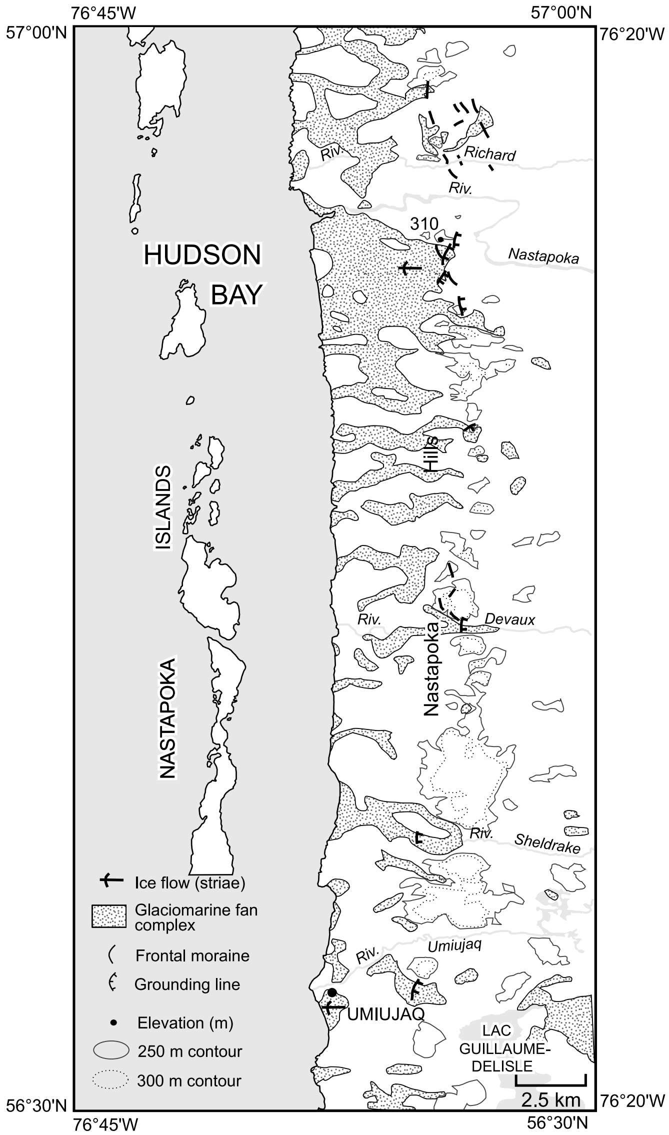
Study area

The Nastapoka Hills have a mean elevation of 280–300 m, with several summits as high as 420 m. The chain of hills extends 4–7 km inland parallel to the curved Hudson Bay coast, in Archean granite-gneiss terrain (Stevenson 1968; Avramtchev 1982). Between the hills and the shoreline, the rolling terrain has a slope of 3°–5°, which continues below sea level. The land drops abruptly east of the hills, forming a basin several kilometres wide with mean elevations < 200 m. Farther east the land rises slowly, giving way to the Canadian Shield peneplain with elevations of 200–300 m (Fig. 4). The Nastapoka Hills are the northward continuation of the Lac Guillaume-Delisle cuesta ridges. The maximum elevation of marine transgression (Tyrrell Sea) is recorded by the limit of wave-washing of ablation till. Several perched glaciofluvial deltas and the highest raised beaches are just a few metres lower than this limit. The marine limit slopes northward from Lac Guillaume-Delisle (270 m) in the south to Rivière Nastapoka (236 m) in the north (Allard and Seguin 1985). The ablation till on hilltops above the marine limit forms a thin and discontinuous veneer of boulders, sand and fines. Rates of postglacial emergence were estimated to have been 0.09 m a^{-1} during the early phase of deglaciation (Lajeunesse 2000; Allard and Seguin 1985).

Methods

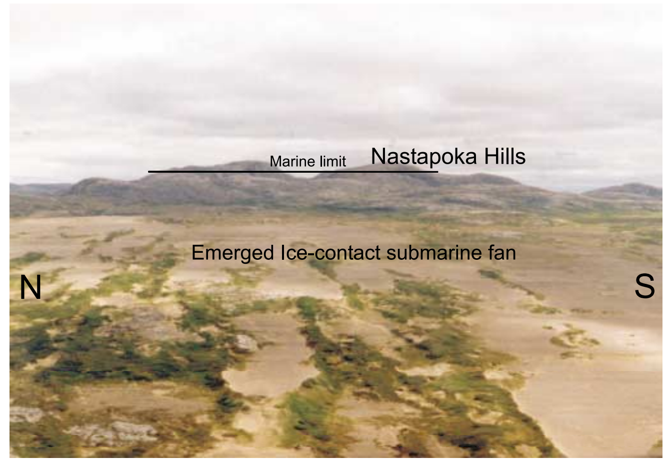
Glacial landforms and Quaternary sediments were mapped on 1 : 20 000 and 1 : 50 000 aerial photographs and during foot and helicopter traverses. Glaciomarine deposits were investigated in sections in gullies (Lajeunesse 2000; Lajeunesse and Allard 2002). Shells recovered in ice-contact, ice-proximal, emergence, and gravity-flow sediments were

Fig. 2. Ice marginal landforms and ice-contact submarine fans, Nastapoka Hills, eastern Hudson Bay.



radiocarbon dated. In total, 31 fossil shell samples from different sites and sections were dated, increasing the likelihood of obtaining the oldest samples in the study area, which yield the dates closest to deglaciation. Elevations of sections and landforms were measured with an electronic pressure-corrected altimeter (0.2 m precision), using a provincial geodetic survey bench mark as reference. We report uncorrected ^{14}C dates here, instead of calendar dates (Stuiver and Polach 1977), to be consistent with previous work (e.g., Hardy 1977). Future workers needing to calibrate our dates to the calendar time scale or planning to apply different reservoir corrections should first correct our dates using calibration curves (e.g., Stuiver and Reimer 1993). To have comparable ages among dates provided by the different laboratories (Isotrace Laboratory, University of Toronto (TO) and Université Laval, ^{14}C Laboratory (UL)), we have substracted 410 years from the TO conventional dates.

Fig. 3. Eastward view of an ice-contact submarine apron (foreground) and the Nastapoka Hills (background).



Nastapoka drift belt

Small frontal moraines lie above the Tyrrell Sea limit on hill tops and grounding-line deposits are located in sills between hills, altogether marking a nearly continuous, north–south-trending ice-front position. The morainic ridges (Fig. 5) that occur on many summits are hardly visible on the small-scale aerial photographs available for the study area. They form irregular, sinuous ridges generally < 1 m high and 3–5 m wide, consisting of gravely and boulder-rich till with a sandy matrix. In some gullies eroded between hills along the lateral contact between Quaternary sediments and bedrock faces, short exhumed segments of similar buried moraines could be seen also. Although small in size and morphologically comparable to De Geer moraines found further east on the shield (Allard and Seguin 1985), they nevertheless mark frontal positions of the glacier that can be followed from hilltop to hilltop.

Grounding lines of the ice terminus are marked by the presence of ice-contact escarpments, occasionally associated with drift bulges in some passes between hills. They are found in alignment with moraines on hilltops, and they constitute the starting points of glaciomarine fans. Inland from the grounding lines (i.e., in the direction of ice-front retreat), thick sequences of silty clay were deposited in valleys below the marine limit. This difference in sediment texture on either side of these features forms a line that is clearly visible on aerial photographs (Fig. 6*a*). In places, overdeepenings occur on the eastern sides of grounding lines (Fig. 6*b*). In Alaska, such a feature has been attributed to subglacial erosion behind the ice terminus (Powell 1981, 1984). Here, we attribute such overdeepenings to lack of sedimentation in the void left after ice retreat. Thick silts and clays located east of these scarps were deposited after retreat of the ice sheet from the Nastapoka Hills. These fine sediments are now distinguished easily by the presence of numerous cryogenic mounds that formed when permafrost aggraded after emergence (Allard et al. 1987).

The ice-contact submarine fans originate from the frontal position along the hill range and extend over the sloping terrain to the sea shore, a distance of about 5 km. They consist of

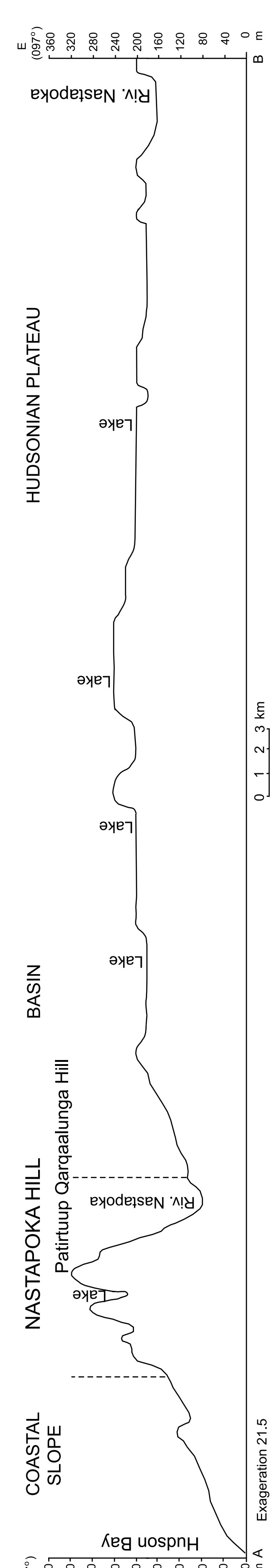
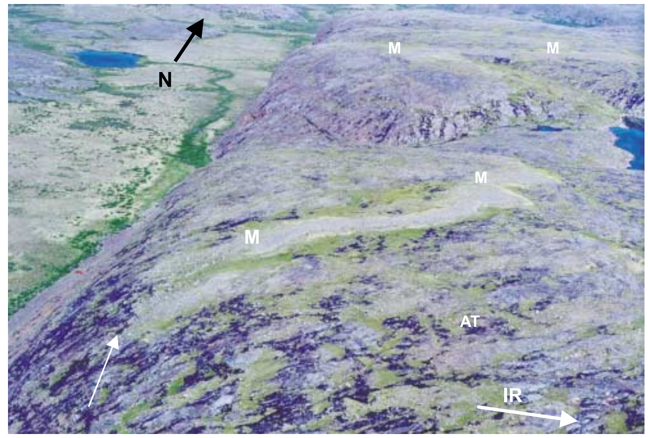


Fig. 4. Topographic profile along Rivière Nastapoka (after Seguin and Allard 1984).

Fig. 5. Frontal moraine (M) consisting of subrounded to subangular boulders with a poor sandy matrix, Nastapoka Hills near Rivière Devaux (AT, ablation till). The ice terminus retreated towards the left side of the photograph (IR, ice retreat direction).



sediments about 25 m thick on average that were deposited subaqueously by the stabilization of the ice terminus on the sea floor. Sedimentation in the fan deposits produced gravel, sand, and silt beds. The fans range in elevation between 220 and 80 m a.s.l. However, acoustic surveys off Rivière Nastapoka reveal the continuation of these fans below the sea level (Lavoie et al. 2002). Several of them are concentrated near Rivière Nastapoka, where they coalesce and form aprons. Such fans also occur in the valley of the Devaux and Umiujaq rivers. The latter is related to a grounding line in Vallée-des-Trois, against the basaltic cuesta ridge at the northern tip of Lac Guillaume-Delisle (Figs. 6*a*, 6*b*). The characteristics and depositional settings of these sediments are described more extensively in Lajeunesse and Allard (2002). Sand beds and laminations dominate these deposits but diamicts, silty sand, and silty clay also occur. Sediment delivery was dominantly by meltwater jets emanating from the ice terminus. These sediments were highly disturbed by submarine gravity flows and local ice surging (Figs. 7*a*–7*c*). Ice-rafted debris is also abundant through the sediments. The coarse fraction consists of Archean lithologies from the Canadian Shield. These ice-contact submarine fans were deposited in relatively deep water (30–150 m) when the Tyrrell Sea was at its highest level (248 m a.s.l.).

The "continuity" of moraines from one summit to the next along the Nastapoka Hills, the ice front grounding lines in the glaciomarine and marine sediments between the hills, and the associated submarine fans together constitute a complex that marks a frontal position of the ice sheet along the hills, just a few kilometres onshore of Hudson Bay. This assemblage of landforms and sediments is rather uncommon as an expression of a frontal position marking a pause in glacial retreat in comparison with known major frontal systems in continental settings. This is due to the fact that most of the deposits were laid down in a submarine proglacial environment, somewhat similar to what has been observed in fjords off active glaciers (e.g., Powell 1981, 1984; Elverhøi et al. 1983; Mackiewicz et al. 1984). Furthermore, most of these sediments were thereafter concealed underneath more recent emergence and post-emergence sediments, such as recent

Table 1. Radiocarbon ages (>6000 ^{14}C BP) from fossil shells, Nastapoka Hills area, eastern Hudson Bay.

Lab No.	Location (Lat., Long.)	Elevevation (m a.s.l.)	Age (^{14}C BP)	1σ limits (^{14}C BP)	Dated species
UL-2346	57°00′18′′N 76°30′28′′W	4	8020 ± 120	8140–7900	*Macoma balthica*
UL-1932	56°51′04′′N 76°27′00′′W	145	7730 ± 120	7850–7610	*Mya truncata*
					Hiatella arctica
					Macoma balthica
UL-1951	56°52′40′′N 76°32′15′′W	108	7360 ± 130	7490–7230	*Hiatella arctica*
TO-7411	56°51′04′′N 76°27′00′′W	154	7290 ± 70	7360–7220	*Mya truncat*
					Macoma Calcarea
UL-1740	56°55′00′′N 76°29′09′′W	115	7210 ± 90	7300–7120	*Hiatella arctica*
UL-1728	56°54′04′′N 76°29′45′′W	110	7160 ± 80	7240–7080	*Hiatella arctica*
UL-1729	56°52′56′′N 76°32′15′′W	93	7030 ± 80	7110–6950	*Hiatella arctica*
UL-1722	56°52′56′′N 76°32′15′′W	95	6920 ± 80	7000–6840	*Hiatella arctica*
UL-1937	56°52′00′′N 76°26′45′′W	132	6910 ± 80	6990–6830	*Mytilus eduli*
					Hiatella arctica
TO-7412	56°55′00′′N 76°29′15′′W	133	6740 ± 90	6830–6650	*Hiatella arctica*
					Buccinum undatum
					Balanus sp.
UQ-547[a]	56°54′00′′N 76°26′30′′W	n.d.	6700 ± 100	6800–6600	*Hiatella arctica*
					Mytilus edulis
UQ-545[a]	56°53′30′′N 76°25′30′′W	120[b]	6600 ± 100	6700–6500	*Hiatella arctica*
					Mytilus edulis
Beta-121551	56°48′12′′N 76°26′15′′W	146	6480 ± 70	6550–6410	*Hiatella arctica*
					Mytilus edulis
TO-7410	56°52′55′′N 76°29′45′′W	108	6460 ± 60	6520–6400	*Hiatella arctica*
					Mytilus edulis
Beta-121552	56°32′30′′N 76°27′30′′W	100	6430 ± 70	6500–6360	*Mytilus edulis*
					Hiatella arctica
UL-1718	56°54′04′′N 76°29′45′′W	112	6740 ± 80	6820–6660	*Hiatella arctica*
					Mytilus edulis
UL-1709	56°52′55′′N 76°29′45′′W	101	6670 ± 90	6760–6580	*Mytilus edulis*
UL-1717	56°52′55′′N 76°29′45′′W	107	6660 ± 80	6740–6580	*Hiatella arctica*
UL-1953	56°52′00′′N 76°26′15′′W	139	6640 ± 80	6720–6540	*Mytilus edulis*
UL-1711	56°52′55′′N 76°29′45′′W	109	6570 ± 70	6640–6500	*Mytilus edulis*
					Hiatella arctica
					Chlamys islandicus
UL-1928	56°51′48′′N 76°26′15′′W	146	6540 ± 80	6620–6460	*Hiatella arctica*
					Mytilus edulis
UL-1918	56°52′55′′N 76°29′45′′W	104	6510 ± 110	6620–6400	*Mytilus edulis*
					Hiatella arctica
UL-1703	56°52′55′′N 76°29′45′′W	109	6420 ± 110	6530–6310	*Mya truncata*
UL-1540	56°52′45′′N 76°29′50′′W	97	6120 ± 80	6200–6040	*Hiatella arctica*
UL-1710	56°52′55′′N 76°29′45′′W	110	6040 ± 110	6150–5930	*Mytilus edulis*
					Hiatella arctica
TO-7409	56°52′55′′N 76°29′45′′W	109	6010 ± 60	6070–5950	*Mya truncata*

Note: Laboratory designation: UQ, Université du Québec à Montréal, Quebec; UL, Laval University ^{14}C Laboratory, Québec, Quebec; CAMS, Lawrence Livermore National Laboratory, University of California, San Francisco, California; TO, Accelerator mass spectrometry at Isotrace Laboratory, University of Toronto, Toronto, Ontario; Beta, Beta Analysis Laboratory, Miami, Flordia. Lat., latitude; Long., longitude.

[a]Allard and Seguin 1985.

[b]The altitude of this sample has been corrected from the previously published data.

fluvial sediments, series of raised beach ridges, ice-rafted boulder fields, and peat soils. Also, their original surficial morphology was further obscured by younger geomorphological processes, such as dune formation and terrain deformation (uneven heave) associated with permafrost aggradation during the late Holocene (Lajeunesse 2000; Allard et al. 1987).

Chronology

Shells recovered in the Nastapoka Hills submarine fans yielded dates ranging from 8020 ± 120 (UL-2346) to 6040 ± 110 (UL-1710) BP (Table 1). Paired valves of *Macoma balthica* recovered from ice-distal silty clay at 4 m a.s.l. provided the oldest date of 8020 ± 120 BP (UL-2346) (Table 1). Paired valves of *Mya truncata*, *Hiatella arctica*, and *Macoma balthica* recovered at 145 m a.s.l. from foreset beds yielded an age of 7730 ± 120 BP (UL-1932). This sample was found at the bottom of a section but was still separated from the underlying bedrock by a few metres of sediments. Sample UL-2346 indicates that the Nastapoka Hills submarine fan deposition and ice stillstand began between 8.1 and 7.9 ka BP. It also provides an average minimal age of 8 ka BP for the 248 m marine limit. Paired valves of *Mya truncata* recovered at 154 m in a disturbed foreset bed, in the same section as UL-1932, dated 7390 ± 70 BP (TO-7411), which indicates that at least 9 m of sediments had already been deposited by 7.4 ka BP.

A sample of *Hiatella arctica* fragments collected in ice-distal silty clay located < 1 km east of the hills gave a date of 7210 ± 90 BP (UL-1740), therefore, setting a minimum age for the frontal complex. Radiocarbon ages as young as 6 ka BP (UL-1710) likely reflect the activity of mass movements on the frontal slopes the submarine fans long after deglaciation. The morphological evidence described earlier in the text suggest an ice stillstand that might have lasted from years to decades.

Another sample of shells of *Hiatella arctica* and *Mytulis edulis* recovered at 185 m a.s.l. from ice-distal silty clay yielded an age of 6700 ± 100 BP (UQ-547) on the east side of the Nastapoka Hills. Many samples of younger shells in distal parts of the contorted sandy and silty layers of the fans also indicate that, as the ice had receded from the frontal position along the hill range and the sea level was dropping, gravity flows were affecting the sedimentary sequence (Lajeunesse and Allard 2002).

Significance of the Nastapoka drift belt

Comparisons and synchronicity with the Sakami moraine

The Nastapoka drift belt is located about 175 km northeast of the northernmost Sakami deposits mapped by Vincent (1974). The Sakami moraine (Fig. 8) marks a stabilization of the QLS ice about 7.9 ka BP (Hardy 1982). It coincides with the eastern limit of Glacial Lake Ojibway. The moraine is 630 km long and forms an arcuate ridge. Although studies have been carried out in the southeastern Hudson Bay region since this moraine was mapped by Lee et al. (1960), its possible extension north of Kuujjuaraapik has remained unknown. However, marine geophysical data indicate that ice-contact deposits lying on the Hudson Bay sea floor (Zevenhuizen 1996) occur northeast of the northermost mapped sector of Sakami moraine, which suggest that they likely belong to this moraine. On land, the moraine consists mainly of glaciofluvial sand and gravel deposited both subaqueously and subaerialy and of few short till ridges oriented perpendicular to eskers and ice flow indicators (Hardy 1982). These deposits are discontinuous and are located between 100 and 380 m a.s.l. Glaciofluvial sediments mainly consist of well-sorted sand and gravel up to 30–50 m thick. A maximal thickness of 80 m has been observed in boreholes. Some silt layers occur through the sandy sequences, mainly at lower elevations. Many submarine fans are attached to the moraine (Hardy 1982). Along the axis of the moraine, marine limit reaches 290 m a.s.l. near Kuujjuaraapik but declines southward to 198 m a.s.l. (Hardy 1977). Marine mollucs from glaciomarine beds on the distal side of the Sakami moraine yielded radiocarbon dates of 7880 ± 160 (QU-122) and 7550 ± 180 BP (QU-124) (Table 2). The approximate beginning of deposition of the moraine was thus between 7.9 and 7.7 ka BP. Another radiocarbon date of 7290 ± 90 BP (GSC-2239) (Table 2) on shells from 13 km east of the moraine indicates that the moraine formed in less than 400 years (Hardy 1982). Based solely on calculations of the equilibrium profile of the ice sheet, Hillaire-Marcel et al. (1981) estimated the duration of this stillstand at about 20 years.

The Nastapoka drift belt also contains both subaerial (above marine limit) and submarine proglacial sediments. Although glaciofluvial sediments of the Sakami moraine (Hardy 1977, 1982; Vincent and Hardy 1977) are somewhat different from the glaciomarine sediments in the Nastapoka Hills (Lajeunesse and Allard 2002), presence of ice-contact submarine fans in both regions indicate as well some similarities in sedimentation dynamics. The dating suggest that the Sakami moraine and the Nastapoka drift belt overlap in time; in fact they are practically synchronous. But as the dating of the Sakami moraine serves as the dating of the rapid breaching of the LIS, the catastrophic drainage of Lake Ojibway, and the invasion of the Tyrrell Sea, what significance does the Nastapoka drift belt takes in this fast sequence of events? The northernmost extension of Lake Objibway immediately before its drainage is a key question.

Possible links with the drainage of Glacial Lake Ojibway

Glaciolacustrine deposits are found on the western side of the Sakami moraine (Hardy 1977, 1982; Vincent and Hardy 1977; Zevenhuizen 1996). On the southeastern Hudson Bay sea floor between Kuujjuaraapik and Manitounuk Sound, one acoustic unit has been defined as either glaciolacustrine or glaciomarine in origin (Josenhans et al. 1988; Bilodeau et al. 1990; Gonthier et al. 1993; Zevenhuizen 1996; Hill et al. 1999). Glaciolacustrine varves have also been observed in one section along Rivière Domanchin, near Manitounuk Sound (Parent and Paradis 1994). Distribution of these varves over this extensive area indicate that the glacial lake extended to the north of Grande-Rivière-de-la-Baleine. However, no evidence of glaciolacustrine sediments was found in sections in the Rivière Nastapoka area. Acoustic surveys off Rivière Nastapoka revealed transparent layered sediments with the same signature as either glaciolacustrine or glaciomarine sediments (Lavoie et al. 2002). The sections on land of what appear the lateral equivalent of the offshore acoustic units belong to the shell-bearing sequence of glaciomarine sediments. Both types of sediments are indistinguishable acoustically and only one short core off Grande-Rivière-de-la-Baleine is reported to contain a basal glaciolacustrine unit (Bilodeau et al. 1990). Hence, no reliable evidence is currently available to assess whether Glacial Lake Ojibway reached the latitude of Rivière Nastapoka before breaching of the ice sheet. If Lake Ojibway ever reached this sector before it drained through Winisk Trough in central Hudson Bay (Josenhans and Zevenhuizen 1990), the glaciolacustrine sediment layer is likely very thin, since this was brief event.

Mechanisms of the Nastapoka Hills stillstand

The high marine limit in the sector (248–230 m) indicates

Table 2. Selected radiocarbon dates for deglaciation of eastern Hudson Bay and James Bay.

	Location	Age (^{14}C BP)	1σ limits (^{14}C BP)	Dated species	Reference
GSC-327	Ivujivik	7350 ± 150	7500–7200	*Mya truncata*	Matthews (1966, 1967)
GX-1070	Mansel Island	7115 ± 100	7215–7015	Undifferentiated shells	Wagner (1967)
GSC-706	Ottawa Islands	7430 ± 180	7610–7250	*Mya truncata* *Hiatella arctica*	Andrews and Falconer (1969)
UQ-956	Akulivik	7700 ± 140	7840–7560	*Hiatella arctica*	Lauriol and Chamberland (unpublished)[a]
UQ-834	Povungnituk	6810 ± 250	7060–6560	*Serripes groenlandicus*	Lauriol and Chamberland (unpublished)[a]
UQ-761	Cap Smith	8040 ± 110	8150–7930	*Hiatella arctica* *Mytilus edulis*	Lauriol and Chamberland (unpublished)[a]
GSC-4332	Cap Smith	6850 ± 110	6960–6740	*Hiatella arctica* *Mytilus edulis*	Redated sample UQ-761
UL-2346	Rivière Nastapoka	8020 ± 120	8140–7900	*Macoma balthica*	Lajeunesse (2000)
UL-1932	Rivière Nastapoka	7730 ± 120	7850–7610	*Mya truncata* *Hiatella arctica* *Macoma balthica*	Lajeunesse (2000)
TO-7411	Rivière Nastapoka	7290 ± 70	7370–7230	*Mya truncata* *Macoma calcarea*	Lajeunesse (2000)
QU-280	Petite-Rivière-de-l a-Baleine	7820 ± 100	7920–7720	Calc. concretions	Hillaire-Marcel (1976)
I-8363	Kuujjuaraapik	8230 ± 135	8365–8095	Calc. concretions	Hillaire-Marcel (1976)
QU-281	Kuujjuaraapik	7940 ± 140	8080–7800	Calc. concretions	Hillaire-Marcel (1976)
I-9005	Kuujjuaraapik	7625 ± 120	7745–7505	*Macoma Balthica*	Hillaire-Marcel (1976)
QU-122	Sakami moraine	7880 ± 160	8040–7720	Undifferentiated shells	Hardy (1977)
QU-124	Sakami moraine	7750 ± 180	8030–7570	Undifferentiated shells	Hardy (1977)
QU-258	Rivière Eastmain	7440 ± 210	7650–7230	Undifferentiated shells	Hardy (1977)
QU-369	Rivière Eastmain	7370 ± 100	7470–7270	Undifferentiated shells	Hardy (1977)
GSC-2239	13 km east of Sakami moraine	7290 ± 90	7380–7200	Undifferentiated shells	Hardy (1977)

Note: Dates are plotted on Fig. 8. Laboratory designation: UQ, Université du Québec à Montréal, Quebec; GSC, Geological Survey of Canada, Ottawa, Ontario; UL, Laval University ^{14}C Laboratory, Québec, Quebec; TO, Accelerator mass spectrometry at Isotrace Laboratory, University of Toronto, Toronto, Ontario; I, Teledyne Isotopes, Westwood, New Jersey; GX, Geochrone Laboratories, Cambridge, Massachusetts. Calc., calcareous.

[a] *In* Gray and Lauriol (1985).

that the Nastapoka Hills stillstand occurred during the early phase of marine trangression of the Tyrrell Sea. Apparently, the Nastapoka Hills drift belt is anchored to the higher relief that acted as pinning points for the previously fast retreating ice sheet in Tyrrell Sea. Presence of the hills also slowed down the rate of ice disintegration by reducing water depths near the ice terminus (Meier and Post 1987; Lajeunesse and Allard 2002). Therefore, in the Nastapoka Hills, where no glaciolacustrine deposits have been observed, the halt in glacial retreat was more likely caused by the presence of a hill range and a regional break of slope that reduced ice ablation by pinning down the ice sheet and by reducing water depths at the ice terminus. In fact this regional break of slope marks the transition between the deeper Tyrrell Sea basin at the time of marine maximum, where ice retreat was dominated by iceberg calving, and the general higher elevation of the continent, where ice became grounded and meltwaters dominated the waning glacial system (as indicated by eskers and sandur-deltas further inland, (Allard and Seguin 1985)), a process change reported in Alaska by Meier and Post (1987). Unless unambiguous evidence of the presence of Glacial Lake Ojibway at the latitude of Rivière Nastapoka is found, it appears most likely that the Nastapoka Hills stillstand occurred after a short period of intensive iceberg calving in Hudson Bay during maximum relative sea level (RSL). Such a scenario would imply that, although the radiocarbon dates of the Nastapoka drift belt and the Sakami moraine are contemporaneous, both systems do not constitute a single one. Sakami moraine would have begun to form first following drainage of Lake Ojibway to allow ice-sheet profile reequilibration. As the breaching probably occurred in the centre of the Hudson Bay basin, the ice front position then must have been far off the Nastapoka coastal area. But, thereafter the glacial margin must have retreated very fast through iceberg calving, probably in a matter of a century at the most, until the halt on higher ground along the coastal chain of hills.

Geometry of the ice margin along eastern Hudson bay

The location of the Nastapoka Hills drift belt suggest a continuation of the arc-shaped Sakami moraine north of Kuujjuaraapik. The Sakami moraine gently curves towards the northeast near the mouth of Grande-Rivière-de-la-Baleine. This suggests that the direction of the ice-margin position

Fig. 6. (*a*) Aerial photograph showing an overdeepening marking the grounding line (arrows) of the ice terminus in Vallée-des-Trois during the Nastapoka Hills stillstand. The left side (west) of the ridge consists of sand and gravel and the right side (east) consists of silty clay affected by cryogenic mounds (CM); (*b*) Horizontal aerial view of the same grounding line (arrow) shown in (*a*). Note the abrupt change in sediment texture, passing from sand to the left (west) to silty clay to the right (east).

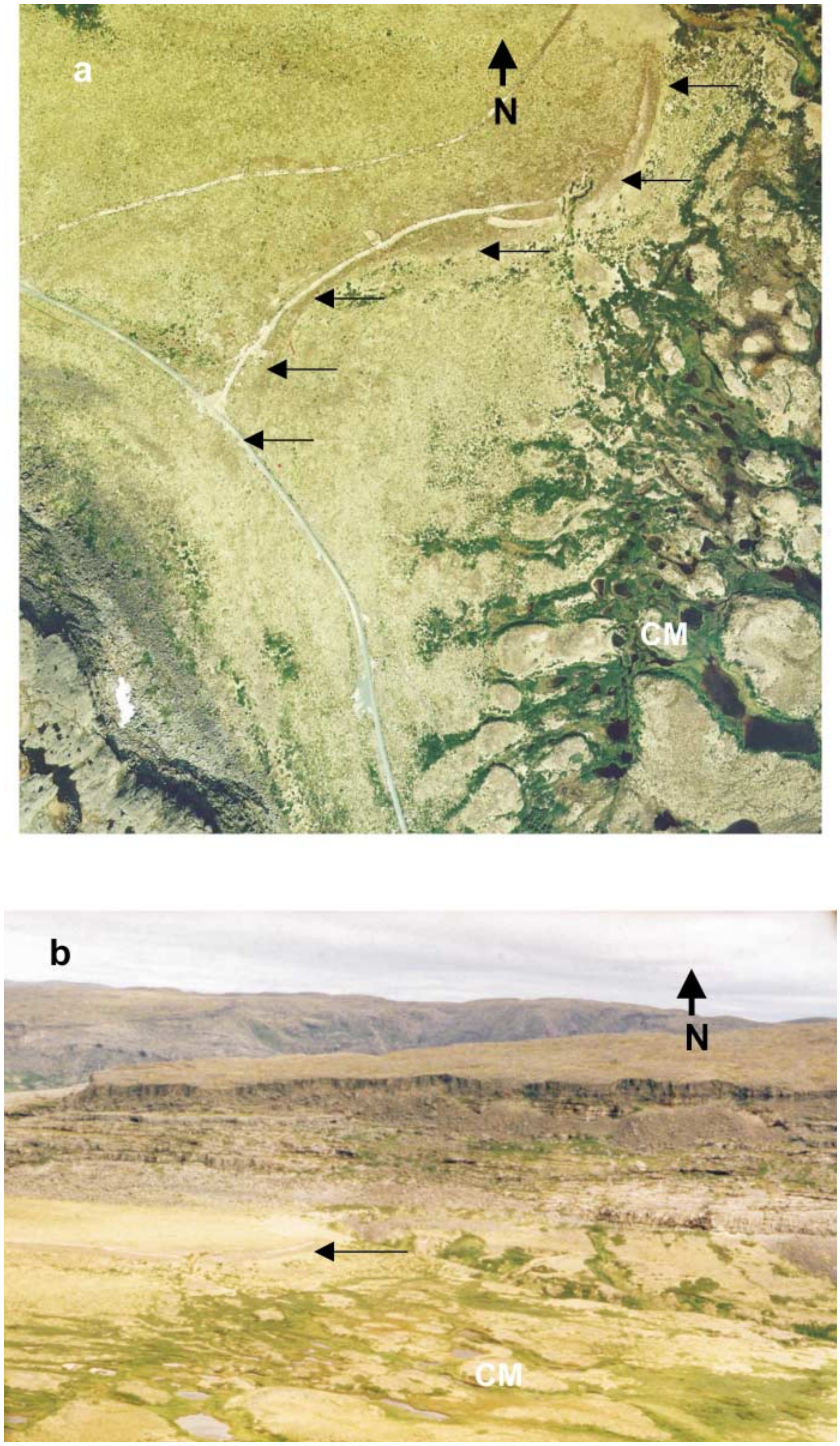
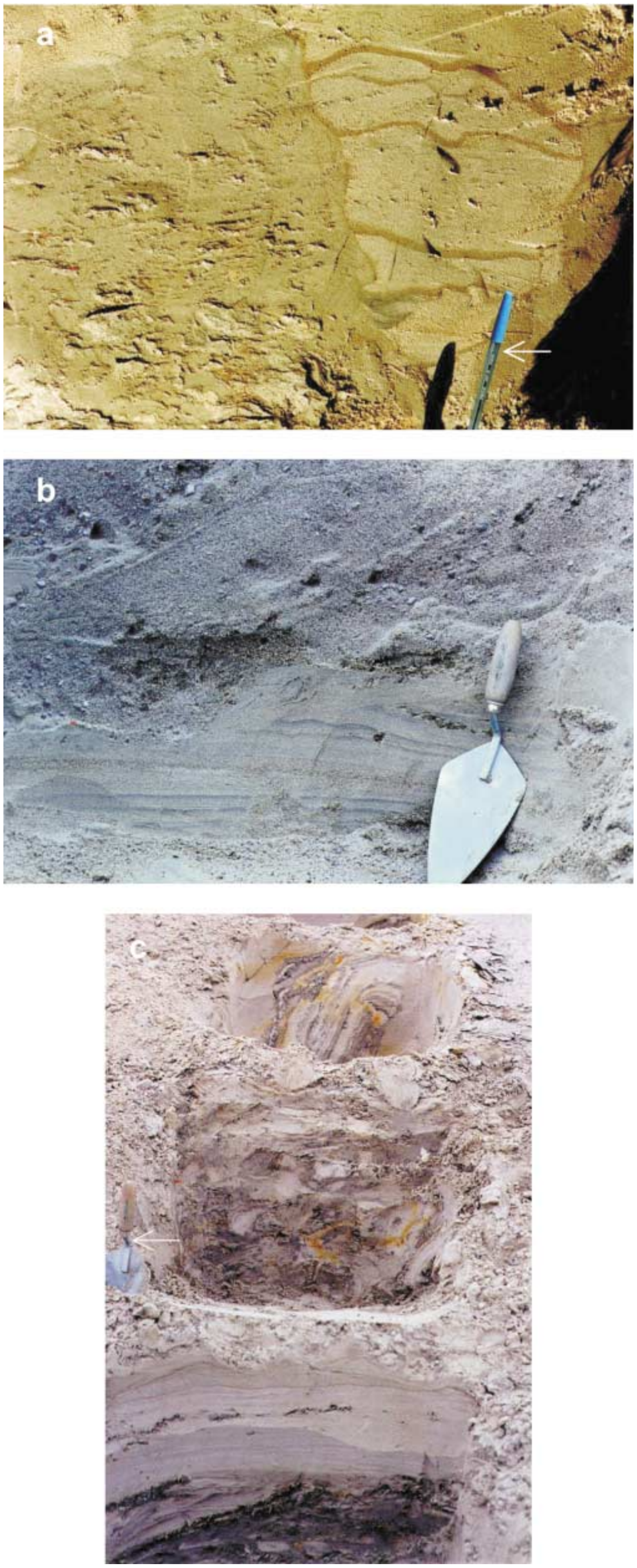


Fig. 7. Sediment deformations in the Nastapoka Hills glaciomarine fan deposits: (*a*) a water escape structure in massive sand (arrow pointed at pen for scale); (*b*) massive sand–gravel bed with erosive lower contact deposited by a debris flow over stratified sand (trowel for scale); (*c*) chaotic bedding in interbedded sand, silt and silty clay (arrow pointing at trowel for scale).



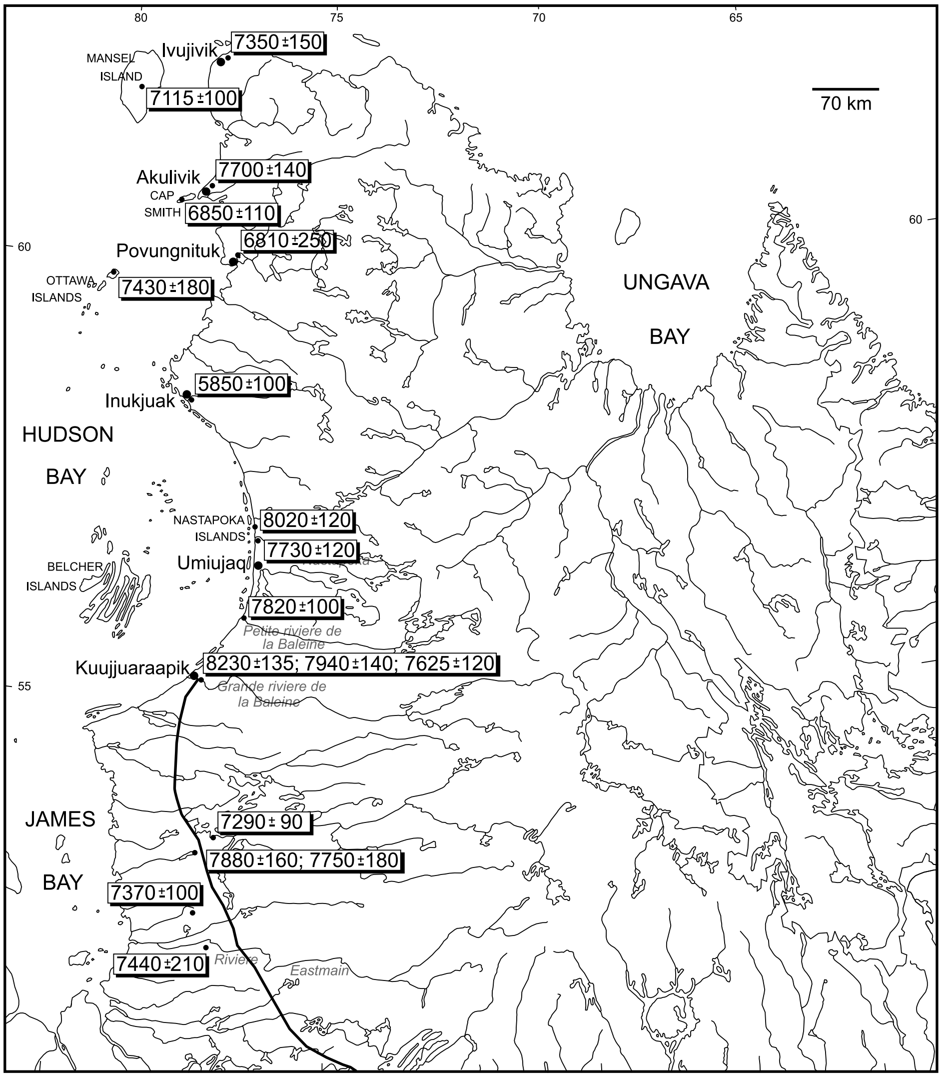
about 8 ka BP at the latitude of Rivière Nastapoka was more likely to continue towards the northeast and follow the actual coastline rather than run directly east of the Belcher Islands, such as reported in previous models (Fig. 9*a*; e.g., Dyke and Prest 1987). Moreover, Zevenhuizen (1996) observed moraines and ice-contact deposits of possible glaciofluvial origin on the southeastern Hudson Bay sea floor, immediately west of Manitounuk Sound (Fig. 9*b*). The location and position of these sediments fit with the general trend and geometry of the Sakami moraine. He suggested that the presence of these deposits in the area could represent the development of a calving bay in southeastern Hudson Bay associated with the Sakami position. However, both the geomorphological evidence and the radiocarbon date of 8020 BP (UL-2346) in the Nastapoka Hills area now indicate that the ice margin reached by that time a position further northeast and was parallel to the coastline, where higher relief occurred.

That 8 ka age suggests that the Nastapoka Hills stillstand was contemporaneous with the deposition of the Sakami moraine. Also, the standard error of shell sample UL-1932 (7730 ± 120 BP; Table 1) overlaps the dates previously associated with the formation of the Sakami moraine in eastern James Bay (7880 ± 160, QU-122; 7750 ± 180, QU-124) by Hardy (1977) (Table 2). These dates suggest that the Nastapoka Hills stillstand share a common span with the duration of the Sakami stillstand. The new data imply a major, but likely short-lived, synchronous event of ice margin stabilization along southeastern Hudson Bay. Samples UL-2346 and UL-1932 also provide new evidence that the area located west of the Nastapoka Hills was deglaciated earlier than reported. The oldest date on shells previously reported in the region was 6700 ± 100 BP (UQ-547; Allard and Seguin 1985). The new radiocarbon ages presented here indicate that the region was deglaciated at least 1000 years earlier. In fact, sample UQ-547 was collected on the eastern slope of the Nastapoka Hills, where ice retreated later.

Previous models have configured the position of the QLS margin in Hudson Bay about 8 ka BP, connecting deglaciation chronological points from slightly east of Kuujjuaraapik (Hillaire-Marcel 1976, 1979) to west of the Ottawa Islands. The new dates associated with the Nastapoka drift belt demonstrate that in fact a very large portion of southeastern Hudson Bay was deglaciated somewhat earlier than suggested by Dyke and Prest (1987). Shells recovered near Akulivik dated 7700 ± 140 BP (UQ-956; Gray and Lauriol 1985; Fig. 5). The deglaciation of the Rivière Nastapoka area also appears older than on the Ottawa Islands, where a date of 7430 ± 180 BP (GSC-706) was reported (Andrews and Falconer 1969). However, the liability of ^{14}C ages in eastern Hudson Bay is likely not optimal, considering the provenance of ages from different laboratories. For example, at Cap Smith, shell sample UQ-761 (8040 ± 110 BP) was re-dated and provided an age of 6850 ± 110 BP (GSC-4332). Also, it is possible that the chronology of the marine shells presented here have been aged by a reservoir effect higher than the mean ocean effect.

Available evidence in eastern Hudson Bay now indicate that the position of the 8 ka BP margin of the QLS extended in an arc-like shape from Lake Mistassini, where the southernmost geomorphological features of the Sakami moraine are found (Hardy 1976) and extended to Manitounuk Strait

Fig. 8. Sakami moraine (after Hardy 1977) and selected radiocarbon dates for deglaciation of eastern James Bay and Hudson Bay.



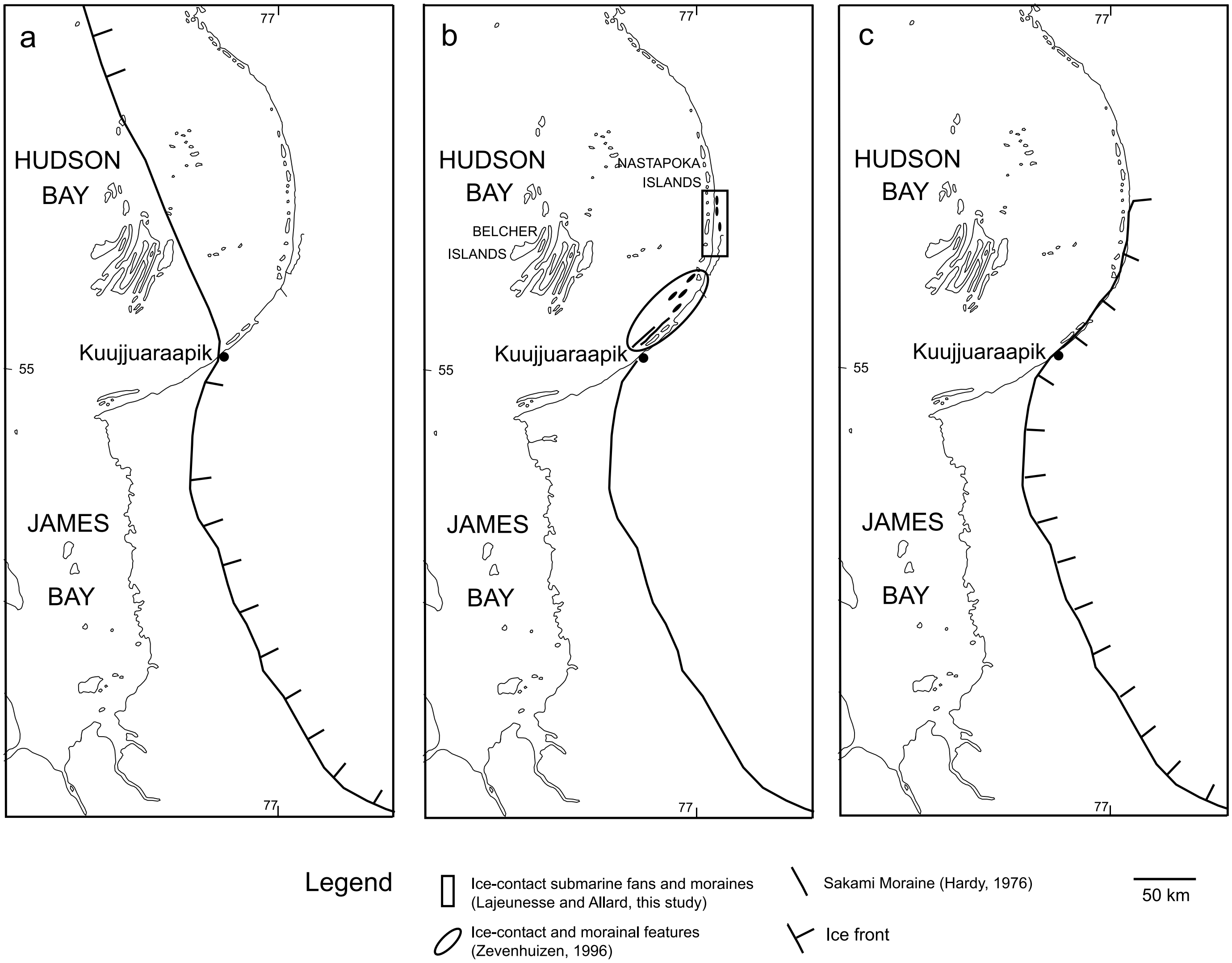
and Petite-Rivière-de-la-Baleine, where contemporaneous submarine ice-contact sediments and landforms were identified (Zevenhuizen 1996; Hill et al. 1999). From there, the margin slightly curved northward to the Nastapoka Hills, where moraines and emerged ice-contact submarine fans occur.

It still remains speculative to trace the 8 ka BP position of the western margin of the QLS in eastern Hudson Bay north of Rivière Nastapoka. However, we argue that it is unlikely that it actually adopted the geometry proposed in the Dyke and Prest (1987) model (Fig. 9*a*) because no major bathymetric restriction exist in this sector of central Hudson Bay. Presence of such a physical obstacle could have allowed a stabilization of the ice margin in the open marine environment by reducing water depth, thus ice ablation. In fact, the deep-water conditions prevailing at that moment at the ice margin in the opened sea enhanced the rates of calving and likely reduced the effective weight of the ice, reducing basal drag and allowing a faster ice transfer (Siegert et al. 2002) from central Quebec–Labrador to central Hudson Bay.

Conclusion

Retreat of the western margin of the Quebec–Labrador sector of the Laurentide Ice Sheet in the Tyrrell Sea was interrupted at about 8 ka ^{14}C BP along the high cuesta ridges and the Nastapoka Hills on the coast of Hudson Bay. This stillstand was either synchronous or slightly out of phase with the deposition of the Sakami moraine. In eastern James Bay and southeastern Hudson Bay, the QLS stillstand was caused by a reequilibration of the ice terminus due to a fall in water level related to the sudden drainage of Glacial Lake Ojibway. In the Rivière Nastapoka area, the stillstand was caused by the presence of a hill range and a regional slope break that acted by grounding the ice sheet and reducing water depths at the ice terminus. However, if Glacial Lake Ojibway ever reached this latitude before it breached through the LIS over Hudson Bay, it is likely that the stillstand could have been caused by the doubling of these two precedent mechanisms. In that case, the Nastapoka and Sakami

Fig. 9. (*a*) Previous interpreted position of the Québec-Labrador sector margin in eastern James Bay and southeastern Hudson Bay at 8000–7900 ^{14}C BP (Dyke and Prest 1987). (*b*) Ice marginal features in eastern James Bay and southeastern Hudson Bay. (*c*) New proposed position of the ice margin in eastern James Bay and southeastern Hudson Bay at 8000 ^{14}C BP.



stillstands would have begun at the exact same moment. In the opposite case, the Nastapoka stillstand would have shortly followed the Sakami stillstand and the two systems would have been practically contemporaneous (within the dating variability of various sites, shell species, and laboratory errors).

The Nastapoka drift belt and Sakami moraine give importance to this stabilization phase of the ice margin about 8 ka BP after a period of extremely rapid collapse and downwasting over Hudson Bay and James Bay. This morainal system constitutes a large-scale ice-marginal geomorphological feature in western Quebec that extended over at least 830 km.

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