Holocene emergence and shoreline delevelling, southern Eureka Sound, High Arctic Canada
L’émersion des terres et l’évolution du littoral à l’Holocène, dans la partie sud de l’Eureka Sound, Haut Arctique canadien
Auftauchen und Entwicklung der Küsten, südlicher Eureka Sound, kanadische Hocharktis

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Résumé de l’article
Cet article présente une reconstitution des changements du niveau marin relatif et de ces conséquences sur le littoral au postglaciaire. Les deltas marins soulevés, les plages et les limites de l’érosion par les vagues qui datent du début de l’Holocène attestent d’une émersion des terres de plus de 150 m dans la région. La limite marine est métachrone et s’est établie successivement au retrait glaciaire. Des contrastes forts dans les courbes du niveau marin relatif et le taux d’émersion initial ont été enregistrés dans la région à l’étude. Dans le Blind Fiord, le niveau marin relatif s’est abaissé de façon continue après la déglaciation. L’émersion initiale s’élevait à ≥ 5 m/siècle. Ce taux contrasta avec les courbes de Starfish Bay et de Irene Bay, où le taux d’émersion initial était de ≤ 1 m/siècle. Les iso-bases tracées sur le littoral de 8,5 ka au Eureka Sound montrent qu’une zone d’émersion plus élevée (≥ 130 m) s’étend le long du chenal et se termine dans le voisinage de l’entrée de la Norwegian Bay. Ce modèle démontre qu’il y a eu une charge glaciaire distincte au-dessus du Eureka Sound durant le dernier maximum glaciaire, ce qui a aussi été démontré par l’entremise d’autres éléments de preuves d’ordre géologique.
HOLOCENE EMERGENCE AND SHORELINE DELEVELLING, SOUTHERN EUREKA SOUND, HIGH ARCTIC CANADA

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ABSTRACT This paper is a reconstruction of postglacial relative sea level change and shoreline delevelling in southern Eureka Sound, High Arctic Canada. Postglacial emergence of up to 150 m is recorded in this area by raised marine deltas, beaches and washing limits that date from the early Holocene. Marine limit is metachronous and formed successively with glacier retreat. Marked contrasts in the form of relative sea level curves and rate of initial emergence are recorded from the study area. In Blind Fiord, relative sea level fell continuously following deglaciation. Initial emergence was characterised by rates of ≥5 m/century. This contrasts with curves from Starfish and Irene bays, where the rate of initial emergence was ≤1 m/century. Isobases drawn on the 8.5 ka shoreline for greater Eureka Sound demonstrate that a cell of highest emergence (≥130 m asl) extends along the length of the channel, and closes in the vicinity of the entrance to Norwegian Bay. This pattern confirms a distinct loading centre over Eureka Sound during the Last Glacial Maximum, and is compatible with independent glacial geological evidence indicating that the thickest ice was centred over the channel during the Late Wisconsinan.

RÉSUMÉ L’émersion des terres et l’évolution du littoral à l’Holocène, dans la partie sud de l’Eureka Sound, Haut Arctique canadien. Cet article présente une reconstitution des changements du niveau marin relatif et de ces conséquences sur le littoral au postglaciaire. Les deltas marins soulevés, les plages et les limites de l’érosion par les vagues qui datent du début de l’Holocène attestent d’une émer- sion des terres de plus de 150 m dans la région. La limite marine est métachrone et s’est établie successivement au retrait gla- ciaire. Des contrastes forts dans les courbes du niveau marin relatif et le taux d’émer- sion initial ont été enregistrés dans la région à l’étude. Dans le Blind Fiord, le niveau marin relatif s’est abaissé de façon continue après la déglaciation. L’émersion initiale s’élevait à ≥ 5 m/siècle. Ce taux contraste avec les courbes de Starfish Bay et de Irene Bay, où le taux d’émergence initial a été enregistré dans la région à l’étude. Dans le Blind Fiord, le niveau marin relatif s’est abaissé de façon continue après la déglaciation. L’émersion initiale s’élevait à ≥5 m/siècle. Ce taux contraste avec les courbes de Starfish Bay et de Irene Bay, où le taux d’émergence initial était de ≤1 m/siècle. Les isobases tracées sur le littoral de 8.5 ka au Eureka Sound montrent qu’une zone d’émersion plus élevée (≥130 m) s’étend le long du chenal et se termine dans le voisinage de l’entrée de la Norwegian Bay. Ce modèle démontre qu’il y a eu une charge glaciaire distincte au-dessus du Eureka Sound durant le dernier maximum gla- ciaire, ce qui a aussi été démontré par l’entre- mise d’autres éléments de preuves d’ordre géologique.


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INTRODUCTION

This paper discusses the postglacial relative sea level history of southern Eureka Sound, High Arctic Canada (Fig. 1), focusing on initial emergence, pattern of shoreline deleveling and implications for former glacier loading. Blake (1970) proposed the existence of the Inuit Nunivak Ice Sheet in the Canadian High Arctic during the Late Wisconsinan on the basis of the pattern of differential postglacial rebound since 5 ka BP. He demonstrated that shorelines of this age in the Queen Elizabeth Islands were highest (>25 m asl) throughout a broad northeast/southwest oriented corridor extending from northern Eureka Sound to Bathurst Island, and he proposed that this emergence reflected a regional ice sheet over the Queen Elizabeth Islands during the Last Glacial Maximum (cf. Walcott, 1972; Tushingham, 1991). In contrast, a markedly different reconstruction, of a restricted Late Wisconsinan glaciation for the same region, was proposed by England (1976a and b) on the basis of glacial geologic data from northeastern Ellesmere Island and an alternative interpretation of the postglacial emergence. These contrasting reconstructions formed the end members in an ensuing debate concerning the extent of ice during the Last Glacial Maximum in the Queen Elizabeth Islands (e.g., Blake, 1992a and b; Blake et al., 1992; de Freitas, 1990; Tushingham, 1991; England, 1987, 1990, 1996, England et al., 1991; Lemmen, 1989; Bell, 1996).


Isobases drawn by England (1992, 1997) on the 8 ka shoreline in Greely Fiord, Ellesmere Island (Fig. 1), exhibit a narrow plunging ridge of maximum emergence, parallel to the regional geological structure. England contrasted this isobase pattern with the broad cells of uplift documented elsewhere in Arctic Canada which record postglacial unloading following removal of the Laurentide Ice Sheet (cf. Andrews, 1970; Dyke, 1984). He also argued that the isobase ridge was difficult to reconcile with a soley glacio-isostatic explanation, as it did not conform to the distribution of the last ice load in the region which was inferred to be restricted (England, 1987, 1990; Bell, 1992, 1996). He therefore proposed a possible neotectonic contribution to Holocene emergence for western Ellesmere Island (England, 1992, 1997).

Both the isobase pattern and slow rate of initial emergence, coupled to independent glacial geological evidence suggesting a restricted ice-cover during the Last Glacial Maximum, were interpreted as being incompatible with a Late Wisconsinan Inuit Nunivak Ice Sheet (England, 1983, 1992, 1997; Bell, 1996). However, more recent fieldwork validates an extensive Late Wisconsinan glacier cover for at least the eastern and southeastern sectors of the Queen Elizabeth Islands (Ellesmere, Axel Heiberg and Devon islands) (Hättestrand and Stroeven, 1996; Bednarski, 1998; England, 1998, 1999; Ó Cofaigh et al., 1998; Ó Cofaigh, 1999; Dyke, 1999; see also Funder, 1989; Funder and Hansen, 1996). Integration of this new glacial geologic evidence with the associated postglacial relative sea level histories of these areas has only recently commenced (Dyke, 1998; England and Ó Cofaigh, 1998).

Glacial geologic and chronologic evidence indicates that Late Wisconsinan glaciation in southern Eureka Sound was characterised by an extensive ice cover (Ó Cofaigh, 1998, 1999; Ó Cofaigh et al., 1998, in press). This paper presents the postglacial relative sea level history associated with removal of that ice load, and it provides new data on the magnitude, timing and pattern of postglacial emergence (Fig. 2). It has three principal objectives: (1) to reconstruct initial postglacial emergence at several sites where the best chronological control is available; (2) to reconstruct the pattern of shoreline deleveling in southern Eureka Sound and to link this data with previously published work to the north and south (England, 1976b, 1992; Bell, 1996; Dyke, 1998, 1999); and (3) to assess implications for former ice sheet loading in the region.

STUDY AREA

Eureka Sound is the inter-island channel, 300 km long and 10-28 km wide, which separates Ellesmere and Axel Heiberg islands (Figs. 1 and 2). Geologically, the study area is dominated by north-northeast striking sedimentary rocks, although igneous rocks outcrop locally (Trettin, 1991). The geological structure dictates a topographic grain of ridges and valleys. Uplands reaching >1000 m asl are dissected by steep-sided fiords and valleys aligned both parallel to bedrock structure (e.g., Blind Fiord) and cross-cutting it (e.g., Bay Fiord). Contemporary glaciers are limited to small, upland ice-caps, although the region is bordered immediately to the east and west by extensive icefields on central Axel Heiberg and Ellesmere islands (Fig. 2).

LATE WISCONSINIAN GLACIATION OF SOUTHERN EUREKA SOUND

During the Late Wisconsinan, southern Eureka Sound supported extensive glaciation, consisting of expanded ice-caps which were coalescent along the length of the channel. Ice-divides were located along the highlands of central Ellesmere and Axel Heiberg islands, from which ice flowed east and west into Eureka Sound, with development of preferential flow along the axes of major fiords (Ó Cofaigh, 1998, 1999; Ó Cofaigh et al., 1998, in press). In Eureka Sound, trunk ice flowed north towards Nansen Sound (cf. Fyles, in Jenness 1962; Bell, 1992; Bednarski, 1998) and south towards Norwegian Bay. Raanes Peninsula supported a local ice-dome which was coalescent with trunk ice in Eureka Sound. Deglaciation of southern Eureka Sound commenced ≥9.2 ka BP.
[9.9 ka calendar years BP] (Ó Cofaigh, 1998, 1999) and was characterised by initial break-up of ice in the channel with subsequent retreat east and west to the former ice-divides. Thus marine limit throughout the study area is time-transgressive and records sequential entry of the sea with ice retreat.

**METHODOLOGY**

**SURVEYING TECHNIQUE AND DEFINITION OF MARINE LIMIT**

The altitude of raised marine features was determined using a Wallace and Tiernan micro-altimeter (accuracy ±2 m). Readings were corrected for fluctuations in atmospheric pressure and site specific temperature. High tide level, commonly demarcated by a well-defined kelp line, was used as the reference datum for sea level. Radiocarbon dates on marine shells and driftwood provide chronological control on the establishment of marine limit and subsequent emergence.

Marine limit is the maximum elevation attained by the sea along a glacioisostatically depressed coastline. Its elevation at a site reflects distance from the former ice margin (which is an indication of ice thickness over the site), date of deglaciation and eustatic sea level rise (Andrews, 1970). Throughout the study area, marine limit was taken to be either: (1) the highest raised marine delta or beach; or (2) the lowest undisturbed till or felsenmeer (washing limits) as commonly marked by a notch cut in till with a well sorted sediment veneer or bedrock below, or by an abrupt textural transition between poorly sorted till/felsenmeer and sorted sediment below; or (3) the highest elevation at which well preserved marine shells were found, which provides a minimum estimate on marine limit.

**MARINE LIMIT: ELEVATION AND PATTERN**

The highest marine limit observed in the study area occurs on the north coast of Stor Island at 145-151 m asl (Fig. 3). Marine limits of >140 m asl also occur along the south coast of Raanes Peninsula between Eureka Sound and Trolf Fiord (Fig. 3). Blind Fiord, Trolf Fiord, Starfish Bay and Jaeger Bay all exhibit a progressive decline in marine limit from mouth to head. For example, in Trolf Fiord, marine limit falls from 143 m asl at the mouth to 98 m asl at the head, whereas in Starfish Bay marine limit decreases from 113 m to 80 m asl.

Along Eureka Sound, north of Hare Bay, marine limit is recorded by deltas at the mouths of several valleys. In inner Trapper’s Cove, ice-contact deltas grade to relative sea levels at 118-120 m asl (Fig. 3). This contrasts with outer Trapper’s Cove and the Eureka Sound coast, where marine limit is defined by poorly-preserved raised beaches at 83 m asl (minimum), and deltas immediately to the south at 99 m asl (Fig. 3). The north coast of Raanes Peninsula is characterised by variable marine limit elevations which range from 76 to 120 m asl in Eureka Sound and central Bay Fiord, respectively, before...
falling progressively to 65-87 m asl at the heads of Strathcona Fiord and Irene Bay (Fig. 3). Regionally, therefore, marine limit exhibits an overall decrease in elevation eastwards from Eureka Sound to the fiord heads. However, this decrease is variable over short distances, a pattern inferred to reflect the metachronous age of marine limit occasioned by ice retreat.

RELATIVE SEA LEVEL CURVES

Emergence data are presented for three sites, Blind Fiord, Starfish Bay and Irene Bay (Fig. 4). At each site, the elevation and age of radiocarbon-dated samples and their associated relative sea levels are given. Figure 5 shows relative sea level curves for the three sites. All radiocarbon dates (including calibrated ages, reported as “cal BP”) are listed in Table I.

The three sites were inundated by ice during the Last Glacial Maximum (Ó Cofaigh, 1998, 1999). Blind Fiord was fed by ice from a local dome centred over Raanes Peninsula. This local ice-dome was coalescent with westerly-flowing, regional ice emanating from a divide under the present-day Prince of Wales Icefield. Starfish and Irene bays were infilled by this regional ice.

BLIND FIORD

At the mouth of Blind Fiord, marine limit is defined by the uppermost raised beach at 138 m asl. A fragment of Mya truncata collected from a beach surface at 128 m asl yielded an Accelerator Mass Spectrometry (AMS) date of 8590±70 BP [9460-9060 cal BP] (TO-5862; Site 1, Fig. 4 and Table I), and provides a minimum age estimate on the 138 m marine limit. In the central fiord, marine limit falls to 133 m asl and two samples provide minimum dates on its formation. A surface fragment of M. truncata from a raised beach at 127 m asl dated 8510±80 BP [9400-8970 cal BP](TO-5612; Site 2, Fig. 4 and Table I). Immediately up-fiord, whole valves and fragments of Hiattella arctica and M. truncata from a raised beach at 119 m asl dated 8550±80 BP [9420-9000 cal BP](GSC-6047; Site 3, Fig. 4 and Table I). Both dates provide minimum age estimates for the 133 m asl marine limit.

The standard errors of these three dates overlap and thus indicate rapid ice-retreat and formation of marine limit through the outer and central fiord. The dates also indicate that at least 5 m of emergence occurred in <100 (14C and calendar years) years assuming that the samples relate to their respective marine limits (138 m and 133 m). If the samples date relative sea levels at their elevations, then emergence could have been as much as 9 m in <100 years (sample elevations at 128 and 119 m asl). Although both GSC-6047 and TO-5612 yielded similar ages for the 133 m marine limit in central Blind Fiord, the former is a bulk date, and thus could contain a mixture of different aged shells. If this sample is excluded from the reconstruction and the emergence rate based on the two AMS dates (TO-5862 and TO-5612), initial emergence is still 5 m in <100 years.

Control on subsequent emergence in Blind Fiord is provided by four dates. In the inner fiord, marine limit is defined by a gravel beach berm at 129 m asl (Fig. 4). A surface fragment of H. arctica from a raised beach at 123 m asl dated 8310±80 BP [9210-8630 cal BP](TO-5608; Site 4, Fig. 4 and Table I) and provides a minimum age estimate on the 129 m marine limit. North of the fiord head, whole valves and fragments of H. arctica and M. truncata from 107 m asl dated 8220±100 BP [9100-8470 cal BP] (GSC-6054; Site 5, Fig. 4 and Table I), which is also a minimum date on local marine limit at 124 m asl. Up-fiord of this site, a marine limit delta at 120 m asl occurs at the mouth of a lateral meltwater channel. Single valves and fragments dominated by M. truncata collected at 95 m asl on the delta foreslope gave a radiocarbon date of 8090 ±110 BP [8950-8370 cal BP] (GSC-5896; Site 6, Fig. 4 and Table I), which is a minimum age for the 120 m marine limit. Finally, paired valves of Astarte borealis and H. arctica from 31 m asl in silt immediately underlaying a delta at 39 m asl in the central fiord dated 5640±110 [6360-5870 cal

1. All dates were calibrated using CALIB 3.0 (Stuiver and Reimer, 1993), and the date range reported here is that which yields 100 % probability when 2σ is used.
BP (GSC-6102; Site 7, Fig. 4 and Table I). It is important to note that although sites at the fiord head are separated from sites in the outer fiord by a distance of ~38 km, the fiord parallels the regional isobases at 8.5 ka BP (see Fig. 6 below), and hence differential postglacial rebound does not compromise treating the dates as a single relative sea level curve. The relative sea level curve for Blind Fiord (Fig. 5A) demonstrates continuous emergence from 8.6 ka BP to present. Initial emergence was \( \geq \) 5 m/century.

**STARFISH BAY**

Deglaciation of Starfish Bay followed the retreat of ice from outer Trold Fiord. At the mouth of Starfish Bay, ice-contact deltas occur at 113 m asl (Fig. 3) and mark a stillstand during retreat of ice into the outer part of the bay. Marine limit drops to 101 m asl along the north shore of the bay where it is defined by a prominent washing limit. A sample of whole valves of *H. arctica* and *M. truncata* collected from glaciomarine silt at 78 m asl dated 7740\( \pm \)90 BP [8410-8000 cal BP] (GSC-6037; Site 8, Fig. 4 and Table I). This is a minimum age for deglaciation and marine limit.

Marine limit in the inner fiord is marked by a bench cut in till at 89 m asl (Fig. 4). At the fiord head, well-developed deltas fronted by extensive glaciomarine silt occur at 86 and 80 m asl (Fig. 4). These deposits mark a major stillstand of the trunk glacier during deglaciation. A sample of *Portlandia arctica* collected from silt at 68 m asl, 3 km west of the fiord head, dated 8710\( \pm \)120 BP [9730-9060 cal BP] (GSC-2719; Site 9, Fig. 4 and Table I) (Hodgson, 1985). Hodgson inferred that this date provided an approximate age for the fiord head deltas. However, because the sample site occurs 3 km west of the deltas, the validity of this inferred relationship is uncertain, and the shells could alternatively correlate with the 89 m bench in the inner fiord. A final sample consisting of a single valve of *H. arctica*, was collected from silt at 71 m asl east (inland) of these deltas. This dated 7240\( \pm \)80
and provides a minimum age for deglaciation and formation of the fiord head deltas.

Therefore, at the fiord head between 8710±120 BP [9730-9060 cal BP] and 7240±80 BP [7910-7570 cal BP], sea level could have fallen by as little as 9 m, based on the assumption that the 8.7 ka BP [9730-9060 cal BP] date relates to the 89 m marine limit in the inner fiord, and the 7.2 ka BP [7910-7570 cal BP] date relates to a relative sea level at 80 m asl (Fig. 5B, “minimum”). This is equivalent to an emergence rate of only 0.6 m/century (based on the calibrated ages). If the 7.2 ka BP [7910-7570 cal BP] shells are related to a relative sea level at 71 m asl (the sample elevation), then emergence increases to 18 m (1 m/century) (Fig. 5B, “maximum”).

IRENE BAY

Prominent marine limit deltas and thick raised marine silt record ice-marginal stabilisation and deposition at the fiord head during deglaciation. Hodgson (1985) reported several radiocarbon dates from this area and these are discussed below. Ice-contact deltas on the south side of inner Irene Bay are graded to relative sea levels of 80 and 92 m asl (Figs. 3 and 4). Whole valves of *P. arctica* were collected at 70-74 m asl from glaciomarine rhythmites capped by the 80 m delta. This sample dated 8820±90 BP [9810-9340 cal BP] (GSC-1978; Site 11, Fig. 4 and Table I). Immediately east of this site, ice-contact deltas with thick pro-delta silt grade to 75 m asl. Whole valves of *H. arctica* and *M. truncata* from 66-70 m asl in this silt dated 7340±170 BP [8170-7500 cal BP] (GSC-3397; Site 12, Fig. 4 and Table I) and provide a minimum age for the 75 m delta. These ages indicate ~5 m of emergence between 8.8 ka BP [9810-9340 cal BP] and 7.3 ka BP [8170-7500 cal BP], equivalent to an emergence rate of only 0.3 m/century, based on the calibrated ages (Fig. 5C, “minimum”). If the standard errors of the dates are considered, the 8.8 ka BP [9810-9340 cal BP] date related to the 92 m delta (the highest marine limit in this part of the fiord) and the 7.3 ka BP [8170-7500 cal BP] date related to a sea level at 70 m asl (the sample elevation), this results in a maximum initial emergence rate of 2 m/century (Fig. 5C, “maximum”).

Additional evidence suggests that emergence remained slow until at least 5.2 ka BP (5.6 ka cal BP). Immediately up-fiord from the 8.8 ka BP site, a marine limit delta is graded to a former relative sea level at 78 m asl. A sample of paired valves of *A. borealis* and *H. arctica* collected by A. Podor from bedded sand on the delta foreslope at 55 m asl dated 5200±70 BP [5840-5460 cal BP] (GSC-5897; Site 13, Fig. 4 and Table I). This provides a minimum age estimate on delta formation. Thus, assuming no elevation measurement error, sea level at 5.2 ka BP [5840-5460 cal BP] must have been at least as high as 55 m asl (Fig. 5C). This indicates that between 7.3 ka BP [8170-7500 cal BP] and 5.2 ka BP [5840-5460 cal BP] a maximum of 20 m of emergence occurred, equivalent to ~1 m/century based on the calibrated ages. A second sample consisting of a piece of drift-wood was recovered from 55 m asl in a raised beach which off-laps the same delta. This dated 6360±100 BP [7400-7010 cal BP] (GSC-5966; Site 14, Fig. 4 and Table I). Given the presence of GSC-5897 from the same elevation and
POSTGLACIAL ISOBASES 8.5 KA BP

Regionally across western Ellesmere and Axel Heiberg islands, many radiocarbon dates of 8500±150 BP [9420-8650 cal BP] are available. In southern Eureka Sound, more dates fall into this interval than any other, and hence it was selected for assessing differential emergence in the study area. These dates form the control points for the isobase pattern in Figure 6 which integrates new shoreline and radiocarbon data from southern Eureka Sound with previously published information from northwestern Ellesmere Island (Bednarski, 1995), northern Eureka Sound (England, 1992; Bell, 1996), western and northeastern Axel Heiberg Island (Lemmen et al., 1994; Bednarski, 1998) and Norwegian Bay (Hodgson, 1985; Hodgson in McNeely, 1989). This isobase reconstruction (Fig. 6) will likely be refined with additional radiocarbon and relative sea level data, particularly from Norwegian Bay.

Isobases drawn on the 8.5 ka shoreline rise in elevation towards Eureka Sound and form an elongate ridge oriented crudely north/south. The highest value (130 m asl) forms a closed cell of maximum emergence centred along the axis of the channel. At its northern end, the ridge extends north-eastwards into Greely Fiord with closure of the 120 m and 130 m isobases (cf. England, 1992). The northwestern extent of the 130 m isobase cell in Nansen Sound is uncertain, as Bednarski (1998) reports high marine limits (150-160 m asl) in this region, but associated dating control is poor. The 8.5 ka shoreline falls to ≤110 m asl in central Hare Fiord (Bednarski, 1995).

At the southern end, the 130 m isobase extends at least as far south as the mouth of Blind Fiord/Bear Corner, and may extend further south onto Byrnes Peninsula and the entrance to Norwegian Bay. Further southwest, on the north coast of Grinnell Peninsula, Devon Island, the 8.5 ka shoreline is <130 m asl (Dyke 1998, 1999). This demonstrates that the 130 m isobase closes to the northwest, and supports the above interpretation of closure in the vicinity of northern Byrnes Peninsula. It also indicates that the 120 m isobase either closes in Norwegian Bay in the vicinity of Graham and eastern Cornwall islands, or continues south-eastwards onto Grinnell Peninsula. Currently, this cannot be resolved given the lack of emergence data from much of Norwegian Bay, and hence the 120 and 110 m isobases are left open to the south.

DISCUSSION

INITIAL POSTGLACIAL EMERGENCE

Marked contrasts in initial postglacial emergence are recorded in the study area. In Blind Fiord, initial emergence was rapid and characterised by rates of ≥5 m/century (cf. Blake, 1975, 1992a; Lemmen et al., 1994). At this site, relative sea level exhibits a continuous fall since deglaciation and marine limit formation (cf. Type A curve of Quinlan and Beaumont, 1981, and Zone 1 of Clark et al., 1978). The emergence history is thus broadly similar to that from other areas of eastern Arctic Canada formerly covered by the Laurentide Ice Sheet. Postglacial rebound in such areas typically exhibits continuous emergence since deglaciation with initial emer-


<table>
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<th>Site</th>
<th>Location</th>
<th>Laboratory dating No.</th>
<th>Material</th>
<th>Age (years BP)</th>
<th>Enclosing material</th>
<th>Sample elev. (m asl)</th>
<th>Related RSL (m asl)</th>
<th>Calibrated age (cal BP)</th>
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<td>TO-5596</td>
<td><em>Hintella arctica</em></td>
<td>7240±80</td>
<td>Silt</td>
<td>71</td>
<td>&gt;71</td>
<td>7910-7570</td>
</tr>
<tr>
<td>11</td>
<td>Ellesmere Island Irene Bay</td>
<td>GSC-1978</td>
<td><em>Portlandia arctica</em></td>
<td>8820±90</td>
<td>Silt</td>
<td>70-74</td>
<td>≥80 (&lt;92)</td>
<td>9810-9340</td>
</tr>
<tr>
<td>12</td>
<td>Ellesmere Island Irene Bay</td>
<td>GSC-3397</td>
<td><em>Hintella arctica</em></td>
<td>7340±170</td>
<td>Surface</td>
<td>66-70</td>
<td>≥70 - ≤75</td>
<td>8170-7500</td>
</tr>
<tr>
<td>13</td>
<td>Ellesmere Island Irene Bay</td>
<td>GSC-5897</td>
<td><em>Astarte borealis,</em> <em>Hintella arctica</em></td>
<td>5200±70</td>
<td>Sand and silt</td>
<td>55</td>
<td>≥55 - ≤78</td>
<td>5840-5460</td>
</tr>
<tr>
<td>14</td>
<td>Ellesmere Island Irene Bay</td>
<td>GSC-5966</td>
<td>Driftwood</td>
<td>6360±100</td>
<td>Sand</td>
<td>55</td>
<td>≥55 - ≤78</td>
<td>7400-7010</td>
</tr>
<tr>
<td>15</td>
<td>Ellesmere Island Irene Bay</td>
<td>GSC-5955</td>
<td>Driftwood</td>
<td>1790±160</td>
<td>Sand</td>
<td>9</td>
<td>9</td>
<td>2050-1340</td>
</tr>
<tr>
<td>16</td>
<td>Ellesmere Island Eureka Sound</td>
<td>TO-2245</td>
<td><em>Hintella arctica</em></td>
<td>8430±70</td>
<td>Silt</td>
<td>101</td>
<td>&gt;101</td>
<td>9350-8920</td>
</tr>
<tr>
<td>17</td>
<td>Ellesmere Island Cañon Fiord</td>
<td>TO-2339</td>
<td><em>Mya truncata</em></td>
<td>8380±80</td>
<td>Sand</td>
<td>98</td>
<td>&gt;98 - ≤114</td>
<td>9310-8740</td>
</tr>
<tr>
<td>18</td>
<td>Ellesmere Island Fosheim Peninsula</td>
<td>GSC-5156</td>
<td><em>Hintella arctica</em></td>
<td>8680±80</td>
<td>Silt</td>
<td>132</td>
<td>≥132 - ≤150</td>
<td>9560-9150</td>
</tr>
<tr>
<td>19</td>
<td>Ellesmere Island Fosheim Peninsula</td>
<td>TO-2241</td>
<td><em>Mya truncata</em></td>
<td>8480±80</td>
<td>Silt</td>
<td>93</td>
<td>&gt;93</td>
<td>9380-8950</td>
</tr>
<tr>
<td>20</td>
<td>Ellesmere Island Fosheim Peninsula</td>
<td>GSC-4708</td>
<td><em>Mya truncata</em></td>
<td>8520±80</td>
<td>Silt</td>
<td>100</td>
<td>&gt;100 - ≤146</td>
<td>9400-8970</td>
</tr>
</tbody>
</table>
TABLE I (cont.)
Holocene radiocarbon dates, greater Eureka Sound

<table>
<thead>
<tr>
<th>Site</th>
<th>Location</th>
<th>Laboratory dating No.</th>
<th>Material</th>
<th>Age (years BP)</th>
<th>Enclosing material</th>
<th>Sample elev. (m asl)</th>
<th>Related RSL (m asl)</th>
<th>Calibrated age (cal BP)</th>
</tr>
</thead>
<tbody>
<tr>
<td>21</td>
<td>Ellesmere Island</td>
<td>GSC-5155</td>
<td>Mya truncata</td>
<td>8570±120</td>
<td>Silt</td>
<td>100</td>
<td>&gt;100 - ≤145</td>
<td>9500-8950</td>
</tr>
<tr>
<td>22</td>
<td>Ellesmere Island</td>
<td>TO-2233</td>
<td>Mya truncata</td>
<td>8440±80</td>
<td>Surface</td>
<td>94</td>
<td>≥94</td>
<td>9370-8900</td>
</tr>
<tr>
<td>23</td>
<td>Ellesmere Island</td>
<td>TO-2229</td>
<td>Mya truncata</td>
<td>8450±80</td>
<td>Silt</td>
<td>94</td>
<td>&gt;94 - ≤149</td>
<td>9370-8920</td>
</tr>
<tr>
<td>24</td>
<td>Ellesmere Island</td>
<td>GSC-2369</td>
<td>Mya truncata</td>
<td>8450±100</td>
<td>Surface</td>
<td>127</td>
<td>≥127 - ≤146</td>
<td>9400-8820</td>
</tr>
<tr>
<td>25</td>
<td>Ellesmere Island</td>
<td>S-2640</td>
<td>Hiatella arctica</td>
<td>8415±130</td>
<td>Silt</td>
<td>88</td>
<td>&gt;88 - ≤139</td>
<td>9390-8660</td>
</tr>
<tr>
<td>26</td>
<td>Ellesmere Island</td>
<td>S-2645</td>
<td>Mya truncata</td>
<td>8465±130</td>
<td>Sand</td>
<td>95</td>
<td>&gt;95 - ≤124</td>
<td>9430-8730</td>
</tr>
<tr>
<td>27</td>
<td>Ellesmere Island</td>
<td>S-2641</td>
<td>Hiatella arctica</td>
<td>8590±130</td>
<td>Silt</td>
<td>88</td>
<td>&gt;88 - ≤110</td>
<td>9540-8940</td>
</tr>
<tr>
<td>28</td>
<td>Ellesmere Island</td>
<td>S-2639</td>
<td>Mya truncata</td>
<td>8370±130</td>
<td>Silt</td>
<td>103-105</td>
<td>&gt;105</td>
<td>9360-8600</td>
</tr>
<tr>
<td>29</td>
<td>Axel Heiberg Is.</td>
<td>GSC-5408</td>
<td>Mya truncata, Hiatella arctica</td>
<td>8390±100</td>
<td>Sand and gravel</td>
<td>37-63</td>
<td>63 - ≤105</td>
<td>9340-8700</td>
</tr>
<tr>
<td>30</td>
<td>Axel Heiberg Is.</td>
<td>GSC-5411</td>
<td>Mya truncata</td>
<td>8430±80</td>
<td>Surface</td>
<td>84-93</td>
<td>93-124</td>
<td>9360-8850</td>
</tr>
<tr>
<td>31</td>
<td>Ellesmere Island</td>
<td>GSC-244</td>
<td>Hiatella arctica, Mya truncata, fragments</td>
<td>8480±140</td>
<td>Surface</td>
<td>116</td>
<td>≥116 - ≤122</td>
<td>9500-9180</td>
</tr>
<tr>
<td>32</td>
<td>Ellesmere Island</td>
<td>AA-23591</td>
<td>Hiatella arctica fragment</td>
<td>8645±60</td>
<td>Gravel</td>
<td>131</td>
<td>131 - ≤143</td>
<td>9500-9180</td>
</tr>
<tr>
<td>33</td>
<td>Ellesmere Island</td>
<td>GSC-840</td>
<td>Hiatella arctica</td>
<td>8590±150</td>
<td>Surface</td>
<td>107</td>
<td>≥107</td>
<td>9430-8600</td>
</tr>
<tr>
<td>34</td>
<td>Ellesmere Island</td>
<td>GSC-2253</td>
<td>Mya truncata</td>
<td>8420±160</td>
<td>Sand</td>
<td>102</td>
<td>&gt;102</td>
<td>9430-8600</td>
</tr>
<tr>
<td>35</td>
<td>Ellesmere Island</td>
<td>TO-2280</td>
<td>Mya truncata</td>
<td>8450±70</td>
<td>Surface</td>
<td>88</td>
<td>≥88</td>
<td>9350-8940</td>
</tr>
</tbody>
</table>


Laboratory designations: GSC = Geological Survey of Canada; TO = IsoTrace Laboratory, University of Toronto; S = Saskatchewan Research Council; AA = University of Arizona. TO and AA samples were dated by accelerator mass spectrometry. These samples were corrected for isotopic fractionation to a base of $\delta^{13}C = -25$‰; a reservoir correction of 410 years was then applied, which is equivalent to correction to a base of $\delta^{13}C = 0$‰. GSC and S samples were dated conventionally and corrected for fractionation to a base of $\delta^{13}C = 0$‰. GSC terrestrial organic samples were dated conventionally and corrected for fractionation to a base of $\delta^{13}C = 0$‰.

Dates were calibrated using CALIB 3.0 (Stuiver and Reimer, 1993) and the calibrated date range reported here is that which yields 100% probability at 2σ.

1960's GSC uncorrected dates (Hodgson, 1985). These dates have not been corrected for isotopic fractionation or a marine reservoir effect. Approximate corrections could be made for isotopic fractionation to a base of $\delta^{13}C = -25$‰ by adding 400-410 years to this uncorrected age (R. McNeely, unpublished communication to GSC clientele, 1991). A similar amount could then be subtracted to account for the marine reservoir effect. However, such a correction has not been applied as the result would be approximately the same as the uncorrected raw date reported here. GSC dates obtained during the course of this study (1990's) typically show differences between raw and corrected (to a base of $\delta^{13}C = 0$‰) ages which are well within the reported standard errors of the individual dates.

However, the Blind Fiord curve contrasts with those from Irene and Starfish bays, where initial emergence rates of ≤1 m/century are recorded. It is important to note that the timing of initial emergence at all three sites is similar. Slow initial emergence in Irene and Starfish bays commenced at 8.8 ka BP [9810-9340 cal BP] and 8.7 ka BP [9730-9060 cal BP] respectively, extending to at least 7.3 ka BP [8170-7500 cal BP] and 7.2 ka BP [7910-7570 cal BP]. However, in Blind Fiord, emergence during this same interval (~8.6-8.0 ka BP; 9460-9060 cal BP to 8950-8370 cal BP) was characterised by rates of ≥5 m/century.

Sea level curves similar to those from Irene and Starfish bays have been presented for several other areas on Ellesmere Island (England, 1983, 1992, 1997). For example, in Greely Fiord, England (1992) documented a period of relative sea level stability at marine limit from 8.8 to 7.8 ka (14C) BP, after which emergence proceeded slowly (2 m/century) until 7.2 ka BP when it increased to 13 m/century. He attributed this arrest in sea level following marine limit formation to a balance between glacioisostatic uplift and eustatic sea level rise. The lack of initial rapid emergence at these sites was considered to be a consequence of a limited Late Wisconsinan glacial cover as inferred from independent glacial geologic evidence (England, 1978, 1990, 1996).
Emergence curves showing a period of relative sea level stability at marine limit have also been presented from the area of the former Barents Sea Ice Sheet, from Spitsbergen (Forman, 1990) and Franz Josef Land (Forman et al., 1996, 1997), and this stability is also attributed to a balance between glacioisostasy and eustasy. These sites are inferred to have sustained a thinner glacial load than at the former ice sheet centre over the northern and western Barents Sea where Type A (Quinlan and Beaumont, 1981) sea level curves are reported (Forman, 1990; Bondevik et al., 1995). It should be noted, however, that ice thicknesses over the sites which exhibit slow initial unloading may have been as much as 1500 m (Forman et al., 1995, 1996).

Thus the slow initial emergence recorded in Irene and Starfish bays may not necessarily be incompatible with an extensive ice cover over these sites during the Late Wisconsinan. A minimum estimate of ice thickness at the Last Glacial Maximum in southern Eureka Sound based on glacial geological evidence is 1200 m (Ó Cofaigh, 1999), and maximum estimates obtained from glaciological modelling are 1500 m or 2000 m (Reeh, 1984). The three emergence curves would therefore imply marked spatial variations in the form and rate of initial emergence, even between sites in close proximity; Blind Fiord and inner Starfish Bay are <45 km apart. Slow initial emergence at the fiord heads of Irene and Starfish bays might therefore reflect a major ice-marginal stillstand during retreat, as marked by the “drift belt” (Hodgson, 1985; Ó Cofaigh, 1998; Ó Cofaigh et al., in press), which locally restrained rebound at both fiord heads.

However, radiocarbon dates indicate that while rapid initial emergence (≥5 m/century) was proceeding in Blind Fiord, Starfish and Irene bays were experiencing slow initial emergence (≤1 m/century). Because of the proximity in the timing of deglaciation and marine limit formation between the three sites, the validity of such marked spatial variations in initial emergence is questionable, as the associated unloading would presumably have been integrated over a wide area (Walcott, 1970; Andrews, 1970). Thus, the argument that the slow emergence in Starfish and Irene bays simply represents the later part of the local emergence history, and that an earlier phase of rapid emergence is not recorded at these sites on account of their later deglaciation (vs. sites closer to Eureka Sound such as Blind Fiord), is negated by the presence of early Holocene radiocarbon dates at both fiord heads.

Obviously, the legitimacy of slow initial emergence in Starfish and Irene bays is dependent upon the validity of two fiord head dates of 8.8 ka BP [9810-9340 cal BP] and 8.7 ka BP [9730-9060 cal BP] on P. arctica which are associated with relative sea levels of 80-92 m asl. This reconstruction would be invalid if the samples relate to higher relative sea levels not observed in the field, or if the true ages of the original samples were actually younger than the reported ages. No evidence was found at either fiord head for higher shorelines above the surveyed marine limit which is well defined by raised deltas and wave-cut benches.

With respect to the dates being erroneously old, Forman and Polyak (1997) have demonstrated from radiocarbon dating of pre-bomb P. arctica that this species can have marine reservoir values as high as 764 years, possibly reflecting either the incorporation of old carbon from surrounding deposits and porewater, or from freshwater inputs by streams or glacial meltwater. A variable reservoir effect for this genus, possibly as high as 700-800 years, could therefore imply that some dates on P. arctica may be too old. Currently the reservoir effect for P. arctica in Eureka Sound is unknown. However, early Holocene radiocarbon dates on P. arctica from eastern Ellesmere and western Axel Heiberg islands (Blake, 1992a; Lemmen et al., 1994) compare closely to dates obtained on other genera from the same sites and units, indicating that not all P. arctica dates are anomalously old. Resolution of this issue can be made through radiocarbon dating of pre-bomb P. arctica from the Canadian High Arctic, and this work is currently in progress (J. England, personal communication 1998).

**POSTGLacial ISOBASES**

Isobases drawn on the 8.5 ka shoreline demonstrate an elongate ridge of emergence, oriented crudely parallel with the axis of Eureka Sound, extending from Grædel Fiord in the north to the entrance to Norwegian Bay in the south. This extends previous reconstructions of postglacial shoreline delevelling in the region (Blake, 1970; England, 1976b, 1992, 1997; Bell, 1996), and is significant in that it demonstrates: (a) that the highest emergence values form a cell over the length of Eureka Sound; and (b) that this highest cell does not appear to extend southwestwards across Norwegian Bay to Grinnell Peninsula on Devon Island (cf. Blake, 1970), but rather closes in the vicinity of the entrance to Norwegian Bay.

Along the Eureka Sound/Nansen Sound fiord system, glacial geologic and chronologic evidence indicates extensive Late Wisconsinan glaciation, which inundated fiords and inter-island channels (Ó Cofaigh, 1998, 1999; Ó Cofaigh et al., 1998, in press; Bednarski, 1998). A similar reconstruction, advocating extensive ice during the Last Glacial Maximum, has recently been presented for Norwegian Bay, Wellington Channel and Devon Island (Hattestrand and Stroeven, 1996; Dyke, 1998, 1999). This glacial geological evidence negates an explanation of this isobase pattern in terms of overlapping peripheral depressions from separate ice masses on Ellesmere and Axel Heiberg islands (cf. England, 1976a). The pattern of shoreline delevelling recorded by the isobases is inferred to be the result of a glacioisostatic response to the unloading which accompanied early Holocene deglaciation (cf. Blake, 1970). This supports earlier reconstructions of maximum Late Wisconsinan loading along Eureka Sound/Nansen Sound (e.g., Blake, 1970; Walcott, 1972).

Modelling of an Innuitian Ice Sheet over the Canadian High Arctic during the Last Glacial Maximum (Reeh, 1984) results in maximum ice thicknesses of 1500 and 2000 m being located west of the main ice-divide, in the vicinity of Fosheim Peninsula/Eureka Sound. In contrast, ice at the modelled main ice-divide, which is located along the highland rim of eastern Ellesmere Island, is ~700-1000 m thick (Reeh, 1984) (current thickness ~500-800 m, Koerner,
1989). Koerner et al. (1987) also propose that the ice at the main divide over the Agassiz Ice Cap during the Wisconsinan was only 200 m thicker than today. Glacial geological data from Eureka Sound indicate a minimum ice thickness of ~1200 m in the channel during the Last Glacial Maximum (O’Coifaigh, 1999). This suggests that the 8.5 ka BP isobase pattern which shows maximum emergence along Eureka Sound reflects the thickest ice load being located there during the Last Glacial Maximum (cf. Blake, 1970). Closure of the highest values to the southwest points to a possible saddle connecting the Eureka Sound loading centre to that proposed in the western part of Norwegian Bay (Hättestrand and Stroeven, 1996; Dyke 1998, 1999).

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