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The Chronostratigraphy of Late Pleistocene Glacial and Periglacial Aeolian Activity in the Tuktoyaktuk Coastlands, NWT, Canada

Mark D. Bateman[1*] and Julian B. Murton[2]

[1] *Sheffield Centre for International Drylands Research, Department of Geography, Winter Street, University of Sheffield, Sheffield S10 2TN, UK*
[2] *Department of Geography, University of Sussex, Brighton BN1 9QJ, UK*

* Corresponding author. *Tel: +44-114-2227929; Fax: +44-114-2797912, E-mail: m.d.bateman@sheffield.ac.uk;*

Abstract

Aeolian periglacial sand deposits are common in the Tuktoyaktuk Coastlands of Western Arctic Canada. Regionally extensive and thick aeolian sand-sheet deposits have been observed in two major stratigraphic settings: within a sand unit characterized by large aeolian dune deposits; and interbedded with glaciofluvial outwash from the Laurentide Ice Sheet (LIS). Small, localized sand sheets have also been observed along the tops of sandy bluffs, within sequences of drained thermokarst lakes deposits and as an involuted veneer above buried basal ice of the LIS.

On the basis of radiocarbon and optically stimulated luminescence (OSL) dates from preserved periglacial aeolian sand sheets and dunes a regional chronostratigraphy is presented which indicates that both extensive dunes and sand sheets accumulated mainly between *c*. 30 and 13 ka. A switch to dominantly sand-sheet aggradation at c. 14–13 ka, with sand sheets forming widely until *c*. 8 ka, is attributed to (a) surface armouring by glacial deposits associated with the advance of the LIS; and (b) amelioration of the climate from cold aridity. An absence of OSL dates between *c*. 8 and 1 ka suggests that sand sheets stabilized during much of the Holocene. Local sand-sheet aggradation during recent centuries has occurred near sandy bluffs and on the floors of drained thermokarst lakes. The OSL dates constrain the maximum extent of the LIS in the Tuktoyaktuk Coastlands to Marine Isotope Stage 2.

1. Introduction

Vegetated, sandy areas of tundra in the Western Arctic of Canada and Alaska are often underlain by stabilized sheets of windblown sand. Because the sand sheets are in remote environments and are often poorly exposed, little is known about their age and formative environmental conditions (Lea and Waythomas, 1990). Addressing these problems is important to improving knowledge of aeolian sedimentation in contemporary arctic regions and palaeoenvironmental interpretation of sand sheets in former cold environments.

Critical to understanding the environmental factors that promote sand-sheet formation are studies of the relationships between sand-sheet aggradation, climate change and local factors such as active-layer depth, vegetation, glaciation and surface sediments. *A priori*, such factors will affect soil moisture, ice cementation, root binding and outwash supply, and these, in turn, will affect the supply, transport and deposition of windblown sand. To evaluate such relationships over timescales applicable to sand-sheet formation, it is essential to date sand-sheet development and compare its chronology with known periods of environmental change.

This paper discusses the timing of and influences on aeolian sand-sheet development during the last glacial-interglacial cycle in the Tuktoyaktuk Coastlands, western Arctic Canada. The objectives are to (1) describe the stratigraphic setting of the sand sheets; (2) establish a chronology of sand-sheet development using luminescence and radiocarbon dating; (3) determine the relationships between the sand-sheet chronology and known regional environmental change in order to infer the controls on sand-sheet formation; and (4) discuss the significance of the aeolian deposits in understanding regional palaeoenvironmental changes. Discussion of the sand-sheet facies, the relationship between

sedimentation and thermal contraction cracking, and comparison with sand sheets reported from sub-arctic and former, mid-latitude cold regions is given by Murton *et al.* (in prep.).

2. Study Area

The Tuktoyaktuk Coastlands form that part of the Arctic Coastal Plain between Amundsen Gulf and the Mackenzie Delta, Northwest Territories (Fig. 1; Rampton, 1988). They are a low-lying region underlain by a thick sequence of Quaternary sediments and located within the zone of continuous permafrost, the maximum thickness of which is *c*. 740 m beneath northeast Richards Island (Taylor *et al.*, 1996). Mean annual air temperatures are *c*. -9°C to -12°C, and mean annual ground temperatures *c*. -6°C to -10°C (Mackay, 1979). The vegetation is that of the low-arctic tundra zone, comprising mainly sedge- and shrub-tundra, with forest-tundra in the south (Ritchie, 1984).

Contemporary aeolian activity in the Tuktoyaktuk Coastlands tends to be localized near sandy beaches, bluffs and drained lakes, the resulting active aeolian sand dunes and sand sheets tending to be of much smaller scale than the buried sand sheets and dunes. Active cliff-top dunes are found along the tops of sandy coastal bluffs, and coastal dunes occur on large sandy spits flanking the northern edge of the Tuktoyaktuk Peninsula (Rampton, 1988). Coastal dunes sometimes overlie Late Holocene lake-shore dunes that formed by deflation of sand from the floors of drained thermokarst lakes and deposition of it around their margins (Ruz, 1993; Ruz and Allard, 1994). Deflation has also formed numerous blowouts amongst cliff-top dunes, leaving lags of wind-polished pebbles (ventifacts) within blowouts developed in pebbly sand (Rampton, 1988). On the Bathurst Peninsula, wind erosion of Holocene aeolian sands has formed yardangs (Rampton, 1988).

Parabolic and linear sand dunes on the northern end of the Tuktoyaktuk Peninsula have resulted from reworking of underlying sand by
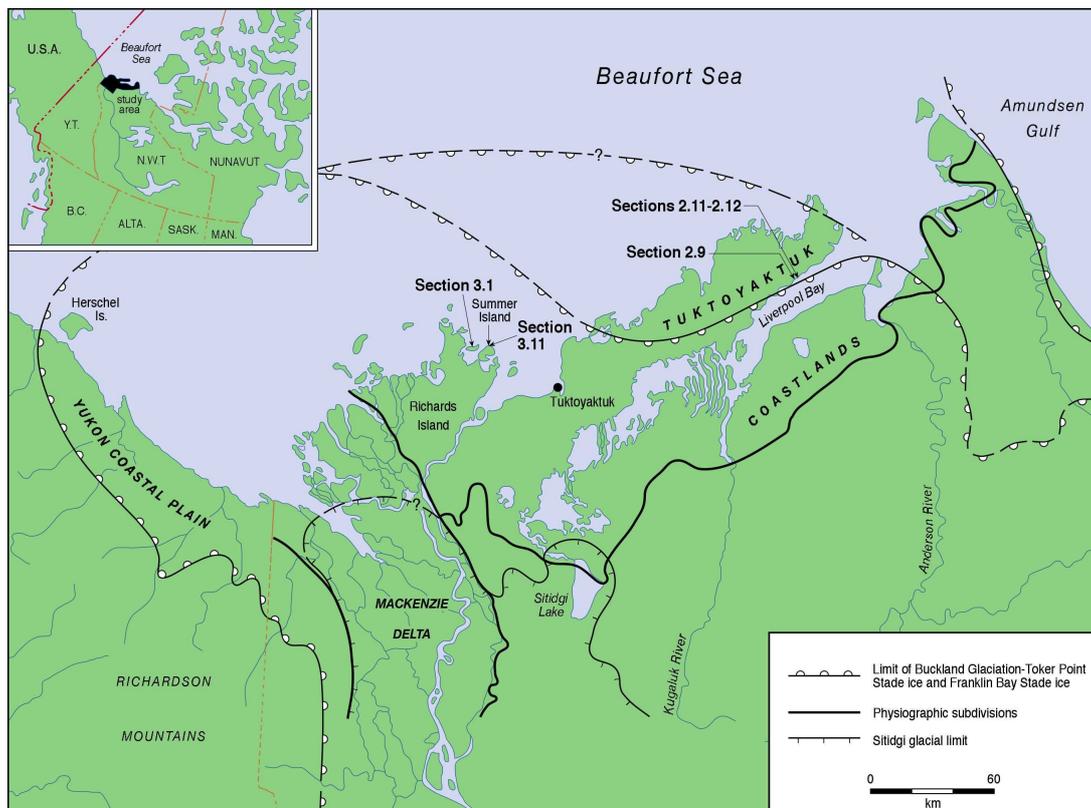

Fig 1: Location map of the Tuktoyaktuk Coastlands including location of sites at on the Tuktoyaktuk Peninsula and in the Summer Island area. Ice limits of the Toker Point Stade and Sitidgi Stade are according to Rampton (1988a). Location of present tree line from Mackay (1963, fig. 70).

winds dominantly from the east, and most of these dunes are now stabilised by lyme grass and willows (Mackay, 1963). The parabolic dunes tend to be distinctly lower (typically only 0.5–3 m high; Michaud and Bégin, 2000) than the buried dunes discussed in this paper (3–15 m high cross sets; Dallimore et al., 1997; Fig. 4). Rampton (1988, p.77) suggested that "most of this activity probably occurred during the Holocene, although few direct dates are available to support this assumption." Michaud and Bégin (2000) obtained a single optical date of 8,600 ± 1,000 yrs (LUX-105gf) from a fossil parabolic dune. The parabolic dunes of the Tuktoyaktuk Peninsula occur within or upon a widespread blanket (i.e. sheet) of aeolian sand and silty sand that is generally 1.5 to 3 m thick and overlies the glaciofluvial Cape Dalhousie Sands (Rampton, 1988, map 1647A).

The Tuktoyaktuk Coastlands were selected for study because (1) sand sheets are well exposed in many coastal bluffs; (2) the area was marginal to the northwest part of the Laurentide Ice Sheet (LIS) on a number of occasions during the Pleistocene and therefore has undergone extensive past environmental change (Fig. 1); and (3) despite climatic amelioration since the Late Glacial, sediments have been extremely well preserved within permafrost.

The study sites are located in two areas of the Tuktoyaktuk Coastlands. The northeastern Tuktoyaktuk Peninsula was examined at more than 30 stratigraphic sections in 1990, 1993 and 2001 along *c.* 20 km of coastal bluffs southwest from Johnson Bay (Fig. 1). Material for sedimentological analyses and for luminescence and radiocarbon dating was collected in 2001 from 5 sections, numbered 2.9–2.13. Second, the area comprising Summer Island, Hadwen Island and, to their south and west, Richards Island (Fig. 1) was logged at numerous coastal sections in 1989–1991, 1993 and 2001. Samples for dating were collected in 2001 from northern Hadwen Island (section 3.1), southern Hadwen Island (section 3.10) and Crumbling Point, on northern Summer Island (section 3.11). All of the dated stratigraphic sections are illustrated in Fig. 2.

3. Stratigraphic setting of the sand sheets

Regional lithostratigraphic units identified by Rampton (1988) and relevant to this study comprise in ascending order: (1) grey sand of the Kidluit Formation (Fm); (2) brown sand of the Kittigazuit Fm; and (3) pebbly clay (diamicton) of the Toker Point Member (Mb) of the Tuktoyaktuk Fm (Rampton, 1988, figs. 19 and 53). The Kidluit Fm has been attributed to deposition on a braided alluvial plain, whereas the Kittigazuit Fm was initially interpreted to be deltaic (Mackay, 1963; Rampton, 1988) but re-interpreted as aeolian dune deposits (Vincent, 1989) eroded primarily from the underlying Kidluit Fm by northwesterly winds (Dallimore *et al.*, 1997). Diamicton of the Toker Point Mb is interpreted as till.

On the northern part of the Tuktoyaktuk Peninsula, Rampton (1988) has identified a lithostratigraphic unit consisting of sand, silty sand and rare gravel ('Cape Dalhousie Sands') beneath an extensive blanket of aeolian sand. He interpreted the Cape Dalhousie Sands as glaciofluvial deposits that accumulated on an outwash plain during the Toker Point Stade and therefore assigned them to the Tuktoyaktuk Fm. As summarised above, the Tuktoyaktuk Fm overlies the Kittigazuit Fm. However, our stratigraphic observations at section 2.13 (Fig. 2c) indicate that the Cape Dalhousie Sands underlie the Kittigazuit Fm, the latter being identified by the large dune foresets that are so characteristic of this formation elsewhere (e.g. section 3.1; Fig. 4). This raises the interesting possibility that the Cape Dalhousie Sands may in fact correlate with the Kidluit Formation.

The sand-sheet deposits described below comprise horizontal to undulating bodies of mainly aeolian sand in which dune deposits with slipfaces are generally absent (cf. Kocurek and Nielson, 1986; Koster, 1988; Schwan, 1988; Lea, 1990). The sand sheets are characteristized by horizontal to gently dipping stratification, and often contain syngenetic sand wedges that have grown vertically with aggradation of the sand (Murton *et al.*, in prep.). Some of the sand-sheet deposits within the Kittigazuit Fm are sandwiched between dune slipface deposits and probably represent the deposits of interdunes, i.e. broadly flat surfaces between dune ridges or mounds (Kocurek, 1981; McKee, 1983).

The sand sheets occur in several stratigraphic and geomorphic settings.

(1)   *Within the Kittigazuit Fm:* The Kittigazuit Fm in the Summer Island area forms a stratigraphic unit that is at least 15–20 m thick (Rampton,

1988) and comprises both large-scale dune slipface deposits (Dallimore *et al.,* 1997) and, less obviously, sand-sheet deposits. The latter are sandwiched between slipface deposits, for example as interdune deposits on northern Hadwen Island (section 3.1; Fig. 2b; Fig. 4). They also overlie slipface deposits, for example on the southwest corner of Summer Island and to the south on Richards Island, where they contain syngenetic sand wedges whose heights can exceed 4–9 m (Murton *et al*., in prep.). Sometimes, however, the precise stratigraphic position of the sand sheets within the Kittigazuit Fm is obscured by slumping, as on southern Hadwen Island, where the syngenetic tops of sand wedges extend into sand-sheet deposits (section 3.10). Overall, the stratigraphy of sand-sheet and dune deposits is complex and variable, as illustrated in sections 2.9 and 2.11–2.13 (Fig 2c).

(2)   *Interbedded with outwash from the LIS:* Sand-sheet deposits are interbedded with glaciofluvial outwash from the LIS on Hadwen

Island and Summer Island (Fig. 1), as previously identified by Murton *et al.* (1997), and on the northeastern Tuktoyaktuk Peninsula. In section 3.1b on northern Hadwen Island (Fig. 2b), most the sedimentary sequence between depths of *c.* 0.5 m and 3.1 m is interpreted as outwash, because it is dominated by pebbly sand with clasts as big as 23 cm in diameter and has a layer of sandy gravel at the base. Within this outwash are sand-sheet deposits (2.30–2.55 m depth) whose horizontal laminae are inversely graded (pin-stripe lamination of Fryberger and Schenk, 1988) and attributed to migrating wind ripples. At Crumbling Point, on northern Summer Island, a sand sheet infills a channel-shaped depression interpreted as a meltwater channel of the LIS (Murton *et al.,* 1997). The sand sheet contains a bed of cross-stratified sand (4 m depth, section 3.11), in places pebbly (Murton *et al*., 1997, fig. 3g), and a number of erosion surfaces overlain by granule-pebble lags, all
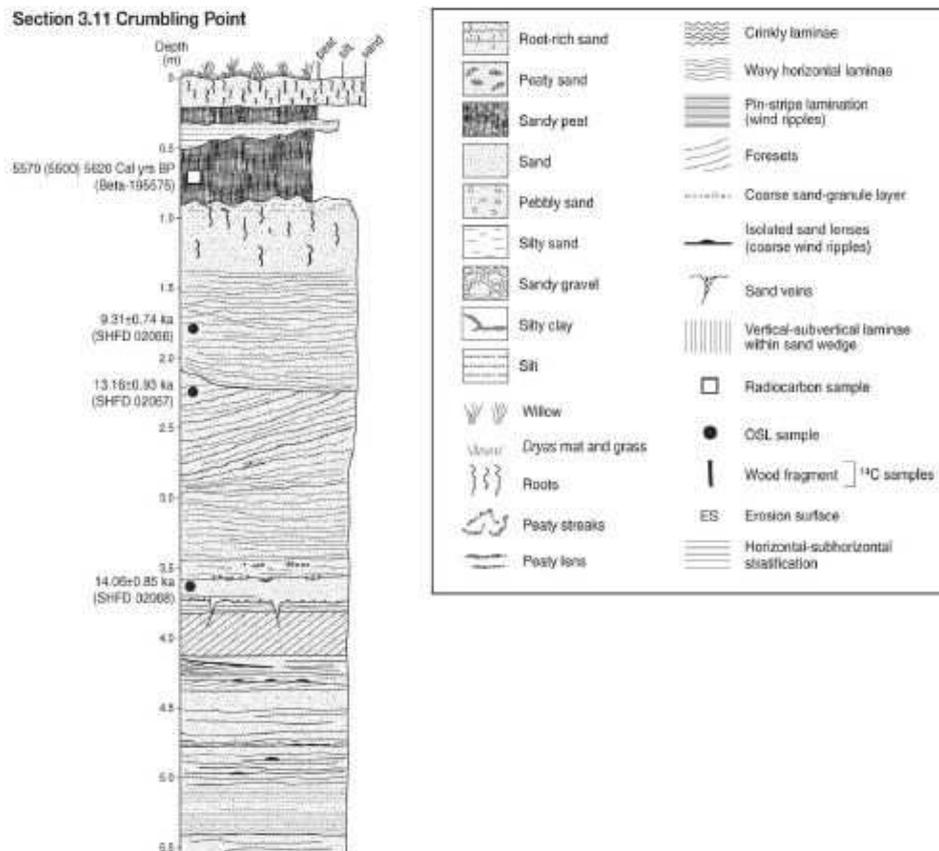

Fig 2: Stratigraphic logs from sample sites in the Tuktoyaktuk Coastlands. (a) Crumbling point. See Figure 1 for locations of sections.
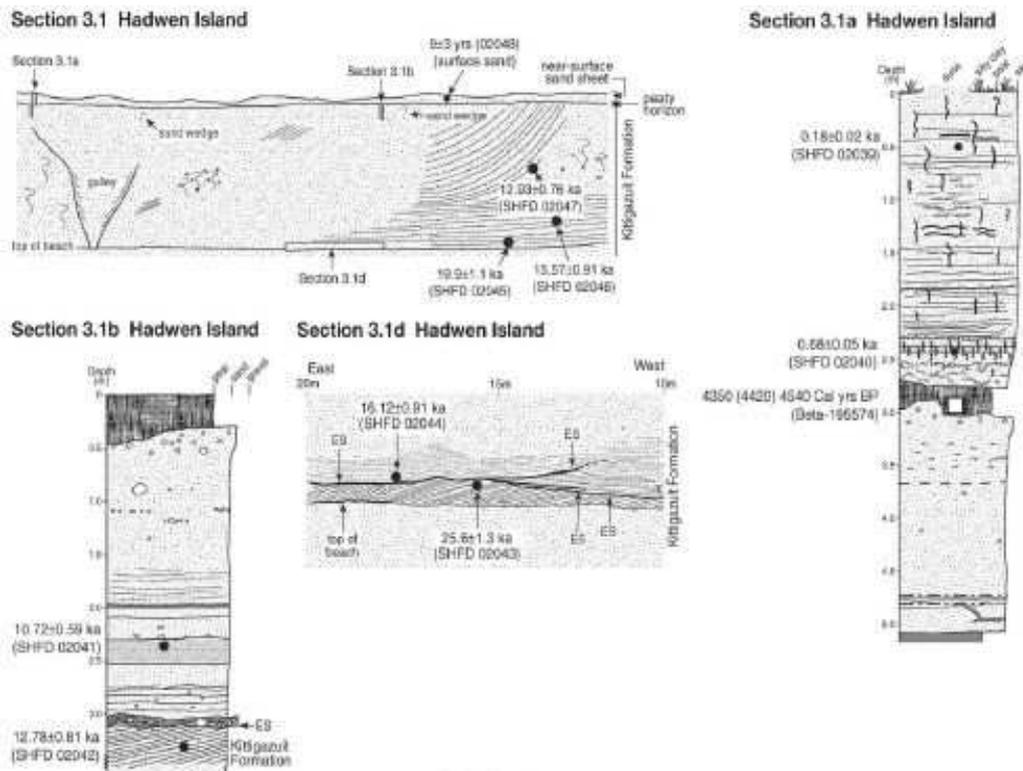
Fig 2 (cont): Stratigraphic logs from sample sites in the Tuktoyaktuk Coastlands. (b) Hadwen Island. See Figure 1 for locations of sections.

features attributed to meltwater activity. In section 2.10, on the northeastern Tuktoyaktuk Peninsula, pebbles and occasional cobbles, overlie two prominent erosion surfaces in a sequence otherwise dominated by aeolian sand (Fig. 2c). The pebbles and cobbles are interpreted as meltwater lag deposits similar to those commonly observed in the northern Richards Island area and central Eskimo Lakes area (e.g. Murton et al., 2005).

(3)   *Involuted within supraglacial melt-out till of the LIS:* The remnants of sand-sheet deposits are commonly observed in thermokarst involutions developed within supraglacial melt-out till of the LIS. In the Summer Island area, the original sand sheet was deposited directly on debris-rich basal ice from the LIS. Only where the thickness of the sand sheet exceeded the maximum depth of active-layer deepening during the subsequent warm early Holocene interval was the sand sheet preserved above the buried ice (Murton *et al.,* 1997, fig. 3e). Elsewhere, the sand sheet was disrupted by soft-sediment deformation during melt-out of the supraglacial till (Murton *et al.,* 1997, 2005).

(4)   *Above the deposits of drained thermokarst lakes:* Sand-sheet deposits often cap sedimentary sequences within drained thermokarst lake basins in the Tuktoyaktuk Coastlands ('vegetation-rich sand' of Murton, 1996). The sand is reworked by wind from the exposed floors of drained lakes and from adjacent sandy bluffs. An example of one such sand sheet is shown in Fig 2c, Section LB2.9.

(5)   *Inland from the tops of sandy bluffs:* Where sand is being released from permafrost in large quantities and significant cliff erosion is currently occurring, some of this sand is blown up the cliff face and deposited on the bluff top. In such localities undulating sheets of aeolian sand with small dunes and numerous blowouts develop. The dune forms comprise small low (<1 m) barchanoid features with clear slip faces and ripple patterns form on unvegetated surfaces (Fig. 3a); nebkha dunes form around clumps of willow and lyme grass (Fig.2b, Section HW3.1a). Vertical sections through such areas indicate that most of the stratification is horizontal to gently dipping, typical of sand-sheet deposits.
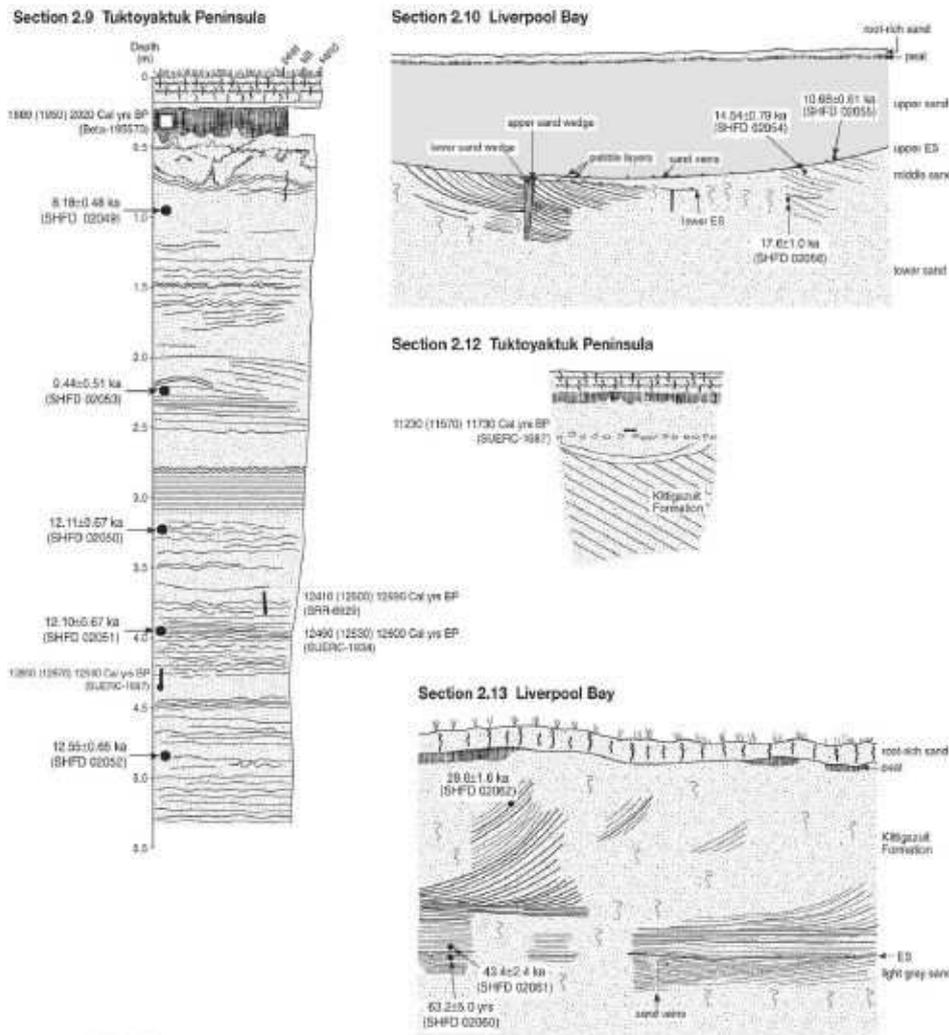
Fig 2 (cont): Stratigraphic logs from sample sites in the Tuktoyaktuk Coastlands. (c) Tuktoyaktuk penisular. See Figure 1 for locations of sections.

4. Geochronology

To evaluate the palaeoenvironmental significance of the periglacial aeolian sediments in the region and in particular the controls on their deposition, it is critical to understand both the age and duration of these sediments. This was achieved through an extensive optically stimulated luminescence (OSL) dating programme with independent validation from radiocarbon dating using samples collected during the 2001 field-season.

4.1 Radiocarbon dating

There are a number of potential problems with establishing a radiocarbon chronology of aeolian sand deposition for the Tuktoyaktuk Coastlands. Firstly, organic material is limited or entirely absent from many of the sedimentary horizons. Secondly, where thick (>10 cm) organic units do exist, these indicate only when aeolian activity was absent; they do not directly date such activity. Finally, the high organic preservation potential in permafrost areas, where the permafrost limits decay, means there is a higher than normal potential for incorporation of recycled older carbon (e.g. Nelson *et al.,* 1988). However, radiocarbon results are deemed useful for providing some independent validation for the OSL derived chronology of aeolian activity and the approximate age of post-aeolian deposits.

(a)

Fig 4: High-angle stratification of the Kittigazuit Fm dunes sitting unconformably on older horizontally bedded sand-sheet deposits, northern Hadwen Island.

A total of six samples were collected from visually undisturbed and laterally continuous peaty units and *in situ* plant remains found in four sites. Inorganic material was largely removed from each sample by floating off the organic material in high-purity de-ionised water and oven-drying. The concentrated organic material was sent for radiocarbon dating either at Beta Analytical or the NERC Radiocarbon Laboratory at East Kilbride. Three of the samples (LB2.9–sample 1, HW3.1 and CP3.11) had insufficient material in excess of 180 microns for a bulk radiocarbon date, and no identifiable macrofossils for accelerator mass spectrometric (AMS) dating although many fine rootlets were present. Analysis of these samples was therefore carried on the <180 micron organic material fraction, which may include an element of modern carbon, especially for the near-surface samples. One sample (LB2.9 sample 3) underwent both bulk and AMS radiocarbon dating on a wood fragment, the results of which closely match, giving some confidence that other bulk dates may be valid (Table 1). Results from the radiocarbon assays along with the calibrated ages can be seen in Table 1 and Figure 2.

4.2 Luminescence dating

OSL dating offers a number of advantages when establishing a Late Pleistocene aeolian chronology. Firstly, it is directly dating the
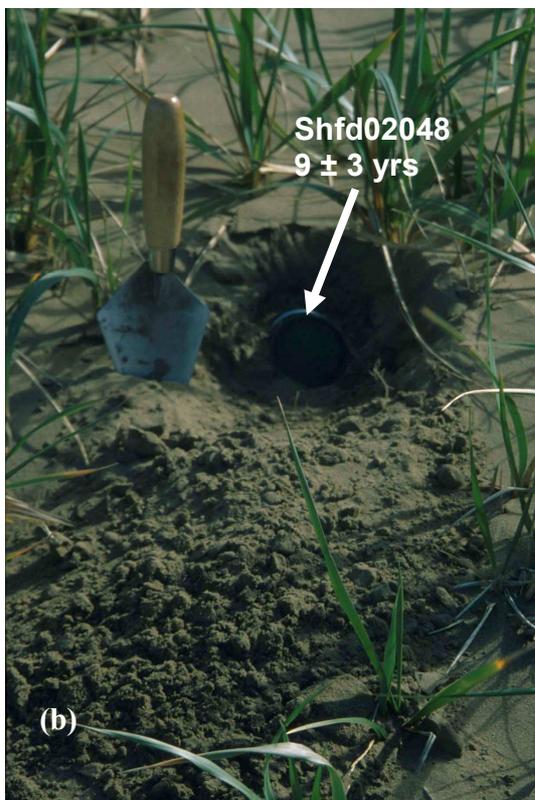
Shfd02048
9 ± 3 yrs

(b)

Fig 3: Modern aeolian activity Hadwen Island. (a) ripple sand of modern cliff-top dunes, (b) OSL sample taken from surface sediment in modern cliff-top dune.

Table 1: Radiocarbon data for Tuktoyaktuk coastland samples as provided by Beta Analytical (Beta) and the NERC Radiocarbon laboratory (SUERC). Ages were calibrated into calendar years using the Calib V3.0 programme of Stuiver and Reimer (1993).

Site Details	No.	Lab Code	Depth (m)	^{14}C Enrichment (% Modern)	δ ^{13}C$_{PDB}$ ± 0.1‰	Age (yrs BP)	Calibrated Age (yrs BP)
Liverpool Bay LB 2.9	1	Beta - 195573	0.25	n/a	-26.9	2020 ± 60	1880 (**1950**) 2020
	2	SRR-6928	3.75	25.59 ± 0.22	-27.0	10949 ± 70	12800 (**12870**) 12940
	3	SRR-6929	4.25	26.80 ± 0.26	-27.4	10578 ± 80 *	12410 (**12500**) 12590
	3	SUERC-1934	4.25	26.70 ± 0.20	-27.9	10609 ± 59 *	12460 (**12530**) 12600
Liverpool Bay LB2.12	1	SUERC-1687		26.54 ± 0.21	-27.6	10074 ± 59	11230 (**11570**) 11730
Hadwen Island HW3.1a	1	Beta-195574	3.00	n/a	-26.7	3990 ± 70	4350 (**4420**) 4540
Crumbling point CP3.11	1	Beta-195575	0.50	n/a	-27.0	4830 ± 50	5570 (**5590**) 5620

* = 2 assays on same wood fragment

aeolian sediments themselves rather than associated, but depositionally and/or palaeoenvironmentally different, material. Secondly, the preferred mineral, quartz, is abundant, allowing strategic sampling both vertically and spatially across the region. Thirdly, OSL is capable of deriving numeric ages beyond the radiocarbon maximum. Finally, the OSL signal, when exposed to sunlight, is reduced to a low background level in only seconds. Thus the potential, especially with aeolian sediments, to erroneously date antecedent events is low. However, the very sensitivity of samples to resetting means that OSL ages may indicate either main periods of aeolian activity or periods of reworking.

A total of 24 OSL samples were taken from 5 sites on the Tuktoyaktuk Peninsula, Hadwen Island and Summer Island (Figs 1 and 2). Samples were collected from freshly exposed vertical sections in opaque PVC tubes whose ends were sealed to form a light-tight container. Sampling avoided pedogenic or cryoturbated sediments and also sediments within 30 cm of a stratigraphic boundary. All samples were dated at the Sheffield Centre for International Drylands Research Luminescence Laboratory using the sample preparation procedure to obtain clean quartz fractions from each sample outlined by Bateman and Catt (1996). The purity of extracted quartz was tested for each sample by stimulation with infra-red light. No samples showed signs of feldspar contamination. OSL from all samples was measured using an upgraded Risø TL-DA-12 reader with optical stimulation provided by a 150 W halogen lamp filtered with GG-420 and SWP interference filters and the resultant luminescence measured by the photomultiplier tube through a Hoya U-340 filter. The single aliquot regeneration (SAR) procedure was used to determine the palaeodose (D_e) using the OSL response to a standardised test dose to monitor sensitivity changes (Murray and Wintle, 2000). OSL measurements were made at 125 ^{o}C with a stimulation time of 80 s. The main preheat used (240 ^{o}C for 10 s) was experimentally derived from a preheat plateaux test. A cut heat of 160 ^{o}C was used prior to the OSL response to the test dose. Five regeneration points were measured within the SAR procedure, including a replicate of the first regeneration point which was used to check that the sensitivity correction procedure was performing adequately. As is consistent with sediments which have been through many cycles of sedimentation (shown by the homogeneity of potassium, uranium and thorium across the region, Table 2), the quartz from all samples responded well to luminescence measurements (Fig. 5). Between 14 and 32 replicate D_e measurements (aliquots) per sample were carried out. Data from aliquots were rejected if an aliquot exhibited poor growth of OSL with laboratory dose or if recycling values fell beyond 1.0 ± 0.1. All replicate D_e data from each sample were plotted as histogram, probability and radial plots (Fig. 5) and further aliquots were rejected if they appeared as outliers (i.e. beyond 2 standard deviations of mean) or if the probability plot showed them to be non-normally distributed (Bateman *et al.* 2003). In this way, it was hoped to have excluded any aliquots whose OSL signal was derived from partially resetting or post-depositional disturbance. At this stage, only an average of 1-2 aliquots per sample were rejected and for 11 samples no aliquots were rejected. Final D_e values were derived using a weighted (by inverse variance) mean with associated standard errors.

Whilst resetting of the OSL signal is very rapid upon sunlight exposure (bleaching), uniform

Table 2: Luminescence data for Tuktoyaktuk coastland samples. Dosimetry as determined by *in situ* gamma spectrometry (Note consistency: K mean = 1.17% σ = 0.08, U mean = 1.56 ppm σ = 0.19 , Th mean = 4.33 ppm σ = 0.52). Total dose rate including palaeomoisture values base on those of present day. Palaeodose data based on the average of multiple replicate OSL measurements and a SAR protocol. Derived absolute ages presented in years from present (2004) with 1 standard deviation errors.

Sample Details			Dose rate Data					Palaeodose	Age
Site	Lab Code	Depth (m)	K (%)	U (ppm)	Th (ppm)	Water Content (%)	Total Dose Rate (Gy/ka)	De (Gy)	(ka)
Liverpool Bay									
LB2.9	Shfd02049	0.95	1.14 ± 0.06	1.41 ± 0.14	4.23 ± 0.42	3.4 ± 5	1.920 ± 0.097	15.72 ± 0.49	**8.18 ± 0.48**
	Shfd02050	2.25	1.11 ± 0.06	1.64± 0.16	4.23 ± 0.42	5.1 ± 5	1.862 ± 0.094	17.76 ± 0.35	**9.44 ± 0.51**
	Shfd02053	3.25	1.13 ± 0.06	1.53± 0.15	4.23 ± 0.42	4.5 ± 5	1.886 ± 0.094	22.86 ± 0.37	**12.11 ± 0.67**
	Shfd02051	3.95	1.07 ± 0.05	1.41± 0.14	4.09 ± 0.41	2.8 ± 5	1.795 ± 0.092	21.75 ± 0.43	**12.10 ± 0.67**
	Shfd02052	4.85	1.09 ± 0.05	1.85± 0.19	4.67 ± 0.47	6.4 ± 5	1.870 ± 0.092	23.47 ± 0.32	**12.55 ± 0.64**
LB2.10	Shfd02055	4.1	1.13 ± 0.06	1.25 ± 0.13	3.82 ± 0.38	3.0 ± 5	1.788 ± 0.094	19.10 ± 0.41	**10.68 ± 0.61**
	Shfd02054	4.6	1.13 ± 0.06	1.60 ± 0.16	4.20 ± 0.42	3.4 ± 5	1.907 ± 0.097	27.73 ± 0.54	**14.54 ± 0.79**
	Shfd02056	5.8	1.32 ± 0.07	1.99 ± 0.20	5.82 ± 0.58	7.7 ± 5	2.18 ± 0.11	38.4 ± 1.1	**17.6 ± 1.0**
LB2.13	Shfd02062	1.57	1.24 ± 0.06	1.46 ± 0.15	4.74 ± 0.47	2.9 ± 5	2.06 ± 0.11	61.0 ± 1.1	**29.6 ± 1.6**
	Shfd02061	3.9	1.19 ± 0.06	1.88 ± 0.19	4.48 ± 0.45	16.1 ± 5	1.755 ± 0.089	76.2 ± 1.6	**43.4 ± 2.4**
	Shfd02060	4.1	1.00 ± 0.05	1.33 ± 0.13	3.15 ± 0.32	2.3 ± 5	1.680 ± 0.086	105.2 ± 2.2	**62.6 ± 3.4**
Hadwen Island HW3.1									
	Shfd02048	0.005	1.20 ± 0.06	1.36 ± 0.14	4.16 ± 0.42	2.2 ± 5	2.01 ± 0.10	0.019 ± 0.007	**0.009 ± 0.003**
	Shfd02039	0.5	1.20 ± 0.06	1.36 ± 0.14	4.16 ± 0.42	3.2 ± 5	1.98 ± 0.10	0.350 ± 0.032	**0.18 ± 0.2**
	Shfd02040	2.45	1.26 ± 0.06	1.46 ± 0.15	4.37 ± 0.44	6.0 ± 5	1.97 0.10	1.336 ± 0.056	**0.68 ± 0.05**
	Shfd02041	2.4	1.21 ± 0.06	1.38 ± 0.14	4.10 ± 0.41	4.3 ± 5	1.918 ± 0.099	20.56 ± 0.35	**10.72 ± 0.59**
	Shfd02042	3.2	1.17 ± 0.06	1.51 ± 0.15	4.61 ± 0.46	5.0 ± 5	1.925 ± 0.098	24.59 ± 0.93	**12.78 ± 0.81**
	Shfd02047	8	1.17 ± 0.06	1.72 ± 0.17	4.62 ± 0.46	4.8 ± 5	1.92 ± 0.10	24.78 ± 0.68	**12.93 ± 0.76**
	Shfd02045	15.55	1.21 ± 0.06	1.75 ± 0.18	5.24 ± 0.52	6.3 ± 5	1.90 ± 0.10	29.22 ± 0.64	**15.57 ± 0.91**
	Shfd02046	16.9	1.10 ± 0.06	1.68 ± 0.17	4.63 ± 0.46	5.6 ± 5	1.797 ± 0.094	35.53 ± 0.84	**19.9 ± 1.1**
	Shfd02044	16.95	1.17 ± 0.06	1.67 ± 0.17	4.53 ± 0.45	7.8 ± 5	1.766 ± 0.096	28.47 ± 0.48	**16.12 ± 0.91**
	Shfd02043	17.45	1.18 ± 0.06	1.78 ± 0.18	4.08 ± 0.41	3.1 ± 5	1.95 ± 0.10	49.80 ± 0.10	**25.6 ± 1.3**
Crumbing Point									
CP3.11	Shfd02066	1.8	1.03 ± 0.05	1.30 ± 0.13	3.77 ± 0.38	4.0 ± 5	1.724 ± 0.087	16.05 0.10	**9.31 ± 0.74**
	Shfd02067	2.3	1.05 ± 0.05	1.35 ± 0.14	3.27 ± 0.33	3.3 ± 5	1.735 ± 0.088	22.8 ± 1.1	**13.16 ± 0.93**
	Shfd02068	3.55	1.07 ± 0.05	1.58 ± 0.16	3.99 ± 0.40	3.8 ± 5	1.828 ± 0.092	25.71 ± 0.85	**14.06 ± 0.85**

exposure of each sand grain cannot be guaranteed, even for sub-aerially transported aeolian sediments in regions within the Arctic Circle, because sediment movement could occur during periods of the year dominated by hours of darkness. Such incomplete bleaching was highlighted by the work of Mejdahl and Funder (1994) in Greenland, although Berger and Anderson (2000), using the less light sensitive TL signal, showed good resetting had taken place at depth within two arctic lakes in Alaska. To test this fundamental question of resetting, one sample (Shfd02048) was collected from near-surface sand (<5 cm depth) from a cliff-top dune which had fresh wind ripples and vegetation regrowth just starting to take place (Fig. 3b).

OSL measurements revealed an extremely low OSL signal, in line with expectations for fully bleached modern sediment (Fig 5), yielding an age of 9 ± 3 years. Whilst a single sample from a cliff top dune cannot taken as a true analogue for Pleistocene sand-sheets it does demonstrate that resetting of sand in arctic aeolian environments is possible. Furthermore, repeated D_e measurements of older samples also showed good reproducibility indicative of sediment which had been fully exposed to sunlight prior to burial (Fig 5). Thus it was concluded that the sediments collected for this study had been fully reset prior to burial, and that the OSL ages from the samples are burial ages.
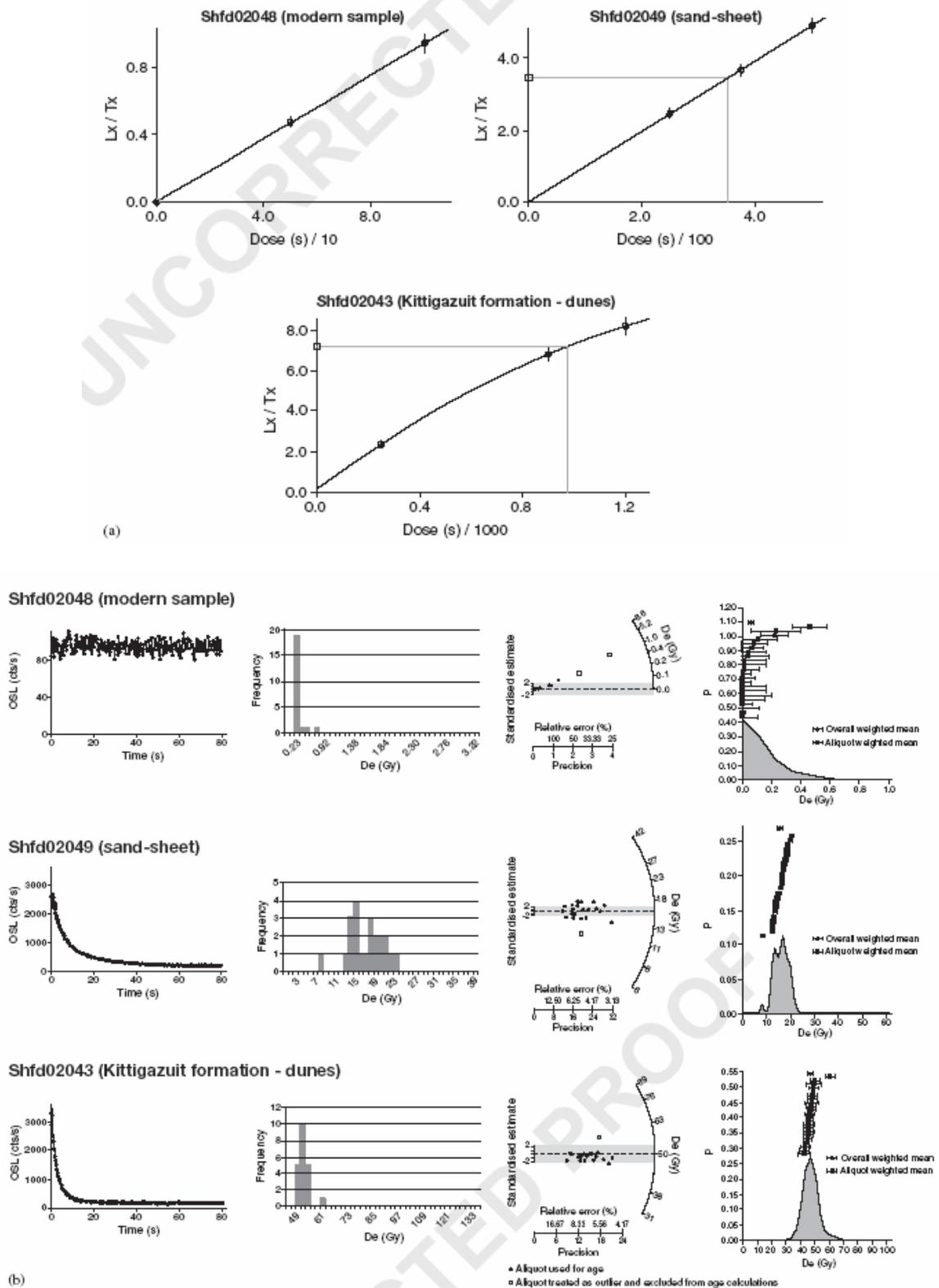
Fig 5: OSL data for a range of depositional environments. (a) single aliquot regeneration growth curves showing good response to laboratory dose (b) OSL decay curves (Note the lack of OSL signal for the modern sample showing full resetting prior to burial) and D_e replicate histograms, radial and probability plots. The high level of replicate reproducibility is indicative of good resetting of the OSL 'time clock' prior to burial.

Dose rates were based on *in situ* gamma spectrometry measurements using an EG&G Micronomad. Palaeo-moisture values were assumed to have been similar to the present day, although errors of $\pm$ 5% were adopted to take into account past fluctuations. That all samples have been preserved through the development of permafrost within them means that moisture fluctuations and leaching or remobilisation of potassium or uranium within the sediment are likely to have been minimal. This inference is supported by an average uranium:thorium ratio of 2.8, which is consistent with naturally occurring concentrations of these isotopes. Both the uranium and thorium decay chains are therefore assumed to be in equilibrium. The cosmic dose-rate contribution was calculated iteratively for samples within each vertical sequence (Prescott and Hutton 1994; Telfer *et al.,* submitted). The final ages in calender years before 2004 are presented in Table 2.

4.2 Validation of chronology

The OSL chronology is thought to be absolute and accurate as, with two exceptions (Shfd02050 and Shfd02046), all OSL ages increase in age with depth and are stratigraphically consistent within sites. As discussed above, there is no reason to presume incomplete bleaching prior to burial, and permafrost will have minimised potential variation in dosimetry through time. Where both radiocarbon and OSL ages are available within the same section they are stratigraphically compatible (Fig. 2) and, in the case of site LB2.9, in agreement within analytical errors. Final verification comes from dates reported by Murton *et al.* (1997) from Crumbling Point (CP3.11). These were produced from a different laboratory using a different luminescence technique (infra-red stimulated luminescence; IRSL) and a different mineral (feldspar). Despite this, the sand-sheet dates from the present study of 14.06 and 13.16 ka (Shfd02067 and Shfd02068) are directly comparable to seven IRSL dates from stratigraphically related sand wedges, which ranged from 13.5 to 14.8 ka.

5. Results and discussion

The resultant dataset spans the last 63 ka and has significance both in terms of palaeo-aeolian activity and also the extent and timing of the last glacial advance of the LIS in the Tuktoyaktuk region.

5.1 Aeolian Chronostratigraphy

Aeolian sediments date to at least the last 43 ka, with most dates falling in the period 30–8 ka. These dates indicate that aeolian activity had commenced by Marine Isotope Stage (MIS) 3, continued throughout MIS 2 and extended into early MIS 1. The paucity of dates before 30 ka may reflect one or more of the following possibilities: (a) limited development of aeolian activity in the region, (b) erosion or reworking of previously deposited aeolian sediments, or (c) sampling bias; further data are needed to resolve this issue. Widespread aeolian activity during MIS 2 (*c.* 24–12 ka; Martinson *et al*., 1987) probably reflects regional aridity caused by a number of coincident events that combined to provide a 'window of opportunity' in which there was sufficient sediment supply, strong winds and little surface moisture. Eustatic sea level fell from *c.* –40 m at 30 ka to *c.* –150 m at 18 ka, which would have reduced the maritime climatic influence to the Tuktoyaktuk Coastlands. It also closed the Bering Strait, forming Beringia (Lambeck and Chappell, 2001), which, according to palaeovegetation reconstructions, is also thought to have been extremely cold and dry between 34 and 26 ka BP in the east (Anderson and Lozhkin, 2001) and very dry between *c.* 30 and 14 ka (Edwards *et al*., 2001; Ager, 2003). Additionally, the Cordilleran Ice Sheet (CIS), which grew from 30 Cal ka onwards (maximum extent *c.* 17.5 Cal ka; Dyke *et al.* 2002; Clague and Menounos, 2005), may have deflected the jet stream south (Edwards *et al.,* 2001) and blocked the passage of moist Pacific air to the northeast of it (Dyke *et al.,* 2002). In short, the supply of moist air to the Tuktoyaktuk Coastlands during MIS 2 would have been severely limited from the regions to the west and north of the LIS.

Abundant sand was already available in the Tuktoyaktuk Coastlands for reworking by wind into dunes and sand sheets. As summarised above, this pre-existing sand consisted of the Kidluit Fm and, at least beneath the northeast Tuktoyaktuk Peninsula, the Cape Dalhousie Sands. In this context, a single date from the Cape Dalhousie Sands beneath the Kittigazuit
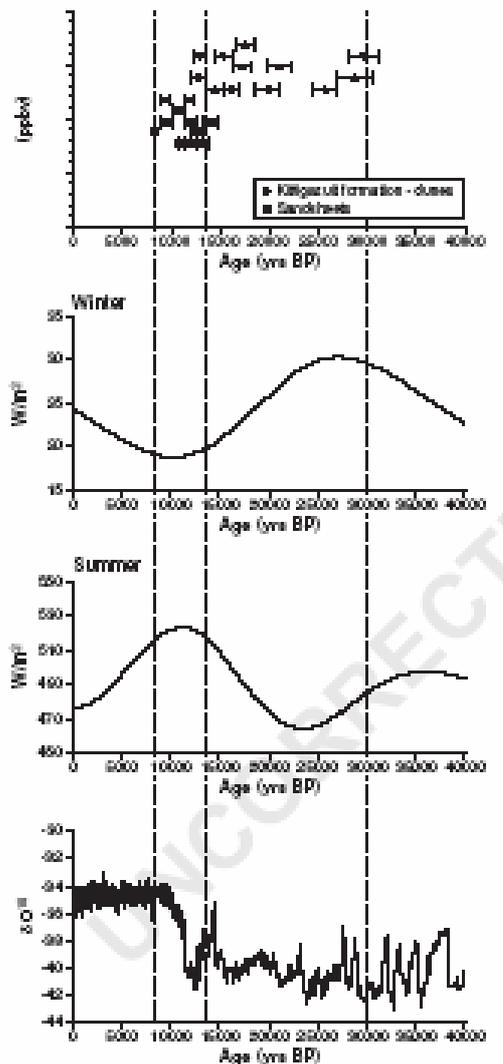
Fig 6: OSL chronology from the different aeolian deposits of the Tuktoyaktuk. Also shown Gisp2 Oxygen Isotope record (Grootes *et al.* 1993) and variation in insolation at 60° N record (Berger and Loutre 1991). Dash lines denote change in dominance of dunes to sand-sheets.

Fm at section 2.13 on the Tuktoyaktuk Peninsula provided the oldest date (Shfd02060; 62.6 ± 3.4 ka) obtained in the present study. Additionally, we cannot discount the possibility that fluvial activity associated with the pre-Last Glacial Maximum (LGM) Porcupine and Anderson river basins (Duk-Rodkin and Hughes, 1994) and the post-LGM Mackenzie and Anderson systems (Lemmen *et al*., 1994; Duk-Rodkin and Lemmen, 2000) delivered sand to the Tuktoyaktuk Coastlands or their fringes, where it was reworked by wind. As discussed below, there is evidence from northeast Richards Island for deglacial meltwater activity interrupting aeolian sedimentation.

Aridity coupled with plentiful sediment cannot account for the extensive aeolian deposits found across the Tuktoyaktuk Coastlands, unless there was sufficient wind. Here lies an enigma, as global circulation models of atmospheric climate during the LGM show that wind vectors in the study area were low throughout the year (G. Bigg, pers. comm. 2005). However, work from Svalbard (Norway) and Alaska (USA) has reported the importance of local katabatic winds in the formation of cold-climate aeolian dunes and sand sheets (e.g. Riezebos *et al.,* 1986; Dijkmans and Koster, 1990; Lea, 1990). The Tuktoyaktuk Coastlands are thought have to been ice-free but increasingly ice-marginal from 30 ka onwards as the LIS expanded northwards (Dyke *et al.,* 2002). It is possible that the Coastlands would therefore have been subject to local katabatic winds coming off the LIS and funnelled northwards by the Richardson mountains to the west. However, this appears to be inconsistent with field evidence of foreset azimuths, which suggests that the dunes were formed by winds from the northwest (Dallimore *et al.,* 1997). Interestingly, winds from the northwest and west are also recorded, in an ice-proximal tundra location between the LIS and CIS in central Alberta, by stabilized sand dunes optically dated to between 16 and 13 ka (Wolfe *et al*., 2005). However, between about 13 and 9 ka, dunes formed under southeasterly katabatic winds coming off the retreating LIS (Wolfe *et al*., 2005). Clearly, further investigations are needed to resolve the nature and timing of wind activity.

During the period from 30 ka to present, the type and amount of aeolian activity appears to have varied through time (Fig. 6). For the purposes of discussion this variation has been separated into phases, although as overlap in figure 6 shows these phases may be time-transgressive.

Phase I: Regional dunes and sand sheets:

Dunes and sand sheets were deposited in the period *c.* 43–13 ka. This extends the period of aeolian activity inferred by Dallimore *et al.* (1997), who ascribed the deposition of the Kittigazuit Fm on northeast Richards Island to the middle Wisconsinan (MIS 3), based on two AMS ^{14}C age estimates of 37 400 ± 810 ^{14}C years BP (CAMS–14839) and 33 710 ± 460 ^{14}C years BP (CAMS–14837) obtained from detrital seeds and moss fragments. These estimates are distinctly

older than all of the OSL dates obtained in the present study from Kittigazuit dune deposits (Fig. 6); only one OSL date (43.4 ± 2.4 ka; Shfd02061), from Kittigazuit sand-sheet deposits on the northeast Tuktoyaktuk Peninsula, is older than the ^{14}C age estimates. The OSL ages indicate that aeolian activity occurred during both the middle and late Wisconsinan, but especially after *c.* 30 ka. Evidence from the northern Yukon Territory and interior Alaska may give clues as to why aeolian activity at the sites examined in the present study was less prevalent prior to 30 ka. According to Anderson and Lozhkin (2001), pollen spectra from sites including those in the Old Crow and Bell basins of the northern Yukon Territory suggest that *Picea* forests developed between *c.* 38 to at least 34 ka BP, during a middle Wisconsinan climatic optimum, when the climate was cooler than present but warmer than during stadial conditions. Similarly Berger (2003) has TL dated a prominent, thick palaeosol at Gold Hill, near Fairbanks, interior Alaska, to *c.* 30 ka, attributing the period of soil formation to a warm interval during late MIS 3 that equates with the Fox thermal event inferred by Hamilton et al. (1988) from the nearby Fox permafrost tunnel. Indeed, he suggested that significant soil formation and climatic warming took place globally during a warm interval *c.* 35–30 ka. Given the uncertainties of radiocarbon age estimates older than *c.* 30 ka, it is not known whether this 35–30 ka interval coincides with the *c.* 38–34 ka warm interval inferred by Anderson and Lozhkin (2001). What is clear, however, is that the two AMS ^{14}C age estimates obtained by Dallimore *et al.* (1997) from the Kittigazuit Fm lie within the *c.* 38–34 ka warm interval, when increased vegetation cover in the Tuktoyaktuk Coastlands limited aeolian activity in comparison to the following period of widespread dune building.

The occurrence of widespread aeolian dunes dating from *c.* 30–13 ka in the Tuktoyaktuk Coastlands is consistent with evidence from neighbouring regions for severe aridity and aeolian activity during the late Wisconsinan (MIS 2). Late Wisconsinan sand sheets and dunes are reported from nonglaciated lowlands in central and southwestern Alaska (Lee and Waythomas, 1990) and from the central Kobuk valley in northwestern Alaska (Dijkmans and Koster, 1990). The Kittigazuit dunes are also similar in size and age to stabilized linear dunes ≤30 m high in a sand sea, informally known as the Ikpikpuk dunes, to the west of the Colville River, northern Alaska (Carter, 1981, 1983, 1993; Dinter *et al.*, 1990).

Fig 7: Upper-most high-angle dune beds of Kittigazuit Fm overturned by overlying Toker Point till, Hadwen Island. Note: right-hand image identical except bedding picked out with trowel marks.

During the cold, arid phase of the Duvanny Yar interval (25–14 ka BP) in northeast Beringia the driest and most hostile Beringian environments were the lowlands flanking the Beaufort Sea, i.e. the Arctic Coastal Plain of northern Alaska and Yukon (Hopkins, 1982; Dinter *et al*., 1990). This area occupied the rainshadow of moisture sources in the Gulf of Alaska and the Bering Sea, and it is likely that pervasive sea-ice cover on the Beaufort Sea combined with a very low glacio-eustatic sea level minimized moisture supply from the north. These same barriers to moisture sources must also have affected the adjacent Tuktoyaktuk Coastlands. Thus the regional aridity along the Beaufort Sea coast during the Duvanny Yar interval was conducive to large-scale aeolian dune activity in areas with abundant, pre-existing sand deposits, including the Tuktoyaktuk Coastlands.

Phase II: Regional sand sheets

Between *c*. 14 and 8 ka aeolian sedimentation at the study sites was dominated not by dunes but by sand-sheet formation. This phase broadly coincides with a postglacial warm interval that, based on palynological and plant macrofossil evidence, lasted from *c*. 13,500 to 8,400 cal. years BP, with rapid warming between *c*. 13,500 and 10,900 cal. years BP (Ritchie, 1984). Ritchie (1984) estimated that mean July air temperatures by *c*. 10,900 cal. years BP were *c*. 3–5°C higher than modern values, and MacDonald (2000) estimated that between *c*. 10100 and 7800 cal. years BP they exceeded modern values by *c*. 1–3°C.
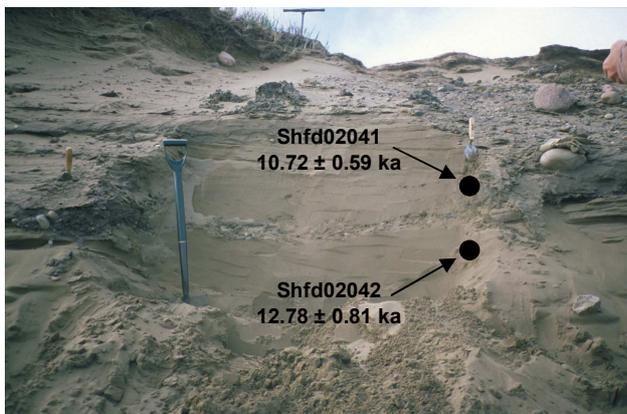

Fig 8: Gravelly meltwater unit above high-angle dune beds of the Kittigazuit Fm and beneath aeolian sand sheet, Hadwen Island.

The switch from dune-dominated aeolian activity to regional sand-sheet formation can be attributed to two causes:

- ***Surface armouring.*** As discussed, aeolian activity during phase I had an abundant sand supply reworked from pre-existing sandy substrates. But as a result of the Late Wisconsinan Toker Point glaciation (discussed below), the sandy surface of the Kittigazuit Fm and other sand units across substantial areas of the Tuktoyaktuk Coastlands became protected by cohesive till (Rampton, 1988), buried basal ice, glacitectonite or gravel (Murton, 2005; Murton *et al.,* 2004, 2005). Such protection is exemplified where Toker Point till overlies Kittigazuit dune deposits on northern Hadwen Island (Fig 7). Deglacial meltwater activity (proglacial flooding) caused significant erosion, reworking some of the Toker Point till to leave a discontinuous gravely lag (Fig 8; Murton *et al.,* 1997). Possibly, glacial meltwater activity itself may also have favoured sand-sheet aggradation, as inferred during the middle to late Holocene in aeolian sand sheets in front of the fluctuating margin of the western Greenland ice sheet (Willemse *et al.,* 2003). In short, the areas of bare sandy substrate exposed to aeolian activity must have diminished substantially, limiting the sediment supply.

- ***Climate change.*** Sand-sheet development appears to have centred on the time of the rapid increases in global temperatures (as shown by the GISP2 Oxygen isotope record) and at the summer insolation maximum (Fig. 6). Regionally this would have translated to increased seasonality and warming during the Last Glacial-Interglacial Transition. The warmest period, based on pollen and macrofossil data, was probably centred on *c*. 11,100 cal. yrs. BP (Ritchie *et al*., 1983). Retreat of the Cordilleran Ice Sheet (CIS), submergence of the Bering land bridge after 13,000 cal. yrs. BP (Saarnisto and Karhu, 2004), and a probable decrease in the extent of summer pack ice in the Beaufort sea would have led to a slight deterioration of regional climate-induced aridity. Large areas of Beringia experienced a rapid vegetation change *c*. 14,000 cal. yrs BP, attributed to a rapid increase in moisture and deeper winter snows (Edwards *et al.,* 2001; Ager, 2003). Thus aeolian transport may have been more limited by seasonal snow cover and/or sediments being more cohesive

through enhanced moisture and vegetation growth.

Although sand-sheet formation between *c.* 14 and 8 ka is inferred at the study sites, parabolic dunes formed elsewhere on the northern Tuktoyaktuk Peninsula, as inferred from the optical date of 8,600 ± 1,000 yrs obtained from one by Michaud and Bégin (2000). The occurrence of this dune type is consistent with a limited sand supply in a vegetated terrain.

Phase III: Regional stabilisation

Based on the absence of OSL dates and the occurrence of three of the radiocarbon dates, the period 8–1 ka was broadly one of sand-sheet stabilization and, at times, peat development. Active-layer deepening in response to climate warming reached its maximum at *c.* 9000 cal. years BP (Burn, 1997), after which gradual climate cooling commenced (Richie, 1984). Tree line advanced over 100 km northward of its present position in the Mackenzie Delta, with spruce trees growing on the Tuktoyaktuk Peninsula by at least *c.* 10,100 cal. years BP; tree line retreated south after 5,300 cal. years BP (Ritchie, 1984). MacDonald (2000) inferred that the region near Tuktoyakak was likely forested between *c.* 10,100 and 7,800 cal. years BP. Such vegetation growth would have helped to stabilize the sand sheets. Stabilization would also have been helped by moister climatic conditions. With the waning of the CIS, moist Pacific air could once again penetrate into the region from the southwest. Between *c.* 12,800 and 7,800 cal. yrs BP eustatic sea level rose by 25 m (Fairbanks, 1989). This, coupled with the re-establishment of the Beaufort Sea gyre, and reduced summer sea ice at this time (Andrews and Dunhill, 2004) would have increased the maritime influence on the climate of the Tuktoyaktuk Coastlands, even though the early Holocene coastline still lay several tens of kilometres to the north of the present coastline (Burn, 1997). With increased moisture would be the probability of enhanced seasonal snow cover, further limiting sediment supply (e.g. Lea, 1990). Evidence for enhanced moisture can been found in eastern Alaska, which shows precipitation similar to present between by 8,500 cal. yrs BP (Edwards *et al.,* 2001). Finally from 8 ka onwards the discharge of water from the Mackenzie River reduced as the LIS withdrew from its headwaters (Lemmen *et al.,* 1994; Dyke*,*

1996; Dyke *et al.,* 2003), reducing the annual sediment recharge to the Coastlands associated with spring melt.

Phase IV: Localised aeolian activity

The final phase of aeolian activity occurs in the period 700 to 9 yrs BP (Shfd02040 – Shfd02039). This latter phase forms localised sand sheets and nehbka dunes above some sandy bluffs. Whilst the preservation potential of these features is low, the lack of observed distinct intercalated sediment in the peat associated with phase III is interpreted as indicating an increase in aeolian activity. Interestingly, the two OSL dates obtained from near the bottom and top of the cliff-top sand sheet on northern Hadwen Island (Fig. 2b; 0.18 ± 0.2 ka – Shfd02039; 0.68 ± 0.05 ka – Shfd02040) almost completely span the period of cold, dry conditions between AD 1330 and the early 1900s (corresponding to the "Little Ice Age") inferred from tree-ring studies by Bégin et al. (2000) at a site 48 km southeast of Tuktoyaktuk. Further dating studies, however, are required to determine if aeolian activity was regionally more extensive during the Little Ice Age.

5.2 Timing of the deglaciation of the LIS in the Mackenzie Delta Area

The precise chronology and limits of the northwestern LIS are poorly understood and constrained (e.g. Dyke *et al.,* 2002). What is certain is that the Tuktoyaktuk Coastlands were glaciated extensively by the LIS during the last (Wisconsinan) cold stage, represented here by the Toker Point Stade, and that the southern part of the Coastlands was glaciated during the later Sitidgi Stade (Rampton, 1988). The age of the Toker Point Stade has been variously assigned to the Early Wisconsinan (>35 ka; Rampton, 1988), to sometime between 33.7 and *c.* 20 ^{14}C ka BP (Dallimore *et al.* 1997) and to as late as *c.* 17 – 14 ka (Mackay and Dallimore, 1992) or shortly before *c.* 14 ka (Murton *et al.,* 1997). As the sites sampled in the present study are either side of Rampton's (1988) Toker Point Stade ice limit (Fig. 1), the OSL data contribute to this discussion in a number of ways.

In many places, ice associated with the Toker Point Stade advanced directly across the dunes of the Kittigazuit Fm, depositing diamicton, forming stratigraphic discontinuities and/or

introducing clasts that have been reworked to form thin gravel lags within the aeolian sands (Figs 7 and 8). As the Kittigazuit Fm pre-dates the glacial advance, it is therefore significant that the average OSL age of the Kittigazuit Fm (as sampled) is 21.2 ± 8.7 ka BP. Whilst not directly dating the advance itself, this adds further weight to the argument for a Late Wisconsinan age for the Toker Point Stade.
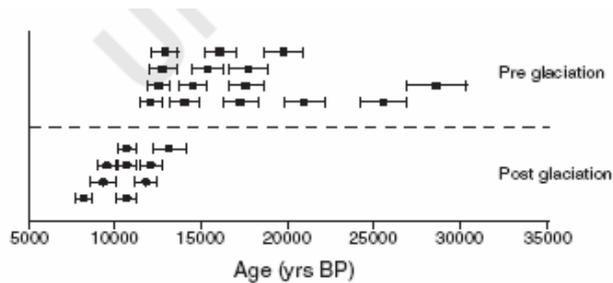

Fig 9: OSL data from stratigraphic horizons which predate glaciation and post-date it. Note temporal overlap between two groups.

The stratigraphic sequences dated did not contain till but contained erosional surfaces and gravelly units attributed to meltwater reworking of Toker Point till. Deglaciation was probably accompanied by proglacial flooding and erosion (Murton *et al.*, 1997). The bracketing dates therefore constrain the timing of meltwater pulses during deglaciation of the northwest LIS (Fig. 8). The temporal overlap in the regional OSL data shows (Fig. 9) that there was no clear regional cessation of sand deposition associated with the Kittigazuit Fm followed by glacial and meltwater sediments and then resumption of aeolian deposition. Instead, the timing of glacial re-advance/ meltwater pulses into the area was non-contemporaneous across the sites, perhaps reflecting a highly dynamic ice front in which meltwater pulses occurred at different points through time. Such glacial dynamism may well reflect the possible existence of palaeo-ice streams in the Mackenzie and Liverpool Bay areas (C.D. Clark, pers. comm. 2005). As Figure 9 shows, the post-meltwater resumption of aeolian activity in the form of sand sheets had occurred across much of the region by 12 ka. The Toker Point Stade must therefore have occurred sometime within the period 21 – 12 ka.

6. Conclusions

During the last 40–60 ka, palaeoenvironmental conditions induced by sea-level falls, ice-sheet growth and climate shifts caused regional aridity in the Tuktoyaktuk Coastlands in which large-scale dune formation and sand-sheet aggradation occurred. Sediment supply is thought to have been bountiful, with pre-existing Kidluit fluvial sand supplemented by sediments seasonally brought into the region by LIS meltwaters coming down the various rivers. At around 14 ka, dune formation gave way to sand-sheet aggradation as regional aridity waned, moisture and snow fall increased, and sediment supply became limited following the final LIS (Toker Point Stade) advance. The latter, on the basis on the OSL data, must date to the Late Wisconsinan in the period between 21 and 12 ka. Soon after the onset of the Holocene the aeolian sand sheets stabilised and the Boreal forest migrated northwards onto the Tuktoyaktuk Peninsula. Over the last few centuries localised aeolian activity has taken place once again.

Acknowledgements

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