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ORIGINAL PAPER

Sedimentary pellets as an ice-cover proxy in a High Arctic ice-covered lake

Jessica D. Tomkins · Scott F. Lamoureux · Dermot Antoniades · Warwick F. Vincent

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Abstract Sediment aggregates ("sedimentary pellets") within the sedimentary record of Lake A (83°00′ N, 75°30′ W), Ellesmere Island, Canada, are used to construct a 1000 year proxy record of ice-cover extent and dynamics on this perennially ice-covered, High Arctic lake. These pellets are interpreted to form during fall or early winter when littoral sediment adheres to ice forming around the lake's periphery or during summer through the development of anchor ice. The sediment likely collects in ice interstices and is concentrated in the upper ice layers through summer surface ice melt and winter basal ice growth. The pellets remain frozen in the ice until a summer or series of summers with reduced ice cover allows for their deposition across the lake basin. Sedimentary pellet frequency within multiple sediment cores is used to develop a chronology of ice-cover fluctuations. This proxy ice-cover record is largely corroborated by a record of unusual sedimentation in Lake A involving iron-rich, dark-orange to red laminae overlying more diffuse laminae with a lighter hue. This sediment sequence is hypothesized to represent years with reduced ice cover through increased chemocline ventilation and iron deposition. During the past millennium, the most notable period of inferred reduced ice cover is ca. 1891 AD to present. Another period of ice cover mobility is suggested ca. 1582–1774 AD, while persistent ice cover is inferred during the 1800s and prior to 1582 AD. The proxy ice-cover record corresponds well with most regional melt-season proxy temperature and paleoecological records, especially during the 1800s and 1900s.

Keywords Sedimentary pellets · Lake ice cover · Sedimentology · Varves · Meromictic lake · High Arctic

This is one of fourteen papers published in a special issue dedicated to reconstructing late Holocene climate change from Arctic lake sediments. The special issue is a contribution to the International Polar Year and was edited by Darrell Kaufman.

J. D. Tomkins (✉) · S. F. Lamoureux
Department of Geography, Queen's University, Kingston, ON, Canada K7L 3N6
e-mail: 0jdt@queensu.ca

D. Antoniades · W. F. Vincent
Centre d'Études Nordiques, Université Laval, Quebec City, QC, Canada G1V 0A6

Introduction

Perennially ice-covered lakes exist in only a few locations in the world, including the McMurdo Dry Valleys of Antarctica (Wharton et al. 1993), northern Greenland (Vincent et al. in press) and northern Ellesmere Island, Canada (Bradley et al. 1996). Such lakes require extremely low temperatures to develop and maintain thick ice covers. A permanent ice cover influences sediment input and deposition (Retelle and Child 1996), light penetration into the water column (Belzile et al. 2001), microbiotic structure (Belzile et al. 2001; Smol and Douglas 2007), water column stratification (Doran et al. 1996), and the preservation of sedimentary structures (Bradley et al. 1996).

Ice cover is influenced by melt-season temperatures and thus, changes in a perennial ice cover can reflect changing regional temperatures (Wharton et al. 1993). High-resolution reconstructions of past summer temperature variability in the Canadian High Arctic are limited to a small number of lacustrine (e.g. Lamoureux and Bradley 1996; Hughen et al. 2000) and ice core (e.g. Fisher and Koerner 1994) records. Many of these studies (e.g. Fisher and Koerner 1994; Hughen et al. 2000; Smith et al. 2004) reported notable warming trends during the past 100–200 years, but only one (Lamoureux and Bradley 1996) involved the analysis of sediments from a perennially ice-covered lake. Additionally, numerous paleoecological studies have identified diatom and invertebrate species shifts in Arctic lakes during the past two centuries (e.g. Michelutti et al. 2003; Antoniades et al. 2005; Smol et al. 2005; Antoniades et al. 2007; Smol and Douglas 2007) that have been attributed to habitat changes associated with climate warming and reductions in summer ice-cover extent.

While biological proxy records of ice-cover variability have been inferred from Arctic sedimentary sequences, reliable physical records of such changes have not yet been developed. Meromictic, perennially ice-covered Lake A, located on the northern coast of Ellesmere Island (Fig. 1), may be in transition from a perennially ice-covered lake to one with seasonal ice cover (Van Hove et al. 2006), and the repercussions of such a change could drastically alter the lake's stratification, sedimentation regime and aquatic communities. In this study, we use mineral grain aggregates ("sedimentary pellets"), an unusual

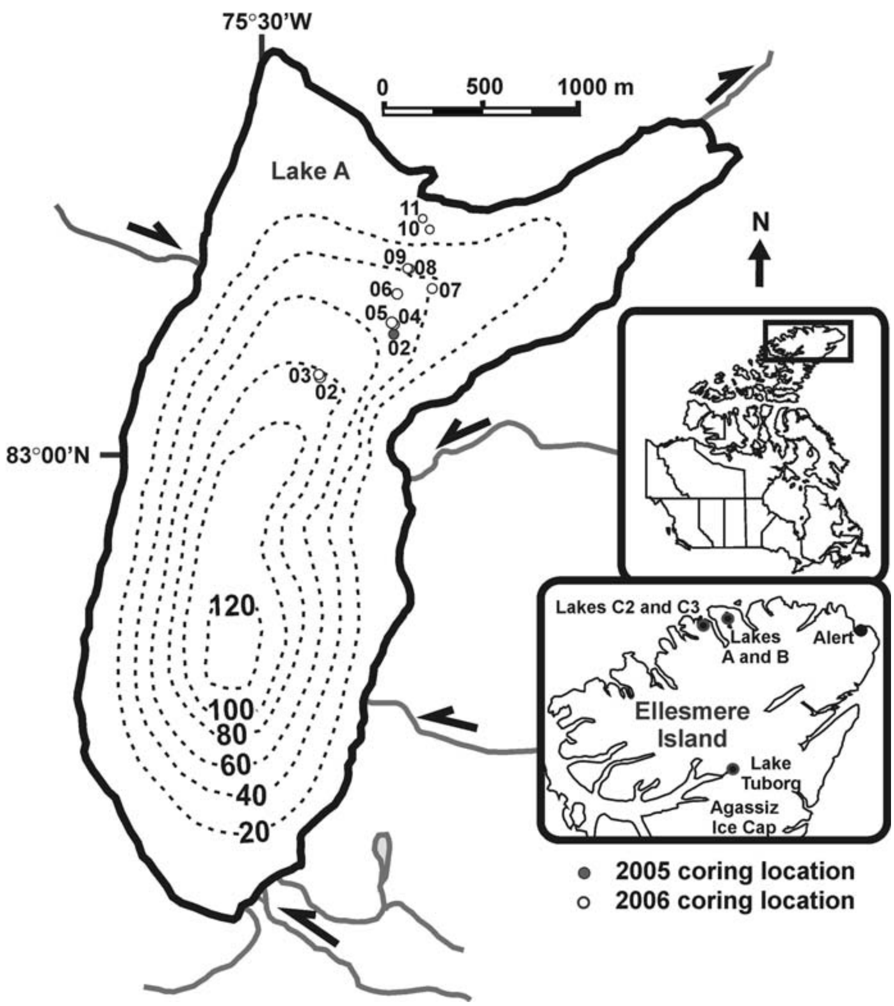

Fig. 1 Lake A bathymetry (contours in m) and core sampling locations. Inset map shows the locations of other paleoclimate records on Ellesmere Island discussed in text. Lake inflows and the only outflow are denoted by black arrows

feature of Lake A's sedimentary record, to develop a proxy record of relative ice-cover extent and, by extension, ice thickness and melt-season temperatures. This record is used to place recent changes into context and understand the history of ice cover variability at Lake A during the past millennium.

Study site

Lake A (unofficial name; 83°00′ N, 75°30′ W), is located ~4 m asl on the northern coast of Ellesmere Island, Nunavut, in Quttinirpaaq National Park (Fig. 1). This 4.9 km^2 lake lies in an unglacierized 37 km^2 catchment within the Challenger Mountains and is underlain by pyroclastic rocks with limestone inclusions of the M'Clintock Formation (Okulitch 1991). Frost-shattered felsenmeer covers much of the landscape and elevation in the catchment rises to 800 m asl. Herbaceous tundra vegetation is sparse and species diversity is low in this extreme environment (Edlund and Alt 1989). Meteorological records for the study area are available from Alert, Nunavut (175 km east; Environment Canada 2008), Parks Canada (1995–2005) and SILA Network (2005–2007; Centre d'Études Nordiques, Université Laval) weather stations on Ward Hunt Island (15 km northeast), and a weather station at Lake A (Centre d'Études Nordiques, Université Laval) that has been operational since 2004. Mean annual temperature and precipitation at Alert are −18°C and 154 mm, respectively (Environment Canada 2008). The melt season is confined to a brief period, mostly during July and August, and mean summer temperature (June-August) is 1.3°C (Environment Canada 2008).

Lake A formed between 2500 and 4450 BP (years before 1950 AD) when isostatic rebound raised the lake above sea level, but it may have been isolated earlier by the growth of the adjacent Ward Hunt Ice Shelf (Lyons and Mielke 1973; Jeffries et al. 1984; England and Stewart 1985). The lake has a maximum known depth of 128 m and is strongly stratified with an anoxic sea water monimolimnion containing hydrogen sulfide underlying a freshwater mixolimnion. A sharp oxycline (10–13 m) and more gradual chemocline (10–25 m) represent the transition zone between the two water masses. The lake has a perennial ice cover and typically forms a moat of open water (usually <25 m wide) along the shoreline during the melt season. While the ice cover has been previously recorded as 2 m thick (Hattersley-Smith et al. 1970; Jeffries et al. 1984; Lamoureux, unpublished data, 1993; Belzile et al. 2001), measurements in 2005 and 2006 revealed a thinner ice cover of 1.5 m ($n = 8$) and 1.3 m ($n = 11$), respectively. Additionally, an ice-off event and a reduced-ice summer were recorded in 2000 and 2003, respectively (D. Mueller, pers. commun. 2007). The ice cover limits wind-induced mixing during the summer, which aids the preservation of stratification in the lake, minimizes disturbance to sediments within the monimolimnion and provides conditions conducive to the formation and preservation of varves.

Methods

Sediment core collection

During spring 2005 and 2006, 21 sediment cores were collected from Lake A using an Aquatic Research Instruments gravity and percussion coring system. Eleven of these cores were analysed for this study (Fig. 1). Sediment cores were kept unfrozen in a vertical position during storage and transport. Cores collected in 2006 were capped with a sodium polyacrylate gel seal to prevent surface disturbance during transport (Tomkins et al. 2008).

Sediment core processing

Sedimentological analyses

All sediment cores were split lengthwise in the laboratory and their surfaces were cleaned with a razor blade. Thin sections were prepared from overlapping sediment slabs (7.0 × 1.5 × 0.3 cm) following the methods of Lamoureux (1994, 2001). Thin sections were scanned at 2400 dpi using a Hewlett-Packard S20 slide scanner for examination using a digital imaging program. They were also assessed using standard light microscopy. Lamina thickness measurements were made using light microscopy (dissecting microscope) and a Quick-Check QC-1000 measurement system mounted on an Acu-Rite Absolute Zero II precision measurement stage. Dr. Raymond Bradley provided access to thin sections created from the sedimentary records of

Lakes A, B, C1, C2 and C3 for the Taconite Inlet Lakes Project (Bradley et al. 1996; Lamoureux and Bradley 1996) for light microscope observations. Sedimentary pellets were counted throughout cores A-02-05 and A-02-06 to A-11-06 on core faces and on thin sections, using a light table.

Grain size was determined for sedimentary pellet material from cores A-02-06 to A-10-06 and the surrounding sediments in core A-04-06 (Table 1). Material from visible sub-rounded sedimentary pellets was sampled to obtain enough material for one grain-size measurement and core A-04-06 was sampled at 1 mm intervals. Light microscopy analysis of the sedimentary pellets indicated the same pellet structure, texture and general composition in all cores and, therefore, combining material from different pellets for the grain size and total organic carbon (TOC) and total nitrogen (TN) measurements was deemed appropriate, given sample-size requirements of the instruments. Core A-11-06 did not have easily discernable sedimentary pellets on the core face for sampling and, thus, pellets from this core were not included in the grain size or TOC and TN measurements. All grain-size samples were treated with repeated applications of 35% hydrogen peroxide at 40°C for up to 3 weeks to completely remove organic material. Prior to measurement, 1–2 ml of sodium hexametaphosphate (38 g l^{-1}) and sodium carbonate (8 g l^{-1}) solution were added to each sample to disperse aggregates during measurement in a Beckman Coulter LS200 grain size analyser. Each sample was measured three successive times with continuous sonication. The third run was retained unaveraged after visual comparison with the other two runs for quality assurance.

Table 1 Particle size, total organic carbon (TOC), total nitrogen (TN), and C:N ratio for the pellets and their surrounding sediment (0–230 mm depth in core A-04-06)

	Sedimentary pellets	Surrounding sediment
Grain size		
Mean (μm)	11.1	7.0
Standard deviation	–	1.9
Median (μm)	15.1	7.8
Clay (%)	16.4	20.8
Very fine silt (%)	22.9	33.5
Fine silt (%)	13.1	19.2
Medium silt (%)	17.4	14.7
Coarse silt (%)	26.9	11.2
Total silt (%)	80.4	78.6
Sand (%)	3.3	0.6
Organic content		
TOC (%)	1.5	1.9
Standard deviation	–	0.2
TN (%)	0.1	0.2
Standard deviation	–	0.0
C:N (atomic) ratio	28.5	15.3
Standard deviation	–	2.4

Total organic carbon and total nitrogen analysis

Total organic carbon (TOC) and total nitrogen (TN) were measured in sedimentary pellet material and throughout the surrounding sediments in core A-04-06 (0.5 cm increments in the upper 3 cm, 1 cm increments below 3 cm) using a Leco CNS-2000 analyser (Table 1). A single sample of material from sedimentary pellets in cores A-02-06 to A-10-06 was collected from the pellet material remaining after grain size sampling. Following standard methods (Meyers and Teranes 2001), all samples were treated with dilute (2 N) hydrochloric acid overnight to dissolve carbonates and centrifuged four times with distilled water to remove the acid. The samples were then freeze-dried and crushed before measurement.

Radioisotope geochronology

Contiguous samples (3 cm^3) were collected from core A-02-05 at 0.5 cm intervals to 6 cm depth for ^{137}Cs determinations. The samples were dried in an oven (60°C) for 24 h and crushed before ^{137}Cs activity was determined by measuring each sample for 80,000 s in an ORTEC high-resolution gamma-counting system, following the methods of Appleby (2001). Samples (0.5 cm^3) were also collected at increasingly larger depth intervals to 8.5 cm for unsupported ^{210}Pb activity. After drying in an oven (60°C) for 24 h, the samples were crushed and sent to MyCore Scientific Inc. (Deep River, Ontario) for ^{210}Pb decay measurements using alpha spectroscopy. Unsupported ^{210}Pb activity was converted into ages within the sedimentary record by using the constant rate of supply (CRS) dating model (Appleby and Oldfield 1978).

Due to a paucity of organic macrofossils, bulk sediment samples were collected from five locations in core A-04-06 for radiocarbon dating. The samples were dried and sent to KECK Carbon Cycle AMS Facility, University of California, Irvine, for accelerated mass spectrometry (AMS) radiocarbon dating and all resulting radiocarbon dates were calibrated using OxCal 4.0 (Bronk Ramsey 1995).

Results[1]

Core sedimentology

Sedimentary facies

The longest, uninterrupted sedimentary records in Lake A are found within the northern portion of the lake at water depths <54 m. This area of the lake does not have slumps like deeper areas with steeper slopes do. Sedimentation rates in Lake A are extremely low (e.g. 0.2 mm $year^{-1}$ in core A-04-06) and the sediment is composed mostly of clay and silt with biogenic components (Tomkins 2008). All cores from this part of the lake contain four sedimentary facies, except for the three shallowest, most distal cores.

The bottom facies, Facies 1, includes massive marine clay and silt (mean grain size: 5.3 µm) and fossil foraminifera, and is inferred to represent sedimentation prior to the lake's isolation from the Arctic Ocean (Tomkins 2008). Facies 2 has a fine texture (mean grain size: 5.6 µm) and higher levels of TOC than other facies (mean: 2.25%). It contains diffuse, wavy laminae, and ovoid clay structures, both possibly from biological processes (Tomkins 2008). Facies 3 has the coarsest mean grain size (mean: 8.1 µm), and represents a transition from diffuse structure to clearly defined microlaminae (Tomkins 2008). This facies contains coarse units interbedded with units containing the ovoid clay structures also noted in Facies 2, and the deepest sedimentary pellets are present within this facies (at 230 mm depth in core A-04-06). The uppermost sediments (0–179 mm in core A-04-06) of the sedimentary sequence form Facies 4, which contains microlaminae, some biogenic deposits and sedimentary pellets (Fig. 2). The microlaminae are typically composed of a silt unit with a clay cap. Mean grain size in this facies is 6.8 µm and TOC is low at 1.9% (Tomkins 2008). Laminae within Facies 4 most commonly show an orange hue (7.96YR 4/8, wet sediment colour), although some laminae are shades of brown (5.01Y 4/4) or have minimal colouring (8.76Y 7/4). Full details of the lake's sedimentology are available in Tomkins (2008).

Sedimentary pellets

Ovoid sediment aggregates, or "sedimentary pellets," of different texture and colouring than the surrounding sediment, are apparent within the uppermost sediments of cores collected from water depths ranging from 18 to 81.5 m (Fig. 2). The sedimentary pellets are also evident in cores from deeper locations in the lake (up to 126 m) collected in 1993 (S. Lamoureux and R. Bradley, unpublished data). Macroscopic pellets (those visible to the unaided eye) are the focus of this study but microscopic pellets were observed at up to 63 × magnification. The smallest pellets are sub-millimetre in size, whereas the largest specimen is 9 × 4 mm (cross-sectional dimensions). Most pellets are composed of grey (1.24 GY 5/0) clastic grains with some black granules and fine detrital organic material. Although based on only one sample, the grain-size analysis results show the pellets to be notably coarser (11.1 µm) than the surrounding sediment (7.0 µm) and to contain higher amounts of medium and coarse silt and sand. They also have lower TOC and TN values and a higher carbon to nitrogen (C:N) ratio (Table 1).

The long axis of each sedimentary pellet lies parallel to laminae horizons and the pellets often warp the sediment below them, sometimes forming load structures. The pellets are soft but appear to be cohesive. Most pellets are solitary, but sometimes multiple pellets are found within the same sedimentary unit. Additionally, pellets within the same unit are sometimes connected by a thin band of the pellet material. Most pellets appear to have been deposited within the coarse unit of each lamina couplet.

While most cores contain sub-rounded sedimentary pellets, those from shallower areas (e.g. core A-11-06,

[1] The pellet-frequency data are available from the World Data Center for Paleoclimatology (ftp://ftp.ncdc.noaa.gov/pub/data/paleo/paleolimnology/northamerica/canada/ellesmere/lake-a2008.txt).

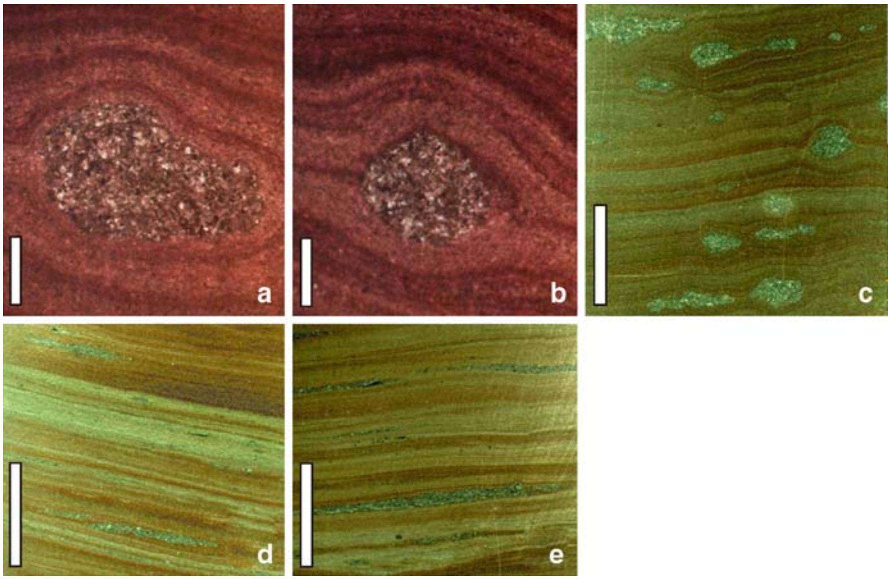

Fig. 2 Examples of pellets within the Lake A sediment. **a** and **b** Larger solitary pellets in core A-04-06 (scale bar = 1 mm). **c** Solitary and chains of pellets in core A-06-06. **d** Flattened pellet material in core A-10-06, and **e** in A-11-06. Scale bars in **c**–**e** are 5 mm in length. Scratch marks on **c** and **e** are artefacts of thin-section preparation

11 m water depth) have pellet material with lensoid profiles that occasionally form nearly continuous units across thin sections (Fig. 2). Additionally, cores A-10-06 (18 m water depth) and A-08-05 (14 m water depth) contain both solitary pellets and chains of pellets, but nearly all have a more flattened profile than pellets in cores from deeper locations (Fig. 2).

Sedimentary chronology

Verification of lamina couplets as varves

Microlamina couplets in Facies 4 typically are composed of a silt unit with a clay cap, which is a simple varve structure. The inferred varve record was independently verified using ^{137}Cs and ^{210}Pb dating. Both radioisotopes are present in extremely low levels within the Lake A sediments (^{137}Cs peak: 0.007 Bq $\cdot$ g^{-1}, ^{210}Pb peak: 0.13 Bq $\cdot$ g^{-1}), which limits the number of dates that could be estimated from the radioisotope decay profiles. However, the profiles developed from Lake A's sediments are adequate for providing age estimates of the uppermost sediments. Due to the presence of disturbed laminae that could not be counted or measured accurately within the uppermost sediments of core A-02-05, the results of the radioisotope measurements were transferred to core A-04-06 using marker beds.

No prominent peak in ^{137}Cs is observed within the sediments to suggest the year of peak ^{137}Cs production (i.e. 1963 AD; Appleby 2001) but the highest activity occurs at the same depth as the estimated 1963 lamina couplet (Fig. 3). These results indicate that the past 40 years of sedimentation are within the uppermost 1.3 cm of core A-04-06. ^{210}Pb dating produced three dates for the sedimentary record and two of these dates (2005 and 1969 AD) correspond with lamina counts (Fig. 3). The one date that did not correspond (1994 AD) is offset by 1 mm. This discrepancy could be caused by transferring the ^{210}Pb results from core A-02-05 to A-04-06. The results indicate that sedimentation since 1969 AD is contained within the uppermost 1.2 cm of core A-04-06, which is consistent with the ^{137}Cs results. The results of the radioisotopic analyses support the interpretation that the microlaminae in Facies 4 are varves and sedimentary structures below the radioisotope-dated sediments suggest that varves continue throughout the facies. A varve-thickness record has been developed from Lake A (Tomkins 2008), but its development and interpretation are beyond the scope of this paper.

Surficial sediments were radiocarbon dated to 3426 cal year BP (3401–3450 calibrated years before 1950 AD; UCIAMS-41208). This reservoir age suggests that radiocarbon ages lower in the sedimentary profile are also affected by old carbon. As an estimate of more realistic ages, the reservoir age was subtracted from radiocarbon age at 8.2–8.4 cm and 13.3–13.4 cm depth, and the resulting ages are

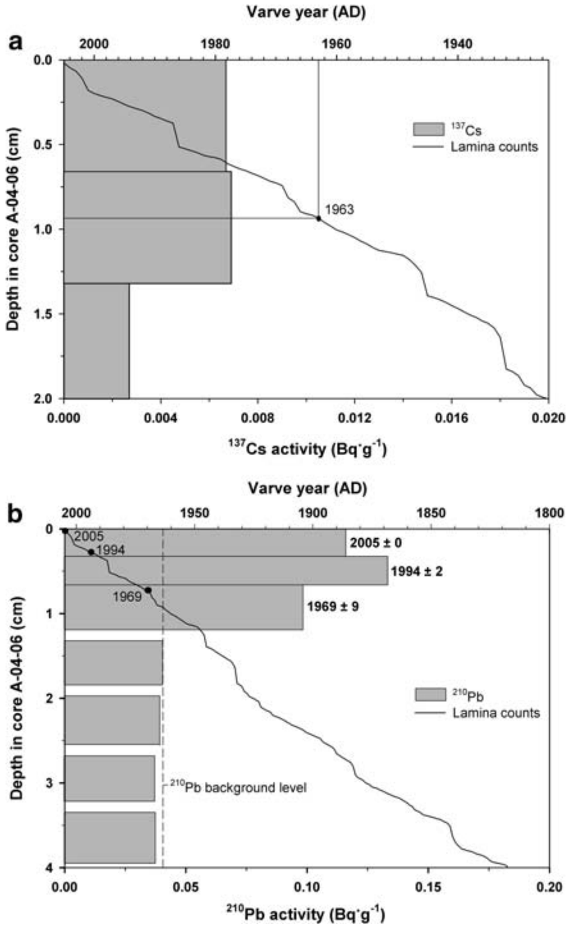

Fig. 3 **a** ^{137}Cs and **b** ^{210}Pb activity and lamina ages with depth in the Lake A sedimentary record. Radionuclide decay was measured in core A-02-05 and lamina counts are from core A-04-06

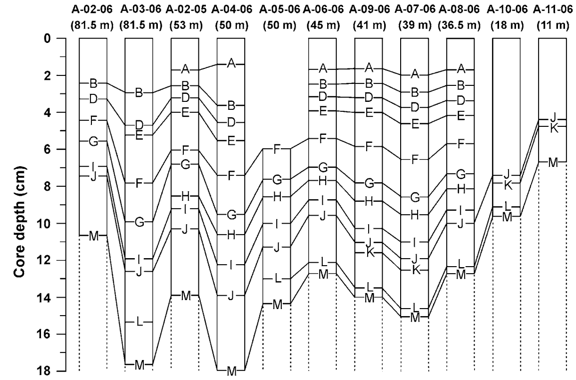

Fig. 4 Marker beds within the varved section of Lake A sediments. The cores are arranged from deepest to shallowest (water depths listed under core identification numbers). Marker bed C was eliminated due to its proximity to marker bed B. The upper 6 cm of core A-05-06 is missing because it was sampled to this depth in the field. The tops of A-02-06 and A-03-06 had disturbance that prevented identification of marker bed A

approximately 990 and 3070 years older than ages estimated using varve counts. The radiocarbon ages from the microlaminated section of the sedimentary record were deemed unreliable due to the strong reservoir effect in the lake and could not be used to support the varve chronology (Tomkins 2008).

Age-depth model construction

Using varve counts and following the methods of Sprowl (1993), an age-depth model was developed for the Lake A sedimentary record. The extremely low sedimentation rate increases the likelihood of errors in counting due to missing varves and, thus, varve counts in each core are considered to be minimum ages. By using multiple cores and common marker beds, missing varves in one core could be accounted for through other cores' varve counts. This method accounted for difficulties that arose in correlating laminae on an interannual basis.

Common marker beds in Facies 4 and the upper part of Facies 3 were identified within cores A-02-06 to A-11-06, and assigned identifying letters (Fig. 4). These distinct units were cross-correlated among all cores collected from the northern part of the lake, except in the two shallowest cores whose marker beds were correlated only at depth (Fig. 4). Low sedimentation rates precluded cross-correlation at an interannual scale but general trends in lamina thickness between marker beds are evident among the cores (Fig. 5). As such, cores were selected for the age-depth model based on varve clarity and the presence of marker beds. Marker bed C was eliminated because of its close proximity to marker bed B. Varves were counted and measured between the marker beds three to eight times to ensure the consistency of counts. Varve counts were stopped at the depth where discrete couplets could no longer be delineated with confidence (marker bed M). The shallowest two cores were excluded from the age-depth model development because they are missing common marker beds above marker bed J.

Summary statistics were calculated for each marker bed section and the maximum varve count per section was retained as the best estimate of the number of varves (Table 2). The value of two standard deviations from the varve count estimate

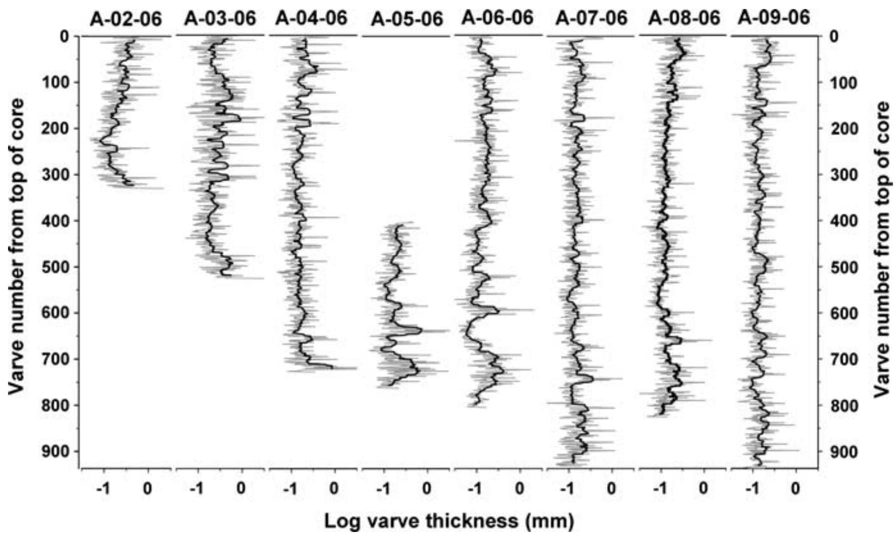

Fig. 5 Skeleton plot of varve thicknesses (log-transformed to enhance display) down to marker bed M with 15-year unweighted moving means from cores A-02-06 to A-09-06. Due to vague structure between marker beds J and M, varve thicknesses from core A-04-06 extend only to marker bed J. Core A-05-06, missing its uppermost 6 cm, is aligned at its top with core A-04-06 using marker bed F

Table 2 Lamina counts using the clearest and longest varve records from Lake A (after Sprowl 1993)

Core ID and water depths (m)	Marker bed											
	A	B	D	E	F	G	H	I	J	K	L	M
A-04-06 (50 m) varve counts[a]	103	62	48	58	131	139	67	87	33			
Repetitions	6	5	6	6	7	7	5	5	5			
A-09-06 (41 m) varve counts[a]	79	65	44	61	132	131	74	104	54	51	103	39
Repetitions	3	3	3	3	3	3	3	3	4	3	3	3
A-07-06 (39 m) varve counts[a]	115	61	51	55	123	138	75	99	39	49	93	34
Repetitions	6	8	6	6	4	4	5	4	3	3	5	3
Number of cores used	3	3	3	3	3	3	3	3	3	2	2	2
Maximum count	115	65	51	61	132	139	75	104	54	51	103	39
Minimum count	79	61	44	55	123	131	67	87	33	49	93	34
Mean count	99	63	48	58	129	136	72	97	42	50	98	37
Standard deviation	18.3	2.1	3.5	3.0	4.9	4.4	4.4	8.7	10.8	1.4	7.1	3.5
Percent deviation	15.9	3.2	6.9	4.9	3.7	3.1	5.8	8.4	20.0	2.8	6.9	9.1
95% CI[b]	36.7	4.2	7.0	6.0	9.9	8.7	8.7	17.5	21.6	2.8	14.1	7.1
Cumulative number of laminae[c]	115	180	231	292	424	563	638	742	796	847	950	989
Cumulative deviation at 95% CI	37	41	48	54	64	72	81	99	120	123	137	144
Estimated error (%)	32	23	21	18	15	13	13	13	15	15	14	15

[a] Counts represent the number of varves between marker beds, starting with the top of the core to the base of marker bed A. Water depths for each core are also listed

[b] Number of laminae representing the 95% confidence interval (CI) based on two standard deviations

[c] Based on the maximum number of counts, assuming that all counts are minimums

represents the 95% confidence interval for each section's counts. By progressively summing the confidence interval values, cumulative errors were assigned to each marker bed level. To minimize potential errors due to differing sedimentation, the three records with the longest, clearly defined varved sections (e.g. A-04-06, A-07-06, and A-09-06) were used to develop the final age model. The highest level of error in this age-depth model is above marker bed D (21–32%) but at the deepest marker bed, the cumulative error is reduced to 15% at the 95% confidence level (Table 2). This value is comparable with the estimated counting error from Elk Lake, Minnesota (12%; Sprowl 1993), but higher than counting errors from other Arctic lakes (e.g. Lake C2 (0.4–10.6%); Lamoureux and Bradley 1996). The resulting Lake A age-depth model covers approximately 1000 years of sedimentation (Fig. 6).

Sedimentary pellet and distinct sedimentation pattern frequency records

Using the age-depth model, an annual sedimentary pellet frequency record (i.e. the number of pellets per year within each marker bed interval) was developed for each core with a varve count record (Fig. 7). Pellet frequency increases at the top of most cores (ca. 1891 AD to present) and ca. 1017–1209, 1264–1367, and 1582–1775 AD. Pellet frequency is reduced in most cores ca 1368–1442 and

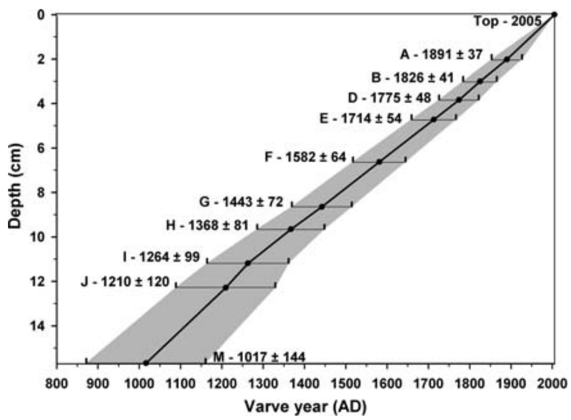

Fig. 6 Age-depth model for the Lake A sedimentary record from cores A-04-06, A-07-06 and A-09-06, including cumulative error (95% confidence interval) for the age at each marker bed (denoted by letters). Depths represent the mean depth of each marker bed from the three cores used in the age-depth model

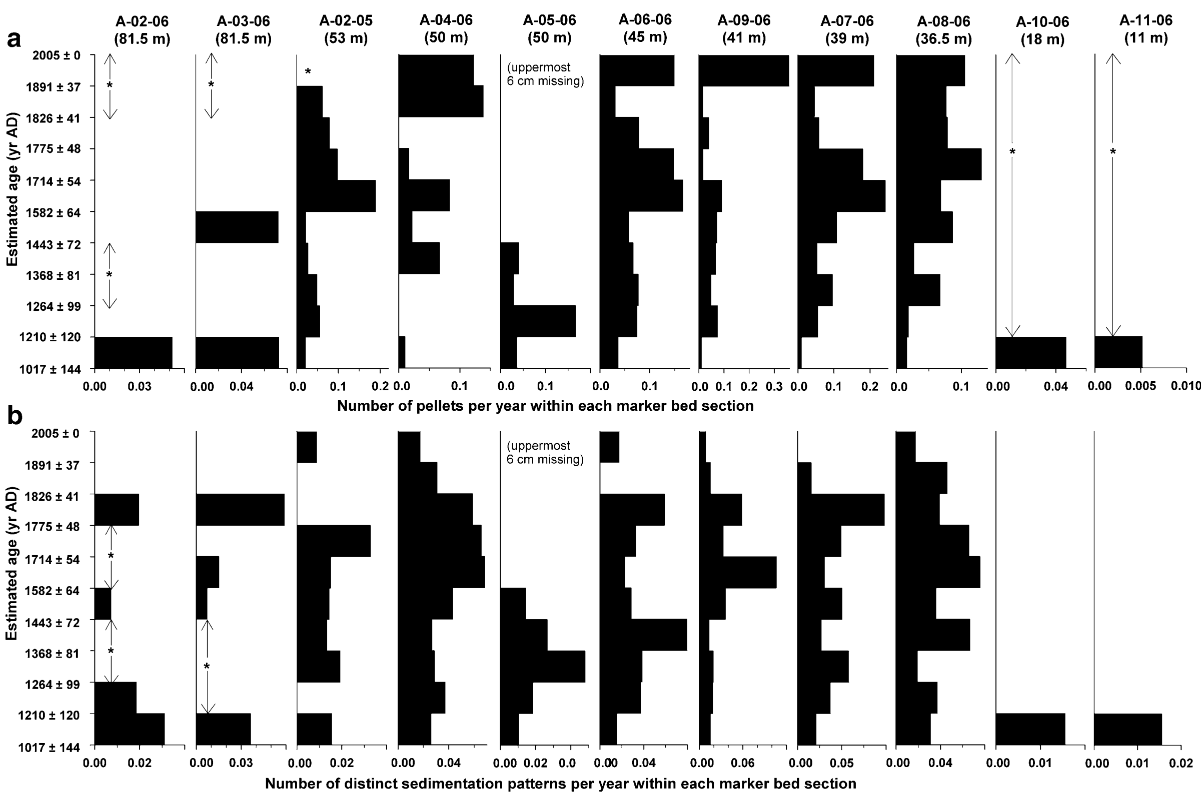

Fig. 7 **a** Annual pellet frequency between marker beds for each core examined. Asterisks (*) indicate pellets are present but disturbed sediment or a missing marker bed prevents the counts from being listed. **b** Annual frequency of the distinct sedimentation pattern of prominent dark-orange laminae overlying laminae with reduced pigmentation. Cores are in order from deepest to shallowest and estimated ages for marker beds and corresponding error estimates are on the y-axis. Note the different x-axis scales

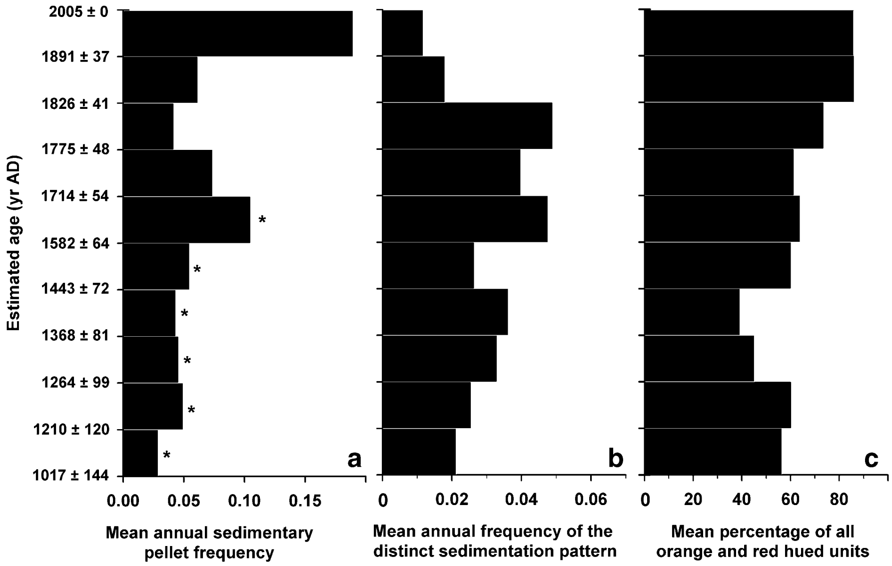

Fig. 8 **a** Mean annual frequency of macroscopic pellets on core faces and thin sections using all cores studied (A-02-05 and A-02-06 to A-11-06). No large pellets are visible below 230 mm in any core. Asterisks (*) indicate large pellets (>1 mm in height) within the section. **b** Mean annual frequency of the pattern of prominent dark-orange laminae overlying laminae with reduced pigmentation, and **c** mean percentage of orange or red-hued units per marker bed interval from all cores studied. Bars represent intervals between marker beds. Estimated ages for each marker bed, derived from the varve-based age-depth model (Fig. 6), are shown with estimated cumulative error

1775–1890 AD. The records were pooled to form a composite record of mean annual pellet frequency for all cores examined (Fig. 8). The highest pellet frequency of the past millennium occurred from ca. 1891 AD to the present, with another period of high prevalence ca. 1582–1775 AD (Fig. 8). Relatively low pellet frequency was notable prior to ca. 1582 AD and from ca. 1775–1890 AD.

Iron precipitates are hypothesized to form in Lake A during strong mixing events that bring oxygen to the upper anoxic waters (Tomkins 2008). To evaluate the frequency of these events, each core was examined for a distinct pattern of dark-orange to red-hued laminae overlying pale, diffuse laminae (i.e. the number of distinct sedimentation patterns per year in each marker bed section) (Fig. 7). These distinct pattern records are not as spatially coherent as the pellet frequency records, but some similarities among the cores were notable. Increased frequency of this unusual sedimentation pattern occurred between ca. 1582 and 1825 AD.

As an additional indicator of the frequency of iron deposition, the percentage of orange or red-hued laminae within each marker-bed section was tallied (i.e. the total number of orange and red laminae compared to all laminae within a given marker-bed section). All orange- and red-hued laminae were included, whether or not they were part of the distinct sedimentation pattern. This was then compared with the mean annual frequency of the distinct sedimentation pattern (i.e. the mean number of distinct patterns per year in each marker-bed section for all cores examined). The two records notably diverge after 1826 AD, when the former record reaches its highest values but the frequency of the unusual sedimentation pattern decreases (Fig. 8).

Discussion

Sedimentary pellet deposition

Sedimentary pellets are widely distributed across Lake A, as evidenced by their presence in cores from various depths. Thus, there are a number of possible transport mechanisms for the pellets. While no single pellet-origin process can be determined at this point, some can be effectively eliminated with the available data.

The sedimentary pellets are not fecal pellets. The sediments from many lakes, including nearby Lake C2 (Zolitschka 1996) and the Beaufort Lakes of eastern Ellesmere Island (Retelle 1986), contain copepod fecal pellets that are described as being mostly inorganic, but they are much smaller (100–500 μm; Smith and Syvitski 1982; Zolitschka 1996) than the Lake A sedimentary pellets. Moreover, fecal pellets may disaggregate while settling through relatively deep water columns (cf. Ferrante and Parker 1977). Based on the available evidence, the Lake A pellets appear unlikely to be derived from copepod activity.

Another possible pellet-formation mechanism involves the aeolian or fluvial transport of hardened aggregates of lacustrine or marine deposits eroded from desiccated sediments (Knight 1999, 2005). These soft sediment clasts could be transported onto a lake's ice cover to become pellets. However, these clasts are commonly described as being much larger (4.5–13 cm long; Knight 1999) than the sedimentary pellets in Lake A, which are also soft and easy to disaggregate.

Some Antarctic perennially ice-covered lakes have a large aeolian component to their sedimentary records due to sediment movement through the ice cover (Nedell et al. 1987; Squyres et al. 1991), and similarly, some Arctic lakes have substantial aeolian contributions to their ice covers that would likely influence the sedimentary record (McKenna Neuman 1990). In Arctic sedimentary records, individual coarse grains (often sand) within finer lacustrine sediments have been inferred to be of aeolian origin (Retelle 1986; Lamoureux 1999; Lamoureux et al. 2002). Individual coarse grains are not common within the Lake A sediments and there was little sediment within the winter snow cover during the 2005 and 2006 field seasons.

Additionally, aeolian sediment transport can lead to the formation of cryoconite holes or similar liquid water inclusions in ice that have a sediment base on which microbial activity may develop (Priscu et al. 1998; Fountain et al. 2004). Cryoconite holes have not been extensively studied in lake-ice covers, but in glaciers, these features have typical diameters ranging from 5 to 145 cm (Säwström et al. 2002; Fountain et al. 2004; Mueller and Pollard 2004). The size of these observed cryoconite holes suggests sediment inclusions, if present, would be larger than the typical pellets at Lake A and no cryoconite holes have been observed in the ice cover of Lake A. Sediment accumulation in the ice cover due aeolian depositional processes remains a possible pellet-origin process at Lake A, but there is limited evidence to support this hypothesis.

Ice-rafting processes are the most likely transport mechanism for the sedimentary pellets, based on the sedimentology of the pellets. Sediment transport via iceberg and sea-ice movement has been recognized as an important process in glacimarine environments (Ovenshine 1970; Gilbert 1990) and in some Arctic lacustrine settings, particularly ice-marginal environments (Smith 2000). Ice-rafted deposits range from coarse diamictons (Smith 2000), drop stones and frozen sediment aggregates (Gilbert 1990) to sedimentary pellets (Ovenshine 1970).

In the ice-rafting hypothesis, littoral sediment could become incorporated into the lake-ice cover by either anchor-ice formation or ice cover contact with the lake bottom. A moat of open water $\sim$25 m wide forms along the shoreline in summer at Lake A. In July 2007, anchor ice was observed to form on the lake bottom within the moat area. Anchor ice incorporates sediment when frazil ice crystals attach to lake bottom sediments and when the ice becomes buoyant, it transports sediment to the ice cover (Reimnitz et al. 1987; Gilbert 1990). Alternatively, if ice cover during fall and winter comes into contact with littoral sediments, sediment could adhere to the base of the ice and eventually be transported throughout the lake, although a layer of sediment, rather than small aggregates, would be likely be initially collected (Smith 2000).

With each of these suggested ice-rafting processes, the sediment would be concentrated in the upper layers of the ice cover through summer surface ablation and winter basal ice growth (Gilbert 1990; Smith 2000). During subsequent high melt years, mobile ice pans would form and release the sediment across the lake. Additionally, the lensoid form of pellets found at shallow depths could be due to sediment first adhering to the ice as a layer (if it were in contact with bottom sediments) and then detaching and settling out in shallow locations. The remaining sediment on the base of the ice cover would be

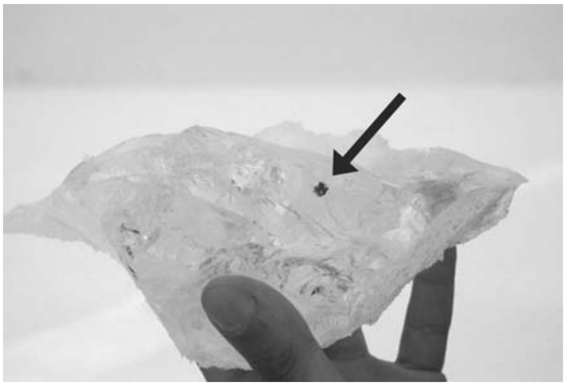

Fig. 9 Pellet-like sediment (indicated with an arrow) within Lake A ice during spring 1999 (reproduced with permission of W.F. Vincent)

incorporated into ice interstices to later form pellets. The cohesion of the sedimentary pellets through the water column could be due to release in a frozen state (cf. Gilbert 1990). Alternatively, the composition of the pellets, including lithic grains possibly coated with a biofilm, suggests that a littoral origin for the pellets could provide the biogenic components of the pellets. Moreover, sediment aggregates within the ice cover's interstices may further act as substrates for microbiota (Priscu et al. 1998), which could aid in maintaining pellet cohesiveness.

Sediment aggregates were observed in Lake A's ice cover (Fig. 9), and Belzile et al. (2001) reported pellets up to 2 cm in diameter within the upper 1 m of the 2 m-thick, multi-year ice during spring 1999 (C. Belzile, pers. commun. 2008). During subsequent reduced ice-cover years, mobile ice pans would release the accumulated sediment to various parts of the lake. Hence, the abundant small pellets near the surface of many cores (Fig. 7) may have resulted from the 2000 AD near ice-off event, when the pellets observed within the multi-year ice cover in spring 1999 would have been released.

The proxy ice-cover record

Limitations of using sedimentary pellets as a proxy for ice cover

As pellet deposition in Lake A is presumably dependent on ice-rafting during high melt years, the frequency of pellets within the sedimentary record can be used as a proxy indicator of past ice conditions. However, the ice cover may not reflect summer climatic conditions in a direct or proportional manner. For example, the warmest year on record at Alert (1998; Environment Canada 2008) did not result in an ice-off event or even notably fragmented ice pans (D. Mueller, pers. commun. 2007). Similarly, during the near ice-off year (2000), summer mean temperature at Alert was the seventh highest on record, spring mean temperature was average and the previous winter mean temperature was below average (Environment Canada 2008). However, it is possible that the high temperatures of 1998 may have preconditioned the ice cover for subsequent melt out in 2000 by altering ice porosity, density, thickness or surface albedo. These factors, in addition to ice type (i.e. first year or multi-year), initial snow cover, and melt rates at the base of the ice cover, may have contributed to the stability and longevity of the ice (Heron and Woo 1994). Further, the high thermal inertia of this lake prevents individual warm years from completely melting the ice cover. The pellet-frequency record cannot therefore provide an accurate representation of ice cover on an annual basis but offers a relatively direct indication of past ice-cover variability and an indirect estimate of melt-season temperatures on a longer timescale, likely decadal.

The temporal resolution of the mean annual pellet-frequency record, due to the summing of the pellets among marker beds, is a limitation. However, the age-depth model provides a much longer and generally higher-resolution chronology than other common dating methods, such as radioisotopic dating (e.g. ^{14}C). The verified varve chronology could not be used to develop an annual pellet-frequency record due to extreme thinness of the Lake A varves. The use of marker beds to constrain approximate 100 year periods allowed for the maximum number of cores to be included in the pellet-frequency record. Some cores with identifiable marker beds but diffuse varves would not have been incorporated into the composite record if an annual timescale had been used.

A third limitation of pellet frequency as a proxy for ice-cover conditions involves the possibility of no pellets being deposited during prolonged periods of relative warmth. As several years would potentially be required for sedimentary pellets to form in the ice cover, the pellet-frequency record assumes that there were no extended periods of seasonal ice cover on Lake A during the past millennium. Regional proxy records of summer temperature suggest that temperatures during the past century are the highest of the past millennium (e.g. Koerner and Fisher 1990), and indicate that conditions were conducive to the maintenance of a perennial ice cover on Lake A throughout the past 1000 years. The lake has been observed with a perennial ice cover in all but one year since observations began.

Comparisons of the proxy ice-cover and iron-rich-lamina records

Variations in sedimentary pellet frequency suggest a relatively persistent ice cover during the early part of the record (ca. 1017–1581 AD), followed by reduced ice cover until ca. 1774 AD (Fig. 8). More persistent ice cover is indicated during ca. 1775–1890 AD, and after ca. 1891 AD, pellet frequency reaches its highest level of the past millennium, although this is largely due to the presence of numerous small, individual and clustered pellets at the top of many sediment cores (Fig. 7). These uppermost pellets likely represent deposition after the near ice-off event of 2000 or reduced ice cover of 2003. Other pellets are evident within the twentieth century sediments, including a cluster of pellets in core A-04-06 corresponding to sedimentation during the 1930s. This was a period of known warmth in the Arctic (Serreze et al. 2000) and was likely a time of substantial Ellesmere ice shelf break-up (Vincent et al. 2001). Additionally, temperature profile modelling in Lake A suggests reduced ice conditions during this time (Vincent et al. 2008). Difficulties in assigning exact ages to individual pellets prevent a more detailed investigation of twentieth-century pellet deposition. However, the pellets at the top of the sedimentary record dominate the pellet-frequency record from 1891 AD to present, and suggest that either the most recent changes in ice cover are more strongly recorded within the sedimentary record or are more substantial than those from the 1930s.

Iron-rich units may be indicative of times with low ice cover and increased wind-induced mixing, which can lead to pyrite formation in meromictic lakes (Dickman 1979; Tomkins 2008). These units are identified as dark-orange to red-hued laminae within the sedimentary record. Most notably, a distinct pattern of dark-orange to red laminae overlying pale, diffuse laminae is suggested to represent a period of persistent anoxia in the upper monimolimnion, followed by years with strong mixing events and associated iron precipitate deposition (Tomkins 2008). However, other processes can lead to iron deposition in the lake, such as terrestrial inputs of iron-rich grains and various aquatic chemical processes that form iron precipitates. Hence, the mean annual frequency of the distinct sedimentation pattern provides only general support for the pellet record and does not represent an independent ice-cover proxy (Fig. 8).

The frequency of mean annual sedimentary pellets and the frequency of distinct sedimentation patterns both suggest the increased occurrence of reduced ice cover summers from the late-1500s to late-1700s (Fig. 8). However, the pellet-frequency record suggests that ice cover was most reduced during the 1900s, while the distinct-sedimentation-pattern record suggests less iron deposition and, by extension, more persistent ice cover. The cause for this discrepancy is likely related to post-depositional iron mobility within the uppermost sediments. A reduced form of iron, such as pyrite, is highly insoluble and will maintain its position within the sediment profile over time, but other forms of iron may be diffused as ferrous ions into the overlying anoxic water (Engstrom and Wright 1984). This iron can be stored within the monimolimnion until reacting with hydrogen sulfide to form pyrite (Engstrom and Wright 1984). In some euxinic water bodies, such as the Cariaco Basin and Black Sea, iron from the upper sediments is commonly diffused into the overlying anoxic water, although trends in sedimentary iron concentrations are related in a complex manner to redox conditions in the water column (Lyons and Berner 1992; Lyons 1997; Yarincik et al. 2000).

Tomkins (2008) showed that pyrite is a common component of the sediments in Lake A, but the relative contribution from other forms of iron to the sediments remains unknown. Therefore, the reasons for the divergence of the distinct dark-orange lamina pattern and pellet-frequency records in recent times can only be speculated upon at this point. Considering that the orange and red-hued varves are most frequent ca. 1891 AD to present but the distinct sedimentation pattern is less frequent during this time (Fig. 8), these results

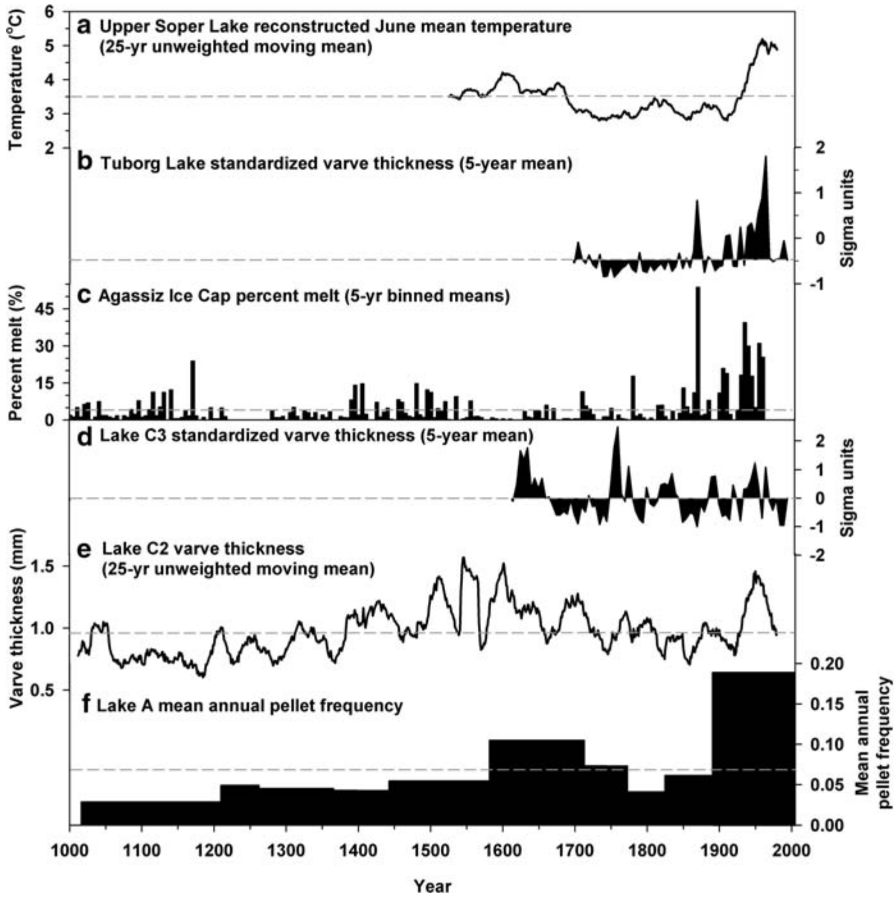

Fig. 10 Comparison of regional climate records including **a** Upper Soper Lake, Baffin Island, reconstructed June mean temperature (Hughen et al. 2000), **b** Tuborg Lake varve thickness standardized to the 1901–1961 mean (Overpeck et al. 1997), **c** Agassiz Ice Cap percent melt (core A84; Fisher and Koerner 1994; Fisher et al. 1995), **d** Lake C3 varve thickness standardized to the AD 1901–1961 mean (Lasca 1997), **e** unfiltered Lake C2 varve thickness (Lamoureux and Bradley 1996), and **f** Lake A mean pellet frequency (average number of pellets in each marker bed section per core). Mean values of each data set are denoted by dashed lines. Data for all locations, except Lake A (this study) and Lake C2 (Lamoureux and Bradley 1996), were obtained from the World Data Center for Paleoclimatology (http://www.ncdc.noaa.gov/paleo/data.html)

may indicate multiple, moderately low ice-cover years that allow deposition of pellets, but few ice-free summers that would be required for higher levels of iron to be deposited due to wind-mixing.

Paleoclimate implications and regional associations

The growth and decay of lake ice is largely controlled by sensible heat, which is related to summer temperature (Heron and Woo 1994). Given this process linkage, regional temperature reconstructions were compared to the proxy ice-cover record from Lake A (Fig. 10). Some discrepancies between the mean annual frequency of sedimentary pellets and regional proxy melt-season temperature records may be due to the differing temporal resolutions and uncertainties of the records. Nonetheless, regional lake records of inferred summer temperature largely corroborate the Lake A proxy ice-over record (Fig. 10).

Varve thickness at nearby Lake C2, an estimate of mean summer temperature (Hardy et al. 1996), suggests below-mean to mean temperatures prior to ca. 1400 AD and above-mean temperatures ca. 1400–1700 AD, followed by relatively cold conditions until the mid-1900s (Lamoureux and Bradley 1996). Similarly, the varve-thickness record from Lake C3, Ellesmere Island (adjacent to Lake C2), suggests warmth during the first half of the 1600s, extended periods of cool summer temperatures during the 1700s and 1800s, and a return to warmer temperatures during the mid-1900s (Lasca 1997). The trends in these two records largely correspond with the Lake A proxy ice-cover record. Additionally, the varve-thickness record from Lake Tuborg, Ellesmere Island, suggests a brief period of above-mean summer temperatures during the 1600s, followed by reduced temperatures during the 1700s and early-1800s, and warmer conditions from 1865 to 1962 AD (Smith et al. 2004). This record also corresponds with the latter part of the Lake A proxy ice-cover record, although twentieth-century warming begins earlier at Lake Tuborg, and sedimentation in the lake is dominated by glacial melt (Smith et al. 2004). The varve-thickness record from Upper Soper Lake, Baffin Island, estimates mean to above-mean June temperatures from the early-1500s to late-1600s (Hughen et al. 2000), which largely corresponds to a period of reduced ice in the Lake A record. Both records indicate relatively cold conditions during the 1800s, followed by strong warming during the 1900s (Hughen et al. 2000).

Climate reconstructions from High Arctic ice caps also show similar phases of inferred warmth to the Lake A proxy ice-cover record, particularly during the 1900s. Warm summer temperatures were inferred from δ^{18}O and summer melt-layer records in ice cores from Agassiz Ice Cap, Ellesmere Island, during the late-1300s, late-1400s, and 1900s (Fisher and Koerner 1983; Fisher et al. 1995) and cold conditions ca. 1600–1900 AD (Fisher et al. 1983). Some of the fluctuations in the Lake A proxy ice-cover record correspond with those in the ice-core records, but the ice cores indicate colder conditions that begin during the 1600s, rather than the late-1700s (Fig. 10). At the Devon Island Ice Cap, warm conditions were inferred from ice-melt features and δ^{18}O ca. 1240 and 1380 AD, and more prolonged warmth was suggested during the 1900s (Paterson et al. 1977). Cold periods were inferred ca. 1300, 1430, 1520, and 1560 AD, with more consistently cold conditions during the 1700s and 1800s (Paterson et al. 1977; Fisher and Koerner 1983; Alt 1985). The Devon Ice Cap records show similarities to the Lake A proxy ice-cover record, particularly during the past two centuries, but differences are evident prior to this period.

Numerous paleoecological records from Arctic lakes are consistent with the most recent part of the Lake A record. On a regional scale, changes in freshwater ecosystems are evident throughout the circumpolar Arctic due to recent climate warming and its inferred influence on ice cover, with numerous ponds and lakes showing unprecedented changes in diatom and chironomid assemblages within the past 200 years, and particularly since 1850 AD (e.g. Douglas et al. 1994; Antoniades et al. 2005; Smol et al. 2005; Antoniades et al. 2007). The changes observed in each of these records are attributed to recent increases in temperature, longer melt seasons, and reduced ice cover, which increases light infiltration into the water column, mixing within the water column and habitat availability (Smol et al. 2005; Smol and Douglas 2007). Ice cover is integral to both physical and biological characteristics of perennially ice-covered lakes and the abiotic and biotic records of ice-cover variability may be linked when both types of records are present within lacustrine sediment. A paucity of diatoms prevented such a comparison in Lake A, but other sites may be suited for using both types of records to assess past ice-cover variability.

The use of pellets as a general ice-cover proxy may be feasible beyond Lake A, as pellets similar to those in Lake A are also common within the sediments in perennially ice-covered Lake B (upstream from Lake A) and Lake C2 (Lamoureux and Bradley 1996, unpublished data). These lakes share many characteristics with Lake A, but are shallower, which suggests that their aquatic communities are likely more sensitive to temperature and ice-cover changes (Smol and Douglas 2007).

Conclusions

The sedimentary pellets in the Lake A sediments are distinct from the surrounding microlaminated, fine-grained sediment, and ice rafting is the most likely process for their deposition. In perennially ice-covered lakes such as Lake A, these pellets may provide a proxy record of reduced ice-cover conditions. The association between pellet frequency and ice-cover variability is supported by field observations and pellets at the top of the sedimentary record after a near ice-off event in 2000. The Lake A proxy ice-cover record indicates that ice mobility in the twentieth century appears to be unprecedented in the past 1000 years, and this proxy record of ice-cover variability is largely corroborated by regional climate reconstructions and paleoecological studies from Arctic lake sediments and ice caps, particularly during the late-1800s and 1900s when increased summer temperatures and reduced ice cover is inferred.

Not all perennially ice-covered lakes have the necessary conditions for pellets to form, but a comparison with sediments from nearby meromictic lakes suggests that similar sedimentological features occur in these settings as well. This physical signal of ice-cover variability could be particularly useful when used in conjunction with biological proxy records of ice-cover changes and associated changes in air temperature. The use of both physical and biological records, where they co-exist in Arctic lakes, could strengthen interpretations of past ice-cover variability by providing independent verification and by identifying weaknesses in individual records.

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