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A 300 year record of environmental change from Lake Tuborg, Ellesmere Island, Nunavut, Canada

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Abstract

Lamination thickness measurements in sediments from Lake Tuborg, northern Ellesmere Island, Canada document an increase in high-energy hydrologic discharge events from ~1865 to 1962. The timing of these events corresponds with evidence for an increase in the amount of melt on the adjacent Agassiz Ice Cap, as recorded in ice cores. There appears to have been a non-linear change in depositional energy resulting from a dramatic increase in Agassiz meltwater discharge, particularly after ~1908. A strong correlation between the Lake Tuborg varve thickness record, the amount of melting on the Agassiz Ice Cap and Eureka 900 mb air temperature records suggests that changes in the height of the freezing level in the atmosphere have affected the extent of summer melting on the Agassiz Ice Cap, leading to high volume discharge events and associated sediment flux to Lake Tuborg.

Introduction

Instrumental meteorological records in the High Arctic are generally limited to <50 years, providing a very limited perspective on climate variability in the region. General circulation model simulations all point to high latitude regions as being sensitive to increases in greenhouse gases from anthropogenic activity (Houghton et al. 2001). Such models raise the question: how have temperatures in the Arctic changed in recent decades, and have recent changes been unusual when placed in a long-term perspective? Answering this question requires the use of climate proxies – natural archives that can be resolved to annual, or near-annual resolution. In the High Arctic context, this boils down to either ice cores or varved sediments.

Ice cores in the Canadian Arctic have been retrieved from the Meighen Ice Cap, Devon Island Ice Cap, Agassiz Ice Cap (Ellesmere Island) and the Penny Ice Cap (Baffin Island) (Koerner 1977; Koerner and Fisher 1990; Fisher et al. 1995, 1996; Grumet et al. 2001; Fisher and Koerner 2003). Some of these ice cores provide excellent temporal resolution for recent centuries, but they are only available from a few locations and they are often difficult to interpret in terms of specific seasonal climatic conditions. Annually laminated lake sediments (varves) are found in many locations in the Arctic, offering the prospect of additional, high-resolution paleoclimatic proxy records (e.g., Zolitschka 1996; Lamoureux and Bradley 1996; Hughen et al. 2000; Lamoureux et al. 2001; Francus et al. 2002). Varved sediments have particular promise for use in High Arctic regions because of the relatively simple seasonality of sediment delivery, and the absence of human influences on catchments, which eliminates several

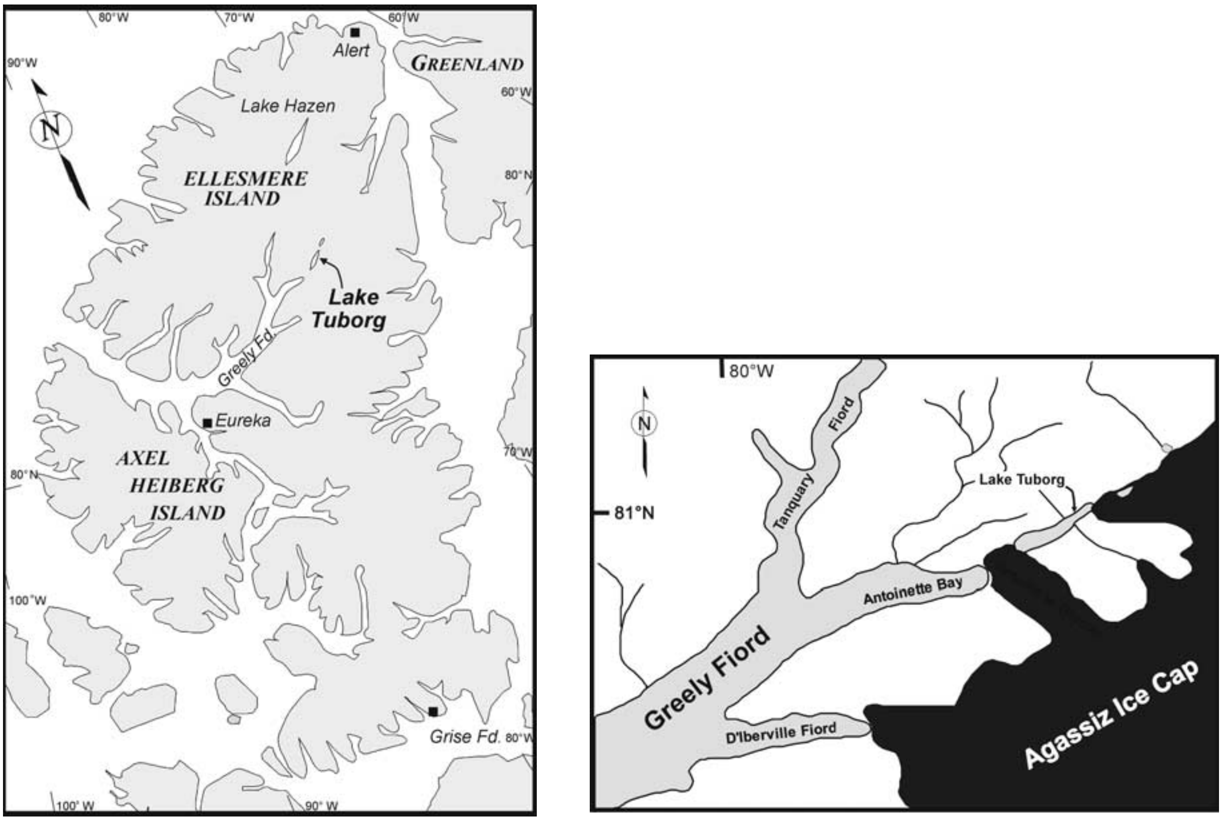

Figure 1. Location of Lake Tuborg, Ellesmere Island.

complications often encountered in varved sediment studies at lower latitudes. In addition, because multiple cores can be recovered fairly easily from each lake, stratigraphic sections can be compared and combined, to both assure chronological accuracy and reduce noise in the records. This is rarely done with ice cores (cf. Fisher and Koerner 1994).

Study area

Lake Tuborg is situated at the head of Antoinette Bay at the north end of the central limb of Greely Fiord (Figure 1). The lake is 20 km long and is isolated from the bay by a tidewater glacier that calves into the lake at the southeastern end. At present, a terminal moraine deposited by the glacier regulates outlet streamflow from Lake Tuborg to Antoinette Bay. The lake itself was created by the advance of this tidewater glacier across the fiord. Today, the lake surface elevation is approximately 10.5 m above sea level. Bathymetric surveys show that the lake has two main basins – a larger southwestern basin up to 155 m deep separated by a sill (at ~35–40 m) from a small northeastern basin, where water depths are up to ~80 m.

Radiocarbon dating of bicarbonate precipitated from anoxic seawater trapped at the base of the lake gave an age of ~3000 yr B.P. (Bowman and Long 1968), suggesting that the ice advance that created the lake occurred some time before that. However, it is not known to what extent samples in the marine bicarbonate might have been contaminated by the addition of older – or younger – material to the mixolimnion following lake isolation, and so the date may in fact be unrelated to the time that Lake Tuborg became isolated from the sea.

As freshwater was discharged into the newly formed lake, stratification of the water column occurred and this condition is still present today in the southwestern basin of the lake (Figure 2). There, beneath ~55 m of fresh water, an extremely

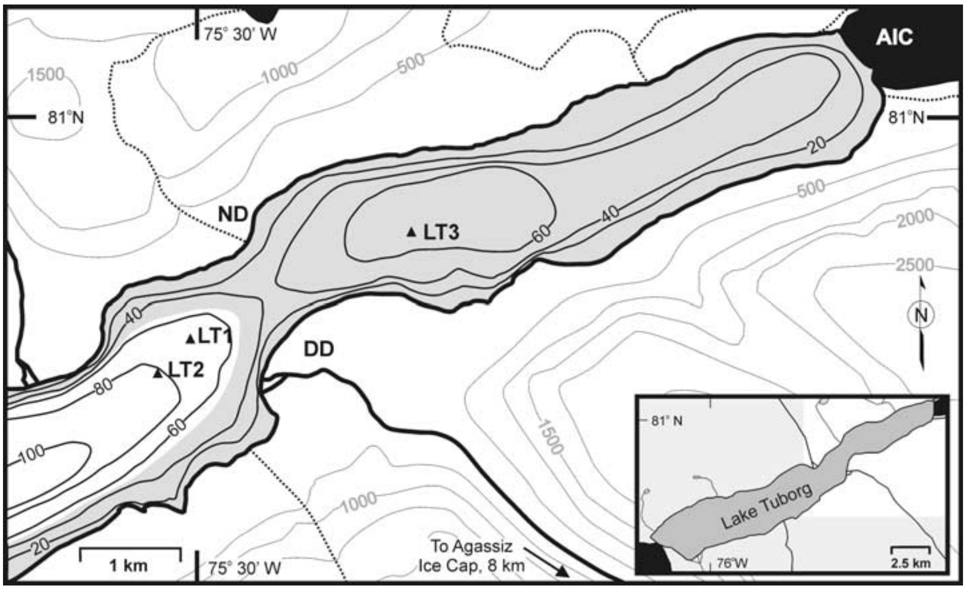

Figure 2. Topographic map of the northeast section of Lake Tuborg. Coring locations are marked with triangle symbols (LT1–3). Isobath interval is 20 m below mean summer lake level. Gray shading areas represents where the water column is entirely fresh; white shading represents areas where the lake is meromictic (below about 55 m). Contours were drawn from 221 GPS-referenced (~2–5 m accuracy) fish-finder soundings. The shoreline is drawn from GPS tracklogs and National Air Photo Library photo A-16721–52. Topographic contour interval is 500 ft (ASL; 152 m). Contours are based on Energy, Mines and Resources Canada 1 : 250,000 topographic maps 340A (Antoinette Glacier) and 340D (Tanquary Fiord). Major tributaries are represented as solid lines, minor tributaries are stippled. Abbreviations: LT1–3, Lake Tuborg cores 1–3; DD, Deception Delta; ND; Nival Delta. Inset shows all of Lake Tuborg, and the portion of the lake represented in the main figure (white shading).

abrupt chemocline isolates the old, anoxic seawater in the monimolimnion. The presence of this anoxic water is conducive to the preservation of laminated sediments as it eliminates disturbance of the sediments from both bioturbation and seasonal overturning. The first modern survey of the limnology of Lake Tuborg region was on 2 June, 1963, by G. Hattersley-Smith and H. Serson, who reported that water of ~25‰ salinity lay at the bottom of the lake. Temperature and salinity profiles taken at that time show the presence of an upper isothermal layer of 27 m above a well-established chemocline (Hattersley-Smith and Serson 1964). Temperature–conductivity–transmissivity profiles taken in 2001 revealed a more complex pattern than documented in the older surveys (Figure 3). The depth of the chemocline in the southwestern portion of Lake Tuborg corresponds to that recorded in 1963, but profiles of the water column in the northeastern basin of the lake show no chemocline (no limnological surveys were made there in 1963). In that basin, the water column is essentially fresh. Flushing of the seawater could simply have occurred over a long period of time, as a result of slow erosion of the chemocline by freshwater entering the lake from large rivers at the northeastern end or more rapidly by the catastrophic drainage of a large glacier-dammed lake that is perched above Lake Tuborg at its northeastern end. This is discussed in more depth later.

Fieldwork

In 1995, fieldwork was carried out at Lake Tuborg in order to retrieve lake sediment cores, monitor the discharge of a meltwater stream from the Agassiz Ice Cap and record meteorological conditions within the watershed (at 800 m elevation and at lake level). The higher elevation site was selected in order to characterize meteorological conditions in the upper parts of the Lake Tuborg Basin, much of which is at or above this elevation. Hydrological and meteorological conditions, and

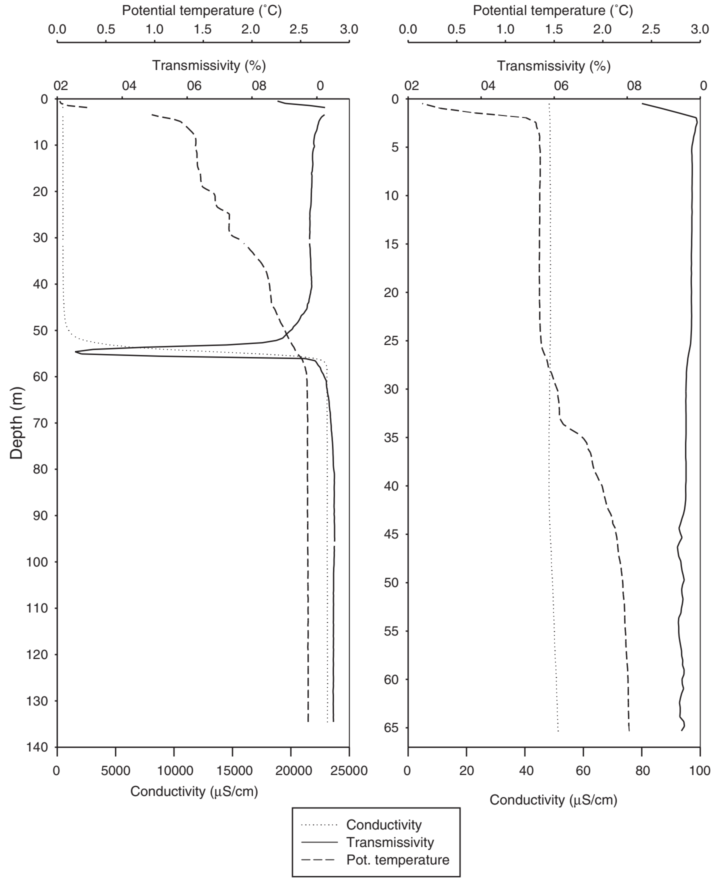

Figure 3. Physical properties of water column at two sites in Lake Tuborg – near LT2 and LT3 on the map. The strong chemocline at the LT2 location is clearly seen.

their relationship to sediment flux during the field season, were reported by Braun et al. (2000). Streamflow and sediment transport into Lake Tuborg were largely a reflection of snowmelt, initially from the lower elevation slopes and later from major slushflow events on the Agassiz Ice Cap. Summer rainfall events had minimal influence on sediment flux to the lake.

During the 1995 field season three cores of >2.4 m length were collected using a percussion corer (Reasoner 1993) together with corresponding surface Glew cores (at the same locations) that retained an intact sediment–water interface. Figure 2 shows the locations of the cores. In general, they were taken from areas of the lake that had a gentle sloping bottom. LT1 and LT2 were recovered (79 and 83.5 m water depth, respectively) in that part of the lake, southwest of the sill, which is strongly meromictic today. Core LT3 was obtained from 77.3 m water depth, in the northeastern basin that is currently not stratified.

Laboratory methods

Following the procedure established by Clark (1988) and modified by Lamoureux (1994), sediment slabs were removed from the face of the split cores utilizing aluminum templates, and then dehydrated using histological grade acetone. Several modifications in procedure were made to minimize costs and laboratory time, and maximize thin section quality. For example, to assure the complete dehydration of samples prior to impregnation, the specific gravity of the acetone rinse was taken with a hydrometer. Gravity measurements were monitored until a 0.7 g cm^{-3} threshold was reached, at which time impregnation could begin.

Following dehydration, sediment slabs were embedded rapidly under a vacuum with a single application of low viscosity epoxy.

Following impregnation, sediment slabs were cut lengthwise using a 10 in. rock saw and then polished using 800 grit abrasive on glass, in order to enhance the contrast between varves. Each polished slab was then scanned at 1200 dpi resolution so that a 16:1 zoom frame could be achieved with adequate clarity. Images were enhanced in order to clarify further laminae details and imported into *NIH Image* where screen calibration techniques were used to facilitate laminae counting. Counting errors were minimized by visually recording varve measurements and measurement axes for further studies. Additional counts changed only slightly, where subjective interpretations of individual laminae varied.

In order to ensure that accurate counts could be justified using this technique of image analysis, laminae counts from selected sediment slabs were compared with counts from corresponding thin sections utilizing an Olympus BH-2 binocular microscope with a calibrated field of view. The results of this experiment indicate that image analysis of the scanned slabs was a valid method of varve identification and quantification, producing counts that were reproducible within ±2 years for each core studied. It is unlikely, however, that microlaminated sediments (varve thicknesses <0.1 mm) can be analyzed effectively in this manner due to the difficulty of discerning individual couplets. Fortunately, in the sediments analyzed, such cases appear to be rare.

Although each core retrieved from Lake Tuborg has the potential of providing a very long record of environmental change, sediment disruption due to friction with the core tube walls precludes the effective measurement of varve thickness for more than the last 300 years. In core 95LT2, the sand content resulted in a significant deformation below varve date 1754. Lower mean grain sizes in cores 95LT1 and 95LT3 resulted in less core disruption and allowed longer time series to be constructed from those cores.

Radiometric dates

Gamma counting of ^{137}Cs identified a peak in ^{137}Cs activity in core 95LT2E at a depth of 6 cm (Figure 4) that we interpret as resulting from atmospheric nuclear weapons testing (Pourchet and Pinglot 1989). This peak correlated well with the varve which was dated as 1965, suggesting an error of ~2 years over 30 years of record. The drop in ^{137}Cs activity to zero at 2 cm depth highlights the effective dilution of radionuclides in the region due to high sedimentation rates. In ideal systems, the ^{137}Cs profile displays peaks in both 1963 and 1986, the latter resulting from the Chernobyl event (Pourchet and Pinglot 1989). The increase in ^{137}Cs activities in the uppermost 2 cm of core 95LT2 may reflect this later increase, but it does not show a distinct peak due to the averaging effects of 1 cm sampling.

Dating the sediments with ^{210}Pb was less successful. Samples taken from core 95LT2P show significant inconsistencies in the ^{210}Pb activity downcore, preventing the use of constant rate of supply (CRS) or constant initial concentration (CIC) dating models (Figure 4). Low fluxes of ^{210}Pb are common at high latitudes causing difficulties in establishing recent chronologies. These problems are undoubtedly compounded at Lake Tuborg due to the large sediment load to the lake and the high sand and silt concentrations. Taken together, these factors preclude the use of the ^{210}Pb data for dating purposes in these cores.

In addition to causing difficulties in the use of ^{210}Pb, the high sedimentation rate and limited vegetation in the watershed of Lake Tuborg resulted in the absence of any datable organic material in the cores. It is hoped that additional fieldwork in the region can retrieve sediments that will aid in the development of a long-term chronology for the Lake Tuborg basin through the use of radiometric dating techniques.

Lake Tuborg varve series

Figure 5 shows the varve thickness measurements in the three cores from Lake Tuborg. Table 1 summarizes their characteristics over the common period of record (1754–1995). The thickest varves were deposited in the proximal basin (as recorded at site LT3) and at LT1. Site LT2 has lower mean sedimentation rates but captures extreme events related to the main outlet stream draining from the Agassiz Ice Cap (cf. Figures 1 and 2).

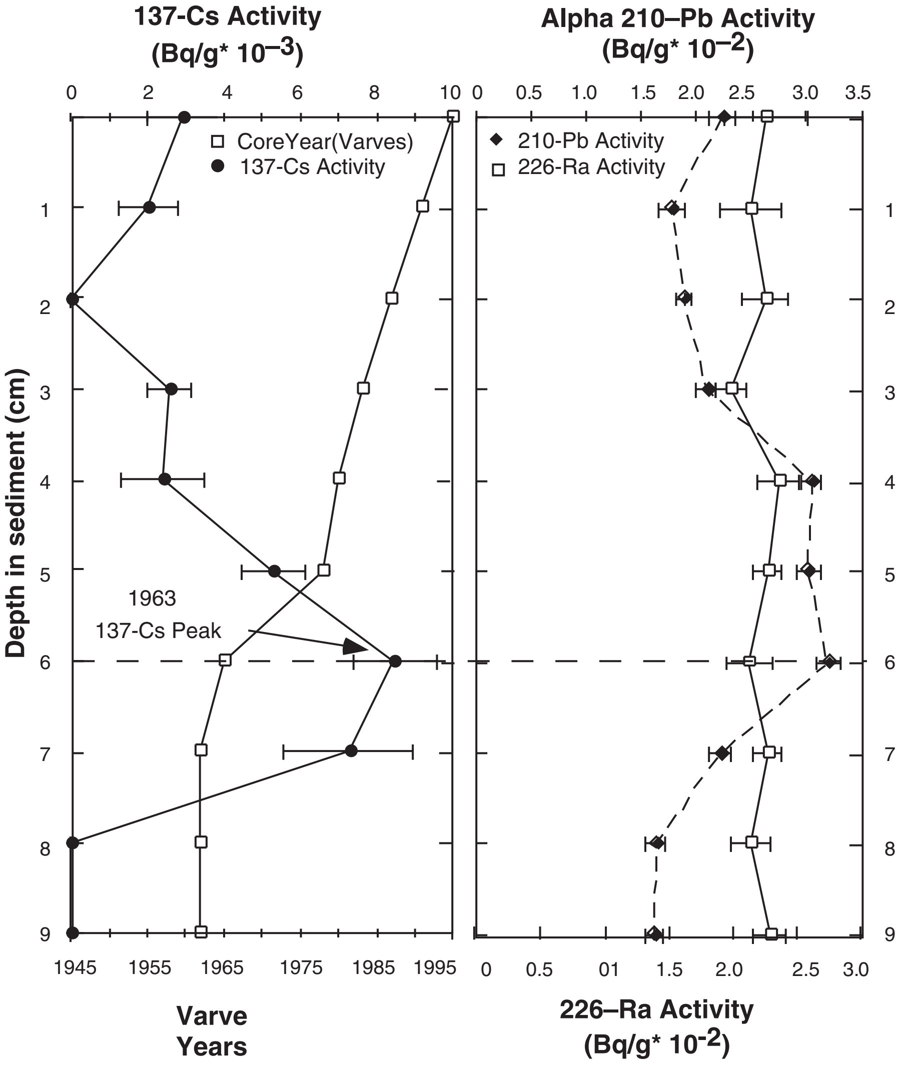

Figure 4. ^{137}Cs and ^{210}Pb records from core 95LT2. Varve counts for the last 50 years of record are also shown.

Catastrophic drainage of pooled meltwater on the Agassiz Ice Cap may have contributed substantially to creating the largest single varve recorded in 95LT2 (and the second largest in 95LT3), in 1962. This year is considered a marker horizon in ice cores from the High Arctic because exceptional warmth during the summer (related to a persistent zone of high pressure over the region) led to melting even at high elevations on many ice caps. This resulted in percolating water re-freezing in the underlying firn, creating a characteristic dense "melt layer" and overall highly negative mass balance conditions (Koerner 1979; Alt 1987). A corresponding peak in varve thickness is clearly represented in the Lake Tuborg varve series, indicating that increased meltwater discharge from the Agassiz Ice Cap was responsible for the deposition of a large amount of sediment to the lake during the summer of 1962. The distinct correspondence of the Lake Tuborg sediment record and the 1962 melt horizon on the Agassiz Ice Cap strengthens the Lake Tuborg chronology and suggests that a strong link exists between meltwater production on the ice cap, and sediment deposition to the lake. We believe that similar high-magnitude drainage events were observed in 1995 when a lake (or lakes) of supraglacial meltwater drained suddenly from the Agassiz Ice Cap to the south (through the

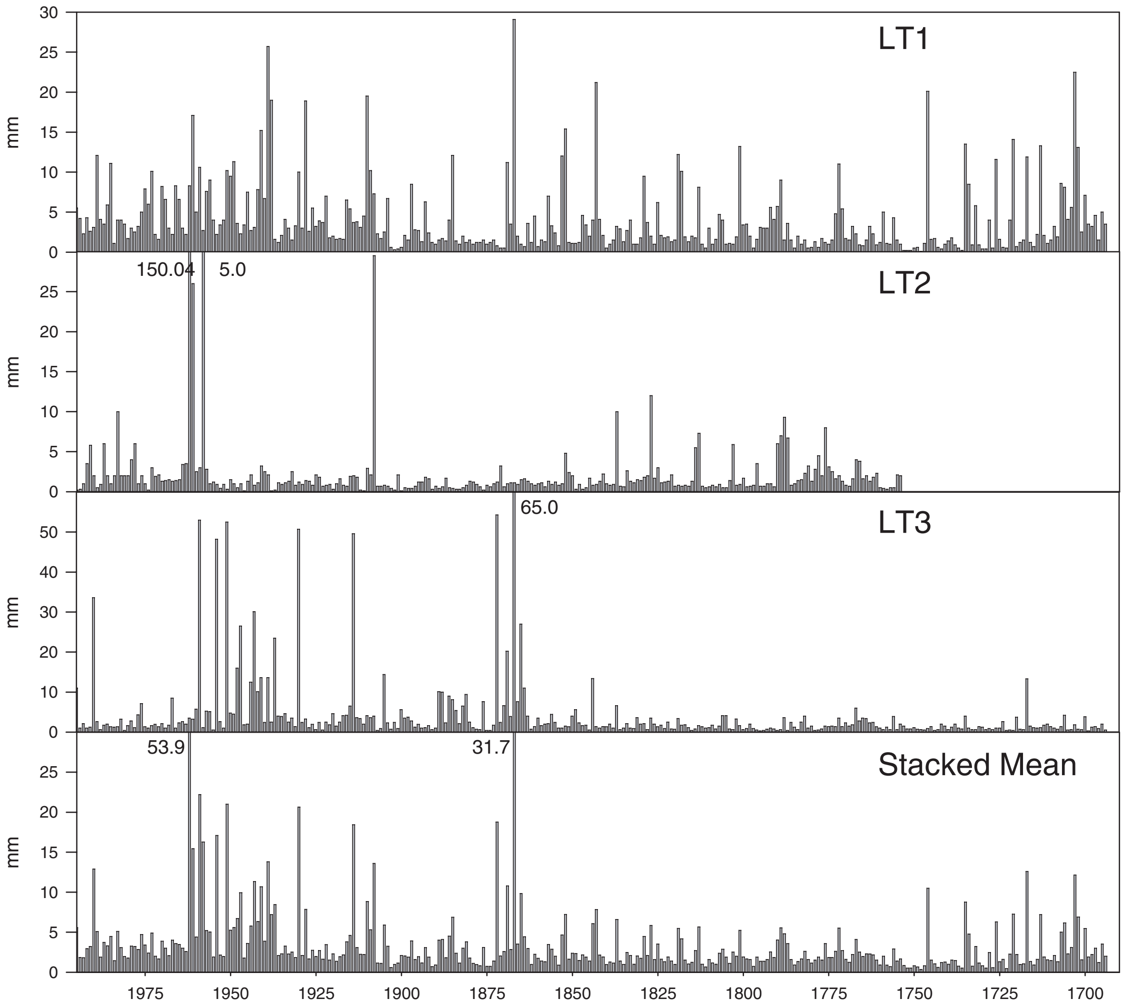

Figure 5. Lake Tuborg varve thickness series for the three cores studied, and for the mean of all three series.

Table 1. Descriptive statistics for Lake Tuborg varve sediments over their common period of record (1754–1995).

Core	*N*	Mean (mm)	Std. dev.
95LT1	242	4.0	4.3
95LT2	242	2.7	10.3
95LT3	242	4.3	9.7
Mean	242	3.8	5.2

northwestward flowing river shown in Figure 2). Discharge more than doubled and suspended sediment flux increased by one order of magnitude in less than an hour during two drainage events. About 70% of the measured suspended sediment transport (recorded up to that time) was transported in association with these events, in <27 h combined (Braun et al. 2000).

Core 95LT3, which came from a site closest to the glacier (and glacier-dammed lake) at the north-eastern end of the lake, also contains thick varves, similar to those in 95LT2, which characterize discrete depositional events in the region. In this core, the thickest single varve was deposited ~1867 (also exceptional in core LT1). Core 95LT1 shows fewer extremes in sediment deposition over the last 300 years, largely due to its location away from major river outlets. It provides a more integrated record of sedimentation in the lake.

To highlight years whose depositional signature was ubiquitous (which we take to reflect regional

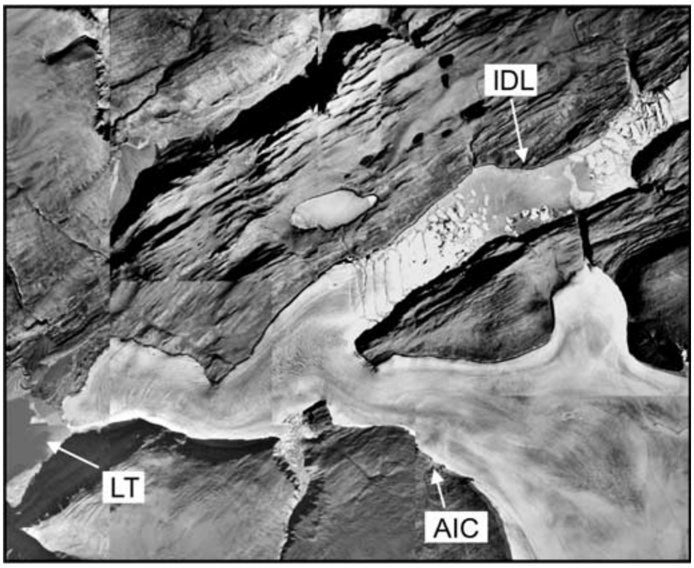

Figure 6. Glacier-dammed lake above northeastern end of Lake Tuborg (Crown copyright 1959). LT is Lake Tuborg; IDL is ice-dammed lake and AIC is Agassiz Ice Cap. Composite of photographs A16687–47 and A16977–63, both taken in August 1959. scale is variable due to parallax; scale shown is at center of photograph.

climatic effects), and dampen variability that may have had more local (non-climatic) control, we created an average Lake Tuborg varve series from the three records (Figure 5). A similar approach was taken in the analysis of ice core records from the Agassiz Ice Cap, to maximize the regional climatic signal to noise ratio (Fisher and Koerner 1994). Given the chronological uncertainty with depth, it may be more appropriate to consider five year running means, as shown in Figure 5, but in either case, it is clear that the last century was characterized by a marked increase in high sedimentation rate events, possibly similar to those observed during the summer field season of 1995 (Braun et al. 2000). In particular, the period from ~1908 to 1962 stands out as a time of numerous high discharge/major sediment flux events. A secondary peak is also apparent, from ~1864 to 1872. The period of least sediment deposition was from ~1736 to 1761 (with one exceptional year in ~1746) though this mean is based on only two cores (95LT1 and 95LT3).

One characteristic of Lake Tuborg that is of particular interest is the presence of a large glacier-dammed lake perched above the northeastern end of Lake Tuborg. In 2003, this lake suddenly drained, discharging all of its volume into Lake Tuborg within a few days and raising the level of Lake Tuborg by ~8 m. We have anecdotal evidence that similar events may also have occurred within the last decade. The upper lake is visible in air photographs of the region taken in 1959 (Figure 6) and it was noticed by Koerner, who flew over it in 1992 (R.M. Koerner, pers. comm.). However, it had drained away entirely by the summer of 1995. Although there is a marginal channel along the northern side of the glacier dam that has evidently carried excess meltwater to Lake Tuborg in the past (when the upper lake exceeded some critical level) the 2003 drainage was clearly sub-glacial and catastrophic. After the lake drained, a tunnel was revealed beneath the glacier that formerly dammed it. This is unlike the well-documented events noted by Maag (1969) who described drainage channels cut along the edge of the Crusoe and Thompson glaciers, Axel Heiberg Island after meltwater accumulated in marginal lakes. In the case of Lake Tuborg, it is possible that the upper pro-glacial lake has filled and drained many times in the past, with the glacier dam acting as a “pressure valve” regulating flow until some critical threshold has been reached. Then, rapid and large magnitude drainage events would have ensued, and such events may have contributed to removal of saline water in the monimolimnion of the northeastern basin of Lake Tuborg. Parts of the lake further

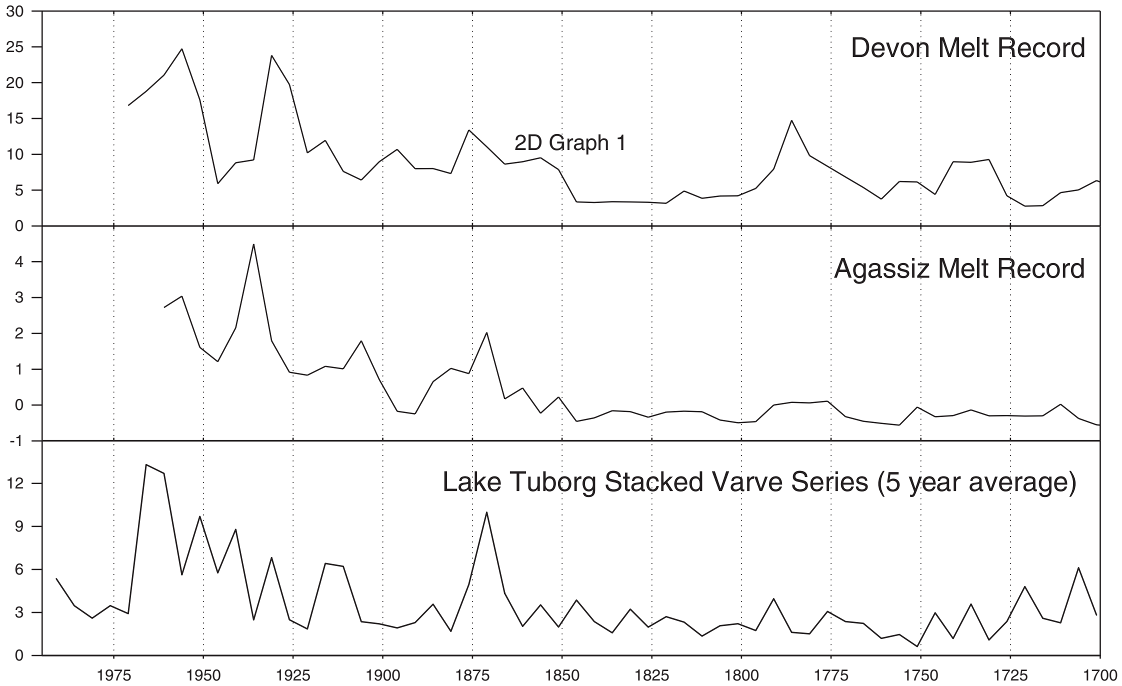

Figure 7. Five year running mean of the Lake Tuborg mean varve record, the Agassiz Ice Cap stacked melt record (cores A77, A79, A84 and A87) and the Devon melt record (Fisher and Koerner 1994).

west would have been somewhat protected from such disturbances by the sill (which extends up to ~38 m depth). We speculate that drainage of marginal, proglacial lakes may be more common when the lake receives exceptional quantities of meltwater in the summer from the adjacent Agassiz Ice Cap, providing additional hydraulic pressure against the "valve" (the glacier dam). Thus, such events and their associated sedimentary signatures may be particularly characteristic of warmer periods in the past.

Paleoclimatic interpretation

A variety of proxy records from the Arctic region have been useful in establishing a Holocene paleoenvironmental history of the Queen Elizabeth Islands (Bradley 1990). In particular, melt records retrieved from the Agassiz (Koerner and Fisher 1990) and Devon Ice Caps (Koerner 1977) provide an important data set for comparison with the Lake Tuborg varve series. The percentage of melt features in an ice core section is an indicator of summer temperatures. Warm summers lead to significant melting on the ice caps, and this would be associated with high runoff and sediment flux to marginal lakes such as Lake Tuborg (cf. Hardy et al. 1996). Thus, the ice core melt record and the Lake Tuborg varve thickness series are physically linked and should provide a similar record of past summer temperatures. Figure 7 shows that they are indeed well correlated ($r = 0.58$, $p < 0.01$). Furthermore, there is a strong correlation between the Lake Tuborg varve series and Eureka melting degree days at 900 mb (corresponding to elevations at the margin of the Agassiz Ice Cap) ($r = 0.47$, $p < 0.01$) suggesting that temperature provides a major control on sedimentation to Lake Tuborg, as verified more quantitatively by Braun et al. (2000).

Broadly, the correspondence between the Lake Tuborg and Agassiz records strengthens the link between meltwater production and sediment deposition to Lake Tuborg and suggests that both systems are responding to changes in ambient summer temperature changes over the period of record.

Of particular interest is the correspondence of melt peaks in the Agassiz Ice Cap and Devon Island Ice Cap ice core records with anomalously thick varves, following the mid-19th century.

Extremely thick varves (>~15 mm) are virtually absent in the Lake Tuborg cores below varve year A.D. 1867, suggesting that the region has undergone significant warming during this period. It is also noteable in the Lake Tuborg varve record that varve thickness dropped sharply after 1962, to levels more typical of the late 19th/early 20th century, indicating that the region experienced periods of significantly greater summer melt event during the past 150 years than in the period from 1963 to 1995 (though this cool regime appears to have ended within the last decade – see Serreze et al. 2000, 2003; Comiso 2002, 2003). A pronounced change in climate following 1962 has been recognized previously in high latitudes and has been attributed to significant changes in atmospheric circulation across the region (Bradley and England 1978; Kahl et al. 1993). In particular, the abrupt lowering of summer freezing levels in the atmosphere in 1963 was associated with the presence of anomalously warm ocean temperatures in the central northern Pacific Ocean, which altered pressure patterns in the Northern Hemisphere and resulted in the onset of this new climatic regime (Bradley 1973; Dronia 1974). Freezing level height directly influences equilibrium line altitudes on ice masses in the High Arctic (Bradley 1975) thereby influencing marginal ablation, runoff and sediment transport to lakes.

Several other studies indicate that the Arctic experienced a pronounced warming trend from the mid-19th century until the mid to late 20th century (Jacoby and D’Arrigo 1989; Lamoureux and Bradley 1996; Overpeck et al. 1997; Hughen et al. 2000) as the Arctic emerged from the “Little Ice Age” (Jones and Bradley 1992; Bradley and Jones 1993). Though the Lake Tuborg varve records only span the last few centuries, the ice core records allow this period to be placed in an even long-term perspective. Composite melt records from two ice cores retrieved from the Agassiz Ice Cap (A87 and A84) demonstrate that, except for a few short intervals, summers over the past 100 years have been warmer than the last ~3500 years (Koerner and Fisher 1990). In addition, it has been argued that the period from 1931 to 1960 was warmer than 80% of the last 3000 years in the High Arctic, making the mid-20th century a period of unprecedented warmth during the late Holocene (Lamoureux and Bradley 1996). This recent warm period is particularly significant in that it follows one of the coldest periods in the entire Holocene (Bradley 2000). The effect of such changes on lacustrine ecosystems in the High Arctic is becoming increasingly apparent (e.g., Douglas et al. 1994; Perren et al. 2002).

Conclusions

The 300 year Lake Tuborg record suggests that the warmest interval of the last few centuries (prior to 1995) was the period from ~1865 to 1962, especially the latter half of this period. During this time, warm summers led to substantial melting on the Agassiz Ice Cap which caused there to be high river discharge and associated sediment flux to Lake Tuborg, resulting in the deposition of thick varves. This finding corroborates other studies that have also noted warming during this period. Cooler summer conditions since 1962 are analogous to those of the late 19th/early 20th century in this region. We do not currently know how the remarkably warm summers of 2000 and 2002 compare with earlier conditions; new cores to place recent events in a long-term perspective are needed.

Both the Agassiz Ice Cap and Lake Tuborg have responded to changes in summer freezing level elevation, as recorded by radiosonde data from Eureka. Corresponding changes in the number of melting degree days at 900 mb provide an additional link between sedimentation to Lake Tuborg and ambient summer temperatures. It is likely that this depositional signature will be found in other High Arctic lakes. Given the correspondence of this event with the 1963 peak in ^{137}Cs, a corresponding drop in varve thickness at this time may provide a useful dating horizon for future varve-based studies in glacially fed basins.

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