Limnology, sedimentology, and hydrology of a jökulhlaup into a meromictic High Arctic lake[1]

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Abstract: A large ice-dammed lake drained catastrophically into Lake Tuborg, Ellesmere Island, beginning on 25 July 2003. Limnological, sedimentological, and hydrological parameters were recorded before, during, and after this event. For several weeks prior to the jökulhlaup, water overtopped the ice-dammed lake and flowed into Lake Tuborg's freshwater basin. A shallow sill separates the freshwater basin from a larger, deeper basin containing ~25 PSU (practical salinity units) salt water. The sill blocked underflows from entering the saltwater basin before the jökulhlaup. The ice-dammed lake drained completely and catastrophically when englacial or subglacial conduits developed, and a glacier portal formed 980 m from the Lake Tuborg shore, marking the beginning of the jökulhlaup. The level of Lake Tuborg increased by 7.6 m in 84 h. This jökulhlaup is the largest known to have occurred in the High Arctic, and the largest witnessed in Canada since 1947. Strata of very cold water flowed above the chemocline for about 14 km, from the sill to the southwest end of the lake. The cold strata turbulently mixed with underlying salt water, allowing for saltwater flocculation of suspended sediment, causing rapid settling. The saltwater layer very slightly freshened and cooled. Close to the sill, near-surface sediments derived from the jökulhlaup are coarse and laminated; however, no erosion occurred toward the distal end of the lake, where a fining upward unit with a coarse base was deposited.

Résumé : Un grand lac retenu par des glaces s'est drainé de manière catastrophique dans le lac Tuborg, de l'île Ellesmere, le 25 juillet 2003. Les paramètres limnologiques, sédimentologiques et hydrologiques ont été mesurés avant, durant et après cet événement. Durant plusieurs semaines avant le jökulhlaup, de l'eau se déversait par-dessus le barrage de glace du lac et s'écoulait dans le bassin d'eau douce du lac Tuborg. Un seuil peu profond sépare le bassin d'eau douce d'un bassin plus grand et plus profond contenant de l'eau à une salinité d'environ 25 USP (unités de salinité pratique). Avant le jökulhlaup, le seuil empêchait des courants de fond d'entrer dans le bassin d'eau salée. Le lac retenu par les glaces s'est drainé complètement et de manière catastrophique lorsque des canalisations dans ou sous la glace se sont développées et qu'un portail glaciaire s'est formé à 980 m des berges du lac Tuborg, marquant ainsi le début du jökulhlaup. Le niveau du lac Tuborg s'est élevé de 7,6 m en 84 heures. Ce jökulhlaup est le plus grand connu qui se soit produit dans le Grand Nord et le plus gros à être vu au Canada depuis 1947. Une couche d'eau très froide coulait par-dessus le chemocline sur une distance d'environ 14 km, depuis le seuil jusqu'à l'extrémité sud-ouest du lac. La couche froide s'est mêlée de manière turbulente avec l'eau salée sous-jacente engendrant une floculation des sédiments en suspension dans l'eau salée et une sédimentation rapide. La couche d'eau salée est devenue légèrement plus douce et plus froide. Les sédiments de surface provenant du jökulhlaup, sont grossiers et laminés à proximité du seuil; toutefois, il ne s'est pas produit d'érosion à l'extrémité distale du lac, où s'est déposée une unité à base grossière et à granodécroissance vers le haut.

[Traduit par la Rédaction]

Introduction

Jökulhlaups (Icelandic for "glacier-burst") have never been monitored in detail entering downstream water bodies. The potential for erosion and sediment bypassing during jökulhlaups is assumed to be great, and this is probably why models describing glacier dam position and jökulhlaup frequency and intensity (e.g., Clague and Evans 1994) have not yet been tested using long sediment archives. If a site could be identified where a characteristic jökulhlaup facies is deposited without producing an unconformity, an uninterrupted record of jökulhlaup frequency could be obtained from a sediment archive.

Lake Tuborg is a large, fiord-type lake on Ellesmere Island in the Canadian High Arctic (Fig. 1) adjacent to the Agassiz Ice Cap. Its bottom sediments are annually laminated (varved) in some locations. It also contains trapped salt water (it is meromictic). A large ice-dammed lake catastrophically drained into Lake Tuborg in July 2003. A limnologic process study was underway in Lake Tuborg at the time of the jökulhlaup.

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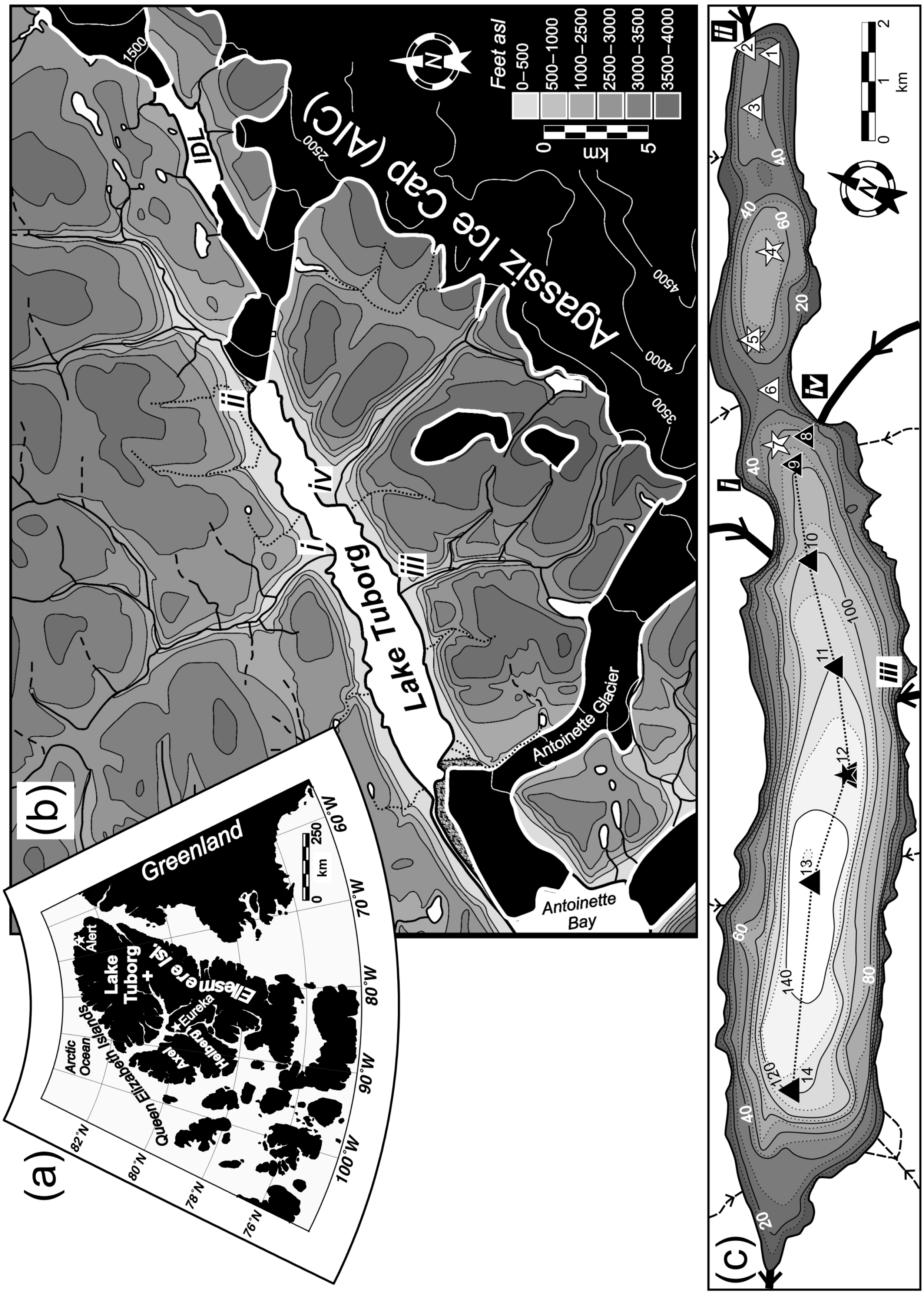


Fig. 1. (*a*) Regional map of Ellesmere Island and surroundings in the Canadian High Arctic. (*b*) Physiography and topography near Lake Tuborg and the ice-dammed lake (IDL). Contour interval is 500 feet (~150 m); shading is feet asl (above sea level). (*c*) Lake Tuborg bathymetry. Solid bathymetric contours are every 20 m. Stars are core locations; triangles are limnologic monitoring stations. Location "*i*" in (*b*) and (*c*) is at 80.98°N, 75.555°W.

Conductivity–temperature–depth (CTD) casts, surface gravity cores, and a lake level record were obtained in the 2001–2003 melt seasons, so unique records of physical lake conditions before, during, and after the jökulhlaup were obtained.

A major objective of this paper is to identify the type and spatial variability of jökulhlaup-derived processes and deposits. These data also provide an opportunity to differentiate jökulhlaup-derived lacustrine sedimentary processes and deposits from those that are not jökulhlaup derived. This is particularly important at Lake Tuborg because there are several relatively small supraglacial lakes on the Agassiz Ice Cap that drain—perhaps annually—as slush flows that quickly and energetically transport large volumes of sediment to the lake (Braun et al. 2000). The data also allow characterization of boundary layer mixing processes along the sharp and strong Lake Tuborg chemocline; these processes have previously only been recorded in estuaries (e.g., Geyer and Smith 1987) and laboratory experiments (e.g., Rimoldi et al. 1996). The regionally unique hydrology and drainage mechanisms of the 2003 jökulhlaup are also described.

Site description

Lake Tuborg was formed about 3000 years ago when Antoinette Glacier advanced and trapped sea water in the lake (Long 1967). The lake, located in a steep-walled 450–600 m deep valley, is at ~11 m asl (above sea level) (Fig. 1*b*). Soils are very thin cryic regosols superimposed on permafrost, and vegetation is very sparse. Lake surface area is 42 km^2, and maximum length and width are 20.9 and 3.4 km, respectively (Fig. 1*c*). There is a large, deep, meromictic basin at the southwest end (maximum depth = ~145 m), and a smaller, shallower freshwater basin at the northeast end of the lake (maximum depth = ~74 m; Fig. 1*c*). A 34 m deep sill separates the two basins. Lake ice cover is nearly perennial.

A major stream enters Lake Tuborg at "*i*" (throughout the text, "*i*" to "*iv*" refer to locations in Fig. 1*b*). It is entirely fed by snowmelt; peak streamflow occurs in late June, and discharge decreases greatly when watershed snow cover is exhausted shortly afterward. Watersheds above "*iii*" and "*iv*" are extensively glacierized, with higher mean elevations than at "*i*". This creates higher discharge duration and amount, and higher sediment transport than at "*i*", but the timing of peak flow is delayed until early to mid-July when peak summer air temperature occurs (Braun et al. 2000). Small supraglacial lakes on the Agassiz Ice Cap quickly drained in July 1995, and discharge and suspended sediment transport briefly, but greatly, increased (Braun et al. 2000). The ice-dammed lake responsible for the 2003 jökulhlaup is part of the watershed above "*ii*", but the lake captures almost all glacial melt from the catchment while filling, so the streamflow regime at "*ii*" is normally nival. CTD casts and visual observations at the glacier terminus near "*ii*" showed no evidence of subglacial meltwater discharge, even during peak summer melt.

The edge of the Agassiz Ice Cap parallels Lake Tuborg at ~900 m asl (Fig. 1*b*). The Agassiz Ice Cap generally is cold based, and discharge is from supraglacial snow and ice melt; however, outlet glaciers near sea level at the northeast and southwest ends of the lake are warm based (Braun et al. 2000). The ice-dammed lake is dammed by a tributary glacier of the Agassiz Ice Cap that branches into a small valley between 300 and 460 m asl (Figs. 1*b*, 2). Bergs calve into the lake at its northeast and southwest ends, and the glacier dam at the southwest end normally blocks all outflow.

Mean annual air temperature at Eureka (Fig. 1*a*; 10 m asl), the closest Meteorological Service of Canada (Environment Canada) weather station to Lake Tuborg, is –20 °C, and the only months with average temperatures greater than 0 °C are June (2.3 °C), July (5.6 °C), and August (3.1 °C).

Methods

Monitoring at Lake Tuborg took place continuously from mid May to mid August in 2001–2003. Coring and limnologic monitoring stations are shown in Fig. 1*c*. Air temperature was recorded every 15 minutes near lake level; precipitation was collected at ground level and manually read. Lake level was recorded every 15 minutes (cf. Reedyk et al. 1997). End of season lake level was marked with cairns to ensure a common lake level datum. The lake level recorder could not be used after 21 July 2003 because of shifting ice pans and rapidly rising lake level, so water height after this date was surveyed with a staff and level. The measurements were made near "*i*".

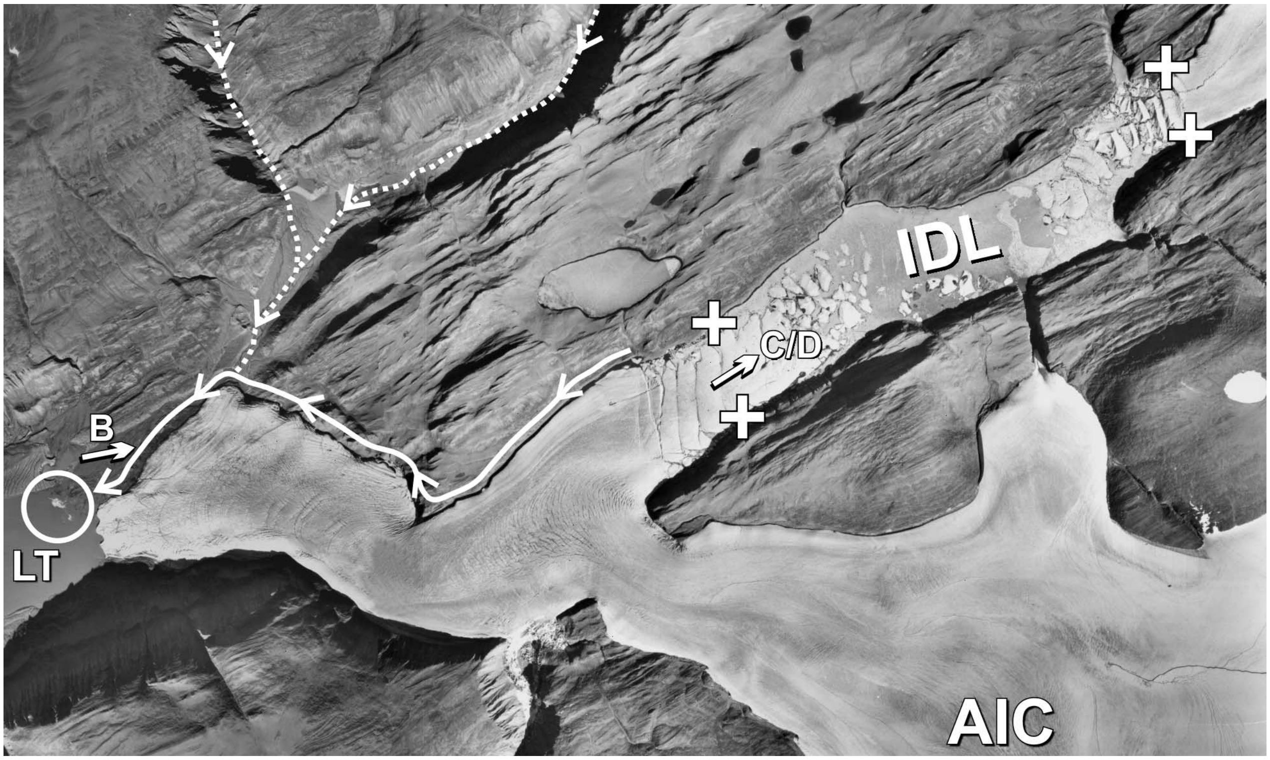
Water column conductivity, temperature, and density were measured with a SBE 19 SEACAT Profiler CTD[3] (Sea-Bird Electronics Inc., Bellevue, Washington, USA). Conductivity is reported as specific conductivity in µS/cm (adjusted by 2%/°C to 25 °C; Ludlam 1996), temperature is reported as potential temperature (T_θ), and density is potential density (σ_θ) referenced to 0 decibars (1 bar = 100 kPa). Brunt-Väisälä frequency (N^2) is a measure of density stratification and resistance to mixing (Wüest and Lorke 2003) and is calculated with 1 m depth bins. CTD casts were plotted in *Ocean Data View* (Schlitzer 2005). A damaged connector prevented the CTD profiler instrument from being lowered deeper than ~80 m in 2003. Dissolved oxygen was measured with a Hydrolab DataSonde 4*a* (Hach Company, Loveland, Colorado, USA).

Transmissivity is measured in percent with a Sea Tech transmissometer (Sea Tech Inc., Corvallis, Oregon). When transmissivity is >10%, it is transformed to suspended sediment concentration (SSC). However, transmissivity from the saltwater layer is not transformed.[4]

Sediment cores were obtained with an Ekman dredge. Thin sections were prepared by following procedures in Francus and Asikainen (2001). Grain size was determined by scanning electron microscopy (SEM) image analysis of thin sections (Francus 1998).[5] Grain size is presented as "equivalent disc diameter" (EDD; Francus 1998). Virtually no microorganisms

[3] See supplementary Table S1 for SEACAT thermistor and conductivity cell accuracies. Supplementary data can be purchased from the Depository of Unpublished Data, CISTI, National Research Council of Canada, Ottawa, ON K1A 0S2, Canada. Data are also available along with the article on the NRC Research Press Web site.
[4] See supplementary Fig. S1.
[5] See supplementary Table S2 for SEM settings and image analysis procedure details.

Fig. 2. The Agassiz Ice Cap (AIC), ice-dammed lake (IDL), and Lake Tuborg (LT) from aerial photographs *a*-16687-47 and *a*-16977-63. Dotted lines mark channels that usually feed the northeast basin of LT, and the solid line represents the IDL overflow route. An "ice raft" in LT on 2 August 1960 (date of the aerial photograph) is circled. Arrows at B and C/D correspond to the locations and directions of photographs shown in Fig. 4. Crosses are the assumed northeast and southwest boundaries of the IDL. Scale is variable due to parallax, but the mosaic is about 15 km × 9 km, and north is up.



were present. Mean grain size was calculated every 500 μm along thin section photomosaics.

Bathymetric soundings (303 measurements) were obtained in Lake Tuborg. Ice-dammed lake dimensions were determined with an aerial photograph (*a*-16687-47, ~1 : 60 000, taken July 1959; Fig. 2).

Results

Climate and weather

From 28 May to 10 August, each year from 2001 to 2003, daily air temperatures at Eureka and Lake Tuborg are highly correlated (r^2 = 0.79, slope = 0.84), with mean air temperature differences within 0.4 °C in all years. In 2001 and 2002, melting degree-days at Lake Tuborg were very similar to the 1948–2002 Eureka mean melting degree-days, but 2003 Lake Tuborg melting degree-days were 23% greater.[6] Much of the anomalous warmth in 2003 occurred in July. Mean Eureka air temperature from 3 to 18 July was 4 °C higher than the 1948–2002 mean for the same period. The year 2003 had the 10th warmest summer on record (1947–2003), with 435 melting degree-days. Rain totalling 37.5 mm fell at Eureka in 2003, the 38th percentile for the 1948–2003 record. Only 28 mm fell at Lake Tuborg during monitoring in 2003, and no rain fell for more than a month prior to the jökulhlaup.

Lake level, discharge, and jökulhlaup observations

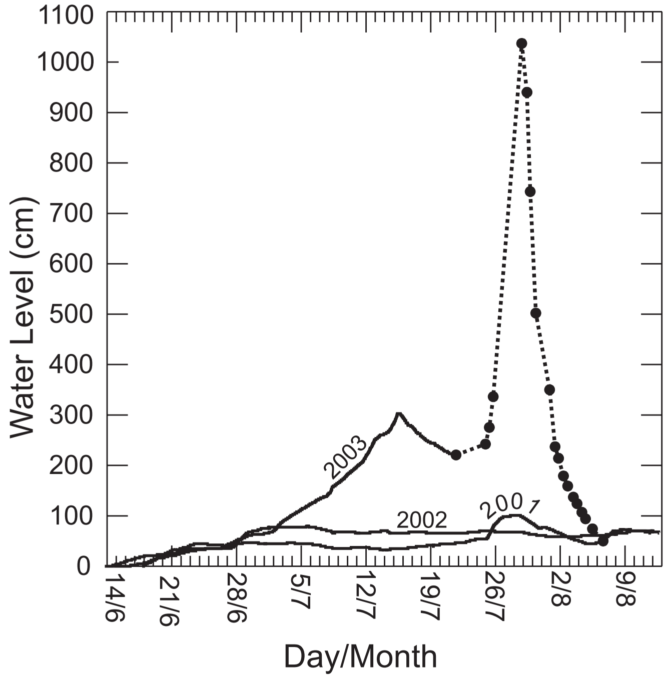
Lake level in all years began increasing in mid-June (Fig. 3) shortly after mean daily air temperature at lake level increased above freezing. Lake level was unusually high beginning in early July 2003. In early to mid-July, the ice-dammed lake overflowed between the Agassiz Ice Cap to the south and bedrock to the north (Fig. 2). Water flowed over an ice ledge, eroding boulder-sized pieces of ice. These ice blocks were then transported to the outwash plain at "*ii*," where they conglomerated into ice rafts (Fig. 2). Sediment-laden ice rafts detached from shore and floated around the freshwater basin as lake level continued to rise (Fig. 4*a*). No "leaking" (Gilbert 1971) of the ice-dammed lake was observed near the glacier terminus. Lake Tuborg lake level decreased by ~0.75 m between 16 and 22 July (Fig. 3) after air temperatures fell following their peak on 4 July 2003.[7]

The ice-dammed lake catastrophically burst, dramatically raising lake level by 7.6 m from 25 July 08:00 to 28 July 20:00 (Fig. 3). To increase Lake Tuborg lake level by this amount, 3.2×10^8 m^3 of water is necessary, providing a minimum estimate of the jökulhlaup volume (see Discussion). The rate of lake level increase requires that average discharge into Lake Tuborg for this period exceeded 1040 m^3s^{-1}.

The jökulhlaup volume can also be estimated assuming the ice-dammed lake morphology is half an ellipsoid. The length,

[6] See supplementary Fig. S2
[7] See supplementary Fig. S2.

Fig. 3. Lake Tuborg 2001–2003 lake levels. Solid lines are automatically logged stilling well data. Circle symbols are lake heights determined with a staff and level.



width, and maximum depth of the ice-dammed lake are 6 km, 1.5 km (Fig. 2), and 116 m, producing a volume of 5.5 × 10^8 m^3. Boundaries from 1959 are broadly similar to those of 2003, based on GPS waypoints obtained at the ice-dammed lake perimeter. Prominent "ice-rims" on the empty ice-dammed lake walls were observed by helicopter on 28 July (Fig. 4*d*). The elevation difference between the uppermost ice-rim and the bottom of the drained ice-dammed lake was 116 m, providing the maximum depth.

Discharge issued from a newly developed glacier portal 980 m from the Lake Tuborg shore (Fig. 4*b*). When the jökulhlaup ended, the portal was ~50 m diameter. The outwash plain near "*ii*" prograded, and Lake Tuborg is now almost completely isolated from the northeast glacier (cf. 1959 photo in Fig. 2). Englacially eroded ice blocks floated as bergs in the freshwater basin during and after the event.

Maximum discharge (Q_{max}) from subglacially draining jökulhlaups is related to the volume drained (V_t, × 10^6 m^3), where $Q_{max} = 46(V_t)^{2/3}$ (Clague and Mathews 1973; Walder and Costa 1996). Q_{max} using the Lake Tuborg lake level V_t is 2200 m^3s^{-1}, and Q_{max} using the V_t from ice-dammed lake morphology is 3150 m^3s^{-1} (see Discussion).

Limnology and sedimentary processes before the jökulhlaup

The northeast, freshwater basin

Before each melt season began, the upper water column (epilimnion) was inversely thermally stratified, and bottom waters (hypolimnion) were relatively warm (e.g., Fig. 5, 30 May cast). The water column was nearly isohaline; bottom water was ~50 mg/L denser than near-surface water (Fig. 5; 30 May cast), and N^2 at the lower thermocline was 4.1×10^{-5} s^{-2}, representing somewhat weak resistance to mixing (cf. Wüest and Lorke 2003).

During the period when the ice-dammed lake was overflowing, the water column was significantly fresher than in the pre-melt season (Fig. 5, 11 July cast). There was a maximum density inversion of 72 mg/L. Water samples at 10 m and near bottom at 30 m had 47 mg/L and 218 mg/L SSC, respectively.

One day before the jökulhlaup began (24 July 2003), two CTD casts were performed on the sill and in the freshwater basin (Fig. 6, stations 5 and 6). Casts were obtained 10 minutes apart, and stations are separated by 1 km. There was no hypolimnion in the freshwater basin. The water column cooled and freshened with depth, producing a density inversion of 88 mg/L. SSC was >23 mg/L through most of the water column. By contrast, on the sill, near-bottom water was significantly warmer, had higher specific conductivity, and was much denser than in the freshwater basin. SSC only reached 16 mg/L in a 5 m thick overflow (Fig. 6, station 6).

The southwest, meromictic basin

Before the jökulhlaup in the saltwater basin, specific conductivity in the upper freshwater layer (mixolimnion) was ~900 µS/cm, and specific conductivity in the deeper saltwater layer (monimolimnion) was ~41 850 µS/cm (~25.01 PSU; Fig. 7). The extremely sharp and strong chemocline creates an abrupt ~20 000 mg/L density contrast between salt and fresh water, and ~0.43 s^{-2} N^2. Wüest and Lorke (2003) cite 0.1 s^{-2} as a typical maximum N^2 in lakes. Before the 2003 melt season began, dissolved oxygen was 10.5–12 mg/L from the surface to 58 m, and dropped to <1 mg/L below 60.5 m.

A sharp decrease to ~5% T_z consistently occurs near the chemocline (Figs. 7, 8, 9). Seven Van Dorn water samples were obtained from 50–59 m in late May 2003, and SSC was <2.5 mg/L in every sample. From the T_z calibration, SSC should have been extremely high[8] (see Discussion).

Response to slush flow events

In 2001, 2002, and 2003, small supraglacial lakes on the Agassiz Ice Cap drained, transporting large volumes of slush and sediment into Lake Tuborg near "*iv*" (cf. Braun et al. 2000). From late May to mid-July 2003, CTD casts were repeatedly performed close to inflow near "*iv*" (Fig. 8*a*, station 8). Before early July, discharge was only from unglaciated terrain at lower watershed elevations, and water column turbidity was <2.5–4 mg/L (Fig. 8*a*). Beginning on 3 July 2003, discharge and sediment transport increased dramatically, and strong interflows formed at 25 and 48 m (Fig. 8*a*). The upper water column became increasingly turbid from the bottom up from 3 to 16 July (Fig. 8*a*).

During peak inflow at "*iv*", several casts were also performed farther from the delta, to determine the extent of processes derived from slush flows in the meromictic basin (Fig. 8*b*, station 9). On 8 July, the transmissometer was overranged below 30 m (Fig. 8*b*), but a near-bottom Van Dorn sample contained 70 mg/L SSC. The next day at the

[8] See supplementary Fig. S1.

Fig. 4. Photographs from Lake Tuborg and the ice-dammed lake (IDL) in 2003. (*a*) Ice raft; arrows point to a large clast and an ice boulder coated in fine-grained sediment. (*b*) The jökulhlaup on 28 July 2003, with the portal circled. (*c*) Full IDL on 18 July. (*d*) Largely empty IDL on 28 July. Note the "ice rims" below the arrow. Locations for (*b*), (*c*), and (*d*) are shown in Fig. 2 as B and C/D. Photos are courtesy of Anders Romundset.

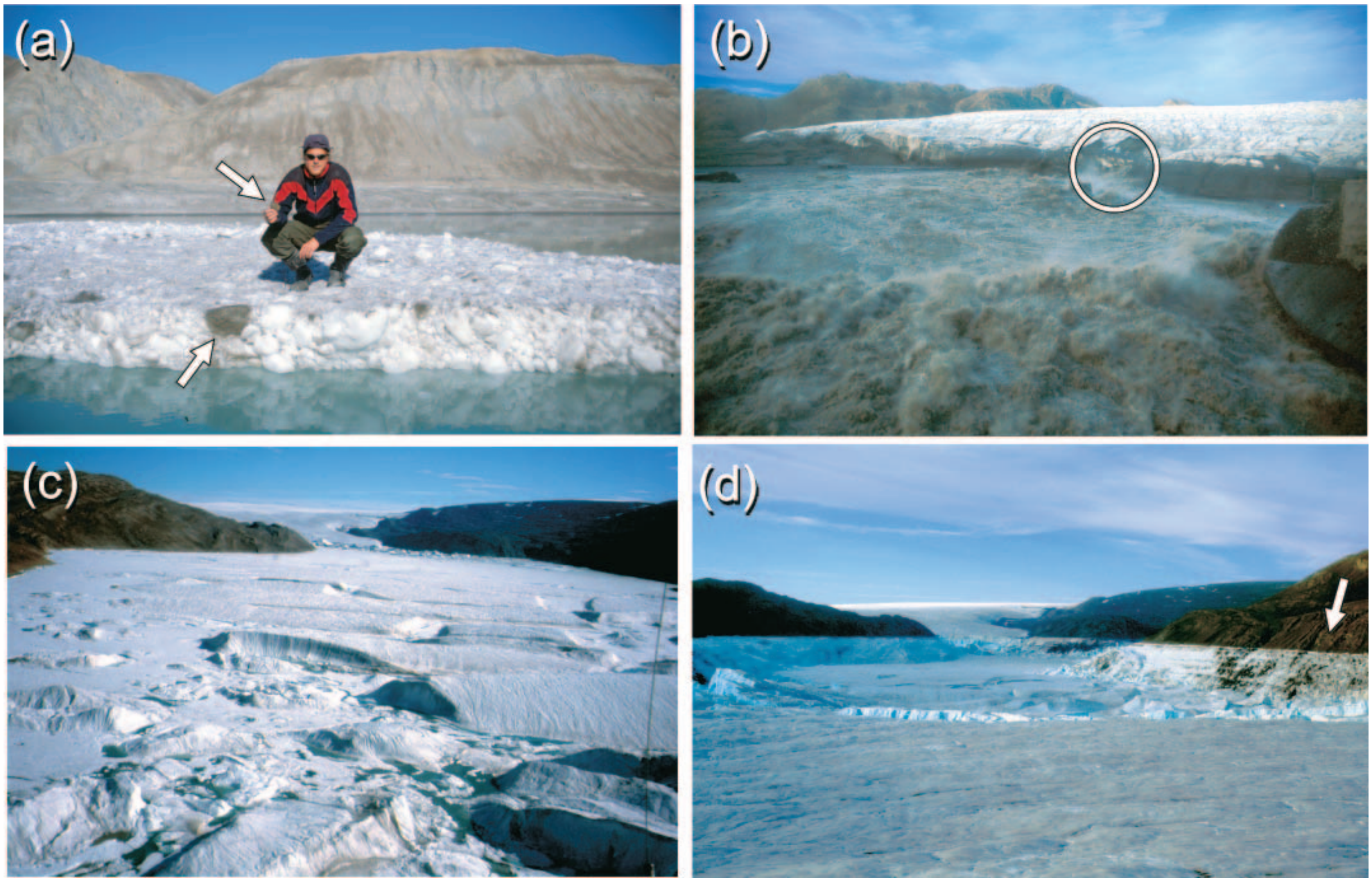
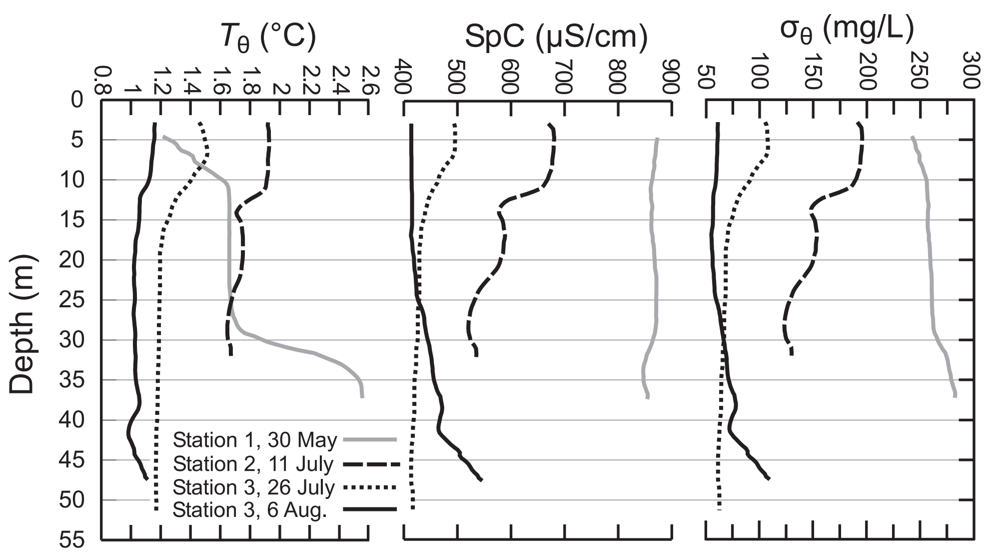


Fig. 5. Freshwater basin CTD casts from before, during, and after the jökulhlaup in 2003. Potential temperature (T_θ), specific conductivity (SpC), and potential density (σ_θ).



same location, the water column was significantly less turbid between 20 and 45 m, and beneath the chemocline (Fig. 8*b*).

Inflow at "*iv*" significantly dropped beginning on 13 July, and continued to wane through the remainder of the melt season. SSC was extremely high at station 8 (near inflow at "*iv*") until 12 July, then the lower water column began to freshen (Fig. 8*a*). The water column cleared from the bottom up, and freshening continued until profiling ended on

Fig. 6. CTD casts from the day before the jökulhlaup began (24 July 2003) in the freshwater basin (station 5) and on the sill (station 6). Potential temperature (T_θ), specific conductivity (SpC), potential density (σ_θ), and suspended sediment concentration (SSC).

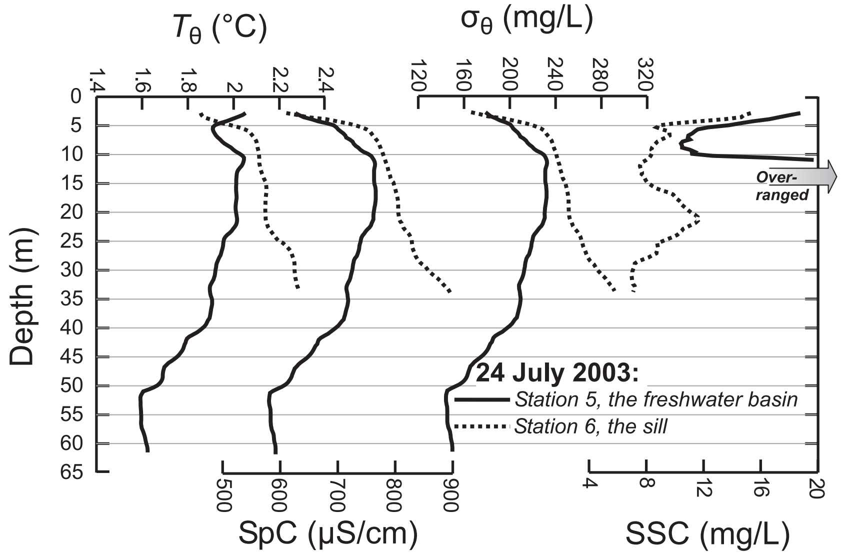
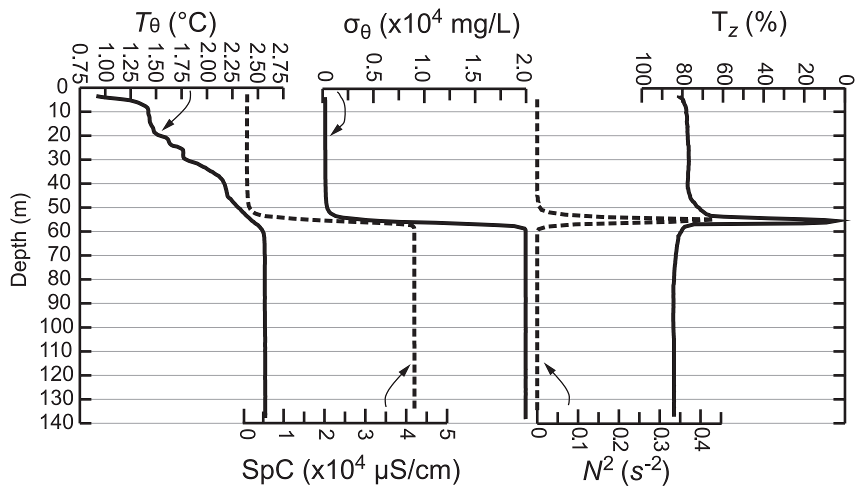


Fig. 7. CTD cast from the meromictic basin (station 13) before the jökulhlaup (10 June 2001): potential temperature (T_θ), specific conductivity (SpC), potential density (σ_θ), Brunt-Väisälä frequency (N^2), and beam transmissivity (T_z; note the reversed *x*-axis).



16 July (Fig. 8*a*). Specific conductivity below the chemocline decreased by ~100 µS/cm near inflow at *"iv"* following the 2003 slush flow events.

Limnology during and after the jökulhlaup

During and after the jökulhlaup in the freshwater basin, near-bottom water samples were very turbid. At station 3 on 26 July, near bottom SSC was 77 mg/L; at station 4 on 28 July, SSC was 1093 mg/L.

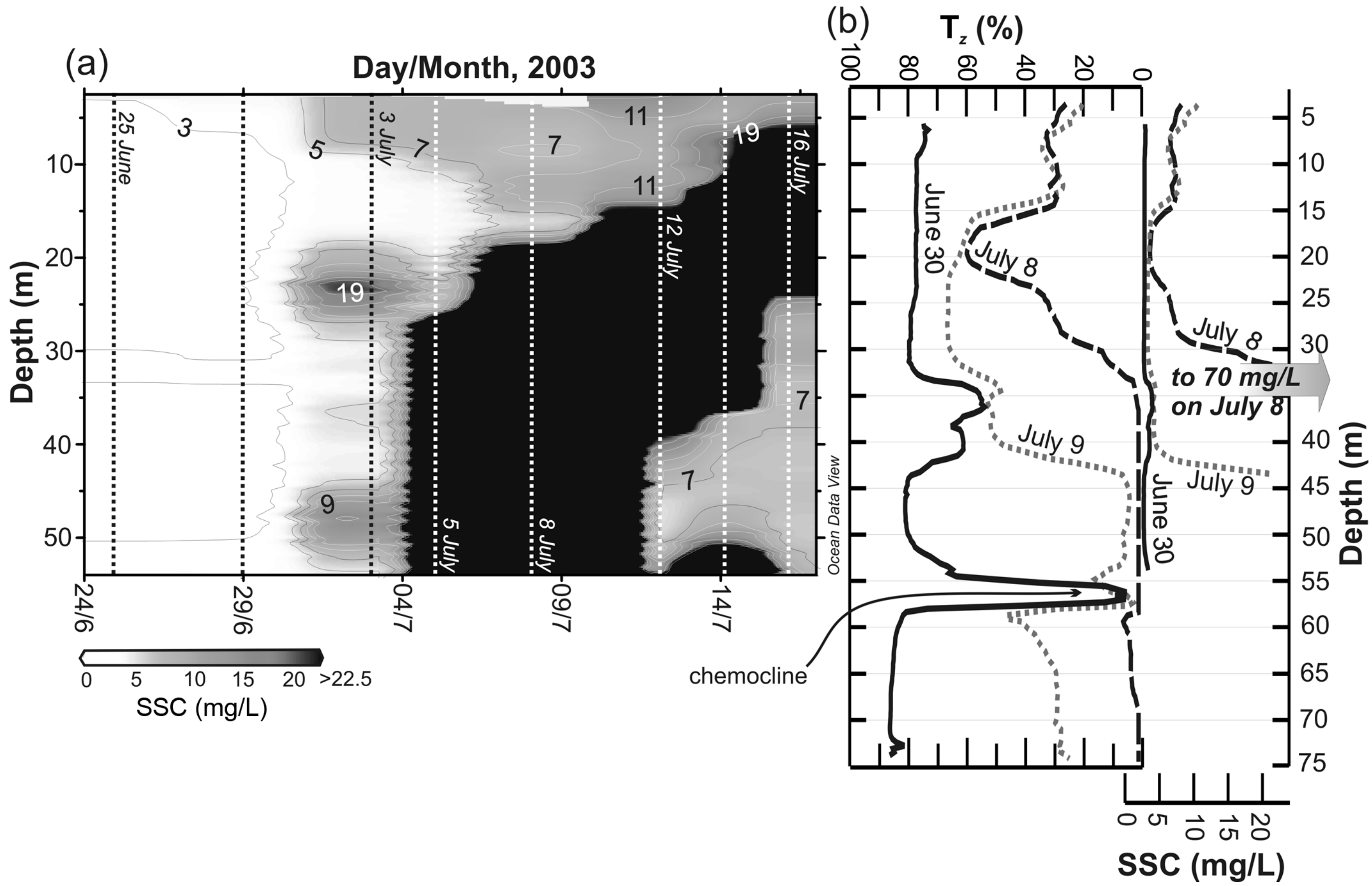
After the jökulhlaup in the saltwater basin, a strong overflow was present in the epilimnion, extending ~9 km from the beginning of a transect near the sill (Figs. 9*a*; 1*c*). Close to the sill, turbid plumes extended as far as ~3 km above and below the chemocline. A very distinct "cold stratum" was present between ~35 m and the chemocline (Fig. 9*b*; cf. Phelps 1996). The epilimnion warmed with distance from the sill.

Pre- and post jökulhlaup temperature and specific conductivity in the meromictic basin are compared in Fig. 10. The "cold stratum" was not present as late as 16 July 2003 (Fig. 10*a*). Although the cold stratum was less thermally dense than the overlying, warmer, epilimnion (Fig. 10*a*), the thermal density deficit was overcome by saltier, more dense water in the lower mixolimnion compared with pre-jökulhlaup conditions (Fig. 10*b*). When the post-jökulhlaup lower mixolimnion is visualized in a vertical section, specific conductivity is seen to slightly decrease with distance from the sill (Fig. 11*b*).

The post-jökulhlaup monimolimnion specific conductivity was ~550 µS/cm lower, and temperature was ~0.04 °C lower compared with pre-jökulhlaup conditions recorded in June 2001 (Figs. 10*c*, 10*d*). As late as 16 July 2003, the monimolimnion was ~2.57 °C, and specific conductivity was ~41 800 µS/cm (Figs. 10*c*, 10*d*).

Downlake temperature and specific conductivity variation in the monimolimnion along the ~10 km saltwater basin transect were extraordinarily low less than two weeks after the jökulhlaup ended (<88 µS/cm and <0.003 °C).

Fig. 8. Limnologic response to the 2003 slush flows. (*a*) Gridded suspended sediment concentration (SSC) at station 8 near inflow at "*iv*." Contour interval is 2 mg/L at ≤9 mg/L and 4 mg/L at >9 mg/L. (*b*) Transmissivity (T_z) and SSC farther from inflow at station 9 on 30 June, 8 July, and 9 July 2003. T_z is not transformed to SSC below the chemocline, or when <10%. T_z decreases to the right.



Dissolved oxygen in the mixolimnion was ~1 mg/L higher in the cold stratum compared with casts from the same location earlier in the melt season. The monimolimnion was not significantly oxygenated by the jökulhlaup.

Near-surface sediments

Before the jökulhlaup, at the end of the period when water was overflowing from the ice-dammed lake (21 July 2003), a surface core was obtained at station 4 in the freshwater basin. At the base of the core, five centimetre-scale units with coarse bases grade upward (Fig. 12*a*, V1–V5). Sediment above ~80 mm is somewhat coarser grained and is less noticeably laminated than sediment below 80 mm. However, three units fine upward, and inter-unit grain size coarsens upwards (Fig. 12*a*, F1–F3).

After the jökulhlaup, on 5 August, another surface core was obtained at the same location (Fig. 12*b*). Laminae are distinctly angled below 95 mm, fine upward, and are relatively coarse. Several dark very fine-grained lenses are present below ~120 mm (Fig. 12*b*, RU), and one has apparently folded internal structure (Fig. 12*b*, RU-F). Sediment is nearly massive above ~95 mm, but very gradually fines upward.

Surface cores were also obtained on the freshwater and saltwater sides of the sill after the jökulhlaup (Figs. 13*a*, 13*b*, stations 5 and 7). Both cores are from roughly the same water depth (62–63 m), and coring stations are 2 km apart. On the freshwater side, sediments are very fine-grained and dense: repeated coring with a weighted Ekman dredge failed to recover more than ~33 mm (Fig. 13*a*). In the core from the saltwater side, there are laminae below ~85 mm, but no clay caps. Texture coarsens upward above 85 mm, but abruptly fines at about 30 mm, above which sediment is massive and slightly fines upward (Fig. 13*b*).

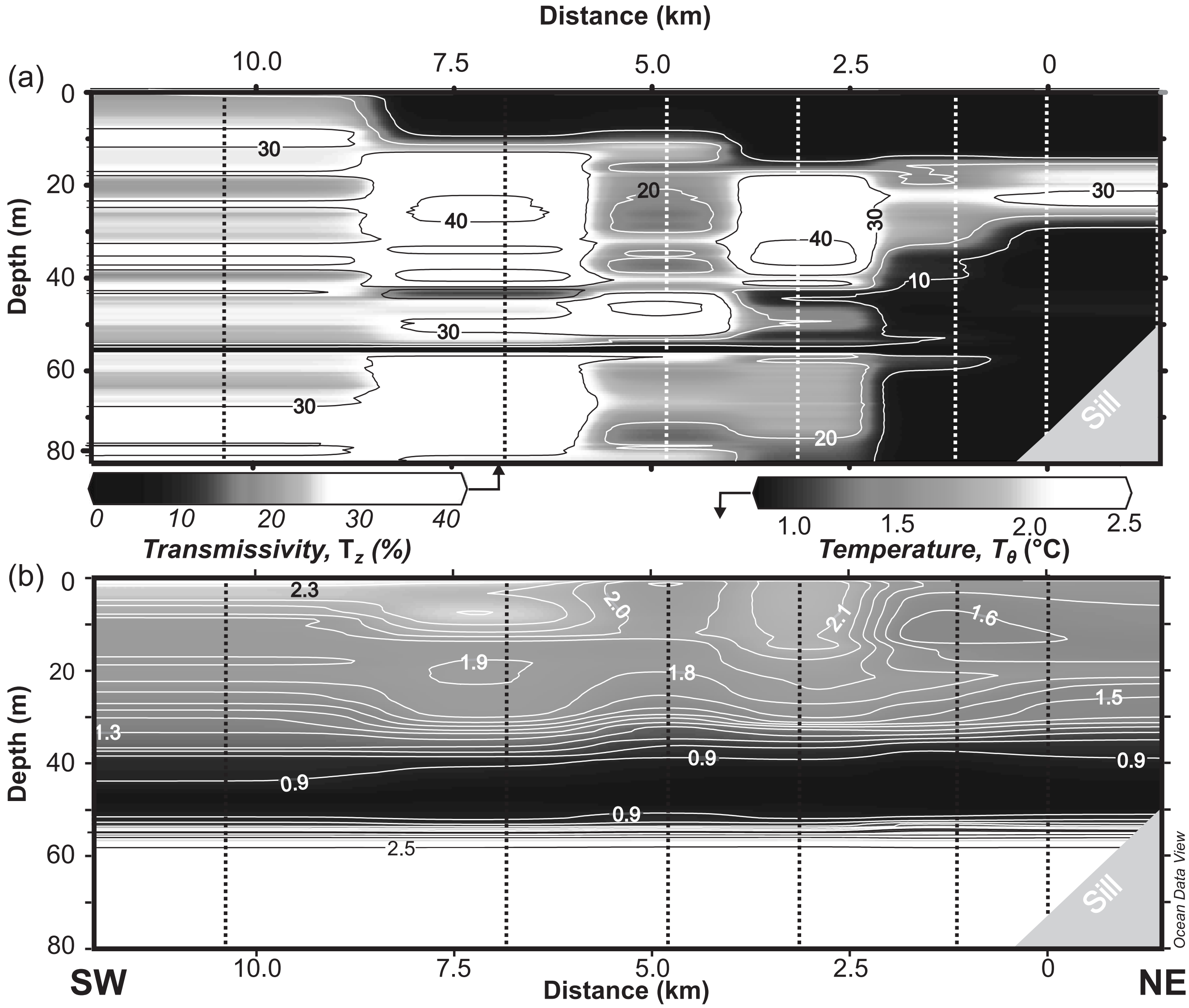
At a much more distal location, four cores were obtained within ~50 m of each other in 2003 in ~135 m water depth (Fig. 14, station 12). All cores, except K, were obtained before the jökulhlaup. The lowermost 60 mm of the stratigraphy (160–220 mm) is largely massive, fine-grained, and slightly fines upward. Sediments above about 160 mm are somewhat coarser, irregularly laminated, and of variable grain size and thickness. Midway through core K, there is a sharp contact, above which sediments are much more coarse and fine upward.

Discussion

Jökulhlaup hydrology

Overtopping of the ice-dammed lake likely contributed greatly to the anomalously high lake levels before the jökulhlaup in early to mid-July 2003 (Fig. 3). The lake level

Fig. 9. Post-jökulhlaup (9 August 2003) vertical sections from stations 9–14 (dotted line in Fig. 1*c*) in the upper 80 m of the meromictic basin. (*a*) transmissivity, T_z, and (*b*) potential temperature, T_θ. Contour interval is 10% for T_z, and 0.1 °C for T_θ. Casts are vertical dashed lines. Isolines are very closely spaced at the chemocline temperature gradient.



decrease after mid-July occurred after air temperature[9] and glacial melt declined, so less overtopping of the ice-dammed lake likely occurred. In 2003, rainfall runoff did not significantly contribute to ice-dammed lake filling or overtopping, nor did it trigger the jökulhlaup.

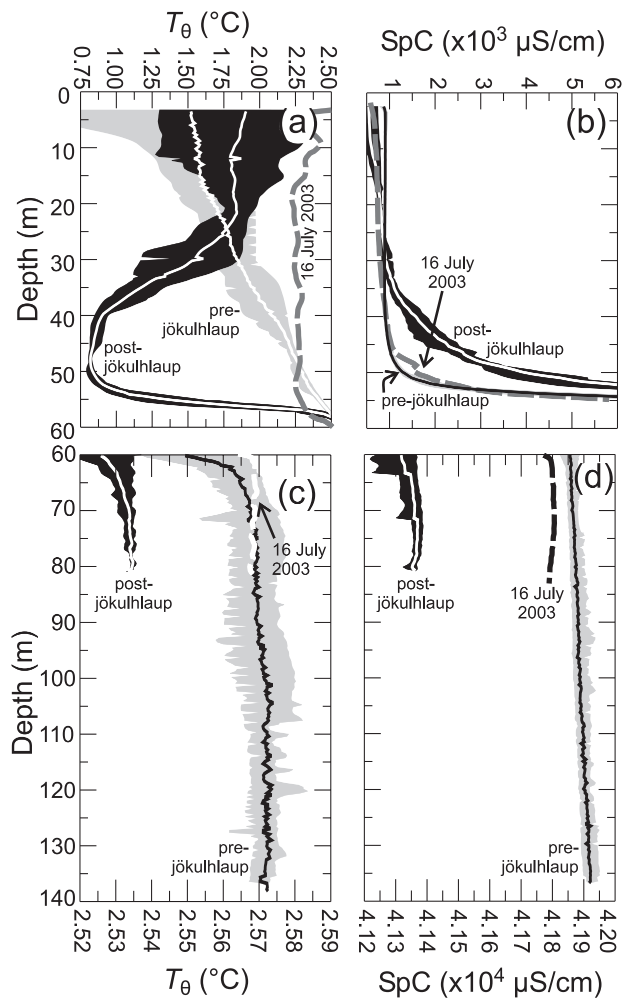
The jökulhlaup was likely triggered by ice-dam flotation, which theoretically occurs when ice-dammed lake depth exceeds 90% of the ice-dam height; however, dam overtopping occurred prior to the jökulhlaup, probably because of dry-based glacier bed adhesion (Roberts 2005). The event likely ended when lake level fell below outlet conduits, the englacial tunnel system was sealed by mechanical blockage, or cryostatic pressure sealed the conduits (Roberts 2005).

Of 125 ice-dammed lakes studied near Expedition Fiord, Axel Heiberg Island (Fig. 1*a*), most drained incompletely by overtopping their dams (Maag 1969). The only other ice-dam flotation jökulhlaup observed in the High Arctic ended by ice-dam resettling, and the lake did not completely drain (Blachut and McCann 1981).

The July 2003 jökulhlaup was the largest recorded in Canada since about 1947 (cf. Clague and Evans 1994; Geertsema and Clague 2005), and the actual volume drained was likely more than the two jökulhlaup volume estimates. Volume calculation using the level of Lake Tuborg does not account for discharge out of Lake Tuborg, which was extreme during the event. Discharge into Lake Tuborg from sources other than the jökulhlaup is also not accounted for. However, watershed snowpack was largely exhausted for about a month prior to the jökulhlaup, and air temperature fell after 19 July, so discharge at "*i*," "*iii*," and "*iv*" was very low. The morphologic

[9] See supplementary Fig. S2.

Fig. 10. Temperature and specific conductivity before and after the jökulhlaup in the saltwater basin. Potential temperature (T_θ ; (*a*, *c*)) and specific conductivity (SpC; *b*, *d*) in the meromictic basin before (grey shading) and after (black shading) the jökulhlaup. Shaded regions encompass maximum and minimum values for each series; solid lines are means. The grey-shaded pre-jökulhlaup series consists of 24 casts from 8 to 24 June 2001, and the black-shaded post-jökulhlaup series consists of seven casts (one from 7 August, and six from 9 August 2003). Post-jökulhlaup casts were completed after lake level regressed, and lake level was very similar in both series (Fig. 3). The thick dashed line is the last cast in the meromictic basin prior to the jökulhlaup, completed on 16 July 2003. Note that abscissa scales on sub-panels showing monimolimnion parameters (*c*, *d*) are very different from abscissa scales showing mixolimnion parameters (*a*, *b*) to better illustrate subtle post-jökulhlaup cooling and freshening of the monimolimnion.



volume estimate is also likely low, since large floating ice wedges (Maag 1969; Blachut and McCann 1981) exist at both ends of the ice-dammed lake, and its length was estimated conservatively (Fig. 2).

The depth of the ice-dammed lake is precise, since ice-rims on the empty lake walls clearly showed maximum lake level (Fig. 4*d*). Ice-rims likely formed as the lake drained, when ice was draped on wave- or ice-cut terraces created in ice-dammed lake filling cycles (Maag 1969).

Underestimating the jökulhlaup volume would result in an underestimation of maximum discharge. However, if the glacier near the ice-dammed lake is cold based, the calculated maximum discharge might be too high, since the Clague–Mathews relationship was determined from warm based jökulhlaups (Walder and Costa 1996).

Processes and deposits in the northeast, freshwater basin

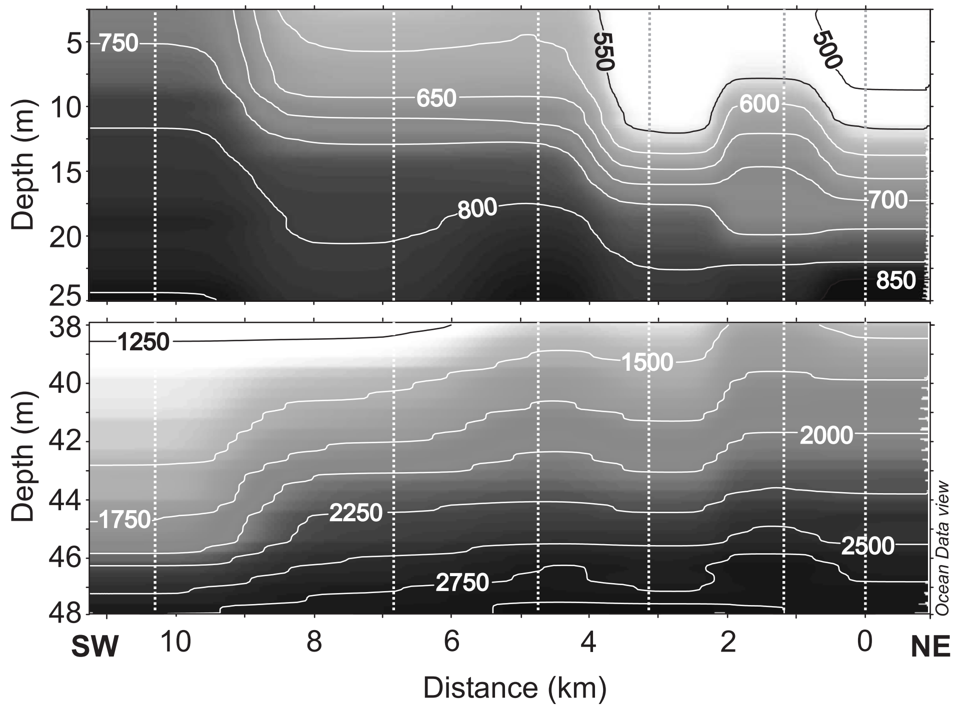
Varves typically accumulate in the freshwater basin. In summer, heightened air temperature raises the Agassiz Ice Cap freezing level, increases glacial discharge and sediment flux to Lake Tuborg (Braun et al. 2000), and turbid sediment plumes disperse sediment through the lake, forming relatively thick and coarse laminae. In winter, fluvial input stops, and clay-sized particles settle very slowly out of suspension, forming a thin clay layer that defines an annual couplet. Water depth at the freshwater basin coring location is deeper than wave base (Håkanson and Jansson 1983), and the temperature gradient between the hypolimnion and metalimnion typically resists mixing (Fig. 5, 30 May cast), allowing undisturbed accumulation of varve couplets. This interpretation is consistent with varve-like structures at the bottom of the pre-jökulhlaup station 4 core (Fig 12*a*) and cesium-137 dated percussion cores from a similar location (Smith et al. 2004).

When water was overflowing from the ice-dammed lake in mid-July, the deepest parts of the northeast basin were filled with cold, fresh, and turbid water (Figs. 5, 6). Bottom water was mixed, removing the weak hypolimnion that was present in May (Fig. 5). This cold, fresh water could only have been from the overflowing ice-dammed lake, since no major streams besides "*ii*" enter the freshwater basin (Fig. 1*b*). Water in the ice-dammed lake was likely near 0 °C, since it is in contact with, and fed almost directly by, the Agassiz Ice Cap, is partially berg and ice-shelf covered, is perennially ice covered, and is about 300 m higher than Lake Tuborg (cf. Maag 1969; Gilbert 1971; Blachut and McCann 1981). Dissolved load and conductivity were also likely low in the ice-dammed lake, since (1) its largest watersheds are highly glacierized; (2) glacial meltwater reaching the lake is very briefly in contact with the ground; and (3) it is above marine limit.

Underflows were present in the freshwater basin two weeks before the jökulhlaup. Their formation was facilitated by the removal of the thermocline. The 11 and 24 July casts are inversely thermally and salinity stratified, producing strong density inversions (Figs. 5, 6) that could only have been compensated for by suspended sediment. This interpretation is supported by highly turbid near-bottom water samples from the freshwater basin while the ice-dammed lake was being overtopped.

Pre-jökulhlaup underflow deposition is recorded in the upper part of the freshwater basin surface core (Fig. 12*a*). Fining upward cycles above 80 mm reflect periods of waning sediment transport and deposition, and were possibly deposited diurnally. Intra-cycle coarsening upward reflects an increase in energy with time, as discharge from the overflowing ice-dammed lake likely increased in early to mid-July (Fig. 3). Amazingly, more sediment was deposited in the deepest part of the freshwater basin in the days to weeks leading up to the jökulhlaup than was deposited in the four preceding years (Fig. 12*a*). Ice-rafted debris rainout also oc-

Fig. 11. Specific conductivity (SpC) vertical sections from after the jökulhlaup on 9 August 2003 (stations 9–14; dotted line in Fig. 1*c*) at two depth ranges. Contour interval is 50 µS/cm in (*a*), and 250 µS/cm in (*b*). Note the greater vertical exaggeration for the lower section. Vertical dashed lines are CTD casts.



curred (Fig. 4*a*). Importantly, the sill separating the saltwater and freshwater basins confined underflows to the freshwater basin prior to the jökulhlaup (Fig. 6, station 6 cast).

Underflows in the freshwater basin during and after the jökulhlaup may initially have been erosive. An unknown amount of sediment separates the pre- and post-jökulhlaup cores at station 4 (Figs. 12*a*, 12*b*). Fine-grained lenses at the base of the post-jökulhlaup core are likely rip-up of semi-consolidated sediment; angled laminae may be cross beds; and texture is coarser than pre-jökulhlaup sediment at the same site (Fig. 12) pointing toward deposition from bottom currents. Closer to the sill, jökulhlaup-derived erosive underflows exposed dense fine-grained sediments (Fig. 13*a*, station 5), probably as flow was constricted, channelized, and accelerated near the southwest end of the freshwater basin (Gee et al. 2001).

Processes and deposits in the southwest, saltwater basin

The monimolimnion is warmed by small amounts of solar radiation that reaches the chemocline through the ice cover (Ludlam 1996; Van Hove et al. 2006). The transmissivity decline at the chemocline (Figs. 7, 8*b*, 9*a*) is probably caused by dissolved sulfur associated with a microbial community or an iron-oxide precipitate (Belzile et al. 2001).

Varve formation would be promoted in the meromictic basin by the anoxic monimolimnion, the high N^2 at the pycnocline, and high circulation rates (Fig. 14, MPJ) relative to many other High Arctic varved lakes. However, it is possible that all fine-grained sediment flocculates and is deposited before winter, preventing the regular formation of clay-caps (Gilbert 2000).

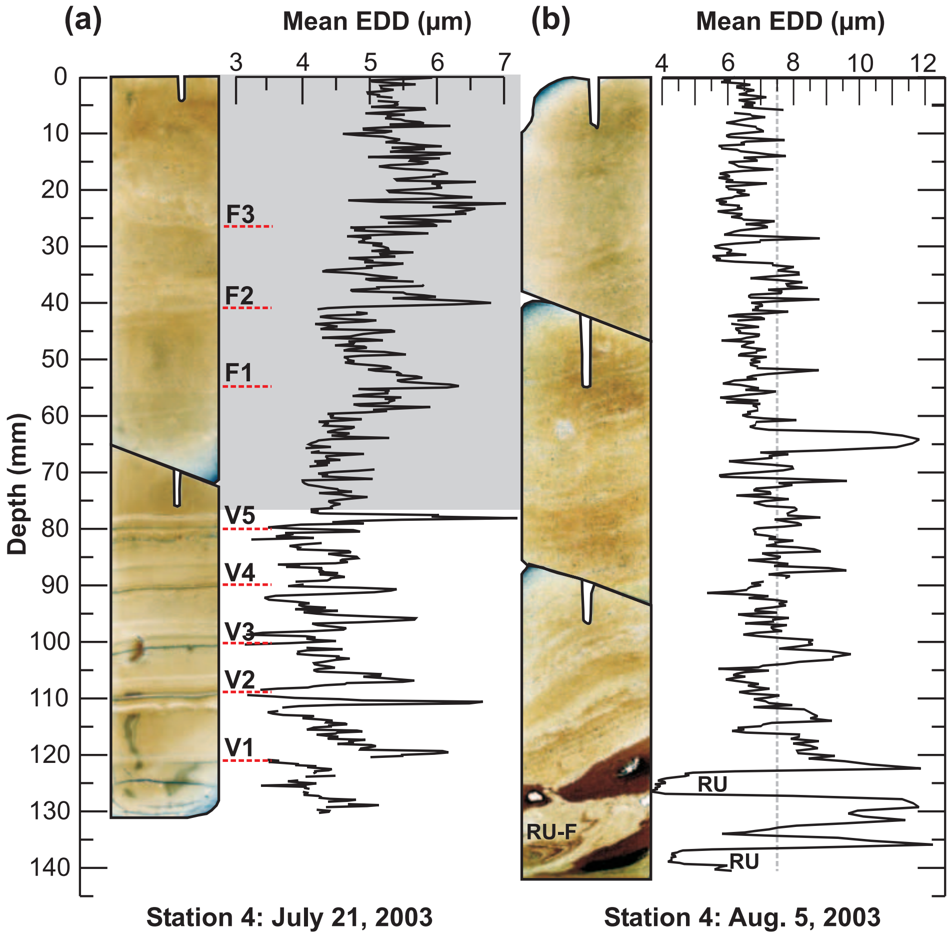
During and after the jökulhlaup, cold ice-dammed lake water flowed through the freshwater basin, overtopped the 34 m sill, then flowed along the chemocline and reached the southwest end of the lake (Figs. 9*b*, 10*a*). Flow on boundary layers promotes the development of internal waves that can eventually oversteepen, forming Kelvin-Helmholtz billows. Billows form when the gradient Richardson number ($R_i = N^2/(du/dz)^2$) is <0.25 (*du/dz* is velocity shear; McCool and Parsons 2004). Billows rapidly collapse due to internal instability, resulting in a thickened boundary layer (Geyer and Smith 1987). This interpretation is consistent with the post-jökulhlaup chemocline thickening (Fig. 10*b*), and the decrease in conductivity with distance slightly above the chemocline (Fig. 11*b*). Flow at the chemocline would have waned with distance, and distal billows would have had smaller length scales. In the Fraser River estuary, chemocline thickening has been recorded under similar physical conditions, and Kelvin-Helmholtz billow length scales decreased distally (Geyer and Smith 1987).

In the meromictic basin, extremely high shear is required for R_i to drop below 0.25 because of the exceptionally high N^2 at the chemocline. This is why the chemocline thickening was subtle (Fig. 10*b*), and monimolimnion cooling and freshening was minimal (Figs. 10*c*, 10*d*).

The Lake Tuborg chemocline remains anomalously sharp compared with most other High Arctic meromictic lakes (Ludlam 1996; Van Hove et al. 2006) because flows along the chemocline during jökulhlaups disturb vertical diffusion by molecular diffusion and wind-generated mixing (Toth and Lerman 1975; Van Hove et al. 2006).

The depth of the Lake Tuborg chemocline is also regionally anomalous, and was thought to be controlled by the ice-contact depth (Ludlam 1996), but shallow bathymetry at both ends of the lake discounts this possibility (Fig. 1*c*). Rather, chemocline depth is a function of the sill depth, shear, N^2, and turbidity.

Fig. 12. Ekman cores obtained from the same location before and after the jökulhlaup. Coring location is station 4, depth is 74 m. (*a*) Thin sections and grain size from a core obtained on 21 July 2003. Sediment that was likely deposited during the period of ice-dammed lake overflow is shaded. V1–V5, clay-caps; F1–F3, bottoms of fining-upward cycles. (*b*) Thin sections and grain size from a core obtained on 5 August 2003. RU, rip-up; RU-F; folded rip-up. Images are flatbed scans of thin sections in cross-polarized light, with contrast and brightness enhanced to better illustrate structures. EDD, equivalent disk diameter (Francus 1998).

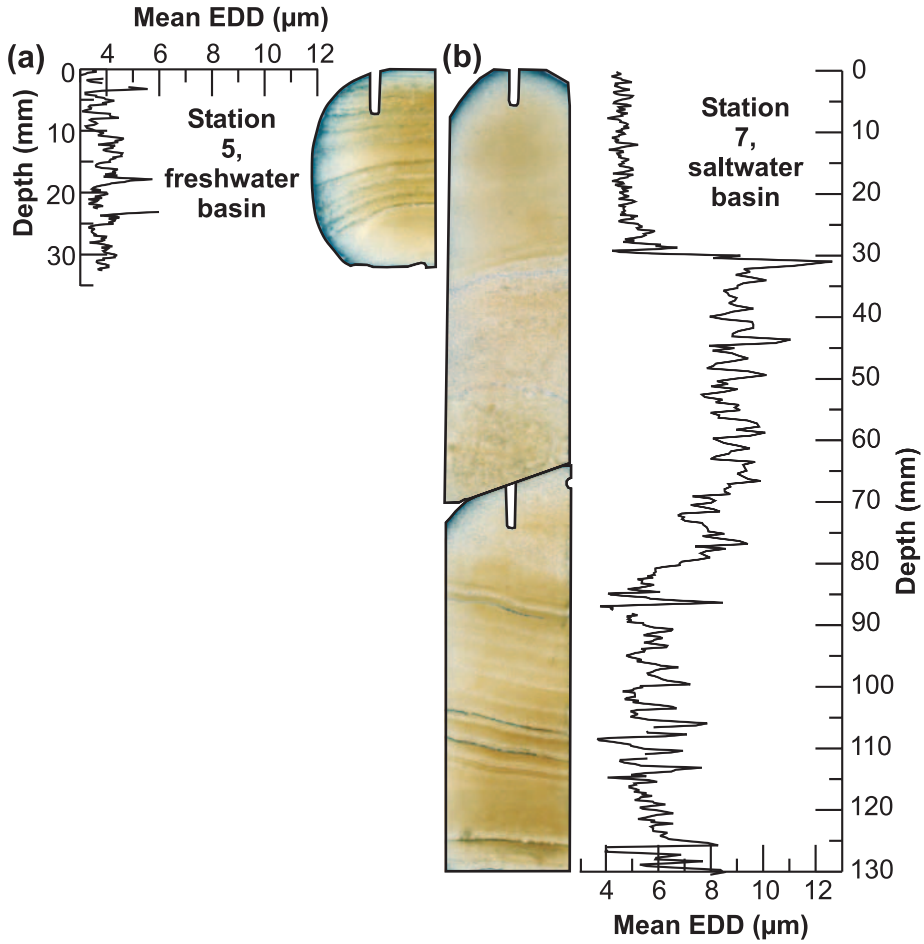


Interflows progressed much less far into the meromictic basin than overflows (Fig. 9*a*). Kelvin-Helmholtz billows would have mixed turbid fresh water with salt water, allowing fine-grained particles to flocculate and drop from suspension.

Jökulhlaup-derived sedimentary processes near the sill in the meromictic basin are inferred from the station 7 core (Fig. 13*b*). The bottom slope is relatively steep, and the chemocline is only 7 m above the lake bottom at this site. Several processes could have deposited the thin, coarse laminae in the core: (1) Downward propagating internal waves could have redistributed sediment. Little energy is required to do this, since the monimolimnion has near zero N^2 (Fig. 7; Smyth, personal communication). (2) A slump could have been triggered either by internal waves on the chemocline hitting the sill (Rimoldi et al. 1996) or by freshly deposited sediment oversteepening the sill. (3) Settling-driven convection. Hyperpycnal flows could have been generated as particles accumulated at the chemocline until their concentration exceeded the density of the monimolimnion (Hoyal et al. 1999; Parsons et al. 2001; McCool and Parsons 2004). It is noted that double-diffusive convection (Parsons et al. 2001) was not possible at the chemocline because the cold stratum was superimposed on the warm monimolimnion—a thermally stable situation at temperatures less than maximum water density. (4) "River-generated" turbidity currents with interstitial fresh water. When turbidity currents were insufficiently thick to overtop the sill, they would have suddenly been restricted to the freshwater basin, and saltwater basin deposition would have been solely from interflows and overflows (Chikita et al. 1996). This is a plausible explanation for the abrupt decrease in grain size above 30 mm depth (Fig. 13*b*). However, classic river-generated turbidity currents with interstitial fresh water require extremely high SSC to overcome density deficits created by ambient salt water, and are globally very rare (Gilbert 2000; Mulder et al. 2003). At Lake Tuborg, turbidity current SSC would have to exceed 20 000 mg/L for SSC to exceed this density deficit (Fig. 7). Although the most turbid near-bottom water sample in the fresh water basin during the jökulhlaup only contained 1100 mg/L, fluvial SSC in jökulhlaups has exceeded 200 000 mg/L (Mulder et al. 2003).

The process(es) responsible for the laminae in the station 7 core did not progress into the deep, distal parts of the saltwater basin. Interflows were not spatially extensive in the

Fig. 13. Ekman cores from both sides of the sill obtained after the jökulhlaup (11 August 2003). Both cores were taken in 62–63 m water depth. (*a*) Thin sections and grain size of a surface core obtained on the freshwater side of the sill at station 5. (*b*) Thin sections and grain size of a surface core obtained on the saltwater side of the sill at station 7. Images are flatbed scans of thin sections in cross-polarized light, with contrast and brightness enhanced to better illustrate structures. EDD, equivalent disk diameter (Francus 1998).



saltwater basin after the jökulhlaup (Fig. 9). The extent of underflows is unknown, but near-bottom processes are inferred from surface core K (Fig. 14). The coarse, mostly massive, fining upward cap (Fig. 14, *J03*) was deposited during and after the jökulhlaup, as it is not present in cores F, A, or B. The cap was deposited non-erosively, and underlying sediments were not disturbed (Fig. 14). The ~90 m between the chemocline and lake bottom would minimize the potential for erosion by downward propagating internal waves. Slump and hyperpycnal flow progression to the coring site is unlikely because of (1) the great distances between the coring location, sill, and jökulhlaup inflow; (2) the 1° average bed slope between the sill and saltwater basin; and (3) the high monimolimnion salinity and density.

The nearly massive structure of the K cap points to deposition by sediment rainout from hypopycnal flow. Sediment was likely rapidly deposited at the site from the upper monimolimnion and lower mixolimnion by flocculated sediment that settled faster than rates predicted by Stokes Law.

Supraglacial lake drainings above "*iv*"

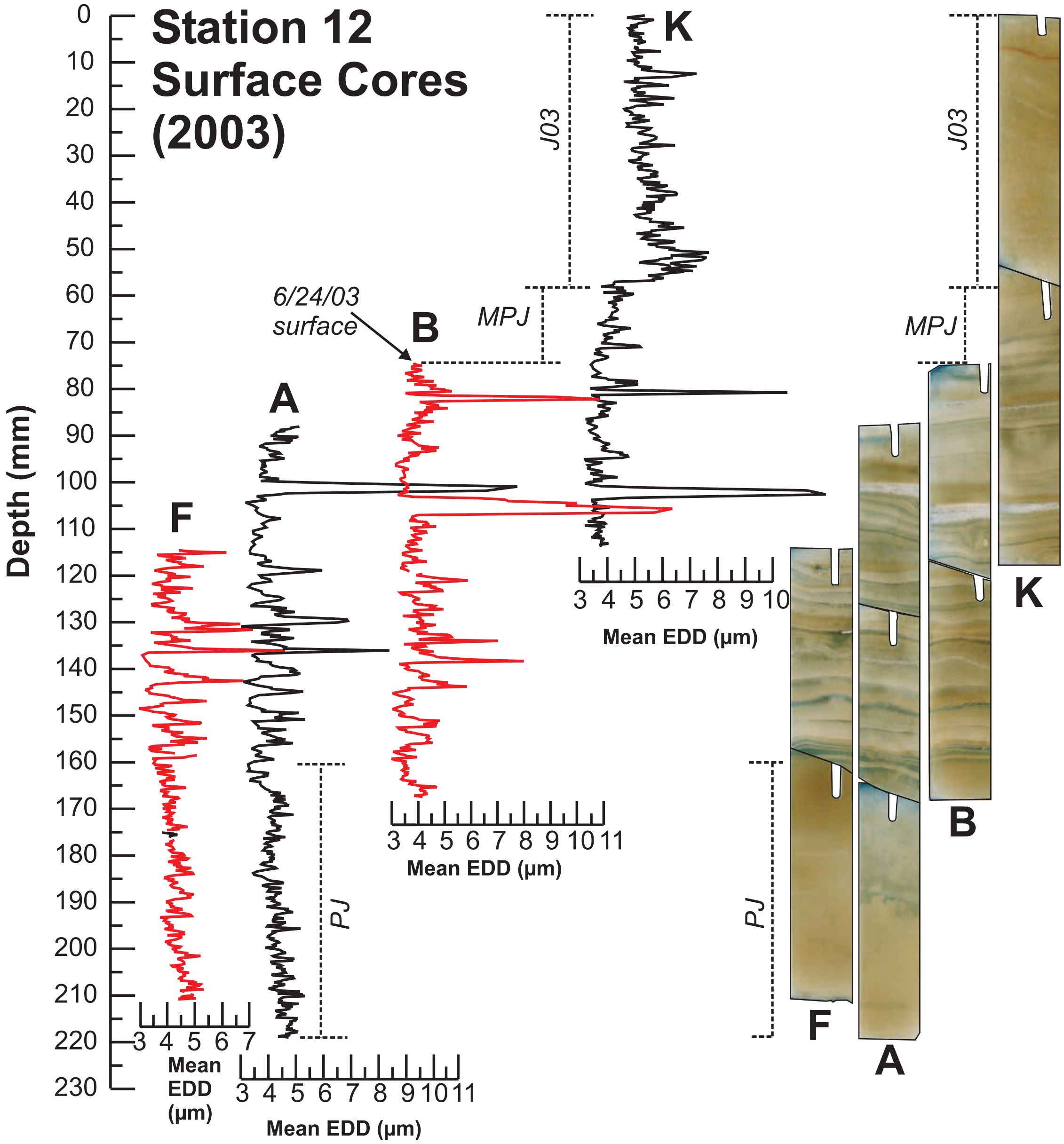
Repeated slush flow events are capable of very slightly freshening the monimolimnion during peak glacial melt in early July (Fig. 10*d*; 16 July 2003 cast). However, sediment plumes associated with these events are not as extensive as jökulhlaup-derived plumes (cf. Figs. 8, 9*a*). Supraglacially derived plumes are not capable of producing the distinctly thick, nearly massive deposit in the post-jökulhlaup core at station 12 (Fig. 14, *J03*), nor of producing the distinct post-jökulhlaup freshening, cooling, and mixing that occurred after the jökulhlaup (Figs. 9, 10, 11).

Supraglacial lakes above "*iv*" are small compared with the ice-dammed lake. The water volume discharged in the entire 45-day 1995 monitoring season (Braun et al. 2000) was 60 and 105 times less than the two ice-dammed lake estimates of the jökulhlaup volume, but only about 20% of the 1995 discharge occurred during slush flow events. Q_{max} in 1995 was ~16 m^3s^{-1} near "*iv*," compared with 2200–3150 m^3s^{-1} from the jökulhlaup.

Previous jökulhlaups

The ice-dammed lake drained in 1993 (R.M. Koerner, personal communication, 1996). Detailed measurements were next made at Lake Tuborg in 1995, when a cold stratum remained in the saltwater basin (Phelps 1996). Observations in 1995 included kettle holes, a glacier portal near "*ii*," and a

Fig. 14. Stratigraphy of Ekman cores and grain size from a ~135 m deep, distal location in the meromictic basin (station 12). All cores, except K, were obtained before the jökulhlaup in 2003. *J03*, sedimentation from the 2003 jökulhlaup; *MPJ*, minimum 2003 sedimentation prior to the jökulhlaup; *PJ*, likely the top of the previous jökulhlaup deposit. The sediment surface in A and F was purposely lost to retrieve cored sediments deeper than ~13 cm (the length of the subsampling tubes). The last core (except K) obtained was B, on 24 June 2003. Images are flatbed scans of thin sections under cross-polarized light, with contrast and brightness enhanced to better illustrate structures. EDD, equivalent disk diameter (Francus 1998).



raised shoreline on marine mud terraces on the north shore of the freshwater basin (M.J. Retelle, personal communication, 2005). It is likely that the lower fining upward sequence in the stratigraphy (Fig. 14, *PJ*) is the upper part a deposit from the 1993 jökulhlaup.

A jökulhlaup also occurred between 1960 and 1963. Air photos from August 1960 show a full, overflowing ice-dammed lake, and an ice-raft near "*ii*" (Fig. 2). Bergs were observed at Lake Tuborg in 1963 (Hattersley-Smith and Serson 1964). There was no cold stratum in 1963 (Hattersley-Smith and Serson 1964), so the jökulhlaup likely occurred closer to 1960 than 1963.

Conclusions

A large and rare jökulhlaup drained into Lake Tuborg in July 2003 while a detailed lake process study was being conducted. The jökulhlaup drainage style, and the effects on the limnology and sedimentology of Lake Tuborg have been described.

Before the jökulhlaup, a weak thermally derived hypolimnion was removed in Lake Tuborg's proximal freshwater basin. This facilitated the formation of hyperpycnal flows. During and after the jökulhlaup, underflows eroded the lake bottom in parts of the freshwater basin, exposing dense, laminated sediments, and creating an unconformity. However, some areas of the freshwater basin experienced high sedimentation rates, and cross-beds and fine-grained rip-ups were deposited.

Alternatively, in the saltwater basin, underflows were limited to steep and proximal areas close to the sill. More distally, jökulhlaup-derived sedimentary processes were not erosive, and a unique thick, coarse, fining upward facies was deposited, probably because of the low bottom slope and the great depth between the chemocline and lake bottom. Underflow run-out distances were limited by rapid deposition from saltwater flocculation.

The monimolimnion has remained remarkably salty for ~3 ka, and through at least three jökulhlaups (1960–1963, 1993, 2003). Mixing of the monimolimnion has been limited by the extremely high N^2 at the chemocline.

Monitoring has also allowed the sedimentary signal and limnologic consequences of jökulhlaups to be disentangled from those of slush flows from supraglacial lakes (Braun et al. 2000). It should therefore be possible to unambiguously identify jökulhlaup deposits in the long core record, and produce a millennial-scale undisturbed reconstruction of jökulhlaup frequency, which would test long-term models of jökulhlaup frequency and magnitude (Clague and Evans 1994).

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