HYDROLOGICAL AND METEOROLOGICAL OBSERVATIONS AT LAKE TUBORG, ELLESMERE ISLAND, NUNAVUT, CANADA[1]

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Abstract: Hydrological and meteorological observations at Lake Tuborg, Ellesmere Island, Nunavut, Canada in 1995 are used to investigate contemporary water and sediment transport processes. Here we describe a new environmental data set for the High Arctic, where such data are scarce. The studied watershed (~460 km^2) ranges in elevation between 63 and ~1900 m asl and is 88% covered by a lobe of the Agassiz Ice Cap. Streamflow and sediment transport were strongly associated with snowmelt runoff, whereas the direct influence of summer precipitation events was negligible. Snowmelt was primarily controlled by synoptic-scale climatic processes. Two high-magnitude pulses of meltwater and slush contributed a significant portion of the measured suspended sediment load to Lake Tuborg. Such events may be associated each year with snowmelt along the Agassiz Ice Cap margin. Additional years of data collection are needed to define the annual and inter-annual variability of the sediment delivery system, particularly with respect to the relative importance of summer rainfall events. Runoff and sediment transport to Lake Tuborg are very likely to increase under climatic warming conditions.

INTRODUCTION

The Canadian High Arctic comprises a vast island archipelago of ~1.3 million km^2 with ~25% glacier cover and nearly perennial snow cover elsewhere (Hodgson, 1989). Climatological data for this remote region are scarce and presently limited to three permanent weather stations, all located near sea level in coastal areas. The interior and mountainous High Arctic climate remains essentially unmonitored (Edlund and Alt, 1989), with the exception of largely unpublished data collected by university-based researchers and Canadian governmental agencies (e.g., Atkinson et al., 2000). Here we present a new High Arctic environmental data set of hydrological and meteorological data from Lake Tuborg, Ellesmere Island, Canada (Fig. 1). We are specifically interested in the extent to which land-to-lake water and sediment transport are related to summer climate. The objectives of the study therefore were: (1) to collect hydrological and meteorological data at sufficient temporal resolution to

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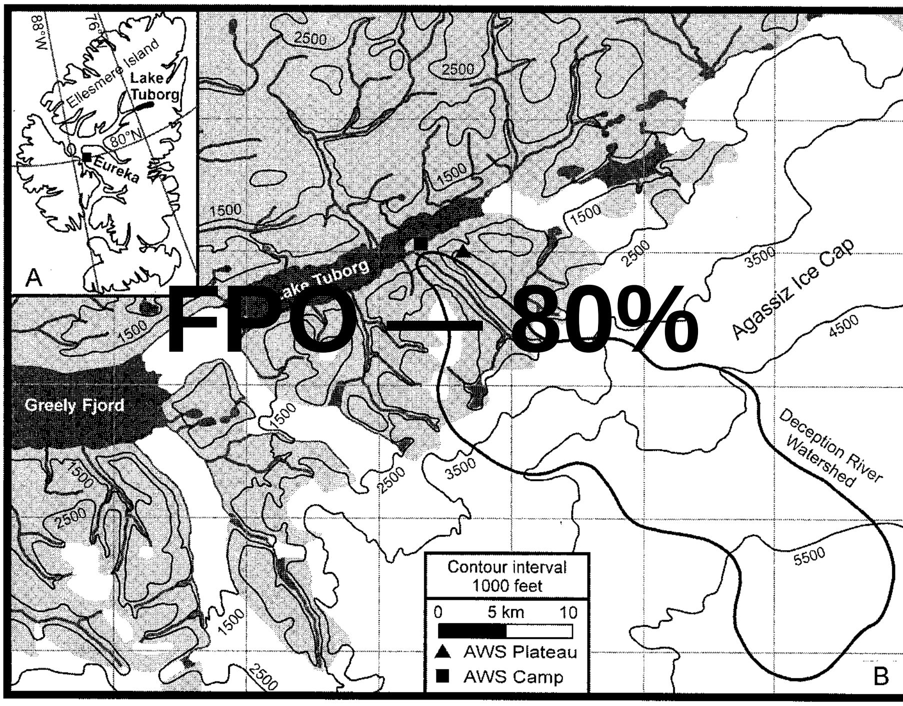

Fig. 1. A. Ellesmere Island, Canada. B. Lake Tuborg and Deception River watershed, Ellesmere Island. Weather station Camp and the hydrological monitoring station were located at the watershed outlet (after NTS Canada map 340D, 1:250,000).

investigate relationships between summer climate, streamflow, and suspended sediment transport; and (2) to place the observations from one summer into a broader regional context. This study is part of our ongoing efforts to quantify the paleoclimate signal preserved by annually laminated (= varved) lake sediments (e.g., Hardy et al., 1996; Smith, 1997; Francus et al., 1999; Braun et al., 2000). In lieu of long hydrological data series, the "calibration" technique typically applied to lake sediments involves statistical associations between laminae variability through time (e.g., varve thickness) and the nearest available climatic data (e.g., mean summer temperature) (e.g., Leonard, 1985; Deslodges, 1993; Hughen et al., 2000). Unfortunately, without a clear understanding of the processes involved in water and sediment transport, spurious associations may be found (cf. Perkins and Sims, 1983; Stihler et al., 1992). Our 1995 field season was primarily reconnaissance work, since the magnitudes and intensities of the fluvial processes in the Lake Tuborg watershed were unknown. Measurements focused on the "Deception River" (unofficial name) watershed, encompassing a portion of the Agassiz Ice Cap (Fig. 1).

STUDY AREA

Lake Tuborg was formed ~3000 years ago (Long, 1967), when an ice advance northward from the Agassiz Ice Cap isolated the inner part of Greely Fjord (Fig. 1). The lake occupies a deep SW-NE–trending valley at 11 m above sea level (asl). The

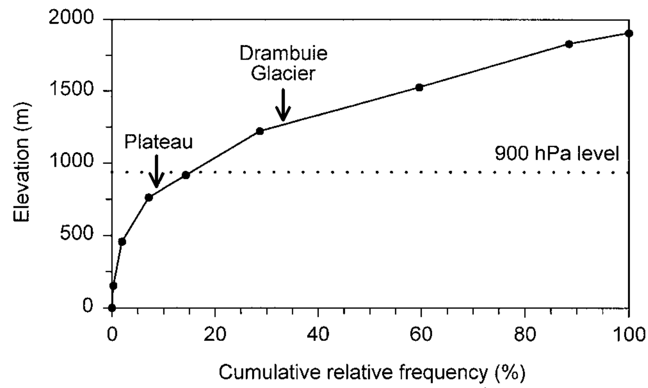

Fig. 2. Relative area-elevation curve for the Deception River watershed, Lake Tuborg, Ellesmere Island (topographic information after NTS Canada map 340D, 1:250,000).

Cape Rawson Mountains surround the lake on the north and south and rise steeply to form a large plateau surface above 800 m asl. The lake is ~20 km long, between 1.5 and 3 km wide, up to 150 m deep, and laminated lake-bottom sediments are preserved due to density stratification (Smith, 1997). Lake Tuborg is a complex hydrological system because meltwater and sediment are derived from three source areas (Fig. 1). Deeply incised, V-shaped valleys drain a large, non-glacierized plateau to the north, whereas two large outlet glaciers extend nearly to, or into, the lake on its eastern and western end. The extent of subglacial meltwater and sediment input is unknown, but field evidence indicates that the glaciers are warm-based at low elevations and subglacial contributions may be significant. Meltwater and sediment also reach the lake from several large lateral meltwater and supraglacial stream channels; evidence exists for episodic jokulhaup-like flood events. Drainage from the south enters Lake Tuborg via proglacial streams originating on the Agassiz Ice Cap margin, including Deception River.

The Deception River watershed (~460 km^2) is 88% glacierized by the Agassiz Ice Cap (Fig. 1). The vertical relief is ~1800 m and over 90% of the watershed is above 800 m asl, either as part of an unglacierized plateau or the ice cap (Fig. 2). It is difficult to ascertain the altitude up to which snowmelt and ablation on the Agassiz Ice Cap actually contributed to runoff (“effective area”) and watershed area; relief, percent glacierization, and hypsometry were calculated on the basis of topographic map interpretation. The apparently cold-based terminus of the watershed outlet glacier was at ~350 m asl where we did not observe subglacial meltwater drainage, nor evidence thereof. All meltwater originated on the ice-cap surface, draining via several deep supraglacial channels; meltwater contribution from lateral meltwater channels was not observed. The proglacial stream follows a fairly straight, V-shaped valley flanked by steep talus slopes, and debauches onto a large fan-delta before entering Lake Tuborg. Numerous erosional surfaces of different ages can be identified on the fan-delta, indicating past changes in stream transport capacity, channel configuration, and/or lake level. The active stream channel is incised several meters into the next older delta surface.

The watershed is underlain by easily erodable folded sandstone and mudrock with minor carbonates (Early Cambrian to earliest Devonian Hazen and Danish River

TABLE 1

Monthly Mean Air Temperature and Total Precipitation for Eureka and Lake Tuborg

Temperature, °C	May	June	July	Aug.	Sept.
Eureka (10 m asl)[1]	-10.7	1.8	5.4	3.3	-8.3
Eureka (10 m asl)[2]	-8.2	3.2	7.3	3.6	-4.1
Eureka (900 hPa)[2]	-10.4	-1.0	4.2	1.0	-5.6
Camp (63 m asl)[2]	*	4.1	*	*	*
Plateau (800 m asl)[2]	*	0.3	*	*	*
Drambuie Glacier (1260 m asl)[3]	-13.3	-4.7	-1.7	-3.6	-12.2
Precipitation, mm					
Eureka (10 m asl)[1]	3.2	5.4	12.1	11.6	9.6
Eureka[2]	7.2	6.5	0.4	11.2	20.2
Camp (63 m asl)[2]	*	10.9	*	*	*

[1]1950–1981 (AES, 1984).
[2]1995 (this study).
[3]1995 (Koerner, Alt, and Labine, pers. comm.).
*Incomplete data.

Formation) (Trettin, 1989). The stream valley slopes contain large amounts of sand and silt and constitute an important source area for sediment during the initial snowmelt period. Large amounts of fine-grained sediment are also available within the proglacial stream channel complex through channel erosion and bedload transport. Frequent, localized casts or drapes of silt, sand, and gravel marked sediment deposition by abandoned channels, indicating channel braiding, channel migration, or past slushflow paths. The channel configuration on the active floodplain became unstable during the observed high flows in 1995, resulting in channel shifts and lateral channel migration on subhourly time scales. Point sources for suspended sediment such as mud flows or debris flows were not observed in the watershed in 1995.

Central Ellesmere Island is protected from the immediate influence of cold Arctic Ocean air by the mountains and high plateaus of northern Ellesmere and Axel Heiberg Island (Edlund and Alt, 1989), and is often referred to as a “Thermal Oasis” (Courtin and Labine, 1977). Precipitation at Eureka (Table 1) is the lowest recorded anywhere in Canada because of the rain-shadow effect. The climate at Eureka is also moderated by the proximity to Eureka Sound (Fig. 1), whereas Lake Tuborg (~200 km northeast) and the interior parts of Ellesmere Island experience a more “continental” climate (Lewkowicz and Wolfe, 1994), albeit possibly influenced by the proximity to the Agassiz Ice Cap.

METHODOLOGY

Meteorology

We established two automated weather stations (AWS) to assess elevational gradients in summer climate. In watersheds with large vertical relief, the high-elevation zones often represent the largest source areas for snowmelt, yet can experience different climatic conditions than low-elevation areas, where research camps are often located (Hardy, 1996). AWS "Camp" (80°58.32' N, 75°23.09' W, 63 m asl, May 18–July 20, 1995) was located near the stream gauging site, whereas AWS "Plateau" (80°58.30' N, 75°14.35' W, 800 m asl, May 28–July 20, 1995) was operated on the upper, non-glacierized plateau (Fig. 1). Measurements at both stations included air temperature, relative humidity, and wind speed/direction. In addition, radiation balance, atmospheric pressure, and precipitation were recorded at Camp. We used data from the "Drambuie Glacier" (unofficial name) AWS (80°49.90' N, 71°55.10' W, 1260 m asl, operated by the Geological Survey of Canada) as a proxy for the meteorological conditions in the glacierized parts of the watershed. This station is located near the mean equilibrium line of Drambuie Glacier (R. M. Koerner, pers. comm.), ~65 km ESE of Lake Tuborg. In addition, climatic data from the nearest Meteorological Service of Canada (MSC) weather station (Eureka, 80°00.0' N, 85°05.6' W, 10 m asl, before 1963 2 m asl) (Fig. 1) were used to place the observations at Lake Tuborg into a regional context. We used rawinsonde soundings at 0000 h UT (1900 h local time) to approximate daily mean air temperature at the 900 hPa level above Eureka (cf. Hardy, 1996).

Hydrology

We collected suspended sediment samples at the stream gauging cross-section using a US DH-48 depth-integrated suspended sediment hand sampler and hand-dip samples (cf. _strem, 1975) when water depth was insufficient for the sampler. We determined two sampling locations laterally in the channel cross-section prior to sample acquisition, using the equal-discharge-increment method (Edwards and Glysson, 1988). Suspended sediment concentration (SSC) was calculated as the mean of the two sampled verticals. We obtained only one sample when it was impossible to safely wade the stream (discharge > ~3.5 m^3 s^{-1}) by reaching as far as possible into the running water with the sampler. Sampling frequency depended on the visually observed rate of change of SSC and varied from 2 to 12 times/day. The samples were vacuum-filtered in the field through individually numbered and pre-weighed Whatman Type WCN 0.45 μm filters and re-weighed in the lab after drying for 60 min. at 105°C. A total of 549 SSC samples, covering a range from 0 to 5.28 g L^{-1}, were collected and processed for 364 sampling times over 47 days. Our sampling frequency was sufficient to resolve daily cycles of SSC and we used linear interpolation to construct an hourly SSC record. We added subjective turbidity observations in a few cases to facilitate interpolation between SSC measurements during the channel break-up phase.

We measured discharge using a Swoffer Model 2100 current meter, employing the velocity-area method (Herschy, 1985). Current meter measurements became too dangerous when discharge was greater than ~3.5 m^3 s^{-1} due to extremely turbulent flow and high bedload transport, and we were forced to estimate discharge by extrapolation of the rating curve. We measured stage using manual staff gauges during the initial channel break-up phase and installed an automatic stage recording system using two Geokon Inc. model 4580 vibrating-wire pressure transducers once the stream had established a stable channel (June 12). The stream channel cross-section became too unstable to record a meaningful stage record after 1900 h July 17, when the stream began to accommodate changes in discharge not by increasing its cross-sectional area of flow and velocity, but by the addition or subtraction of temporary channels into a dynamic braided channel complex. High bedload transport, severe channel bed and bank erosion, frequent channel migration, and extremely rapid discharge fluctuations made it also impossible to measure discharge directly using a current meter or salt-dilution techniques. Braided stream channel systems are common in proglacial environments and are impossible to gauge at high flows using conventional field methods (cf. Wolfe and English, 1995; Smith et al., 1995, 1996). The hydrological monitoring station was destroyed 0600 h July 19, when the stream obliterated its right channel bank. We used SSC as an indirect proxy for streamflow after 1900 h July 17 because of the aforementioned instability of the channel cross-section. Because the relationship between discharge and SSC is often inconsistent, due to sediment supply limitations, we consider the hydrograph after 1900 h July 17 as only a rough approximation of true runoff, but we feel it is the most conservative estimate possible, given the circumstances. Total daily discharge and suspended sediment load were calculated from their respective hourly records using a +9 hour offset from air temperature (best-match position of the time series determined by cross-correlation analysis).

RESULTS

Summer Climate at Lake Tuborg: Temporal, Elevational, and Spatial Variability

The air temperature records from Camp and Plateau exhibited similar temporal patterns throughout the 1995 monitoring period (Fig. 3; Table 2). Temperatures rose above freezing at both elevations for the first time on May 30. They remained positive at Camp for the remainder of the summer, but dropped below freezing for several multi-day periods at Plateau. Temperature inversions between Camp and Plateau occurred for only three days in early June. Temperature differences between Camp and Plateau were the result of elevation, reflecting a gradient of 6°C/1000 m. Air temperatures at Drambuie Glacier corresponded well with those measured at Lake Tuborg (Fig. 3; Table 2), but remained generally below freezing until mid-July. Snow temperatures at Drambuie Glacier (Fig. 3) rose rapidly from July 13 and reached 0°C two days after mean daily air temperature at the site became positive. The higher temperature gradient (8.2°C/1000 m) between Camp and Drambuie Glacier reflects the different surface conditions in the vicinity of each high-elevation station; air temperatures at Plateau were influenced later in the season by turbulent heat transfer from the surrounding snow-free plateau.

Surface air temperatures at Eureka and Camp also followed the same general temporal pattern (Fig. 3; Table 2). This strong association is remarkable, given that the two sites are more than 200 km apart (Fig. 1). Air temperatures at Camp were consistently above those recorded at Eureka (mean difference = 1.4°C; n = 63 days), probably reflecting Lake Tuborg's more inland location (cf. Lewkowicz and Wolfe, 1994). The high-elevation records from Plateau and Drambuie Glacier also corresponded well with the 900 hPa sounding temperature at Eureka (Fig. 3; Table 2). Temperature inversions at Eureka were limited to two short periods in the latter half of the summer (late July/early August). Summer precipitation was low at Eureka and Camp, both in terms of total magnitude and intensity (Fig. 3; Table 1). However, more frequent and

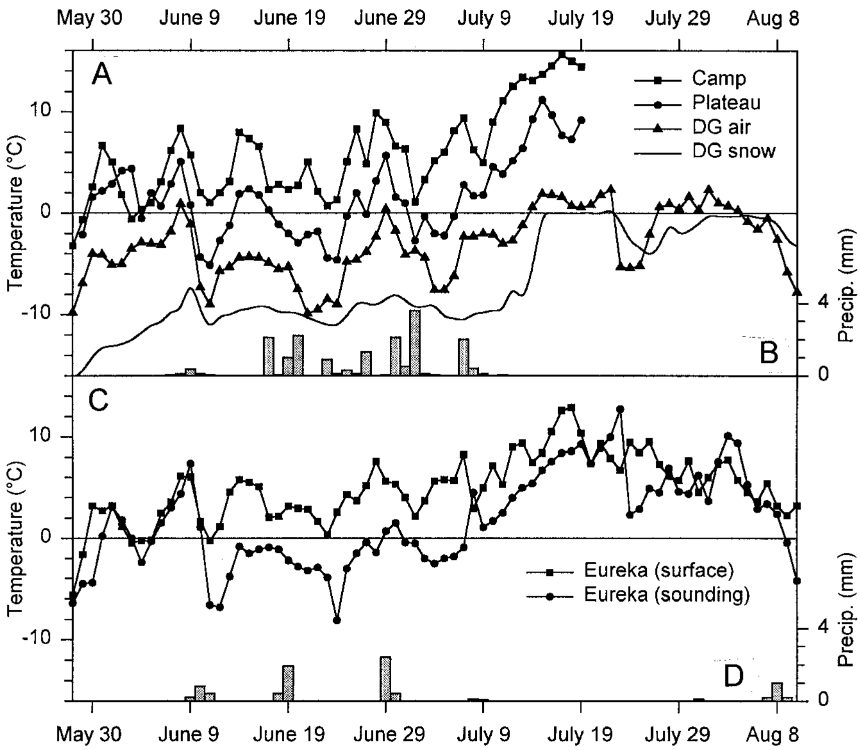

Fig. 3. Summer climate, Lake Tuborg, 1995. A. Daily mean air temperature at Camp, Plateau, Drambuie Glacier (DG air), and Drambuie Glacier snow temperature (DG snow). B. Daily precipitation at Camp. C. Daily mean surface air temperature at Eureka and 900 hPa 0000 h UT Eureka sounding temperature. D. Daily precipitation at Eureka.

TABLE 2

Correlation Coefficient (*R*) between Various Records of Mean Daily Temperature, June 3–July 19, 1995

	Camp	Plateau	Drambuie Glacier	Eureka (surface)
Plateau	0.85	—	—	—
Drambuie Glacier	0.71	0.89	—	—
Eureka (surface)	0.94	0.74	0.66	—
Eureka (900 hPa)	0.76	0.83	0.84	0.72

intense precipitation events were recorded at Lake Tuborg, possibly because of orographic enhancement and additional moisture advection from the east. Atmospheric pressure (not shown) was highly correlated ($r = 0.98$; $n = 63$ days) between Eureka and Camp, indicating that both sites experienced the same weather patterns on a synoptic scale.

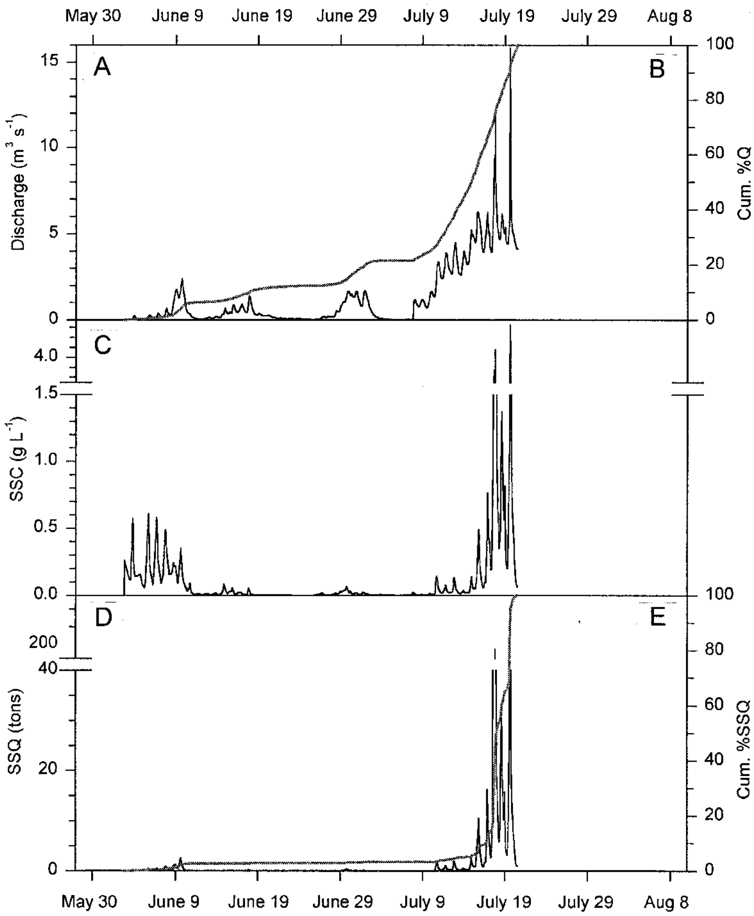

Fig. 4. Streamflow and suspended sediment transport, Lake Tuborg, 1995. A. Hourly discharge (Q). B. Cumulative % Q. C. Hourly suspended sediment concentration (SSC). D. Hourly suspended sediment transport (SSQ). E. Cumulative % SSQ.

Streamflow and Suspended Sediment Transport

Measurable streamflow and suspended sediment transport began 1900 h June 2 and was accompanied by several small, in-channel slushflows (Fig. 4). Discharge and suspended sediment concentration (SSC) varied independently during this initial stream channel break-up period (Figs. 4 and 5) in response to a series of additional upstream slushflows and channel integration processes. Comparatively little sediment was effectively transported during this period, despite relatively high SSC, as a result of low discharge. Nevertheless, suspended sediment transport (SSQ) was higher than on any other day for the next several weeks (Fig. 4).

For the remainder of the measurement period, discharge and SSC co-varied without a time lag and exhibited diurnal cycles, indicating that discharge was governed by the atmospheric energy available for snowmelt and ablation, and that SSC was in turn related to discharge. The scatterplot (Fig. 5) illustrates four distinct clusters,

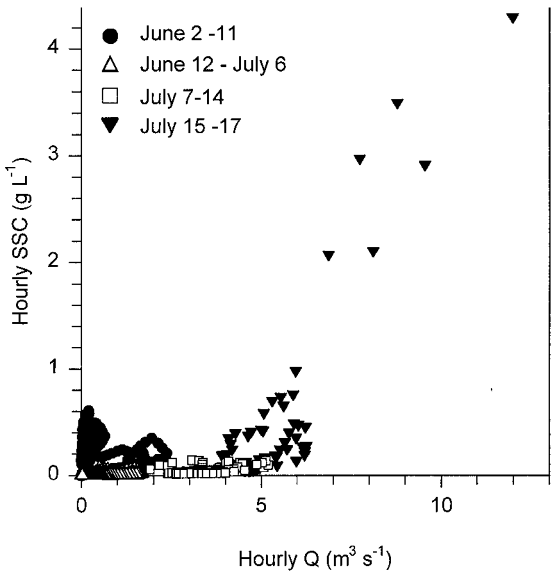

Fig. 5. Relationship between hourly discharge (Q) and hourly suspended sediment concentration (SSC) (1900 h June 2–1900 h July 17).

corresponding approximately to the elevational progression of snowmelt and ablation (see below). It also demonstrates daily and multi-day hysteresis, especially during the first half of the monitoring period, indicative of sediment supply variability. SSQ was minimal due to both low concentrations of suspended sediment and low discharge (Fig. 4) during the middle part of the measurement period (mid-June until early July).

Discharge began to increase continuously on July 7 following the commencement of snowmelt on the lower parts of the Agassiz Ice Cap margin. This increase was punctuated on July 17 and 19 by two high-magnitude drainage events during which discharge more than doubled and SSQ increased by one order of magnitude in less than one hour in each case. Seventy percent of the measured suspended sediment transport, but only 15% of the measured streamflow, were transported in association with these events in less than 27 hours combined (Fig. 4). Both events led to a complete reconfiguration of the braided stream channel system, and the July 19 slushflow also deposited 10–20 cm of snow, slush, and debris on the active floodplain. Over 90% of the measured suspended sediment transport, but less than 50% of the measured streamflow, occurred just during the last five days of measurements (July 15–19) (Fig. 4). Our monitoring efforts ended well before the end of the summer streamflow season when the hydrologic monitoring station was destroyed (0600 h July 19). Summer precipitation events did not lead to significant streamflow responses, and their direct effect on sediment transport was negligible (Figs. 3 and 4).

DISCUSSION

Temporal and Elevational Progression of Snowmelt Runoff

The temporal patterns of air temperature and discharge (Figs. 3 and 4), combined with manual observations of snow cover extent on the plateau and ice cap, document four distinct phases of the temporal and elevational progression of snowmelt and runoff. Snowmelt runoff until mid-June (Phase I) was confined to the immediate valley bottom and other low-elevation areas. Streamflow fluctuations mimicked air temperature variability at Plateau from mid-June until early July (Phase II), suggesting that meltwater originated from the high-elevation slopes and the non-glacierized plateau above 800 m asl. Streamflow almost ceased during two low-temperature intervals in Phase II; although air temperatures remained positive at Camp, the low-elevation meltwater sources had been depleted earlier. Winter snow accumulated on the Agassiz Ice Cap began to contribute to snowmelt runoff in early July and led to a continuous streamflow increase from July 7 (Phase III). Snowmelt had progressed to ~1260 m asl by mid-July, as indicated by the Drambuie Glacier snow temperature record (Fig. 3). We also observed supra-glacial meltwater channels and other indicators of meltwater drainage up to ~1000 m asl on July 20. The snow reservoir in the non-glacierized parts of the watershed was depleted at the same time, and the Agassiz Ice Cap margin turned into the sole source area for meltwater and runoff. The low slope and snow cover of the ice cap margin initially impeded meltwater runoff, which was retained within the ice cap snowpack during a period of rapid melt (July 15–19; Phase IV). The subsequent two-step drainage of this accumulated meltwater led to the two high-magnitude discharge events downstream (Fig. 4).

Relationships between Summer Climate, Streamflow, and Suspended Sediment Transport

The air temperature records from Camp, Plateau, Drambuie Glacier, and Eureka were all associated with measured runoff (Fig. 6). Scatter in the relationship was introduced by: (1) the spatial and elevational variability of snowmelt, accentuated by the watershed hypsometry (Fig. 2); and (2) the variability and development of meltwater routing. Streamflow source areas evolved during the melt season from low-elevation, ice-free to high-elevation, glacierized watershed zones. This vertical progression with time led to a large increase in source area when snowmelt runoff reached an elevation of ~800 m asl (Fig. 2). Meltwater routing imposed variable time lags (1–30 hours based on cross-correlation analysis) between atmospheric energy input to the snowpack, snowmelt (and later ice ablation), and streamflow measured at the basin outlet. Consequently, meteorological data from a single point and elevation cannot fully capture streamflow variability over the entire melt and runoff season. For example, snowmelt runoff on June 8 (Fig. 6) was still confined to the lowest watershed parts, and the warm period from June 5–11 (Fig. 3) merely accelerated ripening of the winter snowpack, while resulting streamflow remained low. In contrast, streamflow source areas observed on July 12 included the non-glacierized plateau as well as the Agassiz Ice Cap margin. As snowmelt progressed to higher elevations, meltwater retention within the glacier snowpack decoupled streamflow from air temperature variability (Figs. 3 and 4). This delay initially dampened the hydrologic response to snowmelt, but subsequently amplified runoff and sediment transport as two high-magnitude drainage events.

Nonetheless, air temperature records from all sites provide a useful proxy for snowmelt and resulting streamflow (Fig. 6), despite these complex internal watershed processes. This suggests that near sea-level observations were representative for the

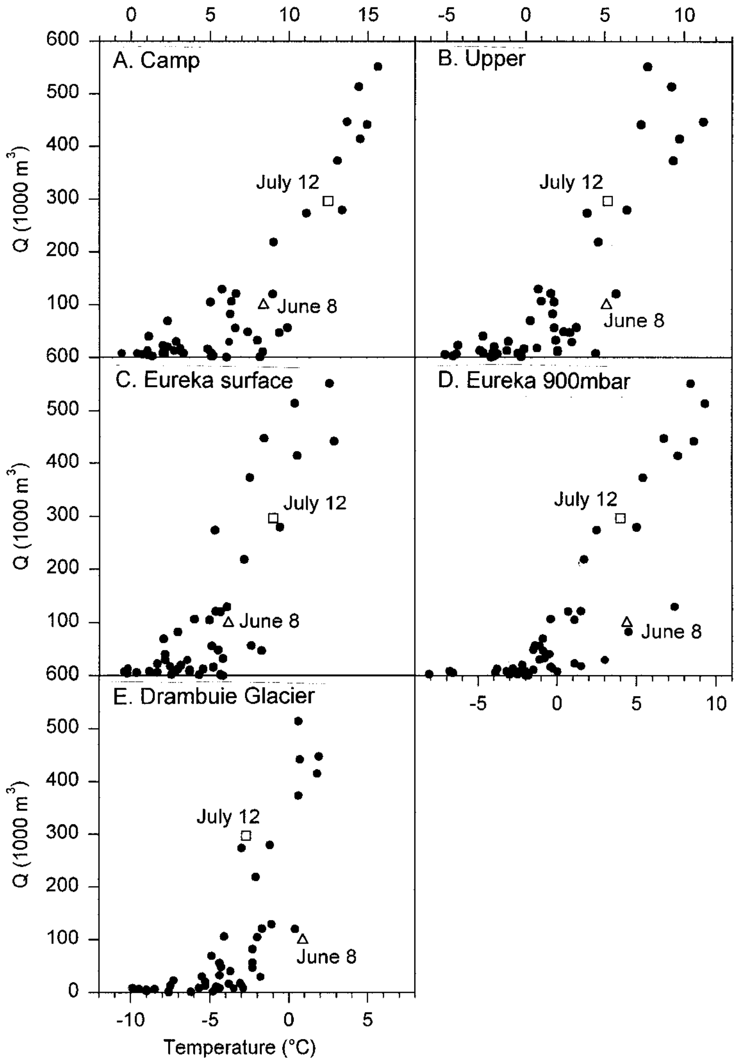

Fig. 6. Relationship between streamflow and summer climate, 1995. Daily runoff (Q) is plotted as a function of daily mean air temperature at (A) Camp, (B) Plateau, (C) Eureka, (E) Drambuie Glacier, and as a function of (D) Eureka 900 hPa 0000h UT sounding temperature. Daily runoff was calculated using a +9 hour offset with respect to air temperature to account for meltwater routing delays. Temperature scale for (A) Camp and (C) Eureka is identical; temperature range is identical for all graphs. June 8 is indicated with an open triangle, July 12 with an open square.

higher elevation zones of the Deception River watershed. It is remarkable how well Eureka surface and sounding data were associated with streamflow, considering the large distance between Lake Tuborg and Eureka (Fig. 1). This indicates that snowmelt and ablation on the Agassiz Ice Cap, and hydrologic input to Lake Tuborg, were less influenced by local meteorological or topographical conditions, but rather were a function of more regional-scale climatic processes.

Suspended sediment transport during our 1995 measurement period was exclusively associated with snowmelt, whereas summer precipitation events had only negligible effects. The Agassiz Ice Cap represents a virtually unlimited meltwater source from snow, firn, and ice. Runoff draining the ice cap margin was sediment-free, indicating the ice cap is merely a source of meltwater, but not sediment. Sediment is eroded by the stream in the extensive proglacial braided channel complex, or comes from the stream valley slopes during snowmelt. A simple degree-day model using Eureka 900 hPa sounding temperature to predict daily runoff and SSQ (June 2 through July 16), applied to the balance of the season, suggests that measured runoff and suspended sediment transport totals were less than 50% of their respective annual totals. The two high-discharge events we observed contributed up to 30% of the annual suspended sediment load and 10% of the annual runoff estimated by the degree-day model. Important caveats to this estimate are that these events are distinctly different hydrological processes (i.e., more effective for sediment transport), and that additional slushflow drainage events may have occurred, from higher elevations on the Agassiz Ice Cap, in late July or in August. These results highlight the importance of slushflow-generated runoff from the Agassiz Ice Cap, in terms of annual sediment transport to the lake. High-magnitude drainage events and slushflows are common on High Arctic glaciers and ice caps (cf. Adams, 1966; Woo, 1993), although their frequency in association with snowmelt runoff initiation along the Agassiz Ice Cap margin remains uncertain.

Summer precipitation events through July 1995 were of insufficient magnitude and intensity to cause any substantial response of streamflow and sediment transport (Figs. 3 and 4). Most events fell as a mix of snow and rain with a transient snowline below ~500 m asl. Hence, precipitation in large parts of the watershed (Fig. 2) fell as snow and was ineffective in generating direct rainfall runoff. High-intensity summer rainfall events in the High Arctic, however, can lead to episodes of high streamflow and sediment transport (e.g., Cogley and McCann, 1976), particularly in glacierized watersheds (Hardy, 1995). In fact, the watersheds described by Cogley and McCann (1976) are very similar to that of the Deception River, in terms of their geographical and geological characteristics, hypsometry, and percentage glacierization.

Hydrological Field Measurements in the High Arctic

This study illustrates the difficulties associated with hydrologic monitoring in remote and unknown environments such as the High Arctic. The Deception River fan-delta was 100% snow covered when we arrived at the field site on May 14, 1995. Snow depth was variable, but particularly deep snowdrifts had accumulated inside the stream channels. This situation made it impossible to assess stream channel location and configuration prior to selecting a suitable site for the camp. This *a priori* limited our options for selecting a suitable monitoring cross-section, as the site needed to be close to camp for logistic and safety reasons.

In braided stream channel systems, one of the fundamental assumptions governing the standard rating-curve approach for discharge measurements (stability of the channel cross-section) becomes invalid as soon as discharge increases enough to activate channel braiding. Furthermore, the rated cross-section often becomes unstable when the channel bed and banks thaw and become subject to fluvial erosion. Conventional field methods such as current meters, stage recorders, or dilution-gauging techniques are inappropriate to measure discharge in a braided channel system. Logistic (and financial) constraints in remote environments such as the High Arctic often preclude the use of more sophisticated techniques or permanent structures. We concur with Lawson (1993), who suggested that new remote sensing techniques (e.g., Smith et al., 1995, 1996) are needed to augment or replace field discharge measurements.

SUMMARY AND CONCLUSIONS

We conducted hydrological and meteorological observations at Lake Tuborg, Ellesmere Island, Canada to develop an understanding of contemporary water and sediment transport processes. Weather observations at Eureka and Lake Tuborg indicate that both sites experienced the same weather on a synoptic scale. Streamflow and suspended sediment transport were exclusively a function of snowmelt runoff, whereas the direct influence of summer precipitation events was negligible in 1995. Snowmelt was primarily controlled by synoptic-scale climatic processes. Two brief, high-magnitude meltwater pulses contributed a significant portion of the annual suspended sediment load to Lake Tuborg. Snowmelt initiation along the Agassiz Ice Cap margin may be accompanied by high-magnitude drainage and slushflow events of annual or sub-annual frequency. Additional years of data collection are needed to define the annual and interannual variability of the sediment delivery system, particularly with respect to the relative importance of summer rainfall events.

Under climatic warming conditions, larger portions of the Agassiz Ice Cap will become potential source areas for meltwater and runoff, and it is likely that water and sediment flux to Lake Tuborg will increase under such a scenario. Likewise, increases in winter snowfall and summer rainfall associated with a warmer climate will lead to increased water and sediment transport to the lake.

LITERATURE

AES. *Eureka, Principal Station Data, Volume 79*. Downsview, Ontario: Atmospheric Environment Services, Environment Canada, 1984.

Adams, W. P. *Ablation and Run-off on the White Glacier, Axel Heiberg Island, Canadian Arctic Archipelago*. Montreal, Quebec: McGill University. Axel Heiberg Island Research Reports, Glaciology, No. 1, 1966.

Atkinson, D. E., B. Alt, and K. Gajewski. “A new database of High Arctic climate data from the Polar Continental Shelf Archives,” *Bulletin of the American Meteorological Society*, Vol. 81, No. 11, 2000, pp. 2621-2629.

Braun, C., D. R. Hardy, R. S. Bradley, and M. J. Retelle. “Streamflow and suspended sediment transport into Lake Sophia, Cornwallis Island, Nunavut, Canada,” *Arctic, Antarctic, and Alpine Research*, Vol. 32, No. 4, 2000, pp. 456-465.

Cogley, J. G. and S. B. McCann. “An exceptional storm and its effects in the Canadian High Arctic,” *Arctic, Antarctic, and Alpine Research*, Vol. 8, 1976, pp. 105-110.

Courtin, G. M. and C. L. Labine. “Microclimatological studies in Truelove Lowland,” in L.C. Bliss, ed., *Truelove Lowland, Devon Island, Canada: A High Arctic Ecosystem.* Edmonton, Alberta: University of Alberta Press, 1977.

Deslodges, J. R. "Varve deposition and the sediment yield record at three small lakes of the southern Canadian Cordillera," *Arctic, Antarctic, and Alpine Research,* Vol. 26, 1994, pp. 130-140.

Edlund, S. A. and B. T. Alt. "Regional congruence of vegetation and summer climate patterns in the Queen Elizabeth Islands, Northwest Territories, Canada," *Arctic,* Vol. 42, 1989, pp. 3-23.

Edwards, T. K. and G. D. Glysson. *Field Methods for Measurement of Fluvial Sediments*. Washington, DC: U.S. Geological Survey Open-File Report 86-531, 1988.

Francus, P., R. S. Bradley, and M. Abbott. "Annually resolved past climate signal from lacustrine varved sediments, Arctic Canada," *EOS (Transactions of the American Geophysical Union),* Vol. 80, Fall Supplement, 1999.

Hardy, D. R. Streamflow and Sediment Transfer from a Mountainous High Arctic Watershed, Northern Ellesmere Island, Canada. Unpublished Ph.D. thesis, University of Massachusetts, Amherst, MA, 1995, 267 pp.

Hardy, D. R. AClimatic influences on streamflow and sediment flux into Lake C2, northern Ellesmere Island, Canada," *Journal of Paleolimnology,* Vol. 16, 1996, pp. 133-149.

Hardy, D. R., R. S. Bradley, and B. Zolitschka. "The climatic signal in varved sediments from Lake C-2, northern Ellesmere," *Journal of Paleolimnology,* Vol. 16, 1996, pp. 227-238.

Herschy, R. W. *Streamflow Measurements.* New York, NY: Elsevier, 1985.

Hodgson, D. A. "Introduction (Quaternary Geology of the Queen Elizabeth Islands), Chapter 6," in R. J. Fulton, ed., *Quaternary Geology of Canada and Greenland.* Ottawa, Ontario: Geologic Survey of Canada, Geology of Canada, No. 1, 1989, pp. 443-448 (also *The Geology of North America.* Boulder, CO: Geological Society of America,Vol. K-1)

Hughen, K., J. T. Overpeck, and R. F. Anderson. "Recent warming in a 500-year paleotemperature record from varved sediments, Upper Soper Lake, Baffin Island," *The Holocene,* Vol. 10, 2000, pp. 9-19.

Lawson, D. E. *Glaciohydrologic and Glaciohydraulic Effects on Runoff and Sediment Yield in Glacierized Basins.* Hanover, NH: U.S. Army Cold Regions Research and Engineering Laboratory, Monograph 93-2, 1993.

Leonard, E. M. "Glaciological and climatic controls on lake sedimentation, Canadian Rocky Mountains," *Zeitschrift fuer Gletscherkunde und Glazialgeologie,* Vol. 21, 1985, pp. 35-42.

Lewkowicz, A. G. and P. M. Wolfe. "Sediment transport in Hot Weather Creek, Ellesmere Island, N.W.T., Canada, 1990-1991," *Arctic, Antarctic, and Alpine Research,* Vol. 26, No. 3, 1994, pp. 213-226.

Long, A. "Age of trapped sea-water at the bottom of Lake Tuborg, Ellesmere Island, N.W.T.," *Transactions American Geophysical Union,* Vol. 48, 1967, p. 136.

Østrem, G. "Sediment transport in glacial meltwater streams," in A.V. Jopling and B. C. McDonald, eds., *Glaciofluvial and glaciolacustrine sedimentation*. Los Angeles, CA: Society for Economic Paleontology and Mineralogy, Special Publication 23, 1975, pp. 101-122.

Perkins, J. A. and J. D. Sims. "Correlations of Alaskan varve thickness with climatic parameters, and use in paleoclimatic reconstructions," *Quaternary Research,* Vol. 20, 1983, pp. 308-321.

Smith, L. C., B. L. Isacks, R. R. Forster, and A. L. Bloom. "Estimation of discharge from three braided rivers using synthetic aperture radar satellite imagery: Potential application to ungaged basins," *Water Resources Research*, Vol. 32, No. 7, 1996, pp. 2021-2034.

Smith, L. C., B. L. Isacks, R. R. Forster, A. L. Bloom, and I. Preuss. "Estimation of discharge from braided glacial rivers using ERS 1 synthetic aperture radar: First results," *Water Resources Research*, Vol. 31, No. 5, 1995, pp. 1325-1329.

Smith, S. V. A Record of Environmental Change Derived from Varved Lakes along the Margin of the Agassiz Ice Cap. Unpublished M.Sc. thesis, University of Massachusetts, Amherst, MA, 1997.

Stihler, S. D., D. B. Stone, and J. E. Beget. "Varve counting vs. tephrachronology and ^{137}Cs and ^{210}Pb dating: A comparative test at Skilak Lake Alaska," *Geology*, Vol. 20, 1992, pp. 1019-1022.

Trettin, H. P. "The Arctic Islands," in A. W. Bally and A. R. Palmer, eds., *The Geology of North America: An Overview*. Boulder, CO: Geological Society of America, Vol. A, 1989, pp. 349-370.

Wolfe, P. M. and M. C. English. "Hydrometeorological relationships in a glacierized catchment in the Canadian High Arctic," *Hydrological Processes*, Vol. 9, 1995, pp. 911-921.

Woo, M.-K. "Northern hydrology," in H. M. French and O. Slaymaker, eds., *Canada's Cold Environments*. Montreal, Quebec and Kingston, Ontario: McGill-Queen's University Press, 1993.