Sedimentation in arctic proglacial lakes: Mittivakkat Glacier, south-east Greenland

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Abstract:
Several sediment cores were collected from two proglacial lakes in the vicinity of Mittivakkat Glacier, south-east Greenland, in order to determine sedimentation rates, estimate sediment yields and identify the dominant sources of the lacustrine sediment. The presence of varves in the ice-dammed Icefall Lake enabled sedimentation rates to be estimated using a combination of X-ray photography and down-core variations in $^{137}$Cs activity. Sedimentation rates for individual cores ranged between 0.52 and 1.06 g cm$^{-2}$ year$^{-1}$, and the average sedimentation rate was estimated to be 0.79 g cm$^{-2}$ year$^{-1}$. Despite considerable down-core variability in annual sedimentation rates, there is no significant trend over the period 1970 to 1994. After correcting for autochthonous organic matter content and trap efficiency, the mean fine-grained minerogenic sediment yield from the 3.8 km$^2$ basin contributing to the lake was estimated to be 327 t km$^{-2}$ year$^{-1}$. Cores were also collected from the topset beds of two small deltas in Icefall Lake. The deposition of coarse-grained sediment on the delta surface was estimated to total in excess of 15 cm over the last c. 40 years. In the larger Lake Kuutuaq, which is located about 5 km from the glacier front and for which the glacier represents a smaller proportion of the contributing catchment, sedimentation rates determined for six cores collected from the centre of the lake, based on their $^{137}$Cs depth profiles, were estimated to range between 0.05 and 0.11 g cm$^{-2}$ year$^{-1}$, and the average was 0.08 g cm$^{-2}$ year$^{-1}$. The longer-term (c. 100–150 years) average sedimentation rate for one of the cores, estimated from its unsupported $^{210}$Pb profile, was 0.10–0.13 g cm$^{-2}$ year$^{-1}$, suggesting that sedimentation rates in this lake have been essentially constant over the last c. 100–150 years. The average fine-grained sediment yield from the 32.4 km$^2$ catchment contributing to the lake was estimated to be 13 t km$^{-2}$ year$^{-1}$. The $^{137}$Cs depth profiles for cores collected from the topset beds of the delta of Lake Kuutuaq indicate that in excess of 27 cm of coarse-grained sediment had accumulated on the delta surface over the last approximately 40 years. Caesium-137 concentrations associated with the most recently deposited (uppermost) fine-grained sediment in both Icefall Lake and Lake Kuutuaq were similar to those measured in fine-grained sediment collected from steep slopes in the immediate proglacial zone, suggesting that this material, rather than contemporary glacial debris, is the most likely source of the sediment deposited in the lakes. This finding is confirmed by the $^{137}$Cs concentrations associated with suspended sediment collected from the Mittivakkat stream, which are very similar to those for proglacial material. Copyright © 2000 John Wiley & Sons, Ltd.

KEY WORDS lake sedimentation; sediment yields; sediment sources; varves; caesium-137; lead-210; Arctic; proglacial; Greenland

INTRODUCTION
In recent years there has been a growing interest in the nature and rates of geomorphological processes operating in arctic and subarctic environments. This interest has been promoted by an increasing awareness

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Contract/grant sponsor: Danish Natural Science Research Council.

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Received 9 December 1998
Accepted 13 July 1999
that such areas are particularly sensitive to climate change (e.g. Koster, 1993; Slaymaker and French, 1993), and the need for information on contemporary processes that can be used for predicting the impact of future changes in climate (e.g. Berger and Iams, 1996; Berger, 1997). In addition, it has been recognized that measurements of suspended sediment transport by proglacial streams can be used to derive information on glacier processes (cf. Hodson et al., 1998) and rates of landscape development in glacialized arctic and subarctic areas (cf. Lawler, 1991). Against this background, information on sediment delivery and transport in Greenland must be seen as very limited. Although Greenland is believed to be a major contributor of sediment to the North Atlantic Ocean, very few studies have been undertaken on the magnitude and temporal variability of sediment inputs to the oceans from glacial and proglacial areas. Furthermore, because of the potential for the exploitation of hydro-electric power, there is also a practical need for information on the sediment loads of rivers in the region.

The lack of empirical data for Greenland reflects the harsh arctic climate, problems associated with accessibility and the high cost of fieldwork. Existing data on sediment delivery and transport relate mainly to short time periods within single years and commonly involve only the late summer period (e.g. Hasholt, 1976, 1992, 1994, 1996). On this basis, suspended sediment yields from glacialized areas in Greenland have been estimated to range between 84 and 1500 t km$^{-2}$ year$^{-1}$, compared with 1 to 56 t km$^{-2}$ year$^{-1}$ from non-glacialized areas. These values are likely to represent minimum values, however, owing to the short measurement periods (cf. Hasholt, 1996). Although the above values provide a general indication of the output of sediment from glacialized and non-glacialized basins, the extent to which they are representative of annual values remains uncertain and little is known about interannual variability in sediment yields and longer term trends. The sedimentary record preserved in the bottom of proglacial lakes provides some scope to remedy this deficiency, although changes in drainage area during glacial advance or deglaciation can complicate the interpretation of longer term trends. Thus, for example, Østrem and Olsen (1987) were able to use sediment cores to estimate the sediment output from the Bondhus Glacier, Norway, during the period 1692 to 1974. In addition to providing information on sediment yields, lakes also represent important intermediate sinks for sediment storage in the proglacial zone and there is a need for more information on sedimentation rates in proglacial lakes in order to understand the delivery of sediment from proglacial environments.

This paper describes an investigation of two proglacial lakes in south-east Greenland which aimed to:

1. use varve couplets and caesium-137 ($^{137}$Cs) measurements (one core was also analysed for unsupported lead-210) on multiple sediment cores to establish core chronologies and determine lacustrine sedimentation rates, and thereby estimate the mass of sediment stored in the lakes and the sediment yields from their contributing catchments;
2. use $^{137}$Cs measurements to identify the main sources of sediment transported through the proglacial zone and deposited in the lakes, and in particular to determine the relative importance of the glacier and the proglacial zone as source areas.

The two study lakes were selected because of their differing contributing catchments, which thus provide an opportunity to examine lake sedimentation, sediment yields and sediment sources under contrasting proglacial conditions. Icefall Lake is located in the immediate ice-marginal zone and the contributing catchment is dominated by the Mittivakkat Glacier and recently deglaciated slopes. Lake Kuutuaq is located approximately 5 km downstream from the glacier front and the glacier represents a much smaller portion of the contributing catchment. Furthermore, in the case of Lake Kuutuaq, there is considerable potential for conveyance losses during sediment transport to the lake.

This study constitutes part of a larger project, aimed at understanding arctic landscape evolution and climate interaction, carried out by the Institute of Geography, University of Copenhagen. The ultimate aim of the larger project is to understand and model geomorphological processes operating in arctic environments, so that the implications of different global climatic change scenarios can be evaluated.
THE STUDY AREA

The study area is located on Ammassalik Island (65°40′N, 37°55′W) in south-east Greenland (Figure 1a) and is delimited by the eastern divide of the (warm-based) Mittivakkat Glacier (surface area 30 km²), a recently deglaciated valley draining the southern flank of the glacier, and an earlier deglaciated valley system draining the northern flank of the glacier, all of which drain into Sermilik Fjord (Figure 1b). The southern outlet drains approximately 7.6 km² of the glacier and feeds the Mittivakkat stream, which flows in a well-developed braided channel across an outwash plain stretching c. 1.5 km from the glacier terminus to the sea, where a substantial delta has formed. The northern outlet drains 8.8 km² of the glacier and the main part of the northern flank drains or calves into the proglacial Icefall Lake (the name refers to a c. 200 m long and 30 to 50 m high glacier front which calves into the lake), which is situated at the western terminus of the glacier. Outflow from Icefall Lake is at the north-east end of the lake (see Figure 1b) and passes through a series of shallow lakes, before turning north-north-west and joining a stream that drains a smaller glacier originating at the top of the Mittivakkat summit (this glacier is separated from the main glacier by nunataks and steep cliffs). This stream then flows into a broad, flat valley and drains into Lake Kuutuaq through a well-developed delta, formed at the eastern end of the lake. The outflow from Lake Kuutuaq finally enters Sermilik Fjord through a gorge-like fissure valley. Fine-grained suspended sediment is the dominant size fraction transported in the proglacial streams, and it has been estimated that suspended sediment contributes c. 72% of the total load transported by the Mittivakkat stream, with bed and dissolved load contributing 22% and 6%, respectively (Hasholt, 1976, 1996).

Geologically, the area is part of the Nagssugtoqides belt, with predominantly gneissic rocks and some basalt inclusions. The landscape is dissected by deep glacial valleys and smaller, geologically-controlled, fissure valleys. The proximity of the Greenland Ice Cap exerts a dominant control on the climate of the locality and results in generally stable cold weather. The nearby town of Ammassalik has an annual mean temperature of −1.2 °C and monthly mean temperatures range from a minimum of −8.2 °C in February to a maximum of 7.5 °C in July. Mean annual precipitation at Ammassalik is 826 mm, but measurements of runoff and precipitation within the study area indicate higher precipitation amounts of the order of 2000 mm year⁻¹. Monthly precipitation totals typically range between 40 and 90 mm, although there is substantial interannual variability (e.g. monthly precipitation for October ranges between 35 and 470 mm). The maximum monthly precipitation occurs in October or November and the minimum in June or July. The area is characterized by a glacio-nival hydrological regime. Melting normally starts in May and reaches a maximum in June or July, although rainstorms can cause short-lived peaks. The runoff from October until the end of April is normally very low, reflecting the cold temperature and the small amount of basal runoff from the glacier, and can cause icing to develop. During winter, foehn winds and low pressure systems tracking along the east coast can generate runoff from the snow surface.

The study lakes

Following an initial reconnaissance survey of lakes in the study area, Icefall and Kuutuaq lakes were selected for the collection of sediment cores. Icefall Lake (Figures 1c and 2a) is located c. 147 m a.m.s.l. and consists of two basins, with maximum water depths of 28 and 35 m for the northern and southern basins, respectively. These basins are separated by a sill with a water depth of c. 10 m. The lake is not illustrated on the 1933 Geodaetic Institute 65 01E map, but it is clearly visible on areal photographs taken in 1972, and was probably formed when the glacier retreated from its Little Ice Age maximum, which was reached around 1895 (O. Humlum, personal communication, 1988). The ice margin in 1994 had retreated c. 100–200 m east and south of the 1972 position. Although Icefall Lake is bounded by the glacier margin to the east, this does not represent a glacier snout in the true sense, and at present most of the catchment bordering the lake is non-glacierized and consists of relatively steep slopes of exposed scree material. Small streams enter the western end of each of the two lake basins, draining non-glacierized catchments of approximately 0.7 km², creating small deltas.
Figure 1. (A) Location of the study area in Greenland. (B) Location map of the study area and lakes: A is Lake Kuutuaq, B is Icefall Lake and C is the location of the suspended sediment sampling site on the Mittivakkat Stream. The contour interval is 50 m and is based on Geodaetic Institute map 65 01E, and the location of Icefall Lake and the streams is based on aerial photographs from 1981. (C) Location of the cores collected from Icefall Lake and the positions of the glacier ice margin in 1972 and 1994. The contour interval is 5 m. (D) Location of the cores collected from Lake Kuutuaq. The delta is denoted by stippling. The dashed line on the delta (to the east of the three cores) marks the approximate boundary between the subaerial (to the east of the line) and submerged (to the west) sections of the delta.
Figure 2. Photographs of: (a) Icefall Lake (also showing the equipment used to collect sediment cores), looking north-east from near core IF7 towards Mittivakkat Glacier (top right); and (b) Lake Kuutuaq looking west towards its outlet
Lake Kuutuaq (Figures 1d and 2b) is about 2 km long and 0.5 km wide, and again consists of two basins, separated by a 30 m deep sill at the narrowest part of the lake. Maximum water depths are 70 and 80 m for the eastern and western basins, respectively. The main stream enters the eastern end of the lake, creating a large delta with steep (c. 0.3 m m$^{-1}$) foreset beds, and becomes braided as it passes through this delta zone.

The residence times of water entering Icefall and Kuutuaq lakes have been estimated to be c. 5–15 and 20–60 days, respectively. Because the distance from the main sediment input to the lake outlet is about 1 km for Icefall Lake and 2 km for Lake Kuutuaq, it is likely that all but the finest clay-sized particles are deposited on the lake beds. Following the procedure documented by Brune (1953), the trap efficiencies of both lakes have been estimated to be 100% for coarse sediment and >95% for fine sediment. Because most of the inflow occurs during the ablation season, the estimates of trap efficiency cited above are likely to be overestimates, as they are based on the ratio of annual capacity and annual inflow. A value within the range 90 ± 10% is believed to be representative of fine-grained sediment in this situation and has, therefore, been assumed for both lakes. It should, however, be recognized that the estimates of sediment mass and sediment yield reported below are sensitive to the value of trap efficiency used.

During the winter period from October until May, the surfaces of both lakes are covered by ice and consequently runoff and sediment supply to the lakes is limited and primarily of subglacial origin. During the summer period, the water column is well mixed and observations have revealed relatively homogeneous concentrations of fine-grained sediment in both lakes. Sediment washed on to the lake ice from the surrounding proglacial area early in the snowmelt period can be transported on ice floes, and may be subsequently deposited on to the lake bed or even pass through the outlet. In addition, at Icefall Lake, the glacier calves into the lake, causing both localized sediment resuspension and deposition (see Figure 9a). In July 1994 several incidents of calving took place, which introduced large blocks of glacier ice into the lake. Coarse englacial and supraglacial material contained in these ice blocks were deposited locally on to the lake bed when they melted (i.e. dropstones) (cf. McManus and Duck, 1988; Schmok and Clarke, 1989). Because Lake Kuutuaq is not ice-dammed, coarse-grained material can be deposited on the lake bed beyond the delta only as a result of melting ice floes, although owing to the steep nature of the foreset beds slumping may also redistribute coarse deltaic sediment to deeper waters just beyond the delta (cf. Gilbert, 1972).

METHODS

Lake coring

Six sediment cores were collected from the bottom of Lake Kuutuaq in 1992 using a 20 kg gravity corer equipped with a 1 m long, 8-04 cm$^2$ surface area, plexiglass tube, and deployed from a platform constructed of oil drums. Six cores were collected from Icefall Lake in 1994 using a modified Nesje corer (Nesje, 1992) (a wire operated piston corer with a surface area of 28-3 cm$^2$). Owing to the anticipated occurrence of varves in Icefall Lake, an Axelsson corer (Axelsson and Håkanson, 1972) was also used to collect cores for X-ray photography (cf. Axelsson and Håndel, 1972). The Axelsson gravity corer uses rectangular core tubes of dimensions 5-7 cm by 2-9 cm (surface area 16-5 cm$^2$) and 1 m long. As access to Icefall Lake was difficult, the coring platform comprised a lightweight raft consisting of four inflatable plastic boats linked together by a frame of lightweight ladders and equipped with a winch mounted on a ladder (Figure 2a). During coring the raft was held in position by ropes stretched across the lake. For both lakes, coring positions were located using a theodolite equipped with an electro-optic distance meter (Kern D502).

Shallow sediment cores were also collected from the submerged portions of the deltas of both lakes using hand operated corers. Three cores (surface area 21-2 cm$^2$) were collected from the submerged portion of the delta at the eastern end of Lake Kuutuaq, and in the case of Icefall Lake a single core (surface area 34 cm$^2$) was collected from each of the two small deltas located where the small streams drain into the western end of each basin (see Figure 1).

The cores collected from Lake Kuutuaq and those collected from Icefall Lake with the Nesje corer were extruded in the field and cut into 1- or 2-cm slices. After siphoning off excess water, the cores collected from
Icefall Lake with the Axelsson corer were transported undisturbed to Ammassalik Hospital, where X-ray photography of the unextruded cores was undertaken (the optimum settings were KV 75 and MAS 12). The cores were subsequently extruded and sectioned into 1-cm slices. The delta cores from both lakes were also extruded and sectioned in the field, although because of the coarse nature of the sediment in these cores, the thickness of each increment varied. All deepwater and delta core slices were transported to Exeter University for further laboratory analysis.

Soil and sediment sampling

Soil and sediment samples were collected from the study area between 1990 and 1998 in order to measure fallout $^{137}$Cs activities. This information was required primarily for identifying the sources of the sediment deposited in the two study lakes and also to assist with the interpretation of the $^{137}$Cs inventories of the lake cores (see Ritchie and McHenry (1990) and Walling and Bradley (1990) for useful reviews on the application of $^{137}$Cs measurements in sediment investigations). Surface samples (>100 g) were collected from potential source areas of fine-grained sediment, such as eroding topsoil, exposed colluvial deposits, sediment deposits around the margins of the lakes, and supraglacial, englacial and subglacial deposits from the Mittivakkat glacier.

In order to determine the total input of $^{137}$Cs from atmospheric fallout to the study area, samples were collected from two different areas of flat, undisturbed grassland, where there was no evidence of significant soil redistribution (subsequently referred to as reference sites). Site 1 was located near to the coast at sea level, and site 2 (which was used mainly as a check on site 1) was located near to, and at the same altitude as, Icefall Lake. At both sites, information on the depth distribution of $^{137}$Cs, which was required to assist with the interpretation of the $^{137}$Cs fallout data, was obtained by collecting soil at 1 or 2 cm increments to depths of >20 cm using either the scraper-plate technique (surface area 450 or 640 cm$^2$) (Campbell et al., 1988) or by combining 1 cm increments from four sectioned cores (surface area 34 cm$^2$) collected within a small area. The reference inventory for site 1 was obtained by calculating the mean total $^{137}$Cs inventory for two scraper-plate samples, two composite sectioned samples (each of four cores) and five additional bulk cores. The reference inventory at site 2 was determined by calculating the total $^{137}$Cs inventory for composite samples derived from four sectioned cores.

Radiometric and laboratory analysis

Samples for radiometric analysis were transported to the Department of Geography at the University of Exeter, where they were air-dried (soils and source materials) or freeze-dried (lake and suspended sediment), disaggregated and sieved to separate the <2 mm fraction prior to analysis. Caesium-137 activities were determined by $\gamma$-ray spectrometry at 661.1 keV, using high purity coaxial germanium detectors. Typical count times were >30 000 s, providing an analytical precision (±2 standard deviations) of between c. ±5 and ±15%. One of the cores from Lake Kuutuaq was also assayed for unsupported lead-210 ($^{210}$Pb) activity by direct gamma spectrometry (cf. Joshi, 1987). All radionuclide data have been standardized to refer to activities in 1998.

The organic matter contents of the sediment samples were calculated from measurements of their organic carbon content, undertaken using a Carlo Erba C/N Analyser. The ultimate (dispersed) particle size composition of the mineral fraction was determined by a combination of sieving (for the >800 $\mu$m fraction) and by using a Coulter LS130 laser granulometer (for the <800 $\mu$m fraction), after standard chemical and ultrasonic preparation.

RESULTS

Caesium-137 inventories and depth distributions at reference sites

No significant input of Chernobyl fallout was detected in the study area (Hasholt and Walling, 1992) and the measurements reported below were, therefore, assumed to relate only to bomb-derived $^{137}$Cs. This
assumption is consistent with the estimated deposition of Chernobyl-derived 137Cs on Ammassalik Island of 36 Bq m$^{-2}$ (Aarkrog, 1994), which would represent only about 2% of the total inventory. Figure 3 presents representative 137Cs depth distributions for each of the two reference sites. Both evidence very high 137Cs concentrations at the surface, with a rapid decrease in concentration with depth, and >95% of the 137Cs is contained in the top 5 cm. The difference in the depth to which 137Cs extends in the two profiles probably reflects differences in soil properties (bulk density, organic matter content, etc.) between the two locations. The extent of downward translocation of 137Cs in these soils is less than that normally encountered in soils in temperate environments (cf. Walling and Quine, 1992; Owens et al., 1996) and this is thought to reflect the reduced biological activity and bioturbation associated with arctic soils. The total inventories of these two profiles are 2126 and 2281 Bq m$^{-2}$. As there was no significant difference in the 137Cs inventories of the samples from the two reference sites, the data were pooled. The 137Cs inventories for some samples at site 1 were significantly higher (>2 standard deviations) than the others (possibly due to drifting snow or concentrated surface runoff and ponding) and were subsequently removed from the data set. The average 137Cs inventory at the two reference sites was 1965 ± 108 Bq m$^{-2}$ (mean ± 2 standard errors of the mean; $\bar{x} \pm 2 \sigma_x$) and the coefficient of variation was 11%.

Sedimentation rates in Icefall Lake

Caesium-137 profiles and inventories. Coring in Icefall Lake proved to be extremely problematic. Despite the use of a core catcher, core recovery was often as low as one in 15 attempts and only six cores (IF1 to IF6) were retrieved successfully from the main lake basins (four from the southern basin and two from the northern basin), and two cores (IF7 and IF8) were also collected from the topset beds of the deltas. Sediment cores were not collected from the area immediately adjacent to the glacier (see Figure 1c), because the sediment deposited here is likely to have been disturbed by calving ice (see Figure 9a), thereby complicating the 137Cs depth profiles and varve layers. Because of this, the core sites were necessarily biased towards the western end of the lake (see Figure 1c). Thus, it is possible that the sediment associated with some of the cores may be biased towards localized inputs from the streams draining non-glacierized subcatchments, and this should be borne in mind when interpreting the results presented below.

The lake sediments in the two main basins generally consisted of distinct laminations of either light-brown or light-grey fine silts and clays, which are characteristic of slackwater deposits in an ice-contact lake, although some of the cores also contained layers of fine sand and occasional coarser sand- and gravel-sized material, which are possibly dropstone deposits derived from melting icebergs. The average organic matter content of the uppermost sediment was 0.91 ± 0.12% ($\bar{x} \pm 2 \sigma_x$) (Table I). The sediments deposited on the delta topset beds generally consisted of massive sands with some gravel-sized material, and the average organic matter content of the uppermost sediment was 0.49 ± 0.14%.
All eight cores were analysed for $^{137}$Cs activity and cores IF2, IF3 and IF6 were also analysed for varves using X-radiography. Figure 4 shows the $^{137}$Cs depth profiles for all eight cores. Cores IF1, IF2 and IF3 were collected from approximately the same location (within $\sim15$ m) in the southern basin. Core IF1 is the longest core at $\sim40$ cm and although there is still significant $^{137}$Cs in the bottom slice, the shape of the $^{137}$Cs profile suggests that this core probably contains most of the $^{137}$Cs profile for this site. The $^{137}$Cs inventory for core IF1 is 14,705 Bq m$^{-2}$ (the highest of all eight cores) and there is a distinct peak in $^{137}$Cs concentration of $\sim100$ mBq g$^{-1}$ at $\sim26$ cm depth, and this peak can be equated with the 1963 $^{137}$Cs fallout maximum (cf. Ritchie and McHenry, 1990). Both the high total $^{137}$Cs inventory ($14,705$ Bq m$^{-2}$ compared with the reference inventory of $1965$ Bq m$^{-2}$) and the depth of the 1963 $^{137}$Cs peak ($\sim26$ cm) suggest that there has been a considerable amount of sedimentation at this sample location. Based on the depth of the 1963 $^{137}$Cs peak in this core, the average sedimentation rate since 1963 is estimated to be 0.84 cm year$^{-1}$ ($1.06$ g cm$^{-2}$ year$^{-1}$ accumulated mass). Cores IF2 and IF3 are much shorter at $\sim7$ and 17 cm, respectively, and are unlikely to contain the full $^{137}$Cs depth profile. Although these two cores are not suitable for estimating sedimentation rates, their $^{137}$Cs profiles and inventories are similar to those for the equivalent sections in core IF1. Because the three cores were collected from within $\sim15$ m of each other, it is suggested that they can be directly compared. Core IF4 was also collected from the southern basin, and the $^{137}$Cs profile for this core compares well with those for the cores described above, although the bottom sediment increment in core IF4 has a relatively high $^{137}$Cs concentration, suggesting that the full $^{137}$Cs profile has not been sampled. However, if it is assumed that the $^{137}$Cs concentration peak at 13 cm depth can be dated to 1963, then the sedimentation rate at this location is 0.42 cm year$^{-1}$ (0.82 g cm$^{-2}$ year$^{-1}$).

Cores IF5 and IF6 were collected from adjacent positions (within $\sim25$ m) in the northern basin. In both cases, the $^{137}$Cs concentrations of the bottom sediment increments are either very low or zero and it has been assumed that both cores contain the full $^{137}$Cs depth profile. In core IF6, the $^{137}$Cs concentration peak at $\sim$
6 cm depth can be equated with the 1963 137Cs fallout peak and the sedimentation rate at this location is 0.19 cm year\(^{-1}\) (0.20 g cm\(^{-2}\) year\(^{-1}\)). In core IF5, there are two pronounced peaks in 137Cs concentration at c. 4 and 12 cm. It is difficult to know which peak equates with 1963. Because 137Cs fallout commenced in the mid-1950s, the peak at 12 cm depth is more likely to represent the 1963 peak than the peak located at c. 4 cm depth (which may reflect a localized input of 137Cs-rich sediment at this sampling point in the 1980s). Given the high 137Cs inventory for this core (11 687 Bq m\(^{-2}\)), which is similar in magnitude to that for core IF1 (14 705 Bq m\(^{-2}\)) and the similarity in the 137Cs concentrations for both cores IF1 and IF5, it is suggested that the 137Cs concentration peak at 12 cm can be equated with the 1963 fallout peak. This gives an average sedimentation rate of 0.39 cm year\(^{-1}\) (0.58 g cm\(^{-2}\) year\(^{-1}\)) for core IF5.

The two cores collected from the deltas at the western end of the two lake basins (IF7 and IF8) are also illustrated in Figure 4. In both cases, the 137Cs concentrations of the sediment are \(5 \times 10^3\) mBq g\(^{-1}\) and this partly reflects the very coarse nature of these sediments (see Table I). The depth distributions of 137Cs in these two cores exhibit considerable variability, primarily reflecting marked variations in the particle size composition of the sediment with depth, and there are no obvious peaks that might be equated with the

Figure 4. Caesium-137 depth profiles for the six deepwater cores (IF1 to IF6) and the two delta cores (IF7 and IF8) retrieved from Icefall Lake. The arrows indicate the bottom of the cores.
1963 fallout peak. Because $^{137}$Cs is still present (albeit at low concentrations) in the bottom sediment increment of each core, it is likely that the full $^{137}$Cs profile has not been sampled. However, it can be inferred from Figure 4 that in excess of 15 and 20 cm of coarse sediment has accumulated at sites IF7 (southern delta) and IF8 (northern delta), respectively, since the onset of $^{137}$Cs fallout in the mid-1950s.

Varves. An initial field inspection of the cores collected from Icefall Lake revealed what appeared to be alternating layers of sediment, although it was difficult to distinguish each layer clearly. Consequently, the three cores collected using the Axelsson corer (cores IF2, IF3 and IF6) were analysed by X-ray photography (also see Hasholt, 1995). Figure 5 reveals alternating lighter (higher density) and darker (lower density) sediment layers, which can be interpreted initially as varves. It is assumed that each varve couplet represents sediment accumulation over 1 year and not intra-annual events (pseudovarves).

In core IF2, five complete varves could be distinguished in the upper c. 4 cm of the core, the oldest being attributed to summer 1989. The thickness of the varves ranges from 6-5 to 11-2 mm and the average is 8-0 mm. For summer layers, the range is from 3-7 to 7-4 mm and the average thickness is 5-4 mm, and for winter layers the range is from 0-9 to 3-7 mm and the average thickness is 2-6 mm. Clasts are present in the 1989 summer layer. In core IF3, the uppermost layers are partially disturbed, but below this in the upper c. 15 cm of sediment is a nearly undisturbed sequence of 23 complete varves, the lowest of which has been
attributed to winter 1970. The varves range in thickness from 2.4 to 21.0 mm and the average thickness is 6.6 mm. For the summer layers the thickness of the varves ranges from 0.9 to 20.0 mm and the average is 5.4 mm, whereas the winter layers range from 0.5 to 2.9 mm and the average is 1.3 mm. As with core IF2, clasts are also present in the 1989 summer layer.

As indicated above, there is consistency in the shape of the $^{137}$Cs profiles for cores IF1 to IF3 (which were collected from adjacent locations). There is also consistency in the varve chronologies for cores IF2 and IF3. It is possible to use the $^{137}$Cs-based chronology for core IF1 to check the varve-based chronology for cores IF2 and IF3, although there are problems associated with differences in the time period on which each chronology is based. The similarity in the chronologies for the three cores suggest that the summer and winter layers in Icefall Lake represent annual varve couplets and not pseudovarves (cf. Østrem and Olsen, 1987) related to seasonal variations in sedimentation.

In core IF6 the uppermost and bottom sediment layers were slightly disturbed. However, 15 varves could be identified in the upper c. 11 cm of this core, dating back to summer 1979. The thickness of these varves ranges between 1.9 and 31.0 mm and the average is 7.5 mm. The summer layers range in thickness between 0.9 and 30.0 mm and the average is 6.1 mm, whereas for the winter layers the range is 0.5 to 2.8 mm and the average is 1.4 mm. Small clasts are present in the 1983 to 1985 and 1994 summer layers. In all three cores, the summer layers are thicker than the winter layers.

There is considerable down-core variation in the thickness of the varves within cores IF2, IF3 and IF6 and this is illustrated in Figure 6, which relates the thickness of the summer and winter layers to summer (June to August) temperatures in the study area for individual years. The thickness of the upper layers of cores IF2 and IF3 are similar and this reflects the close proximity of the two cores. The pattern for core IF6 is, however, generally dissimilar to that for cores IF2 and IF3, both in terms of the temporal pattern of varve thickness and the occurrence of clasts in certain layers. Clasts are probably derived from icefloe, caused by ice calving from the glacier snout, which have subsequently melted and deposited coarse material on the lake bed. The fact that clasts are not found in sediment of the same age for all three cores suggests that ice calving introduces sediment locally to the lake bed. Despite the discrepancy between the varve record for the core from the northern basin and those from the southern basin, there are years (such as 1980, 1986 and 1993) where thick varves can be found in both basins, suggesting that years with abundant sediment supply to the lake will result in high sedimentation rates across the majority of the lake bed. There are no statistically significant relationships (at the 95% confidence level) between average summer temperature and summer layer thickness for any of the cores and this implies that processes other than ablation need to be taken into account in order to explain the variations in the annual sediment input to Icefall Lake. Further research is, however, required to substantiate this finding. Similarly, there are no significant trends (at the 95% confidence level) in varve thickness (or summer temperature) over time for any of the three cores. Thus, using the results from core IF3 (the longest core), there is no evidence of a significant trend in sedimentation rates over the period 1970 to 1994.

Sediment yields. Sedimentation rates in Icefall Lake have been estimated using a combination of $^{137}$Cs measurements and varve chronology. Cores IF1, IF2 and IF3 were collected from similar locations in the southern basin and the estimated mean sedimentation rates for these cores are 1.06 g cm$^{-2}$ year$^{-1}$ (average since 1963 based on $^{137}$Cs), 0.94 g cm$^{-2}$ year$^{-1}$ (average since 1989 based on varves) and 0.84 g cm$^{-2}$ year$^{-1}$ (average since 1970 based on varves), respectively. There is, therefore, good agreement between the sedimentation rate estimates for these three cores. Core IF4 was also collected from the southern basin, and the $^{137}$Cs-based average sedimentation rate is estimated to be 0.82 g cm$^{-2}$ year$^{-1}$. The mean sedimentation rate for all four cores from the southern basin based on a combination of varves and $^{137}$Cs data is 0.92 g cm$^{-2}$ year$^{-1}$. Cores IF5 and IF6 were collected from the northern basin and the estimated average sedimentation rates for these cores, based on $^{137}$Cs measurements, are 0.58 and 0.20 g cm$^{-2}$ year$^{-1}$, respectively. However, there are 15 varve couplets present in the top c. 11 cm of (undisturbed) sediment contained in core IF6, which equates to a sedimentation rate of 0.84 g cm$^{-2}$ year$^{-1}$. The reason for this discrepancy between the
sedimentation rates estimated using the $^{137}$Cs depth profile and varves is not known. It is possible that some of the varves are pseudovarves, but, given the agreement between the varves and the $^{137}$Cs-based chronologies for cores IF1 to IF3, this seems unlikely and would only accentuate the difference between the two approaches. Alternatively, the $^{137}$Cs profile could be disturbed. If the sedimentation rate for core IF6 is assumed to be $0.52 \, \text{g cm}^{-2} \, \text{year}^{-1}$ (the average of the two approaches) then the mean for the two cores collected from the northern basin is $0.55 \, \text{g cm}^{-2} \, \text{year}^{-1}$.

The surface areas of the northern and southern basins are c. 0.14 and 0.08 km$^2$, respectively. If, on the basis of lake bathymetry, it is assumed that the area experiencing significant sedimentation in each basin is 75% of the surface area (sediment is unlikely to be deposited on the steeper parts of the lake bed), then the

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**Figure 6. Variations of the thickness (mm) of the summer and winter varves for three cores (IF2, IF3 and IF6) retrieved from Icefall Lake and of the average summer temperature (June, July and August) for the study area over the period 1968–1994**

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annual deposition of fine-grained sediment on the lake bed is estimated to be c. 578 and 552 t, for the northern and southern basin, respectively, and the total is 1130 t. By correcting for an average organic matter content of 0.9% (assumed to be autochthonous) and a trap efficiency of 90%, and assuming that the contributing area (including both glacier and non-glacierized areas) is 3.8 km², the average specific minerogenic fine-grained sediment yield is estimated to be 327 t km⁻² year⁻¹. Assuming a minimum–maximum range for trap efficiency of 80–100%, the sediment yield range is 295–368 t km⁻² year⁻¹.

Several studies have used the amount of material stored within lake deltas to estimate bedload yields from the contributing catchment (e.g. Lambert, 1982; Duck and McManus, 1994). However, owing to the coarse nature of the sediment and the lack of clearly defined ¹³⁷Cs depth distributions in the two delta cores (see Figure 4), it is not possible to estimate coarse sediment accumulation rates in Icefall Lake with any degree of confidence. Furthermore, because a delta grows both vertically and laterally, a more intensive coring programme would be required to provide reliable estimates of sedimentation rate on the delta. Nevertheless, from the ¹³⁷Cs depth profiles of cores IF7 and IF8, it can be inferred that in excess of 15 cm of coarse sediment has been deposited on the delta surface at these two locations since the start of ¹³⁷Cs fallout in the mid-1950s.

**Sedimentation in Lake Kuutuaq**

*Caesium-137 and unsupported lead-210 profiles.* Core retrieval from Lake Kuutuaq also proved problematic and only six undisturbed deepwater cores (K1 to K5) (two from the western basin and four from the eastern basin) were retrieved successfully (Figure 1d). Unlike Icefall Lake, the cores collected from Lake Kuutuaq are spatially representative of deepwater conditions. Three cores were also collected from the delta (cores K6 to K8). The deepwater lake sediments are relatively homogeneous, both laterally across the sampled area of the lake and with depth, and generally consist of light-grey and light-brown silts and clays with some fine sands. The organic matter content of the uppermost lake sediment is very low and averages 0.49 ± 0.01% (x ± 2 s_x) (Table I). The sediment collected from the delta topset beds is very coarse and consists mainly of sands and fine gravels, and the texture varies with depth. The average organic matter content is 0.06 ± 0.01%.

Owing to the anticipated low sedimentation rates in Lake Kuutuaq, the cores collected were not analysed for varves, but one core (K5) was analysed for unsupported ²¹⁰Pb content in order to estimate the sedimentation rate over the last c. 100 years. In view of the small size of the core tubes, the cores were sectioned at 2-cm increments. Furthermore, the depth increments for two cores (K5a and K5b) were combined to produce a composite core (K5) in order to provide sufficient sediment for ²¹⁰Pb analysis. The ¹³⁷Cs profiles for all cores are similar and are illustrated in Figure 7. In all of the deepwater cores (K1 to K5), the sediment contained in the upper 4 to 6 cm evidences very high ¹³⁷Cs concentrations (often >100 mBq g⁻¹), below which depth concentrations decline markedly. Cores K1 and K2 were collected near to each other in the western basin and in both cores the ¹³⁷Cs peak is located in the 2 to 4 cm depth increment. Based on the shape of the profiles, the ¹³⁷Cs peak is probably located at a depth of between 3 and 4 cm in the two cores. If it is assumed that the peak in ¹³⁷Cs concentration occurs at 4 cm depth, then the average sedimentation rate between 1963 and the time that the cores were collected (1992) is c. 0.14 cm year⁻¹ (0.10 cm year⁻¹ if the peak is located at 3 cm depth). The average sedimentation rates expressed in terms of accumulated mass to 3 and 4 cm are 0.08 and 0.10 g cm⁻² year⁻¹, respectively. Core K5 was collected close to the centre of the eastern basin, and as with cores K1 and K2, the ¹³⁷Cs peak is located in the 2 to 4 cm depth increment. If the ¹³⁷Cs peak is assumed to be at 4 cm, then the average sedimentation rate is 0.14 cm year⁻¹ (0.11 g cm⁻² year⁻¹). Cores K3 and K4 were also collected from the eastern basin, near to the delta, and the ¹³⁷Cs peak in these two cores is located in the uppermost 0 to 2 cm increment. If it is assumed that the ¹³⁷Cs peak is located at 2 cm depth, then the average sedimentation rate since 1963 is 0.07 cm year⁻¹ (c. 0.05 g cm⁻² year⁻¹).

Core K5 has also been analysed for unsupported ²¹⁰Pb and the results are illustrated in Figure 8a. The ²¹⁰Pb depth distribution for this core evidences an approximately exponential decrease in content with increasing depth. Using the CFCS and CF models (cf. Appleby and Oldfield, 1978; Robbins, 1978) it is
Figure 7. Caesium-137 depth profiles for the five deepwater cores (K1 to K5) and three delta cores (K6 to K8) collected from Lake Kuutuaq. The arrows indicate the bottom of the cores.

Figure 8. (a) Unsupported $^{210}\text{Pb}$ depth profile for core K5 from Lake Kuutuaq and (b) the depth–age curves established for this profile based on the CFCS and CF models (see text for details). Also plotted is the location of the 1963 $^{137}\text{Cs}$ peak.
possible to produce an age–depth relationship for this core and this is illustrated in Figure 8b. The average sedimentation rate at this sampling point over the last c. 100–150 years is estimated to range between c. 0.11 and 0.15 cm year\(^{-1}\) (0.10 to 0.13 g cm\(^{-2}\) year\(^{-1}\)). These longer term values are essentially the same as those calculated for the last c. 30 years using \(^{137}\)Cs, and suggest that sedimentation at this site has remained essentially constant over the last c. 100–150 years.

Cores K6 to K8 were collected from the topset beds of the main delta at the eastern end of the lake and the \(^{137}\)Cs depth distributions for these cores are shown in Figure 7. As with the delta cores from Icefall Lake, it is difficult to identify the 1963 \(^{137}\)Cs peak in these cores and, therefore, to estimate sedimentation rates with any degree of confidence. However, the occurrence of appreciable \(^{137}\)Cs concentrations at the base of these three cores implies that in excess of between 27 and 36 cm of coarse-grained sediment (assumed to represent bedload) has accumulated on the delta surface over the last c. 40 years.

**Sediment yields.** It is possible to use the \(^{137}\)Cs-based estimates of sedimentation rate obtained for the deepwater cores K1 to K5 to estimate the fine-grained sediment yield to the lake. As the sedimentation rate in this lake appears to be relatively uniform and the rates estimated for the cores from the centre of each basin are the same, an average value of 0.083 g cm\(^{-2}\) year\(^{-1}\) has been used to calculate the sediment yield. The surface area of the lake is 0.61 km\(^2\), and if it is again assumed that significant sedimentation occurs over 75% of the bed, then the mass of fine-grained sediment deposited annually on the lake bed is estimated to be c. 380 t. By correcting for autochthonous organic matter (mean of cores K1 to K5 is 0.49%) and a trap efficiency of 90%, and based on a contributing catchment area of 32.4 km\(^2\) (including both glacier and nonglacierized areas), the average minerogenic fine-grained sediment yield to the lake over the last c. 30 years is estimated to be c. 13.0 t km\(^{-2}\) year\(^{-1}\). Assuming a minimum–maximum range for trap efficiency of 80–100%, the sediment yield range is 11.7–14.6 t km\(^{-2}\) year\(^{-1}\).

**Sediment sources**

It is possible to compare the \(^{137}\)Cs content of the fine-grained sediment deposited in the study lakes with that of potential source materials (see Table II and Figure 9) to identify its most likely source (cf. Walling and Woodward, 1992). Measurements made on samples of material representative of subglacial and englacial sediment indicate that their \(^{137}\)Cs content is very low (<1 mBq g\(^{-1}\)). These low values reflect the fact that

<table>
<thead>
<tr>
<th>Material</th>
<th>(n)</th>
<th>(^{137})Cs content (mBq g(^{-1}))</th>
</tr>
</thead>
<tbody>
<tr>
<td>Lake sediment cores IF1 to IF6 (top 1 cm)</td>
<td>6</td>
<td>23.3 (13.9) 11.0–51.4</td>
</tr>
<tr>
<td>Delta sediment cores IF7 and IF8 (top 1 cm)</td>
<td>2</td>
<td>5.13 (2.26) 2.87–7.38</td>
</tr>
<tr>
<td>Lake sediment cores K1 to K5 (top 2 cm)</td>
<td>5</td>
<td>103 (20.7) 77.8–132</td>
</tr>
<tr>
<td>Delta sediment cores K6 to K8 (various increments)</td>
<td>3</td>
<td>7.13 (4.04) 3.06–12.6</td>
</tr>
<tr>
<td>Suspended sediment from Mittivakkat stream</td>
<td>4</td>
<td>6.36 (4.59) 3.05–14.2</td>
</tr>
<tr>
<td>Grass-covered soil (top 1 cm)</td>
<td>4</td>
<td>204 (72.9) 111–289</td>
</tr>
<tr>
<td>Proglacial material</td>
<td>10</td>
<td>2.97 (3.27) 0–9.89</td>
</tr>
<tr>
<td>Englacial and subglacial material (englacial cave material and material derived from runoff from melting glacier)</td>
<td>2</td>
<td>0.06 (0.06) 0–0.11</td>
</tr>
<tr>
<td>Supraglacial material (on glacier snout)</td>
<td>8</td>
<td>231 (170) 44.5–480</td>
</tr>
<tr>
<td>Alluvial deposits from unglacierized proglacial area near Icefall Lake</td>
<td>6</td>
<td>7.28 (3.25) 4.75–13.9</td>
</tr>
<tr>
<td>Proglacial material near Lake Kuutuaq</td>
<td>1</td>
<td>98.0</td>
</tr>
</tbody>
</table>
Figure 9. Photographs of the main potential sediment sources: (a) englacial and supraglacial material contained within glacier ice calving into Icefall Lake; and (b) exposed sediment on hillslopes in the proglacial zone adjacent to Icefall Lake (the photograph shows the small subcatchment draining into the south-west of the lake, see Figure 1c)
such material has been protected from direct $^{137}$Cs fallout by the ice. The $^{137}$Cs contents of both supraglacial material and the surface horizons of stable grass-covered soils, in contrast, are relatively high (generally $>100$ mBq g$^{-1}$), and this reflects the fact that these materials have been exposed to direct atmospheric fallout. Samples of material collected from the surface of proglacial areas are characterized by average $^{137}$Cs concentrations of $c. 3$ mBq g$^{-1}$. These relatively low values reflect, firstly, the fact that such material represents a mixture of glacial sediments (i.e. englacial, subglacial and supraglacial sediment) that are being temporarily stored in the proglacial zone and, secondly, that, prior to glacial retreat, some proglacial areas would have been protected from $^{137}$Cs fallout. Similarly, alluvial deposits from non-glacierized areas near to Icefall Lake were found to have an average $^{137}$Cs content of $c. 7$ mBq g$^{-1}$. The recently deposited (i.e. uppermost 0 to 1 cm) sediment in the deepwater cores collected from Icefall Lake is characterized by $^{137}$Cs concentrations that typically range between $c. 11$ and $51$ mBq g$^{-1}$, with a mean of $23$ mBq g$^{-1}$. In the absence of more fingerprint properties (cf. Collins et al., 1997; Walling et al., 1999), it is difficult to determine quantitatively the precise source of the lake sediment, and, given the $^{137}$Cs concentrations of potential source materials listed in Table II, the lake sediment could be derived from a number of possible source combinations. It seems unlikely, however, that en- or subglacial material ($^{137}$Cs concentrations too low) or supraglacial material or grass-covered soil ($^{137}$Cs concentrations too high) individually represent the primary sediment sources. Furthermore, the high organic matter content of grass-covered soil ($\bar{x} = 17.3\%$), compared with lake sediment ($\bar{x} = 0.9\%$), again suggests that it is unlikely to be a major source. Given that the fine-grained sediment samples collected from the proglacial zone are noticeably coarser than the lake sediment (Table I), and recognizing the preferential affinity of $^{137}$Cs for fine-grained material (He and Walling, 1996), fine-grained sediment mantling the steep slopes in the proglacial zone probably represents the dominant source of the sediment deposited in Icefall Lake (Figure 9b). The $^{137}$Cs activities associated with the uppermost sediment from the two delta cores are also relatively low ($\bar{x} = 5.13$ mBq g$^{-1}$), and, given the coarse texture of these deltaic deposits, this sediment is more likely to be derived from proglacial sources.

Although 79% of the contributing area to Icefall Lake is glacial, most of the catchment immediately bordering the lake consists of steep, non-glacierized slopes. These are mantled with large quantities of fine-grained (i.e. <2 mm) mineral material and either feed sediment directly into the lake or supply sediment to small channels such as the two streams entering the western end of the lake (see Figure 9b). Thus, the source tracing results obtained from both the deepwater and delta cores are in broad agreement with the visual importance of proglacial sediment sources bordering the lake. These inferences on sediment sources should, however, be viewed in the context of the location of the sediment cores (which, as described earlier, were necessarily collected from the western end of the lake) relative to the inflowing streams to the west and the more distal location of the glacier front (see Figure 1c).

Table II also lists values of $^{137}$Cs concentration for suspended sediment samples collected as part of the larger Greenland project from Mittivakkat stream, which drains the glacier snout to the south of Icefall Lake (Figure 1b). Such material can be considered to represent actively transported fine-grained sediment, and the area contributing to the stream sampling site is similar in character to that draining into Icefall Lake (i.e. relatively near to the glacier and representing the immediate proglacial foreland zone). When allowance is made for differences in particle size composition (Table I), there is close agreement between the $^{137}$Cs concentrations associated with actively transported suspended sediment, the lake sediment and the material in the proglacial zone, again suggesting that this zone is the main sediment source.

The $^{137}$Cs content of the uppermost sediment in Lake Kuutuaq is considerably higher ($\bar{x} = 103$ mBq g$^{-1}$) than that of the sediment deposited in Icefall Lake. Even allowing for particle size differences (see Table I), subglacial, englacial and supraglacial material can be discounted as major sediment sources. Furthermore, Lake Kuutuaq is over 5 km from the glacier front and the glacier represents $<30\%$ of the catchment contributing to the lake. A composite sample of surface material from the proglacial area near to Lake Kuutuaq had a $^{137}$Cs content of 98 mBq g$^{-1}$. This $^{137}$Cs value is considerably higher than that associated with the proglacial material collected in the vicinity of Icefall Lake, and reflects the greater $^{137}$Cs fallout input.
to surface materials in the area around Lake Kuutuaq, which was ice-free throughout the period of bomb-derived fallout (i.e. since the mid-1950s). Thus, fine-grained proglacial material may again be inferred as the major source for the uppermost deepwater sediment deposited in Lake Kuutuaq.

**DISCUSSION AND CONCLUSION**

The specific suspended sediment yields estimated for the catchments draining into Icefall and Kuutuaq lakes are 327 and 13 t km\(^{-2}\) year\(^{-1}\), respectively. These values are similar to those documented for other arctic and subarctic glacierized catchments (Ostrem et al., 1967; Gurnell et al., 1996; Hodson et al., 1998). They are, however, generally lower than those reported for alpine glacierized catchments or catchments containing predominantly warm-based glaciers influenced by recent volcanic or tectonic activity (i.e. Icelandic catchments), which may be several orders of magnitude higher (Hicks et al., 1990; Collins, 1996; Gurnell et al., 1996). Thus, for example, Hodson et al. (1998) estimated specific suspended sediment yields of 86 and 110 t km\(^{-2}\) year\(^{-1}\) for two gauging sites on the proglacial river that drains the Brøggerbreen Glacier in Svalbard, and Gurnell et al. (1996) report yields of between c. 20 and 2000 t km\(^{-2}\) year\(^{-1}\) for arctic basins world-wide.

The sediment yield from the catchment contributing to Icefall Lake (79% glacier) lies within the range 84 to 1500 t km\(^{-2}\) year\(^{-1}\) reported for other glacierized areas in Greenland (Hasholt, 1996), whereas that for the catchment contributing to Lake Kuutuaq (29% glacier) is within the range 1 to 56 t km\(^{-2}\) year\(^{-1}\) for non-glacierized areas (Hasholt, 1996). The much lower specific sediment yield associated with Lake Kuutuaq reflects the greater opportunity for sediment deposition during transport to the lake, particularly in the lakes within the upper part of the catchment (such as Icefall Lake) and along the proglacial valley, which act as major sediment sinks. This finding confirms previous work in this area by the authors (Hasholt and Walling, 1992; Busskamp and Hasholt, 1996). Consequently, the sediment delivery ratio associated with the proglacial zone in the Lake Kuutuaq catchment is likely to be relatively low. In the case of Icefall Lake, the glacier and many of the hillslopes in the contributing proglacial zone feed sediment directly into the lake (see Figure 9) and there is, therefore, less opportunity for intermediate storage. In consequence, the sediment delivery ratio of the catchment draining to Icefall Lake is likely to be considerably higher than that for Lake Kuutuaq.

On the basis of the evidence for the two study lakes, there appears to be a rapid decline in the sediment yield and, by inference, the sediment delivery ratio with both increasing catchment area (and thus opportunities for sediment storage) and decreasing percentage glacier cover (and thus decreasing sediment supply from glacier sources). It is not possible, however, to establish the relative importance of the influence of catchment area and percentage glacial cover on sediment yields and sediment delivery ratios in this study. Desloges and Gilbert (1998) document a similar decrease in specific sediment yield with decreasing percentage glacier cover for alpine glacier-fed lakes in the southern Canadian Cordillera, whereas Hicks et al. (1990) found no significant relationship with glacier cover for alpine basins in New Zealand and arctic and alpine basins world-wide. Generally, however, the importance of percentage glacier cover, and thus sediment supply from glacial sources, in controlling suspended sediment yield is still uncertain (cf. Gurnell, 1987; Harbor and Warburton, 1993). Desloges and Gilbert (1998) used the catchment area to lake area (CA/LA) ratio as an index of the potential for intermediate storage of sediment. This is because, as the CA/LA ratio increases, the potential for sediment storage in locations upstream of the lake increases and, therefore, the sediment delivery ratio decreases (cf. Dearing and Foster, 1993). The CA/LA ratios for Icefall and Kuutuaq lakes are 17 and 53, respectively, and these values are consistent with the lower sediment delivery ratio inferred for the catchment contributing to Lake Kuutuaq compared with that for Icefall Lake.

A significant quantity of sediment has been deposited in the two study lakes, both within the main lake basins and on the deltas. Consequently, where lakes are present in the proglacial zone, downstream sediment yields, and thus sediment fluxes to the sea, are likely to be considerably reduced.

From the analysis of the variation of varve thickness through time in Icefall Lake and a comparison of the sedimentation rates estimated for Lake Kuutuaq based on \(^{137}\)Cs and unsupported \(^{210}\)Pb, there is no evidence for a significant change in sedimentation rates and thus sediment yield to the lakes over the last c. 24 and
100–150 years, respectively. This relative stability in sedimentation could be seen as somewhat surprising, in view of the known retreat of Mittivakkat Glacier over the past c. 100 years and the associated increase in the area of exposed proglacial material.

The proglacial zone has been identified as the primary source area of the recent fine-grained sediment deposited in the two study lakes. For Icefall Lake, this reflects the importance of sediment supply from the steep slopes immediately bordering the lake, which feed sediment directly into the lake in many places, whereas for Lake Kuutuaq it reflects the small percentage of glacier cover in the contributing catchment and the large distance between the lake and the glacier front (c. 5 km). The findings for the two study lakes are consistent with existing knowledge of the suspended sediment transport dynamics of the Mittivakkat stream (cf. Hasholt, 1976, 1992, 1994), which confirm the importance of rainfall events, and thus of the proglacial zone as a main source area of the transported sediment. Hodson et al. (1998) also found that the majority of fine-grained sediment transported in the meltwater stream from the Brøggerbreen Glacier, Svalbard, was derived from both ice-marginal and proglacial areas, and that subglacial sediment sources were of negligible importance, owing to the limited subglacial drainage associated with this cold-based glacier. As the Mittivakkat Glacier is warm-based, the results reported above suggest that the importance of the proglacial zone as the dominant source area may not be limited to cold-based glacierized arctic catchments.

In this study, the proglacial zone has been identified as both the dominant source area and an important sediment sink. The importance of the proglacial zone in controlling sediment yields also has been documented for alpine glacierized basins (cf. Gurnell, 1987; Harbor and Warburton, 1993; Warburton, 1999). More work, however, is clearly needed to elucidate the temporal and spatial variability of sediment delivery and transport within proglacial zones in arctic and subarctic glacierized basins, and the role of the proglacial zone in influencing downstream sediment yields.

ACKNOWLEDGEMENTS

The financial support provided by the Danish Natural Science Research Council (SNF) and the support of the Universities of Copenhagen and Exeter for the work reported in this paper are gratefully acknowledged. Thanks are also due to Qingping He for assistance with radionuclide analysis, Art Ames for assistance with laboratory analysis and Terry Bacon, Barry Phillips, Kent Poerksen and Andrew Teed for producing the diagrams. Comments from Ian Foster and an anonymous referee have helped to improve the paper.

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