Available online at www.sciencedirect.com
ScienceDirect
Geomorphology 79 (2006) 130–142
GEOMORPHOLOGY
www.elsevier.com/locate/geomorph

The geomorphic and paleoenvironmental record in the sediments of Atlin Lake, northern British Columbia

Robert Gilbert [a,*], Joseph R. Desloges [b], Scott F. Lamoureux [a], Andrea Serink [b], Kyle R. Hodder [a]

[a] *Department of Geography, Queen's University, Kingston, ON, Canada K7L 3N6*
[b] *Department of Geography, University of Toronto, Toronto, ON, Canada M5S 3G3*

Received 9 September 2004; received in revised form 29 September 2005; accepted 29 September 2005
Available online 12 June 2006

Abstract

Atlin Lake in northern British Columbia and southern Yukon Territory is the largest natural lake in the North American Cordillera (791 km^2). Inflow from the Juneau Ice Field delivers large volumes of sediment to the proximal basins of Willison Bay and Llewellyn Inlet. Sediment is distributed by interflow and underflow through these basins. Based on acoustic data, each of these basins contain Holocene deposits about 120 m thick, representing mean annual accumulation since deglaciation of more than 1 cm/a. Cores confirm this, except that the formation of a small lake at the toe of Llewellyn Glacier during about the past 50 years is trapping sediment and has reduced accumulation in Llewellyn Inlet by an order of magnitude. Sills separate these basins from the main body of Atlin Lake and Torres Channel where accumulation is much less, averaging about 1–4 mm/a during the history of the lake. Late Pleistocene glacilacustrine sediment occurs as a thin, patchy deposit and is overlain by up to 10 m of Holocene lacustrine deposits. Unlike other large lakes in the Cordillera with thick late Pleistocene deposits indicating large volumes of sediment contributed by glaciers in the lakes and their basins, the pattern in Atlin Lake documents rapid retreat of glaciers from the lake and much of the drainage basin.

Keywords: Lake; Sediment; Accumulation; Canadian Cordillera; Holocene

1. Introduction

Because Professor Slaymaker's work which is celebrated with this publication was directed to understanding the geomorphology of the Canadian Cordillera, it is appropriate to consider the place of large lakes in documenting the evolution of montane landscapes and the processes associated with them. We do this by describing the sedimentary environment of Atlin Lake (Fig. 1), the largest natural lake in the Cordillera of North America. The sedimentary record of Atlin Lake is compared to those of other large lakes, especially in British Columbia which have been studied since Professor Slaymaker supervised the first detailed research on these lakes more than 30 years ago (Gilbert, 1975).

Lakes provide a potentially continuous archive of environmental changes detailing variations in processes both within the lake and in the contributing watershed (Last and Smol, 2001). Although logistically more challenging to study because of their large area extent and depth, large lakes (arbitrarily defined here as having

* Corresponding author. Fax: +1 613 533 6122.
E-mail address: robert.gilbert@queensu.ca (R. Gilbert).

0169-555X/$ - see front matter © 2006 Elsevier B.V. All rights reserved.
doi:10.1016/j.geomorph.2005.09.021

a surface area greater than 10 km^2) offer several advantages for investigating environmental and paleoenvironmental change. Their large volumes insure that the sediment trapping efficiency approaches 100% (Brune, 1953), and thus even the finest sediment that may be the most diagnostic of climatically controlled processes in the drainage basin is deposited in the lake. The normally greater depth insures that the lake floor is more isolated from more vigorously circulating and wave-mixed surface waters, creating a quiet depositional environment where sediment is less disrupted. Their large areas mean that distinct environments of proximal deltaic processes are well separated from distal environments where fine sediments are found; each represents distinct processes and controls and therefore provide different proxies. As well, areas of the lake floor are away from wave- and ice-generated sediment input near the shore, and particularly from subaerial and subaqueous slope failures along steep lake sides, which may mask environmentally meaningful deposition. Finally, large lakes normally have large drainage basins, which contain diverse sediment sources. While a single source such as glacial meltwater which dominates inflow may be valuable for some assessments, spatially

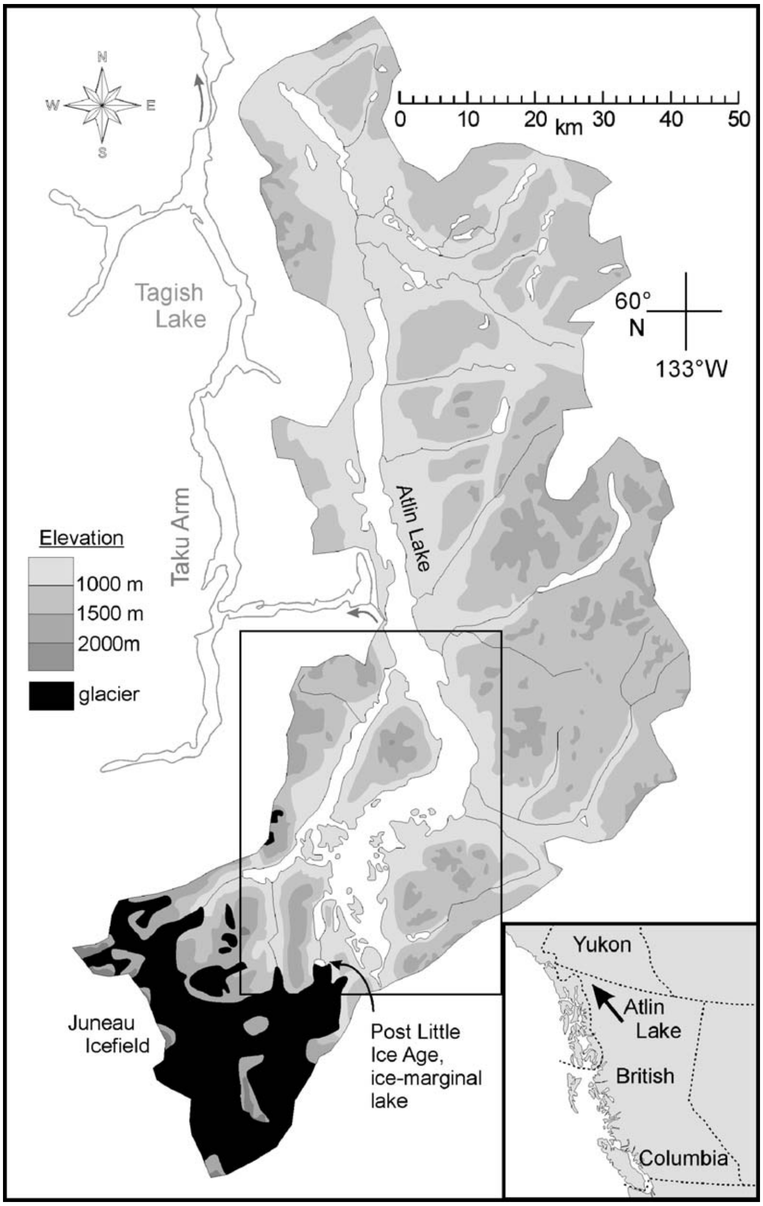

Fig. 1. Map of the location and drainage basin of Atlin Lake. Box outlines Fig. 2.

integrated sources are more inclusive of a range of processes and responses. They reduce the effect of single catastrophic events such as landslides or glacial surges, which may have limited large-scale environmental meaning.

2. Physical setting

Atlin Lake has a surface area of 791 km^2. Its drainage basin (6410 km^2) forms the headwaters of the 3200-km long Yukon River in the northernmost portion of the Coast Range Mountains. Juneau Ice Field (Sprenke et al., 1999) and nearby smaller glaciers occupy 619 km^2 in the southern portion of the drainage basin where maximum elevations reach 2400 m a.s.l. (Fig. 1). Glacial meltwater and sediment dominate the proximal lacustrine environment of its two interconnected arms, Torres Channel–Willison Bay, and Sloko and Llewellyn Inlets (Fig. 2). To the north and east, the drainage basin lies in the northwestern portion of the Stikine Plateau. Although elevations reach 1800 m a.s.l., this region is in the rain shadow of the Coast Mountains; there are no

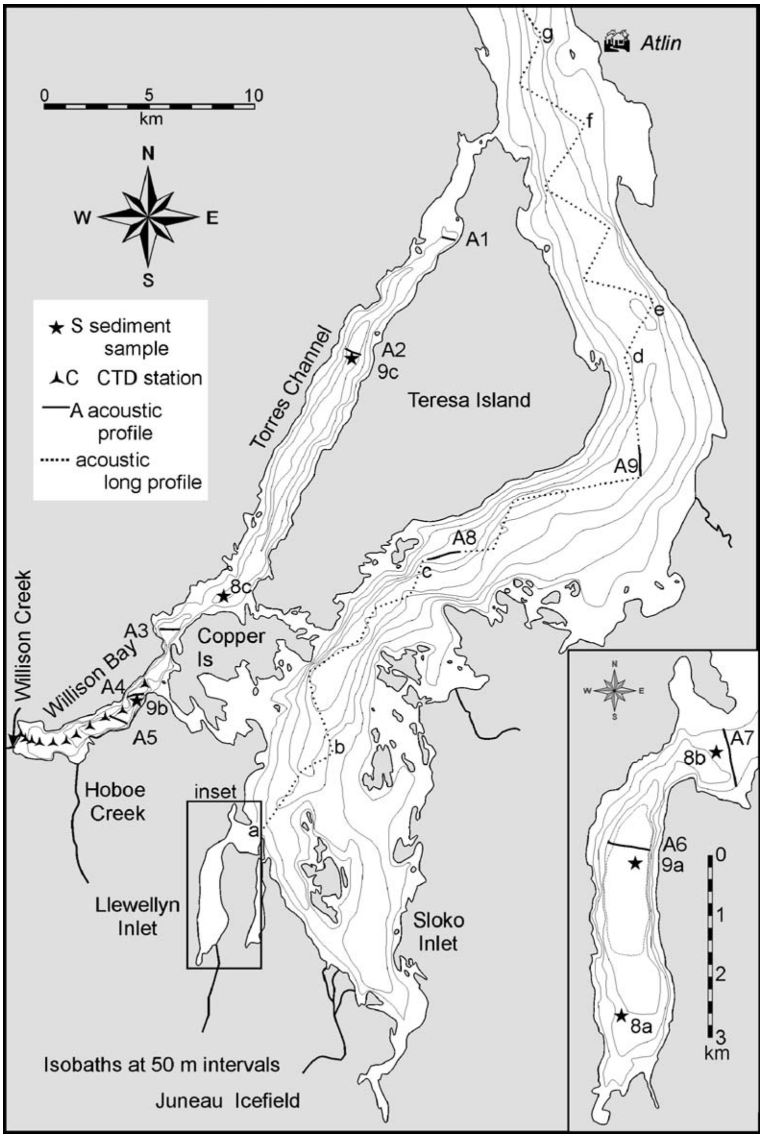

Fig. 2. Bathymetry of Atlin Lake (with detail in Llewellyn Inlet inset) determined from the acoustic survey reported in this paper, supplemented with data from the British Columbia Game Department, Fisheries Management Division, in Sloko Inlet and eastern Atlin Lake. Isobath interval is 50 m with 175 m shown as a dashed line in Llewellyn Inlet. Also shown are the locations of the profiles and displayed acoustic sections (Figs. 5 and 6), and CTD (Fig. 3) and sediment sampling (Figs. 8 and 9) sites.

glaciers, and inflow and sediment input is much less than in the southern portion of the drainage basin.

Water depths reach 166 m in Willison Bay, rising to 80 m on a sill north of Copper Island and to 77 m in Torres Channel (Fig. 2). In Llewellyn Inlet, a large region of flat-lying floor occurs at 178 m depth, with a sill at 98 m at the north end of the inlet. In the main body of Atlin Lake, depths reach a maximum of 289 m. Narrow interconnecting channels north and south of Copper and Teresa Islands and at the north of Torres Channel are less than 10 m deep.

3. Methods

Between 1999 and 2003, we conducted studies of Atlin Lake, including Torres Channel, Willison Bay, Llewellyn Inlet and, to a lesser extent, the main body of Atlin Lake south of the community of Atlin (Fig. 2). A sub-bottom acoustic survey of the lake was conducted using a Benthos Chirp dual-frequency digital system (Gilbert, 2003). Positions recorded using a single GPS receiver accurate to ±5 m were logged at 1-s intervals. A total of 164 km of transects was recorded. Post-processing involved digitizing (a) the depth of the lake floor, (b) facies in the sediment and (c) the acoustically impenetrable surface beneath the sediments at 1-min intervals (average 125 m along track lines). The velocity of sound in water and sediment was assumed to be 1440 m/s, appropriate for the water temperatures recorded in the lake (Kuperman and Lynch, 2004). Even if the sound velocity increases by 1 m/s per metre depth in the sediment, the error in assuming a constant velocity is less than 1.5%. Except in the main body of Atlin Lake where the track lines were widely spaced, the data were mapped by hand to create maps of bathymetry and sediment thickness.

CTD casts recording conductivity (dissolved sediment), temperature and transmissivity (suspended sediment) were made from surface to bottom in a number of sites in the lake using Hydrolab Datasonde and Seabird SeaCat instruments. Suspended sediment concentration (SSC) in mg/L was calculated from turbidity in nephelometric turbidity units (NTU) according to the relation SSC = 0.973 NTU − 2.11 (r^2 = 0.951, n = 29), using samples filtered at 0.45 μm from surface to near bottom waters in both proximal and distal locations in a variety of glacial lakes in British Columbia,

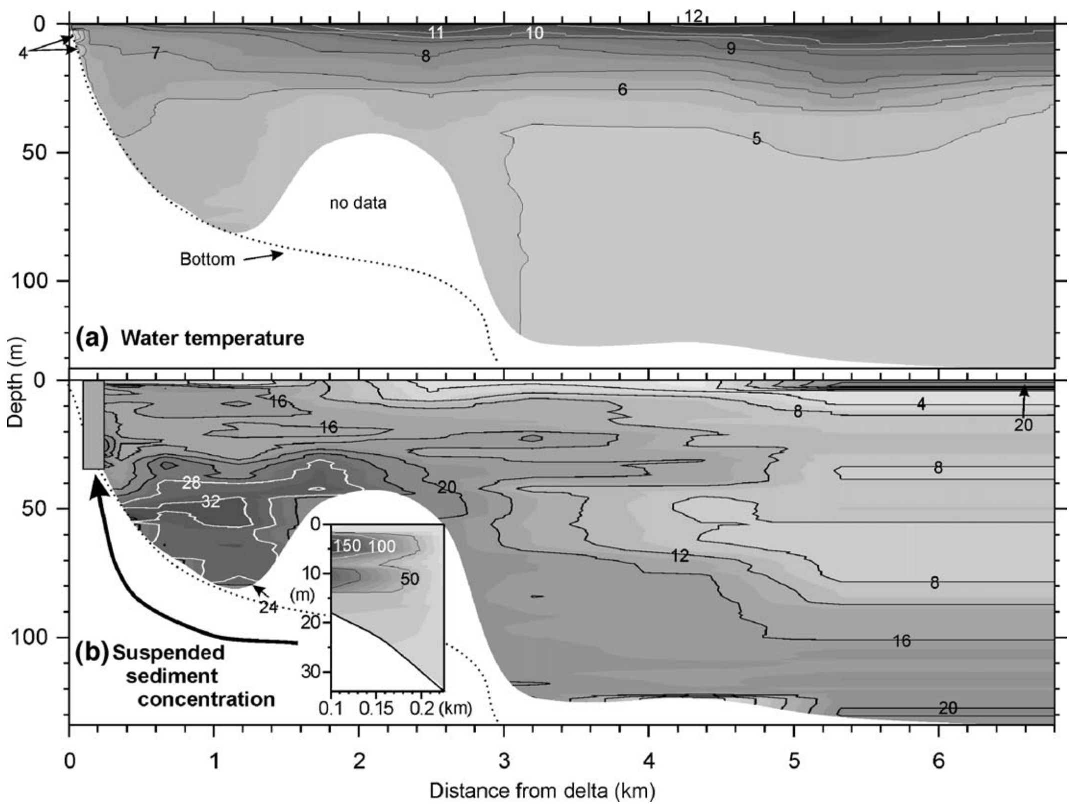

Fig. 3. (a) Water temperature (°C) and (b) suspended sediment concentration (mg/L) in Willison Bay, 8 August 2001. For locations of sampling sites, see Fig. 2.

including Atlin (cf. Gilbert et al., 1997). Short cores were obtained with an Ekman box corer, and 4-cm and 8.9-cm diameter gravity corers. In the laboratory, the cores were split, photographed, logged and sampled for grain-size determination using a Coulter LS 200 laser scattering particle size analyzer. Thin sections were prepared by freeze-drying and embedding samples under low vacuum with a single application of Spurr's low viscosity epoxy resin (Lamoureux, 2003).

4. Results

4.1. Physical limnology

Data on the physical limnology over an extended time are not available for Atlin Lake. However, comparison between our data from July and August 1999 and from 2002 indicates that a series of CTD casts in August 2001 (Fig. 2) are typical of conditions during the height of the melt season for the streams draining the Juneau Ice Field to Willison Bay and Llewellyn and Sloko Inlets (Fig. 2). In Willison Bay, two streams drain the eastern margin of the Juneau Ice Field: Willison Creek entering the western end of the bay and Hoboe Creek entering from the south, 3 km to the east (Fig. 2). Inflowing meltwater in these streams is near 4 °C and cools the proximal lacustrine environment (Fig. 3a), but within several kilometres down-lake, the surface warms to 10 °C or more. A well-mixed epilimnion does not develop in this region of the lake and the temperature decreases to values near 4 °C by about 40 m depth. In the most proximal region, interflows at 5–12 m thick and charged with up to 150 mg/L of suspended sediment (Fig. 3b) occur as a diffuse cloud about 50 m depth, dissipating down-lake as sediment

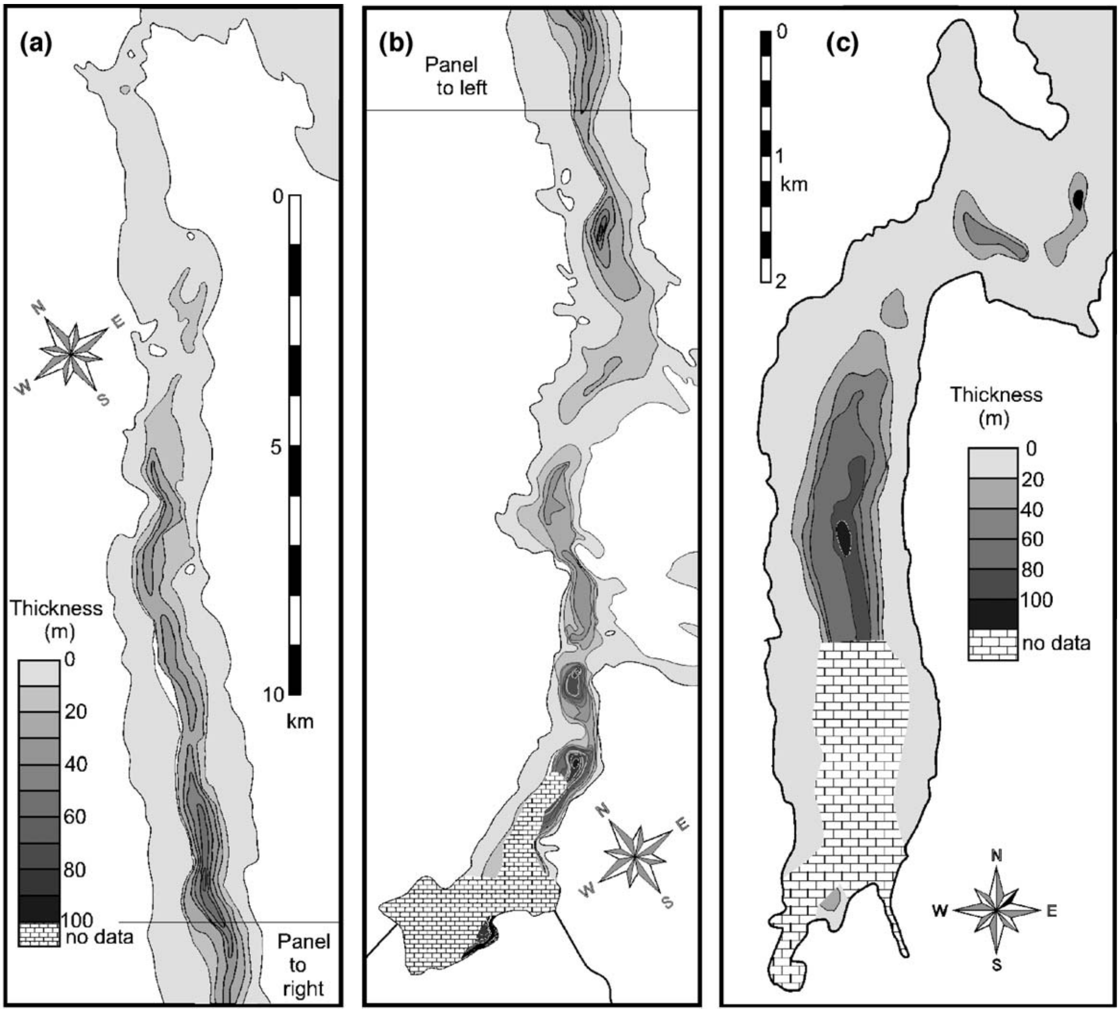

Fig. 4. Thickness of sediment based on sound velocity of 1440 m/s in (a) northern Torres Channel, (b) southern Torres Channel and Willison Bay at the same scale and orientation, and (c) Llewellyn Inlet.

settles from suspension. The concentrations are also sufficient to generate underflow as low density turbidity currents, which spread through this region of the lake. Higher concentrations of up to 20 mg/L in the distal surface water represent inflow from Hoboe Creek that initially deflected along the right hand (south) shore by the Coriolis effect, subsequently directed to mid channel by promontories on the coast.

In Llewellyn Inlet, inflow from the Juneau Ice Field occurs as a single stream entering the southwestern corner of the inlet (Figs. 1 and 2) after passing through a 2-km^2 ice-marginal lake (Fig. 1). The inflow is about the same temperature as in Willison and Hoboe Creeks, and the thermal regime of Llewellyn Inlet is very similar to that in Willison Bay. Surface temperatures increase distally to 12–14 °C at the north end of the inlet. As in Willison Bay, a well-mixed epilimnion is absent and temperatures decrease rapidly downward to about 40 m depth below which they are 4–5 °C. Suspended sediment concentrations in the inflow are up to 0.6 g/L and create a surface plume with concentrations up to 50 mg/L near the point of inflow. The plume spreads down the east (right) side of the inlet. Concentrations in the lake water below about 5 m depth are about 5 mg/L. Some sediment is lost from the plume by settling and observations of temperature through time indicate that turbidity currents are generated from this inflow. Even with these losses of sediment from the surface plume, concentrations near the surface remain in the range 20–25 mg/L at the north end of the inlet.

Another significant inflow from the Juneau Ice Field occurs on the west side of Sloko Inlet (Fig. 2). The pattern of distribution of water and sediment in Sloko Inlet is similar to Llewellyn and Willison except that the inflowing temperature is several degrees warmer and the surface temperatures in the lake are 12–13 °C within 3 km of the delta front. The turbidity data indicate a strong interflow about 20 m depth with concentrations exceeding 35 mg/L. Concentrations above and below are significantly less except that slightly elevated concentrations of 10–15 mg/L near the lake floor indicate that a weak underflow extends about 3 km from the delta front. Beyond 3 km, both turbidity and temperature are nearly uniform with depth except for the warming near the surface as seen in Llewellyn Inlet and Willison Bay.

4.2. Acoustic survey

In Torres Channel and Atlin Lake and in their headwater arms, sediment accumulation is concentrated

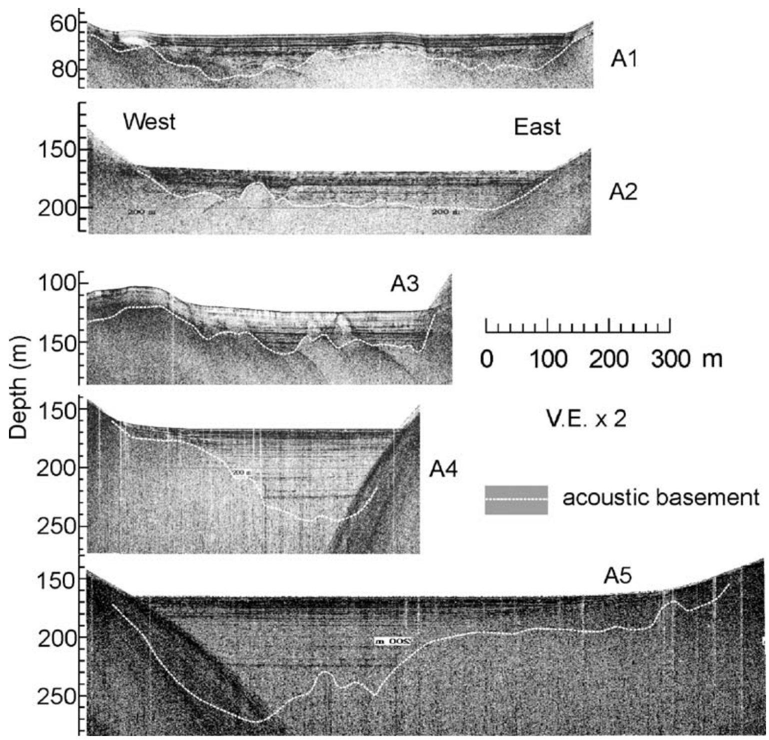

Fig. 5. Subbottom acoustic images of Torres Channel and Willison Bay. Locations are shown on Fig. 2. In each case, the view is looking down-lake (north or northeast). Note the interfering returns from steep side walls in sections A3 and A5.

in the deepest portion of the basins (Fig. 4), with additional sediment occurring as thin veneers along the sides and over sills In the vicinity of inflow to Willison Bay and Llewellyn Inlet, the sediment contains sand in amounts sufficient to limit or preclude penetration of sub-bottom sound as is typical of glacial lakes (Desloges and Gilbert, 1994a, 1995; Gilbert et al., 1997; Desloges and Gilbert, 1998). However, in more distal areas, the acoustic basement of bedrock or compact glacier-proximal sediment can be distinguished almost everywhere (Figs. 5 and 6). The accumulation of lacustrine sediment in the proximal basins exceeds 100 m, which is about the limit of penetration with Chirp acoustics in ideal conditions (e.g. Fig. 6(A6)). A fourth order polynomial curve fit to the bedrock surfaces of the valley sides, above and below water level (Gilbert et al., 1997), allows us to estimate the basement beyond the depth of acoustic penetration and indicates that, in this region (Fig. 4c), the maximum sediment thickness is about 120 m. This is somewhat less than the maxima reported in other large lakes of the Cordillera (Table 1). This suggests a mean rate of accumulation since deglaciation of about 1 cm/a.

The sediment occurs as a single acoustic facies in both proximal basins (Figs. 5(A5) and 6(A6)). It lies unconformably on the irregular basement as a flat-lying, acoustically stratified deposit with closely spaced, nearly horizontal, parallel, reflective surfaces extending laterally and distally for up to several kilometres, and, in some cases, throughout the basin. There is little or no sediment on the side walls of the proximal basins. The way in which these sediments are distributed indicates that deposition has been predominantly from turbidity currents. Because there is no acoustic evidence for large slides or slumps, we infer that any sediment deposited from suspension is reworked and quasi-continuously transported to the deep basin.

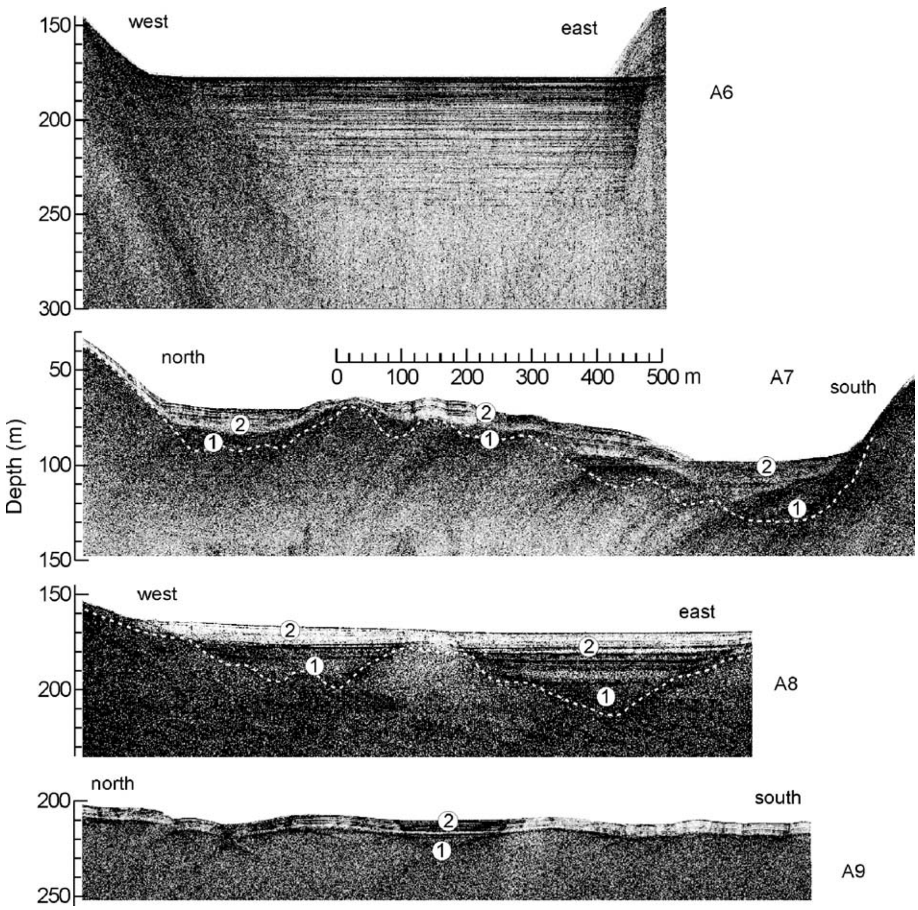

Fig. 6. Subbottom acoustic images of Llewellyn Inlet (A6) and Atlin Lake (A7–A9). Profile A6 extends below the limit of penetration of the instrument and apparent overlap between horizontal and sloping reflectors along the sides if the basin is an artefact of the geometry of sound propagation. Numbers refer to acoustic facies described in the text. Locations are shown on Fig. 2.

Table 1
The sedimentary record of large lakes in the Canadian Cordillera based on acoustic and seismic survey

Lake	Area (km^2)	Drainage basin (km^2)	Area ratio (basin/lake)	Present glacier cover (%)	Maximum water depth (m)	Maximum sediment thickness (m)	Source
Atlin	791	6410	8.10	9.7	289	120	This study
Kluane	432	5426 [a]	12.6	14.3	82	95	Gilbert (unpub.)
Okanagan	369	na [b]	–	<1	238	792 [c]	Eyles et al. (1990, 1991)
Quesnel	272	5264	19.4	0.6	614	70	Gilbert (unpub.)
Harrison	223	7870	35.3	8	270	>80	Desloges and Gilbert (1991)
Chilko	182	1960	10.7	9.9	339	63	Desloges and Gilbert (1998)
Kusawa	142	4070	28.6	4	132	80	Gilbert and Desloges (unpub.)
Shuswap	132	na [b]	–	0	171	800 [b]	Eyles and Mullins (1997)
Bennett	94	3532	37.5	2.2	125	>100	Gilbert and Desloges (unpub.)
Stave	59	883	15.0	3.3	100	40	Gilbert and Desloges (1992)
Bowser	34.5	1449 [d]	42.0	41	119	370	Gilbert et al. (1997)
Meziadin	34	588 [e]	17.3	8.6	133	185	Gilbert and Butler (2004)
Kinaskan	28	1257	44.9	0.75	120	60	Desloges (unpub.)
McDonald	26	454	17.5	0.54	131	150	Mullins et al. (1991)
Lillooet	20.5	3580	174	7.0	137	183	Gilbert (1975) Desloges and Gilbert (1994a)
Kalamalka	25.6	na [b]	–	–	237	272	Mullins et al. (1990)
Kitsumkalum	19.1	1800	94.2	4	138	75	Gilbert and Menunos (unpub.)
Moose	15.3	1640	107	3.2	86	80	Desloges and Gilbert (1995)
Waterton	13.4	614	45.8	>1	145	55	Eyles et al. (2000)

[a] Assuming 50% of the Kaskawalsh Glacier drainage area feeds via Slims River to Kluane Lake.
[b] The area of late Pleistocene glacial drainage responsible for most of the sediment fill was much different than the present drainage.
[c] Based on seismic profiling equipment which is able to penetrate glacial and ice-proximal sediments opaque to 3.5 kHz (Chirp) acoustic profilers and thus the sediment resolved is thicker than recorded with that equipment.
[d] Including the area of drainage from Summit Lake and Salmon Glacier no longer contributing to Bowser drainage.
[e] Including the area of drainage from Strohn Lake and Strohn Glacier no longer contributing to Meziadin drainage.

The sills north and northwest of Copper Island are <80 m deep and those at the northern end of Llewellyn Inlet are about 100 m deep (Fig. 2). They are important in distinguishing the rapid accumulation in the proximal zone from significantly lower accumulation distally. In Torres Channel, the sediment reaches a maximum thickness of just over 60 m in two small basins (Fig. 4a,b) and thins until in the distal 7 km at the north end of the Channel, there are only a few places where sediment is more than 10 m thick. This suggests mean rates of accumulation since deglaciation of about 4 to <1 mm/a. The sediment is concentrated along the thalweg of the channel and, in many places, there is little or no accumulation on the side walls (e.g. Fig. 5(A2)), despite the significantly lower slopes in this region. Again, there is no evidence of sediment slumping off the side walls in large events; sediment accumulated here is quasi-continuously focused to the deep basin.

In the main body of Atlin Lake, the acoustic survey consists of one line, which zig-zags across the thalweg of the basin (Fig. 2). Thus, it is not possible to map the sediment thickness, although the linear results (Fig. 7) document a thin veneer of lacustrine sediment throughout this region of the lake. Nowhere does the total thickness exceed 40 m and in only six small basins does it exceed 30 m. Thus, most of the sediment input from the streams flowing into Llewellyn and Sloko Inlets is trapped by the sills at their northern ends, and does not reach the deepest basin of Atlin Lake.

Through most of the main part of Atlin Lake, two acoustic facies can be distinguished (Figs. 6 and 7). The lower facies 1 is a patchy, thin deposit up to 20 m thick; it is nearly massive and acoustically dense. Layering is poorly defined and irregularly sloping in the southern (proximal) region of the basin (Fig. 6 (A7)). Distally, the layering becomes more uniform and horizontal and the sediments unconformably overlie the basement, although the acoustically dense character is retained (Fig. 6(A8)). These characteristics suggest a brief period of deposition in a glacially proximal lacustrine environment during or immediately following deglacation.

Facies 2 either overlies facies 1 or rests directly on the acoustic basement. It consists of a more continuous

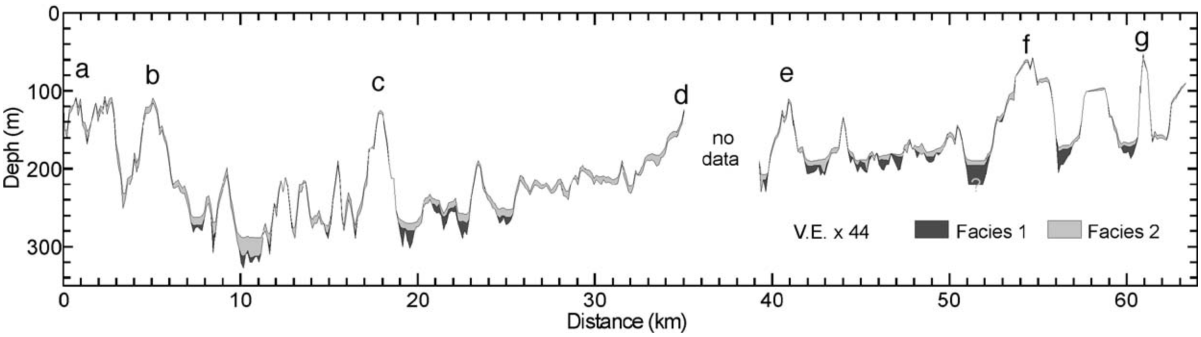

Fig. 7. Profile of sediment thickness in Atlin Lake. Facies 1 and 2 are described in the text. For location, see Fig. 2.

cover of acoustically more transparent sediment, which is well layered and lies conformably on the material below. This suggests that facies 2 was deposited largely from suspension in the water column. It reaches its maximum thickness of 15 m in the deepest basin but the mean thickness is only 6.5 m. Thus, the mean accumulation rates since deglaciation are less than 1 mm/a except in a few isolated locations.

4.3. Sedimentary record

Fig. 8 is typical of the sedimentary structures visible in many of the Ekman, gravity and percussion cores recovered from Atlin Lake. In both proximal and distal areas of Llewellyn Inlet, surficial sediments occur as thin silt-clay couplets that average about 2 mm thick (Fig. 8a). These also are evident in thin sections from the cores (Fig. 9a). Cs-137 dating (Serink, 2004) confirms that these laminae are annual and that this pattern of sedimentation has continued for about the last 50 years. Between about 1850 and 1920, long cores show that sediment accumulation was much greater, averaging 4–9 mm/a (Fig. 8b). These older couplets are structurally more complex, with selected years showing 1–3 mm thick, multiple, normally graded, coarse silt layers deposited during the melt season. Autumn through winter accumulation is significant, forming silt-clay caps of 3–5 mm.

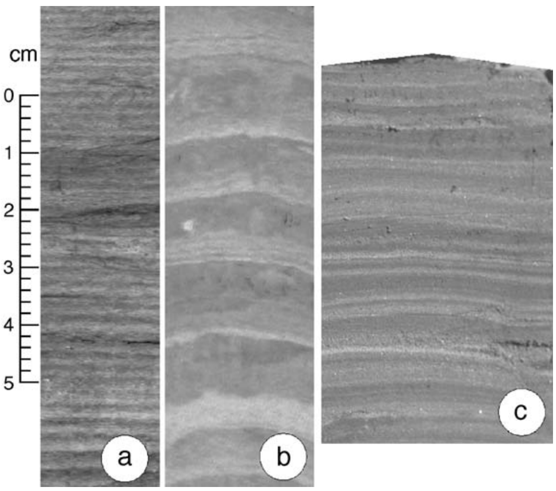

Fig. 8. Photographs of the cleaned surface of samples from Llewellyn Inlet: (a) Ekman core from near the point of inflow showing modern sediment as thin varves formed after the creation of a lake at the toe of Llewellyn Glacier; (b) distal gravity core showing thicker annual accumulation before the opening of the lake; and (c) Ekman core from Torres Channel showing laminations of silt and clay. For locations, see Fig. 2.

The earlier high accumulation rates indicate that Llewellyn Glacier was discharging large quantities of sediment directly into the inlet where turbidity and interflow currents distributed sediment throughout. As well, the rapid retreat of Llewellyn Glacier from its Little Ice Age maximum would have exposed additional sources of sediment, which would have contributed to energetic turbidity currents capable of travelling the length of the inlet (6.5 km) and depositing coarse silt on the outlet sill. This would have required bottom currents to move upslope from about 180-m water depths to at least 110 m. The abrupt and significant decline in sediment availability to the inlet occurred as a result of trapping and storage in the proglacial lake that opened at the toe of the NE margin of the glacier (Fig. 1) as has been reported in other glacial lakes (e.g. Smith, 1981).

In Willison Bay (Fig. 9b), the sediments consist of well-laminated silts (mean grain size 12.7 μm) and alternating with clay laminae about 5 mm thick through the deposit. The silts are consistent with deposition by weak underflows driven by short-term variations in meltwater and sediment delivery from the Juneau Ice Field via Hoboe and Willison Creeks. Although we have no independent dating evidence, the clay may be winter deposits. This suggests that present annual rates of accumulation are about one half of that during the earlier history of the lake based on acoustic evidence.

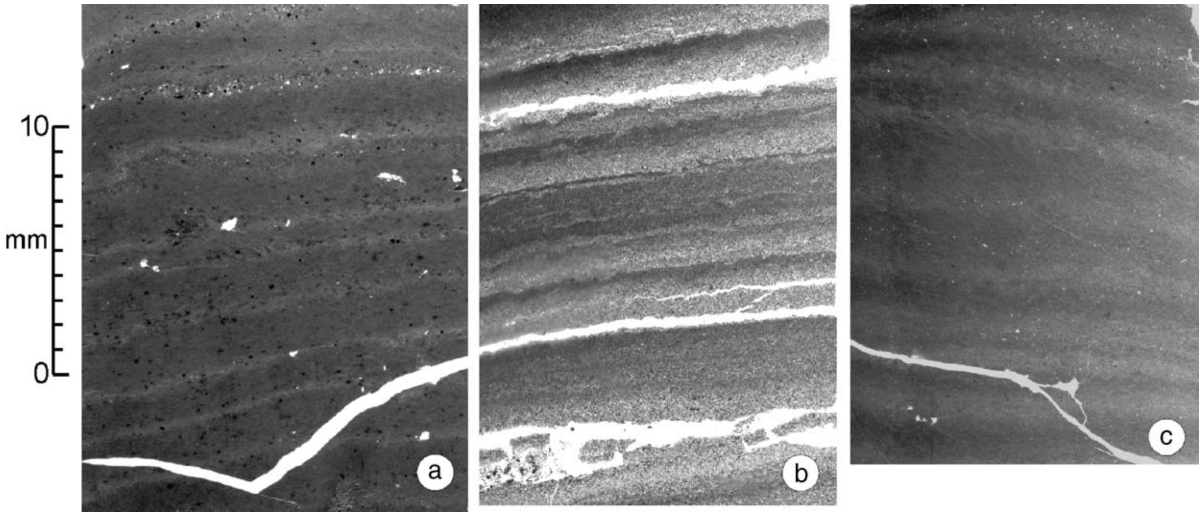

Fig. 9. Scanned thin sections from cores in (a) Llewellyn Inlet, (b) Willison Bay and (c) Torres Channel showing laminated sediments. All scans were obtained on a HP S20 scanner at 2400 dpi. Irregular white areas are cracks formed during drying before impregnation. For locations, see Fig. 2.

In Torres Channel northeast of the sill separating Willison Bay, the sediment is also well laminated (Fig. 8c), probably associated with interflows which irregularly reach this distance down-lake, and possibly occasional more powerful turbidity currents which are able to surmount the sill. However, this deposition is more irregular, and the recognition of annual deposits is not possible. In the more distal regions of Torres Channel, the sediment is uniformly fine textured (mean grain size 4.2 μm) with about equal proportions of silt and clay sizes. The sediment appears massive on visual examination, although in thin section (Fig. 9c), faint, 1–2-mm-thick laminae can be distinguished.

5. Discussion

Large lakes of the Canadian Cordillera contain in their sedimentary records important evidence of post-glacial environmental change as well as evidence to a sub-annual scale of modern aquatic processes and their hydroclimatic controls (Table 1). To compare these lakes with the entire set of lakes in British Columbia, see Schiefer and Klinkenberg (2004).

In Fig. 10, the large lakes in the Canadian Cordillera are divided into three groups: (1) those which contain only small amounts of sediment distributed as discontinuous deposits in the deepest parts of the lake, (2) those which received abundant sediment during late Pleistocene deglaciation but very much less during most of the Holocene and (3) those which contain thick deposits of sediment begun during deglaciation and continued through the Holocene. Of all the large lakes in the Cordillera studied to date, only Atlin, Chilko, Waterton and Quesnel lakes are in the first group, although the much smaller proximal basins of Willison Bay, Llewellyn Inlet and possibly parts of Sloko Inlet are in the third category. There are several reasons for the low rates of accumulation in group 1 lakes. First, these are among the largest lakes and so the sediment is distributed more widely around the lake floor. Note that Atlin and Chilko have the smallest ratios of drainage basin to lake area of the lakes reported in Table 1. But this argument is less convincing when one considers that the large lakes also have relatively large mountainous drainage basins (Table 1) and so should have received relatively larger influxes of sediment during their histories, especially given that larger basins may produce disproportionately greater yields (Church and Slaymaker, 1989). Second, the sediment is trapped before it reaches the lake. This clearly occurs in the proximal basins of Atlin Lake and starves the main lake of sediment. However, except for the proglacial lake formed within the Little Ice Age moraine of Llewellyn Glacier, which has reduced accumulation during the last half century by almost an order of magnitude, there are few lakes in the drainage basin of Atlin Lake (Fig. 1). This is also the case for Chilko Lake where there are few traps in the drainage basin (Desloges and Gilbert, 1994b), although both Atlin and Chilko lakes have large active glaciers in their basins which are currently producing abundant sediment.

The third and most probable reason these four lakes have small sediment accumulations is that they did not receive significant sediment input during deglaciation.

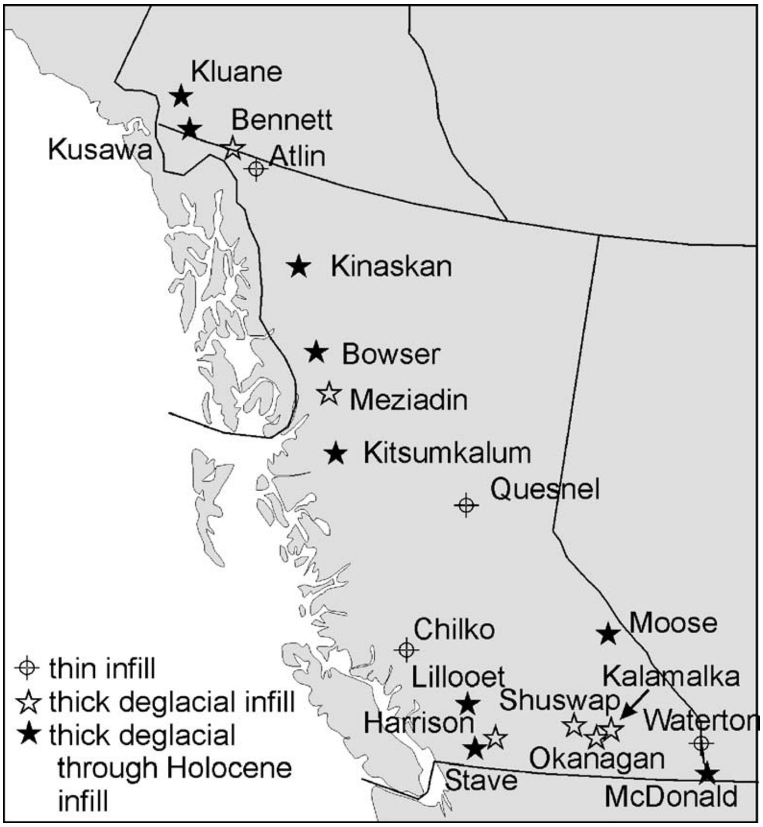

Fig. 10. Locations of large lakes in British Columbia and Yukon Territory referred to in Table 1. Symbols distinguish lakes which contain only thin deposits of sediment, those which received abundant sediment during late Pleistocene deglaciation but significantly less since then, and those which contain thick deposits of sediment begun during deglaciation and continued through the Holocene.

We postulate that the large trunk glaciers which occupied the lake basins (and almost certainly initially carved them by selective linear erosion—Sugden, 1978) removed most or all of the pre-existing soft sediment. But as the glaciers thinned and retreated, they stagnated (Fulton, 1991) so that erosion and sediment delivery to their termini was limited and, more importantly, they retreated very rapidly so that the present lake basin served as a glacier-proximal receiving basin for only short lengths of time. A number of workers (e.g. Souch, 1989) have commented on the apparent rapid disintegration of the Cordilleran ice about 11,000 BP, and Clague and Evans (1993) used the rapid modern retreat of large glaciers in the Yukon as an analogue for late-Pleistocene retreat of the Cordilleran ice sheet.

It is clear that most of the lakes with sediment fills of more than 100 m received much of that input during deglaciation, while large active glaciers terminating in or very near the lakes could have contributed as much as several metres per year of sediment over periods of time amounting up to several centuries (Johnsen and Brennand, 2004). Most of the large lakes, even those with a significant part of the drainage occupied by active glaciers, through the Holocene have deposited from a few millimetres to about 2 cm sediment/year. This is sufficient to accumulate from less than 5 m to a maximum of about 200 m, which, when focused to less than half the area of the lake floor, corresponds with amounts of sediment seen in the basins today (Fig. 4).

The acoustic facies recorded in Atlin Lake correspond to this interpretation. Proximally, there is only one facies because the sedimentary processes dominated by the presence of turbidity currents have continued uninterrupted with varying intensity since the lake was first created by deglaciation. These currents are still active in Willison Bay because there are no upstream traps between the glaciers and the lake. They were also active in Llewellyn Inlet until recently when the small lake in front of Llewellyn Glacier. The deposits are unconformable because turbidity currents flow to the deepest part of the lake basin, depositing the sediment preferentially in this region to form nearly horizontal strata, which mask the irregular topography below. These sediments are well stratified because successive turbidity currents deliver pulses of coarser sediment, which form distinct reflectors due to different acoustic impedance.

In the main body of Atlin Lake, the lower facies 1 represents a similar environment but one which existed only for a short period during deglaciation and so is thin and discontinuous. Facies 2 was deposited mainly from suspension in the water column (and so is conformable to the underlying surface) in an environment where sediment influx was limited throughout the Holocene.

6. Conclusions

Large montane lakes such as Atlin contain in their sedimentary records important environmental and paleoenvironmental evidence from the time of their origin following deglaciation to the present. As shown by the response to changes in upstream sediment traps, they are very responsive to glacial and geomorphic changes in their drainage basins and, as a result, to hydroclimatic forcing. The greatly different amounts of sediment accumulated in these basins during deglaciation are evidence of the glacial and climatic processes operating differently in various regions of the Cordillera. Lakes such as the main body of Atlin Lake, which contain relatively little of this sediment, provide evidence of stagnation of late Pleistocene ice, and especially of very rapid retreat from full glacial cover to glaciers that were at least as reduced as those at present.

The technique most suitable for the investigation of these large lakes combines acoustic or seismic survey with sediment recovery by coring, the former providing the lake-wide evidence of sedimentary processes, and the latter documenting in detail sedimentary processes and deposits at a scale that is commonly sub-annual. The acoustic data determine whether cores are representative of sedimentary conditions (Gilbert, 2003), while the sediments themselves confirm the interpretations from the remote sensing of acoustics. As well, although specialized rotary coring offers the prospect of recovering up to several hundred metres of sediment at great expense, conventional coring technology limits recovery to a few metres of sediment. Even where rates are low, this usually represents only a small portion of the total sediment accumulation.

As our data base of sedimentation in large lakes grows (Table 1), it is apparent that the value of their sediment as environmental proxy depends on better understanding of the processes that determine the production, delivery and deposition of sediment on the lake floor. Otherwise, the link between hydroclimatic control and sedimentary proxy will remain poorly understood and the relation between them is at best inferential, and at worst is misleading or wrong. Much of the research that Professor Slaymaker conducted and supervised was directed to discovering these processes, and sets the groundwork for continued productive study in a globally changing world.

Acknowledgments

This work was supported by research and equipment grants from the Natural Sciences and Engineering Research Council of Canada, the Canadian Foundation for Innovation and the Northern Scientific Training Grant Program. W. Junkin and S. Brooks assisted in the field, and M. Retelle kindly loaned CTD instrumentation in 1999. The Vig family of Atlin supplied the vessel used in some of the work. Thoughtful reviews by T. Fisher, N. Smith and A. Werner improved our work.

References

Brune, G.M., 1953. Trap efficiency of reservoirs. Transactions of the American Geophysical Union 34, 407–418.

Church, M., Slaymaker, O., 1989. Disequilibrium of Holocene sediment yield in glaciated British Columbia. Nature 337, 452–454.

Clague, J.J., Evans, S.G., 1993. Historic retreat of Grand Pacific and Melbern Glaciers, Saint Elias Mountains, Canada: an analogue for decay of the Cordilleran ice sheet at the end of the Pleistocene. Journal of Glaciology 39, 619–624.

Desloges, J.R., Gilbert, R., 1991. Sedimentary record of Harrison Lake: implications for deglaciation in southwestern British Columbia. Canadian Journal of Earth Sciences 28, 800–815.

Desloges, J.R., Gilbert, R., 1994a. Sediment source and hydroclimate inferences from glacial lake sediments: the post-glacial sedimentary record of Lillooet Lake, British Columbia. Journal of Hydrology 159, 375–393.

Desloges, J.R., Gilbert, R., 1994b. The record of extreme hydrological and geomorphological events inferred from glaciolacustrine sediments. In: Olive, L.J., Loughran, R.J., Kesby, J.A. (Eds.), Variability in Stream Erosion and Sediment Transport, International Association of Hydrological Sciences Publication No. 2, vol. 224, pp. 133–142.

Desloges, J.R., Gilbert, R., 1995. The sedimentary record of Moose Lake: implications for glacier activity in the Mount Robson area, British Columbia. Canadian Journal of Earth Sciences 32, 65–78.

Desloges, J.R., Gilbert, R., 1998. Sedimentation in Chilko Lake: a record of the geomorphic environment of the eastern Coast Mountains of British Columbia, Canada. Geomorphology 25, 75–91.

Eyles, N., Mullins, H.T., 1997. Seismic-stratigraphy of Shuswap Lake, British Columbia, Canada. Sedimentary Geology 109, 283–303.

Eyles, N., Mullins, H.T., Hine, A.C., 1990. Thick and fast: sedimentation in a Pleistocene Fiord Lake of British Columbia, Canada. Geology 18, 1153–1157.

Eyles, N., Mullins, H.T., Hine, A.C., 1991. The seismic stratigraphy of Okanagan Lake, British Columbia: a record of rapid deglaciation in a deep 'fjord-lake' basin. Sedimentary Geology 73, 13–42.

Eyles, N., Boyce, J.I., Halfman, J.D., Koseoglu, B., 2000. Seismic stratigraphy of Waterton Lake, a sediment-starved glaciation basin in the Rocky Mountains of Alberta, Canada and Montana, U.S.A. Sedimentary Geology 130, 283–311.

Fulton, R.J., 1991. A conceptual model for the growth and decay of the Cordilleran ice sheet. Géographie Physique et Quaternaire 45, 281–286.

Gilbert, R., 1975. Sedimentation in Lillooet Lake, British Columbia. Canadian Journal of Earth Sciences 12, 1697–1711.

Gilbert, R., 2003. Spatially irregular sedimentation in a small, morphologically complex lake: implications for paleoenvironmenal studies. Journal of Paleolimnology 29, 209–220.

Gilbert, R., Butler, R.D., 2004. The physical limnology and sedimentology of Meziadin Lake, northern British Columbia. Arctic Antarctic and Alpine Research 36, 33–41.

Gilbert, R., Desloges, J.R., 1992. The late Quaternary sedimentary record of Stave Lake, southwestern British Columbia. Canadian Journal of Earth Sciences 29, 1997–2006.

Gilbert, R., Desloges, J.R., Clague, J.J., 1997. The glacilacustrine sedimentary environment of Bowser Lake in the northern Coast Mountains of British Columbia, Canada. Journal of Paleolimnology 17, 331–346.

Johnsen, T.F., Brennand, T.A., 2004. Late glacial lakes in the Thompson Basin, British Columbia: paleogeography and evolution. Canadian Journal of Earth Sciences 41, 1367–1383.

Kuperman, W.A., Lynch, J.F., 2004. Shallow-Water Acoustics: Physics Today, October 2004, pp. 55–61.

Lamoureux, S.R., 2003. Impregnation. In: Middleton, G.V., Church, M.J., Coniglio, M., Hardie, L.A., Longstaff, F.J. (Eds.), Encyclopedia of Sediments and Sedimentary Rocks. Kluwer Academic Publishers, Dordrecht, pp. 374–375.

Last, W.M., Smol, J.P., 2001. Tracking Environmental Change Using Lake Sediments. Kluwer Academic Publishers, Dordrecht. Four volumes.

Mullins, H.T., Eyles, N., Hinchey, E.J., 1990. Seismic reflection investigation of Kalamalka Lake: a 'fiord lake' on the Interior Plateau of southern British Columbia. Canadian Journal of Earth Sciences 27, 1225–1235.

Mullins, H.T., Eyles, N., Hinchey, E.J., 1991. High-resolution seismic stratigraphy of Lake McDonald, Glacier National Park, Montana, U.S.A. Arctic and Alpine Research 23, 311–319.

Schiefer, E., Klinkenberg, B., 2004. The distribution and morphology of lakes and reservoirs in British Columbia: a provincial inventory. Canadian Geographer 48, 345–355.

Serink, A.L., 2004. Glacilacustine sedimentation processes at Llewellyn Inlet, Atlin Lake, British Columbia: implications for Little Ice Age Climate Change. MSc thesis: Toronto, Canada, University of Toronto.

Smith, N.D., 1981. The effect of changing sediment supply on sedimentation in a glacier-fed lake. Arctic and Alpine Research 13, 75–82.

Souch, C., 1989. New radiocarbon dates for early deglaciation from the southeastern Coast Mountains of British Columbia. Canadian Journal of Earth Sciences 26, 2169–2171.

Sprenke, K.F., Miller, M.M., Mcgee, S.R., Adema, G.W., Lang, M., 1999. Canadian landform examples: 36. The high ice plateau of the Juneau Icefield, British Columbia: form and dynamics. Canadian Geographer 43, 99–104.

Sugden, D.E., 1978. Glacial erosion by the Laurentide ice sheet. Journal of Glaciology 20, 367–391.