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The record of an extreme flood in the sediments of montane Lillooet Lake, British Columbia: implications for paleoenvironmental assessment

Robert Gilbert[1,*], Sarah Crookshanks[1], Kyle R. Hodder[1], John Spagnol[2] and Roland B. Stull[2]
*[1]Department of Geography, Queen's University, Kingston, ON, Canada K7L 3N6; [2]Department of Earth and Ocean Sciences, University of British Columbia, Vancouver, BC, Canada V6T 1Z4; *Author for correspondence (e-mail: robert.gilbert@queensu.ca)*

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Abstract

Severe rainfall in mid October, 2003 produced the largest floods in almost a century of record on rivers in the Cordillera of southwestern British Columbia. Sediment deposited in Lillooet Lake as a result of this event is clearly distinguished by stratigraphy, colour, texture, magnetic properties, and organic content. Each of these physical properties is related to the lacustrine processes, especially turbid underflow, that distributed the sediment through the lake. The flood, which lasted less than a week, delivered 8–12 times the amount of sediment that accumulates in most entire years in the deepest, central parts of the lake. Recognition of events of this type in the stratigraphic record offers a means of assessing the changing nature of extreme hydroclimatic events, and their relation to more ubiquitous, lower-energy processes.

Introduction

The physical properties of sediment deposited in glacial lakes provide proxy for environmental change, including precipitation (Desloges and Gilbert 1994), temperature (Leonard 1986; Desloges 1994) and hydrology (Gilbert 1975; Menounos et al. in press). These proxies extend in space and time the short-term instrumental records, and also provide assessment of geomorphic, glacial and geologic processes. However, the value of lacustrine proxy is realised only if the process cascade between the controls in and beyond the drainage basin, through to those in the lake itself is acknowledged, and, ideally, understood. Without causal determination, the relation becomes one of descriptive statistical analysis, which is at best limited, and at worst, spurious.

Important aspects of using proxy include recognising extreme events and understanding their impact on the Earth system compared to normal or average processes (Page et al. 1994; Thorndycraft et al. 1998; Lamoureux et al. 2001; Nesje et al. 2001; Noren et al. 2002). This concept of magnitude and frequency has a long history in geomorphic research (Wolman and Miller 1960), and is especially relevant now as the impact of human-induced global change increases (McCarthy et al. 2001). Changing temperature and precipitation effect changes in earth-surface processes, but the greatest impact may be associated with larger and more frequent extremes in relation to energy in the atmosphere-earth system.

Extreme precipitation in the southwestern mountains of British Columbia during October 2003 generated the largest floods ever recorded in several rivers. The lacustrine sedimentary record of these floods provides an opportunity to assess their geomorphic importance, and to establish criteria for their recognition in the long-term lacustrine record of environmental change.

Study site

The drainage basin of Lillooet Lake is located in the Coast Range Mountains of southwestern British Columbia (Figure 1). Elevations range from 195 m a.s.l. at the lake to over 3000 m in the alpine zone. Large amounts of sediment have been produced (Slaymaker 1993) by glaciers in the drainage basins of Lillooet and Green rivers, and especially associated with volcanism and related colluvial processes during the Holocene (Bovis and Jakob 2000; Friele and Clague 2004; Friele et al. 2005). The delta of Lillooet River has advanced more than 50 km during the Holocene (Friele et al. 2005), and continues with modern rates of 5–20 m/a (Gilbert 1975; Jordan and Slaymaker 1991). Sediments are deposited in Lillooet Lake as varves (Gilbert 2003) that document accumulation on the lake floor of 5–100 mm/a, and provide useful proxy of the hydroclimate of the drainage basin (Gilbert 1975; Desloges and Gilbert 1994).

The discharge of Lillooet River is 65% of the total flow monitored in the three drainage basins tributary to the head of the lake shown in Figure 1. The river exhibits a glacial flow regime (Church and Gilbert 1975) with mean annual discharge of 126 m^3/s, a nival melt peak and a glacial melt peak, typically of 400–600 m^3/s occurring in June and July, respectively. Discharge decreases from August to November and is normally less than 50 m^3/s until April. However, in some years intense tropical Pacific storms create extreme precipitation in western North America (Higgins et al. 2000). The warm rain melts snow at high elevations, augmenting runoff and generating large floods. These autumn events have recently become larger and more common. The recurrence series for Lillooet River discharge (Figure 2) shows that, of the ten largest floods, eight occurred in autumn, and seven occurred in the period from 1975 to 2003. During the previous 56 years of record, in only 9 years did the maximum discharge occur in autumn. This may correlate with the well-documented shift in the Pacific Decadal Oscillation from a cold to a warm phase in 1976 (Mantua et al. 1997). This shift has also been recorded in the change to persistently negative mass balance since 1976 of Place Glacier located in the Birkenhead River basin (Moore and Demuth 2001).

The October 2003 event

Output from the Mesoscale Compressible Community (MC2) numerical weather prediction model, developed by the Numerical Prediction Research branch of the Canadian Meteorological Centre (Laprise et al. 1997), and run daily at the University of British Columbia is presented in Figure 3. It shows the pattern of heavy precipitation that began before midnight on October 15 in southwestern British Columbia resulting from a strong subtropical jet which advected copious amounts of moisture from near Hawaii to British Columbia. Orographic precipitation exceeded 200 mm per day in the mountains of Vancouver Island and the mainland north and east of Vancouver during the tropical storm. On the coast, the event ended on October 18 and coastal stations experienced the incursion of a separate, normal Pacific front on October 19 but in inland areas such as the Lillooet River valley, the tropical storm merged with this event so that by noon on October 19, 313 mm of rain had fallen at Brandywine, 199 mm at Whistler, 159 mm at Cayoosh Creek, and 44 mm at Carpenter Lake. This illustrates the rain shadow created by the Coast Mountains. The modeled results (Figure 3) correspond well with the measured values at a station (Figure 4b). Minor precipitation events occurred on October 20 and 22 during two additional normal frontal storms from the Pacific.

Air temperature increased during and following the tropical storm and a number of records were broken in southwestern British Columbia. At Whistler near the headwaters of Green River (Figure 1) mean daily temperature rose to 6.4 °C on October 21 (Figure 4a). At Brandywine and Carpenter Lake, temperature exceeded 12°C. At Cayoosh Creek, the highest temperature of 10.2°C occurred at noon on October 22. A brief period of mixed rain and snow occurred between about 04:00 and 08:00 on October 16 at Whistler (snow having 1 mm water equivalent). The MC2 model results suggest that, even at high elevations, the

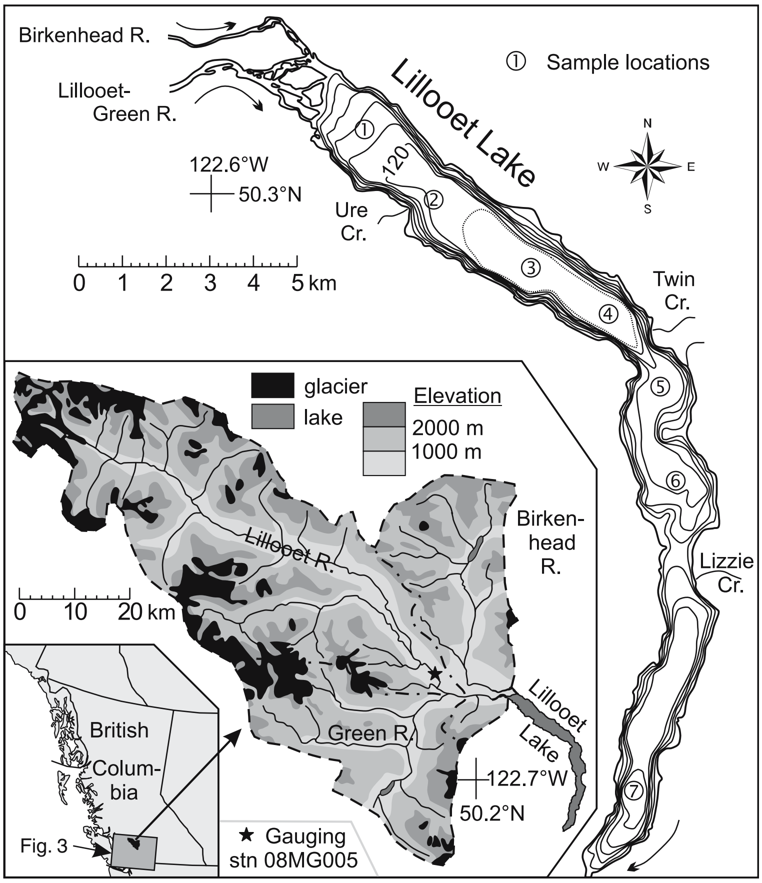

Figure 1. Bathymetry of Lillooet Lake below mean summer level (isobath interval is 20 m; the 130 m isobath is shown as a dashed line) modified from Gilbert (1975) and Desloges and Gilbert (1994). Inset map shows the drainage basins tributary to the head of Lillooet Lake: Birkenhead River drains 680 km^2 of which 0.7% is glacier covered; Lillooet River drains 2240 km^2 of which 12.7% is glacier covered, and Green River drains 823 km^2 of which 10.5% is glacier covered.

rest of the precipitation fell as rain. These temperatures and the warm rain caused widespread melting of the snow already accumulated at high elevations, adding to the runoff.

Discharge in Lillooet River rose rapidly from 60 m^3/s on October 15 to a maximum mean daily discharge of 1370 m^3/s on October 19. Maximum measured instantaneous discharge was 1490 m^3/s, just after midnight (Figure 4c). Secondary peaks occurred on October 21 and 23 in rapid response to the smaller precipitation events on those days, but by October 27, discharge decreased to 200 m^3/s and continued to lessen to normal autumn values thereafter.

Inflow to Lillooet Lake caused the lake level to rise 5.05 m between October 16 and 19 (Figure 4c), about 3.3 m over normal maximum summer levels. The water fell to the preflood level by November 6 but the backwater created by this rise, in combination with the large river discharge, caused flooding to more than 2 m above bank full stage in parts of the lower 14 km of the Lillooet River valley. Sediment discharge data are unavailable, although much smaller floods have produced suspended sediment concentrations in Lillooet River >4 g/l (Gilbert 1975).

Methods

Between June 7 and August 3, 2004, Ekman box cores were taken from a number of locations in Lillooet Lake (Figures 1 and 5). Careful extraction of subsample cores preserved the sediment–water interface. The split cores were logged and photographed when dried to best show sedimentary

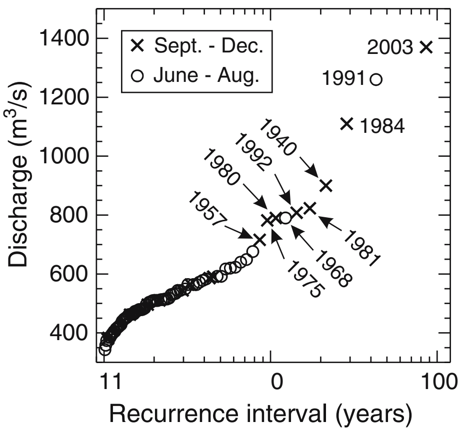

Figure 2. Recurrence series of annual maxima of mean daily discharge in Lillooet River. The period of record is 1914–2003, except 1919–1922. The 10 largest events are indicated by the year in which they occurred. The gauging station is located 14 km upstream from Lillooet Lake (Figure 1). Source: Water Survey of Canada, station 08MG005.

structures. Their sediment was analysed for loss on ignition at 550 °C (LOI) to assess organic content to ± 0.1% (Heiri et al. 2001; Santisteban et al. 2004), magnetic susceptibility (MS) using a Bartington MS2 instrument, and particle size of deflocculated mineral sediment using a Coulter LS 200 particle-size analyzer (the principle of which is described by Beuselinck 1998). Replicate cores from a number of the sites were examined to confirm that the analysed cores represented deposition at their locations.

Results and interpretation

The 10 cores illustrated in Figure 5 document the sedimentary deposits from the 2003 flood in Lillooet Lake. Distal cores (sites 4–7) show that, until the flood, sediment deposition in 2003 was typical of other years. The peak discharge of 560 m^3/s in late June was not unusual (Figure 2), and flow fluctuated and decreased until October. These fluctuations produced sediment pulses that lead to minor laminae best seen at sites 3 and 4, as a result of turbidity current flow as reported previously here (Gilbert 1975, 2003) and in other glacial lakes (Lambert and Hsü 1979; Blass et al. 2003). Sediment consists of about one third to one half clay size and the rest silt; mean grain size less

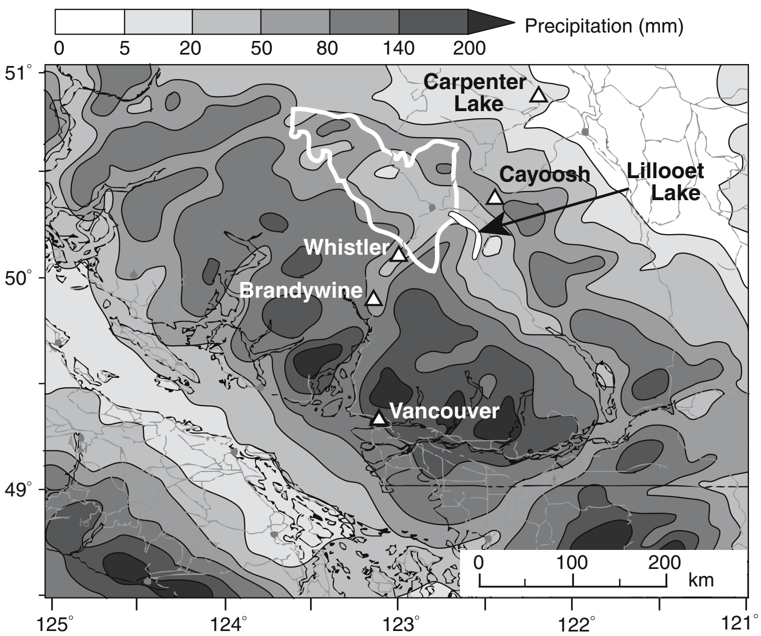

Figure 3. Precipitation during 24 h from 04:00 Pacific Standard Time October 16, 2003 determined from the Mesoscale Compressible Community (MC2) using a 4-km grid (Laprise et al. 1997) produced by the Geophysical Disaster Computational Fluid Dynamics Center at the University of British Columbia. The drainage basin to the head of Lillooet Lake (Figure 1) is outlined in white and the climate stations shown on Figure 4 are marked.

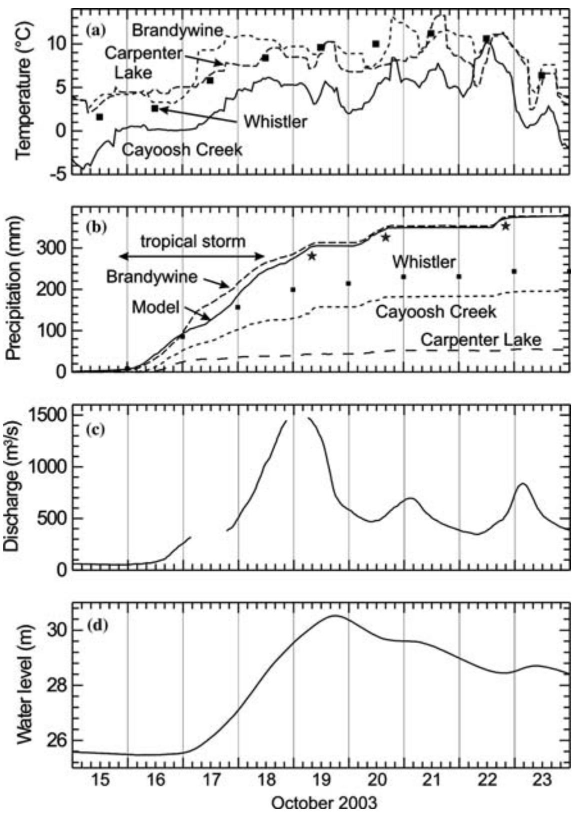

Figure 4. (a) Mean hourly air temperature at Brandywine (395 m a.s.l.), Cayoosh Creek (1350 m a.s.l.) and Carpenter Lake (660 m a.s.l.) (source: British Columbia Ministry of Transportation) and mean daily air temperature at Whistler (658 m a.s.l.) (source: Meteorological Service of Canada). (b) Hourly cumulative precipitation modeled in the Coast Mountains near Brandywine (based on the MC2 model), and measured at Brandywine, Carpenter Lake, and Cayoosh Creek (source: British Columbia Ministry of Transportation) and daily cumulative precipitation measured at Whistler (source: Meteorological Service of Canada) during the tropical storm and subsequent Pacific frontal storms. Stars indicate the Pacific frontal storms that followed the tropical storm. (c) Mean hourly discharge of Lillooet River (Water Survey of Canada, station 08MG005). For location of the recording stations, see Figures 1 and 3. (d) Mean hourly level of Lillooet Lake recorded near the inflow of Twin Creeks (Figure 1) (Water Survey of Canada, station 08MG020).

than 10 μm. Sand comprises less than 1%, except within about 3 km from the inflow.

At site 1 near the base of the foreset deposits of the delta (1 km from the point of inflow, 85 m water depth, slope 2.9°), the flood deposits occur as a 1.7-cm thick layer of more than 80% sand with almost no clay size (Figure 5). Mean grain size exceeds 100 μm and the deposit is very well sorted with no grains coarser than 700 μm. Cross-bedding attests to vigorous turbidity current flow and is one of the few examples of Division C of the Bouma sequence (Shanmugam 1997) reported in a glaciolacustrine setting, others being restricted to Divisions D and E. Observations from an aircraft during the flood (B. Menounos, personal communication, October 2003) document plunging inflow at the river mouth and the presence of a prominent debris line, both evidence of the presence of a strong, persistent turbidity current (Gould 1951) as occurs commonly in Lillooet Lake; Gilbert 1975; Best et al. 2005). The flood deposit at site 1 is considerably thinner than at more distal sites because (1) the turbidity currents were strong enough to transport sediment beyond the site, and (2) the currents may have meandered laterally over the lake floor during their passage down the steeper slopes, but spread more uniformly as they slowed distally. The first has been reported in several glacial lakes (Gilbert and Desloges 1987; Gilbert et al. 1997), including Lillooet Lake (Gilbert 1975), where accumulation is greater in the distal regions of the main basin. The second was proposed as the mechanism accounting for periodic decreases in the measured velocity of turbidity currents (Gilbert and Shaw 1981).

Both the top and bottom of the flood deposit are very sharply distinguished from the sediments above and below due to the sudden onset of the flood, and because, after the flood passed, little deposition occurred subsequently in 2003. The slope in this region is sufficient that the turbid cloud that commonly remains following the current continued to move toward the deeper, more distal basin at velocity sufficient to prevent deposition of its fine-grained sediment at site 1. The winter clay cap occurs about 5 mm above the sand layer, and the deposition above represents much of the accumulation in 2004. After the flood, the lake level and associated backwater in Lillooet River decreased more slowly than the discharge in Lillooet River, so sediment would have been stored in the lower reach of the river and in the delta, not to be flushed out until discharge subsequently rose in 2004. Thus, the 2004 deposit is probably thicker than would be predicted from hydroclimatic control. This lag has significant implications for the application of varve thickness as hydroclimatic proxy.

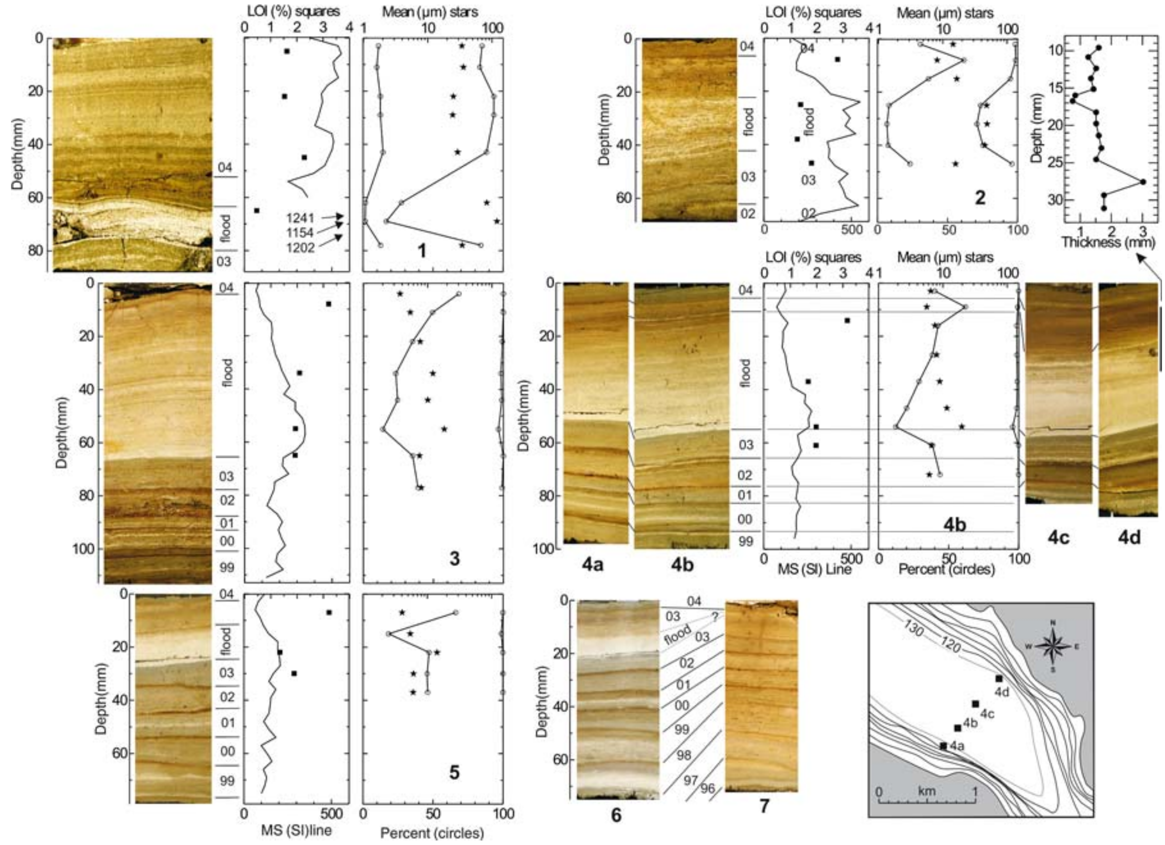

Figure 5. Ekman box cores recovered in summer 2004 shown all at the same scale, and keyed to locations in the lake (Figure 1 and detail of site 4 on the inset map, lower right). Accompanying graphs illustrate loss on ignition (LOI) at 550 °C, magnetic susceptibility (MS) at 3-mm increments, geometric mean particle size, and the per cent sand (>62 μm), silt, and clay size (<4 μm). Graphs are drawn to the same scale. Text to the right of the photographs indicates the year of deposition from 1996 to 2004, and the 2003 flood deposit. Graph in the upper right shows the thickness of rhythmites near the top of the flood deposit in core 4d.

There is no evidence of erosion of the substrate by the turbidity current, although the sample size is too small to make this conclusion definitively. Turbidity current channels have not been found in Lillooet Lake using acoustic mapping, although they have been reported in a few other glacial lakes (Irwin 1980; Lambert and Giovanoli 1988).

The organic content as indicated by LOI of 0.7% is significantly lower in the flood sediment than in other deposits where it ranges from 2 to $>3\%$. This may reflect sediment dominantly from glacial sources including the ice itself and the glacial forefield, but more likely it represents transport beyond this site of the low-density organic matter by the turbidity current (see below). The MS of the flood deposit at site 1 is greater than in other deposits in the lake, including distal flood deposits, because the sands of the lower Lillooet River have a high content of magnetically susceptible minerals, including hematite in sufficient quantities to warrant consideration of placer mining.

Site 2 is located 3 km from the inflow at 123 m depth where a gravel fan built by Ure Creek constricts the lake (Figure 1), and decreases the slope of the lake floor to zero. About 2 cm of silty sand have accumulated due to the flood, with sand making up to 30% and clay size less than 10%. In comparison to the deposit at site 1, the deposit here is thicker, finer textured, horizontally laminated, and contains double the organic content. MS is less than half, reflecting the lower content of heavy minerals and sand. Notable is the accumulation of 15 mm of fine (mean size 8 μm), more organic-rich (2.8%) sediment on top before deposition of the clay cap. As turbid underflow slowed on the flat floor, finer particles were progressively deposited, and the turbid cloud that follows underflow remained in this area until its sediment settled from suspension. Less deposition is recorded in 2004 because the core was taken on June 23, prior to peak inflow.

Sites 3 and 4 in the deepest, flat-floored portion of the lake basin between 6 and 8.5 km from the inflow have the thickest flood deposits, with 48 and 60 mm, respectively. This accumulation, driven by an event lasting only a few days, is 8–12 times thicker than the total normal annual deposit in this region (Gilbert 1975). The flood deposits consist of a graded bed with 3–4% sand at the base, decreasing upward to 0% at the top, with 12% clay size at the base increasing to 40–60% near the top. MS is only slightly higher than in normal deposits, reflecting the small sand content. The turbidity current reaching this area had already deposited most sand from suspension, but still carried a large volume of finer sediment to be deposited after the current weakened and ceased. The lingering turbid cloud and sediment delivered from throughout the water column also contributed to deposition in this region. There is little evidence of lateral variation in the flood deposit at site 4 (Figure 5) because the lake floor is relatively narrow at this location and there is limited opportunity for the influence of the Coriolis effect.

Lamination within the distal flood deposits may relate to pulses in a single turbidity current (Best et al. 2005) lasting much of the period of the flood, or to separate currents generated by the minor discharge peaks on October 21 and 23 (Figure 4c), or to wandering of the current on the featureless, flat lake bottom in this region (cf. Gilbert and Shaw 1981). Previous currents during smaller floods have been shown to last several days (Gilbert 1975), and it is probable that the 2003 current lasted most of the duration of the flood. Nevertheless, the underflow would have been significantly modified as it travelled the 8.5 km from the inflow source, and individual pulses would have merged in the 18–36 h required for this passage assuming a velocity of 0.3 m/s on the foreset slope of the delta (Gilbert and Shaw 1981; Weirich 1986; Best et al. 2005) decreasing to 0.1 m/s or less on the flat floor of the lake. Thus, it is not possible to relate specific lamina to characteristics of the inflow. However, near the top of the flood deposit in cores from site 4, most notably 4d (Figure 5) about 15 rhythmites averaging 1.5 mm thick can be recognised. These may represent subtle fluctuations in sediment input during the waning stage of the flood.

Turbid underflows have been known to rise over a sill on the floor of a glacial lake due to momentum (Chikita et al. 1996), and turbid water from underflows in Lake Mead rises to the surface of the reservoir (Fischer and Smith 1983). Based on sand content of 1.1% and the lower organic content in the flood deposit at site 5, it is probable that the turbidity current in Lillooet Lake reached this point, almost 50 m above the deepest basin in the lake. Alternatively, some of the sediment deposited here may have originated from the small basins of Twin Creeks (Figure 1) which also experienced flooding. However, the flood deposit thickness of about 10 mm, and the texture and MS are similar to the normally deposited material, and there is little except the colour (derived from somewhat higher silt content and lower organic content) to distinguish normal from flood deposits.

At site 6, 12.2 km from the inflow the flood deposits are still recognised, but in the distal basin 20 km from the inflow and separated by a sill at 28 m depth at Lizzie Creek, there is little recognisable record of the flood in sediments. It is possible that these autumn flood events are of too short duration for a significant quantity of the sediment to be spread to the most distal parts of the lake. The rapid decline to normal autumn values of discharge following the flood (Figure 4) may also be a factor in the pattern of sedimentation, in that the main flux of sediment may not be moved downlake by the subsequent inflow as effectively as following summer floods when the inflowing stream carries greater discharge after a flood.

Conclusions and significance

The extreme, but short-lived flood of October 2003 deposited as much as 8–12 times more sediment in the central portion of Lillooet Lake than in most years without autumn floods. Even considering that some of sediment may have originated from short-term storage, for example, in the riverbanks and bars, this observation indicates the importance of rare, basin-wide, catastrophic events driven by hydroclimatic controls. As these events became larger and more frequent in the late 20th century (Figure 2), geomorphic processes, their products, and the impact on human occupance of montane regions may continue to change dramatically.

The colour, stratigraphy, texture, magnetic properties, and organic content all clearly distinguish the flood deposit from other deposits, and should make it relatively easy to recognise earlier floods, including those that occurred before the instrumental record. For example, the 1957 flood produced a deposit similar to that at site 1 and an outlier in the correlation between total annual inflow and varve thickness (Gilbert 1975). However, it is important to recognise that, even in a lake as morphologically simple as Lillooet, the character of the deposit varies greatly depending on location. In some places, the flood record was diminished (site 6) or absent (site 7), and it is possible to miss the significance of major events with an inappropriate sampling strategy.

Thus, recognition of catastrophic events, and assessment of their relative importance, almost certainly depends on (1) replicate sampling, stratified by depositional environment, and (2) understanding of the sedimentary processes that produced and delivered the sediment and created the deposits. In this study, the process aspect is incomplete but it can be inferred from assessment of the magnitude of the climatic and hydrologic event, and from past knowledge of the physical limnology of Lillooet Lake.

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