GLACIER OUTBURST FLOOD ON THE NOEICK RIVER: THE DRAINING OF APE LAKE, BRITISH COLUMBIA, OCTOBER 20, 1984

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PL 1A. Ape Lake basin before Jökulhlaup (photo taken August, 1984). The extent of Fyles Glacier near the turn of the century is shown by a trimline on the right. Photo: J. Desloges

PL 1B. Ape Lake basin two days after the October 20, 1984 outflow. Emergent climax moraines of Ape Glacier (left) and Fyles Glacier (centre) divide the half-emptied lake into three sub-basins. Water levels in the large east basin (f.g.) have been lowered by 20m. Photo: R. Lenci

ABSTRACT

Ape Lake (surface area 2.47 km^2, volume 85.5 x 10^6 m^3, average depth 31.7 m) is located on the eastern side of the Coast Mountains about 390 km northwest of Vancouver, British Columbia. Ape Lake lies at the head of the Noeick River which drains into South Bentinck Arm and the Pacific Ocean. The lake was formed by the Late Neoglacial advance of Fyles Glacier across the Noeick valley, thereby blocking its drainage and forming a lake which spilled over a divide (1395 m a.s.l.) into the Talchako-Bella Coola river system.

Since about 1900, Fyles Glacier has receded between 7.1 and 50.2 m/year while the thickness of ice in the glacier snout has decreased between 1.0 and 3.8 m/year. Between 1964 and 1978, when surrounding glaciers were advancing, Fyles Glacier receded at about 15 m/year while ice thickness decreased 2.2 to 3.8 m/year along the lake margin.

With the continued recession and ablation of Fyles Glacier, the maximum thickness of ice impounding Ape Lake decreased until it was roughly 10-20 percent greater than the maximum depth of water near the ice margin. This condition appears to have been reached in late 1984 when the hydrostatic pressure of water in the lake was sufficient to lift the ice allowing water to seep beneath Fyles Glacier.

About October 19, 1984, subglacial leakage from Ape Lake began to increase rapidly, melting a tunnel beneath Fyles Glacier. As the tunnel enlarged through melting, increasing volumes of water drained from Ape Lake on October 20, creating a large flood on the Noeick River downstream to tidewater. When lake levels had declined about 10 m, a climax end-moraine was exposed which divided the remaining water into two basins. The

west basin continued to drain with peak discharges at the tunnel exit occurring about 1100-1200 hours on October 20. Several hours later the climax end-moraine impounding water in the east basin failed, generating a second flood which peaked at the tunnel exit around 1630-1730 hours. In total, approximately 45.8 $\times 10^6$ m^3, or 54 % of the total lake volume drained beneath Fyles Glacier with maximum peak discharges at the tunnel exit estimated to be between 985 and 1500 m^3/s.

Downstream of Fyles Glacier, the flood destroyed several km of logging road, severely damaged two bridges, razed over 200,000 recently planted trees and probably destroyed a large proportion of the salmon eggs deposited in the river gravel earlier in the year.

The tunnel beneath Fyles Glacier is expected to close within 100-200 days. Once the tunnel is resealed, possibly as early as May, 1985, the lake will refill. The lake could be at or near maximum water level in late August or early September, 1985 after which it could drain at anytime. The lake is expected to drain subglacially, refilling and draining every 1-2 years. Major floods will continue until the thickness of ice damming the valley is 20-30 m lower than ice thickness in 1984. Thus, depending upon the rate of recession and ablation of Fyles Glacier and climatic conditions, major floods can be expected for at least 8 years, probably 14 years and possibly as long as 100 years.

Management options to eliminate or reduce the size of future outburst floods are limited to either excavating a trench in bedrock at the east end of the lake to reduce lake levels or increasing glacier ablation rates through application of cinder or coal dust. It is recommended that no trees be replanted in those areas which are less than 5 m above the river channel and no salmonid enhancement projects should be undertaken until the potential for major outburst floods is reduced.

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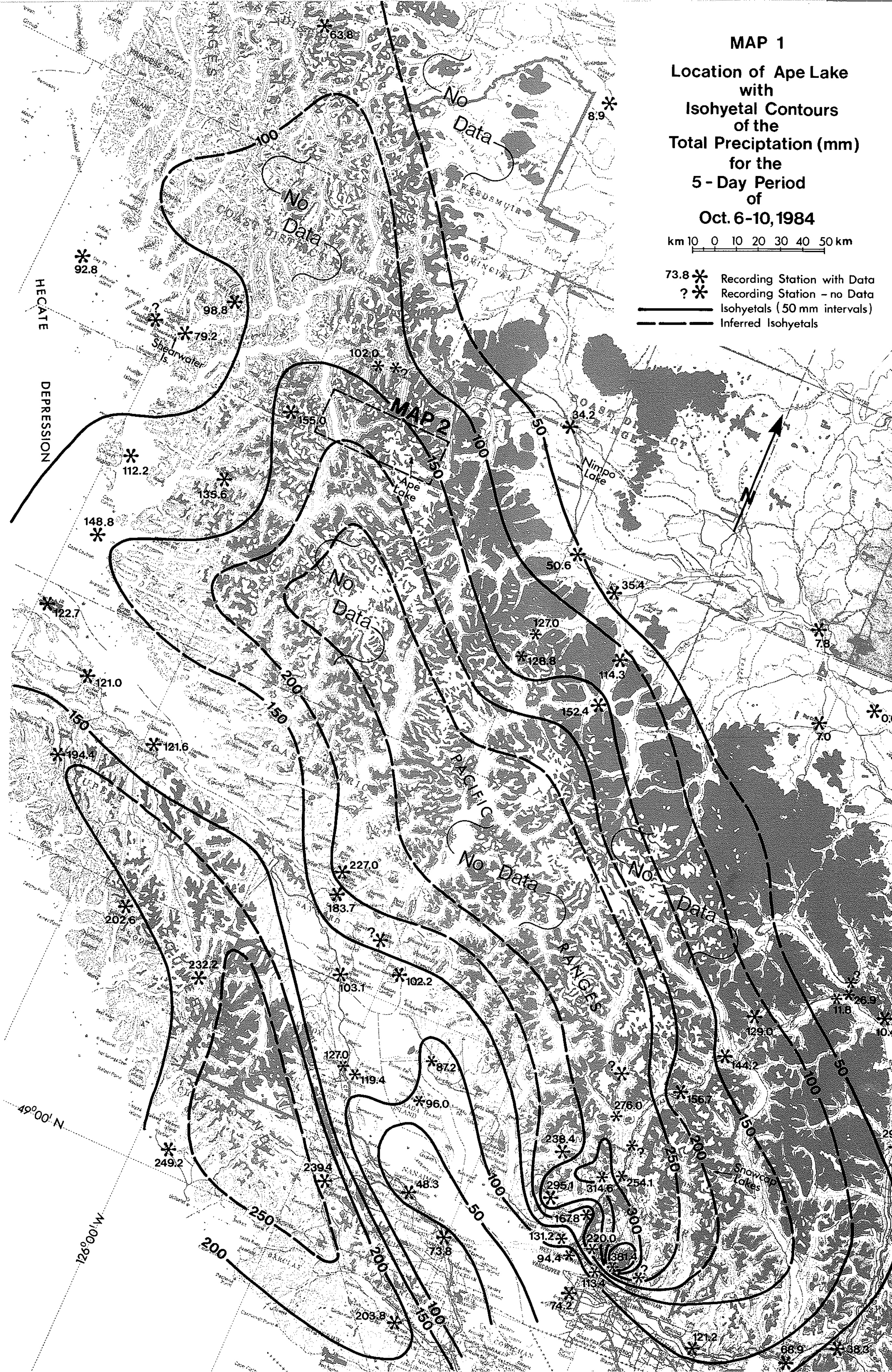
1.0 INTRODUCTION

On October 20, 1984, Ape Lake, located 50 km SE of Bella Coola, British Columbia, quickly drained beneath the snout of Fyles Glacier. The sudden release of about 45.8 million m^3 of water formerly stored in the lake created an unusually large flood on the Noeick River. The flood resulted in extensive erosion, transport and deposition of sediments stored on and adjacent to the floodplain of the Noeick River. The flood also destroyed parts of a newly constructed forestry access road, spur roads, tree plantations and affected the productivity of the local fisheries on the Noeick River downstream of the lake. Debris from the flood covered South Bentinck Arm, affecting access by water for several days, before it dispersed into nearby channels where it continued to be a hazard to boats.

This report contains a compilation of information on Ape Lake and the surrounding area prior to the event, eyewitness accounts of the flood, results of a limited aerial reconnaissance and ground survey after the lake outflow and a preliminary reconstruction of the sequence of events immediately prior to and during the draining of the lake. The report also contains a brief discussion on the likelihood of future floods, possible management alternatives and provides recommendations for future studies.

1.1 Background

Ape Lake lies on the drainage divide between the Noeick and Talchako-Bella Coola watersheds on the eastern side of the axis of the Coast Mountains of British Columbia (Map 1). Prior to the advance of glaciers during the Neoglaciation, the lake basin was probably much smaller, if it was a closed basin at all, and consisted of parts of the headwaters of the Noeick River. With Neoglacial ice advance and specifically, the Late


MAP 1
Location of Ape Lake with Isohyetal Contours of the Total Precipitation (mm) for the 5 - Day Period of Oct. 6-10, 1984
km 10 0 10 20 30 40 50 km
73.8 Recording Station with Data
? Recording Station - no Data
Isohyetals (50 mm intervals)
Inferred Isohyetals
HECATE DEPRESSION
PACIFIC RANGES
MAP 2
Ape Lake
Nimpo Lake
Snowcap Lakes
Shearwater Is
No Data
49°00' N
126°00' W

Neoglacial Advance (Little Ice Age), several glaciers became larger and descended onto the lower valley floor of the Noeick River. In particular, the Fyles, Ape and Noeick glacier complex extended beyond the valley axis, joining Atavist Glacier flowing from the north (Map 1). The resulting transection valley glacier system effectively blocked the drainage of the Ape Lake basin to the Noeick River. Local drainage was backed up, continually enlarging a basin until the lake level reached the height of the divide between the Noeick and Talchako systems (El. _ca_ 1395 m a.s.l). The drainage then spilled to the east via Ape Creek into the Talchako River, a large tributary of the Bella Coola River.

During the period when the Fyles Glacier complex was expanding, the newly formed lake may have emptied periodically but over the last few centuries, while the glacier was in the penultimate and ultimate climax position, the lake probably maintained a constant full pool elevation (_ca_ 1395 m). Large, mature conifers several centuries in age, which have recently been harvested, grew on the floodplain of the Noeick River 20-30 kilometres downstream from the glacially formed lake. The growth of an extensive, mature Sitka Spruce- Western Red Cedar forest on the floodplain suggests that catastrophic drainage such as occurred during the jökulhlaup of October, 1984, seldom, if ever occurred during the last several hundred years.

Since its discovery in 1951, Ape Lake has become a well known and highly popular area for climbing parties, artists and sightseers. Early map makers, who lacked the benefit of aerial photographs, were not aware of Ape Lake. The first recorded observation of the lake was made in 1951 when J. Dudra and P. Schoening saw the lake from the summit of Mt. Saugstad (El. 2908m), some 28 km distant (Dudra, 1952). The former observer recognized the lake as a possible staging point for mountaineering activities on the otherwise inaccessible Monarch Icefield to the south, and immediately began a compilation of a map of the area based on the then recently flown aerial photographs. The lake was positively identified as ice dammed at this time and the map was published following a successful expedition to the area in 1953, which apparently employed the first float plane landing on the lake (Dudra, 1954).

Dudra's account created wide interest and soon after Ape Lake became a favorite staging area for mountaineers, thanks to inexpensive float plane access from either Bella Coola (50 km to the NW) or Nimpo Lake on the Chilcotin Plateau (74 km to the ENE. Numerous camps have been held at the lake by mountaineering clubs since 1964. Commercially guided groups and alpine tours based as far away as California visit the area on an annual basis. In addition, small commercial float plane companies have taken tourists to the lake for general sightseeing. Currently there is one permanent building on the lake shore within registered District Lot No. 95 and there have been repeated suggestions to have the area included in Tweedsmuir Provincial Park located 20 km to the east.

The potential for glacier outburst floods originating from Ape Lake was first recognized by K. Ricker in 1973 while camped on its north shore near the snout of Fyles Glacier. In 1978 he visited an analogous area in southern Garibaldi Park, known as Snowcap Lakes (Map 1), where one or more glacier outburst floods had occurred in the 1930's following the recession of Thundercap Glacier which had formerly occupied the western lake basin at the turn of the century. Ricker (1979) published an interpretation of glacial chronology of the Snowcap Lakes area based on a fairly detailed qualitative analysis of the geomorphology and drew comparisons to Ape Lake, suggesting that the latter would duplicate the recent historic changes shown at Snowcap Lakes.

Beginning in the 1950's, logging of the Noeick valley had been undertaken by Crown Forest Industries Ltd. and other companies starting at tidewater in South Bentinck Arm and gradually moving farther upvalley each year. By the summer of 1983, valley slopes of the Noeick River about 13-15 km downstream of Ape Lake were being logged. In the same year, Bella Coola residents were pointing to the imminent completion of a forestry road which would link the South Bentinck Arm area to the Bella Coola valley by way of a 1100 metre high "Oldegaard Pass". This road would improve access to the Noeick Valley for forestry and public recreational use. Observations of the continued recession of Fyles Glacier in 1983 and the improved access to the lower Noeick valley prompted Ricker (1983) to file a proposal with federal and provincial authorities to assess the potential hazard posed by Ape Lake. Crown Forest Industries Ltd., the company which is logging the Noeick valley, was informed of the potential hazard posed by Ape Lake, and they agreed to make periodic visits to monitor conditions. Unfortunately no funds were available to make a detailed study of the lake and Fyles Glacier.

On October 17, 1984, employees of Crown Forest Industries Ltd. made a final visit to Ape Lake and concluded that the status quo would probably exist for another year. By this date, the melt season had ended and the lake did not show any conspicuous changes which might have occurred in response to the major storm of October 6-10. This storm created large floods in the watersheds on the east side of the Coast Mountains between the Squamish River to the south and the Dean River to the north (Map 1). However only three days later, on October 20th the lake suddenly drained into the Noeick River via a tunnel beneath the snout of Fyles Glacier near its terminus (Plate IB).

The resulting flood in the Noeick valley was first observed at 1100 hours on October 20th by recreational fishermen from Bella Coola who had driven from the Nusatsum basin across "Oldegaard Pass" into the Noeick valley (Map 2). The fishermen drove downstream along the forest access road adjacent to the river to a point about 30 km from South Bentinck Arm where a rapidly disappearing road convinced them to turn back. During the day, other sightseers reached this point, but at 1730 hours a rapidly rising flood crest blocked vehicular traffic 5-6 km upstream Plate VIIIA). On the following day, one party from Crown Forest Industries Ltd. drove to the Noeick valley to investigate while a second party flew over the lake. On October 22 several parties made an aerial reconnaissance to inspect and photograph the effects of the flood between the now half emptied lake and tidewater. Three days of bad weather, which unfortunately included some snowfall to subalpine levels, precluded further field investigations until the arrival of our study team from Vancouver on October 26. However, aircraft pilots did monitor the movement of debris in the fjordland channels throughout the first week following the event. Field surveys and interviews with local residents were conducted during a five day period (October 26-31) but the onset of heavy snow made further field work impractical.

On the afternoon of October 26, three members of the study team made a one hour aerial reconnaissance of the lake basin and Noeick River upstream of km 5 using a video camera and three 35 mm cameras to record the scene. On October 27, three members surveyed a channel cross-section of the Noeick River immediately downstream of Fyles Glacier and made preliminary measurements of water and strand line elevations in the Ape Lake basin. The fourth member of the party drove, via "Oldegaard Pass", into the Noeick valley to make observations of geomorphic changes. The full party returned to the Noeick valley the following day in the company of Messrs. R. Lenci and

R. Clarke, both employees of Crown Forest Industries Ltd., to examine the floodplain and obtain pertinent measurements wherever possible. Due to a severe blizzard to sea level, activity on the 29th was limited to obtaining interviews with eyewitnesses living in Bella Coola. On the 30th, another day was spent at Ape Lake by three members of the party, finishing with a second reconnaissance flight downvalley to tidewater and an interview with the watchman at the vacant South Bentinck logging camp. The field work was concluded with interviews with personnel from Atmospheric Environment Service, B.C. Ministry of Forests and Fisheries and Oceans Canada.

1.2 Glacier Outburst Floods

1.2.1 Distribution and Occurrence

Glacier outburst floods are the result of the sudden release of water stored within or behind a glacier, often leading to high magnitude discharges of short-duration. The largest floods are usually associated with lakes created in valleys which are blocked by glaciers. Such ice-dammed lakes are widely distributed throughout the world wherever glaciers occur. The large, catastrophic floods associated with ice-dammed lakes have occurred in Iceland so frequently that the Icelandic term 'jökulhlaup' is now used internationally to describe them.

Besides the frequent occurrence of glacier outburst floods in Iceland they have been reported to occur frequently in Pakistan (Hewitt, 1982), Norway (Aitkenhead, 1960; Whalley, 1971) and Alaska (Post and Mayo, 1971). Glacier outburst floods have also been recorded in the southern Andes, European Alps and the Caucasus in the Soviet Union.

Within Canada, glacier-dammed lakes occur in the highly glacierized areas of Baffin, Devon, Ellsmere and Axel Heiberg islands (Ricker, 1962), the Coast Mountains of British Columbia and the St. Elias Mountains of the Yukon. A large concentration of glacier-dammed lakes occurs in the St. Elias Mountains, Yukon Territory, and are frequently associated with surging glaciers (Young, 1977). In the Coast Mountains of British Columbia, there are many glacier-dammed lakes although the density of lakes per unit area is not as high as that in either the High Arctic Islands or the Yukon Territory.

Within British Columbia, glacier outburst floods have been recorded on the Stikine River due to the draining of Flood Lake (Kerr, 1948; Clarke and Waldron, 1984), on the Bear River caused by the draining of Strohn Lake near Stewart (Mathews, 1965) and on the Tulsequah River due to periodic drainage of Tulsequah Lake (Marcus, 1960). Perhaps the most notable examples in British Columbia are the large jökulhlaups which have occurred on the Salmon River due to periodic draining of Summit Lake near Granduc Mine north of Stewart (Mathews, 1973).

Prior to 1961, the Salmon Glacier blocked the valley drainage creating Summit Lake which drained northward over a bedrock sill into the Bowser River, a tributary of the Nass River. In late December, 1961, Summit Lake drained rapidly via a tunnel beneath the Salmon Glacier, causing a major flood on the Salmon River. The lake began to fill with spring melt in May, 1962. In May, 1963, the lake was reported to be at least half full, and by autumn of that year it was again over-flowing to the north (Mathews, 1965). The lake was observed to drain again in November, 1965, September, 1967, November, 1968 and August, 1970 (Gilbert, 1972). Since 1970, the lake has drained with increasing frequency although the magnitude of the floods appear to have decreased.

1.2.2 Drainage Mechanisms

Post and Mayo (1971) list seven mechanisms which would lead to the formation of a channel under, through or over the ice resulting in the rapid draining of a glacier-dammed lake. The most common mechanisms which are discussed in the literature are the overtopping of the ice dam and floating of the dam to allow water to escape beneath the glacier.

The obvious mechanism is water overflowing the ice-dam, generally along the glacier margin, leading to the erosion of a channel in the ice. For this to occur however, the maximum depth of water at the dam must be relatively low while the ice controls the elevation of the lowest drainage channel. There are many instances of lakes overflowing the ice. Lake George, a large, ice-dammed lake in Alaska drained annually by this mechanism between 1918 and 1966 (Stone, 1963; Post and Mayo, 1971), ending when the glacier no longer filled the valley. Other examples of overtopping are provided by Ricker (1962) and Liestøl (1956).

Alternately, the drainage may occur through or beneath the ice but it is more difficult to observe directly such instances of englacial or subglacial drainage. Thorarinsson (1939) suggested that water may be able to flow beneath the glacier once the water depth reaches approximately nine-tenths of the maximum ice thickness at the dam. Once a leak is established, the initial opening can be expanded rapidly by melting (Liestøl, 1956; Mathews, 1973). Nye (1976) and Clarke (1982) have provided a detailed theoretical treatment of the process leading to the formation of a tunnel beneath the ice dam. The analyses of Nye and Clarke indicate that thermal energy derived from the conversion of potential energy into heat and energy available in relatively warm lake water is important for the rapid formation of the tunnel, which is assumed to occur at the bed of the glacier.

In most instances, the occurrence of jökulhlaups follows a prolonged period of glacier recession. As valley glaciers recede, areas upstream of the glacier may become ice free while drainage remains blocked by the remaining valley ice. This ice forms a dam, impounding the local drainage and creating a lake with discharge flowing over the basin rim at the location with the lowest elevation which may or may not be formed by the glacier. With continued ablation of the ice, glacier thickness decreases until either the water impounded behind the ice is able to flow over the top or hydrostatic pressure in the lake is sufficient to force a passage beneath the glacier. The occurrence of outburst floods from Summit Lake is an example where the initiation of subglacial drainage is attributed to the recent, marked thinning of Salmon Glacier since 1920.

2.0 DESCRIPTION OF STUDY AREA

2.1 Study Area

Ape Lake (latitude 52° 06' N, longitude 126° 11' W) is located in the Pacific Ranges of the Coast Mountains of British Columbia. The lake, at an elevation of approximately 1395 m a.s.l., is near the drainage divide between the Noeick and Talchako-Bella Coola watersheds east of the axis of the Coast Mountains. Fyles Glacier (Plate IA, IB), which dams the upper valley of the Noeick River forming Ape Lake, is a moderate sized, alpine glacier about 9 km in length which originates in the basin circumscribed by West Jacobsen Mtn., Mongol Mtn., Mt. Dudra and Mt. Fyles (Map 2). Ape Glacier originates on the northern flanks of Mt. Jacobsen; it receded rapidly several decades ago but has recently advanced so that its terminus is about 100 m above Ape Lake.

Downstream of Fyles Glacier, there are two small glacierets in the upper canyon of the Noeick River (Plate IVB) formed by the refreezing of avalanche and icefall debris from a large glacier overhanging the valley (Map 2). Farther downstream, Purgatory Glacier has periodically extended to the valley floor and across the Noeick valley but is presently just above the valley floor (Plate VIB). Beyond this, the Noeick River flows on a broad floodplain with a low gradient for several kilometres (Plate IXC) before entering a narrow, tortuous canyon (Map 2). About 12 kilometres from tidewater, the Noeick River is again flowing in a broad floodplain, entering a short canyon at kilometre 7.0. Downstream of this canyon, the floodplain is wide and relatively flat to tidewater. The Noeick River has formed a delta where it flows into the head of South Bentinck Arm (Plate XC).

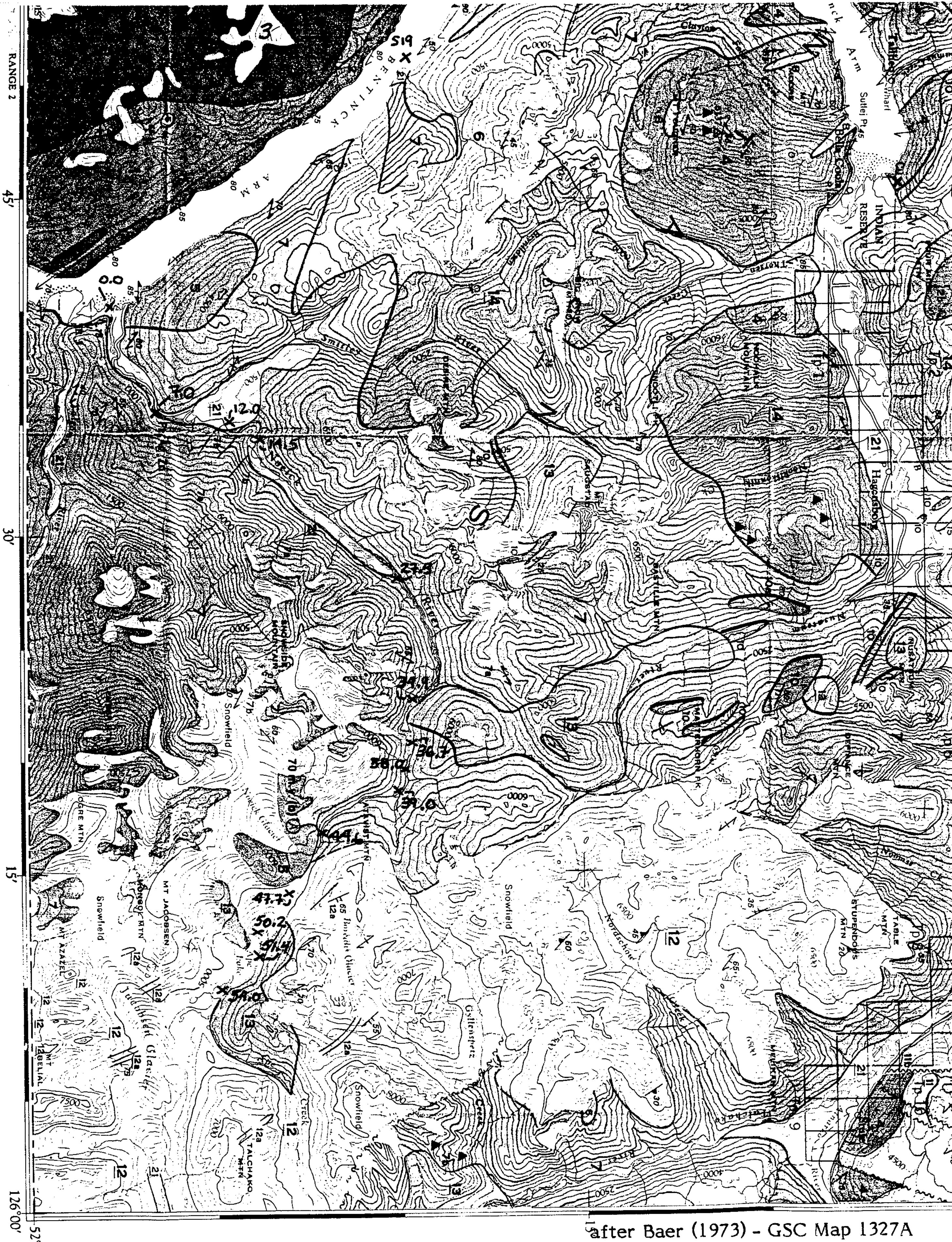
For the purposes of this report, the Noeick River has been subdivided into sections, marked by kilometre posts (K.P.) beginning at tidewater and extending upstream to Ape Lake. The location of various kilometre posts is shown in Figure 1 and Map 2.

The basin contributing runoff and meltwater to Ape Lake has an area of approximately 38 km^2, of which 20 km^2 or 53 % is glacier covered. When full, the surface elevation of the lake is controlled by an outlet over bedrock to the east where discharge flows into Ape Creek, a tributary of the Talchako and Bella Coola rivers.

2.2 Available Information

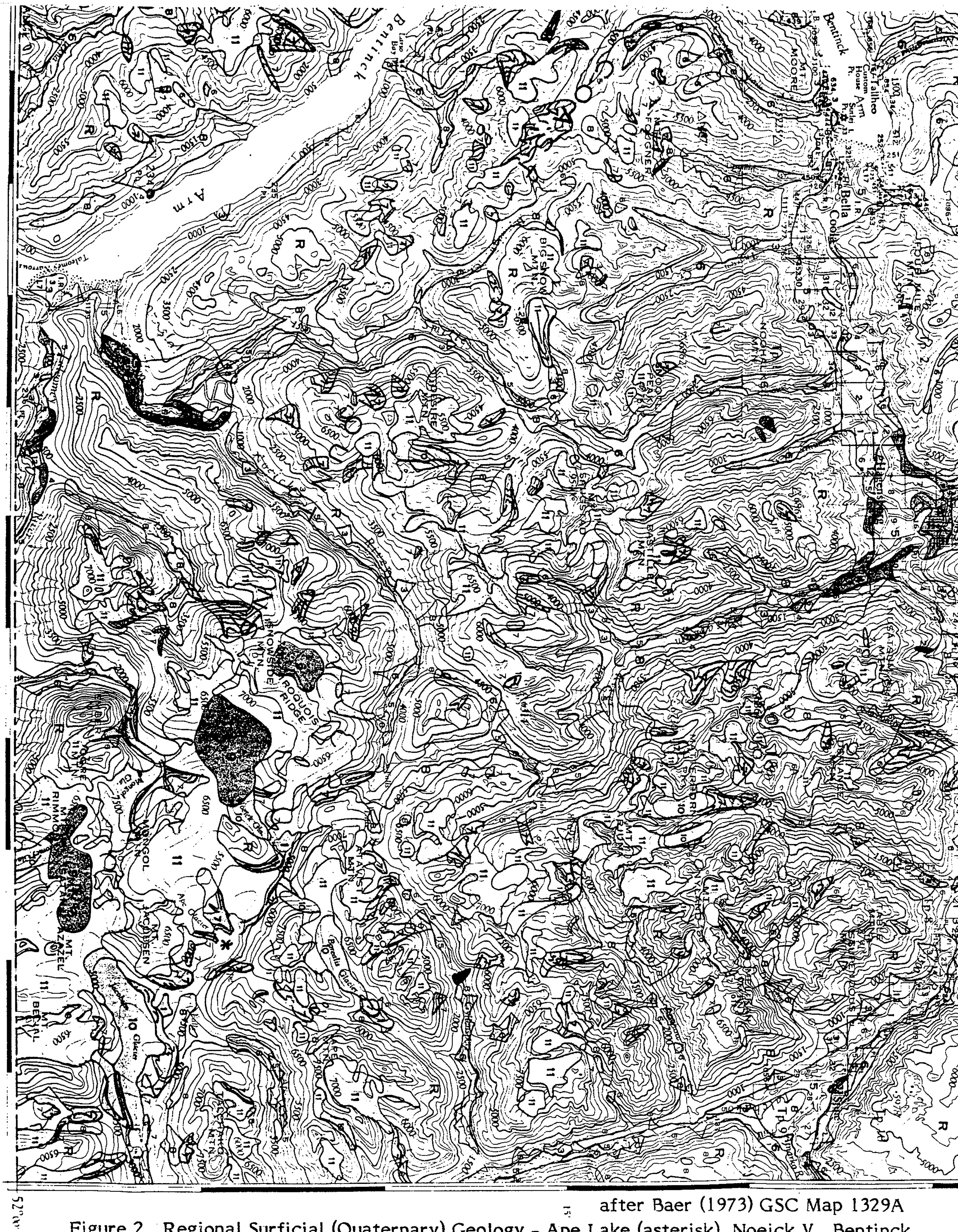
Except for an early reconnaissance of the coastline in 1908 and 1925 by R. Graham and V. Dolmage respectively, (Baer, 1973) little work was done in the study area until 1964 when the Geological Survey of Canada began to map the geology. The local geology is described by Baer (1965, 1967 and 1968) while a subsequent report by Baer (1973) included both bedrock and surficial geology. The surficial geology map which accompanies Baer's report is based upon reconnaissance mapping and does not differentiate Pleistocene and Neoglacial deposits (Figure 2). Ricker (1979) prepared a preliminary sketch map of the extent of Neoglacial moraines in the Ape Lake region and described the major geomorphic features.

Tipper et al. (1979) re-interpreted the bedrock geology presented by Baer (1973) to make it consistent with geologic interpretations of the surrounding areas, particularly the synthesis of a large area to the south of Ape Lake prepared by Roddick et al. (1979).



after Baer (1973) - GSC Map 1327A

Figure 1 Bedrock Geology of Ape Lake, Noeick Valley and South Bentinck Arm Area
x = River Reach Kilometre Posts. 1, 2 = Gneiss; 3, 4, 5, 6 = Coast Plutonic Rocks; 7 = Jurassic Metavolcanic & Diorite Complex (7a); 12 = Lower Cret. Andesites & Slates (12a); 13, 14 = Tertiary Plutons.



after Baer (1973) GSC Map 1329A

Figure 2 Regional Surficial (Quaternary) Geology - Ape Lake (asterisk), Noeick V., Bentinck Fjord Area. 1 = Glacio-lacustrine; 2 = Slides; 3 = Fans, Cones & Deltas; 4 = Terraces; 4a = Terraces?; 5 = Alluvium; 6 = Talus; 7 = Moraines; 8 = Ground Moraine Till; 9 = Ice Caps; 10 = Valley Glaciers; 11 = Glacier undiff.

A regional study of Pleistocene sea levels in adjacent fjords was conducted by Andrews and Retherford (1978). The results of this study have been incorporated into a broad synthesis of coastal Pleistocene sea levels and related features by Clague et al. (1982) and Clague (1983).

An extensive set of aerial photographs of the Ape Lake area are available. Between 1947 and 1978 the Federal and Provincial governments have flown nine sets of photography. Specific details of this photography are presented in Appendix 1.

There are no data on the glaciers of the Noeick Basin other than the information which may be obtained from periodic aerial photographic surveys and the occasional observation by mountaineers.

In 1983, Mr. J. Desloges initiated a study of the Bella Coola basin. In 1984, as part of that study, he visited Ape Lake to obtain sediment cores from the lake basin. As part of this study, 30 echo sounder profiles were made across the lake supplemented with 3 longitudinal profiles in the Ape Glacier embayment and the west basin of the lake. These data were used to prepare a bathymetric map of Ape Lake (Gilbert and Desloges, Unpub. Mss.). This map was subsequently modified to reflect features in the lake basin which were visible after the lake drained (Map 3 - inset diagram).

2.3 Physiography, Bedrock and Quaternary Geology

2.3.1 Physiography

Ape Lake and surrounding drainage systems lie at the northern end of the Pacific Ranges on the east side of the Coast Mountains within the Western System of the Cordilleran region of British Columbia (Holland, 1964). During the Pleistocene

Epoch, the area was covered several times by large ice sheets which had their source areas in the vicinity of the present day Monarch Icefield. The glaciers eroded deep, U-shaped valleys and cirques resulting in relief of up to 3000 m. In the immediate vicinity of Ape Lake glacial erosion has created local relief of up to 1600 m.

2.3.2 Bedrock Geology

The study area lies within the Intermontane Belt just to the east of the Coast Crystalline Complex, both of which lie within the Cordilleran Orogen (Tipper et al., 1979). There are, however, small unaltered plutons (units 13 and 14, Figure 1) underlying the Ape Lake basin and headwaters of the Noeick River. Volcanic rocks of Jurassic and Cretaceous age form the local bedrock. Rocks of Jurassic age have foliations with a westerly trend while the younger Cretaceous volcanics strike NW-SE across the study area. Except for South Bentinck Arm which is a major fault zone, there is little relationship between landforms and the degree of susceptibility of various rocks to erosion (Baer, 1973).

On a local scale the following items should be noted. First, the morainal forefield between Atavist and Fyles glaciers is comprised of volcanic rocks of unit 12 and there are few plutonic clasts from the bedrock underlying Fyles Glacier. Secondly the Noeick River floodplain is overwhelmingly dominated by plutonic clasts as far downstream as km. 34 despite the fact that valleys are eroded in older volcanic rocks of unit 7 (Figure 1).

2.3.3 Quaternary Geology

During the Pleistocene Epoch, one of the spreading centers of the Cordilleran Ice Sheet was located in the Ape Lake area

(Tipper, 1971; Prest et al., 1969). Clague (1983) indicates that the ice sheet surface in the Late Wisconsinan maximum (14,500 to 15,000 years B.P.) was at least 2000 metres above sea level. Baer (1973) in his regional mapping, found evidence that the ice sheet rode over surfaces 2500 metres above sea level similar to maximum present day elevations of the Monarch Icefield. The rapid retreat of the ice sheet beginning about 12,200 years B.P. has been documented by Tipper (1971), Andrews and Retherford (1978) and Clague et al. (1982).

From studies elsewhere in the Coast Mountains the highlights of the recent glacial history of the Noeick valley system can be summarized as follows:

- 8400-5200 yrs. B.P. post glacial warming or Altithermal period
- 3200-2300 yrs B.P. Early Neoglacial glacier advance
- 700-1250 yrs B.P. possible glacier re-advance
- 300-450 yrs B.P. Late Neoglacial advance (Initiation of Little Ice Age)
- 170-240 yrs B.P. 18 th Century re-advance.
- 60-100 yrs B.P. One or more 19 th century "climax" advances and conclusion of the Little Ice Age.

The documentation of these Holocene advances is found in Fulton (1971), Ryder et al. (Mss. Rept), Alley (1976), Mathews (1951), Ricker (1980) and Ricker and Jozsa (1984) among others. The extent of each ice advance is variable and in most cases the older ones were completely obliterated or overridden by the Little Ice Age advances.

2.4 Climate and Hydrology

The climate of the Noeick and Ape Lake basins is strongly influenced by the rugged topography and proximity to the Pacific Ocean. The Coast Mountains, rising to elevations of 3000 m or more, form a barrier to the eastward flow of moist Pacific maritime air. As a result, the area on the windward side of the mountains is subject to heavy precipitation enhanced by strong orographic effects.

Precipitation occurs throughout the year with the largest amounts falling in October, November and December, usually as snow at the higher elevations. No measurements of precipitation have been made in the two basins but mean annual precipitation may be estimated by adding the average May-October rainfall measured at Bella Coola to the average water equivalent snow depth measured over a four year period at two nearby snow courses (station 3C02, Bella Coola, El. 1380 m and station 3C03, Machmell River, El. 1380 m). Using this method, the estimated mean annual precipitation is about 2140-2340 mm. This is lower than the mean annual precipitation shown by Farley (1979) on his map of British Columbia. According to Farley, the headwaters of the Noeick River lie between isopluvial contours with mean annual precipitation of 2500 to 3500 mm.

The Noeick River drains an area of 562 km^2, of which 84 km^2 or 15 % is covered by glaciers (measured from topographic maps based on 1964 air photos). About 53 % of the Ape Lake basin is covered by glacier ice, mainly because of its higher average elevation. No discharge data are available for either the Noeick River or Ape Creek. The mean annual discharge and magnitude of floods can be estimated however, using discharge data obtained at a gauging station on the Nusatsum River which drains the area immediately to the north of the Noeick River headwaters.

Based upon 12 years of data, the mean annual runoff in the Nusatsum basin is 1960 mm. This estimate compares favorably with the estimated annual precipitation noted above. By prorating the discharge according to basin area, the mean annual flood of the Noeick River at the mouth is estimated to be approximately 210 m^3/s. The maximum mean daily discharge of the Noeick River at the mouth with a recurrence interval of 100 years is estimated to be about 460 m^3/s.

2.6 Lake and Glacier Geometry

Ape Lake is located in a flat-bottomed, bowl shaped depression formed by glacial erosion (Map 3). End moraines on the floor of the depression mark the former maximum extent of the glaciers. Table 1 lists the surface area and volume of each sub-basin formed by the end moraines as well as surface area and volume of the entire lake.

Information on the morphometry of Fyles Glacier and other large glaciers in the vicinity of Ape Lake is listed in Table 2.

TABLE 1 APE LAKE MORPHOMETRY

Subbasin	Area km^2	Max Depth m	Volume[1] $\times 10^6$ m^3 Full	Drained	Residual Depth, m	Volume Released $\times 10^6$ m^3 Top 10 m	20 m	Total
West								
E Shoal	0.17	ca. 30	2.077	0.033	ca. 8	1.368	2.004	2.044
W Basin	0.61	57.6	15.752	0.372	0–15.6[2]	5.947	10.939	15.380
Subtotal	0.78	-	17.829	0.405	-	7.315	12.943	17.423
East	1.50	70.6	64.947	38.768	50.5	13.997	26.179	26.179
Ape Gl. Bay	0.19	ca. 30	2.808	0.632	ca. 15	1.598	2.177[3]	2.177
Totals	2.47	-	85.584	39.805	-	22.910	41.299	45.779
Percent of Totals	100	-	100	46.5	-	26.8	48.3	53.5
Surface Area, km^2			2.47	1.27		2.12	1.66	1.27

For locations of subbasins see Map 3, Figure 3 and Plates I and II.

Note 1. Volumes determined using the formula $V=h/3\ (A_1 + A_2 + \{A_1A_2\}^{0.5})$ where h = contour interval, A_1 = area of upper contour and A_2 = area of lower contour line.

Note 2. The west basin drained completely except for a sub-basin formed by a bedrock sill which creates a small pond at the east end of the basin.

Note 3. For the Ape Glacier embayment the maximum drawdown was about 15 m and hence a drawdown of 15 m rather than 20 m was used in computing the volume of water.

TABLE 2 MORPHOMETRY OF GLACIERS IN THE UPPER NOEICK VALLEY

	Glacier				
Parameter[1]	Fyles	Ape[2]	Atavist	Noeick	Purgatory[3]
Area, km^2	20.0	4.6	5.0	8.9	13.9
Axial Length, km	9.2	3.7	4.7	5.3	7.8
Maximum Width, km (Ablation Zone)	3.5	0.8	1.2	1.6	0.6
Highest El., m (Accumulation Zone)	2680	2650	2350	2530	2930
Lowest El., m (Base of Snout)	1320	1460	1570	1460	545
Relief, m	1360	1190	780	1070	2385
Ave. Gradient	0.15	0.32	0.17	0.20	0.31
Approx. Average Elevation of Equilibrium Line in 1978 and 1954	1890	1840	1860	1660	1800
Aspect of Snout	NE	NE	SW	NE	N

Notes:

1. From 1:50,000 Federal Topographic Map 93 D/1, based on 1964 aerial photography. Elevation of equilibrium lines estimated from 1954 and 1978 air photos.
2. Ape Glacier has two snouts; the lower eastern snout is used where appropriate.
3. Based on 1978 snout position; with the exception of the equilibrium line other values estimated using data from Federal topo map.

3.0 STUDY METHODS

3.1 Interviews

As none of the authors were present during the jökulhlaup, it was recognized that observations by local residents would be needed to help reconstruct the sequence of events during the flood. Fortunately there were several people in the Noeick Valley with cameras who recorded part of the event as it was happening. In addition, personnel from Crown Forest Industries Ltd. were cognizant of the flood potential and made observations before and immediately after the event.

A detailed compilation of notes made during each interview is presented in Appendix 2. The most important observations obtained from these interviews are used to reconstruct the sequence of events and are described in section 5.3. Interviews were obtained from the following individuals:

Crown Forest Employees

- Mr. R. Lenci, Forestry Engineer, Bella Coola
- Mr. B. Storry, Operations Engineer, Hagensborg
- Mr. G. Davidson, Foreman, Hagensborg
- Mr. C. Kabel, Camp Watchman, South Bentinck Arm Camp.
- Mr. R.B. Clark, Cruising Supervisor, New Westminster

Wilderness Airlines Ltd, Bella Coola

- Messrs. T. Bonner, C. Schuetze and W. Sissons, Pilots.

Chief Pilot for Dean River Air Services, Nimpo Lake.
Dr. and Mrs. W. Danielson, Dentist in Bella Coola
Mr. K. Danielson, University student, Bella Coola
Mr. D. Dunaway, Editor, The Coast Mountain Courier, Hagensborg
Mr. S. Gascoyne, resident of Hagensborg
Ms. K. McGwiny, Envirocon Ltd., Vancouver, B.C.

Mr. R. Ratcliff, Autobody Mechanic, Hagensborg

Mr. R. Skelly, Pilot, Transwest Helicopter (1965) Ltd, Bella Coola

Mrs. C. Zigler, B.C. Forest Service, Anahim Lake.

3.2 Lake Measurements

A helicopter was used to gain access to the Ape Lake basin and vicinity and to obtain overview photographs of the basin using 35 mm cameras and a video recorder. Elevations of locations easily accessible on foot were made using Paulin altimeters while the relative elevations of more inaccessible locations in the lake basin and on the glacier were measured using the helicopter altimeter.

Several cores of lacustrine sediments were obtained at one site within the Ape Lake basin. The cores are being analyzed by Mr. J. Desloges, Department of Geography, University of British Columbia. Tree ring discs were collected from near the crest of Fyles Glacier moraine and from logs buried in mud at the east end of the Ape Lake basin below full pool water levels (see Figure 3 and Appendix 4).

Although members of the field party were able to get close to the tunnel portals, it proved to difficult to obtain precise measurements of tunnel dimensions so observations were limited to obtaining photographs and estimating tunnel dimensions (see section 4.2.1).

3.3 Glacier Observations and Measurements

3.3.1 Field Work

Glaciological field work consisted of photography, measurement of elevations of the glacier sole and surface in the vicinity of the tunnel entrance and construction of stone cairns near

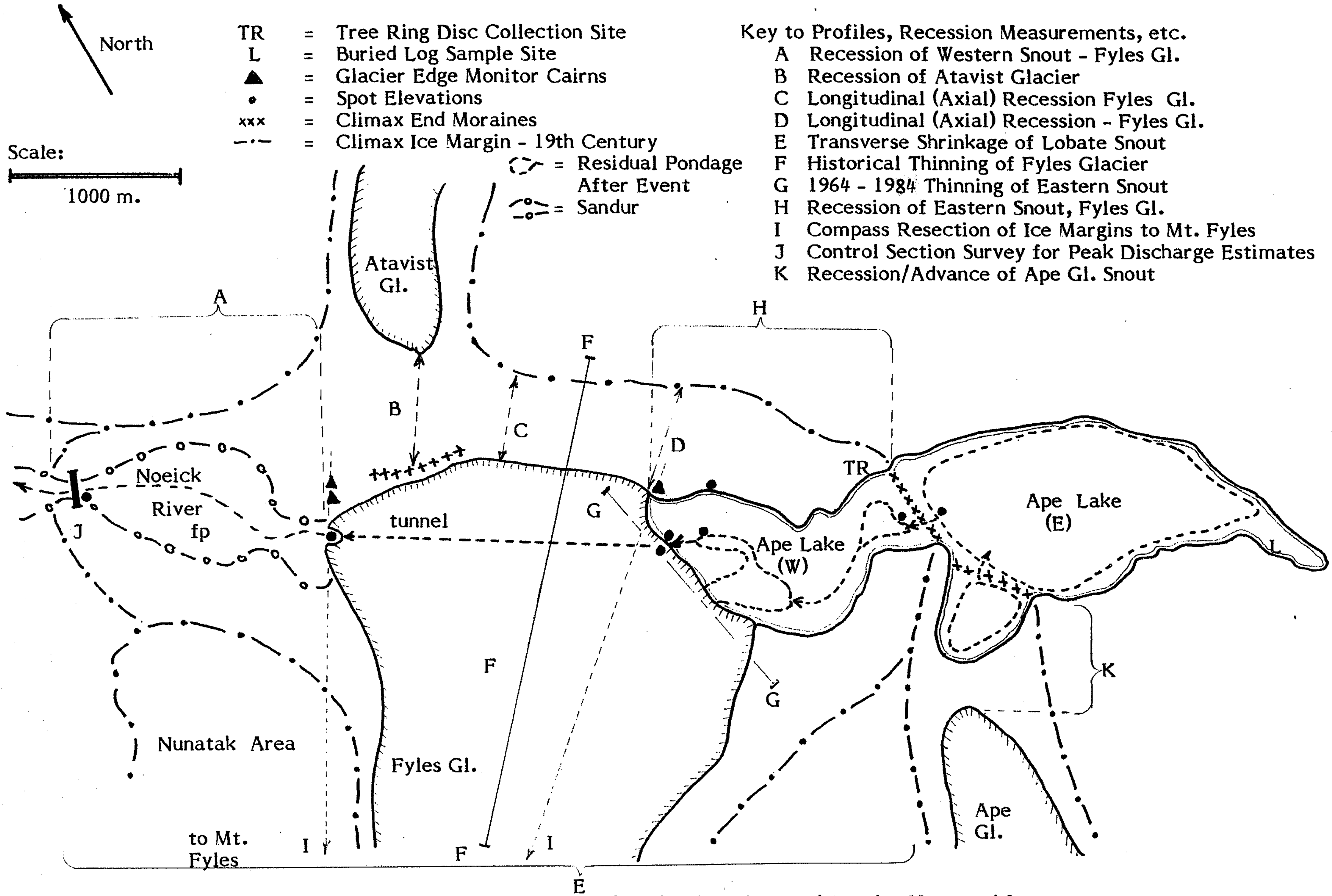


Figure 3 Sketch Diagram of Fyles Glacier - Ape Lake Showing Lines and Locales Measured for Glacier Recession, Peak Discharge and other Miscellaneous Data Collection Pts.

the glacier margin to monitor future retreat (or advance) of Fyles Glacier.

The elevation of the glacier sole and uppermost ice surface above the tunnel were measured using Paulin altimeters, scaled to read to the nearest two feet (0.61 m). All measurements of elevation were made relative to the normal full pool elevation of Ape Lake, 1 395 m a.s.l. High winds during the measurement periods almost certainly affected the accuracy of the readings.

Stone cairns or monitors were constructed near the snout of Fyles Glacier so that compass bearings on Mt. Fyles were approximately parallel to the eastern and western margins of the glacier termini (Plate III B and Map 3). Changes in the position of the termini relative to the cairns can be easily measured by future field parties.

3.3.2 Office Studies

The 1:50,000 scale NTS map sheet (Jacobsen Glacier, 93 D/1) was enlarged to a scale of 1:7500 to provide a working base for plotting the morphometry of the Ape Lake basin and the snout of Fyles Glacier. Aerial photographs taken in 1954 and 1978 were enlarged to the same scale to facilitate plotting of the ice margins. Where vertical trimetrogon photographs flown in 1947 and 1951 are available, the location of the glacier margins was transferred to the map by relating the position of the ice to prominent local features visible in the 1954 photography. The position of ice margins shown in trimetrogon photos were also estimated by comparison with the 1954 photography although the relative accuracy is lower since only the right oblique photographs cover the Ape Lake basin. On the trimetrogon photography, the snout of Ape Glacier coincides with a bedrock outcrop at the lake margin so that the position of the ice margin can be determined relatively accurately. An accurate position of the eastern edge of Fyles Glacier is much more difficult to locate because Ape Lake expands as the glacier retreats and there are fewer ground features to use as reference points.

The amount of retreat or advance of the glacier snout was measured from air photos using two methods. In the first method, the position of the ice margin was plotted on the 1:7500 scale base map for each year of available photography. The distance between successive positions of the ice margin or the distance between the glacier snout and the maximum extent of the glacier inferred from the position of the climax moraine were measured. The change in position divided by the time interval provides an estimate of the annual rate of retreat or advance of the glacier snout. Where irregular ice margins made simple linear measurements of distance difficult, the change in area of ice was measured using a polar planimeter. The measured area was then divided by the width of the glacier and the length of time between observations to estimate the annual rate of advance or retreat. The location of various measurements are shown on Figure 3.

Changes in the thickness of Fyles Glacier snout were estimated from topographic maps and reconstruction of a longitudinal profile of the ice surface (Figure 3). To reconstruct the longitudinal surface cross-section, a profile of the median longitudinal axis of Fyles Glacier (045° Azimuth) was obtained from the 1:50,000 scale topographic maps. The measured longitudinal profile began at the 5200 foot (1585 m) ice surface contour and extended northward to the ice margin (1964 position). The profile was extended upslope to a conspicuous vegetation trimline which demarcates the 19th century glacial maxima at the 4900 foot (1493.5 m) contour (Map 3, Plates IA, IB). The longitudinal profile of the ice surface during the glacial maxima was then estimated by assuming that the glacier snout was at an elevation of 4900 feet, coinciding with the climax trimline, with a similar surface profile to that measured from the topographic map. If projected far enough the slope of the profile would have to decrease somewhat so that the climax and 1964 surfaces have roughly the same elevation near the equilibrium line (Table 2). However, as this point is several kilometres distant and we are only interested in the lower kilometre of the glacier at the snout, no adjustment was made to the profile slope.

Several measurements of the difference in elevation between the two surface profiles were averaged and divided by the length of time between 1964 and the occurrence of the glacial maxima. This provides an estimate of the average rate of thinning. The largest uncertainty in this technique is the date of the glacial maxima which probably occurred sometime between 1856 and 1904. Other estimates of the rate of decrease in ice thickness on the eastern and western snout of Fyles Glacier were made by comparing the position of the 1978 ice margin to the equivalent ice surface contour shown on the topographic map which is based on 1964 air photographs. Field measurements obtained in 1984 were also compared with the equivalent ice surface contour on the topographic map to obtain estimates of the change in vertical elevation. The location of each of these measurements is shown on Figure 3.

3.4 Geomorphology

Because of limited time and difficultly of access to many locations in the Noeick Valley, a large portion of the study was restricted to observations made during helicopter flights on October 26 and 30, supplemented with information obtained from Crown Forest Industries Ltd. Major geomorphic and erosional features were photographed using 35 mm cameras and a video recorder for subsequent study. No landings were made in the valley during these flights.

Using road access, detailed observations and measurements were made of the floodplain between Km 34 and 37 in the vicinity of Purgatory Glacier. Measurements included heights of cut banks, average channel gradients and channel widths using a range finder. Dimensions of large boulders (Plate VIIB), known to have moved during the flood because they were found on top of trees and vegetation, were also measured. Samples of buried wood were also obtained from a fresh exposure of the Purgatory Glacier climax moraine (Plate VIIA).

A helicopter was used to access a location on the Noeick River approximately one kilometre downstream of Fyles Glacier where high banks in erosion-resistant till funneled flood waters through a single channel. Three cross-sections were surveyed using a theodolite to obtain cross-sectional areas, local channel gradients and elevation of the probable maximum height of water levels during the flood. Maximum water levels were estimated from evidence of bank erosion, strand lines and the elevation of stranded ice blocks on floodplain terraces.

Interpretation of air photos taken in 1954 and 1978 was used to check on the amount of floodplain erosion which can be attributed to the glacier outburst flood.

4.0 RESULTS AND OBSERVATIONS

4.1 General

A detailed compilation of notes made during interviews with eyewitnesses is presented in Appendix 2. The salient points of these interviews are described briefly in section 5.3.

4.2 Field Measurements and Observations

4.2.1 Lake Measurements and Observations

After Ape Lake drained, the basin was divided into several smaller sub-basins formed by moraines (Map 3, Plate IB). Water levels at the time of the field visit in October varied from about 15 m below full pool level in the Ape Glacier embayment to about 58 m below maximum water level in the west basin adjacent to Fyles Glacier. The most prominent feature associated with the draining of Ape Lake is the large, water-eroded breach in the climax end-moraine dividing the lake into east and west basins (Plate IIB). In addition, the slopes of the basin have been very unstable and exhibit many collapse features and gullying, most of which apparently occurred after the lake drained. The location of many of these slope failures is shown on Map 3 and illustrated in Plates IIA, IIB, and IIIA.

Except along the climax moraine, slopes of the east basin exhibit sliding and slumping features around the entire basin perimeter. In the shallow embayment at the east end of the basin near the lake outlet, the slumps form a series of small, retrogressive failures (Plate IIA) rather than larger slumps which are common elsewhere around the basin perimeter.

At one site in the eastern arm of the lake just below maximum water level, a newly exposed and frozen section of sediment was examined. The top 0.9 m of the exposure contained about 180 layers of lacustrine silt (based upon an inferred average thickness of 0.5 cm per layer). Underlying this was a 2 m thick deposit of silt which in turn overlay a buried log. A sample of the log was obtained for possible radiocarbon dating by the Geological Survey of Canada (Our sample number 12C, Appendix 4). Sediment cores up to 0.8 m long were obtained in the vicinity of this site for subsequent analysis of the varves and sediment stratigraphy.

The thickness of lacustrine sediments in the east basin indicates that Ape Lake has probably existed for several hundred years or more. Radiocarbon dating of the tree sample obtained from the lake sediments may provide a minimum age on the longevity of Ape Lake.

A well developed strandline about 10 m below full pool elevation was visible at a few locations in the east basin which suggests that the water level was more or less steady at this elevation for a short time (ie. 1-3 hours) when the lake was draining.

Except for a small pool of water near the ice and another near the climax moraine, the west basin drained completely (Plate IIB and IIIA). Water continued to flow from the east basin through the breached moraine into the pool and thence over a bedrock sill into a steep rock chute leading to the mouth of the tunnel beneath Fyles Glacier (Plate IIIA). The tunnel was located on bedrock at the base of the ice. The tunnel mouth was more oval than circular with a height of about 2-3 m and a width of about 5-6 m. The roof of the tunnel was formed of solid ice except for a small portion at the mouth which consisted of collapsed and refrozen ice. The tunnel beneath Fyles Glacier is estimated to be about 1.86 km long with the exit approximately 17.4 m lower than the inlet.

On the slopes of the west basin were long, sinuous, asymmetric ridges up to 1-1.5 m high, roughly parallel to the ice margin (Plate IIIA, IIB, IIC). These features were formed, possibly annually, in a subaqueous environment and are thought to be DeGeer moraines. The sediment texture and asymmetric form of these features are very similar to the description of DeGeer moraines cited by Embleton and King (1975) and Sugden and John (1976). In the same area were buried logs, one of which appeared to have been burnt. Unfortunately because of time constraints we were unable to collect samples.

4.2.2 Glacier Measurements and Observations

4.2.2.1 Chronology of Glacial Maxima

Preliminary estimates of the chronology of the recent maximum extent of Fyles Glacier can be made by counting annual growth rings of trees growing on or near the climax moraine of Fyles Glacier (see Map 3 for locations). The dates of tree colonization provides a minimum age of the maximum extent of Fyles Glacier and hence the presence of Ape Lake at full pool level.

Sample A (Pinus sp.), collected from the area between the lake shoreline and moraine, began growing about 1950. Another Pinus (sample C) obtained from the crest of the moraine had very tight rings at the core and dated to about 1915 A.D. Sample B, (Abies amabilis ?), taken from a site on the outer slopes of the climax moraine, dated to 1914 A.D. Photocopies of the tree discs are shown in Appendix 4. Based upon observations of the time required for conifers to colonize alpine moraines in northern Garibaldi Park (Ricker and Tupper, 1979), it would appear that Fyles Glacier retreated from the moraine between 1856 and 1904 A.D. The bedrock upslope of the moraine on the opposing valley wall (Plate I), however also has a narrow band of successional conifer growth which has not been sampled. This zone could also be part of either a late 19th century advance or an earlier 18th century advance which occurred in many other areas of the western Cordillera.

4.2.2.2 Recent retreat and advance of glaciers in the upper Noeick basin.

An examination of glacier snout positions on successive aerial photographs indicate no unusual behaviour of glaciers in the upper Noeick basin. Up to 1964, most glaciers in the Noeick basin were receding (Tables 3 - 6). Since 1964, several glaciers have advanced, Atavist Glacier has maintained its position while Fyles Glacier has continued to retreat from its glacial maximum. Generally, however, the glaciers in the Noeick basin exhibit a pattern similar to that of other glaciers in the Coast Mountains where many have advanced between 1964 and 1980 (Ommanney, 1984).

Of particular interest are several small glacierets in the upper canyon of the Noeick River (km 41.4 to 44.6) between Purgatory and Fyles glaciers (Plate IVB). Historically, the hanging glacier filled the entire canyon of the Noeick River downstream to km 39.0, effectively blocking the upper Noeick drainage. An examination of air photos indicates that most ice in the canyon had disappeared by 1954 (Table 6). Prior to 1954, during a prolonged period of downwasting of the valley glacier, the Noeick River drained via an ice marginal channel on the northwest side of the canyon wall. Whether the ice completely blocked the valley impounding a lake at some time is unknown.

The glacierets are not plotted on the 1:50,000 scale topographic maps (based on 1964 photography) which suggests

TABLE 3 RECENT FLUCTUATIONS OF THE SNOUT OF FYLES GLACIER

Domain Analyzed (see Fig. 3)	Time Span (A.D. Years)[1]	Horizontal Change (-) = retreat, m (+) = advance, m	Average Rate of Change m/yr	Analysis Methodology[2]
Breadth of Snout (ie lake edge to Noeick Valley outflow)	Climax-1951 [4]	- 1527	-16.1 to -32.5	L,P
	1951 - 1954	- 126	- 42.0	L,P
	1954 - 1964	- 400	- 40.0	L,P,M
	1964 - 1978	- 195	- 13.9	L,M,P
	1978 - 1984 [5]	- 128	- 21.3	L,R,P
Northern Edge-longitudinal axis adjacent to East side of Atavist Glacier[3]	Climax-1954	- 204	- 2.1 to - 4.1	L,P
	1954 - 1964	- 95	- 9.5	L,P,M
	1964 - 1978	- 75	- 5.4	L,M,P
Northeast Corner longitudinal axis[3] (1978-1984, ice to lake edge)	Climax-1954	- 339	- 3.5 to - 6.8	L,P
	1954 - 1964	- 60	- 6.0	L,P,M
	1964 - 1978	- 177	- 12.6	L,M,P
	1978 - 1984 [5]	negligible	neg ±	L,R,P
Eastern Snout (along lake edge)	Climax-1947	ca - 649	- 7.1 to -15.1	L,P(T),B
	1947 - 1951	- 131	- 32.8	L,P
	1951 - 1954	- 132	- 44.0	L,P
	1954 - 1964	- 502	- 50.2	L,P,M
	1964 - 1978 [6]	- 210	- 15.0	A,M,P
	1978 - 1984 [5]	ca - 8	- 1.3	A,P,B,R
Western Snout (Noeick Valley)	Climax-1951 [4]	- 811	- 8.5 to -17.3	L,P
	1951 - 1954	- 37	- 12.3	L,P
	1954 - 1964	- 510	- 51.0	L,P,M
	1964 - 1978 [6]	- 112	- 8.0	A,M,P
	1978 - 1984 [5]	- 123	- 20.5	L,P,R
	or 1978 - 1984	- 97	- 16.1	A,P,R

NOTES TO TABLE 3

Note 1. Date of climax is between 1856 A.D. and 1904 A.D. based on 70 annual rings counted in trees growing on a terminal moraine near Ape Lake. That is, the ring count is adjusted for a time lag of conifer colonization of the site according to studies carried out by Ricker and Tupper (1979)

Note 2. Methodology symbols: L = linear measure, ice tip to ice tip; A = areal measure divided by width of snout yielding average retreat across entire snout; P = measured from aerial photos using a 440 metre base line established across narrows of Ape Lake; M = measure of ice position relative to ice position shown on topographic map; P(T) = aerial trimetrogon (oblique) photograph used to show approximate position of ice; B = 1984 bathymetry map of Ape Lake.

Note 3. This dual set of measurements represents the narrowest and broadest zone of recession along the northern terminus of the glacier. Averaging the two, the recession would read as follows: Climax - 1954 (-272 m or -2.8 to -5.4 m/yr); 1954 - 1964 (- 78 m or - 7.8 m/yr) and 1964 - 1978 (-126 m or - 9m/yr).

Note 4. The western snout of Fyles Glacier coalesced with neighbouring Noeick Glacier during an undated climax which, from vegetative cover, appears to be of an eighteenth century event. A fresher, fluted and hummocky morainal area slightly upvalley from the former is assumed to be 19 th century climax position of the glacier snout and all measurements are taken to this datum.

Note 5. It is assumed that the topographic map outline of the glacier (based on 1964 aerial photographs) is correct. For the western snout of Fyles Glacier the 1964 to 1978 ice tip to tip difference is only about 42 metres; hence the use of areal calculation to document a more realistic picture.

Note 6. The 1984 resection surveys are based on a magnetic compass; the error margin could be considerable.

TABLE 4 RECENT FLUCTUATIONS OF THE SNOUTS OF GLACIERS IN THE VICINITY OF FYLES GLACIER

Glacier	Time Span (A.D. Years) (note 1)	Horizontal Change (-) = retreat, m (+) = advance, m	Average Rate of Change m/yr	Analysis Methodology (note 2)
Ape Glacier	Climax-1947	- 419	- 4.6 to -9.7	L,B,P(T)
	1947 - 1951	- 83	- 20.8	L,P(T),P
	1951 - 1954	- 45	- 15.0	L,P
	1954 - 1964 [3]	- 140	- 14.0	L,P,M
	1964 - 1978	+ 33	+ 2.4	L,M,P
Noeick Glacier	Climax-1954 [4]	-1147	-11.7 to -22.9	L,P
	1954 - 1964 [3]	- 743	- 74.3	L,P,M
	1964 - 1978	+ 294	+ 20.9	L,M,P
Atavist Glacier	Climax-1951 [5]	- 585	- 6.2 to -12.4	L,P
	1951 - 1954	negligible	neg ±	L,P
	1954 - 1964 [3]	- 168	- 16.8	L,P,M
	1964 - 1978	negligible	neg ±	L,M,P

Note 1. Date of climax ice position is assumed to be similar to Fyles Glacier though the smaller glaciers are probably at the older end of the spectrum of 1856 to 1904 A.D.

Note 2. Methodology symbols: see Table 3 on Fyles Glacier Fluctuations for coding descriptions.

Note 3. Extent assumed to be correctly shown on topographic map but left lateral margin of Atavist Glacier appears to be plotted in a much reduced position.

Note 4. Climax position of Noeick Glacier blocked the Noeick Valley to about the same level two or more times in the last few centuries.

Note 5. At the 19th century climax, the extent of Atavist Glacier was limited by the mergence with Fyles Glacier, though morainal lithologies indicate that Atavist reached climax position first.

TABLE 5 PURGATORY GLACIER ICE FRONT FLUCTUATIONS

Period (A.D.)	Time Interval (years)	Horizontal Distance for period (metres)	Average Annual Rate (m/year)
1856-1904 [1] to 1947	91-43	- 1552	- 17.1 to 36.1
1947-1954 [2]	7	negligible	negligible
1954-1964 [3]	10	- 1331	- 133.1
1964-1972 [4]	14	+ 1818	+ 129.9
1978-1984 [5]	6	- 308	- 51.3

Notes:

Note 1. There are two or more climax lateral moraines which terminate in frontal outwash at about the same position downvalley. The inner (and younger) moraine is the 19th century climax tentatively assigned the age of 1856 to 1904 A.D., based upon what is known about Fyles Glacier 11 km upvalley.

Note 2. By comparison of the two years of aerial photos visually, the differences are not visible.

Note 3. The position of the 1964 ice front is taken from the 1:50,000 topographic map. Aerial photos have yet to be inspected to confirm what appears to be an incorrectly located ice terminus.

Note 4. The terminus is floating on a lake measuring 550 m in breadth and 440 m in length.

Note 5. This is the ice position as of October 12, 1984. The flood waters of the jokulhlaup eroded the ice front back another 2 or 3 metres on the following weekend.

TABLE 6 HISTORIC REGIMEN OF GLACIER (OR GLACIERETS) IN THE UPPER CANYON REACH OF NOEICK VALLEY (K.P. 39.0 TO 44.6).

Condition [1]	Length of Icefield in the upper canyon metres	Percent Change metres/year
Climax Extent of Ice in Canyon (ca. A.D. 1856 to 1904) - continous icefill up to 300m thick.	4570	
		- 55.8 % 26 to 51 m/yr
1954 Extent of Ice in Canyon (Two debris cone glacierets) [2]	2020	
		+ 53.6 % 45.1 m/yr
1978 Extent of Ice in Canyon (Three debris cone glacierets) [2]	3100	
		- 35.6 ± 9.7 % ca 184 m/yr
1984 Post-jokulhlaup Extent of Ice in Canyon (two cones)	ca 2000 ± 300	

Notes:

Note 1. The topographic map, based on 1964 air photos, shows no residual glacierets but the photos have yet to be inspected to confirm this.

Note 2. Comparing the two glacierets common to both years of photography, the downvalley cone grew 25.4 % in valley bottom breadth (+9.6 m/yr) whereas the upvalley cone enlarged only 4.7 % (+2.2 m/yr) in the 24 year time span.

that they were very small or non-existent in that year. In subsequent years the glacierets have increased in size, fed by avalanche and icefall debris from the glacier overhanging the valley wall.

The glacierets show conspicous dirt bands which may form annually in the late summer and early autumn due to rock fall. In the case of the downstream glacieret (Plate IVB), photos taken by R. Lenci show thick bands of debris as well as many thin, discontinuous dirt layers between them. Thirteen major bands show a synformal structure with an axis parallel to the valley. There are also at least six more bands conformably wedged in the area between the valley floor axis of the synform and the western valley wall. It is tempting to suggest that these data show a renewal of glacieret activity since 1964 with the major bands indicating annual accumulations. In the case of the upstream glacieret, the exposed ice at the exit of the tunnel formed by the flood shows 18 debris bands tightly folded on a vertical synformal axis running parallel to the canyon floor. However the present glacieret surface truncates the bands and any deposition in 1984 will lie unconformably over older deposits.

During the outburst flood, the glacierets blocked the flow, creating a backwater effect until the water had enlarged pre-existing tunnels sufficiently to convey the large volume of water. Although the glacierets, particularly the one downstream, are not thick, at no time did the water ever flow overtop of the ice. Besides enlarging the tunnels beneath the glacierets and removing a large amount of ice at the ice margin, the flood appears to have been responsible for creating a large collapse feature on the surface of the upstream glacieret (Plate IVB).

Purgatory Glacier, several kilometres further downstream has exhibited large variations in the position of the glacier snout (Table 5). It has extended across the valley floor (El. 550 m) one or more times during the last several centuries as shown by successional forest cover and the position of a climax moraine (Map 2). After a period of general recession between 1856 and 1964, the glacier advanced until by 1978 it had regained the valley floor, covering much of Purgatory Lake. Recently the glacier snout has receded to the edge of the lake (Plate VIB).

4.2.2.3 Thinning of Fyles Glacier

The position of the climax moraine (Plates IA, IB; Maps 2, 3) and data from successive aerial photographs indicates that the thickness of Fyles Glacier snout has been decreasing over the last 70 or more years. Measurements of glacier retreat obtained from air photographs suggests that the thickness of the glacier snout is decreasing at a rate between 1.0 and 3.8 m/year (Table 7). The large range in the estimated decrease of ice thickness is due in part to the variable length of time between photo surveys and also due to the conservative approach used to estimate elevations of the ice surface during the glacial maxima.

The snout of Fyles Glacier is well exposed to solar radiation and glacier ablation is probably high. A recent rate of thinning, 2.2 m/yr, determined from measurements of ice surface elevation in 1984 relative to the equivalent position on the 1964 topographic map (see Map 3), is perhaps the critical value to use in assessing the future size of jökulhlaups and lake levels. This rate appears to be about the same as those measured for roughly comparable glaciers located elsewhere in the Coast Mountains in a similar time period (Muller, 1977; Tupper et al. 1984).

TABLE 7 RECENT THINNING RATES ON FYLES GLACIER TERMINUS
(by comparison of ice position to contour lines)

Domain Estimated (see Fig. 3)	Time Interval (A.D.)	Estimated Ice Thinning for Time Interval feet (metres)	Thinning Rate m/yr
Western Snout [1]	1856/1904 to 1964	500' (150 m)	1.4 to 2.5
	1964 - 1978	100' (30 m)	2.1
Northeastern Front	1856/1904 to 1964	360' (110 m) [2]	1.0 to 1.8
	1964 - 1978	175' (53 m)	3.8[3]
Eastern Snout in Ape Lake	1964 - 1984	141' (43 m) [4]	2.2

Note 1. The values are maximum rates at the tip of the glacier which obviously are less farther upslope on the glacier.

Note 2. Unlike the western snout, this value is based on a profile of ice extent projected upglacier ca 1500 metres from the climax position of the ice edge. It is assumed that the ice profile for the climax extent replicates the 1964 ice surface as taken from the topographic map.

Note 3. As there is little difference between the 1978 and 1984 ice front near the lake edge, the thinning rate may be as low as 2.7 m/yr.

Note 4. Assumes average ice thickness above lake/ice interface of 25.0 m as approximately measured in 1984 during field observations.

4.3.4 Channel Cross-section Surveys

Immediately downstream of Fyles Glacier, the Noeick River flows on a wide outwash plain. Prior to the flood the outwash plain contained a small lake which was subsequently filled with sediments. About one kilometre downstream from the ice margin, the channel is constricted by erosion resistant morainal deposits forming a control section (Plate IVA). Data obtained in surveys of three cross-sections in the control section are presented in Appendix 3. Large blocks of ice, some more than two meters in diameter, on a terrace above the river channel suggest water depths exceeded 5 m in this section.

Rough measurements of the channel cross-section were also obtained further downstream at about km. 35.5 (Plate VIIA). These data are also included in Appendix 3.

4.3.5 Floodplain Measurements and Observations

Measurements and observations of the changes in the floodplain are mostly limited to photography because of time constraints and the large area affected by the flood. Some of the effects of the flood are shown in Plates V through X. Additional information is shown on Map 2.

In the vicinity of Purgatory Lake (Plates VIA and B), the local gradient averages 0.085 m/m. Large volumes of sediment derived from undercutting the Purgatory moraine (Plate VIA and Plate VIIA) were deposited in the channel where local gradients began to decrease. At one location near Purgatory Lake, sediment between two and three meters deep was deposited on top of recently growing vegetation. Further downstream in the vicinity of km. 35.5 where the local gradient was about 0.017 m/m (Plate VIIA) boulders with b-axes up to 150 cm in diameter were found on top of young conifer trees (also Plate VIIB).

5.0 RECONSTRUCTION OF EVENTS

5.1 Conditions Preceding the Draining of Ape Lake

During the summer of 1984, Ape Lake continued to drain eastward over a bedrock sill into Ape Creek and the Talchako system. Photographs of the western snout of Fyles Glacier in the upper Noeick valley show that by August, glacier meltwater had formed a large tunnel where water drained from beneath the glacier. Although some sub-glacial leakage from the lake may have been taking place at this time, the amount was probably very small.

Prior to the draining of Ape Lake, the area was subject to two major storms. The first storm on September 16-17 resulted in 82.9 mm of rain at Bella Coola, equal to about 63 % of the long-term average rainfall for the month of September. The Ape Lake basin probably received a similar amount of rain during this storm.

In a second major storm, Bella Coola recorded 102 mm of rain over a five day period between October 6-10. This particular storm was widespread and resulted in heavy precipitation and large floods throughout southwest British Columbia. An isohyetal map of total precipitation over the five day period during this storm indicates that total precipitation at Ape Lake could have been 150 mm or more (Map 1). Some meteorological stations recorded heavy rainfall on October 12 and 13 but this storm was not as widespread as the one over the Thanksgiving weekend.

While such large amounts of precipitation could have been important in developing and enlarging englacial and subglacial drainage conduits, there is no evidence to indicate that either storm was instrumental in triggering the drainage of Ape Lake. It is possible, as discussed below, that the lake made use of the internal drainage network in Fyles Glacier which had developed earlier in the year.

5.2 Transition from Stable to Cyclic Drainage

The presence of large, mature conifer trees up to 300 years old on the floodplain of the Noeick River suggests that no outburst floods of a magnitude similar to the 1984 event have occurred in at least 300 years. Also, as discussed below in section 5.3, the evidence suggests that the end moraine in Ape Lake formed a continuous dam across the bottom of the lake basin before the lake drained, a condition which is unlikely if the lake had drained since the formation of the moraine. Like other ice dammed lakes which were once stable (see section 1.2), Ape Lake appears to have been stable for several centuries until gradually changing ice conditions led to the sudden draining of the lake.

One explanation for the transition from stable lake levels to cyclic drainage (assuming that the sequence will repeat, see section 6.2) is that before October 1984, ice pressure always exceeded water pressure because the spillway at the east end of the lake prevented water from rising to the level necessary for flotation of the dam. The simplest explanation for such a transition is given by Thorarinsson (1939); subglacial drainage from the lake becomes possible when hydrostatic pressure of water in the lake exceeds the ice overburden pressure in the region of the 'seal'. Glacier retreat (Table 3) and ice ablation (Table 7) has steadily lowered the height of the dam formed by Fyles Glacier and flotation may have occurred in October, 1984 when Ape Lake was full (El 1395 m a.s.l.). Björnsson (1975) and Nye (1976) favor such a mechanism for controlling the drainage of Grimsvotn in Iceland and Clarke (1982) uses a similar model to explain the drainage of Hazard Lake, Yukon Territory. Questions remain about the maximum thickness of the ice seal and the role, if any, that pre-existing englacial/subglacial drainage tunnels played in controlling Ape Lake water levels.

Assuming a density of ice of 900 kg/m^3, Thorarinsson's model (1939) predicts that drainage will occur when the depth of water is approximately 9/10 of the maximum thickness of ice. Nye (1976), however, noted that in the Grimsvotn jökulhlaup, drainage begins when lake levels are 20 m below the level predicted by Thorarinsson's model. Nye attributed this difference to a reduction in ice overburden pressure which arises because the submerged ice does not immediately attain isostatic balance and so acts on the grounded ice at the seal as a buoyant inverted cantilever.

The maximum depth of water adjacent to Fyles Glacier measured in August, 1984 was approximately 58 m (Map 3). Because of practical limitations on the density of echo sounding profiles and the very uneven basin topography near the ice margin, the measurements adjacent to the glacier may underestimate the true maximum depth of water and thus the potential hydrostatic pressure available to lift the ice. During field observations after the drainage it was noted that the glacier is retreating over a steep bedrock knob with the tunnel mouth located in a deep hole on bedrock. Thus it is conceivable that maximum water depths at the edge of the ice could be 5-10 m higher than those measured in August, 1984. Unfortunately the hazards posed by the vertical ice and time constraints in October 1984 did not permit an accurate measurement of maximum depth of water which would have existed when the lake was full.

Based upon the elevation of the ice surface above the tunnel mouth and the elevation of the tunnel exit, the maximum ice thickness is about 95 m (assuming no major depression beneath the ice between the tunnel portals). At the ice-water interface, the ice was measured to be about 82 metres thick. Using Thorarinsson's model, ice thicknesses of between 82 and 95 m would require water depths between 74 and 86 m to cause flotation of the ice. However it is doubtful whether maximum water depth exceeded 65-70 m adjacent to the ice so that flotation could not occur unless another mechanism was operative.

The difference between the water depth required to effect flotation and the estimated water depths available range between 4-21 m. In the case of Grimsvotn, Nye (1976) estimated that the buoyant effect of the submerged ice could result in an effective reduction in the ice overburden pressure of 2 bars, equivalent to approximately 20 m of water. If such an effect occurs at Ape Lake, it would be sufficient to account for the difference between water levels required to lift the ice and those which apparently existed at the time of drainage.

The remaining question is whether the drainage of Ape Lake was precipitated by the lake 'capturing' the subglacial and englacial drainage network of Fyles Glacier which had developed through the spring and summer of 1984. It is generally accepted that meltwater flows in networks of englacial and subglacial conduits in the ablation zone (Stenborg, 1969; Röthlisberger, 1972; Shreve, 1972) although their precise location and geometry have not been elucidated. Studies of water movement, using dye and salt tracers in moulins and crevasses in the ablation zone of many glaciers indicate a well developed drainage network since water arrives at the glacier margins very rapidly (Stenborg, 1969).

The variation in englacial water levels during the summer and autumn may be a secondary factor in controlling when the lake can drain subglacially. Mathews (1964) investigated water levels in South Leduc Glacier and found that they were moderately stable from day to day, although at times they showed slight diurnal fluctuations, interrupted by irregular and catastrophic surges, particularly during periods of rapid snow melt and heavy rain. Hodge (1976) found similar conditions on South Cascade Glacier where the surface defined by the bore-hole water levels during the summer was consistently about two-thirds of the ice thickness. Hodge also found that the water levels in the boreholes decreased throughout the summer until about the end of the ablation season when the trend reversed and water levels rose during October and November. This seasonal variability in water level agrees with observations made by Mathews (1964) and Vivian and Zumstein (1973) and may be caused by the gradual development of conduits through the summer which reach maximum dimensions in late fall and gradually reseal through the winter. Thus cavities and conduits within the glacier may have little or no water in late September and early October due to the high hydraulic capacity of the drainage network.

The height and seasonal variability of water levels in Fyles Glacier may be very similar to that which has been observed in other glaciers. During the summer, high water levels in the glacier prevents water in the lake from connecting with the internal drainage network through collapse of ice cavities or conduits. As water levels decrease towards the end of the ablation season, the hydrostatic pressure of the lake is not offset by hydrostatic pressure of water in the glacier, so that the hydrostatic pressure of water in the lake may cause catastrophic collapse of cavities creating a hydraulic connection to the existing drainage network. This may partially explain why so many lakes drain in the autumn months as noted by Mathews (1973).

Unfortunately there are no data to either support or refute this hypothesis for the drainage of Ape Lake except for photographs taken before and after the flood which show that the jökulhlaup tunnel exit is in the same location as a large meltwater tunnel which had formed earlier in the season. However, the fact that the meltwater discharge and lake discharge drained from the same location may be caused entirely by bedrock topography and have nothing to do with drainage networks which had formed prior to the flood.

5.3 Timing and Duration of the Jökulhlaup

Based upon eyewitness accounts (see Appendix 2), pre- and post-flood photographs and field observations of strandlines in the empty basin and erosion of the end moraine crossing the Ape Lake basin, a tentative chronology of the flood can be established.

A careful aerial reconnaissance of Ape Lake by Crown Forest Industries Ltd. on Wednesday, October 17, seven days after the major Thanksgiving storm, indicated nothing unusual was happening. We also know that the Wilderness Airline pilot did not notice anything unusual such as dramatically lower water levels or an exposed moraine in the centre of the lake during the regular scheduled flight which flew over the lake at about 1300 hours on Saturday, October 20. Eyewitness accounts however, indicate that at km 35, the flood was well under way by 1100 hours on this day although flood debris had not reached South Bentinck Arm at this time. Observations of the lake by the Wilderness Airline pilot and eyewitness accounts of the flood at km 35 place some restrictions on when the flood began and how fast the lake was draining at 1300 hours. It is assumed that between one and two hours is required for the water to travel from the lake downstream to km 35.

During the week of October 13-19, small amounts of water may have been seeping beneath Fyles Glacier but inflow to the lake was more than sufficient to maintain stable lake levels. Significant leakage (ie 10 m^3/s or more) probably began by Thursday, October 18, gradually forming a tunnel beneath Fyles Glacier. By 0700-0900 hours on the morning of October 20, the tunnel was large enough to convey large volumes of water from the lake, causing high water levels downstream at km 35 around 1100 hours. During this initial stage, lake levels were declining at an increasing rate and discharge continued to increase, probably reaching a maximum at the tunnel exit around 1100-1200 hours (Figure 4).

Eyewitnesses in the Noeick Valley observed that water levels were more or less stable between 1100 and 1630 hours at kilometre 30-35. Between 1630-1730 hours, Mr. K. Danielson drove down the forestry access road past kilometre 35 which was now flooded (0.3 m above road level). When he returned about 1830 hours, the road was flooded to an estimated depth of 1.2 m. A photograph taken at this time indicates that maximum water levels had been about 0.45 m higher a short time earlier. Subsequent photographs after the flood indicate that maximum water levels eventually rose about 0.3 m above the previous maximum height at this location (Plate VIII). Eyewitnesses attributed some of changes in water level to effects caused by the river shifting its channel on the floodplain.

By 1300-1400 hours, lake levels probably had declined by up to 10 m, so that the end moraine across the basin became exposed, separating the lake into two basins (Figure 4). The west basin continued to drain beneath the glacier while the moraine impounded water in the east basin for a short time (1-3 hours) as indicated by strand lines which developed during this interval (Map 3).

As the west basin rapidly emptied, there was less hydrostatic pressure to resist the pressure of the water impounded in the

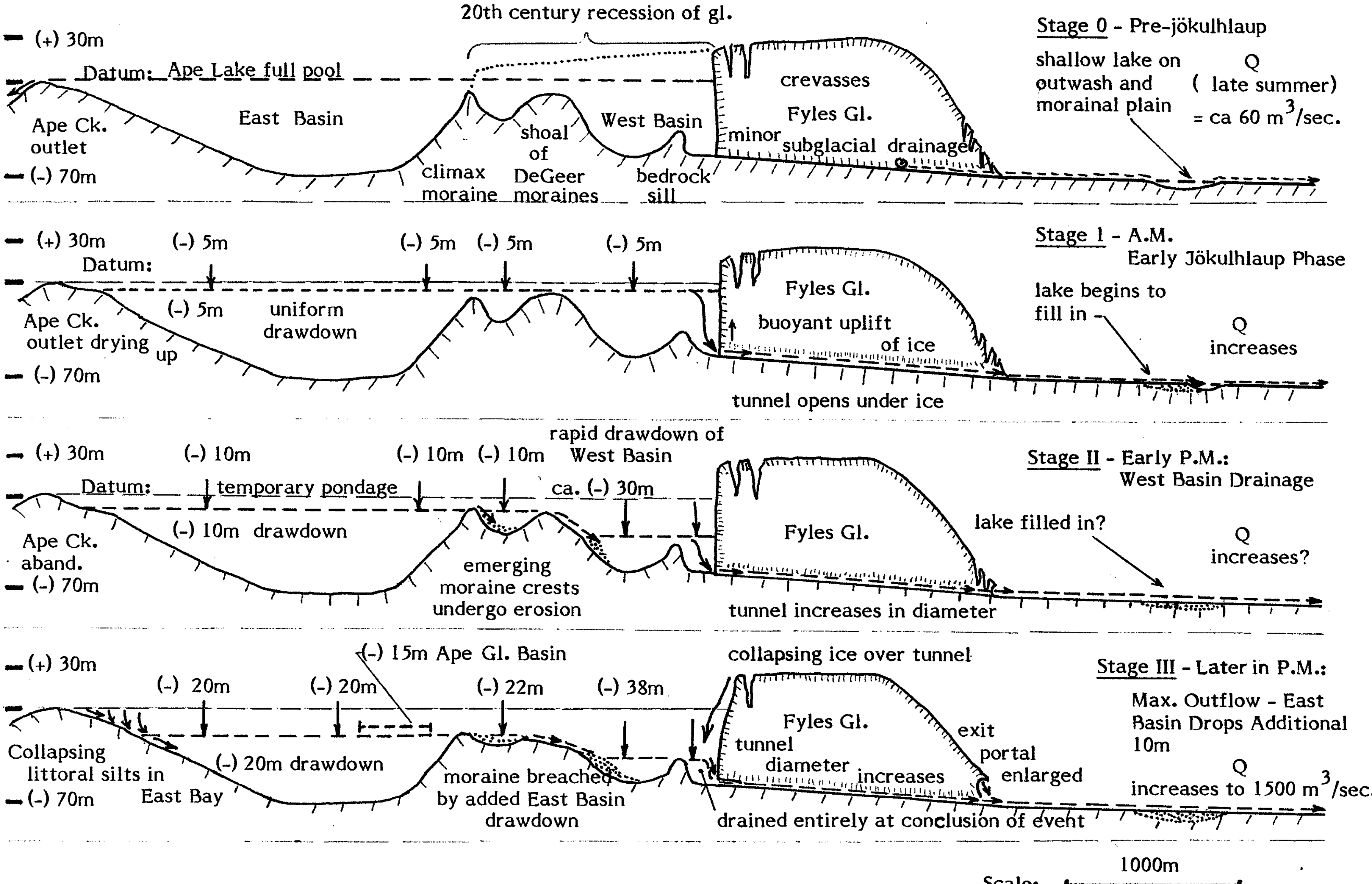


Figure 4 Inferred Sequential Drainage of Ape Lake into Noeick Watershed During 20 October, 1984 Jökulhlaup.

east basin by the moraine. Consequently the saturated moraine failed, probably through piping or overtopping, at about 1530–1630 hours. With the failure of the moraine dam, a large volume of water (Table 1) was released to flow through the tunnel formed earlier in the day (Figure 4).

At 1730 hours, the renewed discharge from the east basin began to arrive at km. 35, raising water levels by nearly two metres in a two hour interval. Up to this time, the river had maintained its normal course, with water entering and exiting Purgatory Lake at km 36 in the pre-existing channel. With the arrival of a new flood wave, water flowed across the floodplain, bypassing Purgatory Lake and continued downstream to tidewater. Peak discharge at the river mouth probably occurred about 2000 - 2130 hours, roughly synchronous with the occurrence of high tide at 2115 hours (Canadian Hydrographic Service, 1983). Consequently water levels at the mouth of the Noeick River were increased further, flooding nearby terraces and the airstrip to a depth of 0.75 m.

The lake was probably close to empty by 2200 hours on Saturday, October 20 with water levels at km 35 observed to be near normal by 1100–1130 hours the following morning (Appendix 2).

5.4 Estimates of Discharge Magnitude

Based on the preceding reconstruction of the flood, it is believed that most of the lake volume drained within 24 hours and possibly in 20 hours or less. During the early stages of the outburst flood, subglacial leakage probably did not exceed normal inflow to the lake. Assuming that a total volume of 45.8×10^6 m^3 of water drained from the lake in 20–24 hours, then the average discharge at the tunnel exit would have been between 540 and 650 m^3/s.

Clague and Mathews (1973) reviewed the literature on the maximum discharge of jökulhlaups and proposed a formula to estimate the peak discharge. Their formula is:

$$Q_{max} = 75\ V_{max}^{0.67}$$

where Q_{max} is the maximum flood discharge in m^3/s and V_{max} is the available water storage (in $m^3 \times 10^6$). The Clague-Mathews formula was modified by Desloges (1984) to include data from Hazard Lake, Yukon Territory and to exclude the Missoula flood. The formula then becomes:

$$Q_{max} = 105.6\ V_{max}^{0.58}$$

Using the modified Clague-Mathews formula and a volume of 45.8 x 10^6 m^3, the peak discharge is estimated to be 970 m^3/s. A plot of Q_{max} vs V_{max} shows that the estimated peak discharge for Ape lake is within the same standard error range as other floods used to construct this relationship (Figure 5, from Desloges, 1984).

As described in Appendix 3, other methods can be used to estimate the magnitude of the peak discharge. The maximum instantaneous discharge at the control section, one kilometre downstream of the tunnel exit, is estimated to be about 1500 m^3/s (Appendix 3, Table 10). At km 35, the peak discharge was estimated by two independent methods. The peak flows were probably in the range 735-896 m^3/s (Table 8 and 10) but could have been as high as 945-1050 m^3/s. The peak discharge at km 35 would be lower than that upstream near the tunnel exit because of channel storage and attentuation of the flood peak. For comparison, the maximum mean daily discharge of the Noeick River at the mouth with a recurrence interval of 100 years is estimated to be about 460 m^3/s. No recurrence interval, however, can be attached to estimates of the magnitude of the

Qmax VERSUS VOLUME OF DRAINAGE FROM GLACIER-DAMMED LAKES

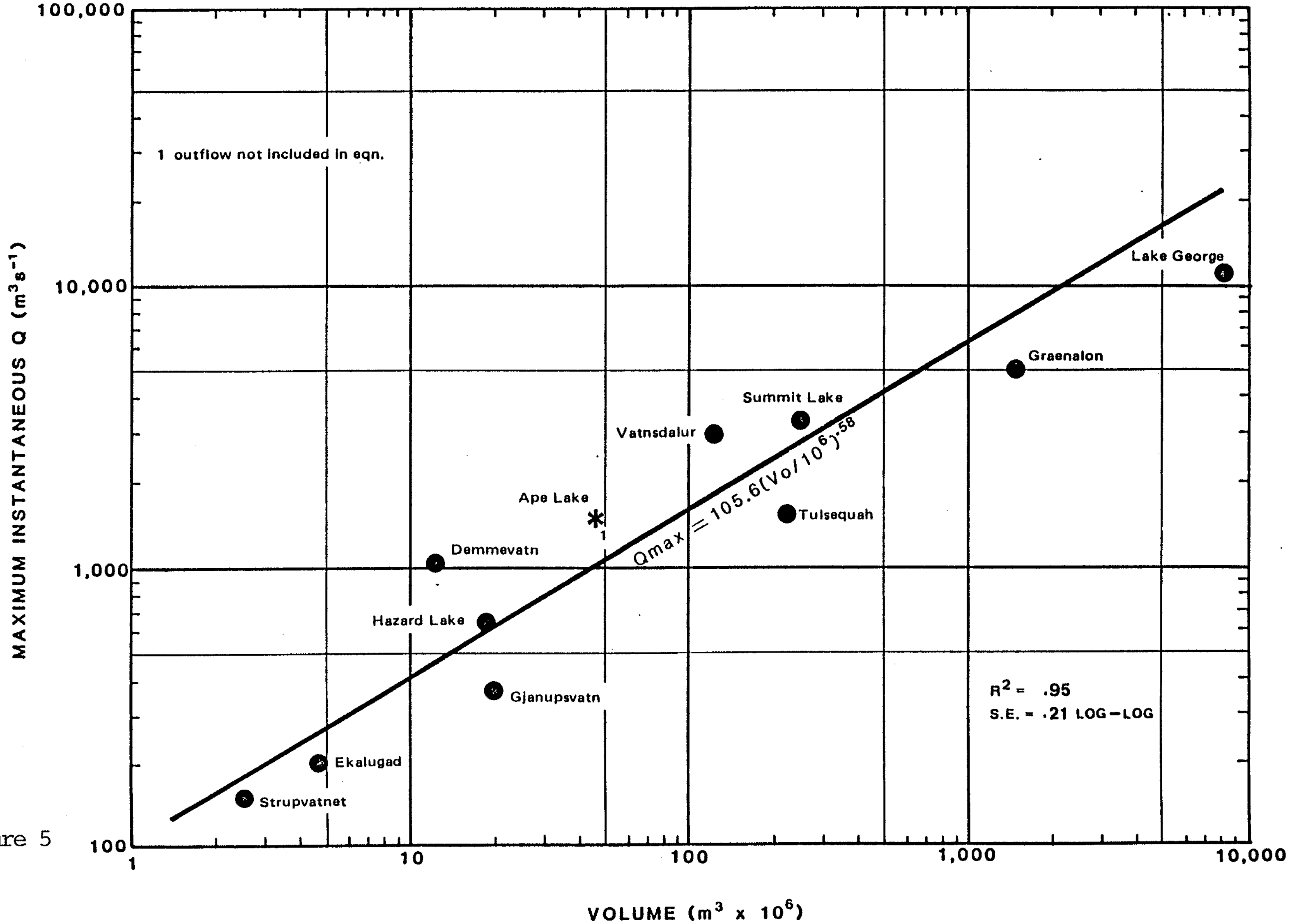


Figure 5

outburst flood.

5.5 Downstream Effects

The rapid discharge of approximately 45.8×10^6 m^3 of water down the Noeick valley resulted in extensive changes in the floodplain and destruction of property and resources downstream of Fyles Glacier. The principal effects of the flood were:

1. Extensive erosion and scour of the river channel along both banks from the snout of Fyles Glacier downstream to tidewater. In the broad floodplain reach (km 14.5 to 27.3, Map 2) however, the main effect of the flood was a reworking of floodplain sediments accompanied by overbank siltation (Plate IXC).
2. Erosion of segments of the forest access road and damage to two bridges spanning the Noeick River.
3. Destruction of large plantations of recently planted and spaced Sitka spruce, Douglas fir and Western Red Cedar trees.
4. Scour and deposition of sediment in areas normally used by spawning salmon which probably resulted in the destruction of most or all of the eggs deposited in the gravel earlier in the year.

The flood may also have created other effects which are much more difficult to quantify. One such effect is the amount of debris flushed into South Bentinck Arm which made boat access impossible for several days. This debris may form semi-submerged 'deadheads', a hazard to boats, floatplanes and fishing gear.

5.5.1 Geomorphic changes

While evidence of geomorphic changes due to erosion, transport and deposition of sediment occurs throughout the length of the

Noeick valley, it is difficult to quantify the changes in any meaningful way with the data available. A brief description of the effects of the flood at several locations however can provide examples of the type and magnitude of geomorphic changes. Notes on such observations and of other geomorphic changes are shown on Map 2.

The effects of the flood were particularly noteworthy in the vicinity of Purgatory Lake. Immediately upstream of the lake, flood waters undercut the high lateral moraine formed by Purgatory Glacier. Eyewitnesses report watching large volumes of the bank, which is 30-50 m high (Plate VIA), falling into the flood waters to be transported downstream. A large volume of sediment was deposited immediately downstream as the slope decreased near Purgatory Lake. Up to 3 m of sediment was deposited upstream of the lake as evidenced by the depth of burial of vegetated surfaces. Immediately downstream, the river scoured and greatly enlarged its channel. Plate VII shows the large cut banks on either side of the river channel, here about 70 m wide. Maximum water depth at this site was at least 4-5 m (Appendix 3).

5.5.2 Forestry Roads, Bridges and Plantations

The forest access road connecting the Nusatsum and Noeick valleys was completed by Crown Forest Industries Ltd. in September, 1984. The access road was constructed on parts of the Noeick floodplain. As a result, in locations where the road was close to the river channel, lateral erosion by the river undercut the banks leading to the destruction of the road. About one kilometre of road was washed out during the flood. Specific locations where the road was damaged are shown on Map 2.

In some locations, the main road can be easily rebuilt either by moving the road alignment slightly or by placing heavy rock fill where the road surface has been eroded. At other locations however, notably at km 18 and 27.5, extensive blasting of bedrock outcrops will be required to reconstruct the road.

In addition to the main access road, several kilometres of local spur roads were destroyed or severely damaged and the main road on the south side of the river was damaged in several short locations (Map 2). Currently there are two bridges across the Noeick River at '5-mile' and '10-mile' (km. 7.0 and 15.5, Map 2). The 10-mile bridge and associated bridge approaches will have to be rebuilt while the 5-mile bridge can probably be repaired (Plate XA and B). The cost of road and bridge repair has been estimated at around $200,000 (Anon, 1984b).

Large areas of the Noeick River floodplain were logged and subsequently replanted with Sitka spruce, Douglas fir and Western Red Cedar. Up to 200,000 of these recently planted trees were destroyed in the flood (R. Lenci, pers. comm.) between km 14 and 35 (Map 2). Some trees were uprooted, some broken by the impact of boulders and others bent over and stripped of foliage by the force of water and then covered with mud and silt.

In addition to the destruction of tree plantations, the flood undermined numerous mature trees which had not been logged along the river bank. At one location, upstream of the 10 mile bridge, approximately one hectare of mature Sitka spruce was destroyed by the flood. Farther upstream between km 30-38, the river undermined its banks so that trees were observed to fall into the river every few minutes, destroying an unknown amount of mature forest. The total damage to roads and and tree plantations, not accounting for the time lag to establish new forest growth, is estimated be be one million dollars (Anon., 1984a).

5.5.3 Fisheries

Available evidence indicates that anadromous fish are unable to negotiate the lower canyon and are thus restricted to the lower 10-12 km of the Noeick River. The upstream migration of anadromous fish on the Smitley River is blocked by a falls approximately 0.5 km upstream from the confluence with the Noeick River (Map 2).

Four species of anadromous fish are known to spawn in the lower Noeick River. About 500 pink (Oncorhynchus gorbuscha) and 500 chum (O. keta) are known to have spawned in the Noeick River during September, 1984. Although upwards of 100 chinook (O. tshawytscha) normally spawn in August and September, none were observed in 1984. According to K. McGwiny (Pers comm.) the jökulhlaup probably coincided with the return of the adult coho (O. kisutch). In recent years the number of adult coho has ranged between 200 and 1500 adults each year. No coho were observed in the Noeick River after the flood so their fate is unknown.

The extensive erosion and transport of sediment which was observed further upstream almost certainly occurred on the lower reaches of the Noeick River. The high rate of sediment transport during the flood probably destroyed most eggs either by scour or by burial.

The impact of one flood on anadromous fish stocks is low to moderate. Pink salmon spawn on a two year cycle with a major run in even years (ie. 1984) and a minor run in odd years (ie. 1985). Stocks of pink salmon may require considerable time to recover from the effects of this one flood since the major run occurred prior to the flood and most eggs deposited in the Noeick River were probably destroyed. In contrast, chum salmon, with a spawning cycle which varies in length between four and six years, should recover much more rapidly. Assuming that the flood destroyed the spawn of 500 pink and 500 chum, then the number of fish available for the commercial fishery will be reduced by about 1000 of each species (Dr. W.E. Ricker, pers. comm.). In the long-term, the Noeick River stocks will probably be reduced further because of the high probability of more, large floods in the near future (see section 6.0).

6.0 FUTURE OF APE LAKE

6.1 Tunnel Closure and Lake Refilling

6.1.1 Mechanisms of Tunnel Closure

Following the drainage of the lake, the magnitude of processes forming and maintaining the tunnel beneath the glacier are greatly reduced and the tunnel gradually begins to close. Several processes, operating at different time scales, may affect the rate of tunnel closure. Four processes which cause tunnel closure with varying degrees of importance are:

1. Collapse of the ice front and sealing of the tunnel entrance with ice blocks.
2. Ice accretion on tunnel walls arising either from accumulations of frazil and pan ice or freezing of water if the ice is below the pressure melting temperature.
3. Plastic deformation of the ice leading to tunnel closure.
4. Closure due to compression of the tunnel caused by forward movement of the glacier.

Where a glacier flows into a lake or other body of water, the morphology of the terminus changes in response to the hydrostatic pressure of the water. The ice will deform until it is in equilibrium with the water, forming a more or less vertical cliff below water level. No hydrostatic pressure resists the forward movement of ice above water level so it continues to move forward until the unsupported ice fractures and collapses under its own weight. Once the lake drains, the vertical ice wall is no longer supported by the hydrostatic pressure of the water. As a result, large portions of the ice face collapse so that ice blocks and debris may block the tunnel entrance. In the case of Fyles Glacier, large portions of the glacier collapsed at both ends of the tunnel but only the size of the tunnel exit appeared to be significantly reduced by the collapsed ice (Plates IIIA, IIIC). Future collapse of the ice face in the spring and summer may result in a decrease in tunnel dimensions at the mouth.

With the onset of sub-zero temperatures, frazil ice generated in the turbulent flow over the steep bedrock controlled section of the west sub-basin may accumulate at constrictions and irregularities in the tunnel, helping to reduce tunnel dimensions. Slush and ice pans from the lake basin could also contribute to a reduction in tunnel dimensions. If the ice is below the pressure melting temperature, water could also freeze on the walls of the tunnel further reducing tunnel dimensions although this is unlikely to occur beneath Fyles Glacier.

While the lake is draining, the walls of the ice tunnel are supported partially by the hydrostatic pressure of the water. Once the lake empties, water is no longer available to maintain the tunnel at its full size and the ice will begin to deform plastically due to the weight of overlying ice. Nye (1953) provides a theoretical analysis of the rate of tunnel closure due to plastic deformation. Hooke (1984) has extended Nye's analysis to incorporate the effects of melting caused by conversion of potential energy of water flowing through a partially filled tunnel to thermal energy. Hooke shows that under certain conditions, the rate of melt due to flowing water is sufficient to offset the rate of closure due to plastic deformation so that the water is able to maintain a tunnel, albeit perhaps with a significantly smaller diameter.

The analyses of Nye (1953) and others assume that the compressive strain rate normal to the tunnel axis is negligible and that the primary mechanism for tunnel closure is through plastic deformation due to shear stresses. Where the tunnel is oriented across the glacier, compression of the tunnel axis caused by glacier movement may be the most important mechanism initiating tunnel closure. Longitudinal velocities at the surface are usually highest near the centre of the glacier and least near the ice margins, depending on the depth of ice. Basal sliding or slip (Sharp, 1954) normally accounts for about 50 % of the total surface velocity of typical valley glaciers in their thicker parts, the remainder being due to internal deformation within the ice (Shreve, 1961; Savage and Paterson, 1963). Kamb and LaChapelle (1964) measured ice velocities in a tunnel at the bed of Blue Glacier and found that most of the differential movement occurred in the lowest 50 cm of ice. Velocities 150 cm above the bed were equal in magnitude and direction to surface velocities. These results suggest that the tunnel beneath Fyles Glacier will be subject to large, differential shear stresses that will lead to the relatively rapid deformation and closure of the tunnel by compression.

Where the glacier lies across a valley with forward movement along the longitudinal axis limited by a valley wall, compression of the glacier snout will increase the rate of tunnel closure over that due solely to normal forward movement of the glacier. This condition occurs at Ape Lake where the forward movement along the longitudinal axis of Fyles Glacier is limited by the opposing valley wall. In response to frontal compression, Fyles Glacier has spread laterally along the valley axis, particularly on the western side. Until recently, the surface of Fyles Glacier was much higher up the valley wall with a major lobe of ice extending down the Noeick valley. With the general retreat of Fyles Glacier, the surface elevation has decreased and the glacier just fills the valley so that frontal compression is probably lower than that which occurred in recent decades.

6.1.2 Rate of Tunnel Closure

The rate of tunnel closure can vary widely depending on the thickness, velocity and temperature of a glacier and the principal mechanism closing the tunnel. In 1965, Summit Lake began to refill about 1 1/2 months after draining beneath Salmon Glacier while in 1967, the lake began to refill within 3 days of draining (Mathews, 1973). Hazard Lake, located in the St. Elias Range Yukon Territory first drained in 1975 through a subglacial tunnel (Clarke, 1982). The tunnel remained open with the lake basin empty through the summer of 1976. During the winter of 1976-77 the tunnel resealed. The lake again filled to its maximum level by July 1977, then drained subglacially around 2-5 August, 1977. Since that time the lake has continued to fill and drain on an annual basis (Clarke, 1982). Following the drainage of Strupvatnet in Norway in 1957, Aitkenhead (1960) reported that the tunnel remained open for at least a month and possibly as long as a year before closing although Whalley (1971) suggests that the low lake levels observed in 1958 were due to a second jökulhlaup prior to the field observation. Based on the behaviour of outburst floods documented in the literature, most subglacial tunnels close soon after the lake drains. One exception to this is reported by Mathews (1965) where Strohn Lake near Stewart, B.C. drained annually through a subglacial tunnel between 1958 and 1962. The tunnel remained open in 1963 and 1964, either because the ice was so thin that plastic deformation was negligible or because the velocity of the glacier snout was very small. As of January 22, 1985, no observable refilling of Ape Lake has occurred (T. Bonner, Wilderness Airlines, pers. comm.).

Of the four processes which may lead to tunnel closure the most plausible mechanism to seal the tunnel beneath Fyles Glacier is the result of glacier flow coupled with compression due to glacier movement which is limited by the valley wall. Closure of the tunnel entrance by ice collapse, while possible, is probably fortuitous and highly unpredictable. Similarly, frazil and pan ice may accumulate inside the tunnel, thus reducing tunnel dimensions, but it is thought that this process is unlikely to be important in completely closing the tunnel.

Following Nye (1976), the rate of tunnel closure due to plastic deformation can be computed using the equation

$$\frac{\partial S}{\partial t} = - K_o S (p_i - p)^n \qquad (1)$$

where S is the cross-sectional area of the tunnel, p_i is the ice overburden pressure, p is the pressure of the water in the tunnel, n is the exponent in the flow law of ice, here taken as n=3 and $K_o = 2B/n^n = 1.6 \times 10^{-25}\ Pa^{-3}s^{-1}$. The ice overburden pressure $p_i = \rho_i g h_i$ where g = acceleration due to gravity, 9.81 m/s^2, ρ_i is the density of ice, 916 kg/m^3 and h_i is the maximum thickness of ice over the tunnel. Once the lake is empty p is assumed to be zero for simplicity (the tunnel may contain water at atmospheric pressure but this will not affect the rate of plastic deformation at the beginning).

By integrating equation (1), it is possible to estimate the decrease in tunnel cross-sectional area for a specified time interval. Assuming that the area equals S_o at t=0, then the integration gives

$$S = S_o \exp(-K_o p_i^n t) \qquad (2)$$

Using a tunnel radius r= 2.5 - 3 m and a maximum ice thickness h = 95 m, the time required for the cross-sectional area to decrease by 90 % is 268 days and for 99% decrease in cross-sectional area 535 days or 1.47 years would be required.

If water continues to flow through the tunnel, the rate of melt resulting from the conversion of potential energy to thermal energy becomes important. The rate of melt, m can be estimated using the equation (Hooke, 1984)

$$m = C_2 \, Q^{3/5} \sin^{6/5} A \qquad (3)$$

where m is the rate of melt in m/s, Q is the tunnel discharge in m^3/s, A is the slope of the subglacial tunnel and $C_2 = 3.73 \times 10^{-5}\ m^{-4/5} s^{-2/5}$, a constant. (Note that equation 1 and 3 can be combined to form a non-linear equation which must be integrated by numerical methods.) Assuming a slope of 0.5^{o}, for a discharge of 1 m^3/s, the rate of melt m = 12 mm/day while for a discharge Q = 0.1 m^3/s, m = 3 mm/day. During the summer the potential discharge is probably sufficient to maintain an open tunnel, although with dimensions considerably smaller than the tunnel created during the outburst flood of October,1984.

The preceding analysis indicates that the subglacial tunnel beneath Fyles Glacier is unlikely to reseal if plastic deformation is the only operative process, so if the tunnel is to close it must be due to the forward movement of the glacier sliding on its bed. Measurements of the average ice velocity of Fyles Glacier have not been made so values must be estimated using data collected from other glaciers. The Fyles Glacier is similar to South Cascade Glacier in Washington which has been studied extensively by Meier and Tangborn. The South Cascade and Fyles glaciers originate at elevations of 2100 and 2680 m, with their respective termini at 1600 and 1320 m and corresponding lengths of 3.4 and 9.2 km. The average surface slope of Fyles Glacier is slightly larger than the average slope of South Cascade Glacier. Meier and Tangborn (1965) found that ice velocities of South Cascade Glacier ranged up to 20 m/year. Based upon the similarities between the two glaciers, Fyles Glacier is estimated to have a minimum velocity of 5 m/yr and a probable velocity between 10 and 20 m/year. This is similar to the range of flow velocities measured by Tupper et al. (1984) on Wedgemount Glacier in Garibaldi Park in zones of compression and extension. However, Wedgemount Glacier is smaller than Fyles Glacier and does not have a lobate terminus so data from Wedgemount Glacier may be less applicable to Fyles Glacier. By comparison, Mathews (1964) measured the surface velocity of the much larger South Leduc Glacier north of Stewart and found that over a 35 day period the velocity averaged 7.0 cm/day (25.6 m/yr) and 9.5 cm/day over a 17 day period in 1961.

Using average surface ice velocities of 2.7-5.5 cm/day (10-20 m/yr) and assuming movement is normal to the tunnel axis, a tunnel 500 cm in diameter would reseal in about 100-200 days. Assuming a winter discharge of 0.1 m^3/s, the rate of ice melt would be about 3 mm/day or 10 % of the rate of tunnel closure expected due to ice movement. This suggests that the tunnel will seal even with water flowing through the tunnel. If the tunnel is not sealed by April-May of 1985, increased summer runoff may be sufficent to maintain the tunnel but it would likely seal the following winter.

6.1.3 Rate of Refilling

The area contributing runoff to Ape Lake is approximately 38 km^2, of which about 20 km^2 is covered by glaciers. The average annual discharge required to fill a volume of 45.8×10^6 m^3 (Table 1) is 1.45 m^3/s, equivalent to a unit runoff of 1200 mm/year, assuming there is no leakage beneath the glacier. This is only about 60 % of the average annual runoff measured at the gauging station on the Nusatsum River and less than 50 % of the estimated annual runoff for the upper Noeick River basin. Thus under average conditions, ample runoff volume is likely to be available to refill the lake to its former level in less than 6 months once the tunnel is closed. About 80 % of the annual runoff, due to snowmelt, glacier melt and rainfall occurs in the months May through October, so that if the tunnel were sealed by the beginning of May, the lake probably would be full by late August or early September.

6.2 Future Jökulhlaups

Although more jökulhlaups are expected from Ape Lake, the exact behavior is difficult to predict until the lake refills and empties through at least one more cycle. Nonetheless, a preliminary assessment of the future behavior of Ape Lake can be made based upon the behavior of similar ice-dammed lakes which drain from time-to-time.

As discussed in previous sections, the drainage tunnel beneath Fyles Glacier could be closed as early as April or May, 1985 so that refilling of the lake can commence with the spring runoff. Under average conditions, the surface elevation of the lake should be near 1395 m a.s.l. by late August or early September with water spilling over the bedrock outlet into Ape Creek and the Talchako system.

Depending on a number of factors such as ice thickness, englacial and subglacial drainage channels, lake depth and lake water temperatures, the lake may drain just before or soon after reaching elevation 1395 m a.s.l. or some indeterminate time may pass before the lake drains again. Unless the 1984 jökulhlaup was triggered by unusual conditions, the lake probably will drain subglacially within a year of refilling. Once the sequence is established, tunnel closure, lake refilling and subsequent subglacial drainage would probably occur on an annual cycle. In the first few years, peak outflows from the lake may be larger than that which occurred in 1984 because the east basin can drain continuously now that the climax moraine has been breached (Plate IIB).

If the net mass balance of Fyles Glacier continues to be negative, the glacier terminus will become thinner, allowing the lake to drain when water levels are below elevation 1395 m a.s.l. When this happens, the glacier rather than the bedrock sill at the east end of the lake controls the surface elevation of the lake. As a result, the magnitude of successive flood discharges will gradually decrease as the glacier recedes and the volume of impounded water decreases. For example, the thickness of the ice could decrease to elevation 1395 m in 7 to 25 years based on the rate of thinning shown in Table 7. If however, the net mass balance of Fyles Glacier becomes positive and the glacier begins to advance, the thickness of the ice damming the lake will increase. An increase in ice thickness over that which existed in 1984, possibly as small as 3-5 m, could be sufficient to prevent subglacial drainage.

Based upon the known basin topography near the glacier and the average slope of the tunnel, it is unlikely that water will ever flow overtop of the ice, even if the ice continues to thin. Because maximum lake depths occur immediately adjacent to the glacier, sufficient hydrostatic pressure should be available to initiate subglacial drainage as the glacier recedes.

Assuming that the lake will drain when maximum water depths are approximately 9/10 of the maximum ice thickness of the snout, the lake will drain at successively lower elevations each year as the thickness of the ice decreases. However, in the absence of site specific data, it is difficult to predict the rate at which ice thickness will decrease. The snout of Fyles Glacier is against the opposing valley wall and ice thickness will decrease through vertical downwasting and ablation and through a reduction in the forward movement of the snout.

The potential exists for large outburst floods to continue until lake levels are 10-20 m lower than maximum 1984 levels. Unlike the situation in 1984, the first few floods are likely to have one major peak, possibly larger than either of the peak discharges in 1984, because a channel eroded through the climax end-moraine allows the water in the lake to drain continuously.

Approximately 50 % of the total volume of water which drained in 1984 was contained in the top 10 m of the lake (Table 1). Once maximum lake levels decline by 10 m, the magnitude of floods will decrease and will become very small once maximum lake levels have declined 20 m below the 1984 maximum levels. This also implies that the ice thickness must also decrease by about 20 m so that the lake drains earlier and earlier at successively lower levels. Note however that as Fyles Glacier recedes, the volume of water in the lake increases, since the volume of ice along the lake margin is replaced with a comparable volume of water.

Assuming that Fyles Glacier continues to recede at a rate similar to that which occurred in the period 1964-1984, the potential for large jökulhlaups will probably last for at least another 8-14 years and possibly as long as 100 years. As noted earlier, other glaciers have been advancing since about 1964. It is possible that Fyles Glacier may respond much slower to climatic changes than smaller, nearby glaciers so that Fyles Glacier may yet show signs of advance and thickening of the terminus. An advance of the snout could temporarily prevent subglacial drainage but outburst floods would probably occur again at a later date.

6.3 Management Options

Attempts have been made to reduce or eliminate outburst floods where they pose a threat to transportation corridors or communities downstream from the lake. Because the draining of Strohn Lake near Stewart, B.C. damaged the road and bridge each year, the B.C. Department of Highways attempted to reduce the magnitude of outburst floods by excavating a trench in outwash materials to lower lake levels. A trench 10 m deep was excavated but two months after lake levels reached the elevation of the trench, the lake drained subglacially (Mathews, 1965). Subsequent attempts to lower lake levels by excavating a trench through the ice were also unsuccessful.

Future jökulhlaups on the Noeick River could be prevented by lowering the level of Ape Lake by about 20 m. A permanent reduction of water level could be achieved by one of three different alternatives:

1. Excavation of a channel in the bedrock sill at the east end of the lake where the overflow spills into Ape Creek or
2. Excavation of a channel around the snout of Fyles Glacier so that water can drain directly into the Noeick River.
3. Application of coal dust or cinder to the ice surface to accelerate ablation, thus reducing the ice thickness and indirectly the volume of water stored in Ape Lake.

Of the first two methods, excavation of a channel to connect the lake to the Ape Creek drainage at a lower elevation is the only one which could provide a permanent solution since any advance of Fyles Glacier would affect a channel around the edge of the glacier. However, while it is technically possible to prevent the occurrence of future jökulhlaups at Ape Lake by excavating a channel, it is unlikely to be economically feasible, due to the location and difficult access.

In the third method, thin layers of a dark material such as cinder, ash or coal dust are applied to the surface of the snow and ice to increase the rate of melt. The dust decreases the surface albedo so that a large proportion of the incoming short-wave radiation is absorbed rather than reflected. The absorption of the energy raises the temperature of the dust and the heat is transferred to the underlying snow and ice by conduction. Using an application of dust, the rate of melt can be increased even when air temperatures are below zero, since the energy to melt the snow and ice is derived from short-wave radiation rather than sensible heat from the atmosphere. To be fully effective, dust probably would have to be applied each spring.

Arnold (1961) found that application of up to one kg/m^2 of fine dust increased the rate of melting about five times over untreated surfaces. Arnold also describes experiments where applications of dust have been used to accelerate melting of snow at airports, ice jams and other conditions. The efficacy of dust to accelerate melting of snow and ice has also been investigated in Russia where the technique is used to increase meltwater from glaciers for use in irrigation (Kotlyakov, 1982). Thus the application of coal dust or similar dark material, possibly by cropdusting aircraft, could be used to accelerate the thinning of the Fyles Glacier snout. Conceivably the ice thickness could be reduced substantially in two or three years, with a corresponding reduction in the magnitude of outburst floods. Artificially increased ablation of the glacier snout however, would change the mass balance of Fyles Glacier, possibly affecting the velocity of the glacier and ultimately the amount of ice which must be lost before the glacier no longer dams Ape Lake.

While these solutions are technically feasible, additional work will be required to assess whether they are both necessary and economical. In the meantime, alternate management strategies are required to minimize the potential for either loss of life and property or damage to resources and facilities on the lower river.

6.3.1 Forestry and Logging

The potential for large outburst floods on the Noeick River probably will affect the construction schedule for forest access roads and the harvesting and reforestation program in the valley. The construction of bridges across the Noeick River which could survive one or more floods may be prohibitively expensive. Since bridges across the river may be frequently destroyed it may be necessary to modify normal logging practices. To minimize costs it may be necessary either to postpone logging operations indefinitely in certain areas or to install a temporary bridge across the river to log as much area as possible before the bridge is destroyed in a subsequent flood. Some of the problems associated with future floods can also be minimized by relocating sections of the main forestry road higher on the valley side.

During the 1984 flood, the river attacked its banks causing a large number of trees to fall into the river, creating large volumes of debris. Future floods will probably continue this process and more merchantable timber on the floodplain will be destroyed. In particular, trees adjacent to the river from KP 30 upstream towards the glacierets are most likely to be destroyed, and stands of timber on the floodplain between KP 10 and 30 could be damaged or destroyed in future floods. Consequently merchantable timber on the floodplain should be harvested as soon as possible.

As indicated on Map 2, upwards of 200,000 recently planted Sitka Spruce, Douglas Fir and Western Red Cedar trees were destroyed or severely damaged in the October flood. Most of these trees were located on the floodplain immediately adjacent to the river. Any trees planted in areas which are less than about 5 m above normal river levels can be expected to suffer extensive damage during future floods. Consequently reforestation efforts should be limited to areas which are at least 5 m above normal river levels, at least until such time when floods are much smaller. Assuming that Fyles Glacier will continue to recede, the magnitude of outburst floods can be expected to decrease so that several decades from now it may be practical to plant the remaining areas of the valley floor.

6.3.2 Fisheries Resources

The only option available for fisheries management is to wait until major floods have ceased before considering enhancement opportunities. In the event that the spawning area on the Smitley River is under utilized, then these fish stocks could be enhanced with negligible effects arising from the occurrence of floods on the Noeick River.

6.3.3 Recreation

The forestry access road into the Noeick valley provides a variety of recreational opportunities for local residents of Bella Coola. The occurrence of aperiodic glacier outburst floods from Ape Lake does pose a hazard to recreational users, particularly those who are unfamiliar with the area. Rising water levels, floating debris and noise caused by future jökulhlaups however, should provide adequate warning under most circumstances to allow people to leave the area. It is possible that access could be blocked by rising water but the general valley topography ensures that high ground above flood waters can be reached quickly. Camping on the floodplain of the Noeick River however, should be discouraged, since campers might be unable to remove themselves from the floodplain in sufficient time to avoid loss of life and property.

Upstream of Purgatory Glacier, the Noeick River flows in a narrow channel with steep, bedrock walls. Hikers and climbers should avoid travel along this portion of the river due to hazards posed by rock fall and avalanche as well as rapidly rising waters of a flood event. Because of the steep valley walls it would be difficult to impossible to escape rapidly rising water should a flood occur.

7.0 CONCLUSIONS AND RECOMMENDATIONS

7.1 Conclusions

1. Ape Lake drained in less than 24 hours, creating two distinct flood peaks. The first peak occurred about 1100-1200 hours on October 20, 1984 as the top 10 m of the lake drained beneath Fyles Glacier. A second peak at the tunnel exit occurred about 1630-1730 hours after the climax end-moraine forming the east sub-basin failed.

Peak discharges were probably in the range of 985 to 1500 m^3/s.

2. The flood created extensive erosion downstream to tidewater and boulders of at least 1.5 m in diameter were mobilized by the flood. Most of the sediment transport and erosion occurred between Ape Lake and km 27.5. Between km 27.5 and 14.5 where the floodplain is particularly wide, the effects of the flood were limited to deposition of an extensive layer of silt and mud where the river overflowed its banks and a slight enlargement of the river channel.

3. Besides destroying more than one kilometre of forest access road and severely damaging two bridges, the flood destroyed more than 200,000 newly planted trees. Several kilometres of spur road were damaged and some merchantable timber was lost.

4. The flood probably destroyed most or all of the salmon eggs deposited in the Noeick River earlier in the year. This one flood could result in a reduction in the commercial catch of about 1000 pink and 1000 chum and an unknown number of chinook and coho.

5. The tunnel beneath Fyles Glacier is expected to reseal within 100-200 days after which the lake will refill. There is

sufficient runoff from the Ape Lake watershed that the lake can be refilled in less than one year.

6. Assuming that Fyles Glacier continues to recede, large outburst floods will probably occur every 1-2 years until the depth of ice at the seal is reduced by 20-30 m. This could occur naturally in as few as 8-14 years or require more than 100 years, depending on climatic conditions. Once the maximum ice thickness has decreased by 20-30 m, the depth of water in Ape Lake will be correspondingly reduced and the magnitude of floods will be much smaller.

7.2 Recommendations

1. Field measurements of flow velocity, ablation rate and thickness of Fyles Glacier terminus should be made so that a more detailed assessment of the future behaviour of Ape Lake can be made. These data can also be used to obtain more reliable estimates of the rateof tunnel closure beneath Fyles Glacier. Since Fyles Glacier may respond more slowly to climatic changes than smaller glaciers nearby, consideration should be given to determining the potential for the glacier to re-advance in the near future.

2. A small scale contour map of Fyles Glacier and Ape Lake basin should be drawn, based on aerial photography flown in 1978. Additional aerial photography of the area would also be helpful in estimating the long-term changes in glacier recession and thinning. This photography would provide information on the amount of erosion which occurred in 1984 and also provide a baseline against which to measure erosion of future outburst floods.

3. Prominent signs should be posted at the South Bentinck Arm airstrip, on the main forestry access road near the airstrip

and at "Oldegaard Pass" warning people of the flood hazard. The signs should include a warning not to camp on the floodplain and advice to leave the area or to seek high ground at least 20 m above the river should a flood occur.

4. Wherever feasible, when reconstructing the forest access road, it should be relocated as far from the river channel as possible or preferably located on the lower slopes of the valley walls to minimize future washouts due to jökulhlaups.

5. Areas of the Noeick River floodplain at an elevation of less than 5 m above normal river levels should not be replanted with trees until the future behavior of Ape Lake and Fyles Glacier are more fully understood.

6. Salmon enhancement projects on the Noeick River are not advisable at this time due to the potential damage future jökulhlaups would cause to salmon stocks.

7. Application of dust on the ablation zone of Fyles Glacier to accelerate ice melt should be considered if warranted by the economic costs of future large floods.

Several other studies, not directly related to the prediction of outburst floods from Ape Lake, should be considered.

1. The flood has created many fresh, erosional scarps in the valley bottom from Ape Lake to tidewater. An examination of these surfaces would provide considerable information on the surficial geology which otherwise would be difficult to obtain and should be undertaken as soon as possible before ravelling obscures the stratigraphy. Subsequent outburst floods would be expected to create additional erosional scarps.

2. An outburst flood transports large quantities of sediment, much of which will be redeposited in the floodplain. However large quantities of sediment will be delivered to the Noeick delta. Accurate hydrographic surveys of the delta before and after an outburst flood could provide data on the magnitude of such sediment transport.

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APPENDIX 1 AERIAL PHOTOGRAPHY OF APE LAKE

The following aerial photography covering the Ape Lake area has been identified:

Year	Description
1947	July 30 Provincial trimetrogon photos (oblique) X-26-1 to X-26-20 (on loan from Dept. Geography, UBC)
1948	Federal vertical photos (93 D-B), scale 1:26,000
1951	Sept. 14 Provincial trimetrogon photos with vertical coverage X-573-36 to X-573-40 (on loan from Dept. Geography, UBC)
1952	Federal photos (93 D-C), scale 1:60,000
1954	Sept. 29 Provincial photography, scale 1:31,560 BC 1846:109-114; BC 1847:54-62, 91-94, 111-112 (on loan from Geological Survey of Canada, Vancouver)
1964	Federal photography (93 D-D), scale 1:50,000
1969	Federal photography (93 D-E), scale 1:43,000
1974	Federal photography (93 D-A), scale 1:60,000 color
1977	August 5 Provincial Government photos of Noeick Delta, scale 1:23000 BC 77105: 148-152. (on loan from Survey Depart. BCIT).
1978	July 24 Provincial photography at two scales flown simultaneously: 1:45,600 and 1:23,000 (on loan from Survey Depart., BCIT). 1:23000 series BC 78124: 204-210, 220-230 BC 78125: 14-23, 35-38, 156-158, 167-176, 200-210 BC 78126: 59-65 1:45,600 series BC 78129: 7-13

APPENDIX 2 INTERVIEW NOTES

The following notes were made by K. Ricker and J. Desloges during interviews with each individual.

1. Crown Forest Industries Ltd.

Messrs. R. Lenci and R. Clark made a reconnaissance flight over Ape Lake on October 17 before the flood and again on October 22 after the flood, obtaining many photographs of the changes caused by the flood. They visited the Noeick River floodplain, making observations of the damage to roads and recently planted trees. Most of their detailed observations are contained in the notes accompanying Map 2.

Mr. G. Davidson was the first person to arrive in the Noeick Valley on October 20. Arriving at 1100 hours, he obtained photographs of the flood in the vicinity of Purgatory Glacier and the road at kilometre 30, paying particular attention to the bank erosion and loss of road. He left the area at 1500 hours, noting that the water level was more or less constant during the time he had been on the scene. Several days later he made a closer inspection of road damage between km 0 and km 7 and measured silt lines 0.75 m above the surface of the airstrip.

Mr. C. Kabel, camp watchman at Bentinck Arm noted debris in the Arm about 1500 hours on October 20. There was no sign of flood water near the airstrip at 1800 hours but when he returned the following day at 1200 hours, the road had been washed out, the airstrip covered in mud and the water level in the Noeick River had already returned to normal.

Mr. B. Storry noted that all measurements of precipitation in the Bella Coola basin and South Bentinck Arm normally made by the logging companies stopped near the end of September, prior to the Thanksgiving storm event.

2. Wilderness Airlines Ltd. operate scheduled flights between Bella Coola and Vancouver and charter flights in the vicinity of Bella Coola, for example to South Bentinck Arm. Weather permitting, the scheduled flight path is over Ape Lake at an elevation of about 3000 m a.s.l. at 1300 hours each day.

On October 19, pilot Tom Bonner flew over the lake and saw nothing unusual. During the regular scheduled flight the following day , pilot Wayne Sissons also saw no unusual activity at the lake. On the same day however (October 20), pilot Carl Schuetze flew over South Bentinck Arm and the Noeick Delta just before and just after 1200 hours. Flood debris and silt plume at the delta were observed during the second flight which was not observed on the earlier flight. From October 20, 1984 to January 21, 1985, pilot Tom Bonner has flown over the lake on numerous occasions. No sign of lake filling was evident during any of these flights.

3. Chief Pilot, Dean River Air Services.

The last flight over Ape Lake before the event occurred in late September. No unusual lake water levels or river discharge was observed. On October 22 and 26, after the event, the east basin of Ape Lake was too turbid to permit safe landings by floatplane.

The pilot noted that the Thanksgiving storm was by far the worst fall storm that he had experienced in the area. Observations on flights after the storm indicated that heavy rains had occurred at higher elevations.

4. Mr. Kent Danielson arrived in the Noeick Valley (Km 35) about 1530 hours on October 20 and proceeded to obtain photographs of the river in flood until 1830 hours. He drove downstream to Km 30 but rising water and road washouts forced a return to Km 35 where he encountered water about 1.4 m deep which he was just able to negotiate in a 4x4 vehicle at 1830 hours.

5. Dr. and Mrs. W. Danielson, driving over "Oldegaard Pass", could hear loud noises from the Noeick Valley, 2-3 km distant. They arrived in the valley about 1530 hours. Noticing that water levels were beginning to rise about 1630 hours, they refrained from driving beyond Km 35.

6. Mr. Dan Dunaway camped on the shores of Purgatory Lake on October 13 and 14, taking photographs of the glacier snout at the water's edge. As publisher and editor of the Coast Mountain Courier, he interviewed many of the eyewitnesses for the newspaper which were subsequently published (Anonymous, 1984a)

7. Mr. S. Gascoyne arrived in the Noeick Valley (Km 35) at 1130 hours during the flood. Around midday, water levels were at least 0.3 m below the lowest point of the road at Km 35 but wet soil and a fresh veneer of silt indicated that water levels had dropped about 0.15 m around noon. While driving along the road adjacent to the river he noted that logs were moving with a velocity of at least 20 kph. The river was actively eroding its bank so that only a few minutes were required for the river to undermine and carry away conifers 25-30 cm in diameter growing on the floodplain.

8. Ms. K. McGwiny and her assistants had worked on the lower Noeick River earlier in the year on a salmon enhancement study for Fisheries and Oceans. The flood removed the staff gauge and water level recorder installed by Envirocon Ltd. Work included measurements of spawning beds before and after the flood.

9. Mr. R. Ratcliff arrived at Km. 35 in the Noeick Valley around 1530 hours. He estimated that the water was flowing as fast as a man could sprint, noting that rooted trees were moving slower with water 'spouting' off of them. He left km 35 at 1630 hours but from other observers he was told that the road had flooded to a depth of approximately 0.3 m by 1730 hours. Between 1630 and 1730 hours, he observed the river eroding the true left bank (Purgatory Moraine, km 36.2 to 36.7) with large amounts of earth and rocks being transported by

the river as it undermined the bank, here 35-45 m high. Throughout the time Mr. Ratcliff was in the area, the entire flow entered Purgatory Lake in existing channels. The lake was a 'whirlpool' of brash and bergybits.

10. Mr. Rob Skelly, Transwest Helicopter pilot made several trips to the Noeick Valley, Ape Lake and South Bentinck Arm area between October 21 and 30. He noted the dispersal of debris in South Bentinck Arm during the course of several days. Initially the debris covered the entire width of South Bentinck Arm, moving en masse to Larso Bay (19km) in 24 to 36 hours before the debris gradually dispersed into various channels around King Island. By October 25, the debris in North Bentinck Arm could be seen from the wharves at Bella Coola. The water in South Bentinck Arm was highly turdid after the flood and a significant plume was still visible around the Noeick delta on October 30, although the Noeick River was relatively clear.

11. Mr. C. Zigler reported that the precipitation at the BCFS station in Anahim Lake was not measured daily during October. Total precipitation for the period between September 29 and October 31 was at least 59.7 mm, most of which fell on Thanksgiving weekend.

12. "Willi" the Hitchhiker. During and after the Thanksgiving weekend this gentleman had been visiting friends living west of Kleena Kleene. Through local "moccasin" telegraph it was recognized that the Thanksgiving storm was unusually intense and of long duration in the Klinaklini Valley. Apparently his hunting partner, based at Klinaklina Lake (40 km SW of Kleena Kleene), had considerable difficulty crossing the usually shallow ford for as long as one week after the storm. Water levels apparently were also very high for the same duration on the upper reaches of the Dean River system to the north.

The following general observations were made by several individuals:

The event could be heard on the Nusatsum side of "Oldegaard Pass", several kilometres from the Noeick River. The noise was deafening beside the river with the movement of boulders in the river clearly audible. Several observers described the valley as "quivering".

At Km 30, brash glacier ice up to football size chunks was still present in the water. Upstream of this point many observers noted large blocks of ice left on the banks. Large blocks of ice up to 2 m in diameter were still visible on the floodplain downstream of the Fyles Glacier a week after the event.

Where the river was actively eroding its bank, collapse was rapid. At the Purgatory moraine (km 36.2 to 36.7), observers noted car sized collapses of till, some with mature conifers, cascading into the river. The erosion of the 10 m wide road allowance at Km 30 was completed in about three hours.

During this time the smell of organic material was very noticeable in the valley and near "Oldegaard Pass".

No observations were made of animals fleeing the scene or fish jumping.

APPENDIX 3 ESTIMATES OF MAXIMUM VELOCITY AND DISCHARGE

In the absence of direct measurements of discharge during the outburst flood, peak flows can be estimated by several techniques. These include empirical formula, reconstruction of hydraulic conditions during the flood and physically based models.

Prediction of flood magnitudes from glacier dammed lakes is not definitive. Clague and Mathews (1973) developed an empirical formula which relates lake volume to peak discharge. Application of the formula to observed outburst floods provides acceptable results. Formula based on regression analysis rather than physical processes however, provide little insight into conditions which occur during the flood.

Nye (1976) developed a model based on the physics of the phenomenon to explain the observed behaviour of the Grimsvotn jökulhlaup in Iceland. Spring and Hutter (1981, 1982) also developed a general model which can be applied to predict the magnitude of outburst floods. They examined the sensitivity of different parameters in modelling the Grimsvotn jökulhlaup. Clarke (1982) adapted Nye's model to simplify calculations of flood magnitude and the simulation of the discharge hydrograph. This model has been calibrated to a number of glacier outburst floods (Clarke and Mathews, 1981; Clarke and Waldron, 1984) and it can be used to predict outflows from Ape Lake.

The Clarke model requires information on Ape Lake and Fyles Glacier which are not easily quantified at this time. It is possible to estimate the parameter values and thus predict peak flows and the discharge hydrograph. The model is not directly applicable to Ape Lake however, because of the presence of the climax moraine which temporarily divided the lake basin when lake levels had declined by about 10 m, thus creating two flood peaks.

To include the range of potential contributions from the East basin and Ape Glacier embayment three possibilites were examined, corresponding to plausible dam elevations:

1. The entire volume of 45.8 million m^3 drained in one continuous event.
2. The top 10 m of the entire lake plus the remaining volume in the west basin drained in a continuous event (30.9×10^3 m^3)
3. The top 5 m of the entire lake plus the remaining volume in the west basin drained (26.2×10^6 m^3)

The modified Clague-Mathews formula (Desloges, 1984) and Clarke model were used to compare estimates of peak discharge for the three scenarios (Table 8).

Some additional comments on the input to the Clarke model are required. The model is sensitive to specification of lake temperature and tunnel roughness. An average temperature of 4°C was assumed to be a reasonable value for late October. The Manning roughness coefficient was varied but the results in Table 8 are for the commonly used value of $n=0.12\ m^{-1/3}s$, which corresponds to a rough tunnel. The ice thickness at the seal was assumed to be 85 m, the height of ice above the tunnel entrance while tunnel length was assumed to be 1860 m.

As shown in Table 8, the modified Clague-Mathews formula and the Clarke model predict peak discharges ranging between a low of 400 m^3/s to a high of 970 m^3/s. Because of the large range in predicted values, independent methods were used to estimate the peak discharge. These methods are based on a reconstruction of hydraulic conditions during the flood as

TABLE 8 COMPUTED PEAK DISCHARGES USING THE CLARKE MODEL

	Case 1	Case 2	Case 3
Volume m^3	45.8×10^6	30.86×10^6	26.21×10^6
Clague-Mathews m^3/s	970	772	702
Clarke model m^3/s	681	483	400
Time to drain hrs	84	79	77

TABLE 9 CROSS-SECTION SURVEY DATA

Control section downstream from Fyles Glacier

	CS#1	CS#2	CS#3	Average
Total Area (m^2)	557	853	789	733
Upper waterline (m^2)	256	326	270	284
Hydraulic radius (m)	3.5	3.3	2.65	3.15
Lower waterline (m^2)	55.5	22.0	31.0	36.0
Hydaulic radius (m)	1.22	0.55	0.94	0.9

Cross-section downstream from Purgatory Glacier

Channel width (m)	70
Channel depth (m)	4.5-5.0
Total area (m^2)	315 - 350
Slope (m/m)	0.017

TABLE 10 RECONSTRUCTED MAXIMUM INSTANTANEOUS DISCHARGE (M^3/S) FROM CROSS-SECTION SURVEYS

Control section

	CS#1	CS#2	CS#3	Average
Upper waterline (m^3/s)	1,480	1,813	1,297	1,530
Lower waterline (m^3/s)	159	37	75	84

Cross-section downstream from Purgatory Glacier

	Depth 4.5 m	Depth 5.0 m
Roughness diameter 3500 mm	752 m^3/s	896 m^3/s
Roughness diameter 4000 mm	735 m^3/s	876 m^3/s

determined from measurements of water level, local gradient and size of sediment known to have moved during the flood.

Table 9 is a summary of survey data collected at two sites; a control section about one kilometre downstream from Fyles Glacier (Plate IVA) and a second site at kilometre 35.5, downstream from Purgatory Lake (Plate VII).

At the control section downstream from Fyles Glacier, two potential trim lines were visible on channel banks. Therefore cross-sectional areas were measured corresponding to each water line and the figures for total area were measured from terrace top positions. Other relevant parameters at this site are slope (0.0104 m/m) and the b-axis dimensions of boulders forming roughness elements in the channel. Examination of the sediment suggested that several of the largest channel boulders may have been derived from bank collapse after the flood surge had passed. Therefore, the largest roughness elements present during the flood were in the range 1000-1500 mm.

At the survey site downstream from Purgatory Lake, channel width was estimated using a rangefinder since it was not possible to cross the Noeick River. At this location, the river overflowed its banks and high water level was well defined by debris and silt deposits in the vegetation. Since the channel was roughly rectangular in shape where the river had eroded its banks, the depth of flow was estimated by measuring the height of the high water line above the channel bed at the edge of the bank. Since the bed of the river in the center of the channel is somewhat lower than the bed near the river bank, the hydraulic radius will be slightly underestimated, thus giving a low value for the peak discharge.

The slope was 0.017 m/m and the b-axis dimensions of boulders forming roughness elements in the channel were in the range

3000 - 4000 mm (Plate VIIB).

Using the Manning-Strickler relation (Henderson, 1966, Church and Gilbert, 1975)

$$Q = A\, R^{2/3}\, S^{1/2} / 0.038 D^{1/6} \qquad \text{where}$$

Q = maximum instantaneous discharge , m^3/s
A = cross sectional area m^2
R = hydraulic radius m
S = slope of the river channel at the bed m/m
D = largest B-axis diameter of sediment in the channel, mm

maximum discharges were estimated for each cross-section listed in Table 9. The results presented in Table 10 were calculated assuming a boulder b-axis diameter of 1500 mm at the control section and 3500 and 4000 mm at the section downstream from Purgatory Lake.

Observations of boulders lying on top of young conifer trees can be used to infer mean velocities and hence peak discharges. Boulders which definitely moved during the flood had a b-axis diameter of about 1.5 m. Galay (1970) assembled data on the largest particle sizes moved under a wide range of mean channel flow velocities. The data are based mainly on reports describing the effects of major floods. Using Galay's data, single exposed boulders with a diameter of 1.5 m would remain stable until mean channel velocities exceeded about 2.7 - 3 m/s. This implies that for a cross-section area of 350 m^2, peak discharge was in the range 945-1050 m^3/s. This approach will probably tend to overestimate the discharge because much material was entrained as the undercut river banks slumped into the channel. Thus the flow merely maintained movement of large clasts rather than initiating movement from the river bed.

Based on the preceding discussion, it is thought that the Clarke model underestimates the peak discharge for Ape Lake. This may be due to incorrect estimates of water temperature, tunnel roughness, volume of available water and maximum thickness of ice. Peak discharges up to 1500 m^3/s may have occurred but based on the available data the best estimate of peak discharge is probably about 1000 m^3/s.

From eyewitness photographs (Plate VIIIA and Plate IXA), the flow was known to have been supercritical, ie the Froude number Fr was greater than one. Mr. S. Gascoyne drove along the side of the river, estimating that the velocity of trees carried by the flow was at least 20 kilometres per hour.

The Froude number is computed from the formula (Henderson, 1966)

$$Fr = V^2/gL \quad \text{or } V = (Fr \ / \ gL\ )^{1/2} \qquad \text{where}$$

Fr = non-dimensional Froude number

V = velocity, m/s

g = acceleration due to gravity, 9.81 m/s^2

L = length dimensions (depth of flow), m

Supercritical conditions occur when Fr = 1. Using Fr = 1 will give a lower bound on maximum velocities. For flow depths of 4 and 5 m, the water velocity required to produce supercritical conditions would be 6.3 to 7.0 m/s respectively (22.5 - 25 km/hr).

APPENDIX 4 SUMMARY OF DENDROCHRONOLOGIC SAMPLING SITES - NOEICK VALLEY AND APE LAKE

Site Number (see Maps 2, 3 for location) — Site Description

Site #8

Retreatal outwash escarpments on Purgatory Neoglacial end moraine (K.P. 35.0). Scarp 6m high from new base level (Plate VIIA). Samples:

- 8a Rooted log 3 m below top of terrace; collected for potential radiocarbon date; submitted to GSC, Ottawa.
- 8b Log under 2 m diameter boulder 4 m below top of the terrace; collected for potential radiocarbon date; submitted to GSC, Ottawa.

Site #10A

Climax Neoglacial lateral moraine, true right bank of Purgatory Glacier. Moraine dissected by Noeick River at K.P. 36.2 to 36.7. Scarp 30-50 m high exposes a single stratum of buried logs 8-15 m below the top surface (Plate VIA). A sample was not obtained due to inaccessibility.

Site #11

Climax Neoglacial end moraine of Fyles Glacier, north shore of Ape Lake below trimline. Low hummocky morainal crests separated from upslope trimline on bedrock by a meltwater channel (Figure 3, Map 3. Samples:

- 11a Live _Pinus_ sp. at shoreline to moraine interface. _ca._ 1931-1950 A.D. (false rings) see photocopy of tree disc.
- 11b Live _Abies amabilis_ (?) on lower slope of moraine facing meltwater channel noted above. _ca._ 1913-1914 A.D. see photocopy of tree disc.

11c Live <u>Pinus</u> sp. on crest of moraine <u>ca.</u> 1914-1915 A.D. see photocopy of tree disc.

All samples were taken at the root-trunk interface.

Site #12

Varved muds below full pool level of Ape Lake in the eastern bay near the Ape Creek outlet (Map 3, Plate IIA). Muds retrogressively slumped into the bay soon after the lake drained exposing 3 to 5 m high backfaces (intact) near shoreline. Samples at three levels below top of the exposure (<u>ca.</u> 1 m below full pool level) are as follows:

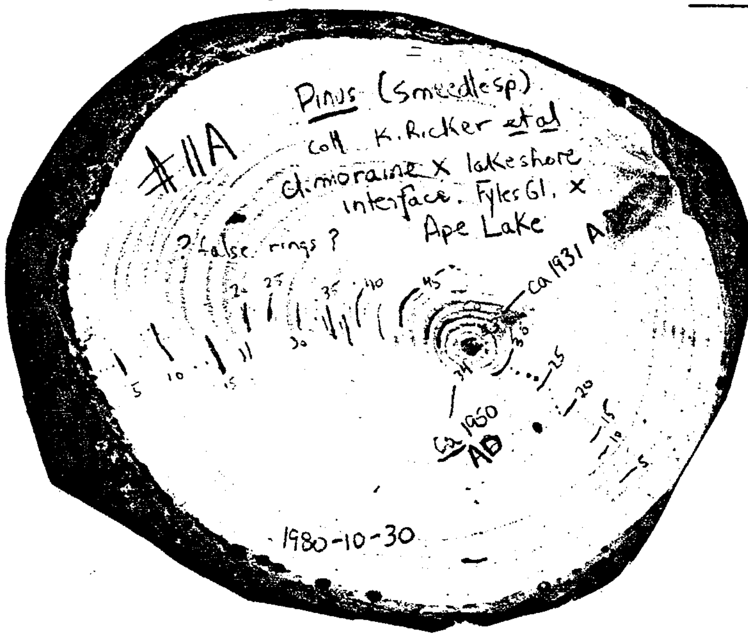
12a Sample of twigs in a varved sequence 0.56 m below the surface.

12b Sample of twigs in a layer at the base of visible varves 0.9 m below the surface.

12c Log protruding 3.0 m from a face formed in homogeneous, massive appearing muds. Sample collected and submitted to GSC in Ottawa for potential radiocarbon dating.

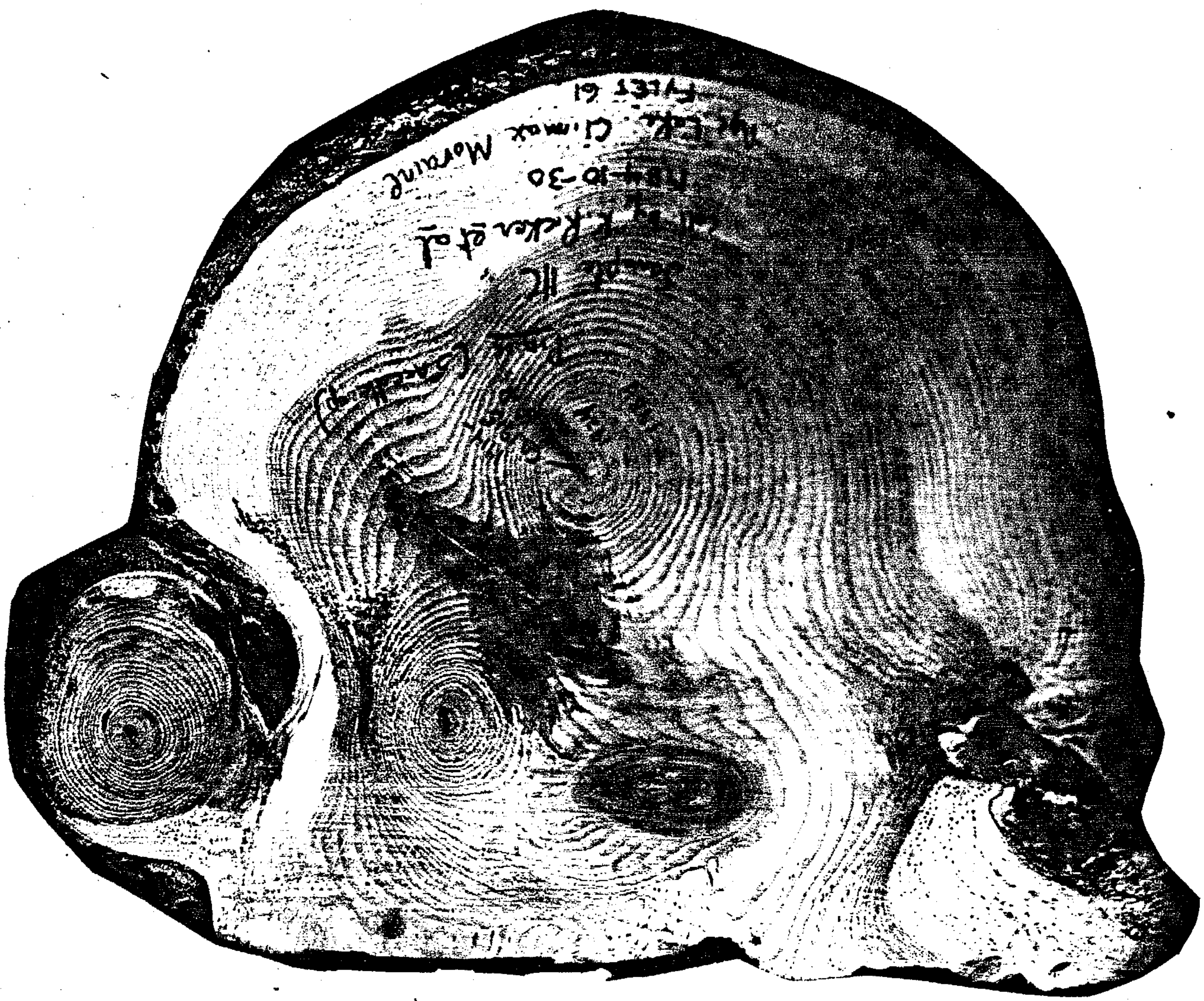
Isolated logs and stumps were seen on the north foreslopes of the west basin of Ape Lake (<u>ca</u>. 20 m below full pool level). No samples were collected because blowing snow obscured the stratigraphic position.

11A Photocopy of tree disc collected at Ape Lake: interface of full pool level shoreline to Fyles Glacier climax moraine. Pinus sp.



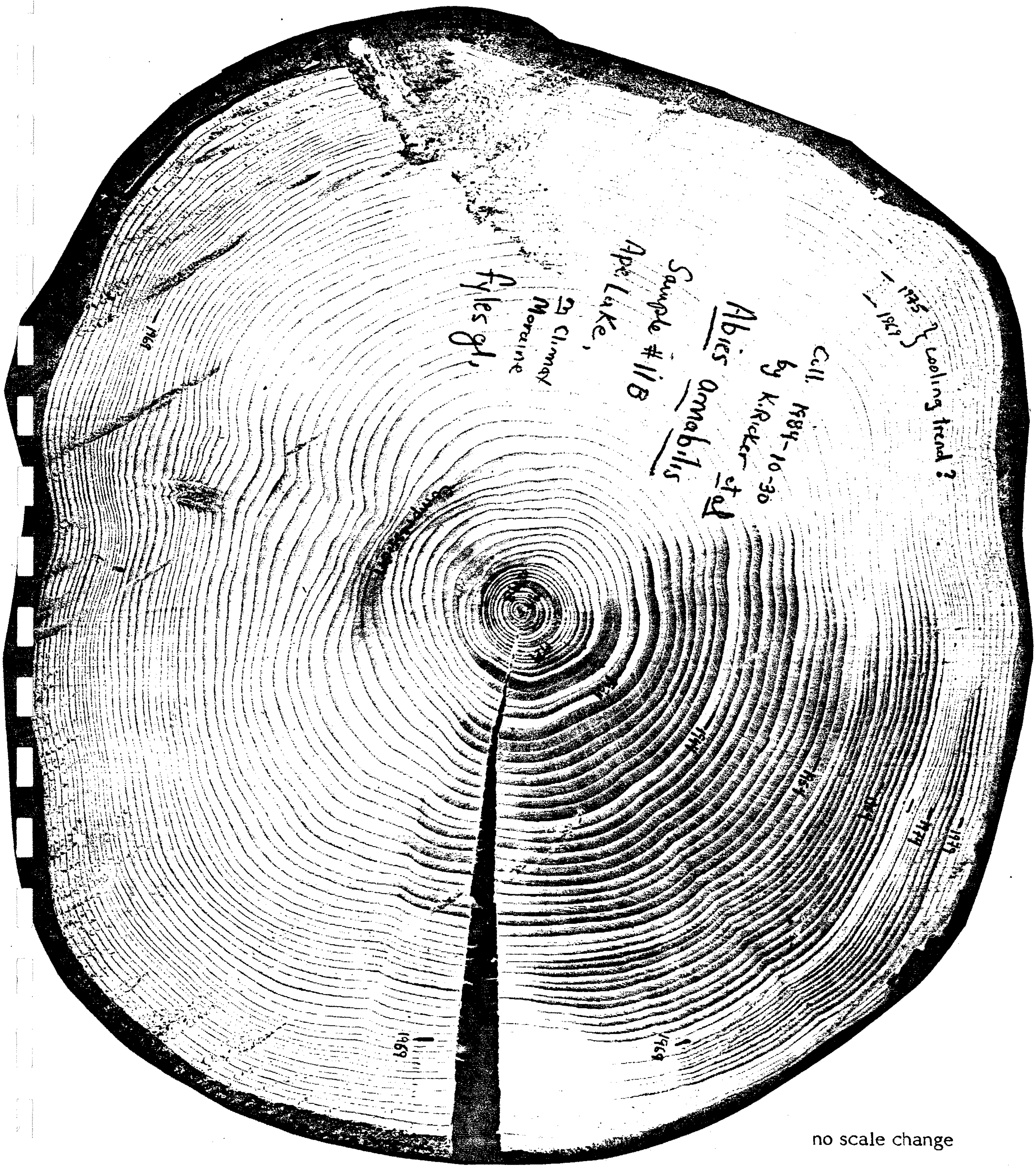
no scale change

11C Photocopy of tree disc collected at Ape Lake on crest of Fyles Glacier climax moraine. Pinus sp.

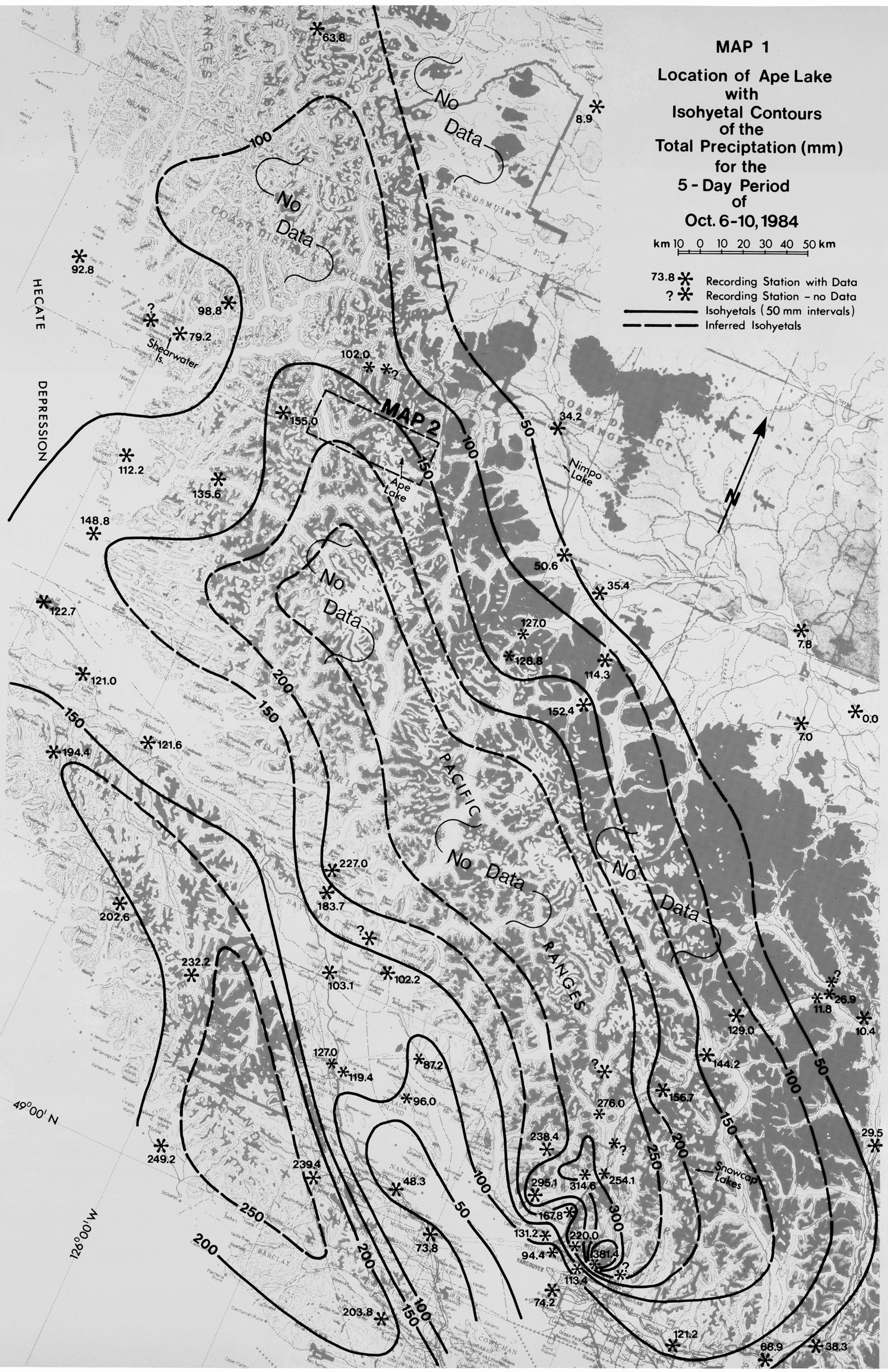


no scale change

11B Photocopy of tree disc collected at Ape Lake on Fyles Glacier Climax Moraine. Abies amabilis?

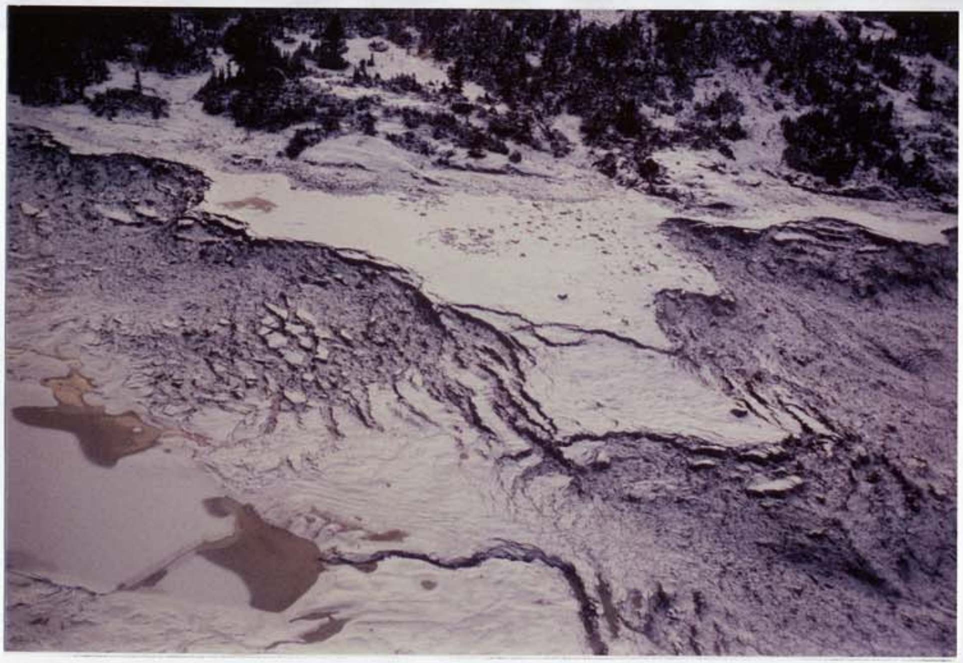


no scale change



MAP 1

Location of Ape Lake with Isohyetal Contours of the Total Preciptation (mm) for the 5 - Day Period of Oct. 6-10, 1984

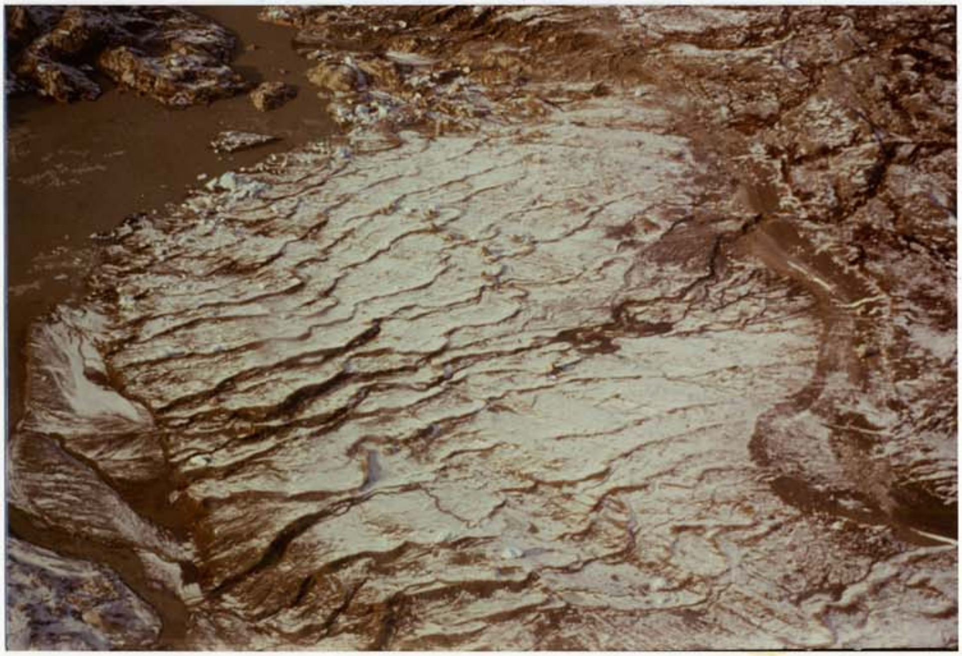


PL 11A.

Retrogressive slumps in varved silts at the east end of Ape Lake. A buried log was extracted from the vertical face in the centre of the photo. K.P. 53.6 Photo: J. Desloges

PL 11B.

A breached climax end moraine (f.g.) separates the east basin (-20m) from the west basin. Note the newly cut channel through shoal of DeGeer moraines (see PL 11C) separating a mid-basin pond (-22m) and standing water adjacent to Fyles Glacier (-38m). Note large washout ravine on delta (right). Photo: R. Lenci



PL 11C

DeGeer moraines in west basin showing ca. 19 years of Fyles Glacier recession between 1947 (lower right) and ca. 1966 (edge of standing water). KP. 50.7 Photo: D. Jones

PL IIIA.

East (lakeside) margin of Fyles Glacier showing: former lake level etched in ice, false tunnel (left), roche moutonee (centre) and actual tunnel entrance below collapsed ice (right centre). Note DeGeer moraines (f.g), ravines cutting the basin foreslope, and full pool level strandlines on right.
K.P. 50.2
Photo: R. Lenci

PL IIIC.

West margin of Fyles Glacier showing an amphitheatre of ice surrounding the tunnel exit (centre) which has subsequently filled with collapsed ice. Ice blocks and boulders are visible on the outwash plain downstream from the tunnel exit. K.P. 47.7
Photo: R. Lenci

PL IIIB.

Monitor cairn constructed on the NE corner of Fyles Glacier to indicate the position of the ice margin on October 30, 1984. Cairn is 32 metres from snout. K.P. 50.2 Photo: K. Ricker

PL IVA. View east towards Fyles Glacier from above Upper Canyon reach. Outflow waters were forced through a constriction (K.P. 46.3) in the river (survey control section): (A) formed by a fluted morainal plain, (B) which also marks the 19th century extent of Fyles Glacier. Note: lateral erosion of channel which forms up to 10m scarp on morainal and outwash deposits. Photo: D. Jones

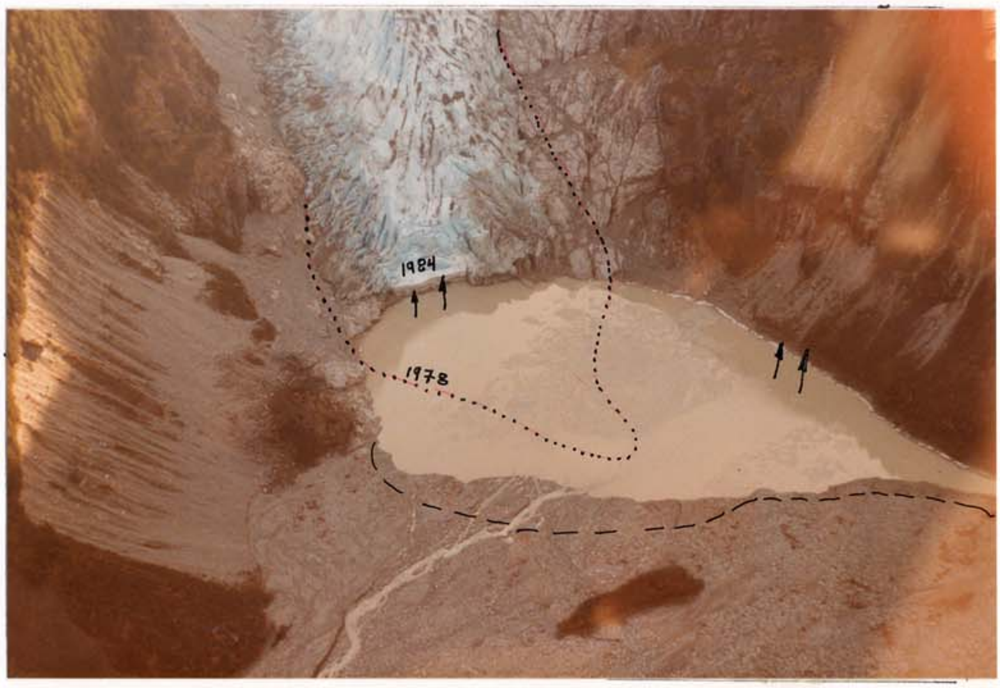
PL IVB.

Upstream view of Noeick River in Upper Canyon reach. The upper debris cone glacieret contains a post-flood collapse pit (arrow) while the bottom end of the lower glacieret has been completely sheared off. Note the large rock slide below and to the left of the lower ice mass, the constriction of the channel around this debris apron, and lateral cutting of the channel in the foreground. Photo: R. Lenci K.P. 39.5-44.0

PL VA. View west along Noeick River (August, 1984) showing Outwash Fan (f.g), Purgatory Lake (m.g) and Bend reach (b.g). Switch back in the road is visible at centre right. This road connects the Noeick Valley with the Nausatsum (Bella Coola) Valley. K.P. 39.0-27.5 Photo: J. Desloges

PL VB. Photo taken from the same vantage point as PL VA (October 30, 1984). Note the erosion scarps in the valley side fans, the stripped valley bottom vegetation and widened foreground channel. Photo: K. Ricker

PL VIA. Cross-section through right lateral moraine of Purgatory Glacier (looking upstream). Section was exposed by erosion from flood waters. Note the stratum of buried organic debris (arrows) suggesting two glacier advances. K.P. 36.2



PL VIB. Overview of Purgatory lake and the terminus of the Purgatory Glacier. A high water trimline (arrow) marks the storage level of transient flood waters in the lake. Debris influx to the lake (dashed line) has probably reduced the lake volume substantially. K.P. 36.0 Photo: R. Lenci

PL VIIA.

An upstream view towards the right lateral moraine (top centre) and subdued end moraine (arrow) of Purgatory Glacier. Near vertical banks (left centre) have formed from the undercutting of the end moraine: an important source of the large bouldery debris seen in the flood channel. K.P. 34.8-36.0 Photo: K. Ricker

PL VIIB.

Buried logs provide evidence for the size of boulders mobilized by flood waters downstream from Purgatory Lake. Mean axial measurements of these boulders are in excess of 150 cm. The undercut end moraine is visible in the background. K.P. 34.9 Photo: D. Jones

PL VIIC.

An erosional scarp (20+m high) at the toe of a valley side outwash alluvial-colluvial fan indicates the severity of flood scour along the valley margins within the Bend reach. Approximate flood water levels reached the low terrace scarp on the left, as shown by eyewitness photos of event. K.P. 34.5 Photo: K. Ricker

PL VIIIA. Noeick River in flood at 1800 hrs. (Oct. 20, 1984) looking downstream from uppermost road washout. Note the turbulence of floodwaters in the middle channel (left centre) and stone line on the road (arrow) showing an earlier high water level. K.P. 34.6
Photo: K. Danielson

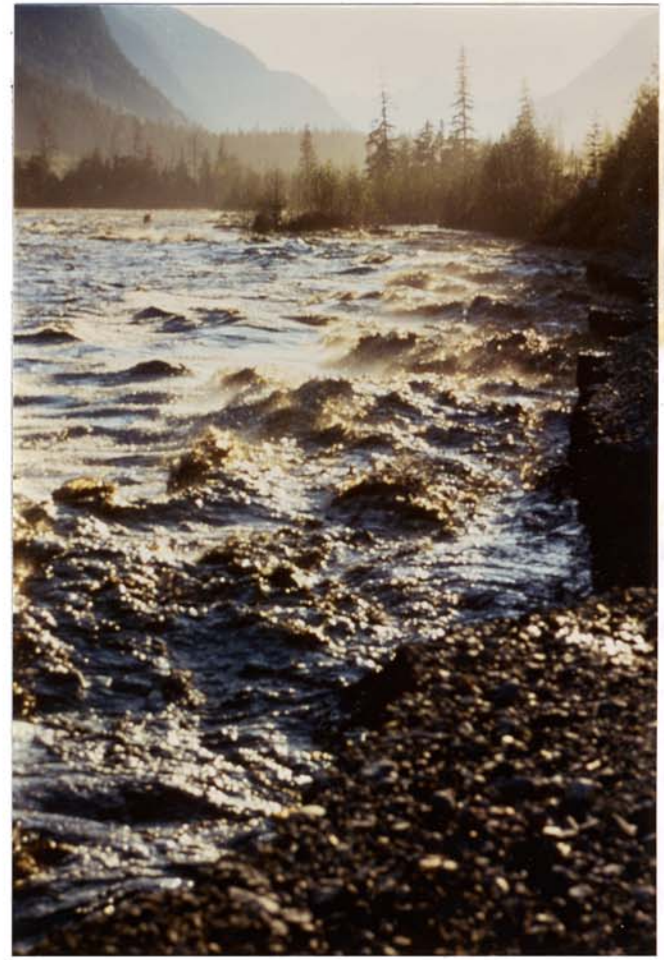
PL VIIIB. Noeick R. looking d/s from the same position as PL VIIIA following the flood. Extensive removal of trees has occurred, and the disappearance of the former high water mark (arrow in VIIIA) suggests water levels may have risen again after the earlier photograph was taken. The scour line on the bedrock (l.c.) rises 5-6m above the flood plain. Photo: K. Ricker

PL IXB

Post-flood overview of road washout shown in PL IXA. Convergence of flood waters within the channel at this point resulted in heavy erosion in the right bank of the river,leaving a large vertical scarp in the valley side fan.
K.P. 30.0(L)-30.5(R)
Photo: K. Ricker

PL IXC.

Overbank deposits of mud and debris cover the remains of an eroded valley flat where extensive tree planting had been completed.
K.P. 19.0
Photo: R. Lenci



PL IXA

View looking downstream at hydraulic jumps forming in flood waters near the road washout in the Bend reach. The photo was taken near peak flow at approximately 1800 hrs on October 20, 1984. K.P. 30.5 Photo: K. Danielson

PL XA

Upstream view of the Noeick River within a bedrock canyon at 5-mile bridge. Vegetation was stripped from levels above the bridge (left) by super elevated floodwaters. The clearspan of the bridge above normal water levels is 12m. K.P. 7.0 Photo: K. Ricker

PL XB.

View south across 5-mile bridge showing scattered organic debris on the bridge deck and upstream erosion (left) of the bridge abutment. Photo: G. Davidson

PL XC.

Upstream view of turbid and debris laden waters at the Noeick River delta approximately 36 hours after the flood peak had passed. Dashed line indicates normal high tide extent. Flood waters occupied both distributary channels including residual overflows onto the airstrip (left) where water depths approached one metre. K.P. 0-3.0 Photo: R. Lenci