GLACIAL HYDROLOGY OF AN ICE-DAMMED LAKE

THE GLACIAL HYDROLOGY OF AN ICE-DAMMED LAKE, ELLESMERE ISLAND, N.W.T.

By

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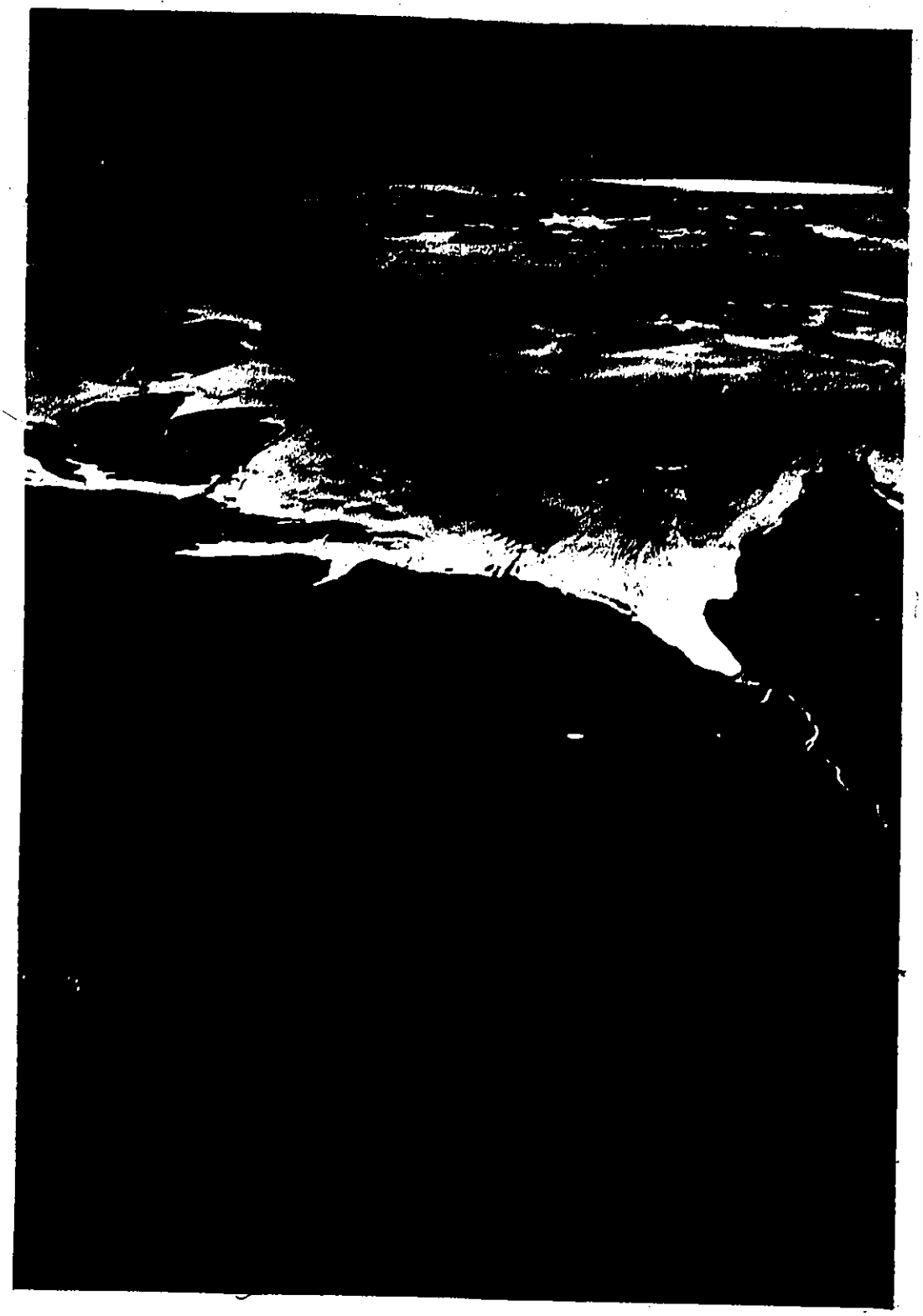

Frontispiece: Oblique air photo of McMaster Lakes

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SCOPE AND CONTENTS:

A series of ice-dammed lakes are ponded against the margin of an ice cap in a high arctic terrain. One of these lakes was investigated in detail, and found to drain catastrophically late in the season. The morphology, thermal regime, and drainage and filling behavior of the lake was monitored. Meteorological data was collected and correlated with the discharge record for the outlet stream from the lake. The damming glacier extends out into the lake basin, and this shelf ice was found to be a critical factor in the drainage of the lake. Finally, a mechanism of lake drainage is proposed to explain the observed behavior.

ABSTRACT

McMaster Lake is a 7.2 km ^{2} ice-dammed lake ponded against the margin of the Ellesmere Ice Cap, Ellesmere Island, Northwest Territories. The lake occupies a deep structural basin, and was found to be amictic, with a near isothermal temperature profile of 0.0 - 0.5° C, and a year round 3 m thick ice cover. The lake was observed over three seasons to lose 3-5 m of water rapidly during the month of August. In 1974 the outflow was monitored in Siphon Creek and the Sverdrup River, and a characteristic jokulhlaup hydrograph with a peak discharge of 85 m^3/sec was observed. Correlation of the temperature, incoming solar radiation, relative humidity, and precipitation data collected at the McMaster Lakes site with Siphon Creek discharge yields a good explanation of variance in discharge prior to the jokulhlaup. The glaciology of the shelving glacier bordering McMaster Lake was found to be instrumental in the lake drainage behavior, and the origins and activity of the ice shelf are documented. Consideration of the available data on McMaster Lake, and a review of existing theories leads to the proposal of a modified barrier flotation mechanism for the drainage of McMaster Lake.

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Finally, my husband Jan contributed field assistance during the 1974 season, his considerable computer programming skills, and general encouragement and advice at all the right times.

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CHAPTER 1

INTRODUCTION

1.1 Rationale and objectives of study

Ice-dammed lakes are recognized as an integral part of the glacial landscape in both the academic and popular literature. Within the last decade, advances in the relatively new field of glacial hydrology have indicated that these lakes are often an important component of a glacier's drainage system. The ice-dammed lake has been utilized as an index of glacio-climatic activity, perhaps overworked as a feature of the Pleistocene era, and recognized as the progenitor of an important geomorphic agent, the jokulhlaup.

In the Canadian high Arctic, glacial hydrology is one of the latest aspects of earth science studies to receive attention, and it is still at a rudimentary level. Arctic hydrology and the geomorphic role of rivers have been the topic of several major studies, Cogley (1971), Church (1972), McCann, Howarth and Cogley, (1972), McCann and Cogley (1974). In these studies, fluvial activity is attributed with a much greater role as a geomorphic agent

than previously supposed. Several of these papers (Church, 1972; McCann and Cogley, 1974) have examined glacial meltwater streams, but the only studies in which the glaciohydrological system is considered as a continuum are the Axel Heiberg Island Expeditions of the 1960's (eg: Muller et al, 1963; Andrews, 1964; Adams, 1966; Maag, 1969).

The initial aim of this study was to investigate the glacial hydrology of a portion of the sub-polar Ellesmere Ice Cap, Ellesmere Island, Northwest Territories (Figure 1.1). The McMaster Lakes study area (78°N, 82° W) is located 25 km beyond the head of Vendom Fiord in south central Ellesmere Island. The presence of several ice-dammed lakes situated along the margin of the ice cap (McCann et al, 1972; Hodgson, 1973) attracted attention, and their study became the main focus of the field program. This set of marginal lakes constitutes the headwaters of one of the major tributaries of the Sverdrup River, draining southward into Vendom Fiord (Figure 1.2).

The specific objectives of this paper are an investigation of the limnology and drainage behavior of a marginal ice-dammed lake, and the derivation of a drainage mechanism for this lake. It is hoped that such a study will provide much needed additional data on the nature and behavior of ice-dammed lakes associated with sub-polar rather than temperate glaciers. Maag's (1969) monograph on this subject stands alone as the sole major reference on

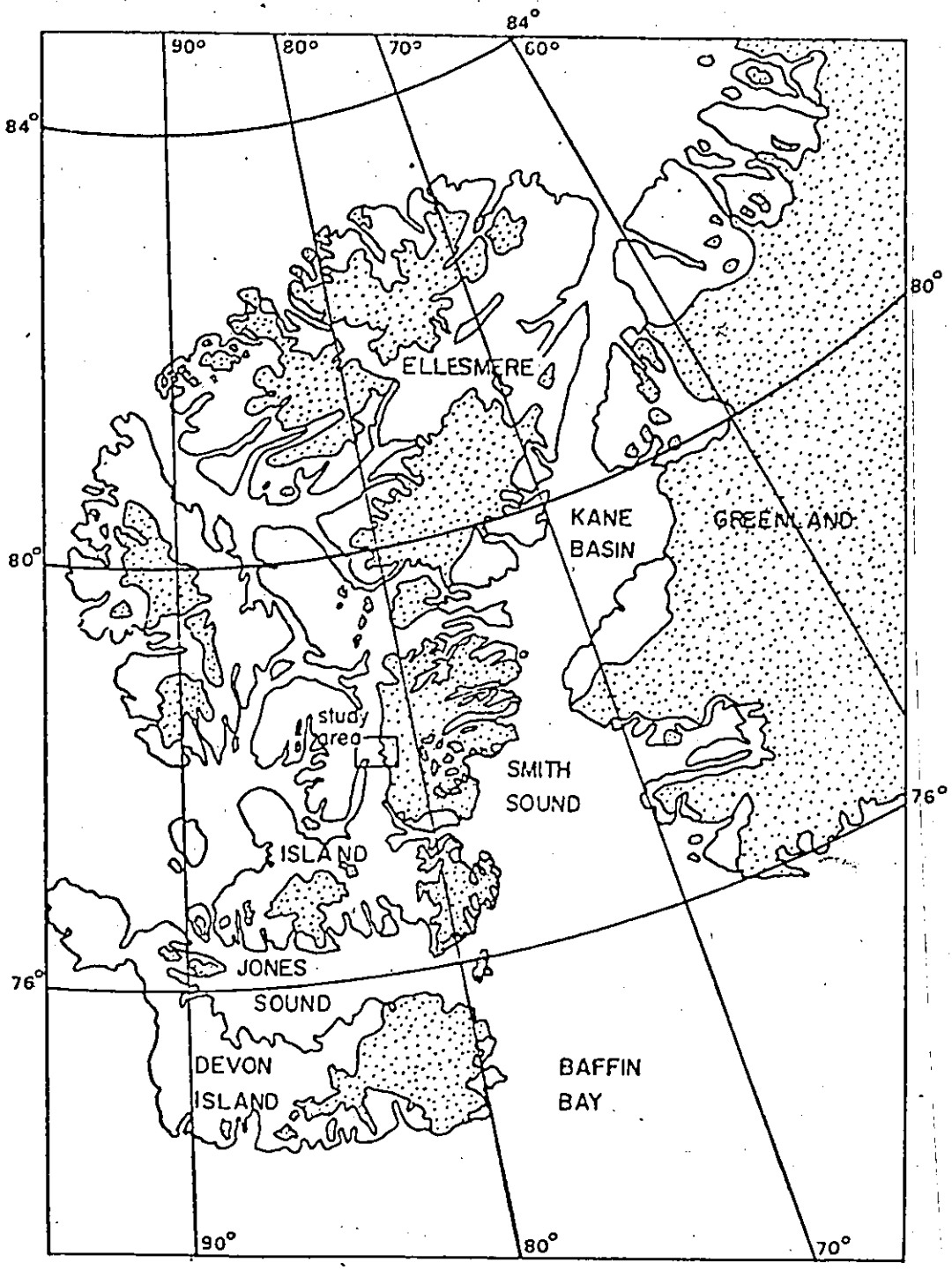


Figure 1.1 Location of the McMaster Lakes study area

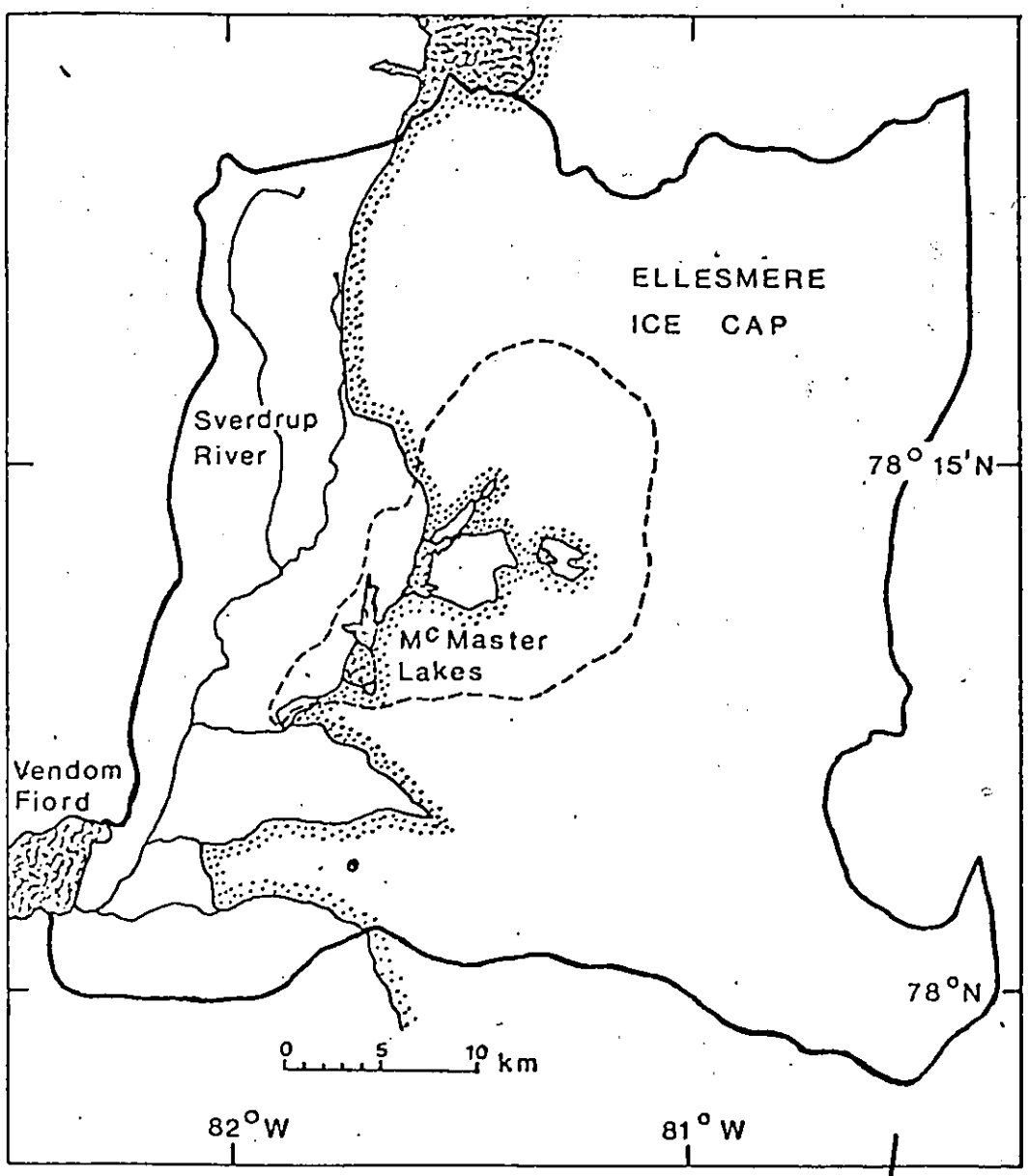


Figure 1.2 The Sverdrup River drainage basin

Arctic ice-dammed lakes at the present time. Consideration of the ice-dammed lake as an integral component of the glacial hydrological system has been provided by Stenborg (1970), but the present study proposes to examine this relationship in greater detail.

1.2 Organization of thesis

The principal findings of the field program are presented in Chapters 4, 5, and 6. These deal with the limnology of McMaster Lake (Chapter 4), the hydrometeorology of the outlet stream and jokulhlaup phenomena (Chapter 5), and those glaciological features relevant to lake drainage (Chapter 6). Chapter 7 brings this data together and postulates a mechanism for the drainage of McMaster Lake. Chapter 3 attempts to establish the study in the context of high arctic glaciological investigations, and the specialized field of ice-dammed lakes. The field area and research procedure is also outlined in Chapter 3.

For the sake of convenience, a list of both official and unofficial place names to be used in the text will be introduced here (Table 1.1), thereby avoiding the constant use of quotation marks with unofficial place names. Most of these unofficial names have been coined by members of the McMaster University field parties to facilitate description of features, and they are not intended to be considered as permanent names.

TABLE 1.1 LIST OF OFFICIAL AND UNOFFICIAL PLACE NAMES USED IN THE TEXT.

OFFICIAL

Vendom Fiord
Ellesmere Ice Cap

UNOFFICIAL

Sverdrup River (also known as Bentham River, Norris, 1963)
Schei River
Schei Glacier
McMaster Lake
Bay
McMaster Glacier
Siphon Creek
Upper Lake
Connecting Ponds
Waterfall Stream - North Channel
- South Channel

CHAPTER 2

GLACIAL HYDROLOGY IN THE CANADIAN HIGH ARCTIC

Considering that the Canadian Arctic archipelago is the third largest glacierized region of the world, after Antarctica and Greenland; 153,200 km^2 of the world total ice cover of 15×10^6 km^2 (Embleton and King, 1968; Flint, 1971); the glaciological literature on the area is surprisingly scant. The literature prior to the 1950's is primarily descriptive, with good summaries of the early work provided by Sharp (1956) and Taylor (1956). The few references to ice-dammed lakes in the Canadian Arctic and other more general works will be discussed as background material for the following chapters.

2.1 General Glaciology and Glacial Hydrology

The remoteness and logistic problems of working in the Canadian high Arctic have restricted the quantity and nature of glaciological research until the last decade. Government or university sponsored expeditions working in the field for only the 2 - 3 month summer period have formed the bulk of the research operations, yielding a sketchy and often intermittent record of glacial activity. The permanent weather stations at Alert, Eureka and Resolute are providing a now 25 year old continuous record of basic

meteological parameters, and automatic recording devices have been used with some success to supply continuous records of various glaciological, hydrological and meteorological parameters.

One of the first groups to make observations on the state of the glaciers of Ellesmere and Axel Heiberg Islands was the 1899 - 1903 Fram Expedition from Norway (Sverdrup, 1904), which provided the first accurate maps of the coastal and ice cap outlines. Other early contributors were the Lady Franklin Bay Expedition (Greely, 1886); MacMillan's party on northern Ellesmere Island and Greenland (MacMillan, 1925); the Oxford University Ellesmere Land Expedition (Humphreys et al, 1936); and Haig-Thomas' expedition to Ellesmere Island. Bentham (1941) made observations on the glaciers of southern Ellesmere Island, noting an apparent stillstand or slight retreat of the ice fronts for the periods 1936-38. Taylor's (1956) monumental 12 volume study of the physical geography of the Queen Elizabeth Islands, Dunbar and Greenaway's (1956) book 'Arctic Canada from the Air', and Sharp's (1956) review of glaciers are the last major papers prepared in the old style, from literature sources and aerial photographs.

The first large scale expedition to provide accurate quantitative data on glaciological and hydrological parameters was the Arctic Institute of North America and Canadian government sponsored expedition to Baffin Island.

Work on the Barnes Ice Cap starting in 1950 included mass balance studies, examination of an unusual situation where superimposed ice is the major form of accumulation, and utilization of new and advanced field techniques. These findings were documented in a series of papers published in the Journal of Glaciology (Baird, 1952; Ward, 1952a; 1952b; Orvig, 1953; Ward, 1953; Ward, 1954; Ward and Baird, 1954; Rothlisberger, 1955; Ward, 1955). Glaciological, hydrological and limnological investigations of large proglacial lakes have continued under the auspices of the Geographical Branch of Mines and Technical Surveys (Andrews, 1963; Sagar, 1966), and now the Glaciology Division of Environment Canada as well as the Geological Survey of Canada (Church, 1972; Holdsworth, 1973; Barnett and Holdsworth, 1974).

The first major glaciological field program carried out on Ellesmere Island was Operation Hazen, organized by the Defense Research Board during the International Geophysical Year, 1957-58 (Hattersley-Smith, 1958). Mass balance studies on Gilman Glacier, northern Ellesmere Island (Hattersley-Smith, 1960a; 1963b) and glacial geomorphological investigations in the Tanquary Fiord area (Christie, 1967; Hattersley-Smith, 1969) confirmed the general belief that Arctic glaciers have been relatively stationary for approximately the last 900 years. The apparently glacial origin of many of the fiord systems of

the Arctic Archipelago proposed by Nichols (1936), is verified by evidence of complete ice cover of most of the Arctic islands up until the time of the Climatic Optimum (6000-4000 years B.P.) (Craig and Fyles, 1960; Hattersley-Smith, 1969). Sounding of the Nansen Sound fiord system separating Ellesmere and Axel Heiberg Islands (Ford and Hattersley-Smith, 1965) reveals classic fiord deepening and staightening of structurally controlled preglacial valleys, and characteristic hanging valleys, truncated rock spurs and fiord mouth sills.

The high Arctic research station established on Axel Heiberg Island by McGill University during the Jacobsen-McGill Arctic Research Expedition 1959-1962 (Muller, 1961; Muller et al, 1963) continues until the present to produce various interrelated glaciological, glaciohydrological and hydrometeorological research reports (Andrews, 1964; Havens, 1964; Adams, 1966; Muller and Roskin-Sharlin, 1967; Ommanney, 1969; Maag, 1969). As will be discussed in subsequent chapters, the work by Adams and Maag provides some of the first quantitative knowledge of melt processes and dynamics of ice-dammed lakes on high arctic glaciers.

The Devon Island Expeditions, 1960-1966, sponsored by the Arctic Institute of North America provided glaciological and meteorological data comparable to the Axel Heiberg work. In addition several field parties wintered over (Koerner et al, 1963) collecting valuable information

on winter ice temperature, ice fabric, and mass balance, as well as a full set of meteorological observations. Detailed glacial micrometeorological data (Holmgren, 1971) and glacier mass budgets (Koerner, 1970) are presented in addition to an excellent study of the glaciohydrological network as a system (Keeler, 1964). Early work on the assessment of ice thickness by geophysical techniques and deep coring of an ice cap was also carried out (Koerner et al, 1963).

Despite the quality of literature cited above, it appears that the glaciology and glacial hydrology of polar and subpolar glaciers lags far behind studies on temperate glaciers. Inclusion of the literature from the more intensively studied Greenland and Antarctic ice masses, and the special case of the Ward Hunt Ice Shelf and associated ice islands (Crary, 1958; 1960; Hattersley - Smith, 1963a; Hattersley - Smith and Serson, 1970; Holdsworth, 1971) adds to the knowledge of Arctic glacial processes, but a gap still exists. This disparity is particularly evident in the field of glacial hydrology and the study of ice-dammed lakes.

2.2 Ice-dammed lakes

A substantive literature on ice-dammed lakes has accumulated over the past century, but the great majority of these references are either brief passing notice in general

glaciological papers or simple descriptions of physical features and events. Cataloguing of the specific characteristics of ice-dammed lakes, their geographical distribution, and some information of such features typifies many pre- 1950 papers, but almost all references include reports of the catastropic emptying and downstream consequences of lake drainage. Review and evaluation of drainage mechanism theories will be reserved until Chapter 7, but a brief general review of the literature precedes the examination of the McMaster Lake example. Other more exhaustive surveys of the literature are provided by Maag (1969), and Blachut and Ballantyne (1976).

The earliest references to ice-dammed lakes are from Iceland and the Alps, where people had become forcibly aware of the catastrophic floods issuing from the glaciers upstream (Thorarinsson, 1939b; Rabot, 1905). The only early writers to attempt explanations of catastrophic drainage were Fabricius (1788) and Palsson (1882); the latter linking Icelandic floods to volcanic eruptions, a theory which persists to the present time (Tryggvason, 1960).

Other papers which perceptively anticipated some of the theories of drainage mechanisms were Rink's (1862) observations on Imaersotoq, Greenland, Russell's (1893) description of the formation of supraglacial and marginal lakes on the Malaspina Glacier, Alaska, and the good set of

observations on the behavior of Marjelen See, Switzerland provided by Munro (1892).

Detailed investigations involving field work on the behavior of ice-dammed lakes was not forthcoming until the 1940's, despite the great importance placed on ice-dammed lakes as significant features of the Pleistocene landscape. The Swedish - Iceland Expeditions to Vatnajokull, Iceland, 1936 - 38, yielded the first accurate documentation of the drainage behavior, flood characteristics and long term changes in the state of Icelandic ice-dammed lakes (Thorarinsson, 1939a; 1939b; Jonsson, 1955). Other major papers which have contributed significantly to the knowledge of ice-dammed lakes are by Liestol (1955), Mathews (1956), Maag (1969), and Gilbert (1969). Advanced field techniques such as sounding and temperature gauging (Gilbert, 1969), and dye tracing (Fisher, 1973) have recently raised ice-dammed lake investigations to the level of a specialized branch of the field of hydrology.

Glacier-dammed lakes are reported from almost all major glacierized regions of the world, with a preponderance of literature on those lakes in the more populuous temperate zones: Iceland (Wright, 1935; Thorarinsson, 1939a; 1939b; 1953; 1957; Arnborg, 1955; Jonsson, 1955; Tryggvason, 1960), Norway (Strom, 1938; Liestol, 1955; Aitkenhead, 1960; Howarth, 1968; Whalley, 1971; 1973), Alaska (Russell, 1893; Stone, 1963a; 1963b; Moravek, 1968; 1973; Reid and Clayton,

1963; Seppala, 1973), British Columbia (Kerr, 1934; Marcus, 1960; Mathews, 1965; 1973; Gilbert, 1969; 1971; 1972) and the Alps (Rabot, 1905; Haefeli, 1970; Bezinge, Perreten and Schafer, 1973). Other glacierized regions of the temperate zones have received much less attention: South America (King, 1934; Helbig, 1935; Nichols and Miller, 1952), the Himalayas (Mason, 1935) and Spitsbergen (Vivian, 1965). Additional references, primarily brief mentions of the presence of ice-dammed lakes are cited by Charlesworth (1957).

Documentation of ice-dammed lakes associated with polar or sub-polar glaciers does not reflect the importance of these features in the high polar landscape. The interiors of the vast Antarctic and Greenland ice sheets are true polar glaciers or the dry snow zone (Paterson, 1969) and therefore no meltwater is available. However, the margins of these ice sheets where they abut mountainous terrain form the sites of many lakes. Fabricius (1788), Palsson (1882), Rink (1862), Weidick (1963), Helk (1966) and Higgins (1970) describe individual lakes in the south coastal region of Greenland, while Røen (1962) provides the most systematic examination to date, with brief descriptions of over 200 freshwater lakes - many of them ice-dammed. With the exception of several lakes reported on South Georgia (Brook, 1971; Stone, 1975), there are few references to Antarctic ice-dammed lakes. A type of supraglacial lake, commonly

termed an ice caldera, is referred to by Aitkenhead (1963) and Koerner (1965) for Grahamland, Antarctica, and by Paige (1968) for the McMurdo Ice Shelf. Koerner cites several earlier expeditions to George VI Sound and the Amery Ice Shelf where similar features, termed ice dolines, were observed. Davis and Nichols (1968) report several large proglacial lakes in the McMurdo Sound area, while the possibility of subglacial lakes was considered by Oswald and Robin (1973).

In the northern hemisphere, smaller masses of subpolar or possibly polar ice with accompanying glacier-dammed lakes exist only on the northern islands of the Canadian Arctic archipelago. Several large proglacial lakes (Lakes Conn, Beiler and Generator Lake) are ponded against the margin of the Barnes Ice Cap, Baffin Island (Holdsworth and McLaren, 1971; Holdsworth, 1973; Barnett and Holdsworth, 1974). The mountainous coastal region of eastern Baffin Island contains many ice-dammed lakes but only Church's (1972) work on Ekalugad Fiord area documents some of these features. No detailed analysis of ice-dammed lakes associated with the Devon Island Ice Cap has been performed. The glacier-dammed lakes of Axel Heiberg Island are the most completely studied set of polar ice-dammed lakes to date. Ricker (1962) and Maag's (1963) preliminary work give brief descriptions of ice-dammed lakes, but Maag's (1969) monograph provides general information on 125 ice-dammed

lakes in a 500 km² glacierized portion of central Axel Heiberg Island, and specific information on 4 ice-dammed lake and glacial drainage systems. Due to many similarities of observation and conclusions, Maag's work is referenced extensively throughout this paper. Ice-dammed lakes on Ellesmere Island are considered separately and in more detail in Chapter 2.3.

2.3 Ice-dammed Lakes on Ellesmere Island

With approximately 77,200 km² of glacier ice covering 60% of the 125,000 km² land surface of a predominantly rugged and mountainous island, innumerable sites for ice-dammed lakes exist on Ellesmere Island. The climatic conditions for the formation of glacier-dammed lakes are also ideal. The presence of sub-polar or polar ice prevents the subglacial or englacial drainage of ponded meltwater. The high density of ice-dammed lakes in Arctic regions has recently been documented (Hattersley-Smith, 1969; Maag, 1969; Wilkins, 1973), contradicting earlier claims by Stone (1963b) for south central Alaska. The great number of these features observed is due to the coincidence of appropriately rugged topography and efficent cold ice dams. The presence of numerous supraglacial lakes and smaller slush ponds uncommon on temperate glaciers attest to the cold thermal regime of Ellesmere glaciers.

The northernmost segment of Ellesmere, Grant Land, is the most mountainous, least accessible and least studied section of the island, and yet it has the greatest potential for ice-dammed lakes. Hattersley-Smith's association with the Defense Research Board of Canada provides the only reports of ice-dammed lakes in Grant Land. An unpublished note by Hattersley-Smith and Serson (1966) and Hattersley-Smith (1969a) discuss several lakes situated in the Tanquary Fiord area and to the north of Lake Hazen. Two of these lakes, Ekblaw and Rollrock Lakes, appear to drain supraglacially or marginally with a less than annual frequency. Christie (1967) discusses the sediments of an earlier Lake Hazen which was dammed to the east by ice descending from the United States Range.

The next of the four major physiographic subdivisions of Ellesmere Island, Grinnell Land, supports the largest continuous ice mass on Ellesmere Island, the Mer de Glace Agassiz. A density of 71 ice-dammed lakes in the 4,000 km² d'Iberville Fiord area was reported by Wilkins (1973). This air photo study found that 40% of all the lakes observed were supraglacial, and over 60% of the lakes were in the smallest size group documented (50,000 - 200,000 m²). The efficacy of cold ice dams is substantiated by such data.

An unusual example of an ectogenic meromictic ice-dammed lake, Lake Tuborg, is situated at the head of Greely

Fiord (Hattersley-Smith and Serson, 1964; 1966; Hattersley-Smith, 1969a). This lake was formed by the recent advance of a glacier cutting off the head of a fiord; the entrapped seawater resulting in the marked density stratification of the lake. The trapped seawater has been radiocarbon dated (Hattersley-Smith and Long, 1967) and used as an indicator of postglacial uplift. This technique has been applied to lakes dammed by the Ward Hunt Ice Shelf (Hattersley-Smith and Serson, 1966), and other non glacial lakes in northern Ellesmere Island (Hattersley-Smith et al, 1970).

There are no references to the ice-dammed lakes of Sverdrup Land, which contains the field area of the present study; except passing notice by Sverdrup (1904) and Humphreys et al (1936) of lakes ponded in Sverdrup Pass, the well traveled overland route from Bache Peninsula to Bay Fiord. The Prince of Wales Ice Field, more recently termed the Ellesmere Ice Cap, covers 18,750 km² or 50% of Sverdrup Land, and Taylor (1956) attributes its existence to the proximity of the anomalously ice free "North Water" located in northern Baffin Bay. The huge piedmont glaciers forming much of the eastern coastline of Sverdrup Land, and the abrupt western limit of the ice cap along the 82nd meridian support this statement.

A survey of the 1959, 30,000 foot flying height air photos reveals ~300 ice-dammed lakes with an approximate minimum size of 50,000 m². This yields a density of ~50

lakes per 100 km^2, a figure comparable to the results obtained by Maag (1969) and Wilkins (1973). This appears to be the highest concentration of ice-dammed lakes reported in the literature for a large area. Substantiation of the hypothesis that polar or sub-polar glacier ice supports a higher density of ice-dammed lakes than temperate ice should be sufficient now for this concept to be accepted. Similarly, the high percentage of supraglacial lakes, slush ponds and surficial drainage networks should indicate the absence of englacial or subglacial drainage systems.

A striking difference in the physiography and underlying bedrock configuration of Sverdrup Land has resulted in a dichotomy of glacier surface expression and ice-dammed lake distribution. The physiography is discussed further in Chapter 3.2, but the patterns of ice-dammed lake occurrence can be considered here. The western margin of the Ellesmere Ice Cap nourishes 6 major proglacial or marginal ice-dammed lakes. Most of these features occupy structurally controlled basins along the contact zone between Precambrian crystalline rocks and Paleozoic sedimentary strata. A few smaller lakes lie in topographic depressions on the ice surface, are proximal to nunataks or are the result of temporary ponding of marginal drainage systems. The group of lakes in the study area, McMaster Lake, Upper Lake and the Connecting Ponds, form a pocket of slightly higher density of lakes in the generally sparse

distribution along the western half of the ice cap. Another of these features, one of the largest ice-dammed lakes on the island, apparently occupies a structural depression at the confluence of three ice lobes (Figure 2.1). This lake was first sighted by members of Sverdrup's party, who stated that "the glacier lake was dammed by a barrier of fine material, which rose to a height of from 18-24 feet above the surface of the water. I saw no outlet from the lake."(p.174, Sverdrup, 1904).The heavily crevassed ice front, presence of ogives and contorted ice structures, and the calving of numerous ice bergs attest to the depth of this lake and the activity of the glacial system in general. As will be seen with the McMaster Lake example, much of the water in this lake appears to be trapped in the basin, with only a smaller percentage of the lake volume actively changing in a season. Due to the low divide between the drainage of this lake into Strathcona Fiord and the Sverdrup River catchment to the south, the possibility, though unexplored, that it may at some time have drained along the ice cap margin and into the Sverdrup River should be kept in mind. Finally, a lake at the head of Makinson Inlet appears to be similar to the Lake Tuborg example mentioned by Hattersley-Smith (1969a), where an advancing glacier has cut off the head of the fiord, possibly trapping saline water.

In direct contrast, the rugged eastern portion of the Ellesmere Ice Cap supports a much higher density of ice-

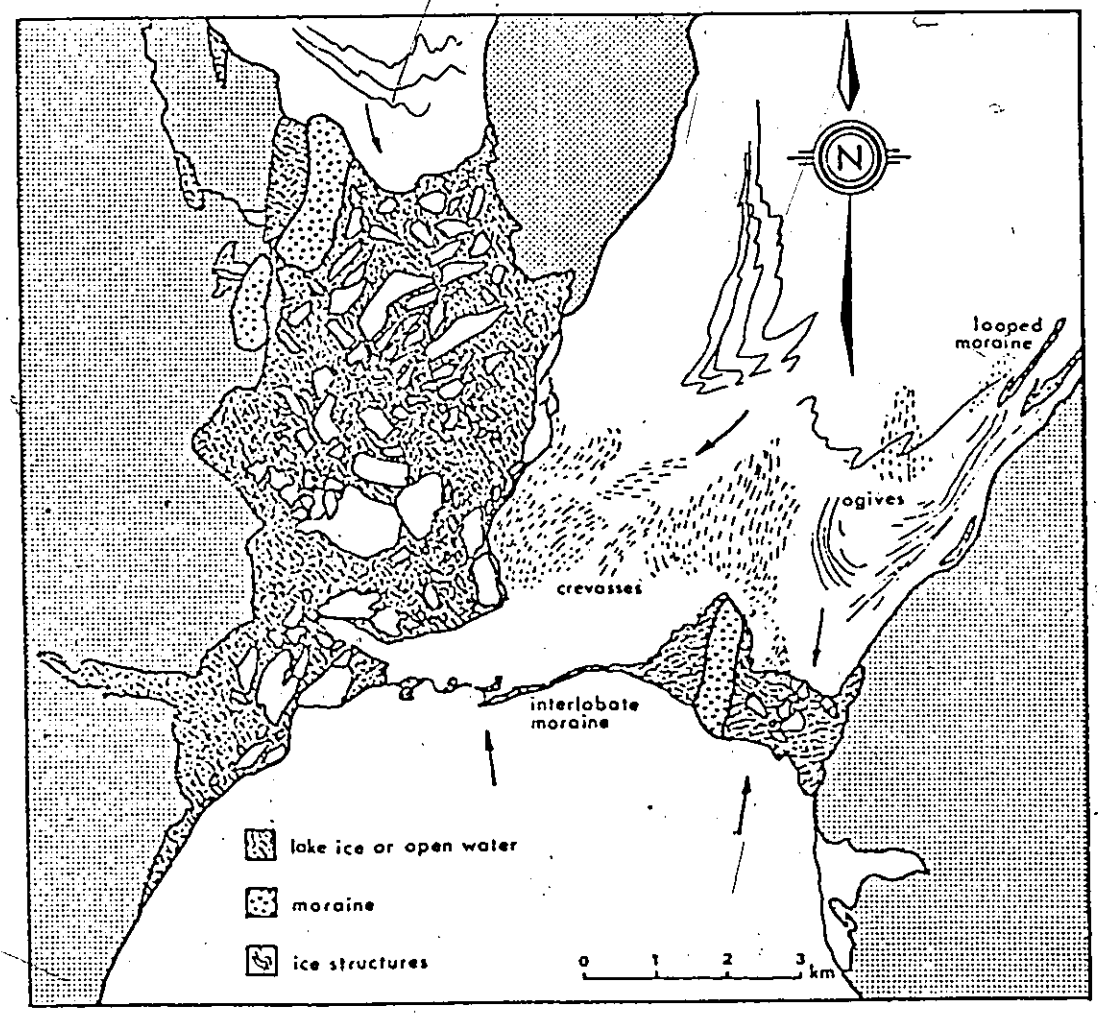


Figure 2.1 Large interglacial lake 25 km. north of McMaster Lakes area

dammed lakes, predominantly in supraglacial or marginal positions, or associated with nunataks. The large trunk valley glaciers with numerous tributary glaciers, such as the Talbot, Cadogan, Ekblaw, Leffert and Stygge Glaciers have large numbers of ice-dammed lakes and other melt phenomena such as slush ponds and crevasse fillings associated with them. A feature common along the eastern margin of Sverdrup Land and possibly indicative of former glacier surges (Hattersley-Smith, 1969b) is the large numbers of small water filled potholes and distorted drainage networks on several of the lower snouts along the Baffin Bay coast. Another possible factor is the proximity of the open North Water, where the formation of superimposed ice is thought to be a major form of accumulation to the lee of the North Water. This source of humidity could provide a good supply of melt water in the summer months, but some accompanying hypothesis must be invoked to explain the surficial irregularities capable of trapping available meltwater.

Finally, the southernmost segment of Ellesmere Island is termed Lincoln Land, and again the ice cap is largest to the east due to the influence of the North Water moisture source. Lakes were identified in this region during glacier inventory mapping for the Glaciology Division, Environment Canada, but there are no published references to the ice-dammed lakes of Lincoln Land.

CHAPTER 3

McMASTER LAKES STUDY AREA AND DATA COLLECTION TECHNIQUES

The McMaster Lakes lie in the headwaters of the Sverdrup River basin (Figure 1.2), which drains a 1630 km² area of barren tundra and 40 km of the margin of the Ellesmere Ice Cap. Glacier ice comprises 77%, or 1250 km² of the Sverdrup basin, and >90% or 150 km² of the McMaster Lakes basin (Figures 3.1, 3.1A). A brief discussion of the historical background, geology, physiography, and general glaciology of the study area is presented, and the field program and data collection techniques are outlined in this chapter.

3.1 Historical Background

Members of the 1899-1903 Norwegian 'Fram' expedition (Sverdrup, 1904) were apparently the first men to see the Vendom Fiord area. Isachsen and Braskerud crossed the Ellesmere Ice Cap from Leffert Glacier on Smith Sound to the large lake (78°30'N;81°30'W) 20 km north of the McMaster Lakes. They observed to the south-east "... a wide, level mountain waste-Braskerudflya..., abutted immediately on the island-ice" (p.174, Sverdrup, 1904). The following year (1901) Fosheim, Shei, Stolz, and Sverdrup sledded overland

Figure 3.1 Vertical airphoto of the McMaster Lakes study area (30,000 ft. flying height)

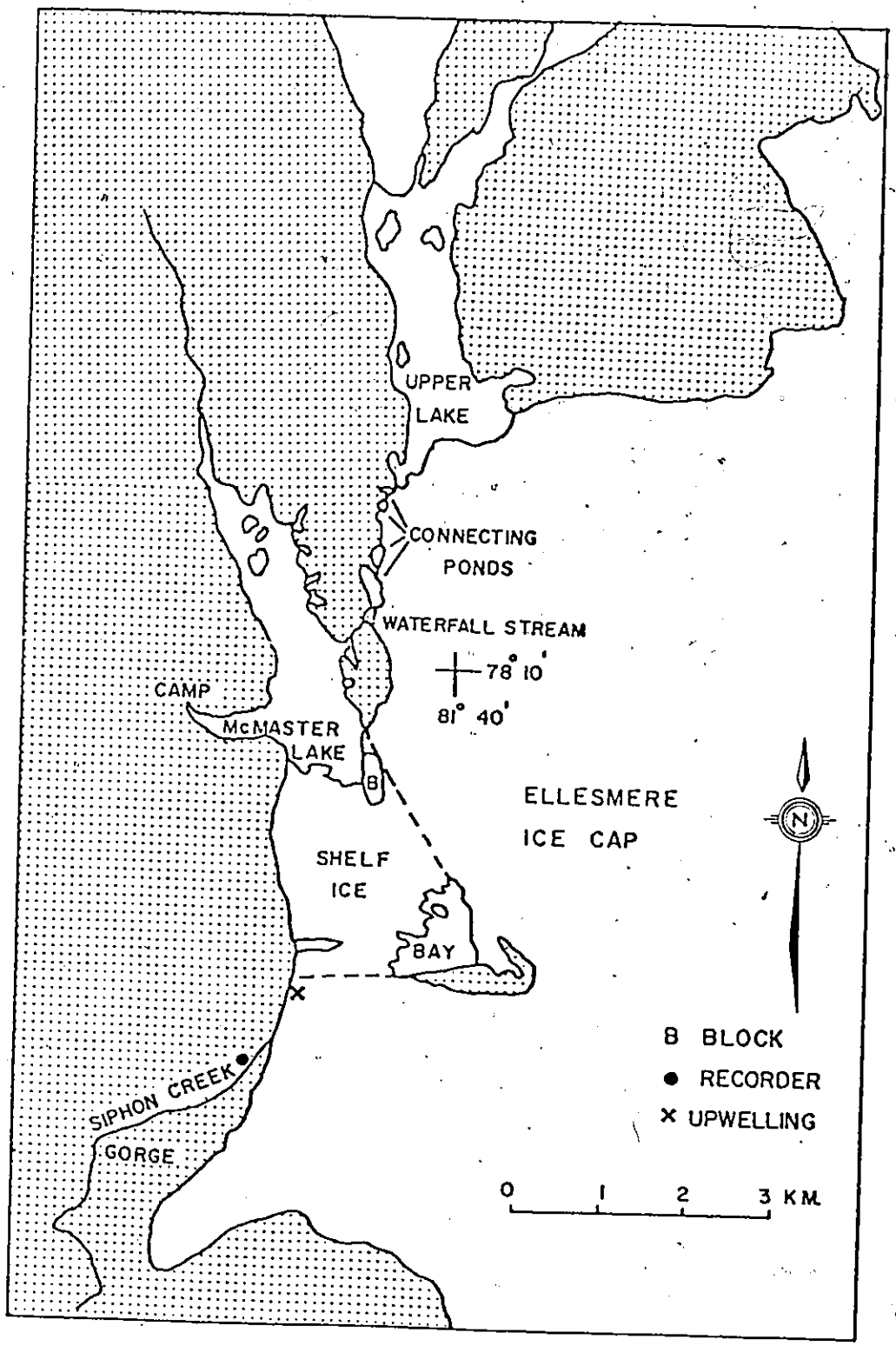


Figure 3.1A Map of the study area with place names

from Goose Fiord on the southwestern corner of Ellesmere Island, and followed Vendom Fiord to within 20 km of the head of the fiord before turning back. In 1935, members of the Oxford University Ellesmere Island Expedition (Shackleton, 1937) travelled southwards from Bay Fiord and claimed to have reached the head of Vendom Fiord. It was proven later, however, that this expedition actually saw Troll Fiord, 80 km to the north-west. During Operation Franklin (Fortier et al., 1963), several members of the expedition visited the Vendom Fiord area and the Braskeruds Plain, while Norris compiled the detailed surficial geology of a small area to the head of Vendom Fiord. There is no other record of exploration in the Vendom Fiord area until the 1970's, when field parties of the Geological Survey of Canada (Hodgson, 1972) and McMaster University (McCann et al., 1972; 1974; 1975) visited the area.

3.2 Geology and Physiography

Seven major geological provinces make up the bedrock geology of the high Arctic (Thorsteinsson and Tozer, 1970). The Vendom Fiord study area lies close to the boundary of the Franklin Miogeosyncline, the Arctic Platform or Lowland, and the Canadian Shield (see Figure 3.2).

The crystalline basement rocks of early Precambrian age belonging to the Baffin-Ellesmere Precambrian belt (Fortier, 1957) and are thought to be continuous with the

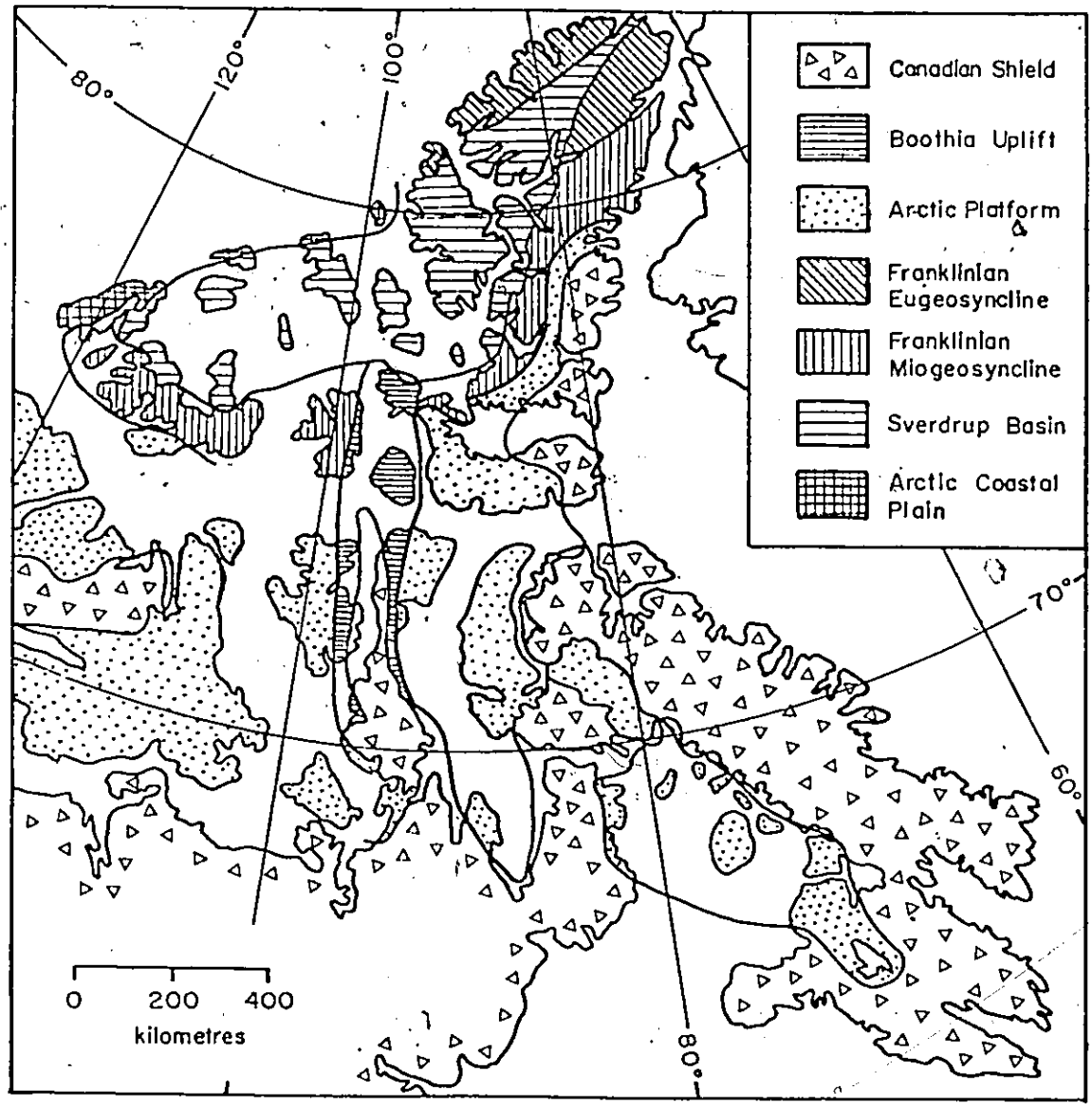


Figure 3.2 The geologic provinces of the Canadian Arctic Archipelago (after Prest, 1970)

mainland Shield. Adjacent to, and apparently conformably overlying the Precambrian basement rocks, flat lying or little disturbed lower Paleozoic strata are exposed. These lowlands represent remnants of the northern continuation of the Inland Plains of the continental mainland (Thorsteinsson and Tozer, 1970).

In the McMaster Lakes basin, rocks ranging in age from Precambrian to Tertiary are found, as well as unconsolidated Quaternary deposits (Figure 3.3). Precambrian gneissic and granitic rocks, apparently subjected to high grade (granulite facies?) metamorphism, outcrop in the extreme eastern portions of the study area, mostly as nunataks in the icecap. Westward dipping outcrops of lower and middle Cambrian and lower Ordovician limestones, dolomites, and minor evaporites outcrop between Upper Lake and McMaster Lake. These strata are probably remnants of the Arctic Platform province, and are separated from the Precambrian and geosynclinal rocks by normal faults.

The lower Paleozoic geosynclinal sediments occur on the east limb of the Vendom syncline, with beds dipping 10 to 60 degrees to the west (Norris; in Fortier et al., 1963). The stratigraphic succession represented ranges from the Lower Ordovician Baumann Fiord formation through the Middle Ordovician Cornwallis Group, to the Middle and Upper Devonian Allen and Reid Bay, and Vendom Fiord formations.

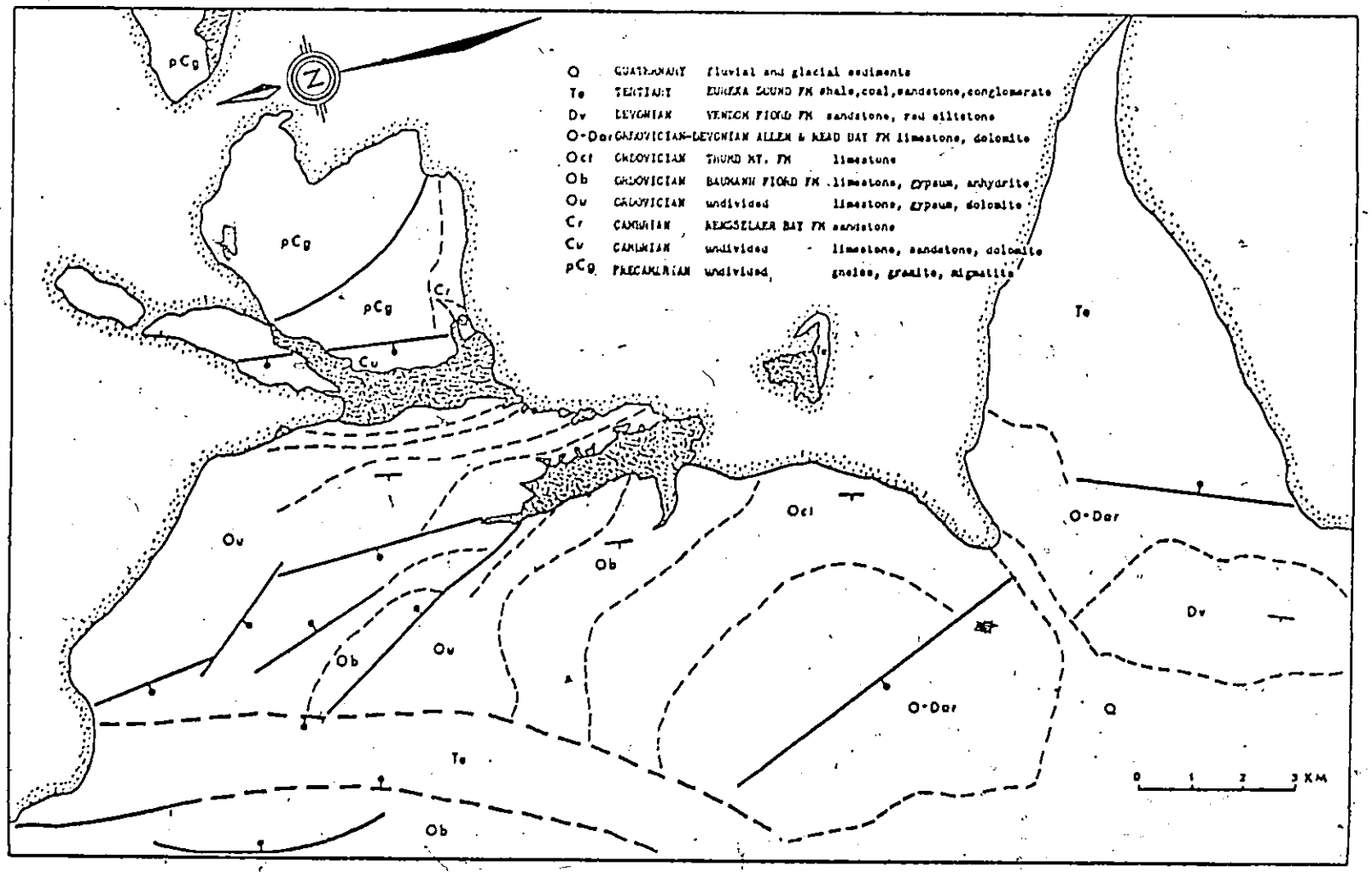


Figure 3.3 Geology of the McMaster Lakes study area

Limestones and dolomites dominate the lower part of the succession, with lesser amounts of evaporites, sandstone, shale, and siltstones, and minor amounts of clastic red-bed Vendom Fiord Formation (Kerr, 1967a). Faults of predominantly north-south strike apparently mark the boundaries between the geosynclinal outcrops and the Tertiary Eureka Sound sediments. Several prominent escarpments of Cornwallis Group limestones and dolomites are found in the study area.

Quaternary material of fluvial and glacial origins is found within the study area, with the Sverdrup River floodplain and large coarse-grained alluvial fans being the dominant features. Glacial erratics of Precambrian crystalline material are present at all elevations, indicating glacial advance over a much larger area at some time in the past. The geomorphic processes represented in the study area include fluvial and periglacial activity, glaciation, mass-movement, nivation, and chemical weathering, in approximate descending order of importance. Stream dissection and cryogenic features dominate the landscape, while active nivation hollows are a common proglacial feature.

The physiography of the study area is primarily a function of the bedrock geology and associated structural features. The McMaster Lakes are set in a structural depression between the gradually sloping ice cap to the east

and the 300-500 m high remnants of the Tertiary erosional surface to the west, the Braskeruds Plain (Hodgson, 1973). McMaster Lake and Upper Lake are elongated, north-south oriented lakes owing their presence to the Ellesmere Ice Cap, which serves to pond the water against the ice margin and provide their main source of nourishment. The steep margins of the lake basins are the dominant relief features of the area, with the notable exception of the anomalously deep McMaster Gorge. This 5 km long, 200-300 m deeply incised canyon appears to be the product of high magnitude flow events, possibly catastrophic drainage.

3.3 Glaciology

The study area occupies a small portion of the western margin of the large, lobed Ellesmere Ice Cap which covers approximately 18,750 km² of the south-central Ellesmere Island land mass, or Sverdrup Land. The configuration of this ice mass closely reflects the underlying topography, with the relatively low relief Paleozoic and Cenozoic plateaux and hills causing the western ice margin to be mainly flat and continuous. In contrast, the southern and eastern margins of the ice cap are much more rugged, with numerous steep-walled nunataks, and large valley and outlet glaciers, reflecting the nature of the underlying mountains of resistant Precambrian strata. This contrast is well illustrated in the Frontispiece, with

the broad, unbroken lobes of the western margin in the foreground; and the rugged eastern coast of Ellesmere Island visible in the distance.

Along most of the western ice margin, including the Vendom Fiord study area, the ice has little topographic expression, and generally low gradient glacier snouts and ice margins. The Schei Glacier lobe, about 10 km long by 3 km wide, is slightly larger in size but typical in configuration and profile of much of the western margin of the Ellesmere Ice Cap. The much smaller McMaster Glacier snout is about 4 km long by 1/2 km wide, and lies in a deep, confined valley with side walls 150-200 m high. The average gradient of the ice cap in the Vendom Fiord area is about 50 m/km, rising from the lowest point, the Schei Glacier snout at 25 m a.s.l. to about 1525 m a.s.l., which marks the east-west divide of the ice cap. The gradient of the McMaster Lakes drainage basin is somewhat steeper, approximately 70 m/km, due to the large, steep outcrop of Precambrian rock.

These topographic features, combined with the low underlying relief, apparently small glacier mass budget, and some well established proglacial vegetation (SALIX ARCTICA) colonies, indicate that the ice front is presently inactive. The greatest measured change was an advance and/or ice fall amounting to 5 m of lateral displacement on the southern margin of the Schei Glacier (Ballantyne, in: McCann et al.,

1975). The smaller McMaster Glacier snout has a steeper than average slope down into the valley below, and as a result, this snout is advancing slightly. The absence of any major proglacial depositional features suggests that the present period of relative inactivity has persisted for a considerable length of time. The relatively sediment free nature of the glacier ice and low ice temperatures probably account in part for the lack of depositional landforms characteristically associated with an ice margin. Several segments of relict terminal or end moraine are evident between 250 and 500 m in front of the Schei Glacier snout, but these apparently represent remnants of a small scale feature. A few examples of medial moraines or possible near-surface bedrock outcrop were found on the ice tongue feeding the north-east arm of Upper Lake.

A small terminal or Thule-Baffin type moraine (Embleton and King, 1968) is actively forming from the melting-out of debris in the McMaster Glacier snout (Figure 3.4). Shear planes or debris layers are generally associated with the more active segments of the ice margin, and most noticably along the southern margin of the Schei Glacier, where large dirt bands are visible (Figure 3.5). In cross-profile, however, these debris planes curve upwards, and many exhibit contorted, broken or sheared patterns. The contortion and fractures can be attributed to the increased internal ice deformation and movement near the

Figure 3.4 Small terminal or shear moraine forming at the McMaster Glacier snout

Figure 3.5 Dirt bands in the Schei Glacier ice front

glacier snout (Paterson, 1969). The controversy of shearing mechanisms (Goldthwait, 1951; Bishop, 1957) versus regelation incorporation of basal debris (Weertman, 1961; Bolton, 1970) for the formation of englacial debris is still unresolved.

A small amount of modern lodgement till, ≤2 m thick, is incorporated into the basal layer of the Schei Glacier, but this was observed in relatively few locations. More commonly, the base of the glacier is frozen directly to a permafrost or bedrock surface. Minor amounts of ablation till are being deposited off the north and west margins of the Schei Glacier, as layers of coarse angular fragments of Precambrian metamorphic and igneous rocks are exposed on the glacier surface. The provenance and distance of travel of the rock is unknown, but the contact of the Precambrian rock province of eastern Ellesmere with the local Paleozoic sediments is 25-50 km east of the Schei Glacier (Thorsteinsson, 1972).

Possibly two of the most important characteristics of an ice sheet are ice temperature, and the amount of meltwater produced. A recurrent theme in this paper is the importance of ice temperature in controlling the nature of the glacial drainage systems and the incidence and behaviour of ice-dammed lakes. Along the 40 km section of ice margin studied at Vendom Fiord, no true examples of englacial or subglacial drainage were noted. Almost all free water

movement is confined to the glacier surface and margins. Examples of sub-marginal drainage (Maag, 1969) are fairly common, where the meltwater erodes a slot under the margin of the ice cap often giving the appearance of an emerging subglacial stream. Similarly, the dominant form of drainage, supraglacial streams, are also capable of eroding into the glacier and appearing again in an apparently englacial position. Examples of supraglacial streams with several levels of water flow were observed in the study area. The possible genesis of such a feature is the downcutting of a surface stream, snowdrift infilling of most of the channel in the winter season, and subsequent generation of flow at two levels; one on the new snow surface, and another reoccupying the lower ice floored stream bed. An analogy to this phenomenon could be drawn from the simultaneous flow of water at several depths in karst regions.

3.4 Data Collection Techniques and Instrumentation

In 1973, the McMaster Lakes camp was occupied for two weeks (July 30 to August 11), when a preliminary reconnaissance was carried out and measurements of the level in McMaster Lake were obtained. In 1974, the same camp was occupied for two months (June 19 - August 21), and the work program involved investigation of the limnology of McMaster Lake, the hydrology of Siphon Creek, the glaciology of the

margin of the ice cap, and adjoining shelf ice, and the collection of standard meteorological data.

A meteorological station was established at the camp site (elevation 380 m a.s.l.), and maintained for the entire field season. Air temperature and relative humidity were recorded continuously by a Lambrecht thermohygrograph; calibrated by wet and dry bulb mercury thermometers (see Figure 3.6) and housed in a Stevenson screen. Incoming solar radiation was obtained from a Casella bimetallic actinograph, which was calibrated against numerous solarimeter readings (Figure 3.7). Precipitation was recorded at the campsite with an Atmospheric Environment Service tipping bucket raingauge. A non-recording M.S.C. type rain gauge was maintained on the glacier surface at 500 m a.s.l. and read weekly. Routine weather observations were made twice daily, including sky condition, percentage of cloud cover, cloud types, wind speed, and wind direction.

Discharge in Siphon Creek was recorded from July 15 to August 9 using a Leupold-Stevens Type F water level recorder, and a siphon type installation as described by Church and Kellerhals (1970) (Figure 3.8). This type of setup was employed in anticipation of rapidly rising water levels likely to occur during a jokulhlaup. A rating curve for the automatic recorder was established using a Price current meter mounted on a wading rod, and the dye dilution technique using Rhodamine-B dye (Figure 3.9). Several

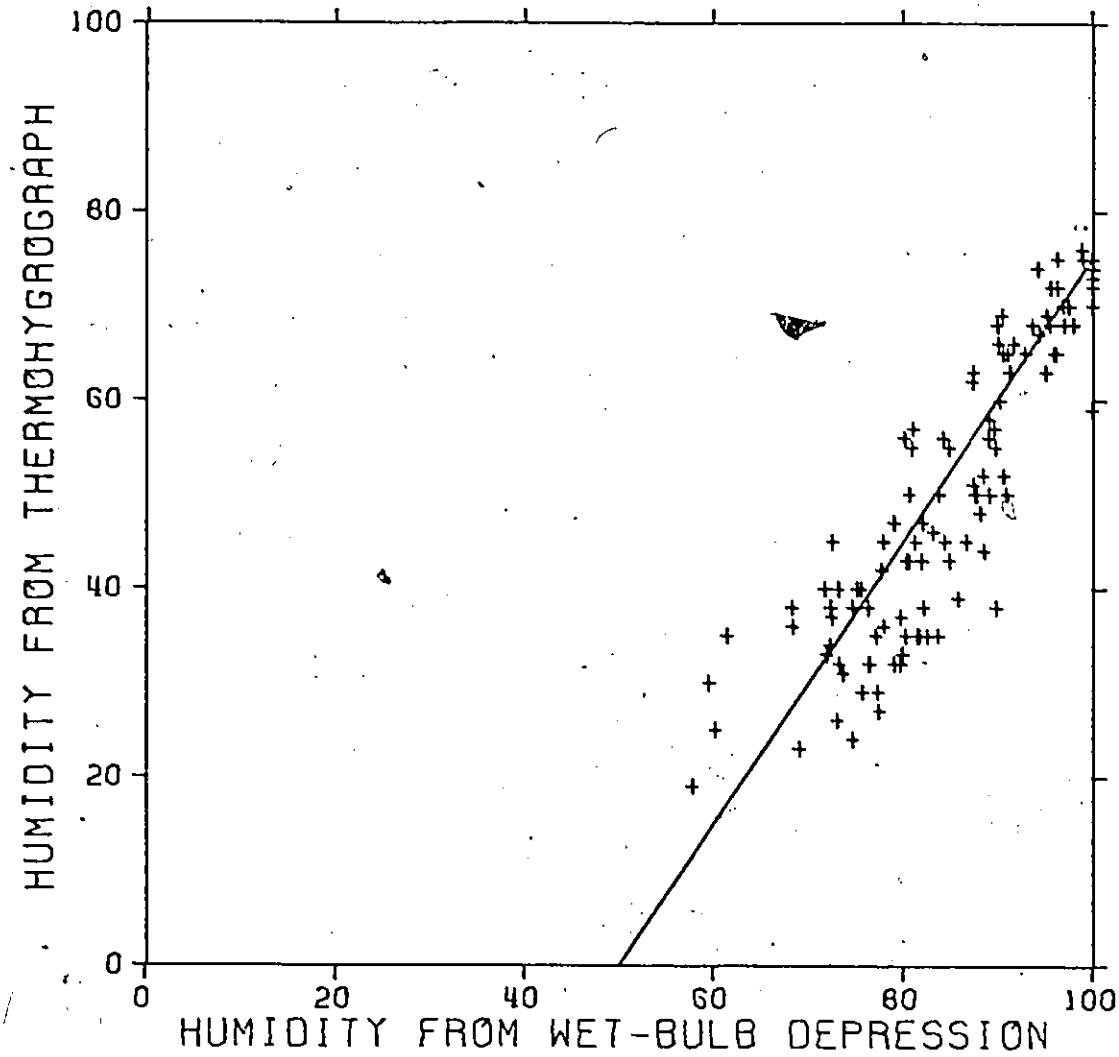


Figure 3.6 Calibration of humidity, McMaster Lakes, 1974.

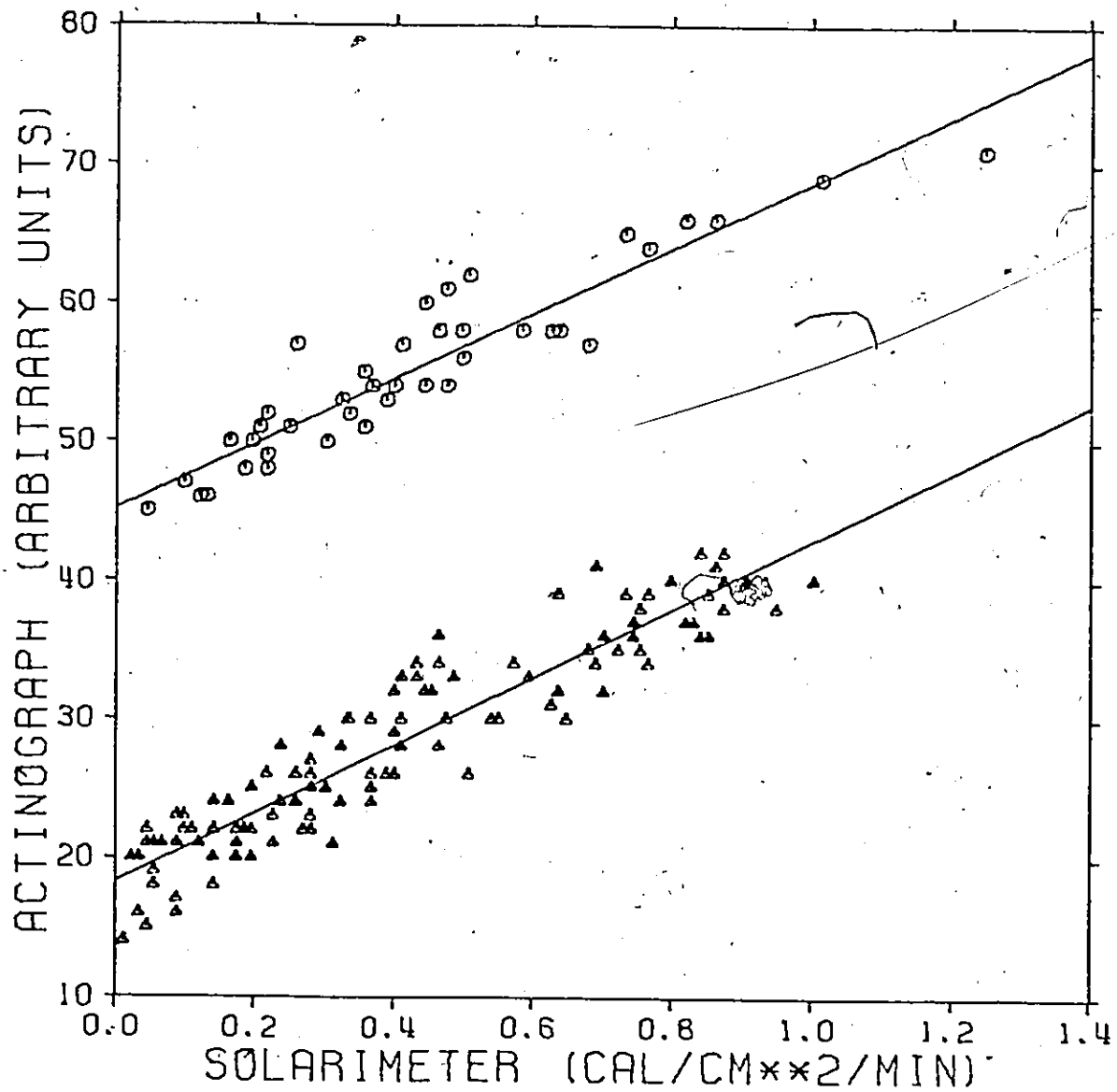


Figure 3.7 Calibration of solar radiation, McMaster Lakes, 1974.
Δ's represent readings taken after instrument reset on June 28.
o's readings taken June 21 - 28.

smaller streams and supraglacial streams were gauged using both the conventional area-velocity and dye dilution methods. A complete discussion of the discharge measurement techniques employed is given by Church and Kellerhals (1970).

The waterlevel of McMaster Lake was monitored twice daily until the jokulhlaup occurred (August 13), after which time water level was observed four times a day. Lake depths were obtained by lowering a weighted line through natural openings and cracks in the lake ice cover, and several holes were drilled with a 3/4" diameter manual SIPRE ice auger. Water temperatures were measured with thermistors soldered to insulated cable and then waterproofed. Resistances were measured with a portable Hewlett-Packard Wheatstone type bridge. The thermistors were calibrated to 0.002°C in a temperature bath at the Heat Flow Laboratory, Earth Physics Branch, EMR in Ottawa (Figure 3.10). The resistance bridge and field procedures were sufficiently sensitive to allow temperatures to be read to an accuracy of 0.01°C.

Ice temperatures were measured using the same thermistor cable and resistance bridge setup described above. Three and four metre deep holes were drilled into glacier ice using the 3/4" diameter SIPRE ice auger for the multi-thermistor cable. A considerable amount of methyl alcohol and constant drilling was necessary to prevent the auger from being frozen into the very cold ice; a lesson

Figure 3.8 Stage recorder installation of Siphon Creek during low flow

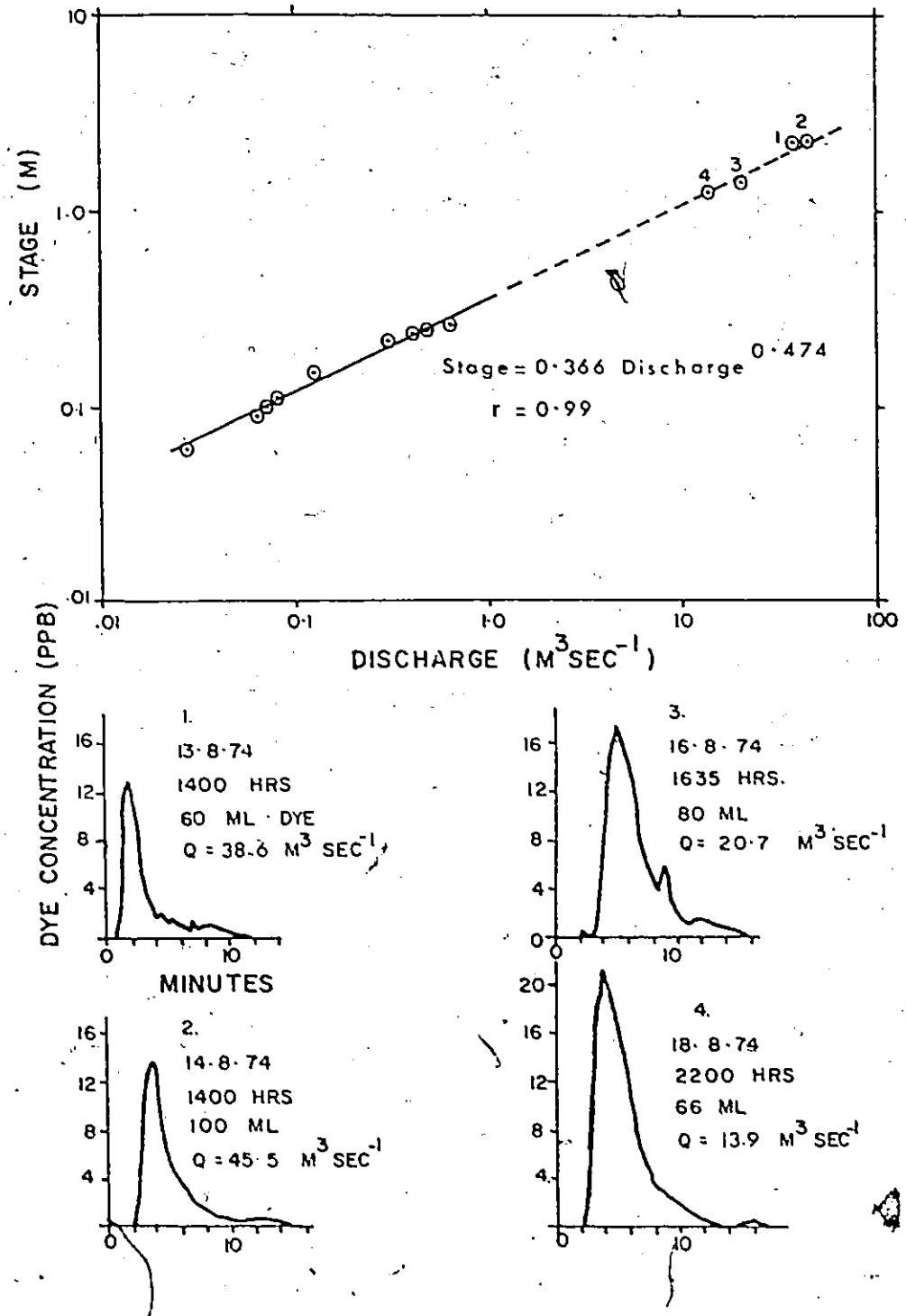


Figure 3.9 Stage - discharge rating curve, Siphon Creek, July August, 1974.

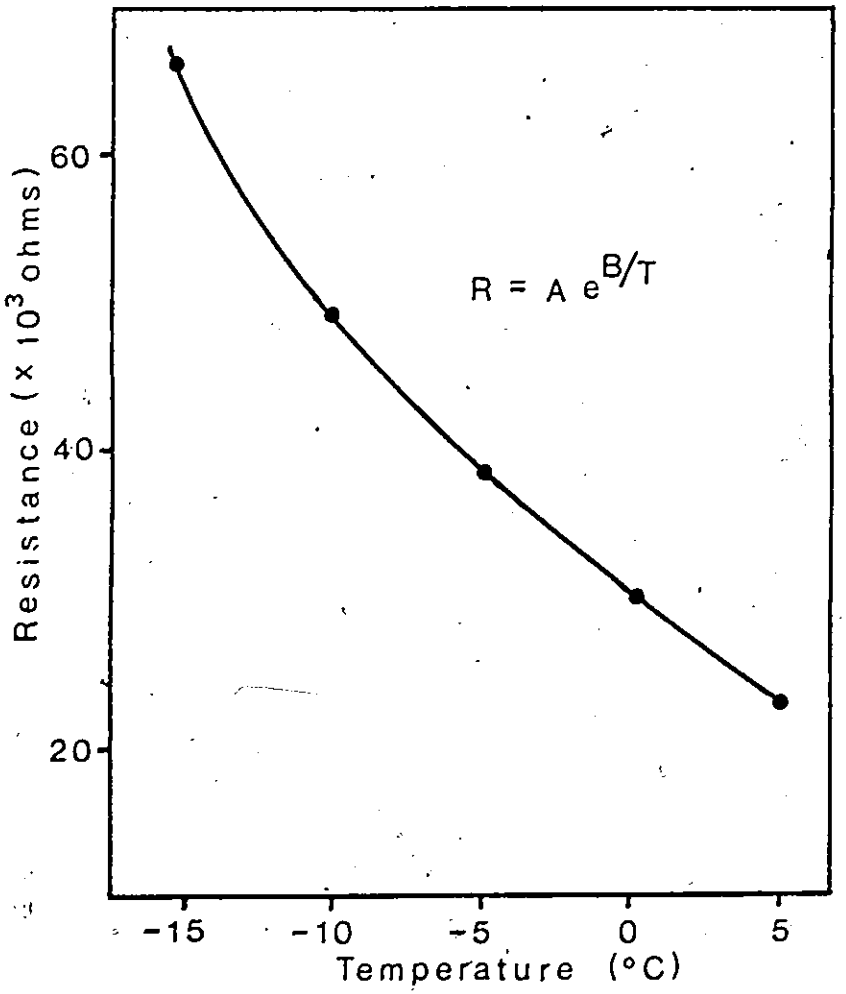


Figure 3.10 Sample calibration of a thermistor

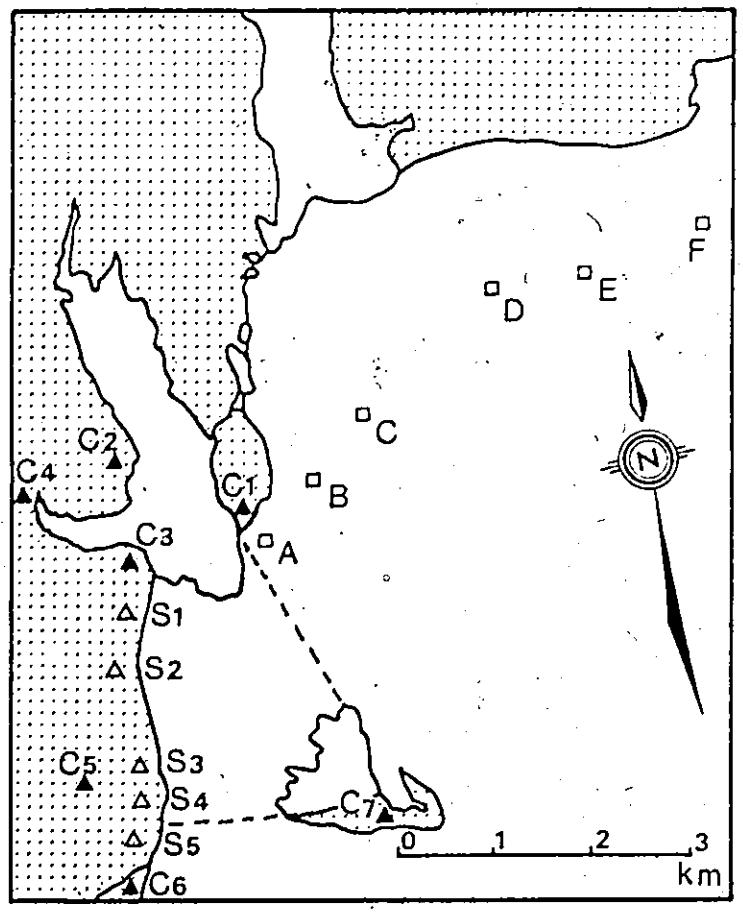


Figure 3.11 Plot of surveying baseline network, ▲ = triangulation cairns, △ = strain net survey stations, □ = ablation stakes

learnt the hard way after digging a 1x1x2 m pit in the ice to extract the only available auger. For ablation measurements, a series of six 1" diameter aluminum stakes were frozen into the ice about 2 m, in holes drilled with a 1 1/4" diameter bit for the SIPRE auger. The lowering of the snow or ice surface relative to the stake top and other snow depths were monitored approximately weekly throughout the field season.

To support the other field programs a network of stone cairns was erected and surveyed with a Wild T-2 theodolite to provide a baseline for other measurements (see Figure 3.11). Surveys of lake levels, ablation stake positions and their seasonal displacement, strand line and shelf ice surface profiles were all tied into the established grid. An extensive network of 1" diameter aluminium stakes was set into the surface of the shelf ice and adjoining ice-cap, and surveyed several times during the season to provide data on the three dimensional displacement of the shelf ice mass.

CHAPTER 4

LIMNOLOGY OF McMASTER LAKE

Despite the fact that limnology is a well established and well documented science (Hutchinson, 1957), research into the nature and behaviour of ice-dammed lakes has been limited. The earliest works are mostly concerned with the remarkable catastrophic drainage events which characterize many ice-dammed lakes (Munro, 1892; Rabot, 1905; Mason, 1935). More complete studies of ice dammed lakes have been made by Thorarinsson (1939), Liestøl (1955), Maag (1969), and Gilbert (1971). The present study is concerned with the origin of the lake basin, the thermal regime of the lake, the filling and drainage behaviour, and the water balance of a high Arctic ice-dammed lake.

4.1 Origin and Morphology of the Lake Basins

The position, orientation and depths of McMaster and Upper Lakes appear to have a strong structural control (Thorsteinsson, 1972). The lakes lie in a major geologic fault zone roughly marking the contact of the Paleozoic sedimentary strata and the undifferentiated Precambrian granites and gneisses (Figure 3.3). The high, near-vertical walls of the basins, widely varying strike and dip angles on opposite shores, and the marked vertical displacement of the

western shore of Upper Lake all indicate the structural origins of these basins (Figure 4.1). The sharp, unrounded cliffs, deep V-shaped valleys, and thick regolith negate the probability of any significant glacial entrenchment of the lake basin. The orientation of the basins at right angles to the main ice mass, and the differential displacement of the shores of Upper Lake imply a structural rather than glacial erosional origin for the lake basins. McMaster and Upper Lake would be considered tectonic basins of the fault trough type (Type 9) in Hutchinson's (1957) classification scheme.

The Ice Cap is responsible for ponding the water in McMaster Lakes to its modern-day levels, but due to the depth of the rock basins, a substantial portion of the water would remain if the glacier retreated from its present position. The height of the damming rock lip which contains the lake and therefore the depth of this non-glacial lake can not be determined precisely without detailed information on the subglacial topography. As will be seen in Chapter 6, the height and nature of the ice dam plays an important part in controlling lake levels. Thus, the bedrock structural geology and the proximity and state of the Ellesmere Ice Cap combine to form the McMaster Lakes. The particular combination of circumstances which produced these lakes would at first seem restrictive and reduce the incidence of

Figure 4.1A View south of McMaster Lake

Figure 4.1B Looking north from the glacier onto Upper Lake

such ice-dammed lakes, but in fact they are quite common along the margins of the Ellesmere Ice Cap.

The main part of McMaster Lake is 3.5 km long by an average 0.8 km wide (2.80 km^2), and the inclusion of the portion of lake under the shelf ice and the Bay gives a total maximum surface area of 7.2 km^2. A network of 53 spot lake depths was used to produce a bathymetric chart of the McMaster Lake basin (Figure 4.2). Conjectural depth contours are continued under the floating shelf ice and into the Bay, on the basis of a small number of spot depths, and the fact that the lake attains its greatest known depth (195 m) at the shelf ice margin (see profile C-C', Figure 4.2). Assuming an average depth of 100 m for the lake, the total maximum volume of McMaster Lake would be about 7x10 m^3. Two spot depths were obtained along the midline of Upper Lake; 81 m in the north-central part of the lake, deepening to 125 m in the south-central region. A trend of deepening towards the south similar to that observed in McMaster Lake can be postulated. The results of depth sounding appear to confirm the view that McMaster Lake is a tectonic basin associated with several fault scarps.

4.2 Thermal Properties of the Lake

The thermal regime of lakes is an important and well documented study (Hutchinson, 1957), but relatively few observations have been made on ice-dammed lakes, despite the

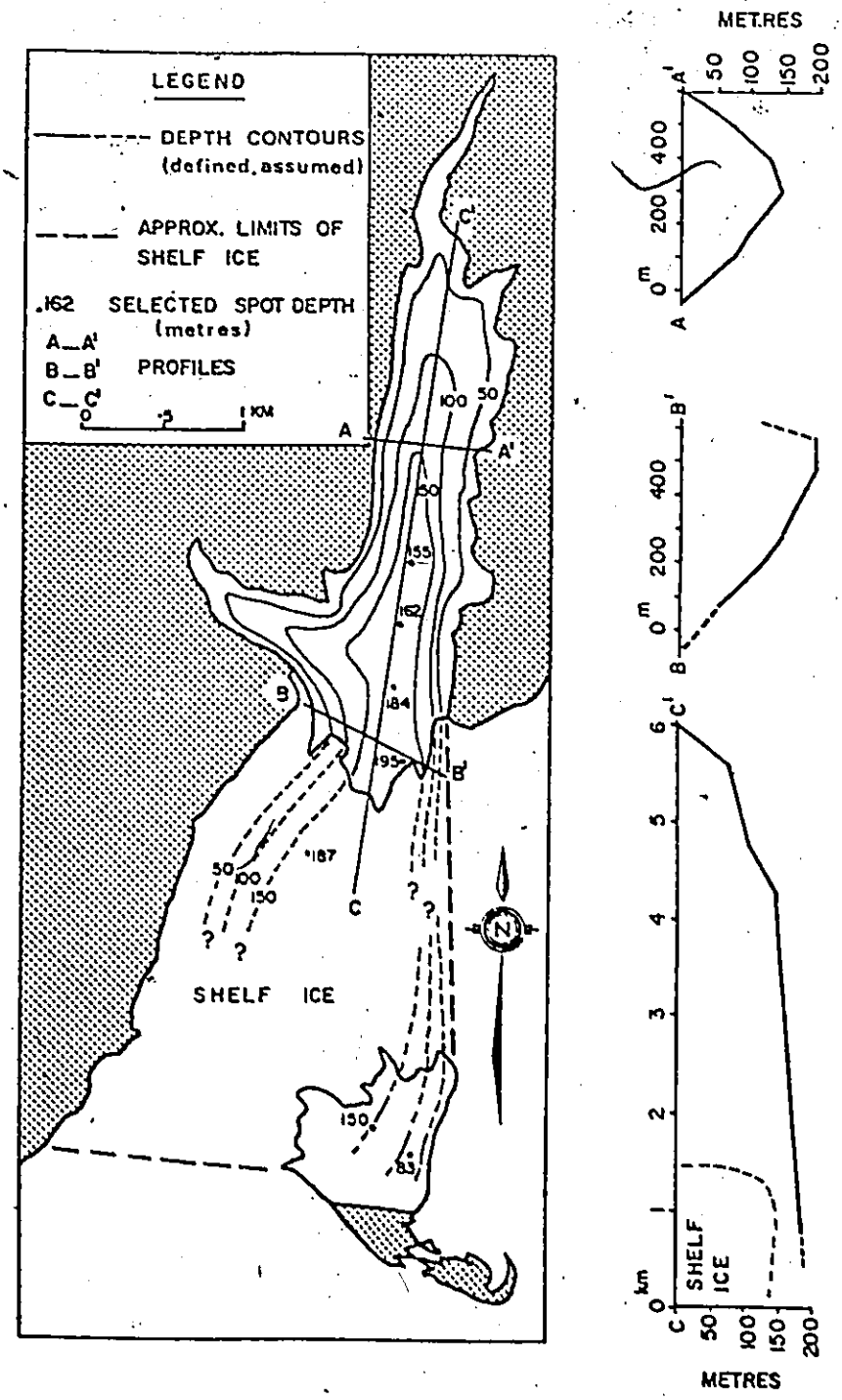


Figure 4.2 Bathymetric map of McMaster Lake, with profiles A - A', B - B', C - C'

apparent significance of water temperature to lake drainage mechanisms (Gilbert, 1972). Liestøl (1955) was the first to consider the importance of temperature measurements in the study of the behaviour of ice-dammed lakes. Other studies include the works of Hattersley-Smith and Serson (1964), Haag (1969), Gilbert (1972), Gustavson (1972), and Mathews (1973). Accurate measurements of lake water temperatures were made in McMaster Lake in view of the lack of quantitative information on the thermal properties of ice-dammed lakes, and particularily because of the recent work by Mathews (1973) and Gilbert (1972). These workers demonstrated the importance of the thermal regime in the behaviour of ice-dammed Summit Lake, British Columbia. As will be discussed in Chapter 7 on drainage mechanisms, a small but critical change in lake water temperature is often sufficient to initiate and/or maintain outlet tunnel enlargement and lake drainage.

Lake water temperature profiles were recorded regularily thoughout the season with a thermistor probe and resistance bridge at various locations and depths in the lake. They indicate near isothermal conditions for the entire depth of the lake, and with no apparent trend over the two month summer season. Temperatures were consistently between 0.0 and 0.5°C, with only minor variations along the entire depth profiles (Figure 4.3).

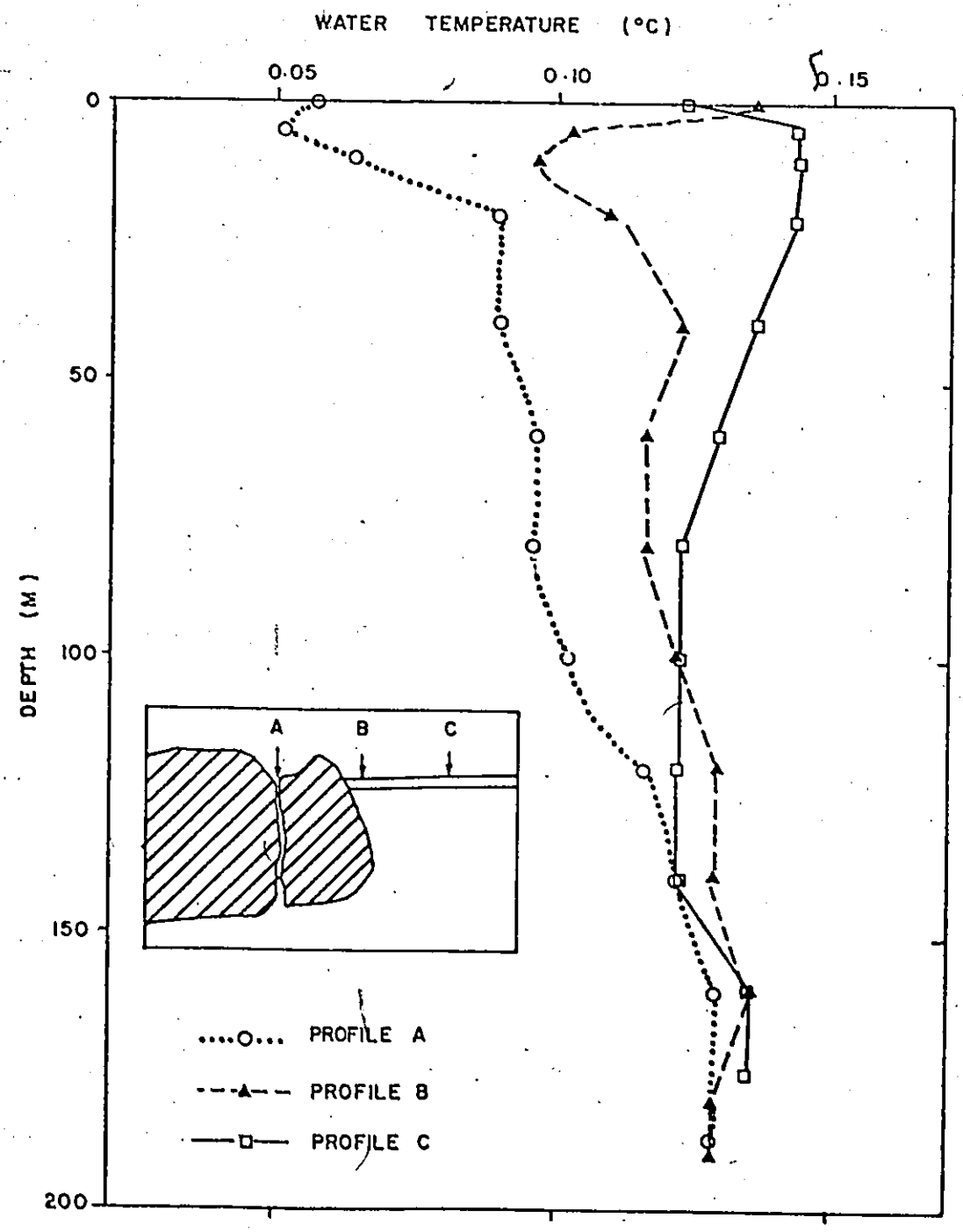


Figure 4.3 Lake water temperature profiles, McMaster Lake.

Factors which control the thermal regime of an ice-dammed lake are contact with the adjacent glacier, low influent water temperatures, short summer season, presence and persistence of lake ice cover, and often sheltered sites. In McMaster Lake all these factors combine to produce and maintain the uniformly low water temperatures. The most significant source of variation was proximity to the large floating shelf-ice mass, which slightly depressed the temperature (Figure 4.3). Profile A was taken through a crack in the shelf ice, and the temperature was approximately 0.05°C lower for the upper 100 m, converging with the other profiles near the apparent base of the shelf ice. Temperature profiles were also taken in Upper Lake, the Bay, and the largest of the Connecting Ponds, with almost identical results to those described for McMaster Lake.

The essentially ephemeral nature of most ice-dammed lakes presents problems when attempting to utilize a thermal classification scheme such as Hutchinson's (1957). The broadly defined categories of amictic, monomictic, and dimictic can be applied to ice-dammed lakes, with most of these lakes falling into the category of cold monomictic. This category is characterized by a very short ice-free period, circulation only at the height of the summer season, a small thermal gradient, little or no development of a thermocline and surface temperatures always below 4°C. The

McMaster Lakes, however, fit the classification of the less common true polar, or amictic lakes,. Amictic lakes are defined by Hutchinson (1957) as having a temperature of less than 4°C year round, essentially isothermal profiles, with no thermocline development, and the lake ice cover persisting the entire year in most seasons. Examples of amictic lakes have been noted in Antarctica (Hutchinson, 1957), Greenland (Katz, 1953; Røen, 1962), and Axel Heiberg Island (Maag, 1969). No references to amictic ice-dammed lakes are found in the Ellesmere Island literature, but this phenomenon appears to be more common than Hutchinson suggests.

The presence of a semi-permanent lake ice cover, a condition included in the definition of an amictic lake, appears to exist for McMaster Lakes. With the exception of open shore leads, a solid lake ice cover was present in all years of airphoto coverage (1950, 1959, 1972, 1973, 1974). The ice thickness was consistently 3 m thick in July, 1974, and approximately 0.5 m of ice was melted off the surface during the summer, with an undetermined but probably negligible (Barnes and Hobbie, 1960) amount being lost off the underside of the ice pack. The lake ice consists of large vertical ice crystals (candles) with crystal size diminishing with depth. The movement of icebergs in McMaster Lake and Upper Lake was traced from the aerial photography (Figure 4.4). The glacier snouts reaching

McMaster and Upper Lakes are not very active, producing only one or two major calving events on the 24 year long photographic record. The movement of icebergs in McMaster Lake indicates that the lake ice cover has cleared at least once in the last 25 years, most likely between 1950 and 1959.

It appears that climatic influences rather than, or in combination with, structural and mechanical forces are responsible for breakup and melt of the ice packs on the McMaster Lakes. In 1973, 2-3 m wide shoreleads were formed around the lake margins, but no cracks developed in the main ice pack. In 1974, shore leads of similar dimensions opened up, but in addition, the ice pack was broken into a series of large cracks, formed when a large block of glacier ice calved into the south end of McMaster Lake. (This event is described in more detail in Chapter 6.4). The pressure wave caused by this event produced a seiche breaking the ice pack on the distal shores of the lake (Figure 4.7), and producing the network of cracks. The three week period following the calving event was characterized by strong (up to 20 m/sec) gusting winds, but the fractured ice pack was not disturbed significantly. In August of 1973 and 1974 jokulhlaupe occurred, resulting in sudden 3-5 m drops in the level of McMaster Lake, which produced rafting of portions of the margin of the ice pack, but again, the main body of the lake ice was not broken up. Apparently, mechanical forces alone

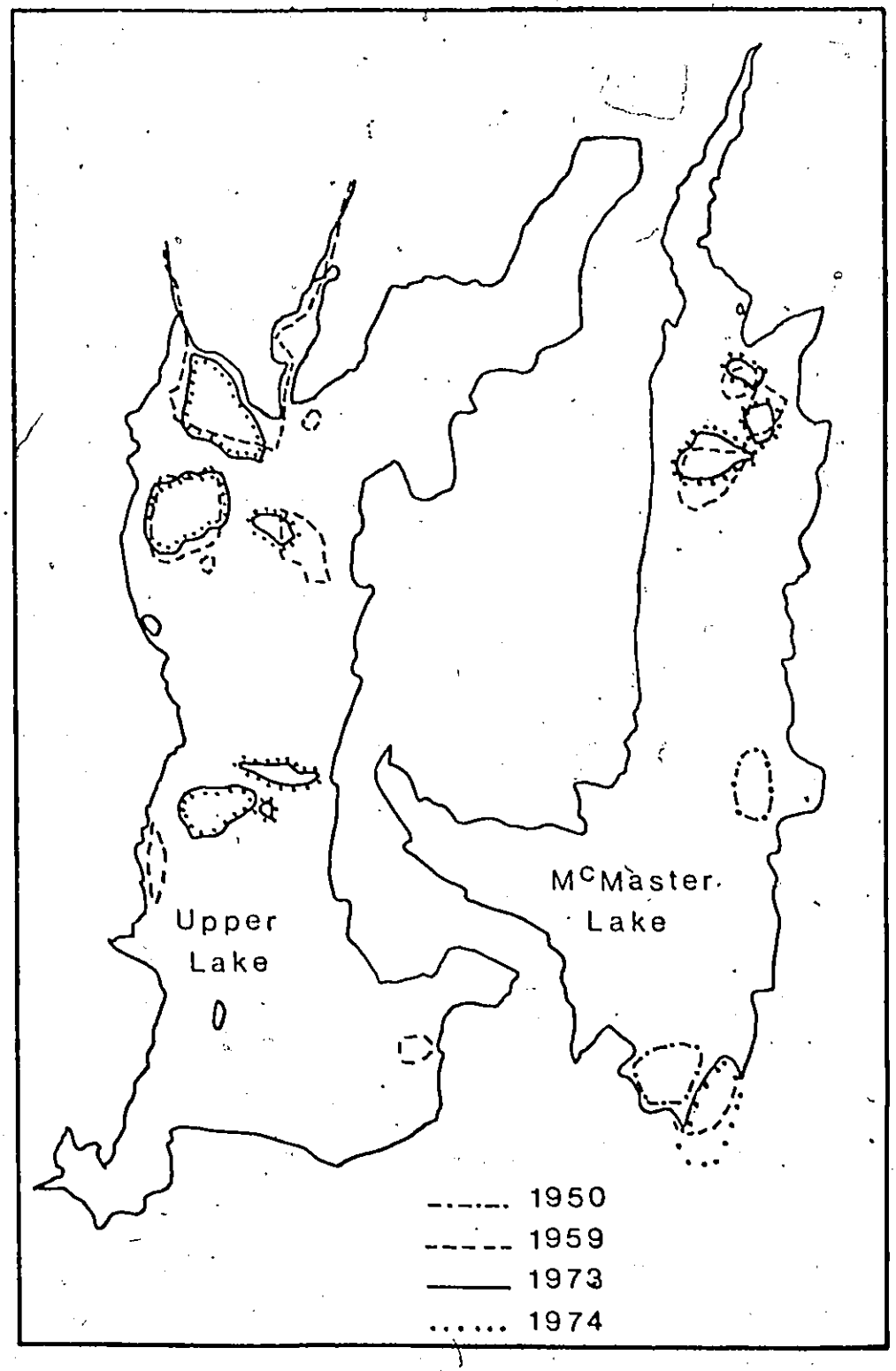


Figure 4.4 Changes in the ice cover of McMaster and Upper Lakes, 1950-1974.

are not sufficient to cause removal of the lake ice pack in the short summer season.

From the climatic records at Eureka alone, there is insufficient evidence to identify any anomalously warm summers, potentially capable of melting lake ice on amictic lakes. From aerial photographs of the McMaster Lakes and other lakes associated with the Ellesmere Island Ice Cap, it appears that complete melt of the ice pack has been an infrequent occurrence.

No sediment analysis of lake or river water was performed at the McMaster Lakes site, but some general comments can be made. Almost all the inputs into McMaster Lake are glacial meltwater streams or direct melt from the glacier front; terrestrial streams are estimated to account for less than 10% of lake inputs. The margin of the ice cap in this region is very clean, with few dirt bands or debris inclusions. As a result, the lake water was observed to be very clear. The water does not appear to have any predominant circulation patterns, with only slight disturbance where clear meltwater streams enter the lake. The depth of the lake and the perennial ice cover precludes any surface disturbances such as wave action. Catastrophic drainage events are often characterized by extremely turbid floodwaters; but this was not the case with the McMaster Lake jokulhlaupe. Water flowing from the lake had apparently very low suspended or solute load concentrations,

one dissolved load sample was taken and a total hardness of 25 ppm was found (Ballantyne, 1975; personal communication). Movement through the bedrock-floored Siphon Creek and gorge entrains little sediment, and suspended sediment values do not increase until the sandur surfaces and floodplain below the gorge are reached.

4.3 Filling and Drainage Behaviour of McMaster Lake.

The behaviour of ice-dammed lakes has been a subject of study for over a century, with a wide variety of theories and ideas as to nature and activities of these phenomena (see Blachut and Ballantyne, 1976 for a discussion of the literature). Some general comments on the major factors involved in determining the behaviour of an ice-dammed lake can be made before moving to the specific discussion of the McMaster Lake example.

The foremost control on the behaviour of ice-dammed lakes is the nature, behaviour, and state of the glacier forming the ice dam. Ice-dammed lake dynamics are closely linked to glacial activity (Liestøl, 1955). Often the conduct of these lakes may not follow any predictable pattern, if the glaciers are actively advancing or retreating, if there are major crevasse fields, or if the lake is situated in a climatically variable location. Most ice-dammed lakes can be demonstrated to be unique in some way in time and space; the end result of an

interrelationship of a complex set of controlling factors. Another control, perhaps almost as significant, is the temperature of the damming glacier. Various authors (Weidick, 1963; Marcus, 1960; Maag, 1969) have suggested that the thermal regime of the ice dam will dictate the mode of drainage of most ice-dammed lakes. In temperate regions, flood waters from ice-dammed lakes generally escape by subglacial or englacial routes. Almost without exception, supraglacial or marginal drainage of ice-dammed lakes in polar or sub-polar ice is the norm. This basic dichotomy will be explored further in Chapter 7.

Changes in the level of McMaster Lake were monitored for a two week period in August 1973, for the entire field season of 1974, and for one day in 1975, by members of the McMaster University field party (Figure 4.5). In 1973, much of the filling behaviour of the lake was not observed and drainage commenced approximately two days after the lakes camp was established. A two week record of continuous drainage reveals a 5 m drop in level from August 1 to August 13; an average rate of 38 cm/day or 1.6 cm/hr. Drainage was initiated in 1973 from a level at least 1 m higher than in 1974, and 12 days earlier. The record high rainfall of July 22, 1973 (Cogley and McCann, 1976) was possibly a factor, but the discussion of burst-triggering mechanisms will be left to Chapter 7.

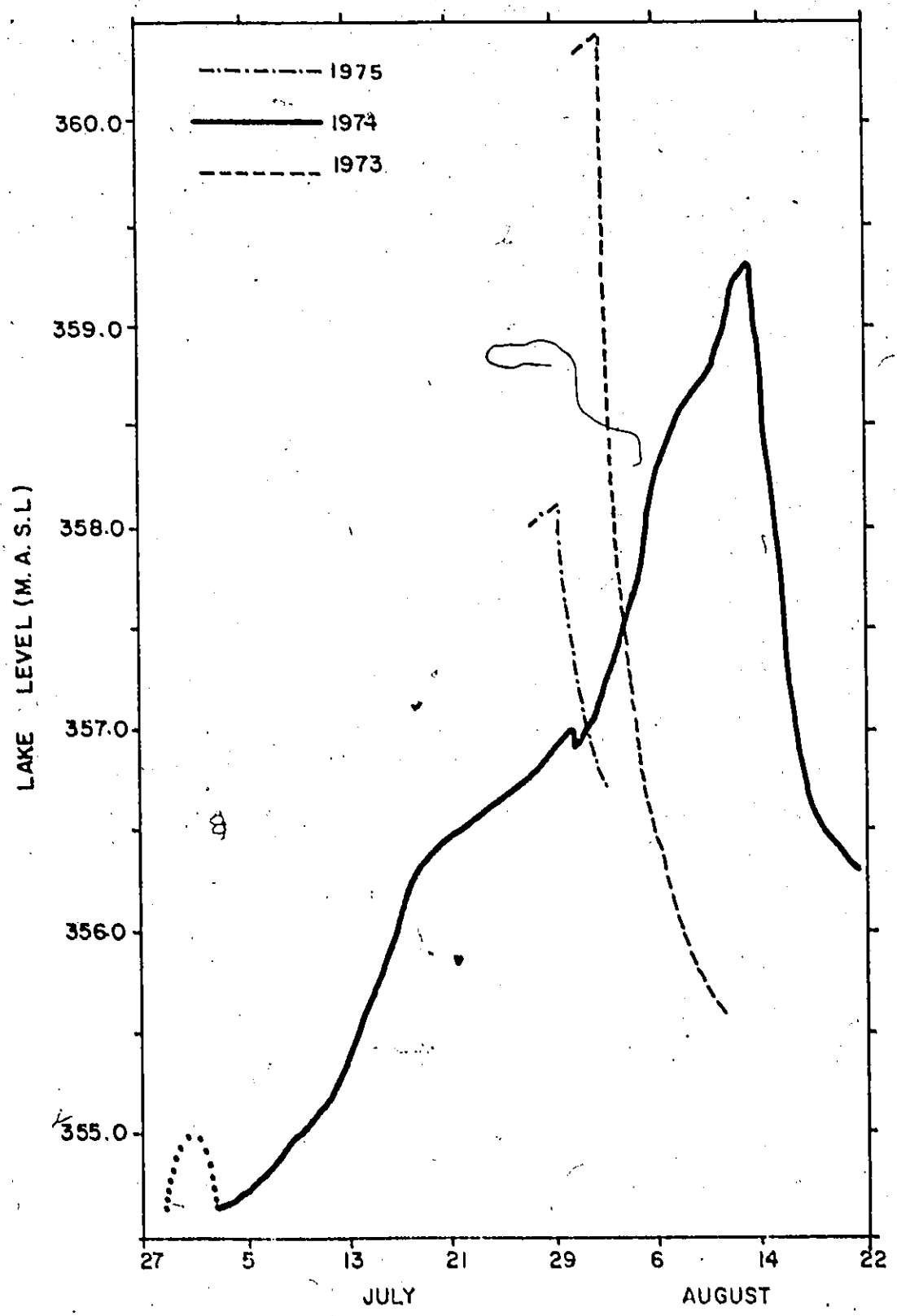


Figure 4.5 Water level fluctuations in McMaster Lake, 1973-1975.

In 1974, lake level fluctuations of McMaster Lake were measured from the start of the melt season on June 27 until August 21. No meltwater was observed to be flowing into the lake basin between the start of the field season (June 20) and June 27. Assuming that the lake drained until the rock base at the head of the outlet stream was reached, and was not interrupted by winter freeze-up, the winter base level of the lake can be considered to be 354.6 m a.s.l. The last reading taken in 1973 was 355.6 m on August 13, which indicates that the level fell another 1 m before winter freeze-up, making the total drop in level in 1973 about 6 m. Given a lake surface area of 7.2 km^2, and assuming vertical shorelines, this represents a total volume discharged during the 1973 jokulhlaup of 42×10^6 m^3, which exceeds the earlier estimates of 13×10^6 m^3 given by Blachut in McCann et al. (1974). This increase is mostly due to the confirmation of the continuation of McMaster Lake under the floating ice shelf, increasing the contributing lake surface area considerably. The assumption that the shorelines around the lake are vertical will cause this value to be somewhat of an overestimate, but the possible error in evaluating the lake surface areas under the floating shelf ice makes calculation of a more detailed volume unrealistic.

The volume of water discharged during the 1974 event was considerably less than in 1973, as can be observed on the Sverdrup River discharge record (Figure 5.4) and on the

graph of water level fluctuations in McMaster Lake (Figure 4.5). The lake commenced drainage at least 1.1 m lower than in 1973, therefore releasing an estimated volume of 34×10^6 m^3. Observations made on July 31, 1975 (S.B. McCann, personal communication) indicate that a maximum lake level of 358.1 m a.s.l. was reached on July 29, 1975; while the level was noted to be 356.9 m on July 31. The 1975 drainage was therefore initiated from a level 1.2 m lower than that observed in 1974; at least 2.3 m lower than in 1973. Again assuming that the lake drained until a winter base level of 354.6 m is reached, an estimated volume of 25×10^6 m^3 was lost from the lake during the 1975 jokulhlaup.

Of potential value in working with ice-dammed lakes, Clague and Mathews (1973) present an equation relating maximum flood discharge and lake storage capacity for 10 major jokulhlaup records. The relationship is defined by:

$$Q_{max} = 75\ V_{max}^{0.67} \qquad (4.1)$$

where Q_{max} = maximum flood discharge in m^3/sec

V_{max} = lake storage capacity in $m^3 \times 10^6$

Applying this relationship to the McMaster Lakes- Sverdrup River data, a Q_{max} of 950 m^3/sec is obtained using the 1973 estimated lake volume of 42×10^6 m^3. For the 1974 situation a V_{max} of 34×10^6 m^3 yields a Q_{max} estimate of 860 m^3/sec. In both examples, there is a wide discrepancy between the measured maximum instantaneous discharge and the values predicted by equation 4.1. Despite allowance for the

factors of distance from the lake reservoir to the recorder, travel time of flood waters, and errors in lake volume and peak discharge values; the McMaster Lake examples fall well outside the 95% confidence intervals constructed for the Clague and Mathews expression. From the record of dropping lake levels in McMaster Lake, the volume changes were computed for 1973, and 1974 (Figure 4.9). These data were in much closer agreement with the Sverdrup River record, once allowance was made for other inputs to the main river (August 2, 1973: volume change 310 m^3/sec in McMaster Lake, 395 m^3/sec at Sverdrup recorder; August 13, 1974: volume change 73 m^3/sec in McMaster Lake, 93 m^3/sec at Sverdrup).

In 1974, a complete record of the filling behaviour of McMaster Lake was obtained (Figure 4.5). No change in lake level was observed pricr to June 27, with the lake ice surface still covered by a substantial snowpack (Figure 4.6). The high temperature and radiation conditions of June 23-27 caused rapid ripening and melt of the lake's snowpack, resulting in standing pools of water on the lake ice surface. THese ponds of water represented an artificial perched water level, until the lake's marginal leads opened, as indicated by the dotted portion of the lake level curve (Figure 4.5) between June 28 and July 2. The average filling rate from the base level on June 28 to the maximum waterlevel prior to the jokulhlaup (August 13) was 10

cm/day, the rate depending on climatic conditions and the behaviour of the influent streams.

Comparison of the lake level plot (Figure 4.5) and the meteorological record, specifically temperature and incoming solar radiation (Figure 5.1), reveals that several periods of good weather with clear skies and high temperatures have a marked effect on lake filling rates. The periods July 10 - 16, and August 1 - 4 are most noticable, where the average filling rate increased to 17 cm/day. The commencement of flow of the Waterfall Stream coincides with these two periods, which is indirectly a function of climatic conditions. A non-climatic occurrence, the calving event of July 31, produced a noticable blip on the lake level chart, the result of the detachment and flotation of a large block (approx. 14×10^6 m^3) of glacier ice. The instantaneous drop of 9 cm in lake level was complicated by the effect of a seiche (Figure 4.8B) which was produced by the calving, but the finer structure of this behavicur was not recorded.

The drainage of McMaster Lake commenced on August 13 in 1974 (Figure 4.8C); the actual mechanism which triggered the jokulhlaup will be discussed in Chapter 7. The lake level fell 2.75 m between August 13 and August 18 at an average rate of 55 cm/day or 2.3 cm/hr. This represents a maximum discharge of 73 m^3/sec on August 13, and an average value of 46 m^3/sec, as obtained from Figure 4.10. The rate

Figure 4.6 Arm of McMaster Lake at campsite prior to spring melt

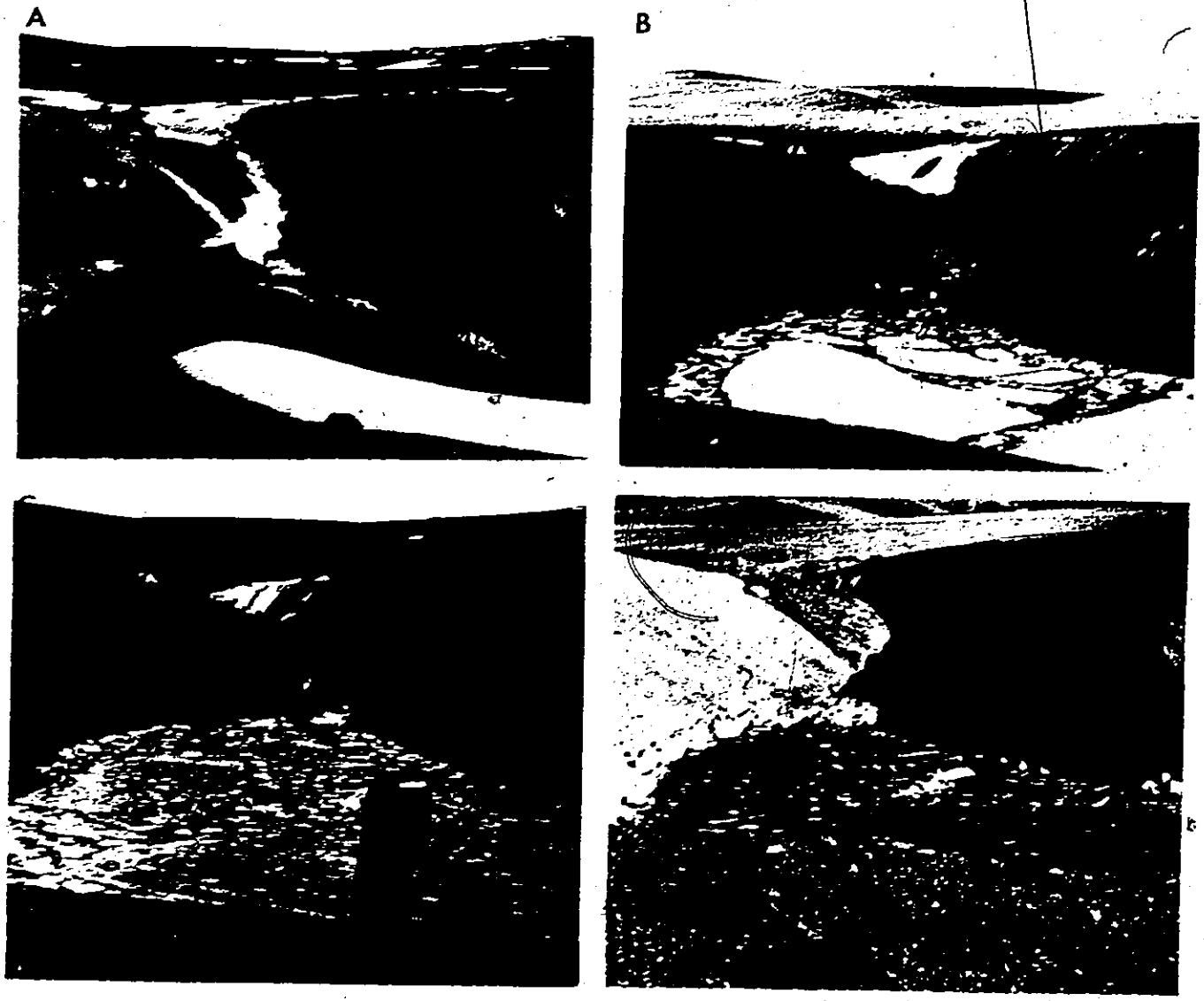
Figure 4.7 Arm of McKaster Lake at campsite looking north.

A) July 12 Water ponded temporarily on the semi-submerged ice pack

B) July 31 Immediately after the seiche. Note the extensive breakup of the margins of the lake pack ice, the rafting of large icebergs at the head of the bay, and the high water mark around the shoreline.

C) August 13 Just after the start of the jokulhlaup, the highest water mark reached in 1974 (359.3 m asl). High water mark in the right middleground is still visible.

D) August 19 Lake level has dropped about 3 m, leaving stranded icebergs. High water mark in right middleground.

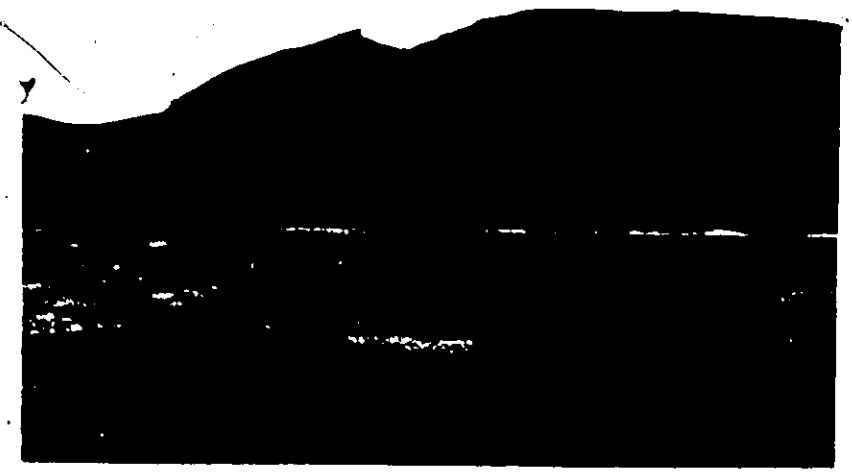


of drainage decreased after August 18 until the end of the measurement period, August 21, to an average rate of 8 cm/day (Figure 4.8D). Assuming that the observed water level in spring prior to the melt season can be considered the lake's base level; in 1974 the lake had a further 1.7 m to drain from the end of the measurement period to winter freeze-up.

Comparison of the drainage events over the three years yields some differences as well as some similarities. The fact that the lake drained in three out of three years of observation will be discussed in Chapter 7. The timing of the start of the jokulhlaup varied over a two week range (1973, August 1; 1974, August 13; 1975, July 29), which is probably controlled by the seasonal climatic trends and the rate of production of meltwater. The record rainfall event of July 22, 1973 (see Cogley; in McCann et al., 1975) probably influenced the rate of lake filling considerably, but there is insufficient climatic data to warrant further speculation. Perhaps the most significant comparison between the three drainage events is the high water mark reached prior to each jokulhlaup. In 1973, the lake level reached was 360.56 m a.s.l., in 1974, 359.30 m, and in 1975, 358.10 m; the difference from year to year being consistently 1.2 m.

Few strandlines were noted in the McMaster Lakes complex, as these features are poorly preserved due to the

A



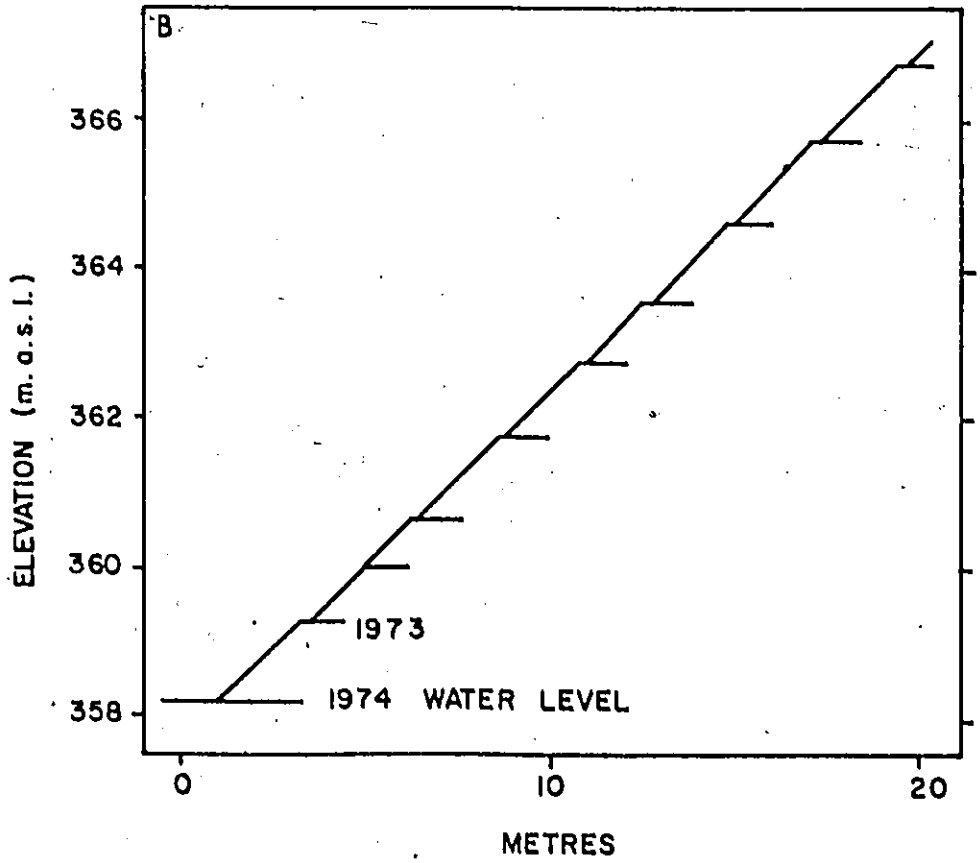


Figure 4.8A Strandlines on McMaster Lake, 4 m. staff rod is on the 1974 high water mark
4.8B Surveyed profile of the strand lines in the above photo

steepness of the shorelines, the coarse regolith, lack of a stabilizing vegetation cover, and mass movement phenomena. No strandlines were observed below the season's high-water line, therefore all lines observed are thought to be former high-water marks. Partial drainage of the lake in 1973, 1974 and 1975 revealed no strandlines below the high-water mark; water movement and the activity of the lake's floating ice pack effectively erasing any water level record. This could indicate that the lake has filled to a progressively lower level after each drainage event for an extended period, including the last three years. At the top of the best preserved sections of strandlines, a noticeable colour and grain size change occurs, possibly marking a former stillstand of water level. One of the best preserved sections of strandlines was observed on the western shore of McMaster Lake (Figure 4.8). With only three years of drainage behaviour recorded for McMaster Lake, little can be postulated regarding the origin and age of such strandline features. However, these findings may reflect the recent history of the behaviour of McMaster Lake as a function of climatic trends or changes in the activity of the ice dam.

4.4 Water Balance of McMaster Lake

During the 1974 field season, an attempt was made to monitor the components of the water balance of the McMaster

Lake system. The system can be expressed in terms of a simple water balance model:

$$V = Qi + P - Qol - Qoj - E \qquad (4.2)$$

where V = volume change of the lake water

Qi = inflow to the lake

P = precipitation on the lake surface

Qcl= outflow by leakage through the ice dam

Qoj= outflow by jokulhlaup

E = evaporation.

Precipitation at the lake site was very low; the cumulative total for the period June 20 to August 20 was 2.8 cm. This represents a negligible input to the total water balance of the lake (average 0.004 m^3/sec) and is not considered further. Similarily, evaporation was probably insignificant as the water temperature was close to 0°C, most of the lake surface was covered with ice, and the average air temperature was generally below 5°C. The term accounting for loss from the system by slow leakage (Qol) is dealt with in more detail in Chapters 5 and 7, but it can be shown to be negligible. The remaining terms of inflow, volume change, and jokuhlaup outflow are plotted on Figure 4.9.

The inputs into McMaster Lake include a small terrestrial runoff component (<5%), consisting of snowmelt runoff and ground-ice melt, in three small streams. Supraglacial runoff and direct melt of the lake ice pack, glacier ice and late lying snow and firn banks accounts for

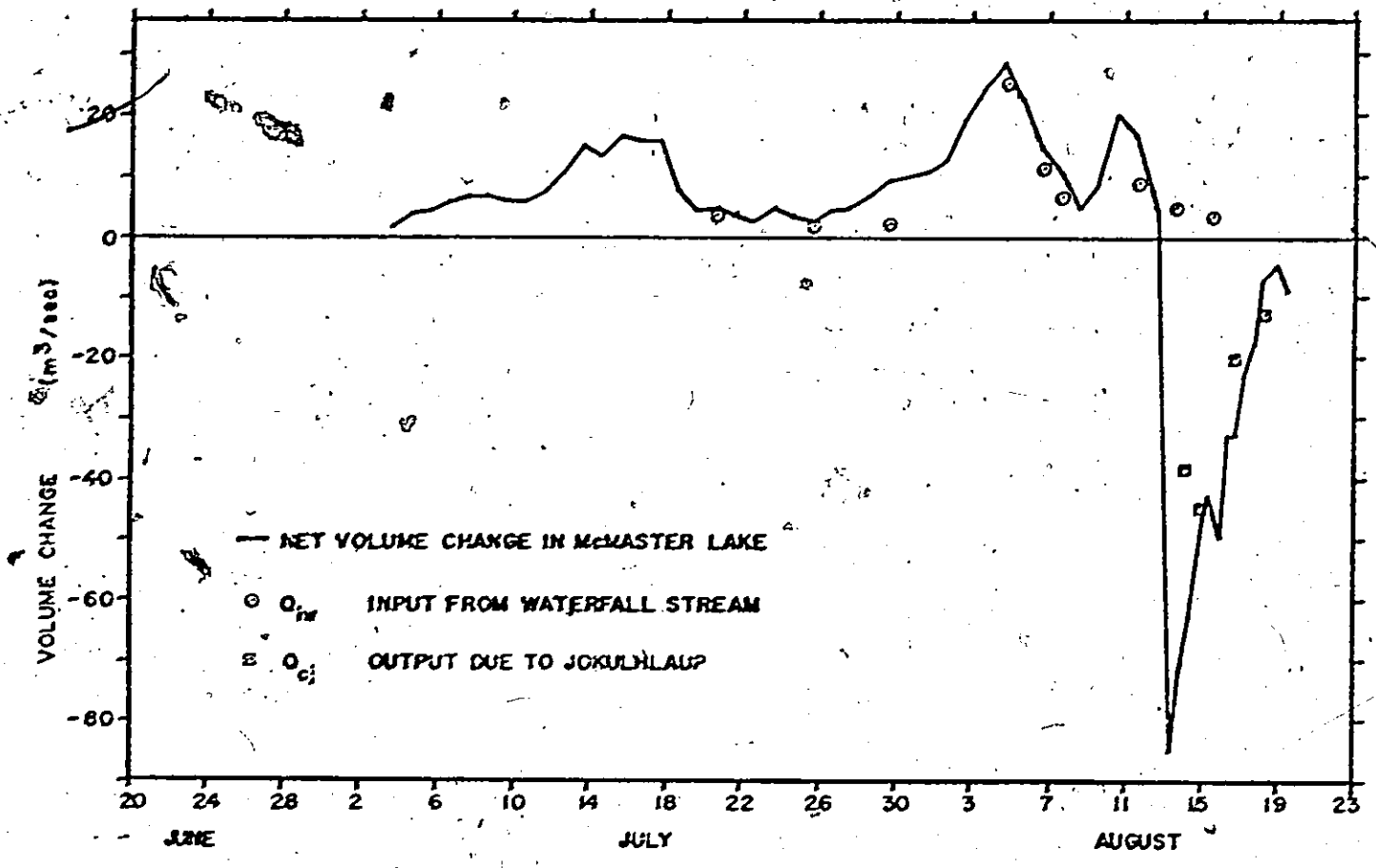


Figure 4.9
Plot of net lake volume changes in McMaster Lake, 1974.

a further 15% of the inputs to McMaster Lake. The main contribution (about 80%) to the lake is from Upper Lake and the Connecting Ponds, which drain through the Waterfall Stream into McMaster Lake. The terrestrial and direct melt components were not continuously monitored due to their complex nature and the lack of manpower and equipment to support such observations. However, spot measurements of these inputs can be shown to respond very closely to climatic parameters, and therefore periods of high radiation and temperature will increase flow from these inputs. This relationship is demonstrated in further detail in Chapter 5 for glacier melt, while the increase in terrestrial flow during good weather was observed in the camp stream. This response is thought to be an indication of an increased melt of the permafrost table, and subsequent runoff.

The proportion of the total input into McMaster Lake contributed by the Waterfall Stream (Q_{iw}) is indicated in Figure 4.9. The spot points of discharge (•) gauged in the Waterfall Stream, superimposed on the plot of volume change indicate that for much of the season prior to the jokulhlaup, the Waterfall Stream is the main input into McMaster Lake. The Waterfall Stream appears to be draining the Connecting Ponds, but a very slow drop in water level and the small size of these ponds indicates another source of water must exist. The water levels in the Connecting Ponds and Upper Lake were surveyed to be identical and were

observed to be dropping at similar rates throughout the season. As no other outlet to Upper Lake was discovered, it would appear that the Upper Lake basin continues under a floating glacier ice shelf (see Chapter 6.4), and is draining continuously and non-catastrophically into McMaster Lake. A weak relationship between meteorological parameters and discharge in the Waterfall Stream was observed on several occasions, indicating a regular outflow from Upper Lake with an additional climatically controlled component derived from supraglacial melt and runoff into the Connecting Ponds. A major control on the discharge of the Waterfall Stream, and therefore a major determinant of the filling behaviour of McMaster Lake is channel availability. The Waterfall Stream has two main channels, and the distribution of snow and ice cover in these channels controls the drainage rates of the Connecting Ponds, and hence, Upper Lake also. The North Channel of the Waterfall Stream commenced flow between 1600 hrs, July 13, and 1400 hrs, July 14 in 1974. Flow continued at a slowly decreasing rate (average 2 m^3/sec) in the North Channel until August 4 or 5, when flow in the South Channel commenced. This channel was not opened until the water in the immediately adjacent Connecting POnd was able to overtop the substantial snowbanks remaining in the South Channel. Flow in the North Channel was greatly reduced, but the total discharge in the

Waterfall Stream increased to 11 m^3/sec, dropping afterwards to 6.6 m^3/sec on August 7, and to 3.1 m^3/sec by August 15.

The change in volume of water (ΔV/day) in McMaster Lake was calculated from lake level data, and is derived:

$$\Delta V/\text{day} = \frac{\Delta \text{lake level/day} \times \text{lake surface area}}{86{,}400 \text{ sec/day}} \qquad (4.3)$$

Net volume change is plotted daily for the period July 2 - August 13, and diurnally after the commencement of the jokulhlaup (Figure 4.9). The first part of the melt season is characterized by the initial snowmelt runoff, which is probably the sole input prior to July 13. The period of July 10 - 16 was marked by high radiation and temperature readings, with temperatures reaching 14.4°C on July 13. On July 13 the Waterfall Stream commenced flow, and from this date it is the main input to McMaster Lake. Good weather conditions of the period August 1 - 4 caused a sharp increase in the input volumes, and probably also triggered the opening of the South Channel of the Waterfall Stream, which caused the lake levels in the Connecting Ponds and Upper Lake to lower more rapidly. On August 13, for the first time in the season, the lake volume change was negative, due to the breaching of the ice-dam and subsequent jokulhlaup. Discharge measurements in the Waterfall Stream after lake levels start to drop indicate that the stream was still discharging an average 5 m^3/sec into McMaster Lake. Addition of this volume to the outflow by jokulhlaup (Qoj =

on Figure 4.9) gauged in Siphon Creek would increase the total volume lost, and bring this volume in closer agreement with the net volume changes.

4.5 Summary

Over three field seasons a substantial amount of information concerning the limnology of the McMaster Lake area was collected. The principal findings are:

1. McMaster and Upper Lakes are structurally controlled basins of the fault trough type. McMaster Lake is dammed by two coalescing lobes of glacier ice, with a large section (about 3.5 km²) of partially floating shelf ice occupying the southern end of the 7.2 km² lake. Maximum depths are 200 m, adjacent to the shelf ice.

2. McMaster Lake appears to be an amictic lake; with water temperatures near isothermal at 0.0-0.5°C, no thermal stratification, and a 3 m thick ice pack covering the lake year-round.

3. Lake levels were observed to drop 5 m between August 1 and August 13, 1973; and 1.2 m between July 29 and July 31, 1975. A complete record of filling and drainage behaviour was obtained in 1974. The melt season commenced on June 27, and lake level rose 4.6 m at an average rate of 10 cm/day between June 28 and August 13. Between August 13 - 18, the lake level fell 3m, at an average 55cm/day, slowing to 8 cm/day after August 18.

4. The water balance of McMaster Lake was computed and it was found that precipitation and evaporation were negligible terms and that no water escaped from the lake prior to the jokulhlaup. McMaster Lake is fed by a terrestrial component (5%), direct snow and glacier melt (15%), and waters from Upper Lake draining via the Waterfall Stream (80%).

CHAPTER 5

HYDROMETEOROLOGY OF A GLACIAL DRAINAGE SYSTEM

The hydrology of a high Arctic glacial system has been studied by very few workers, as noted in Chapter 2.1. The hydrometeorological aspect is covered in more detail, most notably by the longterm studies on Devon Island, Axel Heiberg Island, and Barnes Ice Cap, Baffin Island.

The actual hydrology of the glacier surface, and the interior where applicable, has been a topic of great interest in temperate glaciers, most notably in light of the role of water in recent glacial dynamics theory (Weertman, 1962; Lliboutry, 1971). With a few important exceptions (Adams, 1966; Maag, 1969) this phase of the glacial hydrological system is sadly neglected in polar and sub-polar glaciers. The first three sections of this chapter will deal separately with the meteorology and hydrology of the McMaster Lakes study area, as well as some basic statistical analysis of the hydrometeorological data. The last two sections of the chapter deal more specifically with the glacial hydrologic system, and the hydrologic phenomena peculiar to the drainage of ice-dammed lakes.

5.1 General Climate and Meteorology at McMaster Lakes.

The macroclimatology of the Canadian Arctic Archipelago is described in detail by Hare (1960), Meteorological Branch, Department of Transport (1970), Vowinckel and Orvig (1970), and Hare and Hay (1974). The following brief summary of conditions over south-central Ellesmere Island is extracted from these sources.

From November to May, the high Arctic is affected by a stable anticyclonic system located over the western Arctic, and the general flow of air is from the northwest towards the Iceland-Baffin lows. Temperatures are very low, with little cloud cover (37%) and negligible amounts of precipitation. The brief summer season is characterized by fast moving cyclones of the Arctic frontal belt, bringing the greatest precipitation amounts of the year. These weather systems generally track far to the south of Ellesmere Island, resulting in prevailing westerly winds in the study area. The mean air temperatures for central Ellesmere Island are: -34°C for January; -25°C for April; 6°C for July; and -17°C for October. The mean date of the fall of mean daily temperature to 0°C is September 1, while the mean date of the rise to 0°C is June 15. The mean precipitable water (for a vertical column of air) is 0.2 cm for January, and 1.3 cm for July, indicating a very low supply of moisture to Arctic regions. The mean annual precipitation total for Eureka is 150 mm, but recently these figures have been questioned (Cogley, 1971; Hare and Hay,

1974). The measurement techniques employed, and the practice of disregarding trace precipitation events can lead to significant underestimates of true precipitation values.

The terms of the radiation balance equation:

$$R = RAD\ (1-\alpha) + L\downarrow - L\uparrow \qquad (5.1)$$

where RAD = incoming solar radiation

α = albedo

$L\downarrow$ = incoming longwave radiation

$L\uparrow$ = outgoing longwave radiation

are estimated by Hare and Hay (1974). For south-central Ellesmere Island these values are given in Table 5.1, and from these data it is apparent that the strongly negative winter radiation budget is balanced by the positive summer values, resulting in a mean annual net radiation of about 2000 ly/yr. During the daylight summer months, the solar radiation component is the most important factor in determining values of R due to high transparency and low cloud density in the high Arctic (Gavrilova, 1966), and the longwave term is weakly negative. The reverse is true during the 6 month winter period.

A meteorological station was manned for July and August, 1973-1975 at the base camp site, Vendom Fiord; while a set of observations was made at the McMaster Lakes site during July and August, 1974. The temperature and incoming radiation regimes for McMaster Lakes are presented in Figure 5.1, while cloud cover, relative humidity, and precipitation

TABLE 5.1 RADIATION BALANCE DATA FOR SOUTH-CENTRAL ELLESMERE ISLAND (from Hare and Hay, 1974).

	JAN.	APRIL	JULY	OCT.	ANNUAL
Mean Incoming Solar Radiation (ly/day)	0	340	475	25	
Generalized Surface Albedo (%)	>80	>70	25	60	
Annual Mean Solar Radiation (ly)					75,000
Annual Mean Absorbed Solar Radiation (ly)					40,000
Mean Incoming Long Wave Radiation (ly/day)	325		525		
Mean Outgoing Long Wave Radiation (ly/day)	400		710		
Net Radiation (ly/day)	<-75	-35	225	-65	
Mean Annual Net Radiation (ly/day)					2,000

are plotted in Figure 5.2. Wind speeds and wind directions are given in Figure 5.3. The meteorological records from the base camp, as referred to in following sections, are given in McCann et al. (1975), and Woo (1975).

The temperature pattern in 1974 shows the arrival of above freezing air temperatures to be 10 days later than average. Freezing or near freezing temperatures occurred again several times during the summer season, at a somewhat regular interval of 6 or 7 days. This pattern is also evident on the plots of solar radiation, cloud cover, and relative humidity; and is thought to be a result of the passage of low pressure cells in the Arctic frontal zone to the south of the study area. Church (1972) drew residual correlograms of the runoff for several Baffin Island rivers, and attributed the weak oscillatory effects of 6-day or shorter periods to synoptic weather patterns. The seasonal maximum of 16.0°C was recorded on August 19, and the daily range of temperature varied from 14°C to less than 1°C. Comparison of the McMaster Lakes data with the data collected at the base camp site indicates that the mean temperature is 2°C lower for the Lakes site. The difference in altitude of 375 m and the proximity to the ice cap are responsible for this difference.

Incoming solar radiation was the only component of the radiation balance monitored at the McMaster Lakes site, and the mean incoming solar radiation over the measurement

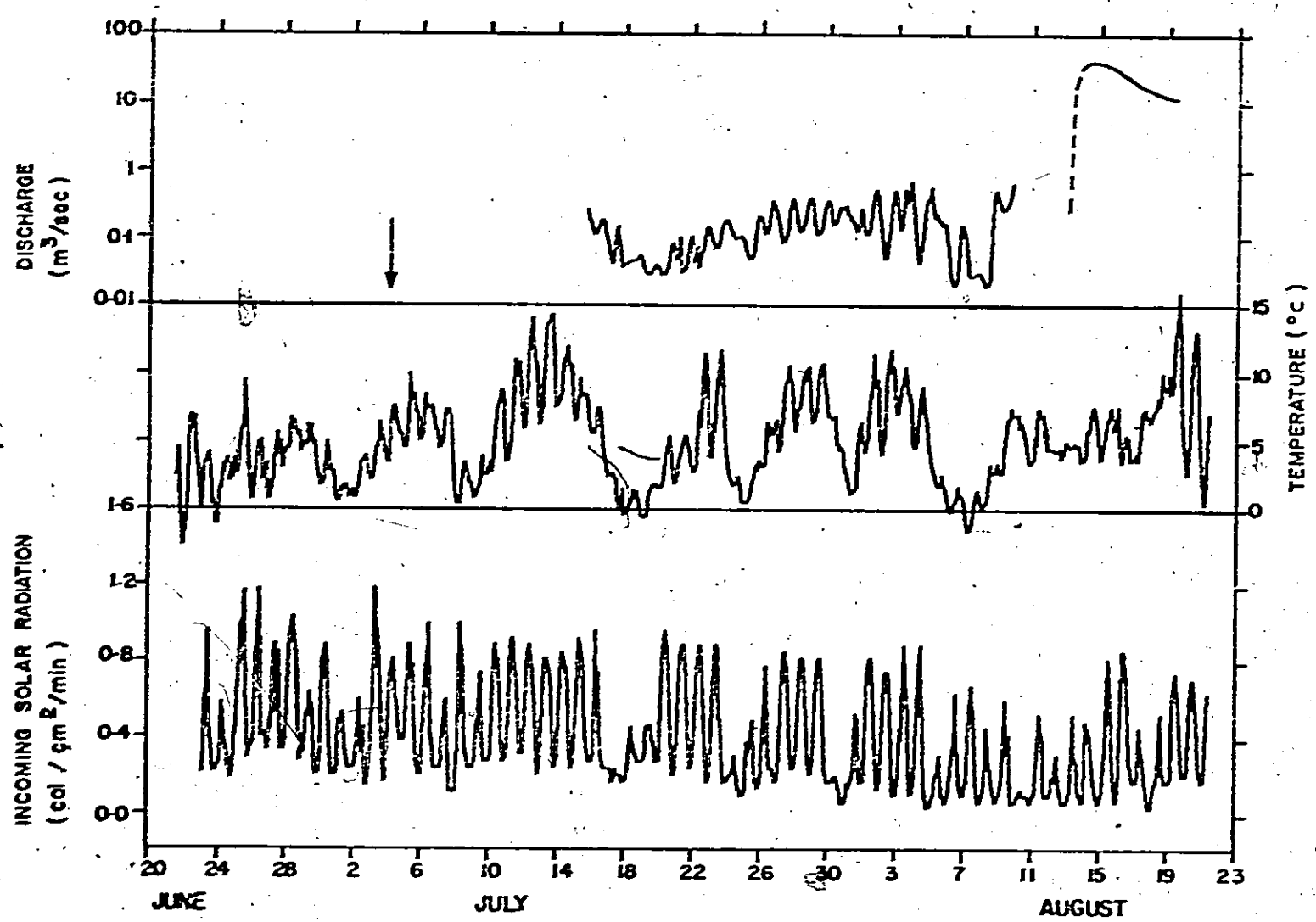


Figure 5.1 Temperature and incoming solar radiation as recorded at the McMaster Lakes camp, 1974
Hydrograph of Siphon Creek, (↓ = start of flow)

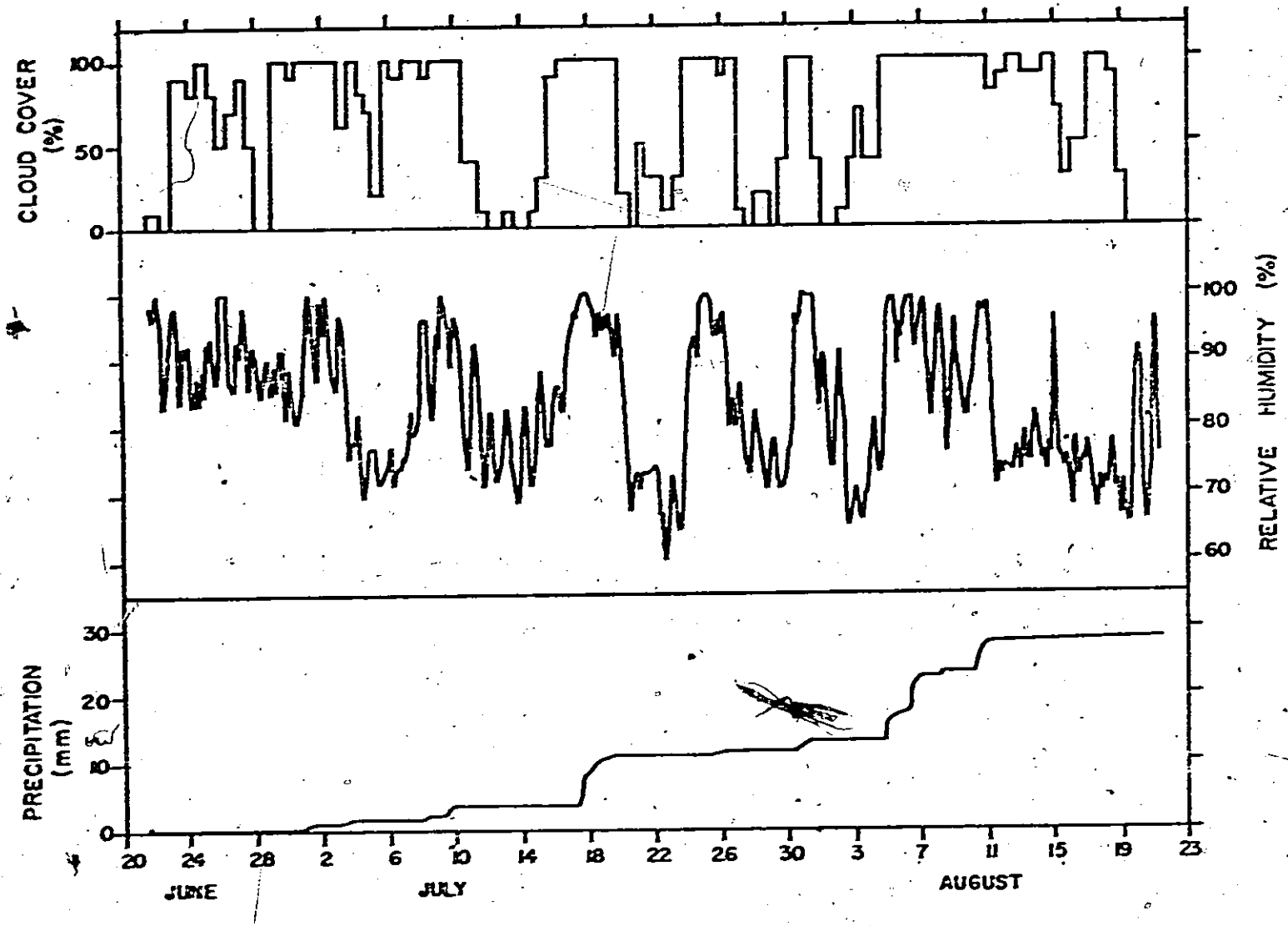


Figure 5.2 Cloud cover, relative humidity, and precipitation at the McMaster Lakes camp, 1974.

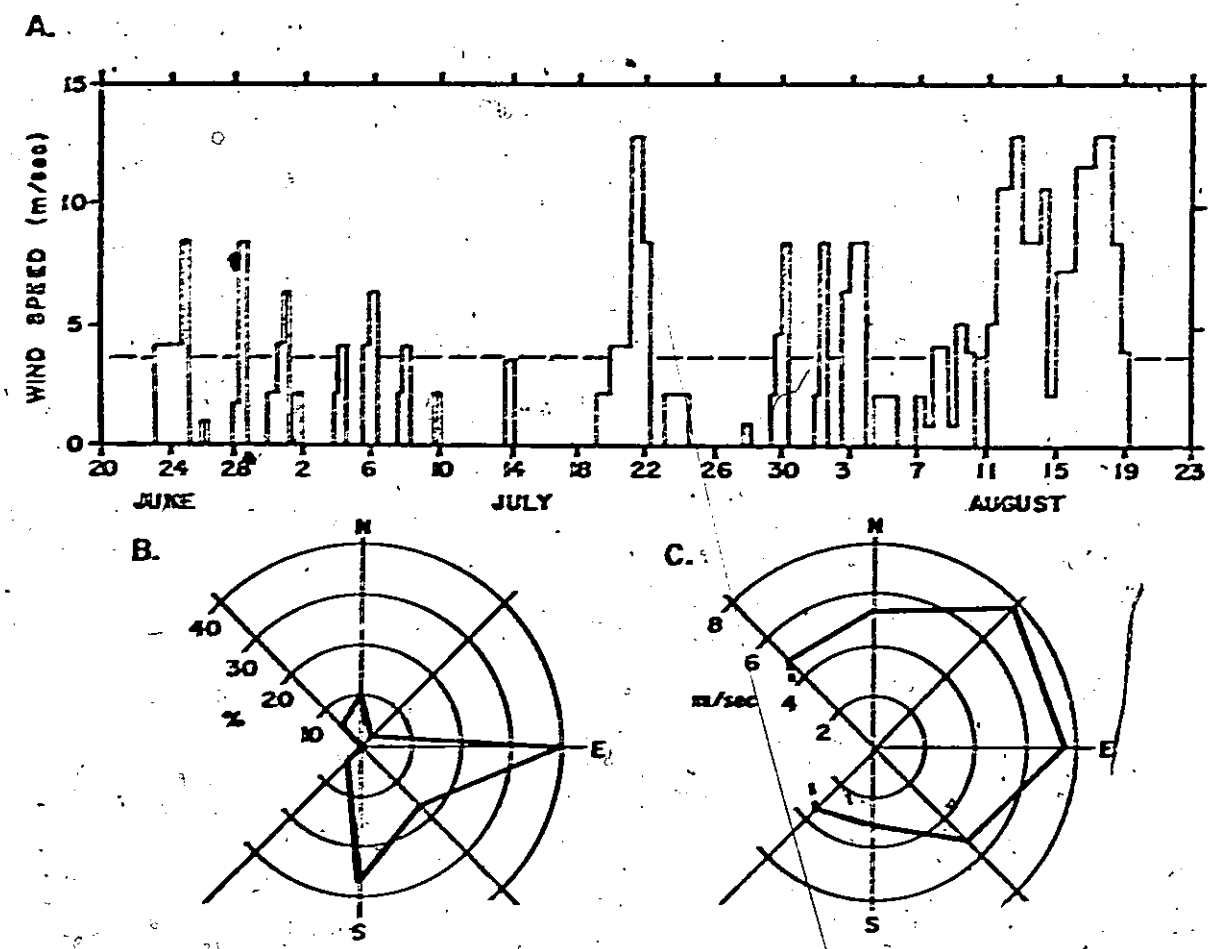


Figure 5.3A Estimated wind speeds at McMaster Lakes camp, 1974
5.3B Wind direction frequency rose of winds greater than 3 m/sec.
5.3C Wind direction rose of average wind speeds

period was 0.38 cal/cm²/min. The maximum value recorded for the season was 1.48 cal/cm²/min on June 27. Incoming solar radiation and net radiation recorded at the base camp, and incoming solar radiation monitored at Mcmaster Lakes showed a slight decline from the maximum in late June. This overall trend is due to the gradual increase of the zenith angle of the sun, from the peak of summer solstice on June 21. Radiation values also correspond in part to the amount of cloud cover, and cloud cover data recorded at McMaster Lakes and Base Camp were very similar (Woo, 1975). On this basis, the seasonal decline in the radiation values in the last two weeks of the study period is accentuated by the almost complete cloud cover (90-100%) recorded for that period.

A total of 28 mm (water equivalent) of precipitation fell at the McMaster Lakes camp (375 m asl) over the two month measurement period. From meteorological data for Eureka (Meteorological Branch, Department of Transport, 1970), and from Hare and Hay (1974), July, August, and September are the stormiest months of the year. Almost half of the precipitation received at McMaster Lakes over the two month period fell in early August. A pronounced orographic effect was noted (Woo, 1975) between the nearby base camp (20 km distant, 10 m asl) and the McMaster Lakes camp; and again over an even shorter horizontal distance, between the McMaster Lakes camp and the rain gauge maintained on the

glacier surface at 500 m asl. With altitude, changes in the amount of precipitation, the number of precipitation events, and the form of precipitation were noted. The precipitation event of July 17-18 demonstrates this effect; where 2.3 mm of rain fell on the base camp site, 7.4 mm (water equivalent) of mixed rain and wet snow was recorded at the McMaster Lakes camp, while 12.7 mm (water equivalent) of snow fell on the glacier surface at 500 m asl. From the data collected at the base camp, 1974 was a drier than normal summer (total rainfall for June 19 to August 21 was 12 mm) when compared to the 30 mm recorded in 1975 (June 23 to August 20).

A total of 111 visual observations of wind speed (using the Beaufort wind scale) and wind direction were made June 21 to August 21, 1974 with an average wind speed of 3.3 m/sec over the summer period (Figure 5.3A). 43% of the observations were recorded as calm, and the remaining readings ranged from 1 to 14 m/sec. A wind rose (Figure 5.3B) of the frequency of wind directions greater than 3 m/sec indicates that 80% of the winds came from the south-east quadrant, with 38% of this total directly from the east. The strongest winds (~7.5 m/sec average) also derive from the eastern quadrant (Figure 5.3c). This effect is apparently due to the site's proximity to the ice cap, in contrast, with the predominantly southerly winds funnelling

up Vendom Fiord which were recorded at the base camp (Woo, 1975).

Between August 8 and 18, 1974 several low-pressure cells were located to the west of Axel Heiberg Island and Ellesmere Island, and high pressure persisted over the Greenland Ice Cap for an unusually long period (Can. Met. Service, synoptic charts). Moderate and variable winds were reported for Cape Herschel on the east coast of Ellesmere Island, while the Vendom Fiord camp reported variable and stronger than normal winds. At McMaster Lakes, broken to overcast sky conditions persisted for the ten day period; while winds were consistently from the south and east, at average wind speeds of 10-13 m/sec, with frequent gusts of 20-25 m/sec. Clouds of blowing snow often obscured the eastern horizon over the ice cap, and lenticular cloud formations were recurrent. The strong winds resulted primarily from the gradient between the stationary low pressure system to the west, and the high centered over the Greenland Ice Cap, but the influence of cold air drainage or katabatic winds (Geiger, 1950) compounded the effect. Temperature inversions are commonly set up over the ice mass, resulting in a katabatic wind inland, and gusty-katabatic winds near the ice sheet margins (Whillans, 1975). In addition to the abnormal winds, the weather conditions of this period were also anomalous, in that the 6-7 day cyclic pattern of variation broke down.

5.2 Hydrology of the Siphon Creek-Sverdrup River System.

Discharge in Siphon Creek was monitored for most of the runoff period in 1974, while Sverdrup River discharge was recorded at the base camp site for the summer seasons of 1973, 1974, and 1975. The hydrographs for Siphon Creek (Figure 5.1) and the Sverdrup River (Figure 5.4) are characterized by a diurnally fluctuating pattern, and a predominantly high Arctic glacial regime (Church, 1974). Beyond this, few generalizations can be made, and the two hydrographs are best discussed separately.

Siphon Creek is a small proglacial stream which drains a 1.0 km² portion of the margin of the Ellesmere Ice Cap (see location map, Figure 3.3). However, the stream underfits the deeply-incised McMaster Gorge which it flows through, and as will be demonstrated, this channel is occasionally the route followed by floodwaters escaping from ice-dammed McMaster Lake. In such instances, Siphon Creek can be considered the outlet of the much larger MacMaster Lakes basin (Figure 3.2). This approximately 150 km² basin consists of McMaster and Upper Lakes, several other smaller ice-dammed lakes, a small (5%) terrestrial component, and a 15 km section of the margin of the Ellesmere Ice Cap. The concept of a two component hydrologic regime will be discussed further in Chapters 5.3 and 5.4.

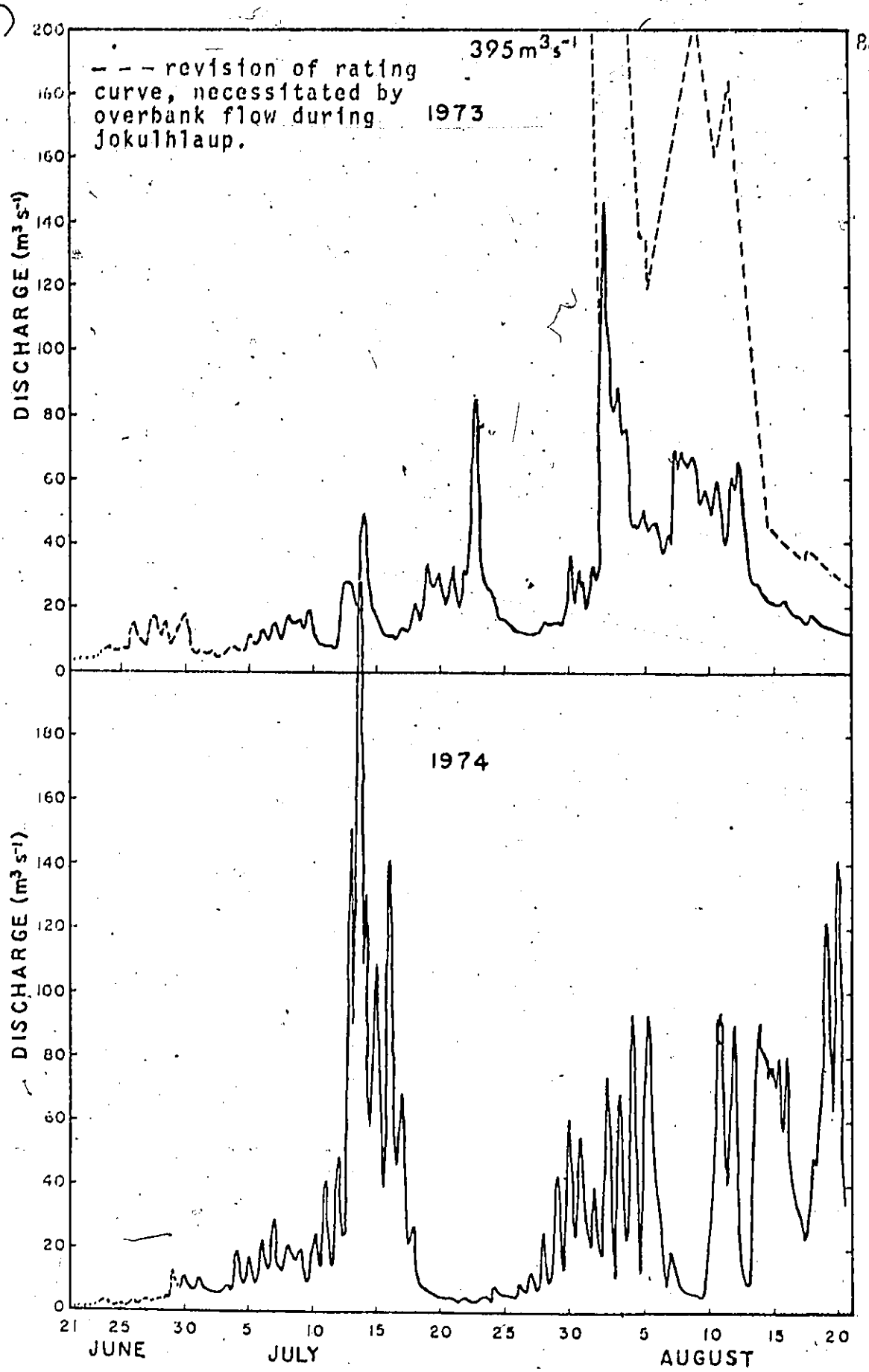


Figure 5.4 Hydrograph of the Sverdrup River

Siphon Creek commenced flow on July 4, 1974 with a small amount of water and slush cutting through a snowdrift filled channel (Figure 5.5A). The stream did not have the capacity to flush the snowbanks from the system, and consequently was continually adjusting its channel bed in the shallow upper reaches of the stream for several weeks. The water cut its way through 2-3 m deep snowdrifts to a bedrock channel by July 15, and a stage recorder was installed. Between July 4 and 15, spot readings of discharge did not exceed 0.3 m^3/sec, so no discrete snowmelt flood was experienced.

Between July 15 and August 13, Siphon Creek exhibited a simple diurnal response to variations in meteorological parameters. The strength of this relationship reflects the high degree of glacierization of the drainage basin, and the proximity of the margin of the ice cap. The maximum discharge for this period (0.68 m^3/sec) was recorded on August 3; while the minimum flow was recorded on August 6 (0.018 m^3/sec). The average time of peak discharge was 1800 hours, and on the average, low flow occurred at 0600 hours. Snowbanks remained along the upper reaches of the stream, and a glacier-margin, lee-slope snowdrift was not completely depleted during this period. At a distance of 1 km from the ice margin, the stream drops over a series of waterfalls to flow in the bottom of the 200 m deep gorge. This middle section of the stream is

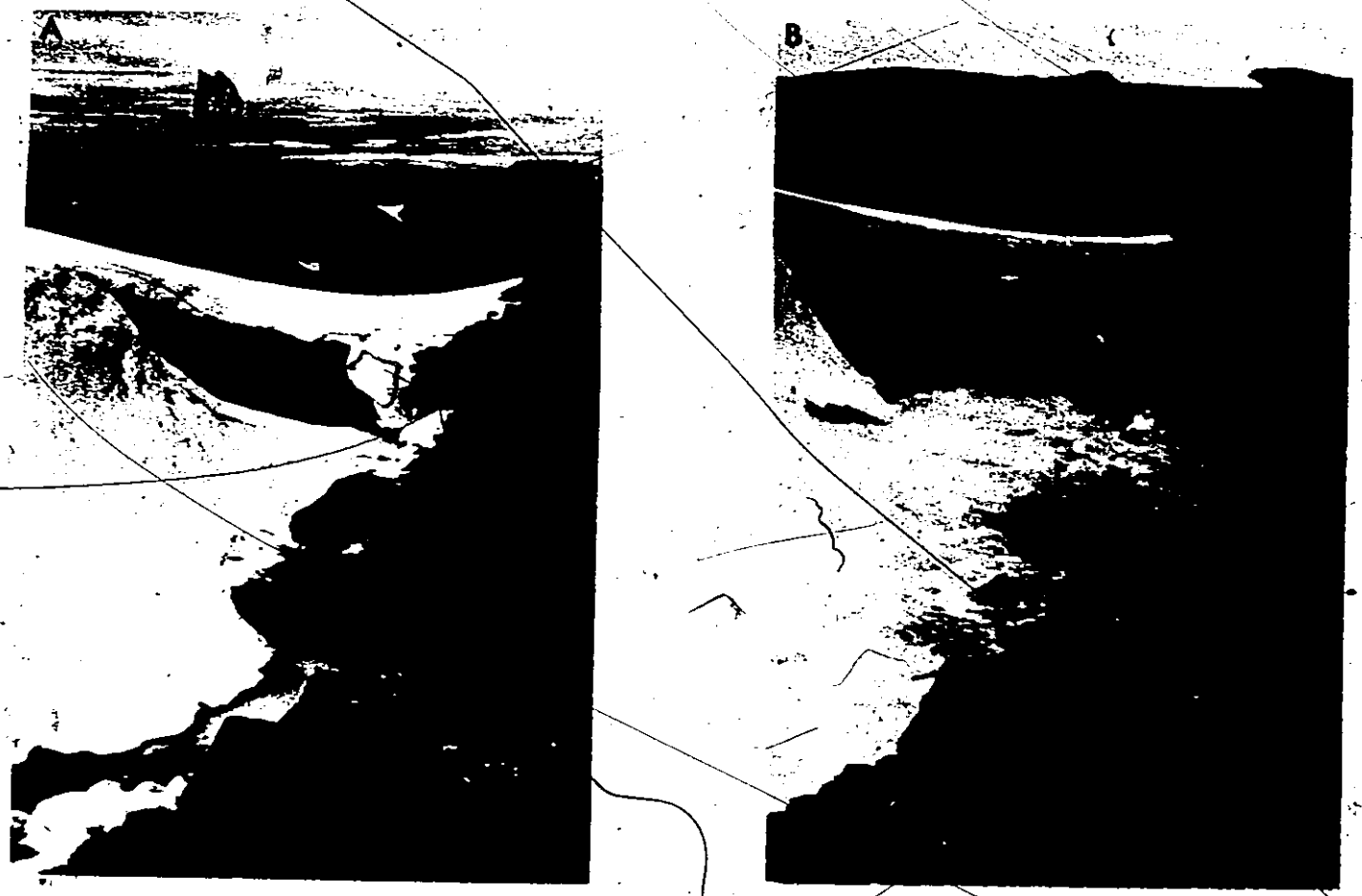


Figure 5.5 Discharge in Siphon Creek 5.5A soon after start of flow (0.2 m^3/s)
5.5B under jokulhlaup conditions (39 m^3/s)

inaccessible due to convexity of the rock walls, but the channel was observed to be free of snow and coarse sediment by mid-July. The lower section of Siphon Creek emerges from the gorge to join meltwater runoff from the McMaster Glacier ice and flow over the large alluvial fan before joining the Sverdrup River system as a tributary.

On August 13, water from McMaster Lake forced an outlet through the glacier ice near the head of Siphon Creek, and the discharge in the creek suddenly increased by two orders of magnitude (Figure 5.5b). Most of the late lying snowbanks were flushed from the system by the initial burst of water, and the automatic stage recording device was destroyed in the flood. Spot dye gauging tests were used to determine jökulhlaup discharge values (Figures 3.9 and 5.1) for the remainder of the field season. The maximum recorded discharge was 45 m^3/sec at 1400 hours on August 14, but the incomplete discharge record is inconclusive. A general trend of gradually decreasing discharge over the two weeks following the jokulhlaup was observed and is substantiated by the Sverdrup River discharge record. The behaviour of this peculiar hydrologic phenomenon is discussed further in Chapter 5.5.

The Sverdrup River is a medium-sized high Arctic basin draining approximately 1630 km^2 of tundra and glacier ice; with the ice cap comprising roughly 75% of the total basin area. The basin areas quoted above are not precise

due to the lack of accurate surveying data, and variability in defining the basin boundaries. The Sverdrup River basin outline in the Ellesmere Ice Cap was delimited on aerial photographs, using the procedures of the Glacier Survey of Canada (Ommanney, Clarkson, and Strome, 1973). This line, however, is the glaciological divide based on ice flow, which does not always coincide with the hydrologic divide based on surface hydrology. For hydrologic purposes, in this instance, the value given for the Sverdrup River catchment area is somewhat overestimated. This uncertainty is further compounded by the variable glacier area contributing to runoff; a function of snowline elevation.

The Sverdrup River flows for about 50 km through a broad, North-South oriented, river valley with numerous tributaries from both glacial and non-glacial headwater sources. The upper half of the valley appears to be structurally controlled and actively eroded by the stream. The lower portion of the valley consists of a broad, flat floodplain with numerous braided channel sections and a small prograding delta being built into the head of Vendom Fiord.

The snowmelt flood on the Sverdrup River occurs in the latter part of June or early July, but the dissimilarity of the 1973 and 1974 hydrographs prevent any generalizations from being made. The onset of warm weather was apparently earlier than normal in 1973; little snow cover remained at

the start of the field season, all major streams were flowing by June 20, and the timing and nature of the snowmelt flood is unknown. In direct contrast, a strong, well defined peak is seen on the 1974 hydrograph, typical in shape and timing if a high Arctic spring snowmelt flood (Church, 1974).

This snowmelt event represented the peak discharge of 1974; with several other minor peaks later in the season due to rainfall response, and glacial-hydrological events. In 1973 no major snowmelt flood was observed, several rainfall response and glacial meltwater floods occurred, and the major event of the season was the jokulhlaup of August 1. The Sverdrup River data for 1975 is not yet available, but from the stage record (S.B. McCann, personal communication) the regime appears to have been roughly similar to the 1973 record, with the exception of a dominant rainfall (or possibly warm weather) response event at the end of the field season. The three year record of this high Arctic glacial stream appears to support the general belief that short term hydrologic records cannot give an unbiased picture of long term climate and hydrology.

5.3 Statistical Analysis of Hydrometeorological Data.

The statistical manipulation of hydrological data for analytical purposes can be subdivided into two main categories: frequency analyses and regression analyses.

Frequency analyses, or the expression of hydrologic data on a probabilistic basis, has been successfully applied to snowmelt and glacial hydrology by Cogley (1971), Gudmundsson and Sigbjarnarson (1972), and Church (1972). The marked diurnal periodicity of a melt hydrograph lends itself to time series analysis, and the covariance and spectral density functions of input (meteorological) and output (hydrological) time series can partially describe the input-output relationship. Cogley (1971) has demonstrated that frequency analysis is a powerful tool in the determination of drainage basin lag times, a parameter which is of interest in Chapter 5.4. However, the data set available for the Siphon Creek system is not suitable for accurate frequency analysis.

Regression analysis is a commonly used technique in glacial hydrology, and short term, predictive deterministic models have been designed, based on simple regression and correlation models. However, it must be kept in mind that there are several inherent problems with the use of regression analysis for hydrometeorologic data.

The basic problems of non-normality, non-randomness and interdependence of the independent variables are common to all statistical analysis. Goodison (1972) examined the possibilities of transformation of hydrometeorological data to obtain normality and linearity, but found that transformation did not improve the statistical fit

significantly, and did not provide physically meaningful results. An issue related specifically to regression analysis is the assumption that there are no errors in the independent variable (Tywoniuk and Wiebe, 1970). Another questionable assumption, in view of the short time series available for analysis, is that the population of the dependent variable is normally distributed about the regression line.

The central problem in multiple regression analysis is the intercorrelation of the predictor variables, and serial correlation of the dependent and independent variables. The non-random elements of hydrologic and meteorologic time series are well known, and to permit meaningful statistical modelling, an autoregression scheme can be used (Matalas, 1966). Discharge, radiation, and temperature show a high degree of serial correlation, while precipitation, and relative humidity show a lesser degree of serial correlation.

Regression analysis of glacial hydrologic data is a relatively new field, with one of the earliest papers (Mathews, 1964) being only ten years old. Several refinements in technique were made by successive workers: Østrem (1966) performed a stepwise multiple regression using temperature, precipitation, and wind velocity, but included no terms for time lags or storage factors. Lang (1968, 1973) paid more attention to seasonal and short term basin

lags by subdividing the melt season into four hydrologically distinct periods, and by testing the correlation coefficients for different time lags of 0 to 5 days. Stenborg (1969, 1970, 1973) employed a multiple regressive scheme in the most comprehensive study of seasonal length time lags or storage effects to date. Goodison (1972) used an autoregressive stepwise multiple regression technique, recognized the problem of intercorrelation, and took lag times into account.

The hydrometeorological data collected in the summer of 1974 was examined using regression analysis to determine the nature of the hydrology of Siphon Creek. The discharge record consists of two distinct flow regimes; an apparently simple low flow glacier melt response, and a catastrophic high magnitude event of short duration. Viewing the seasonal discharge record as a whole, the ice-dammed lake behaves as a major storage factor, which is responsible for the two distinct flow regimes. The regression analysis was performed to test the 'normality', a concept to be discussed in Chapter 5.4, of the discharge of Siphon Creek for the period of July 15 to August 9, 1974. Additionally, this type of analysis will define the relationship between a set of climatic parameters and discharge for a specific region. The geographical location of a glacier and prevailing macroscale climatic conditions strongly control the relative importance of the three main heat sources for the energy

balance (radiation, convection, and conduction). The study area is located in the high Arctic in an area of cold polar climate with low annual precipitation rates; and on that basis, this data differs significantly from the data utilized by other statistical models in the glacial hydrological literature. Another important difference is the presence of sub-polar glacier ice, and its influence on the nature of a glacial drainage system.

A critical factor in obtaining statistically significant results in multiple regression analysis is the selection of predictor variables and the appropriate length of lag times. The variables entered into this analysis are those meteorological parameters monitored during the study period; air temperature, relative humidity, precipitation, incoming solar radiation, and antecedent discharge, to include an autoregressive term. An intercorrelation matrix indicating the relationships between the independent variables and discharge is presented (Table 5.2). The short period of suitable record (July 15 to August 9, 1974) limits consideration of the concept of changing response times throughout the melt season (Stenborg, 1970). There appear to be no majcr longer term trends in the data which would warrant varying lag times over the short record. Such seasonal variations are usually in response to changes in the water storage capacity of the glacier's internal drainage system. The study basin includes sub-polar glacier

ice apparently drained entirely by the network of supraglacial streams, and only a minor snowpack storage term could possibly apply. So, for the period of normal flow, the length of lag of meteorologic variables which control the diurnal variations of discharge are of major concern.

From visual observation of the meteorological records, discharge in Siphon Creek seems to be strongly controlled by the diurnal fluctuations of temperature and radiation (Figure 5.1). When cold, overcast weather prevails, discharge falls off dramatically, and the reverse trend occurs during good weather, precluding the presence of any significant baseflow. This observation is supported by the spot gauging of a single major supraglacial stream draining the shelf ice on July 27, which was found to account for 90% of the total flow of Siphon Creek. Despite reports to the contrary in the literature (Gilbert, 1969; 1972; Fisher, 1973), there does not appear to be any significant leakage of water through the ice dam for McMaster Lake.

The close relationship between discharge and the meteorological variables can also be considered graphically. Figure 5.6 shows the two hourly values for the three meteorological parameters averaged over the two month record, while the two hourly values of discharge were averaged over the 25 day sample period. This plot demonstrates that some positive relationship exists between

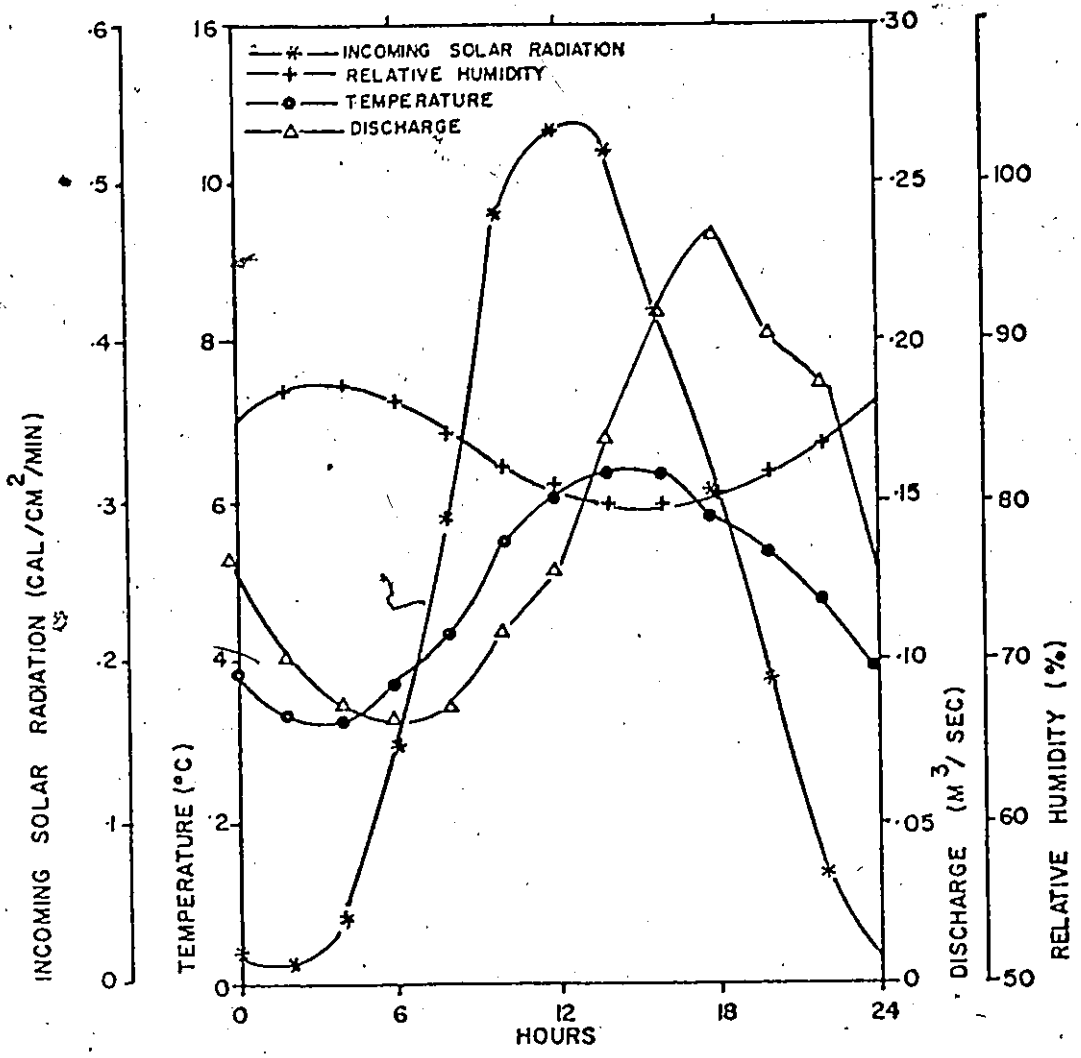


Figure 5.6 Comparison of the average two hourly for temperature, incoming solar radiation, relative humidity, and discharge in Siphon Creek.

discharge, temperature and incoming solar radiation, while an inverse relationship with relative humidity is apparent. On the average, solar radiation peaks sharply at ~1300 hours, followed by a broad temperature high between 1400 and 1600 hours. The plot of relative humidity is roughly a mirror image of temperature, with a similar low trough between 1400 and 1600 hours. Lagging these parameters by 3 to 5 hours, discharge peaks at 1800 hours. Given these visual observations, varying lag times of the meteorological variables were tested in the regression equation, as well as the standard unit of data available, the 24 hour lag. In some test runs, the small basin size and serial correlation of the variables warranted the use of average daily values with no lag. Discharge with a 24 hour lag was included as a predictor variable in some runs, to permit physically meaningful modelling of a non-random, serially correlated hydrologic series (Tywoniuk and Wiebe, 1970).

The general form of the first test run used was:

$$Q_2 = k + k1P_1 + k2P_2 + k3T_1 + k4T_2 + k5RH_1 + k6RH_2 + k7RAD_1 + k8RAD_2 + k9Q_1 \quad (5.2)$$

where: k1,k2,k3,... = regression coefficients

T = mean daily temperature (°C)

P = total daily precipitation (mm)

RH = mean daily relative humidity (%)

RAD = mean daily incoming solar radiation (cal/cm²/min)

Q = total daily discharge (m³/sec)

Table 5.2 Intercorrelation matrix for test run #1

	P_1	P_2	T_1	T_2	RH_1	RH_2	RAD_1	RAD_2	Q_1	Q_2
P_1	1.00	0.08	-0.42	-0.58	0.49	0.46	-0.45	-0.28	-0.27	-0.39
P_2		1.00	0.01	-0.42	0.16	0.49	-0.26	-0.41	0.18	-0.30
T_1			1.00	0.62	-0.85	-0.35	0.62	0.05	0.70	0.41
T_2				1.00	-0.56	-0.82	0.66	0.60	0.51	0.64
RH_1					1.00	0.54	-0.42	-0.17	-0.35	-0.26
RH_2						1.00	-0.99	-0.83	-0.09	-0.28
RAD_1							1.00	0.46	0.02	0.01
RAD_2								1.00	-0.11	-0.06
Q_1									1.00	0.65
Q_2										1.00

Table 5.3 Results of multiple regression analysis

		Regression Coefficients									Multiple Correlation Coefficient
Run#	Const.	P_1	P_2	T_1	T_2	RH_1	RH_2	RAD_1	RAD_2	Q_1	
1	-1400	-570	-1200	-830	2400	120	-240	-43	-420	0.39	0.94
2	-1900			100		21		8.4		0.70	0.85
3	-4400				380		50				0.66
4	-4000			360		47					0.65

Table 5.4 Comparison of partial correlation coefficients

Run#	P_1	P_2	T_1	T_2	RH_1	RH_2	RAD_1	RAD_2	Q_1
1	-0.39	-0.30	0.41	0.64	-0.26	-0.28	0.01	-0.06	0.65
2	-0.01		0.61		-0.35		0.29		0.82
3			0.60	0.61	-0.34	-0.34	0.28		
4	-0.10		0.61		-0.35		0.29		

subscript 2 = daily averages for a 24 hour period

subscript 1 = daily averages for the previous 24 hour period (ie. a 24 hour lag)

The use of all 9 predictor variables results in a multiple correlation coefficient of 0.944. Precipitation and radiation values for both lag times show negative partial correlation coefficients (Table 5.3), and the predictor variables of P_2, RH_1, RH_2, and RAD_2 did not significantly improve the multiple correlation coefficient. Therefore, the best equation (r= 0.94) for mean daily data is:

$$Q_2 = k + k1P_1 + k2T_1 + k3T_2 + k4RAD_1 + k5Q_1 \qquad (5.3)$$

and the estimated discharge values are plotted against actual discharge for comparison (Figure 5.7).

Considering the small size of the study basin, further analysis was performed to investigate the optimum length for averaging values, and the most accurate length of response period between the predictor variables and discharge. A second test run was made, dropping the predictor variables with a 24 hour lag and averaging the remaining variables over different time spans and with different lengths of lag. The equation for test run #2 was:

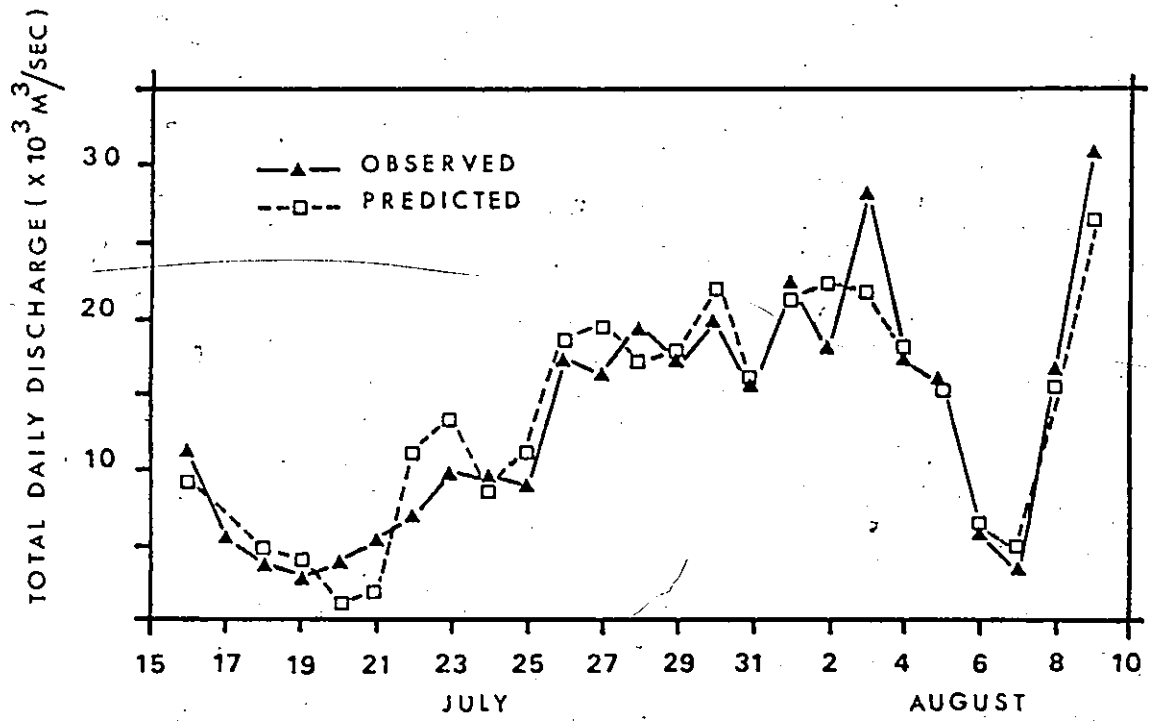
$$Q_2 = k + k1P_1 + k2T_1 + k3RH_1 + k4RAD_1 + k5Q_1 \qquad (5.4)$$

where: P_1 = total 6 hour precipitation with 6 hour lag

T_1 = 4 hour average temperature with 3 hour lag

RH_1 = 4 hour average relative humidity with 3 hour lag

RAD_1 = 4 hour average incoming solar radiation



$$Q_2 = k_1 + k_2 P_2 + k_3 T_1 + k_4 T_2 + k_5 RAD_2 + k_6 Q_1$$

Figure 5.7 Observed versus predicted discharge for Siphon Creek, 1974.

with 5 hour lag

Q_1 = total 6 hour discharge with 4 hour lag

Q_2 = total 6 hour discharge

For this model, the values of all partial correlation coefficients are improved over test run #1 (Table 5.3), but the multiple correlation coefficient dropped to 0.853. This is due partially to the removal of serially correlated predictors, and the fact that the regression program was set to reject those variables which did not significantly improve the multiple correlation coefficient. Over the shorter lag periods the intercorrelation between variables is stronger, causing the improved partial correlation coefficients and the lower multiple correlation coefficient.

In test run #3, precipitation was dropped as well as antecedent runoff, with the remaining independent variables averaged over 6 hour periods with varying lag times. The equation for run #3 was:

$$Q_2 = k + k1T_1 + k2T_2 + k3RH_1 + k4RH_2 + k5RAD_1 \qquad (5.5)$$

where: T_1 = 6 hour average temperature with 4 hour lag

T_2 = 6 hour average temperature with 2 hour lag

RH_1 = 6 hour average relative humidity with 4 hour lag

RH_2 = 6 hour average relative humidity with 2 hour lag

RAD_1 = 6 hour average incoming solar radiation with 4 hour lag

Q_2 = total 6 hour discharge

The multiple correlation coefficient dropped to 0.655 while the standard error of the estimate increased mainly due to the fact that antecedent discharge was not included as a predictor. However, the values of the partial correlation coefficients did not change significantly over the last test run.

A final test run was made, again without antecedent runoff, to further explore the temporal relationship between the important meteorological parameters and discharge. This relationship is given by:

$$Q_2 = k + k1P_1 + k2T_1 + k3RH_1 + k4RAD_1 \qquad (5.6)$$

Where: P_1 = total 6 hour precipitation with 4 hour lag

T_1 = 4 hour average temperature with 3 hour lag

RH_1 = 4 hour average relative humidity with 3 hour lag

RAD_1 = 4 hour average incoming solar radiation with 5 hour lag

The multiple correlation coefficient of test run #4 is 0.655, and only temperature and relative humidity had partial correlation coefficients high enough to be included in the regression equation.

Recalling the objectives of the statistical analysis, it appears that they are adequately met by the multiple regression model. With a 94% explanation of variance in discharge by meteorological variables and antecedent discharge with a 24 hour lag time, it is apparent that the observation that there are no major storage lags in

the Siphon Creek drainage basin for the period July 12 to August 9 is confirmed. The implications of this finding will be discussed further in the next section, but the simplicity and shortness of the meltwater routeway from point of melt to the Siphon Creek gauge make it seem plausible.

The second objective of the model stated that the precise relationship between discharge and the meteorological parameters is primarily a function of geographic location. A variety of different lag times and average values were tested for the meteorological variables. Some tentative conclusions regarding the different predictor variables and the most appropriate lag times for each can be made, and a regional comparison of the relative importance of predictors is also possible.

Quite understandably, antecedent discharge (Q_1), almost irregardless of lag time, is the single most important predictor variable. The shorter the lag time, the better the correlation; as is shown by the increased value of the partial correlation coefficients (Table 5.3) from .650 to . 917 when the discharge was averaged over a 24 hour period and a 6 hour period respectively. The high intercorrelation value (Table 5.4) between Q_1 and Q_2 raises the question of the physical veracity of models containing antecedent discharge, so several test runs were made without this predictor. Air temperature is apparently the single

best meteorological parameter in predicting discharge. From examining the partial correlation coefficients, it appears that the most physically meaningful relationship for Siphon Creek is defined by averaging predictors over a 4 to 6 hour period and lagging the response time by 2 to 3 hours. The strong intercorrelation between temperature and relative humidity (-.0821 for T_2 and RH_2) produces quite high partial correlation coefficients for RH (.346 for test runs #2 and 4), but temperature alone provides adequate explanation of discharge. Incoming solar radiation shows consistently low partial correlation coefficients in all test runs, but the high intercorrelation value metween T_1 and RAD_1 (0.620) is the best explanation of this pattern.

Comparison of these results with those obtained by other workers reveals mostly minor differences, primarily due to variation in geographical location and the attendant microclimate. Jensen and Lang (1973) in a study of the Z'mutt Glacier in the Swiss Alps, found that temperature and relative humidity were the most important predictors, after antecedent runoff. Østrem (1973), in a simple model to predict discharge one or two days in advance, found that temperature, precipitation and wind velocity produced the highest correlations for a Norwegian clacier located in a maritime climatic zone. Lang (1969,1973) did not include antecedent discharge as a variable, and found that temperature, vapour pressure and precipitation were the

important predictors. Goodison (1972), in a detailed study of Peyto Glacier, Alberta, described previous days discharge and temperature as the overall dominant variables, with the role played by other meteorological parameters varying with weather conditions and seasonal (lag) factors.

5.4 Flow regimes in glacial hydrology

A theme prevailing through much of this paper concerns the spatial and temporal variation of glacial flow regimes. Movement of water through, over, under, or beside an ice mass, the seasonal and diurnal timing of this activity, and the significance of the geographic location of the glacier are all relatively unexplored topics in the field of Arctic glacial hydrology. In fact, the entire subject of glacial hydrology is a complex and imperfectly understood science, with three major symposia in the last decade (International Association of Scientific Hydrology, Cambridge, 1969 and Banff, 1972; National Research Council, Quebec City, 1971) covering the bulk of the present knowledge of the subject.

One aspect of the dichotomy between polar and temperate glaciers is the relative simplicity, and therefore predictability, of the polar or sub-polar flow regime. Due to the virtual absence of englacial or subglacial routes, the basic system consists of snowmelt, supraglacial or marginal runoff, and proglacial drainage. As was apparent

in the last section, simple deterministic models of high Arctic drainage systems can yield very high, short term correlation coefficients due to the simplicity of the network. Studies on temperate glaciers often reveal slower and more erratic drainage rates and therefore poorer correlations for short term runoff models (Østrem 1973; Lang, 1973). This is due to the increased complexity and resistance of the internal drainage networks. However, the complication of the time lag factor as perceived by Stenborg (1969;1970), is most notable in the presence of ice-dammed lakes, and affects all drainage systems to some degree.

The concepts of 'normal' and 'abnormal' drainage were first advanced by Maag (1969), in his study of the ice-dammed lakes of Axel Heiberg Island, and in a slightly different context by Church (1972). With this first in-depth study of sub-polar glacial hydraulics, the deeply ingrained bias of the ubiquity of subglacial and englacial channels (Stone, 1963b; Embleton and King, 1968) was seriously questioned. Additional observations of the predominance of supraglacial and marginal drainage networks in the Arctic (Ward, 1952a; Schytt, 1956; Ricker, 1962; Wilkins, 1973) reinforce Maag's ideas. Reports of exceptions to the polar ice--surficial drainage hypothesis can often be reinterpreted (see Chapter 6.1, and Maag) as surficially generated features such as supraglacial or marginal channels and crevasse systems which have reclosed

at the surface due to glacier motion or new snow accumulation. The high density of ice-dammed lakes, especially supraglacial features, reported in the high Arctic (Maag; Wilkins; Hattersley-Smith, 1969a; Blachut and Ballantyne, 1976) is only possible because of the cold glacier ice lacking internal drainage networks.

Maag cites several possible examples of englacial and subglacial drainage in Axel Heiberg Island, and designates them as abnormal, on the basis that cold ice is not usually capable of sustaining internal channelization. The case in point of McMaster Lake, with its apparently sub- or en- glacial drainage, appears to be an unusual Arctic example. No other examples of subglacial or englacial drainage were observed along the 40 km of ice cap margin in the Vendom Fiord study area, or detected on the aerial photographs of the Ellesmere Ice Cap (Chapter 2.3).

The englacial thermal regime of a glacier emerges as the prime determinant in the normality hypothesis. The reverse normal-abnormal relationship apparently holds true for temperate glacier ice, where a substantial literature (Liestøl, 1955; Charlesworth, 1957; Stone, 1963b) reports few examples of major marginal or supraglacial channels, and only then as exceptions to the normal subglacial and englacial processes. Ice temperature, while an exogenous variable in a glacial runoff model, appears to have a critical role in determining the mode of glacial drainage.

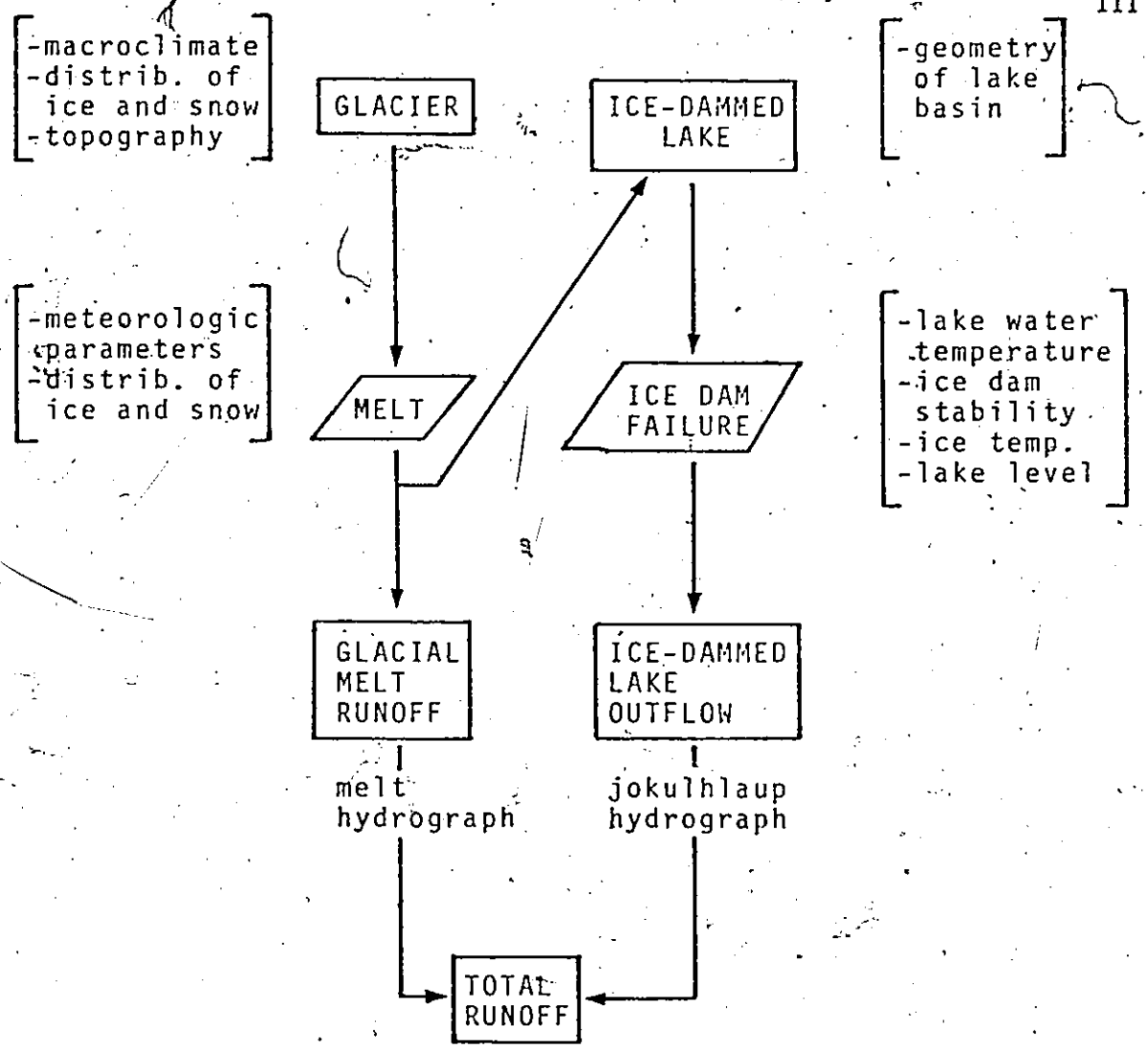


Figure 5.8 Partitioning of the high Arctic hydrograph into two major process-response components (the nival, rainfall and groundwater flow regimes are not included)

[ ] = boundary conditions
□ = physical state
▱ = process
→ = physical flow

From a hydrological viewpoint, the terms 'normal' and 'abnormal' are unsatisfactory, and the concepts can be considered in a simple process-response model (Figure 5.8). The hydrograph of a glaciated watershed can theoretically be partitioned into 5 major process-response components: glacier melt, nival, rainfall, groundwater, and ice-dammed lake outflow. In this instance only the first and last flow regimes are considered, and elements of the Siphon Creek and Sverdrup River hydrographs can be identified as either ice melt response or ice-dammed lake outflow response. Given these observations, the 'normal' flow regime equates to the melt hydrograph component, and 'abnormal' flow can be identified as the jokulhlaup hydrograph component. Differentiation of the jokulhlaup hydrograph on the basis of lake drainage routeway and release mechanism (see Chapter 5.5 and Figure 5.12) permits identification of several distinct jokulhlaup hydrograph patterns.

Examination of the hydrograph on a larger scale reveals finer structures similar to the lake burst outflow component. Shorter term storage of water at some point in the glacial hydrologic system can cause jokulhlaup-like variations on the basic melt hydrograph. To investigate these water storage terms, the movement of water through the glacier can be briefly considered.

Glacial hydrology, strictly defined as "the occurrence and behavior of the liquid phase within the

glacier" Mathews (1970), is generally understood to include the solid and vapour to liquid phase transformations at the snow and ice surface, and the discharge of water from the glacial system. This continuum of elements has rarely been studied as such, and with the exceptions of definable and quantifiable inputs and outputs, the hydrology of a glacier is essentially a black box system.

Detailed consideration of the glacier as a micrometeorological surface has advanced considerably since its inception as a scientific study in the late 1930's. Elucidation of the terms of the energy exchange of a glacier, field investigations and the monitoring of energy fluxes (LaChapelle, 1959; Andrews, 1964; Fohn, 1973), and the comparison of the calculated energy balance and measured ablation rates (Andrews, 1964; Muller and Keeler, 1969; Streten and Wendler, 1968) have raised this section of glacial hydrological investigations to quite a sophisticated level.

The study of water circulation within the glacier is best understood for the upper layers or snowpack. The snowpack, a complex three phase heterogenous system has been modelled using variants of a gravity flow model (Colbeck and Davidson, 1973; Male and Norum, 1971). The ground water analogue model presented by Colbeck and Davidson was found to be sufficiently accurate to have predictive power, but problems such as the spatial heterogeneity of snow, the

presence of ice layers and capillary structures, phase changes within the snowpack, the possible influences of impurities (Lliboutry, 1971), and logistic difficulties in measuring snowpack porosity and permeability have yet to be overcome.

The details of the energy exchange, ablation and snowpack percolation processes are beyond the scope of this paper, but the movement of water through glacier ice can be examined in more detail. This is the true 'black box' element of the glacial drainage system. To date, no detailed microscale examination of the movement of water through firn or ice has been made, while macroscale processes are known to a slightly greater degree. Statistical and mathematical modelling of various components of the drainage network has concentrated on predictive rather than explanatory models, but some parameters can be defined.

Theoretical studies on the thermodynamics of ice-water systems (Nye and Frank, 1973; Nye and Mae, 1972) have demonstrated the existence of a network of waterfilled passages situated along the three grain intersections in polycrystalline temperate ice. The misconceived idea of the 'impermeability' of granular ice at the intergranular level is seriously questioned by Nye and Franks expression for the rate of movement of water travelling through intergranular veins. Their calculations indicate that water movement to

the glacier bed via intergranular spaces amounts to a maximum volume of 1 m^3 of water per m^2 of ice surface in 1 years time. However, the more direct and proportionally much more significant mode of transportation of meltwater through the glacier is via the subglacial and englacial channel network present in most temperate glaciers.

Because of the deformable nature of ice, passages formed in it can expand and contract significantly in response to changes in water pressure on the passage walls relative to the ambient pressure of the enclosing ice. Nye and Frank considered the two pressures to be approximately equal. Shreve (1972) pointed out that melting continually tends to enlarge the passages, which lowers the water pressure and allows increased inward flow of the ice. This also causes the larger passages to grow relative to and at the expense of the smaller ones, creating an arborescent passage network. Modifying Nye and Franks equation by adding a term for surface melting rate, Shreve gives an expression for channel flow, and theoretically proves his concepts of tunnel enlargement by melt, and the enlargement of major passageways into a branching three-dimensional drainage network. Rothlisberger (1972) presents techniques and conclusions similar to Shreve, but tests some of his theory with field data. Campbell and Rasmussen (1973) consider a porous medium flow model for water movement through the glacier which uses Darcy flow laws, but allow

only a single constant term for ice permeability. Derikx (1973) and Derikx and Loijens (1971) also use ground water theory to describe water movement through the glacier, while Mathews (1973) and Gilbert (1972) have developed an expression for channel enlargement by melting, for the special situation of catastrophic drainage of an ice-dammed lake.

Storage of liquid water within the glacier system is a factor acknowledged by most workers in glacial hydrology, but studied in detail by very few. The concepts of variable lag times and long term storage have received considerable attention in the classic hydrologic literature (Laurenson, 1964; Mandeville and O'Donnell, 1973; Askew, 1970), but most of these studies involve terms for generalized basin lag and non specific reservoir storage. Many time lags in the glacial hydrological system are in the order of several days or less, and a majority of these terms can be eliminated over a fairly long time period. Stenborg (1970) however, provides an explanation for a trend in discharge records throughout the melt season in terms of a longer term storage factor. This commonly observed pattern is one of measured discharge being underestimated by calculated discharge in the early part of the melt season, overestimated in the middle of the melt period and equalled or exceeded in the latter part of the season (Figure 5.9). Patterns of relatively constant water storage values for the start of

the melt season, and substantial drops in this term for the mid-season are demonstrated empirically by Elliston (1973) and Krimmel et al (1973) (Figure 5.10).

Golubev (1973) defined glacier storage as:

$$W = S\,H \qquad (5.9)$$

where: W = glacier storage

S = surface area of the glacier

H = the mean head of water in the glacier

When there is no inflow to the system, runoff (Q) from the glacier depends directly on the volume of the accumulated water storage, and changes in volume (dW) correspond to changes in discharge (dQ):

$$dW = \tau \; dQ \qquad (5.10)$$

where: τ = time of decrease of glacier storage with a given steady state flow, or average basin lag time

Since τ is the parameter which defines glacier storage, its relationship is given as:

$$\tau = S / ab\,H^{b-1} \qquad (5.11)$$

where: a = coefficient defining passage size

b = coefficient defining permeability

The average basin lag for a small temperate glacier was found by Golubev to be 2.5 days, which is in general agreement with the results of others (Stenborg, 1970; Lang, 1973). Golubev, however, takes the analysis one step further by considering the basin lags for the accumulation

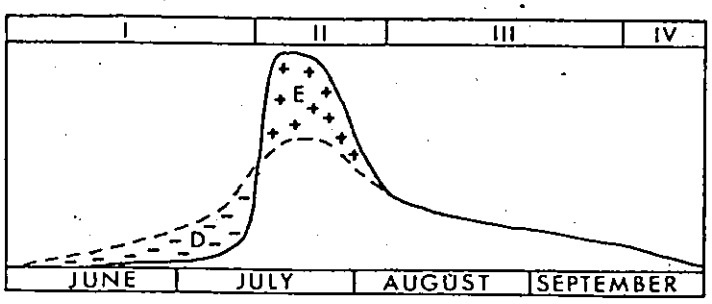


Figure 5.9 The role of storage lags in the seasonal discharge record (Stenborg, 1970)
—— = actual discharge
---- = potential runoff (melt + rain)
D & E= deficit and excess in discharge volumes

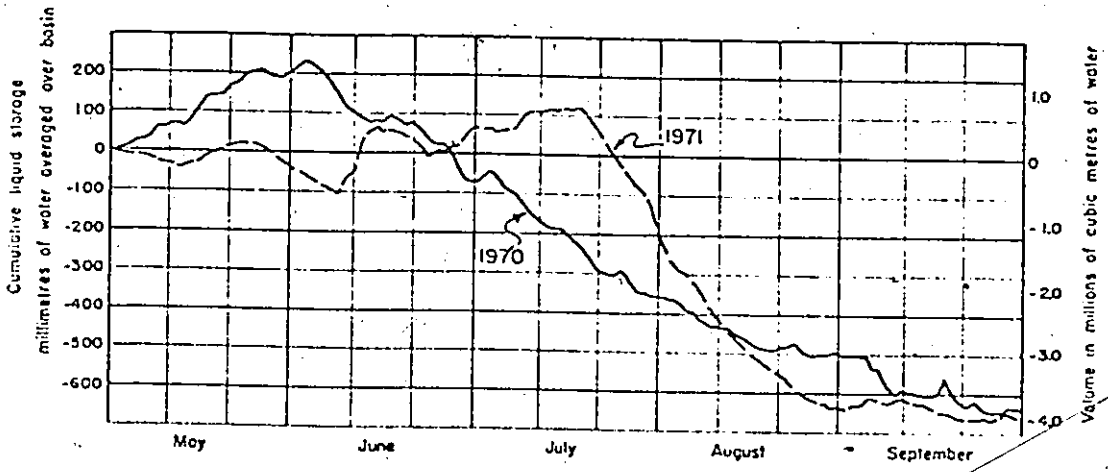


Figure 5.10 Liquid water storage for the South Cascade glacier (Krimmel, Tangborn and Meier, 1973).

and ablation areas separately. This yielded values of 0.125 days (3 hours for the lag time of the bare ice tongue, and 4 days for the lag in the accumulation area).

No model of glacial drainage systems breaks these storage factors down in to individual terms, despite the wide variety of storage features potentially present in the system. The storage terms can be subdivided into two scales, seasonal and longer term delays in runoff, and shorter term, temporary, random occurrence or diurnal fluctuations. These time lags can be expressed in terms of variations on the hydrograph, with the length and size of fluctuation depending on the nature and magnitude of the changes in water supply (Table 5.5).

Tying the glacial hydrological system and storage terms together for presentation purposes has not been presented in even the more sophisticated models such as Derikx and Loijens (1972) and Campbell and Rasmussen (1973). Figure 5.11 is a first attempt at flow charting of the glacial drainage network including all storage terms, and the inherent complexities of the system are readily apparent. Operationalizing such a model would require development of the appropriate mathematical expressions probably using the ground water analogue, a very good and complete set of field data, and in the case of some of the storage lags, methods of quantification which have not yet been worked out. However, the presence and importance of

TABLE 5.5 POSSIBLE STORAGE LAGS ASSOCIATED WITH THE GLACIAL HYDRAULIC SYSTEM.

SHORT TERM STORAGE FACTORS (diurnal to 1 week periods)

1. diurnal variations in meteorological parameters causing temporary storage when meltwater supply exceeds drainage capacity.
2. slush zones at the base of snowpacks or firn layers.
3. temporary storage in crevasses and moulins due to short term blockage of the internal drainage network.
4. collapse of channel walls, snow avalanche, icefall or rockslide.
5. random temporary factors which cause an excess of water supply: rainfall events, extreme temperature trends, and rapid glacier movement.

LONG TERM STORAGE FACTORS (seasonal or longer)

1. winter season causes refreezing of the surface layer, which reduces the intergranular permeability to 0 .
2. accumulations of snowpack over the winter, slow ripening and saturation of the snowpack in spring.
3. glacier flow, the settling of deep firn, destruction of intergranular veins, and englacial channels.
4. complete or partial winter closure of major drainage networks, due to decreased water flow and corresponding increased ice flow. Collapse of passages, refreezing of water to passage walls, silting up of subglacial channels also occurs.
5. presence of ice-dammed lakes in the system at any stage, releasing water in large bursts usually late in the melt season.
6. water trapped in subglacial sediments may recharge.

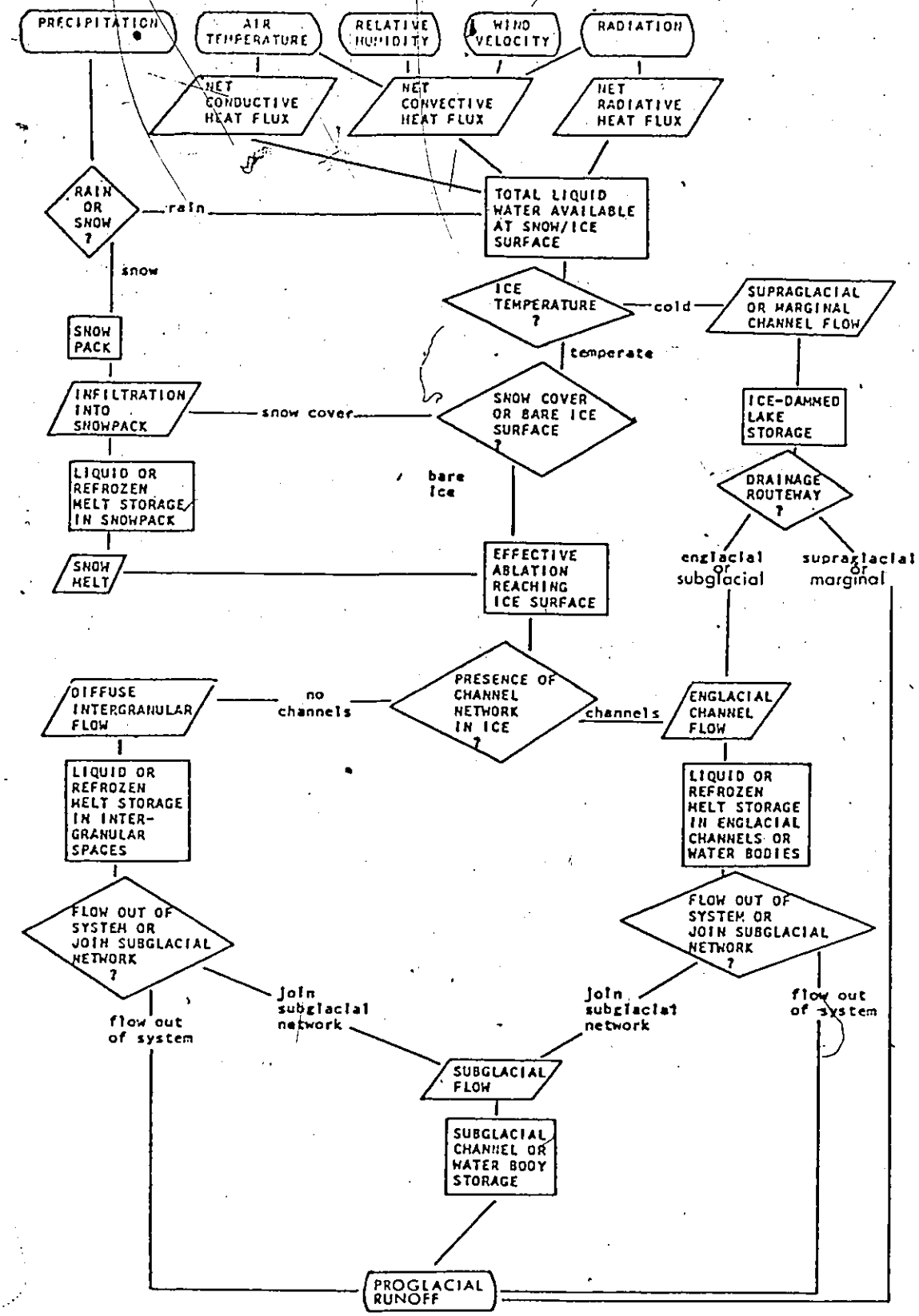


Figure 5.11 Flow chart of the glacial hydrologic system: ⬭ input, output; ▱ process; ▭ physical state; ◇ decision box.

the storage lags in the system has been identified, while decision boxes built into the model allows for differentiation between the surficial and interior drainage networks. Routing of water directly through the glacier model with minimal impedence to flow will result in a statistically predictable melt hydrograph. Storage of water at various stages of the glacial hydrologic system will yield a melt hydrograph compounded with the jokulhlaup hydrograph.

5.5 Jokulhlaupe

Very briefly, the jokulhlaup can be discussed as the unique, high magnitude flow event that it represents from the hydrologic point of view. Its role in the drainage behavior of ice-dammed lakes is discussed in more detail in Chapters 4 and 7.

Glacier outburst floods, glacier torrents, débâcles (Rabot, 1905), shwa (Mason, 1935) and jokulhlaupe (Thorarinsson, 1939b) have long been reported from almost all glacierized portions of the world, but consideration of the magnitude and frequency of these events has been a very recent study. Clague and Mathews (1973) presented historical data on the relationship between the magnitude of flood discharges with lake storage capacity (Equation 4.1), but they made no comments as to the frequency of these events. Because of the extremely high magnitude peak flows

associated with many jokulhlaupe, the concepts of effective forces in geomorphology (Wolman and Miller, 1960) dictate a low frequency for such events. However, consideration of the major storage lags built into many glacial systems raises serious questions of the validity of traditional magnitude-frequency relationships as applied to jokulhlaupe.

Well documented examples of the seasonal catastrophic drainage of ice-dammed lakes exist in the literature (Liestøl, 1955; Marcus, 1960; Stone, 1963a; 1963b; Maag, 1969; Higgins, 1970). With this type of lake, sufficient meltwater fills a lake basin, is stored on a seasonal basis, and an annual drainage event is triggered by any one of a number of possible drainage mechanisms. Irregular catastrophic drainage events are also documented (Thorarinsson, 1939b; Liestøl, 1955; Charlesworth, 1957; Stone, 1963b; Gilbert, 1971; Maag, 1969; Whalley, 1973), where periodicity of drainage varies across an almost continuous spectrum from 1 in 2 years to 1 in 100 years. Various factors are cited in the literature to be reponsible for this wide range of values, and Blachut and Ballantyne (1976) have reviewed this topic thoroughly.

The concepts of frequency-magnitude relationships must be put in a time context; in the long term most lakes are highly ephemeral features. The numerous Swiss lakes reported in the 19th century literature (Rabot, 1905; Charlesworth, 1957) have almost all disappeared with the

rapidly retreating glacier snouts. The record of Lake George, Alaska was over 100 years long when Stone (1963a) studied it, but this well known feature ceased to form in 1966. Many other lakes have such short records of behavior that no long term trends can be detected. Examination of the evidence of mammoth floods in the western United States during the late Pleistocene (Malde, 1968), reveals apparently long lived glacial and pluvial lakes bursting in one single catastrophic event. With such a reconsideration of the time frame, it appears that in the medium-short term (ie- 100 years), jokulhlaupe may act as high frequency-high magnitude events and represent a significant geomorphic force. However, in the long term, even though certain glaciated regions may be significantly altered by catastrophic drainage events, any one individual ice-dammed lake is a short lived feature. On this basis, the magnitude-frequency relationships for normal fluvial processes proposed by Wolman and Miller are perhaps inapplicable to the medium time scale glacial lake events.

Finally, the jokulhlaup displays a very distinctive pattern on the hydrograph, in some ways unlike any other flow event. The apparent randomness of flood wave arrival, unrelated to normal diurnal fluctuations or climatic influences, is easily recognisable for small scale (Maag, 1969; Ballantyne, 1975) or larger size events (Figure 5.4). Additionally, the abruptness of the arrival of the flood

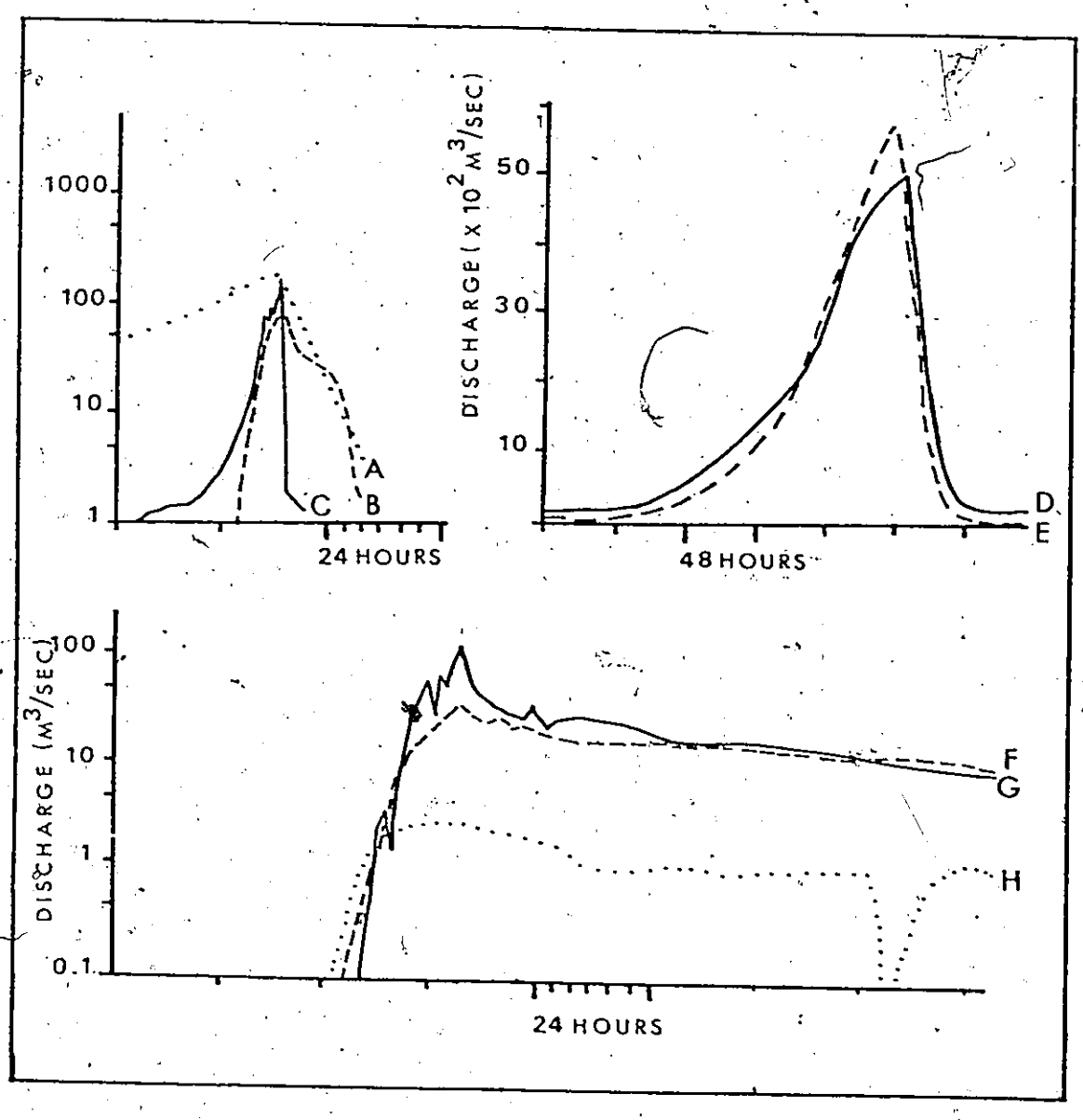


Figure 5.12 Comparison of discharge curves for several ice-dammed lakes

A Gjanupsvatn, Iceland (Thorarinsson, 1957)
B Marjelen See, Switzerland (Lutschg, 1915)
C South River, Baffin Island (Church, 1972)
D Graenalon, Iceland, 1939 (Thorarinsson, 1939b)
E Graenalon, Iceland, 1935 (Thorarinsson, 1939b)
F Between Lake, Axel Heiberg Island, 1961 (Maag, 1969)
G Between Lake, Axel Heiberg Island, 1963 (Maag, 1969)
H Crusoe-Baby Lake, Axel Heiberg Island (Maag, 1969)

wave, despite upstream wave attenuation, is characteristic of a jokulhlaup. Maag (1969) has proposed that the actual shape of the hydrograph is possibly diagnostic of drainage mechanisms. The pattern of supraglacial and marginal drainage is apparently slower and more gradual than subglacial or englacial drainage (Liestøl, 1955; Weidick, 1963; Maag, 1969). Examination of several hydrographs for catastrophically draining ice-dammed lakes (Figure 5.12) reveals a marked difference between polar or sub-polar glacier drainings and temperate glacier lake behavior. Two of the major lake drainage mechanism theories, barrier lifting (Thorarinsson, 1939b; Marcus, 1960) and tunnel enlargement by melting (Liestøl, 1955; Gilbert, 1972) should theoretically have characteristically different drainage curves. Curve C in Figure 5.12 can be explained as a very sharp rise in discharge as the ice barrier is initially floated off the glacier bed, a gradual tapering off as the lake empties and a sudden closing of the outlet as the hydrostatic head required for flotation is no longer available. Curve D is an example of an exponential increase in discharge as the englacial channel network is enlarged by melting and a sharp drop in discharge once the lake basin has emptied and the tunnels start to reclose. Not all recorded jokulhlaupe follow these patterns, but they may provide preminary information of the nature of glacial drainage systems in unstudied basins.

5.6 Summary

Investigation of the glacial hydrologic system and the Siphon Creek-Sverdrup River system in particular yielded the following results:

1. The meteorological record obtained in the Vendom Fiord area from 1973-1975 reveals a typical high Arctic short summer season (mid-June to September) with a high degree of variability between seasons. The McMaster Lakes camp recorded lower average temperatures, higher precipitation and stronger off-glacier winds than the base camp 375 m lower in elevation and 20 km distant.

2. The Sverdrup River hydrograph displays a characteristic high Arctic glacial regime, with a pronounced snowmelt flood and several jokulhlaup events. Siphon Creek experienced no distinct snowmelt event, showed a marked diurnal, low amplitude (maximum Q = 0.68 m^3/sec) discharge between July 4 and August 12, and a sudden two orders of magnitude increase in discharge due to the outflow of water from McMaster Lake.

3. Statistical analysis of the Siphon Creek and meteorological data for the period July 15 to August 9 yields a multiple correlation coefficient of 0.944, indicating a direct input of glacial melt with no major storage lags to Siphon Creek for this period.

4. The concept of 'normal' and 'abnormal' drainage for a particular ice thermal regime (Maag, 1969) is supported on the basis of observations in the Vendom Fiord area, and reconsidered from a hydrologic point of view. The glacial hydrologic system is examined, the major storage lags found in the system are enumerated, and a flow chart of the glacial hydrometeorological network is presented.

5. Jokulhlaupe are considered as unique flow events, and the magnitude-frequency relationships and distinctive jokulhlaup hydrograph are defined.

CHAPTER 6

GLACIAL FEATURES OF THE McMASTER LAKE BASIN

In conjunction with the primarily hydrologic and limnologic field program conducted at McMaster Lakes, some glaciological investigations were also carried out during the 1974 season. Ablation and ice temperatures were monitored at selected sites throughout the 1974 season. Of particular significance to this study of ice-dammed lakes, however, was the nature and behaviour of the floating shelf ice region which was examined in greater detail.

6.1 Ablation

As was noted in Chapter 2, the only major works on glaciology in the Canadian High Arctic were the Axel Heiberg Expeditions 1959-1969 (Muller et al., 1963; Adams, 1966; Maag, 1969) and work on Devon Island (Keeler, 1964; Paterson, 1969). Some of the largest expanses of sub-polar and polar glaciers outside the Antarctic and Greenland Ice Caps are found on Ellesmere Island, and there is little information available on these ice masses apart from the work of Hattersley-Smith (1955, 1960a, 1960b, 1963).

At the McMaster Lakes camp in 1974, net ablation and accumulation was measured every 7-10 days, at a series of 6

stakes roughly evenly spaced down the glacier margin from 391 to 771 m a.s.l. Ablation-accumulation was also monitored on the network of 20 survey stakes on the shelf ice surface, over shorter time intervals. For all stakes, the basic surface lowering technique (Muller and Keeler, 1969) was employed for ablation measurements, and probing to the ice surface was used to measure snow depths. It is realized that this amount of coverage is too low to make general comments (Adams, 1966). Specifically because there was only one stake representing every 60 m or so up the glacier profile, and some lateral extension of the network would have been desirable allowing for an average and more representative value for each altitudinal band.

A sketch map (Figure 6.1) of the study area indicates stake positions, and plots of cumulative net ablation-accumulation in water equivalents, against time. Movement of the temporary snowline up glacier was monitored from the start of the field season, with a snowpack present at all stake locations except Ablation Stake B on June 26. Retreat of the snowline was rapid between June 26 and July 3, due to the high temperature and radiation conditions of the period, and by the end of this time almost all snow (except for prominent drift banks) was removed from the shelf ice surface. The temporary snowline had moved to 640 m a.s.l. by July 10, but after this time the pattern is somewhat complicated by a snowstorm on July 17-18. At the

McMaster Lakes camp (located at 380 m a.s.l.), 7.4 mm (water equivalent) of precipitation fell, while 12.7 mm was recorded at Ablation Stake C (500 m a.s.l.), mostly in the form of wet snow. Due to this and other subsequent snowfalls over the icecap, the temporary snowline did not advance above 650 m a.s.l. in 1974, but the soaked zone of the accumulation area (Muller, 1962) extended above the highest stake (771 m a.s.l.)

Total ice mass lost by August 20 varied from 88 cm at Stake B (432 m a.s.l., on a nearly level surface, to 19 cm at Stake F (771 m a.s.l.). The point DS, on the shelf ice surface, was included to demonstrate the effect of a change in albedo. This stake was located near the upwelling site, and was on a 50% debris covered surface. The mean daily ablation rates for each stake were calculated and plotted against elevation. The point where the curve intersects the horizontal axis when extrapolated represents an estimate of the elevation of the equilibrium line (Muller, 1962) of the glacier. This point is about 960 m a.s.l. for the McMaster Lakes area in 1974.

6.2 Ice Thermal Regime

The concept of differentiating between two major geophysical groups of glacier ice on the basis of ice temperatures was first considered by Lagally (1932) and Ahlmann (1933). These two categories were: 1) Polar or cold

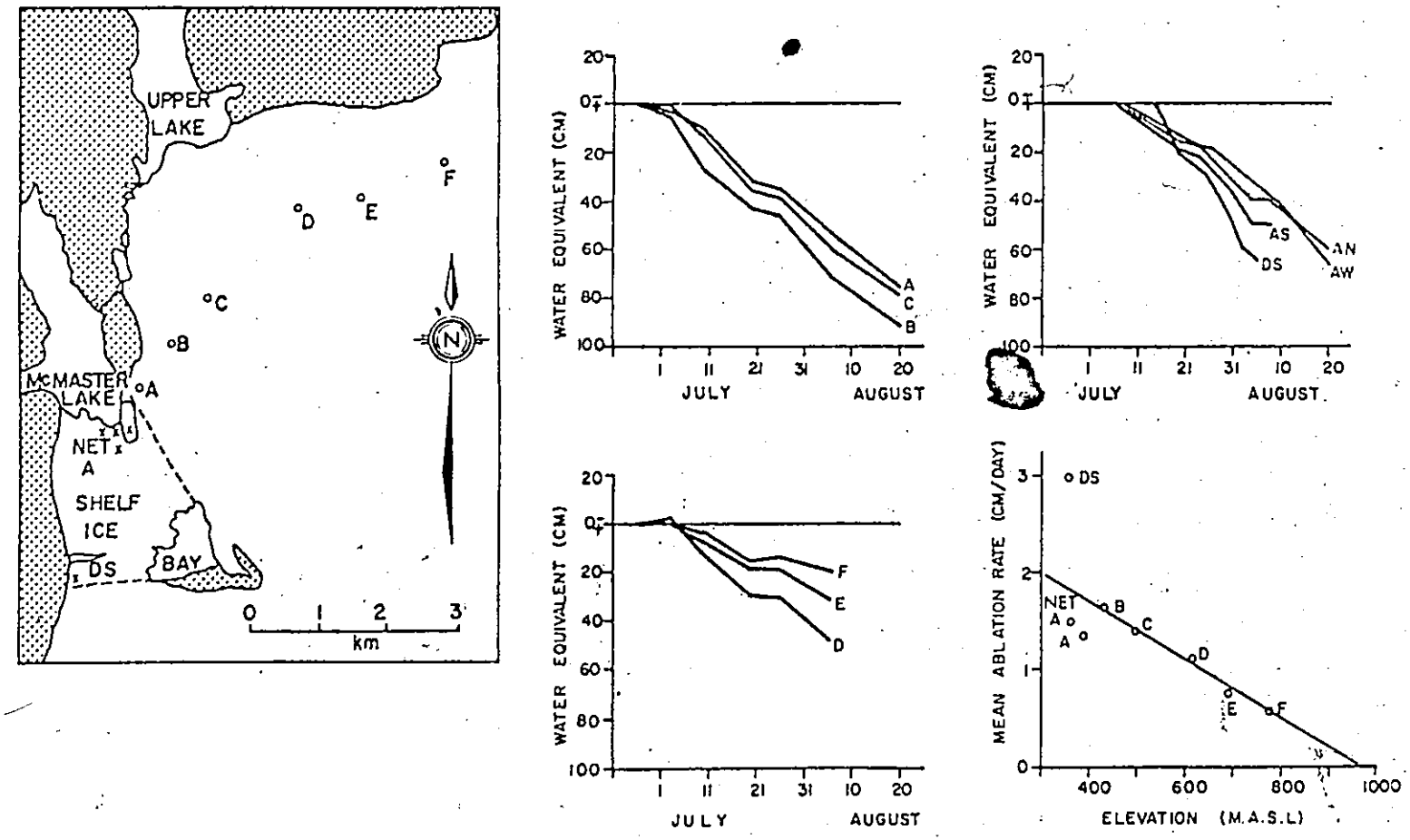


Figure 6.1 Network of ablation stakes monitored in the McMaster Lakes study area, 1974.

glaciers, defined as being perennially below 0°C for the entire depth, except for a shallow surface layer which will respond to seasonal climatic fluctuations ; and 2) Temperate or warm glaciers, which are understood to have temperatures, below a recurrent winter chill layer, consistently at the pressure melting point. Ahlmann also created a third category of sub-polar glaciers, where the penetration of seasonal warmth was restricted to a relatively shallow surface layer at 0°C. Miller (1975) attempts to provide a more precise terminology for definition of the categories, by setting arbitrary englacial temperature limits, obtained from a set of field observations. The categories are: Polar (-10 to -70°C), Sub-polar (-2 to -10°C), Sub-temperate (-0.1 to -2°C), and Temperate (0°C throughout). The significance of the ice thermal regime for the study of ice-dammed lakes has been detailed in the literature (see Chapter 2.2), and will be discussed further in Chapter 7.

The thermal regime of a glacier has been measured accurately for both polar and temperate glaciers (Paterson, 1969), including the 1387 m borehole at Camp Century, Greenland. In general, a negative temperature gradient is found for the main mass of a polar glacier.

At McMaster Lakes in 1974, shallow depth englacial temperatures were recorded at Stake B (432 m a.s.l.) using a set of six thermistor cables of varying length and a resistance bridge (Figure 6.2). The cables were

waterproofed and taped together prior to being inserted into a 3.5 m drill hole (3/4 " SIPRE ice auger) and allowed to freeze into place.

The range of temperatures recorded between the surface and the 3.5 m depth during the 2 month summer season, was 0°C to -17°C. This would indicate that the Ellesmere Ice Cap margin studied is sub-polar in regime. Seasonal temperature fluctuations penetrated well below the deepest thermistor, with a change of 7°C for the 3.25 m depth over the season. The curves are only slightly convergent at 3.25 m, which would imply that the lowest level with seasonal fluctuation can be projected to be on the order of 10-15 m, as recorded on Axel Heiberg Island by Muller et al (1963). The thermal gradient decreases steadily throughout the season, from 5.2C°/m on July 3, 1974 to 4.3C°/m on August 20. In the temperatures measured above the 50 cm level, several positive values were recorded which is explained by the presence of air or water in the top of the drill hole. Similarily, the positive temperature recorded at the 90 cm level on July 20 was thought to be due to an air pocket.

Ice temperatures were also monitored at a site on the shelf ice surface several times during the month of August, but no trend was apparent over the short time span. Little difference was detected between the values obtained for specific depths in the shelf ice and those read on the

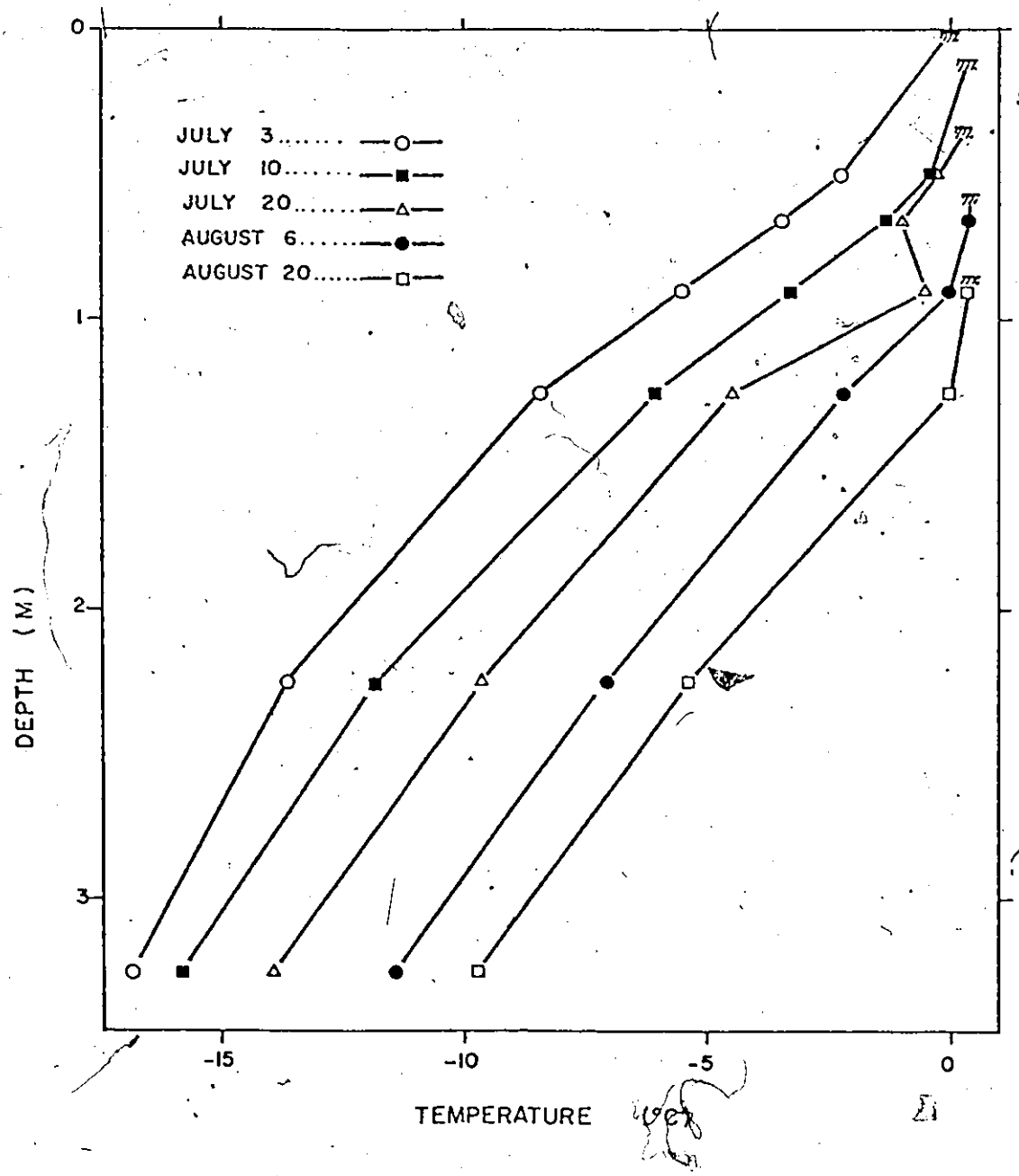


Figure 6.2 Shallow depth ice temperature measurements at Ablation stake B during the 1974 field season.

glacier surface at Stake B. Deeper drilling (i.e. below the level of the seasonal cold wave) might reveal interesting differences between shelf and glacier ice.

6.3 Origins and Behaviour of the Shelf Ice

The study of floating glacier shelf ice has attracted considerable attention, most notably concerning the large ice shelves and floating ice tongues of Antarctica (Debenham, 1948; 1965; Wexler, 1960; Budd, 1966; Holdsworth, 1969; 1974; Thomas, 1974). Other studies include the work on floating ice streams along the northern coasts of Greenland (Koch, 1928; Reeh, 1968), and the Ward Hunt Ice Shelf, northern Ellesmere Island (Crary, 1958, 1960; Hattersley-Smith, 1963a; Dorrer, 1971; Holdsworth, 1971). Little work has been done on the behaviour of glacier ice calving into a lake, the notable exceptions being Holdsworth (1973a, 1973b) and Holdsworth and McLaren (1971). A floating ice shelf and associated calving into ice-dammed McMaster Lake was studied during the 1974 field season.

A section of glacier ice (approximately 3 km² in area) directly to the south of McMaster Lake presents, on first observation, a set of unusual surface configurations, numerous small supraglacial ponds, and a contorted network of clear blue ice veins. The margin of the ice cap appears to be flowing down into a structural basin from the south and the east, around a nunatak in the southeast corner

(Figure 6.3). Due to the depth and extent of McMaster Lake as revealed by sounding, a large portion of this ice is thought to be floating (see Figure 6.4). It is not known exactly where the margins of the lake basin lie, or where the ice shelf is grounded, but the approximate limits of the shelf ice are plotted on Figure 6.3.

Evidence to confirm the theory that the shelf ice is indeed partially floating is provided by survey data. A network of 20 movement and ablation stakes (see Figure 6.3) was located on the shelf ice surface and surveyed several times over the field season, from the established baseline. An average of the stake elevations is presented in Figure 6.5, as a measure of the amount of displacement of portions of the shelf ice throughout the season, relative to water level fluctuations in McMaster Lake. The average elevation of the four Net A stakes, located close to the shelf ice front (Figure 6.5) indicates a close response to water level fluctuations both before and after the jokulhlaup. Changes in the average of Nets B, C, and D stake elevations further back on the shelf ice (Figure 6.5) show a weaker response to lake level changes, probably reflecting a partial grounding of the ice shelf, and proximity to the main glacier mass. The position of the stake nets in the cross-section and the configuration of the shelf ice-lake water and shelf ice bedrock interfaces are given in Figure 6.6. The shelf bottom topography is postulated on the basis of several lake

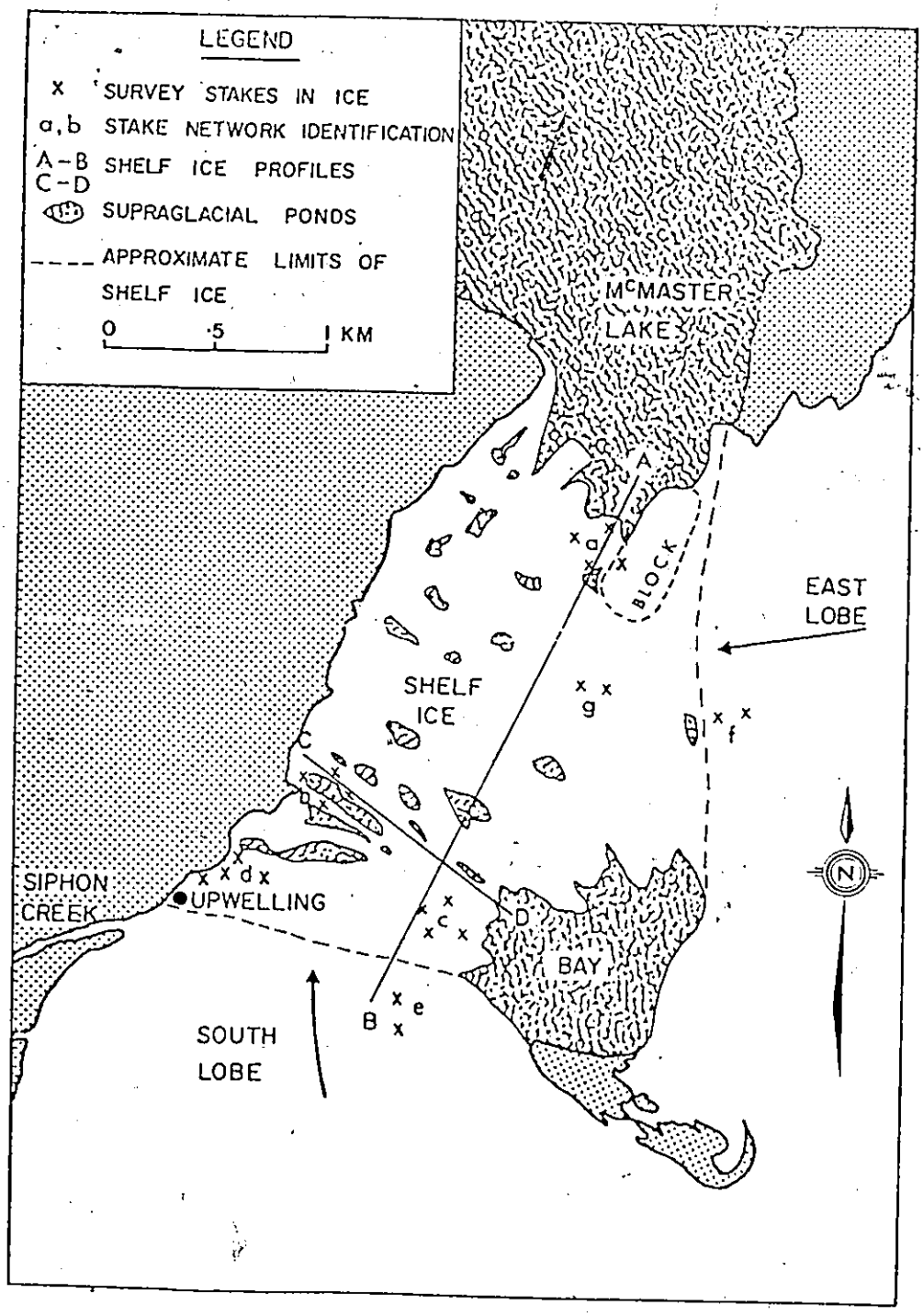


Figure 6.3 General location map of the shelf ice region, McMaster Lake.

Figure 6.4A Low altitude oblique photograph of the shelf ice region

Figure 6.4B Southern portion of the shelf ice. The upwelling site marked with an X

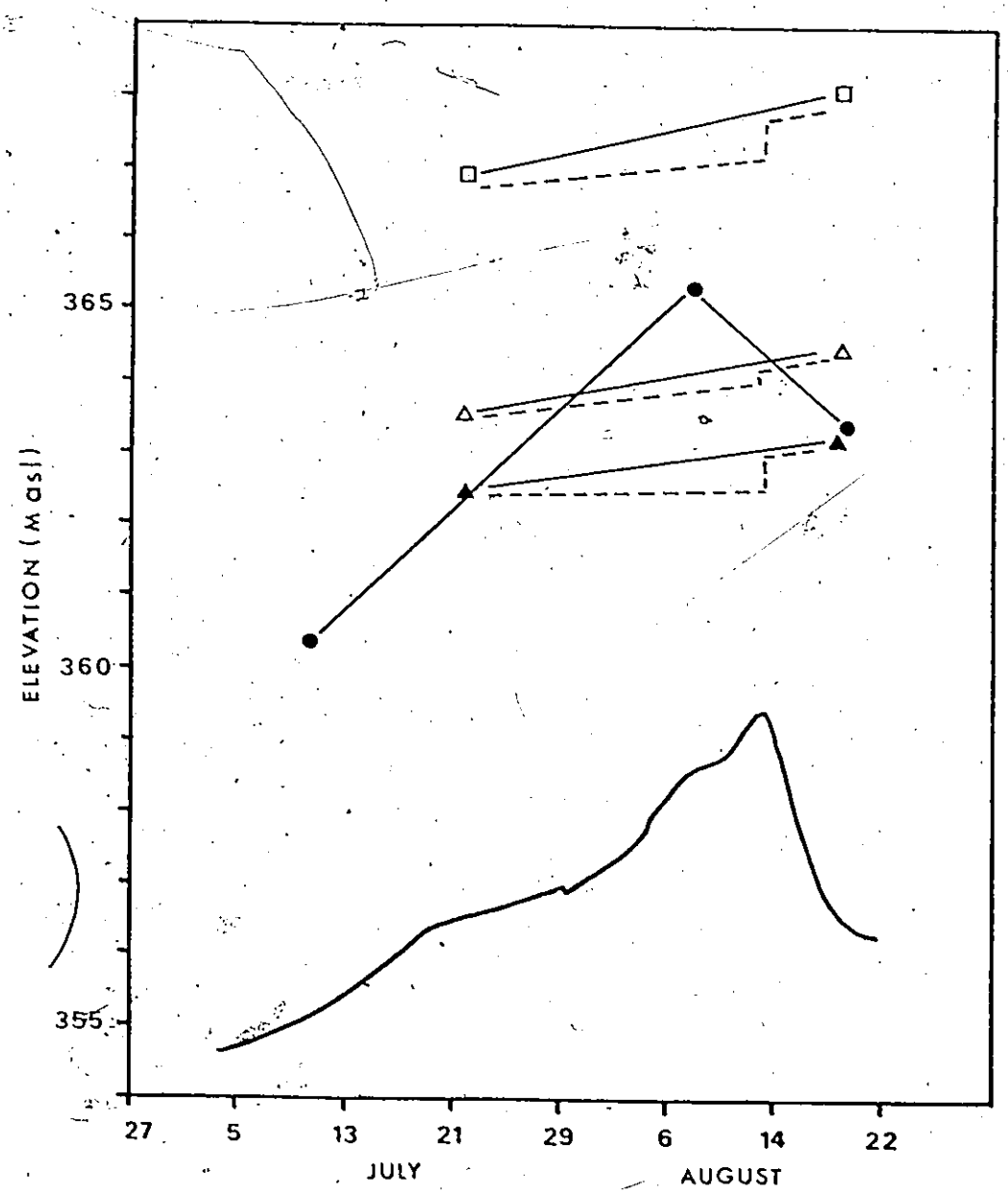


Figure 6.5 Displacement of the shelf ice throughout the season, relative to water level fluctuations. Average of the strain net elevations:
Net A = ● Net B = △ Net C = □ Net D = ▲
solid line connects recorded data points, dashed line postulates rapid displacement with jokulhlaup

depth soundings and the measurement of the height of the freeboard ice.

An additional demonstration that the shelf ice is floating and a possible indication of how portions of the shelf were formed, is the record of the calving of a large slab of glacier ice into the lake. A block of glacier ice 200 m wide by 500 m long calved off the glacier ramp on July 31, 1974 (see Figures 6.4A and 6.8). Profiles across the block were surveyed (Figure 6.7) and it was found that the front end of the block rose an average of 15 m above the shelf - lake ice surface (Figure 6.8B). The back of the block rotated outwards from the glacier front with a maximum horizontal displacement of 10 m, and dropped down 5 m relative to the glacier front. The pressure wave sent through the lake by this event created a network of cracks in the 3 m thick lake ice cover (see Figures 6.4A and 6.8A), and caused a seiche in the distal portions of McMaster Lake.

The mechanisms and factors responsible for the calving are unknown, but the work by Reeh (1968) and Holdsworth (1971, 1973a, 1973b) on shelf ice of similar thicknesses and in similar situations may provide some useful analogies. Studies on the much thicker (500-750 m) and larger shelf ice masses of Antarctica (Budd, 1966; Holdsworth, 1969, 1974; Thomas, 1974b) are not directly applicable, except for some of the theoretical considerations. In general, two main categories of theories

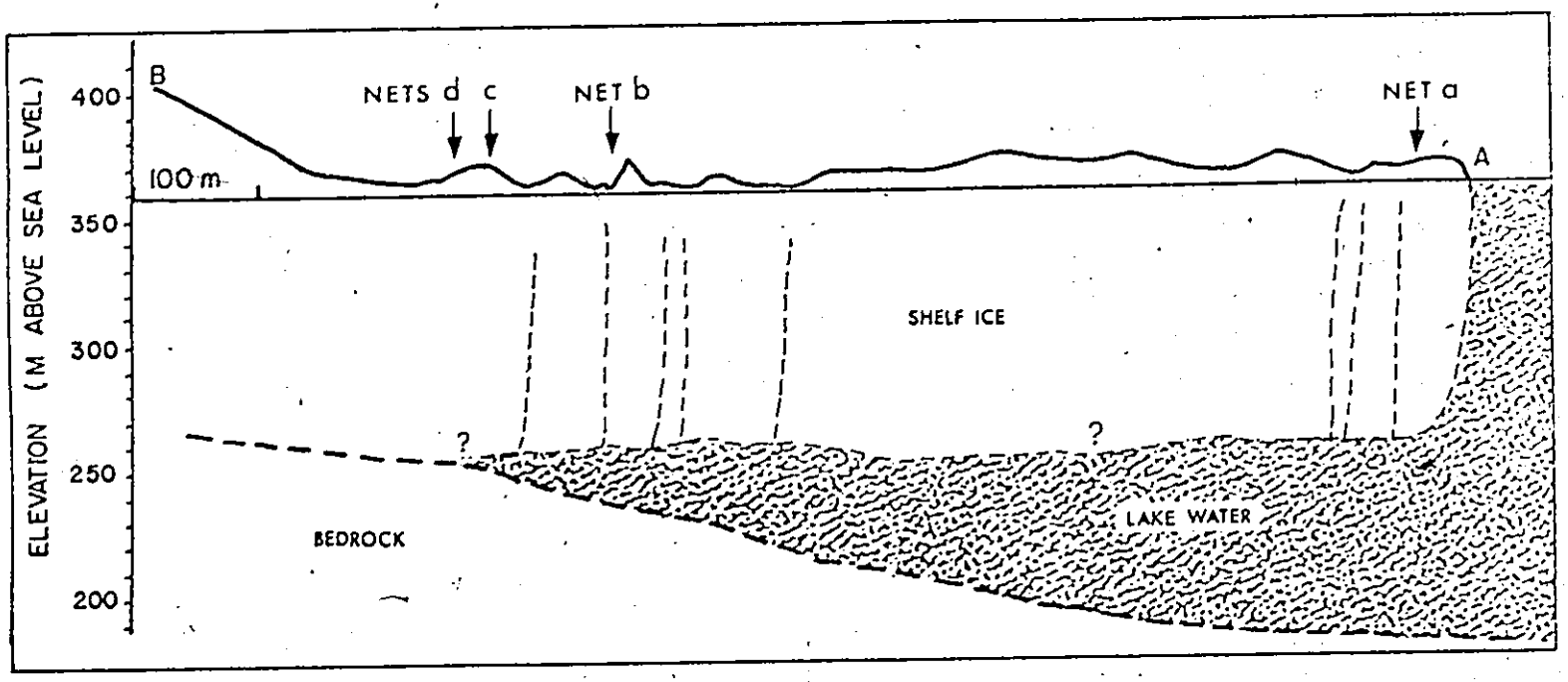


Figure 6.6 Cross-section of the floating ice shelf in McMaster Lake. The A-B profile and strain nets (see Figure 6.3 for location) are accurately surveyed. All other lines are conjectural.

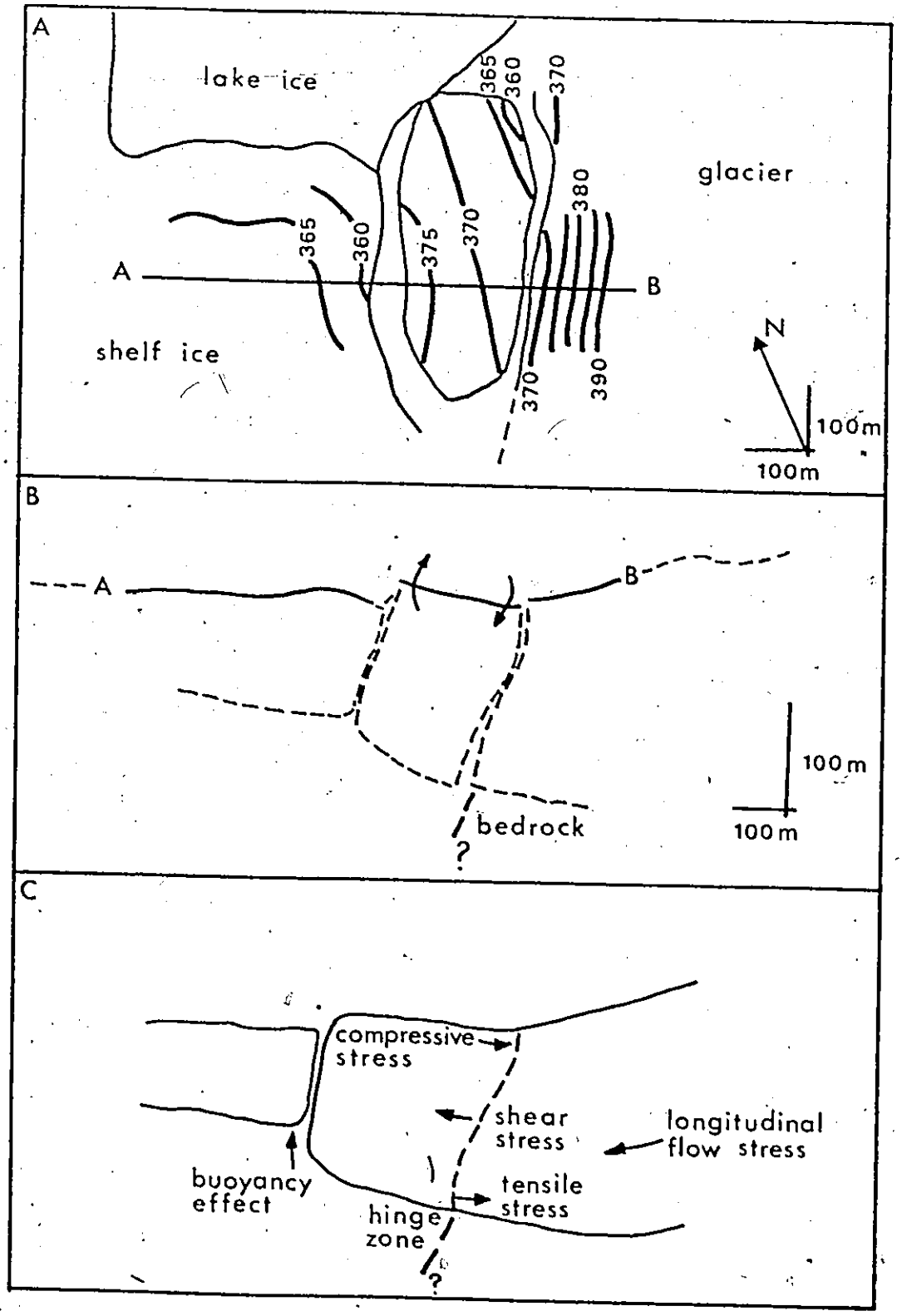


Figure 6.7A Contour map of the calved ice block
6.7B Cross-section profile of the block along line A-B
6.7C Cross-section of the ice ramp, showing stresses prior to the calving event

regarding calving mechanisms have been proposed in the literature: 1) calvings induced by rapid changes in water level, either tidal, storm waves, or lacustrine (Debenham, 1965; Holdsworth, 1973b, 1974), and 2) Reeh-type calving (Reeh, 1968; Holdsworth, 1973b), where, due to the imbalance of water pressure and ice stress, an ice ramp tends to deflect forward and downward until a critical value is reached. These two mechanisms operate on two different time and magnitude scales, and therefore are not directly comparable. According to Reeh's analysis, his theory accounts for the production of small prismatic icebergs. The critical effective stress zone is thought to be built up at a point where ice thickness equals distance from the ice front. This is in direct contrast to the huge tabular icebergs or ice islands (up to 10x10x0.4 km) produced by the Arctic (Hattersley-Smith, 1963) and Antarctic (Holdsworth, 1969) ice shelves.

In the McMaster Lake example, it appears that elements of both mechanisms contributed to the calving event of July 31, 1974. From the photographic record (Figure 4.4) it seems that there has been only one other calving event in the last twenty-five years, most likely between 1950 and 1959. Flow rates have not been measured, but between 1959 and 1974 the ice ramp advanced a distance of approximately 200 m into the lake. There is evidence of Reeh's downcurling of the ice front, as can be seen in Figures 6.8A

6.8A View east across the lake at upthrown face of block
6.8B Close-up of face from the lake ice surface
6.8C Hinge zone between ice ramp and calved block

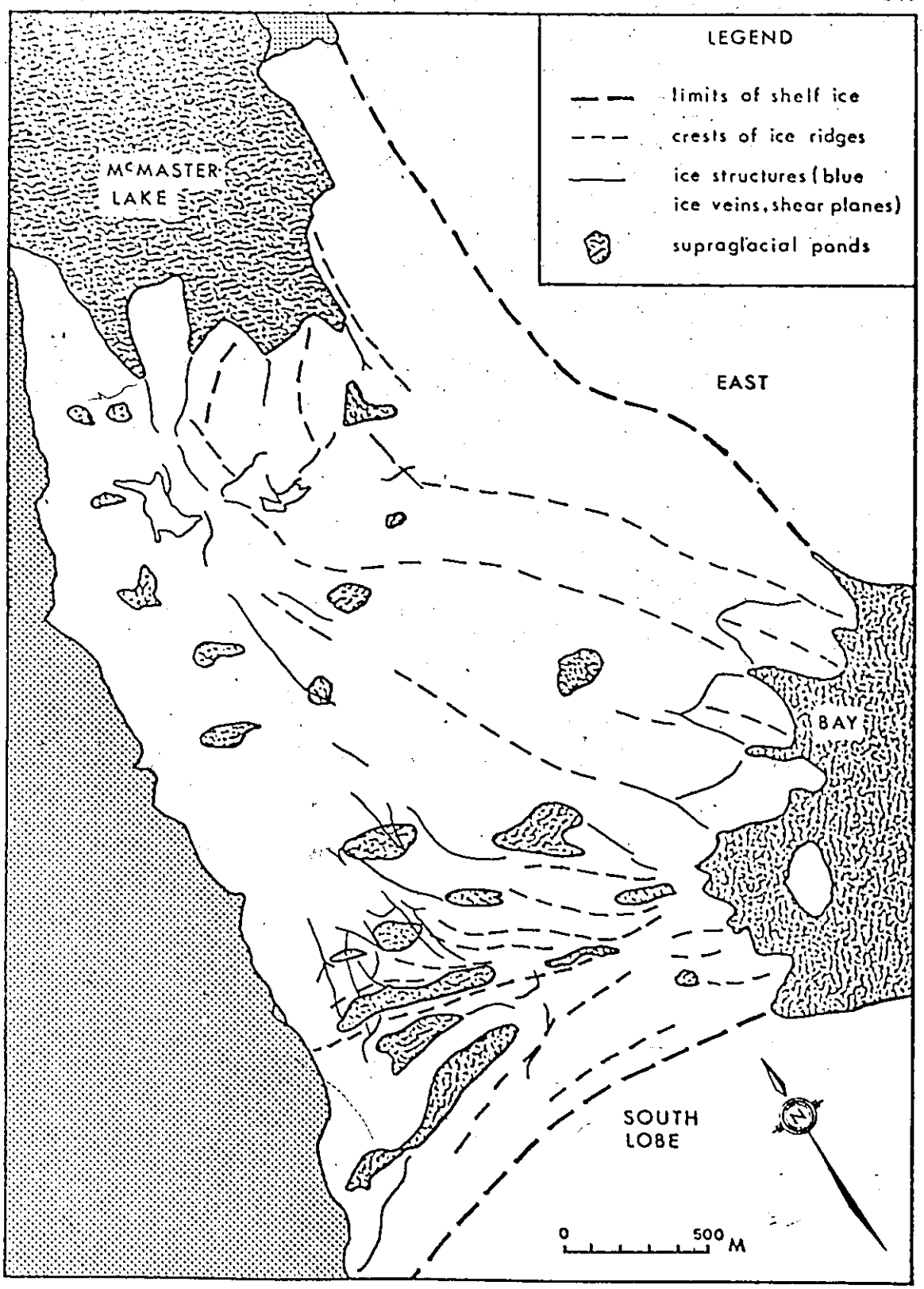


Figure 6.9 Sketch of the shelf icesurface showing structural features and supraglacial ponds.

and 6.8B. For a distance of 20 m back from the face of the iceberg, a layer of blue lake ice and surface snow is superimposed on top of the glacier ice. From the upthrown face of the block (Figure 6.8B), it is evident that the front of the ice ramp was depressed 1-2 m below water level. This would create an upward buoyancy force which would add to the bending (tensile and compressive) and shear stresses already acting upon the floating ice (see Figure 6.7C).

The triggering force which induced failure of the floating ice ramp was the rapid rise of water level in ice-dammed McMaster Lake. The lake level rose at a rate of 10 cm/day for 5 weeks prior to the calving event, increasing the level of the lake by 2.3 m. The added buoyancy force increases the bending stresses sufficiently to initiate failure of the ramp. However, in 1973, the lake rose to a maximum level 2.4 m higher than the failure level in 1974, with no apparent effect. Surface flow between June and August, 1974 was negligible, but basal flow rates must have been high enough between 1973 and 1974, to allow such unstable conditions to be set up, and the calving event to take place. There is insufficient data available to further analyze the bending problem, but from the work of Holdsworth (1973) on the calving of Barnes Ice Cap into Generator Lake, it is apparent that similar mechanisms operated in the McMaster Lake case.

The mode of formation and origin of the shelf ice can be speculated upon by examining the glaciers which feed the shelf, and their flow down into the lake basin. The East Lobe (see Figure 6.3) flows down a moderate gradient (about 50 m/km) crossing no major breaks in slope until reaching the east shore of McMaster Lake. Assuming that the fault scarp forming the shore (Thorsteinsson, 1972) continues south of the main part of the lake, the ice is flowing out into a 200 m deep basin, and is supported only by its hinge zone with the grounded glacier. The previous section demonstrated that medium scale (1-5 m) fluctuations in water level provide the bending stresses necessary to initiate failure of the ice ramp once every 20-25 years.

Examination of the topography of the northern portion of the shelf ice indicates that part of the ice mass is made up of formerly calved blocks held in place by the ambient shelf and lake ice, and refrozen in position (see Figures 6.4A and 6.9). The bands of blue ice indicating the outline of many of these 'resealed' blocks are thought to represent old layers of lake ice, or veins of refrozen meltwater. The pattern of parallel ridges, perpendicular to the direction of ice flow appear to be the 'rolls' referred to by Debenham (1965) and Holdsworth (1974) on several of the Antarctic ice tongues. The troughs between the rolls probably result from the plastic, stretching deformation of the ice as it begins to float. The central 1 km section of

the shelf ice appears to be a solid, unbroken mass which has flowed out across the lake basin for 1-1.5 km to be grounded or supported against the opposite rock slope of the lake basin. Along both margins of this ice tongue are projections or teeth, some of which are thought to be the the 'resealed' calved blocks. The southwest corner of the ice tongue abuts against ice flowing down from the South Lobe, or is partially grounded, as indicated by the heavily contorted ice surface and blue ice vein patterns.

The South Lobe is a minor diversion of flow from the McMaster Glacier which descends into the valley below the gorge. It is of a shallower gradient than the East Lobe and there is less of a break in slope as the ice reaches the lake basin. Little information is available on the configuration of the bedrock basin at this end of the lake. Depths of 150 and 83 m were measured in the Bay, but the continuous ice cover prevented the acquisition of other depth soundings. There is little evidence of calving of this ice front, reflecting a partial grounding of the ice or the confinement of the ice ramp on three sides. The pronounced distortion of the shelf ice in the southern portion of the basin is due to the coalescence of the South and East Lobes, and possible interaction with the bedrock topography.

6.5 Summary

To support the primarily hydrological and limnological field studies, some glaciological investigations were made. Ablation and ice temperature measurements were made as well as shelf ice analysis of McMaster Lake. The principal findings are:

1. Ablation measurements revealed an ice mass loss ranging from 88 cm to 19 cm (water equivalent) between 432 and 771 m a.s.l. respectively for the summer melt period. The melt season was interrupted several times by snowfalls, which showed a marked orographic effect with elevation. The temporary snowline advanced from below the study area (380 m a.s.l.) to 650 m a.s.l. over the summer.

2. Shallow-depth englacial temperatures were monitored, and a negative temperature gradient of 4.8 C°/m was found between the ice surface and 3.5 m depth.

3. A 3 km² section of glacier ice is demonstrated by its close response to lake level fluctuations to be floating in the McMaster Lake basin. The surface configuration, calving behavior and mode of formation of the shelf ice was studied in preparation for discussion of its crucial role in the drainage behavior of McMaster Lake.

CHAPTER 7

ICE-DAMMED LAKE DRAINAGE MECHANISMS

Much of the literature on ice-dammed lakes mentioned in Chapter 2.2 includes some discussion or speculation on the mechanisms responsible for lake drainage. This aspect of glacial hydrology is one of the most interesting and challenging, but there are numerous difficulties associated with its study. The location of many lakes is remote, and few lakes have been monitored on a longterm basis. The suddeness of many catastrophic drainage events is characteristic, but this does not enhance their predictability, and hence pre-burst descriptions and measurements are uncommon. The subglacial or englacial routeways and lake outlets are often inaccessible or in dangerously unstable condition so that direct investigation is commonly impossible. Sufficient data has been collected however, that a number of drainage mechanisms have been postulated. A discussion of the relevant mechanisms suggested in the literature will prepare for a specific description of the drainage events and the proposed means of drainage of McMaster Lake.

7.1 Theories of Drainage Mechanisms

The main ideas of possible relevance to the McMaster Lake example are: 1) barrier lifting, or flotation of the ice dam; 2) pressure deformation of the glacier ice at the lake bottom; 3) melt of a tunnel through or beneath the ice; and 4) multifactor or combination hypotheses. The other main categories of drainage mechanism are barrier overtopping and supraglacial runoff, and drainage associated with tectonic activity. These two concepts are clearly not applicable in the McMaster Lake example, and good reviews of these ideas are provided by Maag(1969), and Thorarinsson (1939b) and Tryggvason (1960) respectively.

1) The theory of barrier flotation is based on the simple relationship between the density of water and ice, where a body of ice will begin to float when surrounding water levels reach nine tenths of the ice thickness. Higgins (1970) refers to Rink (1862), who considered that an ice-dammed lake in Greenland emptied suddenly by lifting the ice barrier when it was full. Strom (1938) briefly mentions the glacier floating to allow escape of floodwaters. However, Thorarinsson (1939b) first formalized this concept, by stating: "a lake...can only be drained subglacially if it can lift the damming ice barrier sufficient for the water to find its way underneath" (p.221, Thorarinsson, 1939b). He gives an expression to predict the thickness (h) of the ice barrier:

$$h = \frac{x - m}{1 - \gamma} \qquad (7.1)$$

where x = difference in altitude between lake and ice dam

m = deduction factor to account for cracks in the ice

γ = specific gravity of the ice.

Thorarinsson substantiates his formula with data from Vatnadalur Lake, Iceland, while making the point that the critical (i.e. lifting) zone of the ice barrier may be some distance from the ice front. As will be seen in Chapter 7.3, this comment is of particular relevance when dealing with the ice shelf of McMaster Lake.

Thorarinsson's theory was soon criticized for neglecting to include a term in his equation for adhesion of the ice to the bedrock surface (Glen, 1954). The other and perhaps more serious comment made by Glen, and Liestøl (1955) was that once drainage is initiated, the water level will fall below 9/10's of the barrier height, and the outlet will close. Refilling of the lake basin to the critical level would trigger another release and an incomplete and irregular discharge would result, rather than the observed massive outflows. Other workers have doubted the applicability of this mechanism in situations where ice-dammed lakes are located in high tributary valleys, where ice thickness far exceeds lake depths, and the lake is a long distance from the glacier snout.

Despite the above mentioned criticisms, elements of Thorarinsson's theory are valid. Many ice-dammed lakes have an adjacent section of glacier ice which is heavily crevassed and at least partially floating, apparently subjected to disturbance prior to and during drainage. This pattern has been observed at Store Brimkjelen, Norway by Howarth (1968), at Phantom Lake, Axel Heiberg Island by Maag (1969), at a lake dammed by Strupbreen, Norway, by Aitkenhead (1960), and in Tulsequah Lake, British Columbia by Marcus (1960). Initiation of drainage when the lake fills to 9/10's the height of its ice dam is common (Lindsay, 1966; Mathews, 1964).

Several workers have taken Thorarinsson's basic idea and expanded upon it, eliminating some of the above mentioned weaknesses. Marcus (1960) has postulated that temporary lifting of the ice barrier is sufficient to allow water to be transformed from a hydrostatic to a hydrodynamic force, capable of opening routeways through or beneath the ice (Figure 7.1). The idea of tapping pre-existing drainage networks or lines of weakness in the ice is also mentioned by Marcus. Perhaps his most important contribution, however, is the concept that barrier lifting may just be a critical triggering factor, rather than the entire release mechanism. Another modification was introduced by Aitkenhead (1960), who suggested that barrier flotation may be followed by uneven settling of the glacier, or by the

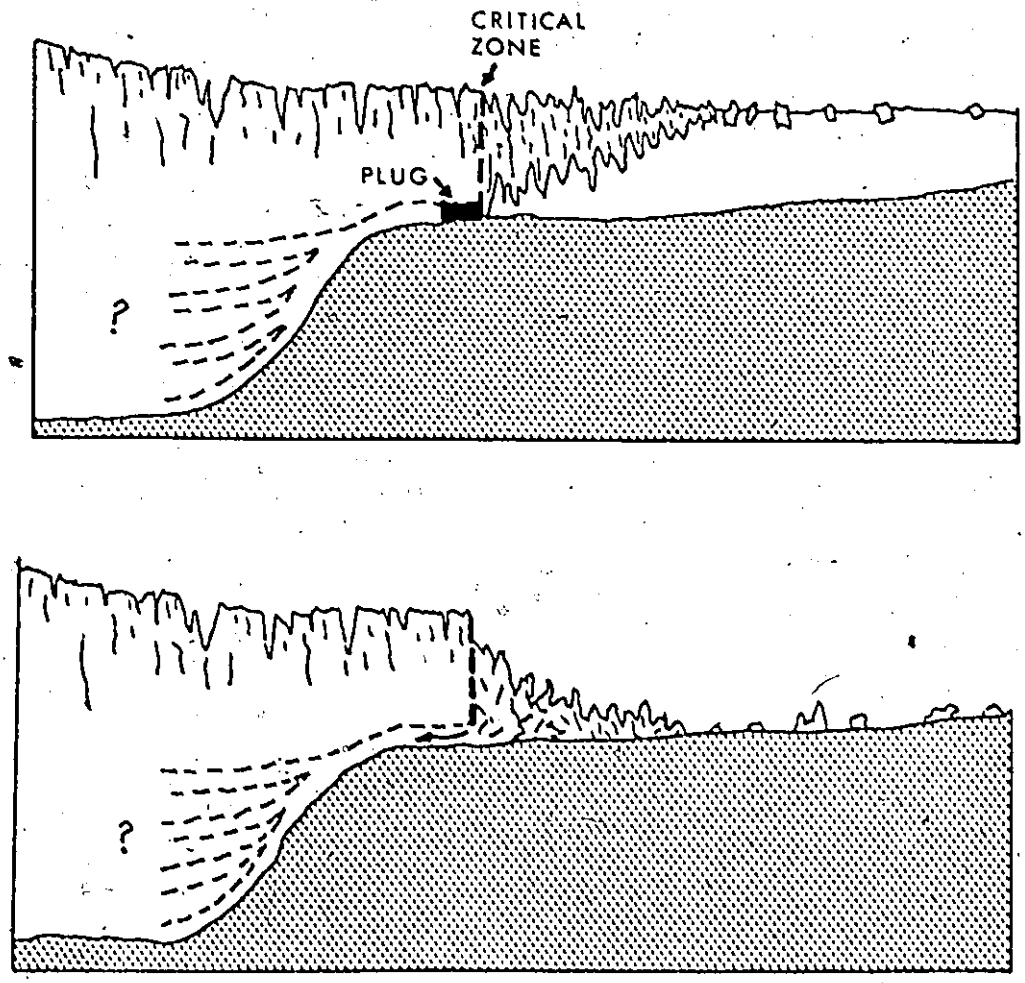


Figure 7.1 Hypothetical cross-section of glacier damming Tulsequah Lake, British Columbia. In the upper view, the lake is filled with water and heavily crevassed ice floats beyond the critical zone. After drainage, lower view, the floating ice mass collapses accounting for the severe ice breakage of the ice front. (Marcus, 1960)

wedging of icebergs in the outlet leaving an opening. Nichols and Miller (1952) realized that tension crevasses may form in the base of a thin ice dam due to flotation. This idea is supported by the recent work on floating ice shelves, as discussed in Chapter 6.3, where the tensile stress and buoyancy forces directly attributable to rising water levels are significant forces in the failure of ice shelves. Thorarinsson's theory clearly still has relevance in a comprehensive drainage mechanism theory, and will be discussed again in the combination theories, 4).

2) From experimental work on the mechanical properties of glacier ice (Glen, 1952; Nye, 1953), the generalized flow law was applied to the problem of ice-dammed lakes and water filled holes in glaciers by Glen (1954). The hydrostatic pressure of water at a depth z_w below the surface behaves as the normal horizontal stress σ_H, and is given as:

$$\sigma_H = \rho_w g \, z_w \qquad (7.2)$$

where ρ_w = density of water

g = acceleration due to gravity

The vertical compressive stress acting on the ice at depth z_i below the surface will be:

$$\sigma_v = \rho_i g \, z_i \qquad (7.3)$$

where ρ_i = density of ice.

If the hole fills with water so that $z_w = z_i$, the hydrostatic pressure of the water will exceed the ambient ice pressure by:

$$\Delta p = (\rho_w - \rho_i)\ g\ z \qquad (7.4)$$

or

$$\Delta p = \sigma_H - \sigma_V \qquad (7.5)$$

With horizontal stress greater than the vertical component, shear stresses will be set up at 45° to the horizontal, in the order of $\Delta p/2$. Results from experiments (Glen, 1953) indicate that shear stresses of greater than 1 bar. cause flow, and therefore a waterfilled hole subjected to a 1 bar stress would enlarge. The depth of failure of the ice is determined by:

$$(\rho_w - \rho_i)\ g\ z = 2 \text{ bars}$$

or

$$z = \frac{2 \times 10^6}{(1-\rho_i)\,980} = 200 \text{ metres} \qquad (7.6)$$

assuming $\rho_i = 0.9$ gm/cm³.

The major dispute with Glen's theory is the depth of hydrostatic head (Equation 7.6) required to initiate flow and create an outlet. Very few lakes are observed to be 200 m deep at the ice barrier. Tulsequah Lake, an example Glen uses, was a maximum of 150 m deep prior to 1959, and Marcus (1960) gives the lake depth as 73 m in 1958. However, Higgins (1970) and Weidick (1963) consider that the drainage of several 120-150 m deep ice-dammed lakes in South

Greenland can be explained by Glen's hypothesis. By definition, the outlet tunnel formed by pressure deformation would occur at the very base of the ice barrier, but incomplete drainage events are quite common. This would imply a tunnel not at the base of the dam, or blockage of the outflow by ice. The rate of enlargement by pressure deformation is not in agreement with observed rapid release times noted for some ice-dammed lakes; and many of these lakes are located a considerable distance from the ice front. Similar to the barrier lifting theory, a drop in water level would close the opening until sufficient pressure is built up again to allow failure of the ice. Liestøl (1955) also notes that tunnels can stay open for a considerable length of time after the lakes have drained. However, as with Thorarinsson's ideas, the pressure deformation theory of Glen might be a contributing factor in a combination drainage mechanism model.

3) Liestøl (1955) and Arnborg (1955) were the first to propose that the thermal influence of lake water can be the mechanism responsible for ice-dammed lake drainage. Liestøl states that if "the water from the lake has in some way forced a small passage beneath the ice, it will, by melting, be able to extend and keep open a tunnel" (p.123, Liestøl, 1955). It was shown theoretically, that water at 1.0°C flowing at 0.1 m^3/sec can melt over 100 m^3 of ice in 24 hours. Liestøl also realized that the potential energy

available in the dammed water will be transformed into heat by friction in the outlet tunnel. The amount of heat supplied (Q) is given by:

$$Q = \frac{H \times W \times 0.9}{427} \text{ kcal/sec} \qquad (7.7)$$

where H = height difference between lake level and tunnel entrance (m)

W = water flow (m^3/sec).

This amount of heat, given H=100 m and W=1 m^3/sec, can melt 250 m^3 of ice in 24 hours, assuming that the total heat surplus is used for melting. A certain amount of heat is added to the water, but is not used for melt. Depending on tunnel length and cross-section, Liestøl acknowledged that tunnel melting was a mechanism to enlarge an existing opening and facilitate lake drainage. He proposed that subglacial cavities or existing subglacial drainage networks provided lake water with an opportunity to start the melting process.

Liestøl's ideas were considered by later workers. Marcus (1960) tested the theory with data from Tulsequah Lake, and found that the theory was able to partially explain the observed behaviour. Aitkenhead (1960), Stone (1963b), Howarth (1968), and Maag (1969) all reference the tunnel melting idea, but it was not until 1969 that Mathews (1973) expanded upon Liestøl's early concepts. Gilbert (1969, 1971, 1972) has done the most to test and elaborate

on Liestøl's theory, using detailed temperature measurements in Summit Lake, British Columbia. Working from Mathews (1973) calculations of velocity of flow, amount of heat generated due to the loss in potential energy, and the heat loss by conduction into the ice, Gilbert (1969) developed expressions to compute tunnel water temperature, and the volume of ice melted. He then compared the lake water temperatures required to explain tunnel development due to melting, with empirical observations in Summit Lake. He found reasonable agreement between calculated and observed values, but postulated that a possible leak beneath the ice dam was present prior to the jokulhlaup. This implies that the tunnel melt theory adequately explains tunnel enlargement and the rapid escape of lake water, but does not provide a necessary triggering mechanism.

4) In the theories discussed previously, most workers qualified their statements by allowing that other factors may be involved in the drainage of ice-dammed lakes. A basic distinction should be made between factors which trigger the commencement of drainage, and the actual mechanism by which water is released from the lake. The pressure deformation theory of Glen, and Liestøl's ideas on tunnel melting constitute escape mechanisms, while ice-dam flotation, volcanic eruptions, and barrier overtopping fit into the first category. In his tunnel melt theory, Liestøl suggested that if water can in some way force a passage

beneath the ice barrier, the tunnel melt process can take over.

However, Marcus (1960) was the first to consider ice-dammed lake drainage as the result of a complex set of interdependent forces, rather than drainage according to a single simplistic principle. For the emptying of Tulsequah Lake, British Columbia, Marcus proposed the following sequence of events: flotation of glacier ice in a critical barrier zone; transformation of the water from a hydrostatic to a hydrodynamic force; clearance or bypassing of ice plugs in subglacial channels; resettling of the ice barrier around a melt-enlarged outlet tunnel; and connection to and escape through an existing network of subglacial and englacial tunnels. Aitkenhead (1960), Stone (1963b), Howarth (1968), and Higgins (1970) suggest various forms of multiple drainage mechanisms, with Thorarinsson-Glen and Thorarinsson-Liestøl combinations predominating.

Recent development in the nature and behaviour of the internal drainage system of a glacier (Stenborg, 1968, 1969, 1970; Haefeli, 1970; Rothlisberger, 1972; Shreve, 1972) (see Chapter 5.5), has led to better understanding and more comprehensive hypotheses of lake drainage mechanisms. Utilization of the existing subglacial or englacial drainage network by escaping floodwaters was mentioned by several workers (Kerr, 1934; Nichols and Miller, 1952; Liestøl, 1955; Marcus, 1960; Howarth, 1968; Higgins, 1970).

Different methods of joining up with the internal drainage system have been envisaged. Gilbert (1969, 1972) compiled a water balance for an ice-dammed lake to detect the presence of a slow leak (3-5 m^3/sec) through the ice dam prior to the jokulhlaup. Fisher (1973) used dye tracing techniques to confirm the existence of this slow leak, the presence of which bypasses the problem of initiating and forming a new tunnel each year.

Whalley (1971) adopted the ideas of Stenborg (1968, 1969) and proposed that the internal drainage system of the glacier itself initiates the jokulhlaup. As the network of internal channels extends through the glacier over the summer season, the base of an ice-dammed lake is thought to be eventually tapped by the drainage system. The hydraulic head of water in the lake is perched above the general piezometric head of the whole glacier (Haefeli, 1970), so once the initial break is made, the pressure will facilitate rapid drainage of the lake. Whalley uses this hypothesis to explain the behaviour of Stupvatnet, and provides an answer to some of the unknowns in other examples; such as Lindsay's (1966) steplike discharge curve, and short term fluctuations in discharge (Mathews, 1965).

Another category of drainage mechanism not considered because of limited applicability, has been suggested by several authors. Kerr (1934) postulated a bedrock tunnel beneath the ice to drain Tulsequah Lake,

while Nichols and Miller (1952) and Carey and Ahmad (1961) consider the possibility of floodwaters escaping through permeable glacial sediments.

Some observations and comments on drainage mechanisms that relate to the glacial-limnological-hydrological system as a whole can be made. These factors are often instrumental in the formation and behaviour of ice-dammed lakes, but are not specific drainage or triggering mechanism theories as discussed previously. Some of these items have been mentioned in Chapters 4 and 6, but will be summarized here briefly.

1) The geophysical nature of the glacier ice often controls the type of ice-dammed lake formed, and the nature of the drainage mechanism involved. A high density of ice-dammed lakes and a supraglacial or marginal mode of drainage appear to be the norm for polar and sub-polar ice (Ricker, 1962; Maag, 1969; Wilkins, 1973). Ice-dammed lakes associated with temperate glaciers, despite frequent mention in the literature, are less common, and subglacial or englacial drainage routes predominate (Embleton and King, 1968). For this reason, overtopping and supraglacial runoff is the most common in much of the Canadian High Arctic and North Greenland.

2) The connection between the regime and behaviour of a glacier, and its associated ice-dammed lakes has been recognized by most workers. Stone (1963b) identifies four

stages of ice-dammed lake formation associated with various positions of glacier snouts in the long Alaskan valleys, while Thorarinsson (1939b) identifies a similar sequence for lakes ponded by Vatnajokull, Iceland. More relevant to the discussion of drainage mechanisms is the idea that glacial activity controls the medium to long-term periodicity of drainage events (Stone, 1963b; Maag, 1969). Many lakes drain annually, but others drain with varying periodicity ranging from one event in two years (Maag) to one in twenty or more years (Thorarinsson, 1939b).

The timing of drainage events is generally restricted to late summer or early fall; with winter emptyings uncommon (Liestøl, 1955; Maag, 1969). This is considered to be a function of the filling behaviour of most lakes, where meltwater production over the summer months fills the lake basin to a critical level and a drainage event is triggered. However, Whalley's (1971) hypothesis of drainage by tapping into the existing glacial channel network introduces a new possibility. A burst may be triggered when the internal drainage network extends to the level and location ice-dammed lake. The increasing capacity of a drainage system as the summer progresses, described by Stenborg (1969), supports this hypothesis.

3) Finally, there is a point which should not be forgotten in the search for a comprehensive and applicable drainage mechanism theory. The complex nature and unique

elements of many ice-dammed lakes must be taken into account. The tendency towards oversimplification of a lake system, and the formulation of general drainage models from only one or two field examples can be misleading. Long, continuous records of several draining lakes, and the accumulation of data from more field areas may eventually lead to a better understanding of the mechanisms of ice-dammed lake drainage.

7.2 Description of McMaster Lake Drainage Events

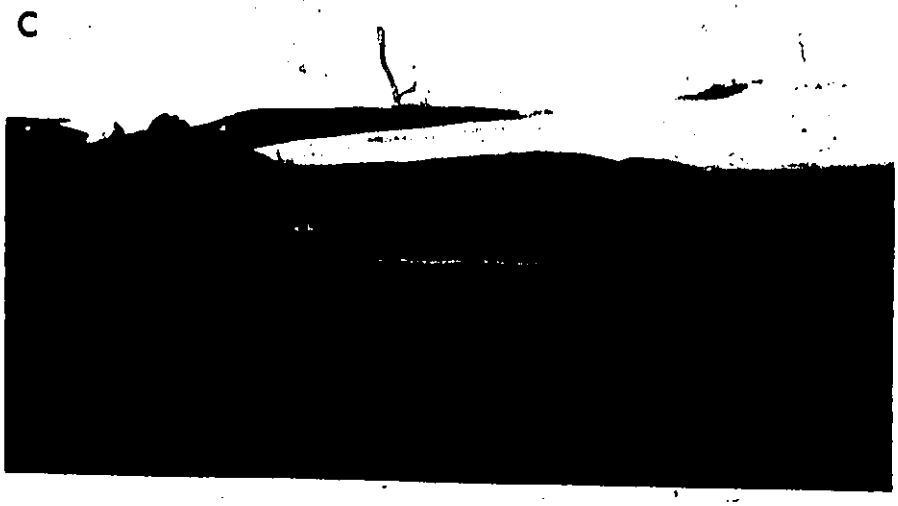
The limnological aspects of the McMaster Lake drainage events are discussed in Chapter 4.3, the hydrological data is considered in Chapter 5, and the contributing glaciological factors are described in Chapter 6. Prior to discussing the postulated drainage mechanisms of McMaster Lake the jokulhlaup is briefly summarized here. Some of the pertinent data regarding the drainage events which will be discussed in Chapter 7.3 are compiled in Table 7.1.

A partial draining of McMaster Lake commenced on August 1, 1973, with the sudden drop of the level of the lake correlated with the rapid increase of discharge in the Sverdrup River. There was evidence of disturbance of the shelf ice surface, changes in the level of supraglacial ponds on the shelf ice, and rapid downcutting of supraglacial channels. Water was upwelling at a site along

A

B

C



7.2A Just after the jokulhlaup of 1973
7.2B July 1, 1974 snowdrifts cover 1973 rafted icebergs
7.2C August 13, 1974 upwelling during jokulhlaup

the ice margin (see Figure 3.3 for location), where there was evidence of a violent and forceful extrusion of water under high pressure (Figure 7.2). Much of the ice margin was broken up in this zone, with displacement of large blocks of ice onto the shore and glacier surface, and the movement of icebergs downstream (Figure 7.2A). Upwelling was occurring along a 30-50 m stretch of ice margin, from a zone of heavily fractured ice. A supraglacial-marginal route was followed for about 100 m, before a low col was crossed into the Siphon Creek-McMaster Gorge system. Information on lake levels and volumes discharged is given in Table 7.1.

In 1974, the lake started to drain on August 13 from a level 1.2 m lower than in 1973. This represents a considerable decrease in available hydrostatic head between the two years, and a possible result of this is the difference in the 1974 mode of upwelling. There was no evidence of forcible extrusion of water, but rather the reopening of existing passageways, with no breakup of the ice margin (Figure 7.2C). The blocks of ice rafted in the 1973 burst still remained at the beginning (Figure 7.2B) and end of the 1974 melt season. A similar marginal routeway was followed until the low westward col into Siphon Creek was met.

Very little information is available on the 1975 drainage event, as the area was only visited for one day by

TABLE 7.1 SUMMARY OF McMASTER LAKE BEHAVIOUR 1973-1975.

YEAR	1973	1974	1975
START OF DRAINAGE	August 1	August 13	July 29
MAXIMUM LAKE LEVEL (m a.s.l.)	360.56	359.30	358.0 (1)
MAXIMUM DISCHARGE (m^3/sec)	310	73	? (2)
TOTAL VOLUME DISCHARGED (m^3)	42×10^6	34×10^6	25×10^6
TYPE OF DRAINAGE	upwelling from beneath shelf ice; marginal runoff	same as 1973	same as 1973
REMARKS	violent creation of new outlet, movement of icebergs downstream	reopening of existing outlet	same as 1974

(1) personal communication, S.B. McCann
(2) extrapolated from Sverdrup discharge record

a field party from the Vendom Fiord camp (S.B. McCann, C.K. Ballantyne, personal communication). The discharge in the Sverdrup river rose abruptly on July 29, 1975 and the McMaster Lakes site was investigated on July 31. The lake level was observed to have dropped from an even lower maximum level than in 1974, and the same upwelling site was reoccupied, with little disturbance of the ice margin.

7.3 Postulated Drainage Mechanism for McMaster Lake

The last point mentioned in Chapter 7.1 must be the first stressed in this discussion. McMaster Lake appears to be unlike many documented examples of ice-dammed lakes in terms of location, the configuration of the ice barrier, and the mode of drainage. However, the applicability of the main drainage mechanism theories discussed previously is considered prior to the postulated drainage hypothesis for McMaster Lake.

Contrary to previous statements (Chapter 2.3) on the apparent normality of supraglacial or simple marginal drainage of lakes dammed by polar or sub-polar glaciers (Maag, 1969), McMaster Lake does not follow this accepted norm. Floodwaters have been observed in three separate events to be upwelling in a marginal position, apparently from an englacial or subglacial source, and leaving the glacier margin after following a short marginal course. There is convincing evidence that supraglacial and marginal

channels predominate in the Vendom Fiord area, so the McMaster Lake example does not seriously challenge the cold ice-supraglacial drainage hypothesis.

The tunnel melting concepts of Liestøl (1955), Gilbert (1969) and Mathews (1973) do not appear to be a significant factor in the drainage of McMaster Lake. All waters in the McMaster Lake system and the outlet were found to be consistently between 0 and 0.1 °C, within the limits of the accuracy of the temperature measurements. Using Gilbert's formula for tunnel melt, it appears that a lake water temperature in excess of 3°C would be required to explain melt enlargement of the McMaster Lake outlet. Additionally, an important gravitational component would not be applicable where water is escaping the system by upwelling from a deep subglacial source. The water temperatures of the lake water probably allow slight enlargement of the outlet by melt, but this factor does not constitute sufficient impetus to initiate a drainage event.

McMaster Lake is one of the few documented examples of an ice-dammed lake satisfying the 150-200 m depth requirement of Glen's (1954) pressure deformation hypothesis. However, by Glen's definition, the stress conditions would be most likely to create an outlet tunnel at the very base of the ice dam. The McMaster Lakes example has an outlet to the ice surface near to or directly above the lake basin boundary, so the vertical compressive

stresses necessary to initiate failure of the ice at depth are not a factor, in the original manner proposed by Glen. However, the role of the stresses induced by increasing lake levels in acting upon the shelf ice mass render some of Glen's concepts applicable to the McMaster Lake event.

The behaviour of the floating shelf ice mass in McMaster Lake (Chapter 6.4), lends considerable validity to Thorarinsson's (1939b) barrier flotation theory. An extensive portion of the shelf ice has been demonstrated to be afloat and responsive to lake level fluctuations. The concept of drainage being triggered when the ice-barrier is temporarily lifted or disturbed (Marcus, 1960) seems to be the trigger mechanism with the most relevance to the McMaster Lake events.

The drainage of McMaster Lake is probably the result of a combination hypothesis. The role of the shelf ice in the drainage event can be hypothesized by looking at the cross-sectional view of the south end of McMaster Lake (Figure 6.6). The shelf ice mass in the southern end of the lake has been heavily structurally deformed, and is subject to high bending stresses in the hinge zone, set up by the fluctuating lake levels. The distorted shelf ice is structurally much weaker than a rigid glacier mass, and lines of weakness would exist preferentially where the most disturbance has occurred. The buoyancy forces resulting from a 5 m rise in lake level over a 40 day period will

place considerable tensile stress on the base of the ice shelf in the hinge zone. In addition, the hydrostatic pressure acting upon the underside of the ice shelf as it attempts to adjust to constantly rising lake levels may be a factor. At a critical water level, preferential deformation along a line of weakness in the hinge zone could create an opening, which could be enlarged slightly due to thermal or mechanical erosion by the escaping floodwaters. This hypothetical sequence of events is presented in Figure 7.3. Looking at the lake basin and the ice dam in cross-section, it can be seen that shelf ice configuration, lake level, and the condition of the outlet vary through the season and from year to year. This mechanism can explain the observed patterns of discharge.

These ideas seem plausible, given the observations from the outlet site. The upwelling (Figure 7.2B) appears to lie along the hinge line between the South Lobe and the shelf ice, as well as in an ice-rock wall marginal position. As indicated in Figure 7.3, there is evidence of a subglacial bedrock ridge or 'riegel', which would effectively prevent downglacier movement of water, and force an outlet to the surface. The discharge in Siphon Creek did not peak until 24 hours after the initial burst rising from 39 m^3/sec on August 13 to 45 m^3/sec on August 14, indicating some enlargement of the opening.

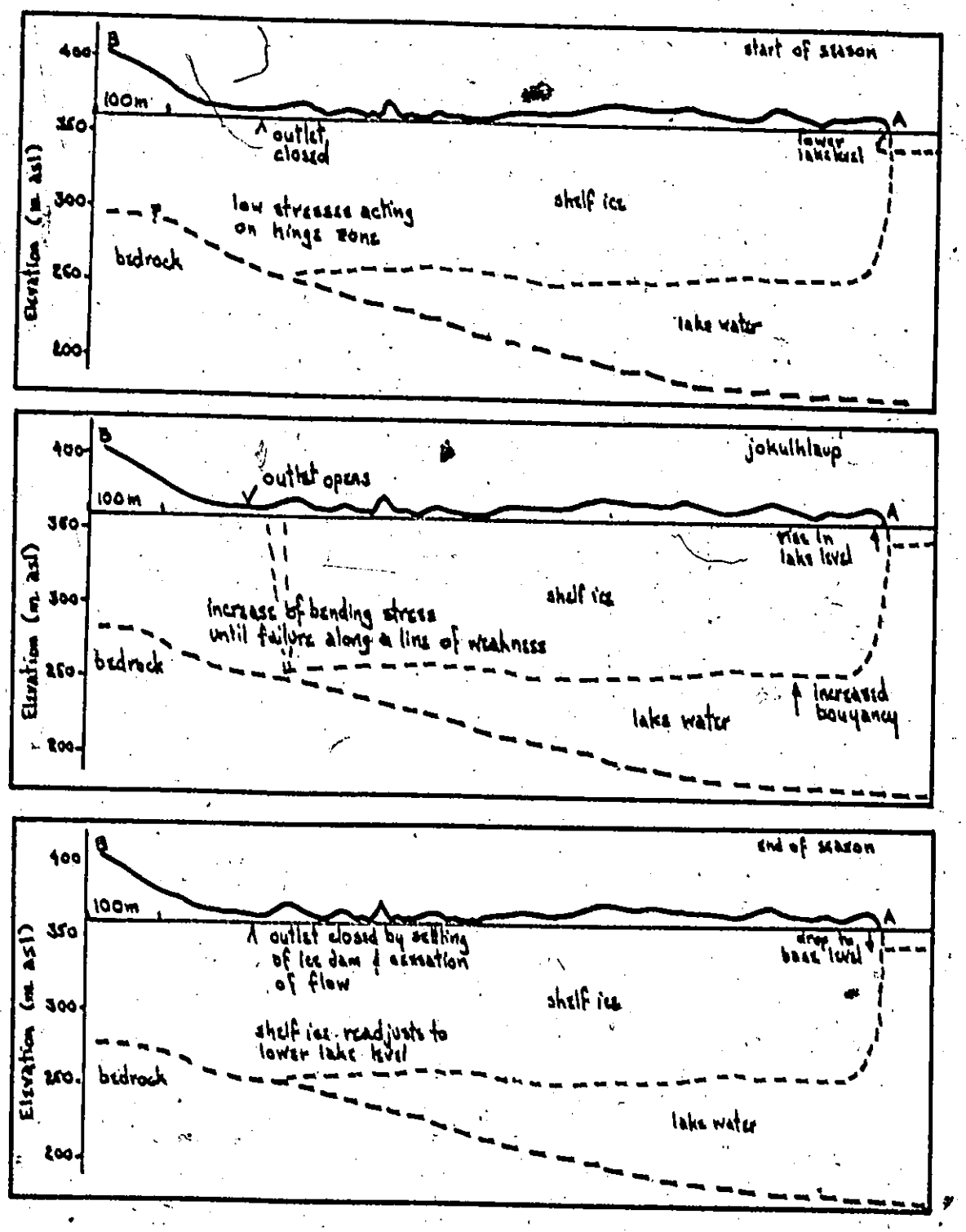


Figure 7.3 Proposed sequence of events in the drainage of McMaster Lake

The frequency and periodicity of drainage events in McMaster Lake might provide insight into the behaviour of the lake. The dates of initiation of drainage varied over a two week range, for the three drainage events observed (see Table 7.1). The timing of the jokulhlaupe is somewhat outside the normal pattern of release in late August or early September noted by other workers (Maag, 1969; Mathews 1973), but this observation can be accounted for when the shortness of the Ellesmere Island melt season is considered. Only one season's record of filling behaviour is available, but the filling rate of the lake and its influence on the shelf ice are thought to control the date of initiation of drainage. The rainstorm event of July 22, 1973 was probably a factor in increasing the filling rate and hastening the onset of lake drainage that year.

The three partial drainage events of McMaster Lake in three sucessive years would indicate that the frequency of these events is high, possibly annual. The summer climatic conditions of the three years showed significant variation, yet all three seasons met the appropriate circumstances required to initiate drainage. However, the differences in the physical characteristics of the three events, most notably in the maximum lake levels reached and at the upwelling site might indicate an irregular drainage frequency.

The maximum lake level reached dropped progressively from 1973 to 1975, by approximately 1.2 m each year. This indicates a slight but steady trend in drainage events. Evidence of the continued lowering of maximum water lines can be drawn from the strandlines on McMaster Lake (Figure 4.8). Apparently linked with the dropping maximum levels are the differences noted in the upwelling site (Chapter 7.2 and Figure 7.2). The extensive breakup and disturbance of the ice margin at the outlet in 1973 indicates that greater pressures were involved (and required) to force an opening through the shelf ice. After the initial violent outburst, and assuming that the winter did not completely destroy and close up the outlet, the 1974 and 1975 events appear to be merely reopening and slight modification of the original passageway. Lower amounts of buoyancy and hydrostatic head are required to trigger the later events, as evidenced by the lower maximum levels. The other factor which may contribute to the lowering maximum lake levels is the thinning of the ice shelf, requiring progressively less buoyancy to trigger a drainage event. Therefore, the pattern of drainage frequency, and to a certain extent, the magnitude of drainage events, is controlled by the structural configuration and activity of the shelf ice barrier.

From the evidence presented, there appear to be three possibilities for the drainage pattern of McMaster Lake:

1) an annual event, acting as a seasonal storage lag, and the normal regulation of one season's runoff;

2) an annual event for the past 11 years (as evidenced by the strandlines), but because of progressively lowering maximum levels this trend might be reaching an end; and

3) an annual event for 3 or 4 years as the lake level drops and the dam progressively closes, followed by a period of non-drainage until the requirements for triggering the event are again met.

The last two possibilities seem the most feasible for McMaster Lake, and there is evidence of a similar sporadic drainage frequency from Iceland (Thorarinsson, 1939b), Norway (Liestøl, 1955), Alaska (Stone, 1963b), and Greenland (Higgins, 1970). In these cases, the activity of the dam is the main factor in controlling drainage frequency. The situation may in fact be an evolving one where drainage patterns are constantly changing in the long term as a function of glacial and macroclimatic trends (Stone, 1963b; Mathews, 1964a). Because of the distinct differences in event magnitude between the three observed (consecutive) years, McMaster Lake does not seem conducive to regular annual drainage.

In conclusion, McMaster Lake drains upward through glacier shelf ice via a structurally formed outlet in an ice-marginal position. The presence and structure of the floating shelf ice and bedrock configuration has created this unusual example of englacial drainage in a high Arctic environment. Without the set of conditions which have generated this uncommon and delicately balanced situation, the lake would probably drain marginally or supraglacially along the McMaster Glacier snout into the valley below. Despite the unique factors involved, various elements of existing theories on drainage mechanisms can be combined to adequately explain the McMaster Lake events. It also points out the complexity of a form of glacial hydrologic system which is generally underestimated and misrepresented in the literature.

CHAPTER 8
CONCLUSIONS

Investigations of an ice-dammed lake in a remote and little known area of the Canadian high arctic have produced valuable data on sub-polar ice-dammed lakes. The principal findings of this research are:

1) McMaster Lake, which occupies a structurally controlled trough, is dammed by a lobe of the sub-polar Ellesmere Ice Cap. Bathymetry reveals a maximum depth of 200 m, the lake water is near isothermal at 0 - 0.5° C, and an 3 m thick ice pack covers this amictic lake year round. Rapid drainage of from 3 - 5 m of lake water was traced to an upwelling marginal pool at the juncture of the ice cap and the floating shelf ice.

2) Discharge in the outlet stream in 1974 was characterized by a lack of a distinct snowmelt flood, a diurnal glacier melt response for one month and a sudden two order of magnitude increase in flow during the jokulhlaup. 94% of the variance in Siphon Creek discharge from July 15 - August 9 is explained by the meteorological parameters and antecedent discharge. The concepts of 'normal' and 'abnormal' drainage of ice-dammed lakes are examined, and redefined in hydrologic terms as ice melt and jokulhlaup components of a process-response flow model.

3) One half of the surface area of McMaster Lake is occupied by a 150 m thick shelf of ice extending out from two lobes of the Ellesmere Ice Cap. This ice mass has been demonstrated to repond closely to lake level fluctuations, and is believed to be instrumental in lake drainage behavior.

4) Of the existing theories of lake drainage mechanisms, Thorarinsson's (1939b) concept of barrier flotation seems the most applicable, in light of the behavior of the shelf ice. Rising lake levels create stresses to be built up in the hinge zone between the glacier and the shelf, resulting in deformation along lines of weakness and the creation of an escape route for the lake water. Drainage continues until a rock sill at the head of the outlet channel is reached, or until the barrier resettles, cutting off lake drainage.

A considerable amount of information was collected on this one ice-dammed lake during a brief reconnaissance and a two month field season, while a three year downstream discharge record is available. Much speculation is based on this data, but few conclusive remarks can be made without further and more detailed observation. Such a study points out the need for continuing monitoring of lake systems and drainage events, most notably in the little known high Arctic.

Long term monitoring is the major necessity for future glacial hydrologic investigations, but numerous possibilities for the direction of research came to mind. To borrow an idea from the study of surging glaciers, instrumentation and detailed surveys of an ice-dammed lake and its ice dam prior to a jokulhlaup could provide information such as the detection of barrier displacement, and the sensing of water movements within the glacial hydrologic system. Compilation of more accurate data of this type would hopefully lead to more verifiable and comprehensive theories of lake drainage mechanisms. Statistical analysis of glaciohydrologic data incorporating some of the more advanced hydrologic models with variable lag times and storage terms could provide further insights to the nature of the glacier's internal plumbing system.

REFERENCES

Adams, W.P., 1966. Ablation and runoff on the White Glacier. Axel Heiberg Island Research Report, Glaciology #1, McGill University, Montreal, 77 pp.

Ahlmann, H.W., 1933. Scientific results of the Swedish-Norwegian Arctic Expedition in the summer of 1931, part 8. Geografiska Annaler, v.15, p.161-216, p.261-295.

Aitkenhead, N., 1960. Observations on the drainage of a glacier-dammed lake in Norway. Journal of Glaciology, v.3, p.607-609.

Aitkenhead, N., 1963. An ice caldera in North-East Grahamland. British Antarctic Survey Bulletin, v.1, p.9-15.

Andrews, R.H., 1964. Meteorology and heat balance of the ablation area, White Glacier. Axel Heiberg Island Research Report, Meteorology #1, McGill University, Montreal, 107pp.

Arnborg, L., 1955. Ice-marginal lakes at Hoffellsjokull. Geografiska Annaler, v.37, p.202-228.

Ankow, A.J., 1970. Derivation of formulae for variable lag time. Journal of Hydrology, v.10. p.225-242.

Aub, C.F., 1969. A supraglacial extension of an ice-dammed lake, Tunsbergdalsbreen, Norway: Comments on P.J. Howarth's paper. Journal of Glaciology, v.8, p.326-327.

Baird, P.D., 1952. The glaciological studies on the Baffin Island Expedition, 1950: Part I Method of Nourishment of the Barnes Ice Cap. Journal of Glaciology, v.2, #11, p.2-8.

Ballantyne, C.K., 1975. Geomorphological and Hydrological investigations in a high Arctic drainage basin. Unpublished M.Sc. Thesis, McMaster University, Hamilton.

Barnes, D.F., and Hobbie, J.E., 1960. Rate of melting the bottom of floating ice. United States Geological Survey, Professional Paper # 400B, p.392-394.

Barnett, D.M. and Holdsworth, G., 1974. Origin, morphology and chronology of sublacustrine moraines, Generator Lake, Baffin Island, NWT, Canada. Canadian Journal of Earth Sciences, v.11, #3, p.380-408.

Bateman, A.H., 1922. Kennecott Glacier of Alaska. Geological Society of America Bulletin, v.32, p.527-539.

Bentham, R., 1941. Structure and glaciers of southern Ellesmere Island. Geographical Journal, v.97, p.36-45.

Bezinge, A., Perreten, J.P. and Schafer, F., 1973. Phenomenon du lac glaciare du Gorner, in: Symposium on the Hydrology of Glaciers. International Union of Geodesy and Geophysics/International Association of Scientific Hydrology, Publication #95, p.65-78.

Bishop, B.C., 1957. Shear moraines in the Thule area, northwest Greenland. United States SIPRE Research Report #17.

Blachut, S.P. and Ballantyne, C.K., 1976. Ice-dammed lakes: A critical review of their nature and behaviour. McMaster University, Department of Geography, Discussion Paper (in preparation).

Boulton, G.S., 1970. On the origin and transport of englacial debris in Svarlbard glaciers. Journal of Glaciology, v.9, #56, p.213-230.

Brook, D., 1971. Scree benches around ice-dammed lakes in South Georgia. British Antarctic Survey Bulletin, v.26, p.31-40.

Budd W.F., 1966. The dynamics of the Amery Ice Shelf. Journal of Glaciology, v.5, #45, p.335-358.

Cadwell, D.H., 1974. Jokulhlaup lakes on existing glaciers as an aid to interpretation in Pleistocene stratigraphy, in: Quarternary Environments, Proceedings of a Symposium, W.C. Mahaney (ed.), Geog. Monographs #5, York University, Toronto, p.137-152.

Campbell, W.J. and Rasmussen, L.A., 1973. The production, flow and distribution of meltwater in a glacier

treated as a porous medium, in: Symposium on the Hydrology of Glaciers, 1969. International Union of Geodesy and Geophysics/International Association of Scientific Hydrology, Publication #95, p.11-27.

Carey, S.W. and Ahmad, N., 1961. Glacial marine sedimentation, in: Geology of the Arctic, Raasch, G.O.(ed.), v.2, p.865-894.

Charlesworth, J.K., 1957. The Quaternary Era, with special reference to its glaciation. E. Arnold Publishers, Ltd., London, 2 vols. 1700pp.

Christie, R.L., 1967. Reconnaissance of the surficial geology of northeastern Ellesmere Island, Arctic Archipelago. Geologic Survey of Canada Bulletin, # 138, 36pp.

Church, M., 1972. Baffin Island sandurs: A study of Arctic fluvial processes. Geological Survey of Canada Bulletin #216, 208pp.

Church, M.A., 1974. Hydrology and permafrost with reference to northern North America, in: Proceedings Workshop Seminar, Canadian National Committee, I.H.D., p.7-20.

Church, M. and Kellerhals, R., 1970. Stream gauging techniques for remote areas using portable equipment. Inland Waters Branch, Department of Energy, Mines, and Resources, Technical Bulletin #25, 90pp.

Clague, J.J. and Mathews, W.H., 1973. The magnitude of jokulhlaups. Journal of Glaciology, v.12, #66, p.501-504.

Cogley, J.G., 1971. Hydrological and Geomorphological Observations on a high latitude basin: Jason's Creek, Devon Island, Northwest Territories. Unpublished M.Sc. Thesis, McMaster University, Hamilton, 131pp.

Cogley, J.G., and McCann, S.B., 1971. Information on a snowmelt runoff system obtained from covariance and spectral analysis, abstract in: Transactions, American Geophysical Union, v.52, #4, p.198.

Cogley, J.G. and McCann, S.B., 1976. An exceptional storm and its effects in the Canadian Arctic. Arctic and Alpine Research, in press.

Colbeck, S.C. and Davidson, G., 1973. Water percolation through homogeneous snow, in: The role of snow and ice in hydrology, Proceedings Banff Symposium, International Association of Scientific Hydrology Publication #107, v.1, p.242-257.

Collet, L.W., 1932. Alpine Lakes. Scottish Geographic Magazine, v.38, p.73-101.

Craig, B.G., and Fyles, J.G., 1960. Pleistocene geology of Arctic Canada, in: Raasch, G.O. (ed.), Geology of the Arctic, v.1, University of Toronto Press, p.403-420.

Crary, A.P., 1958. Arctic ice island and ice shelf studies, Part I. Arctic, v.11, #1, p.3-42.

Crary, A.P., 1960. Arctic ice island and ice shelf studies, Part II. Arctic, v.13, #1, p.32-50.

Davis, J.R. and Nichols, R.L., 1968. The quantity of meltwater in the Marble Point- Gneiss Point area, McMurdo Sound, Antarctica. Journal of Glaciology, v.7, p.313-320.

Debenham, F., 1948. The problem of the great Ross Barrier. Geographical Journal, v.112, #4-6, p.196-218.

Debenham, F., 1965. The glacier tongues of McMurdo Sound. Geographical Journal, v.131, part 3, p.369-71.

de Boer, G., 1949. Ice margin features, Loirbreen, Norway. Journal of Glaciology, v.1, #6, p.332-336.

Demorest, M., 1943. Ice sheets. Geological Society of America Bulletin, v.54, p.363-400.

Dorikx, L., 1973. Glacier discharge simulation by groundwater analogue, in: Symposium on the Hydrology of Glaciers, 1969. International Union of Geodesy and Geophysics/International Association of Scientific Hydrology, Publication #95, p.29-40.

Dorikx, L. and Loijens, H., 1971. Model of runoff from glaciers, in: Runoff from Snow and Ice, National Research Council/Inland Waters Branch, Hydrology Symposium #8 v.1, p.153-199.

Dorrer, E., Movement of the Ward Hunt Ice Shelf, Ellesmere Island, Northwest Territories, Canada. Journal of Glaciology, v.10, #59, p.211-225.

Dunbar, M., and Greenaway, K.R., 1956. Arctic Canada from the Air. Queen's Printer, Ottawa, 541pp.

Dybeck, M.W., 1957. An ice margin lake in Iceland. Geographical Journal, v.123, part 1, p.127-130.

Ellington, G.R., 1973. Water movement throughout the Gornergletcher, in: Symposium on the hydrology of glaciers, 1969. International Union of Geodesy and Geophysics/International Association of Scientific Hydrology Publication #95, p.79-84.

Embleton, C. and King, C.A.M., 1968. Glacial and Periglacial Geomorphology. MacMillan of Canada, Toronto, 608pp.

Eriksson, B.F., 1942. Studies in Northeast Greenland. Geografiska Annaler, v.24, part 4, p.23-50.

Fabricius, O., 1788. Om Driv-Iisen i de Nordlige Vande og Fornemmelig i Davids-Straeder. K. Danske Vidensk. Selsk. Skrifter, v.3, p.65-84.

Fabricius, O., 1955. On the floating ice in the northern waters and particularly in the strait of Davis. Geographical Review, v.45, p.407-415. translation of the 1788 paper.

Fahnestock, R.K., 1963. Morphology of a glacial stream, White River, Mount Ranier, Washington. USGS Professional Paper #422A, 70pp.

Ferryman, M., 1895. The Daemme Vand, or Rembendale Glacier Lake, Norway. Geographical Journal, v.4, p.524.

Fisher, D., 1973. Subglacial leakage of Summit Lake, B.C., by dye determinations, in: Symposium on the Hydrology of Glaciers, 1969. International Union of Geodesy and Geophysics/International Association of Scientific Hydrology, Publication #95, p.111-116.

Flint, R.F., 1971. Glacial and Quaternary Geology. John Wiley and Sons, Inc., Toronto, 892pp.

Fohn, P.M.B., 1973. Short-term snow melt and ablation derived from heat and mass balance measurements. Journal of Glaciology, v.12, p.275-290.

Ford, W.L., and Hattersley-Smith, G., 1965. On the oceanography of the Nansen Sound Fiord system. Arctic, v.18, p.158-171.

Fortier, Y.O., 1957. The Arctic Archipelago: Geology and Economic Minerals of Canada, Economic Geology Series #1, Geological Survey of Canada, 4th edition, p.393-442.

Fortier, Y.O. et al., 1963. Geology of the North-Central Part of the Arctic Archipelago, Northwest Territories (Operation Franklin). Geological Survey of Canada Memoir #320, 671pp.

Gage, M., 1966. Franz Josef Glacier. Ice, v.20, p.26.

Gavrilova, M.K., 1966. Radiation climate of the Arctic: translated from Russian, Israel Program for Scientific Translations, Jerusalem.

Geiger, R., 1950. The climate near the ground (translated from German by Stewart, N. et al). Harvard University Press, Cambridge, Massachussetts, 482pp.

Gilbert, R., 1969. Some aspects of the hydrology of ice-dammed lakes: observations on Summit Lake, British Columbia. Unpublished M.A. Thesis, University of British Columbia.

Gilbert, R., 1971. Observations on Ice-dammed Summit Lake, British Columbia, Canada. Journal of Glaciology, v.10, #60, p.351-356.

Gilbert, R., 1972. Drainings of Ice-dammed Summit Lake, British Columbia. Inland Waters Directorate, Ottawa, Scientific Series #20, 17pp.

Glen, J.W., 1952. Experiments on the deformation of ice. Journal of Glaciology, v.2, #12, p.111-114.

Glen, J.W., 1954. The stability of ice-dammed lakes and other water filled holes in glaciers. Journal of Glaciology, v.2, #15, p.316-318.

Glen, J.W., 1956. Measurement of the deformation of ice in a tunnel at the foot of an ice fall. Journal of Glaciology, v.2, p.735-745.

Goldthwait, R.P., 1951. Development of end moraines in east-central Baffin Island, Journal of Glaciology, v.59, p.567-577.

Golubev, G.N., 1973. Analysis of the runoff and flow routing for a mountain glacier basin, in: Symposium on the Hydrology of Glaciers, 1969. International Union of Geodesy and Geophysics/International Association of Scientific Hydrology publ #95, p.41-50.

Goodison, B., 1972. An analysis of climate and runoff events for Peyto Glacier, Alberta. Inland Waters Directorate, Ottawa, Scientific Series #21, 29pp.

Gray, D.M.(ed.), 1970. Handbook on the principles of hydrology. Canadian National Committee for International Hydrological Decade, Ottawa.

Greely, A.W., 1886. Three Years of Arctic Service; an account of the Lady Franklin Bay Expedition of 1881-84, and the attainment of the farthest north. New York, C.Scribners and Sons.

Gudmundsson, G., and Sigbjarnson, G., 1972. Analysis of glacier runoff and meteorological observations. Journal of Glaciology, v.11, #63, p.303-318.

Gustavson, T.C., 1972. Sedimentation and physical limnology in proglacial Malaspina Lake, Alaska. University of Massachussetts Coastal Research Center, Technical Report #5, 48pp.

Haefeli, R., 1963. Note on the history of the Steingletscher Lake. International Association of Scientific Hydrology Bulletin, v.8, #2, p.123-125.

Haefeli, R., 1970. Changes in the behavior of the Unteraargletscher in the last 125 years. Journal of Glaciology, v.9, #56, p.195-212.

Haig-Thomas, D., 1940. Expedition to Ellesmere Island, 1937-38. Geographical Journal, v.95, p.265-291.

Hare, F.K., 1968. The Arctic. Quarterly Journal of the Royal Meteorological Society, v.94, #402, p.439-459.

Hare, F.K. and Hay, J.E., 1974. The climate of Canada and Alaska, in: Climates of North America (Bryson and Hare, ed.), Elsevier, p.49-192.

Hattersley-Smith, G., 1958. Canadian Operation "Hazen". Polar Record, v.9, #58, p.26-27.

Hattersley-Smith, G., 1960a. Studies of englacial profiles in the Lake Hazen area of northern Ellesmere Island. Journal of Glaciology, v.3, #27, p.610-625.

Hattersley-Smith, G., 1960b. Some remarks on glaciers and climate in northern Ellesmere Island. Geografiska Annaler, v.62, #1, p.45-48.

Hattersley-Smith, G., 1963a. The Ward Hunt Ice Shelf: recent changes in the ice front. Journal of Glaciology, v.4, #34, p.415-424.

Hattersley-Smith, G., 1963b. Climatic inferences from firn studies in northern Ellesmere Island. Geografiska Annaler, v.65, #2-3, p.139-151.

Hattersley-Smith, G., 1969a. Glacial features of Tanquary Fiord and adjoining areas of northern Ellesmere Island, Northwest Territories. Journal of Glaciology, v.8, #52, p.23-50.

Hattersley-Smith, G., 1969b. Recent observations on the surging Otto Glacier, Ellesmere Island. Canadian Journal of Earth Sciences, v.6, #4, part 2, p.883-889.

Hattersley-Smith, G. and Serson, H., 1964. Stratified water of a glacial lake in northern Ellesmere Island. Arctic, v.17, #2, p.108-111.

Hattersley-Smith, G. and Serson, H., 1966. Ice-dammed bodies of water in northern Ellesmere Island. Unpublished note, 6pp.

Hattersley-Smith, G., Keys, J.E., Serson, H., and Mielke, J.E., 1970. Density statified lakes in northern Ellesmere Island. Nature, v.225, #5227, p.55-56.

Hattersley-Smith, G., and Serson, H., 1970. Mass balance of the Ward Hunt ice rise and ice shelf: a 10 year record. Journal of Glaciology, v.9, #56, p.247-252.

Havens, J.M., 1964. Meteorology and heat balance of the accumulation area, McGill Ice Cap. Axel Heiberg Island Research Report, Meteorology #2, McGill university, Montreal, 87pp.

Heinsheimer, G.C., 1954. Zur Hydrologie und Glaziologie des Lago Argentino und Ventisquero Moreno, Argentinien. Zeitschrift fur Gletscherkunde und Glazialgeologie, Bund III, Heft 1, p.33-38.

Helbig, K., 1935. The origin of the Rio Plomo ice dam. Geograpical Journal, v.85, p.41-49.

Helk, J.V., 1966. Glacier mapping in Greenland. Canadian Journal of Earth Sciences, v.3, #6, p.771-774.

Henoch, W.E.S., 1971. Estimate of glaciers secular (1948-1966) volumetric change and its contribution to the discharge in the Upper Saskatchewan river basin. Journal of Hydrology, v.12, p.145-160.

Higgins, A.K., 1970. On some ice-dammed lakes in the Frederikshab District, South-west Greenland, Meddelser Dansk Geologisk Forening, v.19, #4, p.378-397.

Hodge, S., 1974. Variations in the sliding of a temperate glacier. Journal of Glaciology, v.13, #69, p.349-369.

Hodgson, D.A., 1973. Landscape and late-glacial history, head of Vendom Fiord, Ellesmere Island. Geological Survey of Canada Paper #73-1, part B, p.129-136.

Holdsworth, G., 1969. Flexure of a floating ice tongue. Journal of Glaciology, v.8, #54, p.385-397.

Holdsworth, G., 1971. Calving from Ward Hunt Ice Shelf 1961-1962. Canadian Journal of Earth Sciences, v.8, #2, p.299-305.

Holdsworth, G., 1973a. Ice deformation and moraine formation at the margin of and ice cap adjacent to a proglacial lake, in: Research in Polar and Alpine Geomorphology (Fahey and Thompson, ed.), 3rd Guelph Symposium in Geomorphology, p.187-199.

Holdsworth, G., 1973b. Ice calving into the proglacial Generator Lake, Baffin Island, Northwest Territories, Canada. Journal of Glaciology, v.12, #65, p235-250.

Holdsworth, G., 1974. Erebus Glacier Tongue, McMurdo Sound, Antarctica. Journal of Glaciology, v.13, #67, p.27-35.

Holdsworth, G. and McLaren, P., 1971. Proglacial lake studies, Barnes Ice Cap Baffin Island: progress report. Arctic Circular, v.21, #3, p.164-175.

Holmgren, B., 1971. Climate and energy exchange on a sub-polar ice cap in summer. Meteorologiska Intitutionen Uppsala Universitet, Meddelande #11.

Howarth, P.J., 1968a. Geomorphological and Glaciological studies, Eastern Breidamerkurjokull, Iceland. Unpublished Ph.D. Thesis, University of Glasgow.

Howarth, P.J., 1968b. A supraglacial extension of an ice-dammed lake, Tunsbergdalsbreen, Norway. Journal of Glaciology, v.7, #51, p.413-419.

Howarth, P.J. and Price, R.J., 1969. The proglacial lakes of Breidamerkurjokull and Fjallsjokull, Iceland. Geographical Journal, v.135, #4, p.573-581.

Humphreys, N., Shackleton, E., and Moore, A.W., 1936. Oxford University Ellesmere Land Expedition. Geographical Journal, v.87, #5, p.385-443.

Hutchinson, G.E., 1957. A Treatise on Limnology, Volume 1, Wiley and Sons, New York, 1015pp.

Inglefield, E.A., 1853. A summer search for Sir John Franklin; with a peep into the polar basin. T. Harrison, London, 232pp.

Jackson, C.I., 1961. Summer precipitation in the Queen Elizabeth Islands. Folia Geographica Danica, v.9, p.140-153.

Jackson, C.I., 1969. The summer climate of Tanquary Fiord, Northwest Territories. Published in Meteorology #95, Defence Research Board, Canada, 65pp.

Jonson, H., and Lang, H., 1973. Forcasting discharge from a glaciated basin in the Swiss Alps, in: The role of snow and ice in hydrology. Proceedings of the Banff Symposium, 1972, International Association of Scientific Hydrology Publication #107, v.2, p.1047-1051.

Jönsson, J., 1955. On the formation of frontal glacial lakes. Geografiska Annaler, v.37, #3-4, p.229-233.

Kasser, P., 1967. Fluctuations of glaciers 1959-1965. International Committee on Snow and Ice, UNESCO/International Association of Scientific Hydrology, 52pp.

Katz, H.R., 1953. Journey across the nunataks of central East Greenland, 1951. Arctic, v.6, #1, p.3-14.

Keeler, C.M., 1964. Relationship between climate, ablation, and runoff on the Sverdrup Glacier, Devon Island, Northwest Territories, 1963. Arctic Institute of North America, Research Paper #27, Montreal, 127pp.

Kelly, A.G., 1966. The hydraulic design of syphons. Institute of Mechanical Engineering Proceedings, 1965-1966, v.180, part 1, #42, p.981-997.

Kerr, F.A., 1934. The ice dam and floods of Talsekwa, British Columbia. Geographical Review, v.24, p.643-645.

Kerr, J.W., 1967a. Vendom Fiord Formation - a new redbed unit of probable early Middle Devonian (Eifelian) age, Ellesmere Island, Canada. Geological Survey of Canada Paper #67-43, 8pp.

Kerr, J.W., 1967b. Stratigraphy of central and eastern Ellesmere Island, Canada. Geological Survey of Canada Paper #67-27, 155pp.

King, W.D.V.O., 1934. The Mendoza River Flood of 10-11 January 1934. Geographical Journal, v.84, #4, p.321-326.

Kjartansson, G., 1967. Steinsholtshlapid 15 januar 1967. Naturufraedingurinn, Arg. 37, Ht. 3-4, p.120-169.

Klebelsberg, R., 1948. Handbuch der Gletscherkunde und Glazialgeologie. V.1, Springer Verlag, Vienna, 403pp.

Koch, L., 1928. Contribution to the glaciology of North Greenland. Meddelelser om Grønland, Bd. 65, #2.

Koerner, R.M., 1965. An ice caldera near Hope Bay, Trinity Peninsula, Graham Land. British Antarctic Survey Bulletin, v.1, p.37-39.

Koerner, R.M., 1970. The mass balance of the Devon Island Ice Cap, NWT, Canada, 1961-66. Journal of Glaciology, v.9, #57, p.325-336.

Koerner, R.M. et al, 1963. The Devon Island Expedition, 1960-64. Arctic, v.16, #1, p.57-76.

Krimmel, R.M., Tangborn, W.V. and Meier, M.F., 1973. Waterflow through a temperate glacier, in: The role of snow and ice in hydrology, Proceedings of the Banff Symposium 1972, International Association of Scientific Hydrology, Publication #107, v.1, p.401-416.

LaChapelle, E., 1959. Annual mass and energy exchange on the Blue Glacier. Journal of Geophysical Research, v.64, #4, p.443-449.

Lagally, M., 1932. Zur Thermodynamik der Gletscher. Zeitschrift der Gletscherkunde, v.20, p.199-214.

Lang,H., 1968. Relations between glacier runoff and meteorological factors observed on and outside the glacier. International Association of Hydrology, Berne, Publication #79, p.429-439.

Lang, H., 1973. Variations in the relation between glacier discharge and meteorological elements, in: Symposium on the Hydrology of Glaciers, 1969. International Union of Geodesy and Geophysics/International Association of Scientific Hydrology, Publication #95, p.85-94.

Laurenson, E.M., 1964. A catchment storage model for runoff routing. Journal of Hydrology , v.2, p.141-163.

Liestøl, O., 1955. Glacier dammed lakes in Norway. Norsk Geografisk Tidsskrift, Bd.15, Ht. 3-4, p.122-149.

Lindsay, J.F., 1966. Observations on the level of a self draining lake on the Casement Glacier, Alaska. Journal of Glaciology, v.6, #45, p.443-445.

Lliboutry, L.A., 1971. The glacier theory. Advances in Hydroscience, v.7, p.81-167.

Lutschg, O., 1915. Der Marjelensee und seine Abflussverhältnisse. Annalen der Schweiz Landshydrographie, Band I, Bern.

Maag, H.U., 1963. Marginal drainage and ice-dammed lakes, in: Muller, F., et al, Axel Heiberg Island Research Report, Preliminary Report 1961-1962, McGill University, Montreal, p.151-160.

Maag, H.U., 1969. Ice dammed lakes and marginal glacial drainage on Axel Heiberg Island. Axel Heiberg

Island Research Report, McGill University, Montreal, 147pp.

MacMillan, D.B., 1925. Four Years in the White North, 2nd ed. The Medici Society of America Publication, Boston and New York, 428pp.

Malde, H.E., 1968. The catastrophic late Pleistocene flood in the Snake River Plain, Idaho. Geological Society of America Professional Paper #596, 52pp.

Male, D.H. and Norum, D.I., 1971. Movement of water through snow and ice, in: Runoff from Snow and Ice. National Research Council/Inland Waters Branch, Hydrology Symposium #8, v.1, p.83-107.

Mandeville, A.N. and O'Donnell, T., 1973. Introduction of time variance to linear conceptual catchment models. Water Resources Research, v.9, #2, p.298-310.

Marcus, M.G., 1960. Periodic drainage of glacier-dammed Tulsequah Lake, British Columbia. Geographical Review, v.50, #1, p.89-106.

Marcus, M.G., 1968. Effects on glacier dammed lakes in the Chugach and Kenai Mountains in the Great Alaska Earthquake of 1964. National Academy of Sciences, Hydrology, Publication #1603, Volume 3: Hydrology, p.329-347.

Martinec, J., 1970. Recession coefficient in glacier runoff studies. International Association of Scientific Hydrology Bulletin, v.15, #1, p.87-90.

Mason, K., 1935. The study of retreating glaciers. Geographical Journal, v.85, p.24-41.

Matalas, N.C., 1966. Some aspects of time series analysis in hydrologic studies, in: Statistical Methods in Hydrology National Research Council/Inland Waters Branch, Hydrology Symposium #5, p.271-313.

Mathews, W.H., 1963. Discharge of a glacial stream. Berkeley Symposium, 1963, International Union of Geodesy and Geophysics/International Association of Scientific Hydrology, v.63, p.290-300.

Mathews, W.H., 1964. Water pressure under a glacier. Journal of Glaciology, v.5, #38, p.235-240.

Mathews, W.H., 1965. Two self-dumping ice-dammed lakes in British Columbia. Geographical Review, v.55, #1, p.46-52.

Mathews, W.H., 1970. The hydrology of glaciers, in: Glaciers, Proceedings Workshop Seminar, Canadian National Committee for I.H.D., Vancouver, p.31-32.

Mathews, W.H., 1973. Record of two jokulhlaupe, in: Symposium on the Hydrology of Glaciers, 1969. International Union of Geodesy and Geophysics/International Association of Scientific Hydrology, Publication #95, p.99-110.

McCann, S.B., James, W., Cogley, J.G. and Taylor, R.B., 1972. A Hydrological and Coastal Reconnaissance of South-Central Ellesmere Island. Contract Report to the Glaciology Division, Department of the Environment, Canada, 125pp.

McCann, S.B., Cogley, J.G., Woo, M.K. and Blachut, S.P., 1974. Hydrological Studies - South Central Ellesmere Island. Contract Report to the Glaciology Division, Department of the Environment, Canada, 149pp.

McCann, S.B., Cogley, J.G., Blachut, S.P., Ballantyne, C.K. and Bennett, B.G., 1975. Hydrological Studies, South Central Ellesmere Island. Contract Report to the Glaciology Division, Department of the Environment, Canada, 283pp.

Megaham, W.F., Meiman, J.R., and Goodell, B.C., 1967. Net all wave radiation as an index of natural snowmelt and snowmelt accelerated by albedo reducing materials, Proceedings of the International Hydrology Symposium, Fort Collins, Colorado, p.149-156.

Meteorological Branch, Department of Transport, 1970. Climate of the Canadian Arctic. Canadian Hydrography Services; Marine Sciences Branch, Department of Energy, Mines and Resources, Ottawa, 71pp.

Miller, M.M., 1975. Thermo-physical characteristics of glaciers - towards a rational classification. Journal of Glaciology, (in press), Proceedings of the International Symposium on the Thermal Regime of Glaciers and Ice Sheets, Vancouver 1975.

Moravek, J.R., 1968. Some geographical aspects of ice-dammed self-draining lakes: a case study of Casement Lake, Glacier Bay, Alaska. Unpublished M.Sc. Thesis, University of Tennessee, 84pp.

Moravek, J.R., 1973. Some further observations on the behaviour of an ice-dammed, self-draining lake, Glacier Bay, Alaska, USA. Journal of Glaciology, v.12, #66, p.505-507.

Morrison, C.C., 1958. Glaciers and Human Activities (Chapter 1), in: Geographic Study of Mountain Glaciation in the Northern Hemispere, C.C.Morrison (ed.), American Geographical Society, Part 9, p.1-27.

Muller, B.S.(ed.), 1961. Axel Heiberg Island, Preliminary Report, 1959-1960. McGill University, Montreal, 219pp.

Muller, F., 1962. Zonation in the accumulation area of the glaciers of Axel Heiberg Island, Northwest Territories, Canada. Journal of Glaciology, v.4, #33, p.302-318.

Muller, F. et al., 1963. Axel Heiberg Island Research Report, Preliminary Report 1961-1962, McGill University, Montreal.

Muller, F. and Keeler, C.M., 1969. Errors in short-term ablation measurements on melting ice surfaces. Journal of Glaciology, v.8, #52, p.91-104.

Muller, F., and Roskin-Sharlin, N., 1967. A high arctic climate study on Axel Heiberg Island, Canadian Arctic Archipelago, summer, 1961. Axel Heiberg Island Research Reports, McGill University, Montreal, Meteorology #3, 82pp.

Munro, R., 1892. On a remarkable glacier-lake, formed by a branch of the Hardanger-Jokul, near Eidfiord, Norway. Proceedings of the Royal Society of Edinburg, v.20, p.53-62.

Nichols, R.L. and Miller, M.M., 1952. The Moreno Glacier, Lago Argentino, Patagonia Advancing Glaciers and Nearly Simultaneously Retreating Glaciers. Journal of Glaciology, v.2, p.41-50.

Nye, J.F., 1953. The flow law of ice from measurements in glacier tunnels, labratory experiments and the Jungfraufirn borehole experiment. Proceedings of the

Royal Society of London, Series A, v.219, #1139, p.477-489.

Nye, J.F. and Frank, F.C., 1973. Hydrology of the intergranular veins in a temperate glacier, Symposium on the Hydrology of Glaciers, 1969. International Union of Geodesy and Geophysics/International Association of Scientific Hydrology, Publication #95, p.157-161.

Nye, J.F. and Mae, S., 1972.The effect of non-hydrostatic stress on intergranular water veins and lenses in ice. Journal of Glaciology, v.11, #61, p.81-101.

Ommanney, C.S.L., 1969. A study in glacier inventory - The ice masses of Axel Heiberg Island, Canadian Arctic Archipelago. Axel Heiberg Island Research Reports, McGill University, Montreal, Glaciology #3, 105pp.

Ommanney, C.S.L., Clarkson, J. and Strome, M.M., 1973. Information Booklet for the Inventory of Canadian Glaciers. Glacier Inventory Note #4, Inland Waters Directorate, Department of the Environment, Ottawa.

Orvig, S., 1953. The glaciological studies of the Baffin Island Expedition, 1950: Part V On the Variation of the shear stress on the bed of an ice cap. Journal of Glaciology, v.2, #14, p.242-246.

Oswald, G.K.A., and Robin, G.Q., 1973. Lakes beneath the Antarctic ice sheet. Nature, v.245, #5423, p.251-254.

Østrem, G., 1973. Runoff forecasts for highly glacierized basins, in: The role of snow and ice in hydrology, Proceedings of the Banff Symposium 1972, International Association of Scientific Hydrology, Publication #107, v.2, p.1111-1132.

Paige, R.A., 1968. Sub-surface melt pools in the McMurdo Ice Shelf, Antarctica. Journal of Glaciology, v.7, #51, p.511-516.

Palsson, S., 1882. Forsog Til en Physisk Geografisk og Historiske Beskrivelse over de islandske Is-bioerge, I Anledning af en Reise til de fornemste deraf i Aarene 1792 til 1794. Den Norske Turisforenings Aarborg, Kristiania, p.37.

Paterson, W.S.B., 1969. The Physics of Glaciers. Pergamon Press, Toronto, 250pp.

Price, R.J., 1973. Glacial and Fluvioglacial Landforms. Oliver and Boyd, Edinburgh, 242pp.

Post, A., 1967. Effects of the March 1964 Alaska earthquake on glaciers. United States Geological Survey, Professional Paper #544-D, p.1-42.

Rabot, C., 1905. Glacial reservoirs and their outbursts. Geographical Journal, v.25, p.534-547.

Reed, D.W., Johnson, P., and Firth, J.M., 1975. A non-linear rainfall-runoff model, providing for variable lag time. Journal of Hydrology, v.25, p.295-305.

Reeh, N., 1968. On the calving of ice from floating glaciers and ice shelves, Journal of Glaciology, v.7, #50, p.215-232.

Reid, J.R. and Clayton, L., 1963. Observations of rapid water-level fluctuations in ice sink-hole lakes, Martin River Glacier, Alaska. Journal of Glaciology, v.4, p.650-652.

Reynolds, J.H., 1930. Burst of a glacial dam in Iceland. Geographical Journal, v.79, p.241.

Richardson, D., 1968. Glacier outburst floods in the Pacific Northwest. United States Geological Survey Paper 600-D, p.D79-86.

Ricker, K., 1962. Polar ice-dammed lakes. The Canadian Alpine Journal, v.45, p.149-151.

Rink, H., 1862. Om vandets afløb fra det indre af Grønland ved kilder under isen. Naturhistorisk Tidsskrift, v.3, #1, 17pp.

Røen, U., 1962. Studies on freshwater entomostraca in Greenland II, Meddelser om Grønland, Bd. 170, #2.

Rothlisberger, H., 1955. Studies in glacier physics on the Penny Ice Cap, Baffin Island, 1953: Part III Seismic sounding. Journal of Glaciology, v.2, #18, p.539-553.

Rothlisberger, H., 1972. Water pressure in intra-glacial and sub-glacial channels. Journal of Glaciology, v.11, #62, p.177-203.

Russell, I.C., 1893. Malaspina glacier. Journal of Geology, v.1, p.219-245.

Sagar, R.B., 1966. Glaciological and climatological studies, Barnes Ice Cap, Geographical Bulletin, v.8, p.3-47.

Sater, J.E., 1963. The Arctic Basin. Arctic Institute of North America, Montreal, 319pp.

Schytt, V., 1956. Lateral drainage channels along the northern side of Moltka Glacier, northwest Greenland. Geografiska Annaler, v.38, p.64-77.

Schytt, V., 1974. Inland ice sheets - recent and Pleistocene. Geologiska Foreningens i Stockholm Forhhandlingar, v.96, p.299-309.

Seppala, M., 1973. On the formation of small marginal lakes on the Juneau Icefield, S.E. Alaska, USA. Journal of Glaciology, v.12, #65, p.267-273.

Shackleton, E.(ed.), 1937. Arctic Journeys: The Story of the Oxford University Ellesmere Land Expedition 1934-1935. Holder and Stoughton, London, 372pp.

Sharp, R.P., 1947. The Wolf Creek Glaciers, St. Elias Range, Yukon Territory, Geographical Review, v.37, p.26-52.

Sharp, R.P., 1956. Glaciers in the Arctic, Arctic, v.9, #1-2, p.78-117.

Shreve, R.L., 1972. Movement of water in glaciers, Journal of Glaciology, v.11, #62, p.205-214.

Smith, J., 1960. Glacier problems in South Georgia. Journal of Glaciology, v.3, #28, p.707-714.

Solomon, S., 1966. Statistical association between hydrologic variables, in: Statistical Methods in Hydrology, National Research Council/Inland Waters Branch, Hydrology Symposium #5, p.55-125.

Stenborg, T., 1965. Problems concerning winter runoff from glaciers. Geografiska Annaler, v.47A, #3, p.141-184.

Stenborg, T., 1968. Glacier drainage connected with ice structures. Geografiska Annaler, v.50, p.25-53.

Stenborg, T., 1969. Studies of the internal drainage of glaciers. Geografiska Annaler, v.51, p.13-41.

Stenborg, T., 1970. Delay of runoff from a glacier basin. Geografiska Annaler, v.52A, #1, p.1-30.

Stenborg, T., 1973. Some viewpoints on the internal drainage of glaciers, in: Symposium on the Hydrology of Glaciers, 1969. International Union of Geodesy and Geophysics/International Association of Scientific Hydrology, Publication #95, p.117-129.

Stone, K.H., 1963a. The annual emptying of Lake George, Alaska. Arctic, v.16, #1, p.26-40.

Stone, K.H., 1963b. Alaskan ice-dammed lakes. Annals of the American Association of Geographers, v.53, #3, p.332-349.

Stone, P., 1975. Observations on an ice-dammed lake, Hindle Glacier, South Georgia. British Antarctic Survey Bulletin, v.40, p.69-73.

Streten, N.A. and Wendler, G., 1968. The midsummer heat balance of an Alaskan maritime glacier. Journal of Glaciology, v.7, #51, p.431-440.

Strom, K.M., 1938. The catastrophic emptying of a glacier-dammed lake in Norway, 1937. Geologie der Meere und Binnengewasser, Bd. 2, p.443-444.

Sugawara, M., 1967. Runoff analysis and water balance analysis by a series storage type model, in: Proceedings International Hydrology Symposium, Colorado, v.1, p.31-37.

Sverdrup, H.U., 1936. Scientific results of the Norwegian-Swedish Spitzbergen expedition in 1934. IV Results of the meteorological observations on Isachsen's Plateau. Geografiska Annaler, v.18, p.34-47.

Sverdrup, O., 1904. New Land - Four years in the Arctic Regions. 2 Volumes, (translated by E.H. Hearn), Longmans, Green and Company, London.

Taylor, A., 1956. Physical geography of the Queen Elizabeth Islands, Canada. American Geographical Society Contract, New York, 12 volumes.

Thomas, M.K., 1961. A survey of temperatures in the Canadian Arctic, in: Geology of the Arctic, Volume II, (G.O. Raasch, ed.), University of Toronto, Toronto, p.942-955.

Thomas, R.H., 1974a. The creep of ice shelves: theory. Journal of Glaciology, v.12, #64, p.45-53.

Thomas, R.H., 1974b, The creep of ice shelves: interpretation of observed behaviour. Journal of Glaciology, v.12, #64, p.55-70.

Thorarinsson, S., 1939a. Hoffellsjokull, its movements and drainage. Geografiska Annaler, v.21, p.189-215.

Thorarinsson, S., 1939b. The ice-dammed lakes of Iceland with particular reference to their values as indicators of glacier oscillations. Geografiska Annaler, v.21, #3-4, p.216-242.

Thorarinsson, S., 1953. Some new aspects of the Grimsvotn Problem. Journal of Glaciology, v.12, p.267-274.

Thorarinsson, S., 1957. The jokulhlaupe from the Katla area in 1955 compared with other jokulhlaupe in Iceland. Jokull, v.7, p.21-25.

Thorsteinsson, R., 1972. Geology, Strathcona Fiord, District of Franklin. Geological Survey of Canada Map #1307A.

Thorsteinsson, R. and Tozer, E.T., 1970. Geology of the Arctic Archipelago, in: Geology and Economic Minerals of Canada, (R.J.W. Douglas, ed.), Economic Geology Report #1, Geological Survey of Canada, Department of Energy, Mines, and Resources, 5th edition, p.547-590.

Tryggvason, E., 1960. Earthquakes, jokulhlaupe, and subglacial eruptions. Jokull, v.10, p.18-22.

Tywoniuk, N. and Wiebe, K., 1970. Application of Regression Analysis in Hydrology. Inland Waters Branch, Department of Energy, Mines, and Resources, Technical Bulletin #24, 18pp.

Vowinckel, E. and Orvig, S., 1970. The climate of the North Polar Basin, in: Climates of the Polar Regions, (S. Orvig, ed.). World Survey of Climatology, v.14, Elsevier Publishing Company, Amsterdam, p.129-252.

Vivian, R.A., 1965. Glaces mortes et morphologie glaciaire. Revue de Geographie Alpine, Tome LIII, Fasc. 3. p.371-401.

Vivian, R.A., and Zumstein, J., 1973. Hydrologie sous-glaciaire au glacier d'Argentiere (Mount-Blanc, France), in: Symposium on the Hydrology of Glaciers, 1969. International Union of Geodesy and

Geophysics/International Association of Scientific Hydrology, Publication #95, p. 53-64.

Wallen, C.C., 1948. Glacial-meteorological investigations on the Karsa Glacier in Swedish Lapland. Geografiska Annaler, v.30, p.451.

Ward, W.H., 1952a. The glaciological studies of the Baffin Island Expedition, 1950: Part II The physics of deglaciation of Central Baffin Island. Journal of Glaciology, v.2 #11, p.9-23.

Ward, W.H., 1952b. The glaciological studies of the Baffin Island Expedition, 1950: Part III Equipment and techniques. Journal of Glaciology, v.2, #12, p.115-121.

Ward, W.H., 1953. The glaciological studies of the Baffin Island Expedition, 1950: Part IV The exchange at the surface of the Barnes Ice Cap during the ablation period. Journal of Glaciology, v.2, #13, p.158-167.

Ward, W.H., 1954. Studies in glacier physics on the Penny Ice Cap, Baffin Island, 1953: Part II Portable ice-boring equipment, Journal of Glaciology, v.2, #16, p.433-435.

Ward, W.H., 1955. Studies in glacier physics on the Penny Ice Cap, Baffin Island: Part IV The flow of Highway Glacier. Journal of Glaciology, v.2, #18, p.592-600.

Ward, W.H., and Baird, P.D., 1954. Studies on the Penny Ice Cap, Baffin Island, 1953: Part I A description of the Penny Ice Cap, its accumulation and ablation. Journal of Glaciology, v.2, #15, p.342-355.

Weeks, W.F. and Assur, A., 1972. Fracture of lake and sea ice, in: Fracture: an advanced treatise, Volume 7, (H. Liebowitz, ed.), Academic Press, New York, p.879-978.

Weertman, J., 1957. Deformation of floating ice shelves. Journal of Glaciology, v.3, #21, p.38-42.

Weertman, J., 1961. Mechanism for the formation of inner moraines found near the edge of cold ice caps and ice sheets. Journal of Glaciology, v.3, p.965-978.

Weertman, J., 1964. The theory of glacier sliding. Journal of Glaciology, v.5, #39. p.287-303.

Weertman, J., 1974. Stability of the junction of an ice sheet and an ice shelf. Journal of Glaciology, v.13, #67, p.3-11.

Weidick, A., 1963. Ice margin features in the Julianehab District, South Greenland, Meddelelser om Grønland, Bd. 165, #3.

Wexler, H., 1960. Heating and melting of floating ice shelves. Journal of Glaciology, v.3, #27, p.626-645.

Whalley, W.B., 1971. Observations on the drainage of an ice-dammed lake Strupvatnet, Troms, Norway. Norsk Geografisk Tidsskrift, Bd. 25, Ht. 3-4, p.165-174.

Whalley, W.B., 1973. A note on the fluctuations of the level and size of Strupvatnet, Lyngen, Troms and the interpretation of ice loss on Strupbreen. Norsk Geografisk Tidsskrift, Bd. 27, Ht. 1, p.39-45.

Whillans, I.M., 1975. Effect of inversion winds on topographic detail and mass balance of inland ice sheets. Journal of Glaciology, v.14, #70, p.85-90.

Wilkins, S.P. (Blachut), 1973. Some glaciological problems involved in glacier inventory: A case study; D'Iberville Fiord, Ellesmere Island, Northwest Territories. Unpublished B.A. Thesis, McMaster University, 128pp.

Wilson, A.T. and Wellman, H.W. 1962. Lake Vanda: an Antarctic Lake. Nature, v.4860, p.1171-1173.

Wilson, H.P., 1961. The major factors of arctic climate, in: Geology of the Arctic, Volume II, (G.O. Raasch, ed.), University of Toronto Press, Toronto, p.915-930.

Wolman, M.G. and Miller, J.P., 1960. Magnitude and frequency of forces in geomorphic processes. Journal of Geology, v.68, p.54-74.

Woo, M.K., 1975. Active Layer Hydrology, Section B: Hydrological Studies, South Central Ellesmere Island. Contract report to the Glaciology Division, Department of the Environment, Canada, 103pp.

Wright, J., 1935. The Hagvatn Gorge. Geographical Journal, v.86, #3, p.218-234.