ISBN 0-315-53258-0

AN ICE-DAMMED LAKE IN THE ST.ELIAS RANGE SOUTHWESTERN YUKON TERRITORY

WATER BALANCE, PHYSICAL LIMNOLOGY ICE DYNAMICS AND SEDIMENTARY PROCESSES

BY

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A Thesis submitted to the School of Graduate Studies in partial fulfillment of the requirements for the degree of Master of Arts in Geography

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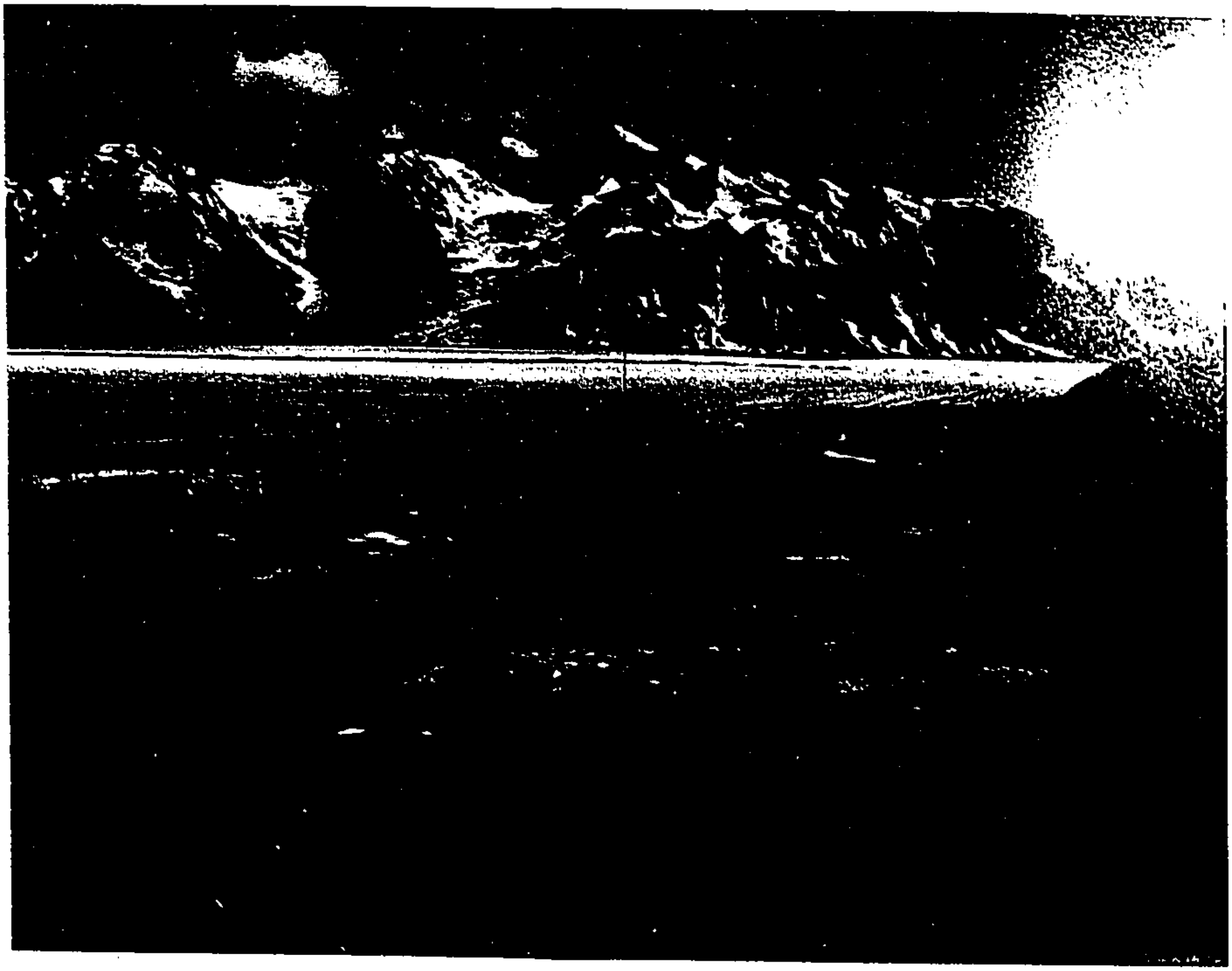

FRONTISPIECE: Overview of the study site, July 1988. Note the ice margin, the stream deltas, and the ice-cored moraine area. See the camp located at the lower right for scale.

ABSTRACT

The formation and drainage of glacier-dammed lakes are a very dramatic element of the hydrological regime of a glacierized region. The objective of this research is to contribute to the understanding of the hydrological and sedimentological processes operating in a glacier-dammed lake in the St. Elias Range, Yukon Territory.

The thesis addresses three major components of ice dammed lake research, the water balance of the lake, some aspects of glacier ice dynamics, and sedimentary sequences formed under conditions of changing water levels.

A systems approach was utilized to study input, storage and output using standard hydrological and sedimentological techniques. Ice dynamics were monitored by survey techniques and by photographic records.

The results indicate that the lake is not completely sealed, but leaks continuously throughout the season. The rates of leakage increase from 0.50 m^3sec^{-1} to 8.61 m^3sec^{-1} during the summer. Drainage is initiated by connection of the lake to the internal drainage network of the Kaskawulsh Glacier, and not by floating of the ice margin. Floating increases both the storage capacity and the leakage rate of the basin but does not cause the onset of drainage. The length of time required to drain the lake completely was 7 days, during which time the rate of drainage increased at an exponential rate, to a maximum value of 78.89 m^3sec^{-1}. However, the time frame in which drainage occurs is known to vary from early in August until mid-September.

The study of the ice margin provided information on the vertical movement during both the filling and draining stages. The ice was found to have lifted during the preceding fall season as well as during the summer season. The amount of lift of the ice margin decreased with increasing distance away from the ice face. Radio-echo sounding quantified the assumed

shape and thickness of the ice margin, confirming that it is a re-entrant of the Kaskawulsh Glacier intruding into a hanging valley.

The analysis of the sedimentary processes operating within a basin and the resulting deposits provided information on the changes caused by fluctuating water levels and annual drainage. The most noticeable factors are that the deltas are constructed in a step-wise fashion moving upstream, with each set in part overlying the previous one, and that the date of drainage affects the amount of erosion and redeposition which occurs. If drainage occurs while discharge is still high, then a reverse set of deposits may be established. Finally, deep water areas are sediment starved during part of the season, since fall drainage removes all fine material, and annual rhythmite deposits therefore do not form. The only material deposited is from summer density-induced underflows.

RÉSUMÉ

La formation et le drainage de lacs de barrage glaciaire sont des éléments spectaculaires du régime hydrologique d'une région glaciaire. L'objectif de cette recherche est de contribuer à la connaissance des processus hydriques et sédimentaires d' un lac de barrage glaciaire situé dans la chaîne des monts St. Elie.

La these comprend trois parties majeures: le bilan hydrique du lac, quelques aspects de la dynamique du glacier, et l'analyse des dépôts mis en place sous des conditions de niveau d'eau variable.

Une approche d'étude de systèmes a été utilisé pour traiter le régime des gains et des pertes en eau et la capacité de retenue du réservoir naturel. Les techniques classiques d'étude hydrique et sédimentologique ont été utilisées. Les mouvements verticaux de la glace ont été surveillés en utilisant des techniques de levés et l'étude de photographies.

Les résultats indiquent que le lac n'est pas complètement scellé, mais s'écoule continuellement. Les vitesses d'écoulement augmentent de 0,50 m^3sec^{-1} à 8,61 m^3sec^{-1} au cours de l'été. Le drainage est initié par la connection du lac avec le réseau interne du glacier Kaskawulsh. Le flottement de la marge glaciaire augmente la capacité de retenue du bassin et la vitesse d'écoulement des eaux mais n'amorce pas le drainage. Le temps nécessaire pour la vidange complète du lac fut de 7 jours, durant lesquels la vitesse du drainage augmenta de façon exponentielle jusqu'à une valeur maximale de 78,89 m^3sec^{-1}. Ce drainage peut survenir du début du mois d'août jusqu'à la fin de septembre.

L'étude de la marge glaciaire a fourni de l'information sur le mouvement vertical du glacier en fonction du changement du niveau d'eau du lac durant la saison estivale. La marge a connu une élévation durant l'automne précédant, ainsi que durant l'été même. La valeur du rehaussement de la marge diminue en s'éloignant du lac. L'utilisation d'un radar a permis de

connaître la forme et l'épaisseur de la glace, confirmant que c'est un lobe du glacier Kaskawulsh dans un bassin d'appoint.

L'analyse des processus sédimentaires et des dépôts qui en résultent a fourni des connaissances sur les changements provoqués par un niveau d'eau variable et par le drainage annuel. Les aspects les plus remarquables sont que l'ensemble des deltas forment des paliers à altitude croissantes construits successivement vers l'amont qui se recouvrent partiellement, et que la date du drainage affecte l'intensité de l'érosion et de re-déposition qui se produit. Si le drainage a lieu quand le débit est encore très haut, une série de dépôts inversés peut se produire. Les sections d'eau profonde sont privées de sédiments pour une partie de la saison, parce que le drainage en automne enlève tout les matériaux fins, et les dépôts rythmiques ne se forment pas. Le seul matériel qui est mis en place provient des courants denses estivaux.

ACKNOWLEDGEMENTS

I would like to extend my sincere thanks to my supervisor, mentor and friend, Peter Johnson, who believed in me and helped me through this research project. His boundless enthusiasm and energy were an inspiration to me. His unfailing willingness to provide a sounding board for ideas, a shoulder to lean on, and endless cups of coffee made all the difference. Many early morning discussions helped to clear up questions, confusion and problems. But most of all, I thank him for introducing me to the most beautiful corner of the world, the Yukon.

To my wonderful mother, whose unfailing support and confidence have made such a difference, go many thanks. The small things she did for me while I was busy with my thesis kept me healthy and happy. I would also like to thank her for her help proof-reading the final product.

Special thanks go to Francis H.M. Jones, formerly of the University of British Columbia for the contribution of his time and equipment. The data provided by his radio-echo sounder and subsequent discussions helped to clarify certain aspects of my research.

Many undergraduate students played a large role in helping me acquire my field data over three summers. For their help, I would like to extend my thanks and wish them well in their own studies and careers.

To G.K.C. Clarke, K. Hewitt, D.A. St.-Onge, A. Williams and others go many thanks for suggestions and advice given during formal and informal discussions, very necessary for writing a comprehensive research paper. Many thanks go also to some of my very special friends for reading my work, and for providing moral support.

Finally, I would like to thank the Department of Indian and Northern Affairs and the

Natural Sciences and Engineering Research Council for their financial support provided through the University of Ottawa's Northern Research Training Program, and through NSERC Operating Grants. I would also like to express my gratitude to Parks Canada for allowing me to carry out my research in Kluane National Park, to the Arctic Institute of North America for the use of their resources, and to the pilot D. Makkonen of Trans North Air for his logistics support and friendly visits.

NOTE

The toponyms 'Surging Glacier', Non-Surging Glacier' and '3rd Glacier' as applied to the glaciers, the streams or the meteorological stations are not official geographic designations. They are used only for convenience.

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CHAPTER I

INTRODUCTION AND PROJECT DEFINITION

1.1 INTRODUCTION

Ice-dammed lakes are an important part of the hydrology of every major glacierized area in the world. An ice-dammed lake may be defined as a substantial body of standing water, located in, on, under or at the margin of a glacier, such that its existence is in some way dependant on damming by glacier ice (Blachut & Ballantyne, 1976). Ice-dammed lakes are by nature relatively transitory and highly dynamic features of a glacial landscape. These bodies of water, whether they are ephemeral surface ponds or large lakes, serve as storage basins. Hydrologically, some of these lakes are important because of the sudden, rapid draining of a large part or all of the stored volume of water after a slow filling. This sudden drainage (or jokûlhlaup) can have serious repercussions in the area into which these lakes drain. Documented cases attest to the destructive force of these sudden discharges: entire alpine valleys have been laid waste, and bridges, roads, other engineering structures, as well as agricultural land and housing have been destroyed.

Ice-dammed lakes are good general indicators of climatic alteration and glacier oscillation. They can form under conditions of general climatic warming and glacier retreat, or they may form as a result of rapid glacier advance. A recent example of a lake created due to a glacier surge is 'Russell Lake', formed by the advance of the Hubbard Glacier in Southeast Alaska. (Anderson, 1986). This glacier blocked a fjord for a period of two months, during which time the volume of water behind the ice dam rose dramatically. This lake drained over a period of hours, after rapid failure of the ice dam due to hydrostatic pressure. Other lakes formed under similar conditions are Lake Donjek, Lake Alsek and 'Hazard Lake' in the Yukon. On the continental glaciation scale, a few examples of lakes contained to some extent by an ice margin are Lake Missoula and Lake Agazziz. At present, the majority of ice-dammed lakes form due to glacier downwasting and retreat, and these will be the focus of this research.

Ice-dammed lakes formed due to glacier downwasting are thought to go through three phases, one of semi-permanent existence, one of frequent drainage, and one of permanent drainage. Each of these phases is thought to be associated with varying conditions of the damming margin. Each phase also has certain sedimentary processes which correspond with the hydrological conditions. The most critical phase of the existence of these ice-dammed lakes, hydrologically speaking, is during the period of frequent drainage, since 'frequent' can range from several times per summer to annually, bi-annually or less often.

The rate of filling of these lakes depends upon a number of factors such as drainage area, topographic limits of the lake itself, effectiveness of the ice barrier, and source of water. The maximum level to which these lakes will fill is dictated by the height of the ice dam. Once full, drainage may be initiated. The exact mechanism of drainage is still unknown, but there are a number of hypotheses which have been proposed.

Sedimentary deposition in proglacial lakes is well documented, and many of the processes operating therein also apply in glacier-dammed lakes. However, the annually draining characteristic of their hydrological regime alters the sedimentary sequences. These lakes drain partially or fully, which interrupts a normal, year-round depositional schedule. The rising and falling water level during filling and draining also affects the formation of the deltaic sedimentary sequences.

1.2 OBJECTIVES

The objective of this research is to contribute to the evaluation of extremes of hydrological systems in alpine areas through the study of a glacier ice-dammed lake with an annual drainage characteristic. Establishing the triggering mechanism for drainage of one particular lake will contribute to a greater understanding of the range of drainage mechanisms operating in these lakes. Furthermore, definition of the role of varying hydrological processes affecting sediment deposition will assist in the modelling of idealized sequences forming under similar conditions. Application of these theoretical sequences to other lakes (present or former), could then help to re-interpret sequences of uncertain genesis.

1.3 HYPOTHESES

1. *Drainage of the ice-dammed lake occurs as a result of floating of the ice dam.* The water level must reach 9/10ths the height of the ice barrier to initiate flotation, and thus trigger drainage. The floating of the ice margin before drainage occurs indicates a secondary triggering mechanism must also be implicated. Possibilities for this secondary mechanism are:

   (A) glacier movement along the bed exerts strain on the ice, and planes of weakness can be opened to provide an escape route for the lake water.

   (B) hydrostatic pressure on the base of the ice forces open a tunnel in the ice.

   (C) warmer, denser bottom-hugging currents provide enough heat to exploit planes of weakness in the ice and develop a tunnel.

2. *The lake is filled primarily by drainage off the Kaskawulsh Glacier.* Maintenance of lake water temperature at less than 1°C will indicate a dominance of the cold water supply.

3. *The basin inflow moves as a high density underflow.* Thermal density stratification of the lake water ensures that most of the suspended sediment load will be deposited in topographically low areas.

4. *The late summer/early fall drainage of the lake will cause the deep water sedimentary sequence to be missing the fine clay layers usually deposited during the winter months.*

5. *A high degree of variability in the sedimentary sequence is caused by daily variations of discharge and sediment input from glacial feeder streams.*

1.4 THE STUDY AREA

The lake is situated at 60°46'N, 139°07'W, on the north side of the Kaskawulsh Glacier in the St. Elias Mountains, Yukon Territory (figure 1). The lake is located at an altitude of 1600 m above sea level, with the surrounding peaks rising to 3100 m a.s.l.

The size of the basin is approximately 48 km^2, that is about 10 km long and 5 km wide, and is fed by three small tributary glaciers. These glaciers occupy 34% of the basin area.

The area was completely covered by ice as recently as 1865 (Denton & Stuiver, 1966). Since then, general climatic amelioration has led to glacier downwasting, and gradual exposure of the valley. This basin is mostly free of ice, since it is south-facing, and receives large amounts of insolation (mean for July 1965: 514.1 langleys) (Gray, 1985). The study area is below the equilibrium line for the Kaskawulsh Glacier, so that all snow melts each summer. On the three small glaciers which feed into the lake, most of the surface snow melts in the summer. Only the upper cirques (or accumulation zones) of these glaciers remain snow-covered. Two of the glaciers are in states of general retreat, and the third is surging at

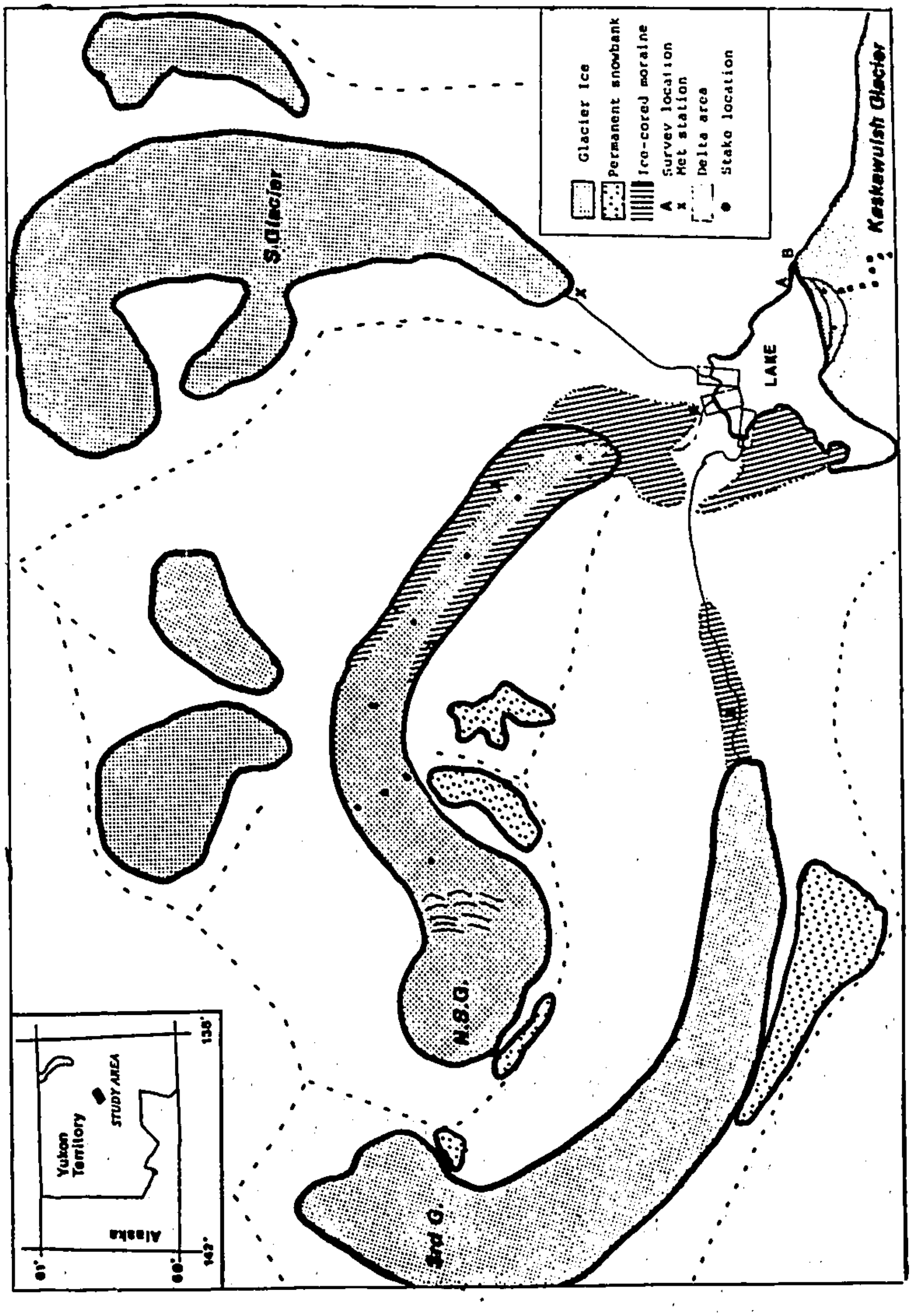

Figure 1: STUDY AREA

present.

The geology of the basin is composed of Paleozoic age rocks. The eastern half of the basin is characterized by volcanics, greywackes and carbonates of the late Cambrian to Permian age, intruded by diorite and granodiorite plutons. The western half of the basin is mostly low grade metamorphosed sedimentary rocks such as sandstones, shales and slates. Quartz veins, found in conjunction with these metasediments, contain galena and iron pyrite (Gray, 1985). Devonian carbonates, located higher in the icefields, are found as erratics in the basin near the ice margin. They were transported by the Kaskawulsh Glacier and deposited in the basin by the re-entrant ice tongue. The whole of the bedrock assemblage in the basin has been deformed at least three times, once prior to the intrusion (late Paleozoic time (270-290 my) to Tertiary), a second at unknown time, and the latest in late Miocene time during the St. Elias uplift (Campbell & Dodds,1978; Gray, 1985).

The climate of the region is highland subarctic, with orographic precipitation on the coastal side of the range, and a tendency towards semi-aridity on the eastern side of the icefields. The temperature is moderate, with the mean daily maximum temperature being above 0°C from March to October. The daily mean temperature is usually in the range of 15-20°C in July and August. The temperature range (both mean and extreme) in this area is moderated, since strong, constant katabatic winds do not allow temperature extremes to develop (Taylor-Barge, 1969).

The total annual precipitation for the site is approximately 800 mm (water equivalent), with 200 mm from the summer and 600 mm from the winter (Gray, 1985). The Atmospheric Environment Service (A.E.S.) records for the nearest stations show that the average amount of snowfall for this area ranges from 76-138 cm/yr (based on 1982-1986 data for Burwash Airport and Haines Junction).

Winds in the area are mostly controlled by the topography. These winds may be part of the regional weather systems, but they are mostly 'glacier winds'. The predominant wind direction at this site is down-glacier; from the south-west, thus from within the icefields (Taylor-Barge, 1969). These winds have speeds which may vary through the day, reaching maximum velocities on summer afternoons when the air/ice temperature difference is greatest (Gray, 1985). These winds bring a lot of cloud, however, most of the precipitation is associated with N.W. winds (Taylor-Barge, 1969).

There are almost no plants or vegetation in the basin. What there is is confined to small patches located on the shoulders of the surrounding ridges. The predominant types of plants seen are *Dryas* spp., *Carex* spp., *Oxytropis* spp. and *Artemisia spp.* (Roy, personal communication, 1987). In general, since there is little soil development, the material remains loose and extremely mobile.

The lake is cold monomictic, meaning it has a limited ice-free period, restricted circulation, a small thermal gradient, little or no thermocline and surface temperatures which are constantly near, at, or below 0°C (Hutchinson, 1957).

The basin (figure 2) is filled with glacial deposits, ranging from moraines to fluvioglacial deltas to glaciolacustrine varved sequences. Some of the till deposits are ice-cored at present, while others show melt-out and collapse features. The east side of the basin is the most active, having been recently renewed by a surge of the cirque glacier. Well defined lateral moraines and push ridges are also present. The western side of the basin shows many stagnant ice features such as kettle-type depressions, mudflows over ice cores and contorted lake sediments overlying till. There are glaciolacustrine sequences at several locations within the basin, often perched atop till mounds and moraine ridges. In addition, with three small cirque glaciers feeding into the lake, there is a large amount of fluvially-reworked material. There are extensive areas of braided channels, and some quite large deltaic features.

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Figure 2: Aerial photograph of the study site.

1.5 SCOPE OF THE INVESTIGATION

The investigation will focus on the drainage basin of the ice-dammed lake. Three major topics will be emphasized, the water balance of the lake, the drainage mechanism and the role of the ice margin, and the sedimentary sequence developed under current hydrological conditions.

Information to be included in the investigation are data collected over two consecutive summer field seasons on :

(1) the size, shape and depth of the lake;

(2) the rate of filling and draining of the lake;

(3) the volume and rate of water input into the basin;

(4) ice margin flotation and drainage mechanisms;

(5) influent stream water temperature and suspended sediment concentration;

(6) sediment dispersal processes;

(7) sedimentary sequences (of current lake deltas); and

(8) general climatic information.

In addition, aerial photographs (A13132, A13133, A22999, A24763), topographic maps (1:250,000), geological maps (G.S.C. map 1134A) and A.E.S. climatic data will provide a fuller understanding of the hydrology of the basin.

The research will be presented in the following format. A discussion of previous documentation on many aspects of the research will be followed by a description and analysis of the methods and techniques used. Chapter four will contain an in-depth analysis of the lake water balance, including the field results, calculations and conclusions. Chapter five will examine the ice movement and margin behavior. The sixth chapter will concentrate on the sedimentary research, both the processes operating within the basin, and the sedimentary sequences formed by these. Subsequently, any general conclusions and suggestions for further research will be elucidated.

CHAPTER II

BACKGROUND RESEARCH

2.1 INTRODUCTION

Research on glacier ice-dammed lakes has shown that the earliest records of the drainage of such a lake are from 1201 A.D. in an Icelandic parish record. Most of the earliest work was carried out by Icelandic researchers, since this was one of the first glaciated areas inhabited. Their work was mostly descriptive, concerned with qualitative narratives of destruction by the catastrophic drainage of these lakes (Thorarinsson, 1939). Following the emergence of the glacial theory in the 1840's, investigations into the subject of ice-dammed lakes were centered on interpretation of landforms and sediments left by now-empty Pleistocene lakes (Blachut & Ballantyne, 1976).

The emphasis in recent research has been on more quantitative aspects of these lakes. Contemporary workers are using detailed observations and measurements to construct hypotheses on lake drainage mechanisms. Studies have been conducted with the intention of establishing general criteria, such as the nature of the retaining ice wall, lake temperature regimes (Campbell, 1968, 1973) water inputs, parameters of subglacial drainage conduits (Rothlisberger, 1972; Nye, 1976), form and frequency of lake drainage, drainage triggering mechanisms, presence of strandlines (Sturm, 1986), evidence of flotation, and detection of leakage prior to drainage (Fisher, 1973). Researchers are now combining physical theory with accurate measurements of such parameters in attempts to model lake behavior (Clague & Mathews, 1973; Blachut & Ballantyne, 1976; Clarke, 1980,1982; Clarke & Mathews, 1981; Clarke, Mathews & Pack, 1984; Clarke & Waldron, 1984).

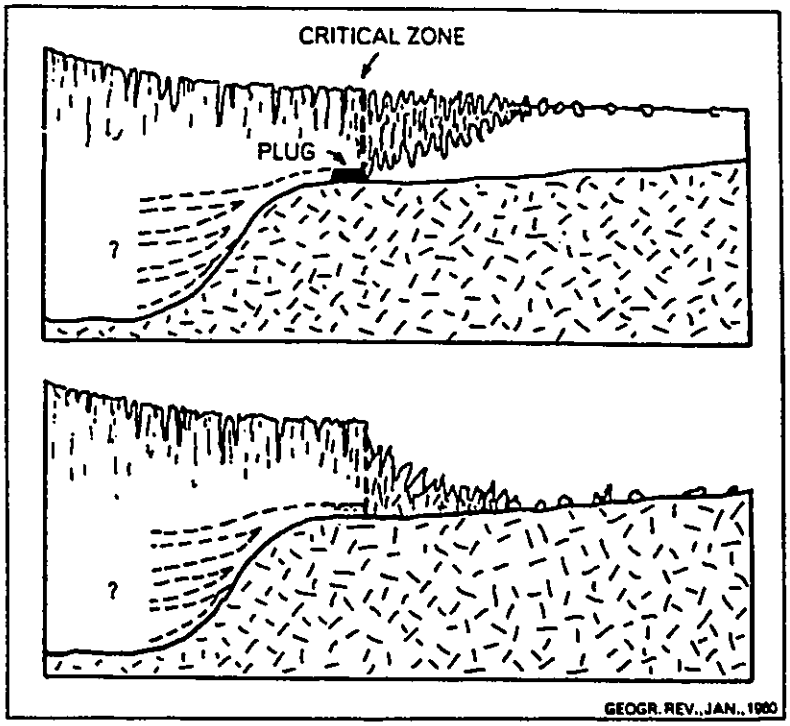
2.2 DRAINAGE MECHANISMS

There are several concepts which have developed as a result of research into the drainage mechanisms of ice-dammed lakes. Hypotheses on drainage of ice-dammed lakes can be placed into four basic categories: 1) barrier flotation (Thorarinsson, 1939); 2) temperature induced tunnel melting (Liestol, 1956) or tunnel enlargement due to pressure deformation (Glen, 1954); 3) geothermal or geophysical factors (tectonic activity)(Thorarinsson, 1953; Tryggvason, 1960; Bjornsson, 1975); and 4) supraglacial overtopping (Maag, 1969).

One of the earliest hypotheses was that of ice margin flotation (Thorarinsson, 1939). The concept supposes that the contained body of water will eventually lift the retaining ice wall away from the lake bed, allowing the water access to subglacial or englacial passageways through which drainage can take place. This mechanism requires the water level to have attained 9/10ths the height of the ice face (Thorarinsson, 1939; Marcus, 1960), and is based on the density differences between ice and water. Several corollaries have been added to supplement this idea. Lindsay (1965) and Whalley (1971) concluded that some extra undefined amount of force was necessary to overcome cohesion of the ice margin to the bedrock, and Thorarinsson (1939) included in his calculations a correction factor for crevassing. This is necessary as it is often evident that the critical zone of the ice barrier is probably some distance back from the ice margin (figure 3) (Aitkenhead, 1960; Marcus, 1960; Howarth, 1968).

Glen (1954) alternatively suggested that margin flotation was a response to a change in the equilibrium of hydrostatic and glaciostatic pressure.

The sagging and fracturing of ice dams after drainage due to the removal of the buoyant support of the water is consistent with a floating mechanism. Marcus (1960) and Sturm & Benson (1985) suggested that the post-drainage ice surface most likely reflects the general



Cross section of the ice barrier. In the upper view, the lake is filled with water, and heavily crevassed ice floats beyond the critical barrier. After drainage (lower view), the floating mass collapses. This accounts for extreme ice breakage in a zone of slight bed disturbance. Note the assumed position of the ice plug.

Figure 3: Ice margin flotation (Marcus, 1960).

shape of the bottom of the basin when the lake has drained.

Another series of hypotheses is concerned with drainage by tunnel opening in response to pressure and temperature factors. Liestol (1956) suggested that movement of a glacier over an uneven basal surface will cause small cavities to open, which can then be exploited by lake water. He proposed that heat available in the water would be transferred to the ice by friction, leading to melting and widening of the drainage passageway. Rothlisberger & Lang, (1987) also stated that the necessary heat for enlarging conduits could be provided by advection from the reservoir and dissipation of potential energy. The source of the heat in the basin may be the influent stream water, which has often been found to be several degrees warmer than the ambient lake water temperature. Gilbert (1972) concluded that warm water is necessary for the onset of the drainage in order to enlarge the tunnel at a more rapid rate than that of tunnel closure from ice overburden pressure. This may be one reason for the timing of many of these lake drainage events. Tufnell (1984) collected results of studies from the Alps, the United States and Norway. These show that a very large percentage of drainage events occur between the beginning of June and the end of September when the water will attain its highest temperatures.

Rothlisberger (1972) quantified the temperature/pressure imbalance within the drainage tunnel. During drainage, frictional melting of the tunnel by turbulent water flow occurs faster than plastic deformation due to pressure by the overlying ice. Mathews (1965, 1973) calculated that the heat transfer to the tunnel walls is nearly 100% in the early stages of the drainage event, but drops off to 35% as drainage progresses. These results from Summit Lake, B.C. are in accordance with Liestol's (1956) findings. Hence, the tunnel opening rate outstrips the closing rate, and drainage can occur.

The closing of conduits depends on the pressure difference between overburden pressure of the ice and water pressure. Rothlisberger (1972) suggested that the time required to close conduits can range from several days to several years. Liestol (1956) concluded from his work in Norway that outlets stay open permanently when the ice dam is less than 50 m thick, as this ice thickness provides insufficient ice overburden pressure to close conduits. This idea has major implications for glaciers in recession.

Glen (1954) approached the problem from a different perspective. He proposed that tunnel opening was brought about by pressure deformation at the base of the ice face. This mechanism requires fairly large water depths in order to invoke the shear stress between the water and ice necessary to open a tunnel in the ice margin. However, very few ice-dammed lakes achieve the necessary depths (Glen, 1954; Whalley, 1971). Furthermore, the rate of channel enlargement is too slow to form all of a passageway by this mechanism alone (Marcus, 1960), since the rate of enlargement by pressure deformation is not in agreement with the observed rapid release times of some lakes.

Another possibility for the development of a subglacial channel is seepage through permeable material below the glacier sole. When the temperature of the water rises in summer, ice melts at the sole and an initial channel can then develop at the contact (Rothlisberger & Lang, 1987).

Stenborg (1968) discussed the idea of the development of internal drainage systems both vertically downwards and laterally as part of a main drainage network within a glacier. According to Whalley (1971), the breaching of the ice dam and tunnel enlargement are insufficient to initiate drainage. Rothlisberger (1972) proposed that contact must be made between the lake and the general glacier drainage system before the ponded water can escape. Weertman (1973) outlined how a water-filled crevasse in a zone of tensile stress could widen downward to the point where such a connection is made with the glacier's main drainage system, initiating drainage. He explained that the terminal region of an ice dam would contain the necessary water, and would be subject to the high tensile stress levels required to widen a crevasse to such an undefined depth. This, coupled with the increasing carrying capacity of the drainage system as the summer progresses, indicates that this may be a possible explanation for lake drainage (Whalley, 1971). However, other factors are known to affect where the water actually flows, such as the flow rate and stress pattern of the damming glacier (Whalley, 1971), local topography, ice conditions and sediment load (Rothlisberger, 1972).

Some ideas on geothermal and geophysical triggers for lake drainage were developed in Iceland, where there is a constant interaction between glaciology and vulcanism. Bjornsson (1975) believed that subglacial volcanic activity melts some of the overlying ice, and lava subsequently infills the resulting cavity. This builds up the hydrostatic pressure, provides abundant heat to melt tunnels,and forces drainage. However, other researchers have suggested that the glacier bursts determine the moment of eruption and not vice versa

(Thorarinsson, 1953). This would be possible as a result of the flood suddenly releasing weight from the earth's surface, therefore initiating earth tremors and volcanic activity (Tryggvason, 1960). Tryggvason (1960) proceeded to establish five possible relationships between tectonic activity and jokulhlaups, which range from one side of the debate to the other. The correlation between tectonic activity and the drainage of ice-dammed lakes is still a matter of controversy. However, Post (1967) showed from his work in Alaska that 38 ice-dammed lakes showed no evidence of disturbance after the 1964 earthquake. Furthermore, regular time intervals between eruptions, which correspond with the typical cyclic behavior of an ice-dammed lake, tend to give more weight to the idea of the glacier burst being the initiation mechanism for the tectonism rather than vice-versa (Bjornsson, 1975).

The last major idea developed is that of ice-dammed lakes draining by supraglacial overtopping and marginal drainage. This hypothesis seems to be most applicable to subpolar or polar glaciers (Blachut & Ballantyne, 1976). This is because the glaciers are usually frozen to their bed, making flotation unlikely. This hypothesis involves the water level rising beyond the retention capacity of the basin, overflowing onto the ice or between the ice and the rock margin, and incising channels or deep gorges into the ice. A different set of triggering factors are involved in this type of drainage mechanism. Heavy precipitation, strong winds or calving icebergs causing waves, avalanches or landslides into the lake, or the outflow of lakes in a chain sequence further upstream have all been proposed (Stone, 1963; Maag, 1969).

One method of determining how drainage occurs is through examination of the resultant discharge pattern seen on a hydrograph. Lakes overflowing through supraglacial or marginal routes will show an initial sharp rise in the discharge record, followed by a gradual drop in the rate of flow. Subglacial or englacial drainage systems, on the other hand, generally show a sharp rise and a sharper, sudden fall in the hydrograph (figure 4) (Thorarinsson, 1939; Maag, 1969; Blachut & Ballantyne, 1976; Young, 1981).

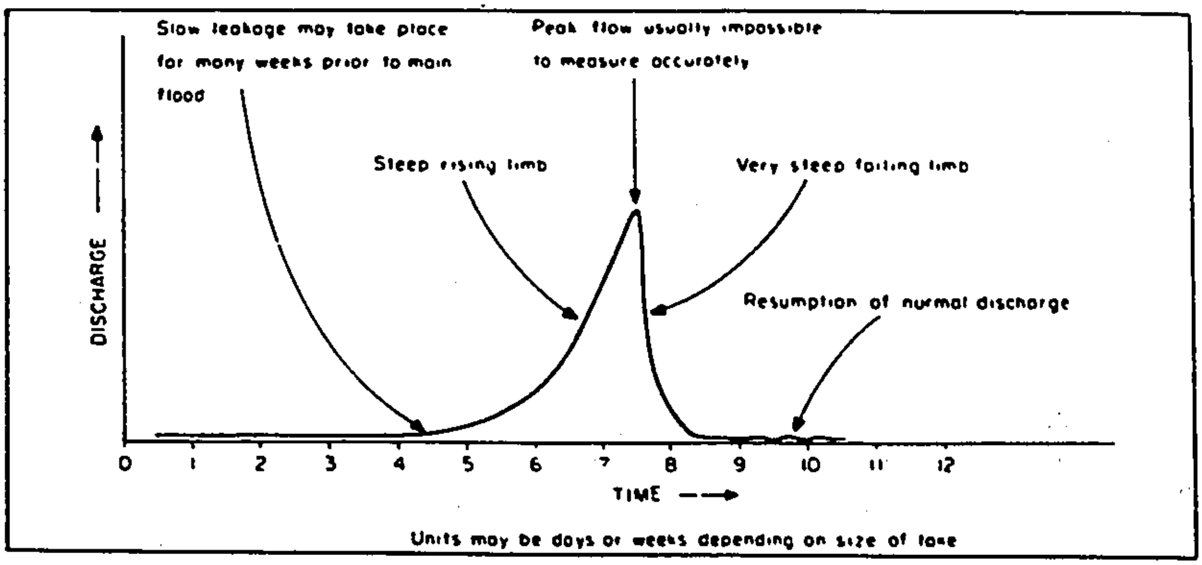


Figure 4: Generalized jokulhlaup hydrograph (Young, 1981).

There is no general consensus on the sequence of events which leads to drainage of ice-dammed lakes. Most workers have, after setting out their own ideas, stated that other factors and/or mechanisms may be partially responsible for drainage. Blachut and Ballantyne (1976) prepared a comprehensive review of the subject, and go so far as to differentiate between factors which trigger the commencement of drainage and the actual mechanism by which water is released from the lake. They assign ice dam flotation, volcanic eruptions and barrier overtopping to the first category and Glen's pressure deformation and Liestol's tunnel melting to the second.

2.3 SEDIMENTATION

Most research on ice-dammed lakes has concentrated on individual case studies rather than general research into this particular environment. These studies were divided fairly evenly between Pleistocene or late-glacial lakes (Fulton, 1965; Ashley, 1975; Shaw, 1975; Shaw & Archer, 1979; Atwater, 1986), and contemporary ice-dammed lakes (Gilbert, 1975; Gustavson, 1975; Smith, 1978,1981; Bradbury & Whiteside, 1980; Gilbert & Shaw, 1981; Weirich, 1985,1986; Liverman, 1987).

The emphasis of most of the work was on the examination of sequences, and subsequent environmental reconstruction for particular locations. Foremost among the techniques used for dating these glaciolacustrine deposits was counting varves. For example, Atwater (1986) used this method to successfully date a section in Northeastern Washington. His results indicated that late-glacial Lake Columbia existed between 15,550 (± 450) yrs BP and 13,050 (± 650) yrs BP. However, as more information on sedimentary processes emerged, it became apparent that the couplet-type deposit did not necessarily represent an annual sequence. For example, turbidity currents also deposit a gradational sequence, referred to as a Bouma sequence (Middleton & Hampton, 1976). These surge current deposits also have an apparent rhythmicity due to the repetition of a randomly occurring process. The time they represent is usually only a few minutes (Ashley, Shaw & Smith, 1985). There are a few distinguishing characteristics between annual rhythmites (varves) and surge rhythmites, but these are often not readily observable.

As a result of these types of problems with differentiation of sediment genesis, the emphasis of sedimentary research in glaciolacustrine environments shifted towards identifying processes and their resultant deposits. There are a few authors who have clearly outlined the processes acting in these sedimentary environments, the factors which control these processes, and the resulting continuum of possible sequences (Middleton & Hampton, 1976;

Selley, 1982; Ashley,Shaw & Smith, 1985).

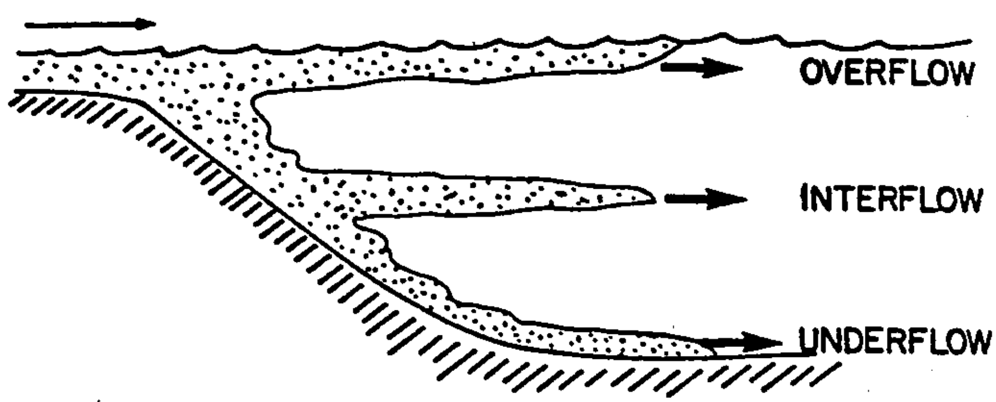
Glaciolacustrine environments are characterized by a wide range of physical processes, and a related range of depositional sequences reflecting these processes. The deposits usually represent a continuum of processes, since more than one process may be operating at the same time.

There are some general physical factors which control the rate and pattern of deposition in ice-dammed lakes. These include water density, suspended sediment concentration, water temperature (both of the lake and the influent stream), current velocity, wind speed and direction as well as the morphometry and topography of the lake basin (Ashley, Shaw & Smith, 1985). The combination of these factors, their individual and collective influence, and changes to any or all of them give rise to the range of processes occurring and the resultant depositional sequence.

Some of the processes operating in ice-dammed lakes are suspension flows (overflow/interflow/underflow)(Ashley, Shaw & Smith, 1985), and sediment gravity flows (turbidity currents, grain flows, debris flows and fluidized sediment flows)(Middleton & Hampton, 1976).

Suspension flows play a major role in sediment dispersal in glacial lakes. Density differences between the lake water and the influent stream water will determine whether the influent plume will travel as a surface current or 'overflow', an 'interflow', or a near-bottom 'underflow'. Inflow density is determined by three variables, temperature, suspended sediment concentration and salinity. Sediment concentration is usually the most important cause of density in glacial streams because it typically varies more rapidly and over wider ranges than temperature or salinity (Ashley, Shaw & Smith, 1985). The greater the density difference between the two water masses, the more energy must be expended in order to mix them. This enhances the tendency of the inflowing plume to be maintained as a discrete density current throughout the lake.

Overflows are the least effective and important sediment dispersal mechanism, since they are easily disrupted by wind or wave action. The method of deposition from an overflow is slow settling out of suspension. Interflows occur where the density of the influent water matches that of some midpoint in the water column. The sediment-laden water sinks to that point and disperses from there. Deposition occurs by slow settling out of suspension. Underflows , however, are one of the dominant distribution mechanisms operating in glacial lakes. These dense currents flow at the base of the water column, and deposit material solely in topographic lows (figure 5)(Ashley, Shaw & Smith, 1985).



Stipled pattern represents turbid inflowing river water whose density is different from all or part of the basin water.

Figure 5: Stratified inflow (Ashley, Shaw & Smith 1985).

Sediment gravity flows are defined as sediment transport in which movement parallel to the bed is provided by the pull of gravity (Middleton & Hampton, 1976). There are four flow types distinguished by the grain support mechanism. These four classes and their support mechanisms are as follows: 1) turbidity currents, supported by fluid turbulence; 2) grain flows, supported by grain to grain interaction; 3) fluidized sediment flows, supported by upward intergranular flow; and 4) debris flows, supported by a matrix made up of fine sediments and fluids (Middleton & Hampton, 1976).

Each of these processes deposits material in a different way which reflects the mode of transportation. Turbidity currents deposit a layered sequence over an erosional basal surface. Each of the layers reflect a drop in competence of the flow, with a decrease in grain size and current structures (Connybeare & Crook, 1968; Middleton & Hampton, 1976; Selley, 1982). Grain flows deposit material by mass emplacement once the stress driving the flow stops. Thus the deposit is usually a massive and ungraded layer of material (Middleton & Hampton, 1976). Fluidized sediment flows deposit material from the bottom up as the pore water is expelled, resulting in dewatering structures being formed (Selley, 1982). Debris flows move material as a massive body, depositing the debris on a non-abraded basal surface. The deposits are massive and ungraded (Selley, 1982). These are the general characteristics of each of these processes if they were to operate independently.

However, sedimentary deposits cannot be divided with ease according to genesis. Firstly, deposits may be formed by several processes acting concurrently, which would tend to either obliterate distinguishing characteristics of any one process or mix all the characteristic features together. Secondly, several types of deposits appear to be very similar, but have markedly different methods of formation. An example of this is rhythmite type deposits, which can either be formed by settling of material from over, inter or underflows in a year long time span (varves), or by rapid deposition of material from slump-generated surge currents in a few minutes (surge deposits)(Ashley, Shaw & Smith, 1985). Further, structures which are generally classified as being syngenetic or post-depositional in temporal origin when associated with one sediment dispersal mechanism, can also be syndepositional when associated with a different mechanism. Thus, it is important to keep in mind that each deposit is a part of a range of deposits, formed by a continuum of processes.

CHAPTER III

METHODS AND TECHNIQUES

3.1 INTRODUCTION

There are one main and two subordinate concepts which act as foci for this research project. The main one of these is the concept of the water balance within a glacierized basin. This concept is broad in scope, including all inputs, outputs and storage capacity for the system. The lake is a very important element of the water balance equation for the basin, since it acts as a storage unit as well as an output regulator. This major section of the water balance concept can be elaborated upon through studies of the concept of varying controls on sediment dispersal and deposition, and, because the lake drains annually, through the concept of ice-dammed lake drainage. Thus, one of the subordinate concepts is the drainage of the ice-dammed lake through a combination of the flotation of the ice margin and the opening of a subglacial drainage conduit by frictional heat. The other concept is the effect of the lake water on the sediment dispersal mechanisms and sediment deposition in an annually draining lake, in terms of its variable level, its temperature, and its movement patterns.

3.2 WATER BALANCE APPROACH

The concept of studying a glacierized basin in terms of its water balance is important in that measuring all sources of input into the system permits the determination of the maximum possible output. The storage capacity of the system is also a very important element, as it has the ability to regulate the rate of output of the system. In addition, understanding the significance of the contribution of each source of input allows the determination of the effect of each of these on the hydrological system as a whole.

This approach required the measurement of all contributing hydrological factors within the glacierized basin. The main areas of concern were stream flow inputs from three small feeder glaciers, climatological inputs (precipitation) and ablation from the retaining glacier. Storage capacity and lake volume, as well as the rates of filling and draining of the lake were also recorded. Several techniques were used to quantify these aspects of the basin water balance.

3.2.1 STREAM INPUT

There are three streams feeding into the lake whose sources are small cirque glaciers in the basin. These glaciers and their proglacial streams are nominally designated the Surging Glacier (S.G.) and Surging Glacier stream; the Non-Surging Glacier (N.S.G.) and Non-Surging Glacier stream; and the 3rd Glacier (3rd G.) and 3rd Glacier stream.

A major source for the stream water is the ablation from the three small cirque glaciers. This was measured by using ablation stakes drilled into the surface of the N.S. Glacier along its length and width. The amount of snow and ice melt at each location was recorded and converted to water equivalences (using measured densities). The amount melted was extrapolated over the whole glacier surface, depending on the snow conditions and surface material (ice vs moraine). This amount was then calculated as a total for runoff from the glacier.

Problems associated with this technique are due to the extrapolation of results from 12 point locations to the whole glacier surface. It is possible that sections of the glacier may not experience the same average amount of melt as do the measured locations, due to microtopographic or exposure considerations, and as such an error margin may be built in to the calculations.

The amount of ablation calculated for the S.G. using these techniques was extrapolated to the other two cirque glaciers, based on snow conditions and surface material. Calculations were not based on sampling, since each of the other two glaciers had a particular characteristic which made direct measurement impossible.

The total contribution of each of the 3 streams was measured using stage recorders, time-lapse cameras and stream gauging. Ott mechanical stage recorders were set up to monitor the fluctuations in water level of the streams in accordance with the procedures outlined in Church & Kellerhals (1970). Two of the streams were equipped with this instrumentation during the 1986 field season (N.S.G. and S.G. streams), and the 3rd G. stream was thus monitored during the 1987 field season. The 1986 field season demonstrated clearly that using a stage recorder system on the S.G. stream was impractical due to the pulsating nature of the discharge. The water level varied by as much as 30 cm within a few minutes, which often endangered the equipment. A 30 cm fluctuation in a stilling well of maximum length 1.2 m could exceed the bounds of the recorder if the water level was high. Further, the stream channel was made up of unconsolidated boulders, which were easily transported with increases in discharge. As a result, the stream frequently migrated by 3-5 m in a few hours, leaving the stage recorder on dry land. Thus, a useful, stable location for the stage recorder was usually short-lived.

The N.S.G. stream presented different problems. The stream flowed out from beneath ice-cored moraines, and tunnel collapses led to major shifts in the location of the main channel. In the space of a few hours, the stream migrated over distances of 200 m. or more. Further downstream, the channel was composed of unconsolidated gravels overlying ice cores. These gravels were extremely mobile due to erosion by the stream channel or melting of the ice core. Thus a stable site for the stage recorder was difficult to locate. The gauging site was moved at least three times during the 6 week field season, so the data collection was interrupted several times.

In order to circumvent these types of problems, a different technique was employed in 1987. Remote sensors in the form of time-lapse cameras were used. These 8 mm cameras were focused on graduated stakes (1 ft. graduations) in the stream channel. The water level was recorded with exposures at 5 minute intervals throughout the 11 week field season. The accuracy of the data collected using this technique is much reduced as insufficient lighting conditions for several hours each night did not permit recording of the water level. Also, distance from the stake (25-30 m) reduced the accuracy of the observation of the level on the gauging staff.

To supplement the water level data acquired with either a stage recorder or a time-lapse camera, stream discharge was measured on a frequent but irregular basis. Discharge readings were taken during lower flow conditions (mid-morning), rising flow conditions (mid-afternoon) and peak flow conditions (early evening) throughout the season. These discharge readings spanned the seasonal variations from snowmelt to icemelt within the basin. In 1987, for the N.S.G. stream, there were 16 days on which data were collected, for a total of 35 readings. For the S.G. stream, there were 15 days and 30 readings, and for the 3rd G. stream, 17 days and 36 readings in total. Stream gauging was carried out most often by wading, according to the standard technique outlined in Church & Kellerhals (1970). During the seasonal peak runoff period, however, salt dilution techniques were used as wading was too dangerous. Thus data acquisition on stream input employed several techniques.

The discharge data was acquired for use jointly by Dr. P.G. Johnson, as part of an ongoing study of discharge regimes from proglacial streams, and by the author for application to this section of the study.

3.2.2 CLIMATOLOGICAL INPUTS

The second aspect of the water balance studied was the climatological inputs (precipitation). The equipment used was standard M.R.I. meteorological recording stations with tipping bucket rain gauges. These machines recorded continually throughout the 1986 field season in 3 locations; at camp, in the S.G. valley and in the N.S.G. valley. In 1987, a fourth site was added in the 3rd G. valley, in order to obtain a more representative data set concerning climatological conditions throughout the basin. The data set acquired was more or less continuous, with a few breaks due to temporary equipment malfunctions.

The data acquired by this equipment provides spot locations for precipitation input, and this is then extrapolated over the basin. This may lead to erroneous calculations, since precipitation is not always of equal intensity and duration everywhere in a basin of 48 km^2 size.

Temperature data was also collected by the M.R.I. stations. This information was used to give a general indication of the amount of melt-days experienced through the season. This information is considered as indirect, since it does not provide a measurable amount of input into the water balance equation. However, it was used to support the ablation data, and to confirm trends seen during the melt season.

3.2.3 ABLATION INPUTS FROM RESTRAINING GLACIER

The third contribution to the hydrological balance is the ablation from the Kaskawulsh Glacier. The technique used to gain an indication of the amount of input was the following. 10 stakes were drilled into the ice surface of the Kaskawulsh glacier. The distance from the top of the stake to the ice surface was recorded on a regular basis throughout the field season. The thickness of the ice lost from the surface at the 10 stakes was averaged in order to negate any microclimatological factors which might have enhanced or prevented melt at any of the locations. This averaged ice thickness loss was multiplied by the surface area of the Kaskawulsh glacier which dips towards the lake. This area was delineated from aerial photographs (A13132-1951; A13133-1951; A22999-1972; A24763-1977). This was in turn converted into water equivalent data.

The accuracy of this data is reduced due to several factors. The first factor is that the stakes used to measure ablation were concentrated within a small area, rather than spread out over the approximately 4 km^2 area of ice surface which appears to drain into the lake. A second factor is that the stakes were moved to be redrilled throughout the season, which can supplement measurement errors.

Further problems associated with the measurement of this data were generated by environmental factors. Sinkholes on the surface of the Kaskawulsh connected to englacial conduits may not feed into the lake but rather into the Kaskawulsh Glacier's main drainage system. There was no way of ascertaining where these conduits drained to in the field. This could result in the loss of a large volume of meltwater being generated by ablation. Some volume of water may also have entered the basin from elsewhere on the Kaskawulsh, for example marginal up-glacier runoff. This addition would also affect the accuracy of the data. Furthermore, the overall surface area of the glacier draining into the lake was determined from aerial photography, so some error may also have been built in through the measurement of this area.

3.2.4 STORAGE CAPACITY AND LAKE VOLUME

To quantify the water balance of a glacierized basin, the storage capacity must be measured. The rates of filling and draining of an ice-dammed lake, when compared to the amount of input into the system, will provide information on the regulating function of the lake on the output from the basin.

There were several techniques used to quantify the water level and lake volume. The water level measurements were collected in the following manner. Two slopes were marked at 1 meter intervals, and the slope angle over each interval was recorded. One site was located on a steep shoulder, where the average slope angle was 32.3°, while the other was located on the shallow slope near camp, where the average slope angle was 10°. The water level was monitored at two sites in order to provide a comparison for the results from either one of the sites. The water level was recorded on a regular basis at both sites up to the maximum filling level, and as long as possible after the onset of drainage. There were problems with the accuracy of the results obtained by using this technique. The most important was the fact that at one site, there were a large number of readings taken, many of these slope angles were fairly small, and the possibility of introduction of error into the calculations increases with the number of entries.

The volume of the lake basin was calculated from the results of a post-drainage survey of the basin. This provided the aerial extent of the basin as well as topographic relief of the lake bed. A map of the basin was drawn from the surveying data and the lake volume calculated. The map was also compared to photographs of the lake at various stages in the filling/draining cycle. This provided a check of the extent of the topographic relief underwater at any given point, since the corresponding water levels were known.

The lake played a major role in the determination of the water balance of the basin. The lake itself was further studied in terms of the stratigraphic and sedimentary controls, and the physical controls leading to drainage. These two major aspects are examined in the following sections.

3.3 CONCEPT OF ICE-DAMMED LAKE DRAINAGE

The drainage of an ice-dammed lake is brought about by a combination of two (or more) factors; a triggering mechanism and a drainage mechanism. With respect to the lake under study, it was thought to be the flotation of the ice dam which acted as a triggering mechanism, while the opening of a subglacial channel by heat transference was the actual drainage mechanism. To study these two, several techniques were employed, but the emphasis of the field work was placed on the flotation of the ice margin.

3.3.1 ICE MARGIN FLOTATION

The first technique used to quantify any vertical movement of the ice margin was surveying. A set of ten stakes were located on the ice surface in a linear pattern of gradually increasing distance from the ice front. These were surveyed from two fixed points on a former ice-contact delta located on the valley side. The surveying was carried out at 4 day intervals during the first 6 weeks of the 1987 field season, then at 3 day intervals thereafter. As drainage of the lake was initiated, more frequent surveying was carried out in order to calibrate the total amount of movement of the ice face. The problems encountered in the surveying were instrument limitations, and difficulties with accessible and stable survey points.

A second technique used to establish the vertical displacement of the ice margin was the use of photography. Panoramas were taken of the ice front at frequent though irregular intervals during the season. These showed the progression of the lifting of the ice face during the filling and draining of the lake. Since not all of the panoramas were taken with the same camera nor from the same location, there was some difficulty in quantifying any changes in the ice face from the photographs. The best results obtained from the use of these photographs was the qualitative agreement of this visual evidence with the calculated results from the surveying data.

A third technique used was time-lapse photography. During the 1986 field season, an 8 mm time-lapse camera recorded the filling sequence of the lake at 50 second intervals. During the 1987 field season, the camera was set up 5 days before drainage was initiated. Filming was done at 2 minute intervals before drainage started, and at 1 minute intervals during drainage. These two film sequences provided further visual evidence of the vertical displacement of the ice face. Quantifying the displacement of the ice face from these films was not possible due to the distances involved between the ice margin and the location of the instrument.

3.3.2 TUNNEL MELTING

The second section of this concept, that of tunnel development due to heat transfer, was based on the ideas put forward by Liestol (1956), Mathews (1965, 1973), Gilbert (1972) and Rothlisberger (1972). The concept is that heat is transferred from a standing body of water to the ice, leading to the widening and lengthening of a plane of weakness in the ice. This would develop into a subglacial drainage conduit which must eventually connect with the main drainage system of the damming glacier.

The technique used to attempt to confirm this mechanism of drainage was the following. Temperature profiles were drawn up from measurements of the water column taken at numerous points within the lake. The temperature was recorded at .5 m intervals to the lake bottom by an Endeco digital thermometer. The temperature conditions at the base of the water column in locations near the ice face were of particular interest, indicating pockets of warm water flowing at depth. The continuity of these warm flows to the area proximal to the ice face indicated heat available to be transferred for tunnel enlarging at the base of the ice face. However, heat for tunnel melting can also be generated by friction from turbulent water passing through an enlarging tunnel during drainage (Drewry, 1986). However, only 2/3 of the frictional heat becomes available for melting (Rothlisberger, 1972). The amount of tunnel melting may be quantified by mathematical formulae. Since many of their variables could not be quantified within the scope of this research these mathematical calculations will not be carried out.

An attempt was made using another method to ascertain whether there were detectable drainage tunnels within or beneath the ice face. A digital impulse radar was utilized in order to scan the lower contact of the ice margin, to try to detect tunnels located there, and to give an indication of the shape and basal conditions of the ice margin.

3.4 SEDIMENT DISPERSAL AND DEPOSITION

There are several ways in which sediment entering a lake from a stream may be dispersed throughout the body of water. Proglacial streams generally exhibit varying densities on both a diurnal and a seasonal basis, which affects the dispersal mechanism. The dispersal mechanism(s) and subsequent lake stratification affect the deposition of sediment throughout the lake basin.

To gain a proper understanding of the processes operating and the resultant depositional sequences, several aspects were monitored: first, the sediment input into the lake; second, the movement of the water within the lake; and third, the depositional results investigated after drainage.

3.4.1 SEDIMENT INPUT

The sediment input from the three tributary streams was quantified through the use of Cygnus automatic liquid samplers and hand sampling. The size of the samples collected was 1 litre for hand samples and one of the Cygnus samplers, and two litres for the second Cygnus sampler. In the case of the 2 litre samples, only 1 litre of the sample was retained in order to maintain a standard volume of water used for filtration. The S.G. stream was sampled hourly during the 11 week field season with a few minor breaks in the data record. The N.S.G. stream was also sampled hourly for the first 6 weeks of the field season, and then from 7:00 hrs to 22:00 hrs for the remaining 5 weeks. The 3rd G. stream was sampled hourly from 08:00 to 21:00 hrs, at approximately 4 day intervals, for a total of 16 times during the season. All three streams were also sampled at 15 minute intervals for an 11 hour period in order to closely monitor the sediment fluctuations in the stream load during one day. This sampling was carried out at the peak runoff period for the season (20/7/87- 3rd G.; 21/7/87- S.G.; 22/7/87- N.S.G.).

There were problems associated with this technique of data collection. The first of these was an environmental problem, in which the machine often froze overnight, due to dropping water and air temperatures. The second of these was the fact that the automatic liquid samplers only collected samples from within the water column, therefore the sediment carried as bedload is not accounted for. Thirdly, there were many problems with maintaining the machines in working order. The vacuum system by which these machines function was not always operating to full capacity, thus only partial samples could be collected. In addition, one of the machines broke down completely during the 1987 field season, so that sampling had to be carried out by hand thereafter. This eliminated the overnight section of the record for the latter part of the summer. Fourthly, some sediment was lost during the filtration process, which employed pressure filters of the type designed for the Norwegian Hydro Electric Board. The filters used Whatman 42 and 44 ashless filter papers, which have a 98%

retention capacity for 2 micron and 2-3 micron particle sizes respectively. Despite this high retention capacity of the filter papers, when too much pressure was applied to the filters, part of the sample could run out along the sealing ring. In this way, some of the sediment could be lost.

The samples collected were to be analyzed using standard grain size analysis techniques, to differentiate the sizes of material by percentage of weight.

This segment of the research was carried out by Dr. P.G. Johnson as part of an ongoing investigation of discharge and sediment regimes from proglacial streams. Some of his results were used by the author. The information on the suspended sediment load was used to gain an understanding of the size ranges of material being carried in suspension by the three streams. This would help to predict which size of material was most likely to be deposited in the deltaic areas. However, the information on the suspended load had to be tempered with some comprehension of the amount of bedload also being transported.

3.4.2 WATER MOVEMENT AND SEDIMENT DISPERSAL

The tracing of movement of the water within the lake basin was carried out by the use of a Marsh McBirney lake current meter. The water current velocity was recorded at .5 m intervals to the bottom. Profiles were drawn from data acquired at numerous locations throughout the basin. These profiles presented a general picture of the location of flows within the water column, and indicated the general stratification of the water column in terms of flowing vs stillstanding water. The location of these flows within the water column gave an indication of the dispersal mechanism for the sediments, hinting at possible depositional patterns after drainage.

Problems experienced with this technique ranged from accuracy of the data (in some cases negative velocities were recorded), to sampling error (in some cases it was impossible to tell where the bottom was). Also, sampling sites were chosen on a random basis due to iceberg distribution. In addition, changes in sampling locations were necessary as the lake filled, to compensate for the changing water depth. If the readings were taken at the same site through the season, then the water level would be different each time, rendering comparison impossible. Furthermore, the equipment only registered the magnitude of the current velocity and not the direction in which the flow was moving. This made it impossible to distinguish between currents originating from the streams, ones originating from subglacial tunnels feeding into the lake, and those resulting from water movement outward beneath the ice margin resulting from constant leakage.

Sediment dispersal within the lake was not easy to document. A few sediment samples from within the lake water were collected, but these were fairly random and could not be considered to be representative of the whole lake. These were collected in an attempt to determine the concentration of sediment at higher flow points within the water column, and to give an indication of the amount of sediment moving in suspension to the various parts of the lake basin. No definitive conclusions can be drawn as to the effectiveness of any technique of dispersal. The only clear indication that sediments were being moved came from photographs of the lake from higher altitudes. These display variations in water color in the lake, which correspond to varying densities in the sediment-laden water.

3.4.3 DEPOSITION OF SEDIMENTS

A post-drainage examination of the lake basin showed which areas contained the greater thickness of lacustrine deposits. These were sampled in order to provide some basic information on the dispersal patterns of the lake. Very detailed information on the exact source of the deposits could be obtained by mineralogical analysis, since the various sections of the basin were characterized by markedly different rock types. Analysis of the lake bottom deposits would allow the determination of the source of these sediments, which would in turn provide evidence of the effectiveness of the dispersal mechanisms of the three streams. However, temporal and financial constraints rendered this analysis beyond the scope of this research.

The major emphasis of this component was put on the deltaic deposits of the three feeder streams. After drainage, the deltaic areas were studied to gain an understanding of the processes operating within deltaic areas, and how these were affected by a changing water level. All 3 stream valleys were photographed to illustrate the sections exposed by the 1986 drainage. Detailed diagrams were drawn, containing information on the grain size, structure and depositional sequence reflecting the filling and draining cycle. The same process was repeated after the 1987 drainage, when a new set of sections was exposed. While specific sections were photographed and recorded for detailed analysis, the deltas were photographed and surveyed in a panoramic format. This permitted not only the examination of specific sections containing structures of interest, but also allowed for the interpretation of the deltaic formation as a whole.

CHAPTER IV

LAKE WATER BALANCE

4.1 INTRODUCTION

The measurement of the water balance of the basin provides the basis for interpreting the dynamics of the lake, and is a first step towards the prediction of lake drainage. It also provides the base data for the assessment of the evolution of the lake if the Kaskawulsh Glacier continues to downwaste.

The three basic components of the water balance, input, output and storage, can each be expressed as an equation for application within the basin.

Input= basin ablation + basin snow melt + precipitation + Kaskawulsh melt 1

Output= f(leakage, evaporation, drainage) 2

Storage= input - output 3

The results of the quantification of all sources of input, output and the holding capacity of the basin will be analyzed and discussed in this chapter.

4.2 INPUT

There are three major sources of water input into the lake basin. These are: 1) three glacierized basins (combination of basin snowmelt and glacier snow and ice melt); 2) precipitation; and 3) the Kaskawulsh Glacier. These can be broken down into a number of components which provide a relatively accurate figure of their contribution to the system.

4.2.1 BASIN ABLATION AND SNOW MELT

Basin ablation and snowmelt quantify the amount of water input into the system from the three sub-basins. Three data sets were collected, ablation measurements from the cirque glaciers, discharge readings and stage records (using both charts and time-lapse films from the streams). Measurement of the ablation gives an estimate of the amount of water which is derived strictly from glacier melt. The stage and discharge records permit the quantification of the input from sources such as snowbank melt, ice-cored moraine melt and water released from subglacial or englacial storage areas.

The input from the sub-basins can be expressed by:

$$Q = A(g) + M(b) + P - E \quad \text{.......... } 4$$

where Q = discharge, P =precipitation, E = evaporation, A(g) = glacier ablation and M(b) = basin snow melt. No measurements of evaporation were taken as evaporation was assumed to be negligeable, and the equation simplified to:

$$Q = A(g) + M(b) + P \quad \text{.......... } 5$$

(A) ABLATION FROM CIRQUE GLACIERS

Basin ablation provides a large proportion of the water generated. It is measured by monitoring the thickness of snow or ice lost from the glacier surface throughout the season, and by converting this thickness into its water equivalent. This loss can then be applied to the area of the glacier, providing a measure of the volume being input into the hydrological system.

Twelve stakes were drilled into the surface of the Non Surging Glacier in a pattern spanning the length and width of the ice mass (figure 6).

The only section of the glacier not measured was the upper part of the cirque. It was inaccessible due to a large ice fall, and was also avalanche fed, making measurement of surface snow thickness difficult and dangerous.

The surface conditions into which the stakes were placed varied with each location. At stakes 1-4, there was snow over the glacier ice until Aug. 1st. Stakes 5-11 (except 6 & 8) were snow covered on June 9th but were successively uncovered by June 29th (table 1). Stakes 6,8 and 12 were located on the lateral moraines, #6 on the east side and #8 and 12 on the west side of the glacier. The debris thickness on the ice cored moraine varied down the length of the glacier. The debris cover at stake 8 was very thin (1 cm), while at stake 12 it was thicker, in the range of 3-5 cm. The debris cover on the moraine on the east side was much thicker; at stake 6, approximately 15-25 cm thick.

The amount of melt at each stake varied considerably over the length of the glacier. At some stakes, the majority of the material melted was snow, while at others it was ice. The water equivalent of the icemelt was calculated using a standard density of .9 for glacier ice.

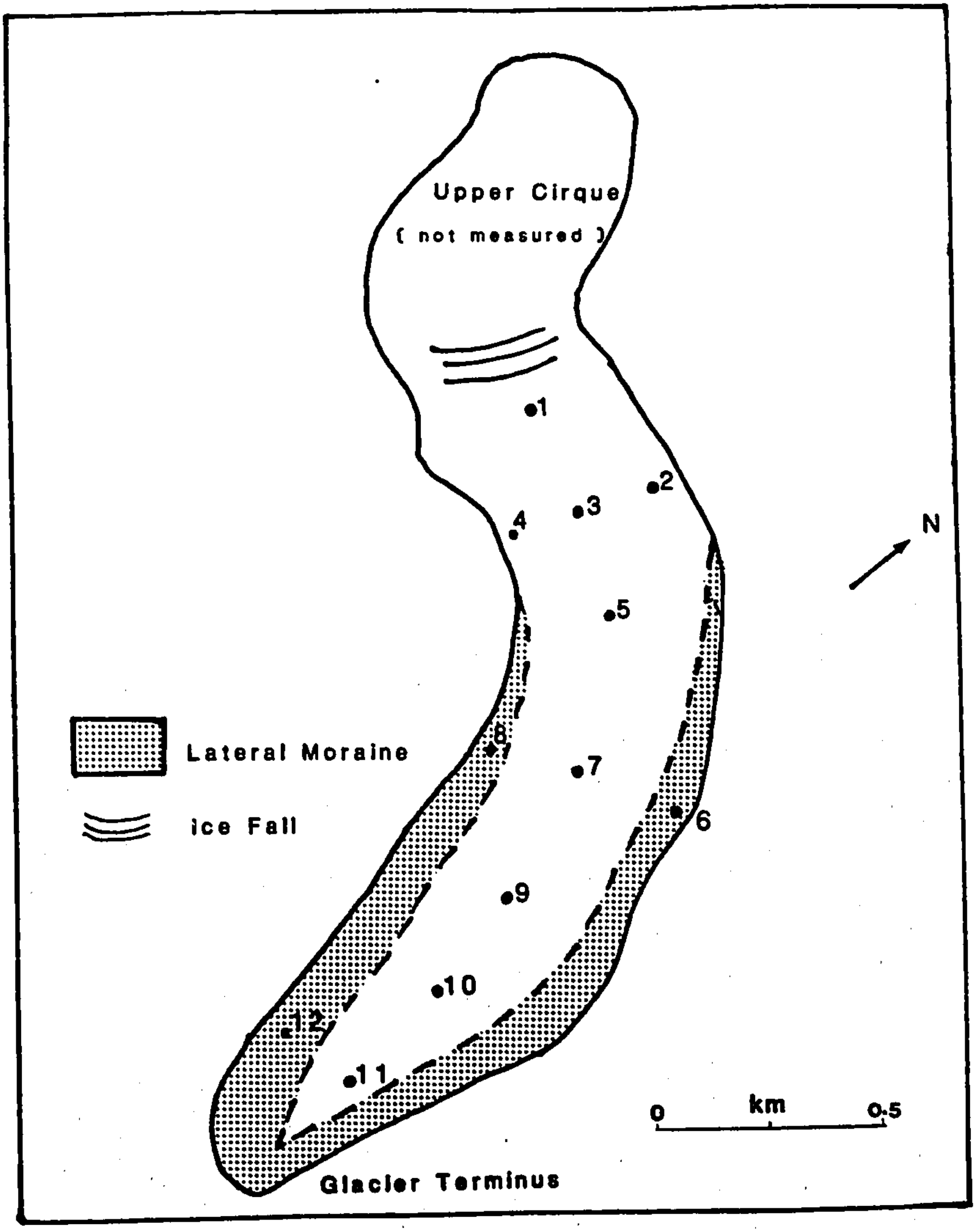


Figure 6: Stake distribution pattern, Non Surging Glacier, 1987.

Table 1: *SURFACE CONDITIONS OF THE NON SURGING GLACIER, SUMMER 1987*

stake date	1	2	3	4	5	6	7	8	9	10	11	12
09/06	S	S	S	S	S	S	S	S	S	S	S	S
15/06	S	S	S	S	S	S	S	S	S	S	S	S
20/06	S	S	S	S	S	M	S	S	S	S	I/s	M
25/05	S	S	S	S	S	M	S	S/s	S/s	I/s	I	M
29/06	S	S	S	S	I/s	M	S	S	I	I	I	M
03/07	S	S	S	S	I	M	I/s	M	I	I	I	M
05/07	S	S	S	S	I	M	I/s	M	I	I	I	M
09/07	S	S	S	S	I	M	I	M	I	I	I	M
16/07	S	S	S	S	I	M	I	M	I	I	I	M
23/07	S	S	S/s	S	I	M	I	M	I	I	I	M
01/08	I	I	I	S	I	M	I	M	I	I	I	M
07/08	I	I	I	S	I	M	I	M	I	I	I	M
13/08	I	I	I	I	I	M	I	M	I	I	I	M

S=snow cover	I=bare ice
M=moraine surface	/s=slush on surface

Snow density measurements were recorded during the season, and water equivalent values calculated from these densities. These measurements accounted for water storage in the snow before runoff occurred.

The net mass loss due to ablation was calculated by dividing the glacier into three zones and determining average ablation figures for each zone. The division of the glacier was made along the line of stakes 2-3-4, which were oriented cross-glacier (figure 6).

The upper section enclosed the area of the main glacier from the dividing line to the ice fall, the whole upper cirque and the two small tributary glaciers on the east side of the valley wall. The conditions at stake 1 were assumed to be representative of the conditions in the upper cirque and on the tributary glaciers. This may have introduced errors in the calculation of the ablation volume. The lower section of the glacier included the zone of glacier surface below the dividing line at stakes 2-3-4. This section was very limited in size due to the presence of extensive moraines along each side which occupy a large area of the narrow valley. The third section contained the extensive ice-cored moraines which extend along both the east and west sides of the glacier.

The area of the N.S.G., including its tributaries and ice-cored moraines was calculated to be 5.19×10^6 m^2. These figures were based on the 1977 aerial photographs (A-24763-18 & 19). It was divided into three areas: 1) upper area including tributaries - 3.51×10^6 m^2; 2) lower area - 0.66×10^6 m^2; and 3) moraines - 1.01×10^6 m^2. The calculations of area were made using a planimeter. A small error may have been introduced by the use of the 1977 imagery.

The water equivalence of the ablation, when averaged for each section and multiplied by the area, provided an estimate of the total volumetric input into the system by the N.S.Glacier. The water equivalent average for the upper section of the glacier was 1.04 m; for the lower section, 2.56 m ;and for the moraine section, 1.48 m. Thus, the total volumetric addition from the N.S.Glacier into the lake was 6.85×10^6 m^3. Because evaporation and sublimation were not calculated, this figure is taken as input to the lake.

The ablation data from the N.S.G. were used to estimate the contribution from the

Surging Glacier and the 3rd Glacier. These estimations were based on the calculated average snow and icemelt from the upper section of the N.S.G. This was employed because the other glaciers were located in the same altitude as the upper N.S.G. Also, snow conditions appeared similar, although no sampling of snow conditions was carried out. The volumetric input estimated by this method for each of the two glaciers was approximately 5.28×10^6 m^3 of water each, since the areas of the two glaciers, based on measurements from 1977 air photos, were very similar. The S.G. at that point was downwasting and retreating, whereas in 1985 it began to surge. This made it difficult to accurately determine the size of it during the 1987 field season, therefore possibly introducing an error factor into the calculations. The 3rd Glacier was slightly smaller in 1987, since this glacier appeared to be slowly retreating and downwasting. No calculations of meltout from moraines were included in either of these two estimates, since the S.Glacier had no ice-cored moraine and the 3rd Glacier's moraines were poorly defined, non-continuous, and seen only on the south side of the valley. This rendered it impossible to estimate their actual extent. Therefore, the measurements of contribution from the other 2 glaciers are less reliable than the N.S.Glacier calculation. However, they provide a reasonable approximation which can be used in the water balance equation calculations.

Hence, the total amount of contribution from the 3 cirque glaciers was estimated at 1.74×10^7 m^3 of water. This does not take into account any error margin, but provides a reasonable figure for the breakdown of the stream input totals into the glacier ablation vs basin snowmelt figures.

(B) DISCHARGE

Discharge measurements provide information on the amount of water being supplied by the streams into the lake. It is measured near the confluence of the streams with the lake to ensure the maximum possible water volume is recorded. The formula used to calculate the discharge for the basin is the following:

$$Q = A(g) + M(b) + P \quad \text{......} \quad 5$$

This formula takes into account all variables which provide water to the lake from the sub-basins.

Seasonal discharge was calculated by taking point discharge readings at random, and maintaining continual stage records. The random discharge values were equated with the corresponding times on the stage records, and the discharge values for the rest of the season were calculated from the stage-discharge relationship.

Discharge readings were taken by wading. This involved measuring the stream channel depth at 0.5 m intervals across the width, recording the velocity of the stream flow in each 0.5 m element and calculating the discharge V x A. Cross sections were measured with each velocity measurement due to the mobile nature of the channel. Readings were taken throughout the 1987 summer on an irregular basis. During June, discharge readings were taken only once a day, usually in the mid-afternoon. During July and August, whenever possible, readings were taken 3 times a day (table 2). These times were midmorning (10:00 - 11:30 hrs), midafternoon (13:30 - 17:00 hrs) and early evening (19:00 - 22:00 hrs), and correspond to low, medium and high stage levels. These three time periods allowed the calibration of the daily runoff curves for different times throughout the season, such as the snowmelt and main icemelt peaks. For the N.S.G., 37 readings were taken on a total of 17 days; for the S.G., 30 readings over 15 days; and for the 3rd Glacier, 36 readings over 17 days. One of the problems with this technique was that with very high discharge, readings were not taken due to safety considerations. Thus, there are no discharge records for peak flow conditions.

Stage records were taken using two different techniques. The 3rd Glacier stream was monitored by a mechanical water level recorder. This extended from June 8th to August

Table 2: DISCHARGE READINGS, SUMMER 1987

	9/6	10/6	11/6	12/6	14/6	24/6	25/6	27/6	28/6	30/6
NSG-m			x				x			x
NSG-a	x	x		x	x		x	x		x
NSG-e							x	x		x
SG-m			x							x
SG-a	x	x		x	x		x			x
SG-e							x			x
3rd-m				x				x		
3rd-a	x	x		x	x	x	x			x
3rd-e							x		x	x

	3/7	7/7	11/7	19/7	23/7	1/8	5/8	8/8	12/8
NSG-m	x	x	x	x		x		x	x
NSG-a	x	x	x	x	x	x	x	x	x
NSG-e	x	x	x	x		x	x	x	x
SG-m	x	x	x	x		x	x	x	x
SG-a	x	x	x	x	x			x	x
SG-e	x	x	x	x				x	x
3rd-m	x	x	x	x		x		x	x
3rd-a	x	x	x	x		x	x	x	x
3rd-e	x	x	x	x		x	x	x	x

m-morning reading, 10:00 - 11:30 hrs
a-afternoon reading, 13:30 - 17:00 hrs
e-evening reading, 19:00 - 22:00 hrs

20th, with a one week period of data interruption from August 3rd to August 10th. There were several other short periods of discontinuity of the record due to equipment malfunction. These occurred throughout the season, but the record in general was reasonably complete. The N.S.G. and S.G. streams were monitored by remote time-lapse cameras, since suitable locations for placement of other recording equipment were not available. Their records were unreliable before the 1st of July, because night-time temperature at or below 0°C caused malfunctions in the cameras. However, from July 1st onwards, the records were relatively reliable, except for late evening or overnight hours due to insufficient light.

The stage and discharge records provided an idea of the discharge regimes from the basin. The discharge curve for the 3 streams followed the normal pattern seen in a glacierized basin, that of diurnal and seasonal variation (figure 7). The diurnal fluctuation was virtually non-existent during minimum flow in late winter, but began to appear in early summer. In the early part of the season, a slow rise in discharge was initiated by enough insolation and sufficiently high daily maximum temperatures to promote ablation. During this time, the variations in stage for most days were of the order of magnitude of 2-3 cm. The discharge which this represented was between 0.15 and 0.5 m^3sec^{-1}, or on average 4.5 x 10^4 m^3day^{-1}. The initial delay in the runoff of the early summer melt occurs for a variety of reasons. The subglacial and englacial tunnel network had decayed by closure and freezing, and had to be reactivated before sufficient discharge could take place. In addition, the first rain and snowmelt was retained in the firn either as superimposed ice or as slush. Stenborg (1970) estimated that this phase can store about 25% of the melted runoff before releasing it later in the summer.

After the 18th of June when the temperature rose sharply, the stage also increased. The daily variation in water level increased to 4 cm, and the discharge rose to between 0.6 and 0.8 m^3sec^{-1} (6.5 x 10^4 m^3day^{-1}). The evening of June 20th showed a marked increase in stage by 10 cm. This may have corresponded with the temperature peak on that day, but changes in weather conditions are more likely to induce fluctuations on the order of days (Sugden & John; 1976). Small scale fluctuations lasting seconds or minutes are thought to be related to the tapping and draining of pockets of water in the glacier. During the last ten days of June, the stage record began showing pulses on a relatively frequent though irregular basis. The magnitude of the pulsations varied, from as little as 1 cm (June 27th) to as much as 10 cm (June 24th, 27th). These pulses were probably due to opening of reservoirs within the glacier and flushing of the ice surface of retained snow/slush.

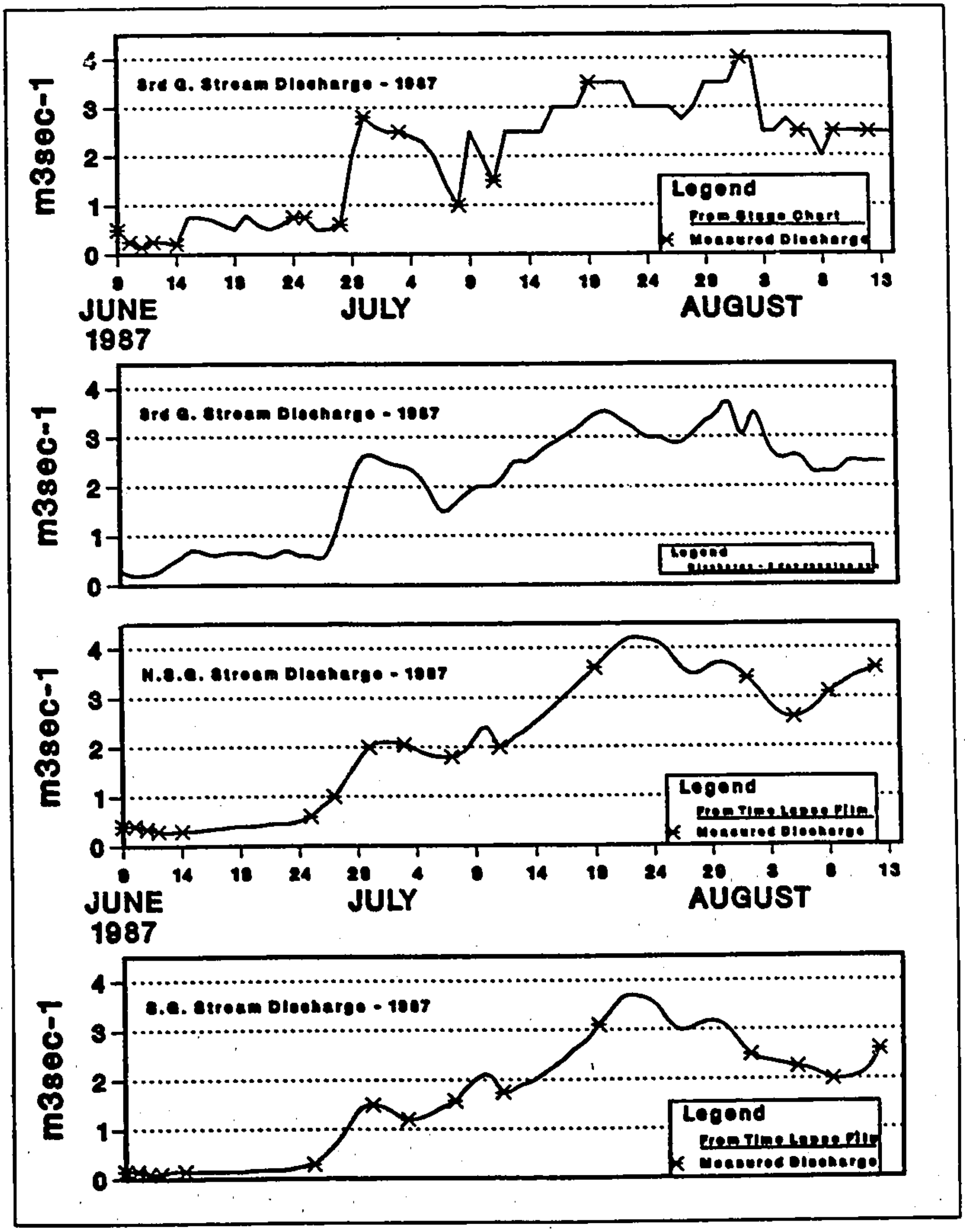


Figure 7: 1987 Discharge curves for all streams.

From the base flow level during the early part of the season, the discharge rose to a first peak on June 30th. This varied from 2.8 m^3sec^{-1} for the 3rd G. Stream to 2.0 m^3sec^{-1} for the N.S.G. Stream to 1.5 m^3sec^{-1} for the S.G. Stream. This first main peak was associated with the loss of the majority of surface snow from the previous winter, a phase of rapid discharge due to the release of water temporarily stored in the snow (Sugden & John; 1976).

After the snowmelt peak, discharge responds to climatic conditions influencing ablation and precipitation (Sugden & John; 1976). This can be seen by matching the discharge curve with the mean temperature curve. The precipitation events recorded during the summer were of such small magnitude that they had no effect on the discharge curve.

From July 1st onwards, the diurnal fluctuations of the stream became more marked. It is well documented that the amplitude of the diurnal fluctuations tends to increase towards late summer when ablation rates are highest (Sugden & John; 1976). A maximum of 16 cm difference in water level between low flow conditions (10:00 hrs) and peak flow conditions (18:00 hrs) was recorded. Discharge also increased correspondingly, rising to between 1.5 and 3.0 m^3sec^{-1}. The average daily discharge during the first 15 days of July was 2.0×10^5 m^3day^{-1} (3rd G. Stream).

In addition to normal diurnal fluctuations, some pulse-type events were recorded on the stage charts. The evening of July 11th showed an increase of stage of 140 cm. This was not just increase in water level, but reflected a relatively large increase in stage coupled with the filling of the stilling well with debris. The filling of the stilling well with gravel was symptomatic of a rise in the bed of the stream. This resulted in an accompanying rise of the water level being recorded. The overall result of the sudden increase in discharge and stage coupled with a channel shift was the dislocation of the recorder. Thus, it was impossible to quantify the actual amount of water level and discharge increase during this flash flood. However, the stage record illustrated the reality of rapid and high magnitude fluctuations in stage and discharge which can damage equipment and render the record unreliable.

From July 15th (when the equipment was recalibrated) until July 25th, there were no more major fluctuations in the stage beyond the normal diurnal variation of 10-12 cm. The discharge rose during this time period. July 25th marked the beginning of a warming period, which was not reflected in the discharge. For the 3rd G. Stream, the discharge had been dropping since the 22nd, and dropped by more than 5.0 x 10^4 m^3day^{-1} to 2.4 x 10^5 m^3day^{-1} on July 27th. It began to rise again after the 27th in response to the period of warmer temperatures which began on July 25th. The lag time of two days between the rise in temperature and the corresponding rise in discharge could not be accounted for by the distance of travel, since the glacier was only 8.5 km long. Reasons for the lag time could be that a lot of snow still on the upper glacier would have retained water, and transit time. The temperature peak in the days around the 25th of July resulted in the icemelt peak for the N.S.G. and S.G. streams, near the 23rd of July. These correspond much more closely to the temperature peak than does the 3rd G. Stream.

On July 28th, another large pulse of 20 cm of stage occurred during a 1 hour period. This corresponded with a temperature peak on that same day. The pulse was possibly caused by several days of melt being temporarily ponded on the glacier, then suddenly released, or a stream relocation event. The discharge curves for all three streams show a slight rise on that day. July 30th showed another disruption on the stage record, but since it resulted in severe damage to the equipment, it was impossible to tell if it represented a channel shift or a genuine pulse in discharge. The rest of the season's stage records were unreliable due to the damage the equipment sustained.

The 3rd G. Stream's discharge curve rose to its icemelt peak on August 1st. This followed 10 days after the main temperature peak of the season and 2 days after a lesser temperature peak. For this stream, discharge peaked with the main temperature peak, but then rose higher again ten days later. For the other streams, there were lesser peaks corresponding to the main peak for the 3rd G stream. Following the icemelt peak, the discharge declined rapidly for the last days of the record. The rest of the season would have seen a continuing decline in the discharge to baseflow once the mean temperature dropped to near or below 0°C.

Problems associated with using a mechanical stage recorder originated from the nature of the stream bed and channel. The stream ran through a gravel outwash area, and tended to migrate in multiple channels over the outwash plain until ice-cored moraines restricted the flow into one channel. As the stream did not completely fill the area between the banks, the main flow often migrated laterally over unconsolidated material, which was easily eroded and redeposited elsewhere. These shifts in the location of the main flow made it difficult to record the stage since the recorder location would be left dry.

There were also problems associated with recording stage using time-lapse photography. The time-lapse camera record was restricted to daylight hours. The cameras and timing devices malfunctioned at cold temperatures so the June record is not sequential. A more continuous record was obtained from July 1st to August 20th. Quantitative data from this record is limited because of channel shifts, destruction of the calibration staff and camera relocation. Drainage pulses and rapid channel migration, however, were obvious on the films.

The evaluation of these records was very subjective. Their main use was the confirmation of similar and different behavior by the 3 streams. The discharge data used for the calculation of the water balance equation came from the 3rd Glacier stream, which was monitored by a mechanical water level recorder. Extrapolation of the 3rd Glacier stream data to the S.G. stream was based on the equal area and altitude range of the two glaciers. The N.S.G. was smaller than the S.G. or the 3rd G., but the difference in size was small and the glacier extended into a lower altitude range. Thus, the discharge for the 3rd stream was also extrapolated to the N.S.G. stream. Errors are acknowledged for these calculations but, as is indicated below, even a large percentage error would not affect the general conclusions.

To conclude, discharge from the basin can be expressed by formula 5: Q = A(g) + M(b) + P. As seen in the preceding section, the total input from ablation of the three glaciers was estimated at 1.74×10^7 m^3 of water, and the total discharge for the summer was estimated at 3.51×10^7 m^3. Thus, the amount of input measured in the discharge value but not accounted for by ablation can be calculated as follows:

$$M(b) + P = Q - A(g) = 1.76 \times 10^7 \text{ m}^3.$$

Slightly more water than that supplied by ablation was added to the basin's discharge by other melt in the basin, and by precipitation.

4.2.2 CLIMATIC CONTRIBUTIONS

There are two main components of climatic data relevant to studies of water balance within a glacierized basin. One of these components' contribution is direct, the other indirect. Precipitation (P in equation 4) is measurable as water added directly to the system. Temperature has an indirect effect in promoting or retarding snow and ice melt.

(A) PRECIPITATION

Precipitation is measured at point locations, and the data recorded is extrapolated for the whole basin.

It was recorded at three stations during the 1986 field season and at four stations during the 1987 field season. One of the rain gauges used in 1987 malfunctioned, so the data set for 1987 was based only on three stations. The main emphasis was placed on the 1987 data, since the water balance was calculated for that season. The data for 1986 is presented to demonstrate the variability of the summer precipitation input.

The three stations at which data were collected in 1987 were the Camp station, the S.G. station, and the 3rd G. station. Figure 8 presents the precipitation record for 1987. Gaps in the data were due to equipment problems at the N.S.G. and S.G. stations. These problems are insignificant in the calculations of the water balance because precipitation was very low.

During June 1987, only two precipitation events were recorded. The first one of these recorded .2 mm per day for four days, the 14th to the 17th, for a total of .8 mm. However, this precipitation registered only on the 3rd G. station, no trace of this precipitation reached the Camp station. The event of June 21-22, however, was of a higher magnitude and recorded at both stations. The total input over the two days was 14.8 mm for Camp station and 13.6 mm for 3rd G. station.

July had more precipitation days, but all the events were of low magnitude. There were two multiple-day events, the 14-15th and the 24-26th, as well as three other single days. Of those 3 single days, only one was of sufficient magnitude to record at all three stations. Camp station reported a total input during July of 6.4 mm, 3rd G. station, 7.0 mm, and S.G. station 5.2 mm (between July 17th and 31st). Thus, in July, less precipitation was recorded than in the 2 day event of June 21-22 1987.

In August, precipitation was recorded at 2 time periods. A one day event of magnitude 1.2 mm registered only at the S.G. station on August 2nd. The second period, August 14-18th, which was the major precipitation event for the season, lasted five days and included both rain and snow. The snow component was mostly the night-time contribution, and would accumulate until the temperature rose in late morning, then melt and record a large volumetric input over a short time span. In the upper reaches of the basin, most of this precipitation fell as snow and persisted for a number of days. Hence, between Aug. 14th and Aug. 18th, the total precipitation recorded was: 1)Camp station - 54.2 mm; 2)3rd G. station - 52.4 mm; and S.G. station - 51.2 mm. However, since the lake had drained completely by Aug.12th, this major precipitation event does not contribute to the calculation of the precipitation input to the lake water balance for 1987.

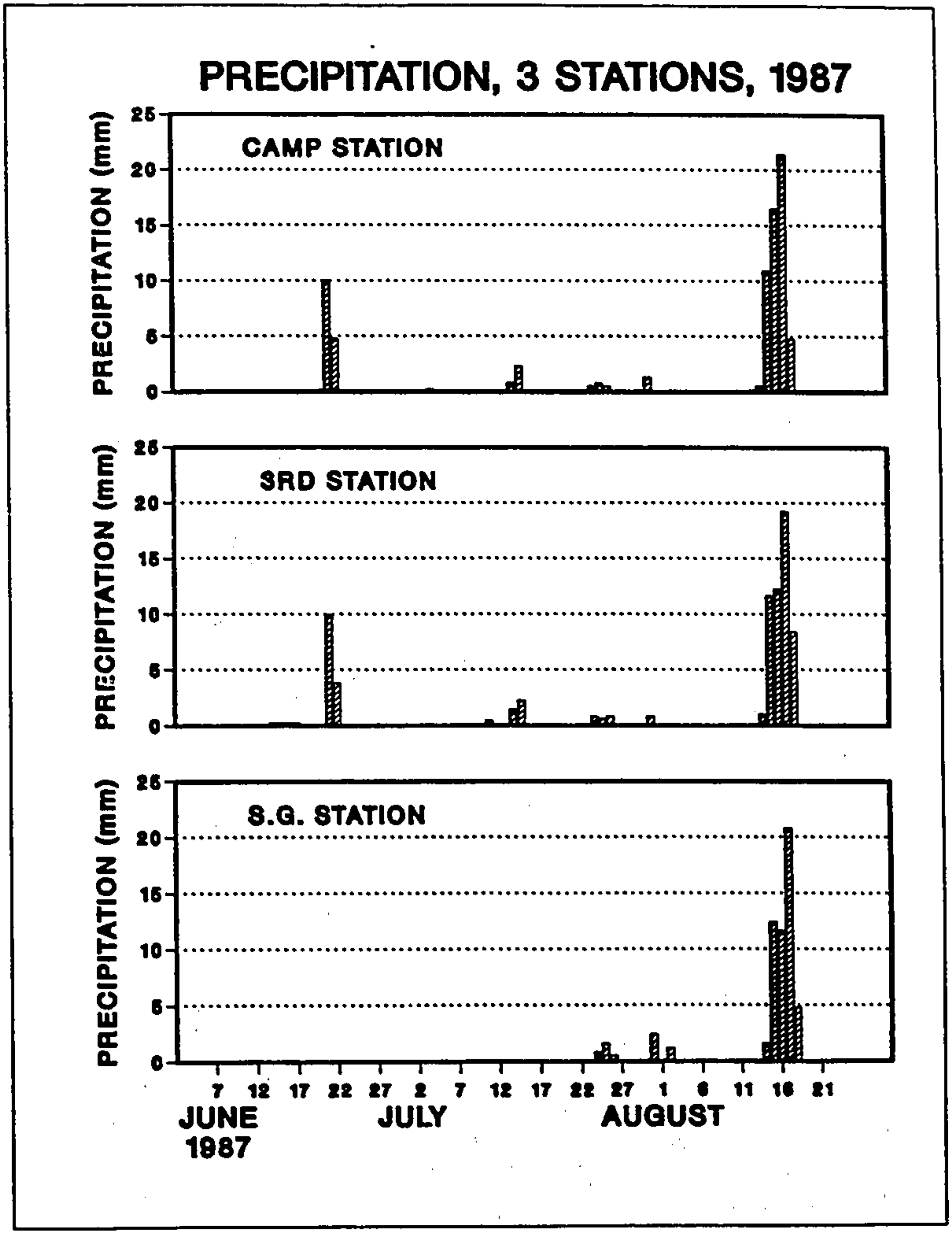


Figure 8: Precipitation, 3 stations, 1987.

The total amount of precipitation added to the system before Aug. 12th was determined by averaging the amount from the 3 stations. The totals for the three stations were: 1)Camp station - 21.2 mm; 2)3rd G. station - 21.4 mm; and 3)S.G. station (July 15-Aug.13) - 6.4 mm. Taking into account the lack of data from the S.G. station for the first 6 weeks of the field season, the average was obtained by dividing the season into 2 segments and averaging each of the two segments, then adding these averages. Thus, the average of 2 stations from June 8th to July 16th was 18.3 mm, and the average for 3 stations from July 17th to Aug. 12th was 4.3 mm. Adding these gave an average for the whole 1987 summer season of 22.6 mm precipitation input into the system.

The volume of input which the precipitation represents was calculated by multiplying the amount of precipitation by the total area of the catchment basin, 48 km^2. An approximation of the volumetric input by precipitation during the 1987 field season is 1.08×10^6 m^3.

Despite the fact that the majority of the basin was underlain by either permafrost or old glacier ice, most of the precipitation probably infiltrated into the ground throughout the basin. This is because there was relatively little precipitation and this was of very low intensity. Thus, precipitation had a negligeable effect on the stream discharge regimes. The only exception to this might have been the June 21-22 rain or snow event which might have temporarily slowed the rate of snow melt.

A comparison of the 1986 and 1987 data shows the variability of precipitation (table 3). In 1986, data was collected only during the month of July except for 4 days into both June and August as start and end points. The average amount of precipitation during July 1986 was 51.5 mm, as compared to the same month in 1987, which had an average of 6.2 mm. Two seasons data can not be used for any general conclusions, but Johnson (personal communication) has found extreme variations between seasons since 1970 and large variations over short distances within seasons.

Table 3: PRECIPITATION INPUTS (mm) 1986 & 1987

Station	June	July	August	Total
Camp 86	n/r	55.1	n/r	55.1
NSG 86	n/r	32.8	n/r	32.8
SG 86	n/r	66.6	n/r	66.6
Camp 87	14.8	6.4	54.2	75.4
SG 87	n/r	5.2	52.4	57.6
3rd G 87	14.4	7.0	52.4	73.8
n/r= no record from the time period				

(B) TEMPERATURE

Temperature is an indirect indicator of water input into the system, since in general warmer temperatures mean higher ablation rates. The temperature, recorded at several locations, provides information on the general temperature regime for the basin. This can be used to estimate the number of melt days.

Temperatures were recorded at 4 stations during the 1987 field season. (figure 9 and Appendix A). The gaps in the data at N.S.G. and 3rd G. stations were caused by equipment malfunctions. Trends in the data are demonstrated by using 3 day running averages which damp daily fluctuations.

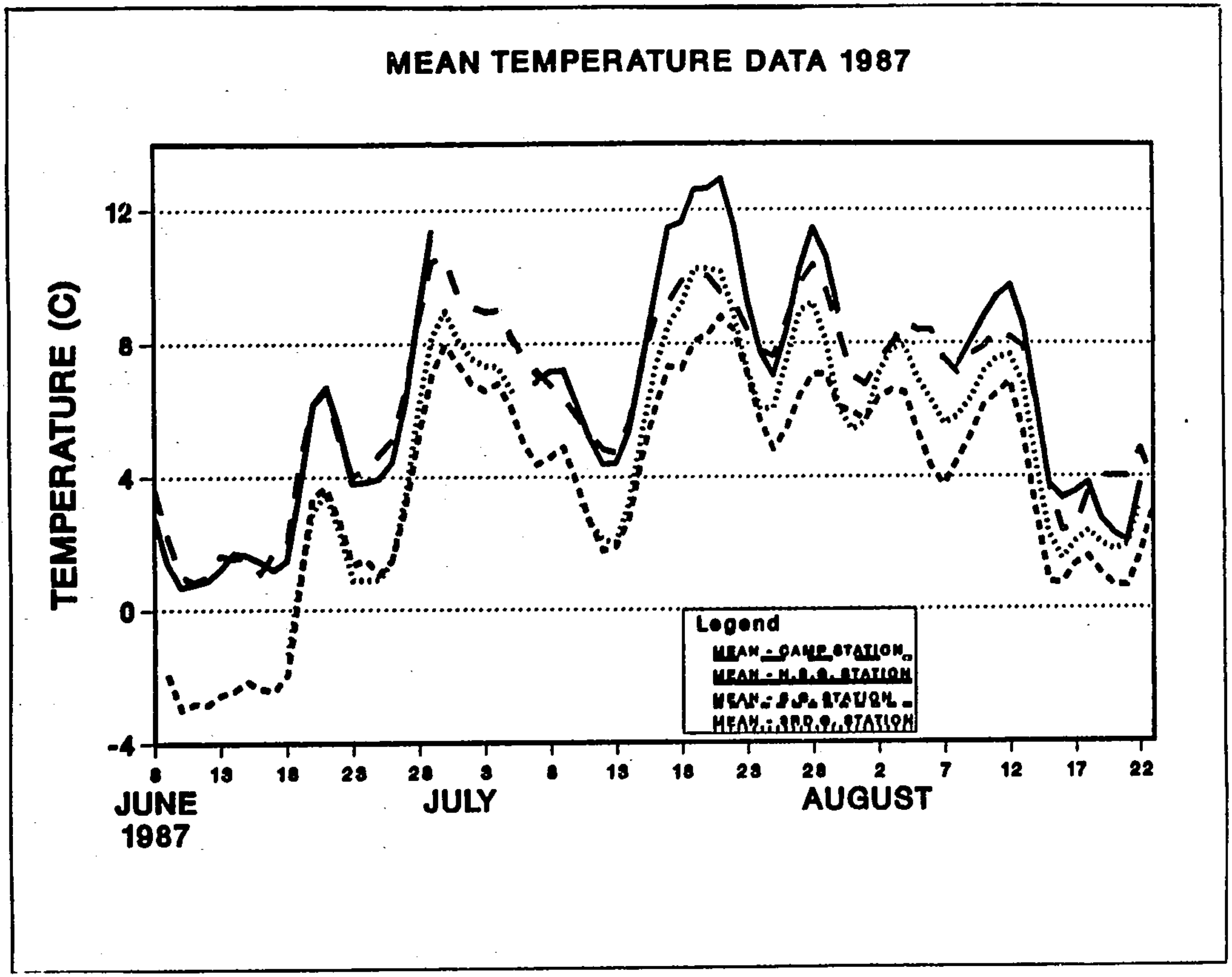


Figure 9: Mean temperature data, 1987.

From June 8th until June 18th, the mean temperature for the basin remained low. For both the N.S.G. and Camp stations, the minimums for this period dropped to below the freezing point while the mean was just above 0°C. For the S.G. station though, the mean stayed below 0°C.

From the 18th of June until August 22nd, the mean temperature at all four stations remained above 0°C but there were a number of fluctuations. The temperature started to rise on June 18th, peaking initially on June 21st at 7°C (Camp) and 4.5°C (N.S.G.). A second peak occurred on June 30th, with values of 8°C (S.G.) to 11°C (N.S.G.). This was followed by a 13 day period of steady decline. Between July 14th and August 12th, the temperature fluctuated over 3-5 day periods with a range of 4°C. The most extreme variations were experienced at the N.S.G. station, and the least extreme variations at the S.G. station. The S.G. station's mean temperature was the lowest overall. Between July 14th and August 12th, with the exception of July 19-22nd, the S.G. station's mean temperature stayed between 4°C and 8°C. The N.S.G. station's mean was the highest throughout the summer. Between July 14th and August 12th, the temperature ranged from 5°C to 14°C, but stayed mostly between 7°C and 12°C. Both the Camp station and the 3rd G. station ranged between 6°C and 10°C during this same time period. Following August 12th, there was a general decrease in temperature with a four-day period of snow and rain. The mean temperature for all four stations was between 0°C and 4°C for the remainder of the field season.

Temperature trends at all four stations were the same but there were differences of a number of degrees between stations. The S.G. station's mean temperature was the lowest due to the proximity of the met station to the Surging Glacier itself. The N.S.G. station data was the highest since the recorder was located on a moraine surface which radiates some heat. The 3rd station recorder was also located on a moraine surface, but was in shadow most of the day due to the morphology and orientation of the valley. The Camp station recorder was subject to 'glacier winds' blowing almost constantly off the Kaskawulsh and across the lake.

This near-constant cool wind neutralized the effects of higher temperatures due to the location of the recorder on a moraine surface.

Although the temperature on the surface of the glaciers was probably lower than at the recorder locations, the temperatures recorded at the 4 stations give temperature trends in the basin. The mean temperature curves provide an estimate of the number of days in which melting was likely to occur. Since the mean temperature for all stations was constantly above 0°C after June 18th, some melting occurred on a daily basis throughout the summer season. It must be acknowledged, however, that melt can also occur with temperatures below 0°C under certain circumstances.

4.2.3 KASKAWULSH ABLATION

Input from the retaining glacier is an important factor in the water balance of a system containing an ice-dammed lake. Measurement of the surface ablation only provides an approximated possible value for this input variable, since there are many other non-quantifiable aspects to be considered. The entire glacier does not feed the lake, only an area including the re-entrant ice tongue and surrounding parts of the main ice body topographically depressed towards the lake. Water may be added through the englacial or subglacial drainage network, as well as via marginal channels. Topographic considerations influence glacier surface drainage patterns, but not all water melted off the surface drains into the catchment basin. Glacial drainage networks are very complex, and surface streams can be diverted by crevasses and moulins to flow in a completely different direction. It is impossible to determine exactly how much water is being input from the retaining glacier, but the measurement of ablation may provide a ballpark figure.

Kaskawulsh Glacier ablation was estimated from measurements taken at 10 stakes. These stakes were located in close proximity to the margin of the lake. They were set up in a pattern moving away from the ice edge at 50 m intervals (figure 10). The stakes were drilled into the ice, and the distance from the top of the pole to the ice surface was recorded approximately every four days throughout the season for a total of 21 readings.

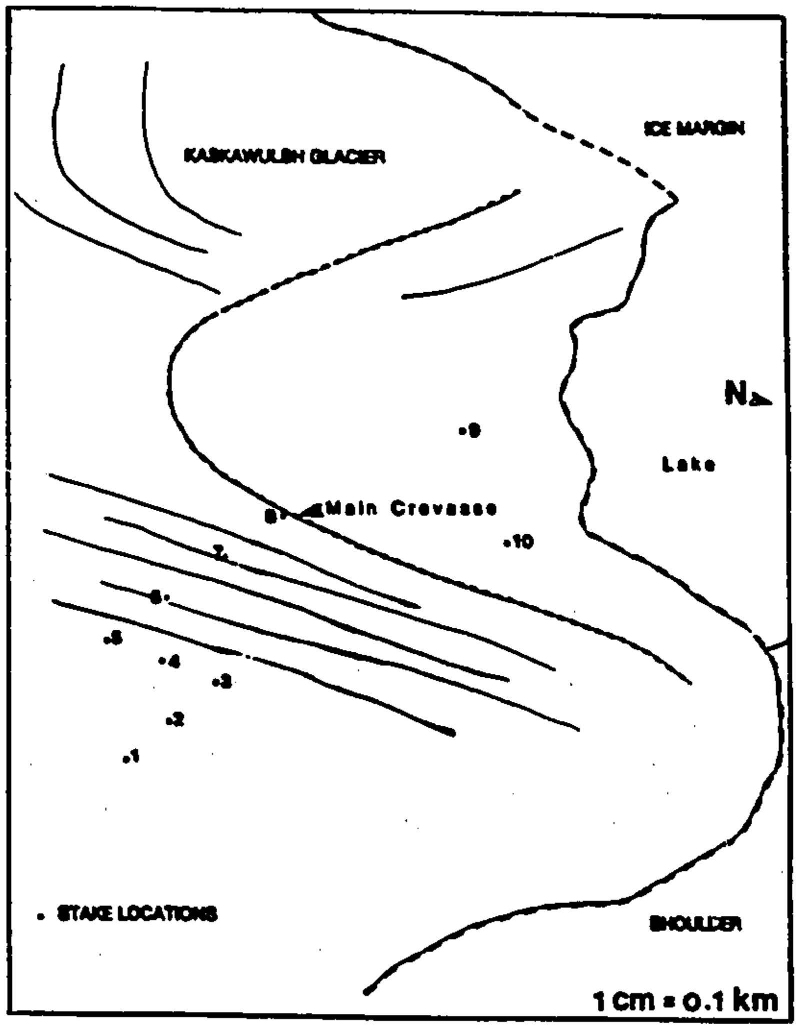


Figure 10: *Stake distribution pattern, Kaskawulsh Glacier.*

At the end of the summer, the total amount of ice melt was calculated for each of the ten stakes. This ranged from 2.33 m (stake 5) to 3.83 m (stake 9), with an average of 3.09 m. There was no snow cover at any of the stakes during the field season. All the stakes were located on the ice surface, not near either a valley wall or a moraine, and had a similar debris cover. There was no major topographic or slope orientation difference which could affect ice melt to any significant extent.

The water equivalent of the ice melt was calculated using a standard density of .9 for glacier ice. Thus, the average amount of ice melt was equal to 2.781 m of water loss. This amount was then multiplied by 18.2 km^2, the area of the glacier determined from aerial photographs which should, under normal topographic considerations, drain into the lake. Thus, 5,061 x 10^4 m^3 of water was the maximum possible contribution of the Kaskawulsh Glacier in terms of melt from the immediate area. However, some of this volume may have been lost due to evaporation or lost into the glacier's main drainage system, and water from further up-glacier may have been added through either marginal channels or englacial tunnels. These volumes were impossible to quantify. Thus the ablation figure calculated could only be used as an estimate of the volume of Kaskawulsh contribution to the lake.

At the beginning of the research project, it was assumed that the contribution from the Kaskawulsh Glacier could be quantified by measuring the basins' input and subtracting this volume from the stored volume in the lake. Any amount not accounted for by stream input could then be assigned as input from the Kaskawulsh Glacier. However, as will be seen later, the input exceeds the stored volume of water, rendering this quantification of the contribution from the Kaskawulsh Glacier impossible. Thus, while it was acknowledged that there was some contribution from the Kaskawulsh Glacier, it seemed more reasonable to not include it in the water balance calculation.

4.2.4 SUMMARY OF INPUT

A(g) was calculated from the measured ablation of the N.S.Glacier, and was then extrapolated to the other two cirque glaciers. The contribution to the water balance from the glaciers was calculated to be 1.74×10^7 m^3.

The precipitation, P, for the basin was calculated by measuring the amount of rain/snow fall at 4 stations and extrapolating this over the area of the basin. This was calculated to be 1.08×10^6 m^3 of water. However, most of the precipitation does not reach the streams. It is absorbed into the ground or evaporated back into the atmosphere. As a result, the precipitation input figure is somewhat deceptive.

The input from the Kaskawulsh Glacier (based on ablation alone) was found to be 5.06×10^7 m^3. However, as was stated previously, this volume will not be included in the final water balance calculation, so the variable Ab(Kask) is not quantified in the formula.

The variable M(b) is the amount of melt from the basin not attributable to melt off the glaciers. This encompasses snowbank and ice-cored moraine melt. This variable is quantified using a second equation, as follows:

$$M(b) = Q - \{A(g) + P\}$$

where Q is the discharge, measured over the summer as 3.51×10^7 m^3. This equation then solves as follows:

$$M(b) = Q - \{A(g) + P\}$$

$$M(b) = 3.51 \times 10^7 - \{1.74 \times 10^7 + 1.08 \times 10^6\}$$

$$M(b) = 3.51 \times 10^7 - 1.85 \times 10^7$$

$$M(b) = 1.66 \times 10^7 \ m^3$$

Therefore, the contribution of the snowmelt from the rest of the basin is almost as large as the contribution made by ablation of the cirque glaciers.

Putting this into our original equation, the input component of the water balance equation is:

I = A(g) + M(b) + P + Ab(Kask) 1

and Q = A(g) + M(b) + P...5

thus I = Q + Ab(Kask). 7

For the basin, input is 3.51 x 10^7 m^3 plus some non-determined amount contributed by the Kaskawulsh Glacier.

4.3 STORAGE CAPACITY OF THE BASIN

The holding capacity of the lake basin is important, since it establishes the storage capacity of the system, and provides a maximum volume which may be released once lake drainage has been initiated. The three main points to be addressed are the volume of the lake, the rate of water input, and the volume retention in the basin.

storage = input - output 3

with output being the continuous loss through either evaporation or leakage.

4.3.1 VOLUME OF LAKE BASIN

The topography of the lake basin was determined by post drainage surveying (figure 11). From this map, maximum volume and rate of volume increase were calculated using a planimeter. The total volume for the lake was calculated as 1.46 x 10^7 m^3. This, however, did not include the volume of water beneath the floating ice margin which was impossible to quantify accurately, but an estimate of 1.00 x 10^6 m^3 was made from the topography. This was determined from the assumption that the collapsed ice margin after drainage reflected the topography of the basin floor (Marcus, 1960; Sturm & Benson, 1985). An overall figure for the storage capacity of the lake was approximately 1.56×10^7 m^3 of water.

4.3.2 INPUT RATE

Input = Q + Ab(Kask)..7

where Q = {A(g) + M(b) + P}...5

The rate of input is the change in the amount of any or all of the above variables with time. All of the variables change in magnitude over the course of the season. The factors which control the changes for these variables will be outlined briefly and an analysis made of the changes in all of the variables on a weekly basis.

(A) KASKAWULSH

The input rate for the contribution from the Kaskawulsh Glacier is expressed as the ablation rate. Variations in the amount of insolation control the ablation of the Kaskawulsh surface, and this is directly related to fluctuations in temperature. The melt rates were calculated from the measurements taken at 3 or 4 day intervals throughout the season. Hence, the ablation figures give an indication of the possible amount of input from this source if the entire volume of meltwater actually entered the lake.

(B) PRECIPITATION

The input rate for precipitation into the basin is relatively small. Because of this, the rate will be examined as totals per week as well as averages.

(C) STREAMS

The stream input rate or discharge rate varies throughout the season, and is directly related to the temperature. Variations in mean temperature in the basin lead to variable discharge rates of the 3 proglacial streams.

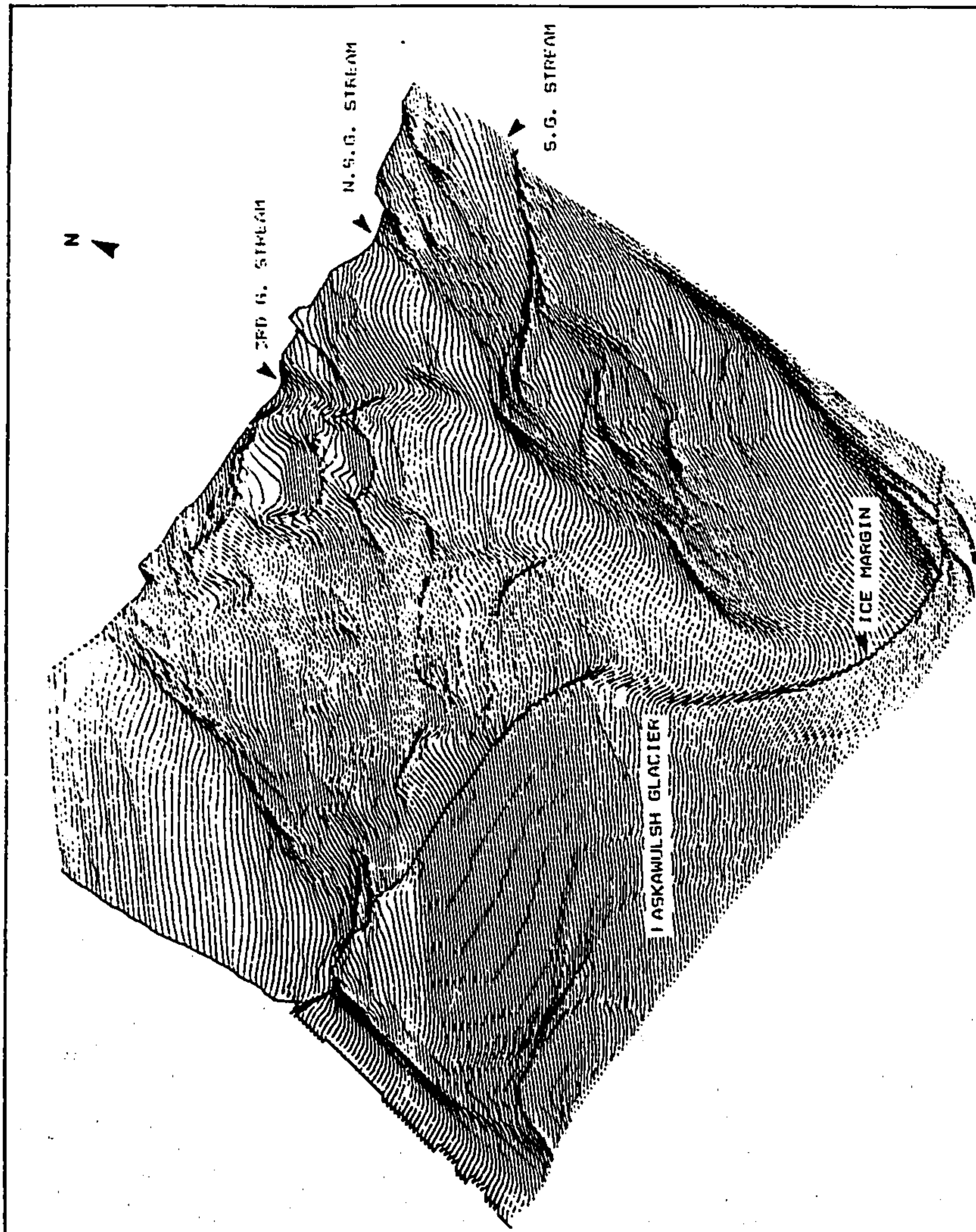


Figure 11: *Topography of the basin.*

Preliminary calculations of stream input were based on the data from the 3rd G. stream stage records and extrapolated to the other 2 streams. This was possible since point discharge readings taken indicate similarities in discharge for the 3 streams.

(D) WEEKLY ANALYSIS

This section examines the changes in the variables on a weekly basis. The numerical values which these changes represent are summarized in table 4.

JUNE

During the first week, June 10-16th, the rate of ablation on the Kaskawulsh was low, with a water equivalent of 3.24×10^6 m total. During this same week, the total precipitation input rate was 1.4×10^4 m^3, which equals 0.023 m^3sec^{-1} average. The stream input rate was also relatively low during most of June, averaging only 1 14 m^3sec^{-1} during this week. This is in response to the relatively low mean temperatures during this period, mostly below 4°C. The overnight temperature dropped below the freezing point, and the daytime highs never exceeded 7°C.

The week of June 17-23 saw an increase in the ablation on the Kaskawulsh to 5.38×10^6 m water equivalent. The precipitation input during this period totalled 6.87×10^5 m^3, occurring over 2 days, June 21-22nd. This was the largest amount of water added during any one week by precipitation during the 1987 season (before Aug. 12th). The stream input averaged 1.87 m^3sec^{-1} during this week. The mean temperature rose to peak at 7°C, then dropped back to 4°C with the 2 day rainy period. However, the early week temperature increase led to greater ablation of the Kaskawulsh and the small basin glaciers.

During the week of June 24-30th, the ablation on the Kaskawulsh increased to 7.45×10^6 m water equivalent. There was no precipitation during this week. The stream input was increasing, averaging 3.30 m^3sec^{-1} during this period. The high ablation and runoff rates were due to the increase in the mean temperature during this period to a peak at 11°C on the 30th.

JULY

The week of July 01-07 showed less ablation on the Kaskawulsh, totalling 5.82 x 10^6 m water equivalent for the week. There was 4,805 m^3 of water added by means of precipitation and 4.05 x 10^6 m^3 of water added by the streams at a rate of approximately 6.70 m^3sec^{-1}. The drop in the ablation rate followed a steady drop in the temperature during the week from 9°C to 7°C. However, the increase in volume being discharged by the streams is probably due to the flushing off of the surface of the small cirque glaciers. This can be seen in table 1, where all but stakes 1-4 were by this time snow-free. Another possibility could be the opening of some subglacial reservoirs and channels.

Between July 8th and 14th, the mean temperature in the basin continued to decline. However, the ablation rate on the Kaskawulsh increased to account for 7.28 x 10^6 m water equivalent, and there was 6.2 x 10^4 m^3 of water added by precipitation at a rate of 0.10 m^3sec^{-1}. Overall, the input rate increased despite the decreasing temperature. This was probably due to lag times between higher temperatures and high melt rates. However, the mean temperature stayed above 4°C and the night time lows never went below 0°C. Hence, it was still warm enough to promote melting within the basin.

The period of July 15-21st brought a marked increase in the mean temperature of the basin. The rate of runoff from the 3 basins rose to 9.43 m^3sec^{-1}, and there was 1.08 x 10^5 m^3 of water input during one day as precipitation. However, the ablation of the Kaskawulsh dropped to only total 5.83 x 10^6 m water equivalent. Hence, the overall rate of input rose slightly during this time.

The mean temperature for the basin fell and then rose again during the July 22-28th week. This is reflected partially by the drop in both the ablation and runoff rates. There was some precipitation, but the overall rate of input was down slightly during this period.

AUGUST

The following week, that of July 29-Aug. 4th, showed a slight drop in the mean temperature but a sharp rise in the ablation of the Kaskawulsh to 6.62 x 10^6 m water equivalent. The rate of input from the streams also rose during this time to 10.04 m^3sec^{-1}. The precipitation input totalled 9.1 x 10^4 m^3. The increased rates of ablation and runoff appeared to be delayed reactions from the temperature peak of July 27-29th.

August 5-11th was a relatively stable week in terms of temperature, with a slight drop mid-week which was recovered before the week's end. The ablation on the Kaskawulsh dropped to 4.22 x 10^6 m water equivalent. The stream input also dropped to 7.30 m^3sec^{-1}. There was no precipitation input during this period. The drop in rates of ablation and runoff while the temperature remained virtually constant reflected a shortening in the length of day, the number of daylight hours and therefore the amount of insolation received.

4.3.3 LAKE FILLING RATE

The filling rate for the lake was established from the recorded changes in the water level throughout the season, and proved to be variable. During the early part of the season, the filling rate slowly increased to 3.47 m^3sec^{-1}. on the 28th of June. This preceded the snowmelt runoff peak by 2 days. This peak rate of filling lasted for 3 days, then dropped as the discharge dropped on July 1st, down to less than a 1.16 m^3sec^{-1}. This paralleled the discharge curve until July 7th when the discharge once again began to rise. The lake filling rate did not start to climb again until July 13th, when the rate of filling rose again to almost 3.47 m^3sec^{-1} by the 21st. This high filling rate corresponded to a preceding peak in discharge. This was followed by another very sharp drop in the rate of filling lasting three days, then a four-day increase to peak at 3.85 m^3sec^{-1} on the 28th of July. This peak, however, did not correspond with a discharge peak, since the discharge was at a low point on that day. From July 28th until the 5th of August when drainage was initiated, the rate of

Table 4: INPUT RATES - 1987

WEEK	STREAMS m^3sec^{-1}	PRECIP. m^3sec^{-1}	KASK.GL. m w.equ.*	TOTAL RATE m^3sec^{-1}
10-16/06	1.14	0.02	3.24×10^6	6.51
17-23/06	1.87	1.14	5.38×10^6	10.77
24-30/06	3.30	—	7.45×10^6	15.62
01-07/07	6.70	0.008	5.82×10^6	16.32
08-14/07	6.27	0.10	7.28×10^6	18.32
15-21/07	9.43	0.18	5.83×10^6	19.07
22-28/07	9.18	0.19	5.03×10^6	17.50
29-04/08	10.04	0.15	6.62×10^6	20.99
05-11/08	7.30	—	4.22×10^6	14.28

* Kaskawulsh ablation figures are water equivalents.

N.B. Total figures include Kaskawulsh ablation data converted to daily flow equivalent rate.

filling decreased very rapidly.

The variable rate of filling was reflected in the rate of increase of the lake volume. The periods of high rates of filling are marked by more rapid increases in the water volume, and periods of low filling rate by slower volume increases, as would be expected.

The rate of filling curve reflects the changes in the rate of input. When comparing the graph of the lake filling rate with the graphs of each of the other input variables, it is noted that the shapes are similar, with peaks at generally the same times. Most noticeably similar are the filling curve and the stream input curve (figure 12). Some peaks are delayed or advanced by a factor of 2 or 3 days though. However, there is a wide gap between the actual curves on the graph, showing that much larger volumes of water are being input than are being stored in the lake, which indicates a large loss to the system.

A series of computer-generated 3-dimensional diagrams were prepared to illustrate the spatial extent of the lake at 1 week intervals during the filling sequence. The initial diagram, drawn up for June 8, 1987 shows the lake concentrated in the deep area up against the eastern half of the ice margin, and not extending into the stream channels or into the western part of the basin. By July 5th, the lake extended well up into the stream channels and across the western part of the basin against the ice face. However, the majority of the water body is still concentrated in the deep section of the basin. By August 4th, the maximum volume and spatial extent were achieved. The lake extended far up into the stream channels, to the point where a large part of their length was drowned. The lake also covered the majority of the western part of the basin towards the 3rd G. stream, and drowned most of the land from the eastern ice face extending back almost to the S.G. stream. A full set of diagrams are contained in Appendix B.

4.3.4. SUMMARY OF STORAGE

When comparing the lake filling rate with the input rate from the streams (Q), it became very apparent that there were major discrepancies between the 2 rates. It was assumed that as the rate of input increased, the rate of filling would increase accordingly, but this did not occur. The rate of input was higher than the rate of filling from the 28th of June onwards. The difference between the two curves ranged from 3.47 to 9.26 m^3sec^{-1}. The two curves mirror each other in terms of high and low points, but it is the magnitude which differs. At the snowmelt peak (June 30th) the discharge is around 7.75 m^3sec^{-1}, while the lake filling rate is only at 3.47 m^3sec^{-1}. The lake filling rate stays below or at 3.47 m^3sec^{-1} for the remainder

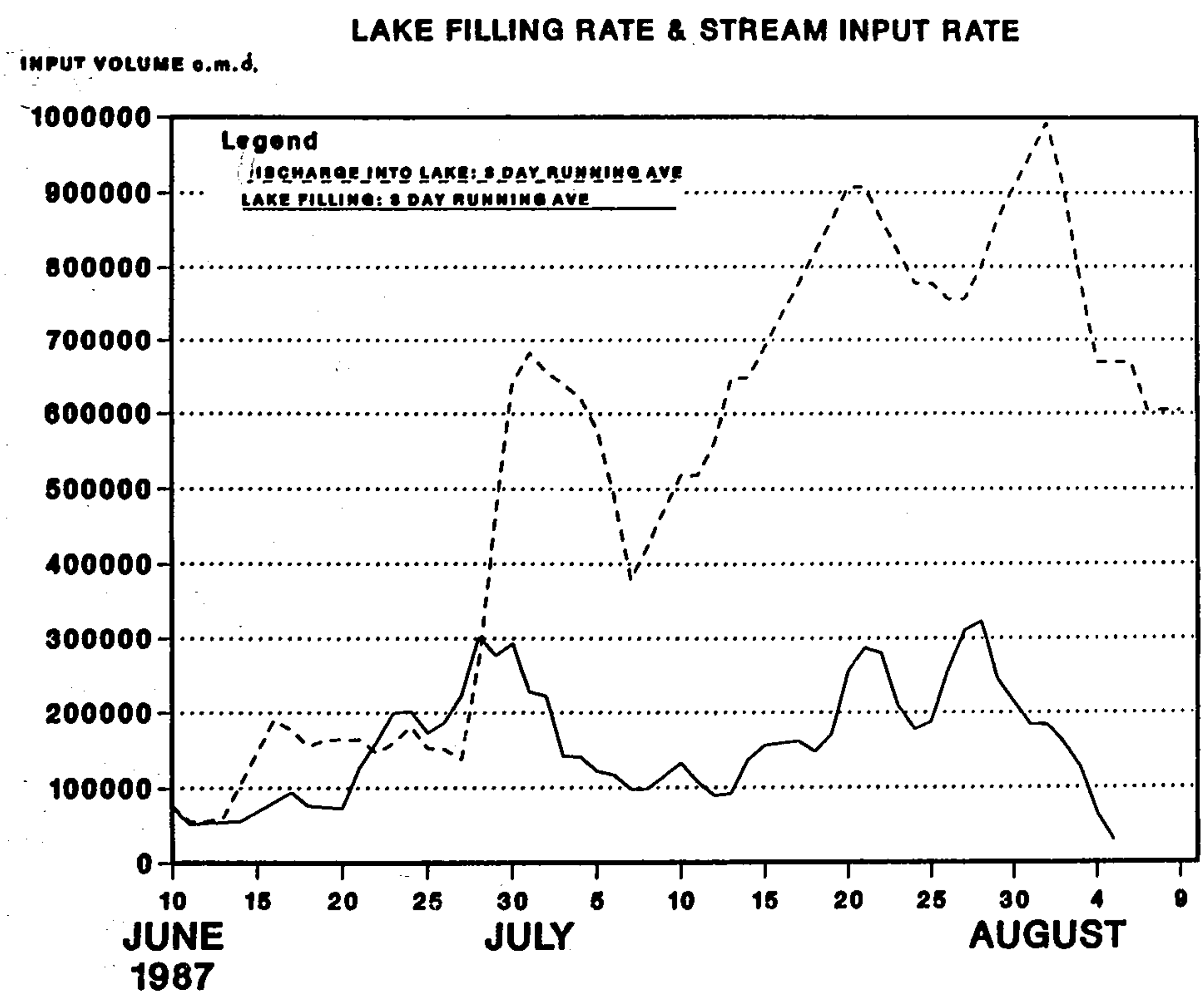


Figure 12: *Lake filling rate vs stream input rate.*

of the season, while the input rate continues to rise to a maximum of 11.57 m^3sec^{-1} on August 1st. While the discharge is rising to this peak (the icemelt peak), the lake filling level is in fact dropping from the 28th of July until the end of the season.

Some of the volume of water being input but not being measured as retained was in fact contributing to the floating of the ice margin. However, as was seen earlier, the estimated total volume of water retained under the ice face at the highest floating point was 1.00 x 10^6 m^3. This does not account for the difference between the volume input and the volume retained.

4.4 OUTPUTS

The outputs from the basin fall into three categories, leakage, evaporation, and drainage. The formula which establishes this is the following:

Output = f(Leakage, Evaporation, Drainage)..2

Evaporation obtained as a residual in a water balance equation is usually less accurate than if monitored by some independent technique such as a form of evaporimeter (Ferguson & Znamensky,1982). In this study, the evaporation was not monitored, so this variable cannot be accurately quantified, and as such is acknowledged as a contribution but discounted in the water balance calculations.

The leakage and drainage are both quantifiable, however. The amount of drainage is the stored volume in addition to the volume still being input during drainage. The leakage was calculated by the formula:

Output = Input - Stored Volume 3

4.4.1 LEAKAGE FROM BASIN

The amount of water being input (seen in section 4.2.4) from the streams was 3.51×10^7 m^3 of water during the season. The maximum volume of the lake was 1.56×10^7 m^3. Using equation 3, the amount of water being lost to the system before drainage can be calculated as 1.95×10^7 m^3.

It is necessary to mention here again that in addition to the volume input from the streams, there was some further amount of water input from the Kaskawulsh Glacier. Thus, the amount of input, and, as a result the amount of output before drainage, is in fact higher than portrayed here. The actual volumetric difference is not calculable, but this provides a minimum figure.

The rate of leakage varied during the season. The rate per week was calculated using the formula:

Output (t2 - t1) = Input (t2 -t1) - Storage (t2 -t1) 8

where t1 and t2 are defined as different time frames. From the weekly rate, this figure was reduced to a rate of leakage per second (figure 13). The rate of leakage was very low during the first two weeks of recording, that is the second 2 weeks of June. The rates were 0.50 and 0.61 m^3sec^{-1} respectively. These figures, though low enough to be experimental error, have been included in the overall evaluation of the leakage. By the last week of June, the rate of leakage had risen to approximately 5.55 m^3sec^{-1}, where it stayed for the next two weeks. The last 2 weeks of July saw an increase to approximately 6.95 m^3sec^{-1}. The final week before drainage of the lake was initiated exhibited an increase in the rate of leakage to 8.61 m^3sec^{-1}.

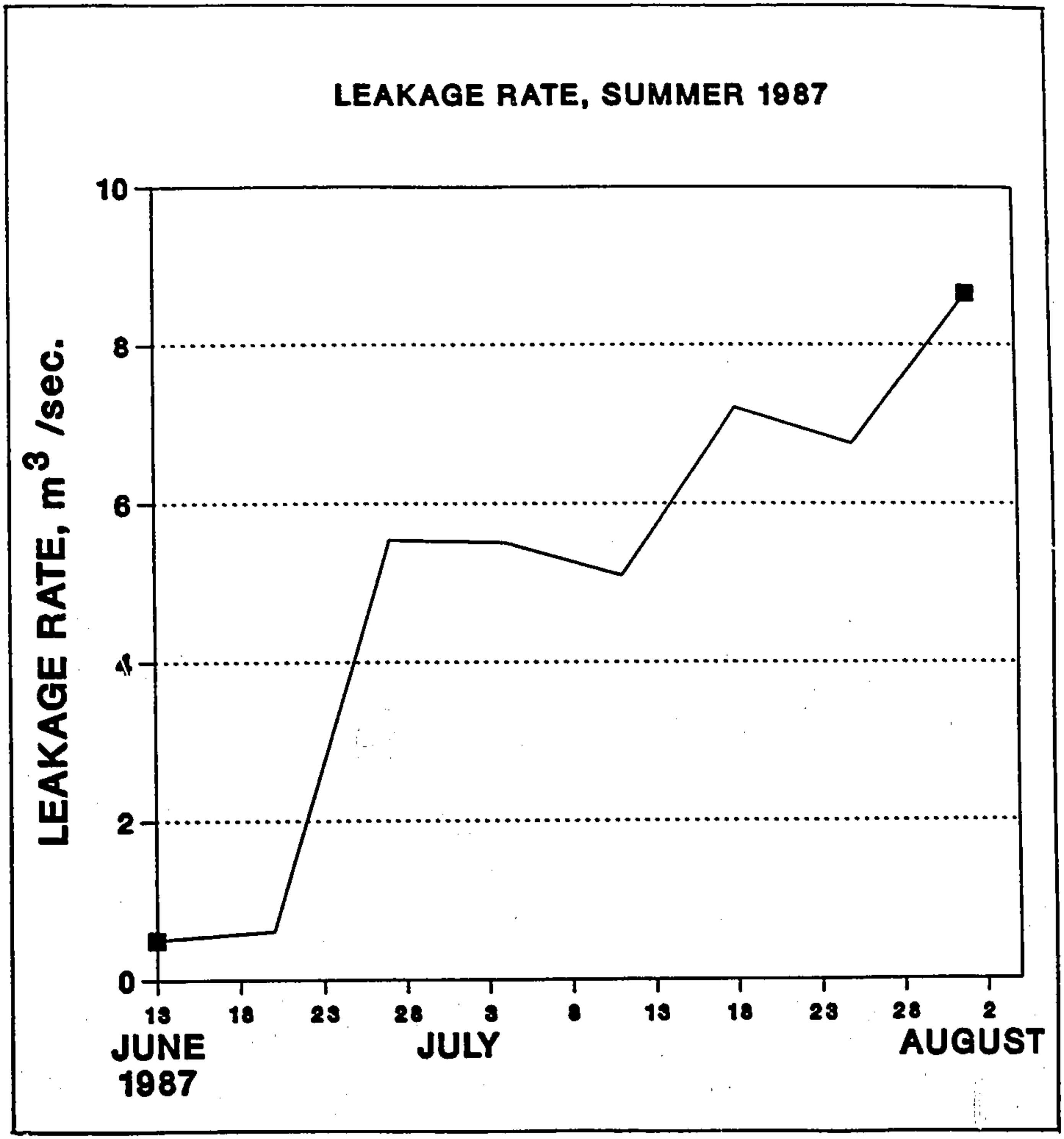


Figure 13: Leakage rates - summer 1987.

4.4.2 DRAINAGE

(A) TIMING OF DRAINAGE

The drainage of the lake occurs in late summer to early fall according to observer reports from the late 1960's onwards. Drainage is assumed to commence when there is a decrease in the water level. However, it must be noted that a drop in water level is recorded only when the leakage rate exceeds the input rate. In 1986, the drainage occurred between Sept 3rd and 10th. It was first noticed that drainage had begun Sept. 5th, 1986. The onset of drainage was observed by one of the local pilots on a regular flypast. During the 1987 drainage, it was noted that drainage had begun 2 days before it was noticeable from the air. It was therefore assumed that the 1986 drainage had also begun 2 days prior to it becoming apparent from the air. Thus, it is assumed that the 1986 drainage actually began on Sept 3rd. Further, the volume of water in the lake basin was roughly equivalent to that stored in 1987, so a similar length of time for drainage can be expected.

In 1987, drainage began on the 5th of August. This was not detectable from the air until the 7th . The lake was completely empty by early morning on the 12th, so a 7 day drainage period is concluded. This coincides with the 1986 information.

(B) VOLUME DRAINED

The total volume stored in the lake was 1.56×10^7 m^3. This included the measured volume of the basin of 1.46×10^7 m^3 and the estimated volume of 1.00×10^6 m^3 stored beneath the floating ice face. However, during drainage, a further volume of water was being input even as drainage was taking place. The input volume from the 3 streams for that week was calculated to be 2.38×10^6 m^3. The Kaskawulsh Glacier was also contributing some amount of water during this time, but again this volume was impossible to quantify. Thus, the total known volume to have drained during the period of Aug 5th to August 12th was 1.79 x 10^7 m^3.

(C) RATE OF DRAINAGE

The rate of drainage of the lake is exponential (figure 14). From the rate of 11.11 m^3sec^{-1} on the 5th of August, the rate climbed relatively slowly over the first five days, reaching only 41.67 m^3sec^{-1} by the 10th. This resulted in only one third of the standing volume of lake water being drained. The increments of increase were small and fairly regular during this time. August 5-6 saw an increase of only 0.97 m^3sec^{-1}, Aug.6-7: 4.44 m^3sec^{-1}, Aug. 7-8: 5.55 m^3sec^{-1}, Aug. 8-9: 8.33 m^3sec^{-1}, and Aug. 9-10: 11.94 m^3sec^{-1}.

The remaining two thirds of the lake water volume drained in the final two days, as the rate of drainage climbed to 62.50 m^3sec^{-1} by the 11th (an increase of 20.00 m^3sec^{-1} from the 10th), and to 78.89 m^3sec^{-1} by the 12th (another 16.67 m^3sec^{-1} increase). The progressive increase in the rate of drainage is a result of the constant widening of the tunnel by frictional melting (Mathews, 1973; Drewry, 1986). It was impossible to measure the size of the tunnel after drainage, since a 10 m wide section of the ice face collapsed into the basin. In addition, since the tunnels would have been located further back under the section of the ice face which was floating, they would not have been visible after the collapse of the ice face.

4.4.3 SUMMARY AND DISCUSSION OF OUTPUT

The drainage of the lake is the final stage in the apparently continual process of output from the basin. As a result of drainage and ice face collapse at the end of the season, the escape routes become blocked. The conduits within the Kaskawulsh Glacier continually deform and close under ice overburden pressure due to the forward movement of the glacier. This would be particularly effective after drainage, since very little or no water would be flowing through the passageways to maintain melting. As a result, the early part of the

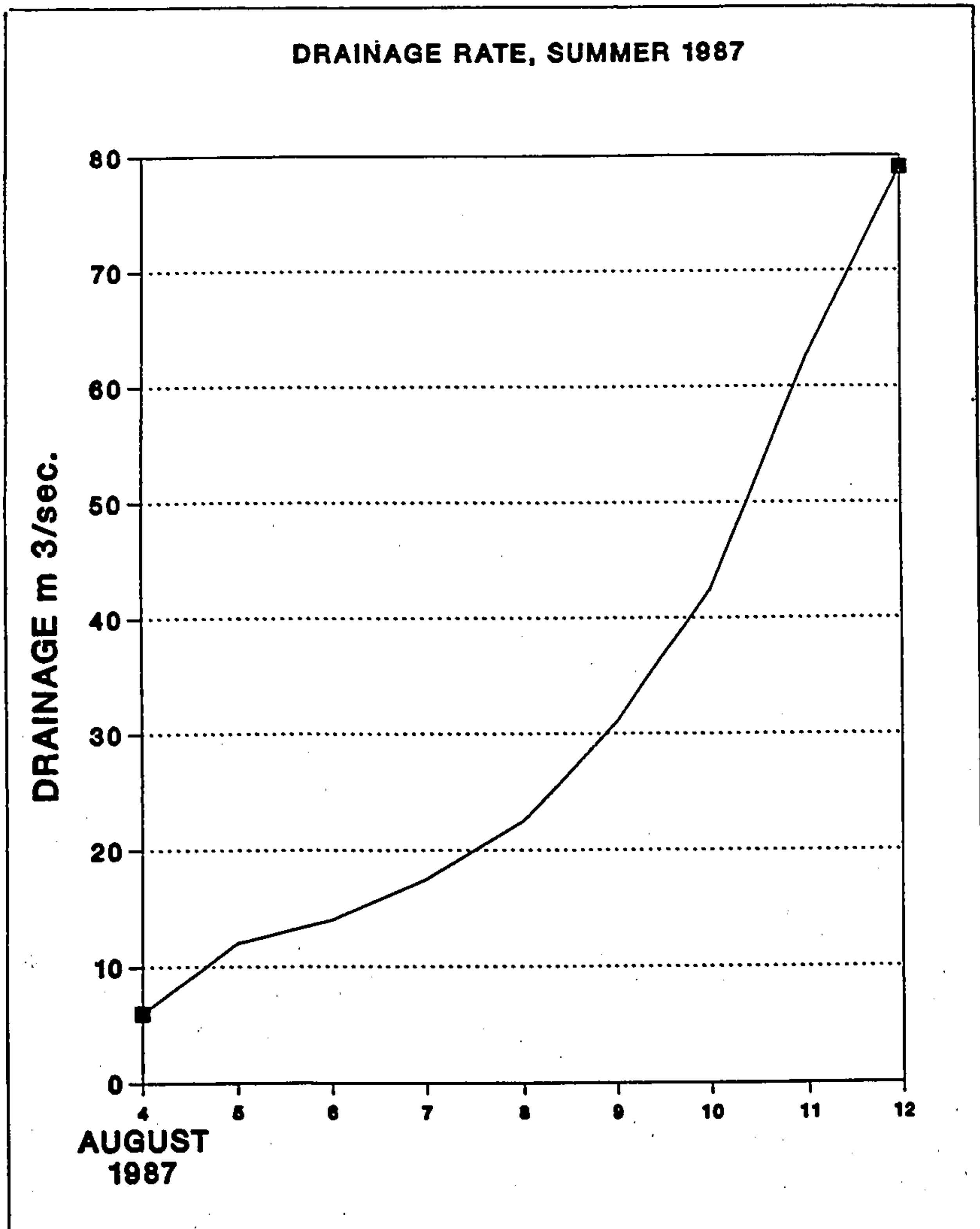


Figure 14: *Rate of drainage, summer 1987.*

following season is devoted to the slow exploitation of former or new channels. The presence of a slow leak bypasses the problem of initiating and forming a new tunnel each year (Blachut & Ballantyne,1976). It is suspected that the drainage system is a combination of Nye and Rothlisberger channels, supplemented by marginal floating and englacial tunnels.

The basal conditions are hypothesized as follows. A Nye channel exists in the bedrock beneath the ice face. This channel may be open year-round, or perhaps freezes up during the winter, or gets blocked by ice debris after the drainage. The channel, however, has a limited capacity, which restricts the amount of water passing through it and escaping the basin (figure 15a). This would correspond with the low rate of leakage experienced until the 20th of June.

A Rothlisberger channel develops above the Nye channel as the season progresses and warmer water is input into the lake. This allows a larger volume of water to pass out of the basin (figure 15b). This channel would be cut by the warmer water flowing along the base of the water column to the ice face. The increase in the size of the opening through which water can escape would correspond to the increase in the rate of leakage after June 20th, up to approximately 5.5 m^3sec^{-1} from the previous 0.5 m^3sec^{-1}.

In addition to the tunnel's continual enlargement due to the heat of friction from the draining water, the ice margin then begins to lift as the lake fills. This allows a larger amount of water access to the drainage passageway (figure 15c). It is also possible that this initial lifting detaches the ice margin from the bedrock. This lifting may be sufficient to allow the water to pass over a rock lip at the edge of the hanging valley into which the ice re-entrant flows (see radio echo sounding results; chap 5). This ice seal with the rock lip may have been restricting the water. This conclusion is supported by the increase of the leakage rate up to approximately 7.0 m^3sec^{-1} in the later part of the summer.

The final stage in this process is thought to be the point at which a connection is made between the passageways and cavities through which the water is draining and the main

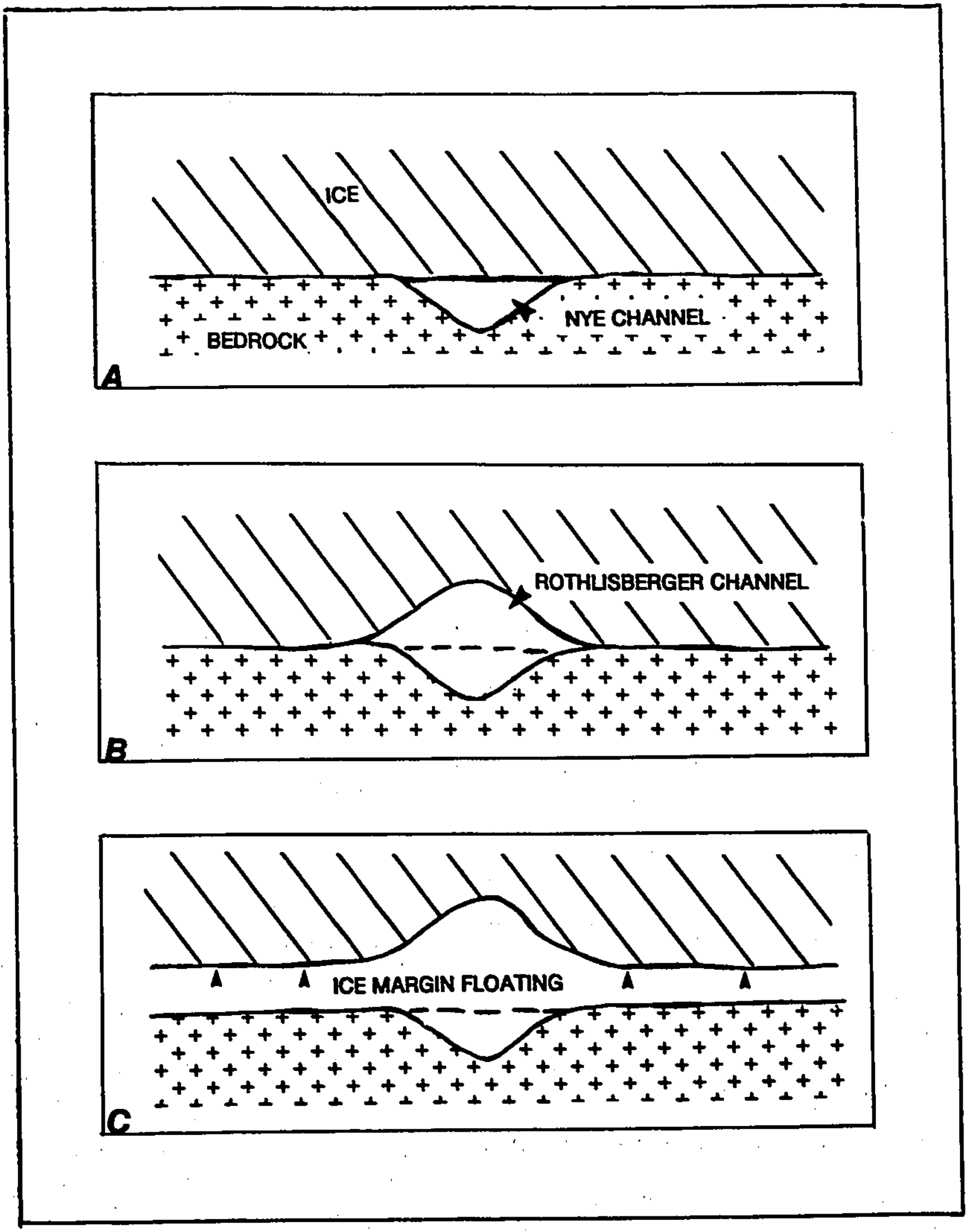


Figure 15: *Stages of the mechanics of drainage.*

channel network of the Kaskawulsh Glacier (Blachut & Ballantyne, 1976). Once this connection has been made, large volumes of water can be evacuated rapidly from the entire system. The opening would then be widened by the volume of water subsequently able to flow through. Thus, the opening of channels which connect to the main drainage network may be part of the triggering mechanism which initiates drainage, and drainage may finally be initiated by the connection being made.

The rate of increase of drainage is exponential (figure 14). The increased amount of water flowing through widens passageways cut through ice, since the larger volume of water passing through the channels means more accessible heat for melting. If the channels are cut through bedrock, some erosion may occur, but the rate of enlargement of the channel would be very slow, insufficient to account for the increasing rate of leakage throughout the season.

The configuration of the lake and retaining ice margin, a re-entrant tongue of ice, imply that while channels may be of the Nye type close to or just under the ice face, most of the length of the channels are cut directly into the ice. It is likely that the drainage conduits connect somewhere into the main glacier drainage network of the Kaskawulsh Glacier.

4.5 GENERAL DISCUSSION AND CONCLUSIONS

Input, output and storage all vary throughout the season. The input rate varies with the temperature and precipitation, since melting requires energy input. Storage is related to the rates of input and output as well as the size of the retaining basin. Output is a function of input and some other factors, thought to be size of the drainage channel, the rate of flotation of the ice margin and the timing of the connection with englacial channels.

CHAPTER V

ICE MOVEMENT AND MARGIN BEHAVIOR

5.1 INTRODUCTION

The retaining ice margin plays an important role in the processes of filling and draining of an ice-dammed lake. It is required in order for the lake to form, and it must be breached for the lake to drain. The drainage process is often tied in with movement and/or disintegration of the ice face.

5.2 GENERAL ICE MARGIN CHARACTERISTICS

The retaining ice wall, a re-entrant of the Kaskawulsh Glacier, is approximately 2.5 km long. The thickness of the ice varies across its length in relation to the topography of the hanging valley into which the ice intrudes. The ice is thicker at the eastern side of the ice margin, as the topography dips to form the deepest point of the lake. This area of the ice margin is marked by a series of arcuate crevasses extending back into the ice with a slightly south-east orientation. When the lake is empty, the sections of ice divided by the arcuate crevasses appear as a series of dropping steps (figure 16). As the lake fills, these sections float sequentially, until the front-most sections are elevated to the height of the back ones. The sections behind and to the sides of the crevassed zone are subject to high stress, as hydrostatic pressure forces the flotation of the front of these sections. If the ice does not break or crevasse, then the whole section flexes slightly until the volume of water can no longer exert enough lifting pressure (figure 17).

Figure 16: *Dropped ice face, June 5, 1987.*

The western side of the ice margin is generally thinner, except for isolated sections (figure 18). Most of it remains grounded throughout the filling sequence. There is little stress-induced fracturing of the margin in this section except for a few long crevasses running perpendicular to the ice face. These have their origin in the stresses induced by the flow of ice into and out of the basin as the Kaskawulsh Glacier flows downvalley.

The floating of the margin depends on hydrostatic pressure, and a water depth equal to 9/10ths the height of the ice face is necessary in order to initiate flotation. In the western section, there is not enough water even at lake-full to float the face. In the eastern section

Figure 17: *Uplifted ice face , July 1986.*

where the deepest part of the lake is located, the ice face is approximately 60 m high, thus requiring 54 m of water to float it. This depth of water is attained very late in the fall preceding drainage. At lake-full in 1987, the water depth at that point was 110 m deep, and only 15 m of that water depth was added between June 1st and August 4th. This exemplifies that the water depth needed to initiate flotation of at least the front sections is reached in the late fall preceding drainage. The volume of water needed to exert sufficient hydrostatic pressure to float the area thought to contain the 'seal' of the ice with the bedrock does not accumulate until well into the summer melt season.

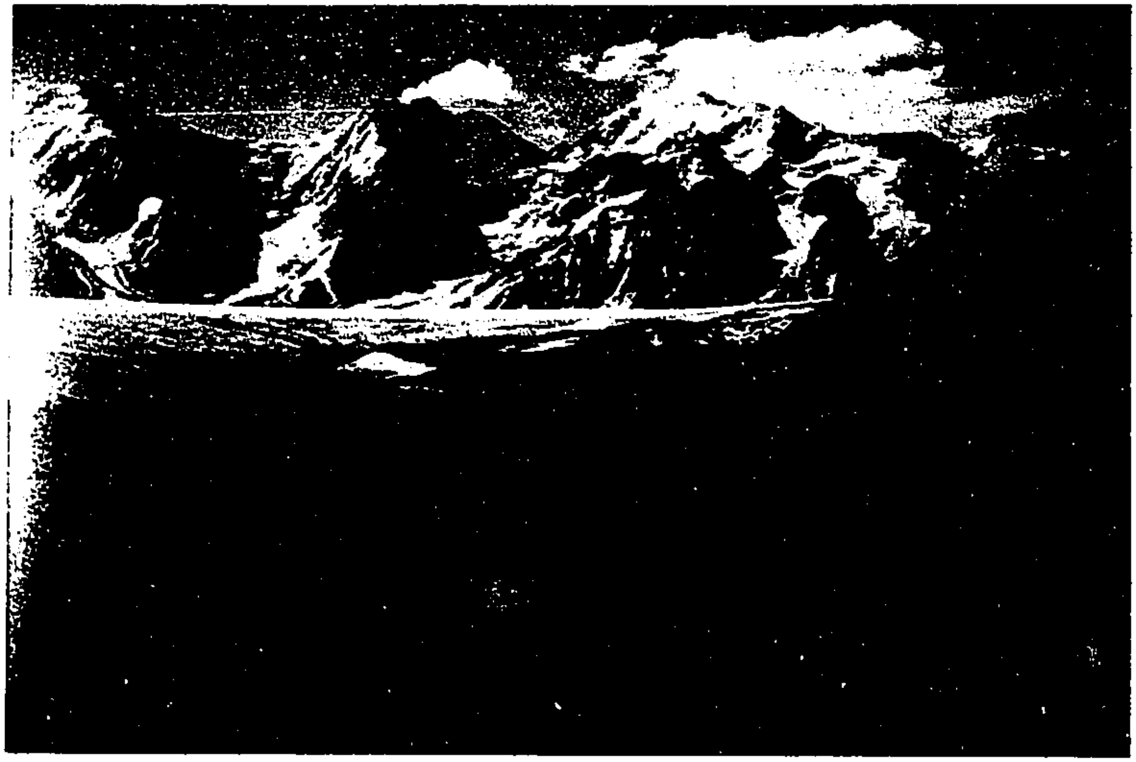


Figure 18: *Western ice margin.*

5.3 RADIO ECHO SOUNDING

A digital impulse radar was used to obtain data on the thickness of the ice and the basal conditions. The work was carried out by F.H.M. Jones of the University of British Columbia. Three profiles were traced across the ice re-entrant (figure 19). These profiles do not change much when adjusted for surficial topography, since this varies only slightly. Profile A (figure 20) shows a gradual increase in thickness from 125 m near the ice margin to 300 m at 800 m along the profile where the signal was lost. This loss of the signal is a result of the ice thickening beyond the design capacity of the instrument. Profile B (figure 20) runs parallel (relatively) to the ice face, and shows a depression in basal topography close to the

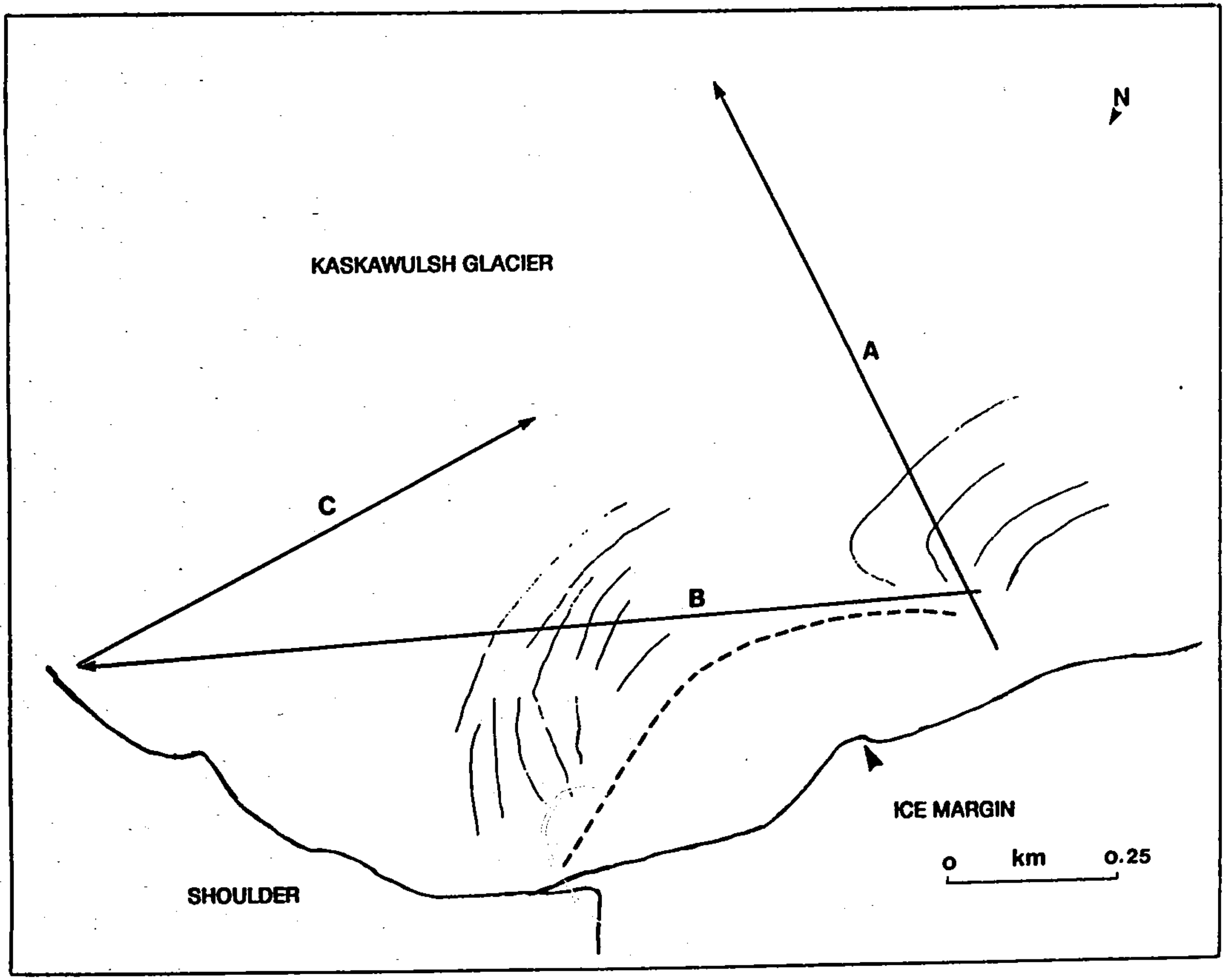


Figure 19: *Location of radio echo sounding profiles.*

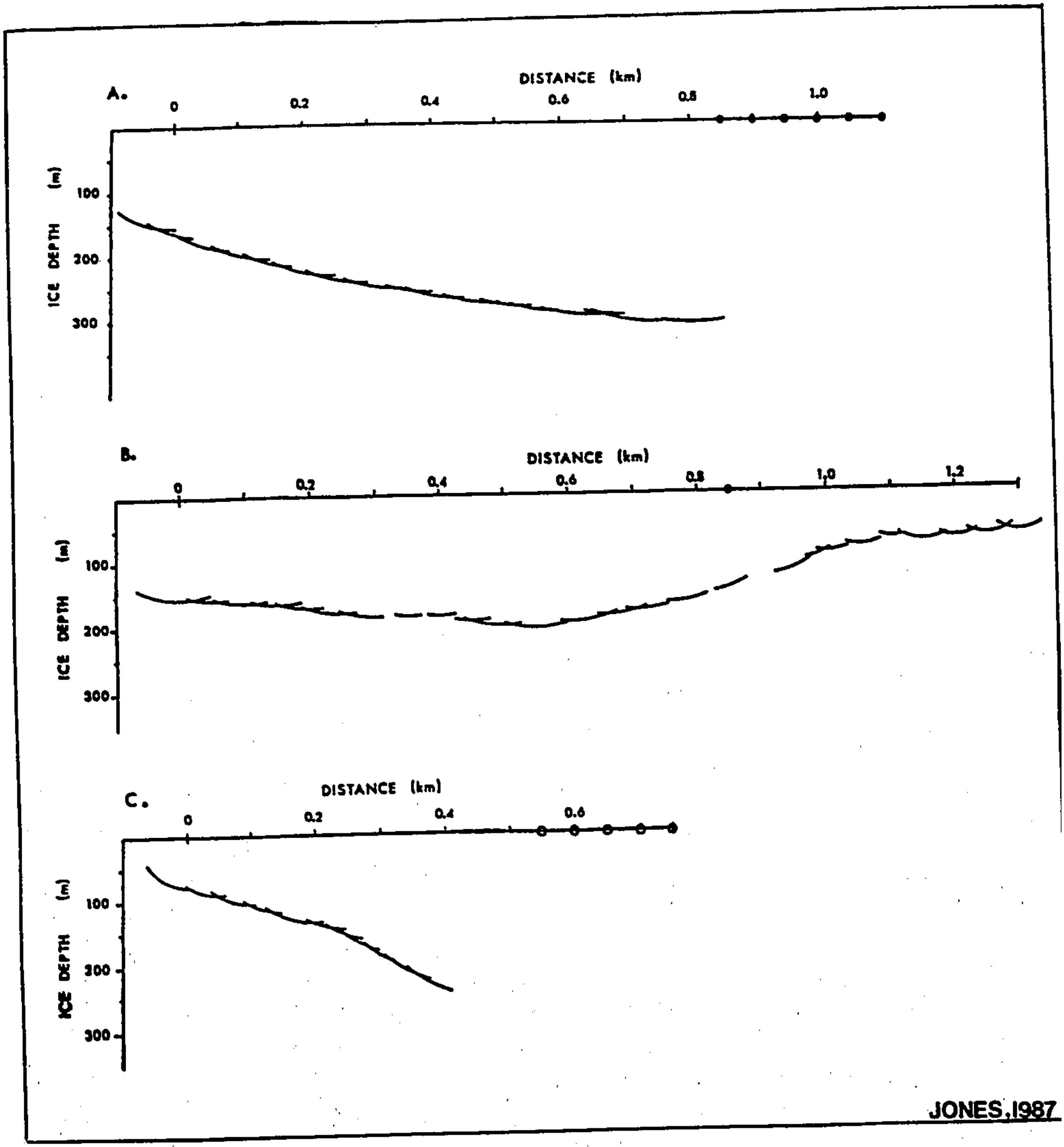


Figure 20: *Radio echo sounding data profiles.*

area where the drainage channels are thought to be located. The ice thickness in this area is approximately 200 m at its maximum point. The profile then shallows near the valley side to less than 100 m ice thickness (figure 20). The double signal seen on the oscilloscope trace in this shallower section may be caused by a water-rich basal zone (Jones, pers. comm. 1987)(figure 21). Water at the base of the ice would distort the return signal. This 'fuzziness' of the basal signal is also seen on profile C in the same area, which emphasizes the probability of a discrete water layer existing between the ice and the underlying bedrock. The third profile also extends beyond the signal identification potential of the equipment. The rapid thickening of the ice in profiles A and C at 800 m and 400 m distance from the ice face demonstrates that the ice margin is merely a re-entrant of the Kaskawulsh Glacier, extending into a hanging valley.

The radio echo sounding work was also performed to investigate if there were any Rothlisberger or Nye channels detectable at the base of or within the ice which could accommodate the evacuation of a large amount of water. The 50 m distance between soundings dictated by the nature of the survey precluded detailed resolution of the bedrock form. It had been planned to conduct the radio-echo sounding survey prior to lake drainage to assist in the determination of the water depth beneath the floating ice margin. This was prevented by early drainage and logistics problems. In addition, the ice margin was severely weakened by the drainage, rendering the ice face hazardous to work near. As a result, the ice thickness readings were taken back beyond the front section, and an estimate of the water depth trapped beneath the frontal section of the ice margin is not possible.

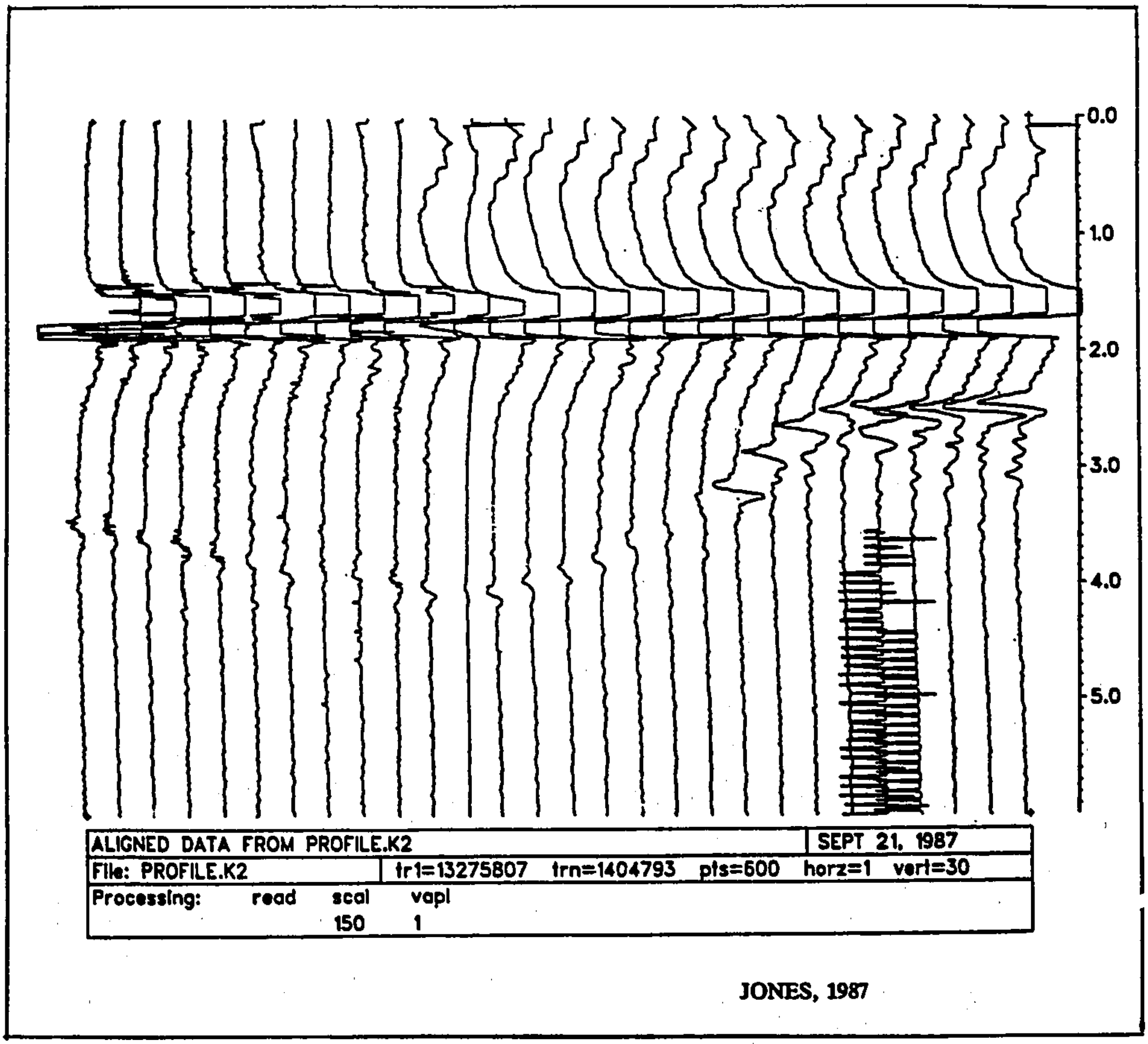


Figure 21: *Oscilloscope trace for profile B.*

5.4 VERTICAL ICE MOVEMENT

Ten stakes were drilled into the Kaskawulsh Glacier in a pattern extending back from the ice face (figure 10, Ch 4). These were re-surveyed every three days to monitor any lifting of the ice margin. Two of these stakes were positioned on the heavily crevassed section near to the ice face. The remaining eight stakes extended towards the center of the Kaskawulsh Glacier beyond the main dividing crevasses.

The survey showed a rise of the ice during lake filling and collapse and settling during and after lake drainage. Indirect evidence of movement was also provided by photography and observations. At the start of the 1987 field season, the winter lake ice was frozen to the ice face. By June 14th, the ice uplift was evident in the slight deformation of the winter ice frozen to the margin. On June 19th, the amount of lift of the winter ice off the lake water surface was approximately 1.5 m (figure 22). The greatest uplift and fall of the ice margin was expected and recorded at stakes 9 and 10. The total amount of uplift between June 11 and Aug 4 was 25.7 m at stake 9 and 24.0 m at stake 10. The drop of the ice margin during and after drainage was 25 m at stake 9 and 34 m at stake 10. It is deduceable from this that some of the early uplift of the ice margin occurred during the initial refilling stages of the lake in the late fall before freeze-up, since the total summer's rise is less than the amount of collapse.

It was hypothesized that the major arcuate crevasse system separated a section of the glacier margin which floated freely from the main mass of the glacier which may or may not have remained in contact with the bed. The survey of stakes 1-8 demonstrated that this section does rise and fall with the filling and draining of the lake. The total amount of displacement decreases with the distance away from the ice face. Stake 8, located at the top of the arcuate crevasse, rose 15 m and fell 10 m. Stake 1, located between 250 and 300 m back from the ice face, was measured to rise 7.0 m and fall 2.25 m.

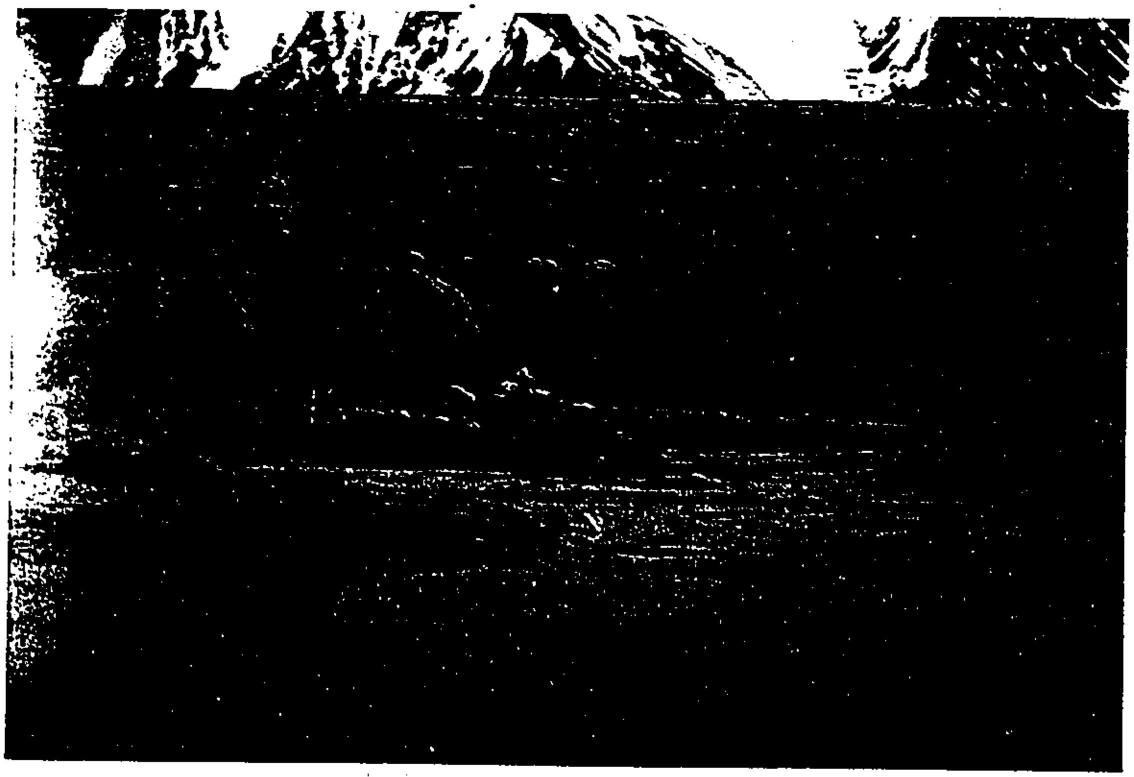


Figure 22: Uplift of winter ice off lake surface, June 19, 1987.

However, when examining the graphs of the ice uplift (Appendix C) it is apparent that the first big increase in the height of the ice is possibly experimental error. This is thought to be because the 6-8 m rise shown on the 18th of June is standard across all stakes, which would seem to indicate some amount of equipment placement error rather than strictly actual movement. The problem of equipment placement error is caused by the survey location. Due to the nature of the area, it was impossible to find a bedrock base on which to establish a permanent survey point. Instead, a set point was delineated on the gravel surface of the moraine, which was reasonably stable. Nonetheless, the non-permanent surface of the survey location tended to cause minor shifts in the placement of the equipment, which registered as changes in the height of the stakes being surveyed. Uplift of the margin later in the season is gradual, and the amount differs with distance from the ice face. Thus, any identical amount of change registered at all stakes is probably an error. Some of the fluctuations in the curves may also be designated as error. Small errors in the instrument set-up can falsely record large amounts of change. However, the graphs do indicate a general trend of ice uplift, more pronounced close to the ice edge. Close examination of some of the graphs permits the interpretation of the dynamics of various parts of the ice face throughout the season.

The section of the margin on which stakes 9 and 10 are located is disconnected from the main ice mass for the most part, except for seasonal freeze-up of the crevasses and along the margins of the arcuate area. In the early part of the season, the front section floats as a whole, and entrains the back section with it. Disconnection of the winter frozen sections is thought to occur around the 20th of June, after which the front section floats nearly independently throughout the rest of the season. These sections are predominantly distinct masses of ice, separated by multiple crevasses and fractures, and they respond individually to hydrostatic pressure, according to their own mass/density balances. Because these blocks are fully flotational during the whole season, they were immediately responsive to the onset of drainage and began collapsing. Stake 10, the furthest forward of the stakes, collapsed to a point 10 m below the height at which it was originally surveyed. This is an indication that a fair amount of uplift had occurred in the previous fall, prior to winter freeze-up.

Stake 8 is located immediately above the major crevasse. The graph of its movement shows an early increase then settling of this section of the ice margin. As previously stated, this rise is probably a combination of measurement error and entrainment of this section of the ice margin by the floating of the frontal sections. Beyond July 1st, the graph demonstrates the slow and steady uplift of the margin induced by increasing water level and increased hydrostatic pressure. A certain amount of collapse is evident during and after drainage, but the stake does not return to the measured starting point, assumed to be the point of grounding. This can be explained in two ways. The first explanation could be that an amount of water and ice debris was trapped beneath the margin, impeding the full settling of the ice onto the bedrock base. However, since the ice thickness in this area was found to be 125 m, a considerable amount of overburden pressure would have been being exerted, which puts into question the validity of this explanation.

A second, more likely explanation, could be that the ice in fact did ground post-drainage, and the 5 m difference is a mathematical accounting for the error in the original rise, mentioned earlier. This would suggest that the original uplift, measured (perhaps erroneously) as 8 m (stake 8) may have been only 3 m. If this is the case, then the graph of movement becomes more realistic (see corrected lines, Appendix C).

Stakes 1-6 show a reasonably smooth uplift through the season, with an increase in magnitude with greater proximity to the lake. Stakes 6-8 show clearly the effect of the increase in lake filling rate between July 20th and 30th, whereas further back into the glacier (stakes 1-5), this effect is mostly damped.

The coincidence of the stake movement with the lake filling and draining indicates that this is a lake-driven effect, rather than a seasonal thickening of the Kaskawulsh Glacier below the confluence of the North and Central Arms. The comparison was carried out using stake 8, since this location was thought to be indicative of the dynamics of the margin, without being either damped by being too far into the margin, or being not representative because of partial or total disconnection. The filling rate and the ice face movement generally mirror each other (figure 23). The most noticeable part of the season for this is from July 15th to 30th. Both the filling rate and the ice movement increase rapidly and fairly dramatically. The ice movement curve also follows the form of the lake water level rise curve (figure 24). The difference between these two is that the water level rises somewhat faster than does the ice. This is to be expected, since the stake being used for comparison is not floating freely, but is attached to the main ice mass, and thus is not free to respond as readily to increased

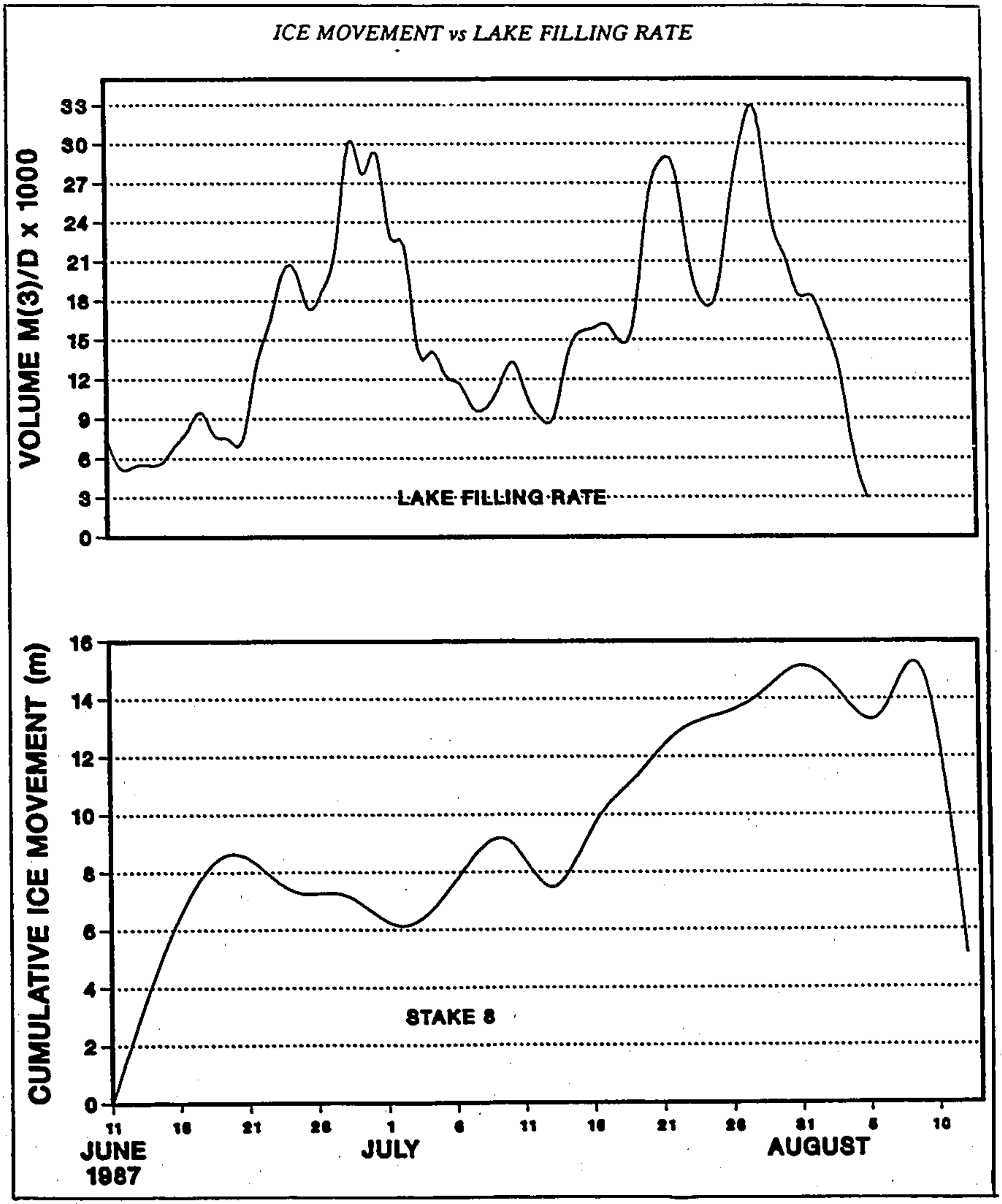


Figure 23: *Vertical ice movement vs lake filling rate.*

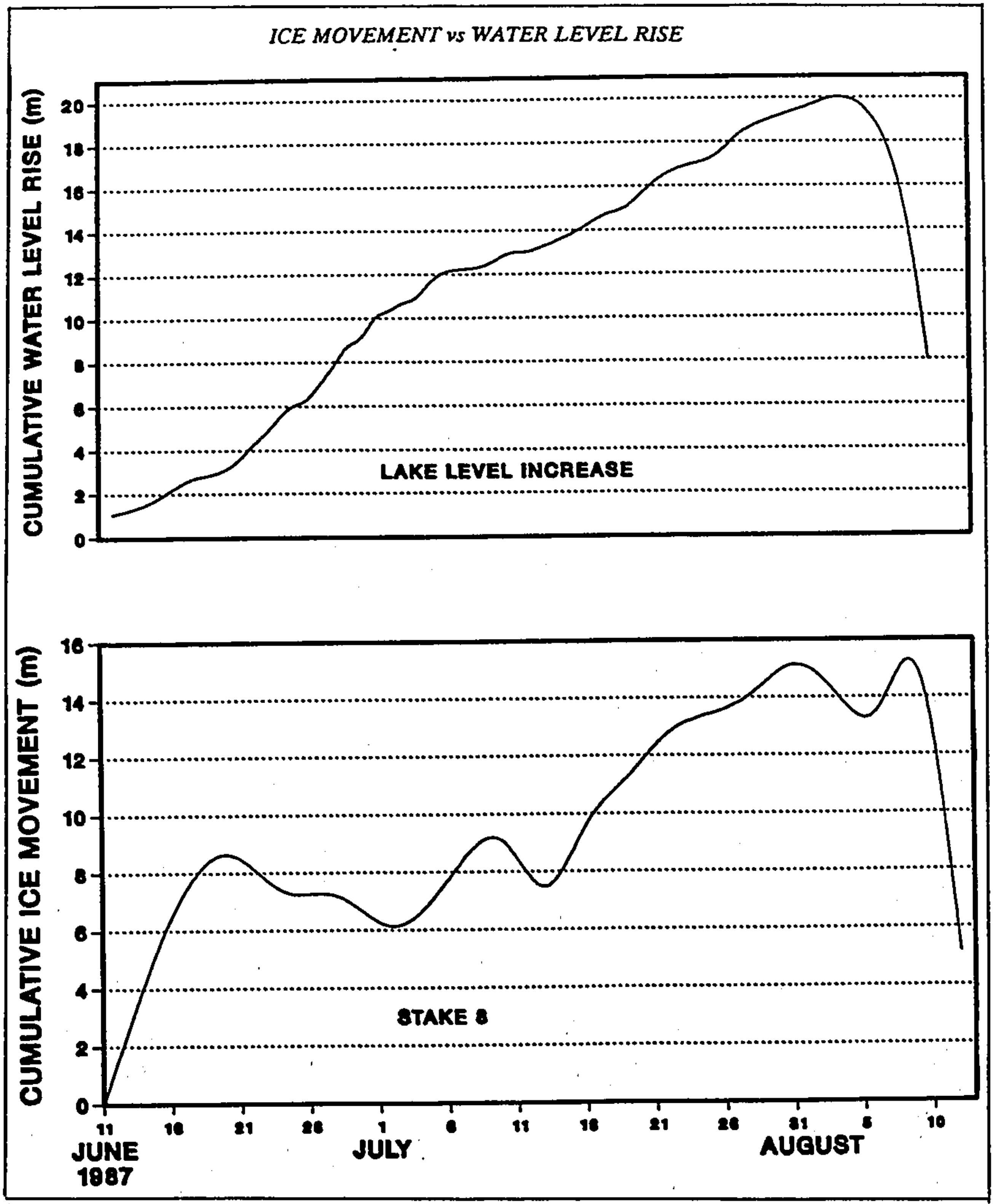


Figure 24: *Vertical ice movement vs water level rise.*

hydrostatic pressure. Examination of the movement of stakes 9 and 10 in comparison with the water level curve would have shown almost identical movement. This is proof that the ice movement is driven by the lake level dynamics rather than seasonal thickening of the margin.

Some small amount of uplift can be explained by summer thickening of the ice mass, since this is a known occurrence with movement of a glacier below an icefall. However, this effect would be most prominent in the stakes farther out into the main ice stream (stakes 1 - 5), and would not have affected the stakes experiencing the most movement.

5.5 PHOTOGRAPHIC EVIDENCE

For a qualitative check on the ice margin behavior, a series of photographic panoramas were taken throughout the season (Appendix D). They demonstrate very clearly the ice margin uplift. The sunken appearance of the amphitheater like section of the margin (June 5th) was modified slowly until there was a minimal height difference across the crevasses by July 23rd. A photograph taken on August 12th after drainage shows the collapsed ice margin. The front sections of the margin can float easier than the back sections because they are mostly disconnected from the main ice mass, which has to deform as the margin floats.

The annual floating and collapsing cycle of the ice margin produces stress fractures in the ice, resulting in the breakup of the glacier. Icebergs in the lake in 1986 indicated some margin breakup in 1985. Photographic evidence after the 1986 drainage shows very little ice calving (figure 25), which was confirmed by the presence of only a few icebergs in 1987. The drainage in 1987 was characterized by the breaking of large sections of the ice margin. A large section broke off on July 22nd, prior to drainage, and the ice face retreated by disintegration by more than 10 m during drainage. The bottom of the basin was not accessible after drainage due to the amount of ice lying in the lake bed, and the very unstable nature of the face itself

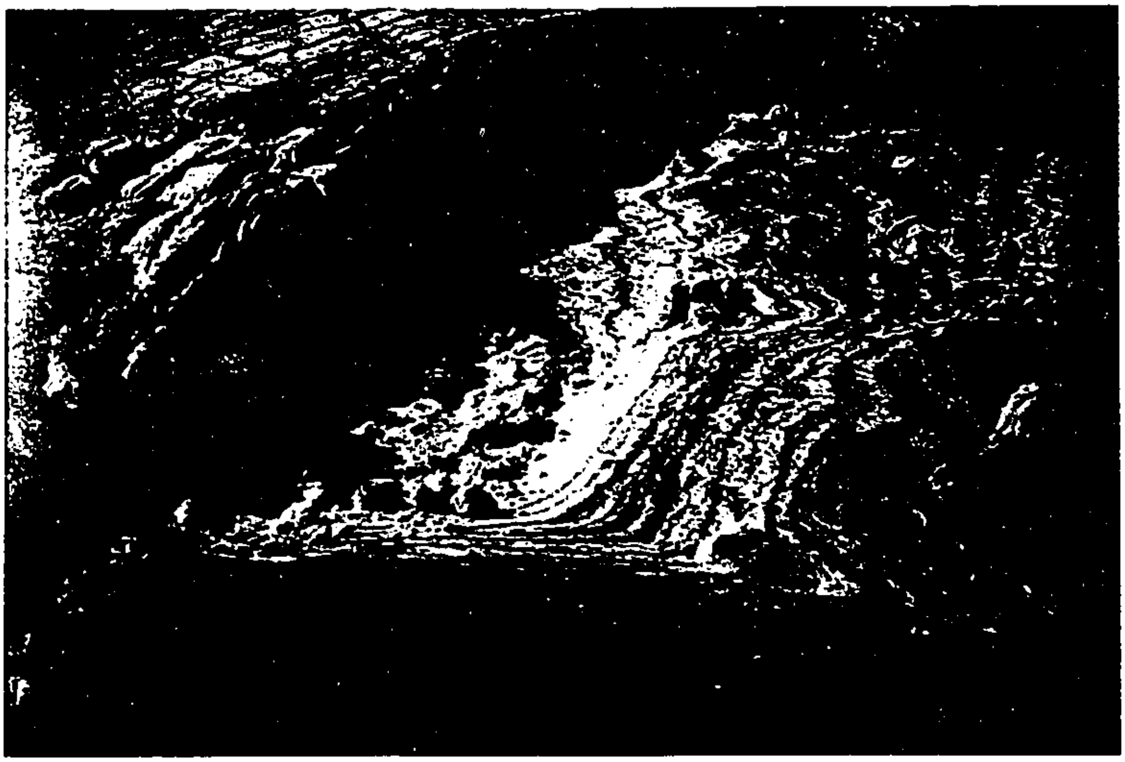


Figure 25: *1986 Drained lake bed - little ice in basin.*

(figure 26). The lake in 1988 was choked with icebergs.

The loss of sections of the ice face due to repeated stressing from several years of floating and collapsing is evident in the air photographic record. Air photos taken post-drainage in 1951 (A13233-37) show the empty lake basin filled with icebergs. The amount of ice is similar to the amount in 1987. Post 1972 drainage coverage (A22999-48) shows very little ice lying in the empty lake bed, slightly more than in 1986 but much less than in 1987. The 1977 air photos (A24763-19) show the lake pre-drainage, so no information can be derived from them. Information provided by a local pilot concurs with the idea of a pattern of

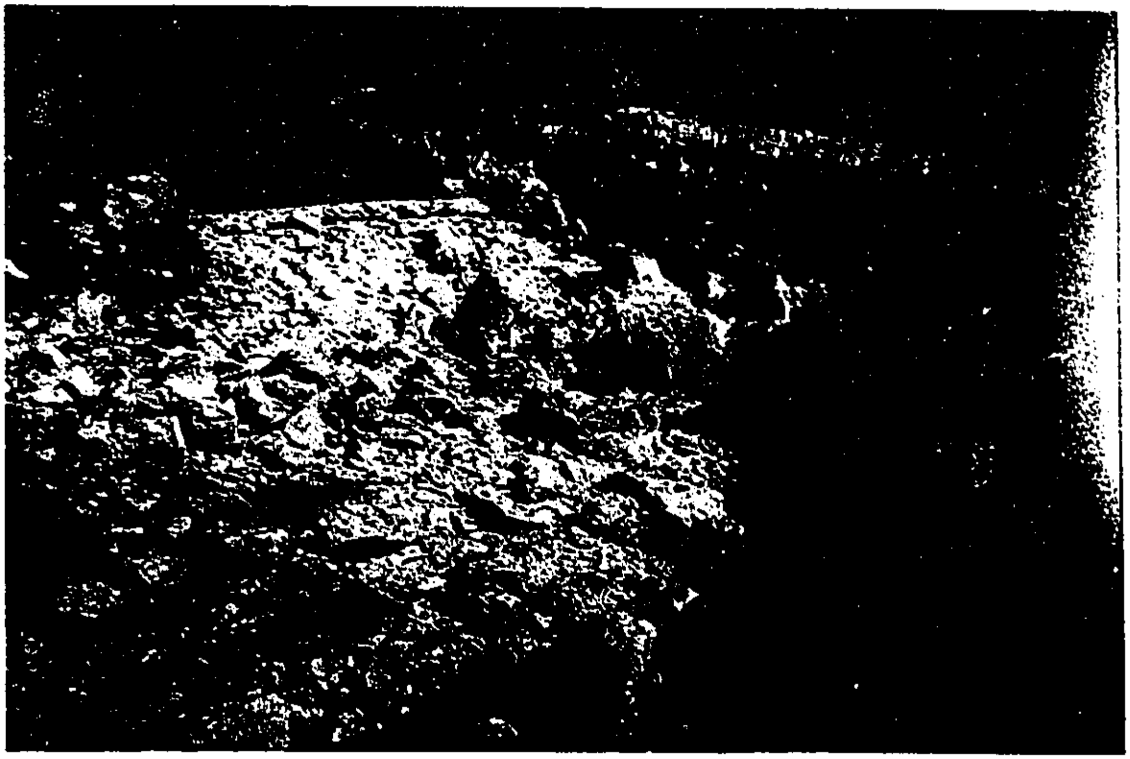


Figure 26: *1987 Drained lake bed - major ice face disintegration.*

years of little break-up intermixed with years of major ice face disintegration. Thus it is apparent that more than one cycle of lifting and falling are necessary to render the ice margin unstable enough to disintegrate substantially during drainage.

The cyclic pattern of years with major ice margin disintegration followed by years with little or no calving is most likely attributable to the movement of ice into the valley. The margin, a re-entrant of the Kaskawulsh Glacier, is constantly being replenished by that glacier. Surface flow rates for the Kaskawulsh Glacier reach a maximum of 152 m/yr for the Central Arm and 215 m/yr for the North Arm (Anderton, 1967). Anderton (1967) found an accelerating velocity profile measuring 137 m/yr at a point 1800 m downglacier from the nunatak along the longitudinal axis of the medial moraine (Loomis, 1970; after Anderton 1967). This point is close to being directly across from the lake, though approximately two miles out. The flow rates for the center of the Kaskawulsh Glacier are relatively high. Those near the margins would be reduced due to friction with the valley sides.

Following two or three years in which ice is being thrust into the basin by the movement of the Kaskawulsh Glacier, a large amount breaks off during drainage. This event is then succeeded by further pushing of ice into the basin. This constantly renews the volume of ice extending into the basin. In accordance with this, there is little evidence of general retreat of the glacier margin. The air photos taken in 1972 and 1977 place the margin at virtually the same position it occupied in 1951. On-site examination indicates that the present location of the ice re-entrant is the same as that delineated by the 1977 air photos.

5.6 CONCLUSIONS

The floating of the ice margin has been found to be only partially responsible for the onset of drainage, (as discussed in chapter 4) since the surveying demonstrates the date of initiation of flotation as being early in June, and the lake does not begin draining until August 4th. Marcus (1960) attributed the retaining of water to a seal just at the edge of the heavily broken zone. On the Kaskawulsh Glacier, this would be approximately at the location of stake 8, just beyond the major arcuate crevasse. However, the data demonstrates that the ice lifts beyond this point long before drainage takes place. This does not mean that the ice does not play some role in the initiation of drainage, rather that it may not be the sealing mechanism. As discussed in chapter 4, it is thought that the lifting does account for an increase in the rate of leakage during the filling sequence (approx June 24th). It is hypothesized that drainage is initiated by a connection being made with the Kaskawulsh

Glacier's drainage system.

The amount of collapse of the ice face after drainage affects the preservation of the tunnel openings beneath and within the ice face. If the tunnels are completely destroyed and/or infilled with debris, the possibility of their being exploited and re-used is reduced. However, leakage throughout the season seems to indicate that some conduits are established early. These channels must be local only, that is leading into the surrounding ice mass, since they do not connect to the main glacier's network. If they did, drainage would be occurring throughout the season, since the water would flow directly out of the basin. The process of tunnel widening, which allows the drainage rate to increase exponentially, would take place in these channels as well, and the basin would not retain any water. The high rate of ice movement in the main part of the Kaskawulsh Glacier ensures that its drainage network is deformed and closed due to stress and overburden pressure. This suggests that the connection of tunnels within the ice mass to the glacier's main drainage network is the trigger for the lake drainage. The connection to the main drainage network is important, since it is unlikely that the water funnels down the rock wall and below a recorded thickness of approximately 1000 m of ice at this point of the Kaskawulsh Glacier (Dewart; 1968). However, this is impossible to test, since the Kaskawulsh Glacier is 30 miles long from the lake to the resurgence point at the terminus, and the main drainage network is subglacial.

The ice is significant to the mechanics of the lake, in that it does affect the retention of the water. However, studies of the ice margin could be better used to understand the dynamics of such a body of ice. The examination of the ice margin has helped to understand the interactions between the ice margin and the lake, in addition to providing information on the thickness of the ice, which allows the confirmation of the nature of the ice re-entrant.

CHAPTER VI

SEDIMENTATION: PROCESSES AND DEPOSITS

6.1 INTRODUCTION

In a periodically or annually draining ice-dammed lake, the sedimentation pattern is different from that of a normal proglacial lake. One of the major factors which affects the depositional regime is the fluctuating water level. Several other factors may affect the processes of sediment distribution throughout the lake, either directly or indirectly. The ice mass in contact with the lake adds large volumes of water at 0°C to the lake, which controls the thermal regime and in turn affects the density contrasts at the inflow positions of the streams. Secondly, sediment input via englacial or subglacial passageways is deposited in different patterns from that of non-contact stream deltas. The currents generated in these passages also produce flows in different directions within the water column, often disrupting the flow patterns of subaerially originating stream underflows. Thirdly, icebergs calving off the ice margin can carry sediment across the lake basin, can disrupt the flow patterns within the water column, and can destroy facies by scouring if they run aground. Lastly, the ephemeral nature of ice-contact lakes often means that a large amount of sediment gets redistributed, sedimentary sequences form incompletely and any deposits found are often distorted due to the fluctuating water levels (Ashley, Shaw & Smith, 1985).

6.2 SEDIMENT INPUT

The sediment input from glacier streams is extremely variable, depending on such factors as time of day, time of season, bedrock type beneath the glacier, amount of movement of the glacier, amount of morainic debris the stream must travel through, and others. These factors combine to ensure that the amount of sediment being carried by the streams cannot be accurately predicted, but may be monitored so that general trends may be delineated.

The sediment input from the three streams was measured during the summers of 1986 and 1987, as part of an ongoing research project being carried out by Dr. P.G. Johnson. The technique used to collect the samples was outlined in chapter 3. Samples from a few days were chosen to provide data on diurnal variations and seasonal changes. These samples include ones from the peak runoff period, as well as early season (low flow). In total, 3 days were highlighted for analysis for the Surging Glacier stream with two of these having both morning and evening samples, and four days were selected for both the Non Surging Glacier stream and the 3rd Glacier stream with two days including morning and evening samples. Bar-graphs illustrating these results are contained in appendix E, but the data is also presented in tabular form (table 6).

The author acknowledges that 5 or 6 samples are statistically insignificant in order to draw any firm, accurate conclusions as to changes in sediment concentration patterns on a seasonal or diurnal basis. However, these samples provide an indication of conditions at several different stages of the seasonal discharge, which allow some basic observations to be made.

Results from the sampling program show that the predominant size range of materials carried by the Surging Glacier stream is less than 63 microns, the silt and clay size range. The next most important is the 125-250 micron size range, or the fine sand size range.

Table 5: SEDIMENT CONCENTRATIONS BY WEIGHT FROM STREAM WATER SAMPLES

Surging Glacier Stream, 1987						
Size Range		7/7/87 10:00	7/7/87 18:00	22/7/87 16:00	5/8/87 07:00	5/8/87 20:00
<63μm		40%	44%	34%	50%	31%
63-90μm		13%	16%	11%	17%	16%
90-125μm		15%	12%	10.5%	18%	12%
125-250μm		23%	22%	20%	7%	23%
250-500μm		10%	6%	17%	4%	12%
>500μm		--	2%	3%	--	6%

Non Surging Glacier Stream, 1987						
Size Range	13/6/87 19:00	7/7/87 10:00	7/7/87 19:00	27/7/87 16:00	4/8/87 10:00	4/8/87 18:00
<63μm	46%	52%	29%	17%	51%	34%
63-90μm	16%	8%	6%	15%	7%	7%
90-125μm	16%	12%	5%	23.5%	5%	28%
125-250μm	11%	14%	28%	21%	14%	17.5%
250-500μm	4%	10%	28%	18%	18%	5%
>500μm	5%	0.5%	2.5%	3%	1%	--

3rd Glacier Stream, 1987						
Size Range	13/6/87 19:00	7/7/87 10:00	7/7/87 18:00	20/7/87 16:30	4/8/87 10:00	4/8/87 18:00
<63μm	25%	18%	37%	61%	52%	51%
63-90μm	27%	18%	10%	8%	6%	10%
90-125μm	28%	32%	9.5%	9%	4%	7%
125-250μm	9%	10%	13%	14%	13%	19%
250-500μm	10%	21%	15.5%	6%	17%	5%
>500μm	--	--	10%	1%	4%	1%

The trends which these samples seem to delineate are as follows: the materials in the coarse and very coarse sand size range are found in higher percentages both later in the day as well as later in the season. These samples were taken at times of higher flow conditions, which means that the water has the capacity to carry larger amounts of materials, and larger size materials. The main size range seen throughout all samples is the silt and clay sized particles. This is expected, as there was a large amount of very fine material being carried by this stream. The presence of material in the very fine sand (63-90 μm) and fine sand size range (90-125μm) in reasonable quantities in all samples leads to the expectation that the material in these size ranges should be present in the post drainage deposits.

The Non Surging Glacier stream is much more variable in the range of material. Results of sample analyses show a much greater presence of the larger materials through all the samples analyzed. The medium sand size range (125-500 μm) and the coarse to very coarse sand size range (>500 μm) are seen in sufficiently large quantities throughout the season. The amount of medium sand sized material increases through the season, as expected due to higher flow velocity, but is variable on a diurnal basis. The evening sample taken early in the season (07/07/87) shows an increase in the percentage by weight of the material in these size ranges from earlier in the day, but the sample taken later in the season (04/08/87) shows a decrease from its morning percentage. The silt and clay size materials (<63 μm) are once again very dominant throughout all samples, with the lowest amount being 17% (27/07/87).

The 3rd Glacier stream samples are generally heavily oriented towards the <63 μm size range. The results of the analyses of the samples from this stream show some changes through the season. In the early part of the season, the second most important size range was the very fine and fine sand size particles (63-125 μm). The 125-500 μm (medium sand size) range increased in importance as the season progressed and discharge rose. The silt and clay size range (<63 μm), which made up an average of 26% early in the season, increased dramatically as the season progressed to account for an average of 54% for the latter part of the summer. The >500 μm (coarse to very coarse sand size) range was present in all samples except for the early season (13/06/87), when the flow velocity would have been insufficient to mobilize this size of particles.

In general, the high concentration of fine material (<63μm) being input into the lake by all three streams can be confirmed through other sources. The changes in the sediment plumes and the definition of the plunge lines reflect the observed increase in importance of the silt and clay size range throughout the season. The concentration of other size ranges is also reflected in the deposits associated with each of the streams.

6.3 DELTAIC DEPOSITIONAL SEQUENCES

The shallow water sequences, in particular the deltas, are of special interest to this study. Changes to the deltaic deposits are associated with changing water levels during the filling and draining of a proglacial lake. A typical delta contains very coarse topset beds, less coarse angled foreset beds, and distally-fining bottomset beds (figure 27) situated in a static location. With the changing water level, a delta is not constructed in one location, but migrates slowly upstream as the water level rises.

At the lowest level, the typical pattern of topset, foreset and bottomset beds is established. As the water level rises, a critical point is reached, and the location of the break between topset and foreset beds retreats rapidly up-delta. The factors controlling the location of this critical point are thought to be a combination of slope angle, water velocity and carrying capacity. As this critical point is reached, the coarser bedload making up the topset beds is deposited further upstream, forcing the formation of a new set of foreset beds. The distally-fining bottomset bed material now drapes the 'lower' delta. This pattern is repeated as the lake fills. The end result is a series of overlying steps on the delta (figure 28). As the water depth increases, the lower sections of the delta are covered with deep water lacustrine deposits. The thickness of these deposits is a function of time and sediment input to the basin.

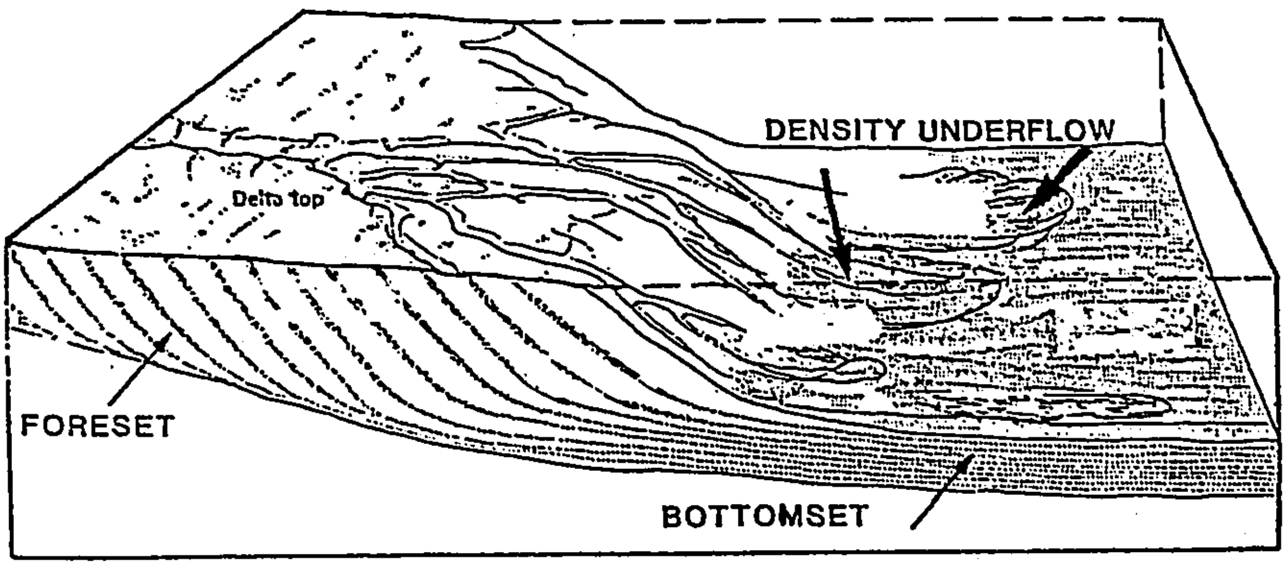


Figure 27: *Typical delta form (Ashley, Shaw & Smith, 1976).*

The typical depositional sequence seen at any point in the deltaic area is crudely laminated coarse gravels overlain by beds fining upwards which exhibit a variety of current structures. The fining sequence contains parallel laminar section overlain by wavy laminations, in turn overlain by ripples and climbing ripples of type A and B (after Jopling and Walker, from Ashley, Shaw & Smith, 1985). This is usually capped by another wavy laminar section overlain by a planar laminar section. Throughout this entire sequence, the material is fining upwards. The top part of the section is usually very fine material associated

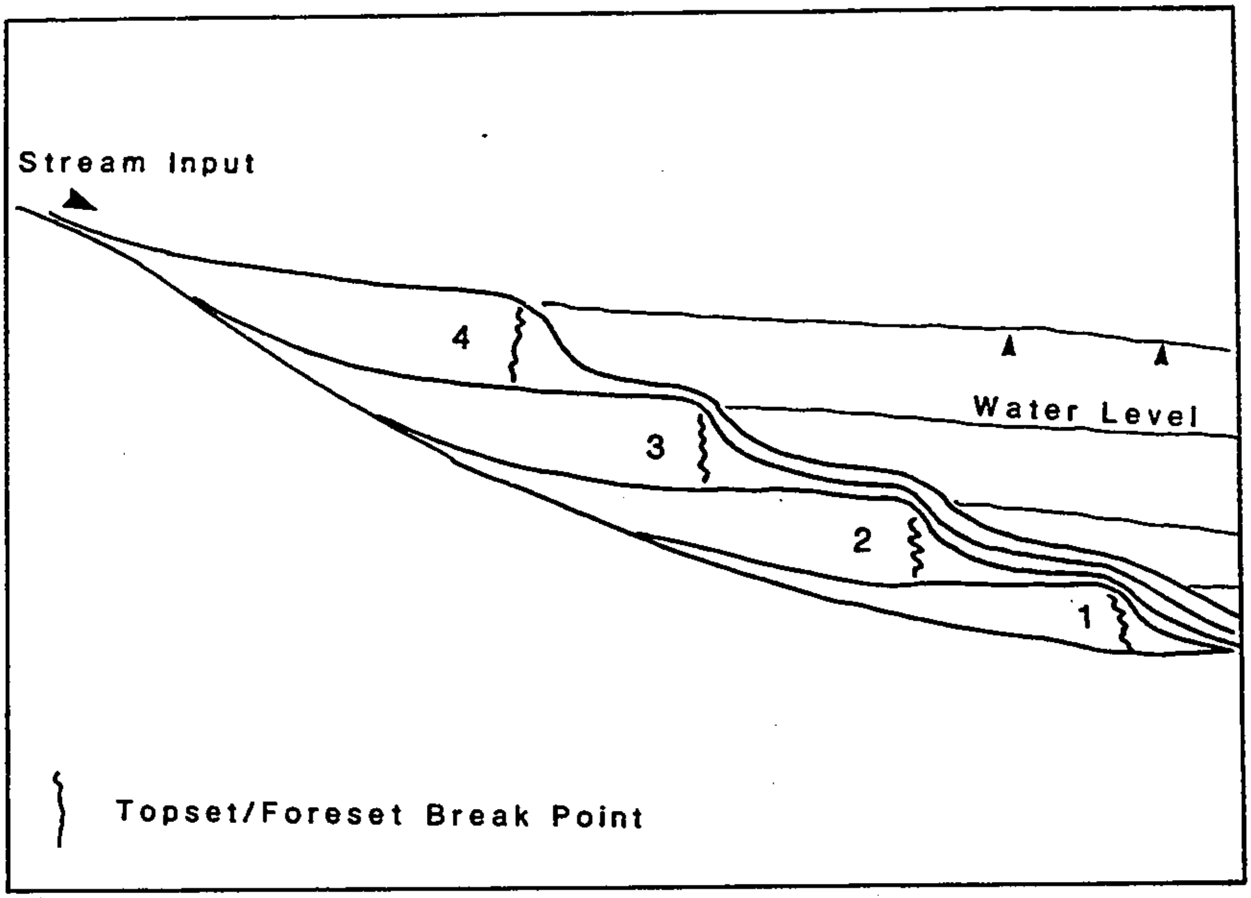


Figure 28: *Step delta form.*

with 'distal' deposition, containing no current structures. Deep water materials are usually deposited in a conforming layer which overtops the whole sequence (figure 29). This typical depositional sequence can be seen throughout the delta.

When drainage is initiated, changes occur in the deltaic area. Drainage is rapid though still long term (7 days). The first change associated with the drop in water level is the alteration in carrying capacity of the stream. Changes in stream gradient allow re-entrainment of suspended load and bedload which is redeposited downstream. The top deltaic sequences are truncated, while the lower sections may have a coarsening upwards sequence overlying the earlier-deposited fining sequence.

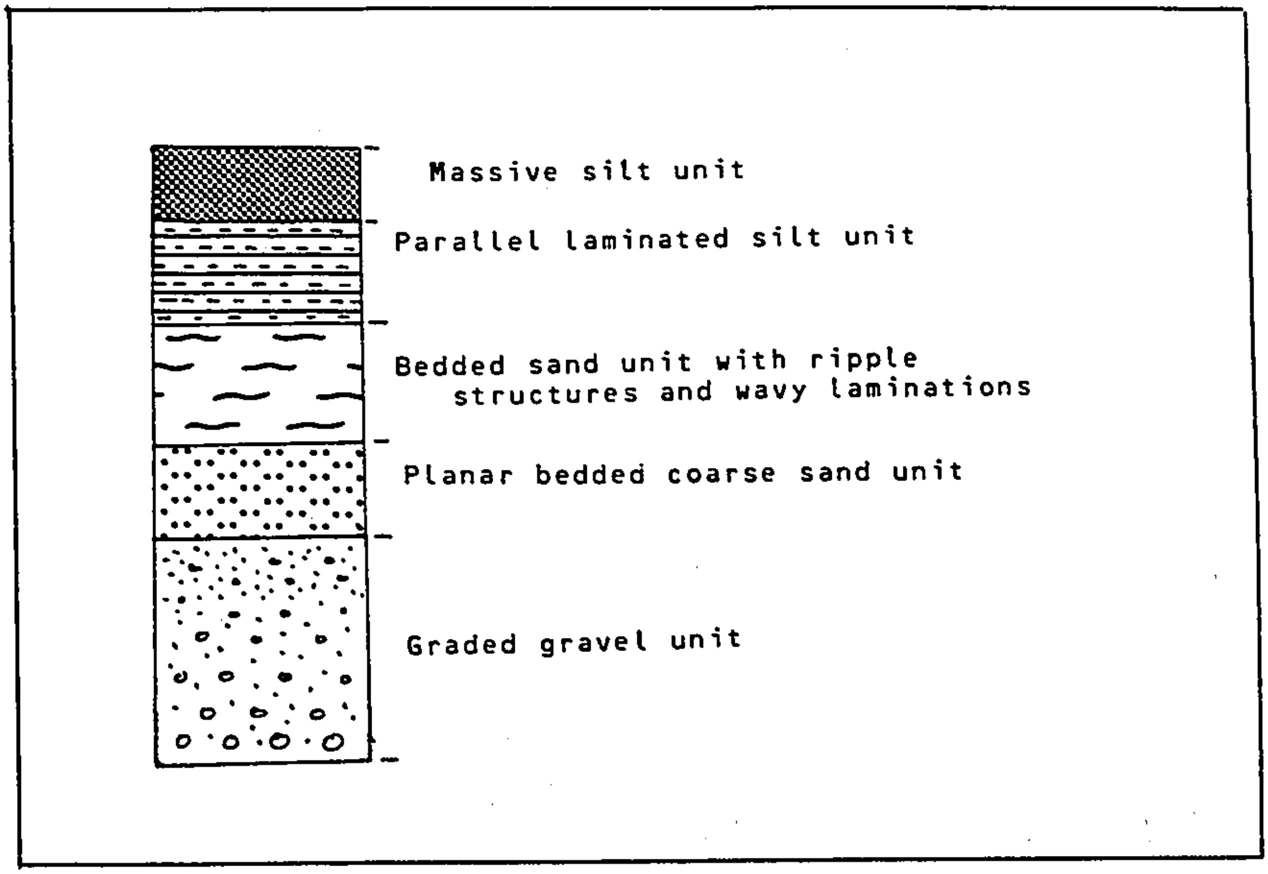


Figure 29: *Typical deltaic sequence*

The volume of material which is re-entrained and redeposited during drainage depends on the stream itself. If drainage occurs early in the season, the stream velocity and discharge are still high, and the capacity for erosion and transport is much greater. As a result, much channel migration will occur, resulting in disturbance to most or all of the upper delta sections, and larger amounts of material being deposited in the lower sections. If drainage occurs late in the season, then stream velocity and discharge are lower, and the erosion and transportation capacities are lower. The result is usually downcutting with little lateral stream migration. This leaves the upper sections undisturbed, and the lower sections have very little capping of the coarsening-upwards deposits.

The two different patterns can be seen in the deposits. In 1986, drainage occurred in early September (3-10), when discharge was low, so there was little disturbance of the filling sequence deposits. In 1987, drainage occurred much earlier (Aug 4-12), which resulted in upper sequence truncation and lower sequence reversed deposition. In 1986, very few sections were exposed, since only a narrow central area was cut by the stream during drainage (figure 30). In 1987, however, a very wide area was affected (figure 31), and remnants of deposits from several years (not necessarily sequential) were exposed.

The geological and glaciological contrasts in the basin are reflected in the composition of the stream sediment loads. The NSG and 3rd streams carry large amounts of material, with bedloads predominantly in the gravel size range. This means that small scale current structures are mostly indistinguishable or absent. Larger scale divisions between topset and foreset beds, and distinctions between fluvial and lacustrine deposits are, however, clearly visible. The deltaic areas from these two streams will be described in general, but only a few sections will be analyzed in detail.

The SG stream transports sediments ranging from cobble and boulder size to silt and sand size. However, the suspended material outweighs the bedload. Deposition on the delta proper is mainly the sand sized or larger material. Small scale current structures are readily seen, as are changes in material size within the depositional sequences. Most of the sedimentary sequences examined in detail will be from this stream. However, the general structure of this delta will also be examined and compared to the others.

The following section will concentrate on the SG sections, and will first distinguish 'classes' of material size and associated structures in order to facilitate the analysis of the individual sections. These facies are based on units set out by Liverman in his work on

Figure 30: *S.G. & N.S.G. deltas, 1986: little disturbance during drainage.*

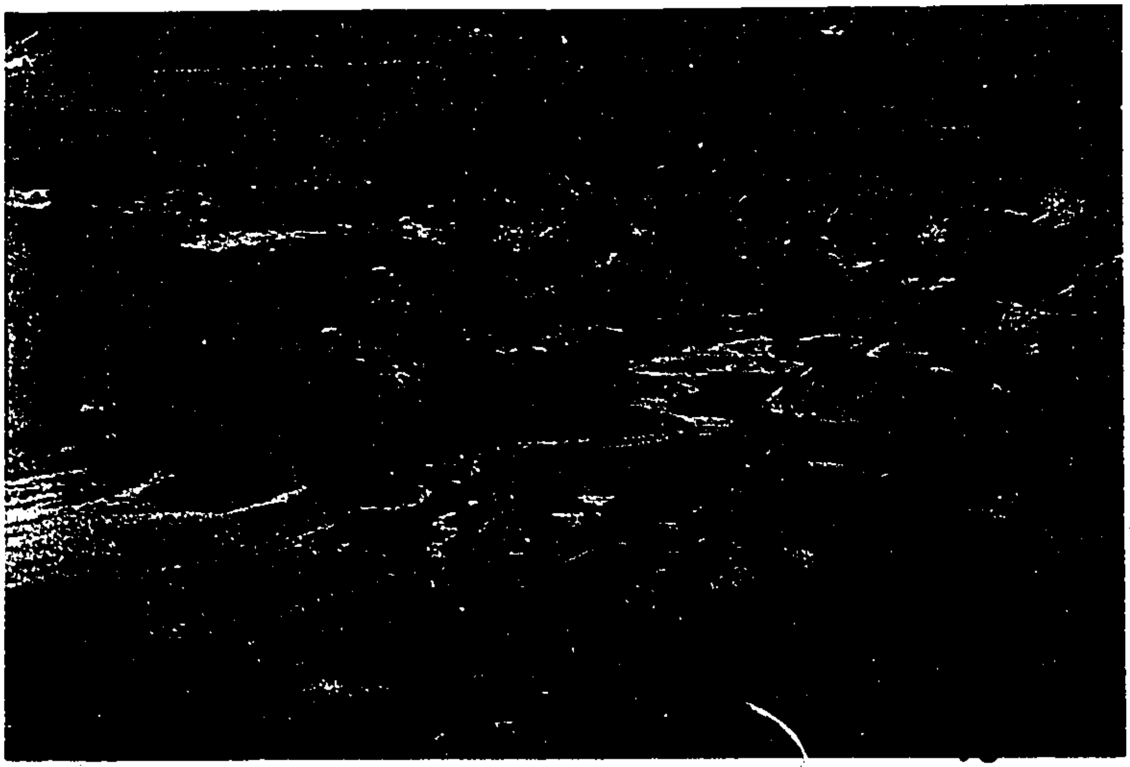


Figure 31: *S.G. delta, 1987: large area affected by drainage.*

Hazard Lake (1980, 1987).

6.3.1 SEDIMENTARY FACIES

Facies 1: Gravels

This facies is characterized by generally poorly sorted, well rounded clasts. These vary in size from boulders to medium to coarse sand matrix. Structures are hard to discern, but horizontal bedding is sometimes visible. These units are usually associated with high velocity, high capacity stream flow and shallow water.

Facies 2: Cross Bedded Sands

This facies is rare, and is characterized by tabular cross bedding, seen in well sorted, medium grained sands.

Facies 3: Planar Bedded to Massive Sands

This unit contains material in the fine gravel to very fine sand range, and is usually well sorted. This unit is most often devoid of internal structure.

Facies 4: Cross Laminated Sands

This unit is composed of well sorted material in the fine sand to silt size range. Structures found in this unit are draped wavy laminations (after Gustavson et al, 1975), and ripples of types A and B (after Jopling & Walker, 1968).

Facies 5: Graded Sands and Silts

Normally graded beds of coarse sand to silt sized material make up this facies. The materials are generally well sorted.

Facies 6: Graded Parallel Laminated Silts

These beds either drape other structures or clasts, or lie flat on top of sedimentary units. Each lamination is normally graded.

Facies 7: Massive Silts

This unit is usually found on top of a normal graded sequence, and the units are often massive (1-3 cm).

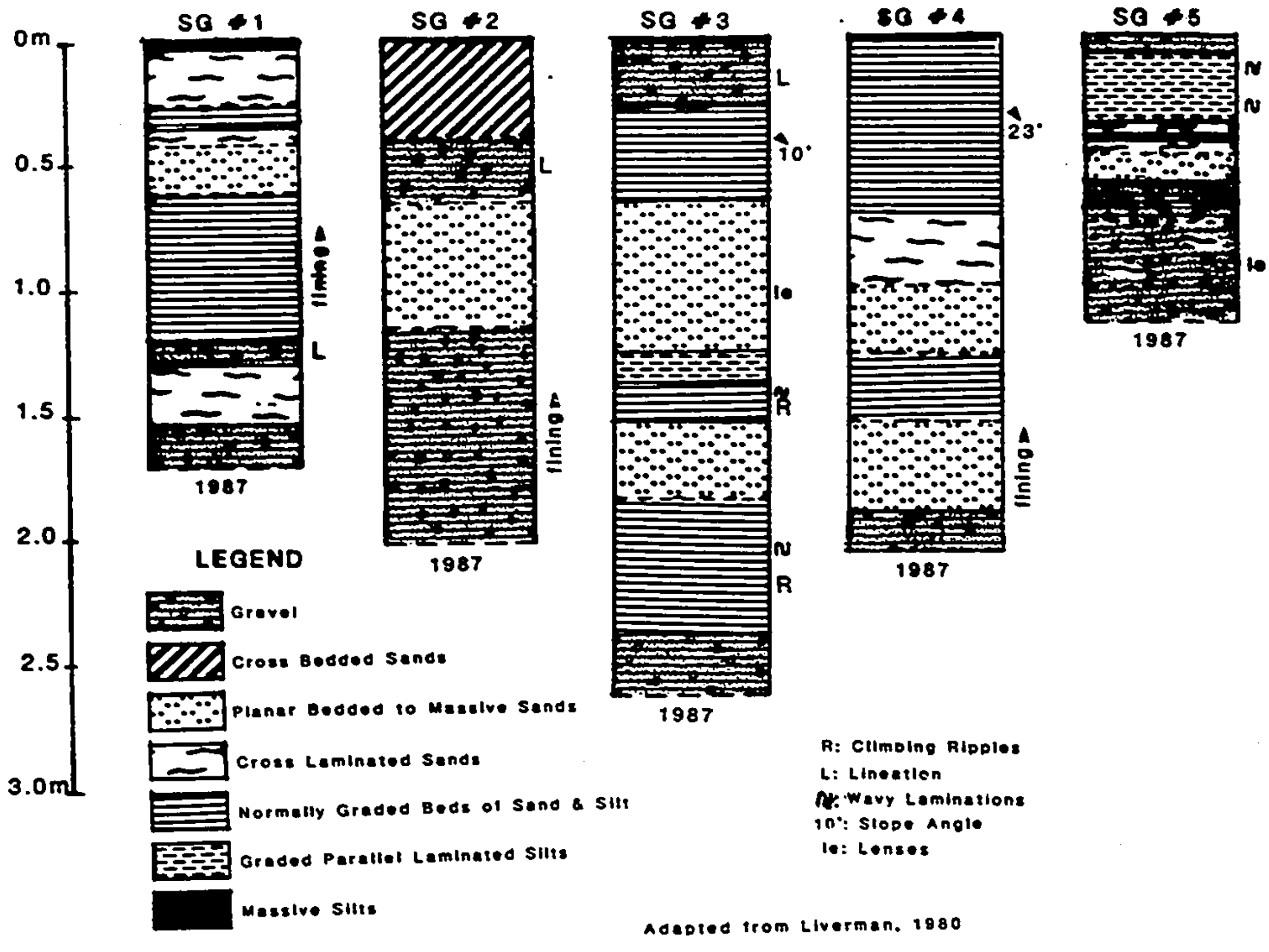
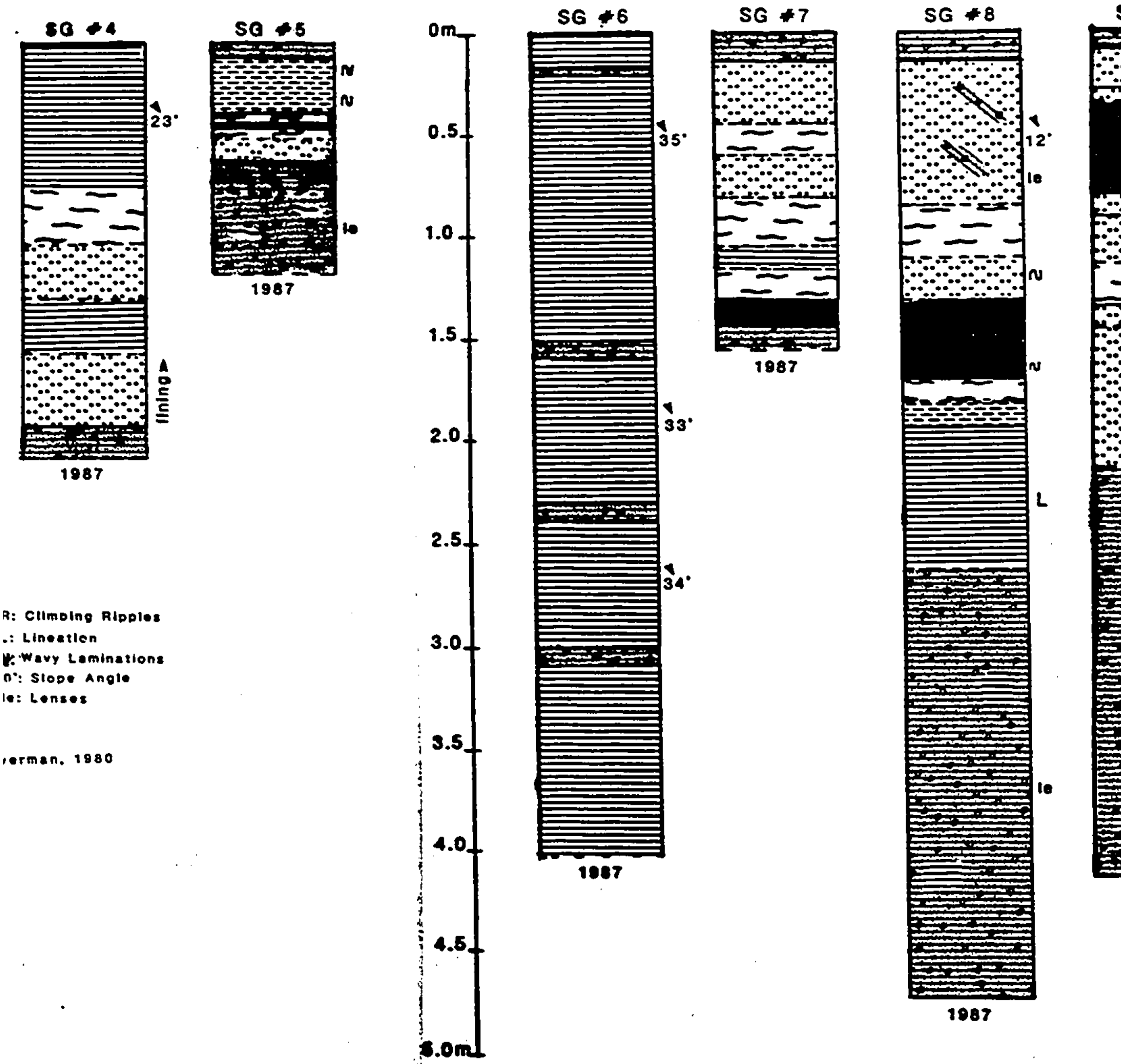


Figure 32: *Surging Glacier deltaic sequences: post 1987 drainage.*


SG #4
SG #5
SG #6
SG #7
SG #8
0m
0.5
1.0
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23°
35°
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fining
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1987
R: Climbing Ripples
.: Lineation
: Wavy Laminations
0°: Slope Angle
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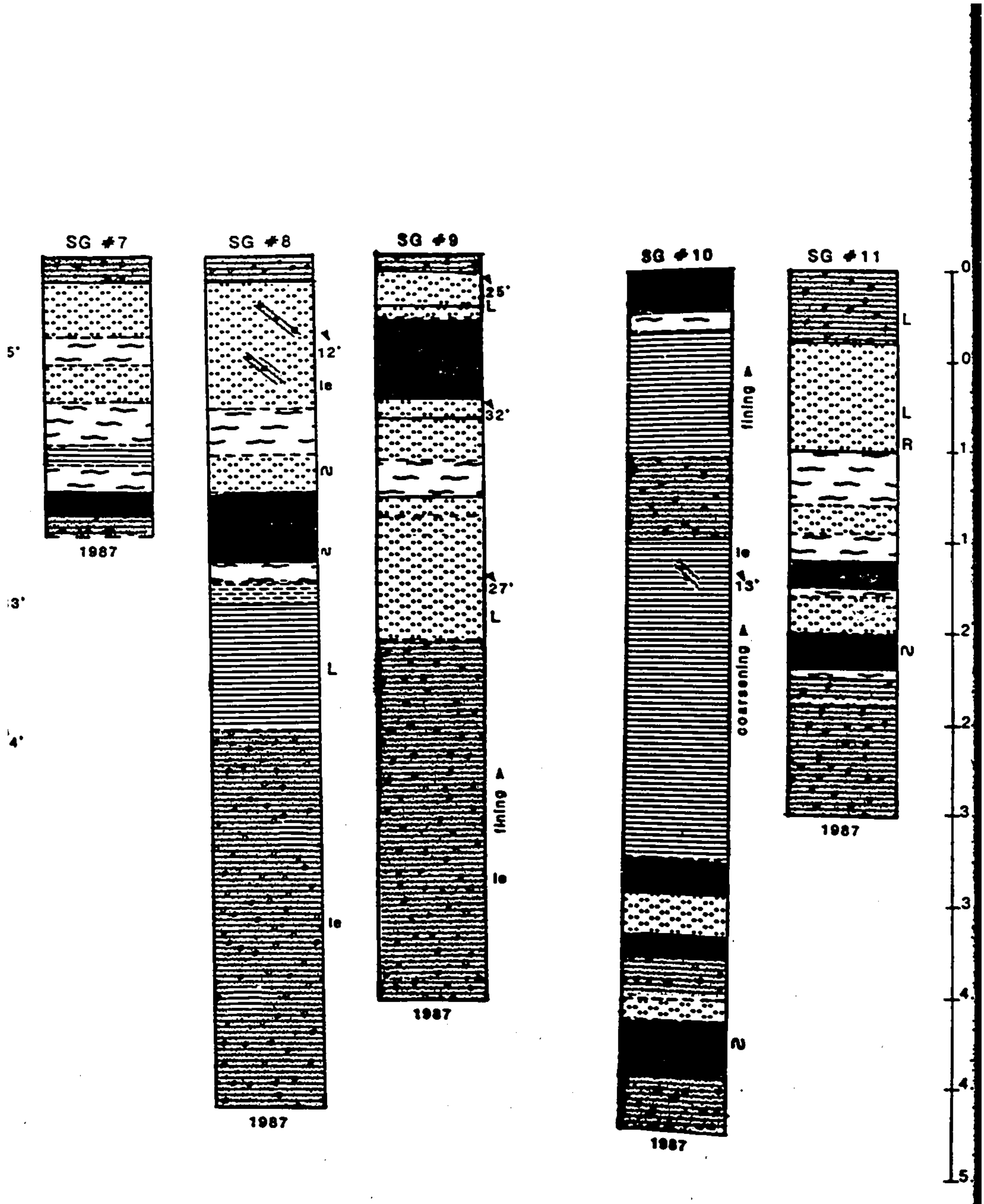

SG #7
SG #8
SG #9
SG #10
SG #11
1987
1987
1987
1987
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25°
L
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32°
27°
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L
fining
coarsening
13°
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R

6.3.2 S.G. STREAM DEPOSITS

11 sections along the east side of the Surging Glacier delta were analysed after the drainage of the lake in 1987. These range in location from the top of the delta to close to the junction with the N.S.G. delta. The thickness of the sections varies from 1 m to almost 5 m. Some sections contain only a fining upwards sequence (higher level), while others contain deep water deposits and reverse sequences overtopping the normal deposits. Some of the sections contain multiple deep-water silt layers, indicating more than 1 year of deposition.

The gravels (facies 1) and the graded sands and silts (facies 5) are the most frequently occurring facies The massive silts (facies 7) are only seen from section 7 to 11, and a reason for the lack of these deposits further upstream is that underflows moving along the delta would have disrupted any settling fine materials. Facies 2, the cross-bedded sands, is only seen in section 2 since this section was cut across the channel flow direction. Facies 1, the gravels, is the basal unit in all sections, and is also the capping unit in all sections except for 2,4,6 and 10. These gravels are deposited very rapidly, since they require a fairly high current velocity to move them. They indicate a proximity to the source, shallow water and rapidly dropping current. These units are usually deposited in the stream channel up from the lake, or in the shallow deltaic area, forming the topset beds. Figure 32 contains the diagrammatic representations of the following sections.

Section 1 shows a deepening water sequence, overlain by a reversed, shallowing deposit. The gravels, alternating with the cross-laminated sands at the base of the section indicate stream flow shifting and channel infilling. As the water deepened, 40 cm of graded sands were layed down. This unit fines upward, showing a decrease in the stream's carrying capacity with the increasing water depth. This then grades into a massive sand unit 20 cm thick. This sand unit is the deepest water, lowest flow condition deposit of this section, as the unit above shows a reversal in the flow conditions and water depth. The overlying 40 cm of graded and cross-laminated sands contain ripples, indicating shallowing water and increasing velocity conditions. The capping unit is the fluvial gravels.

Section 2 was located transverse to the direction of stream flow, and was bounded by two large boulders. The lower unit of the section is gravels, which fine gradually upwards into a 50 cm thick massive sand unit. Immediately above this is a gravel unit, which is a result of a channel shift and the first drop in the water level. This is overtopped by cross-bedded sands, which indicate higher velocity and shallowing water conditions and a cross-current location.

Section 3 is 2.5 m thick, with basal gravels fining upward over 1.2 m. This is capped by graded sands containing ripples and overlain by wavy laminations. Next are massive sands showing faint laminations, and then a graded parallel laminated silt unit. The presence of facies 6 in this section probably indicates that the main current had shifted far enough away for reasonably still water conditions to be established, and for the fines to begin settling out of suspension. These conditions do not last long, since this unit is only 10 cm thick. This is a fairly classic deepening sequence, showing the progression of various types of current structures and grain sizes. An unconformity at the top shows the change in conditions. The next unit is a 60 cm thick massive sand unit. This unit indicates slightly deeper water by the grain size, but does not show much increase in current velocity, since there are no current structures visible within the unit. Moving up the sequence, the next unit is a graded sand unit, dipping downstream with a 10° slope angle. This is a foreset bedding type unit. Overlying the section is a shallow water, high current velocity gravel deposit.

Section 4 is fairly standard in its lower half. The sequence fines upwards from the fluvial gravels at the base, through graded sand beds into a massive sand unit. Above this is a 30 cm thick section of cross-laminated sands containing ripples, climbing ripples, and overlying wavy laminations. This unit is truncated, and overlain by graded sands and silts dipping lakeward at 23°. This 50 cm thick unit is also abruptly truncated on its upper boundary by similar material deposited horizontally rather than at an angle. The presence of the dipping units is an indication of the foreset beds of one of the original set of step-deltas.

Section 5 is a fairly small sequence, measuring 1.14 m in thickness. The lower 40 cm of the section is gravel, gradually fining upwards, but containing isolated lenses of cross-laminated sands. The gravels eventually fine out into a massive sand layer 10 cm thick, containing lenses of fine gravels. These lenses are most likely the result of sediment pulses in the stream flow. Overlying this is a section of cross-laminated sands containing first ripples, then climbing ripples, then finally wavy laminations. This gradual change in the structure represents a deepening of the water and a slowing of the current. Overtopping this concordantly is a unit of graded laminated silts, which is wavy at the base to conform with the underlying sand structures, but gradually smooths out until it is horizontal. This is the first major silt unit (25 cm thick), which indicates that water depths necessary for silt deposition are being achieved. The abrupt transition at the top of this silt unit into a gravel layer shows two posibilities. The first is that the site was isolated from shifting channel deposits with the onset of drainage and thus no reverse sequence was deposited. The second possibility is that when the water was shallow enough, the current removed the shallowing deposits, then replaced them with bedload gravels.

Section 6 is 4.0 m thick, and is mostly foreset bed deposits. The section is mainly graded sand beds, dipping lakeward at 33-35°. Interbedded between these thick units (70 cm - 130 cm) are 10 cm thick gravel layers. These probably originated as debris flows further up the delta.

Section 7 is only 1.5 m thick, and is bottomset-type deposits. The basal unit is gravel, overlain unconformably by 14 cm thick massive structureless silt unit. The rest of the section is a reverse section of deep to shallow water deposits. The section is mainly cross-laminated sands containing current structures. The structures change from wavy laminations to ripples to major climbing ripple sequences. Massive sand units in between the rippled sequence seem to indicate a shift in the water current location rather than a change in water depth. The top 40 cm of the section is a massive sand unit deposited under fairly high current conditions, coarsening upwards into a 15 cm thick capping gravel layer.

Section 8 contains a full normally graded sequence, overlain by silt deposits, and topped by a reverse sequence ending in gravels. The lower section is made up of 2 m of gravels containing isolated lenses of graded coarse sands. This is conformably covered by 70 cm of graded sands, containing lenses of gravels and very fine sands. An abrupt transition leads into a reverse graded silt unit, which then coarsens into a cross-laminated sand unit containing many climbing ripples. The upper boundary marks an abrupt change into deep water massive silt deposits 40 cm thick. The basal part of the silt unit contains some wavy laminations due to the material draping the underlying rippled sand unit. The silt unit gradually coarsens into a massive fine to medium sand unit containing some wavy laminations, indicating a change in water depth and current velocity. This grades conformably into a unit of cross-laminated sands entirely made up of climbing ripples. This unit is conformably topped by a massive sand unit dipping lakeward at 12°, and containing lenses of fine gravels (turbidity flow indicators). The capping unit is the fluvial gravels.

Section 9 resembles section 8 in that the lower 2 m of the section is a gradually fining upward gravel unit containing lenses of sands. The upper transition is gradual as the sequence moves into an 80 cm thick massive coarse sand unit, containing gravel lenses and dipping lakeward at 27°. This is in turn overlain by a 20 cm thick cross-laminated sand unit contain ripples. Above this is a massive sand unit whose lower 3 cm are wavy in response to the underlying structures. This sand layer is unconformably overlain by a 10 cm thick sand unit dipping lakeward at 32°. This unit is part of a set of poorly developed foreset beds. The next unit is a 40 cm thick massive silt layer, which is the deep water deposit for this section. The silt unit coarsens gradually into a sand unit, indicating the switch in direction of water level change. This is unconformably topped by another dipping sand layer (25°), and capped by a gravel layer.

Section 10 is the most complex of the sections on the east side of this delta. It contains 4 separate silt units within the sequence, as well as 3 separate gravels units. The section has a gravel base, overlain unconformably by a 30 cm massive silt layer containing wavy laminations. This unit grades slowly into a massive sand layer, then into a second gravel layer. This appears to be one year's deposits, both the filling and the draining stages. Unconformably overlying this gravel unit is another massive, unstructured silt layer 10 cm thick. This unit is neither underlain nor overlain by conforming deposits. A possible explanation for this may be that an iceberg was grounded on the gravel surface, and did not float until the water was deep enough to also support deposition of fine materials. This would explain the lack of a deepening water sequence, while also accounting for the presence of the deep water silt layer. A 25 cm thick massive sand layer is positioned between this silt layer and another silt unit, with abrupt upper and lower boundaries. The third silt layer is 20 cm thick, and gradually coarsens upwards into a 170 cm thick section of graded silts and sands. This large unit coarsens until fine gravel lenses are included near its upper boundary. The boundary at the top of this unit is abrupt, indicating a major break in the depositional sequence. The reverse grading is the result of a shallowing phase, as the lake drains. The overlying unit is very coarse gravel, which fines slowly into a graded sand bed containing fine gravel lenses. This unit grades in turn into a cross-laminated sand unit containing climbing ripples. This fining upwards pattern is the indication of another filling phase, since the current gradually drops, the grain size decreases, and the structures are present intermittently. This set of deposits is overlain by a 25 cm thick massive silt layer, which drapes the underlying structures. This silt layer is the current year's deep water deposit, and the others are from former years. However, the past deposits are not necessarily from concurrent years, since the stream channel is not always cut into the same area, which leads to selective preservation of various years' deposits.

Section 11 is only 3 m thick, but it contains 2 silt units. The basal unit is gravel, overlain by a thin sand layer, then another thin gravel layer, then another thin cross-laminated sand layer containing climbing ripples. This alternation between gravel and sand units is an indication of shifting channel locations. This set of deposits is draped by a 20 cm thick massive silt layer, whose wavy forms are determined by the lower unit's structures. This silt layer grades up into a massive fine sand body, then into another ripple-filled cross-laminated sand unit. This is a shallowing phase set of deposits. What should follow is a gravel layer, which fines back out with the following season's filling sequence. However, this section moves directly into the next deep water deposits, then grades back up into another shallowing sequence, as follows: The sand layer is overlain by a second 15 cm thick silt layer, thought to be the 1987 deposit. This silt layer reverse grades up into a cross-laminated sand and silt layer containing ripples, then into a massive sand layer, then back into a cross-laminated sand unit containing ripples and lamination. This unit is gradually truncated by a coarse sand unit containing a few laminations, which is in turn capped by a fluvial gravel unit.

6.3.3. N.S.G. STREAM DEPOSITS

The deposits found in the N.S.G. delta are different from the S.G. deposits. They are predominantly fluvial gravels and deep-water silts and sands. The photographs of the delta after the 1986 drainage season show the contact between the two facies (figure 33). After the 1987 drainage, some sections with sands and fine gravels were exposed on this delta (figure 34). The facies of which these units were composed were mostly cross laminated and massive sands. Macro structures seen in this section were the topset, foreset and bottomset beds of one of the delta steps.

The majority of the deltaic sections in the N.S.G. delta are characterized by massive beds of coarse to fine gravels, exhibiting traces of lineation (figure 35). The only variation is the presence of thin beds of fine gravels interspersed within the massive sections of very coarse gravels. These thin laminations are probably the result of a reduction in flow conditions lasting long enough for some of the finer suspended material to be deposited.

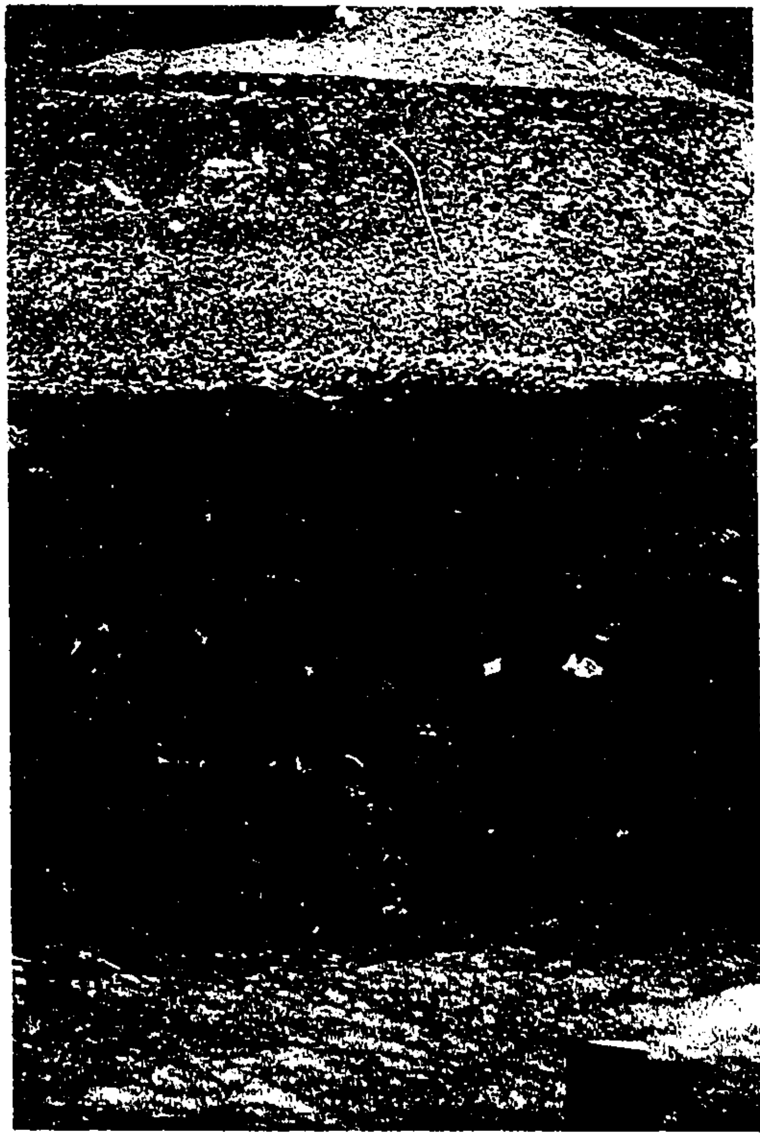

Figure 33: *N.S.G. delta: gravel and silt deposits clearly differentiated.*

Another possibility is that there is a slight drop in the sediment concentration of the stream water. Such sudden changes in the sediment load are well documented through the measurement of the sediment concentration carried out on the streams during two

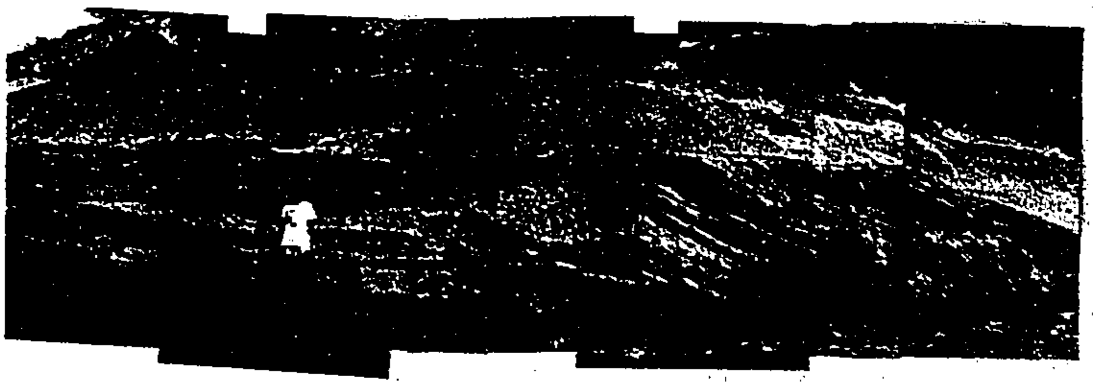

Figure 34: *N.S.G. delta: 1987 panorama.*

consecutive seasons.

The deep water silts contain current structures identical to the ones seen in the S.G. deposits. These are wavy laminations, ripples and climbing ripples, in a normal sequence showing a decreasing flow velocity. The upper layers of this deposit have little or no structure, being laid down beyond the current flow. Thin layers of fine to coarse gravels interbedded within the silts were caused by underflow pulses off the delta front.

Some of the current structures in parts of the delta deposits may have been caused by reverse currents induced by grounded icebergs or by the grounding of icebergs into the delta

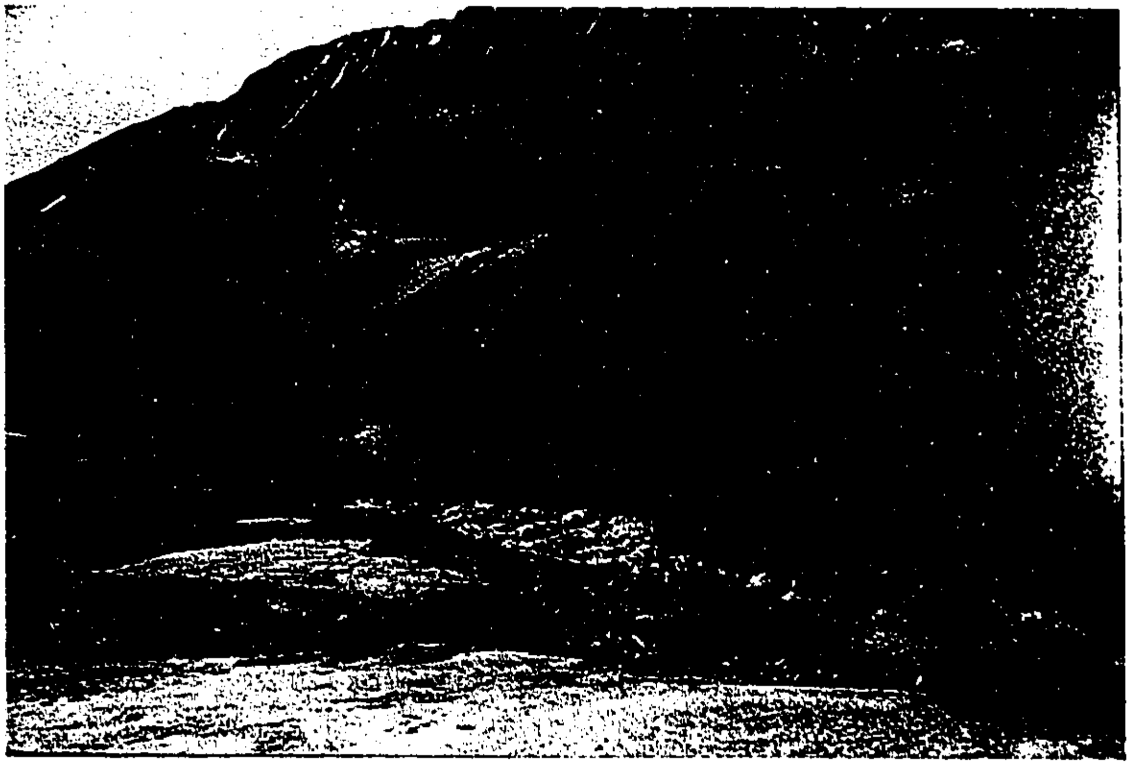

Figure 35: *N.S.G. deltaic gravel deposits.*

sediments.

Figure 36 contains a diagrammatic representation of two depositional sequences which typify this delta. The two units are the coarse gravels and the massive silts. Both section 1 and section 2 have the massive silts with a maximum thickness of 0.5 m. This thickness exceeds that seen on the S.G. delta, since there is more of a transition between the coarse materials on that delta. The predominance of this size material was predicted by its high concentration in the grain size analysis of the stream water. On the N.S.G. delta, since there is generally an absence of medium to fine material within the water column, the deposits go

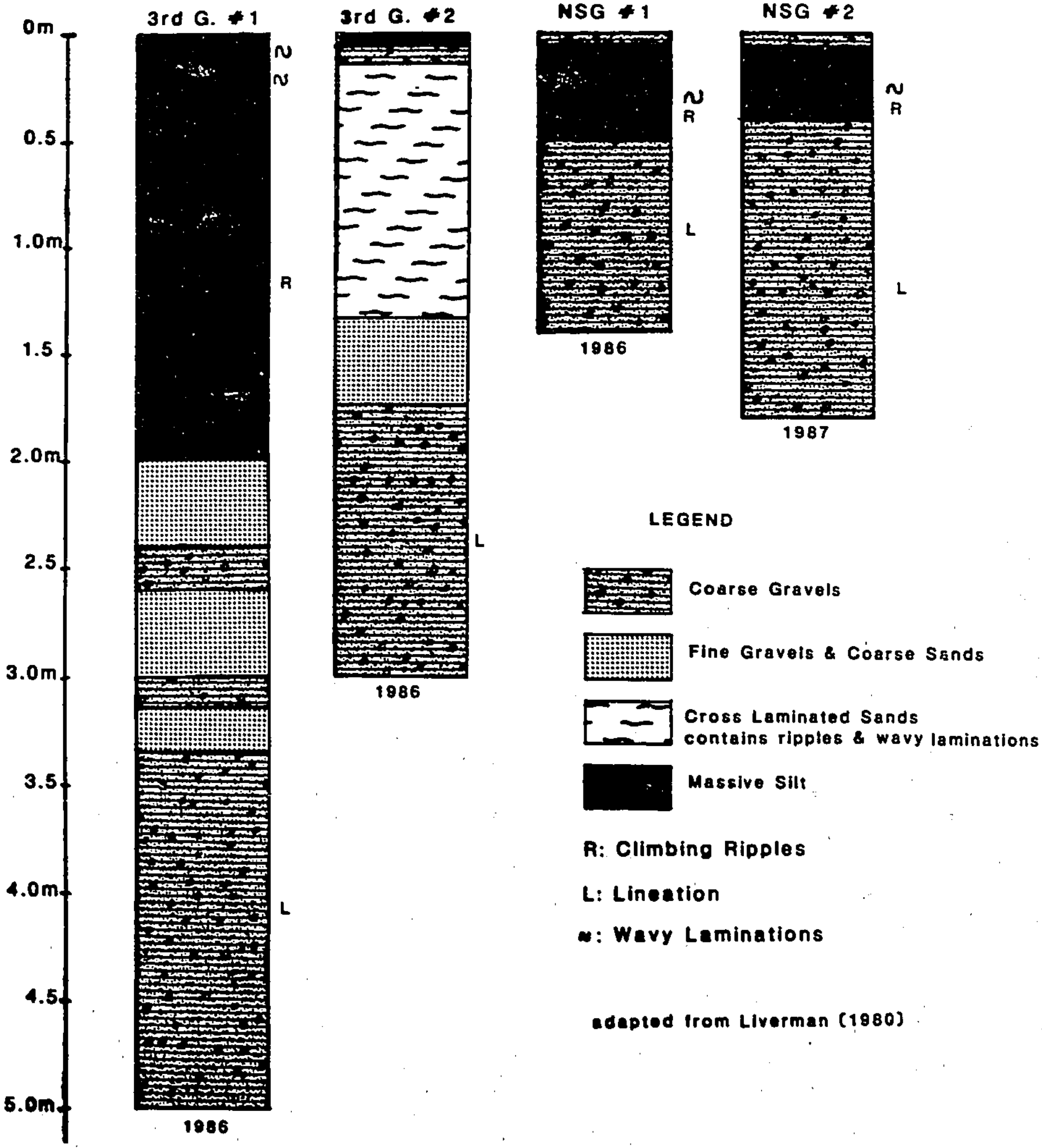

Figure 36: *N.S.G. & 3rd G. stream deposits.*

from coarse gravels almost directly into fine lake silts. Thus, the silt unit is of greater thickness than if other fine depositional units were present.

6.3.4. 3rd G. STREAM DEPOSITS

The deposits from this stream are very similar to the N.S.G. sequences. This is due to similar material size and resulting deposition patterns. The post-1986 drainage deposits are

Figure 37: *3rd Glacier deltaic deposits: post 1986 drainage.*

mostly coarse gravels overlain by thick units of silts (figure 37). Sections which contain intermediate size range materials are concentrated in a sharp elbow of the channel. The channel has a 90° turn close to the location of maximum lake level which seems to have acted as a sediment trap. This turn results in turbulence in the stream flow, which affects the depositional pattern. Some of the sedimentary structures are deformed by slumping (figure

Figure 38: *3RD Glacier sequence: note deformed sand & silt structures.*

38). The post 1987 drainage sections were mostly coarse gravels overlain by a thick silt layer.

Figure 36 presents a diagrammatic illustration of two sections studied after the drainage in 1986. They contain large amount of coarse and fine gravels. Section 1 is characterized by a 2 m thick silt unit with ripples and wavy laminations, the typical current structures which were established in the sand-sized range of materials on the S.G. delta. The absence of the sand-sized material on this delta, as on the N.S.G. delta, forced the deposition of greater thicknesses of the silts, and resulted in their containing the shallowing water current structures. This was predicted by the very high percentage (ave 54%) of material in this size range in the latter part of the season. Section 2 was located in the turn of the stream channel mentioned earlier. This area acted as a sediment trap, which can be seen from the presence of sand sized materials. This unit contains the current structures, which were apparently missing from the silt layer capping this depositional sequence.

6.3.5. GENERAL ANALYSIS

The deltaic deposits reflected the suspended and bed loads of the streams. The former was recorded and the latter was assessed generally by observation in the field. These two factors combine to determine the type of structures and the size of material seen on the deltas.

All three deltas contain a step-type pattern, with the characteristic multiple sets of foreset beds along their length. This indicates that the processes which cause the formation of this type of delta are not dependant on the size of material. The creation of step-type deltas is therefore linked to the water level change, not the material size. The size of material carried by the streams only affects the type of deposits which will form, either containing only the very coarse and very fine materials, or the whole range of sizes. The structures are a function of the velocity of the stream, and will form in whatever size of material is present. This is shown by the formation of current structures in the sand size materials on the Surging Glacier stream delta, and in the silt size materials on the other two deltas.

The general pattern of deposition on any of the steps of the deltas is that of a fining upwards sequence layed down during the filling stage of the lake, and a partial or whole coarsening upwards sequence put in place during the drainage stage. The completeness of either or both of these sequences depends on the location of the section on the delta, since the depositional rate is affected by the time under water, as well as the time of drainage in the hydrological season.

6.4 DEEPER WATER PROCESSES

6.4.1. INTERFLOWS AND UNDERFLOWS

The differentiation of influent streams into discrete plumes is caused by density contrasts determined by differences in temperature, salinity and sediment concentration. The relationship between the densities of the streams and the lake determines which type of inflow pattern is likely to occur.

Interflows are a common sediment distribution mechanism in ice-dammed lakes during periods of lower influent density. These low density periods may be: 1) early or very late in the melt and runoff season; and 2) early in the day before daily runoff starts to pick up. This is based on the general assumption that there is a close correlation between discharge and suspended sediment, which may not always be the case (Collins, personal communication). Interflows occur where the inflow density is intermediate between high density bottom water and low density surface water (Ashley, Shaw & Smith, 1985). The stream water enters the lake, moves towards a 'plunge line', sinks into the water column until it reaches density equilibrium, then spreads throughout the lake at that level along a constant-density surface. The plunge line is the convergence point of the dense lakeward-moving surface plume and the less dense shoreward-moving countercurrent (Ashley, Shaw & Smith, 1985). The location of this point changes with the magnitude of the density difference between the stream and lake water. For example, the plunge line will move closer towards the stream mouth with an increase in the density difference (Weirich, 1986).

Underflows are thought to be the dominant mechanism for introducing and distributing sediment in glacier-fed lakes (Ashley, Shaw & Smith, 1985; Weirich, 1986). Underflows are high density (relative to the lake water) plumes which operate similar to interflows, entering the lake then sinking at a plunge line to flow along the lake bottom. For these flows, density differences depend more on sediment concentration than on thermal differences. These flows have been documented both by direct observation and by interpretation of sediment records.

Two different types of underflows have been identified, density-induced underflows and turbidity currents. The term 'density induced underflow' refers mainly to high density currents generated by underflowing sediment-laden river water which produce quasi-continuous currents (Ashley, Shaw & Smith, 1985). The term 'turbidity current' refers mainly to episodic surge-type currents formed by subaqueous slumping. With the density-induced underflows, all three components (thermal, dissolved sediment, suspended sediment) combine to produce the density difference. With the turbidity currents, the suspended sediment component is the dominant source of the density difference (Weirich, 1986). Continuous feeding underflows have generally lower sediment concentrations and relatively stable rates of input of material. Surge-type underflows are usually associated with a rapid disturbance event such as slumping, which provides a high initial sediment concentration (Weirich, 1986).

Surge-type underflows usually only last for a very short period of time. Continuous feeding underflows, however, have been known to continue for hours or days (Gilbert & Shaw, 1981). However, the current velocity during these extended flows is highly variable. Underflows have also been found to travel long distances (Gilbert, 1975). Reduction of the density (by sediment deposition and mixing with overlying water), and a drop in velocity (due to friction with the lake bottom and low slope angles) leads to the eventual loss of integrity of the flow. However, this can take quite a long time.

6.4.2. FIELD RESULTS

One of the objectives of the data collection program was to determine whether interflows and/or underflows could be detected within the lake by using relatively simple techniques. Data was collected during both the 1986 and 1987 field seasons, concentrating mainly on the stream delta areas and close to the ice margin. The two main types of data collected were water temperature and current velocity measurements. However, supplementary information was used to support the data recorded, such as photographic records of plunge lines as well as thermal and sediment concentration differences between the influent stream water and the lake water.

Evidence gathered indicated that both interflows and underflows occurred. First, there was a temperature difference between the influent stream water and the lake water. The lake is cold monomictic, meaning that it has a limited annual ice-free period, a small thermal gradient, little or no thermocline and surface temperatures which rarely exceed 0°C (Hutchinson, 1957). The preservation of the cold monomictic regime is aided by rapid drainage, high winds, large scale calving, and input of near freezing meltwater (Blachut & Ballantyne, 1976). The lake water rarely exceeded 1.5°C except in the shallow deltaic areas. The influent stream water temperature ranged from 2.7 to 4°C. This accounted for the warmer temperatures recorded in the deltaic zones, since some mixing did occur. Since the point of maximum density for water is 4°C, it was evident that, based solely upon thermal characteristics, the stream water would be denser than the lake water. This thermally-induced density difference is one of the main factors which trigger and maintain interflows and underflows.

Secondly, there was a difference in sediment concentration between the stream and lake water. The sediment load carried by the streams was highly variable, both on a seasonal and diurnal basis. During the early part of the season, suspended sediment concentration was usually < 1 g/l, while later in the season it reached as high as 10 g/l (Johnson, unpublished data).

The Surging Glacier stream's discharge was characterized by pulsations in both the water volume and sediment load, corresponding to shifts in the drainage system of the glacier. The sediment carried by this stream was supplied mainly from englacial and subglacial sources. There was a component of very fine material (glacier flour) in the water which had a very slow settling rate. Samples left to sit for 5-10 days still maintained a milky characteristic, and the filter papers used became clogged very easily.

Both the 3rd Glacier stream and the N.S.G. stream discharges were more constant, without the frequency of pulses of the S.G. stream. Their discharges did vary both seasonally and diurnally, as expected. The 3rd Glacier stream's sediment load originated mainly from mudflows off the valley walls. The Non Surging Glacier stream acquired its sediment load by travelling under and through ice-cored moraines, which often collapsed when parts of the ice core melted.

The difference in suspended sediment concentration between the stream and lake water also creates density differences, which lead to the formation of interflows and underflows. The lake water itself had almost no suspended sediment except at the base of the water column in some of the central areas. This may have been the remnants of dispersed underflows. The only area of the lake with noticeable sediment concentrations was where plumes feeding into the lake showed the approximate location of either englacial or subglacial passageways from the Kaskawulsh Glacier.

Thirdly, there were well-defined plunge lines at all three deltas during most of both summers. Plunge lines are usually seen only at the convergence point of two water masses of different densities. During the summer of 1986 large numbers of icebergs were trapped in the delta areas by onshore winds. These caused interruptions in the normal flow patterns and disruption of the formation of plunge lines. The presence of the icebergs led to a high degree of turbulence and a higher than normal rate of deposition in the deltaic areas, with a slight sediment 'starvation' for the lower reaches of some of the deltas since the icebergs blocked the progress of the stream water into the lake (figure 39). This was most readily observed on the N.S.G. stream delta and the 3rd Glacier stream delta, and only rarely seen on the S.G. stream delta (figure 40). There were few icebergs trapped in the lake basin during the 1987 field season, and normal flow conditions dominated. The plunge lines were either weakly defined or absent during the early part of the 1987 season (early June), because of small sediment concentrations and density differences. As the season progressed and the thermal and sediment concentration differences between the two water masses became more marked, the plunge lines became more distinct (figure 41).

Measurements of current velocities at various locations within the lake basin indicated the presence of interflows and underflows within the water column. In the slightly deeper deltaic sequences, many of the profiles showed very marked current velocity increases at the base, confirming that a large amount of the material coming off the deltas was being moved via turbulent underflows. Near the ice margin, maximum velocities were in the 12-15 cm/s range, while closer to the deltas, velocities of up to 20 cm/s were common. Most of the profiles taken near the ice face indicated an increase in velocity at the base of the profile, consistent with underflows. This is possible, since underflows travelling quite long distances have been recorded elsewhere (Gilbert, 1975), and the maximum distance these flows would have to travel is approximately 1 km. Peaks on the velocity profiles were also seen at several points within the water column, indicating interflows (figure 42).

One problem associated with the water current measurements was the lack of indication

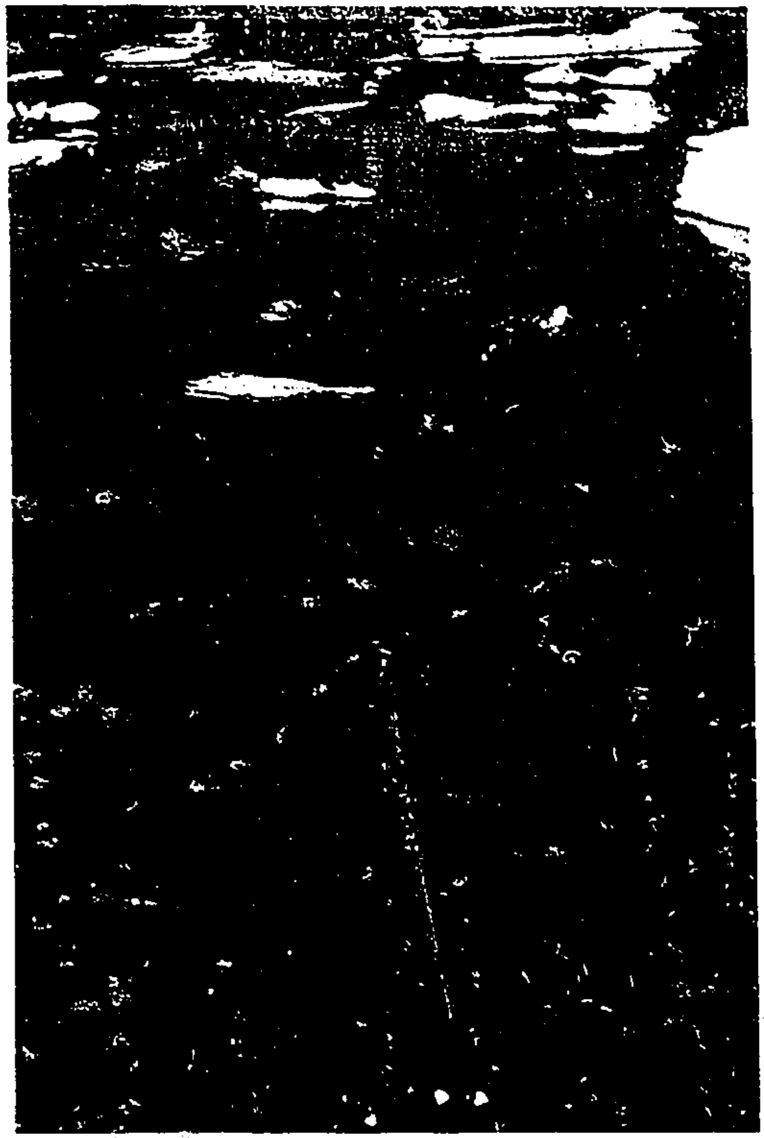

Figure 39: *Close-up of turbulence in the N.S.G. deltaic area due to ice.*

of direction of flow. At the deltas, this was less important, since the direction was assumed to be away from the source. However, in the areas closer to the ice front, some idea of direction of flow would have helped to distinguish between influent currents originating from

Figure 40: *Overview of N.S.G. and 3rd G. deltas, with ice blockage.*

the ice face, remnants of currents flowing from the deltas, and currents formed by water leaving the basin by subglacial leakage. However, this was impossible with the equipment used. The only indication of the possibility of a flow from the ice margin was a sharp change in a temperature profile in an area where sediment plumes could be seen. Unfortunately, a technical problem arose, and the velocity could not be measured to supplement the temperature data and confirm the presence and strength of the suspected current.

Lake temperature profiles showed results fairly consistently around 0.5°C. Temperatures tended to stay at or below 0.5°C until mid July, then increased slightly to between 0.5°C and

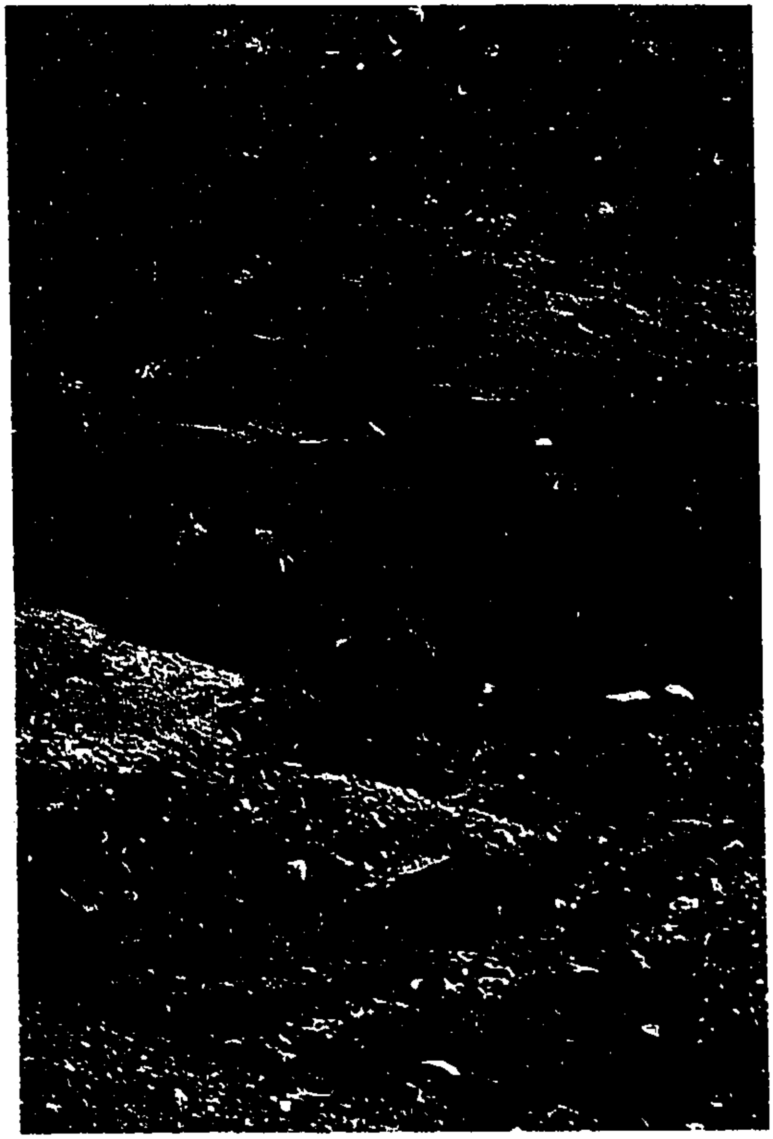

Figure 41: *Plunge line on the Non Surging Glacier delta.*

1.0°C after that. The profiles taken close to the ice margin tended to stay at 0.5°C, with very little variation. Temperature profiles taken closer to the deltas exhibited noticeable warm peaks at the base of the water columns (figure 43). These were caused by underflows moving

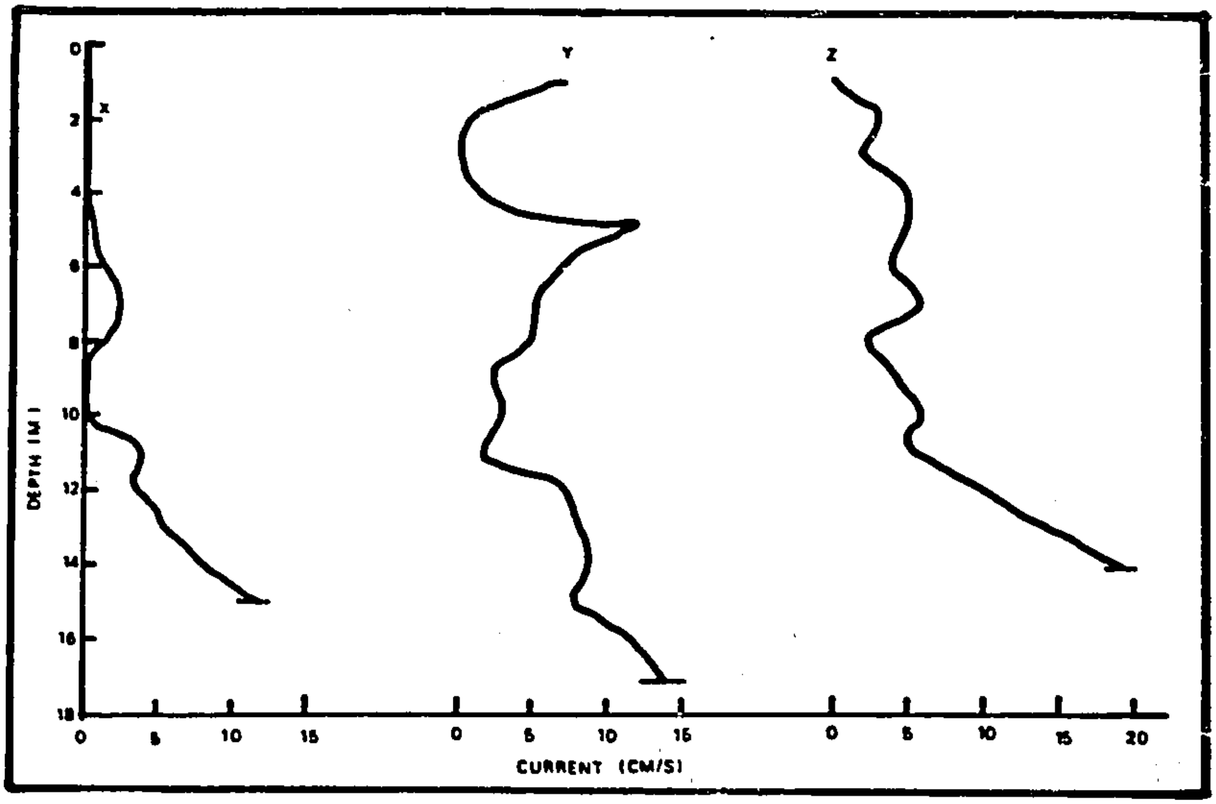

Figure 42: *Representative lake current profiles.*

into the lake. On a few of the profiles taken near the ice margin, temperature peaks registered close to the surface. These were thought to be caused by inflowing water from supraglacial and englacial streams.

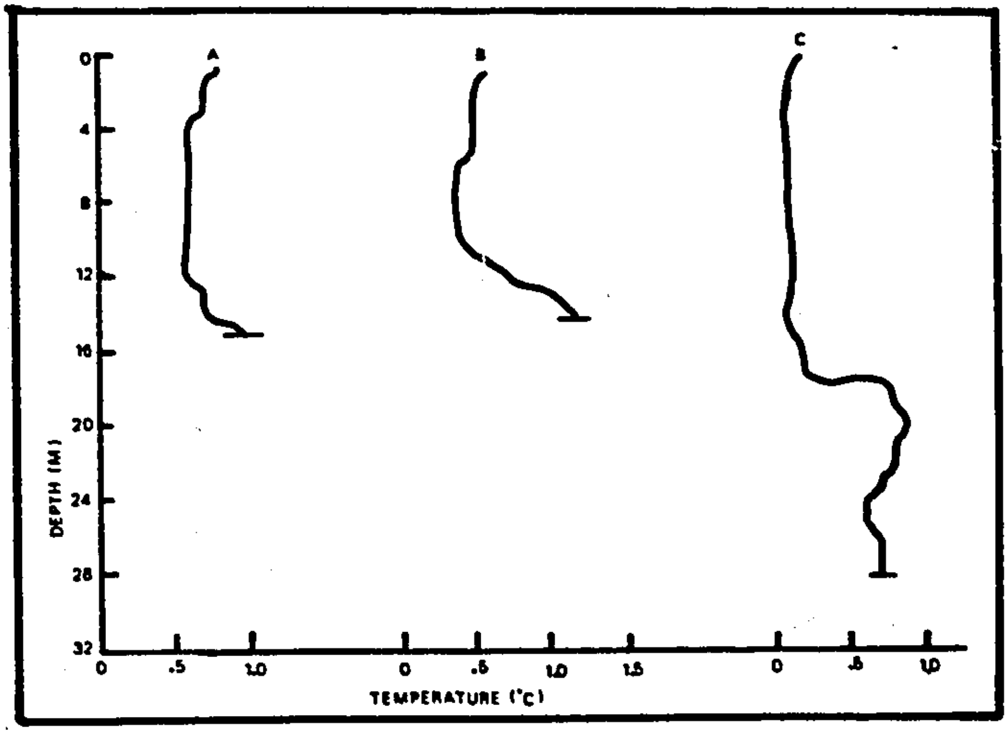

Figure 43: *Representative lake temperature profiles.*

6.4.3. LAKE BASIN TOPOGRAPHY

The basin can be divided into 2 sections, and the dividing line is the main stream channel leading to the ice face (figure 44). On the east half of the basin there is relatively little relief, with the area sloping down from a terminal moraine ridge south of the Surging Glacier stream and from the eastern boundary shoulder towards the ice face. The lake floor slopes gently, if a bit irregularly, towards the ice face, and deepens rapidly close to the face.

The other half of the basin lies west of the main stream channel and the 3rd Glacier stream. There are several relatively large kettle lakes in the north part. The eastern edge is

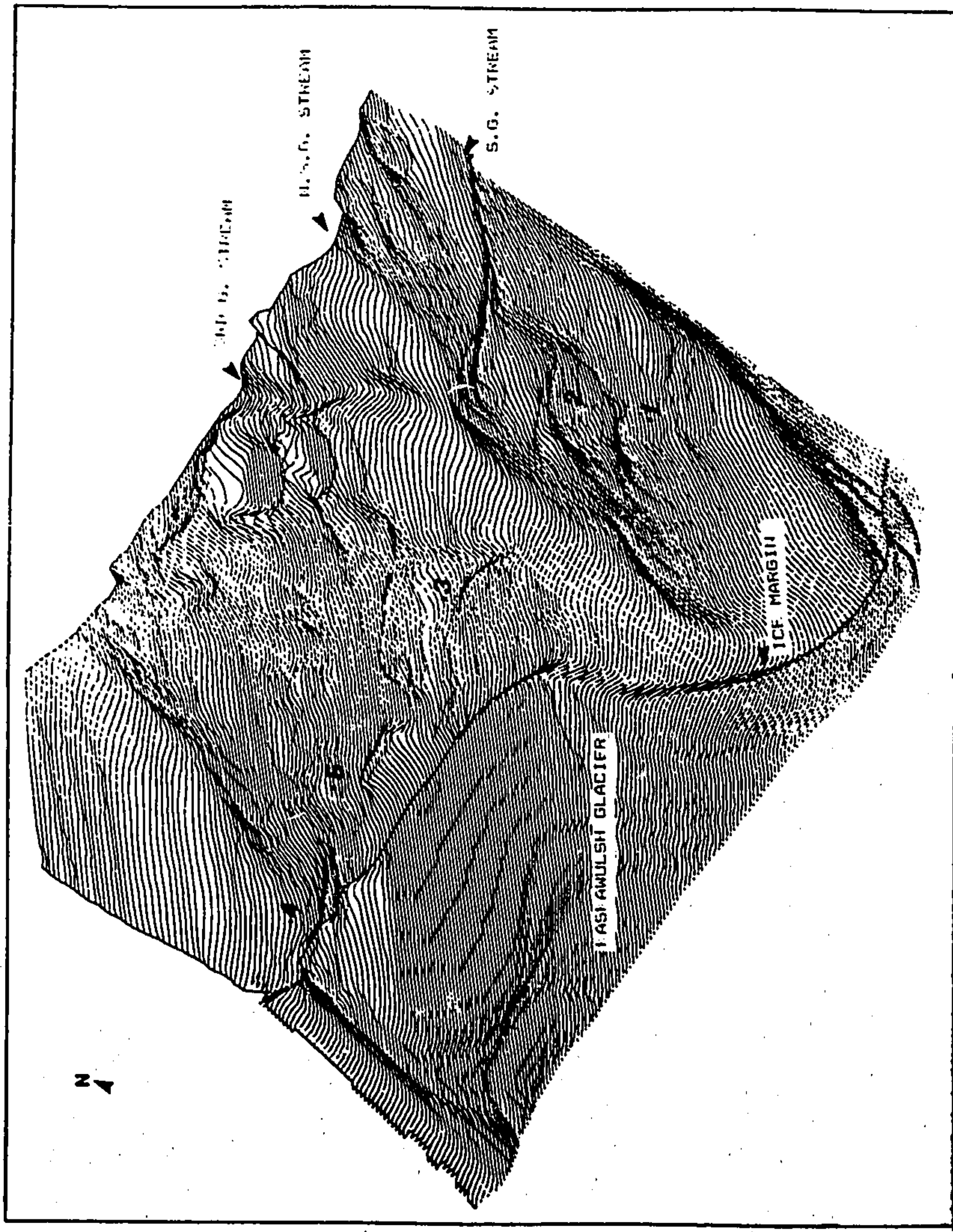

Figure 44: *Lake basin topography, with lake bottom core locations.*

marked by a 15 - 20 m escarpment running along the edge of the main stream channel. The section adjacent to the ice face is relatively flat, without any major depressions. This results in a fairly shallow zone when the lake is full. The only variation in this flat area is a depression in the southwest corner. It is 15 m. deeper than the surrounding area. The western section adjacent to the shoulder is gently undulating, with most of the topographic relief downscaled by ice core melt.

6.4.4. DEEP WATER DEPOSITS

In normal glacier-fed lakes, the deposits in the deeper parts of the lake would be stratified. Rhythmites would combine the coarser material added by turbulent underflows during the summer with the very fine material settled out of suspension during the winter when the lake was frozen over and the input minimal. Lake bottom deposits are mostly multilaminated beds. The closer the lake bottom deposits are to the sediment input source, the more likely they are to contain coarser laminae deposited by underflows, whereas the distal deposits would only contain the finer material deposited out of suspension from overflows and interflows. In deposits where both the coarser and the finer materials are both present, a sharp contact between the 2 units shows the abrupt change in the method of deposition. However, this sharp contact is more likely to be present in proximal areas where large amounts of input occur during the summer.

A factor which must be taken into account when examining lacustrine deposits is the fact that there are rhythmic deposits of similar appearance but of different genesis. These sets of similar deposits are annual rhythmites and slump-generated rhythmites. Annual rhythmites are produced by irregular deposition extending over a year. The contact between the silt and clay beds is sharp, and isolated lenses of silt or clay may be found incorporated within the other material. The silt beds may vary in thickness, but the clay beds are relatively constant. The thickness of the silt beds is dependant upon the mechanism transporting sediments to any location, while the clay beds depend upon basin size and settling time (Ashley, Shaw & Smith, 1985).

6.5.3. DEEP WATER SEDIMENTARY SECTIONS

The deep water sequence, which normally consists of fine winter and coarse summer alternating layers, is disrupted. The fine material carried in suspension would normally settle out once the lake froze over for the winter, and input dropped to almost nil. However, with the drainage occurring in late summer to early winter, this material is lost from the basin. Thus, the thickest layer of clay-sized particles, normally deposited during the winter, would be absent. Any deep-water sequences would probably only contain the coarse summer deposits. The exception to this would be in areas of topographic depressions, where the water would be calm enough to promote some settling during the summer, or which remained as ponds once drainage had occurred. An example of this is a few kettles, annexed during lake-full conditions but containing water year round.

Several deep-water cores were collected after the lake drained. They were taken in topographically isolated basins to either side of the main channel (figure 45). Because of the onset of drainage, and the subsequent downcutting of the streams in the empty lake basin, any deposits located near the main channel were eroded. Thus, the only locations where deep water sediments were preserved were in isolated depressions within the basin.

Five sections collected after drainage in 1987 exhibit a pattern of alternating finer and coarser material. The maximum thickness of the deposits ranged from 14 to 25.5 cm. Theoretically there would be little coarse material in these cores, since these locations would be fed solely by overflows and interflows. Any coarse material found within these cores should be of local origin, probably from small slumps off the slopes of the isolated kettle-type depressions. A seasonal limit cannot be assigned to any of the 'layers' (figure 46).

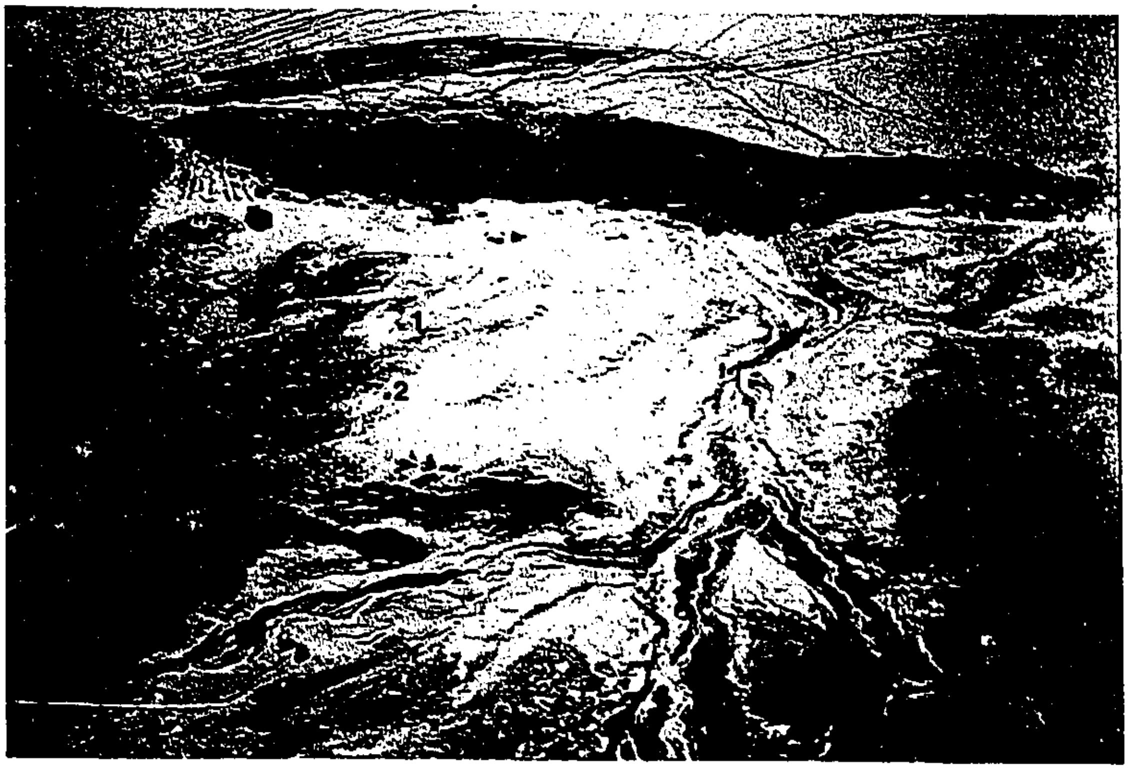

Figure 45: *Sediment distribution post-drainage & deep water sample sites.*

Sample 4 contains alternating layers of light and dark sediments. The size range of material involved is the very fine sand and silt size range. The thickness of the layers ranges from 0.4 to 2.6 cm. There is most probably a seasonal explanation for the alternation of light and dark bands, but there is no way of ascertaining the validity of such a conclusion. There is a slight difference in the material size between the coloured bands, but since grain size analysis was not carried out, it is impossible to accurately quantify the difference in the grain sizes.

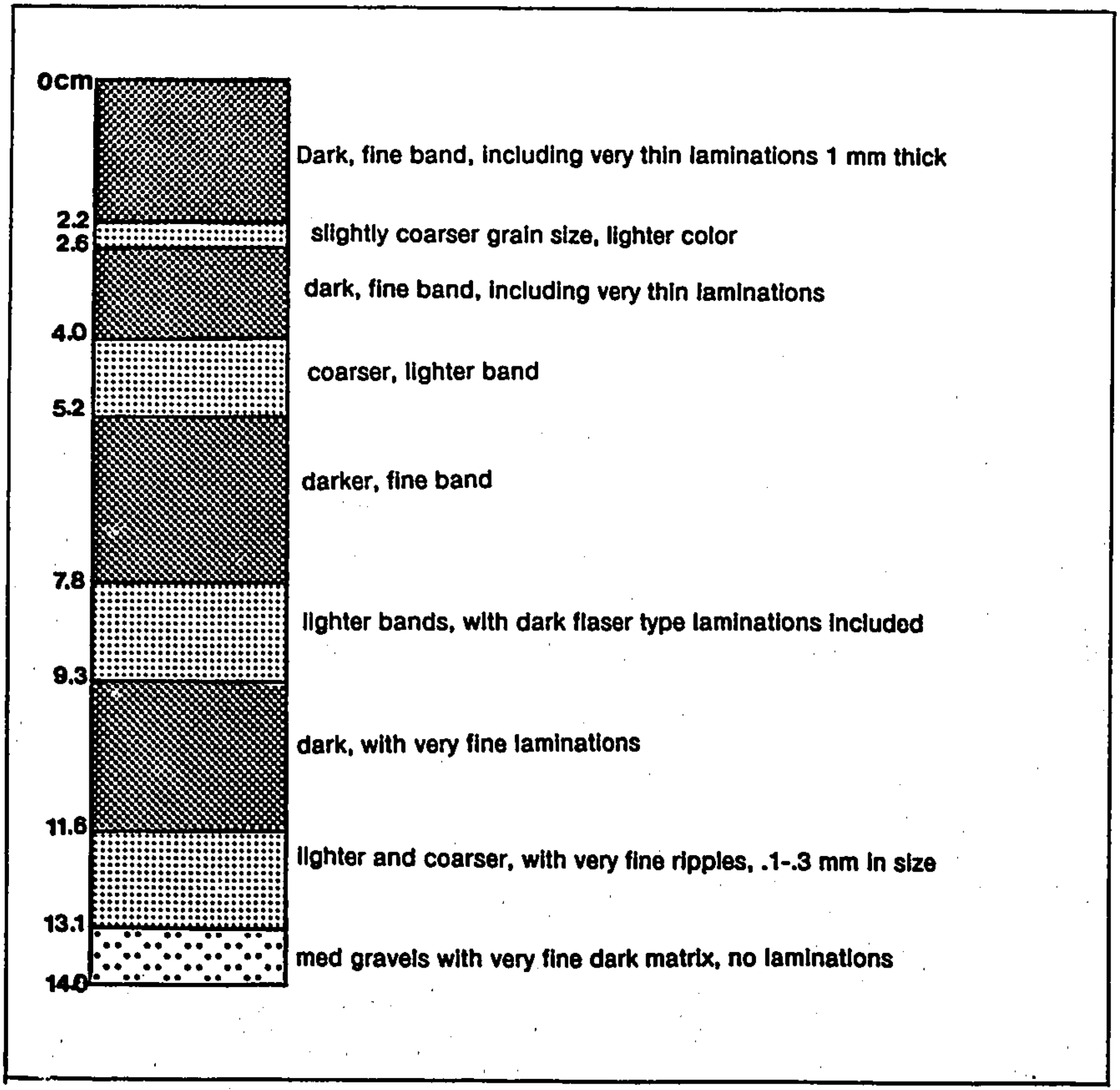

Figure 46: *Deep water sample 4 - post drainage 1987.*

CHAPTER VII

CONCLUSIONS AND IMPLICATIONS

7.1 LAKE HYDROLOGY AND THE ICE FACE

There are a number of conclusions which may be drawn from the study of the hydrology of this ice dammed lake. Several hypotheses outlined at the beginning of this research paper have been invalidated in light of the research results.

The lake was assumed to be a closed storage container which, in response to some triggering device, would drain rapidly on a yearly basis. As the system was thought to be closed, it was assumed that a reasonably accurate calculation of all contributions to the water balance of the basin could be made. Measuring the amount of input from the streams, and calculating the volume of water stored in the lake would allow the quantification of the contribution from the retaining glacier, since this was thought to be responsible for a major part of the lake's volume. The results of the water balance study show that this is not possible, since this is not a closed system. Based solely on the input from the streams, a loss to the system can be calculated. No accounting can be made for any input from the Kaskawulsh Glacier. Some contribution can be inferred by measurement of ablation of its surface. This cannot be used in the water balance equation, since there was no way of measuring the extent of this volume which actually entered the lake.

It was hypothesized that the lake water was supplied primarily by the retaining ice mass. The maintenance of the water temperature at less than 1°C was deemed to indicate this. It was found, however, that the streams can themselves account for more than the amount of water being retained in the basin. The temperature of the water remains below 1°C, but this is due to the proximity of the ice margin and to the presence of large numbers of icebergs in the lake.

The floating of the ice margin was hypothesized to be fully or partially responsible for the initiation of drainage. Contrary to this, results indicate that while it may enhance the ability of the water to access subglacial tunnels or cavities, it was found not to be solely responsible for the onset of drainage. The floating of the ice margin occurs progressively through the season, merely enhancing leakage, until drainage is initiated.

Possibilities for secondary mechanisms associated with drainage were passageways being opened on the sole of the ice margin due to glacier movement, hydrostatic pressure along the base of the ice margin forcing cracks to open, and thermal exploitation of planes of weakness. Evidence suggested that all of these mechanisms may contribute to some extent. There was movement of the ice margin both into and out of the basin, in response to the general pattern of movement of the Kaskawulsh Glacier. There was enough water accumulated in the basin to exert hydrostatic pressure on the base of the ice face. There was also enough of a temperature difference recorded at the base of the water column to initiate melting in any planes of weakness or cracks formed by either of the first two secondary mechanisms.

The results of the research indicate that the mechanism which is responsible for the initiation of drainage is the establishment of a connection between the local subglacial network and the main drainage network in the Kaskawulsh Glacier. The local subglacial network, which is responsible for the seasonal leakage, is established due to the formerly stated mechanisms, hydrostatic pressure, ice movement, and thermal heat transference. The establishment of a connection between the two drainage networks is a result of the Kaskawulsh Glacier's internal drainage network reaching maturity.

7.2 SEDIMENTATION

The sedimentation research carried out during this project has led to some conclusions about the type of deposition in a deltaic area associated with a changing water level. Deltas form as a result of a change in water conditions from a moving water body such as a stream into a standing body of water such as a lake. Normally, the position of this transition point is static, and the delta is then constructed in one location. However, in ice dammed lakes, where the level of the water in the lake is variable, the transition point becomes mobile. As the lake water level rises, the transition point migrates upstream, which then causes the delta to also migrate upstream. This movement results in a particular form of delta being created.

The distinctive set of step-type deltas overlying one another in a sequential fashion are indicative of a rising water level during deposition. The complete reversal of deposits on top of a typical fining upwards sequence is indicative of a rapidly dropping water level.

The micro-structure of the delta levels are similar to most delta forms, since the process of deposition of materials due to loss of carrying capacity is the same in these deltas as in static deltas.

There is a major difference between the lake bottom deposits in annually draining lake bottoms and in non-ephemeral lakes. This difference is that there are no sediments deposited during part of the season. Partial or complete drainage will remove most of the fine grained suspended material from the basin, which means that there can not be the deposition often associated with winter still water conditions. The disruption to the normal depositional sequence is dependant upon the time when drainage occurs. If drainage occurs early in the season, then some of the summer deposits will also be lost. In the lake under study, drainage usually occurs in late summer to early fall. Since the lake only partially refills during the late fall and early winter period, there is little chance of deposition, because most water being input into the basin at this time of year from glacial streams is sediment poor. There is little material in suspension to be deposited under still water conditions.

This differentiates bottom deposits of these lakes from other proglacial lakes. Non-draining proglacial lakes will characteristically contain rhythmite deposits, some of seasonal origin, and some formed from variable sediment input rates. Ice dammed lakes will only have rhythmite-type structures formed from variable sediment input rates, such as pulses in sediment load, turbidity flows or slumps off the delta fronts.

Possible further research which would aid in the complete interpretation of all processes operating in this basin is the petrographic analysis of the lake bottom sediments. This would make possible the establishment of the direction of water movement within the lake. With the geological contrasts in the basin and a third geological type transported into the basin by entrainment by the Kaskawulsh Glacier, there should be sufficient differences in the sediments to make conclusive statements about the patterns of movement and the source of the material. This would enable a full assessment of dispersal mechanisms to be made, and an estimate of the amount of material being contributed by the Kaskawulsh Glacier. Sediment is being input via englacial and supraglacial streams, but subglacial sources are also a possibility. The input of sediment from supraglacial streams is visible in photographs, but is usually too small a quantity to be recorded in a sampling program. However, a comprehensive recording program, using very sensitive recording equipment, set up at a known input site would help to discover how much material is being input from one of these streams.

7.3 IMPLICATIONS FOR THE FUTURE

To conclude this research paper, this next section will focus on changes to the hydrological and glaciological controls which will affect the processes operating in this basin. There are several factors which will remain constant, yet others will change over time, taking into account the present climatological conditions.

The sources of input are constant, but the magnitude of the contribution from each source varies. The storage capacity of the basin remains constant, but the volume contained changes in response to factors such as the rate of input, the rate of leakage, and the initiation time for the drainage. The main factor which controls the amount of water being stored is the rate of leakage. This varies in response to the rate of input, the accessibility to open tunnels, the temperature at the base of the water column, and the rate of lifting of the ice margin.

The amount of input from the 3 cirque glaciers and the ice-cored moraines will change significantly over time. The 3rd Glacier is already stagnant and downwasting. Over time, with the present climatic trends, this process will continue, leading to the disappearance of the entire ice mass. The N.S. Glacier is slowly retreating as well, but at a lesser rate than the 3rd Glacier. The Surging Glacier is of a cyclic surging type, but the maximum extent of the glacier at the peak of the present surge is much less now than in past surges.

This trend indicates an overall decrease in the amount of water available to contribute to the system. Eventually, these glaciers will disappear, unless a major climatic reversal occurs. The ice-cored moraines will be slower to disappear, since their debris cover acts as insulation. However, meltout is already occurring in the southwest corner of the basin, evidenced by the number of mudflows, slumps and collapsed slopes.

The contribution from the Kaskawulsh Glacier will probably increase, since it too is downwasting. With the present trends towards general global warming, the retaining ice margin will downwaste to the point at which it can no longer retain water in the tributary basin. By the time that sufficient melting has occurred to bring the ice margin to this point, perhaps one or more of the cirque glaciers will have disappeared, leaving only very limited output from the tributary basin.

At this point in time, any meltwater will flow directly out of the basin to become incorporated into the glacier's main drainage network. The ice cored basin will be exposed to insolation, and without the insulating water body will melt out rapidly. The stream channels will adjust permanently down to the base level they attain after the lake drains. It is also expected that they will cut down through any deposits sitting on the lip of the hanging valley.

The storage capacity of the basin is relatively constant. There is evidence that in the past the basin's storage volume was greater than at present. This is seen by strandlines located higher up on the valley shoulders than those generated at present. However, this was probably synchronous with a much thicker ice margin. Evidence to support this idea is provided by the ice contact deltas suspended on the main valley wall 50 m above the present ice level.

In the recent past, the volume appears to have remained relatively constant. Only three years of monitoring are the basis for this observation. The water level from 1986 was only 1.2 m higher than in 1987. The amount of water which this represents is relatively insignificant, since there was a full month's difference in input between the two years.

The re-entrant ice margin appears to be relatively stable, fed by movement of the glacier down from the St. Elias Icefields. The maximum extent of the re-entrant has been noted by the presence of erratics brought in from further up the Kaskawulsh Glacier. The rate of loss or retreat of the ice face due to calving has been between 20 and 40 m per year since 1951. However, the re-entrant is also being pushed forward into the basin by the dynamics of movement of the Kaskawulsh Glacier itself. A comparison of the 1951 and 1977 air photographs shows very little change in the location of the ice face.

The preservation of the tunnel openings beneath and within the ice face depends on the condition of the ice margin after drainage. Large amounts of collapse most likely indicate that the tunnels are completely destroyed and/or infilled with debris. This reduces the possibility of their being re-used in consecutive seasons. However, if the leakage occurring at present continues on a permanent basis, it may be that the retention of water within the basin may soon not be possible. Permanently established drainage conduits would ensure a continuous outflow from the basin. However, until significant downwasting of the Kaskawulsh Glacier has occurred, this should not be possible. The Kaskawulsh Glacier, while downwasting slowly, is still dynamic. The high flow rate in the main part of the glacier ensures that any channels are deformed and closed due to stress. This reinforces the idea that maturity of the Kaskawulsh Glacier's drainage system, meaning the opening of the glacier's main drainage network, is necessary for the triggering of drainage.

The study of the water balance of a glacierized basin and all controls on its behaviour can provide information on present conditions, but can also be used to predict changes expected in the future. Many aspects must be monitored over time to properly interpret trends being established. However, studying a basin over a single season provides a general picture of the processes occurring and the factors involved in the dynamics of the hydrology of a glacierized basin.

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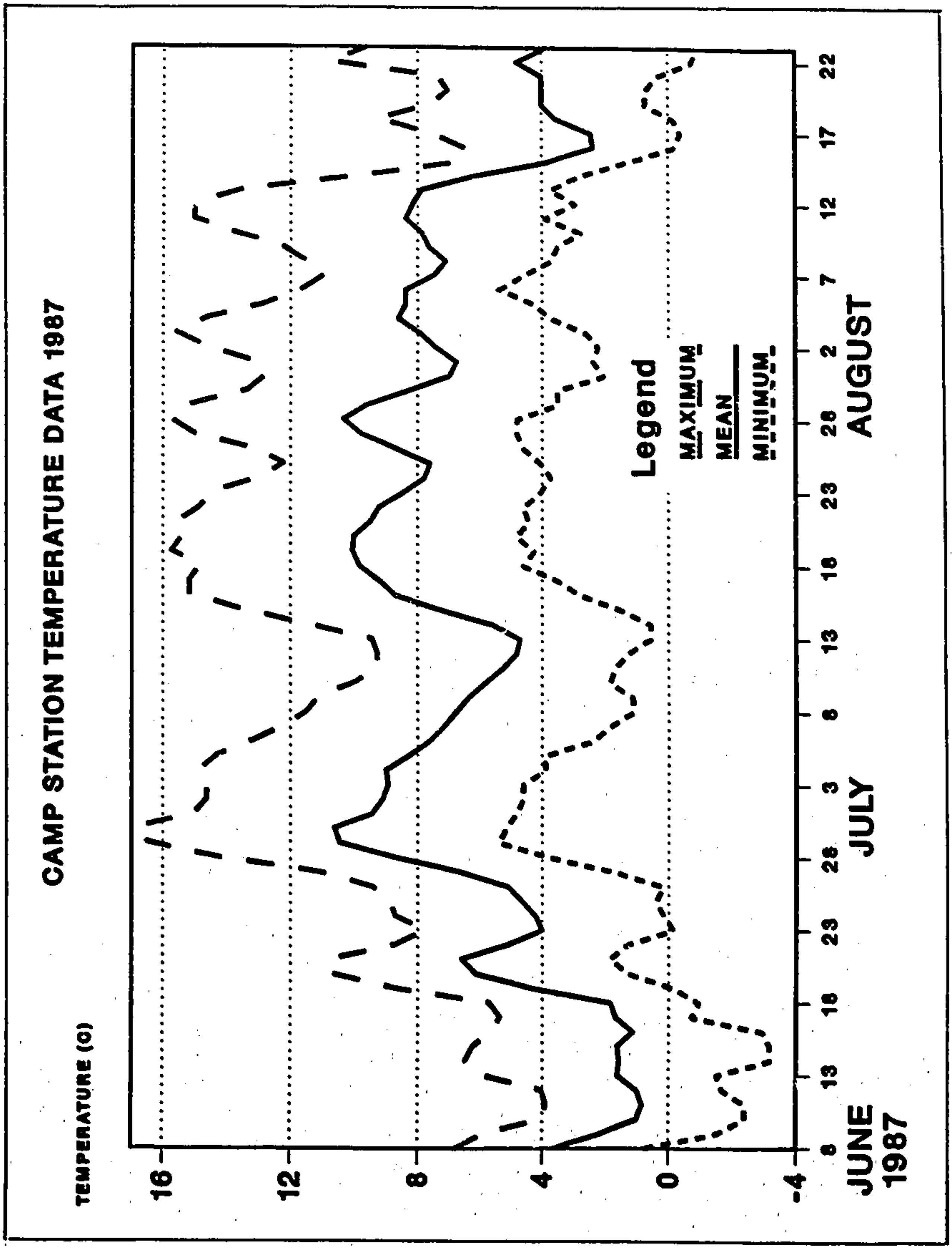
APPENDIX A

TEMPERATURE CURVES FOR ALL STATIONS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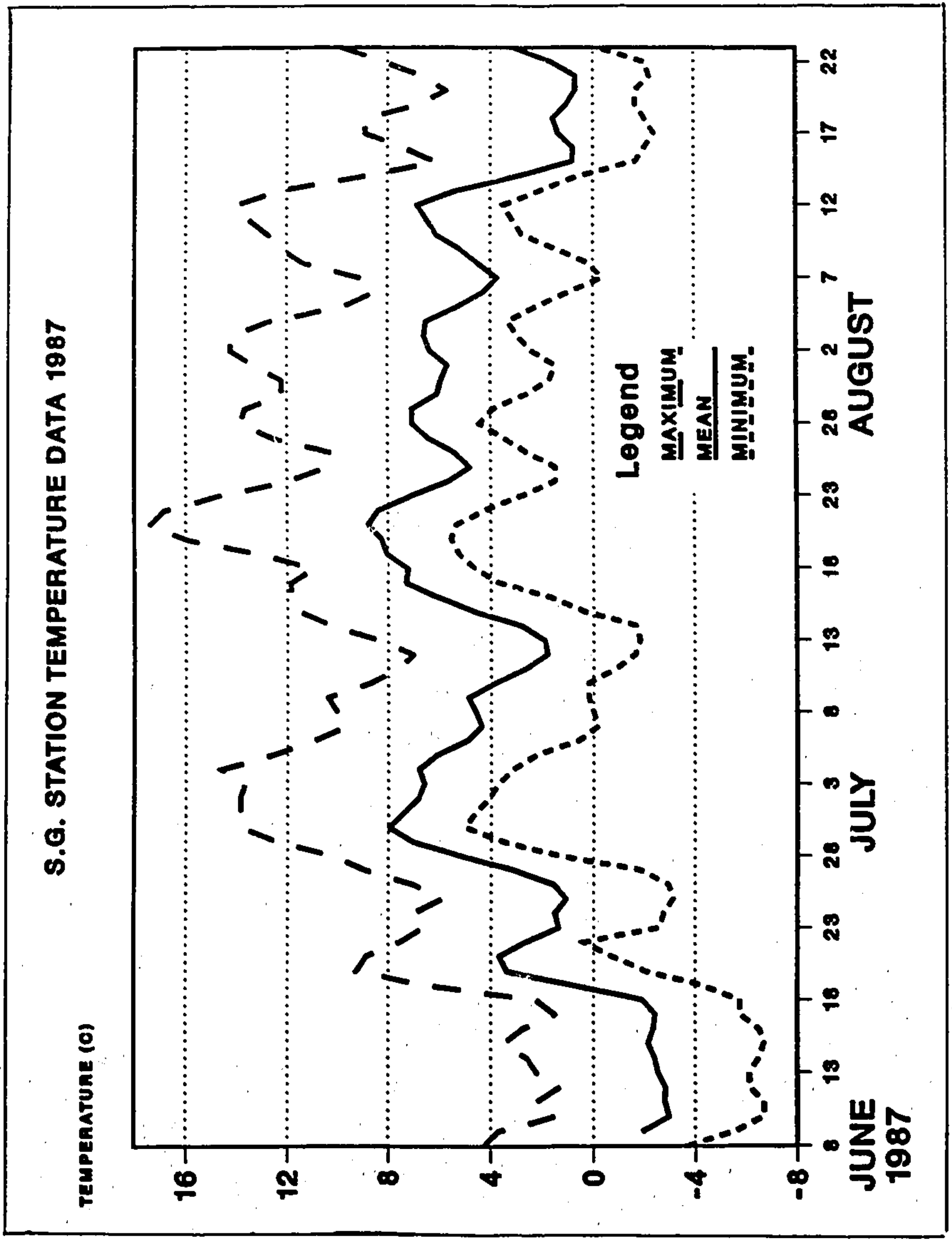
JUNE
1987
JULY
AUGUST
Legend
MAXIMUM
MEAN
MINIMUM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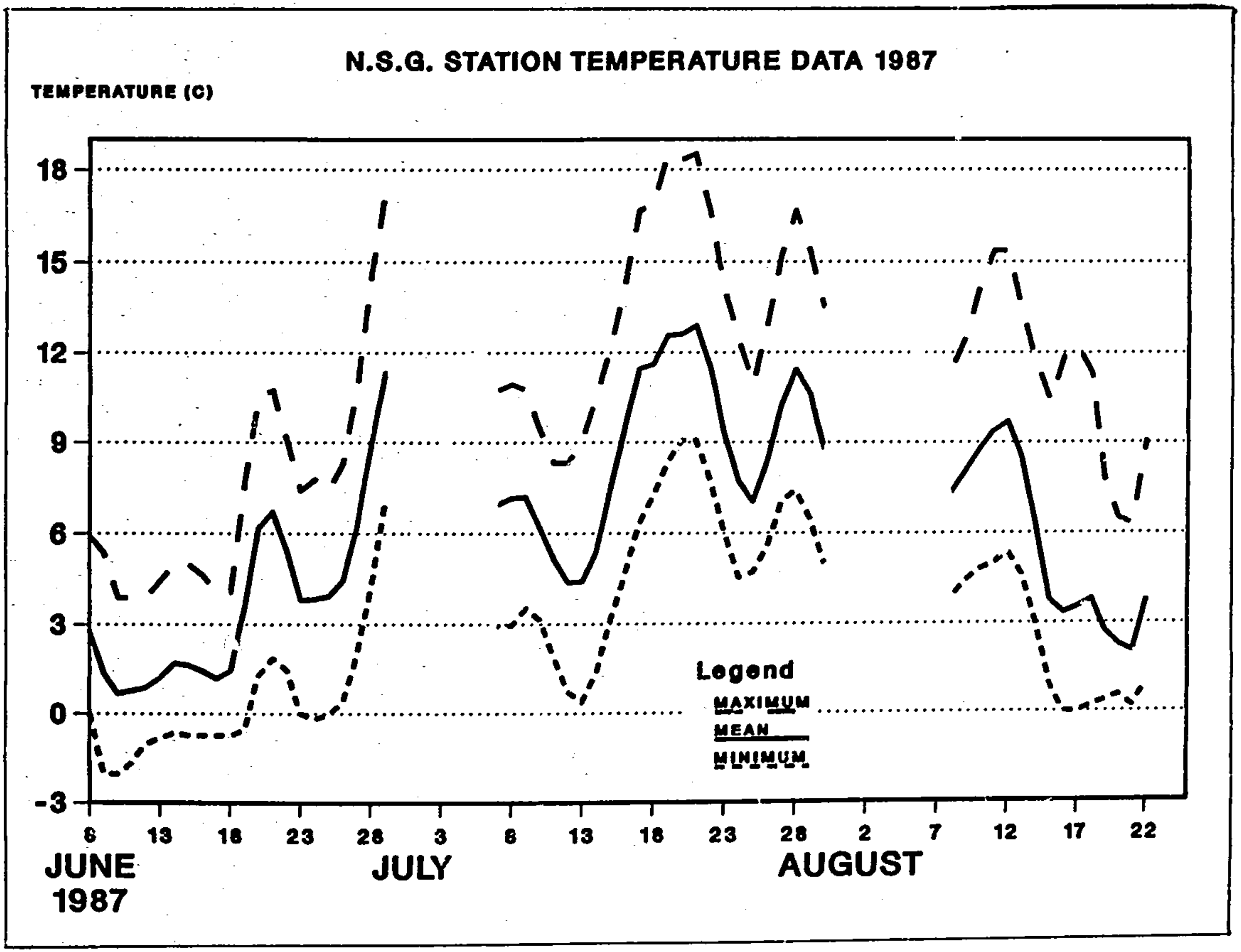
JUNE
1987
JULY
AUGUST
Legend
MAXIMUM
MEAN
MINIMUM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JULY
AUGUST
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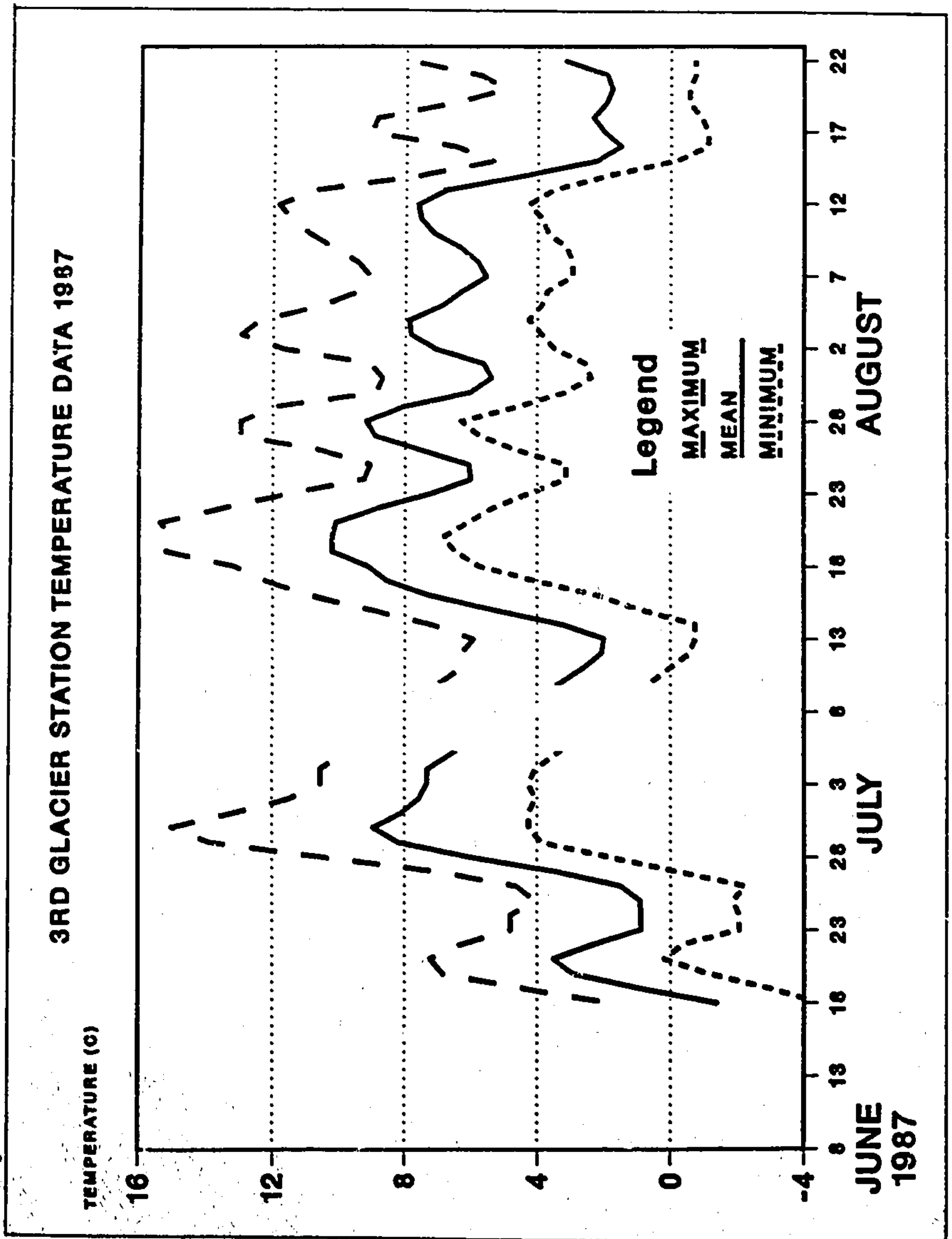
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JUNE 1987
JULY
AUGUST
Legend
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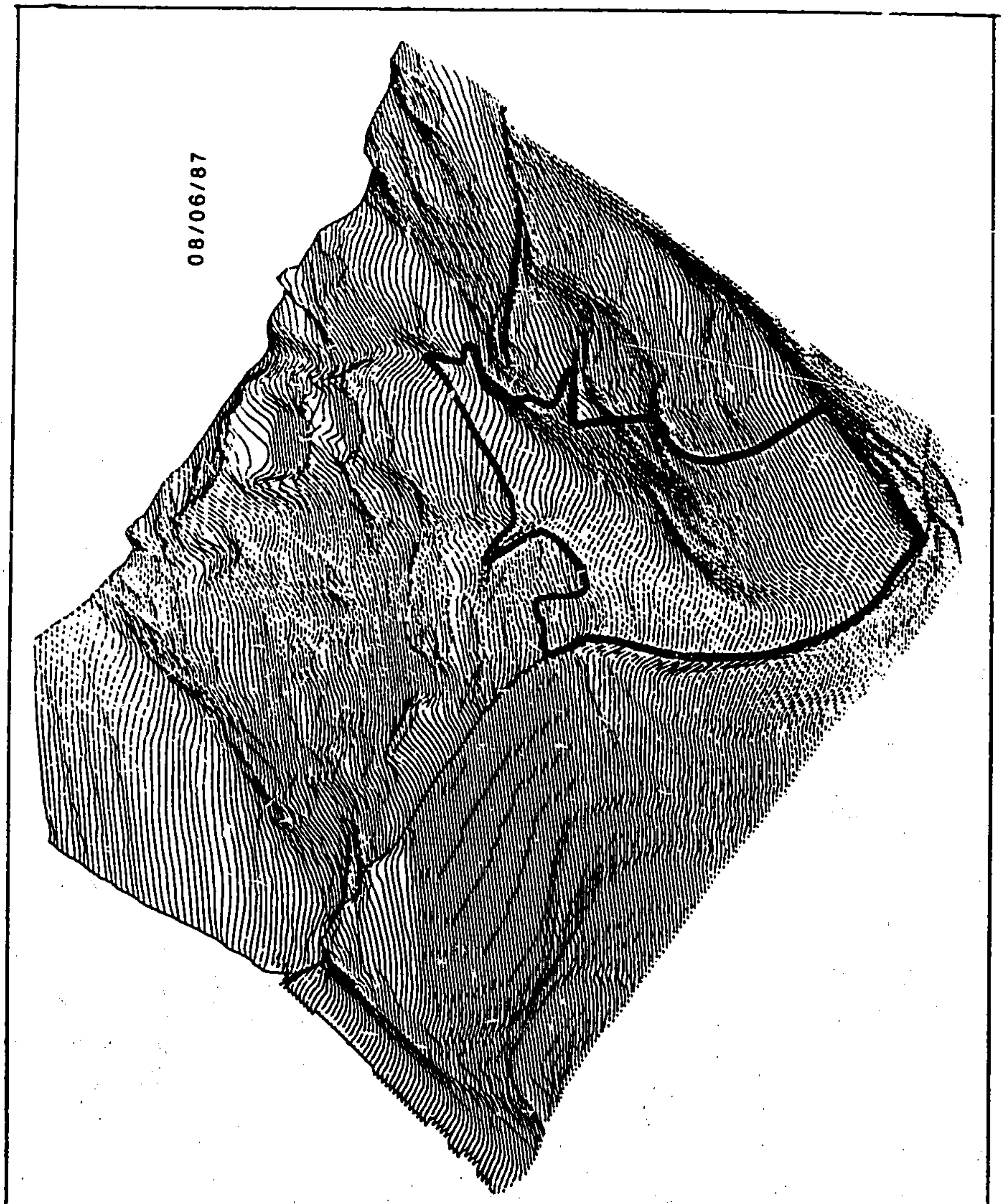
APPENDIX B

WEEKLY DIAGRAMS OF EXTENT OF LAKE, 1987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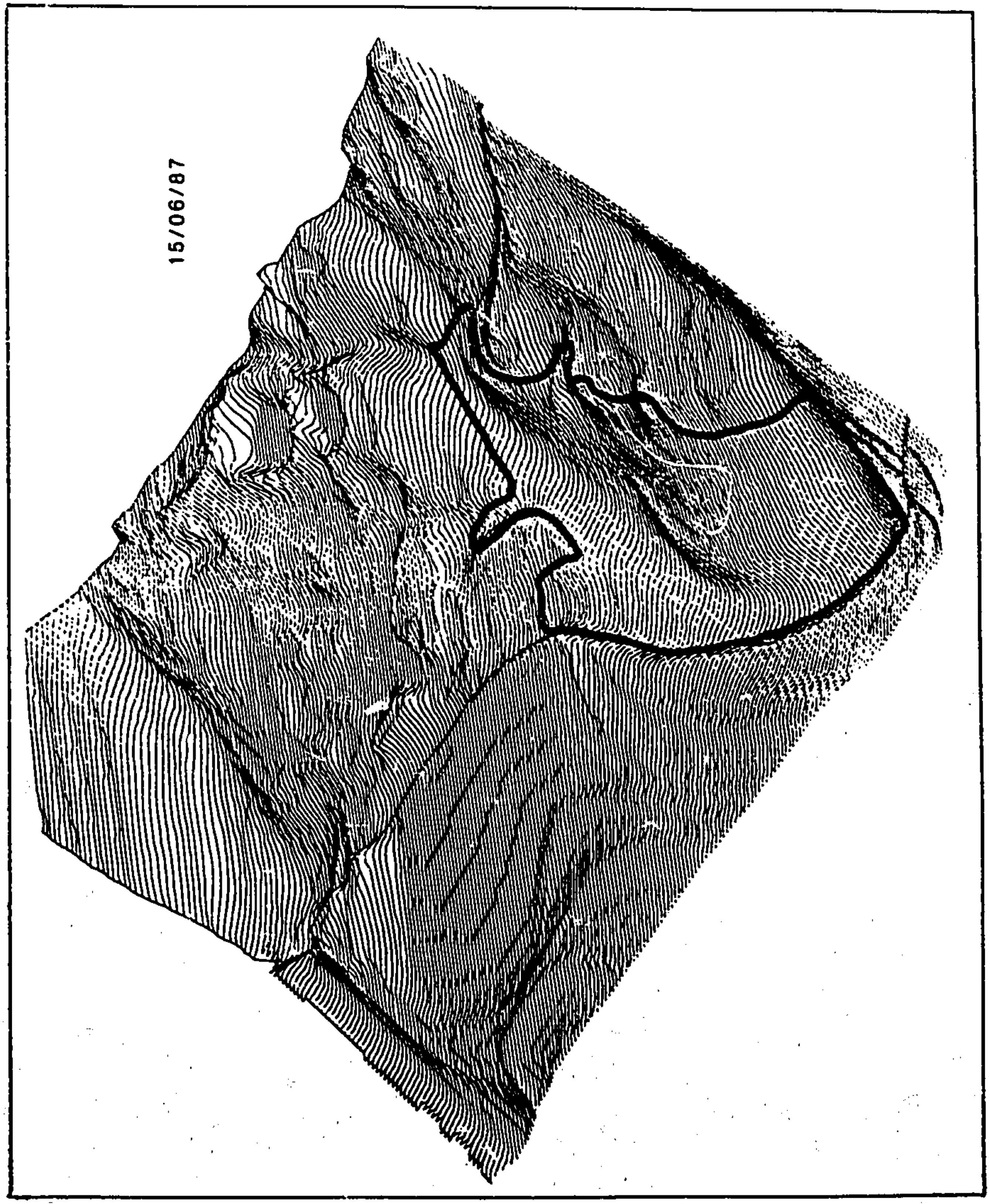
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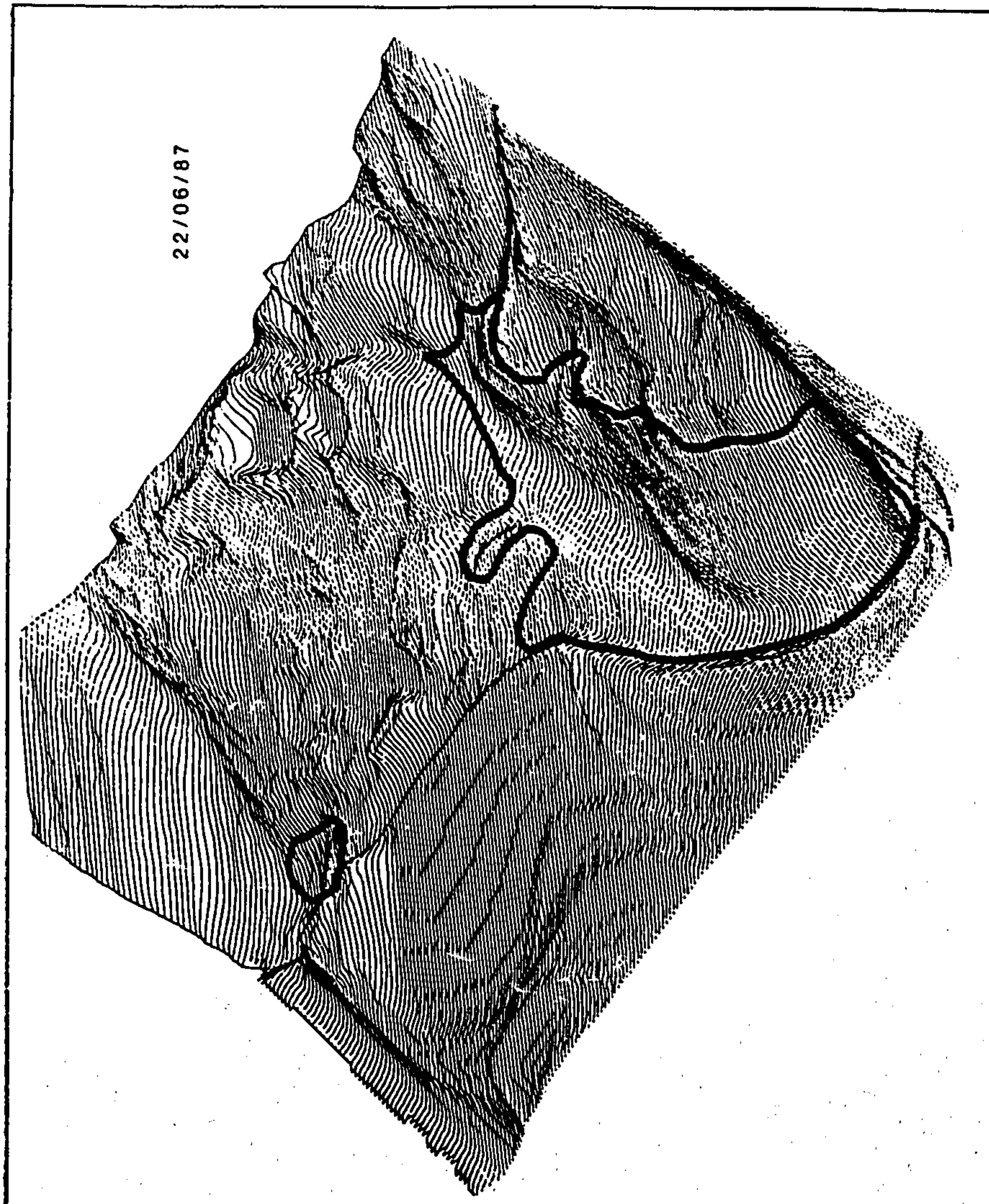
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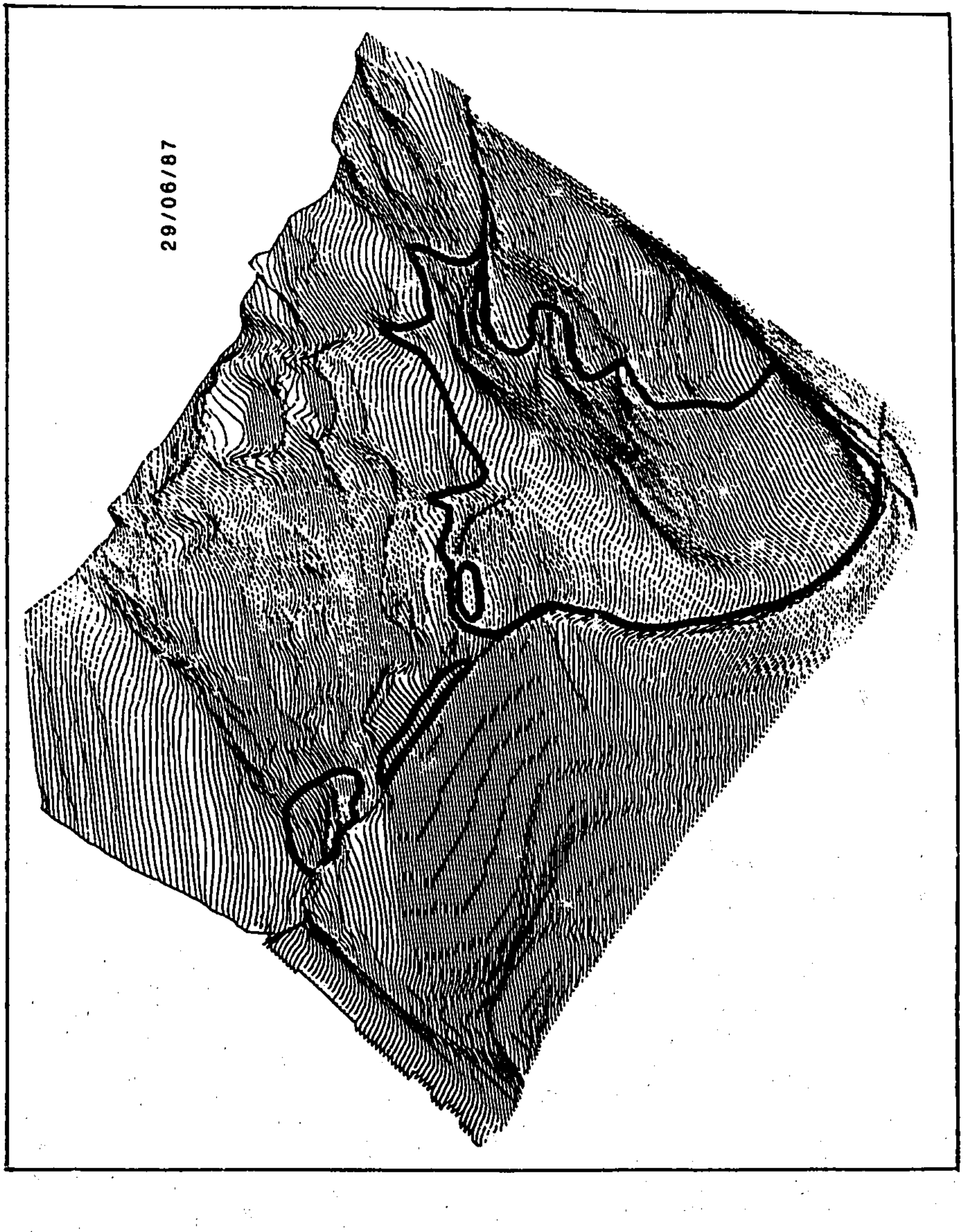
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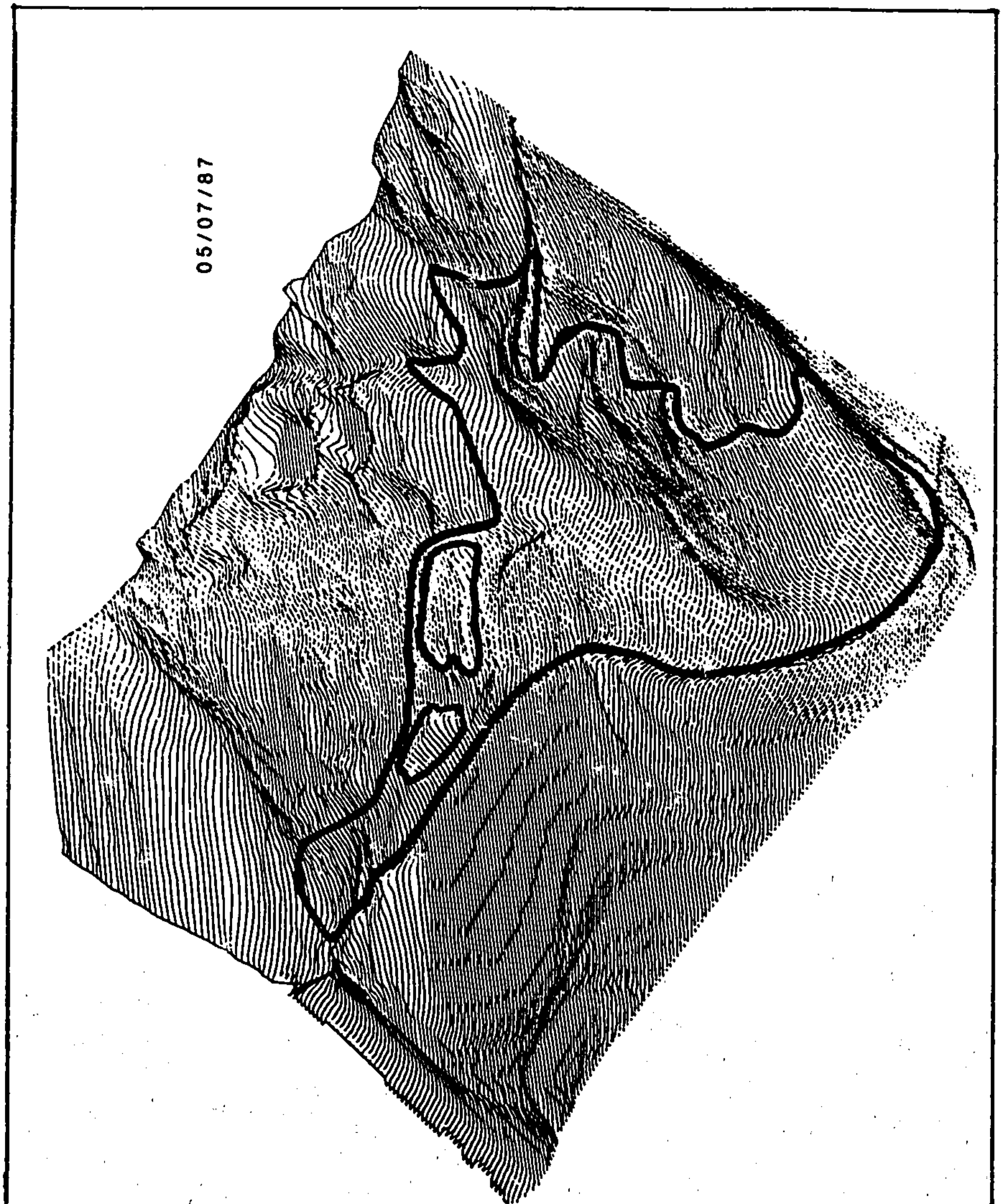
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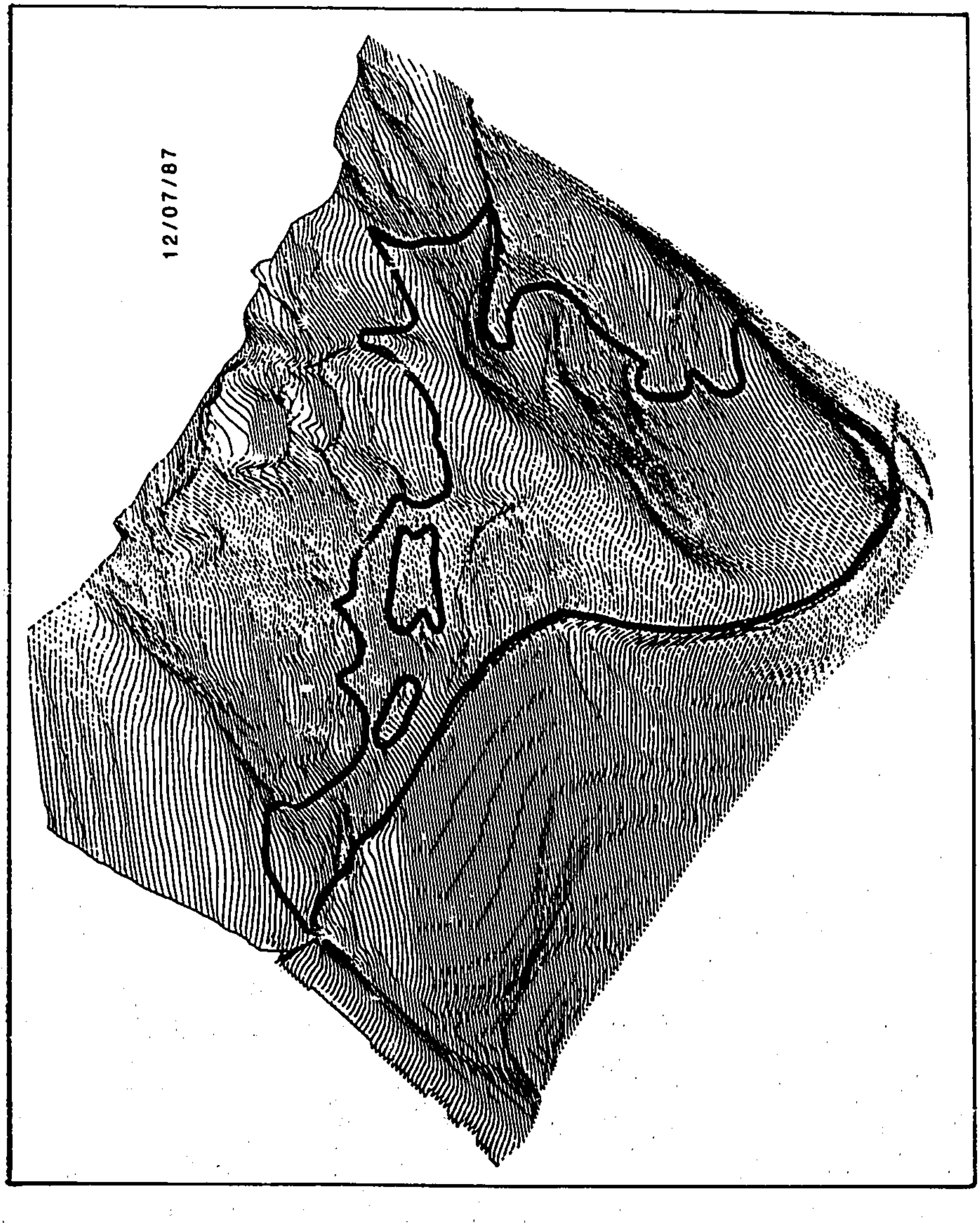
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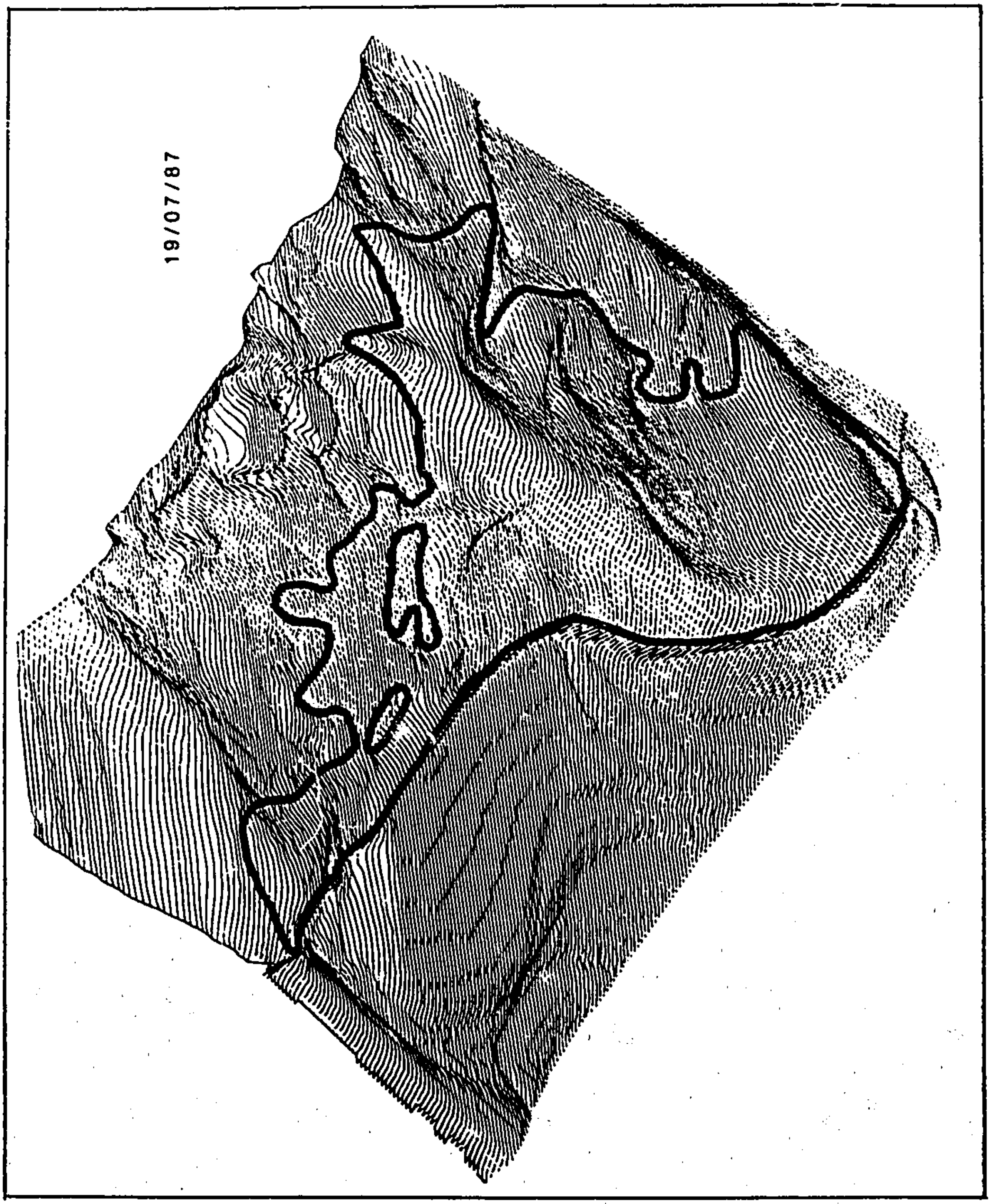
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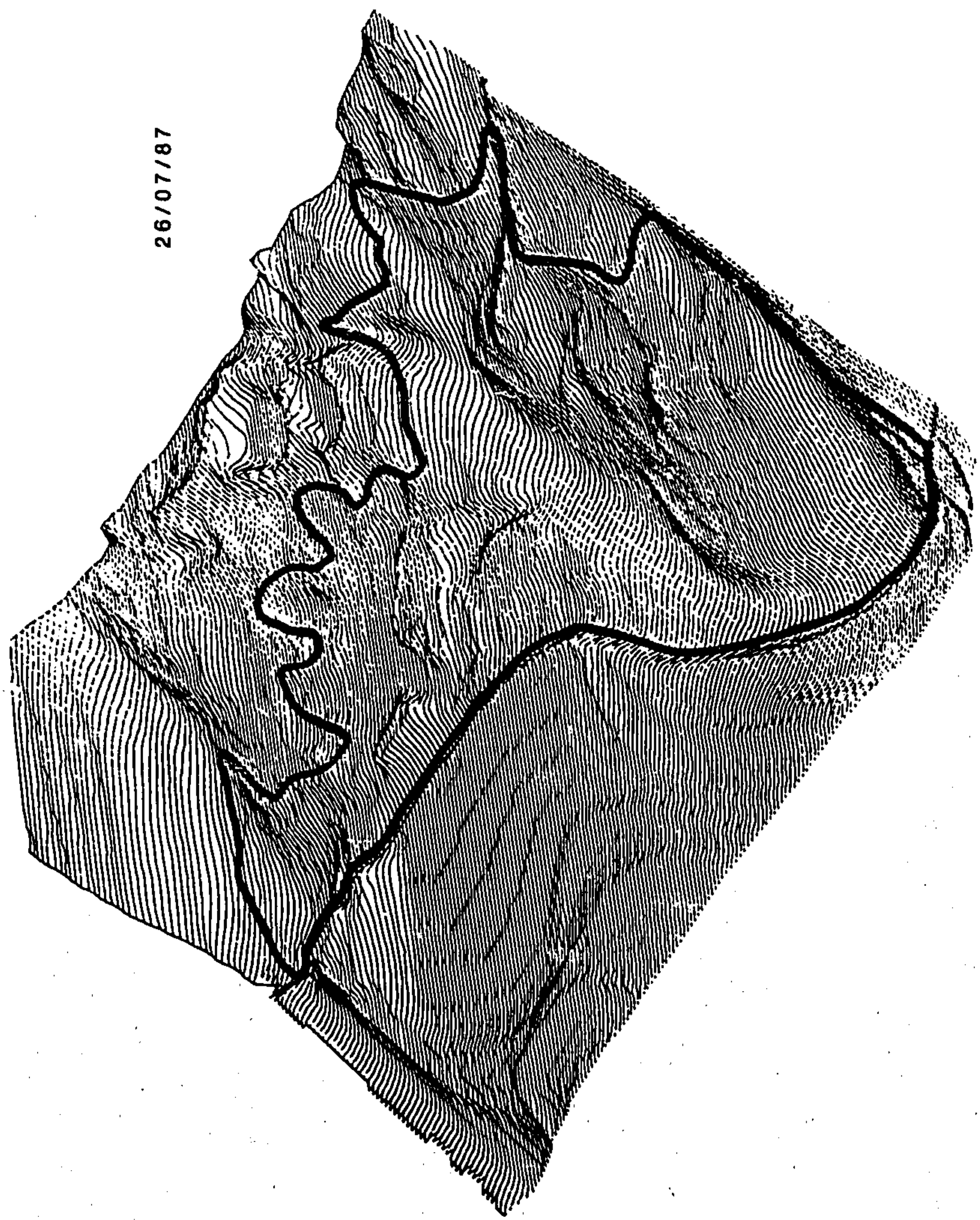
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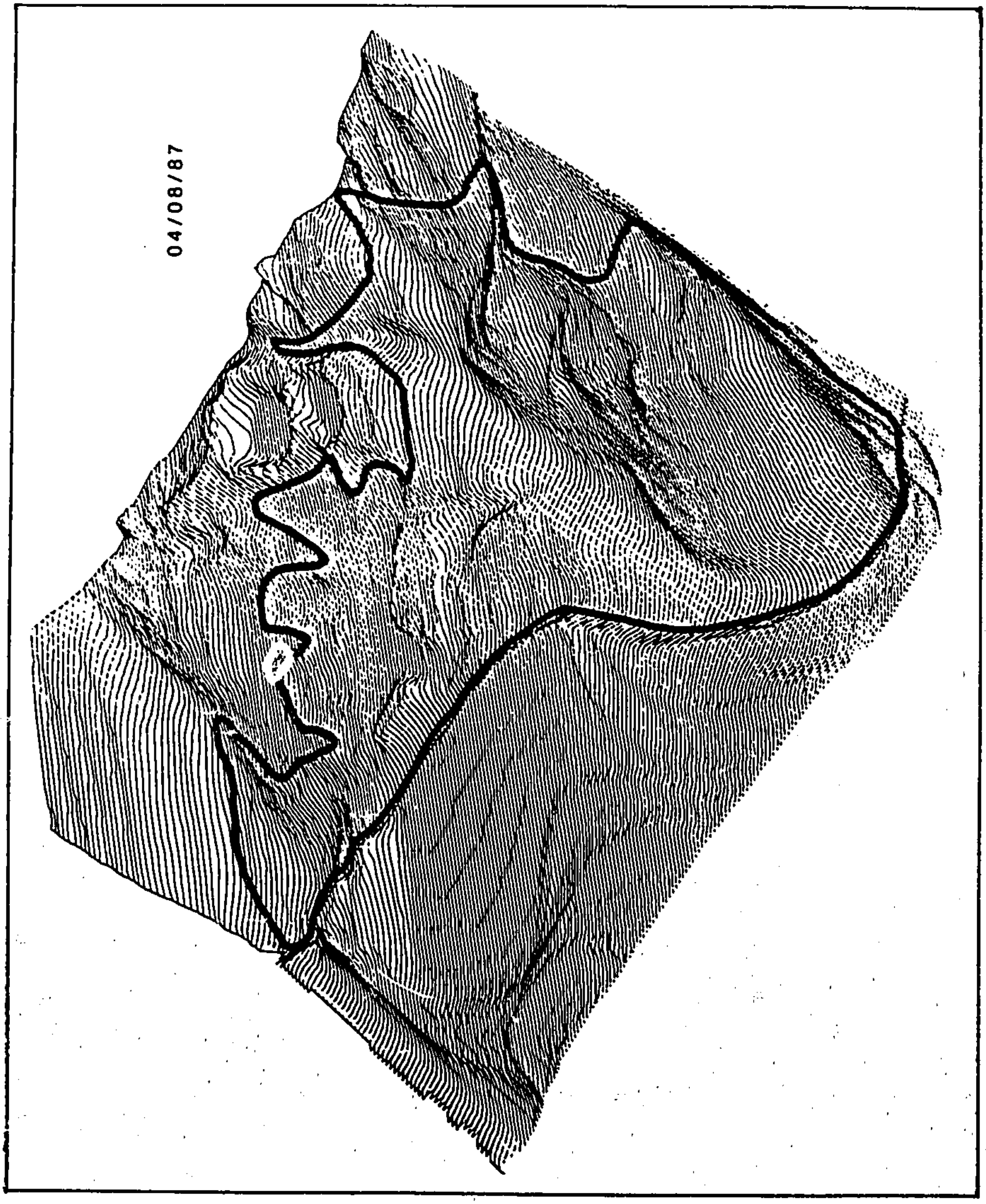
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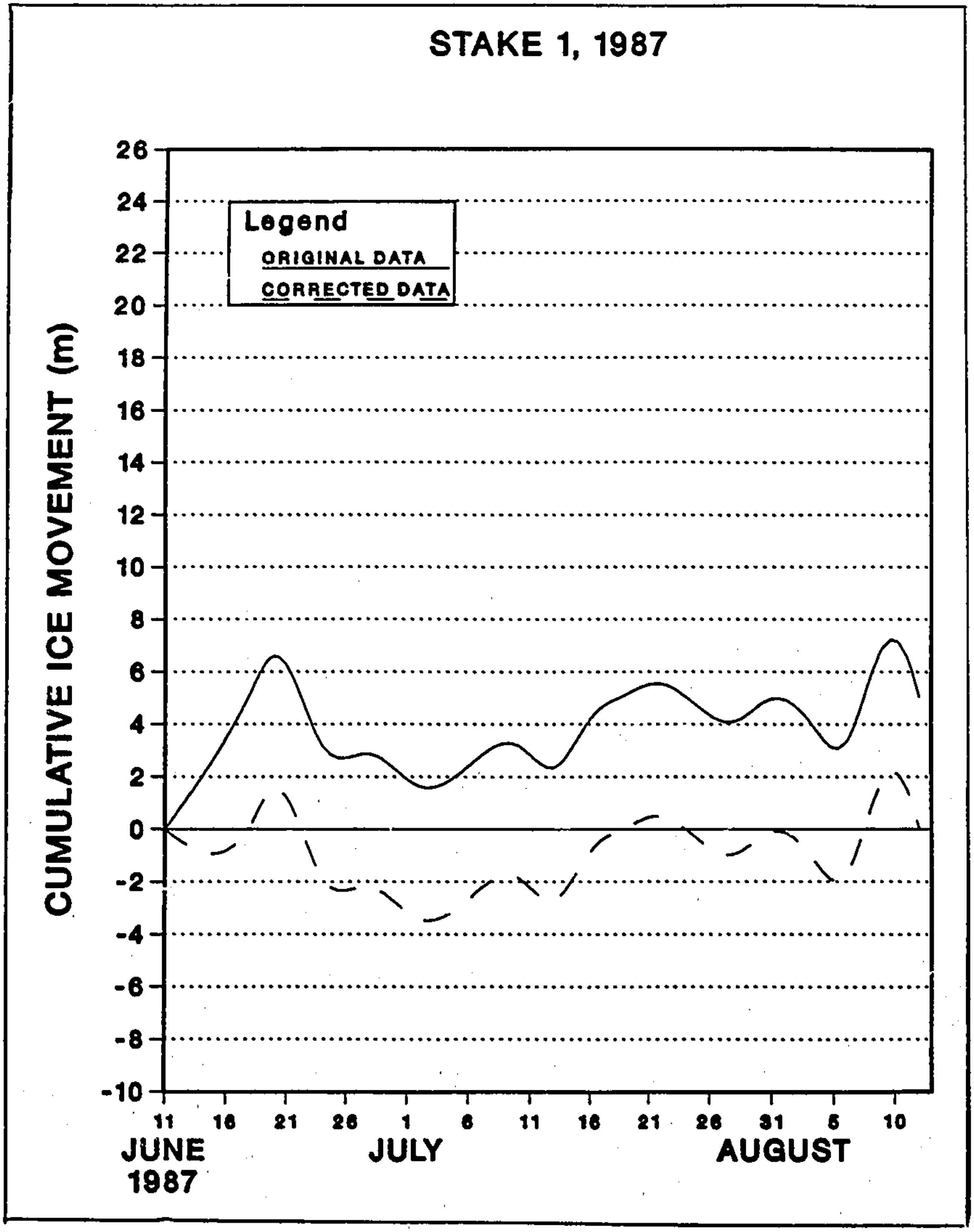
APPENDIX C

GRAPHS OF VERTICAL ICE MOVEMENT, STAKES 1 - 10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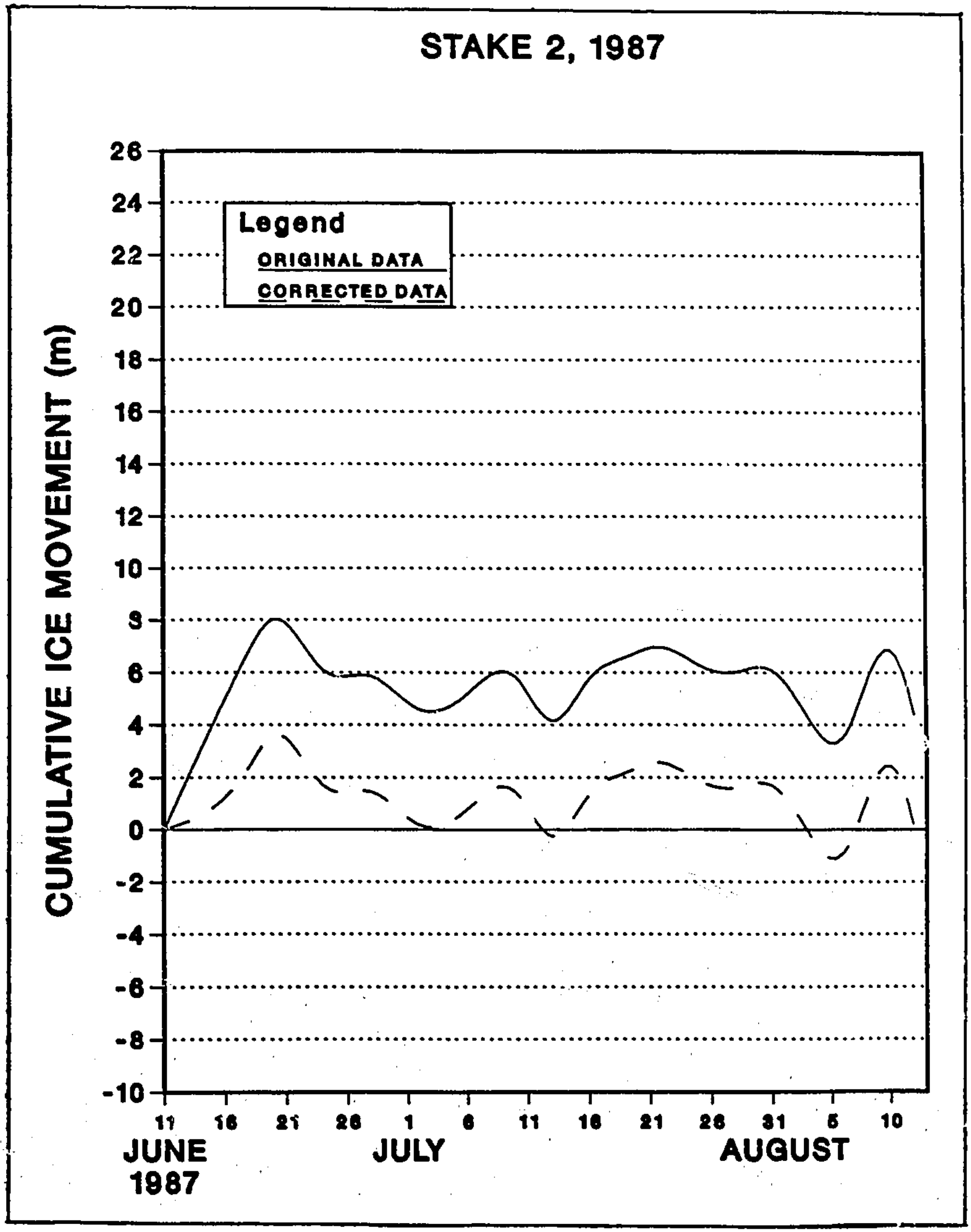
JUNE
1987
JULY
AUGUST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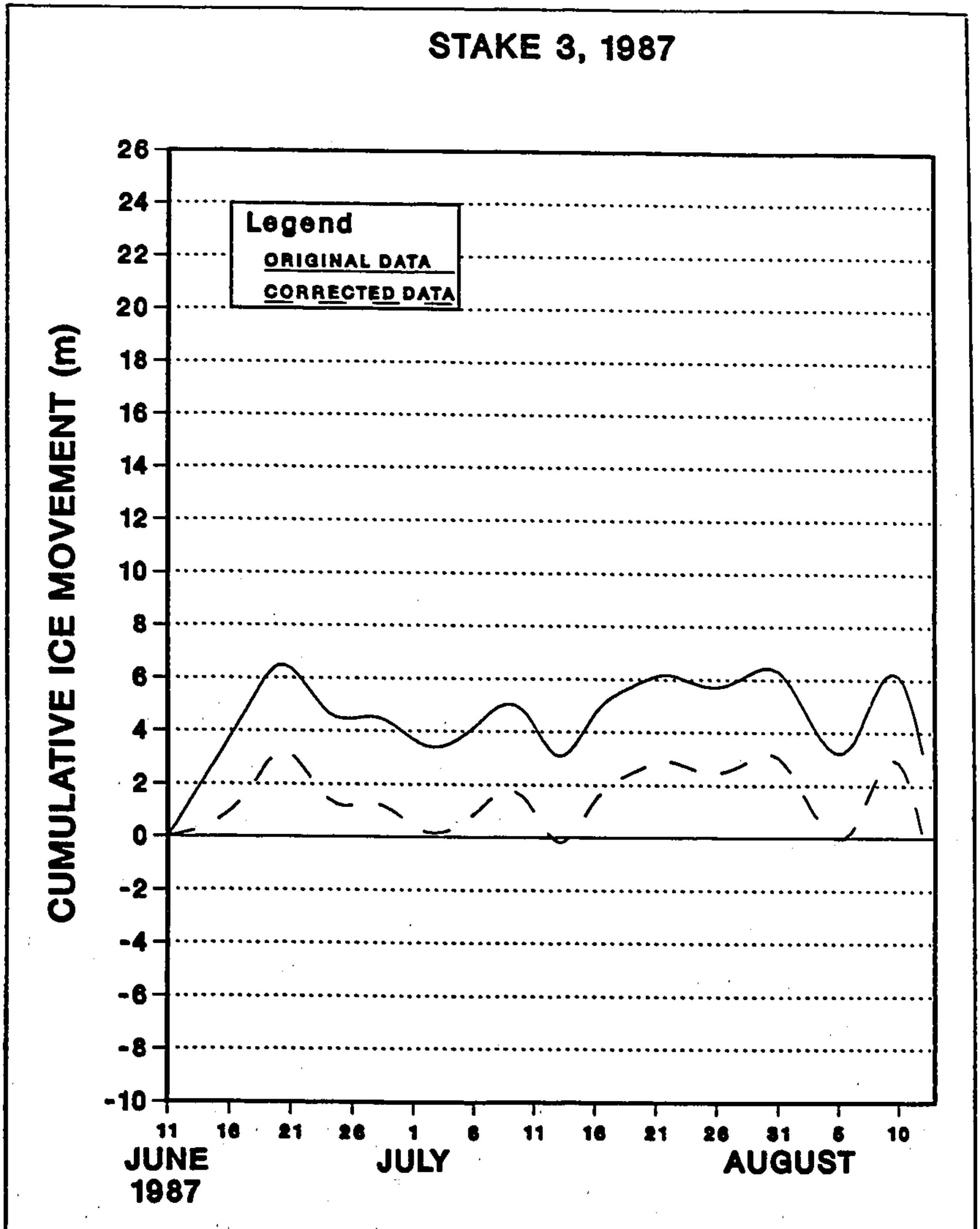
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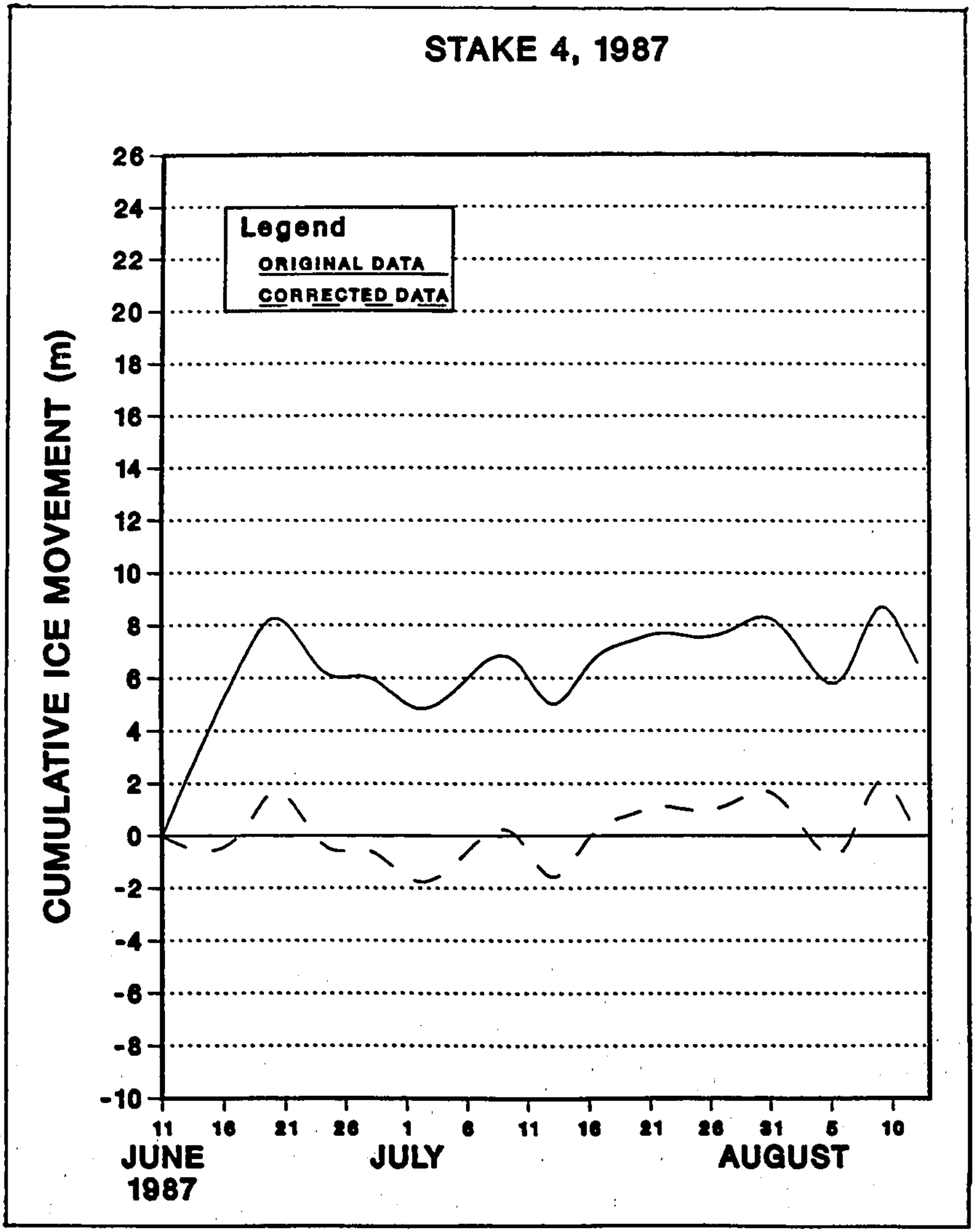
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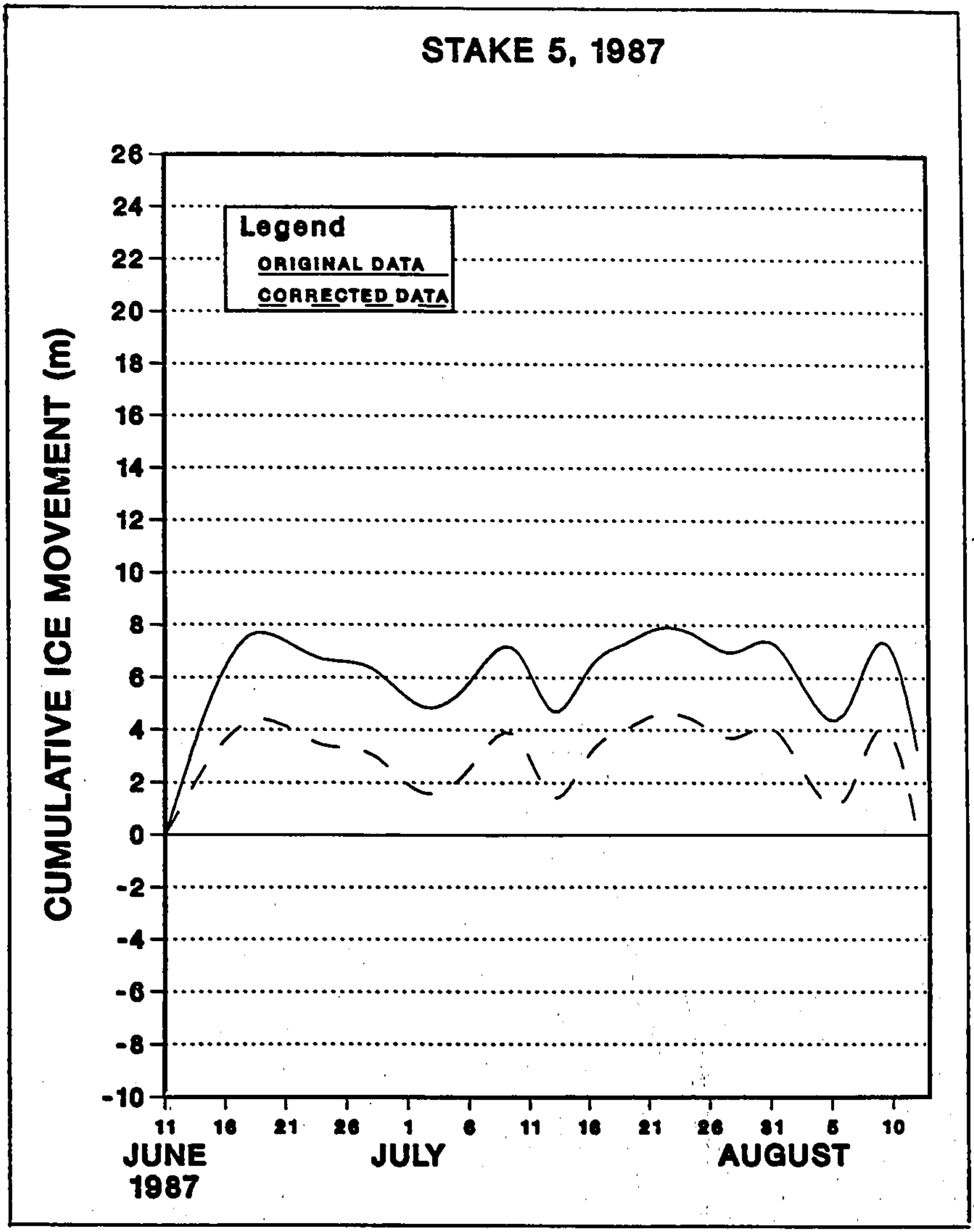
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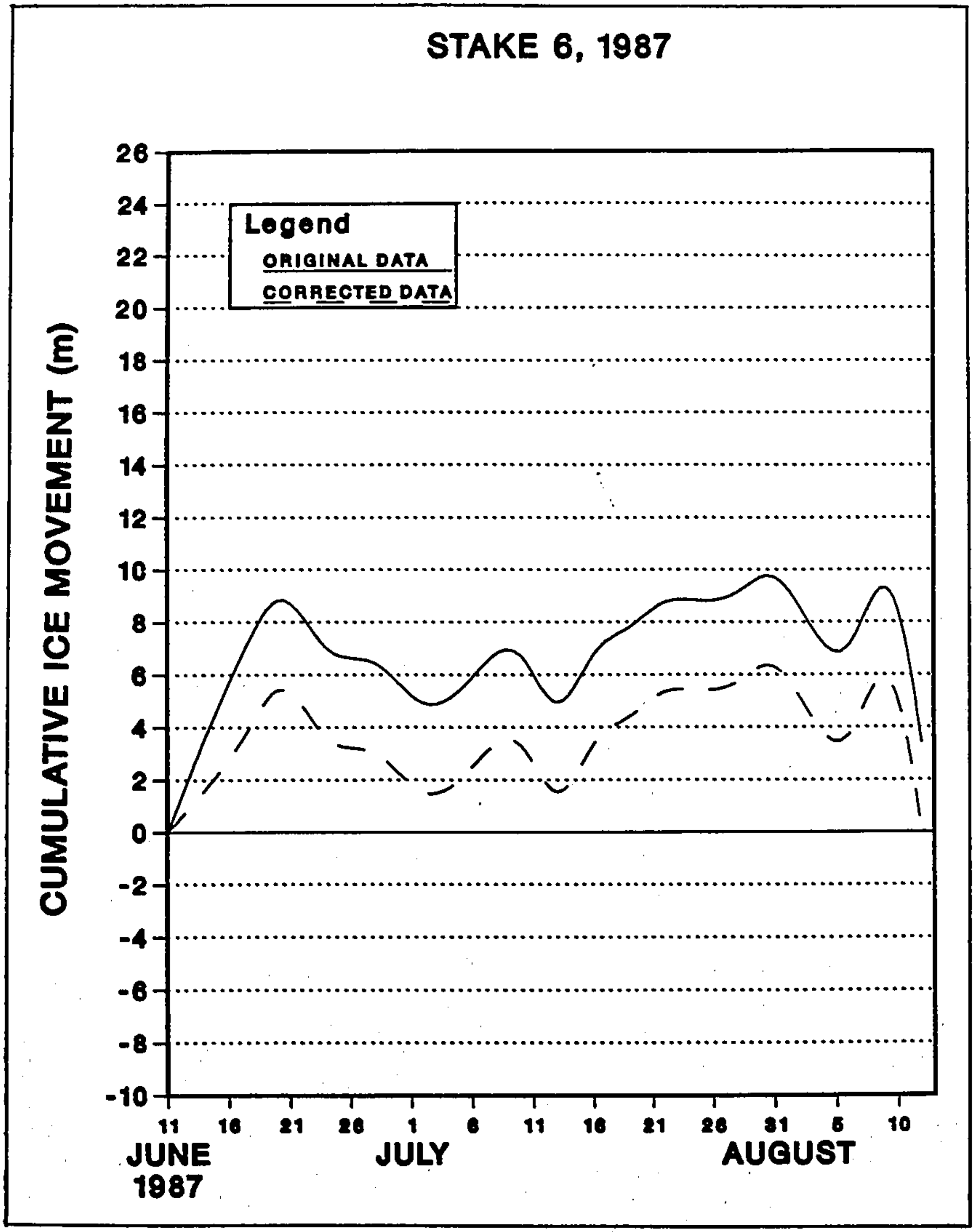
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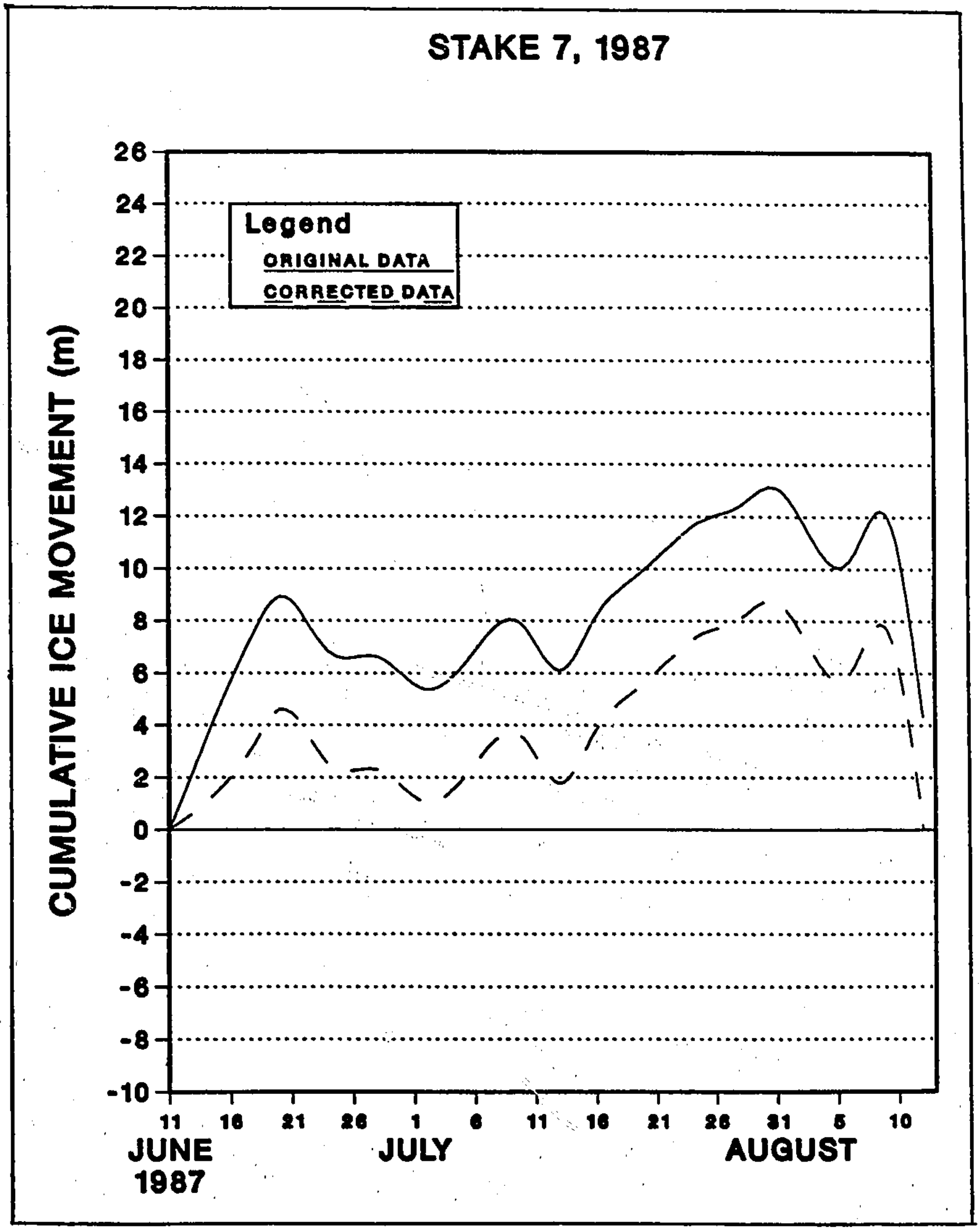
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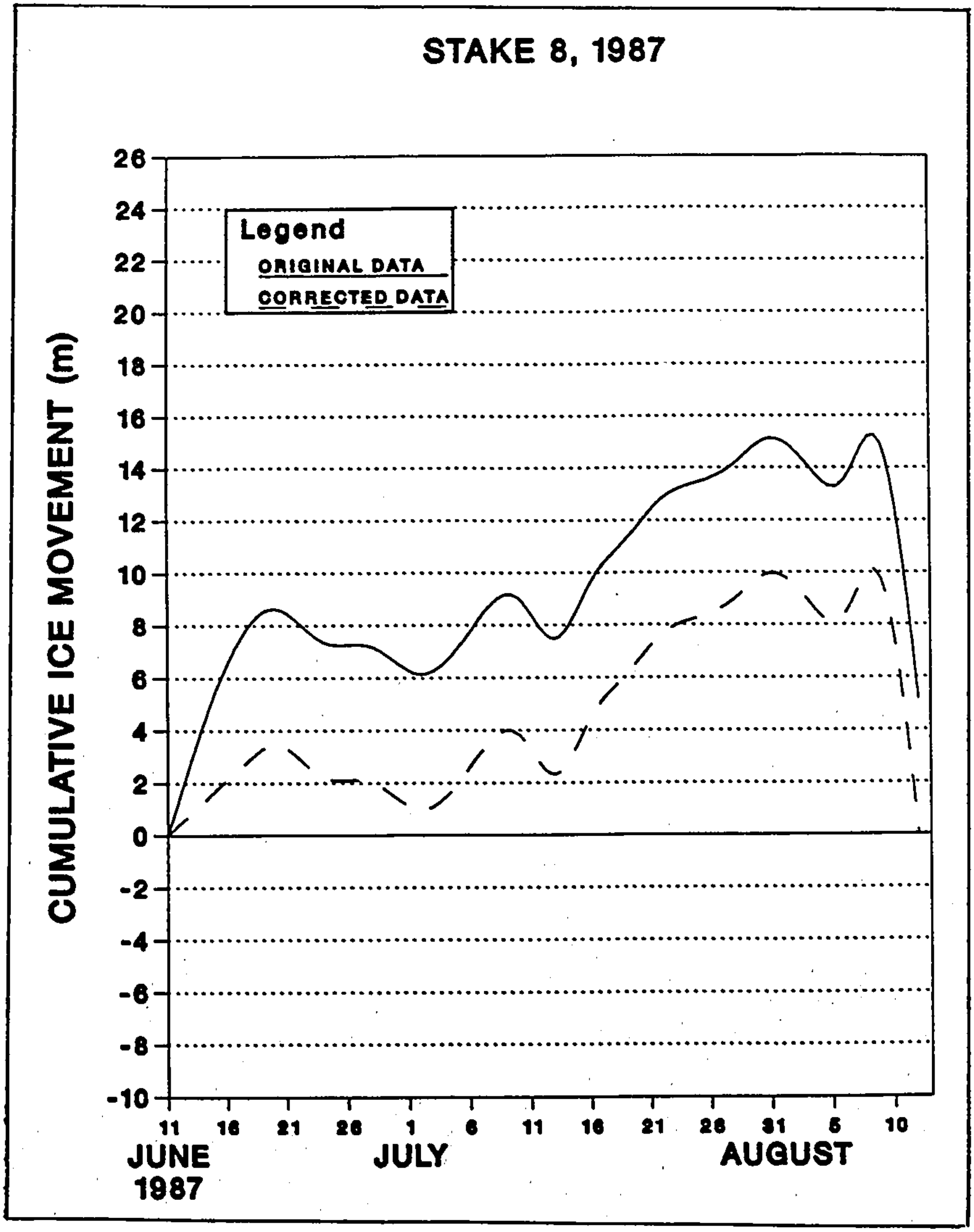
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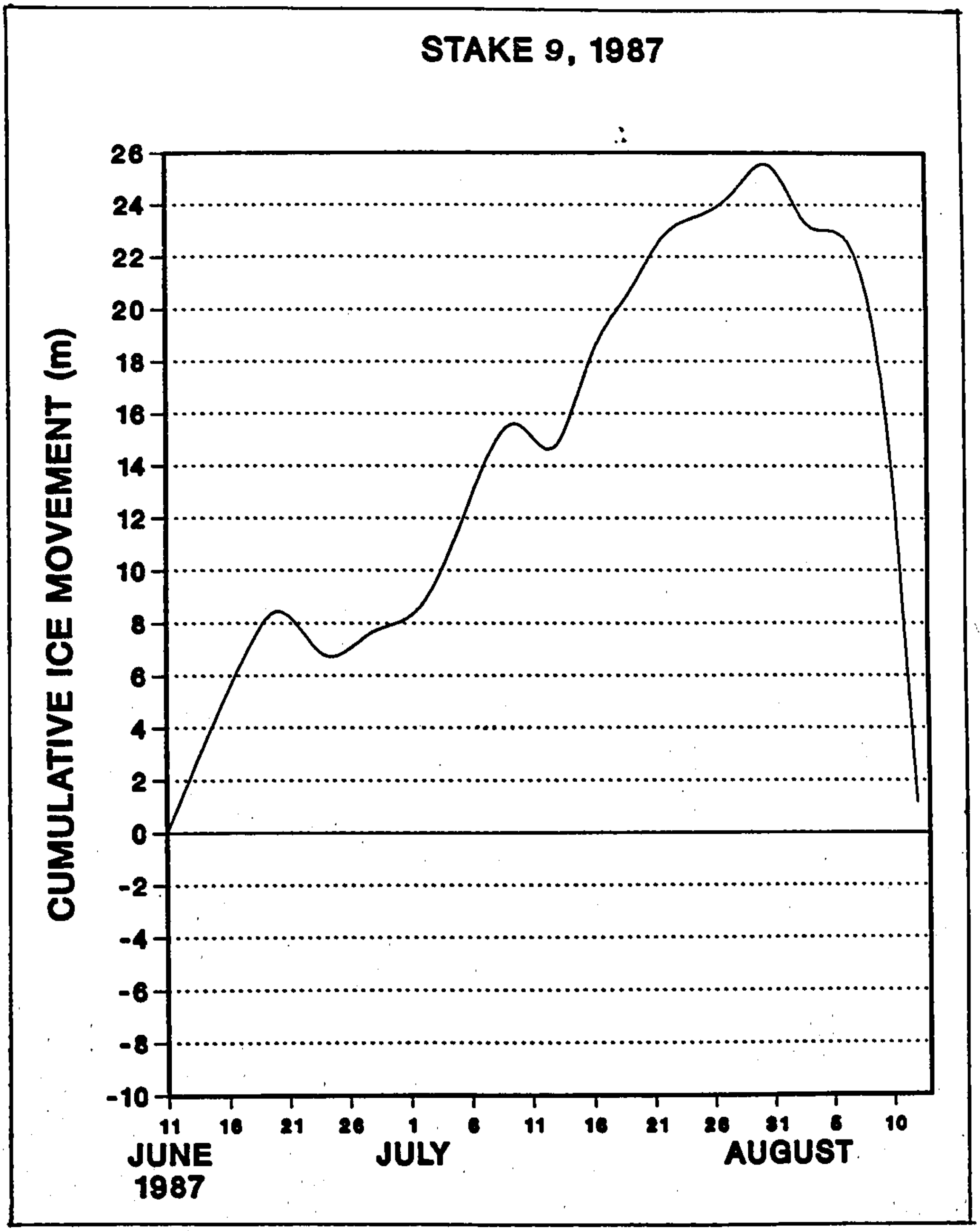
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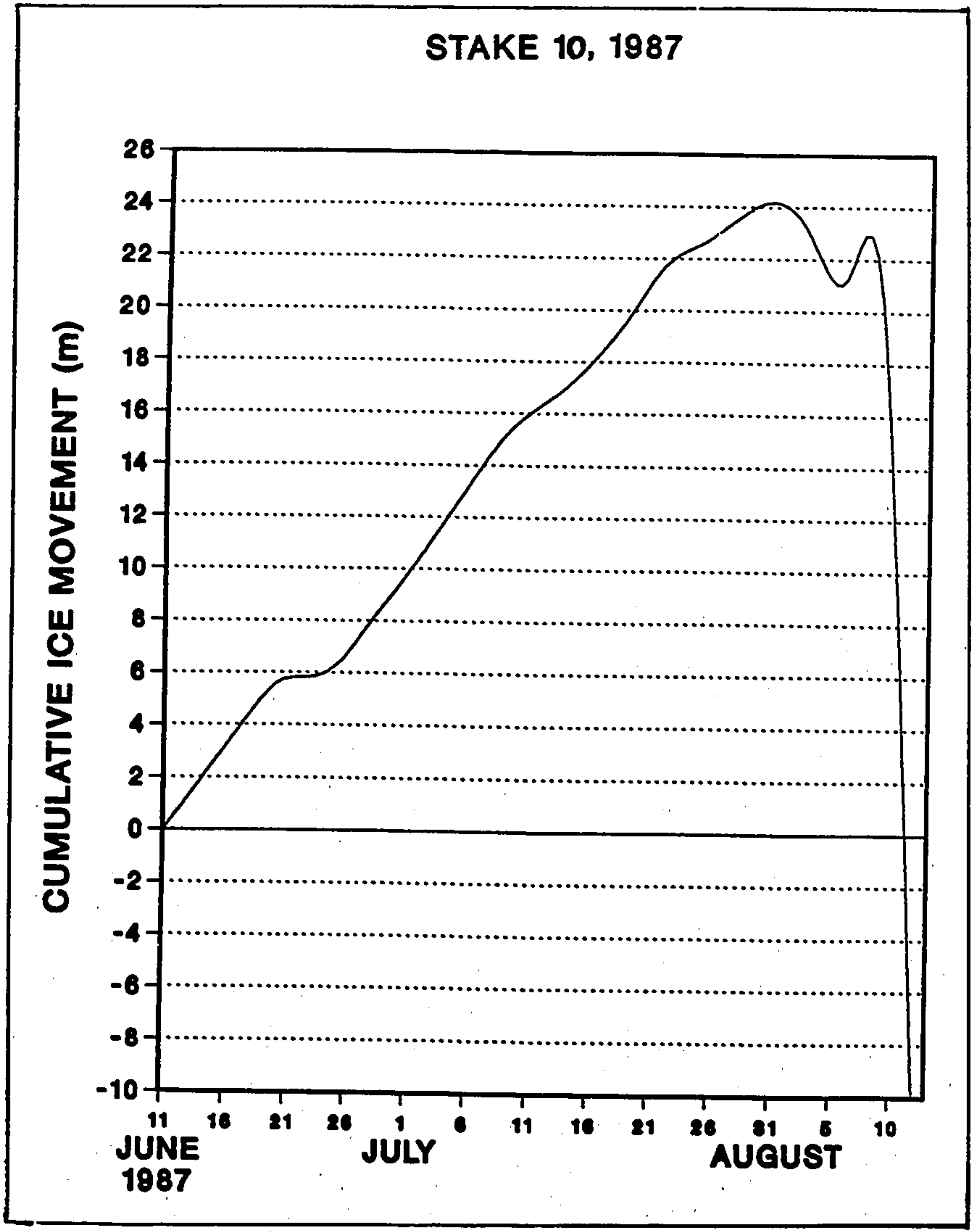
AUGUST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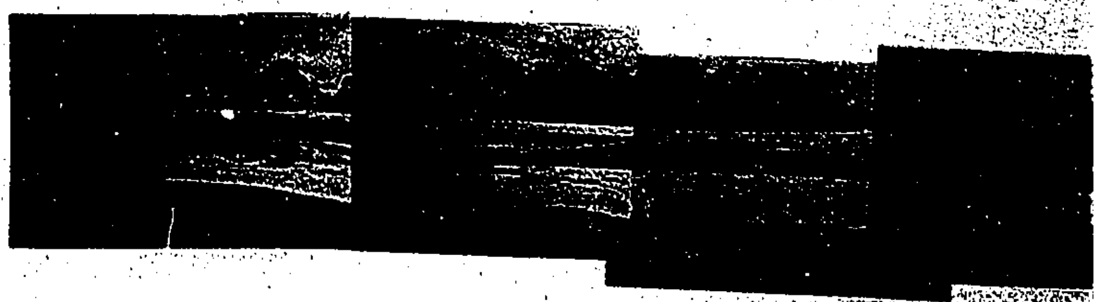
JUNE
1987
JULY
AUGUST

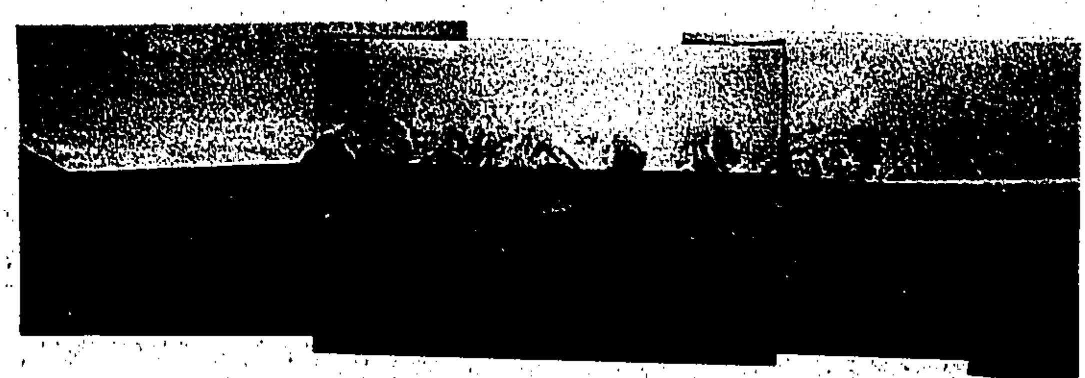
APPENDIX D

PHOTOGRAPHIC PANORAMAS OF THE ICE MARGIN, 1987.

June 6, 1987



June 19, 1987

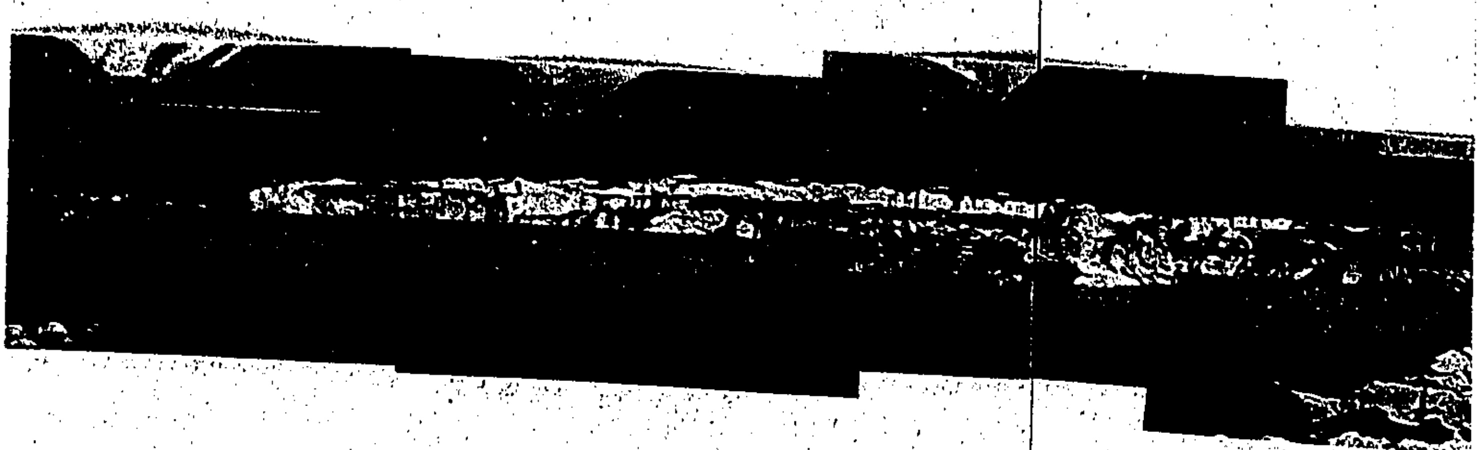


July 8, 1987

July 15, 1987

July 23, 1987

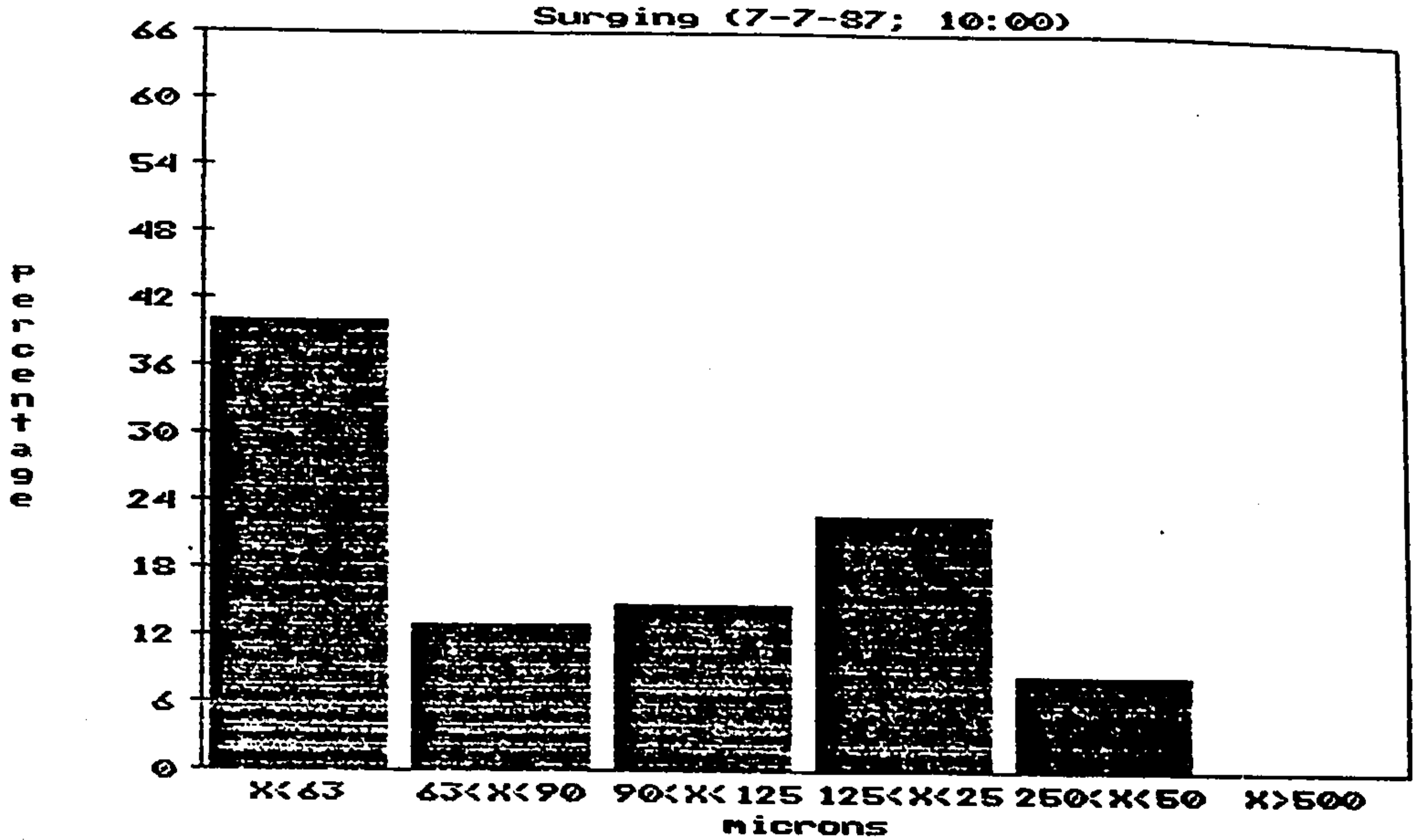
August 5, 1987

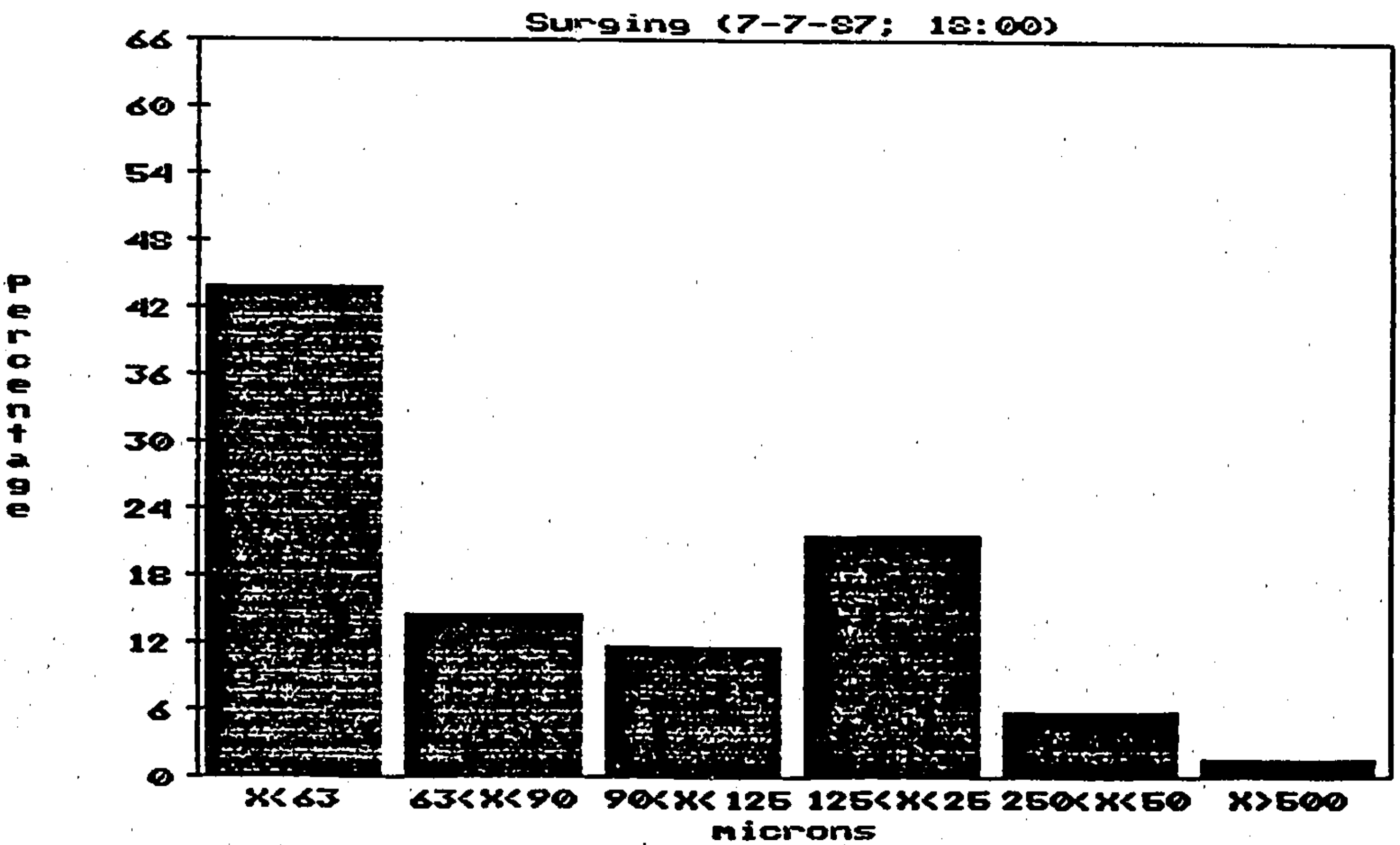


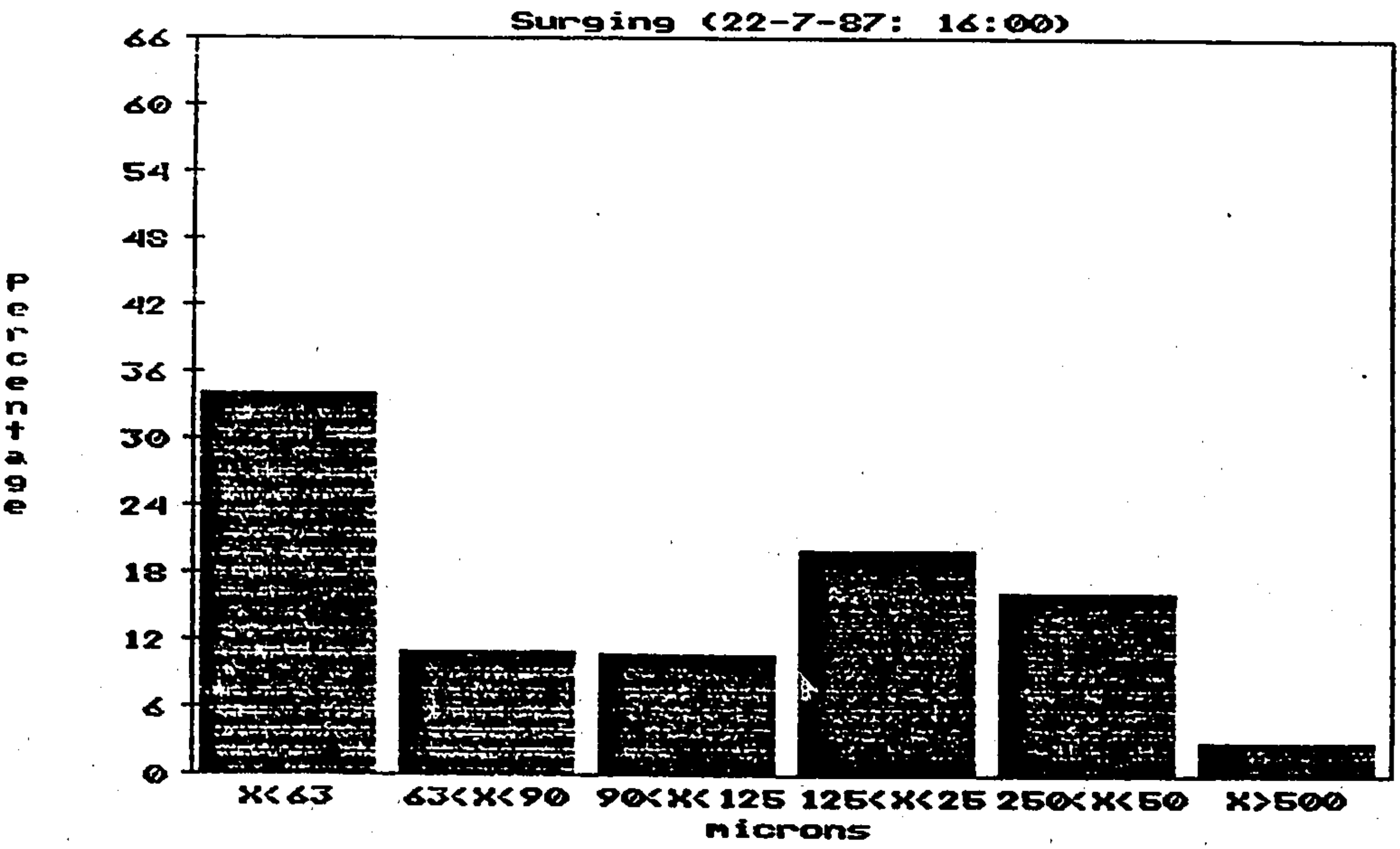
August 9, 1987

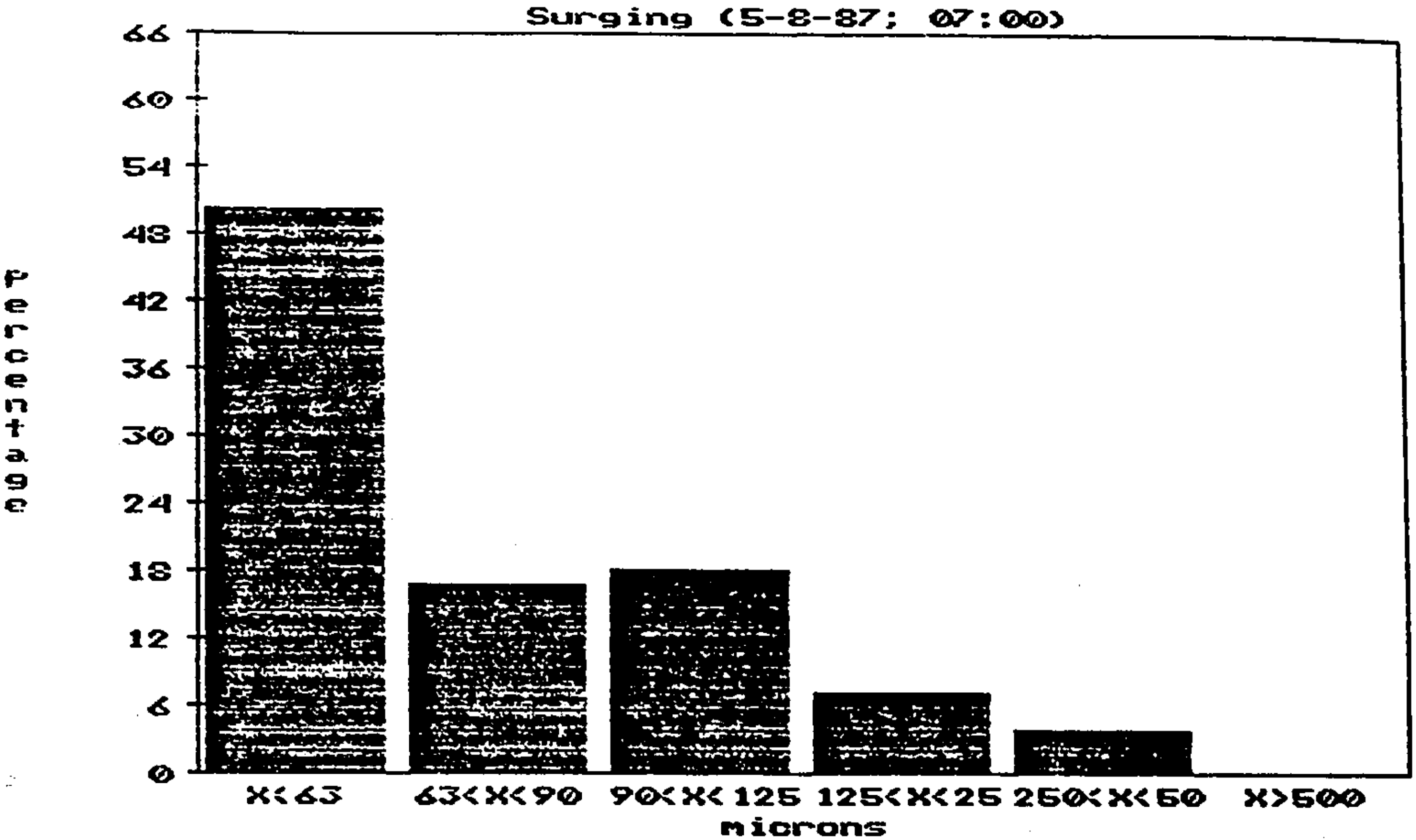
APPENDIX E

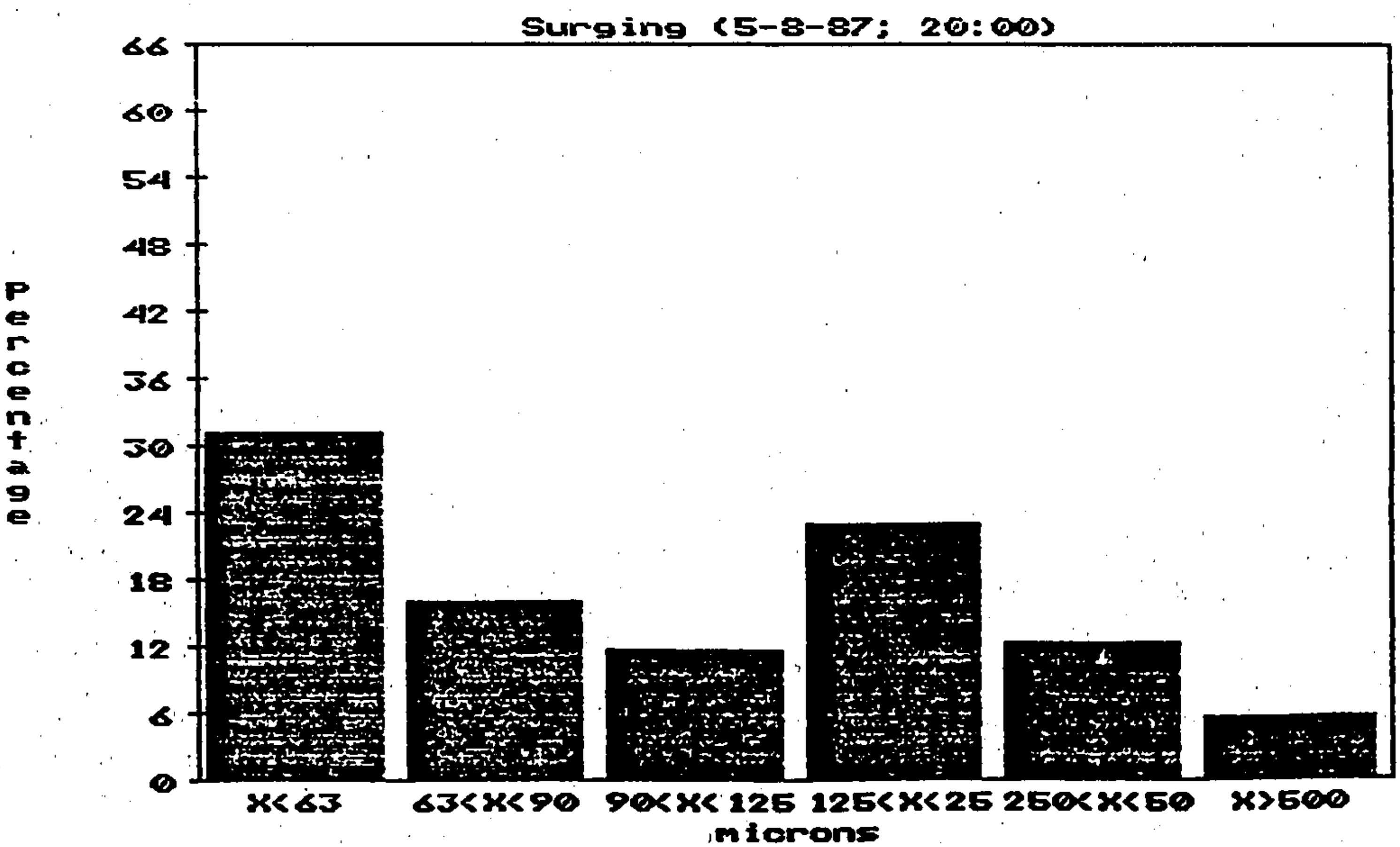
RESULTS OF SELECTED SEDIMENT CONTENT ANALYSES - 3 STREAMS, 1987


Surging (7-7-87; 10:00)
percentage
66
60
54
48
42
36
30
24
18
12
6
0
X<63
63<X<90
90<X<125
125<X<25
250<X<50
X>500
microns


Surging (7-7-87; 18:00)
percentage
66
60
54
48
42
36
30
24
18
12
6
0
X<63
63<X<90
90<X<125
125<X<25
250<X<50
X>500
microns


Surging (22-7-87: 16:00)
Percentage
66
60
54
48
42
36
30
24
18
12
6
0
X<63
63<X<90
90<X<125
125<X<25
250<X<50
X>500
microns


Surging (5-8-87; 07:00)
Percentage
66
60
54
48
42
36
30
24
18
12
6
0
X<63
63<X<90
90<X<125
125<X<25
250<X<50
X>500
microns


Surging (5-8-87; 20:00)
Percentage
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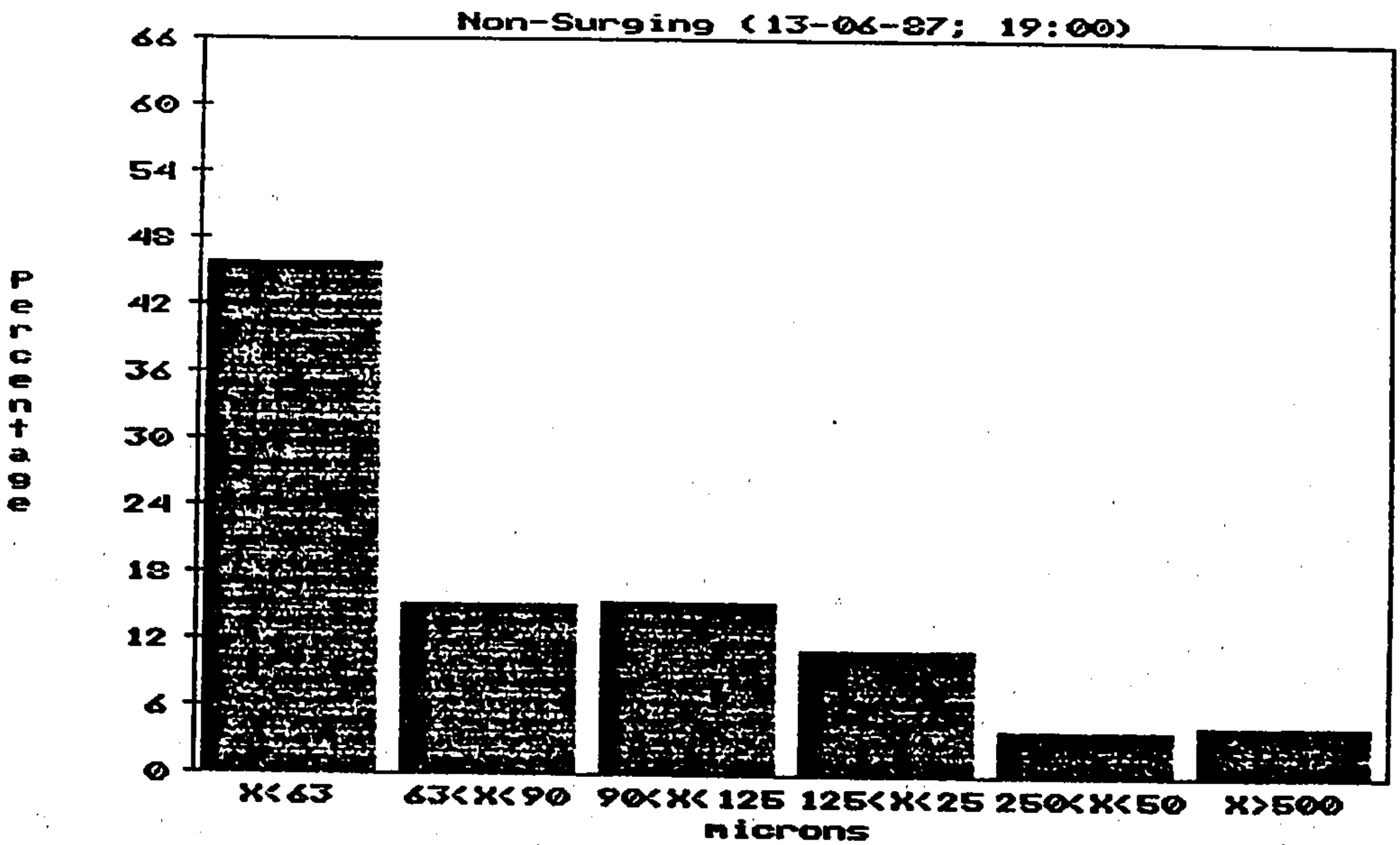
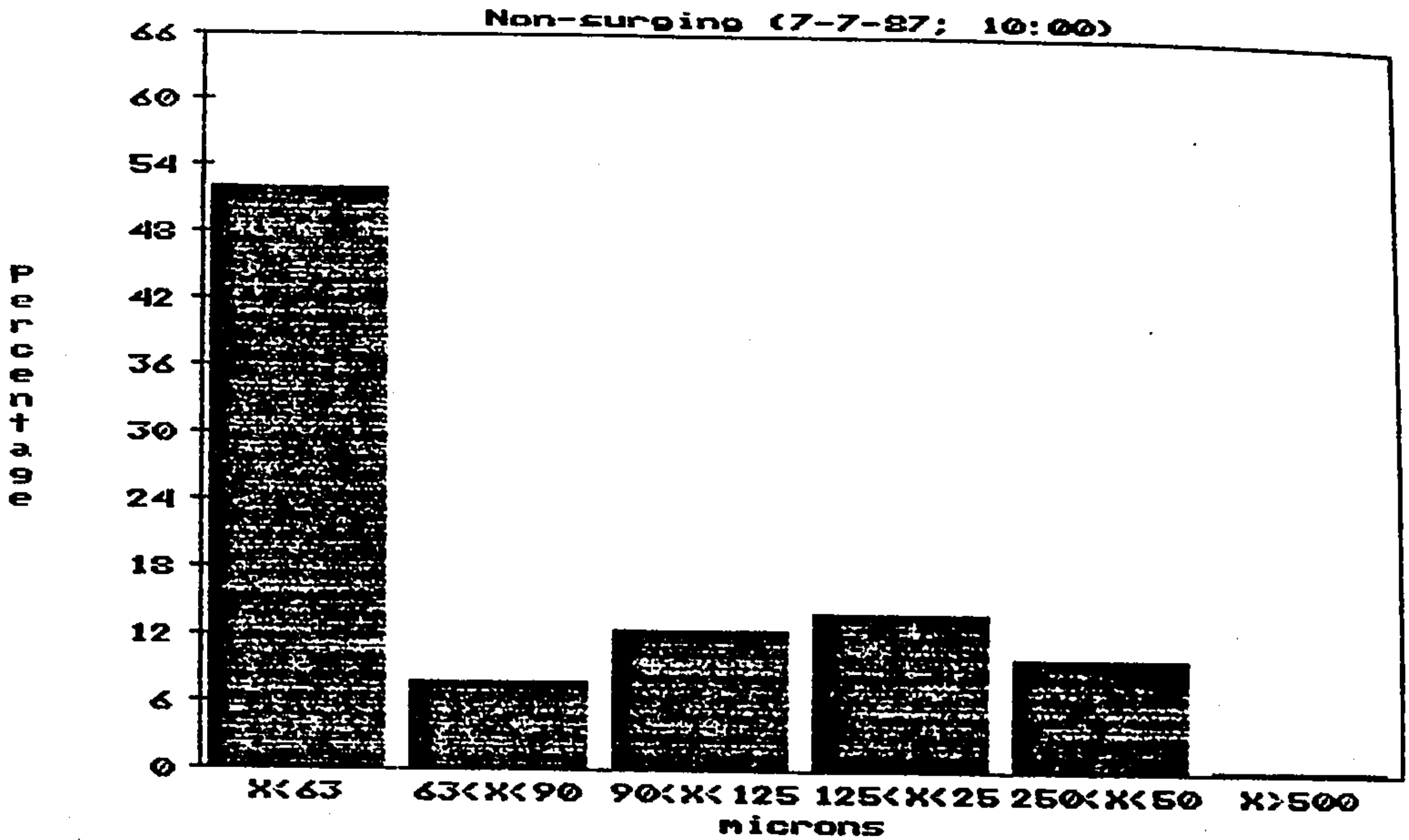
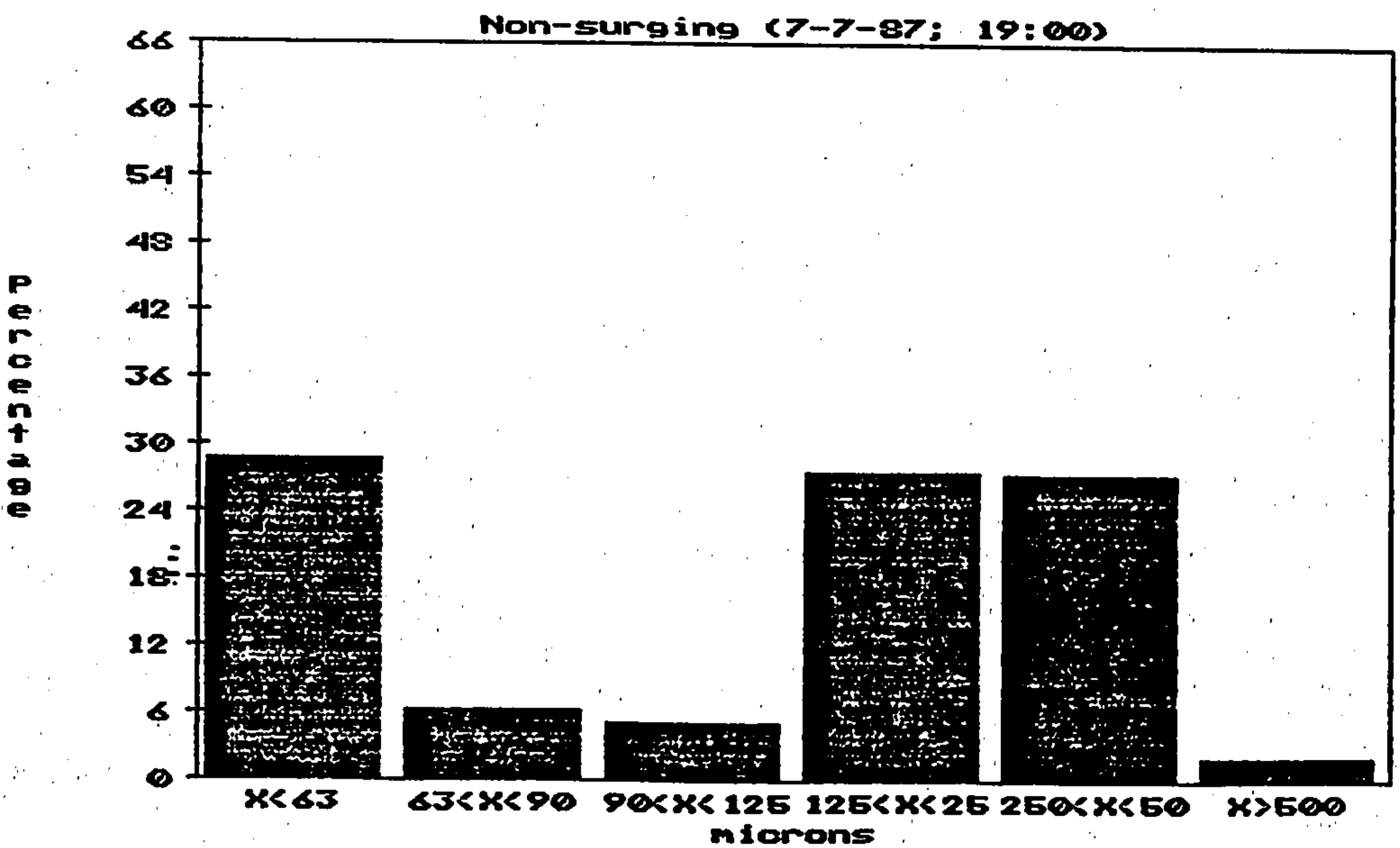
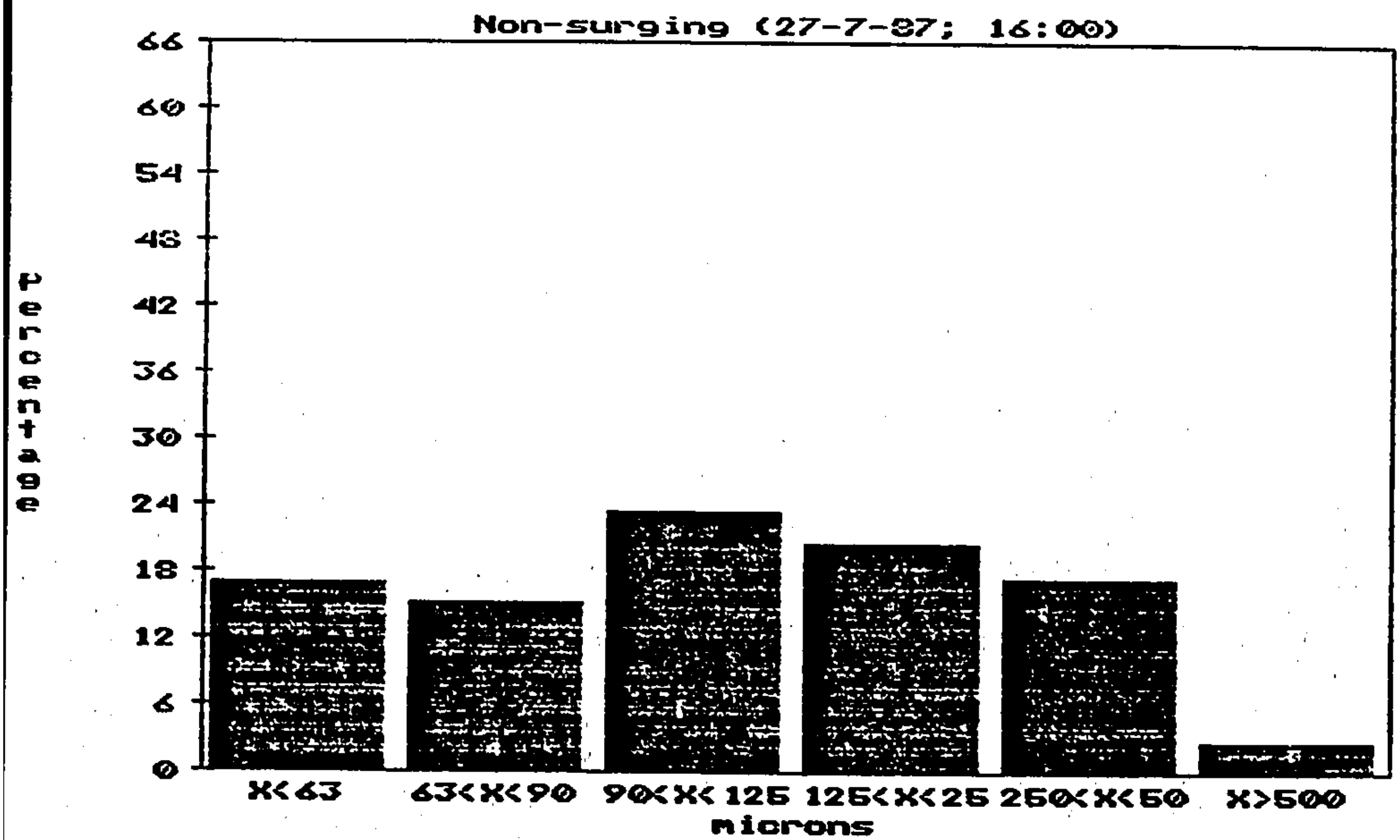
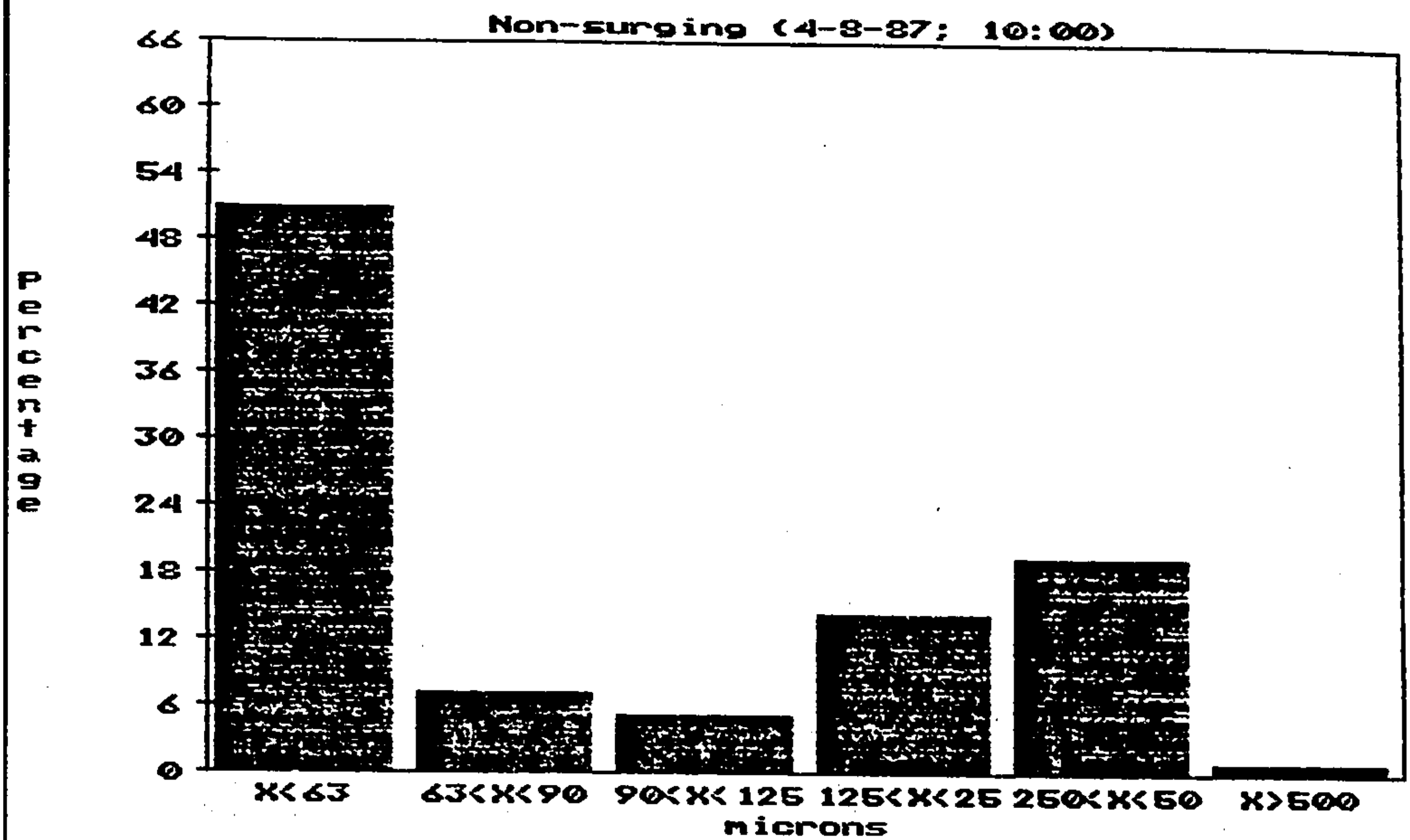
60
54
48
42
36
30
24
18
12
6
0
X<63
63<X<90
90<X<125
125<X<25
250<X<50
X>500
microns

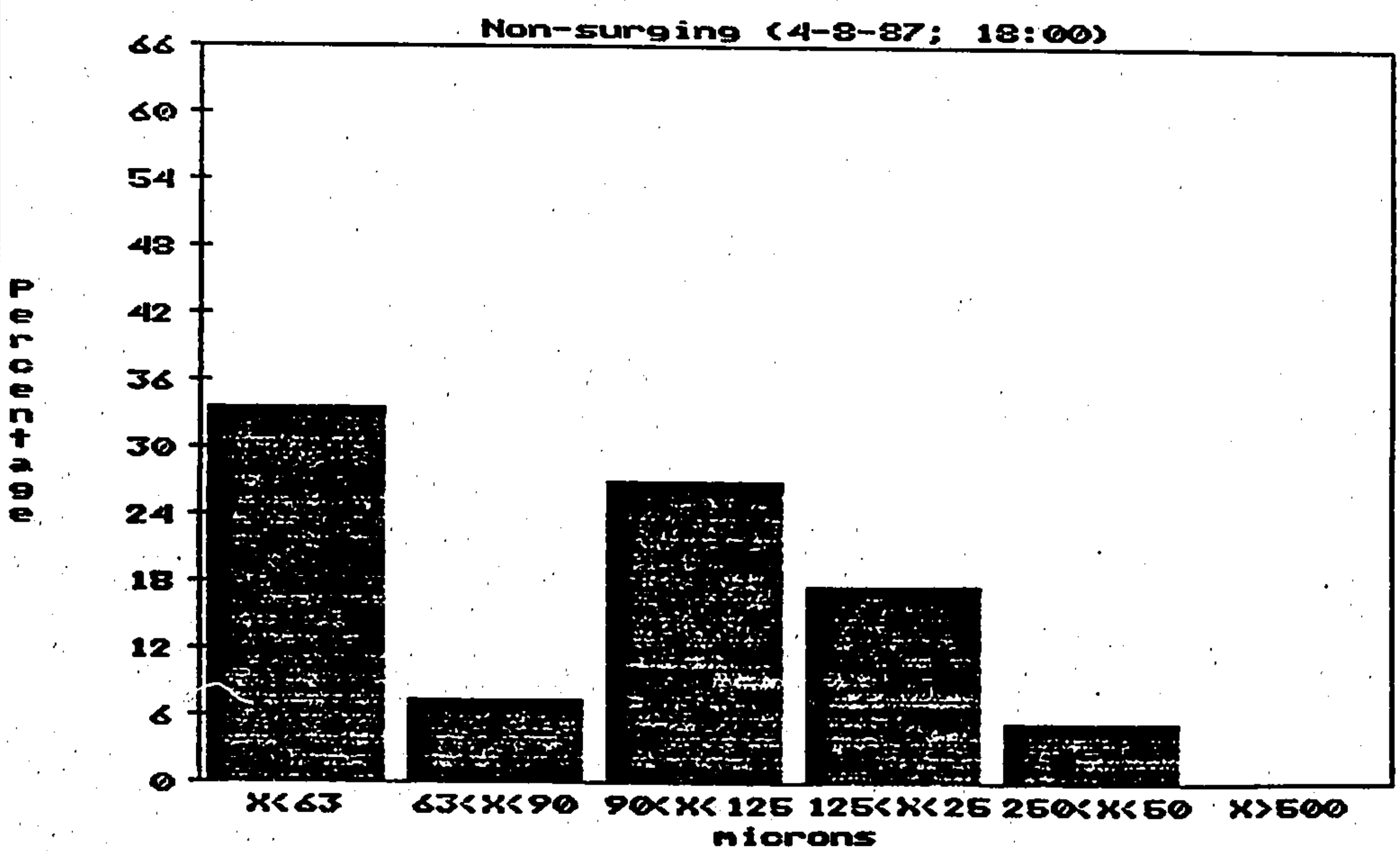

Non-Surging (13-06-87; 19:00)
Percentage
66
60
54
48
42
36
30
24
18
12
6
0
X<63
63<X<90
90<X<125
125<X<25
250<X<50
X>500
microns

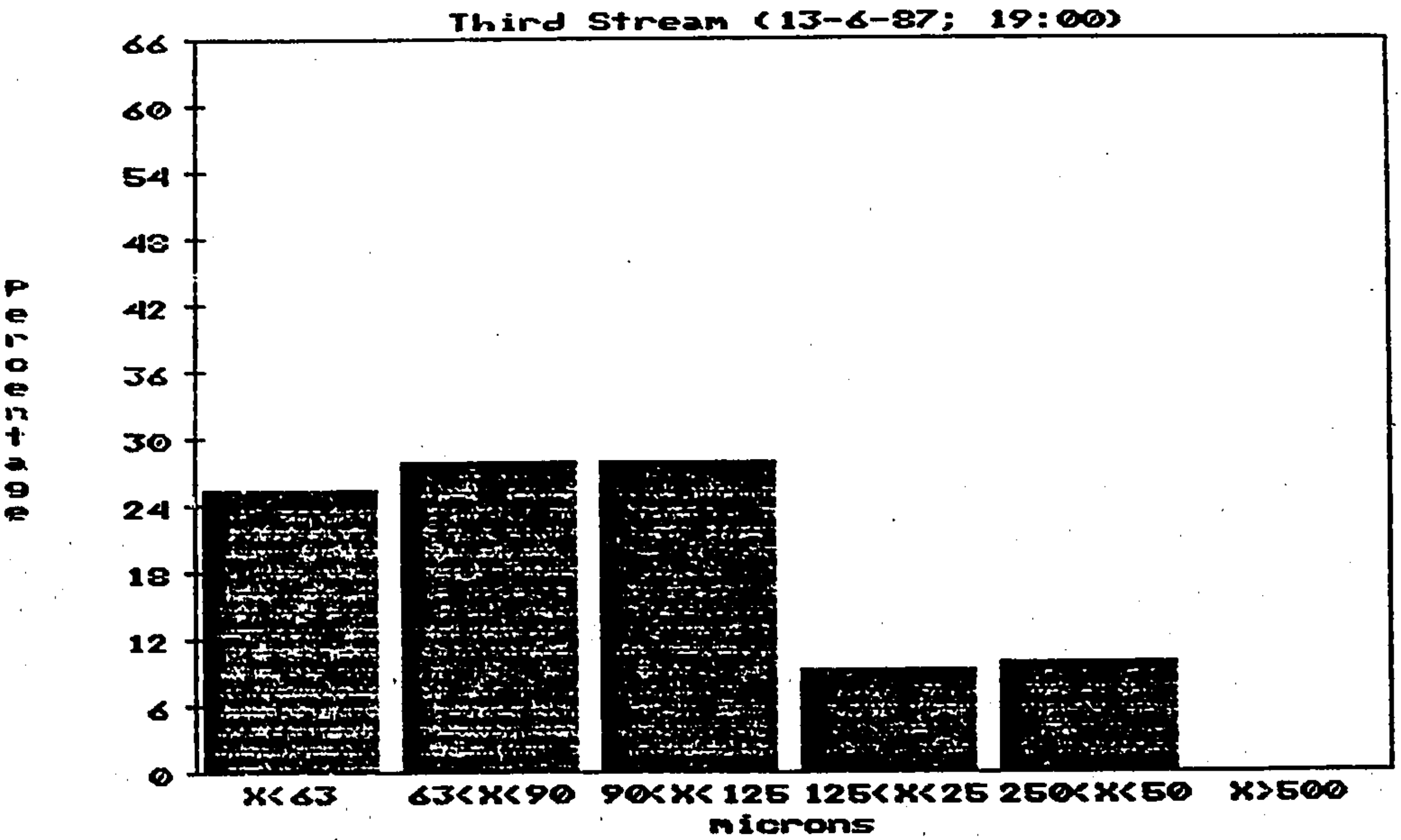

Non-surging (7-7-87; 10:00)
Percentage
66
60
54
48
42
36
30
24
18
12
6
0
X<63
63<X<90
90<X<125
125<X<25
250<X<50
X>500
microns

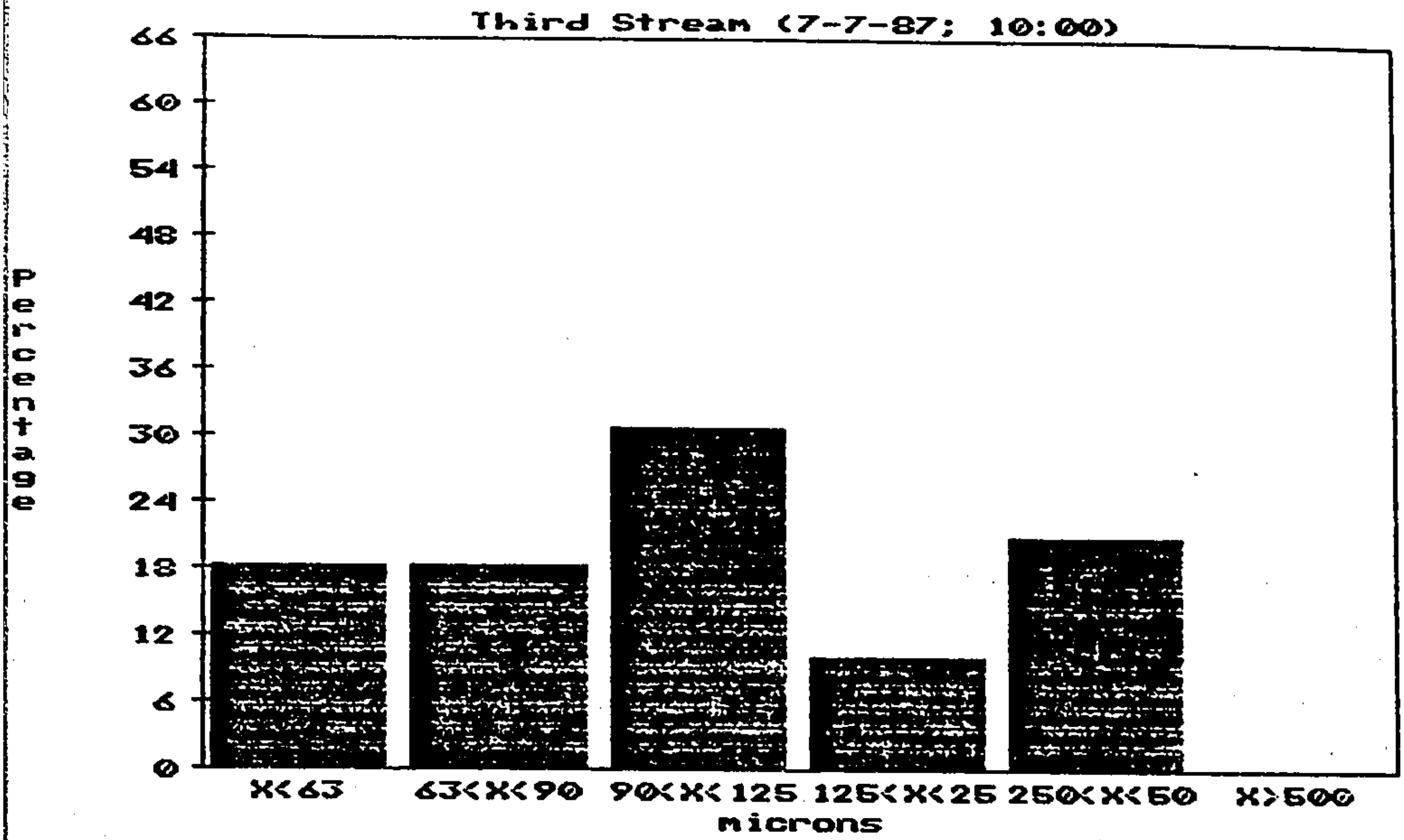

Non-surging (7-7-87; 19:00)
Percentage
66
60
54
48
42
36
30
24
18
12
6
0
X<63
63<X<90
90<X<125
125<X<25
250<X<50
X>500
microns

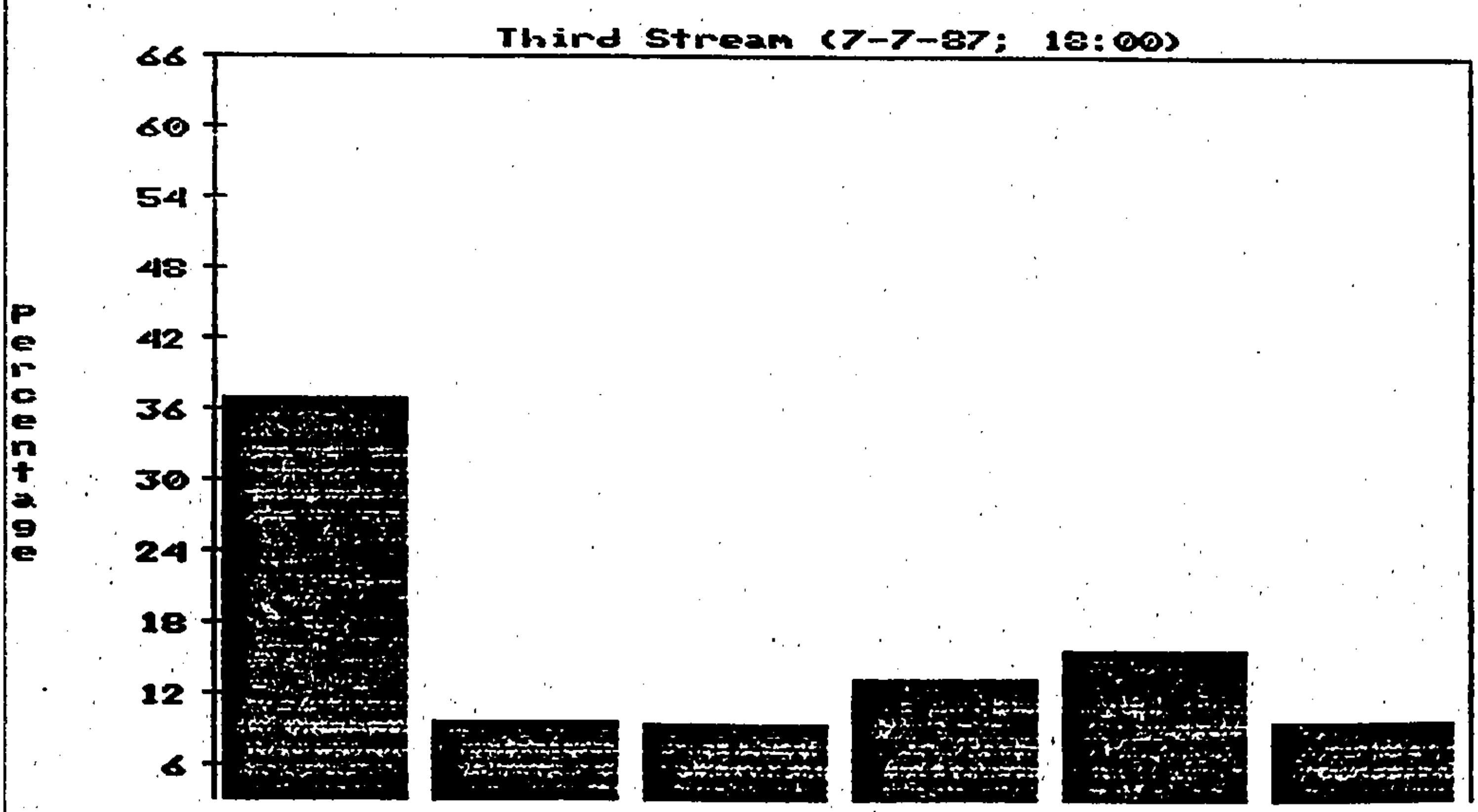

Non-surging (27-7-87; 16:00)
percentage
66
60
54
48
42
36
30
24
18
12
6
0
X<63
63<X<90
90<X<125
125<X<25
250<X<50
X>500
microns

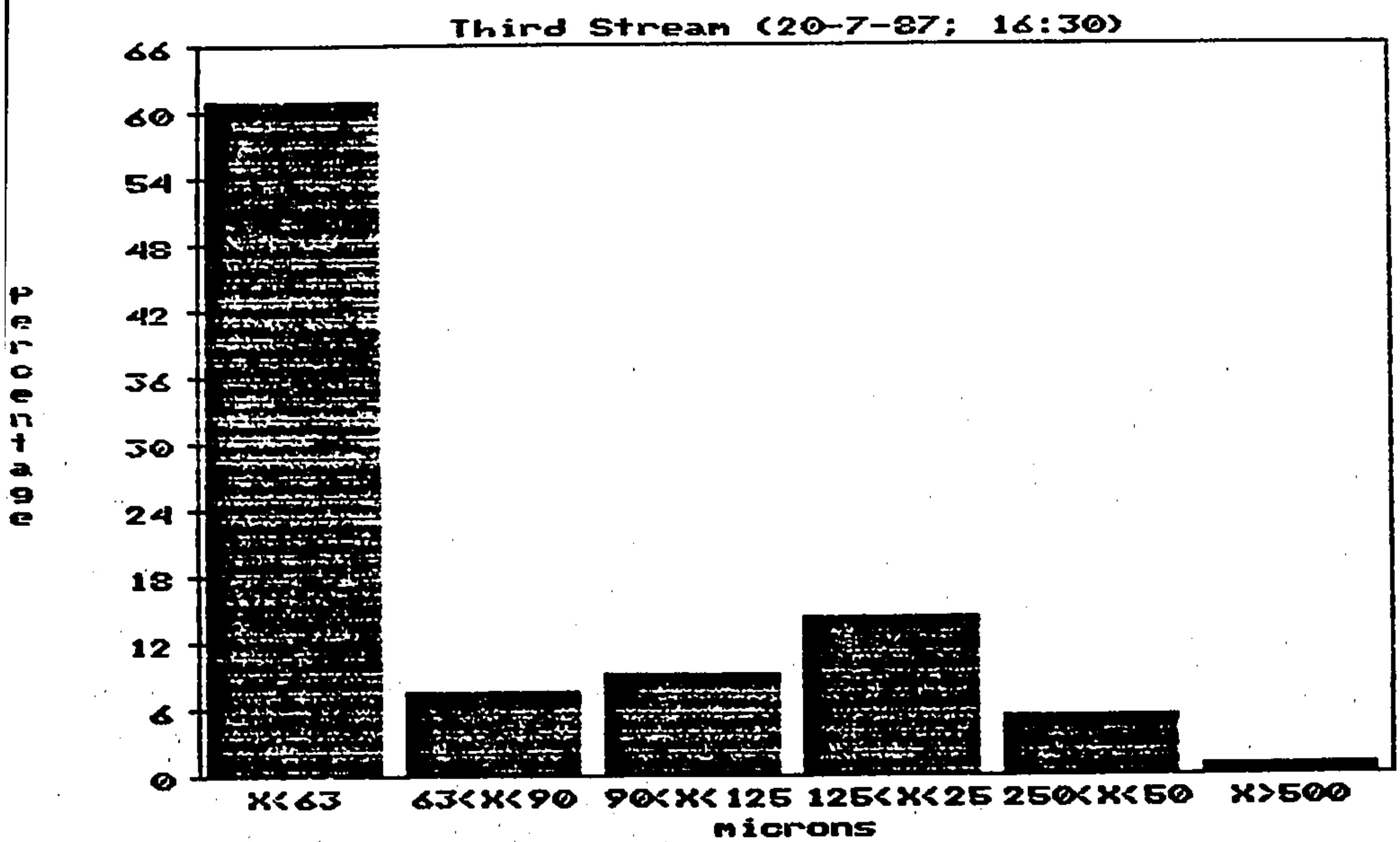

Non-surging (4-8-87; 10:00)
Percentage
66
60
54
48
42
36
30
24
18
12
6
0
X<63
63<X<90
90<X<125
125<X<25
250<X<50
X>500
microns

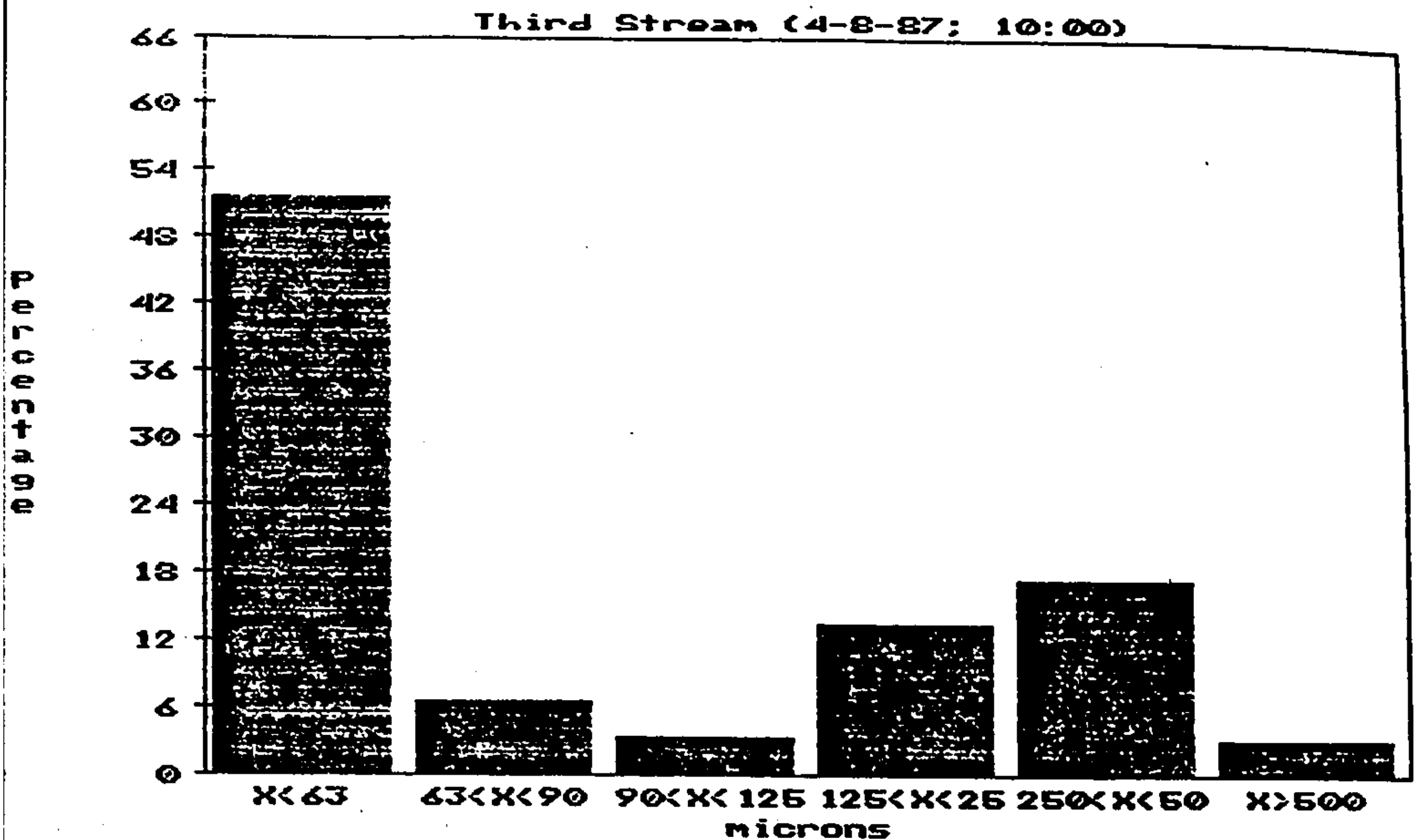

Non-surging (4-8-87; 18:00)
Percentage
66
60
54
48
42
36
30
24
18
12
6
0
X<63
63<X<90
90<X<125
125<X<25
250<X<50
X>500
microns

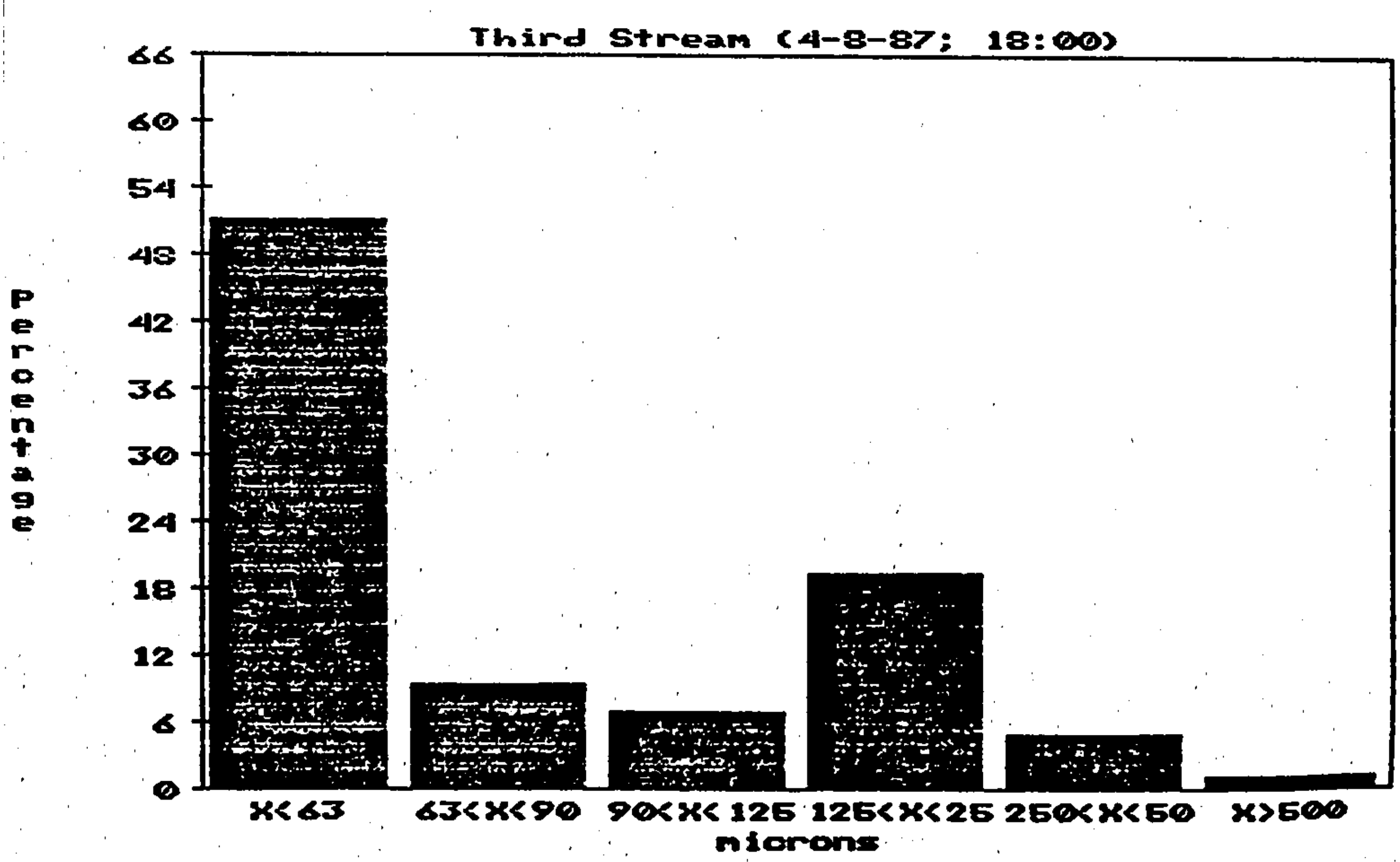

Third Stream (13-6-87; 19:00)
66
60
54
48
42
36
30
24
18
12
6
0
Percentage
X<63
63<X<90
90<X<125
125<X<25
250<X<50
X>500
microns


Third Stream (7-7-87; 10:00)
Percentage
66
60
54
48
42
36
30
24
18
12
6
0
X<63
63<X<90
90<X<125
125<X<25
250<X<50
X>500
microns


Third Stream (7-7-87; 18:00)
Percentage
66
60
54
48
42
36
30
24
18
12
6


Third Stream (20-7-87; 16:30)
percentage
66
60
54
48
42
36
30
24
18
12
6
0
X<63
63<X<90
90<X<125
125<X<25
250<X<50
X>500
microns


Third Stream (4-8-87; 10:00)
Percentage
66
60
54
48
42
36
30
24
18
12
6
0
X<63
63<X<90
90<X<125
125<X<25
250<X<50
X>500
microns


Third Stream (4-8-87; 18:00)
Percentage
66
60
54
48
42
36
30
24
18
12
6
0
X<63
63<X<90
90<X<125
125<X<25
250<X<50
X>500
microns