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ABSTRACT

CONTRIBUTIONS TO THE GEOMORPHOLOGY AND NEOGLACIAL CHRONOLOGY OF THE CATHEDRAL GLACIER SYSTEM ATLIN WILDERNESS PARK, BRITISH COLUMBIA

by

Vernon K. Jones

The Cathedral massif and its glacier systems lie at $59^{\circ} 20'$ N. latitude, $134^{\circ} 5'$ W. longitude, on the continental flank of the Juneau Icefield, near the south end of Atlin Lake in northwest British Columbia. Research was first initiated at this location in 1972.

The aim of this study was to determine the glacio-climatological history of the Cathedral Glacier system during late Neoglacial (Little Ice Age) time. To accomplish this an interdisciplinary approach was used, involving meteorology, climatology, glaciology, geomorphology, and hydrology.

Daily meteorological observations at the Cathedral Glacier station (Camp 29) over the summer field seasons of 1972 - 1974 were compared with similar records from Camp 30 at Atlin, B.C., for the same period. Annual government climatic records from 1907 to the present at Atlin and from Whitehorse and Carcross, Y.T., were combined to form a data base for the continental interior. This was compared with the Alaskan coastal records for the same period developed from records at Juneau and the Juneau Airport on the maritime flank of the Juneau Icefield. The data were subjected to linear and polynomial regression analyses and five-year weighted running means to differentiate longer-term cyclic patterns from annual variations.

The chronological sequence of glacial and glaciofluvial landforms associated with Cathedral Glacier was determined by field study, mapping, and the interpretation of low oblique aerial photography. A time scale was developed with the aid of dendrochronometric and vegetation studies of the moraine and glacial erosion zones. This information was then interpreted with respect to existing climatic records.

The study revealed evidence of three glacial advances beginning with a dominant late Neoglacial extension culminating in 1760 ± 30 years. This advance obliterated nearly all earlier Neoglacial evidence. Intervals between advances since that time appear to be

about 90 years. The last temperature upswing began about 1910. Moraine stabilization, indicating glacial retreat, appears to have lagged temperature rise by about 40 years. Annual mean temperatures peaked in the early 1940's and have trended downward since. If previous patterns are followed, regional temperatures should bottom out in the 1990's. Within the long-term patterns shorter, less regular oscillations occur. Patterns are less identifiable in the precipitation analysis.

Temperature patterns appear to influence glacial behavior through their effect on freezing level elevations and therefore on the zone of maximum accumulation. On a longer time scale englacial temperatures are also affected, which can influence glacial flow. Temperatures also play an important role in the hydrological regime of the glacier. Continued monitoring of the glacial mass balance, the seasonal neve' line, and hydrological discharge is providing further information for glacioclimatic analysis.

The Juneau Icefield is located on the crest of a major long wave ridge position in the circumpolar tropospheric circulation, which controls weather in the temperate zone. Shifts in the mean position of this ridge across the Icefield are amplified by orographic effects of the Alaska - Canada Boundary Ranges. Analysis and comparison of relative climatic patterns on the maritime and continental flanks of the Icefield yield important clues as to the nature and intensity of these tropospheric shifts, and hence can be of significance in understanding variations of atmospheric patterns over time.

Climate is a major factor in the energy and environmental problems facing man, and is a primary control on the provision of global food supplies. The pressing need for better understanding of patterns of variation in climate is intensified by ominous climatic changes in recent years. Regional glacioclimatic investigations such as the program of which this study is a part can add significantly to the understanding of the complex character of global climatic variations.

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OF THE CATHEDRAL GLACIER SYSTEM
ATLIN WILDERNESS PARK, BRITISH COLUMBIA

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Vernon K. Jones

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A scientific study such as this is the product of far more than one person's efforts. Isaac Newton's comment about standing on the shoulders of those who came before is eminently applicable here.

The Juneau Icefield Research Program is primarily the lengthened shadow of one man, Dr. Maynard M. Miller. To Dr. Miller, Mrs. Joan Miller, and to the many staff persons, field aides, and supporters of the Program, whose efforts span nearly thirty years, goes appreciation and a very sincerely felt thanks for their labors. The author has a very personal understanding of the often very difficult conditions encountered in becoming involved with Nature on her own terms; he is also keenly aware of some of the efforts involved throughout the entire year to make possible the few short weeks of summer field work. This report of research reflects the work not only of the author but of many field companions, associates, and other members of the far-flung FGER family.

Dr. Harry McDade kindly provided his personal plane for the initial aerial reconnaissance of the Cathedral area by Dr. Miller in 1971. Initial ground exploration, as well as initial logistics and camp development, was done in the early summer of 1972 by William Lokey and John Schutt. Scientific ideas and glaciological assistance were provided in 1972 and 1973 by Dr. Gordon Warner and the late Dr. Edward M. Little. The loss of Ed Little's gentle personality and of his keen scientific and

philosophical insight was a blow not only to the program but to each of us personally.

Dave Garland, Robert Asher, Howard Langeveld, Frank Stallwood, Lance Miller, Ross Miller, and Richard Heffernan were involved in logistics and in obtaining meteorological records and conducting glaciological surveys in the 1972 and 1973 field seasons. F. Nishio, C. W. Kreidler, and Jacques Guigne' carried out firn-pack and glacio-hydrological surveys in 1973 and 1974. Lee and Kathy Schoen maintained meteorological records at Camp 30 far beyond the normal field season, including the entire winter of 1972-73.

Field companions at Camp 29 in the course of the 1973 field season included Gordon Warner, Vicki Pedone, Scott Burns, Doug Extine, Howie Langeveld, Nancy Norton, Carey Miller, Don Bodean, Dr. Robert Black, Monte Wilson, Don Cadwell, Pete Tomlinson, Dave Henry, Dr. Harry Thompson, Jay Ach, Jamie McNitt, Tom Muirhead, and Bill Martin.

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No discussion of research in the Atlin area is complete without mention of Ann Tallman. Her cheerfulness, generosity, helpfulness, and encouragement over the recent years have earned her a key place in the hearts of the FGER family.

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If any misinterpretations or unresolved erroneous theories remain in this report, the reader is requested to gently point them out. Future researchers are earnestly urged to improve upon the analyses contained herein and bring forward for discussion any new evidence as it becomes available.

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CHAPTER I

INTRODUCTION AND BACKGROUND

Introduction

This study is based upon the premise that a strong relationship exists between the geomorphology of a landscape and the kind and intensity of atmospheric processes which have acted upon the existing structure and material over time. This is especially true in regions of glaciation. The process-oriented analysis of such landscapes is therefore useful in determining climatic history.

The geomorphic character of a landscape in and near a glacial environment is affected by the following inter-related factors:

- 1) the atmospheric processes acting at the surface-atmosphere interface
- 2) the geological processes occurring on and within the surface; these are controlled by atmospheric processes
- 3) the physical nature, material, and structure of the surface, which influences the effects of the atmospheric and geologic processes (exogenetic processes)

The research discussed in this presentation considers the exogenetic processes and their effects on the terrain, especially as concerned with glacierization. The climatic conditions and their geomorphic effects in the research area are studied over the period of existing records. Geomorphic and geobotanic evidence is also examined to determine conditions prior to the existence of climatological records

in the area. The basic aim is to understand both the geomorphology and the past climates of the study area, and their interrelationship.

Purpose, Premises, and Procedures

Statement of Purpose

The primary purpose of this study is to determine the climatic chronology of the Cathedral Glacier system during Neoglacial time, emphasizing the placement and subsequent alteration of the youngest moraine sequences. As these moraines are late Neoglacial in age they are considered to typify the chronology of the Little Ice Age (Matthes, 1949).

Subsidiary aspects of this main purpose are:

- 1) to delineate morphogenetic events which have occurred in Cathedral Valley during Neoglacial time
- 2) to show the relationships of Holocene features to adjacent late Wisconsinan features
- 3) to determine the approximate dates of advance of the Cathedral Glacier and relate them to moraine deposition
- 4) to investigate the atmospheric mechanisms which produced this sequence of glacial landforms

Basic Premises

Interpretations in this study are based on the following obvious but still essential premises:

- 1) that there have been significant and identifiable variations in climate in this area in recent centuries
- 2) That these variations significantly affected the glacial and periglacial processes and therefore are the key to understanding development of the physical landform sequence

The Research Approach

The investigation attempts to be inductive in its approach. Meteorological observations are examined from the Cathedral Glacier Research Station (Camp 29) for the summer field seasons of 1972

through 1974, and compared with the Atlin station (Camp 30) for the same period. Longer-range Canadian and United States government records from nearby and regional weather stations at Atlin, Carcross, Whitehorse, Juneau, and Juneau Airport are analyzed to delineate relative patterns and trends over time. These records extend as far back as 1905, although unfortunately there is not a continuous record at any of the stations involved.

Geomorphic information and background for this study includes ground truth obtained by the writer in field research during the 1973 and 1974 summer field seasons on the Cathedral massif; it also includes one full summer (1971) of related studies on the main Juneau Icefield. Information is also obtained from aerial photography; from meteorological and glaciological data; from a determination of cirque floor elevations (Squyers, 1975); and from access to pertinent lichenometric data (See, 1975). The indispensable assistance and research efforts of other members of the Juneau Icefield Research Program team are readily acknowledged, as is the support of the Foundation for Glacier and Environmental Research, Pacific Science Center, Seattle, Washington.

The empirical and observational results of the field research are combined with a deductive application of some key principles of glacial geology, meteorology, and geomorphology.

Physical Setting of the Research Area

Geographical Position

The primary study area is the Cathedral Glacier system in the Cordilleran region of northwestern British Columbia. It is located at 59°20' N 134°05' W, near the south end of Atlin Lake and about 30

miles northeast of the Alaskan border (Figures 1, 2, 3). This glacier system is one of several small cirque glaciers on the Cathedral massif, a rugged bedrock highland about 1 mile by 3 1/2 miles (1 1/2 x 5 1/2 km) in size on the continental flank of the northern Boundary Range. This is the northernmost range in the Coast Mountains along the international border between west central and Northwestern British Columbia and the panhandle area of Southeastern Alaska.

Relationship to the Juneau Icefield

The Cathedral massif is adjacent to the northern interior periphery of the Juneau Icefield, the fifth largest area of continuous land ice in North America. The Icefield extends along the coastal cordillera from Taku Inlet south of Juneau nearly to White Pass, northeast of Skagway (Figure 2). The massif is separated from the main Juneau Icefield by steep, formerly glacierized valleys with up to nearly 5000 feet (1500 m) of vertical relief.

The Juneau Icefield is composed of approximately 1700 square miles of upland ice and glacierized valleys and plateaus, with barren nunataks protruding through the ice and snow. The highest peaks are Devil's Paw, 8,584 feet (2610 m), on the southeastern edge; and Mt. Nesselrode, 8400 feet (2545 m), in the center of the icefield.

On the maritime flank a number of glaciers such as the Herbert, Mendenhall, Taku, Gilkey, Meade, and the Twin Glaciers reach far down into deep glaciated trenches nearly to tidewater. On the interior side the termini of the Llewellyn, Sloko, Hoboe, and Willison Glaciers extend down to within 2400 to 2800 feet (730-850 m) above sea level, not far from the 2190-foot (670 m) elevation of Atlin Lake. The Taku-Llewellyn glacier system extends some 75 miles as a transection ice surface from Taku Inlet northward across the crestal region at about

Fig. 1 Map of Alaska

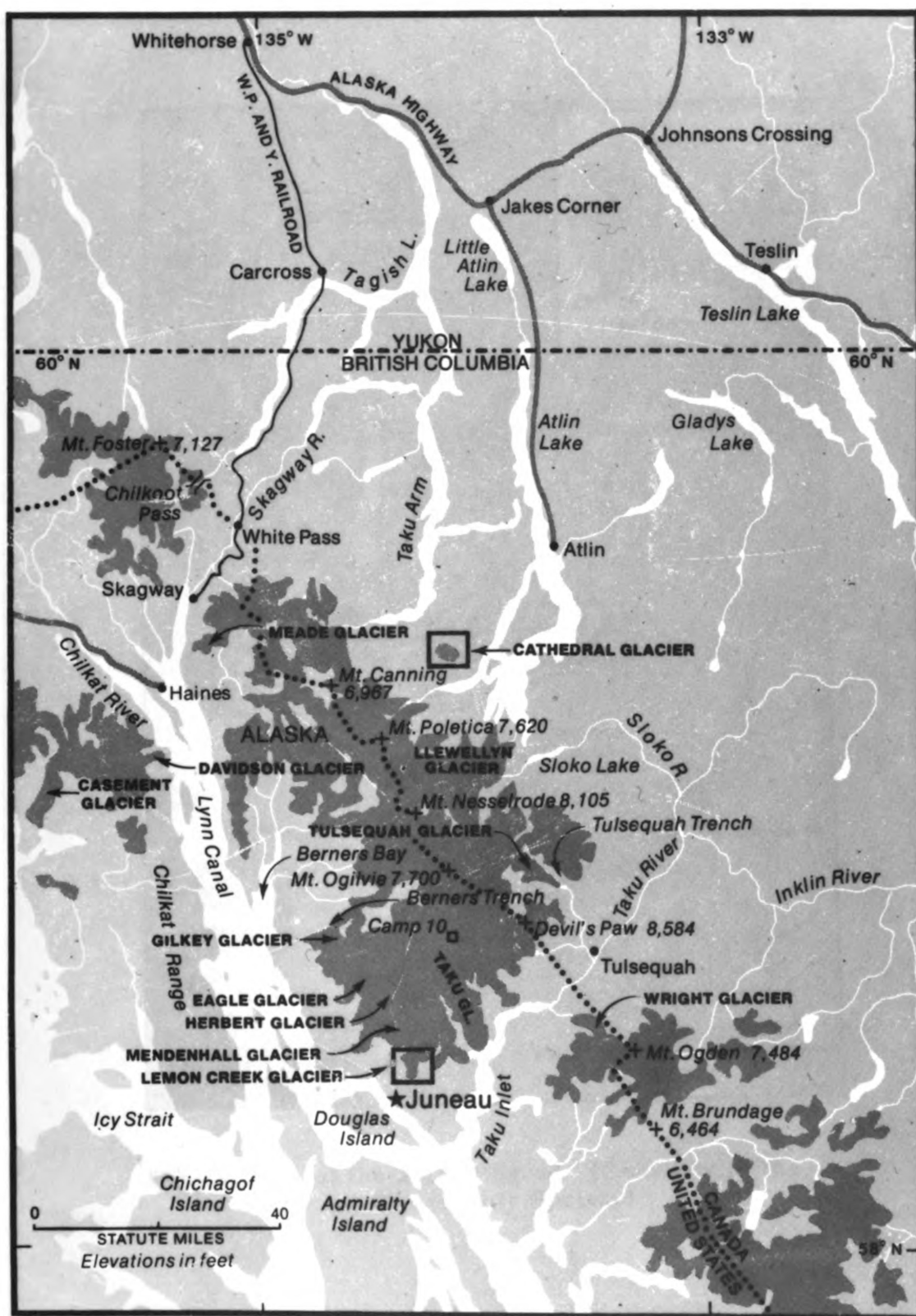


Fig. 2 Map of the Juneau Icefield (Courtesy of the National Geographic Society)

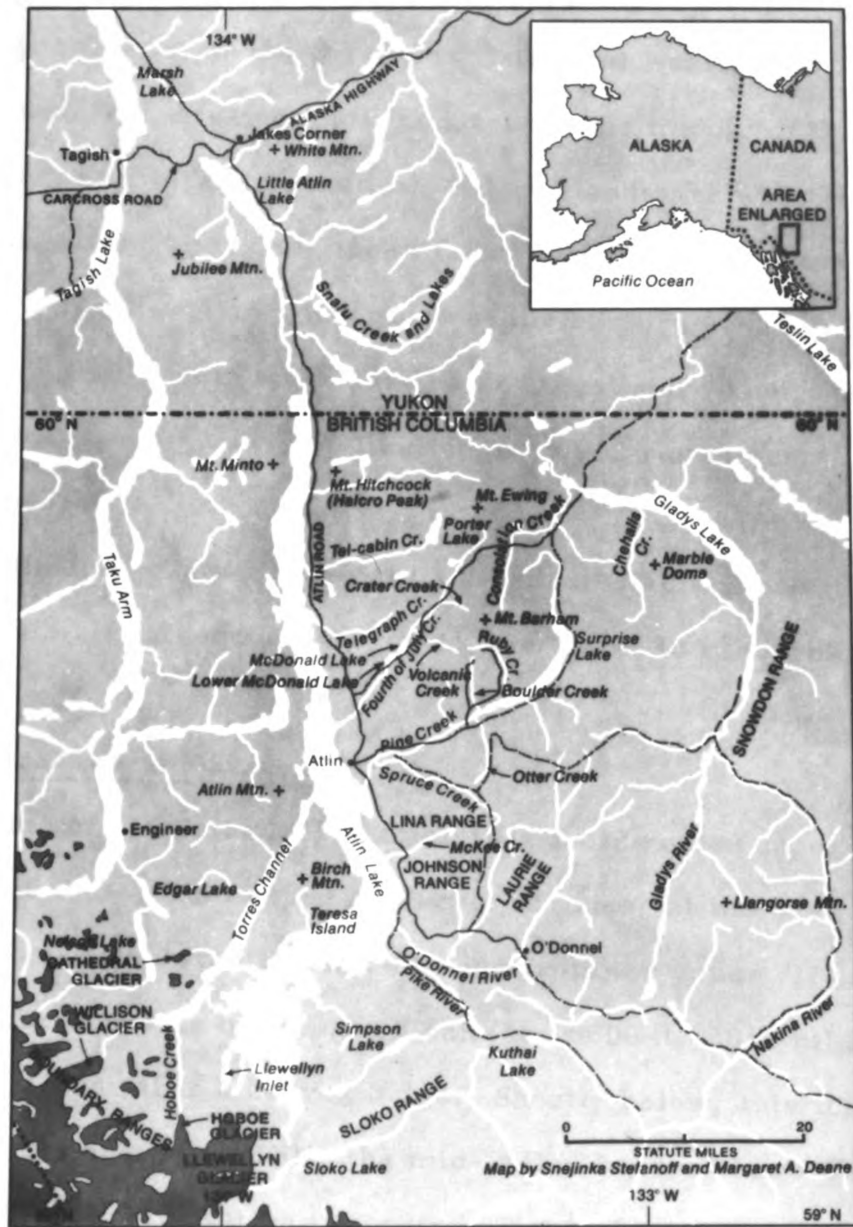


Fig. 3 Map of the Atlin Region (Courtesy of the National Geographic Society)

6200 feet (1890 m) to the Llewellyn terminus in the interior. The highest broad névé of the Llewellyn Glacier lies at about 7000 feet (2130 m) on the flanks of Mt. Nesselrode (8100 ft. 2470 m).

The Cathedral Massif

The highest point on the massif is Cathedral Peak, 6950 feet (2120 m) which is situated approximately in the middle of the massif. Distinct from this is The Cathedral, a local landmark which is made up of the northeast corner of the massif, between Torres Channel and Willison Bay (Figure 4). With a low spire at the eastern end, 3000-foot rock chutes ribbing its precipitous sides, a serrated ridge along its crest, and a western end sloping buttress-like, it is indeed reminiscent of a massive Gothic cathedral.

The Cathedral, Cathedral Peak, Cathedral Glacier, and the Coliseum are easily visible in good weather from the Camp 30 research station in Atlin.

The Cathedral Glacier System

Active ice of the Cathedral Glacier fills a bifurcated upper basin, consisting of non-tandem upper and lower cirques, at the base of Cathedral Peak (6950 ft, 2120 m). As ice from the upper cirque flows past the east side of Splinter Peak (ca. 6600 ft, 2010 m) the ice surface drops off in a convex bulge. Shortly below, this ice terminates just above Met Hill, the mid-valley bedrock knob where the Camp 29 research station is located (Figure 5). Part of the ice from this cirque formerly passed through a narrow depression, now moraine-mantled, to the west of Met Hill. The space between the terminus and this former outlet depression is presently occupied by a small proglacial lake. A small component of ice from the



Figure 4 The Cathedral Massif from Lake Atlin



Figure 5 Cathedral Glacier Valley from Splinter Peak

upper cirque joins ice from the lower cirque and flows to the east of Met Hill.

A rock cleaver of metabasalt separates the upper portions of the two cirques in the headwall area. The nearly level glacier surface of the lower or southern cirque is at approximately 5600 feet (1710 m) elevation. The lower cirque is partially alongside and below the flow area from the upper basin, rather than in the usual tandem configuration. There is only a negligible amount of lateral passage of ice into the lower cirque. Ice from the lower cirque flows steeply past the southeast side of Met Hill to the presently backwasting terminus (Figure 5).

A second bedrock knob, Goat Hill, exists in the lower valley. Ice previously flowed to the northwest of this knob also, to a massive terminal moraine at its lower edge. Main ice flow, however, was past the northeast shoulder of the knob.

As shown in Figures 6 and 9, below Goat Hill is a complex arrangement of terminal moraines, most of which appear quite fresh and unweathered. This unaltered moraine system lies well within a more extensive and relatively well-weathered late-Wisconsinan moraine complex.

The maximum extent of Neoglacial advance is represented by an ice-cored terminal moraine, augmented by talus from Mt. Edward Little (ca. 6100 ft, 2040 m). This moraine continues to advance slowly as a low-angle rock glacier in the lower south side of the valley.

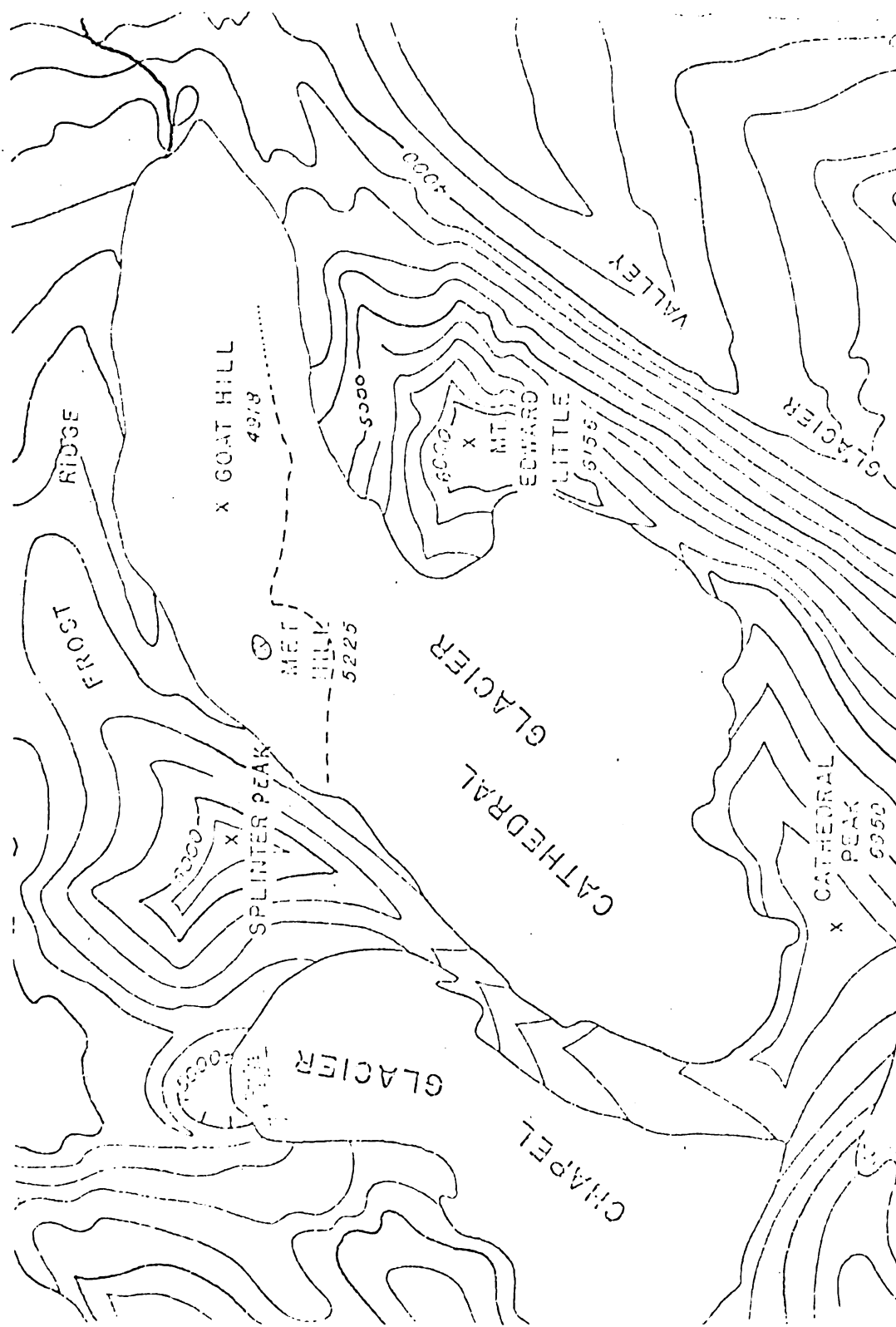


Fig. 6 Topographic Map of the Cathedral Glacier Area (map by C. Cialek)

Key to Figure 9: The Cathedral Massif

- | | |
|---------------------------------|-------------------------------------|
| 1. Met Hill C-29 | 14. Cathedral Glacier, Upper Cirque |
| 2. Present Terminus | 15. Cathedral Glacier, Lower Cirque |
| 3. Goat Hill | 16. Cathedral Peak |
| 4. Lower Valley moraine complex | 17. Banded Glacier |
| 5. Cathedral Creek | 18. Tunafish Canyon |
| 6. Frost Ridge | 19. Snowslide Valley |
| 7. Splinter Peak | 20. Zebra Valley |
| 8. Hidden Cirque | 21. The Coliseum |
| 9. Chapel Valley | 22. Torres Rock Glacier |
| 10. Chapel Glacier | 23. The Cathedral |
| 11. Alcove Glacier | 24. Rock Glacier Valley |
| 12. Balcony Glacier | 25. Mt. Edward Little |
| 13. Southwest Col | 26. Rock Glacier Creek |

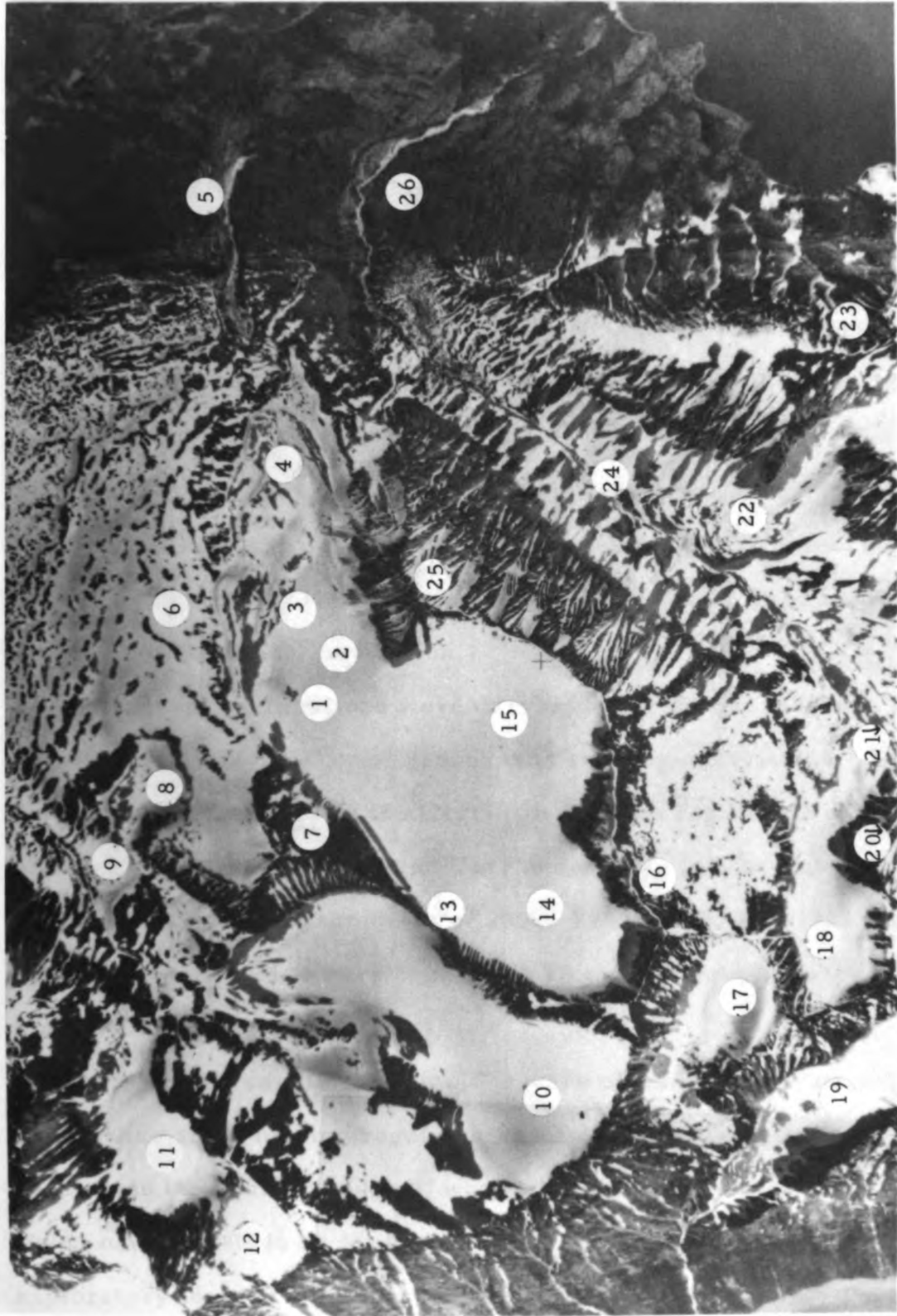


Figure 9 Vertical Aerial Photo of The Cathedral Massif

Comments on Previous Research

The Alaskan Coast

Seagoing explorers in the middle and latter 18th-century described and crudely mapped some of the glaciers along the Alaskan coast. Vitus Bering first viewed Mt. St. Elias and the coastal glaciers in that sector in 1741. In 1786 the LaPerouse expedition discovered Lituya Bay (Klotz, 1899). In 1794 Captain George Vancouver (1801) entered Taku Inlet, with some difficulty owing to extensive drift ice. He passed by the entrance to Glacier Bay, as it was filled with ice. Russian navigators explored the coast between 1788 and 1807 (Tebenkoff, 1848 and 1852), and the Briton, Belcher, in 1837.

The coastal glacier locations and positions were noted and mapped by some of these explorers, not for purposes of glacial research but because glaciers and the many icebergs they spawned were unavoidable facts of life in coastal navigation.

Peripheral observations were carried out from the 1880's to 1913 by numerous individuals and groups with varying motivations. Except for the detailed studies of the Harriman Alaskan Expedition of the 1890's and by I. C. Russell in 1894 and Tarr and Martin for the National Geographic Society in the summers of 1909, 1910, 1911, and 1913 most observations tended to have been made from the decks of ships, or at most not far beyond the beach.

Early in this century the glaciers were popular tourist attractions with cruise ships making regular visits, particularly to the Taku Glacier on the southern edge of what is now known as the Juneau Icefield, named in 1946 by Maynard Miller and colleagues in the initial exploratory team of the Juneau Icefield Research Program. Prior to that time this literally unexplored region was known locally as "the

ice cap". Also, by the time of World War I many persons were visiting the Atlin region to view the Llewellyn Glacier, then touted as the largest in western Canada.

Except for some fragmentary low-level field work done in the 1920's and 1930's (Field, 1932), interest in the Alaskan glaciers waned. The vast upper reaches of the unnamed and unexplored desolation which was the backyard of Alaska's capital city continued to be shunned by white man and native alike.

The Juneau Icefield

Prior to World War II further terminal studies in the Juneau Region were carried out by W. O. Field, Jr., D. B. Lawrence, and M. M. Miller under the auspices of the American Geographic Society (Field, 1947; Lawrence, 1950). Under separate sponsorship this work was extended in 1946 as the Juneau Icefield Research Program (JIRP), directed by Maynard M. Miller of the Geology Department of Columbia University (Miller and Field, 1951 and Miller, 1949). After these low-level studies, plus aerial reconnaissance of the periphery in 1947, it was increasingly clear that the study of the glacier termini alone was entirely inadequate. To understand glacial behavior and regional significance of the glacio-climatic parameters involved, it was also necessary to go into the highland sectors and investigate, by a total systems approach, some of these key glaciers from snout to source.

The first significant scientific penetration of the Juneau Icefield proper was made from the southeast corner between East Twin and West Twin Glaciers in September, 1943. At that time a field weather station was established at 4200 feet (1280 m) elevation near Research Peak on the Twin Glaciers Neve'. The inner icefield was reached

across the Twin Glacier névé to the east, and northwesterly up the Northeast Branch of the Taku Glacier. This initial penetration was made by a field party of six, directed by M. M. Miller and W. R. Latady.

From this beginning nearly 30 years ago a network of 30 permanent research stations and campsites has been extended across the southern and eastern Juneau Icefield. Meteorological observations are taken whenever stations are occupied. The sector to the northeast as far as the shores of Atlin Lake was explored by JIRP personnel as early as 1951. Exploration reached as far as Mt. Bressler (c. 8000 feet or 2440 m) with a field weather station and recording instruments set up at this site in 1962. In 1974 a permanent summer weather station was established at 7200 feet (2200 m) near Mt. Nesselrode (Figure 1). The portion of the Icefield beyond this point to the north and northwest remains virtually unexplored.

The southern maritime flank has thus been studied quite intensively, and synoptic weather data obtained from icefield stations (see JIRP reports, 1950 on). In fact many key concepts and/or methods in field glaciology have had their beginning and/or development within the Juneau Icefield Research Program. Advances have been made in glacier physics, glacier flow and deformation, survey and mapping, palynology, dendrochronology and lichenometry, nunatak soil development, glacier geophysical methods and interpretations, climatic analysis, arctic meteorology, glacier hydrology, snowfield ecology, and the practical subjects of field training and expedition logistics, and general environmental systems education.

The Atlin Lake Region

A major factor in the history of Atlin was gold. The first discovery of gold in paying quantities near Atlin was made in January, 1898. By

that summer Atlin was the center of a gold rush, and geologic and mineralogic information was quickly needed. Thus in 1899 the Geological Survey of Canada sent J. C. Gwillim to carry out a geological and topographic survey of the Atlin District. In 1900 he was assisted by W. H. Boyd. Their resulting publications were in May 1901, a 48-page report (Gwillim, 1901), and in 1902 a geologic map of the region (Gwillim, 1902).

In 1911 a more detailed report was submitted to the Canadian Department of Mines by D. D. Cairnes (1913). In 1955 a detailed geologic map of the Atlin area was presented by J. D. Aitken of the Geological Survey of Canada. Unfortunately this did not include the Cathedral massif, as the area of coverage ended just to the east. While the bedrock geology of the Atlin area has been studied in considerable detail, glacial geology has been dealt with only in a most general way prior to the consideration of this area by the Juneau Icefield Research Program.

As early as 1964 Miller, Lietzke, and Anderson (Anderson, 1970) reviewed the general and glacial geology, as well as the climate and soils of the Atlin region preliminary to a systematic geomorphic, vegetational, and palynological study. One result of this research is a late Pleistocene-Holocene geobotanical chronology for the area. This was later expanded by Miller (Miller and Anderson, 1974) to include supporting glacio-morphic evidences, plus the palynological work done by another JIRP research affiliate, C. J. Heusser (1952, 1960, 1965) in the Taku area on the opposite margin of the Icefield (Figure 10).

Tallman (1972, 1973, 1975) studied glacial deposits and permafrost features in the Fourth of July Creek Valley north and east of Atlin as a contribution to an analysis of the late Pleistocene and Holocene glacio-climatic history of this region.

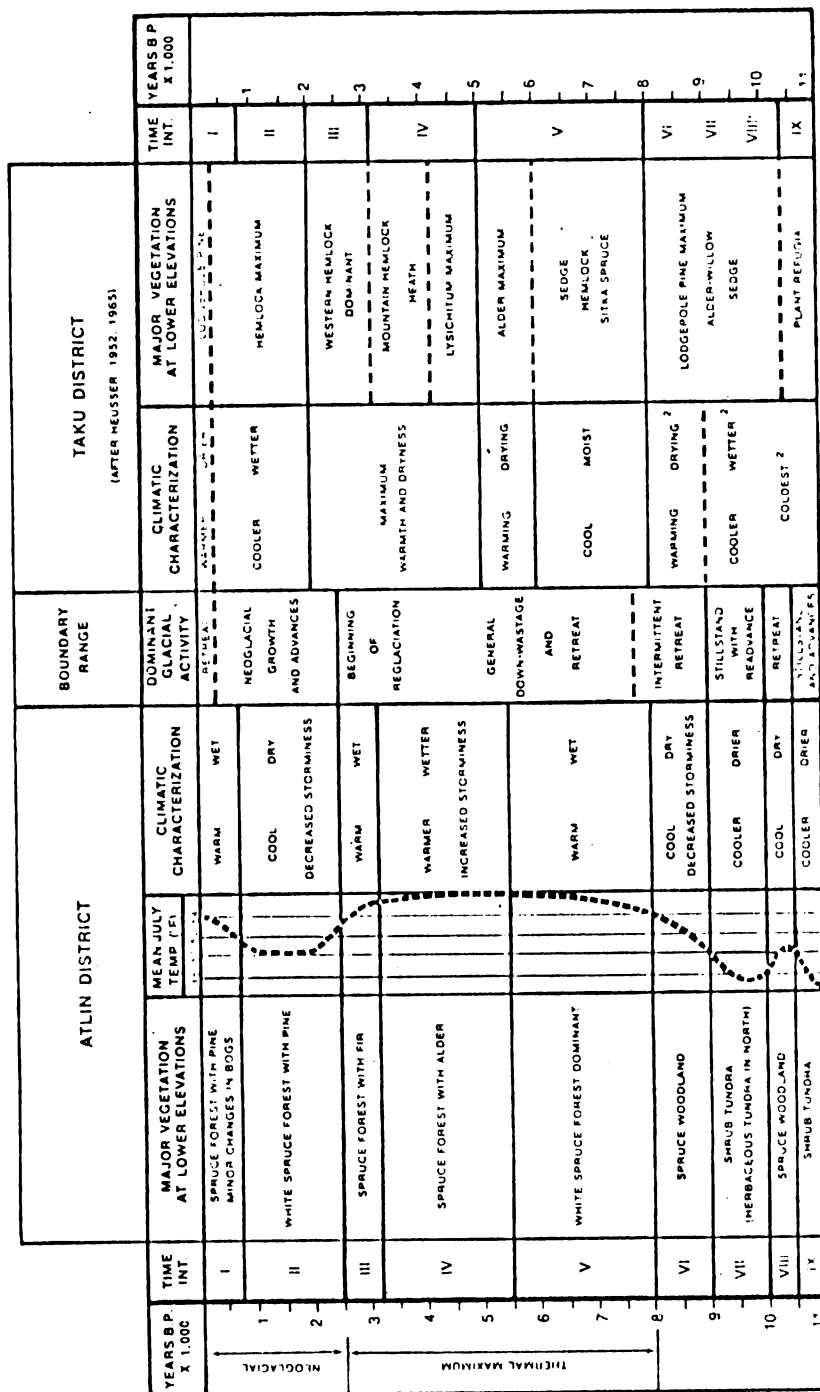
¹ low-neve glaciers² inferred by present authors

FIG. 10 Late Wisconsinan and Holocene glaciobotanical chronology in the Arlin and Taku Districts, Canada-Alaska
(Miller and Anderson, 1974)

The Cathedral Massif

The Report on the Atlin Mining District by Gwillim (1901, 1902) includes the Cathedral massif. Analysis of the geologic map, however, indicates that it was probably mapped from a distance. A few other regional studies mention the Cathedral in passing, with no detail given. JIRP personnel exploring the area in 1972 found a summit register on Cathedral Peak. It contained the names of three Canadian government surveyors, and was dated 1944.

Current Research on the Cathedral Glacier and Environs

In 1971 Miller (1972a) carried out an air reconnaissance of the Cathedral Glacier area as preliminary to development of a research site. Its aspect, area, and nature as a closed basin for hydrologic research provide a very good system for comparison with the Ptarmigan and Lemon Glaciers near Juneau in the maritime sector of the Juneau Icefield (Figure 2).

The Camp 29 field station was established on the bedrock shoulder of Met Hill in 1972 (Figure 5). Facilities consist of a small five-sided building, eight feet on a side, which serves as kitchen, dining and study area. Attached to this is an 8 by 12-foot wing containing bunk space for six, plus a storage loft.

Since the opening of the Cathedral Glacier Research Station in 1972 several long-term research programs have been initiated. A photogrammetric survey begun in 1972 will be continued in 1975 and 1976. Movement and gravity surveys were carried out, especially in 1973. Also in 1973 palynological bog samples were taken within the older moraine area beyond the outermost Neoglacial moraines. A Stevens "A" water recording gage was installed on Cathedral Creek in 1972 and has been in operation each summer since. Despite some equipment malfunction and calibration problems the resulting data have permitted useful

progress reports to be prepared (Miller, 1972a; Nishio, 1973; Guigné, 1974). Three-hourly meteorological observations are taken at all times when the station is occupied. A water-level recorder has been in operation on the proglacial lake near the research station in the summers of 1973 and 1974.

Periglacial studies have also been conducted on patterned ground near the research station. In 1974 a cirque distribution and cirque-floor elevation study was conducted by Squyers (1975) and others. Groundwater studies were initiated in 1974 below the Neoglacial terminal moraine (Zapico, 1974). Lichenometric studies and aerial photography were also carried out in 1974 by Marianne See of the Smithsonian Institution and the University of Alberta (See, 1975).

Continuing studies are under way in the Cathedral area, some of which will be pursued in coming field seasons. It is anticipated that for many years to come this station will be a key link with respect to regional scientific studies on the Juneau Icefield, as well as relating to environmental studies in the Atlin Wilderness Park of the Province of British Columbia.

Geomorphic and Climatic Implications

As the Cathedral Glacier is located on the interior flank of the Juneau Icefield it reflects the continentality of that sector. The Icefield spans a gradient of climatic conditions from the extreme maritimicity of the coastal regions, across an orographic divide which reaches over 8000 feet (2440 m) in this sector, and northeastward to the dry continental interior (Figure 2). This is significant climatologically because the region of the Juneau Icefield is the area over which the North American ridge of the circumpolar westerly circulation fluctuates in the upper troposphere. Thus it represents an interface region during the annual

maximum precipitation period between mean locations of the warm, moist maritime arctic air mass in the Gulf of Alaska (the Aleutian Low) and the cold dry continental polar air mass (Canadian High) of the interior.

These atmospheric factors result in a situation in which there is extreme sensitivity to climatic change. Thus small variations in pattern and intensity of atmospheric processes, amplified by orographic effects, affect the regime of glaciers across the Juneau Icefield.

As we develop an understanding of the relationships between climatic parameters and trends of upper tropospheric ridge locations on and near the North Pacific coast, we should be able to develop a foundation for extending this understanding to previous climatic conditions as recorded by glacial processes. Thus a study and comparison of glacially-formed features on the opposite margins of the Juneau Icefield can provide useful insights into climatic chronology of the region and hopefully of a large segment of the North American sub-arctic.

While the southern maritime flank of the Juneau Icefield Research Program has been studied since 1946, relatively little research was carried out on the continental side until the mid-1960's. This present study represents the first definitive report of research in the area of the Cathedral Massif. As such it provides a foundation in the understanding of present and former regimes of these interior cirque glaciers. As the study incorporates the comparison of climatic conditions over the entire icefield it should provide a beginning of extended analysis of the larger-magnitude climatic patterns affecting the North Pacific coastal region extending some distance into the continental interior.

CHAPTER II

BEDROCK GEOLOGY AND GEOMORPHIC IMPLICATIONS

Lithology

Previous Lithologic Mapping

Because of interest in potential mineral resources, the lithology and structure of the Atlin district have been studied in some detail (Gwillim, 1901 and 1902; Cairnes, 1913; Aitken, 1959). The remoteness and apparent lack of mineral prospects on the Cathedral Massif, however, has resulted in that sector being given only cursory attention. This has meant that very little information is available concerning its bedrock geology.

Hard-rock gold mining was carried on at the Engineer Mine, 15 miles to the northwest (Figure 3) during the 1920's, but by the 1940's that site was abandoned. Some mining development work in copper and molybdenum prospects is under way on Hoboe Creek and in the Willison Glacier Valley 8 to 10 miles southwest of the Cathedral area. These minerals are found in association with granitic or volcanic intrusives, both as lodes and as disseminated deposits injected into the metamorphic complex.

Gwillim's 1902 map shows that the Cathedral Massif above the 3500-foot (1060 m) contour consists of a granite porphyry. This is the same geological association found on Birch Mountain on Teresa Island and the upper part of Atlin Mountain (Figure 3). The lower northeast corners of The Cathedral and Mt. Edward Little, across the bedrock contacts in Cathedral Glacier Valley, and thence toward Edgar Lake he mapped

as "sandstones and conglomerate, chiefly pyroclastic Jurassic (?)" (sic). The dark metavolcanic which forms the crestal summit, including Cathedral Peak, was not noted.

Gwillim did not show glaciers on the Cathedral massif. Rock Glacier Creek was mapped fairly accurately, Cathedral Creek less so, and Chapel Creek not at all. Contour lines are acknowledged to be only approximate. The highest elevation was indicated to be just over 6500 feet (1980 m) in comparison with the present surveyed elevation of 6950 feet (2106 m).

From a distance the granite flanks of The Cathedral, Mt. Edward Little, and Splinter Peak tend to dominate the vista and obscure the central metavolcanic core of the massif. In view of the above-noted inaccuracies it is concluded that the area was initially mapped from a distance. However, considering that Gwillim and Boyd's geological and topographic survey was made over 70 years ago, and covered the entire Atlin region in only two field seasons, their lack of precision should not be unduly criticized.

Aitken's Geological map (1959) is much more accurate and detailed. This survey was based on the government 104N map sheet and thus unfortunately ends just east of the Cathedral area. Because of the similarity with Birch Mountain, however, it is of interest that he mapped that area as a post-lower Jurassic "Coast Intrusion" which is porphyritic and granophyric, petrologic characteristics which are similar to what we find on the Cathedral Massif.

Bedrock Geology from Current Studies

No detailed studies have been made of the bedrock geology of the Cathedral Massif or the adjacent areas. The following discussion is based on reconnaissance-level observations, plus studies of similar areas in the Coast Range.

The highest part of the massif, including Cathedral Peak, is a dark, fine-grained metavolcanic, presumably metabasalt. This rock is possibly related to the metavolcanic series reported by Forbes (1959) in the Devil's Paw area on the eastern flank of the Juneau Icefield. On Cathedral Peak this basic unit gives evidence of having been heavily stressed, with the plagioclase phenocrysts fractured and degenerated by pressure and heat (Flanders, 1974).

The cirques of the upper Cathedral Valley were largely formed in this granodiorite intrusion (probably the same age as the Twin Glacier granodiorite reported as the primary crystalline lithology at the core of the Boundary Range, Miller, 1949; 1963). The up-valley contact zone between these two main lithologies is exposed near the middle of the upper cirque headwall (Figure 7). This contact has an apparent dip of about 70° to the north and disappears under Cathedral Peak. It is exposed again in the headwall of the lower cirque.

Most of the up-valley section of the upper cirque lies in the granodiorite, north of Cathedral Peak proper. At the top of the headwall in this location the granitic bedrock is vertically fractured into large, rough-surfaced slabs, the surfaces of which roughly parallel jointing planes. This mode of fracturing is assumed to be a consequence of the lateral topography in this area, as well as through development of tensile stresses in the crust as upwarping proceeded during the Coast Mountains orogeny in Tertiary time. Downvalley the majority of boulders in the glacial till are of this material, although most are derived from the flanks of Mt. Edward Little and Splinter Peak. Lateral spreading across vertical joint planes dominates those rocks



Figure 7 Lithologic Contact Zone in Headwall of Cathedral Glacier



Figure 8 Cathedral Glacier from Frost Ridge

exposed at the crest of the northwest col above the upper cirque, in a zone composed of the same material.

The contact on the northern edge of the granitic intrusion cuts across the south flank of Splinter Peak and passes northward above the talus slope across the proglacial lake from Camp 29. Close inspection of this contact zone reveals the granodiorite in this sector to be a magmatic emplacement rather than granitized as has been suggested for much of the granodiorite observed on the Juneau Icefield (Gilkey, 1951). In many places the veins interfinger the fractured country rock producing some contact metamorphism. Above this contact the exposed metamorphic rock is highly weathered, especially above the Neoglaciation limit.

Met Hill is composed of the same granodiorite intrusive noted above. Farther down-valley the contact cuts across the mid-point of Goat Hill on a bearing of 100°T . The up-valley part of Goat Hill is the light-colored granitic rock, while the lower part is the dark metavolcanic which in places is highly oxidized and iron-stained. The same dark oxidized facies is exposed across the northeast nose of Mt. Edward Little and the lower northeast face of The Cathedral. To the westward it crosses Splinter Peak, as noted above, and appears in the valley of Chapel Glacier above its Neoglacial moraines. From there toward the west it outcrops in the broad upland above Alcove Valley. The contact zone and change of facies appears to have had little effect on the geomorphology of this highland sector. Joint structure exerts a much more dominant control.

Also exposed in situ on Goat Hill below the igneous contact is a metasedimentary rhythmite, which may serve as a unique index unit in down-valley tills. Here also portions of the bedrock are so iron-rich that fist-sized specimens lying on the very recent glacial till

have stained the adjacent surface for two or more feet downslope. This iron-rich unit may also serve as a till index. Such distinct bedrock units are diagnostically useful in determining provenance of erratics observed in the lower moraines, and they can give evidence as to directions and distances of glacial transport.

Boulders containing tillites or conglomeratic mud-flows have been found in the Wisconsin deposits which flank both sides of lower Cathedral Valley. The origin of at least part of these is a vertically trending bed of clastics exposed near the western end of the northeast ridge of Splinter Peak. A very few pieces of limestone have been found, but their source is unknown. They may be related to the limestone which outcrops south of Willison Inlet toward Hoboe Glacier and Camp 26.

Jointing Structure

Miller (1959) reported four well-developed sets of jointing across the central sector of the Icefield, which have affected the orientation of valley and headwall directions. These four sets are as follows: (Figure 11)

- 1) a primary dominant set, whose strike ranges from 355° to 045° with dips from 80° E to 60° NW.
- 2) a secondary dominant set, with strike running from 315° to 345° T, with dips between 65° and 80° SW.
- 3) a weaker set, striking 260° to 280° T, with a dip close to 90° .
- 4) a least dominant set, nearly horizontal, with dips ranging from 35° south through horizontal to 30° N.

Figure 11 shows schematically the direction and relative dominance of the first three joint sets described above. The fourth set ranges around the horizontal, and thus is not shown on the radial diagram.

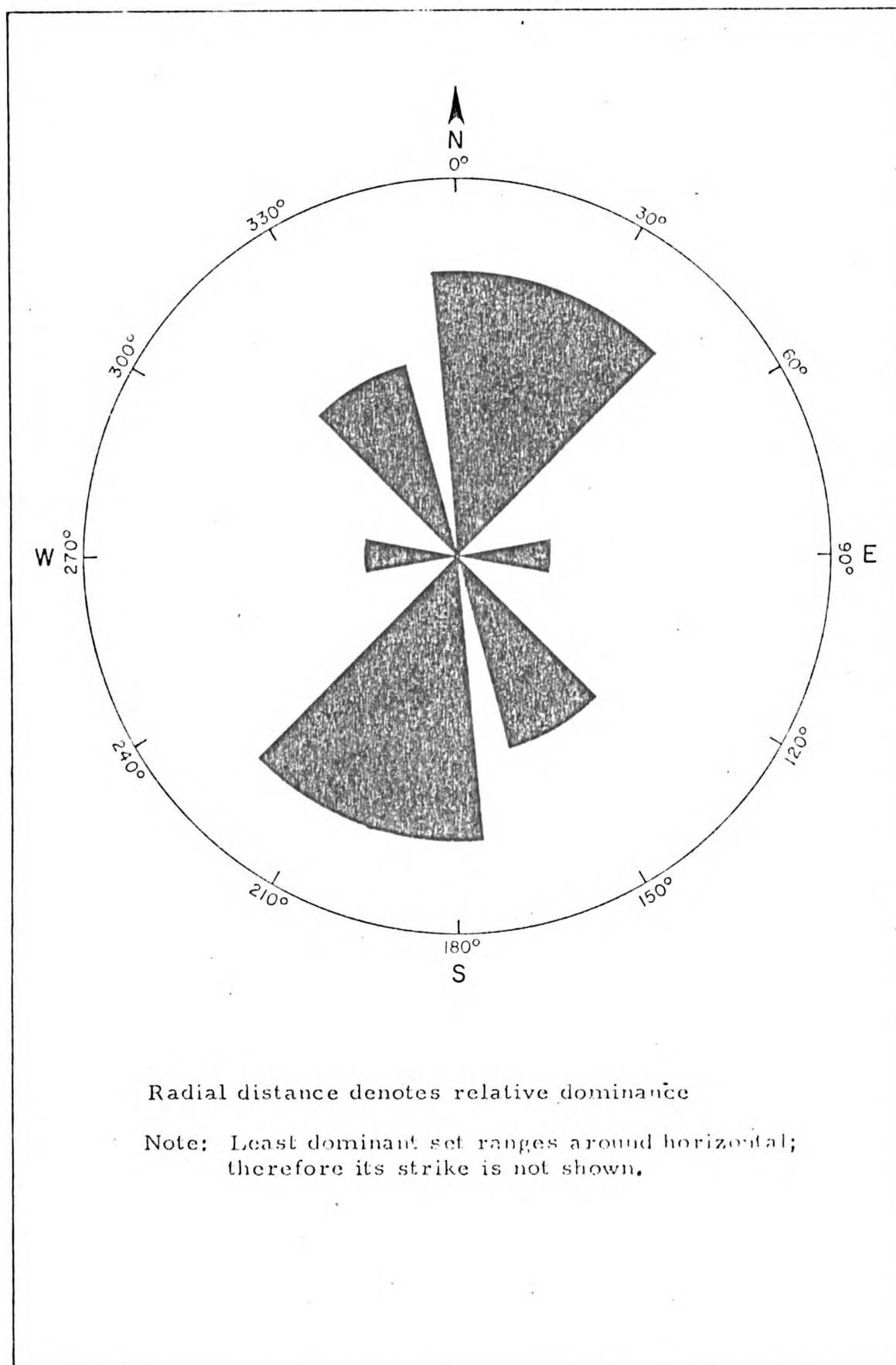


Figure 11 Diagram of the Lithologic Jointing Pattern, Junction Icefield
(adapted from Miller, 1959)

Strike and dip measurements were obtained during the 1972 field season on the northwest col between Cathedral Peak and Splinter Peak (Goodwin, 1973). These measurements are given below (see Figure 12).

- 1) Primary set: Strike 355° to 045° , with most between 15° and 35° ; dip near vertical.
- 2) Secondary set: Strike 265° to 270° ; steep dips.
- 3) Weaker set: Strike 235° to 255° ; dips relatively steep.
- 4) Least dominant set; Strike 295° to 335° ; slight dip NE and SW.

The joint sets in this system, although not directly paralleling those in the joint system of the central Juneau Icefield area as noted previously, are quite comparable with Miller's measurements on the central Juneau Icefield (1963). Egan's measurements on the central and southern Icefield, at C - 18, C - 10, Ptarmigan Glacier, and Norris Glacier areas, show dominance at $330 - 10^{\circ}$ and $220 - 250^{\circ}$ (Egan, 1971). The general similarities in the direction of joint sets suggests that the jointing patterns in the Camp 29 area are related to the stresses affecting the entire northern Boundary Range.

What may possibly be a series of vertical fault fractures cuts across the region from the head of Rock Glacier Valley, across Cathedral Peak, the northwest col, and the western end of Splinter Peak, at about $0 - 5^{\circ}$ T. Some of these are visible in Figure 9. This approximately parallels the strike and dip of the first-order joint set. These features cut linearly across different rock types; display reddish-brown "gouge" along the fracture, suggesting displacement; and leave ridge-top clefts two feet (0.6 m) wide and 8 feet (2.4 m) deep. Further study is suggested in order to determine whether these are actually major jointing planes or true faults.

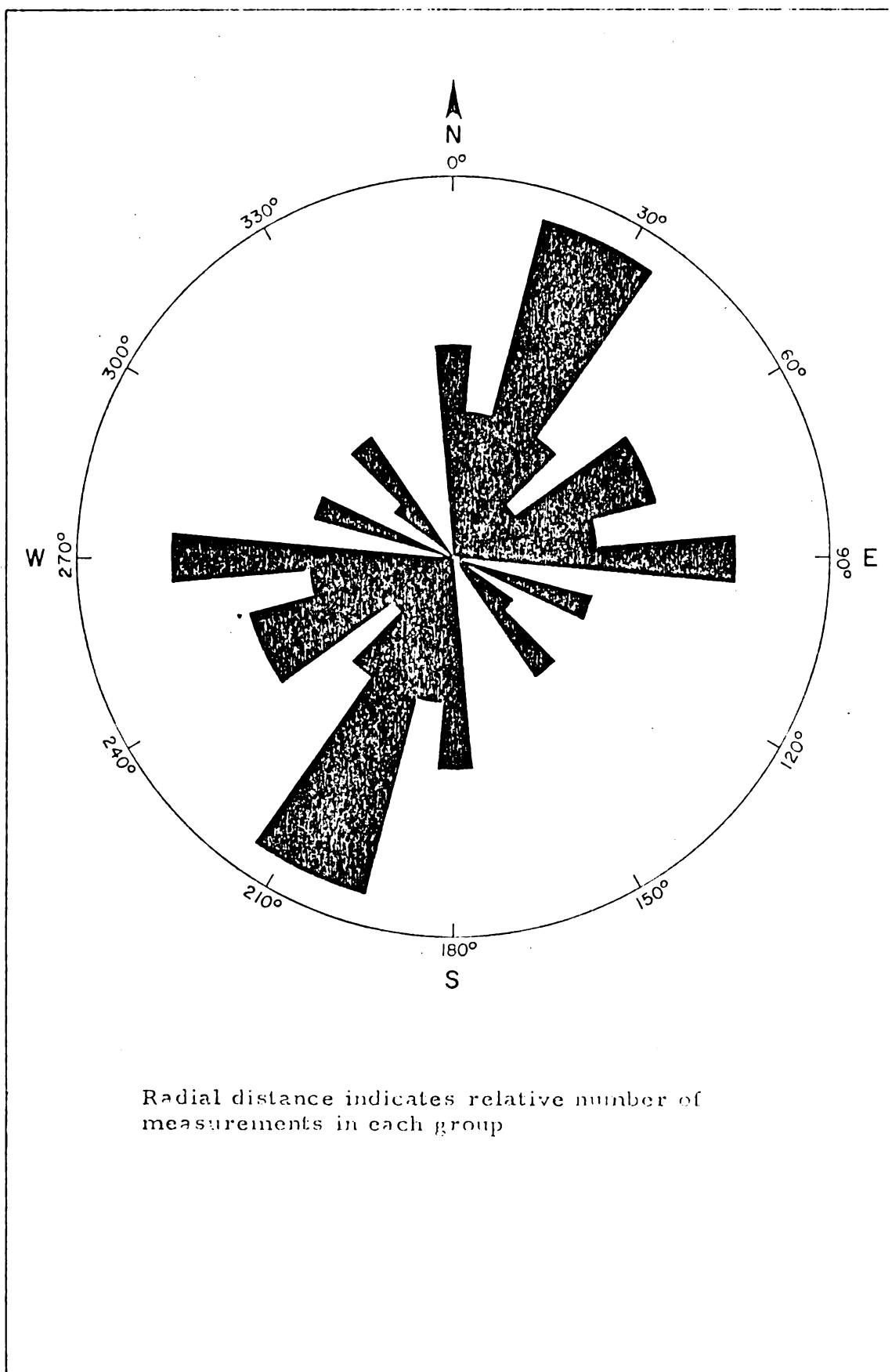


Figure 12 Diagram of Lithologic Jointing Pattern, West Col of Cathedral Glacier System (After Goodwin, 1973)

Geomorphic Relationships

The main axis of the upper Cathedral Valley bears 20° to 25° T. This agrees with the maximum occurrence of the first order joint set as measured on the northeast col. The axis of the lower valley, at 80° to 90° T, agrees with strike of the second order set. It is noted that although the granitic-metavolcanic contact at about 100° T may be related to the 60° deflection of ice flow, the actual contacts between the advancing ice and both Met Hill and Goat Hill are within the granitic intrusion.

In the upper valley the metavolcanic bedrock appears to be more resistant to headwall erosion than the granodiorite. It appears that the granitic rock is more susceptible to joint fracturing. Thus headwall regression would be more rapid in the granodiorite due to slabbing. On the other hand, the granodiorite appears to be more resistant to mechanical corrasion, tending to withstand more effectively any removal by advancing and overriding ice.

A prominent feature on some of the steep bedrock surfaces approximately perpendicular to the first order joint set in the granite is the presence of active rock chutes. On other similarly oriented but less steep surfaces, such rock chutes are absent with the slopes covered by local rock rubble. The rock chutes, where they do occur, are joint-controlled.

CHAPTER III

GLACIAL HISTORY AND GEOMORPHOLOGY

Wisconsinan

Regional Wisconsinan Glaciation

During the maximum Wisconsinan glaciation affecting this region, which is considered to be very early Wisconsinan, the axis of the regional ice center is suggested as being parallel to the present cordilleran crest but ten to twenty miles to the east (Miller, 1964a). This is thought to have covered or nearly covered the 8000 foot (2440 m) peaks at the structural crest of the range. Intensity of subsequent weathering and mass-wasting have removed all direct evidences except for the rounded summit profiles of overridden peaks and relict forms of high cols where transfluent ice flowed toward the coast. Much of the interior ice was probably deflected to the south, and passed through the Taku valley and other major transection valleys farther south. Ice was also deflected to the north through the main valleys of the Tagish-Atlin-Teslin region, excavating and deepening these valleys into what are now the Tagish, Atlin, and Teslin depressions.

The outer limit of maximum Wisconsinan glaciation derived from this part of the Boundary Range northward along the Atlin valley is delineated by a massive terminal moraine complex near Jake's Corner on the Alaska Highway. This is some 87 miles (118 km) north of the present Icefield margin at the Llewellyn Glacier terminus. Mount Minto, an isolated 6913-foot (2100 m) peak in lower Atlin Valley (Figure 3) 30 miles south of the above-mentioned terminal moraines,

appears to be glacially eroded on both lateral margins up to about the 6000 foot (1800 m) level. Above a sharp break in slope and character at that level, the partially concave summit area was apparently not actively overridden by Wisconsinan ice.

Wisconsinan Glaciation on the Cathedral Massif

The extent of Wisconsinan glaciation on the Cathedral Massif is problematical. Ancient, fairly level to slightly concave surfaces of limited extent exist on top of the massif. One such area contains a snowfield near the highest point of The Cathedral. This snowfield feeds a formerly cascading glacier which now, due to ablation of its terminal zone, hangs suspended over the Torres Rock Glacier below. Another upland surface is found northwest of Chapel Glacier and adjacent to the Nelson Lake depression. These surfaces, at an elevation of 6000 to 6500 feet (up to 1970 m) were apparently above the limit of maximum middle to late Wisconsinan valley glaciation in this sector.

High-level ice in Wisconsinan time flowed northeastward out of what is now Willison Bay. This ice spread northward around the northeast flank of The Cathedral, severely eroding this corner into precipitous bedrock cliffs. Glacial debris, carried from or past this scoured zone, trailed off across the mouth of Rock Glacier Valley and was deposited in a massive lateral moraine below the Wisconsinan moraines from Cathedral Valley. Rock Glacier Creek has cut vertically through several hundred feet of this ancient lateral moraine. This segment of the moraine also appears to merge with an eastern valley-mouth moraine once produced by deeper ice moving out of Rock Glacier Valley. There is no evidence of retained deposition of any west-side moraine from this valley. The Willison Glacier detritus also formed a lower-level bench below Cathedral Valley as mentioned previously. Cathedral Creek has deeply notched this moraine as well during Holocene time.

Rock Glacier Valley to the southeast of Cathedral Valley is a deeply incised and geomorphically young valley with steep rock walls on either side deeply incised and ribbed with joint controlled rock chutes. Within this valley is a much younger glacial valley incised into the lower third of the original valley. This is the trunk outlet valley for four major cirque systems. The inner valley was apparently excavated in pre-Holocene time. Its relative age is suggested by the presence of Wisconsinan moraines at the valley mouth.

During the time when glaciation was much more extensive than now, ice in the Chapel Glacier valley deeply incised the sidewall of the northwestern flank of Splinter Peak. This valley is now nearly empty of ice with only a small remnant of Chapel Glacier further up at the head of the valley. The portion of Chapel Valley by Splinter Peak contains late Neoglacial moraines. This flank of Splinter Peak is now incised with precipitous rockfall chutes, cross-cut by the fault lines described earlier. A weathered moraine, similar in age to the Wisconsinan moraine just north of Camp 29, blocks the valley at about the 4600-foot (1400 m) level.

Snowdrift Cirque is notched into the northeast side of Splinter Peak and drains into Chapel Valley. Its northeast side is a smooth and well-rounded valley-wall soil surface obscuring the bedrock. Its south side, in striking contrast, is a rugged and ice-eroded headwall carved into the lower slopes of Splinter Peak. A small remnant of ice remains along this bedrock south wall and in the debris of its 4780-foot (1460 m) floor. This cirque appears to have been formed by ice nourished from snow drifted off the slopes of Splinter Peak by southeast winds. Present accumulation of snow blowing in from the southeast takes place directly up-slope from this point, around the 6100-foot (1848 m) level.

Here ice has never built up deeply enough to begin gravitational movement. The significance of this will be discussed in Chapter VI regarding cirque levels.

These cirque levels are significant geomorphic features of the glacial valleys on the Cathedral massif. They appear to be grouped at four fairly uniform elevations, an observation concurred in by Tallman (1975) and Miller (1975a) from studies in other sectors of the Atlin region. The lower two levels are currently free of ice, and are well-mantled by glacial debris. The upper two cirque levels contain active ice. The nature and climatological significance of these cirque levels will be discussed in Chapter VII.

Pleistocene Glaciation in Cathedral Valley

In Cathedral Valley headward erosion by Cathedral glacier and mass wastage still proceeds, with active rock chutes exposed above the névé in the granodiorite intrusion. On the other hand, much of the upper valley wall on the slopes of Mt. Edward Little and Splinter Peak are merely boulder slopes, with trim-lines indicated by the suppression of lichen growth. On the north flank of Mt. Edward Little, which is the south wall of the lower valley (see Figure 8), there are surface evidences of glacial erosion high on the bedrock walls. This high-level erosion zone is well above a talus-mantled ice slope which presently extends upward from the ice-free valley floor (Figure 5). Here lack of direct solar insolation in the shadow of the mountain appears to sharply decrease the ablation rate. This suggests that the high-level erosion zone in this location results from headwall-type processes as well as being a glacial scour zone.

Moraines are better indicators of chronologic sequences than are erosional features in bedrock. They are more conducive to erosion,

weathering, plant colonization, soil formation, and mass wastage than are bedrock surfaces. Thus in this study specific attention has been given to the depositional stratigraphy of this key valley.

Ancient moraines in the Cathedral Valley extend some 1000 meters beyond the flank of Mt. Edward Little on the east side of the valley. On the west side a long, gently convex ridge extends nearly 1.5 km north-eastward from Splinter Peak. This ridge comprises the north side of the lower valley, and is capped by morainic material which obscures the transition from the metavolcanic and metasedimentary bedrock to the moraine.

The old arcuate lateral and terminal moraine originating from Cathedral Glacier ice during the late Wisconsinan (Figure 5) is from 50 to 400 feet (15-120 m) high, and extends up to a quarter-mile beyond the terminal moraines of Neoglacial age. A proglacial lake about 5 acres in area existed within this moraine system long enough to develop a strandline and permit deposition of lacustrine sediments, portions of which are visible today.

The outermost of the oldest moraines in this terminal area appears to have been distorted by ice flowing eastward across the Edgar Lake transfluence from Taku Arm in the Tagish Lake sector (see Figures 7 and 9). This in turn is partly obscured by the placement of the younger (Wisconsinan) terminal moraine. The chaotic nature of this complex depositional zone indicates that this was a plexal area between early Willison and transfluent Taku Arm ice during middle Wisconsinan time.

The Holocene

Holocene History of the Cathedral Valley

Because of the intensity of late Holocene (Neoglacial) glacial activity, evidence of the early Holocene history of this area is rather sketchy. A

few minor recessional moraines, veneered by thick heath mats, exist in the north central part of the valley below the obvious Neoglacial moraines. Retreat from the maximum extension appears to have been rather rapid, as there is little ground moraine, but most evidence has been destroyed by the intensity and extent of Neoglacial advances.

The period of the past few hundred years has been the most conducive to glacierization of any time since construction of the outer heath-covered moraines which are suggested to represent late Wisconsinan glaciation. This is similar to the situation in the Pacific Coast sector of Washington state and in the Wallowa Mountains of northeastern Oregon. In these locations the overriding of older Neoglacial deposits by younger indicates increasing severity of climates with time. In contrast, decreasing severity of Neoglacial climates with time in the Wyoming and Colorado Rocky Mountains and in the Sierra Nevadas of California is indicated by the successively more restricted extent of younger advances (Kiver, 1968 and 1974).

The chronologic designation of late-Neoglacial time as used in this report is "Little Ice Age" (Matthes, 1949), a terminology which will hereafter be applied to designate the relatively fresh unweathered and unoxidized morainal materials found in the Cathedral Glacier Valley. These stand out conspicuously in the oblique aerial photograph of Figure 53.

CHAPTER IV

REGIONAL CLIMATIC ENVIRONMENT

Southern Alaska Coast and Immediate Interior

The climates of the cordilleran region of northwestern North America show these four fundamental characteristics:

- 1) maritimity - a warm and very humid coastal area with mild winters;
- 2) a heavy single winter precipitation maximum in the maritime zone;
- 3) an abrupt gradient of continentality toward the interior;
- 4) cold, dry winters and mild, dry summers in the interior.

The Maritime Zone

The Japanese Current or Kuroshio (Marr, 1970) brings warm surface water northward in the Gulf of Alaska approximately parallel to the coast between 50° and 60° north latitude. From there it is deflected westward by the configuration of the Alaskan panhandle, and eventually degenerates into a small gyre in the Gulf of Alaska. In January the mean Gulf surface water temperature is about 20°F higher than the mean air temperature (Trewartha, 1961, p. 251). These surface waters experienced significant warming during the 1940's and 1950's, with a cooling trend in progress since the late 1960's (Fleming, 1963, per Miller, 1975, personal communication).

The maritime region of western North America has a distinct single rainfall maximum in the cooler part of the year. The month in which this maximum occurs is a function of latitude. At Anchorage and

Valdez (61-62°N) the maximum occurs in September or even late August. At Juneau (58°) and Ketchikan (55°) the maximum is usually in October.

The coastal precipitation maximum reaches Vancouver, B. C. (49°N) and Seattle, Washington (48°N) by November and finally arrives as far south as Los Angeles (34°) in February (Trewartha, 1961, p. 270).

The relationship between month of heaviest coastal rainfall and latitude is associated with seasonal movement of the tracks of cyclonic centers. During the minimum rainfall season (April-June at Juneau) the sub-tropical anticyclone is displaced poleward, in turn displacing the major mid-latitude cyclones to about 60-65°N. As the sub-tropical high retreats southward the jet stream and cyclonic belts follow, bringing maximum precipitation for the year (Trewartha, 1961, p. 273).

The reverse does not occur in spring with the northward migration of the sub-tropical anticyclone. Rather, the jet stream appears to weaken in the south in late winter, while another develops near the Arctic Circle (Trewartha, 1961, p. 270).

Also during the autumn through spring months the entire length of the north Pacific littoral has heavy and prolonged rainfall, with snowfall at higher elevations along the ranges parallel to the coast. With snowfalls of this magnitude at the higher elevations, intense glacierization results.

The Continental Zone

The gradient of continentality found on transects extending inland from the coast is characterized by:

- 1) heavy precipitation and maximum humidity in the coastal segment relative to the interior

- 2) increased ranges of diurnal and annual temperatures toward the interior
- 3) decreased cloudiness and increased insolation toward the interior

The almost continuous high crest of the cordillera along the coast greatly restricts the penetration and influence of the maritime circulation. Pacific air as such is not excluded by the barrier, but the air that does penetrate to the interior is derived from upper level sources and hence has a low vapor content (Bryson and Hare, 1974, p. 52). The result is a rather abrupt transition from the warm, humid, and temperate coastal climate to the dry, cold, sometimes arctic conditions in the continental interior. This transition occurs within as short a distance as 100 miles on a transect extending inland normal to the coast. This contrasts with some other regions such as the Mediterranean, where strong zonal flow without significant barriers extends oceanic influences as far as 2000 miles inland from the Atlantic Ocean.

The sharp gradient of continentality across the Juneau Icefield results in a semi-arid condition in the interior. For example, at Atlin the mean annual precipitation is 11 inches, compared with 87 inches at Juneau and upwards of 200 inches at Ketchikan and Sitka on the outer coast. Also in the interior winters are dry, having little snowfall, with temperatures dropping to as low as -60°F .

Thus the autumn through spring climate of the interior flank of the cordillera, especially from northern British Columbia into the Yukon, is greatly influenced by a high frequency and a high intensity of anti-cyclones. One result is cold air drainage with high winds characterized by explosive gustiness in the transection valleys (such as the Taku) which connect the interior plateau region with the coast.

Climate of the Juneau Icefield

Climatic Parameters and Their Effects

The climate of the Juneau Icefield is conditioned by the following parameters:

- 1) the coastal maritime situation, intensified by the warm Japanese Current
- 2) the autumn precipitation maximum
- 3) the sub-arctic latitude, between 58° and 60° N
- 4) the presence of a linear coastal cordillera, rising from sea level to 8500 feet in elevation, significantly affecting air mass movements
- 5) the sharp interface between continental and maritime air masses
- 6) the position of the North American ridge in the circum-polar high-level westerlies

The effects of the above parameters on the regional climatic character and its local expressions include:

- 1) the pronounced gradient of continentality as noted previously
- 2) a heavy autumn and winter snow accumulation at higher elevations
- 3) secular changes in freezing levels and identifiable shifts in elevation of maximum snowfall
- 4) secular horizontal shifts in total amounts of precipitation
- 5) secular shifts in the cyclonic tracks along which low-pressure air masses (cyclones) migrate inland from the Gulf of Alaska.
- 6) secular shifts in the patterns and locations of the formation of high pressure cells (anticyclones)
- 7) secular variations in the effective local directions of storm winds moving counterclockwise around the centers of traveling cyclonic storms
- 8) effects of all the foregoing on the variation of glacier regimes on and near the Juneau Icefield

Effects of the Cordillera

There are two primary climatic effects of the presence of the north-northwestward trending Boundary Range. First, as we have seen, the presence of this barrier to humid offshore air physically limits moisture transport to the interior. The adiabatic cooling of the air as it is forced to ascend to higher elevations results in large amounts of condensation and precipitation. This is followed by down-slope air movement across the interior flank of the range which enhances warming and lowers relative humidity, thus decreasing cloudiness and precipitation. These mutually related effects lead to the semi-arid conditions in the Atlin region and the Cathedral Glacier area.

Secondly, cyclonic storm cells, most of which move northeastward out of the Gulf of Alaska, tend to move across the coast in a direction approximately normal to the cordillera. In contrast, subsiding anti-cyclones tend to move southeastward in the interior parallel to the linear trend of these ranges. Thus the interior is shielded from precipitation, while the coastal area is protected from cold air outbreaks, except as cold interior air may drain through trans-range river valleys such as the Taku valley.

Related Climatic Studies

Analysis of mean daily temperatures at six research camps across the Juneau Icefield (Figures 2 and 58) was made with respect to data obtained in the 1971 JIRP summer field season (Jones, 1972). This verifies the pronounced continentality gradient and shows the strong orographic effect. The field data encompass a 130-mile transect from Camp 17 near Juneau, across the crestal zone near Camp 8 and on northward to Camp 30 on the eastern shore of Atlin Lake.

The records reveal that in summer the daily ranges of temperature increase substantially along the transect, with allowance being made for elevation effects.

It was also found that diurnal temperature variations in the summer months were generally parallel at all stations. On a daily basis a sequential movement of warming or cooling trends across the Icefield was not discerned.

Andrews (1972) based a study of solar radiation and duration of sunshine on the 1971 weather records from the Juneau Icefield. His analysis reveals that those stations under continental influence receive considerably larger amounts of sunshine during the summer months than those in the maritime sector. He identifies a regional "sun line" whose position corresponds approximately with the water/ice divide near Camp 8 on the orographic crest of the Juneau Icefield.

Thompson (1972) has made a corollary study of the continentality gradient across the Juneau Icefield. In this he selected two days from the 1971 summer meteorological records from all available Icefield stations. One day represented fair-weather conditions and the other was a stormy day. He found that the gradient appeared in the profiles for both conditions, but was much more pronounced in the fair weather case. In this the sharpest transition was just inland of the orographic divide. On the stormy day the coastal area was of course characterized by extreme maritimity. In this case the maximum gradient was found to lie more than 25 miles inland from the orographic divide, reflecting pronounced inland penetration of the storm belt.

Climate of the Cathedral Glacier Area

The Cathedral Glacier is located in the semi-arid inland zone (Figure 2) at a location where no previous meteorological observations have

been carried out. The nearest weather records are from Atlin, 18 miles (29 km) to the north and 3000 feet (915 m) lower, and from Camp 26, which is situated on the Llewellyn Glacier 22 miles to the southeast.

Cathedral Glacier is a small glacier in a closed basin. Its unique configuration has resulted in a complex morainal history, but one which provides an unusually complete picture of glacio-climatic pulsations in the late Holocene.

Comparison of the meteorological, glaciological, and geomorphic information from this location with that from maritime stations on the southern flank of the Juneau Icefield permits a much more effective analysis of atmospheric behavior as discussed in the following sections.

The Nature and Significance of Tropospheric Controls

The location of the Juneau Icefield at the fluctuating interface between the polar continental anticyclonic high pressure region and the maritime cyclonic mean low-pressure region during the season of maximum precipitation gives the Icefield a distinct and unique sensitivity to shifts in the location and/or relative intensity of these atmospheric phenomena. These in turn are affected by changes in the general circulation of the atmosphere. In addition, the presence and orientation of the coastal cordillera provides additional control and amplification of changing conditions, and elevations sufficient for glaciation make possible the preservation of a natural record of long-term atmospheric conditions.

It must be kept in mind that the North Pacific/Aleutian low pressure system which dominates the Gulf of Alaska and the southeastern Alaska coast during the annual precipitation maximum of the

fall and winter seasons is not a static low pressure condition. Rather, atmospheric conditions in this area are part of the planetary circulation pattern. Long waves of planetary scale (wave-lengths of 60° to 180°) steer the movements of weather patterns, while shorter waves superimposed on and within the mean long wave positions tend to generate and move individual weather systems. The most probable locations, as well as the intensities, of these wave patterns shift over time. Thus the most probable locations and movements of weather patterns, on a frequency distribution basis, may be expected to vary over time.

The "Aleutian Low" represents a condition in which the planetary scale high-level westerlies tend to be moving from a wave trough to a ridge as they traverse the Gulf of Alaska. In the fall and winter seasons low-pressure cyclonic storm systems tend to generate in and move through this area, intensifying as they move from trough to ridge. This may also occur in spring and summer, but storms tend to be farther to the north. These storms, moving in a generally northeast direction, bring clouds and heavy precipitation to the coastal area. As the storms pass the ridge of the tropospheric wave they tend to turn southeastward and deteriorate. If they maintain their coherence until reaching the base of the trough over the United States and begin to move ridgeward again they may reform and/or intensify into rejuvenated full-scale storm systems.

The "polar Canadian High" is a condition in which the high-level westerlies tend to be moving from a wave ridge to a trough. In this situation high pressure anticyclones tend to be generated in western Canada and move southeastward across the general region of Montana.

These "Highs" bring cold fronts across the central United States and help to generate new storms as they encounter warm air systems from the south.

Therefore when we speak of the interface between the Aleutian Low and the Polar High we are actually speaking of the location of the planetary wave ridge crest of the high-level westerlies over the southeast Alaskan Coast in the vicinity of the Juneau Icefield. When the ridge is more frequently located westward of its "normal" position, conditions of continentality move coastward. That is, mean atmospheric pressures are higher, cold air mass characteristics are more dominant, precipitation decreases, freezing-point elevations become lower, and low-level glaciers expand.

When the ridge-crest is more frequently located eastward, maritimicity prevails farther inland, precipitation increases, temperatures are higher, freezing-point elevations become higher, and high-level glaciers expand.

Relationships between relative tropospheric wave ridge positions and precipitation patterns are shown in Figure 13. Air in the area preceding the ridge position, in the west-to-east air flow, tends to experience uplift, condensation, counterclockwise advection of moist air out of the Gulf of Alaska, and a consequent steering and intensification of low pressure storm systems. Air moving from the crest to the following trough tends to experience anticyclogenesis, with subsidence of cold dry upper air, clearing, and clockwise advection of cold dry polar continental air.

Thus it is apparent that a fairly small shift in tropospheric wave location relative to an area on the ground can change that area from a heavy to a light precipitation regime, or vice versa. A persistent



Fig. 13 Relationship of Upper Troposphere Pressure Patterns to. Precipitation (adapted from O'Connor, 1963)

displacement of the mean ridge location for an extended period of time will thus mean consequent changes in glacier mass balance.

Miller's studies (1956, 1964b, 1972b, 1975a) have demonstrated the significance of geographical position and elevation differences in various glaciers in terms of their apparent out-of-phase pulsations in recent centuries. Miller and Anderson (1974) have related this concept to longer-range fluctuations in climate-related glacier behavior along the coast-to-interior climatic gradient discussed earlier in this chapter (Figure 10). Tallman (1975) has related this further to Pleistocene and Holocene glacial fluctuations in the Atlin region. Thus an understanding of atmospheric controls has important paleo-climatic implications and especially with respect to glacial and Holocene history which is the prime concern of the present study.

The glaciers of the Juneau Icefield have been noted to be particularly effective recorders of past climatic conditions (Field and Heusser, 1952). They not only provide a means of correlating past fluctuations with existing historical records, but they also present a history book which goes back far beyond any records or legends of man. An understanding of the chronology which the glaciers have transposed into physiographic features gives unique insight into the past and also into the probable future.

Because the quantity and form of precipitation on the Juneau Icefield are so sensitively affected by the pattern of planetary tropospheric westerlies, fluctuations of the ridge positions over this region are revealed by accumulation changes in these glaciers. Presence of the northwest-trending cordillera accentuates the climatic contrasts across the icefield, in addition to orographically amplifying atmospheric changes.

As present and past glacial oscillations reveal the nature of climatic variations they also indicate the position and character of planetary wave patterns. Thus climatological studies across the Juneau Icefield, of which the Cathedral Glacier program is a part, have significance on a far wider scale.

CHAPTER V

METEOROLOGICAL RECORDS AND ANALYSIS

Introduction

This chapter deals with the presentation and analysis of the glacio-meteorological data related to the study area.

First, existing daily temperature and precipitation records from meteorological observations at the Cathedral Glacier station (Camp 29) are compared with those at the Atlin station (Camp 30). Although these daily records are available only for the summer field seasons of 1972, 1973, and 1974--i.e. since establishment of the Cathedral station--they provide some basis of comparison between the research area and Atlin, 18 miles to the north, where long-term weather records are available.

The official records for Atlin covering about a half century of data are then examined. Because the Atlin record is incomplete, however, a regional data base for this continental interior area is synthesized, using available data from Atlin and from the Whitehorse Airport and Carcross, Y.T. (Figure 2). Similar long-term records for the Juneau coastal area are synthesized from available records from Juneau and from the Juneau airport. The climate on the coastal and interior margins of the Juneau Icefield are compared.

There are two primary reasons for including coastal records in a study which is mainly concerned with the interior. First, little data exist for the Juneau Icefield interior and no records at all exist for the Cathedral area prior to 1972. In contrast, observations have been taken at field sites on the southern flank of the Juneau Icefield for 30 years.

Therefore, if relationships between the coast and interior can be established for recent data and for long-term regional comparisons, we have a much sounder basis for dealing with the past, present, and future situations in the study area.

Second, as has been already demonstrated, a sharp gradient of continentality exists between the coastal sector and the immediate interior. It has also been noted that this gradient is the area over which the tropospheric pressure ridge fluctuates. Therefore similarities and differences in climatic trends across this region have synoptic significance.

Summer Meteorological Records

Sources of the Data

During the summer field season detailed meteorological observations are taken every three hours (0600 to 2100) at each occupied station on the icefield (Appendix A). Thus, except for unavoidable problems such as equipment breakage or malfunction or necessary field operations which prevented the presence of an observer at the scheduled observation time, complete synoptic meteorological records are on file for Camp 29 for the following periods:¹

1972: July 1 - September 19

1973: July 2 - September 18

1974: July 5 - September 12

Similarly, records are available for Camp 30, the JIRP Sub-Arctic Research Station at Atlin. These were obtained by the same standardized procedure, giving a valid comparison of observations at these two stations.

¹ Foundation for Glacier and Environmental Research, Pacific Science Center, Seattle, Washington, 98109

Analysis

Comparison of mean daily temperatures is shown for both stations for the three years of the Camp 29 record (Figure 14). Temperatures generally followed the same trends. Although Camp 29 temperatures ranged from 15°F lower to 4°F higher than Camp 30, during the summer field season C-29 was usually about 5° to 7° cooler.

Precipitation at the two stations is shown in Figures 15 and 16 spanning the period of record for C-29. For the entire period of record over the three summers, Camp 29 had 12.03 inches of precipitation, with 12.09 inches at Camp 30.

Despite the similarity in the total amount of summer rainfall, there were many discrepancies within single storms and on single days. It is noted that precipitation in this region is almost entirely due to the passage of low-pressure systems rather than to convective instability. This is in marked contrast to a region such as the central continental United States, where the largest part of summer precipitation comes from convective instability. In such a situation sharp discrepancies are observed even between adjacent stations in a dense network of rain gages.

From comparison of daily precipitation during the field season we may conclude that rainfall records at the Atlin station provide a fair approximation of summer rainfall at the Cathedral research station. However, we cannot safely extrapolate this conclusion to the remainder of the year, due to vastly different atmospheric conditions during the colder seasons. As an example of variations in winter conditions, see the April 1 snow depth data for Atlin and Log Cabin later in this chapter (Figure 27).

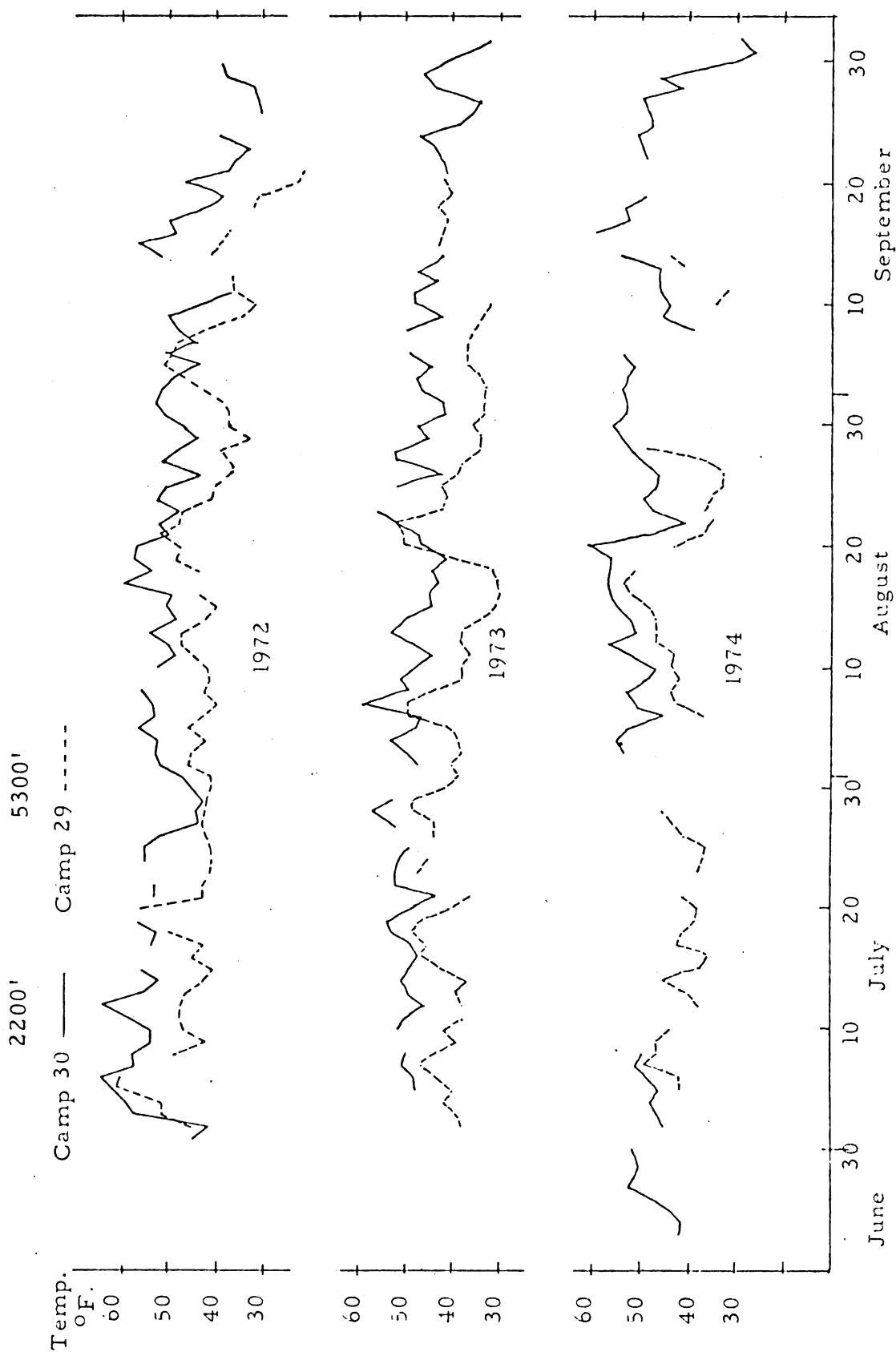


Fig. 14 Daily Temperatures, Camps 29 and 30, 1972 - 74 Summer Field Seasons

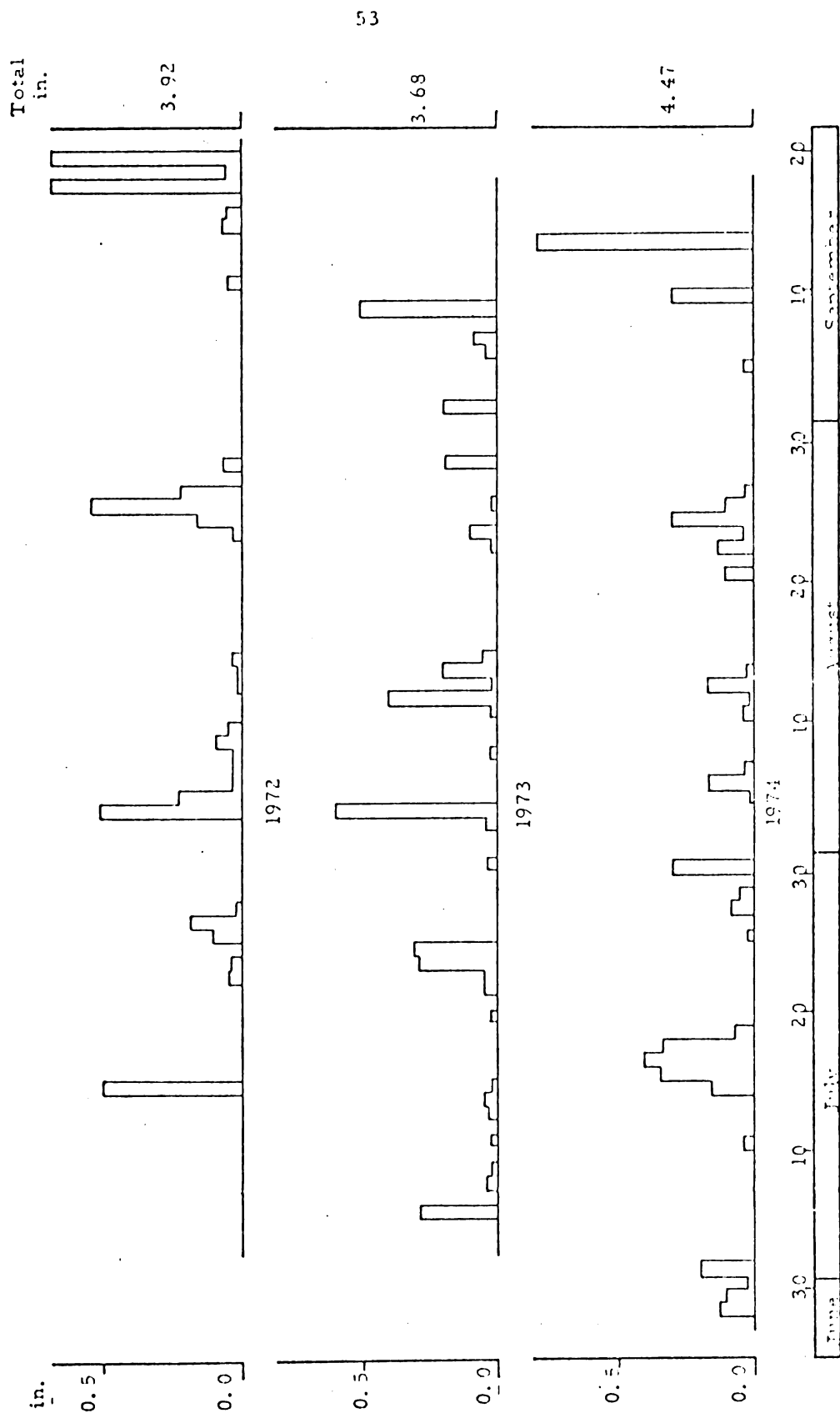


Figure 15 Daily Precipitation, Camp 29 (Cathedral Station), 1972-1974 Summer Field Seasons

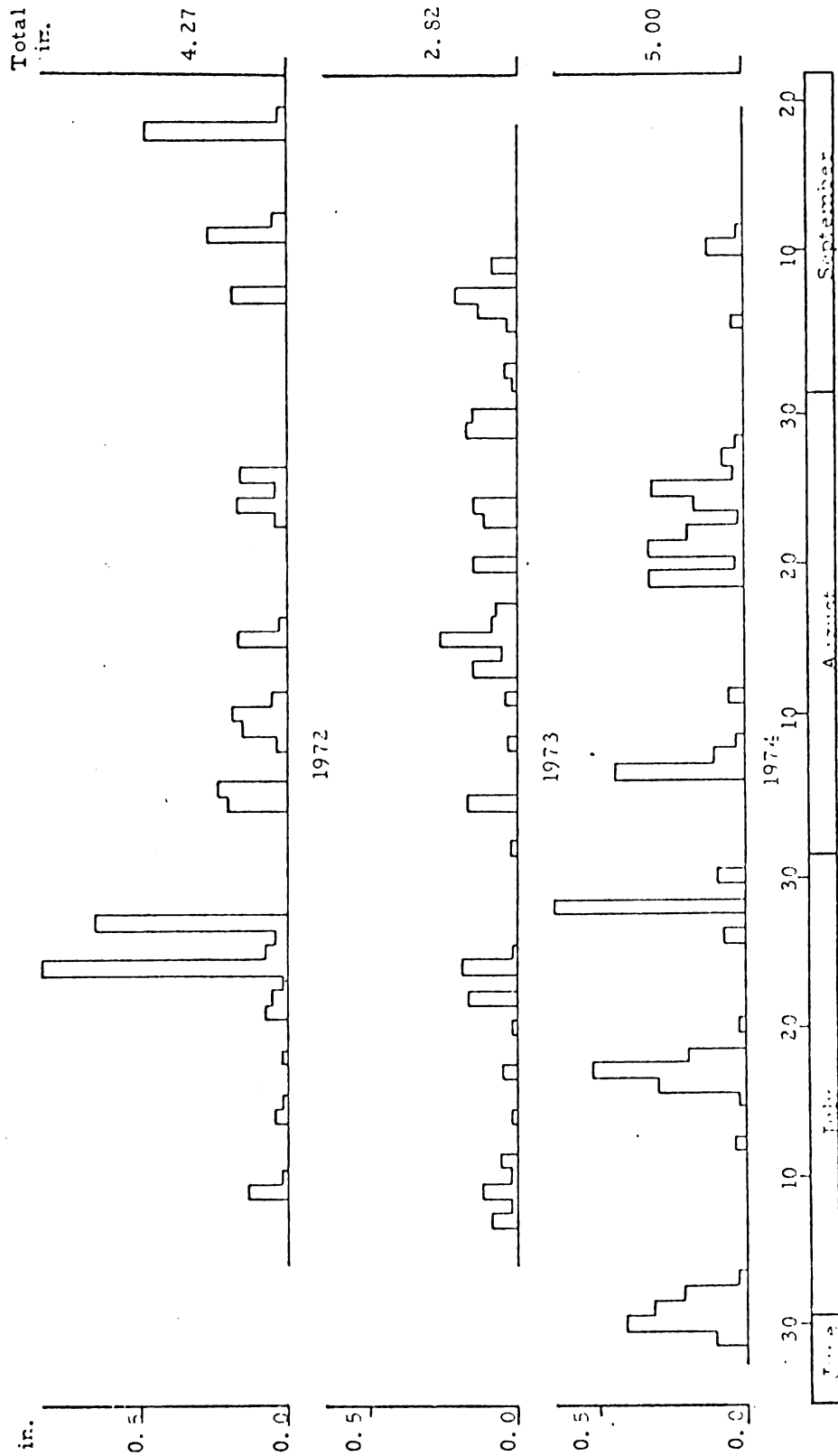


Figure 16 Daily Precipitation, Camp 39 (Aulin Station), 1972-1974 Summer Field Seasons

Long-Term Meteorological Records

Sources of the Data

For this study, official government weather records were obtained for first order stations at Juneau Airport and Whitehorse for the period of record which began in 1942 and 1943 respectively. Records were maintained at the remaining stations by non-professional cooperative observers for varying periods of time, as listed below. Parentheses indicate a fragmental or incomplete record for the years shown (see Figure 2 for site locations).

Atlin, B. C. (government agent office)	1906-45; 1967-71; (1972-74)*
Carcross, Y. T.	(1909-13); 1914-26; (1935-40)
Engineer Mine, B. C.	1926-28; (1929)
Whitehorse Airport, Y. T.	1943-73
Juneau (City) Alaska	(1905-11); 1912-69; (1970-71)
Juneau Airport	(1942-45); 1946-73

At other stations near enough to the study area to be considered directly relevant the periods of record are too short and/or too fragmentary to be useful in the analysis of secular changes involved here.

Data Preparation

Complete records do not exist for any single station in the region. Therefore, in order to have data covering an adequate time span it is necessary to develop composite and representative data sets from existing portions of records from different stations. In doing this we adjust temperature and precipitation data to common equivalents in order to avoid introducing apparent variations which can in fact be the result of combining records from different locations in the area.

.....

* It is also noted that the JIRP station at Atlin has been maintained with full synoptic records obtained over the years 1969-75. These are in more detail and more consistent than the government agent records referenced here.

For example, if precipitation data are obtained from Juneau City records until the United States Weather Bureau Office was established at the airport, and then data are obtained only from airport records, it can appear that the Juneau Area annual precipitation had dropped abruptly by 38.5%, when in fact no such change occurred.

By converting data for each region to long-term equivalents as described here, it is possible to develop a continuous data base for both the coastal and continental interior regions covering the years 1907 through 1973. This particularly pertains to the comparison of mean annual temperatures and total annual precipitation.

To obtain a set of mean annual temperatures which are representative of the immediate interior region near the study area, the means of stations with overlapping records are compared. In this, for example, it is found that Atlin and Whitehorse Airport temperatures are very comparable, while Carcross temperatures average about 1°F lower.

In determining representative coastal temperatures, it is also found that for the period of common record the Juneau Airport temperatures average 2°F below Juneau City temperatures. Therefore where both figures are available the mean is used; otherwise we use city temperature minus 1°F or airport temperature plus 1°F . The consistent discrepancy in mean annual temperatures may be due to: 1) rooftop exposure of the city instrument shelter; 2) the city "heat island" effect; 3) the location of the airport downstream from the Mendenhall Glacier valley, exposing it to katabatic wind effects; and 4) the airport's position, exposed to the open water of Lynn Canal.

It is also noted that since 1959 the average mean annual temperatures of all stations in southeast Alaska followed the Juneau area temperature trend, but with values consistently about 1.5° higher.

Analysis of Temperatures

Temperature records for the immediate continental interior, representing available data nearest to the study area, and for the Juneau area are analyzed by several methods. In this case the data were plotted graphically and processed by computer, and the statistics developed from computer analyses were tested for level of significance (Appendix E).

Frequency distributions were plotted for the mean annual temperatures for both regions (Figure 17). Several observations are noted from this diagram and from the accompanying computed averages.

- 1) There is no overlap between the interior range in mean annual temperature of $25-36^{\circ}\text{F}$ and the coastal range of $38-45^{\circ}\text{F}$.
- 2) The interior averages vary more, with a total range of 12°F , compared with a range of only 8°F on the coast.
- 3) Mean annual coastal temperatures averaged 10.5° warmer.
- 4) The coastal temperature range is continuous and terminates sharply while the interior has outlying extremes.
- 5) The range of mean annual coastal temperatures is skewed to the right, while that for the interior has relatively few years below the modal number.

Actual mean annual temperatures for the continental interior region and for the coastal Juneau region are shown in Figure 18. Two characteristics stand out:

- 1) Sharp oscillations in mean annual temperature from one year to the next. This is more pronounced in the interior.
- 2) The manner in which year-to-year temperature oscillations on opposite sides of the Juneau Icefield are in close agreement. In other words, the temperatures on an annual basis in both regions, though quantitatively different, are quite parallel and in-phase.

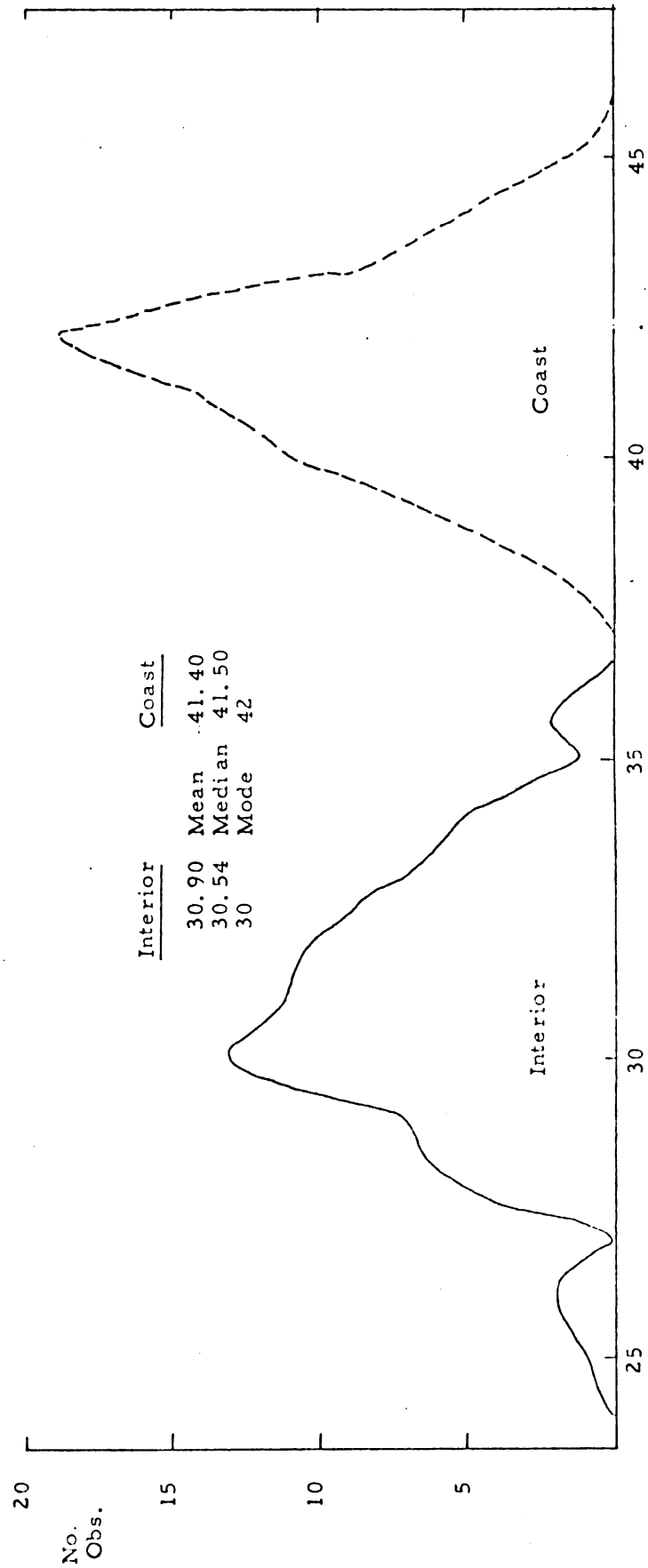


Fig. 17 Frequency Distribution of Mean Annual Temperatures, Coast and Interior, 1907-1973

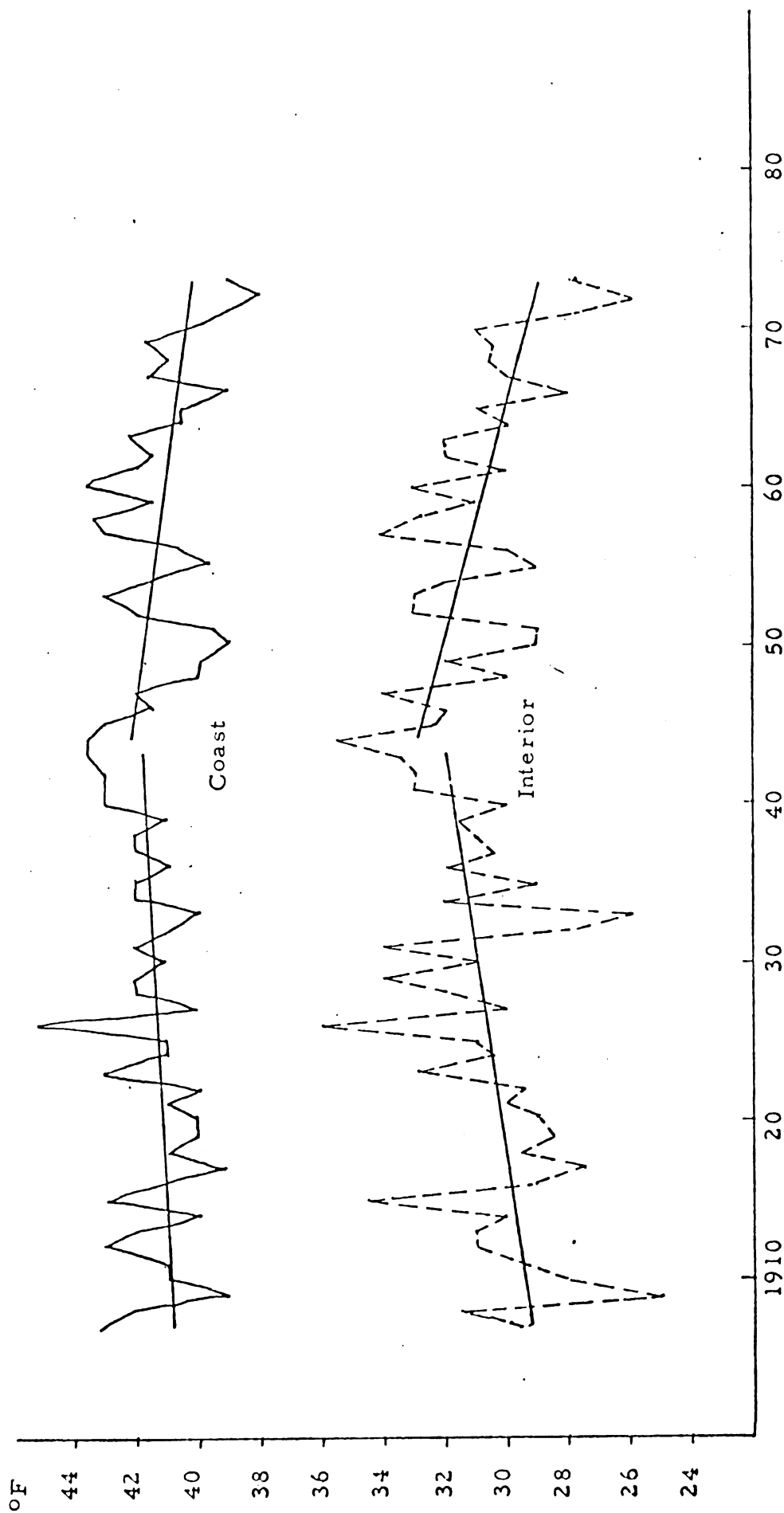


Fig. 18 Annual Mean Temperatures, Coast and Interior, 1907-1973, With Linear Regressions $y=a+bx$

Since year-to-year changes are so extreme, it is necessary to smooth the data in order to determine if trends actually exist. Three methods are used: a weighted moving average; a linear regression; and a polynomial regression.

Weighted Five-year Moving Mean . In this method a weighted average is applied , of the form:

$$\bar{x} = 0.1(x_{n-2}) + 0.2(x_{n-1}) + 0.4(x_n) + 0.2(x_{n+1}) + 0.1(x_{n+2})$$

$\bar{x}$ represents the mean, x the mean annual temperature of a given year, and n indicates the central year of the particular five years being averaged.

The purpose of the weighting was to avoid the Slutsky-Yule effect (Mosteller, et al., 1973). When an un-weighted moving average is applied to a series of random numbers, an apparent periodicity is generated. In research involving a series of temperature data, such an effect can generate a spurious apparent cyclicity or can mask a real pattern which does in fact exist. The 5-year weighted moving average is shown in Figure 19.

Inspection of the data shows a cyclical upward trend in coastal temperatures until the early 1940's, followed by a downward oscillation since that time. Periods of oscillation are roughly between 13 and 17 years in length.

Magnitudes of the oscillations in the weighted 5-year means are also affected by two factors: 1) the level of extreme temperatures and 2) the persistence of departures from the central value. While 1927 was the peak year for temperature it was between two below-normal years. Persistence of above-normal annual temperatures from 1940 through 1947 gave a very positive level in the 5-year weighted average. Similarly, temperatures consistently below normal for 1948 through

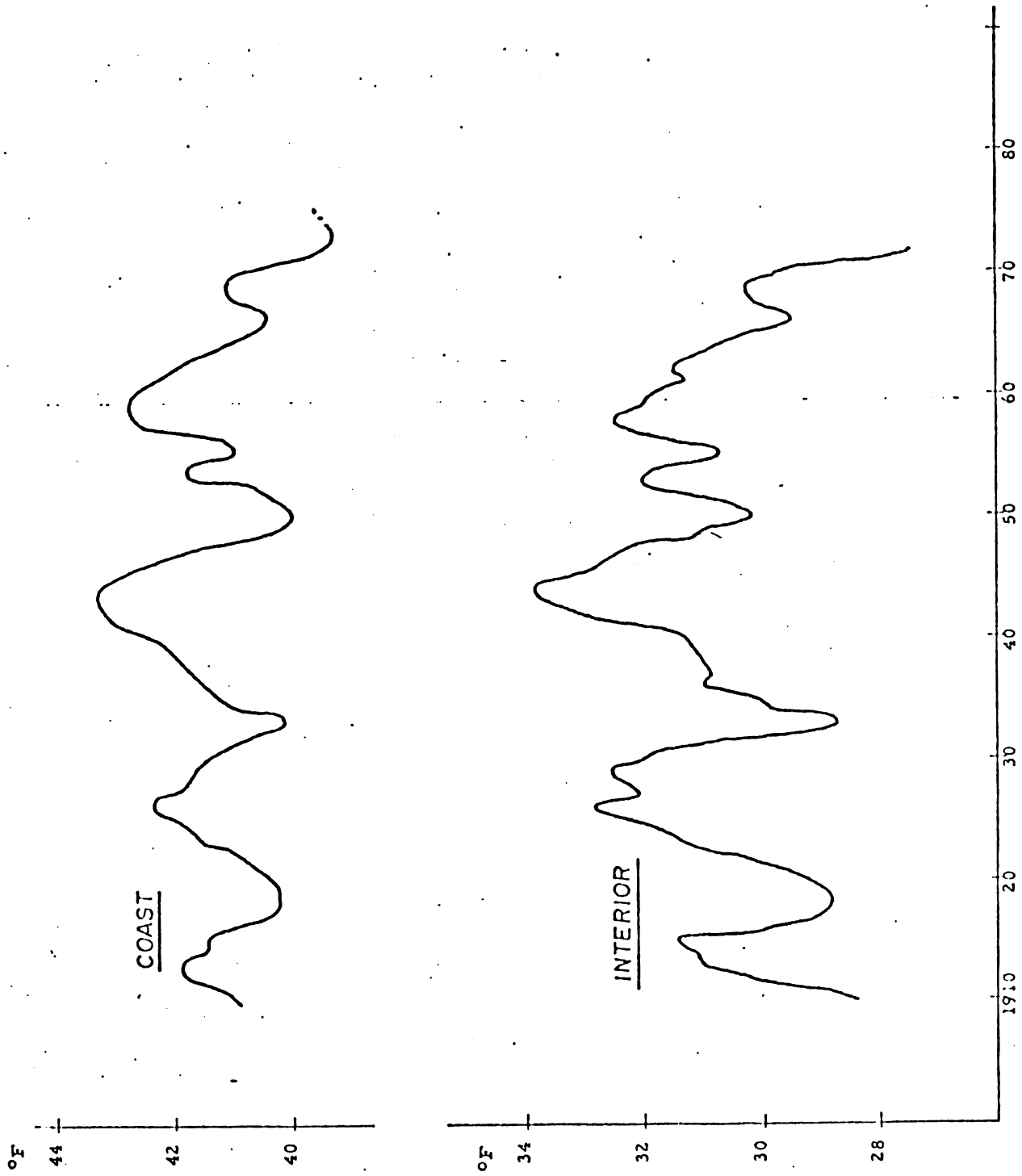


Fig. 19 Temperature, 5-year Weighted Annual Means, Coast and Interior, 1907-1973

1951 without a warm year between gave a low point in the moving average.

Linear Regression Analysis. A relatively simple method (Blaylock, 1960) of determining the rate of change in a variable over time is a simple linear regression of the form:

$$y=a+bx$$

Application of the appropriate statistical formulas provides values for a and b. a is the point at which the regression line crosses the y axis, and b represents the slope of the regression line. x is the value of the independent variable (in this case the year) for which the dependent variable (in this case the predicted temperature) is to be determined. For our purposes, x is the first year in the series and x_n is the last year. The slope of the line (b) and the value of y when $x=0$ provide information as to the magnitude of the change. However, this is applied to net change in the value of the dependent variable from the beginning to the end of the period. A dependent variable could rise to a high value in the middle of its range and then return fairly symmetrically to its original value, and a linear regression could indicate that no change at all had occurred.

Regional temperatures peaked in the early 1940's after oscillating upward since the beginning of the climatic record. After that time they have oscillated downwards. Therefore the linear regression analysis as applied separately to both coastal and interior records is broken into two parts. The first subset for each location includes the years from 1907 to 1943, and the second subset covers the period from 1944 to 1973.

A caveat is in order when applying linear regression techniques to oscillatory data such as these temperature records. The slope of the

regression line may be conditioned by the portions of local oscillations on which the regression line ends. In other words, by starting at the low point of a cycle and ending at the maximum of a later cycle one can exaggerate the slope, while starting at a high point and ending at the minimum of a later cycle will minimize the slope. Where the range of oscillation is greater than the long-term change in trend, as is the case with our data, the latter error may reverse the sign of the slope entirely.

With respect to this study results of the linear regression analysis are shown in Figure 18. Coefficients were computed by a scattergram computer program.

Polynomial Regression Analysis. Since plotting of the 5-year weighted mean temperature data showed an apparent quasi-periodicity, a polynomial regression computer program was applied to the temperature data from the coastal region, and separately to the data from the interior, for the entire period from 1907 through 1973. The predictive equation for the polynomial regression is:

$$y=a+b_1x + b_2x^2 + \dots b_nx^n$$

In this equation a is again the y intercept of the equation and b is the mean slope of that segment of the line. The unit x is the value of the independent variable, while y is the predicted value of the dependent variable (temperature) for that value of x (the year). The exponent of the value x indicates the degree of the equation in that segment (Blalock, 1972). The degree of the equation may be defined as one less than the number of times the value of the slope changes signs. For example, if the slope of the segment b_1x is positive, the slope of the segment b_2x^2 will be negative and the slope of the segment b_3x^3 (third degree) will be positive.

A polynomial regression provides the equation and hence the plot of a line whose formula contains the fewest degrees of x (in essence, changes in sign of the slope of the curve) which are necessary so that the application of an additional degree does not give a significant improvement (decrease) in the sum of the squares of the deviations of the observations from the plotted polynomial regression line.

Results of the polynomial regressions with respect to the temperature fluctuations and trends are shown in Figure 20. This analysis, too, showed a temperature maximum in the early 1940's and a very sharp drop in recent years.

The value plotted for a given year is affected by the observed values of these preceding and following the given year. The computer print-out plots show a tendency for the unsupported end values to go to extremes, therefore the values computed for the first and last three years were deleted.

Since we are dealing with two different series of oscillatory trends of opposite slopes, the polynomial regression analysis was applied separately to the two segments of the data. The results of this polynomial regression agree remarkably well with the five-year weighted moving averages for both the coast and the interior temperatures. This agreement is illustrated in Figure 21. Again unsupported extreme end values are deleted, as explained previously, though this causes a gap in that part of the polynomial regression graph representing the early 1940's.

Observations and Conclusions. The following observations and conclusions regarding temperature patterns emerge from the foregoing analysis:

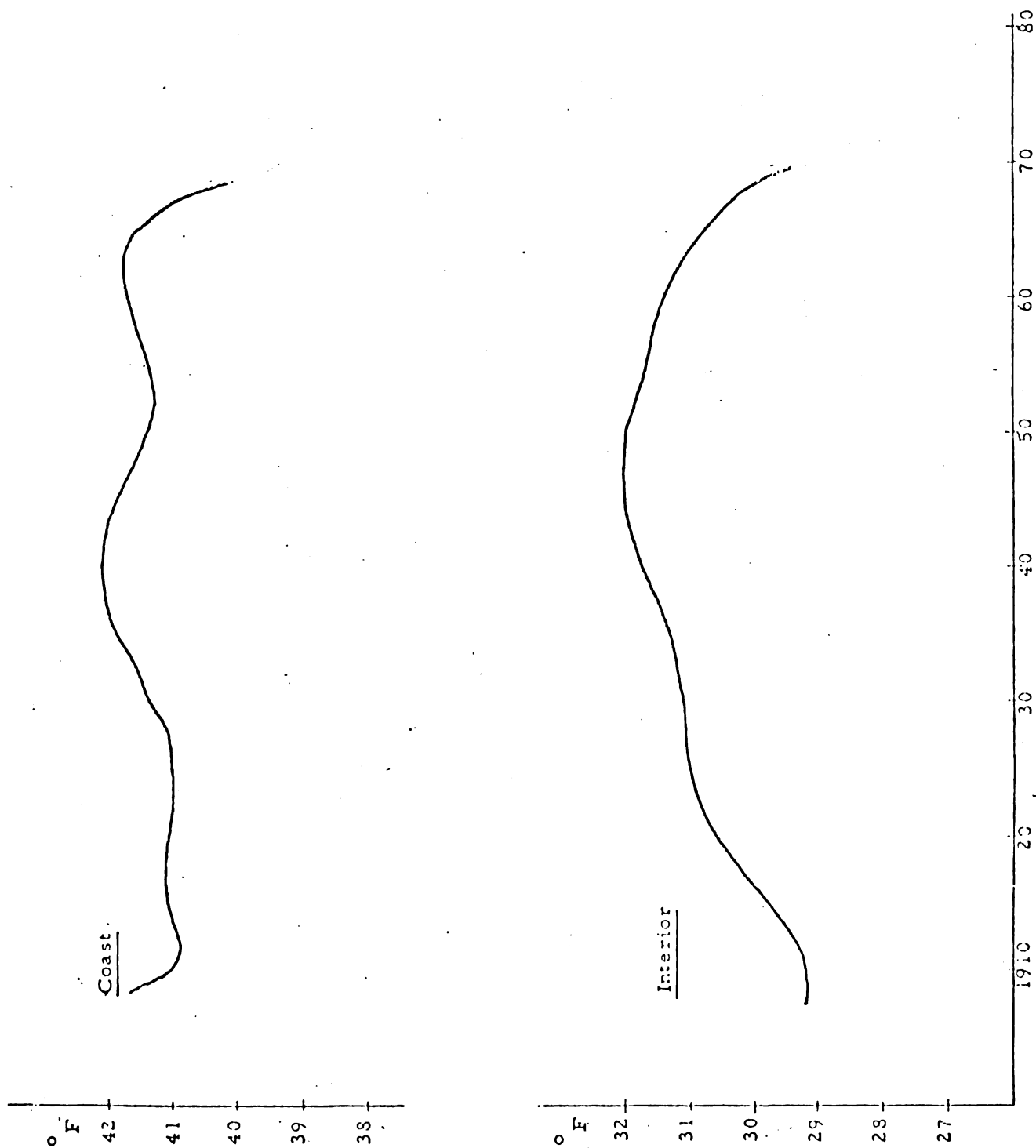


Fig.20 Polynomial Regression of Temperatures, 1907-1973

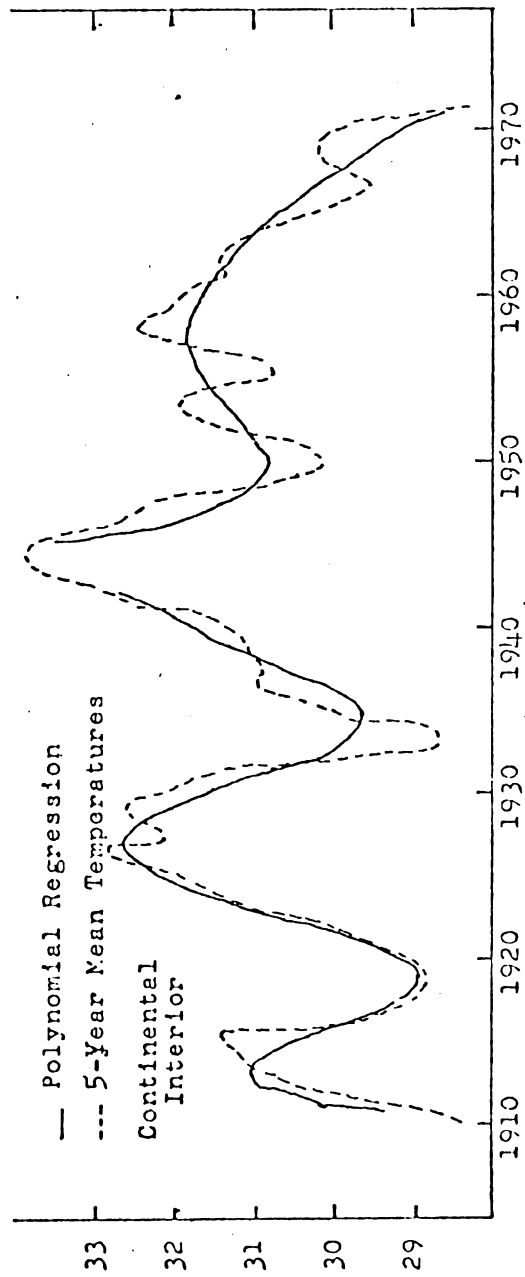
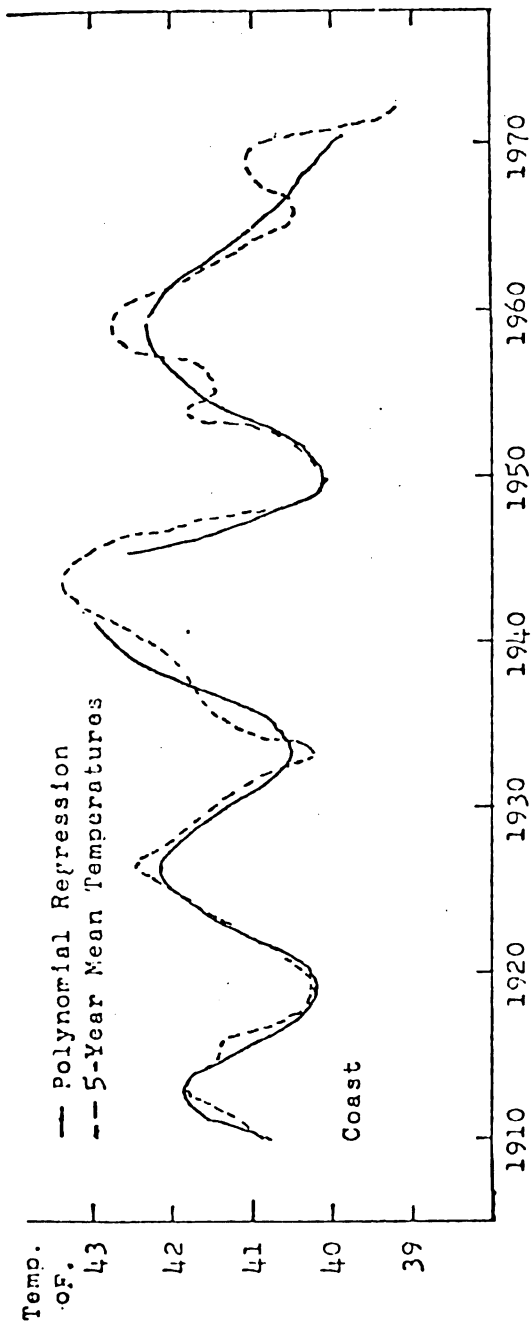


FIG. 21 Polynomial Regression (2-part) and Five-Year Weighted Moving Average of Mean Annual Temperatures, Coast and Continental Interior

1. Annual mean temperatures in the interior are consistently colder than on the coast. Interior means range from 25° to 36° F, with a mean of 30.75° F from 1907 through 1973. Mean annual temperatures in the coastal Juneau area range from 38° to 45° F, with a mean of 41.4° F.
2. Interior temperatures closely parallel those of the coastal region.
3. Visual inspection of annual mean temperatures in the 67 years of record reveals a pattern of about four quasi-cycles with a mean length of about 17 years.
4. A 5-year centrally weighted moving average reveals and corroborates the above patterns more clearly. The first three oscillations are consecutively more pronounced, with the third one peaking in approximately 1944; the fourth is of notably less magnitude.
5. The pattern of the 5-year means is more erratic after the 1950 minimum.
6. Temperatures have been in a downward skid since 1959; the cooling trend has dropped even more sharply since a minor respite in 1967-69.
7. Patterns are somewhat erratic for accurate prediction, but a subjective projection of the visible trend suggests that mean annual temperatures twenty years from now will be far below any yet experienced in this century.

Analysis of Precipitation

Annual precipitation totals for the coast and interior region are shown in Figure 22. Note that it was necessary to use an expanded scale for the interior data to allow effective visual comparison. Precipitation in this interior region averages 11.03 inches per year over the period 1907 through 1974. This is only 12.8% of the 87.1 inches yearly average at Juneau for the same period, clearly illustrating the sharp gradient of continentality across the Juneau Icefield.

The validity of using Juneau City precipitation values as representative of the coastal sector is supported by the close agreement in trends and in means between Juneau City (93.1 inches) and all south-

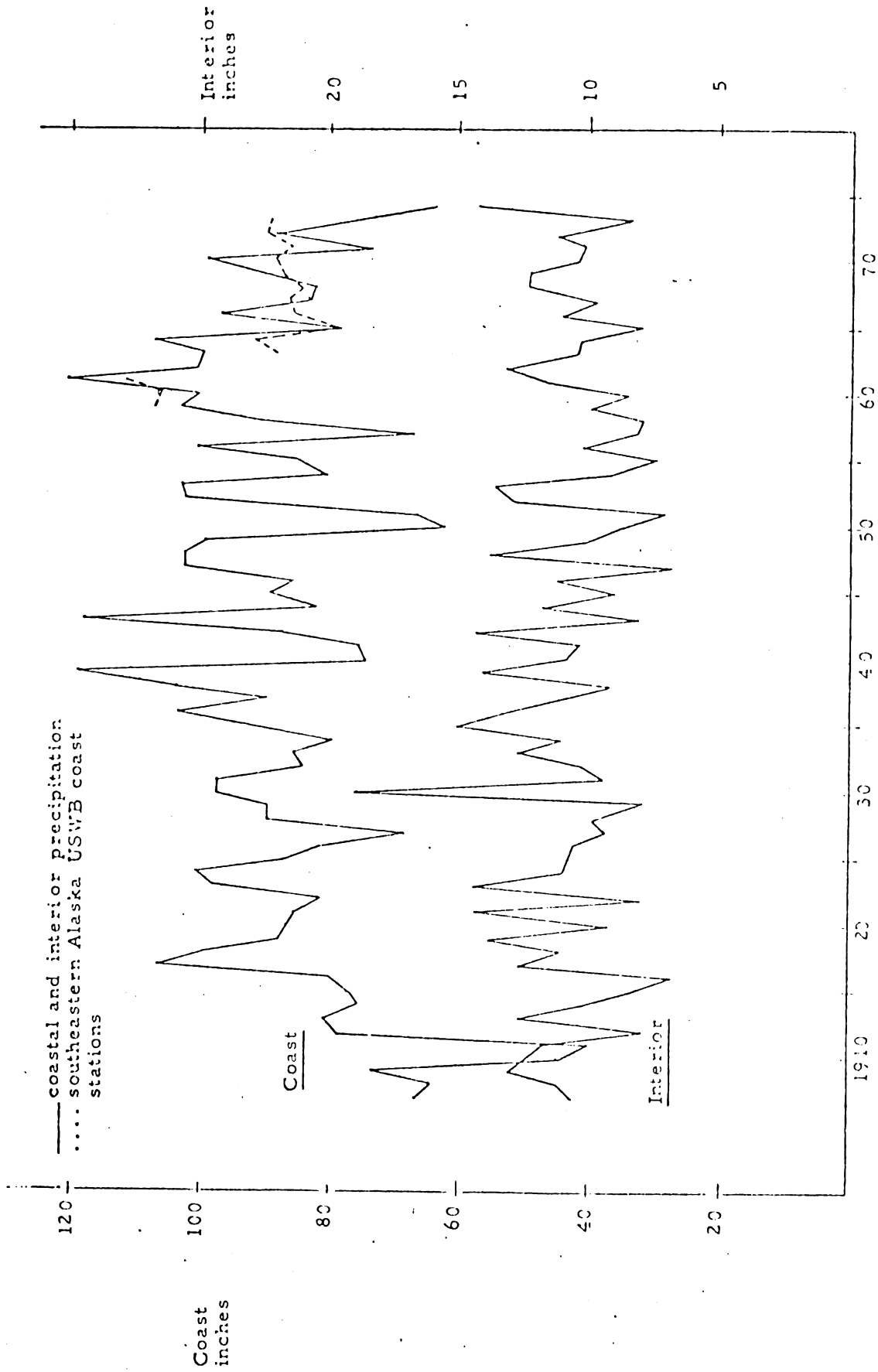


Fig. 22 Annual Precipitation, Coast and Interior, 1907 - 1973

eastern Alaska stations (92.0 inches) for the period 1959-73. In a statistical analysis of precipitation data covering 64 stations along the Alaskan coast for the period 1907-62 Miller (1963) demonstrated correlation coefficients of 0.80 to 0.94, supporting this conclusion.

In order to smooth the data and more clearly reveal any trends, five-year weighted moving means were applied to annual precipitation records by the same formula as was used with the annual mean temperatures. Analysis of Figure 23 reveals no clear-cut trends over time, with the exception of a relatively dry period along the coast around 1910. A roughly approximate seven-year periodicity appears along the coast, except for the 1910-17 period. If the 1965-72 period does in fact appear, it has a sharply restricted maximum. The existence of a seven-year pattern is much less clear in the interior. Oscillations in precipitation in the interior show some slight tendency to be in-phase with patterns along the coast during parts of the record, while other parts show no trend toward parallel variation. There appears to be no tendency toward diametrically out-of-phase variation. Parallel phasing is more pronounced in pattern though not in magnitude toward the end of the study period.

Temperature and Precipitation Relationships

The work of Anderson and Miller (Anderson, 1970 and Miller and Anderson, 1974) has shown some relationship between temperatures and precipitation over the time of the Holocene, a time scale on the order of 10^4 years (Figure 10). In the Taku District near the coast, warmer temperatures have been associated with drier periods. In the interior Atlin District warmer temperatures were associated with wetter conditions and increased storminess. The latter condition would suggest an eastward shift of the tropospheric ridge, with inland penetration of dominating low-pressure cyclonic conditions. From a

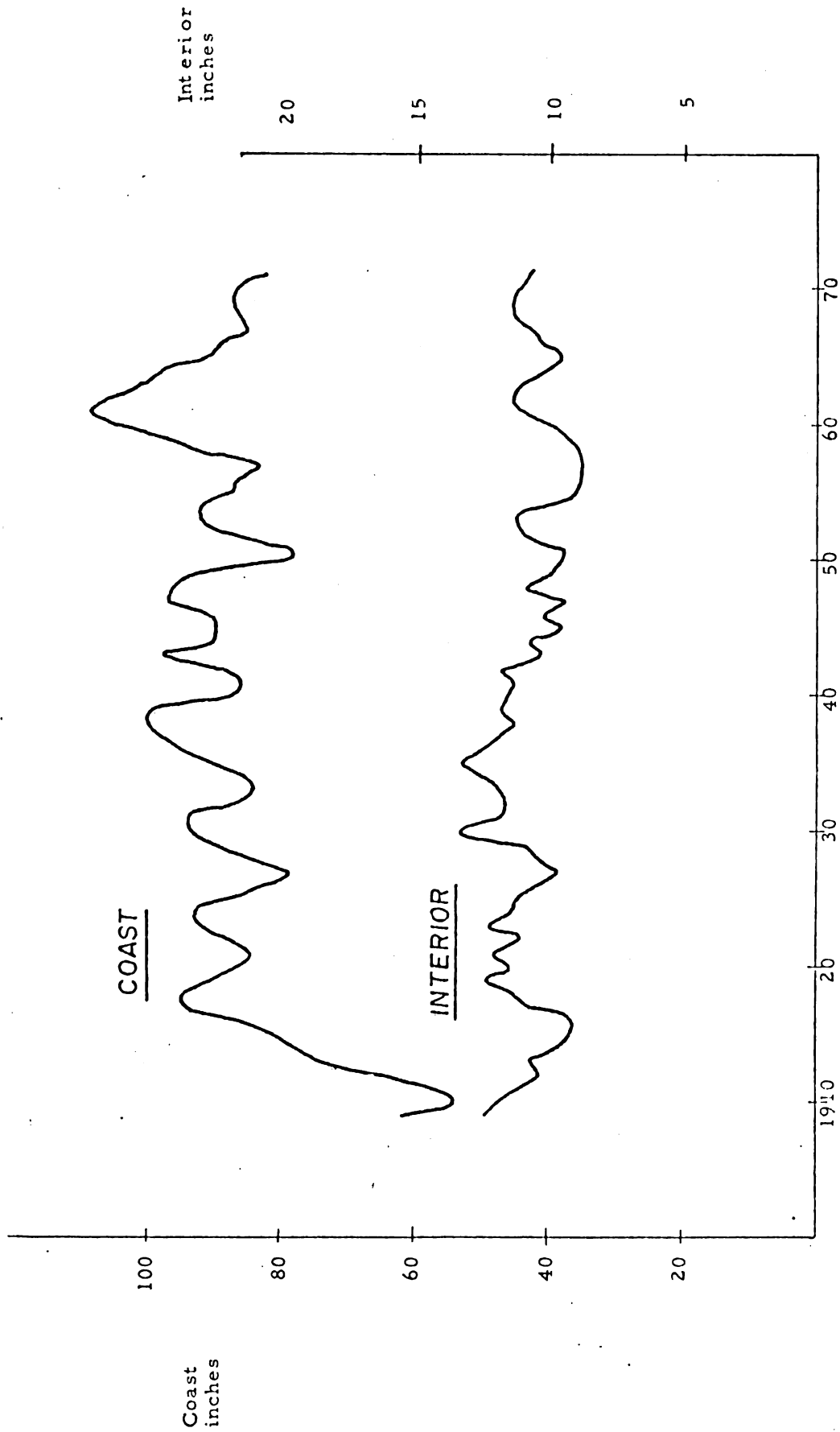


Fig. 23 Annual Precipitation, 5-year Weighted Means, Coast and Interior, 1907-1973

study of the Wisconsin moraine sequence in an interlobate sector of the Atlin region over a longer time span, Tallman (1975) has given some evidences that this shift has indeed taken place during the larger climatic changes of the Pleistocene.

Five-year weighted moving means of temperature and of precipitation were plotted for the coastal region and for the interior (Figures 24 and 25). It may be significant to note that as was the case with temperatures, existing patterns broke down after the peaking of the temperature trends in the early 1940's. Prior to the 1940's temperature and precipitation had a strong tendency to be diametrically out-of-phase. Along the coast since that time there has been a tendency for the patterns to be in-phase.

Figure 26 shows the pattern of snowfall accumulation at the Juneau Airport for 1943 through 1974. There is revealed a seven-year periodicity and a general upward trend over time. It is interesting to note that total precipitation displayed a seven-year periodicity from 1917 to about 1941, when the pattern broke down. A short six-year pattern reappeared briefly one and a half cycles from 1947 to 1957, after which the pattern again broke down. When the five-year snowfall mean is compared with five-year temperature means little relationship appears. Similarly, 5-year average snowfall bears little relationship to total precipitation trends which lag one to two years behind the temperature trend.

In addition to the change in patterns of temperature oscillation and the phase reversal of the temperature-precipitation relationship after the early 1940's, the most recent behavior of temperatures emphasizes the anomaly. An upswing in the 13 to 17 year temperature oscillation, which past patterns suggest as being at a maximum about 1975, has

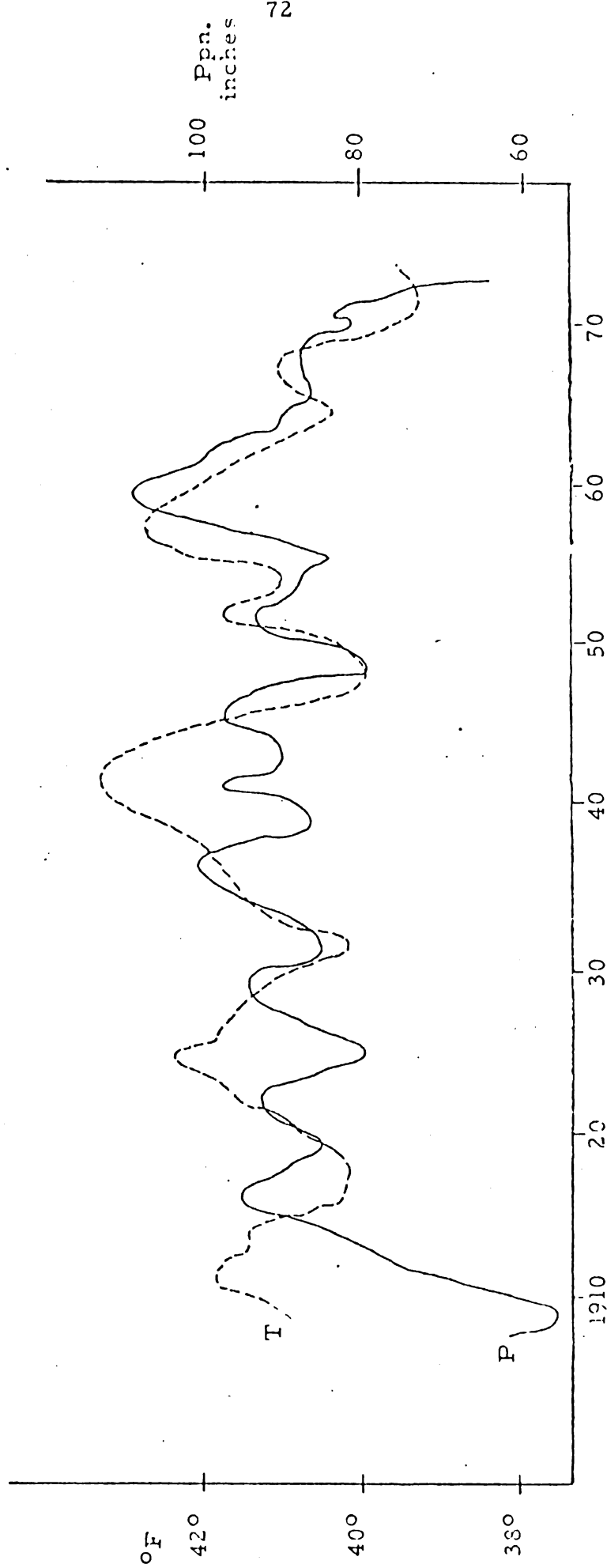


Fig. 24 Annual Temperature and Precipitation, Coast, Five-Year Weighted Means, 1907 - 1973

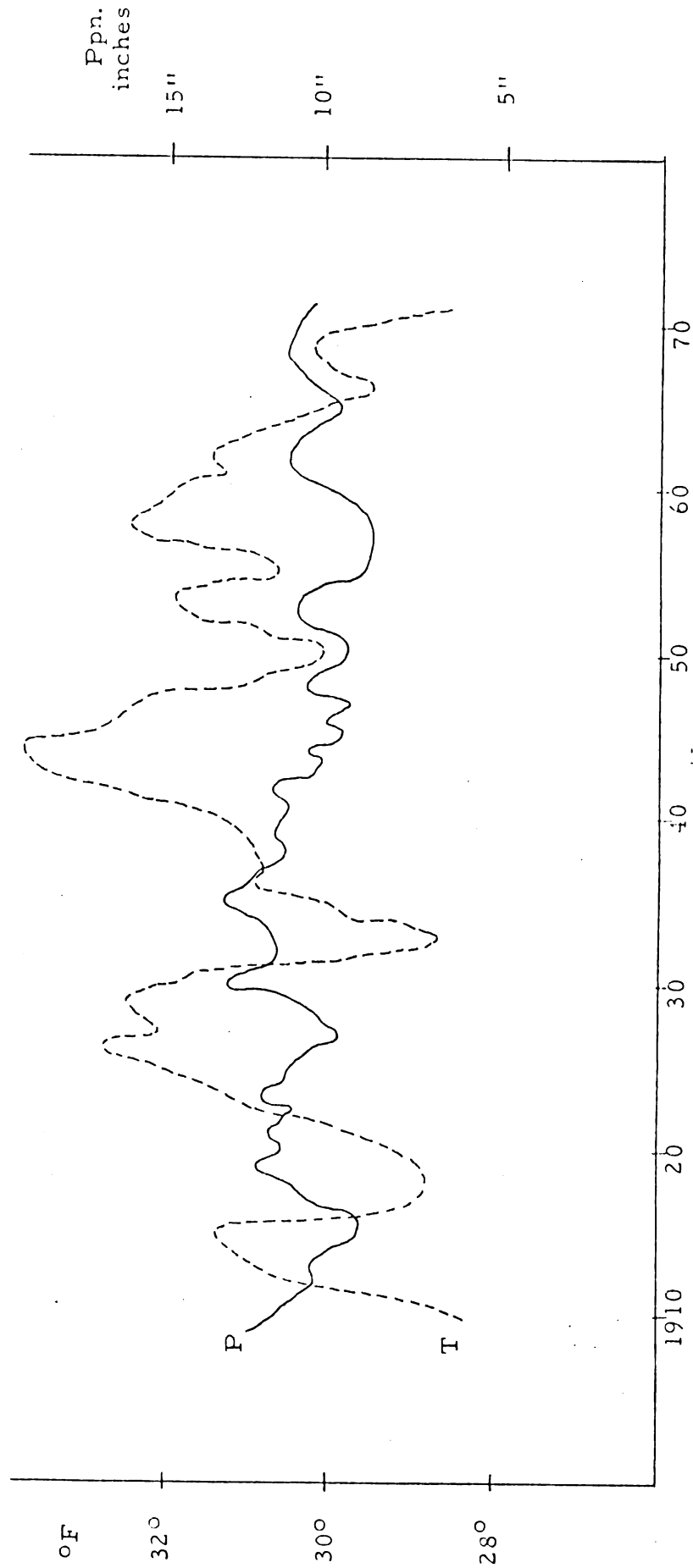


Fig. 25 Annual Temperature and Precipitation, Interior, Five-Year Weighted Means, 1907 - 1973

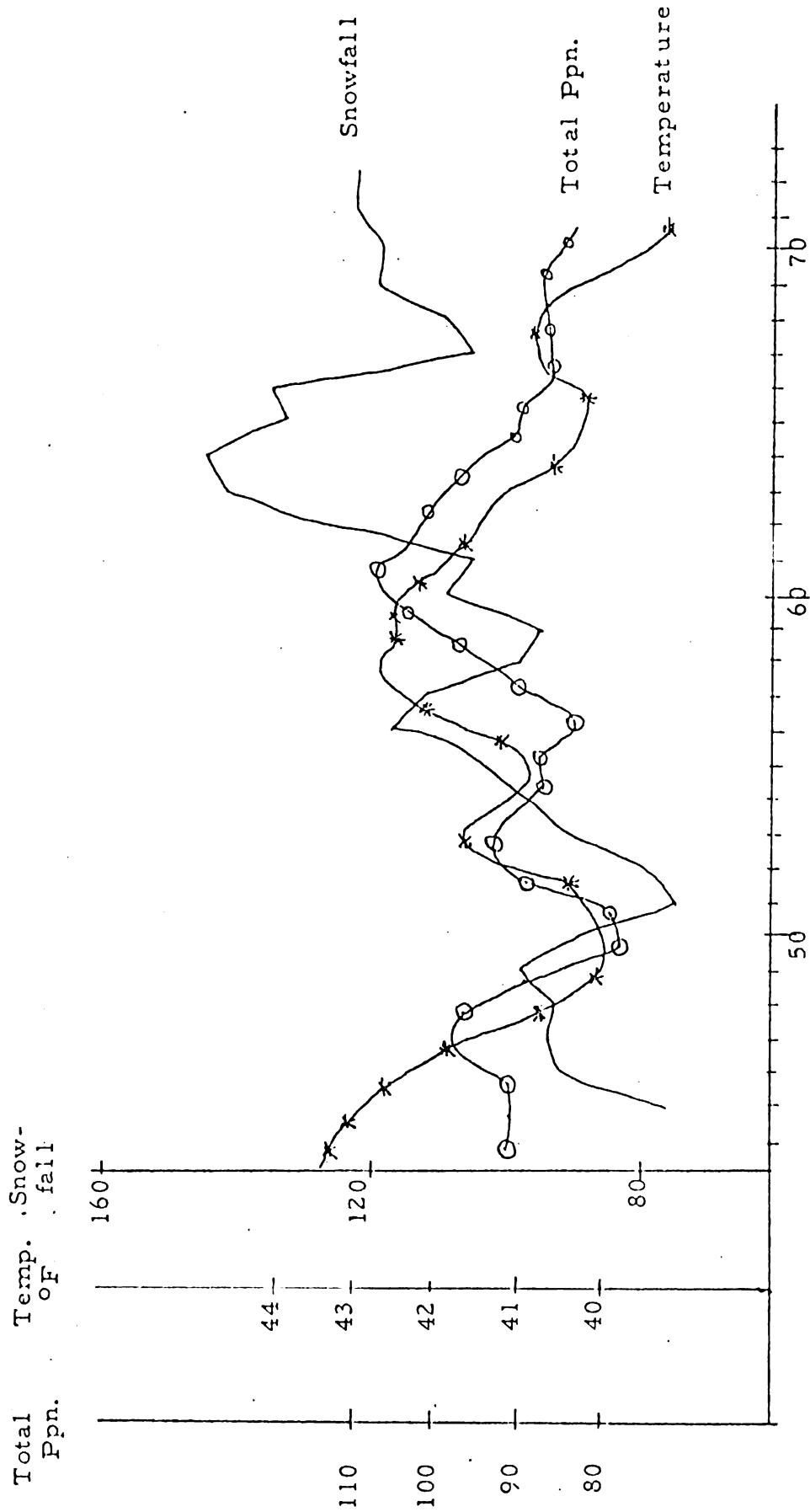


Fig. 26 Five-year Means of Total Snowfall, Juneau Airport, and Total Precipitation and Temperature Coastal Region, 1943-1974

either not clearly begun or else appeared only faintly about 1969 followed by a sharp drop in temperature. Annual means of temperatures over the next several years will be critical in suggesting what paths, anomalous or otherwise, temperatures may be expected to take and also the magnitude of temperature change.

Extreme climatic changes have been painfully obvious on a global scale in the past seven to ten years. Many causes have been suggested for the ominous changes in world climate. Among these are particulate and chemical air pollution, increase in carbon dioxide, atomic bombs, over-grazing, forest-clearing fires, and most recently aerosol cans, with the chlorine in their Freon propellant reacting with and destroying the protective shield of ozone in the very high atmosphere. It may be that man has already set in motion, beyond recall, the forces for his own doom. It may be, on the other hand, that his puny pollutants may have no real effect in comparison with natural processes on a global atmospheric scale.

It may be that dominance has shifted from one atmospheric control to another. Finally, it may be that the atmosphere is merely returning to more normal conditions after a 40-year period of abnormally favorable weather, so that we need to look for a cause for the original "anomaly" rather than for a cause for its end.

Solar Cycles, Temperatures, and Glacier Regimes

Many investigators have noted relationships between weather and the occurrence and periodicity of solar disturbances, i. e. electromagnetic (wave) radiation and charged particle (corpuscular) radiation revealed by sunspot variations (Miller, 1956, 1972b, Willett, 1975 and other publications). To explore this relationship the five-year weighted moving mean of annual average temperatures for the coastal region was

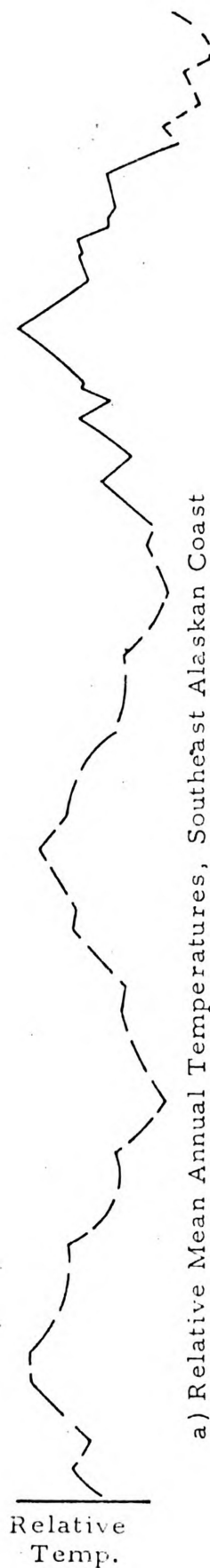
plotted against the Zurich average annual sunspot number (R_z), as shown in Figure 27.

The lack of a discernable relationship between the temperature and sunspot data was supported by a statistical correlation of 0.024. Application of temperature lag times to account for the oceanic heat sink effect provided an even poorer fit. Similarly comparison of short-term sunspot numbers with precipitation patterns, both on the coast and in the interior, showed no closer relationship. It is concluded that any existing short-term relationships between temperatures and the 10.8 year solar cycle are considered to be more complex, requiring a more sophisticated analysis.

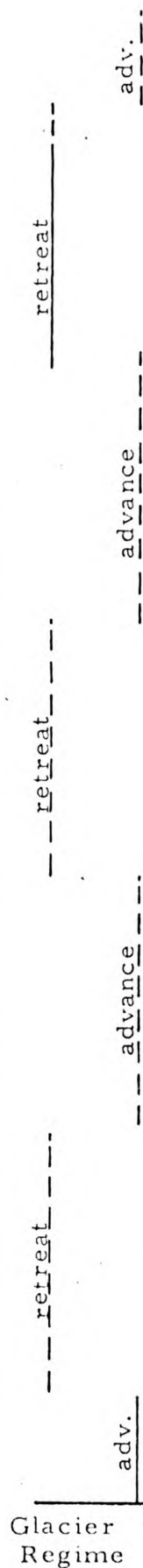
Because of the above, another approach was taken. For example, the lichenometric date of 1760 for the major Little Ice Age advance was used as a benchmark for inferring postulated glacial regimes, along with the known recent behavior of other glaciers on the Juneau Icefield. Also known and inferred temperature patterns in the Juneau Icefield region were plotted. Finally, for comparison, the Zurich sunspot numbers from 1749 through 1974 (Appendix C) were plotted on the same time scale. The results are shown in Figure 27. The following relationships are apparent:

- 1) Periodicity of glacier behavior is very close to 90 years.
- 2) Glacier terminal behavior lags behind temperature patterns by about a half cycle (45 years). The glacier advances during a period of rising temperatures and retreats during the cooling portion of the apparent 90-year cycle.
- 3) Short-term temperature changes do not appear to be related to short-term (10.8 year) solar cycles. The inertia and lack of resolution of finer climatic perturbations in the system may be too great to allow detection of changes at this scale.

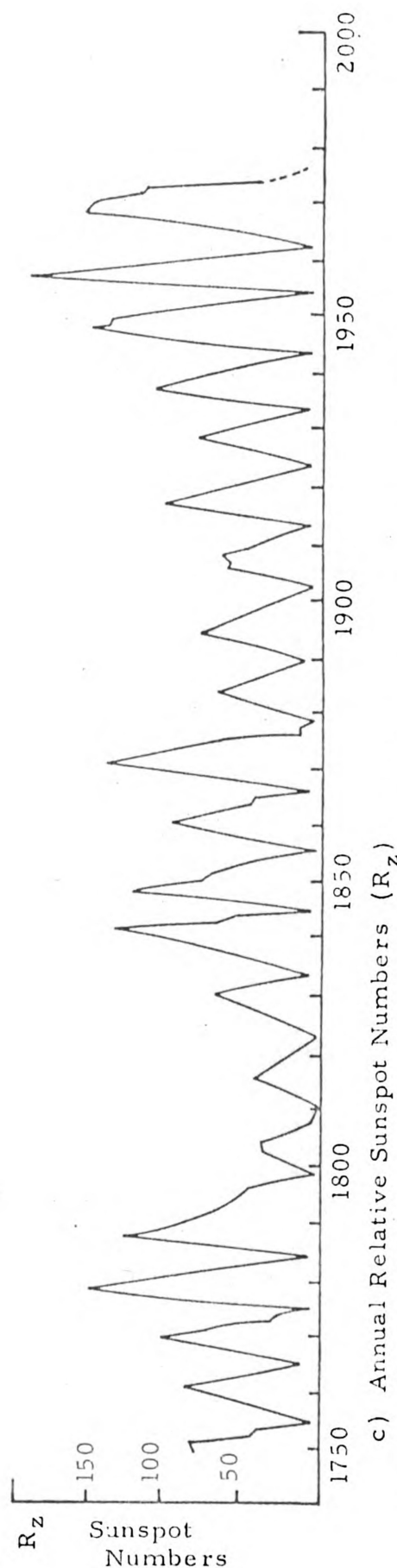
Known ——— Inferred - - - - -



a) Relative Mean Annual Temperatures, Southeast Alaskan Coast



b) Regime of Cathedral Glacier System



c) Annual Relative Sunspot Numbers (R_z)

Fig. 27 Sunspots, Temperatures, and Cathedral Glacier Advance and Retreat, 1750 - 2000

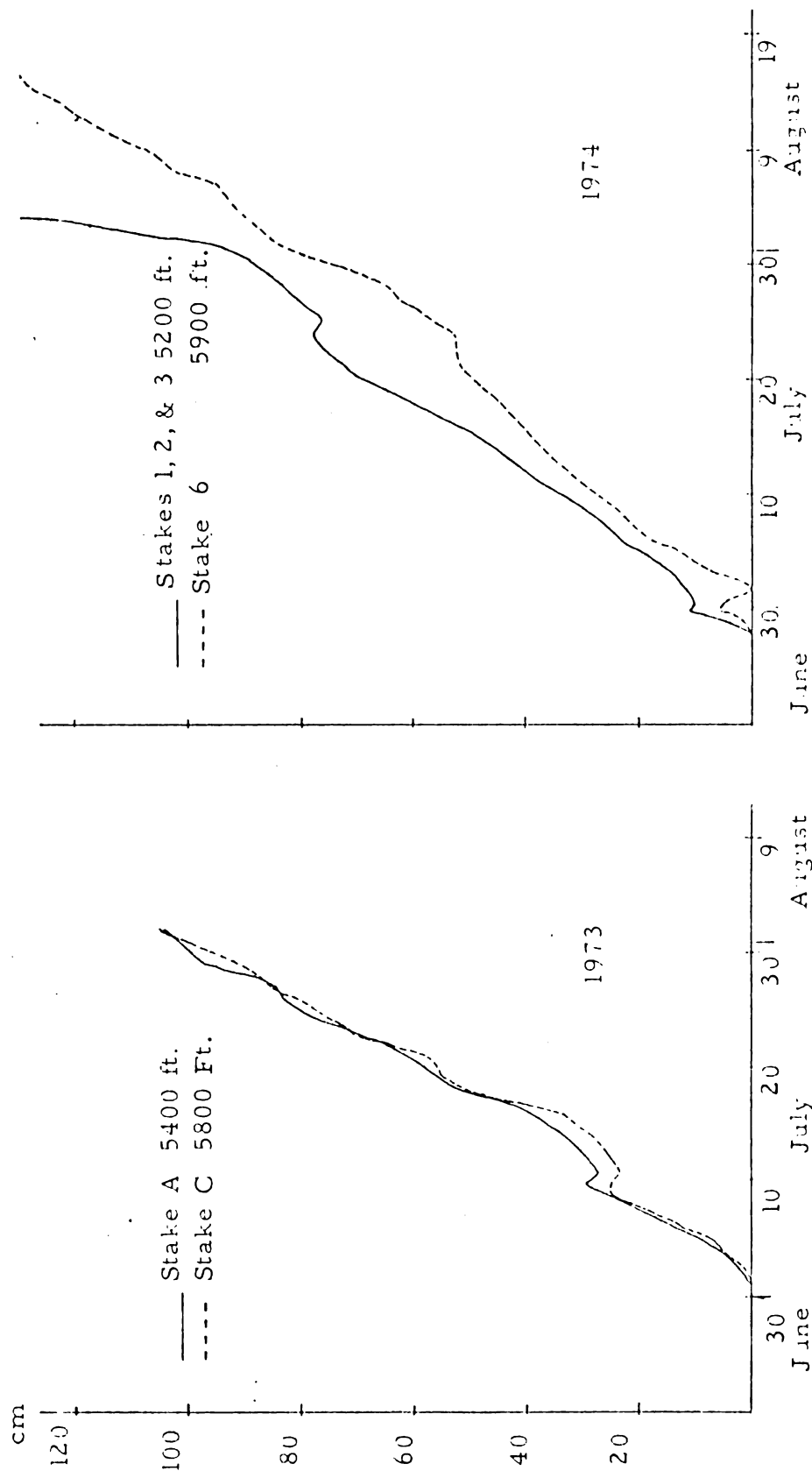


Fig. 28 Cumulative Ablation on Cathedral Glacier. 1973 and 1974

- 4) For the segment of time involved in the Little Ice Age, glacier advances appear to have occurred during and after the minimum portion of the 80-90 year sunspot cycle. Retreats tend to occur during and after the maxima of these "long" cycles.
- 5) Temperature peaks tend to lead the 80-90 year solar peaks by about 15 years. This brings up a tantalizing question concerning what has been previously considered by Miller (1972b) to be a 15-year cooling lag attributed to the heat sink effect of the North Pacific Ocean.

If indeed there is a close solar-temperature relationship, over the time scale of this study there may be a slight drift in the relationships of temperature and glacier regime in comparison to patterns of solar behavior. In the absence of known causal factors, it is suggested that we may be witnessing the in-phase coincidence of two unrelated cycles of slightly differing lengths. Thus on a scale not of the three "long" cycles plotted in Figure 27 but of ten, twenty, or fifty, the relative patterns could be completely reversed.

Relationship to the General Atmospheric Circulation

At this point a highly hypothetical causal mechanism is suggested which is counter to implications of the idea in the preceding paragraph. Solar activity affects the emission of energy from the sun partly in the form of particulate matter (corpuscular radiation). These charged particles are to some extent channelled to the magnetic field, as has been considered in previous JIRP publications (e.g. Miller, 1963, 1972b). The resulting energy-input causes fluctuations of the wave-length of the circumpolar westerlies, changing the location of the tropospheric ridge along the Coastal mountains of Alaska-Canada. The relative locations of the mean low pressure systems in the Gulf of Alaska (Aleutian Low) and areas of anticyclogenesis in the interior (Canadian Polar High) would thus be a reflection of ridge location. In turn

temperature and accumulation changes in the Boundary Range would be affected. Such changes in temperature and precipitation regimes are integrated in the behavior of glaciers and ultimately in the land-forms they produce. But some inconsistencies remain in this mechanism. For example, we might expect increased energy to increase the activity of the westerlies, where in reality, the increased turbulence and meridionality are due to increased thermal contrast which means colder polar areas, with less energy.

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Summary of Temperature and Precipitation Data

	Coast	Interior
Range of Mean Annual Temperatures	38-45°F	25-36°F
Amount of Range in Mean Annual Temp.	8°F	12°F
1907-73 Average of Mean Annual Temp.	41.4°F	30.75°F
1907-73 Average of Annual Precipitation	87.1 in	11.03 in

The wider range in mean temperatures in the interior reflects the continentality gradient. While mean annual temperatures fluctuate sharply from year to year, the fluctuations are usually parallel between the coast and the interior. In contrast, rainfall patterns between the two regions are much less in-phase, and the extent of phasing varies considerably over time.

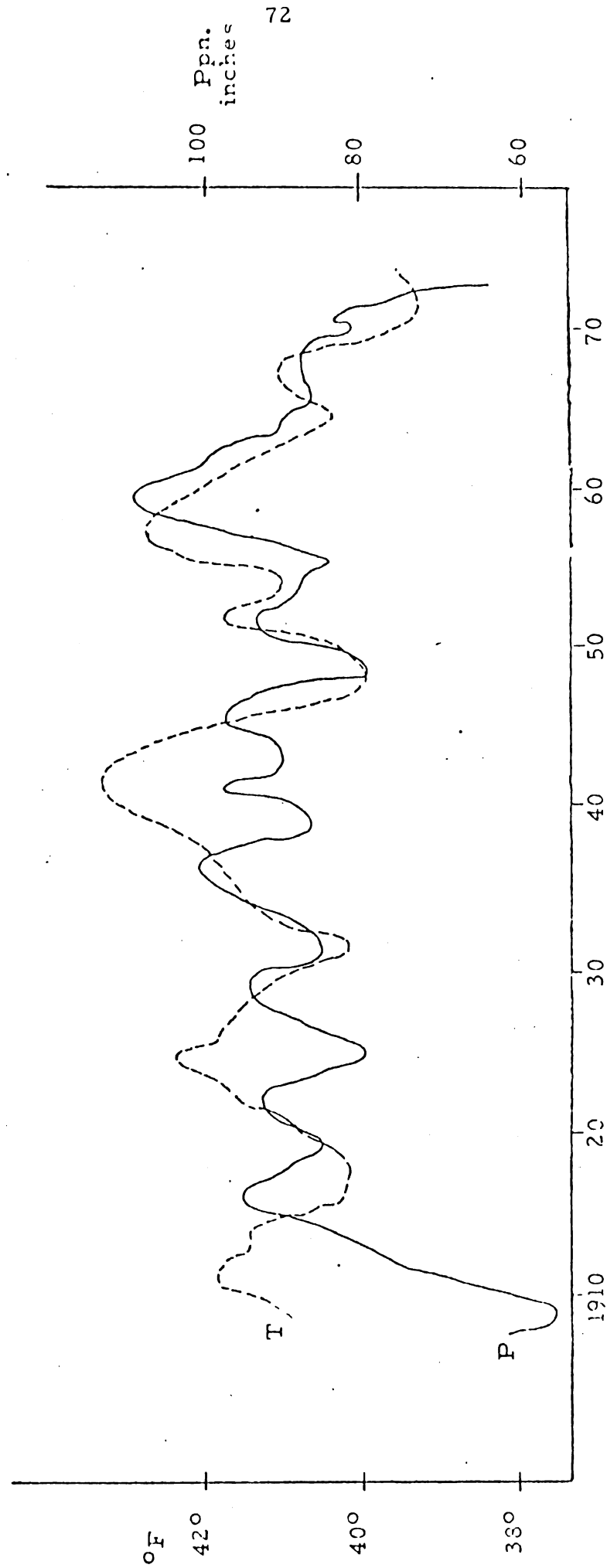


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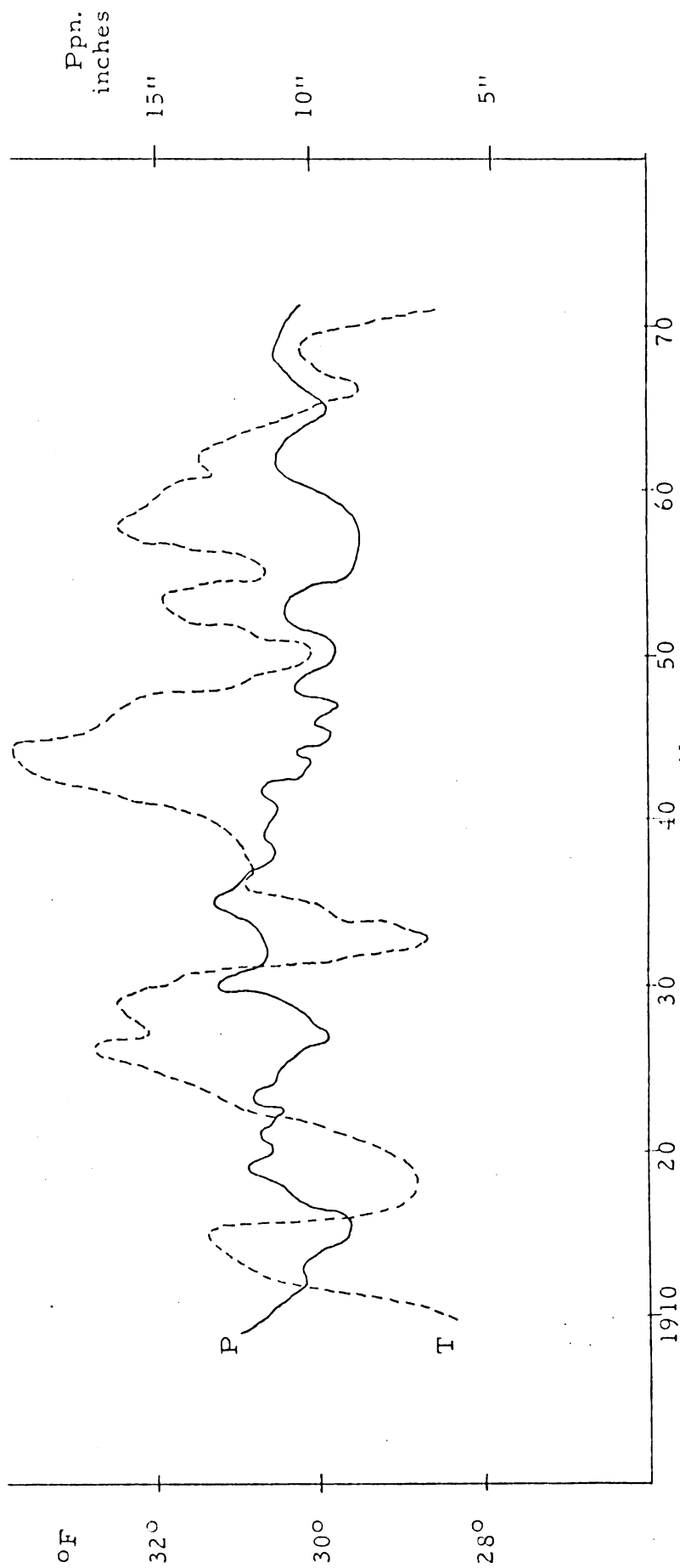


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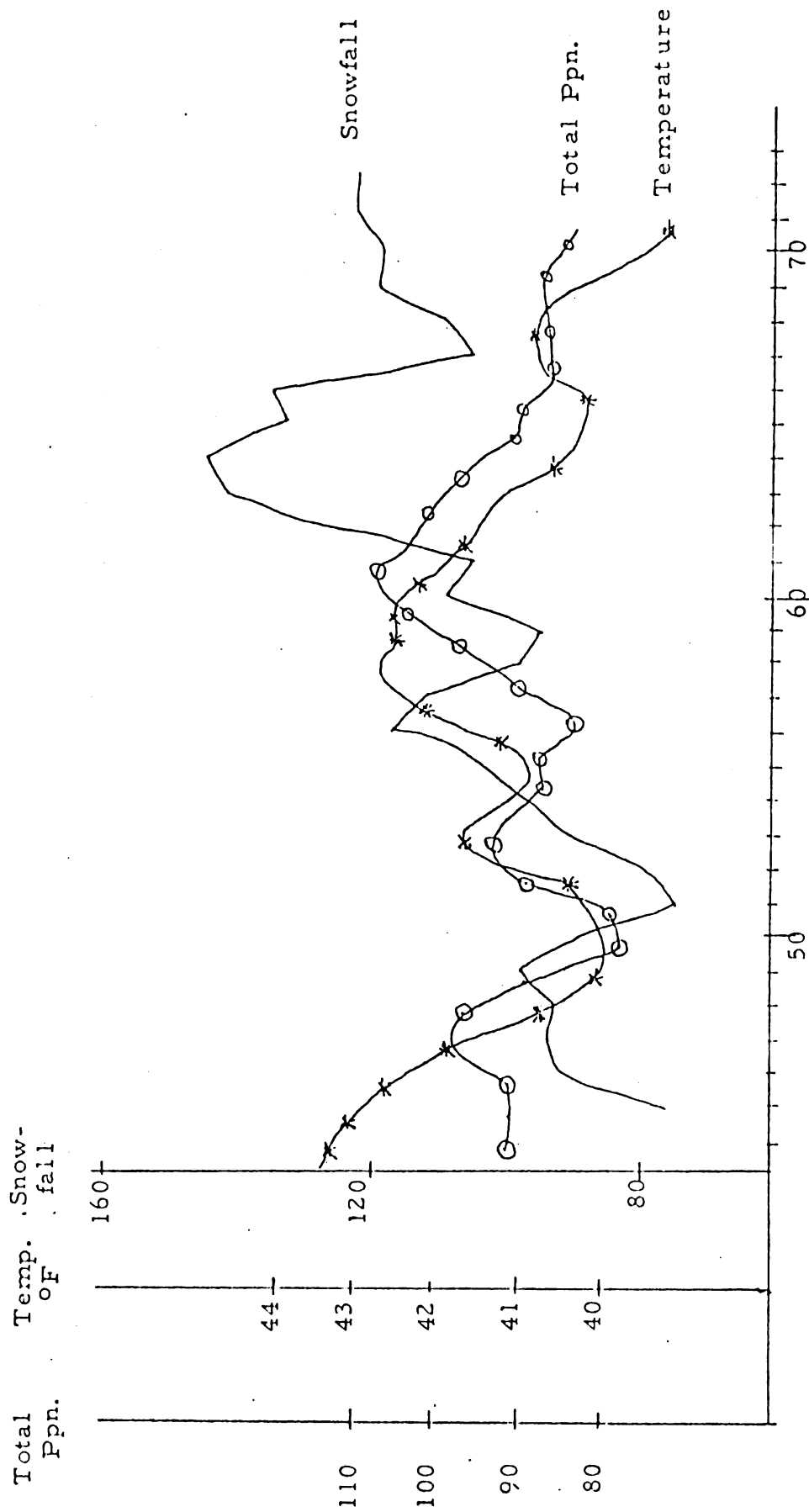


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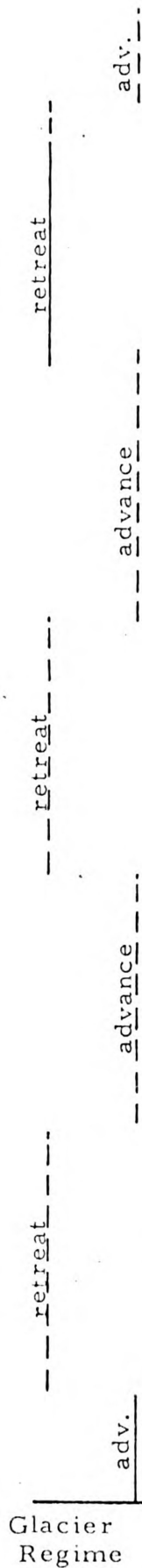
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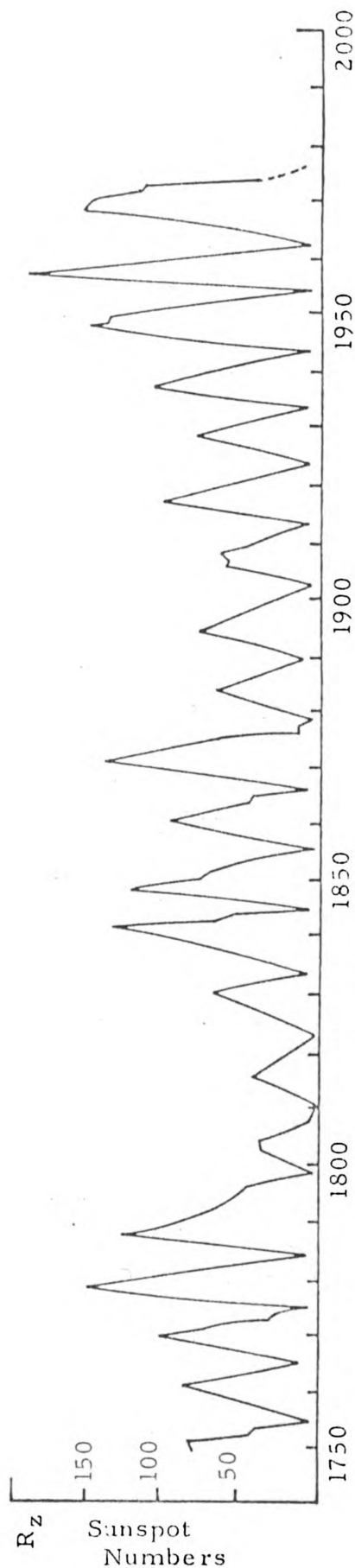
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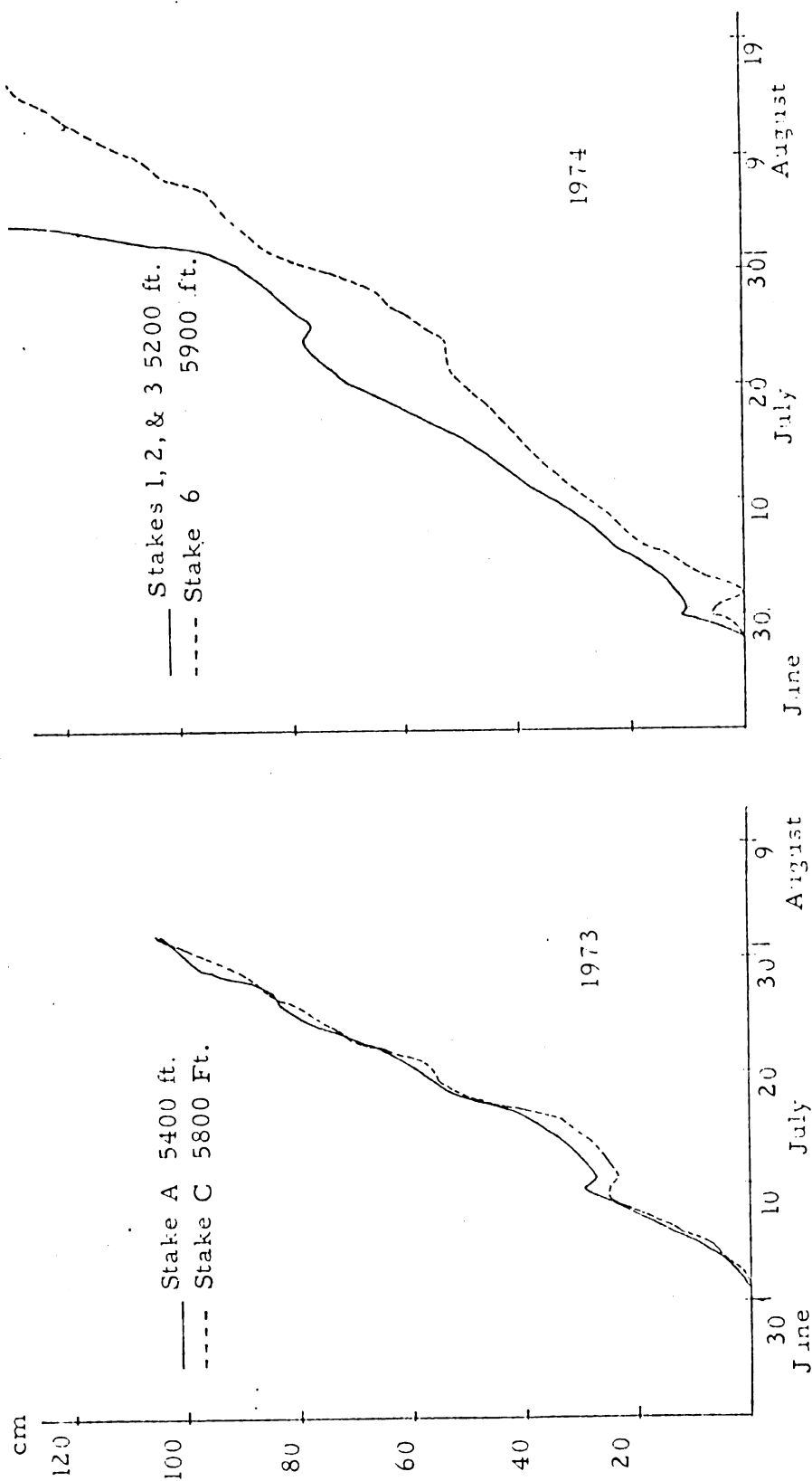


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The analytic techniques applied in this study, including graphic analysis and linear and polynomial regressions, revealed an approximate 17 year periodicity in temperatures and a less distinct 7 year pattern in precipitation. On a longer-term bases, temperatures have generally moved upward from the start of the record (1907) until the early 1940's. Since then they have moved downward, with the rate of decrease accelerating over the past decade.

The most significant observation to be developed from this analysis is that the nature of atmospheric relationships appears to have changed with or shortly following the peaking of mean annual temperature trends in the early 1940's. As patterns of temperature oscillations changed, the phase relationships of temperature and precipitation appear to have nearly reversed. Also, temperatures in the past five to ten years have dropped much more sharply than earlier patterns would suggest, and the tentative outlook is for this trend to continue despite short term upswings. The question arises as to whether this sharp temperature drop since the early 1960's is a natural climatic trend or an anomaly. Field data from the next five years should give a clearer picture of the short term situation, and hopefully provide a clearer idea of what trends can reasonably be expected. A more detailed analysis and correlations of the changes in patterns, related to possible causative factors, must be left to later studies.

CHAPTER VI

FIRN-PACK REGIMEN, ABLATION, AND GLACIO-HYDROLOGY

This chapter deals with the different phases of the annual snow regimen, from accumulation through ablation and run-off. It considers April 1st snow depths at Atlin and Log Cabin, B. C., the nearest available locations of record. Ablation measurements on the Cathedral Glacier surface are discussed, and the elevations of the seasonal névé-lines since 1971 are shown. The nature of the local firn-pack is discussed, including test-pit data, free water content, and diurnal patterns of drainage. Finally, the results of this ablation and drainage are considered, as the total hydrologic run-off is confined to Cathedral Creek and infiltrating groundwater.

Regional Snow-Pack Records

Figure 28 shows snow depths, in inches of water equivalent, on April 1st at official snow courses at Atlin (since 1964) and at Log Cabin (since 1960) monitored by the Water Investigations Branch of the Provincial Government of B. C. (Hydrology division, 1971, and periodic reports). Log Cabin is located at 2880 feet (880 m) elevation on the high plateau between White Pass and Carcross on the Yukon and White Pass Railroad, at the north end of the Boundary Range. Not only is Log Cabin at almost 500 feet (150 m) higher elevation than the Atlin snow course, but it is in the path of maritime air masses which pass north-northeastward up Lynn Canal, over Skagway and on up the valley through White Pass.

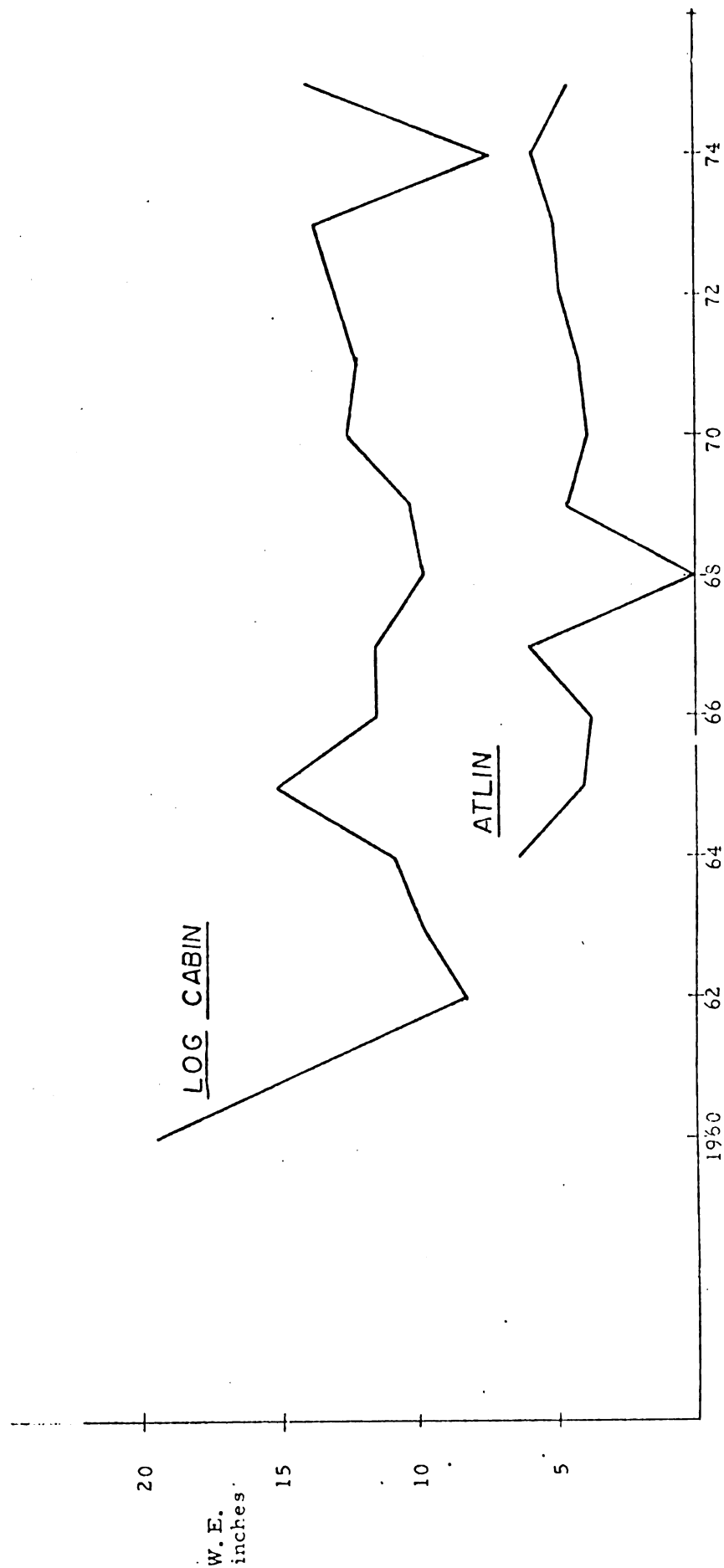


Fig.29 April 1st Snow depths Atlin and Log Cabin; Water Equiv. inches

Though its location is somewhat lower than the Cathedral Glacier, it is observed that a very shallow snow-pack at Log Cabin in the spring of 1974 was followed by an unusually effective ablation season and high névé'-line on the Cathedral massif where ice exposure was much greater than in 1973. The April 1 snow-pack depth at Atlin, at lower elevation, was well above average in 1974. These comparative records should be followed in future years, as another means of monitoring snow regime changes in the region.

Ablation Records and Névé'-line Positions

Cumulative rates of ablation of the winter snow-pack on the Cathedral Glacier are shown in Figure 29. Ablation measurements were taken in the late afternoon or early evening from metal stakes emplaced vertically in the snow. Measurements were taken at a cross-bar placed momentarily on the surface to avoid melting effects (cryoconite depressions) which develop immediately adjacent to ablation stakes.

Stake locations were not precisely the same in 1973 as in 1974. The 1973 locations eventually became exposed below the transient snow-line, with bare ice showing by the beginning of August. Despite the differences in stake position, locations for the two years were not so far apart that the rates of firn ablation were not comparable.

In 1974, a line of three stakes was placed across the glacier between the bedrock berm extending from Met Hill and the similar berm extending from Mt. Edward Little. The Mt. Edward Little berm had not been exposed at all during 1973, and exposure of the Met Hill berm was much greater during 1974 than the previous year. The location of this transect was across the narrowest part of the glacier, and was

approximately at the top of what has already been described as the buried headwall of a relict cirque. These stakes were also used for the annual surface movement survey. One other 1974 stake was placed higher in the upper cirque than in 1973, in an area expected to have positive mass balance.

The measurement stakes melted out at 104 cm on August 1st in 1973. In 1974, the final stake in the lower group melted out at 129 cm of ablation by August 15. As expected, the rate of ablation was slower at the higher stake in 1974. The last 40 cm of snow at the lower sites melted at an accelerating rate; this may have been due to a heavy infiltration of melt-water from above. The glacier surface at this point slopes steeply to the north, so that large quantities of melt-water from above saturated the firn-pack and later broke out as straight supra-glacial streams. Within a few weeks these streams had developed sine-generated forms indicating the rapidity at which the self-regulating glacio-fluvial processes strive for equilibrium on this smooth glacier surface.

The ablation records cited above typify the negative mass balance statistics obtained on the glacier in each of these last three years. This is further corroborated by the following record of relative seasonal and semi-permanent névé'-lines, based on C-29 station elevation of 5300 feet (1620 m).

Cathedral Glacier

	<u>Seasonal Névé'-line</u>	<u>Semi-permanent Névé'-line</u>
1971	5750 ft (1750 m)	5750 ft (1750 m)
1972	5800 ft (1760 m)	5700 ft (1735 m)
1973	5600 ft (1700 m)	5600 ft (1700 m)
1974	5700 ft (1735 m)	5700 ft (1735 m)

Névé'-lines for two dates near the end of the 1974 ablation season are shown in Figures 30 and 31.

The Local Firn-Pack

Englacial temperature measurements to a depth of 18 meters in 1972 indicated that this glacier is thermophysically sub-Polar in its upper head-wall portion i. e. above 6000 feet (1820 m). Below 5700 feet (1730 m) elevation however, it is essentially Temperate ($0^{\circ}\text{C}.$) Thus surface melt-water freely percolates into the firn and flows off the glacier below the equilibrium line (essentially the névé'-line in this case).

Test-pit measurements at the same location as the FWC measurements (5850 ft, 1770 m) showed that the firn changed quite suddenly to bubbly ice at a depth of 130 centimeters (Figure 32). As the test-pit elevation was close to the mean elevation of the 1972 névé'-line the firn-pack depth is representative. This not only gives the annual net gain of the firn-pack (as September 7th was close to the end of the ablation season) but it also suggests that in this glacier the firn changes to glacier ice at a relatively shallow depth. It further implies that this transformation may take place through the re-freezing of melt-water into new ice (superimposed ice) above the previous glacier ice. This observation appears to corroborate the sub-Polar thermophysical character of this headwall section of the glacier. It should be noted, however, that some surface drainage still takes place from this zone via selected channels.

Glacio-Hydrology

Liquid Water Content and Drainage

Free water content measurements (FWC) were made using the calorimetric method in the Cathedral Glacier's surface firn-pack in 1972 (Miller,



Figure 30 Cathedral Glacier from Atlin, B. C., 18 September 1974
(800 mm telephoto by R. Flanders)

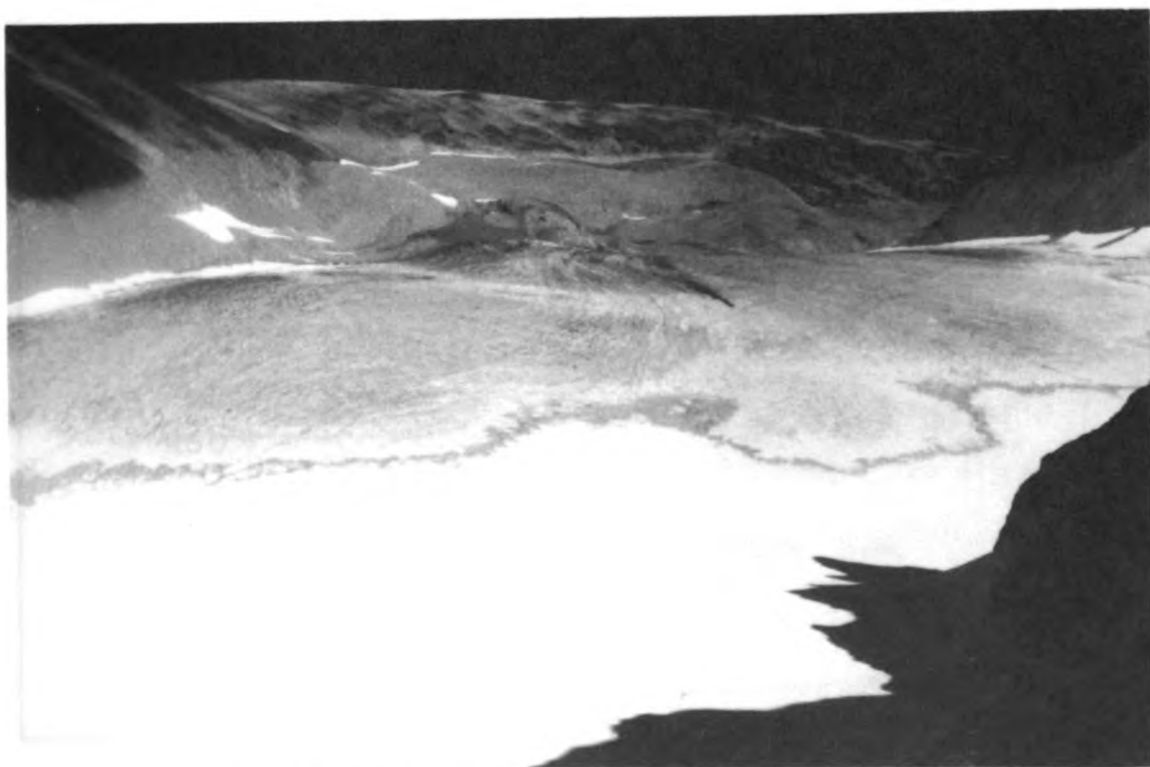


Figure 31 Cathedral Glacier from Cathedral Peak, 12 August 1974

**FIG - 32 FIRM PIT PROFILE AT CATHEDRAL
GLACIER TEST PIT SITE, 5850 Ft. (1770 m)
September 7, 1972.**

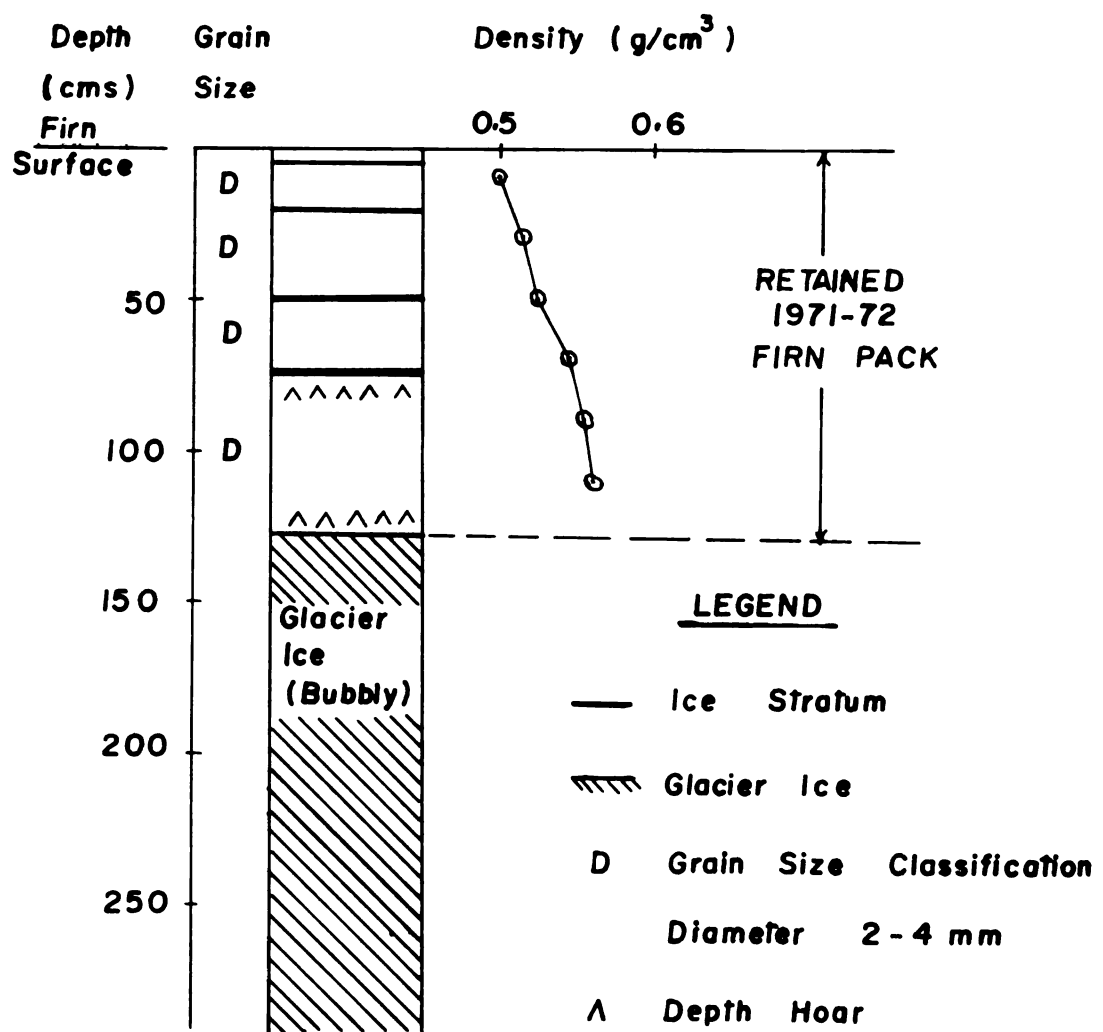


Fig. 32 Firm Pit Profile, Cathedral Glacier, 7 September, 1972

1975b). The FWC was measured at the following depths below the snow and firn surface on September 7, 1972 (Figure 33).

Cathedral Glacier Firn-pack at 5850 feet (1770 m)
on September 7, 1972

0 cm	Max. 13% FWC
20 cm	5% "
50 cm	5% "
100 cm	Fluctuational "

As shown in Figure 33 the FWC seemed to increase with solar radiation shortly before noon and to reach the maximum peak noted above whether the firn was in shade or not. Also revealed by the figure is a strong relationship between FWC and ambient air temperature at the 5850-foot (1770 m) level. FWC decreased within one hour after solar insolation ceased. Also a decrease in the surface level of the pro-glacial lake occurred within 2 to 3 hours after reduction of liquid held in the firn-pack, showing good correlation between the changes in FWC and the rise and fall of the lake. Thus changes in storage of liquid water and run-off in the Cathedral Glacier are shown to be dominantly controlled by variations in diurnal weather. The same may be expected with respect to longer-term climatic trends.

Hydrologic Run-off in 1972

Two types of hydrological measurements have been taken on the Cathedral Glacier system over each summer, one since 1972 and the other since 1973. A water-level recording gage with a stilling chamber was in operation during the 1973 and 1974 field seasons adjacent to a bedrock block on the northwest margin of the pro-glacial lake, just up-valley from Met Hill. While data from this gage have not been analyzed in detail, certain patterns emerge.

First, it has been discovered that substantial run-off does not begin on the Cathedral Glacier until early June, at which time the terminal

FIG- 33 FREE WATER CONTENT OF
CATHEDRAL GLACIER FIRN
5850 Ft. (1770 m)

September 7, 1972.

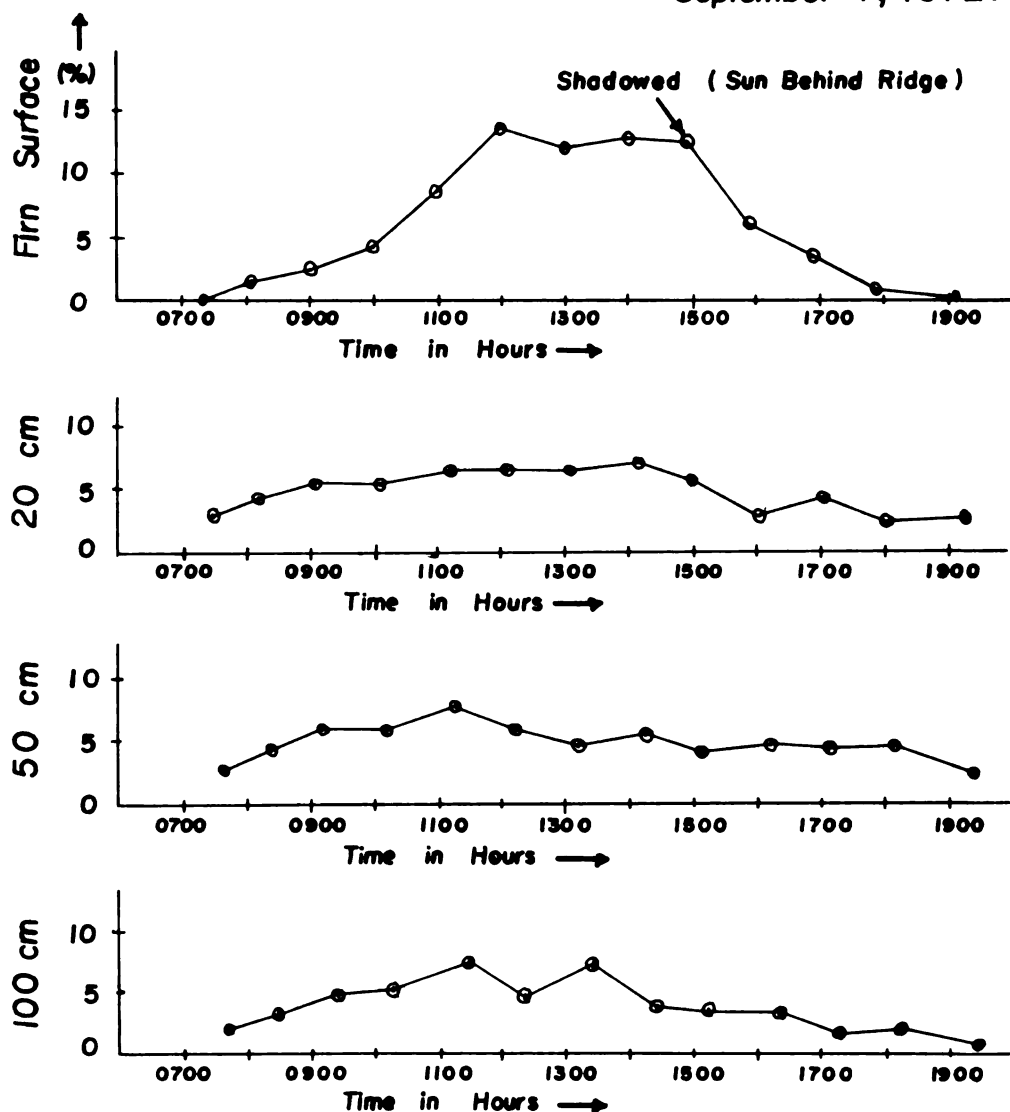


Fig. 33 Free Water Content of Cathedral Glacier Firn, 7 September 1972

lake is usually still frozen. Ice does not go off this lake until about June 20th as its elevation is relatively high, i.e. 5150 feet (1570 m). In fact, even at the relatively low elevation of Atlin Lake (elev. 2190 ft, 670 m) ice often remains in the coves and inlets until the first week in June.

At the terminus of Cathedral Glacier the pro-glacial lake drains to the north of Met Hill. During high water, a vigorous outlet stream rushes through the rocky channel. When the lake is low, little surface flow occurs, but there is some sub-surface drainage through the moraine. Glacial ice from the upper cirque terminates at the up-valley edge of the lake (see Figure 5). As noted below, lake level varies directly with ablation and run-off.

Higher run-off rates relate to the incidence of rainfall, increased solar insolation, and high ambient temperatures. On a good melting day with a high proportion of sunshine, the water level has been observed to increase sharply by noon (see Figure 34). It peaks in mid-afternoon, and decreases in later afternoon even before the sun drops behind Splinter Peak. Such an early decrease in daily melting provides an interesting question. Does the drop in lake level during the night depend solely on temperature? If glacial surface temperature drops below freezing, melting ceases, meltwater drains out of the firn-pack, and the lake level falls by early morning. On warmer nights, ablation and run-off continue and the lake level shows a less pronounced decrease during the night.

The other hydrological instrument employed in this study was a Stevens "A" stream level recorder installed with its float in a stilling chamber in the Cathedral (Elixir) Creek, some 300 yards below

FIG- 34 HYDROLOGICAL TREND DURING A DIURNAL CYCLE
ON THE CATHEDRAL GLACIER

Sept. 7, 1972 .

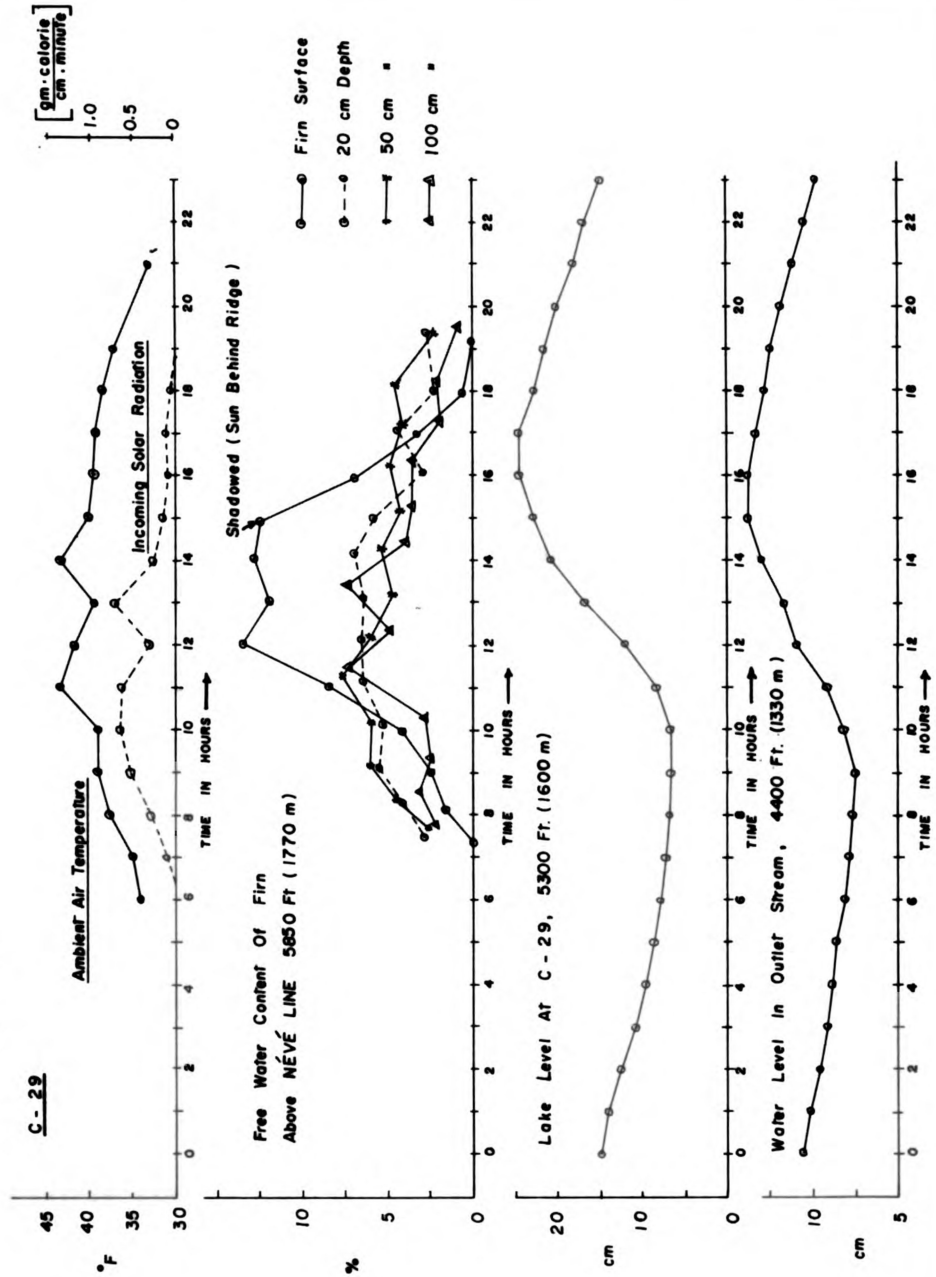


Fig. 34 Hydrological Trend During a Diurnal Cycle on the Cathedral Glacier,
7 September 1972

the main Neoglacial terminal moraine. Here again good correlation is found between ambient air temperature, solar radiation, ablation and run-off (Figure 34) but during each of the past two summers this stream gage appears to have experienced peaks earlier in the day than the lake-shore recorder.

As the stream-gaging station is located about a mile down-stream from the lake-level recorder, this difference in time lag does not appear at first glance to be logical. The lake receives water from much of the upper cirque, and the stream flowing from this lake contributes a small but significant fraction of the total Cathedral Creek flow. A large proportion of the creek flow issues from the base of the glacier terminus below Met Hill, draining all the lower cirque and some melt from the upper cirque. A third but rather small component this late in the season comes from a few remaining lower-valley snow-patches and from slowly melting ice-cores still remaining in some of the moraines.

It is suggested that ablation near and above the seasonal *nevé*-line in early September, when the data in Figure 34 were obtained, results in saturation of the firn-pack, producing a reservoir effect within the firn-pack. In the lower cirque, however, melt-water runs off the surfaces of the exposed glacier ice and drains away in well-established stream channels, taking short cuts through moulins and crevasses to basal streams. Thus it could drain out more quickly and reach the stream recorder down-valley sooner than water from the slushy firn-pack above the marginal lake.

The Cathedral Creek is an ideal closed system for hydrologic run-off and liquid balance studies. The entire run-off from the Cathedral Glacier system is restricted to the Cathedral Creek outflow. Therefore,

any significant changes in ablation and precipitation are directly reflected in the rate of discharge into this creek. The seasonal discharge culminates in a final cessation of flow when the autumn freeze-up takes place, usually in mid to late September.

Practical Problems in the 1974 Discharge Measurements

In 1974 further hydrologic data were obtained. Again discharge was measured at the Stevens "A" gaging station (see Guigné, 1975). Cross-section measurements by Guigné involved using a graduated wading rod combined with Price velocity meter readings. Because of changes in the cross section of Cathedral (Elixir) Creek since the 1972 and 1973 surveys efforts were made to make a discharge rating curve for use with the Stevens data. But this could not be developed in time because of malfunctions of the stream recorder, which were not overcome until late in the field season. Secondly, in this particular summer an unstable stream bed and banks, combined with large and sharply varying volumes of discharge from the glacier resulted in complicating variations in cross-sectional profile even over a few days (see Figure 35). Therefore a discharge rating curve, calibrated with the Stevens recorder, could only provide a rough approximation of actual discharge.

In 1975 these difficulties are to be overcome by the construction of an artificial control structure (weir) providing a fixed cross-section and some means of managing or allowing for bottom deposition and scouring. With this experience and some success in the stabilization of the stream channel at the rating site, subsequent hydrological data can be more accurate and useful.

ELIXIR CREEK PROFILE (Camp 29--ATLIN, B.C. CANADA)

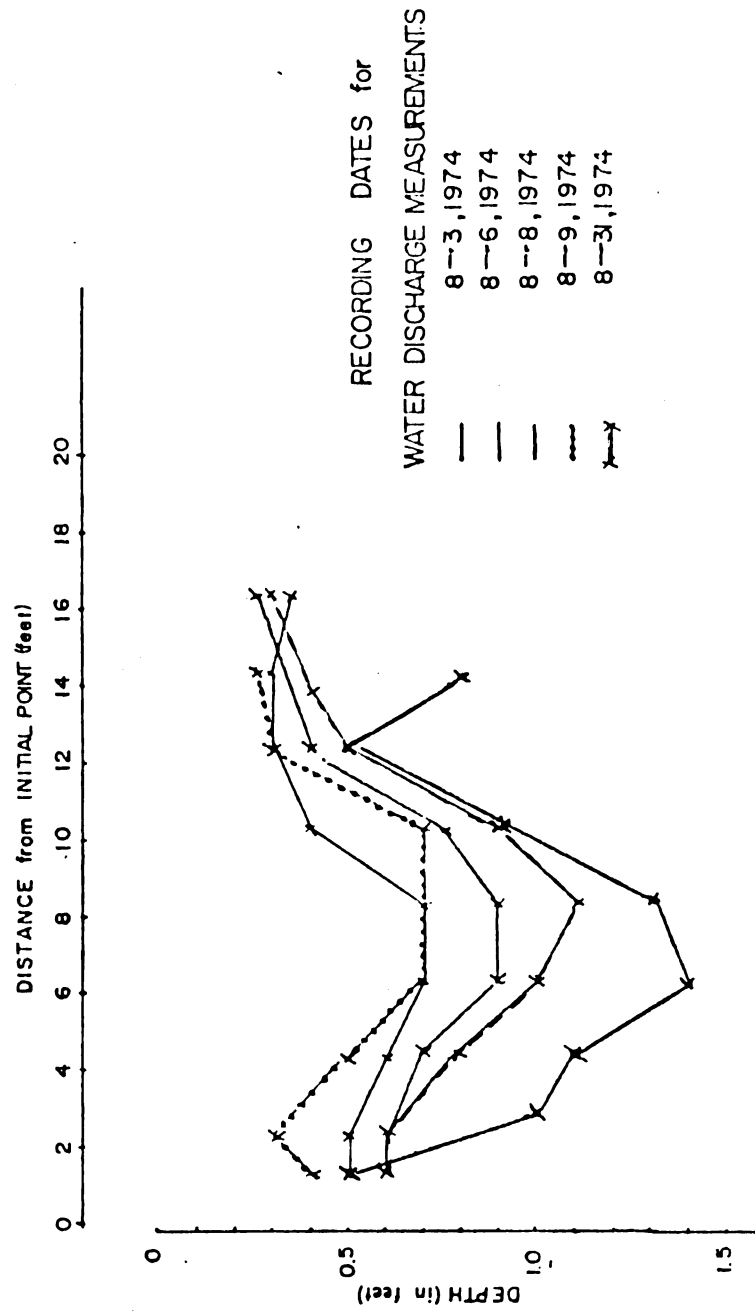


Fig. 35 Cross-Sectional Profile, Cathedral (Elixir) Creek (after Guigne')

An Anomalous Hydrological Surge

The volume of stream flow appears to be related primarily to air temperature. On warm days flow reached high levels by 0300 or before. The occurrence of rainfall also increased run-off, not only from the immediate precipitation but also from warm rain (above 40°F) producing increased melting.

On one day in mid-August of 1974, however, a dramatic increase in stream flow occurred which could not be ascribed to unusually warm temperatures or to precipitation. Great volumes of brown water charged with a heavy load of silt rushed from the lowest point of the glacier terminus below the research station, causing violent and turbulent flow in the overloaded stream channel. This exceptionally heavy flow was far greater than any other observed that summer. It subsided by the following morning. The proglacial lake and its out-flow stream, being above the terminus and beyond Mt Hill, were not affected.

This appears to have been a genuine hydrological surge or glacier burst (Jokulkaup) similar to ones reported by Asher and Miller (1974) on the Lemon Glacier on the Juneau Icefield near Juneau. It is possible that the pronounced change in stream profile at the stream gaging station, occurring between August 9 and 31, as shown in Figure 38, may have been due to increased erosion during this period of catastrophic flow. Thus artificial control structures for gaging purposes, as suggested above, must be designed to withstand and function during unusual extreme events.

The advance party opening Camp 29 for the 1975 summer field season reported that the entire stream gaging station below the

Neoglacial moraines had completely disappeared. It is hoped that the Stevens recorder will be found downstream when the snow melts, and that the gaging station can be reconstructed in an improved form and returned to service. As the gaging station was intact at the time of freeze-up in the fall of 1974, and heavy outflow from the 1975 ablation season had not yet begun, one can only speculate as to the cause of the disappearance.

Conclusions

Conclusions drawn from these observations are that in the present decade the effective ablation season on Cathedral Glacier is of less than three months duration but usually greater than 2 1/2 months. In spite of the long winter season, the net mass balance of the Cathedral Glacier in 1971, 1972, 1973 and 1974 has been slightly negative. From the regional weather records this situation has apparently pertained ever since the 1920's. This negative trend, however, now appears to be reversing. Thus it is imperative that future glacier mass balance and hydro-metric records be acquired annually and without interruption in the years ahead in order to document future effects of this current cooling trend, which is expected to bottom out by the turn of the century.

Snow course accumulation data are as yet too short term to be conclusive, but the situation in 1974 suggests that patterns of snow depth on Cathedral Glacier may be more closely related to Log Cabin, at 2880 feet (880 m) elevation and 45 miles (72 km) northwest of Camp 29, than to the Atlin Lake snow course, at 2400 feet (730 m) and 25 miles (40 km) to the north northeast.

Snow and firn-pack depths, ablation rates, late-season névé lines, and the transient snow-line were all quite different in 1974 than in 1973.

In 1974 bedrock features and moraines were exposed to a much greater extent than in 1973, with much more pronounced down-wasting of ice in the terminal area.

Hydrologic studies are underway in the Cathedral Glacier system. As this is a closed hydrologic basin, improved techniques and equipment will pay dividends in the mass balance studies and other aspects of the liquid and heat balance research.

It is obvious that interpretive problems arise from having too short a period of record at the Cathedral research station to date. Longer time-intervals of data are required to eliminate the effects of short-period variations and to reveal actual trends. But a start has been made in what is projected to be an important long-term program of glacial, climatic, and hydrological research. This station is a key element in synoptic studies for long-term glaciological and meteorological research. Hydrologically this area lies at the headwaters of the Yukon River. Establishment of the Atlin Provincial Wilderness Park, encompassing the Cathedral area, gives far broader implications, however, than only scientific interest.

CHAPTER VII

LOCAL MORPHOGENETIC PROCESSES

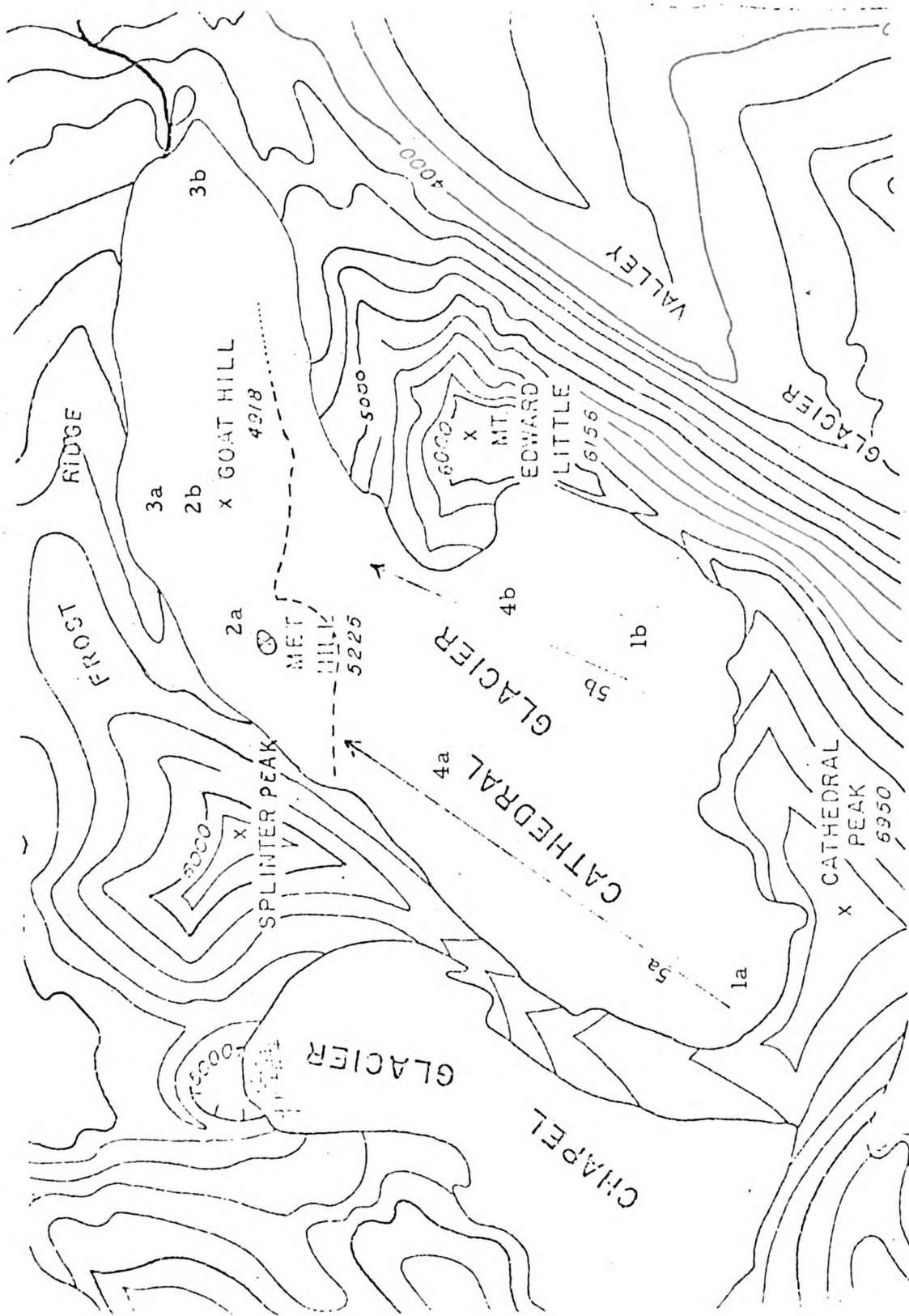
This chapter describes the evolution and evaluation of several concepts relating to the nature and behavior of morphogenetic processes in Cathedral Valley.

Observed Morphological Elements

During the writer's initial field work in the 1973 summer field season five significant morphological elements were observed in the Cathedral System. These are noted as follows, with the cited designations shown in Figure 36.

- 1) the existence of two separate cirque basin accumulation areas (1a and 1b)
- 2) the existence of two mid-valley bedrock knobs (Met Hill, 2a and Goat Hill, 2b) each having exerted strong structural control on glacier flow in the valley
- 3) the existence of one major terminal moraine on the north side of the valley, (3a), and a second one located on the southeast side but farther down-valley, (3b)
- 4) a nearly convex ice surface flowing from the upper (southwest) cirque, (4a), alongside a lower, nearly level ice surface in the lower (southeast) cirque, (4b)
- 5) a significantly longer flow distance from the upper cirque, (5a), than from the lower cirque, (5b)

From these field relationships the idea emerged that the two terminal moraines on opposite sides of the valley represent separate advances from the two separate accumulation areas. Possible mechanisms involved are considered in the following section.



----- 1974 terminus

Fig. 36 Morphological Elements of the Cathedral Glacier System

Concepts of Flow Mechanisms

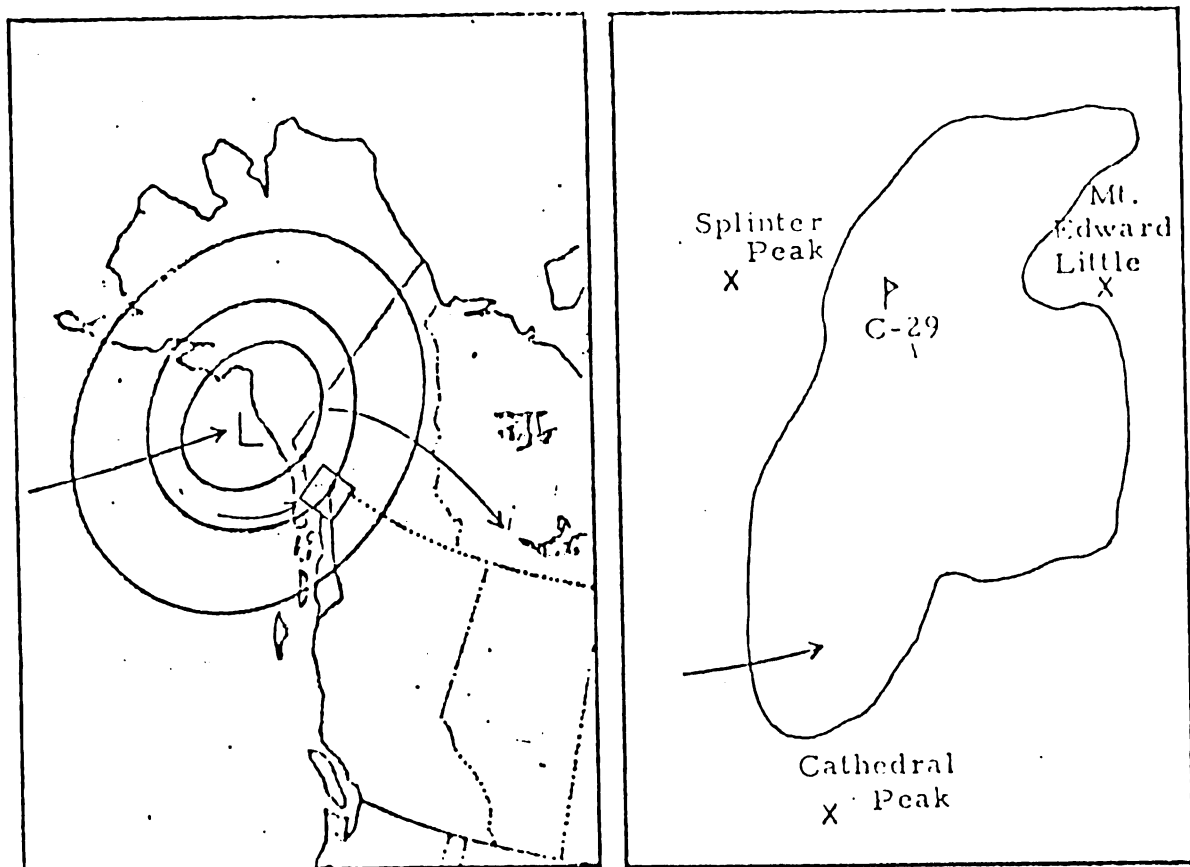
Development of Alternate Flow Mechanism Concepts

Wind Direction Concept

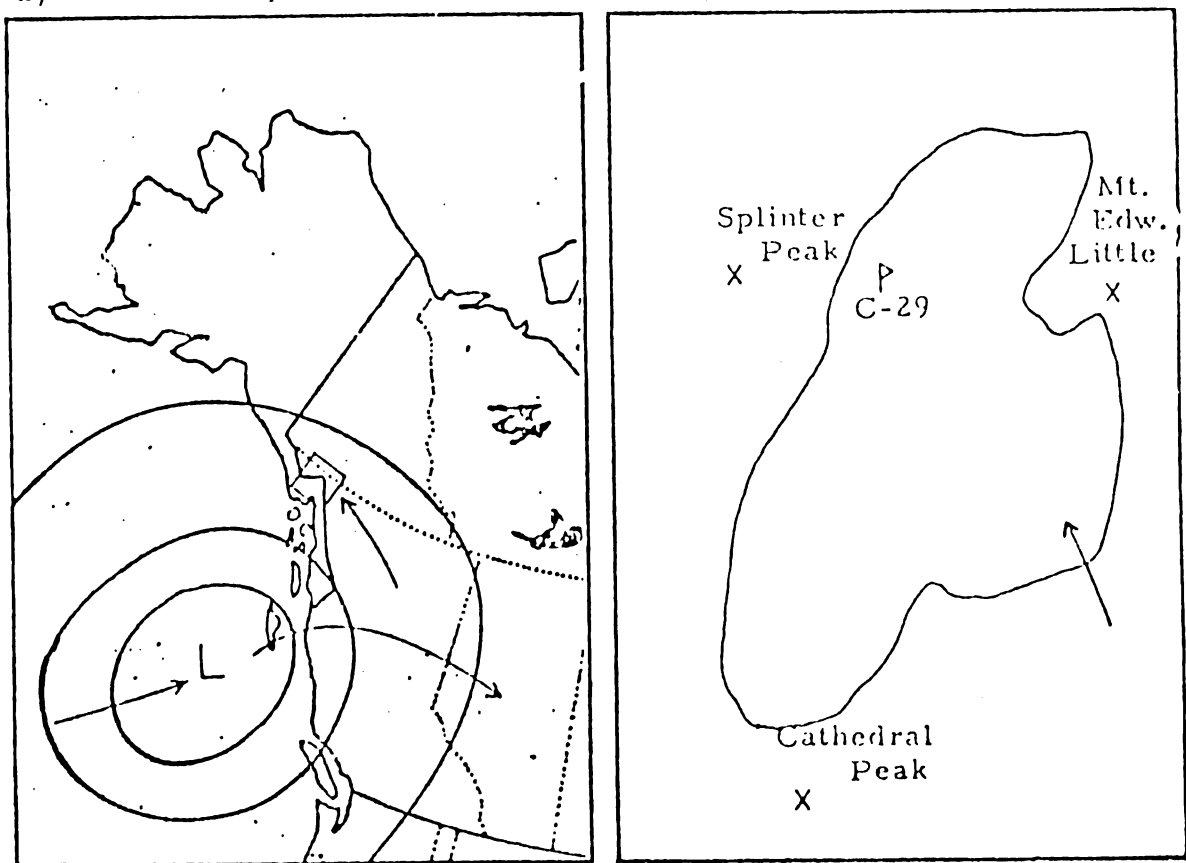
It is possible that differential or out-of-phase glacial flow on the two sides of the lower valley has been due to differing amounts and timing of accumulation in the two cirque-basin accumulation areas. This concept has already been well-documented with respect to the multiple névés nourishing different outflow glaciers on the Juneau Icefield. The prime example has been the presently receding Norris Glacier and the presently advancing Taku Glacier in the Taku District (Field and Miller, 1951; Miller, 1963 and 1973a). The mechanism suggested for this was a shift over time in the mean source directions of snow-bearing storm winds. A southward shift of the dominant mean cyclonic track of the movement of counterclockwise low pressure storm systems could conceivably change the dominant wind direction during snowfall from southwest (upper cirque accumulation) to southeast (lower cirque accumulation), as shown in Figure 37. Each situation produces an asymmetrical accumulation wedge on the across-glacier transect in the main névé zone, as shown schematically in Figure 38.

Mass Balance-Flow Lag Concept

Stereographic aerial photo coverage was obtained after the 1974 field season. Although outdated, the results of this photography appear to show the nature of the dual accumulation areas and differences in flow distance. From ground truth coupled with analysis of the photographs four physiographic zones appear to exist on each side of the valley. These zones represent relatively heavy, light, heavy and again light net accumulation of snow, respectively. Heavy accumulation zones on one side of



a) Storm track passes northward, bringing southwest winds



b) Storm track passes southward, bringing southeast winds

Fig. 37 Storm Wind Directions, Cathedral Glacier

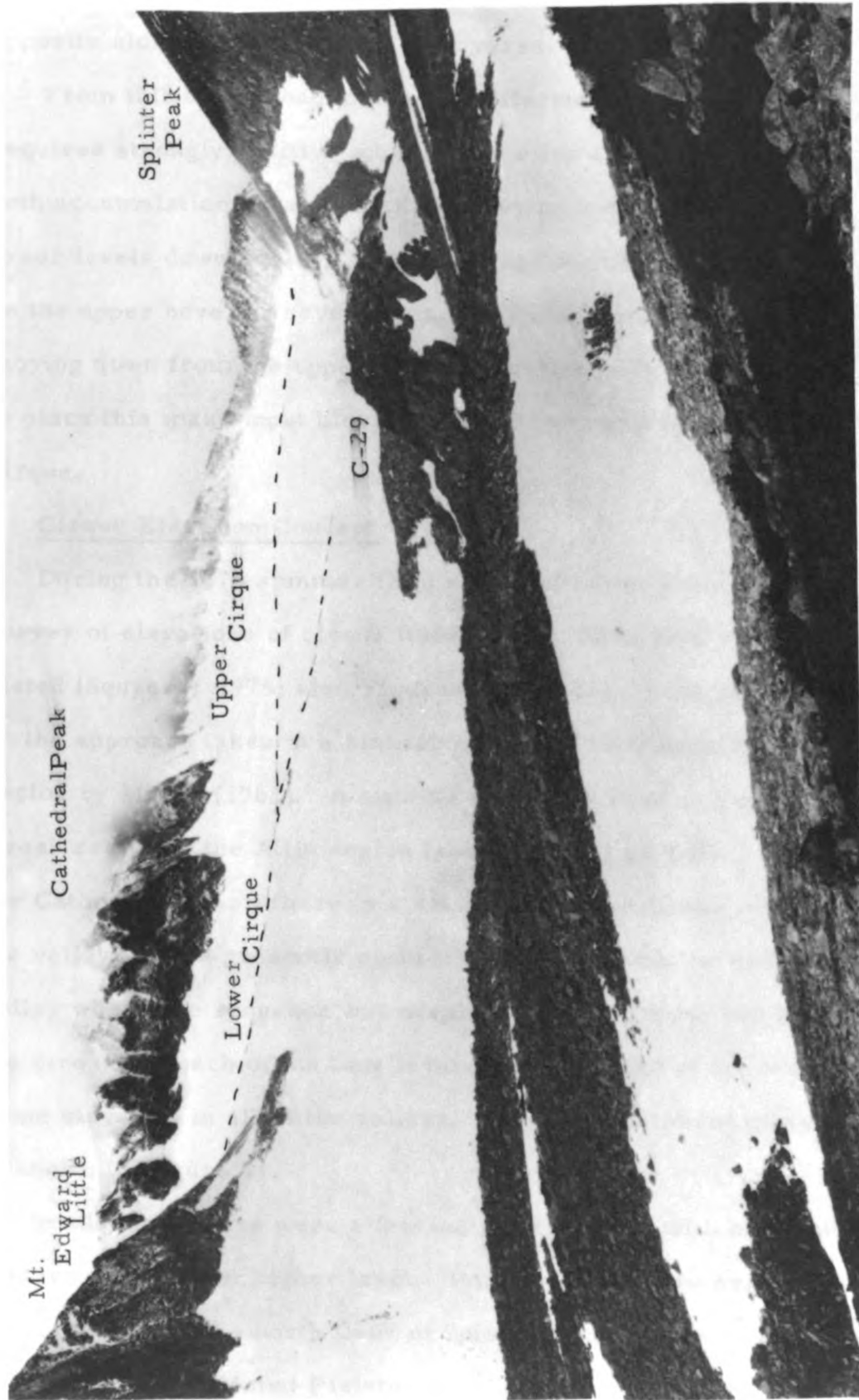


Figure 38 Cathedral Glacier, showing Upper and Lower Cirques and Accumulation Wedges

of the valley were contiguous to light accumulation zones on the opposite side of the valley and vice versa (Figure 39).

From this the second idea as to differential flow emerges. This requires strongly positive mass balance for a period of years over both accumulation areas. As this heavy accumulation moves to lower levels down-valley, a light or negative mass balance prevails on the upper névé for several years. Thus, heavy accumulations moving down from the upper cirque would require a flow lag sufficient to place this mass input alongside a light accumulation from the lower cirque.

Cirque Elevation Concept

During the 1974 summer field season a reconnaissance altimeter survey of elevations of cirque floors on the Cathedral massif was completed (Squyers, 1975; also Figures 40 and 41). This survey was based on the approach taken in a similar survey of cirques in the maritime sector by Miller (1961). A similar study was done in the Fourth of July Creek sector of the Atlin region (see Figure 3) by Tallman (1975). On the Cathedral Massif there is a sequence of four cirque levels in each of the valleys which presently contain glaciers. With the exception of one valley where the sequence was displaced upward some 300 feet (90 m) the cirques at each of the four levels were grouped at approximately the same elevation in all of the valleys. The distribution of these elevations is shown in Figure 41.

In addition, there were a few incipient cirques with contained ice masses at a fifth or higher level. Permanent ice now exists at 6100 feet (1850 m) on the north flank of Splinter Peak. During all of the past history of repeated Pleistocene glaciation on this massif, ice has not previously accumulated in this location in an amount sufficient to

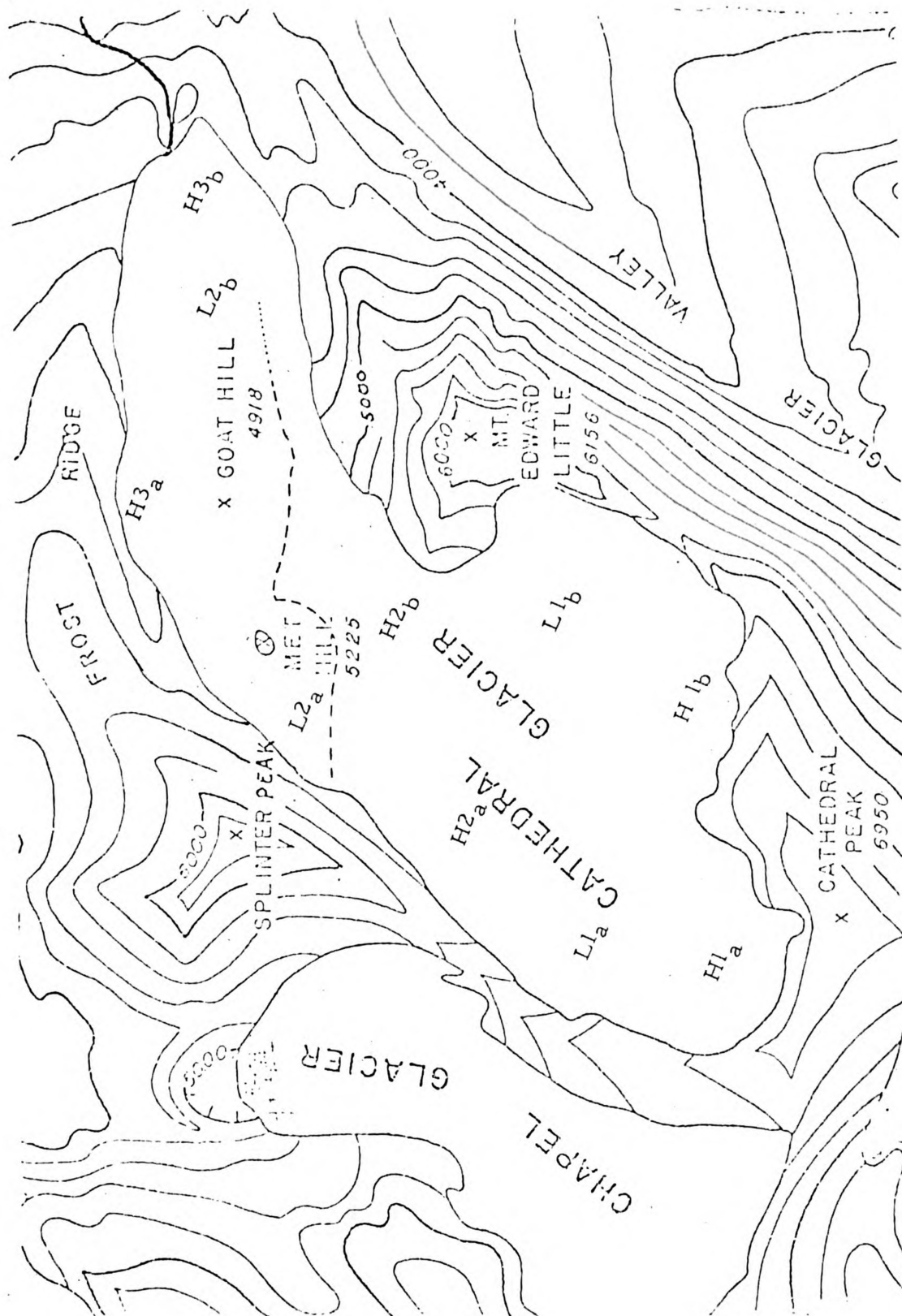


Fig. 39 Apparent Heavy and Light Accumulation Zones, Cathedral Glacier System
 (Note: letters represent post-flow locations)
 ----- 1974 terminus

CIRQUE ELEVATIONS OF THE CATHEDRAL MASSIF

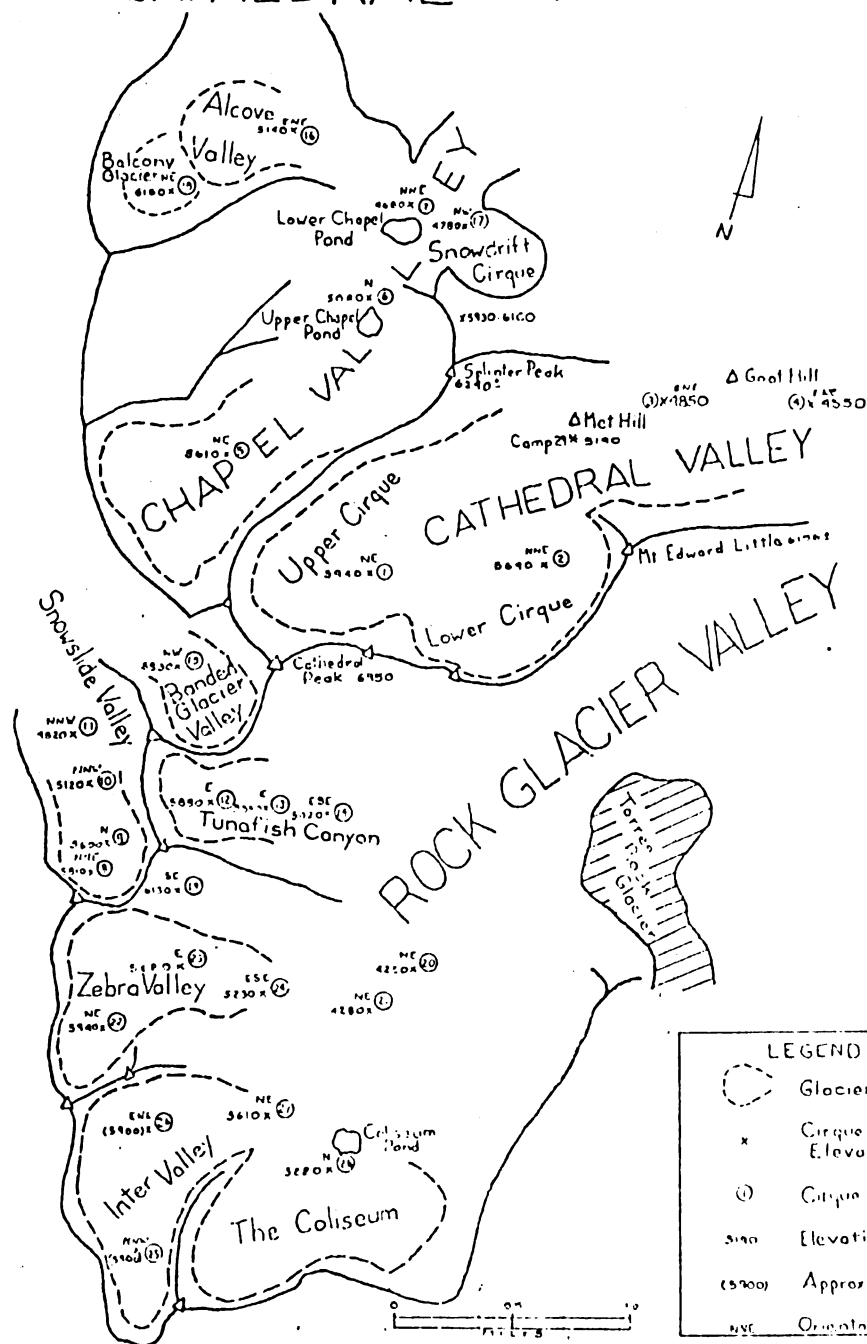


Fig. 40 Cirque Elevations of the Cathedral Massif (modified from Squyers, 1975)

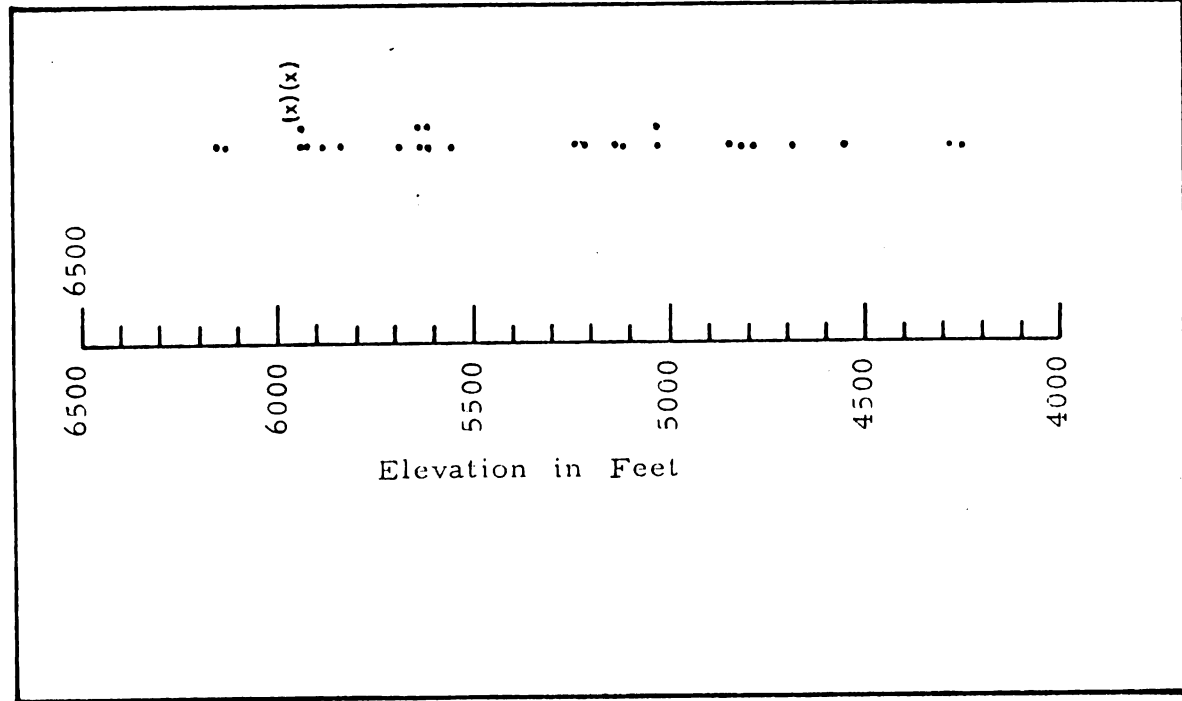


Fig. 41 Cirque Elevation Distribution, Cathedral Massif

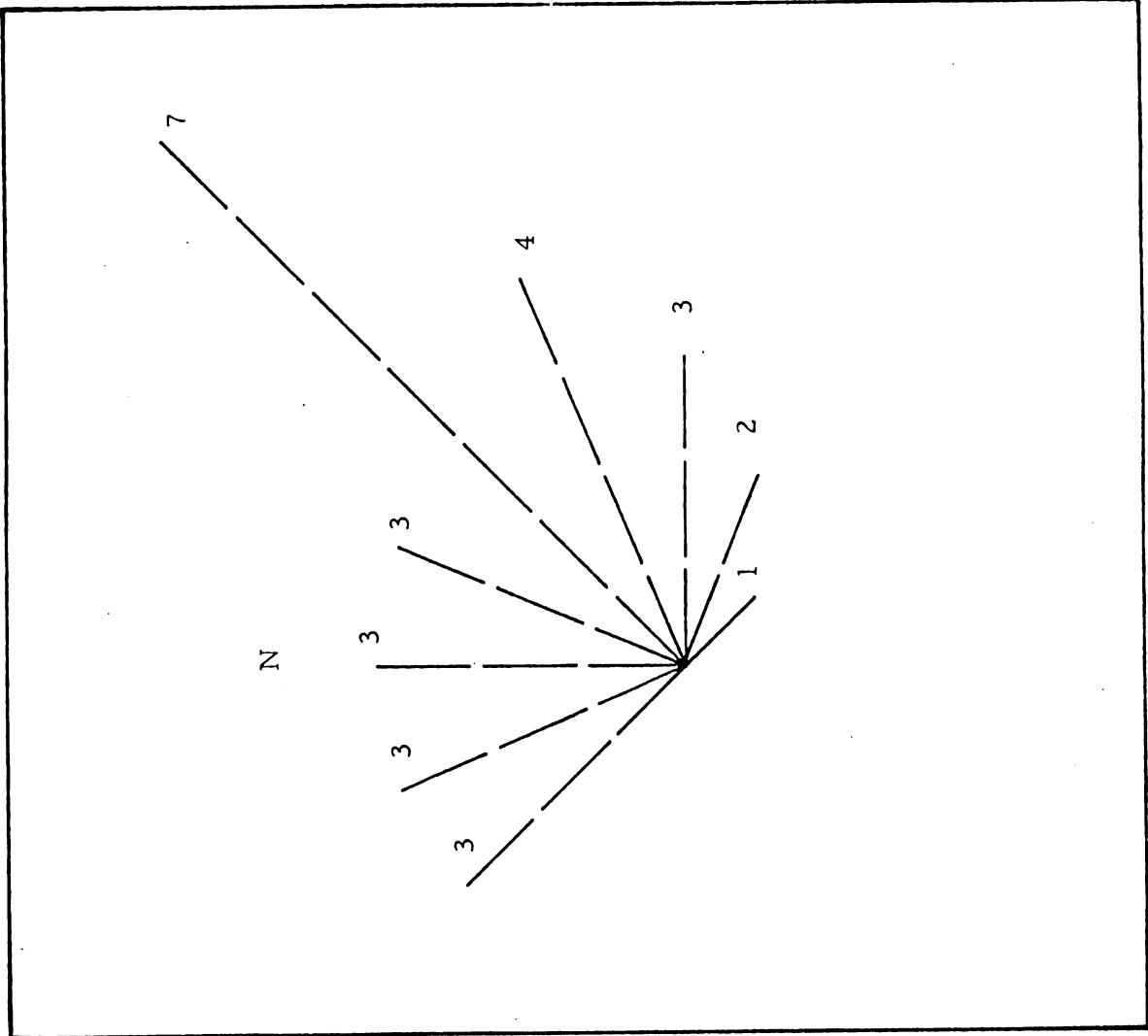


Fig. 42 Cirque Orientation, Cathedral Massif

cause erosion or transport. A previously active cirque exists on the same slope several hundred feet below. Both features appear to have formed from accumulation via winds from the southeast. Two other partially formed incipient cirques exist on the massif at about the 6100-foot (1850 m) elevation.

The cirque elevation study corroborates Miller's (1961) contention that these cirque levels represent elevations of former maximum accumulation, and therefore relate closely to the mean elevation of former freezing levels. Changes in regional temperature regimes would shift the elevation of maximum accumulation from one cirque level to another; a cooling shifting the accumulation downward and a warming trend shifting it upward. Such temperature changes would of course also affect heights of the seasonal and regional névé lines due to changing temperature-elevation effectiveness in the ablation process as well as in the form and magnitude of precipitation.

With a saturated adiabatic lapse rate of $3^{\circ}\text{F}/1000\text{ feet}$, * a 300-foot (90 m) difference in freezing level and thus in elevation of maximum accumulation would require a change in mean temperature of 0.9°F . From 1907 to 1943 the mean annual temperature at Atlin increased by approximately 4°F . Since 1944 it has decreased by the same amount. This would represent a change in freezing level of about 1330 feet.

Linville (1975) suggests that a dry adiabatic lapse rate should be used, since air below the lifting condensation level (LCL) would not be saturated, and hence saturation and therefore precipitation would not be reached below the LCL. With a dry adiabatic lapse rate of $5.5^{\circ}\text{F}/$

.....
 * Because all of the Canadian Weather Service and U.S. National Weather Service data from neighboring stations are in degrees Fahrenheit, in the interest of consistency this unit is employed in this report.

1000 feet, a 300-foot (90 m) difference in the freezing level would require a temperature change of 1.65° F.

In this study area the dominant accumulation level (positive mass balance) at present appears to be in the upper cirque, at about 5900 feet (1800 m). The glacial mass in the lower cirque at 5600 feet (1720 m) at present has a negative annual mass balance. During the late Wisconsinan and earliest Holocene, cirque formation is thought to have occurred at two lower levels. The cirque level study places these two ancient levels at 4100-4400 feet (1250-1340 m) and 4700-4900 feet (1430-1490 m).

Alternating dominance of flow from one presently ice filled cirque to the other, caused by changes in the persistence of mean freezing level, would cause a phasing effect. Thus the resulting ice masses and their related moraine configurations provide physiographic evidence concerning changes in past climatic regimes.

Discussion

By each of the above concepts the Cathedral Glacier system would provide a unique and useful situation for determining past climatic patterns, because of the dual accumulation area and bilaterally-phased flow.

Williams (1975), in study of corrie (cirque) elevations in Baffin Island, points out that changes in snowline for a given temperature change does not depend solely on lapse rate. His study also includes the accumulation gradient, ablation gradient, absorbed direct radiation and snowfall-induced albedo change. Snow accumulation estimated for different periods ranged from 40 to 400% of today's amounts. However, he concludes that the main controlling factor modifying lapse rate effects

was aspect or orientation. The net difference increases with elevation (Williams, 1975, p. 178).

If quantities of precipitation on the Cathedral Massif may reasonably be considered to parallel those at Atlin, as shown in Chapter 5, changes in precipitation over time do not by themselves appear to be of very great magnitude. Variations in direct solar radiation on the glacier surface can on the other hand be quite significant. This is strikingly illustrated by the ice surface in the shade of Mt. Edward Little, where shaded ice slopes steeply down to the empty valley floor. It is also shown on Banded Glacier (Figure 43) where the mid-afternoon summer sun reaches past the end of the south valley wall to melt the north and central portion of the glacier and leave a vertical face on which the annual layers of former retained accumulation are beautifully exposed.

The effect of aspect on glacial valley development on the Cathedral Massif is shown by analysis of the orientation of cirque areas (Figure 42). In this the dominant cirque orientation is northeast, ranging from northwest to southeast and with no valleys oriented on the sector between 135° and 315° T.

Thus while all the major controlling factors considered in this study such as ablation gradients, depth of retained snowfall (firn), and the elevation of the annual névé line are important, the orientation of cirques are of major importance on this massif.

Evaluation of Theories

Several difficulties remain with the idea of a bilaterally-phased advance. One problem is that there is not adequate room to the north side of either mid-valley bedrock knob for the ice from the upper cirque to pass. The result is that the flow down the main part of the valley con-

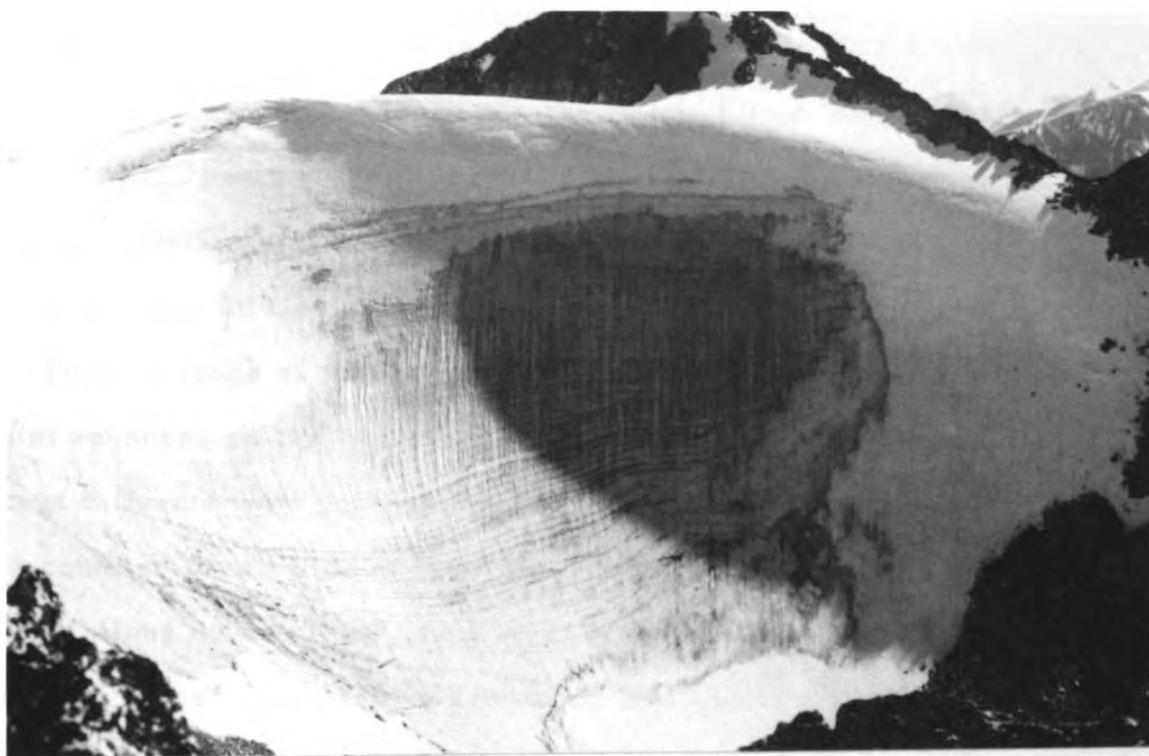


Figure 43 Banded Glacier, Cathedral Massif



Figure 44 Dual Trimlines on the Flank of Mt. Edward Little

tains ice from both cirque sources. This is supported by the existence of a medial moraine of dark matabasaltic rock trailing downvalley from the cleaver between the two upper cirques and down past the south side of Met Hill. From there it is traceable to the terminus, continuing as a linear ablation moraine atop the lighter-colored till and on over the south shoulder of Goat Hill.

The existence of different down-valley termination points of maximum advances on the two sides of the lower valley does not necessarily prove different time periods for these advances. Also all advances were not confined to one side of the medial hills or the other. Very deep penetrations down-valley such as in the 1760's completely overrode Met Hill rather than passing around it, and nearly covered Goat Hill.

A key observation was made during the field studies in 1974. The left lateral moraines on the north side of the lower valley (see Figure 52) have previously been considered to be entirely marginal push moraines formed by ice from the upper cirque bringing debris off of the flank of Splinter Peak. Now it is clear that the lower part of the inner moraine ridges were truncated and disturbed by ice moving out of the lower cirque and thence across-valley below Met Hill but above Goat Hill. A small percentage of boulders remained upright during this disruption and were able to retain their lichen cover, but most were rotated and the lichen obliterated. Also during this process a few boulders were excavated from basal positions, where they had accumulated iron-staining, and were moved to the top of the moraine.

As shown in Figure 9, the valley configuration allows ice moving out of the lower cirque to move directly across the valley below Met Hill, if other ice is not in the way, rather than making an 80° right turn into the lower valley.

The concept of bilaterally phased flow is a key element in the value of the Cathedral Glacier system for reconstructing past climatic regimes in this region. What evidence do we have that this phased flow actually existed?

The primary evidence resides in the bulk and spatial configuration of the present ice masses. The ice surface of the lower cirque basin today is nearly flat along its center. Only its sides and back curve upward to the headwall. At the front the ice spills steeply some 400 feet downward past Met Hill to its present terminus. This frontal zone is augmented by a steep ice slope reaching up the shaded flank of Mt. Edward Little. In contrast, ice from the upper cirque basin does not even reach Met Hill. Its receding terminus ends in surface till shortly up-valley from the hill and above the proglacial lake (see Figures 5 and 6), though a small component of ice still joins with the lower-cirque ice to pass to the south of the nunatak.

Under present climatic conditions the bedrock and till-littered berm extending in a cross-valley direction from Met Hill toward Mt. Edward Little continues to become more exposed. At the terminus above Met Hill, upper-cirque ice is only slightly below the elevation of lower-cirque ice, which continues to rise steeply to the lip of the lower cirque. About 80 meters up-slope from this terminus, however, the ice surface in the upper-cirque slopes upward more steeply. These changes in slope, the ablation on the nose of the slope, and snow drifted at its base give this section the appearance of a convex bulge. Measurements with slope finder indicate a gradient of 10° just above the terminus, changing to 13° on the steep portion, then decreasing to 8° in the up-glacier section. At the top of the steepest slope, ice from the upper-cirque (5650 feet,

1700 m) lies about 100 feet (30 m) above the surface of the lower-cirque ice (5563 feet, 1670 m). Upper-cirque ice continues to slope gradually up for another 100 meters to the base of the headwall slope. Figure 45 schematically shows the relative configuration of these two glacier surfaces.

Other evidence of past surface gradients is a set of trim-lines on the west face of Mt. Edward Little, which show different slopes (Figure 44). A study has been made of the relative angles of these lines, in comparison to a similar set on the east flank of Splinter Peak (Miller, 1975, personal communication). This is suggested to be related to the asymmetrical accumulation wedges caused by secular changes in dominant direction of storm winds. These accumulation wedges are also suggested by the surface configuration depicted in Figure 38.

Analysis of till-boulder locations below Met Hill in regard to rock type and provenance suggests that during most of the recent glacial activity the glacier from the lower cirque has been more dominant than it is at present. A further relevant element of information to be researched by geophysical means is the elevation and configuration of the bedrock floors of these cirques. This will resolve the critical question as to whether the surface elevation of ice in the two cirques is controlled by bedrock base elevation or glacier depth, or both.

A Summary and Synthesis

Three concepts are developed with regard to glacier flow and moraine emplacement in the Cathedral Glacier system. All of these concepts stress the idea of an out-of-phase bilateral flow, based on the recognition of two main névés. As the relationships are somewhat complex these concepts are briefly reviewed as follows:

- 1) There has been an alternating of maximum snow accumulation from one cirque to the other, due to long-term shifts

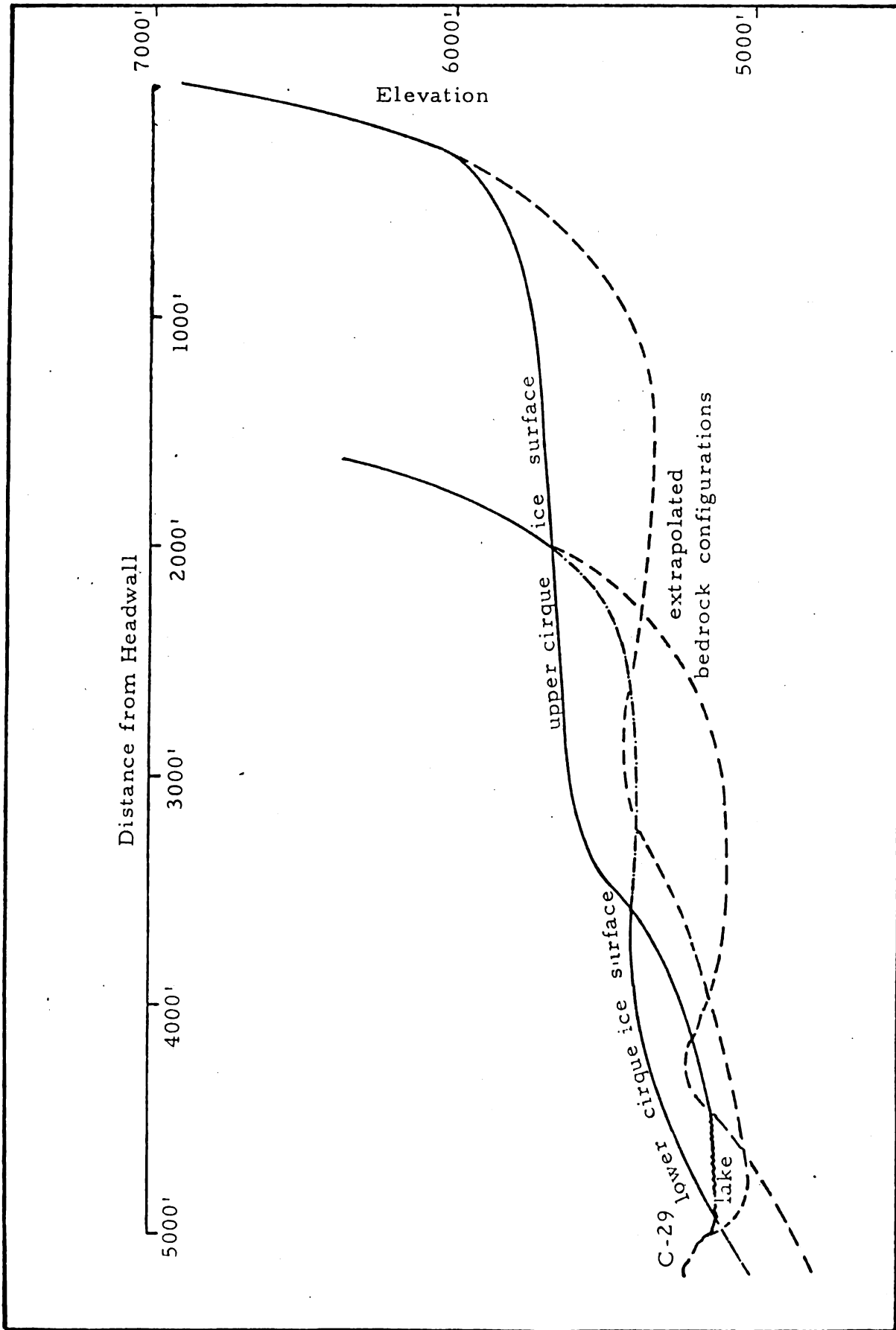


Fig. 45 Longitudinal Section of Upper and Lower Cirques, Cathedral Glacier

in the direction of dominant storm winds and of associated accumulation by wind drifting.

- 2) There has been substantial change in mass balance of this glacier system over a period of years, with phasing caused by differing flow distances (flow lags) from the separate accumulation areas (névés).
- 3) There have been significant vertical changes in the elevation of the mean freezing level due to changes in mean temperature. This in turn has periodically changed the dominant net accumulation from the lower cirque to the higher one, and vice versa.

It would appear that each of these three conditions to some extent has affected the pattern of accumulation. Certainly some shift in the dominance of one névé zone versus another has occurred. Effect of this dominance on moraine patterns has been influenced by the configuration of bedrock features such as Met Hill and Goat Hill. While amounts of total precipitation may have a tendency to vary with temperature changes over time, and temperature levels certainly affect the forms of precipitation, changes in freezing levels are believed to play the dominant role with respect to the patterns of accumulation which have governed the regime of the glacier system.

CHAPTER VIII

THE CATHEDRAL GLACIER MORaine COMPLEX

Introduction

This chapter concerns the description and present interpretation of the morphostratigraphic sequence which is found in the Neoglacial moraine system of the Cathedral Glacier Valley. To some extent the interpretations are subject to modification as more detailed mapping is done and further information becomes available. The description is presented in some detail with the hope that 1) the unusual complexity of this moraine system will be clarified; 2) the reader will be able to visualize the relationships involved, and 3) this will provide an on-the-ground guide and stimulus for further analysis by future researchers and field personnel working out of Camp 29. It is suggested that the reader refer frequently to the sketch map of the moraine system in Figure 46 as well as to the numerous photographs presented in this report in order to maintain a proper orientation as to location, nomenclature, and sequence.

The writer spent nearly three months in various aspects of the ground study at the Cathedral Glacier research station during the 1973 and 1974 summer field seasons. The interpretations are based on this field work and the analysis of other information available and on discussions and exchange of ideas with Dr. Maynard Miller, both in the field and after return to the campus. Analysis of the meteorological information was also augmented by discussion with Dr. Aylmer H. Thompson, JIRP meteorologist, and other staff members and participants in the JIRP during the past three years. Also much useful information has been



Figure 46a Lower Cathedral Valley Moraine System, Phase I

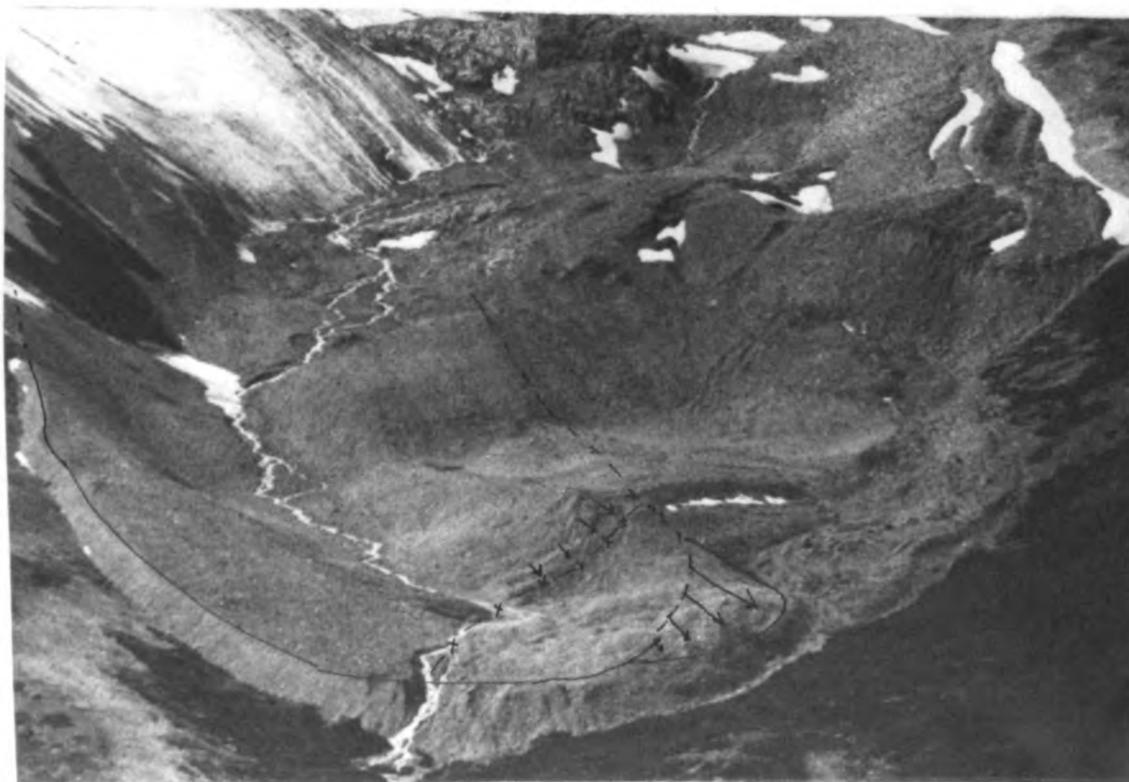


Figure 46b Lower Cathedral Valley Moraine System, Phase II



Figure 46c Lower Cathedral Valley Moraine System, Phase III

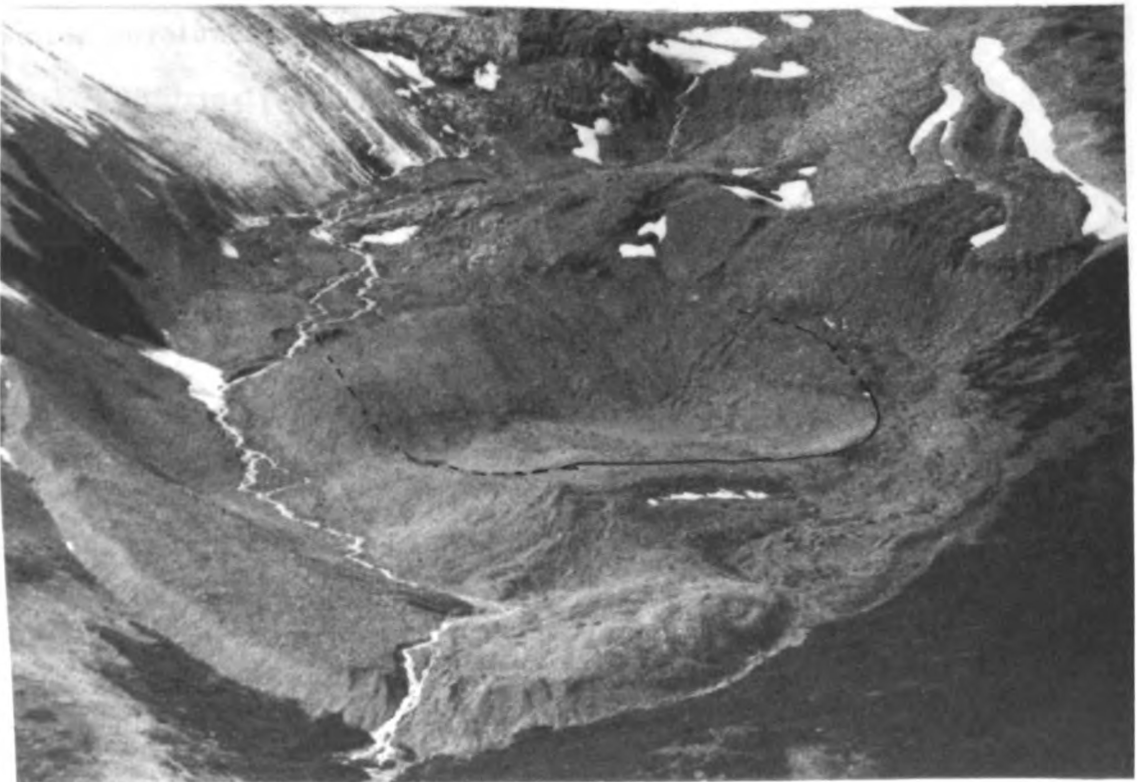


Figure 46d Lower Cathedral Valley Moraine System, Phase IV

derived from low-level aerial photography and lichenometric studies, with significant assistance from Marianne See of the Smithsonian Institution and the University of Alberta.

Lichenometric Studies

Lichenometric data and botanic descriptions are based on a JIRP report on lichenometric studies by Marianne See (1975).

The thallus diameters of Rhizocarpon geographicum and R. macrosporum on morainic debris in Cathedral Valley were measured in July, 1974. Supporting information was obtained from other species of lichen and from vascular plants.

Field work in 1973 and 1974, using gravestones in the Atlin cemetery as a reliable dating base, has established the Atlin lichen factor for Rhizocarpon at 28mm/100 years. This factor is used to convert the thallus diameters into actual dates B. P. (before present). Humidity and temperature are critical factors for these lichen, and year-around weather information would help to relate the Atlin lichen factor to the Cathedral area. Because the lichen factor has been derived from the tombstones in the Atlin cemetery, which is located in a drier but warmer location, use of the Atlin lichen factor will probably not result in precise estimates of time elapsed in the Cathedral area, but dates are considered to be accurate to within 30 years. (Note that if one date is moved 30 years, all others must be adjusted similarly).

Late Neoglacial Advances

Advances Prior to the 18th Century

The strongest advance in the valley since the end of the Wisconsin took place in the mid-18th century. This is commensurate with the conclusions made with respect to Alaskan coastal glaciers by earlier JIRP

researchers (Lawrence, 1950). As in Alaska the advance obliterated nearly all evidences of former advances in the Cathedral vicinity, with three exceptions thus far discovered.

Below the northern terminal moraine is an area of thinly-spread ground and terminal moraines which is stratigraphically older than the vigorous 18th-century moraine which overrides its eastern edge (see Figures 52 and 53). The ice advance represented by this moraine passed to the southeast of Goat Hill, then spread around the base of the hill to the northeast. Similar flow patterns later were caused by blockage of the southeastern part of the valley by relict ice from previous advances. This suggests that an earlier major advance, similar to that of the 18th-century, took place shortly before the one described above. This is consistent with a three-fold pattern of advance noted on the Juneau Icefield (Miller, Egan, and Beschel, 1968; Beschel and Egan, 1965).

The weathering on this moraine, although more advanced than on the younger moraines overriding it, is only a small fraction as extensive as on the late-Wisconsinan moraines lying just to the north.

Lichenometric data on this moraine indicate a maximum exposure date of 1814. However, we have seen that the 18th-century moraine is stratigraphically younger. As an explanation, lichen development may have been inhibited by extensive snow-cover during much of the post-depositional history of the pre-18th century moraine. On the other hand, development of other plant forms is much more advanced here than in any other post-Wisconsinan (Neoglacial) moraine area in the valley. Lush growth of fireweed, erect forms of spruce, mattes of Dryas, and large willow shrubs occur extensively, suggesting a longer time for development of a vegetative cover.

It is noted, however, that vegetation growth is favored by a protected location and thus a favorable microclimate. Also this glacial deposit is fine-textured and thus provides a favorable medium for establishment of vegetation. During much of its exposure it has been well supplied with moisture by through-flowing melt-water streams. It is somewhat shielded from katabatic winds by high moraines and by Goat Hill. The southward-facing Wisconsin moraine near its margin optimizes the warming effects of solar insolation, and serves to reflect and re-radiate heat onto this surface. A slight ridge composing its arcuate outer margin is composed of lag boulders, and so this section does not show the same level of vegetation.

This well-vegetated older moraine is considered to predate the stratigraphically superior 18th-century moraine by no more than a few centuries, despite its much older appearance. This would place it possibly in the 16th century and hence comparable to one found at the Bucher Glacier in the south central portion of the Juneau Icefield by Beschel and Egan (1965).

Other evidence of earlier Neoglacial advances is seen in a small bouldery moraine now being over-ridden by the 18th-century terminus in the form of a low-angle rock glacier. This older moraine extends 2 to 5 meters from beneath debris falling from the face of the advancing moraine. This older moraine consists of 1 to 2 foot (0.5 m) lichenized boulders of granodiorite resting on a substrate of sod-covered morainic material which seemingly represents an even older morainic limit (Figure 48). Samples of in-situ krummholz fragments, as seen in the photo, have been collected and are currently being processed for radio-carbon date (Geochron, 1975, in process).



Figure 48 Older Neoglacial moraine presently being overridden by advancing ice cored 18th Century moraine



Figure 49 Face of advancing ice cored 18th Century moraine

Another evidence is found in a small cobbly moraine edge peeping out from beneath the 18th-century or older lateral push-moraine on the north edge of the valley and somewhat above Goat Hill. The material in this moraine is rounded and stained, having the appearance of uprooted stream-bed materials.

It is concluded that the latter two morainic remnants represent Neoglacial advances prior to the 18th-century and are probably correlates of the 16th-century moraine described above.

The Extensive 18th-Century Advance

The most extensive Neoglacial advance to occur in this valley entirely overrode Met Hill and covered Goat Hill. It left high scour zones along the east flank of Splinter Peak and on the north side of Mt. Edward Little (see left side of Figure 53). It also left two or more successive pronounced lateral moraines along the north edge of the valley, extending from the Splinter Peak scour zone to the northeastern edge of Goat Hill.

The main tongue of this advance followed the topographic low along the southeast side of the valley and did not spread to the northeast around the base of Goat Hill as earlier advances had done. This suggests that the valley was fairly free of ice prior to this advance. It may also indicate lower temperatures resulting in less plasticity of the ice than was the case in earlier advances.

The moraine lobe appears to be still advancing along the gentle slope of the valley floor. This is evidenced by the following:

- a) a continuously fresh slip-off face on the eastern front quadrant (Figure 49). Fine particles are exposed on standing face, while boulders roll to the base of the slope.

- b) curved parallel ridges, 50 to 100 cm in height and 90 to 200 cm apart, on the upper surface of the moraine (Figure 47).

These ridges are interpreted as characteristic flow forms in a rock-glacier flow, and are caused by downslope creep largely due to flow deformation of interstitial ice. A striking example of rock-glacier movement reflected by these surface wave-ridges is shown on Torres Rock Glacier in Rock Glacier Valley just south of Cathedral Valley (Figure 50).

The upper surface of the southeast part of the 18th-century Cathedral Valley moraine is composed of large angular blocks, 60 to 150 cm across, with no fine materials present. With respect to this phenomena, Embleton and King (1968, p. 523-524) note:

"There is general agreement that rock glaciers have an upper crust of angular blocks without interstitial debris (though there may be ice) resting on a much thicker lower layer of angular blocks, sand, silt, and possibly mud (again, with or without ice)."

On this same topic Wahrhaftig and Cox (1959) point out that:

"Transverse ridges, conveying such a strong impression of glacier-like flow, are thought to represent wrinkling of the surface crust or even internal shearing as the motion of the rock glacier is slowed; they are probably not annual features such as glacier ogives."

Embleton and King suggest a rate of rock glacier flow from 0.75 to 1.5 m/yr. with the upper layer moving faster than the lower portion at the terminal face. Wahrhaftig and Cox give rates of surface motion as 1-2 m/yr. These figures are commensurate with those which have been found in recent years by JIRP personnel on the Atlin Rock Glacier (Miller and Swanston, 1975, personal communication).

The ice within the 18th-century moraine, which in effect is now a rock glacier, is relict, presumably retained from the initial advance



Figure 47 Terminal area of Late Neoglacial Moraines, showing Slump Zone
(photo by M. See)



Figure 50 Torres Rock Glacier in Rock Glacier Valley, Cathedral Massif



Figure 51 Lower Cathedral Glacier Valley from Camp 29

over two hundred years ago. This sub-surface ice has been protected from melting by the insulating effect of the surficial rock material. This surface material in the southeast quadrant is derived from talus and avalanching along the north flank of Mt. Edward Little, which here comprises the south wall of the lower Cathedral Glacier Valley.

The up-valley portion of this ice-cored moraine and its associated ice slope grades into the actual glacier surface along the southeast side of the presently receding terminus (Figure 51). As one proceeds down-valley along the glacier surface this ice slope gradually becomes mantled with rock debris from the mountainside. Excavation of a talus slope farther down-valley revealed solid ice at a depth of about 30 cm. The rock material proved to be almost entirely surficial, with none embedded in the underlying ice. At the time melt-water was saturating the mixture of fines and rock fragments, causing sliding and slumping over the ice surface exposed by excavation. This further indicates that much of the ice core in the terminal moraine is primarily relict glacial ice rather than interstitial ice from precipitation and/or melt-water from surface snow, though there is undoubtedly some refreezing of percolation waters between impervious till zones.

The terminal moraine segment to the north of Cathedral Creek, is morphologically different in surface character from the area south of the stream, but it apparently dates from the same advance. This view is supported by the lichenometric measurements. Here the down-valley margin does not display an active slip-off face, but except for slumping effects it fits geometrically into the same arcuate curve as the advancing southeast quadrant.

Boulders in the moraine quadrant southeast of Cathedral Creek tend to be much more angular than material north of the creek. The latter surface is more stable, with a large amount of mixed clay, silt and sand sized particles. Also here few small spruce, willow, and fireweed plants appear.

Exposure date of the southeast quadrant, estimated from lichen-ometric evidence, is 1789 ± 30 years. Exposure date of the area north of the creek is 1769 ± 30 years. Because of the greater stability of the latter area, the earlier date is probably more reliable.

19th and 20th-Century Advances

The probability of one or more advances in the 19th-century is strong, but the evidence has been obscured. The interpretations hinge on differentiation of surface features from the most recent advance from those of earlier advances, and also on recognition of out-of-phase time-stratigraphic relationships discussed in Chapter VII.

This main glacier lobe of this recent advance came forward while the south side of the valley was still occupied to some depth by 18th-century ice. Thus the advance was deflected to the northeast, with the spreading to the north below Goat Hill (see Figure 52).

There is a plexal zone in the area where the southeast margin of the ice in its later advance overrode the northwest side of the 18th-century advance (see Figure 53). Contiguous to this plexal zone are lobate ridges indicating internal flow in the 18th-century moraine (see Figure 47). The cause of this flow is considered later.

The youngest lobate moraine contains a shallow depression which lies behind the subdued terminal region. This depression contained a shallow pro-glacial lake until the terminal moraine was breached to a



Figure 52 Low Oblique Aerial Photo of Cathedral Glacier Valley and Camp 29 (photo by M. See)



Figure 53 Low Oblique Aerial Photo of Cathedral Glacier Valley and Late Neoglacial Moraines
(photo by M. See)

depth of two feet and the lake drained (see Figure 52). The lake persisted long enough, however, to form a wave-built terrace and to rework fines along the shoreline.

The surface of the youngest moraine displays a series of about 30 parallel stripes approximately 2 meters apart, and aligned with the direction of flow (see Figure 52). These stripes are composed of gravel and cobble sized clasts. Similar alignments are found up-valley where an ice advance rode part-way up the up-valley side of Goat Hill (Figure 55).

On this youngest down-valley moraine these furrows are found across most of the moraine surface, extending directly to the terminal edge. The lines are abruptly truncated at the up-valley end, just at the top of the last overridden major moraine below the southeast flank of Goat Hill (see Figure 52).

The most probable explanation of these alignments is that they are flutes in basal till, produced by the latest advance of ice over previous moraines. Such flutings also exist on the youngest moraines of Chapel Glacier (Figure 54). Till fabric studies (Solomon, 1974) indicate a pebble orientation with axes parallel to the trend of the flutings.

The youngest advance below Goat Hill in the northeastern sector terminates atop three or more similar small ridges. It is not clear whether these ridges represent: 1) oscillations of a single 20th-century advance; 2) basal thrust-surface deposits from this advance; or 3) terminal moraines of 19th-century advances.

The stratigraphic sequence in this location is tabulated as follows:

- 1) triple terminal ridges, resting on
- 2) a stratigraphically older moraine terrace, which in turn lies above

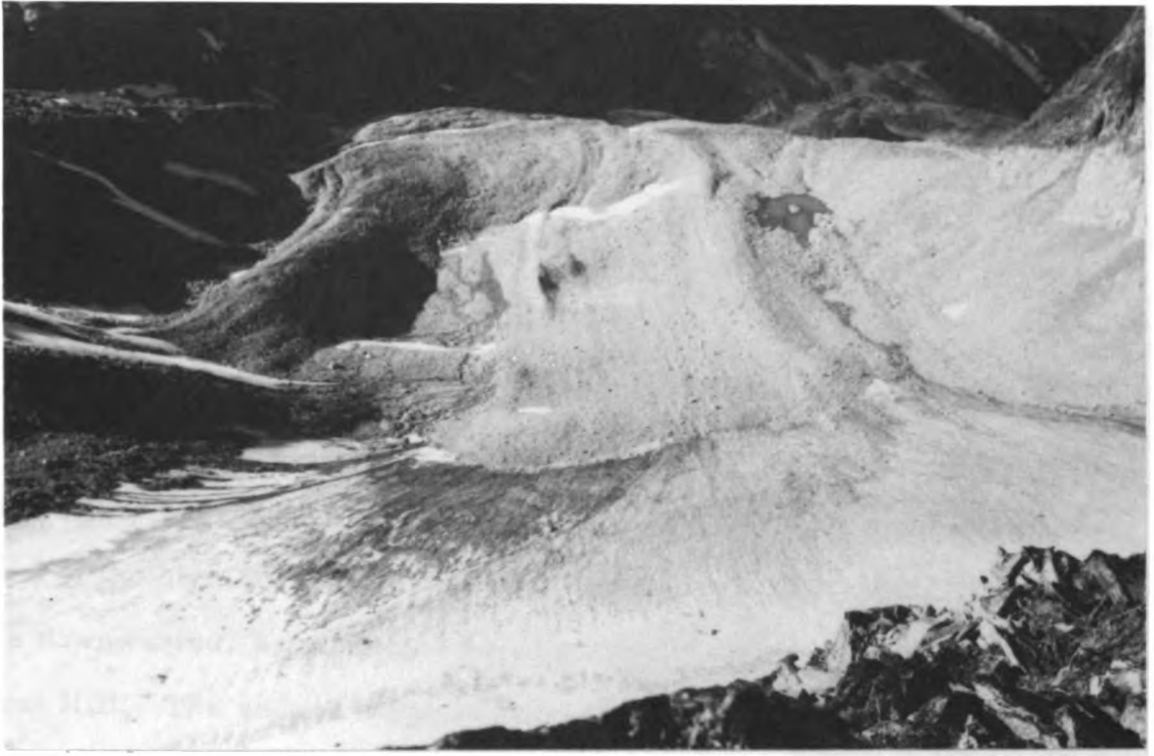


Figure 54 Late Neoglacial Moraines, Upper Chapel Valley



Figure 55 Goat Valley below Camp 29

- 3) the earliest, well-vegetated Neoglacial moraine previously discussed.

The upper limit of the youngest terminal moraine outside of the present terminus occurs rather abruptly at the northern corner of Goat Hill. This is precisely where the flutings are truncated (Figure 52). Up-valley toward the presently retreating terminus there is only a thin ground moraine.

Before the ice receded from the above noted terminal moraine, melt-water was dammed in the valley above Goat Hill. Geomorphic evidence shows that overflow passed north of Goat Hill and cascaded down the face of the high northern terminal moraine (see Figure 52). Then as the ice downwasted, water began flowing around the south shoulder of Goat Hill. The second stage of drainage persisted with development of an approximately 8-foot (2 m +) deep lake in Goat Valley. This is shown by strand lines, a spillway outlet stream channel and kame delta at the base of Met Hill.

As the ice further downwasted, water reached a lower point in the terminal moraine and eventually breached it. (Very recently this stream has cut a new notch some 10 meters to the south).

After the stream began flowing down the center of the ice-cored 18th-century moraine, melting of the moraine's ice core was accelerated beneath the stream. Flowing water is a much more effective heat exchanger than air, thus the central portion of the lobe, which contained the main stream channel, was thermally eroded enough so that a portion of the talus mantle became oversteepened at the angle of repose, making it quite unstable.

Recent Slump Features on the 18th-Century Moraine

After the melt-water stream cut its latest channel below Goat Hill it debouched into the lower center of the 18th-century moraine, with no organized outlet. Rather than pooling on the surface, this large volume of water percolated into the interior of the moraine. When the moraine became saturated a massive slump occurred, involving the entire frontal section of the moraine, with the exception of the southeastern (presumably ice-cored) sector. This slump zone is shown in Figure 47.

The internal portion of the moraine was compressed upon itself. This caused dewatering and a consequent subduing of the surface relief, largely due to the fluidity of saturated fines. As further evidence of this slump shear zones developed along the drier northern margin. Arrows in Figure 46b show the apparent forward movement and the scarp zone.

As this slumping movement removed support from the upper center portion this portion began moving forward. But as it again impinged on materials further downslope the forward motion was halted, with the inertia of the mass behind it causing compressional ridges. These ridges still remain as till garlands, festooned across the center of the main moraine tongue (Figure 47). Meanwhile the outflow water found a new channel along the base of the forward scarp, and flowed easterly to the edge of the moraine. The stream still follows this channel.

Below Goat Hill the area of the lower valley ice-covered by the most recent advance has virtually no vegetation indicating a maximum exposure of only 30 to 40 years. Morphological evidence suggests that the advancing ice in this sector was shallow and contained little surface material to let down onto the basal till. Although the terminal moraine produced by this advance is distinct it is not very large.

At the same time the main ice mass flowing out of the glacier's accumulation area covered the shoulder of Met Hill. It did not override the highest portion where the weather station and meteorological instruments are located. Botanical evidence suggests the area by the station was exposed less than 20 years ago. Only a few primary colonizers are found, including such vascular plants as fireweed (Epilobium latifolium) and dwarf willow. Primary lichen colonizers have not yet become established.

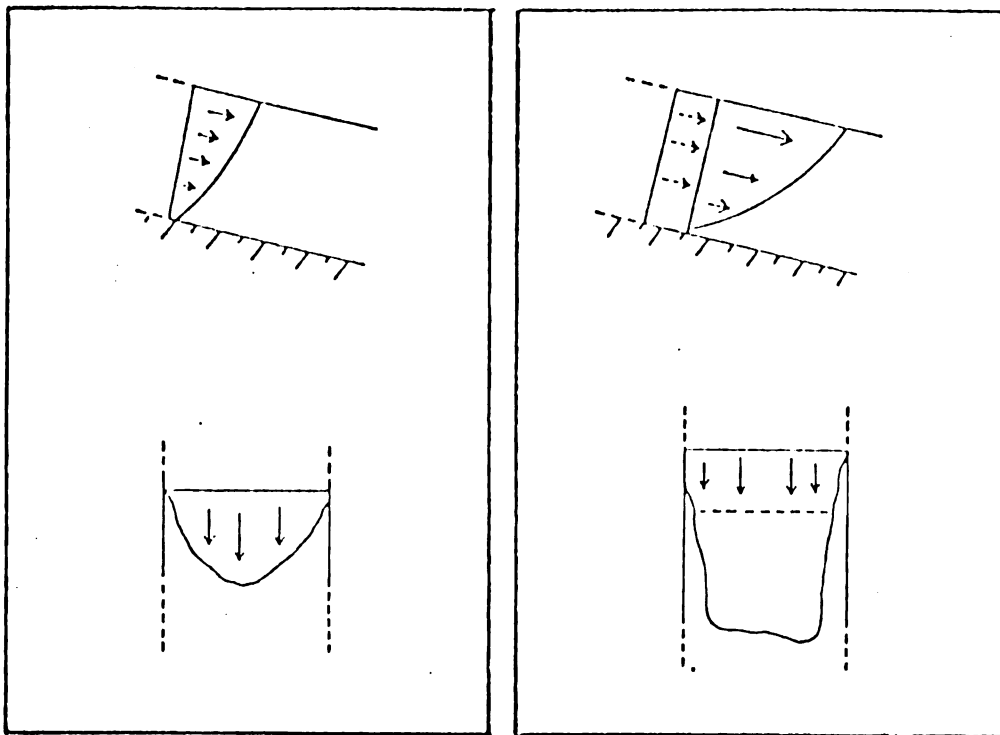
Recent ice impinged against the up-valley (south) side of Met Hill (see Figure 5). It also moved past the south side around the eastern base, and left a push moraine on the north side of the hill. This pulsation filled much of the valley between Met Hill and Goat Hill, and left fluting on the ground moraine as it slid well up onto the southwest side of the latter (Figure 55). This advance excavated the south shoulder of Goat Hill to bedrock before terminating at a level just below the top of the hill (see Figure 57). With this there was some unusual over-thrusting considered later in this chapter.

Mechanics and Modes of Glacier Flow

In order to clarify the flow mechanisms necessary for these interpretations, the mechanics of glacier flow are briefly considered.

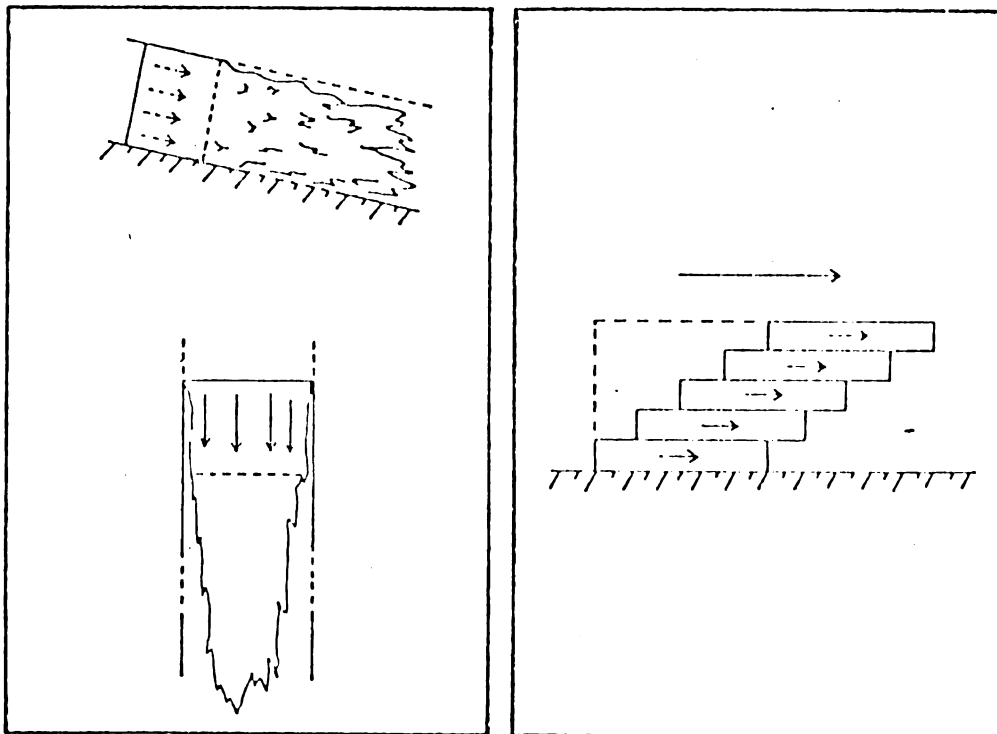
The mass transfer of a valley glacier may be conceptualized in two ways; the first involving mode of flow of the entire glacier, and the second involving discontinuous internal flow, expressed by discrete shearing and over-thrusting.

Glaciers may flow entirely by internal deformation as a quasi-plastic solid; i. e. parabolic flow. In this mode the margins and the base remain essentially in place, while the glacier slowly deforms in response to gravity (Figure 56a). If "viscosity" is low enough in proportion to



a) Parabolic flow

b) Rectilinear (plug) flow



c) Surging flow

d) Internal shearing

Fig. 56 Modes of Glacier Flow (a), b), and c) after Miller, 1973b)

stress, this can result in "caterpillar-track" -type overriding of surface material in its path.

If the down-valley gravitational stresses exceed the ability of glacial ice to deform, the entire ice mass may enter rectilinear or plug flow. In this mode it slides over its bed, excavating and polishing the bed and pushing materials ahead of it (Figure 56b). Most valley glaciers move by a combination of these two modes. Arcuate deformation of crevasse patterns, striation, bedrock polishing, basal till fluting, and the presence of push moraines all indicate that Cathedral Glacier advanced in this manner.

The third mode of glacier flow results when gravitational forces cause the shear stress on the glacier to exceed its ability to relieve the stress by either of the previous two modes. In this situation surging flow occurs, becoming chaotic and disorganized (Figure 56c), with extensive disruption along the lateral margins (Miller, 1973b). Obvious evidence for this type of flow is lacking in our study area. Indeed, it is improbable that Cathedral Glacier has ever approached the mass necessary for this level of stress to occur.

The second main type of glacier motion involves much shearing within the glacier, especially in the basal zone. In this there is not only basal sliding but actual slippage of one plane of ice against another (Figure 56d). To a limited extent this may also occur along the glacier margins during the first type of flow. However, the most significant mode in terminal or icefall areas, where there is compression involved, is basal shear. Greater stress at depth near a terminus or below an icefall, plus resistance of the basal material, may

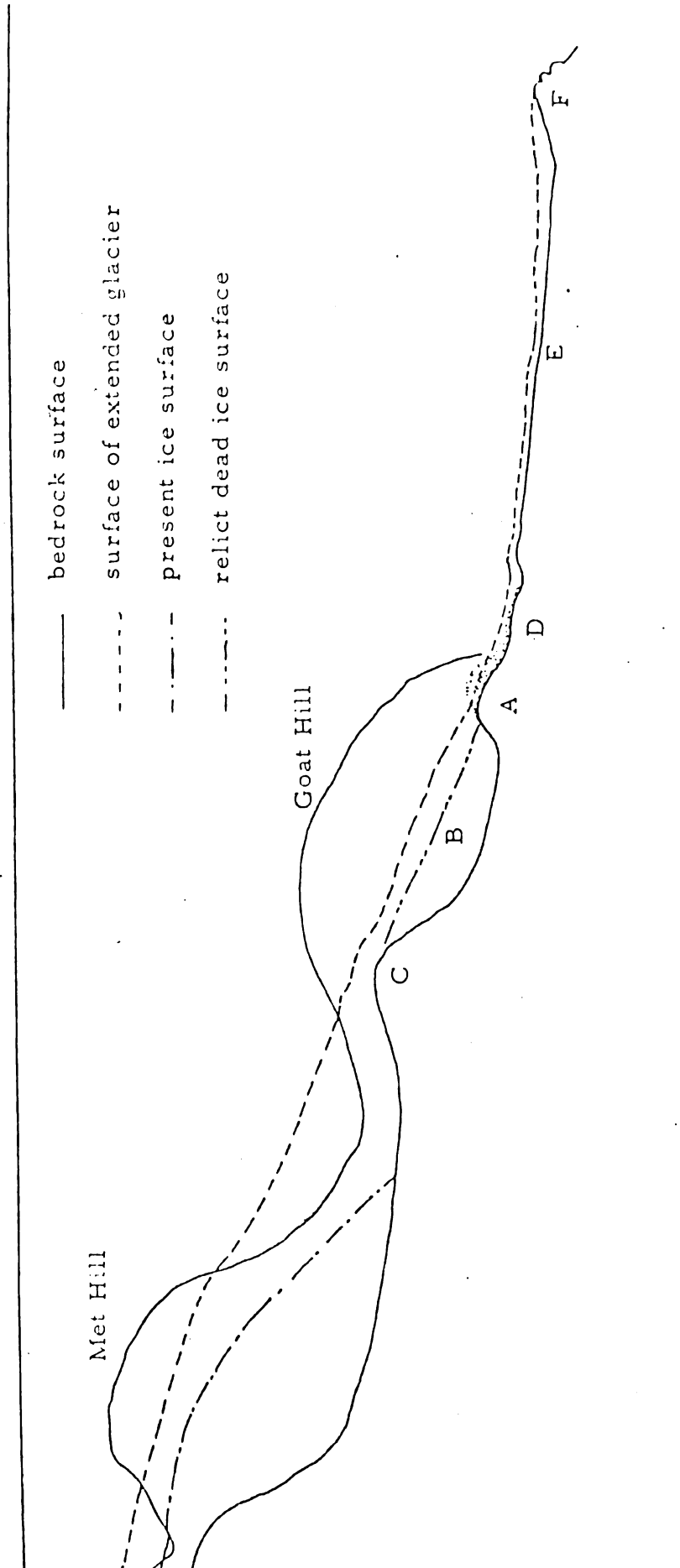


Fig. 57 Schematic Diagram of Overthrust Shearing Mechanism, Lower Cathedral Valley

give a resultant vectorial force causing the basal shear surfaces to slope at a low angle upward. In unconsolidated surface materials or as a result of rotational forces in a cirque basin this may also result in excavation of a depression behind a basal thrust moraine or even behind a bedrock threshold in a cirque. Such an origin is suggested for the formation of the depression in which the present proglacial lake lies, northwest of Met Hill.

A second mode of shearing involves forward movement of surface layers when the mass of sub-surface ice is blocked from advancing and/or protected from stresses from behind.

Exploration of a small marginal ice cave adjacent to the lower part of Met Hill revealed a roof layer of relatively fresh ice about 9 inches (23.5 cm) thick. This cleaner ice arched over an older dirty ice surface, leaving a free air space of 6 to 9 inches (15 -23.5 cm). This represents the over-shearing of new ice thrusting over older debris-entrained former surfaces. This newer ice may have sheared over the top of the cirque headwall just above. Miller (1975) reports the discovery of three such debris-entrained basal ice zones blocking and producing overthrusts on the lower Llewellyn Glacier, and similarly on the Mendenhall Glacier near Juneau.

The above overthrust mechanism is believed to have been in operation during recent advances in the lower valley. The proposed sequence is as follows (see Figure 57).

- 1) Advance to A of the entire glacier mass (dotted line on Figure 57 represents surface), forming the latest massive terminal moraine.
- 2) Still-stand and downwasting, with retention of dead ice in B.
- 3) Advancing stresses on the main glacial mass with the main body of ice blocked by the bedrock shoulder (C)

extending part-way across the valley. Thus the surface layers shear off and thrust over the dead ice in B.

- 4) Shearing off of the tip of the moraine at A, and redepositing it at D.
- 5) Continued advance of the thin overthrust sheet over area E to the terminus at F.
- 6) Downwasting of the overthrust sheet. Also extensive downwasting of the main ice mass, resulting in retreat, and with this recession continuing today.

It is suggested that stages 3, 4, and 5 may have occurred several times within the last two centuries. The blockage and shearing effect may have amplified the effects of the short-term climatic oscillation patterns noted in Chapter V.

Lateral Moraine Patterns

The northern margin of Cathedral Valley is dominated by bold lateral moraines, which are clearly shown in Figures 52 and 53.

In this series the two outer moraines are the most massive, but none are continuous from the terminal zone to the flank of Splinter Peak. A third distinct morainic ridge disappears up-valley in a plexal zone between the second and fourth main ridges. The second ridge is heavily lichen encrusted, including Rhizocarpon geographicum, Alectoria pubescens and Umbilicaria species. The outer (older) ridge is physiographically less dominant and also less lichenized. Perhaps the inhibition of lichen growth by snow-cover explains this.

The outer two lateral moraines may be related to the 18th-century advance. Because of the heavy lichen growth, further lichen studies are desirable to determine if these may indeed represent much earlier Neoglacial advances.

The terminal moraine zone on the north side of the valley consists of a steep embankment reaching from the northeast corner of Goat Hill across

both inner and outer moraine sequences to the outer margin where it abuts against the Wisconsin moraine shown in Figure 52. The face of this embankment is undergoing mass wasting, but does not appear to be advancing. The face of the inner moraines contain darker and more oxidized material than the outer moraine.

The interior face of the fourth and innermost lateral ridge is a chaotic slumped area of huge and blocky granite boulders. An ice core, insulated by the surface detritus, is continuing to melt in the area northeast of Met Hill, causing the inner face of the ridge to collapse.

The 1974 field investigations revealed that the inner portion of this fourth lateral moraine was not emplaced by ice from the upper cirque, but was impinged upon by ice coming cross-valley in the late 19th-century. The last significant advance after this left a low but distinct moraine along the left edge of the valley inside of this fourth ridge (Figure 55). The previous advance truncated the downvalley portion of the third and fourth lateral ridges. Most of the boulders affected were overturned and rearranged, leaving only a small proportion of lichen-colonized boulders still upright. A few boulders on top of the disturbed moraine were iron-stained, as though they had been partly buried in finer materials derived from nearby bedrock composed of iron-rich metavolcanics. This same advance overrode pockets of ice filling parts of the small valley on the north side of Goat Hill well above the terminal moraine. Subsequent melting left small till-free bedrock depressions. Also main outflow streams in this sector cut a notch in the terminal moraine close to Goat Hill, and cascaded down the moraine embankment, leaving abandoned stream channels on its face (see Figure 52).

The southeast flank of Splinter Peak above the present proglacial lake shows a prominent scour zone topped by a trim-line which curves

sharply down to the highest elevation on the outer lateral moraine (see Figure 53). Downvalley from this the lateral moraine slopes off to a much lower level as the valley wall curves away from contact with the moraine. The second lateral moraine at this location parallels the first and shows the same configuration.

Up-valley from the highest elevation on these lateral moraines they loop along the side of Splinter Peak (see Figure 53). During the rapid ablation of recent years the main body of the glacier has downwasted and left the edge of the glacier as an ice-cored lateral moraine, further insulated by debris sliding down from the mountainside. As this ice core continues to melt the slope becomes more and more unstable and the surface more chaotic. Frequent rock-slides take place in this zone on warm summer days as the sub-surface ice melts away. During these warm days a substantial stream pours out from the base of this slope, indicating the presence of its melting core of ice.

At the base of the slope two minor moraine ridges exist within the outer moraines described above. The first (outer) ridge has a reddish-brown color distinct from the other lateral ridges. This is concluded to be a protalus rampart, formed from weathered metavolcanic bedrock falling from the upper slopes of Splinter Peak. The material undoubtedly came from above the granodiorite contact at a time when the lower wall was covered by snow in a cooler period. In contrast, the granitic materials within the other ridges were transported down-valley by moving ice. The second (inner) ridge, about 80 cm high, is composed of a thin layer of fines, washed down from the up-valley lateral moraine, over a core of clear ice.

On the opposite margin of the glacier, the west slope of Mt. Edward Little displays a sharp "snow line", beneath which the Umbilicaria growth has been inhibited (Figure 44). Below this, but above the present glacier surface, slumping of unstable material reveals a slowly melting ice surface underneath one to two feet (30 to 60 cm) of ablation moraine.

The climatic events which formed this complex sequence of geomorphic features are discussed in the following chapter.

CHAPTER IX

LATE HOLOCENE EVENTS

TOWARD A NEOGLACIAL CHRONOLOGY

This section presents a provisional chronology indicated by the moraine sequences described in Chapter VIII. This represents a chain of Neoglacial climatological events which occurred in Cathedral Glacier Valley. This interpretation anticipates modification and refinement upon acquisition of further field information and as additional insights develop from further research by the writer and by future JIRP personnel.

The Chronological Sequence

Phase I: Pre-18th-Century Events

Between the end of the Thermal Maximum Interval (c. 2500-3000 years B. P.) and the major Little Ice Age advance lichenometrically dated as c. 1760, one or more Neoglacial advances reached approximately the same limits as the present Little Ice Age terminal moraines. Only marginal portions of moraines produced by these advances remain exposed, but they are distinct.

The date of 1670 or before is suggested through extrapolation of the 90-year glacial cycle shown in Figure 27. This may have represented a broad advance of temperate ice spreading widely across the valley, suggesting environmental conditions warm enough for increased plasticity of a temperate ice mass. The extent of such an advance is shown in Figure 46a. Alternatively, it could represent two distinctly separate advances. The first would be on the southeast side of the valley, similar

to the Phase II 18th-century advance. The second could have been deflected to the north around the base of Goat Hill by ice from the previous advance, as occurred with the Phase III and IV advances more recently. These ice masses had largely disappeared prior to the Phase II advance described below.

Phase II: The 18th-Century Maximum

The most extensive advance of late Neoglacial (Little Ice Age) time has been dated by See (1975) as having ended and the terminal moraine stabilized by 1760, $\pm$ 30 years. Prior to this advance the lower valley was relatively free of ice, permitting the next ice advance to follow the topographically lowest route downvalley.

Early in this advance ice filled the upper cirques to the present trim-lines, and completely overrode Met Hill and covered Goat Hill. On the north side of the valley the advance halted at the downvalley edge of Goat Hill, leaving the high moraine embankment. The south side limit is shown by the maximum downvalley extension of Neoglacial moraines.

Upon initial inspection the Phase II moraine appears to be stratigraphically superior to the younger Phase III moraine at the northern Phase II margin (see Figures 46b and c). However, the relatively abrupt northern corner of the Phase II moraine is inconsistent with the smooth arcuate form of a normal terminal moraine. This apparent anomaly is attributed to the forward slumping of Phase II morainic materials in the terminal zone, as discussed later.

About 1760 the advance became stable enough for lichen growth to develop. Due to the massiveness of the advance and the heavy cover of detritus from the flank of Mt. Edward Little, the ice downwasted slowly. This ice tongue possessed relief prominent enough to deflect later advances to be built against it. More recently its center section which lies beneath

the meltwater stream has downwasted, allowing mass wastage on the consequently oversteepened slopes to further modify the surface. The glacier with its surficial mantle of rock debris continues to move forward as a low-angle ice-cored rock glacier.

Phase III: The 19th-Century

A 19th-century advance is supported by time-stratigraphic evidence (Figure 46c) but without geobotanic information. It is inferred, on the basis of the apparent 90-year glacio-climatic cycle (see Figure 27) to have occurred circa 1840. Only two small terminal remnants of terminal moraine, visible at the corners of the Phase IV lobe, are ascribed to this moraine (Figure 46 c). At the time of formation the southeast lateral part of this advancing lobe formed a push-moraine ridge atop the northeast flank of the prominent Phase II materials. Photographic evidence also suggests two pulsations of this advance.

Phase IV: The Last Advance

In the most recent advance to occur in the Cathedral Glacier system, the ice tongue overrode the center of the Phase III terminal moraine (Figure 46 d). Its terminal ridge suggests three or more pulsations.

Apparently the Phase IV tongue was a shallow layer of relatively young ice with little entrained or surface debris. This layer sheared over the top of the main terminal glacier mass in the pocket alongside Goat Hill. Fresh and prominent basal flutings on the surface downvalley from this point clearly show the direction and extent of this ice flow.

20th-Century Warming

After an appropriate time lag following the onset of an unusually sharp warming trend early in the 20th-century the main glacier mass began to downwaste. As its terminus and margins receded a thin recessional moraine was left in Goat Valley. The bedrock berm on Met Hill, where

the research station now stands, has been exposed within the past 20 years, and continues to be uncovered. The changes in location of the meltwater stream and a subsequent slump in the 18th-century terminal moraine have been described in Chapter VIII.

A Transect Through History

As the investigator proceeds on foot from the lowest terminal moraine upvalley along the plexal zone left by the above glacial phases he encounters three prominent morainic hills. The first is characterized by three undulations and is the compressional zone of the massive slump which occurred in the Phase II moraine terminal area. The small declivity just upvalley represents the base of the slump scarp.

Above this is another hillocky prominence representing the right-lateral (southeast) portion of the Phase III and Phase IV moraines, produced by material moved along the left (northern) margin of the Phase II ice-cored moraine. This is the area shown on the low oblique aerial photo (Figure 53) as the most obvious plexal juncture.

The third and last morainic hill, above this plexal zone, is adjacent to the eastern corner of Goat Hill. This is the terminal moraine of the last short advance, and is postulated to have been formed in the 1920's or 1930's. The valley floor above this point is covered with thin recessional moraine interspersed with a few areas of fine outwash and kame deposits, indicating rapid retreat of the glacier terminus.

Present Conditions and the Mean Névé-line

Despite a recent downturn in regional temperature (since the 1940's) downwasting and terminal recession continue today in the terminal zone. During the period of maximum temperatures, which peaked in the early 1940's, increased snow accumulation and ice formation occurred in areas at the 6100-foot (1860 m) level which had not experienced much

glacial activity during colder periods. At present the maximum accumulation is in the upper cirque above the 5800-foot (1760 m) level. As we have seen in Chapter VI the mean névé-line in 1971-74 has lain approximately at the 5800-foot (1760 m) level.

The main process taking place in the lower moraine areas today is continued melting of the core ice, causing further oversteepening of slopes and additional slumping and rockfall. Downvalley movement of the detritus-mantled ice mass continues along the south side of the valley where the slope of Mt. Edward Little continues to provide insulating talus and shade from solar insolation. Forward movement (in the mode of a rock glacier) also continues in the southeast quadrant of the 18th-century terminal moraine.

Corners of the bedrock headwall on the relict cirque just below the research station continue to be exposed by downwasting. Also the present proglacial lake above Met Hill continues to expand in size in the upvalley direction as the upper north glacial terminus downwastes under the slightly negative conditions of today's mass balance.

Prognosis: The Next Quarter Century

Location and behavior of the semi-permanent névé-line indicate that downwasting and terminal retreat of Cathedral Glacier should continue for the next several years. Based on meteorological trends in the Atlin area and on the Alaskan coast, however, freezing levels will continue to lower. The elevation of maximum accumulation will very likely reach the lower cirque level within twenty years. By the turn of the century it may even be necessary to relocate the research station to the top of Met Hill, as the 40-foot-lower berm where it now stands may be overridden by ice advancing out of the upper cirque. Also, the mean névé-line should reach the threshold of the lower

cirque, or even below the level of the Camp 29 station, by the mid-1990's. This forecast is based on the lapse rate, extrapolating from the present freezing level at the 5800-foot mean névé-line. After this the névé-line should again retreat up-glacier. However, by then the flow lag should catch up and the entire ice mass may be advancing, with the terminus again pushing ahead. It is suggested that this advance will not reach the magnitude of that in the 1920's.

CHAPTER X

SUMMARY OF CONCLUSIONS

Conclusions drawn from this study are summarized by disciplinary categories.

Geomorphology

The primary control of the landscape configuration on the Cathedral Massif is the structure of the jointing system in the bedrock. Secondary control is cirque aspect (axial orientation), as in this area cirque and glacial valleys have formed only where bedrock structure has allowed ice to be retained on shaded north slopes.

Cirque elevations tend to show a fourfold sequence in independent valleys. The four dominant regional elevations are around 4800, 5200, 5600, and 5900 feet (see Figure 41). In the cirques which head Rock Glacier Valley, the upper three groups of elevations appear. This indicates control by freezing level (elevation of maximum snow accumulation).

Synoptic Meteorological Records for the Summer Field Season, 1972-74

Camp 29 (elevation 5300 feet) temperatures during the summer field season averaged 5° to 7° below those at Camp 30 at Atlin (2190 feet) but generally followed the same trends. Precipitation at Camp 29 during the summer field season was nearly the same in amount and generally similar in pattern to that at Camp 30 (Atlin) which lies 18 miles north.

The meteorological records at Camp 29 require synoptic comparison with other stations on the 130 mile transect across the Juneau Icefield from

Atlin to Juneau (see Figure 2 and Appendix A) and northwesterly 100 miles to Whitehorse, Y. T.

Climatology

A rather obvious conclusion is that climate has varied in the past in this region. It can reasonably be expected to vary again in the future.

Meteorological records from the Atlin region compared with the Juneau region show a sharp gradient of continentality inland from the coast. Mean annual temperatures for 1907 - 1973 in the interior averaged 10.5° F. cooler than on the coast, ranging from 25° to 36° F against coastal temperatures of 38° to 45°. Temperature trends on the coastal side of the Juneau Icefield closely follow patterns in the interior. Mean annual temperatures in both the coastal and interior areas follow a repetitive pattern between 13 and 17 (mean 15) years in length. Long-run average temperatures peaked in the early 1940's. Regression analyses show a distinct downward trend since that time, with less tendency to fit the earlier periodicity. The trend is toward a more pronounced temperature decrease in recent years.

Precipitation in the interior averages 11 inches per year, compared with 87 inches along the coast. Juneau City annual precipitation is representative, in pattern and amount, of average rainfall at sea-level along the Alaskan coast. Distinct patterns do not appear in precipitation records, except for a possible 7-year pattern on the coast.

Prior to the early 1940's short-term cycles of temperature (17 year) and precipitation (7 year) tended to be somewhat out of phase. Since that time they tend to be somewhat in phase. Longer-term periodicities of each tend to be in agreement. The nature of climatic patterns and relationships seems to have changed since the temperature maximum of the early 1940's. It may be that the dominance has shifted from one

atmospheric control to another, or that man's activities (atmospheric pollution) may have affected the climate (Miller, 1973 a). It may on the other hand be only a natural phenomena.

Glacio-Hydrology

Substantial runoff from Cathedral Glacier begins in early June and ceases with a mid to late-September freeze-up. Daily runoff from the glacier is directly correlated with higher ambient temperature, solar insolation and rainfall. Runoff follows a diurnal pattern. Water level in the glacier's proglacial lake tends to decrease in late afternoon. However, volume of flow at the stream gaging station below the Neoglacial moraines tends to peak and decrease before the level of the proglacial lake peaks and begins to fall.

A hydrological surge occurred in mid-August, 1974. This greatly altered the cross-sectional profile of Cathedral Creek at the gaging station, rendering current meter readings unreliable until a cross-section of the stream can be stabilized by weir or other structure.

Glaciology

The Cathedral Glacier regime remains in a negative mass balance, despite recent onset of a distinct regional cooling. Ablation and bare-ice exposures on the Cathedral Glacier at and above Camp 29 were much greater in 1974 than in 1972 or 1973, but the general trend appears to be toward cooling and lowering névé-lines. The present névé-line on the Cathedral Glacier has been around 5800 feet (1780 m). Lateral moraines on the mid and upper parts of the north valley margin are still ice-cored. Melting of this sub-surface ice causes much surface mass wastage and collapse of the moraines.

Changes in glacier flow patterns appear to be intimately related to the changes in directions of local storm winds. This in turn relates to

dependence on changes in location of the interface between low-pressure (cyclonic) and high-pressure (anticyclonic) circulatory cells along the North Pacific Coast of Alaska-Canada.

Glacial Geology

In Neoglacial time glacial advances into the lower Cathedral Glacier valley below Goat Hill consisted of four phases:

- a) an advance shortly previous to the 18th-century
- b) a major advance culminating about 1760
- c) an advance circa 1850
- d) an advance ending in the 1930's

All evidences of early Neoglacial advances have been largely obliterated by the Little Ice Age glaciation.

The timing of these advances agrees with the existence of an approximate 90-year cycle described in Chapter V. An 80 to 90-year glacial cycle has also been observed on the southern portion of the Icefield by Miller (1973). Langway et al (1973), from study of the Camp Century (Greenland) ice core, have found oscillations of 78 and 181 years. Dansgaard et al (1971), from Oxygen-18 values from this ice core show temperature minima at 1430, 1510, 1660, and 1820, with minor cooling around 1895. Assuming a half-cycle flow lag, the Greenland record can be correlated with the Cathedral chronology with the exception of the major advance of the Cathedral Glacier which culminated around 1760. The Camp Century record appears to bear no relation to Borgthorsson's history of Iceland's temperature for the same period (as reported in Alexander, 1974). Further study of the tropospheric ridge locations may provide a possible mechanism for those discrepancies.

Ice from the upper cirque tends to follow the fall-line and flow past the southeast side of Goat Hill to the lower valley. Similarly, ice from

the lower cirque tends to flow directly out of the lower cirque and across the valley between the two bedrock hills. Lateral push moraines on the north side of the valley, originally formed by upper cirque ice overriding Met Hill, were later truncated by lower cirque ice moving across the valley between Met Hill and Goat Hill.

Because of response time and flow lag, advances may be nearly a half-cycle out-of-phase with temperature patterns. They occur mostly during periods of increasing temperatures. Conversely, retreats appear to coincide with periods of falling temperatures. Despite a downturn in regional temperatures since the 1940's, the Cathedral Glacier terminus continues to retreat. However, as the elevation of the freezing level continues to lower in consequence of the regional cooling, the glacier's mass balance over the next decade will become more positive and the névé'-line will approach the level of the research station (ca. 5300 feet). The heavy accumulation from the upper cirque will continue to move downward, and may threaten the research station at its bedrock berm location on lower Met Hill before the turn of the century.

Atmospheric Circulation

In the region of the Juneau Icefield a persistent displacement of the mean position of the tropospheric pressure ridge, through a change in location and path of high and low pressure systems, will result in changes in mass balance of glaciers. Thus changes in former glacier regimes, as recorded by morainic systems in this glacier system and elsewhere in the Alaska-Canada Boundary Range, can provide fundamental clues to former regional and global climatic patterns.

CHAPTER XI

SUGGESTIONS FOR FURTHER RESEARCH

This study has represented only a beginning. As in any reconnaissance investigation in a new area, some of the conclusions must be provisional. The potentials for further research at this locale are unlimited, and the implications of scientific aspects brought into focus are exciting.

A few of the most pressing needs are suggested to carry forward the research efforts already underway. These are categorized in a sequence relating to the glacier system itself, the massif as a whole, and related aspects in the Atlin (Canada) and Taku (Alaska) districts. Finally, a significant teleconnectional study is suggested.

The Cathedral Glacier System

Cross-sectional and longitudinal bedrock profiles obtained by geophysical measurements in the presently ice-filled cirques of the Cathedral Glacier are needed to amplify a gravity survey started in 1972.

High precision glacier movement surveys, and the analysis of previous survey data are needed to determine ice flow at previously triangulated positions. Also needed are measurements of any compressional or tensional changes in the glacier's long profile.

Development of a local topographic map, amplifying a phototheodolite survey initiated in 1972, may be aided by more detailed topographic mapping, by plane table and/or photogrammetric or other means, of the moraine complex in the lower valley.

Continuation of meteorological observations is essential, with more detailed analysis of past and future records and observations and further comparisons with other stations in the Atlin and Taku districts, including the long-term field stations on the Juneau Icefield.

Geophysical investigation should be made of the depth, magnitude, extent, and behavior of the ice-cored moraines in the lower Cathedral Glacier Valley.

More precise measurement of seasonal névé'-lines, and more detailed definition of seasonal vs. semi-permanent névé'-lines, will lead to determination of the mean névé'-line position for the 1971-1981 decade.

Measurement over a period of time of freezing levels (elevation of maximum accumulation) at different seasons and under different conditions, detection if possible, of any secular trends.

Consistent ablation stake measurements should be made at precisely the same locations year after year.

More detailed hydrologic and mass balance investigation of the Cathedral Glacier system will include study of evaporation losses.

More precise determination of the elevations and gradients of lichen trim-lines on Mt. Edward Little and Splinter Peak will aid in evaluation of alternative concepts of flow mechanisms discussed in Chapter VII. Mapping of relative ice movement out of the two adjacent ice-filled cirques of the upper Cathedral Valley will also aid in this.

More detailed lichenometric research, especially on the Phase III moraine segments and on the first and second left lateral push moraines above Goat Hill will provide more accurate dating of the chronological sequence.

Further investigation of the jokulhlaup phenomenon on the Cathedral Glacier, including exploration and mapping of any englacial caves

serving as hydrological channels or reservoirs, is needed to explain the hydrologic surge observed in 1974 and clarify anomalies in stream gaging records.

Detailed investigation of the sinuosity and vertical dimensional geometry of supra-glacial streams on the glacier surface will lead to further understanding of dynamic fluvial geomorphology.

Further englacial strain-rate measurements will amplify the initial micro-strain studies carried out by Warner in 1972 (Warner, 1973).

The Cathedral Massif

Lithological and structural mapping of the massif is needed. The features interpreted in Chapter II as faults should be further studied.

Low-level aerial photography and ground-truth studies of Neoglacial moraines on the other glacier systems of the Cathedral Massif will aid in analysis of the Cathedral moraines and provide a more definite chronology.

More detailed investigation of cirque-levels and cirque-floor positions will assist in evaluating the concept of freezing level as a geomorphic control.

More precise positional and elevational control is required on the Cathedral massif, via an enlarged survey network using high precision survey instruments. In this there is need for accurate base-lines and key bench marks for future reference in the long-term research activities planned for this area.

The Juneau Icefield Region

Continued monitoring of trends in mean annual temperatures is required at coastal and inland stations. In view of the findings of this study regarding trends to date, the next three to five years will be especially significant.

Study of temperature patterns should be done on a seasonal scale. For example, the significance of an exceptionally cold summer in the interior or an unusually dry autumn on the coast should be investigated.

Continuing acquisition of synoptic weather data from all stations on the Juneau Icefield and at Camp 29 on the Cathedral Glacier will amplify the significance and teleconnectional value of the Camp 29 and Camp 30 records.

Further study of available records is needed to reveal the relationship of temperature and precipitation trends to tropospheric ridge positions and to the dominant tracks of cyclonic storms and mean long-term shifts in location of the interior anticyclonic systems.

Teleconnectional Investigations

Further teleconnectional study is required to compare the glacier regimes on the Cathedral massif with those of cirque glaciers in other locations.

There is a need for teleconnectional studies of tropospheric waves and the zonal index. If patterns can be established, study of the Juneau Icefield region can be of increased value with respect to long-term climatological implications on a continental scale.

Climate is a primary control on crop production and thus on the provision of food for the world's people in the present and future. It is also a major factor in energy and environmental problems. The cooling trend now taking place in the polar regions affects precipitation regimes in the hunger belts of the world as well as in the temperate zones. For example, expansion of the northern polar zone causes a southward displacement of the Ferrel circulation cell and thus of the Hadley cell, which in turn suppresses the seasonal northward movement of the monsoon. The result has been drouth and hunger in the northern

edge of the monsoon regions. Expansion of the circumpolar vortex also causes longitudinal displacements of the mean ridge and trough locations in the tropospheric westerlies, with attendant effects on temperature and precipitation in food-producing regions such as the United States corn belt and the U. S. - Canadian and Russian wheat belts.

The Juneau Icefield is located on the crest of the main North American tropospheric ridge. Thus monitoring and analysis of relative climatic conditions on the maritime vs. interior flanks of the Icefield provide an unusually sensitive means of monitoring tropospheric behavior, as well as reconstructing climatic history.

A Final Word

It is hoped that this study will help to encourage further field research in this very interesting area, as well as to help provide a foundation for more detailed future work. This has been a long journey of the mind. In scientific research the significance may lie as much in the journey as in the intended destination. Indeed, the report of this study represents not a destination but a beginning. The writer looks forward with anticipation to interaction with those who may at times continue the journey with him.

The fact that the Cathedral massif lies in the new Atlin Wilderness Park of the Province of British Columbia makes the study of even more interest. It is further significant the Cathedral Glacier and the nearby portions of the Juneau Icefield lie at the ultimate headwaters of the Yukon River, a region of much hydrological and environmental significance for the future. In this the Foundation for Glacier and Environmental Research, with its long-term Juneau Icefield Research Program, will continue to make useful contributions to science and to the training of field personnel.

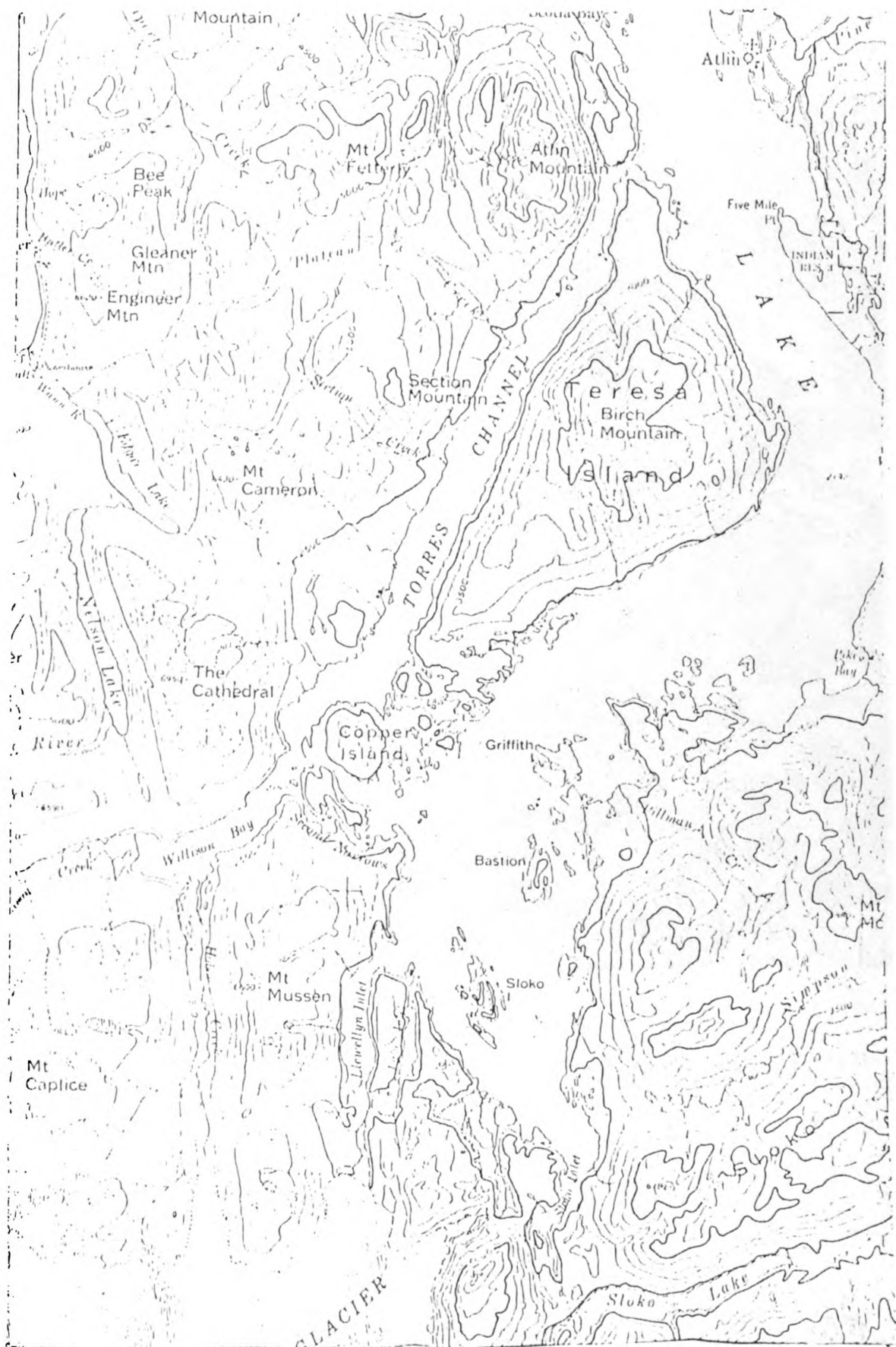


Fig. 58 Topographic Map of the Atlin - Cathedral Region



Fig. 59 Aerial Photo of the Cathedral Glacier System. July 14, 1975
(photo by M. M. Miller, F.G.E.R.)

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APPENDICES

APPENDIX A
GLACIO - METEOROLOGICAL STATIONS

The following are all fully - equipped weather stations in the Juneau Icefield Research Program, with permanent buildings and other facilities for related glaciological research activities. They are occupied all or part of the summer field season.

- Camp 8 Elevation 6300 feet; on western flank of Mt. Moore, overlooking the upper Matthes Glacier just below the international boundary and divide between the southward-flowing Taku Glacier system and the northward-flowing Llewellyn Glacier; sub-maritime to sub-continental climate
- Camp 9 Elevation 5200 feet; on upper Matthes Glacier; sub-maritime to sub-continental climate
- Camp 10 Elevation 4000 feet; on Taku Glacier; maritime to sub-maritime climate. Main central Icefield camp
- Camp 16 Elevation 5000 feet; on intermediate elevation ne've' of Taku Glacier; sub-maritime interior Icefield climatic conditions
- Camp 17 Elevation 4300 feet; on Lemon-Ptarmigan Glacier system; intermediate elevation, maritime climate
- Camp 18 Elevation 5600 feet; on joint ne've' of upper Matthes and Vaughan Lewis Glaciers, beside Vaughan Lewis icefall; sub-maritime to sub-continental high-level conditions
- Camp 19 Elevation 3800 feet; above Gilkey Trench, just below Vaughan Lewis icefall; sub-maritime conditions
- Camp 25 Elevation approximately 7000 feet; on Mt. Nesselrode and the broad ne've' plateau of the Bucher Glacier, part of which also flows into the Llewellyn Glacier drainage; high-elevation continental and Polar conditions
- Camp 26 Elevation 5000 feet; intermediate elevation on Llewellyn Glacier; sub-continental to continental climate
- Camp 29 Elevation c. 5200 feet; in permafrost zone on the Cathedral massif, near terminus of Cathedral Glacier system and

- Camp 29 Elevation c. 5200 feet; in permafrost zone on Cathedral massif, near terminus of Cathedral Glacier and west of Torres Channel, Atlin Lake; continental, semi-arid, sub-Polar conditions
- Camp 30 Elevation 2200 feet; on Atlin Lake, 130 miles north of Camp 17, on the inland flank of of the Boundary Range; continental, semi-arid conditions. Main interior base camp

APPENDIX B

GLOSSARY

Ablation

The combined processes by which a glacier wastes.

Aspect

The direction which a slope faces. Controls exposure to solar insolation.

Adiabatic

A situation in which no heat energy enters or leaves a system

Adiabatic lapse rate

The change in atmospheric temperature with altitude. The lapse rate is $5.5^{\circ}\text{F}/1000$ feet in dry air and $3.0^{\circ}\text{F}/1000$ feet in moist air.

Berm

A bedrock shoulder or terrace, originating from glacial erosion.

Cirque

A deep, steep-walled recess or amphitheater in a mountain, caused by glacial erosion combined with mass wastage processes in intraglacial times.

Continentality

Climatic conditions characterized by little rainfall, low frequency of cloudiness, and large diurnal and annual ranges in temperature. Usually associated with inland areas.

Continentality gradient

The gradual change of climatic conditions, over a geographical transect, from maritimity to continentality.

Equilibrium line

A line connecting points on a glacier surface where the net balance at the end of a balance year is zero. May include superposed ice (refrozen melt) below the seasonal neve' line.

Firn

Compacted, granular material in transition from old snow to glacier ice. Density is from 0.50 to 0.75 g-cm^{-3} .

Glaciated

A surface which has previously been glacierized.

Glacierized

A surface presently covered with glacial ice.

Katabatic wind

Cold air flowing down or down from a glacier surface; caused by the convection of dense cold air.

Lichenometry

A technique for determining the date of exposure of a rock surface from the maximum diameter of lichen.

Little Ice Age

The period of locally colder temperatures, harsher climate, and glacial advances from about 1300 A. D. to the present; followed the warmer period which occurred from the 9th through the 13th centuries.

Maritimity

Climatic conditions characterized by heavy rainfall, cloudiness, and small diurnal and annual ranges in temperature. Usually associated with coastal areas.

Mass balance

The net accumulation or net wastage of a total glacier system during a budget year.

Moraine

A mass of unsorted earth and rock debris transported by a moving glacier and deposited as the glacier melts.

Neoglacial

The period of colder temperatures, harsher climates, and reglaciation from c. 2500-3000 years before present to the present; followed the Thermal Maximum interval.

Ne've' line

The upper limit of bare glacial ice. The seasonal ne've' line is defined for the current ablation season in a given year. The semi-permanent ne've' line is defined as the lowest for the previous ten years.

Regime

The long-term state of health of a glacier. The material balance (budget) of a glacier involving the total accumulation or gross wastage in one or more budget years. The total matter - energy relationship of the glacier, involving the combined mass, liquid, and thermal budget.

Terminal moraine

A moraine formed across the course of a glacier, marking the termination of glacial advance.

Terminus

The outer or distal margin of the lower ablation area of a glacier.

Zonal index

The average pressure difference at sea level between the 35° and 55° latitude circles. This indicates the mean westerly strength of air flow. A high zonal index (8-15 millibars) is related to westerly winds. A low index is less than 3 mb. Indices from 0 to -5 mb indicate an easterly component to wind flow.

Zonal index cycle

The transition from high index to low and back to high. The cycle normally takes 4 to 6 weeks, but may last from 2 to 8 weeks. The cycle lacks a definite periodicity. The index may be high in one portion of the circumpolar pattern and low in another part at the same time. The cycle is less distinct in summer. A high index is characterized by an intense circumpolar vortex which is displaced northward of normal, and long planetary waves with few troughs. A low index relates to a circumpolar vortex expanded to lower latitudes; a higher number of planetary waves, with strongly developed troughs and ridges; and a vigorous north-south exchange of air masses. This causes large negative temperature departures in lower latitudes and positive departures in higher latitudes.

Information regarding the zonal index and index cycle taken from Chang, Jen-hu (1971). Atmospheric Circulation Systems and Climate. Oriental Publishing Company, Honolulu. pp.149-153.

APPENDIX C

RELATIVE MEAN SUNSPOT NUMBERS (Rz)

As supplied by Swiss Federal Observatory, Zurich, Switzerland,
Published in Monthly Weather Review, Vol. 30 et seq.;
and also from World Data Center for Solar-Terrestrial
Physics, NOAA*

Year	Mean (Rz)	Year	Mean (Rz)	Year	Mean (Rz)
1749	80.9	1791	66.6	1833	8.5
1750	83.4	1792	60.0	1834	13.2
1751	47.7	1793	46.9	1835	56.9
1752	47.8	1794	41.0	1836	121.5
1753	30.7	1795	21.3	1837	138.3
1754	12.2	1796	16.0	1838	103.2
1755	9.6	1797	6.4	1839	85.8
1756	10.2	1798	4.1	1840	63.2
1757	32.4	1799	6.8	1841	36.8
1758	47.6	1800	14.5	1842	24.2
1759	54.0	1801	34.0	1843	10.7
1760	62.9	1802	45.0	1844	15.0
1761	85.9	1803	43.1	1845	40.1
1762	61.2	1804	47.5	1846	61.5
1763	45.1	1805	42.2	1847	98.5
1764	36.4	1806	28.1	1848	124.3
1765	20.9	1807	10.1	1849	95.9
1766	11.4	1808	8.1	1850	66.5
1767	37.8	1809	2.5	1851	64.5
1768	69.8	1810	0.0	1852	54.2
1769	106.1	1811	1.4	1853	39.0
1770	100.8	1812	5.0	1854	20.6
1771	81.6	1813	12.2	1855	6.7
1772	66.5	1814	13.9	1856	4.3
1773	34.8	1815	35.4	1857	22.8
1774	30.6	1816	45.8	1858	54.8
1775	7.0	1817	41.1	1859	93.8
1776	19.8	1818	30.4	1860	95.7
1777	92.5	1819	23.9	1861	77.2
1778	154.4	1820	15.7	1862	59.1
1779	125.9	1821	6.6	1863	44.0
1780	84.8	1822	4.0	1864	47.0
1781	68.1	1823	1.8	1865	30.5
1782	38.5	1824	8.5	1866	16.3
1783	22.8	1825	16.6	1867	7.3
1784	10.2	1826	36.3	1868	37.3
1785	24.1	1827	49.7	1869	73.9
1786	82.9	1828	62.5	1870	139.1
1787	132.0	1829	67.0	1871	111.2
1788	130.9	1830	71.0	1872	101.7
1789	118.1	1831	47.8	1873	66.3
1790	89.9	1832	27.5	1874	44.7

* World Data Center for Solar-Terrestrial Physics
Environmental Data Service, NOAA
Boulder, Colorado, USA 80302 (313) 499-1000 Ext. 6467

<u>Year</u>	<u>Mean (Rz)</u>	<u>Year</u>	<u>Mean (Rz)</u>	<u>Year</u>	<u>Mean (Rz)</u>
1875	17.1	1916	55.4	1957	189.8
1876	11.3	1917	103.9	1958	184.6
1877	12.3	1918	80.6	1959	158.7
1878	3.4	1919	63.6	1960	112.3
1879	6.0	1920	38.7	1961	53.9
1880	32.3	1921	24.7	1962	37.6
1881	54.3	1922	14.7	1963	27.9
1882	59.7	1923	5.8	1964	10.2
1883	63.7	1924	16.7	1965	15.1
1884	63.5	1925	44.3	1966	46.9
1885	52.2	1926	63.9	1967	93.7
1886	25.4	1927	69.0	1968	105.9
1887	13.1	1928	77.8	1969	105.6
1888	6.8	1929	65.0	1970	104.2
1889	6.3	1930	35.7	1971	66.6
1890	7.1	1931	21.2	1972	68.9
1891	35.6	1932	11.1	1973	38.0
1892	73.0	1933	5.7	1974	34.5
1893	84.9	1934	8.7	1975	18.3(proj.)
1894	78.0	1935	36.1		
1895	64.0	1936	79.7		
1896	41.8	1937	114.4		
1897	26.2	1938	109.6		
1898	26.7	1939	88.8		
1899	12.1	1940	67.8		
1900	9.5	1941	47.5		
1901	2.7	1942	30.6		
1902	5.0	1943	16.3		
1903	24.4	1944	9.6		
1904	42.0	1945	33.1		
1905	63.5	1946	92.5		
1906	53.8	1947	151.5		
1907	62.0	1948	136.2		
1908	48.5	1949	135.1		
1909	43.9	1950	83.7		
1910	18.6	1951	69.4		
1911	5.7	1952	31.4		
1912	3.6	1953	13.8		
1913	1.4	1954	4.4		
1914	9.6	1955	37.9		
1915	47.4	1956	141.7		

APPENDIX D

MEAN ANNUAL TEMPERATURES AND TOTAL ANNUAL PRECIPITATION AT
 SELECTED STATIONS, SOUTHEAST ALASKA AND CONTINENTAL INTERIOR

Year	Juneau (city)			Juneau Airport			Atlin			Whitehorse Airport			Carcross			Coast			Interior		
	T	P		T	P		T	P		T	P		T	P		T	P		T	P	
1907	44	67					30	11								43	66		30	11	
1908	43						32	11								42	65		32	11	
1909	40	74					26	13					24	11		39	74		25	13	
1910	42	45					28	12					28	12		41	45		23	12	
1911	42						30	12								41	40		30	12	
1912	44	79					33	8					32	3		43	79		31	8	
1913	43	81					32	13								42	81		31	13	
1914	41	75					31	10					29	3		40	75		30	10	
1915	44	77					36	3					33	5		43	77		35	3	
1916	42	80					30	7					28	4		41	80		29	7	
1917	40	107					29	13					26	7		39	107		23	13	

Year	Juneau (city)		Juneau Airport		Atlin		Whitehorse Airport		Carcross		Coast		Interior	
	T	P	T	P	T	P	T	P	T	P	T	P	T	P
1918	42	100			31	11			28	3	41	100	30	11
1919	41	33			30	14			27	10	40	33	29	14
1920	41	37			30	9			28	5	40	37	29	9
1921	42	35			31	14			29	10	41	35	30	14
1922	41	32			31	8			28	6	40	32	30	8
1923	44	98			34	14			32	10	43	93	33	14
1924	42	100			32	11			29	11	41	100	31	11
1925	42	86			32	11			30	10	41	86	31	11
1926	46	31			37	11			35	6	45	81	36	11
1927	41	63			31	9					40	63	30	9
1928	43	39			33	10					42	89	32	10
1929	43	90			35	8					42	90	34	8
1930	42	97			32	19					41	97	31	19
1931	43	99			35	10					42	93	34	10
1932	41	36			29	10					40	95	23	10
1933	39	33			27	13					38	83	26	13

Year	Juneau (city)		Juneau Airport		Atlin		Whitehorse Airport		Carcross		Coast		Interior	
	T	P	T	P	T	P	T	P	T	P	T	P	T	P
1934	43	79			34	11				10	42	79	32	11
1935	43	90			31	15			29	11	42	90	29	15
1936	42	104			35	13			29		41	104	32	13
1937	43	90			32	11			29	7	42	90	31	11
1938	43	104			33	9			29	12	42	104	31	9
1939	42	120			34	14			31		41	120	32	14
1940	44	75			34	11					43	75	30	11
1941	44	75			34	10					43	75	33	10
1942	44	37	43		34	14					44	87	33	14
1943	43	118	44		34	5	33	11			44	118	34	8
1944	45	32	42		36	10	35	14			44	32	36	12
1945	42	39	44		33	8	31	10			43	39	32	9
1946	42	36	41	53			31	11			42	36	32	11
1947	43	102	41	61			33	7			42	102	34	7
1948	41	103	39	64			28	14			40	103	30	14
1949	41	99	39	62			31	10			40	99	32	10
1950	40	62	38	39			27	9			39	62	29	9

Year	Juneau (city)		Juneau Airport		Atlin		Whitehorse Airport		Carcross		Coast		Interior	
	T	P	T	P	T	P	T	P	T	P	T	P	T	P
1951	41	67	38	38			23	7			40	67	29	7
1952	43	103	40	66			32	13			42	103	33	13
1953	44	104	42	55			32	14			43	104	33	14
1954	43	82	40	41			31	9			42	82	32	9
1955	41	85	38	49			23	7			40	35	29	7
1956	43	101	33	63			29	10			41	101	30	10
1957	45	67	41	40			33	8			43	67	34	8
1958	45	91	42	52			32	8			44	91	33	8
1959	43	103	40	56			30	10			42	103	31	10
1960	45	100	42	53			32	9			44	100	33	9
1961	43	121	41	63			29	12			42	121	30	12
1962	43	100	40	62			31	13			42	100	32	13
1963	43	99	41	57			31	11			42	99	32	11
1964	41	107	40	58			29	11			41	107	30	11
1965	42	79	39	48			30	8			41	79	31	8
1966	41	97	37	58			27	11			39	97	28	11

Year	Juneau (city)	Juneau Airport	Atlin	Whitehorse Airport	Carcross	Coast	Interior		
1967	43	34	30	30	10	42	34	30	10
1968	43	83	31	30	12	41	83	31	12
1969	43	86	31	30	10	42	86	31	12
1970		89	31	31	10	40	99	31	11
1971			27	28	10	39	73	28	10
1972				25	11	33	39	26	11
1973				27	9	39	76	28	9
1974									

APPENDIX E

STATISTICAL ANALYSIS OF MEAN ANNUAL TEMPERATURES, COAST AND CONTINENTAL INTERIOR, 1907 - 1973

A test for statistical significance was applied to the data for the segments with increasing temperatures (1907 - 1943) and thesegments with decreasing temperatures (1944 - 1973) of the coast and the interior temperatures. The test used was the F-test of the form:

$$F_{1, n-2} = \frac{r^2}{1 - r^2} (N - 2)$$

Results are tabulated as follows:

<u>Location</u>	<u>Period</u>	<u>N</u>	<u>Change, °F</u>	<u>R</u>	<u>r²</u>	<u>Sig.</u>
Coast	1907-43	37	+1.08	.24	.06	<.05
Coast	1944-73	30	-2.07	-.39	.15	.05
Interior	1907-43	37	+2.69	.35	.12	<.05
Interior	1944-73	30	- 4.06	-.57	.33	.05

The results of this statistical test must be interpreted with some caution. First, changes from one year to the next may be of greater magnitude than the entire change over the period of the regression. Second, the slope of a regression line applied to oscillatory data is conditioned by the stage of the oscillation at the beginning and ending of the period of analysis. This is especially true when long-term trends are of smaller magnitude than the variations contained in the short-term oscillations.

In summary, the change in both coastal and interior temperature trends over the period of the study are statistically significant at the 95% level for the decreasing temperatures from 1944 through 1973. Temperature increase in the interior from 1907 through 1943 falls just short of the 95% level of significance, while temperature increase along the coast during that period is definitely not statistically significant.

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