THE GLACIAL AND PERIGLACIAL GEOMORPHOLOGY OF THE FOURTH OF JULY CREEK VALLEY, ATLIN REGION, CASSIAR DISTRICT, NORTHWESTERN BRITISH COLUMBIA

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This is to certify that the

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ABSTRACT

THE GLACIAL AND PERIGLACIAL GEOMORPHOLOGY OF THE FOURTH OF JULY CREEK VALLEY, ATLIN REGION, CASSIAR DISTRICT, NORTHWESTERN BRITISH COLUMBIA

By

Ann M. Tallman

The Atlin Region of northern British Columbia lies at the northern end of the Wisconsinan Cordilleran Ice-sheet in an area affected by Wisconsinan glaciation from distinctly different provenances. The prime nourishment areas were the Coast Mountains and the adjacent Yukon and Stikine Plateaus in Northern British Columbia. In late stages of deglaciation during warming climatic conditions local ice centers contributed their share.

The study site, the Fourth of July Creek Valley, is situated in an interlobate region where ice from the Cassiar Range and the Boundary Range in out-of-phase fluctuations modified the landscape. Morpho-stratigraphic evidences reveal multiple glaciation sequences identifiable back to the early Wisconsinan. Wisconsinan glaciation was very extensive in the Atlin region and it's earliest phases covered all but a few of the highest ridges and peaks in the region. Evidence for pre-Wisconsinan glaciation is by inference from erosional

remnants on the high ridges of the Boundary Range and from the presence of early drifts well outside of the study area.

A Wisconsinan chronology is presented in which an early Wisconsinan phase and middle Wisconsinan (40,000 ∓ years B. P.) intraglacial are invoked, followed by a middle to late Wisconsinan sequence of two main stages and three lesser ones with intervening intraglacials. These are termed Atlin I, II, III, IV, and V for ice from the Boundary Range and Gladys I, II, III, and IV for ice from the Cassiar source.

The evidence is based on glacial stage positions and an array of related depositional sequences of terminal, lateral and ablation moraines, crevasse fillings, eskers, kames, outwahs trains, glacial-fluvial terraces and drainage patterns resulting from the down-wasting of ice from both provenances. The relative ages of weathering are discussed and inferences attempted with respect to intraglacial conditions. A tentative correlation is made with the glacial stratigraphy and chronologies reported from adjoining regions. Special attention is given to the chronology of the Juneau Icefield and Taku District in Alaska which was the major accumulation area for the Boundary Range ice. In addition, a broader correlation is suggested with the well-developed North American mid-continent chronology.

The evidence presented suggests that global climatic events which led to the major Pleistocene fluctuations, though somewhat out-of-phase, are evidenced in the study area. Where there are significant differences they are considered relative to determining geographical and orographical factors. It is shown that minor fluctuation in mid-continent glaciations do not always appear in this Cordilleran chronology because the high latitude (sub-Arctic) and mountainous terrain tends to produce glaciations earlier than in mid-continental areas, and to retain them for longer periods during intraglacial stages of climatic amelioration.

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CHAPTER I

LOCATION AND PHYSIOGRAPHIC CHARACTER OF THE STUDY AREA

Introduction

Studies of the geomorphology and glacial history of the northern Cordilleran region in British Columbia and the Yukon have been made primarily in connection with the Geological Survey of Canada's natural resources and geology research program (Cairnes, 1913; Kerr, 1934; 1936; Aitken, 1959). Other workers have availed themselves of the logistic advantage of the Alaskan Highway (Fig. 1) to conduct reconnaissance studies of the glacial history along geographically limited sectors of this transect to the north and east of the Atlin region (e.g. Denny, 1952).

The present dissertation concerns such an investigation in a specific locale within this large region where only the most general aspects of Pleistocene glacial stratigraphy have been outlined (Johnston, 1926; Watson and Mathews, 1944; Armstrong and Tipper, 1948; Denny, 1952; Wheeler, 1961; Mulligan, 1963; Miller, 1956, 1963, 1964a and b, 1973; Anderson, 1970; Miller and Anderson, 1974).

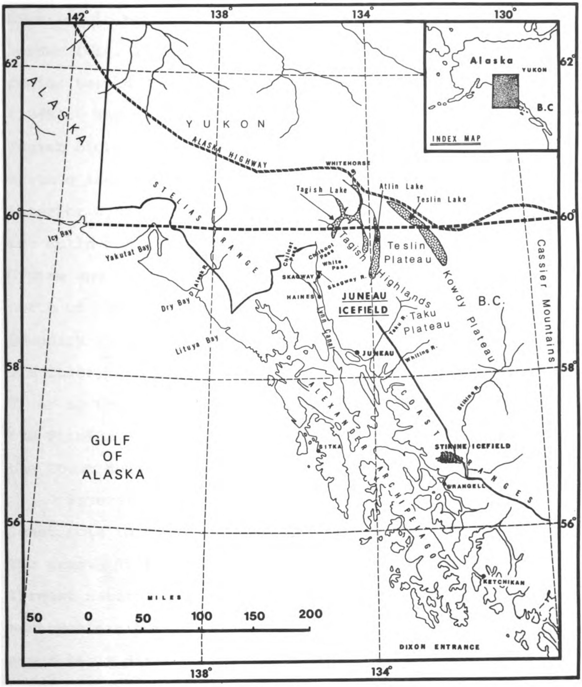

Figure 1. Southeastern Alaska, northwestern British Columbia and southwestern Yukon Territory

Geographical Setting

The study area is in the Atlin sector of the Cassier District in northwestern British Columbia close to the Yukon border (Fig. 2), an area included as part of a physiographic region termed the Canadian Cordilleran Province by Bostock (1948). The major part of the Atlin area lies within the Tagish Highland and the Teslin Plateau. Together these constitute the south central portion of the Yukon Plateau of the Interior System (Holland, 1964). The southwestern portion of the Atlin map area reaches into the Western Physiographic System and the Northern Boundary Range of the Coast Mountains, north of the Taku River (Bostock, op. cit.). The Southern Boundary Range, with certain geologic and physiographic similarities, is considered to extend southward from the Taku River to the area of the Stikine River. Both the Taku and the Stikine Rivers follow antecedent drainage lines across the Coast Mountains to the Gulf of Alaska (Fig. 1).

Several large lakes in the Atlin region play a significant role in the interpretation of the glacial history of the area. Atlin Lake (elev. 2182 feet, 665 meters), the largest natural lake in British Columbia, extends from the southwestern portion of the Atlin map area and continues in a northerly direction for 65 miles (104 kilometers) with its northern 10 miles (6 kilometers) lying in the Yukon Territory (Fig. 3, in pocket). The lake is positioned in a structural as well as physiographic low and has been significantly modified by continental-type glacial lobes which have

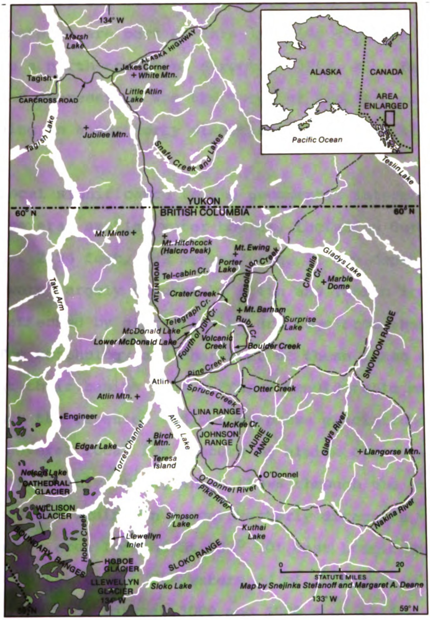

Figure 2. Atlin Region

originated in the Northern Boundary Range in the vicinity of the present-day Juneau Icefield (Fig. 1). Teslin Lake (elev. 2239 feet, 683 meters) and Gladys Lake (elev. 2915 feet, 888 meters) are part of this interior drainage system trending in a north-northwest direction (Fig. 3). Each of these linear depressions was the locus of Pleistocene ice masses moving in a northerly direction from the Stikine Plateau.

Of local significance to the geomorphic interpretations are several smaller water bodies, including Surprise Lake, (elev. 2894 feet, 882 meters)(Fig. 3) the depressions for which were largely scoured by Atlin and Teslin ice or otherwise affected by glaciofluvial drainage from these ice masses. The morphology and character of these lakes also reveal that their basins have suffered considerable modification through the repeated and out-of-phase glaciations relating to several local ice centers of alpine character.

Fourth of July Creek, which flows through the study area, at present drains in a southwesterly direction with its headwaters area located at 133° 20' West Longitude and 59° 42' North Latitude. The watershed of this stream initiates in a zone between 3000 and 6000 feet (915 to 1830 meters) elevation and extends to Atlin Lake at 133° 43' West Longitude and 49° 42' North Latitude. This part of the study area also includes the drainage basin of Consolation Creek, which lies northeast of Fourth of July Creek and includes its outlet to Fish Lake 60° 55' North Latitude and 133° 10' West Longitude (Fig. 3).

The Fourth of July Creek Valley is flanked on the southeast by an unnamed 20 mile (32 kilometers) long subsidiary range, the upper ridges of which attain elevations of 4500 to 6800 feet (1370 to 2075 meters). The major peaks on the ridge line of this range include Mount Leonard (elev. 6000+ feet, 1829 meters), Mount Vaughan (elev. 6000+ feet, 1829 meters) and Mount Barham (elev. 6808 feet, 2075 meters). Mount Ewing (elev. 5410 feet, 1649 meters) is the highest crest on the valley's northwest flank and is part of a 12-mile (19 kilometers) long ridge which separates Fourth of July Creek Valley from the valley leading to Porter Lake (Fig. 3).

CHAPTER II

GEOLOGICAL FRAMEWORK

Previous Research and Allied Studies

Maps of the Canadian National Topographical series cover the Atlin District at a scale of 1:250,000. Only a small part of the western portion, including some of the Atlin Lake sector in the vicinity of the unincorporated village of Atlin, has been photogrammetrically mapped at a scale of 1:50,000. Over sixty years ago, a reconnaissance report of the geology along portions of the Atlin Lake sector was compiled by D. D. Cairnes (1913). Little else, except local studies and brief geological descriptions of mining sites (e.g., Cockfield, W. E., 1925 and Gwillim, I. C., 1901, 1902) was published on this area until the rather recent geological map and report of Aitken (1955, 1959) the field work for which was supported by the Geological Survey of Canada. Part of Aitken's research has also been presented in his Ph.D. thesis (1953). Other recent maps and geological descriptions are available for the Lake Bennett sector, west of Tagish Lake (Christie, 1957), for the Whitehorse area in the adjoining Yukon Territory (Wheeler, 1961), the Teslin area to the north (Mulligan, 1963), Wolf Lake, to the northeast (Poole, 1955) Jennings River to the West

(Gabrielse, 1969), and aspects of vulcanological history of the region to the south (Souther, 1968, 1972). A compilation of these and other local studies is presented by Douglas (1970). Also to be mentioned is the glacial geological and regional geology report by Miller (1956, 1963) relative to the Taku River sector and the Juneau Icefield south of the study area, and that of Gilkey (1951) and Forbes (1959) covering the petrology and structure of the central icefield nunataks and key lithologies found on a transect across the icefield from Juneau to Devil's Paw (8504 feet, 2590 meters).

Main Lithologic and Structural Elements

A review of these reports reveals that three main rock groups dominate the Atlin map area, especially to the south and east. These are the Cache Creek series which is Pennsylvanian and/or Permian in age; the Coast Range batholithic intrusions of Mesozoic and early Tertiary time and the subsequent middle to late-Tertiary volcanic and early-Quaternary volcanic sequences (Aitken, 1959). The northeastern part of the map area contains much limestone and metasediments in the Cache Creek group (Aitken, 1959; Mulligan, 1963). These include cherts, argillites, limestone, greywacke, sandstone and silt-stone. Over the rest of the map area east of Atlin Lake there are predominant exposures of the Cache Creek sequence with more recent intrusives including some Holocene volcanics.

Quartz diorite and diorite intrusions in the Mount McMaster, Mount Llangorse, and Hayes Peak (Fig. 3) sectors are considered by Aitken (1959) to be of the same general age as the coarse-grained igneous stock which has been termed the Fourth of July Creek body. All of these intrusions are considerably unrooted and by comparison of the petrology as well as by structural similarities, they are considered to be of the same age.

A later Mesozoic acidic intrusion came in adjacent to the Fourth of July body and has been referred to by Aitken (1959) as an Alaskite stock which extends on the surface from Surprise Lake to Trout Lake, just south of Gladys Lake. West of Trout Lake is the Snowden Range (Fig. 3) where Alaskite lithology occurs at a location beyond a geomorphologically and lithologically identifiable fault zone found to the east to Trout Lake (Aitken, 1959). A quartz monzonite on Dawson Peak (Fig. 3) is assumed to be the same age as the Alaskite noted above, the criteria being composition, texture and structure. Kerr (1948) has traced a belt of similar intrusives from Surprise Lake to the Taku River map area and from there to the Stikine River region where the intrusion is dated as late-Lower or early-Upper Cretaceous in age. This is quite comparable to the structural and lithologic characteristics in the Stikine area described by Buddington (1927) on the contimental flank of the Southern Boundary Range. Thus revealed is the considerable linear extent of this geological province.

As connoted above, the most predominant intrusives are the Mesozoic Coast Mountain granitic intrusions of batholitic and stock proportions, with associated lithologies affected by permeating metasomatism from these acidic emplacements. By K/Ar dating these granodiorities are known to be some 50 x 10^6 years B.P. (A. Ford, US Geological Survey, personal communication). This granite is particularly abundant in the southern half of the map area adjacent to the higher Coast Mountains, with some outliers in the specific area of this study. One such outlier is the Fourth of July Creek stock or batholithic window which has been referred to as the Fourth of July Creek body. This body is composed of granodiorite and quartz monzonite. It occupies more than 300 square miles (777 square kilometers) in the western sector of the study region. On the southwestern edge of this part of the Coast Range batholith and adjacent to Atlin Lake, this unit occurs as a pink granite, quite porphyritic and with some unusually large phenocrysts.

The southwestern corner of the Atlin map area contains a thick group of predominantly volcanic rocks, the Sloko group. This is an especially unique lithologic assemblage as it includes pyroclastics (dacite and rhyolite) with subordinate andesite and basalt, all segments of which have been well described by Aitken (1959) and Douglas (1970). Other smaller intrusions of less areal importance are also discussed by Aitken (1959) and Souther (1972). Some of these, including dolerites and trap dikes, have been reported in abundance on the Juneau Icefield by Miller (1956, 1963) and Forbes (1959).

Surficial Geology

As for the surficial geology, only brief overviews are suggested in many of the reports noted above, with practically nothing on the Quaternary deposits of the Atlin area per se. It is clear, however, that unconsolidated Quaternary drift covers more than one-half of the map area of this study and thinly mantles the entire Atlin region. Chronosequences in this drift will be of primary consideration in the glacial geological discussions to follow.

CHAPTER III

REGIONAL CLIMATIC PARAMETERS

Introduction

The Atlin district has a continental type climate with some maritime influence from the Alaskan Panhandle partly in consequence of the presence of passes and valleys extending well inland through large sections of the Coast Mountains. The rain shadow effect of the Northern Boundary Range also produces low precipitation on the interior flank of the main axis of the Boundary Range. It should be noted, however, that orographic influences in the interior highlands also cause local increases in precipitation. Though the winters are generally long and cold, temperatures are not as extreme as in the more interior areas of Yukon Territory and north-eastern British Columbia. The summers are short and relatively warm.

Available Weather Records

The Canadian Department of Transport standard synoptic weather records are available for the village of Atlin from 1905 to 1946, with simplified daily observations carried forward by the Government Agent of the B.C. Provincial Government since 1947 (Department of Agriculture, British

Columbian Provincial Government, 1975). Year round three-hourly observations have been taken at the Sub-arctic Research Station of the Foundation for Glacier and Environmental Research in Atlin, B.C., beginning in the summer of 1968. At Carcross, 55 miles (90 kilometers) northwest of Atlin, data have also been recorded from 1907 to 1946, as they have since 1941 in Teslin 60 miles (96 kilometers) northeast. Since 1943 continuous synoptic weather observations have been taken in Whitehorse on the Yukon River 120 miles (183 kilometers) north-northwest of Atlin. Anderson (1970) has tabulated weather data for the Atlin region, including Carcross, Whitehorse, Atlin and Teslin (Table 1).

Temperature and Precipitation Ranges

The mean annual temperature for the village of Atlin is 32°F (0°C) compared to 31°F (-6°C) at Whitehorse and 30°F (-1°C) at Teslin. The range is -54°F to 87°F (-47°C to 31°C) with an average of 85 frost-free days. Total mean annual precipitation averages 11 inches (28 centimeters), with 53 inches (135 centimeters) of snowfall (Porsild, 1951; Kendrew and Kerr, 1955). Seasonal observations by Aitken (1959) and by the writer in the summers of 1969 through 1972, suggest that the climate of the low relief area around the Atlin town site is drier than the surrounding highlands, again reflecting the pronounced effect of topography on precipitation. Temperatures are also lower in the surrounding mountain and higher valley areas. Frost has often been

TABLE I

CLIMATOLOGICAL DATA SUMMARY FOR STATIONS IN THE ATLIN REGION[1]

	Atlin	Carcross	Teslin	Whitehorse
Mean annual temperature	32°F (0°C)	29°F (-1.7°C)	30°F (-1°C)	31°F (-0.6°C)
Absolute range of recorded temperatures	-54° to 87°F -48° to 31°C	--	-63° to 89°F -53° to 32°C	-62° to 91°F -52° to 33°C
Average number of frost-free days	85	54	66	49
Mean annual precipitation	11.3 in 24.9 cm	8.9 in 19.6 cm	12.8 in 28.2 cm	10.6 in 23.3 cm
Mean annual snowfall	53 in 116.6 cm	41.8 in 91.9 cm	55 in 121 cm	45 in 99 cm

1 After Anderson, 1970; data from Porsild (1951) and Kendrew and Kerr (1955).

observed in July and August at elevations of 3500 feet (1067 meters) and above. Snowfall is also common from June to August at and above 6000 feet (1829 meters). It is apparent that much of the weather is strongly controlled by the topography.

Climatic Variations and Secular Shifts in Regional Storm Paths

Of regional significance is a long-term climate-glacier study done by Miller (1963, 1967, 1972, 1973a) and his colleagues on the Juneau Icefield Research Program (JIRP) in the Northern Boundary Range. More recent JIRP reports (Miller, 1975a) and Jones (1975) emphasizes the assessment of Neoglacial to present climatic changes in the Atlin region and relate this to Little Ice Age glacier fluctuations and small cirque glaciers on the Cathedral massif at the head of Atlin Lake. Through these investigations interpretations have been made of recent glaciological data on the Juneau Icefield and its peripheral glaciers, as well as of secular climatic records since the 1840's at a number of coastal stations in Alaska. In particular, temperature trends and periodicities have been correlated with solar energy variations. It has been found that over the past several centuries during periods of high solar activity (maximum sunspot numbers, associated with increased corpuscular radiation) the anticyclonic continental high pressure cell in the Arctic interior has intensified and expanded outward beyond the Coast Range to meet the North Pacific maritime low near the

Alaska coast. During such times, the Atlin sector has experienced generally warmer and drier conditions. With decreased solar energy (reduced charged particle radiation) the clockwise-turning continental air mass contracts and moves inland, with the result that the maritime low moves landward, resulting in increased maritimity. The consequence is more precipitation and related cooling in the Atlin region. The zone of interaction between the high and low pressure cells is the locus of storm paths which are considered to shift in generally northeasterly or southwesterly directions as the major pressure systems are modified by variations in solar energy flux in the atmosphere.

Through the 1950's and early 1960's, these storm paths followed a westerly or coastal orientation, with the interior climate being warmer and drier than it was at the turn of the century when interior ice masses such as the Llewellyn Glacier (Fig. 2), were at their Holocene maximum. Affected by an apparently 90-year cycle which reached its peak in the late 1940's and 1950's, the storm paths at present are shifting into a more interior position. The ultimate interior migration at the end of the current downward swinging hemicycle should tend to shift back shortly after the beginning of the 21st century. This concept has been involved in the interpretation of palynological records from Atlin bogs, covering even larger cycles (Miller and Anderson, 1974). In fact approximately 800 and 2400 year cycles of climatic change have been recently described (Miller, 1973a) which appear to be corroborated by the research of other groups (e.g. Bray, 1971). During the maximum Pleistocene stages, storm paths are considered to have been in a far more easterly position via the mean inland shift of cyclonic centers* (i.e., affecting the region to even stronger maritime influences). It is this which presumably gives rise to the centers of major glaciation well to the east and northeast of the present topographic axis of the Juneau Icefield (Miller, 1964; 1973a).

The position of late-Wisconsinan and Holocene storm paths in the Atlin district and their associated climatic conditions have been discussed in more detail by Anderson (1970) and Miller and Anderson (1974) based on palynological as well as geomorphological evidence relating to the past 10,000 years (Holocene) in the Atlin Valley (Table II).

* It is noted (see glossary) that the term "storm paths" refers to the interpretation of peripheral winds in the interaction zone between the low-pressure maritime cells and high-pressure continental cells along the north Pacific Coast. This is distinct from the "cyclonic tracts" which refers to the movement of the center of low-pressure systems which is normal to the system movement on paths as here defined.

TABLE II

HOLOCENE CHRONOLOGY FOR ATLIN AND TAKU DISTRICTS*

ATLIN DISTRICT / **BOUNDARY RANGE**

YEARS B.P. X 1,000	TIME INT.	MAJOR VEGETATION AT LOWER ELEVATIONS	MEAN JULY TEMP. (°F) 48 50 52 54	CLIMATIC CHARACTERIZATION	DOMINANT GLACIAL ACTIVITY
NEOGLACIAL (0–2.5)	I	SPRUCE FOREST WITH PINE MINOR CHANGES IN BOGS		WARM WET	RETREAT
	II	WHITE SPRUCE FOREST WITH PINE		COOL DRY DECREASED STORMINESS	NEOGLACIAL GROWTH AND ADVANCES
THERMAL MAXIMUM (2.5–8)	III	SPRUCE FOREST WITH FIR		WARM WET	BEGINNING OF REGLACIATION
	IV	SPRUCE FOREST WITH ALDER		WARMER WETTER INCREASED STORMINESS	
	V	WHITE SPRUCE FOREST DOMINANT		WARM WET	GENERAL DOWN-WASTAGE AND RETREAT
	VI	SPRUCE WOODLAND		COOL DRY DECREASED STORMINESS	INTERMITTENT RETREAT
	VII	SHRUB TUNDRA (HERBACEOUS TUNDRA IN NORTH)		COOLER DRIER	STILLSTAND WITH READVANCE
	VIII	SPRUCE WOODLAND		COOL DRY	RETREAT
	IX	SHRUB TUNDRA		COOLER DRIER	STILLSTAND AND ADVANCES

TAKU DISTRICT (AFTER HEUSSER, 1952, 1965)

TIME INT.	CLIMATIC CHARACTERIZATION	MAJOR VEGETATION AT LOWER ELEVATIONS	YEARS B.P. X 1,000
I	WARMER DRIER	LODGEPOLE PINE	
II	COOLER WETTER	HEMLOCK MAXIMUM	1–2
III	MAXIMUM WARMTH AND DRYNESS	WESTERN HEMLOCK DOMINANT	3
IV		MOUNTAIN HEMLOCK HEATH / LYSICHITUM MAXIMUM	4–5
V	WARMING DRYING	ALDER MAXIMUM	6
	COOL MOIST	SEDGE HEMLOCK SITKA SPRUCE	7–8
VI	WARMING DRYING	LODGEPOLE PINE MAXIMUM ALDER-WILLOW SEDGE	9
VII	COOLER WETTER		
VIII	COLDEST		10
IX		PLANT REFUGIA	11

* After Miller and Anderson, 1974

CHAPTER IV

SOILS AND VEGETATION RELATIONSHIPS

Introduction

In this region a relatively wide range of parent material, primarily glacial drift, results in major soil variations. Of particular importance in the Atlin region is the presence of carbonates in glacial debris in restricted areas near limestone outcrops. Thick deposits of outwash material (valley train deposits) are in general well-drained. The less common till deposits, although often quite thick, are much less permeable.

In addition to the effect that drift character and topography have on drainage, the moderate to high relief of the area results in considerable microclimatic variation. It must be remembered that 1,000 feet (305 meters) of elevation is approximately comparable to a climatic shift of 300 miles (500 kilometers) north, or the equivalent of a climatic change to be expected in about 4^o of latitude change.

Micro-climatic conditions are also extremely important in the mountainous areas of this terrain because temperature, precipitation, wind direction and isolation all vary considerably throughout the area. In locations above 3500 feet (1067 meters) it is common to find sporadic permafrost of considerable thickness.

Variations in Soil Maturity

Because of the generally cold winter climate and low rainfall, even in summer, the soil forming processes are slow. This, along with a generally rapid recession of ice from the major Atlin and Gladys Lake valleys makes the soil formation process of less value than expected in the relative dating of glacial sequences of the last major Wisconsinan advances preceding the Holocene. In fact, in the study area the Atlin III and IV and Gladys III and IV moraines are essentially azonal. This means that there is not significant variation in soil maturity on any of the moraines of upper Wisconsinan age. The moraines of Atlin I and Gladys I age, however, are quite well oxidized and can be well distinguished from later moraines by their yellowish (limonitic) oxidation. This oxidized material is particularly well displayed on the upper southfacing flank of Caribou Ridge and at the top of the valley flanks of Porter Lake Valley where slumped morainic debris is revealed by yellowish soil taluses. Similarly, the lower (basal) tills in Boulder Creek, Pine Creek, Spruce Creek and McKee Creek show similar oxidation, quite comparable to oxidation of upper Woodfordian moraines in the mid-continent sequence of the interior Great Lakes region in Michigan and Indiana.

Regional Comparisons

With further respect to the comparison with mid-continent soils, Lietzke and Whiteside (1972), in a paper dealing with the development of spososoils on 5000-6000 foot (1524-1981 meter) ridge topsoils of nunataks on the Juneau Icefield to the south, have found interesting similarities with the glacial soils of Michigan. They report that these soils are characterized by the development of thin microprofiles in that more maritime region which has been substantially wetter than the Atlin area. Specifically they found two distinct breaks in soil development with respect to elevation and hence, presumably, to age. In the central icefield area about 40 to 50 miles (64 to 80 kilometers) inland from Juneau and 80 to 90 miles (128 to 144 kilometers) south of Atlin, the soils were generally young at elevations below 4000 feet (1200 meters) showing little profile development except here and there some leaching and podzolization (spodosols). Above this level, to 5000 feet (1500 meters), buried paleosols were found, with the upper section about the same age as the well weathered and "easily crumbled rock fragments and a strong soil development." Above this to 7000 feet (2100 meters), the surface soils appeared even older. They conclude that in this section "the soil processes may have commenced about the same time as the paleosol formation or earlier."

Above 7000 feet (2100 meters), they found no developed soils, only azonal rubble produced by intense frost action and high winds which have removed the fines. They further note strong similarities between the high-level Coast

Mountain soils in this area of existing glaciation and the typical spodosols of northern Michigan produced by late-Glacial and Holocene soil forming processes.

In this context, in the Atlin area the uppermost till of ground moraine found in the lower Fourth of July Creek Valley near the Atlin road (and also in many road cuts to the south and north toward Atlin and Jake's Corner), display spodosol development, possibly related in age to the younger nunatak soils described above. Lietzke (1969) has found further similarities between the surface soils of Atlin and the upper Great Lakes region, suggesting similar ages...i.e., Valders and Port Huron.

Soils-Vegetation Characteristics With Age

In a study of soils in the Atlin Region, Lietzke (1969) also found soil-vegetation relationships to be quite useful in providing a qualitative assessment of comparative soil parent material. It was found, for example, that well-drained soils with a C_{ca} horizon, or a high pH throughout, support aspen (Populus tremuloids) and Lodgepole pine (Pinus contorta) in proportions which vary with soil texture. Lodgepole pine with a lichen mat (primarily Cladonia) and occasionally aspen, are characteristic of coarse-textures and hence well-drained outwash soils. In a few locations well-drained soils with a thick organic mat support Douglas fir and Alpine fir.

Also it is found that soil development in this region is greater in fir stands than in the areas of Lodgepole Pine. This was borne out by an earlier reconnaissance study of soils between Whitehorse, Y. T., Carcross, Y. T. and Atlin, B. C., in which Lietzke (1969) also reported the soils to be primarily Aridisols or Inceptisols with spodosol (podzol-like) soils occurring at some sites. Palynological and radio-carbon evidence (Anderson, 1970) and from the writer's work in Boulder Creek (discussed later) suggest that all of the surface soils in the lower Atlin valley, north of the Village of Atlin, are not much older than 10,000 years B. P. These spodosols are suggested to be no older than that.

As for differentiating the relative ages of Holocene surfaces in the study area, it is reiterated that the aridity of the region has slowed weathering rates. This is to such an extent that only the soil-vegetation approach has validity, although it is recognized that efforts have been made in other interior maritime regions to use soils alone for distinguishing Neoglacial variations (Mahaney, 1975).

Poorly drained bog areas are also common in the area and in those bogs soils have developed in less than 10,000 years B. P. (Anderson, 1970). In fact, as will be shown, a drained bog in the Boulder Creek Valley has surface soils dated by the C-14 method to be less than 1100 years B. P. (see Ch. VIII, p. 155). When such bogs are at relatively high elevation, they are often sites of sporadic permafrost and unusual frost feature development. This is especially true at elevations above 3000 feet (1067 meters). Dwarf birch and willow along with various grasses and reeds are the predominant bog vegetation. Locales of sporadic permafrost are generally raised and contain fewer grasses and reeds and commonly display a thick mat of lichens. Each of these aspects will be discussed in more detail in later chapters.

CHAPTER V

PRE-WISCONSINAN GLACIATION

Introduction

There is abundant evidence that at least one pre-Wisconsinan glaciation far more extensive than any relating to the Wisconsinan advances covered northwestern British Columbia and adjoining sections of Alaska and the Yukon Territory. Rounded and smoothed highland topography, often seen as remnants on partially serrated ridge crests of the higher peaks in the Coast Range (Miller, 1964a) show the effect of ancient transfluent ice overriding well above the obvious limits of Wisconsinan glaciation. This has also been reported at the highest elevations in the Cassiar Mountains 100 miles (160 kilometers) east of the Atlin region (Johnston, 1926). Considerable evidence is given by Denny (1952) and others that old weathered tills exist north of the Atlin District, well outside of the Wisconsinan glacial limits.

Morphogenetic Phases

The pre-Wisconsinan phases of glaciation is in general referred to as relating to an Intermontane 'Icecap' Glaciation by Miller (1956) and the Continental Ice Sheet Stage by Kerr (1936). This maximum pre-Wisconsinan phase has also been recognized in southwestern British Columbia by Davis and Mathews (1944) who designated it as a Phase IV glaciation. Attained only in pre-Wisconsinan time, this larger sub-continental glaciation presumably covered all summits in the Atlin map area with some radial flow from locally high accumulation areas. The main outflow of this glaciation moved toward the south and southeast and also toward the northwest, with major spillovers into the Skeena River and down the Skagway, Whiting and Taku Rivers in the Boundary Ranges (Fig. 1). There was also significant diffluent dispersal of ice out of the Chilkat, Chilkoot, and White passes into the Lynn Canal area near Haines and Skagway, Alaska (Fig. 1). The pre-Wisconsinan Cordilleran ice sheet in some sectors extended inland over 500 miles (800 kilometers) and so in lateral extent was probably larger than the Greenland ice sheet is today. At its maximum extent it spread eastward in contact with the Laurentide Ice Sheet (Flint, 1957, 1971). A great deal more work should be done on this little known aspect of the glacial geology in this larger region where the Laurentide and Cordilleran ice-sheets were confluent.

Ancient Ice Centers and Outer Limits of Glaciation

The central areas of pre-Wisconsinan accumulation serving as the key ice centers which affected the Atlin area are believed to be not only on the inland flank of the Boundary Ranges but also in the Cassiar Mountains, about 150 miles (240 kilometers) east of the Coast Mountains. The broad region between is considered to have been a linear zone of climatologically controlled accumulation of large magnitude, as opposed to the more local and orographically controlled Wisconsinan ice centers located within the confines of the Boundary Range. Thus in the pre-Wisconsinan glaciation ice radiated in all directions, including a strong flow to the east into Alberta. To the west, its directional movement was largely controlled by the Coast Mountains and the trans-range valleys, particularly in the early waxing and late waning phases. Some of the pre-Wisconsinan ice coalesced with local ice from the high coast mountains, but the main drainage was largely deflected by these mountains in its later flow (Miller, 1964b).

In the Atlin region the local and subsidiary ice center, which added its increment to the major pre-Wisconsinan glaciation and probably to a lesser extent in early Wisconsinan time, was presumably well southeast of the study area of this dissertation. In fact its axis and center in the lesser phases of these ancient glaciations appears to have lain about 10 miles (16 kilometers) or more east of the international border where it crosses the Juneau Icefield and probably adjacent to but somewhat east of Mt. Nelles (8400 feet, 2560 meters) and Devils Paw (8584 feet, 2615 meters). This is noted here because these are the two highest massifs which center the backbone of the Coast Mountains adjoining the map area of this study.

Very little more can be said about details of pre-Wisconsinan glaciation except by reference to Pleistocene stratigraphy investigated elsewhere in Alaska and the Yukon (Karlstrom, 1961; Denton and Stuiver, 1965; Flint, 1971; Péwé, 1975). As these areas are well outside of the locale of the present investigation they are not discussed here, except to note that there is evidence for late Miocene or early Pliocene glaciation in western and southern Alaska (Miller, D. J., 1957; also Flint, loc cit). Also, it should be mentioned that drift of Illinoian age is present in nearly every glaciated area of Alaska (Péwé, 1975), and indeed even possibly in the west flank of the Fairweather Range in S. E. Alaska (Goldthwait, personal communication). In northern Alaska such ages are based on stratigraphic and radiometric evidence near Nome, while in other regions assignment is made on basis of surficial expression and position in the local drift sequence. Weathering profiles on Illinoian drift are thicker and more conspicuous than on Wisconsinan drift (Péwé, et al, 1965). In the Yukon-Tanana Upland of Interior Alaska, the Illinoian snowline was found to lie approximately 1300 to 1500 feet (400-500 meters) below the present snowline. In this interior Alaska region the maximum Wisconsinan snowline was about 560 feet (170 meters) below the present snowline. On the southern Alaska Coast, on the basis of cirque distribution and elevations, (Miller, 1961) the early Wisconsinan snowlines were as much as 2500 feet (762 meters) below present mean late summer snowlines or névé - lines. In the dry interior of east - central Alaska, the Illinoian snowline was found to be about 4100 feet (1250 meters) (Péwé and Burbank, 1960). By extrapolating this type of evidence to the Atlin region, it is again mentioned that the maximum Illinoian and earlier (?) glaciations probably overrode all of the major ranges, including higher peaks in the area.

Tectonic activity in the Coast Range and the interior mountain ranges of British Columbia as well as the air mass circulation have played a significant role in the extent of pre-Wisconsinan glaciation. Along the coast of the Gulf of Alaska, between Cape St. Ellas and Yakutat Bay (Fig. 1), the recent glaciation is nearly as extensive as Wisconsinan maximum glaciation. D. J. Miller (1957) attribures this to higher mountains in recent time than in the Wisconsinan time. The former lower elevations would have permitted more moist air to find its way into the interior ranges, including the ice center locale in the Cassiar Mountains where overriding is evidenced as high as 10,000 feet (Johnston, 1926). Thus it is presumed that pre-Wisconsinan Coast Range elevations may have been sufficiently lower to have allowed greater air mass transfer across them.

The foregoing may also explain the apparent difference in vertical spacing of cirques in the tandem sequences found on the coast compared with those found inland (Ch. VI). A further amplification of these concepts is given by the recognition of Illinoian cirques, with this age designation primarily suggested by weathering profiles in the Yukon-Tanana Upland. Here cirques have been found to be far more extensive on the north facing slopes. This is interpreted by some as evidence for air masses moving in from the north rather than the south-southwest as today (Péwé and Burbank, 1960). This could also be manifestation of a very marginal climate which favored snow accumulation on the north-facing slopes. Another concept discussed recently by Miller (1973a) and also Budyko (1972) is that with colder conditions during pre-Wisconsinan time (and, indeed, even with respect to the Wisconsinan maxima) the Arctic sea ice as well as terrestrial snow cover was much greater in lateral extent. Correspondingly in mountain regions there were lower climatological snowlines. This would mean increased areas where high albedo pertained, thus substantially lowering the incidence of absorbed radiation. This cooling would upset the heat balance of the northwestern sectors of the North American and Eurasian continents, and in turn bring a more maritime influence farther inland.

It should be mentioned in conclusion, that north of the study area there is some possibility of pre-Wisconsinan soils on isolated sectors of the high ridges at the north end of Lake Atlin near Jakes Corner (Fig. 2). This is suggested by the development of Gossan soils well above the moraine-delineated Wisconsinan ice levels (characteristics of these soils are yet to be investigated but they are strongly stained with limonite - possibly, however, this is related to the presence of ferruginuous limestones). For the most part, Wisconsinan glaciation was so intense in the region that it has erased or at least severely altered the pre-Wisconsinan geomorphology. Similarly, periglacial processes of the Holocene have somewhat altered some of the evidences of earlier glacial processes in the study area.

CHAPTER VI

WISCONSINAN GLACIATION

Introduction

Extensive evidence for Wisconsinan glaciation in the study area indicates that ice covered all but a few of the highest peaks in northwestern British Columbia and adjoining areas of Alaska (Miller, 1956, 1964b; Swanston, 1967). The most recent phases of glaciation are middle to late Wisconsinan in age. In reconnaissance studies, based largely on air photo interpretation of the southwestern Yukon region, Hughes et al (1969) have mapped all deposits directly north and contiguous with the Atlin Map area as late Wisconsinan in age. They refer to this as the McConnell glaciation. Assuming continuity in such contiguous sectors, and further assuming that their assessment is valid, this would permit interpretation of the main glaciation which formed the present landscape of the study area to be late-Wisconsinan in age. Except for the presence of deep soils on well-weathered till and associated erratics scattered at a few high elevations, it is tempting to extrapolate this broad assumption to the Atlin region. As will be shown, however, there are inconsistencies in the direct transfer of this overview chronology in the Yukon to the study area in the Atlin region.

An early Wisconsinan or pre-Wisconsinan glaciation has been documented for much of the Cordilleran outside the immediate study area. Rutter, (1975) identifies an extensive early Wisconsinan or pre-Wisconsinan advance in the north-central eastern sector of British Columbia. Denny (1952) recognized an "Old" till in the Summit Lake region of northeastern British Columbia. From a study of cirque and berm levels, Miller (1956, 1961, 1975) interprets an early Wisconsinan glaciation (early Juneau/Atlin stage) in the Juneau Icefield region (Taku District) which far exceeded any of the later Wisconsinan advances. Evidence for this is the existence of low-level cirque basins near sea level and high-level berms, as well as high-elevation erratics on peripheral ridges and relatively small amounts of weathering on high-level tills relative to what should be expected were these upland surfaces mantled with pre-Wisconsinan ground moraine.

Evidences of Early Wisconsinan Glaciation

Within the study area too there is evidence for a higher glaciation than the more recent sequence discussed below. Though more highly weathered than the mid to late Wisconsinan deposits, these high-level tills also do not show the degree of weathering one would expect had they been exposed for the entire Sangamonian interglacial.

The only direct evidence for an early Wisconsinan Maximum is the well-weathered soil mantle on peaks of White

Mountain (Fig. 1) 50 miles (80 kilometers) north of the study area. There, at elevations above 4500 feet (1370 meters), gossan soil development is apparent which may be related to a pre-Wisconsinan surface. The carbonate composition of White Mountain must also be considered, as weathering would be more rapid in this kind of lithology than in the granitic peaks most commonly found within the study area. Further field work in this region is needed for a firm interpretation of an early Wisconsinan maximum throughout the larger northern B. C. - southern Yukon region.

In the Atlin district ice coming out of the Cassiar Mountains covered areas to the north, east and south. During the Wisconsinan glaciations this ice was largely diffluent, i.e. channeled by topographic lows as it flowed in a distributary fashion onto the Stikine, Teslin, Taku and Yukon Plateaus. The part of the Cassiar Lobe which most affected the study area passed northward across the Teslin Plateau and into the Gladys Lake depression (Fig. 2 & 3). In the present report this portion of the Cassiar Lobe is referred to as representing Gladys Ice. During Wisconsinan time the other significant glacial mass affecting the study area was the Coast Range ice-sheet which extended into the Atlin Lake depression from the south. This glacial mass is referred to as representing Atlin Ice.

Evidences for a Middle to Late-Wisconsinan Sequence

The following chronology illustrates mid to late

Wisconsinan glacial activity within the area of investigation. The first stage of both Atlin and Gladys ice is that stage for which there is no field evidence which can clearly delineate the extent of ice within the area of immediate concern. Again, though it should be kept in mind that there was at least one earlier and more extensive Wisconsinan glaciation preceding the glaciations discussed below.

The Fourth of July Creek Valley is divided into three major glacial-morphological sections (Fig. 4). The Lower Fourth of July Creek Valley is from the juncture of the Fourth of July Creek with the Porter Lake Valley, (the location of the Ruffner* Juncture moraine) to its mouth at Atlin Lake. This region contains thick till deposits and is the section which contains McDonald Lake.

The Intermediate Fourth of July Creek Valley extends from the juncture moraine to the confluence of Volcanic Creek and is characterized by glacio-fluvial terraces and by glacio-fluvial erosion in sectors where much drift has been removed.

The area above Volcanic Creek to the drainage divide, is referred to as the Upper Fourth of July Creek Valley and includes the relatively flat plateau at the drainage divide of Fourth of July Creek and Consolation Creek. This region contains a complex esker system, many kettles and some sporadic occurrences of frost mounds.

In Figure 5 the maximum extent of each of the following

* Named for Ruffner mine near moraine complex.

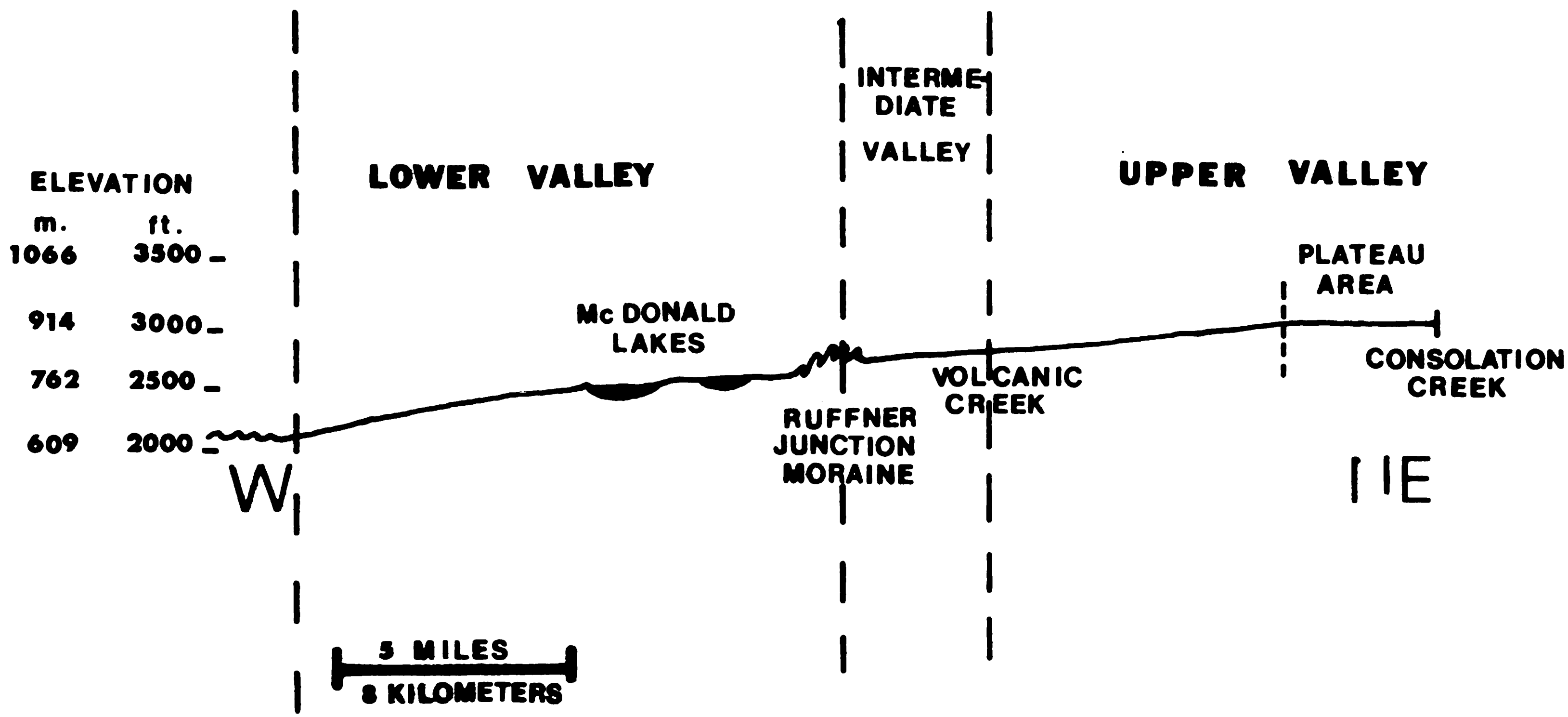

Figure 4. Major divisions of Fourth of July Creek valley

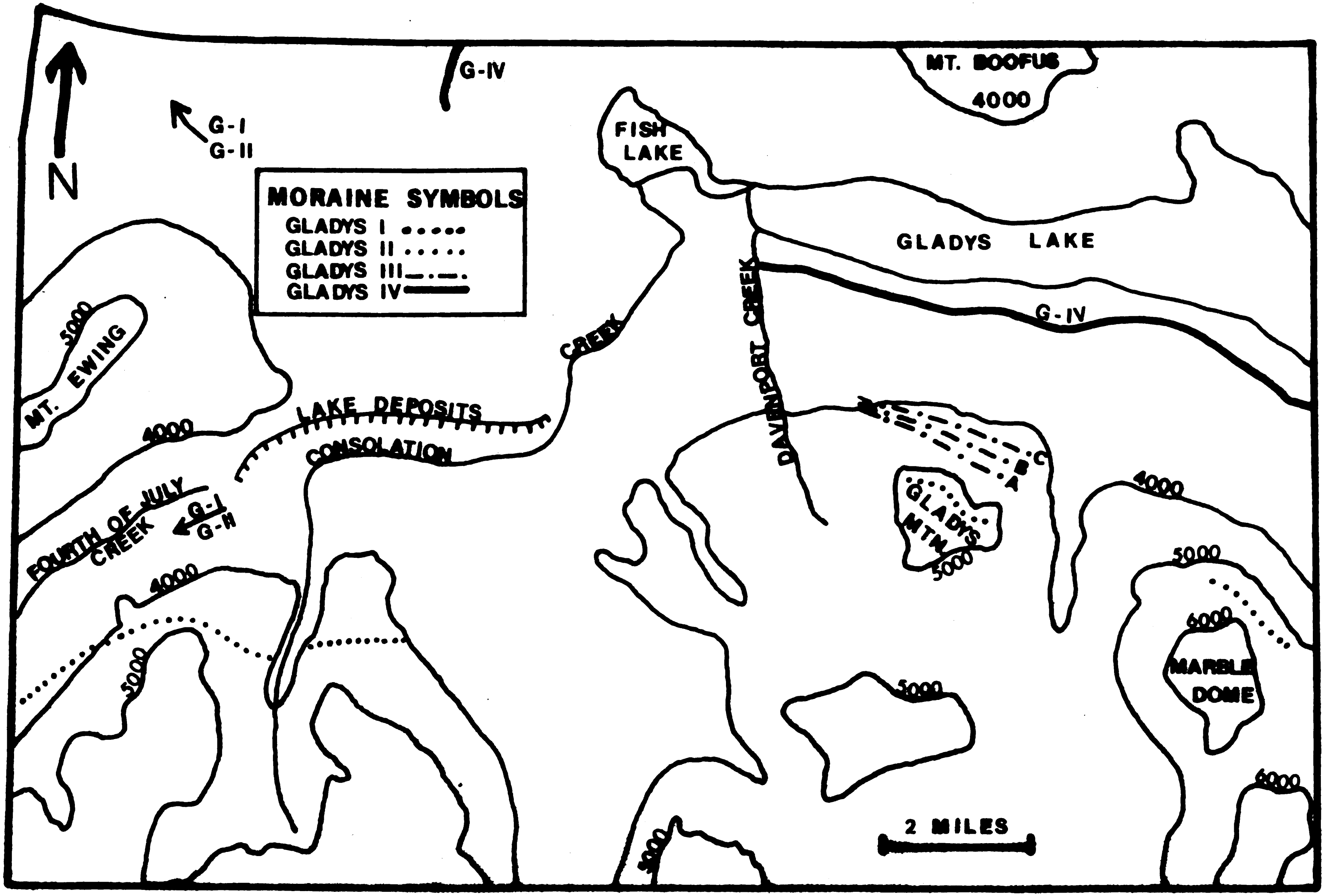

Figure 5a. Maximum extent of Gladys ice

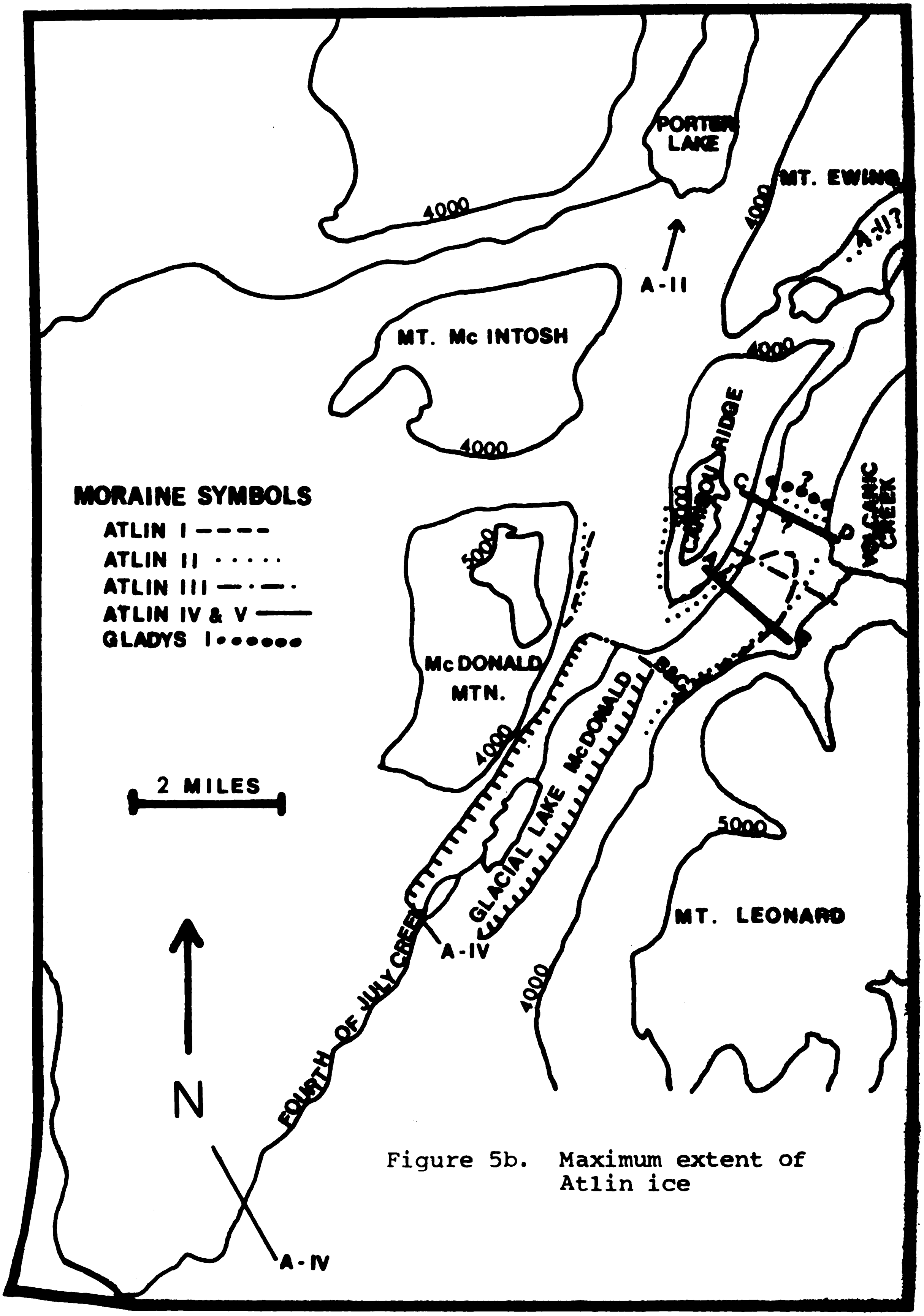

Figure 5b. Maximum extent of Atlin ice

stages is illustrated.

Gladys Ice Moraines

The glacial map of the southern Yukon Territory (Hughes et al, 1969) shows that ice from the Teslin Valley extended north through the Teslin depression where it was joined by ice flowing directly west from high centers in the Cassiar Mountains. The Gladys portion flowed through the northern part of the Atlin map region in the direction North 45° to 70° West (Fig. 3).

Gladys I

Ice of the Gladys I stage completely overrode the summits of Mt. Ewing (5410 feet, 1649 meters) and Mt. Boofus (5060 feet, 1542 meters - as shown in Figure 3). Good evidence for this is shown in Figure 6 where the peak is smoothed with fluted and drumlinoid topography on its flanks. Mass wasting, local cirque development in the waning phases, and subsequent effects of relatively intense glacio-fluvial activity have made it impossible to delineate the true extent of this phase at such type localities as Black Mountain to the northwest (Fig. 7). However, the existence of relatively unweathered high-level till and erratics (Fig. 8) on the top of the mountain between Chehalis and Davenport Creeks (i.e. on Gladys Mountain for the purposes of this study) indicate that Wisconsinan ice of Gladys I Phase did indeed override this peak at an elevation greater than 5500 feet (1676 meters). This state is assigned a middle Wisconsinan age on

Figure 6. Stereo pair of Mount Ewing

Smoothed ridge and summit of Mount Ewing showing molded topography and tor development.

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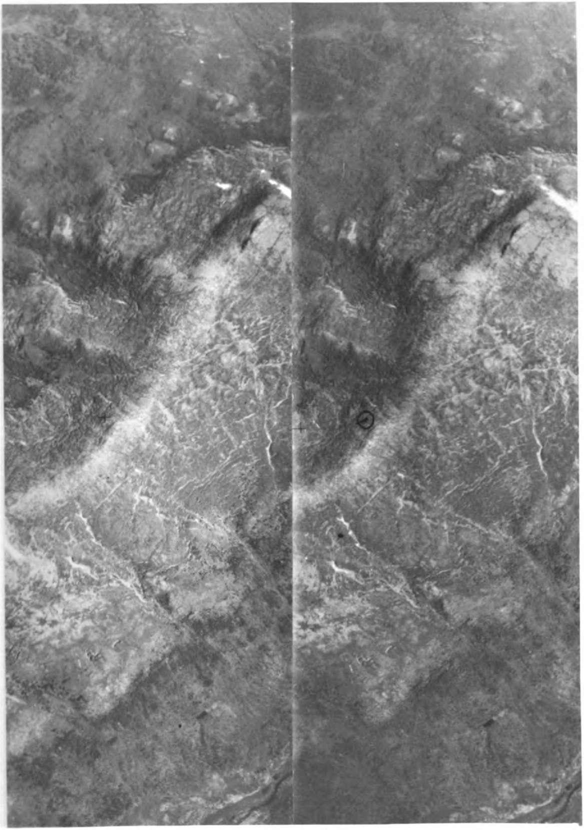

Figure 7. Fluted drift north of Black Mountain

The drumlinoid fluted drift of Gladys Ice cut by run-off channels of Atlin Ice and Gladys Ice. Note truncation by kame terraces of Atlin Ice. Black Mountain is in the lower part of the photograph.

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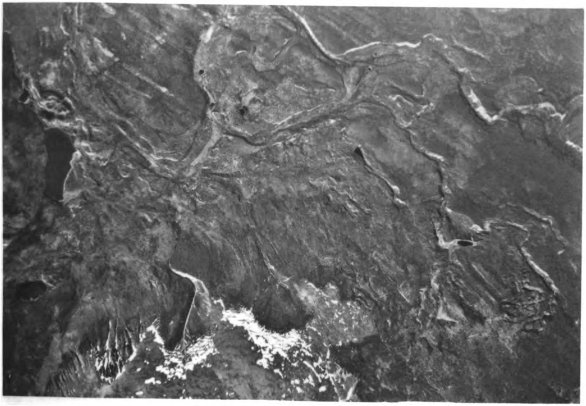
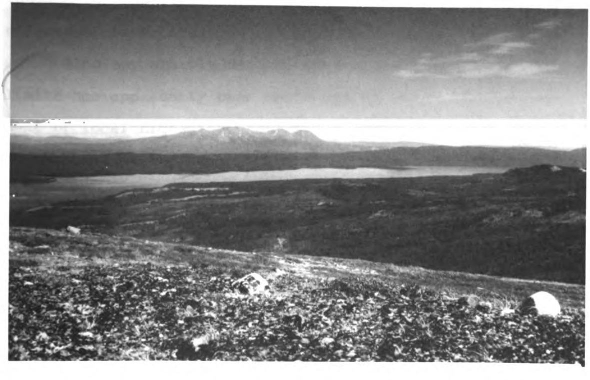
Figure 8. Erratics on Gladys Mountain

Looking southeast across the Gladys depression from summit of Gladys Mountain.





the basis of sequential levels and the great thickness of ice in the Gladys depression which would be required to achieve this level. Geomorphic reasons for this interpretation are considered below.

A more highly weathered pre-Atlin I till above 4500 feet (1372 meters) on Mt. Leonard and Mt. Vaughn is evidence that on Gladys Mountain this high ice phase did not make its way down the Fourth of July Creek Valley at this elevation on the valley walls. Serving as obstructions to this Gladys I ice were the Mt. Leonard, Vaughn, Barham and Edmund massifs (Fig. 3). These channeled diffluent lobes into the Surprise Lake Valley and down the present Fourth of July Creek Valley. Any evidences of terminal positions of these lobate ice tongues in these valleys has been erased by the overriding effects of later ice.

Also any sharply defined morainic evidence of these limits has apparently been destroyed by the later development of local high-elevation valley glaciers out of cirques at the head of the upper Consolation Creek Valley and on the flanks of Mt. Barham and Mt. Edmund. Thick till deposits, however, are present to an elevation of greater than 5000 feet (1524 meters) in spite of the fact that in this sector no clear upper limit of Gladys I ice has been delineated.

On the flank of Mt. Barham, due west of Consolation Creek, at an elevation of slightly more than 4500 feet (1371 meters) evidence for overriding during this and earlier glaciations is shown by deep incisions into a bedrock ridge resulting from sub-glacial streams (Fig. 9). Though modified by Gladys I stage ice, these features were carved by preceding high-level glaciation as well.

The maximum western extent of the Gladys I stage in the northern part of the Atlin map area is also not discernible as more recent kame terraces from later Atlin ice have abruptly truncated the fluted and drumlinoid topography of Gladys I ice and, indeed even those of the Gladys II stage (Fig. 7). Although this truncation has destroyed evidence of the maximum position of Gladys Ice it has actually demonstrated the areal insignificance of Atlin Ice in this region, and its relatively restricted nature after the last major stage of Gladys Ice.

From the degree of surface weathering and the formation of tors on the upper ridges of Mount Ewing (Fig. 6) the Gladys I stage is designated as middle Wisconsinan. This general phase of glaciation corresponds to the Greater Mountain Ice-sheet Phase of Miller (1964b).

Gladys II Stage

Prominent lateral moraines on Gladys Mountain represent a lower phase of Gladys Ice. Here the multiple moraine system at 5200± feet (1585 meters) flanks the eastern side of the mountain and can be traced laterally for several thousand feet. During this phase ice again made its way up the lower Consolation Creek Valley, and as well into the Surprise Lake sector and the upper and intermediate Fourth of July Creek valleys to the west. At this time the plateau

Figure 9. Bedrock incision on the eastern flank of Mount Barham

Figure 10. Moraines on Marble Dome

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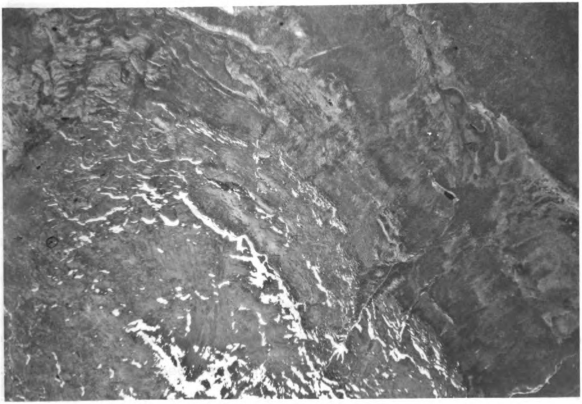

area northwest of upper Fourth of July Creek was filled with ice which in areal extent equalled the coverage experienced in Gladys I time. This possibly suggests a more temperate thermo-physical character than the Gladys I stage...i.e. more mobile ice. No lateral moraines, however, are found on Mt. Ewing. These were presumably destroyed by subsequent mass wasting. The lower phase Gladys II ice was also presumably fed by prominent cirque sources in the highlands near and above 5000 feet (1524 meters). Documenting this phase is a pronounced moraine complex on Marble Dome (Fig. 10), at less than 5500 feet (1676 meters).

Further moraines can be traced to quite a low elevation (i.e. 3000 feet, or 900 meters) in the Fourth of July Creek Valley to a moraine complex above Spear Point Terrace (Fig. 11). From the geometry of these moraines the ice gradient from Gladys Mountain to Spear Point Terrace is interpreted as about 100 feet (30.5 meters) per mile. The moraine is waterworked and densely kettled (Fig. 12). Proximal to the valley wall large fragments of talus cover the moraine. (This moraine complex was also affected by subsequent local ice from high level cirques as discussed later).

There are no lateral moraines in the high-level plateau area northwest of Gladys Lake. This again makes it impossible to do more than speculate the farthest extent of ice into this area during Gladys II time. Ice gradients elsewhere indicate that a glacier sheet covered most of the low-level plateau.

Figure 11. Stereo pair of the intermediate Fourth of July Creek Valley

Volcanic Creek juncture with the Fourth of July Creek is shown in lower part of photograph. Note esker development and fluvial and glacio-fluvial terraces on valley wall. The Spear Point moraine is just below the esker development.

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Figure 12. Kettle in Spear Point moraine complex

This site is just above the confluence of Volcanic Creek in the Spear Point Moraine complex.

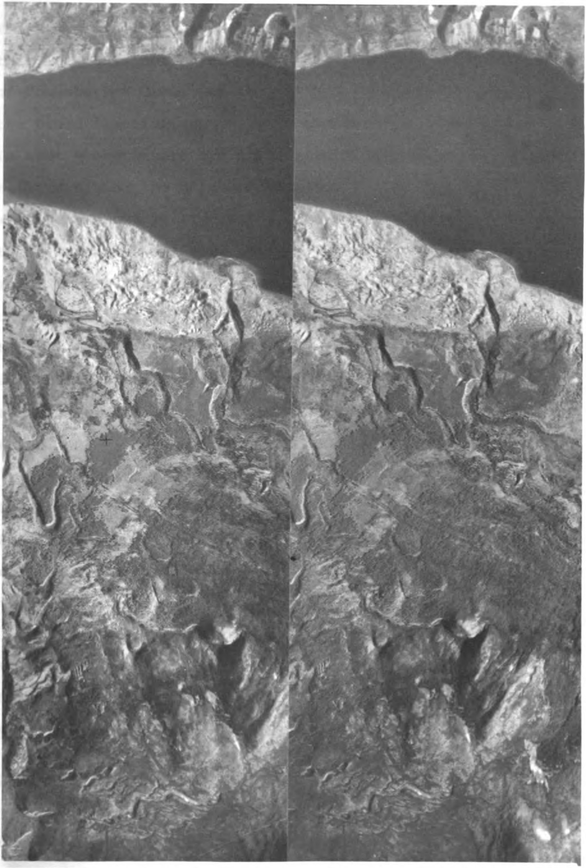
Still a Mountain Ice Sheet, the Gladys II phase represented a major complex of still-stands in the general downwasting of Wisconsinan ice. As such it is considered to relate to the beginning of <u>late-Wisconsinan</u> time.

<u>Gladys III Stage</u>

Although not always apparent on strictly morphostratigraphic grounds, lateral moraine positions are clearly identified for this stage by the distinctiveness of drainage controls on the flanks of Gladys Mountain. Here deeply incised stream channels of strikingly limited extent are found to terminate where the water flowed onto the ice in the Gladys Lake depression (Fig. 13). Morphologically-controlled by the ice position these channels delineate three major levels of downwasting - i.e. at about the 4900 foot (1493 meters) level (designated as Gladys III-A), at 4500 feet (1371 meters) (noted as Gladys III-B), and at 4000 feet (1219 meters) (noted as Gladys III-C). These separate levels are all considered to be parts of one major stage or general ice-level. This is because of the similarity of erosion features associated with each and because of the high gradient and common angle of slope of the ice associated with the three events. In this connection it is worth mentioning that even still-stands can be oscillating to the extent that there are intervals of greater and lesser ablation in consequence of short-term climatic variations.

No moraines for this series are present in the upper Fourth of July Creek or in lower Consolation Creek valleys.

Figure 13. Gladys Ice Moraines on Gladys Mountain



The steep gradient which has been extrapolated suggests that this ice could not have made its way very far up the valley of Consolation Creek and that it was probably confined to the broad flattish region lying northwest of Gladys Lake. In that sector there are three significant underfit streams still draining the northern boundary of the Atlin Map two of which are shown in Figure 14. The broader valleys in which these streams now lie as underfit drainage ways served as major runoff channels in the Gladys III stage. A fourth underfit stream (Fig. 15) lying beyond an intervening 3500 to 4000 foot (1066 to 1219 meter) ridge also served as an important runoff channel for the contiguous Atlin ice and as well for the downwasting of Gladys ice in the Gladys II stage.

The upper Fourth of July Creek and middle Consolation Creek valleys were apparently filled with ice at this time. But local accumulation to the volume of remnant ice left from earlier Gladys phases. The significance of this local provenance of ice is discussed in a latter section.

The extent of the Gladys III glaciation is considered to correspond to Miller's (1964b) Lesser Mountain Ice-sheet Glaciation in the Boundary Range and is presumed as late Wisconsinan in age.

Gladys IV Stage

Also late Wisconsinan in age is a lesser glaciation comparable to the Extended Icefield Glaciation of Miller

Figure 14. Underfit glacial runoff streams

Note the drumlinoid topography and small eskers in eastern portion of photo.

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Figure 15. Atlin Ice and Ice Runoff Channel

Photo northeast of Black Mountain on Yukon border.

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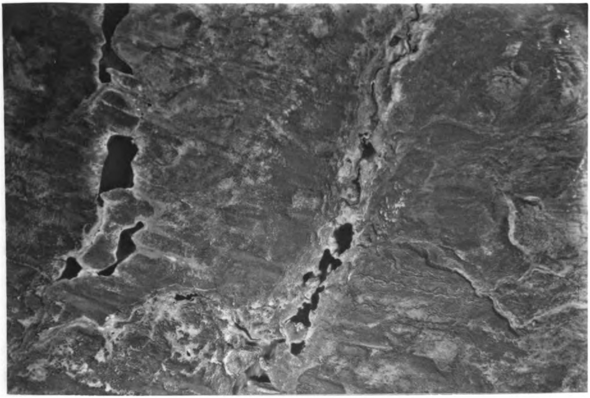


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(1964b). In this glaciation the ice downwasted and stagnated in the Gladys Lake Valley. This is evidenced by a deeply-pitted surface at an elevation of 3000 feet (914 meters) (Fig. 12). Associated deposits of massive proportions are reflections of an intensive fluvial deposition on top of slowly downwasting dead ice. A complex pitted region at Airplane Lake just northwest of Gladys Lake marks the northernmost extent of this important dead-ice zone (Fig. 3).

Atlin Ice Moraines

In middle to late Wisconsinan time, a lobe of Boundary Range ice also filled the Atlin valley. During its down-wasting phases it served to create much of the landform configuration in the vicinity of the village of Atlin and on the adjoining flanks of the main Atlin valley. This glacial mass sent distributory fingers up into all valleys which today are tributaries of the main Atlin valley. The limit of Atlin ice in these peripheral valleys was, of course, controlled by gradients and thicknesses of the main lobe, and as well by thermal properties of the ice, by bed-rock slope of the valley and in some cases by the addition of ice flowing out from local accumulation areas. Except for weathered high-elevation tills on Mt. Leonard, nearly all evidence of a maximum extent of this earlier advance of easterly-flowing Atlin ice into the Fourth of July Creek Valley was completely removed or covered by deposits from the later Gladys ice or by ice relating to Extended Icefield phases originating in the local accumulation centers.

Atlin I Stage

Following what is presumed to have been a pre-classical Wisconsinan glaciation, evidence for the next most extensive advance of Atlin ice is given by a steeply dipping and well-weathered yellow-colored moraine on the main southeastern ridge of Mt. Ewing, referred to in this text as Caribou Ridge (Fig. 3). This moraine extends downward from an elevation of 4300 feet (1310 meters) to 3800 feet (1219 meters) at which elevation it is truncated by a quite un-weathered younger moraine of late Atlin ice. An Atlin I moraine at 4000 feet (1525 meters) can be traced for 2 miles on the eastern side of the mountain south of Mt. McIntosh (presently referred to as Mt. McDonald). A moraine lying at 4500 feet (1372 meters) on the west side of Mt. McDonald further delineates the extent of ice from the main Atlin valley in this early phase.

The steep gradient of the Caribou Ridge moraine indicates that though ice was thicker in the lower Fourth of July Creek Valley than it was in later Wisconsinan stages, it was characterized by a steeper snout. This means it did not make its way into the intermediate Fourth of July valley as far as it did later on - i.e. when it was thinner and hence at lower levels. Once again it is suggested that this stage may have been thermophysically more polar (again the ice being not as mobile) in middle Wisconsinan time. As such it would correspond to the Greater Mountain Ice-sheet Phase in the Boundary Range Morphogenetic Sequence (Miller, 1964b).

The Atlin I ice passing up into Porter Lake valley (Fig. 3) did not have as steep a surface gradient as suggested by the fact that its lateral moraine does not slope as steeply. The total distance which ice flowed up this valley at that time is not known, because its northern limit is obscured by outwash. Also there is no evidence of a Porter Lake lateral moraine on Mt. McIntosh where local cirque activity and mass wastage have been particularly active. During this phase it is clear, however, that main valley ice did flow through the valley between Mt. McIntosh and Mt. McDonald and pass well up through the valley of Tel-Cabin Creek (Figs. 2 and 3).

Atlin II Stage

A much younger ice advance is documented by a lower-level and less weathered moraine on the valley walls in the intermediate Fourth of July Creek valley. Near the juncture of Crater Creek valley this moraine lies at about 3700 feet (1128 meters) elevation and can be traced down to 3000 feet (914 meters) at a point where it is truncated by terraces cut by meltwater from the upper Fourth of July Valley. Its terminal position lies beyond the entry of Volcanic Creek (Fig. 3). The Atlin II stage is the Intermediate Mountain Ice-sheet Phase (Miller, 1964b) and is considered to be of early late-Wisconsinan age. From comparable weathering it is suggested that the bold lateral Wisconsinan moraine at 5300 feet (1615 meters) elevation near the Camp 29 research station in the Cathedral Glacier Valley (Figs. 2 and 3) represents the same phase of glaciation (Jones, 1974, 1975). From that location the ice would have had a surface gradient of about 50 feet (15 meters) per mile.

Ice channeled through the Porter Lake valley reached a point about 1½ miles (2.4 kilometers) beyond Porter Lake (Fig. 3) where a moraine complex truncates the fluted topography produced in Gladys II time. This is one of few localities where there is evidence that there was contemporaneous and contiguous Atlin Ice and Gladys Ice. This total glacier mass included Atlin ice which made its way up the Tel-Cabin Creek Valley and through the valley south of Mt. McIntosh.

Other major intrusions of Atlin ice in this stage were between Halcro Peak (Mt. Hitchcock) and Mt. Carter to a position three miles (4.8 kilometers) north of Indian Creek (northwestern corner of map in Fig. 3). Here it is again indicated that Atlin II Stage ice met Gladys II Stage ice in an area where molded drumlinoid topography produced by Atlin Ice merges with the landforms produced by the most westerly limit of Gladys Ice. During the downwasting of these ice masses, meltwater drained north across the area and resulted in fluvial dissection of this drumlinoid and fluted terrain which had initially been produced by ice masses of the Gladys I and II stages (Fig. 14).

Although the originating ice-sheet of the Atlin II stage can be considered morphogenetically as relating to a Greater Mountain Ice-sheet Phase in the source regions of the Boundary Range, the confinement of this ice to major valleys with some diffluence through minor higher-level valleys in the Atlin region leaves room for misinterpretation of the magnitude of the glaciation unless the total regional picture is kept in mind.

Atlin III Stage

In the waning stage of Wisconsinan ice in the intermediate and lower Fourth of July Creek valleys three closely associated moraines near and just below the juncture of the Porter Lake valley are grouped into the Atlin III Stage. These are referred to as the Ruffner moraine complex. Morphogenetically they are bold terminal and lateral moraine remnants. They are also well displayed on the flanks of Mt. Leonard.

The first phase, Atlin IIIA, is represented by a sandy moraine (Figs. 16 and 17) with its highest section in this locale being at 3000 feet (914 meters). This moraine is closely associated with what is designated as Atlin IIIB, which is the juncture moraine complex (Fig. 18). In this situation Atlin IIIA ice again moved up the Fourth of July Creek valley to the mouth of Volcanic Creek, but this time at a much lower elevation. A pitted ground moraine representing this stage is shown in Figure 17.

The juncture moraine complex played a significant role in the late-Wisconsinan drainage of the Fourth of July Creek. During this stage the ice front continued to block off the lower Fourth of July Creek valley from its inter-

Figure 16. Atlin III moraine complex

Looking northeast up the Fourth of July Creek Valley from sandy moraine near the Ruffner mine.

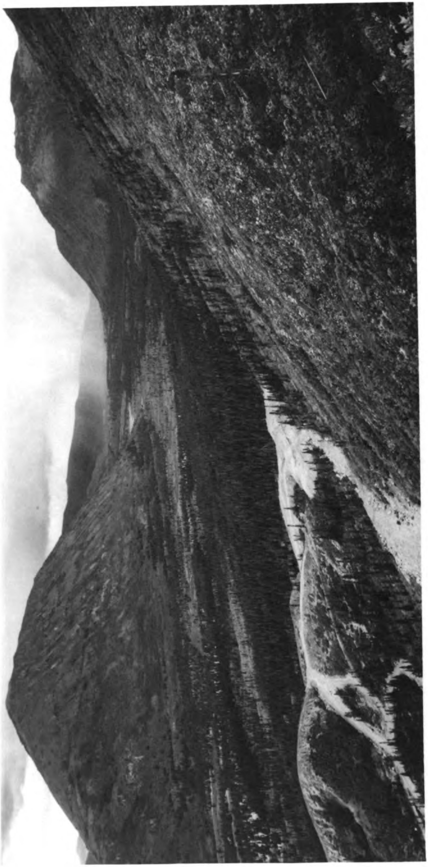

Figure 17. Aerial stereo pair of Fourth of July Creek Valley with Porter Lake Valley.

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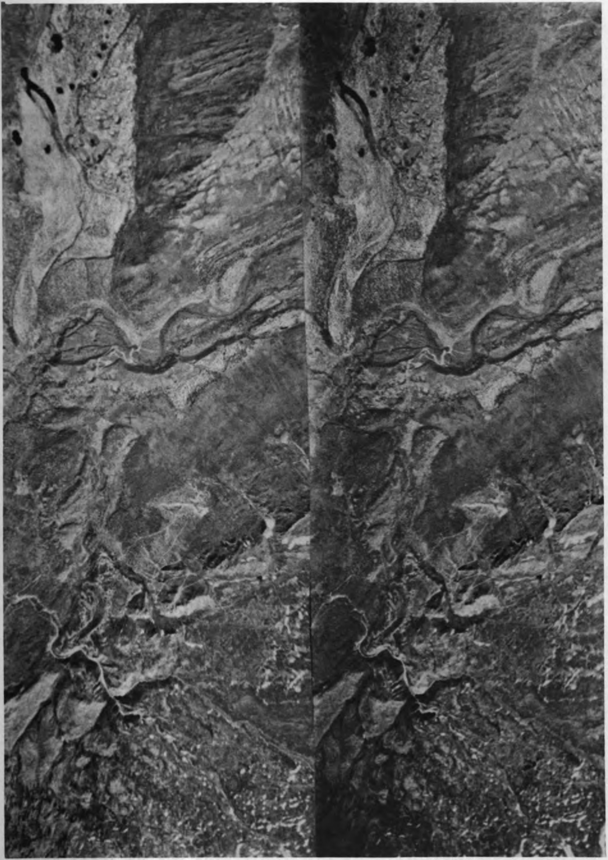

mediate sector until the Porter Lake valley opened to reverse its pre-glacial drainage and permit channeling of meltwater into the Porter Lake depression. For some time ice-cored moraines remained here, veneered with outwash deposited by streams from melting ice in the upper Fourth of July Creek valley. There was some deposition too from feeder glaciers in re-activated high-level cirque systems, as discussed later.

As the ice receded from the juncture zone the massive kame-moraine complex continued to deflect meltwater through Porter Lake valley resulting in deposition of thick morainic material southeast of McDonald Lake. This is designated as the Atlin IIIC moraine, much of which was well-washed by meltwater. Low-level lateral moraines on the southeastern side of McDonald Mountain represent this final phase of glaciation in the juncture area, which we are reminded separates the lower Fourth of July Creek Valley segment from the intermediate valley segment.

The juncture moraine is composed of water-worked till and contains many gently sloping yet still relatively smooth areas dramatically punctuated by kettles and pits revealing that during deposition there was a large volume of meltwater coming out of the upper Fourth of July Creek valley. Some of this outwash, coming from different provenances, buried dead ice zones and so also became incorporated in the juncture moraine. Though no identifiable lake deposits are found today in the intermediate Fourth of July Creek valley above the juncture moraine, an ice-dammed lake is presumed to have existed there until the glacier receding below the juncture opened up flow into Porter Lake valley. A series of fluvial terraces to the southwest and in the Porter Lake valley itself reflect variations in fluvial activity associated with the final stages of deglaciation in the intermediate and upper valley and in the adjoining highlands.

The gradient from upper Fourth of July Creek to the juncture moraine today is about 50 feet (15.25 meters) per mile. This gradient is sufficient to have flushed out drift in the western part of the intermediate Fourth of July Creek Valley. The present gradient from the juncture moraine to Porter Lake is at most 15 feet (4.6 meters) per mile, which of course resulted in extensive valley train deposits being laid down in the Porter Lake Valley. The higher-level fluvial terraces of the Porter Lake Valley are also densely pitted indicating the presence of much dead ice in the valley during the period when glacio-fluvial deposition began.

Correlation of these terraces is discussed in a later section. Overall, this phase of glaciation is considered to correspond to a waning phase which altered with the downwasting of an Intermediate Mountain Ice-sheet (Miller, 1964b) in the Boundary Range.

<u>Atlin IV Stage</u>

A large recessional moraine in lower Fourth of July Creek Valley lies just below Lower McDonald Lake (Fig. 3).

This is evidence for the last significant pulsation of Atlin ice in the Fourth of July Creek Valley. Deposition of this moraine impounded water and formed Glacial Lake McDonald, (Fig. 5) a body of water which is contiguous to the large kame moraine at the juncture of Porter Lake Valley and the intermediate Fourth of July Creek Valley. This recessional moraine and the glacial still-stand it represented was followed by broad deposition of ground moraine during rather continuous and rapid downwasting of Atlin valley ice in late-Glacial time.

During this thinning and retreat drainage of Glacial Lake McDonald was initially through the juncture moraine (Ruffner) which is diagrammatically illustrated on page As ice downwasted and retreated Glacial Lake McDonald drained out via the Fourth of July Creek to a level comparable to the high water elevation of present Atlin Lake and hence was probably reduced to near its present day extent. The reversal of this drainage, which was likely both supra and sub-glacial, was assuredly representative of what was happening to all distributaries in the Atlin glaciation system during the intervals of final downwasting of Atlin Valley ice.

Glacial Lake McDonald was also produced by wasting ice blocks in morainic material near the valley juncture. Groundwater produced from ice in these deposits had to drain to the southwest, while contemporaneous surficial drainage from the upper Fourth of July Creek valley was diverted through Porter

Lake valley.

Atlin V Stage

Evidence for the final phase of Atlin ice in the overall Fourth of July Creek valley is shown by pitted deposits on the gentle floor of the main Atlin valley near the present-day mouth of the Fourth of July Creek (Fig. 3). These deposits lie at an elevation of about 2250 feet (785 meters). That the final retreat of main Atlin valley ice in this region involved systematic downwasting and frontal recession is shown by the presence of drumlinoid hills below the confluence of Two John Creek and Fourth of July Creek (Fig. 18). It is also revealed by the lack of dead-ice features above 2250 feet (785 meters) and by the existence of fluted topography in places down to an elevation of 2500 feet (762 meters). Other than very low-level kettles and associated pits in glacio-fluvial mantles and a few eskers, the area lacks many striking evidences of this ice stagnation.

Local Cirque Activity

One of the most striking features of the Fourth of July Creek valley is the greater thickness of glacial drift on the southeast side of the valley and on the flanks of Mt. Leonard, Mt. Vaughn, Mt. Barham, and on the Mt. Edmund massif. This is evidence that during all of the glacial phases described glaciers originating within nearby high-elevation cirque basins were active and indeed as cirque-headed alpine glaciers played an ultimate role in the volume

Figure 18. Mouth of the Fourth of July Creek.

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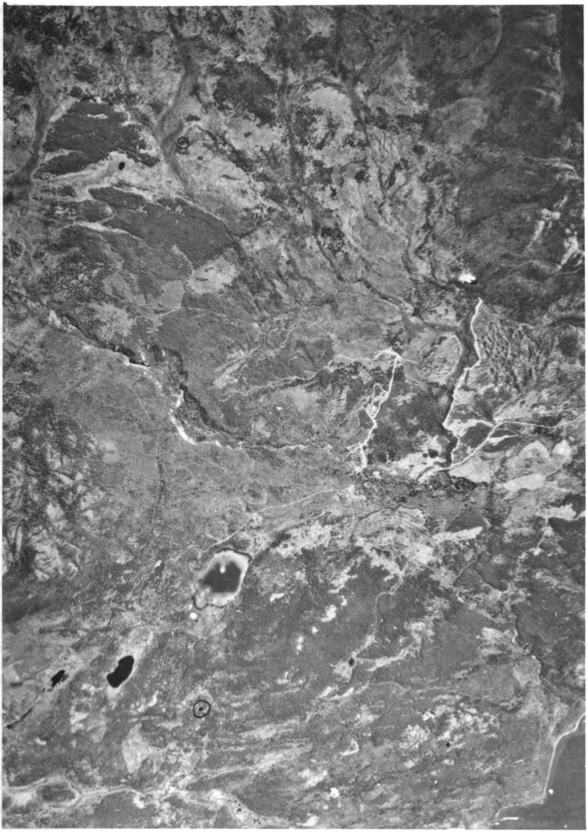

of highland ice involved, and consequently also in the provenance of some of the glacial drift in the lower valleys.

No terminal moraines of high-level cirques were found in the main Fourth of July Creek or Consolation Creek valleys, only ground moraine. Thus, it is concluded that the maximum development of cirque glaciers took place during phases of Mountain Ice-sheet Glaciation when Atlin and/or Gladys Ice filled these valleys. In the waning phases of the glaciation which filled these deep valleys with ice, glaciers from the cirque sources thickened, re-advanced and coalesced, eventually to downwaste again leaving classic dead-ice features behind.

That retracted glaciers occupied the cirques even after all ice had melted from the main lower Atlin and Gladys Lake valleys is evidenced by glacio-fluvial erosion of the youngest lateral moraines in these main valleys, and as well by alluvial fan deposition related to much greater water flow than at any time during the Holocene or at present. This is particularly well illustrated where Consolation Creek reaches the lower gradient of the water divide of the Fourth of July and Consolation Creeks. Here a large alluvial fan greater than one-quarter square mile in area (0.4 kilometers squared) is densely dissected by abandoned channels. The composite of erosional and depositional evidences which has been discussed above confirms that there was intensive cirque activity in the waning phases of

Wisconsinan glaciation. It is also probable that there was re-occupation of some of the high level cirques in Anderson's cool-dry zone II (Table II) 750 to 2500 years B. P., equated to the early Neoglacial, and in the late-Neoglacial Little Ice Age of 300 to 500 years B. P. (Matthes, 1949).

A Sequence of Tandem Cirques

A detailed inventory of cirque levels and the tandem arrangements where they occur in the study area is rendered difficult by the 500 foot (152 meter) contour interval of the Atlin map.* A cirque distribution is obtained, however, by integrating map and air-photo interpretations, but again this is made imprecise by the large contour interval which exists on available maps. Helping the interpretations, however, is the availability of larger-scale (1:50,000) maps in the Atlin Village to Teresa Island sector (Canada Dept. of Mines and Technical Surveys, 1956) where, although out of the study area, comparable cirques can be measured at 100-foot contour intervals. In general, therefore, four distinct cirque levels can be identified. Their average elevations are at 6000 feet (1829 meters), 5500 feet (1676 meters), 4900 feet (1494 meters) and 4100 feet (1249 meters).

In Table III the cirque levels are correlated with levels of coastal cirques in southeastern Alaska as determined in previous studies by Miller (1960) and Swanston (1967). Numbers are assigned on the basis of Miller's

* Canadian Department of Mines and Technical Surveys, 1954; also shown as Figure 3 in pocket.

TABLE III

COMPARISON OF CIRQUE ELEVATIONS IN THE ATLIN REGION
AND SUGGESTED REGIONAL CORRELATION

Southeastern Alaska

Cirque Level	Juneau Icefield and the Taku District Miller (1956)	Prince of Wales Island Swanston (1967)
C 1	300 ft (90 m)	0.500 ft (0-150 m)
C 2	1000 ft (305 m)	650-950 ft (200-290 m)
C 3	1800 ft (550 m) King Salmon-Port Huron?	1050-1350 ft (320-410 m)
C 4	2500 ft (760 m) Tulsequah-Early Valders?	1450-1950 ft (440-590 m)
C 5	3200 ft (975 m) Sittakany-Late Valders?	2050-2650 ft (625-805 m)
C 6	3900 ft (1190 m) Early Holocene and Neoglacial	
C 7	4600 ft (1400 m) Thermal Maximum	

TABLE III

COMPARISON OF CIRQUE ELEVATIONS IN THE ATLIN REGION AND SUGGESTED REGIONAL CORRELATION

Atlin Map Area

Cirque Level	Fourth of July Creek Region	Cathedral Range Miller (1975) Jones (1975)
C 1	______	______
C 2	______	______
C 3	4100 ft (1249 m)	______
C 4	4900 ft (1494 m)	4500 ft (1370 m)
C 5	5500 ft (1676 m)	5100 ft (1550 m)
C 6	6000 ft (1829 m)	5800 ft (1770 m)
C 7		6500 ft (1980 m)

(op. cit.) original designations. As to be expected with this inland positioning the series in the study area averages 2275 feet (693 meters) above corresponding levels on the Juneau Icefield, which, interestingly enough, is close to the difference in elevation today between the mean névé-lines on the coastal versus the continental sides of the Juneau Icefield (Miller, 1975a).

Precipitation patterns in the study area were orographically controlled in the waning stages of Wisconsinan glaciation. The cirque level at the mean elevation of maximum snowfall during the mid-Holocene (Thermal Maximum) is not represented as there are no peaks in the immediate study area above the projected elevation of 6875 feet (2095 meters). In the study area one semi-permanent snow-field was found at an elevation about 6200 feet (1890 meters) which, allied with ice-filled cirques at the same elevation on the Cathedral Massif (Jones, 1975), supports the concept of occupation of cirques in this region at the projected elevation of 6875 feet (2095 meters) during the mid-Holocene.*

Effects of Late-Glacial Climatic Ameliorations and Rising Freezing Levels

The plateau area, which is the drainage divide for the Fourth of July Creek and Consolation Creek, lies at an elevation of 3200 to 3400 feet (975 to 1036 meters). During the

* Of interest is that this elevation (5th interior level) does indeed support a set of higher cirque basins in the vicinity of Camp 26, on the continental flank of the Juneau Icefield - i.e. some 15 miles southwest of Llewellyn Inlet at the head of Atlin Lake (Fig. 2).

Atlin I and Gladys I stages this sector served as a major accumulation zone adding substantially to ice masses from both source areas. In the waning phases ice from the main nourishment zones of these two glaciations thinned, as described in a previous section, but a large volume of later-stage ice from Mt. Barham and Mt. Edmund continued to feed the plateau.

During each climatic amelioration as freezing levels (see glossary) rose and zones of snowfall migrated to higher and higher cirque levels the plateau ice stagnated and down-wasted forming esker complexes and ground moraine kettles, that is on the plateau till plain. In a final phase associated with this deglaciation a rather extensive ice-dammed lake formed (Fig. 5). In several localities subsequent stream cutting during the Holocene exposed silt-clay lacustrine deposits up to at least 15 feet (4.5 meters) thick. No true varved structures were found in these lake sediments. At one exposure lacustrine deposits drape an esker, truncated by post-Glacial erosion. Also these dissected lake deposits lie conformably on and beside sandy, kettled moraines with numerous eskers and crevasse-filling features present.

It is of interest that the ice-dammed lake in this plateau sector formed while ice was still present in most of this high area. Final drainage of the lake followed lowering and recession of Gladys Lake ice to the Gladys III position.

As noted earlier, the thickest drift deposits are confined as an assymetrical wedge on the south side of the upper and intermediate sections of the Fourth of July Creek Valley. This drift was largely deposited by ice from the high-level cirque-headed glaciers to the south. The extent of this cirque ice is impossible to determine except to note that it materially added to the relict Atlin and Gladys Ice resting in the upper Fourth of July Creek Valley. In the final deglaciation all of this ice downwasted together some 10,000 years B. P. (see later discussion with respect to dating methods). It was during this time that the englacial eskers developed, the complex esker system of which is shown in the sketch map of Figure 19.

Two opposing tributary patterns are notable in the esker system and indicate two directions of flow. From this we see that the subglacial drainage divide was some 3 miles (4.8 kilometers) south of today's water divide. The explanation lies in the englacial (ice contact) rather than subglacial nature of some of these drainage ways.

The Fourth of July plateau was the most continuously glaciated region in the whole study area. During the onset of major Wisconsinan glaciation as the névé-line rose high-level cirques increasingly fed the plateau. This was followed by invasions of the main ice-sheet during Atlin and Gladys Ice times. The final phase of glaciation also was strongly influenced by nourishment from the local cirque centers. In the plateau area, however, the extent of

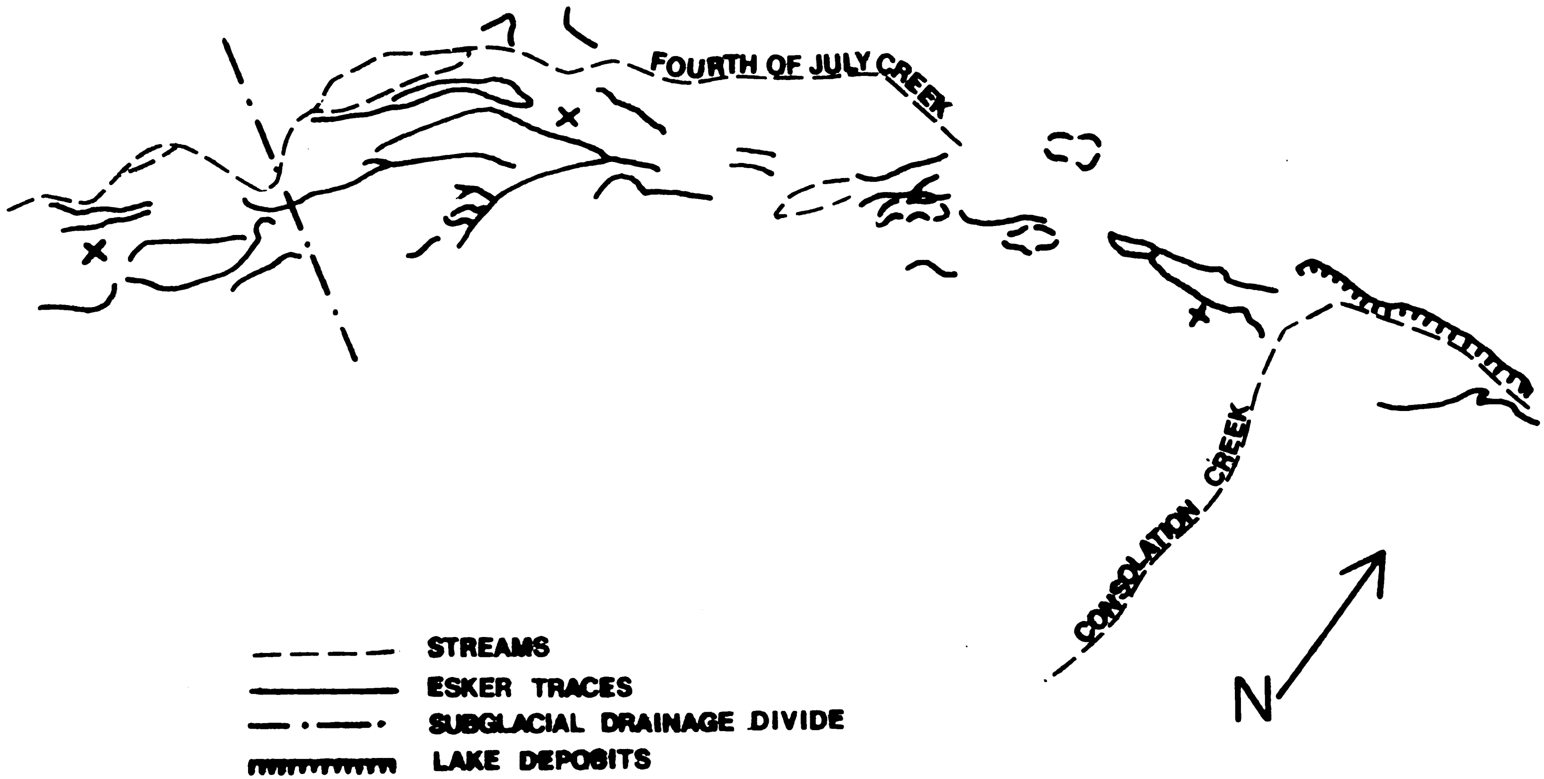

Figure 19. Sketch map of upper Fourth of July Creek Valley

glaciation between early and middle-Wisconsinan time is not known because the evidence was so completely erased by the intensity of late-Wisconsinan glaciation.

Radiocarbon Dates Relevant to Deglaciation

In the upper Fourth of July Creek area alder twigs at the base of a thick peat layer in a peat plateau at about the 3500 foot (1970 meter) level have been collected and dated by radio-carbon technique. A 9315 ± 540 C-14 years B. P. age (Geochron, 1972) on these samples gives a minimum date for the plateau's deglaciation, i.e. that of the upper Fourth of July Creek Valley. A C-14 date was also obtained from basal organic material in a bog along Fourth of July Creek 4 miles (6.4 kilometers) farther down valley and at an elevation of 2900 feet (880 meters). This date is 8050 ± 430 C-14 years B. P. (Geochron, 1973). In this case, however, the lowest contact between organic and inorganic sediments was not reached, suggesting that the beginning of peat formation at this locality was even greater than 8050 C-14 years B. P. In view of the first sample noted above the onset of vegetation following deglaciation is suggested to have taken place somewhere between 9000 and 10,000 years B. P.

History of Glacio-Fluvial Terraces in the Porter Lake Valley

The late-Glacial and Holocene glacio-fluvial history of the Porter Lake Valley is well recorded in an unusually complete set of paired and non-paired fluvial terraces near the junction of the lower Fourth of July Creek Valley and Porter Lake valley (Fig. 17). On foot traverses across this zone a detailed topographic and textural transect was produced. The observations were based on aneroid altimeter readings referenced to the Atlin meteorological station on the shore of Atlin Lake. Details of this transect are given on line AB which is the cross-sectional plot in Figure 20, shown also in plan on Figures 3 and 5b. Differential elevations are plotted as absolutes, and by use of a Wallace and Tiernan surveying altimeter the elevations of each terrace on this transect are indicated to within 25 feet (7.6 meters) of actual elevation above lake level, corrected to a base datum at mean sea level.

During the Atlin III stage, the terminus of Atlin ice was at the juncture of Fourth of July Creek Valley where, for some time, it remained in contact with the narrow drift-filled valley of Porter Lake and the intermediate Fourth of July Creek Valley segment. Evidence for this is given by the massive dimensions of the terminal moraine and the abundance of mixed drift the concentration of which has been described for that sector. It is presumed that ice-cored moraines filled even more of the valley following downwastage of the preceding advance of Atlin II ice. Not much ground moraine is visible today, however, several segments remain and are veneered with outwash from the intermediate Fourth of July Creek valley. These remnants are seen

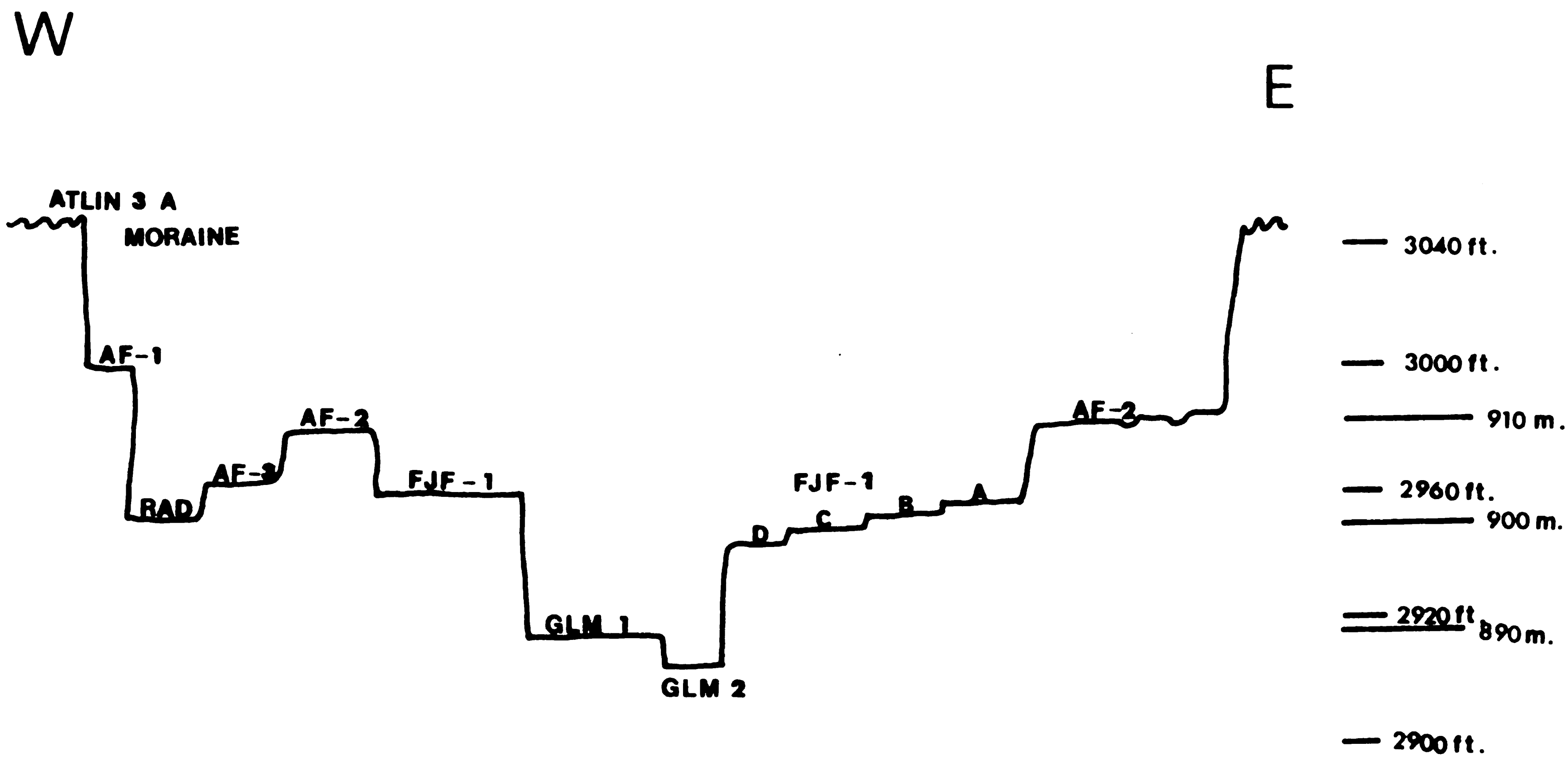

Figure 20. Porter Lake terraces

as a pitted surface on the floor of Porter Lake valley about 2 miles (3 kilometers) northeast of the juncture moraines. The main part of this remnant shows up well in the air photo of Figure 17. This is a boldly pitted terrace of outwash-veneered ground moraine and is graded to the highest fluvial terrace in the proximal zone. It is considered to be Atlin III in age, because the Atlin IIIA glacier tongue contributed previously to the building of the juncture moraine complex. The apparent sequence of subsequent events is shown in Table IV. This table relates to the following discussion, which depicts the total chrono-sequence which was involved in the fluvial terraces shown in the transect sketch of Figure 20 and the allied transect in Figure 21.

The highest and first terrace, Atlin Fluvial I (AF-I) is about 55 feet (16.8 meters) below the moraine laid down on McDonald Mountain and Caribou Ridge during Atlin IIIA time. The only remnant of this oldest level as indicated in Table IV and the cross sectional plot, lies on the eastern flank of McDonald Mountain. This first level was produced by a combination of runoff from Atlin ice and upper Fourth of July ice plus drainage from the local cirque glaciers at higher levels, in consequence of cirque build-ups when freezing levels rose substantially in Atlin IV time.

The next terrace was Atlin Fluvial II (AF-II in Table IVA) which was mapped 20 feet (6 meters) lower. This represents a stage of significant runoff and down-cutting as the Atlin ice experienced a still-stand at the valley

TABLE IV A

FLUVIAL TERRACE SEQUENCE IN PORTER LAKE VALLEY

Feature and Description	Designation*	Approximate Elevation in ft(meters)	Fluvial Provenance	Pertinent Details
Small unpaired terrace	AF-I	3000	Runoff from downwasting Atlin ice and upper Fourth of July Creek Valley	The highest identifiable terrace representing the final significant event when Atlin ice was at its 3A position; initiating the Ruffner moraine system. In this stage the valley train was about a mile across representing a copious flow of water including periodic flooding.
Wide paired terraces	AF-II	2980	Same as AF-I	Represents a major event with Atlin ice terminous remaining in the juncture area building the Ruffner moraine complex with Atlin 3B ice. Again copious fluvial outflow and mantling moraines with outwash thus producing a kame moraine.

* AF connotes Atlin Fluvial; RAD connotes Reversed Atlin.

TABLE IV A

FLUVIAL TERRACE SEQUENCE IN PORTER LAKE VALLEY

Feature and Description	Designation*	Approximate Elevation in ft(meters)	Fluvial Provenance	Pertinent Details
Narrow unpaired terrace	AF-III	2965	Drainage of early Glacial Lake McDonald	Represents Atlin 3C age, retreated phase, initiated early ice dammed Glacial Lake McDonald.
A shortened narrow channel, reversed drainage toward Atlin Lake	RAD	2954	Melting ice blocks from AF-III terrace	Drainage reversal resulting from melting near surface ice blocks from dead ice of Atlin III age. Partial collapse of ice contact faces into Glacial Lake McDonald.

* AF connotes Atlin Fluvial; RAD connotes Reversed Atlin

TABLE IV B

FLUVIAL TERRACE SEQUENCE
PORTER LAKE VALLEY B

Feature and Description	Designation*	Approximate Elevation in ft(meters)	Fluvial Provenance	Pertinent Details
Wide paired terrace with eastern segment having 3 subsequent stages	FJF-IA FJF-IB FJF-IC FJF-ID	2960 2955 2950 2945	Minor runoff from Atlin ice plus meltwater from upper Fourth of July Creek valley and cirque glaciers	Marks the end of significant runoff from Atlin III ice as it retreated to the Atlin IV position. Lesser climatic variations are presumed to be responsible for the multiple terrace levels.
Relatively wide deeply entrenched channel with curving form	GLM-I	2913	Main out-flow of Glacial Lake McDonald	Further ice-block melting in Ruffner moraine complex producing sufficient lowering to permit drainage of Glacial Lake McDonald.

* FJF connotes Fourth of July Fluvial; GLM connotes Glacial Lake McDonald

TABLE IV B

FLUVIAL TERRACE SEQUENCE
PORTER LAKE VALLEY B

Feature and Description	Designation*	Approximate Elevation in ft(meters)	Fluvial Provenance	Pertinent Details
Small entrenched channel on inside of meander of GLM I surface	GLM-II	2905	Final drainage of Glacial Lake McDonald into Porter Lake Valley	Represents a relatively short interval when Glacial Lake McDonald rose to a stage permitting drainage through GLM I, downcutting it further on the eastern side. Notch too abrupt to be considered a point deep on the meander scar.

* FJF connotes Fourth of July Fluvial; GLM connotes Glacial Lake McDonald

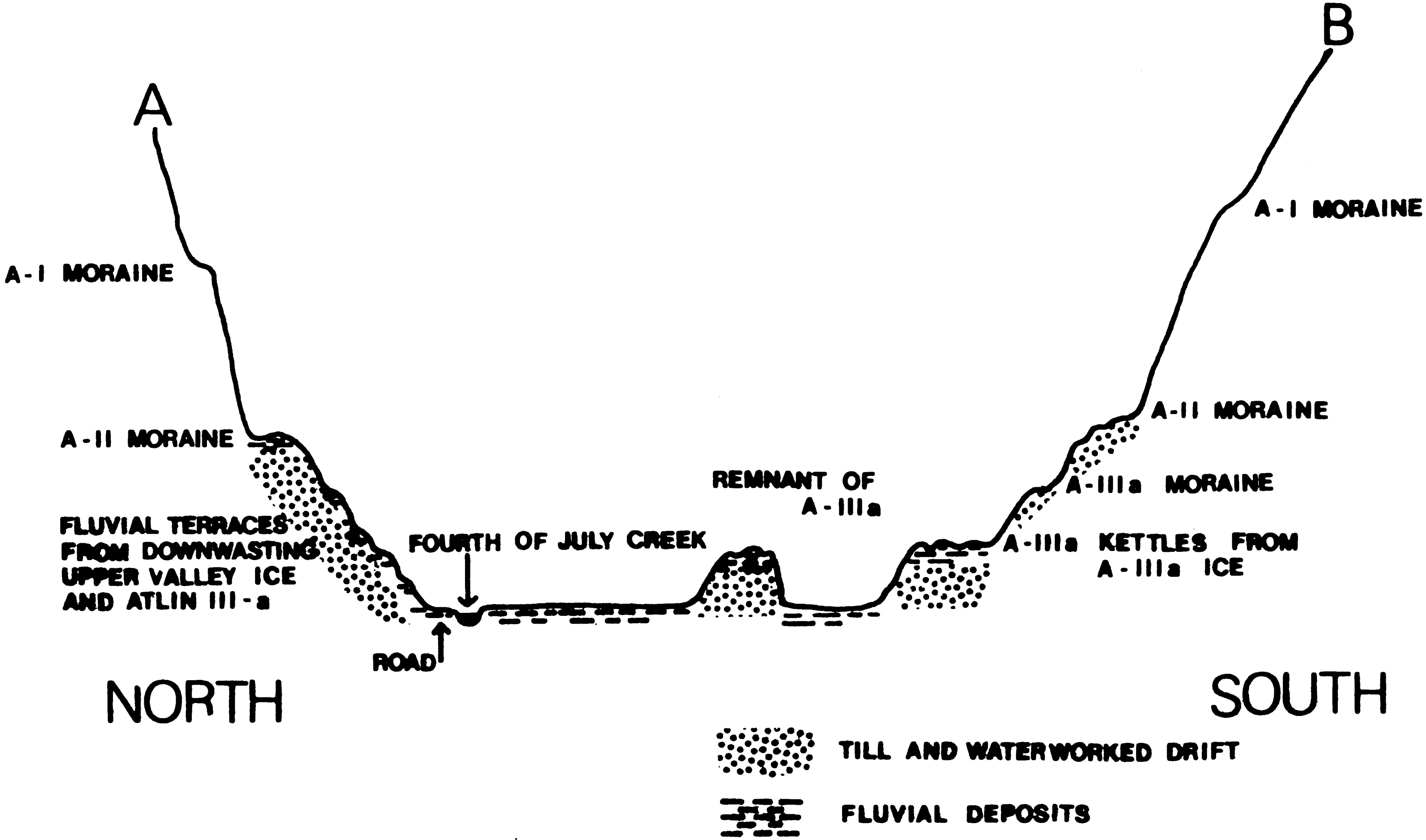

Figure 21. Intermediate Fourth of July Creek transect

juncture. This resulted in final construction of the massive morainic-outwash complex in the Ruffner system. This represents the Atlin IIIB glacial stage, and is correlated with the lower pitted terrace two miles (3 kilometers) farther down the Porter Lake valley (Figs. 3 and 20). This terrace is well shown in the aerial photo view of Figure 17.

As downwasting progressed, assuredly a large volume of water was derived both from melting ice in the upper Fourth of July Creek valley and from that in the retreating Atlin ice in the lower Fourth of July Creek Valley. For a short time, the meltwaters which were dammed in a rather restricted early Glacial Lake McDonald (Fig. 17) succeeded in draining to the north. This outflow from the melt-back of Atlin III ice cut Atlin Fluvial terrace III (AF-III), which we see in Figure 20 on the west side of the valley. As the Atlin ice receded farther and Glacial Lake McDonald further enlarged, the elevation of the lake apparently fell enough that the AF-III terrace level became abandoned. This resulted in a drainage reversal in the immediate area, and formed the reversed Atlin drainage terrace (RAD). This sequence of events was accomplished by headward erosion and slumping as ice blocks which had been buried by the outwash and ice-cored moraines melted out. Thus, this channel (RAD in Fig. 20) was occupied only during the time the local ice source survived. Evidence for this reversal of drainage is given by the fact that the gradient of the now abandoned channel is directed downward toward the southwest.

With respect to the foregoing, water draining southwestward out of the intermediate Fourth of July Creek Valley (i.e. from remaining Gladys and local plateau ice) was sufficiently deflected by the juncture moraine that it curved back to the northeast. As it flowed around the bend, it formed a meander and cut a new channel, designated in the transect as Fourth of July Fluvial I (4JF-1). This, in turn, left an unpaired terrace remnant noted as AF-2. The two distinct streams, AF-3 and 4JF-1, then joined and flowed into the Porter Lake valley. Water levels were by then quite low and the stream very well entrenched. This means it cut its way deeply through the west side of the now densely pitted AF-1 and AF-2 terrace seen to the north of the juncture (Fig. 17).

There are four closely spaced terraces on the east side of the cross section which are designated as 4JF-1 A, B, C, and D. These are all associated with 4JF-1 on the west side and the highest, A, is considered a paired terrace of that level.

Unequivocally, the stages of 4JF-1 represent fluctuations in runoff from what might be termed local ice sources in the upper Fourth of July Creek valley. It was during this time, corresponding to the Atlin III C Stage, that vast amounts of runoff came from the intermediate and upper Fourth of July Creek Valley. But the thick deposits of ice-filled drift which still lay in the Fourth of July Creek Valley, much restricted this flow along the southeast side of the valley and produced terrace remnants on the sides of Caribou Ridge in the intermediate Fourth of July Creek Valley (Fig. 21).

Entrenchment of the drainage in GLM (Glacial Lake McDonald) time is considered a direct result of the severe downwasting and ultimate collapse of the ice-cored moraine complex at the juncture. This was then followed immediately by the drainage of Glacial Lake McDonald (Fig. 5) which emptied into Porter Lake Valley. The bold and decisive terrace remnants seen today suggest that this was a short and catastrophic episode and one which, in fact, initiated the present day drainage of Fourth of July Creek in quite the opposite direction. The final channel level in McDonald Fluvial II time (GLM-II) represents another short-lived still-stand impounding Glacial Lake McDonald, during which time it again drained north into Porter Lake Valley. These neatly paired terrace levels (Fig. 20) correspond to the glacial condition in Atlin IV time when the then steep and receding terminus of Atlin ice not only produced the meltwater for this pro-glacial lake but as well dammed its runoff thus forming Glacial Lake McDonald. As the ice in this terminus retreated farther to the Atlin V position, which was well down-valley and at a much lower elevation toward the southwest, we may assume that Glacial Lake McDonald decreased in depth and ultimately drained southwestward into the present lower Fourth of July Creek Valley.

Today's channel in the Fourth of July Creek thus was opened. This episode marked the end of all late-Wisconsinan glacial drainage through Porter Lake Valley. The reason for assigning a late-Wisconsinan time table to these events is fully discussed in the following sections on the glacial chronology.

As for modification of the geomorphology of the Porter Lake Valley, the terraces and channels described have suffered little change in Holocene time except via the eventual melting out of ice cores and the final development of a vegetative cover. The relatively small amount of precipitation in this region, plus the coarse sedimentological texture which produced good drainage quality in the surface deposits have prevented any further channeled runoff in these valleys, except for the presence of an intermittent stream just north of McDonald Mountain between it and Mt. McIntosh (Fig. 3).

During the sequence of events described in the foregoing it is apparent that the drainage of Porter Lake was essentially controlled by the presence of a natural outlet valley seen on the map (Figs. 2 and 3) as the valley of Tel-Cabin Creek. It is also clear that early high strand lines along Porter Lake extended north and south of the shore limits of the present lake.

Much of the early drainage was also through the lower and broader outlet valley of Indian Creek, from which in all probability the main meltwater regressed directly on to the Atlin Ice in the main Atlin valley, or at least was impounded against it, forming huge side-valley moats north and south of

Halcro (Hitchcock) Peak and Black Mountain (Fig. 3).

During these high ice levels, it is also possible that some of the meltwater drainage was through the valley between Mt. McIntosh and McDonald Mountain as well as from Tel-Cabin Creek, both of which are the next east-west trending valleys to the north. There is, in fact, geomorphic evidence that the McIntosh valley outflow drained across the present water divide. At lower levels, however, the drainage reversed when ice-cores in the drift of the valley melted out, resulting in a lower elevation than the present bedrock divide between Porter Lake and Atlin Lake.

By the time that terraces AF-3 and 4JF were developed the water had reached a sufficiently low elevation that all further drainage was westward via Tel-Cabin Creek. It is suggested that the multiple terraces of 4JF-1 time may well reflect local base-level changes because Tel-Cabin Creek, via the melting of ice, assumed the main role of drainage in this sector and so lowered the level of Porter Lake almost to its present stage.

The lower Fourth of July Creek lacks post-glacial notching and intrenchment from fluvial drainage. This is interpreted as further evidence for diversion of early melt-water through the Porter Lake Valley followed by the temporary base-level control of Glacial Lake McDonald. Thus during the time of most ice melting in the upper Fourth of July Creek Valley there was no channeled run-off in the lower valley. Post-Glacial drainage in the time span since has been relatively minor in this arid region.

Moraines and Glacio-Fluvial Terraces in the Intermediate Fourth of July Creek Valley

The glacial history in the intermediate and upper Fourth of July Creek Valley is recorded by the presence of lateral moraines and glacio-fluvial terraces in both the upper and intermediate valleys. Two composite schematic transects are presented along lines AB and CD in Figures 21 and 22. The general location of these transects is shown on Figure 5b (page 38) and also on Figure 3.

The two most prominent moraines in the lower valley are those of Atlin II and Atlin III ice, each composed of till and veneered with glacio-fluvial sediments. The higher lateral moraine of Atlin II age has a distinct channel between the moraine crest and the valley side of Caribou Ridge. Subsequent to Atlin II time, there was lateral drainage on the north side of the valley as evidenced by pronounced glacio-fluvial terraces on the Atlin III-A lateral moraine. The final downwasting of Atlin III-A ice is represented by a pitted terrace on the south edge of the valley floor (Fig. 17). An isolated remnant of this terrace remains in the center of the valley indicating that this part of the intermediate Fourth of July Creek Valley was once filled with a much greater volume of drift which was later flushed out by the vigorous glacial streams cascading as torrents out of the intermediate valley.

The Atlin III-A terraces are related to the earliest terraces in the Porter Lake Valley (i.e. AF-1 and II and 4JF-1 in Fig. 20). Though at least four such terraces can be traced for several miles, there are other terrace remnants probably relating to the individual stages of 4JF-1 in Porter Lake. Because of slumping and poor elevation control it is impossible to correlate these terraces as individual units.

It should be noted too that there are small fluvial terraces of Holocene age in the present Fourth of July Creek Valley. Figure 22 illustrates the drift deposits in the lower part of the upper Fourth of July Creek Valley. Here eskers become prominent as well as the more massive till deposits associated with the Gladys I and II phases and with the downwasting of local plateau ice. The south wall of the valley contains thickly concentrated pitted outwash mantling a kettle-holed moraine which has been covered with post-Glacial talus off the flank of the unnamed massif to the southeast. The northwest side of the valley also contains a moraine remnant of Atlin II ice which is thought to have been in contact with the late-Glacial ice-cap glaciation formed from combined plateau and Gladys Ice. Evidence for this is a strong linear moraine continuing up valley for several miles.

The glacio-fluvial terraces of Atlin III-A ice first appear just above Volcanic Creek. It is mentioned here that these may be from remaining Atlin II ice combined with plateau ice when Atlin III ice was in the lower valley.

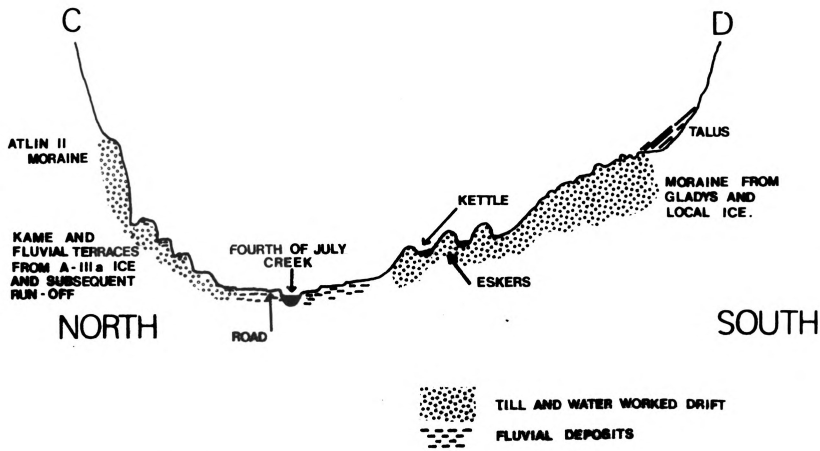

Figure 22. Lower-upper Fourth of July Creek Valley

There is no evidence on the south valley wall that Atlin III-A ice made its way this far up valley.

Finally, it is noted that lesser (younger) terraces in the Fourth of July Creek sequence are also present in the lower section of the upper Fourth of July Creek valley.

Esker-Kame Complex in Upper Fourth of July Creek Valley

The upper Fourth of July Creek Valley is filled with a large and well-preserved esker complex extending just above the Spear Point out-wash mantled moraine system for about 12 miles (19 kilometers) into the Consolation Creek Valley. The terminal zone of this esker complex is shown in Figure 11. The esker system forms a reticulated network of narrow and steep-sided ridges of the classical esker form, none exceeding a length of 2 miles (3.2 kilometers) (photo in Fig. 27). In places other and more non-descript deposits of irregular form cross-cut the eskers, some of the cross-cutting features being elongated and others not, thus making a unique group of overlapping water-deposited or water-worked features. This kind of esker complex has been described by Armstrong and Tipper (1948) as compound eskers and later by Tipper (1971) as esker complexes. In this area they have sometimes been referred to as esker-kame "swarms" (Miller, 1975a). The axes of the ridges and enclosed depressions follow the elongate form of the major eskers in the system and are sub-parallel to the present valley drainage.

The esker "swarm" was laid down in meltwater channels in and under the downwasting ice in the upper Fourth of July Creek Valley. The major continuous eskers, diagrammatically shown in Figure 19 reveal two directions of tributary flow with a small non-esker zone between. This delineates a subglacial drainage divide some distance to the west of and below the present day water divide.

The morphologies of the ridges in the esker complex are divided into three main kinds. The first represents the classic englacial or subglacial esker and in this region these are less than 25 feet (7.5 meters) high and not more than 20 feet (6 meters) wide. They are composed of very coarse and sub-stratified sand to boulders. Almost all of the clastic material shows the strong effects of fluvial sorting, water-scouring and erosion.

A second type is a larger esker-like ridge, 30 to 50 feet (9 to 15 meters) high, containing a higher percentage of sand and gravel-sized sediments, hence with much less coarse material. It is thought that these particular "eskers" were probably formed in open channels between ice walls or as water-laid deposits in wide crevasses during the waning or dead-ice phase of the plateau glaciation. Technically these should be considered as kames.

The third type of morphology involves smaller ridges which are usually not parallel to the main esker axes in the system. These are interpreted as crevasse fillings. They have textures more like the larger open channel "eskers" or kames. The probability is that these are not totally stratified features, but formed as ice-contact deposits which suffered much slumping and redistribution during the lowering of the bottom of crevasses in the final stage of melting.

The area north of Gladys Lake also contains a series of esker complexes similar to those in the upper Fourth of July Creek Valley. Here, however, the esker "swarms" are quite linear with fewer overlapping non-descript forms such as the irregular kames and crevasse fills found in the plateau section. These, however, are related to the main meltwater channels of Gladys IV ice during final downwasting and decay of the ice.

Since esker complexes form in downwasting ice they cannot be associated with vigorous movement. Therefore, this interpretation is in order, i.e. when we find such an array of deposits it indeed connotes wasting ice, changing into a dead-ice phase. For this reason the tributary relationship, in the case of the upper Fourth of July Creek Valley eskers, has been helpful not only in determining englacial and subglacial drainage but, indeed, has served to delineate the final downwasting of plateau glaciation in Gladys IV and Atlin IV and V time.

Valley Asymmetry and Abandoned Channels in the Gladys Lake Depression

Asymmetrical valleys are common in the study area, particularly on Gladys Mountain. As such they represent a unique study in themselves, for the asymmetry appears to controvert the law of divides. The asymmetry concerns angle of repose, relative erosion and slope stability. In the Gladys Lake depression, the abrupt beginning of some tributary streams on the north slope suggests that a thin mass of ice remained on these slopes for some time after the main glacier surface had lowered (e.g. see upper Davenport Creek, Fig. 23). From this, meltwater runoff cut through thick till deposits, resulting in deeply incised channels. The nature of the drainage indicates interestingly enough that many of the gulley streams washed out onto the downwasting surface of the Gladys IV stage glacier in the Gladys depression. The evidence for this is the sharp truncation of Gladys IV ice by the pitted moraine i.e. the strong presence of pit degression surrounded by stratified material.

It is also of significance that the south-facing slopes of these small streams are steeper than the north-facing slopes. Being close to the angle of repose, these south-facing slopes do not support much vegetation. Surprisingly, they have not been susceptible to extensive delevelling by mass wastage. The reason may be the permeable nature of the drift which is comprised mainly of sands and gravels through which saturating ground water readily moves. Also because of differences in insolation, the north-facing slopes are much drier and less affected by cryoturbation and other forms of mass wasting. Even where the outside of a meander has a north-facing slope, it retains a much gentler gradient than the south-facing cliff above the in-

Figure 23. Head of Gladys Lake

Note Consolation Creek flowing into Fish Lake and Davenport Creek flowing into Gladys Lake.

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side of meanders. Low annual precipitation of this region abets the lack of vegetation on the south-facing slopes where, of course, available ground water is slight and summer evaporation great. This has resulted in relatively denser vegetation on the north-facing slopes which have now stabilized their gradients to relatively gentle relief.

Another manifestation of marginal precipitation allied to slope gradient is found on many of the larger eskers, and on the crevasse fillings and non-descript kames in the upper Fourth of July Creek Valley. Here again, steep south-facing slopes are less vegetated and north-facing slopes gentle. All of these features too are composed of highly permeable sand and gravels.

CHAPTER VII

PERIGLACIAL ENVIRONMENT

Periglacial processes have played a large role in the evolution of Holocene geomorphic features in the area of consideration. Some processes are active today while others are revealed by relict forms that show no evidence of present day or even Little Ice Age activity, that is for at least the past four to five centuries.

Solifluction Zones

Frost action and mass wastage have greatly modified valley flanks in the Fourth of July Creek area. Although frost solifluction is still an active process in the tundra zone above timberline (c 4000 ft., 1212 meters) and at higher elevations, talus on steep slopes is common. At lower levels the talus is heavily lichen-covered, indicating relative inactivity and stabilization today. Anderson's study (1970) of Holocene environments in the Atlin region (Table II) defines a relatively cool/dry period 1000-2500 years B. P., the early Neoglacial. This is the classical condition of a periglacial climate.* Thus it is presumed

* The dry/cold <u>periglacial condition</u> of the Atlin region is the antithesis of the wet/cold <u>glacial condition</u> typifying most of the Juneau and Stikine Icefields and the Alaskan Coast.

that frost action and associated mass wasting processes, as well as other periglacial processes, were intensified during this time.

Today no evidence of solifluction is found at elevations below timberline. Above this zone (4000 feet, 1212 meters) numerous solifluction terraces and associated flow-lobes are present.

Nivation Hollows

No semi-permanent snowfields or firn patches are found in the drainage basins of Fourth of July Creek or Consolation Creek valleys. One such local firn patch, however, occurs on Ruby Mountain at an elevation 6000 feet (1830 meters). This pocket of residual firn, lies in the lee of a col and its persistence through recent years makes it, in effect, a glacieret. An abandoned addit of a tungsten scheelite prospect about 400 feet (120 meters) below this glacieret contains large stalactites of ice up to 2 feet (0.6 meters) long and 1 foot (0.3 meters) in diameter. Some 16 feet (5 meters) within the addit, ice crystals extend up to 2 feet (0.6 meters) from the walls and ceiling. Large hexagonal crystals of sublimation ice reach 6 inches (15 centimeters) in diameter. Clearly this is in the permafrost zone, i.e., above 5500 feet (1575 meters).

As we have seen in the foregoing discussion of climatic parameters, at the higher elevations of this region the mean annual temperature is well below freezing, thus producing appropriate conditions for permafrost development below the surface and for nivation processes at the surface. In fact, at the permafrost seep level, about 5500 feet (1670 meters), the mean annual temperature, extrapolated from a prevailing lapse rate of 3.3°F/1000 feet, is 21°F (-9°C). Therefore, in this region a minor climatic deterioration can readily produce a re-occupation of high-level nivation hollows and cirques.

Within the study area, a number of nivation hollows are present at elevations down to 5000 feet (1525 meters)... i.e., on Mt. Vaughan (Fig. 24) just above the juncture of the Fourth of July Creek Valley and the valley to Porter Lake. These hollows are strikingly apparent because of the lack of lichen on rocks formerly covered by snow and/or ice. In view of the very fresh surfaces in the bare rock areas and rubble zones exposed by de-nivation it is apparent that the removal of firn from these nivation hollows does not relate in time to Anderson's Zone II (Table II) but rather to a very recent warm period following a general cooling. As this condition is cyclic and the climate is now returning to a colder phase (Miller, 1972; Miller and Anderson, 1974a), these sensitive firn patches and glacierets should be monitored. They will certainly respond to the continuing cooling, and in fact already there is re-development of nivation patches in many of the recently-bare depressions at and above timberline, a trend currently observed on the Alaskan Coast as well (JIRP and also the U. S. Forest

Figure 24. Nivation hollows on Mt. Vaughn. Photo by M. Miller.

Figure 25. Relict stone circles in upper Fourth of July Creek valley. Photo by M. Miller

Service, communication from the Forestry Sciences Laboratory in Juneau).

It is of regional significance, therefore, that even in the last five summers an increasing number of snow patches have been observed on the hills east of Atlin, with the late-summer firn being retained as a base for further accumulation in the on-coming autumn and winter seasons.

A number of sorted and unsorted stone circles, stone stripes and stone polygons are found in the study area. These are both as relict forms and as ones active today. The best developed <u>relict</u> stone circles are found in the intermediate Fourth of July Creek Valley at an elevation of about 3200 feet (975 meters) (Fig. 25). These circles, about 3 feet (1 meter) in diameter, are now often under water during the summer and are positioned well inside of the swarm of eskers shown in Figure 19. They are considered to be relicts from the earliest Holocene (ca. 7000-9000 years B. P.) but they also may have been affected by cooler, drier climate of the early Neoglacial 2500 years to 1000 years B. P. The rationale for an early Holocene age assignment is the fact that similar relict stone circles have been found at 2500 feet (760 meters) on the Alaskan coast and have been with some confidence assessed as very early Holocene or late-Glacial in age (Miller, 1975a). Regardless of vintage, it is clear that increased temperatures and higher water tables succeeded in terminating the growth of these forms.

At elevations above 5000 feet (1524 meters) quite active and well-defined circles are found. The circles at or above the 5000 feet elevation, are seen to contain frost-heaved particles no larger than pebble size. Most of these circles average a half meter in width. Those at 6000 feet and above are 4 to 5 (1.2-1.5 meters) across and have been seen in abundance from the air. Details of the texture and morphology of these features where they occur in the study area are not yet known, however, some studies have been carried out on comparable features at 5000 to 7000 feet (1524 - 2121 meters) elevation on Teresa Island near Atlin (Buttrick, 1974) and at Camp 29 on the Cathedral Massif west of Torres Channel (Jones, 1975)

Palsas and Related Frost Mounds

The Atlin district lies directly south of the discontinuous permafrost limit (Brown, 1970). In spite of this, perennially frozen ground has been found to be common in poorly-drained bogs at elevations considerably below the permafrost seep level. For example, in the study area this is well evidenced by the presence of circular to elongate mounds of peat and/or mineral sediments containing a partially frozen core. It is of special interest that these mounds are found as much as 2000 to 2500 feet (610 - 670 meters) below the local permafrost level (seep level). The common occurrence of frost mounds puts the Atlin district in the sporadic permafrost zone.

Frost mounds of the type common to the study area have been investigated extensively in Scandinavian countries and are given the Finnish name palsa (Maarleveld, 1965; Seppala, 1972a and 1972b; and others). Larger areas of frost mounds showing raised relief are referred to as palsa plateaus or plate palsas (Kujansuu, 1969). Several of these are found in both the intermediate and upper Fourth of July Creek valleys (Figs. 26 and 27).

Origin of Palsas

The origin of palsas and palsa plateaus is dependent on temperature, water supply and amount of snow cover as well as on physical properties of the sediments involved. The process involves a peat bog which completely freezes over in the winter and forms domes from ice lenses and frost action at random locations below the surface of the bog. Once the bog water freezes this freezing extends into the saturated peat and underlying silt. The raised peat mound then grows and becomes drier. Containing many air pockets in the vegetation it also becomes a better insulator to atmospheric heat in the summer months.

The initial exposure of peat and its subsequent drying seems to be the most critical step in palsa formation. Ideally increased precipitation in the fall and decreased evapo-transpiration leads to increased moisture in the raised mound in the winter. Because of wind scour, snow is less likely to accumulate on the raised palsa mounds than the surrounding bog, thus decreasing its insulation in the

Figure 26. Palsas in the upper Fourth of July Creek Valley

Figure 27. Eskers and peat plateau in the Fourth of July Creek Valley

winter months and actually causing an increased cold penetration, with consequent enlargement of the extent of frozen ground and its allied ice inclusions. Resulting too is a consequent increase in palsa height. As the palsa grows, the peat and/or mineral soil breaks away from the sides and eventually forms large cracks across the mound sometimes exposing the frozen core. The height a palsa reaches and the length of time it can exist are functions of the peat thickness and the consistency of the peat or mineral sediments exposed, as well as of the climatic conditions.

Palsas commonly occur in bogs where the mean annual temperature is less than 33.6^{O}F (1^{O}C). They are recorded in northern Sweden where the mean annual temperature remains below 32^{O}F (0^{O}C) during more than 200 days per year and are present only where precipitation during the period of November to April is less than 14 inches (31 centimeters) (Lundquist, 1962).

Character of the Atlin Palsas

Three prototypical localities lie in the study area and there are many others in the Atlin region e.g. at the head of O'Donnel River (Fig. 3). The first locality dealt with in this investigation is the bog sector at the drainage divide of Fourth of July Creek and Consolation Creek. Here there are two well-developed and active palsas and two sectors of raised peat plateaus (Fig. 26 and 27). The extent and depth of sub-surface ice in these frost mounds were studied using the electrical resistivity technique. Specifics of this geophysical method are described later. Elevation of this particular palsa area extends from 3300 to 3400 feet (1005 to 1035 meters).

Another sector was investigated one mile (1.6 kilometers) southwest of the highland zone and at about 50 feet (15 meters) lower elevation. Here several small frost blisters were studied (Fig. 19) which appeared to represent the beginning of palsa formation. Here the ground was frozen 8 to 12 inches (25 to 30 centimeters) below the surface for a thickness of 10 to 12 inches (25 to 30 centimeters). These mounds are not considered to be remnants of deflated palsas as there is no evidence of the characteristic vegetation of the well-drained and mature palsa surface. Observations over a three-year period have shown only minor fluctuations in the depth of permafrost in these palsas and in their heights above water table. One year when the bog water level was unusually high the frozen material was much less in both aerial and vertical extent. This points out the critical affect the water table has on palsa formation. As freezing takes place at or near the water table, a rise of the water table can easily destroy small palsas.

The third palsa locality is still within the esker complex but about 1½ miles (2.4 kilometers) farther down the valley, and another 50 feet (15 meters) lower in elevation than the frost blisters. Here a very large and well-developed palsa was found to be raised about 10 feet (3 meters) above the bog surface. Its size was 150 feet (45 meters) by 40 feet (12 meters) (Fig. 19). The depth to permafrost was variable but averaged 15 inches (38 centimeters). Active sloughing was seen to be occurring in places on the sides of the palsa but much of the flanking area was well vegetated indicating that vigorous active growth is not at present taking place. The fact that all of these palsas lie well inside of the late-Glacial esker swarm connotes an age of less than 10,000 years. Confirmation of this is provided by C-14 samples discussed later.

The electrical resistivity method was used to determine the depth and thickness of permafrost in the highest palsas and peat plateaus of the water divide sector described above.

Electrical Resistivity Assessment of the Fourth of July Creek Palsas

The freezing of interstitial water has little or no effect on the density, magnetism, and radioactivity of rocks or soil, but this can result in large changes in the velocities of compressional seismic waves and in the measured electrical conductivity characteristics (Timur, 1968). As seismic and electrical geophysical methods have been used extensively in permafrost delineation (Barnes, 1963), the electrical method was applied in this interpretation.

Before considering the geophysical data some basic considerations are reviewed. First, the resistivity or inverse conductivity of sediments is a function of the amount and electrolytic nature of water present in the interstices. It decreases as the water content increases. Conversely when water freezes in the sediments, there is a marked increase in resistivity. Also the resistivity of pure ice increases with a decrease in temperature. Keller (1966) has shown that the resistivity of a porous sandstone increased by 184 times when the temperature is decreased from 20°C to -12°C.

In frozen sediments, the salinity of the water also has a complex effect on resistivity. Salts lower the freezing point and when freezing begins salts migrate to unfrozen water, thus increasing the salinity there and decreasing the resistivity of the liquid water present (Keller and Frischknecht, 1966).

Resistivity techniques have been used to map upper permafrost surfaces (MacKay, 1970). The depth to permafrost can be determined quite accurately because the primary control of resistivity is the change of water to ice at the permafrost-active layer interface. Frischknecht and Stanley (1970) were able to get good depth soundings in areas not seriously affected by lateral variations and the thickness of permafrost was determined where the current completely penetrated the frozen ground. Because of the multiple controlling variables within the permafrost (i.e. relative amounts of ice, the electrolytic nature of water and temperature) resistivity measurements cannot be used for positive identification of sediment type at depth. Only with ample direct observational data can the real nature of sediment character be inferred from resistivity measurements.

In areas of relatively thin and sporadic permafrost, the resistivity method offers a quick and inexpensive reconnaissance method not only for mapping permafrost surfaces but also for determining thickness of the sub-surface ice. It is particularly useful in palsa investigations (Tallman, 1973).

Within the highest swampy area of the upper Fourth of July Creek Valley, the most active palsas have a height of about 10 feet (3 meters) above the bog surface (Fig. 26). Each palsa surface is covered with about 24 inches (60 centimeters) of peat and the depth to permafrost varies from 12 to 30 inches (30 to 75 centimeters) where frozen peat or silt and clay is encountered. At about 4 feet (1.2 meters) below the surface, lenses of ice 2 centimeters thick are common and the quantity and thickness of ice lenses increases with depth. Excavation was only made to a depth of 6 feet (1.8 meters) from the palsa surface and did not reach the water table.

Also present is a larger and well-defined peat plateau (Fig. 27). This had a broad flat surface 6 feet (1.8 meters) above the general level of the bog. This plateau was found to have a thinner layer of peat ranging from 6 to 12 inches (15 to 30 centimeters). Here the depth to permafrost was

12 to 18 inches (30 to 45 centimeters). In the upper four feet (1.2 meters) it lacked the isolated lenses of ice found in the active palsa.

Other slightly raised permafrost areas, with less well-defined limits and more undulating surfaces, are also present in the investigation area. These areas are prominent because of the distinctive vegetation on the well-drained permafrost sites.

Details of the Resistivity Method and Computational Technique

The resistivity measurements were made using an expanding Wenner configuration (Fig. 28) with four equally-spaced electrodes. The potential difference was measured between two points on the ground, B and C. A center electrode (Lee partition) was added to measure the potential difference between B and E and C and E, or the right and left half of the array (Fig. 29). These additional readings permitted comparison of the apparent resistivity on either side and also served as a check on accuracy of the measurements. In general, little variation was found in the left and right half of the configuration, and only the apparent resistivity in the full circuit is plotted for analysis.

A direct current instrument was used which gave the value of $\frac{V}{I}$ where V is the voltage across the potential electrodes, and I is the current flowing through the ground. Both forward and reverse current readings were taken, and the average value was used to calculate the apparent resistivity

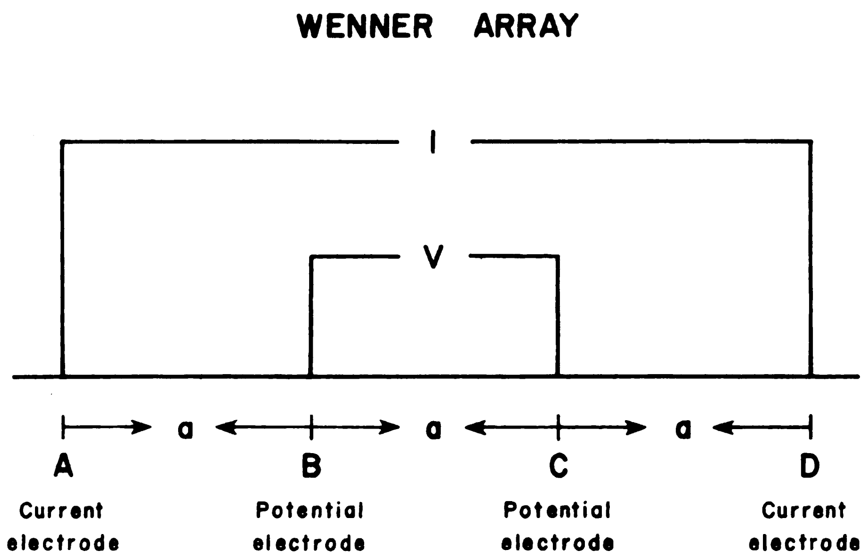

Figure 28. Wenner Array

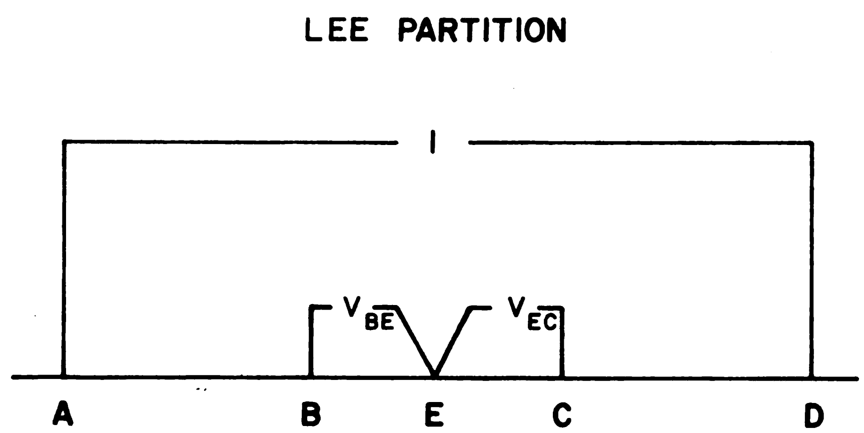

Figure 29. Lee Partition

from the formula:

$$A = 2\pi a \frac{V}{I}$$

where: A = apparent resistivity

a = "a" spacing of Wenner array

π = 3.1416

and: $\frac{V}{I} = \frac{\text{potential difference}}{\text{current}}$, read directly from the meter.

The "a" spacing is assumed to be the depth of current penetration.

In all, six profiles of apparent resistivity were plotted against "a" spacing. Their general location is shown on Figure 30. Profiles 1 and 2 (Fig. 31) are in locations where there is no surficial evidence of permafrost. Profile 1 is across a well-drained area of coarse sand and gravel. Profile 2 is across the bog with a surficial layer of damp to saturated peat. The apparent resistivity was greater in the well-drained profile but, with increasing depth the two profiles were very similar. It must be kept in mind that apparent resistivity at depth is an average of the resistivity of the entire sub-surface from ground to depth of current penetration. There is no indication of a burried high resistance layer in either profile.

Profile 3 (Fig. 32) was made across an active palsa mound about 120 feet (36 meters) long. The Wenner array was expanded from a 2 to 40 feet (0.6 to a 12 meter) "a" spacing. By stopping at the edge of the palsa, complications due to change in relative elevation and the effect of electrodes in

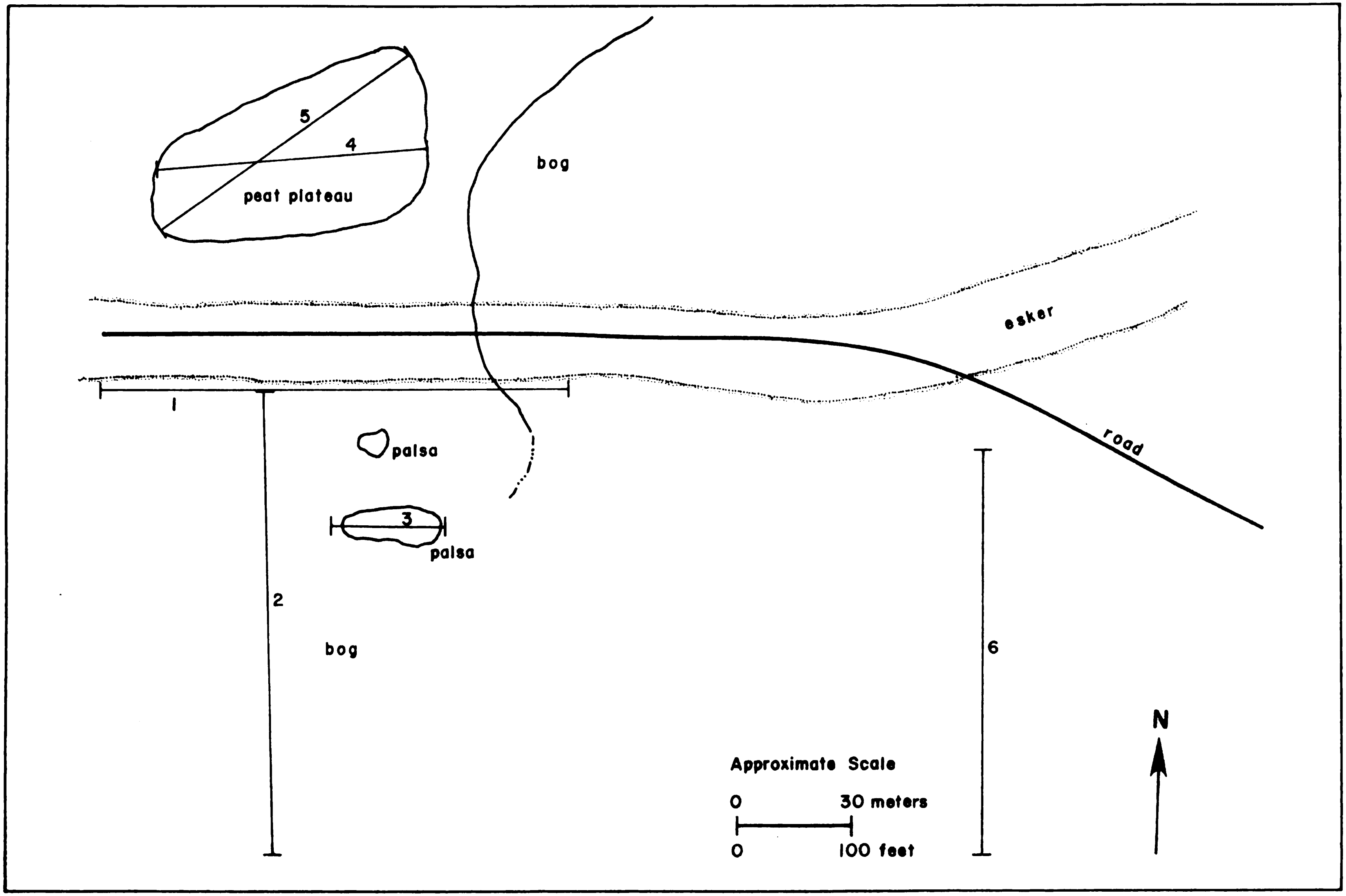

Figure 30. Map of resistivity profiles

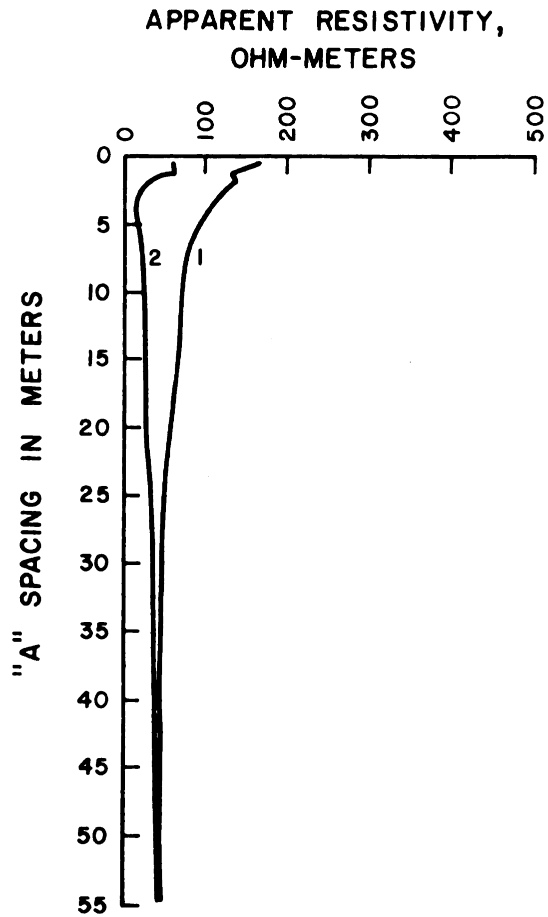

Figure 31. Apparent resistivity, profiles 1 and 2

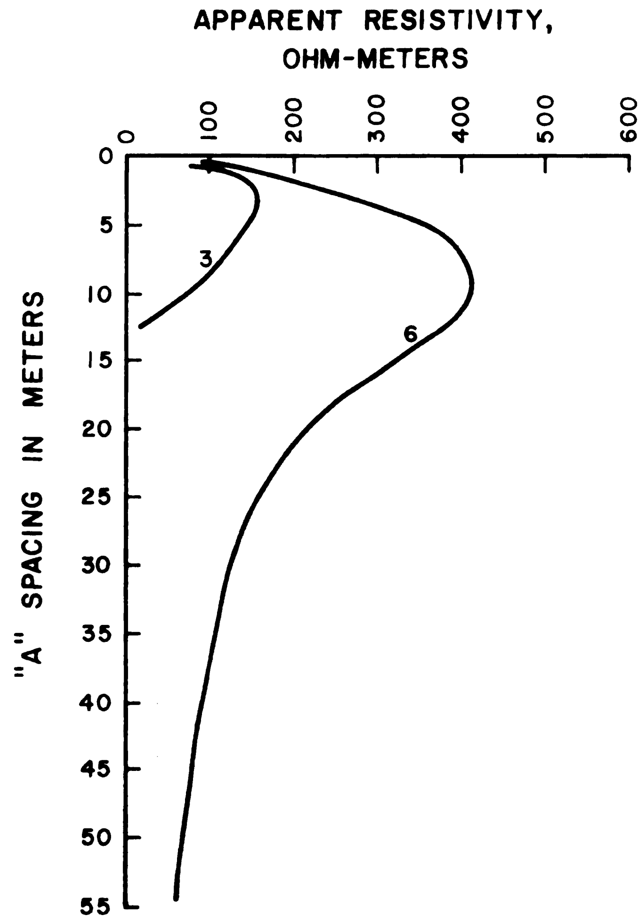

Figure 32. Apparent resistivity, profiles 3 and 6

the surrounding bog water was eliminated. The sharp decrease in apparent resistivity in profile 3 indicates that the limited "a" spacing was sufficient to penetrate completely the layer of high resistance.

Profiles 4 and 5 (Fig. 33) were both made diagonally across the highest peat plateau in this sector. Again, it was possible to expand the array to the limits of the mound and have sufficient current penetration to go beyond the high resistivity layer. It should be noted that apparent resistivity was greater in profile 5 indicating a higher resistivity at depth and/or a greater thickness. A high resistivity could be due to differences in temperature as well as the quantity of ice at depth.

Finally, profile 6 (Fig. 32) was across a slightly raised hummocky area. Test-pits dug here showed a wide variation in the thickness of the active layer and also considerable lateral variation in sediment texture.

<u>Interpretation of the Geophysical Data on Sub-surface Ice</u>

Profiles 3 to 6 all showed high resistance layers at depth. The maximum apparent resistivity in each profile varied greatly. This variation is not only a function of the actual resistivity at depth but it is also dependent on the ratio of the thickness of the frozen layer. As successive measurements are taken, each portion added at the bottom decreases in its relation to the entire mass being measured. Clearly, resistivity changes near the surface cause a much

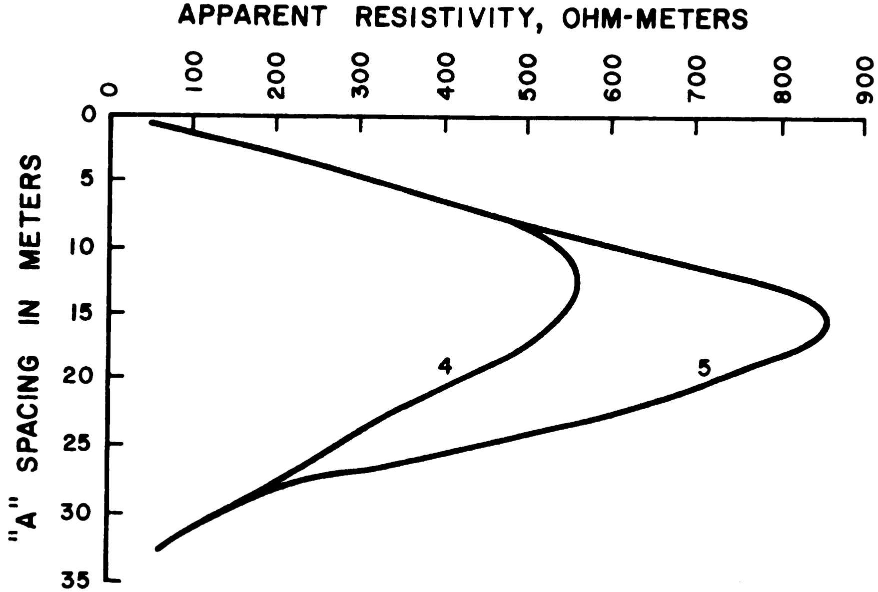

Figure 33. Apparent resistivity, profiles 4 and 5

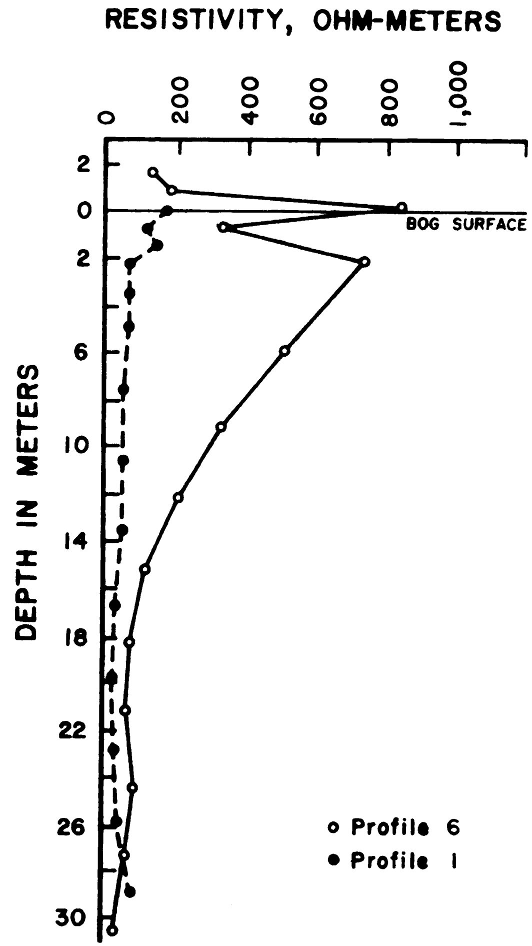

Figure 34. Resistivity, profiles 1 and 6

greater change in apparent resistivity than it would deeper in the substrate.

The apparent resistivity values were plotted against "a" spacings on log paper and compared to Mooney and Wetzel's (1956) theoretical curves for 1, 2, 3, and 4 layers. Because of the gradational variation of resistivity with depth in the frozen layer, it was not possible to match the curves accurately.

The method employed for interpretation was the Barnes layer method (Barnes, 1954), devised to minimize the masking effect of the overlying material. Assuming an "a" spacing of 3 feet (1 meter) the first increment measures the resistivity of the volume of material to a depth of 3 feet (1 meter). The 6 feet (2 meters) increment measures resistivity to a depth of 6 feet (2 meters) and includes the previous 3 feet (1 meter) increment plus an additional 3 feet (1 meter) layer. This can be considered as two resistors in a parallel circuit where the conductance of one resistor is known (the top 3 foot (1 meter) increment) and the total resistance is known (the 6 feet (2 meter) increment). It is thus possible to solve for conductivity of the bottom 1 m layer by the following formula:

$$\frac{1}{R_n} = \frac{1}{R_n} - \frac{1}{R_{n-1}}$$

where: $\frac{1}{R_n}$ = layer conductance of a given increment, in mhos

$\frac{1}{R_n}$ = total conductance between ground surface and bottom of given increment, in mhos and: $\frac{1}{R_{n-1}}$ = total conductance between ground surface and the bottom of the increment just above given increment, in mhos

Layer resistivities were calculated for all six profiles using the smallest "a" spacing available. These are plotted in Figures 34 to 36 at the mid-point of each layer.

The layer resistivities calculated for profiles 1 and 2 differed very little. Profile 1 (Fig. 34) showed a slightly higher resistivity value near the surface due to drier surficial material. Only profile 1 is plotted and used to aid interpretation in profiles where permafrost is present. The sharp decrease in resistivity in the 2 to 4 feet (0.6 to 1.2 meters) layer is interpreted as delineating the water table in this high bog sector.

Profile 3 (Fig. 35), which transects the most active palsa is plotted 9 feet (3 meters) above the bog surface. Here the maximum resistivity is above the bog surface and falls off to a value comparable to that of unfrozen sediments in this area at a depth less than 6 feet (2 meters) below the bog surface. This suggests that freezing is currently taking place at or slightly below the water table.

The layer resistivities of the peat plateau (profiles 4 and 5) are shown in Figure 36. Again, the resistivities increase rapidly below the first 2 feet (0.6 meter) layer and remain within the frozen ground range to a depth of 55 to 70 feet (17 to 21 meters). The wide variation of resistivities within the permafrost could be a function of differences in porosity, temperature variation, pockets

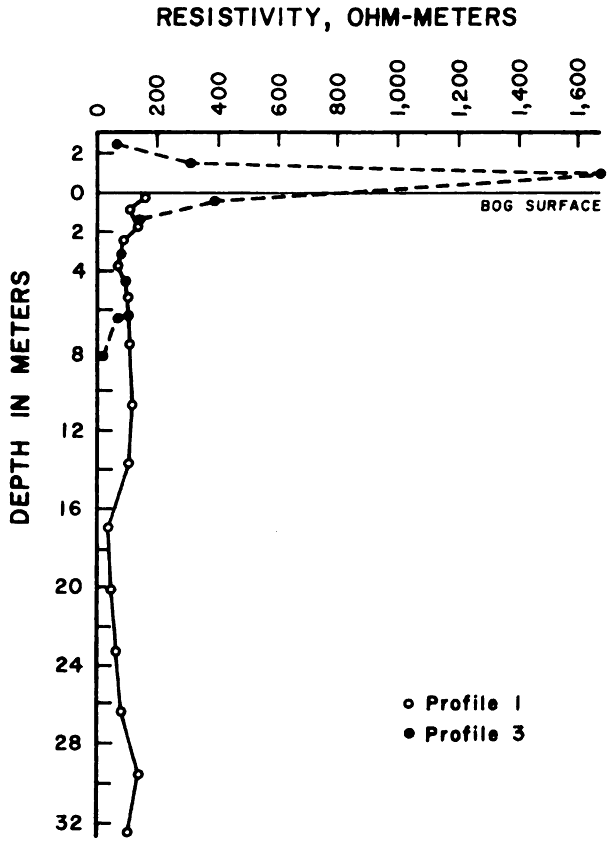

Figure 35. Resistivity, profiles 1 and 3

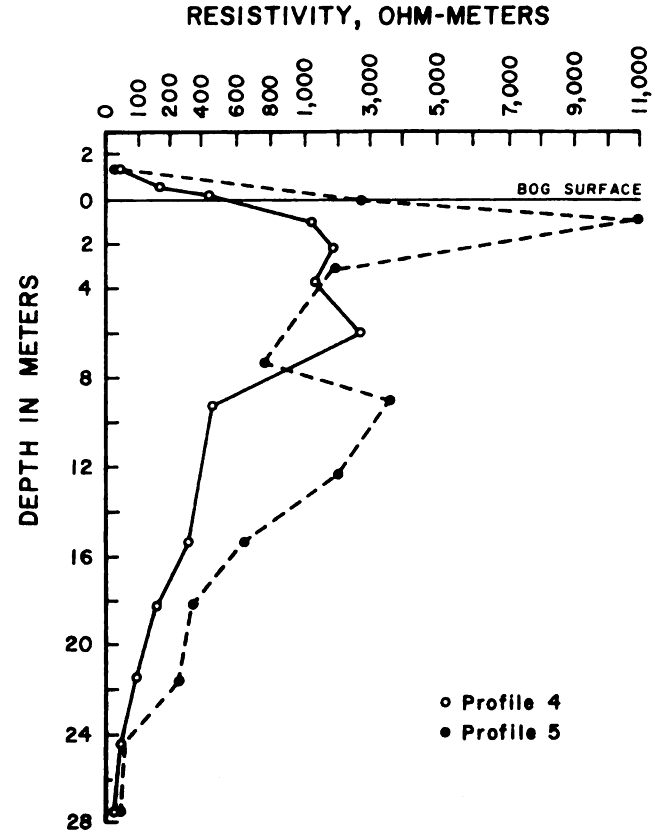

Figure 36. Resistivity, profiles 4 and 5

of unfrozen ground, and/or isolated ice masses. The substantially greater thickness of permafrost in the peat plateau sample suggests a different genesis than that of the round palsa denoted in Figure 35.

The final profile (6 in Fig. 34) again shows a marked increase in resistivity but at the 4 to 6 feet (1.2 to 1.8 meters) layer. Here the resistivity at depth was not as great as in the other permafrost profiles but was still far greater than in non-permafrost areas, and well within the range of resistivities of standard frozen ground. The wide variation in resistivities within the permafrost range is considered to reflect the uneven distribution of frozen sediments found in the pits dug. The smooth decline of resistivities, however, makes it more difficult to determine the lower permafrost boundary. Comparison with the non-permafrost profile suggests the boundary to be between 30 to 40 feet (10 and 13 meters) below the main bog surface.

In general it is known that resistivity ranges within permafrost vary greatly. It has not been the purpose of this study to try to correlate resistivity ranges with the specific sediment type with which we are here concerned, but rather to delineate the depth and extent of the subsurface freezing in this sector at considerably lower elevation than the regional permafrost level. In a geomorphological sense, one usually has a knowledge of the underlying sediments, as we do here, so that this helps the interpretation.

It is now clear that resistivity readings taken in non-permafrost sectors can add useful information about the substrate and be valuable to the interpretation of depths of localized permafrost profiles. With this information, we now have a clearer understanding of the delineation of sporadic permafrost pockets in this discontinuous permafrost zone at intermediate elevations in the Atlin region.

Significance of the Atlin Palsas and Related Considerations

The plate palsas or palsa plateaus described and discussed in this study are considered to be examples of relict permafrost. Their presence in this region suggests that they are likely the last remains of what must have been a far greater extent of permafrost. In the same vein, with the present short-term cooling trends to the end of this century (Miller, 1972), they appear to reflect a re-expansion of the permafrost zone.

That actively growing palsas are forming at this elevation and latitude under present climatic conditions is indeed significant, as no palsas have heretofore been reported at or below this parallel in the Canadian sub-Arctic. Also made clear by this study is the fact that the initial exposure of peat by differential upwarp is a critical step in palsa formation. Furthermore, these observations suggest that the level of bog water tables in palsa areas is extremely important to the genesis of such features. For example high water tables may prevent sufficient drainage in the upwarped peat and result in rapid melting of subsurface ice in summer. If the water tables are too low there is not sufficient water transport through the sediments for the thin interstitial ice layers and thicker ice lenses to form.

Textural variations within such bogs may also exert an important control over the transport of capillary water and hence over the initial differential upwarp. In this sense that part of the study carried out to determine the textural variations in unfrozen bog sediments and palsa sediments were somewhat inconclusive. In a follow-up program it is anticipated that more concentrated research aimed at specific characteristics can be accomplished on this interesting aspect of palsa formation.

CHAPTER VIII

A WISCONSINAN CHRONOLOGY

Keys to the Interpretation

These investigations in the Atlin area reveal three major glacial phases followed by three lesser phases in the interior region of the Alaska-Canada Coast Range. Each represents a significant and identifiable glacio-climatic event which took place during Wisconsinan time at the northern end of the Cordilleran Ice Sheet. Undoubtedly, there are others that have not been identified, but the main lines of the sequence seem clear. It should also be noted that these glaciations are the result of generally contemporaneous accumulations in two distinctly separate but contiguous regions, one in the Cassiar Range and its flanking plateaux* to the east of the study area and the other in the Boundary Range to the south and southwest. Discussion of the recognized phases of glaciation and their designations into stages follows. The key to understanding the complex interrelationships and to the interpretations which follow is the recognition and appreciation of the different provenances of the ice involved. A conceptual model for this situation is considered in the final chapter

* The Teslin, Kawdy and Taku Plateaux. (Bostock, 1948). As a part of the greater interior plateau.

based on certain conclusions derived from the long-term glaciological investigations which have been carried out on the Juneau Icefield by JIRP personnel.*

The provisional chart of the Wisconsinan and Holocene stratigraphic sequence and glacial chronology in the Northern Boundary Range, Alaska-Canada is presented in Table V in the pocket of this volume. It should be appreciated by now, that in the study area many details and evidences of the actual chronology involved have been eradicated by the intensity of later events. The outline does, however, provide a helpful guide to understanding the complex glacio-geomorphologies of the three segments of the Fourth of July Creek Valley with which we have been particularly concerned, and which are so well distinguished by different glacial and fluvial limits. The chart is also most useful in delineating the Pleistocene chronology of the Atlin region at large. Conversely, study of the geomorphology of both the smaller and larger areas has abetted development of the regional chronology.

For some key interpretations, evidences from outside the immediate study area and even beyond the Atlin region are invoked. Thus an effort is made in Table V to compare the chrono-sequence with those found in the Taku District immediately to the south and west and in other adjacent sections of British Columbia, the Yukon and Alaska. Lastly, some teleconnectional inferences are drawn through comparison

* The Juneau Icefield Research Program.

with the Wisconsinan chronology of the Rocky Mountain region, the Puget Sound region of Washington State, and the classic mid-continent region of the Great Lakes. The Puget Sound chronology is perhaps of more direct significance in that it is to the southern end of the Cordilleran Ice-sheet what this study and its suggested chronology are to the northern end of the Cordilleran Ice-sheet. Allied, of course, with this is the Taku District chronology, including the interpretations from the Tulsequah and southern Atlin region (Miller, 1956).

Pre-Atlin I Glaciation

An early Wisconsinan glaciation is documented by well weathered till and relatively deep glacial soils on the higher summits within the Fourth of July Creek study areas. In addition, a peat horizon has been dated at ± 31,000 C-14 years B. P. in the valley of Boulder Creek on the south flank of Mount Leonard (Fig. 3) (Radiocarbon, 1976; also Institute of Marine Sciences, University of Alaska Radiocarbon Dating Lab, sample AU-59, 1975). In this connection Miller (1975a) has reported a C-14 date of ± 36,000 on wood fragments at a bedrock-till interface in McKee Creek 10 miles south of Atlin. These organic fragments are beneath the lowest weathered till in the Vesnover Mine and probably can be correlated with the Boulder Creek peat noted above (see Radiocarbon, 1976; and University of Alaska, 1975, sample AU-114).

Beneath the above-noted peat horizon is well-weathered glacio-fluvial and till material which is presumed to correlate with tills found as ground moraine on the higher elevation slopes and ridges in the immediate study area. Also erratics have been found on summits high above the substantiated upper glacial limits of subsequent advances. Added to this is the strong development of tors on high ridges in the Atlin region, such as on Sentinel Mountain at 6300 feet (1920 meters) elevation in the adjacent area south of the Atlin townsite (Fig. 3). All of this attests to the relative antiquity of this extensive early glaciation. From the Boulder Creek and McKee Creek C-14 dates and the general degree of weathering on the highest-level glacial deposits, the pre-Atlin I phase is considered to be <u>early Wisconsinan</u>, and probably greater than 40,000 years B.P. Specifically it is suggested as probably lower Altonian in the mid-continent chronology. This interpretation is consistent with the Alaska composite of Péwé (1975) and the Yukon Shakwak glaciation of Denton and Stuiver (1967) noted in Table V. It is also consistent with oxygen-isotope ratio trends over the past 100,000 years revealed by deep ice cores in the Greenland Ice-sheet (Langway, et al, 1973, and Dansgaard, et al, 1973).

Much emphasis has been put on the Boulder Creek and McKee Creek sites in the age assignment of this phase. Therefore, in the interest of scientific integrity it must be kept in mind that these dates are considered to be near the maximum range of C-14 dating and, indeed, that the pre-Atlin I phase could be even older than suggested.* On the other hand the amount of weathering on high-level drifts in the Fourth of July Creek area is much less than would be expected if the drift had been exposed for the entire Sangamonian. This may be explained, of course, by the presence of cool-dry conditions in this region even during Sangamonian time, conditions which may have impeded or slowed down weathering rates. Though the character of the Sangamonian in this region is not well understood, it is nevertheless assumed that there must have been long periods of warmer climatic conditions that would have produced more extensive weathering than is found in the cited drifts.

Comparison with the Taku District Chronology in the Alaskan Panhandle and Southern Atlin Area

Of particular significance to the study area is the Quaternary glacial record or the Juneau Icefield and its peripheral sectors which were the source area for the Atlin Ice. On the basis of extensive and moderately well-weathered drift and recognized cirque sequences, the pre-Atlin phase is correlated with Miller's (1956) early Atlin stage (Early Juneau/Atlin stage in Miller, 1975a). The Cl level cirque system, Table III (also see Miller, 1961), is suggested to have been initiated in late-Illinoian or early-Wisconsinan

* Further samples of the critical Boulder Creek peat are currently undergoing isotopic interpretation by Geochron (Kreuger Enterprises) in Cambridge, Mass., and will be reported upon in the final publication of this report.

time. Evidence for interglacial climates on the Juneau Icefield have been destroyed by later glaciations and the only real evidence for an interglacial following this stage is a change in elevation of active cirques in consequence of a rise in elevation of the freezing level (a warming condition). This interpretation is somewhat restricted in that cirque levels were occupied at many different times during the Pleistocene and their Wisconsinan initiation is difficult to correlate as to time, thus forcing the analysis to emphasize only the sequential development. For these reasons the lowest cirque level is correlated to the most extensive Atlin region glaciations and assigned an age on the basis of known C-14 datings. These refer not only to the Boulder Creek/McKee Creek information in the Atlin region, but also to pre-Woodfordian C-14 dates reported from the southwestern Yukon region by Denton and Stuiver (1967).

Time of Onset of Wisconsinan Glaciation

Because of the mountainous character of this region it is probable that the early Wisconsinan glaciation began here before the end of the classical mid-continent Sangamonian age (ca. 70-90,000 years B. P.) and that it existed for a longer period of time than in more southerly regions (perhaps even up to 30,000 years). Such an earlier onset of glaciation in this northern Cordilleran region is suggested in view of the earlier accumulation at higher elevations as freezing levels fall. It is also suggested because of the higher latitude position - i.e. in this sub-Arctic region.

Details for fluctuations within this phase would have been removed by subsequent glaciation, as noted diagrammatically in Table V, but by inference it is believed that major glacials and intraglacials undoubtedly existed during this early Wisconsinan time frame. This is also consistent with the suggested divisions of the Wisconsinan derived from Greenland ice cores (Dansgaard, et al, 1969, 1971).

Very little work that can be correlated with Gladys Ice has been done in the region east of the study area. For the north central British Columbia area Rutter, (1975) has documented a very early advance, which he suggests as possibly pre-Wisconsinan. This may actually be a correlate of the early Wisconsinan phase shown here for the Atlin-Taku region. It should be noted, thought, that in the area of Gladys drift evidence for a pre-Atlin I phase has been completely removed or covered by the effects of extensive Gladys I glaciation.

Regardless of details, it is certain that the Greater Mountain Ice-sheets which were involved in the pre-Atlin I glaciations were the most extensive ice sheets which have affected the Atlin region during Wisconsinan time.

Boulder Creek Intraglacial

The evidence used in this study that there was a substantial retreat of ice from the Atlin region sometime between 35 and 50 thousand years ago is based on the cited Boulder Creek/McKee Creek C-14 dates, abetted by the C-14 dates of Denton and Stuiver (1967) relating to the Silver non-glacial of the southwestern Yukon Territory. That this stage actually existed is not questioned, only its precise date and details of related intraglacial variations. The dating would be less tentative if more locations with organics had been found.

A tentative correlation with the Port Talbot Intraglacial (Flint, 1971; Terasmae and Dreimanis, 1975) of some 50 to 60,000 years B. P. in the Great Lakes region and Easterbrook's (1967) corresponding Salmon Springs Interglacial (Intraglacial would be a more appropriate designation) of the Puget Lowland is made on the basis of the above-cited radio-isotope ages. Miller (1975a) has also recognized the existence of this intraglacial on the Juneau Icefield (see Table V) and for that sector considers it to have been of short duration and relatively cold. He has suggested this on the basis that sub-Arctic mountain environments have been involved throughout the Pleistocene, making even the intraglacial conditions more severe than along the southern periphery of the Cordilleran and Laurentide Ice-sheets of the Pleistocene.

Atlin I - Gladys I Glaciation

This stage of glaciation followed the Boulder Creek Intraglacial and represents the first glaciation in the region for which there is in-valley morphological evidence. Atlin I Ice was thicker and assumed to be more thermophysically polar in character. This is inferred from the steeply dipping gradient of its terminal moraine remnants. The rationale is that ice flows more slowly at temperatures below the pressure-melting point, as demonstrated in the laboratory by Glen (1955) by Nye (1953) and by Miller (1956, 1975a) with field measurements on the Juneau Icefield. This results in steep terminal snouts. (In fact in Greenland and the Antarctic such termini today are found to be greatly oversteepened compared to those of temperate glaciers). This suggests that cold high-level ice came into the region from the Juneau Icefield sector to the south as well as from local high-elevation accumulation. Extensive ice from the Cassiar source to the southwest and as well much accumulation in local highlands in that sector indicates that in this part of the Pleistocene the Arctic Front had moved well east of its mean present position. This would have more readily permitted maritime air to make its way into the interior highlands. During this major glaciation ice is thought to have completely filled the valleys between the Coast Range and the Cassiar Mountains, resulting in a very extensive accumulation on the Teslin and Taku Plateaux (Bostock, 1948, and also Fig. 1 of this present report).

The Atlin I - Gladys I Stage is correlated with the upper Woodfordian of the mid-continent and Salmon Springs Stage II in the Puget Lowland. The Boutellier nonglacial in southwestern Yukon Territory is C-14 dated by Denton and Stuiver (1967) at between 30 and 40 thousand years B. P. As such it agrees well with the Boulder Creek/McKee Creek C-14 dates and hence appears to have been contemporaneous with the early part of Atlin I - Gladys I glaciation. Only with more complete and absolute dates from the Atlin region, can the apparent out of phase relationship found in the study area be fully understood. Also because of some uncertainty in the radiocarbon dates it is even possible that the beginning of Atlin I was earlier than represented on the correlation chart (Table V).

The more highly weathered moraines for Atlin I ice than those of Gladys I age suggest that Atlin Ice built up as the Arctic Front moved inland and was influenced by increased maritimity, and retreated while Gladys I ice continued to build up primarily from out-flow from the less maritime sector of the broad region in the Cassiar Mountains. It is suggested that the Gladys I build-up correlates morphogenetically with the McConnell Glaciation in the southeastern Yukon Territory (Hughes et al., 1969). Both of these glaciations had an easterly source area and the Gladys I ice lobe certainly extended well north into the Yukon Territory beyond the present Yukon-British Columbia border (Figs. 1 and 2).

There is no clear evidence that Gladys I ice completely withdrew from the Gladys Lake depression between Gladys I and Gladys II time. Another possibility is to correlate the Gladys I stage with Atlin II (v. Table V) which subsequent research in the area south and east of Gladys Lake may support.

Evidence from the Juneau Icefield suggests that the Atlin I-Gladys I stage is a part of the lower Gastineau/ Sloko glaciation (Miller, 1956) which was also morphogenetically a Greater Mountain Ice-sheet. Evidence for this stage is present only in the sequence of erosional landforms of cirques and berms, particularly the initiation of low-elevation cirques (C-2, Miller, 1956, 1961, 1975a) on the Alaskan coast at a mean elevation of 1100 feet (300 meters).

Also in the Juneau Icefield sector, no depositional evidence exists for the Farmdalian Interglacial as the extensive late Wisconsinan glaciation was so intense that it removed all depositional evidence of any preceding deglaciation phase in the presently glacierized areas. From the character of Atlin I moraines it is further suggested that the originating ice in this glacial phase included a build up of thermophysically Polar ice which late in the stage, in consequence of climatic amelioration, probably changed its character to a more sub-Polar ice mass. Subsequent to this, it may be presumed to have been followed by build up of a more Temperate ice character in the Boundary Range, leaving no remnants whatsoever of the former colder phase. Such warmer conditions naturally resulted in retreat of ice from the Fourth of July Creek Valley, eventually to be followed by re-invasion of ice from the main Atlin valley during the Atlin II stage.

A Suggested Pine Creek Intraglacial

As discussed above, the more extensive weathering of Atlin I drift indicates a relatively important intraglacial in the area of Atlin drift which would make it the equivalent of the Farmdalian intraglacial in the mid-continent chronology, and of the Olympia Interglacial of the Puget Lowlands. For the Atlin and Taku areas this has been termed the Pine Creek Interglacial, in recognition of a substantial weathering profile on drift sheets exposed in the valleys of Pine Creek, Spruce Creek, Boulder Creek and McKee Creek in the Atlin valley. No time correlate of the Pine Creek Intraglacial has been designated in either the studies of Denton and Stuiver (1967) or Hughes et al., (1969). From this lack of evidence for an intraglacial found to date in adjoining regions of the Yukon, as well as on the Juneau Icefield* and in the Gladys Lake sectors, it is assumed that this phase, if indeed the correlation proves to be valid, was not as long nor as warm as the Farmdalian Intraglacial in the mid-continent region.

Atlin II-Gladys II Glaciation

The advanced position of Atlin II ice and the indicated lower level of Gladys II ice suggests that the major accumulation area in this phase was the Boundary Range especially its

* Except to cite the sub-stages (moraine entities) in the Sloko Stage noted for the Taku District in Table V.

higher elevation névés as zones of maximum snow accumulation. Although still an important accumulation area, the Cassiar Mountain region and its bordering Yukon and Stikine Plateaus, did not foster as extensive a glaciation. This is based on the recognition that this stage was characterized by a major still-stand at an elevation lower than that reached by Gladys I ice. Continued field work east of the study area could produce evidence for a more extensive retreat and re-advance of Gladys ice at this time. The corresponding maximum advance of Atlin II ice within the Fourth of July Creek Valley is also not clear, because of the blocking effect of Gladys II ice plus the locally derived ice from adjacent cirque basins. Thus all of this interpretation has to stem from the observed moraine sequence.

In the foregoing context, there are also moraines on the southeast flank of Mount Ewing which appear to be related to Atlin Ice. Since the rest of the moraines in the upper Fourth of July Creek Valley area are related to the Gladys and/or local plateau ice, it is suggested that some of this was ice nourished from the large 4000 foot (1300 meter) cirque between Caribou Ridge and Mount Ewing. This cirque basin is presently filled with a tarn lake, as shown in Figure 3.

The Gladys II stage is correlated with the upper Glatineau/Sloko stage of the Greater Mountain Ice-sheet in the Taku District and hence to the lower to middle Woodfordian of the mid-continent stratigraphy. In this stage, the Puget Sound area was not glaciated as early or as extensively and both the Evans Stade and the Vashon Stade, separated by an interstadial, (Table V) are correlated within the Atlin II - Gladys II time period. The Puget Lowland, of course, was much farther south and as well the glacier termini there were a much greater distance from their source areas. Are we to conclude that the Boundary Range and the Cassiar highlands accumulation areas were not as sensitive to climatic perturbations which produced the distinct stages within the Woodfordian in the mid-continent area and in the Puget Lowlands? This question is yet to be revealed.

Work done in the southern Yukon Territory delineates no substage from maximum Atlin I to the end of the Wisconsinan. The Naptowne Glaciation of Cook Inlet (Karlstrom, 1964) may however, have a correlative stage in the Skilak Lake stage noted in Table V.

Again, it may be pointed out that variation of climatic conditions is not so clearly manifested in the drift records described in this dissertation because evidence for earlier fluctuations were so often erradicated by the next maximum advance and also modified by the effects of later alpine ice. It is also conceivable that climatic variation in the accumulation zones only caused minor changes in accumulation levels and not in the total amount of ice channeled into the Fourth of July Creek Valley.

A Late - Wisconsinan Climatic Amelioration

During and following the demise of the Atlin II glaciation, there was an extensive development of glacio-fluvial terraces produced by melt waters flowing in massive quantities from the downwasting Gladys II ice-sheet in the upper Fourth of July Creek Valley. This is illustrated by the highest fluvial terrace shown in Figure 11 for the intermediate valley section of the Fourth of July Creek Valley. The position of this related "intraglacial" is also suggested schematically in Table V. Following this time and the related sequence of glacio-fluvial events there was a return to cooler and wetter conditions, which in turn led to the major resurgence of ice in Atlin II and Gladys II time.

Atlin III, IV and V - Gladys III and IV Glaciations

Because of the youthful topography and minimal weathering found on the glacio-fluvial terraces the Atlin III, IV and V stages and the Gladys III and IV stages are considered to be late-Wisconsinan in age. The evolution of this sequence is indicated as follows.

A rather extensive retreat of Atlin II ice preceded the Atlin III phase during which time the massive kame moraine (Ruffner moraine - outwash complex) was built at the juncture of the Porter Lake Valley and the Fourth of July Creek Valley (Fig. 17). Following this, Gladys III ice downwasted and retreated from the valley to the northeast, the melt-waters scouring several large runoff channels as the ice retreated. Three well-defined glacier levels lie about 500 feet (152 meters) apart but the steep gradient of their projected surfaces, as well as their quite similar morphology and weathering, indicate that the represented substages of Gladys III were very close in time. They have thus been grouped as one stage.

The Ruffner moraine - outwash complex at the valley junction is correlated with the zone of massive embankment moraines near Jakes Corner on the Alaskan Highway. These lie some 60 miles (100 kilometers) north of the village of Atlin. Extrapolating from Carbon - 14 dates in bog sediments, Anderson (1970) has suggested that the Jakes Corner moraines are of maximum late Wisconsinan age and hence were built by 10,200 years B. P. Thus a similar age is assigned to the Ruffner Moraine Complex. Dates from the upper Fourth of July Creek plateau indicate that alder was growing 9800 C-14 years B. P. It is to be noted that both of these dates are minimum. With respect to the mid-continent stratigraphy and that of the Colorado Front Range, this stage (termed Atlin III) is correlated with the late Woodfordian or the Port Huron stage in Michigan and the Middle Pinedale Stage in the Rockies. The McConnell Glaciation and Kluane Glaciation of the eastern and south western Yukon Territory respectively were present the entire period from Atlin/Gladys II to final deglaciation at the end of Atlin V time. Data from Cook Inlet (Karlstrom, 1964) suggest that the Skilak Lake glaciation ended about the same time as the final stage of the Wisconsinan and was followed by an early Holocene glaciation, the Tanya (Table V).

In the Boundary Range, the Intermediate Mountain Icesheet glaciation, called the Douglas/Inklin stage by Miller (Table V) is thought to be equivalent to the Atlin III - Gladys III glacial resurgence. Evidences for this stage in the Juneau Icefield region are the modification of cirque levels 2 and 3 at the 1100 and 1800 foot (300 and 545 meter) elevations, the presence of a pronounced erosional berm (level 3) in the highland and the existence of notable moraine embankments and terraces in the Inklin junction region of the upper Taku River Valley (Miller, 1963).

The final downwasting and retreat of Atlin III ice is documented by a recessional moraine complex near Lower McDonald Lake. This is considered to correlate with a huge kame terrace complex at the lower end (north) of Atlin Lake, (Fig. 2) 30 miles (48 kilometers) north of the Fourth of July Creek mouth. By this time, the Atlin Ice was confined to the main Atlin valley, with a flattish gradient of only 16 feet per mile (3 meters per kilometer). From the study of bog sedimentation and palynology, Anderson (1970) has dated the Atlin Lake kame terrace at Mile 24 on the Atlin Road, some 30 miles (48 kilometers) north of the Fourth of July Creek Valley, at 9000 to 10,000 years B. P., which would correlate to the Sumas Stage in the Puget Sound region of Washington State and the waning Valders Stage of the mid-continent stratigraphy in the Great Lakes region of the mid-continent Laurentide Glaciation. The final Wisconsinan glaciation in the coastal and inland sector of the Taku District (Juneau Icefield) is designated the Salmon Creek/Zohini stage by Miller, again based on deltaic terraces in the Salmon Creek valley near Juneau and on moraine and kame terrace sequences near Zohini Creek in the upper Taki River tributary sector of the southeastern Atlin region (Tables V and VI). Thus Atlin IV - Gladys IV would be a correlate of the lower Salmon Creek/Zohini stage, all of this being associated with a lesser Mountain Ice-sheet in the highlands of the Coast Mountains. The primary evidence for this stage on the Juneau Icefield itself is the occupation and erosion of Cirque level 4 at 2500 feet (757 meters) and the erosion of berm level 4 in the wide-strathed glacial-filled valleys, such as occupied by the present-day Taku and Llewellyn Glaciers. Evidences at lower elevations along the Alaska coast and in the Taku Valley and the southeastern Atlin region lie in the presence of gravels stratigraphically above marine till and diamictons in the Coastal area, and corresponding pebble terraces at the mouths of interior valleys in the upper Taku River valley. Upper Salmon Creek/ Zohini stage deposits are represented by undifferentiated outwash gravels and kame terraces at Tulsequah and Zohini Creek where the Taku River crosses the Alaska-British Columbia border (Fig. 1 and Miller, 1956).

More precise teleconnectional study of the latest

Wisconsinan features in the adjoining Taku River region could be profitable as the sequences do appear similar to that described in the Atlin IV and V stages. Some of the Taku sequences may, however, predate the early Holocene (Miller, personal communication).

Once confined to their immediate valleys, near the study area of this report, Gladys and Atlin Ice downwasted and retreated rapidly. This has been shown by Anderson, (1970) and also by the general lack of recessional moraines in the immediate valleys. Atlin Ice retreated to near the present glacial position of the Llewellyn and Willison Glaciers, just south of the head of Atlin Lake, (Figs. 1 and 3). Following this, all Gladys Ice, which was by then completely devoid of a source area, disappeared by down-wasting and stagnation. This dessication left large kettle ponds and pit lakes, as well as massive deposits of melt-water sedimentation (largely sands and gravels) in the area between Gladys and Teslin Lakes (Fig. 3).

CHAPTER IX

THE HOLOCENE

The Last 10 to 11,000 Years

Holocene climatic trends as discussed by Anderson (1970), Miller and Anderson (1974) and Miller (1975a) show a rather continuous warming trend from about 9500 years B. P. to about 6000 B. P. This culminated in the Thermal Maximum (Table II) when mean annual temperatures were some 2°F (3°C) higher than present. During this interval, the most intensive of periglacial processes migrated to higher elevations and the greatly retracted Juneau Icefield received its maximum accumulation only on the highest neve and at cirque levels 5 and 6 (i.e., at 3200 and 3900 feet; 975 and 1190 meters as shown in Table III). The latter is some 700 feet (210 meters) higher than the lowest cirque filled with ice today. Evidence for the Thermal Maximum in the Atlin region is also given by a recent C-14 date of 3472 ± 87 years B. P. on a silty peat horizon in the foreland area about a mile below the Cathedral Glacier terminus (Miller 1975a; Jones, 1975; also see Radiocarbon, 1976, Vol. 18, No. 1, sample AU-77A). This site lies between two recessional moraines of early Holocene age.

Near the end of the Thermal Maximum, the Arctic front maintained an average position well west of the inner channel-ways of the Alexander Archipelago and the southeastern Alaska coast. Then too, wetter conditions and increased storminess prevailed on the continental flank of the Boundary Range and well over into the Atlin region.

The Thermal Maximum (Hypsithermal Interval or Climatic Optimum of some writers) was followed by a major cooling trend as the Arctic Front once again moved inland. This was coincident with a decreased storminess and drier conditions accompanied by general cooling in the Atlin district. Concurrently, on the coast, relatively cooler and wetter conditions dominated the Juneau Icefield and the Taku District. The nature of these out-of-phase climatic trends has been documented and explained by Miller and Anderson in their 1974 report on glacial and palynological changes in this region during the approximately 10,000 years of the Holocene.*

From these studies it is also apparent that the early Neoglacial time interval from about 2500 to 1200 years B. P. (600 B. C. to 700 A. D.) was as much as 4°F (2°C) cooler than today. This was a period of extremely intense periglacial activity in the upper Fourth of July Creek Valley as described from the field evidence discussed in Chapter VII. Then a short warm interval persisted from about the end of

* Defining the Holocene as the interval since the last major glacial or large-magnitude climatic event. On this definition the Holocene may represent quite different time spans in different regions of the globe.

the 7th century (1200 years B. P.) to the 13th century (750 B. P.) (Miller and Egan, 1968). This is now well-documented for the Alaskan coast by significant C-14 dates on overridden trees at the terminus of Mendenhall Glacier (1,197 ± 153 years B. P.) and of the Taku Glacier (970 ± 100 years B. P.) at sites studied by Miller (1975a); also see Radiocarbon 1976, sample AU-100).

Since about the 14th century, the Atlin region has experienced a relatively and wetter climate compared to the short warm interval of the Middle Ages noted above. With these cooler conditions, in the Atlin region nivation hollows have been periodically filled with firn as the lesser climatic oscillations continued.

Quaternary Peat Deposits in Boulder Creek Valley

Carbon 14 dates obtained in the Boulder Creek Valley further substantiate the late Holocene climatic changes. In Figure 37 is illustrated the position of two late Holocene organic horizons stratigraphically above an organic horizon representing the beginning of the Holocene and which lies on the latest Wisconsinan drift surface comprised of unweathered ground moraine and basal till. The organics are well-formed intercallated with lacustrine and glacio-fluvial layers.

The lower of these late-Holocene peat horizons is dated 2770 C-14 years B. P. (Geochron, 1973) which correlates fairly well with the Cathedral Glacier peat bog date of 3472 ± 87 C-14 years B. P. It represents the end of the Thermal Maximum and beginning of Anderson's (1970) cool dry

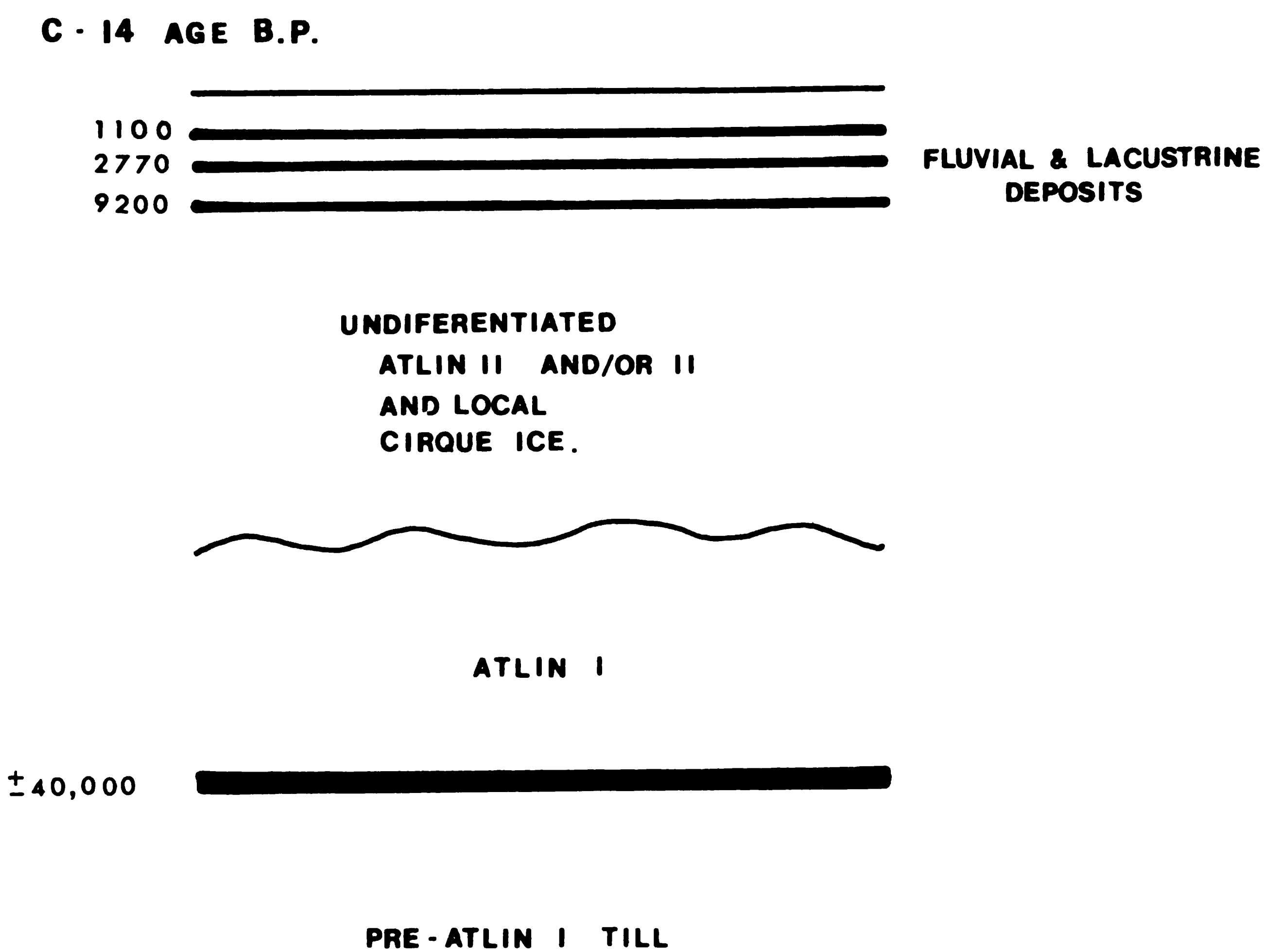

Figure 37. Organic horizons in lower Boulder Creek valley

Neoglacial period. As such it is related in time to Neoglacial advances of the Mendenhall Glacier near Juneau dated by several selected samples which pre-date the Neoglacial advance at 1800 to 2200 C-14 years B. P. (Heusser, 1952; Miller, 1956; Cross, 1968; Miller, 1975a). The upper horizon at 1100 C-14 years B. P. (Geochron, 1973) correlates well with the later Mendenhall buried forest (1197 ± 153 years B. P.) near Juneau, cited in the previous section. It also may possibly correlate with an overridden forest of mid-Neoglacial age near the Davidson Glacier on the western side of Lynn Canal some 10 miles southeast of Skagway, Alaska (Egan, 1971; and Fig. 1).

Because of these teleconnections it is most probable that the two upper peat horizons in Boulder Creek Valley are associated with regional climatic change, though one could also suspect some relationship to changes in drainage of Boulder Creek and associated bogs along the shores of Surprise Lake (Fig. 1). Today the vegetation above the upper peat horizons is well drained and dry. Therefore it is certain that at least locally wetter conditions existed during the formation of the peat and that this was followed by fluvial deposition.

Increased nivation and perhaps occupation of high-level cirques by renewed glaciation could have produced the run-off necessary for fluvial deposition after each peat-producing period. Further stratigraphic work in this and other valleys in the area should lead to a refined chronology as it applies to the Holocene.

Based on the evidence cited in Chapters VI and VIII, the following tabulation (Table VI) of the Glacio-Chronology reviews the Wisconsinan events in the Fourth of July Creek Valley. By implication it also refers to the broader Atlin Region and the adjacent Taku District to the south and west. In combination this summary represents the late-Pleistocene chronology of the northern end of the Wisconsinan Cordilleran Ice Sheet.

TABLE VI

LATE-PLEISTOCENE GLACIO-CHRONOLOGIC SUMMARY FOR THE ATLIN REGION AND TAKU DISTRICT

Approx. Yrs. B.P. x 1000	Atlin Region (this study)	Taku District (Miller, 1956)
0		
2	Nivation and palsa development	Neoglacial
4		
6	Thermal Maximum	Thermal Maximum
8		
10	Intense peri-glacial	High cirque glaciation
12	Atlin IV & V Gladys IV glaciation	Salmon Creek/ Zohini stage, with 3 substages
14	Atlin III-Gladys III glaciation	Douglas/Inklin stage
16	Climatic Amelioration	Intraglacial
18	Atlin II-Gladys II glaciation	Gastineau/Sloko stage
20		
30	Pine Creek Intraglacial Atlin I-Gladys I glaciation	
40	Boulder Creek Intraglacial	Intraglacial
50		
100	Pre-Atlin I glaciation	Early Juneau/ Atlin Stage

CHAPTER X

SOME SUMMARIZING CONCLUSIONS

The correlations attempted in the study area were done with the axiom that if large-magnitude continent-wide and even world-wide climatic perturbations were responsible for major glacial advances and retreats in the mid-continent region of the United States and Canada then major atmospheric perturbations should also have been recorded in the sequence of glacial landforms in the Pacific Cordilleran region.

Problems in Teleconnection

The reader was reminded early in this report that the main climatic and/or glacial divisions might be expected to be out-of-phase in the two localities considered, i.e., the Atlin Valley and the Gladys Valley provenance zones. This should be particularly apparent in the early phases of the glacial sequence when it is expected that major accumulation would begin somewhat earlier in the highlands of the Boundary Range. Thus the chrono-relationship (Table V) suggests that the earliest Wisconsinan glaciation actually began in late Sangamonian time. That is to say major accumulations were initiated in the regions of highest elevation and especially near the coast where ready sources of moisture prevailed.

Both of these prerequisites are fulfilled in the Alaska-Canada Cordillera. Although there are no time-stratigraphic boundaries known, the Boulder Creek (Port Talbot or "equivalent") intraglacial is inferred to have been shorter in the Cordilleran region than revealed in the mid-continent chronology. The reason for this conclusion is again stated to be an earlier build-up of ice during re-glaciation and a retarded recession at the end of the early Wisconsinan (pre-classical) glacial stage...i.e., just prior to Gladys I glaciation.

Even in terms of sequence, there is limited evidence for a correlate of the Farmdalian intraglacial in the Cordilleran region, except for some notable weathering of low-elevation tills in the Atlin Valley. This is found on a basal till widely exposed on the sluiced down-banks of McKee Creek, Pine and Spruce Creeks and in the Boulder Creek and Otter Creek areas where intensive gold mining during and since the Gold Rush days has exhumed this key horizon. In each case, this till lies stratigraphically below a younger and unweathered surface till. Because of particularly note-worthy limonitic weathering of the lower unit in the Pine Creek area (near the old mining town of Discovery) the name Pine Creek Intraglacial is suggested in Tables V and VI, to represent a retreatal phase between Atlin I and Atlin II glaciation.

During this time of ameliorating climate in the mid-continent region melting of the Laurentide glacial sheet was greatly enhanced by increased precipitation in the summer months as maritime sub-tropical air masses moved up from the Gulf of Mexico as they do today. The frontal position of these storms was presumably very much controlled by position of the ice margin and by the cold high pressure air masses which were associated with the ice sheet beyond this margin. With these increased temperatures and precipitation, the ice receded and Farmdalian soils developed (Ruhe, 1975).

In the region of the Cordilleran ice sheet, however, there were strong geographic and orographic influences on the progress of deglaciation. For example, in spite of global atmospheric warming, the result in the highlands of the northern Cordellera was increased high-level accumulation. This continued to nourish glaciers in the Alaskan Panhandle and in the interior regions of British Columbia and the Yukon. Because of the high elevation and relief in this region, the global climatic warming was insufficient to cause much reduction in the upland ice masses in this Cordilleran region during the Farmdalian time-span. Only the low-elevation valley glaciers suffered wasting, such as those filling the valleys of McKee Creek, Pine Creek, Spruce Creek, Otter Creek and Boulder Creek, as previously discussed. The relative locations of these key sites are indicated on the map of Figure 1 and shown in topographic detail in Figure 3. In fact, it is implicit here that there was probably not a true intraglacial manifested in the interlobate sector of the highland study area where there was such an interplay between the Atlin and Gladys ice and where even in the waning and waxing phases local cirque glaciers added substantially to the ice-cover.

In the final stages of Wisconsinan glaciation (Atlin III, IV and V time) quite different conditions appear to have existed. Certainly there is strong evidence that the ice retreat was rapid. From the time-stratigraphic record given by C-14 dates, it appears that these stages were concurrent with those of the mid-continent deglaciation. Miller (1975a) has also suggested this for the adjacent Taku District and the Southeastern Atlin region, noting that "the main morpho-erosional and morpho-stratigraphic sequences, abetted by available time-stratigraphic evidence, are subdivided into an Early Juneau/Atlin (early Wisconsinan) stage and a Gastineau/Sloko (mid-Wisconsinan) stage, followed by two late Wisconsinan stages...the Douglas/Inklin and Salmon Creek/Zohini stages. On the basis of sequence, the latter two should be equivalents of the Port Huron and Valders glaciations of the mid-continent chronology." This observation distinguishes the glacio-climatic character at the end of the Wisconsinan from the apparently quite different nature and teleconnectional relationship of earlier stages within the Wisconsinan.

As has been previously shown, in the northern Cordilleran region the nature of the pre-Atlin I phase of glaciation (early-Wisconsinan) is still obscure and hardly understood and, indeed, in this region it may never be. But there is abundant evidence for the Atlin I and Atlin II glaciations, and for the final wasting of stages in Atlin III, IV and V time.

As for the Pine Creek Intraglacial (ca. 13,000 to 14,000 B. P.), it is reiterated that this is only characterized in the lower valleys of the Atlin region. The high plateau sector of the upper Fourth of July Creek valley conversely experienced a build-up, or at the least was affected by a long period of still-stand of ice. Hence on evidence from the study area, a classical type of intraglacial cannot be distinguished. In fact, just the opposite apparently took place, with effective increases in local glaciation in the high cirque-headed valleys. This situation and its basic significance in terms of the causal factors involved when there are simultanious advances and retreats of glacier lobes have been discussed at length in the early Juneau Icefield Research Program reports. Some of these illustrate particularly well the nature of so-called anomalous advances on the Taku Glacier in contra-distinction to the retreatal activities of the termini of nearby Norris Glacier, the Twin Glaciers, and the Herbert and Mendenhall Glaciers on the Juneau Icefield (Miller, 1956, 1963, 1973a, 1973b).

Conceptual Model for Out-of-Phase Pleistocene Glacial Variations of Large Magnitude

In view of the classic mid-continent Wisconsinan chronology to which reference has been made in the correlation chart of Table V, it is of interest that large-magnitude out-of-phase variations with time-stratigraphic significance have occurred in the regime of the Des Moines, Michigan, Huron and Erie glacial lobes during the Pleistocene (Flint, 1971; Embleton and King, 1974). This has long perplexed glacial geologists, especially those without experience in the behavior of existing glaciers.

A significant contribution to the understanding of the out-of-phase phenomenon has been made by the research teams of the Juneau Icefield Research Program which, in the mid 1940's, began to look at the total system activity of glaciers with multiple neves and separate nourishment provenances. A detailed analysis of this problem, exemplifying the behavior of the Norris (receding) and Taku (advancing) glaciers, was first presented by Miller (1956, 1963). In these, the effects of shifting zones of maximum snowfall and changing freezing levels through time have been delineated. Since its beginning the study has been based on solid empirical evidence through a number of years of systematic sampling of net accumulation changes in different sectors and at different elevations on the Juneau Icefield. Because the icefield is geographically adjacent to the study area dealt with in this report these findings are of special interest.

Subsequently, Anderson (1970), Miller and Anderson (1975, 1974a and 1974b), Jones (1975) and Miller (1973, 1975a) and Miller (1975c) have elaborated on the conceptual model based on recent and current regime trends on the Juneau

Icefield. This has been done by extending these ideas backward through the Holocene to explain the seemingly enigmatic out-of-phase climatic characteristics in the Atlin area compared to the Taku District on the Alaskan coast. Details of these out-of-phase climatic differences have been noted in Table II.

The results of the writer's current studies in the Fourth of July Creek Valley tend to substantiate the validity of this conceptual model in terms of its application to those realms of the Wisconsinan time-scale which predate the Holocene. The conceptual model has also helped to explain the nature of the out-of-phase oscillations which the glacio-morphostratigraphy and the glacio-fluvial record gives in the Porter Lake Valley and in the intermediate segment of the Fourth of July Creek Valley. This includes that most significant interlobate zone above and below the Ruffner Moraine Complex. Thus the fundamental interpretations of this study have leaned on key results of the cited previous investigations of Holocene stratigraphy in the Atlin valley and on the Neoglacial and Little Ice Age research on the Juneau Icefield proper. But also the current study has built upon these foundation concepts, and it is hoped that they had added some verification to them.

It is also sincerely hoped that this report will serve as a ledge upon which other researchers may stand as further detailed investigations unfold.

CHAPTER XI

FURTHER RESEARCH PROSPECTS AND ADDITIONAL STUDIES

Although this study was initially focused on the relatively small arena of the Fourth of July Creek Valley, it became apparent early in the investigation that it was essential to look beyond the immediate field area in order to come to grips with some of the basic problems involved. This is why, a few observations from surrounding areas and from the larger adjoining region have been included. Because the considerations involved this larger area which much of which is characterized by inaccessibility, high relief, limited exposures and, in some cases even ground cover of dense undergrowth, it has been difficult to encompass. The constraints are not nearly as severe as those occasioned by the extremely rugged terrain and dense mantling of rain forest which obscures so much of the glacial geology in the Alaskan coastal sectors of the Taku District. Still, because of this situation and in terms of the chronological interpretation, the results of this study must be looked upon as fundamentally reconnaissance in nature. But some satisfaction is derived from the fact that an important step has been made, for in any new region reconnaissance geology is essential before detailed work can follow. This difficulty has been clearly recognized in other studies referenced from the Alaska-British Columbia-Yukon region - e.g. those of Johnson, Denny, Hughes, Karlstrom, Denton and Stuiver, Rutter, Miller et al.

Although some details of the chronology may not stand severe scrutiny of future investigations, the main tenets of the chronology should remain valid. Some confidence stems from the gross similarity of the conclusions with the main lines of interpretation found in adjoining regions - albeit some of the interpretations from other regions are quite incomplete. The hope is that interest will now be generated to encourage further work in other critical valleys of the Cassiar and Coast Mountains and in the Atlin region. Such investigations should include a detailed analysis of soils, and of soils-vegetation relationships along the line of the reconnaissance work of Anderson (1970) and Lietzke (1969) in this area. Also further time-stratigraphic information must be diligently searched for because, to date, too few buried peat horizons have been located. When such additional research is accomplished, hopefully more complete teleconnectional correlations can be made from valley to valley.

In the broader context it is the writer's belief that this reconnaissance study has established a framework for the Wisconsinan stratigraphy of the region from which future workers can begin to attack the more specific details. The most urgent needs for amplifying these results is additional control on dating of the glacio-climatic events, and further soil-stratigraphic evidence. Therefore, even though we are a long way from where we were in this investigation several years ago, there is still much to be done and many questions to be answered. As more absolute time relationships become available, however, some of the key questions should be answered and the recognized gaps in the chronology closed.

APPENDIX

GLOSSARY

Diffluent ice: The condition of glaciation when ice is confined to major valleys.

Freezing level: The elevation at which permanent snow accumulates. This level rises with increased temperatures and fall with temperature lowering.

Holocene: The time interval since the last major glacial or large-magnitude climatic event. The Holocene may represent quite different time spans in different regions of the globe.

Kame moraine: A moraine which is mantled by ice contact glacio-fluvial drift.

Storm paths: Peripheral winds in the interaction zone between the low-pressure maritime cells and high-pressure continental cells along the north Pacific Coast. This is distinct from the "cyclonic tracts" which refers to the movement of the center of low-pressure systems which is normal to the storm paths.

Transfluent ice: The condition of glaciation when all available cols are used and ice spilling from one valley to another, ice.

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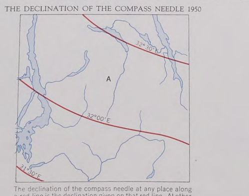
Military users, refer to this map as: Référence de la carte pour usage militaire:	SERIES A 502 SÉRIE MAP 104 N CARTE EDITION 1 MCE ÉDITION

THE DECLINATION OF THE COMPASS NEEDLE 1950



The declination of the compass needle at any place along a red line is the declination given on that red line. At other places the declination is between those given on the neighbouring red lines; thus at the place marked A the declination is between 32°00′E. and 32°30′E. The declination of the compass needle is decreasing 4′ annually.

Produced by the Surveys and Mapping Branch, Department of Mines and Technical Surveys, 1953, from air photos taken in 1951.
Printed by the Army Survey Establishment, R.C.E.
Department of National Defence, 1954

REFERENCE

Road, Hard Surface, All Weather — More than 2 Lanes; 2 Lanes Route No (2); Less than 2 Lanes
Road, Loose Surface, All Weather — 2 Lanes or More; Less than 2 Lanes; Dry Weather
Wagon Road, Cart Track etc
Trail or Portage
Boundary, International
Boundary Mon
Boundary, Provincial
Survey Mon
Boundary, County or District
Bench Mark — BM 2239
Boundary, Indian Reserve, Park etc
Triangulation Sta
Lot and Lot Number — 2091
Spot Elevation (in feet) — 2915
Power Transmission Line
Telephone, Trunk Route
Railway, Standard Gauge — Multiple Track; Station; Abandoned; Single Track

ATLIN

BRITISH COLUMBIA

Scale 1:250,000

1 Inch to 4 Miles Approximately

Miles 5 0 5 10 15 20 Miles

Kilometres 5 0 5 10 15 20 25 30 Kilometres

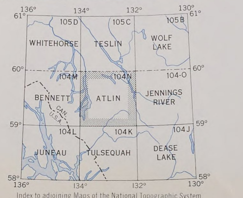
Copies may be obtained from the Canada Map Office, Department of Energy, Mines and Resources, Ottawa, or your nearest map dealer.

Ces cartes sont en vente au Bureau des Cartes du Canada, ministère de l'Énergie, des Mines et des Ressources, Ottawa, ou chez le vendeur le plus près.

Transverse Mercator Projection.
Contour interval 500 Feet
All Elevations in Feet above Mean Sea Level.

REFERENCE

Building
Fire Lookout Tower
Contours, Elevation — 2500
School
Wireless Station
Contours, Approximate — 2000
Post Office — P
Mine
Contours, Depression — 2500
Church
Cliff
Glacial Wash
Stream, Indefinite or Unsurveyed
Wooded Areas
Stream, Intermittent
Navigable Canal — Navigable
Stream, in Dry River Bed
Rapids and Falls — R F
Braided Stream
Ferry — Fy
Marsh or Swamp
Dam
Marsh or Swamp, in water
Lighthouse
Glacier or Snowfield
Landing Ground
Sand or Mud — Sand Mud
Seaplane Anchorage



Index to adjoining Maps of the National Topographic System

GRID ZONE DESIGNATION: 8V 100,000 M. SQUARE IDENTIFICATION NB PB / NA PA (660, 60)	TO GIVE A STANDARD REFERENCE ON THIS SHEET TO NEAREST 1,000 METERS			
	SAMPLE POINT: SCHOOL			
	1. Read letters identifying 100,000 meter square in which the point lies:	NB		
	2. Locate first VERTICAL grid line to LEFT of point and read LARGE figure labeling the line either in the top or bottom margin, or on the line itself: Estimate tenths from grid line to point:		7 3	
	3. Locate first HORIZONTAL grid line BELOW point and read LARGE figure labeling the line either in the left or right margin, or on the line itself: Estimate tenths from grid line to point:			0 4
IGNORE the SMALLER figures of any grid number; these are for finding the full coordinates. Use ONLY the LARGER figures of the grid number; example: 570000	SAMPLE REFERENCE:	NB 73 04		
	If reporting beyond 18° in any direction, prefix Grid Zone Designation, as:	8VNB7304		

TEN THOUSAND METRE
UNIVERSAL TRANSVERSE MERCATOR GRID
ZONE 8

ATLIN B.C.
SHEET 104N
FIRST EDITION

TABLE 5

TENTATIVE WISCONSINAN CHRONOLOGY

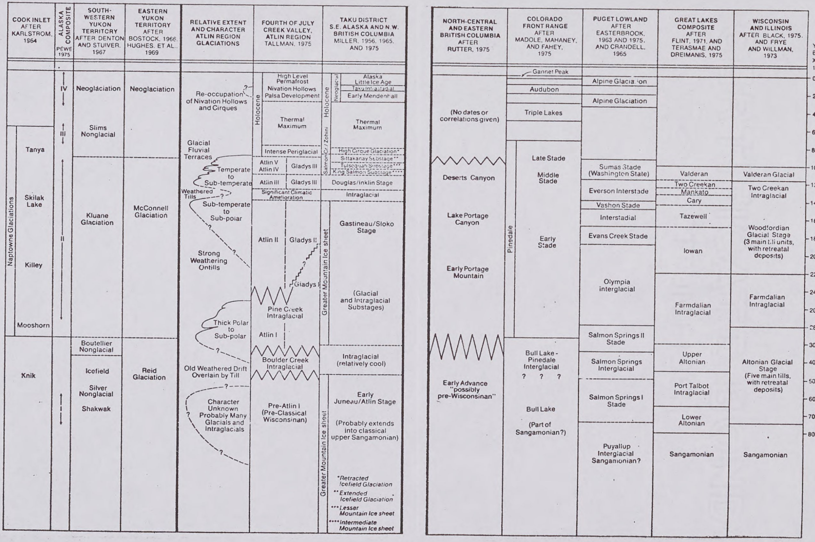


Table 5

Pocket has: Fig. 3 & Table 5

127
634
THS
T

SUP