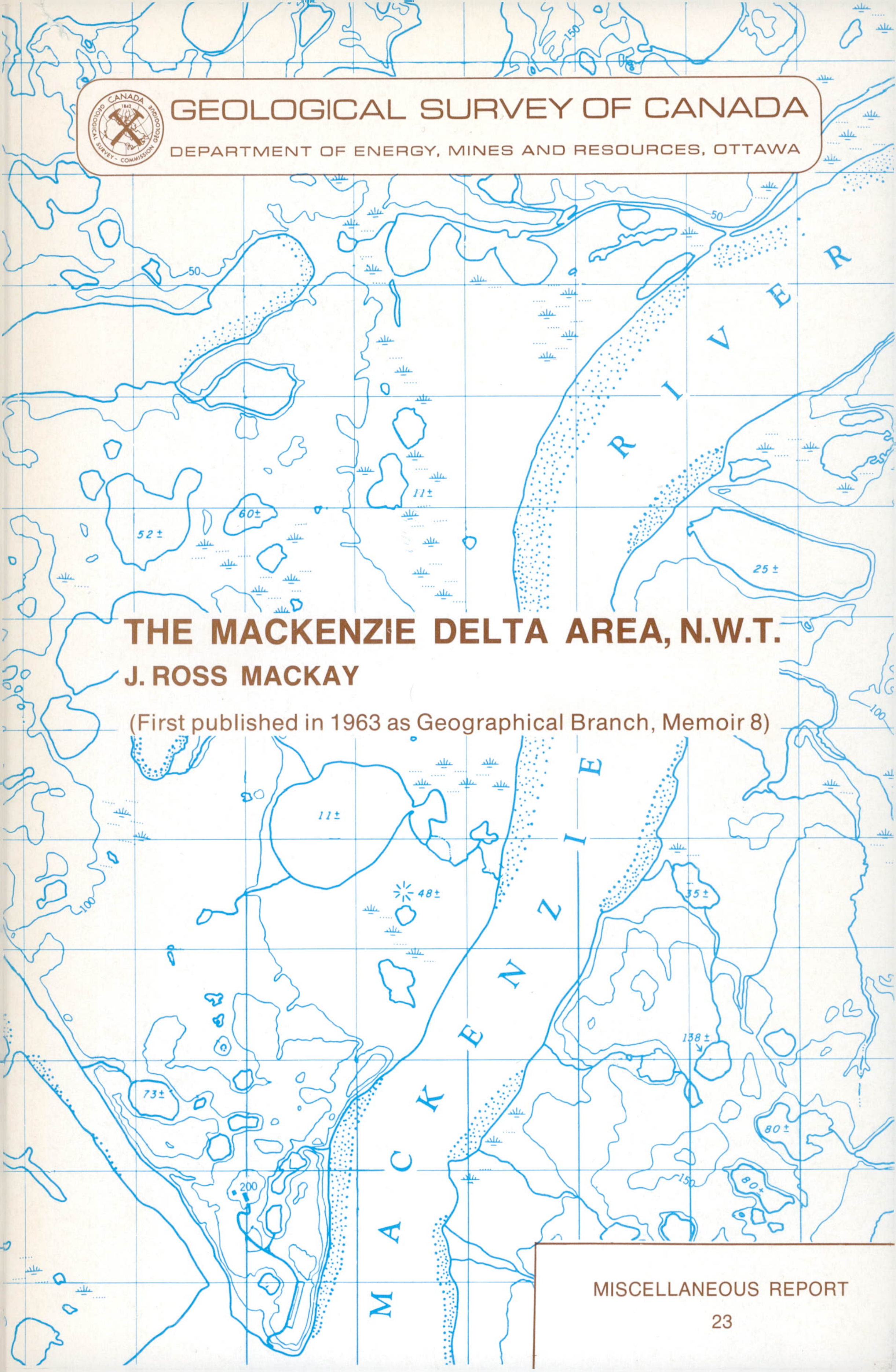
GEOLOGICAL SURVEY OF CANADA
DEPARTMENT OF ENERGY, MINES AND RESOURCES, OTTAWA



THE MACKENZIE DELTA AREA, N.W.T.

J. ROSS MACKAY

(First published in 1963 as Geographical Branch, Memoir 8)

MISCELLANEOUS REPORT
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Foreword

This report, published in 1963, has been out of print for some years but because of current interest in this part of the Canadian North, especially its possible use as a transportation corridor, it is timely that it be reissued. In October 1967, the programs of the Geographical Branch, under whose auspices Dr. Mackay's research was carried out, were transferred to other branches of the department. Studies in physical geography were continued by the Geological Survey and several of our recent bulletins present the results of studies initiated by the Geographical Branch. It is thus appropriate that this reprint be published as a Geological Survey report.

It has not been possible to use original photographs and line drawings in this reprint, which has resulted in some loss of detail. Certain of the demographic data, especially those in Chapter VI, are out of date after the elapse of more than a decade but to facilitate the reissue of this report no attempt has been made to include newer statistics derived from the 1971 census.

In his preface to this report, Dr. N. L. Nicholson, then Director of the Geographical Branch, wrote in 1963, "This memoir presents the results of surveys carried out over eight years, five summers of which the author spent in the field.The Mackenzie Delta is the largest modern delta in Canada and provides a natural laboratory for carrying out scientific studies and for testing arctic and subarctic techniques of investigation." The Delta is still an area of great interest and still serves as a laboratory for many scientific activities. I am pleased that Dr. Mackay's report on this important area is once more available to the general public.

D. J. McLaren,
Director.

Ottawa, July 1974

PREFACE

The Mackenzie Delta is the largest modern delta in Canada and provides a natural laboratory for carrying out scientific studies and for testing arctic and subarctic techniques of investigation.

This memoir presents the results of surveys carried out over eight years, five summers of which the author spent in the field. The field work was supplemented by airphoto interpretation and the most modern methods of statistical analysis were applied to much of the data obtained.

N. L. Nicholson,
Director, Geographical Branch

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CHAPTER I

INTRODUCTION

1. The Mackenzie Delta area is the centrum of population, economic activity, and transportation facilities in extreme northwestern Canada. Most of the area north of 68° 45'N is treeless, and so is wholly in the arctic. The settlements of Aklavik, Inuvik, and Reindeer Station are in the subarctic. The population of over 2,000 is about equally divided between those Eskimo, Indian, and white trappers who live mainly on the floodplain of the Mackenzie Delta and those who spend most of each year in the settlements of Aklavik, Inuvik, Reindeer Station, and Tuktoyaktuk. Although the total population is only slightly more than half of the estimated native population of 150 years ago, there are certain pressing demographic problems of population increase and overpopulation. The economy of the dispersed trapping population is unstable, as it depends upon fluctuating market prices and catches of fur-bearing animals, principally the muskrat and mink in the delta, and white fox along the coast. In the absence of roads and railways, the Mackenzie River has come to serve as a major transportation route, not only for the riverine settlements but also for those along the Arctic coast to the east and west of the delta.

For the purpose of this report, the Mackenzie Delta area is taken as nearly equivalent to the area covered by the Port Brabant map-sheet (National Topographic Series) on a scale of 8 miles to the inch. The area is, therefore, similar in size and shape to that covered in the report on the adjoining 8-mile Anderson River map-sheet to the east (Mackay, 1958a). As small portions of the Mackenzie Delta lie south and west of the Port Brabant map-sheet these areas are included in discussions when relevant.

The present account is based primarily on field work carried out in the summer field seasons of 1954 to 1958, and 1960. The areas observed at first hand were the Mackenzie Delta and adjacent terrain to the east and west; the coast; the full length of the Eskimo Lakes-Sitidgi Lake region; and the Anderson River upstream to its forks at the junction with Carnwath River. Information from the inland area between Sitidgi Lake and Anderson River has been based on air photo interpretation only. Published reports for this area are unavailable.

The emphasis in this report is on the physical geography of the area, because it was the principal objective of the field work. The human geography of the delta has been studied in part by those involved in the relocation of Aklavik and by others in recent years.

Acknowledgments

2. The writer would like to acknowledge the help of government officials, residents of the delta area, and assistants for many varied types of aid rendered over the years. In particular, K. H. Lang, of Aklavik, has been most helpful. D. K. MacKay and B. J. Morrison have carried out many of the computations presented in this report.

Historical Cartography

3. Mapping of the Mackenzie Delta area commenced with Alexander Mackenzie's historic voyage to the sea in 1789. Within 40 years, Franklin and Richardson had completed the coastal exploration of the areas to the west and east of the delta. The Anderson River was explored by MacFarlane in 1857 and Petitot reached Sitidgi Lake in 1869. Twenty years later, Sainville charted the Eskimo Lakes. Mapping of the area between Sitidgi Lake and the Anderson River has come from aerial surveys. Scientific literature has been very sporadic in publication and uneven in topical and areal coverage.

Mackenzie 1789

4. Descent of the Mackenzie River: Alexander Mackenzie, famous fur-trader and explorer, descended the river which now bears his name to the seaward perimeter of its delta in the summer of 1789. Mackenzie was born at Stornoway, Scotland, in 1764 (Wade, 1927, p. 18; some authorities state 1763), so that he was only in his mid-twenties when he accomplished his first major exploration, while in the employ of the North West Company. On June 3, 1789, Mackenzie's party left Fort Chipewyan in birch bark canoes. The party included four French-Canadians (two with their wives), a young German, and a group of Indians. Passing down the Slave River into Great Slave Lake, the party, after numerous difficulties, found the outlet of Great Slave Lake and started down the Mackenzie River on June 29. On July 10, Mackenzie writes: "The land is low on both sides of the river, except these mountains, whose base is distant about ten miles; here the river widens, and runs through various channels, formed by islands, some of which are without a tree, and little more than banks of mud and sand . . ." (Mackenzie, 1801, reprint edition, p. 256). From the description, it is clear that the party had reached the southern apex of the Mackenzie Delta, near Point Separation. Thus, on the morning of the twelfth day of travel from Great Slave Lake, the party had covered about 915 miles, or an average of about 70 to 80 miles per day. The speed of travel was, of course, increased by the current of the river which averages some 3 miles per hour. The estimated surface flow travel time (*162*)* of water from Great Slave Lake to Point Separation is about 13 days

*Italicized numbers in parentheses throughout the text are cross-indices to similarly numbered paragraphs.

or nearly the same as Mackenzie's over-all travel time. If the party spent an average of 12 hours each day travelling by canoe, and if the resultant of the canoe and current speeds are additive, then the equivalent still-water distance paddled each day was approximately 40 miles, at the rate of slightly over 3 miles per hour.

5. In the Mackenzie Delta: Mackenzie's route through the delta is impossible to reconstruct from his map, as it is so generalized, but a good estimate may be made of it from his descriptions and a knowledge of deltaic conditions (Figure 1). The islands downstream from Point Separation were assuredly present in 1789, because Franklin's map, based on his travels of 1825-1826, show island groups much as they are today. Moreover, the islands antedate the spruce trees, which are well over 100 years in age, growing on them. Downstream from Point Separation, where the river splits into large and small channels, Mackenzie was faced with the selection of the best channel to follow. His guide preferred the easternmost channel (East Channel), because there would be less likelihood of meeting with hostile Eskimos. Mackenzie on the other hand, preferred and followed the Middle Channel, because it was larger, trended from south to north, and so suggested that it emptied into the Hyperborean Sea.

Having decided upon taking the Middle Channel, the course was west by north for 6 miles and northwest by west for 32 miles. All course directions given by Mackenzie were magnetic, so that the variation of the compass in 1789 must be considered. Franklin, in 1825, found that his magnetic variations were about 15° farther east than those of Mackenzie, of which one in the delta was N 36°E. Using a variation of about 36° east, the first 6 miles was probably measured from the takeoff of East Channel, and the next 32 miles along the roughly north-south section of Middle Channel. An observation during the day gave 67° 47′ N. The magnetic course was then north-northwest 4 miles, northwest 3 miles, northeast 2 miles, northwest by west 3 miles, and northeast 2 miles. The description of this meandering section of the river corresponds reasonably well with the former course of the river around the large meander bend at 68° 15′ N. The meander neck cutoff probably did not exist in 1789, because Franklin's map of 1828 shows the bend, without cutoff, and he even took an observation for latitude at the bend; moreover, Sainville's map (1898) shows the main channel going around the loop and only the beginnings of a cutoff. Camp for the night was apparently made shortly after rounding the bend. That night, Mackenzie observed the midnight sun for the first time.

The next day, July 11, the party left early in the morning (about 4:00 a.m.), steering northwest (i.e., about true north) along a serpentine course. At about 7:00 a.m., they saw a ridge of high land, which would be the Caribou Hills, to the east. In some three hours of paddling, the party would have travelled about 20 miles. If they had started from a camp near the meander loop, they would have reached a position in the vicinity of Oniak Channel, from where the Caribou Hills stand out prominently as seen from Middle Channel. The party landed at noon, having travelled for 5 hours since first observing the Caribou Hills. By assuming an over-all speed of slightly over 6 miles per hour, the distance travelled would be a little over 30 miles. This would place the position just north of 68°

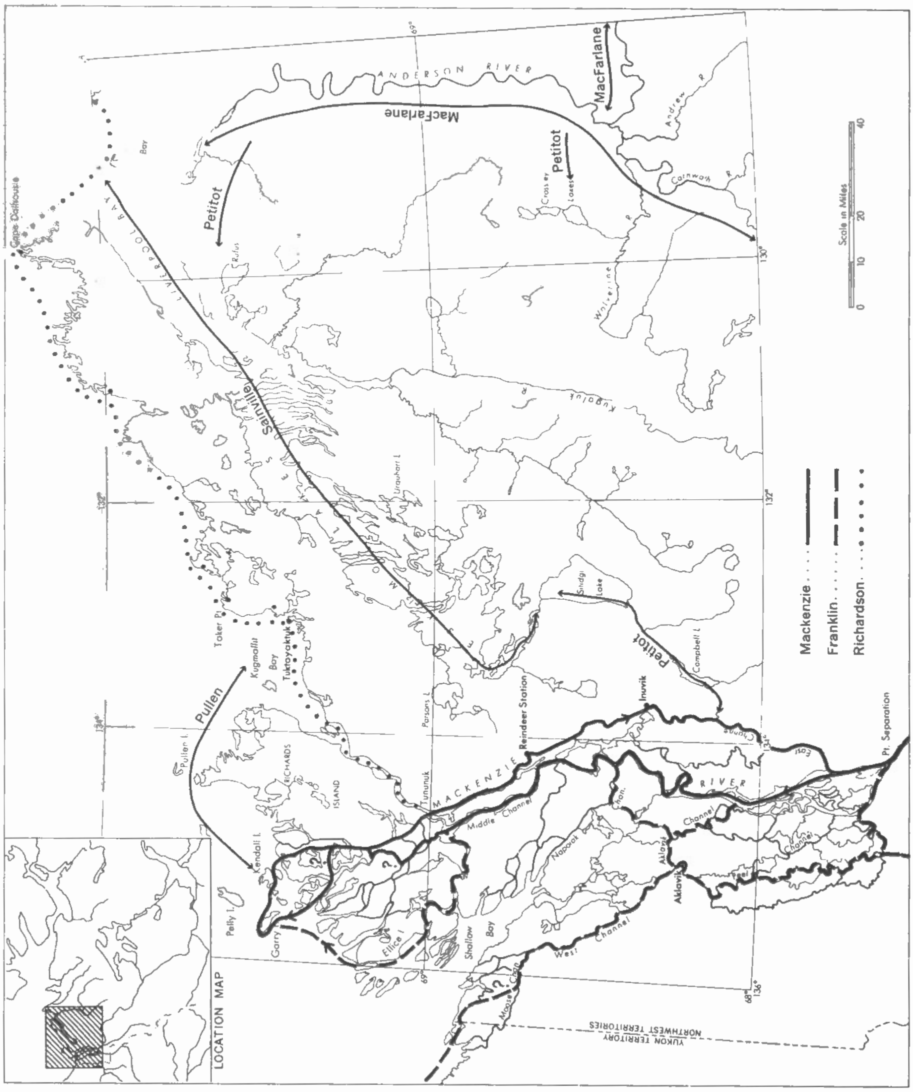

Figure 1. Exploration and location map.

50′ N. The description of the river, downstream from the noon landing, reads: "From this place for about five miles, the river widens, it then flows in a variety of narrow meandering channels, amongst low islands, enlivened with no trees, but a few dwarf willows". (Mackenzie, 1801, reprint edition, p. 260). The description applies, at least now, to the stretch between 68° 52′ N and 68° 56′ N.

Mackenzie does not state what channel he took to leave the broad, two-mile-wide lake-like expansion of the river at 68° 56′ N. It is certain that Mackenzie did not take Reindeer Channel, which trends southwest in a direction diametrically opposite to that described by him. Likewise, there is no evidence that he took the channel, which initially flows eastward, to Tununuk. At the present time, there is a third channel leading northwards. It was probably in existence in 1789, for Franklin (1828, p. 31), writing of his 1825 trip, states: "An attentive perusal of Sir Alexander Mackenzie's Narrative leads me to the conclusion, that it was this northern branch which that traveller pursued in his voyage to Whale Island". All evidence suggests, therefore, that Mackenzie took the northern channel. There are no clues as to where camp was made that night.

At 10:00 a.m. on July 12, the party landed where the adjacent land was high, covered with short grass and flowers, and snow was seen in the valleys. The soil on the hills was described as yellow clay mixed with stones, a contrast to the stoneless alluvial soils of the delta. There is no doubt that Mackenzie had reached the higher Pleistocene delta, (*125*) either on the west side of Richards Island, or on one of the eroded outliers nearby. At the present time, the north-flowing channel swings to within a mile of Richards Island at 69° 09′ N; a major channel leads directly to Richards Island at 69° 10′ N; and another major channel follows the western side of the island at that latitude.

Upon re-embarking, Mackenzie had difficulty in deciding what route to follow, but as the stream set to the west, he went "to an high point, at the distance of about eight miles, which we conjectured to be an island; but, on approaching it, we perceived it to be connected with the shore by a low neck of land. I now took an observation which gave 69.1 North latitude". (Mackenzie, 1801, reprint edition, p. 265). It appears as if Mackenzie first entered the large series of lakes, which are silting up and were certainly much larger in Mackenzie's time, between 69° 11′ N and 69° 17′ N and about 135° 00′ W. After finding his way out of the lakes, he headed for the higher land of the detached portion of the Pleistocene delta at about 69° 15′ N to 69° 20′ N where altitudes rise from 50 to 100 feet above sea level (Figure 37, outlier IIIa). Mackenzie continues: "From the [high] point that has just been mentioned [Pleistocene delta], we continued the same course for the Westernmost point of an high island, and the Westernmost land in sight, at the distance of fifteen miles." Such a bearing and distance from the above location would have taken Mackenzie to Garry Island which is not only right in direction and distance, but is also high, much of the island rising over 150 feet in altitude. The shallow depths encountered by Mackenzie were also typical of the area in Franklin's time (1825) and are today. Mackenzie and an Indian (the English Chief) climbed to the highest part of the island (200 feet) and found any further course blocked by ice, which extended from the southwest (by compass) to the eastward.

6. Whale Island: Mackenzie called the island on which they landed on July 12 Whale Island after the white whales which he observed (*177*). A rise and fall of the water showed a tidal effect, thus indicating to Mackenzie that he had reached the sea.

The exact location of Whale Island is hardly open to debate, as it is almost certainly Garry Island. Mackenzie obtained two latitude observations on Whale Island which were 69° 14′ N and 69° 07′ N. For well over 100 years, Whale Island was shown on maps in a position west of Richards Island in the latitude given by Mackenzie, that is, considerably south of Garry Island. There is no land there which meets the description of Whale Island. Mackenzie states that "Whale Island was the westernmost land in sight"; it was 15 miles from a point of high land; and the "lake" was open to the west. Quite clearly, Whale Island stood alone, isolated, and far out to sea. In 1825, Franklin sailed to Garry Island, passing within 10 miles of the plotted position of Whale Island. As the area traversed by Franklin was shoal, with a maze of islands as at present, no isolated island remnant of the Pleistocene delta (i.e. Whale Island) could have stood in a large open bay 36 years earlier.

The Garry Island of Franklin is doubtless the Whale Island of Mackenzie, and both explorers took nearly the same route to reach it.[1] Garry Island meets all the major descriptions of Whale Island except for latitude, and Mackenzie realized that his survey was inaccurate: "In this voyage, I was not only without the necessary books and instruments, but also felt myself deficient in the sciences of astronomy and navigation. . ." (Mackenzie, 1801, reprint edition, Preface, p. ix). Garry Island lies at the terminus of a logical interpretation of Mackenzie's route. Mackenzie stated that Whale Island was 7 leagues long, not more than 0.5 miles wide, and east and west by compass: Garry Island is 7 miles long, 0.5 to 2 miles wide, and it trends northwest-southeast, which agrees with east and west by Mackenzie's compass when allowance is made for magnetic variation.

If Whale Island was indeed Garry Island, then Mackenzie had ill-luck in not encountering salt or brackish water. On Franklin's visit, there was salt water at Garry Island, but the boundary between salt and fresh water is transient, depending upon the discharge of the Mackenzie River, tidal currents, and winds.

Mackenzie saw two small islands in the ice, to the northwest by compass from Whale (Garry) Island. There are no islands in the described location today. It is quite possible that there were small islands, which have since been removed by wave action. Alternatively, Mackenzie might have seen "dirty" ice covered with debris from river break-up (*158*).

As Whale Island is probably Garry Island, the two latitude observations of Mackenzie were about 18′ too low. If the error were consistent, and 18′ is added

1. The initial suggestion that the Whale Island of Mackenzie was the Garry Island of Franklin was made to the writer by Dr. J. K. Stager (University of British Columbia). After completion of this Memoir, an article by Bredin (1962) has appeared presenting the same conclusion, although the reasoning is somewhat different.

to Mackenzie's two other observations in the delta, the results would be 69° 19′N for the high point 15 miles from Whale Island and 68° 05′ N for an observation on July 10. The figure of 69° 19′ N for the high point agrees well with the topography and that of 68° 05′ N for the route of Mackenzie. The variation of the compass, as given by Mackenzie, appears too small when applied to the direction of the channels travelled.

7. *Return Journey:* The period of July 13 to 15 was spent at Whale Island and on July 16 the party set out on their return journey. They first sailed among the islands and then ascended the river. At 2:00 p.m. the water was quite shallow, the bottom always being less than a paddle length. Such shallow areas are typical of the distributary mouths. On both this day and the following, Mackenzie mentions passing recently occupied Eskimo encampments and he discusses river depths. These suggest that a different, and certainly more easterly, channel was taken on the return route.

On July 17 the party landed "upon a small round island, close to the Eastern shore, which possessed somewhat of a sacred character . . .", which is the island south of Tununuk (*125*). The route just northwest of Tununuk may be that marked on Figure 1. Although there is no through channel at low water at present, geomorphic evidence shows the existence of a former channel which still carries water at spate (*93*). Mackenzie took the East Channel back instead of the Middle Channel. This is clear from his description of the high land (Caribou Hills) along the river. As detail is largely absent from Mackenzie's description of his return route, it is very difficult to make even an enlightened guess of his daily mileage. On July 18, Mackenzie refers to the red earth with which natives bedaub themselves. This is probably the red material left as a result of spontaneous combustion, the burnt beds occurring south of Reindeer Station along the west slope of the Caribou Hills (*138*).

On July 20, they passed a river, where they expected to meet some of the natives. Judging by earlier accounts, this was a river which natives used as a route to the sea. It was probably Gull (Campbell) River which leads to Campbell Lake and a low-level portage route to Sitidgi Lake and then via the Eskimo Lakes to the sea. Mackenzie's descriptions of the lower heights of the hills (in comparison with the higher Caribou Hills to the north), and also the vegetation of spruce and birch to the summits, is descriptive of the hills by Gull River. At 10:00 a.m. on July 21 the party left the delta at Point Separation and again entered the main channel of the Mackenzie River.

The return trip of 175 miles from Whale (Garry) Island to Point Separation took nearly five and a half days. The daily average was excellent because the party was canoeing upstream against a current of about 2 miles per hour. Thirty-three days after leaving Point Separation, the party reached Great Slave Lake, thus averaging about 28 miles per day against the current. They reached Fort Chipewyan on September 12, having gone to the mouth of the Mackenzie and back in 102 days.

Mackenzie's contribution to knowledge of the Mackenzie Delta area was its discovery, and not its accurate description or mapping. He was young, inexperienced in the scientific aspect of exploration, and travelled long tedious hours whenever weather permitted. Mackenzie has been criticized for not pressing on until he encountered salt water, but he has also been defended by men of the calibre of Franklin. As the shallow waters off Whale (Garry) Island are very dangerous to navigate in bad weather and fogs are common when ice is near, Mackenzie showed excellent judgment and self-discipline in returning when he did.

Period of 1789-1825

8. In the period between Mackenzie's descent to the mouth of the Mackenzie and that of Franklin in 1825, fur-trading spread northward along the Mackenzie system. In 1799, two explorers (Sutherland and Livingstone) were killed in the delta, probably not far from the junction of Middle and Oniak Channels (Franklin, 1828, p. 30). In 1809, a Mr. Clarke reached the delta (Wernerian Natural History Society, Edinburgh, 1822). There seem to have been no additions to geographical knowledge made by explorers between Mackenzie and Franklin, because a map of the delta drawn in 1821 (Wentzel, 1821) shows no change from Mackenzie's crude map.

Franklin 1825-1826

9. Franklin 1825: In the first quarter of the nineteenth century, following the publication of Mackenzie's journals, there was much scientific, commercial and political interest in the exploration of the unmapped Arctic coast to the east and west of the Mackenzie Delta. Captain John Franklin of the Royal Navy, who had journeyed in 1821 to the mouth of the Coppermine River in company with Dr. John Richardson, proposed to the British Government "a plan for an Expedition overland to the mouth of the Mackenzie River, and thence, by sea, to the northwestern extremity of America, with the combined object, also, of surveying the coast between the Mackenzie and Coppermine Rivers" (Franklin, 1828, p. IX). The expedition was soon approved, and Franklin was placed in command.

The plans for the expedition called for a descent of the Mackenzie River in 1826, with Franklin going west along the coast to Kotzebue's Inlet to rendezvous with a boat, the H.M.S. *Blossom,* and for Dr. Richardson and party to travel east to Coppermine River. However, as Franklin's party reached the Great Bear Lake area early enough in the summer of 1825 for initial explorations, arrangements were made for Franklin, accompanied by Mr. E. N. Kendall (surveyor) to make a quick trip to the sea in order to obtain information from Eskimos on coastal conditions.

On August 12, Franklin reached the southern part of the delta. He proceeded down the Middle Channel, following Mackenzie's route, but mapping it in greater detail. Franklin's route diverged from that of Mackenzie's when he took Reindeer Channel instead of continuing northward (Figure 1). Franklin skirted the west side of Ellice Island following the coast to its northernmost point. From there, he headed for an island which loomed blue to the northeast, passing from fresh to brackish water before reaching it (Garry Island). He had thus succeeded, unlike Mackenzie, in reaching saline water. After a short stay on Garry Island, the party headed back for Ellice Island and retraced their course upstream. On the 19th, while travelling up Middle Channel in the fog, they inadvertently took a westerly channel, probably that of the Napoiak-Schooner-Aklavik Channel, which eventually led them back to Middle Channel. If this is in fact the route which they followed, it seems peculiar that they did not observe they were moving with the current when they entered Napoiak Channel, and yet there do not appear to be any other channels which they could have taken. By the 20th, they were out of the delta.

As a result of Franklin's 1825 reconnaissance trip to the coast and return, Middle Channel was more accurately mapped, and Reindeer Channel, Ellice, Garry, Kendall, and Pelly Islands were explored.

10. Franklin 1826: Franklin's and Richardson's parties left their base of Fort Franklin on Great Bear Lake, in June 1826, and by July 3 they had reached the Mackenzie Delta. They camped that night at Point Separation on the left bank of the river "at the head of a branch that flowed towards the Rocky Mountains". Later references by Richardson (Franklin, 1828, p. 188; Richardson, 1851, p. 223-227), and the 1828 map of Franklin, all show Point Separation on the left (west) bank of the Mackenzie, and not on the east bank, as in present maps (Figure 1). Leaving Point Separation on the 4th, Franklin travelled in H.M.S. *Lion,* and Lt. George Back in H.M.S. *Reliance,* each with a crew of seven. Franklin's descriptions of the channels are difficult to follow, but he seems to have taken the southern (unnamed) upstream continuation of Aklavik Channel. He then probably descended Aklavik Channel down to West Channel, the description of this portion being reasonably clear. West Channel was followed nearly to its mouth and then part of Ministicoog Channel (?) to Shoalwater Bay. Seeing some Eskimo tents on Tent Island, the party approached, but the boats grounded when still a mile from shore. At first the Eskimos were friendly, but after their rapaciousness was whetted in learning of the material contents of the boats, the *Lion* and *Reliance* were dragged to the south side of the river and pillaged. Franklin and party were fortunate to escape massacre at the spot aptly named Pillage Point.

The party continued west along the Alaskan coast, going as far as weather and ice conditions permitted (beyond Barter and Flaxman Islands). When they reached the point of no return on August 18, they began to retrace their steps. The journey upstream in the delta was similar to that downstream between Pillage Point and the junction of Aklavik and West Channel. The party missed the turnoff to the east at Aklavik Channel and continued upstream, possibly in the Husky Channel, but more probably along Peel-Phillips Channels, and then up the Peel River a short distance. Discovering their error, they descended the Peel River and reached Point Separation again on September 3.

In his voyage of 1826, Franklin mapped several channels on the west side of the delta and the coast for nearly 400 miles towards Point Barrow, a remarkable, though disappointing, achievement,—he did not attain his objective of making contact with H.M.S. *Blossom* and the Bering Sea.

Richardson 1826 and 1848

11. Richardson 1826: After leaving Franklin's party on July 4, Richardson with five men in H.M.S. *Dolphin* and Kendall (surveyor) with five in H.M.S. *Union,* set out to map "the coast between the Mackenzie and Coppermine Rivers, and to return from the latter overland to Great Bear Lake" (Franklin, 1828, p. 188). The party proceeded by way of East Channel which they mapped, since Mackenzie had not done so. The party continued past Tununuk and then took the hitherto uncharted lower course of East Channel to the sea. They camped the night of July 6 at Lucas Point. At Point Encounter the party, as happened with Franklin's at Pillage Point, narrowly escaped serious pillaging and bodily harm from Eskimos. By the 8th, the party had reached Refuge Cove, in the shelter of Topkak Point. Passing Toker Point and Warren Point, Richardson states "we crossed the mouth of a large river, the water being muddy and fresh for a breadth of three miles, and the sounding lead was let down to a depth of five fathoms, without striking the bottom. This river is, perhaps, a branch of the Mackenzie, and falls into a bay, on which I have bestowed the name of my esteemed friend, Copland Hutchison" (Franklin, 1828, p. 214). As we know now, Richardson made one of his few mapping errors here, as there is no such river. He doubtless crossed an unusually deep channel, with a current of Mackenzie water. Proceeding along the coast, the party arrived at Cape Dalhousie on July 16 and, finding the land trending to the southeast, they headed in that direction, reaching Nicholson Island (actually Nicholson Pensinsula). From the summit of the peninsula they could see a body of water (Anderson River) to the south, and, to the southwest, the Eskimo Lakes whose existence was known to all previous explorers. Leaving Nicholson Peninsula on July 17, the party continued their explorations, and after many difficulties completed the objectives of their coastal surveys.

The eastern detachment, with the excellent surveying of Kendall, mapped East Channel and the coast with what must be regarded as high precision when their difficulties are considered. The principal topographic features were named, and Richardson made numerous detailed observations on the geology, flora, and fauna of the region.

12. Richardson 1848: In 1848, Richardson joined in the vast and prolonged search for the lost expedition of Sir John Franklin. Richardson descended the Mackenzie, revisited Point Separation, and then again took the East Channel as he did in 1826. Richardson's narrative (Richardson, 1851) drew upon his earlier observations in considerable degree, but he added a number of topographical and geological observations from his later visit. The course being very similar to that followed in 1826, little cartographical information was added to the area under discussion. The party continued eastward, without finding any trace of Franklin.

Pullen and Hooper 1849-1850

13. The search for Sir John Franklin, which took Richardson in 1848 along his earlier route of 1826, also resulted in further exploration. On behalf of the Admiralty, Pullen and Hooper sailed from England via Alaska in 1849 and ascended the Mackenzie River. In 1850 they went down the Mackenzie and then via Garry Island and north of Richards Island to Cape Bathurst (Hooper, 1853; Pullen, 1947). Heavy ice prevented them from proceeding beyond Cape Bathurst, and they retraced their steps to the delta. On their outward journey they observed new islands which they named and investigated on their return journey. They landed on, and named, Rae and Hooper Islands. They observed Pullen Island and took bearings on it but did not land. With the area north of Richards Island having thus been traversed, all of the coast was partially mapped by 1850.

MacFarlane 1857-1866

14. Roderick Ross MacFarlane, like his predecessor Alexander Mackenzie, was born at Stornoway (Deignan, 1947) and served the Hudson's Bay Company from 1852 to 1894. For 15 years, the Athabaska District was under his control (Kerr, 1953). MacFarlane was an indefatigable collector of museum material (much being sent to the Smithsonian Institution), his collector's interest partly explaining his desire for travel. His collections made important contributions to knowledge of natural history and ethnography.

In 1857, MacFarlane left Fort Good Hope for the Anderson River. He descended the Iroquois and Lockhart (Carnwath) Rivers to their junction (the "forks") with Anderson River which he followed northwards, passing first Indian and later Eskimo encampments en route. When he was within a few miles of the mouth he was turned back by hostile Eskimos. His arduous return route was overland to the forks at 68° 25′ N where he secured a canoe and was able to proceed upstream on the Anderson River to continue his explorations. Two years later, MacFarlane finally reached the mouth of the river (MacFarlane, 1905, p. 681) at Wood Bay.

In 1861, MacFarlane established Fort Anderson on the right bank of the river some 35 miles downstream from the forks. Although Fort Anderson was abandoned in 1866, MacFarlane crossed four times to Franklin Bay to collect specimens for the Smithsonian Institution. The presence of Fort Anderson was an encouragement for others, such as Petitot, to venture into the area. MacFarlane has left us no map of his travels, but his publications, specimens, and topographical observations recorded in Petitot's 1875 map have been, for nearly a century, the standard reference for the Anderson River area (MacFarlane, 1890-91, 1892, 1905, 1908; Preble, 1908).

Petitot 1865-1872

15. Father Emile Petitot was one of the most widely travelled missionaries of his day. Although many missionaries explored in frontier areas, few wrote of their travels as assiduously as Petitot, and most of their information is lost forever. Petitot made exploratory trips from Fort Good Hope, where he was stationed, into the large unexplored area lying between Great Bear Lake, the Arctic coast to the north, and the Mackenzie River to the west (Petitot, 1875a; 1875b; 1876; 1887; 1889; 1893).

Although Petitot was neither a surveyor nor a scientific explorer, he was fascinated with the country through which he travelled. He was interested in, and often critical of, the maps and comments of others, yet he drew freely upon them in the preparation of his own maps. His map of 1875 shows numerous rivers, lakes, and mountains whose inferred locations were based primarily on Indian and Eskimo information, hearsay, and imagination.

Petitot followed MacFarlane's route down the Anderson River in 1865. Using the Anderson River as a starting point for his explorations, he walked northwest from near its mouth almost to the Eskimo Lakes without, however, sighting them. From Fort Anderson he journeyed west to beyond Crossley Lakes and southeasterly to Simpson Lake. In the Mackenzie Delta area, Petitot crossed the portage route from the Mackenzie system to the Eskimo Lakes via Gull River, Campbell Lake, and Sitidgi Lake in 1869.

Petitot had strong convictions on the presence, or absence, of certain topographic features when he compiled his 1875 map which influenced cartographers for many years. Among other things, Petitot greatly under-estimated and discounted the reported size of the Eskimo Lakes. He insisted that two large rivers discharged into Liverpool Bay when evidence from other explorers indicated only one, the Begh'ula-tessé of Richardson or the Anderson River. On the other hand, Petitot was exploring in an inland area far removed from surveyed points established along the major water bodies (Arctic coast, Mackenzie River, and Great Bear Lake) so that positions were difficult to estimate. Petitot also contributed much to the ethnography of the area.

Sainville 1889-1894

16. The French explorer, Count de Sainville, spent four years in the lower Mackenzie area (Russell, 1898, p. 138; Sainville, 1898). He had surveying instruments with him, and his map of 1898 provides us with the first detailed information on the Eskimo Lakes. He mapped Sitidgi Lake and all of the Eskimo Lakes as far as Liverpool Bay. Curiously enough, however, his observations for longitude are quite erroneous, because he places Warren Point and Hutchison Bay at the east end of Liverpool Bay, in the relative position of Cape Dalhousie. No land is shown to the east of Warren Point, although Richardson had mapped the coast sixty years earlier. The estuarine mouth of Kugaluk River is modestly named "Port de Sainville" although the depths then (3 feet) and now provide a suitable harbor only for rowboats and canoes. Despite the earlier descents of Anderson River by MacFarlane and Petitot, and the publication of Petitot's 1875 map, the mouth of the Anderson River is sketched in only on the basis of Eskimo reports. Garry, Kendall, Pelly, and Pullen Islands are located, but "Whale Island" is missing.

Stone 1898

17. In 1898, A. J. Stone, an American naturalist collecting for the American Museum of Natural History, travelled along the Arctic coast east of Richards Island as far as Darnley Bay (Stone, 1900). As so often happened with early explorers, one person was unfamiliar with the work of another and Stone was no exception. He mapped a large river entering East Channel near Reindeer Station and named it Jessup River. As there is no river of any consequence entering East Channel anywhere near the indicated position of Jessup

River, perhaps Stone was misled by the apparent size of a drowned creek mouth under winter conditions. Stone, as Petitot did before him, also discounted the size of the Eskimo Lakes. He transferred Nicholson Peninsula from its accurately plotted position at the mouth of Anderson River to an unnamed imaginary island off Maitland Point. On the positive side, Stone did point out some errors in Petitot's map of 1875, such as the non-existence of Petitot's "Fleuve MacFarlane" flowing into Liverpool Bay east of Anderson River.

Harrison 1905-1907

18. A. H. Harrison spent about two years surveying and mapping in the Mackenzie Delta area, often under extremely adverse physical conditions (Harrison, 1908). Like Sainville before him, he took the portage route from Campbell Lake (referred to as Long Lake) to Sitidgi Lake (referred to as the First Eskimo Lake), where he wintered by the outlet creek. Harrison's map of the Eskimo Lakes is basically similar to that of Sainville to which no reference is made. Likewise, Harrison appears to have disregarded Richardson's 1826 survey of the coast north of the Eskimo Lakes, although some names (e.g., Warren Point) are included, but grossly misplaced. It seems rather peculiar that Harrison ignored most of the early and fairly accurate coastal maps. Moreover, by Harrison's time, whalers were well acquainted with the main coastal features, and Harrison had met them at Herschel and Baillie Islands. By 1907, when Harrison left, most of the principal geographic features of the Mackenzie Delta, the coast, and the Anderson River were mapped. Recent precision in maps has come from air photography. Nevertheless, few geographic names have been added in the past 100 years.

CHAPTER II

PHYSICAL GEOGRAPHY

19. The landscapes of the Mackenzie Delta area are a reflection of the geomorphic history of the region. For an understanding of the landscape of the region and the surficial deposits, the Quaternary history is most important. In this chapter, the Pleistocene deposits, glaciation, post-glacial history, and principal geomorphic processes are considered. Some aspects of physical geography are deferred to the chapter on Regional Physical Geography in order to provide a more unified discussion of the regions in question.

Pleistocene Deposits

20. Distribution: Pleistocene fluvial, deltaic, and estuarine sediments are found along the northern coast. Most of Richards, Garry, Kendall, Pelly, Hooper, and Pullen Islands, the area north of the Caribou Hills, Tuktoyaktuk Peninsula, the southern coast of the Eskimo Lakes, Nicholson Peninsula, and Cape Bathurst, are believed to be composed largely of Pleistocene sediments. A marine Pleistocene formation mantles the Yukon coast west of the delta, and also that of the coast east of Cape Bathurst (O'Neill, 1924, pp. 29A-33A).

Most of the Pleistocene sediments lie below an altitude of about 150 to 200 feet. The few higher areas can be explained as a result of glacial ice thrusting or the growth of thick tabular ground ice sheets which may add 20 to 30 feet to the altitude of the terrain. With few exceptions, the blanket of glacial deposits is too thin to modify the relative relief of the underlying sediments. The regional slope of the terrain is towards the north.

Along the flanks of the Caribou Hills, the Pleistocene sediments may terminate in a bluff of erosional origin. The bluff, which may be subdued, marks the erosional contact of the Caribou Hills with the younger Pleistocene sediments, in the same way that sediments of the modern Mackenzie Delta are being deposited along an erosional contact between Inuvik and Reindeer Station. To the northeast of Sitidgi Lake, the limit of the Pleistocene sediments may lie along a fault-line scarp. East of 132° 00′ W, and south of the Eskimo Lakes, there is no distinct topographic break marking the inland limit of the Pleistocene sediments.

21. East Channel Area: Most of the exposed sections are along freshly cut river banks and sea cliffs, because slumping quickly obscures the stratigraphic succession. In the Mackenzie Delta area, the southern limit of the continuous Pleistocene beds is at 68° 52′ N, on the right bank of East Channel, where the younger Pleistocene sediments abut with an erosional contact against the higher and older Caribou Hills to the south (Mackay, 1956a, pp. 7-9). The first 15 feet above river level when examined was obscured by slumping. From 15 to 25 feet above river level there is sand with silty lenses, iron staining, and wood fragments. Above this there is 5 feet of clay and silt and then an abrupt contact with the peaty layer above. The 2-to-3-foot-thick layer of peat includes branches up to 2 inches in diameter and a few logs up to 5 inches in diameter. Some of the logs are partly carbonized, like charred wood. The peaty material is locally cemented with iron oxides. In places, the peaty layer may be interbedded with silt and sand containing shells. At 60 feet above river level there is a second continuous peat bed from 2 to 3 feet thick. From the 60-foot level to 130 feet there is silt interbedded with peat and lenses of sand and gravel, with vegetation-covered slumped material from 130 feet up to the top at 150 feet. The succession described above is typical of the deposits which are being built in the modern Mackenzie Delta. The peaty beds with shells and water-worn driftwood are very similar to those exposed along cut banks of the Mackenzie Delta. No plants in the peaty layers were seen in situ with roots penetrating into a soil horizon below, nor were any casts of the ice-wedges of tundra polygons seen. The thick deposits over the basal peat imply deposition in a subsiding area or rising sea.

22. A specimen of peat, obtained from the lower portion of the above section at an altitude of about 18 feet above sea level had a radiocarbon age of more than 44,000 years (L-522A). A pollen analysis (by Dr. J. Terasmae, Geological Survey of Canada) yielded 61 per cent tree pollen and 39 per cent non-tree pollen. The composition of the tree pollen was: 11 per cent pine, 3 per cent black spruce, 33 per cent white spruce, 6 per cent birch, 35 per cent alder, 11 per cent willow, and 0.6 per cent tamarack. Peat from a nearby section, at a similar if not identical stratigraphic position, has been considered as of interglacial age on the basis of palaeobotanical and palynological study (Terasmae, 1959). An interpretation of the evidence indicates that the vegetation at the time of peat accumulation was similar to that of the present, if not indicative of slightly warmer conditions. The presence of tamarack cones and spores, about 40 miles north of the present local limit for tamarack, suggests a slightly more favourable climate.

23. Cut bluffs along the right bank of East Channel from 68° 52′ N to Tununuk, and along both banks of East Channel from Tununuk to Kugmallit Bay, expose brown sands and silts, with peat beds being observed only south of Tununuk. The sands and silts are nearly horizontally stratified although many beds show current crossbedding. The deposits are remarkably stonefree, but contain many wood fragments ranging from fresh-looking, water-worn chips, twigs, and branches, to water-worn slabs of carbonized wood. Mammoth teeth and tusks have been found in the bluffs up to an altitude of at least 50 feet. A bison skull (probably *Bison crassicornis*) was picked up by natives on a beach 4 miles southwest of Tuktoyaktuk. The skull and horns show no evidence of abrasion from distant transport by water or ice.

24. Inuvik: At Inuvik (Figure 73) on the south side of Boot Creek three-quarters of a mile east of East Channel, a specimen of wood obtained from cross-bedded sands has been dated at over 39,000 years (G.S.C.-29). The specimen was at an altitude of 100 feet above sea level and beneath an estimated 30 feet of overburden which had been removed for construction purposes. Below the cross-bedded sands containing the specimen was 5 feet of tabular ground ice and then gravel containing carbonized wood. The stratigraphy of the Inuvik site is difficult to interpret, despite the large number of test pits which have been dug in the area for construction purposes. One reason for the difficulty of interpretation is the presence of ground ice, varying from massive clear bubbly white ice to icy silt. Ice sheets occur at depths of 6 to 10 feet from the surface, with individual sheets being as much as 20 feet thick. A common succession, from the surface down, is: (a) 0.5 feet of organic matter or turf; (b) 5 to 8 feet of coarse brownish gravel containing some 3-to-4-foot Shield boulders, with the upper part being somewhat shingly; (c) at a depth of about 8 feet below the surface, there is a 1-to-2-foot iron-cemented hardpan beneath which the gravel is unweathered and grey, being lithologically and texturally similar to the brownish gravels above the hardpan; (d) a grey sand or grey silt lies beneath the grey gravel—ice lensing is common in the silt; and (e) silts, sands, and gravels below to an undetermined depth. Horizon (a) represents the postglacial accumulation of organic matter. The gravels of horizon (b) may be of glaciofluvial origin laid down between ice to the west and the higher land to the east. Horizon (d) represents shallow water deposition as the sands show crossbedding typical of deposition in pools. The sand contains small pieces of driftwood and some drift coal. The radiocarbon-dated wood came from horizon (d). If the driftwood was not redeposited old wood, then an interglacial age is suggested for horizon (d).

25. Richards Island: Most of Richards Island is believed composed of stonefree sands, silts, and clays with fresh and carbonized water-worn wood fragments. Silty clays and clays appear to occur more frequently towards the northern end of Richards Island and the offshore islands (Garry, Kendall, Pelly, Hooper, and Pullen) than farther to the south. Fossils found in Kendall Island include: *Macoma balthica, Natica* sp. (probably *N. clausa*), *Yoldia arctica*, and *Elphidium* sp. (identifications by Dr. F. J. E. Wagner, Geological Survey of Canada). The species are typically Pleistocene.

26. Tuktoyaktuk Area: In the Tuktoyaktuk area, cuts show clean stonefree sand, with thin silty laminations and fresh to carbonized wood. An unusual boulder bed was seen on the north side of 30-foot-high Port Brabant Island (Figure 75) at the harbor entrance of Tuktoyaktuk in the summer of 1954. The boulder bed averaged one to two feet in thickness and was followed discontinuously for 180 feet from an altitude of 20 feet above sea level to sea level; the bed apparently continued below sea level. The boulder bed was underlain by a foot of sandy silty clay. The boulder bed was composed of small and large boulders, some exceeding a diameter of 3 feet, in a matrix of sandy silts. Many of the larger boulders were typical crystalline Shield erratics with striated and faceted surfaces. The smaller boulders, such as those less than one foot in diameter, were exceptionally well rounded. The boulder bed was overlain by several feet of a poorly laminated sticky silty clay, above which was interstratified sand and silt containing water-worn carbonized wood, especially along the bedding planes. The island surface is remarkably horizontal, despite the presence of the tilted boulder bed and lenses of ground ice. No till or gravel occurs at the top of the bluff. The boulder bed does not seem to be buried till, because there is no disturbance of the underlying sand, the bed is not compacted, and some "balanced" boulders protrude into the overlying laminated silty clay whose laminations abut against them. The heterogeneity of the boulder bed suggests accumulation of ice-slumped debris in a depression where fine-grained sediments could be deposited later. The boulder bed resembles present accumulations of slumped debris below melting ground ice lenses *(63)*. Whatever the origin of the bed may be, glacially derived Shield boulders lie beneath stratified sands and clays.

27. In a study of pingos near Tuktoyaktuk, Müller (1959, p. 80; 1962) has obtained a number of radiocarbon dates for his sections, which are given here in terms of altitudes prior to pingo growth. At the Ibyuk pingo a radiocarbon age for driftwood above glacial till and accumulated some 18 feet below present sea level is 12,000 ± 300 years (S-69); at a depth of about 40 feet, and below glacial till, driftwood has been dated at 28,000 ± 2,000 years (Be-49); and one at a depth of about 50 feet at over 33,000 years (L-300A). These dates not only confirm the Pleistocene age of the lower sediments, but also provide narrow limits for the last ice advance. However, too much reliance should not be placed upon uncorroborated single dates.

28. Coast: The coast from Tuktoyaktuk to Cape Dalhousie is low, with altitudes generally within 30 feet of sea level. A few tilted sand and gravel beds occur south of Toker Point, but otherwise cuts observed at Warren Point, Point Atkinson, the east side of McKinley Bay, and Cape Dalhousie are in stonefree sands and silts.

29. Eskimo Lakes Sections: The sediments exposed along the cut bluffs of the highly indented Eskimo Lake coasts are similar to those already described, except for a far greater development of deltaic foreset bedding. Several cut bluffs of the first Eskimo Lake (first lake north of Sitidgi Lake) show interstratified silts, sands, gravels and peat, as in the section described for East Channel *(21)*. An excellent section is exposed along the north shore at 68° 46′ N, 133° 15′ W, in the bluff of a 155-foot-high flat-topped terrace. The lowest 30 feet is of silt and clay, with some sand. From 30 to 65 feet above lake (i.e., sea) level, the sediments are sands and silts, in places horizontally stratified, well laminated, or crossbedded.

Bits of carbonized wood, branches up to 2 inches in diameter, and peat beds from 2 to 6 inches thick are interstratified with the sand and silt. A 3-foot-thick compressed peat bed occurs at an altitude of 65 feet above the lake. From 65 to 155 feet in altitude, sands and gravels are exposed, the upper part being outwash material. A palynological report on a single sample of the peat indicates that climatic conditions were probably rather similar to the present (analysis by Dr. J. Terasmae, Geological Survey of Canada). The pollen and spore assemblage is compatible with an interglacial age for the buried peat bed. The tree pollen differs from that north of Reindeer Station *(22)* in the abundance of birch, rather than alder, and the absence of pine pollen.

30. Throughout the full length of the Eskimo Lakes, extensive foreset bedding is present. Individual foreset beds frequently extend through a 50-foot vertical range, thus showing the minimum depth into which delta deposition took place. The major source of sediment was probably a Pleistocene Mackenzie River, supplemented with flow from southern tributaries. The descriptions below are given from west to east.

At 68° 52′ N, 133° 15′ W, foreset beds over 50 feet in thickness are overlain by outwash gravels. In the coastal area south of Tuktoyaktuk, horizontal greyish sands and sandy silts are most commonly exposed in the bluffs, which are usually below 50 feet in height. Driftwood and gravel beds, containing Shield pebbles, are locally interstratified with the sands.

From about 132° 00′ W to 131° 15′ W, the sands are mainly greyish, without the brownish colour so characteristic of the sands of Liverpool Bay to the east. In this area occur grey sand with excellent foreset bedding, flattened driftwood logs, and gravel lenses with Shield boulders. Some beds are cut by shear planes; slickensided planes are confined mainly to fine-grained beds. From about 131° 15′ W to about 130° 30′ W greyish sands 10 to 15 feet thick with foreset bedding dipping north to east tend to be overlain with a sharp contact by brown sands. Organic matter in the grey sand is abundant along bedding planes.

Along the north side of Liverpool Bay, east of 130° 30′ W, coastal bluffs decrease gradually from about 60 to 70 feet in height to 15 feet or less. Gravel may occur at the tops of the higher bluffs but most are entirely of sands, many showing both topset and foreset bedding, with apparent dips ranging from north to east.

Along the south side of Liverpool Bay from about 130° 30′ W to 129° 35′ W, foreset beds of fine brown sands are very numerous. Individual beds of 50 feet vertical extent are common (Figure 2). Apparent dips are towards the northeast. Topset beds, with gravel containing driftwood and Shield boulders may overlie the foreset beds with their contact at about 55 feet or higher. The total thickness of sands and gravels over the foreset bedding may be 60 feet. Glacial till occurs as a veneer on top of the bluffs.

The south side of Liverpool Bay between Nicholson Peninsula and 129° 35′ W is predominantly of horizontal sands and silts with many ice lenses and much ice lens slumping.

Figure 2

Foreset bedding of sand along the south coast of Liverpool Bay at 69°40'N, 130°22'W, July 29, 1960. The bluff is 115 feet high. The foreset beds are about 55 feet thick. They are overlain by topset beds of sand and gravel (containing Shield rocks up to 6 inches diameter) with driftwood and then till. Logs and stumps up to 8 inches in diameter are found on adjacent hill slopes. (Geog. Br. JRM-60-3-4).

31. Nicholson Peninsula: Nicholson Peninsula is composed of sands, silts, and clays with ground ice segregations occurring abundantly in several localities. The heights attained by the sediments are greater than in adjacent areas of similar material, the anomalous altitudes and presence of extensive shear planes, tilted, folded, and overturned beds being attributed to glacial ice-thrust directed from the west (Mackay, 1956b). The marine pelecypod, *Yoldia arctica,* is abundant in the clays and silty clays. Fresh-appearing driftwood branches and logs, some of which are flattened, abound in some steeply dipping sandy beds to an altitude of over 100 feet. The sediments may include estuarine and marine facies of the fluvial and deltaic deposits of the Tuktoyaktuk Peninsula-Eskimo Lakes area. This is further suggested by the Pleistocene marine clays which occur along the east side of Liverpool Bay. At Cape Bathurst, opposite to Baillie Island, shells from a 25-foot clay bluff (identifications and comments by Dr. F. J. E. Wagner, Geological Survey of Canada) are:

Foraminifera

Quinqueloculina seminula (Linné)
Globulina glacialis Cushman & Ozawa
Elphidium clavatum Cushman
Elphidium frigidum Cushman
Elphidium orbiculare (Brady)
Elphidiella groenlandica (Cushman)

Pelecypoda

Yoldia arctica (Gray)

Ostracoda

Cyprideis sorbyana (Jones)
Cytheridea pencillata Brady
Cytheris sp.
Cytheropteron sp.

The species are found in Pleistocene deposits along the Arctic coast and are common in the Arctic seas today. The assemblage probably dates from late Pleistocene time. The depth of water in which the organisms lived is estimated at between 15 and 20 fathoms.

32. Grey Sands: In the region about Richards Island, cuts frequently expose near sea level a distinct light grey sand beneath contrasting brownish sands which form the greater part of the sediments. Wherever exposed, the contact is horizontal and very sharp, the transition from grey to brown being knife-edge. On the east side of "Corral Bay" (the first bay due west of Kidluit Bay) grey sand is exposed from sea level to an altitude of 40 feet. The bedding appears horizontal. Organic debris, like that found at present high water mark with wood chips, twigs, bark, etc., is locally abundant in the sand, as are flattened fresh-looking driftwood logs up to one foot in diameter and a few slabs of water-worn drift coal, similar to that found elsewhere. Near Kittigazuit, 20 feet of light grey sands are overlain with brown sands; by the abandoned Loran Station on East Channel, the lowest 30 to 40 feet is of light grey sand; at the mouth of Pete's Creek, on the south side, the bluff of East Channel is cut into light grey sand with carbonized wood; at Tununuk, crossbedded grey sands are exposed for 15 feet above river level; and at Kendall and Garry Islands, grey sand is also exposed.

The widespread distribution of brownish sands over grey sands, with sharp contacts varying from about 15 to 40 feet above sea level, indicates a consistent difference in the nature of the material. Both sands are unconsolidated, nearly stonefree, horizontally stratified with some current crossbedding, and contain fresh-looking driftwood as well as drift coal. It is not known if the contact is erosional; no sign of weathering or a buried soil profile was observed along the contact. The greyish sands resemble sands which might have been derived from erosion of the Caribou Hills.

33. Grain Size Distribution: The textural variation in grain size of the Pleistocene sediments is quite small. Although gravel and silty clay are present, by far the greater number of exposures are almost entirely composed of sand. In a "sampling" of over 30 sandy beds, from Richards Island to Nicholson Peninsula, most of the sand was in the medium to fine sand range (0.1 to 0.5 mm).* The percentage of silt and clay-sized particles is small, usually only a fraction of a per cent in the samples tested. Judging from the exposures of the deformed beds at Nicholson Peninsula, finer-grained sediments may underlie the sands.

*Note: The textural classes used in this report are: gravel (> 1 mm); coarse sand (0.5 to 1.0 mm); medium sand (0.25 to 0.50 mm); fine sand (0.10 to 0.25 mm); very fine sand (0.05 to 0.10 mm); silt (0.005 to 0.050 mm); and clay (< 0.005 mm).

34. Mammoth Remains: The dispersed distribution of mammoth tusks, teeth, and probably bison horns indicates a Pleistocene age of the sediments. Numerous writers have pointed out that the mammoth and great bison are known only from the Pleistocene in North America, mainly from the last half of the Pleistocene if not restricted only to the Wisconsin (Farrand, 1961; Hibbard, 1958, pp. 20-22; Simpson, 1947, pp. 621-622). An interglacial age has been suggested for the mammoths and great bison of Alaska-Yukon (Johnston, 1933, pp. 33-34).

Sainville (1898, p. 300) reports: "Je découvris une tête de cet animal [mammoth] en parfait état de préservation, récemment tombée d'une falaise où je vis 30 pieds au-dessus du niveau de l'eau et 15 pieds au-dessous de la mousse la place qu'elle avait occupée, les autres os avaient disparu précédemment". The location of the find was in the central portion of the Eskimo Lakes. Mammoth bones have been reported between Cape Bathurst and Herschel Island. Mammoth tusks and teeth have been found at Cape Bathurst (Mackay, 1958a, p. 25); near Stanton and up the Anderson River; at the mouth of Kugaluk River; and along East Channel north of Tununuk to Kittigazuit (Mackay, 1956a, p. 9). Mammoth tusks have been reported (without confirmation) from the lower Wolverine River area.

35. Age of Deposits: The sediments are believed to be of interglacial age as shown by radiocarbon dates, mammoth and great bison remains, and palaeobotanical evidence. Whether the Pleistocene deposits grade down into Tertiary sediments is unknown. The sediments with peat appear to have been laid down, in part, in a period of rising sea level. The thick foreset beds could only have been deposited along an actively migrating delta front in water at least 50 feet deep. Sea level stood over 100 feet higher than at present when the upper beds were deposited. More than one period of glaciation is suggested by the distribution of Shield pebbles in foreset beds.

The correlation between the Pleistocene deposits of the Mackenzie Delta area and the Beaufort formation of the northwest side of the Arctic Archipelago and the Gubik formation of the Arctic Coastal Plain of northern Alaska is unknown. The Beaufort formation (Tozer, 1956, pp. 25-28) of unconsolidated sands and gravels is considered to be of late Tertiary and perhaps early Pleistocene age. The Gubik formation, probably of late Pleistocene age, consists of unconsolidated marine and non-marine gravel, sand, silt, and clay.

Glaciation

36. The greater part of the Mackenzie Delta area has been glaciated. The ice did not advance westward beyond the front of the Richardson Mountains, and there is the possibility that the Cape Dalhousie area was unglaciated during the last ice advance. Direct evidence for late Wisconsin glaciation is based upon a single section at Ibyuk pingo, near Tuktoyaktuk *(27)*. Müller (1962) has described a till underlain by driftwood 28,000 ±2,000 years (Be-49) and overlain by driftwood 12,000 ±300 years (S-69) old. The till is considered by Müller to have been deposited subaqueously; however, it might also be slumped till resulting from the melting of a ground ice sheet, as similar-appearing "tills" are being formed by slumping at present (Figure 20). Craig and Fyles (1961) suggest that the coastal area might have been overriden by Laurentide glacial ice, possibly during a pre-Wisconsin glaciation. However, available evidence is in favor of at least Wisconsin glaciation of the area. Further study will probably give evidence of multiple glaciation as in Alaska.

37. Limits of Glaciation: The limits of glaciation are uncertain, just as they are imperfectly known for the mainland to the east and the Arctic Islands to the northeast. To the west of the delta, ice pushed to the Richardson Mountains and along the coast to just beyond Herschel Island (Bostock, 1948; Mackay and others, 1961a, p. 26). The westernmost feature (a submarine ridge) recognizable as probably of glacial origin is 25 miles due west of Herschel Island. To the north, field evidence for glaciation has been seen on Garry, Kendall, and northern Richards Island. Air photo interpretation suggests that Hooper and Pullen Islands were glaciated. Glacial ice probably covered all the northern coast of Tuktoyaktuk Peninsula as far east as Point Atkinson.

The northeastern part of Tuktoyaktuk Peninsula may have been unglaciated during the last ice advance, but it should be stressed that field investigations were limited to the east side of McKinley Bay, Cape Dalhousie, and the north coast of Liverpool Bay. Air photographs show no recognizable glacier features north of the inferred glacial limit (Figure 3) but many to the south. Erratics are very few at Point Atkinson; moreover some stones along the shore can be directly associated with transport in the root system of driftwood logs. The east side of McKinley Bay and the Cape Dalhousie beaches are stonefree. On the north side of Liverpool Bay, erratics are numerous west of 129° 55′W but sparse or even absent to the east. The lack of erratics along the shore may result from efficient removal by sea ice transport, but their absence (?) was also observed during inland traverses. Recognition of the limit of glaciation in the Cape Dalhousie area is confused by the fact that postglacial submergence may have obliterated any glacial features present. During submergence, erratics could have been ice-rafted to any area below the marine limit. Erratics are present on the east side of Wood Bay to an altitude of over 100 feet, but glacial features are not recognizable on air photographs of areas farther inland.

A large abandoned river channel links the Eskimo Lakes with McKinley Bay (Figure 4). The southern half of the channel shows glacial modification, the northern half is seemingly unmodified. In the extreme south, the channel vanishes a mile from the Eskimo Lakes. In proceeding northwards, the abandoned channel becomes increasingly better preserved and distinct with long lakes marking the old course. Glacial deposits, which appear in the southern part of the channel, decrease to the north, so that meander scars on slipoff slopes are glacially unscarred by 69° 43′N. From 69° 46′ N to McKinley Bay the channel scarps are 2 to 3 miles apart. The eastern side is a single, continuous, and distinct bluff with its base dropping from an altitude of more than 50 feet at 69° 48′ N to sea level. The western side is composed of three distinct bluffs at 69° 46′ N, their bases lying between 50 and 100 feet in altitude. The lowest bluff, which matches the eastern one in altitude, is locally subdued and modified by dunes blown by east

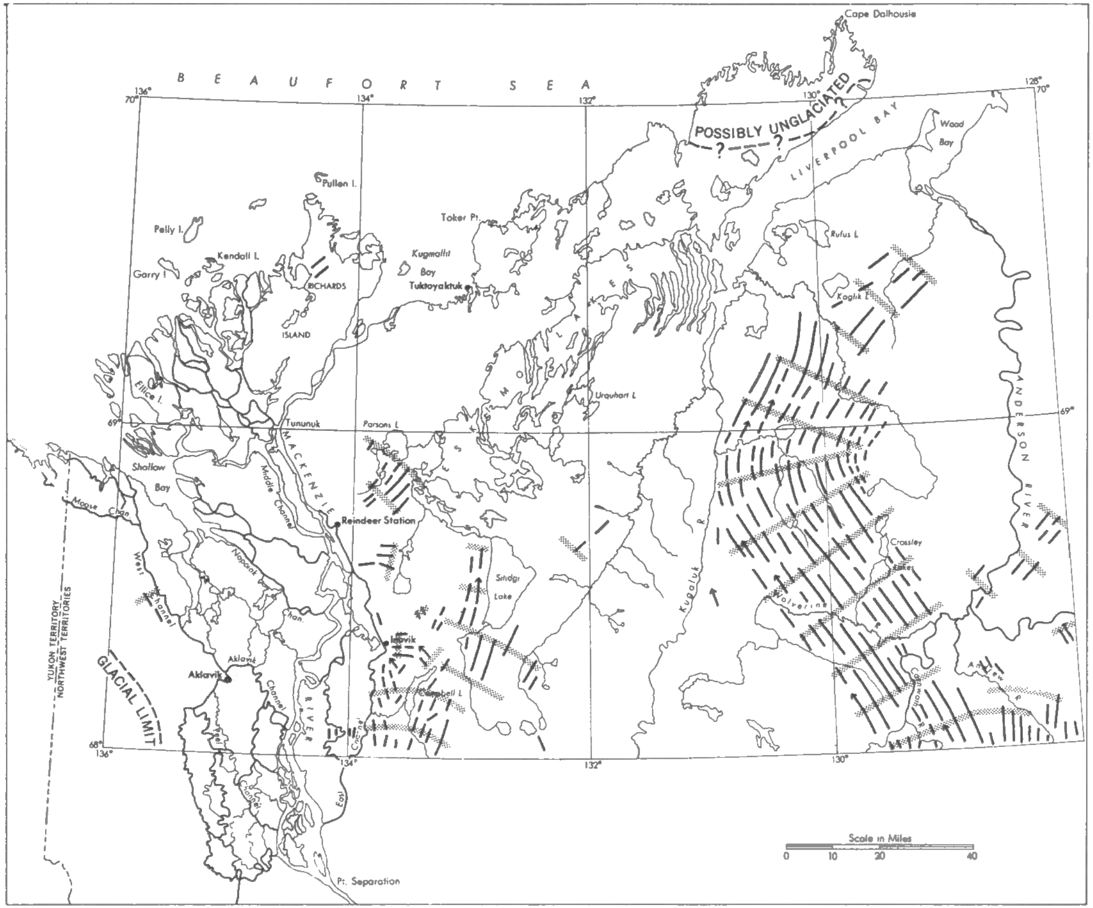

Figure 3. Glacial lineation. The solid lines show the trend of glacial fluting; the arrows show direction as derived from crag-and-tail topography. The shaded bars are normal to the fluting. The limit of glaciation in the southwest is after Bostock (1948).

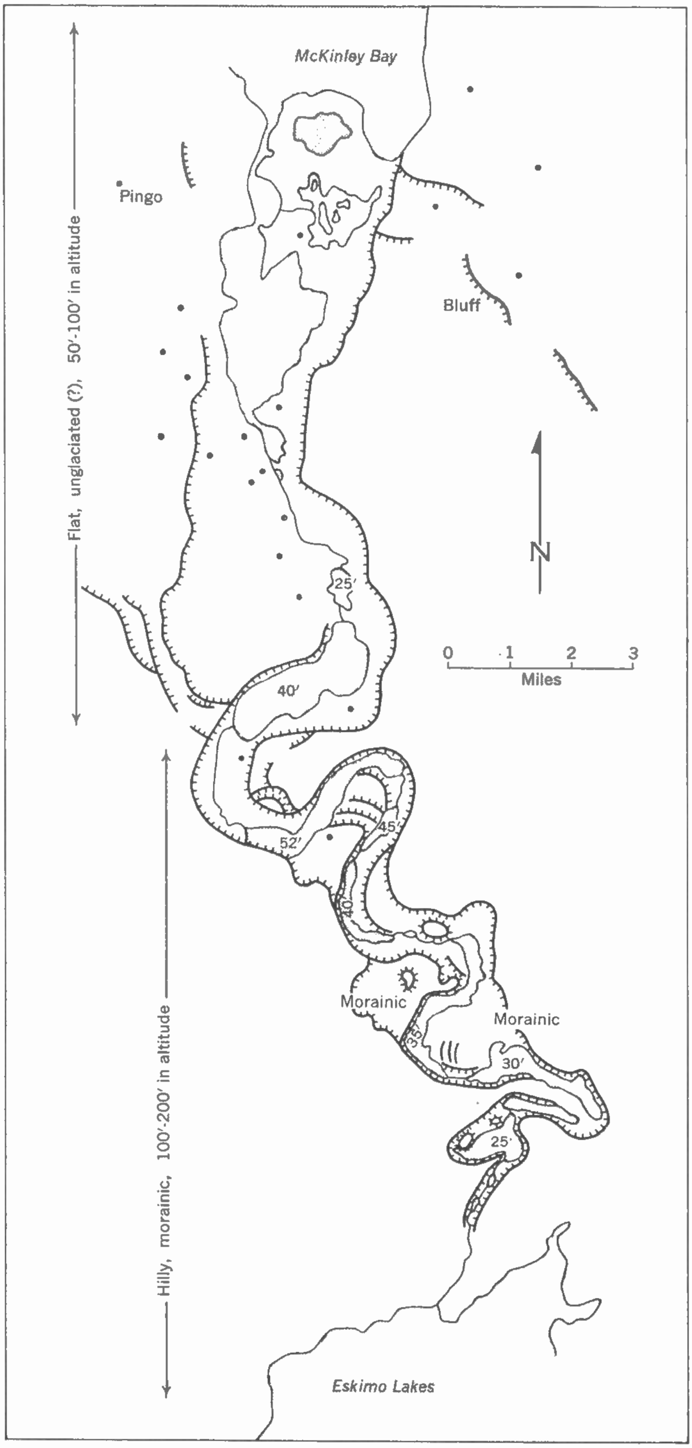

Figure 4. Abandoned glacially modified channel connecting the Eskimo Lakes with McKinley Bay.

winds. As the northern half of the channel is glacially unmodified, the limit of the last ice-advance is placed midway between the Eskimo Lakes and McKinley Bay. The relatively straight bluffs north of 69° 45′N may have been modified during postglacial submergence.

38. Glacial Lineation and Ice Movement: Glacial fluting, drumlinized topography, crag-and-tail ridges, and striae are confined to two principal belts, one on the east side of the Mackenzie Delta, the other to the west of Anderson River (Figure 3). As those features which are preserved were probably formed during the closing phases of deglaciation, they may be used as indicators of the pattern of late glacial ice movement.

Ice moved down the Mackenzie River valley, in a northwesterly direction, from near Arctic Red River to Point Separation. The ice was blocked on the west by the barrier of the Richardson Mountains. Erratics occur on Mount Goodenough to an elevation of about 3,000 feet (Camsell, 1906, p. 40; Bostock, 1948, pp. 37-38; Gabrielse, 1957, pp. 3-4) and at 2,000 feet on the top of Mount Gifford. Ice movement was north-northwesterly, along the front of the Richardson Mountains, and then northwesterly along the Arctic plateau and Arctic coastal plain to slightly beyond Herschel Island. Along the coastal plain west of the delta, the marginal movement of ice during the period of retreat was inland at right angles to the coast.

As the ice entered the apex of the Mackenzie Delta, movement was north as shown by striations and flutings on bedrock outcrops on either side of Kalinek Channel, about 20 miles south-southwest of Inuvik. Striations are still well preserved on some adjacent outcrops in several channels. Although the striations are found below the yearly flood stage, they have suffered little attrition from erosion by river sediment or abrasion by river ice.

The movement of ice near Inuvik was complex. Ice pushed along the Campbell-Sitidgi Lake depression, as shown by flutings and striations. However, at some unknown but probably late period, ice movement curved from the north end of Campbell Lake towards Inuvik as revealed by crag-and-tail topography and fluting in both bedrock and till. The sharp curve leads through a low saddle, 200 to 250 feet in altitude, with terrain several hundred feet higher to the north and south. In the rough terrain south of Inuvik, the bare rocky hill tops of the fault blocks are strongly lineated, whereas the lower drift-covered slopes are fluted. Minor differences in relief, of the order of several hundred feet, seem to have played important roles in controlling ice movement, thus suggesting a thin cover at the time of glacial fluting.

The fluting in the Anderson River area is signally well developed with individual lineaments often traceable for many miles. Along the upper Anderson River, east of 128° 00′ W, the movement was towards the northwest (Mackay, 1958a, p. 27). A splaying-out of the ice occurred along the Carnwath and Andrew Rivers. The fluting west of the Anderson River is highly suggestive of streamlined flow swinging through an arc. To illustrate, the lines which have been drawn in Figure 3 normal to the flutings resemble successive positions of a moving wave front. At about 68° 55′ N, the ice-front lines curve rapidly, as if pivoting around a centre north of Crossley Lakes. The pivot centre is an area of confused lake-strewn topography, the fluting gradually disappearing towards the centre. The swing in direction can hardly be related to local topography, because of very minor differences in relief. Most of the fluted region lies between 300 to 800 feet altitude, there being some association between flatness and perfection of fluting (Figure 38). Perhaps the curving was associated with the spreading-out of ice to the west as it cleared the topographic barrier of the Richardson Mountains.

39. Ground Moraine: The recognition of till is often very difficult in certain areas which are known to have been glaciated from other evidence. In some rocky areas, such as the rocky hills south of Inuvik, the only remnant of any initial till cover may be a few scattered erratics. In other areas, frost heaving and soil movement have so churned the surface deposits that till and underlying material have been thoroughly intermixed. In the Eskimo Lakes-Tuktoyaktuk Peninsula-Richards Island areas, numerous cleanly sectioned bluffs were examined. Although the terrain is known to have been glaciated, and the underlying Pleistocene deposits are predominantly of sand, only in a small percentage of the exposures could till be found over the sands. The upper several feet usually contain a higher percentage of fines and a few stones, but where stones are lacking, the field identification of till is hazardous. The preceding remarks also apply to other areas studied. In no region were thick deposits of till seen.

40. End Moraines: Few end moraines have been identified. Several end moraines occur in the Campbell-Sitidgi Lake area and to the southeast. They are minor features, up to 3 miles long, usually with slightly steeper southerly ice contact faces. A series of small, closely spaced ridges between East Channel and the southern end of Campbell Lake may be annual moraines.

Morainic topography, believed to reflect terminal ice conditions, forms an irregular belt, 5 to 10 miles wide, on the north side of the Eskimo Lakes (*130*). The morainic topography is bordered to the north by patches of pitted outwash. The topography is hilly. Numerous sand and gravel hillocks, being windblown and drier than adjacent areas, show up as light spots on air photographs. Altitudes range from 100 to 200 feet above sea level. The lakes in the morainic belt are irregular in shape and surface drainage is poor. Although the topography is morainic, the amount of glacial drift is believed to be small (*see*: Downie and others, 1953).

41. Eskers: In comparison with many other glaciated areas in the Northwest Territories, eskers are sparse. The sparseness of eskers may be a partial reflection of the paucity of coarse granular materials suitable for esker formation. The largest and longest esker lies midway between Urquhart Lake and the Eskimo Lakes. The esker, which has both single and multiple ridges, reaches a maximum height of 170 feet; with gaps included, it is 20 miles long. Scattered short single-crested eskers up to 5 miles in length occupy the low broad Campbell-Sitidgi Lake depression. These are believed to have formed during ice retreat when a lobe lay in the depression (Figure 9). Some eskers grade into kames or kame terraces. Several complex groups of subparallel ridges trending northeast-southwest across Richards Island may be eskers. The individual ridges are 5 to 50 feet high. Their material, to a depth of several feet, ranges from sand to gravel. Although Shield pebbles are present, they comprise less than one percent of the total. Some of the eskers merge into outwash plains; portions shown on Figure 5 as eskers may have been laid down subaerially as crevasse fillings.

42. Outwash and Kame Deposits: Outwash and kame deposits are widely distributed, but are most numerous where glacial drainage was channeled through large and small Pleistocene valleys (Figure 5). As the separation between outwash and kame deposits is often difficult, they have been combined in Figure 5. On Richards Island, glaciofluvial deposits (eskers, kames, and outwash) form a series of sand and gravel ridges and flat-topped uplands lying in a discontinuous belt to the north of Tununuk. Several small patches of outwash are on the east side of East Channel downstream from Tununuk.

At the time of glacier retreat, when a lobe filled the depression now occupied by the modern Mackenzie Delta, streams flowing off the glacier and from the Caribou Hills built kames against the ice, spreading outwash deposits beyond. Such deposits are found on the east side of East Channel between Reindeer Station and Tununuk. The upper sands and gravels of Inuvik (*24*) and those a few miles downstream were laid down, in part, as a kame complex between ice to the west and the higher land to the east.

43. A large expanse of pitted outwash was deposited in a proglacial lake to the north of Sitidgi Lake. At the time of deposition, ice lay to the west in the Mackenzie Delta and to the south, thus blocking southward drainage (Figure 9, line B). The outwash deposits are of reddish-brown sands with some gravel. They are not much over 5 to 10 feet in thickness. As the underlying deposits are contrasting grey sands, the outwash sands stand out as prominent reddish-brown capping beds. Initial deposition of the sands was into a body of water standing slightly over 150 feet in altitude, as flat-topped outwash plains occur at about 155 feet. In order for deposition to be possible, ice must have occupied much of the present water area, because some drainage was apparently over ice. Two sources of sediment can be recognized. Firstly, three large postglacial melt-water channels, originating in the higher land lying between Reindeer Station and Inuvik, fed their waters into the outwash complex. The lower portions of the channels merge into outwash aprons at 150 to 200 feet in altitude. Secondly, the outwash area lying north-northeast of Sitidgi Lake was also nourished by northerly flow from a channel which can be traced upstream nearly to Sitidgi Lake. Although extensive flat areas of outwash occur, much is pitted.

44. Outwash and kame deposits occupy the valley bottoms and sides of large Pleistocene channels of the ancestral Anderson River system. One large channel links the upper Kugaluk (at 68° 00′N) via Hyndman Lake with the Wolverine River (Figure 63). The large lakes south of, and including, Hyndman Lake are glacially modified channels blocked by drift and glaciofluvial deposits. The amount of disturbance and channel plugging can be shown by lake altitudes along the upper Wolverine River; in a channel distance of 35 miles, altitudes

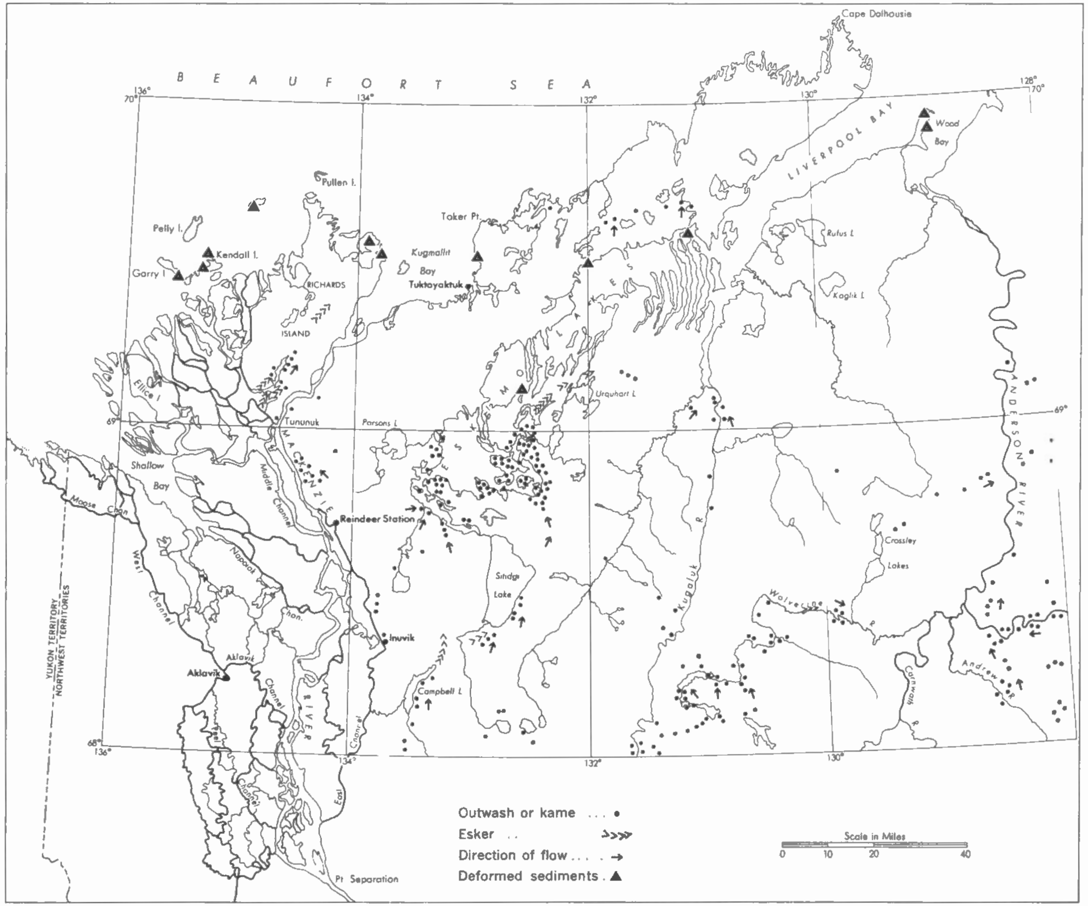

Figure 5. Glacial features: outwash or kames; eskers; direction of flow in drainage channels; locations of deformed sediments.

decrease from 689, 539, 528, 475, to 359 feet. The drop of 350 feet represents an average descent of 10 feet to the mile, a figure greatly in excess of any reasonable river gradient for the Pleistocene river. A few kames were built where melt-water streams flowed into ice-blocked portions of the Pleistocene valley.

Small patches of outwash, kames, and eskers lie on the flat upland into which Kugaluk River has incised its channel, south of 69° 00′N. The presence of glaciofluvial deposits shows that the Kugaluk River in this stretch is of post-glacial age. Pitted outwash and kames are present along Andrew River, particularly on the right side. The outwash deposits are about 350 feet in altitude, the kamey areas 400 to 450 feet.

The Anderson River, above the forks, is at an altitude of slightly over 100 feet and terraces rise 200 feet higher to about 300 feet. Some of the terraces contain steep-sided kettle lakes. As the lake bottoms lie well below the flat terrace levels, and as continuous flow marks can be observed on the terrace flats on both sides of the depressions, it is clear that ice occupied the depressions while terrace materials were first deposited and then eroded. The present Anderson River channel at the forks is bypassed by a Pleistocene channel, at an altitude of 300 feet (i.e., some 200 feet above the present river) about 5 miles east of the forks (*147*). Glaciofluvial deposits fill part of the bottom of the Pleistocene channel and kames occur on its sides. Some tributary streams, entering Anderson River downstream from the forks, flow partly in oversized Pleistocene valleys containing some glaciofluvial deposits.

45. Deformation by Glacier Ice: Pleistocene sediments, deformed by glacier ice-thrust, have been observed in coastal exposures from Herschel Island on the west to Nicholson Peninsula on the east (Mackay, 1956b; 1959). Ice-thrusting may have been facilitated by the development of high neutral stresses in pore water which would have weakened the shearing strength of the soil at the base of a permafrost (?) layer beneath glacier ice (Mackay, 1959, p. 20; Mathews and Mackay, 1960). The deformed beds (Figure 6) are recognizable in the field by slickensided shear planes, titled beds, and folds. On air photographs, lineation resulting from titled beds may show as subparallel patterns of ridges, hydrography, and gully erosion. Of course, not all dipping beds have been glacially

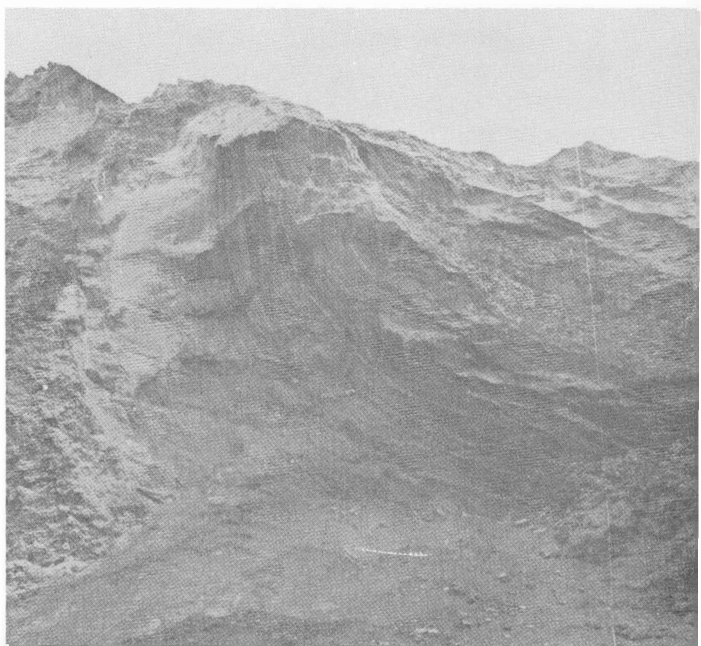

Figure 6

Deformed beds on the west side of Nicholson Peninsula, July 16, 1955. (Geog. Br. JRM-55-2-10).

deformed; there is foreset bedding, ground ice disturbance of bedding, and some beds are tilted by slumping.

On the southeast side of Garry Island, shear planes cut across the bedding planes of a greyish silt. The shear planes dip 25° to 30° to the southwest, at right angles to the faint lineation on the island surface. When Franklin visited Garry Island in 1825, he observed highly inclined beds. Lineation typical of ice-thrust features is present in the Kidluit Bay area and on Hooper and Kendall Islands, the latter also having slickensided silt. Deformed and tilted sands occur north of Topkak Point, midway between Toker Point and Tuktoyaktuk. Tilted beds and shear planes are also found in the Eskimo Lakes.

Deformed beds are abundant in Nicholson Peninsula. When the peninsula was studied in the summer of 1955, tilted beds were very much in evidence along fresh coastal exposures. When the same area was revisited in 1960, extensive slumping had obscured most of the stratigraphic succession and few of the former tilted beds were seen, although there were several new exposures. On the east side of Nicholson Peninsula numerous closely spaced nearly horizontal shear planes are crossed by two sets of slickensides, one trending N 63° E, the other N 118° E. The first set of slickensides has the finest striations, many being needle-like in sharpness. The second less distinct curving set is on a metallic, remolded smooth surface. The bluffs, with the shear planes, are hard-packed. They are relatively resistant to cliff recession and stand in vertical cliffs.

The beds which have proven most susceptible to shearing are in fine-grained sediments. Three samples have been tested for grain size distribution, one from Nicholson Peninsula and two from deformed beds west of the Mackenzie Delta at Stokes Point and Herschel Island. The grain size distributions in per cent are:

	Fine Sand .25 to 0.1mm	Very fine sand 0.1 to 0.05mm	Silt 0.05 to 0.005mm	Clay <.005mm
Nicholson Peninsula	52	22	26	0
Stokes Point		2	40	58
Herschel Island		1	23	76

The sediments from which the specimens were obtained have closely spaced shear planes, often less than two inches apart.

46. *Glacial Drainage Channels:* Glacial drainage channels are widely distributed, but their classification into sub-glacial, sub-lateral and overflow channels, etc., cannot be made without more field study. Many abandoned channels which resemble glacial drainage channels can be shown to result from stream capture in easily eroded sediments, so that identification of abandoned channels is often difficult. Moreover, as the natural flow of the major rivers (e.g., the Kugaluk and Anderson) is towards the north, they served initially as glacial drainage channels.

There is good evidence to suggest that glacial drainage was carried by East Channel downstream from Tununuk (Figure 7). The evidence is based on high-level meander scars, more than 50 feet above sea level, between Tununuk and the sea but with no matching high-level channel upstream from Tununuk. The

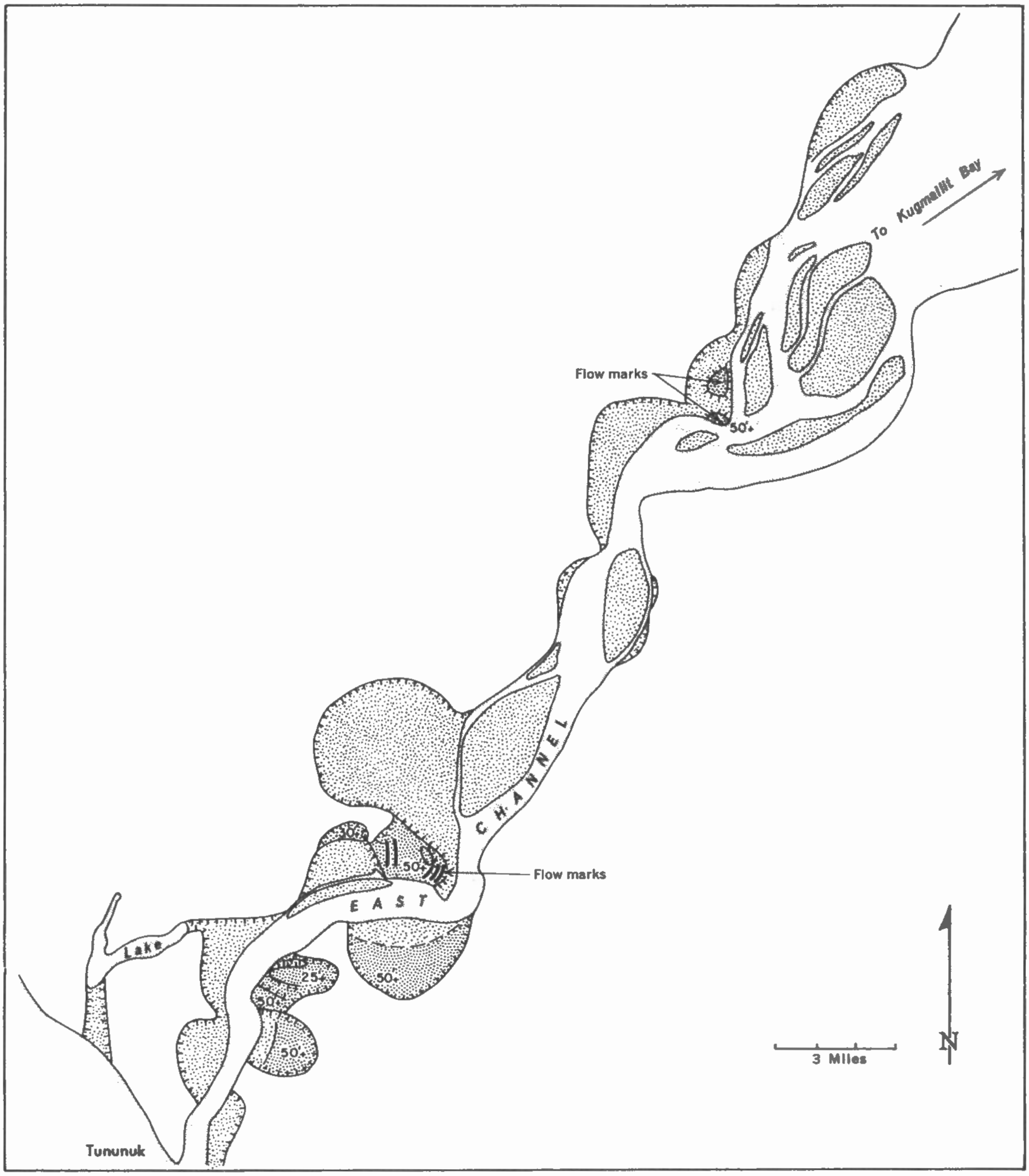

Figure 7. East Channel between Tununuk and Kugmallit Bay. The high-level meander scars, in the darkest stippled pattern, have smaller radii of curvature than the present channel.

missing upstream channel may have been removed by erosion, differential tilting, or glacial recession. It seems unlikely that the channel has been removed by erosion, because the modern Mackenzie Delta between Tununuk and Point Separation, 100 miles to the south, is 40 miles wide and all the sediments above sea level and in places to a depth of over 150 feet are of modern alluvium. Removal by postglacial erosion would, therefore, have involved sediments up to 150 feet thick, 100 miles long and 40 miles wide. Differential tilting would have to be unusually high to remove the channel. A third alternative is that the channel lay partially or wholly on a lobe of glacier ice lying in the Mackenzie Delta trough. The high-level meander scars are much smaller than those of East Channel, and, therefore, the flow was doubtless smaller. The mean radius of curvature of the high-level meander scars is about 3,700 feet, as measured to the outer bank. If the ratio between mean meander radius and width was roughly the same in the past as for present channels, the width of the high-level river was about 800 to 1,200 feet, or half the present width of East Channel in the same area. The presence of flow marks at an altitude of 50 feet within a few miles of Kugmallit Bay is a good indication that, at the time of flow, sea level was also higher.

On the east side of East Channel, midway between Reindeer Station and Tununuk, a melt-water channel at 250 feet is matched by patches of outwash at the same altitude. Melt-water streams enlarged or cut broad valleys in the Caribou Hills, particularly east of Reindeer Station. As the channels commenced at altitudes of 250 to 400 feet and discharged into the Eskimo Lakes area to build outwash plains at 150 to 200 feet, the eastern side of the Caribou Hills was apparently ice-free before the western side. The highest drainage channel is to the north, the lowest to the south, so that a southerly ice front retreat may be assumed.

In the Caribou Hills, north of Reindeer Station, a series of elongated lakes are spaced through a distance of 25 miles (from 68° 40′ N, 133° 50′ W to 68° 55′ N, 134° 10′ W) of which 16 miles are in lakes. The lake altitudes decrease from 375 feet in the south to 175 feet in the north, thus suggesting an old channel partially blocked by morainic and glaciofluvial material. The lakes lie along a rise in the Caribou Hills and so may represent an old melt-water channel or, possibly, an interglacial East Channel, as the northern portion is confluent with the Pleistocene deposits.

An 8-mile-long channel, 50 to 100 feet deep, trends due east from Inuvik (Figure 73). Boot Creek, which enters the delta at Inuvik, flows in the channel, except for the last mile. The highest point or saddle of the channel is at 250 feet. Through flow, irrespective of direction, could only have occurred if ice blocked the Inuvik-Campbell Lake area to the south. The channel may have served as a drainage channel.

On the west side of the Mackenzie Delta, several long drainage channels trend along the eastern slope of the Richardson Mountains and Arctic Plateau. They were probably eroded when a lobe of ice lay to the east in the Mackenzie Delta. The longest channel is now occupied by two rivers flowing north and south, respectively, the divide being at 300 feet.

In the area between Sitidgi Lake and Anderson River, there are numerous abandoned channels. Some of the abandoned channels once carried glacial drainage, but many have been abandoned as a consequence of postglacial drainage adjustments. Portions of the abandoned channels are dammed by slumps, alluvial fans, or deltas to form chains of long sinuous lakes.

Near the mouth of Kugaluk River, gravels rest on bedrock at an altitude of 50 feet and river terraces rise up to 70 feet above sea level. The gravels are interpreted as pitted outwash and kame terrace deposits. If this interpretation is correct, melt-waters were discharged through the estuarine mouth when sea level was higher than at present. The Kugaluk River between 68° 15′ N and 69° 00′ N is considered mainly postglacial in age and not a glacial drainage channel (*see:* Glacial Map of Canada, 1958). The trend of a number of abandoned channels (e.g., at 68° 30′ N and 68° 47′ N) is across the Kugaluk valley, showing that the Kugaluk River was established postglacially as a result of drainage adjustments. Moreover, the present river fits its valley, which is not appreciably oversized. The headwaters of the Kugaluk River join a series of large Pleistocene valleys partially infilled with glaciofluvial deposits. The highest melt-water channels reach an altitude of 800 feet, but most of the flow was at an altitude of 450 feet when melt-waters escaped northwards down the Kugaluk River.

The Wolverine River, which flows in a Pleistocene valley, carried varying amounts of melt-water, but there is no evidence for a large through discharge. The Carnwath and Andrew Rivers are bordered by flights of terraces, some of which were probably associated with melt-water discharge.

47. Proglacial Lakes: Proglacial lakes were uncommon during deglaciation, because the regional slope of the land was to the north, in the natural direction of drainage. In a few areas, proglacial lakes were impounded by higher land, or else ice blocked drainage outlets. When ice occupied the Mackenzie Delta trough, some lakes were probably impounded in high-level valleys on the east slope of the Richardson Mountains. Such lakes would have fluctuated in size during glacier advance and retreat, the evidence for them being very poorly preserved, although high-level overflow channels and deltas are present to the west along the Yukon coast. Some lakes were impounded between the retreating ice and the Caribou Hills on the east, as shown by outwash, dead ice terrain, and melt-water channels (*42*).

The largest proglacial lakes were formed in the southwestern part of the Eskimo Lakes. The highest beach-like ridges are at an altitude of 190 to 210 feet (Figure 8). If the gravel ridges are beaches of a proglacial lake, then an ice dam lay to the east (line A in Figure 9), otherwise the waters would have drained northwards through lower outlets by Parsons Lake or to the northeast. Cut bluffs, strandlines, and wavebuilt terraces are found at all levels below 150 feet. Cut bluffs, some traceable with hardly a break for many miles, are the most typical remnants of former shorelines. However, downhill creep has subdued the sharpness of detail along some bluffs, so that precise water levels cannot be measured. Extensive pitted outwash deposits are present in the area due north and northwest of Sitidgi Lake (Figure 5) at altitudes of about 150 feet. As

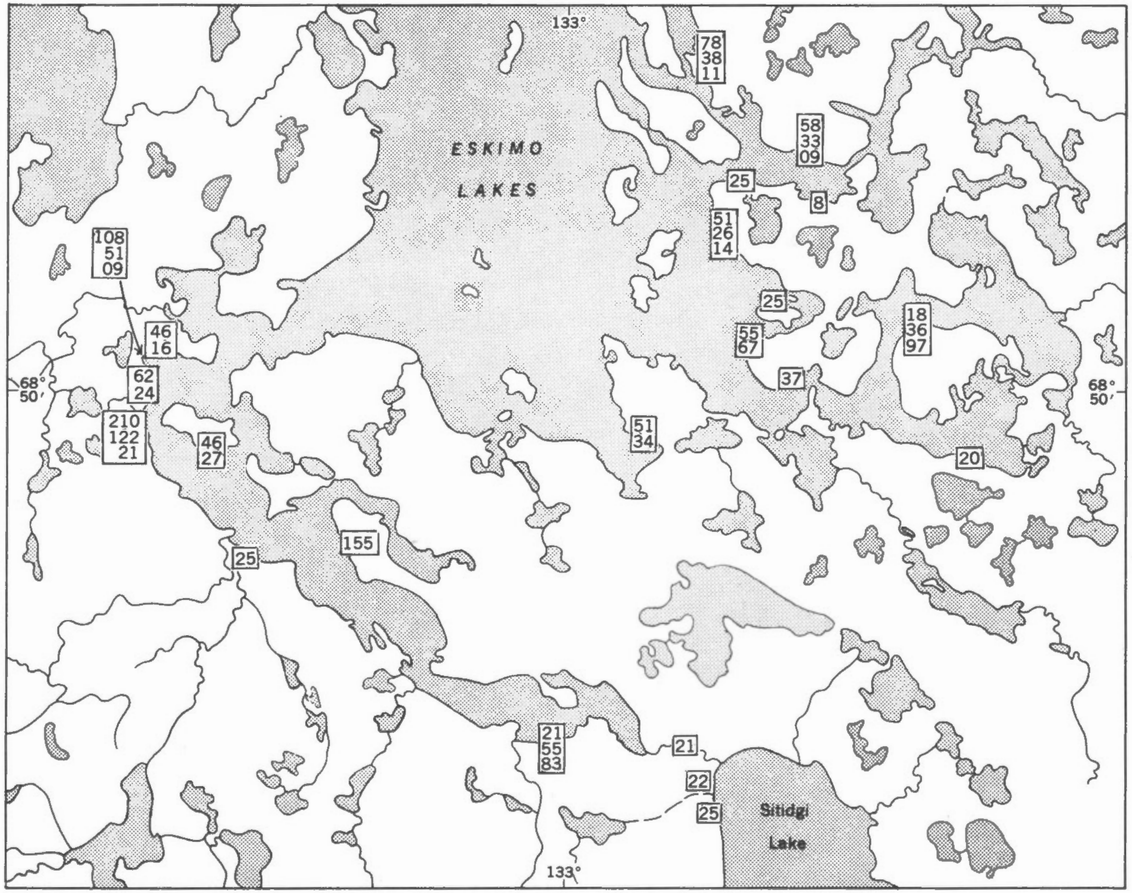

Figure 8. Old water levels, mainly of proglacial lakes, in the southern Eskimo Lakes-Sitidgi Lake area. Altitudes are in feet above sea level.

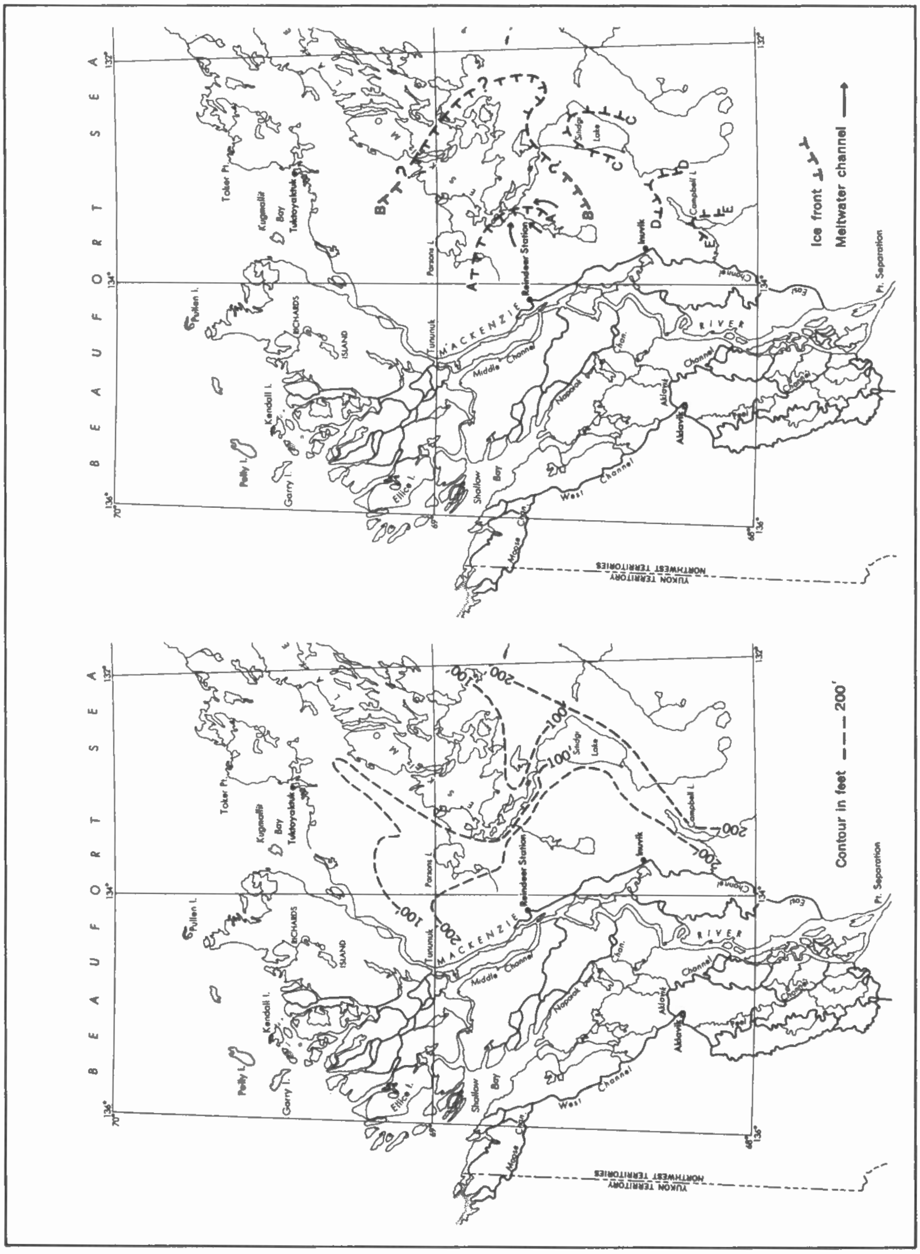

Figure 9. Controlling altitudes of proglacial lakes and inferred positions of ice fronts.

the source of some of the outwash sediment was from the Caribou Hills melt-water channels, glacier ice probably lay to the west (in the Mackenzie Delta trough), to the south, and east (Figure 9, line B), acting both as a source of melt-water and as an impounding dam. As the ice front withdrew, uncovering successively lower outlets, the levels of the lakes dropped accordingly. A fairly extensive proglacial lake level is represented by cut bluffs at about 55 feet. The lake was apparently dammed on the south by ice in Sitidgi Lake, because matching bluffs are absent there. The eastward extent of the water body is unknown. There may have been an ice barrier across the Eskimo Lakes or else one or more of the finger-like peninsulas may have extended across to block drainage (Figure 62). If one or more of the finger-like peninsulas were joined across their transverse channels, while maintaining their approximate heights, a 55-foot rise in lake level would flood the two sets of fingers south of Tuktoyaktuk, but possibly not the set of peninsulas at the mouth of Kugaluk River, assuming no major differential tilting. On the other hand, it is evident that sea waters might have entered the Eskimo Lakes through Liverpool Bay if uncovering by ice coincided with the inferred postglacial submergence. With further retreat of the ice front, a lobe occupied Sitidgi Lake (Figure 9, line C), as shown by kamey deposits laid down between the sloping ice front and the eastern hillside of Sitidgi Lake. If this interpretation is correct, the slope of the ice front (against the hillside of Sitidgi Lake) was about 150 feet to the mile. A moraine with ice contact face to the south (Figure 9, line D) shows the position of the ice between Sitidgi and Campbell Lake. A lobe later occupied Campbell Lake (Figure 9, line E), as shown by kames on the east side. The kamey areas of Sitidgi and Campbell Lakes are on the east sides, which may mean that the ice did not form lobes in the Sitidgi-Campbell Lake depression but maintained a northeast-southwest trending front.

When the ice had melted back sufficiently to uncover the southwestern end of Campbell Lake, waters were free to drain into the Mackenzie Delta area, when it was ice-free, over the Sitidgi-Campbell Lake threshold at about 30 feet above present sea level (Figure 10). For a time, this outlet may have governed water levels in the Eskimo Lakes, as a 30- to 35-foot water level is very extensive.

The most widespread old lake bluff is about 25 feet above present lake level. In the south, the bluff is from 20 to 25 feet above the level of Sitidgi Lake which, in turn, is about 5 feet above sea level. The bluff nearly encircles Sitidgi Lake, except in the southwest where it disappears in the ill-drained flats between Sitidgi and Campbell Lakes. The bluffs are continuous on both sides of the creek flowing from Sitidgi Lake into the Eskimo Lakes, so the two lake systems were once continuous. A prominent bluff at 22 to 25 feet above lake level occurs in the Eskimo Lake centred on 69° 10′N, 132° 40′W. The north-south group of finger-peninsulas, centred on 69° 13′N, 132° 27′W, are crossed by bluffs at about 22 feet above lake level. Bluffs at about 25 feet above lake level are present in the Eskimo Lake due east of Tuktoyaktuk. Therefore, Sitidgi Lake was formerly part of an enlarged lake system when the water level was about 20 to 25 feet above the present. There is no available evidence either for, or against, a sea level con-

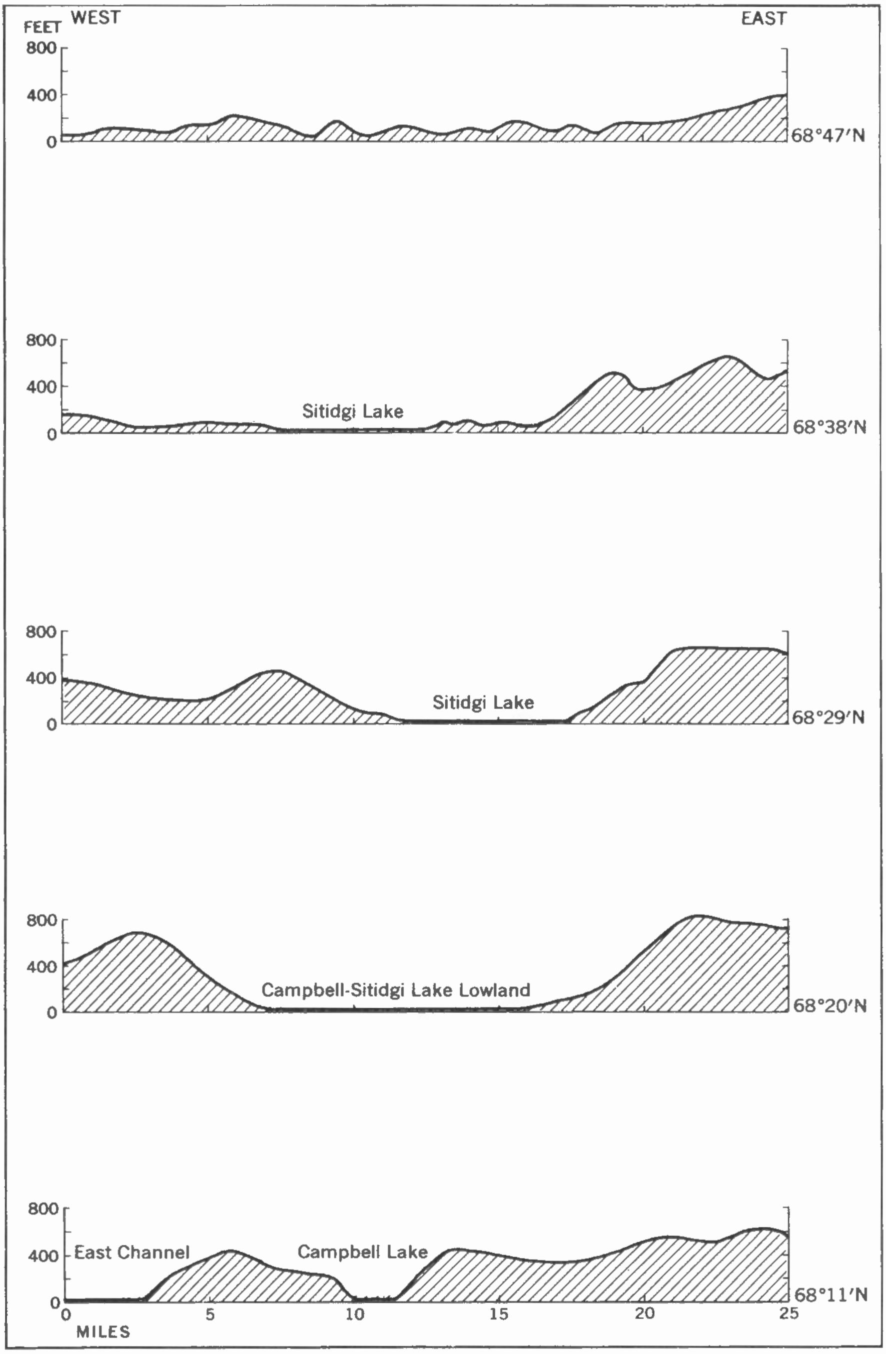

Figure 10. Profiles across the Pleistocene (?) channel connecting the ancestral Mackenzie River via Campbell and Sitidgi Lakes to the Eskimo Lakes. The top profile, along 68°47'N, is between 132°02'W and 133°02'W.

nection of the lake with Liverpool Bay as at present. There has certainly been considerable postglacial erosion of the channels, particularly in places where the width is less than half a mile. "Thumb Island" (the island lying 4 miles southwest of Campbell Island) was formerly joined to the northern coast, judging by the geometric trajectory of terraces on mainland and island. "Thumb Island" may also have been joined to the peninsula on the west side of Kugaluk estuary which is crossed by an abandoned channel two miles long and 500 feet wide. The channel is at an altitude of about 23 feet above sea level, and it is continuous with bluffs and terraces of Kugaluk estuary at the same altitude. Therefore, the 25-foot bluff of the Eskimo Lakes may have been formed when sea level was 25 feet above the present. At this stage, proglacial conditions were then definitely ended.

Postglacial Changes in Sea Level

48. Submergence: There is good evidence for postglacial submergence, emergence, and then possibly submergence again in the Mackenzie Delta area. However, the coast lacks the striking flights of raised beaches which have come to be associated with emergence in the Eastern Arctic. The reasons are several: the Mackenzie coastal area is low, the sediments are unconsolidated, coastal recession is rapid, and postglacial submergence was minor, so that conditions favoring the formation and preservation of beaches have been poor.

The principal evidence for submergence at the time of deglaciation comes from estuarine river terraces. On the east side of Liverpool Bay, the estuarine part of Harrowby Bay is bordered by three prominent terraces at altitudes of about 25, 60, and 100 feet above sea level (Mackay, 1957a, p. 7; 1958a, p. 43). The two lowest terraces are of sand and gravel; the uppermost terrace is partially cut into soft grey shales. The terraces are very flat. The 100-foot terrace can be followed for at least 20 miles on both the north and south sides of Harrowby Bay. There is the possibility that the Harrowby Bay area escaped glaciation during the last ice advance, so the highest terraces, at least, may have been formed when adjacent areas were under ice. The lowest terraces, however, lie within the altitudinal range of terraces in nearby glaciated areas. The Harrowby Bay terraces are believed to have been formed slightly above sea level in an estuary opening to the sea.

At Stanton, a terrace at 75 feet may represent an old water level. On the east side of Wood Bay, south of Stanton, a cut bluff stands about 40 feet in altitude. In the estuarine portion of the Anderson valley, for a distance of 20 miles upstream from its mouth, pronounced terraces with flow marks occur up to an altitude of 75 feet above sea level. The proximity of terraces and erosional channels near to Wood Bay suggests that sea level was higher when they were formed.

The estuarine portion of Kugaluk River, north from 69° 15′ N, has sandy terraces extending for several miles at altitudes of 20 to 40 feet. The terraces are matched by cut bluffs. As the estuarine portion is from 3 to 5 miles wide, the terraces are up to a mile wide, and the terraces consist of sands, deposition probably occurred fairly close to sea level. However, a radiocarbon date for a piece of wood at a depth of 10 feet from the top of a 40-foot bluff is greater than 38,000 years (I-482; Geographical Branch JRM-60-W1). No pebbles or till were found on top of the bluff, which was of grey crossbedded sand with wood chips, scattered Shield-type pebbles, and a few Cretaceous rocks carried from the south. If the wood reflects the age of the sediments, and has not been derived from older deposits, then the terrace material would be of an interglacial rather than postglacial age. Although the terrace material might be of interglacial age, the form of the terrace, the absence of a cover of till, and its youthfulness all suggest that it has been modified at least during a period of postglacial submergence.

Terraces along East Channel have already been discussed (46). In the Kittigazuit area, extensive flat-topped terraces at 40 to 50 feet match cut bluffs at 50 to 55 feet above sea level.

There is no evidence, west of the Mackenzie Delta, for a submergence of the order of 100 to 200 feet. Samples of soil from the top of a hill at Shingle Point (altitude of 195 feet), King Point (altitude of 180 feet), and Kay Point (altitude of 200 feet) were all barren when examined for microfossils (analyses by Dr. F. J. E. Wagner, Geological Survey of Canada). In the delta area, the following soil samples were all barren: sample from 25 miles north-northwest of Aklavik at an altitude of 123 feet; sample from the top of Hendrickson Island (altitude of 10 feet); from a grey sandy silt (altitude of 75 feet) beneath about 20 feet of gravel at Inuvik; and sample from the Eskimo Lakes (68° 57′ N, 132° 52′ W) at 15 feet in altitude.

The cumulative evidence from the terraces points to a relative submergence of 50 feet at the time when the coastal area became ice-free.

At Ibyuk pingo near Tuktoyaktuk a radiocarbon age of a sample of driftwood at 7 feet below the top of the pingo is 12,000±300 years (S-69; Müller, 1962). If the updomed pingo were to be returned to its original position, the driftwood would be 15 to 20 feet below present sea level (*27*). The driftwood was most likely deposited close to sea level of the time. A similar situation existed at 69° 32′ N, 130° 55′ W in the Eskimo Lakes where a wood fragment 15 feet from the top of a pingo has been dated at 10,800±300 years (I-483, Geographical Branch JRM-60-W2). If the pingo material were returned to its original position, it would lie 15 to 20 feet below present sea level, thus indicating the relative position of sea level at time of deposition, assuming, of course, the age of the wood to reflect the time of deposition. The two radiocarbon dates from the pingos agree in indicating a relative rise of the land of about 75 feet between deglaciation and 11,000 years ago. Continued uplift produced relative emergent conditions during which time streams cut their valleys which now appear drowned.

49. Emergence: The coastal area from Richards Island to Cape Dalhousie is a highly indented drowned shoreline. For example, the settlement of Tuktoyaktuk is at the mouth of a drowned valley (Figure 75). At the seacoast, the entrances to the harbor are shoal (about 15 to 25 feet) due to infilling caused by coastal erosion. However, as one proceeds up the estuary, depths rapidly increase to about 50 feet, and then gradually decrease at an average rate of 8 feet per mile. Former emergence of greater than 50 feet is indicated.

On the right side of East Channel, most of the streams draining the Caribou Hills terminate in elongated lakes which indicate drowning. Many long profiles of the streams, when projected, pass 20 to 50 feet below the level of East Channel. For instance, soundings along the first drowned lake, 5 miles southeast of Tununuk, show a depth of 35 feet some 2 miles from East Channel. The distal ends of such lakes are shoal from East Channel sedimentation; otherwise, the elongated lakes have bottom profiles continuous with the subaerial stream profiles.

Along the north side of Liverpool Bay, west of 130° 00′ W, several drowned valleys were sounded. All had shallow thresholds where the drowned valleys reached Liverpool Bay, but were deeper farther upstream. The deepest sounding was 94 feet. The Kugaluk estuary, where it is a mile wide, has depths over 30 feet. The digitate peninsulas at 69° 30′ N, 131° 00′ W, are separated by deep channels. The channels have long stretches from 100 to 150 feet deep. Although there is a fast tidal current through the channels, the depths are believed to be too great for tidal scour alone (*132*).

To the north of Tununuk, there are two irregular elongated lakes, a smaller southern lake, and a larger northern lake, locally called "Yaya Lake". At a distance of 14 miles due north of Tununuk, "Yaya Lake" has a depth of at least 130 feet. Inasmuch as narrow portions of the lake 10 miles north of Tununuk are only 15 to 20 feet deep, the extreme depths are not believed due entirely to postglacial erosion.

In a drill hole located 5 miles southwest of Inuvik, in the modern Mackenzie Delta, the succession of deposits from top to bottom is: 0 to 100 feet, thinly stratified sandy silt with layers of decomposed organic material throughout; 100 to 180 feet, fine to medium sand with thin layers of organic material in the upper 40 feet; 180 to 240 feet, very dense clay-sized soil, free of stones except for the last few feet (Johnston and Brown, 1961). The stony material at the bottom may be till; the uppermost 180 feet appears to be sediment deposited close to, or below, sea level. Freezing of the ground, with ice segregation, and also compaction, may alter the details of the altitudes of the vertical succession, but even allowing for this, the deposits suggest a former water level considerably below the present. A radiocarbon date of 6,900$\pm$110 (G.S.C.—54) has been obtained for organic material from 125.5 feet below the surface for bore hole number 3 (Johnston and Brown, 1961). As the ground surface is 15 to 20 feet above sea level, the specimen came from a depth of about 105 to 110 feet below present sea level. The depth of water in which the organic material was deposited can only be inferred, but, judging from conditions now prevailing, a depth of less than 50 feet seems reasonable.

50. Eustatic Rise: After the period of low sea level, relative to the land, eustatic rise in sea level caused a drowning of the stream valleys to result in the embayed coast from Richards Island to Cape Dalhousie, and the estuarine mouths of the Anderson and Kugaluk Rivers. The land, however, has remained at its present altitude, or above, relative to sea level for over 8,000 years, as shown by radiocarbon dates for two peat deposits. At 69° 12′ N, 132° 27′ W, wood was obtained from the bottom of a 7.5-foot peat deposit at an altitude of 3 feet above lake level. The underlying material was lake bottom sand with organic matter. A dating for the wood is 7,400±200 years (G.S.C.—16). As the peat shows no evidence of burial, sea level could hardly have stood higher than at present since the peat began to accumulate. At Inuvik, a natural exposure of peat occurs at Twin Lakes (Figure 73). A specimen of wood from the bottom of the 12-foot-thick deposit, which is about 20 to 25 feet above sea level, has been dated at 8,200±300 years (G.S.C.—25). Thus, the two radiocarbon dates for peat deposits close to sea level support the view that for the past 8,000 years sea level has been at, or below, its present position in relation to the land.

51. Submergence: The last episode may be a slight submergence of 10, 20 or more feet, although the evidence is speculative and inconclusive. For example, in the low alluvial islands north of Shallow Bay, sectioned pingos expose peaty beds uplifted from 10 to 20 feet; that is, from a depth of 5 to 15 feet below present sea level (Figures 35 and 36). The peaty beds are of the type now forming above sea level in some portions of the alluvial islands (*86*). If the below-sea-level position of the beds is not due to settlement or compaction, submergence is suggested. The south shore of Shallow Bay is being eroded back; although this may reflect a change in depositional environments, it may also reflect submergence. The radiocarbon date for the wood specimen from the bore hole in the Delta (*49*) suggests that sea level may have stood lower than at present when the wood was deposited about 6,900 years ago. As eustatic sea level was at about —50 feet compared with present sea level (Curray, 1961; Hopkins, 1959; McFarlan, 1961), there would be no submergence if the wood was deposited in water about 55 to 60 feet deep. Such a depth seems too great, and slight submergence may be indicated. However, too much reliance should not be placed on conclusions based on one radiocarbon date.

Sedimentation from the Mackenzie distributaries has resulted in shoal conditions offshore, and, therefore, in the infilling of any nearshore valleys which had been eroded during a period of emergence. The areas of active sedimentation are seaward of the distributary mouths west of Richards Island and in Kugmallit Bay. The water off the northwest side of the delta, seaward of the active distributary mouths north of Shallow Bay, is very shallow. Depths 20 miles offshore seldom exceed 20 feet. Depths north of Richards Island are greater, being 20 feet just north of Pullen Island. Kugmallit Bay itself is less than 20 feet deep, but north of Kugmallit Bay a slight valley is traceable to a depth of 200 feet, with a suggestion of smaller valleys north of Hutchison and McKinley Bays. These valleys may have been cut during the period of emergence. There is no distinct valley north of Liverpool Bay, but there is a pronounced seaward bulge of the submarine contours at depths of 100 to 200 feet. The bulge suggests sedimentation from Liverpool Bay flow. During a period of emergence, water from the Kugaluk, Anderson, and Horton (via Harrowby Bay) Rivers may have united into one large postglacial river to empty into the sea northwest of Cape Bathurst. The bulge of the submarine contours may be that of a delta. It might also mark the site of earlier delta buildings.

Submergence of the Mackenzie Delta and immediate vicinity might be caused partly by the load of sediment deposited in the delta (*see* Richards, 1950a; 1950b). The depth of delta sediments is unknown, but bore-holes 5 miles southwest of Inuvik reached bedrock at about 250 feet (Johnston and Brown, 1961). Channel depths exceeding 100 feet occur locally on West Channel, Middle Channel, and on an unnamed channel 7 miles west of Tununuk. As it seems most unlikely that the channels have been cut into bedrock, either the delta sediments are over 100 feet thick, or else the channels are cut into easily eroded sediments such as those of the Caribou Hills. There is the good probability, therefore, that the delta sediments are at least 100 to 250 feet thick over much of the delta. Whether such a load of sediment could cause downwarping is problematical.

In Figure 11 the generalized position of past sea level has been based upon Hopkins (1959) with a modification showing a still-stand for the last 5,000 years (Curray, 1961; McFarlan, 1961). The triangles show the inferred altitudes of zero bench marks, now at sea level, for the time period indicated. Point number 1 is for bore hole number 3 (Johnston and Brown, 1961) just southwest of Inuvik (*49*). On the assumption that the age of the specimen dated at 6,900 years reflects the time of deposition, then sea level was about 50 feet below that of the present time. As the specimen is from a depth of about 105 to 110 feet below present sea level, it would have been deposited in about 55 to 60 feet of water if the Mackenzie Delta has remained at its present level since then. If the depth of accumulation were appreciably less than 55 to 60 feet below sea level, then slight submergence of the delta may be inferred. Point 2 is for the base of the peat deposit (*50*) in the Eskimo Lakes and point 3 for that at Inuvik (*50*). As the peat deposits have been above sea level for about 7,400 and 8,200 years, respectively, present zero altitude bench marks would have been on the dotted lines in the past. Point 4 is for the pingo at the east end of the Eskimo Lakes (*48*) and point 5 is for Ibyuk pingo, for which there are two dates (*27*). Taking point 4, there are terraces in the area about 50 feet above sea level and drowned valleys of 100 to 150 feet. Therefore, if a bench mark now at sea level is followed back in time, it lay close to line A when drowned valleys were cut, then at point 4 when the wood was deposited, then on line B when the 50-foot terraces were formed. For Ibyuk pingo, point 5, nearly the same sequence holds true. When the drowned delta valleys, Tuktoyaktuk sound, etc. were being eroded, point 5 was well above sea level near line A; when the wood in the pingo was deposited, a present zero bench mark was just above sea level at point 5; and when the 50-foot terraces of East Channel were eroded, a present zero bench mark was on line B.

The assembled evidence, which, it should be stressed, is based on few radiocarbon dates, suggests that the Mackenzie Delta trough was at an altitude of −200 to −300 feet relative to present sea level prior to 10,000 or 12,000 years ago. The land rose faster than sea level during which period rivers incised deep valleys. The last episode may have been one of slight submergence, there

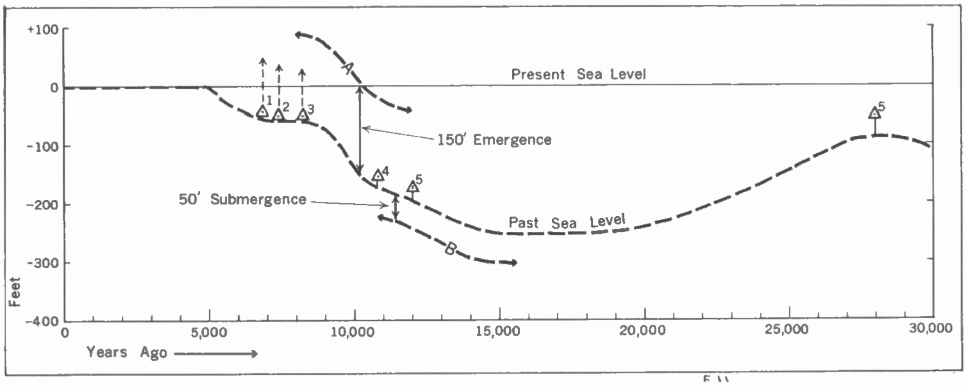

Figure 11. Graph of data concerning changes in sea level.

being the possibility of an actual sinking of the Mackenzie Delta trough and adjoining area. There is insufficient evidence to assess any effects of differential tilting, except to note that points 4 and 5 appear to lie on different uplift curves.

Fluvial Action

52. Stream erosion is rapid, when the paucity of precipitation is considered as a criterion. Although the total precipitation in the form of snow and rain is only about 10 inches a year (*151*), the precipitation effectiveness is concentrated into the four months of June to September. The greatest amount of stream erosion is accomplished by flood waters fed from melting snow and thawing soil in late May and early June. High water levels on the larger rivers such as the Mackenzie, Kugaluk, and Anderson may rise over 20 feet within a few days, especially at break-up. River ice is frequently shoved 5 to 10 feet above the highest water level and, like a bulldozer, it scrapes and shoves the ground beneath it. River boulders may become faceted and striated by the passage of rock-armored river ice (Mackay, 1958b, p. 44, Figure 2).

The eastern slopes of the Richardson Mountains are deeply gullied and crossed by incised rivers. For example, Donna (Willow) River leaves the Richardson Mountains in a postglacial gorge 250 feet deep. To the northwest, the Blow River and others occupy postglacial valleys over half a mile wide at the bottom and as much as 200 feet deep. The streams which drain the west slope of the Caribou Hills are mainly senile and aggrading, because of submergence and deposition in the Delta. The Kugaluk, Anderson, and other rivers have all cut deeply in parts of their courses.

Wind Action

53. The geomorphic effects of wind action are diverse indeed. Wind action has caused the orientation of lakes, built parabolic dunes, "ribbed" shallow lakes, and abraded rocks. As most of the region has a continuous vegetation cover, active dunes and drifting sands are atypical of the area, despite the prevalence of sandy sediments.

The direct and indirect effects of winds on lakes are many. The most unexpected is the orientation of lakes with their long axes transverse to the prevailing winds (*54*). At time of break-up, lake ice may be blown by winds to produce ice-shove heaps along the shores. If the heaps are of sand, they may be washed away by the first storm, but if of gravel, ridges may remain as semi-permanent features.

In hundreds of the large shallow lakes of the Mackenzie Delta, northwest winds have contributed to the formation of a ribbed pattern of northeast-southwest trending ridges (Figure 43). The sedgy ridges are minor features, a foot or so in height and several hundred to a thousand yards long. The ridges extend lakeward, from opposite northeast and southwest shores, like outstretched fingers of two hands. The northern limit of the ribbed lakes coincides closely with the limit of spruce and drop in floodplain level. The low outer islands and the southwest side of Shallow Bay have few or no ribbed lakes. The ribs, which are perpendicular to the prevailing wind direction, are believed to be formed primarily by wave action associated with a gradually falling summer water level, and not with ice-shove.

Cliff-head dunes are common along north-south trending sandy bluffs or where a sandy beach provides a source of sand. In the Kittigazuit area west of Tuktoyaktuk, some cliff-head dunes have been built by west winds. From Tuktoyaktuk to Cape Dalhousie, although dunes formed by west winds do occur, they are less numerous than dunes formed by east winds on the west sides of lakes (Figure 12). The dunes, whether built by east or west winds, trend almost exactly east-west. The source of sand is usually the bottom of a partially drained lake or a broad sandy beach. Lyme grass and a few willows tend to stabilize the sand and, consequently, haystack hillocks surrounded by blowout depressions are common.

Crescentic to hairpin dunes, with the convex end facing west and the limbs pointing east, are distributed throughout the northeastern part of Tuktoyaktuk Peninsula (Figures 12 and 13). The dunes are typically 300 to 600 yards long, 10 to 30 yards wide, and several yards high. Some dunes have the curved portion blown out, so that the two limbs are detached and angle in; if much is blown away, the limbs are subparallel. The steeper slopes are usually towards the west. In a few places, dunes overlap from westward migration. The dunes are probably parabolic dunes (Figure 61) formed by east winds, rather than barchanes built by west winds. The dune sands are poorly graded, with about 90 per cent fine sand and 10 per cent very fine sand. The grain size distribution differs little from that of the underlying sediments which, in general, are also poorly graded. The parabolic dunes are fixed dunes, bound down with a vegetation cover of lichens, avens, ground birch, heather, low willow, cranberry, sedges, and other plants.

The time of year when the loose sand is most mobile is unknown. As an opinion, unsupported by much field evidence, fair-weather summer easterly winds blow sand more readily than rain-bearing westerly winds which dampen the sand to hold it. The contribution of winter winds to sand drift can only be inferred, but natives have commented on the frequent darkening of snow from blown sand.

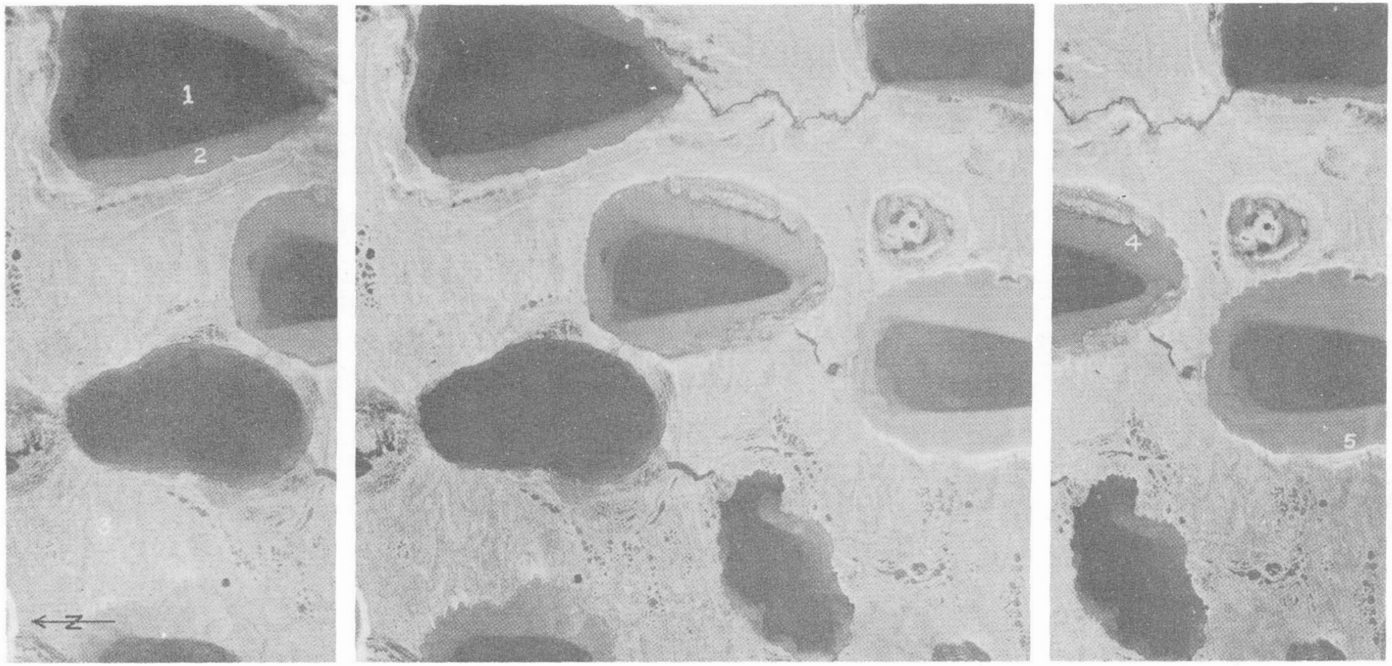

Figure 12. Stereo triplet of oriented lakes centred at 70°00′N, 130°25′W. The lakes are oriented north-south. 1—deep central trough; 2—sublacustrine bench; 3—crescentic (parabolic) dunes; 4—ripples on sublacustrine bench; 5—blowouts on west shore of lake. (RCAF A 12699—418 to 420).

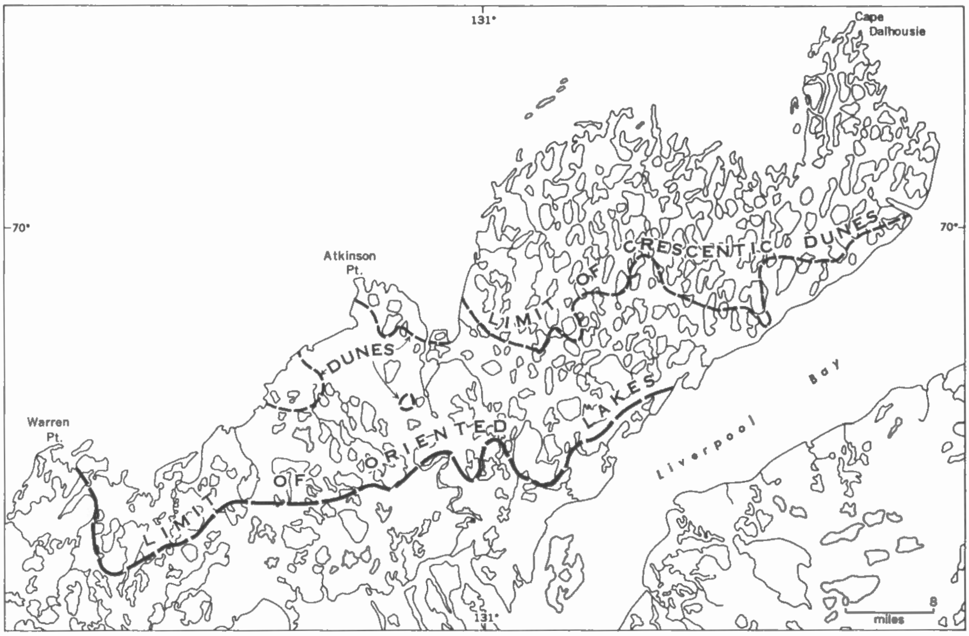

Figure 13. Limits of crescentic (parabolic) dunes and oriented lakes.

Wind abrasion has been observed in sandstones of the Arctic plateau foothills 25 miles northwest of Aklavik. As much as 4 inches of sandstone has been blasted away from around protruding pebbles or even a lichen-protected spot. Since there is no source of sand near the abraded rock and the surrounding snow is sand-free, the abrasive material may be wind-driven snow at low temperatures.

Oriented Lakes

54. Introduction: A large group of oriented lakes, with north-south-trending long axes and oval shapes, occurs at the eastern end of Tuktoyaktuk Peninsula, east of Warren Point (Figure 13). Geometric symmetry in lakes is so unusual that oriented lakes, wherever found, stir speculation as to their origin. The oriented lakes are scattered throughout a low coastal plain mostly below 50 feet in altitude (*129*). The southern limit of the oriented lakes is along the zone where the coastal plain merges into the higher land to the south. Only a few isolated oriented lakes lie beyond the limit shown on Figure 13. To the east of Cape Dalhousie, similar oriented lakes are present near Cape Bathurst (Mackay, 1956c; 1958a, pp. 74-81). All the oriented lakes are in areas of unconsolidated sand with a tundra vegetation. The lakes vary in length from 100 feet to several miles. The coalescence of small lakes may result in a segmented larger lake, or eventually in a symmetrical large lake.

55. Orientation: The axial trend of the lakes is close to true north-south, but there are slight local and areal variations. A sampling of 88 of the most symmetrical oriented lakes gave an average trend of N 06° E. The extreme axial orientations were N 09° W and N 25° E. Of the 88 lakes sampled, 27 were west of 130° 30′ W; 30 were between 130° 00′ W and 130° 30′ W; and 31 were east of 130° 00′ W. The mean axial trends and ranges were: western group, mean of N 02° E and range of N 09° W to N 08° E; central group, mean of N 05° E and range of N 01° W to N 11° E; and eastern group, mean of N 11° E and range of N 01° W to N 25° E. The data shows that there is a clockwise shift in axial trend with progression eastwards.

56. *Shape:* The oriented lakes have two "shapes", namely that of the shoreline which is typically egg-shaped, and that of the deep central portion which is triangular (Figure 12). In order to analyze the shapes of the lakes and the central aeeps, four basic shapes, which seemed to suit best the majority of lake shapes, were selected (Figure 14). The four shapes were the oval or egg shape (Figure 14A); the ellipse (Figure 14B); the "triangle" (Figure 14C); and the lemniscate

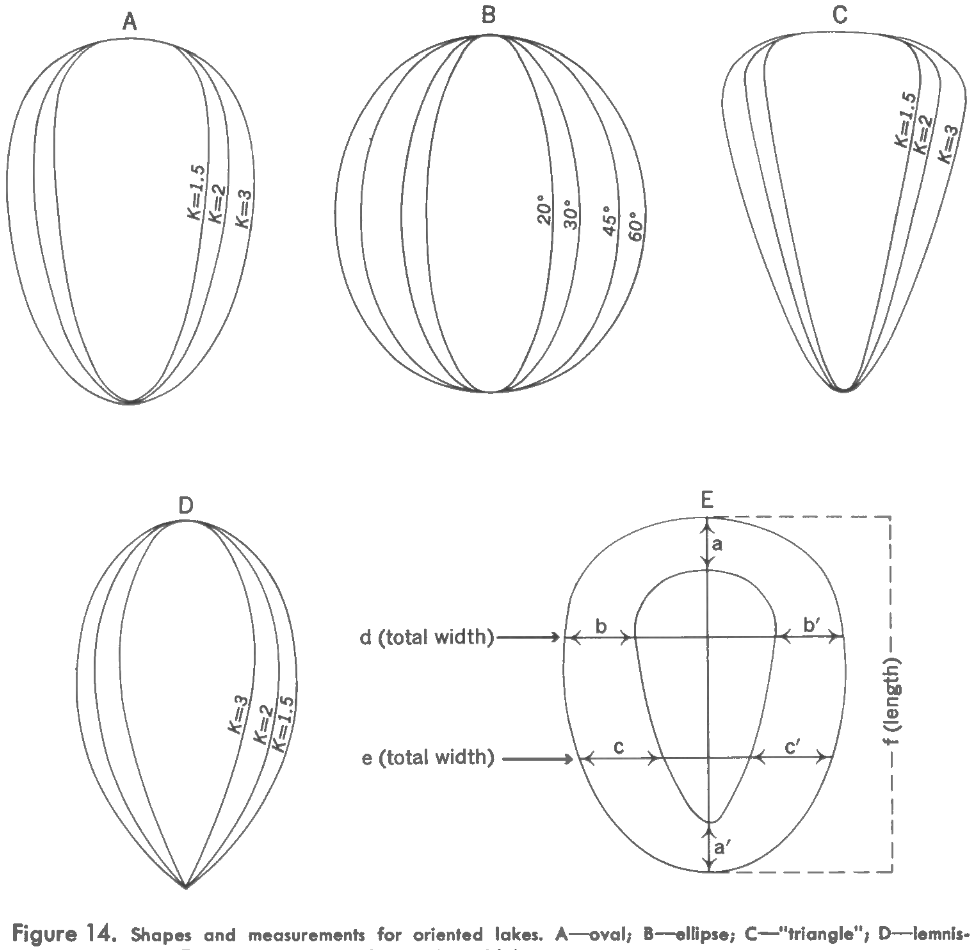

Figure 14. Shapes and measurements for oriented lakes. A—oval; B—ellipse; C—"triangle"; D—lemniscate; E—measurements made on oriented lakes.

(Figure 14D). Ellipses were drawn for angles of 20°, 30°, 45°, and 60°. The lemniscate, in polar co-ordinates with unit length, is $r = \cos k\,\theta$. Figure 14D was drawn with k values of 1.5, 2, and 3.

An oval or egg shape which would resemble closely many of the oriented lakes was more difficult to obtain. An oval derived from combining part of an ellipse with an arc of a circle or with another ellipse produced shapes which did not match lake shapes. A solution was obtained by deriving a curve from a circle and lemniscate. The ordinate for the oval is taken as the vertical distance between unit circle inscribed around a unit lemniscate, giving

$$x = \cos\theta \cos k\theta$$

$$y = \sqrt{\cos\theta\ \cos k\theta\,(1 - \cos\theta\ \cos k\theta)} - \sin\theta\ \cos k\theta$$

Since many of the oriented lakes taper to a point from a nearly straight top, being more or less triangular, a logarithmic transformation of the oval (Figure 14A) was used—by plotting it on semilogarithmic paper with the x values on the logarithmic scale—to obtain Figure 14C.

Eighty-eight oriented lakes were classified according to the 13 shapes of Figure 14. This was done by optically projecting and adjusting air photographs of the lakes to a fixed scale, and then visually selecting the closest match for the east and west sides of both the shoreline and the deep central trough, which shows up clearly on air photographs. The measurements shown in Figure 14E were obtained directly from the air photographs. Other measurements included axial trends and those of the submarine "ripple" marks occurring on the benches (Figure 12).

As Table 1 shows, about half of the lakes are bilaterally symmetrical, with similar east and west shorelines. Oval shapes are the commonest, followed by triangular, elliptical, and lemniscate. There is a marked prevalence of tapering shapes, narrow end pointing south. Some of the north ends are straighter than represented by the triangular shape, others are slightly heart-shaped. To adequately represent all possible variants, however, would unnecessarily cumber the classification system. The right sides of the skew lakes have a tendency to be broader than the left.

The general relationships between length and width (Figure 14E) of the lakes are given by:

Equation	r	S_y	S_x	
$d = .193 + .51\,f$	.84	.278	.457	(56-1)
$e = .158 + .45\,f$	.85	.241	.457	(56-2)
$d = .041 + 1.09\,e$	.95	.278	.241	(56-3)

The coefficient of correlation at the one per cent level of significance is $r = .27$, so all are significant; S_y and S_x are the standard deviations of the dependent and independent variables.

Table 1

Shorelines of Oriented Lakes

East Shoreline

West Shoreline		Shape	Lemniscate (D) k			Oval (A) k			Triangle (C) k			Ellipse (B) degrees				
			3	2	1.5	1.5	2	3	1.5	2	3	20°	30°	45°	60°	Total
Lemniscate (D)	k	3														0
		2														0
		1.5			4											4
Oval (A)	k	1.5		1											1	2
		2					6									6
		3						25			6				11	42
Triangle (C)	k	1.5							2							2
		2								1	1					2
		3					1	7	1		5					14
Ellipse (B)	degrees	20°			1							1				2
		30°						1					1	1		3
		45°									1					1
		60°						2			2				6	10
Total			0	1	5	0	7	35	2	2	15	1	1	1	18	88

The central deep troughs (Table 2) of the lakes have a high concentration of oval and triangular shapes, with lemniscate and elliptical shapes barely represented. As with the shorelines, the north ends may be heart shaped, the tendency being far more strongly developed in the troughs.

A χ^2 (chi-square) test shows that the respective groupings of oval-triangular shapes both for the shorelines and deep central troughs are significant at the one per cent level.

The general relationships between the sizes and shapes of the sublacustrine benches and outer shorelines are given by:

Equation	r	S_y	S_x	
$a = .021 + .67\,a'$	.60	.064	.057	(56-4)
$b = .006 + .69\,b'$	.77	.057	.064	(56-5)
$c = .017 + .67\,c'$	.80	.066	.079	(56-6)

As with the widths and lengths, all coefficients of correlation are significant at the one per cent level. The sublacustrine bench tends to be more consistent in width east-west than north-south.

Table 2

Central Deeps of Oriented Lakes

East Side

Shape			Lemniscate (D) k			Oval (A) k			Triangle (C) k			Ellipse (B) degrees				
			3	2	1.5	1.5	2	3	1.5	2	3	20°	30°	45°	60°	Total
West Side: Lemniscate (D)	k	3								1						1
		2														0
		1.5									1					1
Oval (A)	k	1.5				4	1		1							6
		2				1	6	1		2	2					12
		3					1	8		2	2					13
Triangle (C)	k	1.5				1			5	2				1		9
		2					3	1	2	7	2				1	16
		3			1		2	6	1		16					26
Ellipse (B)	degrees	20°										1				1
		30°											1			1
		45°						2								2
		60°														0
Total			0	0	1	6	13	18	9	14	23	1	1	1	1	88

The sublacustrine platform or bench is strongly "ripple-marked" on air photographs (Figure 12). The ripples may be as much as 50 feet from crest to crest and of 6-inch amplitude. The ripple marks are least common on the north bench. Ripple marks which trend north-south are mainly in the southeast part; northeast-southwest ripples in the southwest and west part; east-west ripples are on all the benches but are concentrated on the northeast and northwest parts; northwest-southeast ripples are on the east and southeast parts. Intersecting ripple patterns are frequent. Although the ripple patterns may merely reflect the wind direction of the last storm, they do suggest the direction of wave refraction and longshore currents. The shallow submarine platform is only a few feet deep. The central trough is relatively deep; it is believed to exceed the winter ice thickness in most of the lakes. Therefore, thermokarst effects (*122*) may accentuate the depth of the central trough in a growing oriented lake.

57. Origin: The Carolina Bays of North Carolina, South Carolina, and Georgia were the first extensive group of oriented lakes to be studied in North America. Despite their accessibility and the efforts of numerous researchers, theory after theory of their origin has been proposed, attacked, and discredited. The theories have ranged from a cataclysmic shower of meteorites, through gyroscopic eddy action, upwelling of artesian springs, solution, erosion from Pleistocene winds blowing parallel to the long axes, to the action of hovering fishes while spawning. With the advent of air-photographic coverage of remote areas, other groups of oriented lakes were soon discovered. Cabot (1947) described a large group of oriented lakes on the Alaskan coastal plain near Point Barrow. The lake orientation was attributed to summer winds blowing from the Cordilleran ice-cap, which existed some thousands of years ago, parallel with the long axes of the lakes. Black and Barksdale (1949), after detailed field studies, came to the same conclusion as Cabot, namely that prevailing Pleistocene winds in the direction of the long axes of the lakes were the chief factors which controlled their orientation. Livingstone (1954) took a stand opposed to Cabot, Black, and Barksdale. He suggested that the Alaskan lakes owed their elongation to modern processes, primarily that of longshore rip currents on the lake ends, resulting from winds blowing across the lakes. The oriented lakes of Tuktoyaktuk Peninsula and the Liverpool Bay area were described in 1956 (Mackay, 1956c); their origin and those of the Alaskan lakes were attributed to cross winds. Rex (1961) has reviewed the history and theory of oriented lakes, especially the Alaskan lakes, and has applied hydrodynamic principles which resolve the origin in favour of the cross wind theory. Other areas of oriented lakes are known in Canada; for example, those on Baffin Island (Dunbar and Greenway, 1956, p. 132-134).

Although it is conceivable that the oriented lakes of Tuktoyaktuk Peninsula, Point Barrow, and Baffin Island have differing origins, their remarkable similarity and mode of occurrence argue against this. The Tuktoyaktuk Peninsula and Point Barrow Lakes are strikingly alike in appearance. Both groups of lakes occupy flat featureless coastal plains; some lakes are at sea level. Their shore material is unconsolidated. The lakes of Alaska are in marine silts and sands of Pleistocene age, those of the Tuktoyaktuk Peninsula-Liverpool Bay area are of Pleistocene and Recent sands and silts. In both areas, the prevailing winds (which are nearly opposed in direction) blow across the minor lake axes and longitudinal sand dunes parallel the minor axes. The Alaskan and Tuktoyaktuk Peninsula-Liverpool Bay lakes are in the tundra, where neither trees nor bushes break the sweep of the wind. Other characteristics, such as size, depth, shape, and shore features are much in common. Although the Baffin Island lakes have not been examined in the field, their appearance on air photographs is suggestive of those just discussed.

58. Equilibrium Forms: There is strong evidence to support the view that oriented lakes are in equilibrium with forces now operating. Some oriented lakes have developed within irregularly shaped depressions through infilling by tundra polygons which rim the lake shores. Other oriented lakes have been partially drained, so that young lakes overlap onto older ones, with no change in orientation. Just as some oriented lakes are shrinking, by ingrowth of tundra polygons, so others may enlarge by erosion or coalescence, yet all preserve the same orientation. In particular, small ponds, no more than 200 feet across, are now growing (Mackay, 1956c). Therefore, orientation seems in balance with present conditions.

Origins requiring special conditions, such as buried ice bodies (Brochu, 1957; Mackay, 1957b), meteorite scars, or tidal eddies are excluded, because any acceptable theory must explain the fact that oriented lakes are in dynamic equilibrium with their environment, reacting to changes in it.

Wind action is an obvious source of energy for lake orientation. To have the prevailing winds and sand dunes at right angles to the long axes of the lakes in the Mackenzie Delta area might be a coincidence, but when the same holds for the Alaskan lakes, the coincidence seems too much to accept.

Bruun (1953) has shown that the equilibrium form of a shoreline of finite length composed of readily moved unconsolidated material is a cycloid. Rex (1961) has utilized the theory of Bruun, together with other relevant hydrodynamic theory, to demonstrate that wind action can account for the Alaskan oriented lakes. The maximum littoral drift, from theoretical and empirical studies, is at an angle of 50° between the deep water angle of the waves and the normal to the shoreline (*see* also: Tanner, 1962; Zenkovitch, 1959).

In the modern alluvium of the Mackenzie Delta and in the Pleistocene sediments to the east, there are many lakes with elongated bays exposed to only one wind direction. An examination of over 100 such bays, of large and small lakes, shows that an equilibrium shape approaching that of the cycloid (Figure 15A) tends to form at the ends of the bays, irrespective of the compass orientation of the bay or the directions of the prevailing winds. Most of the equilibrium bays are

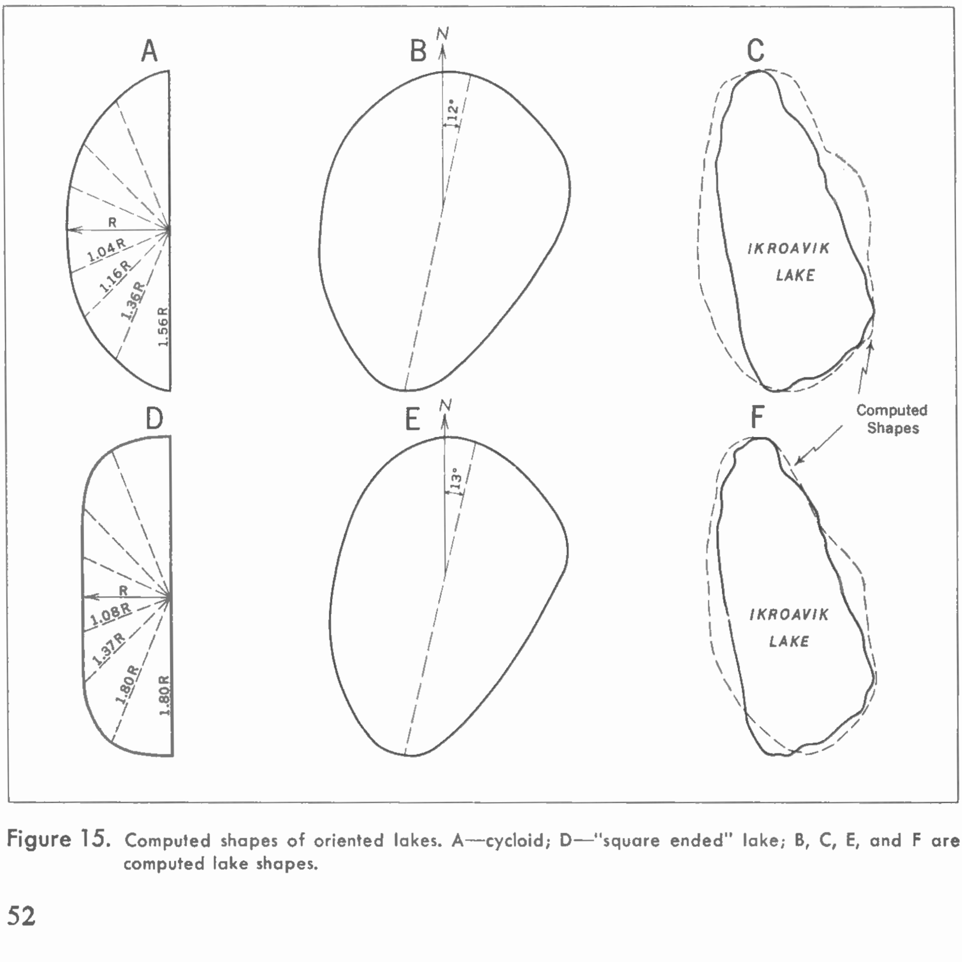

Figure 15. Computed shapes of oriented lakes. A—cycloid; D—"square ended" lake; B, C, E, and F are computed lake shapes.

smooth in outline, even though portions of the shorelines include erosional and depositional features (Figure 30). This shows that ice-shove does not form the smoothly rounded embayments, because eroded cliffs are just as much a part of the bays as are sand bars. The equilibrium bays depart from the shape of a cycloid in tending to have the greatest erosion, and hence curvature, at an angle of 60° to 70° between the deep water waves and the normal to the shoreline (Figures 15D and 30). The sides of the bays may be eroded enough to form slight pockets. Some bays are nearly square-ended, with rounded corners. Thus there is little doubt that wind action is the agent responsible for rounding the bayheads.

59. Wind Effectiveness: The effectiveness of wind in rounding a bay will depend upon numerous factors. Firstly, there are variables involving the wind itself, such as its velocity, steadiness, and duration. Secondly, there are the variables not directly related to the wind, such as the fetch, beach slope, shore materials, and the presence or absence of an ice cover. The height of a wave is nearly proportional to the square of wind velocity (V) and as high waves have much more energy than low waves, the effectiveness of a wind may depend, in general, on some power of the velocity. Moreover, weak breezes may not raise waves of sufficient height to erode and transport material, so there may be a lower threshold wind velocity (v) below which winds are ineffectual. In small lakes of short fetch, the influences of weak and strong winds are similar (Zenkovitch, 1959). For larger lakes, where the fetch is several thousand feet, strong winds are far more effective than weak winds.

On the basis of the preceding assumptions, the resultant effect (R) of a wind from a given direction may be approximated by:

$$R = \sum_i f_i \, (V_i - v)^n$$

where i is the period of record, here taken on a monthly basis; f_i is the frequency for month i, and n is a positive exponent (*see* Landsberg, 1956). For the oriented lakes, the open period is from early June to the end of September ($i = 6$ to 9). The threshold velocity (v) and the exponent (n) must be estimated.

If it is assumed that winds from each of the 16 compass directions recorded in climatic records tend to develop curved bays, approximately cycloidal, then an oriented lake may be viewed as the summation or integration form of 16 cycloids, each of different size. To test this theory, let us assume that the diameter of the generating circle of the cycloid is equal to the resultant (R) for a given wind direction. Then the vector lengths of the cycloid (Figure 15A) at angles of 22.5°, 45.0°, 67.5°, and 90° are about 1.04 R, 1.16 R, 1.36 R, and 1.56 R. Thus a resultant wind, blowing in one of the 16 compass directions, contributes a total of 9 vectors, weighted as just given, towards an over-all lake shape. The summation around the compass rose can then be viewed as an integrated shape embracing the effects of all winds.

In order to test the method, threshold velocities (v) of 0 and 5 miles per hour were used, and exponent values (n) of 1, 2, and 3. More elaborate treatment was considered superfluous, because the wind velocity records for Tuktoyaktuk and Nicholson Peninsula, the nearest stations, are short.

Figure 15B shows the integrated lake shape based upon wind velocities at Nicholson Peninsula with v = 0 and n = 2. The theoretical lake shape of Figure 15B may be compared with a mean lake shape computed from equations in section 56, given a lake length of one mile.

	Mean Lake Shape (from equations)	*Computed Lake Shape* (from 16 cycloids)
Length (given)	1.00 mile	1.00 mile
Width d	0.70 mile	0.73 mile
Width e	0.61 mile	0.62 mile
Axis	N 11°E	N 12°E

The agreement between the mean lake shape and the computed lake shape is excellent. As the wind data for the computed lake shape is for Nicholson Peninsula, the orientation for the eastern group of lakes (*55*) is used for comparison.

If shapes other than the cycloid are used, then the weighting factors change. However, as long as the shape is elongated normal to the wind direction producing it, oval lakes oriented slightly east of north tend to result. For instance, if the common "square-ended" shape (Figure 15D) is used with weights of R, 1.08 R, 1.37 R, 1.80 R, and 1.80 R, v = 0, n = 2, the resulting lake is slightly narrower than that derived from the cycloid (Figure 15E). If a more circular shape than the cycloid is used, with weights of R, 1.04 R, 1.08 R, 1.20 R, and 1.08 R, the resulting lake is slightly more elliptical than that derived from the cycloid. If a threshold velocity (v) of 5 miles per hour is used with values of n of either 1 or 2, the lake shape still remains oval with an orientation agreeing well with field conditions.

In order to corroborate the theory, the method was applied to Ikroavik Lake, Point Barrow, Alaska (data for shape and winds from Carson and Hussey, 1960). Figure 15C shows a comparison of the computed and actual lake shore for v = 0 and n = 2; Figure 15F shows the same for v = 5 and n = 2. The overall agreement in lake shape and orientation is good.

The predominant winds for the open season at Nicholson Peninsula and Tuktoyaktuk, as given over a one- to two-year period, are from the east and northwest. The eastern and western resultant winds, computed in terms of kinetic energy units $(\text{miles/hour})^2$ (*see*: Rex, 1961, p. 1024) are N 82°E and N 81°W for Nicholson Peninsula and N 78°E and N 75°W for Tuktoyaktuk. The obtuse angles between eastern and western resultants are 163° for Nicholson Peninsula and 153° for Tuktoyaktuk. When the prevailing winds are nearly opposed, as in the above cases, the application of equilibrium shapes can readily be seen to result in oval, triangular, and heart-shaped lakes.

The preceding analysis which associates lake orientation and shape with an equilibrium shape, approximated by the cycloid, gives a close agreement in lake shape between theory and actuality. However, the precise mechanism of lake orientation remains unexplained. The amount of littoral transport, the aspect of a two-cell circulation, the possibility of thermal effects in a permafrost region, the preference for vegetation growth under favored micro-climatic and topographic conditions, and the effect of lake ice on lake orientation need further study.

Coastal Features

60. Coastal recession is rapid. The process is well known to local inhabitants, many of whom have personal recollections of coastal retreat amounting to several hundred feet and the disappearance of offshore islands. An excellent illustration is afforded by retreat at King Point to the west of the delta where coastal photographs taken in 1905-06 by Amundsen and in 1913 by Chipman (Canadian Arctic Expedition, 1913-18) were compared with the coastal bluffs as seen in 1957. In the 50-year period, the average rate of retreat was from 1 to 2 feet a year (Mackay, 1960, p. 23).

Coastal erosion is aided by at least three factors: the coastal material consists predominantly of easily eroded sands; numerous horizontal ice sheets facilitate thermal erosion; and the rise in water level associated with violent westerly storms helps to compensate for the small tidal range by removing eroded and slumped debris. Recent submergence of the Mackenzie Delta may also be a contributory factor.

The coast of the Mackenzie Delta from the delta of Blow River to the mouth of Napoiak Channel is being cut back rapidly, in some areas 5 to 10 feet a summer. The north and northeast coast of Richards Island is likewise receding. Persons long familiar with Tuktoyaktuk tell of coastal recession there. Coastal retreat at Point Atkinson was sufficiently large for Richardson (1851, p. 254) to notice the changes between his visits of 1826 and 1848. During one exceptionally violent storm, in August 1955, parts of the sandhills at Point Atkinson were cut back 10 to 15 feet. The spits extending northeast and southwest from Point Atkinson have migrated southeastward concomitant with retreat of the sand hills. Anchorages and channels formerly used by coastal boats have been completely altered or obliterated within the brief span of 30 years. At Cape Dalhousie, an island 20 feet high and several hundred feet across is reported to have been washed away within the past 20 years; all that now remains is a sandbar.

Storm beaches, from 3 to 5 feet above sea level, have been built in many low coastal areas between Kittigazuit and Cape Dalhousie. In places, storm beaches produced by westerly winds merge into parabolic dunes built by easterly winds. Storm beaches which are close to sea level, and therefore, frequently flooded, usually support a cover of lyme grass, a few low willows, and sedges. The higher less frequently inundated beaches have little lyme grass, but a fairly continuous cover of lichens, avens, ground birch, willows, sedges, etc.

The recognition of storm beaches and the limits of flooding is made easy by the widespread distribution of driftwood. Most of the driftwood lodged along the coast from the Mackenzie Delta to Cape Dalhousie has come down the Mackenzie River, the primary source being the Liard River (Kindle, 1921; Giddings, 1952). Most logs are of spruce in an excellent state of preservation, although they lack branches and the roots have been worn away to stubs. Logs may exceed 2 feet in diameter and 50 feet in length. The concentrations of driftwood show up as faint whitish lines on air photographs, so that the inner limit of marine inundation can be plotted by the limit of driftwood (Figure 16). Because sea level is raised highest by westerly storms, driftwood becomes lodged farthest inland on low coasts with a westerly exposure, excellent examples being the coast to the southwest of Point Atkinson and Toker Point where driftwood has been washed inland as far as 2 miles. On the west side of Richards Island, driftwood is lodged in wide belts along many bays. Driftwood logs are scattered over the low alluvial islands of the modern Mackenzie Delta and at flood limit on both sides of Shallow Bay. Logs abound on bars, spits, and beaches. In general, driftwood will lodge with the trunk pointing in the direction of movement, the roots upstream. Thus, driftwood logs may be used as indicators for flow in areas where currents change direction during storms, as during submergence of portions of the Mackenzie Delta.

Figure 16

Driftwood along a storm beach on the east side of McKinley Bay, August 23, 1960. The driftwood is 8 to 9 feet above mean sea level and from 300 to 1,000 feet inland. The area just below the driftwood has sedges, willows, avens on 6-inch hummocks, and low-centred polygons. The driftwood is on the slope of a 3-foot rise of the ground. The higher area within 20 feet of the driftwood is dry with 6-inch hummocks; about 70 per cent is in avens, 20 per cent in sedges, 10 per cent lichens with a scattering of low willows. The hills on the skyline are hummocky sand dunes which rise to a maximum height of 27 feet above sea level. (Geog. Br. JRM-55-12-2).

Few Mackenzie River-transported logs are found along the shores of Liverpool Bay. Some spruce and poplar are carried to sea by the Anderson and Kugaluk Rivers, but the amount is negligible. Driftwood in the Eskimo Lakes is mainly of willow and alder, and is likewise sparse. Usually only a thin line of wood chips and small branches accumulates at the high water mark.

The effect of sea ice on coastal recession is negligible, although the presence of an ice cover is obviously of importance in reducing the effective duration and fetch of wave action. Ice-shoved ridges and heaps may be pushed up by the bulldozing action of windblown ice. Large boulders are shoved around by ice and tend to be concentrated on exposed points. Considerable seasonal variation in the abundance of boulders has been observed. At Tuktoyaktuk and Nicholson Peninsula, numerous large boulders over 2 feet in diameter were observed to have been removed within a period of 5 years, probably by the combined processes of coast recession and ice action.

Weathering

61. The visible effects of weathering are subdued and inconspicuous, particularly in exposures of unconsolidated sediments. In areas of limestone, as near Inuvik, furrowed and fretted solution surfaces have formed with a relief of roughly half an inch or less. An examination of stones in ground squirrel burrow heaps shows that some stones have acquired a lower coating of calcite in the time elapsed since the burrows were dug.

A slight profile development is visible in numerous sections both in the treed and treeless areas (*see* Leahey, 1947). Light-toned ashy podzolic-like horizons are present south of the tree line, as at Inuvik, and to the north as at Point Atkinson where a leached layer occurs in sands at a depth of 2 feet. In several sandy terraces at the mouth of Kugaluk River, ashy layers lie 10 inches below the surface. At Inuvik, there is a 1- to 2-foot iron cemented hardpan at a depth of about 8 feet in coarse gravels (*24*). Although the gravels above and below the hardpan are lithologically similar, they look different because those above are weathered brown, those below are unweathered and grey. As the hardpan and the weathered gravels immediately above it are in permafrost, the hardpan presumably formed when the active layer was deeper. A thicker active layer could be attributed to a period before insulating organic matter accumulated at the surface or to milder climatic conditions. Similar hardpan layers occur elsewhere, as in outwash gravels in the Eskimo Lakes.

Frost-shattering of all types of boulders, including Shield erratics, appears to be most rapid along gentle beaches where the boulders are frequently wetted.

Solifluction and Soil Creep

62. The importance of solifluction can be greatly overestimated if the "flow" in "soil flow" is literally interpreted. A very careful search must be made before a typical solifluction feature, such as a solifluction lobe, can be found. There is, however, a gradual downslope movement of the surface material, just as in soil creep in temperate climates.

In extensive areas south of the tree line and east of the Mackenzie Delta, downslope stripes show up clearly on air photographs. The striping, which was studied in the field only in the Anderson River Valley and Sitidgi Lake areas, arises from vegetation differences between better drained swells and poorer drained swales which trend downslope. The relief differences between swells and swales are minor, being at most only a few feet. The widths of the individual swells and swales are upwards of 10 to 20 feet; lengths reach a maximum of over half a mile. The drier swells support spruce, willow, ground birch, blueberry, etc., whereas the wetter swales are sedgy. The vegetation differences register as alternating light and dark stripes on air photographs, and as they trend downslope, they resemble hachures. Whether the striping should be classed as a solifluction feature is debatable, because drainage seems to play a very important role in its development. Where seen in the field, striping formed on slopes as low as 2°.

Bun-shaped mounds 2 to 5 feet in diameter and up to a foot in height are common elements of the ground surface in both tundra and forest areas (Figure 17). The mounds are characteristic of mixed soils in which the silt fraction exceeds

Figure 17

Excavation through the middle of a hummock at 68°58′N, 132°52′ W, on August 16, 1960. The excavation was made to frozen ground which was 27 inches deep beneath the hummock centre. The brunton rests on frozen ground which is seen to plunge down steeply towards the hummock centre. The horizontal pencil in the hummock is at the same level as the brunton. The strings mark the boundary between mineral soil and organic matter. Note how organic material occurs along the frost surface. (Geog. Br. JRM-60-7-9).

10 to 20 per cent. The tops of the mounds may be bare or turfy. Usually the amount of organic matter increases towards the inter-mound depressions where there is fresh moss at the surface, decomposed moss at a depth of several inches.

The inter-mound organic material may extend downwards as a curtain uniting with an organic layer roughly paralleling the permafrost surface. The organic layer may range from a mere darkening of the soil to a layer of sphagnum moss a foot or more thick. The permafrost surface is deepest beneath the mound centres and shallowest beneath the depressions, so that the details of the permafrost surface reverse those of the ground surface (Figures 18 and 19). The burial of surface-accumulated organic matter is believed to take place by the gradual downslope movement of the mounds, aided by frost action (Mackay, 1958c). Thus, organic layers are "plastered" onto the permafrost surface. The process described above is believed to act slowly over a period of thousands of years (Mackay and others, 1961, p. 46).

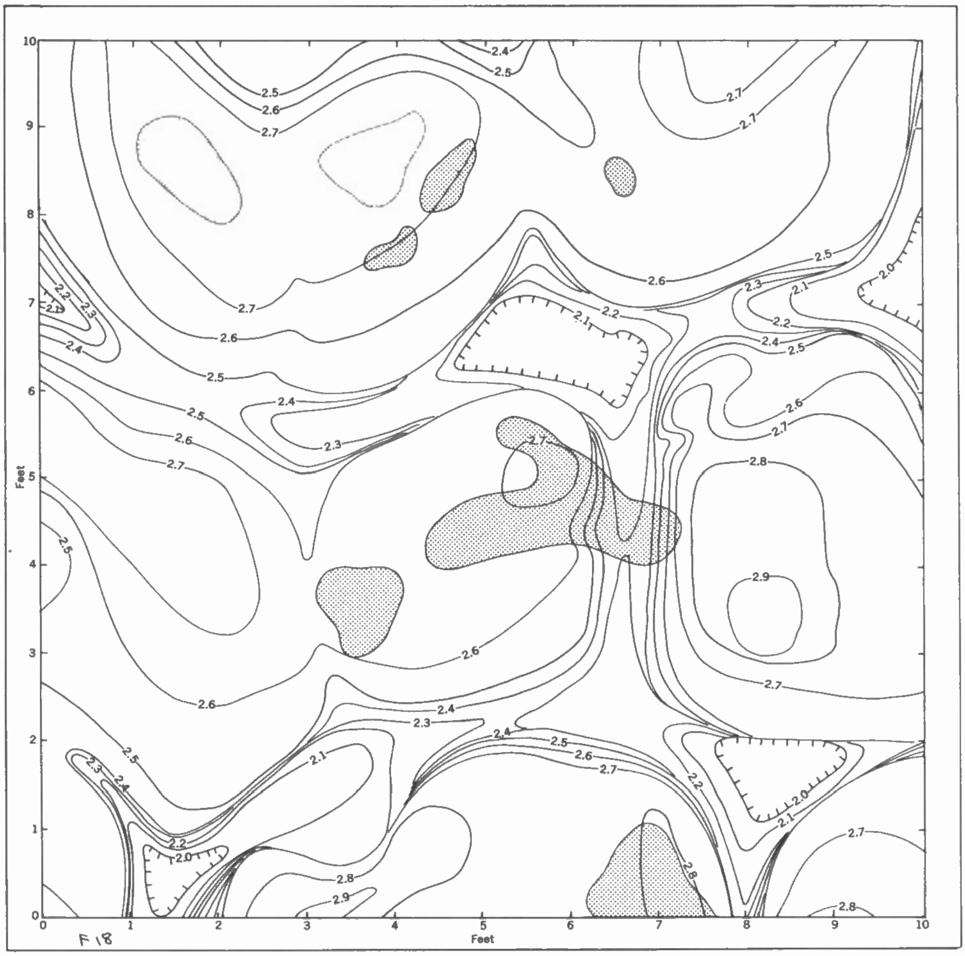

Figure 18. Contour map of ground surface at 69°30′N, 139°05′W on August 5, 1957. The contour interval is 0.1 foot. Stippled areas represent bare ground.

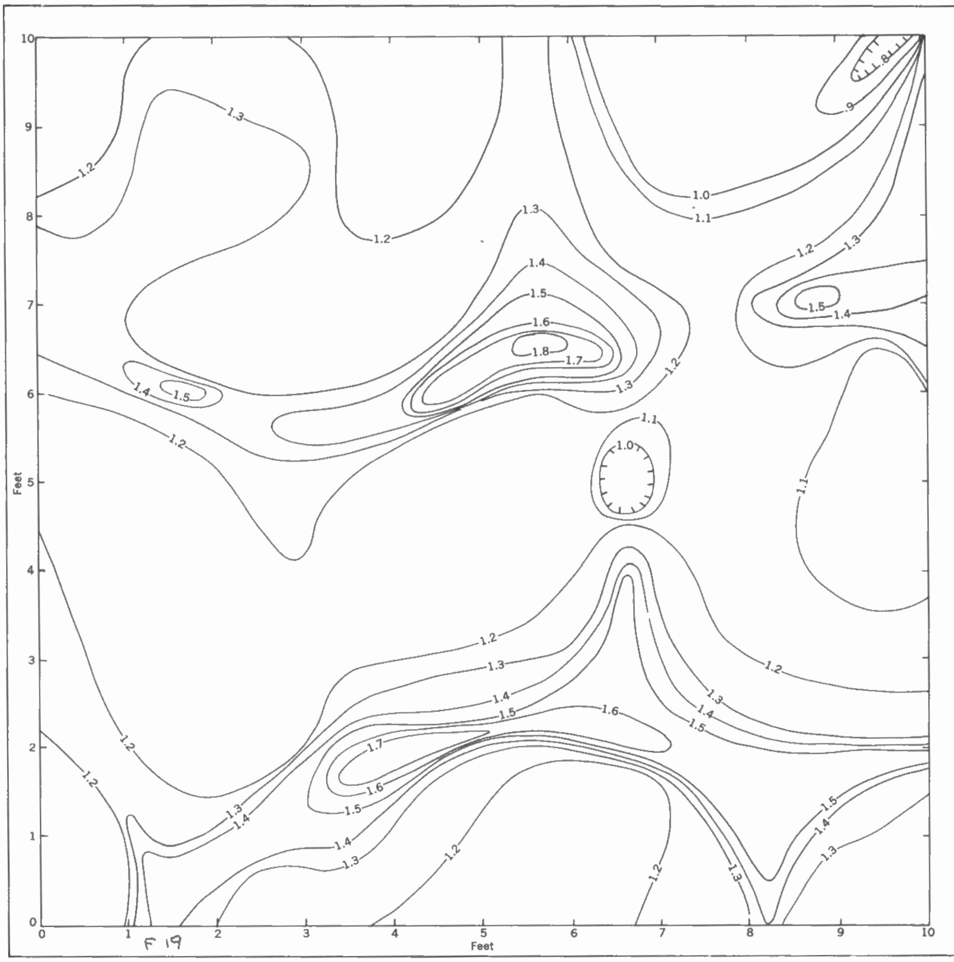

Figure 19. Contour map of the frost table, directly beneath the surface area of Figure 18, on August 5, 1957. Contour interval 0.1 foot. *Note:* A high frost table typically occurs beneath low spots on the ground.

Ground Ice

63. Ground-ice segregations are notable for their variety and wide-ranging distribution (Figure 21). Ground ice is found as a bonding cement, as veinlets and stringers, as clear vertical ice-wedges, as dirty banded tabular ice-sheets, and as plugshaped pingo ice-cores.

In the Mackenzie Delta, the ice content (per cent of ice to dry soil, on a weight basis) may exceed 50 per cent for depths of several tens of feet (Pihlainen and others, 1956). However, the ice content at greater depths may be quite small (Johnston and Brown, 1961).

Massive, tabular, horizontal sheets of ground ice (Figure 20) from several feet to over 20 feet thick underlie many tens of square miles of the region of

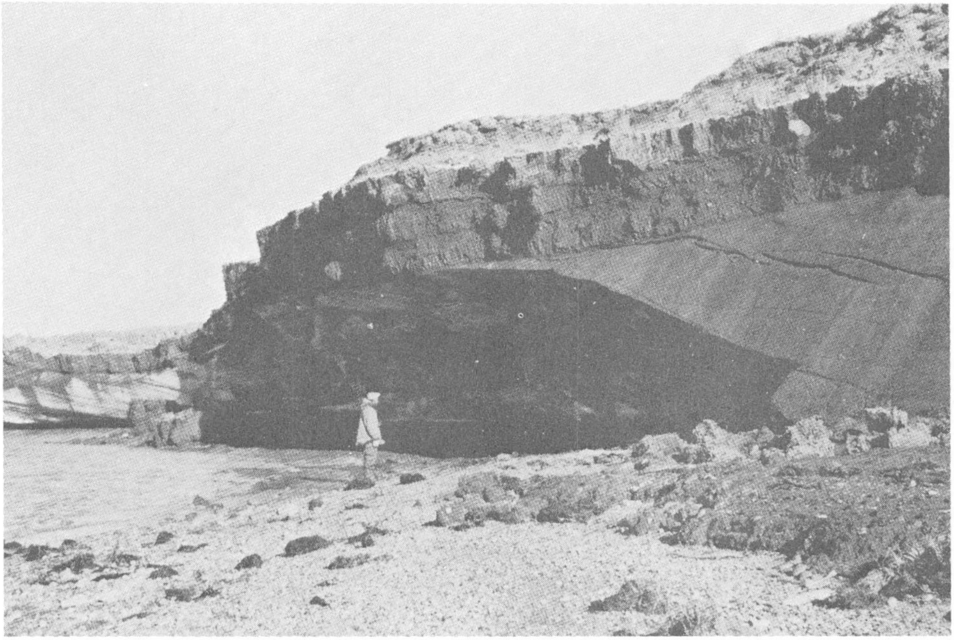

Figure 20. Massive ground ice exposed at sea level along the coast 4 miles southwest of Tuktoyaktuk on July 13, 1954; by July, 1955, all of the ice-cave had collapsed. The ground ice sheet was 15 to 20 feet thick, banded, and of fairly clean ice with a slope of about 50°. A 7-foot section of three boulder-silty-clay beds containing undecomposed organic matter (moss, roots, willows, etc.) overlaid the ice. The boulders were partly of Shield type and striated. The material had been transported by slumping from a higher 5-foot-thick ice lens, located 100 feet inland. (Geog. Br. R-3-4).

Pleistocene sediments and are also present, though less abundantly, elsewhere (Figure 21). The tops of the ice-sheets may coincide with the base of the active layer which is about 1 to 2 feet below the surface, or they may occur at depth. Although the ice may be whitish, most exposures look darkly metallic, partly because of a thin veneer of fines washed from above. Much of the ice is dirty in the sense that mineral soil is scattered throughout the ice and may even exceed it in volume. The included mineral soil is usually fine-grained. For example, a typical soil in which a large sheet of ground ice formed was medium sand 24 per cent, fine sand 36 per cent, very fine sand 15 per cent, and silt 25 per cent. Ice with much included mineral matter tends to be strongly banded. The banding may be horizontal, tilted, or even show folds 5 to 10 feet in amplitude. Ice-wedges which penetrate into ice-sheets have always been the younger of the two ice features (Figures 22). The white ice-wedges protrude because of differential melting. In one exposure, the banding of an ice-sheet continued through a large ice-wedge as if some replacement took place as the ice-wedge grew (Figure 27).

Melting of ice-sheets takes place most readily along disturbed slopes, such as a coastal bluff or river valley. As the ice melts, it maintains a steep slope of about 50°. The overburden drops or slides down the ice face and then moves along as a soggy mudflow. Mud levees are common in the mudflows. Large crescentic

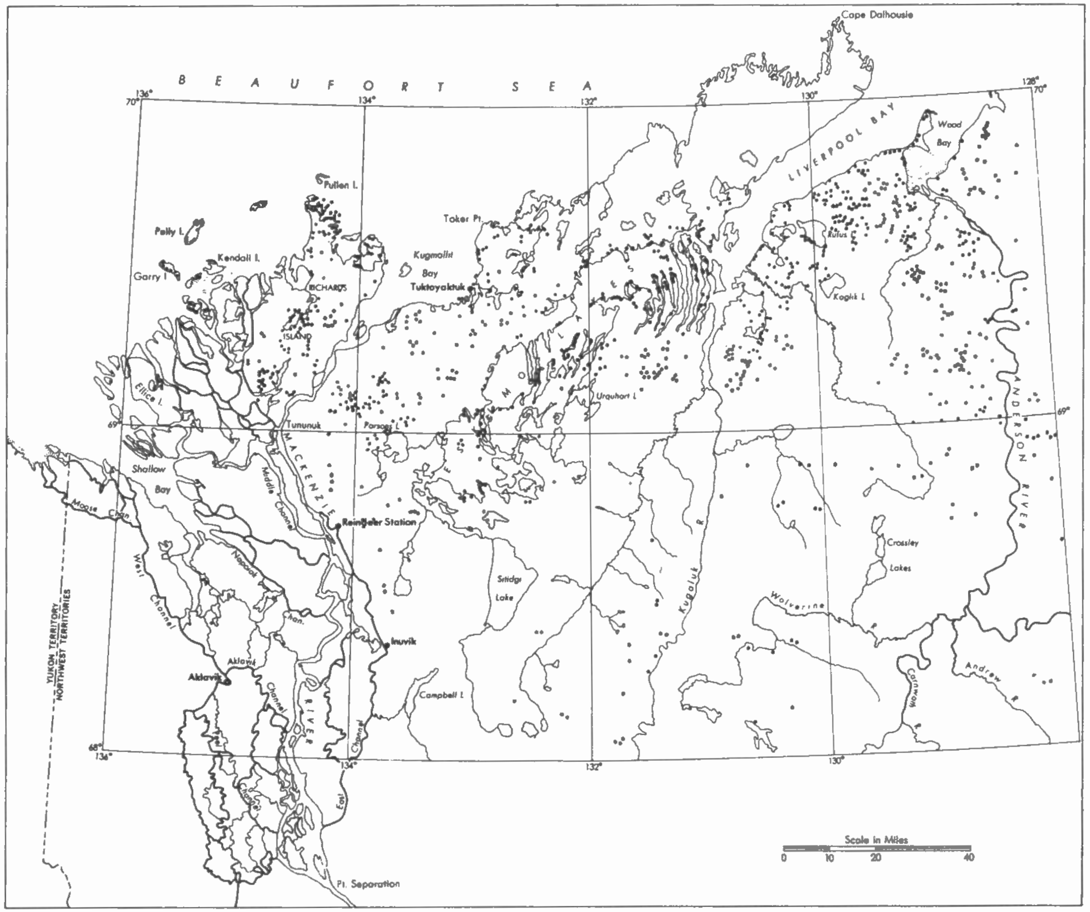

Figure 21. Ground-ice slumps, as mapped from air photographs. In the area from Richards Island to Liverpool Bay, each dot represents one or more ice slumps; in the rest of the map area, each dot usually represents the site of only one slump.

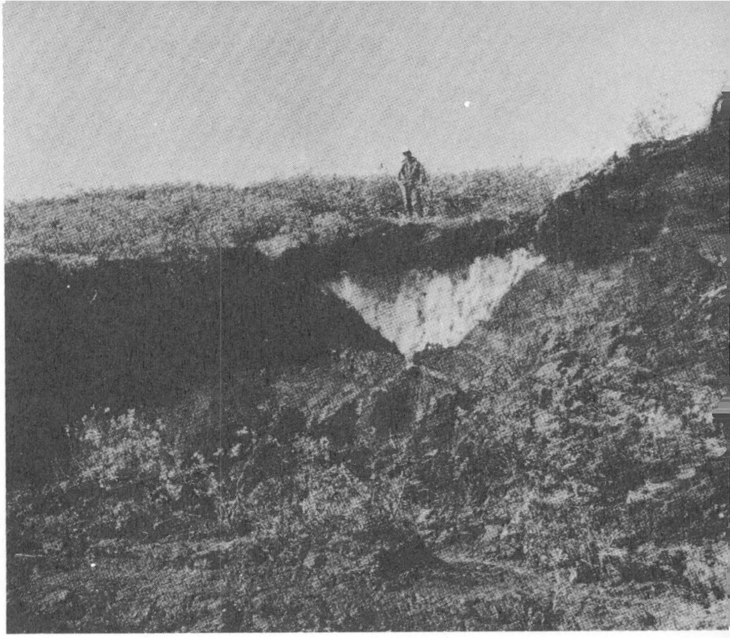

Figure 22

Ground-ice slump with large ice-wedge. Several feet of peat overlie the ground ice in the top right hand part. Frozen silts, with a high ice content, abut against the left side of the ice-wedge. Clods of earth with willows and other plants tumble down the retreating scarp to become buried in the slumped debris. Location: 69°31'N, 132°00'W. August 10, 1960. (Geog. Br. JRM-60-5-5).

scars mark the final termination or burial of an ice-sheet. The scars may be ribbed, like seats in an amphitheatre, with the ribs marking fluctuating periods of melting. Scars resulting from ground ice melting are easily distinguishable from landslide scars by the relative paucity of slumped material. However, the mudflows resulting from the melting of ice-sheets are puzzling features when first seen, because they may show a crude stratification of mixed soils, boulders, and fresh organic remains preserved in newly formed perennially frozen soil (Figure 20).

The ice-sheets are believed to have formed during the progressive downward aggradation of permafrost in fine-grained soils in an open system, where fresh water moved to the bottom of the freezing plane, as new ice formed. Some ice-sheets may be sill-like injection features formed similar to pingo ice (*80*). If, as seems probable, the ice-sheets were formed by aggradation of permafrost from the surface downwards, then the banding may represent seasonal or annual freezing effects. In the case of an ice-sheet with little foreign matter, the thickness of ice (E) formed in time t with the ice surface at temperature T may be estimated from Stefan's solution for the freezing of thin ice (Ingersoll and others, 1954, pp. 194-198),

$$E = \sqrt{\frac{-2Tkt}{L\rho}}$$

where k is the conductivity of the ice, L the latent heat of fusion, and ρ is the density. Using $k = 5.2 \times 10^{-3}$, $\rho = 0.92$ and $L = 80$ cgs units for ice,

$$E = .012\sqrt{-Tt}$$

It is obvious that if the temperature remained constant, the thickness of the ice-sheet would increase with the square root of time, if geothermal effects are neglected. If the ground surface temperature were to change (e.g., decrease), the

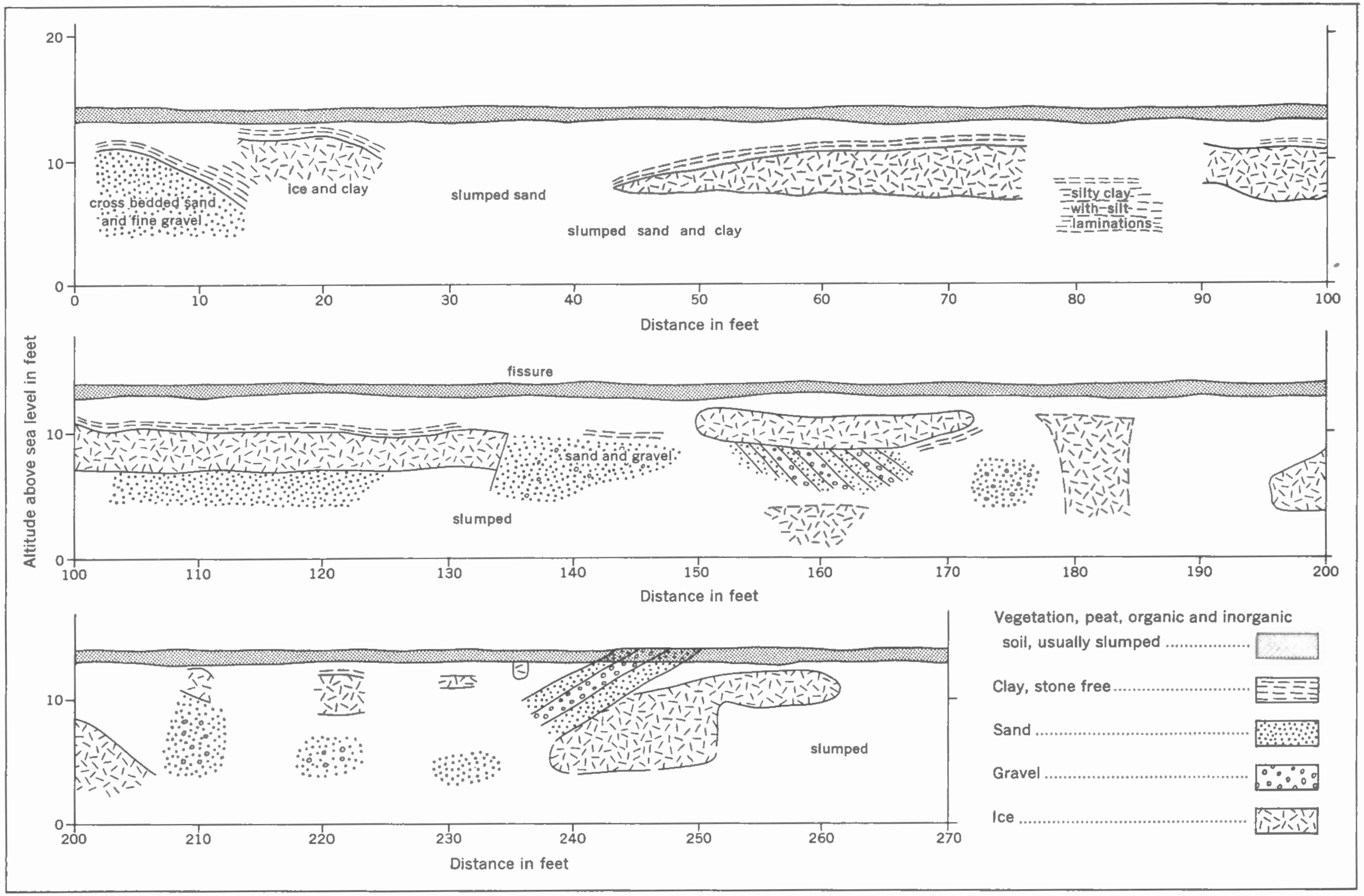

Figure 23. A coastal exposure to the northeast of Tuktoyatuk, July 15, 1954. The horizontal ice-sheets contained considerable fine-grained soil, whereas the ice-wedges (e.g., that at 180 feet) were of clear whitish ice. Wherever ice-wedges intersected the horizontal sheets in coastal exposures, the ice-wedges, which were the younger features, stood out in relief from differential melting.

functional relationship would also change. As an illustration, with T of $-0.1°C$, it would take about 200 years to freeze 10 feet of ground ice; with T of $-1°C$, about 20 years; and with T of $-5°C$, about 4 years. With one more year of freezing, the thickness of ice at the bottom would increase about .3 inches for $T = -0.1°C$; 2.75 inches for $T = -1°C$; and 14 inches for $T = -5°C$. The rate of heat radiation from the interior of the earth is about 40 cal cm $^{-2}$ per year. The heat of fusion of ice is about 70 cal cm $^{-3}$, so that the rate of melting from geothermal heat might approximate 0.5 cm per year, or about 0.20 inches per year. Therefore, at a constant temperature slightly below freezing (e.g. $T = -0.1°C$) the retarding effect of geothermal heat would be significant. As the bands in ice-sheets at a depth of 10 feet below the tops of the ice-sheets may be only several inches apart, the bands can hardly represent yearly accretions at temperatures such as $-5°C$, although they might be seasonal bands. Some bands are of dirt, and probably result from laminations in the sediments. Perhaps it is worthwhile mentioning that overburden pressure can prevent ice lensing (Penner, 1959). Therefore, ice lensing to form ice-sheets beneath drift may have occurred subsequent to ice retreat, because the weight of glacier ice might have been sufficient to prevent lensing.

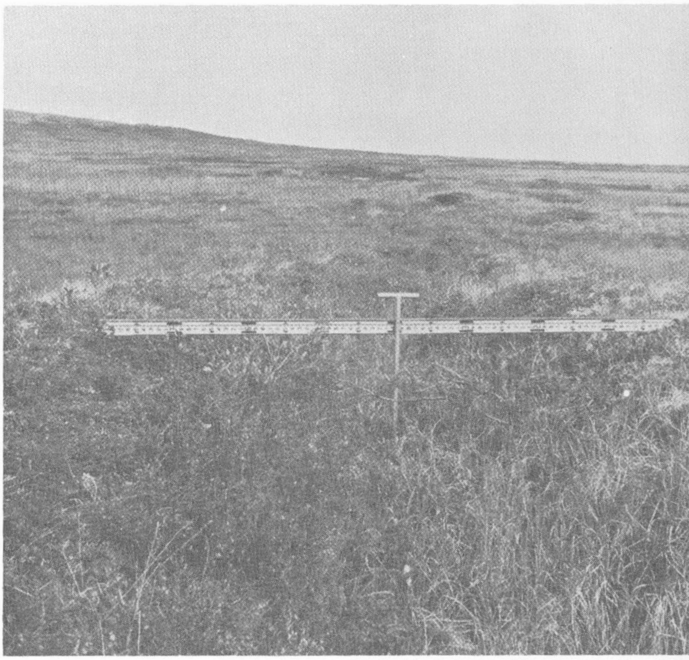

Figure 24

View along a fissure of a tundra polygon (shown in section in Figure 25C) at 69°31′N, 130°53′W, on July 31, 1960. The 3-foot auger is pushed down to frozen ground. Sedges grow in the fissure; a willow partly obscures the auger; ground birch are in the lower left; the light toned patch at the right end of the stadia rod has many lichens. The pond on the right is in the centre of a low-centred polygon with many sedges. A pingo is in the background. (Geog. Br. JRM-60-3-6).

Patterned Ground

64. Tundra polygons are the most conspicuous type of patterned ground. The polygons are numerous in the southern part of the Mackenzie Delta, the area of Pleistocene sediments, the Caribou Hills, sand and gravel terraces, and

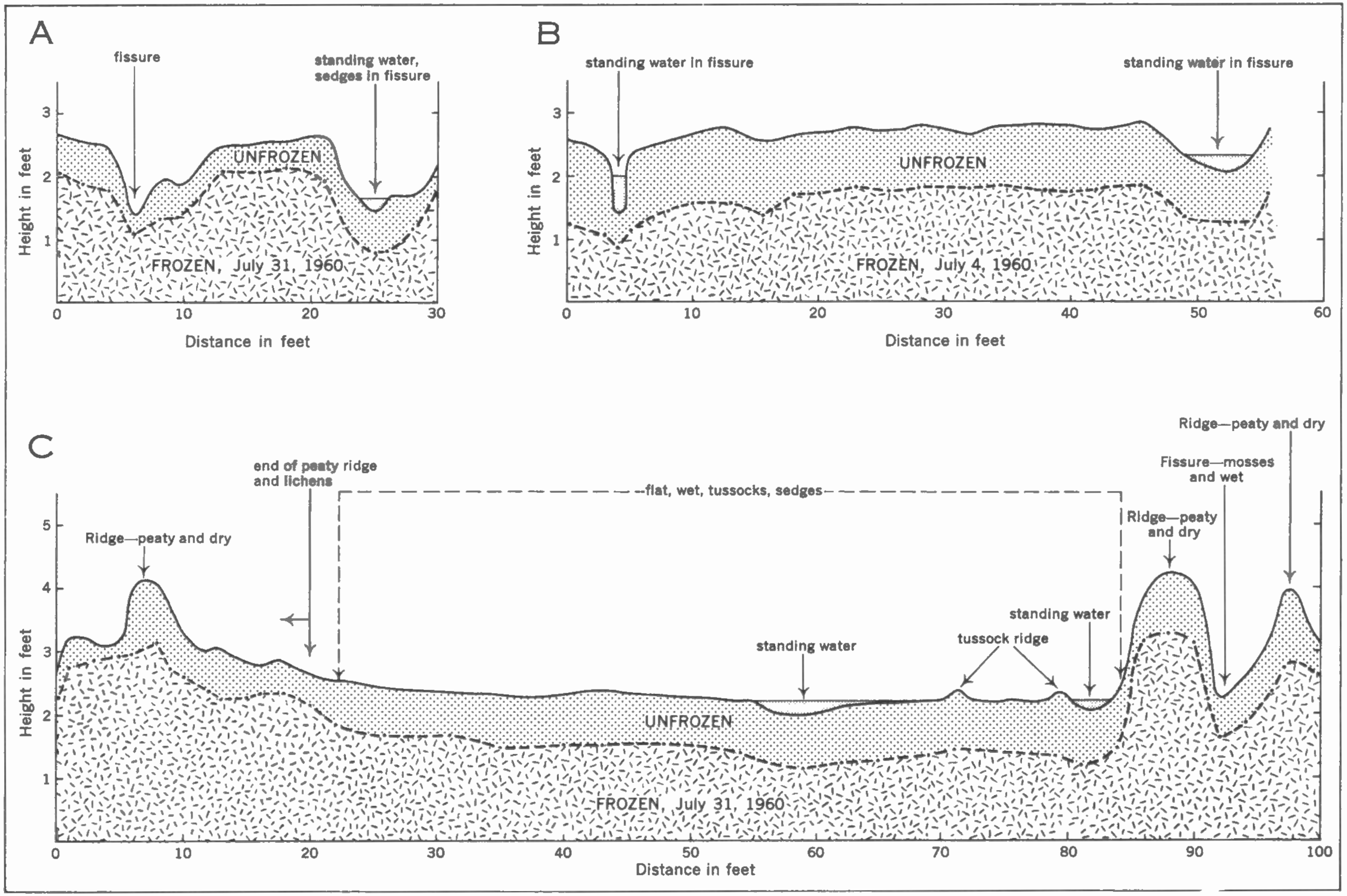

Figure 25. Profiles across tundra polygons in the Eskimo Lakes area. A—high-centred peaty polygon; B—polygon subject to periodic inundation; C—low-centred polygon.

in poorly drained areas, particularly former lake bottoms and sedgy depressions. Tundra polygons range from about 20 to 100 feet in diameter, their size being dependent on the climate, rheological properties of the soil, and stage of growth. The ice-wedges which underlie the bounding fissures of tundra polygons form in response to thermal tension set up in the frozen ground (active layer and permafrost) during the winter (Leffingwell, 1923; Lachenbruch, 1960a and 1960b). Water, freezing in the vertical contraction cracks, forms tapering ice-wedges. Recurrent fractures cause the wedges to grow, particularly towards the top where they may flare out (Figure 22) to a width exceeding 15 feet. Lateral growth of the ice-wedges is associated with the formation of bounding parallel ridges. In the initial stage of growth, the ridges enclose, or partially enclose, the flat terrain in between to give a saucer shape to the young low-centred polygons (Figure 24 and 25C). The raised rims are peaty and dry, with a cover of plants such as the crowberry, cranberry, labrador tea, ground birch, willow, cloudberry, lichen, moss, grass, and sedge. The central flats have sedge, tussocks (which may stand isolated or in ridges), fernweed, sourdock, moss, a few willows, etc. As the ridges widen, encroaching on the centre, peaty mounds develop, the end result being a high-centred bun-shaped peaty polygon (Figure 25A). In the process of transition from low- to high-centred polygons, subdivision of the low-centred polygons frequently occurs. High-centred polygons, therefore, are on the average smaller than low-centred polygons. The transition from low- to high-centred polygons is shown in Figure 25 where A and C are high- and low-centred polygons close by a pingo at the eastern end of the Eskimo Lakes. Figure 25C shows a typical low-centred polygon in the bottom of a drained lake. The fissures when observed in summer were wet with sedges and mosses, the bounding ridges were peaty and dry. The low centre, between the ridges, had pools of standing water with several tussock ridges just protruding above.

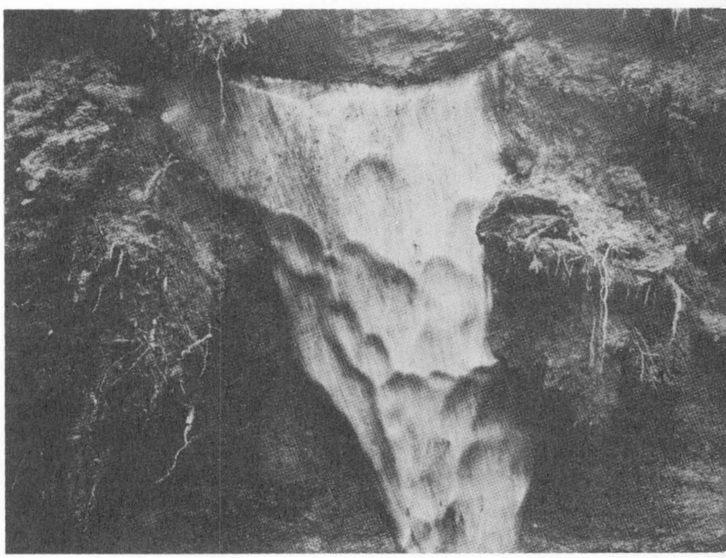

Figure 26

Ice-wedge of clean white bubbly ice showing vertical banding and scalloped surface caused by differential melting. The top of the wedge is about 18 inches below the ground surface. Location: 69° 07′N, 132°35′W. August 12, 1960. (Geog. Br. JRM-60-5-11).

Figure 27

Ice-wedge intersecting a ground ice sheet, which is the banded material in the lower half of the photo. The bands of the ice sheet apparently continue through the ice-wedge, suggesting "replacement". Location: 69°07'N, 132°35'W. August 12, 1960. (Geog. Br. JRM-60-6-1).

The active layer of the ridges was mainly in peat, whereas that of the low centre was in soil with minor amounts of organic matter. Figure 25A shows a profile across a high-centred polygon, a few hundred feet from the low-centred polygon, on the side slope of a pingo. The high-centred polygon was much smaller. The fissures of the high-centred polygon were wet, sedgy, and mossy but the interfissure area was dry with a growth of ground birch, labrador tea, willow, cranberry, crowberry, lichens, sedges, etc. The active layer was entirely in peat, which was at least several feet thick. The transition from low-centred polygons to high-centred ones is accomplished through a widening of the peaty ridges, the growth of peat and peaty mounds in the low centres, and the subdivision of the larger polygons into smaller ones. The growth of peat causes a rise in the ground surface and a thinning of the active layer.

The low-centred polygons are extremely abundant in the flats of Tuktoyaktuk Peninsula, particularly near Point Atkinson where they extend unbroken for miles. The high-centred polygons are less common, occurring frequently on the lower slopes of pingos, in old lake beds, and in abandoned channels.

The tundra polygons which develop on low coastal flats below high water level, and are therefore subjected to periodic flooding, have a different appearance from the low- and high-centred polygons discussed above. These polygons tend to have poorly developed bounding ridges, or even no ridges (Figure 25B). The surfaces are sedgy. Peat may accumulate to a depth of 1 to 2 feet over the mineral soil, which is usually a sand or silt. The active layer tends to be deeper than in normal low-centred and high-centred polygons. These coastal polygons, being liable to frequent inundation, have little chance of growing into high-centred polygons.

Tundra polygons are found in the Mackenzie Delta, being particularly abundant in the southern part (Figure 28). The polygons are easily recognizable on air photographs, even in the presence of a nearly continuous forest cover (Figure 72). Some polygon ice-wedges are exposed by rapid bank erosion at break-up and subsequent draw-down effects, but burial is sufficiently rapid and exposures so few that many long-time residents of the delta, when questioned, have been unaware of the ice-wedges. The largest ice-wedge seen was 8 to 9 feet across, the top of the wedge lying 4 feet below the surface which was 16 feet above river level. In a horizontal distance of 500 feet, there were at least 6 other ice-wedges exposed along the cut bank; the combined width of the wedges exceeded 30 feet. The ice-wedges extended below river level to an unknown depth. Spruce trees, over 100 years old, grew above the wedges. An ice-wedge 8 feet across might be at least several thousand years old if rates of growth are roughly comparable to those elsewhere (Black, 1952, Müller, 1962).

The sorted forms of patterned ground, such as stone circles and polygons, are absent in the Mackenzie Delta, because the sediments are stone-free. Such forms are also very scarce in the Caribou Hills and coastal areas where stones in sufficient concentration are rare, although other factors requisite to their formation may be lacking.

Unsorted circles, with bare soil surrounded by vegetation, are wide-spread in silty soils with considerable clay in areas such as the foothills of the Richardson Mountains, Caribou Hills, and silty surface soils of the coast.

String bogs occupy a few depressions and swales in the area south and east of Sitidgi Lake, but are not numerous.

Pingos of the Pleistocene Mackenzie Delta Area

65. Pingos are ice-cored hills, which are typically conical in shape. They are relatively stable intrapermafrost features, normally enduring for hundreds to thousands of years, thus differing from the smaller seasonal frost blisters or icing-mounds which form above the permafrost surface in winter. Pingos are abundant and widely distributed in the Mackenzie Delta area in: (a) the distal portion of the modern Mackenzie Delta, and (b) the area of Pleistocene deposits to the east (Figure 29). The pingos in the two areas differ in age, size, and details of formation and so are discussed separately.

The Eskimo word *pingo,* applied locally by natives to conical mounds in the Mackenzie Delta area, was proposed by Porsild (1938) as a technical term for ice-cored hills. The term is now widely adopted. The Mackenzie Delta area has one of the world's greatest concentration of pingos.

66. Distribution: There are over 1,400 pingos in the Mackenzie Delta area between the west side of the delta, at the Yukon-Northwest Territories boundary, and Nicholson Peninsula on the east. The area involved is a northeast-southwest coastal zone some 200 miles long and 50 miles wide. The pingos occurring along the Yukon Coast, west of the Mackenzie Delta, may not exceed ten. To the east of Nicholson Peninsula, pingos are very few (Mackay, 1958 a, pp. 54-56).

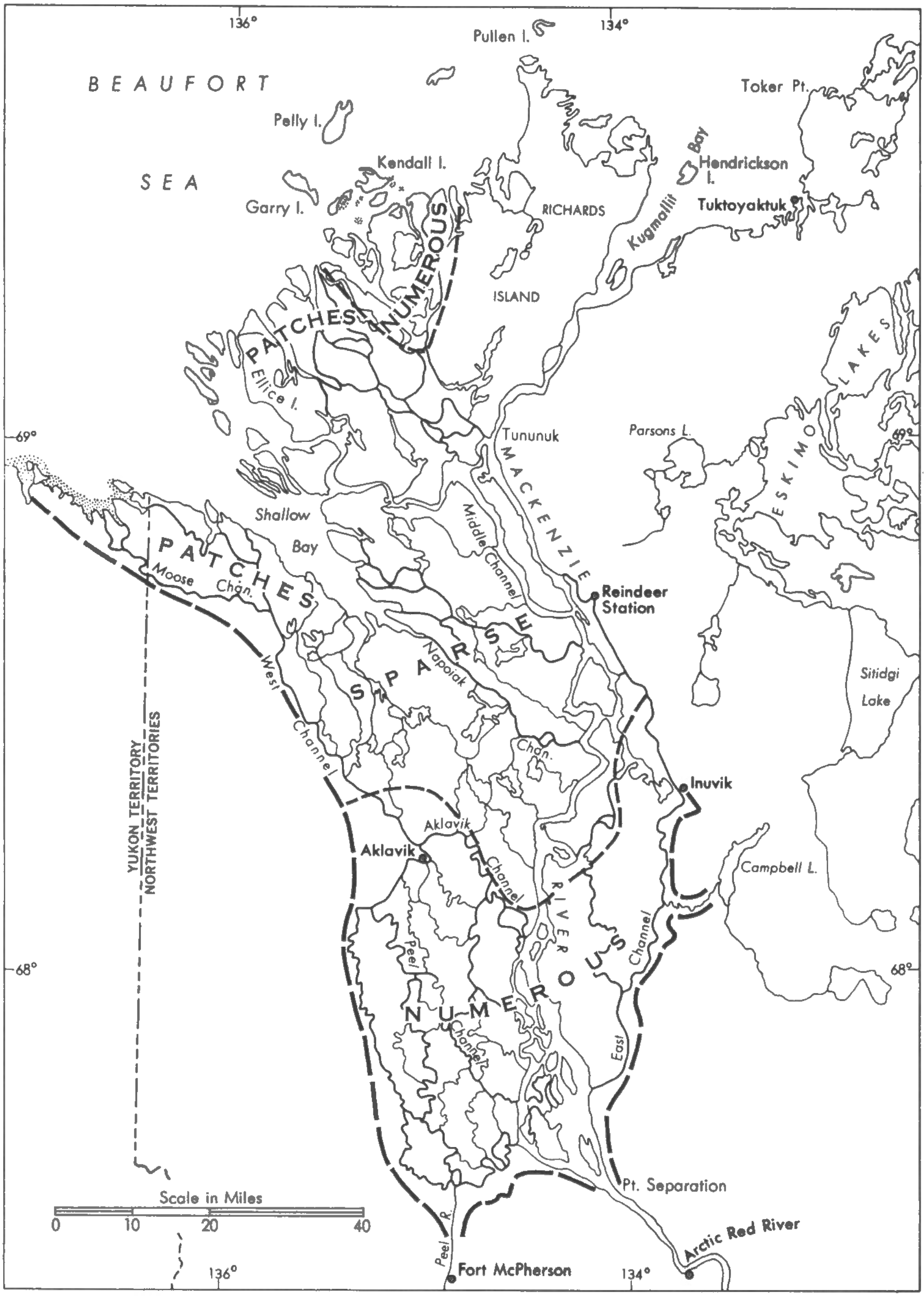

Figure 28. Distribution of tundra polygons in the Mackenzie Delta as mapped from air photographs.

The largest group of pingos, numbering over 1,350, lies in a belt extending east from Richards Island through Tuktoyaktuk Peninsula to Cape Dalhousie; this group includes the pingos on the south side of the Eskimo Lakes. Stager (1956) has plotted the locations of about 1,380 pingos in this area, and the writer about 1,350, the slight discrepancy arising from the definition of size used to delimit the smallest pingo. As the figures should be considered a minimum, rather than a maximum, the total number of pingos is approximately 1,350 to 1,400. This group of pingos is in an area of Pleistocene sands and silts with rolling relief. A second group of pingos occurs on the seaward portion of the modern Mackenzie Delta below storm level. A third group of about 10 pingos is found on an older portion of the modern Mackenzie Delta near the lower course of West Channel. There are several pingo-like hills along the eastern slopes of the Richardson Mountains at 68°29′N and near Mount Goodenough (Fraser, 1956), and one or more pingos in the southern portion of the Port Brabant map sheet of the National Topographic System.

67. Occurrence in Depressions: The most distinctive terrain characteristic of the pingos is their occurrence in flat low-lying areas. With very few exceptions, the low areas are present or former lake basins or channels with poor drainage. No feature, which unequivocally could be identified as a pingo, has been observed either in a large lake (e.g. one mile in diameter) surrounded by water considerably deeper than the winter thickness of ice, or on top of a positive relief feature, such as a mesa-like tableland.

The correlation between areas of pingo concentration with completely or almost completely drained lakes is borne out by their irregular distribution. Most drained or partially drained lakes are found in areas where a river system, even though poorly integrated, has developed in postglacial time, or else where coastal recession has initiated drainage. Areas with interior or no visible surface drainage tend to have few old lakebeds.

On Richards Island, the main concentration of pingos is on the west half where drainage is most complete (Figure 29). The eastern and northeastern part has numerous relatively deep lakes with few streams, and consequently a minimum area of old lakebeds or shallow lakes. On the east side of East Channel, between Tununuk and Pete's Creek, the concentrations are related to stream drainage. The high density of pingos, about 10 per 10 square miles, between Kittigazuit and Toker Point, is in a coastal region where cliff recession and stream action have partially drained large areas. Tuktoyaktuk Peninsula, east of Toker Point, has a pingo density averaging less than 1 pingo per 10 square miles. Much of the region is low and flat, with either a very poorly integrated drainage system, or no visible surface drainage. Thus, the conditions suitable for pingo growth are largely absent. Pingo concentrations along the north and south coasts of the Eskimo Lakes are closely associated with shore-cliff recession and stream action.

There is a general relationship between the density of depressions "suitable" for pingo growth and the number of pingos present. To illustrate this, a random sample of 25 air photographs was studied. The area analyzed on each sample photograph was about 30 square miles, the total sampled area covering about 750

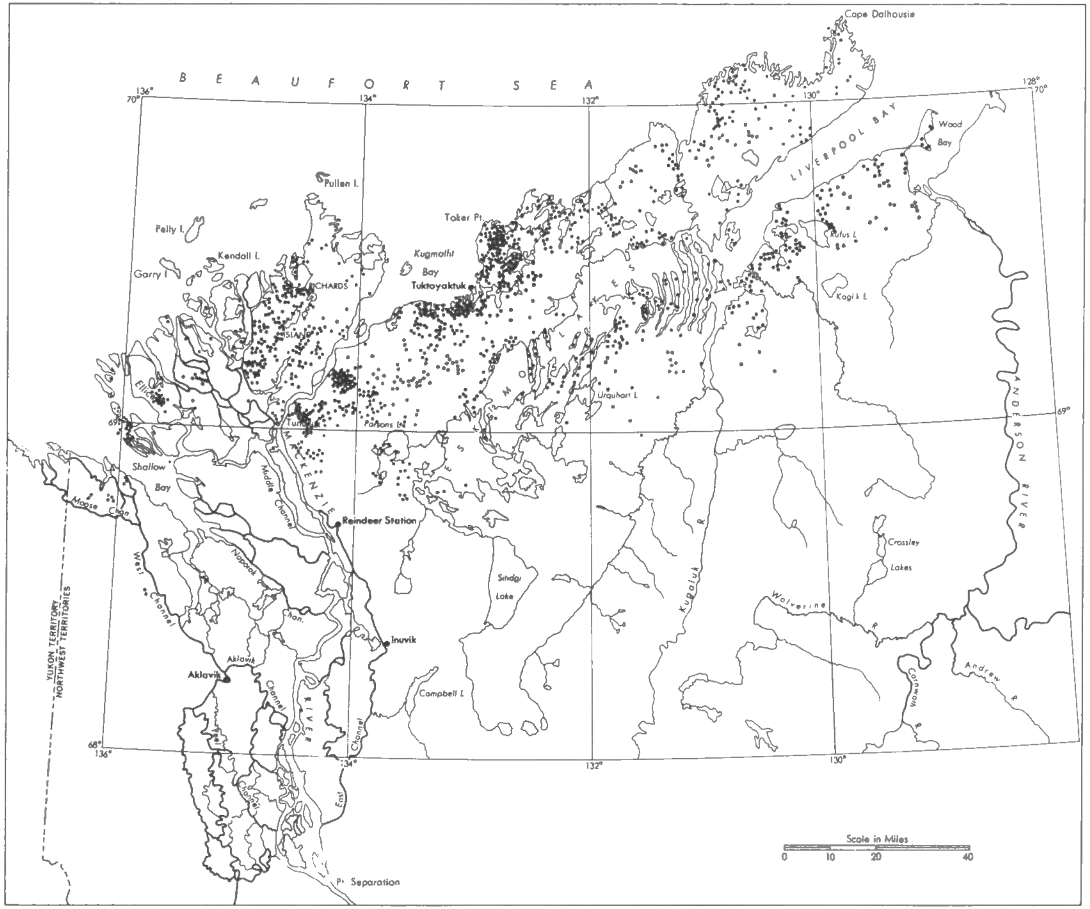

Figure 29. Distribution map of pingos as mapped from air photographs.

square miles. For each 30 square miles, the limits of all depressions containing pingos were outlined, the boundaries following as closely as possible the estimated extent of the depressions which generated the pingos. Then all similarly appearing areas without pingos on the air photographs were marked. The 25 points can be represented reasonably well—for the range of data—by the line

$$Y = 10^{.071x - .283}$$

where y is the pingo density per 10 square miles and x is the depression density per 10 square miles. Although a certain degree of subjectivity was unavoidably involved in the sampling procedure, the results suggest that the probability of finding a nearby pingo, if one pingo is selected at random, increases non-linearly with the depression density or number of "suitable" pingo depressions.

68. Shape: The typical pingo rises as a solitary hill, ranging from round through elliptical to oval in ground plan. Basal diameters vary from about 100 to 2,000 feet and heights from about 10 to 150 feet. The highest pingos have intermediate sized ground diameters of about 500 to 700 feet. As diameters increase beyond 500 to 700 feet, pingo altitudes tend to decrease, so that the broadest and longest pingos become merely large bulge-like swellings or protuberances.

The outline of a pingo tends to be simple and smooth. Oval shapes are most common, but many are elliptical or nearly circular. The rare crenulated outlines are usually the result of the coalescence of two or more pingos.

In vertical cross-section, most pingos are asymmetric. Even the classic and most photographed pingo, Ibyuk pingo southwest of Tuktoyaktuk, is symmetrical only when viewed from one direction; the variation in slope between opposite sides approaches 20°. The steepest pingo slopes rarely exceed 45°. The gentlest slopes, which may be only several degrees, are found on small incipient pingos but most commonly on large bulge-like swellings with diameters of more than 1,000 feet. The bulge-like swellings suggest that lateral growth has taken place at the expense of vertical growth.

The side slopes of pingos are either straight or convex upward. The majority of pingos—76 per cent according to Stager (1956, p. 16)—have smooth rounded summits. Most summits of the largest pingos are ruptured, with star-shaped craters surrounded by cuestas which seemingly open out as rupturing progresses, like the petals of a budding flower.

69. Overburden Thickness: The thickness of the sediment over the ice-core varies considerably and appears to be directly related to the size of the pingo, with small pingos having the thinnest cover of sediments, large pingos the thickest. For example, on the east side of McKinley Bay, a violent storm (in the summer of 1955) cleaned off the face of a partially sectioned pingo to expose the ice-core. About three-fifths of the pingo had previously been cut away. The original pingo was probably oval, about 300 feet long and 25 to 30 feet high. The cut face of the pingo showed an ice-core of clear, white to bluish, bubbly ice, with crystals 1 to 1.5 inches in diameter, beneath 4 to 5 feet of brownish sand. The sand at the contact with the ice was dark grey. Part of the overlying sands were lacustrine, as fresh water shells and peat were present. Of the five known examples in which the overburden thickness may be estimated with reasonable accuracy, the cover is about one-half to one-third of the pingo height. The greatest variation in cover thickness, as determined from collapsed pingos, is in irregularly shaped pingos, or those with asymmetrically located ice-cores.

70. Type of Sediments: All of the pingos are believed to have developed in Pleistocene interglacial sands and silts locally veneered with glacial drift and postglacial sands, silts, and organic matter rarely totalling more than 10 to 20 feet in thickness. Inasmuch as most pingos have grown in lake depressions, organic matter interbedded with fine-grained lake sediments is typical of the top few feet of a pingo. The lake sediments may overlie postglacial material or glacial drift, some of which is reworked.

At a variable depth, which is estimated at about 10 to 20 feet, the underlying sediments are believed to consist predominantly of sands to an undetermined depth. Although no stratigraphic section has been observed directly beneath a pingo ice-core, many tens of miles of coastal sections in which pingos occur have been examined in the field, there being every reason to believe that the sediments observed are identical with those beneath nearby pingos. Over 30 "samples" of the sediments from Richards Island to Nicholson Peninsula have been tested for grain size. With very few exceptions, the sediments are at least 95 per cent in the sand range, most specimens testing 99 per cent sand, of predominantly fine- to medium-grain size. The silt fraction rarely exceeds one or two per cent. A few laminae are locally present; silty clays may occur at depth. However, pingos appear to have developed in sandy material which is too coarse-grained to be susceptible to extensive ground ice-sheet formation.

71. Ice-Cores: Drilling records and exposures of naturally sectioned and collapsed pingos suggest that their ice-cores generally resemble the gross outer shapes of the pingos but they have somewhat steeper sides. The bottom of the pingo ice is inferred to lie close to, or below, the level of the flat terrain around the pingo base.

Much information on the size and shape of pingo ice-cores has been obtained from a study of collapsed pingos. When pingo ice melts, usually through summit rupturing of the protective cover but also through lateral cutting along a bluff, the overburden cover along with slumped side material infills part of the depression left by melting of the core. The bottom of the collapsed central depression—which usually contains a lake—does not represent the bottom of the pingo ice-core but stands at least as much above it as there is infilling by overburden material. The bottoms of ponds in the collapsed pingos are not often appreciably above the bases of the pingos; in some, the bottoms may be 5, 10 or more feet below the surrounding flat terrain. When allowance is made for slumped and collapsed material, most pingo ice-cores are inferred to bottom below the level of the adjacent terrain.

There seems to be a good possibility that some pingos have more than one ice-core. Pingos, apparently compound in form, have been observed with one

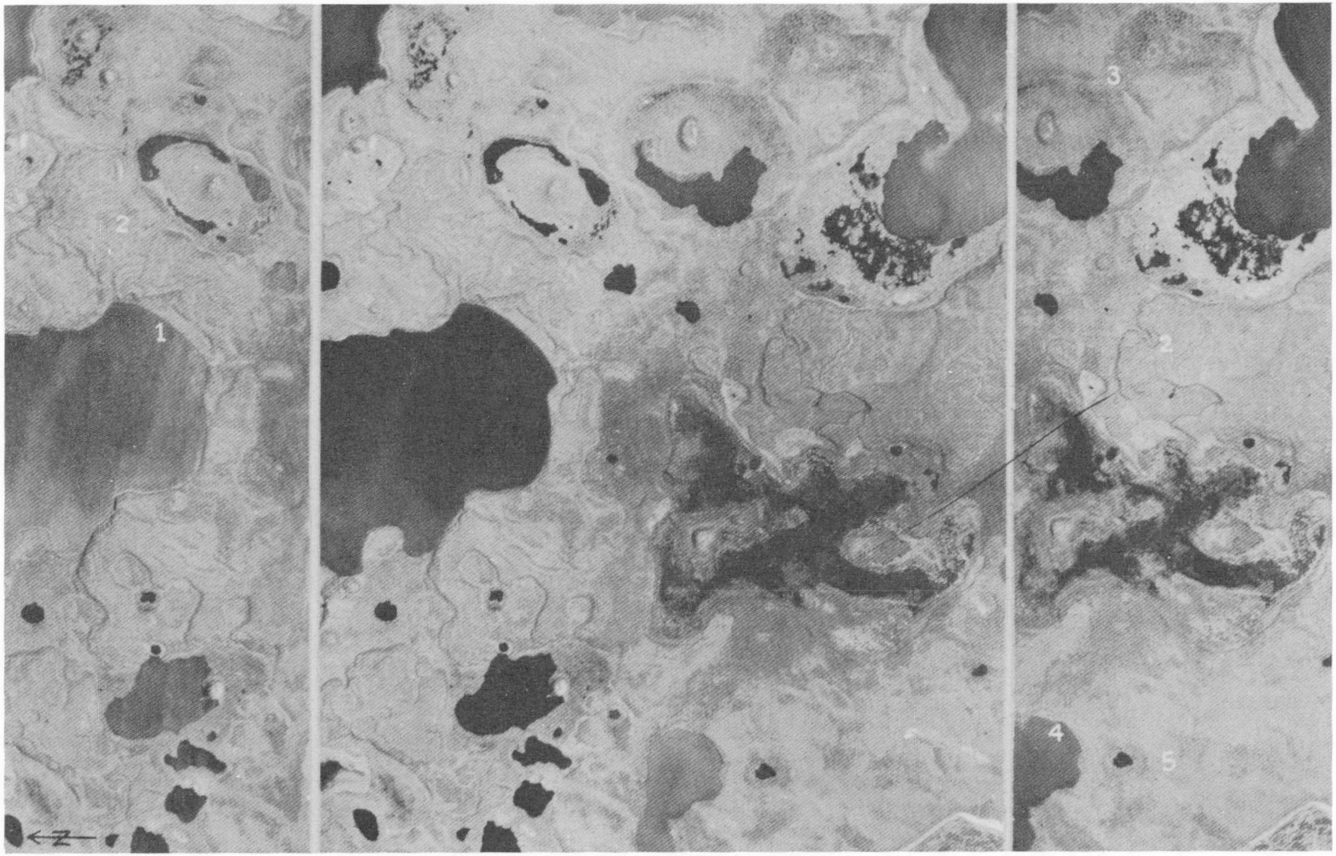

Figure 30. Stereo triplet of equilibrium bays, pingos, and other features. The area is northeast of Tuktoyaktuk centred at 69°32′N, 132°50′W. 1—equilibrium bay; note that the symmetry of the bay is identical where the shore is against both high and low land. 2—involuted terrain. 3—high-centred polygons. 4—equilibrium bay like location 1. 5—the small lake midway between numbers 4 and 5 has a tiny equilibrim bay similar in shape to 1 and 4. Ten pingos can be recognized easily. (RCAF A 12902-120 to 122).

collapsed ice-core (for example, RCAF photo A 12760-20) adjacent to an isolated, uncollapsed portion. The physical appearance of such a compound pingo strongly suggests two separate ice-cores; some may have three cores.

72. *Theories of origin:* Only two basic theories have been seriously proposed for the origin of the Pleistocene Mackenzie Delta area pingos. True, the early Arctic explorer Richardson (1851) who observed the pingos in 1826 and again in 1848, considered them to be a peculiar type of sand hill, but his opinion was presented more in the nature of a comment rather than as a reasoned theory. Gussow (1954) has suggested that pingos are piercement domes which have formed as a result of a geostatic load on thick buried ice of Pleistocene origin. The theory does not seem to be tenable (Mackay, 1958 pp. 54-56) primarily because pingos are superficial features formed in response to local conditions.

The most generally accepted theory on the origin of pingos has been proposed by Porsild (1938, p. 55) who states they "... were formed by local upheaval due to expansion following the progressive downward freezing of a body or lens of water or semi-fluid mud or silt enclosed between bedrock and the frozen surface soil, much in the way in which the cork of a bottle filled with water is pushed up by the expansion of the water when freezing." Porsild points out that the bottoms of the larger lakes are unfrozen, but will freeze when lakes become sufficiently shallow to freeze solidly to the bottom in winter. The rapid infilling of lakes which formed a "closed system" is attributed primarily to sedimentation.

Müller (1959), in his detailed analysis of the Mackenzie "closed system" type of pingo, accepts Porsild's basic theory, including the shoaling of lakes resulting from infilling. The growth of a pingo is depicted as follows: (1) a relatively large (e.g., more than 1000 feet in diameter) and deep (e.g., more than 90 feet) lake will have an unfrozen central portion surrounded by permafrost; (2) gradual infilling of the lake by organic matter and sediments simultaneously reduces the lake size and depth; (3) permafrost forms, ab initio, at the bottom of the lake concurrent with infilling, to create an impermeable cap on the unfrozen material beneath; (4) the inward growth of permafrost at the sides of the unfrozen central core causes a hydrostatic pressure through volume expansion on freezing; (5) the superfluous water is trapped in a "closed system" created by the impermeable capping of permafrost at the lake bottom, the advancing permafrost surface at the sides, and saturated soil at depth; and (6) the expelled water, taking the path of least resistance in a closed system, forms an ice-core near the geometric lake centre.

A consideration of the origin of pingos as proposed by Porsild, and elaborated upon by Müller, requires a knowledge of permafrost conditions beneath lakes.

73. *Depth of Unfrozen Ground Beneath Lakes:* If depths in a lake exceed that of the winter ice thickness, the subjacent bottom sediments in the deeper portions will be unfrozen, irrespective of the presence or absence of permafrost at greater depth, because the thermal capacity of the unfrozen pool of water will prevent the subjacent bottom sediments from freezing. The thickness of lake ice in the Mackenzie Delta area is estimated to range from about 3 to 4 feet in the south to 5 to 6 feet—some estimates range to 7 feet—along the coast, with con-

siderable local variation depending on factors such as the severity of the winter and the nature of the snow cover. Therefore, lakes deeper than 4 to 8 feet, depending upon local conditions, may be expected to have unfrozen bottom sediments.

The depth of unfrozen ground beneath a winter lake pool depends on factors such as: lake area, shape, depth, bottom configuration, and water temperature; the mean ground temperature and its past thermal history; the nature of the soils and disturbances induced by nearby lakes.

The only record for the depth of permafrost beneath lakes in the Mackenzie Delta area has come from drilling operations of the National Research Council (Johnston and Brown, 1961) some 5 miles southwest of Inuvik. The drill holes are in the modern alluvial delta where disturbances are introduced by adjacent bodies of water and the processes of delta sedimentation. Nevertheless, in a shallow lake some 900 feet in diameter, the centre was unfrozen to a depth of at least 240 feet, the permafrost surface plunging at a high angle from the lake edge towards the centre. Away from the lake, permafrost at least 260 feet thick was encountered.

74. Ground Temperatures: There are no continuous long-term ground temperature records for the pingo area. North of Tununuk, a pingo at 69° 02′N and 134° 25′W was drilled in 1954 and thermocouples installed to a depth of 25 feet (Pihlainen and others, 1956). Ground temperature at 25 feet was —4.1°C for August 29, 1954 and August 15, 1955. At the Ibyuk (Crater Summit) pingo near Tuktoyaktuk the temperature on July 1, 1955, at a depth of 40 feet in the ice was —4.4°C and that at nearby Sitiyok pingo on June 30, 1955, at 36 feet was —8.2°C (Müller, 1959, pp. 94-95). As the level of zero annual amplitude for seasonal temperature disturbances may be over 50 feet, the surfaces of pingos are not plane, and the diffusivity of pingo ice is higher than that for adjacent soils, the ground temperatures cannot be representative of normal horizontal ground conditions. However, they do permit rough estimates of mean ground temperatures at the base of the active layer.

At Inuvik, the mean ground temperature at a depth of 47 feet for three years of record is about —3.4°C (Brown, 1960, p. 170). As mean ground temperatures tend to increase with depth because of geothermal heat effects, the mean temperature at the top of permafrost at Inuvik might be as low as —5°C and farther north in the area with pingos the temperatures might be as low as —6°C or —7°C.

75. Mean Lake Temperatures: The water in arctic lakes of relatively shallow depths, such as 5 to 20 feet, is probably isothermal both vertically and horizontally during most of the ice-free period of July to September. This has been observed for the lakes in the Point Barrow area, and judging from similar conditions, there should be no appreciable summer stratification in the Mackenzie Delta area. If conditions similar to the Point Barrow area are applicable to the Mackenzie Delta region, mean monthly water temperatures for the ice-free months of July to September might be expected to range from slightly over 0°C to 10°C or more. As the water temperatures in an unfrozen pool must be 0°C or above during the winter months, the mean annual water temperature of an unfrozen pool can hardly be less than about 2°C.

76. Three-Dimensional Heat Conduction beneath Lakes: In view of the paucity of data on permafrost beneath lakes, a theoretical approach may give information on permafrost surfaces. In the ensuing discussion, it should be obvious that certain simplifying assumptions must be made to keep the study within bounds. Theoretical aspects of the general problem of three-dimensional heat conduction in a semi-infinite medium, disturbed by surface effects, are treated by Carslaw and Jaeger (1960, pp. 255-296), Birch (1950), Lachenbruch (1957; 1959) and others. In particular, Lachenbruch (1957) has discussed theoretical aspects of three-dimensional heat conduction in permafrost beneath heated buildings. The same theory may be applied to lakes, by treating them as heated buildings, and thus the temperature disturbances in the ground beneath and around them may be estimated. In Lachenbruch's study, the ground is assumed to be homogeneous, although the effects of periodic heat flow in a stratified medium have also been examined (Lachenbruch, 1959). The effects of latent heat are also neglected, but this has no relevancy if only steady state or equilibrium conditions are being investigated. Seasonal disturbances may also be estimated but may be omitted in a study of pingo formation.

The symbols, given below, are mainly based upon Lachenbruch (1957):

- θ temperature in °C
- t time since initiation of temperature disturbance in lake U
- α thermal diffusivity (cm^2 per sec)
- U finite region of circular lake of plane $z = 0$
- T_G mean annual temperature ($T_G < 0$) of undisturbed permafrost region outside U of plane $z = 0$
- T_L mean annual temperature ($T_L > 0$) of the lake (U)
- T temperature under the centre of a circular lake
- B constant temperature difference between a point inside the disturbed region (the lake, U) of plane $z = 0$ and the mean annual ground temperature of the undisturbed region outside the lake; i.e. $B = T_L - T_G$
- I geothermal gradient in cm per °C
- R radius of circular lake

The following equations, with original references, are discussed in greater detail in Mackay (1962).

The principal disturbance (θ) induced under the vertex of a circular lake "placed" on the ground may be shown to be (Lachenbruch, 1957, equation 30)

$$\theta\,(0,0,z,t) = B\left[\operatorname{erfc}\frac{z}{2\sqrt{\alpha t}} - \frac{z}{\sqrt{z^2+R^2}}\operatorname{erfc}\frac{\sqrt{z^2+R^2}}{2\sqrt{\alpha t}}\right] \tag{76-1}$$

Under steady state or equilibrium conditions, equation 76-1 becomes

$$\theta\,\underset{t\to\infty}{(0,0,z,t)} = B\left[1 - \frac{z}{\sqrt{z^2+R^2}}\right] = (T_L - T_G)\left[1 - \frac{z}{\sqrt{z^2+R^2}}\right] \tag{76-2}$$

It is interesting to note that equation 76-2 is completely independent of the physical constants of the ground, such as the thermal diffusivity and water content, so long as steady state conditions exist.

The temperature (T) under the centre of a circular lake, including geothermal effects is

$$T = T_G + (T_L - T_G)\left[1 - \frac{z}{\sqrt{z^2+R^2}}\right] + \frac{z}{I} \qquad (76\text{-}3)$$

The principal disturbance (θ) at depth z beneath the centre of a lake with a winter frozen annulus of width R_2—R_1 and an unfrozen central portion of radius R_1 is

$$\theta\,(0,0,z,t) = B\left[\operatorname{erfc}\frac{z}{2\sqrt{\alpha t}} - \frac{z}{\sqrt{z^2+R_1}}\operatorname{erfc}\frac{\sqrt{z^2+R_1}}{2\sqrt{\alpha t}}\right] +$$

$$+\frac{B}{2}\left[\frac{z}{\sqrt{z^2+R_1}}\operatorname{erfc}\frac{\sqrt{z^2+R_1}}{2\sqrt{\alpha t}} - \frac{z}{\sqrt{z^2+R_2}}\operatorname{erfc}\frac{\sqrt{z^2+R_2}}{2\sqrt{\alpha t}}\right] \qquad (76\text{-}4)$$

Under steady state conditions, equation 76-4 becomes:

$$\theta(0,0,z,t)_{t\to\infty} = \frac{B}{2}\left[2 - \frac{z}{\sqrt{z^2+R_1}} - \frac{z}{\sqrt{z^2+R_2}}\right] \qquad (76\text{-}5)$$

The ground temperature (including geothermal effects) under the centre of the lake is then (*see* equation 76-3)

$$T = T_G + \frac{(T_L - T_G)}{2}\left[2 - \frac{z}{\sqrt{z^2+R_1}} - \frac{z}{\sqrt{z^2+R_2}}\right] + \frac{z}{I} \qquad (76\text{-}6)$$

If there is a permafrost surface beneath the centre of the lake—in which case there must be an upper and lower surface—then the temperature is 0°C. Therefore, a permafrost surface occurs at a depth z where the sum of the undisturbed temperature and that of the principal disturbance induced by the lake is zero, i.e., T in equation 76-3 is zero

$$(T_L - T_G)\left[1 - \frac{z}{\sqrt{z^2+R^2}}\right] + T_G + \frac{z}{I} = 0 \qquad (76\text{-}7)$$

Equation 76-7 is not a single valued function of z; therefore it is unsuited for direct determination of the top and bottom of permafrost and the equation may not be satisfied by a real value of z, as would be the case in a lake too large for permafrost to occur beneath the centre under the specified conditions.

It is evident that as a lake increases in radius from small to large, a critical radius will be reached beyond which no permafrost will be present beneath the lake centre. The radius of a lake with the top (bottom) of permafrost at specified z is obtainable from equation 76-7 by rearranging terms

$$R = \left[\frac{(IBz)^2}{(IB + IT_G + z)^2} - z^2 \right]^{\frac{1}{2}} \qquad (76\text{-}8)$$

If permafrost exists beneath the centre of a lake, and the lake radius is gradually increased, with steady state conditions prevailing, a critical radius will be reached when the permafrost lens beneath the lake "opens up", so that all the material directly beneath the lake centre is above 0°C. The critical radius is reached when

$$z = I\,[(B^2T_L)^{\frac{1}{3}} - T_L] \qquad (76\text{-}9)$$

Equation 76-9 gives the maximum depth to permafrost, at the critical radius where permafrost "opens" beneath the centre of a lake under steady state conditions. It is interesting to note that neither R nor the physical constants of the soil enter into equation 76-9. The maximum depth to permafrost depends only upon the interrelationships among the geothermal gradient, mean temperature of the lake, and mean temperature of the undisturbed ground. Once z is determined from equation 76-9, substitution into equation 76-8 gives the required radius of the lake under steady state conditions.

Figures 31, 32, and 33 have been computed for equation 76-5 for depths (z) of 100, 200, and 300 metres for various combinations of R_1 and R_2, the isolines being in units of B. (Note: so far as accuracy is concerned, metres can be read as yards). Thus, in Figure 31, using R_1 of 150 metres and R_2 of 200 metres, the disturbance at a depth of 100 metres is about .50 B; from Figure 32 at 200 metres, about .25 B; and from Figure 33 at 300 metres about .135 B. If $B=10°C$, $T_G = -7°C$, $T_L = +3°C$, are used, the disturbances at 100, 200, and 300 metres are about 5.0°C, 2.5°C, and 1.35°C. Figures 31, 32, and 33 make it possible to estimate steady state disturbances beneath lakes of varying sizes with varying seasonally frozen annuli. Geothermal effects can be added to the warming disturbance of the lake to obtain an estimated ground temperature.

Because information on the ground temperatures at the time of glacial retreat is unknown, it seems pointless to attempt to analyze transient effects in detail. For example, the ground surface beneath the ice may have been unfrozen, or it may have been frozen to a depth of tens of yards. If the ground were unfrozen, then freezing would have to be considered; if frozen, then thawing beneath a deep lake would take place. Latent heat effects would also be involved, its role being to delay thawing or freezing, depending upon temperature conditions. An estimate for deglaciation of the pingo area from a radiocarbon date is placed at over 12,000 years and probably near 15,000 years ago (Müller, 1962). Although the dates of formation of the pingos are unknown, available evidence, to be presented later, suggests that at least 5,000 to 10,000 years may have elapsed after deglaciation before many of the pingos were formed. The period is too short for steady state

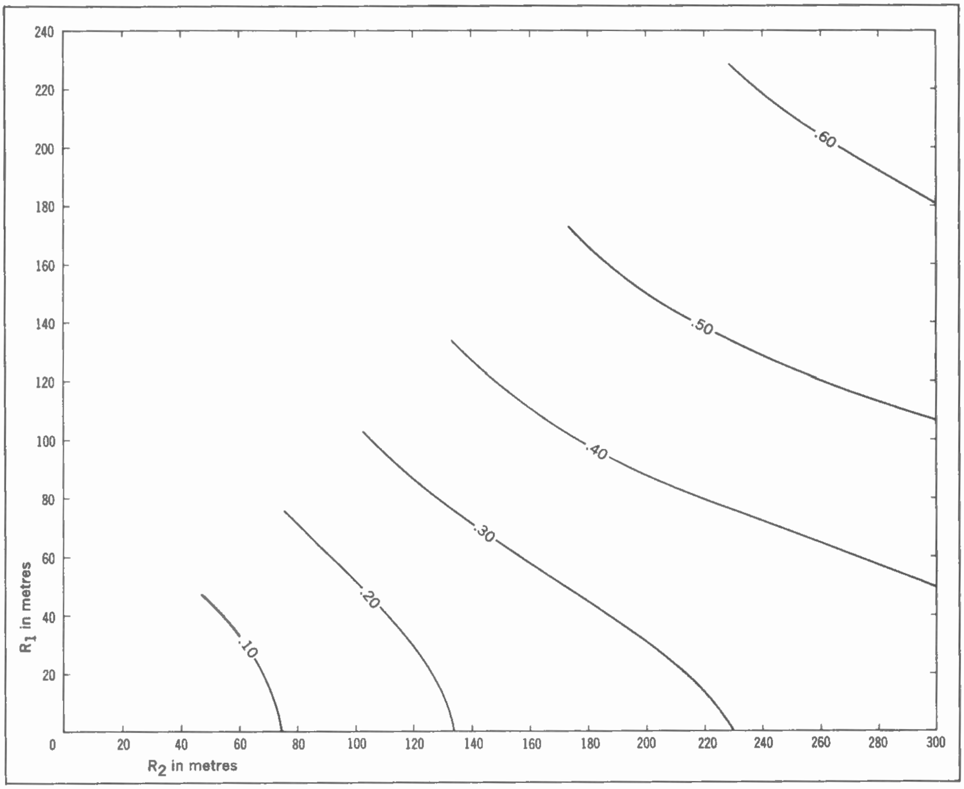

Figure 31. Nomogram for the disturbance, at a depth of 100 metres beneath the vertex of a circular lake expressed in units of B. In this figure, and also in 32 and 33, yards can freely be substituted for metres without loss of accuracy

conditions to pertain, even assuming a constant environment, but it seems long enough for the depths to permafrost to approximate equilibrium conditions closely enough that the geomorphic discussion of pingo formation is not affected.

77. Permafrost Aggradation in Unconsolidated Lake Bottom Sediments: Pingos are believed to grow as a consequence of the penetration of permafrost into unconsolidated lake bottom sediments. The rapidity of downward penetration of the freezing plane, which is the lower permafrost surface, will depend upon such factors as the climate, physical properties of the sediments, topography, vegetation cover, presence of nearby water bodies, and so forth. In the present discussion, only the thickness of bottom sediment corresponding to the cover over pingo ice-cores will be considered.

Prior to downward penetration of permafrost in unfrozen lake bottom sediments, all the sediments would be saturated. Therefore, the latent heat of fusion would be important in delaying the downward penetration of the freezing plane, although it has no effect on the equilibrium temperature profile. To simplify the problem, let us first assume that the unfrozen sediment, at the instant that perma-

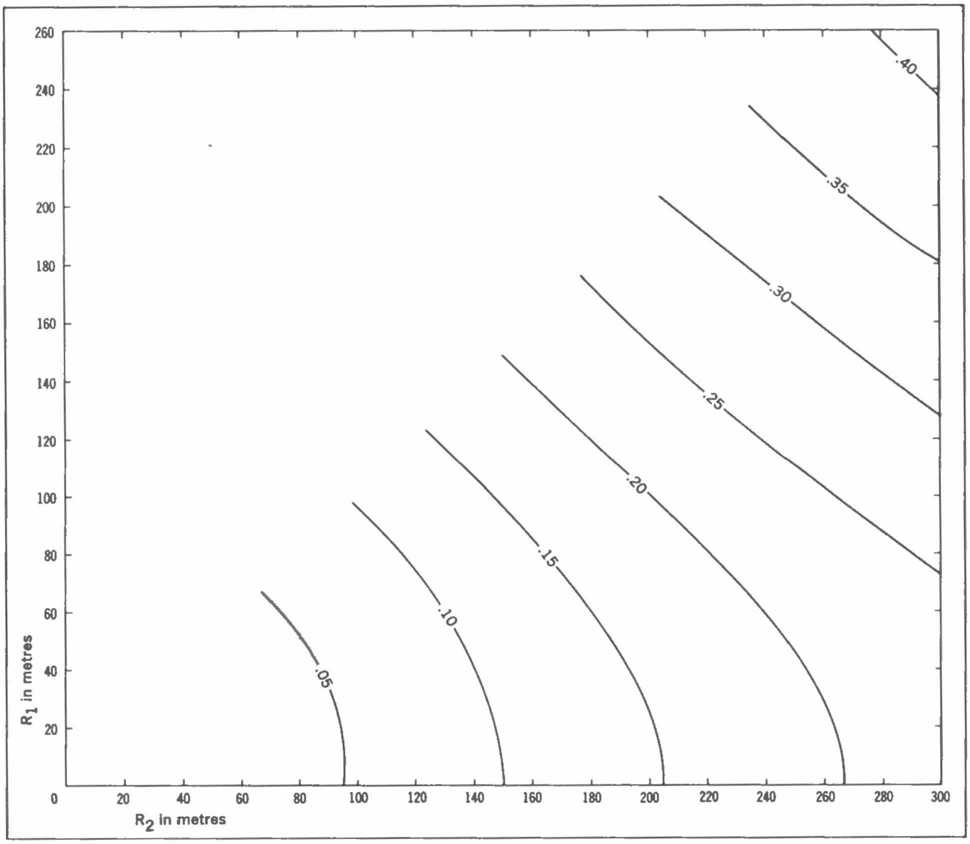

Figure 32. Nomogram for the disturbance, at a depth of 200 metres beneath the vertex of a circular lake expressed in units of B.

frost begins to develop in the lake centre, is at 0°C for a depth of 10 or 15 yards. This does not seem to be an unreasonable assumption, because the mean bottom temperature for a depth of 10 or 15 yards would normally be close to the mean lake water temperature which, at commencement of permafrost, would be about 0°C.

If the preceding assumptions are valid, then the general approach to the rate of freezing may be derived from theoretical studies involving change of state in the formation of ice, viz. Stefan's and Neumann's solutions (Carslaw and Jaeger, 1960, pp. 282-286; Ingersoll and others, 1954, pp. 190-199; See also Redozubov, 1946). The approach used below is adapted from that of Terzaghi (1952, pp. 39-44) for the rate of thawing beneath heated buildings. All units are in the cgs system:

T_T mean surface temperature at the top of permafrost

η porosity of the frozen ground

L 80 cal gm^{-1}, heat of fusion of ice

k_s average thermal conductivity of frozen ground

k_i 5.3×10^{-3} cal cm^{-1} sec^{-1} (°C)$^{-1}$, average thermal conductivity of ice

The thickness (E) of frozen soil may be shown to be about

$$E \backsim \sqrt{\frac{-2T_T k_s t}{72\eta}} \tag{77-1}$$

The thermal conductivity of frozen soil is highly variable, but ranges are from about 3.0×10^{-3} for frozen organic silty clay, 5.0×10^{-3} for a saturated frozen clay, 5.3×10^{-3} for pure ice, 6.0×10^{-3} for an icy silt, to 9.0×10^{-3} cgs units for a dense frozen sand (Lachenbruch, 1959, p. 10; Terzaghi, 1952, p. 9). As most of the material above the pingo is sand, the soil may range from about 6.0×10^{-3} to 9.0×10^{-3} cgs units.

From equation 77-1, it is evident that the depths of penetration of permafrost in two soils for the same time and temperature are about proportional to the square root of the conductivity divided by porosity for each soil. A substitution

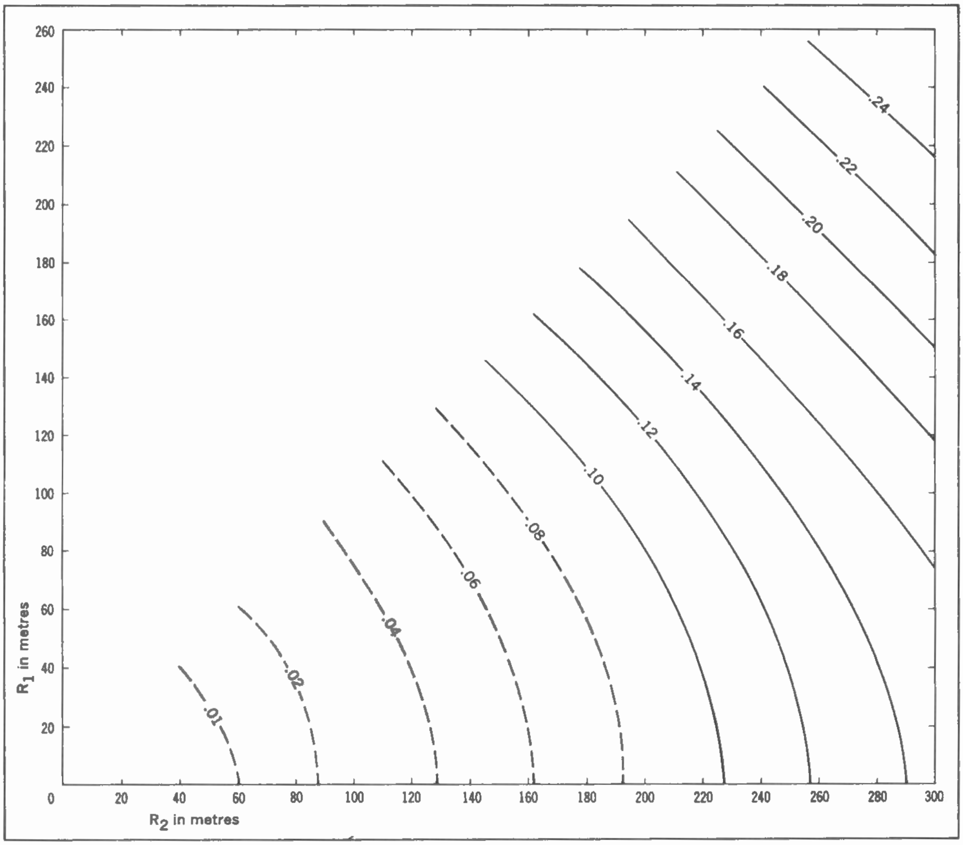

Figure 33. Nomogram for the disturbance, at a depth of 300 metres beneath the vertex of a circular lake, expressed in units of B.

of the range of values which might be expected for ice and frozen soil indicates that the thickness of ice (Eɪ) which might develop would be from about one-third to two-thirds that of frozen sand (Es) under similar temperature and time conditions and a saturated soil with a high porosity would freeze much more slowly than a soil with low porosity.

Under actual field conditions, three factors would seem to operate in combination to produce a dome-shaped lower permafrost surface at the site of pingo growth during the aggradation of permafrost in a lake basin. Firstly: if in an otherwise uniform, flat, homogeneous lake bottom area, ice formation commenced at one spot during the downward penetration of permafrost, the penetration of the lower permafrost surface would be slowed at the site of ice segregation so that a dome-shaped lower surface would tend to result. Secondly: most lakes are deepest towards the centre, where pingo growth normally occurs, and shallowest near the shores. Thus, if the water gradually shoals, and permafrost begins to aggrade in an unfrozen lake basin, permafrost will have had the longest time to grow near the shore, the shortest in the centre, thus contributing towards a dome-shaped lower surface. Thirdly: a pingo protruding above the lake bottom alters the original flat lake bottom topography. The exact effects are impossible to specify, since ground temperatures, position of the freezing plane, size of pingo, snow cover, etc., are involved. However, in general, the formation of a mound would be expected to result in an upward bending of isotherm surfaces beneath the mound.

78. Hydrostatic Pressure: The generally accepted theories for pingo formation depend on updoming by hydrostatic pressure. The hydrostatic pressure may result from a hydraulic head associated with slopes (East Greenland type of pingo) or pressure resulting from volume expansion of water on freezing as permafrost advances into a closed unfrozen pocket (Mackenzie type). Laboratory and field experiments carried out on the freezing of saturated sands show that excess pore water will be squeezed out if the permeability exceeds about 5 inches per day, provided the surplus water is free to escape (Balduzzi, 1959; *Foundations of Geocryology,* Vol. 11, 1959, pp. 33-35; Kosmachev, 1953; Petrov, 1934). Pressures may attain tens of atmospheres.

If V is the volume of unfrozen material in a closed system and η the average porosity, then the volume of expelled pore water (Ve) cannot exceed about a tenth the volume of ice.

$$Ve < 0.1\,\eta\,V$$

79. Development of a Closed System: The development of a closed system whereby expelled pore water is trapped under pressure would seem to require only the growth of a continuous permafrost seal on the lake bottom (Figure 34B). Expelled pore water could not escape upward through the impermeable permafrost seal; it could not go sideways because of permafrost extending out from the shore beneath the unfrozen sediments; and it could not escape downwards, because of either the presence of permafrost, saturated soils, impermeable sediments, or a combination of them.

80. Origin: Pingos are believed to have formed primarily in lake basins with unfrozen central portions (Figure 34A) which subsequently acquired a surface layer of permafrost to form a closed system. The initiation of permafrost along a lake bottom would seem to require either exposure of the bottom to air, as by drainage, or freezing of winter ice to the bottom for a sufficient duration each year so that mean bottom temperatures were below 0°C (Figure 34B).

Exposure of a lake bottom to air temperatures has usually resulted from draining due to stream erosion or indirectly from coastal recession. Freezing of ice to the bottom could be initiated in many ways, besides drainage, such as from a climate change with lessened precipitation and greater evaporation or from infilling. The process of infilling in causing the formation of permafrost is believed to be relatively unimportant (*see* Porsild, 1938; Müller, 1959, p. 99).

For reasons previously given, a new permafrost cover beneath a lake bottom would tend to be thinnest in the centre, with a dome-shaped lower permafrost surface (Figure 34B). With downward aggradation of the permafrost cover, and a rise of the permafrost surface beneath the unfrozen sediment, water would tend to be expelled from pores as a result of volume expansion due to freezing. Thus, hydrostatic pressure is believed to have been relieved by uplift of the permafrost layer where it was thinnest, namely at the top of the dome-shaped permafrost surface. If such were the case, then the thickness of the overburden over the ice-cores would indicate the approximate thickness of permafrost when pingo growth began.

In the larger lake basins, the central portions could be completely unfrozen at all depths, although permafrost would surround the unfrozen centre (Figure 34A). If a permafrost surface cover were to develop, the unfrozen portion would be shaped like an hourglass. Inward migration of the permafrost neck of the hourglass could produce two closed systems. Therefore, pore water could be expelled from a number of "different" permafrost surfaces.

Lakes with irregular bottoms might have more than one deep pool, so that more than one dome-shaped lower permafrost surface might develop. Multiple pingos are believed to have grown in such lakes. Likewise, if the deeper pools were off-centre, the pingos would also be off-centre.

As shown in the theoretical discussion of permafrost, the rate of formation might be taken as proportional to the square root of time (equation 77-1). Thus the rate of downward aggradation of permafrost would be relatively rapid when permafrost was thin and the temperature gradient high, but it would be relatively slow, when permafrost had extended down to greater depths. Therefore the expulsion of pore water, from downward freezing, would gradually slow down from this cause alone. In addition, the volume of unfrozen sediment would normally decrease with depth, so that equal increments of permafrost growth might not contribute equal amounts of expelled pore water.

Pingo ice-cores seem, in general, to be of nearly pure ice, quite unlike the "dirty" tabular ice sheets which may occur nearby. The implication is that pingo ice forms from the freezing of a "pool" of water. However, there is no reason to believe that pools of water of the sizes of the ice-cores ever existed at one time,

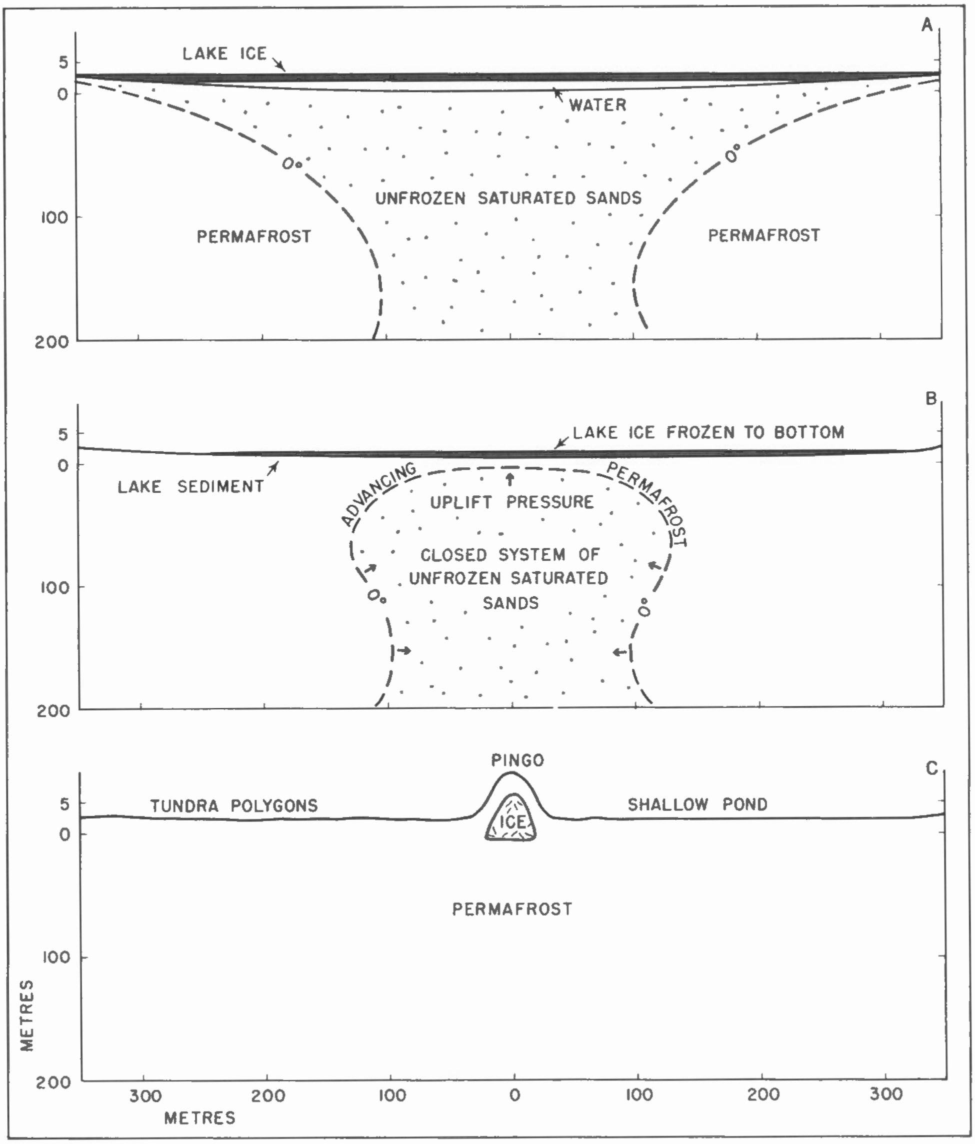

Figure 34. Schematic origin of a pingo. In diagrams A, B, and C a vertical exaggeration of 5X has been used for the height above zero in order to show the lake ice and the open pool of water. A—a broad shallow lake has an open pool of water in winter with a frozen annulus around it. No permafrost lies beneath the centre of the lake. B—prolonged shoaling has caused the lake ice to freeze to the bottom in winter and induced downward aggradation of permafrost. Infilling has raised the lake bottom a small distance. The deepest part of the lake, which has the thinnest permafrost, is gradually domed up to relieve the hydrostatic pressure. C—the pingo ice-core, being within permafrost, is a stable feature. The old lake bottom is occupied by tundra polygons and shallow ponds. Because of scale changes in the diagram, the volume of the ice-core should not be construed as showing a direct relationship to the initial volume of unfrozen material.

except, perhaps, in the smallest pingos. Rather, pingo ice-cores are believed to have grown slowly over many years or decades by the constant freezing of a replenished "pool".

The rate of pingo growth cannot be rapid, except in very special cases. There are reports from Russia (Suslov, 1961, p. 144) of a growth of 5 to 7 feet for small pingos in 20 years with larger pingos reputed to develop in several dozen years, but no verifiable reports are available for the Mackenzie area. However, a consideration of the volume of pingo ice can lead to an estimate of the volume of unfrozen material required to produce the pingo, and hence to the downward penetration of permafrost. For example, if the ice-core in Ibyuk pingo is assumed to be a right cone about 40 yards high with a base 70 yards in radius, the volume would be approximately 200,000 cubic yards. If this represented a 10 per cent volume expansion of soil with 30 per cent porosity, the required volume of unfrozen material would have been about 7,000,000 cubic yards. Let us further assume that the unfrozen material was in a circular lake and that it was in the shape of a right cone with a slope of 45°. If so, the radius and "depth" of the cone would have been 190 yards. The freezing of such an unfrozen cone would have taken many decades. If the radius of the cone was taken as 300 yards, the "depth" of the cone would have been 75 yards. Again, freezing to such a depth, to which should be added the frozen layer which existed prior to the development of doming, would be long.

81. Age: The great majority of the pingos are probably hundreds, if not thousands of years old. As long as permafrost is thicker than the height of a pingo, and the surface cover remains intact, a pingo should be a relatively permanent fixture of the postglacial landscape. Under present climatic conditions, a pingo should collapse only through breaching of the protective overburden. Although the Eskimo names for some pingos suggest growth, as "the one that is growing" or "the poor thing that is getting to be a pingo" (Porsild, 1938, p. 52) there are, so far as is known, no reliable accounts of pingo growth either witnessed by Eskimos, or traceable back to a specific period. The pingo appearing in Richardson's book of 1851 appears unchanged today.

The vegetation cover of pingos with humus, old willows, and other plants attests to an age of at least several hundred years. The accumulation of humus is very slow, so that the presence of a turf mat at the surface may indicate an age of several thousand years. Tundra polygons, with ice-wedges exceeding 6 feet in width, are common features of large and small pingos. In general, the pattern of tundra polygons is continuous from the pingo to the surrounding area, showing that formation of the polygonal pattern had taken place prior to or concomitant with pingo growth. Studies of ice-wedge growth suggest a rate of about 1.5 to 3 feet per thousand years (Black, 1952) for the Point Barrow area and if rates of growth in the Mackenzie area are approximately the same, the widths of the larger ice-wedges point to ages of at least several thousand years. Some pingos are surrounded by high-centered peaty polygons whose peat feathers out against the pingo sides. The feathering-out of the peat, as shown in sectioned pingos, suggests that the peat accumulated after the pingo had grown. Two radiocarbon dates for peat deposits, one near Inuvik (GSC-25) and the other for the mid-Eskimo Lakes area (GSC-16) give an average rate of growth of about 1 to 1.5 feet per thousand years. On this basis, a typical sectioned pingo in the Eskimo Lakes (69° 32′ N, 131° 41′ W) is three to five thousand years old. Similar estimates can be made for many other pingos. Along the Arctic coast and along that of the Eskimo Lakes, coastal recession has cut back into numerous lakes which contain pingos. As the pingos must have formed prior to cliff recession, which may amount to several thousand feet, again an old pingo age is suggested. Müller (1962) places the minimum ages of two pingos near Tuktoyaktuk at 7,000 and 4,000 years.

The assembled evidence from vegetation types, humus thickness, widths of ice-wedges, peat accumulation, and cliff recession all suggest an age of a few thousand years for the majority of the pingos. Such an age would be compatible with a period of cooling following the postglacial thermal maximum, or even earlier growth. To the east, at 64° 19′ N, 102° 41′ W, Craig (1959) has described a pingo whose formation may have been due to the marked cooling of climate following the postglacial thermal maximum. The growth of some pingos in the U.S.S.R. has also been ascribed to the same period of cooling (Grave, 1956). Although pingos in the Pleistocene Mackenzie Delta area might be growing today, all the evidence indicates that the great majority of the pingos have been stable landscape features for many hundreds of years.

Pingos of the Modern Mackenzie Delta

82. The low seaward islands of the modern Mackenzie Delta have several concentrations of small pingo hillocks and pingo ridges. These hillocks and ridges are of considerable geomorphic interest, because: they superficially resemble the Mississippi delta mud lumps, which have long been a unique geomorphic feature unreported from any other place in the world (Morgan, 1952); they occur to the west of possibly the world's greatest concentration of pingos; they have formed in a region of aggrading permafrost; and the islands on which they have grown are inundated yearly or every few years.

83. Distribution: The pingos lie within 15 miles of the sea in a belt between Richards Island and the mainland to the southwest (Figure 29). More than 80 pingos have been plotted from air photographs, with 40 having been seen in the field. The outlines of many wave-eroded pingos, horizontally sectioned flush with the surrounding island surface, are detectable on air photographs. Therefore, the total number of pingo forms exceeds one hundred.

The pingos north of Shallow Bay are the most numerous, numbering over 70. They occur in three main groups: the first on the distal alluvial islands which form the northwest side of Shallow Bay; the second on Ellice Island; and the third in a nameless area centred at 69° 09′ N, 135° 20′ W. The pingos are on low alluvial islands or shoals below flood level. Flooding may take place at break-up, or when water levels are raised during violent westerly storms. The islands which are mainly within several feet of sea level, are so monotonously flat that even an incipient pingo, 5 feet high, looms like a hill.

The 10 or more pingos south of Shallow Bay are in an older part of the modern delta where sedimentation is less active, floodplains are higher, and bush vegetation (willow-alder) is much thicker than in the alluvial islands to the north (Figure 70).

84. Occurrence in Depressions: Unlike the pingos of the Pleistocene Mackenzie Delta area, the hillock type of pingo has little direct association with existing lake basins. This may arise, in part, because the lakes are so shallow and ephemeral that they are liable to rapid infilling by sedimentation or destruction by drainage. Therefore, if a pingo were initially in a basin, there might be no such recognizable relationship today. However, pingos have grown sufficiently often in lake basins to indicate that such sites are favorable for pingo generation.

There is a close agreement between the ridge type of pingo and drainage channels. On air photographs, the ridge type of pingo clearly occupies a drainage channel, although the agreement is less obvious in the field. Usually, the trace of an infilled channel will occur at one or both ends of a ridge. The linear alignments of series of slightly elongated pingos also show growth in abandoned channels.

A few pingos are on islands which are so flat and low that the base of the active layer is below sea level. Such low islands have no lakes. Some pingos may occur on shoals, below normal high tide level, but this opinion is subject to verification as the water levels could not be observed long enough to determine if the areas were truly shoals.

85. Shape: The hillock type of pingo is up to 25 feet in height and 200 feet in diameter. The typical outline is oval to circular. The larger pingos tend to have ruptured or furrowed summits. A few pingos are surrounded by a gentle swale or moat, which may be water-filled.

The ridge type of pingo varies from a few tens of feet in length to well over 1,000 feet. The longest ridge pingo is about 2,000 feet long and up to 25 feet high. The ridge crest is undulating, with a swale running lengthwise along the crest, as if it represents the beginnings of a longitudinal rupture. The ridge averages 150 to 200 feet in width, with slopes of 20 to 30 degrees. It curves slightly. A single line of alders, locally broadening to a width of 20 feet, parallels the base of the ridge.

86. Type of Sediments: The surface material of the islands—where examined—to a depth of several feet, is predominantly in the very fine sand range with only a few per cent of either coarser sand or finer silt. Stones are absent. The sediments are deposited at flood time when the islands are completely inundated except for the pingos which protrude as islands in a shallow sea 20 or 30 miles from the nearest dry land. The distributary channels, between the islands, tend to have finer-grained sediments with muddy pockets. One sample tested had about 50 per cent in the silt-clay range.

The upper several feet of the pingos examined in the field were of a very fine sand to silt often with peaty partings; that is, of a finer grain size than the surface material of the adjacent areas. Material from a depth of 10 to 20 feet has been raised to the surface, by pingo growth, and frequently exposed by wave cutting. It is difficult to estimate the precise depth from which material has been raised, because the beds exposed near the surface have been disturbed by frost heaving. Nevertheless, at a depth of 10 to 20 feet the beds appear to be of interbedded peat and sands. A section through a bisected pingo (Figures 35 and 36) was as follows: the top of the pingo was 11 feet above low water level or 8 feet above the general level of the island. The top several feet of the pingo was composed of a very fine grey sand, in places slightly silty; the material was somewhat finer than that of the island surface. The horizontal section was exceptionally good. Concentric semi-circles of interstratified peat and fine grey sand enclosed a central mud pool of silty clay. The structure was that which might be expected of a partially sectioned dome. The beds of peat and silty-clay stood out in relief as miniature cuestas each from 1 to 5 inches high and dipping towards the periphery. As the surface dips were, in general, steeper than those at a depth of 1 to 2 feet, possibly frost heaving had accentuated the dips near the surface. Radial faults, transverse to the strike, were spaced every 10 to 20 feet. Displacements usually amounted to only a few inches. The peat beds are of a type now forming in sedgy areas where driftwood fragments accumulate. At location 1, Figure 35, a sample tested: 10 per cent fine to coarse sand, 70 per cent very fine sand, and 20 per cent silt. At location 2, about 5 per cent was fine to coarse sand, 10 per cent very fine sand, and 85 per cent silt to clay; at location 3, about 5 per cent was fine to coarse sand, 25 per cent very fine sand, and 70 per cent silt. The grain size distribution shows that the soils have sufficient fines to be susceptible to ice lensing (Balduzzi, 1959). Sample 1 contained the ostracod genus *Candona,* indicating fresh or only slightly brackish water; a few unidentifiable minute shell fragments were found in sample 2, and no microfossils in sample 3 (analyses by Dr. F. J. E. Wagner, Geological Survey of Canada). The material exposed in the horizontal section differs from that of the alluvial island surface in having both peaty and silty clay beds. The formation of peaty beds would require a cover of ground vegetation which is lacking on the island at present. Similar peaty beds are now being formed only on higher alluvial islands with a growth of sedges, grasses, horsetails, etc.; for example, on islands rising about 5 to 7 feet above low tide level in the area immediately to the south, across Shallow Bay. The cut bluffs of these islands are composed of a very fine sand with silty partings interstratified with thin peaty beds of grasses, sedges, horsetails, and a few willow leaves. Nowhere was silty clay observed similar to that in the mud pool. The only silty clay encountered in the general region was in sea bottom muddy patches from which specimens were picked up with the sounding lead.

If the lowest peaty beds were returned to their original positions, they would lie 10 to 20 feet below the island surface, or about 7 to 17 feet below sea level. As the peat beds probably accumulated at least several feet above sea level of the time, a submergence of roughly 10 to 20 feet, or compaction of the sediments, is suggested (*51*).

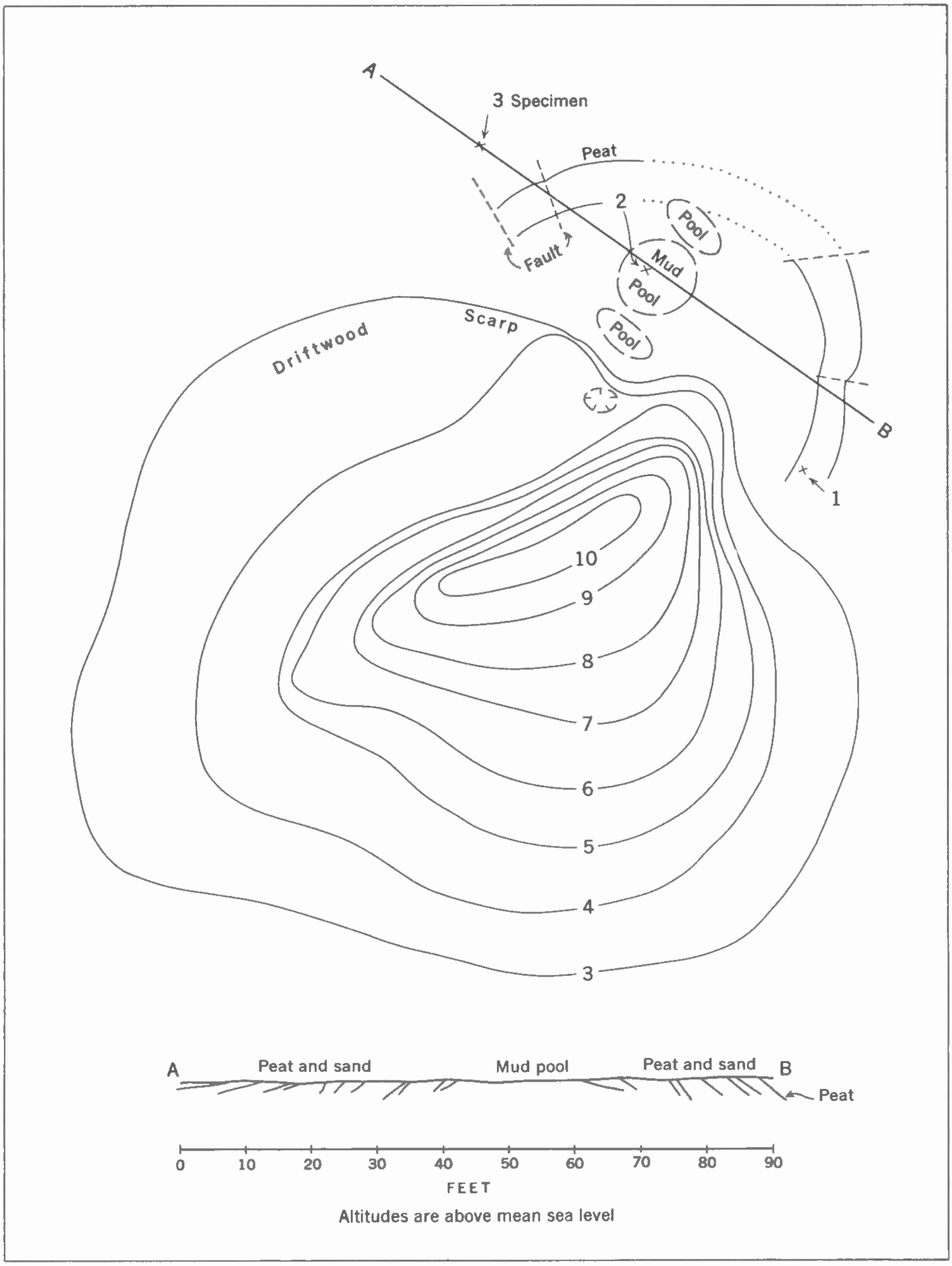

Figure 35. Partially sectioned pingo in the Mackenzie Delta at 68°55′N, 135°57′W, surveyed June 19, 1958. The area along the line of section A-B has been wave-eroded. Only the major faults are shown.

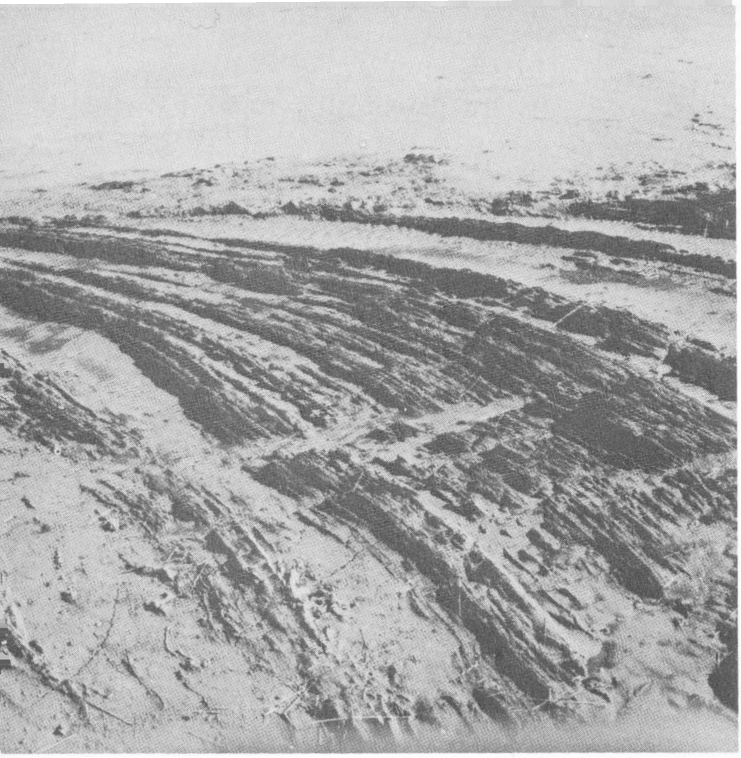

Figure 36

Section through pingo shown in Figure 35. Note the tilted beds. August 27, 1957. (Geog. Br. photo).

87. Origin: On first examination, the pingos might be considered similar to the Mississippi mud lumps. The formation of a Mississippi mud lump is attributed to the static load—on underlying plastic clay—of bar sediments that accumulate most rapidly at flood periods; clay is squeezed laterally and locally breaks through upwards to form the mud lumps (Hilgard, 1871, 1906; Morgan, 1952; Shaw, 1913). The Mississippi mud lumps have been observed rising as islands in relatively deep water. The dimensions and structures of the Mississippi mud lumps are similar to those of the hillock type of pingo. In a way, the mud lumps are formed somewhat like the East Greenland type of pingo with the movement of plastic mud taking the place of water under hydraulic gradient.

Unlike the Mississippi mud lumps, the pingos have grown in islands or shoals and not in association with deep water bar deposits. Thus, the basic mechanism for developing a differential geostatic load is missing with the pingos. Moreover, the pingos have grown in an area where permafrost is inferred to be aggrading downwards, ab initio, and some pingos are known to have ice-cores. Therefore, the origin of the pingos differs in important respects from that of the Mississippi mud lumps.

No data are available on the thickness and extent of permafrost beneath the alluvial islands. However, there is no question of the presence of permafrost because of the widespread distribution of tundra polygons, ice-cored pingos, frozen ground at shallow depth in late summer, and so forth. Permafrost should be aggrading in all the larger young alluvial islands. Inasmuch as most of the distributary channels which meander among the islands are considerably shallower than the winter ice thickness, and shoals extend for hundreds of feet offshore, deep-water warming effects should be negligible. In mid-summer, frozen ground—

not necessarily permafrost—is often encountered in shallow channels within 2 feet of the bottom.

The pingos which are associated with lake or channel depressions may have formed in a manner analogous to those of the Pleistocene area. In an alluvial island where ground is being built above sea level, permafrost would aggrade downwards most rapidly in portions exposed to air temperatures. Shallow lakes and channels would be expected to have no permafrost or thin permafrost in the deeper parts. A channel only 200 feet wide and 4 feet deep in the centre might have enough thermal capacity to prevent freezing of the subjacent bottom sediments. Consequently, infilling or shoaling of lakes and channels should initiate a cover of permafrost on the lake or channel bottoms. The presence of very fine sand and silt in the overburden of many pingos suggests that infilling of lakes and abandoned channels by flooding has been the most important cause of shoaling, because the sediments are finer-grained than the island surfaces.

Once a permafrost surface formed on a lake or channel bottom, a closed system might be formed in one of two ways. Firstly, fine-grained soils, at a moderate depth (e.g., 20 to 30 feet) might have had sufficient impermeability to prevent the rapid escape of expelled pore water. Fine-grained soils, such as silty clays, are known to occur at depth and such soils may have had such a low coefficient of permeability as to have restricted the downward and lateral escape of pore water to the sea. Secondly, as permafrost grew adjacent to lakes and channels, permafrost would tend to extend inwards, at depth, beneath them. For example, given mean ground temperatures of about —5°C, permafrost should be able to extend, ab initio, beneath the smaller lakes and channels within a hundred years to provide an impermeable lower surface. After a closed system was formed, hydrostatic pressure could build the pingos similar to those of the Pleistocene delta area.

The smallness of the pingos contrasts with the much larger pingos of the Pleistocene areas. In a seemingly typical 15-foot-high pingo, the overburden over the ice-core was 5 feet thick. If the ice-core were conical, being 15 feet high and 15 feet in radius, an unfrozen cone about 10 feet deep and 100 feet in radius could have supplied sufficient expelled pore water on the basis of 30 per cent porosity and 10 per cent volume expansion, or an unfrozen cone 50 feet deep and 50 feet in radius would have sufficed. Even if the porosity and proportion of expelled water were far less than assumed, the volume of unfrozen sediment beneath shallow lakes and channels could reasonably be assumed compatible with the sizes of the pingos. As the pingos grew from expulsion of pore water, the adjacent island surface would, at the same time, undergo sedimentation. Thus the poor association between pingos and depressions may be explained as the result of sedimentation.

88. Age: The pingos range in age from young pingos, which may be in active growth, to stable pingos probably hundreds of years old. There is no transitional belt between the small pingos of the modern delta and the much larger pingos of the Pleistocene deposits. The pingos of the modern delta are much younger.

Some pingos are overgrown with vegetation and have willows in summit depressions exceeding 4 inches in diameter and small poplar on the sides. The ages of the shrubs are frequently over 50 years. On the other hand, some pingos of equal size have little to no vegetation, not even grasses and sedges. In 1825, Franklin camped at Ellice Island (69° 3′ 45″ N, 135° 44′ 57″ W) and saw three conical hills where there is, at present, a major cluster of pingos. The area was described as flat although there are now 20 pingos nearby. There is no assurance that the pingos Franklin saw exist today, but as the pingo groups are widely scattered, there seems little doubt that Franklin described pingos in an area where pingos now occur and have grown since his visit. Ellice Island was in existence over 150 years ago, so the island surface has been exposed to subaerial temperatures for at least that long. Although the area is periodically flooded and the region may be slowly subsiding, nevertheless, ground temperatures are probably sufficiently low for permafrost to have penetrated to a depth of tens of feet. Thus, the chances of new pingos growing in areas without depressions seem slight.

Many of the pingos are ephemeral and are either eroded before they reach full growth, or else are eroded soon afterwards. When flood waters inundate the islands, the bases of the pingos are exposed to wave attack. Partially sectioned pingos are numerous. The eroded portions are usually light-toned on air photographs because of differences in sediment and the general absence of vegetation. By using these features as criteria, many pingos, eroded to island level, can be recognized on air photographs.

CHAPTER III

REGIONAL PHYSICAL GEOGRAPHY

88. The Mackenzie Delta area possesses a broad variety of landscapes ranging from high mountains to flat alluvial plains. The contrasting scenery has resulted partly from differences in rock type, structure, and geomorphic evolution. The area has been subdivided into physiographic regions which have a certain degree of homogeneity among themselves (Figure 37). The boundaries are sharp between some regions, as, for example, that between the Mackenzie Delta and the Caribou Hills. Other boundaries are transitional.

The Mackenzie Delta physiographic region is discussed in the greatest detail, not only because of its intrinsic geomorphic interest, but also because more field work was carried out in the delta than elsewhere. Systematic topics, such as Mackenzie Delta channel geometry and floodplains, are reserved for analysis in this chapter on Regional Physical Geography rather than in Chapter II on Physical Geography because the features are peculiar to the delta and help in the understanding of its uniqueness. Some of the physiographic regions were not seen in the field, the analysis of them being based only on air photograph and map interpretation.

Western Uplands: Region I

89. The western uplands are part of the arctic mountain area of the eastern system of the Canadian Cordillera (Bostock, 1948, 1961). By the local inhabitants, the western uplands are referred to as the Richardson Mountains. The uplands, when seen from Aklavik, provide a welcome change to the delta scenery. The region is unimportant for minerals, although a small coal mine has operated intermittently on the northwest side of the delta. The coal has been taken by barge to Aklavik for local use, but shallow channels have made water transportation difficult.

The western uplands have very few lakes (Figure 39), partly because of the relief, but also because the area was barely touched by glaciation. There are scattered lakes in the piedmont (Ic) but the hilly areas are nearly lakeless. The higher areas (Ia and Ib) are tundra-covered, the piedmont has scrub tundra, with some trees in the southern part (Figure 70).

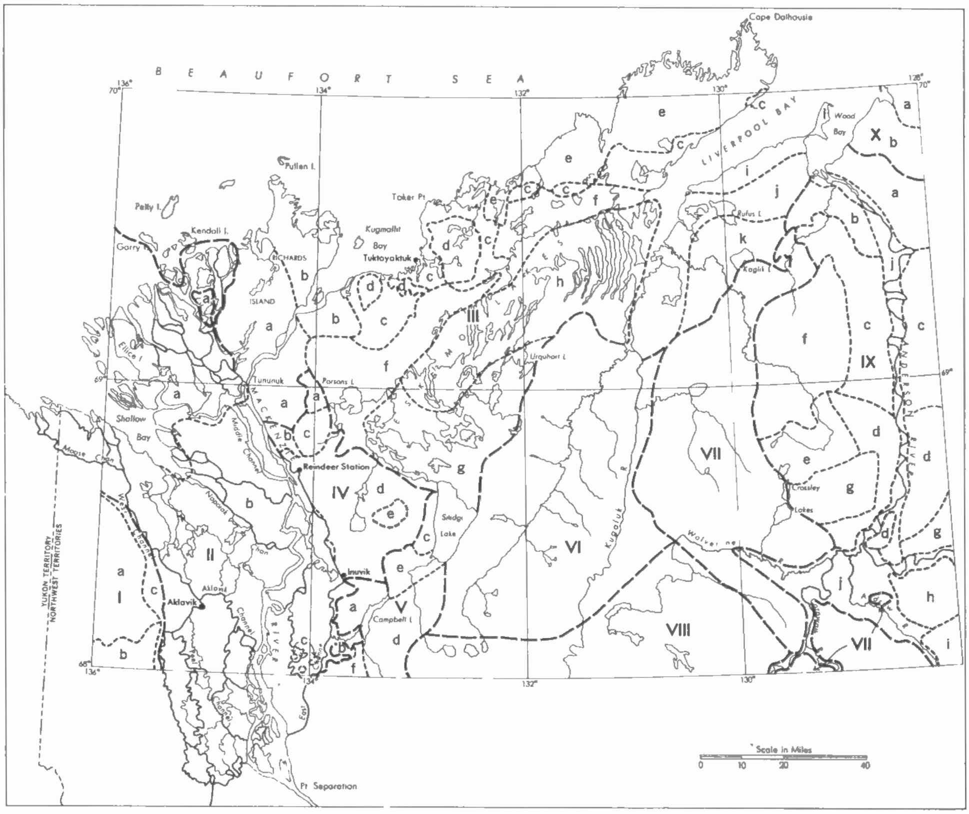

Figure 37. Physiographic regions, based mainly on photo interpretation.

The arctic plateau (Ia) blends in the north into a narrow coastal plain and in the south into the Aklavik range of the Richardson Mountains. Altitudes range from about 1,500 down to 1,000 feet. The sediments of the Richardson Mountains are largely Mesozoic (primarily Cretaceous) in age. The Richardson Mountains were uplifted in the widespread early Tertiary alpine orogeny which produced the Rocky Mountains to the south (Jeletzky, 1961; Martin, 1961). Although there had been earlier movements, the major structural features of the Richardson Mountains were formed at that time. Jeletzky (1958, 1961) has mapped in detail their eastern slope. The principal faults are northerly, northeasterly, and northwesterly, thus subdividing the eastern slope into fault blocks of varying size. Jeletzky suggests that some of the major faults of the eastern slope appear to be still active, citing as evidence epicenters of recent earthquakes.

The Aklavik Range (Ib) overlooks the Mackenzie Delta. Its northeastern extremity is marked by Mount Gifford (2,000 feet) which is locally called Red Mountain. Donna River, flowing around the north end of Aklavik Range, has cut a flat-floored valley about 500 feet deep. The Aklavik Range, from 68° 00′ N to 68° 11′ N, rises nearly 1,000 feet within a horizontal distance of one mile as a straight north-south scarp above the piedmont (Ic). The scarp (Figure 38) is so long, straight, and steep that a number of observers have interpreted it

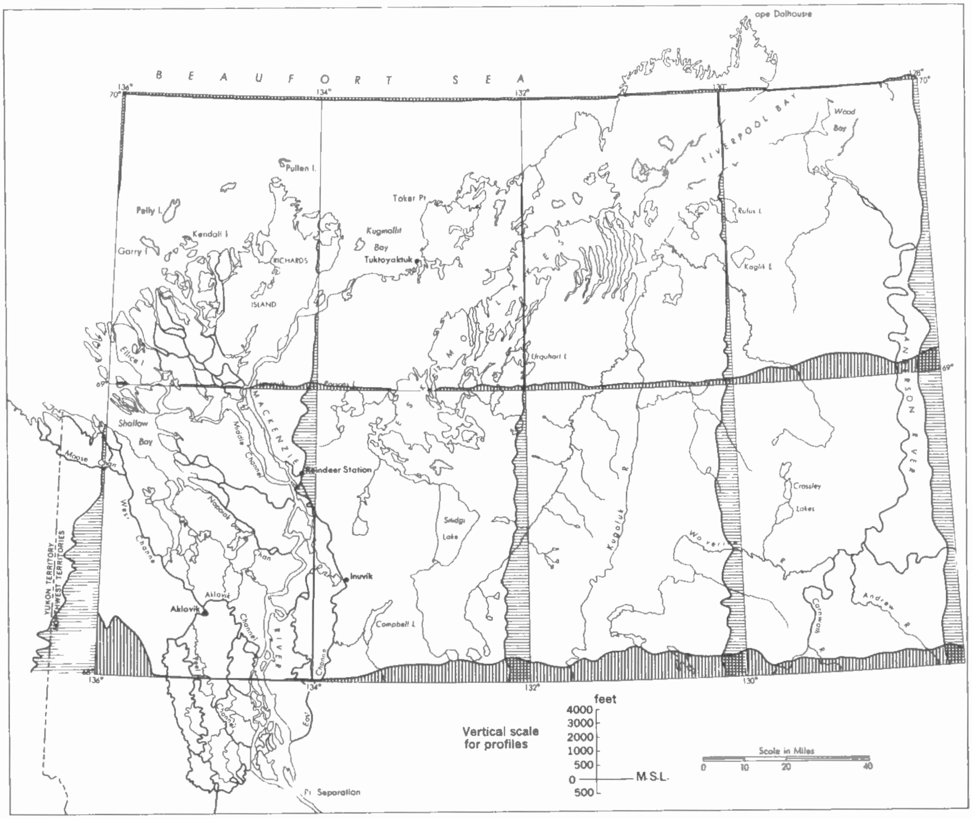

Figure 38. Profiles showing heights and depths along meridians and parallels. Note the scale change above 1,000 feet.

as a fault or fault-line scarp. However, the scarp does not coincide with any major fault or fold but cuts across numerous faults. Jeletzky (1961, p. 572) has interpreted the scarp as a postglacial feature, formed by uplift of the Richardson Mountains with downcutting by the prehistoric Husky Channel, which now flows on the west side of the Mackenzie Delta. However, there is ample evidence to show that the Richardson Mountains were in existence during the Pleistocene, for they formed a barrier to glacier movement and features associated with glaciation, such as till, kames, moraines, and melt-water channels abound on the eastern slopes. The straight eastern scarp of the Richardson Mountains is probably of erosional origin cut by the ancestral Mackenzie River. Glacier ice, moving north along the trough now occupied by the Mackenzie Delta, straightened and smoothed off the eastern scarp. Postglacial erosion has notched gorges back rapidly into the scarp face. Many streams have knick points near their headwaters with gentle gradients and tundra slopes above, steep gradients, gullies, and bare slopes below. Streams flowing from the Aklavik Range have deposited a series of coalescing alluvial fans with gentle gradients of 50 to 100 feet to the mile. In this area, south of 68° 11′ N, the piedmont is 1 to 2 miles wide. To the north of 68° 11′ N, the piedmont broadens to a maximum width of 5 miles. There is no scarp comparable to that in the south and alluvial fans are few. The scarp is

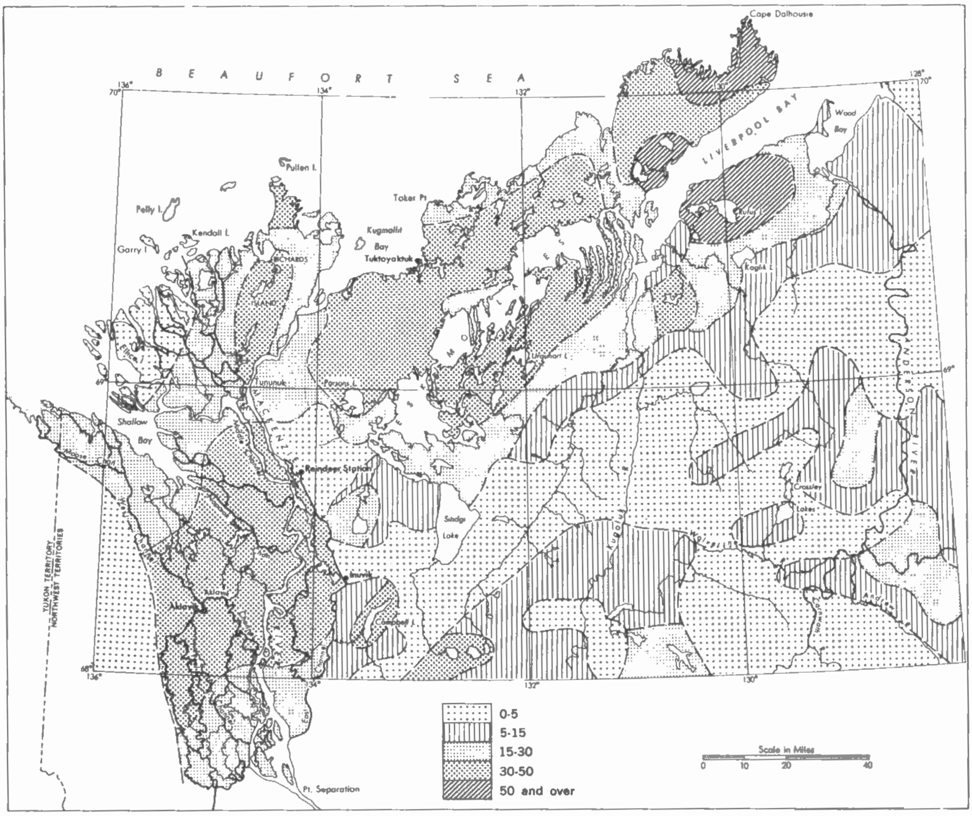

Figure 39. Per cent of area in lakes as determined from a systematic sampling of air photographs.

rarely more than 500 feet high. It is discontinuous, being sectioned by valleys penetrating far inland. The piedmont rises in a gentle slope from West Channel to about 250 feet in altitude within a mile, and then gradually rises inland to about the 500 foot contour to meet the arctic plateau.

Mackenzie Delta: Region II

90. The Mackenzie Delta with its maze of lakes and channels defies easy description. The delta is shaped like the upturned left palm and fingers of a hand with the fingers pointing northward. The wrist is at Point Separation. The indentation between thumb and first finger corresponds to Shallow Bay, the fingers representing the distributary channels. The north-south length is 130 miles and the width about 40 miles. The area, including land and water, is about 4,700 square miles, or a third the area of Vancouver Island. The delta is composite in origin. The greater part has been built up by sedimentation from the Mackenzie River, but the southwest part receives sediment from the Peel and Rat rivers. As the southern part of the delta has already been described (Henoch, 1960a, 1961) the major emphasis will be given to the portion north of 68° 00′ N.

The Mackenzie Delta can be subdivided into three regions, the alluvial islands (IIa), the main delta (IIb), and the rocky islands (IIc). In the northeastern part of the alluvial islands there are several outliers of region IIIa, the principal one being enclosed in a heavy dashed line. Descriptions applying to areas IIa and IIb will be found in subsequent discussions. The easternmost rocky island has a maximum height of nearly 100 feet above the delta; the others are much lower.

91. Boundaries: The southern apex of the delta is at Point Separation. On the east side, the boundary from Point Separation to 68° 00′ N is as far distant as 3 miles east of the right bank of the main Mackenzie River, and then to the east of East Channel. From 68° 00′ N to near Inuvik, the boundary is within a mile of East Channel, except for 3 alluvial re-entrants into the eastern hills. The first re-entrant leads to an unnamed lake at 68° 03′ N, 133° 43′ W; the second and third are infilled valleys connecting with Campbell and Dolomite (Trout) lakes. From 2 miles south of Inuvik to 3 miles north, the boundary is within several hundred yards of the right bank of East Channel. From 68° 23′ N to 68° 35′ N the boundary is irregular. It is no more than 3 miles east of East Channel. From 68° 35′ N to Tununuk, the boundary is at the clearly defined base first of the Caribou Hills to 68° 52′ N and then at that of the Pleistocene area. The boundary from Tununuk to the sea is along the contact of the low alluvial sediments with the higher Pleistocene region. As there are outliers of the Pleistocene deposits, some of which have been eroded close to sea level, the boundary is not clear-cut.

On the west side of the delta, the boundary is irregular in outline, with a few alluvium filled re-entrant valleys. As there is a marked topographic rise from the delta to the higher land on the west, the boundary is sharp.

92. Mackenzie River: The Mackenzie River is the longest in Canada and one of the 10 longest in the world. The length to the remotest tributary, Finlay River, is 2,635 miles. The length from its head in Great Slave Lake to its mouth at the Beaufort Sea in 1,071 miles. The Mackenzie Delta is the gift of the Mackenzie River.

The discharge of the Mackenzie River may exceed 300,000 cubic feet per second (cfs) during the summer, with a peak of over 500,000 cfs at break-up in late May or early June. The late winter and early spring discharge may be below 100,000 cfs. The Mackenzie River water issuing from Great Slave Lake is clear with very little suspended matter. During the summer, tributary rivers such as the Liard feed sediment into the Mackenzie, and as the Mackenzie also picks up sediment en route, the waters carry a load increasing with distance downstream. Comparatively little data are available on the load of the river, but the suspended matter in summer at upstream stations, such as Fort Simpson, ranges from about 10 parts per million (ppm) to 100 ppm (Thomas, 1957). In the summer of 1958, water samples were collected in the Mackenzie Delta, the analyses being carried out by J. Ungar (1958). The suspended matter of East Channel at Reindeer Station on the given dates in parts per million was: July 1, 25 ppm; July 15, 23 ppm; August 1, 128 ppm; August 15, 210 ppm; August 31, 31 ppm. Samples from other channels gave similar results. The residue on evaporation ranged from

175 to 250 ppm, the range being slightly lower in 1960 (Thomas, 1961). Thus, although the coffee-colored, unpalatable-looking Mackenzie water is turbid in summer, and is particularly roily in the swirls, the suspended load is far below that of innumerable other rivers of the world. No estimate can be made of the bottom load.

Point Separation, at the southern apex of the delta, is at mile 912 from Great Slave Lake. The river is a mile wide at Point Separation, but it divides and subdivides downstream so that the seaward discharge is through a total width of about 12 linear miles of channels; about two-thirds is fed by water from the Middle Channel of the Mackenzie, and a sixth each by East and West Channels, the latter including much Peel River water.

The water which flows past Point Separation takes not only different routes, but different lengths and times of flow to reach the sea. In Figure 40 the isolines show the approximate flow distances in miles from Point Separation as measured along the major channels with interpolation between them. Some channels have much steeper gradients than others. For example, at mile 73 on Middle Channel, some water flows down Napoiak Channel reaching sea level at Shallow Bay at about mile 110; water also flows down Middle Channel to Tununuk where some goes down East Channel to reach the sea at about mile 150 to 155, whereas some goes west of Tununuk to reach the sea at mile 170. Even if allowances are made for shifting sea levels, resulting from the small tidal effects, it is clear that water takes very devious routes of unequal lengths and gradients to reach the sea. Moreover, as there is good evidence to suggest that the larger channels have varied little from their present size for 150 years, the channels are in near equilibrium, with neither additions nor abstractions of water. The largest channels tend to have the gentlest gradients, and are efficient channels.

The importance of Peel River to flow and sedimentation in the western part of the delta is considerable. It has been estimated (K. H. Lang, personal communication) that over 90 per cent of the water passing Aklavik derives from the Peel River and the mountain rivers flowing into Husky Channel. This water is diluted with Mackenzie water where Aklavik Channel enters West Channel. The discharge entering the southwest side of Shallow Bay is estimated to contain about 15 per cent Mackenzie water.

Drainage Areas

93. Drainage Area I: This area receives its water from Middle Channel. At Tununuk, some of the Middle Channel water goes northeast into East Channel, the remainder northwest into the channel leading to Kendall Island (Figure 41). The East Channel current is stronger. The main channel is the large unnamed channel leading to Kendall Island. At 7.5 miles northwest of Tununuk, the height of land leading to "Yaya Lake" (*49*) is only about 6 or 7 feet above late summer channel level. In times of break-up, water from Kendall Island channel may flow into "Yaya Lake" and a through channel is in process of formation. The channel connecting "Yaya Lake" with Harry Channel and the sea is a reversing channel 10 feet deep at low water. The predominant high water flow is into Harry Channel. However, a moderate west wind will cause a reversal of flow, carrying muddy Mackenzie water into the clear lake.

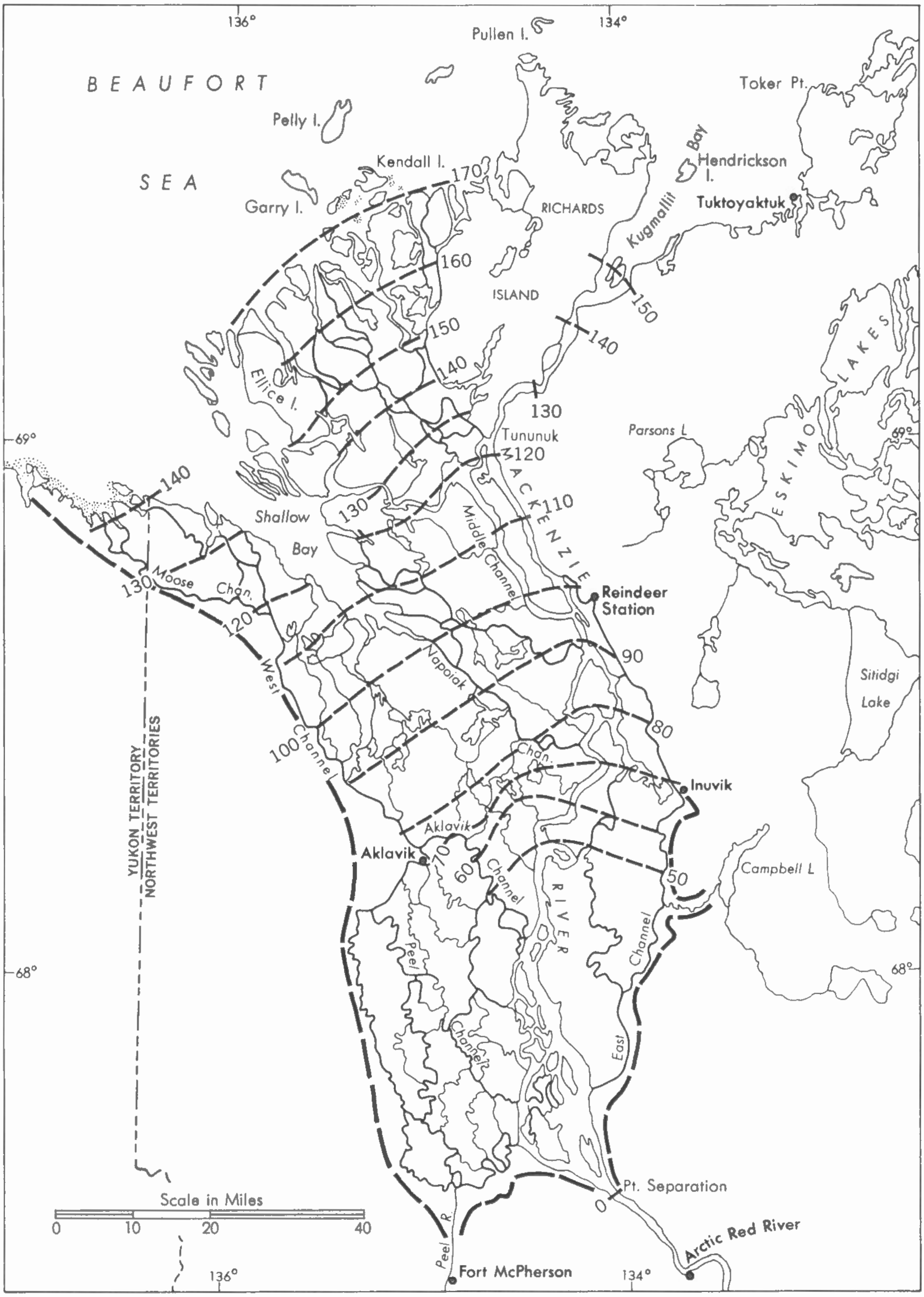

Figure 40. Flow travel distance in miles of Mackenzie River water from Point Separation.

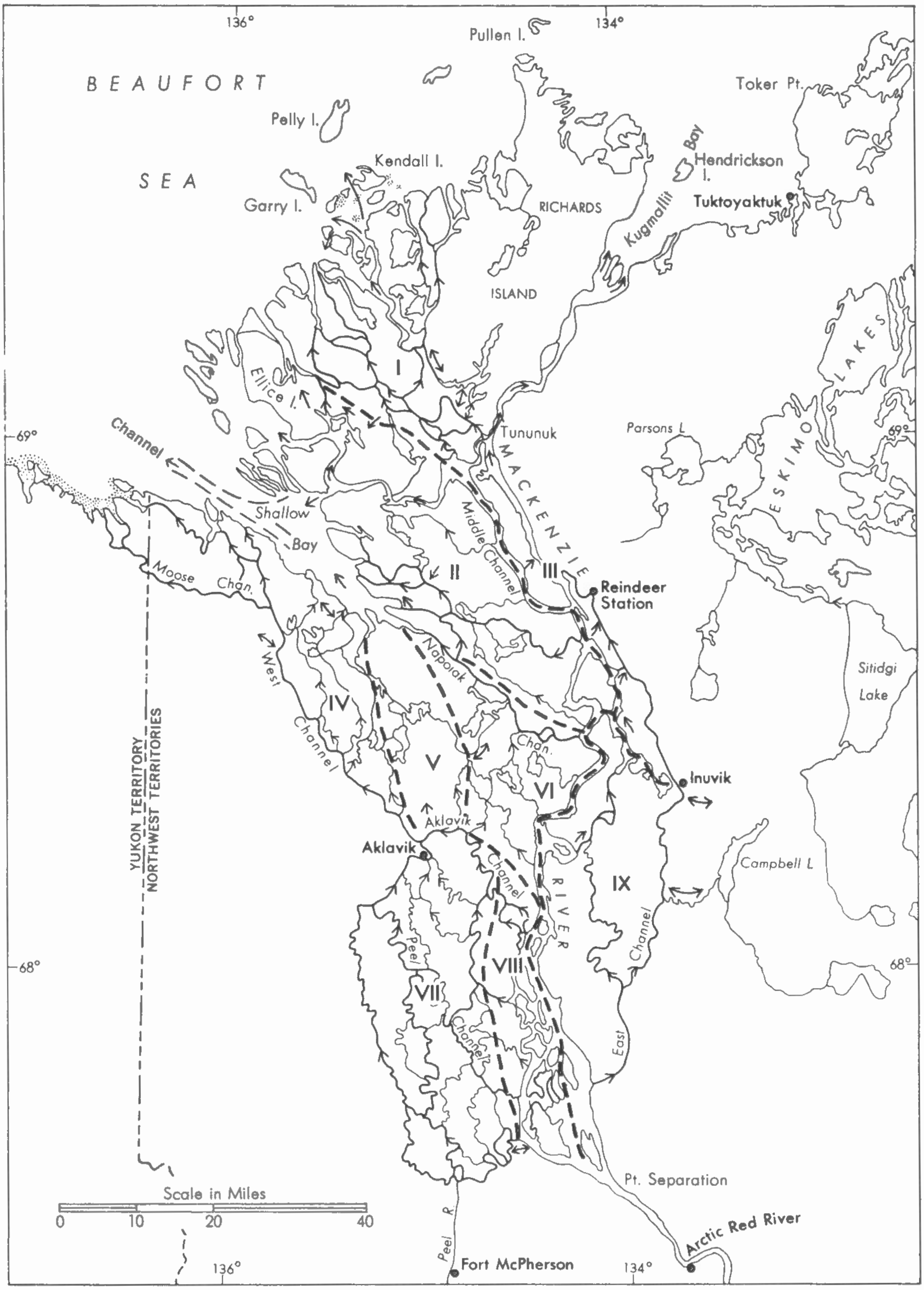

Figure 41. Drainage areas of the Mackenzie Delta. A single arrow shows the principal direction of flow; double arrows indicate reversing flow.

Kendall Island channel is about half a mile wide, broadening into several estuarine mouths which are shoal with mud flats and sand bars. The channel mouth leading directly to Kendall Island winds through extensive mud flats at low tide, but is several miles wide at high tide. The westerly channel, reaching the sea at about 69° 21′ N, 135° 33′ W, was sounded where it was 2 miles wide. All depths were less than 3 feet, except for a narrow channel 200 feet wide and as much as 5 feet deep. The bottom was hard and sandy. However, flood waters rise at least 8.5 feet above low water level, thus inundating all the alluvial islands. The waters far seaward of the alluvial islands are very shallow and often are not deep enough to float an empty canoe.

94. *Drainage Area II:* The flow in this area is from Middle Channel, discharging via large and small channels into the north side of Shallow Bay. Reindeer Channel, the northernmost large channel, was traversed by Franklin in 1825. and the channel has the same major bends today. Depths vary considerably, from shallow to over 100 feet. Reindeer Channel leads into the "submarine channel" in Shallow Bay with depths of 15 to 20 feet. The channels on the west side of Ellice Island are very shallow. Channels half a mile wide at high tide may be impassable for a canoe with a one-foot drop in water at low tide. The channels south of Reindeer Channel tend to be long, narrow and sinuous, typical of forested areas of the delta. The exception is a broad stretch, centered at 68° 40′ N, 134° 50′ W, where a large channel, over half a mile wide, is silting up.

95. *Drainage Area III:* From Inuvik to Tununuk, the right bank of East Channel is never far from the Caribou Hills or the upland of the Pleistocene sediments. The channel is not free to meander on the east and is almost straight. To the west, Middle Channel flows nearly parallel to, and within 5 miles of, East Channel. East Channel receives considerable flow from Middle Channel through 5 interconnecting cross channels, the last being at Tununuk; consequently, the flow of East Channel downstream from Tununuk is much greater than at Inuvik. Widths and depths likewise increase downstream from a fifth of a mile and 10 to 20 feet near Inuvik, to over half a mile and over 100 feet in places north of Tununuk.

East Channel has a shoal, island-dotted, estuarine mouth where it empties into Kugmallit Bay. The alluvial islands stand only 2 to 3 feet above river level. The southern part of Kugmallit Bay is less than 5 feet deep, and a buoyed channel with a 5-foot limiting depth is used by boats. Deeper channels, exceeding 25 feet, are present, but their full direction and extent are unknown. At 69° 17′ N, 134° 06′ W, there is an archaeological site with charred wood, whale bones, etc., on the top of a 25-foot bluff for a distance of 150 feet. At 69° 17′ N, 133° 55′ W, across the river, is another site at 25 to 30 feet above sea level. As there are alluvial islands and shallow channels downstream from the sites, thus restricting the upstream travel of white whales, perhaps the sites were occupied when the channel was less blocked by islands.

96. *Drainage Area IV:* This area is characterized by flow from West Channel and its distributaries into the south side of Shallow Bay. About 2 miles downstream from Aklavik, the Peel and Aklavik channels join to form West Channel. Peel

Channel water, which comes mainly from Peel River, is frequently less turbid than the Mackenzie-fed Aklavik Channel. The flow is estimated to contain 85 per cent Peel and mountain river water, 15 per cent Mackenzie water (personal communication, K. H. Lang). A broad shoal is being built into West Channel, presumably by Aklavik Channel sediment, downstream from its entrance.

West Channel flows against higher land along part of its course and is not entirely free to meander (*106*). As it sends out large right-bank distributaries, but receives few tributaries, its volume diminishes downstream. Numerous cross profiles surveyed across West Channel show a highly varied profile with deep narrow stretches 1,000 feet wide and 50 to 70 feet deep interspersed with broad shallow stretches 2,500 feet wide and 10 to 20 feet deep.

The major right-bank distributaries, such as Leland and Tiktalik channels, reach Shallow Bay without estuarine mouths, unlike those on the opposite side of Shallow Bay, because the south shore of Shallow Bay is a receding higher part of the delta. Tiktalik Channel is used by boats in preference to the West Channel route to Shallow Bay. Tiktalik and other channels may reverse flow at their mouths with a high tide or raised water level in Shallow Bay.

From Little Moose Channel (68° 33′ N) downstream, left-bank distributaries leave West Channel. These channels are flowing through a new area of the delta with large lakes. The channels are narrow and, for the most part, have little sinuosity. Reversals of flow occur. On the extreme west, Mackenzie channels intermingle their waters with that of the Blow River distributaries to form an interconnected network (Mackay, 1960).

97. Drainage Area V: Discharge in this area is primarily from Aklavik Channel via Jamieson and Taylor channels into Shallow Bay. Interconnecting channels, some with reversing flow, link the area to the two adjacent ones. Taylor Channel formerly carried a much larger volume of water (Figure 59); portions of the present channel are inactive. The flow is estimated at about half Peel River and mountain river water and half Mackenzie water.

98. Drainage Area VI: This area might be compared to a river drainage basin, Napoiak Channel being the river. Unlike most delta channels, Napoiak Channel receives more tributaries than it sends out distributaries, so that its size increases. The Napoiak Channel leaves Middle Channel with a width of 200 yards and a swift current, considerably faster than that of most delta channels. Other tributaries, which leave Middle Channel, join Napoiak Channel. The southern tributaries flow through an area with large shallow first-generation lakes (*111*). The flow on the west side of the area is estimated at about one quarter Peel and mountain river water.

Schooner Channel, which enters Napoiak Channel from Aklavik Channel, is an important boat channel for boats going between Aklavik and Reindeer Station. Like most tributary channels, it has a shallow threshold (less than 10 feet deep) where it leaves the main channel (Aklavik Channel) but depths descend to 15 to 20 feet within 200 feet.

Napoiak waters flow into Shallow Bay, which is aptly named. Where it is about 4 miles wide, soundings made in the summer of 1957 every 150 feet with a lead line gave a maximum depth of 6 feet and a minimum depth of 5 feet, except for the near-shore areas. Farther seaward, there is a "submarine channel" with a seaward depth of at least 30 feet (Mackay, 1960). The bottom of Shallow Bay, including the "submarine channel" varies from hard sand to soft mud.

99. Drainage Area VII: Although this area is within the Mackenzie Delta alluvial plain, sedimentation is primarily from the Peel and Rat rivers. Reversals of flow occur between the Mackenzie and Peel rivers in the extreme south of the delta, especially at break-up. Reversals take place in the narrow small channels at the "Indian Houses" at the mouth of the Peel River where it enters the Mackenzie River. The current is usually slack. The flow in Husky Channel is partly derived from the mountain rivers. Most of the flow in the remainder of the drainage area is from the Peel River.

100. Drainage Area VIII: Flow in this area feeds into drainage area VII. South of Aklavik Channel, left-bank distributaries of Middle Channel carry Mackenzie water into drainage area VII, particularly into Esau Channel, a sluggish shallow channel, which was formerly a major channel leading into Aklavik Channel (Figure 59). The Aklavik Channel, between its mouth at Middle Channel and the take-off of Schooner Channel, is a sinuous, meandering river.

101. Drainage Area IX: All the flow in this area is from the Mackenzie River, with the exception of that derived from small rivers entering East Channel. Rocky islands lie on both sides of Kalinek Channel, near its junction with East Channel, and bedrock outcrops along several channels. This is exceptional for delta channels. Areas of new lakes occur southwest of Inuvik (Figure 73).

Channel Characteristics

102. Types: In the intricate anastomosing network of channels and interconnecting lakes, a number of characteristic channel types predominate. The channels are not all of the distributary type, like the branching veins of a leaf. Type (1): Distributary channels diminish in volume as they send off distributaries. Good examples are Middle Channel, West Channel, and East Channel downstream to 68° 30′ N. Type (2): River channels receive tributaries, as do normal rivers. Napoiak and East Channel (north of 68° 30′ N) are examples. Type (3): Network channels link the main distributary and tributary systems. Examples are Oniak Channel linking Middle with East Channel, segments of Schooner Channel, and Enoch Channel. Most of the network channels are from 200 to 1,000 feet wide and from a mile to 10 or 15 miles long. Two network channels may unite to form one larger channel or one may subdivide into two. The channels rarely meander, in the geomorphic sense, although they may be sinuous with meander-like bends. The channels wander. Many lakes "open" onto the channels, or the channels may "flow" through lakes. Type (4): Lake channels interconnect lakes with other lakes or with other types of channels. These lake channels are usually the smallest of all channels, with widths of from

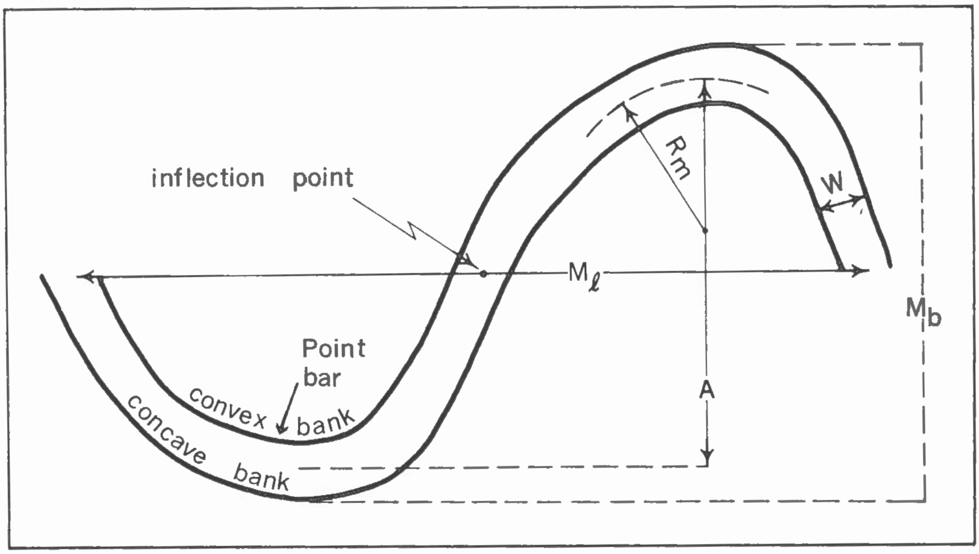

Figure 42. Terminology for a river meander. M_l is the mean meander length; R_m the mean radius of curvature; M_b the mean meander breadth; W is the mean width; and A is the mean amplitude, which is a width (W) smaller than M_b.

a few feet to about 100 feet. The lake channels reverse, with water flowing into the lakes at high water level, and out again at low water. The reversals occur most commonly after break-up, but also take place following any pronounced rise in river level whether from heavy precipitation or surges accompanying storms. Type (5): Reversing channels have a frequent and sufficiently strong reversing flow to affect the channel characteristics. A channel with a strong reversing flow tends to have uniform sides with a poor development of both undercut and slipoff slopes. The flow of Gull River, which connects East Channel with Campbell Lake, is an example. The inflow at break-up may raise the level of Campbell Lake by 20 feet, and a delta is being built into the lake. Inflow or outflow from Campbell Lake may reverse frequently during one summer. Conditions of inflow and outflow from Dolomite (Trout) Lake are similar Reversing channels are well known to delta inhabitants. Indications of reversals of flow may be given by bent willows leaning opposite to the momentary direction of flow, traces left by the ploughing and scraping of river ice, and the attitude of driftwood logs on the banks, which tend to lodge with the roots pointing upstream, the trunk downstream. Type (6): Despite the small tidal range of about one foot, some tidal channels with reversing flows occur along the distal part of the delta. For example, at the "Whitefish Station" on the west side of Mackenzie Delta (*185*), tidal observations were kept for three days on a channel a third of a mile from the sea. The current reversed once or twice a day. Tidal channels appear to have U-shaped cross profiles, both above and below water level. In this respect their cross profiles are similar to reversing channels.

103. Channel Geometry: The geometry of rivers, especially meandering rivers, has been studied intensively by many investigators working in the laboratory and in the field on both canals and rivers. These studies have demonstrated that certain relations tend to exist in meandering rivers, whether they are large or small, or whether they flow on bedrock, unconsolidated material, or glacier ice. Well-defined linear relationships often exist between the meander length (M_l), the meander breadth (M_b), mean radius of curvature (R_m), and channel width (W) (Figure 42). Meander length ranges from 7 to 10 times the channel width; meander length is about 5 times the radius of curvature; the ratio of radius of curvature to width ranges from 1 to over 4; but the ratios involving amplitude show little consistency (Leopold and Wolman, 1960). River width, depth, and velocity are related by power functions to discharge. The measurements are normally for bankfull or flood conditions, in which the product of mean width, depth, and velocity give the discharge. The meander length and amplitude (which is essentially the same as meander breadth) have also been shown proportional to the square root of discharge.

Meandering rivers fall into two main types: spilling rivers and incised rivers (Blench, 1957; Inglis, 1949, p. 145). In spilling or floodplain rivers, the point bar of the meander is inundated at high water with a short-circuiting of some channel flow across it. In incised rivers the flow is in a well-defined channel with relatively steep banks, so that flood waters are confined in a channel of nearly constant and uniform width.

In the analysis of Mackenzie Delta channels, measurements must be based, of necessity, largely on air photographs, and the water levels at the time of photography are unknown. As all of the photographs used were taken after break-up, the measurements apply, in general, to mid-summer levels. If the measurements were made for flood conditions, the river width would be greater, but the meander length, breadth, and radius of curvature would remain essentially unchanged. Unfortunately no discharge data are available. Despite the inadequacy of the statistical data, the results are consistent enough when analyzed to give the general characteristics of the geometry of the Mackenzie channels. The analyses have been based on a sampling of 39 large and small channels ranging from the smallest, 110 feet in width, to the largest, which was Middle Channel with a width of nearly a mile. Equations 1 to 5 are derived from regression analysis, assuming the exponent of the independent variable as unity; r is the coefficient of correlation.

Equation	r	
$M_l = 13.6W$	.87	(103-1)
$M_l = 3.7R_m$	.94	(103-2)
$M_b = 7.5W$	.83	(103-3)
$R_m = 3.6W$	.89	(103-4)
$M_l = 1.8M_b$	.76	(103-5)

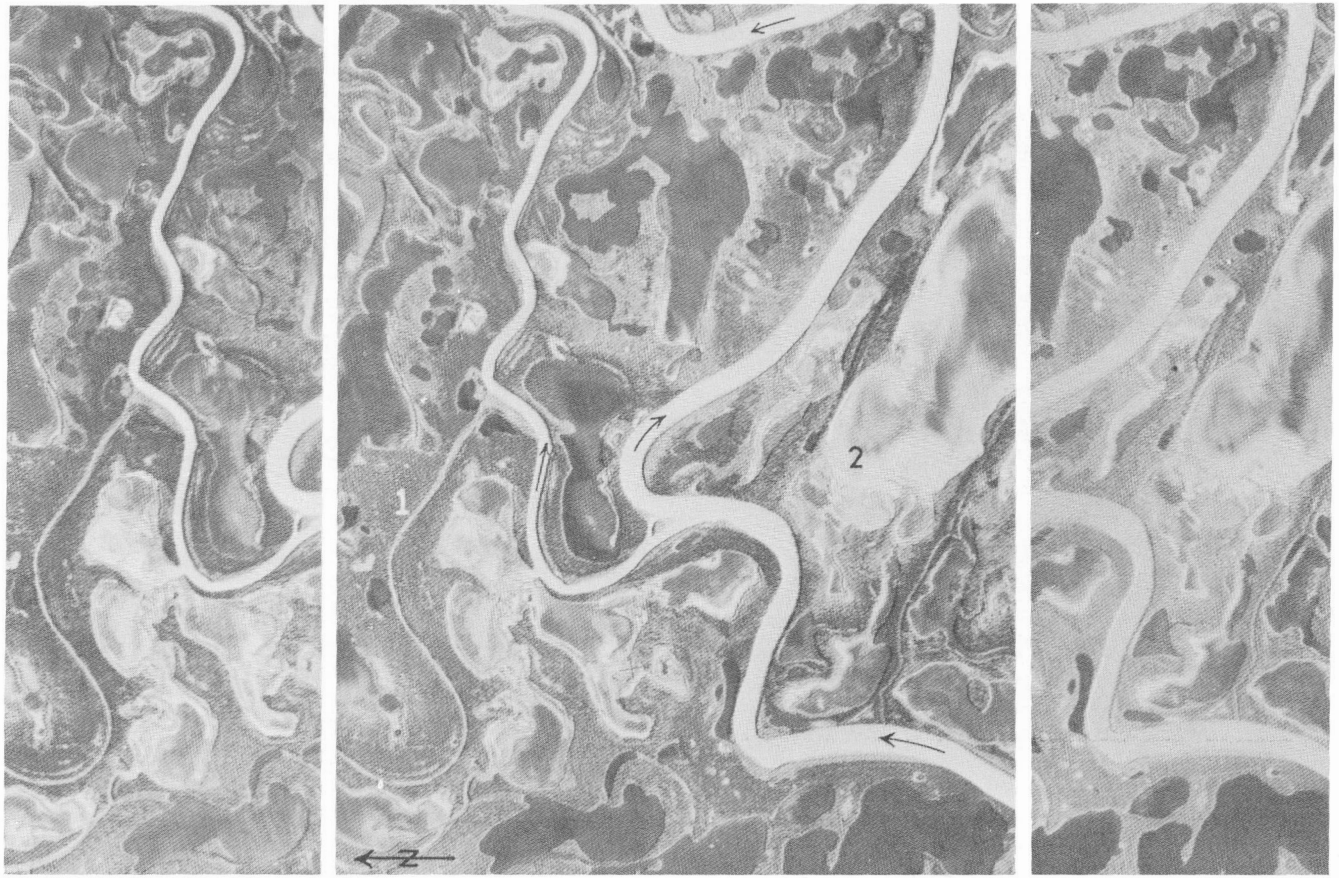

Figure 43. Stereo triplet of Mackenzie Delta centred at 68° 17′N, 134°05′W. 1—abandoned channel. 2—shallow lake with ribbing commencing to form at the northwest end, the ribbing being less pronounced than at the southeast end, off the photograph. The dark tall trees along the channels are spruce, the lower bushy vegetation is willow-alder. Whitish patches around lakeshores have sedges. (RCAF photos A 12918—138 to 140).

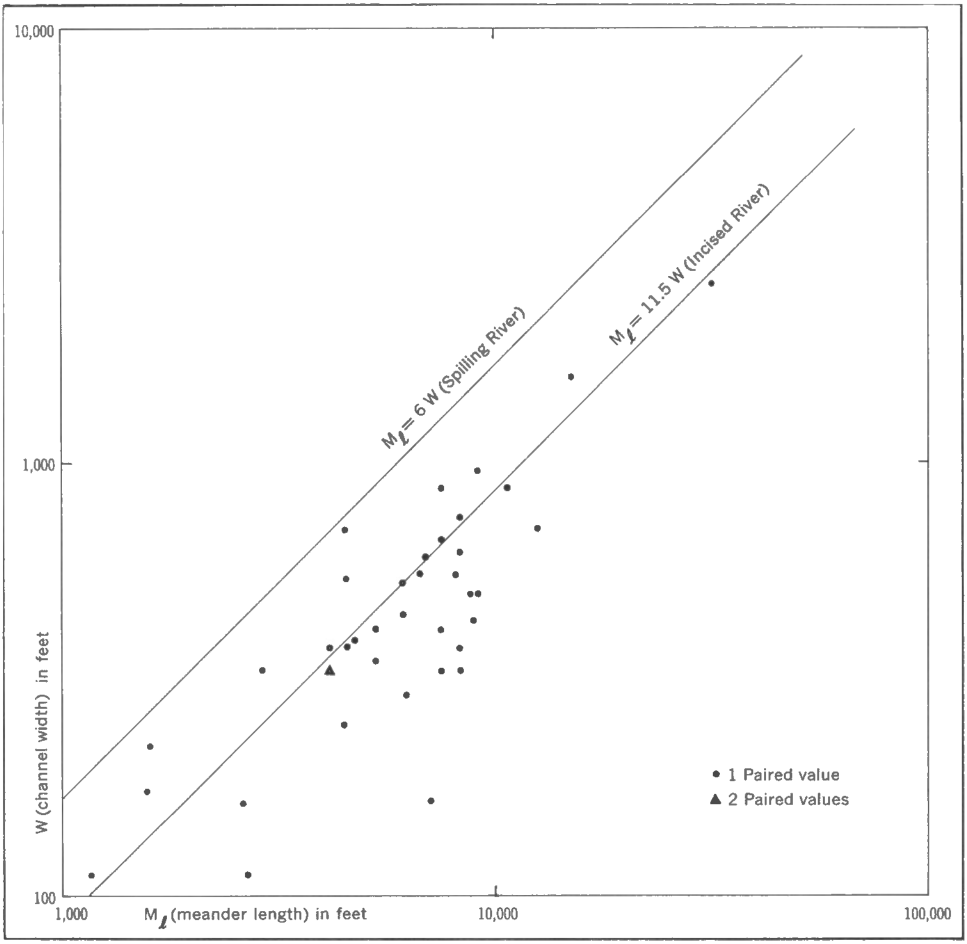

Figure 44. Graph of channel width and meander length.

As the one per cent significance level for r is .40, all the values are highly significant. If least squares regression equations are calculated to determine the exponent of each independent variable which gives the best fit, that for W in equation 1 is .79; for R_m in equation 2 is 1.03; for W in equation 3 is .75; for W in equation 4 is .73; and for M_b in equation 5 is .86. Although only the exponent in equation 2 is close to unity, the spread does not seem excessive, in view of the general roughness of the measurements.

In Figures 44, 45, and 46, data for spilling and incised rivers have been added for purposes of comparison, most of the data coming from Inglis (1949, p. 145); the values for spilling and incised rivers should be considered only as rough approximations. In Figure 44 the points tend to be scattered slightly below the line

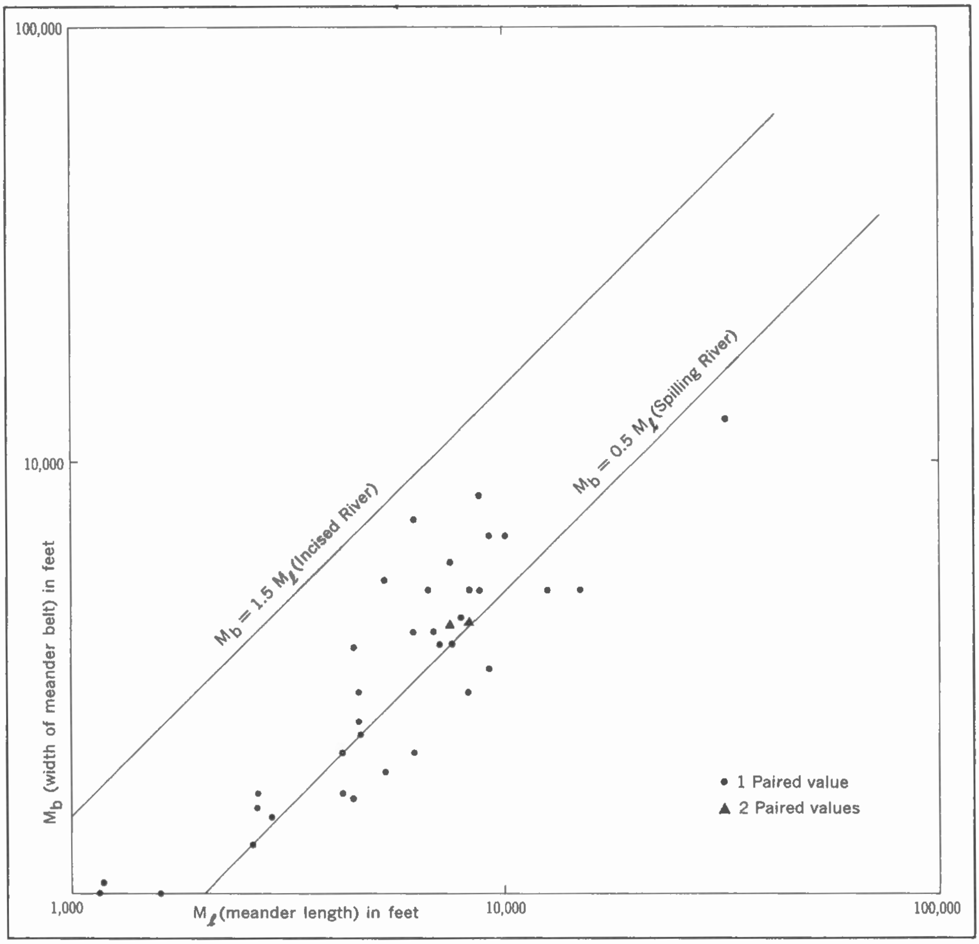

Figure 45. Graph of width of meander belt and meander length.

for incised rivers for both large and small channels. In Figure 45, the scatter of points approaches that of spilling rivers. In Figure 46, the points are far removed from the trend of either incised or spilling rivers. Thus, the Mackenzie Delta channels do not fall into the category of either spilling or incised rivers. Even if allowance is made for a considerable underestimation of width, still the Mackenzie channels do not conform precisely to river geometry elsewhere. The probable reason is that each channel behaves not as a single river, but as several rivers, depending on the time of year. At break-up, for example, ice-jams cause diversions, reversals of flow, and overtopping of levees. Thus, the flood flow may be little related to late summer conditions. In winter, many of the smaller channels freeze to the bottom and flow in them ceases. Consequently, the summer and winter flow networks are not identical.

In Figure 47, the radius of curvature is plotted against width. The central line is that of equation 4, the upper and lower lines values for R_m of 6W and 2W. With most rivers, the scatter of points would lie below the central line, so the Mackenzie channels tend to be narrow in relation to the radius of curvature. With further analyses, and a stratification of the channels into homogeneous types, a smaller scatter of points should occur. The agreement in channel geometry for certain types of channels with rivers elsewhere would likely prove close.

104. Cross-Sections: The channel cross-sections are varied in shape, depending upon the channel type. Numerous cross-sections were surveyed in the field. Altitudes were accurately measured with a telescopic alidade; depths were obtained by sounding with a lead line at constant time intervals from a boat run at slow speed.

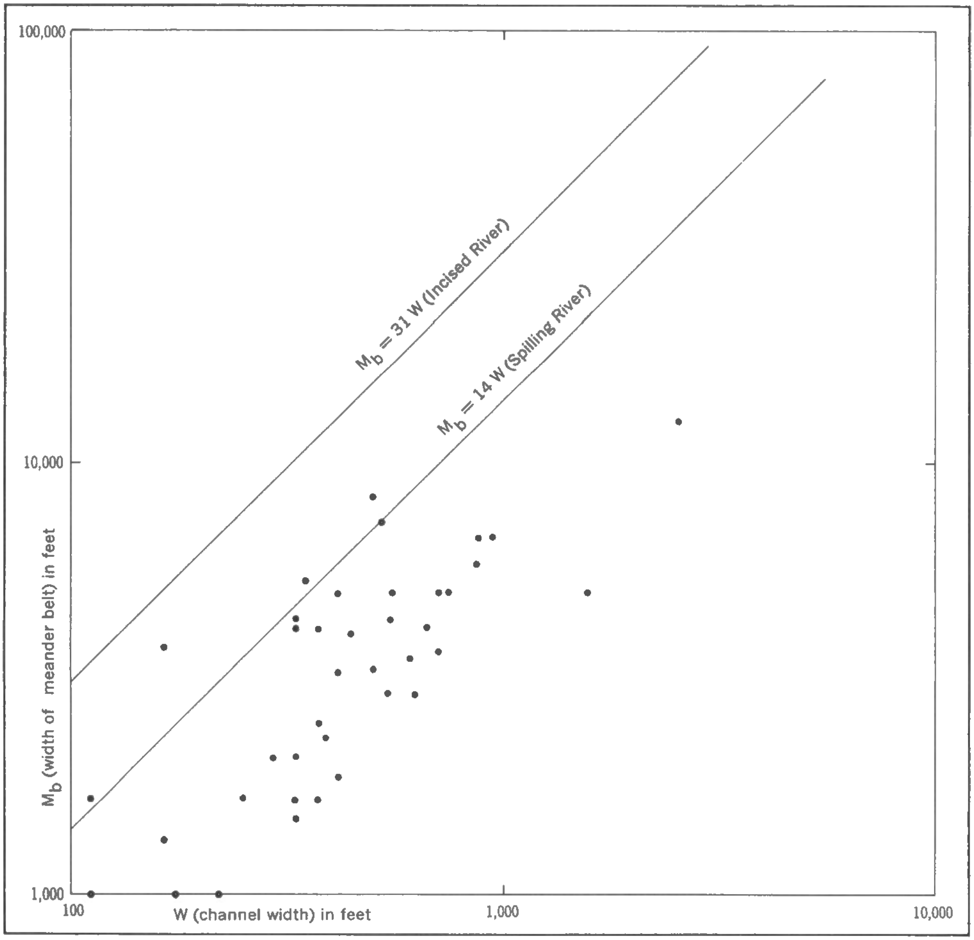

Figure 46. Graph of width of meander belt and channel width.

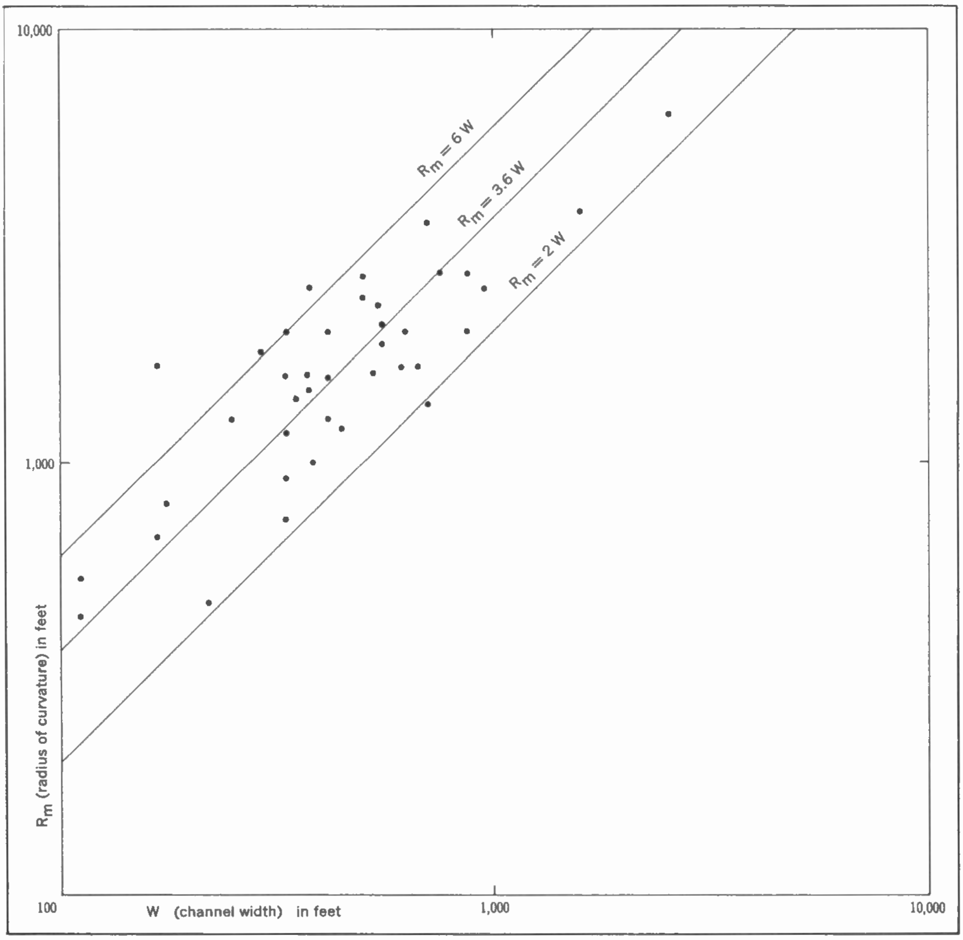

Figure 47. Graph of radius of curvature and channel width.

At meander bends, cross-sections through the axis show an undulating profile of alternate swells and swales on the point bar (Figure 50, Nos. 1, 2; Figure 51, No. 8). The undercut slope is usually marked by a levee which maintains its altitude and backslope as the bank is eroded. Cross-sections along straight to gently curving reaches frequently show steep cut banks on one or both sides of the channel, with a shoal bench close to low water level. On the larger channels, the benches may be from 50 to 400 feet wide, and from a few inches to several feet deep (Figure 50, No. 3). The bottom profile of many of the smaller channels which are 100 to 200 feet across are smooth and broadly U-shaped (Figure 50, No. 1; Figure 51, Nos. 11, 13). Such U-shaped cross sections are also typical of reversing channels (Figure 50, No. 2; Figure 51, No. 11; Figure 52, No. 20) and tidal channels. Most of the smaller channels which are near to the sea are U-shaped (Figure 51, No. 7). Some of the larger channels, especially West Channel which was sounded at numerous locations, have highly irregular bottoms, at least at low water when they were sounded (Figure 50, Nos. 3, 4, 5). As there is scour and fill in all channels between high and low water, high-water profiles are unknown.

As the channel mouths are approached, most channels become very shoal. In Figure 51, No. 8, a cross-section of Tiktalik Channel is shown a mile from its mouth. The current at the time of sounding was about 2.5 miles per hour and the bottom profile was highly irregular. Figure 51, No. 6 shows the profile where Tiktalik Channel emptied into Shallow Bay. The width of the channel had quadrupled, but depths were much less. The shore of Shallow Bay is being rapidly cut back at the mouth of Tiktalik channel, as much as 5 to 10 feet in a single summer storm, but the channel maintains its depth right down to its mouth. The Tiktalik Channel is reported to sustain a good flow beneath the winter ice.

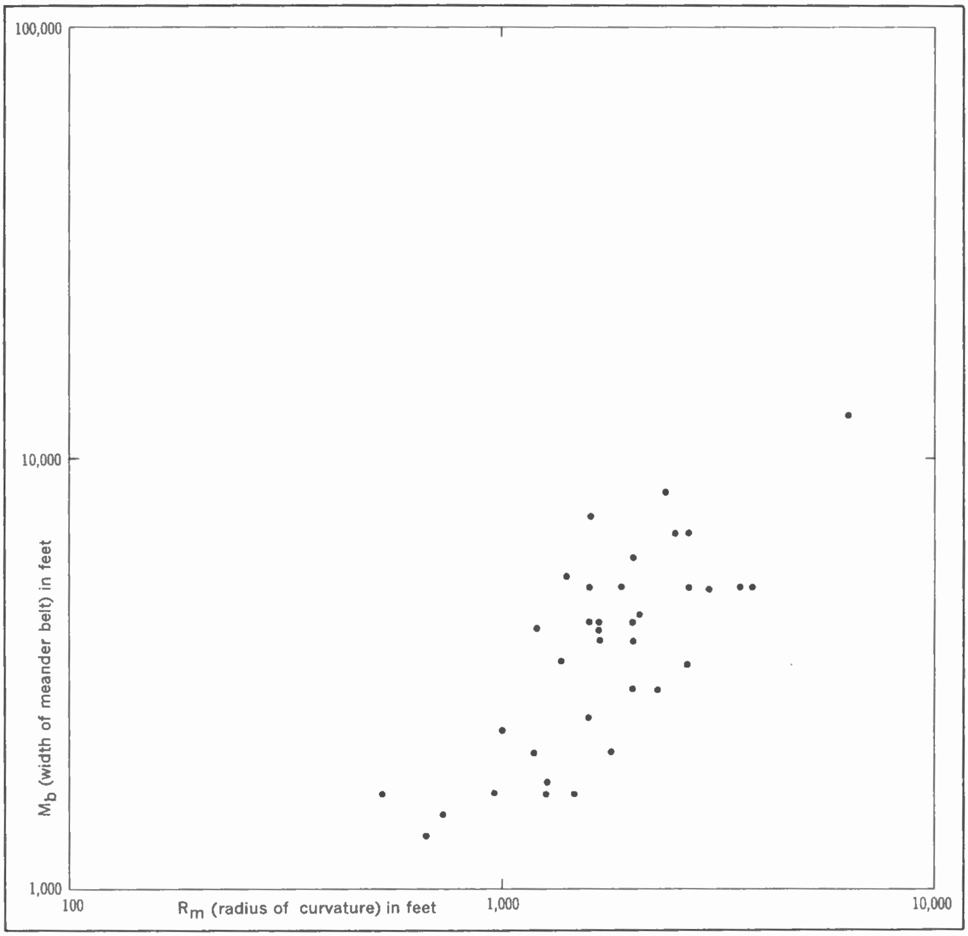

Figure 48. Graph of width of meander belt and radius of curvature.

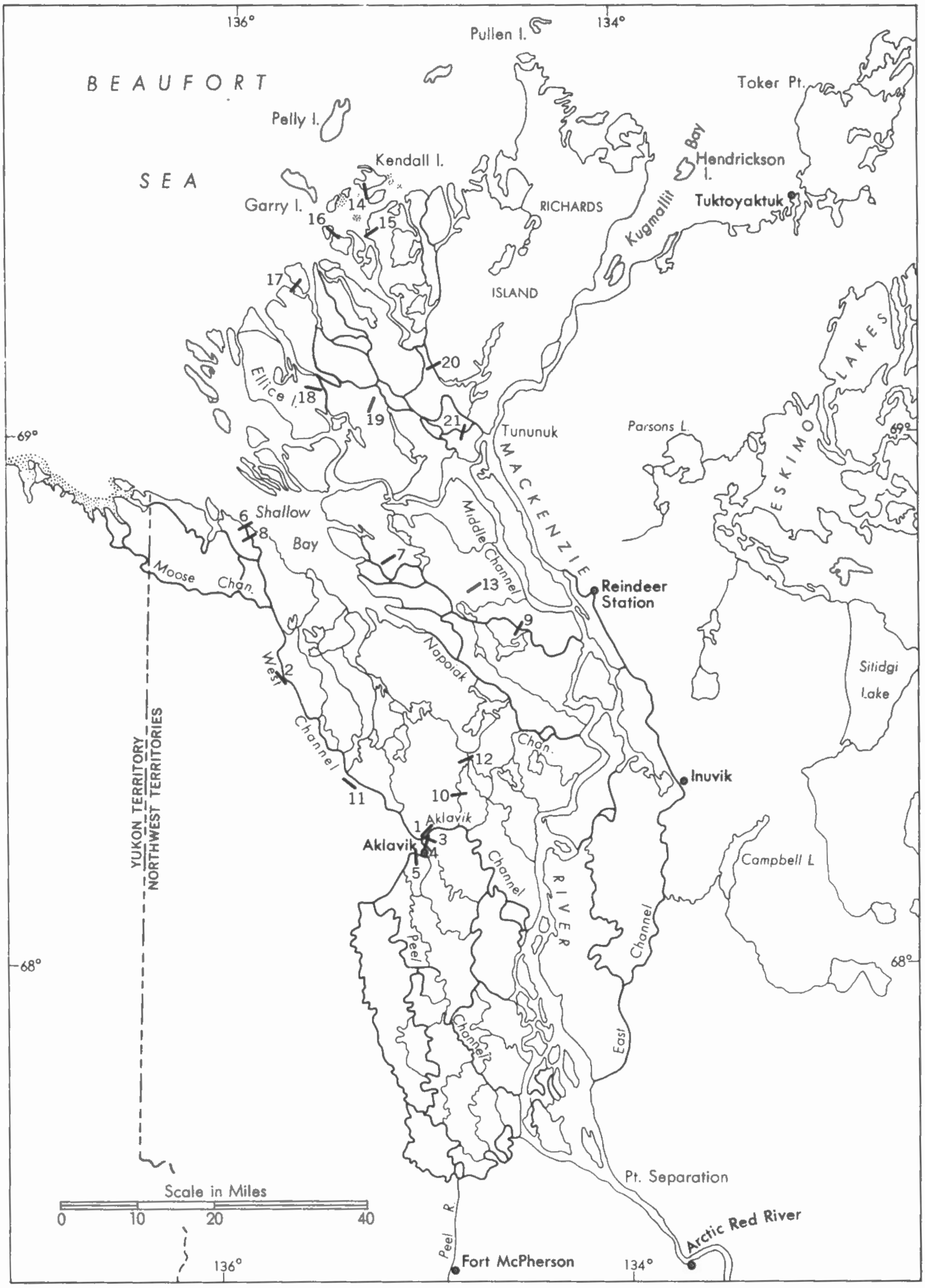

Figure 49. Location map for channel cross-sections (Figures 50, 51, 52).

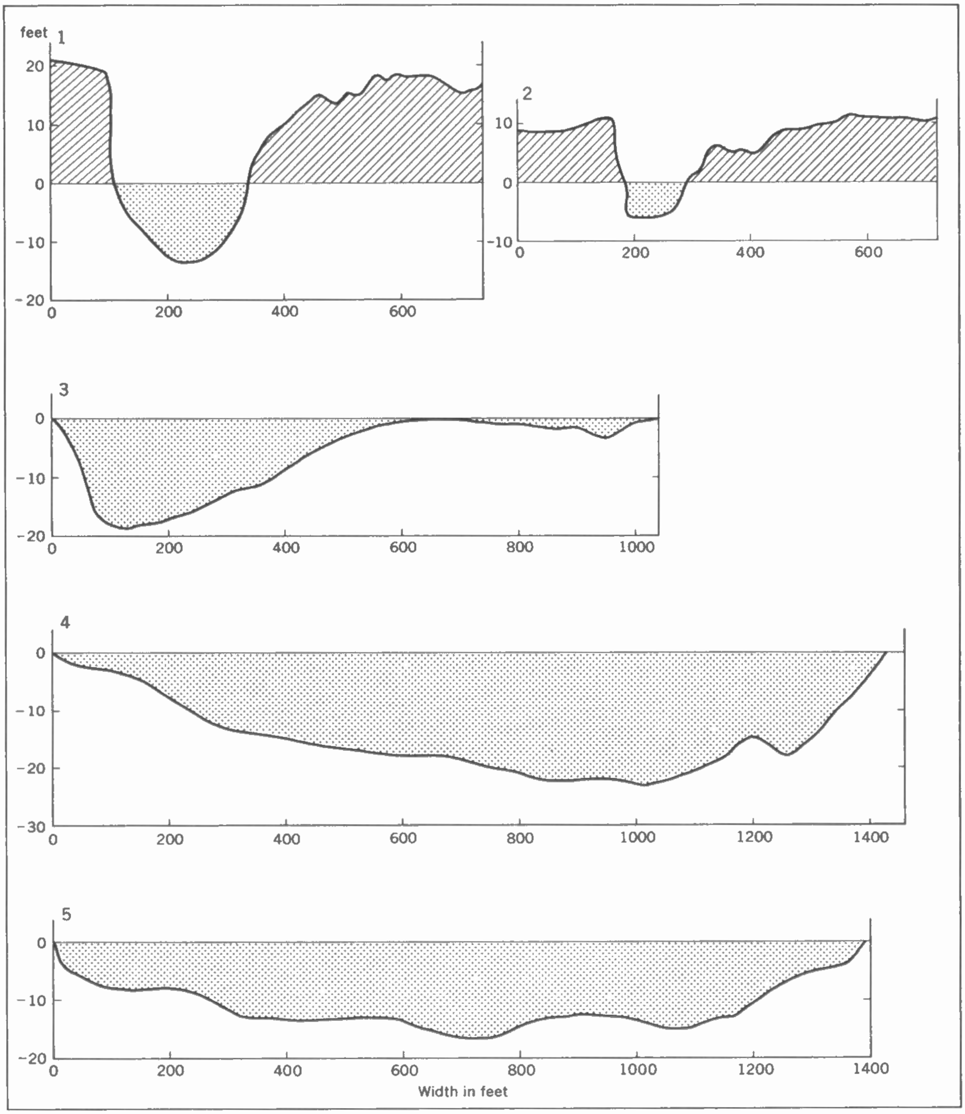

Figure 50. Channel cross-sections. Note the vertical exaggeration.

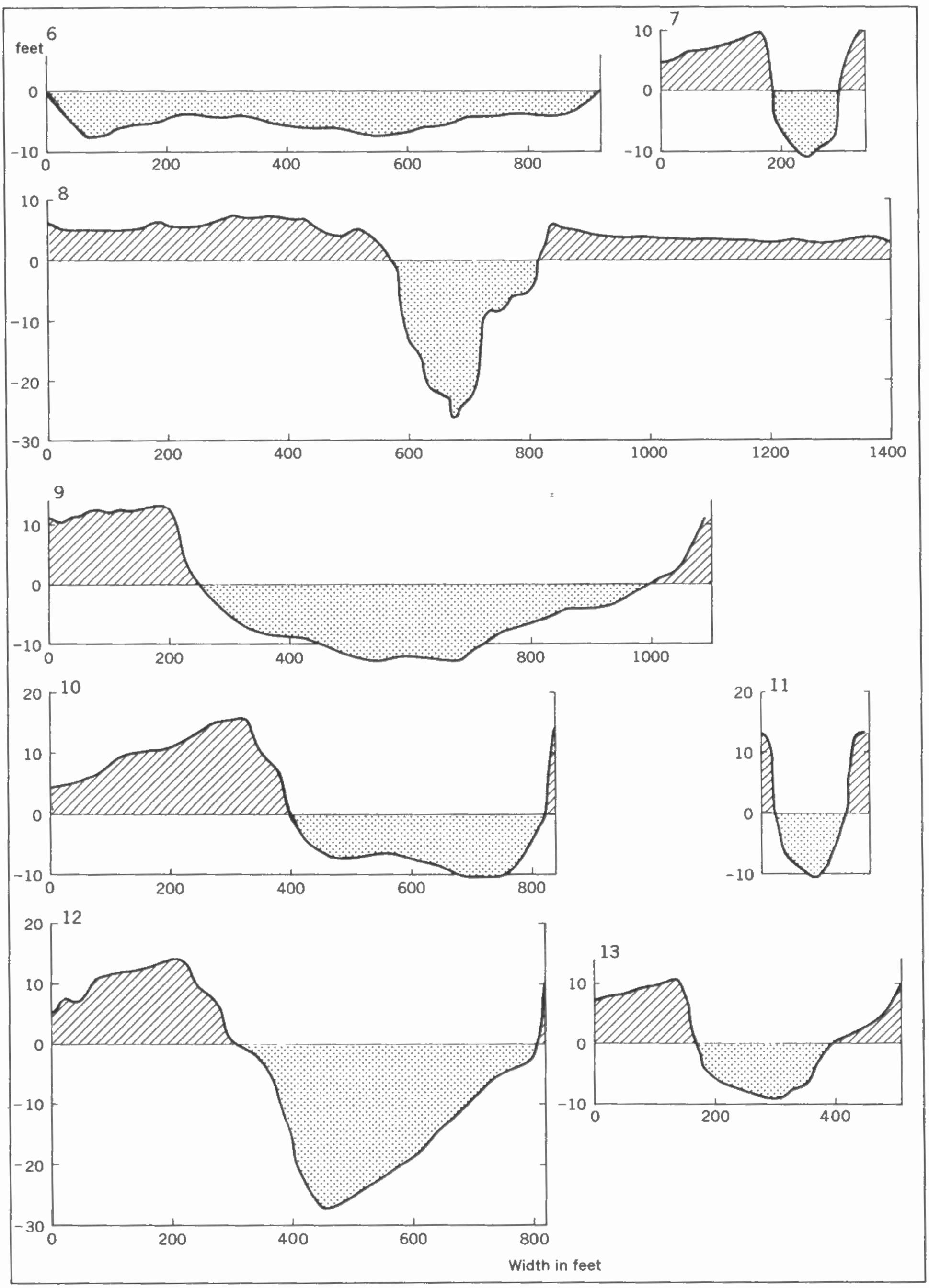

Figure 51. Channel cross-sections. Note the vertical exaggeration.

At flood time, the cross-sectional area (A_f) of channels is much greater than that at low water (A_l). The hydraulic radius (the ratio of the cross-sectional water area to the wetted perimeter) also changes from that at flood (H_f) to low water (H_l). The data, given below, show the relationships, as determined by regression analysis, for 19 channels of varied sizes for which profiles were available for flood levels on both banks. Measurements are in feet and square feet; r is the coefficient of correlation.

	r	
$A_f = 21 + 1.7\ A_l$	.95	(104-1)
$H_f = 13 + .00032\ A_f$	.68	(104-2)
$H_l = 6.8 + .00062\ A_l$	.80	(104-3)

The one per cent level of significance for the coefficient of correlation (r) is 0.58, so all are significant. The best agreement is between cross-sectional area at flood and low water. The cross-sectional area at flood is usually 1.5 to 3.0 times that at low water. The hydraulic radius, which reflects the average depth,

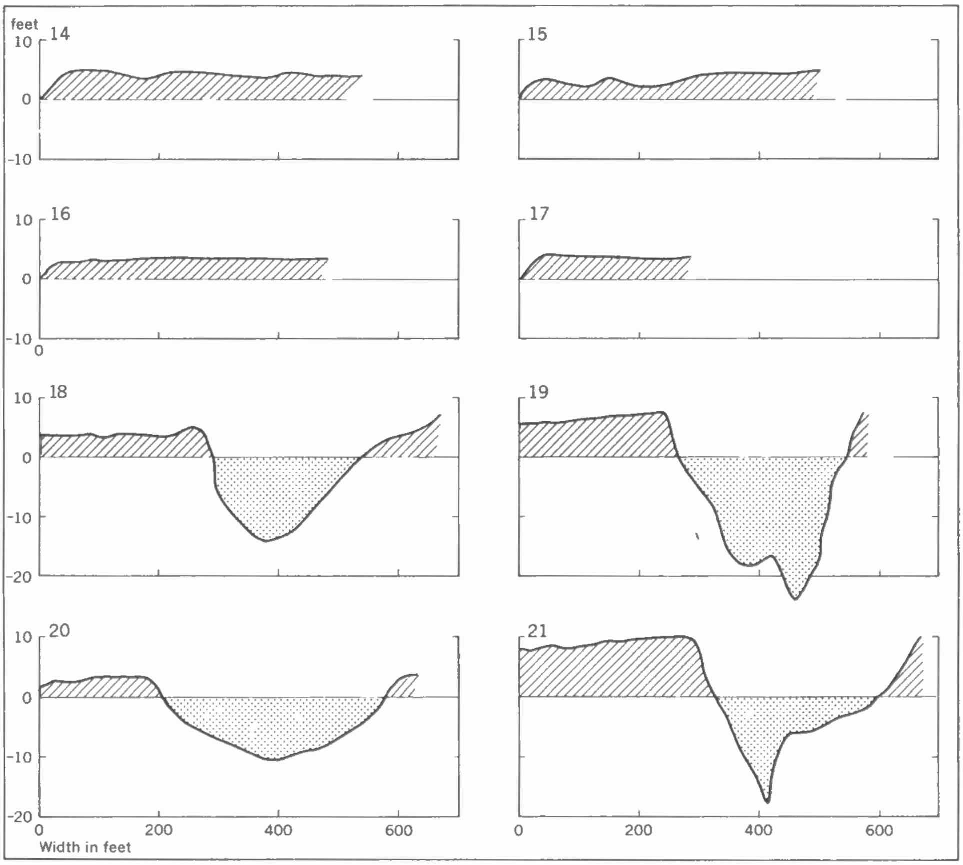

Figure 52. Channel cross-sections and levees. Note the vertical exaggeration.

shows considerable variation even for channels of the same cross-sectional area. Possibly an analysis based on a curvilinear relationship, using channel width as a variable also, would be preferable. However, the number of measurements are at present insufficient to justify any more than a preliminary analysis.

105. Harmonic Analysis: Break-up in many northern rivers is accompanied by the formation of ice-jams, frequently at sharp river bends. Thus, the normal shape of a bend, such as that of a Mackenzie Delta channel, might be affected by flood conditions. In order to study the geometric symmetry of a large channel liable to ice-jams, a 45-mile-long stretch of the large Middle Channel from 68° 15′ N to 68° 41′ N was subjected to harmonic analysis. That is, the shape of each meander is expressed by a finite number of sine terms. The inflection points of the Middle Channel were first marked, there being seven successive bends (note: two bends were omitted because channel bifurcations made it difficult to determine the precise limits of the bends). The harmonic analyses were carried out for 12 equidistant points for the concave bank taken in the direction of flow. The first three harmonics were found to describe the curves quite accurately. The time angle wave peaks for the first three harmonics (X_1, X_2, X_3) are shown below:

Curve No.	X_1 max	X_2 max	X_3 max
1	161°	73°	44°
2	114°	47°	25°
3	170°	79°	49°
4	153°	72°	49°
5	156°	73°	36°
6	163°	75°	44°
7	115°	57°	32°

All the time (phase) angle wave peaks for the first, second, and third harmonics are less than 180°, 90°, and 60° respectively. This means that the seven meander bends are all asymmetric downstream, a conclusion by no means visually evident from an inspection of the curves of Middle Channel. The bends reflect therefore, accentuated erosion of the concave banks and deposition on the convex banks downstream from the axes of the bends. It is interesting to note that curves 2, 4, and 7, which are on the right bank of the river, have smaller phase angles for the first maximum than the left bank curves 1, 3, 5, and 6, the same holding for the second harmonic. Whether this is by chance, or related to other factors such as coriolis force, prevailing wind direction, or exposure is problematical (*see* Williams, 1952). The asymmetry of the Middle Channel bends is characteristic of meandering rivers in general, although an harmonic analysis of some laboratory channels suggests that the asymmetry of the Middle Channel bends is more than might be expected.

106. Variance Spectra: Most of the channels display a bewildering variety of shapes, whether viewed from the air, on maps, or on air photographs. A few appear to have well-developed symmetrical meanders of textbook perfection, others have considerable sinuosity but with little regularity in the sizes of the bends, but most seem to wander haphazardly without evident control. To add to the complexity, entrance or take-off of tributaries and distributaries produce changes in channel geometry downstream from the junctions. In order to analyze the channel sinuosities, the method of variance spectra (Panofsky and Brier, 1958, pp. 140-147; Bryson and Dutton, 1961) was used. The channel sinuosities were treated as a stationary time series and variance spectrum analysis (a statistically sophisticated method of analyzing the cyclical behavior of a time series) was applied to the channel shapes.*

The channels subjected to variance spectrum analysis were:

1. Channel from 68° 24′ N, 135° 37′ W to 68° 45′ N, 135° 37′ W. Channel has tributaries and distributaries.

2. Channel from 68° 52′ N, 135° 11′ W to 68° 50′ N, 135° 24′ W. Channel has neither tributaries nor distributaries.

3. Channel from 68° 18′ N, 135° 07′ W to 68° 25′ N, 135° 15′ W. Channel has neither tributaries nor distributaries.

4. East Channel from 68° 00′ N, 133° 58′ W to 68° 12′ N, 133° 50′ W. Channel has tributaries and distributaries and is not free to meander on the east side.

5. Channel from 68° 33′ N, 134° 25′ W to 68° 39′ N, 134° 44′ W. Channel has tributaries and distributaries.

6. Channel from 68° 39′ N, 134° 32′ W to 68° 37′ N, 134° 16′ W. Channel has neither tributaries nor distributaries.

7. West Channel from 68° 15′ N, 135° 03′ W to 68° 45′ N, 135° 51′ W. Channel sends off distributaries and locally is not free to meander on west.

8. Napoiak Channel from 68° 29′ N, 134° 32′ W to 68° 39′ N, 135° 06′ W. Channel has tributaries and distributaries.

9. Middle Channel from 67° 45′ N, 134° 25′ W to 68° 56′ N, 134° 50′ W. Channel has tributaries and distributaries but does not vary greatly in volume.

Channels 2, 3, 6, 8, and 9 show good power-law relations and appear to have "regular" sinuosities. Channels 2, 3, and 6 have neither tributaries nor distributaries; channels 8 and 9 have tributaries and distributaries, but they tend to counterbalance each other. Channels 1 and 5, which have tributaries and distributaries, have negative spectral estimates and no "regular" sinuosities. Channel 4 (East Channel) has an up-and-down pattern and channel 7 (West Channel) one with zero slope. These two channels, which are not fully free to move laterally and send out distributaries, also lack regularity. Although more channels need to be studied, it is evident that those channels which are little altered by tributaries and distributaries have "regular" sinuosities even though they look just as irregular as other channels. Thus the method of variance spectrum analysis gives information on the dynamic behavior of the delta channels.

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*The computations were carried out on a computer at the University of Wisconsin through the cooperation of R. A. Bryson and J. A. Dutton, Department of Meteorology. A fuller joint paper is in preparation.

107. Channel Junctions: There are many hundreds of major and minor channel junctions within the delta. The relative positions of the channel junctions are not at random, as a cursory study of a map suggests, but they are systematically disposed both in their specific location on channel bends and also areally.

Figures 53 and 54 have been plotted from the principal channels shown on the 2-mile-to-the-inch hydrographic charts (Nos. 6386 and 6387) covering the delta. The lowest density (Figure 53) of channel junctions (less than 5 per 100 square miles) is in the southwestern part of the delta where conditions are governed largely by Peel River and mountain river flow. This is an area of high lake density, locally exceeding 10 per square mile (Mackay, 1956a, p. 4) but few interconnecting channels. In contrast, the southern side of Shallow Bay has a high channel junction density (more than 20 per 100 square miles) but the number of lakes per square mile averages less than 5. The zone along Middle Channel has an intermediate channel junction density (10 to 15 per 100 square miles) and an intermediate lake density of 5 to 10 per square mile. In Figure 54 the "average" distance in miles between channel junctions is shown. The "average" distance has been computed on the basis of an even hexagonal distribution of channel junctions. The map shows a general northward decrease in distance between channel junctions with the exception of the islands undergoing active sedimentation north of Shallow Bay.

Channel junctions of meandering channels, or those with well-developed sinuosity, are nearly always along the concave (undercut) banks (Figure 42). Very few channel junctions are found on the convex banks. The most important factor determining the location of channel junctions is probably the path followed by the bed-load. In general, the bed-load travels downstream from convex bend to convex bend. That is, it tends to move towards the centre of curvature. Therefore, a natural or artificial diversion on a convex bend will tend to entrain more sediment than one on a concave bend, and will often plug up. For this reason, artificially assisted cut-offs or channel diversions in rivers and canals are typically located on concave banks (Blench, 1957, p. 88-93; Lindner, 1953; Matthes, 1933). In the case of the delta channels, most of the junctions are clearly located on the concave bends. Although a few do occur on convex bends, those which could be examined in the field were in each instance on convex bends which were being undercut by a swing of the current. In other words, the channel flow pattern was reversing the positions of concave and convex banks. Such reversals may result from any major alteration to a channel flow, such as an upstream channel diversion. In Figure 55 the results of a measurement of the position of 100 channel junctions are plotted. The measurements were made on 10 representative meandering channels of varying sizes. Although there is obviously some subjectivity in the location of the axes of bends and the centres of curvature, the figure probably gives a good indication of the position of channel junctions in relation to channel bends. About two-thirds of the junctions occur within a 60° arc subtended at the centre of the radius of curvature. No statistically significant differences were found in the behavior of large and small channels; of right and left hand banks; and of channel junctions involving either distributary or tributary flow. In many of the

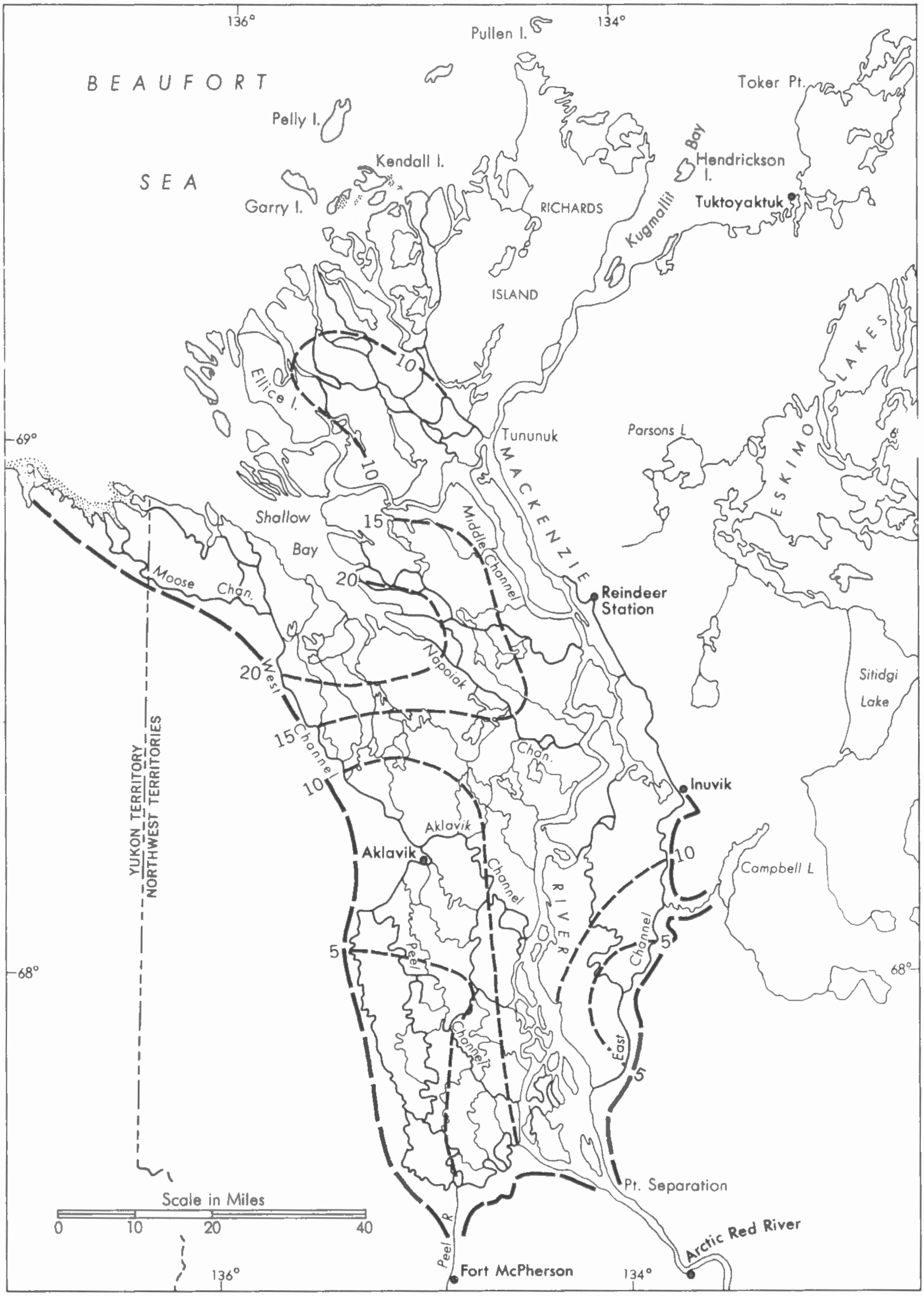

Figure 53. Density of major channel junctions per one hundred square miles.

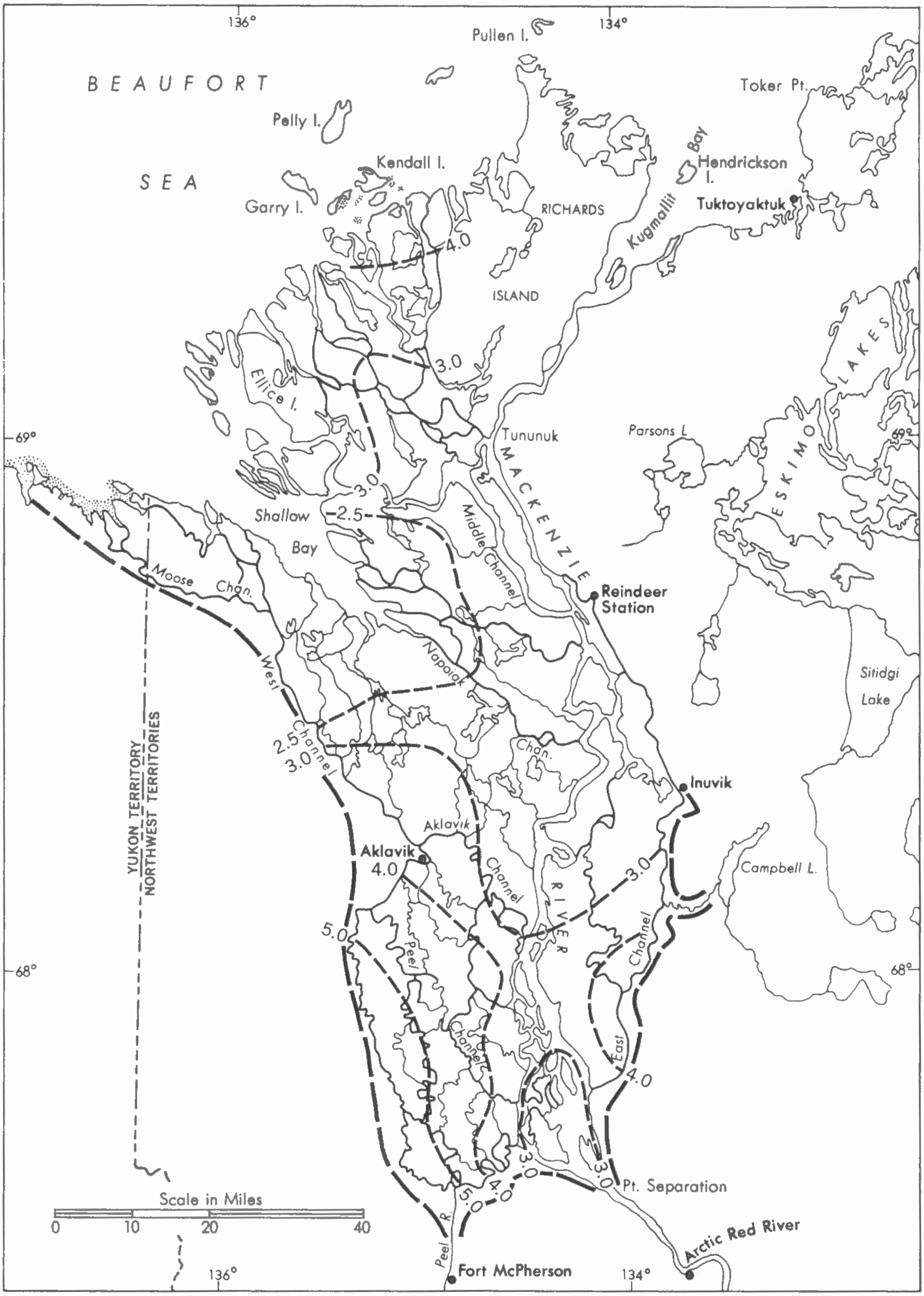

Figure 54. Mean distance in miles between channel junctions.

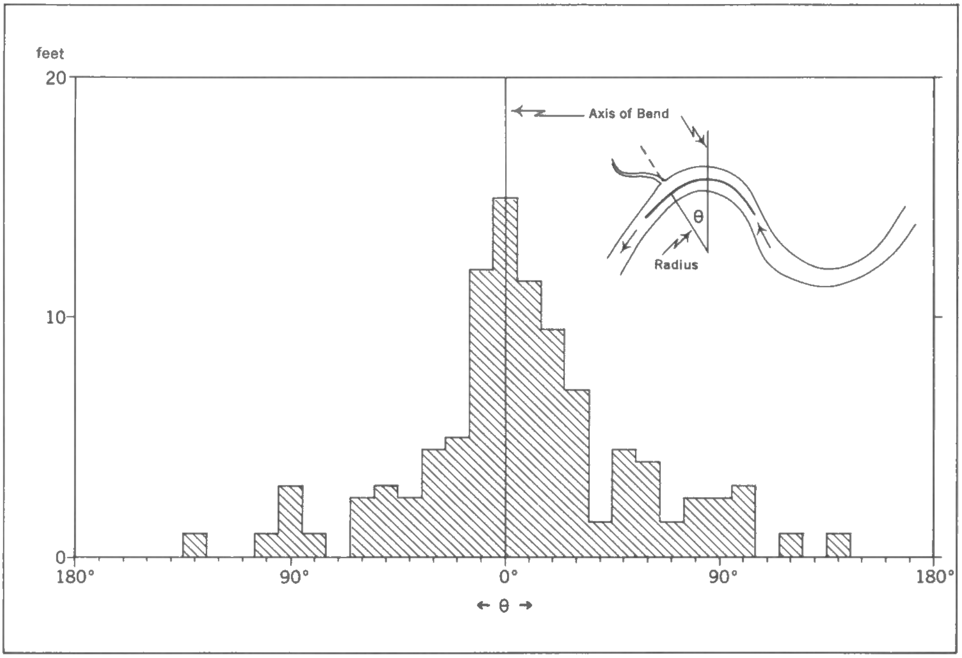

Figure 55. Histogram of junction take-off or entrance angles of 100 channels.

junctions, the altitude of the bed of the distributary channel may be well above that of the main channel, consequently there might be less tendency for the distributary channel to plug up from entrainment of bed-load in the main channel.

108. At every channel junction, the volume of flow in one channel is equal to the sum of the volume of flow in the other two. Thus, in the inset in Figure 56, the flow in the large single channel is equal to the combined flow of the two smaller channels. If Q_1 is the flow in the main channel and Q_2 and Q_3 in the distributary channels, then

$$Q_1 = Q_2 + Q_3 \qquad (108\text{-}1)$$

Numerous studies of canals, rivers, and ephemeral streams have shown a consistent relation between width and discharge, viz:

$$W = a\,Q^b \qquad (108\text{-}2)$$

The value of the exponent b has been established as 0.5 for average downstream relations and about 0.26 at average at-a-station relations (Leopold and Miller, 1956). That is, at discharge of constant frequency of occurrence, width increases downstream as the square root of discharge. However, with a changing discharge at one cross-section of a river, width increases approximately as the .26 power of discharge. In the case of the bifurcating channels, only one width is measurable for each cross-section, so at-a-station cross-section data cannot be compared.

From equations 1 and 2, using b=0.5,

$$\left(\frac{W_1}{a_1}\right)^2 \backsim \left(\frac{W_2}{a_2}\right)^2 + \left(\frac{W_3}{a_3}\right)^2 \qquad (108\text{-}3)$$

If channels W_2 and W_3 are similar, with a_2 nearly equal to a_3, then

$$W^2 \backsim k\,(W^2 + W^2) \qquad (108\text{-}4)$$

where k is a constant.

A regression line was computed for equation 4 based on 35 bifurcations, the line being

$$W^2 = 1.37\,(W^2 + W^2) \qquad (108\text{-}5)$$

The coefficient of correlation (r) is .85 and it is significant at the one per cent level. Other values of b were tried, but as the widths of the channels could not be measured with high accuracy, differences in results were not considered significant. In Figure 56, W_1 is plotted against the sum of W_2 and W_3. The scatter of the points shows that the width of a main channel tends to be slightly smaller than the combined widths of the bifurcating channels. Therefore, the depth of the main channel would normally exceed that of the bifurcating channels in order for it to carry the larger flow within a narrower width, other things being equal.

109. It is often difficult to determine whether a Y-junction is of the tributary or distributary type. This is especially true in the lower part of the delta where the direction of flow is not always apparent. In general, where the Y-junction is of the distributary type, the sides of the main channel at the place of branching are low, somewhat like a slipoff slope. The point of bifurcation is usually rounded, undercut, and has a sand-bar. In the tributary type, the sandspit occurs, but the base of the V tends to be less rounded and the main channel lacks the slipoff slopes.

110. Floodplains and Levees: All of the delta is susceptible to flooding which may take place at break-up or associated with storm surges in coastal areas. In some of the older parts of the delta, flooding is of the "one in a century" type and so rare an occurrence that local inhabitants may never have experienced a flood. As an illustration, for many years prior to 1961, some portions of the delta had been flood-free for so long that they were considered nearly safe from floods. However, a combination of ice-jams and rain resulted in the inundation of large parts of the delta, including Aklavik. In Aklavik, silt was deposited to a maximum depth of 5 feet along part of the shore of Peel Channel. The shoreline was built out some 50 feet, and silt accumulated in all low places in the town (R.C. Timmins, personal communication). The rise in water level was almost 20 feet.

The maximum heights of the floodplains, as measured to the tops of the levees, decrease in a step-like fashion from south to north. The estimated average heights (from field surveys) of levees above the late summer low river levels, are shown in Figure 57. So far as can be determined, there are two rapid drops in levee levels.

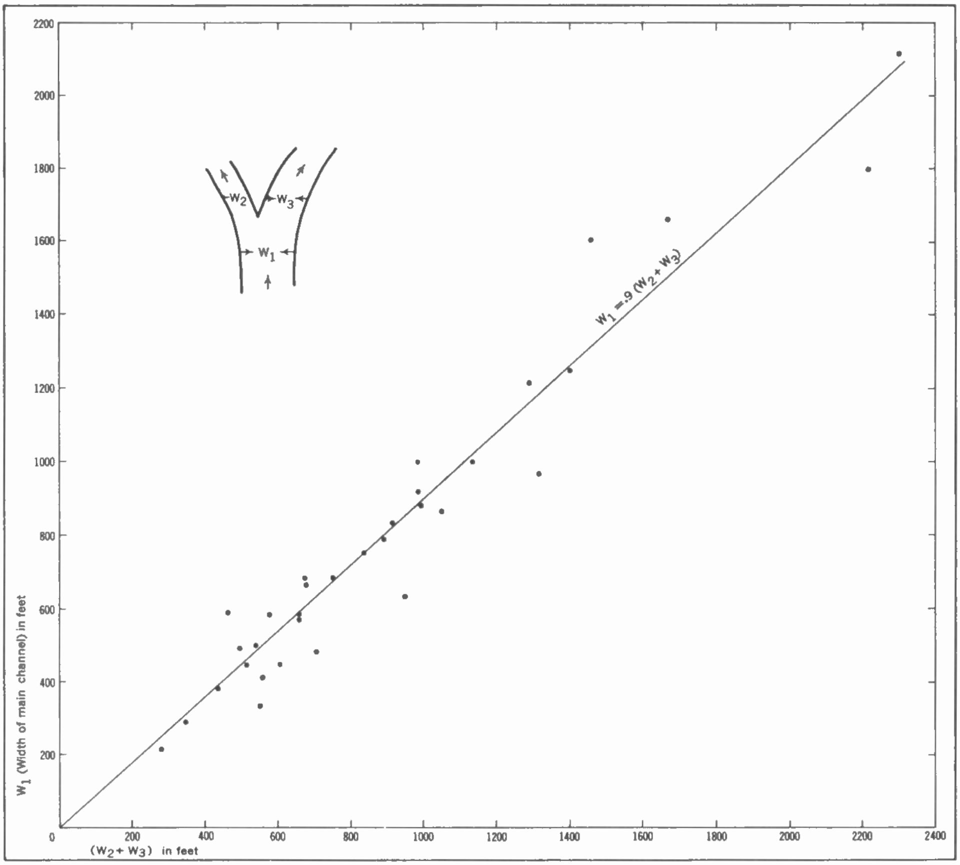

Figure 56. Graph of width of main and sum of minor channels at Y-junctions.

The first is just to the north of the 20-foot isoline where levee heights drop to about 15 feet in less than 10 miles. The second levee descent takes place at the 10-foot isoline where heights drop several feet in a few miles. The second drop corresponds closely with the limit of spruce, most of which are on levees at least 12 feet above low water level. The cause of the first drop in levee level is unknown. The lower level seems to be associated with the limit of break-up-induced flooding. North of the 10-foot isoline, much of the flooding is related to both storm and break-up conditions.

The older and higher floodplains, found in the southern half of the delta, have poorly developed levees. The channel banks are mainly in spruce forest, the ground surface is vegetated, uneven, and hummocky. Such areas may be flooded infrequently, but little sediment is deposited. Levees are common in the northern half of the delta. The crest of the levee tends to be close to the channel, usually within 20 to 50 feet of the low water edge of the channel. The levee crest is usually only several feet wide. The backslopes range from 1° to 2° (Figure 50, No. 2; Figure 51, Nos. 7, 8, 10, 12, 13; Figure 52, Nos. 19, 21).

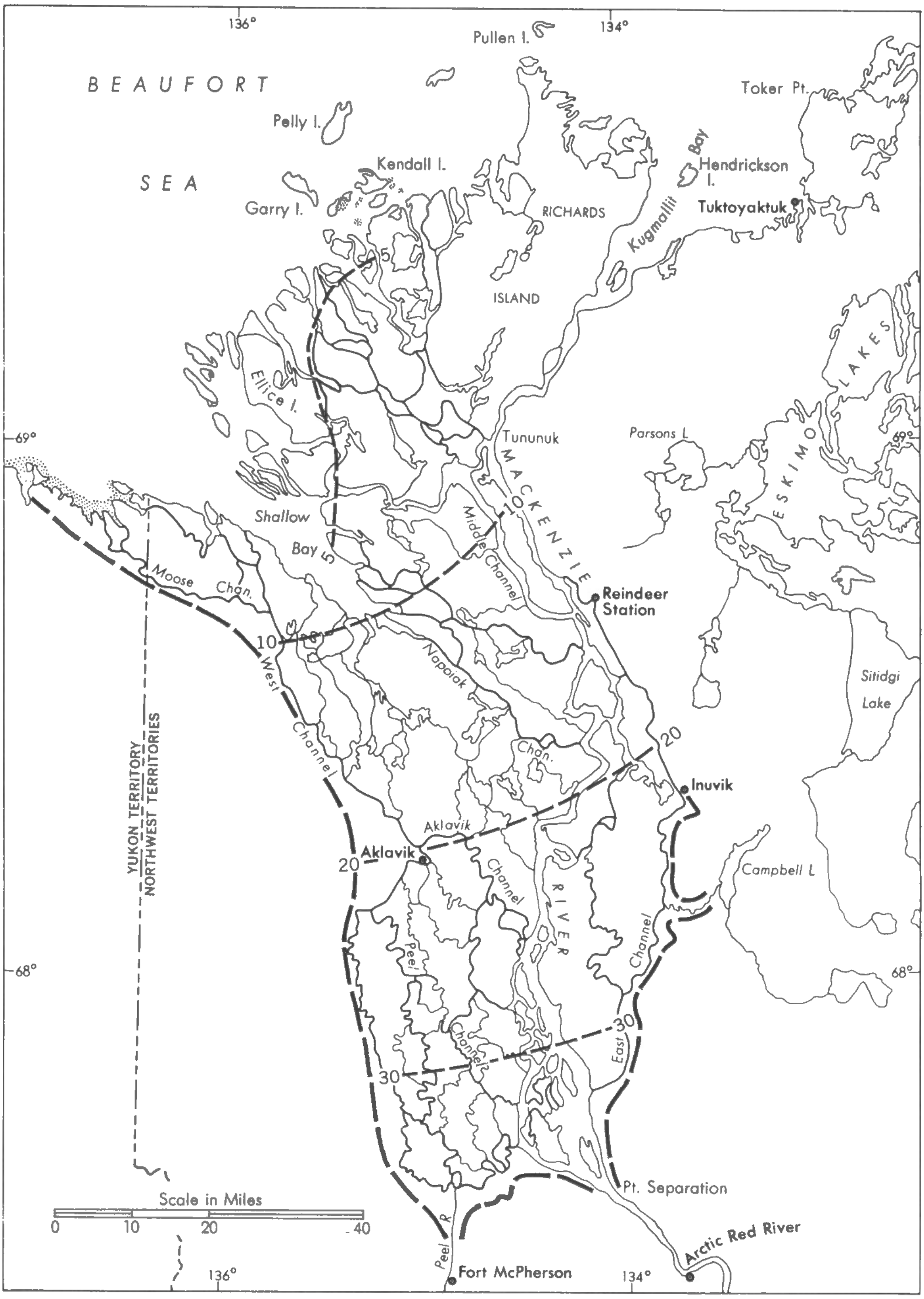

Figure 57. Estimated heights in feet of levees above late summer low-water levels.

Figure 58 shows a detailed survey of a floodplain surface at 68° 27′ N, 134° 19′ W. The edge of the area surveyed was about 20 feet from a cut bank; the datum on the map was 13 feet above river level at time of survey. There was no intervening levee between the map area and the river. Fresh mud burying a discontinuous cover of leaf litter showed that flooding took place one or two years previously. Mud cones had accumulated at the bases of the spruce, willow, alder—including living and dead plants—and stumps. Some of the mud cones were a foot high. The cones tended to be streamlined, being longest and highest on the leeward sides as related to flood currents. For a distance of 200 feet from the river, the tops of the mud cones varied no more than 0.5 feet in altitude, the low spots very little more. Figure 58 shows that in a forested area, which is subject to frequent flooding as shown by the absence of ground vegetation, the terrain is far from flat. As sedimentation proceeds, willow, poplar, and spruce must send out lateral roots, or else die, because the permafrost surface rises with sedimentation. In the surveyed area of Figure 58, stumps of trees buried to a depth of several feet were exposed in the cut bank. Elsewhere, live spruce have been seen growing with over 6 feet of silt having accumulated since growth commenced.

In general, levee construction tends to keep pace with bank recession, even along undercut slopes. The amount of sedimentation after overtopping varies considerably, from thin films to a quarter of an inch or more.

Levees occur along the channels which flow into Shallow Bay where they reach the sea. For instance, levees at the mouths of Tiktalik (Figure 51, No. 8) and West channels rise over 6.5 feet above sea level. Inundation of the levees and the adjoining floodplains which terminate along the coast could only occur if sea level were raised.

Levees are present along some of the channels of the low alluvial islands lying between Ellice and Kendall islands (Figure 52, Nos. 18, 19). However, levee development is virtually absent in areas less than 4 feet above sea level (Figure 52, Nos. 14, 15, 16, 17).

111. Channel Shifts: On a planimetric map, channels appear so sinuous that they immediately suggest meandering rivers which sweep back and forth across the delta, as a meandering channel does in its valley. Most channels do not meander back and forth: they wander. The migrating cut-and-fill river meanders with oxbow lakes and narrow necks are not typical of the delta channels. There are exceptions, of course, such as the upper part of Aklavik Channel, the central part of Middle Channel, and a number of channels in the southwestern part of the delta.

Figure 59 shows the principal areas of channel shifting which are recent enough to map from air photographs. Most of the major channel shifts, south of the tree line, are shown for the past several hundred years, the age being estimated from the vegetation cover. Spruce trees 200 to 300 years old are very common along channel banks. Giddings (1947) reports living trees up to 560 years old in areas near the tree line. In well-treed areas, therefore, temporal and spatial limitations to channel shifting can be determined by the ages of the trees. Where spruce grow on point bar deposits, channel shifting preceded tree growth. Spruce trees lining both banks of a channel show that it has been narrower than the distance between

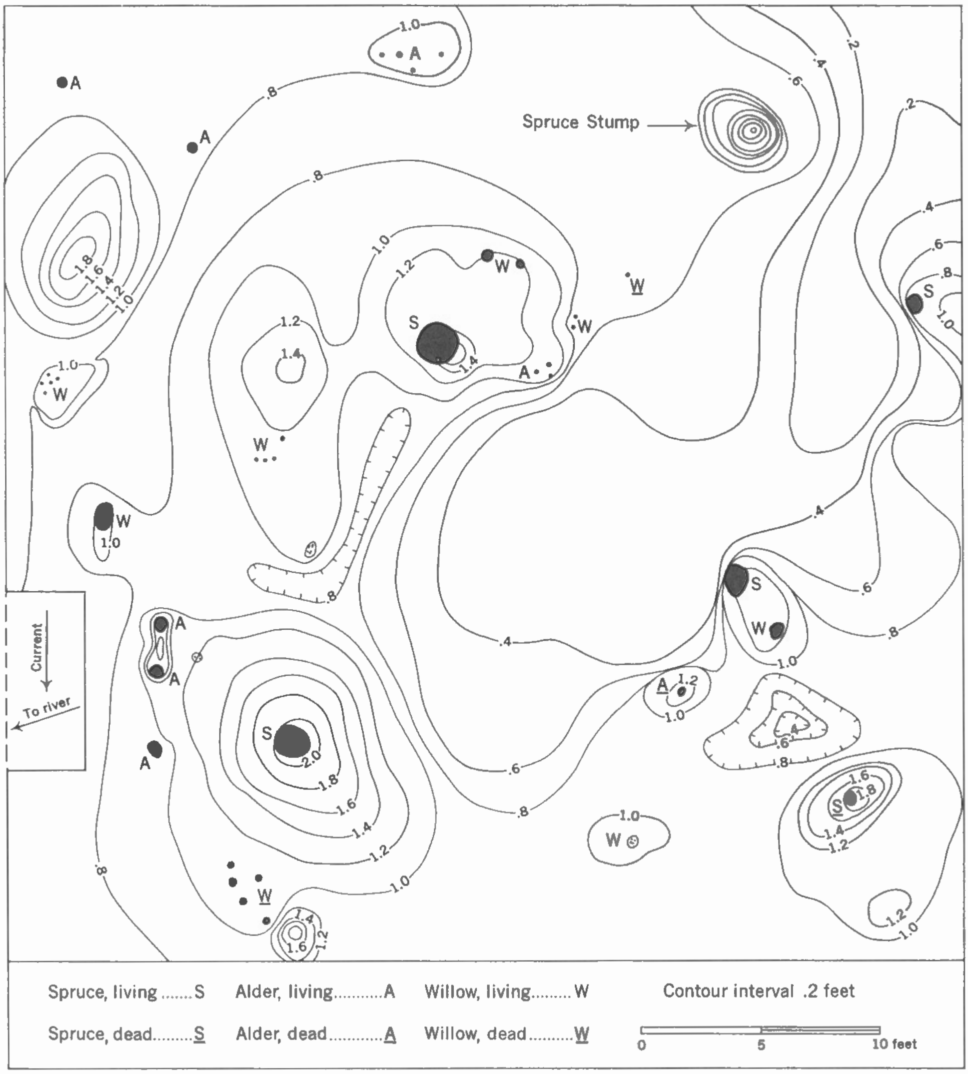

Figure 58. Contour map of floodplain at 68°27′N, 134°19′W.

the trees ever since growth commenced. The evidence derived from the age of forests is against any extensive channel shifting during the past several hundred years. Large tundra polygon ice-wedges also indicate no channel shifting for hundreds to thousands (?) of years where they occur.

The existence of hundreds of small and large irregularly shaped lakes, many with bottoms below low water level in the channels, also implies little channel shifting. Some lakes may be thermokarst features, resulting partially from the thawing of ground ice, but most lakes are unfilled depressions lying between the higher levees of bounding channels. For example, there are several large shallow lakes to the west of East Channel, opposite to Inuvik, where the process of channel growth may

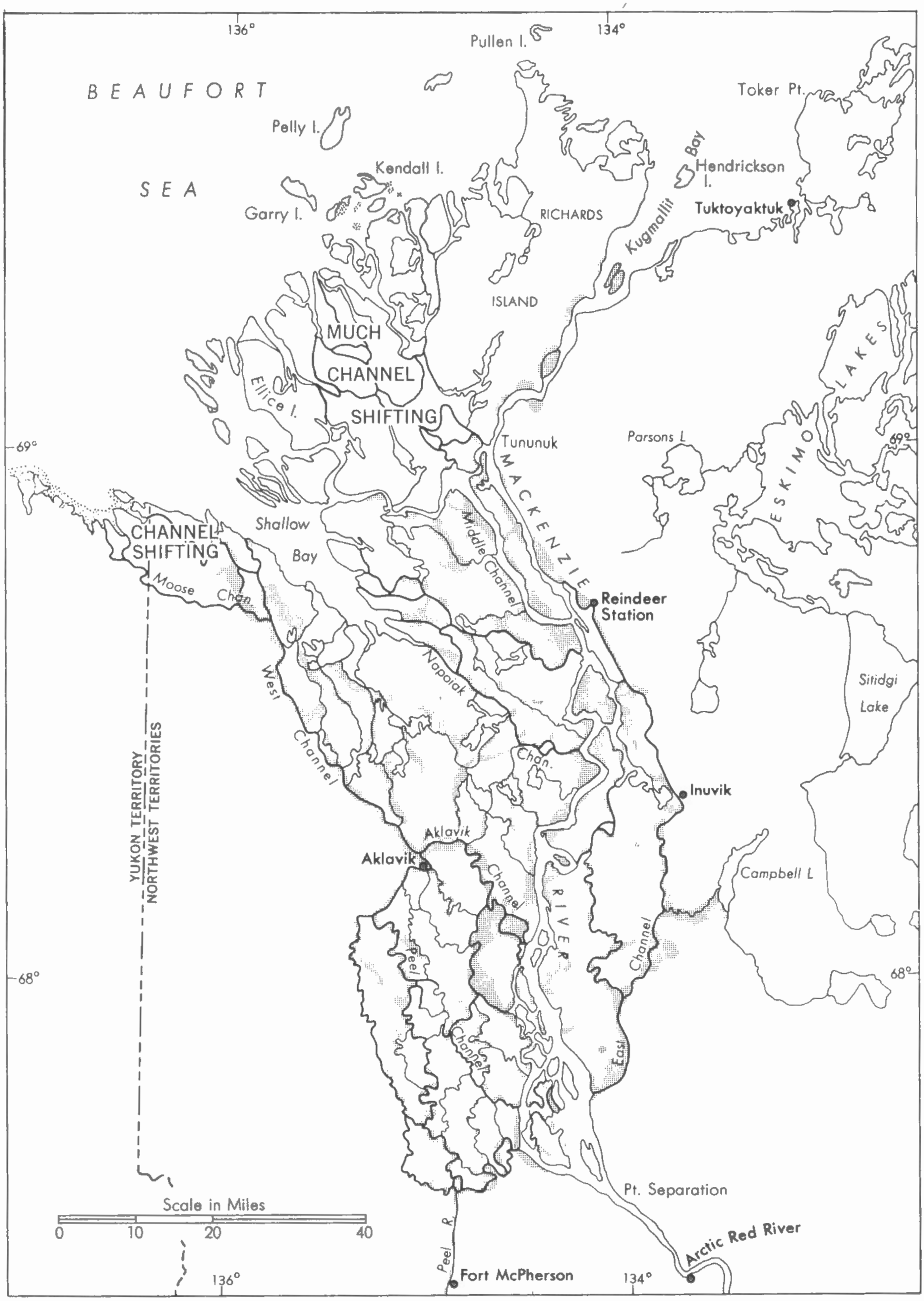

Figure 59. Principal channel shifts mapped from air photographs.

be observed. Areas with what appear to be first-generation broad shallow lakes also occur south of horseshoe bend of Middle Channel (68° 15′ N), in the area from Aklavik northeast to Napoiak Channel, south of Moose Channel, and elsewhere. At periods of high water, parallel levees are gradually extended into many lakes (Figure 73) until eventually channels develop through the lakes, thereby subdividing them. The lakes are also infilled by vegetation, the accumulation of driftwood, sediments, and the freezing of interstitial water. It is evident that no channels have migrated across most of the areas now occupied by large irregular lakes because if they had done so, the lakes could not exist in their present form. With aggradation, a channel could eventually sweep across an area formerly occupied by lakes, but at the present time, the bottoms of many lakes, as in the Aklavik area, are below sea level. Therefore, as long as there has been relatively little land-sea level changes, the existence of lake bottoms at or below sea level shows that channels have migrated little.

Some channel shifting does, of course, occur. Following the high waters of spring break-up, stretches of some channels retreat rapidly. As undercutting and removal of sediment takes place most easily at water level, overhangs develop, with the ground surface bound together by a curtain of turf. Locally bank recession of 10 to 20 feet may result from several weeks' undercutting in early summer. Along the left bank of Peel Channel, where it flows past Aklavik (Figure 72) the estimated bank recession for the 1940-1957 period is over 100 feet, or several feet per year. Much of the sediment load in the delta may be merely that of delta-derived sediment in transport between a cut bank and a nearby point bar, lake, or floodplain.

The large horseshoe bend of Middle Channel (68° 15′ N) is a cutoff meander loop. A comparison of Franklin's map, plotted from his 1825 trip, with the present river shows that he went around the bend (Figure 1) presumably because the cutoff did not exist. Sainville's map (1898) shows the loop as the main channel, with a small channel crossing at the meander neck. There appears to be a good probability, therefore, that the horseshoe bend has been formed in the past 100 years.

The sediments which are deposited after flooding on the point bars tend to be fine-grained. Although any estimate is bound to be very general, most of the point bar deposits sampled tested about 25 to 50 per cent in the silt range.

Lakes

112. The Mackenzie Delta is truly a land of lakes (Figure 39). Perhaps nowhere else in Canada is there an area of comparable size with as many lakes, which number in the thousands. Many of the lakes, particularly those in the southern portion of the delta, are no more than tiny ponds (Figure 43), but towards the seaward perimeter, lakes 2 to 5 miles long abound. The typical lake is irregular in outline, with an indented shoreline. The irregularity of shoreline is partially related to the age of the lake and the bordering vegetation. A newly forming shallow lake on a nearly vegetation-free island tends to have a smooth shore; an old deep lake tends to have an irregular shoreline.

Figure 60

A typical Mackenzie Delta lake of the Aklavik area taken on June 24, 1954, as the water level was gradually dropping, having subsided over 5 feet since break-up. The leafless bushes in the water and on the lakeshore are willows, with spruce on higher ground. (Geog. Br. photo).

Those lakes which flood yearly are either connected with channels or have low closures. These lakes receive yearly additions of silty water if flooding results from a rise of the Mackenzie-Peel rivers rather than from seaward inundation along the distal part of the delta. In the early summer, following break-up, those lakes which have been flooded are easily recognized by their muddy waters which contrast with the unflooded clear lakes whose suspended sediment has had at least one winter in which to settle out.

Delta lakes are shallow. Few exceed 10 feet in depth, when measured at low water. Depths at high water are much greater, and can be estimated by noting the high water marks on lakeshore willows. The high water marks show up as broad bands encircling the lakes.

The delta lakes can be classified into at least five broad types, primarily on a genetic basis.

113. Abandoned Channel Lakes: In the delta, with its hundreds of miles of channels, individual loops or long stretches may become abandoned (Figure 43). The abandonment of a channel may take place in a season, such as on the rare occasion where the neck of a loop is breached, or the process may extend over many years, where gradual changes cause less and less current to flow through a channel until flow ceases. Whether cessation of flow is slow or rapid, eventually both ends of an abandoned channel tend to silt up, in much the same way as plugs form in meander cutoffs.

Abandoned channels are easily recognizable in the field, on air photographs, and maps. A long-abandoned channel soon becomes segmented (Figure 74). Silting up is often at the inflection points, so that the abandoned system resembles a link of sausages. An abandoned channel of recent occurrence may be seen in the loop of Aklavik Channel severed by cutting of Middle Channel. Large abandoned channels may continue to carry a greatly reduced volume of water, with underfit dimensions resulting. Thus a small crooked channel may link two long, wide, and straight reaches of a large abandoned channel.

Abandoned channels, even when silted up at the ends, may carry flood waters and so form part of an integrated flood drainage network. With periodic accretions of sediment, the channels become shoal, the banks silt up so that former undercut and slipoff banks are obscured, and aquatic vegetation takes over along the sloping banks. Even when the topographic form of the channel is nearly gone, vegetation differences may betray its presence.

114. Arcuate (Point Bar) Lakes: Migrating channels may leave a swell and swale ribbed pattern, marking flood deposition on convex point bars (Figure 72). The swells and swales are subparallel, their curvature matching the radius of the flood channel which built them. The smaller swales may be occupied by shallow ephemeral lakes, whereas the larger swales resemble small abandoned dead-end channels. Sizes range from a length of a few tens of feet to 2 miles. Good examples may be seen along the left bank of East Channel, after it branches from Middle Channel; along the unnamed channel lying between 68° 00′ N and 68° 10′ N and 134° 40′ W and 134° 50′ W; and along the last 5 miles of Aklavik Channel before it joins West Channel.

Arcuate lakes differ from abandoned channel lakes in origin and shape. Arcuate channel lakes are in series and subtend angles which rarely exceed 90° whereas abandoned channels may oscillate through several cycles.

Arcuate lakes are short-lived in comparison with many other delta lakes. Being shallow, they are easily reduced in depth and area by deposition of sediment. They also decrease in size by the encroachment of aquatic vegetation succeeded by willow, alder, and spruce (in the southern part of the delta). Finally, only the topographic form of the swell and swale pattern, as revealed by vegetation differences, may remain.

115. Floodplain Lakes: Well over 90 per cent of all the delta lakes are floodplain lakes which occupy depressions rimmed by the higher land along the channels (Figure 43). The process of lake formation is observable in its initial stages in the low outer alluvial islands and that of lake obliteration in the large lakes to the south. The floodplain lakes are divisible into three main types: newly formed island lakes occurring on the seaward alluvial flats; lakes with closure; and open lakes with channel flow.

116. The seaward section of the lower delta, which lies north of Shallow Bay, is subject to inundation from both channel overflow and a rise of sea level. This belt is about 20 miles wide. The highest altitudes—excluding pingos—are estimated at less than 6 feet above sea level. Indeed, the coastal 10-mile wide zone is mostly below 4 feet in altitude. Where distributary channels reach the sea, a shoal frequently grows directly off the channel mouth. The distributary channel tends to bifurcate around the shoal which evolves into a horseshoe shape between the two branches of the channels. Strickland (1940) has referred to this type of shoal as a "dwip" and attributes it to reversing tidal currents (*see* also Bates, 1953, pp. 2158-2159). Each arm of the dwip tends to increase in length and height. Should the arms join, a lake may form in the depression between them. Storm waves increase the heights of the seaward sides of the bars. Channel flooding also adds sediment to build the land higher adjacent to the channels, although precise levelling shows that true levees do not form (*110*). Broad distributary mouths, which may be over 2 miles wide, also silt up, to increase the breadths of the islands whose shores form their banks. Streamlined shoals and islands also grow in estuarine mouths, such as that of East Channel near Kittigazuit. The end result of sedimentation, whether in the form of dwips, new islands, or the increase in area of existing islands, is to form new alluvial areas, several feet above sea level. The interiors of these new areas, being farther removed from channel overflow, tend to remain as basins, often with lakes.

The island lakes are shallow. As the lake bottoms are usually above sea level and the islands are generally less than 5 or 6 feet above sea level, most lakes are no more than several feet deep. The shallowest lakes freeze to the bottom in winter and permafrost probably forms beneath them. Frozen ground may lie sufficiently close to lake bottoms in summer that it provides firm footing for walking. Shorelines are normally smooth in outline where the fluctuating lake edge terminates against bare soil; otherwise, the growth of grasses, sedges, horsetails, rushes, and other aquatic plants tends to produce indented to serrated shorelines.

117. Lakes with closure form when the alluvial islands are built high enough for levees to become established along channel banks (Figure 72). As the rate of infilling by sedimentation and growth of organic matter is generally less rapid than levee growth, for a period most lakes are deepened and enlarged. The process does not continue indefinitely. The bigger lakes become partitioned into smaller lakes by infilling to increase lake density but decrease mean lake size. Moreover, the swing of a channel may breach a lake or an outflow channel may partially drain it. Those lakes with a low closure tend to flood annually, whereas those with a high closure flood infrequently.

118. Numerous floodplain lakes have no closure, in the sense that they open out to, are joined with, or are crossed by channels. Many of these lakes form from the breaching of lakes with closure. These lakes receive yearly additions of sediment (*119*). Their areas and depths fluctuate, being greatest at break-up and decreasing thereafter (Figure 60). Flow into or out of the lakes responds to channel levels.

A common sight for a person travelling along a channel is to see a sag in the otherwise horizontal bank level. As viewed from a boat in the channel, the floodplain inland of the sag is usually without trees or bushes, thus a sign of a lake. The sag is caused by lateral cutting into the floodplain backslope of a lake basin. With further cutting, the lake is joined to the channel.

The flow of a channel into a lake is often accompanied by delta building into the lake (Figure 73). The delta may be cuspate, or marked by parallel levees extending far into the lake. Differences in turbidity between lake and channel waters may show up on air photographs as jet flow through the levees. The predominantly northwest winds cause asymmetry of the deltas, irrespective of the directions of entry of the channels into the lakes. In those lakes with two connecting channels, through flow may occur. Levees which form along the course of the through flow eventually subdivide the lake and thus create a new full-fledged delta channel.

119. Estimates of the rapidity of lake sedimentation can be sketchy, at the very best. In May, 1959, two sediment pans, measuring 8 by 8 inches on the bottom and with 2-inch sides, were placed (through the co-operation of K. H. Lang) in two interconnected lakes near the junction of Peel River and Phillips Channel. On September 25, 1959, the pans were examined. The first pan, located in the center of a lake half a mile from Peel Channel, was so covered with mud that it could not be found. The second pan, placed over a mile from Peel Channel, had sediment about .02 inches in thickness. These results suggest that for the lakes under examination, 1 to 2 inches of sediment may have been deposited in the first lake, the thickness decreasing rapidly to .02 inches in the remotest lake.

120. In the summer of 1956, observations were made on 48 lakes by J. E. Bryant and P. Jolicoeur (personal communication). From 12 to 26 July, 32 of the lakes were studied within a radius of 10 miles of Aklavik, and from 10 to 14 August, another 16 lakes near 67° 45′ N, 134° 55′ W, in an area near the junction of Husky and Phillips channels. Large and small lakes were sampled, ranging from those covering more than 200 acres to tiny lakes an acre in area. No information is available on whether specific lakes had closure or were joined to channels. Data on the lakes are given below:

	Mean Depth (*at stated distance from shore*)			Standard Deviation of Depths (*at stated distance from shore*)		
	15′	50′	Centre of Lake	15′	50′	Centre of Lake
Aklavik Lakes	2.6′	4.4′	5.6′	1.0′	2.1′	2.1′
Husky—Phillips Channel Lakes	2.3′	4.5′	5.0′	1.2′	2.8′	2.7′

On a statistical basis, using the "t" test between means at a 95 per cent confidence level, there is no significant difference in the mean lake depths of the two groups. However, the Aklavik lakes were sounded earlier in the summer, when water levels are normally higher, so that the differences are probably greater than shown by the table. The variations of depth, as suggested by the standard deviations, are small. The ranges in depth are given below:

	Ranges in Depth (*at stated distance from shore*)		
	15′	50′	Centre of Lake
Aklavik Lakes	1–6′	1–11′	1–11′
Husky—Phillips Channel Lakes	1–4′	5–12′	2–12′

The typical lakes, as exemplified by the Aklavik and Husky-Phillips channel lakes, are then shallow, with mean depths in the center of about 5 to 6 feet.

Maximum depths rarely exceed 10 feet. However, depths in excess of the winter ice thickness must be present in a large number of lakes, because the delta is an important muskrat trapping region, and muskrats live in lakes with unfrozen pools (Stevens, 1953).

121. The over-all aspect of the delta lakes results from a contest between levee building, infilling, thermokarst melting, and lateral cutting. Levee building is probably the major control. This may be illustrated by comparing the river levels, at low water, with floodplain heights. In 1954, the Topographical Survey of Canada established a line of levels from Arctic Red River to the coast. As the survey was carried out in June and July the water levels were higher than normal late summer and winter levels. The altitudes of the bench marks, which were usually located close to the ground, are believed to represent quite accurately the heights of the floodplains above sea level. Where channel levels were not surveyed, estimates of water levels derived from linear interpolation have been given. According to the table, given below, the differences between bench marks and channel levels diminish from 30 feet in the southern part of the delta to less than 5 feet at the coast. The data agree reasonably well with the heights of levees above channel levels as measured in the field (Figure 57). Consequently, the possibility for a deep lake to occur is greater in the southern part of the delta because the range in altitude of levees above the channels is the greatest.

Location		Bench Mark (*in feet*)	Water Level (*in feet*)
67°39′N	134°07′W	47	15
67°45′N	134°12′W	41	13
67°54′N	134°25′W	33	12 (est.)
67°59′N	134°28′W	29	11 (est.)
68°12′N	134°44′W	28	10 (est.)
Aklavik		22	5
68°21′N	135°20′W	16	4 (est.)
68°29′N	135°31′W	16	3.7
68°33′N	135°32′W	15	3.0
68°45′N	135°35′W	6	1.5 (Shallow Bay)

122. Thermokarst Lakes: Thermokarst or cave-in lakes are those which occupy depressions resulting from subsidence caused by thawing of ground ice. The delta has few of the cave-in lakes typical of some other subarctic areas (Wallace, 1948). These lakes are recognizable by leaning or "drunken" trees tilted over the banks of lakes, often with submerged, or partially submerged, trees offshore. The leaning trees tilt lakewards, as bank recession engulfs the lakeshores. The partially submerged trees, which have continued to grow since bank recession dropped them into the enlarging lake, have bends below their upright crowns, an indication of the amount of growth since submergence. The bends indicate the positions of the tree tops after they came to stable rest on the lake floors. Estimates of the rate of bank recession can be made by dating the ages of the erect crowns.

Thermokarst lakes grow best in soils with a high ice content, particularly fine-grained and organic soils. Once equilibrium conditions are disturbed, as by a rupture of the surface cover, bank erosion, or a marked increase in lake depth, thawing of ground ice may take place. The thawing can result in further bank recession and in deepening. This may be illustrated by using Aklavik soils (Pihlainen and others, 1956) as an example. Ice contents at a depth of 10 to 30 feet are nearly constant at 50 per cent (weight of ice to dry soil). In a small shallow lake, in which the winter ice freezes to the bottom, permafrost will normally lie within a few feet of the lake bottom. If the lake becomes deeper and larger, so that the central portion remains unfrozen in winter, the permafrost surface will degrade. A melting of 10, 20, or 30 feet of frozen soil might then result in appreciable settlement and lake deepening. Conversely, a shoaling of a lake may cause a rise in the permafrost surface, the formation of ground ice in situ adding a further shoaling to the lake.

123. Dammed Lakes: Along the land boundary of the delta, streams entering the delta may be partially dammed to impound lakes (Figure 73). As the delta floodplains aggrade, the dammed lakes impart an estuarine appearance to some river mouths. Although the estuarine aspect of most of the dammed lakes reflects a former period of delta emergence, nevertheless, any significant rise in floodplain level will create the same general effect.

Pleistocene Coastlands: Region III

124. The coastlands of the Beaufort Sea between Kendall Island on the west and Cape Dalhousie on the east consist mainly of Pleistocene fluvial and deltaic deposits. The southern limit is about 10 to 20 miles south of the Eskimo Lakes. Most of the area lies below an altitude of 200 feet, with about 50 per cent below 100 feet (Figure 38). The highest altitudes are in southern Richards Island, the areas adjacent to the Caribou Hills, and in an irregular belt on the north side of the Eskimo Lakes between Parsons Lake and Campbell Island.

The Pleistocene coastlands have numerous lakes which cover over 15 per cent of the surface (Figure 39). With minor exceptions, pingos are found throughout the entire area. Large tabular sheets of ground ice are equally widely distributed, omitting, however, the portion of Tuktoyaktuk Peninsula east of Hutchison Bay, where they are virtually absent. The vegetation ranges from tundra along the coast through scrub tundra farther inland to patches of open woodland in the Sitidgi Lake area (Figure 70).

125. Tununuk Low Hills: Section III a: The Tununuk low hills include the western part of Richards Island, the outliers in the Mackenzie Delta; Kendall, Garry, Pelly, Rae, and Hooper islands; Tununuk Island south of Tununuk; and a strip of land south of East Channel. With the exception of the outliers, altitudes range from 100 to 200 feet above sea level.

The outliers in the Mackenzie Delta are eroded detached portions of Richards Island. The outliers lie east of a line joining Tununuk with the northwest point of Garry Island. Considerable areas are flat, with a predominant altitude of about

50 feet. Garry Island, rising up to 200 feet, stands isolated out-to-sea. Kendall Island, on the other hand, is nearly united to the Mackenzie Delta by island-building. Pelly and Hooper islands rise to about 100 and 200 feet, respectively. Both have many ground ice formations and their coasts are receding rapidly. Tiny Rae Island, 50 feet high and a third of a square mile in area, is also being eroded away.

The western side of Richards Island is irregular, with broad drained flats and drowned valleys. Piles of driftwood have been swept into bays and lodged along cut bluffs. In the southern part of Richards Island there is a long estuarine lake ("Yaya Lake") which resembles a drowned river. As glaciofluvial gravel ridges drape down the banks to within a few feet of lake level, the lake trough is not entirely a postglacially cut and drowned valley. A bench 45 feet in altitude partially surrounds the lake. South of "Yaya Lake", there is a V-shaped unnamed lake which is connected to "Kendall Island Channel" on the south and separated from "Yaya Lake" by relatively high land on the north, but is nearly joined to East Channel on the east. A rise of roughly 10 to 15 feet would permit flood waters to connect East Channel with "Kendall Island Channel" via the V-shaped unnamed lake (Figure 7). If the lake formerly served as a channel, it seems unusual that the northern arm of the lake was not infilled with sediment and sealed off.

One third of a mile south of the southernmost extremity of Richards Island, there is a remnant of Pleistocene deposits rising prominently as a 200-yard-long 75-foot-high island in the middle of East Channel. This island was observed by Mackenzie, who mentioned that it "possessed somewhat of a sacred character" to the Eskimo, an observation also made later by Richardson.

The area south of East Channel has a discontinuous west-facing erosional (?) bluff at 150 feet in altitude some 5 to 10 miles inland. The boundary zone with the Caribou Hills is marked by old drainage channels, glaciofluvial deposits, and dead ice terrain. Holmes and Pete's creeks are aggrading their drowned mouths. Draining of land by Pete's Creek has assisted in the growth of a pingo field.

126. Kittigazuit Low Hills: Section III b: The Kittigazuit low hills section differs from the Tununuk section to the west in the presence of relatively deep lakes with smooth shorelines. There is a slight northeast-southwest predominant trend in lake orientation, which may have been caused by glaciation. As flat drained depressions are sparse, so also are pingos. A drowned interglacial (?) valley has its mouth at the abandoned site of Kittigazuit.

For convenience, Pullen and Hendrickson islands are included in this section. Pullen Island rises to about 125 feet and is being rapidly eroded. Hendrickson Island is mostly below 15 feet, except for the pingos. The material is sand.

127. Undifferentiated Coastlands: Section III c: The undifferentiated coastlands comprise 6 separate areas undistinguished by any single feature. The areas have medium to large lakes, flats with pingos, and low hills. The coastal areas along the Beaufort Sea are highly indented, but spits are extending across many bays to smooth off the shoreline. Small patches of pitted outwash and dead ice terrain occur in areas adjoining the morainic hills (III f). The eastern subsection, between McKinley and Liverpool bays, has several large lakes showing some orientation and also drowned valleys along the Liverpool Bay coast. A small section fronts on Liverpool Bay, due south of Cape Dalhousie and northwest of Nicholson Peninsula.

128. Involuted Hills: Section III d: The involuted hills resemble, on air photographs, the wrinkled skin of a well-dried prune (Figure 30). The wrinkles are curving to branching ridges, ranging up to several hundred yards in length, several score yards in width, and 20 feet in height. When seen from a lower altitude in the field, some involuted slopes appear to rise in a succession of terraces. The three principal areas of involuted hills are 50 to 200 feet high with a relief of 100 to 150 feet. The involutions are on the flat hill tops and not in the intervening depressions.

The involutions are caused by the development of extensive ground ice, both tabular ice-sheets and tundra polygon ice-wedges, along with partial melting and slumping. The ridges are most prominent on the sides of the hills where there has been the greatest amount of thermal erosion. Some of the ridges resemble small ice-cored eskers which gradually disappear as the ice-cores melt. The ground ice has grown in situ in fine-grained soil, probably till or reworked till, judging by the overburden cover with striated erratics.

129. McKinley Bay Coastal Plain: Section III e: This area is mostly less than 50 feet in altitude, with a narrow coastal strip lying below flood level. Large and small oriented lakes are numerous (Figure 12). Parabolic and blown-out parabolic dunes are also common (Figure 61). There are numerous pingos but few ground ice-sheets, or at least visible signs of them. The westernmost part, south of Hutchison Bay, is a low flat embayment surrounded by terrain rising up to 150 feet in altitude. From Hutchison Bay to McKinley Bay, three quarters of the surface is a flat featureless tundra polygon plain. The transition from flat plains to the morainic hills in the south is at 50 to 75 feet, the topographic change being abrupt. Fixed longitudinal and parabolic dunes are present in the Point Atkinson region.

Oriented lakes and parabolic dunes are abundant east of McKinley Bay. Large tundra polygon flats are few. The southern boundary is 50 to 100 feet in altitude. In places, low sandy ridges, a few feet high and up to a mile long, parallel the boundary on the north (coastal plain) side of Section III c. The ridges show up as white strips on air photographs. They may represent beach (?) ridges.

The coast is indented, particularly east of McKinley Bay where coastal recession of the oriented lake plain has caused an interfingering of land and water. The bays are remnants of oriented lakes. Offshore waters are shoal, so that spits and bars are numerous. Offshore bars parallel the coast at distances up to several miles principally to the southwest of Point Atkinson, on the east side of McKinley Bay, and along the coast from Cape Dalhousie south. Some shoals lie 5 to 10 miles from shore.

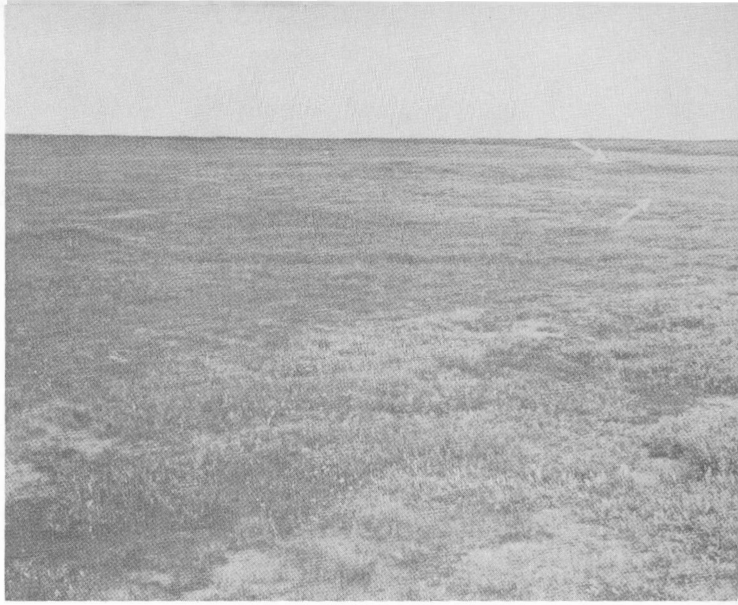

Figure 61

Terrain on the north side of Liverpool Bay at 69°53′N, 129°55′W, taken August 6, 1960. The low plants with large leaves (foreground) are willows; other vegetation includes grasses, sedges, avens, arctic heather, mosses, and lichens. The surface material is a fine sand. The limbs of a small 3-foot-high parabolic dune are marked with arrows. (Geog. Br. JRM-60-4-9).

130. Morainic Hills: Section III f: The morainic hills are in two separate sections, but both are characterized by a higher and rougher topography than is typical of the area of Pleistocene sediments. Some of the region may be underlain by pre-Pleistocene sediments, particularly in the southern part near Parsons Lake. Numerous hills rise up to 200 feet, many being capped with sand and gravel hillocks of glacial origin. In several areas, as near Parsons Lake, fluting occurs. Lakes are irregular in shape; drained lakes are infrequent, so that pingos are sparse. There are few ground ice slumps. Interior drainage is common.

131. Pitted Outwash Plains: Section III g: The pitted outwash plains are composed of many flat-topped mesa-like areas interspersed among numerous large and small irregularly shaped highly indented lakes. Most of the area lies below an altitude of 250 feet. The outwash sands and gravels are reddish brown, usually less than 15 feet thick, and form a capping over Pleistocene sands and gravels. Dead ice and morainic terrain are interspersed with patches of outwash, kettle holes, and glaciofluvial deposits to make a complex jumbled landscape. The lake shores have wave-cut terraces which are most numerous within 100 feet of sea level. Considerable erosionally induced slumping of tundra polygons along the edges of the flat-topped plains has fashioned a series of terracettes which drape down the slope parallel to the contour. Although the sediments are mainly Pleistocene, older deposits may occur in the south and east. The region is almost park-like in its vegetation, with open stands of spruce.

The creek connecting Sitidgi Lake to the Eskimo Lakes is deep enough to allow boats drawing several feet to enter Sitidgi Lake. The floor of the Eskimo Lakes area in III g has an irregular bottom. A depth of 170 feet was obtained midway between the mouth of "Stanley Creek" and the opposite shore.

132. Eskimo Lakes Area: Section III h: The Eskimo Lakes are partitioned into several large water bodies by arcuate peninsulas, projecting into the lake from north and south shores. These peninsulas are puzzling features of uncertain origin. There are three main finger groups; from west to east, the first is centered at 132° 55′ W, the second at 132° 25′ W, and the third at 131° 00′ W (Figure 62).

The first finger group comprises several subparallel ridges which are less than 100 feet in altitude. Portions of two ridges are drowned to form curving islands. A north-south channel, narrowing to a width of 150 yards in the south, is the only through connection. The shallow channel is from 10 to 25 feet deep. As tidal effects in this part of the Eskimo Lakes are negligible, tide rips and tidal scour are absent. The channels resemble drowned river valleys. Depths in the Eskimo Lake between the first and second fingers are under 30 feet, at least in the southern half. The bottom is fairly flat.

The second finger system has two long arcuate peninsulas, and four principal segmented ridges. The interfinger channels terminate in bays in the north, but on the south, they grade through long narrow "drowned valleys" into strings of lakes occupying oversized glacially modified valleys through which no streams of consequence now flow. The over-all finger-channel pattern is arcuate, concave to the west. The two prominent north-south peninsulas are mostly under 50 feet in altitude. However, all the interchannel ridges increase in height southwards to altitudes of 100 to 200 feet, until the ridge-channel pattern disappears. There is only one through channel and it narrows to 250 yards. Depths in the principal north-south channel and the mile-long east-west channel are remarkably uniform, varying only between 20 to 25 feet. In the channels, the tide is about 6 inches. Strong tidal currents sweep through some channels; standing waves, several feet high, have been observed on calm days.

In the large lake between the second and third set of fingers, the hydrographic pattern of the south shore shows a series of long "drowned" valleys decreasing in size to the southwest as they merge into lakes and "dry" valleys. Lake depths of 30 to 75 feet indicate an uneven bottom, as far east as 131° 50′ W. The very uneven bottom of the easternmost part may indicate ridges and troughs. The greatest depth exceeds 200 feet.

The third set of fingers is the most symmetrical. Like the others, the fingers terminate to the north but may be traced far inland to the south, some being 30 miles long. The peninsulas are 50 to 150 feet in altitude. The channels are relatively deep. Where channel widths exceed half a mile, depths range from 50 to 150 feet. The wider northern parts of the channels are over 100 feet deep.

The tide in the western part of Liverpool Bay is about 1.5 to 2 feet; it is about 6 inches in the Eskimo Lake between the second and third fingers. The area of the lake is about 340 square miles. The flow into the lake, via the third set of fingers, is through a channel 0.9 miles wide with a mean depth of 50 feet. If the flow through the channel reverses every 6 hours, and raises (or lowers) the 340-square-mile lake some 6 inches in the process, then the flow averages about 200,000 cfs. Even if the flow figure is several times too high, the flow through the channels is still comparable to that of the Liard River in summer. At storm time, the water in the western side of Liverpool Bay may be raised several feet above high tide level, and this causes a rise in the Eskimo Lakes of a lesser amount. The maximum rise in the large 340-square-mile lake, discussed above, is estimated from beach

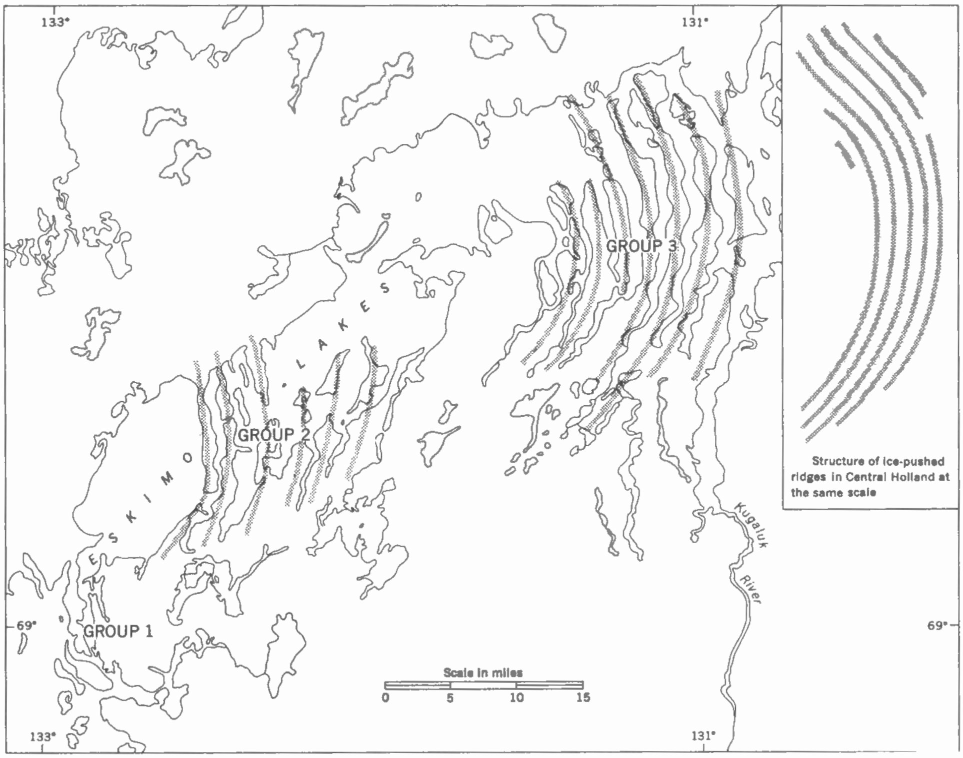

Figure 62. Eskimo Lakes area. The inset map shows the structural trend of ice-pushed ridges in Central Holland at the same scale (after Pannekoek, 1956 and Rutten, 1960).

debris at about 2 to 3 feet above high tide level. Therefore, a strong current may flow through the narrow channels during storms when sea level is raised in Liverpool Bay. The scouring effect of the currents on the channel bottom topography is unknown.

133. Any discussion on the origin of the arcuate fingers must be speculative. However, the arcuate shape of the ridges, their trend which is nearly normal to the inferred trend of the last ice advance, and the presence of shear planes all suggest that glacial ice-thrusting may have helped to form the ridges. An open anticline with apparent dips of 5 to 10° is exposed for a distance of 300 yards in a wave-cut section along the end of a peninsula at 132° 32′ W (Figure 5). The material is of interstratified sands, silts, and silty clay. Several poorly defined peninsulas are present along the north shore, one at 132° 00′ W, showing many shear planes cutting across sand and gravel beds. Shear planes cut stratified sands and silts at the north end of the most northerly extending peninsula. Arcuate ice-pushed ridges, of nearly identical size and shape, have been formed by ice-shove in Pleistocene unconsolidated sediment (Pannekoek, 1956, pp. 88-90; Rutten, 1960) and similar ridges have been reported from elsewhere. There is no evidence to indicate that the ridges, if ice-pushed, were formed during the last ice advance.

On the contrary, as the southern extensions of some interfinger channels resemble glacially modified river valleys, the finger system may predate the last ice advance.

134. Coastal Area: Section III i: The coastal area rises from Liverpool Bay in bluffs 25 to 100 feet high, the terrain inland ranging between 50 and 150 feet in altitude. If bluff exposures are a representative sample of the deposits, the sediments are composed of foreset sands, overlain by topset sands and gravels with some driftwood (Figure 2). Offshore areas are shoal; depths may be only 1 to 2 feet at a distance of several hundred yards from shore. Sandspits extend west from any promontories. The lakes, like those of III b, have smooth shores although irregular shapes. Valley mouths are drowned and nearly enclosed by bars.

135. Rufus Lake Area: Section III j: This section is believed to mark the change from Pleistocene to Cretaceous (?) sediments inland, the evidence being based partly on photo interpretation. The area east of Rufus Lake has rolling hills with large and partially drained lakes. The area southwest of Rufus Lake is similar, except for more irregularly shaped lakes.

The mouth of Kugaluk River is an estuary with depths of about 35 feet or less. A small delta is building out into the estuary. Several terrace levels occur on either side of the estuary.

136. Kaglik Lake Area: Section III k: This section is believed to be mainly in pre-Pleistocene sediments. There are numerous irregularly shaped lakes. Pingos are absent, in contrast to areas IIIi and IIIj to the north. Ground ice slumps occur, but not in abundance. This section is transitional to region VII.

137. Nicholson Peninsula Area: Section III l: Nicholson Peninsula (Mackay, 1956 b; 1958 b, pp. 49-51) is a topographic anomaly as it rises to 300 feet, which is considerably higher than adjacent areas of Pleistocene sediments. The exceptional height is attributed to glacial ice-thrust. The northern half is hilly, the southern half is flat and low. The exposed north and northwest coasts are being eroded quite rapidly. Large ground ice-sheets, when exposed, contribute towards excessive recession of the north coast. The south side is low with banks 10 to 15 feet high showing many exposures of laminated sand and silty sand with twigs, shells, organic rich laminae containing horsetails, sphagnum moss, pebbles, etc. The sections look similar to those exposed along cut banks of alluvial deltaic islands in the area. The isthmus at Nicholson Peninsula is 50 feet wide at its narrowest part. Storm waves sweep right over the isthmus, which may soon become just a gravel bar.

Caribou Hills: Region IV

138. The Caribou Hills are a rolling to hilly area cut by broad melt-water channels. The age of the sediments is unknown, although it has been suggested that they resemble Tertiary lignite-bearing formations from other parts of Northern Canada and Alaska (Kellaway, 1956), but they may be older. The beds are of poorly consolidated gravels, sands, and silts with thin seams of lignite, some of which are burnt. The sediments erode easily, particularly where the western slope is steepened by undercutting of East Channel. In general, lakes are few and cover less than 15 per cent of the terrain, a strong contrast to regions II and III. Pingos are absent and ground ice slumps are few.

Section IVa is transitional between regions III and IV. The boundary runs along an indistinct discontinuous bluff whose base is close to 200 feet in altitude. Section IVb is also transitional. Altitudes are lower than in the hills to the south but higher than in section IIIa to the north. Section IVc is hilly, rising up to 850 feet and thus several hundred feet above adjacent areas.

Section IVd comprises the major part of the Caribou Hills. From Inuvik to 68° 35′ N, there is a gentle slope, about a mile wide between East Channel and the 150-foot contour, above which the land rises as a bluff to altitudes of 300 to 400 feet. Kame deposits and several melt-water channels lie along the west slope which is covered with spruce and many deciduous trees. From 68° 35′ N to Reindeer Station, East Channel flows against the Caribou Hills, the western slope being gullied (Figure 74). The hills are higher than to the south, with altitudes of 500 to 800 feet occurring within less than a mile of East Channel. All the material appears to consist of unconsolidated or poorly consolidated sands, silts, and gravels, with a few seams of lignite. Many burnt and iron-stained beds are present on hill slopes south of Reindeer Station. The western slopes of the Caribou Hills are locally eroded into a ribbed pattern by evenly spaced gullies. Many streams are aggrading in their lower courses. Spruce and paper birch go part-way up the slopes. North of Reindeer Station the western part of the Caribou Hills decreases in altitude, the western slope is cut into facets, and the hills become barer. The inland areas are rolling, crossed by melt-water channels, and treeless.

Section IVe, like IVc, is a slightly higher hilly area which rises to 650 feet, being several hundred feet above adjacent areas. The eastern boundary is along a steep bluff whose base is at 250 feet altitude.

Campbell Lake Hills: Region V

139. The Campbell Lake hills are mainly an upland area where bedrock either outcrops in escarpments, or else is close to the surface. The principal exception is the flat area southwest of Campbell Lake where altitudes range from 50 to 150 feet. There is much fluting in both bedrock and ground moraine. Kames and kame terraces are numerous on hill slopes. Abrupt swings in fluting show rapid directional changes of glacial flow. The vegetation cover is open woodland with large white spruce in favored areas, spindly black spruce in some wet portions, a few tamarack, scattered paper birch, and poplar.

140. Section Va is the rocky and hilly area around Dolomite (Trout) Lake near the Inuvik airport. The oldest known rocks in the Mackenzie Delta area, possibly Proterozoic but more probably Cambrian to Silurian, are exposed near Inuvik in an arch (the Aklavik Arch) with flanking younger Palaeozoic and Cretaceous (?) rocks on both sides (Brandon, 1961, p. 22; Jeletzky, 1961, p. 539; Sproule, 1961, p. 1146). The rocks include colorful grey to purple shales and greenish to reddish dolomites. Devonian limestones are exposed south of Inuvik (e.g. near Campbell Lake) and in outliers within the delta (IIc) where they form rocky islands nearly submerged in a sea of alluvium. Outcrops are also present as minor navigational hazards in several channels.

Bedrock hills rise 300 to 400 feet south of Inuvik as far as Campbell Lake. The rocks are cut by subparallel northeast—southwest faults which are reflected in a series of elongated lakes and scarps 100 to 200 feet high. Alluvium-filled re-entrants extend into the hills, for example, those of Dolomite and Campbell (Gull or Long) lakes. The hills are glacially fluted, the terrain near Inuvik being covered with a thin veneer of drift, but there is much bare rock to the south. The Inuvik airport is located on bedrock on the northeast side of Dolomite Lake with crushed rock being used as the foundation. Small deposits of lead and zinc have been reported in the hills.

141. Section Vb is a low, flat, fluted area, 50 to 75 feet high and underlain by bedrock. Section Vc on the west side of Sitidgi Lake has bedrock escarpments and rocky hills up to a maximum altitude of about 450 feet. The escarpments trend nearly parallel to the lake shore. The base of the major escarpment lies close to the 150-foot contour line. The escarpment and the hills show evidence of glacial action in crag-and-tail topography and fluting. Section Vd is rocky where it fronts onto Campbell Lake, with excellent fluting. Eastwards, the rolling terrain is 500 to 600 feet high.

142. The Campbell-Sitidgi Lake lowland (section Ve) is part of the old course of a Pleistocene (?) river which flowed through the Campbell-Sitidgi Lake depression (Figure 10). The entire area is low and flat, with the divide between Campbell and Sitidgi lakes estimated at about 30 feet above sea level. The soils are predominantly silty, the drainage poor, and shallow lakes abound. The land on either side of the lowland rises gradually to heights of 400 to 800 feet. Glacial deposits, including kames and recessional moraines, occur both on the lowland flats and slopes. As the deposits show no evidence of recent erosion, there was probably no appreciable postglacial flow through the lowland.

Campbell Lake, which has also been known as Gull Lake and Long Lake, occupies part of a glacially modified interglacial (?) channel whose location is probably controlled by faulting. Rocky cliffs, partly in Devonian limestone, parallel both sides. The southern end of the lake is shallow, where inflow from East Channel has almost sealed off the southern third of the lake by means of a delta.

The Campbell-Sitidgi Lake lowland has long been used as a portage route to the Eskimo Lakes. It is doubtless the route mentioned to Alexander Mackenzie by natives. Although the portage trail is nearly overgrown by vegetation and has suffered from disuse, there are stretches with cross logs still preserved, laid like a corduroy road, over which heavy whaleboats and canoes used to be dragged. The portage route leads from the creek flowing into the north end of Campbell Lake through six lakes into the creek flowing into the south end of Sitidgi Lake. The route was frequently used to reach Anderson River, in preference to the more exposed coastal route. Sitidgi Lake is a large shallow lake; the maximum depth reported so far is 50 feet. Section Vf is a low flat area 50 to 100 feet in altitude.

Unfluted Plains: Region VI

143. The unfluted plains include the drainage basin of the "Miner River" (local name) and much of that of Kugaluk River. Although there are a few hills in the area, the terrain is mainly flattish, sloping from an altitude of 1000 feet in the south to about 100 feet in the north (Figure 38). Major glacial features are conspicuously absent. Only several small areas are fluted, eskers and kames are few, and other glacially formed topographic features are meager. The "Miner River" meanders in a shallow valley. The valley of Kugaluk River south of 68° 45′ N is deeply incised in a postglacial gorge 200 to 300 feet deep.

In the area east of Sitidgi Lake, shales, siltstones, and poorly indurated sandstones of Mesozoic age may overlie Devonian sediments and also extend north towards the Eskimo Lakes. They occur, for example, near the mouth of Kugaluk River and along the lower Anderson River (region IX) downstream from the forks (Martin, 1961, p. 450, Fig. 10; McGill and Loranger, 1961, p. 517). The Mesozoic sediments are rapidly eroded, but their full distribution in the area is unknown.

Fluted Plains: Region VII

144. The fluted plains form a broad crescent, 20 to 30 miles wide, to the west of Anderson River. The plains are strongly fluted, with individual linears being traceable for many miles (Figure 3). The southern third, south of Wolverine River, lies between 500 and 700 feet altitude. The southwestern side terminates against higher rocky terrain which reaches a maximum altitude of about 1,100 feet. In this higher border area, where nearly horizontal rocky escarpments occur, ridges and crag-and-tail topography are present. Wolverine River flows in a large Pleistocene valley with a rocky scarp, well over 100 feet high, along the left or outer bank where it makes its big bend into the fluted plains (Figure 63). There is no continuous scarp on the right or inner part of the bend. The depression occupied by Wolverine River has considerable fill; kame deltas have been built into it.

The northern two thirds of the fluted plains slopes from an altitude of 500 to 600 feet near Wolverine River to less than 200 feet in the north. The area is flat, broken only by a few rolling hills, rocky escarpments, and cuestas with crag-and-tail topography. Bedrock is close to the surface. On the west side of the plains, between 68° 30′ N and 68° 45′ N, there are three south-north trending systems of linear lakes and partially infilled abandoned channels. The right-hand tributary of Kugaluk River, entering from the east at 68° 55′ N, flows in a low, poorly drained, Pleistocene (?) valley. North of about 69° 10′ N, the fluting becomes subdued, lakes are more numerous, and the boundary is indefinite.

Hyndman Lake Hills: Region VIII

145. The Hyndman Lake hills are a distinctive region with higher terrain than adjacent areas, rocky escarpments, large Pleistocene valleys, and numerous lakes. Many hills in the southern part have summit altitudes of 1000 to 1500 feet. East of the Mackenzie Delta and extending from the Point Separation—Campbell Lake area to the Wolverine—upper Anderson River area, Devonian rocks are known to occur either at the surface or beneath younger sediments (Kindle, 1916, p. 246; Lenz, 1961; Martin, 1961; Warren, 1944). As Devonian rocks are relatively resistant to denudation processes, they tend to form hills, scarps, and narrow-walled valleys. The areal extent of such rocks is unknown.

A large Pleistocene valley connects the headwaters of the Kugaluk and Wolverine rivers (Figure 63). The valley, including its tributaries, is partially infilled with undifferentiated drift, kames, dead ice deposits, and eskers which have produced extensive lake-strewn valley flats and ponding to form large lakes, Hyndman Lake being the best example. At location 1 (Figure 63) the Pleistocene valley has been blocked with fill which reaches an altitude of about 650 feet. The postglacial Kugaluk River has not flowed through the saddle at location 1, but has taken a course to the west, leaving flights of terraces from 600 down to 400 feet in altitude. At location 2, eroded outwash or deltaic deposits infill a broad Pleistocene valley up to an altitude of 450 feet. The flat-floored valley between locations 3 and 4 is nearly 2 miles wide and partly infilled with glaciofluvial deposits. There is no evidence of any large postglacial flow in the valley. Glaciofluvial sediments and other fill occur from location 4 to Hyndman Lake. Kame deposits are banked against the valley sides in this and other parts of the Pleistocene valley, suggesting that ice lingered here after the upland areas were bare. The valley is narrowest at location 5. An extensive terrace, at an altitude of 500 feet, is at location 6. The big bend of Wolverine River (location 7) and downstream is cut into Pleistocene (?) fill. Although details of the Pleistocene valley have been greatly modified by glaciation, the former direction of river flow seems to have been from west to east. Thus flow from locations 1 and 2 may have joined at location 3 then flowed via location 4 into the valley now occupied by Wolverine River. In general, the valley is 2 to 3 miles in width, except for a narrow part at location 5. Judging from the width of the valley and the large sweeping bends, the size of the river which occupied the valley would be nearer that of the present Anderson River than that of Kugaluk River.

Anderson River Uplands: Region IX

146. The Anderson River uplands are a complex area of diversified relief. The eastern part extends into the Anderson River map-area (Mackay, 1958a, pp. 84-88). The underlying rocks are mainly easily weathered and eroded Cretaceous (?) sediments.

The terrain on the right side of Anderson River (section IXa) north of about 69° 26′ N is composed of flat to rolling hills (Figure 67) which descend from 500 feet in the south to 200 feet in the north. There is a scattering of lakes, a few of which are drained. At 128° 10′ W, a large north-south partially

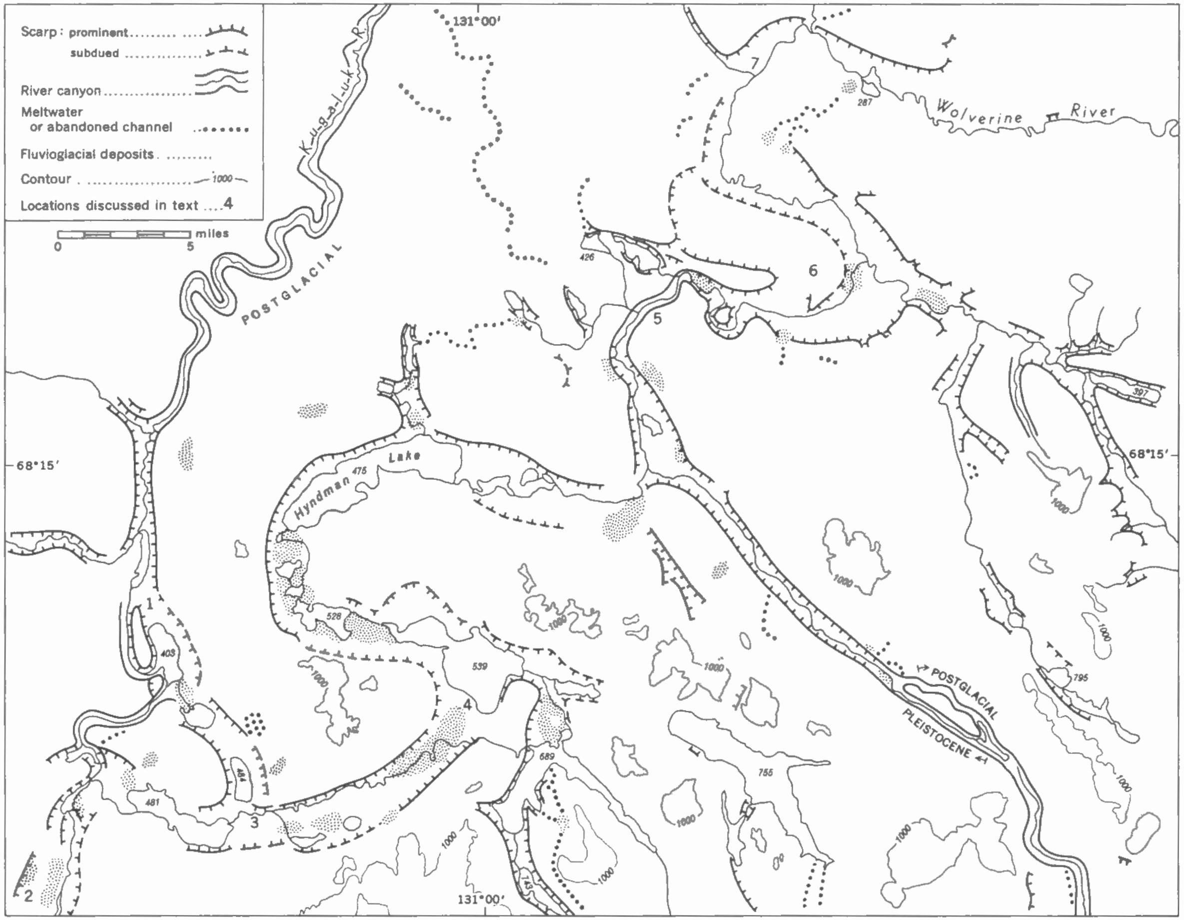

Figure 63. Details of the geomorphology of the Hyndman Lake area.

abandoned drainage channel can be traced for 12 miles. The southern end has been abandoned by stream capture. Section IXb, to the left side of the Anderson River, is flat to rolling. Drained lakes and flats are numerous. Section IXc has a flat upland surface (400 to 700 feet in altitude) which has been deeply incised by gorges of tributary rivers spaced roughly 5 miles apart. Even some streams which lack surface flow in late summer have cut steep-sided valleys several hundred feet deep. The rapidity of erosion has frequently caused stream capture. Lakes are few and small.

Figure 64

View looking east across the Anderson River at 69°16′N. When seen in August, 1955, lignite seams were burning over a linear distance of several hundred feet and burnt seams were exposed nearby. Note the logs (indicated with an arrow) marking flood waters nearly 25 feet above river level. (Geog. Br. JRM-55-7-10).

The flat, lake-strewn area of section IXd lies at an altitude of 600 to 800 feet. It is characterized by a large number of sub-parallel small streams which trend southwest-northeast. Most of the streams are 3 to 5 miles long, originate from a lake, and flow through one or more lakes before entering a main stream. The majority of the streams flow from southwest to northeast. The trend may result from structural control, glacial fluting, or both. Major tributary streams have deeply incised gorges. A few streams and elongated lakes lie in Pleistocene valleys which are recognizable as such by infill of glaciofluvial deposits. Some of the Pleistocene valleys are nearly a mile wide and 15 miles long. As the valley bottoms are from 400 to 600 feet above sea level, even within several miles of Anderson River, and have gentler gradients than present Anderson tributaries, the valleys were probably eroded when the ancestral Anderson River flowed at a higher level. Both glacial melt-water channels and those abandoned by stream capture are also present.

Section IXe is flat. The regional slope of the terrain is to the southwest. Altitudes drop gradually from 800 to 700 feet in the first 15 miles, and then within 3 miles, from 700 to 450 feet at Crossley Lakes. Section IXf is hilly with altitudes which exceed 1,000 feet. Some streams draining to the Anderson River have cut deep gorges in older Pleistocene valleys. Stream capture has frequently taken place. Lakes are relatively sparse. Most of Section IXg is rolling, with altitudes of 700 to 900 feet. Lakes abound. The principal streams draining to the Anderson River are underfit, as they either occupy portions of Pleistocene valleys, glacial melt-water channels, or have lost volume from stream capture.

Section IXh south of the upper Anderson River is a land of numerous large and small irregularly-shaped lakes. The northern half is flat and 350 to 400 feet in altitude; the southern half is hilly and 400 to 600 feet in altitude. Section IXi is hilly, with altitudes of 500 to 1,000 feet. Between 128° 00′ W and 128° 10′ W, there are two large meridional-trending melt-water channels which occupy enlarged Pleistocene (?) valleys. The largest channel is 150 to 200 feet deep and half a mile wide. The channels lead into the Andrew River which has postglacial terraces up to an altitude of 450 feet. The western third of the area is glacially fluted.

147. Anderson Valley: Section IXj: All the major tributaries of the Anderson River, with the exception of Wolverine River, are bordered by flights of terraces whose flow marks and radii of curvature indicate that the rivers were formerly larger. A few of the upper terraces are of Pleistocene age, but most of the lower terraces are postglacial.

The Iroquois River, from 68° 00′ N to its junction with the Carnwath River, has a deep flat-floored valley one mile wide. Flow marks are found at altitudes of 250 to 350 feet, that is, 50 to 100 feet above present river level.

The Carnwath River upstream from the entrance of Iroquois River flows in a broad one-mile-wide valley, with flow marks over 300 feet in altitude, or 50 to 100 feet above the river. The combined flow of the postglacial Iroquois and Carnwath rivers has cut a valley from 1 to 2 miles wide. The radii of some of the slipoff slope flow marks are nearly a mile, this being at least twice the radii of the largest meanders of the present river. High-level terraces descend from an altitude of over 300 feet at the junction of the Iroquois and Carnwath rivers to about 250 feet at the Andrew River junction; i.e., the highest terraces are 150 feet above river level.

The Carnwath River north of 68° 14′ N, Andrew River north of 68° 06′ N, and Wolverine River east of 130° 00′ W, all flow in channels cut below a flat glaciated plain, whose altitude ranges from 250 to 350 feet. The higher land of Region VII rises above the plain along a sharp boundary at an altitude of 300 to 400 feet. The plain is dotted with numerous lakes and marshy areas, with small hills several tens of feet in height. The plain is believed to be a strath terrace of Pleistocene rivers. The surface material appears to be alluvium and drift over bedrock which outcrops at an altitude of about 250 feet. The upper limit of postglacial flow is between 300 and 350 feet, with postglacial erosion exceeding 100 feet along extensive channel sections.

Andrew River from 68° 00′ N to 68° 06′ N has cut a 50- to 100-foot-deep valley into a flat fluted plain. North of 68° 06′ N Andrew River flows at the bottom of a 1 to 2 mile wide valley cut below the broad flat Pleistocene plain discussed above. The south and east sides of the plain are partially covered with glaciofluvial deposits including pitted outwash and kames. The upper limit of the Pleistocene plain, on the east, is at 400 feet. On the left side of Andrew River, at 68° 22′ N, there is a large cut-off river bend with flow marks on the slipoff slope rising to slightly over 300 feet. The size of the bend is about double that of the bends produced by the combined flow of the Carnwath and Andrew rivers, so that glacial melt-waters were probably involved.

From 128° 00′ W to the forks of the Anderson River, terraces rise from the Anderson River level to 300 feet, the river level at the forks being at 100 feet. The highest terraces, at 250 to 300 feet, may be of Pleistocene age, as some appear partially filled with kamey deposits. Postglacial flow marks occur at altitudes below 275 feet. As previously mentioned (*44*) kettle lakes are found in some terraces in locations such that glacier ice probably persisted until river flow was at 150 feet above present sea level. This suggests that the terrace deposits infill a Pleistocene valley.

Five miles east of the forks a 10-mile-long abandoned Pleistocene channel trends from south to north, thus bypassing the present forks. The floor of the channel, which is at an altitude of 300 feet, averages slightly less than a mile in width. Dead ice terrain and kames lie in the bottom of the channel and along the sides. There is no evidence for any postglacial flow through the channel. A southward projection of the abandoned channel across the Anderson Valley would meet the Pleistocene (?) strath terrace at an accordant level, suggesting a genetic relationship.

North of the forks, the valley ranges from 1 to 4 miles in width. Terraces are found at all levels, decreasing from an upper altitude of 250 feet at the forks to 75 feet at the Anderson River delta. Steep cut bluffs, 200 to 400 feet high, form much of the valley sides. Several small high-level meander scars occur at altitudes of 400 to 600 feet, far above the lower terraces. The scars could have been cut by only a small river, such as the size of Andrew River, and may represent marginal flow when ice occupied the Anderson Valley.

Excellent incised meanders occur between 69° 05′ N and 69° 27′ N. The slipoff slopes have continuous terraces which show at least 150 feet of postglacial cutting. Eskimo portage routes used to cross at the meander necks.

Along the estuarine portion of the Anderson delta, commencing at 69° 32′ N, extensive terraces lie at 75 feet and below. They indicate that sea level was higher when they were formed, as the land opens out to the sea with no intervening higher ground. From the apex of the delta to the sea, a distance of 20 miles, the river flows on the right (east) side of a gradually widening floodplain. Terraces

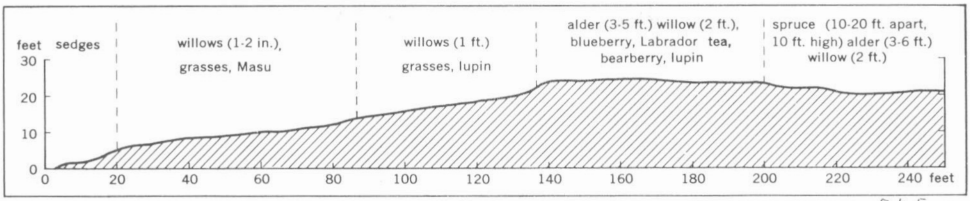

Figure 65. Profile of left bank of Anderson River floodplain at 69°12′N, 128°20′W. Altitudes are above river level on August 6, 1955. Yearly flood waters are estimated to rise to the 15- or 20-foot levels. The spruce grow on the backslope above annual flooding.

border both banks. The delta is composed of low alluvial islands, channels, and lakes. At the mouth, the alluvial flats are 4 miles wide. The Anderson delta is building northwards into the shallow waters of Wood Bay. Tidal flats, exposed at low water for a distance of 1 to 2 miles from the mouth, are covered to a depth of 5 feet or more at extreme high water. Anderson River water can usually be detected by salinity and color as far north as the northeast sandspit of Nicholson Peninsula where the water may, at times, be fresh enough to drink.

The banks and channel of Anderson River are of mud and sand from the mouth to about 69° 16′ N; upstream, gravelly banks and channel bottom predominate. At time of break-up, water levels rise 10 to 25 feet, thus floodplains stand well above low water levels (Figure 65). Boulder pavements, striated and hard-packed by the passage of river ice, are especially common along the middle reaches of the Anderson River.

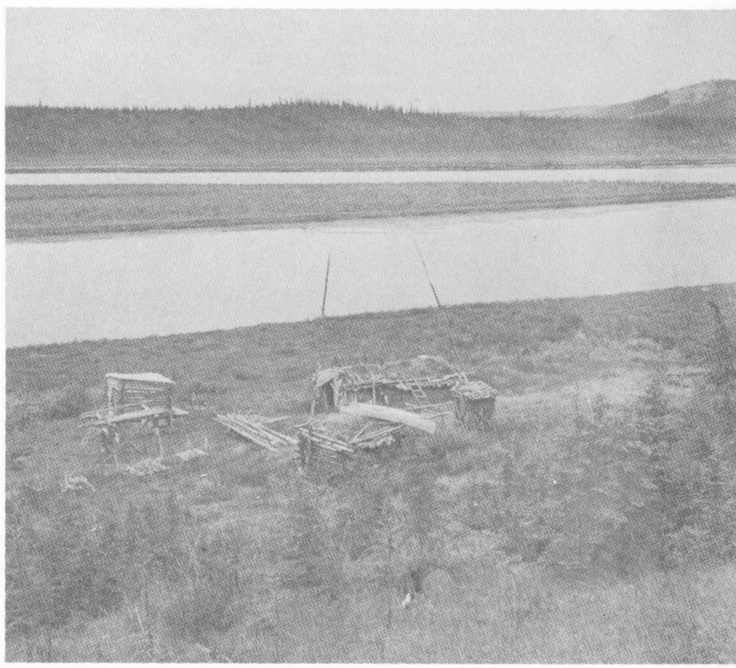

Figure 66

Trapper's cabin just upstream from the forks of the Anderson River showing an island of the Anderson River in the background. The Carnwath River flows at the foot of the hill at the right background. The log cabin (center) measured 11 by 14 feet and had an attached shed. A 7 × 7 storage shed raised 4.5 feet off the ground is to the left of the cabin. Three dog kennels are close to the storage shed. August 9, 1955. (Geog. Br. JRM-55-9-4).

Only a general account can be given of the bedrock along the Anderson Valley. At the forks (Figure 66) a grey to brown fissile Cretaceous shale is exposed in the hills and along steep, undercut banks. The shales weather rapidly into light grey muds. From about 10 miles downstream of the forks nearly to 69° 10′ N, a dark grey shale with thin beds of iron-cemented sandstone is exposed in many sections. Slumping and landsliding are prevalent in weathered shale where slopes are steep. From 69° 10′ N to about 69° 31′ N, there is a formation in which spontaneous combustion has taken place. Brick-red and yellow-ochre burnt areas extend from near river level to over 300 feet above the river, and grade laterally into unburnt sections of laminated shale with much carbonaceous material which is easily ignited with a match. Active burning in 1955 was observed in a ravine on the east side of the river at 69° 25′ N (Figure 64).

In a canoe trip made in 1955 from the mouth of Anderson River to the forks and return, no tabular sheets of ground ice were seen, in contrast to the abundance of such ground ice along the Liverpool Bay coast to the north.

Mason River Upland: Region X

148. The Mason River upland has a general altitude estimated at 100 to 200 feet above sea level. The coast terminates in wave-cut bluffs with narrow gravel beaches. Terraces up to an estimated 50 to 75 feet border the coast to the north and south of the mouth of Mason River. The area north of Mason River (section Xa) is mainly flat, with a few drained lakes. The area south of Mason River (section Xb) has numerous drained lakes and poorly drained flats. Mason River and its tributaries meander in broad flats which show evidence of much drainage. The meanders do not form a succession of smoothly curving bends, but wander aimlessly back and forth, impinging on one another with considerable irregularities in meander shapes and sizes. The lack of symmetry may be attributed to a highly variable discharge. The mouth of Mason River is estuarine, with sedimentation and delta formation occurring. Several uneroded remnants of higher land stand, as islands, in the delta.

Figure 67

Terrain south of Stanton showing the rolling topography covered with willows 6 to 18 inches high, avens, sedges, and grasses. August 3, 1955. (Geog. Br. JRM-55-6-6).

CHAPTER IV

CLIMATE

Meteorology

149. The coastal portion of the Mackenzie Delta area lies in the arctic, the southern portion in the subarctic. Irrespective of the specific criteria used, the outer islands of the modern Mackenzie Delta north of Reindeer Channel, Richards Island, Tuktoyaktuk Peninsula, and a 10- to 20-mile-wide coastal belt on the south side of the Eskimo Lakes are in the arctic. Aklavik and Inuvik are in the subarctic. On the basis of a tundra-forest delimitation, both are over 30 miles south of the tree-line and so are in the subarctic. The location of Reindeer Station is transitional. If a climatic definition is used, such as the mean temperature for the warmest month of less than 50°F, then the July mean temperature of 56.4°F (1926-1950 period) excludes Aklavik from the arctic. The record for Inuvik is too short to provide reliable means, but the July normal is doubtless over 50°F and very close to that of Aklavik. The July and August mean temperatures for Tuktoyaktuk (1948-1960 period) are close to 50°F, although on a vegetation basis, Tuktoyaktuk is clearly in the tundra.

The meteorological record for the area varies considerably. Fairly continuous records are available for Aklavik from 1926 and for Tuktoyaktuk since 1948. With the establishment of the Distant Early Warning (DEW) line along the coast, information of varying duration is now available for Shingle Point to the west, and for Tununuk, Atkinson Point, and Nicholson Peninsula. In the ensuing discussion of climatic elements, the non-comparability of data should be continually borne in mind.

150. Temperatures: The coldest month of the year is usually February for arctic stations in Canada, because coastal areas in early winter tend to be kept warmer than interior mainland areas by radiation from water beneath the ice (Thomas, 1960, p. 4). At Tuktoyaktuk, February normally averages several degrees colder than January, with means of —20°F for January and —24°F for February. The other coastal stations (Shingle Point, Atkinson Point, and Nicholson Peninsula) have the coldest month in January, as do the inland stations at Tununuk and Aklavik. In addition, winter temperatures in the Aklavik-Inuvik area, the most densely settled region, tend to be slightly higher than the Fort Good Hope area to the south. This also applies to extreme temperatures which range from a low of —62°F at Aklavik to —79°F at Fort Good Hope. The monthly rise of temperature from winter to summer is rapid, because the mean annual range is about 75°F. The highest temperature on record for Aklavik is 93°F; for the several years of record, Shingle Point and Tununuk have experienced temperatures in the low 80's; Atkinson Point and Nicholson Peninsula have reached 79°F. Therefore, extreme maximum temperatures throughout the area probably exceed 80°F. These temperatures are higher than those along the arctic coast to the east. Freezing temperatures may occur along the coast during any summer month, but are uncommon inland in July and early August.

151. Precipitation: The annual total precipitation is low, varying from 11 inches at Fort McPherson, through about 10 inches at Aklavik, to less than 10 inches along the coast. Most of the precipitation in the summer months comes in the form of rain, the maximum usually occurring in July or August. Although the rainfall is light, all stations have recorded downpours of over .5 inches of rain in 24 hours, and some over 1.0 inches. Thunderstorms, though infrequent, occur. Along the coast, snow, sleet, and driving rain may be expected anytime in the summer. Snowfall is light. The measurement of snowfall is difficult and the results are notoriously unreliable, but since 1947 careful observations have been carried out on snow cover characteristics at Aklavik (Williams, 1957). In the period from 1947 to 1954, the maximum snow depth at Aklavik was 16 inches, it being understood, of course, that the depth does not refer to concentrations in snow drifts. For the three consecutive years of 1949 to 1952, depths at Aklavik where below 5 inches. By extrapolation from Aklavik-Fort McPherson-Fort Good Hope data, average snow depths along the coast may be only a few inches in winter. However, winds redistribute the snow to create bare areas and snowdrifts, thus giving the impression of a heavy snowfall to some travellers.

152. Fog: Fogs are common in the summer along the coast. If the winds are onshore, thick fog banks may roll many miles inland as far south as the Eskimo Lakes. Prolonged winds may blow fog inland to Aklavik-Inuvik. Fogs may recur day after day but often dissipate towards noon. Radiation fogs may form inland, and, where air drainage is strong, the fogs may flow downslope like a misty river.

153. Wind: The prevailing wind at Aklavik throughout the year is from the northwest. At Tuktoyaktuk, for a two-year period of record, the prevailing winds were from the northwest and east-southeast, with mean wind speeds of about 10 mph throughout the year. At Nicholson Peninsula the prevailing winds, over a two-year period, were from the west, with mean wind speeds of about 15 to 20 mph throughout the year. Strong winds of gale force may blow and be a hazard to small coastal boats. Winds exceding 70 mph have been measured along the coast.

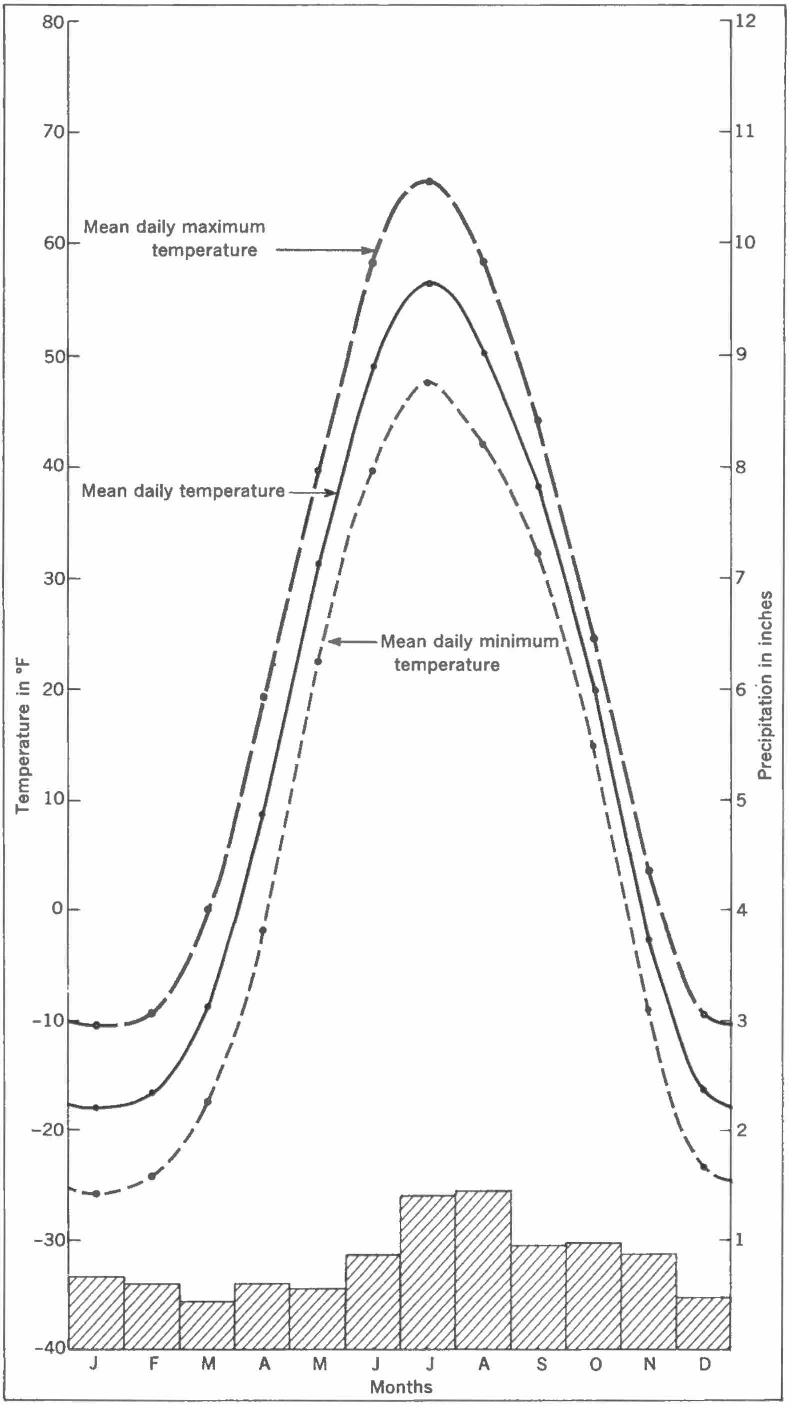

Figure 68. Climatic graph for Aklavik. Scale for precipitation on right.

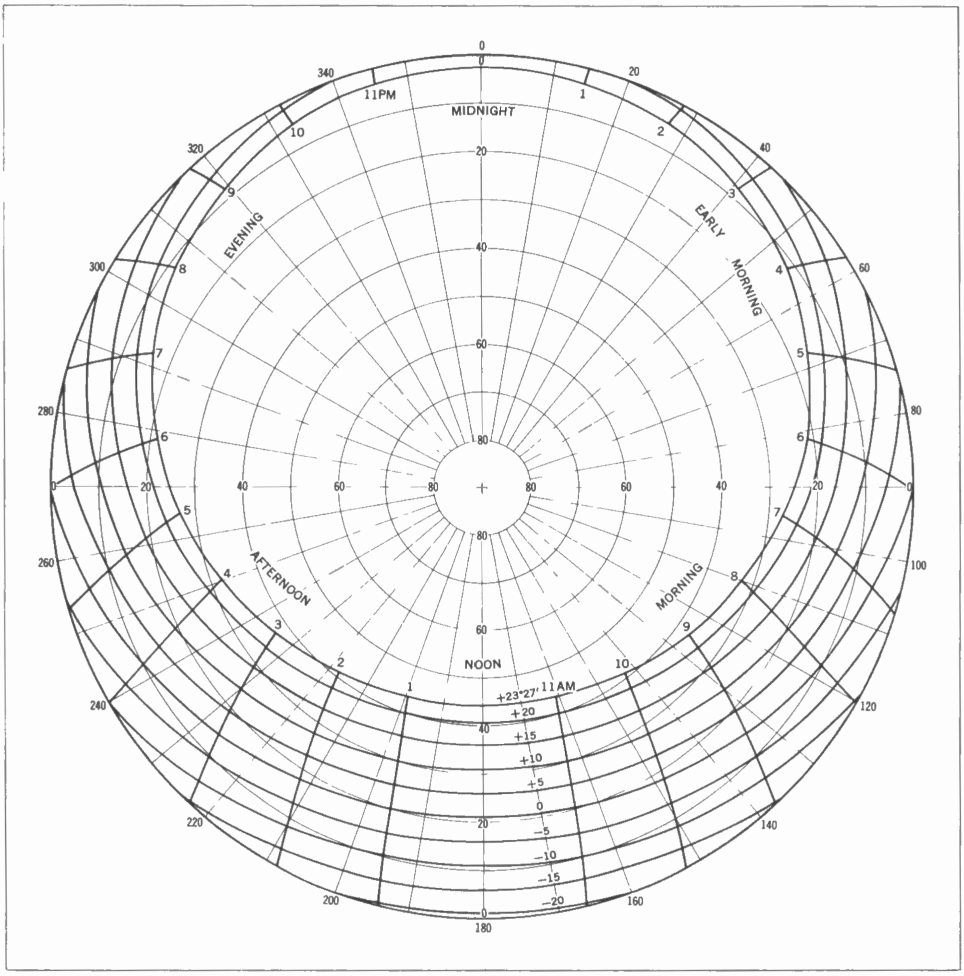

Figure 69. Graph of solar altitude and azimuth for latitude 69°N.

154. In Figure 69 the solar altitude and azimuth have been computed for 69° 00′ N, the graph being similar in interpretation to those in the Smithsonian Meteorological Tables (1958, pp. 497-506). The graph may be used for all settlements in the delta area, within the accuracy with which it can be read. The approximate calendar dates with given declinations are:

Declination	*Approximate Date*
+23° 27′	June 22
+20°	May 21, July 24
+15°	May 1, August 12
+10°	April 16, August 28
+5°	April 3, September 10
0°	March 21, September 23
−5°	March 8, October 6

Declination	*Approximate Date*
−10°	February 23, October 20
−15°	February 9, November 3
−20°	January 21, November 22
−23° 27′	December 22

To interpret Figure 69, the declination for the date on which information is sought is obtained. The intersection of the curved heavy line corresponding to the given declination with the concentric circles (light lines) gives the altitude of the sun above the horizon for specified times of day. The direction of the sun, for any time, is obtained by following the time-line to the periphery. For example, when the declination is +23° 27′, the sun is about 44° above the horizon at noon; in the early morning at about 3:40 a.m. it is 10° above the horizon in the northeast. When the declination is —15°, the sun is about 6° above the horizon at noon; it rises at 9 a.m. in the southeast and sets at 3 p.m. in the southwest. There is a brief period when the sun remains below the horizon in winter.

Break-Up And Freeze-Up

155. All the water bodies are frozen over in winter, so that break-up and freeze-up are important phenomena in the lives of the inhabitants and the economy of the area. The data used in the following study have come mainly from unpublished sources with the assistance of government officers, missionaries, the Royal Canadian Mounted Police, traders, and private individuals. The data are of varying reliability; when two independent sets of data have been obtained for one location, the variation has at times been sizeable. Nevertheless, the dates recorded appear to be of sufficient statistical reliability to serve the needs of quantitative analyses (Mackay, 1961a; 1961b; 1963 in press). Mean dates for break-up and freeze-up are as follows: (Mackay, 1961b, p. 1129).

MEAN FREEZE-UP AND BREAK-UP DATES, 1946-55

Location	*Distance from Great Slave Lake*	*Freeze-up*	*Break-up (un-adjusted)*	*Break-up (adjusted)*
Fort Providence	50	Nov. 24.9	May 18.6	May 18.6
Fort Simpson—				
Mackenzie above Fort Simpson	208	Nov. 27.5	May 15.4	May 15.4
Mackenzie below Fort Simpson	218		May 11.5	May 11.5
Liard River			May 6.1	May 6.1
Fort Norman	513	Nov. 15.3	May 14.2	May 14.2
Norman Wells	565	Nov. 10.9	May 15.1	May 15.1
Fort Good Hope				
Ramparts	680	Nov. 5.6*	May 22.9	May 22.9
Settlement	684	Nov. 12.8	May 15.0	May 15.0
Arctic Red River Settlement—				
Arctic Red River	898	Oct. 8.7	May 25.1	May 25.1
Mackenzie River	898	Nov. 1.5	May 26.8	May 24
Lang Trading Post	970	Oct. 9.3**	May 29.1	May 26
Aklavik	999	Oct. 9.4**	June 0.1	May 28
Reindeer Station	1,004	Oct. 18.1	June 2.5	May 27

*Seven years of record.
**Nine years of record.

156. Break-up: The definition of break-up varies according to the water body and interests of the observer. In the case of rivers, break-up is usually related to the first movement of ice in the spring, or else to the final passage of river ice, except for a few drifting cakes. Definitions for lakes, harbors, and the sea coast are less precise, but often refer to the general clearing of ice from the standing body of water. In the Mackenzie River system, break-up from Fort Providence to Fort Good Hope has usually been defined as the first ice movement, even though it may later jam and movement temporarily ceases. At Arctic Red River, break-up is taken when the main river ice moves past, even though it may jam downstream. At Lang Trading Post (15 miles south of Aklavik, at the junction of Phillips Channel with the middle Peel Channel) break-up is dated at the time when Peel driftwood passes the Lang Trading Post (personal communication, K. H. Lang). In some years, the river may be clear in front of the Trading Post for a day or more but be jammed upstream. When the driftwood passes, the Peel Channel is clear of Peel River ice. In many years, Mackenzie River ice runs for days after the driftwood passes. Navigation is not fully open on Peel Channel until most of the Mackenzie ice passes the mouth of the Peel Channel near Point Separation. At Aklavik, break-up dates are usually given for the time when the river is clear of ice. At Reindeer Station, break-up is when the ice clears out of the river in front of the settlement; that is the main ice, not the scattered floes which drift past for a day or so after the main ice has gone. As the preceding discussion shows, break-up has different meanings along the Mackenzie River. Therefore, when break-up dates are compared for prognostic or analytic purposes, the differences in definition should be borne in mind. In the table, dates in the column on the right have been adjusted to give the first movement of ice.

The process of break-up in a river depends on many factors, some of which are of more importance in one year than another (Brown, 1957). Ice thickness influences the rapidity of break-up, with its thickness dependent upon the severity of the winter, thaws, time and duration of the snow cover, and so forth. In the 1953-1960 period, river ice thickness at Aklavik ranged from 5 to 7 feet (Meteorological Branch, Department of Transport, 1959a; 1961), but in areas with drifted snow and mild winters, the thickness may be only 3 to 4 feet, or less. In the spring, side channels may open along the shore. Melting snows and precipitation cause a rise in water level. The larger channels in the delta maintain a year-round flow, but many of the smaller channels freeze to the bottom. Thus the distribution pattern for the influx of melt-waters may be complicated. In time, river ice is lifted, cracked, spread, and carried downstream. Ice-jams may temporarily form, especially at sharp bends or at channel junctions.

In the Mackenzie Delta, break-up is complicated because the Mackenzie and Peel rivers, together with innumerable interconnecting channel-lake systems, are involved. Mackenzie River channels receive flow influenced by climatic and hydrologic conditions a thousand miles upstream, whereas Peel River channels receive water from the nearby mountain tributaries of the Peel system. Other smaller rivers also play their part, notably Rat River on the west. Rat River, fed by melting snows from the Richardson Mountains, contributes a large volume of water to Husky Channel. The time of break-up for the Mackenzie Channels is generally later than for the mountain river channels of Peel and Rat River. Thus, the southwestern portion of the delta breaks up earlier, on the average, than the remainder.

Prognosis of break-up depends on many factors. Methods which have been employed successfully elsewhere range from simple measures, such as mean monthly temperatures and cumulated degree days, to those involving long range forecasting on a sophisticated basis. Detailed analyses of break-up in the delta cannot be made until more data on snow cover, hydrologic conditions, and meteorological factors are available, so that the following discussion is more in the nature of preliminary observations.

In most north-flowing rivers, break-up tends to progress downstream, with the headwaters breaking up first, and the northerly portions the last. This is not, however, wholly true with the Mackenzie River and so it is in contradistinction with the progress of break-up in some other north-flowing rivers. The progress of break-up along the Mackenzie River has been studied by standard statistical procedures, and the 1961 spring break-up was observed and recorded through the co-operation of Pacific Western Airlines. The results show that break-up tends to occur nearly synchronously between Fort Simpson and Fort Good Hope, in some seasons the downstream stations breaking up first and in others the upstream stations. Afterwards, the delta area breaks up.

157. Lang Trading Post Break-up: Break-up is influenced mainly by Peel River floodwaters, which are released about 10 days before those of the Mackenzie River. There is a good correlation between break-up at Lang Trading Post and that at Fort McPherson, on the lower Peel River. If X_0 is the date of break-up (counting from April 30) and X_1 is that for Fort McPherson, based on 15 years of data in the period 1939-1958, then:

$$X_0 = 15.4 + .62\,X_1 \qquad (157\text{-}1)$$

The coefficient of correlation ($r = .72$) is significant at the one per cent level and the computed standard deviation is 2.1 days. However, the coefficient of correlation between break-up at Lang Trading Post and the Mackenzie at Arctic Red River is low and not significant, even though Arctic Red River is at the same latitude as Fort McPherson. A multiple regression analysis using both Fort McPherson (X_1) and the Mackenzie at Arctic Red River (X_2) break-up gives

$$X_0 = 10 + .31\,X_1 + .42\,X_2 \qquad (157\text{-}2)$$

The improvement in adding the effect of the Mackenzie River is slight, because the coefficient of correlation only increases from $r = .72$ to $r = .76$ and the computed standard deviation is reduced only to 1.9 days. Thus, break-up at Lang Trading Post seems most related to Peel River, not Mackenzie River, conditions.

158. Aklavik Break-up: At break-up the initial flow is primarily from the Rat River-Husky Channel and the Peel River. Once the ice has moved at Aklavik, Mackenzie River ice, which is locally termed "black ice" because of a surface discoloration of drifted dirt from the exposed banks of the Mackenzie River, may move past for some time. Break-up at Aklavik is 2 to 3 days after that at Lang Trading Post, and nearly two weeks later than Fort McPherson.

159. Reindeer Station Break-up: Break-up at Reindeer Station is controlled by the Mackenzie River, not like Aklavik and Lang Trading Post by the Peel River. For the period 1937 to 1958, inclusive, the dates of break-up counting from April 30 at Reindeer Station (X_0) and with Fort Good Hope (X_1), Ramparts (X_2) at Fort Good Hope, and Fort Simpson (X_3) are:

Equation	r	S_c	
$X_o = 24 + .61\ X_1$	.72	2.7	(159-1)
$X_o = 24 + .38\ X_2$	.48	3.4	(159-2)
$X_o = 29 + .22\ X_3$	.30	3.7	(159-3)

where r is the coefficient of correlation and S_c is the computed standard deviation in days. At the one per cent level of significance ($r = .54$), only break-up at Fort Good Hope is significantly correlated with that at Reindeer Station. It is interesting to note the poor correlation of break-up at Reindeer Station with that at the Ramparts, just above Fort Good Hope, where break-up is delayed by ice-jams. Break-up at Fort Good Hope normally takes place two to three weeks before that at Reindeer Station.

160. Tuktoyaktuk Break-up: The harbor at Tuktoyaktuk usually breaks up in the latter third of June after much of Kugmallit Bay is ice-free.

161. Freeze-up: The definitions of freeze-up, like those of break-up, are determined by local usage. For rivers and channels, freeze-up is usually defined as the date when the river freezes over and the ice stops moving or jams, although open holes may persist for some time afterwards and in a few stretches may never freeze all winter. Some of the historical records include dates on the appearance of the first skim or drift ice, which may come from far upstream.

In the case of a standing body of water where current and tidal effects are negligible, the cooling of the water body is a function of local environment. For instance, the freezing of an enclosed delta lake is attributable to local factors. However, the water flowing in a nearby channel past the lake may be moving at a velocity of 2 to 3 miles per hour. Consequently, its temperature at any given moment is a function of conditions upstream. Because a great proportion of the water comes from the Mackenzie, the chilling of the water preceding freeze-up may have taken place over a period of several weeks far upstream. In the Mackenzie Delta, the water which comes from the Mackenzie River has had a long flow history; that which enters from the Peel and Rat rivers, a short history.

The Mackenzie River proper takes its source in Great Slave Lake, a large deep lake. According to Rawson (1950) the annual inflow—and hence outflow—from all sources into Great Slave Lake is less than one per cent of its volume. As Great Slave Lake is deep, the mean lake temperature changes slowly; it is, in essence, an integrator of past water temperatures. In the autumn, lake waters in the shallow western portion of Great Slave Lake, where the Mackenzie outlet takes off, get thoroughly mixed by strong winds and circulation due to density differences. It seems probable that by the end of September, complete circulation takes place (Rawson, 1947; 1950) so that waters leaving the lake are close to 39°F or lower. Liard River is the first major tributary of the Mackenzie River downstream from Fort Providence. As the Liard River water in the fall is colder than the Mackenzie River water at Fort Simpson, Mackenzie River water is chilled in the autumn (before freeze-up) by the addition of Liard water. This is also probably the case for the other mountain rivers which respond more rapidly to low autumn temperatures and other factors favoring cooling than the main Mackenzie River, as shown by their earlier dates of freeze-up. In 1960, water temperatures were taken along the Mackenzie and some of its tributaries. In September-October, Liard water was about 1°F colder than the Mackenzie; the Bear River in September about 5°F; Arctic Red River, about 5°F; and the Peel River at Fort McPherson was about 5° to 9°F colder than the Mackenzie in the same latitude. The cumulative discharge of the colder tributary rivers on the Mackenzie results, therefore, in an accelerated northward lowering of Mackenzie water temperatures.

162. In the open season, the water of the Mackenzie River is estimated to take a flow travel time of 15 to 20 days from the outlet of Great Slave Lake to Beaufort Sea. The estimate of the flow travel time has been based on the log of the patrol boat "RCMP Spalding" which travelled in June 1960, under the command of Supt. W. G. Fraser, from Fort Smith to Aklavik to Inuvik. The boat was run at constant engine speed. On the assumption that the actual speed of the boat was the sum of the still water boat speed and the current speed, an estimate was made of 15 days for the surface flow of the river between Great Slave Lake and the sea. As the velocity along the boat course is likely to exceed the mean river velocity, by a factor of 10 to possibly 30 per cent, a mean flow travel time of 15 to 20 days seems probable. Therefore, the Mackenzie River water which freezes in the delta in October is chilled to near its freezing point in its downstream journey of two to three weeks from Great Slave Lake. Although local factors determine the specific date of freeze-up, upstream factors lower the water temperature to nearly 32°F.

163. In view of the lack of detailed meteorological data for the Mackenzie River, an attempt can be made to relate freeze-up to air temperatures alone, although obviously this is not causally correct. Rodhe (1952) has shown that for a standing body of water, the water temperature (τ_n) on day n is a function of the water temperature (T_0) at a past date, and the intervening air temperatures (T_v) at time (t_v) at period $_v$. If the starting date for the computations is taken relatively far back, the function is approximated by

$$\tau_n \backsim \sum_{v=1}^{n} T_v \, e^{-k(t_n - t_v)}$$

At freeze-up, τ_n would be 32°F.

In order to test for k values, τ series, with k values of .00, .01, .10, .20, .30, and .40 were calculated for Reindeer Station (Mackay, 1963, in press). The value of each τ series was calculated for the actual date of freeze-up at Reindeer Station and the standard error for each k value computed. In order to make the τ series comparable, a constant was added so that the sum of the constant and mean τ value on freeze-up was zero. The standard error may be expressed in either temperature or in time, the two not being comparable. When the standard error is expressed in temperature units, the deviation from 32°F on freeze-up day may be obtained, for any year, from a single τ series terminating on the day of freeze-up. When the standard error is expressed in days, the deviation is the number of days between freeze-up and the date when the τ series first reaches 32°F. Using a flow travel time of 13 days from Fort Providence to Reindeer Station, results were:

REINDEER STATION

(Standard errors in °F for n = 13)

			k			
Period	0.00	0.01	0.10	0.20	0.30	0.40
1946-1955	2.90	2.90	2.98	3.62	4.17	4.86
1941-1955	3.85	3.84	3.51	3.40	3.72	3.92
1941-1955 .... (omitting 1944, 1947)	2.12	2.10	1.78	1.80	2.25	2.82

The smallest standard error for the period 1946-1955 was for k of 0.00 and 0.01; for 1941-1955, k was 0.20; and for 1941-1955, when the erratic years of 1944 and 1947 were omitted, the smallest errors were with k of 0.10 and 0.20, the difference between the two being negligible. The results suggest that the smallest standard errors are associated with small values of k, such as those of k = .20 or less. When the standard error is expressed in days, similar values of k are obtained. With more meteorological and hydrological data, much better results could be derived.

The method of correlating accumulated degree-days below freezing with freeze-up at Reindeer Station is unsatisfactory, if one climatic station is used. This is because the water which freezes at Reindeer Station has had a prolonged period of chilling uninfluenced by local weather. For example, using Aklavik climatic data, as Aklavik is the nearest weather station, coefficients of correlation did not exceed r = .27 for the two 10-year periods 1938 to 1947 and 1948 to 1957, both when frost sums and frost sums less those above freezing were used. As $r = .63$ at the 5 per cent level of significance, the correlations are low and not significant.

Freeze-ups at Aklavik and Lang Trading Post differ from that at Inuvik and Reindeer Station in the minor influence of Mackenzie River water. As the water freezing at Lang Trading Post and at Aklavik comes from streams which are relatively near to the delta, freeze-up is more closely related to local conditions than for Mackenzie-fed channels. Correlations with frost sums are higher than for Reindeer Station, but are not significant at the 5 per cent level. Likewise, there is no significant correlation between frost sums at Aklavik and freeze-up at Tuktoyaktuk for the available period of record.

CHAPTER V

VEGETATION

164. The northern portion of the Mackenzie Delta area is in the tundra, the southern portion in the boreal forest. The boundary zone between tundra and treed areas may be either sharp or transitional. In general, the sequence from north to south is: tundra; tundra with scrub willow and ground birch; scrub willow and ground birch; woodland and tundra with much scrub willow and ground birch; open woodland; and continuous woodland (Figure 70). The specific names of the plants, corresponding to the common names used here, are given in Polunin (1959).

165. Tundra Communities: The word tundra has been loosely used to describe many vegetation communities and thus lacks precise definition. Here it designates the vegetation, beyond the tree-line, which has wet tundra with many sedgy tussocks on poorly drained flats, a drier tundra including lichen and moss heath on the better-drained areas, and thickets of willow, alder, and ground birch on valleys and slopes.

Sedgy tussock flats are characteristic of ill-drained terrain and so occur around many lakes, on the bottoms of drained lakes, and along coastal lowlands. The sedges, usually called cotton-grasses, grow luxuriantly in wet areas, particularly on low-centered polygons and around muddy pools (Figure 24). The most continuous area is in the flats at McKinley Bay.

The higher land south of Tuktoyaktuk and Richards Island is, in general, rolling and well-drained, so that a drier type of tundra vegetation prevails. In soils where the silt-to-clay fraction exceeds 25 per cent, bare mud boils surrounded by barrens of mountain avens with associated sedges, grasses, herbs, and low willows tend to predominate. In sandy areas, a dry tundra vegetation is prevalent (Figure 61). Gravelly outwash and eskers, which are dry and well-drained, have sparse vegetation. Such areas look bald and bare from a distance. Besides grasses, sedges, lichens, and mosses, the gravel slopes in early summer are often colorful with the wild crocus, rose, and saxifrage (three-toothed or prickly saxifrage).

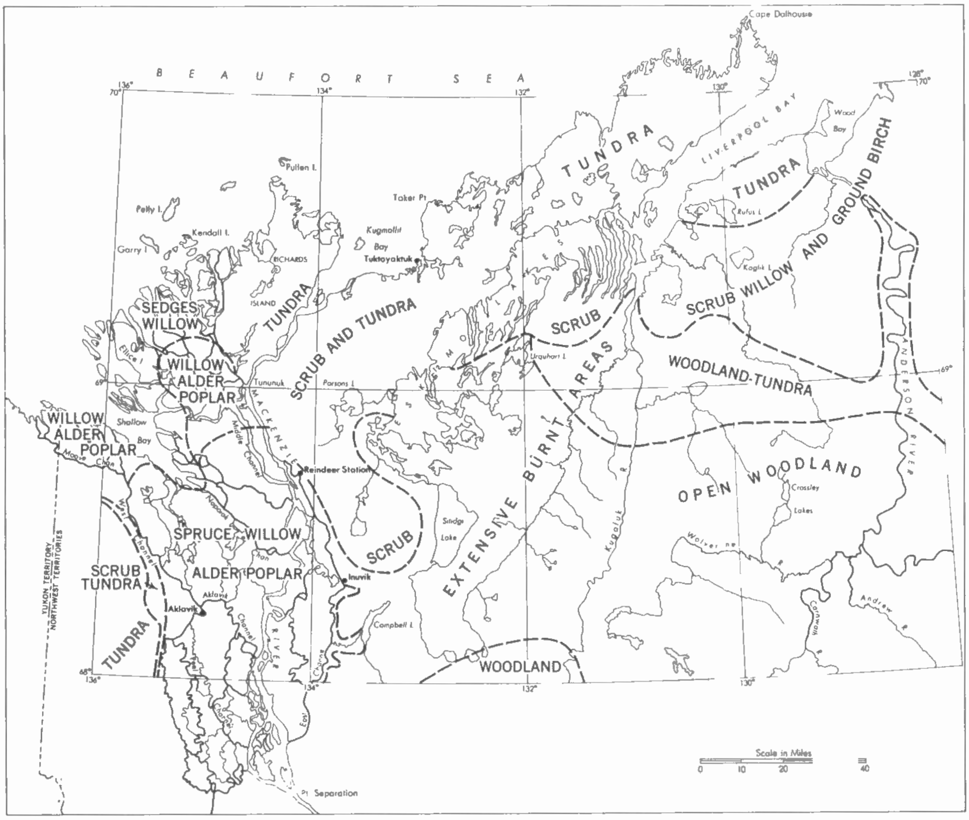

Figure 70. Vegetation regions.

Willows are found chiefly along stream banks, lake shores, seepage channels on hill slopes, old mud slump scars, and the sides of hummocks. Willows grow throughout the tundra area, but alders do not extend far beyond the tree-line. Alders tend to grow in moist soil where there is a winter protective snow cover. Thus, near Tuktoyaktuk, clumps of alder may grow where there is a seepage line down a slope or in a protected ravine. Ground birch grows in isolated clumps but may completely cover high-centred peaty polygons together with labrador tea, crowberry, bearberry, cloudberry, mosses, and lichens. Late snow-bank slopes may appear dark, from a distance, because of an abundance of the white arctic bell-heather.

The vegetation of the Richardson Mountains is predominantly of tundra, although trees grow in some sheltered valleys. There is considerable variation in the tundra vegetation, because some areas are rocky, steep, and nearly bare, whereas others are flat, poorly drained, and covered with wet tundra vegetation.

166. Scrub Willow and Ground Birch: Scrub vegetation, characterized by willow and ground birch, occupies a transitional zone between the tree-line and the tundra (Figure 67). Near the tree-line, the willows are in clumps 5 to 7 feet tall, interspersed with a lower growth of ground birch. Many heaths (cranberry, crowberry, labrador tea, rhododendron, bog-laurel, and heather), lupins, avens, sedges, grasses, lichens, and so forth thrive in association with the willow and ground birch. The ground surface is usually broken into hummocks several feet in diameter and a foot or so in height (Figure 17). As a rough approximation the surface cover in a randomly selected area might be: willows, 20 per cent; ground birch, 30 per cent; heaths and other plants, 50 per cent. Scrub willow and ground birch are found over much of the Caribou Hills interspersed with a few trees near the tree-line and grading into tundra on rolling hills (Figure 74). The assemblage is widespread on the south side of the Eskimo Lakes, in its peninsulas and islands, and in favored spots on the north side of the Eskimo Lakes. The vegetation type becomes sparse on the south side of Liverpool Bay, at Nicholson Peninsula, and north of Stanton.

The piedmont slope of the western uplands has a grassy-sedgy tundra vegetation with much scrub (willow, alder, birch), heaths, and straggling spruce and poplar extending up the slopes in the south. The distributary pattern of the alluvial fans is outlined by concentrations of alder and willow along the water courses.

167. Woodland-Tundra: The northern limit of the woodland-tundra is placed at the inferred limit of spruce, although isolated plants go far beyond the indicated limit. For example, a few prostrate spruce grow on Richards Island, 30 miles north of the "limit" of spruce and also near Kittigazuit. Spruce, though by far the most numerous, are not the only trees in the woodland-tundra zone, as scattered poplar, tamarack, and possibly paper birch are present. Spruce trees are mainly in the valleys and on slopes with tundra or scrub willow and ground birch on the hilltops. Ill-drained flats have much sedgy vegetation. Spruce extend northwards down the valley of Anderson River. In sheltered places, as at latitude 69° 00′ N, spruce reach heights of 40 feet; at 69° 15′ N, spruce are found on many bluffs and slopes but the higher hills are treeless and covered with scrub willow and ground birch (Figure 64); at 69° 25′ N, spruce are on slightly drier ridges but the hills are in willow-birch scrub. The northernmost spruce seen in the valley were at 69° 39′ N, 10 miles south of the river mouth, but a young stand of spruce may grow near Stanton, according to an unconfirmed report.

Spruce formerly grew north of their present "limit", judging by fairly numerous logs which are found up to 30 miles beyond the tree-line. For instance, near Kittigazuit spruce stumps, over 8 inches in diameter, are buried beneath recent peat and windblown sand; logs have been found near Tuktoyaktuk; in the Eskimo Lakes many dead spruce occur north of the tree-line; and near Stanton spruce logs up to 8 inches in diameter occur in peat bogs. Willow and alder, of large size, also are found in association with the spruce; beaver-chewed logs occur even at the north end of Kendall Island.

168. Open Woodland: Open woodland covers most of the area east of Campbell-Sitidgi Lake. The predominant tree is the white spruce which grows in valleys, up many slopes, and as isolated, widely dispersed trees on many hill summits. The word "open" refers to the vegetation cover between the open stands of spruce. The

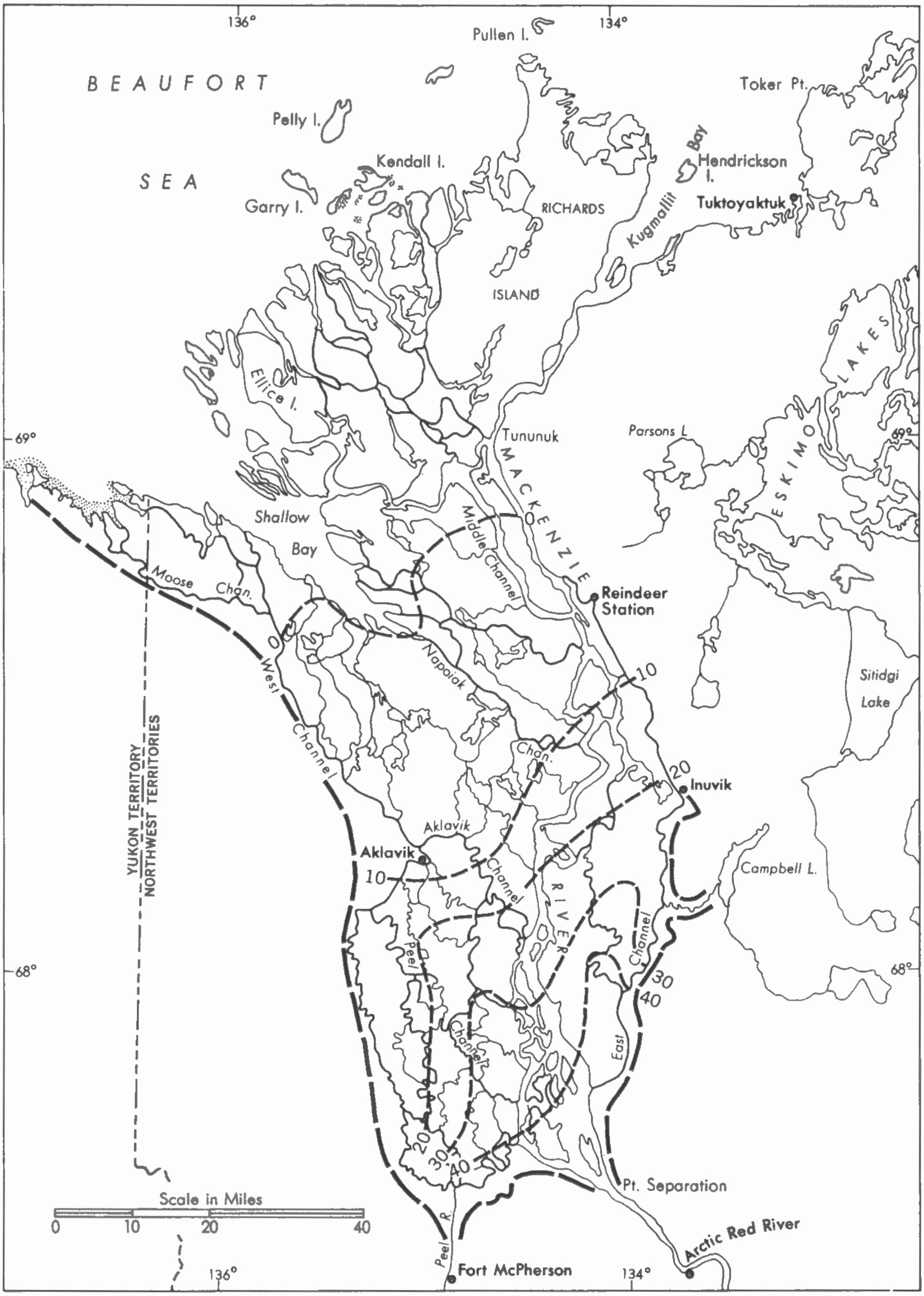

Figure 71. Per cent of Mackenzie Delta in spruce as measured from air photographs.

open cover may be a luxuriant growth of shrubs (willow, birch, alder, and buffaloberry), heaths, and a thick ground mat of grasses, sedges, herbs, lichens, and mosses. The spruce trees may exceed a foot in diameter and an age of 300 years. Hummocky ground is common.

Between Kugaluk River and Crossley Lakes, many tens of square miles of terrain have a downslope vegetation pattern of striped ground (*62*) resembling hachures, where spruce trend in lines alternating with lichen-rich scrub-covered vegetation stripes. Similar striped slopes are present on the terraces surrounding Sitidgi Lake.

Spruce grow along the western slopes of the Caribou Hills, gradually becoming sparser and sparser to the north, until the effective limit of spruce is about 20 miles downstream from Reindeer Station (Figure 74). Spruce thrive best on the north-facing slopes of the gullies which dissect the steep western slope of the Caribou Hills. Spindly black spruce grow in some flattish bogs. In the Dolomite-Campbell lakes area, near Inuvik, tamarack grow on rocky open slopes and in depressions with spruce. In the Anderson Valley, tamarack are most numerous on terraces at, or just above, the floodplain; they extend down the valley to about 69°07′ N, not as far north as the spruce.

Balsam poplar are scattered in small stands, particularly along south-facing gravelly slopes. Aspen poplar seem less common, but appear to have nearly the same range as balsam poplar. Paper birch are common on the west slopes of the Caribou Hills (Figure 73) as far north as Reindeer Station. The birch will grow on thin soil, less than 2 feet deep, over bedrock. In the Anderson Valley, paper birch grow above the floodplain level and extend northwards to about 68° 49′ N.

In poorly drained flats, polygonal ground and string bogs may develop. In many flats, as in the Campbell-Sitidgi lakes lowland and in those southwest of Campbell Lake, dense willow thickets cover endless square miles of poorly drained land.

Fires have burnt over hundreds of square miles of open woodland, woodland-tundra, and scrub vegetation. Large burnt over areas occur between the Caribou Hills and Kugaluk River. Old and young burnt areas are easily recognizable on air photographs by tonal and textural contrasts and straight-lined vegetation boundaries. Most of the burnt-over areas trend northwest—southeast and have apparently burnt to the southeast, being fanned by northwest winds. Fires appear to have burnt most often in treed areas with thick peat. The disruption of the surface cover causes local thickening of the active layer, melting of ground ice, and topsy-turvy blackened trees precariously balanced in a morass of collapsed hummocks.

169. Woodland: The woodland area has nearly continuous forest stands interspersed with scrub. The forest stands are not limited to the valleys but spread up slopes and even onto some summits. The area of continuous forest cover is quite small.

170. Peat: In the tundra, peat deposits are usually associated with tundra polygons. Cross-sections show that peat, with much sphagnum moss, is found in association with most low-centered tundra polygons (Figure 22). Large blobs and discontinuous layers of sphagnum moss up to 2 feet thick underlie the soil of many

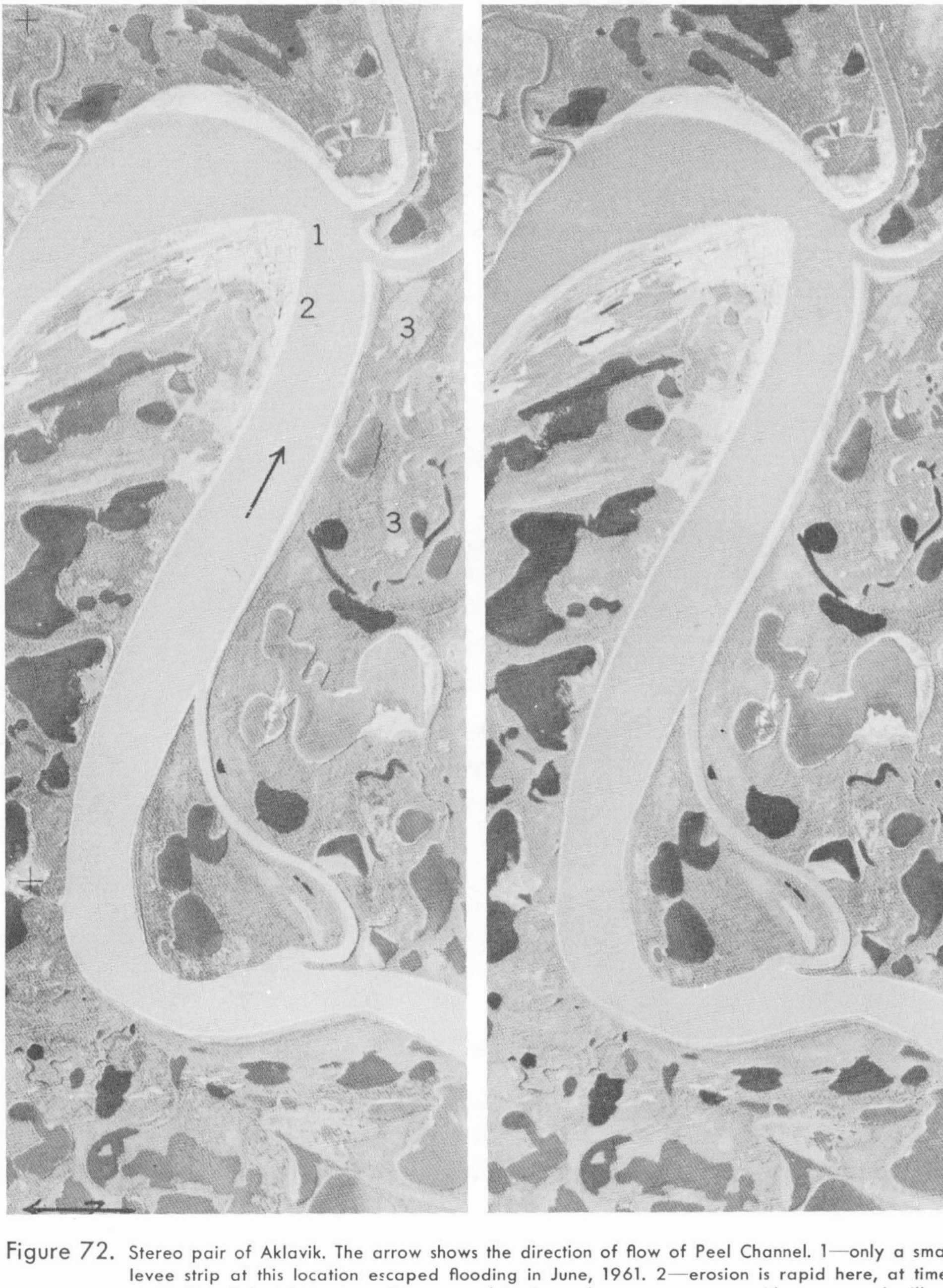

Figure 72. Stereo pair of Aklavik. The arrow shows the direction of flow of Peel Channel. 1—only a small levee strip at this location escaped flooding in June, 1961. 2—erosion is rapid here, at times exceeding 5 feet in a year. 3—areas of tundra polygons overgrown with spruce and willow-alder. Two small channels (Enoch and Pokiak) enter Peel Channel opposite Aklavik. Enoch Channel is the westerly one closest to location 3. Small settlements are on the right bank of both channels. (RCAF A 12848—211 and 212).

tundra polygons, usually at the bottom of the active layer, adjacent to the ice-wedges, or above any tabular ground ice sheets which are present. The burial of surface-accumulated sphagnum moss is believed due to soil movement (62).

Thick peat beds comprise the surface matter of high-centered polygons. The peat may contain logs and branches of spruce, willow, and alder inter-stratified with other organic matter. As an impression, tree and shrub remains are most numerous near the bottoms of the peat bogs. The rate of accumulation of peat can be estimated for two localities. At Inuvik (Figure 73) the bottom of a 12-foot peat deposit has been dated at 8200 ± 300 years (G.S.C. -25) and that of a 7.5-foot peat bed in the Eskimo Lakes at 7400 ± 200 years (G.S.C. -16). As the specimens were obtained from cleaned-off faces of the deposits, the actual thicknesses might be slightly different because of slumping and other causes. The figures give average rates of accumulation of about .7 and 1.0 foot per thousand years.

170. Mackenzie Delta Spruce, Willow, Alder, Poplar: A traveller sailing down a delta channel gains the impression of an endless forest as trees line most of the higher banks (Figures 43; 71). From the air, however, the view is entirely different, because the forested areas are shown to be narrow ribbons along the channels with the remainder of the delta in willows, alders, sedgy flats, and lakes. The distribution of vegetation is controlled, to a considerable degree, by fluctuations of river level, particularly that associated with flooding and sedimentation. In addition, there are vegetation changes which respond to a rise in the level of the permafrost surface induced by sedimentation, erosion, and growth of vegetation. The vegetation types along channels are so responsive to these conditions that the altitudes at which a distinctive vegetation type—for example, a pure stand of alders—appear on opposite river banks usually agree within a vertical limit of several inches.

After the water level has dropped following spring break-up, the gentle lower slopes of channel banks near water level soon appear bright green from a growth of horsetails, sedges, and grasses. These lower slopes and flats are sites of annual sedimentation or, at times, erosion. The swamp horsetail, *Equisetum fluviatile,* (all specific identifications of plants are by W. J. Cody, Science Services, Department of Agriculture) usually grows closest to channel level and is succeeded upslope by the aquatic sedge (*Carex aquatilis*), water-oats (*Colpodium fulvum*), common horsetail (*Equisetum arvense*), and willows. The variation in willow size and species is considerable. The first willows above the channel level tend to grow to nearly uniform height in any one area, for example, 2 to 4 feet high. Common species are *Salix alaxensis, Salix farrae,* and *Salix pulchra.* Willows continue to grow by sending out fresh roots as the stems are buried by sedimentation, provided it is not excessive.

Alder thickets appear just above the limit of annual flooding, judging by the distribution of fresh sediment in the spring. Alder thickets thus lie slightly below levee level. If driftwood is present, it tends to be concentrated in the vertical belt occupied by nearly pure stands of alder. The larger alder reach

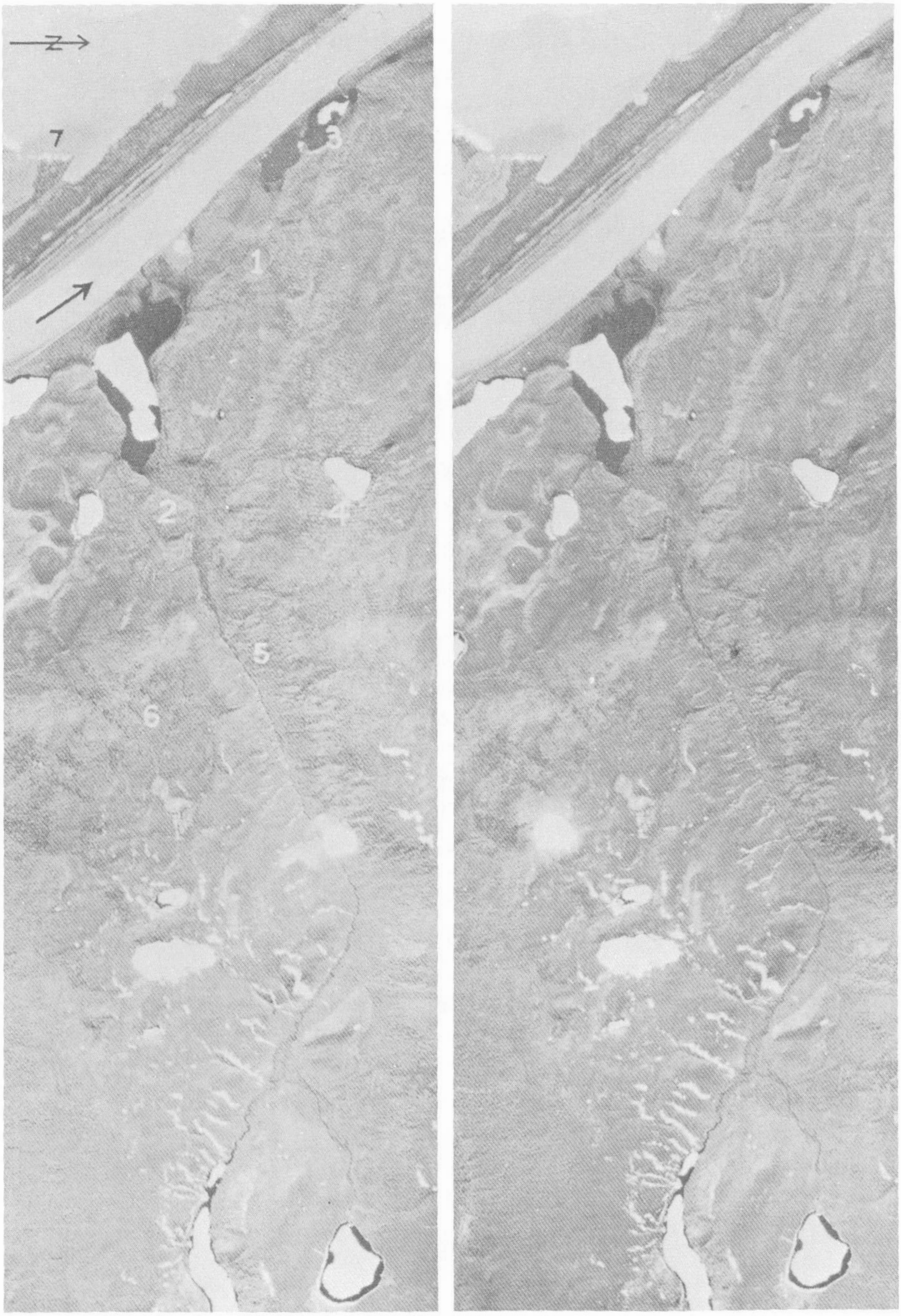

Figure 73. Stereo pair of the site of Inuvik prior to construction of the new town. East Channel flows in the direction of the arrow. 1—Inuvik is built on the terrace. The willowy trees are paper birch. 2—location of specimen of wood with radiocarbon date over 39,000 years (G.S.C.—29). 3—location of peat deposit and wood specimen with radiocarbon date of 8,200 ± 300 years (G.S.C.—25). 4—water supply and storage lake. 5—Boot Creek. 6—spruce terminating near the hill top. 7—levees being built by inflow from location 7 into the next lake. (RCAF A 13406—282 and 283).

heights of 25 feet or more and diameters exceeding 6 inches. Alder growth merges with that of willows to give willow-alder thickets.

Balsam poplar usually grow above the level of the alder stands and below that of spruce, which occupy the highest ground. Poplar thus may precede spruce as colonizers of river banks. Sandy point bar deposits are favored sites for poplars. Poplars are able to maintain growth, by sending out roots, as sedimentation and a rise in the permafrost surface occur. Near the northern limit of spruce, poplars grow to a height of 25 feet, a diameter of 8 inches, and an age of over 100 years. Few stands of poplar are more than 200 feet long and 50 feet wide. Paper birch have not been observed growing in the Mackenzie Delta alluvium.

White spruce is the most abundant tree, black spruce is less common (Porsild, 1945, p. 5). Rarely is a stand of spruce so thick that a lower story is absent. The understory is usually of willow-alder thickets in areas where spruce are far apart; where spruce are close and branches touch, the ground cover varies from bare mud of a recent flood to a cover of heaths, herbs, and grasses.

The distribution of white spruce is closely associated with flood conditions. Along any given channel cross section, the lowest altitude of spruce on both banks is nearly the same, being usually within a few inches. The seaward limit of spruce corresponds to a levee height of about 10 feet above sea level (Figure 57). The precise limit is believed due more to edaphic and geomorphic factors of soil, drainage, and flooding than to microclimatic factors. Exposures in cut banks show that some spruce have survived, by sending out fresh roots, during the time when 6 or more feet of sediment were deposited and the permafrost surface rose accordingly. If sedimentation is too rapid, the spruce trees die, topple over, and break near ground level. The stumps may, in time, become concentrated near low water level by bank erosion and give the erroneous impression of the exhumation of a buried forest.

Spruce trees may reach venerable ages which exceed 500 years (Giddings, 1947). As a very rough estimate, trees 8 inches in diameter may be 75 years old; 12 inches in diameter, 200 years; and 15 inches in diameter, 250 years or older. Large trees indicate a relative degree of stability of the channels which they border. The only tamarack seen in the middle and lower Mackenzie Delta were growing on the rocky islands in the delta and not on alluvial material.

All of the lakes which are interconnected with main channels have their shores flooded when the waters in the channels rise. Such lakes are usually bordered by willow thickets (Figure 60). Other lakes, without connecting channels, are only inundated when flood waters overtop levee banks. The shores of lakes which rarely flood are peaty, as there is little addition of alluvium. Sedges, grasses, pondweeds, mares-tail, and other flora grow along the shores.

172. Mackenzie Delta Willow, Alder, Poplar: The alluvial islands north of the limit of spruce have levee heights of 5 to 10 feet above sea level. Here, poplars grow in scattered small clumps on levees well beyond the limit of spruce and several feet below annual flood levels. Alders are also found throughout the area, but diminish in size and numbers along with a decrease in floodplain altitude. Willows grow along many channel banks and on the island flats. The low spots with standing water or the shores of shallow lakes sustain a growth of marshy vegetation (water-oats and the aquatic sedge), pondweeds, mares-tail, etc.

173. Mackenzie Delta Sedges and Willow: The low seaward alluvial islands are sedgy with tussocks in poorly drained areas by stagnant pools. A few small willows, horsetails, grasses, etc. favor slightly higher ground. Alder and poplar are absent. The islands which are a foot or so above low tide level have much bare soil. Sedges grow below high tide level, even along the banks which are constantly soggy. The alluvial island vegetation helps to trap sediment at flood time.

CHAPTER VI

HUMAN GEOGRAPHY

174. When Franklin and Richardson explored the Mackenzie Delta and coastlands in the summer of 1826, many Eskimos were encountered. Again, in 1848, Richardson met with large numbers of Eskimos at Kittigazuit. Stefansson (1913, p. 452) estimates the population in 1848 as roughly 1,000 at Kittigazuit and 500 at Point Atkinson with a total population exceeding 4,000 for the entire Mackenzie Delta area; other estimates are lower, Jenness (1958, p. 422) giving 2,000 for 1826.

Indian hunting territory overlapped onto that of the Eskimos. The Indians were forest people, who ventured beyond the forest into the barrens in winter for caribou and musk-oxen while the Eskimos were away sealing along the coast. In summer, Eskimos sought caribou in the barren lands vacated by the Indians. Several bands of the Hare Indians frequented the upper Anderson River (Osgood, 1932) and the Loucheux inhabited the basin of Peel River to its junction with the Mackenzie River (Jenness, 1958, p. 399).

Hundreds of Eskimos and Indians perished from diseases introduced by Europeans. From about 1890 to 1910, a large number of whaling ships operated in the Western Arctic with many unfortunate consequences for the natives. In addition to spreading disease, they upset the delicately balanced native economy by destroying local self-sufficiency and encouraging a cash-trapping mode of life. Jenness (1958, p. 422) estimates that in 1929, only about 12 Eskimos were really native to the district between the Alaska—Yukon boundary and Cape Bathurst, the remainder of the existing population of 800 having migrated from Alaska. By the late 1950's and early 1960's the native population was restricted almost entirely to the vicinity of the Mackenzie Delta and Tuktoyaktuk. The coastland (with the exception of DEW line stations), the Eskimo Lakes, the area to the south of the Eskimo Lakes, and the Anderson Valley were uninhabited in both summer and winter, with very few exceptions (*see* Gajda, 1960).

The population of the delta and adjacent areas in the 1950-to-1960 period was over 2,000 of whom about 50 per cent were Eskimo, 30 per cent white, and 20 per cent Indian. The Eskimos are concentrated in the lower portion of the delta and the shores of Kugmallit Bay, whereas the Indians are in the upper part of the delta and at Fort McPherson, an Indian settlement (Henoch, 1961). The whites are mainly in the settlements, although some are trappers.

The basic economy of the delta has been founded on fur trapping. The most important fur-bearing animal is the muskrat, which breeds in the myriad lakes and, in lesser numbers, in the channels (Stevens, 1953). In the eleven-year 1950-to-1960 period inclusive, the total return was about 1,792,000 pelts with an annual range from 64,000 to 284,000 pelts (Black, 1961, p. 80). The average number of trappers was about 300 and the average value per trapper about $380. In the 10-year period of 1950 to 1959 inclusive, about 11,000 mink were caught by an average yearly number of 120 trappers who received an average yearly value of $140.00 (Black, 1961, p. 69). Clearly, trapping has not supplied an adequate cash income in recent years. Some trappers add to their cash income by seeking partial or seasonal employment, supplementing it by cutting logs, selling a few fish, and so on.

The homes of trappers are situated on the highest ground along channels which permit the best landing facilities for boats and protection from floods. Naturally, the sites are usually on levees by cut banks. The cabins and storage sheds are constructed of logs, with sod being frequently placed on the roofs. Each trapper normally has a team of dogs whose insatiable appetite is an economic millstone for the trapper. Large numbers of fish are required for dog feed. The fish are caught in gill nets all year and by jigging through the ice in winter. Whitefish, inconnu, and pike are abundant in the delta. Many trappers go to the coast in summer for the white whale hunt and a few to the Richardson Mountains in the fall for caribou (Munroe, 1953). The delta is an important breeding ground for waterfowl (Barry, 1960; Porsild, 1943) such as the Canada goose, Pacific brant, snow goose, white-fronted goose, baldpate, mallard, and pintail, large numbers of which are shot for food in the autumn.

Trappers along the coast are able to use driftwood for cabins. Fish caught include those found in brackish water and lakes. Crooked backs, herring, whitefish, inconnu, loche, and arctic char are all caught in gill nets. Ringed and bearded seal are fairly common. The bearded seal, which may ascend some of the delta channels, is found along the coast and in the Eskimo Lakes.

175. Aklavik: The town of Aklavik ("Place of the Brown Bear" or "Place of the Barren Land Grizzly Bear") is on the left bank of Peel Channel where it makes a sharp U-turn (Figure 72.) Two smaller channels (Enoch and Pokiak) enter the right bank of Peel Channel opposite Aklavik. A small Hudson's Bay Company fur trading post was established in 1912 at Pokiak Point, opposite Aklavik, and later relocated in Aklavik. Over the years, a small town grew up with government administrative offices, medical services, schools, stores, and so forth. By 1930 the population of Aklavik and the surrounding area had reached 400, and by 1950, about 1,500. The population of Aklavik fluctuates seasonally, being highest in the summer. The population of the Aklavik area in 1961 was roughly 600 to 700 people, of which the native population constituted about 90 per cent.

Aklavik is built on a floodplain at an altitude of about 21 feet above sea level, or about 18 feet above low water level. As Aklavik is built on a floodplain, the entire area is susceptible to flooding. In the break-up of 1961, for example, exceptionally severe flooding was experienced at Aklavik and other parts of the delta with a visual estimate made from an airplane of 98 per cent under water (R.C. Timmins, personal communication). At the period of maximum flooding, only a level strip about 100 yards long remained as dry land (Figure 72), the remainder of the settlement being under water. Silt was deposited to a maximum depth of 5 feet, with an average depth of 2 feet. The shore on the downstream bend of the point bar (convex bank) built out about 50 feet. Although the flooding in 1961 was exceptionally severe, it demonstrated clearly the disadvantage of a floodplain site for Aklavik.

Quite apart from the flooding problem, Peel Channel is also eroding its left bank, especially along the stretch by the Roman Catholic mission (Figure 72). The average yearly rate of recession, over a period of 30 years, is estimated at several feet. Some of the eroded sediment is swept yearly around the convex bank to be redeposited on and downstream from the government wharf.

For many reasons, of which the flooding and erosional hazards were only two, the Government of Canada decided in 1953 to "move" the settlement to a new site. A relocation survey was made in 1954 and a new site ("East Three", later named Inuvik) was selected on the east side of the delta, about 33 miles to the east-northeast (Merrill and others, 1960).

Aklavik is in a period of transition and its future function remains uncertain (Clairmont, 1962; Spence, 1961). However, as long as there are trappers in the delta, Aklavik will provide service facilities for its environs. Efforts are being made to encourage local industry, such as the manufacture of "fashion" fur garments. Aklavik, unlike Inuvik, has the flavor of a northern pioneer settlement and will long be remembered for its terminus as a jumping-off place, experiments at dairying, the luxuriant vegetable garden of the Roman Catholic mission, the annual visits of the "Banksislanders", the arrival of the spring "banana boat", and so forth.

176. Inuvik: Inuvik or the "Place of Man" (*see* Nuna, 1961, p. 18 for another meaning, viz. "The Happy Man" in the western Eskimo dialect) is the new government-planned town built on the east side of the delta. Selection of the site was made in 1954 and construction completed by 1961. From start to finish, every effort was made to ensure that Inuvik would be able to serve its intended function of a modern northern town (Baird, 1960; Merrill and others, 1960; Sullivan, 1960).

Inuvik is built between the right bank of East Channel and the foothills a mile to the east (Figure 73). The town is along the river on undulating terrain which is terminated inland by a bluff 100 to 150 feet in altitude. Permafrost and ground ice conditions have been taken into account in the construction of buildings, most of which rest on piles so as to minimize disturbance of the thermal regime of the ground.

East Channel provides water transportation for Inuvik and an all-weather airstrip is used for commercial flights. The town has modern services with central heating, running water, electricity, sewers, taxis, hotel, theatre, liquor store, restaurants, government administration offices, hospital, churches, and excellent educational facilities which draw students from long distances. Inuvik is used as a seasonal base for government field parties and some commercial enterprises (e.g. oil exploration). The real problem of Inuvik is the provision of employment for the relatively large population of boys and girls who cannot, under present economic conditions, become trappers. The 1961 population of the Inuvik area was 1800, including 500 school children in residence (Department of Northern Affairs and National Resources, Settlement Survey, 1961).

177. Kendall Island: Kendall Island is the site of an abandoned settlement located on the extreme northwest peninsula of the island. When seen in 1958, all that remained of the settlement were 6 abandoned log cabins. Natives still make periodic visits to Kendall Island for whaling and trapping, but it formerly was a much more important area for white whale hunting and fox trapping. It even boasted of a store! There is a reasonably sheltered anchorage for small boats in the cove formed by the western extremity of the peninsula. Only shallow-draft boats can reach Kendall Island via the Mackenzie River or from the sea. Kendall Island is in a shallow part of Mackenzie Bay, like the southern part of Kugmallit Bay, where white whale can be killed easily. Snow geese nest in large numbers on the low alluvial islands south of Kendall Island and were formerly killed for food.

178. Kidluit Bay: Kidluit Bay is on the northeastern part of Richards Island, due west of Hendrickson Island. Only shallow-draft boats can enter the little harbor. Reindeer herders hold an annual roundup in the corral at the bay. There are the ruins of a few old Eskimo houses dug into high-centered polygons and some graves near the eastern point of the bay (*see* MacNeish, 1956). Except for the period of the reindeer roundup, only a few herder families live in the vicinity of Kidluit Bay.

179. Kittigazuit: Kittigazuit is a name which appears in bold letters on many maps, but the Eskimo settlement has been virtually abandoned for decades, although the seasonal population of the area about a hundred years ago was possibly a thousand. Only several old buildings and a few graves remain at the map location (69° 20′ N and 133° 42′ W) of Kittigazuit. The old site is on an island, separated from the mainland by a drowned valley.

180. Point Atkinson: When Richardson visited Point Atkinson in 1826 and 1848, it was the site of an important Eskimo village whose population, although fluctuating considerably was about 500 people. As with other coastal settlements, the population declined during the early 1900's until there were only about 10 inhabitants in 1924 (Ostermann, 1942, p. 43). With few exceptions, no trappers have wintered at Point Atkinson in recent years. There is a harbour for small boats at the base of the long sandspit, so that the site formerly served as a stopping-off place or wintering ground for trappers. A few old graves, the last remnants of a once large archaeological site, are all that is left to record the former settlement. Coastal recession is now removing these few remains (Mathiassen, 1930, pp. 7-19; Mackay, 1957a, pp. 7-9).

181. Reindeer Station: Reindeer Station is a small settlement of about 80 people at the foot of the Caribou Hills, built just above floodplain level, on the right bank of East Channel (Figure 74). Reindeer Station has been the field headquarters for the government-sponsored reindeer herds. Reindeer were brought from Alaska (1929-1935) to the Mackenzie Delta area with the intention of establishing a government herd which would form the nucleus from which native herds would be cut out. Unfortunately, the experiment, despite considerable effort, has not proved successful in creating native herds. In the twenty-year period 1938-1958, for example, the total number of reindeer showed little absolute increase; the population fluctuated from a low of 5,000 in 1938 to 9,200 in 1943 (Krebs, 1961).

182. Stanton: The abandoned settlement of Stanton is at the foot of a bluff on the northeast side of Wood Bay. The settlement was formerly a small Roman Catholic mission founded in 1937. When the mission was founded, only a white trapper and his family lived there, but natives came later, so that about 5 Eskimo families soon lived at or near Stanton (Mackay, 1958b). There is no harbor at Stanton, the nearest shelter being at either the northeast or southwest sandspits of Nicholson Peninsula (Mackay, 1957a, pp. 10-11). However, the site was chosen partly because there was good fishing and driftwood nearby. Stanton was abandoned by 1954.

The short life of the settlement at Stanton is symptomatic of the depopulation of the Anderson Valley area. When the river was explored by MacFarlane in 1857, the total Eskimo and Indian population was about 600 (Mackay, 1958b, p. 39). The Hudson's Bay Company founded Fort Anderson, on the right bank of the river 35 miles northeast of the forks in 1861 for the Eskimo trade, but it was abandoned in 1866. Since then, the population of the Anderson Valley gradually dwindled until only 5 to 10 families lived in the valley by the early 1940's. A short-lived attempt was made to introduce reindeer at the mouth of the Anderson River in 1938, but a disastrous drowning of some herders resulted in much loss of the herd and a transfer of the population to the Tuktoyaktuk region. The names of some of the herders (Stanley L. Mason, reindeer supervisor, 1936-1944; Peter Kaglik, manager of native reindeer herd no. 2 between the Anderson and Horton Rivers, and Charles Rufus, manager of native reindeer herd No. 1, near the Anderson River mouth), are commemorated in the local toponymy. By the mid-1950's, only one trapper remained in the valley, and he has since left. Under present fur prices and regulations, no person can be financially independent as a trapper in the Anderson Valley.

183. Tuktoyaktuk: Tuktoyaktuk is the best deep water harbor (*49*) between Herschel Island on the west and Cape Bathurst on the east. It is the nearest harbor suitable for ocean-going boats to the Mackenzie Delta. Tuktoyaktuk serves, therefore, as a transfer point for cargo shipped down the Mackenzie River for distribution to settlements in the Western Arctic. Boats of 15 foot draft can enter the harbor.

The settlement at Tuktoyaktuk (formerly called Port Brabant) is on the east side of a north-south peninsula which is 500 to 1,000 feet wide (Figure

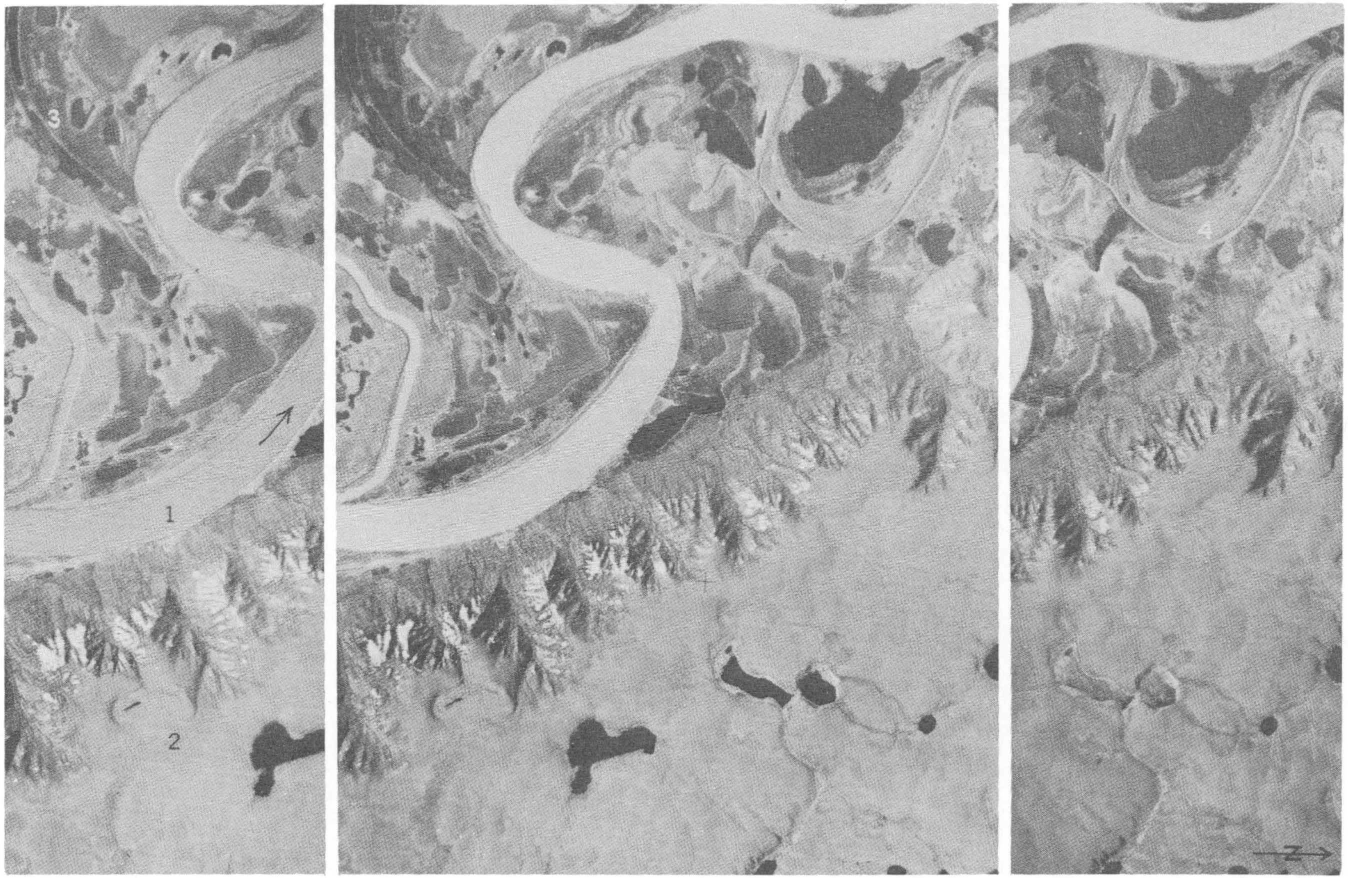

Figure 74. Stereo triplet of Reindeer Station. The direction of flow of East Channel is indicated by the arrow. 1—Reindeer Station. 2—top of Caribou Hills; tundra and scrub vegetation. 3—abandoned channel lined with spruce. 4—abandoned channel lined with willow and willow-alder. (RCAF A 12918—150 to 152).

75). The north end of the peninsula terminates in an east-facing gravelly spit. The coast is retreating slowly under constant wave attack, and the harbor approach must be made with caution. The population of Tuktoyaktuk has varied considerably, with periods of dormancy, decline, and recent increase. From 1900 to 1955, the native population of Tuktoyaktuk and its neighboring area ranged from about 20 to 100. The Hudson's Bay Company established a post at Tuktoyaktuk in 1934. In the ensuing years, with the coming of missions (Roman Catholic and Anglican) and service facilities provided by the government (RCMP, school, etc.) the population gradually increased. With the construction and maintenance of a radar station at Tuktoyaktuk, numbers of Eskimos moved to the settlement. Since 1955, the native population, which is mainly Eskimo, has been several hundred.

Some of the Eskimos obtain full, partial, or seasonal employment in the settlement (Ferguson, 1961). A few are engaged in trapping and reindeer herding (Toker Point to Warren Point). Fish provide a ready food supply for man and dog. White whales are shot from mid-July to mid-August in the shallow waters of Kugmallit Bay.

184. Whitefish Station: Whitefish Station (69° 23′ N and 133° 38′ W) is at the northern end of a drowned river valley which permits through passage for shallow-draft boats drawing 2 to 3 feet. Whitefish Station is used as a white whaling station in the summer. The white whales (beluga) are hunted in the shallow waters of Kugmallit Bay where depths range from 5 to 10 feet. The whales are shot when they surface, are secured by harpooning, and towed to Whitefish Station where they are cut up. The meat and blubber are processed for human and dog food. White whaling is a gala occasion, especially when the hunt is good and numerous boats have gathered from the delta and nearby areas.

185. "Whitefish Station" at Blow River: On the west side of the Mackenzie Delta at Shoalwater Bay there is a second white whaling station often referred to as "Whitefish Station" (Mackay, 1960, pp. 29-30). The entrance to the channel harbor is shoal and difficult to locate. The campsite is usually occupied by a few families in July and early August. There are no buildings—just drying racks and modest accessories for processing whale blubber and meat. The campsite may be flooded in a bad storm.

186. Population Problems: There is no doubt that the Mackenzie Delta area is suffering from a pressing problem of over-population in relation to average income and employment opportunities. The rate of population increase is higher than the national average. As is well known, the income from trapping is totally inadequate to provide an acceptable standard of living to a trapper family, unless supplemented by substantial outside income. The trapping economy is too unstable, dependent as it is on fluctuations of fur-bearing animals, prices, and fashions, to provide a sound economic basis for the area. The problem is not peculiar to the Mackenzie Delta area but is shared by other northern regions (Jenness, 1961). It is hoped that the full potential of the Mackenzie Delta area, in both human and natural resources, may be soon realized for the benefit of all.

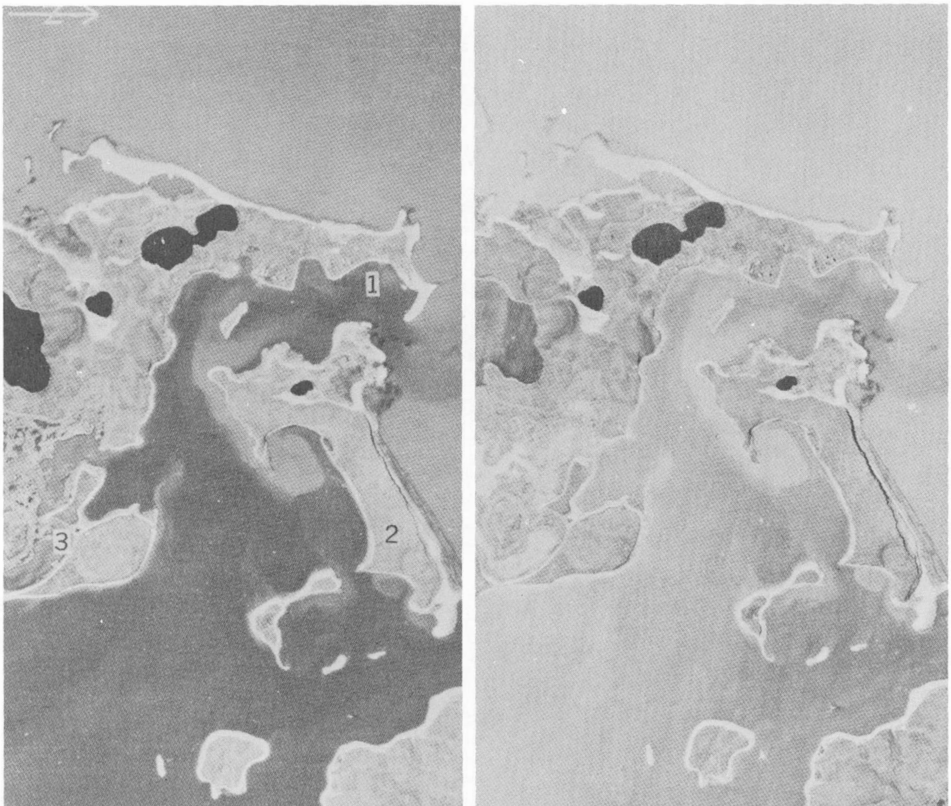

Figure 75. Stereo pair of Tuktoyaktuk. 1—settlement. 2—Port Brabant Island with ground ice slump north of location 2. 3—pingo to the south (left) of location 3. (RCAF A 12902—131 and 132).

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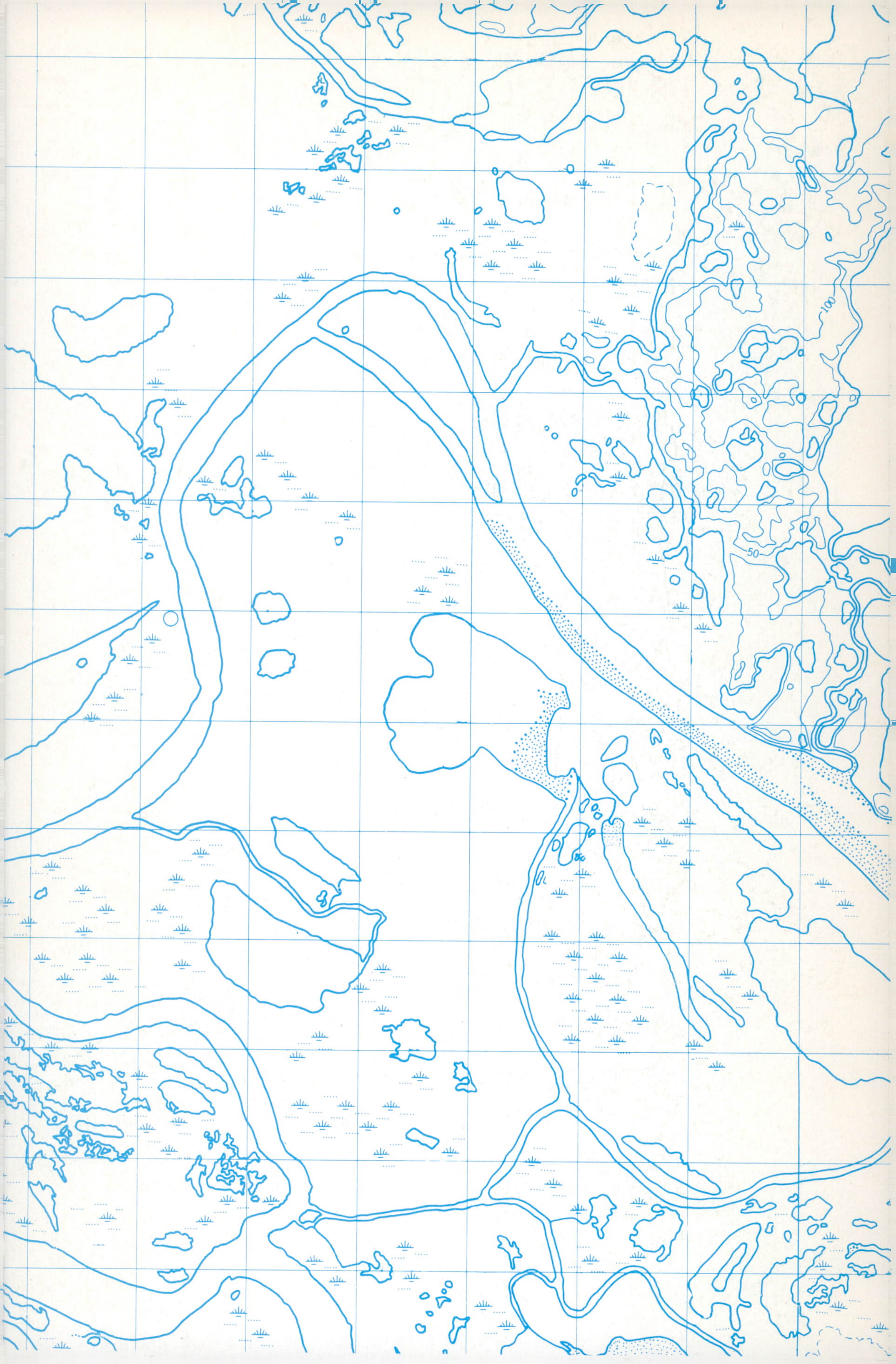
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