GEOLOGICAL SURVEY OF CANADA
PAPER 87-29

SURFICIAL GEOLOGY AND GEOMORPHOLOGY, AISHIHIK LAKE, YUKON TERRITORY

Owen L. Hughes

1990

Energy, Mines and Resources Canada
Énergie, Mines et Ressources Canada

Canada

THE ENERGY OF OUR RESOURCES
THE POWER OF OUR IDEAS

GEOLOGICAL SURVEY OF CANADA
PAPER 87-29

SURFICIAL GEOLOGY AND GEOMORPHOLOGY, AISHIHIK LAKE, YUKON TERRITORY

Owen L. Hughes

1990

Available in Canada through

authorized bookstore agents and other bookstores

or by mail from

Canadian Government Publishing Centre
Supply and Services Canada
Ottawa, Canada K1A 0S9

and from

Geological Survey of Canada offices:

601 Booth Street
Ottawa, Canada K1A 0E8

3303-33rd Street N.W.,
Calgary, Alberta T2L 2A7

100 West Pender Street
Vancouver, B.C. V6B 1R8

A deposit copy of this publication is also available for reference
in public libraries across Canada

Cat. No. M44-87/29E
ISBN 0-660-13249-4

Price subject to change without notice

Cover photo

Shorelines of a glacial lake impounded on Aishihik Upland at the maximum of McConnell Glaciation. Thermokarst ponds on the valley floor are developed on ice-rich glacial lake sediments (site 9, Fig. 3, p. 11).

CONTENTS

SURFICIAL GEOLOGY AND GEOMORPHOLOGY, AISHIHIK LAKE, YUKON TERRITORY

Abstract

Physiographic character of the Aishihik Lake map area ranges from the relatively low relief of Lewes Plateau in the eastern part of the map area to the deeply dissected and mountainous terrain of Nisling Range and Ruby Range within Kluane Plateau in the western and southwestern parts of the map area.

The eastern part was glaciated by the Cordilleran Ice Sheet, and the central and southern parts by a large piedmont glacier derived from St. Elias Mountains. The higher parts of Ruby Range supported local ice caps and cirque glaciers. Most of Klondike Plateau within the map area, together with Nisling Range, escaped glaciation. The known glacial deposits of the map area appear to be related to the Early Wisconsinan Reid Glaciation and the Late Wisconsinan McConnell Glaciation. It is unlikely, however, that the entire area could have escaped earlier glaciations which are recognized in Kluane and Carmacks map areas to the southwest and north, respectively.

Glacial deposition north of Aishihik Lake and glacial erosion at the south end of the lake reversed the former northward drainage of Aishihik and Sekulmun lakes. Glacial lake sediments and abandonned shorelines provide evidence of numerous glacial lakes, of which glacial Lake Sekulmun-Aishihik was the largest, during the maximum stand and subsequent retreatal stages of McConnell ice.

Periglacial features include large forms, such as tors on ridges of porphyritic quartz monzonite, and cryoplanation terraces on unglaciated ridges above treeline. Smaller features such as solifluction lobes and sorted circles occur above treeline throughout the area.

Glacial deposits afford large supplies of aggregate and other construction material, whereas unglaciated areas have deficient supplies.

The map area lies within the discontinuous permafrost zone. Permafrost is common on north-facing slopes and in fine grained glaciolacustrine or alluvial sediments on valley floors. Permafrost severely limits construction in limited areas of ice-rich fine grained sediments, and imposes moderate to severe limitations where it occurs in other materials.

Small amounts of placer gold have been produced in the map area, mainly from creeks tributary to Jarvis River. The placers are the product of postglacial concentration from glacial deposits.

Résumé

La région cartographique du lac Aishihik présente un visage géomorphologique varié; elle s'étend des reliefs peu élevés du plateau de Lewes, dans la partie est de la carte, jusqu'au terrain montagneux parsemé de profonds ravins que sont les chaînons Nisling et Ruby, sur le plateau de Kluane, dans les parties ouest et sud-ouest de la carte.

L'inlandsis de la Cordillère est à l'origine de la glaciation de la partie est de la carte, tandis qu'un vaste glacier de piémont provenant du massif de St-Élie est à l'origine de la glaciation du centre et de la partie sud de la région illustrée sur la carte. Les calottes glaciaires et glaciers de cirque locaux se sont accumulés dans les parties supérieures. La majeure partie du plateau de Klondike dans la région à l'étude et le chaînon Nisling ont échappé à la glaciation. Les dépôts glaciaires répertoriés dans la région cartographiée semblent dater de la glaciation de Reid, survenue durant le Wisconsinien inférieur, et de la glaciation de McConnell, survenue durant le Wisconsinien supérieur. Il est peu probable toutefois que la région cartographiée ait échappé aux glaciations antérieures, observées sur les cartes des régions de Klaune et de Carmacks, situées respectivement au sud-ouest et au nord.

L'accumulation des dépôts glaciaires au nord du lac Aishihik et l'érosion glaciaire survenue à l'extrémité sud du lac ont orienté vers le sud, et non plus vers le nord, le drainage des lacs Aishihik et Sekulmun. Des sédiments lacustres de nature glaciaire et d'anciennes lignes de rivage prouvent que de nombreux lacs d'origine glaciaire, dont le plus grand était le lac Sekulmun-Aishihik, parsemaient la région durant la période d'étendue maximale des glaces et au cours des étapes du recul de la glace de McConnell.

Parmi les vestiges périglaciaires de grande taille, on note la présence de buttes rocheuses mises à nue par l'érosion sur des crêtes de monzonite quartzifère porphyrique ainsi que des terrasses de cryoplanation observées sur des crêtes épangnées par la glaciation aux altitudes supérieures à la limite des arbres. Des éléments de moindre taille comme des lobes de solifluxion et des sols polygonaux se manifestent dans toute la région, aux altitudes supérieures à la limite des arbres.

Les dépôts glaciaires fournissent de vastes quantités d'agrégats et d'autres matériaux de construction, tandis que les régions épangnées par les glaces ne recèlent pas de telles ressources.

La région cartographiée fait partie de la zone à pergéliso discontinu. Le pergélisol se manifeste surtout sur les pentes orientées vers le nord ainsi que dans les lits de sédiments glacio-lacustres ou alluviaux à grains fins des vallées. Dans les rares endroits caractérisés par la présence de sédiments à grains fins à forte teneur en glace, le pergélisol restreint considérablement les possibilités de construction tandis que dans les endroits où ce dernier se trouve mélangé à d'autres matériaux, il restreint modérément ou considérablement de telles possibilités de construction.

De petites quantités d'or alluvionnaire ont été découvertes dans la région cartographié, surtout dans des ruisseaux tributaires de la rivière Jarvis. La concentration post-glaciaire des dépôts glaciaires est à l'origine de ces gîtes alluvionnaires.

INTRODUCTION

Aishihik Lake map area lies in southwestern Yukon, embracing parts of three subdivisions of Yukon Plateau: Klondike, Lewes, and Kluane Plateaus (Fig. 1). The surficial geology and geomorphology of the area were investigated during part of the 1966 and 1967 field seasons and briefly in 1978. In 1966, traverses were made by vehicle along the road that extends from Alaska Highway to the now abandoned Aishihik Airport; boat traverses were made along the shores of Aishihik and Sekulmun lakes; and back-packing traverses were used to extend observations beyond the road and lakes. In 1967 both helicopter and fixed wing aircraft were used to establish fly camps in remoter parts of the area. The helicopter was also used to a limited extent for traversing, mainly to establish the limits of glacial erratics and other glacial deposits. Geological observations were widely scattered and extensive areas remain for which there are no ground data. Accordingly, surficial geology Maps 20, 21, 22, 23-1987 are in large part the result of air-photo interpretation, with minimal ground checking.

For information on access, climate, vegetation, wildlife and history, readers are referred to a concise summary by Tempelman-Kluit (1974, p. 1-5).

Acknowledgments

Able assistance was provided in the field by J.T. Gray and J. Bruce in 1966, by J. Look, D. Reimer, and R. Klaubert

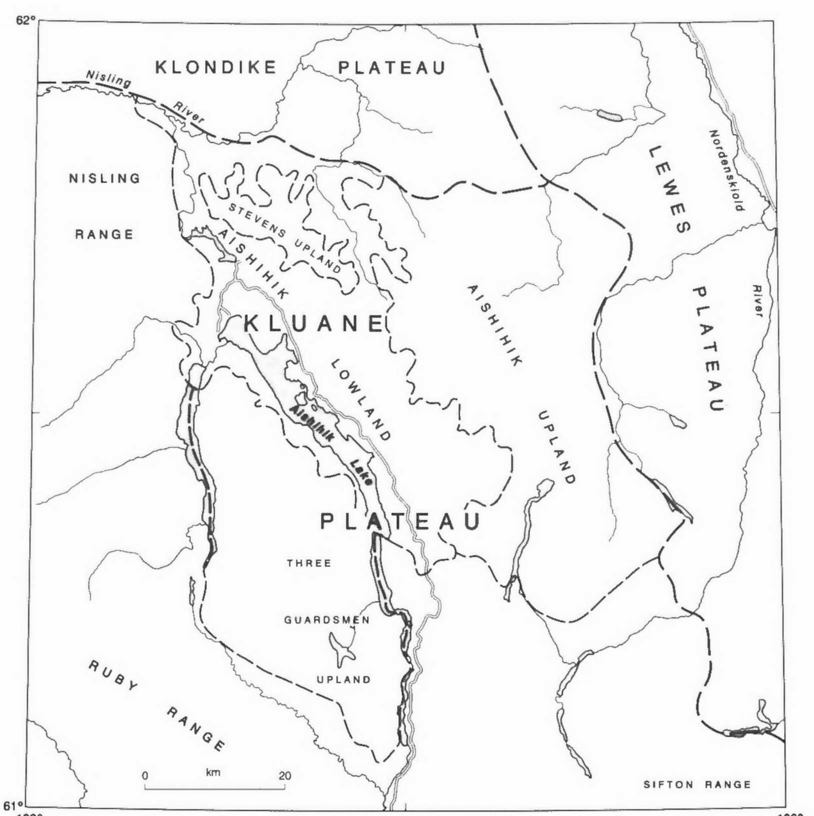

Figure 1. Physiographic subdivisions of Aishihik Lake map area (modified from Bostock, 1948).

in 1967, and by Y. Decoste in 1978. At various times V.N. Rampton (Terrain Analysis and Mapping Services Ltd.) and S.R. Morison (Exploration and Geological Services, Indian and Northern Affairs Canada) have joined me in field work in Aishihik and bordering map areas and have provided much useful discussion of problems of correlation and chronology. D.J. Tempelman-Kluit provided generous assistance in identifying rock types represented in the glacial deposits and indicating their probable sources.

Information on recent placer activity was provided by G.W. Gilbert, Mining Inspector, Department of Indian Affairs and Northern Development, Whitehorse. D. Campbell and L. Sushelnitzky assisted in compilation of manuscript maps from airphotos, and Sushelnitzky drafted the final manuscript maps.

PHYSIOGRAPHIC SETTING

Aishihik map area lies in southwestern Yukon, embracing parts of the mainly glaciated Lewes and Kluane plateaus and part of the mainly unglaciated Klondike Plateau (Fig. 1). The Lewes Plateau section, along the east side of the map area, is part of a broad depression between Pelly Mountains on the east and the generally higher Klondike and Kluane plateaus to the west (Bostock, 1948, p. 65).

Generalized geology is given in Figure 2. The lowest part of Lewes Plateau within the map area, Nordenskiold Valley, is underlain by sandstone, conglomerate, and minor shale of the Lower and Middle Jurassic Laberge Group (Tempelman-Kluit, 1974) and Upper Jurassic to Lower Cretaceous Tantalus Formation. The westward rise to adjoining plateau elements is underlain mainly by Triassic(?) massive green volcanic rocks, plus hornblende granodiorite and porphyritic quartz monzonite of Early to Middle Jurassic age.

That part of Kluane Plateau within the map area is bisected by the Central Aishihik Metamorphic Belt, comprising schist with marble, that extends from the west side of Hutshi Lakes northwestward across Three Guardsmen Upland to and beyond Nisling River (Tempelman-Kluit, 1974, Fig. 13). To the east of the metamorphic belt the plateau is underlain mainly by hornblende granodiorite and pink quartz monzonite of Triassic(?) age and acid tuff of Eocene or younger age. The southwestern corner of the Aishihik Lake map area is underlain by hornfelsed schist. The highest peaks within the study area, including Mount Bark (elevation 2384 m) are developed on these rocks. Triassic(?) Ruby Range granodiorite (**TR**gd), Eocene Coffee Creek granite (**T**g), and Eocene fledspar porphyry (**T**fp) underlie successive areas to the north.

A lowland area within Kluane Plateau designated Aishihik Basin by Bostock (1948), lying east and north of Aishihik Lake is here called Aishihik Lowland. It is extended northward to include low areas of thick drift along Stevens Creek and the middle reaches of McIntosh Creek. The lowland as designated surrounds a small upland called Stevens Upland. The upland area to the east of Aishihik Lowland is designated as Aishihik Upland. As delimited in Figure 1, it includes the southeastern extremity of Klondike Plateau as originally defined by Bostock.

Klondike Plateau in the northwestern part of the map area is developed mainly on biotite schist and amphibolite of the Central Aishihik Metamorphic Belt, and porphyritic quartz monzonite of Early or Middle Jurassic age. A prominent ridge lying within Klondike Plateau in the northwest extremity of the map area is capped by tuff and tuff breccia of the Eocene Mount Nansen Group.

There are marked differences in general landscape aspect within the map area. The differences are attributable to the preglacial topography as determined by bedrock lithology and structure, and subsequent modification (or absence of it) by glacial erosion during the Pleistocene. That part of Klondike Plateau within the study area, plus the northern part of Nisling Range in Kluane Plateau lie at the southwestern extremity of a remarkably uniform area of dissected plateau that includes the remainder of Klondike Plateau and its continuation into Alaska — the Yukon-Tanana Upland — an area some 790 km long and up to 165 km wide (Wahrhaftig, 1965, p. 24, Plate 1). Except for occasional peaks that were high enough to support cirque glaciers, the region is unglaciated. Bostock's (1948, p. 69) description of Klondike Plateau in general is applicable to that part lying within Aishihik Lake map area, and to the northern part of Nisling Range: "The topography is a maze of deep, narrow valleys separated by long, smooth-topped ridges whose elevations are very uniform and which are remnants of an old uplifted erosion surface. This surface shows gentle undulations rising here and there along converging ridges to culminate in monadnocks that consist of dome-like eminences or groups of relatively smooth-sloped mountains". Local relief ranges from 460 to 760 m; the highest peak, in northern Nisling Range, is 1935 m. The ridge crests are typically gently rounded, except for some of the highest ridges where cryoplanation has produced giant steps. The ridge crests and slopes are typically mantled by bedrock detritus, with bedrock outcropping only in tors that stand conspicuously on the ridge crests, or in the risers that separate cryoplanation terraces. Lakes, which are abundant in the glaciated interior of Yukon Territory, are lacking, except for oxbow lakes in alluvial floodplains, and small lakes and ponds within areas of thermokarst alluvial sediments.

Lewes Plateau within the map area is generally lower and has lower relief than adjacent parts of Kluane and Klondike plateaus. The lowest part of the area, generally along Nordenskiold River, is underlain by Mesozoic sedimentary rocks (**J**L, L**K**T) of the Whitehorse Trough Fold Belt (Tempelman-Kluit, 1974, Fig. 13 and Map 17-1973). The remainder is underlain by massive green volcanics (**TR**vb) and hornblende granodiorite (**TR**gdm) of Triassic (?) age, together with porphyritic quartz monzonite (**M**qmp) of Jurassic age in the north and Little Ridge Volcanics (**T**LR) of Eocene or younger age in the south. Even the highest hills at slightly above 1525 m were overridden by successive advances of the Cordilleran ice sheet, giving the entire area a glacially smoothed and subdued appearance. Except for a few canyon-like meltwater channels incised deeply into bedrock, the valleys are broadly rounded to flat, with thick drift cover on the valley floors. Small lakes, mostly occupying depressions in the drift or parts of disused meltwater channels, are scattered throughout the area.

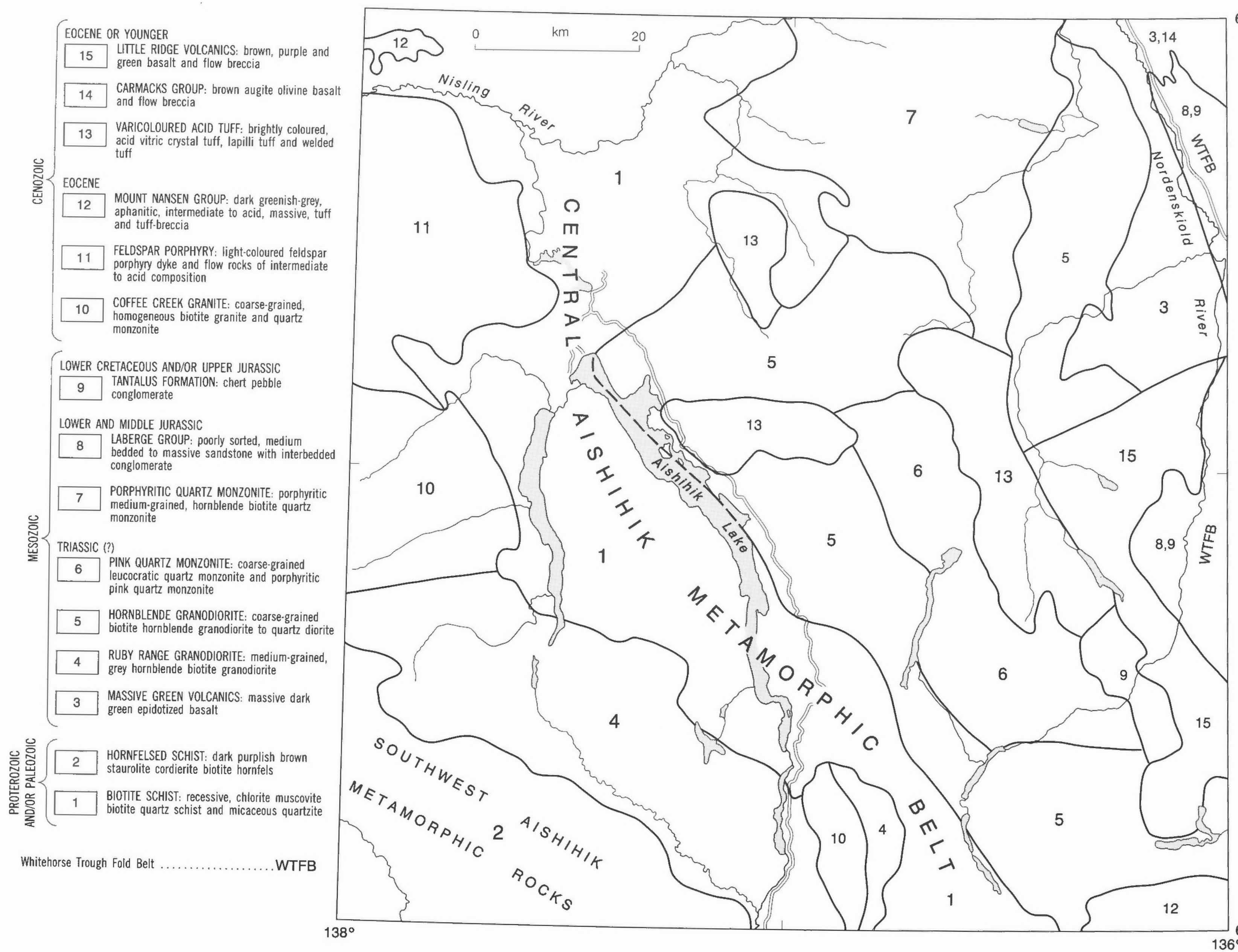

Figure 2. Generalized bedrock geology of Aishihik Lake (after Tempelman-Kluit, 1974, Map 17-1973 and Fig. 13).

In contrast to Klondike and Lewes plateaus, each with its characteristic but oft-repeated landforms, Kluane Plateau is markedly varied. In the southwest corner of the study area, a prominent chain of peaks developed on hornfelsed schist rises abruptly above a drumlinized till plain drained by Jarvis River. The highest peak in the chain, Mount Bark, supports the only permanent ice, a small "glacierette", within the study area. Many peaks, however, supported active cirque glaciers during the McConnell and earlier glaciations. The chain is divided into a series of links by deep trough-like glacially eroded valleys that provide passes through the chain with elevations between 1340 and 1525 m.

To the north of the chain and west of the West Aishihik-Sekulmun Valley is an area of deeply entrenched plateau developed on Ruby Range granodiorite and Coffee Creek granite. The blocks between the main trenches are larger than the "links" of the chain to the south, and the upland surfaces are flatter and more plateau-like. A few cirques supported glaciers during the McConnell Glaciation and earlier glaciations, but ice caps with outlet glaciers leading to the entrenched valleys were more important contributors of ice.

At the southern end of Three Guardsmen Upland, a chain of peaks developed on Ruby Range granodiorite repeats at somewhat reduced scale the chain to the southwest. As in the latter chain, glacially eroded trough-like valleys, some with elongated lakes, provide passes through the chain. The lower northern part of the upland is underlain mainly by biotite schist of the Central Aishihik Metamorphic Belt.

The southeastern part of Aishihik map area comprises mountainous plateau elements separated by broad valleys and areas of lower plateau. All but the highest peaks have been glacially rounded. Northwards along Aishihik Upland there is gradual reduction in both elevation and relief, and at the northern extremity a transition occurs into terrain with little or no glacial modification like that of Klondike Plateau. Most of Stevens Upland was overridden by the Cordilleran Ice Sheet, but glacial erosion was less intense than in areas farther south where glacier ice was thicker.

GLACIAL LIMITS AND FLOW PATTERNS

Bostock (1966, p. 1-2) inferred four distinct advances of the Cordilleran Ice Sheet in the Carmacks map area to the north of Aishihik map area; he named these the Nansen (oldest), Klaza, Reid, and McConnell. In Aishihik map area (Fig. 3, in pocket), the limits of McConnell Glaciation are defined by prominent moraines, ice contact glaciofluvial deposits, and meltwater channels along the eastern flank of Aishihik Upland, along the western flank of that upland, the southwest flank of Stevens Upland, and to the west of Sekulmun Lake. In the southern part of the map area, where high plateau elements stood above the Cordilleran Ice Sheet as nunataks, the limit is less well defined. Moraines and other ice marginal features marking the limit of Reid Glaciation are subdued in comparison with those of McConnell Glaciation and much less continuous. In general, the Reid limit lies only a few kilometres beyond the limit of the later McConnell Glaciation, whereas to the north, in Carmacks and McQuesten map areas, the limits are typically several tens of kilometres apart. In the same map areas drift of the older Klaza and Nansen glaciations, as mapped by Bostock, extends several tens of kilometres beyond the Reid limit, whereas in Aishihik map area erratics have been found only short distances beyond visible ice marginal features of Reid age, so that it is uncertain whether they belong to the Reid or an older drift.

During the McConnell Glaciation, the ice that invaded the eastern part of the study area, essentially Lewes Plateau, was derived from the Cassiar lobe of the Cordilleran Ice Sheet (Wheeler, 1961, p. 10; Hughes et al., 1969, p. 2). Ice of that lobe flowed northwesterly with minimal topographic control, except near the margin of the ice sheet, where numerous tongues projected into valleys along the east side of Aishihik Upland. The southern and central parts were invaded by ice from the Coast and St. Elias mountains that followed tortuous anastomosing pathways through the mountainous plateau area in the south, to form the Aishihik sublobe in Aishihik Lowland. A substantial flow moved northward into the area via the headwaters of Nordenskiold River, merged with westerly moving ice of the Cassiar lobe, then followed a northwesterly path into Aishihik Lowland. A narrow tongue that extended up Long Lake merged with the Cassiar lobe to isolate a nunatak area east of the lake. Ice moving northward into the area via the lower reaches of Aishihik Valley divided to flow along Aishihik Lake, along West Aishihik Valley, and by anastomosing pathways through Three Guardsmen Upland. Drumlins indicate strong northwesterly flow of ice in Jarvis River valley, and a major distributary from that flow moved through the low Shutdunman Pass to merge with ice moving northwesterly along West Aishihik Valley. Ice also moved northward through Twelfth of July Pass, was augmented by local ice, but failed to join ice that flowed via Shutdunmun Pass and West Aishihik Valley.

The extent to which ice moved northward through the higher passes (Ruby, Granite, West Bark, East Bark, and others) of Ruby Range during McConnell Glaciation remains uncertain for several reasons. Firstly, as discussed below, the upper limit of McConnell ice is difficult to determine where the ice impinged against high surfaces. Secondly, cirques and ice fields in the higher parts of the range contributed ice that extended into the passes, so that fresh glacial landforms within the passes would not of themselves signify through-flowing ice. Detailed study of the lithology of glacial deposits will be required to distinguish the respective deposits of through-flowing ice, local ice, and confluent through-flowing and local ice. Finally, the presence of erratics is not diagnostic because they may have been transported into the passes during the earlier and more extensive Reid Glaciation or possibly a still older glaciation. There is a need for development of criteria for distinguishing between different ages of erratics.

Ice from cirques and ice caps in Albert Creek drainage to the north was restricted to headwater valleys. Nisling Range in the northwestern part of the map area is mostly unglaciated, but relatively well preserved end moraines are found downvalley from a cluster of cirques. Uncertainty remains, however, as to whether the moraines are of McConnell or Reid age.

The McConnell limit is well defined along the east side of Aishihik Upland by almost continuous moraines and other ice marginal features north of about 61°21'N, and along both sides of Aishihik sublobe north of about the same latitude. To the south, where ice wrapped around mountainous nunatak areas and filtered through gaps in the Ruby Range, the limit is much less well defined. In general, the area where the upper limit of the ice is poorly defined corresponds with where the surface of the ice sheet was at an elevation of about 1585 m and higher. A general lack of moraines and ice marginal channels at the higher levels suggests that the equilibrium line elevation on the ice surface was at about 1585 m; above that level there was little meltwater to cut ice marginal channels and little supraglacial debris to form moraines. In the general absence of moraines and meltwater channels, the upper limit of McConnell ice can be inferred from the level to which bedrock spurs are freshly rounded, or the limit to which rock detritus has been stripped off by glacial scouring. Placement of the limit on that basis, whether by airphoto interpretation or in the field, is subjective, depending as it does on original configuration of the spurs and the character of the bedrock, particularly its resistance to glacial erosion and the rate at which it has yielded bedrock detritus since retreat of the ice. Upper limits of the ice estimated from such criteria can differ by a few hundred metres over short distances in situations where, by analogy with modern ice sheets, little difference in elevation would be expected. Because of the unreliability of the criteria, limits based on them are shown on the accompanying 1 : 100 000 scale maps only where they appear convincing. In Figure 3, however, the limits are extrapolated widely to show the general size and shape of nunataks that stood above the ice sheet.

During Reid Glaciation, ice flow patterns are assumed to have been essentially the same as for McConnell Glaciation, although much of the evidence was erased by the later glaciation. The ice was thicker and hence more extensive, and it can be assumed that minor topographic features exerted less control on ice movement than during McConnell Glaciation.

In Carmacks map area immediately north of the study area, Bostock mapped limits of Klaza and Nansen glaciations considerably beyond the limit of Reid Glaciation. Evidence for the older more extensive glaciations comprised occurrences of till and undifferentiated drift deposits, together with stream reversals attributed to glacial damming (Bostock, 1966, p. 2). The limits of the older glaciation, if extrapolated southward into Aishihik Lake map area, would encompass an extensive area beyond the limits of Reid Glaciation as shown in Map 23-1987. Limited ground checking failed to produce glacial erratics or other evidence of glaciation within the extrapolated limits of the older glaciations. Such negative evidence is, however, inconclusive, particularly as much of the area in question has forest or shrub cover and lacks exposures of surficial deposits, making the search for erratics very difficult. Intensive ground study will be required to resolve the contradictory evidence from the two contiguous areas.

STRATIGRAPHY

Exposures that exhibit multiple stratigraphic units are uncommon in Aishihik Lake map area except along West Aishihik River, along Aishihik River above the mouth of the West Aishihik, and along Dixie Creek. An isolated exposure occurs on a tributary to Kirkland Creek, and another in the upper reaches of Nisling River.

Aishihik and West Aishihik valleys

In the lower reaches of West Aishihik River a till (Unit 5, of Fig. 4) is prominently exposed in bluffs to about 5.3 km upstream from the river mouth as measured along the valley trend, but is lacking farther upstream. In exposures between km 3.7 and km 5.2 the till is underlain by sediments (Unit 4), which consist of massive silt with minor interbedded silt and clay. Lenses of stony diamicton and zones with highly deformed bedding are common. Although typical varves are lacking, Unit 4 is judged to be glaciolacustrine in origin. Between km 3.7 and km 7.7 Unit 3, comprising coarse sand to very coarse boulder gravel, is exposed below the glaciolacustrine sediments. Locally the till (Unit 5) is overlain by gravel or by interbedded glaciolacustrine sediments and diamicton, overlain in turn by silt and silty clay, mostly massive, with a few lenses of diamicton (Unit 6). At most exposures there is an Orthic Eutric Brunisol up to 20 cm thick at the top of the glaciolacustrine sediments. Also at most exposures the Brunisol is overlain by an accumulation of cliff-top loess 2 to 15 m thick (Unit 7) with numerous thin organic horizons, few weakly developed Bm horizons, and a layer of White River Tephra 2 to 3 cm thick that typically occurs at about the mid-point of the loess thickness.

At km 5.9, a second till (Unit 2) is exposed near river level and is overlain by gravel similar to that of Unit 3 described above. The till is similar in texture and lithology to the previously described Unit 5, but appears to be a distinct and stratigraphically lower unit that is not exposed elsewhere in the middle and lower reaches of the river. It is probably correlative with till exposed at km 32.6, which is underlain by gravel (Unit 1) and overlain by glaciofluvial sediments of combined Units 4 and 6. Exposures are generally lacking betweem km 7.7 and km 20. Between km 20 and 29, West Aishihik River appears to be incised entirely in glaciolacustrine sediments that are equivalent to Units 4 and 6 but without the intervening till (Unit 5). If the till at km 5.9 is correlative with the till at km 32.6, then Units 1 and 2 might lie below the base of exposures situated between the two localities. At km 1.4 on Aishihik River and at other nearby exposures, till judged to be correlative with Unit 5 is underlain and overlain by glaciolacustrine sediments judged to be correlative with Units 4 and 6, respectively.

No paleosol, organic deposit, nor other evidence of a prolonged depositional hiatus was found in the sections along West Aishihik River or the sections on Aishihik River above the mouth of the former. Hence it is assumed here that all the exposed sediments relate to advance and retreatal stages of McConnell Glaciation. The lower till (and at km 32.6, the underlying gravel) was deposited by ice of an advancing sublobe. During retreat of the sublobe, West Aishihik and Aishihik valleys were occupied by arms of gla-

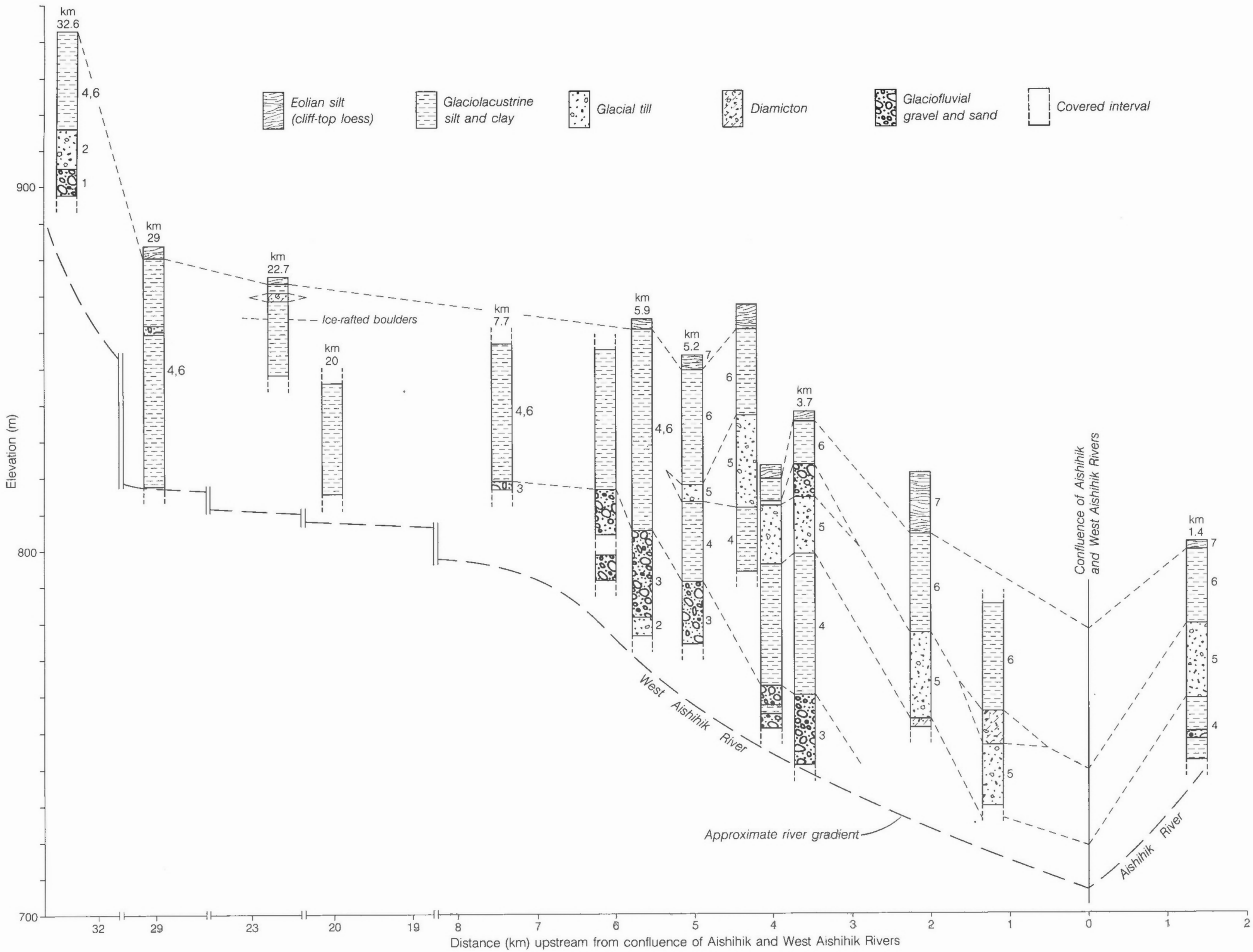

Figure 4. Quaternary stratigraphy, West Aishihik and Aishihik rivers, Yukon Territory.

cial Lake Sekulmun-Aishihik (see *Glacial lakes*) in which the glaciolacustrine sediments were deposited. Glacial Lake Sekulmun-Aishihik expanded southward in step with ice retreat. The gravel (Unit 3) overlying till Unit 2 must then have been deposited into the glacial lake subaqueously, presumably near the outlets of subglacial channels that discharged into the lake at the level of the valley floor. The generally coarse and poorly sorted character of the gravel is compatible with such an origin. When the lobe had retreated to some point south of the junction of the two valleys, it readvanced to km 5.2 in West Aishihik Valley and an undetermined distance beyond km 1.4 in Aishihik Valley, depositing the upper till, and finally retreated south out of the study area.

The common occurrence of contorted silt horizons in the glaciolacustrine sediments can be attributed to subaqueous slumping of the sediments on the rather steep side slopes. The likewise common occurrence of diamicton lenses is attributable to a number of causes, such as flow of ablation debris from the glacier surface, rafting by debris-laden ice, subaqueous slumping, or mudflows originating on steep slopes that bordered the glacial lake. Interpretation of individual occurrences of diamicton in West Aishihik Valley is complicated by the fact that ice flowed not only along the main valley but was joined by ice that flowed northward through the lower passes between Jarvis and West Aishihik valleys. The contribution of ice tongues in those passes to the glacial lake sedimentation in West Aishihik Valley during the retreat-readvance-further retreat stage cannot be assessed from present data, but may have been considerable. Advance of one of the tongues into the glacial lake, possibly during readvance in West Aishihik Valley, would explain the occurrence of very large ice-rafted boulders within glaciolacustrine silt at and near km 22.7.

Glaciolacustrine sediments (Units 4 and 6) at km 1.4 and nearby exposures on Aishihik River differ from those in West Aishihik Valley in that they are mainly distinctly varved silt and silty clay rather than mainly massive silt. The difference suggests slow progressive sedimentation at the Aishihik River localities and rapid sedimentation in West Aishihik Valley, presumably because glacial meltwater was directed mainly into the latter. Intercalations of coarse gravel were noted within varved sediments of Unit 4 at km 1.4 and nearby exposures. These might represent brief periods when the main meltwater discharge was directed into Aishihik Valley but detailed data on the lithology and sedimentary structures which might confirm this conjecture are not available.

Dixie Creek

Pleistocene sediments are exposed intermittently for 5 km along the middle reaches of Dixie Creek and tributary gullies. The greatest exposed thickness, almost 120 m, is at site 1 (Fig. 3), where Dixie Creek opens into Jarvis Valley. Table 1 is a composite section measured in partial exposures extending for about 400 m along the right bank of the creek.

Table 1. Dixie Creek section, stratigraphy: Stream-cut bluff on the right bank of Dixie Creek, approximately 7 km upstream from its confluence with McKinley Creek (61°02.1′N, 137°41.5′W; site 1, Fig. 3)

Unit	Deposit	Thickness (m)	Cumulative height above creek level (m)
9	Silt; contains thin (2-5 cm) organic silt zones; White River tephra 1-2 cm thick, 1 m above base	5.0	119.1
8	Silt, eolian; thin brunisol at top	1.0	114.1
7	Gravel, coarse	6.0	113.1
6	Gravel, coarse, with boulders to 25 cm, grading upwards into diamicton with thin discontinuous silt lenses	6.0	107.1
5	Till, dark grey; stony, sandy silt matrix; discontinuous lens of compact silty gravel 3.7 m thick in middle	25.1	101.1
4	Silt, sand, with discontinuous gravel lenses; 15 cm of sharply laminated silt at top	20.3	76.0
3	Till, dark grey; stony, sandy silt matrix	3.1	55.7
	Concealed	27.0	52.6
2	Till, dark grey; stony, sandy silt matrix; slight oxidation of joint faces	18.9	25.6
1	Silt, silty clay; in basal part, silty clay layers 2.5-8 cm thick alternate with silt partings 1-2 mm thick; gradational upwards into alternating silt and clay laminae 1-2 cm thick		
	Concealed to creek level	1.7	1.7

Unit 1 consists of glaciolacustrine silt and clay, deposited in a glacial lake impounded by an ice tongue that was advancing northwestward in Jarvis Valley. With further ice advance, the site was overridden, and till of Unit 2 was deposited. Despite a concealed interval between Units 2 and 3, they are judged to be parts of a single thick till deposited during a single ice advance. The dominantly silty to sandy character of Unit 4 is compatible with relatively rapid deposition into a shallow ice marginal lake during a period when the ice tongue had retreated to expose the site but remained nearby. Unit 4 contains no evidence of weathering, indicating that a readvance that produced till of Unit 5 followed closely on deposition of Unit 4. Unit 6 is mainly glaciofluvial gravel deposited after retreat of the ice from the immediate site. Diamicton layers in the upper part of the unit are mudflows, derived either from the nearby ice tongue or from slopes immediately to the north. Unit 7 is similar but lacks the diamicton layers.

Unit 8 is typical of the rather thin loess deposits that mantle the valley floors and moderate slopes of the region. The Brunisol at the top of the loess indicates a prolonged interval before deposition of Unit 9, which is cliff-top loess characteristic of virtually all extensive exposures in the region. The loess is carried by wind from the bare exposure and deposited within a few tens of metres of the cliff top. The White River Tephra is part of the eastern lobe of that tephra, radiocarbon-dated at about 1220 BP (Hughes et al., 1972, p. 20). Thus the lower 1 m of the unit was deposited prior to about 1220 BP, and the upper 4 m subsequently. Organic silt layers within the loess are incipient soils, probably representing intervals when all or part of the exposure became vegetated, to be re-exposed later due to lateral cutting of Dixie Creek.

Dixie Creek section lacks weathering horizons (soils) or organic horizons that would indicate a significant hiatus between deposition of lacustrine sediments of Unit 1 and loess of Unit 8. Accordingly, the entire section, except the cliff-top loess, is assigned to the McConnell Glaciation and a subsequent period of glacial retreat. The retreat recorded by Unit 4 and readvance recorded by Unit 5 may have been synchronous with a similar advance and retreat recorded in the Aishihik-West Aishihik sections. As suggested in discussion of the latter area, a tongue of ice may have extended, during the readvance, from Jarvis Valley into an arm of glacial Lake Sekulmun-Aishihik occupying West Aishihik Valley. Icebergs calving from that ice tongue may have been responsible for deposition of ice-rafted boulders within glacial lake silt at or near km 21 in West Aishihik Valley.

In several exposures upstream from site 1 (Fig. 3), till correlative with Unit 5 of the latter locality is overlain by glaciolacustrine silt and clay up to 10 m thick. These sediments indicate that during final retreat of ice from Jarvis Valley, the ice impounded a small glacial lake in upper Dixie Creek Valley.

Kirkland Creek

The section (site 2, Fig. 3) is an isolated bluff exposure about 300 m long. When examined in 1967 and 1978, the exposure was much slumped and only gross units were measured (Table 2).

Although the section contains three tills with intervening gravel units, no weathering horizons (soils) or organic deposits were found that would demonstrate significant interglacial or interstadial intervals. Till of Unit 1 and gravel of Unit 2 have a distinctly browner cast than overlying units. The browner cast is characteristic of Reid drift in central Yukon and suggests, but does not prove, a Reid age for Units 1 and 2. Till Units 3 and 5 are similar in general aspect, and together with gravel of Unit 4 are judged to indicate an initial McConnell advance, a short-lived retreat from the site and a readvance, as suggested by stratigraphy in Aishihik and West Aishihik valleys and at Dixie Creek.

There is considerable variation in pebble lithology of the respective units, although volcanic rocks predominate throughout. Most of the volcanic pebbles are probably derived from an area of "massive green volcanics" (TRvb) lying east of the locality (Fig. 2; Tempelman-Kluit, 1974); others may be from an area of Little Ridge volcanics (TLR) to the southeast or an area of tuff with minor flow rocks (Tvr) to the southwest and south. Except for till of Unit 3, a sample of which contained only volcanic pebbles, the units all contain granitoid rocks plus minor amounts of chert, sandstone, and conglomerate from the Whitehorse Trough Fold Belt.

Table 2. Kirkland Creek section, stratigraphy: Stream-cut bluff on the right bank of a northwest-flowing tributary of Kirkland Creek (62°38.9′N, 136°25.2′W; site 2, Fig. 3)

Unit	Deposit	Thickness (m)	Cumulative thickness from base (m)
7	Silt, contains thin (2-5 cm) organic silt zones throughout; White River tephra 10-12 cm thick as undulating band 5.2-6.1 m above base	8.9	115.0
6	Gravel; discontinuous laterally	0.5	106.1
5	Till; greyish brown; sandy matrix with abundant boulders to 1 m diameter; discontinuous silt lenses in upper 1 m	11.6	105.6
4	Gravel, pebbly sand light olive brown; lens of silt 60 cm thick at base	14.0	94.0
3	Till, olive; sandy silt matrix with minor clay; abundant pebbles, cobbles, and boulders	25.0	80.0
2	Gravel, brown; mainly sandy but with bouldery layers 0.3-1.5 m thick	50.0	55.0
1	Till, brown; silty sand matrix with minor clay; sparse pebbles, cobbles	5.0+	5.0

The granitoid rocks, although not studied petrographically, are probably the pink quartz monzonite (TRqm) and porphyritic quartz monzonite (Mqmp) of Tempelman-Kluit (1974). Both rock types are found as secondary components within areas of hornblende granodiorite (TRgdm), the map unit within which the locality is situated; indeed, pink granitoid rock outcrops in the creek bottom at the locality.

Granitoid rocks constitute 33% of pebbles in till of Unit 1. They are lacking in till of Unit 3 and constitute only 3% in till of Unit 5. The intervening gravel, Units 2 and 4, contain 9 and 24% of granitoid rocks, respectively. The high proportion of granitoid rocks in till of Unit 1 as compared with till of Units 3 and 5 suggests a significant difference in direction of ice flow during ice advances that deposited the respective tills, specifically movement from the north during deposition of Unit 1, and from the southeast during deposition of Units 3 and 5. This explanation does not account for the high granitoid content of Unit 4, which is presumed to be retreatal outwash associated with till of Unit 3 or advance outwash associated with till of Unit 5 — tills with little or no granitoid component. It is possible that all the inferred ice movements were from the southeast. Till of Units 3 and 5 could be devoid of or low in granitoid rocks because the local source of those rocks was concealed beneath older units when advances responsible for those tills took place. The relatively high proportion of granitoid rocks in Unit 4 could be explained by local incision of meltwater channels into local granitoid bedrock.

Upper Nisling River

Although the locality (site. 3, Fig. 3) lies some 250 m north of the north boundary of Aishihik map area, a description is included herein because stratigraphic sections are rare in the unglaciated areas. At this site, Nisling River flows in a broad valley floored by glaciofluvial sand and gravel and alluvial deposits of Nisling River. The glaciofluvial sediments entered the headwaters of Nisling River via a meltwater channel that leads from a glacial limit of Reid age. Terrain on either side of the valley is part of the dissected Klondike Plateau, developed on porphyritic quartz monzonite (Mqmp) of Mesozoic age (Fig. 2; Tempelman-Kluit, 1974). Tributary streams display simple dendritic patterns, with no evidence of glacial derangement of drainage, and glacial erratics appear to be lacking except in the glaciofluvial gravel. These observations indicate that the locality is unglaciated, although extrapolation of glacial limits shown by Bostock (1966, Fig. 1) would suggest that the locality was glaciated during one or more early advances (see *Glacial limits and flow patterns*).

The exposure is cut into the toe of an alluvial fan-apron, the surface of which grades upslope into a colluvial blanket and then into bedrock with a discontinuous colluvial veneer. The fine gravel that constitutes most of the section (Fig. 5)

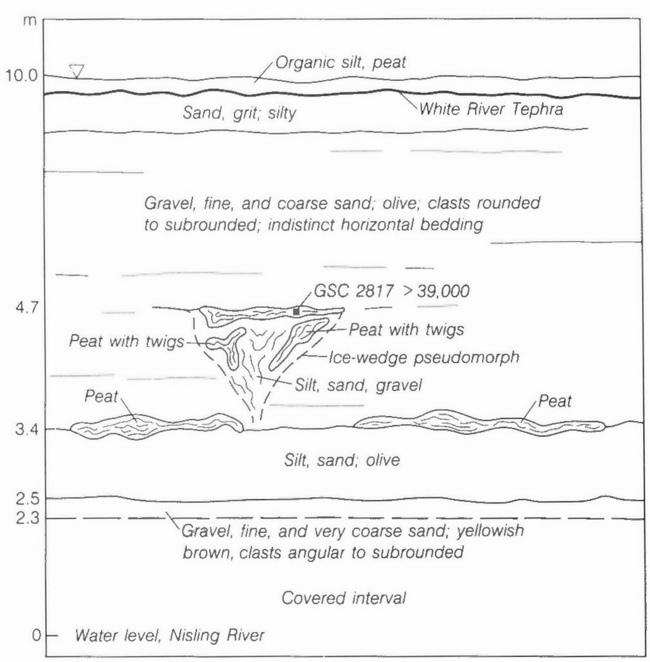

Figure 5. Nisling River section, right bank of Nisling River, approximately 4.7 km (direct distance above the mouth of Victoria Creek (62?00.2'N, 136?59.5'W in 115 I/2).

consists of angular to subrounded coarse sand to small pebble-size clasts of quartz, feldspar or both quartz and feldspar, plus rare clasts of fine grained granodiorite. The sediments are almost entirely derived from weathering of the porphyritic quartz monzonite, which as noted by Tempelman-Kluit (1974, p. 33) disintegrates readily to produce coarse sand.

A discontinuous layer of black partly humified peat about 3.4 m above creek level marks a hiatus in fan deposition, at least at that point. Another hiatus is marked by the top of an ice-wedge cast at about 4.7 m. The cast is filled with peat, sand, silt, and gravel. No peat, sand, or silt is found lateral to the top of the cast. If these formerly extended laterally, they must have been removed by minor erosional truncation prior to renewed deposition of gravel.

Peat from the ice-wedge cast, comprising twigs, sedge, and unidentified plant remains, has been dated as >39 000 BP (GSC-2817; Lowdon and Blake, 1979, p. 29). The peat at 3.4 m and the ice-wedge cast each represent periods of several centuries, perhaps millennia, of stability with no accretion of the fan at that point. Given the limited lateral extent of the exposure (about 50 m), however, it is impossible to determine whether these horizons represent regional climatically controlled interruptions of deposition or merely shifts of the locus of deposition on the fan.

Seasonal contraction cracking of the ground required to produce ice wedges has been thought to require a mean annual temperature considerably colder than the present day mean of -4.2°C recorded at Aishihik Airport (for example, Péwé, 1966, p. 79 and Hughes et al., 1983, p. 12). That assumption appears invalid in the light of new data on ice wedges from near Fairbanks, Alaska (Hamilton et al., 1983). There, ice wedges with apparent widths up to 1 m developed under mean annual temperatures that probably were close to those of the Fairbanks area today (-3.5°C). Hence, mean annual temperature when the ice wedge of the Nisling River site formed was not necessarily colder than that of today.

Thawing of the ice wedge to produce an ice-wedge cast may have resulted from climatic warming. However, ice wedges, particularly near their southern limit of distribution, are subject to destruction due to minor changes in factors controlling the ground temperature regimen, such as loss of vegetation cover or an insulating peat cover due to fire.

The fan sediments themselves afford possible evidence as to climatic conditions during aggradation of fan-aprons of the area. Little or no aggradation is taking place today, nor has there been significant aggradation at the immediate locality since deposition of the White River Tephra about 1220 years ago (Hughes et al., 1972, p. 20). Erosion in the source areas of the fan-aprons is strongly inhibited by boreal forest or subalpine shrub vegetation. Given the small size of most of the drainage basins that contributed to the fan-aprons, active erosion and transport of sediment would seem to require a near-absence of vegetation; under such conditions, sheet wash and flash floods would have been effective agents of erosion, transportation, and deposition of the fan sediments. The appropriate conditions last prevailed in the region during the McConnell Glaciation, when areas beyond the McConnell limit were subjected to extensive wind erosion (Foscolos et al., 1977, p. 4) that could have taken place only on surfaces free or nearly free of vegetation. Thus it seems reasonable to correlate deposition of much of the sediment above the ice-wedge pseudomorph with the McConnell Glaciation. Finite radiocarbon dates are lacking for the locality, and in any case the lower age range for McConnell Glaciation remains to be determined (see *Regional correlation and chronology*) so that the suggested correlation is of necessity imprecise and tentative. The age of the sediments below the top of the ice-wedge cast remains so uncertain that no speculation is warranted.

GLACIAL LAKES

During the McConnell Glaciation numerous glacial lakes undoubtedly formed as advancing ice of the Cordilleran Ice Sheet, together with glaciers originating in Coast and St. Elias mountains, blocked and diverted parts of the existing drainage. Lakes that formed within the areal limits of advance were progressively overridden, and the sediments deposited in them were either eroded by the advancing ice or buried beneath till. Glacial lake sediments deposited during the initial McConnell advance are known only from Dixie Creek (see *Stratigraphy*). Undetected subtill sediments may occur in any valleys that, given the prevailing topography and pattern of ice movement, could have been dammed and then overridden. In general, these would be the same valleys that were occupied by glacial lakes during the retreatal stage of McConnell Glaciation, with a major exception: the depressions occupied by Sekulmun and Aishihik lakes, which constituted a large part of glacial Lake Sekulmun-Aishihik during the retreatal stage, probably drained northward freely during the advancing stage (see *Drainage changes*). Thus no precursor glacial lake stage would be expected.

Glacial lakes must have formed during advance and retreatal stages of Reid and other glaciations, in settings comparable to those of glacial lakes of McConnell age. However, shorelines or glacial lake sediments that can be referred to pre-McConnell glaciations of the area have not been identified.

Glacial lakes impounded at the limit of McConnell advance

Glacial lakes that were impounded beyond the limit of McConnell Glaciation, as well as those that developed later as ice withdrew from the area, have been inferred from glacial lake shorelines, glaciolacustrine sediments, or, in some cases, from both shorelines and sediments. All strandlines associated with former glacial lakes of McConnell age exhibit preferred distribution on slopes with southerly aspect. The preferred distribution may be the product of both stronger initial development and better subsequent preservation. Dominant southerly winds during the McConnell maximum are indicated by the orientation of U-shaped dunes superposed on drift of Reid age in Carmacks (115 I) and McQuesten (115 P) map areas to the north. U-shaped dunes in Dezadeash map area (115 A) to the south and Whitehorse map area (105 D) to the southeast, formed during retreat of McConnell ice, also indicate dominant southerly winds (Klassen, 1978; Rampton, 1981). Although dunes are lacking in Aishihik Lake map area, it can be assumed from the evidence from adjacent areas that dominant southerly winds prevailed during the maximum and retreatal stages of McConnell Glaciation. Accordingly, lake shorelines should have been most strongly developed on southerly slopes, with weaker development on northerly lee slopes. All the shorelines lie between 1250 and 1310 m elevation; within that range of elevation slopes of southerly aspect are typically permafrost free, well drained, and devoid of solifluction features. Slopes of northerly aspect are commonly perennially frozen, imperfectly drained, and exhibit solifluction features. Thus southerly slopes, favoured for development of shorelines, were also most suited for retention of them.

A former glacial lake in Aishihik Upland (site 9, Fig. 3) is unusual in that at maximum lake level it was impounded between ice of the Aishihik sublobe on the west and ice of the main Cordilleran Ice Sheet on the east. Multiple shorelines of the former lake are sharply developed and readily visible in airphotos on slopes with southerly aspect. Only a few faint shoreline traces occur on slopes of northerly aspect.

Shorelines of former glacial lakes at sites 10 and 11 (Fig. 3) are closely and rather regularly spaced. The shorelines suggest that lakes were lowered gradually as their respective outlets were eroded. The glacial lake at site 9 probably discharged northward briefly via a channel incised across the ridge that separated glacial lakes at sites 9 and 10, but closely spaced shorelines below the level of that channel indicate subsequent gradual lowering; at the lower levels the lake must have drained around the east end of the ridge between glacial lakes at sites 9 and 10, either along or beneath the margin of the impounding ice, and thence into the glacial lake at site 10. The ultimate control of lower levels of the lake at site 9 may, therefore, have been the eroding outlet level of the lake at site 10.

Glacial lake sediments occur in a restricted area below the 1160 m level of the lake at site 9. A section comprising 4.1 m of sediments exposed in the steep bank of a thermokarst pond, plus 11.0 m of sediment penetrated by a portable diamond drill, is described in Table 3. The varved sediments of Unit 4, which extend below the limit of boring, are interpreted as deposits of a deep water stage of the lake, and Unit 3 as sediments deposited in the shallowing, shrinking lake. Units 2 and 1 were deposited in a shallow residual lake or pond after surrounding slopes had been invaded by shrub vegetation. The White River tephra is thicker than would be inferred from an isopach map prepared by Bostock (1952, Fig. 1), suggesting some concentration of the tephra in a pond or lake.

Other of the smaller former glacial lakes of the area contain patches of glaciolacustrine sediment of mappable size, and others contain remnant patches that are mapped in complex units together with morainic and colluvial deposits. Still others contain no glaciolacustrine sediments at the surface, but such sediments may be masked by glaciofluvial sediments deposited after lake drainage, or by later colluvium or alluvium.

For most of the minor glacial lakes in the area, the manner in which they were impounded by glacial ice is readily evident. However, although glaciolacustrine sediments blanket gentle slopes on either side of the more northerly of the Hutshi Lakes (Map 20-1987; site 12, Fig 3), it is not obvious how a lake may have been impounded in that valley. Regular southward retreat of an ice tongue in the southern part of the valley is indicated by multiple ice marginal channels incised into slopes on both sides of the valley. A plug of ice in that part of the valley north of Mount Cooper would be required to impound a glacial lake in the valley if one did indeed occupy the full valley width. An extensive area of hummocky morainic and glaciofluvial deposits on the south side of the valley north of Mount Cooper may mark the position of stagnant ice that temporarily occupied the full width of the valley and served as a plug. Alternatively, glacial lakes may have developed on either side of the valley lateral to stagnant ice that occupied the valley axis. In this interpretation the central part of the valley would not have been occupied by a glacial lake, and no plugging mechanism need be invoked.

Glacial Lake Sekulmun-Aishihik

Glacial Lake Sekulmun-Aishihik, by far the largest to exist in the area, came into existence during the retreatal stage of McConnell Glaciation and persisted through a minor readvance in the vicinity of the confluence of Aishihik and West Aishihik rivers. As described in the section *Drainage changes*, the basins of Sekulmun and Aishihik lakes formerly drained northward to Nisling River, and the glacial lake, like the present lake, owed its existence to extensive deposits of McConnell age drift north of the lake. Former shorelines of the glacial lake are mostly isolated short segments. The most prominent shoreline feature is an arcuate scarp at the north end of Aishihik Lake. From the northern extremity of the scarp, the former outlet at an elevation of about 930 m leads northward via Polecat and Stevens lakes.

The fragmentary nature of the shoreline traces, particularly in the south, makes reconstruction of former water planes impossible. However, considering the highest traces at any given locality only, there is a definite inclination upward to the south. On the basis of aneroid determinations the highest discernible former levels rise at a rate of about 1.14 m/km in the northern part of the former lake, increasing to 2.28 m/km in the southern part (Hughes, 1967, p. 49). Within the study area, the highest shoreline of glacial Lake Sekulmun-Aishihik occurs on the slope of West Aishihik Valley (Map 21-1987) at about 1080 m[1] and faint traces occur at about the same level on the east side of Aishihik Valley 2 km south of Aishihik map area (Rampton, 1981). Rampton mapped shorelines at about 1130 m on the east side of Aishihik Valley about 9 km south of the study area, and others at elevations of 1280 m and above still farther south. The latter are in a reentrant where a small glacial lake could have been impounded above the level of

[1] Elevation estimated from topographic maps with 100 foot contour interval and then converted to SI units.

Table 3. Unnamed section, stratigraphy: (Loc. 9, Fig. 2): East side of a small thermokarst pond in a valley draining eastward to Kirkland Creek (61°29.5′N, 136°43.5′W; site 9, Fig. 3)

Unit	Deposit	Thickness (m)	Cumulative thickness below surface (m)
1	Silt, gritty at base and with thin sand lenses near top; small pelecypods (*Sphaerium*) 0.6 m above base; White River tephra up to 40 cm thick immediately below surface vegetation mat	1.6	1.6
2	Silt, sandy, in part clayey; organic with irregular peaty stringers and lenses containing twigs and charcoal	1.6	3.2
3	Silt, clayey, minor gritty layers; segregated ice in lenses up to 12 cm thick constitutes about 40% by volume of bulk sediment[1]	8.8	12.0
4	Silt, clay, varved, with varve couplets 2-2.5 cm thick; segregated ice in lenses up to 4.5 cm thick constitutes about 35% by volume of bulk sediment	3.1	15.1

[1] The section below 4.1 m was obtained by boring using a Winkie drill to drive a 7.3 cm I.D. casing with a diamond-set casing shoe. Estimates of ice content are approximate because of poor core recovery.

the main lake early in the retreatal stage. The elevation of 1130 m is therefore taken as the highest Sekulmun-Aishihik shoreline. The shoreline is about 196 m above the present day level of the Sekulmun-Aishihik outlet, indicating a differential isostatic rebound of that amount in southern Aishihik Valley.

The maximum thickness of ice in lower Aishihik Valley and adjacent parts of Dezadeash Valley was about 1065 m. The average thickness was much less, perhaps about 710 m because local high surfaces stood as nunataks above the ice. Assuming a crustal density of 3.0 and perfect isostasy, potential postglacial uplift would be 237 m. In the outlet region, maximum ice thickness was about 305 m, average thickness about 200 m, and potential isostatic uplift about 67 m. The difference in potential uplift for the two extremities of the lake would be 170 m, 26 m less than the observed differential uplift. The difference could be attributed to several possible sources of error: inaccuracy in determination of shoreline levels by aneraoid altimeter and in determination of glacial limit levels from maps with 100 foot contours, arbitrary assumption of the crustal density factor and assumption of perfect local isostasy. The last assumption ignores the flexural parameter of the lithosphere, which is of greatest importance near the ice margin, as in this case. In any case, the crude observations and assumptions are intended only to show that the differential uplift is of the expected order of magnitude.

The transition from glacial Lake Sekulmun-Aishihik, with shorelines up to 1130 m, to glacial Lake Dezadeash, with shorelines near 855 m and lower (Rampton, 1981), has not been studied in detail. Multiple shorelines between 1130 m and 855 m near the confluence of Aishihik and Dezadeash valleys suggest that the transition was progressive or stepwise rather than catastrophic.

Sediments

Glaciolacustrine sediments occur in an irregular, rather narrow belt around the shore of the northern part of Aishihik Lake (Map 22-1987) and locally along the shore of the southern part of that lake and bordering Sekulmun Lake.

At the time of this study, the few exposures of Quaternary deposits along the shores of the lakes were mostly slumped and none provided a complete section of the glaciolacustrine succession. Examination of several partial sections showed that the glaciolacustrine sediments typically rest on till or coarse gravel. The lowest sediments consist of massive to indistinctly bedded silt transitional upward into varves comprising 15 to 20 cm of silt and 1 to 1.5 cm of silty clay. The varves thin upward, mainly at the expense of the silt, and are overlain by indistinctly bedded silt and locally fine grained sand. At the north end of Aishihik the uppermost 2 m of a 15 m-high bluff includes at the base 1.1 m of silty clay containing segregated ice as veins 6 to 12 mm thick. Near the top of the silty clay is an irregular layer of organic detritus dated 7170 ± 140 BP (GSC-755, Lowdon and Blake, 1970, p. 75). The silty clay is overlain by 60 cm of peat and organic silt containing molluscs, 6 cm of dark brown to black soil, 20 cm of White River Tephra, and 6 cm of loess (the last three varying considerably in thickness along the bluff). The sequence records deposition in a pond or shallow lake following drainage of glacial Lake Sekulmun-Aishihik, drainage of the pond and development of a soil, and subsequent deposition of White River Tephra and cliff-top loess.

A control structure was placed at the south end of Aishihik Lake in 1975, as part of the Aishihik River power development. Although the lake level was raised only about 0.5 m above its natural maximum, some old exposures were freshened and a few others initiated, exposing ice-rich glaciolacustrine sediments that slumped or flowed on thawing. A systematic re-examination of the exposures, however, was not carried out.

Only a small part of the total available volume of the former glacial lake in Sekulmun and Aishihik basins was filled with sediments. Both lakes remained rather deep (maximum known depth of Sekulmun Lake is 47 m, that of Aishihik Lake is 122 m; P. Etherton, personal communication, 1985) and fiord-like, especially in their southern parts. West Aishihik Valley and that part of Aishihik Valley beginning below Canyon Lake southward to Dezadeash Valley, appear to contain much more continuous and thicker glaciolacustrine sediments, although it should be noted that geophysical or other data are lacking on possible extent of glaciolacustrine deposits on the floors of Sekulmun and Aishihik lakes.

From evidence available, it appears that retreat of ice from Aishihik and Sekulmun basins was initially slow, accompanied by extensive deposition of glaciolacustrine sediments at the north end of the lake, then increasingly rapid as arms of the lake opened southward. A halt and readvance after the ice had retreated nearly to Dezadeash Valley (see *Stratigraphy, Aishihik and West Aishihik valleys*) allowed time for deposition of the large volume of sediment in West Aishihik Valley. In West Aishihik Valley, significant sediment may have been contributed by meltwater discharging northeastward through passes from Jarvis Valley.

DRAINAGE CHANGES

General

Glaciation resulted in numerous drainage changes within and adjacent to Aishihik Lake study area. The changes range considerably in scope but all involve one or more of the following processes: 1) lowering of pre-existing divides by glacial erosion so that with ice retreat all or part of one drainge basin was added to another; 2) deposition of large volumes of drift in formerly through-draining valleys so that the drift masses constituted divides after ice retreat; 3) diversion of streams flowing from ice-free areas to flow along the ice margin; such diversions became permanent where ice marginal channels became deeply incised or ice marginal moraines blocked re-establishment of former courses following ice retreat; 4) blockage of valleys that drained toward the ice margin from ice-free areas, causing impoundment of lakes that overflowed divides in headwater areas. Downcutting of the glacial lake outlets, in some cases augmented by construction of drift plugs in the lower reaches of the valleys, resulted in permanent diversion; 5) establishment of consequent streams on drift-covered surfaces in positions that no longer coincide exactly with the preglacial courses.

Aishihik Lowland

By far the most extensive drainage change inferred in Aishihik map area (Fig. 3) was reversal of the formerly northward drainage of Aishihik Lowland via Nisling, White, and Yukon rivers to Bering Sea, to its present drainage via Aishihik, Dezadeash, and Alsek rivers to the Pacific Ocean. The reversal illustrates the combined effects of glacial erosion and deposition of a drift barrier. The northward convergence of the valleys containing Aishihik and Sekulmun lakes toward a broad valley leading to Nisling River strongly suggests former northward drainage. The fiord-like character of the southern end of Aishihik Lake and Canyon Lake to the south indicates extensive overdeepening by glacial erosion to form the U-shaped trough by which Aishihik Lowland now drains southward. The probable extent of glacial deepening appears to be of the order of many tens of metres yet a rise of less than 20 m in the lake level would restore the northward drainage that prevailed during the glacial Lake Sekulmun-Aishihik stage. Moreover, the valley north of Aishihik Lake contains an extensive fill of glacial drift (till, outwash gravel, minor glaciolacustrine sediments) that extends to and along Nisling River. The depth to bedrock below the drift fill has not been determined. In an exposure at the north end of Aishihik Lake, however, glaciolacustrine sediments extend down to and presumably well below lake level, suggesting that a buried bedrock channel extends northward at a level below that of the modern lake, and that northward drainage would be restored if the drift were removed.

Despite uncertainties in estimating the extent of glacial erosion at the south end of the lake, and the lack of data on the thickness of drift fill to the north, it is clear that erosion and a drift barrier combined to cause the reversal. Although deposition of the drift fill clearly relates to an early stage of retreat following the McConnell Glaciation, erosion at the present outlet was almost certainly accomplished in stages during McConnell, Reid, and perhaps one or more pre-Reid glaciations. Given the small difference in elevation between the present outlet and the potential northern outlet, it seems likely that the drainage reversal was first accomplished when McConnell ice retreated from the south end of Aishihik Lake.

Mackintosh and Tahte creeks

The anomalous courses of Mackintosh Creek and its largest tributary, Tahte Creek, illustrate several processes and ages of diversion. The farthest headwater tributaries of Mackintosh Creek — streams that formerly drained toward Aishihik Lake — are dammed by moraine of McConnell age (site 4a, Fig. 3) to form Mackintosh Lake and an unnamed lake to the northwest. The drainage now discharges northward, then westward through a canyon (site 4b, Fig. 3) which was first incised across a local divide by meltwater during Reid Glaciation and deepened by further meltwater discharge during McConnell Glaciation. From the west end of the canyon, it flows north in a pass eroded during Reid Glaciation and further eroded during McConnell Glaciation, then northward in a broad glaciated valley. In the lower reaches of that valley, the creek makes an abrupt 120° turn to flow "up" the valley occupied by Tahte Creek, and joins Tahte Creek to flow northward through a deeply incised canyon (site 5, Fig. 3) into the neighbouring valley to the north. The early history of the canyon at site 5 is unclear. It lies beyond discernible ice marginal features of Reid age and beyond any reasonable projection of the Reid glacial limit. It also lies within an extensive area in the north-central part of Aishihik map area where, on the basis of scattered checks, glacial erratics are lacking, apparently ruling out the possibility that the channel was initiated during a pre-Reid glaciation. Most likely, a glacial lake was impounded west of the canyon position during Reid Glaciation, initiating an eastward outlet that had cut down to near its present level to drain the glacial lake before the impounding ice retreated. Deposition of drift in the former northwestward continuation of Mackintosh Creek prevented resumption of the former course.

The southwestern headwaters of Tahte Creek were diverted from Mackintosh Creek during Reid Glaciation via an incised meltwater channel (site 6a, Fig. 3) and were maintained in that course by a prominent moraine (site 6b, Fig. 3). The southeastern tributary includes part of a former tributary of Kirkland Creek, captured by an incised meltwater channel (site 7, Fig. 3).

Nisling River

Beginning 10 km below the mouth of Mackintosh Creek, Nisling River flows for about 5 km in a canyon incised into the upland on the north side of the valley (site 8, Fig. 3). The canyon was initiated as an ice marginal channel when ice abutted against the upland, but the timing of its development is uncertain. Incisement of the channel may have begun during a pre-Reid glaciation, although evidence for pre-Reid glaciation in the vicinity is lacking. It may also have begun during the advancing stage of Reid Glaciation when ice first abutted against the upland. If so, the channel would have been blocked as the ice thickened to attain an elevation of about 1130 m, and would have been reoccupied during retreat following the maximum of Reid Glaciation.

There are no ice marginal channels on the upland surface at 1130 m — the maximum level of Reid Glaciation. Local drainage and meltwater from the east and southeast (including the Mackintosh-Tahte drainage) must have circumvented the ice-blocked segment of Nisling River by another route. The lowest route around the blocked segment of Nisling Valley is northwestward via the head of Lonely Creek into headwaters of Klaza River (site 13, Fig. 3), which follows a glacially deranged course northwestward then southwestward to join Nisling River. According to Bostock (1966, p. 2) much of the Klaza drainage was acquired by glacial diversion, probably during Nansen glaciation, of drainage formerly tributary to Lonely Creek. The broad divide area between the head of Lonely Creek and Klaza River (site 13, Fig. 3) lacks a sharply defined channel that might be expected if it served as a spillway during the Reid Glaciation. Broad low-angle coalescent fans extend toward the valley axis from both sides of the valley, with extensive organic deposits between. These deposits may obscure the outlines of a former channel.

CRYOPLANATION TERRACES, TORS, AND OTHER PERIGLACIAL FEATURES

Cryoplanation terraces

Cryoplanation terraces, also called altiplanation terraces, goletz terraces, cryoturbation terraces, and nivation terraces (St-Onge, 1965, p. 12; Embleton and King, 1975, p. 159-160; Washburn, 1979, p. 237) are widespread in the unglaciated parts of Yukon, from Buckland Hills in the northern part of the territory (Rampton, 1982, p. 19) at least as far south as southern Aishihik Lake map area. Their distribution is thus somewhat more extensive both to the north and to the south than shown by Reger and Péwé (1976, Fig. 5). In southwestern Carmacks map area and northwestern Aishihik Lake map area (Map 22-1987), well developed sets of cryoplanation terraces are found on most of the high ridges developed on volcanic rocks of Eocene Mount Nansen Group (Fig. 6) and locally on biotite schist and feldspar porphyry. In the southern part of Aishihik Lake map area, they are found on high surfaces developed on Mount Nansen Group rocks and on Ruby Range granodiorite.

Individual terraces are up to 600 m long and 450 m wide, with gently sloping (1 to 5°) surfaces. Typically they occur

Figure 6. Cryoplanation terraces developed on volcanic rocks of Eocene Mount Nansen Group, northwestern extremity of Aishihik map area. Residual remnants (tumps) are marked by arrows. GSC113137

in sets, with the individual terraces separated by steep scarps 5 to 40 m or more high. Extensive cryoplanation summit flats (Demek, 1969, p. 7), up to 1800 m long and 500 m wide, are also found in the area. Some of the highest terraces and summit flats are surmounted by tumps (Demek, 1969, p. 7), pyramidal or conical hillocks up to 75 m high. In the northwestern part of the study area, the terraces and summit flats range in elevation from 1370 to 1890 m, the highest surface in that part of the map area. In the southern part of the study area, where the highest peak is Mount Bark at (2219 m), they range between 1770 and 2135 m. The upper limit of white spruce (Picea glauca), the treeline species of the area, is about 1250 to 1310 m, just below the lower limit of cryoplanation terraces. Thus the terraces are found from the lower limit of alpine tundra[1] upward into high fell-field where herbaceous plants are sparse or lacking.

The lowest terraces at 1370 m have almost continuous sedge-shrub-moss cover. Vegetation cover decreases upward to about 1830 m, above which herbaceous plants and mosses are sparse, with vegetation consisting mainly of crustose and foliaceous lichens on rock detritus. The terraces are mantled by 1 or 2 m of coarse rock detritus with some interstitial finer material. Active layer thickness is probably within the range 50 cm to greater than 150 cm reported by Price (1971, Fig. 5, 8, 10, 12) for upper elevations in Ruby Range immediately west of the study area. There, active layer thickness was found to vary considerably with slope aspect, microtopography, and vegetation cover.

[1] Price (1973, p. 235) quoting Love (1970) referred to the same zone in Ruby Range immediately west of Aishihik Lake map area as "subarctic alpine tundra". Love (1970, p. 68), however, emphasized the close parallels between the subarctic and subalpine zones, and did not suggest the hybrid term subarctic alpine.

Polygons exhibiting varying degrees of sorting are common. Vegetation cover indicates that up to about 1830 m elevation the polygons are mostly inactive. At higher levels, mineral soil is commonly exposed in polygon centres, indicating at least a low level of activity. Active movement of rock detritus across the terraces by gelifluction seems inimical to development of sorted polygons, much less to stabilization of them. The cryoplanation terraces are therefore judged to be essentially stable fossil forms.

The cryoplanation terraces of northwestern Aishihik Lake map area lie well beyond the all-time limit of glaciation. In the southern part of the map area the terraces are restricted to mountainous plateau areas that stood as nunataks above the ice sheet during the last Wisconsinan McConnell Glaciation; indeed, they lie above evidence of glaciation of any age. No glacial erratics were noted on the cryoplanation terraces of such nunatak areas, although they were found at levels as little as 30 m below. Presence of cryoplanation terraces seems, in the absence of contrary evidence, to be an acceptable field criterion for recognition of unglaciated surfaces, at least in Aishihik Lake study area, as well as in the Mayo map area, about 250 km to the north-northeast (Hughes, 1983a-d). The lack of observations of erratics on cryoplanation terraces is, however, negative evidence that can be accepted only after a much more intensive search for erractics is carried out than has been conducted to date. The processes of nivation and gelifluction by which the terraces are thought to have formed are clearly unfavourable for retention of erratics, so that noncryoplanated surfaces at the same elevation should also be examined.

Despite uncertainties in the negative evidence which suggests that cryoplanated surfaces of the area have not been glaciated, glacial history appears to offer an explanation of why cryoplanation terraces extend down to an elevation of

1370 m in the nonglaciated northwestern part of the map area, but are not recognized below about 1770 m in the intensively glaciated southern part of the area.

It should be noted here that the disparity in levels is not satisfactorily explained by the rise in the lower limits of cryoplanation terraces from north to south within Yukon Territory. The lower limit is about 610 m in Driftwood Hills east of Old Crow Plain in northern Yukon (Hughes et al., 1981), rising to 1370 m in northwestern Aishihik map area, a rate of rise of about 1.14 m/km. The abrupt rise from 1370 m to about 1770 m from northwestern to southern Aishihik Lake map area, a rate of about 4.4 m/km seems clearly outside the regional trend.

It is appealing, therefore, to explain the high lower limit of cryoplanation terraces on the nunataks in the southern part of the area by supposing that they were developed at lower levels during a preglacial or interglacial interval, but were obliterated by a subsequent glaciation. Uncertainties in defining the upper limits of respective glaciations on the slopes of nunataks in the southern part of the area (see *Glacial limits and flow patterns*) precludes an attempt to determine when cryoplanation was last active in the area, except to note that cryoplanation terraces are lacking below the Late Wisconsinan McConnell limit wherever that limit is clearly defined. The cryoplanation terraces of adjacent Alaska are restricted to pre-Wisconsinan surfaces (Reger and Péwé, 1976) but there, sharply defined terraces occur on surfaces thought to have been glaciated in Illinoian time.

The processes by which the cryoplanation terraces were formed, and the climatic conditions under which the processes operated, remain uncertain. North American and European workers are in general agreement that the features are the product of a periglacial climate, that is, the climate prevailing in proximity to glaciers, or in broader usage, any cold climate. As summarized by Embleton and King (1975, p. 1-3) and by Washburn (1979, p. 2-4), periglacial climate embraces a wide range in annual temperature and precipitation. As noted by Washburn "the diagnostic criterion is a climate characterized by significant frost action and snow-free ground for part of the year". There is general agreement, as summarized by Embleton and King (1975, p. 163) and Washburn (1979, p. 240) that cryoplanation terraces begin as nivation hollows. The original hollow enlarges by the nivation process acting on a retreating scarp, with the resultant debris being transported across the broadening terrace by a complex of processes, of which gelifluction is probably the most important. Tumps are the remnants of a rock mass that has been encroached on from all sides by retreating scarps. Reger and Péwé (1976, p. 105) supported the same origin for Alaskan cryoplanation terraces. They suggested mean summer temperatures between 2° and 6°C during the period of formation of the terraces compared with present mean summer temperatures up to 10°C, and concluded that "under these conditions nivation was more effective in causing scarp recession because the frequency of freezing and thawing of bedrock was greater than now and seasonal snow banks lasted longer each summer, providing more meltwater for frost action during the critical fall freeze-up. Permafrost was shallower and more widespread than now and mass-movement processes were more effective".

That lowering of mean summer temperatures of the region would indeed increase the frequency of freezing and thawing remains to be demonstrated experimentally. Further, if gelifluction is assumed to be the main mode of transport of debris across the terrace treads, then it is difficult to argue that cooler summers would reinitiate active debris transport at cryoplanation sites in Aishihik Lake map area. The thickness of the active layer is commonly less than the diameter (up to 1 m or more) of the largest blocks found on cryoplanation terraces developed on, for example, rocks of Mount Nansen Group. Present active layer thickness would seem therefore the minimum at which such blocky debris could be moved across the terrace tread by gelifluction. Reduction of active layer thickness would serve to bind the larger blocks into the perennially frozen substrate, effectively prohibiting gelifluction. It could be argued therefore that the cryoplanation process would be favoured by an increase, rather than a decrease, of summer temperature, but perhaps more importantly by an increase in precipitation, in order to maintain the active layer in a saturated state.

It is unlikely that the origin and climatic significance of the cryoplanation terraces of Aishihik and adjacent areas can be resolved by study of the existing stable fossil forms. Instead, quantitative process studies are required at sites where terraces of comparable scale are being formed on similar rocks.

Tors

For present discussion a definition of the term "tor", originated by Pullan (1959, p. 54) and adopted by Washburn (1979, p. 78), is used: "A tor is an exposure of rock in situ upstanding on all sides from the surrounding slopes and it is formed by the differential weathering of a rock bed and the removal of the debris by mass movement". In the unglaciated part of Aishihik Lake map area many hundreds of such features occur, ranging from 1 to 2 m in height and 4 to 5 m in maximum dimension, to 18 or 20 m high, and 50 m or more in maximum dimension, that most geomorphologists would call tors. In both glaciated and unglaciated parts of the area, there are also castellated outcrops on steep gullied, actively wasting slopes and on serrate ridges that most would exclude as tors, such features being commonplace throughout the Cordilleran region. No definitive criteria have been proposed that would provide a clear division between tors and "just outcrop". In this report the term "tor" is restricted (so far as possible from airphoto interpretation and limited ground checking) to prominent individual outcrops of rock surrounded by a more or less continuous blanket of rock weathering products. This excludes prominent crags on slopes and ridges that comprise bedrock with or without a thin discontinuous veneer of rock detritus. Specifically excluded as tors are the relatively rare conical or pyramidal hills surmounting cryoplanation terraces, which herein are termed "tumps" (see *Cryoplanation terraces*). Also excluded are prominent outcrops on ridges that were clearly produced by incision of closely spaced glacial meltwater channels across the ridge crest.

Tors are commonly regarded as periglacial features, yet from perusal of literature on tors as summarized by Washburn (1979, p. 78-79) and Embleton and King (1975,

p. 163-167), it is clear that there is considerable diversity of opinion as to the mechanism of tor formation and the climatic conditions under which they were formed. Hypotheses involve either a single stage, in which there is concurrent differential weathering and removal to leave more resistant rock upstanding, or two stages, one of deep differential weathering during a warm preglacial or interglacial interval and a later stage of exhumation under periglacial conditions.

In the present study, observations of tors were incidental to mapping of surficial deposits and attempting to define the limits of successive glaciations. No new definitive evidence is offered on the origin or age of tors of the area. However, Aishihik Lake map area and areas to the north, which encompass the outer limits of successive glaciations ranging in age from more than one million years to Late Wisconsinan, and which display great diversity in bedrock lithology and terain, afford unusual opportunities for study of tors, warranting a summary description of their occurrence herein as a guide to future research.

Tors are most abundant and most spectacular in the unglaciated northern end of Aishihik Upland and the southeastern extremity of Klondike Plateau, where they are developed on Jurassic(?) prophyritic quartz monzonite (Tempelman-Kluit, 1974). The tors are typically aligned along the rounded ridges of the dissected plateau surface, providing the only bedrock outcrop. The ridges and slopes are blanketed by orange weathering coarse sand (grus), derived from the monzonite. Tors are also abundant along ridges in areas of biotite schist and occur widely but less abundantly in areas of amphibolite with interfoliated schist and gneiss and in areas of Triassic(?) hornblende granodiorite. They are rare in areas of Eocene feldspar porphyry, areas of tuff and tuff breccia of Mount Nansen Group and areas of Eocene and younger varicoloured acid tuff.

Within the unglaciated areas, where glacial history can be ruled out as a factor in the distribution of tors, bedrock lithology is a general controlling factor, with tors occurring on medium to coarse grained igneous and metamorphic rocks. Among the former, however, there are no obvious consistent mineralogical, textural, or structural characteristics that distinguish those rocks most favourable to tor formation from those less favourable. The two rock types most favourable to tor development, the Jurassic porphyritic quartz monzonite and the biotite schist, have relatively dense dendritic drainage systems, with rounded but relatively narrow interfluve ridges; drainage of the less favourable rock types, such as Triassic(?) hornblende granodiorite and pink quartz monzonite, is less dense with broad, flattish interfluves. The difference in drainage pattern may be simply another expression of the same mineralogical, textural, and structural characteristics that make rock types favourable or otherwise for tor formation. It can be argued, however, that narrow ridge crests provide preferred sites for tor formation, and hence that rock type influences tor formation by predetermining favourable topographic form, rather than or perhaps in addition to influencing tor formation directly.

Because neither the age of the tors nor the climatic conditions under which they formed are known, only broad speculation is possible as to the role of the ridge form in localizing tor development. The ridge as a whole can be viewed as a first order product of differential weathering and erosion within the landscape and the tors as a second order product of differential weathering and erosion along the ridges. Once formed, the ridges become well drained sites, deficient in moisture with respect to seepage slopes below, or with respect to the flattish interfluves of less dissected terrain. With relatively low moisture available for whatever combination of chemical and mechanical weathering prevailed during tor formation, minor differences in weathering susceptibility of rocks forming the ridges may have been exaggerated. Ridge sites may further favour tor formation, in that weathering products are more readily shed from them than from flattish interfluves.

Distribution of the tors differs markedly between the northern part of Ruby Range underlain by Triassic(?) Ruby Range granodiorite and Eocene Coffee Creek granite, and the southern part, underlain by hornfelsed schist (Tempelman-Kluit, 1974). The northern part consists of undulating plateau carved into large blocks by deep, steep-sided glaciated valleys. There, tors are sparsely distributed between elevations of 1675 and 2040 m, with most of them situated at or near plateau margins overlooking the deep valleys. Castellated outcrops are common on the steep valley walls, particularly those sections that are sharply gullied. Such castellated outcrops are not notably different from those found on very steep slopes throughout the Cordilleran region, and hence are not included as tors in this discussion. The features considered to be tors are all above the level attained during McConnell Glaciation by ice derived from St. Elias Mountain to the south of the study area, or outside the limits of local ice fields and their outlet glaciers. Glacial erratics were found around the base of one tor (at 60°19.3'N, 137°49.6'W) and in a pit formed by tafoni weathering on top of the tor. The erratics must have been emplaced during the Reid or an older glaciation. Although the tor is relatively solid and coherent compared with many in the area, it could not have survived glaciation in its present form. It is possible, however, that this and some other tors in the area may have formed in a preglacial or interglacial interval, survived glaciation as roche moutonnée forms, and were subsequently modified to their present state (Hughes et al., 1972, p. 29).

On the hornfelsed schist of southwestern Ruby Range, tors occur mainly on two types of sites: on rather steeply inclined ridges along the southwest flank of the range, and on broadly rounded ridges within the high interior part of the range. Ridges of the first type display intense glacial scouring and rounding up to about 1525 m. Effects of glacial scouring (as seen on airphotos) decrease markedly above about 1525 m and outcrops become increasingly castellated and tor-like. Sharp ridges with tor-like outcrops extend almost to 1980 m, and a few tors are found on rounded ridges up to 2075 m. There are jagged outcrops on arêtes at still higher elevations west of Bark Pass, that are not considered to be tors.

The tors or tor-like outcrops of the steeply inclined ridges are above the level of scouring by McConnell age ice that moved along Jarvis Valley and extended distributary tongues northward through Ruby Range. The rounded ridges with tors between 1980 and 2075 m lie within the altitudinal zone of local ice accumulation in cirques and small ice fields, during McConnell Glaciation. However, the ridges formed local divides that were probably not ice-covered, or at least did not undergo ice erosion. The upper limit of Reid ice that moved through passes in Ruby Range has not been determined. It is probable that McConnell and Reid upper limits converge southward toward the source region in St. Elias Mountains, so that on the south flank of Ruby Range the Reid upper limit may be only slightly above the McConnell upper limit. The pattern of local ice accumulation during Reid Glaciation was probably similar to that during McConnell Glaciation. If the above assumptions are valid, most of the tors of southwest Ruby Range are on sites unaffected by either St. Elias or local ice during Reid and McConnell glaciations.

In central Yukon Territory, north of Aishihik Lake, Bostock (1966) found that tors (castellated outcrops) were widely distributed in unglaciated terrain and in areas of pre-Reid glaciation, but were lacking within the limits of Reid Glaciation. This relationship holds for the northern part of Aishihik map area where the Reid limit can be defined with some confidence, but cannot be demonstrated in the southern part of the area, where the Reid limit is, in general, highly speculative. Restriction of tors to pre-Reid surfaces suggests that in central Yukon the climatic conditions required for tor formations have been lacking since culmination of Reid Glaciation.

Pedological studies of the respective drifts (Foscolos et al., 1977) show that weathering of Reid drift (and by inference weathering of rock surfaces scoured during Reid Glaciation as well) is much less intensive than weathering of pre-Reid drift. The absence of a history of deep weathering is the most obvious factor distinguishing surfaces that lack tors from those where they are present, suggesting that deep weathering is an essential part of the tor-forming process. If future studies should provide a reliable basis for distinguishing the Reid limit in Ruby Range, however, it would be important to determine whether some of the tors (or tor-like forms) there lie within that limit. It may be that tors or tor-like forms can develop on some rock types, such as the hornfelsed schist of that range, without an interval of deep weathering.

Other periglacial features

Open-system pingos occur throughout central Yukon (Hughes, 1969). They are most common in the dissected unglaciated Klondike Plateau and in bordering areas of pre-Reid glaciation, but are found locally in areas of Reid and McConnell glaciations. Only two have been noted in the Aishihik Lake area — on Klondike Plateau north of Nisling River.

Closed-system pingos are rare in Yukon Territory. Only one small pingo has been noted in Aishihik Lake map area; it rises from the centre of an elliptical marshy area in the former outlet channel of glacial Lake Sekulmun-Aishihik. The setting is characteristic for closed-system pingos (Mackay, 1963, p. 71) and it is on that basis that the pingo is judged to be of the closed-system type.

The tundra polygons that typically occur in flat areas with peat cover farther north in Yukon Territory, are lacking in Aishihik Lake map area. Remarkably uniform orthogonal polygons, however, occur on fine grained gravel of fans in the lowermost reaches of Victoria and Nansen creeks, and a small tributary between them. Less regular polygonal forms occur on glaciofluvial gravel of Reid age along Nisling River and McGregor Creek. The gravel fans are mainly perennially frozen, and the Reid glaciofluvial gravel is in part perennially frozen, but it is not known whether ice wedges that gave rise to the polygons remain or not.

Smaller scale periglacial features, particularly stone circles or nets that occur on cryoplanation terraces and other flattish surfaces, and turf-banked soliflution lobes that occur on slopes, are common above treeline. Such features have been described by Price (1971, 1973) from Ruby Range immediately west of Aishihik Lake map area.

QUATERNARY HISTORY AND CHRONOLOGY

Correlation within Aishihik Lake map area

Of four glaciations recognized by Bostock (1966) in central Yukon Territory, evidence for only the latest (McConnell) and penultimate (Reid) glaciations has been found in Aishihik map area. Although in general ice marginal features of McConnell age are sharp and well preserved, and those of Reid age are relatively subdued, there is considerable range in degree of preservation of each depending upon texture of deposits, angle and aspect of slope, presence or absence of permafrost, among other factors. Also, differences in vegetation, from boreal forest to alpine tundra, influence subjective estimates of degree of preservation. For these reasons, degree of preservation is not in all cases adequate of itself to distinguish McConnell and Reid ice marginal features. Age assignment presents few problems where ice marginal features are continuous or nearly so. However, where the features are discontinuous (a common condition along the Reid limit), heavy reliance is placed on linking features in such a way that the inferred glacial limit is consistent with the surface slope and pattern of flow of the former ice sheets as inferred from all available evidence. The principle applies equally to correlation of glacial limits on nunataks but practical application becomes increasingly difficult with increasing distance of the nunatak from the former ice margin, and is further complicated in Ruby Range by former local ice sources within the nunataks. Correlation of moraines formed by ice derived from isolated local centres in Nisling Range is based of necessity on landform preservation alone. At several localities in central Yukon Territory north of Aishihik Lake map area, Bostock (1966, Fig. 1, p. 11) mapped ice marginal features intermediate in position between Reid and McConnell limits, and suggested that they may represent an important readvance after the Reid maximum. Comparable uncorrelated moraines occur between Reid and McConnell limits in Aishihik Lake map area.

In the valley of Mackintosh Creek upstream from the abrupt bend eastward, two prominent and rather well preserved moraines loops (site 6c, Fig. 3) mark significant readvances of a northwesterly moving ice tongue. Another moraine loop northwest of the abrupt eastward bend of Tahte Creek marks the readvance of a southeasterly moving ice tongue. Ice that formed the two tongues moved northward mainly by way of the gap between Stevens and Aishihik uplands, bifurcating to form the two tongues (Fig. 3). That two ice tongues with a common source appear to have had different patterns of retreat and readvance suggests nonsynchronous marginal fluctuations during the retreatal stage of Reid Glaciation, and suggests the further possibility that the maximum position was not attained synchronously. The diverse sources of ice affecting Aishihik Lake map area would enhance that possibility.

Uncorrelated moraines also occur west of the north end of Long Lake (Map 20-1987; site 9, Fig. 3) between moraines attributable to an ice tongue occupying Long Lake and ice of Aishihik lobe. The moraines appear subdued compared with nearby moraines of McConnell age and may therefore be of Reid age as mapped. On the other hand they may be of McConnell age, with the subdued form attributable to deposition in the glacial lake at site 9 (Fig. 3) and modification during progressive lowering of the lake. A segment of lateral moraines lies above the McConnell limit as interpolated on the east flank of a nunatak area, 10 km southeast of Giltana Lake (Map 20-1987). Ground data from the locality are insufficient to determine whether the moraine segment marks the Reid limit or the limit of a late Reid readvance.

Moraines marking retreatal positions of McConnell ice are common along the northwest limit of the Aishihik lobe, but are rare elsewhere. A succession of moraine segments suggests pulsating retreat of an ice tongue that extended northwesterly up a valley between Sekulmun Lake and Isaac Creek. On the other hand, a broad belt of unoriented rolling to hummocky moraine (Mx) within the northeastern limit of the lobe suggests that ice of that part of the lobe reached its maximum, then wasted as dead ice.

Regional correlation and chronology

The early Pleistocene history of Aishihik Lake map area is obscure. Bostock (1966) inferred two glaciations, Klaza and Nansen (older) in the adjoining Carmacks map area (115 I) to the north, that are older and more extensive than the Reid and McConnell glaciations recognizable in Aishihik Lake map area. However, the lack of firm evidence for pre-Reid glaciation in Aishihik Lake map area, either in stratigraphic section or at the surface beyond the Reid limit, leaves no basis for speculation on the early glacial history. Careful search for evidence of glaciation beyond the Reid limit in northern Aishihik map area, together with careful reexamination of Bostock's evidence for placement of the limits of Klaza and Nansen glaciations in Carmacks map area, is needed if conflicting evidence from the two areas is to be resolved.

Reid and McConnell glaciations were defined on the basis of moraines and other ice marginal features at type localities in the McQuesten (115P) map area (north of Carmacks). Bostock (1966, Fig. 1) mapped the limits of the respective glaciations from the type localities southward to 62°N by tracing more or less continuous ice marginal features on airphotos. Hughes et al. (1969) extended the interpretation through Aishihik Lake map area west and north to Kluane Lake (115 G, F($E^{1}/_{2}$)) and Snag (115 J, K($E^{1}/_{2}$)) map areas to link with the limits of Mirror Creek and McCauley glaciations as defined by Rampton (1969, 1971, p. 280-286).

There are no radiocarbon dates from Aishihik Lake map area that relate to Reid Glaciation. A date from near the type Reid locality indicates retreat of the Cordilleran Ice Sheet from the area before 42 900 BP (GSC-524; Lowdon and Blake, 1968, p. 228); if, as seems probable, dated wood from beneath McConnell till is of post-Reid age, then ice of the Reid advance had retreated some 70 km from its maximum position before 46 580 BP (GSC-331; Dyck et al., 1966, p. 19; Hughes et al., 1969). The onset of retreat may be still older than indicated by radiocarbon dates. GSC-534 was based on wood from a volcanic tephra now known to be the Sheep Creek Tephra. Uranium series dates on bones associated with the tephra indicate an age for the tephra, and hence a minimum age for retreat following Reid glaciation, of about 80 000 years (Hamilton and Bischoff, 1984). The evidence is similar from southwestern Yukon, where organic material from above Mirror Creek deposits has produced dates to greater than 38 000 BP and material from beneath McCauley till, presumed to be of post-Mirror Creek age, has produced dates as old as 48 000 ± 1300 BP (GSC-732; Lowdon and Blake, 1970, p. 79; Rampton, 1971, Table 2, p. 294). The dated deposits may be even older if, as suggested, the dated material was contaminated by modern rootlets. Finite dates from subtill organic deposits have been obtained at Silver Creek near the south end of Kluane Lake (Denton and Stuiver, 1967, Fig. 7, p. 505; Hughes et al., 1972); dates on the organic material range from about 37 000 to 30 000 BP. Palynology and bryology of the deposits indicate nonforested alpine conditions (Schweger and Janssens, 1980) at a site now covered by boreal forest. The deposits cannot therefore represent the whole of the Reid-McConnell nonglacial interval, because soils developed on Reid drift indicate that the interval was somewhat warmer and more humid than at present (Foscolos et al., 1977, p. 11). It is possible that the deposits represent the cooler early or late stage of the nonglacial interval, but it is equally possible that it represents restricted mid-McConnell (mid-McCauley) retreat, perhaps very local in extent or, if of wider extent, not yet recognized elsewhere in Yukon (Hughes et al., 1972, p. 3).

A radiocarbon date on organic deposits above McConnell till near the northern limit of the Aishihik lobe indicates that ice had retreated from the limit by 9660 ± 150 BP (GSC-749; Lowdon and Blake, 1970, p. 75). Retreat probably began considerably before that time, as in Wellesley basin to the northwest, retreat had begun by 13 600 years ago (Rampton, 1971). A date of 7170 ± 140 BP (GSC-775, Lowdon and Blake, 1970, p. 75) is minimum for the end of glacial Lake Sekulmun-Aishihik, but that event, too, probably took place considerably earlier.

Although as much as 17% of Aishihik Lake map area may not have been glaciated, much of that area is upland with thin discontinuous colluvial cover. The most extensive and probably also the thickest deposits of the unglaciated area are the fan-apron deposits of parts of Nisling Valley and some of its tributaries. The depositional history of the fan sediments, inferred from the Nisling River locality (see *Stratigraphy*) remains uncertain. If, as suggested, accretion of the fans took place under conditions of little or no vegetation, then the last significant accretion took place during the McConnell Glaciation.

ECONOMIC GEOLOGY

Construction materials

Gravel and sand of glaciofluvial origin are widespread and abundant throughout the glaciated parts of the map area. They are also found in several valleys in the unglaciated part of Aishihik Upland that carried glacial meltwater from the ice fronts, and beyond the limit of glaciation in Nisling Valley. Glaciofluvial deposits occur only locally in Nisling Range and are generally lacking in Klondike Plateau north and west of Nisling River. In the former area, the floodplains and low terraces of several large streams afford potential gravel sources. In the latter area, mixed colluvial and low-angle alluvial fan deposits typically extend from both sides of the valley to the valley axis, with minimal development of floodplains in which well sorted aggregate could be expected. Sand and minor gravel, however, occur in terraces and the floodplain of upper Schist Creek.

Moderately well sorted beach gravels are associated with some of the glacial lake shorelines. The beach deposits are probably shallow but may be worth investigating where only small volumes of gravel are required and other sources are not ready to hand.

The lithology of glaciofluvial, alluvial, and beach deposits of the area varies greatly from place to place, reflecting the diversity of the bedrock of the region. Detailed study of individual deposits will be required before the gravel can be used for specific purposes such as concrete aggregate.

Deposits of Reid age have relatively thick (1 m or more) soil profiles, in which clasts of schistose and phyllitic rocks are intensely shattered and those of crystalline rocks are weakened by weathering. In such deposits it may be necessary to remove the uppermost metre or so of material, depending on the intended use.

The glaciofluvial deposits are typically well drained and free of permafrost, hence permafrost is unlikely to be a limiting factor in extraction of gravel and sand. Permafrost is known to occur locally where silt or peat occupy channels or depressions in the surface; if necessary, such areas can be stripped in advance of exploitation to permit solar thaw. Alluvial sand and gravel are free of permafrost within the active, vegetation-free channels of streams, but permafrost is common where the sand and gravel are overlain by thick silt or peat. Alluvial deposits should be a "last resort" source of aggregate because of potential stream siltation.

Till of the area typically contains 20% or more coarse clasts (+4 mm) in a matrix of sandy silt with only minor clay. That of moraine ridges and hummocks is commonly very coarse, with 40% or more coarse clasts in a silty sand matrix. Most of the till is suitable for common fill, and over much of the glaciated part of the area, till can be found that provides a suitable mixture of coarse fraction and silt and clay binder for road surfacing. Where till occurs in ridges and hummocks, or on moderate to steep south-facing slopes, permafrost is unlikely to limit exploitation. On north-facing slopes and valley floors, permafrost plus locally high ice content may impose serious limitations. Where more suitable material is unavailable, it may be practical to strip perennially frozen till, allowing it to thaw and drain before use.

Glaciolacustrine deposits constitute the main source of fine grained material for use as impermeable cores in dams and berms or for mixing with gravel for road surfacing. Large supplies are available in west Aishihik Valley and the lower reaches of Aishihik Valley. Smaller deposits occur elsewhere within the former extent of glacial Lake Sekulmun-Aishihik. Glaciolacustrine sediments also occur within some but not all of the sites of other former glacial lakes. Although the floor of Nordenskiold Valley below Kirkland Creek is mapped as silty alluvial floodplain with thermokarst topography (siApk), glaciolacustrine silt and clay may occur at a depth of 1.5 to 2 m beneath alluvial cover.

Permafrost is widespread in the glaciolacustrine sediments of the region, and segregated ice up to 40% or more by volume is common (Hughes and van Everdingen, 1978, p. 6; Klassen, 1979). The high ice content would place major practical and environmental constraints on extraction of the deposits.

Location of facilities

Surficial deposits of the map area vary widely in suitability for location of facilities, from the glaciofluvial deposits (best) to the thermokarst glaciolacustrine and alluvial deposits (worst). The glaciofluvial deposits are mainly free of permafrost or if perennially frozen, are stable when thawed. The larger areas of glaciofluvial plain and terrace (Gp, Gt) afford potential sites for facilities such as airfields and industrial developments that require large areas of stable soils. In the extensive areas of ridged, hummocky, or pitted glaciofluvial deposits (Gr, Gh, Gx) topography would impose significant constraints on most types of construction.

Although there is little experience on which to base judgment, it is likely that the larger areas of glaciofluvial deposits would yield substantial groundwater supplies from wells a few metres or tens of metres deep. The high permeability of the deposits would permit use of septic tanks with field drains, but at the same time would pose the problem of contamination of water supplies by infiltration of polluted surface water.

Extensive areas of ridged and/or hummocky moraine (Mr, Mh, Mx) appear to be free of permafrost and suitable for location of roads and other facilities, but with inherent topographic constraints. Areas of moraine blanket on moderate to steep south-facing slopes are commonly free of permafrost, but with the probability of permafrost increasing with elevation. On gentle slopes, on valley floors, and on north-facing slopes, moraine blanket should be treated as permanently frozen and potentially unstable when thawed until site investigations prove otherwise. The limited areas of rolling moraine (Mm) and drumlinoid till plain (Md) should be treated in the same way. The distribution of permafrost in moraine veneer (Mv) with respect to slope aspect and elevation is the same as for moraine blanket. Problems related to thaw instability should be less because the deposits are thinner. It should be noted, however, that the distinction between moraine blanket and veneer from airphotos is imprecise. Locally thick morainic deposits may exist in any area mapped as moraine veneer.

Despite the mountainous character of the region, most of the major streams have low gradients, hence the floodplains and low terraces (Ap, Apk) are mostly silty and poorly drained and commonly include extensive organic deposits. Permafrost and soils with high ice content are prevalent throughout floodplains having thermokarst ponds (Apk) and may also be common in those lacking such ponds. Some alluvial fans, for example on the east side of Nordenskiold Valley in the northeastern corner of Map 23-1987, are free of permafrost in their upper parts where the sediments are coarse, but permafrost with segregated ground ice occurs in fine grained sediments at the outer periphery of those fans. In unglaciated parts of the map area (Maps 22, 23-1987), valley sides are characteristically mantled by colluvial deposits (Cb) that grade imperceptibly toward the valley axis into fan aprons comprising mainly organic silt (Af). Few data are available on distribution of permafrost and ground ice on such slopes (mainly mapped as a complex, Cb/Af). Slopes of similar appearance in Klondike district are known from placer mining openings to be mainly perennially frozen, and high ice contents are common in the organic silts of the lower slopes.

In general, the alluvial deposits should be avoided as far as possible because of poor drainage, potential flooding, and the possibility of subsidence due to degradation of permafrost.

The glaciolacustrine sediments of the area are mainly perennially frozen, but with a wide range in segregated ground ice content. Sediments deposited in the glacial lake at site 9 (Table 3; see *Glacial lakes*) contain segregated ice up to about 40% by volume. There, thawing and subsidence of ice-rich sediments has produced thermokarst ponds, but extensive areas of glaciolacustrine sediments, notably along West Aishihik and Aishihik rivers, lack thermokarst features, suggesting a much lower ice content. Gullied silt bluffs along the northeast side of West Aishihik River, which are now nonfrozen for an unknown but considerable distance behind the bluff faces, retain slopes up to 70°; it seems unlikely that the silt contained much segregated ice when in the frozen state. This conclusion is supported by borings made along Aishihik River prior to construction of the generating station (Map 21-1987). In most of the borings the silt was frozen and poorly to well bonded, but contained only occasional inclusions of visible ice. A possible explanation of the low ice content is that Aishihik and West Aishihik rivers were incised rapidly through the sediments after drainage of glacial Lake Sekulmun-Aishihik, permitting drainage of the sediments prior to establishment of permafrost. The high ice content of site 9 is probably more typical of glaciolacustrine sediments of the area, and all occurrences should be treated as potentially ice-rich until proven otherwise. Regardless of ice content, glaciolacustrine sediments are highly susceptible to erosion. Construction should be planned to minimize removal of vegetation or concentration of drainage.

In the unglaciated parts of Yukon Territory, many pioneer roads were located on interconnecting ridges that could be followed for great distances with minimal road construction. The mostly treeless rounded ridges are typically covered by a thin mantle of rock detritus with only occasional outcrops of rock as tors or "castles", providing well drained and stable surfaces. The ridge routes obviated the need for bridges or culverts, avoided flood hazards, and perhaps most importantly, avoided the perennially frozen and commonly ice-rich sediments of the valley floors. For those reasons routes following interconnecting ridges would be preferable to valley routes for mining access roads in the unglaciated northern part of Aishihik Lake map area.

Whereas airfields and campsites can usually be located on stable surficial materials, most linear facilities such as roads, though carefully located, must cross intervals of perennially frozen and potentially troublesome terrain. The current practice is to cross such areas by advancing fill over the undisturbed, hand- cleared surface. Ideally, the fill should be thick enough that permafrost invades the base of the fill. If areas of glaciolacustrine and alluvial deposits with thermokarst ponds must be crossed, alignment should remain as far as possible from pond margins, even though individual ponds may display no evidence of active subsidence. The factors that control thermokarst subsidence remain poorly understood; rapid enlargement of ponds sometimes occurs with no apparent cause, so that no reliable rules can be offered for construction in thermokarst areas.

Placer deposits

Placer gold has been produced from several creeks in the southwestern corner of the map area. The creeks, all tributaries of Jarvis River, rise in the southwestern Aishihik metamorphic belt (Tempelman-Kluit, 1974; Fig. 1). Ruby Creek, the most productive, rises in a narrow glaciated pass floored by glacial drift and colluvium. The middle reaches of the creek are incised through the drift and a few metres into bedrock, but in the lower reaches the stream gravels lie on thick glacial drift. The creeks of the region were staked during a rush in 1903, and according to Cairnes (1915, p. 17) Ruby Creek had yielded $6000 to $8000 (300 to 400 ounces) by 1914, almost all from the section of the creek that is incised into bedrock. Twelfth of July, McKinley and Dixie creeks also produced at least small quantities of gold. Creeks of the region have seen only sporadic activity since, except for Twelfth of July Creek. Production of 6210 ounces of gold ascribed to Fourth of July Creek for the period 1978-84 apparently came from this Twelfth of July tributary (Debicki and Gilbert, 1986, Table 9).

According to Cairnes (1915), the gravel of those creeks and the contained gold is all derived from stream reworking of glacial deposits, with gold being unevenly distributed in the gravels. Gold in the glacial deposits was presumably incorporated by glacial action from preglacial placers that in turn were derived from rocks of the metamorphic belt. Cairnes believed that gold placers might be preserved in channels beneath the glacial deposits, but warned of the large outlay of time and capital required to search for such channels, and the possibility that at any one point the old gravels may have been swept away and their gold contents dispersed. In discussing Ruby Creek, he noted that the section of the creek within the valley of Jarvis River held little promise because of the probability that glacier ice moving up the valley had scattered the preglacial gravels. The same reasoning would apply to lower sections of McKinley and Dixie creeks. It is important to emphasize that all the tributary valleys were also glaciated, though perhaps not as intensely as Jarvis Valley, so that search for fortuitously preserved preglacial gravels would be an extremely high-risk enterprise.

With the recent increase in the price of gold, there has been some staking on Ruby, Twelfth of July, and Fireworks creeks and Jarvis River, but with no significant actual mining reported. A placer mine was operated on an unnamed creek at a point about 3.2 km south of the east end of Buffalo Lake as recently as 1980. Gold values are reported to have been low. The creek is incised about 100 m into glacial drift, from which the gold was probably reconcentrated. Mining was also conducted in 1980 on a small tributary that enters Kirkland Creek from the east at 61°32.1'N. This creek is also incised into drift, from which any gold present would likely have been reconcentrated.

The potential for discovery of economic placers in the glaciated parts of Aishihik map area is low because, as in the case of the Jarvis River tributaries, whatever placers existed preglacially have been eroded and dispersed, or buried beneath thick drift cover. Placers are most likely to have survived glacial erosion in small valleys oriented normal to the main flow patterns of glacial ice. Shallow seismic equipment may prove effective and inexpensive for defining buried channels in such valleys, but expensive drilling will be required to prove the existence of economic placers. High mining costs can be expected where placers lie beneath thick drift, particularly if the drift consists of bouldery till.

Placer gold has not been reported from unglaciated parts of the area, although Bostock (1936, p. 51) noted placer mining equipment, but little evidence of mining, on the west fork of Schist Creek, a tributary of Nisling River that lies partly within the map area. The most productive placers in Yukon Territory occur within areas of schist bedrock, so that the schist terrain north of Nisling River would appear to be the most promising target within the unglaciated part of the area.

Geochemical exploration

Aishihik map area comprises a wide range of surficial geological materials that differ greatly in the degree to which they reflect the occurrence of potentially economic metallic deposits. Unlike most parts of Canada, where the surficial deposits are mainly fresh glacigenic materials of Late Wisconsinan age, the surficial materials of central Yukon range in age from pre-Quaternary to Late Wisconsinan. These deposits therefore have a remarkable range of weathering histories that greatly complicates the interpretation of geochemical data.

Of the widespread surficial materials, colluvium occurring on slopes on which colluvial processes are active provides the clearest geochemical response. An anomaly is likely to reflect a source immediately upslope. In the unglaciated area, however, colluvial deposits may be deeply weathered, with complete removal of metallic ions to a level below that usually sampled in geochemical surveys. Among the glacigenic materials, the glacial deposits (till) provide the clearest response; however admixture of far travelled material with local material can reduce a possible geochemical response greatly. An anomaly is likely to reflect a source immediately upstream in the direction of the latest ice movement. Tills of Reid age have brunisolic soils to depths of 90 cm or more (Tarnocai et al., 1985) from which metallic ions may be partly leached. Thin deposits of loess (wind-borne silt) are widespread on valley sides and uplands of the area, especially adjacent to the larger glaciofluvial plains. The silt of the loess has undergone glacial, fluvial, and then wind transport, and is a poor reflector of possible economic deposits. Geochemical samples should always be taken from beneath such a loess cover.

The glaciofluvial deposits have undergone glacial transport followed by fluvial transport. Only large, high grade metallic deposits are likely to be reflected by geochemical anomalies in these materials. Glaciolacustrine deposits have likewise undergone both glacial and fluvial transport and have comparable limitations for geochemical exploration. Among the fluvial deposits, only the alluvial sediments of small streams (mostly too small to delineate at the present map scale) are likely to yield significant anomalies. Anomalies in such sediments are likely to reflect anomalies in the surficial deposits traversed by the stream. For such anomalies, further sampling should be directed to areas lateral to and upstream from the original anomaly.

REFERENCES

Bostock, H.S.

1936: Carmacks district, Yukon; Geological Survey of Canada, Memoir 189, 67 p.

1948: Physiography of the Canadian Cordillera, with special reference to the area north of the fifty-fifth parallel; Geological Survey of Canada, Memoir 247, 106 p.

1952: Geology of northwest Shakwak Valley, Yukon Territory; Geological Survey of Canada, Memoir 267, 54 p.

1966: Notes on glaciation in central Yukon Territory; Geological Survey of Canada, Paper 65-36, 18 p.

Cairnes, D.D.

1915: Exploration in southwestern Yukon; Geological Survey of Canada, Summary Report for 1914, p. 10-33; reprinted in Bostock (1957): Yukon Territory, Selected Field Reports of the Geological Survey of Canada, 1898 to 1933; Geological Survey of Canada, Memoir 284, p. 354-380.

Debicki, R.L. and Gilbert, G.W.

1986: Yukon Placer Mining Industry 1983-1984; Indian and Northern Affairs, Canada, Whitehorse, Yukon, 157 p.

Demek, J.
1969: Cryoplanation terraces: their geographical distribution, genesis and development; Ceskoslovenska Akademia Ved, Rozpravy, Rada Matematickych a Prirodnich Ved, Rocnik 79, Sesit 4.

Denton, G.H. and Stuiver, M.
1967: Late Pleistocene glacial stratigraphy and chronology, northeastern St. Elias Mountains, Yukon Territory, Canada; Geological Society of America, Bulletin, v. 78, p. 485-510.

Dyck, W., Lowdon, J.A., Fyles, J.G., and Blake, W., Jr.
1966: Geological Survey of Canada radiocarbon dates V; Geological Survey of Canada, Paper 66-48, 32 p.

Embleton, C. and King, C.A.M.
1975: Periglacial geomorphology; in Glacial and Periglacial Geomorphology, Second edition, volume 2; Edward Arnold (Publishers) Ltd., London, 203 p.

Foscolos, A.E., Rutter, N.W., and Hughes, O.L.
1977: The use of pedological studies in interpreting the Quaternary history of central Yukon; Geological Survey of Canada, Bulletin 271, 48 p.

Hamilton, T.D., Ager, T.A., and Robinson, S.W.
1983: Late Holocene ice wedges near Fairbanks, Alaska, U.S.A.: environmental setting and history of growth; Arctic and Alpine Research, v. 15, no. 2, p. 157-168.

Hamilton, T.D. and Bischoff, J.L.
1984: Uranium series dating of fossil bones from the Canyon Creek vertebrate locality in central Alaska; United States Geological Survey, Circular 939, p. 26-29.

Hughes, O.L.
1967: Surficial geology studies, Aishihik Lake map-area; in Report of Activities, Part A, Geological Survey of Canada, Paper 67-1, Part A, p. 48-49.
1969: Distribution of open-system pingos in central Yukon Territory with respect to glacial limits; Geological Survey of Canada, Paper 69-34, 8 p.
1983a: Surficial geology and geomorphology, Big Kalzas Lake, Yukon Territory; Geological Survey of Canada, Map 2-1982.
1983b: Surficial geology and geomorphology, Grey Hunter Peak, Yukon Territory; Geological Survey of Canada, Map 3-1982.
1983c: Surficial geology and geomorphology, Janet Lake, Yukon Territory; Geological Survey of Canada, Map 4-1982.
1983d: Surficial geology and geomorphology, Mount Edwards, Yukon Territory; Geological Survey of Canada, Map 5-1982.

Hughes, O.L. and van Everdingen, R.O.
1978: Central Yukon-Alaska; Third International Conference on Permafrost (Edmonton, Alberta), Field trip No. 1.

Hughes, O.L., Campbell, R.B., Muller, J.E., and Wheeler, J.O.
1969: Glacial limits and flow patterns, Yukon Territory, south of 65° north latitude; Geological Survey of Canada, Paper 68-34, 9 p.

Hughes, O.L., Harington, C.R., Janssens, J.A., Matthews, J.V., Jr., Morlan, R.E., Rutter, N.W., and Schweger, C.E.
1981: Upper Pleistocene stratigraphy, paleoecology and archaeology of the northern Yukon interior, eastern Beringia.1. Bonnet Plume Basin; Arctic, v. 4, p. 329-365.

Hughes, O.L., Rampton, V.N., and Rutter, N.W.
1972: Quaternary geology and geomorphology, southern and central Yukon (northern Canada); XXIV International Geological Congress (Montreal), Guidebook, Field Excursion All, 58 p.

Hughes, O.L., van Everdingen, R.O., and Tarnocai, C.
1983: Regional setting-physiography and geology; in Northern Yukon Territory and Mackenzie Delta, ed. J.M. French and J.A. Heginbottom; Fourth International Conference on Permafrost (Fairbanks, Alaska), Guidebook 3, p. 5-34.

Klassen, R.W.
1978: Surficial geology, southern Yukon; maps at scale 1:100 000 with extended legend; Geological Survey of Canada, Open File 539.
1979: Thermokarst terrain near Whitehorse, Yukon Territory; in Report of Activities, Part A, Geological Survey of Canada, Paper 79-1A, p. 385-388.

Lowdon, J.A. and Blake, W. Jr.
1968: Geological Survey of Canada radiocarbon dates VII; Geological Survey of Canada, Paper 68-2, Part B, 39 p.
1970: Geological Survey of Canada radiocarbon dates IX; Geological Survey of Canada, Paper 70-2, Part B, 86 p.
1979: Geological Survey of Canada radiocarbon dates XIX; Geological Survey of Canada, Paper 79-7, 58 p.

Love, D.
1970: Subarctic and subalpine: where and what?; Arctic and Alpine Research, v. 2, no. 1, p. 63-73.

Mackay, J.R.
1963: The Mackenzie Delta area, N.W.T.; Canada Department of Mines and Technical Surveys, Geographical Branch, Memoir 8, 202 p.

Péwé, T.L.
1966: Ice-wedges in Alaska — classification, distribution and climatic significance; Proceedings, International Conference on Permafrost, 11-15 November, 1963, Lafayette, Indiana; National Academy of Sciences — National Research Council, Publication no. 1287, p. 76-81.

Price, L.W.
1971: Vegetation, microtopography, and depth of active layer on different exposures in subarctic alpine tundra; Ecology, v. 52, no. 4, p. 638-647.
1973: Rates of mass wasting in the Ruby Range, Yukon Territory; North American Contribution, Second International Permafrost Conference, National Academy of Sciences, Washington, D.C., p. 235-245.

Pullan, R.A.
1959: Tors; Scottish Geographical Magazine, v. 75, p. 51-55.

Rampton, V.N.
1969: Pleistocene geology of the Snag-Klutlan area, southwestern Yukon Territory, Canada; unpublished PhD dissertation, University of Minnesota, Minneapolis, Minnesota.
1971: Late Pleistocene glaciations of the Snag-Klutlan area, Yukon Territory; Arctic, v. 24, p. 277-300.
1981: Surficial geology and geomorphology, Taye Lake, Yukon Territory, Geological Survey of Canada, Map 14-1981, scale 1:100 000.
1982: Quaternary geology of the Yukon Coastal Plain; Geological Survey of Canada, Bulletin 317, 49 p.

Reger, R.D. and Péwé, T.L.
1976: Cryoplanation terraces: indicators of a permafrost environment; Quaternary Research, v. 6, no. 1, p. 99-109.

Schweger, C.E., and Janssens, J.A.P.
1980: Paleoecology of the Boutellier nonglacial interval, St. Elias Mountains, Yukon Territory, Canada; Arctic and Alpine Research, v. 12, no. 3, p. 309-317.

St-Onge, D.A.
1965: La géomorphologie de Île Ellef Ringnes, Territoires du Nord-Ouest, Canada; Direction de la Géographie, Ministère de Mines et des relevés techniques, Étude géographique no 38, Ottawa.

Tarnocai, C., Smith, S., and Hughes, O.L.
1985: Soil development on Quaternary deposits of various ages in central Yukon Territory; in Current Research, Part A, Geological Survey of Canada, Paper 85-1A, p. 229-238.

Tempelman-Kluit, D.J.
1974: Reconnaissance geology of Aishihik Lake, Snag and part of Stewart River map-areas, west-central Yukon; Geological Survey of Canada, Paper 73-41, 97 p., Map 17-1973.

Wahrhaftig, C.
1965: Physiographic divisions of Alaska; United States Geological Survey, Professional Paper 482, 50 p.

Washburn, A.L.
1979: Geocryology: A Survey of Periglacial Processes and Environments; Edward Arnold (Publishers) Ltd., London, 406 p.

Wheeler, J.O.
1961: Whitehorse map-area, Yukon Territory; Geological Survey of Canada, Memoir 312, 156 p.