GEOLOGICAL SURVEY OF CANADA
COMMISSION GÉOLOGIQUE DU CANADA

BULLETIN 331

RETREAT OF THE LAST ICE SHEETS IN NORTHEASTERN BRITISH COLUMBIA AND ADJACENT ALBERTA

W.H. Mathews

Energy, Mines and
Resources Canada

Énergie, Mines et
Ressources Canada

1980

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Available in Canada through

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Hull, Québec, Canada K1A 0S9

and from

Geological Survey of Canada
601 Booth Street
Ottawa, Canada K1A 0E8

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

Cat. No.	M42-331E	Canada:	$4.00
ISBN -- 0-660-10595-0		Other countries:	$4.80

Price subject to change without notice

Critical Reader

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Original manuscript submitted: 1977 - 01 - 11
Approved for publication: 1978 - 12 - 21

Preface

The area described contains the contact zone of the two major North American ice sheets and has been the centre of much scientific interest. This report provides information on the relationship of the ice sheets at the time of their maximum extent and during their retreat.

The interpretation of late glacial events will aid further mapping of the surficial deposits of the region and will be of benefit in understanding their genesis. These deposits play an important role in agricultural and forest productivity, and affect foundation conditions for roads, pipelines and buildings. The report may aid research in other fields such as archeology and zoology, by shedding light on the possible migration of man and animals during the critical period at the end of the ice age.

D.J. McLaren
Director General
Geological Survey of Canada

Ottawa, February 1979

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Abstract

The last ice sheets to have covered northeastern British Columbia and adjacent Alberta have left a clear record of their presence and movement in the existing landforms and surficial deposits: grooves and drumlins, meltwater channels, erratics and drift sheets, and traces of former ice dammed lakes. Aerial photograph and map interpretation, supplemented by limited data on surficial materials, has been used to reconstruct the history of the ice retreat.

Three glacier systems were involved: the Laurentide Ice Sheet originating over the Canadian Shield, the Cordilleran Ice Sheet originating in the interior of British Columbia, and a system of valley glaciers from the northern Rocky Mountains (between 57° and 59° N).

At the climax of glaciation all of the area was ice covered, except for high summits and ridges of the northern Rocky Mountains. The earliest exposure of land east of the mountain front between 56 and 60 N was close to the present-day Alaska Highway where relatively high-level lakes and meltwater channels were developed. A series of large lakes, dammed by Laurentide ice, then developed in Peace River valley, with water levels falling as retreating ice exposed lower outlets. The extent of these lakes has been interpreted from the distribution of shorelines, bottom deposits, and tributary and outlet channels. The disposition of Laurentide ice farther north is inferred from ice flow patterns, meltwater channels, and glaciological considerations. The contemporaneous positions of Cordilleran and Rocky Mountain ice fronts, in some stages, can be inferred from the elevations of lakes or of spillways controlled by Laurentide ice. Laurentide ice withdrew more rapidly from the area than did Cordilleran ice, and a late local ice advance from the northern Rocky Mountains is indicated.

Most of the stages of retreat discussed date from about 13 500 to 10 000 years B.P., but the earliest stages may relate to an earlier glaciation.

Résumé

Les dernières nappes glaciaires qui ont recouvert le nord-est de la Colombie-Britannique et les terrains proches situés en Alberta ont laissé des traces évidentes de leur existence et de leur passage sur les formes de relief et les dépôts de surface actuels: cannelures et drumlins, chenaux proglaciaires, blocs erratiques et épandage de drift, et enfin vestiges d'anciens lacs de barrage glaciaire. Pour reconstituer l'historique de la ségression glaciaire, on a fait appel à la photointerprétation et à l'interprétation de cartes classiques, en même temps qu'aux données limitées fournies par l'étude des matériaux de surface.

On doit considérer trois grands ensemble glaciaires: la nappe glaciaire Laurentide issue du Bouclier canadien, la nappe glaciaire de la Cordillère qui a pris naissance dans l'intérieur des terres en Colombie-Britannique, et un réseau de glaciers de type alpin formés dans le nord des montagnes Rocheuses (entre 57° et 59° N).

A l'apogée de la glaciation, toute la région s'est trouvée recouverte par les glaciers, excepté les sommets des plus hautes crêtes des montagnes Rocheuses septentrionales. A l'est du front montagneux, entre 56 et 60 N, les toutes premières zones dégagées sont proches du secteur actuel de la grand-route de l'Alaska, où se sont formés des lacs glaciaires et des chenaux proglaciaires à assez grande altitude. Une série de grands lacs glaciaires, dont les eaux avaient été retenues par les glaces de la nappe Laurentide, se sont alors constitués dans la vallée de la rivière de la Paix (Peace River); leur niveau s'est abaissé au fur à mesure que le recul des glaces a libéré des déversoirs. L'étude de la distribution des lignes de rivage, de celle des sédiments lacustres profonds et des déversoirs principaux et secondaires a permis d'évaluer l'étendue qu'ont pu atteindre ces lacs. Plus au nord, pour connaître quelle a été l'extension de la nappe Laurentide, on a dû considérer les traces de l'écoulement des glaces, les chenaux proglaciaires et diverses autres particularités glaciaires. On peut sans doute aussi estimer la position du front glaciaire de la Cordillère et des montagnes Rocheuses au cours de certains stades glaciaires d'après l'altitude d'anciens lacs ou déversoirs, dont l'existence dépendait de la situation de la nappe Laurentide. Cette nappe s'est retirée beaucoup plus rapidement de la région que la nappe de la Cordillère, et il semble qu'il y ait localement eu une avancée tardive des glaces à partir des montagnes Rocheuses septentrionales.

La plupart des phases de retrait dont il est question se sont produites pendant un intervalle de temps daté à 13 500 - 10 000 ans, mais les phases les plus anciennes sont peut-être en relation avec une des plus anciennes glaciations.

RETREAT OF THE LAST ICE SHEETS IN NORTHEASTERN BRITISH COLUMBIA AND ADJACENT ALBERTA

INTRODUCTION

Information on landforms and surficial deposits left by the last ice sheets in northeastern British Columbia and adjacent Alberta (Fig. 1) is now sufficient to warrant an analysis of the history of ice retreat in more detail than the broad studies by Prest (1969, 1970) and Bryson et al. (1969). This paper is designed to complement one by St-Onge (1972) for central Alberta.

The ice sheets have left a clear record of their presence and movement in the existing topography. The meltwater streams flowing from these sheets during their retreat likewise have left widespread marks. The distribution of former ice dammed lakes is less clearly defined in the topography, but a record of their former presence is found in the surficial deposits. Although the mapping of these deposits as yet is limited, what data exist can be combined with the more complete geomorphic record to permit a reasonable interpretation of the pattern of movement of the last ice sheets and the configuration of their margins during their retreat. This interpretation, in turn, will aid further field mapping of the surficial deposits by indicating, for example, the sites and directions of drainage if not the precise extent of former ice dammed lakes. It should aid, too, the understanding of the origin of these deposits, which play an important role in determining agricultural and forest productivity and which determine foundation conditions for roads, bridges, pipelines, and buildings. Finally it may shed light on the possibilities of migration for animals and man during a critical period at the end of the last ice age.

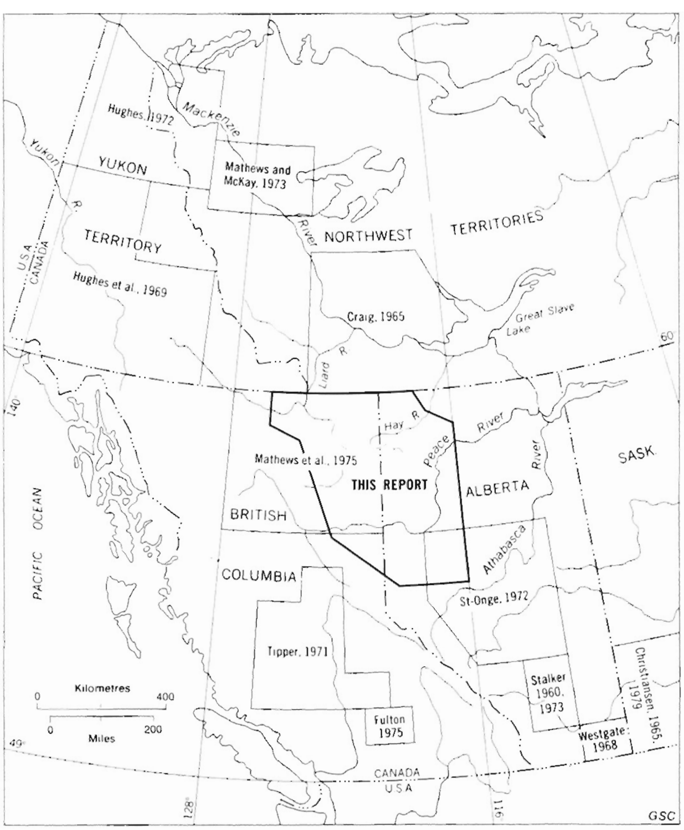


Figure 1. Index map showing location of study area and of other regional studies in western Canada.

The present study is an extension of field investigations conducted in the Charlie Lake map area, some 13 560 km^2 (5300 square miles) near Fort St. John, British Columbia (Mathews, 1978), supplemented by reconnaissance surveys along near the Alaska Highway northward as far as Liard River (mile 495, Alaska Highway). An examination of aerial photographs of British Columbia northeast of the Rocky Mountains and of pertinent areas in adjacent Alberta was undertaken in the National Air Photo Library, Ottawa, during the winter of 1971-72 and was completed in Vancouver in the fall of 1973. Additional information has been gleaned from reports and maps, published and unpublished, of the Soil Surveys of British Columbia and Alberta (Lindsay et al., 1963), and from what little has been published by geologists concerning surficial materials.

Part of the record of the former glaciers, meltwater streams, etc. has been incorporated with data from farther west for the preparation of a glacial map of Beatton River map area (Mathews et al., 1975) (Fig. 2).

Acknowledgments

Field work during the summers of 1971 and 1972 was supported by the Geological Survey of Canada, and office facilities were provided by this organization in Ottawa during the winter of 1971-72. Additional support has been provided by the Killam Foundation and by the Department of Geological Sciences, University of British Columbia. Special help in the field was provided by Mr. A. Anderson, Lands Inspector, then at Fort St. John, by Messrs T. Lord and A. Green of the Canada Soil Survey, and by the late A.J. Luckhurst, British Columbia Department of Agriculture. Messrs J. Calvert and P. Pintus ably assisted in the field. Critical comments on the report have been kindly offered by Dr. R.J. Fulton.

The Three Glacier Systems

Three glacier systems extended into northeastern British Columbia at the Wisconsinan climax: the Laurentide Ice Sheet from the east, the Cordilleran Ice Sheet from the distant west, and a local system of coalescent valley glaciers from the northern Rocky Mountains directly west (Fig. 2, 3). This latter system hitherto has been regarded as part of the Cordilleran Ice Sheet, but in terms of both ice movement and lithology of its depositional products it can be distinguished.

Laurentide deposits are recognizable by their significant content of erratics from the Canadian Shield, notably of red granite and red gneiss. Very rare occurrences of erratics from the Shield, however, are found in till and outwash derived from Cordilleran ice. It is suspected that these previously had been rafted westerly across proglacial lakes from the Laurentide Ice Sheet to the Rocky Mountain front and locally into the mountain valleys (Beach and Spivak, 1943, p. 375). Conceivably, too, some of the erratics of eastern provenance may have been derived from an undiscovered till laid down by an early Laurentide ice sheet which succeeded in reaching farther west than did

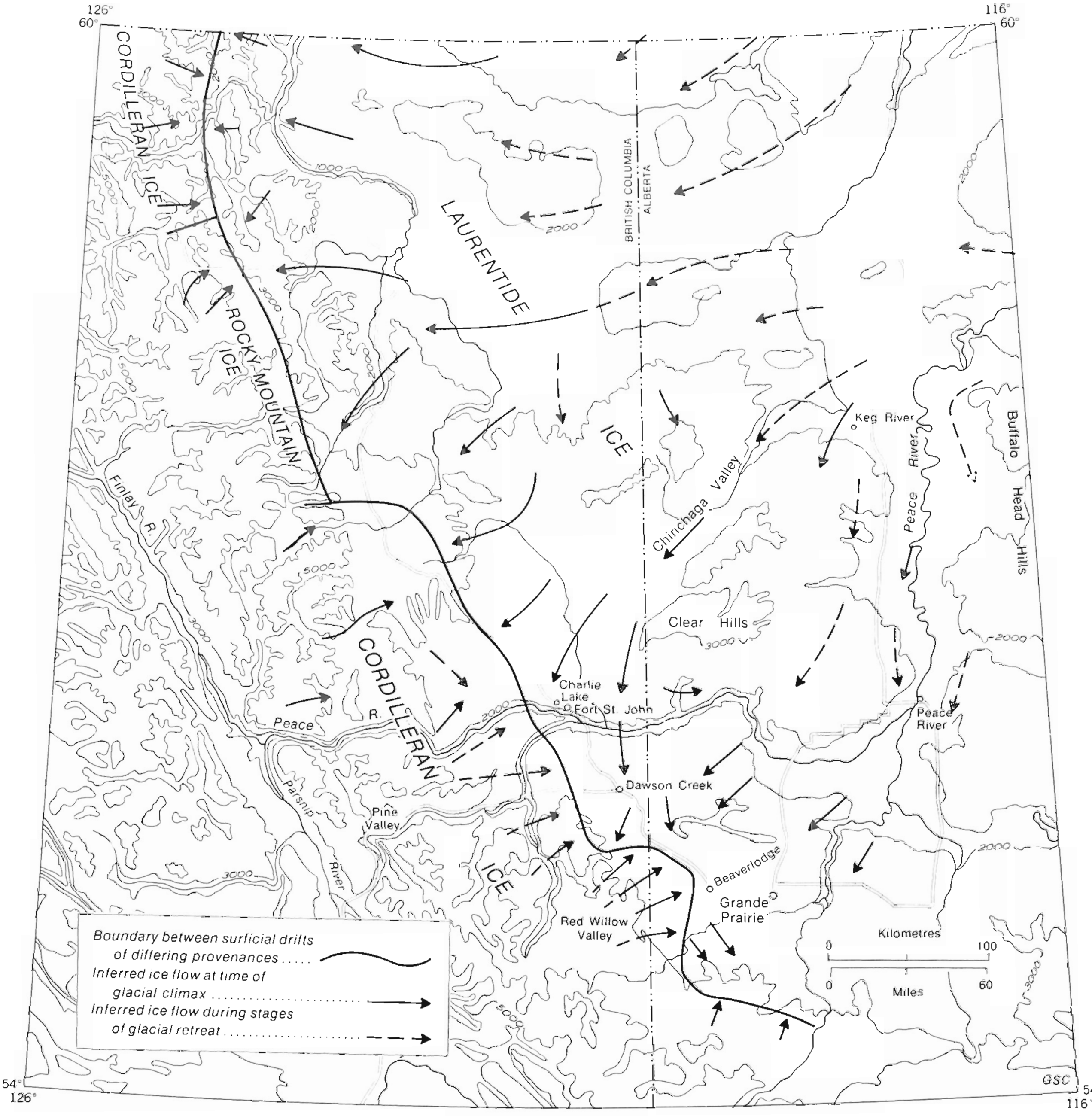


Figure 3. *Pattern of ice flow.*

Wisconsinan Laurentide ice (e.g. Roed, 1975, p. 1501). A component of erratics from the Canadian Shield that is significantly higher than any found in Cordilleran drift, together with a lack of erratics diagnostic of Cordilleran drift, is required to identify drift of Laurentide origin.

Cordilleran deposits contain, among other rock types, grey granitic fragments from plutons west of the Rocky Mountain Trench, together with quartzite pebbles and boulders from the vicinity of the Trench itself. The latter erratics are not, however, unique to Cordilleran deposits; significant amounts of quartzite occur in Laurentide till, apparently derived from preglacial or interglacial fluvial deposits such as are widespread on the plateaus of northwestern Alberta (C. Hage, personal communication, 1951; Green and Mellon, 1962; Vonhof, 1967; Roed, 1975, p. 1496-7). More diagnostic of a western source for the till are small pebbles and flakes of low grade schist and of slate, derived from Hadrynian beds near the Rocky Mountain Trench. The schist and slate are sufficiently friable in a fluvial environment that probably nowhere do they occur in pebble size in the preglacial or interglacial gravels east of the Rocky Mountains.

The deposits of the glacier complex of the northern Rocky Mountains lack the plutonic rocks from either the Cordilleran or Shield source but bear, instead, a local assemblage of sedimentary rocks from the Precambrian to Triassic succession (Taylor and Scott, 1973) together with minor diabase. Chert, quartzite, and carbonate rocks are locally abundant.

Outer Ice Limits

The boundaries between the tills from these three sources at or immediately following the climax of Wisconsinan glaciation are indicated in Figure 2. These have been located for the most part by pebble counts in fresh cuts in surface till along roadsides. The western limit of Laurentide drift has been placed where counts of red granite or gneiss drop off abruptly westward to a concentration at which no more than one or two clasts can be located during a careful, one minute search. The eastern limit of Cordilleran ice similarly has been placed where counts of schist or slate fragments drop abruptly eastward to nil. Rarely do the two limits, as identified by these criteria, differ by more than 8 km in any one transect.

Along the north bank of Peace River numerous stratigraphic sections show a single and continuous near-surface till sheet attributable to Wisconsinan glaciation. This one sheet has Cordilleran lithology in the west and Laurentide lithology in the east. It seems likely that contemporaneous ice sheets from west and east coalesced when till deposition was underway, during or shortly after the climax of glaciation. Minor changes in the regime of the two ice sheets at this stage could have caused shifts in the line of junction, thereby accounting for the lack of a sharp boundary here between their till deposits (Mathews, 1978, p. 9).

Rocky Mountain ice and Laurentide ice seem to have been contemporaneous in the vicinity of the Alaska Highway west of Fort Nelson. Thus in North Tetsa Valley (near mile 390, Alaska Highway) meltwater channels formed along the edge of Rocky Mountain ice have been graded to high base levels, believed to have been established by a Laurentide ice dam only a few kilometres to the east. Other high-level meltwater diversions farther south in the eastern part of the Rocky Mountains seem to have been caused either by Laurentide ice or by Rocky Mountain glaciers which had been backed up to great depth by Laurentide ice.

The drift limits, as mapped, are not reconcilable everywhere with contemporaneous ice movement from east and west. The abrupt easterly salient of Cordilleran drift near Sikanni Chief River, for example, can best be interpreted as a product of a late advance of Cordilleran ice into an area which had been occupied earlier by Laurentide ice.

The upper limits of the ice are poorly known since these limits are determinable only where mountain tops have protruded as nunataks. Few investigations have been undertaken within the mountains in pursuit of the critical evidence. The problem is compounded by the difficulty in distinguishing late Wisconsinan ice limits from possible earlier and higher limits.

Sentinel Peak (54°54'N, 121°57'W) is reported to bear pitted and angular quartz erratics (Reimchen, personal communication, 1975) on its summit at 2515 m (8250 ft). Reimchen and Rutter (1972) suggested that these erratics may be contemporaneous with an early Wisconsinan or Illinoian (?) drift in the central part of the Dawson Creek area, 100 km to the northeast. By implication, Wisconsinan ice may have failed to overtop Sentinel Peak and remove the erratics.

Mount St. George (58°36'N, 124°40'W) bears erratics to within a few metres of its summit. Siltstone erratics at 2130 and 2165 m (7000 and 7100 ft) above sea level show a weathered surface with microrelief and staining about 2 to 3 mm in depth; it is possible that this could be a product of post-Wisconsinan processes. At an elevation of about 2060 m (6750 ft) a diabase erratic shows scarcely 1 mm of staining, comparable in depth to that of similar erratics found at much lower levels which may be ascribed to Wisconsinan glaciation. Thus although no clear-cut late Wisconsinan limit can be identified here, it seems to have been above the 2060 m (6750 ft) level.

Elsewhere in the northern Rocky Mountains, erratics, glacial grooves, meltwater channels, and rounded ridge crests visible in aerial photographs occur up to elevations of from 1770 to 2010 m (5800 to 6600 ft) (Fig. 2) and mark minimum levels reached by the ice probably, but not necessarily, in Wisconsinan time.

The ridge north of Mount St. Paul (about 58°42'N, 124°45'W) bears several large erratics of conglomerate at about 1675 m (5500 ft) elevation on both west and east slopes (G.T. Taylor, personal communication, 1972). These have been ascribed to a source in the Lower Cambrian Atan Group some 50 km to the west, near the head of a valley whose floor lies close to 1070 m (3500 ft) elevation. If Pleistocene glacier profiles in mountainous areas (Mathews, 1974), such as those of the Puget Sound-Georgia Strait area, are applied here, the ice surface at the source would have been 300 m (1000 ft) higher than where the erratics were being deposited.

Times of Ice Retreat

The stages of retreat discussed below refer to the last ice to have occupied the sites in question as recorded by the uppermost till, by clearly defined meltwater channels, or by glaciolacustrine beds at the surface. It had been assumed, when preparing this report, in part because of the freshness of the record and the lack of deep weathering, that these were of late or 'Classical' Wisconsinan age. The radiocarbon record (see section on Geochronology) generally supports this conclusion. Thus the earliest dates from lake or pond deposits above the uppermost till in the Swan Hills area of Alberta (St-Onge, 1972) and in the Dawson Creek area fall in the interval 9960 to 13 580 years.

Two sites, however, now give discordant radiocarbon dates. High in the Saddle Hills of Alberta (55°31'N, 119°25'W) a lakebottom core has yielded dates of >30 000 years and 17 570 years, with no clear sign of any later interruption by glaciation (White et al., 1979). Redating of a tusk from the Portage Mountain moraine (see section on Geochronology) has, moreover, given a date of 25 800 years in place of earlier dates of 11 600 and 7670 years B.P. Although other explanations may account for these discordant dates (e.g., contamination by preservative in the tusk or redeposition of the tusk), a distinct possibility must now be entertained that the outermost and highest areas escaped late Wisconsinan glaciation and hence displays marks of mid or early Wisconsinan glaciation. If this is the case, the limits of Classical Wisconsinan ice have still to be located within this region.

DATA AND INTERPRETATION

Ice Flow Patterns

Patterns of drumlinoids and glacial grooves are widespread but not universal in northeastern British Columbia (Fig. 2). Gaps between the patterns can exist where the grooved terrain is covered by younger materials, such as ablation debris or lake beds; in other places, for example in the hilly area up to 30 km east of the Rocky Mountain front, grooving of the terrain may never have occurred. It is tempting to look upon the assemblage of patterns as recording ice movement at some instant in time, perhaps at the climax of glaciation, but experience suggests that this assumption is unjustified. The pattern at each locality should be looked upon, instead, as registering the local ice flow at times of last active movement – earlier near the ice limits and on upper parts of plateaus than nearer the ice sources or in valley bottoms (Fig. 3).

At any time ice sheets can be expected to slope from source to terminus essentially in the direction of flow. Junctions of two ice sheets should be marked by a trough on the ice surface. Except locally where the two ice masses impinge head on, the axis of the trough should plunge in the common direction of ice flow of the two adjoining ice sheets. As the moving ice of each sheet approaches the mutual boundary, their directions of flow should swing into parallelism and towards this common flow direction.

Sharp discordances between adjacent patterns of grooves can be taken as a sign of noncontemporaneity of the ice movements responsible for their development. Two striking examples of this relationship are found west of Beaverlodge and Dawson Creek (Fig. 3) where two lobes of Cordilleran ice, each marked by a striking fan-shaped pattern of grooves and drumlinoids, abruptly truncate an earlier southerly to southeasterly trending pattern of grooves attributable to Laurentide ice. The fact that the outer limits of the fan patterns are essentially normal to the local directions of Cordilleran ice flow indicates, moreover, that this flow was unobstructed by any contemporaneous Laurentide ice.

Topography has had a strong influence on ice flow patterns. Thus two major gaps in the eastern Rocky Mountains, the Pine and Redwillow valleys, seem to have funnelled the flow of Cordilleran ice to the apices of the two lobes described above. Similar ice lobes marked by fan-shaped patterns of grooves occur farther south at the mouth of Athabasca Valley (Roed, 1975) and North Saskatchewan Valley (Boydell et al., 1974) where again Cordilleran ice flowed in quantity from the mountains out onto the plains.

The influence of topography also is expressed clearly in the flow pattern of Laurentide ice which moved generally westerly and southerly up the main valleys. Noteworthy is an abrupt bend in the indicated flow of ice in Peace River valley around the northwest corner of Buffalo Head Hills (Fig. 3). Significant, too, are the curving patterns at the southwestern and southeastern corners of the Clear Hills where ice flowed from both sides into a 'shadow area' in the lee of the hills. The strong influence of bottom topography on ice flow here implies that ice thickness at the time would have been little, if at all greater than the relief of the substratum, some 600 m (2000 ft). Independent evidence (Mathews, 1974, p. 40; Fig. 4) indicates that the surface slope of the ice was remarkably low, only about 90 m (300 ft) in 100 km at about this stage: hence it would not be unexpected for the top of the hills to have protruded above ice then streaming around them and perhaps coalescing on the lee side.

The flow pattern also is clearly influenced on a still larger scale by the disposition of source areas for the ice and the sites of its dissipation (Mackay, 1965). The general westerly to southwesterly flow of Laurentide ice evidently has been determined by a regional slope of the surface from the summit of an ice dome in the vicinity of Hudson Bay (Andrews and Barnett, 1972; Walcott, 1972; Paterson, 1972, Shilts et al., 1979). The southerly deflection of Laurentide ice near the Rocky Mountain front, e.g. near Dawson Creek, may represent influence of topography, the obstruction by easterly moving Cordilleran ice, the flow towards major ablation areas far to the southeast, or some combination of these effects. A similar swing of the ice flow pattern from west to north near Fort Simpson, Northwest Territories (Craig, 1965) clearly marks flow towards a major ablation area around lower Mackenzie River.

The ice flow patterns have aided in the reconstruction of ice front positions during stages of retreat. In this it is presumed that ice on the plains formed a series of lobes, with convex-up cross-profiles, which terminated on flat ground at right angles to the local direction of flow as shown by drumlins and grooves. Where an ice stream was confined to a broad valley between high plateaus it is further presumed that the ice lobe would be symmetrical, rising at any one time to similar heights on opposing points on the valley walls. The evidence of meltwater channels (see below) indicates that the forward component of ice slope along the valleysides was low; accordingly, the ice margins here no longer would have been at right angles, or even nearly so, to the local direction of flow.

A record of moraines, at least those with clear-cut topographic expression, is particularly meagre in northeastern British Columbia and adjacent Alberta. A curved ridge some 5 km long, concave to the west, crossing the Rocky Mountain Portage, 11 km west of Hudson Hope (see p. 11), obviously was developed at the front of an ice tongue extending easterly to the mouth of Peace River valley. Excavations into this ridge to obtain fill for the nearby W.A.C. Bennett dam, however, showed that it consists of well washed gravel and sand and should be considered a kame moraine. No end moraines formed by Laurentide ice have been identified in either field work or aerial photograph interpretation other than the 'Fish Creek moraine' described by Henderson (1959, p. 25) in the Sturgeon Lake area (55°20' to 55°25'N, 117°15' to 117°25'W). A series of transverse ridges in the Milligan Hills (57°20' to 57°25'N, 120°30' to 121°10'W) have been interpreted as ice-thrust features on the basis of reports

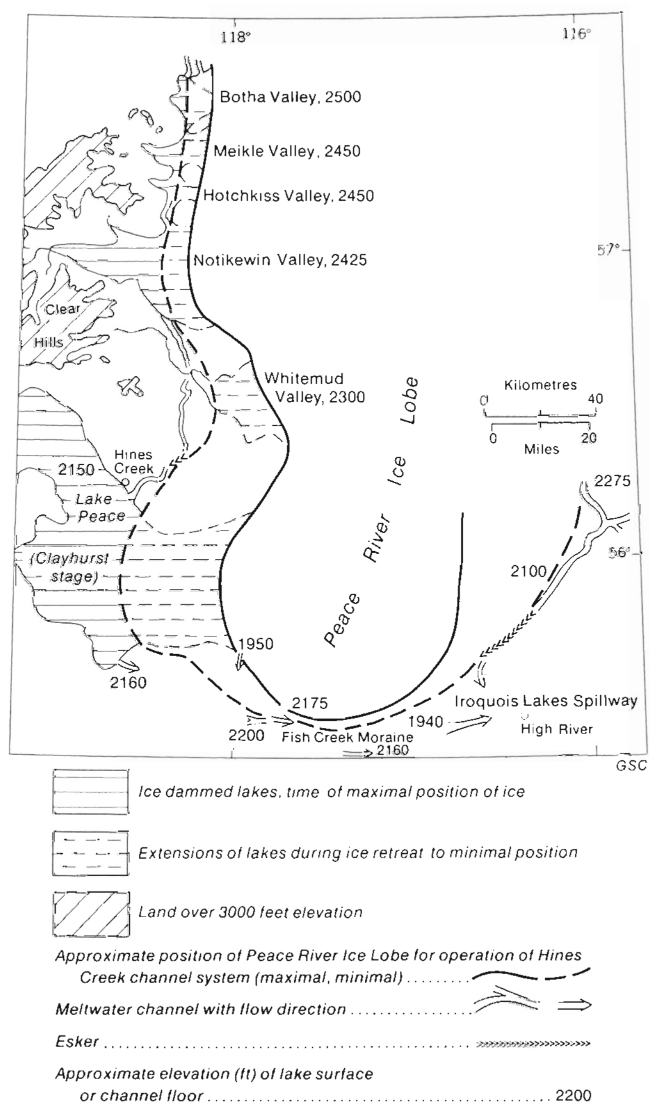


Figure 4. *Hines Creek channel system and contemporaneous ice front positions.*

(A. Green, personal communication, 1971) that they contain tilted shales. As a result of this limited topographic record of moraines, these features have not been used in the interpretation of ice flow or of ice front positions.

Provenance of drift may offer a further clue on ice flow within individual glacier systems, but as yet only a few examples can be cited from this area. Occurrence of oil-impregnated dolomite erratics at Pouce Coupe, British Columbia, has been attributed to a source near Pine Point on the south shore of Great Slave Lake, 650 km to the northeast (Gravenor and Bayrock, 1955). Boulders of pink to white quartzose sandstone attributable to the Athabasca Formation of northeastern Alberta and northwestern Saskatchewan are found in Laurentide drift of the Beaverlodge (Jones, 1961) and Fort St. John areas (Mathews, 1963, p. 8) and in the Charlie Lake map area east of the Doig-Beatton divide. The distribution of these erratics is in keeping with the general southwesterly flow of the Laurentide Ice Sheet here. Their northwesterly limit may correspond to a sinuous flow line extending back along Chinchaga (Fig. 3) and lower Peace valleys to Athabasca Lake, the northern limit of the Athabasca Formation.

Fluvioglacial Record

Glacial meltwater has left a particularly conspicuous topographic record in the form of channels up to 1.5 km wide (Ekwan Creek-Fontas River) and 90 m deep (Fontas River) and braided channel networks up to 8 km across (between Fort Nelson River and Klua Creek). These channels are associated with glaciofluvial sands or gravels (Valentine, 1971, p. 15; Mathews, 1978, p. 11). The large channels have flat or terraced floors and steep, parallel walls which can be traced in a more or less sinuous course for up to 160 km (Ekwan Creek to Fort Nelson). Some of the large channels are now occupied by major streams, e.g. Fort Nelson River, and here the floors of the channels have been incised or modified by the modern river. Smaller meltwater channels also have been modified by such processes as landsliding or talus development from their walls and partly have been infilled by alluvial fans from small tributaries. Many of these channels thereby have been converted into a chain of lakes or swamps. One diagnostic feature of the channels, large or small, is that any stream now flowing in them tends to be much narrower and commonly much more sinuous than the channel itself; in short these channels are occupied by 'misfit' or more specifically by 'underfit' streams.

A second characteristic of many channels is that they extend upslope maintaining their full width to a head at a low point in a drainage divide. These are the so-called 'direct overflows' (Kendall, 1902) or 'sadelskåror' (i.e. col gullies) (Mannerfelt, 1945), a product of streams fed directly by ice or by ice dammed lakes which spill across the low points in a divide to pour actively <u>down</u> the distal ice free slopes. Other channels course <u>across</u> a slope, nearly parallel to its contours, again maintaining full width from upper to lower terminus. Their locations would be inexplicable in the absence of a glacier; however, they can be ascribed to diversion of meltwater more or less along the contact between the bare ground, sloping towards the ice, and the ice itself, sloping in the opposite direction. These are the 'lateral overflows' of Kendall (1902) and the 'lateral skvalrannor' (lateral drainage channels) of Mannerfelt (1945).

Direction of flow for the long meltwater channels is obvious from the topography. For some of the short channels, where there may be but little change in elevation between head and lower end, the former direction of flow may be uncertain. In some examples where obstructions have been built near the terminus, the present flow direction may be the reverse of the original. The present elevation of the head of the channel in all probability is below that of the initial spill point by an amount determined by erosion during active discharge of meltwater. The lower end of the channel may have changed during active life, too, particularly with continued downslope withdrawal of the ice or with a drop in water level of the lake into which the channel drained. The lower end of many, but by no means all, channels at their terminal stages of activity may be recognized in aerial photographs as a deltaic accumulation which may or may not have been incised by a smaller stream surviving in the same channel.

The elevations of the head and lower ends of the channels may be estimated from topographic maps, providing these ends are reasonably well defined in aerial photographs. Ideally such elevations can be determined to somewhat less than one contour interval. In practice such confidence cannot be assigned everywhere because of lack of adequate vertical control or because of different optical models used in the photogrammetric mapping of adjacent map areas. For example, at 59°00'N, 124°00'W the floor of a former meltwater channel has been variously mapped as lying at from 1725 ± 25 to 1875 ± 25 feet (525 ± 8 to 570 ± 8 m) above sea level.

The correlation of successive meltwater channels developing simultaneously can aid enormously in interpreting contemporaneous ice front positions. Such correlations, however, must follow certain constraints: (1) any one channel can be linked in a downstream direction only with a second channel heading at an elevation no higher than that of the lower end of the first; (2) the second channel must be at least as large as the first; and (3) just as the first channel must have drained <u>towards</u> the second, so the second must have drained <u>away</u> from the first. Gaps between the first and second channel may mark an ice dammed lake, a place where meltwater was draining over or through the ice, or some combination of lake and ice. Where the gap was the site of a lake, the difference in elevation between the near ends of the channels should be minimal; such may or may not have been the case where water was flowing over or through ice depending on the dimensions of the watercourse. Where the gap was occupied by ice, the possibility exists for the development of one or more eskers.

Reconstruction of ice front positions from single or serial meltwater channels presumes that for any channel to be occupied by meltwater (1) this channel was not effectively blocked by ice anywhere downstream to an elevation above the channel floor and (2) all alternative drainage routes were ice dammed at that time. It is assumed, unless there is evidence to the contrary, that the ice was essentially impervious and sloped continuously towards its margin and hence that the passage of large volumes of meltwater through or over the ice, other than at its outermost margin, was precluded. This general rule is not true everywhere as is shown in the Petitot River basin of northeasternmost British Columbia by an alternation of channels and eskers for distances of 15 to 25 km.

A good example of the use of meltwater channels in interpreting ice front positions can be found east of Clear Hills, Alberta (Fig. 4). Here a channel system can be traced southward from upper Hotchkiss Valley at 750 m (2450 ft) elevation (if not from the north side of Botha Valley, 50 km farther north). A large gap, presumably once occupied by an ice marginal lake, occurs at Notikewin Valley, but south of this a channel can be traced for 70 km, with but one short interruption at the crossing of Whitemud Creek and another at an esker, to 655 (2150 ft) elevation near the town of Hines Creek. At this point the channel dies out on what is believed to be a delta on the shore of a large ice dammed lake in Peace River valley. Two possible outlets exist for this lake, at 658 and 663 m (2160 and 2175 ft) at points 65 km south and 100 km southeast of Hines Creek. The latter spillway is part of a system at the front of the Fish Creek ice lobe (Henderson, 1959, p. 34-35). Figure 4 illustrates the maximal and minimal positions of the ice lobe that would permit development of a system of linked meltwater channels as described here. Any advance significantly beyond the maximal position is likely to have blocked the system; any retreat beyond the minimal position would have opened lower outlets to the lakes linking the channel segments.

Expectable symmetry of the Peace River ice lobe during formation of the channels described above is confirmed by a similar channel-esker system, some 80 km long, on the east side of the lobe. The general southward flow of the ice lobe is, moreover, consistent with the pattern of grooves and drumlins developed on its bed. The low gradient of the meltwater system along the west side of the lobe, declining 90 m in 100 km, is indicative of a remarkably low surface slope to the ice lobe itself.

Henderson (1959, p. 75-76) suggested that this ice lobe had retreated sufficiently far to drain any ice dammed lake in Peace Valley before readvancing perhaps 240 km to a limit at the spillways 65 and 100 km southeast of Hines Creek. If so, it is likely that the meltwater systems have recorded the maximum stage of this readvance. The low gradient, moreover, may reflect a glacial surge rather than normal ice flow (Mathews, 1974).

Another ice marginal system of meltwater channels, likewise with a very low overall gradient, has been identified at the east end of Milligan Hills, heading at the British Columbia-Alberta border (57°28.5'N, 120°00'W) at 858 m (2815 ft) elevation and extending southwesterly for 26 km, declining approximately 50 m (165 ft) in this distance. This system may have continued southwesterly for another 22 km along a linear esker network at about 790 m (2600 ft) elevation. Here, too, the possibility of an ice surge may be offered as an explanation of the low ice-surface gradient (Mathews, 1974).

Glaciolacustrine Record

The best record of the presence of a former ice dammed lake is the slackwater deposit left on its floor. Even this record, however, presents its problems. The base of a glaciolacustrine deposit commonly contains scattered stones, in places so abundant that the material becomes almost indistinguishable from till deposited directly from ice. Only a vague stratification may serve to identify it as a lacustrine sediment. Presumably the stones have been dropped either from icebergs or, in some cases, from the ice front itself. At a greater distance from the ice front the proportion of stones can be expected to be less. In many sections the stony basal glaciolacustrine sediment gives way upward to less stony and more clearly stratified, even varved, material, presumably as the ice front withdrew from the site and the concentration of icebergs was reduced. It follows that the large, long-lived lakes should be identified readily by these thick distal deposits, but the small lakes, which are probably also short-lived, may be represented only by proximal deposits not easily distinguished from underlying till in the absence of fresh and clean exposures.

The glaciolacustrine deposits characteristically have a high content of silt or clay. Laurentide ice moved across large areas of dark grey Cretaceous shale and its erosion products are particularly rich in dark clay. The deposits in the deep parts of some lake basins, however, may be richer in silt, or even sand, than those of shallower water or of isolated embayments (Mathews, 1978, p. 10).

Identification of sites of former lakes by ground-based studies is relatively straightforward, although the presence of lake sediments does not establish the lake surface elevation, the precise distance to the nearest ice front, or the location of the lake outlet. The lateral distribution and variation of the sediment as seen on the ground may help to answer these questions although the possibility that the deposit is composite, representing a series of overlapping lakes, with successive levels and outlets, cannot be precluded.

Identification of the sites of former lakes on the basis of airphoto interpretation is difficult, although the thick clay deposits overlying massive till may be recognizable in actively retreating river-cut slopes by the dark tone of the clay deposit, by the gullied slopes breaking off into steeper and paler cliffs marking the underlying till, and by the high incidence of soil slumps. The generally featureless lake-bottom topography, surviving on the top of the clay or silt deposits, is suggestive but not diagnostic in distinguishing them, for example, from fluted or drumlinized till topography exposed where lake deposits are thin or absent. A general absence of meltwater channels over an extensive area is likewise suggestive of a former lake.

Shoreline features are commonly inconspicuous. On the ground a belt of gravelly soils following the contours commonly marks the former shore. A small proportion of these is sufficiently well developed to exhibit topographic form or to influence vegetation to the degree that they can be recognized through careful airphoto examination. In places, however, neither distinctive soils nor landforms are present, let alone visible in airphotos, where interpolation of data indicates lake shores should have been present. The best developed shorelines are found on moderate slopes and on rocky to gravelly substrates – sandstone, stony tills, and fluvioglacial deposits. Prolonged exposure to vigorous wind and wave action during the time that the ice dammed lake persisted at a particular level presumably aided in the development of the shore feature. Relatively few localities, however, met these ideal requirements and shorelines visible in airphotos are limited. The best examples are (1) near 610 m (2000 ft) elevation for a 50 km stretch northwest and southwest of the town of Peace River, Alberta, in an area identified as having a fluvioglacial substratum (Scheelar and Odynsky, 1968); (2) near 670 m (2200 ft) elevation north of Peace River intermittently for 40 km east and 11 km west of the British Columbia-Alberta boundary; and (3) at 335 m (1100 ft) elevation for about 70 km southeast, southwest, and northwest of Zama Lake, Alberta (this may record a lake persisting into postglacial time).

Abandoned deltas, where a meltwater channel entered a former lake, are only locally conspicuous in this area. Most meltwater channels have not produced clear-cut deltaic forms at their downstream terminations, possibly because the volume of coarse sediments discharged from them was too small or the slope on which this was deposited was too gentle. Dune fields, which to the southeast so commonly mark deltas (St-Onge, 1972), here are few in number and limited in size. Careful ground mapping, however, particularly at the lower end of large channels, undoubtedly will reveal many more deltas than have been identified so far.

A small but significant tilt to abandoned shorelines has been detected by detailed surveys in the Charlie Lake map area (Mathews, 1978, p. 12) and has been ascribed to differential glacioisostatic rebound. Data from the area show a component of rise to the west of about 0.4 m/km (2 ft/mile) and to the south of between 0 and 0.4 m/km 0 and 2 ft/mile). It is presumed that the tilt has been dominated by the response to unloading of late-melting Cordilleran ice to the west. Craig (1965, p. 74-75, his Fig. 2) indicated a general rise of shorelines in the Great Slave Lake area, Northwest Territories (Fig. 1) from west to east of about 0.5 m/km (2.5 ft/mile). Here the response seems to have been dominated by the easterly retreating Laurentide ice. Between these two areas tilting is presumed to diminish to values near zero. Corrections to relative elevations and slopes of meltwater channels because of this later isostatic tilting are considered to be negligible. On the other hand, for large ice dammed lakes, which once extended for up to 350 km in an east-west or northeast-southwest direction at right angles to the retreating ice fronts, the shore feature can be expected to show significant differences in elevation between their extremities.

Figure 5A-H. Position of ice front, lakes and channels during stages of retreat.

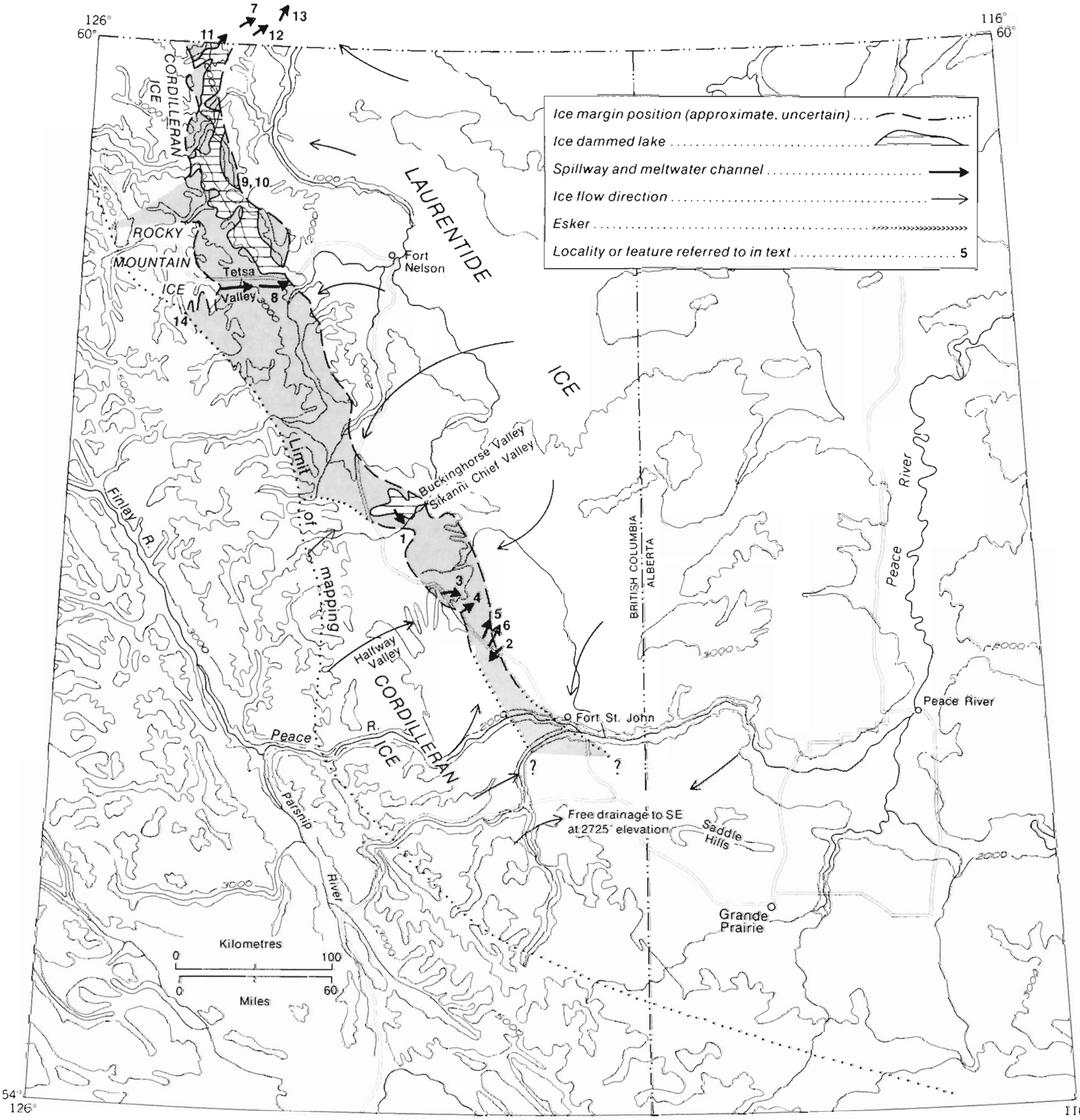


Figure 5A. Phase 1. Early ice diversions.

ICE RETREAT

Introduction

The following history of ice retreat is derived from the physical evidence, visible in aerial photographs and locally from ground studies, interpreted, as outlined above, in terms of ice front positions at various stages of withdrawal (Fig. 5A-H). The history is simplified, perhaps oversimplified, in that it shows only continual retreat of ice. Local and temporary readvances, such as the one reaching the Fish Creek moraine of Henderson (1959, p. 75-76), are difficult to evaluate even with careful ground studies. A corollary of the assumption of uninterrupted ice retreat is that the lakes in any one basin should become progressively lower with time. That this is not true everywhere in the region is shown by the occurrence of an abandoned delta near mile 92 on the Alaska Highway which had become completely covered by lacustrine sediment and was discovered only in a trench dug for a gas pipeline (Mathews, 1978, p. 11). Thus, this reconstruction is intended for use as a guide rather than as the final answer.

Phase 1 (Fig. 5A)

The highest and presumably the earliest signs of exposure of the land by Laurentide ice retreat are found along the Alaska Highway near mile 168 and west of mile 179, north of Sikanni Chief River. A veneer of lacustrine sediments occurs here up to 1130 m (3700 ft) elevation. No shoreline feature has been found and limited areas of land exist at or above this elevation in the vicinity, hence it is likely that the lake was bounded to a large extent by icy walls.

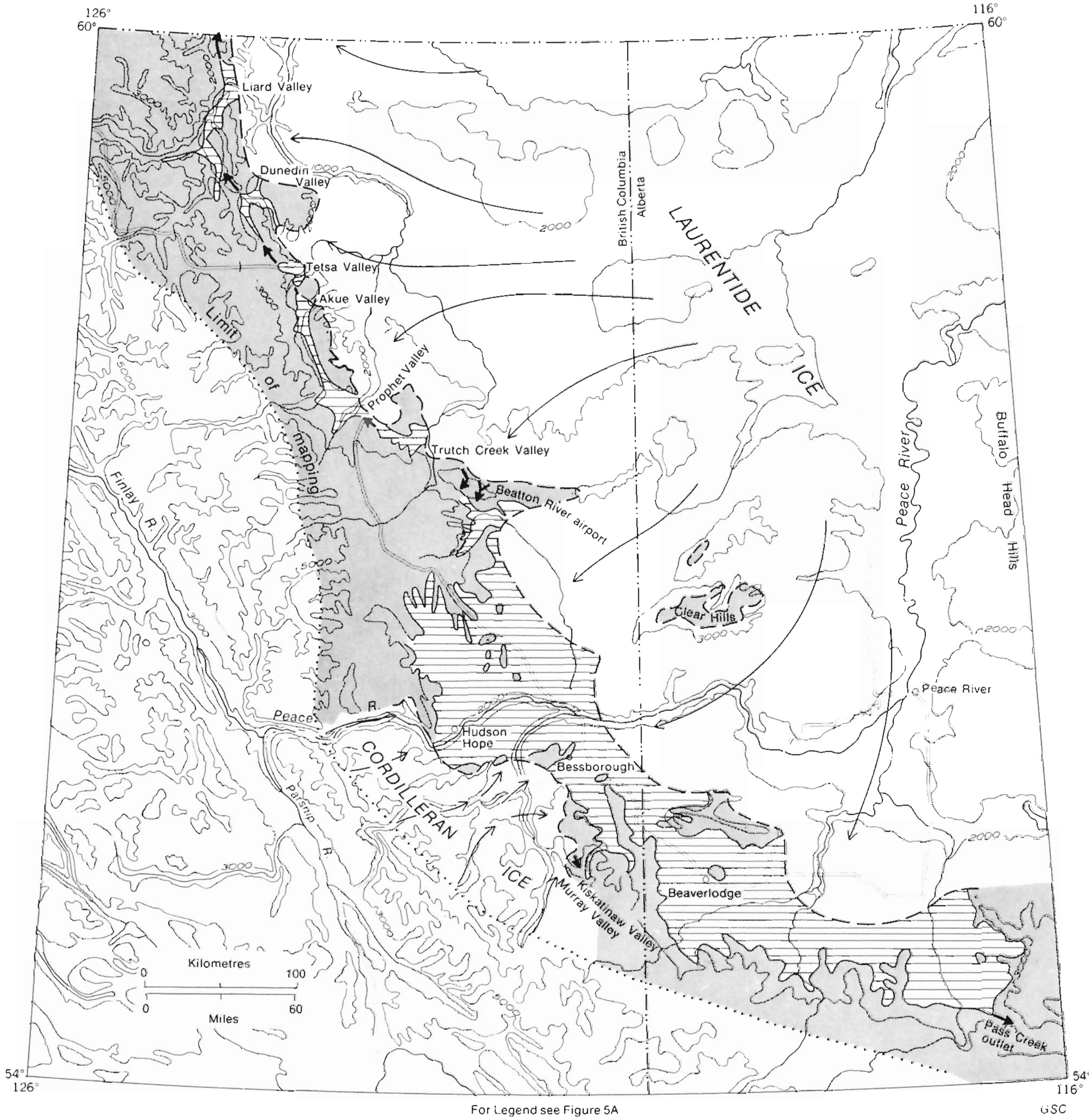


Figure 5B. *Phase 2. Glacial Lake Peace, Bessborough stage.*

A meltwater channel (1 of Fig. 5A) 10 km east of mile 174 and heading at 1020 m (3350 ft) a.s.l., apparently drained southeastward. For this spillway to have functioned the easterly sloping Buckinghorse Valley, a few kilometres to the north, would have had to be blocked at the time by Laurentide ice.

The abandoned delta (2 of Fig. 5A) near mile 92 (referred to above) is a record of early stream diversion westward from the Laurentide Ice Sheet into a temporary lake at 860 m (2825 ft) elevation. The submergence of this delta and its mantling by glaciolacustrine sediment probably took place shortly thereafter.

Two notches in the Halfway-Beatton divide, one at mile 122 on the Alaska Highway (3 of Fig. 5A) with a spillway elevation of about 970 m (3175 ft), the second at mile 112 (4 of Fig. 5A) with a spillway elevation of 910 m (2975 ft) apparently were occupied and enlarged by easterly draining meltwater and accordingly are ascribed to Cordilleran ice. Two other notches, at miles 94 and 97 (5 and 6 of Fig. 5A), both at 895 m (2775 ft) elevation, also may have been occupied and enlarged by meltwater; here, too, the Cordilleran ice, crowding close to the ridge crest, may have been responsible for the diversion.

Farther north, west of Fort Nelson, the earliest record of Laurentide ice retreat is a deposit of dark varved clay and silt up to at least 775 m (2540 ft) a.s.l. at mile 367 and to about 820 m (2700 ft) a.s.l. near mile 360. The spillway holding a lake to this level is not clearly identified. The nearest known meltwater channel close to this elevation lies about 755 m (2470 ft) a.s.l. on the east side of the

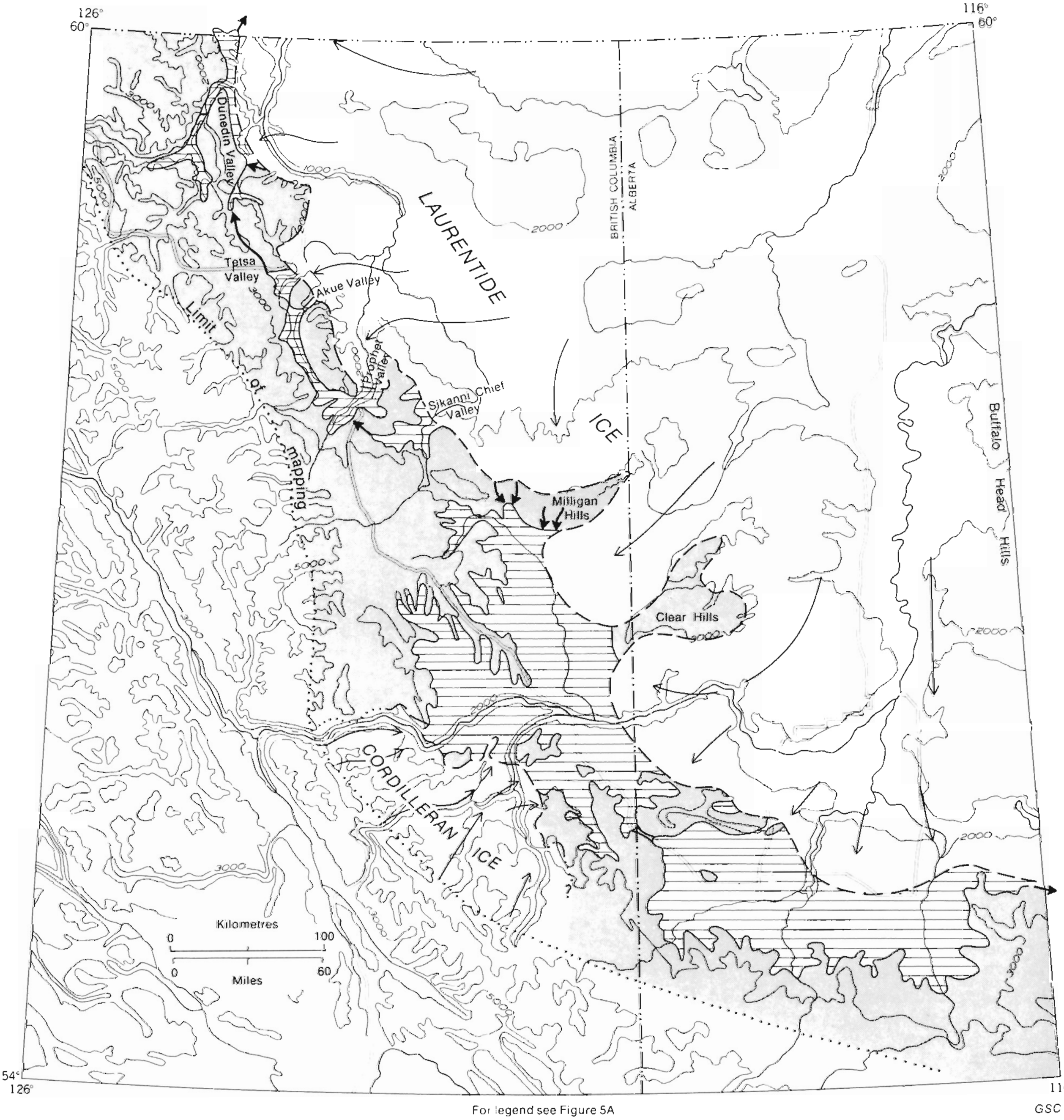


Figure 5C. *Phase 3. Glacial Lake Peace, unnamed stage.*

La Biche Range in Yukon Territory, 135 to 160 km to the north (7 of Fig. 5A). This feature is close to the outermost limit of the Laurentide Ice Sheet, although it is almost 600 m (2000 ft) below its uppermost recorded limit. If this is the spillway for the lake in the Fort Nelson area, it is likely that a proglacial lake extended along the lower valleys of Dunedin, Toad, Liard, and Beaver rivers, bounded on the east, in the Foothills belt, by Laurentide ice and on the west by Cordilleran ice. This spillway, in turn, may have drained northwards along the western bases of the Liard Mountains and Nahanni and Camsell ranges to Mackenzie River.

A younger meltwater channel (8 of Fig. 5A) transects the glaciolacustrine deposits along the Alaska Highway at mile 362. Starting at Tetsa River at 665 m (2190 ft) a.s.l., this channel evidently drained northward via the valley of Mill and McLennan creeks to Dunedin River, thence via gaps at 570 and 555 m (1875 and 1825 ft) (9 and 10 of Fig. 5A) to what was probably an ice dammed lake in lower Toad and Liard rivers. Presumably Liard River was still blocked by Laurentide ice near the eastern limit of the Foothills; if so, the meltwater may have escaped from this lake by a series of channels, first at 60°00'N, 124°45'W, elevation 520 m (1700 ft) to Beaver Valley (11 of Fig. 5A); thence at 60°07'N, 124°14'W (Fig. 2), elevation 460 m (1500 ft) to La Biche Valley (12 of Fig. 5A), and at 60°12'N, 124°03'W, elevation 460 m (1500 ft) to Kotaneelee Valley (13 of Fig. 5A). Beyond this point the water then may have flowed west of the front ranges of the Mackenzie Mountains or found an unrecorded route into or over Laurentide ice to lower Mackenzie River.

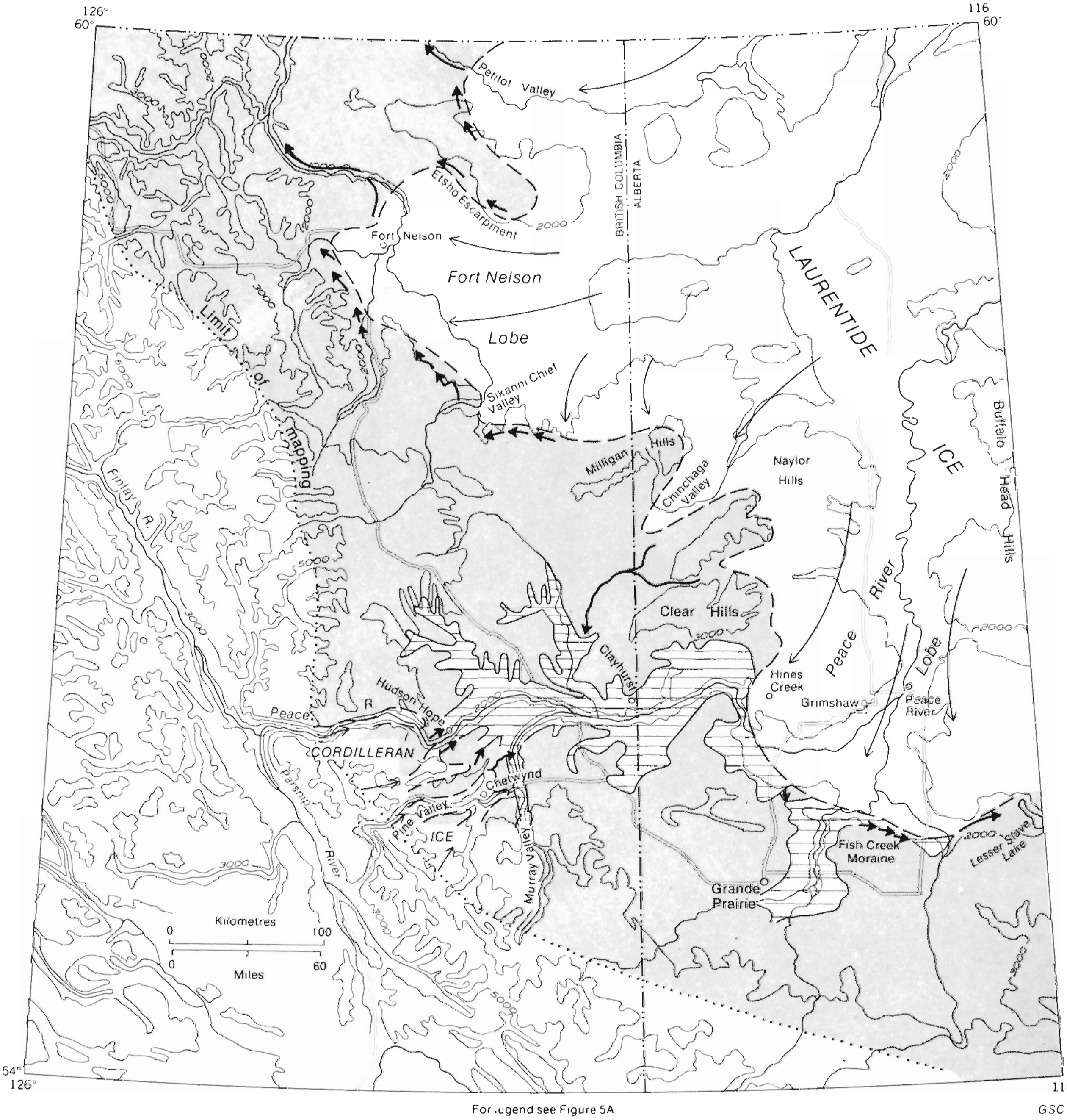
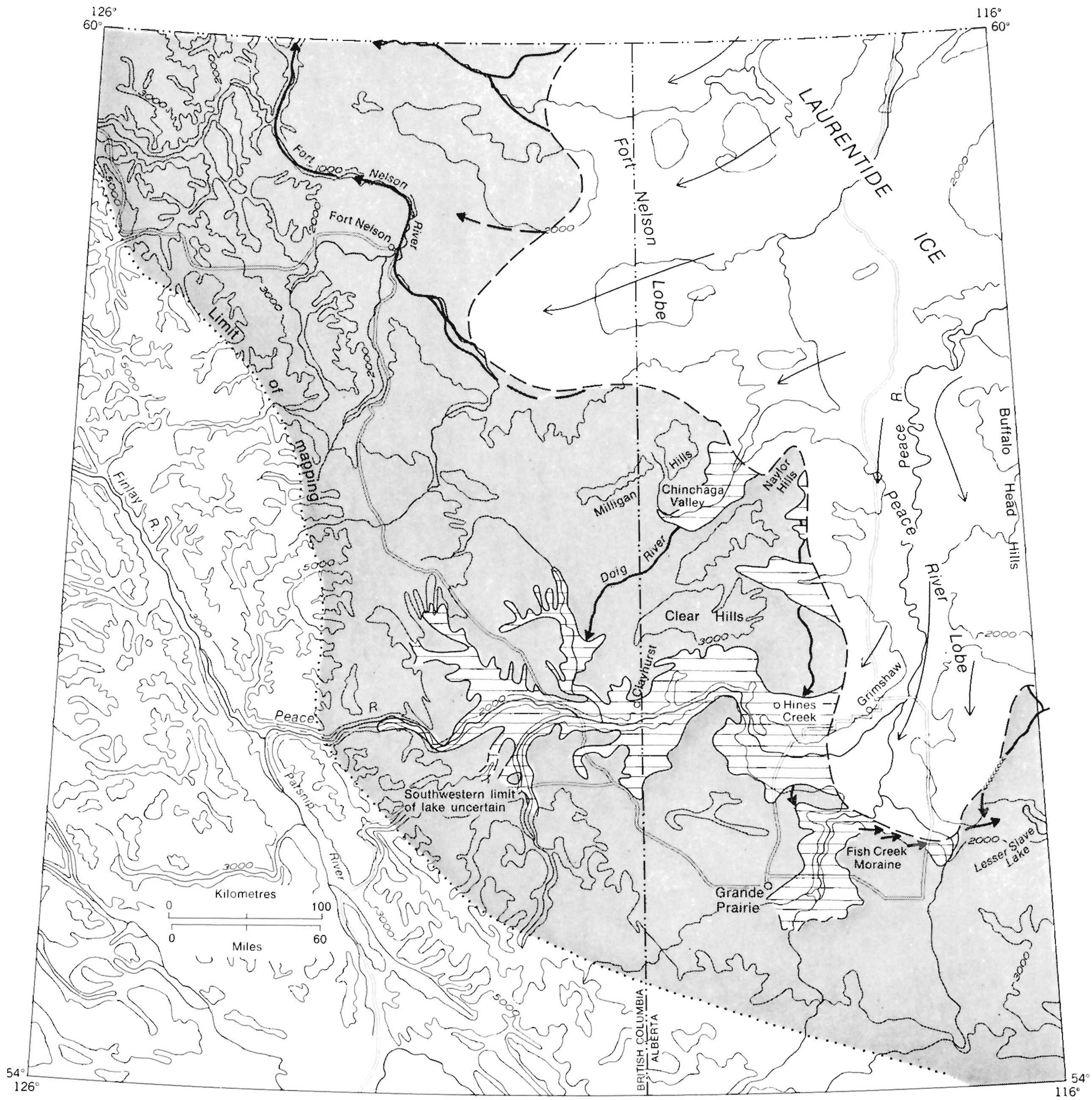


Figure 5D. *Phase 4A. Glacial Lake Peace, early Clayhurst stage.*

Within the northern Rocky Mountains high meltwater channels occur up to 1585 m (5200 ft) elevation 2 km south of mile 395 on the Alaska Highway (14 of Fig. 5A; Fig. 6). These drain easterly for the next 7 km, terminating at 1370 m (4500 ft) elevation. Near mile 390 a group of kames marks the head of a nearly continuous series of terraces on one or the other side of the North Tetsa River, declining from about 1265 m (4150 ft) a.s.l. here to 930 m (3050 ft) at mile 380 where it converges on the present river. The terrace marks a proglacial valley train which postdated the Laurentide climax as this ice sheet at some stage had contributed erratics as far west as mile 384 and as high as 1220 m (4000 ft) a.s.l.

The Liard lobe of the Cordilleran Ice Sheet in the extreme northwestern part of the area already seems to have withdrawn from any contact with Laurentide ice at this stage, back at least as far as 59°15'N, 125°15'W (Fig. 2), and to 60°15'N, 124°50'W. Only from here and from points farther to the west was ice-diverted meltwater able to descend to levels of less than 760 m (2500 ft) a.s.l.

Rocky Mountain ice at about or shortly after this stage seems to have reached only to the north end of Muncho Lake (mile 464 on the Alaska Highway) (see Fig. 2) where an ice-contact face (875 m, 2870 ft elevation) marks the upstream end of a major valley train. This valley train extends north to about mile 475, overlying a series of lacustrine beds at 805 m (2635 ft). The water from the former lake in which these beds were deposited is believed to have escaped easterly over a pass 8 km east of 480, at an altitude of about 790 m (2600 ft), thence flowing northeasterly towards Liard River. Damming to form the lake and obstruction of the outlet stream as it approached Liard River seem to have been accomplished by Cordilleran ice.



For legend see Figure 5A

Figure 5E. *Phase 4B. Glacial Lake Peace, late Clayhurst stage.*

Phase 2: Glacial Lake Peace, Bessborough Stage (Fig. 5B)

The area between the retreating Laurentide and Cordilleran ice sheets south of 57°N was covered mostly by waters of an ice dammed lake to which the name 'glacial Lake Peace - Bessborough stage' (Fig. 5B) is recommended (Mathews, 1978, p. 15)[1]. Records of this stage of the lake are well developed near the settlement of Bessborough and consist of four closely spaced shorelines at about 830 m (2725 ft). Correlative shorelines have been found at intervals from Beaverlodge, Alberta (Jones, 1961) north almost to 57°N (Mathews, 1978, his Fig. 6), a distance of 250 km. It is presumed that isostatic tilting, comparable if not greater than that affecting later shorelines, would have produced a westerly rise in the shorelines of this lake stage. Lacustrine beds up to at least 790 m (2600 ft) and the flat top to the kame moraine in the Rocky Mountains Portage (Fig. 7) at 885 and 858 m (2905 and 2815 ft) probably mark the western extension of this lake. On the other hand evidence of such high lake levels is lacking some 50 km east of the Alaska

[1] The terms Lake Bessborough and Lake Rycroft were proposed informally at a field conference on soils correlation in 1951, for ice dammed lakes represented by shorelines at about 840 and 685 m (2750 and 2250 ft) elevation at Bessborough, British Columbia, and Rycroft, Alberta; respectively. Henderson (1959, p. 74) adopted the term Lake Rycroft but, unfortunately, attributed it to the upper rather than the lower of the two lakes then known. As thus applied, "Lake Rycroft" would have had no shoreline anywhere near the town after which it was named, and indeed the site of Rycroft may have remained ice covered until after the disappearance of this lake. Meanwhile, Taylor (1958) formally had introduced the name Lake Peace, a term which already had been used informally by C. Hage in unpublished reports for Shell Oil Co. St-Onge (1967) used the term "Lake Rycroft" for several stages of ice retreat "before the Lesser Slave Lake area became ice free". To avoid continued confusion, therefore, it is recommended here that the name Lake Rycroft be abandoned and that at least for the time being the term Bessborough be considered as a stage of Lake Peace.

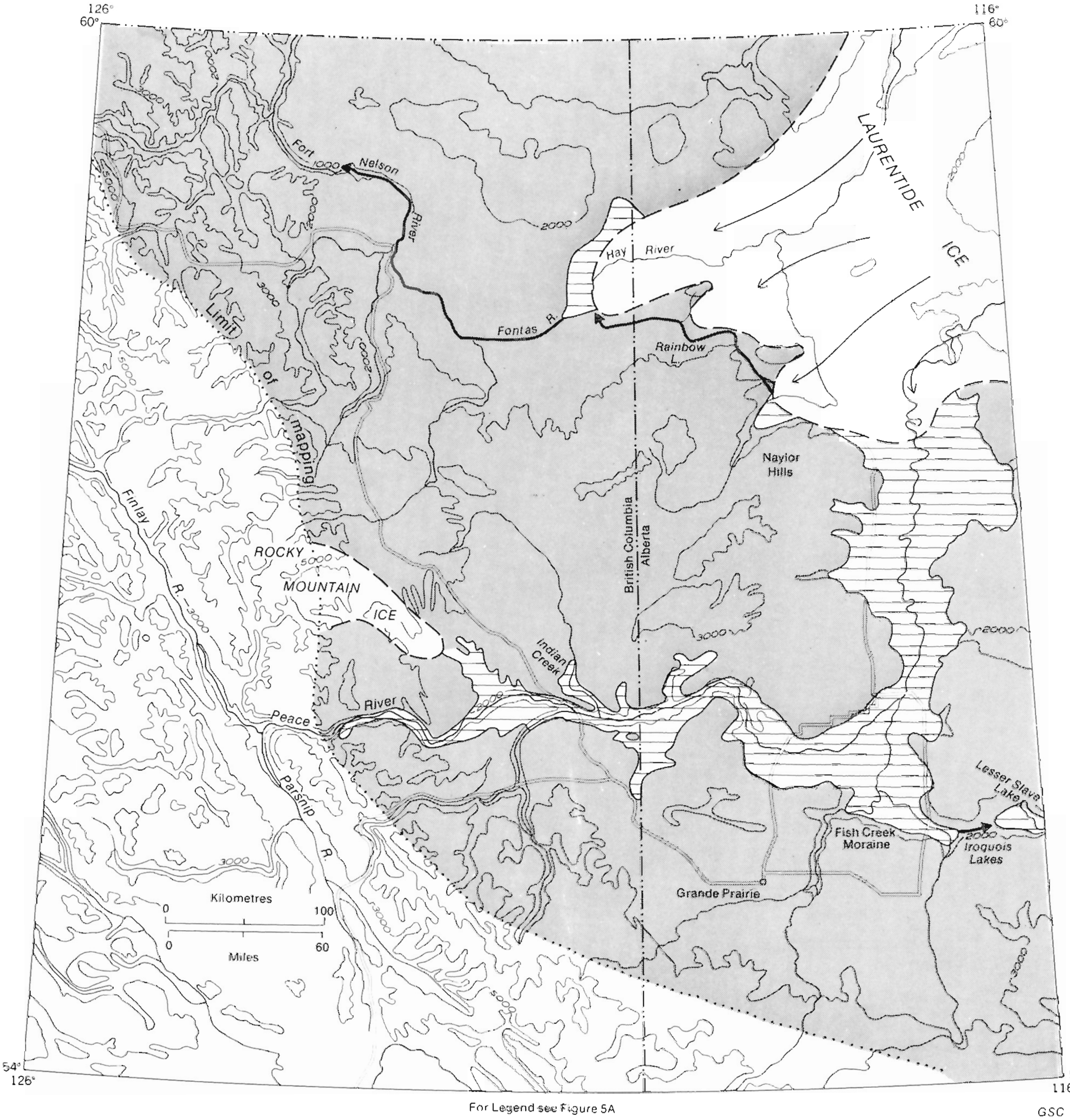


Figure 5F. *Phase 5. Glacial Lake Peace, Indian Creek stage.*

Highway where uplands rising to barely 760 m (2500 ft) a.s.l. bear no lacustrine sediment, and where ice-diverted meltwater channels descend to elevations well below 760 m (2500 ft), as near Beatton River airport. It is inferred, therefore, that a single large lake existed southeast from 57°20'N to beyond Beaverlodge, Alberta bounded on the east and northeast by the Laurentide ice front. On the grounds that this lake, with its relatively well developed strandlines restricted to a limited vertical range, must have had an overland outlet - rather than one over or through ice - a search has been made for an appropriate spillway. Such a spillway exists at the head of Pass Creek in the Iosegun Lake map area (St-Onge, 1967). Association of the Bessborough stage with this outlet implies a correlation of this lake stage with one or both of Lake Iosegun I or Lake Iosegun II of St-Onge (1972, p. 5-6) and with "Lake Rycroft" of Henderson (1959, p. 74).

Cordilleran ice at the Bessborough stage of Lake Peace evidently reached the mountain front, building the kame moraine into standing water at the mouth of Rocky Mountain Portage, 11 km west of Hudson Hope (Fig. 7). Farther south an ice tongue occupying Murray River valley diverted meltwater across to the head of the west fork of Kiskatinaw River via a meltwater channel heading and terminating between 880 and 850 m (2900 and 2800 ft).

Near mile 210 on the Alaska Highway and 160 km north of Peace River, a local history of ice diversion is particularly intriguing. Here a broad preglacial or interglacial valley occupied by upper Minaker River, flowing northeasterly, turns abruptly to the southeast where it is occupied in turn by Beaver Creek, draining northwesterly, and Trutch Creek, flowing easterly (Fig. 8 inset). The north side of this valley, east of Grassy Mountain, is scored over a 10 km length by

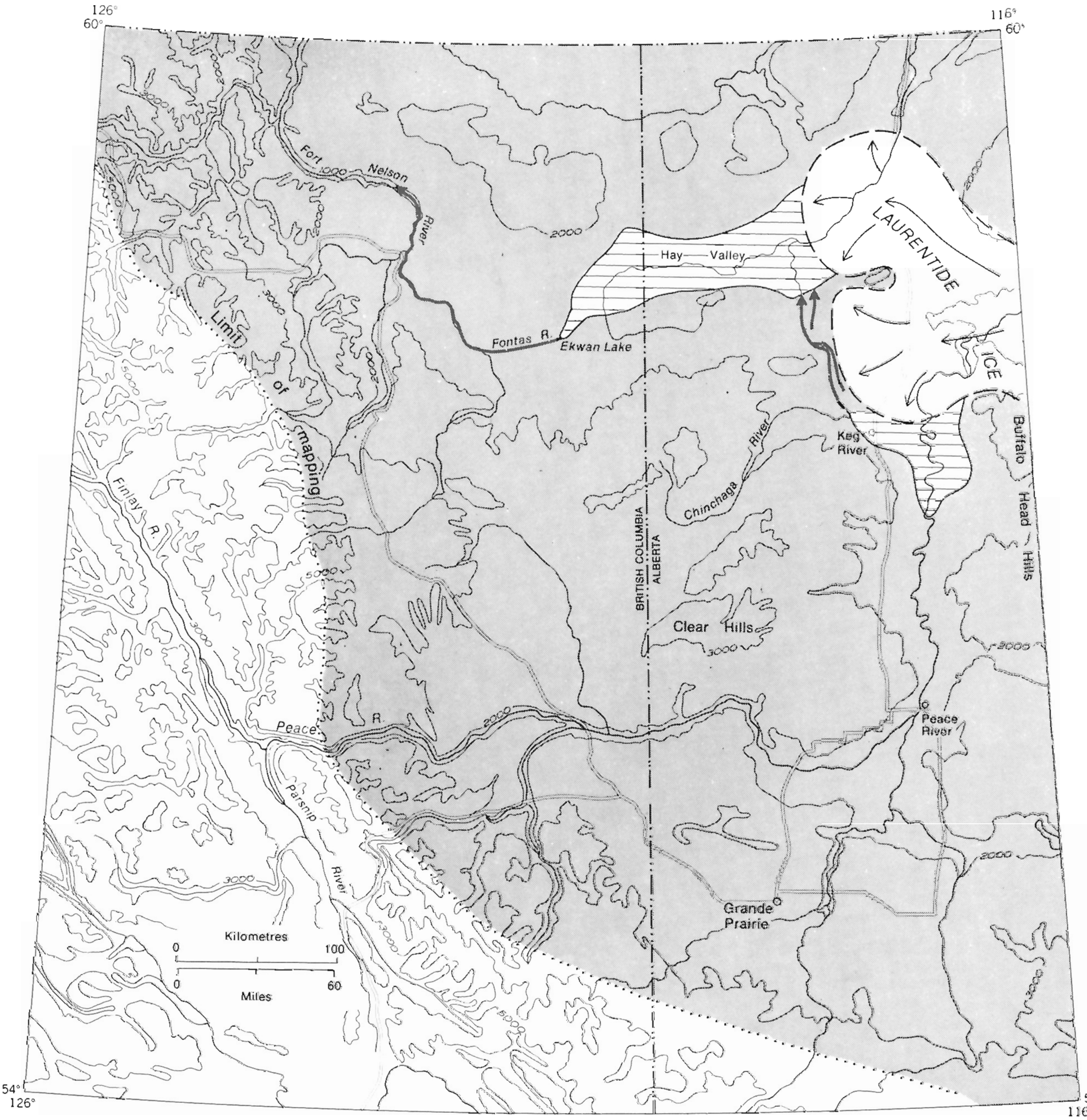
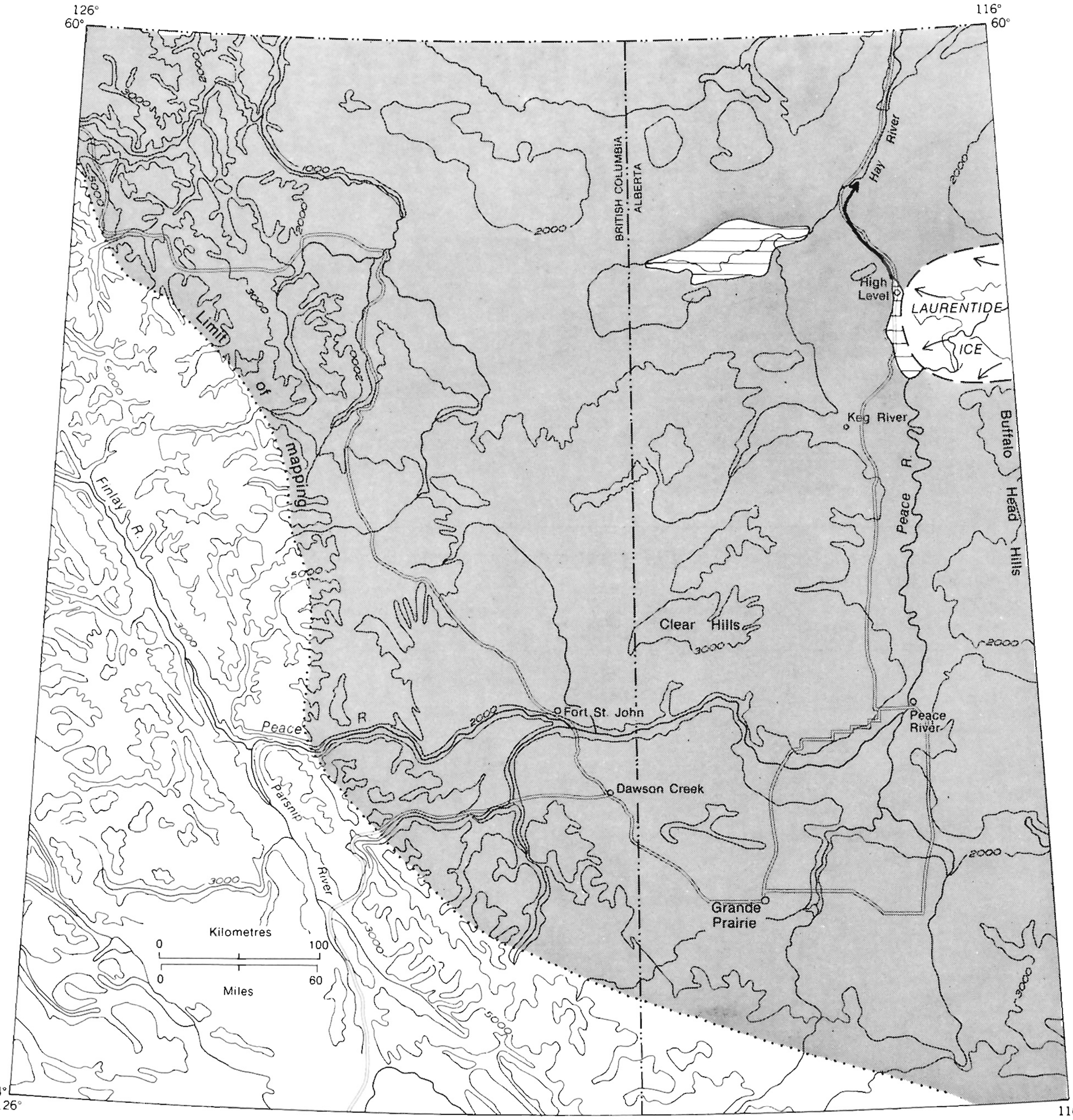


Figure 5G. *Phase 6. Glacial Lake Peace, Keg River stage.*

eight meltwater channels, draining southerly (Fig. 8). The ridge southwest of Grassy Mountain is transected by a steep-sided water gap, obviously youthful, through which passes the present Minaker River.

The eight meltwater channels head on or near the ridge crest and evidently were fed by drainage from a tongue of Laurentide ice to the north. At least one channel (1 of Fig. 8) heads on the south-facing slope and must have been fed directly from an ice lobe overtopping the divide. Channel 6, heading in a pair of lateral overflows, evidently was fed from an ice lobe terminating on the north-facing slope which diverted water first along the ice margin then through a saddle to the south slope. Channel 7, heading at the lowest point on the divide, may have received its water from an ice dammed lake on the north slope rather than being fed directly from the ice. The eight channels terminate, at varying elevations, part way down the slope presumably where they entered a body of standing water. Assuming the level of this body of water dropped continuously with time, channels 1, 2, and 3, terminating at about 805 to 808 m (2640 to 2650 ft) a.s.l. would be the oldest; channel 7, terminating at about 760 m (2490 ft), would have been the last to have functioned. A lake with surface elevation of 760 to 810 m (2490 to 2650 ft) could have existed only if the Beaver-Trutch section of the valley were effectively dammed on the east by Laurentide ice and if the ridge to the west had been unbreached below the lake level. The outlet of this lake is believed to have been initiated at the site of the water gap by ice diversion across a pre-existing saddle at about 840 m



For legend see Figure 5A

Figure 5H. *Phase 7. Initiation of High Level spillway.*

(2750 ft). It then would have become progressively deepened by meltwater from the northern channels (1-8 Fig. 8), from the ice dam to the east, together with normal runoff from the upper Minaker basin to the south. Once the water gap had been deepened below the 740 m (2425 ft) level of the Beaver-Trutch pass, the lake would be almost completely drained, and Minaker River would have been firmly established in its new course to Prophet River.

To the west of the former Minaker-Prophet divide another ice dammed lake existed which apparently submerged the Prophet-Muskwa divide 25 km west of mile 220 and extended another 65 km northward along the front of the Rocky Mountain Foothills to another sill. At the time the water could not flow northeasterly down the broad low valley of Akue Creek, presumably because this was plugged by Laurentide ice, and instead it spilled across the sill to the lower part of Tetsa River 6 km south of mile 350. This spillway was cut to sufficient depth to survive into postglacial time as the present course of Muskwa River. When it was first being cut, however, the easterly extension of Tetsa Valley also was blocked by ice, and the meltwater presumably escaped to the north via the Mill-McLennan spillway described above. From there the water entered an ice dammed lake in Dunedin Valley to be diverted in turn northwesterly to another lake in Liard Valley ponded to the level of the 520 m (1700 ft) spillway at 60°N. From that point it would have flowed along the front of the Mackenzie Mountains ultimately to join Mackenzie River.

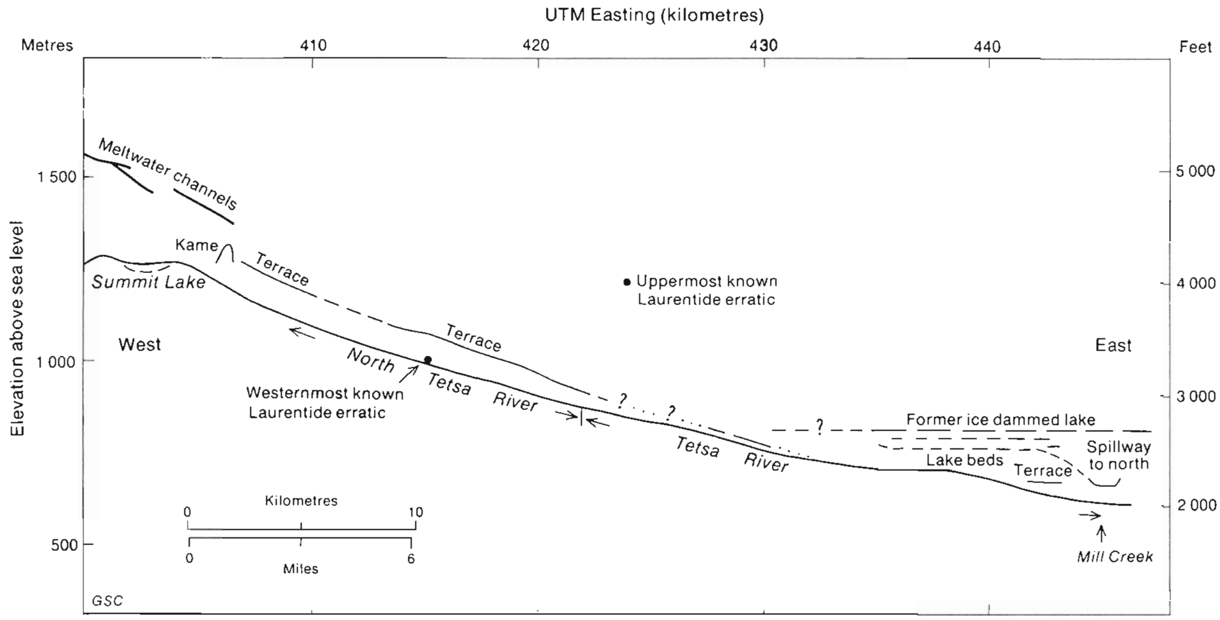


Figure 6. *Longitudinal (east-west) profile of Tetsa Valley between Summit Lake on the west and the Mill McLennan spillway on the east.*

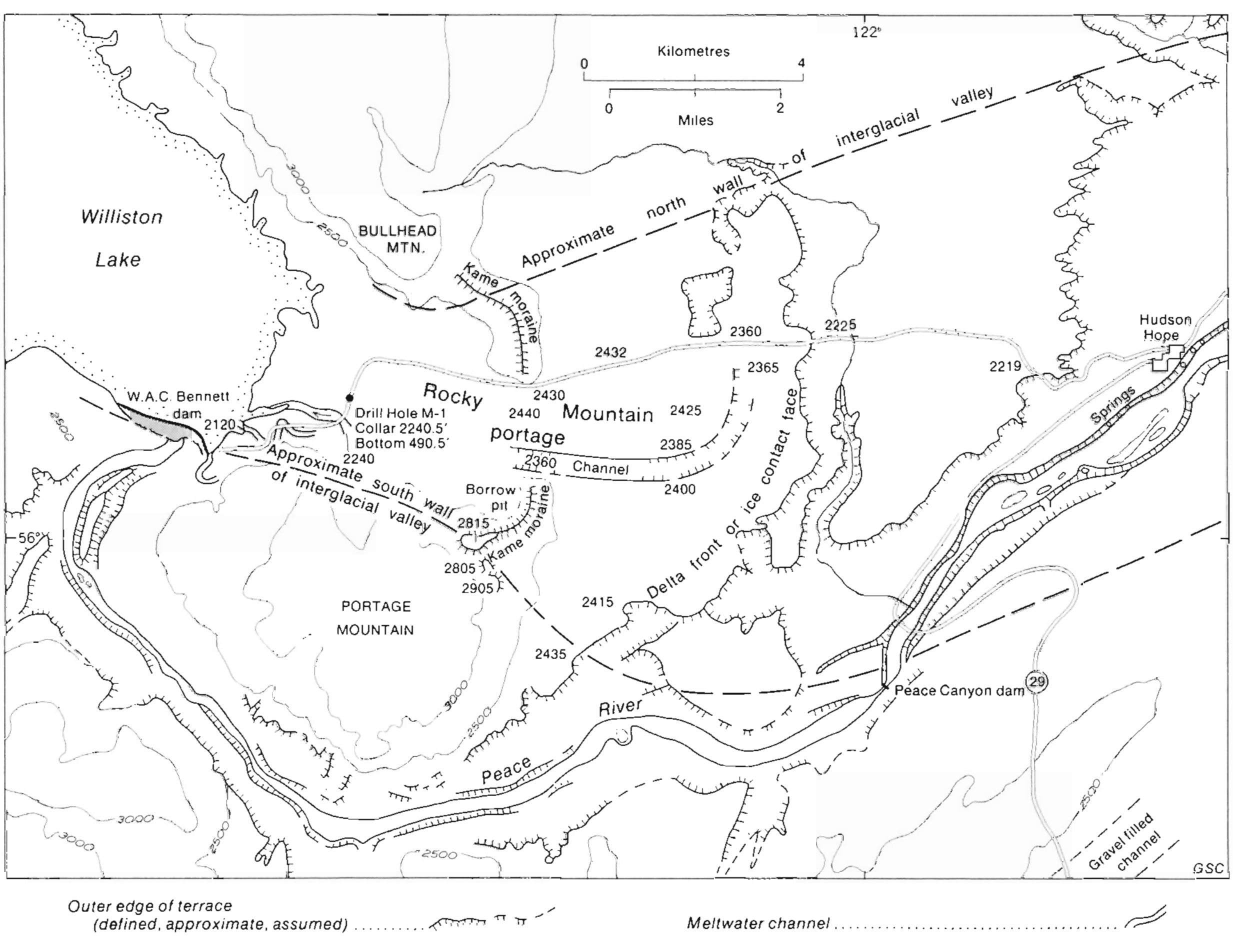


Figure 7. *Peace River Canyon – Hudson Hope area, British Columbia, showing the former (interglacial) Peace River valley, late glacial kame moraine and meltwater channel in the portage, and postglacial trench of Peace River.*

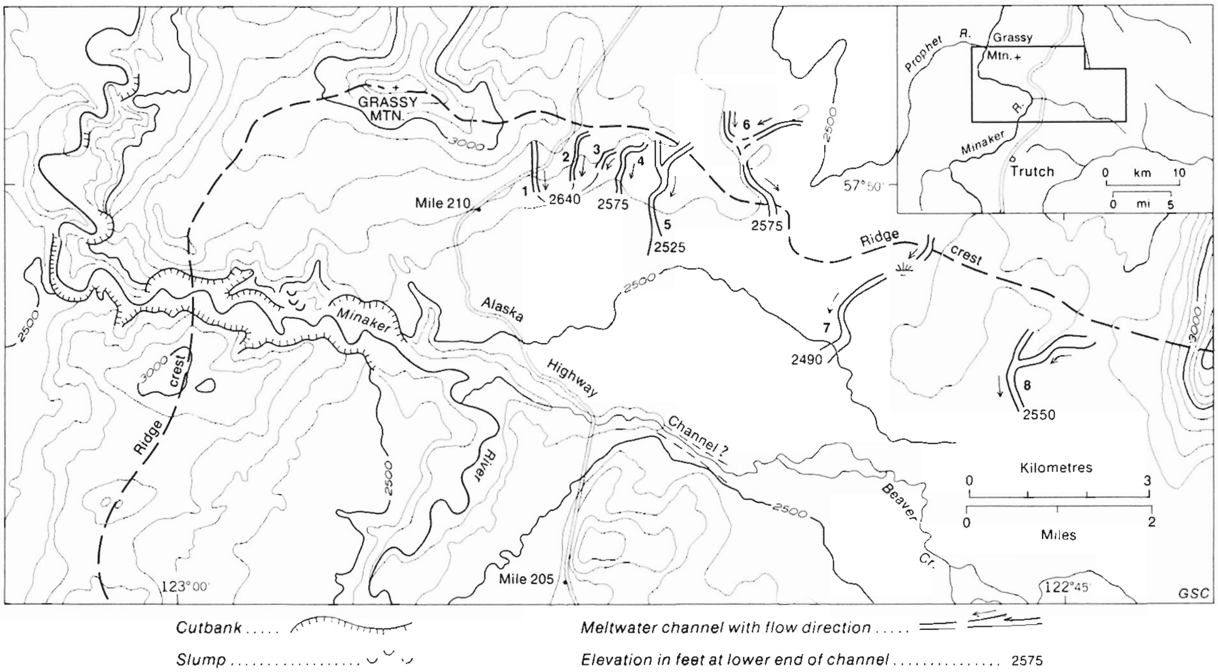


Figure 8. *Minaker River diversion, near Trutch, British Columbia.*

The position of the Rocky Mountain ice and of the ice front of the Liard lobe of the Cordilleran ice sheet at this phase is not known.

Phase 3: Glacial Lake Peace, unnamed stage (Fig. 5C)

An unnamed stage of glacial Lake Peace is marked by shorelines in the Charlie Lake map area (Mathews, 1978, p. 15) at about the 800 m (2625 ft) level. In the absence of known shorelines between this level and the Bessborough level 30 m (100 ft) higher, it is presumed that the opening of a new and distinctly lower outlet, rather than continued downcutting of the Pass Creek spillway, was responsible for the change. Conceivably, the new lake stage in the Charlie Lake area corresponds to Lake Iosegun III (St-Onge, 1972, p. 7). Alternatively the basin in the north may have been separated from Lake Iosegun III by a sill a few kilometres east of the Alberta-British Columbia boundary at 55°25'N. The water level in the northern basin seems to have persisted during the uncovering of the western part of the Milligan Hills where several ice-diverted meltwater channels terminate abruptly at about the 790 m (2600 ft) level.

Assuming that the ice front rose to comparable heights 50 km northwest of the intakes to the Milligan Hills channels, Sikanni Chief Valley still would have been blocked high enough to divert water westerly via Trutch Valley and the Minaker gap into the proglacial lake in the Prophet-Muskwa Valley, thence northerly as in Phase 2 towards Mackenzie River. Retreat of Laurentide ice since Phase 2 could have led to some expansion of the proglacial lake easterly down the Prophet, Akue, and Tetsa-Muskwa valleys. Withdrawal of ice from the mountain front in the Liard re-entrant, moreover, could have opened a new and lower route for the water from Dunedin Valley into a lake ponded to about the 460 m (1500 ft) level of a spillway at 60°07'N, 124°15'W extending northward to La Biche Valley.

Phases 4A and 4B: Glacial Lake Peace, Clayhurst Stage (Fig. 5D, 5E)

The Clayhurst stage of glacial Lake Peace was a relatively long-lived stage marked in many places by a series of closely spaced shorelines. Near the British Columbia-Alberta boundary at Clayhurst the shorelines are found between 660 and 690 m (2170 and 2260 ft) elevation. Farther west the shorelines are found at progressively higher elevations (Mathews, 1978, his Fig. 6) indicating a tilt to the east of about 0.4 m/km (2 ft/mile). An abandoned channel in the Rocky Mountain Portage (Fig. 7), 10 km west of Hudson Hope, terminates in a delta at about 740 m (2420 ft) a.s.l. which is believed to correspond to one of the higher shorelines of the series. Glaciolacustrine deposits and kame terraces extending to slightly above 730 m (2400 ft) on either side of Peace River for 65 km west of the Portage may correspond to one of the younger shorelines. Another abandoned channel, with broad sweeping meanders and a flat floor 0.5 km wide, can be traced from the Pine Valley arm of the former lake southwesterly 19 km to a well defined ice-contact face, 11 km northeast of Chetwynd, marking a former limit of the Pine Valley glacier. Conceivably this ice may have withdrawn many kilometres, leaving a lake spilling through the abandoned channel, before a new and lower escape route was made available easterly to Murray Valley (Hughes, 1967, p. 21-22).

Strandlines at about 655 m (2150 ft) elevation can be identified in aerial photographs and extend intermittently from Cardinal Lake, 6 km west of Grimshaw, Alberta, southwesterly and westerly for about 27 km. The south-western edge of a nearby dune field and termination of the Hines Creek channel, both at about 655 m (2150 ft) elevation, probably mark another part of the shore.

The inferred outlet, or outlets, for the Clayhurst stage of glacial Lake Peace can be found along the line of the Fish Creek moraine (Henderson, 1959, p. 25-26) where there is a series of large channels at 650 to 680 m (2125 to 2240 ft) elevation (Fig. 4). Meltwater escaping around the ice front here would have drained easterly via the valley of Lesser Slave Lake.

It is at this stage of Lake Peace that the Hines Creek meltwater system (Fig. 4), described above, must have functioned, and the record of this channel indicates that contemporaneous ice must have just buried the north end of the Naylor Hills, covering all ground up to the slightly above 760 m (2500 ft) elevation. The eastern end of the Milligan Hills, 50 km to the west-northwest, then would have been overriden by ice to a comparable elevation, and the northeasterly draining Chinchaga Valley, between the two sets of hills, would have been dammed. In keeping with this conclusion is the record of an overflow channel leaving the southwestern end of Chinchaga Valley, draining via Doig River, and terminating at about 670 m (2200 ft) in an abandoned delta at a Clayhurst shore.

The ice margin during this time must have dipped gradually westerly along the north slope of the Milligan Hills where a series of westerly draining channels is obvious. One such system declines gradually from about 700 m (2300 ft) a.s.l. at 57°40'N, 120°50'W to about 585 m (1925 ft) at 58°10'N, 122°20'W, a distance of 88 km. This conceivably may link with another system of channels starting at 58°17'N, 123°00'W, elevation 585 m (1925 ft), 65 km south of Forth Nelson, and draining northerly and northwesterly for 55 km to terminate at the 510 m (1675 ft) level. Whether the water from here drained on to the northwest via a gap in the Poplar Hills at 59°00'N and 124°00'W, or by a clearly defined channel system at the east end of these hills, is unknown. The former gap, according to some topographic maps, is significantly higher than 510 m (1675 ft) and shows no clear marks of Wisconsinan stream flow. Drainage via the channel system east of the Poplar Hills would necessitate an abnormal form for the ice front. In either case the meltwater would have found escape to lower Fort Nelson River, thence to Liard River, and northward towards the Arctic Ocean.

Interpretation of the ice limit northeastward from Fort Nelson presents problems. It has been assumed for the purpose of illustration (Fig. 5D) that the ice margin hugged the foot of the Etsho escarpment, rising at its eastern end to a level comparable to that on the Milligan Hills to the south, thence bowing northwestward into Petitot Valley. The complex of channels and eskers here, however, suggests ice stagnation and a coherent ice front may not have existed at the time.

With continued retreat of ice in the Fort Nelson basin (Fig. 5E), marginal channels low on the Milligan Hills would have taken over the function of their higher counterparts. These seem to have flowed into a braided system of channels extending along and west of Fort Nelson River from the mouth of Sikanni Chief River almost to the site of Fort Nelson where the drainage became concentrated into a single watercourse followed by the present river.

It may seem inconsistent in Figures 5D and 5E to interpret a significant retreat of ice in the Fort Nelson basin at the same time that ice remains more or less stationary at the Fish Creek moraine to the southeast. Clearly the correlation of ice front positions between the two areas lacks good control. Moreover the two ice lobes indeed may have behaved differently. Henderson (1959, p. 75-76) for example, suggested that the Fish Creek moraine marks the limit of a significant readvance following a period when ice dammed lakes of Peace River had been drained below 555 m (1825 ft). Conceivably there could be a complex history to the Peace River ice lobe (Fig. 5E) during the Clayhurst stage(s) with almost no retreat of this ice front during the same period in which the Fort Nelson lobe withdrew 80 km.

Phase 5: Glacial Lake Peace, Indian Creek Stage (Fig. 5F)

The last of the major ice dammed lakes known to have existed in Peace River valley is referred to as the Indian Creek stage of Lake Peace. This stage is marked by a single, moderately well developed strandline about 30 m (100 ft) below strandlines of the Clayhurst stage. The probable outlet is the Iroquois Lakes spillway (elevation 590 m, 1940 ft) near High Prairie, Alberta, which leads to Lesser Slave Lake and Athabasca River drainage. The front of the ice dam in Peace River valley must have been located somewhere north of the Fish Creek moraine but no farther than the present site of Keg River, Alberta, 225 km to the north. This stage of Lake Peace is known to have extended at its maximum westerly past 122°45'W where extensive kame terraces are concentrated at 730 m (2400 ft) elevation (Mathews, 1946, p. 18).

Termination of this lake would have occurred with the retreat of Laurentide ice northeasterly from the site of Keg River, allowing escape of water westerly around the north end of Naylor Hills into Chinchaga Valley and thence by a well developed meltwater channel northwesterly along upper Hay River and Rainbow Lake to 120°10'W where a deltaic (?) deposit near 425 m (1400 ft) elevation marks an early lake in the Hay River basin. This lake evidently drained westerly via Fontas and Fort Nelson rivers to Liard and Mackenzie drainage.

A possible repetition of the Indian Creek stage is suggested by the presence of a well developed strandline cut into earth mounds near Indian Creek (Mathews, 1978, p. 15). Although the precise origin of the mounds remains problematical, their common association with local calcareous slough deposits suggests that they developed on an exposed land surface following the disappearance of the early ice dammed lakes. A readvance of ice near Keg River and a temporary restoration of the Indian Creek stage thus may be indicated.

A late advance of ice from the Rocky Mountains is revealed by a sandy outwash apron which can be traced down Halfway Valley (Mathews, 1978, p. 12) following drainage of the Indian Creek stage.

Phase 6: Glacial Lake Peace, Keg River Stage (Fig. 5G)

A late, ice-diverted drainage is recorded by a major meltwater system following lower Chinchaga River northward from the settlement of Keg River, Alberta. The ice responsible for this diversion must have blocked Peace River valley up to about the 425 m (1400 ft) level, creating the relatively small water body of glacial Lake Peace, Keg River stage, extending south to approximately 57°20'N. The Chinchaga channel dies out at about 380 m (1250 ft) elevation where it likely entered a second ice dammed lake in Hay Valley, a lake that almost certainly overflowed westerly to Fontas and Fort Nelson rivers across a spillway which ultimately retreated to the east end of Ekwan Lake (elevation 380 m, 1250 ft).

Phase 7 (Fig. 5H)

With the withdrawal of ice from the Keg River area northeast to the vicinity of High Level, Alberta, some 65 km away, a new spillway at 325 m (1075 ft) elevation was opened from Peace to lower Hay valleys. At this stage almost no ice-ponded water survived within the area mapped here. One lake, however, persisted for a time in upper Hay Valley near 118°00'W. The lake is now marked by a well developed shoreline close to 335 m (1100 ft). Progressive downcutting of the outlet since has exposed most of the lake floor, leaving a great swamp and a series of relatively small water bodies.

Table 1

Radiocarbon dates from northeastern British Columbia and northwestern Alberta bearing on the glacial history

Laboratory Number	^{14}C Date years B.P.	Location	Elevation m (ft)	Material
Wat-361	>30 000	55°34'N, 119°25'W	870 (2860)	Organic matter
I-4878	27 400 ± 850	55°43'N, 117°38'W	400 (1310)	Wood, peat
GSC-2034	27 400 ± 580	56°09'N, 120°43'W	455 (1500)	Tooth apatite
GSC-573	25 940 ± 380	56°18'N, 124°21'W		Plant
GSC-2859	25 800 ± 320	56°00'N, 122°07'W	760 (2500)	Tusk collagen
I-2244A	≥11 600	56°00'N, 122°07'W	760 (2500)	Tusk collagen
Wat-406	17 570 ± 650	55°34'N, 119°25'W	870 (2860)	Organic matter
GSC-698	13 580 ± 260	54°21'N, 117°01'W	820 (2700)	Small shells
GSC-694	13 510 ± 230	54°21'N, 117°01'W	820 (2700)	Large clean shells
GSC-508	12 190 ± 350	54°21'N, 117°01'W	820 (2700)	Freshwater gastropods
GSC-1049	11 400 ± 190	54°44'N, 112°24'W	625 (2050)	Gyttja
GSC-1654	10 400 ± 170	55°59'N, 120°16'W	655 (2150)	Freshwater shells
GSC-1548	9 960 ± 170	55°59'N, 120°15'W	670 (2200)	Freshwater shells
GSC-1497	9 280 ± 200	55°48'N, 123°38'W		Sheep skull

Designations for radiocarbon dating laboratories:

GSC - Geological Survey of Canada, Ottawa, Ontario
Wat - Department of Earth Sciences, University of Waterloo, Waterloo, Ontario
I - Teledyne Isotopes, Westwood, New Jersey

GEOCHRONOLOGY

Radiocarbon dates pertinent to the history outlined here are listed in Table 1.

Three dates (I-4878, 27 400; GSC-2034, 27 400; and GSC-573, 25 940) are from beds below the youngest till indicating a later Wisconsinan glaciation at their sites. The core samples dated >30 000 years (Wat-361) and 17 570 years (Wat-406), on the other hand, had no younger till cover (White et al., 1979), and it is possible that their site at Boone Lake, high in the Saddle Hills of Alberta, escaped the late Wisconsinan glaciation.

A special problem is presented by the two dates GSC-2859 and I-2244A from bone collagen of a tusk, from the Portage Mountain kame-moraine, found in 1966 during construction of the W.A.C. Bennett dam. When the tusk was first discovered, a sample was submitted to Isotopes Inc., and a date of 7670 ± 170 years (I-2244) was obtained from the carbonate fraction. Suspicion was then raised that carbonate-bearing groundwaters had contaminated the tusk, and the organic (collagen) fraction remaining from the first determination was dated. This yielded a date of 11 600 ± 1000 years (I-2244A). The weight of collagen utilized, however, was so small that no estimate of counting precision could be made, and the quoted age "represents a minimum value" (J. Buckley, Teledyne Isotopes, personal communication, 1972 to W. Blake, Jr.). The remainder of the tusk was coated with a preservative (Krylon?) and placed on display at the dam. In 1978 still another sample was obtained from the same tusk, by carefully selecting core material which, although partially imbedded in plaster used to fill the hollow interior, had hopefully escaped impregnation by the preservative. The collagen obtained in this way yielded the 25 800 ± 320 year age (GSC-2859). This date from a proglacial deposit near the Cordilleran ice limit is difficult to reconcile with the nearly contemporaneous date (GSC-573) from nonglacial sand (Rutter, 1977, p. 19, 20) 135 km to the west and much closer to the source of Cordilleran ice. This discrepancy raises the question of some contamination, notwithstanding the care taken in the latest sampling of the tusk, or even of the chance of redeposition of the tusk. Alternatively, however, it raises the possibility that the Cordilleran climax at Portage Mountain was early in Classical Wisconsinan time and that Cordilleran glacial deposits still farther to the east are significantly older. The date, moreover, throws in doubt the previous interpretation that the tusk, the kame-moraine, and the apparently contemporaneous Bessborough stage of Lake Peace were of late Classical Wisconsinan age (Mathews, 1978, p. 17 and Fig. 7, p. 14).

St-Onge (1972, Fig. 3) indicates that drainage of Lesser Slave valley occurred about 11 000 years ago, a date established, in part, by GSC-1049. According to the pattern of deglaciation outlined here this would correspond to the end of the Clayhurst stage or a later stage of Lake Peace. The last two dates shown in Table 1 are believed to postdate the withdrawal of ice dammed waters (Mathews, 1978, p. 13) from their sites at 655 and 670 m (2150 and 2200 ft) a.s.l., and hence to postdate the Clayhurst stage of Lake Peace. This stage seems to have terminated more than 10 000 radiocarbon years ago.

Finally, an estimate of the number of varves in the late glacial lacustrine succession near Fort St. John (Mathews, 1978, p. 11) indicates that sedimentation during the Bessborough through Clayhurst stages of Lake Peace involved at least 300 years.

PERMANENT STREAM DIVERSIONS

Perhaps the best example of stream diversion associated with deglacial history is that of Peace River at the eastern front of the Rocky Mountains (Beach and Spivak, 1943; Mathews, 1946, p. 18), the site of the Rocky Mountain Portage and the W.A.C. Bennett dam. Clearly the main valley of the Peace, enlarged, though probably not initiated, by glaciation, formerly extended through the Rocky Mountain Portage (Fig. 7), and it now can be traced in subsurface records east and northeast past Hudson Hope (Mathews, 1964). A hole (M-1, Fig. 7) drilled in the Portage (Rutter, 1977, p. 13) reveals several gravel sheets indicating fluviatile (or glaciofluvial) action during valley infilling (this hole terminated in unconsolidated fill). The final filling of the valley by well sorted gravels of a kame moraine built the ground surface above that of a shallower valley leading southward and eastward around Portage Mountain. When the Cordilleran ice retreated, this valley first was occupied by an arm of the Clayhurst and Indian Creek stages of Lake Peace, extending westerly to the wasting ice front. Later, with the draining of Lake Peace, Indian Creek stage, the shallower valley was occupied by Peace River, which in time cut through lacustrine sediments into bedrock, creating the Peace River canyon at the head of which the W.A.C. Bennett dam now has been built.

Nevis Creek, and possibly Besa River as well, 160 km north of Peace River Portage (Fig. 2, 9), has been diverted northwards from an easterly trending valley at 57°28'N, some 8 km west of the mountain front, apparently by a fill of unconsolidated sediments - probably silts, judging from the gently sloping slump topography now masked by continuous vegetation - directly east of the point of diversion. The northerly flowing channel of Nevis Creek has been cut almost 120 m into shales and has become thoroughly stabilized in its new course.

The Minaker River diversion (Fig. 8) is another example in which Laurentide ice played the dominant role in stream diversion.

Muskwa River, north of Akue Creek (Fig. 2) is also a stream diverted at the end of the last glaciation.

Fontas River has been reversed for a 65 km stretch between the Fort Nelson-Hay River divide, at the east end of Ekwan Lake, almost to Sikanni Chief River.

Two similar but older examples of diversions may be mentioned here, even though they are not related to the latest glaciation: (1) Nevis Creek, 12 km south of Besa River, turns abruptly northward (Fig. 9) from its large easterly draining valley (57°21'N), some 10 km west of the mountain front, and 6 km upstream from the diversion described above. Downstream it occupies a valley cut in shale and occupied by glacial and fluvioglacial fill at levels 100 m (350 ft) below that of the bedrock sill in the easterly, abandoned continuation of the valley. (2) Sikanni Chief River, near 122°W, formerly may have turned southeasterly through a windgap at the heads of Coal and Laprise creeks into upper Beatton River (Fig. 2). If so, the presence of chert gravel from the Rocky Mountains in the interglacial valley of Beatton River at Rose Prairie (Mathews, 1963, p. 7) may be explained; at present Beatton River drainage no longer includes any pre-Cretaceous outcrops from whence the chert originally must have come.

INTERPRETATION OF SURFICIAL MATERIALS ON THE BASIS OF THE RECORD OF ICE RETREAT

The history of ice retreat, as outlined above, can provide additional clues, beyond those used in establishing the history, concerning the surficial materials to be expected throughout the area.

Individual meltwater channels, for example, can be expected to be floored with a veneer of gravel left from the reworking of the deposits into which the channel was cut. The floors of larger and long-lived channels, moreover, may have a significant depth of gravel which has been winnowed by reworking, washed down from the upstream sections of the channels, or in some cases contributed directly from glacier ice at the head or margin. One such channel, that of Doig River leading from Chinchaga Valley to the Clayhurst stage of Lake Peace in Beatton River valley, is floored by a 6 m-thick mantle of gravel which is an important local source of construction material. The downstream termination of a channel where it entered a former ice dammed lake to form a delta is the part most likely to contain thick deposits of coarse aggregate.

Areas with meltwater channels essentially would have been drained on or very shortly after withdrawal of the ice and therefore may be expected to have little or no sediment covering the freshly exposed glacial deposits. Marginal ice dammed lakes may have developed locally in these areas but are likely to have been small and short lived. Sediments deposited within the lakes can be expected to be thin and to show, by a relatively high proportion of stones, the influence of the nearby ice. Areas where meltwater channels are rare or lacking, by contrast, may have been the sites of large, former ice dammed lakes and may be expected to be mantled with fine textured and frost-susceptible sediments deposited on low, flat parts of the basin floor. Where the reconstructed history indicates that ice dammed lakes persisted at the same spot for two or more phases of ice retreat, an opportunity probably existed for the accumulation of considerable depth of lacustrine sediment. The area near Fort St. John and Taylor, British Columbia provides such an example; here an average depth of postglacial lacustrine beds of clay, silt, and locally, sand is about 16 m (55 ft). The stony material that was dropped directly from the ice front or carried out from it by icebergs typically is concentrated in the so-called "proximal" sites of a large lake, either (1) at the base of the lacustrine succession laid down immediately following withdrawal of the ice or (2) throughout the succession in areas close to the ice front when the lake was finally drained (here no opportunity existed for burial of proximal beds by stone-poor "distal" sediments).

The lithology of the till, channel deposits, or lake sediments at any particular site of interest can be inferred from the pattern of ice movement and water flow, relative to

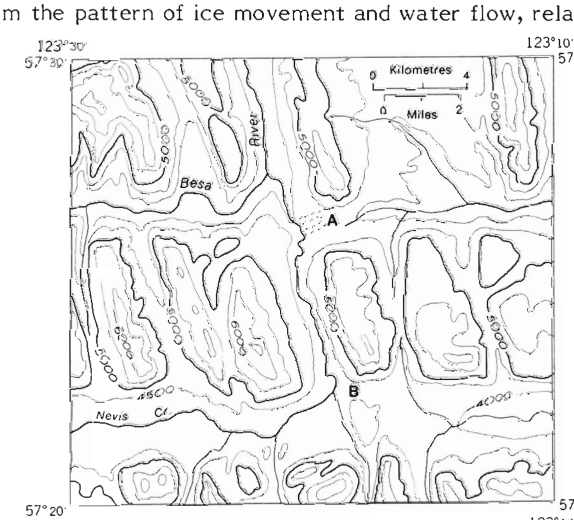


Figure 9. *Nevis Creek – Besa River area, British Columbia, showing at A the site of a late glacial diversion of Nevis Creek (and Besa River?) by valley infill (stippled) and, at B, the site of an earlier diversion of Nevis Creek.*

bedrock sources. Thus, for example, tills deposited in the western part of the area and derived from the Cordillera can be expected to contain numerous pebbles and boulders of quartzite and sandstone in a fine matrix containing significant amounts of carbonate derived from limestone of the Rocky Mountains. By contrast, till laid down by Laurentide ice can be expected to contain scattered pebbles, cobbles, and boulders of resistant rocks from the Canadian Shield, notably gneiss and granite, scattered through an abundant matrix rich in clay derived from the erosion of Cretaceous shales over which the Laurentide ice had passed. Gravels from the channel and deltaic deposits of the meltwater streams should have much the same lithology as does the coarse fraction of the tills from the same source area. Lacustrine deposits, by the same token, generally should be lithologically similar to the finer fractions of the Laurentide or Cordilleran debris, depending on the source, or sources, of meltwater feeding the lake; mixtures from two sources may occur locally.

ARCHEOLOGICAL IMPLICATIONS

The area considered here, lying as it does astride the so-called 'ice free corridor' between Cordilleran and Laurentide ice sheets, is of concern to the archeologist studying the passage of early man from Alaska to southern Canada and points beyond. Whether man did, in fact, use this corridor as his main access route, rather than the west coast of North America or some intermediate course, in this original migration remains an open question. It appears, however, that the corridor may have been closed for a time during the climax of the last glaciation by local coalescence of the two ice sheets between about 52°15'N in southern Alberta (Harris and Boydell, 1972, p. 49) and 60°10'N in the southeastern corner of Yukon Territory (this study; see White et. al., 1979). When coalescence of the ice first occurred is unknown, but it may have been after the two 27 400 year radiocarbon dates (Table 1). For how long the corridor remained closed and whether man could have overcome the obstacle by travelling from nunatak to nunatak also remains unknown. Continuity of the corridor would have been regained by withdrawal of ice in the high plateau between Peace and Fort Nelson rivers shortly before phase 2 of the glacial history outlined above, and migration would probably have been physically manageable, assuming a supply of food, by the time of the Bessborough stage of glacial Lake Peace.

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