THE UNIVERSITY OF ALBERTA

Quaternary Sedimentology and Stratigraphy, Peel Plateau and Richardson Mountains, Yukon and N.W.T.

by

Norman Rhoderick Catto

A THESIS

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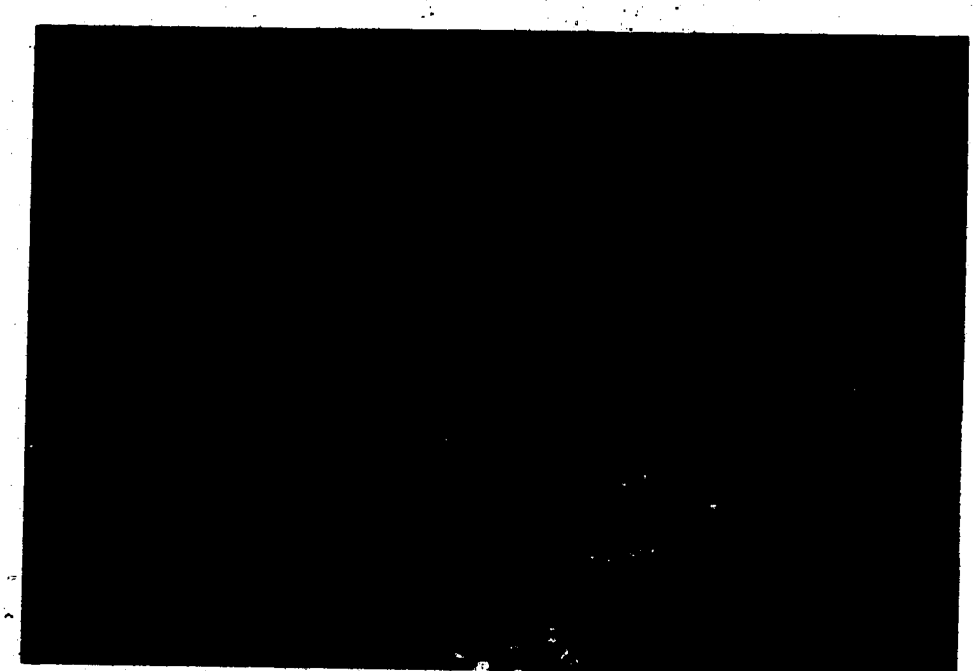
Plate 1 Frontispiece

The Peel Plateau and the foothills of the Richardson Mountains. The photograph was taken looking west along the Dempster Highway 50 km southwest of Fort McPherson.



THE UNIVERSITY OF ALBERTA

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The undersigned certify that they have read, and recommend to the Faculty of Graduate Studies and Research, for acceptance, a thesis entitled Quaternary Sedimentology and Stratigraphy, Peel Plateau and Richardson Mountains, Yukon and N.W.T. submitted by Norman Rhoderick Catto in partial fulfilment of the requirements for the degree of Doctor of Philosophy.

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Date..February 17, 1986..........

Abstract

Quaternary sediments exposed throughout the Peel Plateau and Richardson Mountains in the western portion of the Northwest Territories and northeastern Yukon record a succession of glacial, fluvial, and lacustrine events. The oldest unconsolidated sediments are preglacial fluvial gravels of unknown age, exposed along tributaries of the Peel River. The oldest glacial event in the region is recorded by the intensely weathered granite-bearing Brown Bear gravels exposed along the Peel River. These sediments are also of unknown age.

A second glaciation is documented by till exposed along the Snake River, and by glaciotectonically deformed sediments in the central Peel Plateau. This event is correlated to the Deception lacustrine sediments in the Bonnet Plume Basin, and pre-dates 40,000 B.P.

The final glacial event was the most areally extensive to affect the region. Sediments formed during this event are correlated to the Hungry Creek and Buckland glaciations recognized west and north of the Peel Plateau. This glaciation commenced after 31,000 B.P. Deglaciation commenced at some time prior to 22,000 B.P., and resulted in the development of lacustrine sequences in the Rat and Caribou River valleys.

The region was deglaciated by 12,000 B.P. The postglacial environment has been characterised by the development of braided-meandering streams, periglacial

activity, and the re-establishment of tundra and taiga vegetation.

Acknowledgements

The advice, guidance and encouragement of Dr. N.W. Rutter at all stages of this project is gratefully acknowledged.

I was especially fortunate to be able to work with Dr. Owen Hughes of the Geological Survey of Canada. Somehow, he managed to tolerate the antics of a cheechako, and kept the project under control.

Consultations with many researchers greatly improved this thesis. In particular, discussions with C.E. Schweger and J.V. Matthews Jr. clarified several puzzling aspects of the paleoenvironmental succession. The interest expressed by many other Quaternary researchers - too many to name - is greatly appreciated. Thank you all.

Discussions with all of my colleagues in the Quaternary research group of the University of Alberta served to focus mu ideas. Dave Liverman and Doug Schnurrenberger performed yeoman service in reading early drafts of this thesis.

Constructive reviews of this thesis by John England, Fran Hein, Frank Schwartz and Jean-Serge Vincent led to an improved final product.

I had the benefit of excellent field and laboratory assistance from John Kulig, Margaret Dittrick, and Olwen Wirth. Navigation of the Peel River was made possible through the river sense of Simon Snowshoe, of Fort McPherson. The camaraderie of our field camps and laboratory teams made the most tedious work enjoyable.

Jenni Blaxley typed this opus, after she overcame the obsolescent scruples of the author. Linda Halsey provided drafting services.

Finally, my deepest thanks to Heather Whitehead, who tolerated all the vagrancies and temperamental outbursts of the author, and somehow maintained patience and still remains my wife.

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Catto, Norman R., "Quaternary Sedimentology and Stratigraphy, Peel Plateau and Richardson Mountains, Yukon and N.W.T."

p. 4 l. 2 Reads: "in the spring and are generally by mid-day"
Should Read: "in the spring and are generally high by mid-day"

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I. Introduction

The Richardson Mountains, Peel Plateau, and Peel Plain are located in the northeastern Yukon and northwestern District of Mackenzie, N.W.T. (Figure 1). With the completion of the Dempster Highway, the region has become more accessible and therefore more open to development. At the same time, public interest and concern for developmental, environmental, and social issues has led to a heightened awareness of northwestern Canada. Investigation of the Quaternary geology of the region provides basic information necessary to any evaluation of the potential and problems of development.

The objectives addressed by this research are:

1. To describe and analyse the sedimentology, mineralogy, and stratigraphy of Quaternary deposits exposed along the major rivers of the region, and throughout the area,
2. To correlate these exposures with surface geomorphological features, especially meltwater channels and moraines marking former ice-front positions,
3. To describe the geological and climatic/vegetative/faunal environments present during the Quaternary,
4. To establish a chronology of Quaternary events and environments in the region,
5. To correlate this chronology with previously investigated deposits and described events in the Bonnet Plume Basin to the southwest, the Bell, Old Crow, and

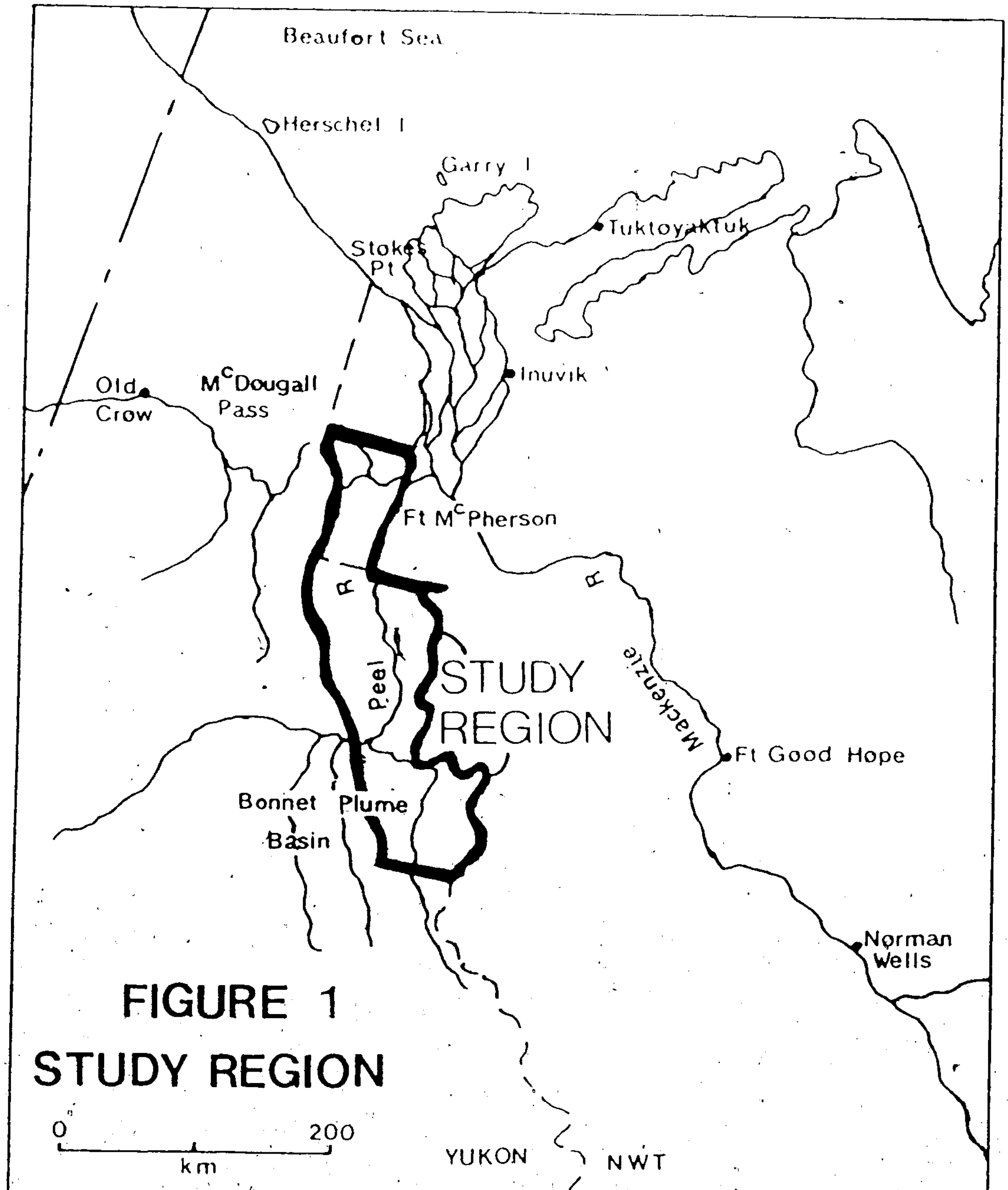


FIGURE 1
STUDY REGION

Bluefish Basins to the west, the Mackenzie Delta and Valley to the east, and the Beaufort Sea coastal plain to the north.

Understanding of the Richardson Mountains - Peel Plateau Peel Plain area would enable events in these surrounding regions to be linked.

A. Location of the Study Region

The study region consists of the Peel Plateau, excluding the Bonnet Plume Basin but including the western Peel Plain, and the eastern slopes of the Richardson Mountains, in the Yukon Territory and District of Mackenzie, Northwest Territories (Figure 2, in pocket).

The total area encompassed is approximately 24,000 km², with 17,500 km² in Yukon and 6,500 km² in District of Mackenzie. The only permanent settlement within the region is Fort McPherson, N.W.T., with a winter population of approximately 800. Road access to the village is provided by the Dempster Highway, the only public roadway in the region.

B. Climate

The study region has a continental climate characterized by cold, dry winters and short summers. Mean July temperatures throughout the area range from 12° to 15°C, while January mean values are -29°C at Inuvik and as low as -35°C in the interior (Burns 1974). The mean annual temperature ranges from -10°C at Inuvik to -15°C in the

Richardson Mountains. Temperatures undergo a rapid rise in the spring and are generally by mid-day throughout the summer, reaching 25°C and commonly exceeding 28°C (Britton 1957). Temperatures decline equally rapidly in the autumn. These seasonal temperature fluctuations produce a short frost-free season, varying from 37 days at Canoe Lake, west of Mount Gifford (Lambert 1968) to 50 days at Inuvik (Burns 1974).

Precipitation along the eastern slope of the Richardson Mountains and on the Peel Plateau is 250 mm/yr, of which 120-130 mm is rain and the remainder snow (Burns 1974). Snowfall usually is confined to the summer and autumn months, as little precipitation of any type falls between early November and March. West-facing slopes are often free of snow two weeks or more before east-facing slopes (Lambert 1968). During the growing season, precipitation is usually slightly in excess of plant transpiration (Mather and Thornthwaite 1956), which is indicative of semi-arid conditions.

The prevailing wind in the Mackenzie Valley area is northerly (Burns 1974). The strongest winds and storms tend to move southerly to southeasterly up the Peel River valley from the Mackenzie Delta and Beaufort Sea throughout May - August. These winds gradually swing to move easterly by mid-autumn, effectively bypassing the western part of the Peel Plateau. In the Richardson Mountains, the prevailing winds are westerly and northwesterly (Lambert 1968), driven by thermal convection during the summer over the Old Crow Basin. During the winter, radiation cooling establishes a cold, dry high pressure system over the central Mackenzie Valley, effectively preventing any precipitation in the Peel Plateau (Environment Canada 1973).

C. Permafrost

Both continuous and discontinuous permafrost, as defined by Brown (1967) and Washburn (1980) are found in the study region. The border between the zones of continuous and discontinuous permafrost trends east-west across the Peel Plateau, south of Fort McPherson and Arctic Red River (Brown, 1967; Zoltai and Tarnocai, 1975) as illustrated on Figure 3. In the Richardson Mountains, permafrost is essentially continuous. The estimated thickness of the permafrost in the continous zone of the Peel Plateau is in excess of 300 m (Judge 1973). Active layer thicknesses vary from 0.5 m in fine clay sediments to 4 m in some coarse fluvial gravels.

Throughout the discontinuous zone, permafrost is widespread. Thus, frozen and unfrozen sites can exist in proximity to each other. Whether a particular site in the discontinuous zone is permafrost-free depends upon the microclimate, vegetation, and sediment type.

Streams, through their heat energy can create zones of talik along meandering reaches (Crampton 1979), as well undercut bases of bluffs leaving an overhanging bank above

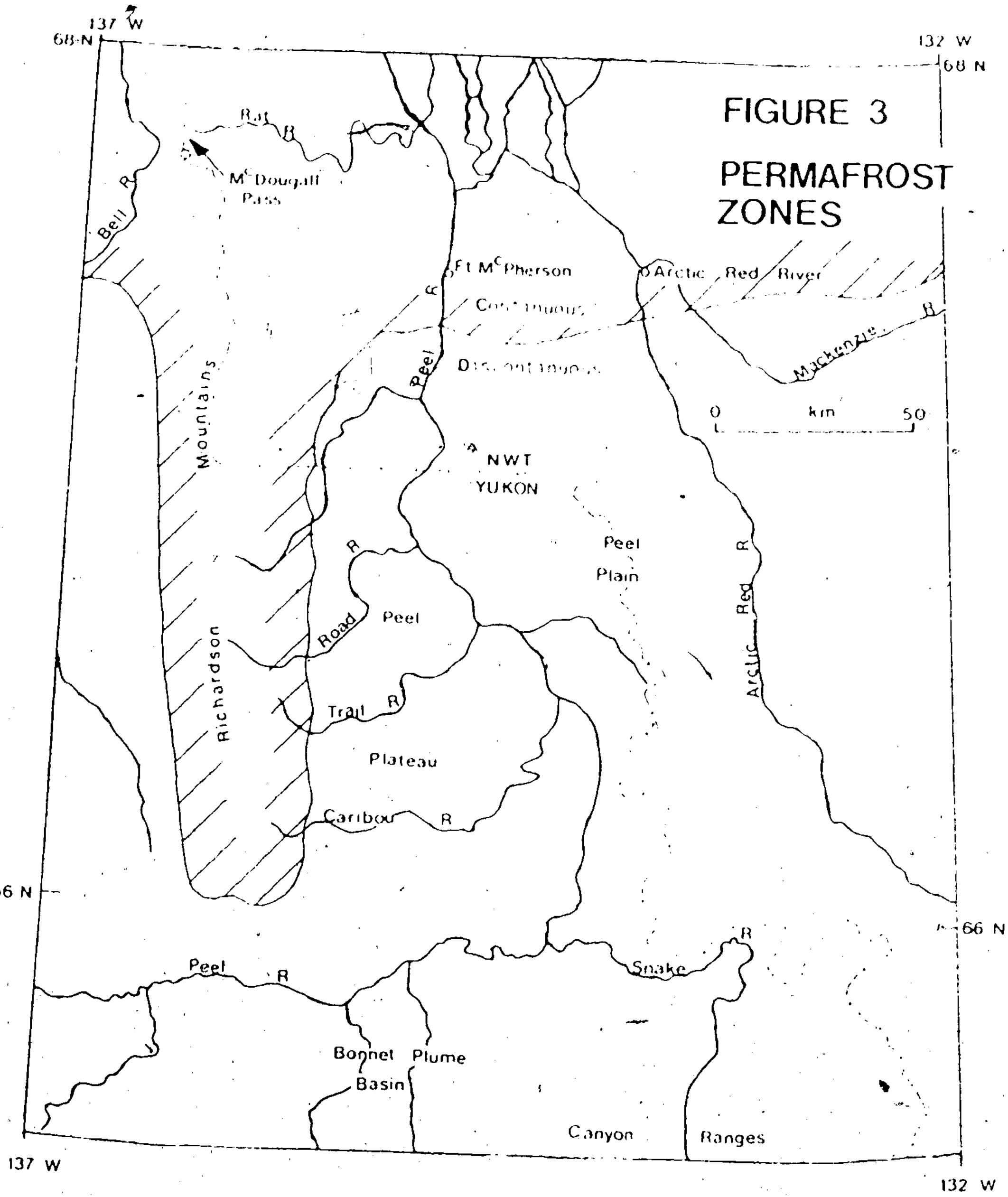


FIGURE 3

PERMAFROST ZONES

the river level. The undercut zones are here referred to as thermal niches. Any thawing of the permafrost, either through climatic change or removal of vegetation, creates the potential for mass movement. Numerous retrogressive thaw-flow features are present throughout the Peel Plateau, as a result of permafrost degradation.

In addition to the thaw-flow features, the permafrost regions of the Peel Plateau are characterized by extensive fields of small-scale cryoturbation features and earth hummocks; lenses of segregated ice developed in silty sediments; and thermokarst lakes, with both orthogonal and curved margins. Polygonal ground is rare on the Plateau, but does occur on the Peel Plain, especially in drained thermokarst lake basins. Permafrost features within the Richardson Mountains are confined to thaw-flows, gelifluction lobes, stone stripes and nets, felsenmeer, and scattered earth hummocks on flat-topped ridges and plateaux.

D. Soils

Soil classification in northern Canada is still evolving and no one system has gained general acceptance. The Canadian Soil Survey Committee (1978) recognizes the Cryosolic Order of soils, defined as those soils formed in either mineral or organic materials that have permanently frozen ground (permafrost) either within 1 m of the surface (Static and Organic Cryosols, respectively), or within 2 m of the surface if more than one-third of the pedon has been strongly cryoturbated (Turbic Cryosols). Most of the soils in the study region can thus be classed as Cryosols. At some locations, the active layer has deepened sufficiently (more than 2 m) to enable Eutric Brunisolic soils to form. This most commonly occurs along entrenched river valleys. Regosols are also present in a variety of environments, such as floodplains and mountain slopes.

E. Vegetation

The study region is divided into two major vegetation zones. The majority of the Peel Plateau and Peel Plain lie within the Boreal Forest Region of the lower Mackenzie Valley (Rowe 1972), although much of the territory is devoid of trees. The dominant communities are open evergreen sclerophyll/ deciduous forests (Hettinger et al 1973), with open forests of Picea mariana (black spruce) and Larix laricina (tamarack) with associated Betula glandulosa (Shrub birch) and Alnus crispa (alder). The understory vegetation is dominated by Ericaceae (heath) and Sphagnum. Low lying areas are occupied by dwarf scrub communities, with Cyperaceae (sedge), Sphagnum, and scattered Alnus, Picea mariana and Salix (willow). Well-drained sites support Picea glauca (white spruce) and Betula papyrifera (white birch).

The higher areas of the Peel Plateau, and the foothills of the Richardson Mountains and Canyon Ranges, are termed the Alpine Forest-Tundra Region (Rowe 1972). The communities are predominantly orthophyll scrub, with Salix, Betula *glandulosa*, Ericaceae, *Saxifraga* and *Artemisia* (Hettinger *et al* 1973). Unstable slopes are colonized by a discontinuous cover of grasses. This zone represents a transitional area between the boreal forest of the Peel Plateau and the open ground in the higher part of the Richardson Mountains.

II. Bedrock Geology

The bedrock geology of the Peel Plateau and Plain is characterized by complexly deformed Precambrian, Palaeozoic, and early Mesozoic sediments overlain by flat-lying to gently folded Cretaceous rock. In the Richardson Mountains and Canyon Ranges, the older sediments are exposed. The ages of the rock units range from Aphebian to Tertiary (Figure 4 and Table 1).

A. Peel Plateau and Plain

The oldest rock exposed in the Peel Plateau and Plain are the shale, siltstone and sandstone units of the Devonian Imperial Formation (Norris 1981c, d, e). Unconformably overlying the Imperial Formation are several Lower Cretaceous siltstone and sandstone units (Table 1). The youngest of these is the Arctic Red Formation, the areally most extensive unit in the Peel Plateau and Plain.

East of the Snake River, and in the Trevor Range, the Arctic Red Formation is overlain by the resistant siliceous sandstone of the Middle Cretaceous Trevor Formation (Norris 1982a, b). The youngest unit within the Peel Plateau is the Upper Cretaceous Boundary Creek Formation consisting of marine shale and siltstone with bituminous beds and bentonitic strata (Norris 1981 b).

Figure 4

Bedrock Geology

Tertiary Clastic Rocks

Middle Cretaceous Trevor Formation Limestone

Lower Cretaceous Clastic Rocks (primarily Arctic Red Formation Siltstone and Shale

Lower Mesozoic and Palaeozoic Clastic Rocks

Early Palaeozoic Rocks (primarily Road River Group, Argillite, Chert, and Breccia)

Cambrian Dolomite and Limestone

Hadrynian and Helikian Rocks (primarily Rapitan Group Jasper-Hematite and Diamictite, and Katherine Group Quartzite and Argillite

Major Faults

Many units with outcrops of limited aerial extent are not shown. The delineation of units is based on Norris (1981 a, b, c, d, e; 1982 a, b)

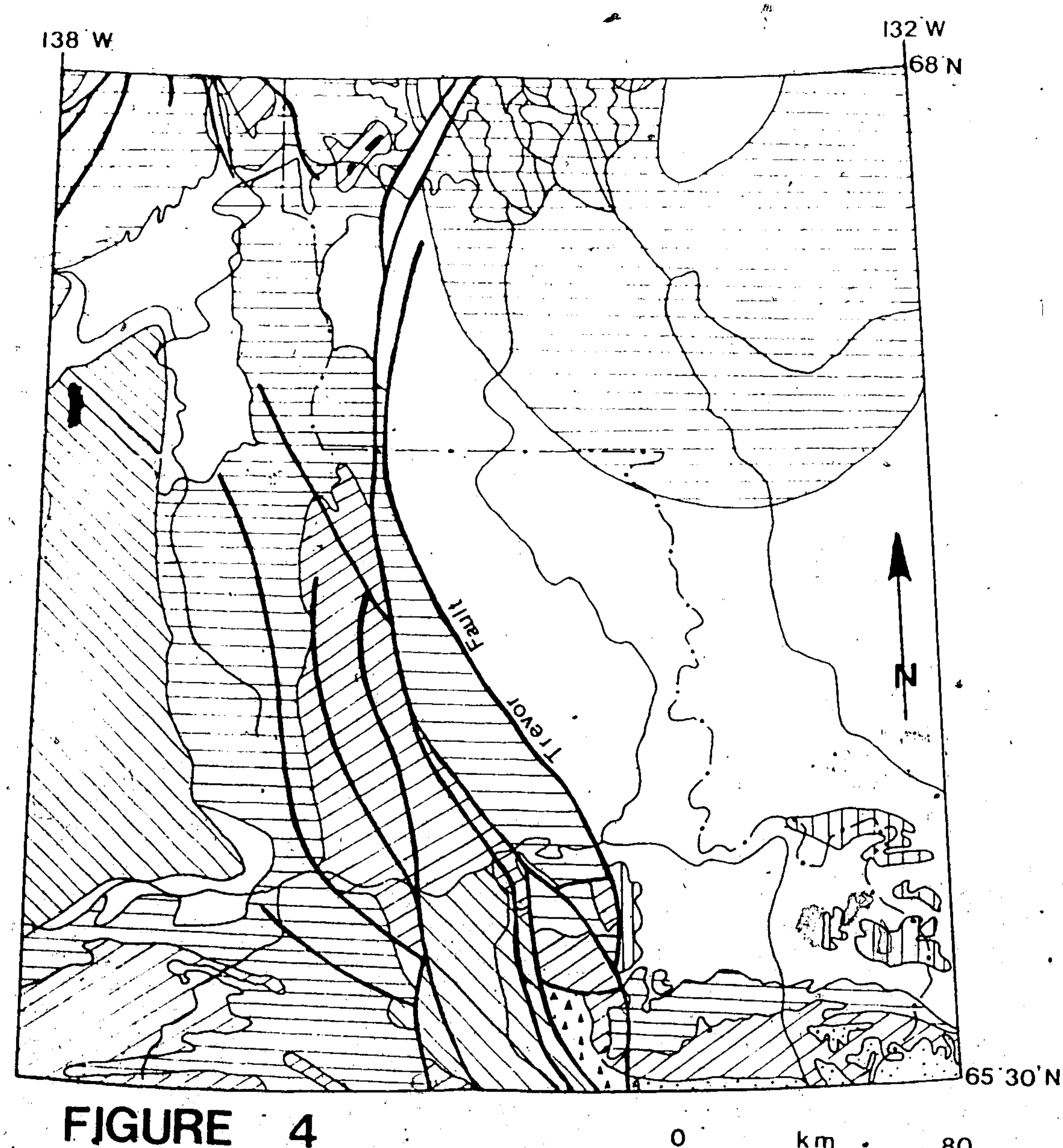


FIGURE 4

BEDROCK GEOLOGY

Table 1: Principal bedrock units, study region.

Unit	Age	Lithology
Bonnet Plume Formation	Tertiary - Late Cretaceous	Clastics
Eagle Plain Fm.	Late Cretaceous	Clastics
Boundary Creek Fm.	Late Cretaceous	Shale/ siltstone
Trevor Fm.	Middle Cretaceous	Sandstone
Arctic Red Fm.	Early Cretaceous	Siltstone/ shale
Martin House Fm.	Early Cretaceous	Sandstone
Rat River Fm.	Early Cretaceous	Clastics
Mt. Goodenough Fm.	Early Cretaceous	Shale/ sandstone
Martin Creek Fm.	Early Cretaceous	Sandstone
McGuire Fm.	Early Cretaceous	Sandstone
North Branch Fm.	Early Cretaceous-Jurassic	Sandstone
Husky Fm.	Early Cretaceous-Jurassic	Shale/ siltstone
Bug Creek Gp.	Jurassic	Clastics
Shublik Fm.	Triassic	Limestone
Boundary Creek Fm	Permian	Clastics
unnamed	Mississippian	Shale/ siltstone
Tuttle Fm.	Mississippian	Clastics
Imperial Fm.	Devonian	Shale/ siltstone
Canol Fm.	Devonian	Shale
Road River Gp.	Early Devonian - Late Cambrian	Shale, siltstone, limestone, breccia, argillite

Mt. Kindle Fm.	Early Silurian-Late Ordovician	Dolomite/ shale
Franklin Mtn. Fm.	Early Ordovician-Late Cambrian	Dolomite/ shale
Slats Creek Fm.	Middle Cambrian	Sandstone
Illtyd Fm.	Early Cambrian	Dolomite/ sandstone
Rapitan Gp.	Hadrynian	Diamictite, jasper-hematite, shale
Katherine Gp.	Helikian	quartzite, dolomite, shale
Quartet Gp.	Aphebian	Quartzite/ argillite

B. Richardson Mountains

The boundary between the Richardson Mountains and Peel Plateau is marked by a series of major normal faults, which trend northwest in the southern part of the region and northeast in the Rat River area. (Young et al 1976; Norris 1981b, d).

The Yukon (southern) area of the Richardson Mountains is characterized by a north-south trending anticlinal structure, with normal faults parallel to the limbs (Hume 1954; Norris 1981c, d). The exposed core of the anticline is occupied by Middle Cambrian sandstones and siltstones. This unit is flanked by the Upper Cambrian-Lower Devonian Road River Group, a thick sequence of graptolitic shale, argillite, siliceous and cherty siltstone and limestone, and cherty sharpstone breccia. Clasts of these units are frequently a major component of both preglacial and Quaternary sediments.

The Road River Formation is overlain by a succession of shales, sandstones, and conglomerates which extend east to the Trevor Fault (Hume 1954; Norris 1981d). To the east of the Caribou Fault, outcrops of Upper Triassic skeletal limestone (Shublik Formation) are present (Norris 1981d). In the vicinity of the headwaters of the Caribou River, faulting has resulted in the elevation of a small area of the Aphebian Quartet Group (argillite and quartzite) and the Lower Cambrian Illtyd Formation (micrite and microsparite) to the surface. Clasts derived from the carbonate units are readily recognizable in Quaternary sediments.

The N.W.T. (northern) area of the Richardson Mountains is characterized by an abundance of normal faults and folds. The oldest units definitively recognized (Norris 1981a, b) are Cambrian dolomites and limestones exposed in the headwaters of the Bell River, and Road River Group units east of Scho Creek. A small exposure of Helikian quartzite in the Barrier River drainage basin was tentatively identified by Norris (1981b). The Devonian Imperial Formation outcrops along Sheep Creek and the Dempster Highway.

Scattered outcrops of marine Permian clastic rocks are present east and north of Scho Creek and in McDougall Pass. Overlying these units unconformably are quartzites, sandstones, siltstones and shales of the Jurassic Bug Creek Formation (Poulton et al 1982), and clastic rocks of the Jurassic/Lower Cretaceous Husky Formation and the North Branch Formation (Jeletzky 1980; Norris 1981a, b). These formations outcrop ~~extensively~~ north of the Rat River, and in McDougall Pass and its approaches.

Lower Cretaceous clastic rocks outcrop along the Yukon/N.W.T. border and north of the Rat River.

C. Canyon Ranges

The Canyon Ranges south of the Peel Plateau are characterized by east-west trending fold axes and faults. The oldest rock exposed is the Aphebian Quartet Group, composed of siliceous argillite, quartzite, breccia, slate, and phyllite (Norris 1982a, b). It is overlain unconformably by the Helikian Katherine Group (quartzite), Little Dal Formation (dolomite and algal stromatolites), and the Hadrynian Rapitan Group (diamictite, sandstone, siltstone, and banded jasper-hematite containing copper sulphides) (Eisbacher 1981; Norris 1982a, b; Carriere et al 1981). Although the Precambrian rocks do not outcrop extensively in the western Canyon Ranges, clasts derived from these units are common constituents of clastic sediments exposed on the Peel Plateau.

The northern part of the Canyon Ranges is composed of cherty dolomite of the Upper Cambrian/Lower Ordovician Franklin Mountain Formation, marine dolomite of the Upper Ordovician/Lower Silurian Mount Kindle Formation, and exposures of the Road River Formation and Devonian clastic sediments (Norris 1982b).

The Trevor Range is characterized by a north-south trending system of anticline and faults (Norris 1982a). In the core of the range, Ordovician and Silurian carbonate strata are exposed, flanked on the west by shale exposures of the Canol Formation and on the east by the Trevor Fault. Fault-bounded exposures of the Road River Group are present along the western slopes of the Trevor Range.

D. Adjacent Areas

Clasts derived from the Upper Cretaceous Eagle Plain Formation (sandstone, siltstone, shale and lignite) exposed in the Bell Basin to the west of the Richardson Mountains indicate eastward transport when found in Quaternary sediments in the study region, as do clasts derived from the Upper Cretaceous/ Tertiary Bonnet Plume Formation exposed in the Bonnet Plume Basin (conglomerate, sandstone, siltstone, mudstone, and lignite). In addition, sediments derived wholly or partially from continental ice sheets originating in the Canadian Shield to the east and southeast, contain clasts of acidic igneous and medium and high grade metamorphic rocks.

III. Physiography

The physiography of northern Yukon and northwestern District of Mackenzie has been discussed previously by Bostock (1948, 1970), Douglas et al (1963), and Rampton (1982). Four main physiographic units - Richardson Mountains, Peel Plateau, Peel Plain, and Canyon Ranges - are present in the study area. Figure 5 depicts the distribution of the major physiographic units.

A. Richardson Mountains

The Richardson Mountains consist of a series of north-south trending ridges, often smoothed and flat-topped, separated by broad valleys. The mountain belt is 35 km wide in the headwater area of the Caribou River, and narrows to 15 km in the vicinity of the headwaters of the Vittrekwa River. The belt widens to the north of the Vittrekwa River, and is 80 km wide along latitude 68° N.

Relief throughout the range is generally less than 500 m, and the topography is more similar to that of foothills rather than a major mountain chain (Bostock 1948). The highest point in the vicinity of the study region is Mount Sittichinli (1574 m), located 700 m west of the Yukon/N.W.T. border (67° 11' N, 136° 15' W). The lowest point along the drainage divide is Summit Lake in McDougall Pass (320 m) at 67° 42' N, 136° 27' W.

Along the eastern edge of the mountains bordering the Mackenzie Delta, an escarpment 700-750 m high marks the

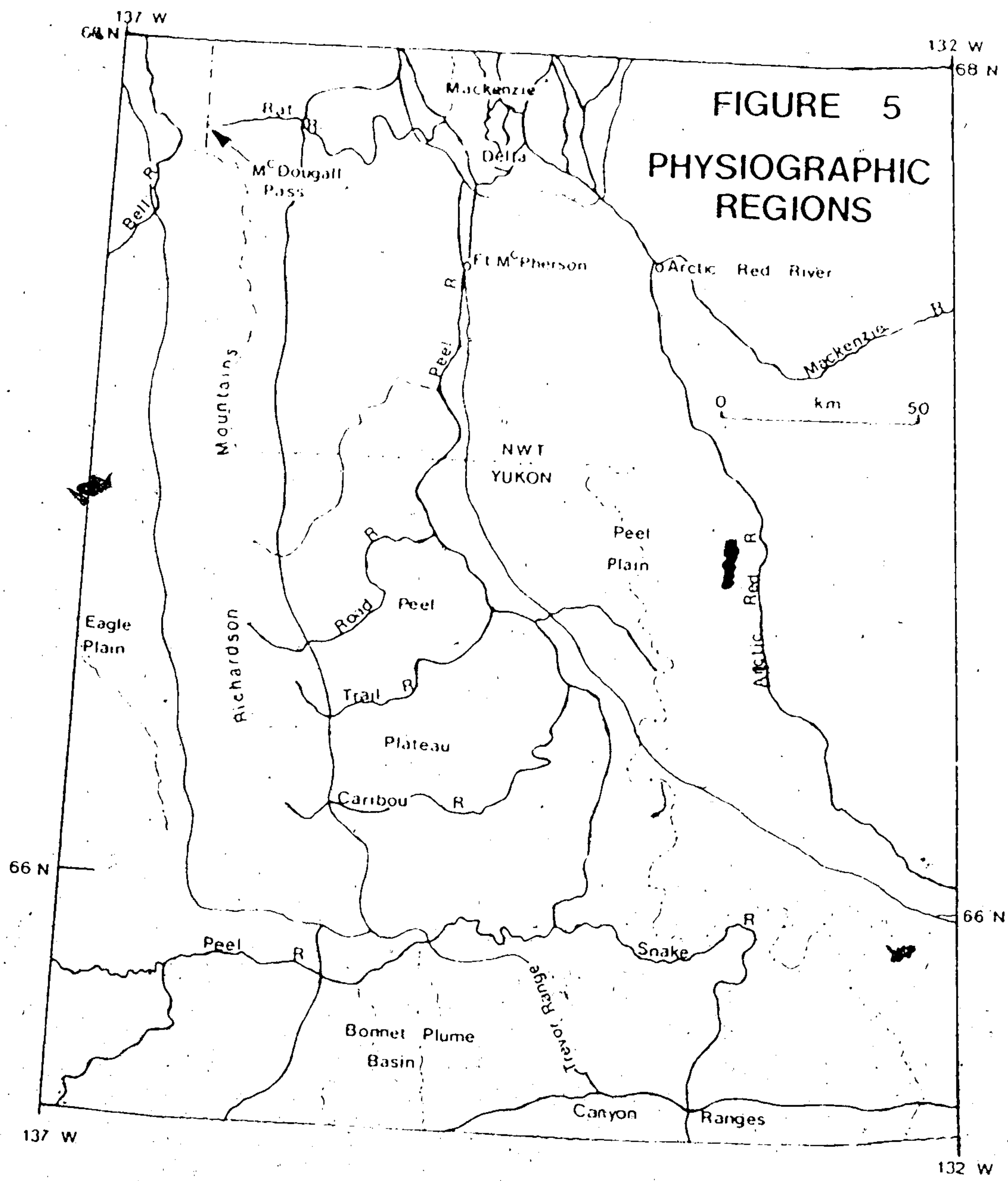


FIGURE 5
PHYSIOGRAPHIC REGIONS
137 W
68 N
132 W
66 N
Mackenzie
Delta
Rat R
McDougall Pass
Bell R
Ft McPherson
Arctic Red River
Mackenzie R
0 km 50
Peel R
NWT
YUKON
Peel Plain
Arctic Red R
Mountains
Richardson
Eagle Plain
Peel
Road R
Trail R
Plateau
Caribou R
Peel R
Snake R
Trevor Range
Bonnet Plume Basin
Canyon
Ranges

mountain front. The highest summit in the vicinity of the escarpment is Mount Goodenough (981 m).

Ridges and drainage divides generally trend north-south or northwest-southeast in the southern part of the Richardson Mountains. The orientation of the ridges and divides reflects the structural geology of the region and the stratigraphic succession. Resistant quartzites and sandstones form the ridges, whereas weathering of siltstones and shales and faulting are responsible for creating the valleys. The asymmetry of the fold limbs of the bedrock structure is reflected in the asymmetric ridge profiles, with the eastern sides commonly being steeper than the western sides. The major river valleys are broad but not deeply entrenched, and trend east-west. The drainage patterns are dendritic or sub-parallel, with local examples of parallel, pinnate, and trellis drainage present adjacent to major or sharp ridges. Relief throughout the headwaters of the Caribou, Road, Trail, and Vittrekwa Rivers is generally less than 400 m.

The dominant feature of the northern part of the Richardson Mountains is McDougall Pass, a U-shaped valley 1 km wide at Summit Lake. Local relief in the pass is 350 m. Ridges in the area of the pass are aligned northeast-southwest, approximately parallel to the valley and are controlled by fold orientations and sedimentary lithology. Truncated bedrock spurs are evident along both valley walls.

North of McDougall Pass, the ridge systems vary in alignment from northwest-southeast to northeast-southwest. The major valleys are aligned north-south and are occupied by moderately entrenched streams. Relief throughout the region does not exceed 550 m. In the eastern ranges, the valleys are broader and are generally aligned east-west or northeast-southwest, and no conspicuous north-south valleys are present. Drainage in both areas is dendritic to subparallel.

B. Peel Plateau

The Peel Plateau, as defined by Bostock (1948), is a broad, gently eastward-sloping surface ranging from 650 m elevation ASL in the west to 100-150 m elevation ASL in the east. It is approximately triangular in shape and is bordered on the west by the Richardson Mountains and Porcupine Plateau, on the south by the Canyon Ranges and Wernecke Mountains, and on the east by the Peel Plain (Figures 2 and 5). The north-south distance across the plateau is approximately 250 km, and the east-west distance along 66° N is approximately 150 km. The Bonnet Plume Basin is considered to be part of the Porcupine Plateau, and the summits of the Trevor Range mark the boundary between the physiographic units (Bostock 1970).

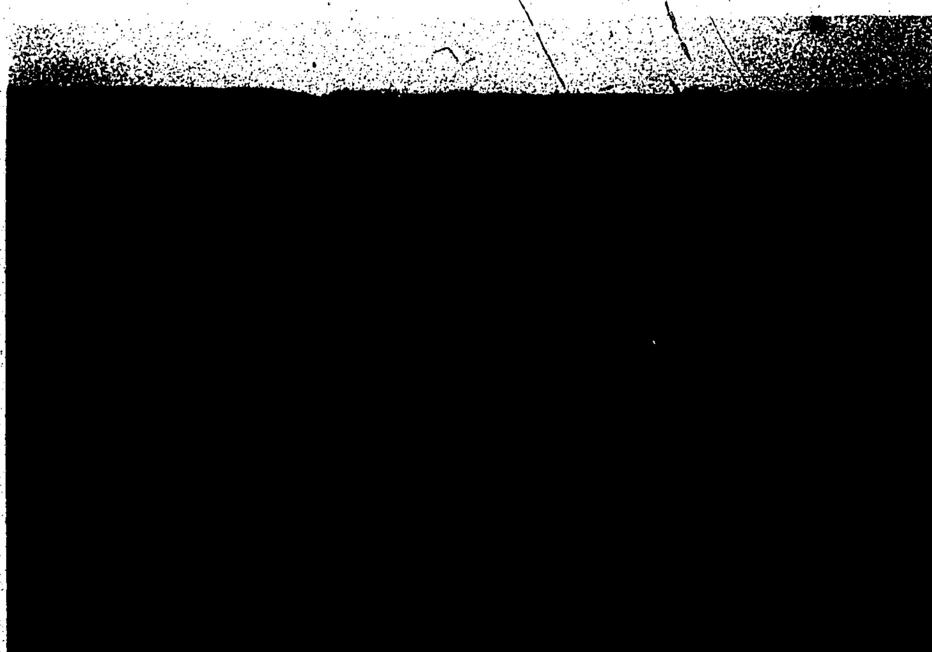
The surface of the plateau is generally even, with relief of less than 100 m. The Peel River and its major tributaries, however, have cut deep canyons into the

Plate 2 (top)

The Peel Plateau, looking east from the Richardson Mountains near the headwaters of the Vittrekwa River. The Dempster Highway is visible in the middle ground.

Plate 3 (bottom)

The Caribou River has incised the Peel Plateau to a depth of 50 m 40 km southwest of its confluence with the Peel River. East of the Trevor Fault, the depth of incision reaches 150 m. This reach is intermediate between a braided and a braided-meandering stream. The Richardson Mountains are visible in the background.



Cretaceous Arctic Red Formation shale and siltstone. The Peel River is entrenched 200-300 m below the surface of the plateau throughout its lower reaches below its confluence with the Wind River, and its position cross-cutting the plateau at almost 90° azimuth to the regional slope is suggestive of an origin as a glaciomarginal meltwater channel (Hughes et al 1981). A second major channel is located 20-25 km west of and parallel to the Peel River between the Arctic Circle and the Rat River (Hughes 1972).

In the northern part of the plateau, the major river channels are entrenched to a maximum depth of 230 m as far west as the foothills of the Richardson Mountains. In the central part the streams are entrenched only in the Arctic Red Formation shale and siltstone exposed east of the Trevor Fault. In both areas, the drainage pattern is dendritic to subparallel, and the streams generally flow to the east and northeast. Local areas of deranged drainage with abundant shallow lakes are present north and southeast of the Caribou River.

The southern part of the Peel Plateau is traversed by the Snake River, a braided-meandering stream. South of 65° 40' N and 65° 58' N, the river has eroded a canyon to 140 m depth through weak Cretaceous Arctic Red Formation shale and siltstone. The valley in this reach is generally wide, (maximum 3.2 km), and broad meanders are common. In both areas, the tributary pattern is predominantly dendritic and well-integrated, with few lakes. Remnants of meltwater channels are present, generally oriented towards the west (Hughes 1972). The most prominent of these originates from the Cranswick River (Hughes 1972).

The Snake River abruptly turns to the west at 65° 58' N, 133° 09' W. Downstream from this point, the valley is narrow (1 km), deep (190 m), and characterized by few, tight meanders and longitudinal bars. High-level meltwater channels parallel the Snake River along the southern side of the valley. The northern side is flanked by undulating glacial deposits (Hughes 1972), with a relief of 10-25 m.

The Peel Plateau is separated from the Bonnet Plume Basin to the west by the Trevor Range, a north-south trending series of flat-topped sandstone ridges. Drainage is subparallel to dendritic. The highest point of the range is approximately 817 m. Although Bostock (1948, p. 30) suggested that the summits of the range remained unglaciated, no evidence of nunatak development or local ice-marginal positions around summits was documented by Hughes (1972).

C. Peel Plain

The Peel Plain is part of the Interior Plains region (Douglas et al 1963; Bostock 1970). It forms the northeastern part of the study region. Here, it is characteristically low (maximum elevation 95 m), with little relief. The plain slopes gently to the northeast, towards the Arctic Red and Mackenzie Rivers, but away from the adjacent Peel River. It is underlain by Cretaceous and Devonian siltstone and shale.

East of Fort McPherson, and along the Sainville River, the plain contains many shallow, discontinuous abandoned channels formed by westward and northward meltwater flow (Hughes 1972). Shallow, unoriented lakes and deranged drainage also characterise this area.

The Satah River basin, in contrast, is largely devoid of lakes, and the drainage here is integrated. The general trend of the streams is northwesterly, and shallow abandoned channels oriented in this direction are common. Throughout this region, the drainage divide between the Peel and Arctic Red Rivers is very low, and it is not marked by discernable topographic features.

D. Canyon Ranges

The Canyon Ranges are a rugged series of sharp northwest-southeast trending ridges, separated by broad, U-shaped valleys emanating from the Backbone Ranges to the south (Bostock 1948). Small areas of plateaux are present. The ranges are composed of a variety of sedimentary lithologies of Mesozoic, Paleozoic and Precambrian ages. The highest point within the study region is 1750 m (65° 29' N, 132° 39' W), and local relief exceeds 800 m.

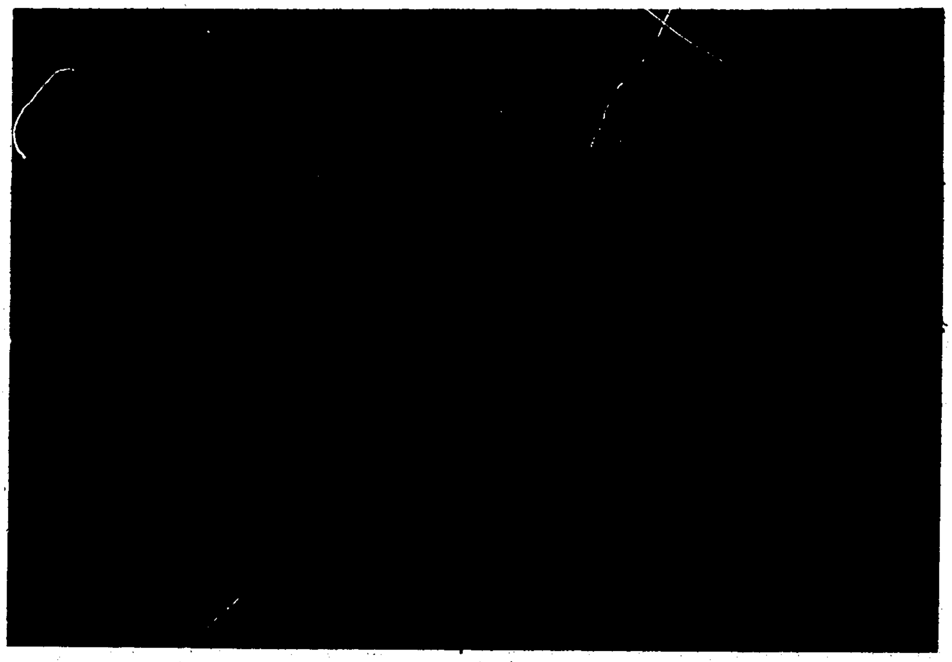
The minor streams have incised V-shaped valleys with dendritic to subparallel drainage systems. Cirques are developed only on the highest peaks in the southern part of

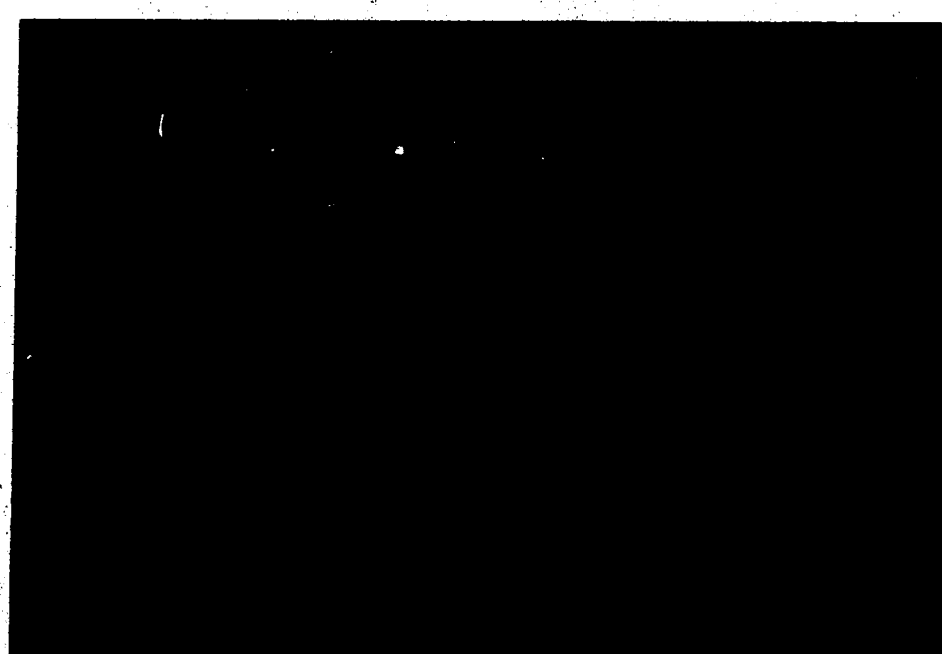
Plate 4 (top)

The northern flank of the eastern approach to McDougall Pass. The strata in the foreground are Permian sandstones and the Jurassic Husky Formation shales and sandstones. The Cretaceous Bug Creek Group clastic rocks and the Mount Goodenough Formation shales and siltstones are visible in the background. The strata dip gently to the northwest, and are exposed along the southern limb of a northeast-southwest trending anticline.

Plate 5 (bottom)

The Canyon Ranges, in the headwater areas of the Snake River (approximately 65 °N, 133 ° W). The strata exposed represent the Helikian Rapitan Group, Katherine Group, and Quartet Group. Alpine glaciation has formed numerous cirques, aretes, and horns (background), and the terrain resembles a ploughed field.





the range and are generally oriented towards the northeast. The headwalls and margins of these cirques are extensively weathered and in some cases rockfalls have occurred.

The Canyon Ranges are separated from the Peel Plateau to the north by a west-east oriented channel which borders the mountain front. This channel was carved by westward-flowing meltwater, and has a maximum depth of 50 m and a maximum width of 200 m. The absence of Canadian Shield erratics south of this channel indicates that it represents the maximum known position of ice originating from the Canadian Shield (Hughes 1972; Hughes et al 1981).

IV. Previous Work

The Richardson Mountains were discovered and named by Sir John Franklin in 1825 for Sir John Richardson, the expedition's physician, biologist and geologist. Franklin was also responsible for naming the Peel River after Sir Robert Peel, a prominent British politician interested in Canadian affairs, but Franklin's 1825 party and subsequent expeditions did not investigate either the mountains or the river. Exploration of the area commenced in 1839 with the establishment of Fort McPherson by Hudson's Bay Company traders John Bell and Murdoch McPherson. Bell followed the Peel River south to its confluence with the Snake River, and then proceeded up the Snake River to its headwaters at approximately 64° 130' W(Isbister 1845). Both Bell and Isbister considered the Snake River to represent the main channel. It was not until 1893, when Count V.E. de Sainville reached the Bonnet Plume Basin, that the course of the Peel River was accurately delineated (de Sainville 1898).

The Hudson's Bay Company established LaPierre House (1843-44) in the Bell Basin to the west, and supplied it via an overland route (the "Winter Portage") from Fort McPherson. This route was traversed and described by Alexander Murray (1848). In 1872, trader James McDougall travelled through the low divide between the Bell and Rat drainage basins, subsequently named McDougall Pass. The routes of these early explorations are depicted in Figure 6.

Figure 6

Early Explorations in the Peel Plateau-Richardson Mountains Region

B---Bell, 1839

A---Murray, 1844

M---McConnell, 1888

O---Ogilvie, 1888

S---de Sainville, 1893

C---Camsell, 1905

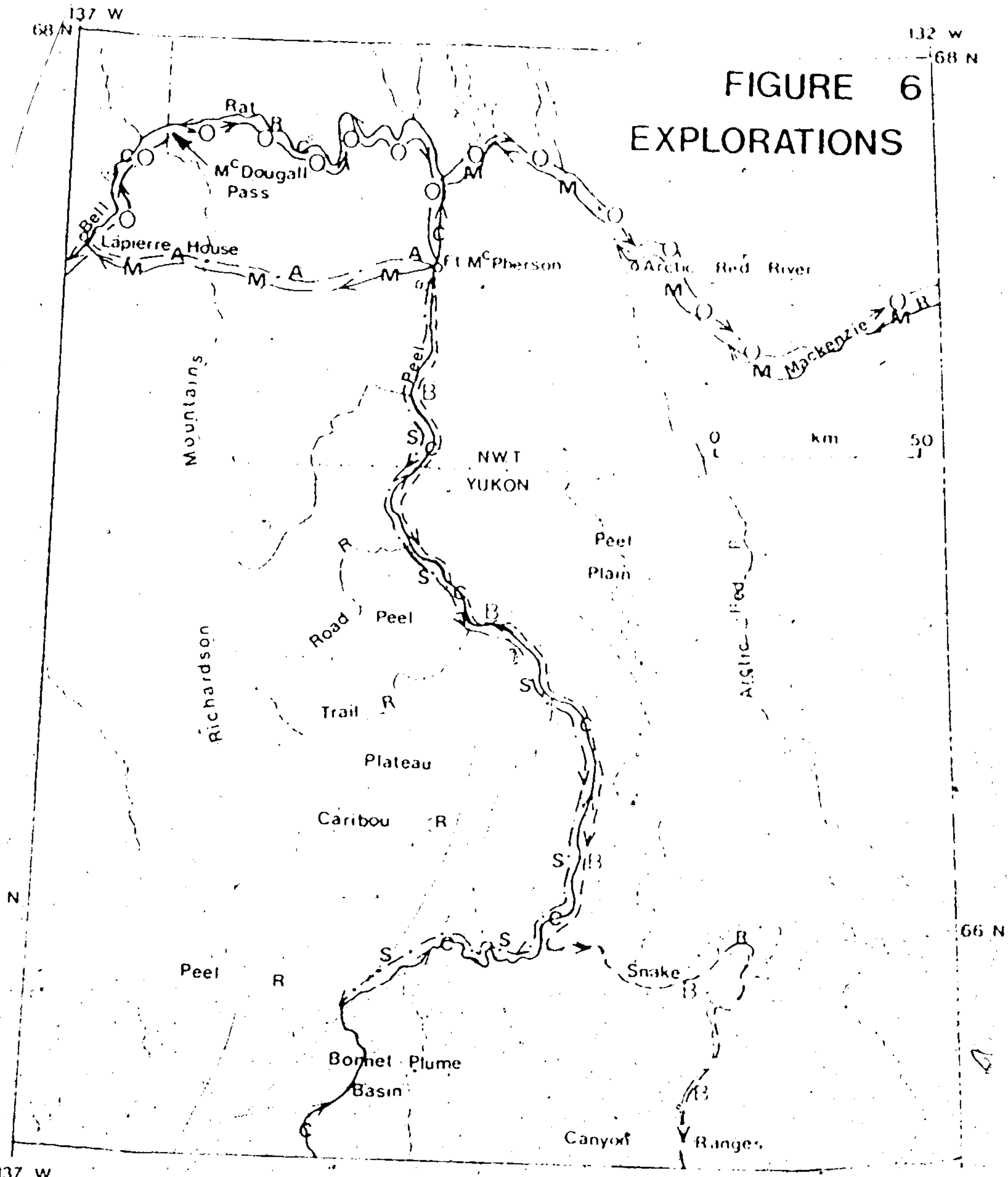


FIGURE 6
EXPLORATIONS
137 W
68 N
132 W
68 N
Rat
McDougall Pass
Bell
Lapierre House
Ft McPherson
Arctic Red River
Mackenzie
Peel
Mountains
NWT
YUKON
Peel Plain
0 km 50
Road Peel
Trail R
Richardson
Plateau
Caribou R
Arctic Red
66 N
66 N
Snake
Peel R
Bonnet Plume Basin
Canyon
Ranges
137 W
132 W

Both these routes across the Richardson Mountains were more thoroughly explored during 1888. R.G. McConnell (1891) travelled westward along the "Winter Portage", noting the presence of a terrace and abundant earth hummocks west of Fort McPherson. He also noted that glacial deposits were not present west of the Richardson Mountains, nor were there any signs of alpine glaciation. William Ogilvie (1889) travelled east along the Bell River, through McDougall Pass, and to Fort McPherson along the Rat River (which he referred to as the Trout River), noting the general characteristics of the sediments exposed. During the Klondike gold rush of 1897-99, both these trails were used by hopeful miners. In addition, many travelled south along the Peel River into the Bonnet Plume Basin, and from there moved southwest across the Wernecke Mountains to the Yukon drainage system.

Camsell (1906; Camsell and Malcolm 1921) travelled downstream along the Peel River from the Bonnet Plume Basin (Figure 6). Camsell recognized glacial features in the Peel Plateau, but believed that all of the ice originated from the Mackenzie Mountains to the south. He also reported an unsorted deposit on the flanks of Mount Goodenough, and suggested that an alpine glacier was responsible for its formation (Camsell 1906, p.40). These tentative conclusions, based on scattered observations along the Peel River, were accepted without reservation until 1948 (eg. Flint 1947; Wickenden 1947).

Bostock (1948) described the physiography of the northern Canadian Cordillera, including the current study region. The delineation of the ice flow directions and marginal positions was greatly aided by aerial photography, which enabled a clear picture of the area away from the major rivers to be gained for the first time. In addition, improved understanding of the bedrock geology (Hume 1954) enabled recognition of the source areas of the clasts contained in the deposits. Bostock showed that the Richardson Mountains were devoid of alpine glacial features, and that the glaciers emanating from the Backbone Ranges were confined to the main valleys in the Canyon Ranges to the north. Ice flow in the Bonnet Plume Basin and the southern Peel Plateau was to the west, while in the northern Peel Plateau north to northwest prevailed.

Hughes began conducting investigations of the Peel Plateau-Richardson Mountains area in 1962. The subsequent reports (Hughes 1963, 1970, 1972; Hughes and Hodgson 1972; Hughes et al 1972, 1973a, 1973b, 1974) described the basic surficial geology of the region and delineated ice-front positions. Based on these studies, Hughes and his co-workers recognized two distinct continental glacial events. The most extensive event terminated along a line extending from the western end of McDougall Pass (Little Bell River) southward, slightly to the east of the drainage divide between the Beaufort Sea and the Bering Sea along the crest of the Richardson Mountains, to the headwaters of Mountain Creek

(Figure 7). The margin extended as a series of lobes up Doll Creek, the Peel River, and Hungry Creek. To the south, the margin bordered the Wernecke Mountains in the west and the Canyon Ranges in the east. Along this border, tills deposited by ice sheets originating in the Canadian Shield are interstratified with till deposited by glaciers debouching from the Wernecke and Mackenzie Mountains. The Richardson Mountains were not of sufficient elevation to produce local glaciers at ~~any~~ time during the Quaternary, except possibly immediately surrounding the summit of Mount Millen (Hughes 1972), west of the study area.

On the bases of ^{14}C dates of >38,600 B.P. (GSC-120, Dyck and Fyles 1964, p. 173) and >38,300 B.P. (GSC-204, Dyck et al 1965, p. 38) obtained from organic detritus and wood underlying the oldest till identified along the lower Rat River at section HH 62-107 (67° 39' N, 135° 28' W), the maximum advance was considered to predate the Late Wisconsinan (Hughes 1972). An additional corroborating date was obtained from a terrace on the Snake River within the glaciated area, approximately 135 km east of the maximum ice-front position. There, organic rich silt from a terrace not overlain by till was ^{14}C dated at >31,000 B.P. (GSC-181, Dyck et al 1965, p. 38). Wood believed to be from this locality was subsequently ^{14}C dated at >35,000 B.P. (GSC-2956, Hughes et al 1981).

A second advance delineated by Hughes (1972) formed a belt of moraines, meltwater channels, and outwash plains to

Figure

Ice Front Positions

Maximum position recognized by Hughes (1972). This position was considered to be of pre-Late Wisconsinan age by Hughes (1972). Hughes _et al_ (1981) considered this position to be of Late Wisconsinan age. Rampton (1982) correlated the Buckland position (in the Yukon Coastal Plain) to position 1, and considered it to be pre-Late Wisconsinan.

2---Late Wisconsinan position recognized by Hughes (1972). This position was considered to represent the maximum extent of Late Wisconsinan glaciation by Hughes (1972) and Rampton (1982). Hughes _et al_ (1981) considered this position to represent a recessional phase of Late Wisconsinan glaciation.

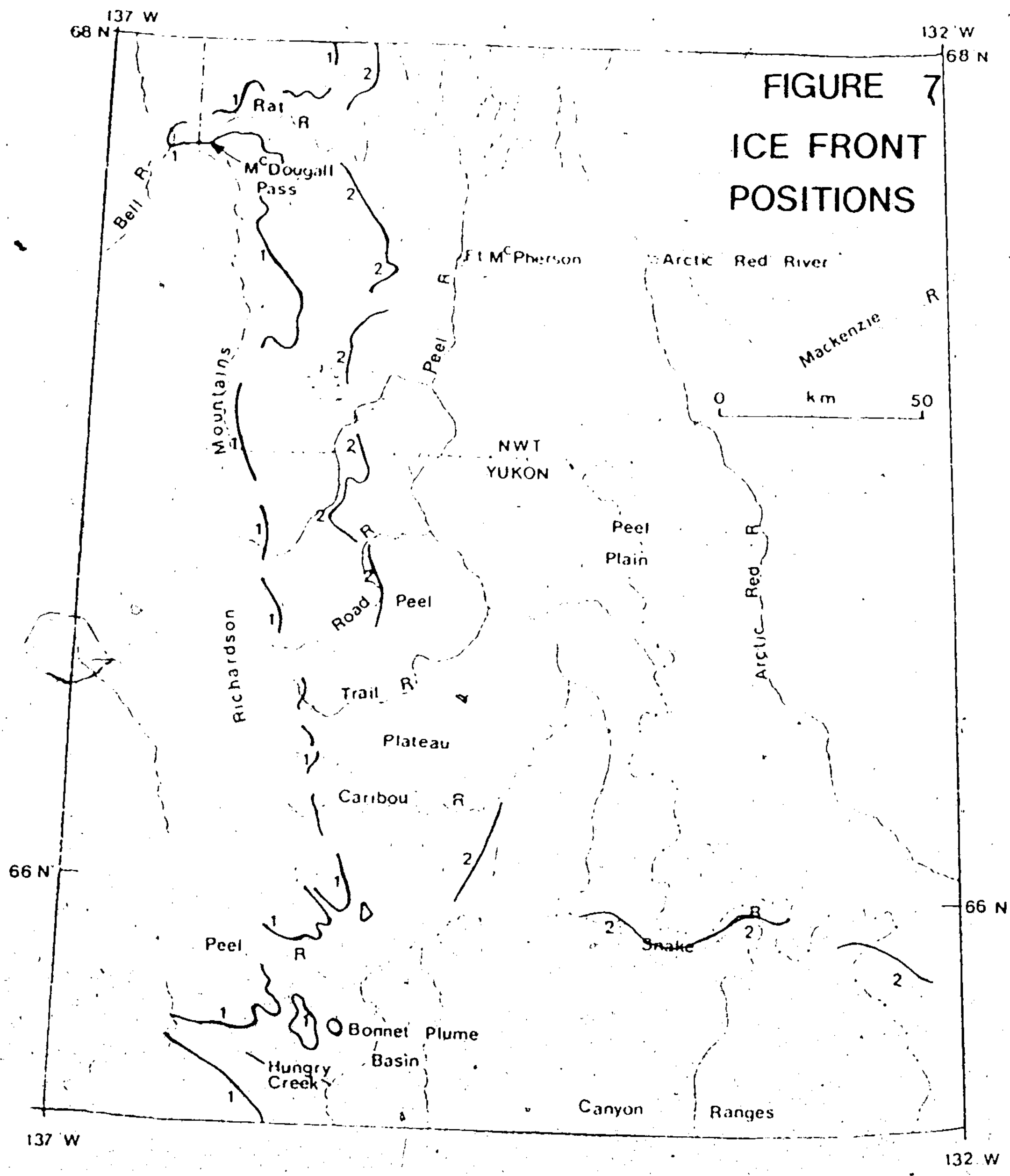


FIGURE 7
ICE FRONT POSITIONS
137 W
68 N
132 W
68 N
Rat R
Bell R
McDougall Pass
Ft McPherson
Arctic Red River
Mackenzie R
Peel R
0 km 50
NWT
YUKON
Mountains
Richardson
Peel Plain
Arctic Red R
Peel
Road
Trail R
Plateau
Caribou R
66 N
66 N
Peel R
Snake R
Bonnet Plume Basin
Hungry Creek
Canyon Ranges
137 W
132 W

the east of the maximum glacial event (Figure 7). This limit can be easily recognized from the northern boundary of the study area to the Trail River (in the vicinity of the location where it crosses the Arctic Circle). From this point south, the ice margin is uncertain. In the southern Peel Plateau, the margin is represented by a belt of moraines along the northern side of the Snake River (Hughes, 1972). This second glacial advance was considered to be of Late Wisconsinan age.

This chronological outline, involving one pre-Late Wisconsinan and one Late Wisconsinan event, is supported by data obtained from the Yukon Coastal plain by Rampton (1970, 1974, 1982). Rampton recognized two glacial limits: the Buckland, along the northern flank of the Buckland Hills and Richardson Mountains, considered to be Early Wisconsinan; and the Late Wisconsinan, confined to the Mackenzie Delta and the vicinity of Sitidgi Lake. The Buckland Glaciation limit corresponds to the maximum limit of glaciation recognized along the Yukon coastal Plain by Hughes (1972).

At Stokes Point, along the Beaufort Sea, Buckland drift is overlain by autochthonous peat dated at 22,400± 240 B.P. (GSC-1262, Lowden and Blake 1976). Marine pelecypod shell fragments from proglacial pitted outwash sediments on Garry Island and the vicinity of Kendall Island have been dated at >35,000 (GSC-562) and >37,000 B.P. (GSC-690) respectively (Lowdon _et al_ 1971). Both these localities are believed to lie within the margins of a glacial event correlated to the

Buckland limit (Rampton 1982).

The dates obtained are all subject to question, however. Autochthonous peat from the Stokes Point area has recently been dated at 7,510±100 B.P. (GSC-3747). This suggests that the 22,400±240 BP date may be incorrect, although some uncertainity remains because of the difficulty in identifying the precise peat unit dated initially (J-S Vincent, Geological Survey of Canada, personal communication, 1985). The shell fragments may have been affected by the inclusion of detrital carbonate, which would produce anomalously old ^{14}C dates. In addition, the presence of the fragments in proglacial outwash suggests that some transportation of the material has occurred (J-S. Vincent, Geological Survey of Canada, personal comminication 1985). Therefore, the shell fragment dates may not reflect the time of deposition of the sediment. The uncertainties surrounding all the age determinations of the Buckland limit and correlative limits in the Mackenzie Delta are thus considerable, and are sufficient to prevent any confident assessment of the age of these events.

Hughes' (1972) Late Wisconsinan maximum position is considered to represent a retreat phase of the Buckland ice by Rampton (1982), and is termed the Sabine limit. This position is not recognized along the eastern flanks of the Richardson Mountains. Peat overlying till of the Sabine phase has been dated at 14,400±180 B.P. (GSC-1792, Rampton 1982), but this date is considered to be considerably younger than the time of ice retreat from this position by Rampton (1984, personal communication). The Late Wisconsinan glacial maximum in the Mackenzie Delta Sitidgi Lake area is considered by Rampton (1982, p.25) to have occurred c. 13,000 B.P..

Hughes et al (1981) have revised this chronological sequence, however, based upon research in the Bonnet Plume Basin. Wood from a silt unit underlying till attributed to the maximum Quaternary glacial advance in the Hungry Creek area has yielded a ^{14}C date of 36,900±300 B.P. (GSC-2422). Because the section is located in the western Bonnet Plume Basin, adjacent to the maximum ice-front position, the date if reliable suggests that the maximum ice advance in this region occurred since 36,900±300 B.P.. As the ice-front position to the west of this site is correlated to the maximum position west of McDougall Pass (Hughes 1972), a Late Wisconsinan age is implied for the maximum glacial limit throughout the region, and by extension, for the Buckland limit of Rampton (1982). This event is termed the "Hungry Creek Glaciation" (Hughes et al 1981). The ice front positions to the east are considered to represent retreat phases during the Late Wisconsinan. Some of these may be correlative to the Late Wisconsinan maximum position recognized in the Mackenzie Delta by Rampton (1982), and to the Sitidgi Lake Stade and Tutsieta Lake Phase, c. 13,000 B.P., recognized by Hughes (Personal Communication, 1985).

Lacustrine and fluvial sediments from a event predating the Hungry Creek Glaciation were also recognized by Hughes et al (1981) in the Bonnet Plume Basin. The lacustrine sediments are interpreted to have been deposited in a lake in the Bonnet Plume Basin impounded by a glacial ice front located east of the basin. This event has subsequently been termed the "Deception Glaciation" (Hughes, Personal Communication, 1984). No surface expression of the Deception Glaciation maximum has been recognized.

Thus, a conflict exists between the chronological interpretations of Hughes et al (1981) and Rampton (1982). The Quaternary glacial maximum in the Bonnet Plume Basin is considered to be Late Wisconsinan by Hughes et al (1981), whereas the Quaternary glacial maximum in the Yukon Coastal Plain is considered to pre-date the Late Wisconsinan by Rampton (1982). Both Hughes et al (1981) and Rampton (1982, 1985 personal communication) accept the continuity of this ice limit, as originally proposed by Hughes (1972). Shell fragments from sections within the limit have been dated at >35,000 and >37,000 BP, supporting a pre-Late Wisconsinan age for the maximum, but these dates are not wholly reliable. Conversely, wood from a sub-till silt unit in the Bonnet Plume Basin dated at 36,900±300 BP suggests that the maximum glacial advance in this area postdates 36,900 BP. The Late Wisconsinan maximum position recognized by Rampton (1982) in the Mackenzie Delta is considered to represent a pause in the retreat of Late Wisconsinan ice by Hughes

(1985, personal communication). Resolution of the problem requires either re-assessment of the ^{14}C dates previously obtained, or revision of the correlation of ice-front positions between the Bonnet Plume Basin and McDougall Pass areas, or consideration of the nature of glacial and related activity in the area. It is apparent that some of the ^{14}C dates determined previously may not accurately indicate the time of deposition of the sediments. It is also conceivable that the assumption of a single synchronous Quaternary glacial maximum in the Bonnet Plume Basin, along the eastern flank of the Richardson Mountains, and in the Yukon Coastal Plain may be in error.

V. Methodology

A. Field Research

Field research conducted in 1981 and 1982 involved the systematic description, sampling and tentative identification of all Quaternary sedimentary units, as well as several unconsolidated sedimentary units of unknown Cenozoic age, at 66 localities (Figure 2 (in pocket) and Appendix 1). Sampling intervals varied from ~1 per 10 m (colluvium and some river gravels) to 1 per 5 cm or less (peats, organic-rich sediments, and fine-grained material). Fabric analyses were performed on all diamictons tentatively identified in the field as tills, on all diamictons of uncertain origin, and on selected colluvial, lacustrine, and fluvial diamictons and gravels. Current directions (both aqueous and aeolian) were obtained from clast orientations, sedimentary structures, and erosional surfaces. At each site, the vegetation assemblage at the surface was described.

In addition to the investigations at the principal sections, reconnaissance surveys were conducted throughout the region by helicopter, and sections observed <u>en route</u> briefly described (Figure 2). The Peel River was traversed by boat from Fort McPherson to Peel River Canyon, downstream from Bonnet Plume Basin, in the 1981 field season. Numerous modern fluvial, colluvial and aeolian exposures were also investigated (Appendix 2).

B. Laboratory Research

Laboratory investigations were conducted to establish the sedimentological and mineralogical parameters of the deposits and to determine the region's palaeoenvironmental history. The first objective entailed textural analyses, conducted utilizing the hydrometer-sieve (ASTM 1964) and pipette (Griffiths 1967) methods. Mineralogical analyses of the sand, coarse silt, and granule size fractions were performed using binocular and petrographic microscopes. Granule, pebble, and cobble lithologies, as well as roundnesses, sphericities, and shapes were determined for all coarse-grained samples. Carbonate concentrations of the silt and clay fractions were initially determined using the Chittick technique (Dreimanis 1962), but inconsistencies in the results limited the reliability of the values obtained.

Palaeoenvironments were elucidated by analysis of palynomorphs, diatoms, mollusca, arthropoda, and plant macrofossils. Details of the methodology are discussed in Appendix 5.

C. Statement of Approach

Quaternary deposits in the study region include glacial, lacustrine, fluvial, aeolian, colluvial, and paludal sediments. Several sedimentary processes commonly act to produce different lithologic units in the same or similar environments.

The approach taken in this thesis involves the division of the region into three main areas: the southern part, surrounding the Snake River and the upper Peel River; the central part, consisting of the terrain in the vicinity of the Caribou River and Brown Bear Creek; and the northern part, consisting primarily of the Rat River Valley and McDougall Pass. These areas are illustrated in Figure 8.

Within each area, the sedimentology of the Quaternary deposits is discussed in terms of the environmental and stratigraphic settings of the units. The units in each area will be considered in stratigraphic succession. Discussion of the three areas will be followed by a summary chapter correlating the stratigraphic units and geological events.

Many distinct environments, as represented by a multitude of sedimentary beds and units, have influenced the development of the study region during the Quaternary. The spatial and temporal separation of the sedimentary events make the formulation of general models applicable to the entire region difficult in many cases. Formal designation of each distinct sedimentary unit as a "facies" (cf. Middleton 1978; Walker 1984) would result in a vast number of such designated facies. Most of these facies would be characteristic of only a single specific environment within the study region. Although analysis of individual sedimentary exposures was greatly facilitated by using an approach based on facies recognition, formal designation of a large number of facies encompassing many environments

Figure 8

Divisions of the Study Region

The study region has been divided into three areas for convenience of discussion. These are: the Snake River watershed and Upper Peel River area; the Caribou River-Brown Bear Creek area; and the Rat River-McDougall Pass area.

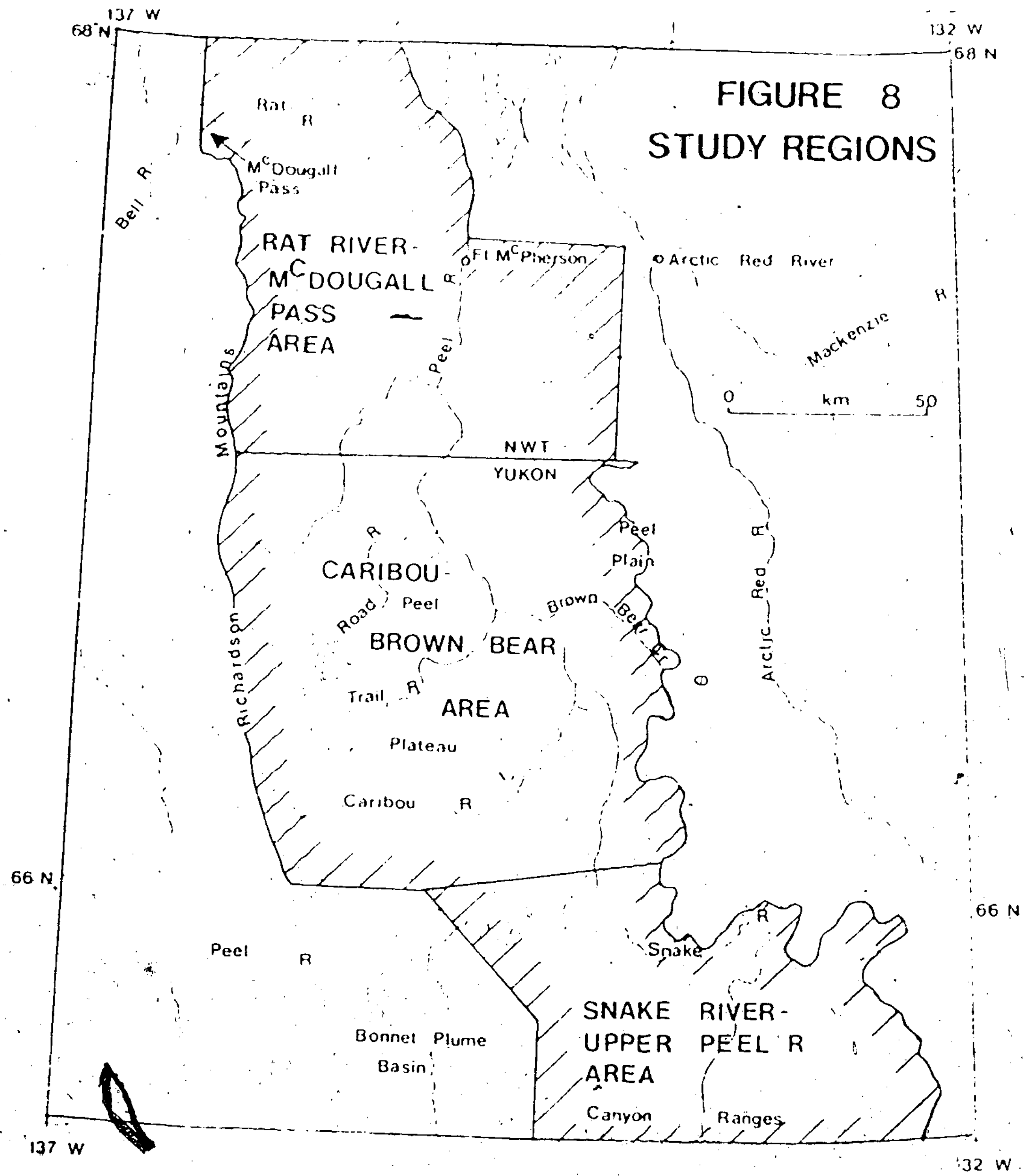


FIGURE 8
STUDY REGIONS
137 W
68 N
132 W
68 N
Rat R
M^c Dougall Pass
Bell R
RAT RIVER-
M^C DOUGALL
PASS
AREA
Ft M^c Pherson
Peel R
Arctic Red River
Mackenzie R
0 km 50
Mountains
NWT
YUKON
Peel Plain
CARIBOU-
BROWN BEAR
AREA
Road Peel Trail R
Plateau
Caribou R
Richardson
Brown Bear R
Arctic Red R
66 N
66 N
Peel R
Snake R
Bonnet Plume Basin
SNAKE RIVER-
UPPER PEEL R
AREA
Canyon Ranges
137 W
132 W

would induce confusion and hamper stratigraphic interpretation over an area as broad as the study region. Consequently, the approach adopted in this thesis involves the analysis of sedimentary environments and the formulation of environmental models without formal designation of facies.

VI. Snake River Watershed and Upper Peel River Area

A. Introduction

The Snake River watershed and the upper Peel River area represent the southern part of the study region. Exposures of sediment are largely confined to the valley walls of the two major streams, although minor outcrops are present in the surrounding plateau.

Glacigenic deposits blanket the Peel Plateau in this region, and extend westward into the Trevor Range and southward into the Canyon Ranges. Geomorphic, sedimentologic, and lithologic characteristics serve to delineate the fluvial, glacial, and lacustrine episodes which affected the region. The distribution of surficial deposits in the area is illustrated in Figure 9.

The sedimentary record encompasses the Quaternary period from pre-glacial time to the Holocene. In the following sections of the thesis, the deposits will be discussed in chronological order.

B. Preglacial Sediments

At several locations along the Snake River, shale and siltstone of the Arctic Red Formation is overlain by gravel and sand beds (Figure 10). These coarse deposits are dominated by orthoquartzite, feldspathic sandstone, siltstone and chert clasts. Clast surfaces are frequently stained with iron and managanese oxides, and coated with

Figure 9

Quaternary Geology, Snake River Upper Peel River

(after Hughes 1972; Hughes et al 1972, 1973, 1974; and observations by the author)

Holocene Marsh and Bog, standing water and organic sediments

Holocene Fluvial Deposits along major active streams, predominantly sands and gravels

Holocene Alluvial Fan Deposits, gravels and diamictons

Holocene and Pleistocene Abandoned Channel Deposits, sand, clay, and organic sediments

Pleistocene Glacial Diamicton, with streamlined features showing direction of glacial flow

Pleistocene Glacial Diamicton, hummocky terrain

Pleistocene Glacial Diamicton, with organic sediment blanket or veneer

Pleistocene Glaciofluvial Deposits, sand and gravel

Pleistocene Eskers, sand and gravel

Pleistocene Glacial Deposits, scattered erratics and thin discontinuous diamicton cover

X Principal Exposures discussed in text

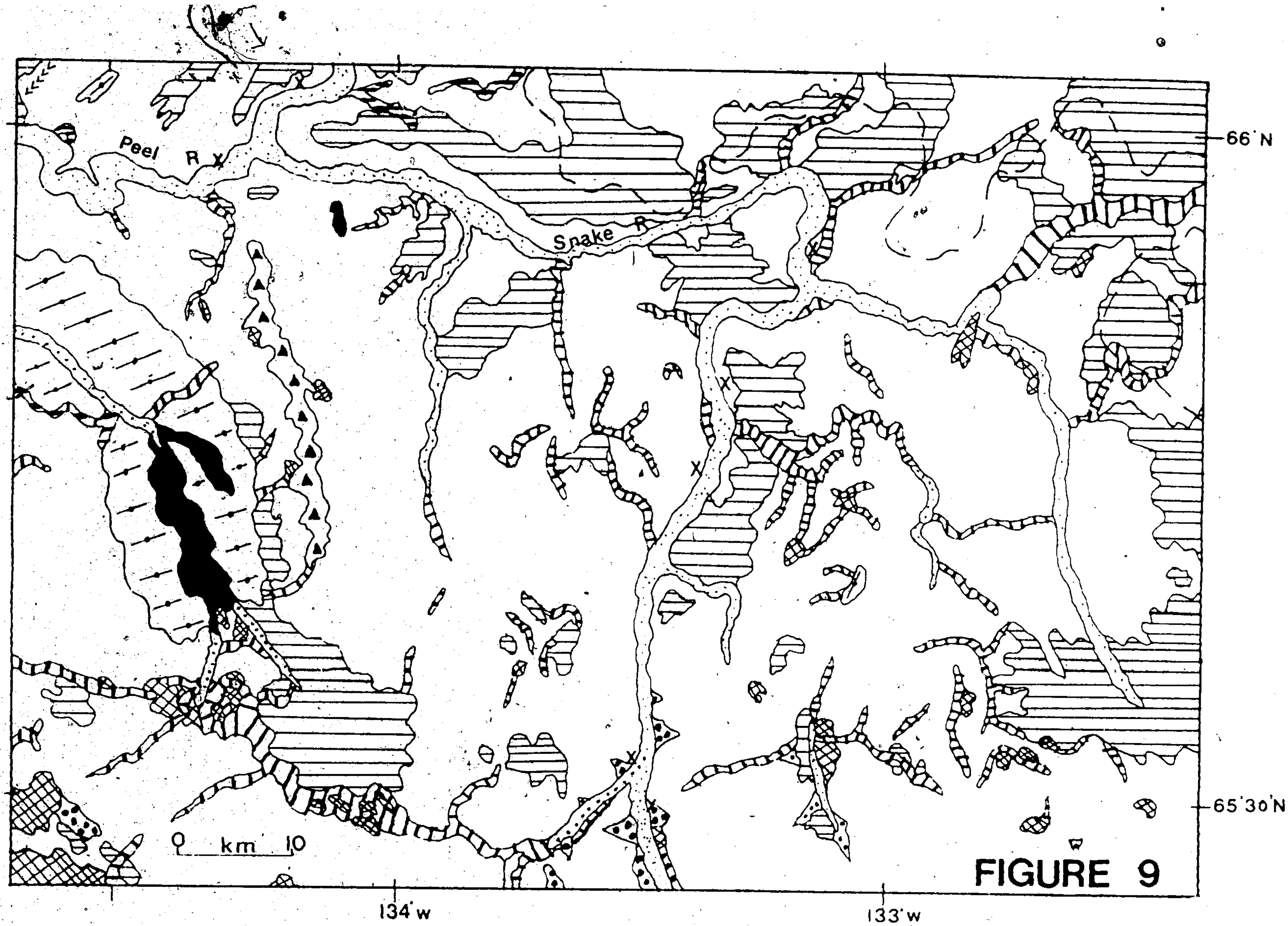


FIGURE 9

Figure 10

Preglacial Sediments, Snake River Area

Preglacial Sediments are exposed at the sections indicated with an X. See text for discussion.

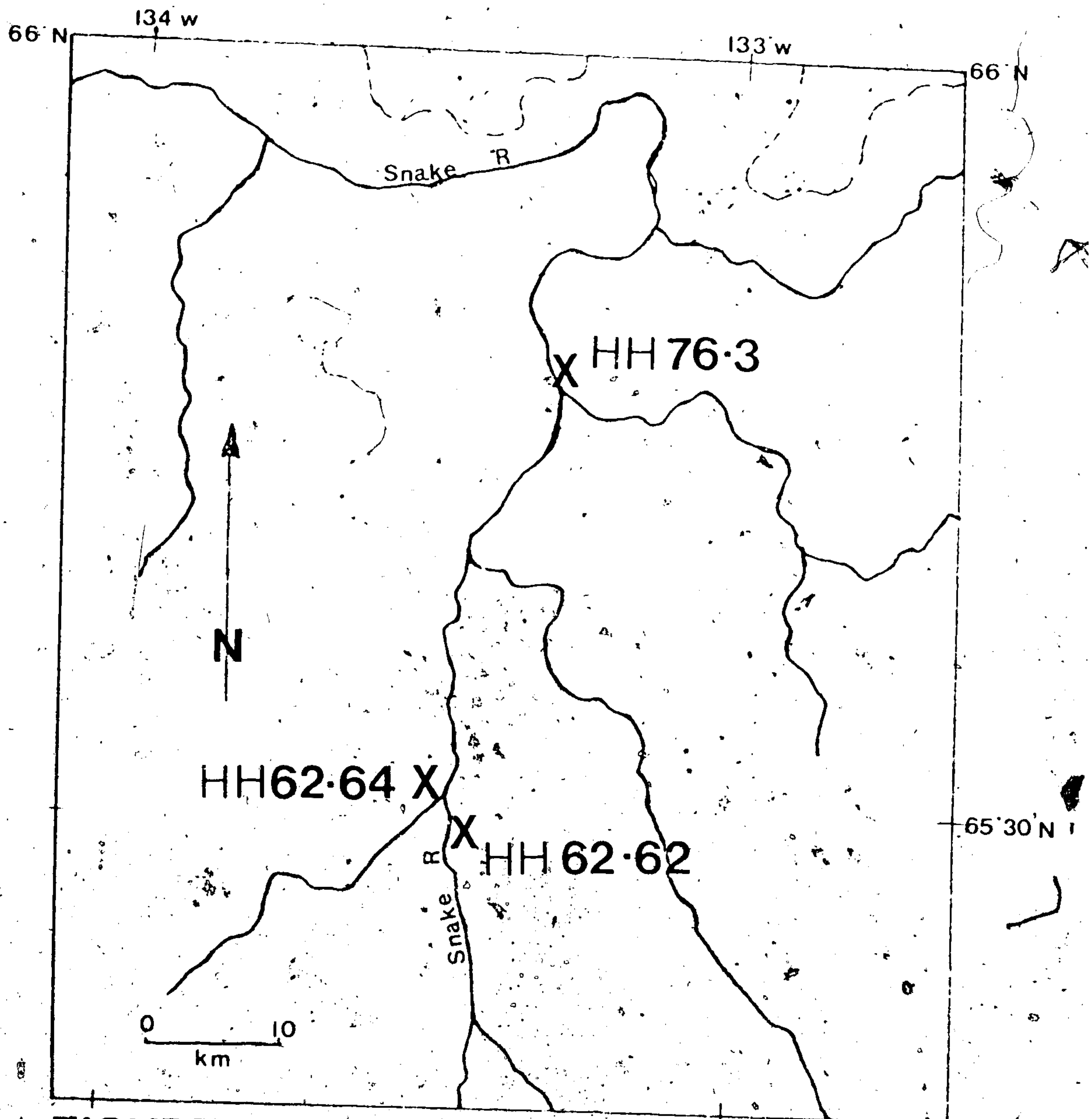

66°N
134°W
133°W
66°N
Snake R
HH 76·3
N
HH62·64
HH 62·62
Snake R
65°30'N
0
10
km

FIGURE. 10

calcium and magnesium carbonate. The sediments contain no granitic or medium and high-grade metamorphic clasts, and all of the clasts analysed were derived from rock units exposed within the Snake River drainage basin (Table 2). These mineralogical data indicate that the sediments were deposited prior to the initial continental glaciation to affect the Snake River region. The stratigraphic position of the sediments, directly overlying bedrock and overlain by glacigenic deposits and fluvial deposits containing granitic clasts, also suggests a preglacial age (Figure 11).

The sedimentary sequences are dominated by structureless or crudely horizontally stratified pebble and granule gravel beds, with or without finer matrix material. These sequences are similar to those identified in longitudinal bars developed in modern alpine braided streams (Smith 1974; Boothroyd and Ashley 1975, dominant facies of Scott type stream of Miall 1977, 1978) and also resemble longitidinal bar and channel sediments developed in the modern braided streams of the Richardson Mountains (Appendix 4). Isolated beds with tabular foreset cross-strata composed of coarse sand, granules and pebbles are also present. These strata are interpreted as longitudinal or lateral bar deposits. The textural variations were produced by avalanching of minor migrating dune bedforms (Smith 1972, Steel and Thompson 1983, Rust 1984). Similar sequences characterize the lateral bars of the modern braided-meandering streams of the area (Appendix 4). The

Table 2

Preglacial sediments, Snake River area, clast lithology and mineralogy.

A) Pebbles (6 analyses)

Lithology	Maximum	Mean	Minimum
	%	%	%
orthoquartzite	71	58	41
feldspathic sandstone	25	16	0
siltstone	27	11	8
chert	15	8	0
argillite	31	6	0
jasper-hematite	3	1	0
diabase	1	<1	0

Table 2 (continued)

B) Coarse sand (0.50-1.00 mm), 8 analyses

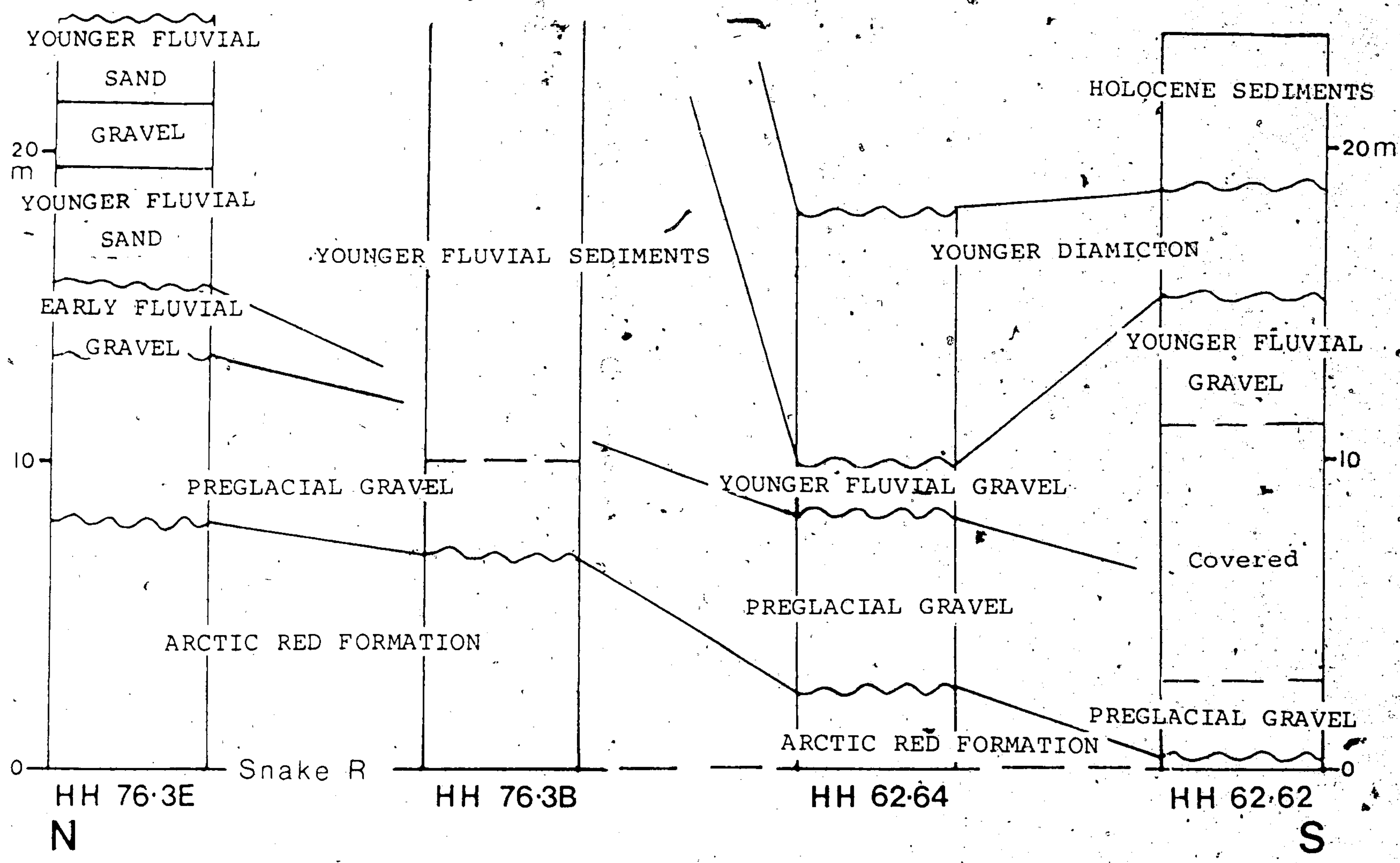
Lithology	Maximum	Mean	Minimum
	%	%	%
quartz	90	65	50
feldspar	30	20	5
siltstone fragments	30	6	0
chert	20	10	5
calcite	5	<1	0
hematite	3	<1	0
dolomite	<1	<1	0

Figure 11

Stratigraphic Position of Preglacial Sediments, Snake River Area.

The major outcrops containing preglacial fluvial sediments are illustrated, and stratigraphic correlations are indicated. Only the bases of sections HH 76-3B and HH 76-3E are illustrated. The datum is the Snake River. See Figure 10 for section locations.

FIGURE 11


YOUNGER FLUVIAL
SAND
GRAVEL
20
m
YOUNGER FLUVIAL
SAND
EARLY FLUVIAL
GRAVEL
10
PREGLACIAL GRAVEL
ARCTIC RED FORMATION
0
Snake R
YOUNGER FLUVIAL SEDIMENTS
YOUNGER DIAMICTON
YOUNGER FLUVIAL GRAVEL
PREGLACIAL GRAVEL
ARCTIC RED FORMATION
HOLOCENE SEDIMENTS
20m
YOUNGER FLUVIAL
GRAVEL
10
Covered
PREGLACIAL GRAVEL
0
HH 76·3E
HH 76·3B
HH 62·64
HH 62·62
N
S

orientations of the clasts suggest that flow was generally towards the north.

The deposits are devoid of palynomorphs and other environmental indicators. No means of determining the precise time of deposition is available.

The preglacial fluvial sediments are confined to the Snake River valley south of 65° 50' N. This distribution, combined with the palaeocurrent indicators, suggests that the preglacial Snake River flowed northeast or north into the Peel Plain and Mackenzie River. The absence of preglacial sediments in the valley of the westward-flowing lower part of the Snake River and the valley of the northward-flowing part of the Peel River indicates that these channels were excavated after the initial glaciation of the region.

C. Early Diamicton

At location HH76-3a(82)(Figure 12), a silty gravel diamicton overlies Arctic Red Formation Shale and underlies fluvial sediments (Figure 13). The 1.4 m thick diamicton is unstratified, and contains irregular pockets and lenses of sand. The matrix of the unit is predominatly silt and clay. The pebble content of the diamicton is 0.5%, dominated by orthoquartzite clasts (Table 3). Other clasts derived from the Canadian Shield are also present. Some limestone clasts were striated. No iron-oxide staining or carbonate coating of clasts was observed. The clasts showed no preferential

Figure 12

Early Fluvial Sediments, Snake River Area

Early Fluvial Sediments are exposed at the sections indicated with an X. See text for discussion.

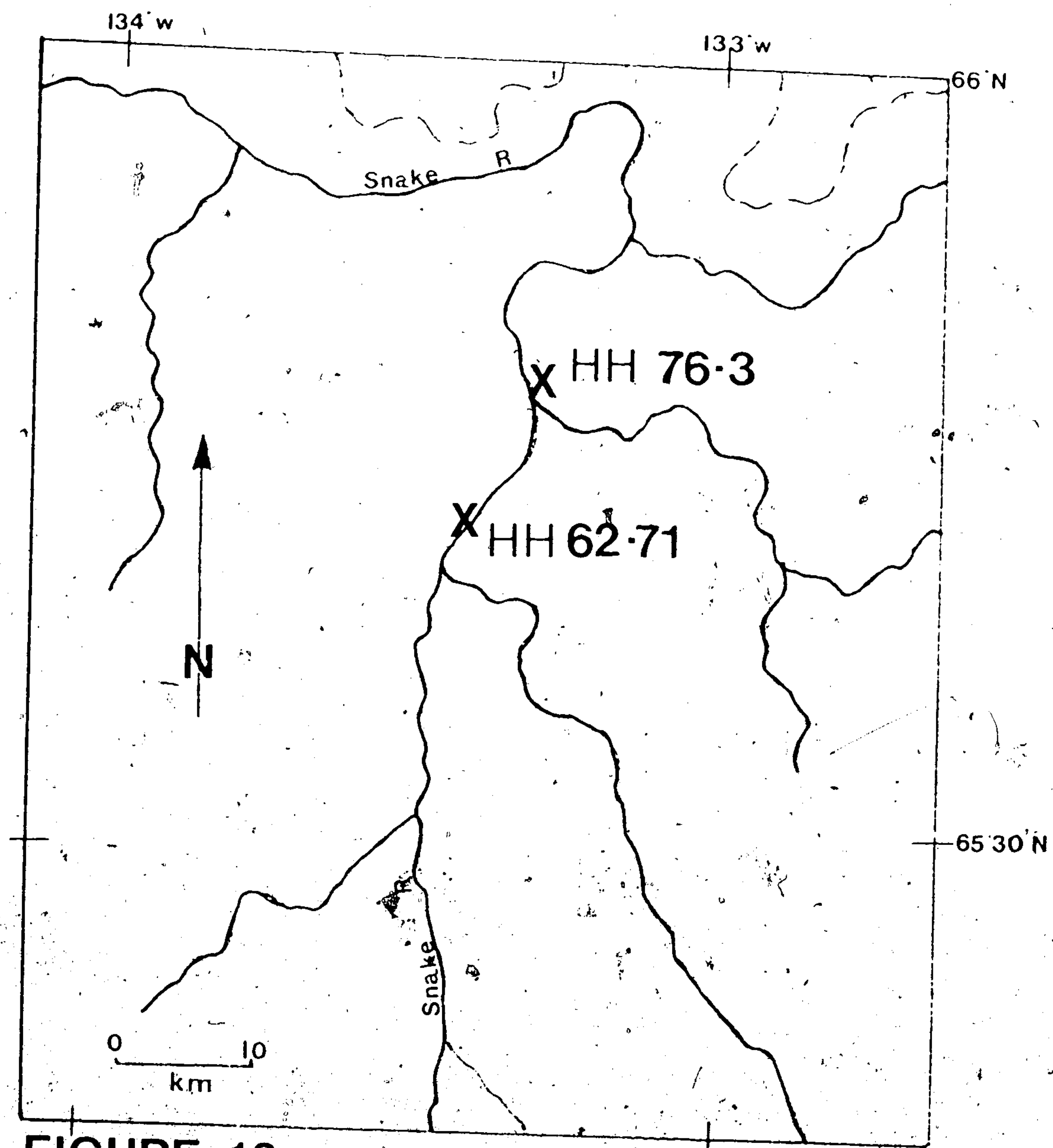


FIGURE 12

Plate 6 (top)

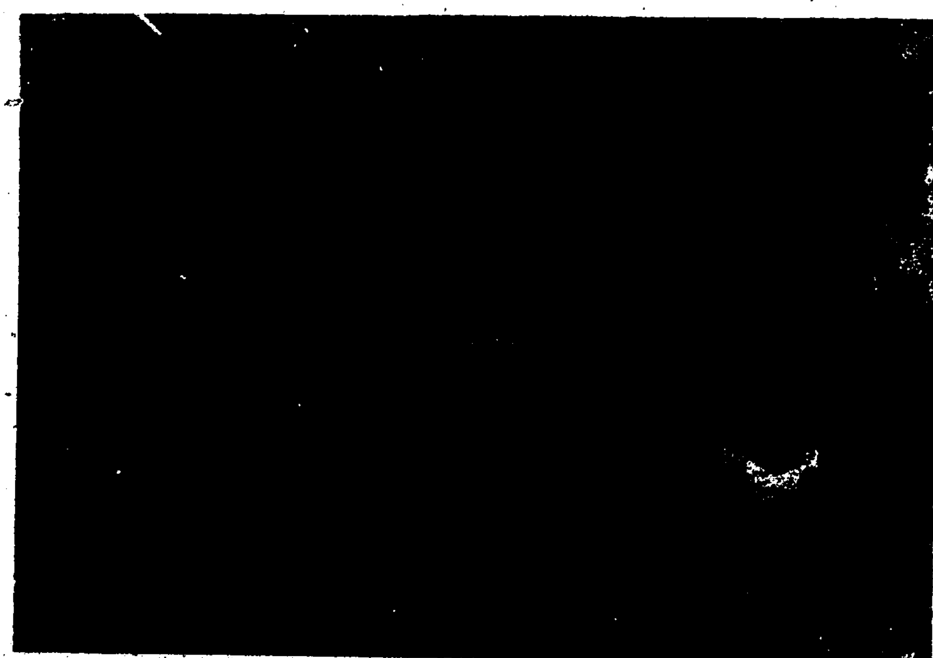
Early Diamicton, HH 76-3a

The unit is interpreted as a mass-movement deposit. Similar exposures of Holocene sediments are common throughout the region.

Plate 7 (bottom)

Oldest Glacigenic Diamicton, HH 62-71

The matrix of the oldest glacigenic diamicton at HH 62-71 is predominantly silt and clay (note the shardlike weathering on exposed surfaces). Discontinuous, poorly-defined sand and gravel lenses are present adjacent to the larger clasts. The clasts are oriented northwest-southeast. The white quartzite cobble in the foreground is 8 cm long.



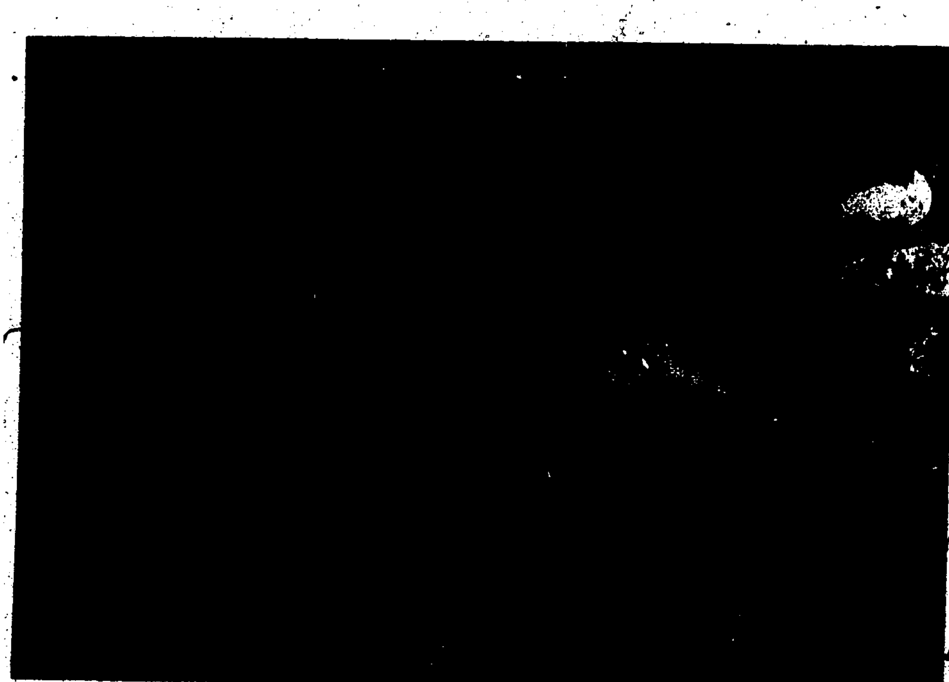


Figure 13

Stratigraphic Position of Early Fluvial Sediments, Snake River Area

The basal parts of the major sections containing Early Fluvial Sediments are illustrated, with stratigraphic correlations. The datum is the base of the measured sections. See Figure 12 for section locations.

FIGURE 13

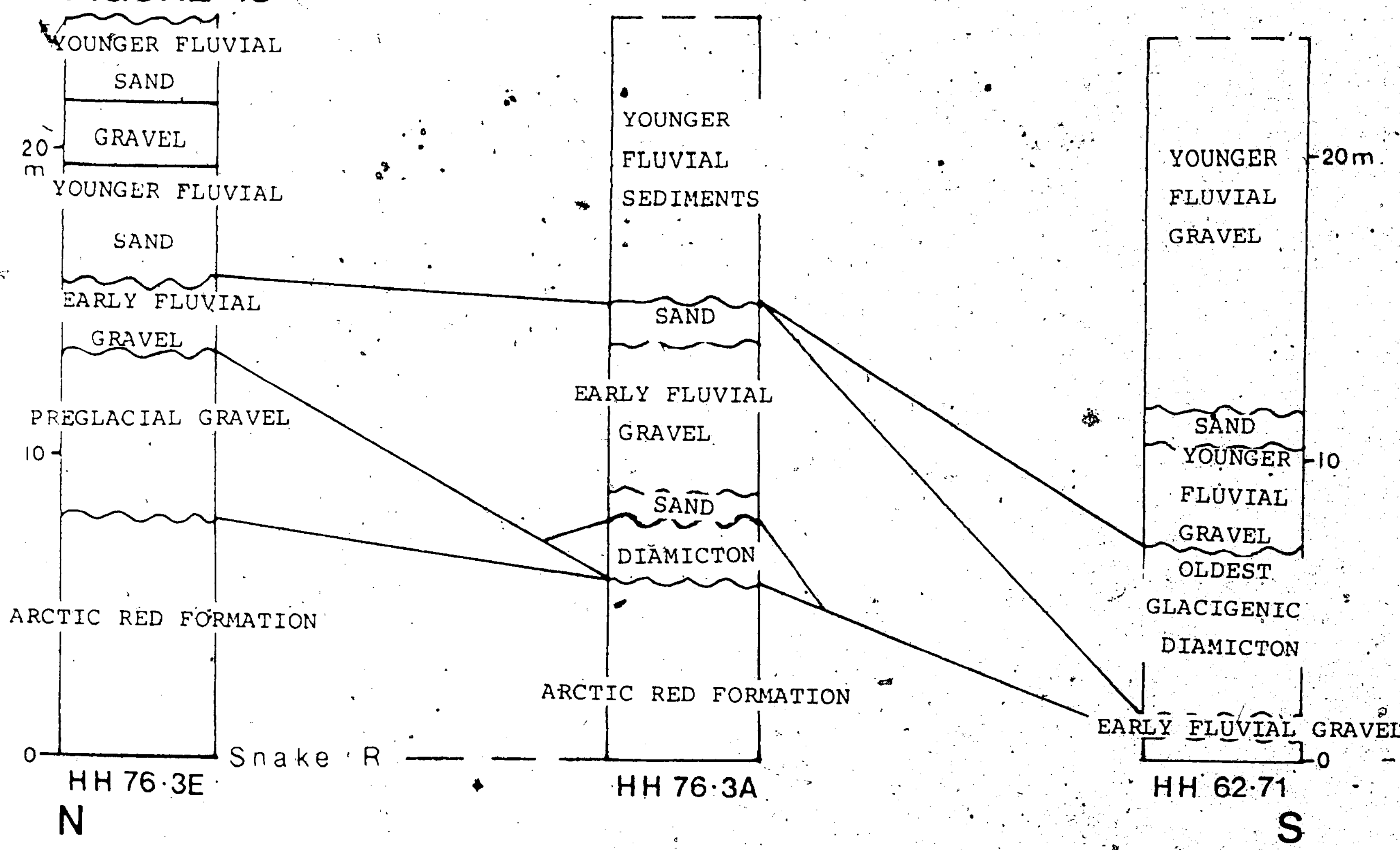

YOUNGER FLUVIAL
SAND
GRAVEL
20 m
YOUNGER FLUVIAL
SAND
EARLY FLUVIAL
GRAVEL
PREGLACIAL GRAVEL
10
ARCTIC RED FORMATION
0
Snake R
HH 76·3E
N
YOUNGER
FLUVIAL
SEDIMENTS
SAND
EARLY FLUVIAL
GRAVEL
SAND
DIAMICTON
ARCTIC RED FORMATION
HH 76·3A
YOUNGER
FLUVIAL
GRAVEL
20 m
SAND
YOUNGER
10
FLUVIAL
GRAVEL
OLDEST
GLACIGENIC
DIAMICTON
EARLY FLUVIAL GRAVEL
0
HH 62·71
S

Table 3

Early diamicton, section HH 76-3a, Snake River area; clast lithology and mineralogy

A) Pebbles and granules

Lithology	Coarse pebbles (1 analysis)	Granules (mean of 4 analyses)
	%	%
orthoquartzite	76	60
siltstone	7	8
argillite	7	8
feldspathic sandstone	3	6
chert	<1	6
jasper /hematite	3	3
limestone	1	6
dolomite	1	1
diabase	<1	<1
granite	0	1

Table 3 (continued)

B) Coarse Sand ([illegible] mm), major minerals ([illegible])

Lithology	Maximum	Mean	Minimum
quartz	[illegible]	[illegible]	[illegible]
feldspar	[illegible]	[illegible]	[illegible]
argillite fragments	[illegible]	[illegible]	[illegible]
chert	[illegible]	[illegible]	[illegible]
calcite	6	2	[illegible]
hematite	[illegible]	[illegible]	[illegible]
dolomite	[illegible]	2	1
chlorite	[illegible]	[illegible]	[illegible]
hornblende	[illegible]	2	1
pyroxene	[illegible]	[illegible]	[illegible]

Table 3 (continued)

(C) Coarse Sand Size Minerals

Mineral	Number of analyses containing mineral
Ankerite	3
Apatite	4
Biotite	4
Chromite	2
Cordierite	2
Corundum	1
Fluorite	4
Gypsum	2
Ilmenite	1
Kyanite	3
Magnetite	4
Muscovite	2
Pyrite	2
Pyrolusite	1
Rutile	3
Scapolite	2
Sphene	4
Tourmaline	4
Zircon	2

alignment.

This unit is tentatively interpreted as a mud flow deposit. The lack of oriented fabric, the irregular sand lenses, and the texture of the matrix, are all typical of modern mudflow deposits exposed in the region. The striated limestone clasts and granitic clasts, however, indicate that the deposit post-dates the initial glaciation of the area.

The deposit could represent a supraglacial ablation till or a till which has been disturbed by colluviation subsequent to its deposition. The presence of the diamicton in a single exposure suggests interpretation of the unit as a localized mudflow event. The unit does, however, represent the oldest deposit which contains evidence of glaciation of the region.

D. Early Fluvial Sediments

Fluvial sediments post-dating the preglacial deposits are present at several localities along the Snake River (Figure 12). Two of these localities, HH-62-71(82) and HH 76-3(82), were selected for detailed investigation. The stratigraphic relationship between the sediments from these outcrops is presented in Figure 13.

The sediments are predominantly sand and gravel, with clasts derived predominantly from the Snake River drainage basin (Table 4). Granitic clasts and medium and high grade metamorphic rock clasts are present in low concentrations within the sediments, however. These clast assemblages

Table 4

Early fluvial sediments, Snake River area, clast lithology and mineralogy.

A. Pebbles (10 analyses)

Lithology	Maximum	Mean	Minimum
orthoquartzite	74	65	60
feldspathic sandstone	21	4	2
siltstone	21	17	15
chert	8	<1	0
argillite	26	12	8
jasper-hematite	4	1	0
limestone	6	<1	0
dolomite	3	<1	0
granite	2	<1	0

Table 4 (continued)

B) Granules (7 analyses)

Lithology	Maximum	Mean	Minimum
orthoquartzite	56	45	[illegible]
feldspathic sandstone	16	12	8
siltstone	32	22	15
chert	5	3	2
argillite	24	6	1
jasper-hematite	6	3	1
limestone	6	5	4
dolomite	2	2	[illegible]
granite	3	[illegible]	<1
granodiorite	[illegible]	<1	0
diabase	[illegible]	<1	<1
gneiss	<1	<1	0
schist	<1	<1	0

Table 4 (continued)

(c) Coarse sand (0.42 - 1.00mm), major minerals (10 analyses)

Lithology	Maximum	Mean	Minimum
quartz	55	40	30
feldspar	35	30	20
argillite fragments	40	14	6
chert	18	8	4
calcite	6	3	1
hematite	2	1	<1
dolomite	2	1	1
hornblende	2	1	<1

Table 4 (continued)

D) Coarse sand, trace minerals.

Mineral	Number of analyses containing mineral
Andalusite	5
Ankerite	14
Apatite	12
Biotite	16
Chlorite	13
Chromite	11
Cordierite	4
Corundum	6
Epidote	7
Fluorite	14
Garnet	12
Gypsum	2
Ilmenite	5
Kyanite	4
Magnetite	16
Muscovite	16
Pyrite	13
Pyrolusite	2
Pyroxene	14
Rhodonite	11
Rutile	8
Siderite	12
Sillimanite	3

Table 4 (continued)

Mineral	Number of analyses containing mineral
Tourmaline	12
Wollastonite	10
Xenotime	1
Zircon	2

indicate that the sediments were deposited after the initial glaciation from the Canadian Shield, although no pre-existing glacial sediments were detected in the Snake River basin.

The lithologic successions within these sediments suggest that the environment of deposition was a braided-meandering stream, similar to the modern Snake River. The assemblages are dominated by crudely stratified poorly to moderately sorted gravel and sand, with planar tabular cross-beds. The upper parts of the erosionally bounded gravel sequences are characteristically formed as laterally-accumulating diagonal or longitudinal bars in riffle zones of braided and braided-meandering streams (eg. Smith 1974, Miall 1978, Schwartz 1978). Similar sequences were observed in longitudinal bars along the modern Snake River (Appendix 4).

Pebbles and cobbles are transported during periods of moderately high flow, either as diffuse gravel sheets (Hein and Walker 1977) or as imbrication clusters (Martini 1977). The rolling, sliding, and rotating finer pebbles and granules were forces into interstices in the framework formed by cobbles deposited during higher energy flow, a process termed "contact imbrication" (Blacknell 1981, 1982). The resulting fabric reflects the combined influences of the initial rolling process and the subsequent jamming of slightly smaller clasts into the original framework.

If the stream flow levels decline rapidly, the sediment will be preserved as an open-worked, matrix-free deposit. If the levels decline more slowly, however, finer sediments transported in suspension during the lower flow stages can infiltrate into the open-worked structure, producing a very poorly sorted gravel with coarse clasts in a fine matrix (Eynon and Walker 1974; Beschta and Jackson 1979, Frostick et al 1984).

The deposits are derived of palynomorphs and other environmental indicators, and no means of determining the precise time of deposition is available. The sediments are presumed to postdate the initial incursion of glacial ice from the Canadian Shield, because they contain granitic and medium and high grade metamorphic fragments.

E. Oldest Glacigenic Diamicton

The oldest glacigenic diamicton has been identified only within the Snake River valley, and is here informally designated "Snake River till". At locality HH 62-71 (82) (65° 46' N, 133° 19' W), where it is well-exposed, the till unit is 5.9 m thick (Figure 14, Plate 7). At other localities, the till reaches a thickness of 10 m (Figure 15).

The till at HH 62-71 (82) is coarse with as much as 46% sand and 5% cobbles, pebbles and granules, the majority of which are subrounded. Sedimentary rock clasts are dominant (Table 5). The proportion of quartzite clasts is low and

Figure 14

HH 62-71 (82)

HH 62-71 (82) is the location at which the Oldest Glacigenic Diamicton ('Snake River till') is best exposed. The figure illustrates the stratigraphic position of the diamicton in relation to the Early Fluvial and Younger Fluvial Sediments. The base of the section is at 270 m asl.

FIGURE 14
HH 62-71(82)
1:200

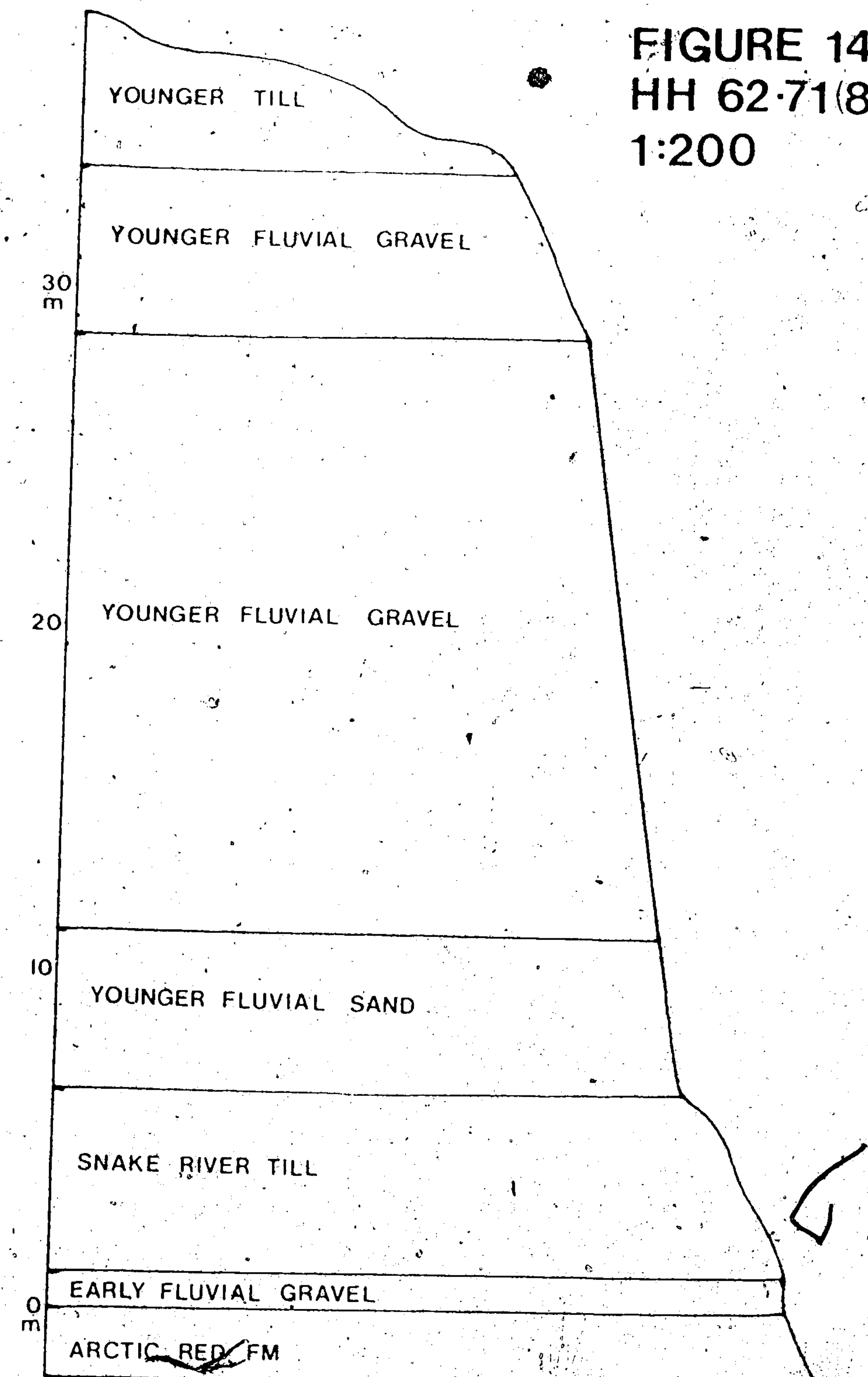


Figure 15

Oldest Glacigenic Diamicton, Snake River Area

The Oldest Glacigenic Diamicton is exposed at several locations along the Snake River. The most detailed investigation was conducted at location HH 62-71 (82). See text for discussion.

S---Location of HH 62-71 (82)

s---other exposures of Oldest Glacigenic Diamicton

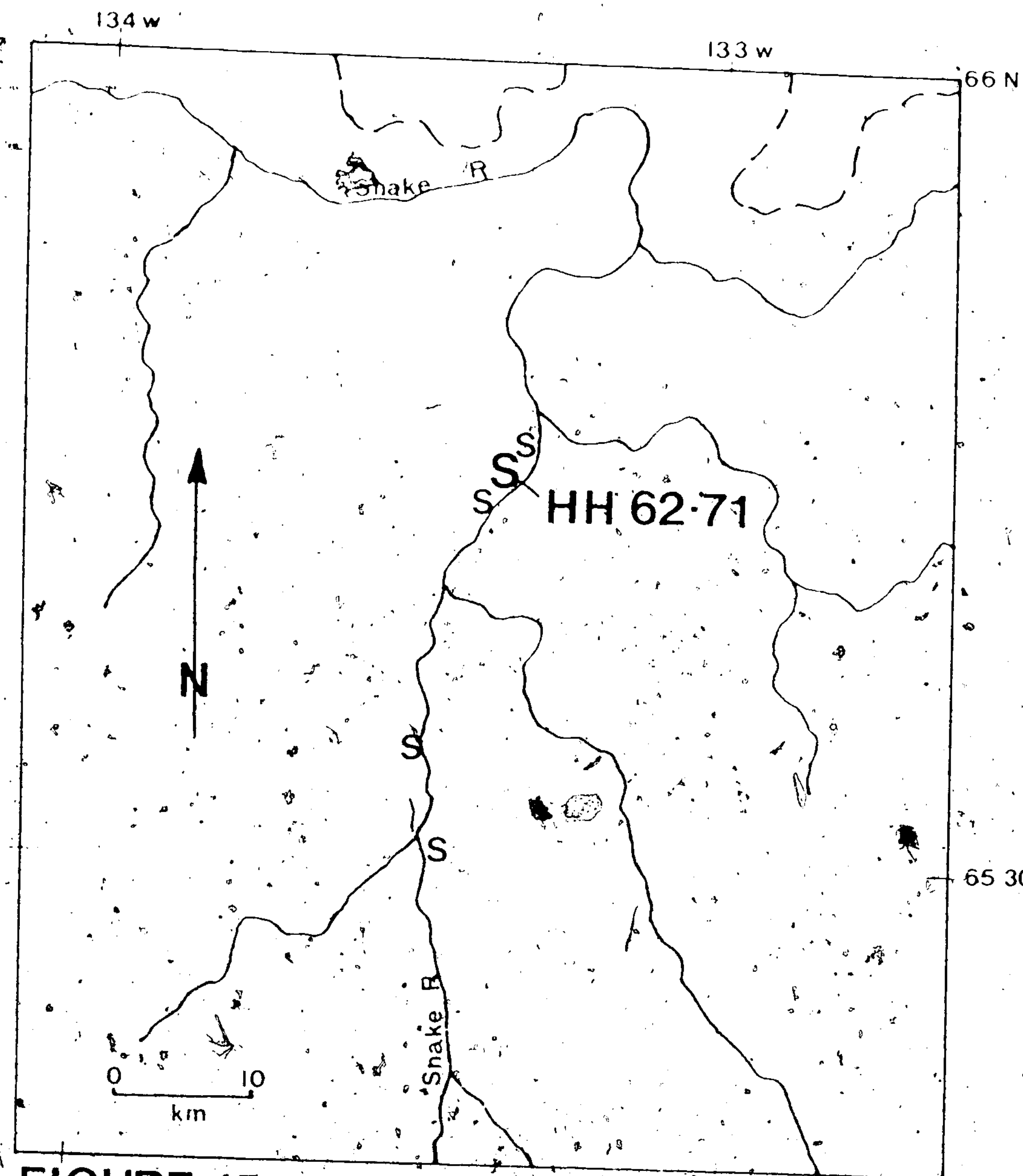


FIGURE 15

Table 5

Oldest glaciogenic diamicton, 4H of [illegible]: [illegible] lithology and mineralogy. Lower refers to the mean of [illegible] samples obtained 1m above lower contact. Middle refers to mean of [illegible] samples obtained 3m above lower contact, and Upper refers to mean of [illegible] samples obtained 5m above lower contact.

A) Coarse pebbles (1 analysis)

Lithology	Percent
orthoquartzite	40
feldspathic sandstone	0
siltstone	15
chert	12
argillite	14
jasper-hematite	10
limestone	3
dolomite	0
granite	0
diabase	6
gneiss	0
quartz conglomerate	0
metaquartzite	0

Table 5 (continued)

B. Fine pebbles and granules

Lithology	Lower	Middle	Upper
orthoquartzite	[illegible]	[illegible]	6
feldspathic sandstone	[illegible]	[illegible]	20
siltstone	16	20	28
chert	8	7	4
argillite	12	6	5
jasper-hematite	6	5	3
limestone	[illegible]	8	20
dolomite	2	6	5
granite	1	2	2
diabase	7	6	5
gneiss	<1	<1	<1
quartz conglomerate	0	1	<1
metaquartzite	2	3	<1

Table 5 (continued)

(?) Coarse sand ([illegible] [illegible]mm), major minerals

Lithology	Lower	Middle	Upper
quartz	34	[illegible]	[illegible]
feldspar	[illegible]	[illegible]	[illegible]
argillite fragments	26	[illegible]	[illegible]
chert	1[illegible]	[illegible]	[illegible]
calcite	2	2	[illegible]
hematite	[illegible]	1	1
dolomite	2	[illegible]	[illegible]
hornblende	3	6	[illegible]

Table 5 (continued)

Diffuse Sand, trace minerals

Mineral	Lower # of analyses containing mineral	Middle # of analyses containing mineral	Upper # of analyses containing mineral
Apatite	[illegible]	[illegible]	[illegible]
Biotite	3	[illegible]	[illegible]
Chlorite	3	[illegible]	3
Chromite	1	3	2
Corundum	0	0	2
Epidote	1	1	1
Fluorite	2	3	3
Garnet	3	3	3
Gypsum	2	0	2
Ilmenite	0	1	3
Kyanite	0	2	3
Magnetite	3	3	3
Muscovite	2	2	2
Pyrite	3	1	2
Pyroxene	2	3	3
Rhodonite	0	3	3
Rutile	1	3	2
Siderite	2	1	1
Sillimanite	0	1	0
Sphene	1	3	3
Tourmaline	2	3	3

Mineral	Lower	Middle	Upper
	# of analyses containing mineral	# of analyses containing mineral	# of analyses containing mineral
Wollastonite	0		

declines stratigraphically upward. Fresh granite, diabase and gneiss pebbles and granules are also present. Serpentinite clasts were not observed. Angular blocks of Arctic Red Formation shale are present in the basal 2 m of the till unit.

Lenses of medium grained sand are present throughout the till, the largest observed being 30 cm x 10 cm x 8 cm. The lenses are randomly oriented, although they are more numerous and larger in the basal part of the unit. The lenses are generally structureless, although some contorted laminations are present. Contacts between the diamicton and the lenses are sharp and irregular. The lenses displayed no consistent morphology. The pebble fabric is bimodal, with the major mode oriented at 290-110 and the minor mode oriented 325-145.

The diamicton at HH 62-71 (82) is interpreted as a basal melt-out till. The presence of unlithified sand lenses, the well-developed preferred orientation of the fabric, the predominatly subrounded large clasts, and the preservation of angular clasts of weakly consolidated sediment (such as the Arctic Red Formation Shale) are all typical of sediments interpreted as basal melt-out tills (eg. Lawson 1979; Shaw 1982; Haldorsen and Shaw 1982; Catto 1984).

The bimodal fabric observed may be due either to shifting in the direction of ice-flow, or to the effects of local subglacial topography, or to intraglacial flow regimes. Re-orientation of pebble fabric can occur as a result of glacial override (MacClintock and Dreimanis 1964; Ramsden and Westgate 1971; Catto et al 1982), but no independent evidence of multiple events exists. Younger tills present in the area differ in fabric orientation from either of the modes preserved in the "Snake River till" exposed at HH 82-71 (1982). At this locality, the till is overlain by 28 m of undisturbed fluvial sediment (Figure 14). The bimodal fabric is therefore considered to reflect variation in motion of the ice responsible for depositing the till, rather than being a product of subsequent glacial override and or re-orientation.

Till correlative to the Snake River deposits has not been found west of the Trevor Range. Glaciolacustrine sediments predating the final ice advance in the Bonnet Plume Basin have been recognized by Hughes et al (1981). These sediments were produced in a lake impounded in the Bonnet Plume Basin by ice which blocked eastward drainage north of the Trevor Range. Hughes (Geological Survey of Canada, personal communication, 1984) has informally termed this event the "Deception Glaciation", named for the outcrop of glaciolacustrine sediments in the vicinity of Mount Deception, along the Wind River and Hungry Creek.

The "Snake River till" is here considered to be correlative to the "Deception" glaciolacustrine sediments. The known areal distribution of the till coincides with the position which would be occupied by an ice sheet capable of impounding the Bonnet Plume Basin. This correlation must be regarded as tentative, however, as no outcrops demonstrating the relationship between the two units have yet been detected.

F. Peel River Lacustrine Sediments

An exposure of lacustrine sediments is present at location HHC 81-7 (66° 03' N, 134° 11' W), along the north bank of the Peel River east of its confluence with the Snake River. The base of the exposed section stands at 310 m A.S.L. A total of 13.36 m of lacustrine sediment is exposed (Figure 16).

Description

Unit A

The basal 10.7 m of the section (unit A, Figure 16) consists of alternating strata of fine sand and silty sand (59 cm to 265 cm thick) and clayey silt (3 cm to 26 cm thick). Successive strata tend to decrease in thickness. The fine sand and sandy silt units are primarily normally graded, although intervals of reverse grading are present within individual strata. The layers contain a variety of sedimentary structures, including load contortions, flame and spoon structures, and ripple-drift cross-laminations (primarily type B of Jopling and Walker 1968, with some type S).

Figure 16

Laçustrine Sediments, HHC 81-7

The succession at HHC 81-7 consists of four units, lettered A through D. Typical parts of each unit are illustrated in the inset diagrams. See text for further discussion.

Sand

Silt

Clayey Silt

Clay

Swing-dash Laminations

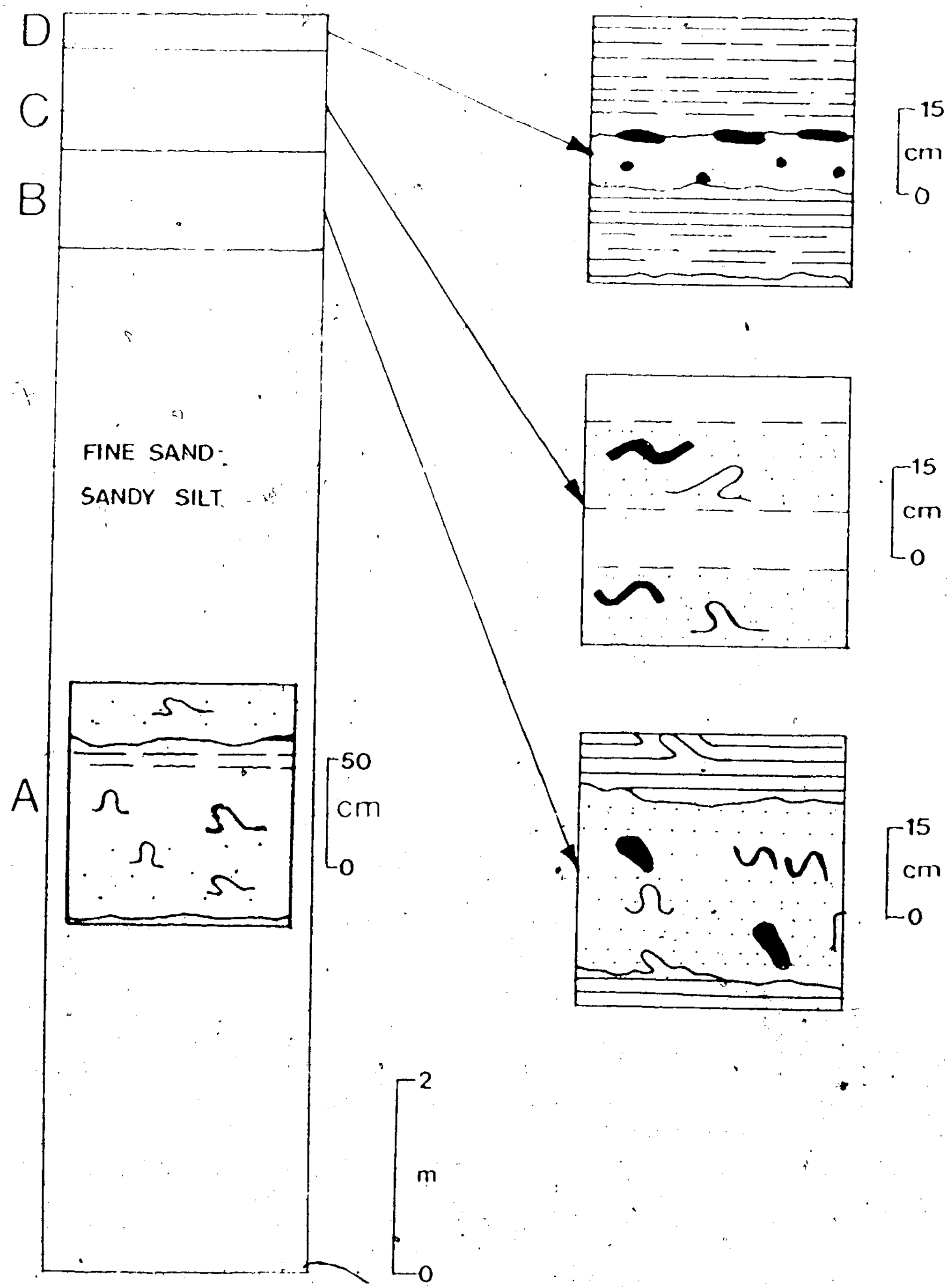
Rip-up Clasts

Recumbent Drag-Folds

Isolated Granules and Coarse Sand in Coarse Silt

Discoid Shale Pebbles

FIGURE 16
SEDIMENTS, HHC 81-7



Discontinuous, irregularly curved or sinusoidal laminations of clay 1-5 cm in length and [illegible] 0.5 cm in width are scattered throughout the strata. The laminations are oriented with the eastern portion convex up, and the western portion concave up. As the features resemble swing-dashes in form, they are here termed "swing-dash laminations". The laminations are invariably composed of finer sediment than the enclosing clayey silt, and are internally structureless. All of these features indicate that the direction of current and sediment movement was towards the west throughout the sedimentary sequence.

Also present in the sediments are teardrop shaped silt and fine sand clast aggregates oriented vertically, and rare horizontal laminations. A single structureless silty clay biconvex lens (9 cm maximum thickness), with an irregular lower contact and a distinct, conformable upper contact was present within a sandy stratum. The clayey silt units are either structureless or contain fine laminations defined by fine sand and coarse silt monolayers. Contacts between fine sand and clayey silt layers are either irregularly gradational or sharp, planar and erosional. Clayey silt layers invariably possess an upper erosional contact.

Unit B

The overlying 1.06 m of sediment (unit B, Figure 16) is characterised by silty sand, clayey silt, and silty clay layers separated by irregular, erosional contacts. The clayey silt strata contain recumbent drag folds and spoon structures with orientations suggesting westward flow. Horizontal laminations of silt, fine sand lenses and laminations and ball-and-pillow structures are also present. The silty sand stratum contains irregular load casts, silt and clay rip-up clasts, contorted laminations, dewatering structures, and boudinaged silt clasts. The silty clay unit is structureless.

Unit C

The overlying 1.07 m sequence (unit C, Figure 16) consists of a series of fining-upward and coarsening upward silt and sand units, separated by gradational contacts. Recumbent drag fold, spoon-shaped silt structures, and swing-dash laminations in the base of the sequence indicate westward flow, but similar flow indicators in the upper part of the sequence show that the direction shifted to the east and northeast. The alignment of aggregated fine sand and silt clasts in sandy units also suggests eastward flow. The shift in flow directions is gradual.

Unit D

The uppermost 41 cm of the exposed section (unit D, Figure 16) consists of two layers of clayey silt separated along erosional contacts by a coarse silt layer containing granules. A layer of discoid shale pebbles separates the upper surface of the coarse silt from the overlying clayey silt. The sediments are overlain by at least 25 m of colluviated clayey silt, which contains trace amounts of pebbles derived from the Canadian Shield.

Swing-dash Laminations

The fine texture of the laminations indicates formation during the periods of slowest flow. These structures are interpreted to be deposits draped over ripple forms developed on the silty clay by sluggish currents. The swing-dashes are similar in some respects to the flaser laminations described by Reineck and Wunderlich (1968) and Terwindt and Breusers (1972, 1982), but the two features are not identical. Flasers are generally preserved only in the lee of ripple forms, and usually are developed overlying sandy sediment. The quasi-sinusoidal form of most swing-dash laminations indicates that the draped sediment over the upper stoss area of the ripples was at least partially preserved.

The curved and irregular upper surfaces of the swing-dash laminations indicates that some erosion and modification of the original infill form is involved. In a high energy environment, the fine sediment infill would be prone to reworking and transportation. Under the low energy conditions of this lacustrine regime, reworking takes the form of modification of the upper surface into a quasi-ripple feature. Preservation of the upper stoss area may reflect migration and climbing of the ripple forms.

Interpretation

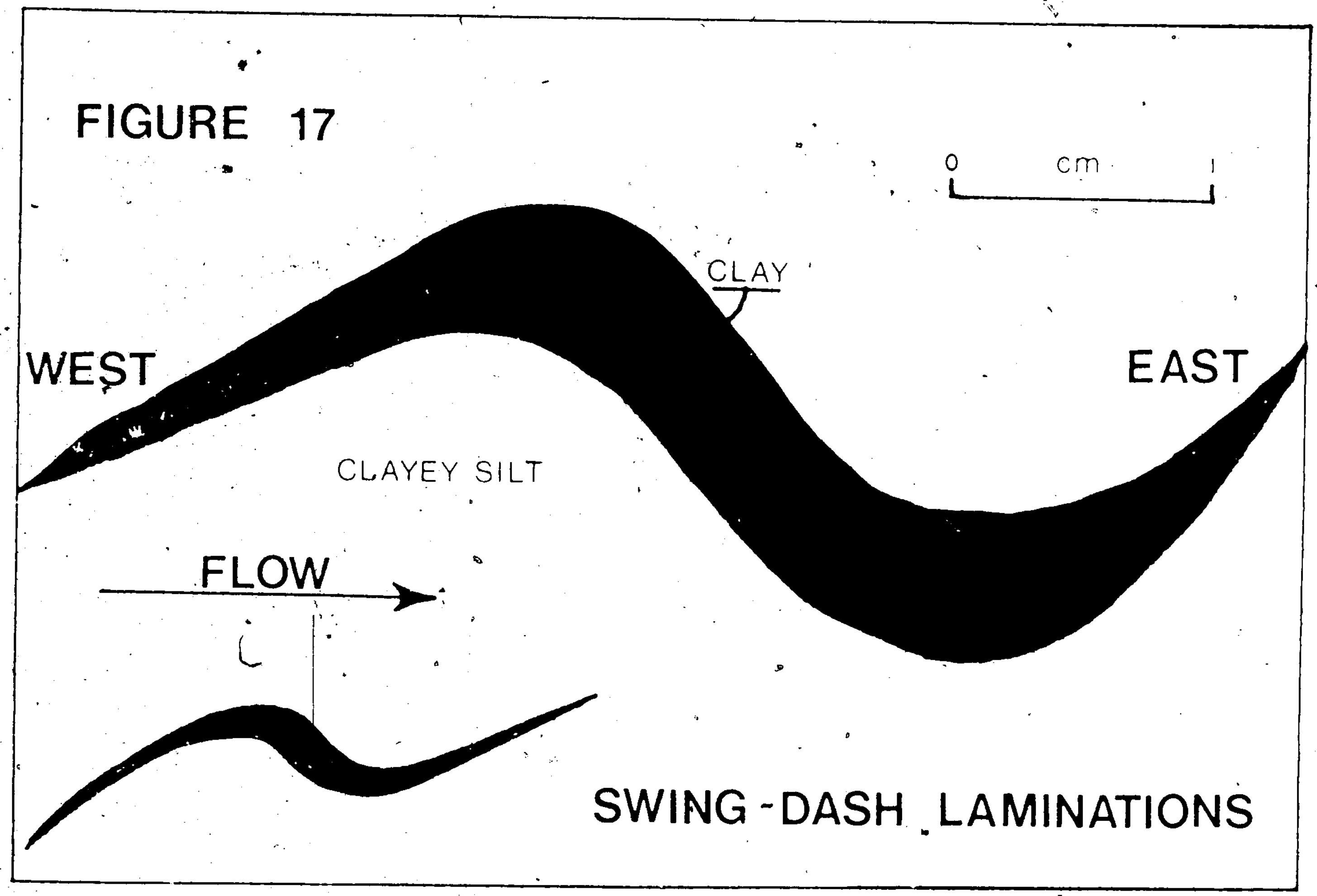
The sediments were deposited in a lake dammed by ice to the east, that slowly retreated throughout the depositional

Figure [illegible]

Swing-Dash Laminations

The figure shows two swing-dash laminations contained in clayey silt of the upper part of Unit C, HHC 81-7. These laminations were produced by flow from west to east. The laminations are composed of clay, and are intercalated in clayey silt. See text for discussion.

FIGURE 17



0
cm
1
CLAY
WEST
EAST
CLAYEY SILT
FLOW
SWING-DASH LAMINATIONS

period. Initially, the ice front was adjacent to the site of HHC 81-7 (Figure 18). Sediment-laden plumes of glacial meltwater discharged across the lake bottom from the east, depositing the sediments of unit A with loading structures, en *echelon* faults and recumbent drag folds, sedimentary structures typical of proximal lacustrine environments (McKee and Goldberg 1969). Ripple-drift cross-lamination and the texture of the sediments indicate that flow proceeded towards the west at 15 - 20 cm/sec.

As the ice withdrew towards the east, flow velocities recorded at HHC 81-7 declined, and the thickness of sediment deposited by each successive plume decreased. Suspension sedimentation deposits of clayey silt and silty clay gradually increased in proportion to the decline in the strength of the turbid underflow events, as seen in unit B. Loading of the silty sands by the settling of fine sediments permitted the formation of dewatering structures and ball-and-pillow features. The flow direction throughout this period remained westerly, suggesting that drainage was *via* the upper Peel River valley westward into the Bonnet Plume Basin.

Further eastward retreat of the ice eventually permitted northward and eastward drainage of the lake over the plateau to the west of the modern Peel River valley (Figure 19). This period of the lake's development is represented by the upper part of unit C, with structures produced by eastward flow. The transition between westward

Figure 18

Stage 1, Lacustrine Events, HHG 81 7

Silt

Clay

Ripples

Swing-dash Laminations

Lacustrine Currents.

Lacustrine Sediment Flows

See text for discussion.

FIGURE 18 STAGE 1. HHC 81-7

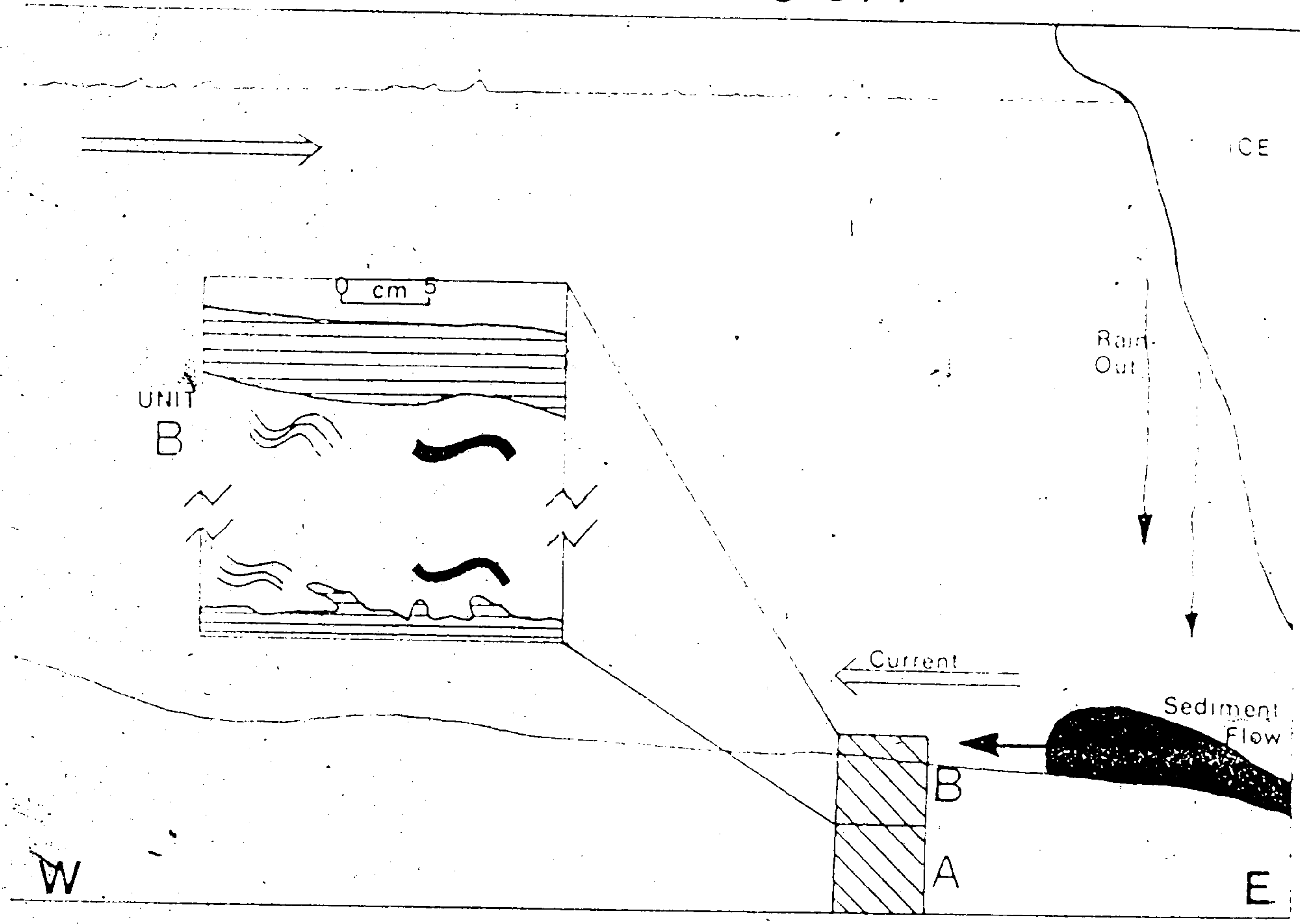


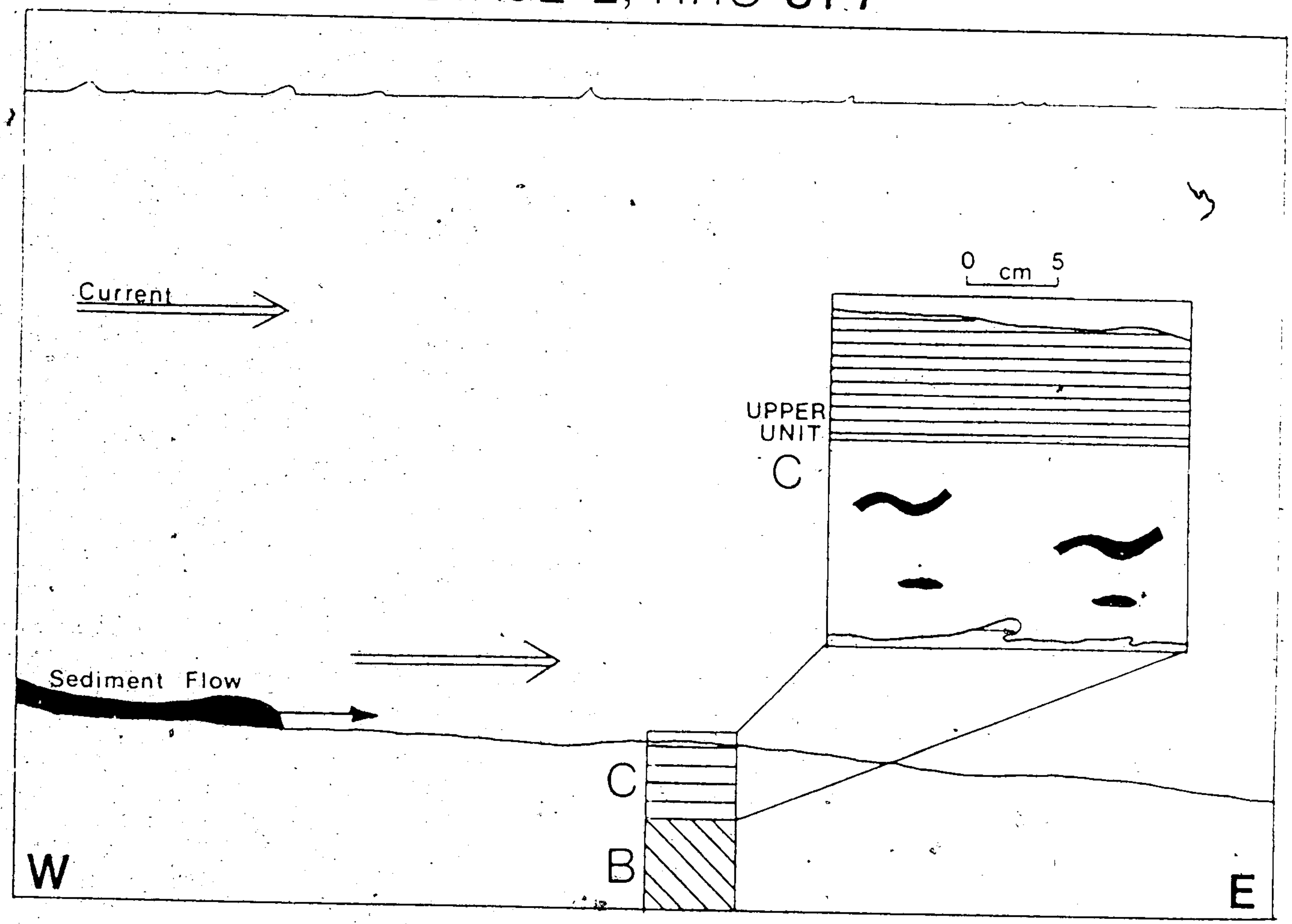
Figure 19

Stage 2, Lacustrine events, HHC 81-[illegible]

During the deposition of the upper part of Unit C, flow was towards the east. Unit C is composed of silt layers with swing-dash laminations, overlain by clay. The upper contacts of the clay layers are erosional. See text for further discussion.

Legend as for Figure 16.

FIGURE 19 STAGE 2, HHC 81-7



and eastward flow was gradual, indicating that ice remained in the lower Peel valley northeast of the lake at this time.

Withdrawal of the ice from the lower Peel valley resulted in the drainage of the lake. A drainage event is indicated by the erosional surface and discoid shale pebble layer present in unit D in the uppermost part of the section. The overlying clayey silt unit may represent a temporary refilling of the lake induced by a seasonal readvance of the ice, or it may simply be an infilled fluvial channel or colluvial erosion surface.

Although the lacustrine sediments are overlain by colluvium containing granitic and gneissic clasts, the position of section HHC 81-7 marginal to the Peel River valley indicates that the colluvial material may have originated from upslope exposures of older units. The age of the lacustrine sediment cannot therefore be assessed at present. No other lithological units are present in the exposure. The Peel River lacustrine sediments could represent an eastern arm of the lake in the Bonnet Plume Basin, in which the "Deception" lacustrine sediments were deposited (Hughes et al. 1981). There is no evidence currently available to support or refute such a conjectural opinion, however.

G. Younger Fluvial Sediments

Fluvial sediments which postdate the "Snake River till" and predate younger glacigenic sediments are present at several locations along the Snake River (Figure 20). Three of these locations, HH62-62(82), HH62-71(82), and HH76-3(82) were selected for detailed investigation. The stratigraphic relationships among the sediments from these outcrops are presented in Figure 21. Two additional exposures, HH62-74a(82) and HH62-74b (82), also display sequences of fluvial sediments. The absence of correlatable units and datable material at either exposure precludes the determination of the stratigraphic position of either sediment assemblage.

The sediments are predominantly sand and gravel, with associated clay and silt beds. The clasts are derived predominantly from the Snake River drainage basin, although granitic and medium- and high-grade metamorphic rock clasts are present (Table 6).

The sediments were deposited in braided-meandering rivers similar to the modern Snake River. The most complete braided-meadering stream complex is present in six exposures along the east bank of the Snake River at locality HH76-3(82). The complex includes primary channel gravels, lateral bar deposits, back-channel sands, silts and clays, bar-chute sands and gravels, scour pool infill deposits, confluence bar deposits, and aeolian deposits(including loess and adhesion structures), as illustrated in Figure 22.

Figure 20

Younger Fluvial Sediments, Snake River Area

Younger Fluvial Sediments are exposed at the sections indicated with an X. See text for discussion.

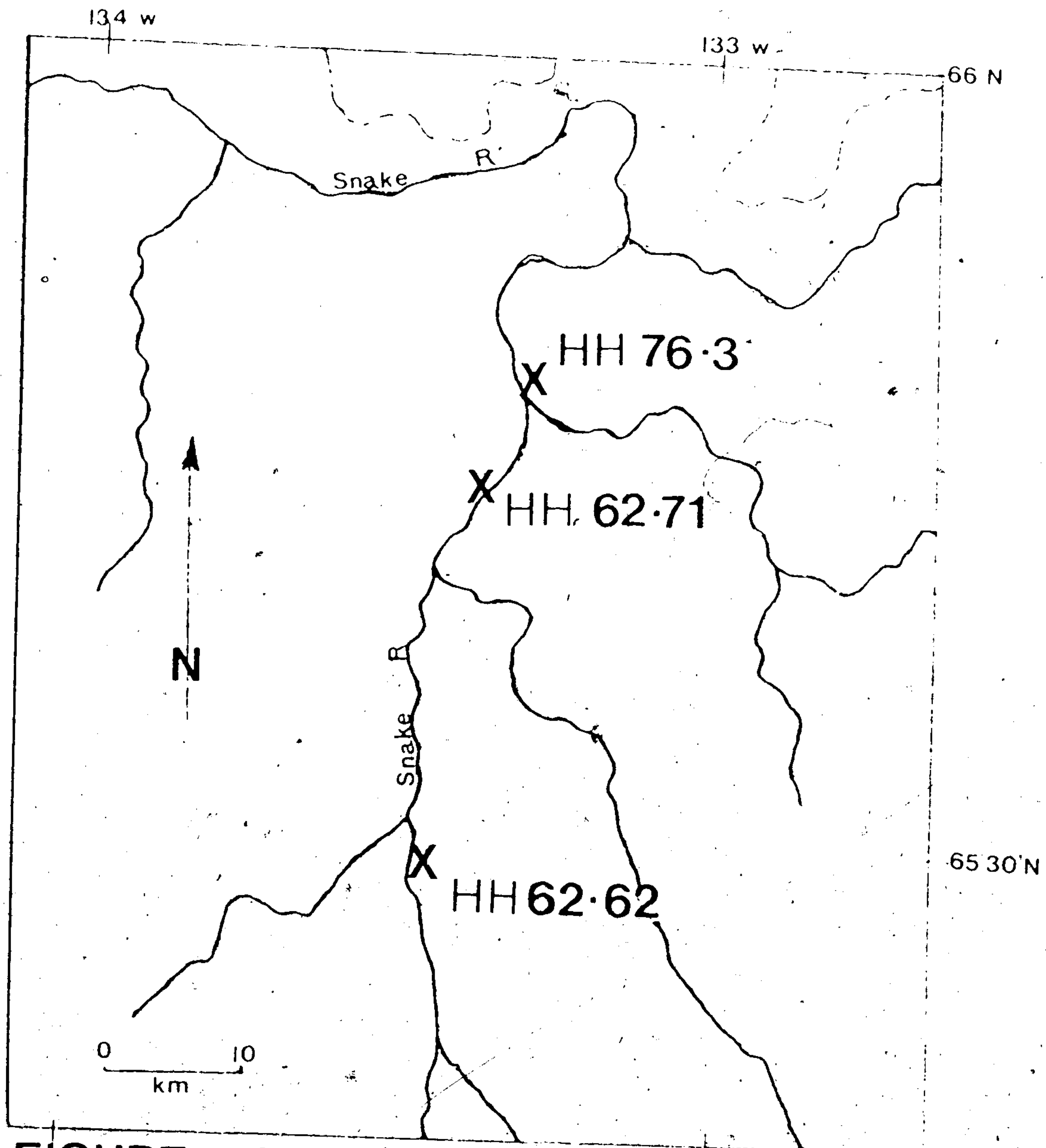

134 w
133 w
66 N
65 30'N
Snake R
Snake R
HH 76·3
HH 62·71
HH 62·62
N
0
10
km

FIGURE 20

Figure 21

Stratigraphic Position of Younger Fluvial Sediments, Snake River Area

The principal sections at which Younger Fluvial Sediments are exposed are illustrated. Stratigraphic correlations among the sections are indicated. Locations of the sections are illustrated in Figure 20.

A---location of wood ^{14}C dated at >39,000 B.P. (GSC-3697)

B---location of wood ^{14}C dated at >40,000 B.P. (GSC-3800)

D---Disturbed and Colluviated Sediment

ED---Early Diamicton

EFS---Early Fluvial Sediment, sand and silt

EFG---Early Fluvial Sediment, gravel

HS---Holocene Fluvial sand

HG---Holocene Gravel, Alluvial Fan and Fluvial

KA---Cretaceous Arctic Red Formation Shale

OGD---Oldest Glacigenic Diamicton

PG---Preglacial Fluvial Gravel

YFG---Younger Fluvial Sediment, gravel

YFS---Younger Fluvial Sediment, sand

YFU---Younger Fluvial Sediment, silt and clay

YGD---Younger Glacigenic Diamicton

FIGURE 21

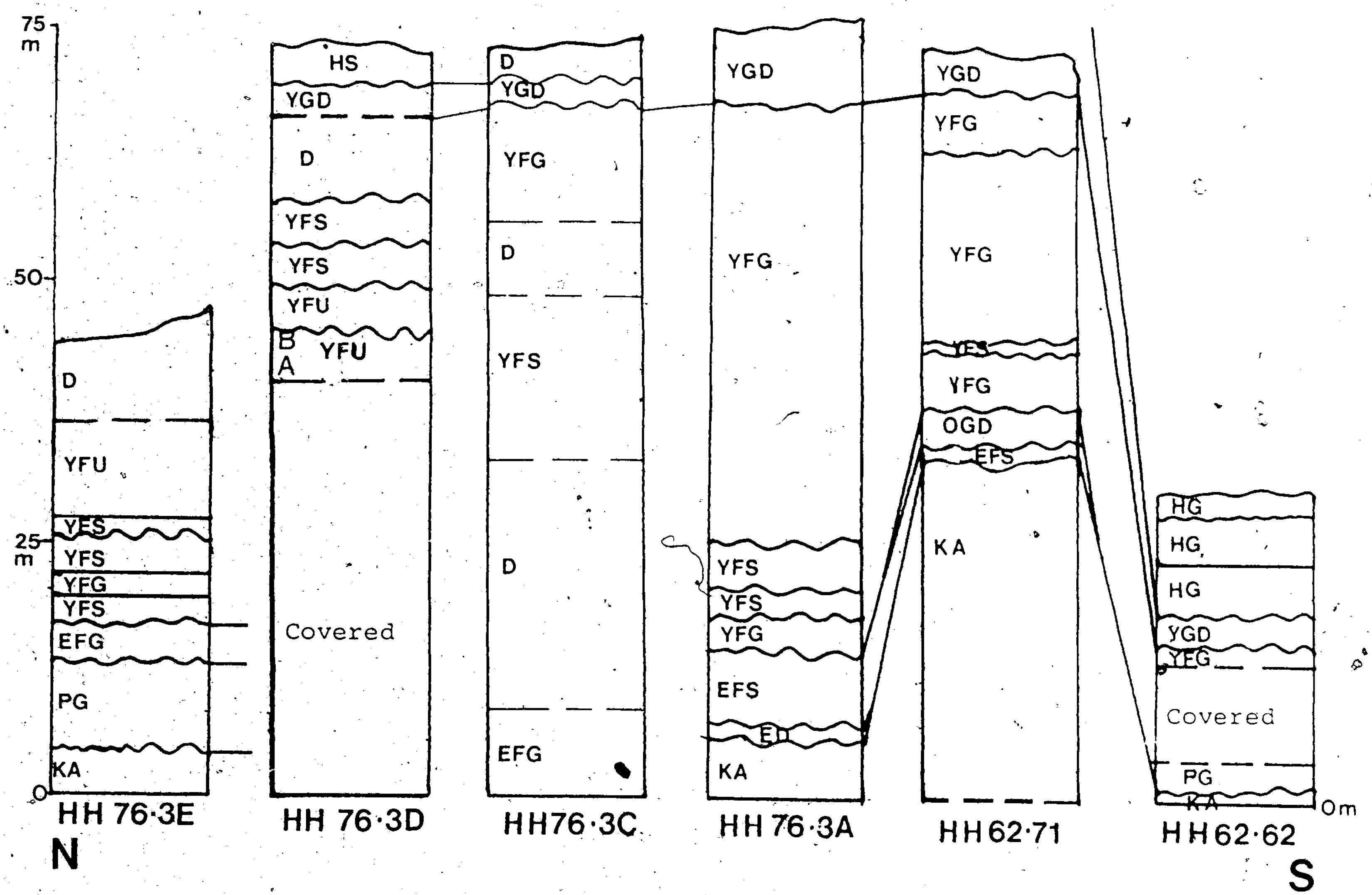

75 m
50
25 m
0
HS
YGD
D
YFS
YFU
B
A
Covered
YFG
YFS
EFG
PG
KA
OGD
EFS
ED
HG
0m
HH 76·3E
HH 76·3D
HH76·3C
HH 76·3A
HH62·71
HH62·62
N
S

Table 6

Younger fluvial sediments, Snake River area, clast lithology and mineralogy.

A) Pebbles (20 analyses)

Lithology	Maximum	Mean	Minimum
	%	%	%
orthoquartzite	75	55	44
feldspathic sandstone	21	15	6
siltstone	22	10	0
chert	10	3	1
argillite	29	12	10
jasper-hematite	16	3	0
limestone	11	2	0
dolomite	6	<1	0
granite	2	<1	<1
diabase	8	1	<1
granodiorite	<1	<1	0
gneiss	<1	<1	0

Table 6 (continued)

B) Granules (15 analyses)

Lithology	Maximum	Mean	Minimum
orthoquartzite	70	52	45
feldspathic sandstone	25	12	8
siltstone	20	8	3
chert	15	4	3
argillite	30	20	15
jasper-hematite	12	1	0
limestone	10	< 1	0
dolomite	5	< 1	0
granite	3	2	1
granodiorite	< 1	< 1	0
diabase	8	3	2
gneiss	< 1	< 1	0
metaquartzite	2	< 1	0
schist	< 1	< 1	0

Table 6 (continued)

c) Coarse Sand (0.42 - 1.00mm), major minerals (34 analyses)

Lithology	Maximum	Mean	Minimum
quartz	50	40	[illegible]
feldspar	50	35	30
argillite fragments	20	10	[illegible]
chert	15	8	5
calcite	4	1	<1
hematite	2	<1	<1
dolomite	2	<1	<1
hornblende	8	6	4
magnetite	1	<1	<1

Table 6 (continued)

D) Coarse sand, trace minerals.

Mineral	Number of analyses containing mineral
Actinolite	25
Andalusite	16
Ankerite	15
Apatite	29
Biotite	34
Bornite	16
Chlorite	32
Chromite	17
Cordierite	6
Corundum	28
Covellite	15
Epidote	11
Fluorite	30
Garnet	9
Gypsum	19
Ilmenite	23
Kyanite	16
Leucoxene	11
Magnesite	5
Muscovite	30
Pyrite	26
Pyrolusite	2
Pyroxene	31

Table 6 (continued)

Mineral	Number of analyses containing mineral
Rhodonite	16
Rutile	17
Siderite	15
Sphene	27
Tourmaline	33
Xenotime	4
Zircon	15

Figure 22

Location HH 76-3

The stratigraphic relationships among the six sections exposed at HH 76-3 are illustrated. The datum is the Snake River.

A---location of wood ^{14}C dated at >39,000 B.P. (GSC-3697)

B---location of wood ^{14}C dated at >40,000 B.P. (GSC-3800)

C---location of organic material ^{14}C dated at 8,510 ±70 B.P. (GSC-3695)

D---Disturbed and Colluviated Material
ED---Early Diamicton
EFG---Early Fluvial Sediments, gravel
FS---Fluvial Sand, age uncertain
FU---Fluvial Silt, age uncertain
HS---Holocene Fluvial Sand
KA---Cretaceous Arctic Red Formation Shale
PG---Preglacial Fluvial Gravel
YFG---Younger Fluvial Sediments, gravel
YFS---Younger Fluvial Sediments, sand
YGD---Younger Glacigenic Diamicton
YGD---Younger Glacigenic Diamicton
ZZ---Youngest Diamicton

FIGURE 22

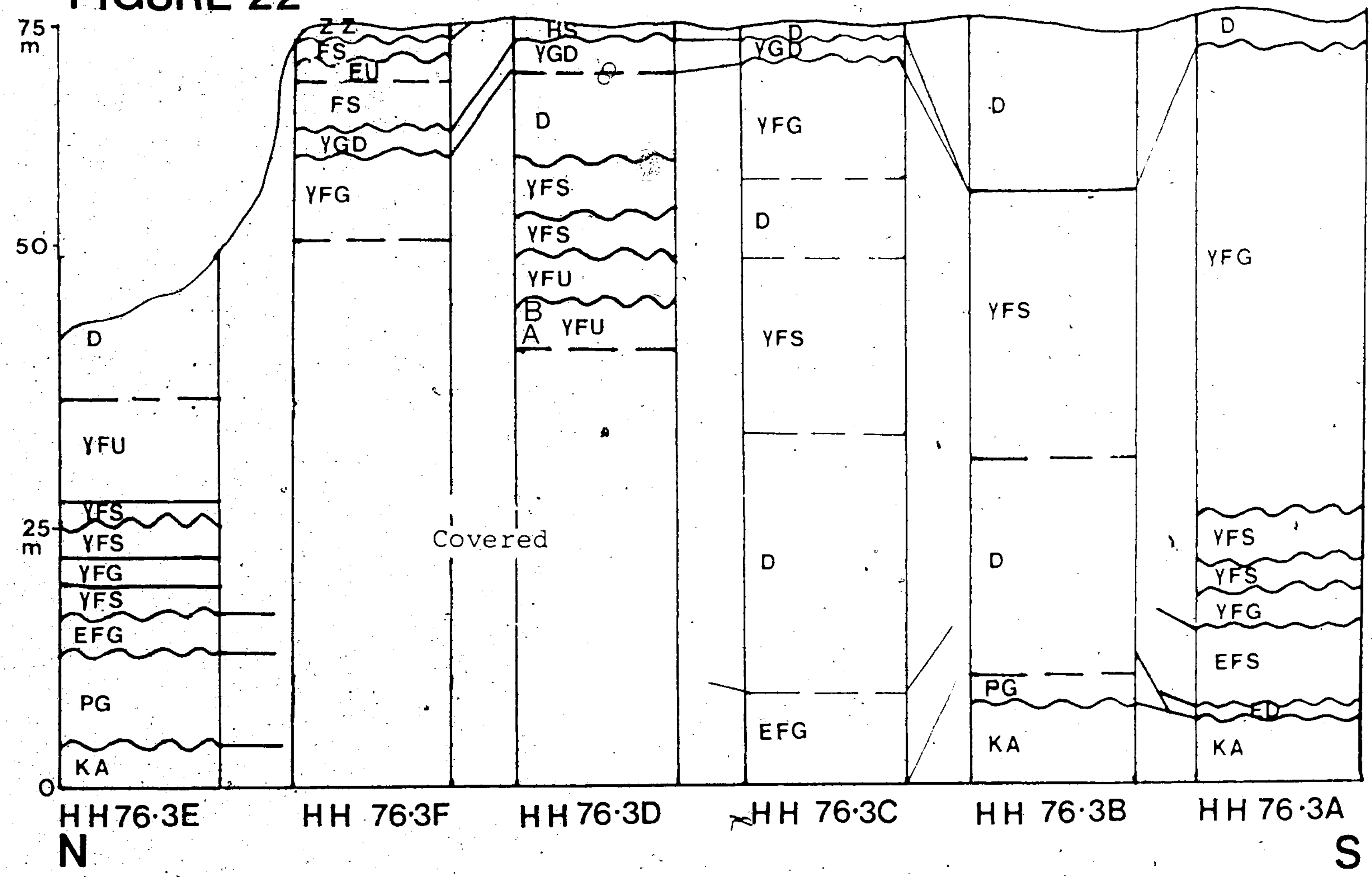

75 m
50
25 m
0
HH76·3E
HH 76·3F
HH76·3D
HH 76·3C
HH 76·3B
HH 76·3A
N
S
Covered
D
YFU
YFS
YFG
EFG
PG
KA
Z Z
FS
EU
YGD
HS
A
B
EFS

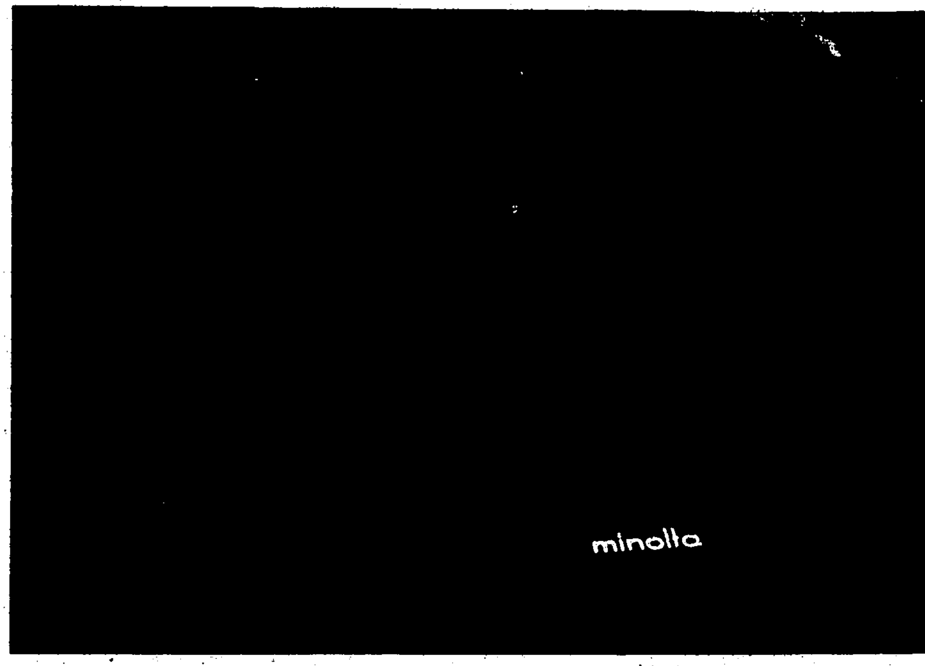
Plate 8 (top)

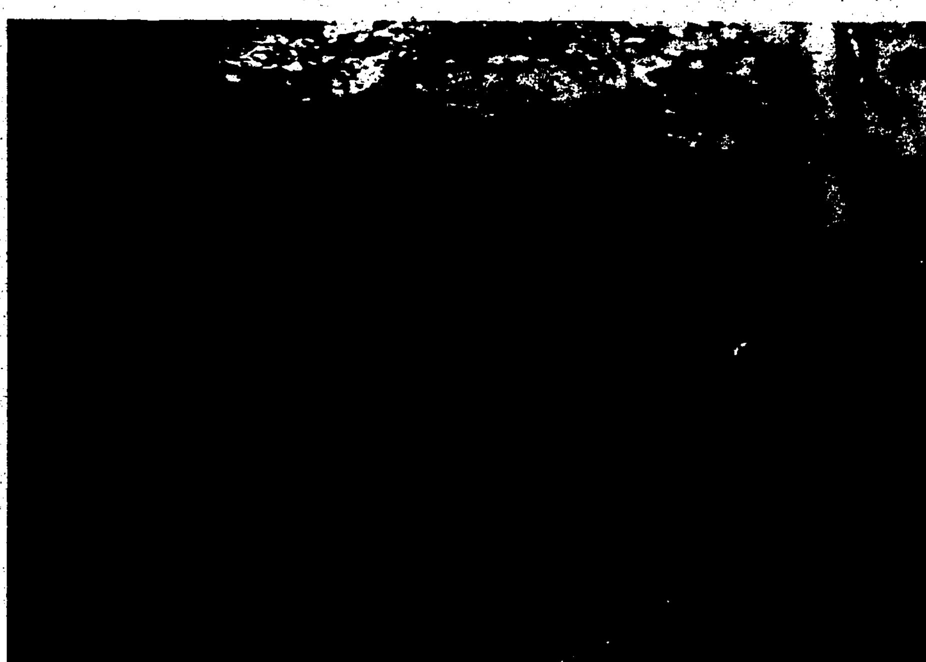
A part of the Younger Fluvial Sediment unit, HH76-3a. The lowest beds exposed (1) are a back-channel complex, dominated by sand. Overlying these beds is a sequence which fines upwards (2), interpreted as a deposit formed in a subsiduary channel. This sequence is succeeded unconformably by a second back-bar channel or abandoned channel sequence, dominated by sand and silt (3). An erosional event was followed by deposition of gravels on the foreslope of a lateral bar (4). Colluviated material caps the sequence. The entire succession is 20 m thick.

Plate 9 (bottom)

Large clast in medium--grained sand, Back--Bar Channel sediments, HH 76-3d. The clast has deformed the underlying sediments and laminae, suggesting that it was dropped into the sand. The cobble could have been carried on seasonal river ice. The cobble is 8 cm long. The position of the cobble in the back--bar channel succession is illustrated in Figure 23.

7





The primary channel gravel units are similar to those described previously in the preglacial and early fluvial successions. The sediments are predominantly structureless or crudely horizontally stratified cobble and pebble gravel beds, with isolated tabular cross-stratified granule beds.

The lateral bar sequences are characterized by interbedded horizontal coarse sand and granule units 2-5 cm thick, which grade laterally into tabular foreset cross-strata composed of coarse sand, granules, and pebbles. The pebbles are aligned transverse to the foresets, with intermediate axes oriented down the foreset slope. Trough cross-beds (Pi type of Allen, 1963a) are present in some coarse sand and granule beds.

The bar sequences are overlain by trough cross-bedded medium sand units 20-40 cm thick, which grade upwards to ripple cross-laminated fine sand and silt units 5-10 cm thick. Transverse aeolian ripple and aeolian adhesion ripple cross-laminations (Hunter 1973; Kocurek 1981) characterise the upper parts of the fine sand and silt. Scattered dreikanter are present along the upper surfaces of the bar sequences.

The lateral bar sequences are very similar to those currently forming along the Snake River (Appendix 4). The scale of the bar foreslope deposits suggests that a much greater volume of water and deeper scouring and pooling characterized the older stream, however.

Most secondary channels in braided-meandering systems are seasonal, occupied only during peak flood season. These channels have been referred to as "side channels" (Smith 1972; Eynon and Walker 1974) and "slough" and "inner channels" by Bluck (1976). In this study, the channels are termed "back-bar channels", reflecting their position relative to the main thalweg and the large lateral longitudinal and point bars developed along it.

Sedimentary sequences interpreted as back bar channel deposits (Figure 23) record gradually declining flow. The basal unit consists of well to moderately sorted fine cobble and coarse pebble gravel, with little matrix material. This unit grades vertically upwards into medium-fine pebble gravel with coarse-medium sand. Trough cross-stratification (Theta of Allen 1963a) is frequently present. The gradational upper contact of the coarse grained gravel is produced by infiltration of medium sand into the open framework of the pebbles and cobbles (Frostick et al 1984). Irregular mud intraclasts are frequently present. Similar sequences have been interpreted as minor channel deposits by many researchers (eg. Smith 1972; Shelton and Noble 1974; Ramos and Friend 1982; Kirk 1983; Johnson 1984).

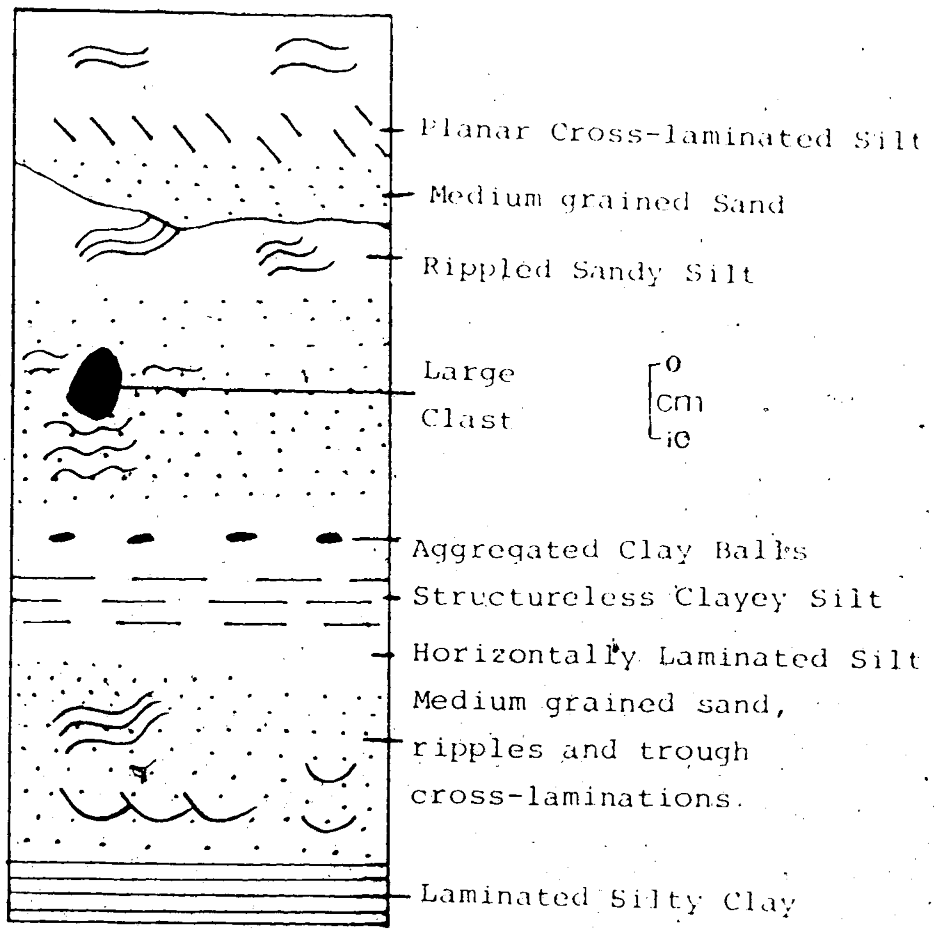
The trough cross-stratified coarse-medium sands are gradationally or abruptly overlain by medium-fine sands (Figure 23). Horizontal laminations, trough cross-laminations, ripple and climbing-ripple laminations, and contorted laminations are all present in various

Figure 23

Back-Bar Channel Assemblage, HH 76-3D

A typical back bar channel assemblage from the Younger Fluvial Sediments at HH 76-3D is illustrated. The back-bar channel successions are discussed in the text. Modern examples are discussed in Appendix 4. See also Plate 9.

FIGURE 23


Planar Cross-laminated Silt
Medium grained Sand
Rippled Sandy Silt
Large Clast
0
cm
10
Aggregated Clay Balls
Structureless Clayey Silt
Horizontally Laminated Silt
Medium grained sand, ripples and trough cross-laminations.
Laminated Silty Clay

deposits. Commonly, the medium-fine sand units are contained in complexes also incorporating finely horizontally-laminated sandy and clayey silt, rippled and planar cross-laminated sandy silt, micro-laminated silty clay, and structureless clayey silt and clay. Aggregated clay balls are rarely present as thin (less than 3 cm) horizontal strata. Isolated large clasts underlain by contorted sediments are also rarely present. These clasts may have been dropped on to the sediments by ablating seasonal river ice (Plate 9).

The contacts between units are generally erosional where horizontal or trough cross-laminated sand is overlain by silt or clay, or where medium sand overlies finer units. Gradational contacts occur between clayey silt and clay units, and between rippled sand and sandy silt.

The sands, silts, and clays are seldom arranged in consistent fining upwards or coarsening-upwards sequences. Instead, the vertical succession in most modern channels and in older deposits appears to be random. The units are thin (maximum 40 cm) and are laterally discontinuous. Sand units compose 20 - 50 per cent of the complex sediments, with silts representing 30 - 60 per cent.

The complexes are interpreted as the results of fluctuating energy levels in the channels, combined with ripple and dune migration and input of fine sediment from bank caving. The laterally and vertically discontinuous deposits indicate that flow and energy levels varied considerably and inconsistently. Once a back-bar channel is established, it will be re-occupied during each flood season. High energy events capable of transporting gravel would be relatively uncommon, however, as much of the energy of the floodwaters would be dissipated during flow over the bar adjacent to the main channel.

The back-bar channel sediments are very similar to the deposits found in the back-bar channels of the modern Snake River and other braided-meandering streams on the Peel Plateau (Appendix 4). The only significant difference between the modern and older deposits is the organic content. The beds of fine organic detritus characteristic of the modern sediments are not present in the older sedimentary record. Combined with the absence of marginal, bar-flank, and bar-head branch and log jam deposits (Appendix 4), this lack of organic material indicates that vegetation was scarce during the deposition of the sediments. No palynomorphs were recovered from any of the back-bar channel sediments.

Minor channels 0.5 - 1.5 m deep and 1 - 4 m wide are developed on the surface of the modern lateral bars. These channels are termed "bar-chute channels" (McGowen and Garner 1970; Bluck 1976; Blacknell 1982).

Sequences of erosionally-bound granular gravel, sand and silt similar to those present in the fluvial complex exposed at HH 76-3 (82) (Figure 24). The basal unit of the sequence is well-sorted granular gravel, 20 - 40 cm thick.

Figure 24

Bar-Chute Channel Assemblage, HH 76-3C

A typical bar-chute channel assemblage from the Younger Fluvial Sediments at HH 76-3C is illustrated. The successions consist of fining-upward sequences of granular gravel, sand, and structureless silt, locally capped with a sandy silt veneer. See text for discussion.

FIGURE 24

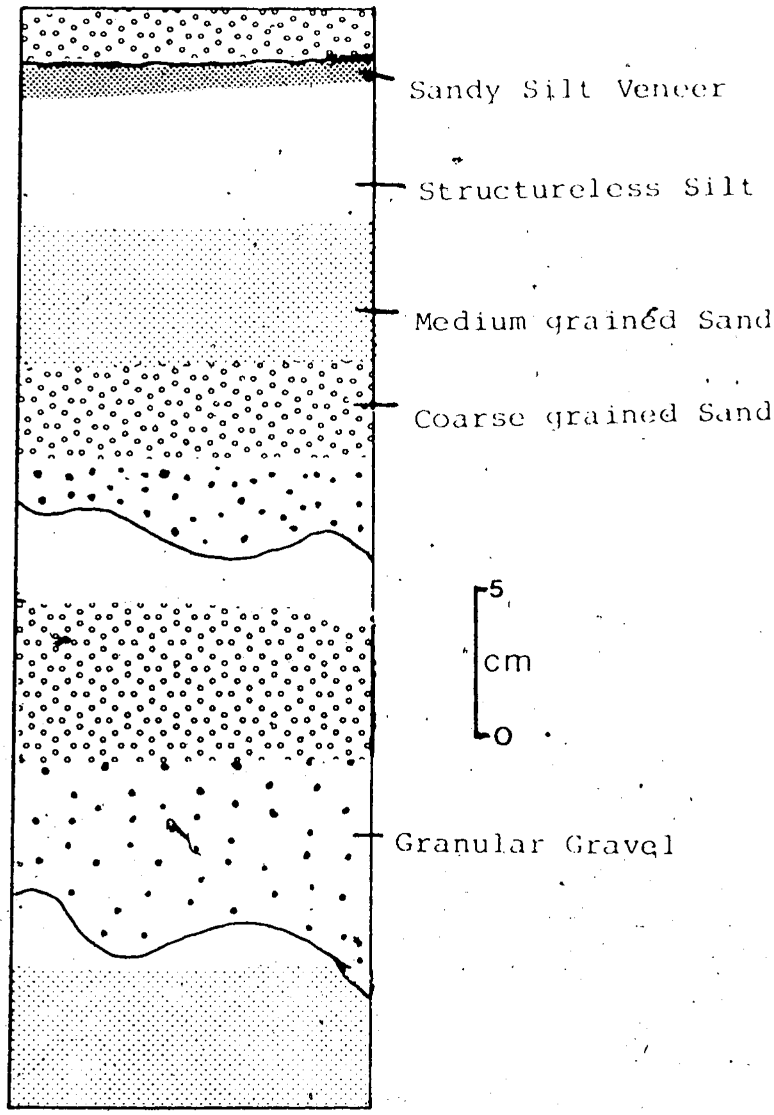
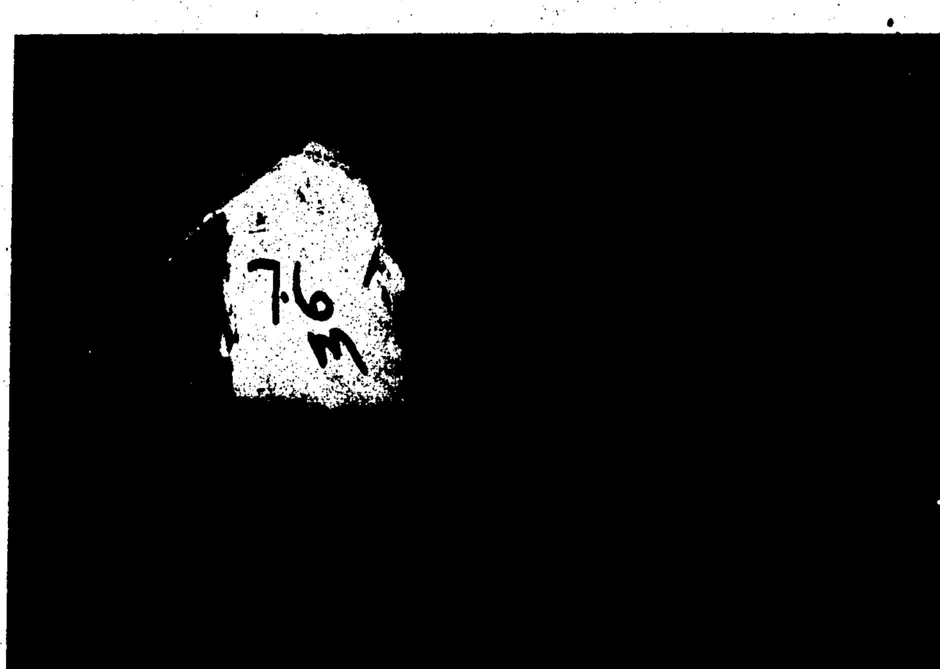

Sandy Silt Veneer
Structureless Silt
Medium grained Sand
Coarse grained Sand
5
cm
0
Granular Gravel

Plate 10 (top)

Bar Chute Channel-Back Channel Confluence Zone. The transition from coarse sand (base) to medium-fine sand (top) is illustrated. Note the distortion of medium-grained sand laminations in the chute channel unit (centre right), produced by overriding flow in the back-bar channel.

Plate 11 (bottom)

Rippled medium-coarse grained silt, Backflow Bar Sequence, HH 76-3d. The ripples were produced by backflow from the main channel. The overlying material is colluvium, probably derived from the Younger Glacigenic Diamicton.


7.6
m

The gravel is gradationally overlain by structureless moderately to well sorted medium to coarse-grained sand units 10-20 cm thick. The sands grade upwards into poorly sorted structureless silts, 5-15 cm thick. The silt units generally fine upwards, but are often capped by a veneer of sandy silt. Adhesion ripples (Hunter 1973; Kocurek 1981) are infrequently present in the silt strata. The silt units are overlain along erosional contacts by gravel or sand.

The sequences were produced by steadily decreasing flow in the bar-chute channels, as indicated by the gradational contacts between the units. Abandonment of the channels permitted the development of the aeolian adhesion ripples and the deposition of loess, producing the sandy silt veneers.

At section HH 76-3c (82), a different sedimentary sequence is preserved. The basal unit consists of massive, poorly sorted silt and clay, interpreted as a slough infill deposit. Similar sediments are present in sloughs associated with modern braided-meandering streams in the Peel Plateau region (Appendix 4).

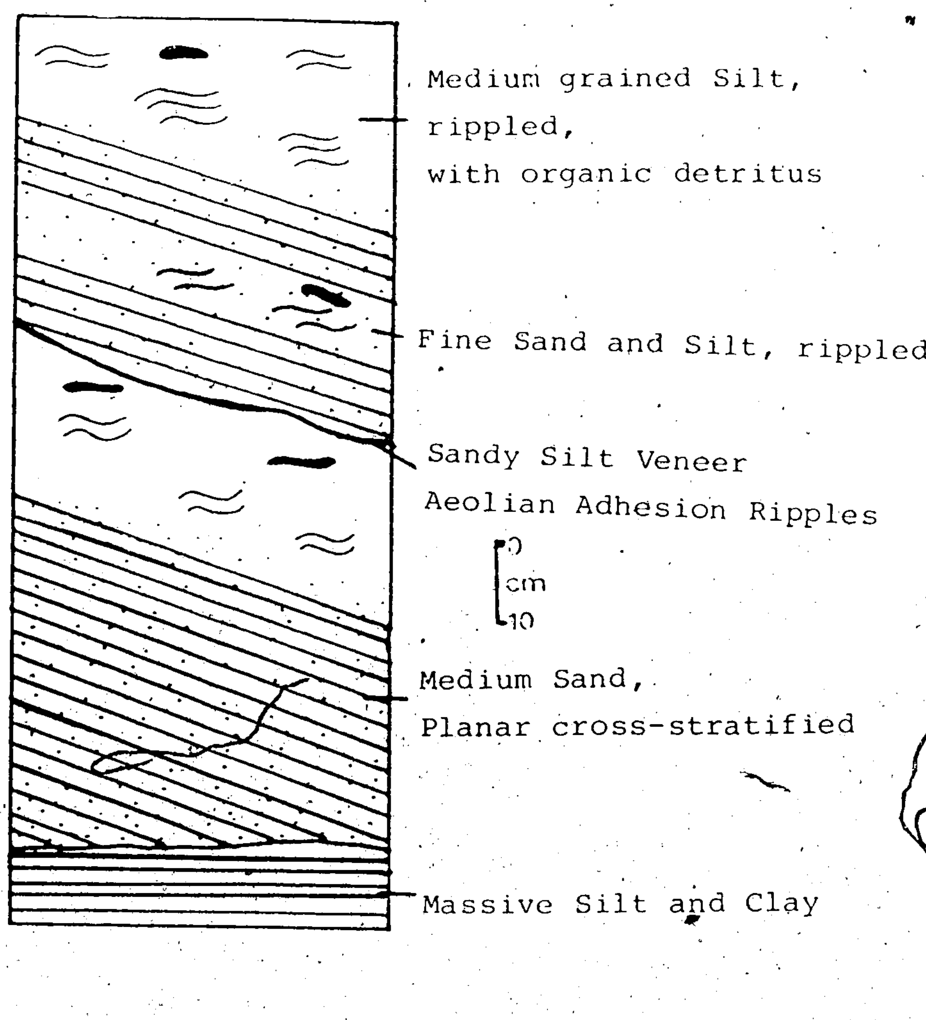
Overlying the massive fine sediments are gently-inclined (5°-15°) planar cross-stratified medium sand beds, ranging from 15 - 40 cm in thickness (Figure 25). These sand beds are interstratified with asymmetric ripple cross-laminated fine sand and coarse silt, draped with medium silt.

Figure 25

Backflow Bar Sequence, HH 76-3C

The figure illustrates a typical backflow bar sequence from HH 76-3C. Massive silt and clay, deposited in the slough channel, is overlain along an erosional contact by planar cross-bedded medium sand and rippled fine sand and silt. Medium-grained silt overlies the sand deposits. The sequences are capped by sandy silt veneers, which commonly contain aeolian adhesion ripples. See text for further discussion.

FIGURE 25


Medium grained Silt,
rippled,
with organic detritus
Fine Sand and Silt, rippled
Sandy Silt Veneer
Aeolian Adhesion Ripples
0
cm
10
Medium Sand,
Planar cross-stratified
Massive Silt and Clay

The sequences preserved are similar to those present in crescentic bars developed in the confluence zone of the main and back-bar channels (Appendix 4). On the surfaces of the modern bars, asymmetric ripples and cross strata indicate that the bars are constructed by water flowing from the main channel into the back-bar channel. The orientations of the cross-strata on the bar foreslopes also indicate flow from the main channel. The sedimentary sequence exposed at HH 76-3c (82) is erosionally-bounded, and therefore the orientations of the cross-strata cannot be compared with those developed in the main channel at the time of deposition. The similarity of the sequence to those of modern backflow bars, however, suggests similarities in genesis.

A composite model of the braided-meandering stream environment prevalent during deposition of the younger fluvial sediments is presented as Figure 26 (see also Plate 12). The sediments preserved are dominated by primary channel gravel units, lateral bar sequences, and back-bar channel deposits. Bar-chute channels and confluence bar deposits are also preserved.

Although preservation of complete lateral cross-sections is rare, comparison of the exposures with modern braided-meandering stream sequences has revealed sufficient similarities to consider the latter to be a modern analogue for the fluvial sediments. The streams which produced the sediments were characterised by higher volumes

Figure 26

Braided Meandering Stream Environments, Section HH 76-[illegible]

The environment of deposition of the Younger Fluvial Sediments at HH 76-[illegible] was a braided-meandering stream, similar to the modern Snake River. Sediments were deposited in the main channel, in back-bar channels, in bar-chute channels, in sloughs, and in confluence bars. These sub-environments are illustrated in Figure 26. A plan view of a modern braided-meandering stream reach is depicted in Figure A4-11, Appendix 4. See text and Appendix 4 for further discussion.

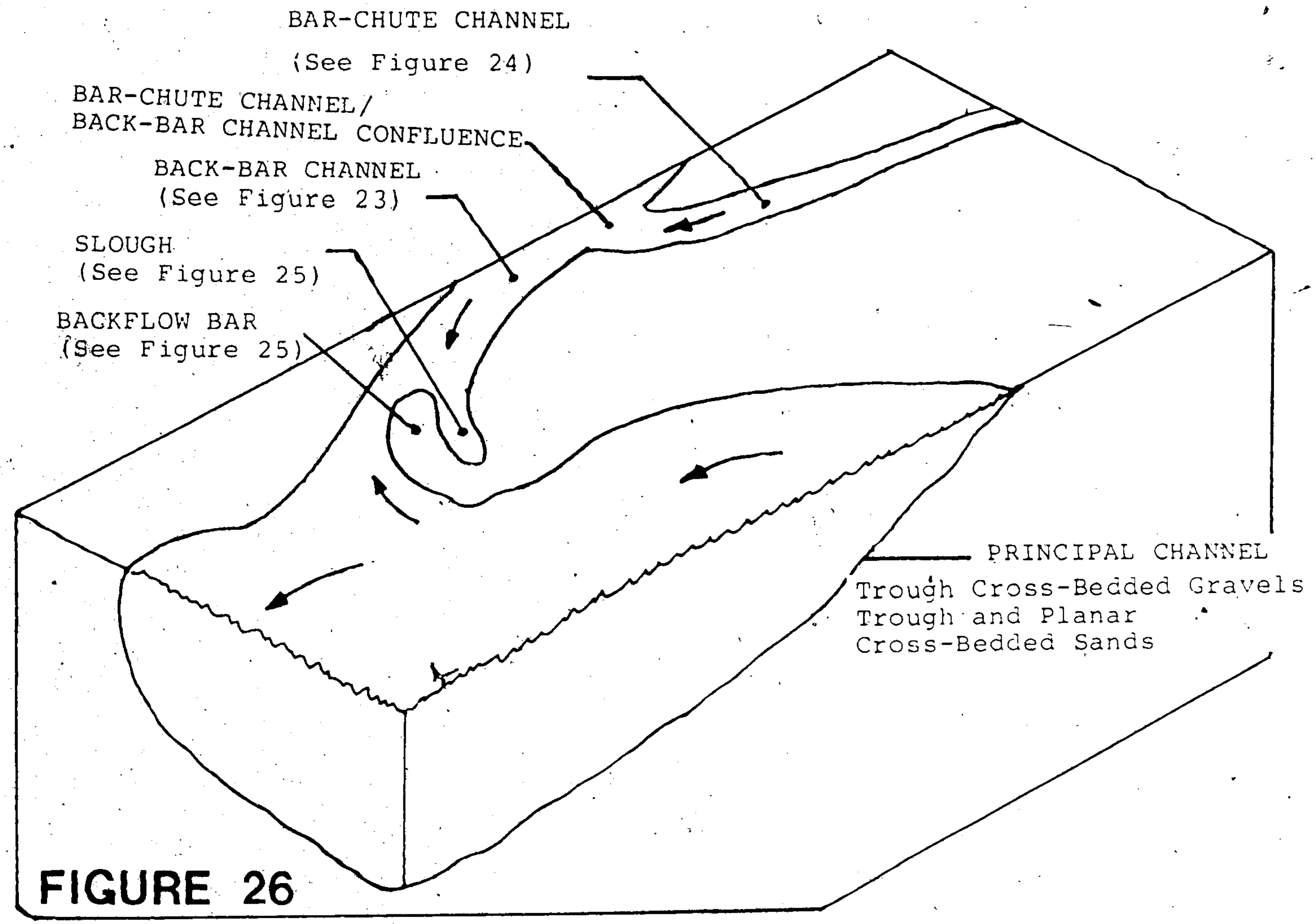
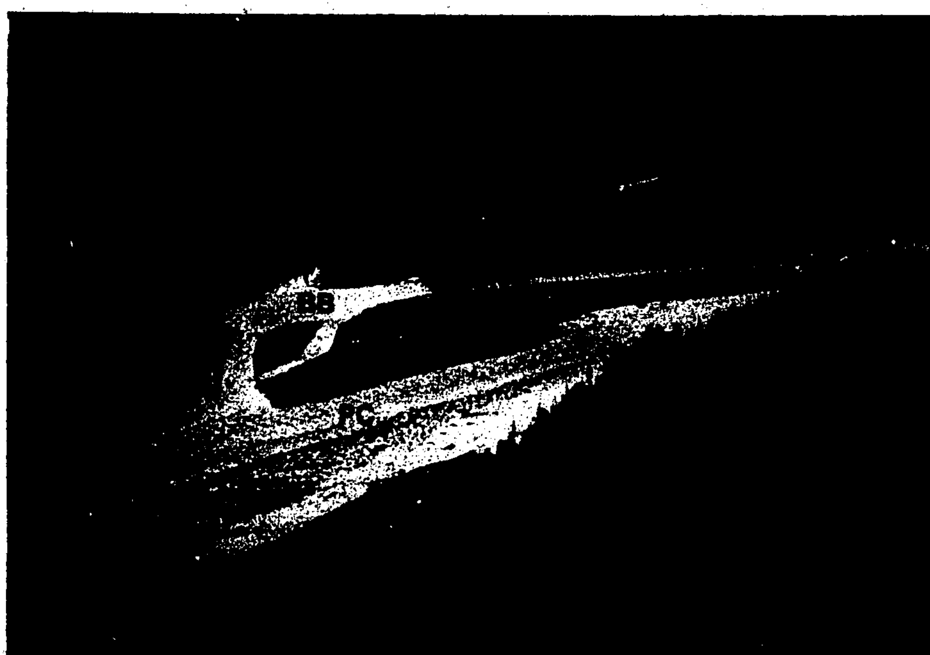

BAR-CHUTE CHANNEL
(See Figure 24)
BAR-CHUTE CHANNEL/
BACK-BAR CHANNEL CONFLUENCE
BACK-BAR CHANNEL
(See Figure 23)
SLOUGH
(See Figure 25)
BACKFLOW BAR
(See Figure 25)
PRINCIPAL CHANNEL
Trough Cross-Bedded Gravels
Trough and Planar
Cross-Bedded Sands

FIGURE 26

Plate 12

A Braided Meandering Reach, Snake River. Illustrated are the principal channel (PC), a subsidiary channel (SC), a back bar channel (BBC), and two lateral bars (LB). Bar-Chute channels are marked by accumulations of logs and shallow depressions on the lateral bar in the centre ground. The highest part of a crescentic backflow bar (BB) is barely visible at the confluence of the back ~~bar~~ channel and the subsiduary channel. See text and Appendix 4 for further discussion.



and flow rates than the modern Snake River. Scarcity of vegetation during the period of deposition is suggested by the absence of branch and log jam deposits and the lack of palynomorphs in any of the sediments.

Two fragments of Picea wood obtained from primary channel sediments 2.5 m and 8.0 m above the base of the fluvial sediments exposed at section HH 76-3d(82) have been ^{14}C dated (Figures 21 and 22). The wood fragments were dated at >39,000 B.P. (GSC-3697) and >40,000 B.P. (GSC-3800) respectively. The dates suggest that the sediments may have been deposited prior to 39,000 B.P. This conclusion cannot be regarded as definitive, however, as the dates reflect the ages of the Picea fragments rather than the times of deposition. Reworking of older fluvial deposits could result in the inclusion of old wood fragments in younger sediments.

H. Younger Glacigenic Diamicton

The younger glacigenic diamicton has been identified throughout the southern Peel Plateau, and is well-exposed along the Snake and Peel Rivers (Figures 27 and 28). Basal melt-out till and supraglacial ablation till have been recognized in several outcrops.

Basal Melt-out till

Basal melt-out till is characteristically silty, with very low concentrations of sand and large clasts. The proportion of quartzite clasts is high, fluctuating

Figure 27

Ice-Flow Direction Indicators, Southern Peel PlateauRichardson Mountains Region.

The principal exposures of the Younger Glacigenic Diamicton are indicated.

↗ Till Fabric Analysis, showing dominant mode

↗ Streamlined Features, direction of ice movement

FIGURE 27 ICE-FLOW INDICATORS

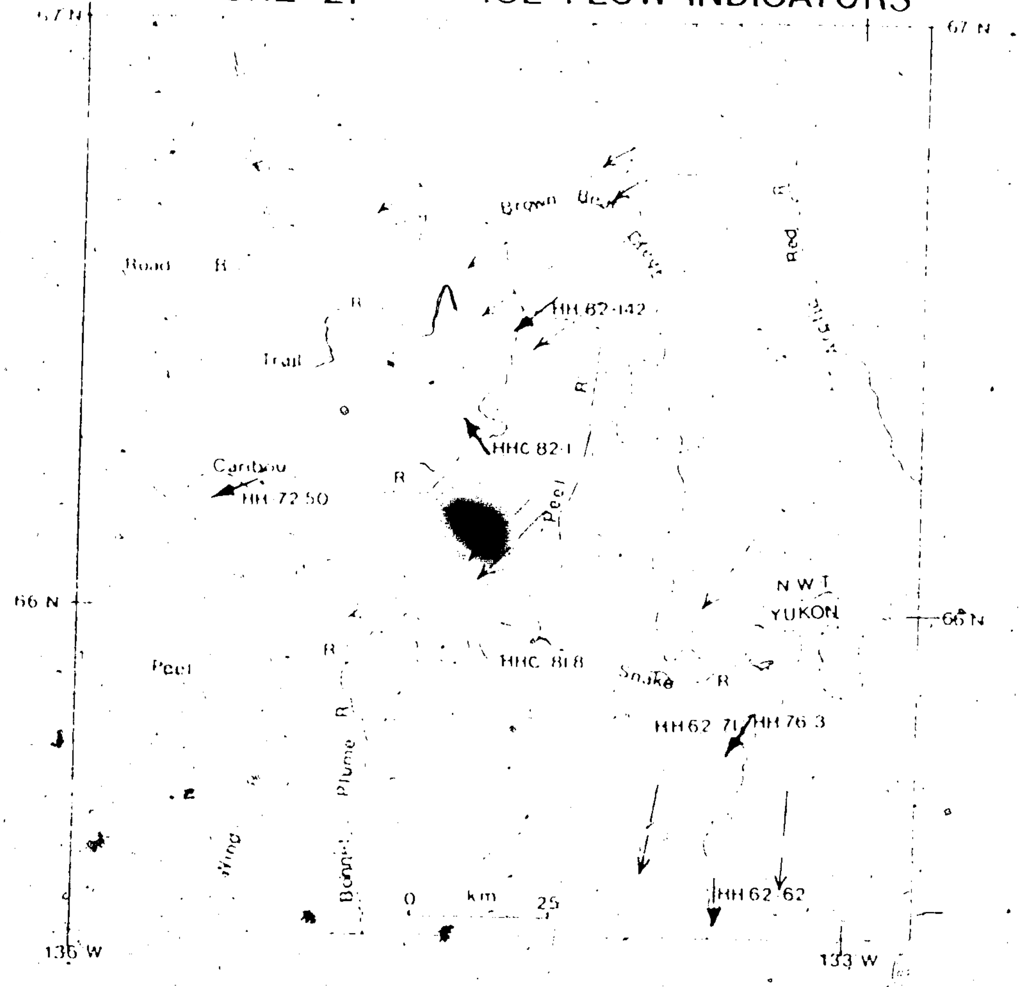


Figure 28

Stratigraphic Position of Younger Glacigenic Diamicton (Hungry Creek Till), Snake River Area

The stratigraphic position of the Younger Glacigenic Diamicton in exposures HH 62-62, HH 62-64, and HH 76-3F, and in the upper parts of HH 62-71, HH 76-3C, and HH 76-3D, is illustrated. The locations of the exposures are depicted in Figure 27.

C---location of organic material ^{14}C dated at 8,510 ±70 B.P. (GSC-3695)

D---Disturbed and Colluviated material

HS---Holocene Fluvial sand

HG---Holocene gravel, Alluvial Fan and Fluvial

HW---Holocene silt and organic detrital beds

KA---Cretaceous Arctic Red Formation Shale

PG---Preglacial Fluvial gravel

SS---Fluvial Sand and Silt, age uncertain

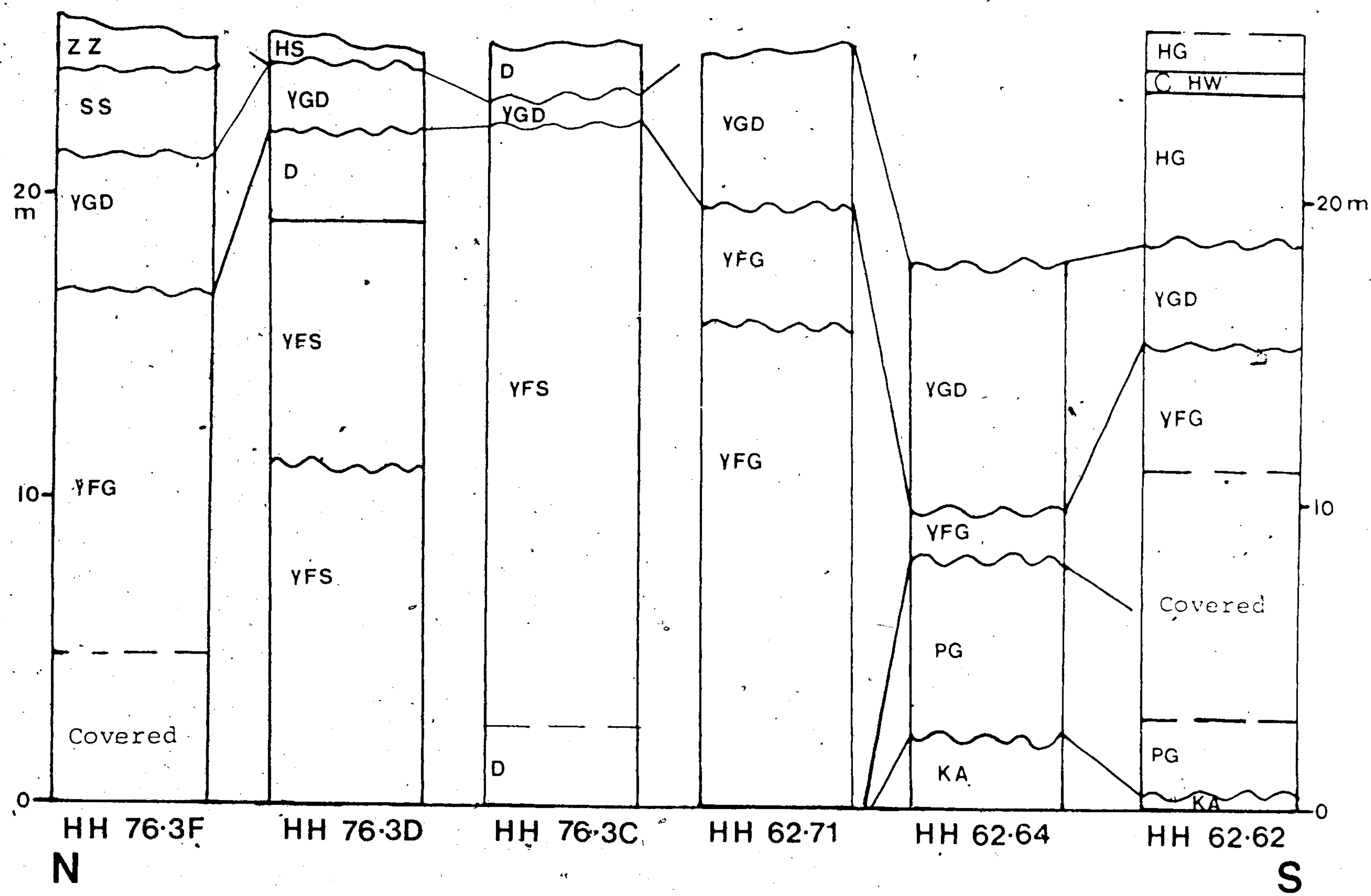
YFG---Younger Fluvial Sediments, gravel

YFS---Younger Fluvial Sediments, sand

YGD---Younger Glacigenic Diamicton

ZZ---Youngest Diamicton

FIGURE 28



20 m
10
0
ZZ
SS
YGD
YFG
Covered
HS
YGD
D
YES
YFS
D
YGD
YFS
D
YGD
YFG
YFG
YGD
YFG
PG
KA
HG
C HW
HG
YGD
YFG
Covered
PG
KA
20 m
10
0
HH 76·3F
HH 76·3D
HH 76·3C
HH 62·71
HH 62·64
HH 62·62
N
S

irregularly throughout the vertical extent of several exposures (Table 7). The concentrations of granule and coarse sand orthoquartzite clasts do not show a consistent pattern. The till contains fresh granite, diabase, gneiss, basalt, and pyroxenite granules and pebbles. Angular blocks of Arctic Red Formation shale and horizontally bedded and laminated unlithified sand; and granules and pebble-sized blocks of red shale and lignite are also present. Vertical zonation in the mineralogical and lithological assemblages is common. Figure 29 illustrates the variations of these parameters in a typical exposure, located at section HH 62-62 (82) (65° 28' N, 132° 26' W, Figure 27).

This exposure also displayed sand lenses, as well as ice-thrust structures and shear plane traces indicating ice flow towards the south.

The pebble fabric is unimodal in all exposures investigated, and is generally oriented south-southwest (170-220), a weak or negligible transverse component in the Snake River watershed (Figure 27). In the upper Peel River area, streamlined features suggest westerly flow.

These sediments are interpreted as basal melt-out tills, formed during ablation of an ice-sheet. The presence of angular unsorted blocks of weakly consolidated Arctic Red Formation shale, red Bonnet Plume Formation shale, lignite, and unlithified horizontally bedded sand indicates that lodgment of the clasts has not occurred. The subrounded to rounded large clasts and the absence of bullet- or

Table 7

Younger glacigenic diamicton, Snake River area, clast lithology and mineralogy.

A) Pebbles (8 analyses)

Lithology	Maximum	Mean	Minimum
	%	%	%
orthoquartzite	85	58	39
feldspathic sandstone	18	10	<1
siltstone	20	12	3
chert	2	1	0
argillite	30	15	3
jasper-hematite	5	2	1
limestone	2	<1	0
granite	6	<1	0
basalt	<1	<1	0
pyroxenite	<1	<1	0
gneiss	2	<1	0

Table 7 (continued)

B) Granules (10 analyses)

Lithology	Maximum	Mean	Minimum
	%	%	%
orthoquartzite	70	50	30
feldspathic sandstone	20	12	1
siltstone	22	10	1
chert	4	2	0
argillite	40	20	5
jasper-hematite	6	3	2
limestone	4	2	1
granite	7	< 1	1
granodiorite	< 1	< 1	0
diabase	2	< 1	0
gneiss	1	< 1	0
basalt	< 1	< 1	0
gabbro	< 1	< 1	0

Table 7 (continued)

C) Coarse sand (0.42 - 1.00mm), major minerals (16 analyses).

Lithology	Maximum	Mean	Minimum
	%	%	%
quartz	65	40	25
feldspar	50	30	20
argillite fragments	40	16	2
chert	10	4	1
calcite	10	2	1
hematite	3	< 1	< 1
dolomite	6	1	< 1
hornblende	8	4	2

Table 7 (continued)

D) Coarse sand, trace minerals.

Mineral	Number of analyses containing mineral
Actinolite	[illegible]
Andalusite	3
Ankerite	10
Apatite	14
Biotite	16
Bornite	2
Chlorite	14
Chromite	6
Cordierite	7
Corundum	16
Covellite	[illegible]
Epidote	3
Fluorite	14
Garnet	7
Gypsum	11
Ilmenite	6
Kyanite	12
Leucoxene	1
Magnetite	16
Muscovite	14
Pyrite	7
Pyrolusite	0
Pyroxene	16

Table 7 (continued)

Mineral	Number of analyses containing mineral
Rhodonite	4
Sphene	10
Tourmaline	16
Wollastonite	14
Zircon	6

Figure 29

Variations in Sedimentary Parameters in Till, HH 62 62 (82)

x---Sample Number

sed---clasts derived from sedimentary bedrock

i/m---clasts derived from igneous and metamorphic bedrock

feldsp----feldspar

ch---chert

[illegible]---carbonate minerals

hm---heavy minerals (s.g. >2.65, excluding carbonate minerals)

FIGURE 29 TILL, HH 62-62 (82)

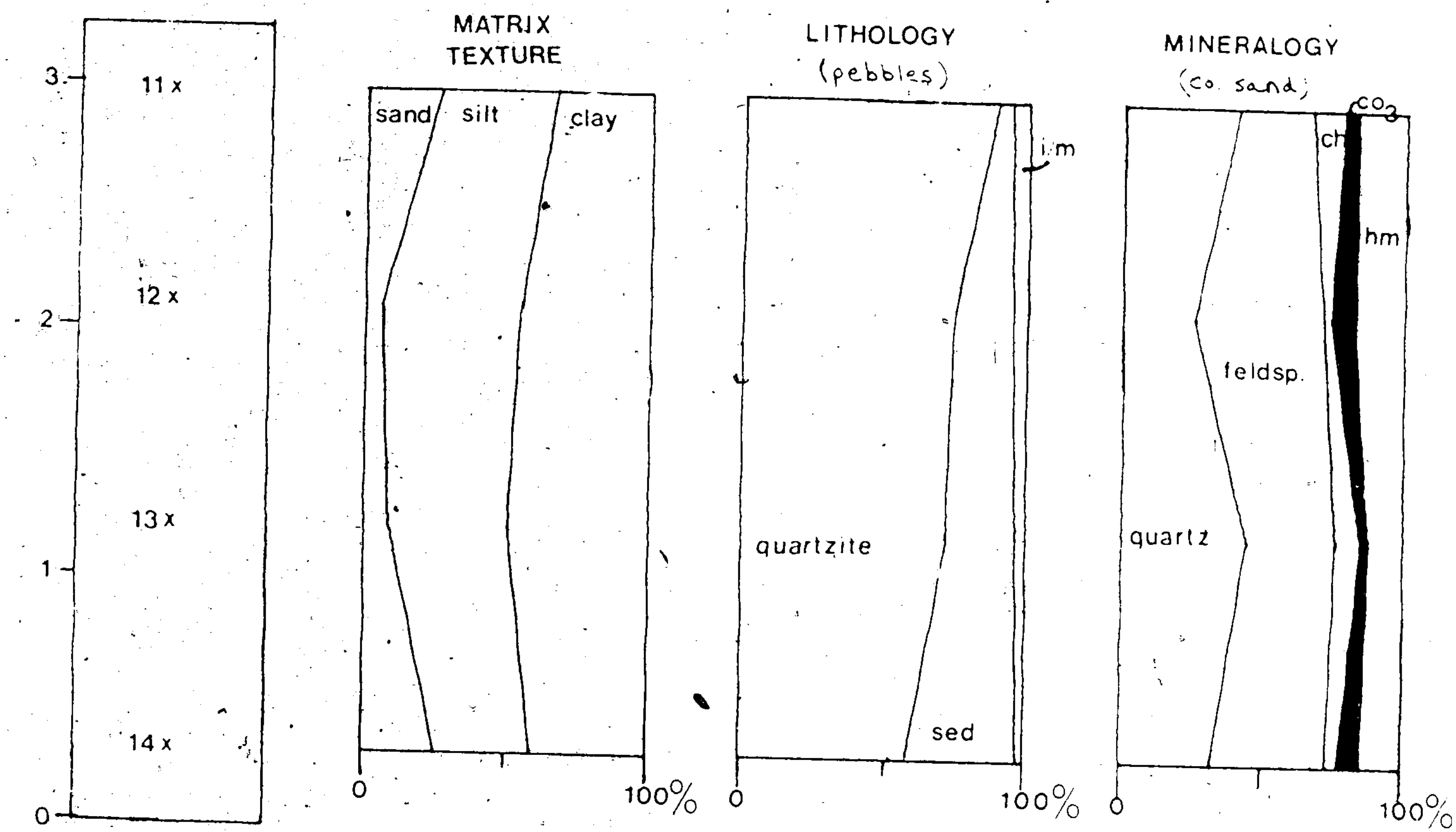
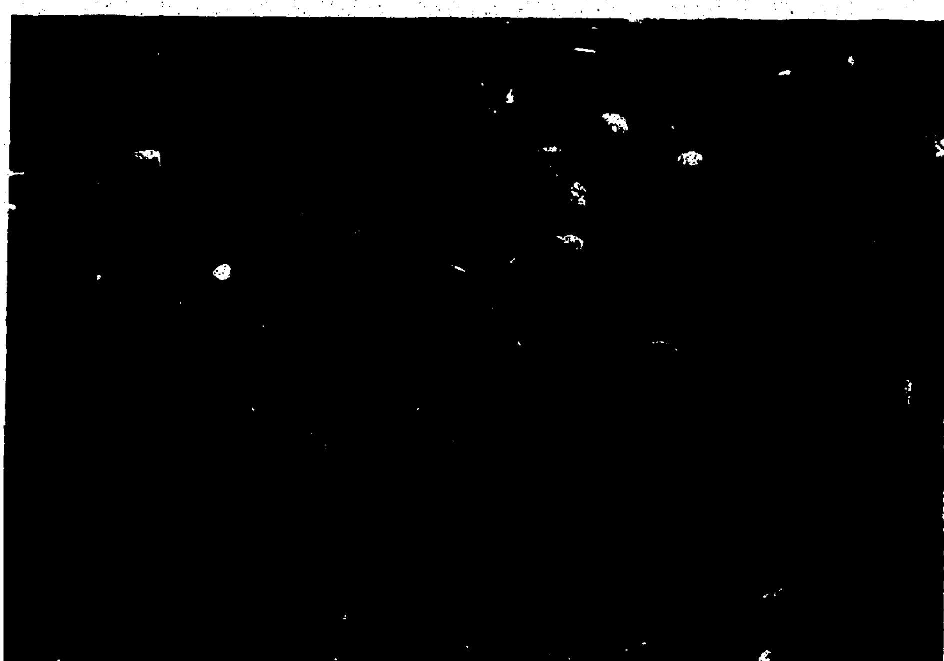


Plate 13 (top)

Younger Glacigenic diamicton, HH 76-3d. The diamicton is interpreted as a basal melt-out till, produced by an ice sheet which flowed to the south-southwest. Scattered sand lenses occur in the till unit.

Plate 14 (bottom)

Youngest Diamicton, HH 76-3f. The unit could be interpreted either as a till or mass-movement deposit. See text for discussion.



barrel-shaped pebbles commonly produced by lodgment (Boulton 1978; Kruger 1984) also suggest deposition by basal melt-out, as does the strongly oriented fabric (Lawson 1979; Haldorsen and Shaw 1982). Variation of the fabric orientation with elevation in the deposit exposed at HH 62-62 (82) is attributed to fluctuations in ice motion direction. The pronounced lithological and mineralogical vertical zonation observed in some exposures also is compatible with basal ablation of successive strata of ice.

The general absence of sand lenses, usually regarded as an important supporting criterion for basal melt-out till identification (eg. Harrison 1957; Kruger 1979; Shaw 1982) is probably a consequence of the general scarcity of sand throughout the deposits. Sand was either absent from the original sediment load transported by the ice, or it has been selectively concentrated at the glacier base and/or margins. Development of a network of subglacial and englacial streams (cf. Walder and Hallet 1979; Collins 1979) would cause the sand and coarser components to be selectively deposited in the channels, leaving the surrounding glacier base with reduced sand concentrations. The area occupied by the channels in modern glacial systems is small compared to the total basal area (Walder and Hallet 1979), and so channel sediments would be less abundant and thus likely to be exposed than would the marginal zone sediments. Examples of well-sorted subglacial channel fill sands surrounded by structureless silty till developed under basal melt-out conditions have been documented by Shaw (1982) and Catto (1984). The high concentrations of quartzite indicate that fluvial sediments along the Snake River valley were incorporated into the glacier sole during flow. The sediments underlying these tills contain sand, both as discrete beds and as matrix material within the gravels, and it is unlikely that sand was absent from the glacier base.

The high proportions of coarse quartzite clasts, compared to the low amounts present in the older, underlying till, can be explained by postulating different flow directions for the two ice sheets. Flow during the second event was along the Snake River valley, and thus the overriding glacier sole was exposed to a lengthy succession of quartzite-bearing fluvial sediments. The quartzite concentrations therefore were produced through incorporation of previously deposited fluvial material. During the first event, the direction of flow was oblique to the valley, and thus the glacier had less opportunity to incorporate fluvially deposited quartzite. Exposures of till on the Peel Plateau outside the immediate vicinity of the Snake River would not be expected to contain as much quartzite as the valley deposits. Therefore, while the quartzite concentrations are useful for distinguishing tills of the two events within the Snake River valley, they would be less useful in the surrounding region.

The presence of lignite and red shale clasts in the till indicates that sediment has been previously transported from the Bonnet Plume Basin, the nearest source of lignite (Norris 1982a). The shale clasts contain fragments of the Palaeocene organisms Osmundia and Sequoites, as identified by C.R. Stelck (University of Alberta, personal communication, 1984), also indicating Bonnet Plume provenance. The angular nature, size, and fragility of the clasts suggests that extensive fluvial transport did not occur. Blocks of lignite and red shale could have fallen or slid onto seasonal river ice, and then subsequently been transported east from the Bonnet Plume Basin through rafting during spring breakup. This mechanism may account for the presence in glacial sediments of angular blocks of friable material derived from downice areas.

Supraglacial ablation deposits

Recognition of supraglacially-produced ablation deposits and differentiation of these from postglacial colluvium is difficult in the Peel Plateau. Identification of diamicton exposures as supraglacial ablation till is based primarily on the absence of well-oriented fabric and the high proportion of distally-derived clasts, such as igneous and high-grade metamorphic rocks and minerals. The potential for mis-identification or overlooking of supraglacial till is high, and the sediment is therefore of little value as a correlative tool. It is probable that some of the exposures rejected for intensive study because of apparent colluviation may represent supraglacially produced sediments undisturbed since deposition. Distinguishing between a silty diamicton ~~which~~ underwent colluviation on the glacier surface and one which was colluviated after glaciation is at best a tenuous exercise.

The younger glacigenic diamicton deposits in the Snake River watershed and the Upper Peel River are correlated to exposures west of the Trevor Range and in the Bonnet Plume basin documented by Hughes (1972) and Hughes <u>et al</u> (1981). Correlation is based upon stratigraphic position and the relationship of the tills to the regional geomorphology and ice-flow direction indicators (Figure 27). Wood fragments contained in fluvial sediments beneath this till have been ^{14}C dated at >39,000 (GSC-3697) and >40,000 (GSC-3800).

Hughes (1972) also correlated these tills between the Snake River area and the Bonnet Plume Basin. The unit has been termed the "Hungry Creek Till" (Hughes <u>et al</u> 1981) and is defined on the basis of an exposure along Hungry Creek in the Bonnet Plume Basin. The younger till in the Snake River area stratigraphic succession can therefore be considered as correlative to the Hungry Creek Till of Hughes <u>et al</u> (1981).

Streamlined Features

A major field of drumlins and drumlinoid features is located south of the Peel River west of the Trevor Range (Figure 24). Isolated drumlinoids are scattered throughout the Snake River Watershed.

The majority of the streamlined forms are drumlinoids, rather than well-fromed drumlins. The features range in length from 20 - 150 m, in width from 10 to 50 m, and are commonly less than 10 m high. Shapes are generally elliptical, the length:width ratios ranging from 1.2:1 to 4.5:1. Slopes parallel to the long axes are low, generally less than 5°. The slopes normal to the long axes are also generally less than 5°. For many features, the difference in slope angle between slopes parallel to the long axis is insufficient to confidently designate stoss and lee sides. No detailed investigations of drumlin and drumlinoid internal structure and composition were conducted.

Drumlinoid forms showing orientation indicate glacial flow was to the west in the Upper Peel River area and to the southwest and south in the Snake River Watershed. No cross-cutting features were noted, suggesting that each field was formed during a single glacial event.

Glaciofluvial Deposits and Landforms

Glaciofluvial deposits compose a very minor part of the sediments in the study region, and glaciofluvial landforms are uncommon. The major features have been previously delineated by Hughes (1972), as shown in Figure 9. Discontinuous kame terraces are present along the northern margin of the Canyon Ranges. Irregular kames are also present east of the Snake River, adjacent to the Trevor

Range.

No detailed investigations of the internal structure of the glaciofluvial landforms were attempted. The landforms examined are composed primarily of gravel, with associated sand and silt. Clasts derived from the Canadian Shield and other distal areas are common. Sedimentary structures within the glaciofluvial deposits indicate formation in high energy braided stream environments.

I. Youngest Diamicton

At section HH 76-3f (82) (65° 48' N, 133° 18.5' W) diamicton overlies fluvial sediment which in turn overlies the younger till unit exposed throughout the Snake River valley (Figure 30). The diamicton varies in thickness between 0.76 and 1.7 m, and is structureless and apparently colluviated. The concentration of granitic pebbles and granules is extremely low (approximately 1: 1200). The diamicton contains striated pebbles of micrite and greywacke, and chert and orthoquartzite pebbles with crescentic fractures. The surface of the deposit coincides with the level of the plateau.

This unit could be interpreted as either a till or as a mudflow deposit. The clast lithology and sand mineralogy are sufficently different from the underlying till to indicate that the two units are distinct (Table 8), and therefore the upper diamicton does not represent a mudflow generated from a topographically higher exposure of the stratigraphically

Figure [illegible]

HH [illegible]

This section contains the only exposure [illegible] Diamicton. The diamicton overlies clay and fluvial sediment which in turn overlies the Younger Glaciogenic Diamicton (Hungry Creek Till). The base of the section is located at approximately 330 m asl. See text for discussion.

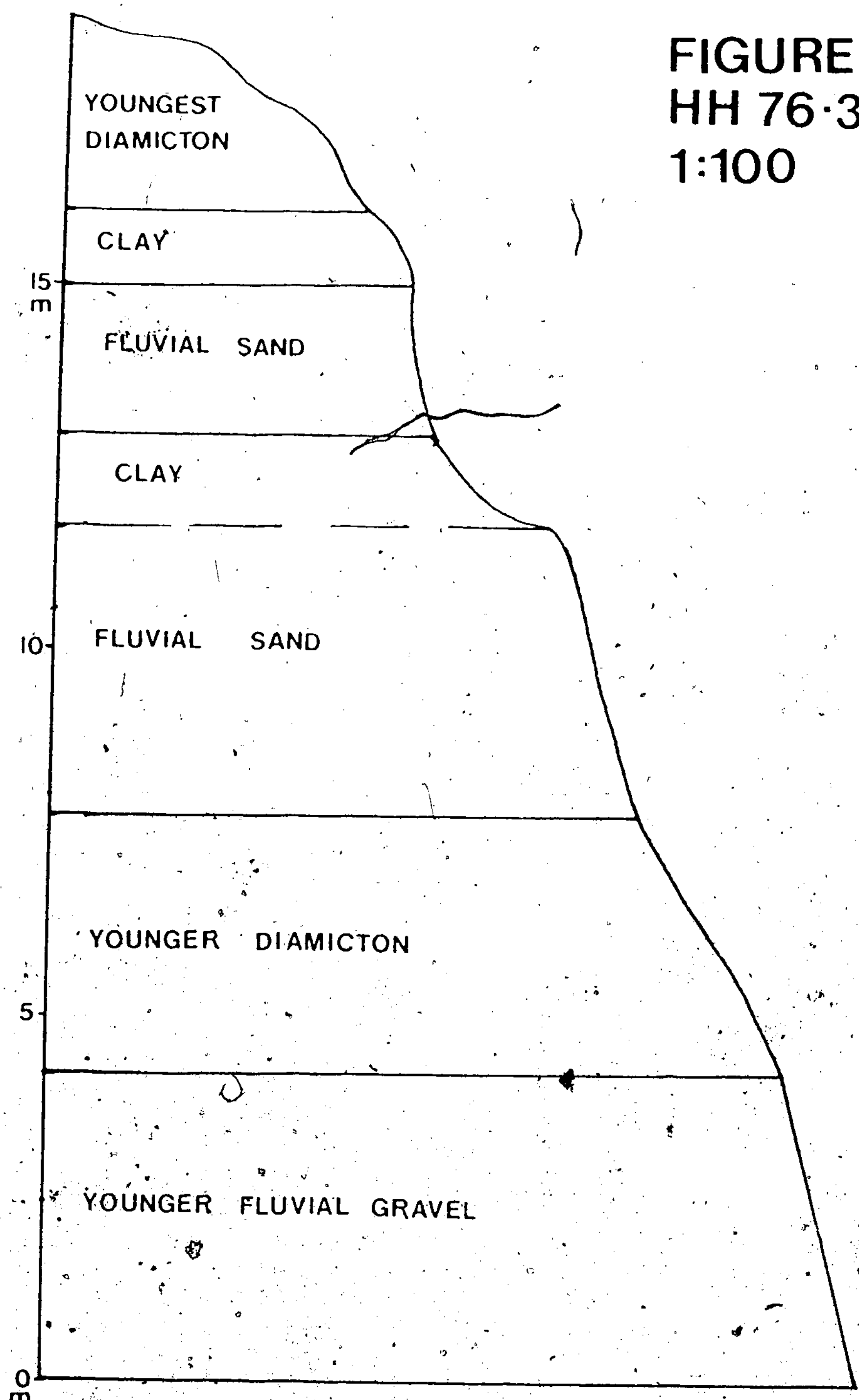


FIGURE 30
HH 76·3F(82)
1:100

Table 8

Youngest diamicton, location HH 76-3f(82), clast lithology and mineralogy.

A) Granules

Lithology	HH 76-3f, mean of 2 analyses	Younger diamicton[1], mean of 10 analyses
orthoquartzite	30	50
feldspathic sandstone	24	12
siltstone	8	10
chert	8	2
argillite	3	20
jasper-hematite	1	3
limestone	18	2
dolomite	6	0
granite	<1	<1
diabase	<1	<1
granodiorite	<1	<1
gneiss	<1	<1
gabbro	0	<1
basalt	<1	<1
lignite	1	0

[1]Refer to Table 7

Table 8 (continued)

B) Coarse sand (0.42 - 1.00mm); major minerals

Lithology	HH 76-3f, mean of 2 analyses	Younger diamicton¹, mean of 16 analyses
quartz	25	40
feldspar	53	30
argillite fragments	1	16
chert	8	4
calcite	8	2
hematite	<1	<1
dolomite	2	1
hornblende	2	4

¹ refer to Table 7

lower diamicton. The topographic position of the upper diamicton surface at plateau level also militates against this possibility. The sediment therefore represents a separate depositional episode.

If the unit is interpreted as till, it represents a minor readvance (15 - 20 km maximum) during the retreat of the glacier responsible for the deposition of the underlying Hungry Creek Till.

Interpreting the unit as a mudflow deposit implies a highly localised sediment source on an adjacent ice sheet. The striated and crescentic fractured pebbles have undergone prolonged traction and have been abraded by resistant clasts and/or substrate. These conditions are unlikely to occur during debris flow (Lowe 1979, 1982; Nardin et al 1979), but remobilization of material previously scored during glacial transport cannot be excluded. The low concentrations of granitic and other distal provenance clasts indicate that if a mudflow did occur, it probably originated at or near the base of the ice front and incorporated little or no material derived from the surface of the glacier.

J. Holocene Sediments

Holocene sedimentation in the Snake River Watershed and Upper Peel River area is dominated by braided-meandering stream deposits in the Snake River, braided-meandering and ingrown meandering stream deposits formed along the Peel River, and alluvial fan development. These environments are discussed in Appendices 4C, 4D, and 3 respectively.

The surface of the Peel Plateau is covered by a thin veneer of organic sediments and cryosolic soils developed on the Younger diamicton unit. No locations suitable for palynological analysis were noted in the Peel Plateau.

K. Summary

The sedimentary exposures of the Snake River Watershed and Upper Peel River area contain a record of preglacial fluvial activity, early glacial fluvial and colluvial activity, and two distinct glacial events separated by non-glacial conditions. A third glacial event may be represented at a single locality, HH 76-3f (82). A composite stratigraphic column for the region is presented in Figure 31.

The earliest sediments preserved are preglacial fluvial gravel and sand beds, containing no granitic clasts and showing indications of extensive weathering. The sediments were deposited in braided-meandering stream environments.

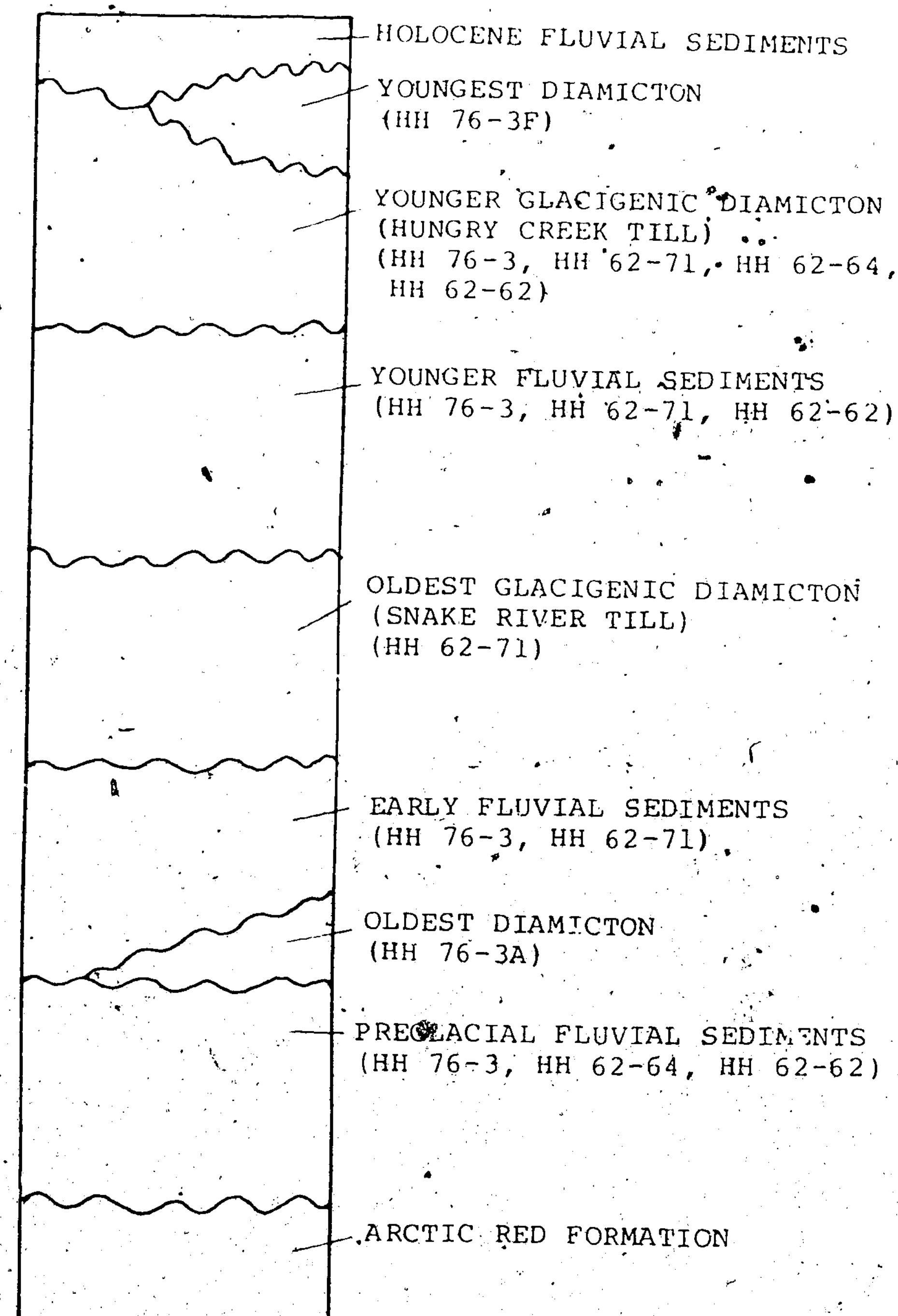
The overlying units contain granitic and high- and medium-grade metamorphic clasts, and therefore post-date the initial glaciation of the region from the Canadian Shield. The sequence is dominated by braided-meandering stream sediments. A single exposure of a diamicton unit directly overlying Arctic Red Formation shale is tentatively interpreted as a mudflow deposit.

Figure 37

Composite Stratigraphic Column, Snake River-Upper Peel River Area

The column is schematic, and the thicknesses of the units as illustrated do not represent either the thicknesses of sediment deposited or the amount of time represnted by each unit. The stratigraphic position of the Peel River Lacustrine Sediments is unknown, and therefore these sediments are not represented on the column. See text for further discussion.

FIGURE 31


HOLOCENE FLUVIAL SEDIMENTS
YOUNGEST DIAMICTON
(HH 76-3F)
YOUNGER GLACIGENIC DIAMICTON
(HUNGRY CREEK TILL)
(HH 76-3, HH 62-71, HH 62-64,
HH 62-62)
YOUNGER FLUVIAL SEDIMENTS
(HH 76-3, HH 62-71, HH 62-62)
OLDEST GLACIGENIC DIAMICTON
(SNAKE RIVER TILL)
(HH 62-71)
EARLY FLUVIAL SEDIMENTS
(HH 76-3, HH 62-71)
OLDEST DIAMICTON
(HH 76-3A)
PREGLACIAL FLUVIAL SEDIMENTS
(HH 76-3, HH 62-64, HH 62-62)
ARCTIC RED FORMATION

The oldest glacigenic diamicton overlies the early fluvial sediments, and is, in turn overlain by younger fluvial sediments. This diamicton was deposited by a glacier flowing towards the northwest, and is informally designated as the "Snake River till". The diamicton is coarse textured and has a low concentration of quartzite clasts. It is tentatively correlated to the lacustrine sediments of the "Deception Glaciation" recognized in the Bonnet Plume Basin by Hughes <u>et al</u>. (1981).

Lacustrine sediments exposed at locality HHC 81-7 along the Upper Peel River could represent an eastward extension of the "Deception" lacustrine event from the Bonnet Plume Basin. The stratigraphic position of this unit is unclear at present, however.

The younger fluvial sediments were deposited in braided-meandering streams similar to the modern Snake River. Two fragments of <u>Picea</u> wood obtained from the sequence have been ^{14}C dated at >39,000 B.P. (GSC-3697 [illegible] >40,000 B.P. (GSC-3800). These dates cannot be regarded as conclusive, however, as the time deposition of thefragments is not necessarily reflected by the age of the wood.

A younger glacigenic diamicton overlies the fluvial sediments. This till was deposited by a glacier flowing towards the southwest. The till is silty, and exposures along the Snake River contain high concentrations of quartzite clasts. The till is correlated to the "Hungry Creek Till" recognized by Hughes <u>et al</u>. in the Bonnet Plume

Basin. The till is exposed on the surface throughout much of the area, and drumlins and other streamlined features are associated with it.

The youngest diamicton in the Snake River Valley represents either a local readvance or mudflow event subsequent to the commencement of final glacial retreat. This unit is exposed only at locality HH 76-3f (82).

Holocene events in the area were dominated by valley cutting by major streams, deposition of fluvial sediments and the development of alluvial fans. Deposition of organic material and cryosol development on the Younger diamicton occurred on the upland surfaces of the Peel Plateau.

VII. Caribou River Watershed and Brown Bear Creek area

A. Introduction

The Caribou River Watershed and Brown Bear Creek area represent the central part of the study region (Figure 8). Exposures of the sediment are largely confined to the valley walls of the Caribou River and the east bank of the Peel River between the Caribou and Trail Rivers, although minor outcrops are present in the surrounding plateau.

Glacigenic deposits cover the Peel Plain and Peel Plateau in this region, and extend westward to the flanks of the Richardson Mountains. The distribution of surficial deposits in the area is illustrated in Figure 32.

The sedimentary record encompasses the Quaternary period from pre-glacial time to the Holocene. In the following sections of the thesis, the deposits will be discussed in chronological order.

B. Preglacial Sediments

Shale and siltstone of the Arctic Red Formation is overlain by gravel at several locations along the Caribou, Trail and Road Rivers (Figure 33). The gravel is dominated by orthoquartzite and other clasts derived from the Richardson Mountains, and no granitic or high-grade metamorphic clasts are present (Table 9). These data indicate that the sediments were deposited prior to the initial glaciation of the area.

Figure 32

Quaternary Geology, Cariboo River: Brown Bear Creek Area (after Hughes 1972; Hughes et al. [illegible], [illegible], [illegible]; and observations by the author)

Holocene Bog, standing water and organic sediments

Holocene Fluvial Deposits along major active streams, primarily sands and gravels

Holocene and Pleistocene Abandoned Channel Deposits, sand, clay, and organic sediments

Pleistocene Glacial Diamicton, with streamlined features showing orientation of glacial flow

Pleistocene Glacial Diamicton, hummocky terrain

Pleistocene Glacial Diamicton, with organic sediment veneer and blanket

Pleistocene Glacigenic Ridge and Valley Complex

Pleistocene Glaciofluvial Deposits, sand and gravel

Pleistocene Esker, sand and gravel

Pleistocene Glacial Deposits, scattered erratics and thin discontinuous diamicton cover

X Principal sections discussed in text

FIGURE 32

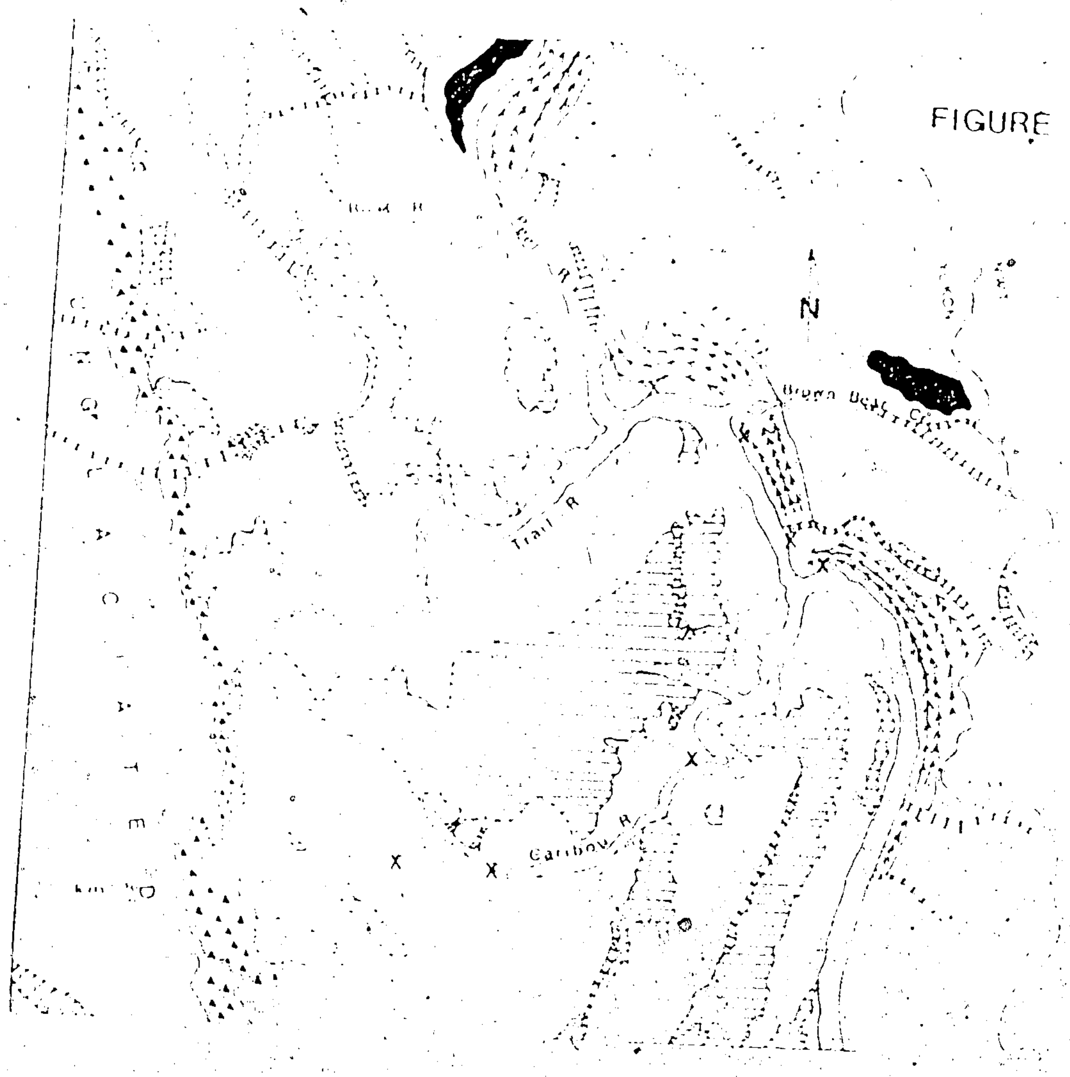


Figure 33

Locations of Preglacial Sediment Exposures, Caribou River Area

X---sections containing preglacial sediment exposures investigated in detail

x---other exposures of preglacial sediment

FIGURE 33

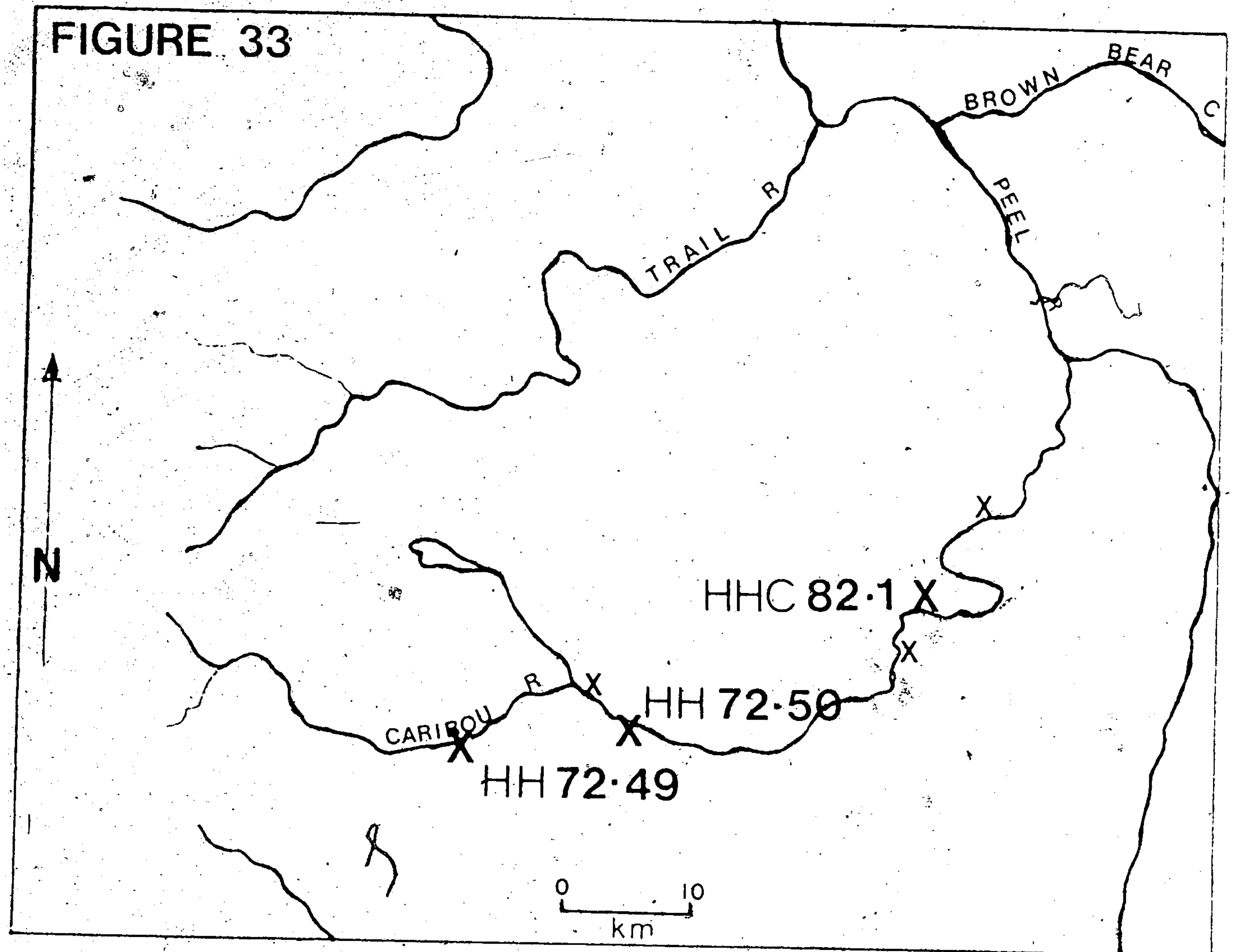

BROWN
BEAR
C
PEEL
R
TRAIL
R
N
HHC 82·1
HH 72·50
HH 72·49
CARIBOU
R
0
10
km

Table 9

Preglacial sediments, Caribou River area, clast lithology and mineralogy.

A) Pebbles and granules (7 analyses)

Lithology	Maximum	Mean	Minimum
	%	%	%
orthoquartzite	80	56	2
feldspathic sandstone	11	1	0
siltstone/shale	95	35	20
chert	8	1	0
argillite	32	7	0

B) Coarse sand (0.42 - 1.00mm), major minerals (8 analyses)

Lithology	Maximum	Mean	Minimum
	%	%	%
quartz	85	44	5
feldspar	38	13	3
argillite fragments	80	36	6
chert	16	6	2
calcite	2	<1	0
dolomite	<1	<1	0

The gravels are composed of structureless or crudely horizontally stratified pebble and cobble beds. The sequences are similar to those interpreted as longitudinal bar deposits by Miall (1977, 1978). Similar assemblages were observed in the modern braided-meandering stream lateral and longitudinal bars of the region (Appendix 4). The orientation of the clasts suggests that flow was generally towards the east.

A single palynological assemblage was obtained from the preglacial sediments exposed at HH 72-50 (82) (Figure 34). The sample contained largely pre-Quaternary grains (Table 10). The remaining 17.4% of the sample was composed of grains which could be either Tertiary or Quaternary. All of the grains had been transported and subjected to some abrasion, and many were fragmented. The identifiable portion of the assemblage is dominated by Lycopodium, Cystopteris, Cyperaceae, and Gramineae with minor amounts of Sphagnum, Botrychium, and bisaccate fragments. The latter grain segments resemble Picea, but no definite identification could be made. This assemblage is not believed to accurately reflect the botanical community in existence at the time of deposition.

The preglacial fluvial sediments are confined to the tributary valleys of the Peel River. The absence of preglacial sediments in the Peel River Valley indicates that this channel was excavated after the initial glaciation of the region.

Figure 34

HH 72-50 (82)

The stratigraphy of HH 72-50 is illustrated.

E---unit expanded on Figure 34A

p.6 --- palynological analysis, with sample number

Figure 34A (following page)

Units 5, 8, and 9, HH 72-50 (82)

Subunits of unit 8 are indicated as 8a, 8b, etc.

7v---degree of decomposition, von Post scale

FIGURE 34

HH 72-50 (82)

66°15'N 134°58'W

380 m ASL at base

1:80

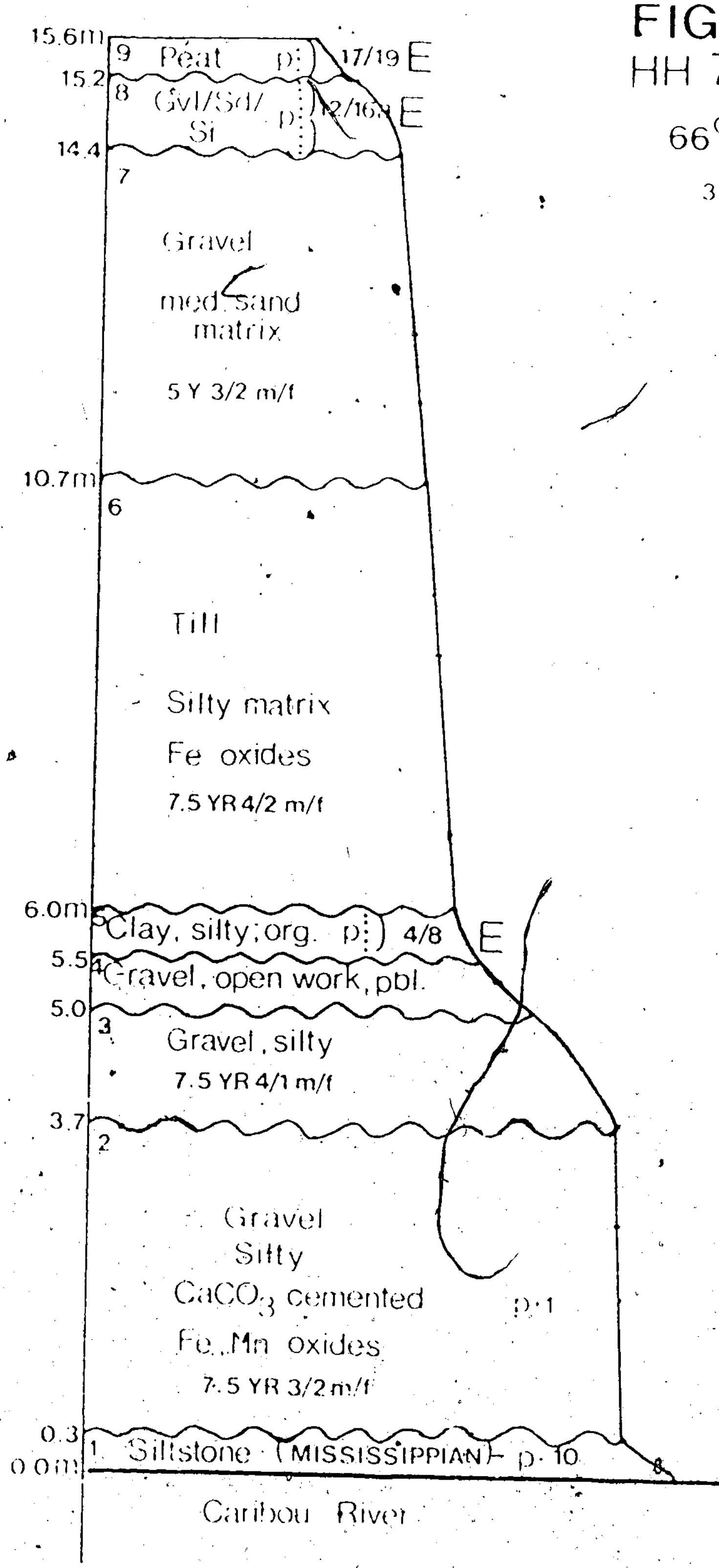


FIGURE 34A
HH 72-50 (82)
Units 5, 8, and 9
1 : 10

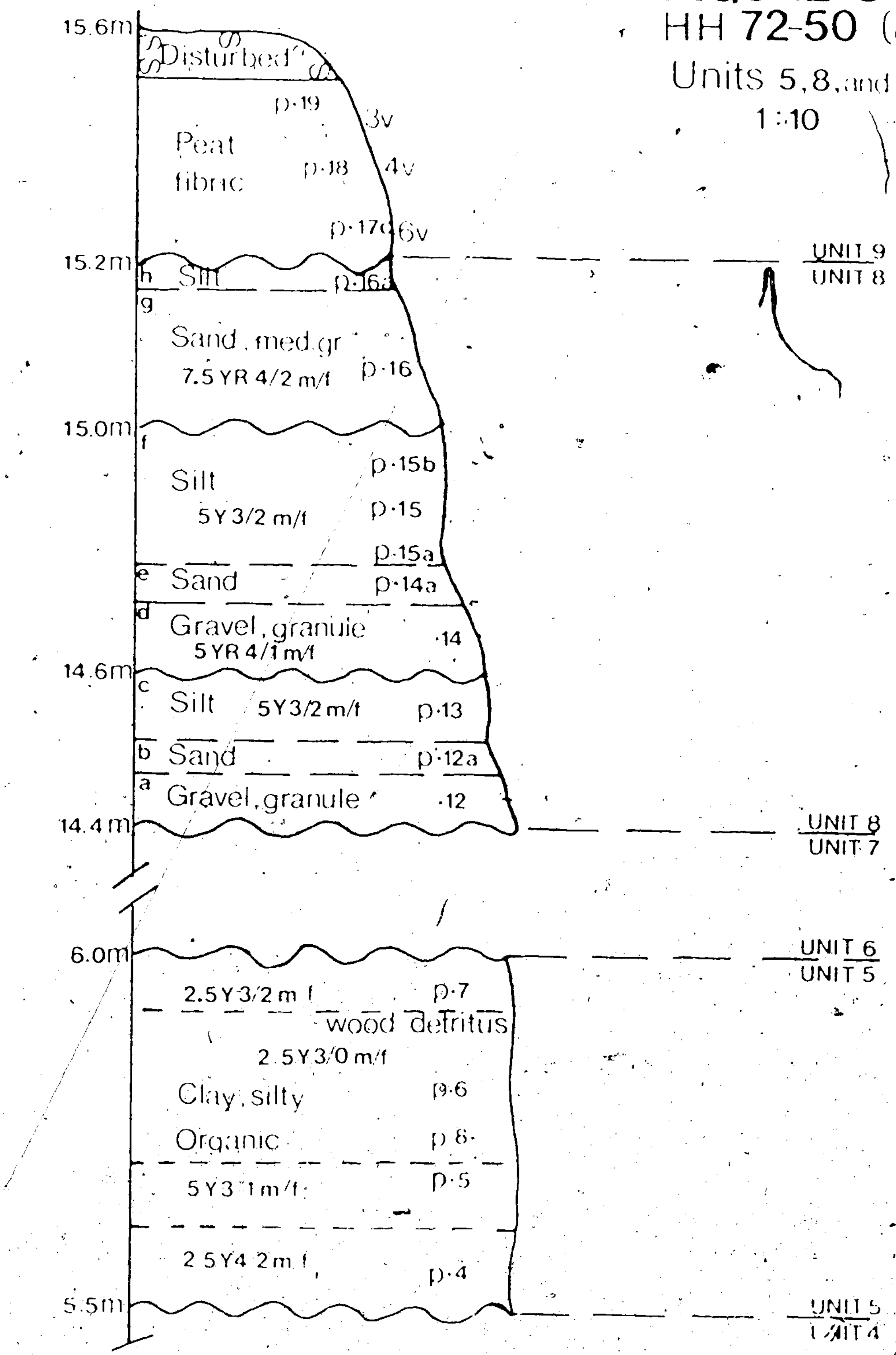


Table 10

Quaternary-Tertiary palynomorph counts, sample 1, pre-glacial sediments, HH 72-20 (82)

Taxon	Percentage
Lycopodium	5.3
Cystopteris	4.5
Cyperaceae	3.9
Gramineae	2.3
Sphagnum	1.0
Botrychium	0.2
Bisaccate fragments	0.2
Total	17.4
Pre Tertiary grains	82.6
Unidentified and indeterminate	102

C. "Brown Bear" Sediments

The "Brown Bear" Sediments are exposed along the Peel River in the vicinity of Brown Bear Creek (Figure 35, Plate 15). The sediments overlie Cretaceous Arctic Red Formation shale, except where incorporated into glacigenic ridge-and-valley complexes (as discussed below). Exposure is generally poor.

The sediments are predominately gravel, with granules, sand, and silt interbeds and lenses. All of the clasts are weathered and coated with iron and/or manganese oxides and calcium and magnesium carbonate, and many have weathering rinds. Thin saprolite layers are rarely present. The gravels are dominated by orthoquartzite, feldspathic sandstone, and shale clasts, but granitic and gneissic clasts are present (Table 11). Most of the heavy minerals originally within these clasts have been altered, and disaggregated masses of quartz and feldspar crystals are common.

The "Brown Bear" sediments appear to represent the remains of a braided-meandering sequence which has been extensively altered by erosion and weathering. Lateral bar assemblages dominate the sediments. and bar-chute and back-bar channel deposits similar to those of modern braided-meandering rivers in the study region (Appendix 4) are present.

Early Quaternary glaciation of the region is indicated by the presence of weathered granitic and gneissic clasts. No tills or other glacigenic deposits associated with these

Figure 9.

Locations of "Brown Bear" Sediment exposures, Brown Bear Creek Area

[illegible] Sections containing "Brown Bear" sediment exposures investigated in detail

[illegible] other exposures of "Brown Bear" sediment

FIGURE 35

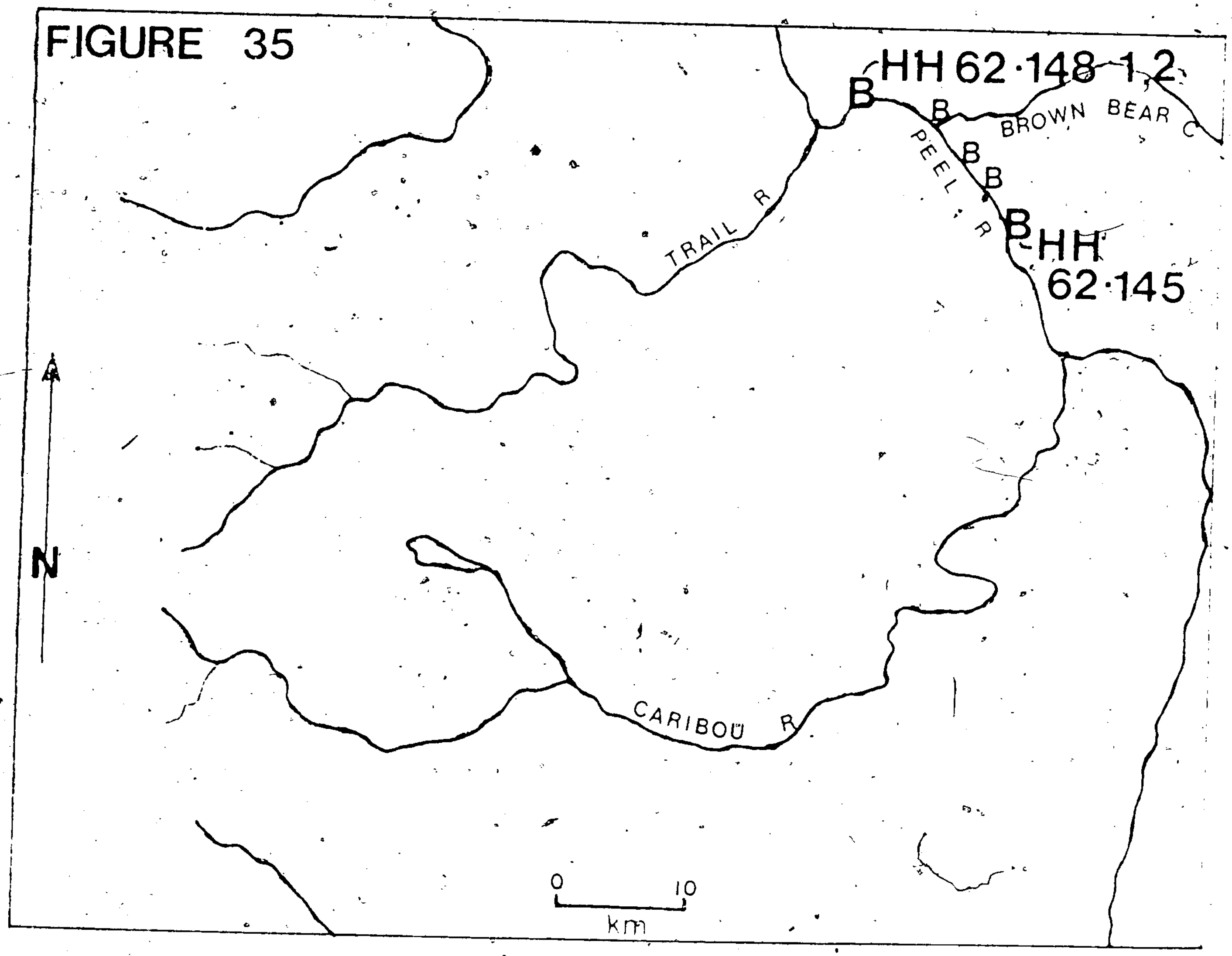

HH 62·148-1,2
B
B
B
B
B
HH
62·145
BROWN BEAR C
PEEL R
TRAIL R
CARIBOU R
N
0
10
km

Plate [illegible]

"Brown Bear" Fluvial sediments, BB 62-145. This bed is the least weathered exposure of "Brown Bear" sediment present in the study region. The large clasts are predominantly quartzite, but granitic clasts and heavy minerals derived from igneous and metamorphic rocks are present in the matrix.

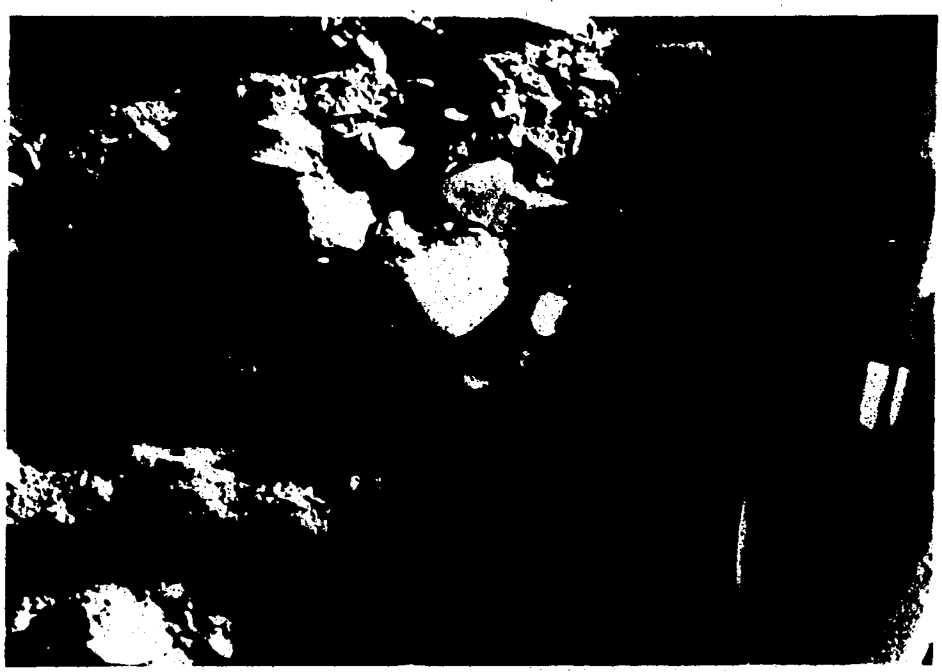


Table 11

Brown Bear sediments, clast lithology and mineralogy.

A) Pebbles (6 analyses)

Lithology	Maximum	Mean	Minimum
	%	%	%
orthoquartzite	65	56	45
feldspathic sandstone	10	4	3
siltstone/shale	30	21	12
chert	10	4	3
argillite	16	9	6
jasper-hematite	3	1	0
granite	1	< 1	< 1
diabase	5	4	2
gneiss	< 1	< 1	0

Table 11 (continued)

B) Granules (6 analyses)

Lithology	Maximum	Mean	Minimum
	%	%	%
orthoquartzite	56	48	44
feldspathic sandstone	6	3	0
siltstone/shale	32	19	10
chert	11	6	4
argillite	26	16	7
jasper-hematite	3	2	1
granite	2	< 1	< 1
diabase	7	3	2
gneiss	< 1	< 1	0
schist	2	< 1	0

Table 11 (continued)

(c) Coarse sand (0.42 - [illegible].00mm), major minerals (6 analyses)

Lithology	Maximum	Mean	Minimum
	%	%	%
quartz	75	60	50
feldspar	35	26	17
argillite fragments	12	7	6
chert	15	5	3
calcite	2	< 1	< 1
hematite	1	< 1	< 1
hornblende	2	< 1	< 1

Table 11 (continued)

D) Coarse sand, trace minerals.

Mineral	Number of analyses containing mineral
Apatite	1
Biotite	2
Chlorite	6
Corundum	5
Dolomitee	2
Fluorite	3
Garnet	2
Gypsum	5
Magnetite	3
Muscovite	4
Pyroxene	1
Sphene	3
Tourmaline	1

gravels have been recognized. The areal extent of the glaciation is unknown.

The age of the sediments cannot be quantitatively ascertained at present. Saprolite development and disintegration of granitic clasts are usually considered to signify extended periods of weathering (Boyer and Pheasant 1974; Dyke 1979), as is the development of weathering rinds (Porter 1975; Colman and Pierce 1981). Although temporal estimates from weathering phenomena contain large uncertainties due to the influence of many climatic and lithologic factors (Brookes 1985), the slowness of chemical weathering under periglacial conditions (Loughnan 1969) and the uniformity of the alteration phenomena throughout the "Brown Bear" depoits suggests an old, early glacial age. The absence of correlative glacigenic and other deposits prevents assessment of the amount of time between glaciation and deposition of the sedimentary sequence.

D. Younger Fluvial Sediments

Fluvial sedimens which postdate the "Brown Bear" sediments and predate younger glacigenic sediments are present at section HH 72-50 (82) along the Caribou River (Figure 34), and at locations along the Peel River (Figures 36 and 37). The maximum thickness of the sediments is 50.1 m, at location HH 62-142 (81).

The sediments are predominately gravel and sand, with associated clay and silt beds. The clasts are derived igure 36

Younger Fluvial Sediment Exposures, Cariboo River Brown Bear Creek Area

X---Sections containing Younger Fluvial Sediment nvestigated in detail

x---other exposures of Younger Fluvial Sediment

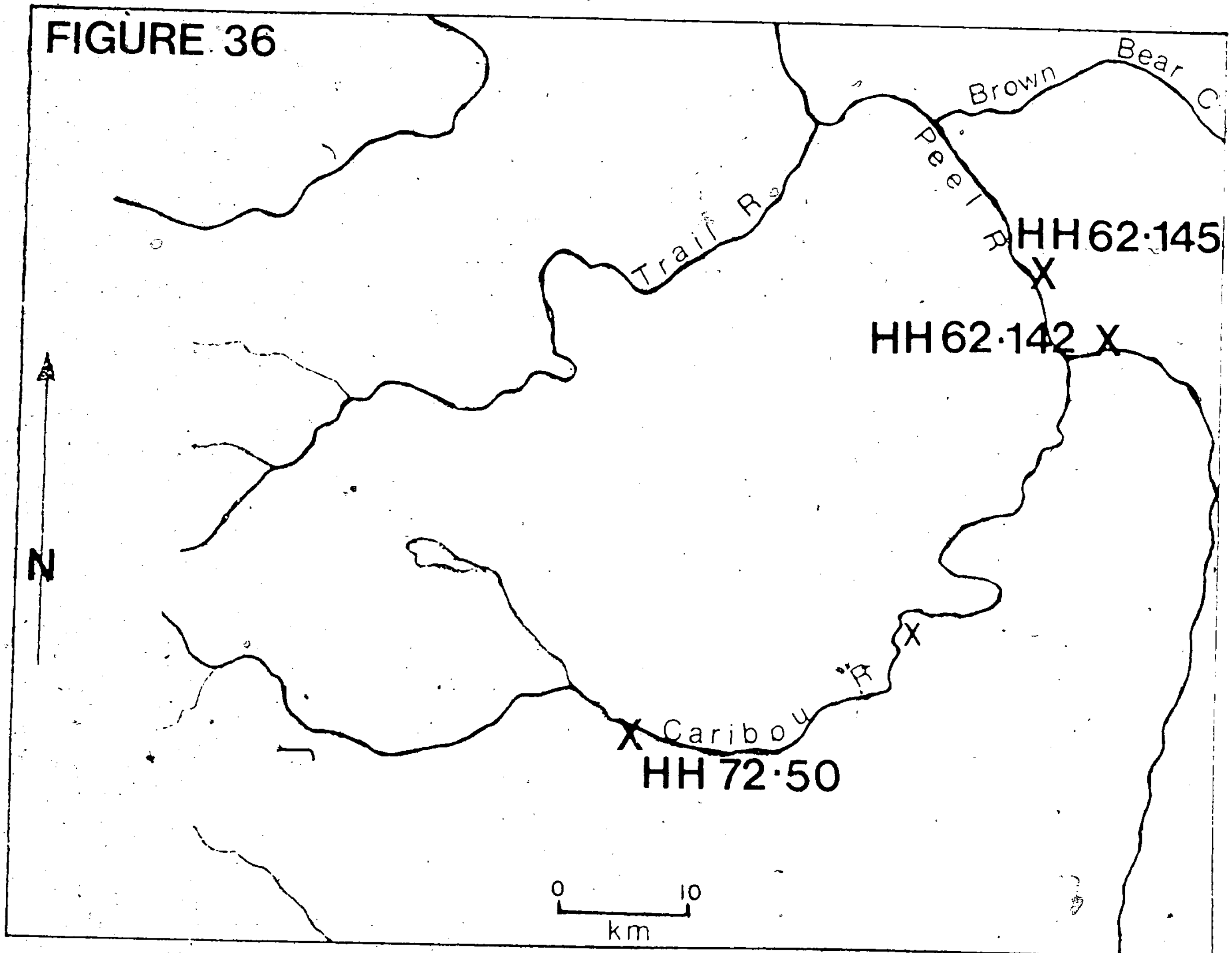

Brown
Bear C
Peel R
Trail R
HH 62·145
HH 62·142
Caribou R
HH 72·50
N
0
10
km

FIGURE 36

Figure 37

Stratigraphic Position of Younger Fluvial Sediments, Caribou River--Brown Bear Creek Area

The principal sections where the Younger Fluvial Sediments are exposed are illustrated.

HF---Holocene Fluvial sands and silts

HG---Holocene Fluvial gravels

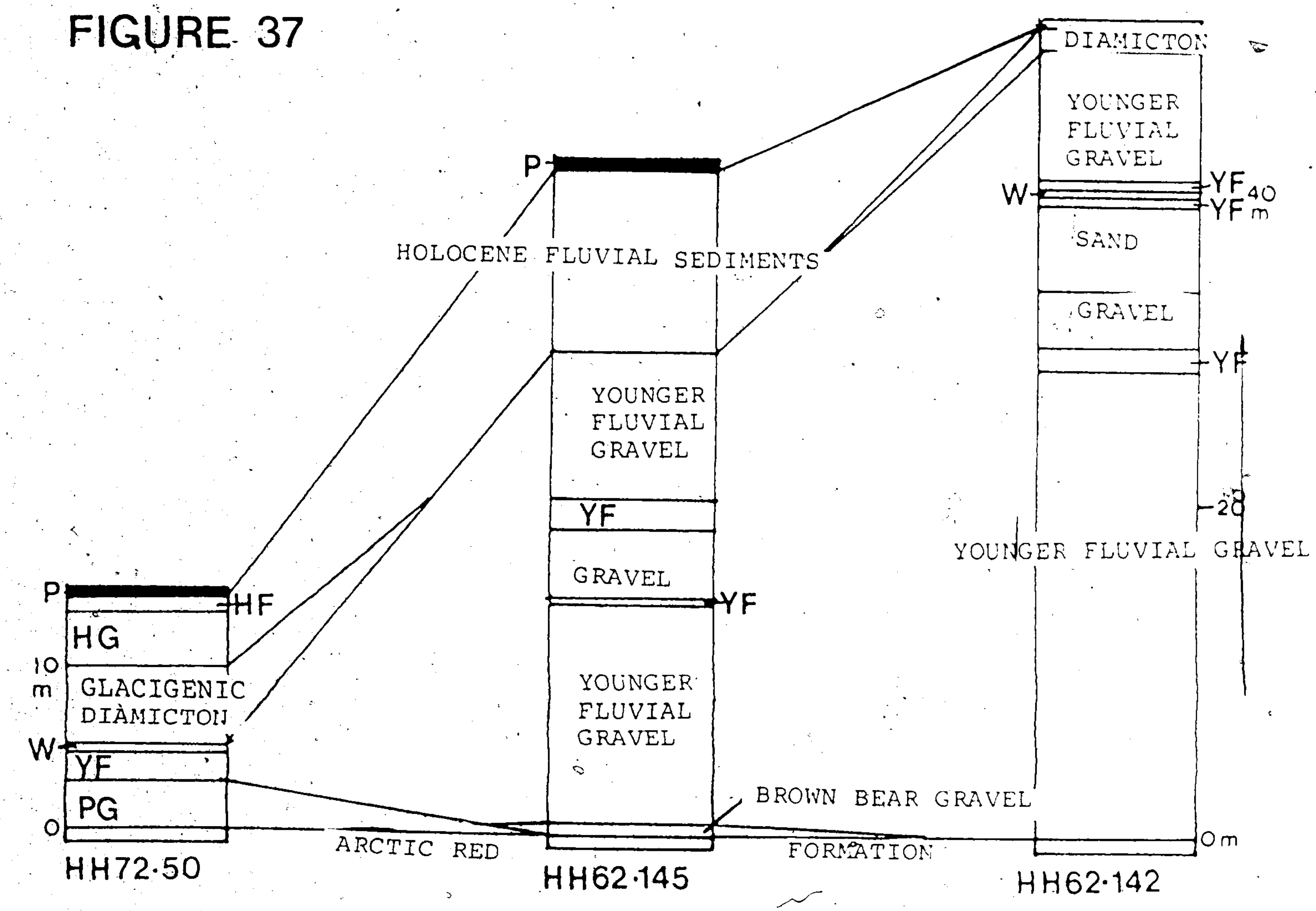
P---Holocene Peat

W---Younger Fluvial Sediments with Palynomorph Assemblages

YF---Younger Fluvial Sediments

Datum: top of Arctic Red Formation

FIGURE 37



P
HF
HG
10
m
GLACIGENIC
DIAMICTON
W
YF
PG
0
HH72·50
HOLOCENE FLUVIAL SEDIMENTS
YOUNGER
FLUVIAL
GRAVEL
YF
GRAVEL
YF
YOUNGER
FLUVIAL
GRAVEL
BROWN BEAR GRAVEL
ARCTIC RED
FORMATION
HH62·145
DIAMICTON
YOUNGER
FLUVIAL
GRAVEL
YF
40
m
W
YF
SAND
GRAVEL
YF
20
YOUNGER FLUVIAL GRAVEL
0 m
HH62·142

primarily from the Caribou and Peel River drainage basins, although granitic and medium- and high-grade metamorphic rock clasts are present (Table 12). Bands of crushed and folded shale are present at the bases of some exposures along the Peel River (Figure 37), suggesting glacio-tectonic deformation.

The sediments were deposited in braided-meandering rivers similar to the modern Peel River and Caribou River. The sequences represent primary channels, lateral and longitudinal bars, back-bar channels, bar-chute channels, and scour pools. Similar depositional sequences have been recognized and described in the modern fluvial environments (Appendix 4), and are also represented in the younger fluvial and early fluvial sediments present along the Snake River (as described previously). A model of the environment of deposition has been previously presented as Figure 26 (see also Figure A4-11). The sediments preserved in the exposures along the Peel River are dominated by primary channel gravels, lateral bar sands and gravels, and back-bar channels sequences (Figure 37), whereas the succession at HH 72-50 (82) consists of back-bar channel, bar-chute channel, and slough deposits (Figure 34).

Two exposures of the younger fluvial sediments were analysed to determine the palynomorph contents. A silty clay bed (Unit #5) present at HH 72-50 (82) (Figures 34 and 36) yielded five samples with sufficient palynomorphs for analysis. The sediment was deposited in an abandoned back

Table 12

Younger fluvial sediments, Caribou-Brown Bear area, clast lithology and mineralogy.

A) Pebbles (33 analyses)

Lithology	Maximum	Mean	Minimum
	%	%	%
orthoquartzite	68	46	26
feldspathic sandstone	32	18	5
siltstone/shale	16	4	0
chert	21	8	0
argillite	21	10	1
jasper-hematite	6	2	0
limestone	11	<1	0
dolomite	21	1	0
granite	27	7	2
diabase	3	1	<1
granodiorite	8	<1	<1
gneiss	7	<1	<1
diorite	7	<1	0
andesite	1	<1	0
basalt	<1	<1	0
gabbro	<1	<1	0
olivine porphyry	<1	<1	0
pyroxenite	<1	<1	0

Table 12 (continued)

B) Granules (36 analyses)

Lithology	Maximum	Mean	Minimum
	%	%	%
orthoquartzite	72	48	31
feldspathic sandstone	31	17	5
siltstone/shale	16	4	0
chert	27	10	0
argillite	26	13	6
jasper-hematite	8	3	0
limestone	6	<1	0
dolomite	5	<1	0
granite	20	5	3
granodiorite	6	<1	<1
diabase	3	<1	0
gneiss	3	<1	0
diorite	2	<1	0
andesite	<1	<1	0
basalt	<1	<1	0
gabbro	<1	<1	0

Table 12 (continued)

C) Coarse Sand (0.42 - 1.00mm), major minerals (62 analyses)

Lithology	Maximum	Mean	Minimum
	%	%	%
quartz	75	53	33
feldspar	45	22	10
argillite/siltstone	45	8	0
chert	17	8	2
calcite	8	2	<1
hematite	2	<1	0
dolomite	3	<1	0
hornblende	5	1	<1

Table 12 (continued)

D) Coarse sand, trace minerals.

Mineral	Number of analyses containing mineral
Actinolite	16
Allanite	8
Anatase	5
Andalusite	4
Ankerite	29
Apatite	48
Biotite	60
Chlorite	54
Chromite	36
Cordierite	21
Corundum	37
Epidote	55
Fluorite	53
Garnet	36
Gypsum	60
Ilmenite	31
Kyanite	22
Leucoxene	16
Magnesite	15
Magnetite	59
Muscovite	50
Pyrite	26
Pyrolusite	9

Table 12 (continued)

Mineral	Number of analyses containing mineral
Pyroxene	44
Rhodochrosite	37
Rhodonite	19
Rutile	40
Siderite	46
Sillimanite	11
Sphene	49
Tourmaline	39
Wollastonite	57
Xenotime	3
Zircon	35

channel or slough (Plate 16).

Palynomorph assemblages for the five samples from Unit #5 are treated as a single pollen zone, A (Figure 38, in pocket). The dominant arboreal taxon is Picea, which increases from 5.2% at the base to a maximum value of 46.7%, and then declines sharply to 16.2% at the top of the zone. Pinus and Larix are present in minor amounts. The arboreal-shrub component is dominated by Betula, with associated Alnus type crispa and minor Alnus type incana and Corylus. These taxa show a distribution pattern similar to that of Picea. Shrubs form a relatively minor proportion of the pollen assemblages, with Ledum, Vaccinium, and Shepherdia canadensis being the chief taxa represented. Herbs, especially Gramineae and Cyperaceae, are an important constituent. The distribution of these grains is the inverse of that of the arboreal and arboreal-shrub palynomorphs. Other herb pollen present includes Chenopodium, Cruciferae, and Rubus chamaemorus. Pteridophyta also vary inversely in proportion with arboreal taxa. The two dominant genera, Cystopteris and Lycopodium, differ in that Lycopodium spores gradually decline in numbers vertically through the unit, whereas Cystoperis reaches its maximum in the upper portion of the zone. The unit is devoid of aquatic pollen, and Sphagnum is the only bryophyte represented.

Zone A can be interpreted to represent the progressive evolution of a vegetation assemblage developed on low, gravelly floodplain terraces with some input from the

Plate 16

Younger Fluvial Sediment, BH 72-50, Palynomorph-Bearing Unit.

The sediment is silty clay of unit 5, deposited in a slough associated with a braided-meandering stream. Palynomorphs obtained from these sediments indicate a climate warmer than that currently prevalent in the Peel Plateau. See Figures 34 and 38.


minolta

surrounding upland plateau community. It represents a meld of communities similar to the Mixed Deciduous/Evergreen Forest and Microphyllous Evergreen Dwarf Scrub which commonly occur currently in the Peel Plateau region (Hettinger et al 1973).

The Picea component was initially composed entirely of P. type mariana, but the amount of P. type glauca grains gradually increases with elevation. This indicates that the floodplain had stablized and that the banks were dry enough to enable P. glauca to begin to establish itself. The proportions of Picea grains are consistent with the gradual establishment of a mixed forest in the area. Picea does not tend to be over-represented in the pollen spectrum, as the effects of its relatively efficient aerial distribution are conterbalanced by the tendency of the genus, especially P. mariana, to reproduce vegetatively. The absence of Larix from the spectrum is probably due to its easily degraded pollen, rather than its absence from the environment.

The deciduous portion of the arboreal-shrub assemblage is dominated by shrub Betula. In this instance Betula appears to be over-represented in the spectrum when compared to the other deciduous shrub taxa. Alnus would be expected to dominate the arboral-shrub component of a floodplain environment, but Alnus pollen is present only as subordinate to Betula. The absence of two other arboral-shrub taxa, Populus and Salix, is a noteworthy feature of the assemblages. Populus pollen is extremely susceptible to crumpling and degradation, and the taxon is capable of vegetative reproduction. The complete absence of Salix grains suggests that the flooding frequency of the area was low, providing other taxa with the opportunity to establish themselves along the river banks. Phenological differences are sufficiently minor in the modern environment (Ritchie 1977) to indicate that segregation of the grains due to differential production schedules could not account for the distribution observed.

Corylus, Pinus, and Cornus canadensis all require warmer climatic conditions than those currently existing at HH 72-50 (82). The presence of small amounts of Pinus and Cornus pollen can be attributed to either minor climatic amelioration, or to sporatic long distance aeolian transport. Corylus, however, is not currently extant in the Yukon or Northwest Territories (Porsild and Cody 1980), so the degree of climatic amelioration required for it to extend its range to the Caribou River is considerable. The small number of Corylus grains observed suggests that the taxon was not extant in the region at the time of deposition, although it may have been growing to the north of its present maximum range. Lichti-Federovitch (1974) regarded Corylus grains in the lower sediments at Old Crow as indicative of the occupation of the area by the southern boreal forest, while Schweger (1982) considered grains found in alluvial sediments in Alaska to be indicative of long-range transport. No Corylus macrofossils have been recovered from north-east Beringia to the author's knowledge.

The understory vegetation record reflects the gradual establishment of a relatively stable community on infrequently flooded coarse-textured sediments, with Shepherdia canadensis, Cerastium, Empetrum, Sagina, Arnica and Lycopodium. These genera, together with Ledum and Vaccinium, are all found in the understory of the Mixed Deciduous/Evergreen Forest. Open, dry portions of the bars and the surrounding sandy and gravelly slopes were colonized by Artemisia, Shepherdia, and Equisetum.

The wetter back-channel margins were dominated by moisture-preferring taxa such as Cyperaceae, Myrica Gale, and Sphagnum. These taxa are most probably over-represented in the spectra, because of their proximity to the depositional site, while some valley slope and snowbank taxa (eg. Cassiope) are probably under-represented. The under-representation of the upland and valley slope communities may reflect the absence of large-scale sheetwash events capable of carrying the pollen to the fluvial system.

The gradual development of the assemblage is reflected in the increases in Picea pollen and the decline in herbaceous palynomorph content. This indicates that Picea was establishing itself on the floodplain and bars to the detriment of the herbs and ferns. However, in the uppermost portion of the zone, high herb concentrations and low arboreal and arboreal-shrub values suggest that the area's vegetation had become rejuvenated to a large degree. Simultaneously, the continued decrease in Cyperaceae, the drop in the Sphagnum fraction, and the absence of Myrica Gale pollen indicate that water-table changes or tundra development are unlikely agents of the changes.

Fire is the most probable explanation of the changes. Tundra and taiga fires effectively rejuvenate the vegetation by destroying the Picea and Betula forests, allowing the understory vegetation to dominate. The vegetation which initially recolonizes the burned area varies, but it is often heavily dominated by lichens and bryophytes (usually under-represented in palynological spectra), Betula glandulosa, Salix, and components of the pre-fire understory. Fire occurence in tundra areas is characterised by frequent small events with occasional large fires (Wein 1974; Hall et al 1978; Johnson 1979). Consequently, fire events probably cannot be correlated over large distances in the tundra.

In summary, Unit #5 presents a picture of gradual evergreen forest expansion over stable floodplain sediments, interrupted by a single fire event, as recorded in the spectra of zone A. The climate indicated is similar to or slightly milder than that prevailing on the Peel Plateau today (Ritchie 1984).

At location HH 62-142 (81), a 6 cm thick stratum of structureless silty clay 41.8m above the base of the exposure (Figure 37) contained sufficient palynomorphs for analysis. The clay was deposited in a back-bar channel. The palynological assemblage (Table 13) is dominated by *Picea* (61.8%), chiefly *glauca* type, with associated Cyperaceae. No other taxon exceeds 2.4% of the total. Minor amounts of *Betula*, *Vaccinium*, *Ledum*, *Artemisia*, Cruciferae, *Epilobium*, *Myrica Gale*, *Sphagnum*, and *Lycopodium* are present. This unit grades vertically into a second silty clay (31% silt, 69% clay) 5 cm thick (Figure 37). The spectrum of this stratum is dominated by *Betula* (45%), with associated Cyperaceae (21.6%) and *Picea* (8.7%). Gramineae, *Rubus chamaemorus*, *Ledum*, *Artemisia*, *Sagina*, *Polemonium*, Cruciferae, and *Chenopodium* are present in lower amounts.

These two assemblages illustrate the difficulties involved in interpreting a fluvial back-channel palynological succession. The two sedimentary units appear penecontemporaneous, but contain substantially different assemblages. These differences can be attributed in part to the effects of fluvial transport (cf. Catto 1985). The lower stratum contains 30% coarse silt and sand, and is relatively enriched in the large *Picea* grains compared to the upper unit. The finer sediment is enriched in the smaller grains, such as *Betula*, and the more delicate, thin-walled palynomorphs (eg. Gramineae, *Larix*, and *Populus*). Robust grains (*Lycopodium*) are concentrated in the coarser unit.

The great variation in *Picea* and *Betula* concentrations, howver, indicates that hydrodynamic factors alone are incapable of explaining the differences. It seems probable

Table 13

Palynological counts, younger fluvial sediments, section HH 62-142 (81). All taxa included in pollen sum

Taxon	Lower clay	Upper clay
	%	%
Picea	61.8	8.7
Pinus	0.4	0.0
Abies	0.4	0.0
Betula	1.7	45.0
Salix	0.0	1.0
Populus	0.0	0.5
Larix	0.0	0.5
Ledum	0.8	1.1
Vaccinium	1.0	0.5
Arctostaphylous rubra	0.4	0.0
Myrica Gale	0.8	0.5
Cyperaceae	26.0	21.6
Gramineae	0.4	6.3
Artemisia	0.8	2.6
Arnica	0.2	0.5
Sagina	0.0	1.6
Cruciferae	0.8	1.1
Saxifraga	0.0	1.7
Chenopodium	0.4	1.1
Linnaea borealis	0.0	0.5
Polemonium	0.0	1.1

Taxon	Lower clay	Upper clay
	%	%
Epilobium	0.2	0.6
Ribes	0.0	0.5
Rubus chamaemorus	0.0	0.7
Rosa	0.0	0.5
Lycopodium	2.4	0.0
Cystopteris	0.4	1.7
Equisetum	0.0	0.5
Sphagnum	1.2	0.0
Myriophyllum	0.0	0.5

that the two assemblages do represent distinct communities. The lower spectrum may reflect a regional community, with the palynomorphs having been transported to the site by wind and spring flooding. The upper spectrum is interpreted to represent a local community developed on a channel margin or stabilized bar surface, dominated by Betula shrubs with dry-ground taxa such as Sagina, Arnica and Epilobium. The presence of the characteristically under-represented taxa Salix and Populus also suggests an assemblage derived from a local bar surface or channel margin environment.

The two assemblages considered jointly are indicative of an environment very similar to that presently influencing the Peel Plateau (Ritchie 1984). The combined spectra suggest that the stabilized bar surfaces and margins were occupied by Betula-Salix shrub communities, while the upland areas along the river valley were bordered by fringes of Picea glauca open woodlands. The major difference between this assemblage and current Peel River vegetation is the absence of Alnus from the fossil spectra. Since the climatic conditions indicated by the assemblages were suitable for the growth of Alnus, its absence implies that it was not extant in the region at the time. The single grains of diploxylon Pinus and Abies (type balsamea) are assumed to be distally-derived.

The sediments are overlain by 4 gravel and sand strata totalling 9.05 m. These units in turn are overlain by 1.1 m of silty clay matrix till (Figure 37).

No materials suitable for dating were recovered from any of the younger fluvial sediment exposures along the Peel River. A sample from the silty clay bed at HH 72-50 (82) (unit #5) proved to have insufficient organic matter for ^{14}C dating (W. Blake Jr, Geological Survey of Canada, personal communication, 1985). Therefore, no quantitative age assessment for the younger fluvial sediments is possible.

E. Younger Glacigenic Sediments and Landforms

Till

Glacigenic diamicton caps the exposures present along the Peel River between Chiltree and George Creeks (Figures 39 and 40). Till with similar clast lithological and mineralogical composition is exposed along the Caribou River (Figure 40, Table 14). Similar diamicton sediment is present over the surface of the Peel Plateau and Peel Plain in the vicinity of the Caribou River, the Peel River, and Brown Bear Creek. The similarities in composition, and in the stratigraphic position of the till exposures, suggests that the sediments represent a single glacial advance.

An outcrop representative of the till exposed along the Caribou River, and along the east bank of the Peel River between Chiltree and George Creeks, is present at HH 72-50 (82) (66° 15' N, 134° 57' W) (Figure 34). Here, 4.7 m of structureless silty till stands in moderately steep exposures. The till unconformably overlies organic silt, and

Figure 39

Glacigenic Diamicton Exposures, Cariboo River - Brown Bear Creek Area

X-- sections containing Glacigenic Diamicton investigated in detail

x---other exposures of Glacigenic Diamicton

FIGURE 39

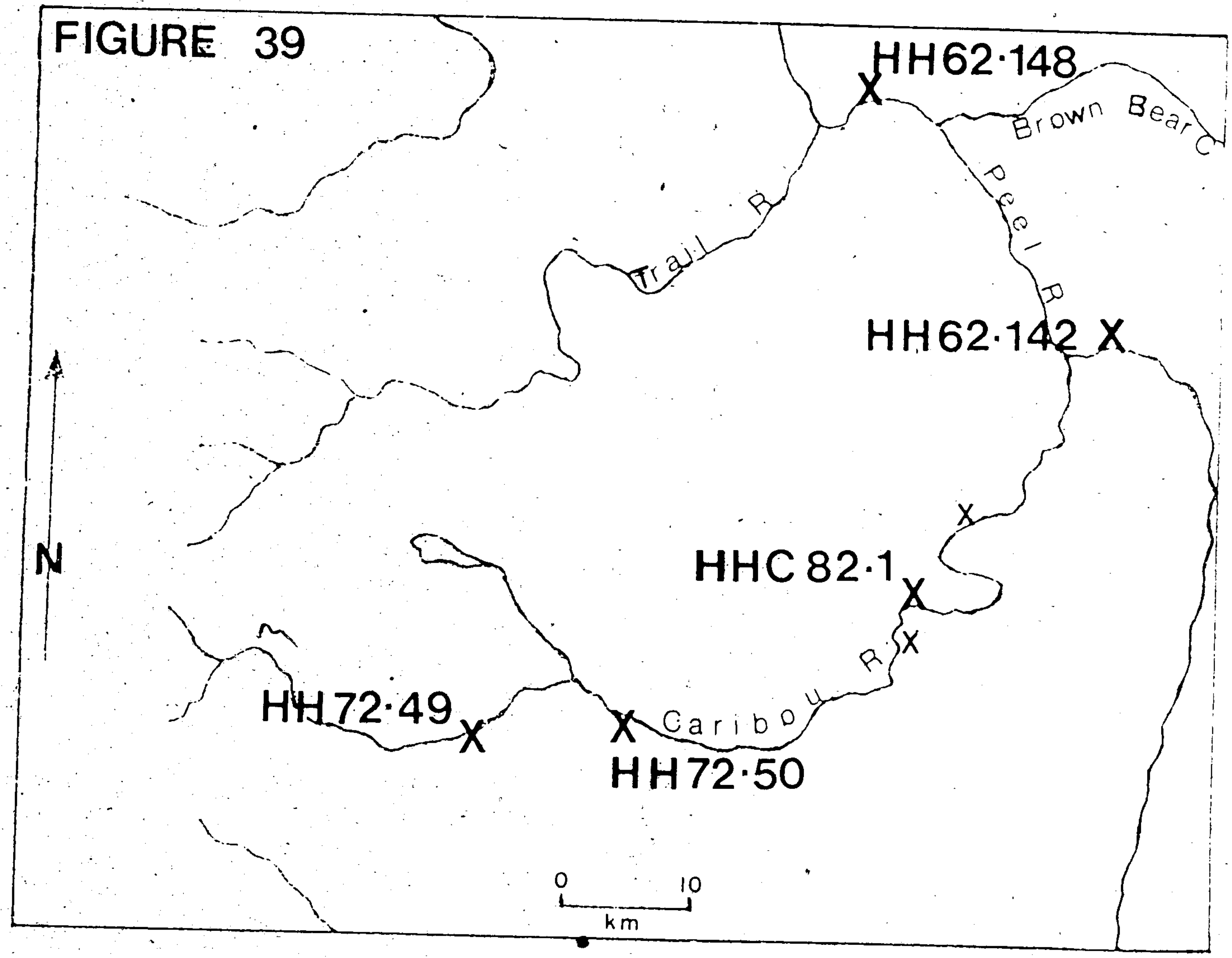

HH62·148
Brown Bear C
Peel R
Trail R
HH62·142
HHC82·1
Caribou R
HH72·49
HH72·50
N
0
10
km

Figure 40

Stratigraphic Position of Glacigenic Diamicton, Caribou River--Brown Bear Area

The principal sections where the glacigenic diamicton is exposed are illustrated.

HF--Holocene Fluvial sands and silts

HG---Holocene Fluvial gravels

P---Holocene Peat

W---Younger Fluvial Sediments with Palynomorph Assemblages

YF---Younger Fluvial Sediments

Datum: top of Arctic Red Formation

FIGURE 40

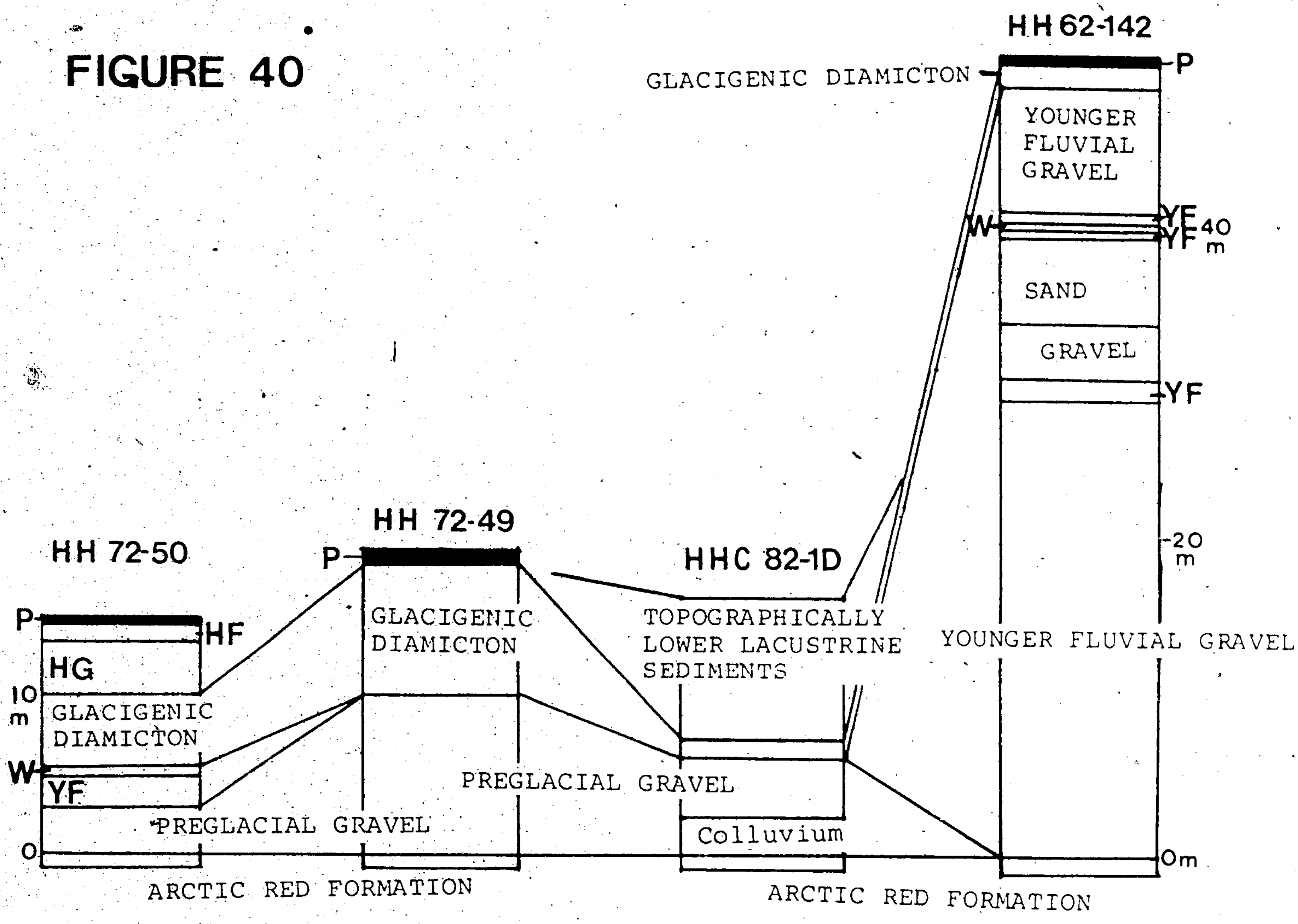

HH 72-50
P
HF
HG
10 m
GLACIGENIC DIAMICTON
W
YF
PREGLACIAL GRAVEL
0
ARCTIC RED FORMATION
HH 72-49
P
GLACIGENIC DIAMICTON
PREGLACIAL GRAVEL
HHC 82-1D
TOPOGRAPHICALLY LOWER LACUSTRINE SEDIMENTS
Colluvium
ARCTIC RED FORMATION
HH 62-142
GLACIGENIC DIAMICTON
P
YOUNGER FLUVIAL GRAVEL
YF
W
YF
40 m
SAND
GRAVEL
YF
20 m
YOUNGER FLUVIAL GRAVEL
0m

Table 14

Till, Caribou River-Brown Bear Creek area, clast lithology and mineralogy.

A) Pebbles (17 analyses)

Lithology	Maximum	Mean	Minimum
	%	%	%
orthoquartzite	56	41	21
feldspathic sandstone	30	14	6
siltstone/shale	33	17	10
chert	8	5	3
argillite	27	16	8
jasper-hematite	1	<1	0
limestone	8	3	1
dolomite	2	<1	0
granite	<1	<1	<1
diabase	3	1	0
granodiorite	<1	<1	0
gneiss	<1	<1	0
phyllite	<1	<1	0
iron carbonate	2	<1	0
schist	<1	<1	0
basalt	<1	<1	0
gabbro	<1	<1	0

Table 14 (continued)

B) Granules (18 analyses)

Lithology	Maximum	Mean	Minimum
	%	%	%
orthoquartzite	54	31	15
feldspathic sandstone	38	21	6
siltstone/shale	61	32	15
chert	6	2	0
argillite	18	8	3
jasper-hematite	1	<1	0
limestone	8	2	0
dolomite	1	<1	0
granite	1	<1	0
granodiorite	<1	<1	0
diabase	<1	<1	0
gneiss	<1	<1	0
iron carbonate	1	<1	0
phyllite	<1	<1	0
andesite	<1	<1	0
basalt	<1	<1	0
gabbro	<1	<1	0
schist	<1	<1	0

Table 14 (continued)

C) Coarse sand (0.42 - 1.00mm), major minerals (18 analyses)

Lithology	Maximum	Mean	Minimum
	%	%	
quartz	55	43	3[illegible]
feldspar	50	42	36
argillite fragments	20	5	0
chert	10	6	3
calcite	3	1	<1
hematite	2	<1	0
dolomite	1	<1	0
hornblende	5	2	1

Table 14 (continued)

b) Coarse sand, trace minerals.

Mineral	Number of analyses containing mineral
Actinolite	12
Andalusite	[illegible]
Ankerite	9
Apatite	13
Biotite	18
Chlorite	17
Chromite	6
Cordierite	3
Corundum	9
Epidote	10
Fluorite	12
Garnet	4
Gypsum	18
Ilmenite	8
Kyanite	7
Lignite	14
Magnesite	3
Magnetite	18
Muscovite	16
Pyrite	10
Pyroxene	17
Rhodochrosite	12

Table 14 (continued)

Mineral	Number of analyses containing mineral
Rhodonite	[illegible]
Rutile	[illegible]
Siderite	[illegible]
Sillimanite	[illegible]
Sphene	8
Tourmaline	[illegible]
Wollastonite	16
Zircon	[illegible]

is overlain along an erosional contact by well-sorted, matrix-supported pebble gravel. Significant differences exist between the lithologies of the pebble and granule fractions, and between the lower and upper parts of the till. The lower pebbles are dominated by orthoquartzite, siltstone, greywacke, and siliceous argillite, whereas the granule fraction is predominantly siltstone, greywacke, and orthoquartzite, with little siliceous argillite (Table 15). Both fractions of the lower portion of the till are devoid of igneous and high-grade metamorphic clasts, suggesting that the provenance of this material is predominantly local. Both fractions lack ultramafic clasts, and both contain clasts of phyllite (or slate). Although the phyllite could originate from the Canadian Shield to the east, phyllite of the Quartet Group is exposed 15 to 20 km west of the section along the Caribou River (Norris 1981d). In the absence of any other clasts of Canadian Shield provenance, the second alternative is considered to be more probable.

No stratigraphic break or lithological contrast between the lower and upper parts of the till were observed in the field at HH 72-50 (82). Nevertheless, there are lithological differences in the clast assemblages. The upper part of the till contains both pebbles and granules of Canadian Shield provenance, including granite, gabbro, granodiorite, diabase, basalt, gneiss and schist, but excluding serpentinite and peridotite. The concentration of Canadian Shield material was estimated as 1 clast per 800 pebbles and

Table 15.

Till, HH 72-50 (82), pebble and granule lithology.

A) Pebbles

Lithology	Lower[1]	Upper[2]
		%
orthoquartzite	32	45
feldspathic sandstone	20	10
siltstone/ shale	23	16
chert	5	6
argillite	16	18
jasper-hematite	0	<1
limestone	5	2
dolomite	<1	<1
granite	0	<1
diabase	0	1
granodiorite	0	<1
phyllite	<1	<1
iron carbonate	0	1
schist	0	0
gabbro	0	<1
basalt	0	<1
gneiss	0	<1

[1]mean for 2 analyses

[2]mean for 3 analyses

Table 15 (continued)

B) Granules

Lithology	Lower[1]	Upper[2]
	%	%
orthoquartzite	25	36
feldspathic sandstone	27	20
siltstone/ shale	40	28
chert	1	2
argillite	7	9
jasper-hematite	0	< 1
limestone	1	4
dolomite	0	< 1
granite	0	< 1
diabase	0	0
granodiorite	0	< 1
phyllite	< 1	< 1
iron carbonate	0	0
schist	0	< 1
gabbro	0	0
basalt	0	< 1
gneiss	0	0

[1]mean for 2 analyses

[2]mean for 4 analyses

granules. The pebble fraction is dominated by orthoquartzite, siliceous argillite, siltstone and greywacke, while the granule fraction is richer in siltstone and greywacke and contains less orthoquartzite and siliceous argillite (Table 15). Similar lithological variations, both among different size fractions and with vertical position were detected in other till exposures along the Caribou River. Lateral variations were also observed.

The clast fabric of the till at HH 72-50 (82) is strongly oriented east-west, with a median of 257-077 and a mode of 265-085. The clast content indicates that the regional direction of glacial flow was westerly, a conclusion confirmed by other fabric analyses and by regional geomorphology (Figures 32 and 39). The till is apparently devoid of stratification, shearing and deformation structures, sand lenses, and dewatering features.

The strongly oriented fabric of this till, and the conformance of the fabric and other ice-flow indicators, indicate that the sediment was deposited directly from ice, without significant flowage or slumping. Unimodal fabric distributions are characteristic of both basally- deposited melt-out and lodgment till (Harrison 1957; Lawson 1979). The till at HH 72-50 (82) is therefore considered to be basal.

Similar till exposed at sections HHC 82-1d, 1e, 1f, and 1g (66° 22' N, 134° 20' W)(Figure 39), along the north bank of the Caribou River has a strongly-developed northwest-trending (310-320) fabric. This pattern, with minor or no southwesterly component, could represent: a) a transversely-oriented distribution produced by subglacial shearing; b) local flow variations within a single glacier; or c) a distinct local advance directed to the northwest.

Transverse fabric orientations have been experimentally produced from sheared dispersions (Rees, 1983), and are commonly found in association with shear plane traces and drag folds in till. The absence of these features at HHC 82-1d, 1e, 1f and 1g does not preclude shearing. Microfabric determinations could lead to the recognition of shear plane surfaces too small and discontinuous to be observed in outcrop (cf. Harrison 1957).

Local flow variations within the glacier could also produce macrofabrics at variance with the regional orientation (Harrison 1957; Young 1969; Andrews and Smith 1970). The shift in direction could be produced either by glaciological conditions within the ice sheet (such as lateral temperature gradients) or by topographical obstructions. A minor topographic feature may have a large influence on glacier flow especially during the early stages of an advance (when the ice is thinner), resulting in the production of fabric which bears no relation to the regional flow direction subsequently imposed on the area.

The possibility that the northwest-trending fabric was produced by a local readvance cannot be conclusively eliminated at present. Geomorphic, mineralogic, or lithologic evidence to support this interpretation does not exist however. The production of fabric transverse to regional flow is therefore attributed to either shearing within the ice mass, or local variation in flow, or both.

The till exposed along the Caribou River at both HHC 82-1 and HH 72-50 (82) is considered to be produced as a result of basal melt-out. The clast shape distribution is consistent with that expected from either a basal melt-out or lodgment till (Drake 1972; Haldorsen 1981; Lister 1981; Sharp 1982). The distinct lithological stratification of the coarser fraction, with the lower portion enriched in local material, is also characteristic of undisturbed basal deposition from an internally stratified glacier. The textural composition of the till is similar to that reported from melt-out tills derived from predominantly siltstone and silty sandstone bedrock and sediments (eg. Shaw 1982; Catto 1984). The absence of sand lenses at HH 72-50 (82), a sedimentary feature commonly found in melt-out tills, (Harrison 1957; Kruger 1979; Haldorsen and Shaw 1982) may be an artifact of the limited lateral exposure of the section. Alternatively, basal conditions during glacial activity at this location may have precluded the devlopment of subglacial channels and cavities necessary for the production of sand lenses. A wide range of basal conditions commonly exists beneath modern glaciers, some of which are not conducive to the formation of stable subglacial channels and cavities (Collins 1979; Hodge 1979; Walder and Hallet

1979; Vivian 1980). Consequently, the distribution of sand lenses would not be expected to be uniform throughout a till deposits or within a drainage basin. Other exposures of till along the Caribou River do contain sand lenses and laminae.

The similarities in stratigraphic position; the progressive changes in clast lithology reflecting the local bedrock; the heavy mineral assemblages; and the relationship between the till exposures and the glacially-produced geomorphic features, indicate that the till exposures of the Caribou River area and the portion of the east bank of the Peel River between Chi Itree and George Creeks were produced by a single glacial event. The exposures at HHC 82-1 represent a possible exception to this conclusion, based primarily on the differences in clast fabric. The absence of any stratigraphically younger glacigenic sediments indicates that this event was the last glaciation to affect the region.

Streamlined Features

Major fields of streamlined drumlinoids and drumlins are located east of the Peel River in the vicinity of Brown Bear Creek and the Satah River (Figure 32). Isolated drumlinoids are present in the vicinity of Nihtal Git Creek, and southeast of the Road River. The streamlined forms indicate that ice flow was to the west and southwest throughout the region. No cross-cutting forms were observed, suggesting that the features were produced during a single glacial event.

Glaciofluvial Features

Small kames are present throughout the Peel Plateau (Figure 32). A single esker is present in the vicinity of the Satah River. The glaciofluvial features are minor components of the geomorphology of the Peel Plateau and Peel Plain, and no investigation of these sediments was conducted.

F. Glacigenic Ridge and Valley Complexes

Description

Several series of ridge and valley complexes approximately parallel to the Peel River are present east of the stream (Figure 41). South of the confluence of the Peel and Caribou Rivers, the ridges and valleys are aligned north-northeast to south-southwest. In the vicinity of the Caribou River confluence, the complex is oriented as a northeastward-convex arcuate belt, with the southern end trending north-northwest to south-southeast and the northern part trending west-northwest to east-southeast. A similarly-oriented arcuate belt is present opposite the confluence of the Trail and Peel Rivers (Plate 17). North of the confluence, the ridges and valleys parallel the eastern limit of the Peel valley north to a position opposite Shiltee Rock. The width of the belt of ridge-and-valley

Figure 41

Location of Glacigenic Ridge-and-Valley Complexes

● Areas of Ridge-and-Valley Complexes

FIGURE 41

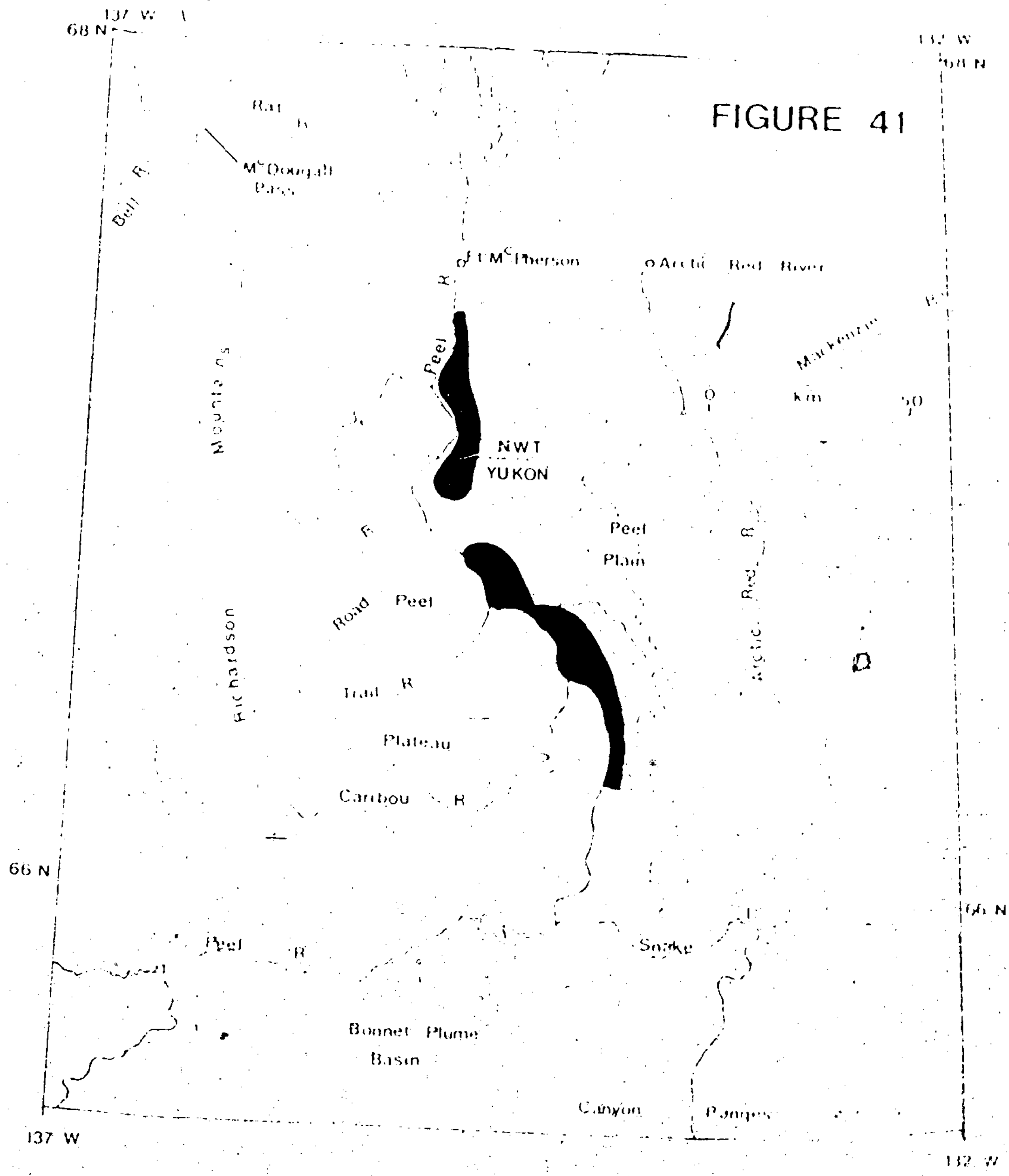
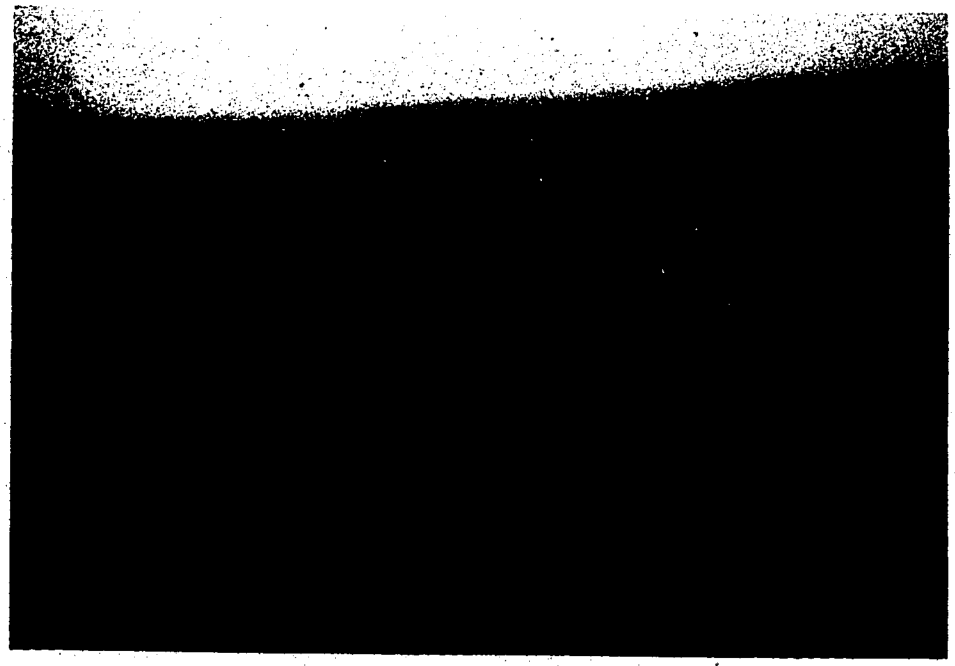

137 W
68 N
132 W
68 N
Rat R
McDougall Pass
Bell R
Ft McPherson
Arctic Red River
Mackenzie R
Peel R
0
Km
50
NWT
YUKON
Richardson Mountains
Peel Plain
Arctic Red R
Road
Peel
Trail R
Plateau
Caribou R
66 N
66 N
Peel R
Snake
Bonnet Plume Basin
Canyon
Ranges
137 W
132 W

Plate 17 (top)

Aerial View of Glaciogenic Ridge and Valley Complex, vicinity of HH 62-145. The photograph was taken looking south, and the Peel River is visible along the western side of the complex. Note that the valleys have been modified by meltwater channel activity. HH 62-145 is located at the top of the westernmost exposed face.

Plate 18 (bottom)

Deformation adjacent to a Granodiorite Boulder, HH 62-148. The deformation was produced by thrusting due to southwestward-flowing ice. See Figure 47 for further details.



complexes reaches a maximum of 12 km. In the southern part of its extent, the belt's eastern margin is indistinct. North of 66° 35' N latitude, however, the eastern margin is marked by an escarpment with a maximum height of 90 m.

A profile through the ridge and valley complexes is presented in Figure 42. The floors of the valleys are generally 120-200 m above the level of the Peel River, although several deeply incised channels are present (such as that occupied by Cooking Rocks Creek). Channel widths range between 50 m and 500 m. The valleys consist of segments 10 km to 15 km in length, rather than being lengthy, uniform depressions.

Streamlined drumlinoid features are developed on some valley floors. The drumlinoids are 20 m to 120 m in length, 10 m to 40 m wide, and commonly less than 10 m high. Orientation of the features indicates ice movement towards the west and southwest, approximately normal to the ridge and valley complexes. The drumlinoids examined were composed of till, but many of the features remain uninvestigated.

The ridges stand 10 m to 70 m above the valley floors. Ridge segments are commonly 300 m to 600 m broad and 2 km to 6 km long. Although some ridges are symmetrical in cross-section (Figure 42), most display a pronounced asymmetry, with the western slopes steeper (mean slope angle 17.1°) than those on the eastern sides (mean slope angle 6.4°. The majority of the ridge crests are flat, and no drumlinoid features are present. Table 16 summarizes the

Figure 42

Cross-Sections Through Glaciogenic Ridges, Brown Bear Creek Area

Letters refer to ridges described in Table 16.

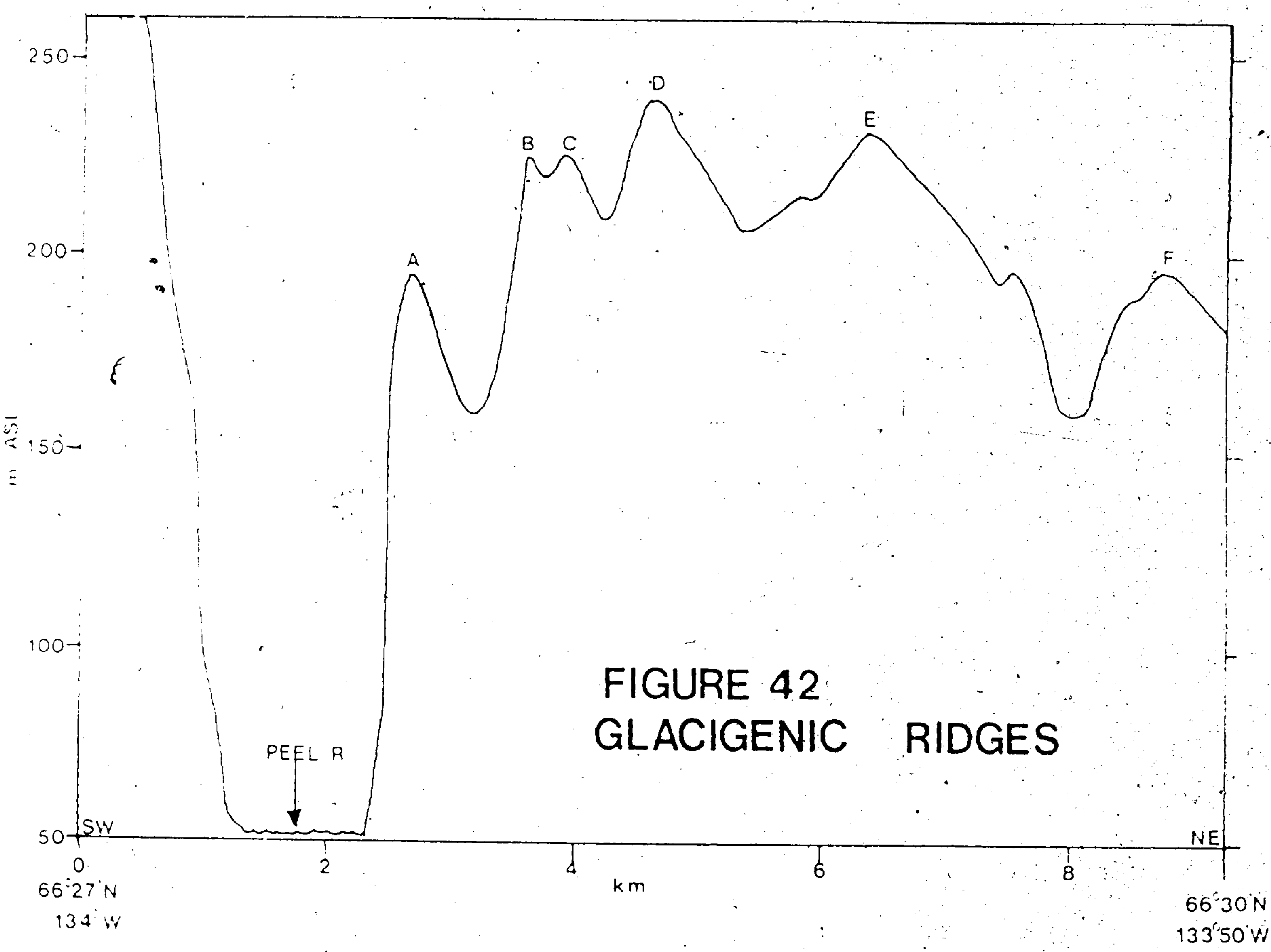


FIGURE 42
GLACIGENIC RIDGES

geomorphic properties of several ridge segments.

The bedrock throughout the area is Cretaceous Arctic Red Formation shale (Norris 1981e). No pronounced textural fluctuations were noted in outcrops of this unit examined along the Peel River. The shale dips gently to the west, and no major structural features or discontinuties are present in the vicinity of the Peel River.

The internal structures in the ridges show that extensive deformation of the sediments has occurred. Contacts between the Arctic Red Formation shale and overlying sediments commonly show a sawtooth surface, with northeast-trending slopes plunging at 15° to 25° and southwest slopes at 40° to 55°. Interstratification of crushed shale and sediment layers is characteristic of the northeast-trending inclines. At some locations, thin (3-5 cm) layers of crushed shale striking northwest-southeast, inclined at angles of 15° to 25° towards the northeast, and connected at the northeastern ends to the bedrock are present in the overlying sediments. These crushed shale bands are confined to the basal 5 m of sediment, and are locally overlain by 40 m to 50 m of undisturbed sediment.

A sediment complex exposed in a ridge 5 km south of the confluence of the Trail and Peel River (locations HH 62-148-81, 1 and 2, 66° 40' N, 134° 32' W, illustrates several types of deformation (Figure 43). The basal unit is the Cretaceous Arctic Red Formation shale, badly weathered and friable where exposed. Within the upper 15 m of the

Table 16

Morphology of glacigenic ridges, Brown Bear Creek area.

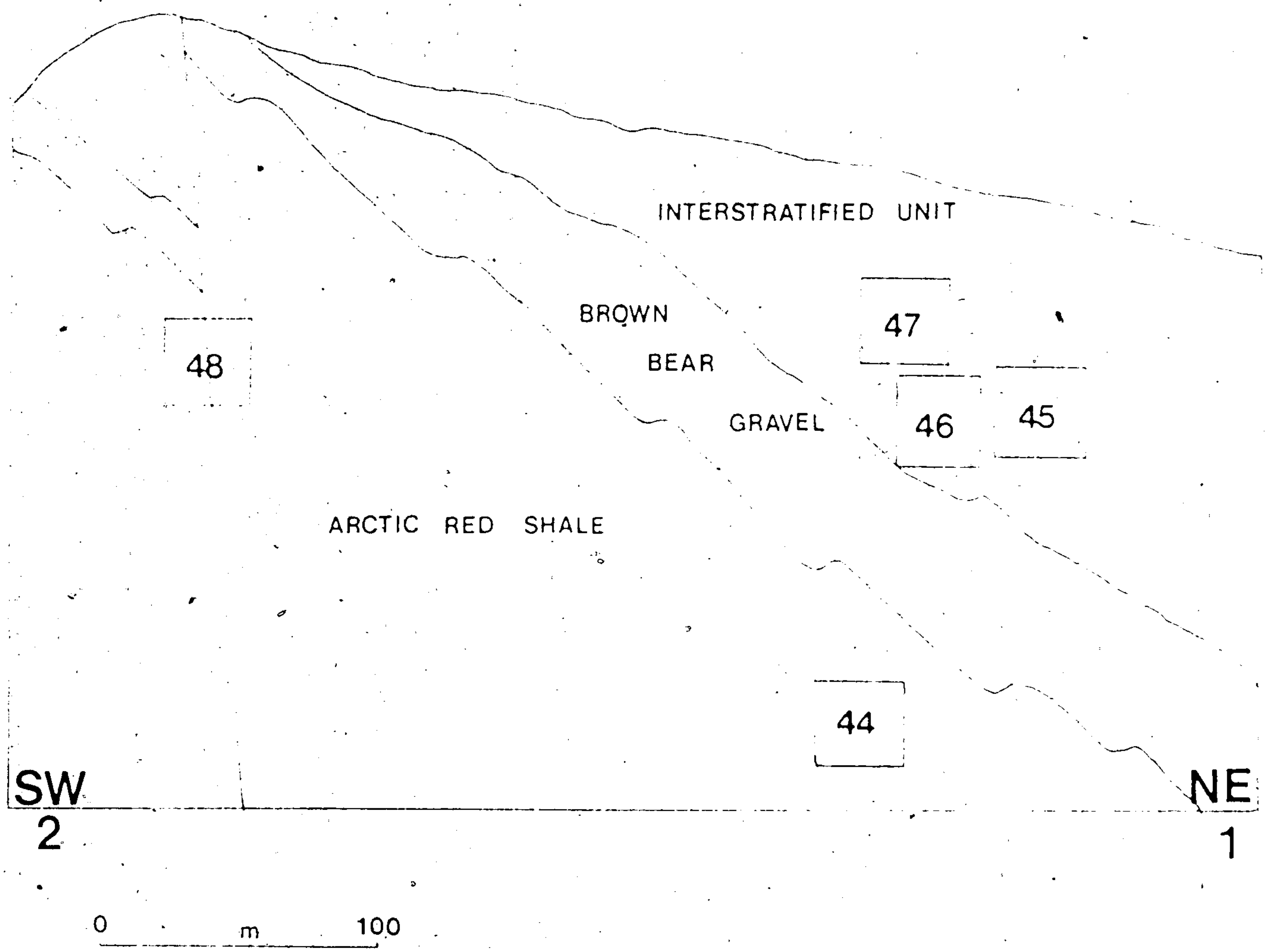
Ridge	Angle of west slope	Angle of east slope	Ratio	Azimuth trend
A	0.508	0.127	4.00	331
B	0.305	0.101	3.02	319
C	0.077	0.038	2.02	324
D	0.133	0.076	1.75	326
E	0.041	0.024	1.71	330
F	0.071	0.017	4.18	325
mean for 25 ridges	0.308	0.112	2.75	325

Figure [illegible]

Glaciogenic Ridge, HH 62-148 81-1 and 3

A cross-section through a glaciogenic ridge is illustrated. Details of the internal structure are illustrated in Figures 44, 45, 46, 47, and 48, the locations of which are indicated.

FIGURE 43
GLACIGENIC RIDGE, HH.62·148 81·1 & 2

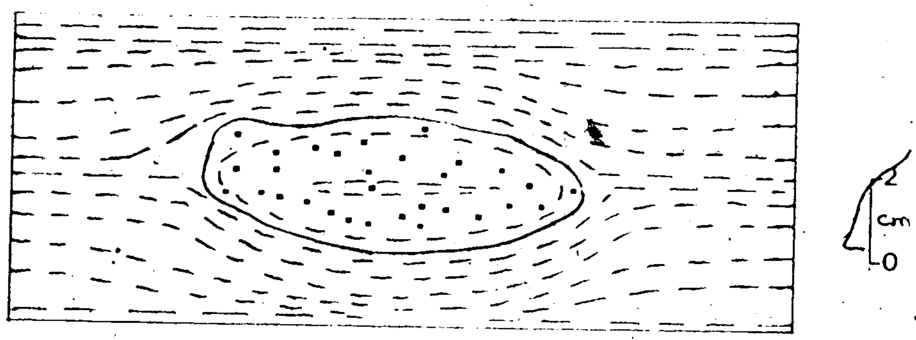


shale are several ovoid masses of brownish-yellow (10 YR 6/8 m/f) fine to medium sand, measuring (modal) 10 cm x 4 cm x cm (Figure 44). The lenses are aligned parallel to the shale bedding planes, with the long axes oriented northeast-southwest. Some of the lenses are massive, but most show a concentric structure when moist. No textural contrasts within the lenses were apparent. The lenses contain basic and acidic igneous and metamorphic rock fragments. Shale strata are deformed adjacent to the lenses, and crushed shale particles are present within the sand bodies.

Overlying the shale along a sawtooth contact is a bed of brownish-yellow (10 YR 6/6 m/f) "Brown Bear" gravel 30 cm to 80 cm thick. Pebbles contained within the gravel are aligned north-northeast to south-southwest, and are imbricated 5° to 10° towards the north-northeast. Within the gravel are spherical to ovoid lenses 10 cm in diameter (modal) of medium-grained sand, similar to those present within the Arctic Red Formation shale. Many of these sand lenses contain minute fragments of shale, arranged concentrically.

The gravel is overlain along a highly irregular contact by a unit composed of interstratified crushed shale layers, tabular fragments of shale, brownish-yellow gravel, dark grey (10 YR 4/1 m/f) clayey silt and dark grey (10 YR 4/1 m/f) well-sorted granule and fine pebble gravel with fresh granitic clasts (Figure 45).

FIGURE 44



SAND LENS, HH 62-148

Legend

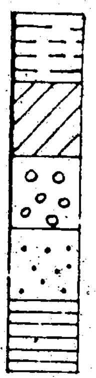
Crushed and disturbed Arctic Red Shale

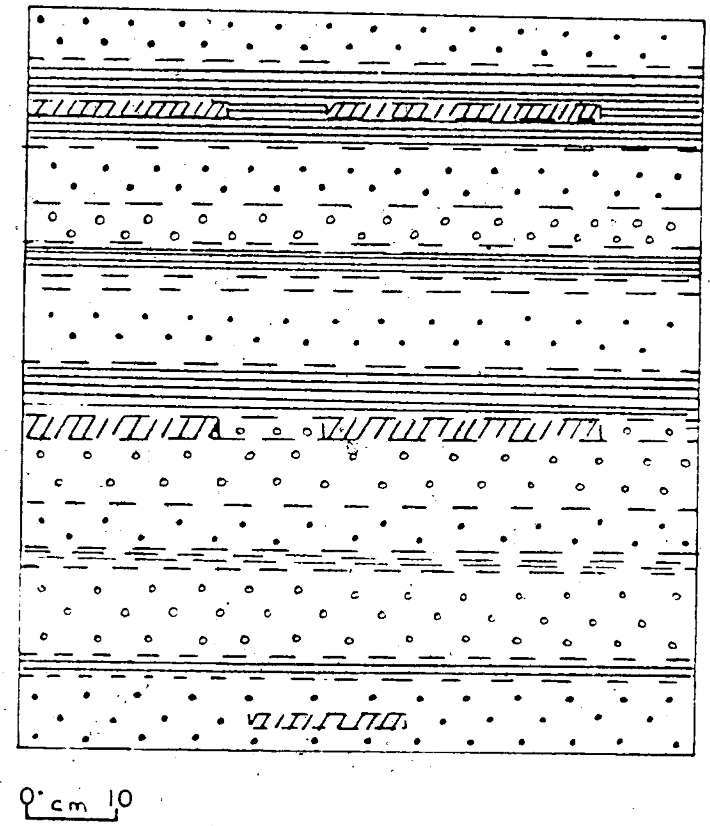
Ovate sand lens, with interbedded crushed shale

See Figure 43 for location of feature.

FIGURE 45
INTERSTRATIFICATION, HH 62-148

Legend





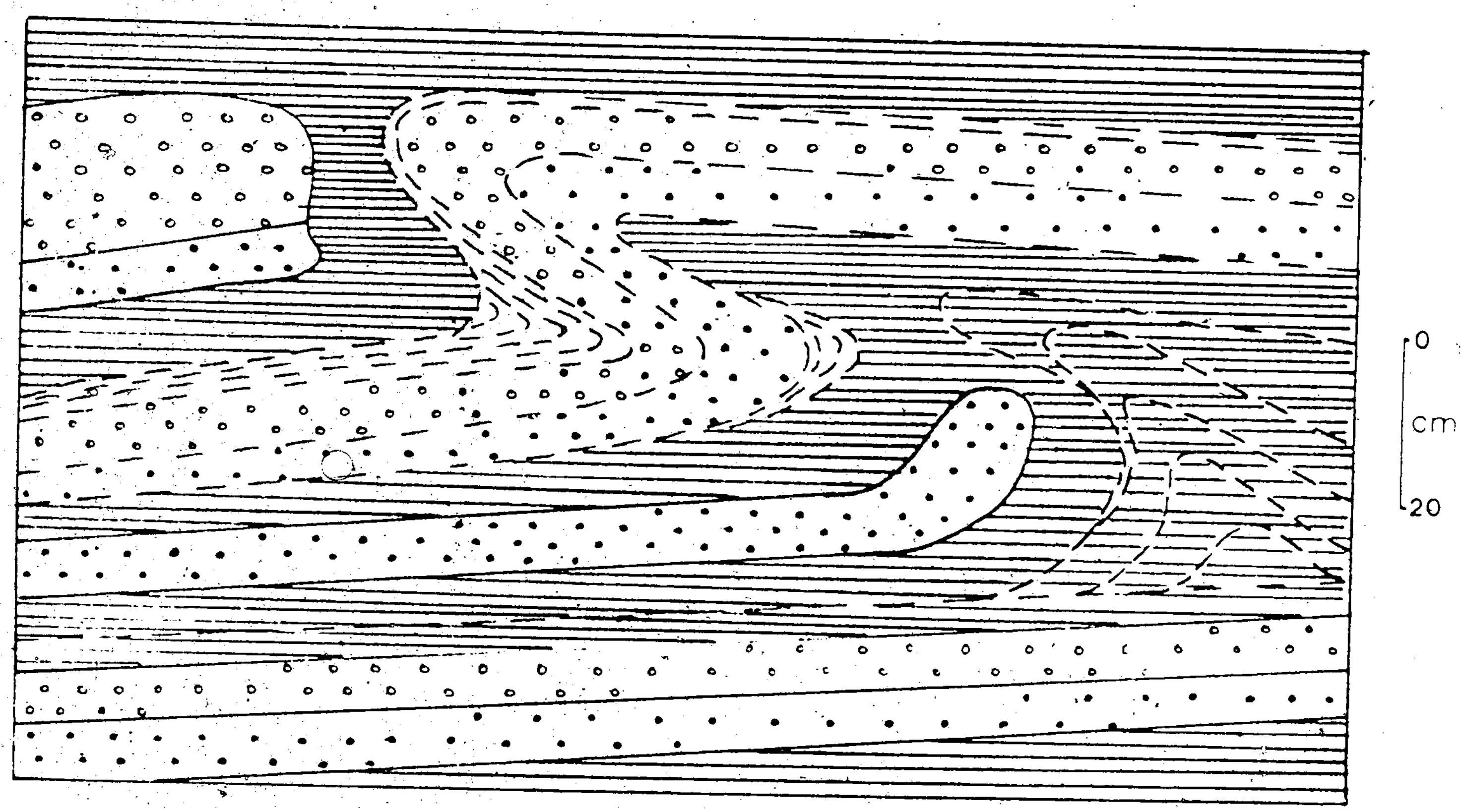
See Figure 43 for location of feature

These strata are not arranged in a discernable cycle, and vary in thickness from 2 mm to 10 cm. The total thickness of the interstratified unit ranges from 50 cm to 155 cm. Individual strata are parallel to each other, and are approximately horizontal (some plunge at low angles to the east-northeast).

Locally, groups of strata are deformed into tight recumbent or overturned folds (Figure 46), indicative of a force applied from the east or northeast. Some laminations in the basal limbs of these folds are truncated, terminating in bulbous structures. Platy shale clasts are aligned with their short axes at 90° to the fold axial planes. Pebbles within the gravel units are aligned with their long axes trending 065-095 azimuth and plunging 20° to 30° northeast.

Also present within the deformed, interstratified unit are orthoquartzite and granitic cobbles and boulders. These clasts are subrounded, of moderate to high sphericity, and have undergone little chemical weathering, as shown by the presence of unaltered hornblende, actinolite and biotite. The cobbles and boulders are embedded throughout the interstratified unit, and the strata are compressed beneath the large clasts and draped over them (Figure 47, Plate 18). To the east or northeast of each large clast is a zone of disrupted strata, often containing fractured pebbles with horizontal long axes. Adjacent to the west or southwest side of the large clasts are one or more contorted laminations, folded parallel to the surface of the large clasts. The

FIGURE 46 DEFORMATION, HH 62-148



Legend as for Figure 45

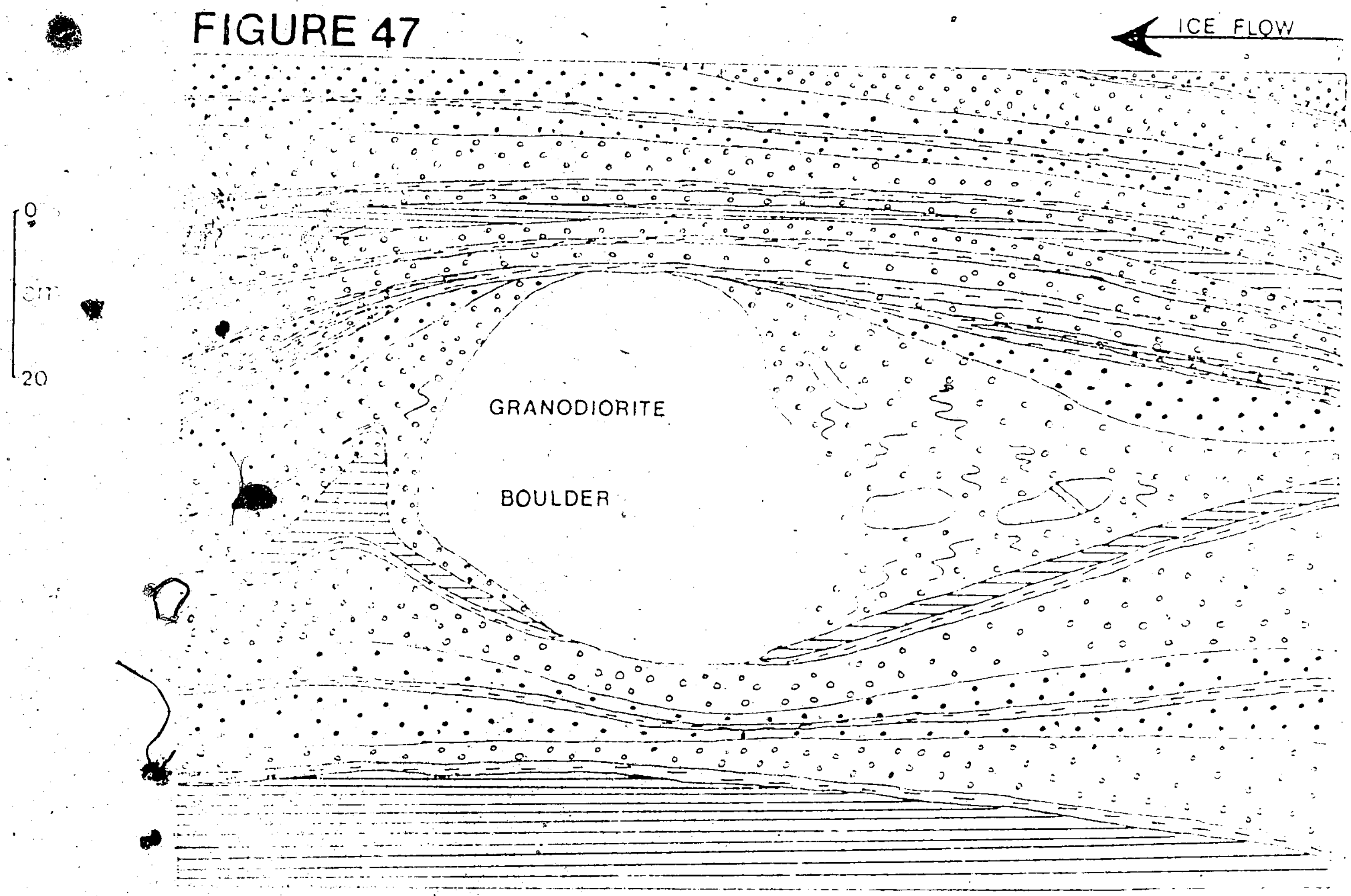
See Figure 43 for Location of feature

Figure 47

Deformation adjacent to Granodiorite Boulder, HH 5[illegible] 14[illegible]

Legend as for Figure 45. See Figure 43 for location of feature; also see Plate 18.

FIGURE 47



folds extend to a height of approximately half the vertical diameter of the adjacent large clasts.

A reverse fault inclined at 60° E separates the western and eastern parts of the section (Figure 48). The vertical displacement along the fault is approximately 2 m. The fault surface is marked by a layer of crushed shale. Pebbles in gravel strata displaced by the fault are aligned parallel to the fault plane in its vicinity. Stratified sediments in the footwall block are deformed into tight recumbent and overturned folds.

Interpretation

The asymmetry of the ridges and the orientation of the ridge and valley complexes normal to the direction of ice advance suggests that the features were formed by glaciotectonic activity.

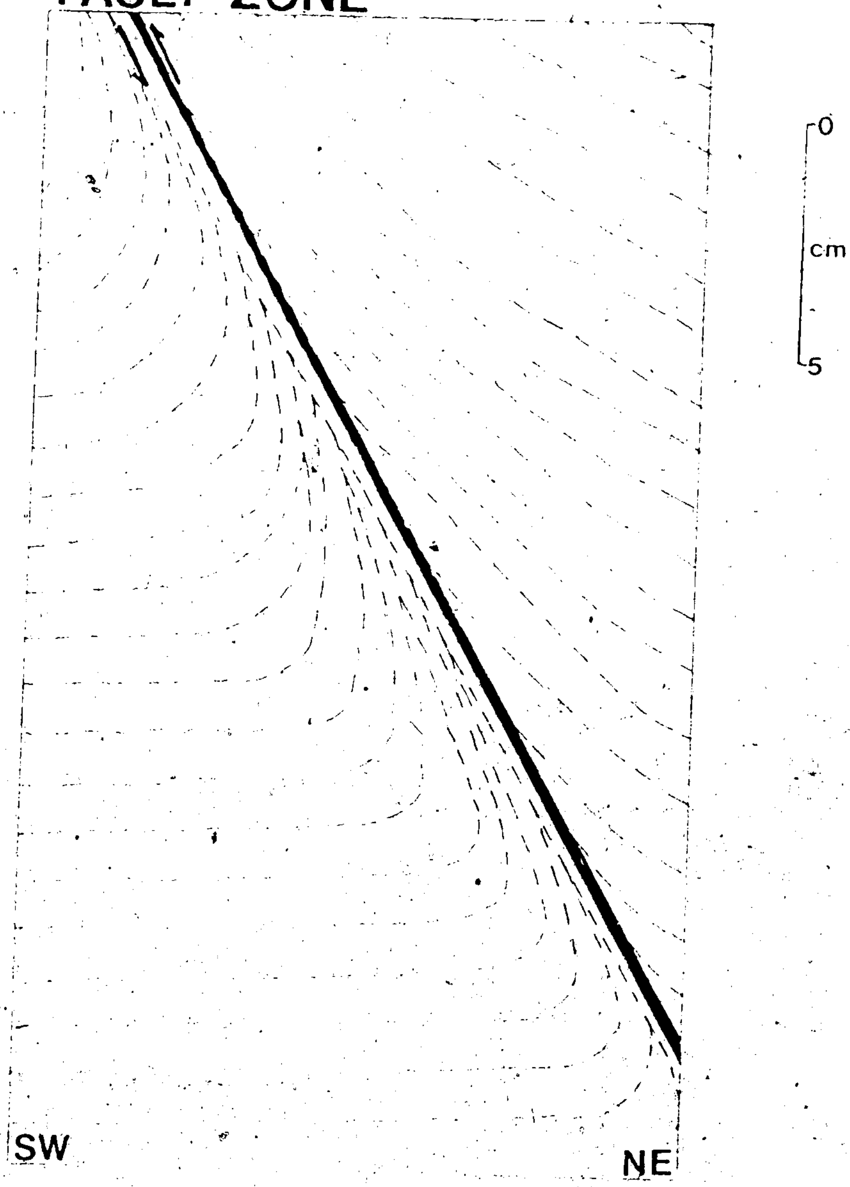
Controversy exists concerning the deformation process for glaciotectonic features. Thrusting induced by excess pore water pressure beneath the marginal zone of the glacier, has been postulated (Moran 1971; Clayton and Moran 1974; Moran <u>et al</u> 1980; Bluemle and Clayton 1984). Other researchers (Kupsch 1962; Berthelsen 1979; van der Wateren 1981) have cited evidence suggesting proglacial thrusting and ice-pushing of arcuate ridges. Sheared and folded material produced by penetrative deformation beneath ice sheets is commonly associated with thrusted sediment as well (Banham 1977; Aber 1979). Polygenetic features, produced by

Figure 48

Detail of fault zone, HH 62-148 81-2.

See Figure 43 for location of feature.

FIGURE 48
FAULT ZONE



combinations of these three processes, have been recognized by Aber (1982). Proglacial thrusting of permafrost terrain in modern Arctic environments has been documented by several researchers (e.g., Kalin 1971; Klassen 1982).

i) Internal Structure

The internal structures of the ridges contain ample indications of shearing and penetrative deformation. The stratigraphic position of the deformed sediments and the presence of igneous and metamorphic rock fragments indicate that glacial ice provided the deforming force. The sawtooth contacts, crushed shale units, and recumbent folds indicate deformation by a southwest-moving force. Large shear planes are marked by the inclined crushed shale units, and by high-angle reverse faults with drag folds developed in the footwall block along the fault contact. Shearing and penetrative deformation are also indicated by the complex stacking and folding of weathered brownish-yellow gravel, fresh granular grey gravel, clayey silt, and shale fragment layers. Pebble orientations parallel to the direction of motion indicate that secondary reworking during flow has re-oriented the fabric from the transverse configuration usually produced by shearing (eg. Rees 1983). This suggests that penetrative deformation was active in shaping the deposit.

Penetrative deformation is also indicated by the ovoid sand lenses within the weathered brownish-yellow gravel and

Arctic Red Formation shale strata at HH 62-148-81-1 and 2. The presence of fresh granitic clasts within these sediments indicates that the lenses were inserted into the strata after deposition of the latter. The ovoid form and the deformation of the surrounding strata suggest that the sands were initially injected as tabular beds or ribbons and were subsequently boudinaged to form the lenses. Injection could have occurred along bedding planes in the shale, or along shear zones initiated at a bed protruberance (cf. Broster _et al_ 1979). Ice lens development along planes of weakness may also have acted to provide suitable conduits for injection of sediment. The concentric structures noted in many of the lenses suggest that boudinage occurred as a result of ductile deformation, resulting in failure by necking, rather than simple brittle deformation. Ductile deformation would enable the laminations and crushed shale layers to be bent into a concentric pattern, whereas brittle deformation would result in a series of lenses with the laminations undisturbed.

The development of these structures in sand and gravel units indicates that the sediments were frozen at the time of deformation. Unconsolidated medium and coarse sands and gravels would not be deformed in a ductile manner. From the absence of silt and clay particles, and the lack of evidence for selective winnowing, it is inferred that interstitial ice was present during deformation.

ii) Spatial Arrangement

Belts of ice-pushed ridges, or thrusted ridges produced in ice-marginal positions, are commonly oriented with the concave sides toward the ice margin in plan view. The eastwardly-convex arcuate ridge and valley belts along the Peel River are oriented in the opposite sense from the inferred direction of ice motion. Although an alternating convex and concave pattern could develop from local lobate advances, the absence of a distinctly concave ridge and valley segment in the area between the convex belts, and the lack of independent evidence for strongly lobate ice flow, makes this interpretation problematic.

iii) Environment of Formation

The spatial arrangement of the ridge and valley complexes and the internal structures of the ridges are most compatible with formation beneath an active ice sheet, in a narrow marginal zone. Aber (1982) proposed that thrusting was initiated proglacially by high pore water pressures generated along a basal permafrost plane of décollement, and subglacial shearing and penetrative deformation upon glacial override acted to modify the features. Subglacial shearing was accentuated by the thrust features, which acted as topographic obstacles to basal flow. Under these conditions, proglacial thrusting would continue to be propagated along the direction of ice advance until either a)the sediments in the proglacial area changed sufficiently to prevent the development of a basal permafrost décollement plane; or b) the thrusted ridges served as a sufficient topographical obstacle to hamper or prevent glacial override; or c) the topography in the proglacial zone varied sufficiently to cause a change in the basal flow conditions of the ice or to prevent the buildup of high pore water pressures. The absence of ridges and glaciodynamic deformation features west of the Peel River indicates that some change in glacial behaviour did occur.

The presence of a basal permafrost plane of décollement is not required for thrusting. Underthrusting has been documented in thick premafrost layers under arctic conditions (e.g. Kalin 1971). Therefore, changes in sediment type would not appear to be critical to initiate or halt thrusting. The Arctic Red Formation in the area is composed of shale and lacks coarse strata. Consequently, sedimentological changes in the proglacial zone were not the cause of cessation of thrusting.

The maximum height of the ridges above the valley floors is 70 m. Although relief of this scale is sufficient to create perturbations in flow in the basal zone of a glacier, it is insufficient to cause a prolonged halt to the flow of a continental ice sheet. No evidence of a prolonged stabilization of the margin adjacent to or northeast of the ridge and valley complexes is present. The presence of drumlinoid features in the valley bottoms, and the penetrative deformation features, indicate that the area was overridden at least once following the initial phases of ridge formation.

The remaining hypothesis is that a topographic feature in the proglacial zone acted to prevent the propagation of thrusting. A deep channel along the course of the modern Peel River would constitute such an obstacle. The presence of badly weathered granite-bearing gravels, both disturbed and undisturbed, indicates that fluvial sedimentation had occurred in the Peel valley prior to the glaciotectonic events recorded in the ridges. Therefore, it is possible that the Peel River was formed marginal to an ice-sheet pre-dating the glacier responsible for the creation of the ridges, and that consequently the ancestral valley acted to halt the propagation of either proglacial or subglacial thrusting and shearing. The absence of preglacial sediments from the Peel River valley precludes a preglacial origin for the ancestral channel.

The development of the ridge and valley complexes would also be aided by any pre-existing vertically-oriented zones or planes of weakness. In the absence of tectonic disturbances and textural contrasts in the Arctic Red Formation shale, a possible mechanism for producing such zone would entail the formation of subparallel meltwater channels. Proglacial streams flowing along the margins of a retreating ice front would excavate a series of subparallel channels (Figure 49). These channels would then act as nucleii for either subglacial or proglacial thrusting during

Figure 49

Genesis of Ridge-and-Valley Complexes

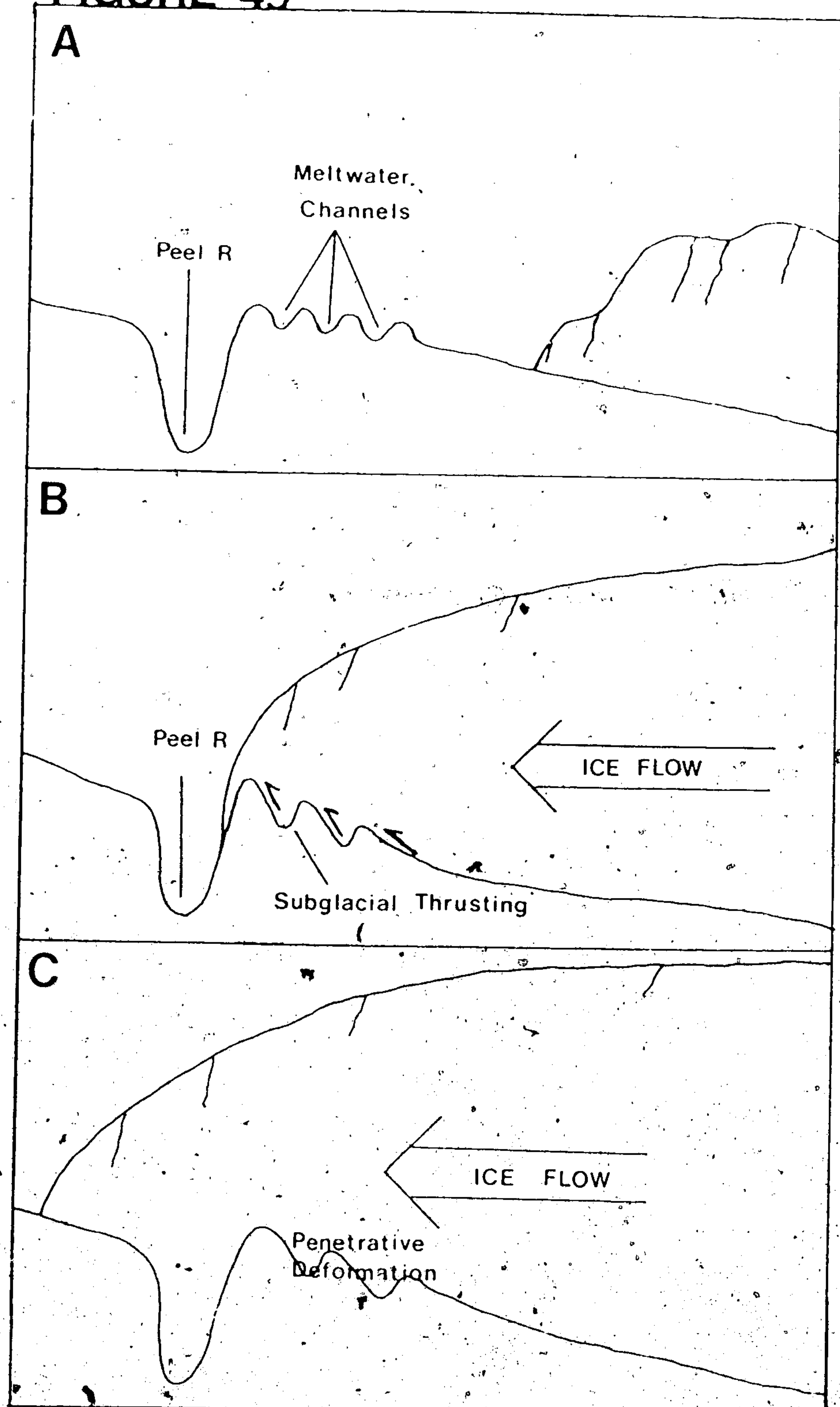
A---Development of Meltwater Channels during initial Deglaciation

B---Meltwater Channels serve as locus for glacial thrusting during readvance

C---Continued Glacial Override leads to penetrative deformation of the thrusted ridges

See text for discussion.

FIGURE 49


A
Meltwater
Channels
Peel R
B
Peel R
ICE FLOW
Subglacial Thrusting
C
ICE FLOW
Penetrative
Deformation

a subsequent glacial event. The ridges produced would not be expected to be oriented concavely in plan view towards the subsequent ice sheet, because the zones of weakness along which they developed would reflect the alignment of the previous retreating glacier. The subparallel arrangement of the valleys, and their conformance to local shifts in the alignment of the Peel River valley, add support to this hypothesis. Although the channels were re-occupied by streams following deglaciation, the asymmetry of the northeast and southwest-facing slopes and the presence of the drumlinoids indicates that the postglacial streams inherited the valleys, rather than being responsible for their formation.

Continued glacial override of the ridge and valley complexes sheared and folded the sediments in the ridges and produced penetrative deformation structures. Some modification of the original ridge form also occurred.

At several locations along the Peel River, crushed shale bands and other deformation structures are overlain by as much as 50 m of undisturbed "Younger Fluvial" sediments (containing Picea and Betula palynomorphs) capped by till (Figure 37).

This sedimentary succession suggests that a significant period elasped between initial glacial deformation and the final overriding, and that the sequence records two distinct glacial events which postdate the formation of the "Brown Bear" gravels.

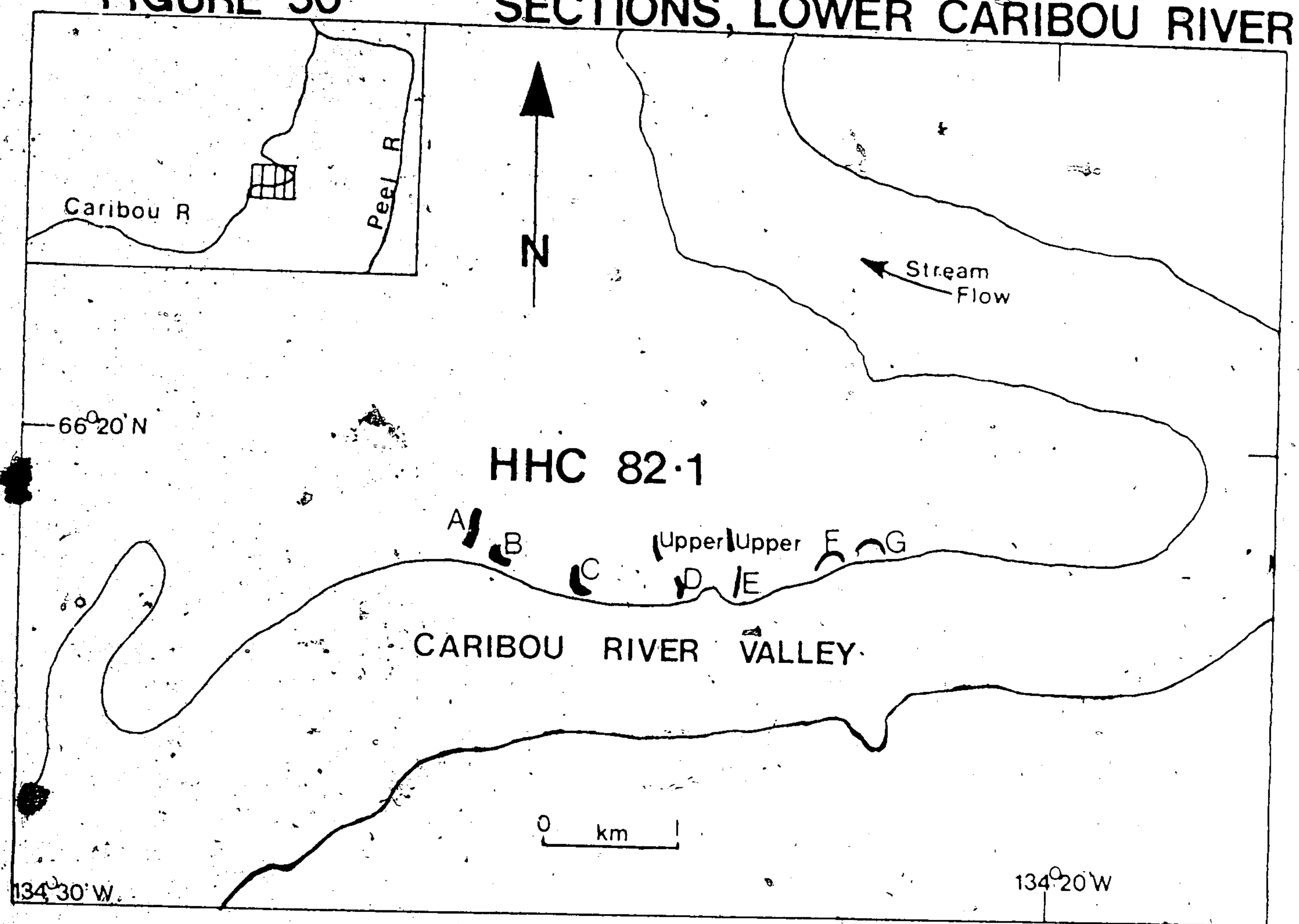
Each glacial retreat left the ridge and valley complexes exposed to fluvial and periglacial modification. The Peel River valley was re-occupied, and the inter-ridge valleys supported minor meltwater streams for short periods. The absence of a prominent moraine system east of the ridge and valley complexes suggests that the ice receded gradually and consistently during the final deglaciation episode, without a major stillstand.

G. Lacustrine Sediments

Lacustrine sediments are present along the lower Caribou River at a series of exposures located at 66° 22' N, 134° 19-20' W. (Figure 50). Undisturbed sediments at these exposures display sequences of coupleted laminations which can be correlated between outcrops. Correlation of these laminations is based on lamination thickness, repetition of vertical arrangements, included sedimentary structures, and elevation (cf. DeGeer 1942; Sauramo 1923; Antevs 1925; Hughes 1965), with due regard for the potential for lateral facies change. The alignment of the sections along the north wall of the valley suggests that facies changes parallel to the modern valley are more likely to be preserved than those representing successive micro-environments developed normal to the valley.

The coupleted sediments can be divided into two principal sequences. The topographically lower sequence is separated from the upper sequence by colluviated material.

FIGURE 50 SECTIONS, LOWER CARIBOU RIVER



The two sequences are not present together in any vertical exposure (Figure 51). Each sequence will be considered separately.

Topographically Lower Sequence

i) Description

The base of the 188 couplets comprising the topographically lower sequence (Figure 52) is exposed at section HHC 82-1c. Couplets #26 to #69 at this section can be correlated to the basal part of the sequence exposed at HHC 82-1a (couplets #1 to #44) and HHC 82-1b (couplets #1 to #10). The base of the lower sequence is therefore represented by the 69 couplets exposed at section HHC 82-1c, plus couplets #45 to #54 exposed at section HHC 82-1a.

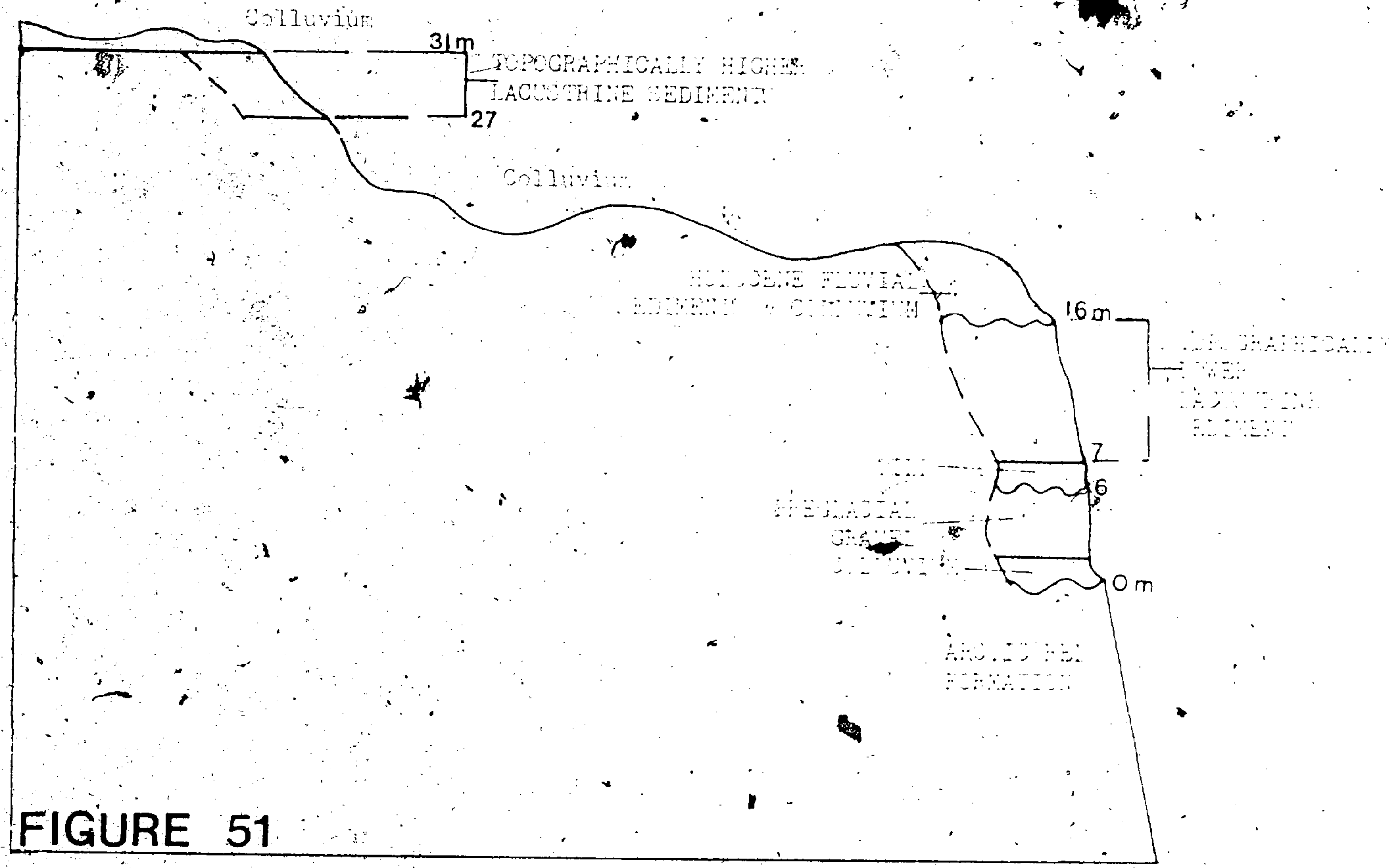
These couplets are characterized by silt and clay layers of almost equal thickness (Plate 19). Silt members range from 0.5 to 2.3 cm (mode 1.5 cm) and clay members range from 0.6 to 1.9 cm (mode 1.2 cm). Silt layers contain clay laminations (1 mm or less thick), ripple-drift cross-lamination (types B and S of Jopling and Walker, 1968), and oriented clay rip-up clasts indicative of eastward flow. The clay layers are generally structureless, but rarely contain fine laminations.

These sediments are succeeded vertically by 8 couplets (#55 to #62 at HHC 82-1a; #21 to #28 at HHC 82-1b), characterized by thicker (2-3 cm) silt layers with an average of 3 fine clay laminations per layer, and

Figure 51

Topographically Lower and Upper Lacustrine Sequences, HHC 82-1d.

The relationship between the Lower and Upper Sequences is illustrated. No stratigraphic conclusions are possible at present. See text for discussion. Datum: base of measured section, 240 m asl.



Colluvium
31m
TOPOGRAPHICALLY HIGHER
LACUSTRINE SEDIMENT
27
Colluvium
HOLOCENE FLUVIAL
16m
7
6
0m

FIGURE 51

Figure 52

Correlation of Couplets, HHC 82-1a, 1b, and 1c.

The figure illustrates the correlations made among the numbered couplets exposed at sections HHC 82-1a, 1b, and 1c. Couplet #1 from HHC 82-1a is correlated to couplet #26 from HHC 82-1c; couplet #35 from HHC 82-1a is correlated to couplet #1 from HHC 82-1b and couplet #60 from HHC 82-1c. The lacustrine sequence is discussed in detail in the text.

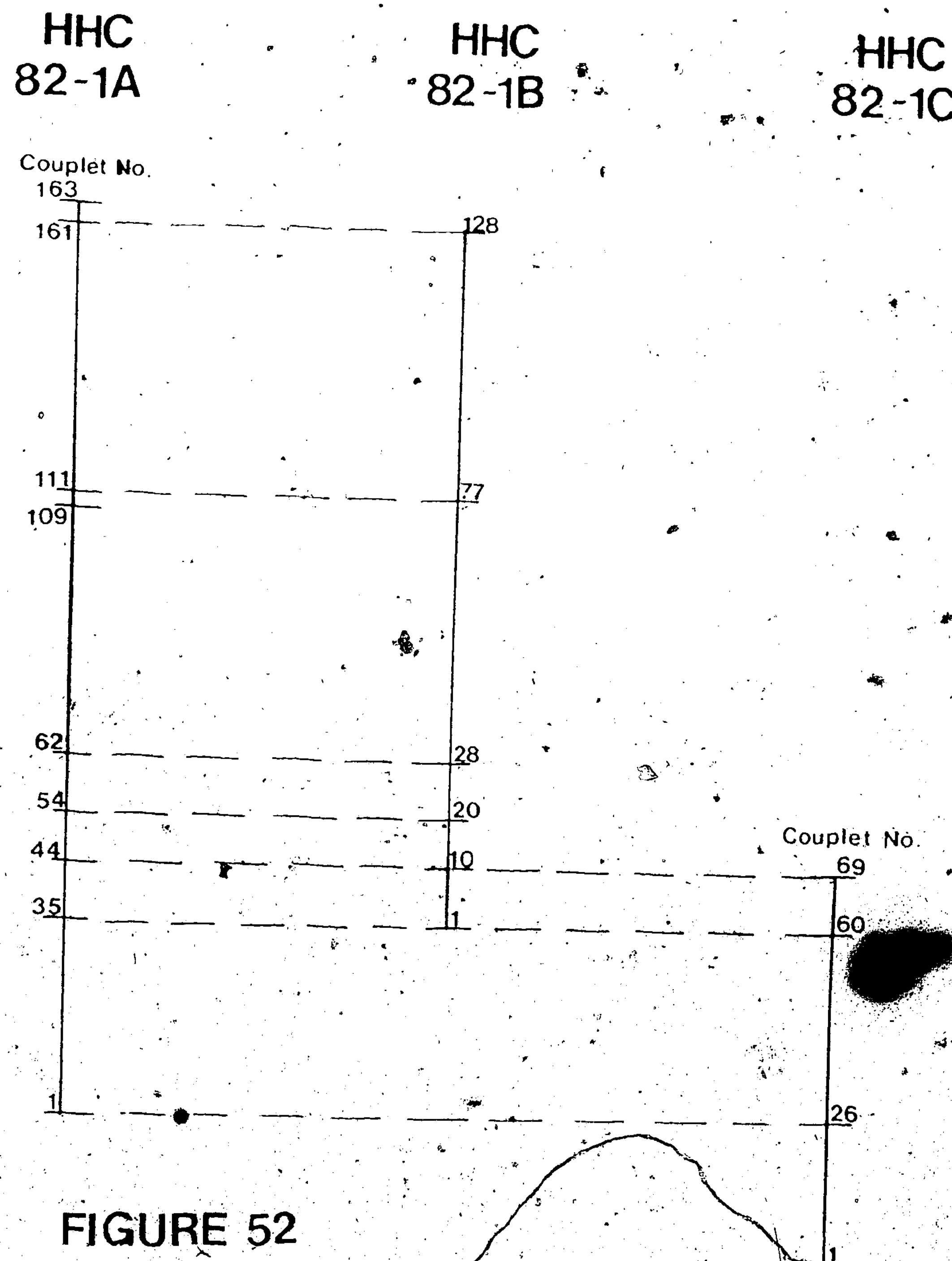
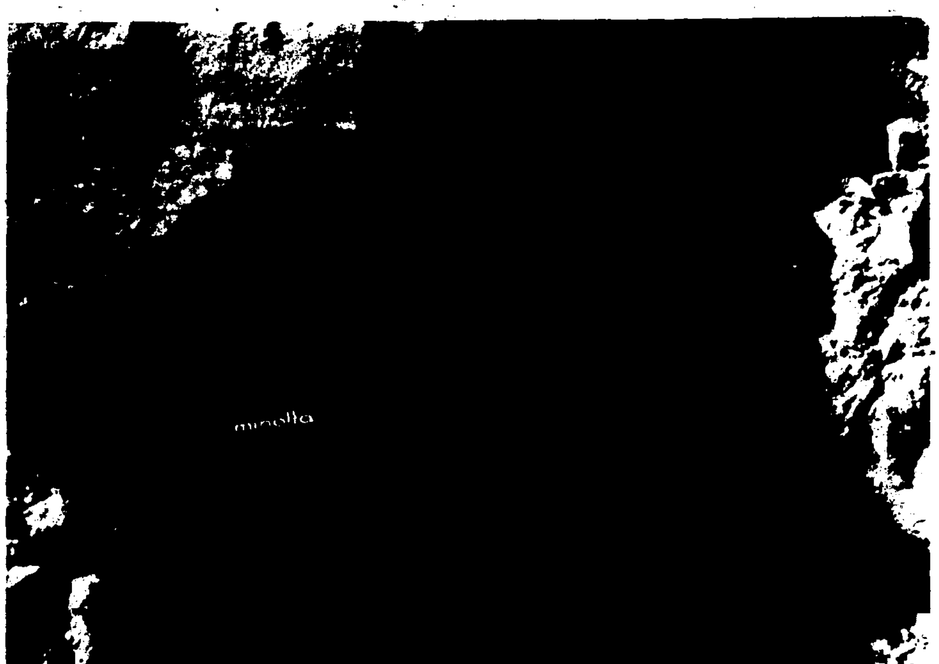


FIGURE 52

CORRELATION of COUPLETS, HHC 82-1A, B, & C

Plate 19

Couplets 49-55, HHC 82 1a. The lower couplets in the vicinity of the lens cap are characterised by silt and clay layers of approximately equal thickness. Couplet [illegible] is characterised by a thick silt member (prominent layer at upper left) and a thinner clay member. Note the presence of fine clay laminations within the silt members.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structureless clay layers 1 cm thick. Contacts between the silt and clay members are gradational, whereas contacts overlying the clays are sharp and planar. Ripple drift cross lamination (type B) within the silt member of couplet #60 (HHC 82-1a) indicates a northerly flow direction (mode 005). An easterly flow direction is indicated by type S ripple-drift cross lamination in other silt members at HHC 82-1b.

The overlying succession consists of 46 couplets at HHC 82-1a (#63 to #108) and 48 couplets at HHC 82-1b (#29 to #76), characterized by thick clay layers (1.8 cm modal, 3 cm maximum) gradationally overlying thin silt units (2 mm modal, 5 mm maximum). Both the clay and silt members are structureless (Plate 22). At HHC 82-1a, these couplets are overlain along an erosional contact by two unusual units. Couplet #109 consists of a basal silt layer 2 mm thick, striking 032 azimuth and dipping 16° ESE, overlain gradationally by a wedge-shaped clay layer, thickening towards the south. The clay is overlain along an abrupt, erosional contact by an irregular silt layer 1.7 - 3.2 cm thick, with a horizontal planar upper contact with clay. The couplets above the clay member of couplet #110 are horizontal. The inclined couplets do not crop out at section HHC 82-1b.

The overlying couplets (#111 to #163 at HHC 82-1a, #77 to #128 at HHC 82-1b) consist of thick silt layers and thin clay layers (Plates 23 and 24). The silts gradually thicken

Plate 20 (top)

The lens cap lies on the contact between couplets 54 and 55, HHC 82-1a. The lower couplets have silt and clay layers of equal thickness, whereas the upper couplets have thicker silt and thinner clay layers. See text for discussion.

Plate 21 (bottom)

A closer view of couplets 56 and 57, HHC 82-1a. The lens cap lies along the contact.

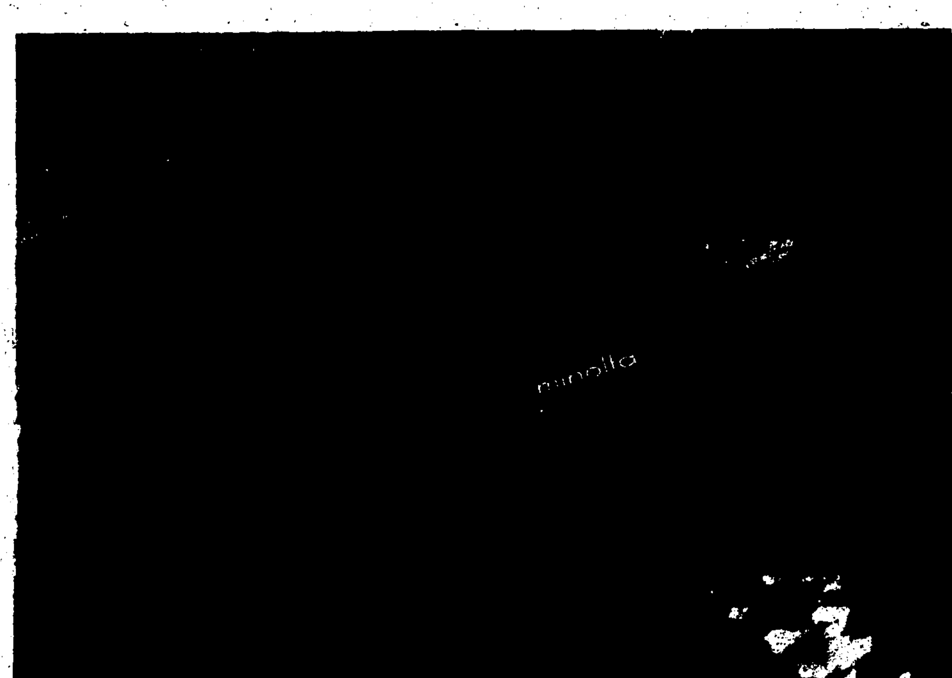

Plate 2[illegible].

Couplets 62-77, HHC 82-1a. The lens cap rests on the contact between couplet 62, characterised by a thick silt member and a thinner clay member, and couplet 63, characterised by a thin clayey silt member overlain by silty clay. The silty clay members appear chocolate brown because they retain moisture more readily. Contacts between the couplets are commonly obscure, and are often marked by horizontal partings or joints.

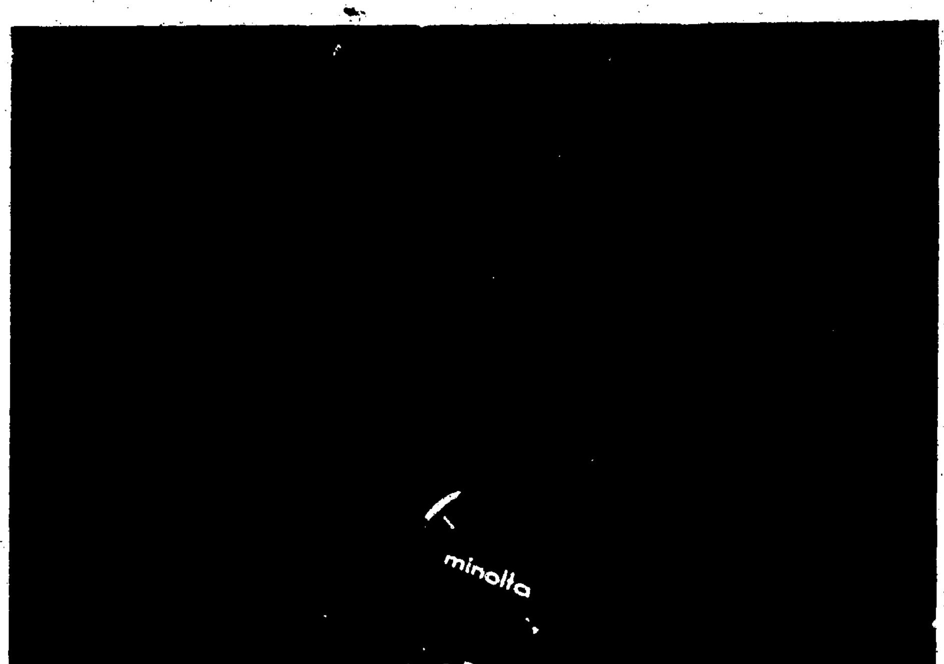
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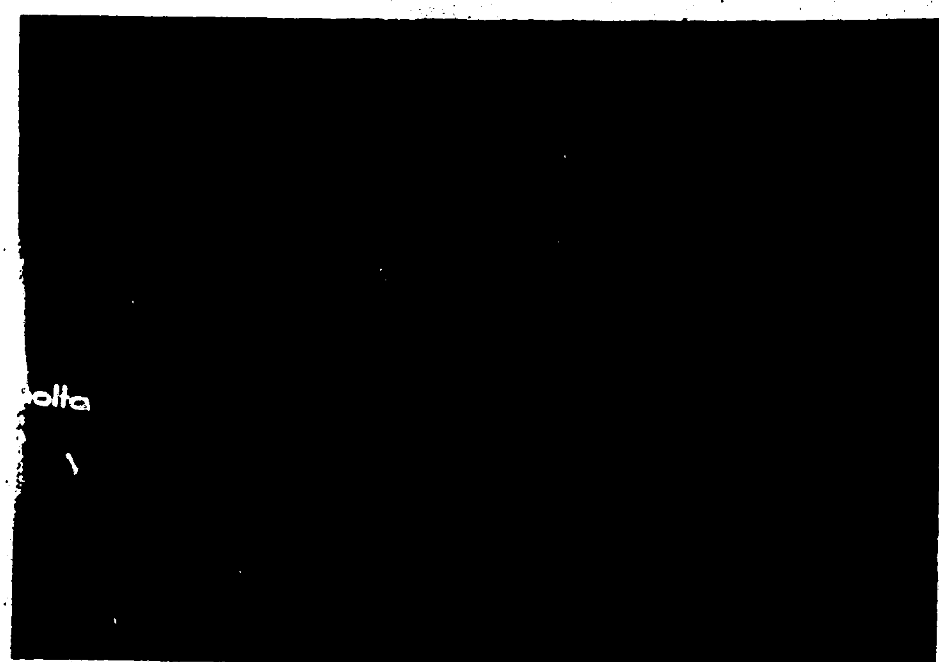
Plate 24 (top)

Couplets 127 and 128, HHC 82-1a. The lens cap rests on the silty clay member of couplet 127. The couplets are characterised by thick silt and thinner clay members.

Plate 24 (bottom)

Another view of the upper part of the lacustrine succession at HHC 82-1a. The upper contact of the silty clay member of couplet 127 is located at the base of the trowel handle.



vertically from a modal value of 4 cm at the base to 8 cm at the top of the sequence. The clay units correspondingly decline in thickness from 1 cm (modal) at the base to 0.8 cm (modal) at the top. Silt layers contain horizontal fine clay laminations, which are convoluted at the base of the silts. Ripple-drift cross-laminations (predominantly type B of Jopling and Walker 1968), loading structures, flame structures, and spoon-shaped inclusions of clay (to 2 cm in length) are also present in the basal parts of some silt layers. The orientations of these structures suggest flow towards the east and northeast.

The couplet sequence is truncated at sections HHC 82-1a, 1b, and 1c by fluvial sediments at an elevation of approximately 250 m asl, 96 m above the level of the Caribou River. Disturbed laminated clay and silt present at sections HHC 82-1d and 1e may be partially equivalent to this sequence.

ii) Interpretation of Couplet Sequences

The couplets of silt and clay can be interpreted, on the basis of sedimentary structure and unit succession, as the products of turbid underflows and other lacustrine currents generated to the west of the exposure locations. The successive changes in couplet thickness and stratification reflect the changing nature of the lake basin.

Basal Couplets (HHC 82-1c and #1 to #54, HHC 82-1a)

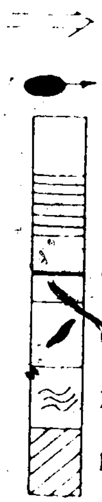
The basal 79 couplets represent proximal deposition in a shallow lacustrine environment dominated by sporadic eastward-flowing turbid underflows (Figure 53). The presence of ripple drift cross-laminations in the silt layers indicate that the flows travelled at velocities of approximately 10 cm/sec (Sorby 1908; Reed 1966). Oriented clay rip-up clasts indicate flow sufficiently vigourous to cause erosion, but of limited duration. The fragility of the clay clasts precludes lengthy transport. The presence of multiple fine laminations of clay within the silt layers suggests that conditions suitable for the initiation of turbid underflows persisted for periods sufficient to permit several episodes capable of transporting sediment. Multiple storm events during a specific season, or sequential influxes of spring runoff from tributary streams could act to initiate several temporally closely-spaced turbid underflow events. The resulting deposits would reflect the multiple flows as a series of silt units separated by thin clay layers representing short periods of quiescence and settling (Figure 53). If the events generating the turbid underflows were temporally limited, the silt strata would be succeeded gradationally by finer sediment. This sedimentation pattern has been observed in several modern lakes (eg. Gilbert 1975; Smith 1978; Lambert and Hsu 1979; Gilbert and Church 1983).

Couplets #55 to 62, HHC 82-1a

Figure 53

Generation of Couplets #1-#54, HHC 82-1a

The couplets were formed by eastward-moving sediment flows. The sedimentary succession for a typical couplet, #34, is illustrated. See text for discussion; see also Plate 19.



Lacustrine Currents

Sublacustrine Sediment Flows

Silt

Silty Clay

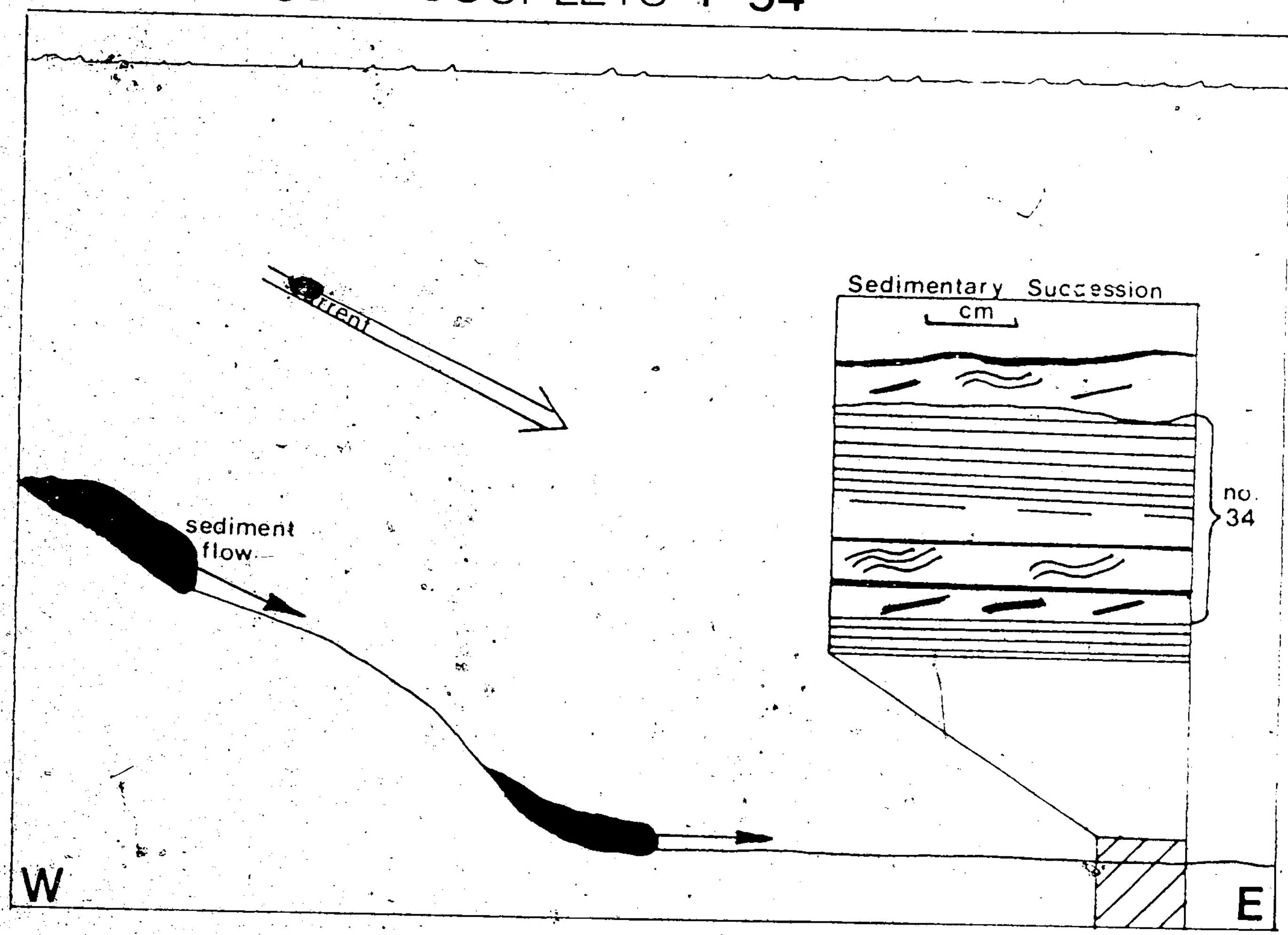
Clay laminae

Clay Rip-up Clasts

Ripples

Previously deposited couplets (#1-#33)

FIGURE 53 COUPLETS 1-54



The eight overlying couplets were produced either in more proximal conditions or under the influence of stronger flows than the underlying sediments (Figure 54). Proximal conditions are indicated by the thick silt members with minor interlaminated clays. Fluctuations in the flow directions recorded at HHC 82-1a and 1b may reflect differing initiation points for underflows. Collapse of local sublacustrine scarps or rises (Shaw and Archer 1978), failure of nearshore sediment (Siegenthaler et al 1984), or sediment surges from fluvial influx (Lambert and Hsü 1979) originating from different tributaries could be responsible for generating flows with different travel directions. In a proximal situation, where the energy levels of the flows are relatively high, local bathymetric obstacles would not be expected to initiate substantial deviations in flow direction, although in distal situations such effects do occur (eg. Kuenen 1951). This stage of lake development is depicted schematically in Figure 54.

Couplets #63-108, HHC 82-1a

The sequence of 46 couplets at HHC 82-1a and 48 couplets at HHC 82-1b, with the thick structureless clay units and thin silt layers, represents a considerable change in the depositional environment from the underlying units. Evidence of slump-generated or runoff-generated underflows is not present. Instead, the sequence is dominated by clay deposited from suspension, with minor silt laminations formed from slow bottom currents (Figure 55).

Figure 54

Generation of Couplets #55–#62, HHC 82-1a

The couplets were formed by eastward moving sediment flows. The sedimentary succession for a typical couplet, #58, is illustrated. See text for discussion; see also Plates 20 and 21.

Previously deposited couplets #55–#57

Previously deposited couplets #1–#54

other symbols as for Figure 53

FIGURE 54 COUPLETS 55-62

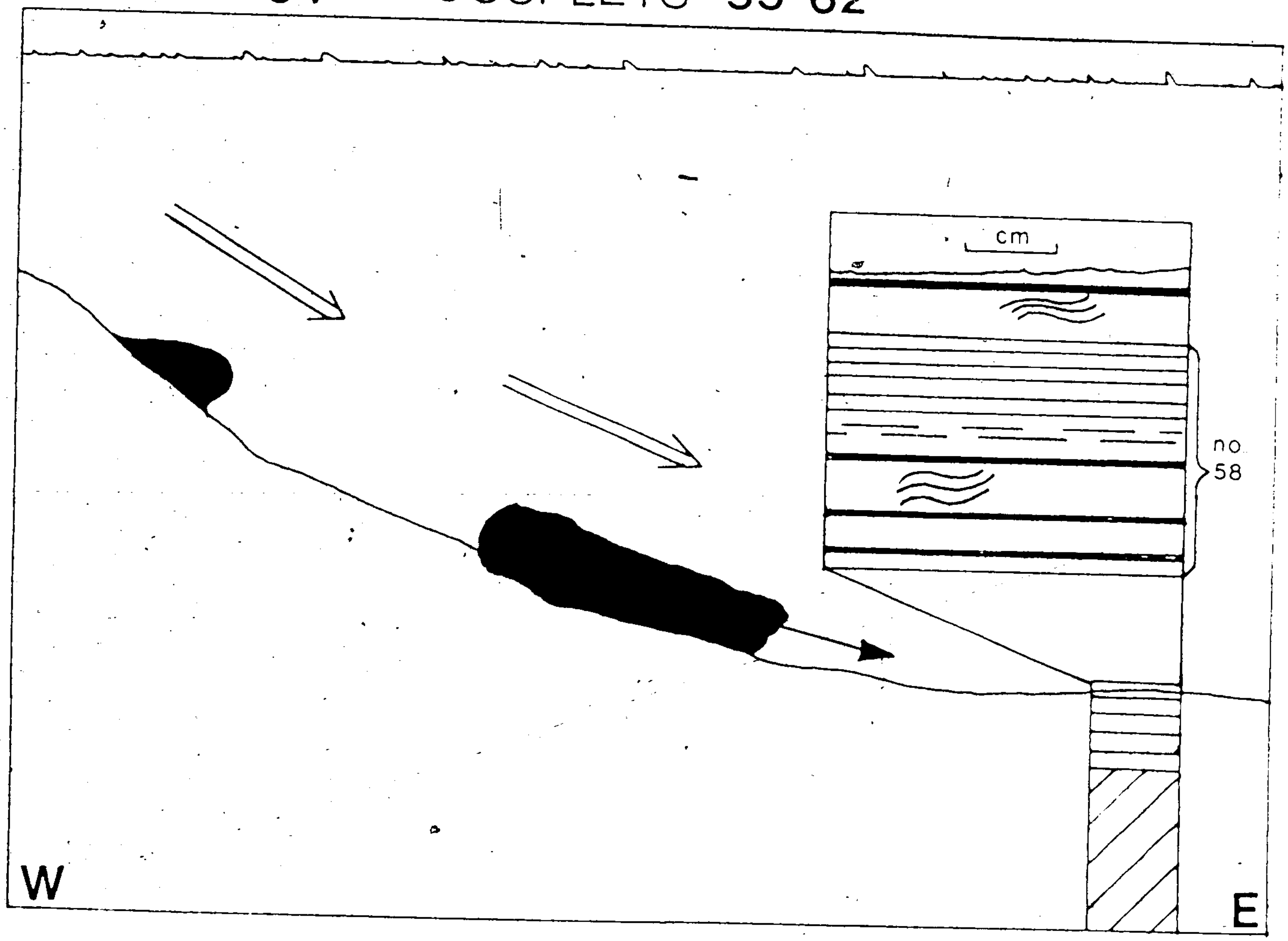


Figure 5[illegible]

Generation of Couplets #63-#108, HHC 82 1a

The couplets were formed by eastward trending currents generated by katabatic winds in a shallow lake. The sedimentary succession for a typical couplet, #78, is illustrated. See text for discussion; also see Plate 22.

Previously Deposited couplets #63-#77

Previously Deposited couplets #55-#62

Previously Deposited couplets #1-#54

Wind Direction

Wind-Generated Lacustrine Currents

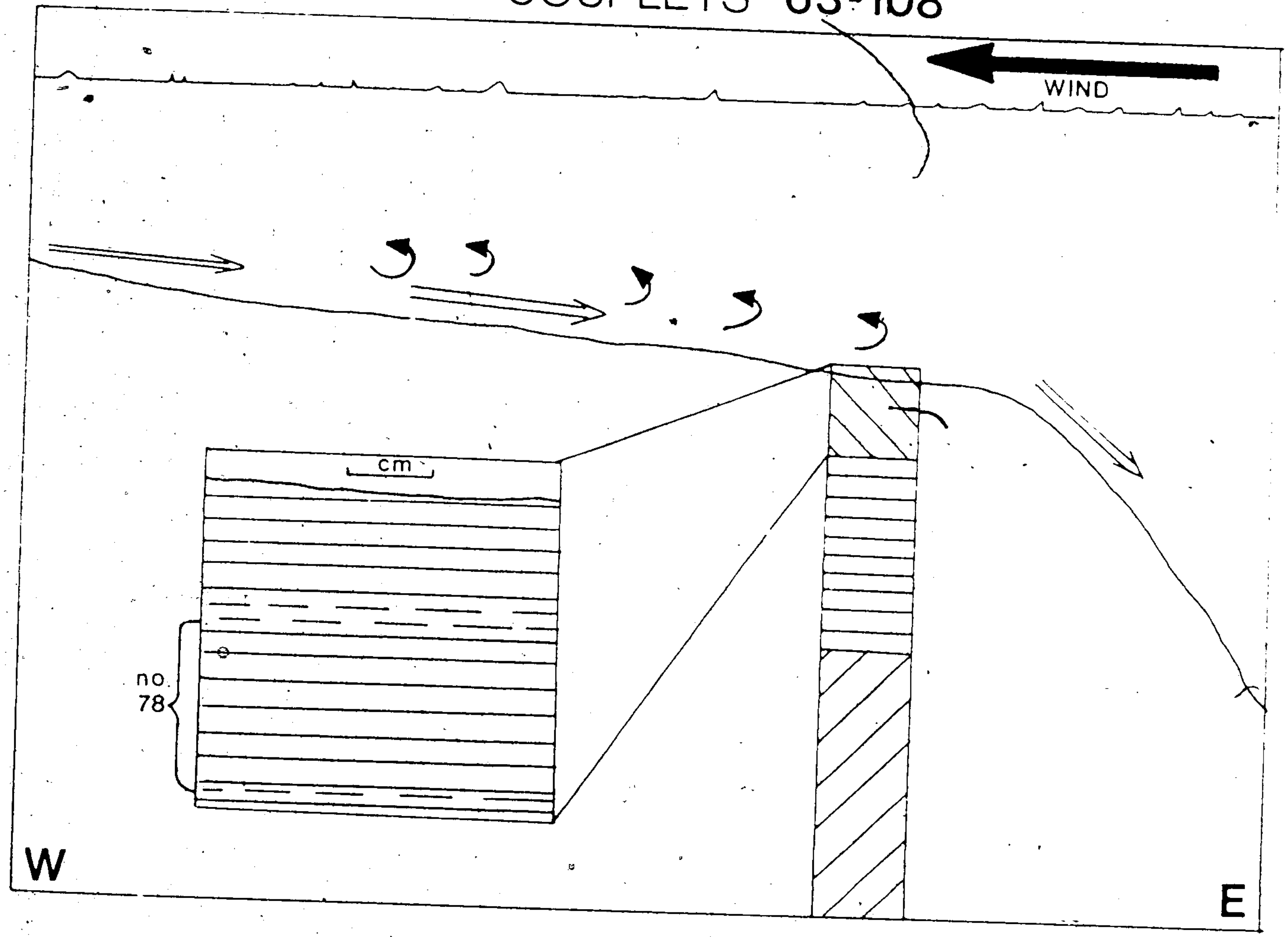
Vertical Turbulence Currents

Fine Sediment in Suspension

Clayey Silt and Silt

Silty Clay and Clay

FIGURE 55 COUPLETS 63-108



Gilbert and Shaw (1981) interpreted a similar sequence formed in Sunwapta Lake, Alberta, to have been produced by settling of suspension load clay combined with diurnal or subseasonal bottom countercurrents induced by katabatic winds. High concentrations of suspended sediment in the shallow (maximum depth 12 m) lake effectively prevented the development of turbid underflows, forcing glaciofluvial influx to enter the lake as inflows and overflows. Katabatic winds moving from the adjacent Athabasca Glacier with velocities of approximately 25 km/hour generated lake bottom countercurrents and vertical turbulence through frictional interaction with the lake surface. Repetitive diurnal or subseasonal bottom current events would result in the production of fine laminations of coarse silt, separated by layers of mixed medium silt and finer sediments, deposited during quiescent periods.

This mechanism can only dominate lacustrine sedimentation in a shallow lake with a very high suspended sediment load. Continued sediment deposition in nearshore areas of the Caribou River lake could act to build up a shallow platform or bank, over which conditions would be suitable for the generation of countercurrents and the deposition of sediment as outlined above (Figure 55). The thickness of the silt laminae, and the mean particle size (0.035 mm), are both greater than the respective values recorded for Sunwapta Lake by Gilbert and Shaw (1981), suggesting that the counter current velocities and vertical turbulence in the Caribou system were larger than their Sunwapta counterparts. Stronger currents would be generated by stronger winds, or could indicate a greater depth than 12 m. The absence of shorelines in the Caribou valley, and the impossibility of assessing palaeowind velocities in this region devoid of coarse grained aeolian sediments, makes a selection between these alternatives impossible at present.

Couplets #109 and #110, HHC 82-1a

The overlying couplets at section HHC 82-1a, #109 and #110, attest to another change in the environment (Figure 56). The sloping, erosional base of couplet #109 suggests relatively rapid flow and deposition on a sloping surface. This surface was then infilled and levelled, as indicated by the wedge form and less steeply dipping upper surface of the clay member of couplet #109. The overlying silt and clay members of couplet #110 resemble the sediments of the lower part of the couplet sequence, and suggest that conditions again were suitable for deposition from turbid underflows.

The silt and clay units of the underlying couplet sequence are plastic and prone to slumping and flowage. A shallow bank of sediments would tend to fail along its margins, either by flow or rotational slumping (Figure 56). Removal of overlying sediments by rotational slumping would produce a curved upper surface in proximal areas, asymptotically declining to the horizontal in distal positions. Removal of overlying sediments by flow could produce either a near-horizontal or gently-sloping upper

Figure 56

Generation of Couplets #109 and #110, HHC 82-1a

The couplets were formed as a result of a mass movement event. The structure of couplets #109 and #110 is illustrated. Correlative couplets were not deposited at section HHC 82-1b. See text for discussion.

Previously Deposited Couplets #63-#108

Previously Deposited couplets #55-#62

Lacustrine Currents

Lacustrine Sediment Flows

Turbulence on Surface of Sediment Flows

Clay

Silt

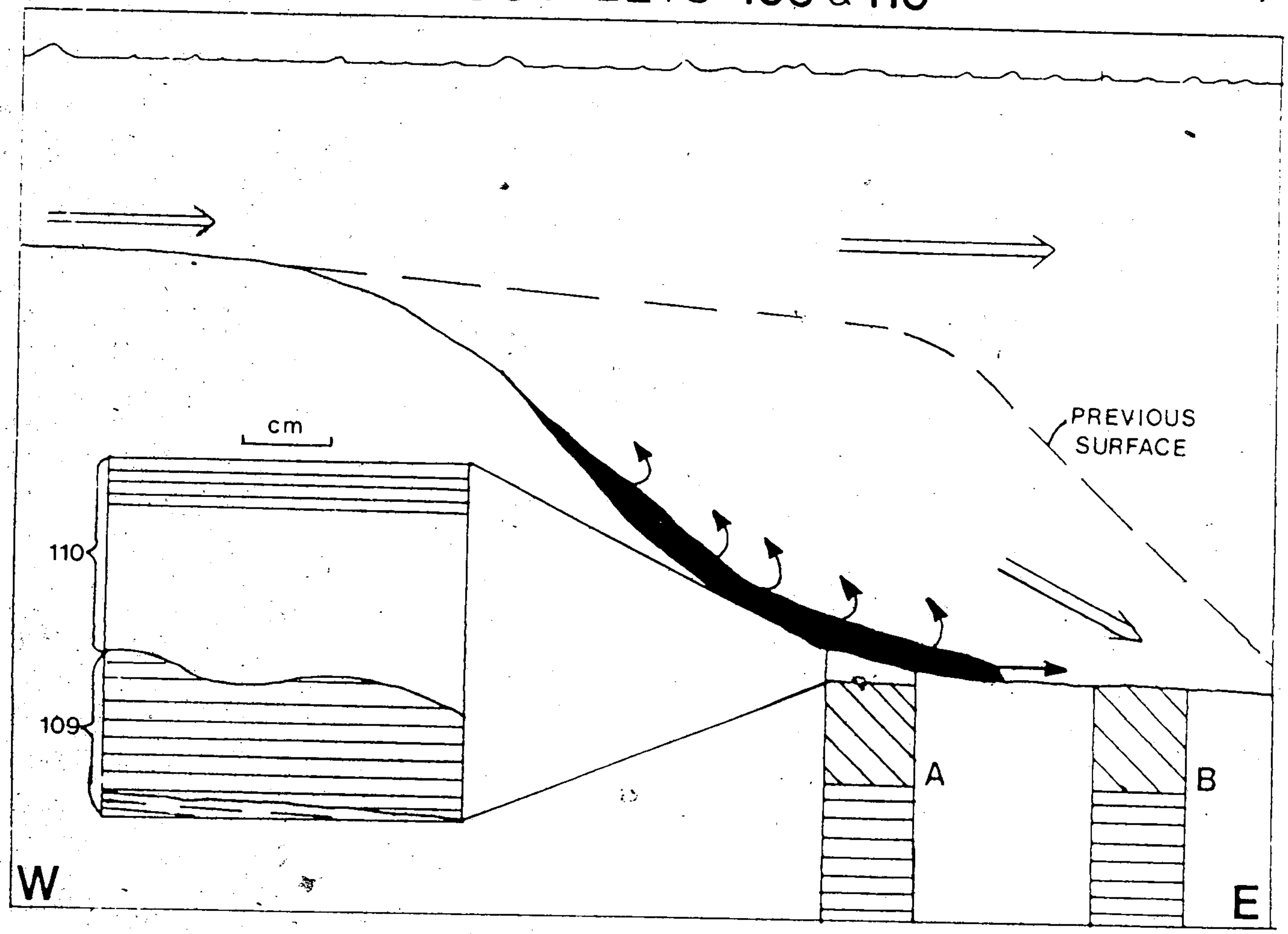
Clayey Silt

A---Schematic Location of HHC 82-1a

B---Schematic Location of HHC 82-1b

Previous configuration of Lake Bottom

FIGURE 56 COUPLETS 109 & 110



surface. The sloping surface created by mass movement would be infilled either by stagnation of the upper part of the slump or flow or settling of suspended load.

The erosional, sloping configuration of the basal contact of the silt member of couplet #109 at HHC 82-1a, and the wedge-shaped infill of clay above, suggest that these units were deposited over an erosional surface formed by slumping or flow of a shallow sediment bank. The direction of mass movement was towards the east-southeast. Section HHC 82-1a was located some distance from the source of the disturbance, whereas section HHC 82-1b was more distal (Figure 56). The absence of laminated blocks of disturbed sediment or abundant rip-up clasts suggests that the mass movement may have been a flow rather than a slump. Slumping in the proximal zone, however, may have initiated sediment flows in the intermediate and distal areas. The material disturbed by slumping may also have lost cohesion and become sufficiently unstable to flow subsequent to the original event. Small-scale occurrences of flow after slumping of ice-rich silts were observed in the modern terrestrial environment.

Couplets #111-#163, HHC 82-1a

Mass movements would be expected to generate turbid underflows downslope from their occurrences (Hyne <u>et al</u>., 1973; Shaw and Archer, 1978; Siegenthaler <u>et al</u>., 1984). The deposition of the succession of couplets with thick silt and thin clay members above the mass movement deposits could therefore be interpreted to be the product of turbidity currents generated by the disturbance. Persistance of the pattern, however, requires that turbid underflows continue for some time after the initial mass movement (Figure 57). Conditions in the area must have been suitable for the generation of successive turbidity currents similar to those recorded in the lower part of the lacustrine sequence. Slumping or flow of the bank would act to increase the slope of the lake bottom and increase the depth of the lake (locally), conditions which would promote the development of turbidity currents. Repeated mass movements along a retreating bank margin could also induce turbid underflows and prevent a return to the countercurrent-dominated style of sedimentation.

The progressive increase in silt member thickness in the upper part of the sequence indicates that the underflows were steadily becoming more proximal in character, suggesting recession of the lake (Figure 57). The sequence is abruptly truncated by fluvial sediment, so the details of the recessive phase cannot be completely ascertained.

iii) Development and Duration

Impoundment of a lake in the Caribou valley during the Late Pleistocene requires the presence of glacial ice to the east, blocking the outlets across the regional slope to the Peel River. Shorelines associated with the lake were not observed. This is probably a consequence of the orientation of the Caribou valley, parallel to both the regional winds

Figure 57

Generation of Couplets #111-#163, HHC 82-1a

The couplets were formed by eastward moving sediment flows. The sedimentary succession for a typical couplet, #132, is illustrated. See text for discussion; also see Plates 23 and 24.

Previously Deposited couplets #111-#131

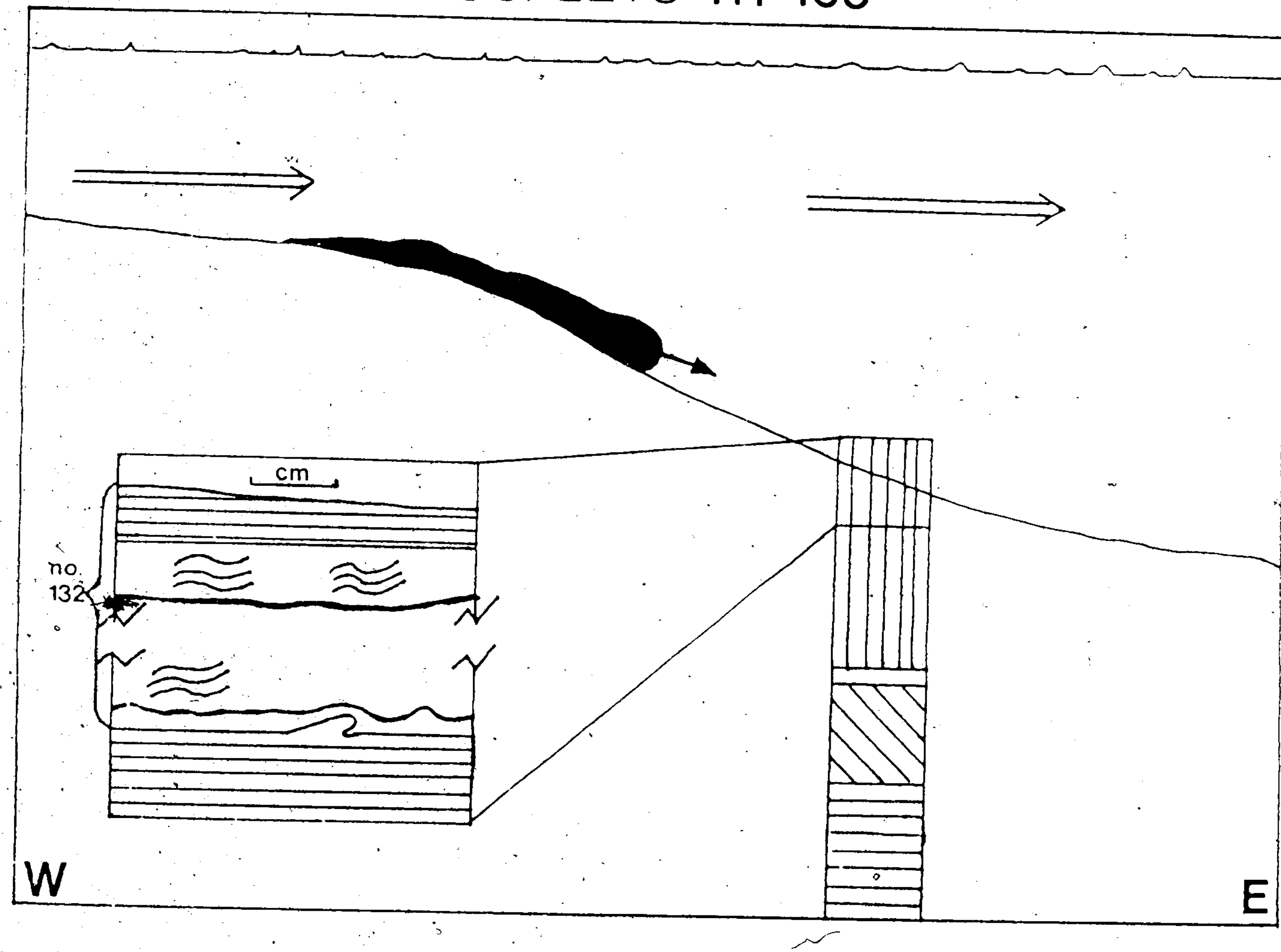
Previously Deposited couplets #109 and #110

Previously Deposited couplets #63-#108

Previously Deposited couplets #55-#62

Other symbols as for Figure 53

FIGURE 57 COUPLETS 111-163



(Burns 1974) and any katabatic winds generated from the glacier surface. The dominant flow direction of the lake bottom currents was easterly, down the regional slope (possibly accentuated by glacioisostatic depression) and towards the ice front. This pattern of flow explains why little coarse material, commonly abundant in glacial lakes proximal to an ice front, is found in the sediments. Drainage of the lake was probably towards the southwest, parallel to the ice front, although no unequivocable evidence is present.

Interpreting the laminated couplets generated as turbid underflow currents as "varves" (cf. DeGeer, 1912; Sauramo, 1923: Antevs, 1925) indicates that these phases of lacustrine sedimentation encompassed 140 years. The strict seasonality of couplet deposition in lacustrine environments has been challenged by several researchers, however, (eg. Gilbert 1975; Lambert and Hsu 1979; Gilbert and Church 1983) and the formation of multiple laminated units within a single season has been observed. It is also possible that circumstances could temporarily interrupt the deposition of couplets in all or part of the lake. Consequently, the 140 year duration indicated by strict application of varve chronological theory remains at best an approximation.

The thick clay/thin silt couplets also pose a chronological problem. Gilbert and Shaw (1981) suggested that similar couplets in Sunwapta Lake could represent diurnal events. This indicates a minimum depositional period of only 48 days for the sequence present at HHC 82-1b, a mean accumulation rate of 2 cm of sediment/day. Such a high rate of deposition is unlikely. More probably, the 48 couplets at HHC 82-1b represent 48 katabatic wind events of sufficient magnitude to initiate bottom countercurrent flow. Estimates based on the grain size and the thickness of the silty laminations suggest that wind velocities of 25 to 30 km/hr would be sufficient to initiate flow. Although this value must be regarded as only an approximation, given the large uncertainties concerning the bathymetry and fetch of the lake, it suggests that conditions suitable to generate bottom countercurrent flow would be frequently achieved during the seasonal periods of open water.

Specific conclusions about the duration of the lake in the Caribou River Valley are therfore not possible. The evidence available indicates that the lake existed for a short period during deglaciation, on the order of one or two hundred years.

Topographically Higher Sequence

A second lacustrine coupleted sequence is preserved at sections HHC 82-1d and HHC 82-1e (Figure 50). This sequence is composed of 93 couplets: the basal 87 are exposed at HHC 82-1d, and the upper 44 at HHC 82-1e (Figure 58). These sediments are topographically higher than the couplet sequence discussed previously, but because they are exposed north of the main valley wall and are isolated by colluvium,

Figure 56

Correlation of Couplets, Sections HHC 82-1d and HHC 82-1e

The figure illustrates the correlation made between the numbered couplets exposed at sections HHC 82-1d and HHC 82-1e. Couplet #50 from HHC 82-1d is correlated to couplet #1 from HHC 82-1e. A typical couplet sequence, couplet #74 from HHC 82-1d, is also illustrated. See text for further discussion; also see Plate 25.

Silt

Clayey Silt

Silty Clay and Clay

Ripples

Swing-dash Laminations

Rip-up Clasts

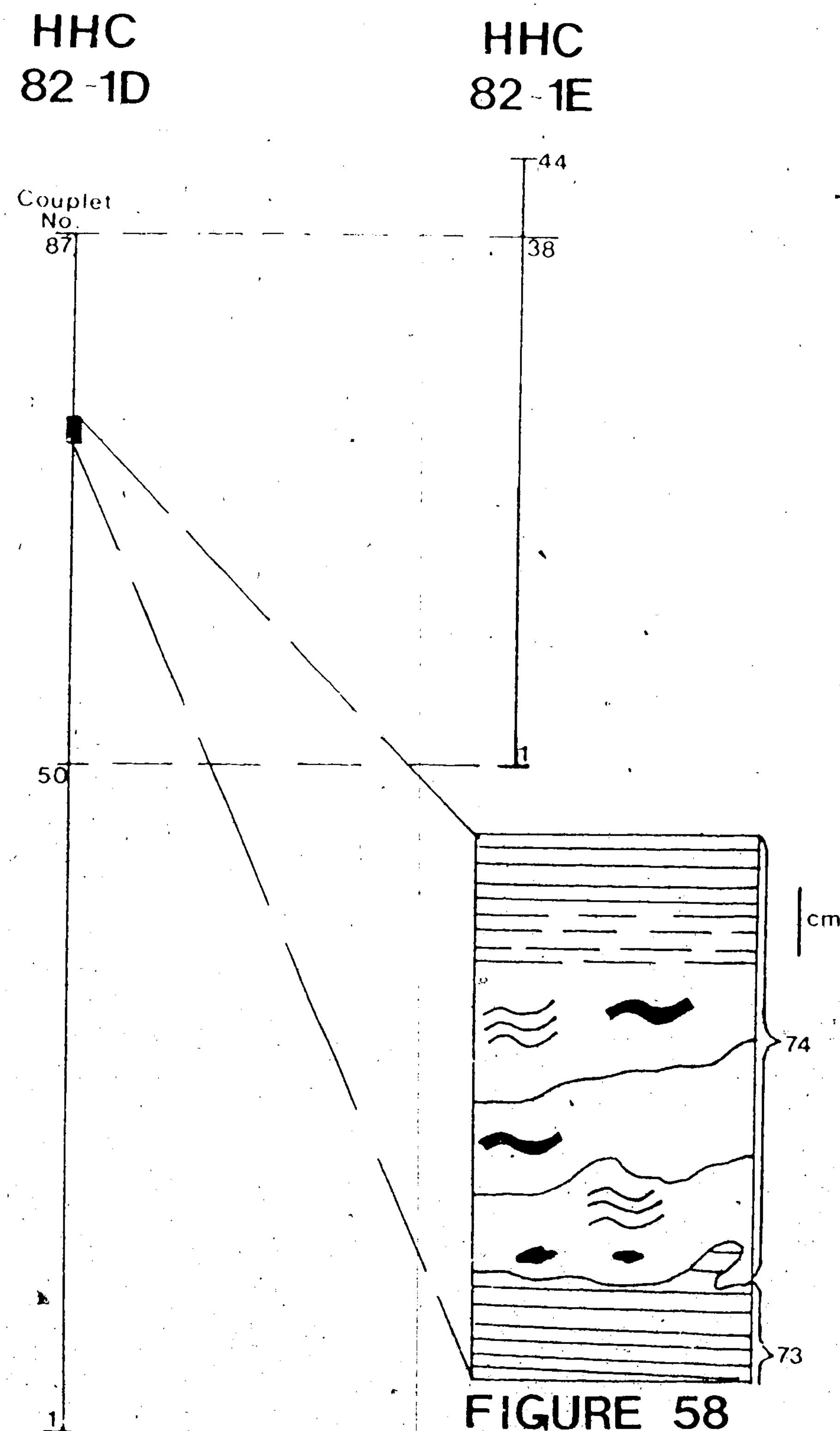


FIGURE 58

no stratigraphic assessment of the relative positions of the two couplet sequences is possible.

i) Description

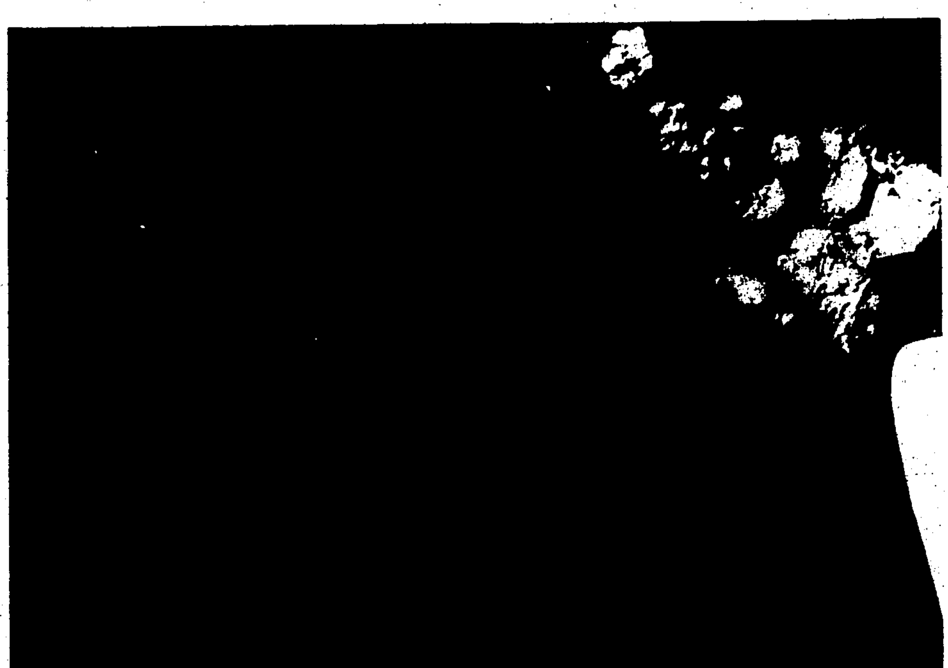
The couplets are dominated by the lower silt members (1-6 cm thick), which increase in thickness vertically (Plate 25). The silts contain regular planar laminations of fine clay (1-2 mm thick) which are often kinked and convoluted. Loading structures, ripple-drift cross-laminations (types B and S of Jopling and Walker 1968), and swing-dash laminations suggest that the direction of current flow was towards the northeast. The clay members of the couplets are thin (0.4 - 2 cm) and structureless. Contacts between silt and clay members are gradational, whereas contacts between couplets are erosional. Granule-sized clasts of clay are commonly present in the basal part of silt members.

ii) Interpretation

The sequence of 93 couplets was produced in a proximal situation, as indicated by the thick silt layers. The presence of ripple-drift cross-laminations and swing-dash laminations produced by draping of fine sediment over ripples indicates that bedload transport was involved in the formation of the coarser units. The regular fining-upwards sequences, truncated by successive events, are typical of lacustrine strata deposited by turbidity currents (eg.

Plate 25

Topographically Upper Lacustrine Sequence, HHC 82-1e, couplets 16-23. Trowel handle is 15 cm long. See text for discussion.



Kuehen 1951; Ashley 1972; Gustavson 1975). The progressive vertical increase in thickness of the silt members indicates that the lake was shallowing throughout deposition of the sequence.

The flow indicators in the sediments suggest that the currents moved generally towards the northeast and east. This direction coincides with the regional slope, but not with the local slope, indicating flow parallel to the modern Caribou River valley rather than into the valley area. Isostatic depression of the terrain adjacent to a retreating ice sheet may have created an increased northeastward slope, facilitating turbidity current flow in that direction.

The lack of couplets correlative to the topographically lower sequence, and the isolation of the sequence amidst colluvium, prevents reliable correlation of the sediment to any other exposure at present.

H. Holocene Sediments and Environments

Holocene sedimentation in the Caribou River Watershed and Brown Bear Creek area is dominated by braided-meandering stream deposits formed along the Caribou and Peel Rivers, braided stream deposits formed along the flanks of the Richardson Mountains, and meandering stream deposits formed along small streams in the Peel Plateau and Peel Plain. These environments are discussed in Appendix 4.

Palynological assemblages from eight sites show a gradual pattern of climatic amelioration throughout the early and middle Holocene, superimposed upon local conditions related to aspect and fluvial activity. In many instances, local problems of hydrodynamics, alteration of the permafrost profile by river erosion and channel shifting, and the local microclimates created in the immediate vicinity of the streams completely control the nature of the palynomorph assemblages. The strength of the local influences makes the assembly of regional pictures difficult. The assemblages can be divided into two categories: those deposited in fluvial sediments, where fluvial processes dominate; and those present in autochthonous peat and lacustrine deposits.

Palynological Assemblages in Fluvial Sediments

i) HHC 82-1a

This section is located along the north bank of the Caribou River (Figure 50, 52).

The base of the section is composed of Cretaceous Arctic Red Formation siltstone, which extends 80 m above the level of the Caribou River. The bedrock is overlain by 10 m of covered and colluviated sediment. This unit is overlain by 5.75 m of lacustrine sediment, composed of 163 couplets (Figure 52). A sample taken from the silt member couplet 158 proved devoid of palynomorphs and diatoms.

The overlying unit, a finely laminated clayey silt (21-26% sand, 23-32% clay) 30 cm thick, was also barren. This unit is interpreted to represent either a tundra pond or an abandoned back-channel slough.

These units are overlain along an abrupt planar contact by a 16.1 m thick complex of clay, silt, sand, and organic lenses, beds, pods, and laminations, all of fluvial origin. The members of this complex are all vertically and laterally discontinous, and the greatest thickness achieved by any bed is 42 cm.

The sand units are iron- and manganese-oxide stained. Horizontal and ripple-drift laminations (easterly flow) are present. The sands comprise 20% of the unit, and are dominantly quartz and feldspar. Metamorphic and igneous minerals, such as garnet, magnetite, zircon, apatite, wollastonite, tourmaline, fluorite, kyanite, chromite, cordierite, and anatase occur in trace amounts.

Silt members comprise 50-60% of the complex. They are commonly horizontally laminated, less regularly ripple cross-laminated, planar cross-laminated (easterly flow), convoluted, or laminated with clayey swing-dash forms. Load structures indicating easterly flow and peardrop clay balls are also present. The silts are generally calcareous and are usually iron-stained.

The clay beds form 25-30% of the complex, and are extensively iron and manganese-stained. Two basic types of clay units are present. The majority of the clays are finely laminated or structureless bands approximately 10 cm thick. Organic detritus is not associated with the majority of these clays. The second type of clay stratum consists of an aggregation of clay balls, including both spherical and teardrop types. These units are usually 3 cm thick or less.

Organic detrital beds form a very small constituent of the complex. They are usually found as a coating over rippled sand or silt, bedded immediately above heavy-mineral laminations in sand, or as discrete laminae in the finer silt beds. Most of the material is very fine detritus. Twigs present in the beds are occasionally oriented parallel to the inferred current direction (easterly) but are more commonly random in orientation.

A total of 23 samples from this unit were processed for palynomorphs: 4 from sand members, 8 from silts, 4 from clays, and 7 from organic detrital layers. Four of the organic detrital sample, and one each of the sand, silt, and clay samples contained sufficient palynomorphs for counting (Table 17). Five other samples contained lesser amounts of pollen and spores. Twelve of the samples were processed for diatoms, but none were observed.

The palynonologcal assmeblages in the fluvial complex sediments are dominated by Picea (primarily mariana type) and Betula (shrub type), as well as Cyperaceae and Gramineae. Minor amounts of Salix, Larix, and Alnus are present in the arboreal shrub fraction. Pinus, noted in sample #29, was probably not an inhabitant of the area at this time. All Pinus pollen observed was abraded, suggesting long-distance transport. The shrub component is relatively minor, and is dominated by Empetrum nigrum and the

Table 17

Palynological counts, Quaternary taxa, HHC 82-1a. All taxa included in pollen sum

	Organic		detritus		sand	silt	clay
Sample #	21	[illegible]	34	36	30	28	[illegible]
Taxon	%	%	%	%	%	%	%
Picea	19.8	20.0	28.6	20.0	15.1	27.9	19.5
Pinus	0.0	2.2	0.0	0.0	0.0	0.0	0.0
Betula	18.9	6.5	19.0	4.3	8.1	16.3	14.3
Alnus ty. *crispa*	0.0	0.0	0.9	0.0	0.1	0.2	0.0
Salix	0.0	0.0	1.0	0.0	0.0	0.5	1.6
Empetrum	1.6	2.7	0.0	1.7	3.0	1.8	1.0
Larix	0.0	0.0	0.0	1.7	0.0	0.0	0.3
Shepherdia	0.0	0.0	0.0	0.0	0.0	0.0	0.7
Ledum	1.0	1.1	1.0	0.0	2.4	1.2	0.3
Vaccinium	1.1	1.0	1.0	0.0	2.9	1.2	1.0
Andromeda	0.0	0.0	0.0	0.0	0.2	0.0	0.0
Myrica	1.0	0.0	3.8	0.0	0.3	0.0	3.2
Cyperaceae	24.7	16.8	15.8	20.0	37.9	18.8	13.4
Gramineae	19.9	20.5	9.5	12.2	3.4	16.0	17.6
Artemisia	3.9	15.7	4.8	1.7	3.7	5.0	3.9
Arnica	1.0	0.0	1.9	0.0	0.3	0.0	0.7
Sausserea	0.0	0.0	0.0	0.0	0.0	0.0	0.3
Cruciferae	0.0	0.0	0.0	1.7	0.0	0.2	0.0
Dryas	0.0	0.0	1.0	0.0	0.0	0.0	0.0
Chenopodium	0.0	0.0	0.0	0.0	1.1	0.0	0.3

Table 17

(continued)

	Organic		detritus		sand	silt	clay
Sample #	21	29	34	36	30	28	33
Taxon							
Cerastium	0.2	0.5	0.0	0.9	0.0	0.0	0.0
Polystichium	0.0	0.0	0.0	0.0	0.2	0.0	0.7
Epilobium	1.5	1.6	2.9	0.9	1.1	1.5	5.2
Rubus chamaemorus	0.0	0.0	1.0	0.0	0.0	0.0	0.3
Lycopodium	1.2	0.5	0.0	2.6	11.4	1.5	1.3
Cystopteris	1.3	4.9	3.8	27.8	2.0	0.9	12.7
Equisetum	0.6	4.9	0.0	0.0	1.1	1.2	0.7
Sphagnum	2.1	1.1	3.8	3.5	5.9	5.6	0.7

Ericaceae. In the fine-grained samples, shrubs comprise 1.7-5.8% of the spectrum, as compared to 8.6% of the spectrum of the single sand sample. The herb fraction of the sand sample is dominated by Cyperaceae (37.9%), with minor amounts of Gramineae, Chenopodium, Artemisia, Epilobium, and Arnica grains. The fine-grained sample spectra are co-dominated by Gramineae and Cyperaceae, with associated Epilobium, Artemisia, and Arnica, as well as minor amounts of other pollen. The pteridophyte fraction in the sand sample is dominated by Lycopodium, while in the finer samples Cystoperis spores are dominant. Some of the grains of both genera may have been derived from the adjacent Cretaceous bedrock. Sphagnum spores are present in varying amounts in all samples. Pre-Quaternary grains are present in all the samples. The concentrations varied between 10 and 25% in the fine-grained samples and exceeded 40% in the sand sample.

No consistent vertical trend was observed in the spectral variation, and the oscillations in palynomorph concentrations present can be ascribed to variation due to fluvial transport (Catto 1985). It is likely that Lycopodium, Ledum, Vaccinium, Empetrum nigrum, and Cyperaceae are over-represented in the sand sample spectrum. Conversely, Picea, Betula, Gramineae, and Cystopteris are probably over-represented in the finer samples. Reworking of material from the bedrock has also caused fluctuations in the concentration of Cystoperis.

With these limitations in mind, the assemblages can be interpreted to represent a cold riverbank/back channel tundra environment, with scattered groves of Picea and Betula shrubs covering stable areas along the river system. Open bar areas were colonized by Shepherdia canadensis, Cerastium, Artemisia, Epilobium, Arnica, and Lycopodium, while the slough and back-channel margin areas supported Ledum, Vaccinium, Salix, Larix, Myrica Gale, Rubus chamaemorus, and Aster. Alnus may have been present, but the small proportions of the palynomorphs suggest that these grains were transported by westerly winds from Kamchatka or southeastern Alaska. This hypothesis is more compatible with the regional picture reconstructed by Hopkins et al (1981), and by Ritchie (1984).

The fluvial complex is overlain by 1.4 m of medium-fine grained sand. The lower portion of this unit is planar cross-laminated, with the laminations oriented 040. The upper portion is horizontally laminated. A sample from the upper portion of the sand was devoid of palynomorphs. This unit is interpreted as an aeolian deposit. Overlying the sand is 2.7 m of colluviated sand-silt-clay (35% sand, 42% silt, 23% clay) which extends to the top of the section.

ii) HHC 82-1c

This section is located along the north bank of the Caribou River, approximately 120 m east of section HHC 82-1a (Figure 50).

The stratigraphy exposed at HHC 82-1c is very similar to that of HHC 82-1a. The basal 82.5 m is Cretaceous siltstone. This is overlain by 10.7 m of covered material. The basal Quaternary unit exposed is a lacustrine silt and clay sequence, 251 cm thick, composed of 69 couplets (Figure 52). A sample obtained from the silt member of couplet 34 was processed for palynomorphs, but proved to be barren. Overlying the lacustrine sediments is a 45 cm thick bed of clayey silt (50% silt, 47% clay) which was devoid of palynomorphs. This bed in interpreted as a pond deposit, not related to the lacustrine episode represented by the couplet sequence.

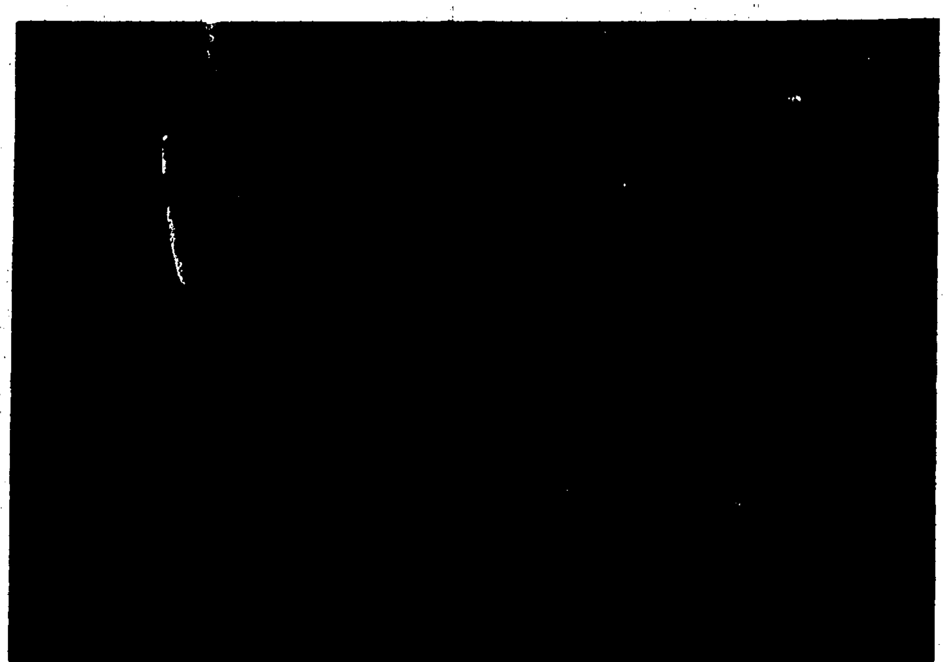
These units are overlain unconformably by a complex of sand, silt, and detrital organic beds with wood fragments. The complex totals 15 m in thickness, and is similar to the fluvial complex exposed at HHC 82-1a. A fragment of Salix (sample #35) obtained from 2.3 m below the upper contact of the complex has been ^{14}C dated at 9,780±110 B.P. (GSC-3573). A total of 19 samples from this unit were analysed for pollen. None proved to be sufficiently rich for counting, although all contained Picea, Betula, Cyperaceae, and Gramineae grains. Traces of Empetrum nigrum (2 samples), Ledum(8), Vaccinium(4), Arctostaphylous rubra/alpina(2), Aster(1), Equisetum (2), and Sphagnum (13) grains were observed, along with Lycopodium and Cystoperis grains and pre-Quaternary taxa. The assemblages are thus similar to those noted in the lithologically correlative fluvial

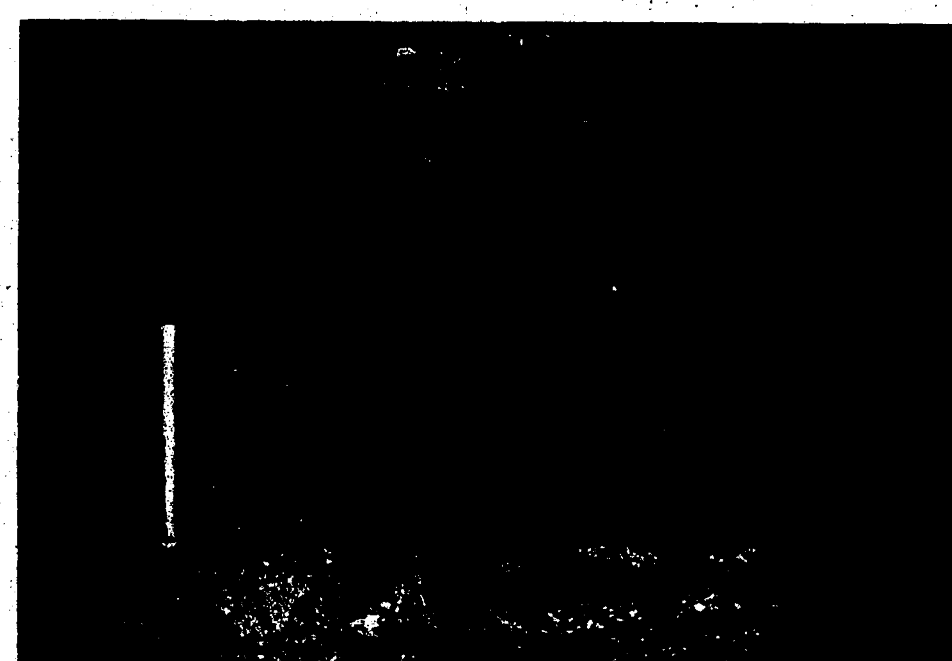
Plate 26 (top)

Back-Bar Channnel Sediments, HHC 8. 1a. These sediments are predominantly rippled sand and silt, with thin clay members. See text for discussion.

Plate 27 (bottom)

Organic detrital unit in fluvial sediments, HHC 82-1c. A Salix fragment from this bed was ^{14}C dated at 9,780 ± 110 B.P. (GSC-3573), and represents a maximum age for these sediments. The sediments outcrop 109 m above the Caribou River.





complex of 82-1a. Four samples were checked for diatoms with negative results.

Overlying the fluvial complex is a horizontally laminated sand unit with occasional silty layers, with a maximum thickness of 2.6 m. Small ripple laminated and trough cross-laminated horizons indicate that flow was to the east. A single sample from a slightly organic silt layer within this unit proved to be barren. This unit is tentatively interpreted as a bar-tail slough fill deposit. Overlying the sand unconformably is 4.1 m of colluvium, which extends to the top of the section.

iii) HH 62-148(81-6c)

HH 62-148(81-6c) is a complex of overbank, back-channel, bar-slope and slough deposits situated on the south bank of the Peel River (Figure 2, in pocket). The basal unit exposed on 10 August 1981 was a silt (12% sand, 24% clay), horizontally laminated, with fine organic detritus along the laminations (Figure 59). This unit is interpreted as a back-channel fill or slough fill deposit. A sample from this unit contained insufficient palynomorphs for counting. The pollen from 12 taxa were observed, including Picea, Betula, Populus, Larix, Salix, Alnus type crispa, Cyperaceae, Gramineae, Linnaea borealis, Cruciferae, Saxifraga, and Viola. As well, Sphagnum spores were noted. The absence of Ericaceae grains may be due to their tendency to be concentrated hydrodynamically with the sand fraction. No diatoms were present in the sample.

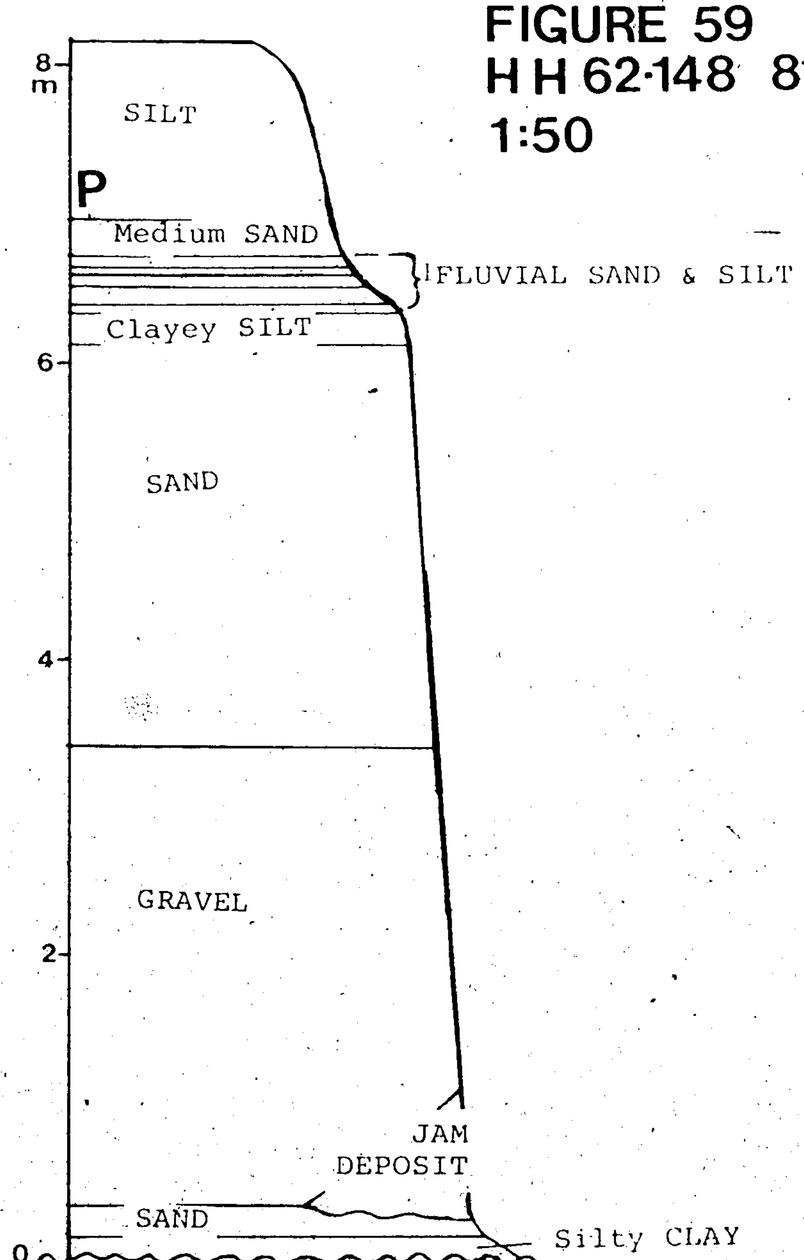
Figure 59

HH 62 148 81-06

The section is located along the eastern bank of the Peel River (66°40'N, 134°32'W). The base of the section is 25m ASL. Fluvial sediments exposed at this locality contain palynomorphs. A single sample contained an adequate number of palynomorphs for counting and interpretation. See text for further discussion.

P- location of palynomorph sample.

FIGURE 59
H H 62·148 81·6C
1:50



Abruptly overlying this unit is a 21 cm thick fine-grained sand stratum, with horizontal laminations and indistinct eastward-trending planar cross-laminations. This unit contained several small wood lenses, interpreted as marginal jam deposits, and a larger accumulation of branches, twigs, and organic detritus. interpreted as a bar head jam deposit. This jam extends 6 m laterally along the outcrop, reaches a maximum thickness of 90 cm, and extends at least 1 m into the face of the exposure. The wood fragments are predominantly Picea, although Salix, Alnus, and Betula are also present. The largest fragment observed was 130 cm long and 12 cm in diameter. The larger branches and trunks are coated with iron oxide and iron carbonate. A sample of organic detritus from the bar-head deposit was analysed for palynomorphs. Although Picea, Betula, Alnus, Ledum, Cruciferae, Chenopodium, and Cyperaceae grains were noted, the concentrations proved insufficient to warrant counting.

Overlying the sand along an abrupt contact is a pebble gravel with a medium to fine sand matrix. The unit is 3.1 m thick(Figure 59). The pebbles are imbricated, indicating eastwardly flow.

This gravel is overlain along a gradational contact by 2.7 m of medium-grained sand, with interbeds of medium pebbles-granules. The sand contains planar cross-laminations of varying orientations and horizontal laminations. Pebbles within the interbeds are imbricated in a variety of orientations. A sample from this unit contained no palynomorphs.

Overlying this sand abruptly is a very finely laminated sandy silt (30% sand, 18% clay) 20 cm thick. Fine organic detritus was present along the lamination planes. The stratum contained a few grains of Picea and Cyperaceae. The silt is in turn overlain by 3 cm of structureless medium-grained sand and 12 cm of sand-silt-clay (26% sand, 50% silt, 24% clay). The sand-silt-clay unit was sampled and possessed a small number of Picea grains and a single Vaccinium pollen.

This unit is abruptly overlain by 8 cm of structureless fine-grained sand (19% silt) which is devoid of palynomorphs. The sand grades upwards into 4.5 cm of finely laminated silt. This sand-silt couplet is overlain unconformably by a second couplet 7 cm thick.

The overlying unit is a coarse-medium grained sand, 25 cm thick, with horizontal laminations defined by accumulations of fine organic detritus. Contained within the sand spectrum is a wood deposit, formed along the margins of the channel. The sand contained Picea, Betula, Cyperaceae, Chenopodium, Ledum, and Arctostaphylous uva-ursi palynomorphs.

Overlying the sand is a structureless silt (18% clay) 1.2 m thick. This unit contains many randomly oriented small accumulations of Picea, Alnus, and Betula fragments. The silt is interpreted as a bar slope and veneer deposit formed during the waning stages of the flooding of a previously vegetated island. The wood accumulations are classified as bar-flank jam deposits. A palynological sample obtained from the silt was dominated by Cyperaceae (Table 18). The arboreal fraction was predominantly Picea (both types), with associated Betula and minor Larix. Salix, Alnus type crispa, A. type incana and Populus were present in small amounts. The shrub fraction is relatively minor, and is dominated by the Ericaceae genera and Empetrum nigrum. Herbs, with the exception of Cyperaceae, are present only in very low concentrations. Equisetum spores are also present. The assemblage also contains small amounts of Myriophyllum, as well as Sphagnum spores. No pre-Quaternary grains were present.

Although the number of grains in all but one of the samples was insufficient for counting, the qualitative data suggest that the palynomorphs have undergone hydrodynamic sorting, similar to that documented for sections along the Caribou River (Catto 1985). This depositional pattern, combined with the paucity of grains, limits the interpretations possible. The assemblage in the upper silt unit (Table 18) is interpreted to represent a deposit formed by flushing of back-channels and sloughs during flooding. The sediments and palynomorphs transported by the flood waters were deposited, along with some locally-derived pollen, when the spring flood was ebbing. The assemblage probably occupied the margins and floor of a temporary

Table 18

Palynological counts, Quaternary taxa, BH 62-148 (81), silt unit (sample 34)

Taxon	Percentage
<u>Picea</u>	26.9
<u>Betula</u>	2.5
<u>Salix</u>	0.[illegible]
<u>Populus</u>	0.4
<u>Alnus</u> type <u>crispa</u>	2.2
<u>Alnus</u> type <u>incana</u>	0.4
<u>Larix</u>	0.7
<u>Empetrum nigrum</u>	1.4
<u>Ledum</u>	2.9
<u>Vaccinium</u>	1.6
<u>Arctostaphylous rubra/ alpina</u>	0.3
<u>Cassiope</u> type <u>tetragona</u>	0.4
<u>Andromeda</u> type <u>Polifolia</u>	0.3
<u>Myrica Gale</u>	0.1
Cyperaceae	53.3
Gramineae	0.7
<u>Artemisia</u>	0.1
<u>Viola</u>	0.1
<u>Dryas</u>	0.1
Cruciferae	1.0
<u>Saxifraga</u>	0.1
<u>Rubus chamaemorus</u>	0.1
<u>Cystopteris</u>	0.8

Table 18 (continued)

Taxon	Percentage
Equisetum	0.5
Sphagnum	2.3
Myriophyllum	0.4

abandoned back-channel. The area was still moist to wet, as indicated by the dominance of Cyperaceae, and the presence of the Ericaceae genera, Viola, Rubus chamaemorus, and Myrica Gale. Dry ground taxa, such as Artemisia, are present in very low concentrations.

The assemblage is contained in fluvial deposits within the modern floodplain of the Peel River. It is therefore considered to be Late Holocene, in the absence of any chronological data.

iv) HH 72-50 (82)

Holocene fluvial sediments (unit #8) exposed at HH 72-50 (82)(Figure 34) contain assemblages of palynomorphs segregated by fluvial activity (Figure 38). The palynology of these sediments has been discussed previously (Catto 1985), and a brief summary is presented here.

The sediments of unit #8 are treated as a single pollen zone, B, which is subdivided into six subzones (B1 to B6), each representing a distinct lithological and palynological entity. In all subzones, the dominant arboreal taxon is Picea. Larix is present in minor amounts in two samples. The arboreal-shrub component is dominated by Betula, with associated Alnus, type crispa and minor Salix pollen. There is a distinct difference between the proportions of arboreal and arboreal-shrub pollen present in the sand sub-units and the silt strata (Table 19). The mean for the three sand samples is 47.6%, compared with a value of 63.0% for the five silt samples. The individual taxa Picea, Betula, and

Table 19

Palynomorph distribution, upper strata, HH 72-50 (82).

A) Sand strata (6 samples)

Taxon	Minimum	Maximum	Mean
	%	%	%
Picea	44.7	49.2	46.5
Betula	0.4	1.3	0.7
Alnus crispa	0.0	0.2	0.1
Populus	0.0	0.0	0.0
Empetrum nigrum	0.8	1.9	1.5
Ledum	2.6	3.0	2.9
Vaccinium	1.1	1.6	1.4
Gramineae	0.3	0.4	0.3
Cyperacaea	19.3	33.6	28.3
Chenopodium	1.1	2.2	1.5
Lycopodium	3.1	9.5	6.1
Total grains counted	387	482	426

Table 19 (continued)

B)Coarse silt strata (transitional between sand and medium silt strata, 2 samples).

Taxon	Minimum	Maximum	Mean
Picea	48.8	54.6	51.[illegible]
Betula	2.0	2.4	2.2
Alnus crispa	1.3	1.3	1.3
Populus	0.1	0.4	0.3
Empetrum nigrum	1.3	1.5	1.4
Ledum	2.4	2.6	2.5
Vaccinium	1.0	1.5	1.3
Gramineae	1.5	2.5	2.0
Cyperacaea	20.0	29.5	24.8
Chenopodium	0.1	0.1	0.1
Lycopodium	0.6	1.0	0.8
Total grains counted	390	463	427

Table 19 (continued)

C)Medium silt strata (8 samples)

Taxon	Minimum	Maximum	Mean
	%	%	%
Picea	55.0	64.2	59.6
Betula	2.0	5.2	4.1
Alnus crispa	0.1	1.9	1.3
Populus	0.0	0.6	0.3
Empetrum nigrum	0.6	1.6	1.2
Ledum	1.5	2.3	1.9
Vaccinium	0.9	1.3	1.1
Gramineae	1.5	3.0	2.4
Cyperacaea	13.8	20.2	17.4
Chenopodium	0.1	0.1	0.1
Lycopodium'	0.4	6.0	1.7'
Total grains counted	417	1246	682

'Modal value 0.6%

Alnus all show similar variations. Shrubs form a lesser component of the total assemblage (3.2-7.2%). Ericaceae dominate the assemblage, along with Empetrum nigrum and Myrica Gale. The sand units contain a slightly higher proportion of the shrub palynomorphs than do the silt units (mean 6.5% versus 5.1%).

Herbaceous taxa are dominated by Cyperaceae grains, with minor amounts of Gramineae, Artemisia, Chenopodium, Cerastrium, Cruciferae, Dryas, Epilobium, and Arnica. The concentration of Cyperaceae grains is greater in the sand sub-units (mean 28.3%) than in the silts (mean 19.8%). A similar distribution pattern characterises Chenopodium. Gramineae pollen, however, is more concentrated in the silt layers (Table 19). The only aquatic pollen present is from Myriophyllum.

Pteridophyte spores are dominated by Lycopodium, with lesser Cystoperis. The total pteridophyte fraction varies between 3.0% and 12.4%. Lycopodium is preferentially concentrated in the sands (6.1% vs. 1.7%), while Cystoperis is preferentially contained in the silt strata (3.0% vs. 1.4% in sands). Sphagnum spores vary irregularly between 2.3% and 4.7%.

The sand and silt strata thus show two different palynological spectra (Table 19). Picea, Betula, Alnus type crispa, and Gramineae are more concentrated in the silt units. In contrast, the sand strata contain higher concentrations of Ericaceae, Empetrum nigrum, Cyperaceae, Chenopodium and Lycopodium, lesser amounts of Picea, and very little Betula, Alnus and Gramineae.

Two considerations are potentially involved in analysing the distribution of the pollen. The grains may have been sorted and segregated during fluvial transport and deposition, or the differences may reflect botanical and ecological factors.

If the genera produced pollen at different times, a segregated distribution could develop in a fluvial succession. In this instance, the coarser sediments deposited during the height of the spring flood season would be enriched in the early pollenating species, while the silt units would be dominated by summer pollen producers. Palynomorph distributions of this kind have been observed in seasonally-laminated lacustrine sediments (Terasmae 1963; Tippett 1964). Thus, if phenology was a critical factor, the sand units at HH 72-50 would be enriched in Alnus, Betula and Picea, the first genera to pollenate in the region (Ritchie 1977), and would be deficient relative to the silts in the late-season pollenators Artemisia, Ledum, and Vaccinium. Many of the Cyperaceae species are also late-season pollenators. The actual distribution in the sediments, however, is the reverse of this theoretical model. It is apparent that phenological differences cannot explain the distribution pattern observed.

Since the sandy units were deposited relatively rapidly during periods of moderate flow, the palynological assemblage should be less diverse and more local in nature. This conclusion is supported by the relative importance of Cyperaceae in the sand spectra, and the lesser number of species distant from the immediate braided stream chute channel environment, such as Picea glauca. The low concentrations of Alnus crispa and Gramineae, and the importance of Ledum, Vaccinium, and Lycopodium, however, are not in accord with this model. The presence of these upland acidic bog and tundra genera in the assemblages indicates that the physical processes of transport and deposition are dominantly responsible for the observed distribution of palynomorphs.

The absence of sharp or erosional contacts within the gravel/sand/silt cycles suggests that sedimentation was continuous, proceeding gradually from traction load lodgment to settling of suspension load through still water. This succession of depositional modes can account for the distribution of pollen grains within the sequence.

Brush and Brush (1972) and Brush and DeFries (1981) have investigated the transportation of pollen by water in laboratory flumes and estuaries, respectively. These studies indicate that Betula grains remain in suspension for long periods, and hence tend to be concentrated in the suspended load. Conifer grains are initially effective floaters, but tend to be incorporated into the bedload after their surfaces are wetted during long periods of transport. Alnus pollen were preferentially concentrated in the bottomset beds produced in the flume (Brush and Brush, 1972), while Artemisia and Typha were concentrated in the topset beds and Chenopodium in the foreset beds. Conifer and Betula grains were preferentially deposited in the topset and bottomset beds.

The results of the study of HH 72-50 (82) concur with the experimental data of Brush and Brush (1972). Picea, Betula, and Alnus crispa are concentrated in the silt beds produced by the settling of the suspended load. The sand strata, produced from mixed bedload and suspended load, contain more Chenopodium. The depletion in the concentration of Picea in the sands is reflected in the percentage increase in Cyperaceae, Ledum, Vaccinium, and Empetrum nigrum.

A single grain of Picea type glauca, with an equivalent spherical diameter of 75×10^{-3}mm, when fully wetted, is calculated to fall at a rate of approximately 5×10^{-1}mm/s, using the equation of Baba and Komar (1981). The small fall velocity would be achieved by a quartz grain with an equivalent spherical diameter of 25×10^{-3}mm, (5.3 phi). Picea glauca grains would therefore be expected to be concentrated in the coarse to medium silt fraction of a deposit produced by settling, which is in agreement with the data from HH 72-50 (82).

Single grains of Betula and Alnus type crispa will fall at the rates of 1.5×10^{-1}mm/s and 4.0×10^{-2}mm/s respectively, velocities equivalent to those of spherical quartz grains between 1.4×10^{-2} mm and 0.5×10^{-2} mm diameter (6.1 to 7.5 phi). These pollen grains would therefore be expected to be deposited along with medium to fine silt. If the grains of these genera were transported as tetrads, the fall velocities would be increased, and the grains would settle along with the coarse to medium silt and Picea pollen. Alternatively, the pollen could have been deposited along with adhering fine silt, clay and colloidal matter. Such aggregations deviate markedly from Stokes' Law behaviour during settling (Chase 1979). Since the concentration of tetrad arboreal-shrub grains noted in the spectra was very low, the second explanation is preferred.

The apparent percentage concentration in the sand sub-units of Cyperaceae, Ledum, Vaccinium, and Empetrum nigrum are largely functions of the fluctuations of the number of Picea grains, rather than being due to large changes in the numbers of these palynomorphs. The number of Cyperaceae species, their variation in phenology, and the proximity of suitable botanical environments to the depositional site suggest that the site could conceivably have received Cyperaceae pollen throughout the sedimentation period, making segregation induced by settling unlikely.

The calculated fall velocities of the Ericaceae tetrads are similar to those of single Picea type glauca grains, suggesting that these grains would be deposited with coarse to medium silt if settling after a single seasonal influx was the dominant mode of depositon. The observed distribution indicates that a substantial proportion of the tetrads was deposited from the bedload portion of the transported sediment. These upland grains may have been initially transported to the stream by snow melt runoff. The tetrad grains of _Typha_, investigated by Brush and Brush (1972b), differ from Ericaceae pollen in their density and shape. _Typha_ pollen would be expected to remain in suspension or floating longer than would terrestrial Ericaceae or _Empetrum nigrum_ tetrads.

Within the bedload zone, higher energy levels and abundant clast-pollen collisions would act to reduce the concentration of fragile, thin-walled palynomorphs, and thus accentuate the importance of thick-walled, robust grains in the pollen assemblage. This accounts for the higher percentages of the robust spores of _Lycopodium_ and the lesser proportions of the fragile grains Gramineae and _Populus_ in the sand strata.

Zone B can be interpreted as a mixed deciduous/evergreen forest community. _Picea_ type _glauca_ dominates the palynomorph assemblages, suggesting that the community had developed on a stable floodplain/bar complex, with no disruption due to flooding. The deciduous portion of the forest, composed largely of _Betula_ and _Alnus_, also indicates a mature and stable assemblage, although the absence of _Populus_ pollen is not conclusive due to its fragile nature. The scarcity of open ground is indicated by the low proportions of Gramineae.

The understory vegetation contained Shepherdia canadensis, Empetrum nigrum, Cerastium, Arnica, Epilobium, and Anemone in minor amounts, all indicative of coarse-textured sediments. The principal taxa associated with the mixed forest, Ledum and Vaccinium, are also present. The surface of the floodplain was probably undulatory, with numerous wet slough areas and abandoned channels which served as habitats for moisture-preferring taxa such as Aster, Myrica Gale, Andromeda type Polifolia, Cyperaceae, and Ranunculus .

The valley-marginal vegetation, presumably consisting of Artemisia and snowbank species such as Cassiope tetragona, is poorly represented. Myriophyllum, which prefers slow moving or stagnant streams and ponds, was probably transported from these areas during high flow events which flushed out the relatively inactive channels. The pollen of this genus is well adapted for water transport.

The climate suggested by zone B is similar to that currently prevailing on the Peel Plateau (Burns, 1973; Ritchie, 1984). The only species in the spectrum which are not probably extant are Cornus canadensis and Salicornia rubra (Porsild and Cody, 1980). The presence of pollen of Cornus at the site could result either from a slight climatic amelioration or aeolian transport. Salicornia rubra is an unlikely consituent of a northeast Beringian assemblage developed on non-saline fluvial gravel and sand;

and it is therefore assumed that the single grain present in each of two samples was transported from a distal provenance by wind.

No ^{14}C dates were obtained from the fluvial sediments at HH 72-50 (82). The base of the overlying peat unit has been dated at 2290±50 B.P. (GSC-3690).

Assemblages in Autochthonous Peat and Lacustrine Sediments

i) HH 72-49 (82)

This section is located on the south bank of the Caribou River, 50 km southwest of the confluence of the Caribou and Peel Rivers, Yukon (Figure 2).

The base of the section is composed of Mississippian shale (Norris 1981). This shale is overlain unconformably by gravel, 8.6 m thick. The gravel is overlain along an erosional contact by till, correlated to till exposures throughout the Caribou River area (Figure 40).

Overlying the till is a peat deposit, 132 cm thick. The degree of decomposition varies from 2 (Von Post Scale) at the surface to 7 at the base. The upper portion of the unit contains abundant <u>Picea</u> cones and needles. This peat deposit was designated as unit #4, and was thoroughly analysed for palynomorphs.

A total of 28 samples were processed. Samples 1 through 14 were taken at 10 cm intervals, beginning 4 cm below the surface (#1) to the base at 132 cm (#14). An additional sample, #15, was obtained from a clay band 2 cm thick 72 cm below the surface. A second series of samples, numbered #1a through #13a, were taken at 10 cm intervals, beginning 10 cm below the surface (#1a), to 130 cm below the surface (#13a). Thus, the greatest vertical separation between samples is 6 cm. The unit is subdivided into 5 pollen zones, lettered A through E in chronological order (Figure 60, in pocket).

The basal 6 assemblages are included in pollen zone A. The arboreal component of this zone is minor, and is dominated by Picea. Arboreal-shrub taxa are dominated by Betula (shrub type) pollen, with minor Salix and Populus. These three taxa comprise 21.6-30.2% of the assemblage, gradually increasing in proportion with decreasing age. Shrubs are dominated by Empetrum nigrum, Ledum, Vaccinium, and Myrica Gale. Shepherdia canadensis is present in the older samples, and the Ericaceae pollen content increases with decreasing sample age. Herbaceous taxa grains are dominated by Gramineae and Cyperaceae. These two families comprise between 40.3% (#14) and 29.4% (#11a) of the spectra, decreasing with decreasing age. Other herbs present include Artemisia, Arnica, Epilobium, Polemonium, and Sagina. Pteridophyta are dominated by Cystopteris and Lycopodium, and Sphagnum is the only bryophyte present.

The palynomorph assemblage suggests an upland environment adjacent to or beyond the treeline. The vegetation community envisaged contains elements of the Seasonal Orthophyll Meadow, Open Deciduous Orthophyll Scrub, and Seasonal Grass/Herb Steppe communities of Hettinger et al (1975). Seasonal orthophyll meadows are found in poorly drained areas, and are dominated by Cyperaceae and shrub Betula. These taxa also dominate orthophyll scrub communities, found primarily on soliflucted or similarly disturbed areas. Such communities also contain Ledum and Vaccinium. Both these assemblages indicate wet to moist conditions, in an unstable to semi-stable area, under colder climatic conditions than currently prevail. The presence of these Richardson Mountain assemblages on the Peel Plateau, together with the absence of Alnus, indicate a relatively cold climate.

The low proportions of Picea (chiefly mariana type) can be interpreted in two ways. Either the grains were derived from distal provenances through aeolian transport, or the region supported scattered stands of Picea mariana, possibly vegetatively reproducing in many instances. The few P. glauca type grains present (3-12%) were probably transported from distant regions. The presence of Myrica Gale, Linnaea borealis, Saxifraga, Lupinus, Polemonium, Aster and Cystopteris all indicate a very moist environment, such as a wet boggy unstable area or a tundra pond shoreline. The palynomorphs considered above, therefore, suggest an upland environment characterised by abundant surface or near-surface water, possibly an area adjacent to a thermokarst pond.

A significant component of the assemblage, however, suggests that dry conditions prevailed in some areas. Taxa such as *Chenopodium*, *Shepherdia canadensis*, *Sagina*, *Silene*, *Epilobium*, *Artemisia* and *Arnica* require dry terrain and prefer sands and coarser sediments. The same requirements are exhibited by many Gramineae species. This assemblage could have inhabited unstable bedrock slopes in the vicinity, filling the niche currently occupied by the Seasonal Grass and Herb Steppe communities in the Richardson Mountains. The assemblage differs from this modern analogue, however, since it contains shrub taxa such as *Shepherdia canadensis*. The assemblage may also have occupied exposed river bars and sand flats.

Diatom analysis revealed fragments of *Navicula* and *Fragilaria*. These taxa provide no critical palaeoenvironmental information.

It is apparent that no modern vegetation assemblage adequately mirrors the spectra of zone A. The terrain consisted of an unstable Cyperaceae wetland bordering a pond adjacent to the Caribou River, surrounded by sparsely vegetated slopes dominated by Gramineae and other herbs with scattered shrubs. The climate was colder than at present. The degree of representation of the slope communities suggests that the overall pollen production and amount of vegetation cover was sufficiently low to prevent the masking of the slope-derived palynomorphs by the *Betula*-Cyperaceae communities immediately surrounding the depositional site.

The ^{14}C date of 12400±120 B.P. (GSC-3691) obtained from sample #14 supports the hypothesis that the site was under the influence of a relatively rigourous climate during the formation of zone A. The presence of small amounts of *Picea* pollen and the absence of *Alnus* pollen are both compatible with the migration patterns outlined by Hopkins *et al* (1981).

The spectra of zone B are dominated by arboreal-shrub and herbaceous pollen (Figure 60, in pocket). The arboreal component decreases from 18.0% (#11) to 8.6% (#9), with *Picea* the dominant taxon. The arboreal-shrub component, primarily *Betula*, varies from 25.4% to 30.7% of the total spectrum, and genrally increases as the arboreal component decreases. Shrubs, dominated by *Ledum*, *Vaccinium*, *Empetrum*, and *Myrica Gale*, tend to decrease in proportion with decreasing sample age, although the decline is relatively low (13.2% to 9.6%). Herbaceous taxa are dominated by Cyperaceae and Gramineae grains, with *Artemisia* sub-dominant. The herbaceous fraction varies between 18.9% and 30.0%, the variations mirroring the fluctuations in the concentration of Cyperaceae pollen. Gramineae and *Artemisia* remain relatively constant throughout zone B. Pteridophyta are dominated by *Lycopodium* spores, and *Sphagnum* is the only bryophyte present.

Plant macrofossils observed in sample #10 by J.V. Matthews Jr. (Geological Survey of Canada, unpublished macrofossil report 83-27), were characterised by the aquatic taxa *Potamogeton*, *Myriophyllum* and *Hippuris*. *Betula* type *glaudulosa* seeds and *Picea* needles were also present. The arthropod component provides little information about the regional environment (unpublished arthropod report 83-20). The ostracod Cyclocypris ampla was present in samples #11, 10a, 10, and 9. This species has previously been reported from an early Holocene marl west of Fort McPherson (Delorme et al, 1977).

The assemblage suggests an environment characterised by warmer temperatures than zone A, with more extensive cover by deciduous shrubs (Betula type glandulosa), scattered Picea mariana and rare Picea glauca, and a closed shrub understory dominated by Ledum and Vaccinium, with associated Empetrum nigrum, Arctostaphylous rubra/alpina, and Rubus chamaemorus. Wet depressions and the margins of the pond supported a community dominated by Cyperaceae, with associated Sphagnum, Myrica Gale, Linnaea borealis, Aster, Saxifraga, Polemonium and Cystopteris. The relative palynomorph influx from the adjacent dry, open upland slopes and plateaux declined, as indicated by the decrease in Artemisia, Arnica, Sagina, and Chenopodium, and the absence of Silene. This change probably reflects the increased productivity of the lowland community, as compared to the more consistent pollen production rates of the upland assemblage. The major successional change evident in the zone is the gradual increase of Betula (both arboreal and shrub) and the corresponding decreases in Picea type mariana and Sphagnum palynomorph concentrations. These fluctuations may indicate a decline in the local water table (or permafrost table). The community depicted above resembles that of the Open Evergreen Sclerophyll Deciduous Forest, as described by Hettinger et al (1973), with some exceptions. Alnus type crispa is still a very minor component of zone B, whereas it is an important and sometimes dominant constituent in the modern community. This contrast probably is due to the delayed migration of Alnus crispa to the Peel Plateau's western areas after deglaciation. The distribution pattern suggested by Hopkins et al (1981) and Ritchie (1984) indicates that Alnus crispa was a relative latecomer to the region after deglaciation.

The seven samples obtained between 54 cm and 80 cm below the surface are considered as palynological zone C (Figure 60). The arboreal component of this zone (mainly Picea mariana) varies from 1.2% to 6.8%, with the exception of the somewhat anomalous sample #15. The arboreal-shrub component, dominantly Betula type shrub with associated arboreal Betula and Salix, fluctuates from 38.2% to 45.5%, and is one of the major components of the assemblage. Shrubs form a minor component of the spectrum, Myrica Gale, Ledum, and Empetrum nigrum being the principal taxa.

Herbaceous taxa are the second major component of the spectrum. Gramineae values are high (15.7% to 18.6%), with the exception of sample #15. Cyperaceae is also an important constituent. Lesser amounts of Artemisia, Sagina, Cruciferae, Saxifraga, Rubus chamaemorus, Epilobium, Polemonium, Arnica and Aster are also present. Pteridophyta form a minor component of the assemblage, and are dominated by Cystopteris. Sample #15 is anomalous with respect to these taxa as well, as it posseses a Lycopodium content of 14.6%. Minor amounts of aquatic taxa (Potamogeton and Myriophyllum) and the bryophyte Sphagnum are also present.

The environmental conditions indicated by this assemblage are similar to those prevailing during the deposition of Zone B. The two zones differ in that the proportion of arboreal Betula has increased, the amount of understory vegetation has declined (especially the Ericaceae), and the proportion of Gramineae has increased considerably. The assemblage suggests an open deciduous forest dominated by shrub Betula, Alnus crispa, Picea mariana, and Larix and Salix trees and/or shrubs, with open areas colonized by Gramineae with associated Epilobium, and wet depressions inhabited by Cyperaceae with associated Myrica Gale, Rubus chamaemorus, Polemonium, and Cystopteris. The dry upland slope component, dominated by Artemisia with Arnica, Shepherdia canadensis, Chenopodium, Sagina, Silene, and Linum Lewisii, appears to be slightly more prominent in Zone C than in Zone B.

The assemblage reflects the modern open deciduous/evergreen forests which develop on recently burned land, the chief difference being the low proportion of Alnus crispa. Fires act to remove the conifers and permit the expansion of arboreal-shrub genera, as well as shade-intolerant taxa such as Epilobium. By removing the closed-shrub understory, fires also increase soil instability, leading to an increase in the amount of mass movement and thus producing undulating surfaces with many wet depressions.

Sample #15, as mentioned above, is anomalous. This sample was taken from a poorly sorted clay layer 2 cm thick, 72 cm below the surface. Several of the pollen grains in the sample were present as aggregated clusters, especially *Picea*, and the diversity of the assemblage is limited compared to the adjacent samples. The characteristics, together with the sedimentology of the deposit, suggest that the clay layer was formed as a result of bank caving and subsequent flowage and/or sliding of the clay during the pollenation season, probably during the spring thaw. Aggregated clusters of grains usually indicate that anthers of the plants concerned have been incorporated into the sediment. The absence of diatoms in Zone C suggests that the pond may have been ephemeral at this time. The bank caving was the result of differential thermally-generated subsidence of the surrounding peat. Such deposits would contain a palynological record biased in favour of the taxa fortuitously present in the small area immediately adjacent to the pond. Although the phenological differences are not great between the taxa of the region, the record preserved in an ephemeral pond or bank cave sediment would accentuate any differences which did not exist by preserving an essentially momentary distribution.

The presence of shallow pond deposits is a further indication that the area may have been burned. Burning would remove the shrub understory cover and accentuate the development of undulatory surfaces, with shallow ponds and moist depressions. It could also promote bank caving of pre-existing ponds by destroying the root systems of the understory vegetation.

The samples between 10 cm and 50 cm below the surface are considered as palynological zone D (Figure 60). The arboreal component generally increases with decreasing age of the samples. Picea type mariana is dominant, with associated P. type glauca, Larix, and Juniperus. The arboreal-shrub component is chiefly composed of Betula shrubs, with associated arboreal Betula and Salix, Populus and Alnus type crispa. It varies in proportion from 30.0% to 43.8%. The shrub component is relatively minor, and is predominantly Myrica, Empetrum, Ledum, Vaccinium and Arctostaphylous rubra/alpina pollen. Herbaceous pollen forms a major component of the assemblage, primarily Cyperaceae and Gramineae. The proportions of Cyperaceae remain relatively stable, whereas Gramineae declines in proportion to the increase in arboreal taxa. A wide variety of herbaceous pollen is present, including Artemisia, Sagina, Saxifraga, Rubus chamaemorus, Lupinus, Viola, Polemonium, Epilobium, Arnica and Cruciferae. The proportions of Pteridophyta fluctuate irregularly throughout zone D, with Cystopteris and Lycopodium being co-dominant. Aquatic taxa and bryophytes are represented by small amounts of *Potamogeton*, *Myriophyllum* and *Sphagnum* palynomorphs.

Plant macrofossils observed by J.V. Matthews Jr. (Report 83-29) include *Picea* needles, *Betula* seeds, and the seeds of the aquatic and shoreline taxa *Potamogeton*, *Carex*, *Ranunculus*, and *Hippuris*. Insect fossils examined by J.V. Matthews Jr. provided little palaeoenvironmental information (Unpublished fossil arthropod report 83-22, Geological Survey of Canada).

Sample #3 has been radiocarbon dated at 7370± 80 B.P. (GSC-3822). The small proportion of *Alnus* type *crispa* grains may be derived from isolated stands, such as the Twin Lakes site discussed by M. Kuc (in Fyles *et al*, 1972), or may reflect long-distance aeolian transport.

Zone D represents the re-establishment of the open evergreen sclerophyll deciduous forest over the formerly burned terrain. The open spaces occupied by Gramineae gradually declined in the area as *Picea* and *Betula* groves expanded. The understory vegetation developed into a closed shrub cover, dominated by *Ledum*, *Vaccinium*, and *Arctostaphylous rubra/alpina*. Moister depressions were occupied by *Rubus chamaemorous*, *Myrica Gale*, *Cassiope tetragona*, *Cornus canadensis*, *Saxifraga*, *Lupinus*, *Polemonium* and *Cystoperis*, while the drier areas supported *Shepherdia canadensis*, *Sagina*, *Artemisia*, *Arnica*, and *Equistum*. The inreasing maturity of the forest through time is further indicated by the gradual replacement of shrub *Betula* and

Picea mariana by arboreal *Betula*, *Picea glauca* and possibly *Alnus crispa* evident in the uppermost samples. The influx of palynomorphs from the upland vegetation decreased in proportion, probably due to increased vegetation cover, and pollen production by the local taxa. The climatic conditions apparently did not vary sufficiently to alter the basic successional colonization pattern depicted by the palynological spectra of zones C and D.

Palynomorphs counts for the two uppermost samples are combined as zone E (Figure 60). The dominant component of the spectra is the arboreal fraction, increasing upwards from 39.7% to 47.5%. *Picea* type *glauca* is the most abundant palynomorph, comprising approximately 75% of all *Picea* grains. The amounts of *Betula* tree and shrub type pollen are approximately equal, and *Alnus* type *crispa* is the sub-dominant arboreal-shrub taxon. The shrub component is predominantly *Ledum* and *Vaccinium*, with *Arctostaphylous rubra/alpina* and associated *Empetrum nigrum* and minor *Myrica Gale*. The herb assemblage is less diverse than those of zones C and D, with little *Artemisia* and *Arnica* being present. Pteridophyta, aquatic taxa, and *Sphagnum* palynomorphs are present only in minor amounts. Diatoms were not detected.

Zone E represents the establishment of an open evergreen/deciduous forest on the site. The understory vegetation, dominated by *Ledum*, *Vaccinium* and *Arctostaphylous rubra/alpina*, and the arboreal vegetation

(*Picea glauca*) are typical of those found in modern communities. Wetter areas were occupied by Cyperaceae and other moisture-preferring taxa. The community occupying the site presently is very similar to that represented by the spectra of zone E.

The similarity of the modern assemblage represented by zone E to the modern vegetation suggests that a considerable time elapsed between the deposition of the upper portion of zone D (^{14}C dated at 7370±80 B.P.) and zone E. The thermokarst depression may have been completely filled in during the interval between the two zones, although no vertical discontinuity in the sediment was observed.

ii) HH 72-50 (82)

The youngest unit (#9) exposed at HH 72-50 (82) is a fibric peat 36 cm thick (Figure 34). Three samples were processed for palynomorphs, the lowermost of which (#17) has been ^{14}C dated at 2290±50 B.P. (GSC-3690). The palynological assemblages are depicted in Figure 38 (in pocket), and are combined as a single pollen zone, C.

The dominant arboreal taxon is *Picea*, which increases regularly in pollen concentration from 12.9% in the basal sample to 39.4% in the uppermost sample. *Larix* is present in minor amounts. The arboreal-shrub component is predominantly *Betula* and *Alnus* type *crispa*, with minor *Alnus* type *incana*, *Salix* and *Populus*. Shrubs form a lesser but significant component of the assemblage, increasing from 6.0% at the base of the zone to 18.8% in the uppermost sample. *Ledum*, Vactinium, Empetrum nigrum, and Myrica Gale dominate, with minor amounts of Cassiope type tetragona, Arctostaphylous, Andromeda type Polifolia, Pyrola and Shepherdia canadensis also present. Herbs, especially Cyperaceae, are an important constituent. The concentration of Cyperaceae varies inversely with the arboreal, arboreal-shrub, and shrub components, from a maximum of 58.2% at the base of the zone to a minimum of 21.0% in the uppermost sample. Gramineae, Silene, Cruciferae, Rosaceae, Epilobium, Artemisia, Arnica, Sausseria, and Viola grains were also recorded. Pteridophyta and aquatic taxa are minor components of the spectrum. The former are dominated by Cystopteris, while Myriophyllium is the only aquatic taxon present. The sole bryophyte represented is Sphagnum, which declines in spore concentration from 10.1% at the base to only 0.8% in the uppermost sample.

Plant macrofossils from sample #18 were examined by J.V. Matthews Jr (Unpublished plant macrofossil report 83-35, Geological Survey of Canada). Carex seeds, Picea needles, and fragments of Sphagnum were noted, as well as insect fragments of the order Hymenoptera.

Zone C can be interpreted as a vegetative assemblage produced during the progressive drying of a Cyperaceae (Carex?) bog, suggesting a decline in the water table and/or permafrost degradation. Initially the bog was extremely wet, with Cyperaceae dominant and Sphagnum sub-dominant. The Picea present in this sample was mainly type mariana, and the understory vegetation contained moisture-preferring plants such as Myrica Gale, along with limited amounts of Ledum and Vaccinium. As the bog dried, the diversity of the vegetation increased. Picea glauca began to occupy the area, and P. mariana, Alnus crispa, Ledum, Vaccinium and arboreal Betula all increased in abundance. Betula shrub cover decreased at this time, possibly due to competition from Alnus shrubs.

The area remained moist, as shown by the continued importance of Picea mariana, the presence of Cyperaceae, Ledum, Vaccinium, Myrica Gale, Cassiope type tetragona, Arctostaphylous rubra/alpina, Andromeda type Polifolia and Viola pollen, and the low concentrations of Gramineae and Artemisia grains. Continued decline of the soil moisture content led to conditions suitable for forming the mixed Deciduous/Evergreen forest seen at HH 72-50 (82) today.

The base of zone C (sample #17) has been ^{14}C dated at 2290±50 B.P. (GSC-3690), suggesting that organic accumulation at this site continued throughout the Late Holocene. This conclusion is not in complete accordance with the data of Zoltai and Tarnocai (1975). These authors suggested that Sphagnum peat accumulation at Mackenzie Valley sites ceased 2710-2650 B.P., and that extensive permafrost development began at this time. At HH 72-50 (82), permafrost degradation occurred prior to 2290±50 B.P., forming the initial pond fringed by the Cyperaceae-dominated assemblage of sample #17. This anomalous, late-developing basin was probably caused by local permafrost degradation induced by the the nearby Caribou River.

iii) HH 62-144 (81)

This section is located along the east bank of the Peel River, 3 km upstream from the Peel and Caribou Rivers (Figure 2, in pocket). The basal part of the section consists of fluvial sediments overlain by till (Figure 40), as discussed previously.

The till is overlain by 12 cm of peat containing fragments of *Picea*, *Betula*, and *Alnus* . A sample obtained from this sediment was dominated by *Picea* type *glauca*, with associated Cyperaceae, *Betula*, *Alnus* type *crispa*, *Ledum*, and *Vaccinium* and lesser amounts of Gramineae, *Larix*, *Arctostaphylous rubra/incana*, *Alnus* type *incana* and *Sphagnum* (Table 20). Minor amounts of *Salix*, *Cassiope* type *Tetragona* and *Empetrum nigrum* are also present.

The assemblage is typical of an open evergreen/deciduous forest, very similar to the community presently occupying the site. The peat probably developed quite recently, within the last 2000 years. Due to the potential for contamination from modern penetrating roots, the peat unit was not ^{14}C dated.

iv) HH 62-145 (81)

This section is located on the east bank of the Peel River 6 km downstream from its confluence with the Caribou River (Figure 2).

Table 20

Palynological counts, Quaternary taxa, peat unit, section HH 62-142 (81). All taxa included in pollen sum

Taxon	Percentage
Picea	49.7
Betula	7.1
Alnus type *crispa*	4.1
Alnus type *incana*	1.1
Salix	0.2
Larix	1.3
Ledum	4.8
Vaccinium	7.1
Arctostaphylous rubra	1.9
A. uva-ursi	0.2
Andromeda	0.2
Cassiope	0.7
Empetrum nigrum	0.9
Myrica Gale	0.2
Cyperaceae	15.2
Gramineae	1.7
Cruciferae	0.2
Saxifraga	0.2
Epilobium	0.2
Viola	0.2
Lycopodium	0.9
Sphagnum	2.0

The base of the section is the Cretaceous Arctic Red Formation shale, which extends 55.1 m above the river level (13 August 1981). The bedrock is overlain by 36 fluvial gravel, sand and silt units, totalling 42.2 m in thickness.

Palynological analysis was attempted on four samples from silt and fine sand beds in the 42.2 m sequence. All proved to be barren. Fragments of Pisidium valves were noted in a single gravel bed (sample #16, 75 m level), but specific identification could not be made. Thus, no palaeoecological conclusions can be formulated for this part of the section.

The gravel sequence is overlain by a clayey silt stratum (52% silt, 48% clay) 6 m thick. This unit proved to be devoid of palynomorphs. The clayey silt is overlain by a silty clay unit (38% silt, 62% clay) 32 cm thick. A sample (#5) from this unit revealed a palynological assemblage dominated by Picea (Table 21). Arboreal Betula and Alnus type crispa are the major components of the arboreal-shrub fraction, with associated Populus and Alnus type incana. The shrub component is dominated by Ledum and Vaccinium, with Empetrum nigrum, Myrica Gale, Andromeda type Polifolia , and Arctostaphylous rubra/alpina. Cyperaceae dominates the herbaceous taxa, Gramineae and Cruciferae being the only other herbs represented. Small proportions of Myriophyllum and Sphagnum are present, but no Pteridophyta spores were detected. The sample also contains fragments of Fragilaria and Navicula.

Table 21

Palynological counts, Quaternary taxa, section HH 62-145 (81). All taxa included in pollen sum

Taxon	Sample #5	Sample #6
	%	%
Picea	40.1	50.1
Pinus	0.0	0.2
Betula	8.3	7.9
Alnus type crispa	7.2	4.3
Alnus type incana	2.2	1.8
Salix	0.0	0.3
Populus	0.4	0.3
Empetrum nigrum	2.9	1.6
Ledum	7.6	8.0
Vaccinium	6.5	5.7
Arctostaphylous rubra	2.5	2.0
A. uva-ursi	0.0	0.2
Cassiope type tetragona	0.0	0.7
Andromeda type Polifolia	0.4	0.0
Myrica Gale	0.9	0.0
Cyperaceae	16.1	11.6
Gramineae	2.0	2.0
Cruciferae	0.4	0.8
Epilobium	0.0	0.2

Table 21 (continued)

Taxon	Sample #5	Sample #6
	%	%
Lycopodium	0.0	1.5
Cystopteris	0.0	0.2
Sphagnum	2.0	1.8
Myriophyllum	0.7	0.0

The silty clay grades upwards into a clayey silt stratum (48% silt, 43% clay) 15 cm thick. Clasts larger than 0.015 mm (6 phi) comprise 25% of the sediment. The fine sand fraction is predominantly quartz and feldspar, but contains 20-25% heavy minerals (most notably hornblende, magnetite, pyrite, corundum, and biotite). The stratum also contains fine granite and orthoquartzite pebbles. Many of these clasts appear to be ventifacts. This unit contains an assemblage in which Picea grains comprise 50.1% of the total (Table 21). A single Pinus type Banksiana grain was observed. Its presence is attributed to long-distance transport. The arboreal-shrub fraction is dominated by Betula (chiefly arboreal) and Alnus type crispa, with associated Alnus type incana, Salix and Populus. Ledum and Vaccinium comprise the majority of the shrub component, with Empetrum nigrum, Arctostaphylous rubra/alpina, and Cassiope type tetragona grains also present. Herbaceous taxa are dominated by Cyperaceae, with minor Gramineae, Cruciferae, and Epilobium pollen. Pteridophyta are represented by Lycopodium and Cystopteris spores, and Sphagnum palynomorphs are also present. This unit is overlain by the litter mat, produced by the modern vegetation community, which includes Picea glauca, Betula, Alnus crispa, Ledum groenlandicum, Vaccinium uliginosum, V. vitis-idaea, Empetrum nigrum, Arctostaphylous rubra/alpina, Salix reticulata and Lupinus.

The succession can be interpreted to represent communities produced during the drying of a Cyperaceae bounded wetland or tundra pond, as the Carex assemblage was replaced by a Picea glauca community. Standing water, either a shallow lake or a series of puddles formed by degrading permafrost occupied the site during the deposition of the lower clayey silt unit. The few diatom fragments present do not provide ample information for an assessment of the limnological characteristics. Several species of Fragilaria and Navicula are found in modern lakes in the Mackenzie Delta region (Koivo and Ritchie, 1978). The absence of pollen in the lower clayey silt cannot be explained.

As the surface water disappeared, the vegetation pattern also changed. The area initially remained moist, as shown by the presence of Cyperaceae, Ledum, Vaccinium, Myrica Gale, Arctostaphylous rubra/alpina, and Andromeda, as well as the absence of Artemisia in the silty clay unit. The presence of some standing water at this time is indicated by Myriophyllum and the diatoms. The grains were derived almost exclusively from the area immediately surrounding the depositional site. The absence of coarse silt and fine sand from the stratum suggests that the sediments were locally-derived, and lack appreciable amounts of aeolian-transported material.

The disappearance of standing water and a general decline of the soil moisture content led to the establishment of an open evergreen forest in the vicinity, dominated by Picea glauca, with Ledum as the dominant understory taxon. The exposed pond sediment was subjected to aeolian reworking and loess deposition, as shown by the presence of heavy-mineral rich fine sand and coarse silt, and the ventifacts. The site thus remained unvegetated for a period after the deposition of the standing water. Wetter areas still existed in the vicinity, as indicated by the presence of Cyperaceae and Arctostaphylous rubra/alpina in the pollen spectrum. Eventually, Picea glauca colonized the open site, producing the modern community observed. The palaeoenvironmental data therefore documents the gradual drying of the tundra pond (or wetland) and the establishment of an open woodland community.

No material suitable for dating is present in any of the sedimentary strata. The stratigraphic position of the units, and their thinness, suggests a very young age, possibly 2000 B.P. or less. The drying of the area was probably initiated by very local conditions, possibly related to permafrost degradation induced by valley cutting by the Peel River or by mass movement of the underlying or flanking gravels.

v) HH 62-147 (81)

This section is located on the east bank of the Peel River 80 km south of Fort McPherson (Figure 2).

The three samples analysed from this locality were obtained from a highly organic silt and gyttja deposit 6.8 m thick, which becomes progressively more fibric towards the surface. The unit directly overlies the Cretaceous Arctic Red Formation shale. Samples were taken 10 cm, 60 cm and 200 cm above the basal contact of sediment with the bedrock. Additional sampling was not possible due to an overhanging peat cap present at the top of the exposure. Wood lenses are present throughout the sediment, increasing in frequency towards the base. The base of the deposit is assumed to be early Holocene.

The palynological assemblage of the basal (10 cm) sample is dominated by Gramineae, Artemisia, and Cyperaceae (Table 22). Betula is the dominant arboreal-shrub taxon, with Alnus and Salix present in small amounts. Ericaceae, chiefly Ledum and Arctostaphylous rubra/alpina grains, comprise the shrub component. Herbs present include Saxifraga, Cruaifereae, and Caryophyllaceae (chiefly Sagina and Silene types). The high concentrations of all the herbs, and the complete absence of Picea palynomorphs, indicate that an open tundra environment prevailed in the region. The sedimentology of the deposit suggests that the depositional environment was a thermokarst pond or lake. This conclusion concurs with that reached by J.V. Matthews Jr. through analysis of the plant macrofossil (Unpublished Report 82-7) and arthropod (Unpublished Report 82-7) assemblages. Macrofossils of Carex, Betula glandulosa, Ranunculus, Potentilla, and other aquatic and shoreline plants were observed. The arthropod assemblage contains aquatic beetles (Cyphon, Stenus, Gymnusa, Olophrum and Lathrobium), as well as dragonfly mandibles. Matthews states that the latter indicate a low arctic environment (personal communication,

Table 22

Palynological counts, Quaternary taxa, section HH 62-147 (81). All taxa included in pollen sum

Taxon	10 cm sample	60 cm sample	2.0 m sample
Picea	0.0	16.2	30.3
Betula	14.0	13.0	9.9
Salix	1.4	0.3	0.3
Alnus type *crispa*	1.8	1.2	0.2
Empetrum nigrum	0.0	6.9	11.4
Larix	0.0	0.0	0.2
Ledum	1.4	2.4	3.3
Vaccinium	0.2	1.6	1.9
Arctostaphylous rubra	0.7	1.4	1.0
Andromeda type *Polifolia*	0.0	0.3	0.0
Shepherdia canadensis	0.0	0.0	0.2
Cyperaceae	16.2	11.0	7.2
Gramineae	25.7	18.9	11.4
Artemisia	17.6	7.5	4.5
Silene	2.4	2.6	2.1
Sagina	2.8	3.3	3.1
Cerastium	0.2	0.0	0.7
Cruciferae	4.5	5.0	3.8
Saxifraga	7.2	6.2	6.3

Table 22 (continued)

Taxon	10 cm sample	60 cm sample	2.0 m sample
	%	%	%
Chenopodium	1.4	1.5	1.5
Rosa	0.0	0.0	0.5
Lycopodium	1.0	0.0	0.0
Cystopteris	1.5	0.0	0.0
Equisetum	0.0	0.3	0.3
Sphagnum	0.2	0.1	0.1

1982). The abundance of wood fragments supports this assessment.

The diatom assemblage was dominated by Fragilaria pinnata with associated Stephanodiscus and Navicula and a trace amount of Melosira. Such assemblages are common throughout the Mackenzie Delta region, and are not preferentially in any geographic location or limnologic environment (Koivo and Ritchie, 1978).

The intermediate (60 cm) sample assemblage reflects a transition from the open tundra pond to a forested river valley environment. The decline in the percentages of Gramineae, Cyperaceae, and Artemisia is matched by rises in the Picea, Empetrum nigrum and Ericaceae palynomorph content. The arbroeal-shrub taxa pollen concentrations decline slightly. The proportions of Saxifraga, Cruciferae, and Caryophyllaceae remain high. The presence of dry-ground taxa such as Equisetum and Shepherdia canadensis indicate that the site was receiving material derived from the slopes flanking the Peel River. The high percentage change in Picea compared to the stability of the concentration of Betula pollen also suggests that aeolian transport of grains from river bank and slope areas was an important factor in the development of the pollen assemblage. Possibly, Picea, Equisetum and Shepherdia canadensis grains were blown upslope by winds channelled along the Peel River valley, and were deposited in the pond located near the valley edge. This hypothesis is supported by the abundance of fine quartz sand and silt contained in both the upper samples, which resemble loess in texture and grain morphology. Fragments of the diatoms Fragilaria pinnata and Navicula, were present.

The trend towards forestation of the valley margins is evident in the spectrum of the uppermost (2 m) sample. Gramineae, Artemisia, and Cyperaceae reach their minimum values, while Picea, Empetrum, Ledum, Vaccinium and Arctostaphylous rubra/alpina increase and Larix pollen appears (Table 22). Alnus, Betula, and Salix concentrations decline somewhat, however, indicating that the component of the spectrum derived from aeolian transport of the valley-marginal vegetation increased in importance. The presence of some open, unstable slopes in the area is indicated by the Shepherdia canadensis, Equisetum and Rosa pollen. Fragments of the diatoms Fragilaria pinnata and Navicula were also observed.

The assemblages can thus be interpreted to represent the gradual development of valley flank vegetation along the Peel River, as reflected in the sediments trapped in a thermokarst pond on the open tundra upland. The low proportions of Alnus suggest an early Holocene age for the base of the unit. The absence of Picea in the basal sample can be reconciled with the postulated early Holocene age if this taxon had been temporarily removed from the immediate area (possibly by fire). The presence of even minor amounts of Alnus is incompatible with the early post-glacial age that a complete absence of Picea would imply.

Environmental Succession

The problems of interpreting fluvially-deposited palynological assemblages, and the lack of chronological data from several locations, limit the interpretation of the regional environmental succession.

The oldest postglacial environment at section HH 72-49 (82) was colder and moister than that currently prevalent in the Peel Plateau. The palynological assemblages suggest upland environments, adjacent to or beyond treeline, similar to conditions prevalent along the flanks of the Richardson Mountains today (Ritchie, 1984). The vegetation community envisaged contains elements of the modern seasonal Orthophyll Meadow, Open Deciduous Orthophyll Scrub, and Seasonal Grass/Herb Steppe communities. The peat containing the oldest assemblage has been dated at 12400±120 B.P. (GSC-3691).

Gradual climate amelioration is indicated by the assemblages of palynological zones B, C, and D from HH 72-49 (82) (Figure 60, in pocket). These assemblages represent the establishment of Open Evergreen Sclerophyll Deciduous Forest, dominated by Picea trees and Betula shrubs with scattered Alnus shrubs. Fire events punctuated the sequence. The youngest sediments of zone D have been ^{14}C dated at 7370±80 B.P. (GSC-3822). Alnus, therefore, may have been extant during the early Holocene.

The pattern of climatic amelioration is similar in general terms to that proposed by Ritchie (1984). The specific details of the early Holocene succession differ from the model proposed by Ritchie (1984) however.

The scarcity of palynologically analysed exposures, compared to the theoretical requirements suggested by Faegri and Iversen (1975) and Ritchie (1983), indicates that the significance of these differences and the degree of local variability cannot be reliably assessed at this time.

The younger Holocene successions, such as those preserved at HH 72-50 (82) (Figure 38) and HH 62-147 (81) (Table 22), record the effects of infilling of tundra ponds and the development of valley marginal vegetation throughout the last 2000 years. The assemblages are characterized by the gradual replacement of Cyperaceae and Gramineae by arboreal vegetation. The single ^{14}C dated peat unit, that at HH 72-50 (82), began to form 2290±50 B.P. (GSC-3690). Peat accumulation at this site thus occurred after this date, in contrast to the regional scenario developed by Zoltai and Tarnocai (1975). These authors sugggested that Sphagnum accumulation at other Mackenzie Valley sites ceased 2710 - 2650 B.P. The later development of the HH 72-50 (82) tundra pond was probably caused by local permafrost degradation induced by the nearby Caribou River.

I. Summary

The sedimentary exposures of the Caribou River Watershed and Brown Bear Creek area contain a record of preglacial fluvial activity; an early glacial event, represented by the weathered granite-bearing "Brown Bear" sediments; two later glacial events, represented by glacigenic ridge-and-valley complexes along the Peel River; lacustrine sediments, formed in the Caribou River valley after deglaciation; and Holocene fluvial and autochthonous peat deposits, some of which contain palynomorphs. A composite stratigraphic column is presented in Figure 6-1.

The earliest sediments preserved are preglacial fluvial gravels, formed in eastward-flowing braided-meandering streams. The preglacial gravels are confined to the tributary valleys of the Peel River. The absence of preglacial sediments in the Peel River valley indicates that this channel was excavated after the initial glaciation of the region.

The "Brown Bear" sediments are exposed along the Peel River in the vicinity of Brown Bear Creek. The unit represents a braided-meandering stream complex which has been extensively weathered. The presence of granitic clasts indicates ~~that~~ the unit was deposited after the initial glaciation of the region. The degree of weathering of the granitic clasts suggests an old, early glacial age for the deposit.

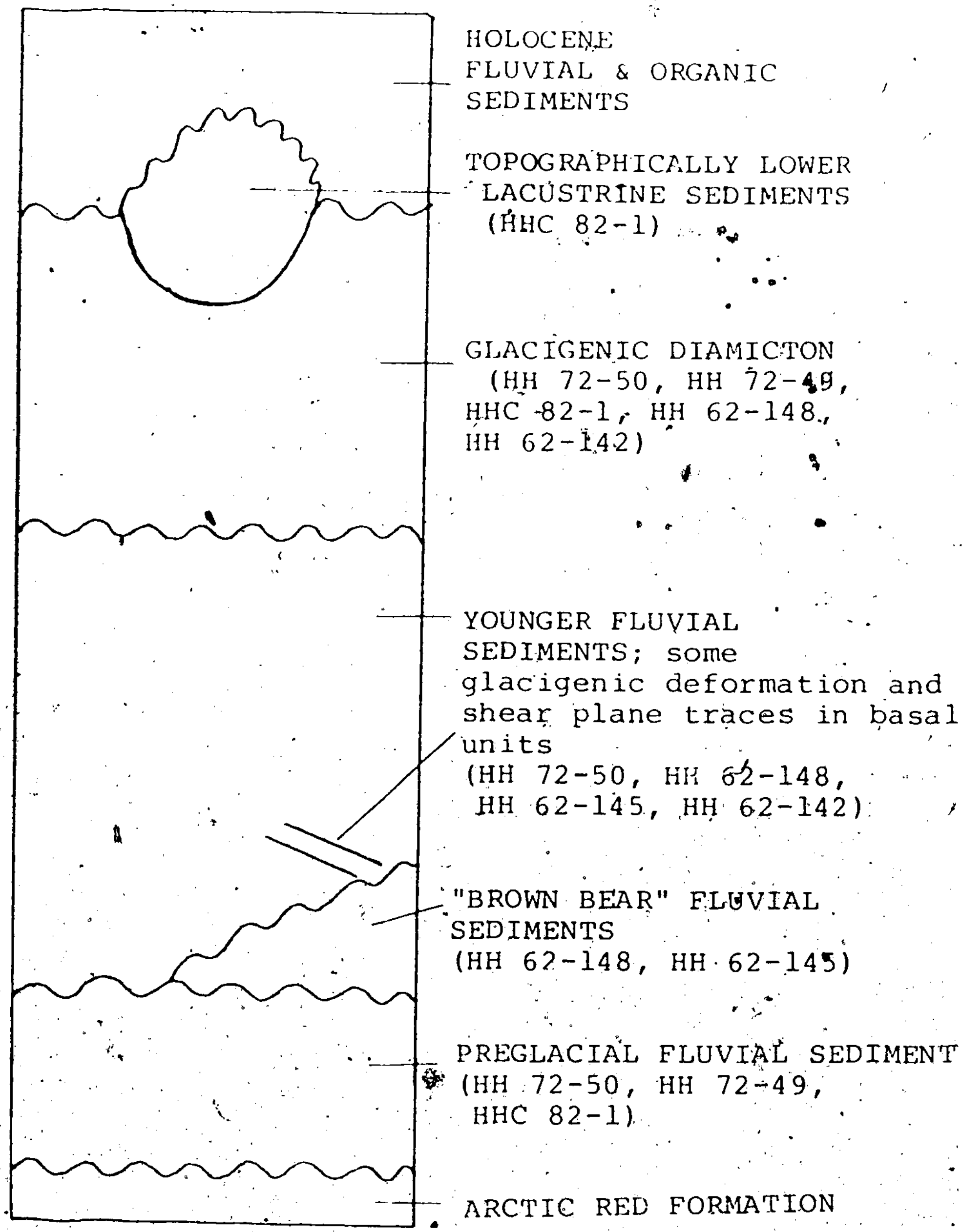
Fluvial sediments which post-date the "Brown Bear" sediments are exposed along the Caribou and Peel Rivers. These units were deposited in braided-meandering streams. Palynological assemblages preserved at HH 72-50 (82) suggest environmental conditions very similar to those currently

Figure 61

Composite Stratigraphic Column, Caribou River-Brown Bear Creek Area

The column is schematic, and the thicknesses of the units do not represent either the thicknesses of sediment deposited or the amount of time represented by each unit. The topographically lower Caribou River lacustrine sediments occur only in the lower Caribou River Valley. The stratigraphic position of the topographically higher lacustrine sediments is unknown, and therefore these sediments are not represented on the column. See text for further discussion.

FIGURE 61


HOLOCENE FLUVIAL & ORGANIC SEDIMENTS
TOPOGRAPHICALLY LOWER LACUSTRINE SEDIMENTS (HHC 82-1)
GLACIGENIC DIAMICTON (HH 72-50, HH 72-49, HHC 82-1, HH 62-148, HH 62-142)
YOUNGER FLUVIAL SEDIMENTS; some glacigenic deformation and shear plane traces in basal units (HH 72-50, HH 62-148, HH 62-145, HH 62-142)
"BROWN BEAR" FLUVIAL SEDIMENTS (HH 62-148, HH 62-145)
PREGLACIAL FLUVIAL SEDIMENT (HH 72-50, HH 72-49, HHC 82-1)
ARCTIC RED FORMATION

prevalent in the region.

Glacigenic ridge-and-valley complexes are present east of the Peel River. The successions within the complexes indicate that two distinct glacial events affected the region. The complexes were formed by thrusting induced by southwestward moving glaciers. Glacigenic sediments and landforms from the younger glacial event are present throughout the Caribou River - Brown Bear Creek area. The direction of ice movement was towards the west and southwest.

Deglaciation of the Caribou River watershed led to the impoundment of a lake in the eastern part of the valley. The lacustrine succession records several environments, all characterised by eastward-flowing currents originating in the deglaciated area. A topographically-higher sequence of lacustrine sediments cannot be assigned a stratigraphic position at present.

Deglaciation of the upper Caribou River valley [illegible]red before 12400±120 B.P. (GSC-3691), as indicated by a [illegible]c date from autochthonous organic sediment at HH 72-49 (82). Gradual climatic amelioration occurred throughout the early Holocene, as indicated by the palynological assemblages. The latter part of the Holocene was characterised by a climate similar to that of the present. The palynological assemblages record the effects of the infilling of tundra ponds and the development of valley marginal vegetation.

THE UNIVERSITY OF ALBERTA

Quaternary Sedimentology and Stratigraphy, Peel Plateau and Richardson Mountains, Yukon and N.W.T.

by

Norman Rhoderick Catto

VOLUME 2

of

A THESIS

SUBMITTED TO THE FACULTY OF GRADUATE STUDIES AND RESEARCH IN PARTIAL FULFILMENT OF THE REQUIREMENTS FOR THE DEGREE OF Doctor of Philosophy

Department of Geology

EDMONTON, ALBERTA

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VOLUME 2

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VIII. Rat River Valley and McDougall Pass Area

A. Introduction

The Rat River Valley and McDougall Pass area represent the northern part of the study region (Figure 8). Exposures of sediment are largely confined to the Rat River valley. The distribution of surficial deposits in the area is illustrated in Figure 62.

The sedimentary record encompasses the Quaternary period from pre-glacial time to the Holocene. In the following sections of the thesis, the deposits will be discussed in chronological order.

B. Preglacial Sediments

A total of sixteen exposures were documented along the west bank of the Rat River between 67° 39' and 40' N and 135° 28' and 30' W, as illustrated in Figure 63. Of these, four sections proved to have the greatest stratigraphic and sedimentologic significance, and were intensely investigated during the summers of 1981 and 1982. Section HH 62-107-81-1 is the most southerly of the exposures. Sections 82-2c, 81-2a, and 81-3a are 900 m, 1000m and 2400 m north of 81-1, respectively.

Section HH 62-107-81-2a is the focal point for discussion of the exposures of the area, as it contains the most complete palynological succession. The stratigraphic relationships among the sections are presented in Figure 64.

Figure 62

Quaternary Geology, Rat River-McDougall Pass Area (after Hughes 1972; Hughes et al 1972, 1973, 1974; and observations by the author)

Holocene Bog, standing water and organic sediments

Holocene Fluvial Deposits along major active streams, sands and gravels

Holocene and Pleistocene abandoned channel deposits, sand, clay, and organic sediments

Pleistocene and Holocene alluvial fan deposits

Pleistocene glacial diamicton, with streamlined features showing orientation of glacial flow

Pleistocene glacial diamicton, hummocky terrain

Pleistocene glacial diamicton, with organic sediment veneer

Pleistocene glacigenic ridge and valley complex

Pleistocene glaciofluvial deposits, sand and gravel

Pleistocene esker, sand and gravel

Pleistocene glacial deposits, scattered erratics and thin discontinuous diamicton cover

X---Principal sections discussed in text

FIGURE 62

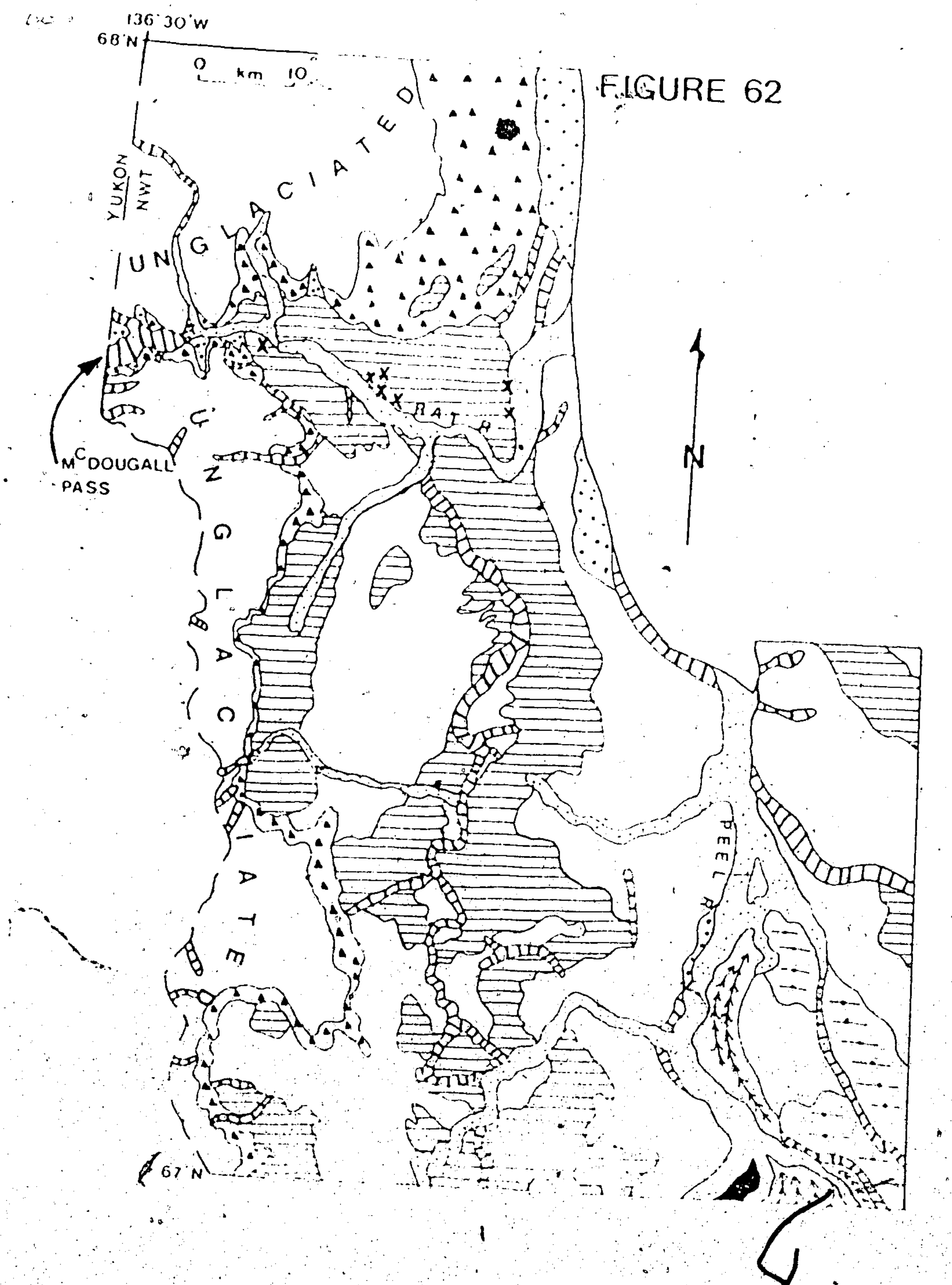

136°30'W
68°N
0 km 10
YUKON
NWT
UNGLACIATED
M^C DOUGALL PASS
UNGLACIATED
RAT R.
N
PEEL R.
67°N

FIGURE 63

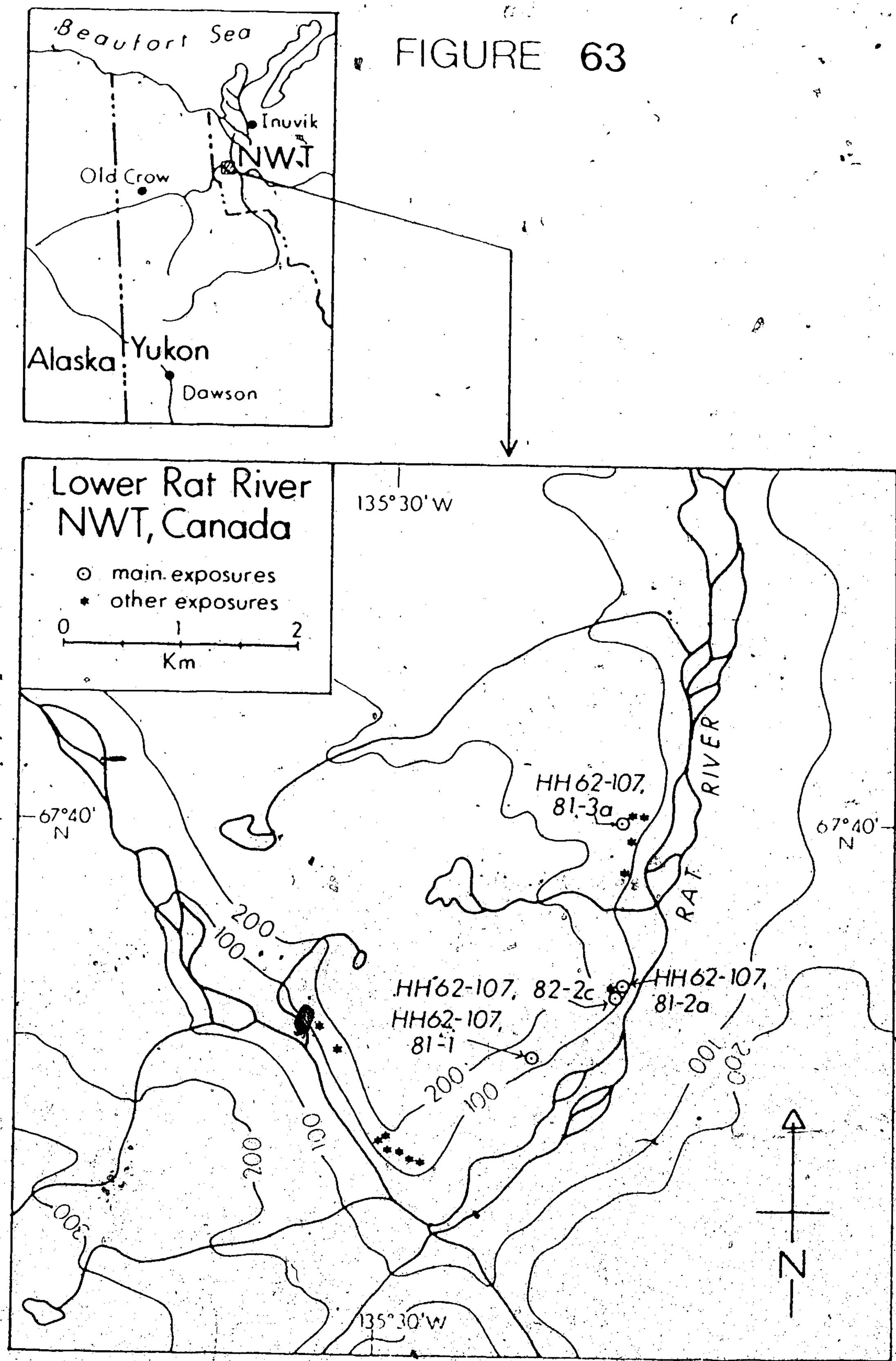
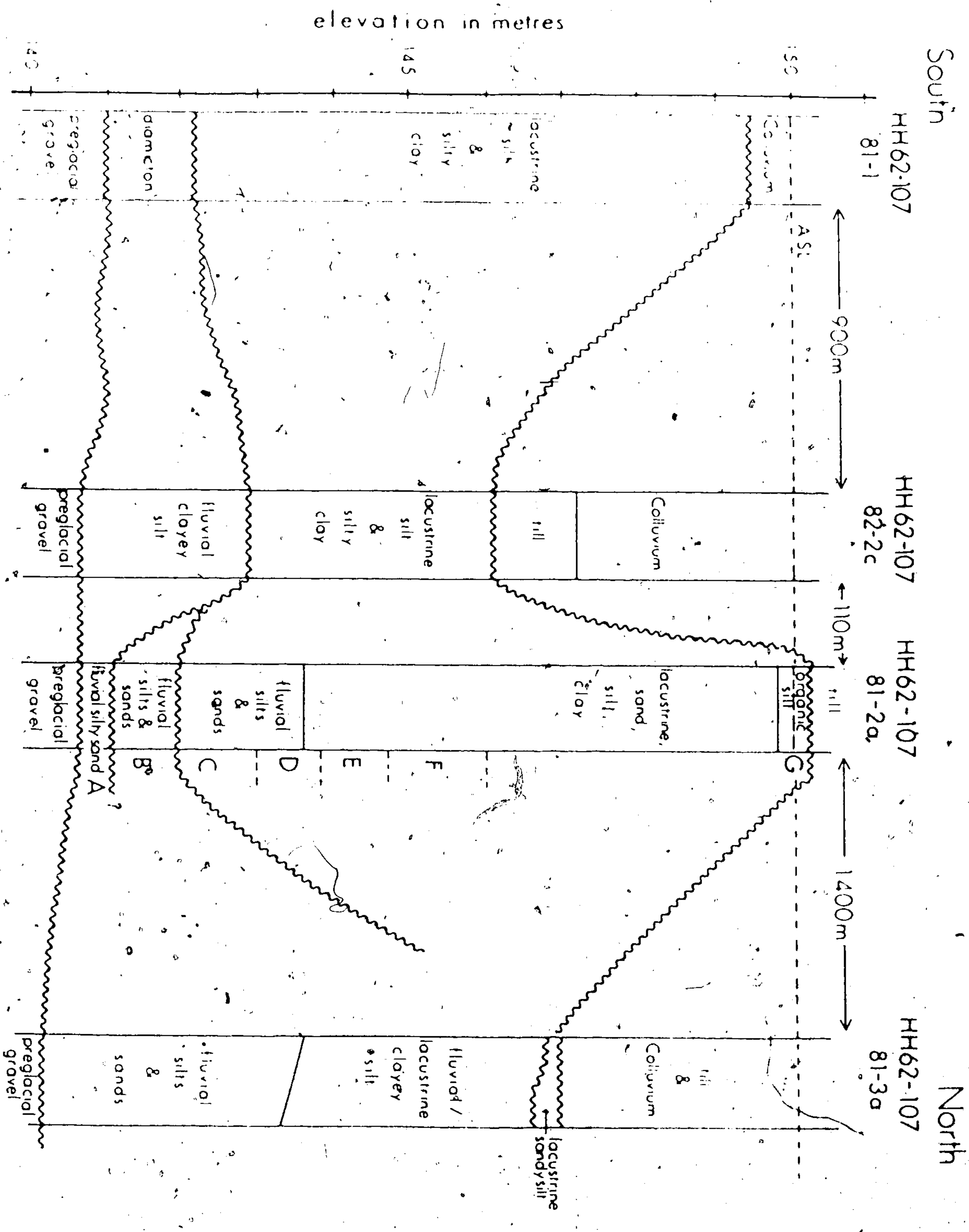

Beaufort Sea
Inuvik
NWT
Old Crow
Alaska
Yukon
Dawson
Lower Rat River
NWT, Canada
main exposures
other exposures
0
1
2
Km
135°30'W
67°40' N
HH 62-107, 81-3a
RIVER
RAT
HH62-107, 82-2c
HH62-107, 81-2a
HH62-107, 81-1
200
100
300
N

Figure 64

Correlations, HH 62-107 area

The stratigraphic relationships between the exposures are illustrated. Palynological zones defined from assemblages contained in exposure HH 62-107 81-2a are indicated by letters A through G. See text for further discussion.



elevation in metres
140
145
150
South
North
HH62-107 81-1
HH62-107 82-2c
HH62-107 81-2a
HH62-107 81-3a
900m
110m
1400m
ASL
Colluvium
lacustrine silt & silty clay
diamicton
preglacial gravel
till
fluvial clayey silt
lacustrine sand, silt, clay
organic silt
fluvial silts & sands
fluvial silts & sands
fluvial silty sand
A
B
C
D
E
F
G
till & Colluvium
fluvial / lacustrine clayey silt
fluvial silts & sands
lacustrine sandy silt

FIGURE 64

The basal preglacial gravel has a maximum thickness of 18 m (Plate 28). It overlies Cretaceous shale at all exposures where it is present. The gravel is dominated by orthoquartzite and other clasts derived from the Rat River watershed, and no granitic or metamorphic clasts are present (Table 23). These data indicate that the sediments were deposited prior to the initial glaciation of the area.

The gravels are composed of structureless or crudely horizontally stratified pebble and cobble beds. The sequences are similar to those formed in lateral and longitudinal bars developed by modern braided-meandering streams in the area (Appendix 4). The orientations of the clasts suggest that flow was generally towards the east.

C. Early Sediments, Lower Rat River Area

Zone A

At section HH 62-107 81-2a, the basal preglacial gravel is overlain by a 40 cm thick unit consisting of fluvial silty sand (54% sand, 18% clay) at the base, fining upward to sandy silt (32% sand, 8% clay) near the upper contact (Figures 63 and 64). Two samples from this sediment were processed for palynomorphs. The spectra are considered as a single pollen zone A (Figure 65, in pocket).

The assemblages are dominated by spores, particularly _Lycopodium_, _Sphagnum_, _Equisetum_, and _Polypodiaceae_. Poorly preserved Cyperaceae, Gramineae, _Picea_, _Saxifraga_, _Salix_,

Plate 29 (top)

Preglacial Gravel, HH 62-107-81-2a. The gravel is devoid of igneous and metamorphic rock clasts.

Plate 29 (bottom)

Cryoturbated Diamicton, HH 62-107-81-1. The dark zones within the diamicton are masses of peat. Analyses have revealed the presence of tundra palynomorphs in the peat inclusions. Note that many of the larger clasts are oriented vertically. See text for discussion; also see Figure 66. Ice axe is 1 m long.

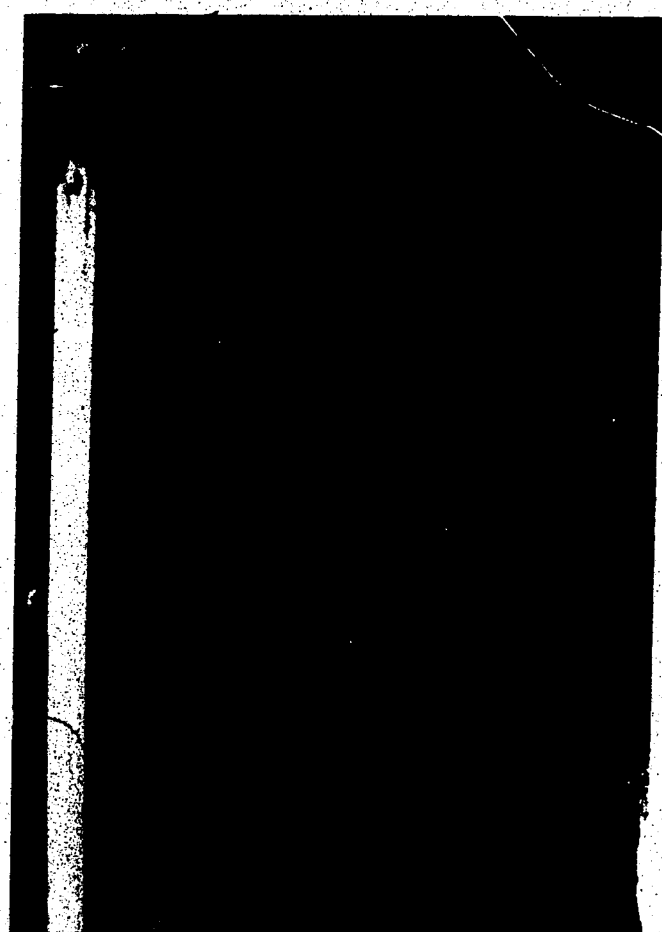


Table 23

Preglacial sediments, Rat River valley, clast lithology and mineralogy.

A) Pebbles (7 analyses)

Lithology	Maximum	Mean	Minimum
	%	%	%
orthoquartzite	98	93	83
feldspathic sandstone	2	1	0
siltstone/ shale	1	< 1	0
argillite	15	4	< 1
limestone	1	< 1	0
iron carbonate	2	< 1	0

B) Granules (7 analyses)

Lithology	Maximum	Mean	Minimum
	%	%	%
orthoquartzite	98	91	80
feldspathic sandstone	4	1	0
siltstone/ shale	1	< 1	0
argillite	17	6	< 1
limestone	< 1	< 1	0
iron carbonate	1	< 1	0

Table 23 (continued)

C) Coarse sand (0.42 - 1.00mm), (6 analyses)

Lithology	Maximum	Mean	Minimum
	%	%	%
quartz	98	89	85
feldspar	15	5	2
argillite fragments	12	3	0
chert	5	2	0
calcite	1	<1	0
dolomite	<1	<1	0

Cruciferae, Ericaceae, Betula, Sagina type, and Silene -type pollen are also present. This assemblage probably reflects the effects of differential preservation of the spores. The pre-Quaternary component which can be definitively recognized reaches 45 % of the assemblages. Some of the spores may represent pre-Quaternary taxa.

At section 81-1, the preglacial gravel is overlain by a sandy silt matrix diamicton varying in thickness from 0.6 to 1.2 m (Figures 64 and 66, Plate 29). No granites, gneisses, or other Canadian Shield-derived pebbles were present. The matrix, however, contains apatite, fluorite, garnet, magnetite, ilmenite, wollastonite, and epidote (Table 24). This unit therefore postdates the first glaciation of the area. The diamicton contains lenses of silt similar in composition to the overlying unit, as well as several detrital organic beds and lenses. Sedimentary structures in the stratum suggest cryoturbation (Figure 66).

Three samples from this stratum were investigated for palynomorphs. Sample 1-8, obtained from disseminated organic detritus within the matrix of the diamicton 35 cm above the basal contact, contained an assemblage dominated by Cyperaceae and Gramineae, with lesser amounts of Picea (both mariana and glauca type), Betula, Vaccinium, Ledum, Artemisia, Salix, and pteridophyte spores. The diversity of the assemblage is very low, and pre-Quaternary spores (primarily Cretaceous) comprise approximately 55 % of the sample. Samples 1-9 and 1-9a, obtained from organic detrital

Figure 66

Diamicton, HH 62-107 81-1

The diamicton unit illustrated crops out above preglacial gravel. The unit is interpreted as a cryoturbated deposit. See text for further discussion; also see Plate 29.

DIAMICTON, HH 62-107 81-1
FIGURE 66

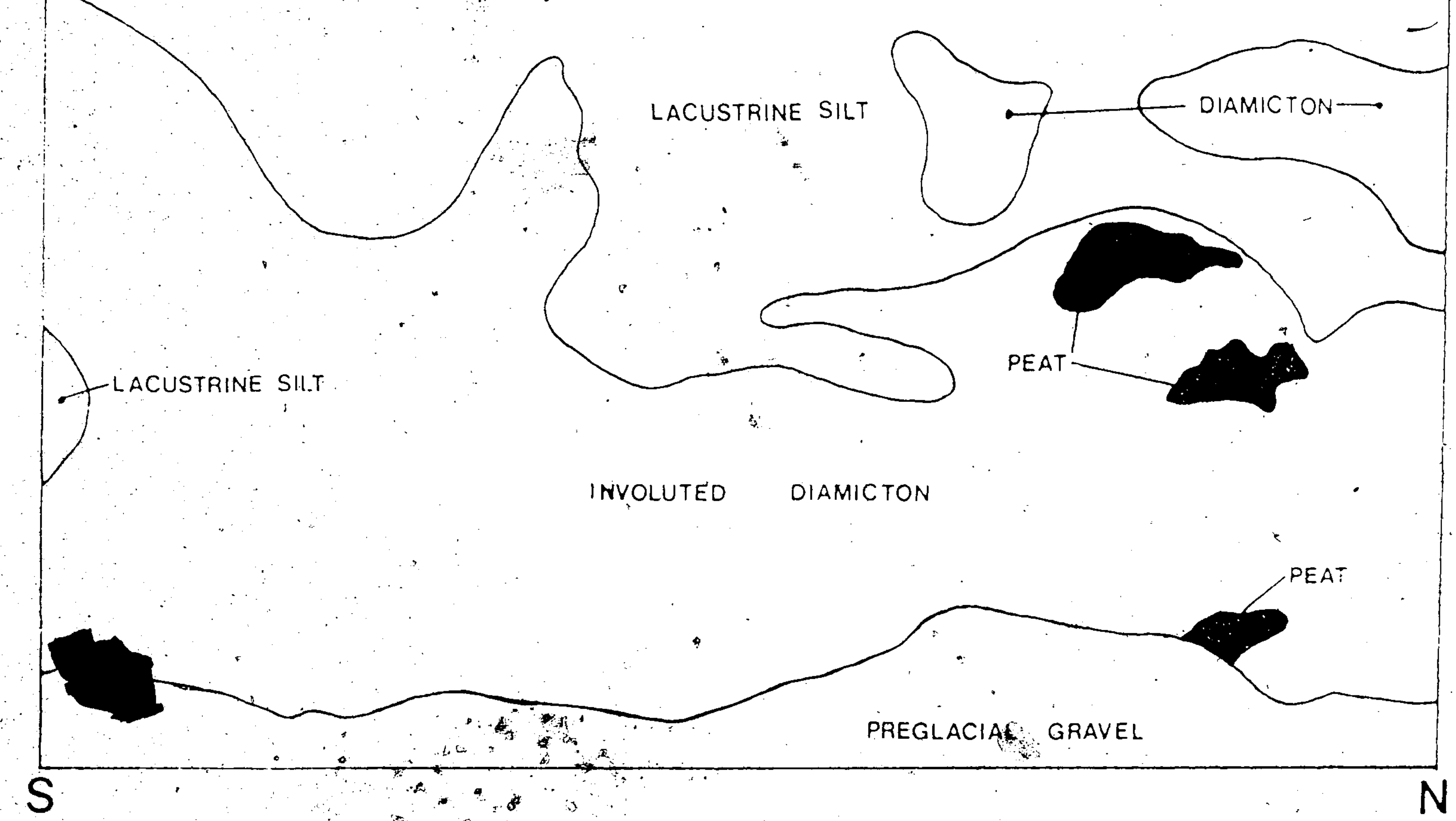


Table 24

Early diamicton, Lower Rat River, HH 62-107 (81-1), clast lithology and mineralogy.

A) Pebbles (mean of 3 analyses)and granules (mean of 3 analyses).

Lithology	Pebbles	Granules
	%	%
orthoquartzite	92	91
feldspathic sandstone	4	2
siltstone/ shale	<1	2
argillite	3	2
limestone	<1	<1
iron carbonate	<1	<1

B) Coarse sand (0.42-1.00 mm), major minerals

Lithology	Mean of 3 analyses
	%
quartz	88
feldspar	8
argillite/ siltstone	1
chert	2
calcite	<1
dolomite	<1

Table 24 (continued)

C) Coarse sand, trace minerals (3 analyses).

Mineral	Number of analyses containing mineral
Apatite	2
Biotite	3
Chlorite	3
Epidote	2
Fluorite	3
Garnet	2
Gypsum	3
Hornblende	3
Ilmenite	3
Magnetite	3
Wollastonite	3
Jasper-hematite	1
Vivianite	3

lenses 45 cm and 65 cm above the basal contact, are dominated by Gramineae grains. Other major constituents are Cyperaceae, Picea, Artemisia, Saxifraga, and Chenopodium. Lesser amounts of Caryophyllaceae (Cerastium -type), Betula, Salix, Shepherdia canadensis, , and Ericaceae (Ledum-type?) grains are also present. Pre-Quaternary spores (Early Cretaceous) comprise approximately 25 % of the sample assemblages. Preservation of the grains in all three samples was fair to moderate, although these grains proved to be susceptible to degradation subsequent to processing.

At section 82-2c, 900 m north of section 81-1, the preglacial gravel is overlain by 2 m of clayey silt-silty clay, which contains some fine-grained sand lenses. This unit was deposited in back-bar channels and sloughs of a braided-meandering stream, and is correlative to the basal fluvial sand exposed at 81-2a. Seven samples from the unit exposed at 82-2c were processed, and two of these (2c-6 and 2c-8), obtained 0.75 m and 1.25 m above the basal contact, proved to have sufficient palynomorphs for counting.

Both samples are dominated by spores, particularly Polypodiaceae, Lycopodium, and Equisetum. Poorly preserved Gramineae and Cyperaceae grains, and trace amounts of Cruciferae and Saxifraga, are also present. These assemblages reflect the combined influences of differential preservation and the incorporation of an uncertain number of pre-Quaternary spores. The component recognised as pre-Quaternary is in excess of 70 % for both samples.

The assemblages at all three sections can be interpreted to represent a low arctic tundra environment, with scattered stands of Picea and arboreal Betula, small amounts of shrub Betula and Salix, no Alnus, and large expanses of open tussock tundra. Small fragments of Picea wood testify to its presence on or near the sites. Assemblages from such environments are usually dominated by Cyperaceae and Gramineae (eg. Colinvaux 1967, Ritchie 1982, 1984). The extremely high percentage of Gramineae grains in the upper two 81-1 samples indicates that these taxa were the major form of vegetation at the time of deposition. The cryoturbation structures in the sediment at 81-1 are also consistent with the hypothesis of an open tundra setting. Cryoturbation is probably primarily responsible for the introduction of detritus and slope wash from surrounding exposures of bedrock playing minor roles..

Cyperaceae seeds and Coleoptera from sample 1-9a have been examined by Matthews (Geological Survey of Canada, personal communication, 1983). Small mammal feces and mammal bones were also observed, but were not identified. No definite conclusions were possible, although the presence of the moss beetle Ochthebius is unusual in tundra assemblages.

The reliability of the environmental interpretation is limited by the similarity of the pollen of the various genera within the families Gramineae and Cyperaceae. As the families contain genera capable of surviving in many climatic regimes (Colinvaux 1967), interpretation based largely on their presence and the absence of other taxa is highly tenuous.

Zone B

An unconformity separates the sediment of zone A from a fluvial complex of overbank, back-bar channel, slough, and scour pool clayey and sandy silts (10-38 % sand, 8-26 % clay), and bar sands at HH 62-107-81-2a (Plate 30). Also included in the complex is a bar-head jam deposit and minor flank deposits. This unit varies in thickness from 2.3 to 2.7 m. Fragments of Pisidium, ivory, and Picea cones and twigs were found in the bar head and flank jam deposits.

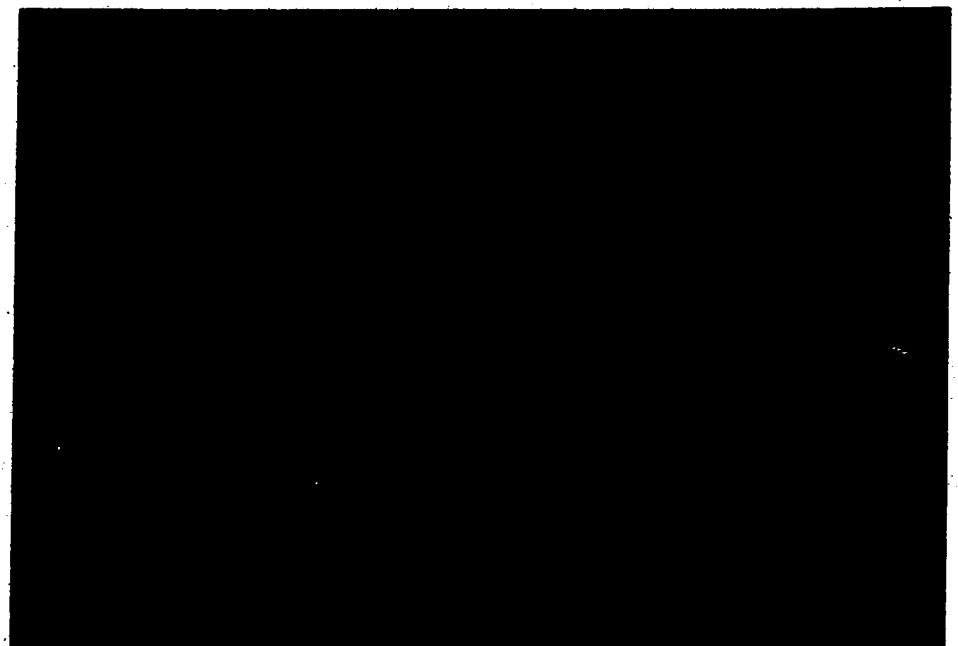
Three palynological zones were distinguished from the 18 samples obtained from this unit (Figure 65, in pocket). The basal zone, B, is rich in Picea mariana and Picea glauca (Plate 31). Diploxylon Pinus grains are also present. The arboreal-shrub fraction is dominated by Betula (mainly tree), with Alnus type crispa, Salix, Populus, and trace amounts of Alnus type incana and Corylus. The shrub component includes Ledum, Vaccinium, Andromeda, Empetrum nigrum, Cornus, and Pyrola. The herb fraction is dominated by Cyperaceae and Gramineae, with lesser amounts of Chenopodium, Cerastium, Sagina, Ranunculus, Rubus, Saxifraga, Epilobium, Viola, Artemisia, and Cruciferae. The pteridophyte spores are predominantly Lycopodium, some of which are reworked Cretaceous examples. Sphagnum is the only bryophyte present.

Plate 30 (top)

Contact between palynomorph zones A and B, HH 62-107-81-2a. The thin gravel layer separates sediments containing tundra palynomorphs (zone A) from sediments containing *Pinus*, *Corylus* , and other pollen indicative of warmer conditions (zone B). See text for discussion. The red part of the shovel is 15 cm long.

Plate 31 (bottom)

A part of palynomorph zone B, HH 62-107-81-2a. Palynological samples were obtained from a back-bar channel assemblage (left side of photograph, with sample locations marked with flagging tape). The assemblage was deposited in the lee of a bar-head jam deposit (right side of photograph, sample location 8a). Note the abundance of logs and other organic detritus.



Coarser-grained strata within the zone contain higher percentages of Lycopodium, Chenopodium, and Ericaceae grains. The finer strata are richer in the arboreal and arboreal-shrub palynomorphs.

No sediments containing similar assemblages were observed at 81-1 and 81-2c. At the northernmost section, 81-3a, however, the upper part of a fluvial stratum (Figure 64, Plate 32) contained both palynological and other environmental indicators suggesting that the unit is correlative to the fluvial deposits at 81-2a. Scattered throughout the stratum are sandy silt and sand lenses containing Pisidium nitidum, Valvata, and Lymnaea shells. Picea seeds and Larix cones were also observed, along with decayed deciduous wood fragments. Three samples from the upper part of the clayey silt contained small amounts of Picea, arboreal Betula, Alnus type crispa, and Cyperaceae pollen, although in insufficient quantities for counting. One of the samples contained a single Pinus grain. A left femur of the ground squirrel Sphermophilus, identified by C. R. Harington (Palaeobiology Division, National Museum of Natural Sciences, Canada), was recovered from this stratum (personal communication, 1983).

The plant macrofossil assemblages are diverse, containing seeds of Carex pseudo-cyperus , a taxon not currently extant in the Northwest Territories (J. V. Matthews Jr., Geological Survey of Canada, personal communication, 1983). The assemblages also contain more

Plate 32

Fluvial sand and silt, HH 62-107-81-3a. The succession consists of a series of fining-upward silty sand and sandy silt units, deposited in a back-bar channel, overlain by medium-grained sand and sandy silt. The sharp, irregular contact between the lower sandy silt and the overlying medium-grained sand suggests that rapid deposition of the sand bed occurred. The protrusions of silt into the overlying sand unit indicate that loading of the silt accompanied the deposition of the sand. The texture of the sandy silt directly below the sand is similar to that of loess deposits in the area, suggesting that subaerial exposure occurred. The development of a thin crust of ice or compacted snow can initiate the formation of small irregularities in exposed sediment layers. Subsequent loading during spring flooding acts to enhance these protrusions, producing irregular contacts. Similar structures have been observed in modern sediments.


minolta

Alnus fragments than Betula (which is absent from some samples, more Carex diandra type and Glyceria than are commonly present in northern samples, and few Potamogeton fragments. In addition, an unusual Picea needle present in sample 3a-39 could represent a southern species. The insect components of sample 3a-39 contains elytkona of the weevil Lixellus, as well as fragments of many aquatic and nearshore taxa.

Corylus, Pinus, and Cornus canadensis all require warmer climatic conditions than those currently existing along the lower Rat River. The presence of small amounts of Pinus and Cornus pollen can be attributed to either minor climatic ameliorations or to sporatic long -distance aeolian transport. Corylus is not currently extant in the Yukon or N.W.T. (Porsild and Cody 1980), so the degree of climatic amelioration required for it to extend its range to the Rat River is considerable. The small number of Corylus grains observed suggests that the taxon was not extant in the region at the time of deposition, although it may have been growing to the north of its present maximum range. Lichti-Federovich (1974) regarded Corylus grains in the sediments at Old Crow as indicative of the occupation of the area by the southern boreal forest, while Schweger (1982) considered grains found in alluvial sediments in Alaska to be indicative of long-distance aeolian transport. No Corylus macrofossils have been recovered from the northeast Beringia region.

The concentrations of arboreal and arboreal-shrub pollen are also indicative of warmer climates. The terrain in the vicinity of the depositonal site was moist, providing habitats for Ledum, Vaccinium, Andromeda, Ranunculus, and Viola. Drier areas (such as exposed bars) supported Artemisia, Sagina, and Epilobium. The river banks were colonized by a southern boreal forest similar to that of northern Alberta and southern Mackenzie District, while the upland areas supported a taiga assemblage. This environment is also suggested by the macrofossil and arthropod assemblages present in the sediments of 81-3a. Lichti-Federovich (1974) and McCourt (1982) described palynological assemblages indicative of southern boreal forest conditions dominated by Betula, Picea, and Pinus, with scattered Alnus and Corylus shrubs in the Old Crow region to the west. Differences in the concentrations of conifer and Cyperaceae pollen between the assemblages were attributed to site location relative to the centre of the basin by McCourt (1982). He suggested that sites in the basin ~~centre~~ (such as those investigated by Lichti-Federovich) should receive a greater proportion of arboreal pollen than sites on the basin margin (such as that investigated by McCourt).

Zone B at HH 62-107-81-2a is similar in some respects to those reported from the Old Crow region. However, the low concentrations of Betula and Alnus pollen in Zone B suggest a mature Picea taiga forest on upland areas, intersperced with low, moist clearings supporting grasses and sedges, combined with a boreal assemblage in the floodplain and riverbank areas. Similar assemblages are common today in the southern Mackenzie district (Rowe 1972, Ritchie 1984), although Corylus is absent.

Zone C

The sediments containing the palynomorphic assemblages of zone C at 81-2a are separated from the underlying zone B sediments by an irregular iron-oxidised lamination which shows no definite sign of representing a major erosional event (Figure 64). The assemblages of zone C, however, are chartacterised by high concentrations of Cyperaceae, Gramineae, and Artemisia , and low concentrations of arboreal, arboreal-shrub, and pteridophyte palynomorphs (Figure 65, in pocket). The arboreal and arboreal-shrub components are dominated by Picea (mariana-type), and Betula (shrub) respectively. The shrub fraction is chiefly Andromeda type Polifolia and Ledum pollen. The herb comopnent contains Cérastium, Sagina, Silene, Chenopodium, Saxifraga, Potentilla, Epilobium, Rubus chamaemorus, , and Cruciferae pollen. The spore fraction is small and is dominated by Polypodiaceae. The aquatic taxon Potamogeton is represented in some samples.

Sample 2a-11b, from a pool developed below a bar-head jam deposit, was examined by Matthews (Geological Survey of Canada, personal communication, 1983). The plant macrofossils noted included _Picea_ needles and seeds, _Sparganium_, _Carex_, _Potamogeton_, _Betula glandulosa_, and _Potentilla_ seeds, and _Salix_ capsules. A _Bidens cernua_ seed was also noted. Arthropods noted by Matthews included Carabidae, Dytiscidae, Gryinidae, Curculionidae, and _Cercyon_. Fragmented mammal bones and small mammal feces were also present.

The pollen, plant, and insect fossils suggest an environment similar to today's, with forested river valleys bordered by arctic tundra uplands. Temporary scour pools, back-channel ponds, and sloughs provided a habitat for _Potamogeton_ , and the banks of the pools and back-bar channels were colonized by Cyperaceae, _Sparaginium_, _Salix_, _Betula glandulosa_, and the arthropoda. Wet depressions were inhabited by _Ledum_, _Andromeda Polifolia_, and _Rubus chamaemorus_ . Exposed bar surfaces supported _Chenopodium_, _Sagina_, _Silene_, and _Epilobium_ .

The low concentrations of _Picea_ and the other arboreal taxa suggest that the assemblages were local in derivation. The variety within the pollen spectra and the arthropod taxa indicate a climate not more severe than at present. The presence of _Bidens cernua_ , a southern species, argues against a tundra environment which might be inferred from the scarcity of arboreal pollen. The similarity of the assemblages of zone C to those of modern forest/tundra transition areas, however, clearly distinguishes this zone from the more southerly boreal forest conditions represented by zone B.

Zone D

Zone C is transitional into the overlying palynological unit, zone D. This zone encompasses the upper part of the fluvial complex at 81-2a and also extends through the overlying sedimentary unit, a fining-upwards sandy to clayey silt 30 to 70 cm thick (Figure 64). This unit is interpreted as a thaw-lake deposit, formed in a floodplain shortly after abandonment of the area by a braided-meandering stream.

The assemblages of zone D are dominated by *Picea*, largely *mariana* type, Cyperaceae, and Gramineae (Figure 65). *Larix* and *Pinus* grains are present in trace amounts. The arboreal-shrub fraction is small, and is dominated by *Betula* shrubs with *Alnus* type *crispa*, *Populus*, *Salix*, and arboreal *Betula*. The only shrub pollen represents Ericaceae, chiefly *Vaccinium* and *Ledum* with small amounts of *Andromeda*. In addition to Cyperaceae and Gramineae, *Cerastium*, *Sagina*, *Saxifraga*, *Epilobium*, *Artemisia*, Cruciferae, and *Chenopodium* are present in small amounts. The uppermost samples (from the lacustrine deposits) are especially rich in Cyperaceae. Polypodiaceae, *Lycopodium*, and *Sphagnum* spores are also present.

The assemblages of zone D can be interpreted in three ways: a) as a result of gradual climatic amelioration; b) as assemblages more representative of the surrounding landscape than those of zone C; or c) as a result of a shift in the adjacent river channel and the stabilization of the valley side and slope environment.

The main differences between zone D and the underlying zone C are the increase in Picea concentration and the corresponding decreases in Caryophyllaceae, Saxifraga, Artemisia, and Rubus chamaemorus . As the assemblages of zone C suggest a climate similar to that of today, continued amelioration would be expected to produce a rise in the concentrations of either or both of the arboreal-shrub taxa, Alnus and Betula . The proportions of these taxa in the spectra remain approximately constant, however. The low concentrations of Pinus indicate that it was not extant at the site at this time. Therefore, no firm evidencce to support the hypothesis of climatic amelioration exists.

Long-distance transport of pollen grains would tend to increase the proportions of upland taxa present, thus resulting in a decrease in the site-specific taxa. The elevation of the site with respect to the surrounding plateau, however, suggests that extensive aeolian transport of grains would result in increases in the low arctic tundra component, represented by Gramineae, Cyperaceae, and Artemisia . Long-distance fluvial transport may in part be responsible for the increased concentration of palynomorphs capable of floating for extended periods, such as Picea .

Shifting of the adjacent river channel provides a partial explanation for the changes observed between these zones. Stabilization of the bars and back-channel areas would lead to colonization by *Picea* and a gradual disappearance of the unstable exposed coarse sediment areas favoured by Caryophyllaceae and *Artemisia* . Simultaneously, the wet low-lying areas preferred by *Andromeda* and *Rubus chamaemorus* would gradually drain, thus reducing the suitable habitats for these taxa. The establishment of a thaw lake on the site is also compatible with the gradual shifting of the river channel. In the absence of large fluctuations in the concentrations of the pollen of most of the taxa, therefore, channel shifting is considered to be a more reasonable explanation of the variations in the spectra than are either of the other two hypotheses.

Zone E

Overlying the sediments of zone D at 81-2a is a sequence of lacustrine sands, silts, and clays 6.3 m thick (Figure 64). All of these sediments were deposited in environments influenced by permafrost. The site gradually changed from a back-channel pond, occasionally flooded by water from the main stream, to an isolated thermokarst lake.

The assemblages obtained from the lower 70 cm of the unit are grouped as zone E (Figure 65, in pocket). These spectra are dominated by Cyperaceae and Gramineae. *Picea* is the only arboreal taxon present, and the arboreal-shrub fraction, containing *Betula*, *Alnus* type *crispa*, and *Salix* , is also small. The shrub component is chiefly *Andromeda*. The herb fraction contains *Cerastium*, *Sagina*, *Silene*, *Chenopodium*, *Saxifraga*, *Artemisia*, *Epilobium*, *Potentilla*, *Rubus*, and Cruciferae. Pteridophyta are dominated by *Lycopodium* and Polypodiaceae. Some of these spores are pre-Quaternary, and their presence along with the increased concentrations of *Chenopodium* indicate that fluvial transport of palynomorphs was an important factor in the development of the assemblage.

Similar assemblages were noted in the upper part of the silty clay and silt stratum at 81-1. Cyperaceae and Gramineae pollen were dominant. Cretaceous spores comprise approximately 15 % of the assemblage.

The sediments of zone E were formed in a back-channel pond or slough developed in a taiga-low arctic tundra transition area, under the influence of a climate similar to today's. The assemblages are similar to those of zone C. The low concentrations of *Picea* pollen are due to the unsuitability of the immediate environment for *Picea* growth, and also probably due to periodic fires. Charred needles are present in the sediment, and charred air-lofted cone fragments were also observed. The arthropod fossils in this sediment also indicate a taiga or low arctic tundra sedge wetland (J. V. Matthews Jr., Geological Survey of Canada, personal communication).

Zone F

The overlying 5.6 m of lacustrine sediment at HH 62-107-81-2a contained palynomorphs only in the lower 2 m.

Fragments of Picea wood, ivory, and Pisidium shells are present. The palynomorph assemblages from this sediment are considered as zone F (Figure 65, in pocket).

The assemblages are dominated by Picea (mainly mariana), Cyperaceae, and Gramineae. The arboreal-shrub fraction, comprised of Betula, Alnus type crispa, and Salix, is small. Vaccinium and Ledum are the only shrub grains present. Small amounts of Cerastium, Cruciferae, Silene, Chenopodium, Saxifraga, Artemisia, and Epilobium are also present. Pteridophyta are dominated by Lycopodium.

The spectra are similar to those of zone D, and developed adjacent to a relatively stable thermokarst lake in a taiga or low arctic tundra environment similar to that prevalent in the region today. Plant macrofossils and arthropods examined by Matthews (Geological Survey of Canada, personal communication, 1983) support this conclusion. The variations between zones E and F were caused by changes in the local topographic setting and sedimentological environment, rather than by climatic change.

Zone G

The upper 3.6 m of the lacustrine unit is devoid of palynomorphs and diatoms. Pisidium nitidum shells were present 3.5 m above the basal contact.

The lacustrine sediments are overlain at 81-2a by 45 cm of organic-rich silt containing thin, discontinuous beds of organic material (Figure 64). This sediment was deposited on the infilled lake site under tundra conditions. The silt is a combination of reworked lacustrine material and loess. The palynomorphic assemblages obtained from two samples of this unit are considered as zone G (Figure 65, in pocket).

The assemblages are dominated by Cyperaceae, Gramineae, and *Artemisia*, with associated *Saxifraga*, Cruciferae, and *Cerastium*-type pollen. Minor amounts of *Picea* (*mariana* type) and *Betula* (shrub type) grains were also present. The assemblages have low diversities, and the relative pollen influx is low compared to samples from the older, underlying zones. These characteristics indicate that the environment at the time of deposition was cold, dry, open tundra. The arthropod fossils present also support this conclusion. Abundant small mammal feces indicate that the material is probably autochthonous. Sample HH 62-107-81-2a from this unit has been ^{14}C-dated at > 43,000 B.P. (GSC-3359). The tundra sediments are overlain by till, which extends to the surface (Figure 64).

Summary

The palaeoenvironmental succession indicated at the four principal HH 62-107 exposures commenced with open tundra conditions (zone A). After a period of erosion, warmer conditions prevailed. The Rat River valley was colonized by a southern boreal forest, similar to that of northern Alberta and southern Mackenzie District (zone B). A second hiatus separated this period from a colder event, during which forest/ tundra transition conditions similar to those of today prevailed in the area (zone C). Changes in the local environment produced by channel migration and lake development caused fluctuations in the distribution of the taiga and low arctic tundra taxa (zones D, E, and F). The final event recorded was the development of open tundra conditions in the area, at some time before 43,000 years B.P. (zone G).

The chronological data indicate that the deposits were formed prior to the final glaciation of the Rat River area, and also predate the exposures of sediments in the Bonnet Plume Basin described by Hughes et al (1981). The period of warm climate represented by the assemblages of zone B is tentatively correlated to that represented by similar assemblages documented by Lichti-Federovich (1974) and McCourt (1982), in the absence of absolute dates. The age of the underlying tundra zone (A) is unknown. The overlying tundra-taiga transition zones (C-F) were developed under conditions similar to those prevalent in the Rat River area during the Holocene, but no similar assemblages predating 43,000 B.P. have been reported from northern Yukon or western Northwest Territories. The youngest zone, G, is indicative of tundra conditions.

D. Early Lacustrine Phase, Rat Valley

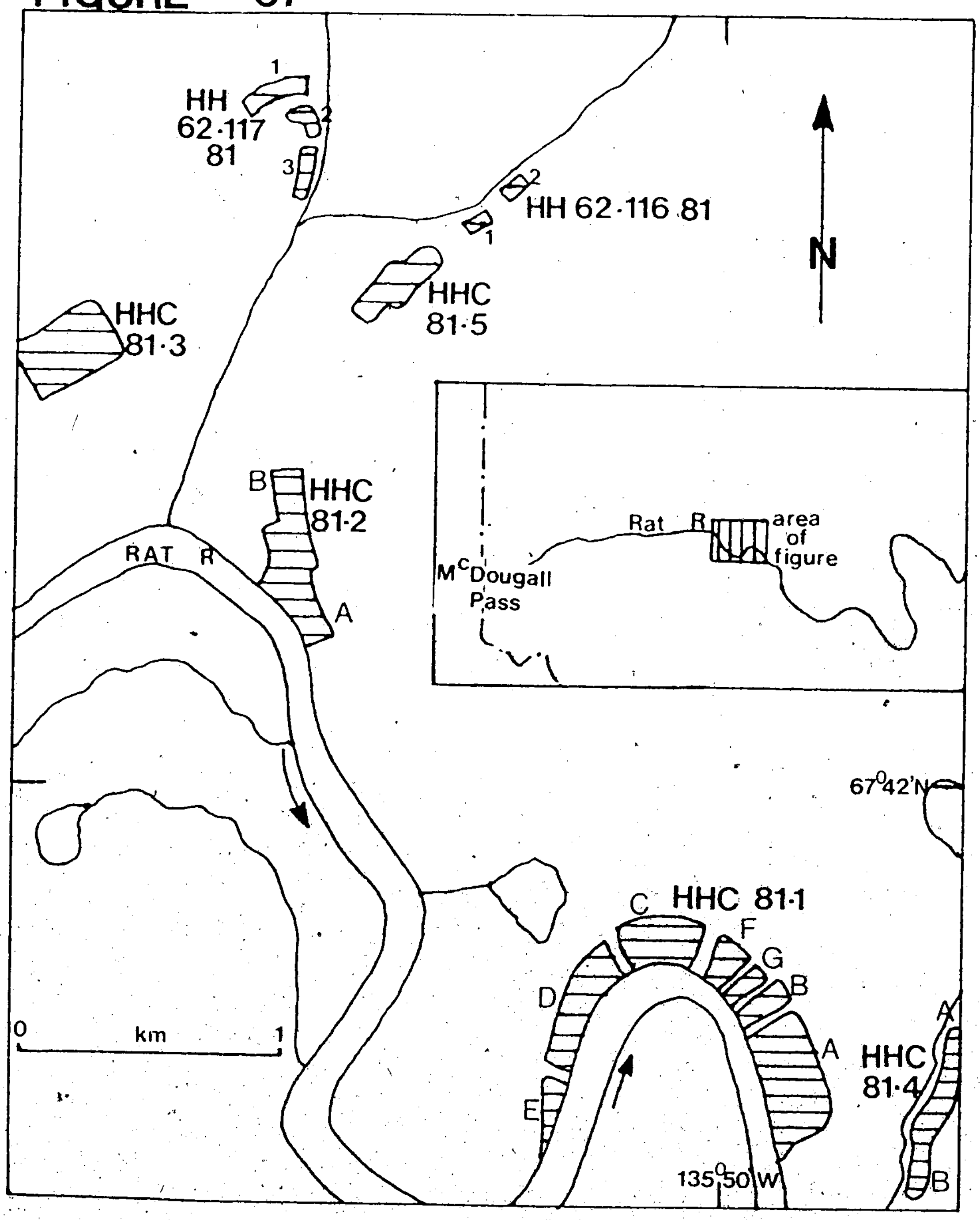
Two stages of lacustrine sedimentation have been identified in the upper drainage basin of the Rat River. The initial stage is represented by deposits of clay, silty clay, and clayey silt exposed at the bases of sections HHC 81-1b, 1c, 1d, 1f, and 1g along the Rat River (Figure 67). The sediments are exposed to 200 m asl elevation, where the unit is truncated by fluvial strata. The maximum thickness of the sediments of the initial lacustrine stage is 11.3 m.

Description

Figure 68 illustrates a composite section of the lacustrine sediments. The basal unit is composed of structureless clay and silty clay, stratified into horizontal sub-units 20-30 cm thick by silty partings. The clay is plastic, slippery, and contains 20-33% interstitial ice (by volume). Discontinuous, irregularly curved or sinusoidal laminations of clay 1-5 cm in length and 0.1- 0.5 cm in width are scattered throughout the strata. The laminations are oriented with the eastern part convex-up, and the western part concave-up. The laminations are invariably composed of finer sediment than the enclosing silty clay, and are internally structureless.

The silty partings between clay layers are 1-5 cm thick, and often incorporate a basal fine sand layer one or two grains thick. Clasts of Rapitan Group jasper-hematite and iron sulphide minerals are present in the sand horizons.

FIGURE 67



SECTIONS, UPPER RAT R. VALLEY

Figure 68

Lower Lacustrine Sequence, Rat River Valley

A composite section of the Lower Lacustrine sequence is illustrated. The sequence consists of a basal clay unit, with silt partings and swing-dash laminations, grading upwards to finely laminated clay; structureless silt; and a diamicton complex, exposed only at HHC 81-1C, 1D, and 1G. See text for further discussion.

LOWER LACUSTRINE SEQUENCE, RAT R. VALLEY

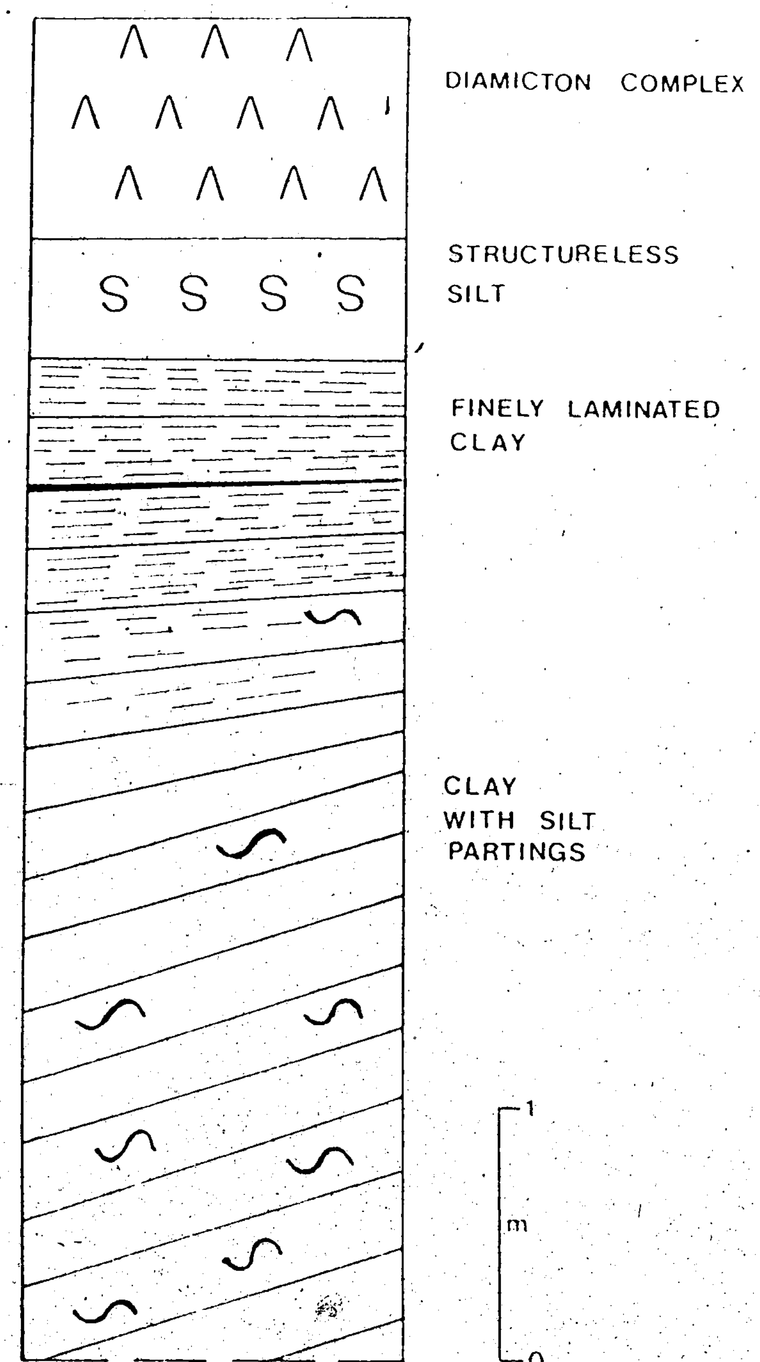


FIGURE 68

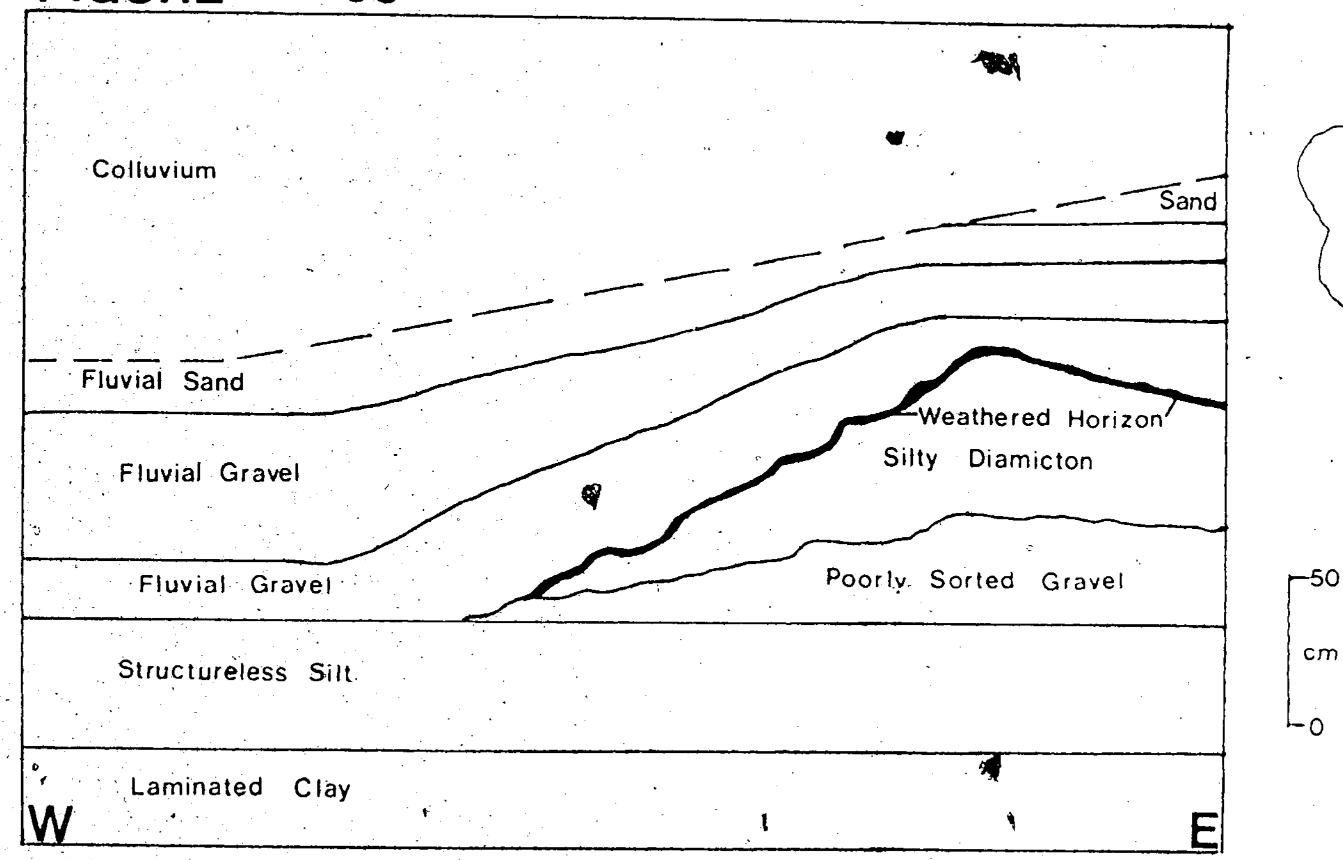
The parting planes strike at 220-227 azimuth, and dip at 10° to 12° W.

The central part of the lacustrine unit is similar to the base, but the clay layers show more stratification. The clays are finely stratified with silty laminae 1 mm thick, spaced at intervals of 0.5 to 1.5 cm. The silt laminae are generally horizontal and planar, although wedge-shaped laminae dipping to the west at 2° - 4° were also observed. Swing-dash laminations are present, but are less common than in the basal part. The contact between the basal and central parts of the unit is gradational.

The uppermost unit is composed of structureless silt, prone to liquifaction and flowage. The contact between the silt and the underlying clay is sharp and planar, but no indication of erosion or subaerial exposure of the clay was detected. Laminations in the clay are not truncated by the contact surface. Clasts of granitic rock, garnet, sphene, apatite and kyanite are present in the sand fraction.

The silt is overlain along a gradational contact at section HHC 81-1c by poorly sorted silty fluvial gravel, which in turn is overlain along an erosional contact by a silty diamicton (Figure 69). At sections HHC 81-1d and 1g, the diamicton overlies stratified clay of the central part of the lacustrine unit. The diamicton is internally structureless, displays a random fabric, and is irregular in shape. At 81-1c, it is wedge-shaped, thickening towards the east. At 81-1d and 1g, the diamicton is 23-25 cm thick, with

FIGURE 69



LOWER LACUSTRINE SEQUENCE, HHC 81·1C

erosional lower and upper contacts.

The upper 1-3 cm of the diamicton is weathered and iron-oxidized The carbonate content of this zone is less than that of the remainder of the diamicton. Fine vertical jointing is present. The lacustrine sediments and the diamicton are unconformably overlain by fluvial sands and gravels.

Interpretation

i)Clay and Silt Strata

The transition from clay layers separated by silty partings (basal and central units) to structureless silt (upper unit) could be produced either by progradation of the sediment source or by progressive shallowing of the lake. Initially, the lake was dominated by sedimentation from suspension of fine silt and clay, producing the structureless beds. The silty partings between the beds represent downslope flow of material from an eastern source, as indicated by the westward dip of the silt layers. Progressive decreases of flow velocity with time in individual events are indicated by the grading from fine sand to fine silt in the strata. Flow velocities of 10 cm/sec or less would be sufficient to transport this fine material (Sundborg, 1956; Shields; as modified by Blatt et al., 1972). These data suggest that the silt partings represent deposition from slow, low-density turbidity currents directed towards the west. The fine silt and clay units were deposited from suspension during the intervals separating turbidity current events.

Similar sequences in lacustrine and marine sediments have been interpreted as "distal" turbidites (eg. Kuenen, 1951; Piper, 1978; Sturm and Matter, 1978; Stow and Bowen, 1980; Chough et al, 1984; Shiki and Yamozeki, 1985). The units may be considered "distal" in the sense that they were formed under the influence of weak and waning currents. The topography of the valley and surrounding area, however, suggests that deposition may have occurred at a relatively short distance from the source of the flows. The westward flow direction is opposed by the regional slope, and a flow initiated from a sediment source to the east would rapidly lose energy and dissipate as it moved upslope. The most probable position for the initiation of the flows, therefore, is in the immediate vicinity of the site, where local west-trending lake bottom slopes opposed to the regional slope permitted downslope movement of the flows.

The consistency of spacing of the silty laminations may indicate that the flows were periodic, although no independent evidence to support this interpretation exists. Periodic flows could be generated by regular fluctuations in sediment loading and current patterns in the source area, such as would be induced by seasonal runoff. Regularly-spaced coarse and fine sediment layers (rhythmites) in "proximal" lacustrine situations, or in those characterized by higher sediment load in the water, are commonly interpreted to represent seasonal fluctuations in the flow (eg. DeGeer, 1912; Hughes, 1965; Gilbert, 1975).

Shallowing of the lake basin or progradation of the sediment source is indicated by the increasing silt content. In a shallow, sediment-rich lake, normally graded turbidite strata would not develop, because the supply of sediment to the lake bottom would be continuous. Such lakes tend to produce massive strata (Gilbert and Shaw, 1981). The continuous sediment rain would also effectively mask the activity of bottom currents, making the preservation of ripple forms and swing-dash laminations unlikely. This transition is indicated by the change in sediment character between the lower and central units of the lacustrine succession. It is probable that shallowing and progradation were both occurring during the final stages of lacustrine sedimentation. Gradual dissipation of the lake waters is indicated by the gradational transition from lacustrine silts to fluvial gravels visible at HHC 81-1c.

ii) Silty Diamicton

The silty diamicton is interpreted to be a subaerial debris flow deposit, probably produced by retrogressive thawing of sediments similar to the stratigraphically underlying lacustrine silts and clays, with admixed fluvial gravel. The similarity of the clast mineralogies of the diamicton, underlying fluvial gravel and overlying fluvial sediments (Table 25) indicates that the clast assemblage of

Table 25

Fluvial sediments and diamicton, HHC 81-1, clast lithology and mineralogy.

A) Pebbles

Lithology	Poorly sorted fluvial gravel[1] %	Diamicton[2] %	Overlying fluvial sediments[3] %
orthoquartzite	65	70	66
feldspathic sandstone	10	10	11
siltstone/ shale	6	6	4
chert	3	2	2
argillite	10	11	12
jasper-hematite	< 1	< 1	< 1
limestone	< 1	< 1	< 1
dolomite	< 1	< 1	< 1
granite	2	1	2
diabase	< 1	0	< 1
metaquartzite	< 1	< 1	< 1
gneiss	< 1	< 1	< 1
gabbro	< 1	< 1	< 1

[1]mean of 6 analyses

[2]mean of 4 analyses

[3]mean of 10 analyses

Table 25 (continued)

B) Coarse sand (0.42 - 1.00mm), major minerals.

Lithology	Poorly sorted fluvial gravel[1]	Diamicton[2]	Overlying fluvial sediments[3]
	%	%	%
quartz	51	50	53
feldspar	27	30	27
argillite/ siltstone	6	10	9
chert	4	3	5
calcite	1	1	1
hematite	< 1	< 1	< 1
dolomite	< 1	< 1	< 1
hornblende	3	2	3
lignite	1	< 1	< 1

[1]mean of 6 analyses

[2]mean of 4 analyses

[3]mean of 10 analyses

Table 25 (continued)

C) Coarse sand, trace minerals.

Lithology	Poorly sorted fluvial gravel	Diamicton	Overlying fluvial sediments
Apatite	3	2	6
Biotite	4	3	7
Chlorite	5	4	8
Chromite	3	1	3
Corundum	4	3	8
Epidote	6	4	10
Garnet	2	1	3
Gypsum	6	4	9
Ilmenite	2	1	3
Kyanite	3	2	6
Magnetite	4	3	8
Muscovite	2	1	4
Pyrite	1	0	1
Pyroxene	3	4	6
Rutile	3	2	6
Serpentine	2	1	3
Sphene	6	3	8
Tourmaline	3	2	5
Vivianite	2	0	1
Wollastonite	3	3	7
Zircon	1	0	0

the diamicton was derived from local Quaternary sediments. This implies that the deposit was not formed as a result of englacial or supraglacial activity. The random fabric and lack of internal structure are typical of modern retrogressive thaw-flow deposits throughout the area, including several developed from lacustrine sediments along the Rat River.

iii) Summary

The sequence of events at the HHC 81-1 localities commenced with lacustrine deposition by turbidity current flow. Shallowing of the lake and progradation of the sediment source produced a transition from laminated fine sediment to massive silt. Drainage of the lake was followed by fluvial activity. At a later time, the lacustrine and fluvial sediments were covered by retrogressive thaw debris flow deposits. Renewed fluvial activity followed this event.

The presence of granitic and metamorphic mineral clasts in the massive silt subunit, and the presence of Rapitan Group jasper-hematite and iron sulphides in the sandy horizons of the laminated strata, indicate that the lacustrine sediments post-date the first glaciation of the area. Maintenance of a lake in the Rat River valley requires the presence of glacial ice to the east. It is therefore considered that the early lacustrine phase in the Rat River valley was produced by impoundment of water during a glacial event, prior to the advance responsible for the formation of the second lacustrine phase.

The relative times of formation of the early lacustrine sequence and the fluvial assemblage exposed at HH 62-107 cannot be definitely assessed at present. The necessity for an ice dam during the period of lacustrine sedimentation, however, suggests that this event may have followed the development of the palynological assemblages.

E. Fluvial Sediments, Upper Rat Valley

Fluvial sediments overlie the lower lacustrine sequence at locations HHC 81-1c and 1d (Figure 69). The sediments are granite-bearing pebble gravel with a poorly sorted sand matrix containing quartz, feldspar, chert, hornblende, sphene, magnetite, fluorite, corundum and lignite (Table 25). The maximum thickness of the units at HHC 81-1c and 1d is 80 cm. Correlative fluvial sediments are present at location HHC 81-1a, to the east, and location HHC 81-2, to the west (Figures 67 and 70). At both these locations, the fluvial sediments are overlain by sediments of the later lacustrine phase (discussed below).

The gravels were deposited in a primary stream channel. The direction of flow was easterly, as indicated by the imbrication of the large clasts. No dateable material or palynomorphs were recovered from the strata.

Figure 70

Stratigraphy of Principal Exposures, Upper Rat River Valley

The stratigraphic relationships among the 8 main exposures in the Rat River Valley are depicted. See text for discussion.

FS--Fluvial Sediments, post-dating the Lower Lacustrine sediments and Diamicton

HD--Holocene Colluvium and Disturbed Material

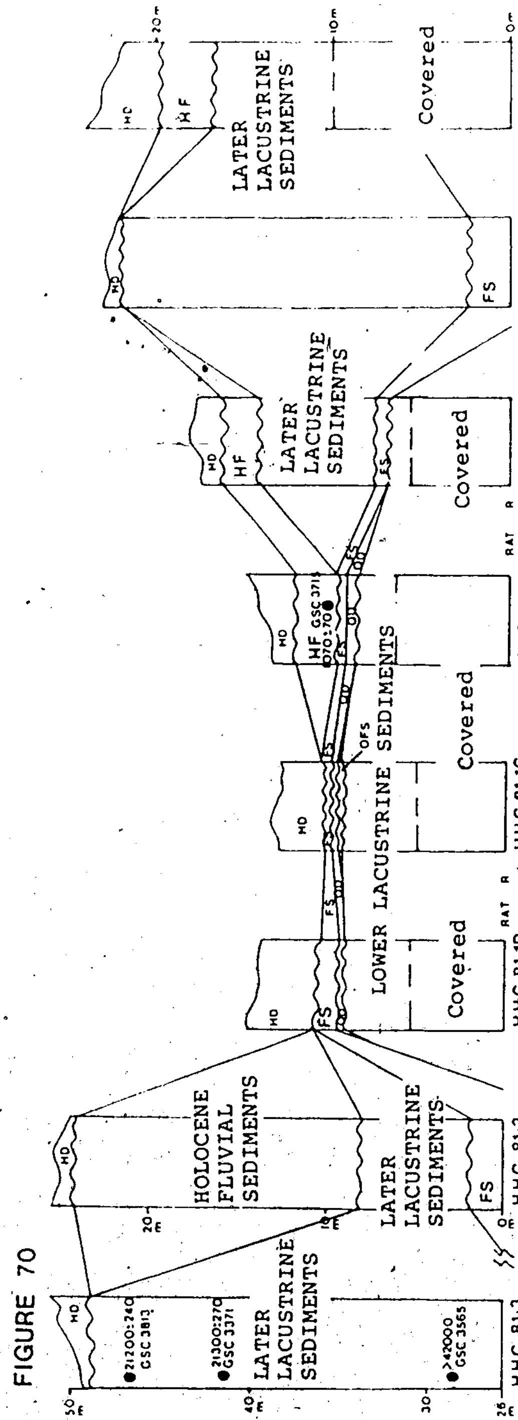
HF--Holocene Fluvial Sediments

OFS--Fluvial Sediments, pre-dating the Quaternary Diamicton and post-dating the Lower Lacustrine Sediments

QD--Quaternary Diamicton Complex

.--Location of ^{14}C dated sample

FIGURE 70



HD
21200±240 GSC 3813
21300±270 GSC 3371
LATER LACUSTRINE SEDIMENTS
>42000 GSC 3565
HHC 81-3
HOLOCENE FLUVIAL SEDIMENTS
LATER LACUSTRINE SEDIMENTS
FS
HHC 81-2
LOWER LACUSTRINE SEDIMENTS
Covered
HHC 81-1D
HHC 81-1C
HF
HHC 81-1G
HHC 81-1B
HHC 81-1A
HHC 81-4

F. Glacigenic Sediments and Landforms

Till

Exposures of till suitable for detailed sedimentologic investigation were not detected in the Rat River drainage basin, and none contained more than a single diamicton. Pebbles and granules identified in several exposures of colluviated diamicton included granite, gneiss, granodiorite, basalt, schist, arkose, and Rapitan Group jasper-hematite, in addition to a variety of sedimentary clast types derived predominantly from the Rat River drainage basin (Table 26). The heavy mineral assemblages were notable for containing igneous and metamorphic minerals such as fluorite, apatite, corundum, spinel, zircon, wollastonite, and xenotime. Fragments of serpentinite and serpentinized peridotite were also present in some samples. The clasts are fresh in appearance.

Although these exposures are unsuitable for the sedimentologic characterization of the till of the Rat Valley, the assemblages of clasts do provide information concerning the provenance. The presence of ultramafic clasts suggests that the glacier which originally transported this material flowed from the Bear Province of the Canadian Shield to the east-southeast. The nearest source of ultramafic rock to the study area is the Muskox complex east of Great Bear Lake (Baragar and Donaldson, 1973).

Table 26

Glacigenic diamicton, Rat River Valley, clast lithology and mineralogy.

A) Pebbles and granules (10 analyses)

Lithology	Maximum	Mean	Minimum
	%	%	%
orthoquartzite	88	68	45
feldspathic sandstone	40	10	3
siltstone	10	3	1
chert	6	2	0
argillite	35	10	5
jasper-hematite	1	< 1	0
limestone	3	1	0
dolomite	< 1	< 1	0
granite	2	< 1	< 1
diabase	2	< 1	0
granodiorite	< 1	< 1	< 1
lignite	1	< 1	0
basalt	< 1	< 1	0
peridotite	1	< 1	< 1
schist	< 1	< 1	0
serpentinite	< 1	< 1	0
gneiss	< 1	< 1	< 1

Table 26 (continued)

B) Coarse sand (0.42 - 1.00mm), major minerals (10 analyses)

Lithology	Maximum	Mean	Minimum
	%	%	%
quartz	68	50	37
feldspar	45	29	13
argillite/ siltstone	27	13	6
chert	11	4	2
calcite	1	<1	<1
dolomite	<1	<1	0
hornblende	6	2	<1

Table 26 (continued)

C) Coarse sand, trace minerals.

Mineral	Number of analyses containing mineral
Actinolite	2
Apatite	10
Biotite	10
Chlorite	10
Chromite	1
Cordierite	4
Corundum	9
Epidote	10
Fluorite	10
Garnet	4
Gypsum	6
Ilmenite	7
Jasper	10
Kyanite	8
Lignite	6
Magnetite	10
Muscovite	8
Olivene	1
Pyrite	3
Pyroxene	4
Rutile	6
Serpentine	7
Sphene	10

Table 26 (continued)

Mineral	Number of analyses containing mineral
Spinel	6
Staurolite	3
Tourmaline	5
Vivianite	1
Wollastonite	8
Xenotime	6
Zircon	7

Glacigenic Landforms

Isolated drumlinoids are present in the McDougall Pass area (Figure 62). The orientation of the features indicates west-southwestward glacial flow. Drumlinoid forms indicative of westward ice flow are also present in the Peel Plain southwest of Arctic Red River.

Discontinuous kame terraces are present along the eastern margin of the Aklavik Range, in the vicinity of Mt. Goodenough (Hughes, 1972). Irregular small kames are present in the McDougall Pass area, and in the vicinity of Nerejo Lake. Minor eskers are also present in the Nerejo Lake area (Figure 62).

A major meltwater channel system, parallel and 25-30 km west of the Peel River, marks a glacial stillstand position (Figure 62). Exposures of glacigenic sediment are not present along the meltwater channels. The clast lithology and mineralogy of till exposures east and west of the stillstand position are similar (Table 26).

This position was considered to mark the Late Wisconsinan maximum by Highes (1972) and Rampton (1982) (see Figure 7). Hughes <u>et al</u> (1981) considered the position to represent a recessional phase of the Late Wisconsinan glaciation. The absence of suitable exposures for investigation prevents direct assessment of the age of the glacigenic deposits west of the stillstand position.

Regional Significance

The similarity of the mineralogical and lithological assemblages through the Rat valley and the absence of exposures with multiple diamictons suggest that the till in the Rat valley was produced by a single glacial event. A subsurface diamicton exposed at HH 62-107-81-1 (67°39' N, 135°28' W), considered to be a till by Hughes (1972), was re-investigated by Hughes and the author in 1981 and is now interpreted as a cryoturbated soil horizon, as discussed previously.

If the diamicton exposures throughout the Rat Valley and in McDougall Pass do represent a single glacial event, the fresh nature of the clasts and the capping position of the sediment on the Peel Plateau (Figure 62) suggest that the glaciation occurred relatively recently. This conclusion supports a young age for the maximum glacial event, as suggested by Hughes <u>et al</u> (1981). The exposures studied, however, do not provide any data which can be used to estimate a date for the initiation of the last glaciation of the region. The only dated material underlying till in the region, autochthonous organic matter from HH 62-107-81-2a, is >43000 B.P. (GSC-3359). Consequently, no definite conclusions can be drawn from the till exposures of the Rat River Valley.

G. Later Lacustrine Phase, western Rat Valley

Description

In the western Rat valley, a later phase of lacustrine sedimentation is represented by sediments exposed at HHC 81-1,2,3, and 4 (Figure 67). These sediments overlie the early lacustrine and fluvial sequences (Figure 70). Till units were not observed in these outcrops.

A composite section of the lacustrine sediments present in the western Rat Valley is presented in Figure 71. The principal units are fine sand-clay sequences, a sand-silt-clay complex, and a thick couplet sequence.

Fine Sand-clay Sequences (unit A)

The basal unit of the later lacustrine phase is represented by fining-upward sequences of fine and silty sand, silt, and clay with interbedded medium-fine sand lenses (Figure 72). Each sequence is terminated along a planar erosional contact.

The basal fine sand and silty sand members contain planar cross-laminations, outlined by shale fragments and lignite. The lignite clasts are concentrated near the bases of the laminations. Within the fine sand units, the cross-laminations have erosional upper and lower surfaces and dip at angles of 8-12°. The angle of dip along the length of each lamination surface is constant. These cross-lamination packages are 3-5 cm thick. The flow direction inferred from the lamination orientations is

Figure 71

Upper Lacustrine Sequence, Rat River Valley

The three principal sedimentary units of the Upper Lacustrine sequence are:

a) Fine Sand - Clay sequences, with ripple cross-lamination and draped ripples indicating eastwardly flow;

b) A Sand-Silt-Clay complex, composed of fining-upward sequences with erosional contacts. Sand members contain ripple cross-laminations. A gravel bed is exposed at one section, HHC 81-3;

c)A Couplet sequence, composed of normally-graded coarse (silt, sandy silt, or silty sand) and fine (clay, silty clay, or clayey silt) members. Rip-up clasts, load structures, and isolated trough cross-laminations are present. Reverse graded couplets are also present throughout the sequence.

See text for further discussion.

Sand

Silt

Clay

Ripple Cross-Laminations

Rip-up Clasts

Diapiric Loading Structures

Recumbent Drag-Folds

Isolated Trough Cross-Laminations

UPPER LACUSTRINE SEQUENCE, RAT R. VALLEY

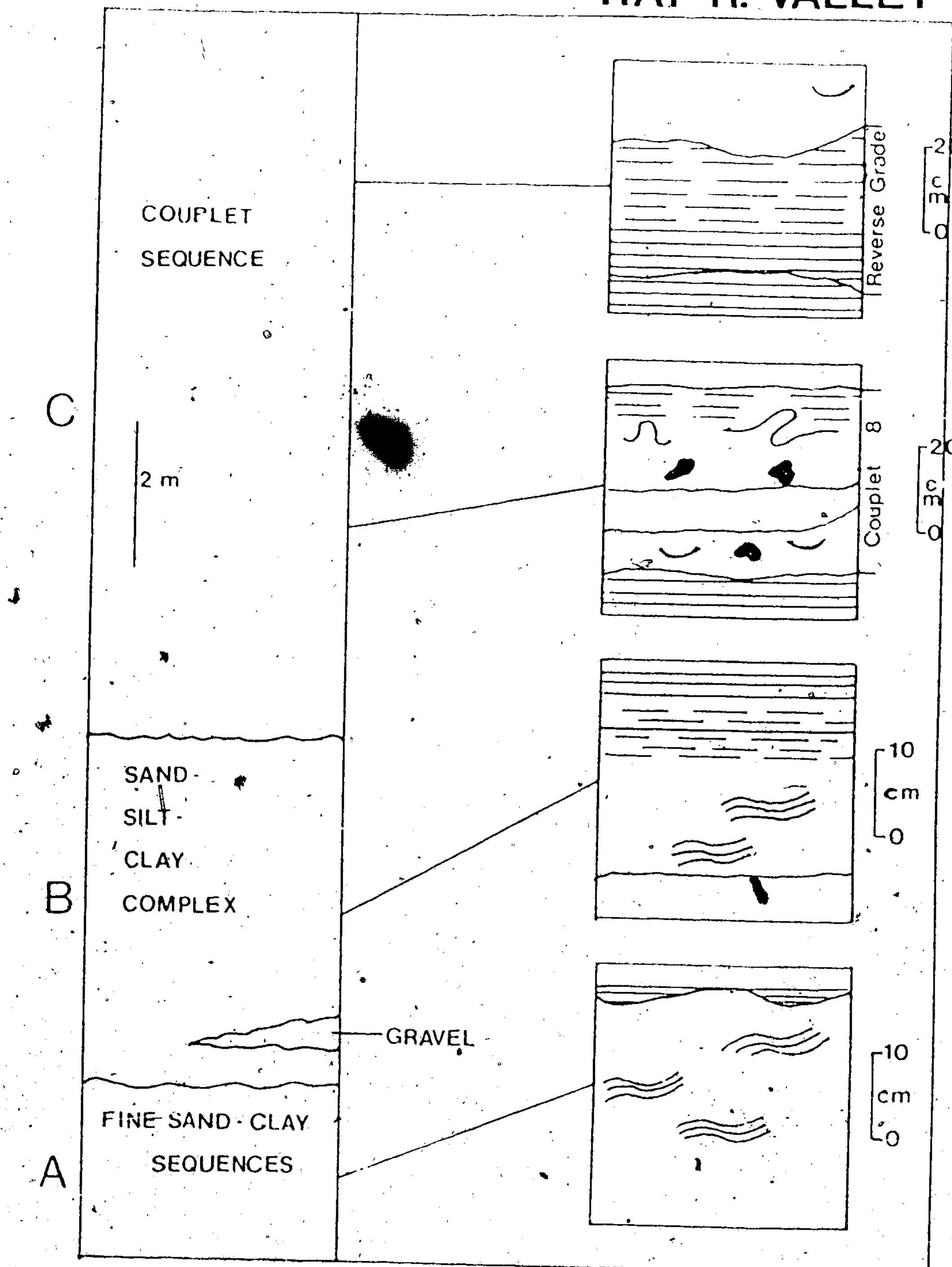


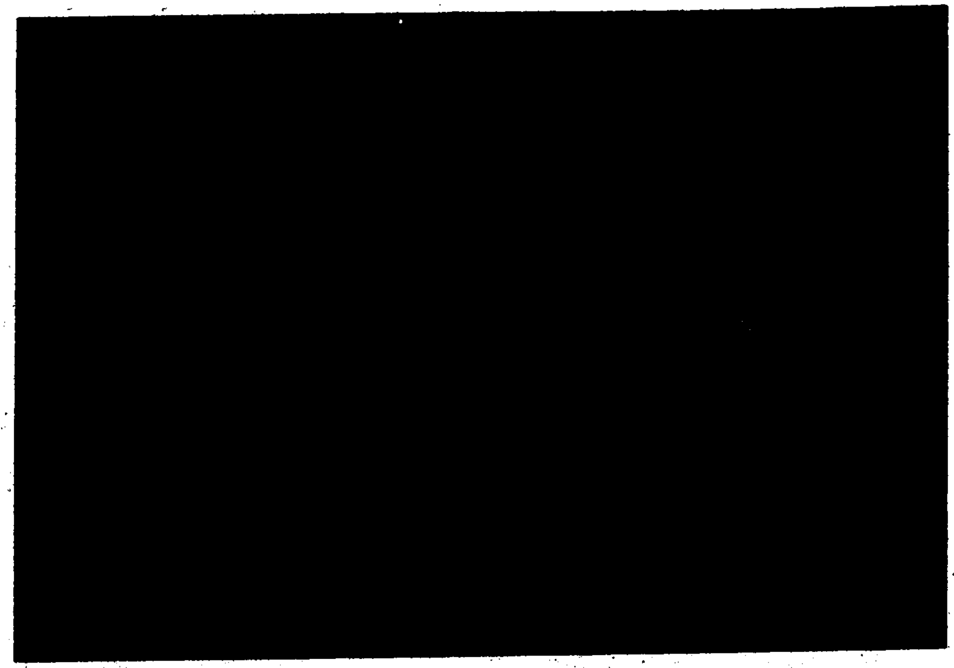
FIGURE 71

Plate 33 (top)

Contact between fluvial sand (bottom) and Later Lacustrine Clay (top), HHC 81-2. The clay contains fragments of Valvata shells.

Plate 34 (bottom)

Contact between fluvial sand (bottom) and Later Lacustrine Laminated Silt (top), HHC 81-1b. See text for discussion.



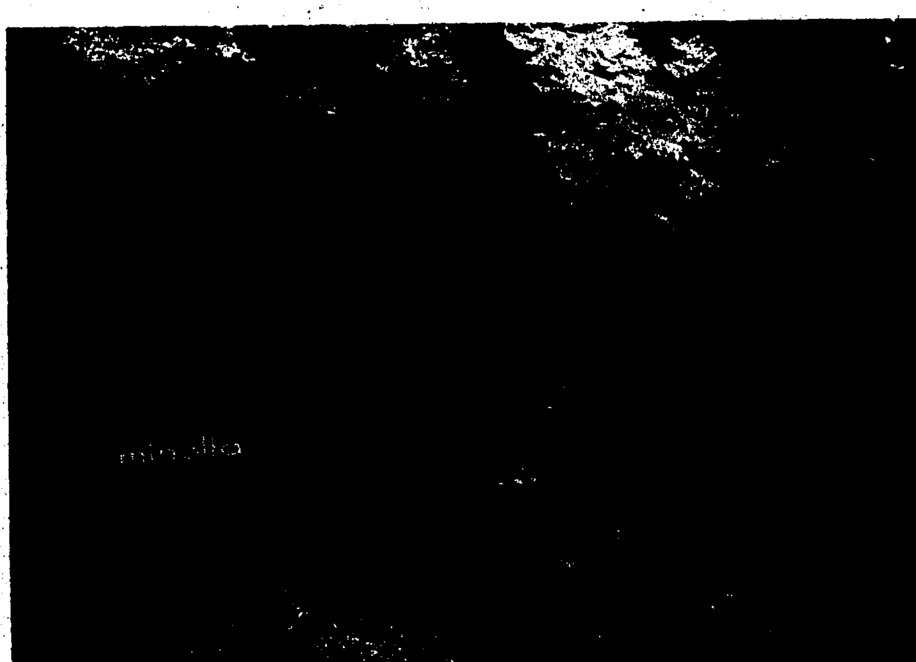


Figure 72

Typical Bedding Sequence, Unit A, Fine Sand - Clay Sequences, Upper Lacustrine Sequence, Rat River Valley

The sequence consists of beds of fine sand, silt and clay. Contacts between sand beds are generally erosional, and the finer sediments are overlain unconformably by sand units. Thin lenses of clay are present along the upper contact of some sand units. See text for further discussion.

Fine Sand

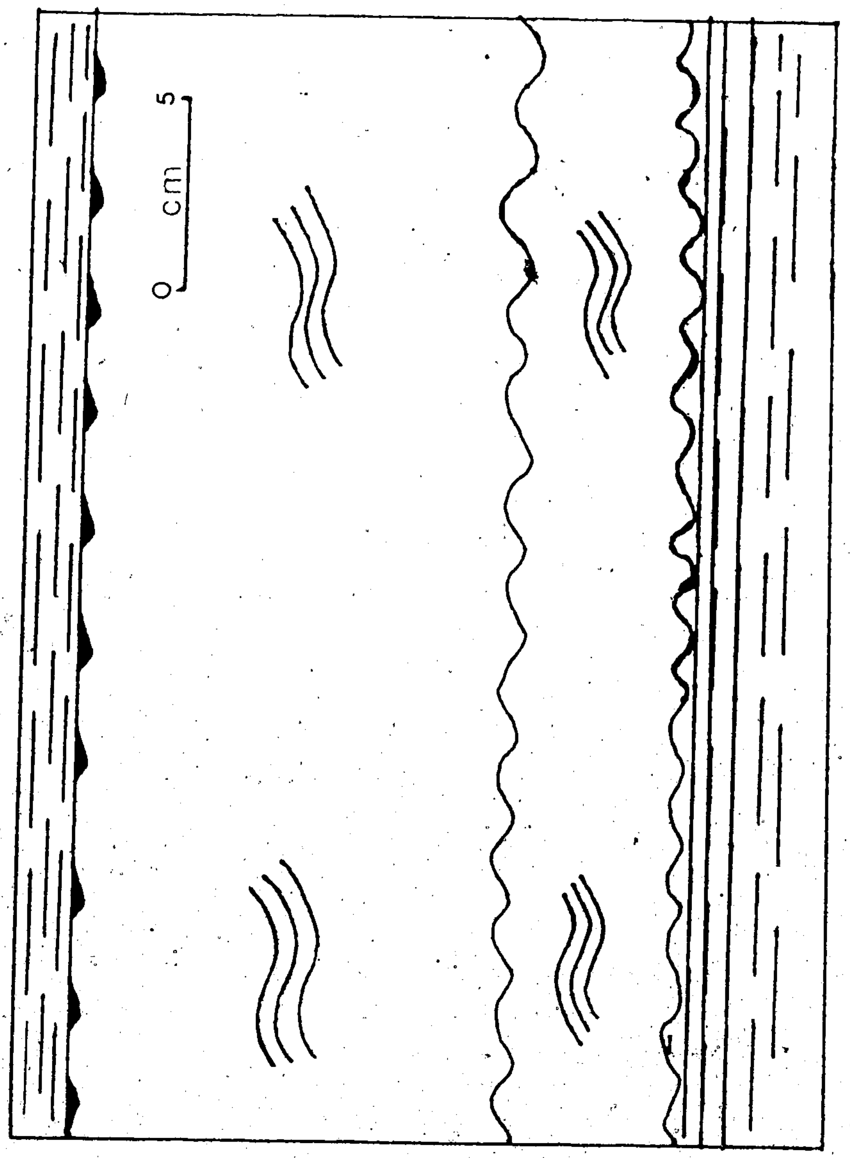
Silt

Clay

Clay Lenses

Ripple Cross-Lamination

FIGURE 72



generally easterly (070-090). Each package of laminations displays a single flow direction. In the medium and fine-grained sand lenses, poorly defined cross-laminations with erosional upper contacts and dips of 5°-8° also suggest easterly flow.

The upward decrease in grain size from the fine sand to silty sand and sandy silt is accompanied by structural changes in the cross-laminations. In the silty units, the cross-laminations display curved surfaces (asymmetrical and sinusoidal), with dips varying from less than 1° to 10°. The bases of successive lamination packages are progressively offset towards the east, indicating that the bedforms were climbing and drifting in that direction. The packages are 1-3 cm in thickness, and have lengths of 5-10 cm. Lignite clasts and micaceous minerals are not present on the crests of the curved laminations. Although the inferred palaeocurrent is easterly, fluctuation among individual packages between 070 and 115 is common. Within a single stratum, the orientation of the laminations varies laterally between 30°and 56° azimuth. Each set of cross-laminations was deposited by a single current, and the variation in orientation of successive beds indicates that the current direction was not consistent through time.

Contacts between the sandy silt and fine silt and clay members of the sequences are gradational at some locations and abrupt at others. Where the contacts are abrupt, the lower surfaces of the clay and fine silt units are irregular, but the bases along the southwest sides are commonly inclined at a steeper angle (5°-10°) than are those along the northeast sides (2°-5°) (Figure 73). This configuration suggests that the clay and fine silt were deposited from suspension and were draped over ripples developed in the underlying silty sand and sandy silt. The covering sediments are much finer (modal 7.0 - 9.0 phi) than the underlying sediments (modal 5.8 - 4.6 phi). The fine silt and clay is commonly present only above the low points in the sandy silt's upper surface. Above the higher points and around the crests of the bedforms, the sandy silt is directly overlain by fine sand along a planar erosional contact.

In locations where the sandy silt/fine sediment contact is gradational, it is marked by a gradual decrease in texture to medium silt (to mode 5.0 phi) and the absence of cross-laminations. This medium silt may be overlain by either fine silt and clay or may be overlain along an erosional contact by fine sand of the overlying sequence.

Sand-Silt-Clay Complex (unit B)

The fining-upward sequences are succeeded vertically by a complex consisting of coarse, medium, and fine sand units, and silt and clay strata (Figure 74, Plate 35). A consistent vertical pattern is not present, but relationships between the textural members are predictable.

The sand, silt and clay beds associated with the complex are commonly bounded by erosional contacts. Although

Figure 73

Clay Lenses, Unit A, Upper Lacustrine Sequence, Rat River Valley

The form of the clay lenses suggests that they were produced by the infilling of hollows between ripples developed in the underlying sand and silt. See text for further discussion.

Legend as for Figure 71.

FIGURE 73

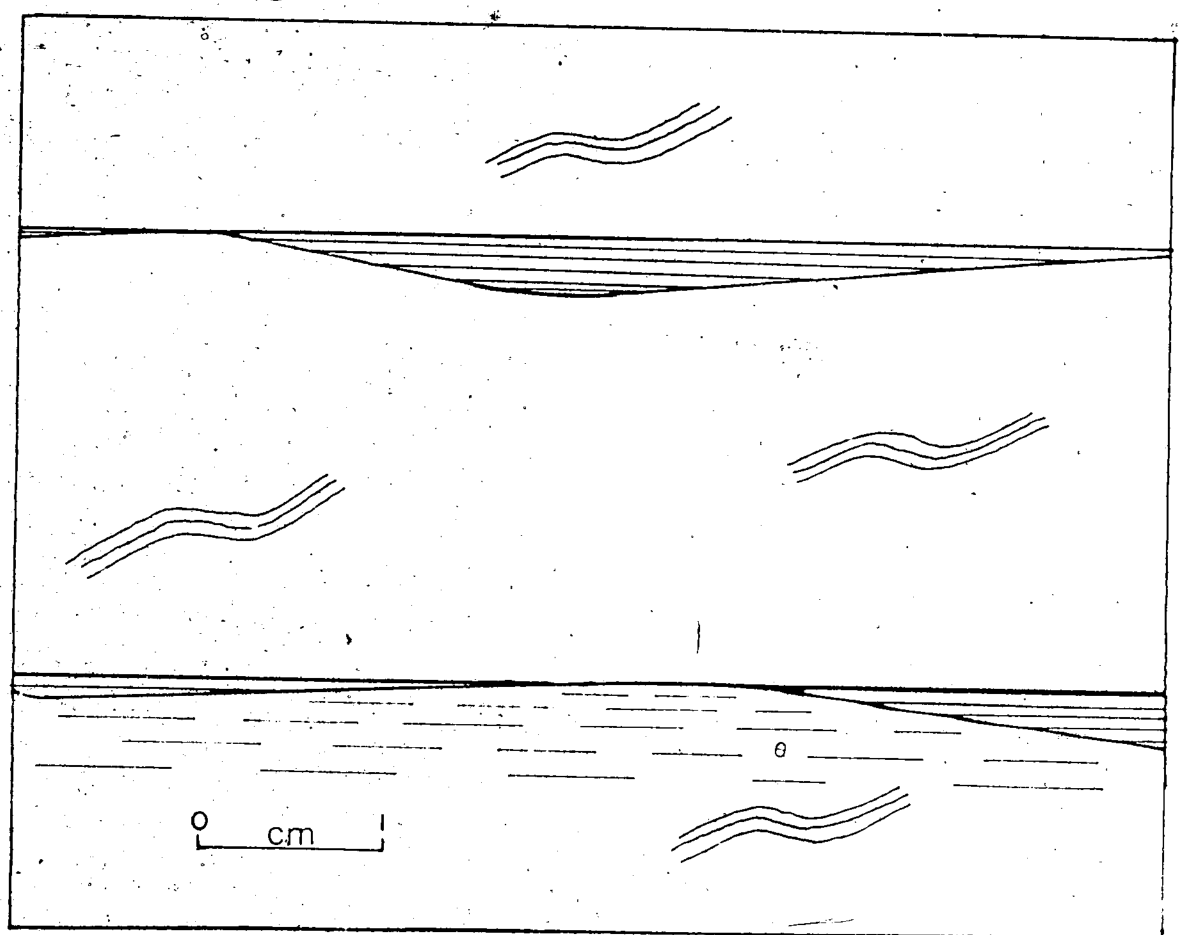


Figure 74

Typical Bedding Sequence, Unit B, Upper Lacustrine Séquence, Rat River Valley

Unit B is characterised by sand beds which grade upwards into silt and clay units. The basal contacts of the sand beds are commonly erosional. Textural fluctuations between silt and clay occurs in the upper parts of the sequences. See text for discussion; see also Plate 35.

Legend as for Figure 71.

FIGURE 74

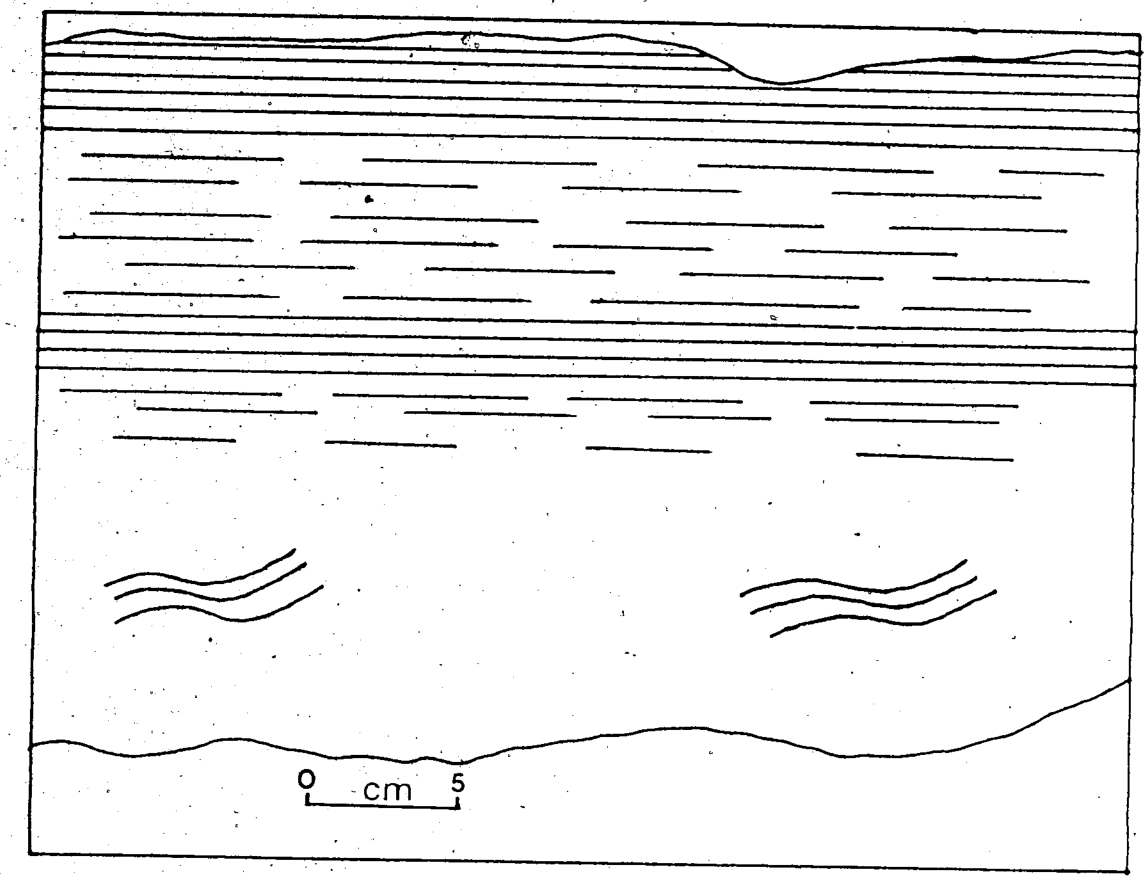


Plate 35

Later Lacustrine Sediments, Unit B, HHC 81-3. The sequence is dominantly composed of sand beds (light brown), grading upwards to silt and silty clay (grey). Thicker fine-grained units (base of photograph) are composed of alternating bands of clay and silt, and may appear homogeneous when observed in outcrop.

the units are arranged in different vertical successions, fining-upward sequences consisting of sand abruptly overlain by silty sand, silt and/or clay can generally be recognized. In this discussion, the texturally-distinct components will initially be considered separately.

The basal sand members are moderately to well-sorted, coarse to medium-grained (0.74 phi - 1.76 phi median diameter). and vary in thickness from 7 to 78 cm. They are dominantly quartz, feldspar, chert, hornblende, biotite, magnetite, and siltstone fragments (Table 27). Trace amounts of apatite, chromite, wollastonite, actinolite, cordierite, staurolite, rutile, corundum, bornite, and covellite, as well as other heavy minerals, are also present. Talc schist fragments and vivianite granules and sand grains were also observed.

The coarse-grained sands are generally structureless, although indistinct horizontal laminations are present locally. The medium-grained sand units, however, commonly contain planar cross-laminations and/or horizontal laminations. The planar cross-laminations dip toward the east at angles of 8°-12°, with the dip remaining constant for each lamination. Both the upper and lower surfaces are erosional. The cross-lamination packages average 4 cm in vertical thickness. The orientation of the laminations is clustered within 15° azimuth for each stratum. The inferred flow direction is easterly. The horizontal laminations are spaced vertically 1-2 mm apart, and are defined by heavy

Table 27

Basal sand members, HHC 81-3, clast mineralogy.

A) Coarse and medium sand (0.25 - 1.00mm), major minerals (19 analyses)

Lithology	Maximum	Mean	Minimum
	%	%	%
quartz	60	48	33
feldspar	27	14	8
argillite/ siltstone	30	12	6
chert	16	10	5
calcite	2	<1	<1
biotite	2	1	<1
hornblende	8	4	2
magnetite	3	1	<1

Table 27 (continued)

B) Coarse sand, trace minerals (19 analyses).

Mineral	Number of analyses containing mineral
Actinolite	15
Apatite	18
Bornite	11
Chlorite	16
Chromite	15
Cordierite	13
Corundum	17
Covellite	8
Cuprite	3
Epidote	10
Fluorite	13
Garnet	6
Ilmenite	8
Kyanite	12
Lignite	13
Muscovite	8
Pyrite	6
Pyroxene	10
Rutile	13
Serpentine	8
Sphene	13
Staurolite	9
Talc schist	12

Table 27 (continued)

Mineral	Number of analyses containing mineral
Tourmaline	16
Vivianite	8
Wollastonite	12
Zircon	4

mineral mono- or di-layers. These laminations are less common than the cross-laminations, and where both types are present the horizontal laminations underlie the cross-laminations. Vertically oriented peardrop-shaped and spherical mud clasts, and fragments of Picea sp. twigs lacking bark, are also present in some strata.

Fine sand, silty sand, and (rarely) sandy silt form the bases of some fining-upward sequences and overlie the coarse and medium-grained sand members of the other sequences along erosional and/or abrupt contacts. The thicknesses of the fine sand units range between 3 and 83 cm. The grain size decreases upwards regularly in these units, from a modal median of 2.5 phi at the bases to 3.7 phi at the tops of the strata. The bulk mineralogy of these units is similar to that of the coarser sand members, but the heavy mineral assemblage is characteristically less diverse (Table 28). Oxide and sulphide minerals are commonly absent, and vivianite is not present.

Curved and sinusoidal cross-laminations are present in the lower parts of the fine sand and silty sand beds. Dip angles vary from less than 1° to 10° along the lamination surfaces. The asymmetric laminations have steeper eastern slopes than western slopes. Although the orientations of the laminations vary within each stratum, the general dip of the individual laminations and the inclination of the lamination sets indicate that eastward to southeastward flow prevailed during deposition. The cross-laminated packages have

Table 28

Fine sand members, HHC 81-3, clast mineralogy.

A) Major minerals (12 analyses)

Lithology	Maximum	Mean	Minimum
	%	%	%
quartz	60	49	30
feldspar	37	21	12
argillite/ siltstone	15	9	3
chert	15	11	6
calcite	3	2	1
biotite	1	< 1	< 1
hornblende	6	3	1
magnetite	< 1	< 1	0

Table 28 (continued)

B) Trace minerals (12 analyses).

Mineral	Number of analyses containing mineral
Actinolite	3
Apatite	6
Chlorite	8
Chromite	2
Cordierite	3
Corundum	6
Epidote	4
Fluorite	6
Kyanite	3
Lignite	10
Muscovite	6
Pyroxene	6
Rutile	4
Serpentine	6
Sphene	8
Staurolite	1
Tourmaline	6
Wollastonite	5

vertical thicknesses of 5 to 25 cm, and inclinations of 1° to 6° towards the west. The asymmetrically-curved laminations have amplitudes of 1.5-3 cm and wavelengths of 4-7 cm. The sinusoidal laminations are shallower (amplitudes 1-2 cm) and longer (5-10 cm).

The curved laminations grade upward into fine horizontal laminations defined by hornblende-biotite and shale fragment monolayers spaced at 2mm intervals. The biotite grains are concentrated in the troughs of the cross-laminations in the transitional zones between the lamination types. Locally, thick laminations (0.5 mm-1 mm) are preferentially enriched in biotite along the bases and contain more hornblende in the upper parts. Most laminations, however, are defined by a mixed monolayer of hornblende, biotite and shale fragments. The grain size within the monolayers varies from fine sand to medium silt, and the biotite grains are generally coarser than the median texture of the monolayers. The upper contacts of the fine sand, silty sand, and sandy silt units are invariably abrupt.

The fine silt and clay members of the sequence are characteristically structureless, although fine horizontal laminations defined by shale fragments and medium silt are present in some strata. Both structureless and laminated clay units tend to break in vertically-oriented shards (usually thicker at the top than the base) when dry, but are cohesive when moist. Fine silts are prone to flow and locally contain ice lenses up to 5 mm thick. Clay units are apparently devoid of lensose ice, and do not exhibit flow.

At one section, HHC 81-3, a gravel bed is contained within the complex (Figures 70, 71, and 75). It is 32 cm thick, is separated from the uppermost fining-upward sequence of the underlying succession by 51 cm of medium-grained sand, and has a basal elevation of 218 m. The gravel unit is moderately sorted, open-worked deposits of moderately rounded, bladed and discoid granules. Granite, Rapitan Group jasper-hematite, diabase, argillite, chert, feldspathic sandstone, and orthoquartzite granules are present in the stratum (Table 29). Thin irregular lenses of planar cross-laminated medium-grained sand, composed primarily of quartz, feldspar, and chert, are interbedded with the gravel. Trace amounts of metamorphic and igneous minerals, such as epidote, actinolite, kyanite, corundum, wollastonite, chromite, serpentine, bornite, and covellite, and talc schist fragments are also present. This mineralogy indicates that the sediment postdates the initial glaciation of the region. The planar cross-laminations in the sand lenses dip towards the east at angles of 8° to 12°, and have erosional upper and lower surfaces. These cross-lamination sets are 1-3 cm in vertical thickness. Stacked sets of cross-laminations were not observed. The orientation of the cross-laminations, and the imbrication of the granules in the gravel, indicate that the flow direction was easterly.

Figure 75

Gravel Bed, Unit B, Upper Lacustrine Sequence, HHC 81-3

The gravel bed is 32 cm thick, and is erosionally bounded by sands. Fragments of <u>Picea</u> wood are present in the gravel. See Figure 71 for the stratigraphic position of the unit. The gravel is discussed further in the text.

Granule Gravel

Sand

Ripple Cross-Laminations

P Location of <u>Picea</u> fragment, ^{14}C dated at >42,000 (GSC - 3565)

FIGURE 75

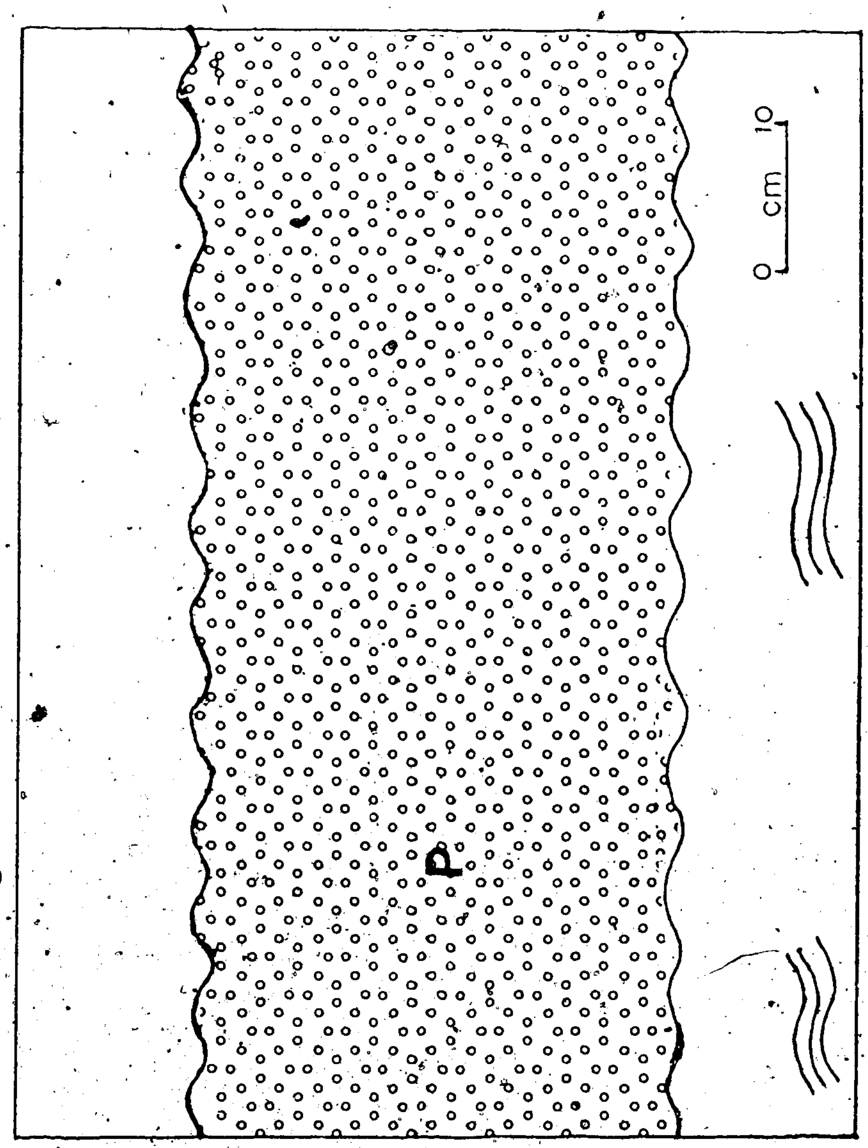


Table 29

Granule gravel bed, HHC 81-3, clast lithology and mineralogy.

A) Granules

Lithology	Mean of 3 samples
	%
orthoquartzite	38
feldspathic sandstone	11
siltstone/shale	6
chert	15
argillite	20
jasper-hematite	<1
limestone	7
granite	1
diabase	1
metaquartzite	<1

B) Coarse and medium sand (0.25 - 1.00mm), major minerals

Lithology	Coarse sand matrix[1]	Medium sand matrix[2]	Sand lens[3]
	%	%	%
quartz	50	43	41
feldspar	22	27	32
argillite/siltstone	15	8	6
chert	8	10	11
calcite	<1	3	3
hornblende	<1	3	3

[1]mean of 3 analyses

[2]mean of 3 analyses

[3]mean of 2 analyses

Table 29 (continued)

C) Coarse and medium sand, trace minerals (8 analyses).

Mineral	Number of analyses containing mineral
Actinolite	5
Apatite	4
Biotite	6
Bornite	6
Chlorite	4
Chromite	6
Cordierite	1
Corundum	8
Covellite	6
Cuprite	2
Epidote	8
Fluorite	5
Garnet	3
Gypsum	4
Ilmenite	3
Kyanite	6
Lignite	7
Magnetite	5
Muscovite	4
Pyrite	3
Pyroxene	6
Rutile	3
Serpentine	7

Table 29 (continued)

Mineral	Number of analyses containing mineral
Sphene	3
Staurolite	2
Talc schist	5
Tourmaline	4
Vivianite	4
Wollastonite	2
Zircon	2

The gravel stratum also contained several twigs and fragments of Picea sp. All of the fragments lacked bark, and were straight without branches. Ends were extensively abraded to rounded, smoothed forms, and the surfaces of the fragments were scored by longitudinal grooves 0.5 [illegible] 1.0 mm in depth. The largest fragment present, 1[illegible] cm long, 1.5 cm in diameter, and 11.5 g in mass, was ^{14}C dated at >42,000 B.P. (GSC-3565) (Figure 70).

The degree of abrasion evident on the surfaces of the Picea fragments, combined with their presence in a gravel unit within a lacustrine gravity flow sequence, suggests that the >42,000 date does not accurately reflect the sediment's time of deposition.

Couplet Sequence (unit C)

The gravel-sand-silt-clay complex is overlain by lacustrine sediments deposited as a series of couplets composed of a coarse member (silt, sandy silt, or silty sand) which grades vertically into finer sediment (clay, silty clay, or clayey silt) (Figure 76). The upper contact of each couplet is abrupt and erosional. The couplets tend to thin and coarsen with decreasing stratigraphic age. Commonly, the decreases in thickness observed at a particular section are primarily due to decreases in thickness of the fine members.

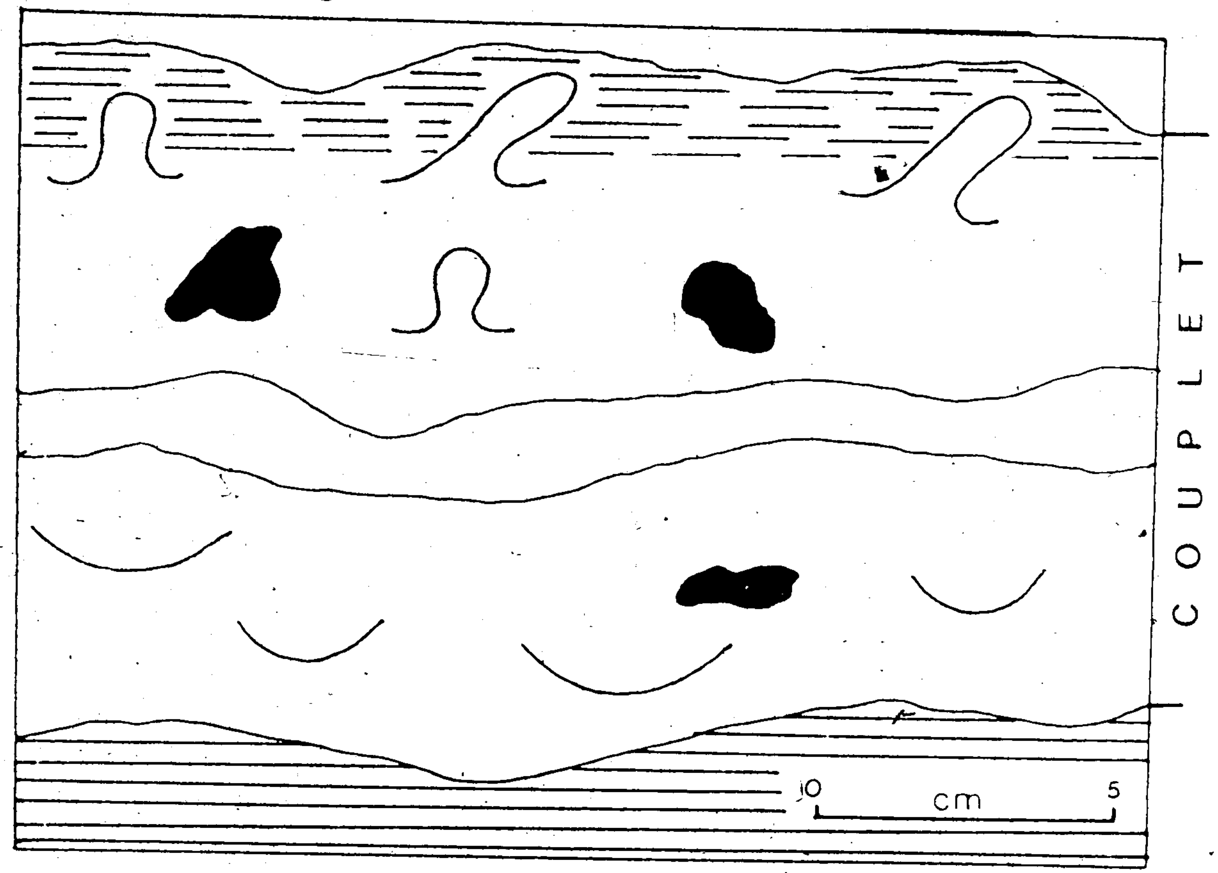
The sandy members are generally planar cross-laminated. The laminations dip at angles of 8°-12°, have erosional

Figure 76

Typical Bedding Sequence, Unit C, Upper Lacustrine Sequence, Rat River Valley

A typical couplet of unit C consists of a basal member of sand, containing rip-up clasts and isolated trough cross-laminations. Massive sand interbeds may also be present. The upper part of the sand member usually contains recumbent drag-folds and diapiric loading structures. The overlying silt or clay member of the couplet is generally structureless. Contacts between silt or clay members and the overlying sand member are erosional. See text for further discussion. Legend as for Figure 71.

FIGURE 76



lower contacts, and are spaced at intervals of 4-10 mm. Packages of laminations range in thickness from 1.5 to 8 cm. The upper contacts of the planar cross-laminations are gradational with sinusoidal or asymmetrically-curved cross-laminated sandy silt, horizontally-laminated sandy silt and silt, or massive sandy silt. Horizontally-laminated silt is draped over cross-laminated sands in some couplets. Orientations of the cross-laminations suggest eastward to east-southeastward flow, with variations in azimuth between 25° to 70° within individual members. Several coarse members contain more than one cross-laminated horizon. Isolated trough cross-laminations, also indicative of eastward flow, are also present in the basal parts of some coarse members. These trough cross-laminations are 1-4 cm deep, 2-7 cm wide, and are parabolic in outline. Generally, they are concentrated in the lower part of the couplet sequence.

Several sand members contain irregularly-shaped clay and silt clasts, 1-2 cm in length, 0.5-1 cm in depth, and 0.5-1 cm in thickness. The fine sediment clasts are not concentrated in distinct horizons, but are scattered throughout the sand units. The clasts are randomly oriented. Some clasts have been modified to quasi-sinusoidal forms (swing-dash forms), but most are completely irregular, without rounded or abraded edges.

Isolated granite and sedimentary pebbles and cobbles were present in some sand members. The stones are subrounded, and are oriented with the long axes vertical and the tapered ends (if present) pointing upwards. The underlying strata are commonly disturbed beneath the clasts to a depth of 2 cm or less. Overlying strata are draped over the clasts, while strata bordering the sides of the pebbles and cobbles are deflected around the obstructions.

The fine members of the couplets (silts and clays) are sporadically horizontally laminated, but generally are massive in freshly-exposed sections. Dried silt and clay members break in irregular shards. The fine units contain no clasts coarser than medium-grained sand.

Both fine and coarse couplet members display load casts, diapiric structures, spoon structures, flame structures, recumbent drag-folds, and _en échelon_ normal faults. All of these features are indicative of sediment loading by subaqueous gravity or current flow. The features are confined to the upper 2-5 cm of sand members, but occur throughout the finer members. Sand units which underlie silt and clay bands less than 5 cm thick are often deformed in combined structures with the fine sediment. The majority of the loading structures indicate that current flow was towards the east, although isolated examples suggesting southeast or northeast flow are present. Structures suggesting westward flow were not observed.

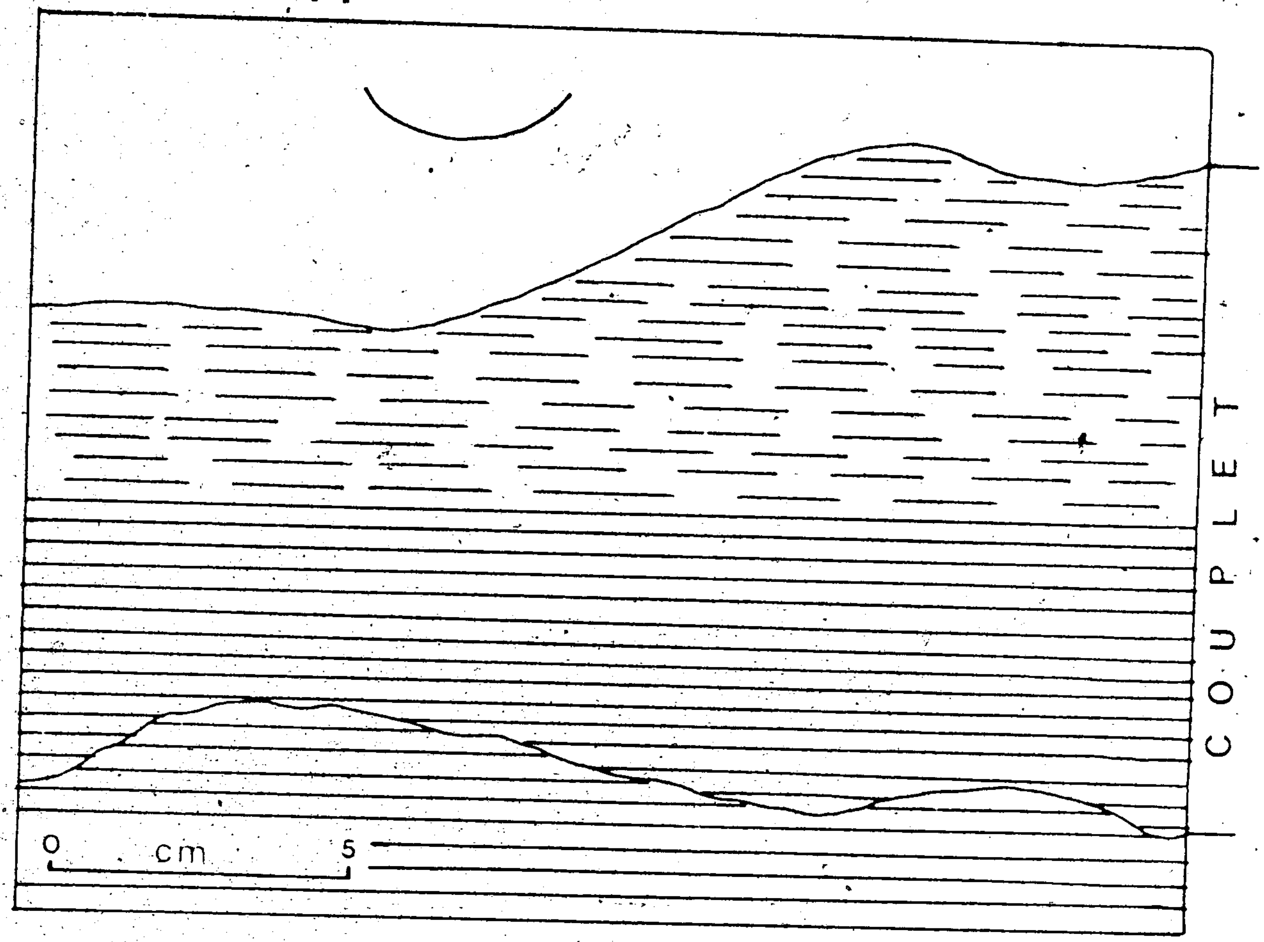
Reverse-graded couplets occur sporadically in the upper part of the sequence (Figure 77). The basal members of the couplets are structureless clay or clayey silt, moderately to poorly sorted, and devoid of sand-sized clasts. These

Figure 77

Typical Bedding Sequence, Unit C, Reverse Graded Couplet, Upper Lacustrine Sequence, Rat River Valley

Reverse graded couplets occur sporadically throughout unit C. A typical couplet consists of a basal member of clay, which grades upwards into clayey silt. The members are structureless. The upper and lower contacts of the reverse graded couplets are erosional. See text for further discussion. Legend as for Figure 71.

FIGURE 77



members grade upward into structureless or finely horizontally laminated medium silt, moderately to poorly sorted, containing 10%-35% sand. The contacts underlying the clay units and overlying the silt units are erosional. Reverse graded couplets occur singly or (very rarely) in groups of two. Where they overlie normally-graded couplets, the basal contact is invariably sharp.

Thin lenses and layers (1-2 cm thick) of organic silt are present in the uppermost parts of some sand members within the couplet sequence. The organic detritus is composed of very fine, fibrous and matted plant fragments. Admixed lignite and amber (0.01-1%) of Cretaceous and Tertiary age derived from the Eagle Plain Formation, exposed in the Bell Basin, are also present.

A sample, 81-33a, was obtained from the uppermost organic horizon exposed at HHC 81-3 in 1981, 50 cm below the top of the setion visible on 26 July 1981 (Figure 70). It yielded a date of 21,300 ±270 B.P. (GSC-3371). An additional sample. 82-78, was collected on 8 July 1982 from the uppermost organic layer exposed at HHC 81-3 at that time, located 4.3 m above the level of 81-33a and 1.2 m below the base of a colluvial layer. It yielded a date of 21,200 ±240 B.P. (GSC-3813).

The consistency and relative stratigraphic position of the ^{14}C determinations indicates that contamination by detrital lignite was probably not significant. The Rat valley was therefore occupied by lacustrine waters until at least 21,200±240 B.P.

Palaeoenvironmental Information

Palaeoenvironmental information from the lacustrine sediments is scarce. Assemblages of plant macrofossils and arthropod remains present in detrital organic layers are highly diverse, and represent mixtures of floodplain, tundra, and fell-field communities (J. V. Matthews Jr., Geological Survey of Canada, personal communication, 1983). These diverse assemblages contrast strongly with the paucity of palynomorphs. Lacustrine samples commonly contain only minor amounts of Picea, Gramineae, and Cyperaceae pollen, along with pre-Quaternary taxa. These data suggest that the assemblages were transported to the depositional site from the lakeshore by sublacustrine gravity flows and bottom currents. The scarcity of palynomorphs reflects the differences in hydrodynamic competence between the plant macrofossil/arthropod assemblages and the pollen.

Palynological assemblages present in the upper part of the couplet sequence suggest that low arctic tundra conditions were prevalent during deposition. The only sample with sufficient palynomorphs for counting was obtained from the organic layer dated at 21,200 ± 240 B.P. (GSC-3813) exposed at section HHC 81-3 (82).

The assemblage present is dominated by Cystopteris, Cyperaceae, Gramineae, and Betula (shrub type), with associated Equisetum, Sphagnum, Polystichum, Botrychium, and Artemisia (Table 30). Many of the spores were abraded indicating that some may have been derived from the Cretaceous bedrock of the region. The assemblage has been altered by aquatic transport biochemical corrosion, and degradation by fungi, limiting the ecological interpretation possible.

The spectrum suggests a low arctic tundra environment, with a claimate somewhat colder than that prevailing in the area today. The very low proportion of Picea type glauca pollen suggests that these grains were brought to the site by long distance aeolian transport, and therefore P. glauca was not native to the site. P. type mariana grains are also relatively rare, but the ability of this taxon to reproduce vegetatively makes conclusions about its presence or absence less certain.

Interpretation

The deposits are interpreted to represent successive sublacustrine gravity flows, combined with lacustrine bottom currents induced by winds and density stratification within the water mass. The predominant direction of flow was towards the east. Lake waters covered the entire upper Rat River valley throughout the sedimentation period.

Fine Sand-Clay Sequences (Unit A)

A) Planar cross-lamination formation and ripple development

Table 30

Palynological counts, Quaternary taxa, detrital organic horizon, section HHC 81-3 (82). All taxa included in pollen sum

Taxon	Percentage
Picea	7.0
Betula	15.6
Ledum	0.5
Shepherdia canadensis	0.5
Cyperaceae	17.6
Gramineae	13.7
Chenopodium	0.5
Artemisia	4.9
Cystopteris	19.5
Polystichium	5.9
Equisetum	9.3
Botrychium	3.9
Lycopodium	1.0
Sphagnum	6.3

The planar cross-laminations contained in the basal fine sand and silty sand members are similar to Allen's (1963a) type Mu cross-strata, formed by of the migration of asymmetrical ripples with essentially straight crests (Ashley et al, 1982). Erosion of the stoss side must exceed deposition, and hence the volume of sediment received from suspension must be less than the volume of the bedform. The mineralogical segregation of the lignite clasts, however, precludes direct classification of the cross-strata as type Mu.

The concentration of the lignite clasts at the base, and the erosive contacts, are characteristic of Jopling and Walker's (1968) type A ripple drift cross-laminations, formed by currents with low suspended loads flowing over non-cohesive beds. The absence of climbing structures in these laminations, and their planar depositional surfaces (as opposed to the concave-up surfaces recognized by Jopling and Walker) prevents direct assignment of the strata as type A, however.

The concave-upward surfaces recognized by Jopling and Walker (1968) may have been produced by migration of linguoid ripples, which has been shown to produce cross-laminations with concave-upward surfaces (Hamblin, 1961; Allen, 1963b). The absence of stoss-side preservation and the planar erosional bases, however, preclude classification of these structures as either Allen's (1963a) Kappa or Nu cross-strata, respectively. The cross-laminations are thus closely related to both Allen's (1963a) type Mu and Jopling and Walker's (1968) type A end-members, but are not identical with either.

Downcurrent climbing of the ripples is influenced by the rate of migration of material onto the slipface, and the amount of sediment supplied by bedload and suspension transport (Stanley, 1974). The absence of climbing structures in the basal strata indicates that the amount of sediment supplied was relatively low, and that supply was curtailed after a short period of time. The lack of fluctuation in current direction evident in individual cross-strata sets also suggests a short period of flow, without significant channelling induced by the developing bedforms. A short pulse would allow little time for bedform migration after initial construction, thus hindering the development of climbing ripple structures.

Hydrodynamic considerations suggest that the flow rate was very low. Estimates of average flow velocity based on the relationships presented by Sundborg (1956) and Shields (as modified and cited by Blatt et al 1972) suggest maximum values bwetween 15 and 20 cm/sec. Similar velocities were believed to be required for the initiation of fine sand rippling by Sorby (1908), Rees (1966) and Jopling and Walker (1968).

The bed thickness preserved necessarily represents a minimum height for the original bedform. Migrating bedforms, especially those propogating at high velocities, commonly produce cross-stratified beds with thicknesses less than the original bedform height (Rubin and Hunter, 1982). Although the inferred current velocity of 15 to 20 cm/sec is low, some of the upper lee side material was eroded during the transport and deposition processes.

The definition of the laminations by lignite and shale, rather than by heavy minerals, is due to erosion of the stoss side of the ripples, in addition to the low rate of flow. Along a typical fine sand ripple's surface, quartz and feldspar grains are transported by turbulent eddies from the separation zone in the trough to the crest of the next bedform. There, the grains are deposited on the crest, and roll down into the separation zone of the next ripple. Heavy mineral grains of equal diameter are also swept from the separation zones in the ripple troughs by turbulent eddies, but are redeposited on the upper part of the stoss side of the ripples because these grains are denser than quartz and thus require a more competent flow to be transported. This process results in an accumulation of heavy minerals on the upper stoss side of the ripples (McQuivey and Keefer, 1969). As this part of the ripple is subject to constant erosion, heavy mineral concentrations are not persistent. Instead, the heavy minerals are gradually incorporated into the lee-side laminations.

The low specific gravity of the lignite fragments (mean 1.24) results in hydrodynamic separation of the lignite from the remainder of the mineral assemblage during flow. The specific weight of the lignite in the flow medium is 6 to [illegible] times less than the specific weight of the bulk assemblage. Consequently, the lignite grains remain in transport during the initial stages of ripple formation, and are only deposited as the current velocity subsequently [illegible]. The shale fragments are somewhat coarser than the remainder of the clasts, and tend to be deposited at the base of successive laminations.

The cross-laminations in the the overlying silty sand units are transitional types, with some preservation of the stoss side and concave-upward surfaces on the lee side. They are intermediate between Allen's (1963a) types Kappa and Mu, and between Jopling and Walker's (1968) types A and B. Stanley (1974) reported similar cross-strata in Quaternary lacustrine silts in Nebraska, and inferred discontinuous deposition with low to moderate volumes of bedload transport. The presence of lignite clasts in the separation zones of the ripples indicates that draping of the silty sand occurred, and therefore the bedload concentration could not have been high. The transition from planar cross-laminations to curved-surface ripple-drift laminations indicates a gradual decrease in the ratio of suspended to traction load (Jopling and Walker, 1968; Ashley et al., 1982). This transition was probably produced by a decrease in flow velocity (and hence competence).

The silt units display cross-laminations similar to Jopling and Walker's (1968) type B (asymmetric) and sinusoidal ripple laminations (type S of Allen, 1973), and to Allen's (1963a) types Kappa (formed by linguoid ripples) and Lambda(formed by straight ripples). Similar type B ripples were interpreted by Stanley (1974) and Ashley et al., (1982) to represent continuous deposition on both stoss and lee faces. The transition between type B and type S ripples was ascribed to increasing bed cohesiveness and decreasing traction load by Jopling and Walker (1968), and to simple harmonic variation of flow velocity by Stanley (1974), Gustavson et al., (1975), Hunter (1977), and Ashley et al (1982) considered draped lamination to be the result of suspended sediment fallout, independent of bed cohesion. The relative textures of the sediments exposed in the Rat Valley outcrops, however, suggest that the infiltration of silt into a pre-formed ripple structure would greatly increase the ripple's resistance to erosion, and therefore cohesive effects cannot be completely disregarded for these particular ripples. The concentration of lignite and micaceous minerals in the troughs of the ripples, and the relative enrichment of heavy minerals on the crest and stoss, are consequences of their specific weights.

The dimensions of the asymmetrical ripples, and the texture of the incorporated materials, suggest flow velocities from 7-10 cm/sec (Sorby 1908; Rees, 1966). Current flow was influenced by the pre-existing bedforms, producing greater variations in cross-lamination orientation. In addition, irregularities induced in ripple alignment by pre-existing bedforms (especially in the case of linguoid ripples) caused the cross-lamination orientations to vary.

The sinusoidal (type S) ripples formed on cohesive bed surfaces under the influence of a minimal traction load (Jopling and Walker, 1968; Gustavson et al., 1975). Silt and clay settling out of suspension filled voids in the sandy silt structure, increasing bed cohesion and thus preventing erosion of the sinusoidal form. This increase in cohesion reduced the ability of the current to erode the bedform. Addition of a concentration of 20% of medium-fine silt to a ripple formed under the influence of a peak velocity of 10 cm/sec may produce a bedform capable of resisting erosion by currents twice or three times as rapid, according to the relationships proposed by Sundborg (1956). Consequently, if a sinusoidal ripple can be established and if the overlying water mass carries even a low suspended sediment load, preservation of the form is likely.

Sinusoidal ripple forms are also commonly associated with wave activity (eg. Campbell, 1966; Tanner, 1967; Boersma, 1970; Miller and Komar, 1980). Although sinusoidal ripples produced by wave activity tend to be symmetrical, because of the dominance of oscillatory motion, asymmetrical forms can also develop in nearshore environments (cf. P. Allen, 1981). The absence of biconvex forms, such as those described by Clifton (1976) and Allen (1981), and on a larger scale by Harms <u>et al</u>., (1975) and Dott and Bourgeois

(1982) indicates that high energy surf zone or storm generated conditions were not present.

The asymmetric, irregularly spaced, high amplitude ripples present in the sands suggest unidirectional current flow (Tanner, 1967). These characteristics are more evident when the effect of infilling of the troughs by suspension load material is discounted, as suggested by Tanner (1967). Since the stoss side laminations are preserved, the original bedform height was probably not significantly greater than the preserved bed thickness.

Asymmetric wave ripples are oriented towards the palaeo-lake shoreline (P. Allen, 1981). The topography of the upper Rat River valley indicates that the lake was linearly aligned along an east-west axis. For sites in the centre of the valley, the nearest shorelines would be to the north and/or south. The easterly flow direction indicated by the ripple cross-laminations is therefore incompatible with formation by shoaling waves. Although the prevailing winds are presently westerly (Lambert, 1968) and were probably funnelled along the valley during the lake's existence, the absence of a suitable topographic obstruction to the east of the sites eliminates the possibility of eastward-shoaling wave development.

The absence of shoreline or subaerial features in sections located centrally in the valley also militates against a swash-zone or wave interpretation. Wave ripples are not characteristically formed in water of greater depth than approximately one-half of the wavelength (cf. Lamb, 1945; Christopher et al., 1952; Smith and Sinclair, 1972).

The overlying fine silt and clay units were deposited by gradual settling of material from suspension. The draping of the ripple structures at some locations and the gradational contacts in others suggest that the suspended sediment load component was not derived from settling through the entire water mass, but resulted from settling of material carried in suspension directly above the bedforms. The thinness of the suspended sediment deposits indicates that little sediment entered the lake through overflows or interflows.

B. Formation of Sequences

The sand, silt, and clay sequences are interpreted to have formed from sublacustrine bottom currents flowing towards the east (Figure 78). The ripple structures indicate that the traction load carried by the currents was initially minor, and declined throughout the sedimentation process. The intimate association of traction load and suspension load deposits indicates that each sequence formed as a result of a single event. The absence of coarse material in the strata, and the lack of deep scouring and large rip-up clasts, suggests slowly-moving low-density fluid currents of low viscosity and moderate turbulence. Low density turbidity currents are the only form of subaqueous flow which meet the hydrodynamic requirements, as specified by Lowe (1979).

Figure 78

Postulated Glaciolacustrine Environment, Upper Rat River Valley

Lacustrine Currents

Surface Runoff

Sublacustrine Debris Flow Movement

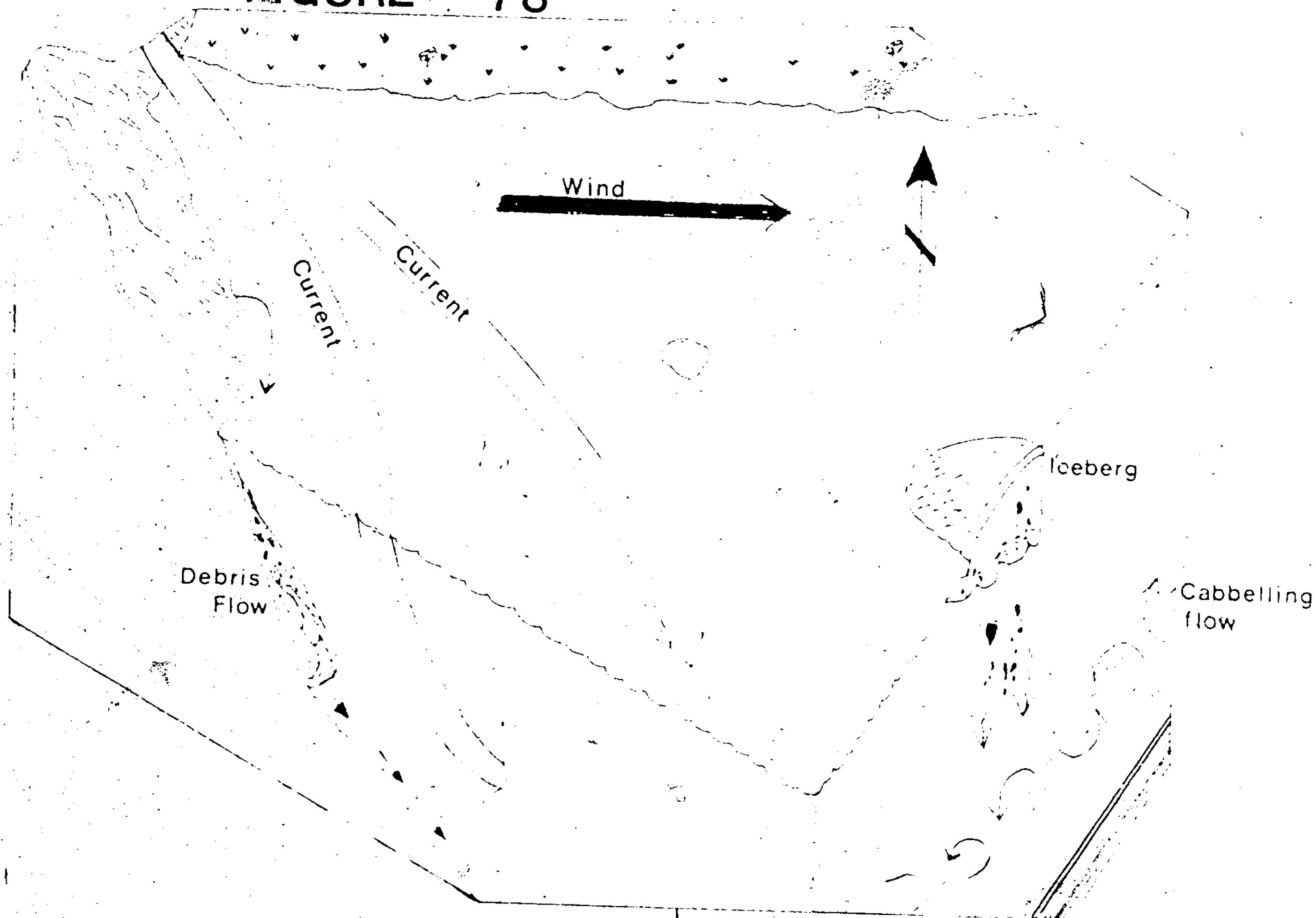
Rainout from Icebergs

Cabbeling Flow

Wind Direction

See text for discussion.

FIGURE 78



Turbidity currents have been frequently invoked to explain fining-upward sequences deposited in glacial lakes (eg. Kuenen, 1951; Banerjee, 1973; Gustavson, 1975; Shaw and Archer, 1978) and also in non-glacial lakes (eg. Lambert and Hsu, 1979). The sequence of ripple cross-laminated fine sand and silty sand overlain by massive or thinly laminated silt is similar to the Tc-Td couplets of distal turbidites recognized by Bouma (1962). Differences in ripple drift cross-lamination form can be attributed to variations in ripple morphology and fluctuations in the suspended sediment and traction loads. The transition from traction load to suspension load deposition is a common feature of low-density turbidity currents (Lowe, 1982).

Hydrodynamic calculations based on bed thicknesses (Sadler, 1982) and the inferred slope along the path of the currents (Komar, 1969; Normark and Dickson, 1976), utilizing the constraints on velocity and drag derived from the ripple structures and sediment textures indicate that the flows were diffuse but relatively thin, and were very low in traction and suspended sediment load. Increased slope angles due to isostatic depression of the eastern part of the basin may have allowed thicker, denser currents to flow.

No single mode of origin can satisfactorily explain all turbidity currents generated in the lacustrine environment. Suggested causes include sediment surges from fluvial influx (Lambert and Hsu, 1979; Pharo and Carmack, 1979), repacking of material on upper delta or fan slopes (Normack and

Dickson, 1976), failure of deltaic sediment or nearshore sediment (Fulton and Pullen, 1969; Shaw and Archer, 1978; Siegenthaler et al, 1984) or lake bottom sediment (Hyne et al, 1973; Shaw and Archer, 1978), bottom countercurrents induced by winds (Smith, 1978) and jokulhlaups (Shaw et al., 1978).

The orientation of the flows, parallel to slightly oblique to the axis of the Rat Valley, suggests that slumping of nearshore or shoreline sediment in the vicinity of the depositional site was not responsible for generating the turbidity currents. The absence of extensive deltaic deposits along the margins of the valley and in McDougall Pass similarily precludes deltaic influxes as major initiating mechanisms. Jokulhlaup flow influxes are not compatible with the inferred easterly direction of flows.

The three mechanisms most likely to generate turbidity currents are fluvial influxes from alluvial fans, failure of lake bottom sediment on moderate slopes, and bottom currents generated by wind activity on the surface. All three of these processes were probably active on the lake.

The prevalent winds in the area are westerly (Lambert, 1968) and currents generated in the shallowest part of the lake in McDougall Pass would tend to flow to the east, a direction co-inciding with the topographic slope (Figure 78). The presence of lignite derived from the Eagle Plain Formation in the sediments suggests that a connection existed between the lacustrine waters and the Bell Basin *via*

McDougall Pass, although individual flows initiated in the western Bell Basin are unlikely to have transgressed the height of land. Unconsolidated material on the eastern slope would be disturbed and incorporated into the flows. Other unconsolidated sediments and weathered fine clastic bedrock would also be susceptible to disturbance.

Underflows would be most significant during the early winter. Cold, dense water from the McDougall Pass area would flow along the bottom of the lake, as the larger water body gradually equilibrated with the atmosphere. The flow would be deflected by the Coriolis Force towards the right (south) side of the lake. Similar circulation patterns have been proposed for Kamloops Lake, B.C., by Pharo and Carmack (1979)and Carmack et al., (1979), and for Lake Algonquin by Gravenor and Coyle (1985).

As the temperature of the inflowing water continued to decline, the water mass in the main body of the lake would begin to mix with the eastward-flowing water to produce a dense basal layer. Mixing of the inflowing water and the main water mass commonly is initiated at the base of the main stratum (eg. Ivey and Boyce, 1982). When cold river water is involved, the process is termed "cabbelling" (Carmack et al., 1979). This dense layer of water effectively hinders underflow activity. Little contribution from fluvial sources would be expected during the winter, as precipitation is currently minimal from November to March (Burns, 1974).

In the spring, the water flowing from the McDougall Pass area would rapidly increase in temperature, lessening the density and effectively preventing underflows from this source. Turbidity current activity would be confined to slump-generated flows and those derived from local fluvial sources. In most environments, the reduction in the density of the inflowing stream water due to warming is greater than the effect of suspended and traction sediment load. Consequently, the stream waters are less dense than lake waters, and overflows result (Smith, 1978; Carmack et al., 1979). Sudden snowmelt could generate flash flooding, however, which would produce stream flows with concentrations of sediment sufficient to cause underflows and turbidity currents. In addition, fluvially transported sediment deposited in nearshore areas would be prone to reworking. Overflow conditions would prevail until autumn, when cooling of the inflow would again result in the generation of underflows and turbidity currents.

The majority of the sediments were therefore deposited in the late autumn and early winter, although isolated events triggered by slumping or nearshore erosion probably occurred throughout the year. The small amount of colloidally dispersed material retained in suspension during the initial early winter influx could be incorporated into the waning stages of turbidity current flow (as suggested by Azmon, 1981) or could remain in suspension indefinitely.

Sand Silt-Clay Complex (Unit B)

The overlying sediments share many characteristics with the underlying material, suggesting similarities in genesis. Many of the conclusions derived in preceding paragraphs are thus equally applicable to analysis of these sediments. This discussion, therefore, will focus on those units which differ significantly in characteristics and/or genesis from the strata previously interpreted.

A. Granule Gravel Bed

Two interpretations can be proposed for the well-sorted granule gravel bed exposed at HHC 81-3. The deposit could have resulted from an eastward-trending viscous grain flow, or from a sandy high density fluidized flow.

Grain flows are propagated by the dispersive pressures generated through clast-clast interactions (Middleton and Hampton, 1976; Lowe, 1976). When the concentration of debris is low, the grain flow moves as a viscous fluid, until friction along the base leads to stagnation or "freezing" of the flow (Nardin et al., 1979, Lowe, 1982). Gravel-dominated flows characteristically produce massive deposits with reverse grading at the base (Middleton, 1970; Lash, 1984), and clasts aligned parallel to the flow direction and imbricated upflow (Walker, 1975; Nardin et al., 1979).

The granule deposit at HHC 81-3 is strongly imbricated in the upflow direction, and the clasts are aligned parallel to the inferred current. The Picea sp. fragments are also aligned in an east-west orientation. Reverse grading of the clasts was not observed within the gravel bed, but the largest *Picea* sp. fragments were concentrated near the top of the unit. Although the gravel is underlain by medium-grained sand, the 51 cm thickness of the unit suggests that it is not a basal member of a grain flow deposit. Lowe (1976) has shown that sand grain flows require slopes at or near the critical angle of repose, and that the resulting flows are less than 5 cm thick.

The absence of fine clastic material and the presence of planar cross-laminated sands within the gravel bed are criteria not commonly associated with grain flow deposits. Fluidization or reworking by slow-moving currents could remove some of the fine material, but these processes would be expected to produce a deposit with an upwardly decreasing matrix concentration. This was not observed at HHC 81-3. The cross-laminated sands could not develop under true viscous grain flow conditions, and are likely to be preserved unaltered if extensive fluidization of the deposit occurred. If the grain flow activity was involved in the deposition of the granule bed, other processes must have modified the initial sediment distribution.

An alternative interpretation suggests that the granule bed and the underlying sand unit were emplaced by a sandy high density fluidized flow. Deposition from these flows occurs in four stages (Lowe, 1982). The initial phase is traction sedimentation, represented by low-angle planar cross-stratification or parallel stratification produced in made by migrating dunes and plane beds. Increasing flow unsteadiness leads to the concentration of suspended sediment near the bed, producing high dispersive pressures and developing a traction carpet. This layer is characterized by thin horizontal laminations, often with reverse grading. When the energy level declines in the unsteady flow, it stagnates or "freezes", producing a massive ungraded deposit commonly dominated by granules. Finally, sedimentation of suspended clasts on the now stagnant flow commences, producing massive or normally-graded sand units. Complete sequences are on the order of 1 m in thickness (Lowe, 1982).

The deposits at HHC 81-3 correspond in most details to the succession outlined by Lowe (1982). The medium sand beneath the granule layer is planar cross-laminated and the scale of the structures is compatible with dune development. Horizontal laminations are present in the upper part of the sand unit, suggesting a transition from steady traction sedimentation to unsteady flow. Deposition of the granule layer by traction carpet freezing is compatible with its well-sorted nature and well-defined imbrication. The cross-laminated sand strata within the granule bed indicate that tractional motion persisted in some parts of the flow. Possibly, the flow did not stagnate simultaneously along its entire base. Stagnation in one area would permit traction to continue in others for a short period, until the effect of the newly-created bed obstacle influenced the motion of flow. The granule bed is overlain by horizontally laminated sand, which could have settled out of suspension upon termination of the sandy fluidized flow.

4.5. Sand Strata

The presence of horizontally laminated coarse sand layers and heavy mineral cross-laminated medium sand strata is indicative of increased sediment concentration in the sublacustrine flows. Flows with high initial sediment concentrations are more proximal than those with smaller loads. The prevalence of curved and sinusoidal cross-laminations in the fine sand and silty sand units, however, indicates that the flows responsible were relatively low-density turbidity currents, although denser than the currents that formed the underlying fining-upward sequences.

The heavy mineral assemblages contained within the sand strata originated due to glacial transport of the clasts. The minerals could have become incorporated into the lacustrine sediment through reworking of pre-existing glacially-derived deposits, or could have been deposited on the lake bottom by rainout from calved ice or seasonal lake ice. No indication of westward sublacustrine currents originating from the ice front to the east was observed. The presence of vertically-oriented peardrop-shaped and spherical mud clasts suggests that some sediment rain-out did occur during the deposition of the unit. The shape and dimensions of these mud clasts are not compatible with extended transport (Smith, 1972).

C. Vivianite Grains

The rounded vivianite grains, and their concentration only in coarse and medium grained sand units, indicate a detrital origin for these clasts. Vivianite is usually formed in anaerobic sediments (Nriagu and Dell, 1974; Emerson and Widmer, 1978), although authigenic occurrences have been reported in phosphorous-rich aerobic sediments (Hearn et al., 1983), and in association with fossil wood (Reed, 1946) and bone (Zwann and Kortenbout van der Sluys, 1971). Formation of authigenic vivianite in sediments requires the presence of phosphate, from organic or inorganic sources. No inorganic phosphate-bearing strata have been identified in the Rat River Valley, and apatite is present only in trace amounts. The Picea fragments present in the sediment do not contain authigenic vivianite. Vivianite is present as detrital grains in sediments deposited prior to 43,000 B.P. in the lower Rat Valley (Section HH 62-107 81-2a, discussed previously). Vivianite grains reported in lacustrine sediments from the Bluefish Basin were probably formed in the lacustrine sediments, although many of the grains have been subjected to transportation. Authigenic vivianite associated with a caribou molar has been reported from the McDougall Pass area (Jorg Mattner, University of Clausthal, F.R.G., personal communication, 1985). The detrital vivianite grains in the coarse lacustrine sediments could therefore have been derived from detritus in pre-existing local deposits, could have been transported from the Bluefish Basin, or could have been derived from isolated local authigenic occurrences.

Couplet Sequence (Unit C)

The upper part of the lacustrine deposits is characterized by repetition of couplets of coarse and fine sediment. The scale and nature of the bedding structures are similar to those preserved in the ~~underlying~~ sediments, and therefore the mode of deposition was probably similar. The differences between the underlying sediments and the couplet sequence are the presence of additional current and load structures, the regularity of spacing of the couplet strata, and the incorporation of detrital organic lenses.

A. Current And Load Structures

The presence of trough cross-laminations in the bases of some coarse couplet members indicates dune development and thus higher energy than the ripple laminations. This increase in energy is also indicated by the presence of irregularly-shaped clay and silt rip-up clasts. The low durability of these clasts (Smith 1972) suggests that they were transported for very short distances. The flows were probably a series of short, rapid pulses, rather than continuous streams. Pulsating flow and irregular, sudden loading of underlying sediment is also suggested by the load casts, diapiric structures, spoon structures, flame structures, recumbent drag folds, and *en échelon* faults. All of these features commonly formed as a result of rapid sediment loading (McFee and Goldberg, 1969). The variations in orientation among the structures indicate fluctuations in current direction, which would be expected from discrete flow pulses. These features all indicate low-density turbidity currents, initiated by bottom or nearshore sediment slumping or seasonal water mass flow. The reverse graded couplets may have been formed by isolated sandy high density fluidized flows, which commonly are transformed laterally into turbidity currents (Lowe, 1982).

The isolated pebbles and cobbles were probably transported to the sites by icebergs and/or winter lake ice. The deformation structures beneath the pebbles and the vertical orientations are more readily explained by vertical rather than horizontal transport through the water mass. The subrounded nature of the pebbles and cobbles suggests glacial or fluvial abrasion. Clasts derived from avalanches or rock falls are generally angular (Luckman, 1975).

B. Couplet Spacing

The regularity of the couplet spacing and the progressive coarsening upward pattern suggest that the controls on the initiation of the currents were less localized. The pattern indicates progressive shallowing of the basin, either by progradation of deltas and alluvial fans or by declining water levels and receding of the lake shorelines from the north and south sides of the valley. The absence of extensive progradational fluvial sequences on the margins of the valley makes the latter hypothesis more probable.

C. Organic Detritus Layers

The presence of the organic detrital layers in the sediments may be due to hydrodynamic sorting of nearshore sediments (including plant and arthropod fragments) during the process of turbid flow. Nelson (1982) noted sequences of cross-laminated sand overlain by parallel-laminated coarse silt containing beds of plant fragments offshore of the Yukon River delta. He attributed the development of these sequences to storm-generated bottom currents, combined with storm-wave reworking and liquifaction of nearshore sediments. These sediment assemblages are very similar to those attributed to marine turbidity currents by Bouma (1962), although the turbidite sequences investigated by Bouma lacked plant fragments.

The sediments described by Nelson (1982) were deposited by sediment-laden currents which, despite their genesis by storm activity, are hydrodynamically apparently identical to the turbidity currents generated in deeper marine environments. The mode of genesis does not affect the classification of the event, and consequently these storm-derived flows can be classed as turbidity currents.

The presence of terrestrial organic detritus in the Rat Valley sediments indicates that nearshore material, probably initially transported by rivers and alluvial fan streams and sheetwash, and possibly carried by lake or river ice, was reworked in the sublacustrine environment. The flows could have beeen triggered by storm events. The orientation of the current indicators, however, indicates that the flows travelled along the axis of the valley, parallel to the prevailing wind direction. The general scarcity of pollen in the sediments is consistent with sediment deposition and influx during the early winter.

D. Jokulhlaup outflow?

Sand-silt couplets have been suggested as evidence for rapid drainage of glacially-dammed lakes (eg. Borell and Offenberg, 1955; Walker, 1967; Tauber, 1970). The "drainage varves" consist of a basal sand member, moderately to well-sorted, which may contain cross-laminations but is commonly horizontally laminated (Walker, 1967), overlain by finely horizontally laminated silt. The sequences may contain dewatering structures, and often show indications of slumping or creep. The upper surfaces of the silts are often weathered, dessicated, and scoured, and may be coated with loess or show signs of frost infiltration or ice wedging.

Rapid outflow from large glacially-dammed lakes produces channels and other erosional features in the emptied basin (Pardee, 1942; Malde, 1968; Hardy, 1976; Kehew, 1982). Similar features have been noted in drained reservoirs (Bishop et al., 1984). In the downstream areas, pre-existing channels may be deepened, or new channels incised in a dense network (Bretz et al., 1956; Baker, 1973), and the floodwaters often deposit stratified material in tributary valleys (Waitt, 1980; Atwater, 1984). The sudden drainage of smaller glacial lakes may have little or no impact on the previously deposited sediments in areas removed from the outlet (Liverman, 1981).

Chambers (1971) described sedimentary sequences developed as a result of the periodic drainage and refilling of Lake Missoula, Montana. The transitions from festoon cross-bedded sand and gravel to rippled sands to rhythmites ("varves" of Chambers) represented basin filling sequences, combined with delta progradation, rather than basin outflow. The weathered zones at the top of each varved unit are the only representation of the overflow period. These are overlain by fluvial sands and gravels, which are succeeded by wave-rippled cross-laminated sand, indicating re-establishment of the lake in the basin (Chambers, 1971).

The absence of large-scale erosional features, the lack of deep scouring along sand/clay contacts, and the absence of weathered zones or frost disturbance zones within the sedimentary sequence, all suggest that the central part of the basin remained under lake water throughout the sedimentation period, and that rapid, strong currents were not involved in the formation of the sequence. This conclusion does not preclude periodic partial drainage of the lake, but it does indicate that complete drainage along the base of an ice dam, as demonstrated for Lake Missoula by Bretz et al (1956), did not occur. The lake sediments therefore provide no direct information concerning the nature or mechanics of lake drainage, except to indicate that this drainage was never so complete during the period represented by the sedimentary succession as to expose the entire basin.

The pattern of sedimentation is most compatible with deposition from low-density turbidity currents generated by mass movement of nearshore and lake bottom sediments, by seasonal influxes of fluvially-transported material, and by water circulation patterns in the basin. The predominant flow direction was down the modern topographic slope, from west to east.

H. Later Lacustrine phase, Eastern Upper Rat Valley

Description

In the eastern part of the upper Rat River Valley, at sections HHC 81-1a and HHC 81-4 (Figure 67), a different style of lacustrine sedimentation is observed. Rhythmically-bedded medium sand strata 6-20 cm thick are separated by thin beds and laminae of sandy silt and silty clay (Figure 79, Plates 36 and 37). At section HHC 81-1a, 108 sand-fine sediment rhythmic cycles are present. The

Figure [illegible]:

Later Lacustrine Sediments, HHC 81 1a

The sedimentary succession at HHC 81 1a consists of 108 sand-fine sediment couplets deposited during the Later Lacustrine Phase, overlying fluvial gravel. The internal structures of a typical couplet, #48, are illustrated in the inset figure. The couplets dip to the west at angles of 12° to 17°. See text for further discussion; see also Plates 36, 37, and 38.

Medium Sand

Silt

Silty Clay

Type A (erosional-stoss) climbing ripples

Type B (depositional-stoss) climbing ripples

Teardrop clay ball

Spherical clay ball

Swing-dash Lamination

Shadow Structure

Ball-and-Pillow Structure

FIGURE 79 SEDIMENTS, HHC 81·1A

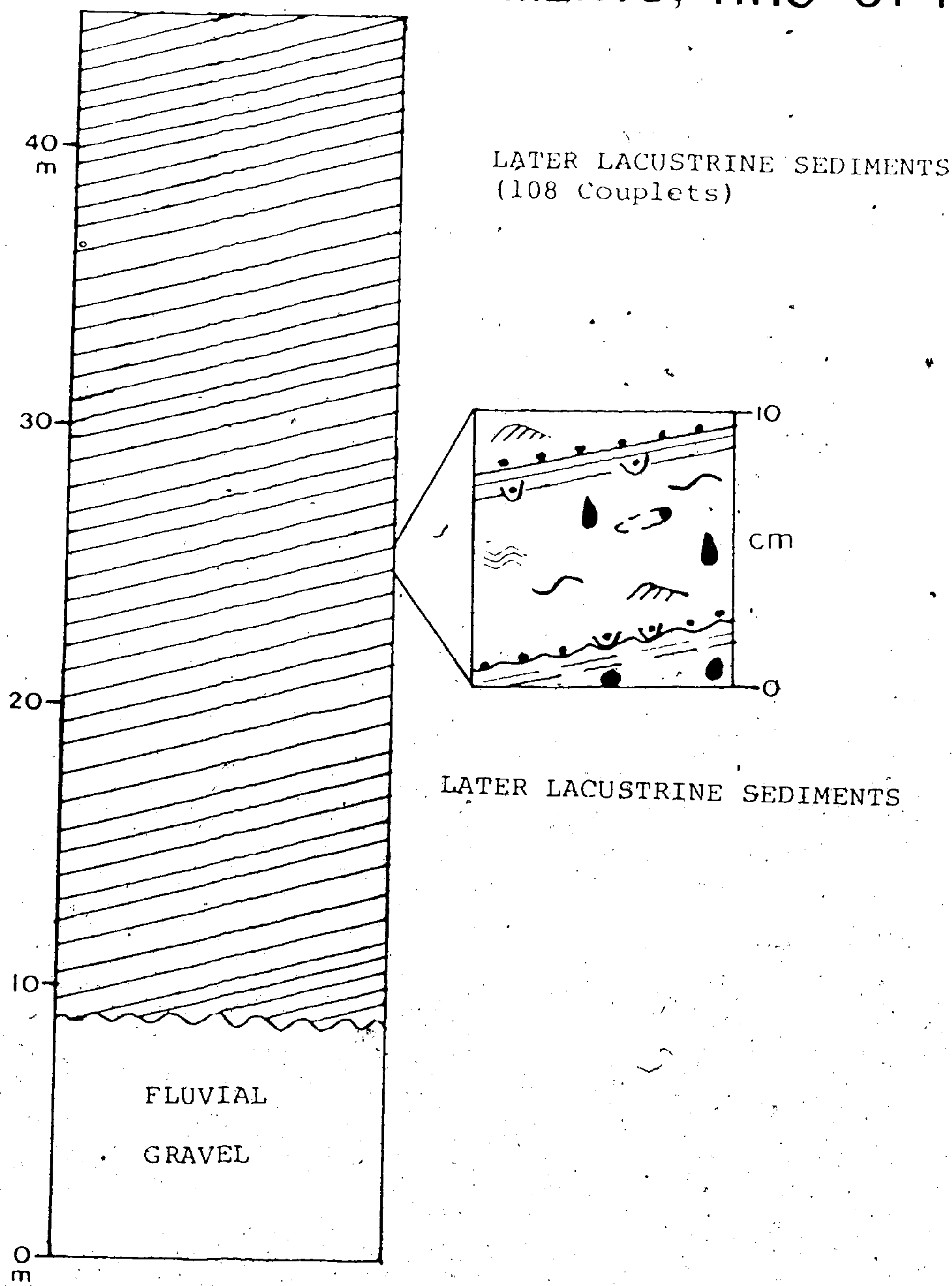
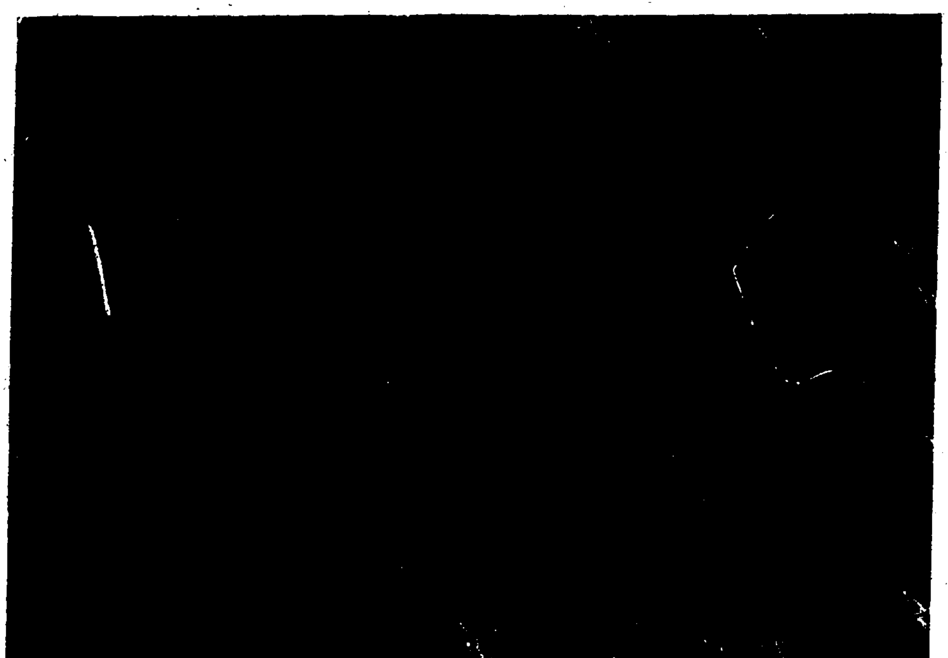


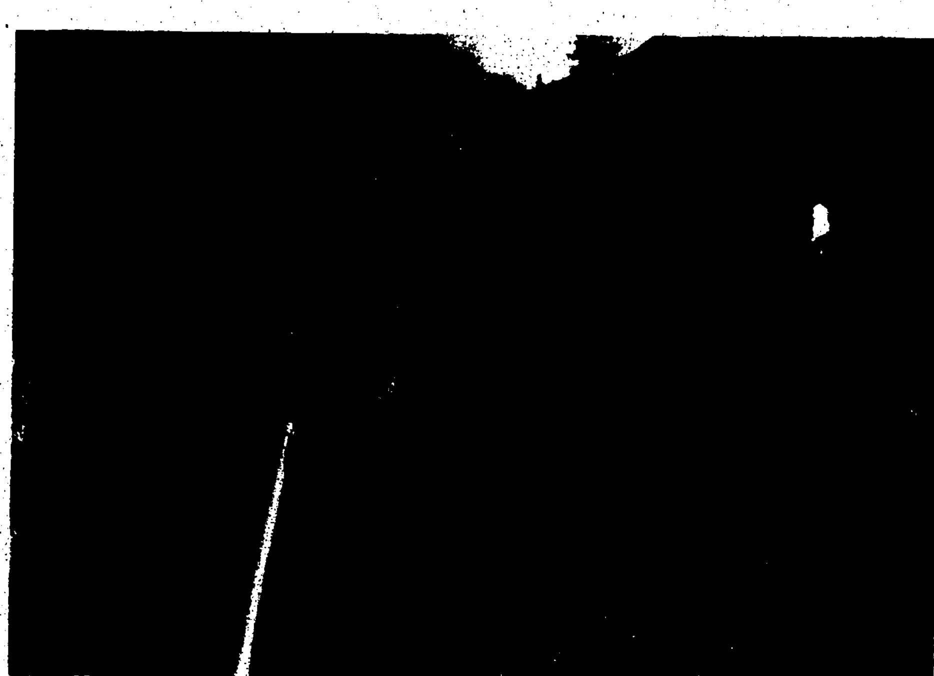
Plate 36 (top)

Later Lacustrine Sequence, Eastern Part, HHC 81-1a.. The sequence is composed of thick sand members which grade upwards to thinner silt and clay members.

Plate 37 (bottom)

Later Lacustrine Sequence, Eastern Part, HHC 81-1a. The couplets dip to the west at 12 ° to 17 °. See text for further discussion.





bedding and texture of the sediments thicken and coarsen upwards. Coarse sand mono- or di-layers are present along the bases of the medium sand units. The contacts below the sand units are erosional, while the upper contacts are gradational. Small-scale ball-and-pillow structures are present along both the upper and lower contacts of the sand units. Where not affected by ball-and-pillow development, the contacts are planar and dip 12°-17° (modal 14°) towards the northwest.

Within the sand layers, ripple-drift cross-lamination (types A and B), trough cross-lamination, flute casts, swing-dash laminations formed by draping of ripples and subsequent reworking, and shallow structures behind large clasts all indicate northwesterly flow, parallel to the dip of the bedding surfaces (Plate 38).

In many of the sand layers, small balls (0.7 cm modal diameter) of clay and clayey silt are present. Most commonly, the balls are elongated parallel to the bedding, often tapering towards the northwest (Figure 80a). Other balls are spherical or subspherical, rarely sitting amidst undisturbed sand laminations (Figure 80b) but more commonly with shallow structures indicating divergence around the obstacle by northwestward flowing water (Figure 80c). Some balls are aligned vertically, with a tapered upward pointing end and a rounded base (Figure 80d). The underlying lamination is frequently distorted around these tapered clasts, here termed "teardrop balls". The basic teardrop

Plate 38

Internal Structure, Sand member of Couplet 33, HHC 81-1a. The current flow was towards the west.

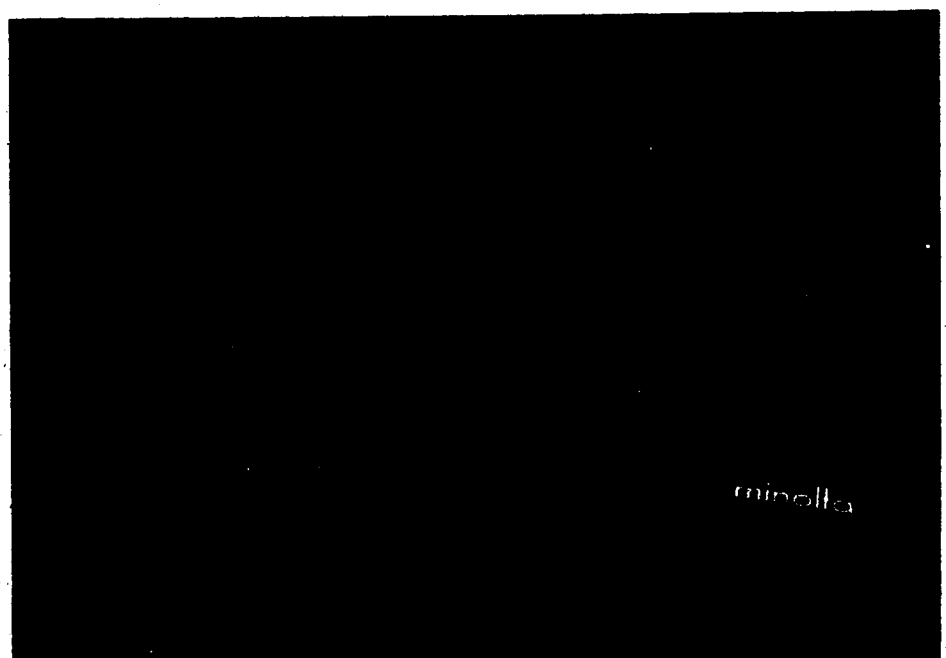

minolta

Figure 80

Clay Ball Types

A---Elongate Balls

B---Spherical Balls

C---Spherical Balls and Shadow Structures

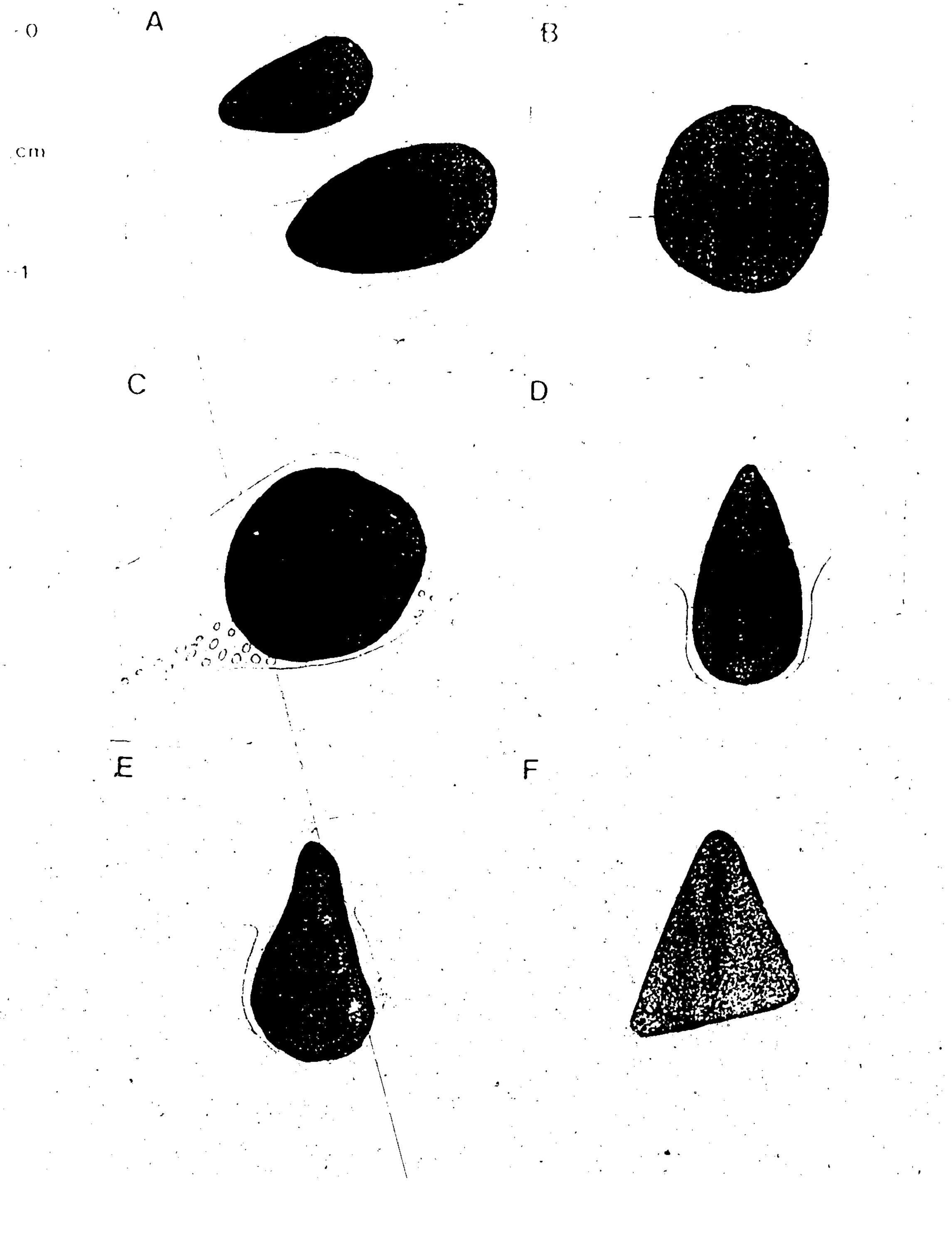
D---Teardrop Balls

E---Peardrop Balls

F---Chocolate Drop Balls

See text for discussion.

FIGURE 80 CLAY BALL TYPES



morphology may be modified to produce an elongated tapered neck, sometimes curved in the direction of flow ("peardrop balls", Figure 80e); or the base of the clay ball may be flat and the underlying laminations undistorted ("chocolate drop balls", Figure 80f). Within a particular sand stratum, one of the above clay ball types is usually dominant.

In isolated fine sand and silt horizons within the sequence exposed at HHC 81-1a, small-scale planar herrringbone cross-lamination was observed. Flow directions for pairs of these cross-laminations were opposed by approximately 180°, and neither direction generally co-incided with the northwestern orientation suggested by the inclination of the bedding surfaces or the sedimentary structures within the sand units.

Interpretation

The deposits were formed by northwestward discharge from streams emanating from the ice front to the east. The northwestward dipping beds are interpreted as foresets produced subaqueously in the marginal area of the lake. Flow velocities varied between 5 and 25 cm/sec.

The rhythmic bedding, the systematic increase in bed thickness vertically, and the vertically coarsening texture of the deposits suggest that the sediment body prograded towards the northwest and that flow was periodic, possibly seasonal. Seasonally-controlled outflow of glacial meltwater could produce the sedimentation pattern observed.

Similar deposits have been interpreted as glaciolacustrine delta sediments by Gustavson et al (1975), Shaw (1975), Cohen ([illegible]), and Thomas (1984a, 1984b). Deposits interpreted as fan delta sediments (Postma et al 1983) and underflow fan sediments (Fenton et al., 1983, Kehew and Clayton, 1983) also show similarities to the HHC 81-1a sediments. Differences between the sediments of the distal zones of deltas, fan deltas, and underflow fans are minimal, and the chief distinctions are based on the position of the upper foreset beds relative to lake level and the development or absence of topset beds. Although underflow fans tend to have shallower foreset slopes and more debris flow sequences than deltas, these distinctions are strictly relative.

The erosional nature of the upper contact of the foreset sequence, and the absence of exposures of proximal lacustrine sediment east of HHC 81-1a and HHC 81-4, mean that the presence or absence of topset beds cannot be determined. The inferred proximal situation of the sediment body, however, suggests a short, steep sediment cone developed from immediately adjacent ice. This environmental situation is more conducive to the development of an 2 underflow fan than a Gilbert-type (1890) delta.

The variety and number of clay and clayey silt balls indicate that the deposits were proximal, as these clasts would not be capable of surviving transportation for long distances (cf. Smith, 1972). The clay may have been transported to the lakeshore by glacial ice, or may have been derived from the older lacustrine deposits. The teardrop-shaped, vertically oriented clay clasts were transported on the lake surface, either by brash ice calving or seasonal lake ice, and were subsequently deposited by settling after ablation. The different shapes of the clay clasts reflect variations in cohesion of sediment within the clasts and in the substrate, with cohesive clasts resisting elongation and cohesive substrates resisting deformation upon impact. The strength of bottom currents also has an impact upon clast shape.

The mode of formation of the herringbone cross-laminations recorded in some fine sand strata at HHC 81-1a remains enigmatic. The variance between the flow directions and the northwestward flow trend that dominates the deposit, and the inconsistency among the herringbone flows, suggest that the cause of the alternating flows was relatively minor. Small channels could have developed on the margins of the fan or delta at variance to the main flow direction during peak influx. When the influx declined, these channels could have served as pathways for minor wind-driven lacustrine currents and waves, thus producing a pair of diametrically opposed cross-laminations.

The ice-marginal lacustrine environment is illustrated in Figure 81. Flow into the Rat Valley caused the development of the sequences formed by eastward-moving turbid underflows. Meltwater discharge from the ice front

Figure 51

Postulated Glaciolacustrine Environment, Ice Marginal Area, Upper Rat River Valley (Eastern Portion)

Regional Wind Direction

Katabatic Wind Direction

Sublacustrine Debris Flow

Sediment Rain-out

Debris apron

Ice Flow Direction

See text for discussion.

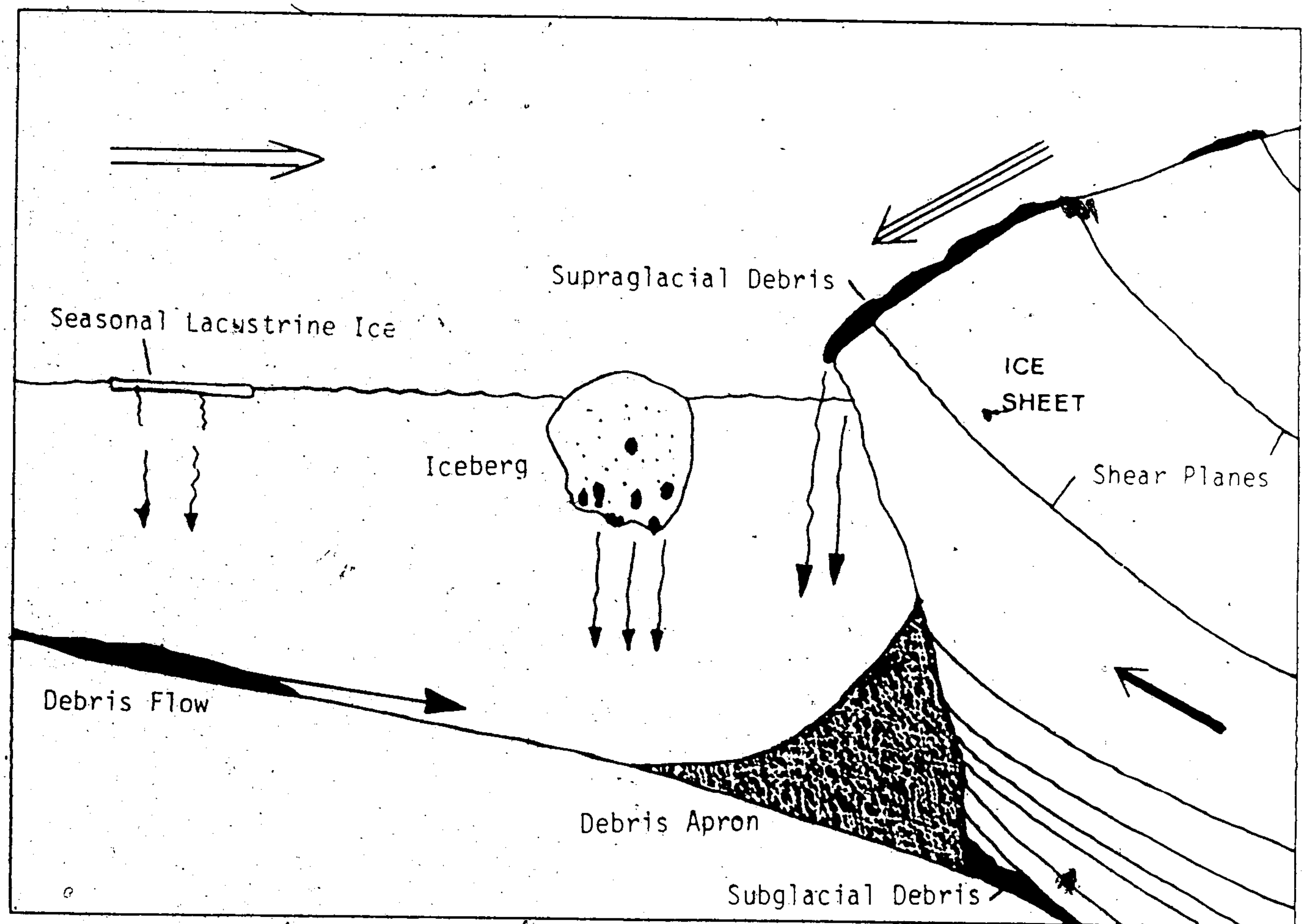

Supraglacial Debris
Seasonal Lacustrine Ice
ICE
SHEET
Iceberg
Shear Planes
Debris Flow
Debris Apron
Subglacial Debris

FIGURE 81

constructed the sediment cone (either as an underflow fan or a delta) exposed at HHC 81-1a and HHC 81-4.

Regional Significance

Formation of a large lake by ice damming in the Rat River basin would be expected to influence events and drainage patterns in the Bell, Bluefish and Old Crow Basins to the west. The maximum level of inundation recorded in McDougall Pass is 365 m (Figure 62), which corresponds to the level of the highest shoreline recorded in Old Crow and Bluefish Basins (O.L. Hughes, Geological Survey of Canada, personal communication, 1984; P.L. Waters, University of Alberta, personal communication 1984). During the time of maximum lake levels in the Rat Valley, a system of interconnected basins existed throughout the northern Yukon. At the time of maximum glaciation, drainage through the Rat Valley to the east was initially blocked, and water from the western basins discharged west via the Ramparts of the Porcupine River (Thorson and Dixon, 1983, Stage 3; O.L. Hughes, Geological Survey of Canada, personal communication, 1985).

The topography of the region north of the Rat River effectively precluded the development of outlets distal to the ice margin, except through McDougall Pass towards the west. The record preserved in the Bell, Bluefish, and Old Crow Basin sediments, however, indicates that eastward drainage was rapidly established (Thorson and Dixon, 1983;

O.L. Hughes, Geological Survey of Canada, personal communication, 1985). The only potential eastward outlet route from the Rat valley is adjacent to the ice front, along the eastern margin of the Richardson Mountains. This area is marked by a steep east-facing escarpment, the base of which is covered by extensive alluvial fans (Legget et al., 1966). The precise position of the outlet channel therefore cannot be determined. It probably abutted the ice front as far north as Shoalwater Bay, and may have been connected with the large meltwater channel on the Yukon Coastal Plain that terminates at Phillips Bay. A channel immediately adjacent to the ice margin would be susceptible to periodic damming during minor or even seasonal readvances, and thus local jokulhlaups may have occurred in the downstream areas.

The sedimentary succession in the Bell Basin suggests repeated drainage and infilling of a large lake (O.L. Hughes, Geological Survey of Canada, and J.V. Matthews, Jr., Geological Survey of Canada, personal communication, 1985). An intervening high flow stage between low flow events in the Porcupine River sediments is also recorded by Thorson and Dixon (1983). These events have been attributed to periodic (possibly seasonal) outflows through the eastern outlet developed during the later lacustrine phase in the Rat valley. Although no evidence for periodic drainage of the valley was observed in the sedimentary successions exposed, partial drainage of the lake induced by seasonal retreat of the ice front along the outlet is not precluded by the sedimentary record. If the lake had a water surface elevation of 365 m (the maximum level recorded in McDougall Pass), location HHC 81-3 would lie approximately 150 m below the water surface, disregarding the effects of glacioisostatic depression. Considerable quantities of water could be regularly drained from the lake through a high-level channel (or channels) to the east without significantly influencing the sedimentation pattern in deep-water locations. The exposure of water-saturated sediments in McDougall Pass through periodic drawdowns might act to induce slumping in the western region of the lake, and hence generate turbid underflows. Maintenance of lacustrine conditions in the deepest section of the upper Rat valley is therefore not incompatible with periodic discharge through the eastern outlet.

The ^{14}C dates of 21,000 ±270 B.P. (GSC-3371) and 21,200 ±240 B.P. (GSC-3813) obtained from detrital organic beds at HHC 81-3 suggest that this location was covered by lake waters c. 21,000 B.P. Although the detrital organic beds provide only a maximum age for the lacustrine sediments, the degree of preservation of the organic matter and the stratigraphic positions of the dated beds indicate that the dates represent a reasonable estimate for the time of deposition of the lacustrine sediments.

I. Postglacial and Holocene Sediments and Environments

Fluvial Sediments

Scattered throughout the Rat River valley are exposures of fluvial sediments (Figure 62). The exposures are small, laterally discontinuous, and provide no critical stratigraphic, sedimentologic or palaeoenvironmental information.

Sequences in the upper Rat Valley area are similar to those developed in the modern meandering Rat River (Appendix 4). An abandoned channel sequence at HHC 81-1g exposed directly above older fluvial gravels (Figure 70), contains abundant wood fragments near its base. These fragments were tentatively identified as Picea, and a sample (81-1g-11) was submitted for ^{14}C dating. The result of 1070±70 B.P. (GSC-3715) indicates that this sediment is a recent fill deposit formed in an abandoned meander loop truncated by the Rat River. The river has therefore initially aggraded at least 2 m and subsequently dropped at least 11.5 m since 1070 B.P. The dated sample was identified as Alnus by R.J. Mott (Geological Survey of Canada). J.V.Matthews Jr. has subsequently identified a twig and needles from the sample as Picea (personal communication, 1984).

Along the lower part of the Rat River, Holocene fluvial sequences indicate braided-meandering stream environments similar to the modern lower Rat River (Appendix 4). The exposures are only present adjacent to the modern braidplain.

Palynological Analysis

Only one location proved to have sufficient palynomorphs for analysis. This location, HHC 81-6, is located along the south bank of the Rat River near the west end of Horn Lake (Figure 62).

The section consists of two exposures of peat, which overlie a terraced kame of outwash gravel. The upper terrace surface was 55 m above river level on 3 August 1981, and is capped by a thin layer of peat and transported organic detritus. The lower terrace surface is approximately 35 m above river level, and is capped by a layer of peat varying from 10 cm to 210 cm in thickness. This terrace is flat-topped, with the exception of a depressed area at the base of the upper terrace, located approximately 100 m east of the thickest exposed portion of the peat cap. This depression extends to a maximum depth of 5 m and represents the remnant of a slightly lower terrace. A thin peat cap (maximum thickness 30 cm) overlies this depression. The outwash gravel contains a variety of lithologies, including clasts derived from the Bonnet Plume Formation that contain fragments of <u>Osmundia</u> and <u>Sequoites</u> or <u>Elatocladus</u> (C.R. Stelck, University of Alberta, Personal communication, 1983).

Samples were taken from an organic detrital lens on the surface of the upper terrace (#1), from the peat cap on the lowermost terrace (#3), and at 15 cm intervals from the 210 cm thick peat cap at the edge of the 35 m terrrace (#4 - #17). All of these samples proved to contain adequate palynomorphs for counting. ^{14}C dates obtained for samples #4 and #17 indicate that the main peat exposure was formed between 8280±110 B.P. (GSC-3399, #4) and 990±50 B.P. (GSC-3797, #17).

The detrital lens sample (#1) was dominated by Cyperaceae (34%), Picea (mariana type), Empetrum nigrum, and Vaccinium. Shrub Betula and Alnus type crispa grains are present in lesser amounts, along with Gramineae and Saxifraga. This assemblage suggests a low arctic tundra environment with scattered Picea stands and minor areas of shrub cover. The relatively low percentages of Artemisia and Gramineae indicate a moist environment and suggest that the assemblage was derived very locally, possibly from the shoreline area of a shallow pond developed on the gravel. The absence of upland vegetation palynomorphs (such as Cassiope type tetragona and Dryas) also suggests a local source. This spectrum is similar to those of zone A of the main exposure, as described below.

The 14 samples of the main exposure can be divided into 4 pollen zones (Figure 82, in pocket). The basal zone, A, encompasses the two lowermost samples. These assemblages are characterised by high concentrations of Cyperaceae, Gramineae, and Picea, low to moderate concentrations of Betula, Alnus type crispa, and Artemisia , and low amounts of Ericaceae pollen. The arboreal fraction is exclusively Picea (60% mariana type), while the arboreal-shrub fraction contains minor amounts of Alnus type incana, Salix, and Populus. Betula is present in excess of Alnus throughout the zone. The shrub fraction is relatively small, and is dominated by Empetrum nigrum with Ledum and Vaccinium subordinate to Andromeda type Polifolia. Shepherdia canadensis is present in trace amounts. The herbaceous component is quantitatively the most important, and is dominated by Cyperaceae and Gramineae, with Cruciferae and Saxifraga pollen present in excess of Artemisia. Caryophyllaceae (Sagina and Silene types) and Rubus chamaemorus are also present. The spore fraction is dominated by Sphagnum, Equisetum, and Lycopodium. No aquatic pollen is present.

Analysis of sample #4 by J.V. Matthews Jr. (Geological Survey of Canada, unpublished Plant macrofossil Report 81-16) revealed Empetrum nigrum, Andromeda Polifolia, and Carex seeds and rhizomes tentatively assigned to Equisetum. These plant macrofossils suggest an acidic bog, a conclusion that agrees with the palynological spectrum. The regional environment was characterised by stands of Picea with Ledum and Vaccinium-dominated understory vegetation, separated by open areas occupied by Cyperaceae and Gramineae, with occasional Betula (glandulosa) shrubs around wet depressions. Dry areas, on exposed gravel or bedrock, were occupied by Equisetum, Caryophyllaceae, and Shepherdia canadensis.

The postulated presence of Alnus in the region at 8280±110 B.P. is compatible with the reported find of Alnus twigs at Twin Lakes, N.W.T., by M. Kuc (as cited by Fyles et al 1972).

The overlying zone B, represented by the interval between 45 cm and 105 cm above the base of the unit, is characterised by high percentages of Picea (chiefly glauca), Betula, and Alnus type crispa (Figure 82). Larix is also present in minor amounts. The arboreal-shrub fraction is the largest component, with Betula and Alnus type crispa comprising approximately equal proportions. Alnus type incana, Salix, and Populus are present in minor amounts. The shrub component is dominated by Empetrum nigrum, with Andromeda type Polifolia and minor Ledum, Vaccinium, Arctostaphylous rubra/alpina and Myrica Gale. The herb component is comprised mainly of Cyperaceae and Gramineae grains. Sagina, Cruciferae, Rubus chamaemorus, Saxifraga, Chenopodium and Artemisia are also present. The pteridophyte component is dominated by Lycopodium and Equisetum, while Sphagnum is the only bryophyte represented.

The reduced concentrations of the herb component and the increases in the arboreal-shrub and shrub fractions indicate a general spread of woodland and thicket vegetation. The increase in vegetation cover resulted in an increase in the locally-derived pollen component, as the woodland and thicket areas acted as traps for the open tundra grains. Thus, the spectra of zone B are enriched in the bog taxa (such as Empetrum nigrum). The shrub thickets were probably dominated by Alnus, as Betula is usually over-represented in pollen spectra (eg. Pennington, 1980). The assemblages are indicative of a climate very similar to that presently prevailing in the region.

The portion of the peat between 120 cm and 180 cm above the base of the unit is characterised by high concentrations of Picea and Alnus type crispa pollen, with Cyperaceae, Gramineae, and Caryophyllaceae (Sagina, Silene), and low concentrations of Betula. These samples are considered as palynological zone C (Figure 82).

The arboreal fraction is dominated by Picea (mainly glauca), with minor amounts of Larix. The arboreal-shrub component is comprised of Alnus type incana, Salix and Populus in addition to Alnus type crispa and lesser amounts of Betula (chiefly shrub). The shrub fraction is very small, and consists chiefly of Empetrum nigrum, with Andromeda type Polifolia, Ledum, Vaccinium, Arctostaphylous uva-ursi, and Myrica Gale present in trace quantities. The large herbaceous component is dominated by Cyperaceae and Gramineae, with Cruciferae, Sagina, Silene, Saxifraga, Arnica, Rubus chamaemorus, Artemisia and Polemonium. The high concentration of Cruciferae grains noted at the 180 cm level probably does not reflect the relative importance of this taxon in the vegetation pattern of the region. Many of the Cruciferae grains were present as aggregated clusters, suggesting that they originated from plants extremely close to the depositional site. One or more Cruciferae anthers may have been incorporated into the peat.

The Pteridophyta fraction is dominated by Lycopodium and Equisetum. Sphagnum is the only bryophyte present, and no aquatic pollen was detected.

The general decline in moisture-preferring taxa (such as Empetrum, Andromeda, Sphagnum and Rubus chamaemorus) and the replacement of Betula by Alnus indicate that the immediate area became drier at this time. The regional climate may not have undergone significant change, however. Picea, Cyperaceae and Gramineae concentrations remained approximately constant, suggesting that the distribution of wooded and open areas remained relatively unaltered. The fluctuations were probably induced by a local lowering of the water table, possibly as a result of downcutting and lateral migration by the adjacent Rat River.

The uppermost pollen zone D, comprises the top 30 cm of the peat (Figure 82, in pocket). The arboreal fraction is dominated by Picea type glauca , with lesser P. mariana and minor Larix . The arboreal-shrub component is dominated by Alnus type crispa, with subordinate Betula (shrub) and minor arboreal Betula, Alnus type incana and Salix pollen. The shrub fraction is dominated by Vaccinium, Ledum andEmpetrum nigrum, with minor amounts of Andromeda type Polifolia, Arctostaphyllous rubra/alpina, A. uva-ursi and Cassiope type tetragona. Cyperaceae and Gramineae are present in approximately equal amounts in the herbaceous fraction, along with Cruciferae, Saqina, Cerastium, Viola, Saxifraga, Chenopodium, Rubus chamaemorus, and very minor traces of Artemisia. Equisetum, Lycopodium and Sphagnum spores are also present.

The increase in Rubus chamaemorus, Empetrum nigrum, Ledum, and Vaccinium, and the presence of Viola and Andromeda pollen, indicate an increase in moisture in the immediate area of deposition. This could indicate a local rise in the water table, or it could be indicative of a regional trend towards the aggradation of permafrost. Zoltai and Tarnocai (1975) have documented such an aggradational period in the Mackenzie Valley, beginning approximately 2600 B.P. The ^{14}C date of 990±50 obtained from the uppermost sample is compatible with this regional picture. However, the topographic situation of the site, along the Rat River Valley, the presence of a thick deposit of permeable gravel under the peat, and the absence of ice lenses in the sediment suggest that local factors may have been more important in the development of the site than was regional permafrost aggradation. The increase in Alnus pollen and the stability of the concentration of Betula grains throughout the zone imply that the region was characterised by areas with relatively dry slopes, sand flats, and exposed bar surfaces, conditions which could be produced by local shifts in the channel pattern of the Rat River as well as by regional climatic alteration.

Sample #3, from the lowermost peat cap, shows an assemblage dominated by Picea type glauca (Table 31). Larix is also present. The arboreal-shrub component is relatively small and is predominantly Alnus type crispa, with lesser Betula and minor Alnus type incana, Salix and Populus. The shrub fraction is dominated by Ericaceae, principally Vaccinium and Andromeda type Polifolia, with minor Empetrum nigrum. The herbaceous component is large, with Cyperaceae, Gramineae and Rubus chamaemorus prominently represented. Cruciferae, Saxifraga, Cerastium, Viola, and Chenopodium are also present. The large Pteridophyta fraction is dominated by Lycopodium, with small amounts of Equisetum and Cystopteris. Sphagnum spores are also present.

The asseblage is characteristic of Rubus chamaemorus bogs developed adjacent to Picea upland stands, and reflects the current vegetation of the immediate area. Consequently, it is correlated to the uppermost portion of the palynological stratigraphic succession of the principal exposure. It may be slightly younger than 900±50 B.P., and the high proportion of palynomorphs apparently derived from the immediate vicinity of the site suggests that it reflects the climate and vegetation of only the very local area.

J. Summary

The sedimentary exposures of the Rat River valley contain a record of preglacial fluvial activity; fluvial and lacustrine sediments which postdate the initial glaciation

Table 31

Palynological counts, Quaternary taxa, sample #3, section HHC 81-6. All taxa included in pollen sum

Taxon	Percentage
Picea	28.7
Pinus	0.0
Abies	0.0
Betula	6.0
Alnus type *crispa*	9.3
Alnus type *incana*	1.0
Salix	1.0
Populus	0.5
Larix	0.5
Ledum	0.5
Vaccinium	1.5
Andromeda	1.5
Empetrum nigrum	0.5
Cyperaceae	11.6
Gramineae	7.6
Rubus chamaemorus	4.8
Cruciferae	1.5
Saxifraga	1.0
Cerastium	1.0
Chenopodium	1.5
Equisetum	1.5
Cystopteris	1.5
Viola	0.5

Lycopodium	12.3
Sphagnum	2.8

of the region; a single glacial till; lacustrine sediments formed in an ice-dammed lake; and Holocene fluvial and autochthonous peat deposits. A composite stratigraphic column is presented in Figure 83.

Preglacial gravel is exposed only along the lower Rat River. The gravel was deposited as lateral and longitudinal bars by an eastward-flowing braided-meandering stream.

Fluvial and lacustrine sediments containing palynomorphs overlie the preglacial gravels. The oldest palynomorphic assemblages suggest a low arctic tundra environment. The presence of a cryoturbated diamicton unit also suggests tundra conditions. The sediments were deposited after the inital glaciation of the area, as indicated by the clast lithologies.

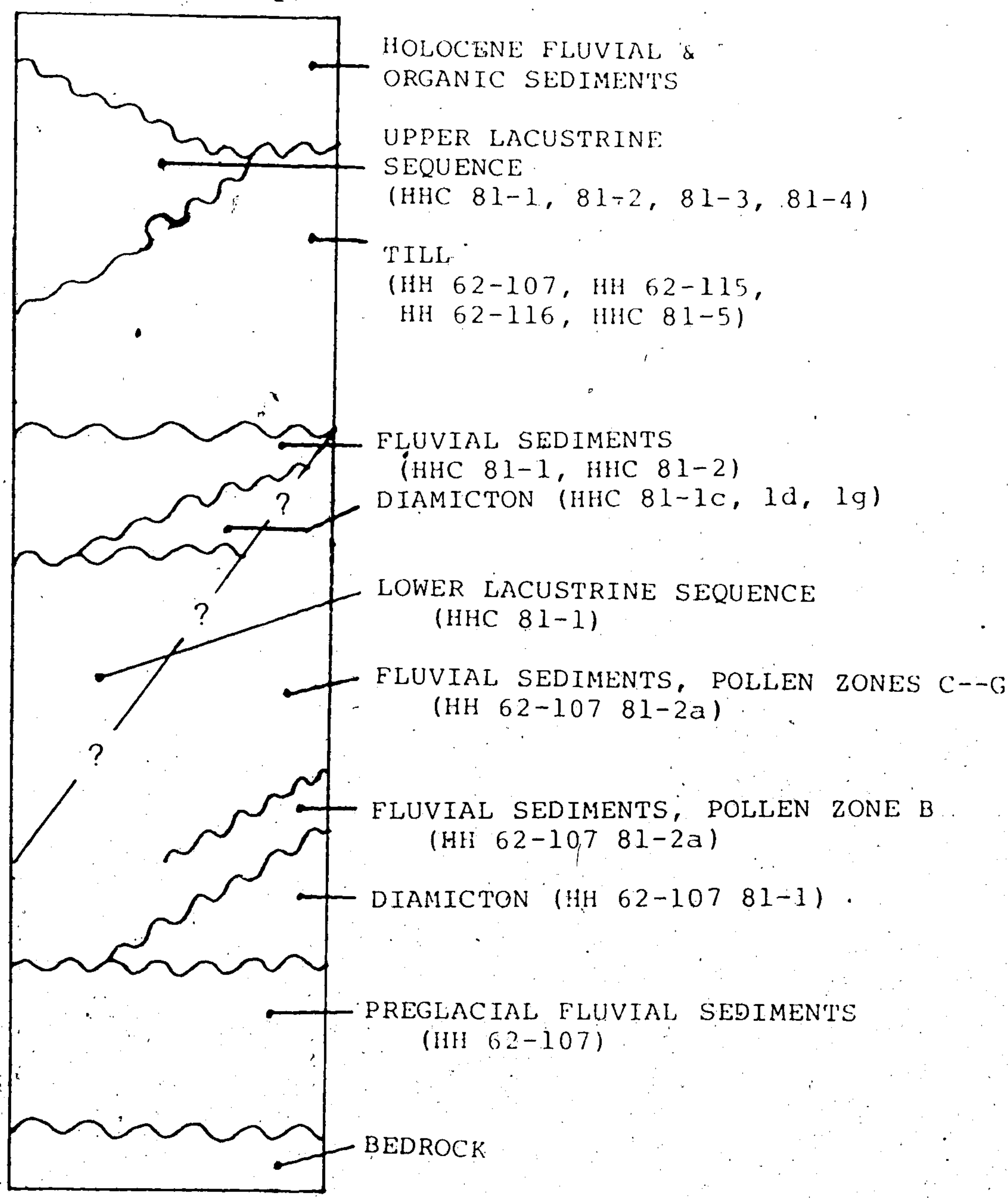
A period characterised by a climate warmer than that of the present is indicated by palynomorph assemblages containing <u>Pinus</u>, <u>Corylus</u> and <u>Cornus</u> grains. This event may be correlative to the warm interval recognized in the palynomorphic succession by Litchi-Federovich (1974) and McCourt (1982). Successive assemblages indicative of climates similar to that of the present Rat River valley are separated from the older warm period assemblages by an erosional contact. The youngest palynological zone, indicative of tundra conditions, is associated with autochthonous organic sediments which have been ^{14}C dated at >43000 B.P. (GSC-3359).

Figure 83

Composite Stratigraphic Column, Rat River-McDougall Pass Area

The column is schematic, and the thicknesses of the units as illustrated do not represent either the thicknesses of sediment deposited or the amount of time represented by each unit. The relationship between the pollen bearing sediments at HH 62-107 and the Lower Lacustrine Sequence exposed at HHC 81-1 is uncertain, but both units predate the till exposed in the Rat River Valley. See text for further discussion.

FIGURE 83


HOLOCENE FLUVIAL &
ORGANIC SEDIMENTS
UPPER LACUSTRINE
SEQUENCE
(HHC 81-1, 81-2, 81-3, 81-4)
TILL
(HH 62-107, HH 62-115,
HH 62-116, HHC 81-5)
FLUVIAL SEDIMENTS
(HHC 81-1, HHC 81-2)
DIAMICTON (HHC 81-1c, 1d, 1g)
?
?
?
LOWER LACUSTRINE SEQUENCE
(HHC 81-1)
FLUVIAL SEDIMENTS, POLLEN ZONES C--G
(HH 62-107 81-2a)
FLUVIAL SEDIMENTS, POLLEN ZONE B
(HH 62-107 81-2a)
DIAMICTON (HH 62-107 81-1)
PREGLACIAL FLUVIAL SEDIMENTS
(HH 62-107)
BEDROCK

An early lacustrine sedimentary sequence present in the upper Rat Valley was produced by glacial impoundment. Although the relative times of formation of the early lacustrine sequence and the palynological succession cannot be established, impoundment of the lake required the presence of glacial ice in the lower Rat Valley. The early lacustrine sequence therefore may have been deposited after the deposition of the palynomorph assemblages. This interpretation is highly conjectural, however, as no definitive evidence is available.

Fluvial sediments were deposited overlying the early lacustrine sediments at location HHC 81-1. These sediments are correlated to fluvial beds underlying later lacustrine sediments at HHC 81-1a and HHC 81-2. These sediments indicate that a period or non-glacial conditions separated the early and later lacustrine phases in the upper Rat River valley.

Exposures throughout the Rat River valley do not contain more than one glacial till. The similarities of the clast lithologies and mineralogies in the till exposures investigated suggest that only a single till is represented, and therefore all of the sediments were deposited during a single glaciation. The freshness of the heavy minerals, and the presence of till at or near the surface of the majority of the region, suggest that the glacial event occurred relatively recently. The time of commencement of the glaciation is unknown. Autochthonous organic sediments underlying till at HH 62-107 81-2a have been ^{14}C dated at >43,000 B.P. (GSC-3359).

Lacustrine sediments produced by eastward-flowing turbid underflows are present in the western part of the Rat River valley. Detrital organic beds within the lacustrine succession have been ^{14}C dated at 21300±270 B.P. (GSC-3371) and 21200±240 B.P. (GSC-3813). The preservation of the organic matter suggests that these dates do represent the approximate time of deposition of the beds. Correlative ice-marginal deposits are present at sections HHC 81-1a and HHC 81-4. Till and lacustrine sediments are not present together in any exposure.

The later lacustrine phase is correlated to lacustrine events in the Bell, Bluefish and Old Crow Basins. These basins drained eastwards through McDougall Pass into the Rat River Valley. Drainage of the lake was to the northeast, probably paralleling the ice front as far north as Shoalwater Bay.

The oldest postglacial organic sediment is autochthonous peat formed 8280±110 B.P. (GSC-3399) at HHC 81-6, in the approaches to McDougall Pass. Climatic conditions were similar to those prevalent in the region today, as indicated by the palynomorph assemblages. *Alnus* was established in the area at this time. Local conditions controlled the vegetation development at location HHC 81-6 throughout the Holocene. At other locations, meandering and braided-meandering streams deposited sediments.

There is considerable uncertainty regarding the Quaternary record in the Rat River Valley. The relationship between the early fluvial palynomorph-bearing succession at HH 62-107 and the early lacustrine phase in the upper Rat River Valley is unclear. The time of the commencement of the last glaciation is unknown. Although the evidence available suggests that the last glaciation was also the most areally extensive, this conclusion must be regarded as extremely tentative. Although much information concerning the mode of lacustrine sedimentation has been garnered, the duration of the lacustrine phases is still unknown. These problems indicate that more research is required in this region. The general absence of suitable exposures has greatly hampered the elucidation of the Quaternary history. Future research should combine intensive sampling of surficial deposits with subsurface investigations.

IX. Quaternary History and Chronology

The Quaternary sediments of the Peel Plateau and Richardson Mountains provide a record of three glacial events and associated fluvial and lacustrine activity. Correlations between the Snake River Watershed - Upper Peel River area and the Caribou River Watershed - Brown Bear Creek area can be established for most sedimentary units. Correlations between the Rat River Valley - McDougall Pass area and the two areas to the south, however, are for the most part highly tentative and uncertain. The correlations are based primarily on stratigraphic position and ^{14}C dated material (Table 32). Table 33 summarizes the principal events and correlations throughout the region. The composite stratigraphic columns for the Snake River Watershed - Upper Peel River area (Figure 31), the Caribou River Watershed-Brown Bear Creek area (Figure 61), and the Rat River - McDougall Pass area (Figure 83), are combined and presented in Figure 84.

Preglacial fluvial deposits of uncertain age are present along the major tributaries of the Peel River. The absence of granitic clasts and the lack of other clasts derived from outside the drainage basins of the tributary streams indicate that the sediments were deposited prior to the initial glacial advance from the Canadian Shield. The distribution of these sediments, along the Rat, Caribou and Snake Rivers, indicates that drainage was towards the east from the Richardson Mountains in the southern and central

Table 32

^{14}C determinations, Peel Plateau region.

Laboratory number	Location	Material and site description	Date
GSC-3359	HH 62-407 81-2a 67°39 5'N 135°28'W	Autochthonous fecal remains and organic material in unit with tundra palynomorphs (pollen zone G); underlies till. See Chapter VIII C	>43,[illegible]
GSC-3565	HHC 81-2 67°43'N 135°51'W	*Picea* fragment in lacustrine sediment flow granule gravel deposit. See Chapter VIII G	>42,[illegible]
GSC-3697, GSC-3800	HH76-3d 65°48'N 133°18 5 W	*Picea* fragments in fluvial deposit; underlies younger glacigenic diamicton in Snake River valle. See Chapter VI G	39,[illegible]

GSC-3371	HHC 81-3 67°43'N 135°51'W	Organic detritus in lacustrine sediment flow deposit, 8m below surface. Contained in Later Lacustrine sediment, Rat River valley. See Chapter VIII.G.	[illegible]
GSC-3813	HHC 81-3 67°43'N 135°51'W	Organic detritus in lacustrine sediment flow deposit, 2m below surface. Contained in Later Lacustrine sediments, Rat River valley. See Chapter VIII.G.	[illegible]
GSC-3691	HH 72-49 66°13'N 135°11'W	Peat at base of peat unit overlying glaciogenic sediment (pollen zone A). See Chapter VIII.H.	[illegible]
GSC-3573	HHC 82-1c 66°22'N 134°2[illegible]'W	Salix fragment in fluvial sediment overlying stratigraphically Later lacustrine sediment, [illegible] River valley. See Chapter VIII.H.	[illegible]

GSC-3695	HH62-62	65°28'N 132°26'W	Organic detritus in alluvial fan channel silt deposit; overlies till. See Chapter VII J	[illegible]
GSC-3399	HHC 81-6	67°44'N 136°03'W	Peat at base of peat unit overlying glaciofluvial gravel. See Chapter VIII I	9,280± [illegible]
GSC-3822	HH 72-49	66°13'N 135°11'W	Peat at top of palynological zone D. See Chapter VII H	7,3[illegible]±[illegible]
GSC-3690	HH 72-50	66°15'N 134°58'W	Peat at base of palynological zone C, overlying fluvial sediments. See Chapter VII H	[illegible],230±[illegible]
GSC-3715	HHC 81-1g	67°42'N 135°49'W	Wood fragment from fluvial sediments overlying mass flow diamicton and Lower Lacustrine sediments Rat River Valley. See Chapter VIII I	[illegible]
GSC-3797	HHC 81-6	67°44'N 136°03'W	Peat, 15 cm below surface, upper part of palynological zone D. See Chapter VIII I	[illegible]90±[illegible]

Table 33

Quaternary Events, Peel Plateau- Richardson Mountains

Time BP	Bonnet Plume Basin	Snake River Valley	Caribou River - Central Peel Plateau	Rat River Valley	Beaufort Sea - Southern Coast
0					
8000		vegetation established		vegetation established	
10,000			extensive fluvial downcutting begins		
11,000		deglaciation complete	deglaciation complete, Central Peel Plateau		
12,400			establishment of vegetation, Caribou Valley		
			lacustrine event, Caribou Valley		
15,000					

16,000	vegetation established deglaciation complete?			deglaciation complete?	deglaciation complete?
21,000		deglaciation begins	deglaciation begins	later lacustrine phase, w. Rat River Valley	
36,900	Hungry Creek Glaciation maximum age of Hungry Creek organic beds	Hungry Creek Glaciation	Hungry Creek Glaciation	unnamed glaciation	withdrawal of Buckland ice begins? Buckland Glaciation?
Late Quaternary	"Deception" lacustrine event	"Snake River" Glaciation	"Snake River" Glaciation	early lacustrine events??	Buckland Glaciation?
			Peel lacustrine event		pre-Buckland events

Middle Quaternary			"Brown Bear" glaciation and fluvial events	oldest palynological successions at HH 82 [illegible]	pre-[illegible] events
Early Quaternary or Late Tertiary	preglacial fluvial events	preglacial fluvial events	preglacial fluvial events	preglacial fluvial events	preglacial fluvial events

Peel Plateau, and towards the northeast from the Canyon Ranges in the southern Peel Plateau. The absence of preglacial sediments from the Peel River valley indicates that this channel was not excavated until after the initial glacial event.

Early Quaternary glaciation of the region is indicated by the presence of weathered granitic and gneissic clasts in the "Brown Bear" fluvial sediments exposed along the Peel River. No tills or other glacigenic deposits associated with these gravels have been recognized. The areal extent of the glaciation is unknown. The age of the sediments cannot be quantitatively ascertained at present. The saprolite development, the disintegration of the granitic clasts, and the development of weathering rinds on the quartzite clasts within the fluvial sediment suggest great antiquity and extensive weathering. The absence of correlative glacigenic and other deposits prevents assessment of the amount of time between glaciation and deposition of the fluvial sedimentary sequence.

Early fluvial activity postdating the "Brown Bear" sediments is recorded in the Snake River valley. These sediments overlie a mudflow diamicton at a single exposure (HH 76-3a). At other locations, the fluvial sediments overlie bedrock. The unit is considered to postdate the Brown Bear sediments because it shows no indication of chemical weathering.

Figure 84

Stratigraphic Correlations, Peel Plateau Region

Correlations among the three principal areas-the Snake River-Upper Peel River area; the Caribou River-Brown Bear Creek area; and the Rat River-McDougall Pass area- are illustrated. The diagram is schematic, and the unit thicknesses illustrated do not represent either the thicknesses of the sedimentary units or the amount of time involved in their deposition. See text for discussion.

ED--Early Diamicton, Snake River valley, HH 76-3a

FS--Fluvial Sediments, Rat River valley, HHC 81-1, 81-2

LLS--Lower Lacustrine Sediments, Rat River valley, HHC 81-1

QD--Quaternary Diamicton, Rat River valley, HHC 81-1c, 1d, 1g, mass-movement deposit

TLLS--Topographically Lower Lacustrine Sediments, Caribou River valley, HHC 82-1

ULS--Upper Lacustrine Sediments, Rat River valley, HHC 81-1, HHC 81-2, HHC 81-3, HHC 81-4

ZZ--Youngest Diamicton, Snake River valley, HH 76-3f

S--Shear Planes and Glacigenic Deformation Structures, Brown Bear Creek area, HH 62-145

Schematic Position of ^{14}C dated materials (See Table 32 for details)

a-- GSC-3359 >43,000
b-- GSC-3565 >42,000
c-- GSC-3697 >39,000
d-- GSC-3800 >40,000
e-- GSC-3371 21,300 ±270
f-- GSC-3813 21,200 ±240
g-- GSC-3691 12,400 ±120
h-- GSC-3573 9,780 ±110
i-- GSC-3695 8,510 ±70
j-- GSC-3399 8,280 ±110
k-- GSC-3822 7,370 ±80
l-- GSC-3690 2,290 ±50
m-- GSC-3715 1,070 ±70
n-- GSC-3797 990 ±50

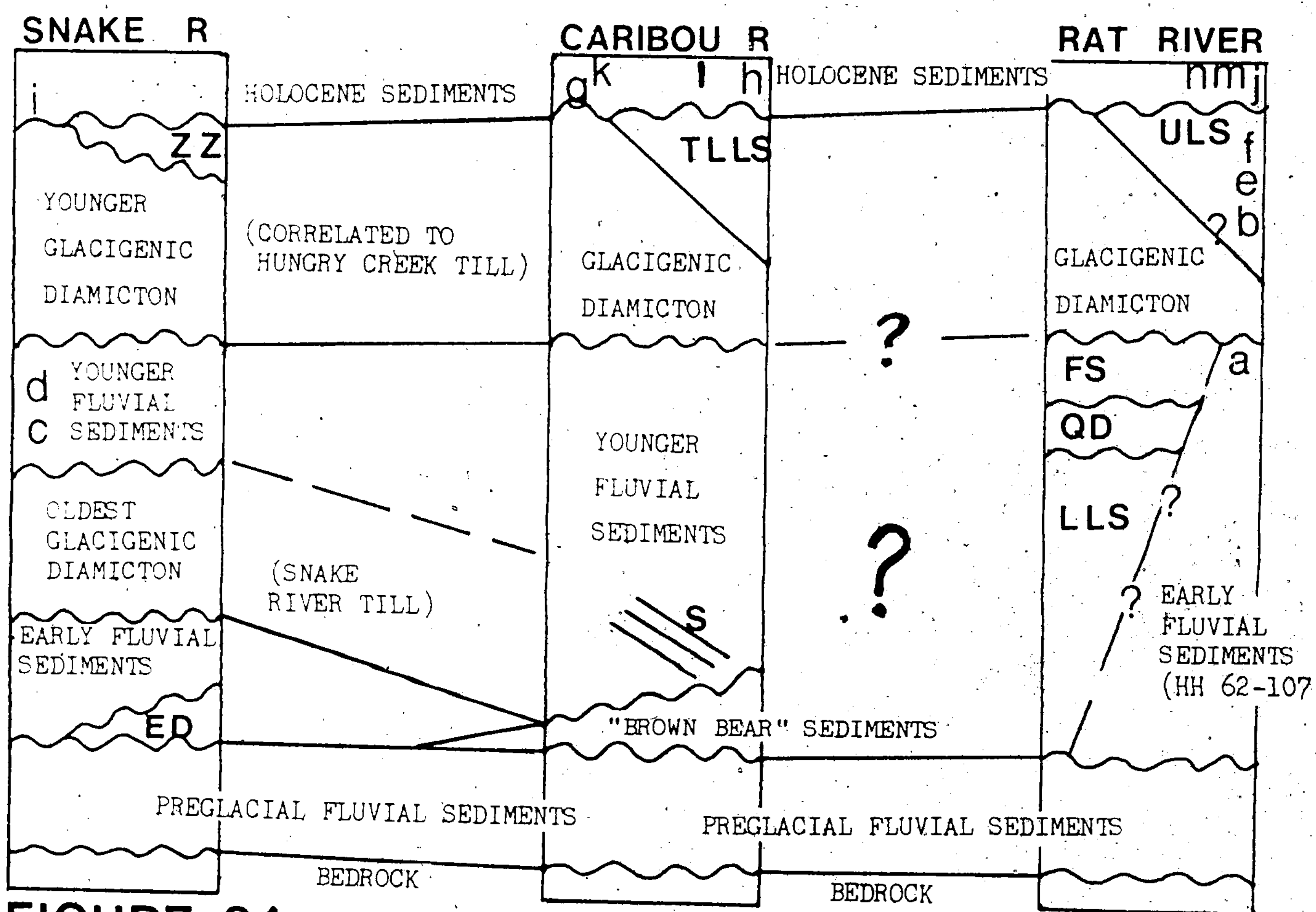

SNAKE R
CARIBOU R
RAT RIVER
HOLOCENE SEDIMENTS
HOLOCENE SEDIMENTS
i
ZZ
g
k
l
h
TLLS
hmj
ULS
f
e
?b
YOUNGER GLACIGENIC DIAMICTON
(CORRELATED TO HUNGRY CREEK TILL)
GLACIGENIC DIAMICTON
GLACIGENIC DIAMICTON
?
d
c
YOUNGER FLUVIAL SEDIMENTS
a
FS
QD
LLS
?
YOUNGER FLUVIAL SEDIMENTS
?
OLDEST GLACIGENIC DIAMICTON
(SNAKE RIVER TILL)
s
?
EARLY FLUVIAL SEDIMENTS (HH 62-107)
EARLY FLUVIAL SEDIMENTS
ED
"BROWN BEAR" SEDIMENTS
PREGLACIAL FLUVIAL SEDIMENTS
PREGLACIAL FLUVIAL SEDIMENTS
BEDROCK
BEDROCK

FIGURE 84

A second glacial advance flowing to the west-northwest from the Canadian Shield resulted in the deposition of the "Snake River" till in the southern Peel Plateau. The absence of correlative till in the Bonnet Plume Basin suggests that the advance terminated east of the summits of the Trevor Range, and blocked eastward drainage from the Bonnet Plume Basin (Figure 85).

The second glacial event was probably responsible for shearing and thrusting the "Brown Bear" and younger fluvial sediments along the Peel River, producing the shear plane traces found at the bases of several exposures. No till from this glaciation has been identified in the central Peel Plateau, and correlation between central and southern Peel Plateau is therefore tenuous.

The lacustrine sediments present along the Peel River near its confluence with the Snake, could have been deposited during retreat of this glacier. The absence of glacigenic sediment and dateable materials in the lacustrine exposures preclude firm correlations.

The "Deception" lacustrine sediments at the Hungry Creek locality in the western Bonnet Plume Basin (Hughes <u>et al</u>, 1981) are overlain by organic-rich beds. Detrital material within these beds has been ^{14}C dated at 36,900±300 B.P. (GSC-2422). This date represents a maximum age for the organic-rich beds, and suggests that the underlying lacustrine sediments are older. The "Deception" sediments cannot be definitively correlated to either the "Snake

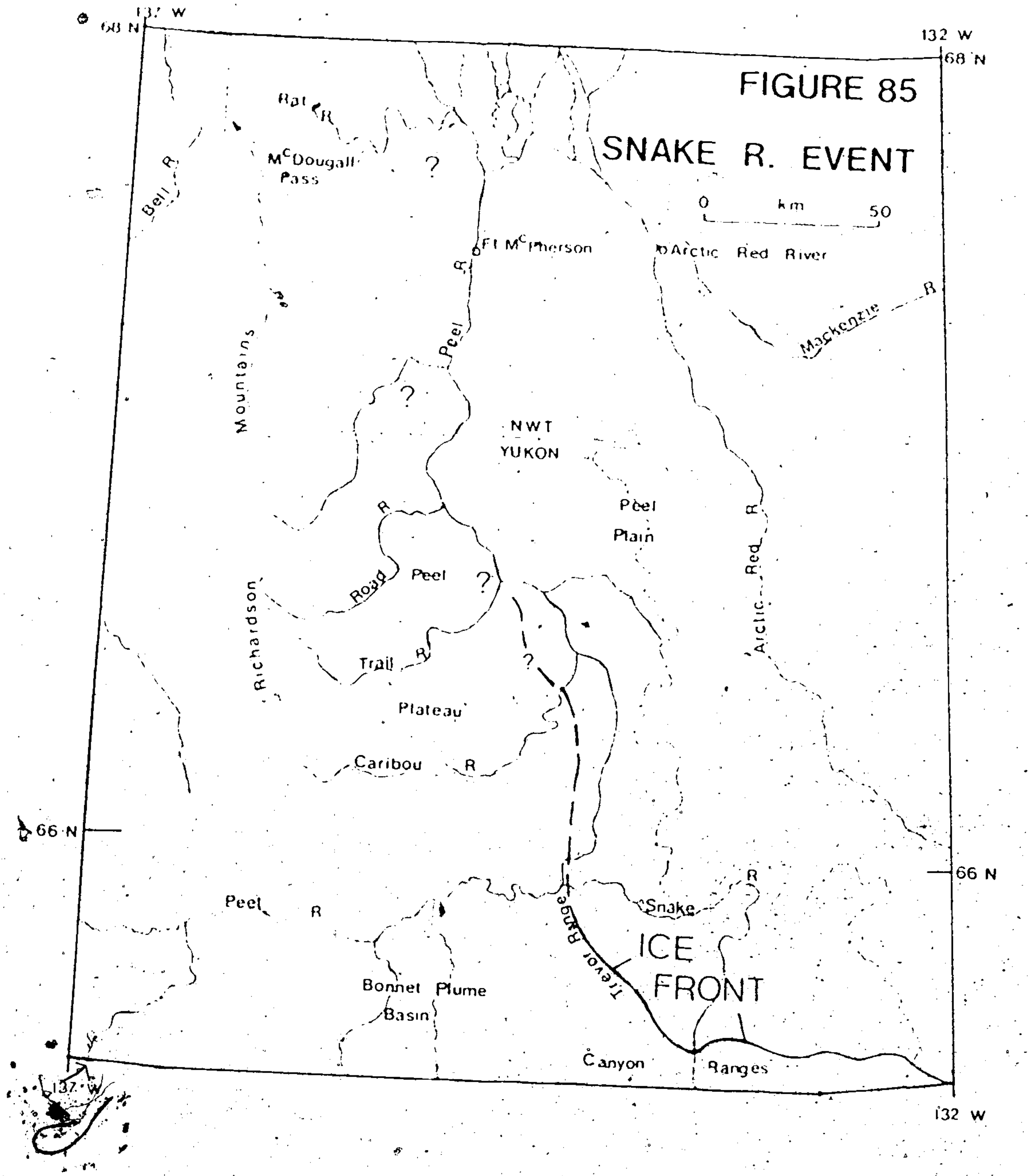


FIGURE 85
SNAKE R. EVENT
0 km 50
137 W
68 N
132 W
68 N
66 N
66 N
137 W
132 W
Rat R
McDougall Pass
Bell R
Mountains
Ft McPherson
Arctic Red River
Peel R
Mackenzie R
NWT
YUKON
Peel Plain
Arctic Red R
Road R
Peel
Richardson
Trail R
Plateau
Caribou R
Peel R
Snake R
Trevor Range
ICE FRONT
Bonnet Plume Basin
Canyon Ranges
?
?
?
?

River" till or the Peel lacustrine sediments. The extent of the older glacial event and the lakes impounded by it, therefore, remain obscure at present.

Younger fluvial sediments are present along the Snake River (Figure 84). Two fragments of Picea wood obtained from the sequence have been ^{14}C dated at >39000 B.P. (GSC-3697) and >40000 B.P. (GSC-3800) (Table 32). Although these wood fragments are clearly detrital, and hence not necessarily representative of the time of deposition of the fluvial sediment, the dates do suggest that the younger fluvial sediments may be of pre-Late Wisconsinan age. The stratigraphic position of these sediments overlying the "Snake River" till thus suggests that this till is of pre-Late Wisconsinan age also.

Events in the Rat River valley cannot be reliably correlated to events in the central and southern regions.

Palynological assemblages in fluvial and lacustrine sediments exposed along the lower Rat River show a succession of environments. Early tundra conditions, indicated by grass- and herb-dominated assemblages and cryoturbated sediments, were succeeded by a climate warmer than that currently prevalent. After an erosional hiatus, forest-tundra transition climatic conditions similar to those of today prevailed in the area. This climate was in turn succeeded by a tundra event. Autochthonous organic sediment from the bed containing tundra palynomorphs at HH 62-107-81-2a has been ^{14}C-dated at >43,000 B.P. (GSC-3359,

Table 32), indicating that the succession is of pre-Late Wisconsinan age. The similarities between the warm climate assemblages and those reported from mid- to early Quaternary sediments in the Old Crow area by Lichti-Federovich (1974) and McCourt (1982) suggest that the Rat River sediments may be considerably older than the mid-Wisconsinan.

Early lacustrine and fluvial sediments are present along the upper Rat River valley (Figure 84). Although the relative times of formation of the early lacustrine sequence and the palynomorph-bearing sequence in the lower Rat River valley cannot be reliably assessed at present, the existence of a lake in the upper Rat River valley necessitates an ice front in the lower valley. Thus, it would appear that the lacustrine sediments postdate the deposition of the palynomorph succession, especially those assemblages indicating boreal forest conditions. No correlation between these units and climatic periods and those noted in the central and southern Peel Plateau, is possible at present.

The youngest glacial event resulted in the deposition of till throughout the central and southern Peel Plateau, and the deposition of the "Hungry Creek Till" in the Bonnet Plume Basin. This glacial advance represents the maximum recorded extent of ice in the study region. The direction of ice flow was southwest to west-southwest in the central Peel Plateau, west-southwest in the headwaters of the Caribou River and north of the Trevor Range, and south-southwest to south in the Snake River valley. Formation of the ridge-and-valley complexes was completed during this glaciation.

The single till exposed in the Rat River valley is tentatively correlated to the youngest glacial event recognized in the central and southern Peel Plateau, primarily on the bases of stratigraphic position and geomorphology. The maximum position reached by the ice front in the western approaches of McDougall Pass extends south to connect with the maximum position reached by the youngest glacial advance in the central Peel Plateau, and also connects with the Buckland glacial maximum recognized north of the study region by Rampton (1981) (Figure 86).

The "Hungry Creek Till" was deposited after the organic-rich beds dated at 36,900 ± 300 B.P. (GSC-2422, Hughes et al 1981). This date, if assumed to be finite, therefore represents a maximum for the commencement of the final glacial event. The lack of reliable dates for the Buckland glacial maximum to the north (as discussed in Chapter IV, "Previous Work"), precludes any firm chronological assessment of the glacial event.

Deposition of the later lacustrine sequence in the upper Rat River valley was initiated by the final glacial advance. Blockage of the Rat River resulted in the diversion of drainage westward into the Bell Basin and the Bluefish Basin. This resulted in the formation of a single large lake with a surface elevation of 365 m covering the Bell, Bluefish, and Old Crow Basins and the upper Rat River

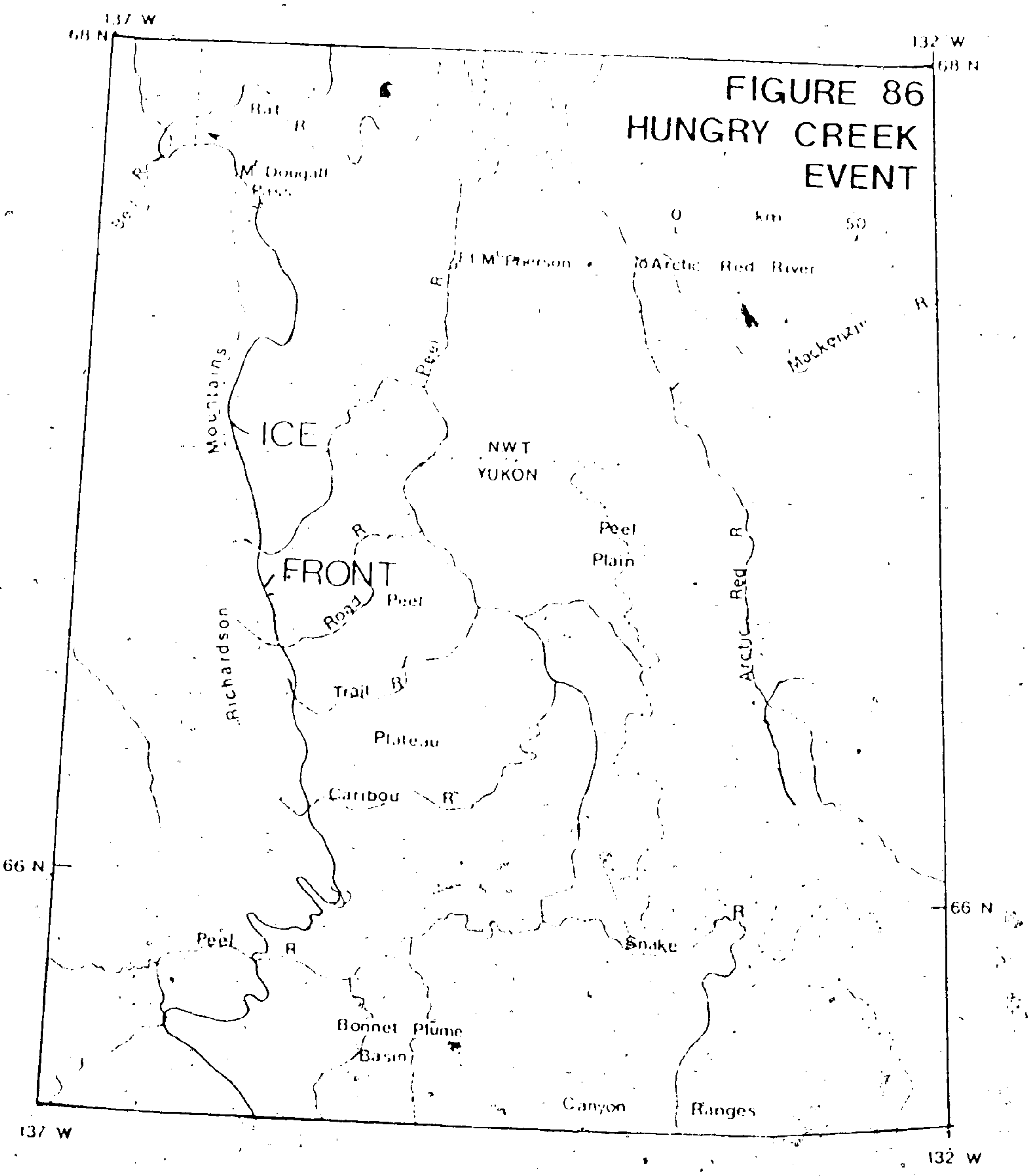


FIGURE 86
HUNGRY CREEK
EVENT
137 W
68 N
132 W
68 N
66 N
66 N
137 W
132 W
0 km 50
Rat R
McDougall Pass
Ft McPherson
Arctic Red River
Mackenzie
R
Peel
R
Mountains
ICE
FRONT
NWT
YUKON
Peel Plain
Arctic Red R
Richardson
Road
Peel
Trail R
Plateau
Caribou R
Peel R
Snake R
Bonnet Plume
Basin
Canyon Ranges

valley, and high westward discharges through the Ramparts of the Porcupine River (Hughes 1972; Hughes et al 1981; Thorson and Dixon 1983). Detrital wood fragments deposited prior to downcutting by the Porcupine River in the Ramparts area have been ^{14}C-dated at 31,840 ± 855 B.P. (BETA-1824, Thorson and Dixon 1983). This date therefore is a maximum estimate for the initiation of westward discharge. The subsequent alternating lake levels in the Bluefish and Old Crow Basins associated with eastward discharge adjacent to the ice front probably began shortly thereafter and continued for several thousand years.

Organic detritus from the upper part of the couplet sequence along the Rat River (HHC 81-3, Figures 70 and 71) has been ^{14}C-dated at 21,300 ±270 B.P. (GSC-3371) and 21,200 ± 240 B.P. (GSC-3813)(Table 32). These dates were determined through analysis of organic detritus obtained 7.6 m and 11.9 m above the base of the sequence, respectively.

A fragment of Picea wood from the lower part of the lacustrine sediments was ^{14}C-dated at >42,000 B.P. (GSC-3565). The degree of abrasion evident on the surfaces of the fragment, combined with its occurrence in a gravel unit within a lacustrine gravity flow sequence, indicates that the >42,000 B.P. date does not accurately reflect the sediment's time of deposition.

The consistency and relative stratigraphic position of the younger determinations (Table 32 and Figure 70) indicate that contamination by detrital lignite was probably not significant. The upper Rat River valley was therefore occupied by a lake until at least 21,200 ± 240 B.P. Drainage of the lake followed the retreat of the ice margin to the east.

The apparent continuity of the glacial limit between the Beaufort Sea and the Bonnet Plume Basin suggests that the limit represents a single (although possibly time-transgressive) glacial maximum. The absence of exposures of glacigenic sediment along the limit between McDougall Pass and the headwaters of the Caribou River, and the lack of units that can be correlated between the Rat River area and the central and southern Peel Plateau, mean that the continuity of the limit can only be inferred using geomorphological criteria.

The source of the glacial ice in the Rat River valley was to the east, in the Canadian Shield, as indicated by the granitic and gneissic clasts. The presence of ultramafic clasts suggests that the ice may have overridden the Muskox complex east of Great Bear Lake (Barager and Donaldson 1973). In the Caribou River watershed and Snake River valley, the presence of granitic clasts also indicates an ice source in the Canadian Shield. The clast assemblages in the southern and central Peel Plateau do not contain similar ultramafic clasts, however. The possibility of distinct glacial lobes in the northern and southern parts of the Peel Plateau is therefore suggested by the differences in mineralogy between the sediments, the patterns of glacial flow, and the lack of correlatability between the regions. Further research is necessary to assess this hypothesis.

Retreat positions, marked by meltwater channels and belts of undulating and hummocky moraine, are present throughout the Peel Plateau. No conclusive evidence for glacial readvances is present, although minor events may have occurred in both the Caribou River valley (as represented by the sediments at HHC 82-1d, 1e, 1f, and 1g), and the Snake River valley (as represented by the youngest diamicton at HH 76-3b).

Glacial retreat throughout the study region resulted in the impoundment of lakes against the receding ice front. The topographically lower lacustrine sequence exposed along the Caribou River was formed after withdrawal of the ice from the upper Caribou River valley, c. 13,000 B.P. The time of re-establishment of vegetation in the vicinity of section HH 72-49, ^{14}C-dated at 12,400 ± 120 B.P. (GSC-3691, Table 32), represents an absolute minimum for the deglaciation of the region. This phase of glacial retreat is correlated to the Sitidgi Lake Stade of Hughes (Geological Survey of Canada, personal communication).

Withdrawal of the ice from the region was probably completed by 12,000 B.P. Deglaciation of the Fort Good Hope area to the southeast is known to have occurred before 11,600 B.P. (Mathews and Mackay 1973). Establishment of the modern river valleys in the Peel Plateau began shortly after deglaciation, but initially proceeded slowly. A ^{14}C date of

9780 ± 110 B.P. (GSC-3573) (Table 32) obtained from wood in fluvial sediments 109 m above the level of the Caribou River indicates that the majority of fluvial downcutting did not occur until after deglaciation of the area.

A gradual pattern of climatic amelioration occurred throughout the early and middle Holocene, superimposed upon local conditions related to aspect and fluvial activity. The area was initially colonized by herbaceous vegetation, c. 12,400 B.P., as indicated by the palynomorphic succession at HH 72-49 (Figure 60). <u>Picea</u> was established in the region before c. 11,000 B.P., with arboreal <u>Betula</u> following shortly thereafter, but not becoming dominant. <u>Alnus</u> may have been in the region by 7370 B.P., as indicated by the record at HH 72-49.

Little evidence for regional climatic change during the late Holocene is present. It is probable that local climatic factors influenced these valley-marginal sites investigated to such extents that regional variations were completely obscured. The number of sites investigated at present is not sufficient to permit the formulation of models of regional climatic fluctuations except in broad chronological terms.

X. Summary

1. The Peel Plateau-Richardson Mountains Region was glaciated at least three times during the Quaternary. The initial glaciation deposited sediments which were reworked to form the 'Brown Bear' fluvial deposits. The second glaciation deposited the 'Snake River till' in the Snake River watershed, and was responsible for the initial development of the ridge-and-valley complexes along the Peel River. These two glacial events were not recognized in the Rat River Valley. The most recent glaciation resulted in the deposition of the "Hungry Creek Till" in the Bonnet Plume Basin, and correlative tills and other glacigenic sediments in the Snake River, Caribou-Brown Bear, and Rat River areas. Correlation between the southern areas and the Rat River Valley is tenuous, and further work is required.

2. Lacustrine sediments formed during the deglaciation of the area were deposited in lakes abutting the ice front. The majority of the sediments originated from the deglaciated areas to the west of the lakes, rather than from the ice margin. The patterns of sedimentation predominantly reflect deposition from underflows.

3. Braided, braided-meandering, meandering, ingrown meandering, and straight streams are present in the region. Each of these stream types has produced a distinctive assemblage of sediments. With the exception of the staight reach, all of the river types present in the region currently are also represented in the stratigraphic record.

4. Palynomorph assemblages reflect the influences of local sedimentary environmental settings, as well as regional climatic conditions. A period characterised by a climate more moderate than today's prevailed prior to the last glaciation of the region, as indicated by *Pinus* and other warm-temperature pollen in sediments at HH 72-50 and HH 62-107. Postglacial palynomorph assemblages were influenced primarily by the local geological setting, rather than by the regional climate.

5. Future research must address the problem of correlation between the northern and southern sections of the Peel Plateau. The lack of success in correlation may be attributable to the existence of separate lobes of ice during the final glacial advance. Study of the area south of Great Bear Lake (Camsell River) will help to resolve this problem. The provenance of many potentially diagnostic clasts cannot be determined at present, and future research in the regions to the east and southeast is required to resolve this problem. Such research would also clarify the relationship between the ice in the southern and central Mackenzie River Valley and that in the western District of Keewatin.

6. Problems which were not addressed in detail by this work include loess characteristics, periglacial processes, and paludal sedimentation. These aspects of the Quaternary geology of the Peel Plateau are of potential interest, and should be addressed by future work.

7. Additional palynological investigations should be conducted for the purpose of constructing a regional climatic picture, both for Holocene and pre-Holocene time.

8. The chronology of events in the region, and throughout the Beaufort Sea area, remains regrettably unclear. Many of the ^{14}C dates which have been obtained are susceptible to question. Many events remain undated. The chronological data available cannot be resolved into a single scenario of events at the present time. Further research, both in the Peel Plateau-Richardson Mountains Region and in the surrounding regions, is required.

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XII. Appendix 1: Location of principal sections

Section	Latitude(N)	Longitude(W)
1. HH 62-62(82)	65°26' N	132°26' W
2. HH 62-64(82)	65°30' N	132°27' W
3. HH 62-71(82)	65°46' N	133°19' W
4. HH 62-74a(82)	65°53' N	133°08' W
5. HH 62-74b(82)	65°53' N	133°08' W
6. HH 62-107(81-1)	67°39.5' N	135°28' W
7. HH 62-107(81-2a)	67°39.5' N	135°28' W
8. HH 62-107(81-2b)	67°39.5' N	135°28' W
9. HH 62-107(81-2c)	67°39.5' N	135°28' W
10. HH 62-107(81-3a)	67°40' N	135°28' W
11. HH 62-107(81-3b)	67°40' N	135°28' W
12. HH 62-107(81-3c)	67°40' N	135°28' W
13. HH 62-107(81-3d)	67°40' N	135°28' W
14. HH 62-107(81-3e)	67°40' N	135°28' W
15. HH 62-107(81-4a)	67°39' N	135°30' W
16. HH 62-107(81-4b)	67°39' N	135°30' W
17. HH 62-107(81-4c)	67°39' N	135°30' W
18. HH 62-107(81-4d)	67°39' N	135°30' W
19. HH 62-107(81-4e)	67°39' N	135°30' W
20. HH 62-107(81-5a)	67°39.5' N	135°30' W
21. HH 62-107(81-5b)	67°39.5' N	135°30' W
22. HH 62-107(81-6)	67°39' N	135°28' W
23. HH 62-115(81)	67°40' N	135°37' W
24. HH 62-116(81)	67°42' N	135°48' W
25. HH 62-117(81)	67°43' N	135°51' W
26. HH 62-142(81)	66°30' N	134°08' W
27. HH 62-145(81)	66°33' N	134°14' W
28. HH 62-147(81)	66°38' N	134°23' W
29. HH 62-148(81-1)	66°40' N	134°32' W
30. HH 62-148(81-2)	66°40' N	134°32' W
31. HH 62-148(81-3)	66°40' N	134°32' W
32. HH 62-148(81-4)	66°40' N	134°32' W
33. HH 62-148(81-5)	66°40' N	134°32' W
34. HH 62-148(81-6a)	66°40' N	134°32' W
35. HH 62-148(81-6b)	66°40' N	134°32' W
36. HH 62-148(81-6c)	66°40' N	134°32' W
37. HH 76-3(82-a)	65°48' N	133°18.5' W
38. HH 76-3(82-b)	65°48' N	133°18.5' W
39. HH 76-3(82-c)	65°48' N	133°18.5' W
40. HH 76-3(82-d)	65°48' N	133°18.5' W
41. HH 76-3(82-e)	65°48' N	133°18.5' W
42. HH 76-3(82-f)	65°48' N	133°18.5' W
43. HH 76-3(82-g)	65°48' N	133°18.5' W
44. HHC (81-1a)	67°42' N	135°49' W
45. HHC (81-1b)	67°42' N	135°49' W
46. HHC (81-1c)	67°42' N	135°49' W
47. HHC (81-1d)	67°42' N	135°49' W
48. HHC (81-1e)	67°42' N	135°49' W
49. HHC (81-1f)	67°42' N	135°49' W
50. HHC (81-1g)	67°42' N	135°49' W
51. HHC (81-2)	67°43' N	135°49.5' W
52. HHC (81-3)	67°43' N	135°51' W

53. HHC (81-4)	67°42' N	135°48' W
54. HHC (81-5)	67°43' N	135°48' W
55. HHC (81-6)	67°44' N	136°03' W
56. HHC (81-7)	66°03' N	134°11' W
57. HHC (81-8)	65°57' N	134°15' W
58. HHC (82-1a)	66°22' N	134°20' W
59. HHC (82-1b)	66°22' N	134°20' W
60. HHC (82-1c)	66°22' N	134°20' W
61. HHC (82-1d)	66°22' N	134°20' W
62. HHC (82-1e)	66°22' N	134°20' W
63. HHC (82-1f)	66°22' N	134°20' W
64. HHC (82-1g)	66°22' N	134°20' W
65. HH 72-49	66°13' N	135°11' W
66. HH 72-50	66°15' N	134°57' W

XIII. Appendix 2: Holocene sedimentary sections

Alluvial fans	Aklavik Range	67°58' N 135°40' W
		67°59'N 135°40' W
		68°00'N 135°40'W
	McDougall pass	67°41'N 136°16'W
		67°42'N 136°22'W
		67°42'N 136°26'W
		67°43'N 136°20'W
		Snake River Area
		65°26'N 132°24'W
		65°27'N 132°25'W
		65°28'N 132°26'W
		65°29'N 132°23'W
Loess	McDougall Pass	67°41'N 136°16'W
		67°43'N 136°20'W
		67°44'N 136°20'W
	Dempster Highway	67°04'N 136°14'W
	Canyon Ranges	65°25'N 132°24'W
		65°26'N 132°24'W
Alpine braided streams	Tributary to Vittrekwa River	67°02'N 136°06'W
	Sheep Creek	67°41'N 136°16'W
	Barrier River	67°28'N 136°06'W
	Tributary to Rat River	67°46'N 135°51'W
	Tributary to Caribou River	66°11'N 136°10'W
Meandering Streams	Tetlit Creek	66°43'N 135°32'W
	Nihtal Git Creek	66°03'N 134°25'W
	Brown Bear Creek	66°40'N 134°24'W
	Rat River	67°40'-44'N 135°42'-55'W
Braided-meandering streams	Rat River	67°38'-41'N 135°27'-30'W
	Peel River	66°30'N 134°08'W
	Peel River	66°33'N 134°14'W
	Peel River	66°40'N 134°32'W
	Vittrekwa River	67°10'N 135°17'W
	Caribou River	66°13'N 135°11'W
	Caribou River	66°15'N 134°57'W
	Caribou River	66°22'N 134°20'W
	Rat River	67°44'N 136°03'W
	Snake River	65°28'N 132°26'W
	Snake River	65°30'N 132°26'W
	Snake River	65°48'N 133°18.5'W
	Snake River	65°53'N 133°08'W
Ingrown meandering streams	Peel River	65°56'-58'N 135°25'-35'W
Lower reach, Peel River		67°15'N 134°52'W
		67°22'N 134°51'W

XIV. Appendix 3: Alpine Sub-Aerial Deposits and Landforms

A. Alluvial Fans

Alluvial fans are not generally areally significant along the eastern flank of the Richardson Mountains. Extensive fans are present along the eastern escarpment of the Aklavik Range, and minor fans are developed along the flanks of McDougall Pass. Fans are also present along the margins of the Snake River in the northern portion of the Canyon Ranges. Pleistocene and early Holocene alluvial fan deposits are confined to the regions of modern fan development. The sedimentation pattern of the fans is controlled by climatic and lithological factors.

Aklavik Range Fans

The alluvial fans developed along the Aklavik Range escarpment form a series of steep, narrow, overlapping aprons. Major fans range in length from 2 km to 4 km, and maximum widths vary from 700 m to 3 km. Fan head gradients are typically 0.04 to 0.10, declining to 0.005 to 0.020 in ~~the toe~~ region. Portions of the fans occupied by active streams show even, regular concave profiles, with no terracing. The areas without active streams are characterized by undulating topography, and commonly exhibit subparallel ridges oriented at right angles to the fan slope. The fan margins are convex, with no digitate or lobate projections. Braided stream channels are present along the entire length of most of the fans, although the toe areas are affected by periodic sheetflow. The trunk channels terminate within the fan head region, thus forming 'intersection points' (as defined by Hooke, 1967). Frost blisters (to 1 m high) and high-centre tundra polygons are frequently present in inactive zones.

Vegetation is primarily confined to inactive stream channels in the midfan region, and in the fan head region of large fans. The fan head zones of small fans are usually covered with vegetation, and midfan zones may be completely vegetated as well. Depressions between the subparallel ridges of inactive fan segments are also commonly vegetated, as are the toe regions of large fans.

Exposures of fan sediment are poor and of limited vertical extent. The surface of the fans is composed primarily of non-plastic silt and fine angular shale particles, derived from the Cretaceous bedrock exposed along the escarpment. Scattered sandstone and granitic granules are dispersed throughout the deposits. Sediments in the channel areas are coarser than those in marginal zones, but are similiar in lithology and clast shape. The subparallel ridges developed on inactive zones are also composed of silt and angular shale fragments.

Legget et al., (1966) investigated two similiar fans developed along the Aklavik Range escarpment north of 68°N. These fans were also dominated by silt derived from Cretaceous bedrock, combined with thin layers of mudstone clasts, scattered coarse sand grains and sandstone pebbles, and organic matter derived from the grass and sedge vegetation of the fan surfaces. Boreholes penetrated in excess of 12 m (40') of silt without reaching the base of the fan sediment. Segregated ice layers were present within the fan complexes, and frost blisters and polygonal ground were developed on the fan surface.

Interpretation

The Aklavik Range fans are dominated by subaerial mass movement processes. The large proportion of silt and clay aids in the propagation of debris flows, either by increasing the cohesive strength (Rodine and Johnson, 1976) or by generating excess pore pressure through low permeability (Pierson, 1981). Internal deformation by shearing during flow was probably non-existent or insignificant, as suggested by the non-plastic nature of the sediment.

The undulating surface of inactive regions of these fans is indicative of multiple mass movements. Individual debris flows tend to have smooth upper surfaces (Bull, 1977), and therefore the undulations indicate multiple events. Multiple events could be triggered by successive heavy rainfalls (Bull, 1977), but the climatic pattern of the region suggests that such events are rare. Alternatively, rapid snowmelt and slush avalanches could trigger the flows (Caine, 1969; Theakstone, 1982).

The presence of permafrost throughout the fan sediments, as indicated by the frost blisters and polygonal ground developed on the surfaces and the segregated ice lenses noted by Legget et al. (1966), has a large influence on mass movement processes. Saturated silt resting on the permafrost table is highly prone to downslope movement by gelifluction, and the required gradient for motion initiation can be less than 0.02 (St-Onge, 1965). This gradient is exceeded at almost all points on the Aklavik Range fans. The subparallel ridges and lack of pronounced lobate projections or digitate margins suggest slow moving, highly viscous flows, or creeping motion as would be produced by gelifluction of non-plastic material (Washburn, 1980).

The deposits produced by rapid debris flow and slow gelifluction of silt and shale fragments would both be dominated by massive, matrix-supported poorly-sorted diamictons with fabric either random or oriented parallel to the gradient (Enos, 1977; Heward, 1978; Nardin et al., 1979; Kerr, 1984; French, 1971; Washburn, 1980). In the absence of intensive research on well-exposed sections, distinguishing between these two processes is not possible. Both rapid, snowmelt-induced debris flows and slow gelifluction probably occur, in addition to frost creep. Degradation of segregated ice lenses may also contribute both to debris flow activity and the formation of irregularly undulating topography.

Stream activity is a lesser component in the development of the Aklavik Range fans. The sediment within the channels differs little from the mass movement material, primarily because of the uniformity of the source material. The absence of textural contrasts obscures the recognition of distinctive fluvial facies, such as those proposed for alluvial fans by Hooke (1967), Boothroyd and Ashley (1975), and Miall (1978). This is the result of the highly seasonal nature of stream flow and the consequent annual exposure of the stream bed to subaerial conditions, which accentuate the effects of frost weathering and shattering.

Stream channels cause degradation of the permafrost directly beneath the thalwegs. Upon abondonment of the channel, this thawed area is moist and sheltered, and forms a suitable environment for plant growth. Once the vegetation is established, the probability of mass movement is reduced, and thus abandoned channels quickly become occupied by ribbons of vegetation. In the toe regions of the larger fans, and over the entire surface of some small fans, channels are less well-defined or absent. Vegetation is established here as a result of sheetflow, which transports silt and organic detritus at slow, steady rates during spring and summer. The sheetflow may be initiated by precipitation, runoff from low viscosity debris flows, or degradation of segregated ice.

McDougall Pass and Canyon Range Fans

The fans flanking McDougall Pass, and those discharging into the Snake River from the northern Canyon Ranges, form distinct, non-coalescing landforms (Plates 39 and 40). These fans range in width from less than 1 km to 3 km, and have maximum widths between 0.5 km and 4 km. Gradients decline from 0.03 - 0.15 in the fan head areas to 0.004 - 0.02 in the toe regions. Concave, regular profiles are commonly developed in the active portions of the fans, whereas the inactive portions may show irregular breaks in slope which are not consistent in elevation across the fan surface. Some inactive zones do not show pronounced breaks in slope. Subparallel ridges and strongly undulating topography are generally not present in active or inactive zones.

The McDougall Pass fans are mainly inactive, and entrenchment by braided streams is common. The fan margins are irregularly convex, without pronounced digitation. subsidiary fans, developed downstream from the downcut fan, occur only rarely. Vegetation covers the majority of the fan surfaces, except in areas where coarse gravel is exposed.

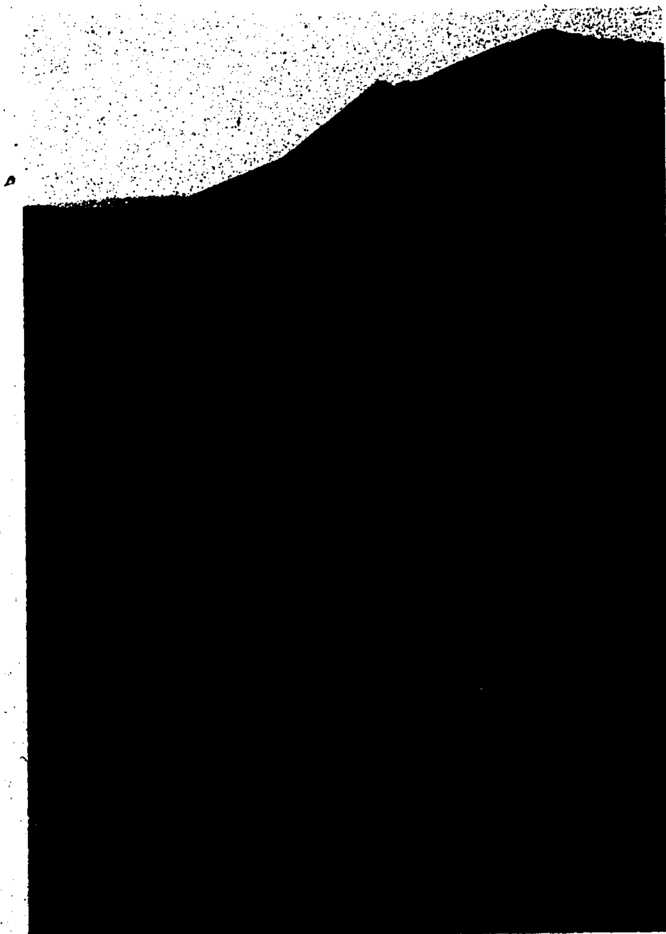
Exposures in the McDougall Pass area fans are generally poor. Extensive slumping along downcut intrafan streams has obscured the internal structure of the fan sediments. The most extensively exposed sequence of fan sediments, located along Sheep Creek, is dominated by locally-derived gravel, with sandy and silty beds. The units appear to be laterally continuous parallel to the creek. Sorting in the gravel units is poor, and the clasts are angular. Tabular shale fragments are common. The fan toe is situated at approximately 310 m A.S.L., about 20 m above the Rat River.

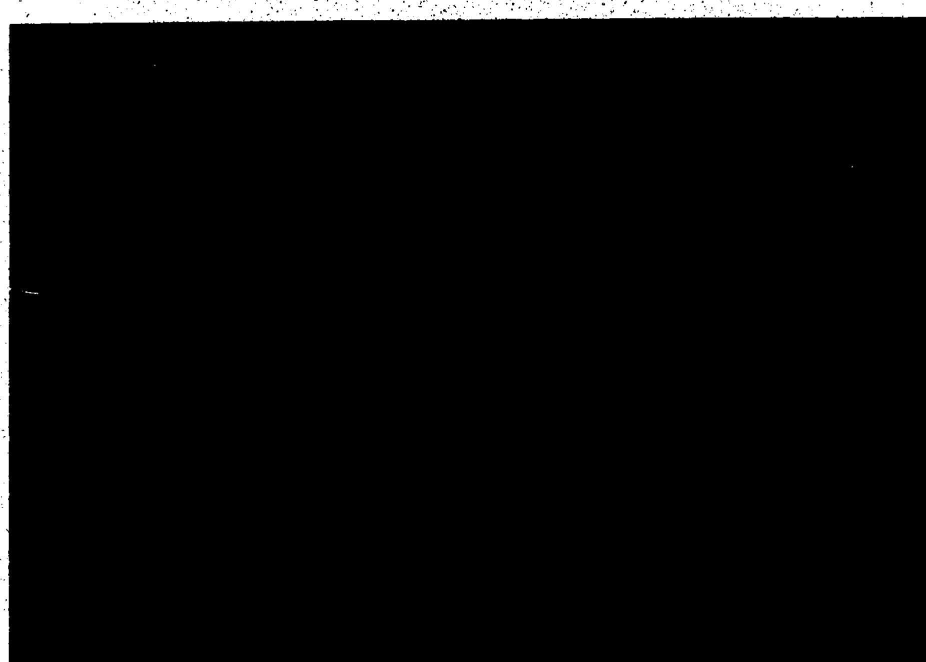
Plate 39 (top)

Solifluction-dominated Alluvial Fan, Sheep Creek Area, eastern approach to McDougall Pass.

Plate 40 (bottom)

Alluvial Fan, Section HH 62-62, Snake River Valley.





In contrast, the fans bordering the Snake River are largely active. Entrenchment by the trunk channel terminates in the fan head zone. The fan margins are convex, without digitation or lobate projections. Commonly, the convex margin is symmetrical about the axis of the fan's drainage system. subsidiary fans and extensively downcut channels are absent. Vegetation is confined to abandoned channels and the toe areas of larger fans, and the entire fan surface is frequently devoid of vascular plants.

Exposures of sediment within the Snake River fans are extensive. Most of the fans extend to the level of the Snake River, and lateral erosion by the stream has resulted in several exposures of sediment in temporarily inactive zones. A single fan at locality HH 62-62 (82) (65° 28' N, 132° 26' W) was selected for investigation in detail, and other fans were observed briefly.

In the toe area, the sedimentary succession is dominated by matrix-supported structureless gravel beds 10 cm to 2 m thick, without consistent imbrication. Bed contacts are generally planar. Clast sizes vary from granules to coarse pebbles and the clasts are subrounded and blade-shaped. The clasts are prodominantly derived from the adjacent Devonian shale and carbonate bedrock, but jasper-hematite, othoquartzite, chert, arkose, basic volcanic rock, and granitic granules are also present. The matrix is coarse to medium-grained sand, with less than 1% silt and clay. The sand grains are similar lithologically to the larger clasts.

Interbedded within the massive gravel units are moderately sorted beds of imbricated gravel, clast- or matrix-supported, with clasts imbricated up-fan. Poorly developed fabric, with long axes aligned roughly perpendicular to fan slope, is commonly present. The imbricated units are 10 cm to 1 m thick, have planar contacts, and are laterally continuous. The matrix is coarse to medium sand, well-graded and moderately sorted, with less than 5% silt and clay. Lithologically, the deposits are dominated by locally-derived clasts, although orthoquartzite, chert, jasper-hematite, and basic volcanic rock fragments were observed. Granitic clasts were not detected. Both clasts and matrix are dominated by subrounded to rounded, bladed to roller-shaped particles.

Scattered throughout both the massive and imbricated gravel beds are normally graded lenses of sediment. The lenses are trough or scoop-shaped, 10 cm to 80 cm thick, 20 cm to 1.2 m in width, and range in length from 50 cm to in excess of 3.5 m, when exposed in three dimensions. Grading in the lenses is cyclic, from granules at the base to fine sand and silt at the top of each cycle. Individual cycles are 8 cm to 15 cm thick, and lenses commonly display multiple successions. The contacts beneath each granule bed are erosional whereas other contacts in the cycles are gradational. Organic-rich detrital laminations are present within some silt layers. Authigenic calcium and magnesium carbonate and sulphate crystals are present in the coarser sediments.

The lenses are oriented with their long axis inclined downfan and approximately parallel to the maximum gradient of the fan. The gradients of the basal contact surfaces of the lenses are approximately 0.01 to 0.02.

Silty clay and clayey silt beds and lenses are also present within the massive gravel units. A single silt lens was noted interbedded within an imbricated gravel bed. The clay and silt units vary in maximum thickness from 10 cm to 40 cm, and in some instances extend across the entire width of the outcrop. The sediments are moderately to poorly sorted, but contain no clasts larger than fine sand. Iron-oxidation features are present in the upper portions of the strata.

Laminations and thin beds of silt rich in organic detritus are commonly present within the finer sediments. These organic units are continuous acroos the width of the clay and silt ~~strata~~, and maintain constant thickness. The detritus is generally fine, although small twigs (often flattened) are commonly present.

Palynological analysis of two samples obtained from locality HH 62-62 (82) revealed assemblages which had undergone extensive chemical corrosion, and had been attacked by fungi and bacteria, as indicated by the scar types (criteria of Elsik, 1966, 1971), and the presence of fungal spores. Deterioration of the grains was unusually complete. The remnant assemblages were dominated by Lycopodium grains and other spores. Pre-Quaternary material comprised at least 60% of the spectra. Further details are presented in Appendix 2.

A sample from the horizon analysed for palynomorphs has been ^{14}C dated at 8510±70 B.P. (GSC-3695). The possible presence of small amounts of contemporaneous rootlets (W. Blake, Jr., Geological Survey of Canada, personal communication) suggests that this date represents a minimum age for sediment deposition. No rootlets were observed during sampling of the deposit, however, although fungal filaments were detected during palynological analysis.

In addition to the imbricated gravel beds and sorted lenses of finer sediment, coarse, very poorly sorted matrix-supported gravel beds 1 - 3 m thick are also interbedded with the massive gravel strata. Some of these units are sufficiently poorly sorted to be considered diamictons. Clast sizes range from granules to boulders. The clasts are angular to subangular and generally have low sphericity, although a variety of shapes are present. Local lithologies predominate, although the orthoquartzite content is high (maximum 41%). The matrix is composed of subangular coarse sand to silt grains of similar lithology. Imbrication is absent, and the fabric is either random or weakly developed, with long axes trending downfan. These units tend to be laterally discontinuous. No sand or silt lenses are present.

All three gravel facies observed in the toe areas were also present in the mid-fan and fan head zones. The proportion of imbricated gravel and massive, matrix-supported, rounded clast gravel relative to the coarse, poorly-sorted gravel/diamicton facies decreases in the up-fan direction. The modal clast size and proportion of the matrix material remains constant. The matrix and clasts are dominated lithologically by material derived from the surrounding bedrock, and clasts from outside the immediate drainage basin are extremely rare. Transverse gravel ribs are present in stream channels, but were actively being eroded in August 1982.

In the midfan zone, sand and granule gravel lenses are rare, discontinuous parallel to the fan slope, and are confined within the massive, rounded clast gravel units. The internal structure of these lenses is similar to those in the toe area, but normally only a single fining-upward sequence is preserved. The maximum number of cycles observed in a single lens was three. No silty clay or organic rich beds or lenses were observed.

The fan head areas lack fine-grained sediment lenses. Sand is present in the base of the active trunk channels as apparently discrete ribbons and lenses. Fine organic detritus was observed in depressions in the sand, and in the lee of cobbles and boulders.

Interpretation

Limited exposures of the McDougall Pass fans make any interpretation tenuous. The texture and clast shape suggest that the bulk of the sediment was transported by debris flow. The sorted sandy and silty units may have been deposited in ephemeral channels. On the limited data available, these sediments appear to be similar in most aspects to the better-exposed fan sediments bordering the Snake River.

The fan head and midfan portions of the active fans along the Snake River are dominated by mass movement processes, whereas stream sedimentation is more significant at the toes. The very coarse matrix-supported poorly sorted gravel and diamicton units which are dominant in the fan head areas and subdominant in the toes are similar to strata interpreted as alluvial fan debris flow deposits by Hooke (1967), Bull (1977), Heward (1978), and Miall (1978), among other researchers. The absence of imbrication, the angular, bladed clasts, and the lack of sorted lenses all indicate that extensive fluvial reworking has not occurred. The fabric is commonly random, but where developed it is oriented with the long axes parallel to the slope, configurations which have been commonly reported from deposits interpreted to be formed by debris flows (Kerr, 1984; Shultz, 1984). The increases in thickness, areal extent, and clast size in the upfan direction are typical for debris-flow deposits (Bull, 1977). The laterally discontinuous structure of the units suggests deposition by many discrete debris flows.

The characteristics of the strata are compatible with deposition from either pseudoplastic flows or low viscosity Bingham plastic flows, considering the criteria proposed by Enos (1977), Lowe (1979), Nardin _et al._, (1979), and Schultz (1984). This ambiguity indicates that the motion varied between viscous and inertial flow locally, possibly in response to minor changes in slope, substrate permeability, and fine sediment content.

The matrix-supported structureless gravel units which comprise the bulk of the toe zone sediments and lesser proportions of the midfan and fan head successions are interpreted to be stream deposits. The very low proportion of silt and clay is incompatible with a debris-flow origin. The non-ordered fabric was produced by fluctuations in stream flow direction, bed armouring effects, and collapse of banks cut by shallow ephemeral channels on the fan surfaces. Similar channel fill deposits were noted by Bull (1977) and Kerr (1984). The planar nature of the basal contacts precludes deposition in deep channels.

The vertical decrease in lithologic variability reflects a change in sediment source for the fans. Initially, glacial and fluvial sediment along the Snake River was incorporated into the fans, producing beds of varied lithology. As the fans prograded, this Cenozoic sediment was overstepped by material derived from the immediately adjacent bedrock. This 'inverse stratigraphy' derived from successive erosion of older deposits has been reported from other fan successions by Laming (1966) and Heward (1978).

Structureless coarse gravel deposits devoid of fine sediments developed on arid California fans were interpreted as sieve deposits by Hooke (1967), formed when flowing water removed fine material from gravel and coarser beds. These deposits were formed on steep slopes, and are dominantly composed of blocky clasts, such as those produced by a strongly jointed resistant outcrop. For the Snake River fans, the presence of the coarse deposits on the gently sloping distal areas; the high proportion of matirx material; and the subrounded clasts are all incompatible with formation by sieving.

The moderately sorted, imbricated gravels are also interpreted as streamflow deposits. The absence of scoured contacts and the apparent lateral continuity of the deposits suggests that flow occurred in shallow, constantly shifting, ephemeral channels or (for some thin units) sheetfloods (sense of McGee, 1897 and Hogg, 1982). Current climatic conditions in the Canyon Ranges make high capacity sheetfloods unlikely, however. The ~~matrix~~ material may have been incorporated by infiltration during declining water stages after peak discharge (Wasson, 1977). The low concentration of fine sediment suggests that slurry flow did not occur.

The presence of transverse ribs in the youngest gravel deposits of the fan head area suggest that peak water flow was sufficient to form gravel antidunes (Shaw and Kellerhals, 1977; Koster, 1978). The structures were being actively reworked in August 1982, however, suggesting that the preservation potential is relatively low. Transverse ribs were confined to the trunk channels, and water flow rates in the marginal zones were probably not sufficient for antidune development. No transverse ribs deposits could be recognized in exposures.

The normally graded granule to silt lenses are interpreted to be ephemeral channel deposits. The trough or scoop-shape of the lenses: the low-angle downfan inclination: the fine detrital organic layers; the erosional contacts beneath the granule strata; and the cyclic multistorey bedding, all suggest deposition in temporary channels developed on the fan surfaces. Similar structures have been noted in the Scala Dei Group in northeastern Spain (Allen et al., 1983). The absence of these lenses from the fan head area, and the single cycles noted in the midfan zone, reflect the relative stability of the trunk channel and primary distributaries. Sands and silts deposited in these channels during low water stages would be prone to reworking and downfan transportation.

The silty clay and detrital organic lenses can also be considered as channel fill deposits formed during the waning stages of water flow. The laterally continuous beds, however, represent deposition from sheetflow or sheetflood activity. The texture of the sediments and the interbedding of fine organic detritus are more typical of slow, periodic sheetflow events (as described by Horton, 1945 and Moss and Walker 1968), than of rapid, high-intensity sheetfloods (as described by McGee, 1897; and Rahn, 1967). The presence of vegetation on the toes of active fans is also more compatible with the sheetflow hypothesis. Variations in moisture content of the sediment are indicated by the oxidised horizons, the authigenic carbonate and sulphate minerals, and the degradation of the palynomorph assemblage obtained from HH 62-62 (82).

The general downfan trend exhibited by the Snake River fans is a transition from dominant mass-movement processes in the fan heads to stream flow and sheet flow processes in the toes. The very coarse, massive gravel and diamicton units are succeeded downfan by massive matrix-supported gravel units lacking in silt and clay; imbricated gravels; graded channel fill deposits; and fine sheetflow sediments. The mineralogical and lithological compositions reflect successive reworking and erosion of glacial and Cenozoic fluvial deposits, followed by erosion of the bedrock.

Summarial Discussion

The alluvial fans developed along the eastern escarpment of the Aklavik Range, flanking McDougall Pass, and in the Snake River valley, were all formed by combinations of subaerial mass movement and fluvial processes. The relative importance of these processes depended (and still depends) upon the topographic setting, type of sediment available, and degree of fan activity.

In the immediate post-glacial period, ablation of the ice sheet permitted the development of fans in McDougall Pass. Growth and maintenance of these fans depended upon the increased rainfall generated locally as a consequence of glacial ablation. Ryder (1971) and Church and Ryder (1972) used the term "paraglacial" to describe non-glacial processes directly conditioned by glaciation. Since fan development in the McDougall Pass was initiated as a result of conditions engendered by the glacier, and since cessation of these conditions has resulted in the termination of fan development and the commencement of active stream downcutting, the McDougall Pass fans are paraglacial.

The Snake River fans were initially supplied with glacial and fluvial debris. These fans could therefore be considered paraglacial in the sedimentological sense. Continuation of fan development to the present, however, indicates that the immediate post-glacial climate was not essential for fan growth, and so the Snake River fans are not paraglacial in the climatic sense. Similarly, the development of the Aklavik Range fans did not depend on the post-glacial climate. The lack of exposures at the bases of these fans prevents the assessment of the importance of glacially- and fluvially-derived sediment in fan development.

Schumm (1977) proposed that alluvial fans could be classified according to climatic conditions under which they developed. "Arid" or "dry" fans, as exemplified by deposits described by Bull (1964, 1977), Denny (1967), and Hooke (1967), are dominated by debris flow sediments, although stream flow, sheetflow, and sieve deposits are also presnt and frequently significant. "Humid" or "wet" fans are considered to lack debris flow deposits by Schumm (1977), and sedimentation patterns were suggested to resemble those of proximal braided streams. Although some humid-region fans are composed primarily of stream deposits (eg. Boothroyd and Ashley, 1975; Boothroyd and Nummedal, 1978), others show pronounced downfan zonation (eg. Cherven, 1984). In local cases, humid fans may be dominated by mudflows (Winder, 1965; Ryder, 1971). Miall (1978) proposed two models for fan sedimentation: the arid "Trollheim" type (based on the work of Hooke , 1967); and the humid "Scott" type (based on the work of Gustavson, 1974; and Boothroyd and Ashley, 1975). Miall suggested that the models should be regarded as end-members of a continuum.

With 250 mm of precipitation annually, falling almost exclusively during the summer months, and a mean annual temperature of - 10°C (Burns, 1974), the climate of Inuvik can be classified as subhumid to semiarid. The mountainous regions are colder and drier than Inuvik, and the fans thus developed primarily under semi-arid conditions with summer water influx. The availability of water is increased locally through degradatión of segregated ice during the summer, especially where the silt content of the fans is high.

Under these conditions, streamflow deposition is dominant in fan-toe areas where the fan is developed from silt-poor material, as is the case along the Snake River. In the fanhead and mid-fan zones, debris flow activity dominates. Fans built from silty bedrock, as is the situation along the Aklavik Range escarpment, are dominated throughout their lengths by debris flows and/or gelifluction deposits. Gelifluction does not appear to be significant in gravel-dominated fans. Degradation of ground ice in finer sediments creates surface irregularities, lubricates loose debris on the surface, and promotes sheetflow.

The dominant control on the mode of fan development and sedimentation is the composition and texture of the source material, rather than the aridity of the climate. Although the fans cannot be considered as typical "Trollheim" type arid-region landforms, the overall preponderance of sediment gravity flow deposits indicates that these processes are dominant. Stream flow plays an ancillary role in the proximal regions of the fans, and is dominant only in the toes of coarse-sediment landforms.

B. Other Deposits and Landforms

The alpine sub-aerial environment in the Richardson Mountains is characterized by extensive felsenmeers developed from highly jointed bedrock, stone stripes and solifluction features developed on fine-grained bedrock, scattered organic tundra deposits, and thin veneers of loess in exposed areas. Colluvial sediments are present along steeply sloping valleys.

The periglacial features are more extensively developed in unglaciated areas. The development of these features is dependent upon the interaction of climatic, bedrock and hydrologic factors. Analysis of the periglacial features was not the primary focus of this study, and extensive research would be required to assess the importance of the processes and factors involved.

Loess deposits in the alpine areas are poorly to very poorly sorted, containing coarse and medium-grained sand, silt, and clay. Clasts are predominantly subangular to angular, have low to moderate sphericity, and are bladed to discoid. The mineralogy of the deposits is similar to that of the adjacent bedrock. Distally-derived clasts, such as igneous and metamorphic minerals, are present in trace ~~amounts in~~ unglaciated-area loess, and compose 3-10% of loess deposited in glaciated areas. Trace amounts of coarse sand-sixed lignite fragments are present in loess in McDougall Pass and in deposits bordering the Dempster Highway near the Yukon/N.W.T. border. The maximum thickness of loess noted in alpine areas was 30 cm.

The texture of the loess deposits suggests that extensive cryoturbic mixing with the underlying sediments has occurred. Similar very poorly sorted sediment in the Cypress Hills of Alberta was interpreted as cryoturbated loess by Westgate (1968) and Catto (1983). The mineralogical distribution is compatible with aeolian transport from the west, as suggested by the low proportions of igneous and metamorphic minerals in the unglaciated zone loess. Accurate assessment of the transport direction would require extensive investigation, involving the determination of progressive changes in deposit thickness and texture, mineralogical variations and grain orientations.

XV. Appendix 4 - Fluvial Sediments

Sediments representing a variety of fluvial and related environments are well-exposed along the Peel River and its tributaries. Four river types can be recognized in both the modern environment and late Cenozoic sedimentary record: 1) alpine braided, 2) plateau/plain meandering, 3) plateau braided-meandering, and 4) ingrown meandering. The lower reaches of the Peel River constitute a fifth type of stream that has not yet been recognized in older sediments. The sedimentology and development of each river type will be considered separately.

A. Alpine Braided Reaches

Several of the headwater reaches of the modern streams along the eastern slope of the Richardson Mountains and the northern slope of the Canyon Ranges are braided. Most are seasonally ephemeral, and large diurnal fluctuations in flow are normal, with peak flow occurring in the late afternoon. Typical gradients for 10 alpine braided streams are presented in Table 4-1: the modal value is 20 m/km. The sinuosity indices (Leopold and Wolman, 1957) are low (modal 1.2). Typical values calculated from aerial photographs for the braiding index (Brice, 1964) range between 1.4 and 2.8. Braiding parameter values (Rust, 1978) are 2.6 to 5.3.

A tributary stream to the Caribou River, here designated 6-south Caribou, is an example of an alpine braided stream. The stream was first observed on 10 July

Table 4-1: Channel Properties of Braided Stream Reaches

Reach	Gradient (m/m)	Sinuosity *	Braiding Index **	Braiding Parameter ***	Date of Photo
5-N. Vittrekwa					
67°00'-11' N	0.02	1.20	1.9	3.2	July 1972
136°02'-09' W					
8-N. Vittrekwa					
67°02' N	0.05	1.16	1.6	3.1	July 1972
136°05'-08' W					
Sheep Creek					
67°39'-43' N	0.02	1.10	2.5	2.5	Aug. 1954
136°12'-16' W	0.02	1.10	1.9	2.7	July 1972
Barrier R.					
67°28' N	0.04	1.20	1.4	3.0	July 1972
136°05'-10' W					
4-N. Rat					
67°45'-52' N	0.22	1.25	1.8	4.0	June 1971
135°51'-53' W					

cont. Table 4-1

Reach	Gradient (m/m)	Sinuosity *	Braiding Index **	Braiding Parameter ***	Date of Photo
5-N. Vittrekwa					
67°05'-06' N	0.02	1.10	2.9	4.3	Aug. 1954
135°38'-42' W	0.02	1.10	2.5	5.4	July 1972
6-S. Caribou					
66°11'-13' N	0.04	1.20	1.8	2.6	Aug. 1977
136°10' W					
4-N. Trail					
66°26'-27' N	0.03	1.20	1.7	2.9	Aug. 1977
135°07'-10' W					
4-S. Road					
66°30'-34' N	0.02	1.08	2.3	3.3	Aug. 1977
135°50'-52' W					
Road River					
66°34' N	0.02	1.13	2.2	3.6	Aug. 1977
135°50'-55' W					

* after Leopold and Wolman 1957

** after Bryce 1965

*** after Rust 1978

1982, during a light to moderate rainfall which had commenced approximately 6 hours prior to observation. The channels consisted of a succession of shallow zones (riffles) adjacent to gravel bars and pools in the areas where channels joined. The maximum channel depth was 1.4 m. The sinuosity of the channels was low. The gravel and coarse sand was not in motion, and the crests of the bars were composed of unstratified coarse granules and pebbles without finer matrix material. Isolated Equisetum(horsetail) plants were also present on the bar crests. The bars were predominantly parallel to the valley and channel trend, although some transverse forms were present in the upper portion of the reach. Straight-crested ripples were migrating along the active channels.

The stream was observed again on 11 July 1982, approximately 20 hours after rainfall had ceased. At that time, the maximum channel depth was 80 cm, and most of the channels were inactive (Figure A4-1). The channel floors were covered with patches of straight-crested and linguoid asymmetrical ripples, interspersed with coarse granule and pebble gravel infilled with fine sand. No granules or pebbles were present in the areas of ripple development. The confluence areas (pools) were veneered with massive fine and medium-grained sand. In some locations, small foresets to a thickness of 10 cm were deposited around the upstream margins of the pools. Strata within these foresets were 0.5 to 3 cm thick, and were inclined at 15° to 30°. Silt was

FIGURE A4 I

6 -- South Caribou stream reach, as observed on 11 July 1982.

SS Active Stream Channel

C Confluence Zone

O O O Exposed Gravel

Exposed Stright-Crested Ripples

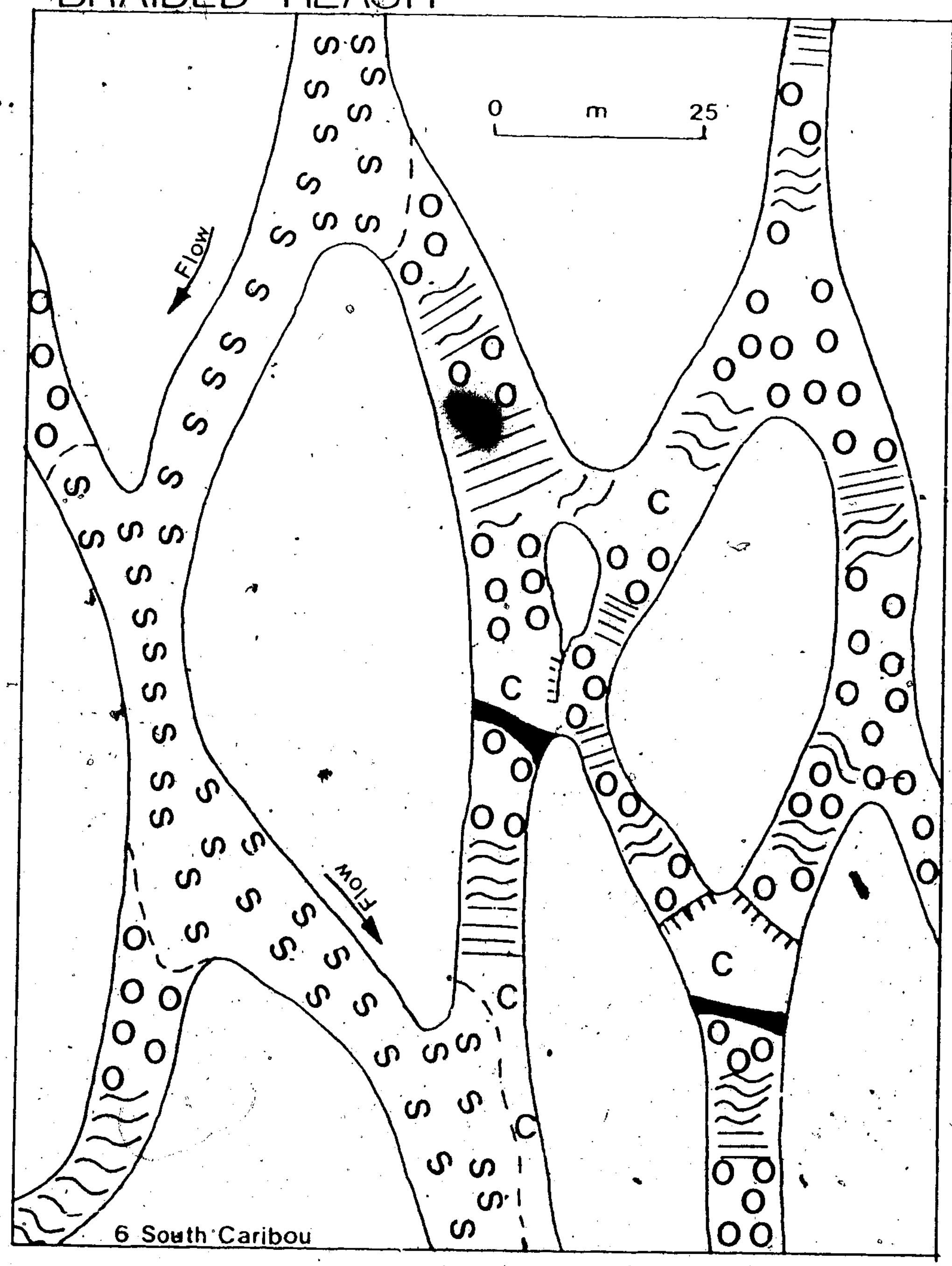
Exposed Linquoid Ripples

Exposed Foreset Beds

Exposed Silt

See text for discussion.

BRAIDED REACH



present only as thin (less than 1 cm thick) veneers in the downstream portions of the pools, and no organic detritus was noted.

The stream was observed briefly from the air on 25 July 1982, after 14 days of little precipitation. At that time, only 2 channels were active, and all of the tributary valleys and flanking slopes appeared dry.

These observations illustrate the inaccuracies inherent in temporally-limited study of alpine braided streams. The bedforms observed represent a composite relict assemblage derived from previous stream flow phases, a situation common to a braided stream environment (eg. Allen, 1973a; Koster, 1978). Calculations of statistical measures of stream form, such as sinuosity and braiding parameters, are subject to similar inaccuracies. The figures presented in Table 4-1, determined from aerial photography, may therefore not be indicative of the 'true' or 'seasonally-adjusted' nature of the streams.

The sedimentary sequence preserved consists of longitudinal bar and coarse channel sediments, with interspersed transverse bar deposits, patches of rippled medium and fine sand, and pool fill sequences of massive sands, foresets, and thin silt veneers (Figure A4-1). The coarse deposits are generally massive, although crude horizontal bedding may be present. Channel sediments contain strata of coarse granules and pebbles infilled with matrices of medium and fine sand, indicating two periods of flow

(Beschta and Jackson, 1979; Frostick et al., 1984). The absence of matrix material on the bar crests indicates that the bars were submerged only during periods of peak flow. The orientation of the clast long axes is generally transverse to the bar long axes, suggesting alignment transverse or oblique to flow. Transverse fabrics have frequently been reported from braided stream deposits (eg. Hein and Walker, 1977). These bars are composed of massive or crudely horizontally stratified gravel, with or without matrix (facies D of Smith (1974); Boothroyd and Ashley (1975); dominant facies of 'Scott' type of Miall (1977, 1978)), and usually represent the majority of the sedimentary succession in the alpine braided stream headwater environment. Transverse forms usually indicate lesser discharge and areas largerly influenced by aggradation. The dominance of longitudinal bars in a stream with relatively minor and highly ephemeral discharge suggests that downstream migration of gravel has taken precedence over aggradation during bar construction. This situation could result from flashy discharge generated by snowmelt or slush avalanching, followed by a sharp decline in flow rate, causing rapid stagnation of the coarse traction load. The bars produced during the initial high flow would thus be starved of sediment, and the aggradation process would not proceed long enough to extensively modify the bedforms.

Peak flows in the early spring would traverse frozen channel bottoms exposed throughout the winter, covered with a veneer of sand and silt. Consequently, the initial granule and pebble load would be transported as a sliding sheet over the frozen sand and silt bed, with relatively little frictional resistance from the substrate and little interference from protrusions from the channel floor. Without high frictional resistance and minor bathymetric obstructions which could serve as nucleating points for bar development, downstream migration of granules and pebbles would be enhanced and aggradation retarded.

Downstream migration of large clasts and longitudinal bar construction are thus favoured by the climatic conditions prevalent in the study area. Discharge rates and slopes in the braided reaches are insufficient to permit the development of transverse ribs and other antidunal structures.

The rippled sand units are similar to those observed in low energy areas of larger alpine streams (eg. facies B of Smith (1974). The absence of sediments representative of intermediate flow conditions (such as trough cross-stratified sands) indicates that the streams are essentially ephemeral.

The preservation of pool-fill sequences beneath pebble and granule gravel beds also reflects the transitory nature of the high flow events. During peak flows, the gravel was transported over the fine sediment without inducing erosion (cf. Foley, 1977; Allen, 1983a). Freezing of the exposed silt during the winter aided the overpassing process.

Pre-modern sediments of the alpine braided stream environment are sporadically exposed in the foothills of the Richardson Mountains and Canyon Ranges. No similar sediments were detected in sections from the Peel Plateau and Peel Plain. The facies exposed in the alpine areas are similar to those present within the modern stream channel and bar systems. Structureless or crudely horizontally stratified pebble and granule gravel representing longitudinal bars and channels dominates the deposits. Lesser amounts of rippled and massive channel-fill sand and pool-fill sequences are also present. The deposits are invariably found in settings typical for modern alpine braided streams, and provide no critical palaeoenvironmental or palaeogeographical information.

B. Plateau/Plain Meandering Reaches

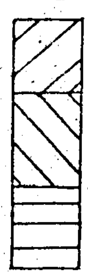
Meandering stream reaches are present in the western portion of the Peel Plateau, and in the Peel Plain. Channel morphology parameters are presented in Table 4-2. Gradients along the plateau reaches are higher than those for the Peel Plain streams. The streams are all characterised by high sinuosity values and low braiding indicies and parameters.

Figure A4-2 depicts a portion of a modern meandering reach. The diagram is schematic and composite. The scale, distribution, and number of features are variable, depending

FIGURE A4 2

Meandering Reach, Schematic Diagram

1 -- Main Channel
2 -- Partially Abandoned Channel
3 -- Avulsion Channel
4 -- Infilled Channel
5 -- Remnant Lake
6 -- Floodplain Tributary, developed in abandoned channel
7 -- Tight Bend Point Bar
8 -- Gentle Bend Point Bar
9 -- Bar Crown
10 - Flank Drift Jam Deposit
11 - Isolated Longitudinal Bar
12 - Erosional Inside Bank
13 - Spring Bank Caving Zone
14 - Spring and Summer Bank Caving Zone
15 - Retrogressive Thaw-Flow Chutes
16 - Retrogressive Thaw-Flow Debris
17 - Low Floodplain with arboreal vegetation
18 - Low Floodplain without arboreal vegetation
19 - High Floodplain
20 - Valley Margin
21 - Tundra Upland
22 - Tributary Stream
T - area with arboreal vegetation
d - area veneered with organic detrital drift



Moisture at or near surface throughout year

Surface dry during summer months

Caving Zones

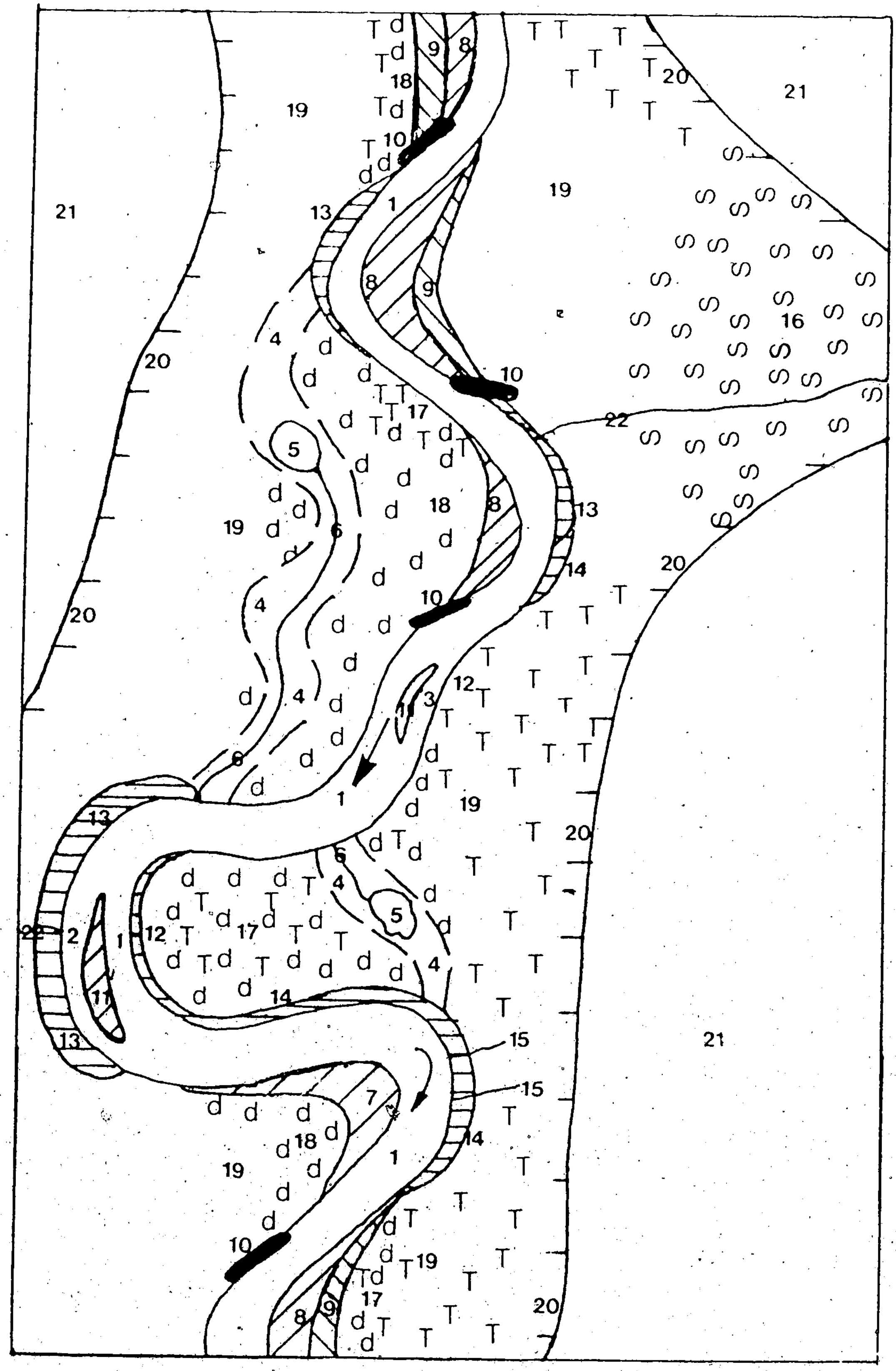

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Table 4-2: Channel Properties of Meandering Stream Reaches

Reach	Gradient (m/m)	Sinuosity *	Braiding Index **	Braiding Parameter ***	Date of Photo
Tetlit Creek 66°43' N 135°30'-38' W	0.19	1.94	0.1	0.2	July 1978
Satah River 66°56'-59' N 134°20'-29' W	0.002	1.63	0.1	0.1	June 1971
Nihtal Git Cr. 66°03'-06' N 134°23'-28' W	0.009	1.70	0.1	0.1	June 1971
Brown Bear Cr. 66°40'-42' N 134°00'-15' W	0.001	1.88	0.1	0.1	Aug. 1973
Brown Bear Cr. 66°39'-40' N 134°15'-25' W	0.002	2.19	0.1	0.1	July 1978

cont. Table 4-2

Reach	Gradient (m/m)	Sinuosity *	Braiding Index **	Braiding Parameter ***	Date of Photo
Rat River					
67°40'-44' N	0.005	1.80	0.4	0.5	Apr. 1954
135°42'-55' W	0.005	1.90	0.3	0.7	June 1971

* after Leopold and Wolman 1957

** after Bryce 1965

*** after Rust 1978

upon the consistency and magnitude of discharge, sediment load, and valley width and depth relative to stream dimensions. Plateau reaches are characterised by fluctuating discharges, high sand and gravel sediment loads, and relatively narrow, deep valleys. Peel Plain reaches generally are subject to less fluctuation in discharge during the spring and summer seasons, are dominated by fine sand and silt loads, and flow in wider, shallower valleys.

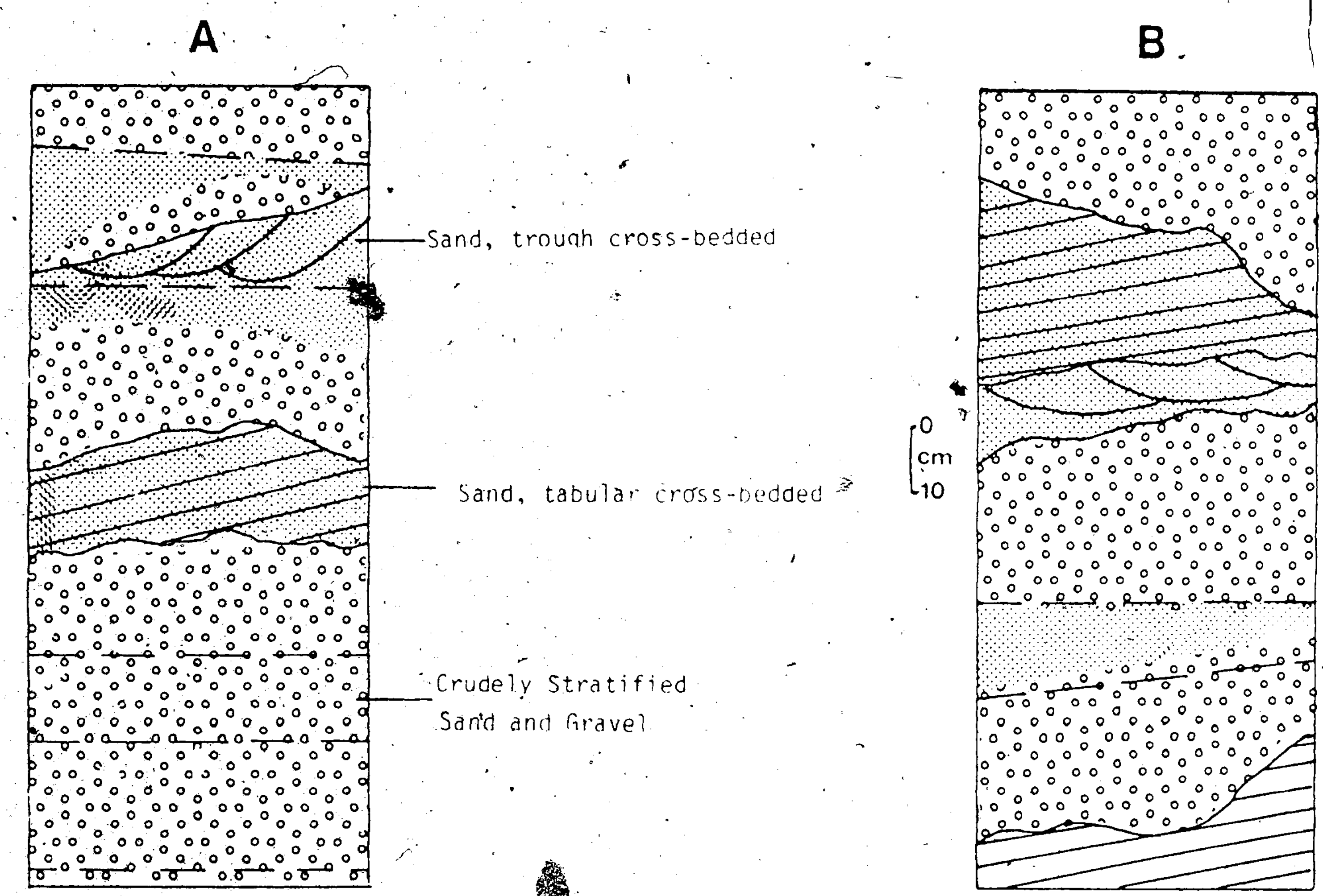
The main channel contains the coarsest sediment in the river system. Sections through abandoned channels (Figure A17.3) are dominated by crudely stratified sand and gravel beds, with secondary erosionally-bound planar tabular cross-beds (Theta type of Allen, 1963a). The flow directions inferred from the trough cross-beds, when the effects of exposure orientation are considered (Decelles <u>et al.</u>, 1983), as well as those determined from the planar cross-beds, coincide with the orientation of the channels. Sections from different positions with respect to thalweg curvature show no marked variations in grain size, in contrast to the findings of Jackson (1978). Continuous reworking during high flow stages, and abundant input from older gravel strata, prevent the development of textural zonation in the bends. Zonation may also be prevented in tight bends by the maintenance of a high-velocity zone along the inside bank of the entire meander curve, coupled with helicoidal secondary flow. Such conditions are common where high volume flow occurs around tight meander bends (Jackson, 1975; Bridge and

Figure A4-3: Main Channel Sedimentary Sequences, Meandering River Reaches

A -- Section from abandoned main channel segment modern Rat River, 67°42' N, 135°51' W

B -- Section from HHC 81-4

See text for discussion


A
B
Sand, trough cross-bedded
Sand, tabular cross-bedded
Crudely Stratified
Sand and Gravel
0
cm
10

Jarvis, 1982).

Channel segments along the margins of the point bars are asymmetric in plan view, with the apparent radius of curvature decreasing progressively in the downstream direction. The thalweg length from the upstream inflection point of the meander to the intersection with the cutbank is 65-90% of the length from the intersection point to the downstream inflection point. The channels intersect the cutbanks at angles between 20° and 90°.

Secondary channels are not common. Channels partially abandoned by avulsion are more frequent than secondary channels created by avulsion, as the latter appear to rapidly capture the flow of the stream and consequently become primary channels.

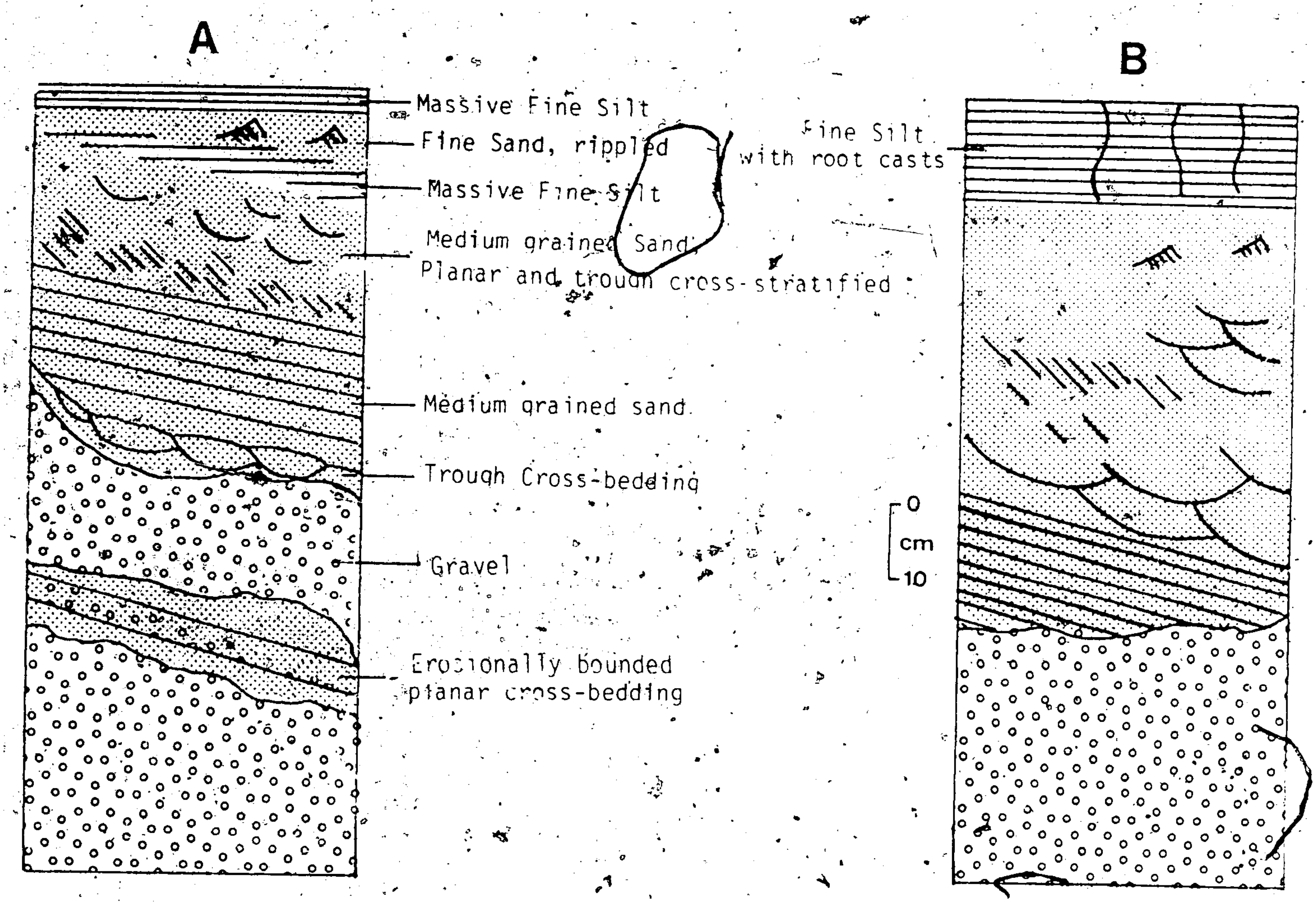
Abandoned channels (Figure A4-4) show the effects of gradually diminishing flow in the sedimentary record. The crudely horizontally stratified and planar cross-stratified coarse sediment is overlain gradationally by planar cross-beds with erosional contacts along the bedding surfaces (type Omikron of Allen, 1963a) and by scoop-shaped trough cross-beds with erosional bounding surfaces (type Pi of Allen, 1963a). This sediment grades vertically into finer sediment (medium sand to coarse silt) which may contain small-scale versions of the underlying cross-stratification (types Mu and Nu of Allen, 1963a). Ripple-drift cross-laminations with lee sides preserved exclusively (type A of Jopling and Walker, 1968; type Kappa of Allen, 1963a)

Figure A4-4: Avulsion Channel Sedimentary Sequence, Meandering River Reaches

A -- Section from Abandoned Avulsion Channel, modern Rat River, 67°42' N, 135°51' W

B -- Section from HH 62-148, 81-6, 66°40' N, 134°32' W

See text for discussion


A
B
Massive Fine Silt
Fine Sand, rippled
Massive Fine Silt
Medium grained Sand,
Planar and trough cross-stratified
Medium grained sand
Trough Cross-bedding
Gravel
Erosionally bounded
planar cross-bedding
Fine Silt
with root casts
0
cm
10

are commonly preserved in the upper portion of this unit. Structureless medium to fine silt interbeds are also contained within the upper portion.

The cross-stratified units are capped by massive medium silt and finer sediments, which often contain abundant root casts (cf. Wing, 1984) and other bioturbation structures. The concentration of organic matter is low, although laminations composed of organic-rich detritus are present in the lower portion of the fine sediment units. The organic content in modern abandoned channels is inversely related to the development of arboreal vegetation. Trees act to stabilize the fine sediment, block or hamper sheet flow across the abandoned channel during flood periods, and produce less fine organic detritus than do understory plants or grasses. Fine sediment infills colonized by arboreal plants thus tend to be poor in organic content. Holland and Buck (1982) reported the development of comparable strata under temperate climatic conditions. Small ice lenses are present in sediments distal from the main channel.

The cross stratified silts and massive fine sediments were produced by the infilling of a sinuous channel. Sediments developed under different climatic regimes investigated by Allen (1970) and Van der Meulen (1982) are similar in several respects, and have also been interpreted as sinuous channel infill deposits.

In deeply scoured areas of the abandoned channels, particularly in the crossover zone between meander bends, depressions are occupied by small lakes (Figure A4-2). These lakes may be ephemeral or perennial. Most of the ephemeral lakes develop in areas no occupied by arboreal vegetation. The sediments are texturally fine silt and clay, lack bioturbation structures, and frequently contain dessication-crack horizons infilled with loess (Figure A4-5a). Distinct organic detrital layers are absent.

Perennial lakes generally occur in association with arboreal vegetation. Sediments deposited in these lakes are dominantly fine silt and clay, lack bioturbation structures, and do not contain dessication features or loess (Figure A4-5b). Thin layers of organic detritus were observed in some exposures.

Both ephemeral and perennial lakes are generally connected to the main channel by small floodplain tributaries. Sediments in these tributaries are primarily fine sand and silt, usually massive but infrequently sinusoidally-rippled, with interbedded lenses of fine organic detritus. The tributaries tend to follow the course of larger abandoned channels, even in cases where the large channels are infilled.

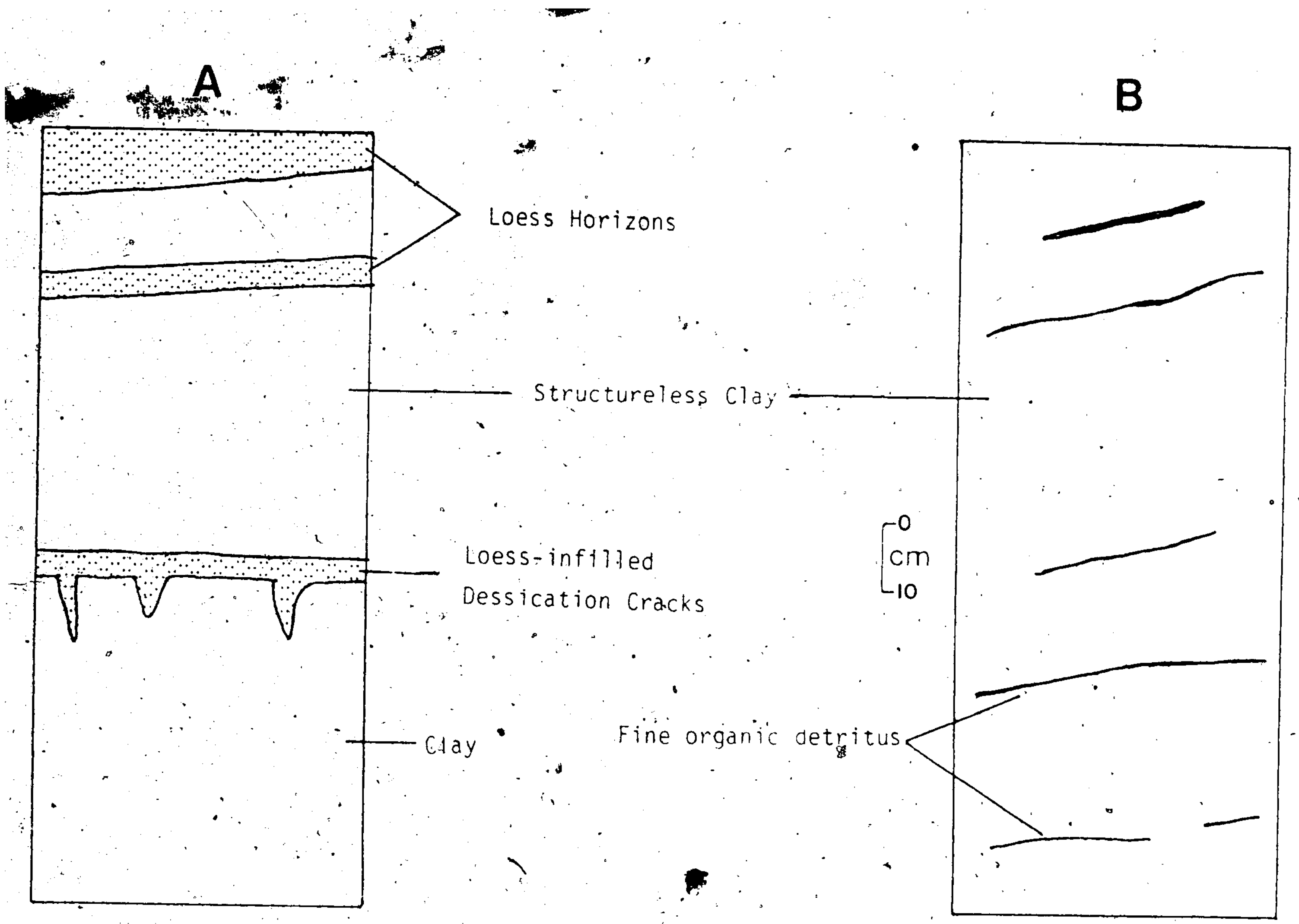
The dominant geomorphic features of the meandering stream reaches are point bars. Development of a point bar depends on the velocity and volume of stream flow, and on the radius of curvature of the meander. In the study region, distinctive forms and strata have developed along bends with high and low radii of curvature.

Figure A4-5: Remnant Lake Sedimentary Sequences, Meandering Reaches

A -- Ephemeral Lake Sequence, modern Rat River, 67°43' N, 135°55' W

B -- Abandoned Perennial Lake Sequence, Rat River, 67°42' N, 135°52' W

See text for further discussion

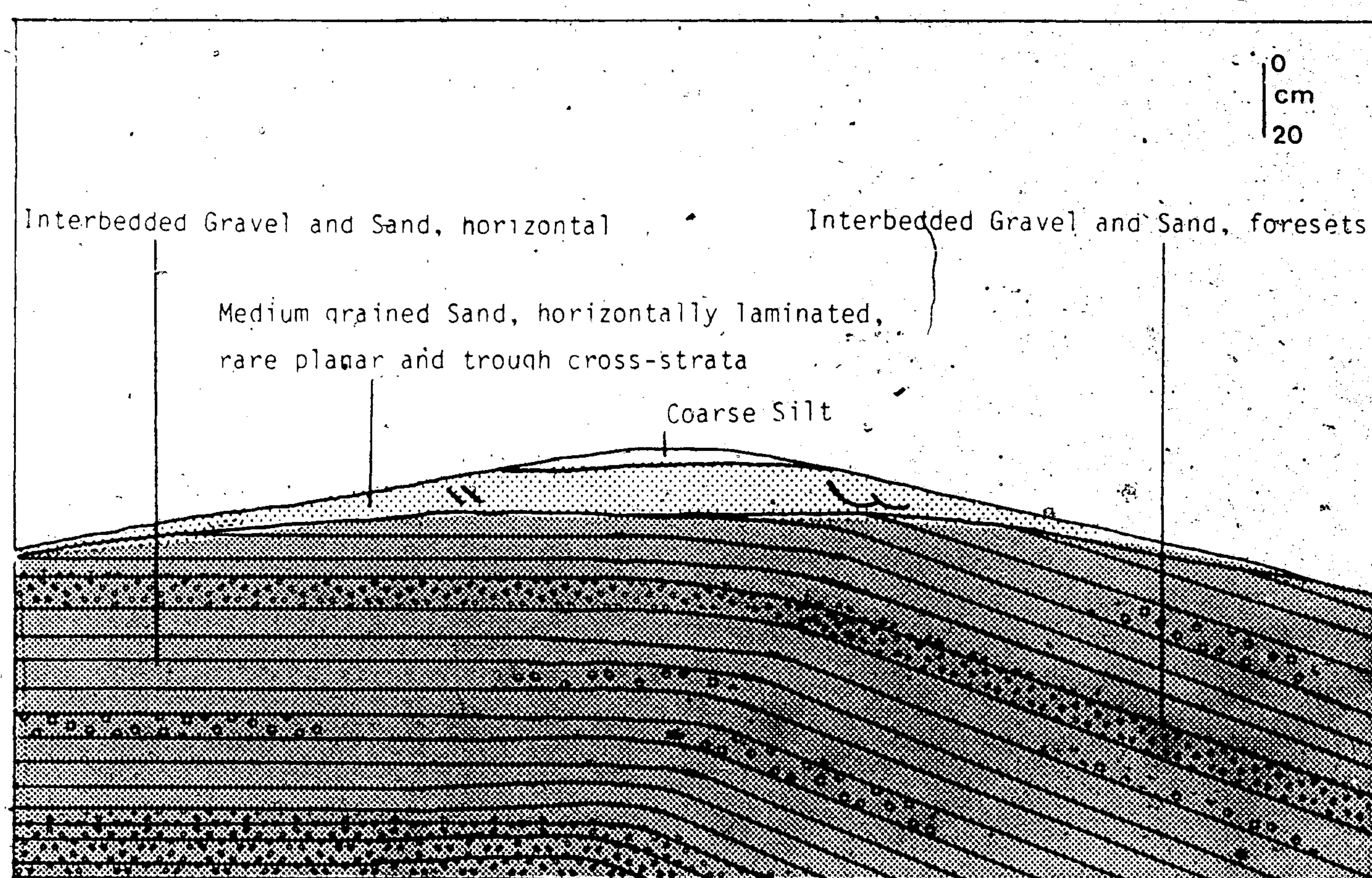

A
B
Loess Horizons
Structureless Clay
Loess-infilled
Dessication Cracks
Clay
0
cm
10
Fine organic detritus

The bars developed along tight meander bends have steep (8° to 15° approx.) downstream slopes and shallower (5° to 10° approx.) upstream slopes (Figure A4-6). The slope inflection line is approximately central on the bar. The upper strata of the bars remain moist throughout the year, but vegetation is confined to small patches of Equisetum (horsetail) and grasses, with rare Arnica. The basal zone of a typical bar is composed of interbedded gravel and sand strata: horizontal or gently upstream-dipping planar foresets of coarse to medium sand with interbedded pebble lag layers in the upstream portion, grading laterally downstream to tabular foreset cross-beds (type Gamma of Allen, 1963a) composed of sand, granules, and pebbles (as illustrated in Figure A4-6). The upstream sediments prograde in the downstream direction. The dip directions of the foresets are aligned at 35° to 75° to the thalwegs of the main channels, but are approximately parallel to the net flow direction of the steams. Pebble fabrics in both the upstream and downstream sections are dominantly transverse to the bar slopes.

The bars are capped by medium sand units, either horizontally bedded (2-5 cm thick), planar cross-laminated (type Lambda of Allen, 1963a), or trough cross-laminated (type Nu of Allen, 1963a). Horizontally bedded sediments are most frequently observed. Thin fine sand and coarse silt strata are rarely present (Figure A4-6).

Figure A4-6: Schematic Diagram, Tight Meander Bend Point Bar, Meandering Reaches

See text for discussion

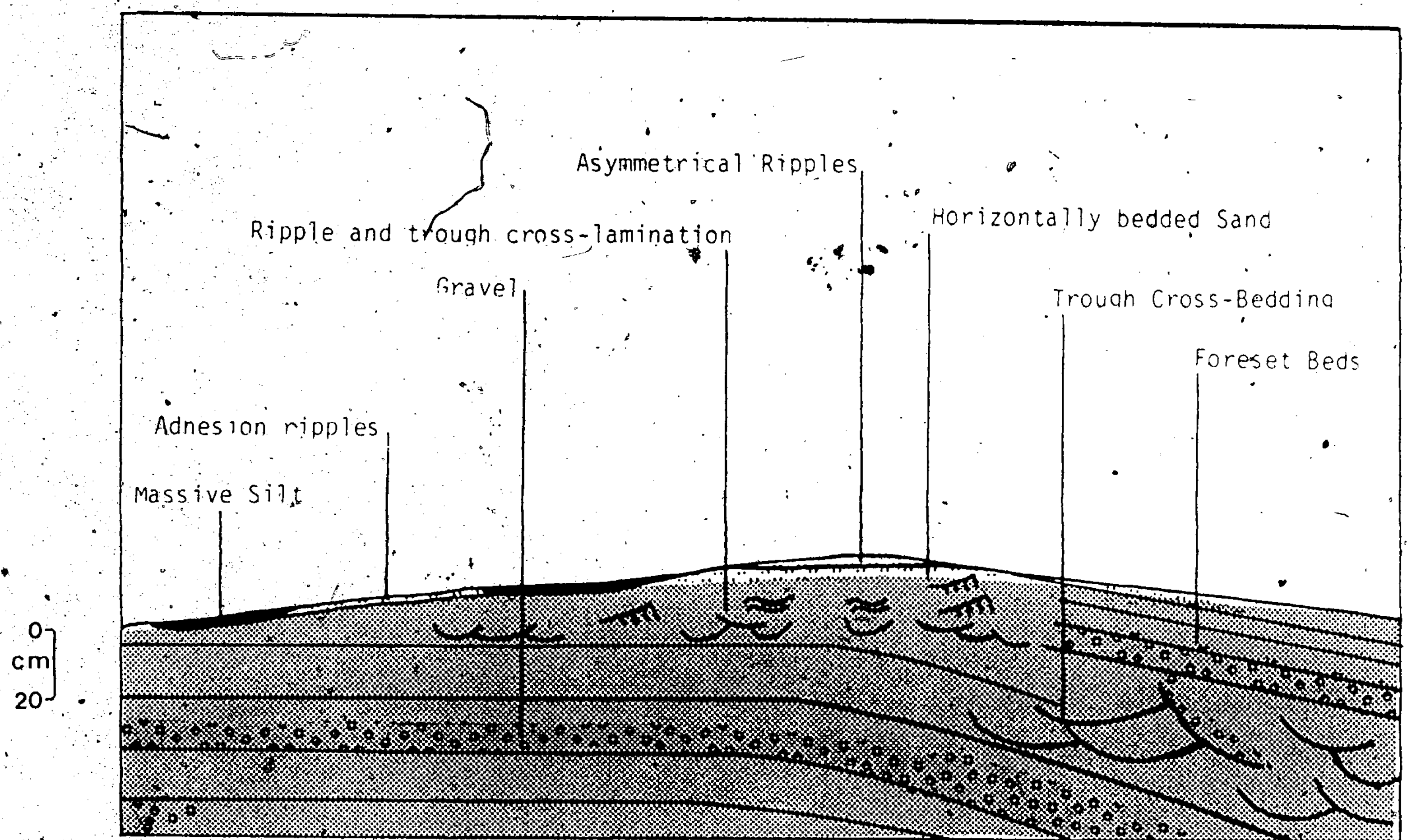

0
cm
20
Interbedded Gravel and Sand, horizontal
Interbedded Gravel and Sand, foresets
Medium grained Sand, horizontally laminated,
rare planar and trough cross-strata
Coarse Silt

The bars developed along gentle meander bends have shallow upstream and downstream slopes (3° - 8° approx.), the downstream slopes usually being steeper (Figure A4-7). The upstream slopes are 1.2 X to 1.5 X longer than the downstream slopes. The surface strata adjacent to the river remain moist throughout the year, and are sparsely vegetated with grasses. The upper portions of the bars are dry during the summer months, and are either sparsely covered with Equisetum arvense or are unvegetated.

The basal upstream zone of a typical gentle meander bend bar is composed of interbedded gravel and sand strata, either as horizontal beds 5-20 cm thick or gently upstream-dipping foresets of medium sand with interbedded coarse sand and granule deposits, similar to those strata developed in tight meander point bars. The sediments grade laterally downstream into tabular foreset cross-beds and scoop-shaped trough cross-beds (types Gamma and Pi of Allen, 1963a), composed of sand, granules, and pebbles. Pebble monolayers are commonly interbedded in coarse sand and gravel layers. The dip directions of the foresets are aligned at 45° to 90° to the thalwegs of the main channels, and differ from the net flow direction of the streams by as much as 50°. Pebble fabrics in both the upstream and downstream sections are dominantly transverse to the bar slopes, although substantial oblique flow components were noted in the downstream sediments.

Figure A4-7: Schematic Diagram, Gentle Bend Meander Point Bar, Meandering Reaches

See text for discussion



Asymmetrical Ripples
Horizontally bedded Sand
Ripple and trough cross-lamination
Gravel
Trough Cross-Bedding
Foreset Beds
Adhesion ripples
Massive Silt
0
cm
20

The bars are capped by medium to fine sand units 10-50 cm thick. The sands are commonly trough cross-bedded (type Nu of Allen, 1963a), along the bases, grading upwards to ripple-drift cross-laminations (types A and B, Jopling and Walker, 1968). Horizontally bedded and laminated sands are present in the upper portions of the capping units. Adhesion ripples (Kocurek, 1981) are developed on some moist areas adjacent to the active channels. Massive fine silt and sand are patchily distributed over the moist areas. Thin laminae of fine organic detritus are also present, and concentration of organic detritus are developed in the lee of adhesion ripples.

The bar crown area (Figure A4-7) lacks adhesion ripples, fine sand, silt, and fine organic detritus. Aeolian asymmetrical ripples, indicative of eastward-moving winds, are developed in fine sand on the bar crown areas. The ripple indices and ripple symmetry indices for these features lie within the ranges proposed for aeolian ripples by Tanner (1967). Microchannels excavated by rill flow are oriented parallel to the slope of the bar crown.

The point bars developed along gentle and tight meander bends thus differ in several aspects. In the stratigraphic record, tight bend pint bars could be recognized by their moderate-angle downstream-dipping planar foreset beds, their strongly oriented pebble fabrics, and the absence of thick fine sand and coarse silt strata. Gentle bend point bar deposits would contain foresets inclined at lower angles downstream, trough cross-beds in the basal portion of the downstream zone, moderately oriented pebble fabrics, and thicker coarse silt and fine sand units. Marginal and bar crown areas might be distinguished by ripple form and organic detritus content. These distinctions cannot be applied with certainty in all cases, however. The bars flanking modern streams in the Peel Plateau and Peel Plain vary considerably in morphology and radius of curvature, and similar variation would be expected for bars developed by ancient streams.

The differences in orientation among the bar foresets, clast fabrics, thalwegs, and net stream flow directions are produced as consequences of peak flow conditions. At peak flow, as much as one-third of the volume of a meandering stream may be transported across the point bar necks (Jackson, 1975). Tight meander bend bars, because they are higher and steeper than gentle bend bars, are covered only during the peak flow stages. Flow divergence across the bar surfaces is limited by the narrowness of the necks, while divergence around the bar is accentuated by the steepness and height. Consequently, foreset development and pebble fabrics reflect the general downslope direction of flow, rather than the course of the thalweg, and the foresets are therefore oriented downvalley (parallel to the net stream flow). The pebble fabrics are oriented transverse to the flow, as would be expected in a meandering stream of moderate velocity (cf. Johanssen, 1963). The absence of influence from low flow stages is shown by the scarcity of fine sediments.

In contrast, point bars developed along gentle bends are covered during diminishing flow periods, because these bars are lower and have shallower slopes. Helicoidal secondary flow (Hooke, 1975) and/or transverse gradients in the rate of sediment transport (Edwards et al., 1983) act to initiate divergence of flow across the bar surfaces. As stream flow slackens, divergence of flow across the bar surfaces increases, and some clasts may be re-oriented. Consequently, the component of fabric generated by movement of clasts towards the thalweg oblique to the slope of the bar would be greater for the gentle bend bars than for the tight meander bars, resulting in the preservation of fabrics with substantial oblique components. Divergence of flow during lowering stages would also be expected to produce foreset beds oriented at variance to both the thalweg and the net stream flow direction, as is observed in gentle bend point bars. Flow divergence may also contribute to the preservation of finer sediment units on the surfaces of gentle bend point bars (Milne, 1982).

Similar deposits along modern rivers and in ancient successions have been interpreted as point bars developed in meandering streams dominated by sand and gravel by many researchers (eg. Allen, 1970; McGowen and Garner, 1970; Gustavson, 1978; Jackson, 1978; Levey, 1978; Blacknell, 1981; Van der Meulen, 1982; Johnson, 1984). Distinctions between tight and gentle bend point bars are not commonly suggested, possibly because distinguishing the two types can be difficult and seldom is critical in interpretation. If the flow directions of ancient streams are to be inferred from fabric and foreset bed orientations, some knowledge of the degree of divergence between these factors must be obtained and applied. As the orientations of fabrics and foresets developed on gentle bend point bars may differ considerably from the net stream flow direction, recognition of this sub-environment is important. Recognition of bar head sequences of upstream-dipping foresets is also important in palaeoflow reconstructions. Seen in isolation, such foresets may be easily mistaken for beds developed on the lee side of point bars (Taylor, et al., 1971).

Flank jam deposits of large logs and smaller fragments of organic drift are developed along the outer bank of meander bends, typically downstream from the inflection points (Figure A4-2). Logs in the deposits are oriented at low (azimuth) angles oblique to the thalweg, and are flat-lying. The jam deposits are usually not extensive.

Nanson (1981) proposed that dead trees (among other factors) acted as nuclei for scroll bar development along the Beatton River of northeastern British Columbia. The absence of dead trees and large logs in the bar crown areas, and the absence of jam deposits along the upstream margins of forested point bars, indicate that in the Peel Plateau and Peel Plain streams most of the large organic detritus is transported along the thalweg and that little is stranded on the higher portions of the river valley system. This distribution of organic matter could result from two causes. The logs may remain in transport throughout the higher stages of the flood season, and hence do not become stranded on the point bars. The small diameters of tree trunks on the Peel Plateau, compared to the larger trees of northeastern British Columbia, suggest that transport may occur throughout the spring and summer. Alternatively, the logs may be preferentially introduced into the stream during the lower late spring and summer flow stages. This could occur as a result of summer caving of forested banks, or it may be due to beaver activity. Preferential introduction and extended transport are probably jointly responsible for the absence of stranded logs on the bar crowns.

The development of jams on the downstream edges of point bars acts to anchor the bars and prevents growth parallel to the stream thalweg. The bar responds by expanding laterally into the channel, and the jam becomes isolated on the downstream side. This pattern is apparent in the foreset and ripple drift orientations of sediments surrounding such jams. Flank jams thus act to decrease the radii of curvature of meander bends, and increase the sinuosity of the streams.

Isolated remnants of point bars truncated by avulsion channels are present as longitudinal bars (feature #11, Figure A4-2). These bars are usually modified into a teardrop shape, with the tapered end downstream. This configuration provides the minimum drag and the least resistance to flow (Komar 1983). Deposition on the bar surface is confined to veneers of medium sand and finer sediments.

Erosion occurs along both the inner and outer banks of meander bends, although the outer cut banks are more susceptible. Inner bank erosion generally occurs along avulsion channels, especially those cut through forested or moss-covered areas. Along forested outer banks, caving occurs throughout the spring and summer seasons. Non-forested outer banks generally are subject to caving during intermediate flow stages only, although erosion continues throughout the summer.

Non-forested cut banks are generally covered with grass and low shrubs. Those plants possess extensive, intricate root systems which serve to anchor the sediment. They also tend to grow in better-drained areas, with little or no interstitial ice (Figure A4-2). In contrast, forested cut banks on the Peel Plateau and Peel Plain usually support blankets of _Sphagnum_, and tend to be less well-drained. Interstitial ice is present in finer-grained sediments. Removal of a portion of one of these cut banks results in the exposure of sediment susceptible to thermal erosion, overlain by a cap of matted vegetation. Under these conditions, bank caving is prompted by rotational slumping along the base (Figure A4-8), as occurs where cohesive sediment is underlain by readily erodible material (Turnbull et al., 1966; Laury, 1971). Similar effects have been observed in a non-permafrost environment where interstitial ice is not involved (Murgatroyd and Ternan, 1983). The instability of forested cut banks in the study region is a consequence of the climate and vegetation type. In other environments, forested cut banks and slopes are more stable than their unforested counterparts (eg. Riestenberg and Sovonick-Dunford, 1983). Once the lower surface of a forested cut bank is exposed, erosion can continue to occur even after the river level drops if the sediment is unstable or ice-rich. As a result, caving may persist in these areas throughout the summer.

Debris flows and slides generated by retrogressive retreat of slopes upon thawing are important in modifying the sedimentation pattern of meandering streams flanked by older lacustrine deposits, silty till, or friable Cretaceous shale. Small amounts of silt and coarse clay are steadily injected into the streams by retrogressive thaw-flow chutes developed in steep cut banks (Figure A4-9). Accumulations of sediment collect at the heads of the chutes, and when the weight at the head exceeds the bearing capacity of the basal sediment, it flows or slides. Input into the stream therefore occurs as a series of sediment pulses, rather than as a continuous flow. Pebbles in the thaw-flow deposits are oriented downslope with low plunges (2° - 5°). The sides and bases of the chutes are grooved to depths of 0.5 - 3 cm, and

Figure A4-8: Rotational Slumping along River Bank, Rat River

See text for discussion

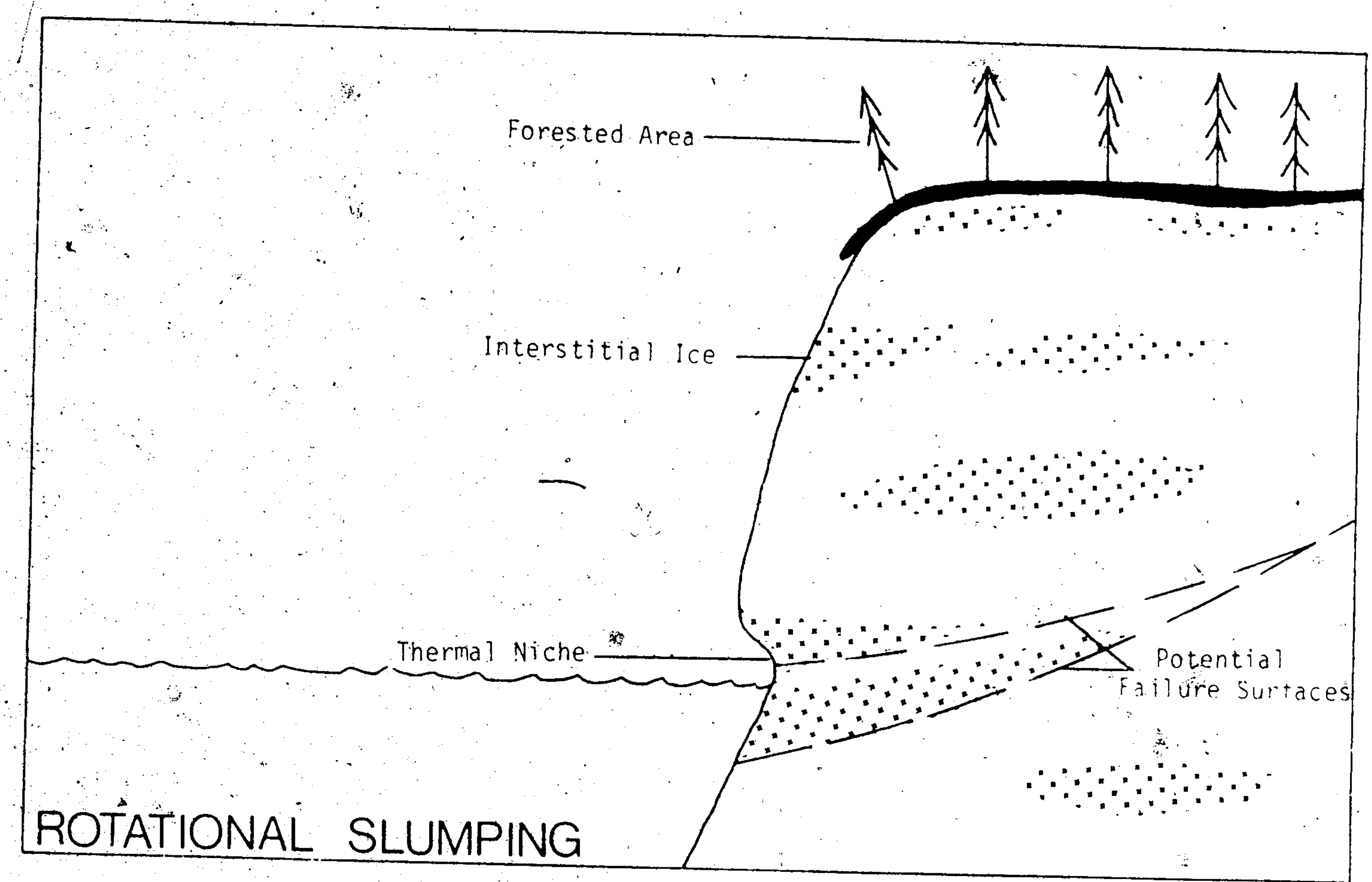
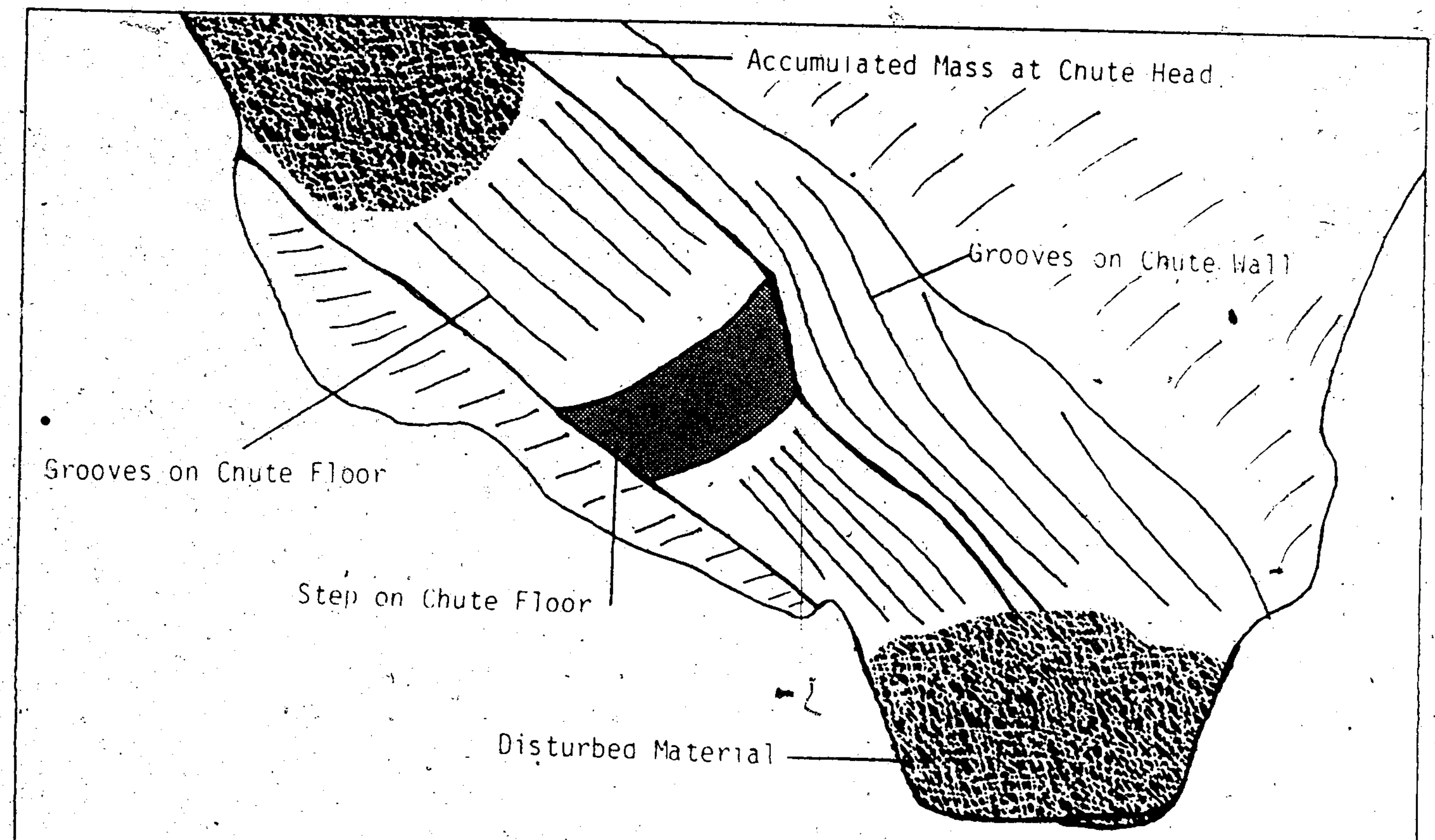

Forested Area
Interstitial Ice
Thermal Niche
Potential
Failure Surfaces
ROTATIONAL SLUMPING

Figure A4-9: Schematic Diagram, Retrogressive Thaw-Flow Chute

See text for discussion



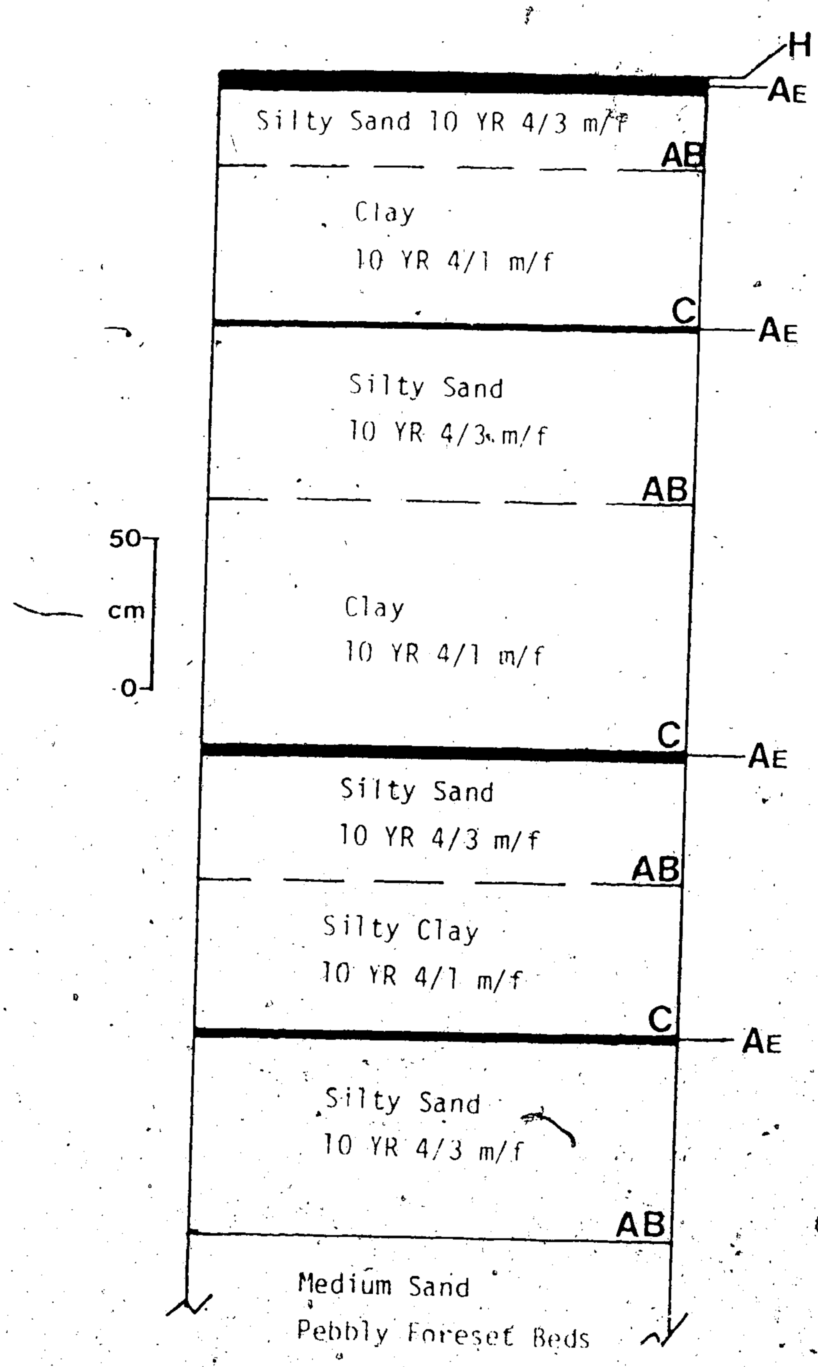
RETROGRESSIVE THAW-FLOW CHUTE

the side grooves parallel the configuration of the bases of the chutes.

Development of the thaw-flow chutes enhances the erosion of the cut banks. The input of silt to the stream system throughout the warm months provides a supply of fine clastic material. Although the sediment is frequently delivered to the stream as coherent blocks, no large aggregations of silt were observed in channel sediments. The large blocks may be completely disaggregated in the high velocity zones of the thalweg adjacent to the cut banks.

Larger retrogressive thaw debris flows and slides also occur where unstable silts flank river valleys. The deposits formed by the mass movements consist of structureless silt or silty clay, grading upwards to sandy silt or silty sand. At some locations, multiple events separated by thin Brunisolic Ae or Regosolic A soil horizons can be recognized in distal areas (Figure A4-10). The disturbed sediments form arcuate cones and fans, and can cause diversion of the river channel. Point bars developed downstream from the disturbance often contain interbedded silt layers within the foreset beds, forming the alternating sand/silt epsilon cross-strata associated with meandering channels excavated in fine sediment (eg. Allen, 1963a; Jackson, 1978; Mossop and Flach, 1983). The lack of such cross-strata in most point bar deposits in the study region suggests that the debris flows and slides are areally restricted.

FIGURE A4-10
SOIL HORIZONS, HH 62-116 81-1



The floodplains are occupied by a variety of floral communities. Viereck (1970) noted a successional sequence of vegetation communities along the terraces of the Chena River in Alaska, where the pioneer arboreal Salix (willow) and Alnus (alder) community is gradually replaced by a mixed community dominated by Picea glauca (white spruce), with Populus (poplar) and Betula papyrifera (white birch). Simular communities on the Peel Plateau have been described by Rowe (1972) and Hettinger et al. (1973). The vegetation on the low floodplain traps fine sediment and prevents excessive erosion during spring floods, leading to the formation of massive silt layers similar to those formed in abandoned channels. Sedimentation on the high floodplain is confined to thin layers of silt, representing overbank deposits of exceptional flood events. The development of arboreal vegetation is controlled by aspect, frequency of gelifluction and retrogressive thaw debris flows and slides, and the nutrient content of the parent material.

Tributary streams rework the fine grained deposits on the floodplain and bar surfaces. Small tributaries often produce a thin lens-shaped lag deposit of granules and pebbles, oriented transverse to flow in the tributary stream. The orientations of these lenses bear no relation to the flow direction in the principal stream, and isolated lenses of pebbles and granules with orientations differing from those induced by flow in the main channel are found interbedded in point bar and floodplain sediments.

Backflow from the main channel into the mouths of tributaries may occur during the waning stages of the spring flood season. This results in the production of two opposed sets of cross-laminations, the first representing the bacckflow direction and the second the subsequent return flow to the main channel, coupled with flow in the tributary. The pattern of cross-lamination resembles the herringbone structures reported from tidal environments (Coleman and Wright, 1975; Clifton, 1983).

The meandering stream environments present on the Peel Plateau and Peel Plain are characterized by sedimentary successions representing abandoned channels, tight and gentle bend point bars, and bar flank jams.

Examples of rivers which can be classified as graveliferous sand bed streams (Jackson's type #4, 1978) or as sand bed streams lacking gravel (Jackson's type #3, 1978) are prevalent on the Peel Plateau and Peel Plain.

C. Plateau Braided-Meandering Reaches

Most of the reaches of the principal streams of the Peel Plateau exhibit characteristics associated with both braided and meandering rivers. These reaches can be considered to be intermediate between braided stream environments, such as those described by Miall (1977, 1978) and Rust (1978), and meandering stream environments, such as those documented by Allen (1965), Hickin and Nanson (1975), Gustavson (1978), and Jackson (1978). Intermediate stream types developed under different climatic, physiographic, and geologic conditions have been recognised by several researchers (eg. Blench, 1969; Shelton and Noble, 1974; Bluck, 1976; Schwartz, 1978; Allen, 1983b).

Channel geometry parameters for several braided-meandering reaches are presented in Table 4-3. Gradients range between 1.1 and 21 m/km. The modal value for sinuosity (Leopold and Wolman, 1957) is 1.35. Typical values for the braiding index (Brice, 1964) range between 1.2 and 1.9, while braiding parameter values (Rust, 1978) are 1.8 to 4.2.

Figure A4-11 depicts a portion of a modern braided-meandering reach. The diagram is schematic and composite. The scale, distribution, and number of features are variable, depending upon the consistency and magnitude of discharge, sediment load, and pre-existing valley width.

The reaches are characterised by a single main channel of moderate sinuosity. Two co-dominant channels are present in some reaches, but this situation is uncommon. The width of the main channel is constant.

Sections through abandoned segments of large channels (Figure A4-12) reveal sediment assemblages dominated by crudely stratified poorly to moderately sorted, well-rounded to sub-rounded gravel and sand. Trough cross-beds (Theta type of Allen, 1963a) and rare erosionally-bound planar tabular cross-beds (Beta type of Allen, 1963a) are also present. The flow directions inferred from the cross-beds

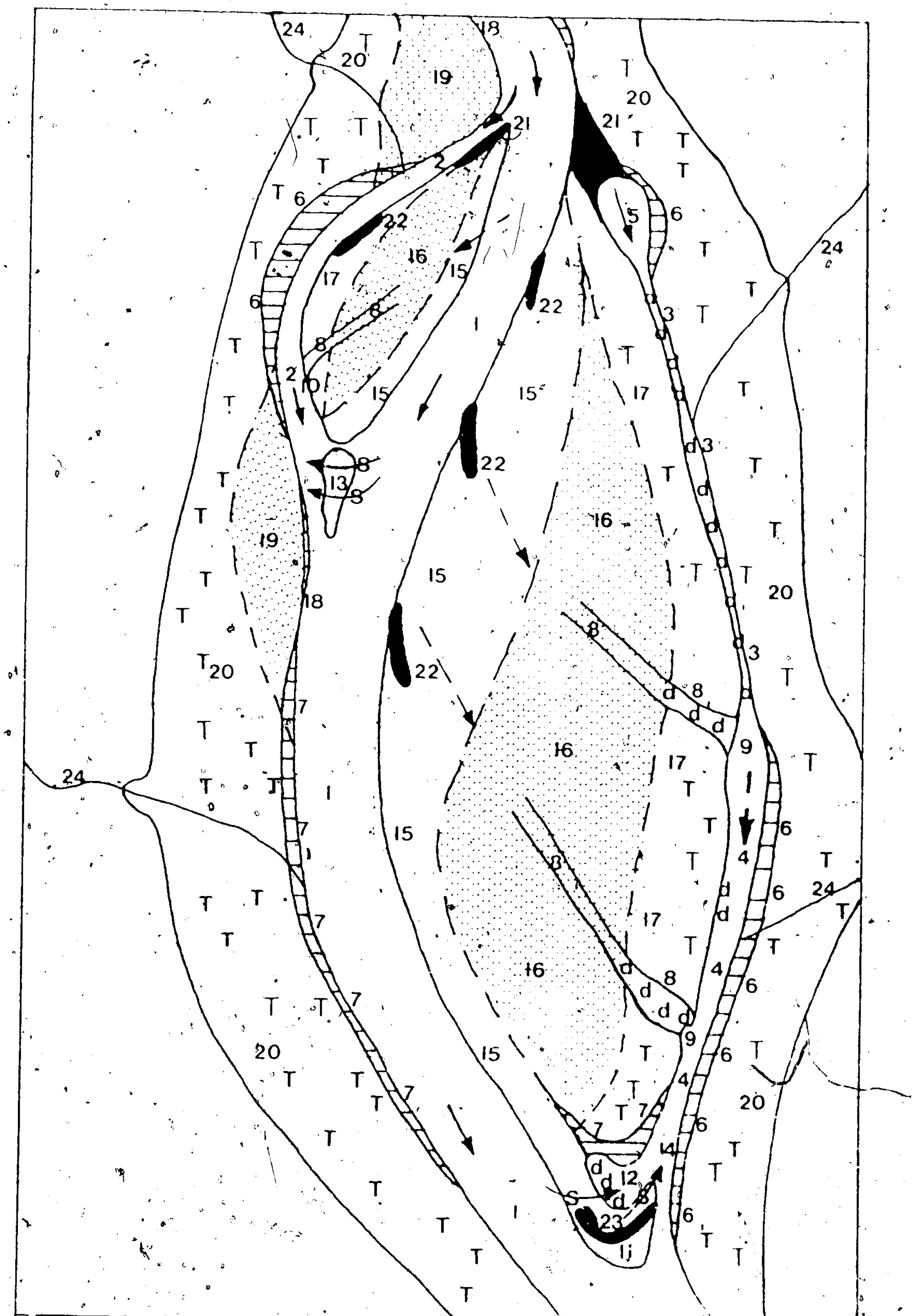
FIGURE A4 II

Braided-Meandering Reach, Schematic Diagram

1 -- Main Channel
2 -- Subsidiary Channel
3 -- Seasonal back-bar Channel, upstream zone
4 -- Seasonal back-bar Channel, downstream zone
5 -- Scour Pool
6 -- Spring Bank Caving Zone
7 -- Spring and Summer Bank Caving Zone
8 -- Bar-Chute Channel
9 -- Bar-Chute/Back-Bar Confluence Zone
10 - Bar-Chute/Subsiduary Channel Confluence Zone
11 - Backflow Crescentic Bar
12 - Backflow Slough
13 - Subsiduary Channel-mouth Bar
14 - Backflow Zone
15 - Bar Foreslope
16 - Bar Crown
17 - Bar Backslope
18 - Channel margin
19 - Exposed marginal Sand Flat
20 - Vegetated Floodbelt
21 - Bar Head Drift Jam Deposit
22 - Bar Flank Drift Jam Deposit
23 - Bar Tail Drift Jam Deposit
24 - Tributaries
T - Arboreal vegetation
d - area veneered with organic detrital drift

Surface dry during summer months

Bank Caving Zones


24
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23
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Figure A4-12: Sedimentary Sequences, Large Channels, Braided/Meandering Reaches

A -- Rat River, 67°40' N, 135°29' W

B -- HH 76-3 82-c, 65°48' N, 133°18.5' W

See text for discussion

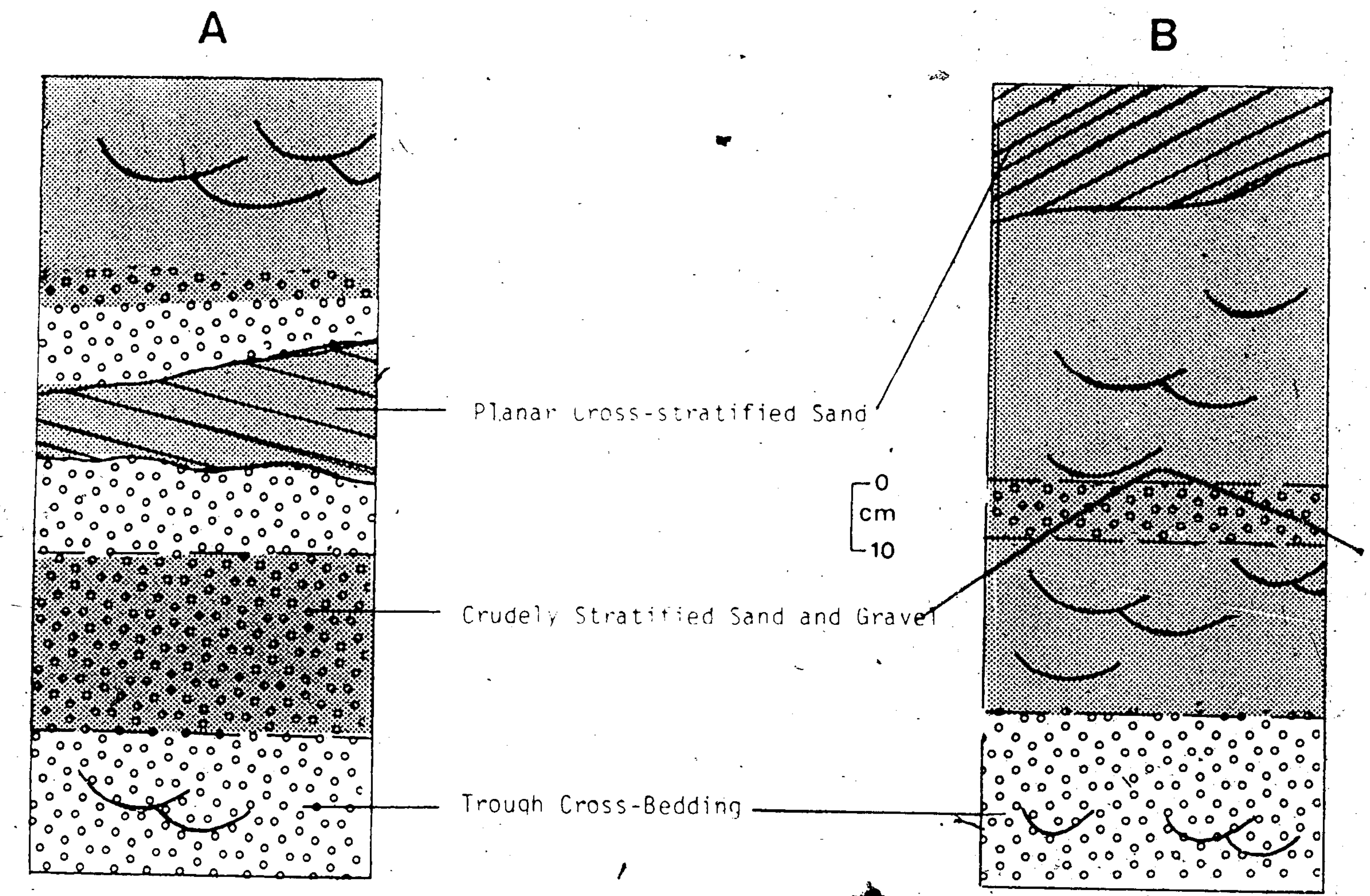

A
B
Planar Cross-stratified Sand
Crudely Stratified Sand and Gravel
Trough Cross-Bedding
0
cm
10

Table 4-3: Channel Properties of Braided Meandering Stream Reaches

Reach	Gradient (m/m)	Sinuosity *	Braiding Index **	Braiding Parameter ***	Date of Photo
Stony Creek					
67°20'-22' N	0.021	1.28	1.4	2.2	July 1970
135°00'-06' W					
Vittrekwa R.					
67°07'-11' N	0.006	1.35	1.2	2.9	July 1970
135°15'-25' W	0.006	1.41	1.0	2.4	June 1971
Rat River					
67°38'-41' N	0.008	1.31	1.3	2.4	Apr. 1954
135°27'-30' W	0.008	1.35	1.4	2.2	June 1971
Barrier R.					
67°36'-40' N	0.010	1.37	1.2	1.9	June 1971
135°40'-45' W					
Peel River					
66°15'-30' N	0.001	1.35	1.7	4.2	June 1971
133°51'-134°W					

cont. Table 4-3

Reach	Gradient (m/m)	Sinuosity *	Braiding Index **	Braiding Parameter ***	Date of Photo
Trail River					

67°32'-35' N 134°45'-55' W	0.005	1.26	1.5	2.6	Aug. 1977
Caribou River					
66°13'-14' N 134°40-55' W	0.004	1.54	1.5	2.2	Aug. 1977
Caribou River					
66°18'-19' N 134°19'-28' W	0.004	1.37	1.6	2.6	Aug. 1977
Road River					
66°38'-42' N 135°15'-20' W	0.006	1.38	1.5	3.1	July 1977
Chii Nahil Cr.					
66°46'-49' N 135°43'-48' W	0.013	1.23	1.5	2.4	July 1978

cont. Table 4-3

Reach	Gradient (m/m)	Sinuosity *	Braiding Index **	Braiding Parameter ***	Date of Photo
Aghoo Creek					
66°00'-05' N 134°39'-42' W	0.011	1.41	1.2	1.8	Aug. 1977

* after Leopold and Wolman 1957

** after Bryce 1965

*** after Rust 1978

vary from the inferred thalweg of the channels by 20° or more, where the trends of the channels are visible. The orientation of the clast fabric is commonly oblique to the flow direction, as has been reported in braided stream sediments by other workers (eg., Cant and Walker, 1978; Schwartz, 1978).

Massive, poorly-sorted gravels with a framework of large clasts infilled with inorganic silt characteristically form as laterally accumulating diagonal or longitudinal bars in riffle zones (Smith, 1974; Miall, 1978). Pebbles and cobbles are transported during periods of moderately high flow, either as diffuse gravel sheets (Hein and Walker, 1977), or as imbrication clusters (Martini, 1977). The well-rounded to subrounded nature of the larger clasts indicates that the stream power during flow was insufficient to transport the majority by saltation, a process which results in angular surfaces formed by chipping and fracturing of the clasts against each other (Pittman and Ovenshine, 1968; Moss <u>et al.</u>, 1973). Instead, the clast apparently travelled as rolling or sliding clusters, whose combined coefficient of drag would be lower than that of individual particles (Martini, 1977; Brayshaw <u>et al.</u>, 1983). Any diminution in the flow velocity would initiate lodgment of the largest clasts, which then would form obstacles to the motion of the other particles in the clusters. The rolling, sliding, and rotating pebbles were forced into interstices in the framework formed by the previously deposited cobbles, a process termed "contact imbrication" (Blacknell, 1981, 1982). The resulting fabric reflects the combined influence of the initial rolling process (with b-axes generally parallel to the flow direction), and the subsequent jamming of slightly smaller clasts and those following into the original framework, disrupting it in some instances.

The insertion of coarse clasts into the interstices acts to produce an armoured stream bed, resistant to erosion or alteration at subsequent flow stages (Whittaker and Jaeggi, 1982). If the stream level drops rapidly, the sediment will be preserved as an open-worked, matrix-free gravel deposit. If the stream drops more slowly, finer sediments transported in suspension during the lower flow stages can infiltrate into the open-worked structure, producing a very poorly sorted gravel, with coarse clasts in a fine matrix (Eynon and Walker, 1974; Beschta and Jackson, 1979; Frostick et al., 1984). Slow decline also allows some intermediate clasts to be winnowed from the armoured bed during the initial stages (Carling and Reader, 1982).

Since the bar deposit serves as a bedform obstacle, flow separation occurs which in turn induces deposition of the suspended sediment load. Thus, the initial open-worked deposit is transformed into a poorly sorted silt matrix gravel. Several such events are commonly preserved in channel deposits in the Peel Plateau. No records of the intermediate flow stages are preserved because the armoured bedforms were resistant to extensive erosion and alteration during these periods.

Nelson and Scott (1962) suggested that organic particles could also be trapped in the interstices of gravel deposits. Frostick et al. (1984) reported that organic detritus was absent from matrix infill deposits. The sediments investigated in the Peel Plateau region contained no significant amounts of organic matter. In a channel where flow is maintained throughout the year, the low specific gravity of the organic detritus would serve to keep the particles in suspension.

Abandoned subsiduary channel sequences are dominated by moderately sorted gravel, composed of medium to coarse pebbles with scattered cobbles in a medium sand matrix. Crude horizontal stratification is characteristically dominant, and cross-stratification generally absent. Imbrication and clast-fabric orientation are variable with respect to the inferred thalweg directions. Clasts tend to be aligned obliquely or at right angles to the net flow direction, but polymodal fabrics are common. Despite the inexactness of the determination, however, a general flow direction co-incident with the trend of the thalweg can be assumed. Imbication is usually more closely aligned to flow direction than is the modal clast fabric.

Coarse-grained imbricated lateral or diagonal bar sediments without cross-stratification and with little or no fine-grained material are characteristic of high-sinuosity gravel-bed braided streams (Bluck, 1971, 1976; Jackson, 1978). These sediments tend to accumulate in riffle or mid-reach sections of the channel with the finer sediments accumulating in the intervening pools (Milne, 1982). The absence of cross-stratification and textural differentiation amongst the sand fractions indicates that the stream power was sufficient to transport the sand as a homogeneous population (Smith, 1974), and that subsequent low flow stages were unable to disturb the previously deposited sediment. This pattern is characteristic of streams with substantial diurnal or seasonal fluctuation in flow (Smith, 1974; Miall, 1977; Rust, 1978).

Most secondary channels in the braided-meandering system are seasonal, occupied only during peak flood season. These channels have been referred to as "side channels" (Smith, 1972; Eynon and Walker, 1974) and "slough" and "inner channels" by Bluck (1976). In this study, the channels are termed "back-bar channels", reflecting their position relative to the main thalweg and the large lateral longitudinal and point bars developed along it.

The back-bar channels are 1-3 m deep during the maximum spring flow. The upstream portion of the channel is narrow and shallow, bordered ~~by~~ matted organic debris. At locations where the channel banks are bordered by <u>Alnus</u> scrub or are forested, the channel is straight, contains abundant organic detritus, and is little affected by aeolian reworking during the dry season. Where the margins are unvegetated or vegetated with open scrub, the channel tends to bifurcate and aeolián features are well-developed. Downstream, the channel widens and deepens as it is joined by the bar-chute channels.

Sedimentary sequences in the back-bar channels record gradually declining flow. The basal deposit consists of well to moderately sorted fine cobble and coarse pebble gravel, with little matrix material, similar to that found in the perennial subsiduary channels. This unit grades vertically upwards into medium-fine pebble gravel with coarse-medium sand. Trough cross-stratification (Theta of Allen, 1963a) is frequently present, and planar tabular cross-stratification (Gamma and Beta of Allen, 1963a) rarely occurs. The gradational upper contact of the coarse grained gravel is produced by infiltration of medium sand into the open framework of the pebbles and cobbles (Frostick et al., 1984). Irregular mud intraclasts are frequently present. Similar sequences have been interpreted as minor channel deposits by many researchers (eg. Smith, 1972; Shelton and Noble, 1974; Ramos and Friend, 1982; Kirk, 1983; Johnson, 1984).

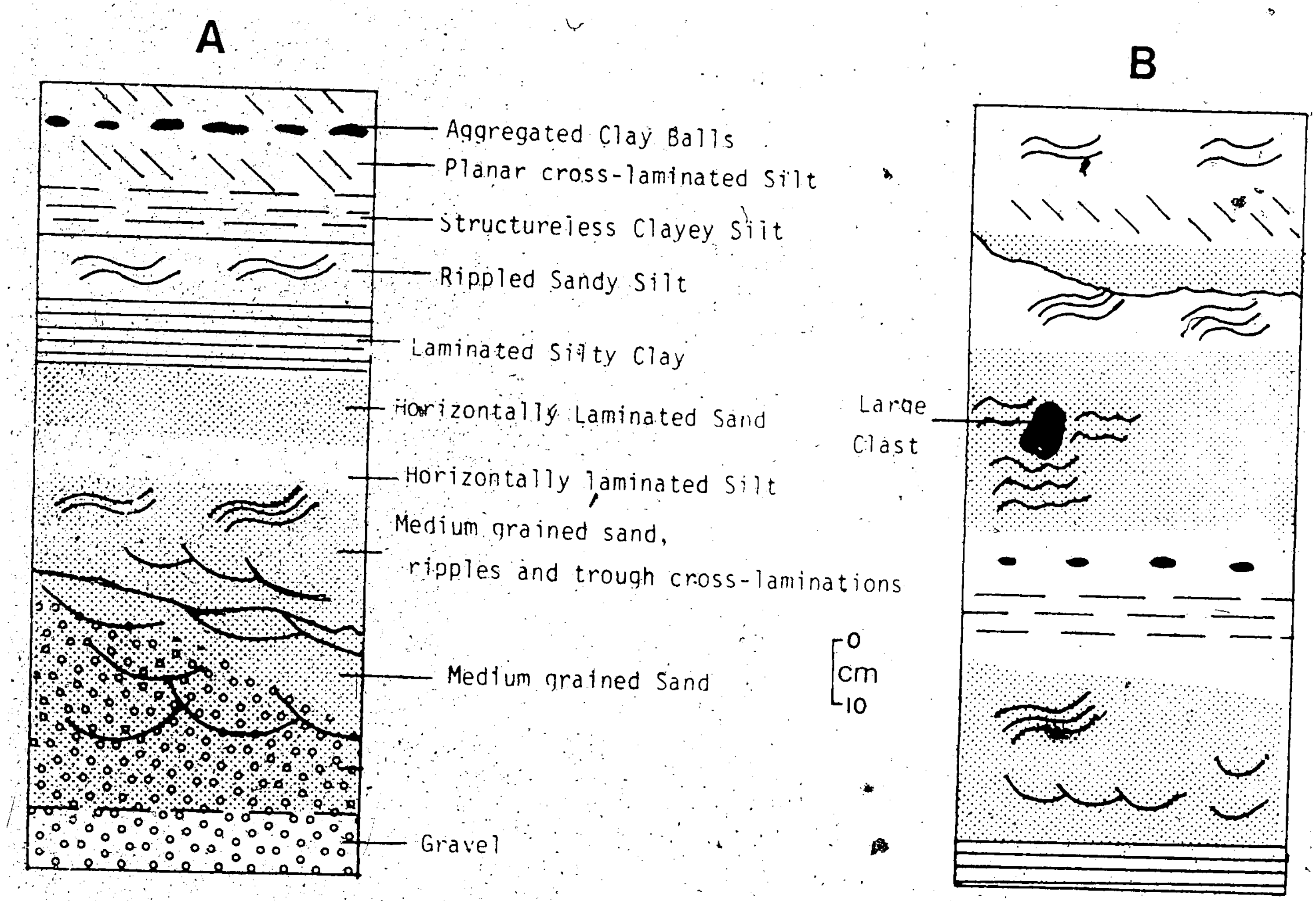
The trough cross-stratified coarse-medium sands are gradationally or abruptly overlain by medium-fine sands (Figure A4-13). Horizontal laminations, trough cross-laminations, ripple and climbing-ripple laminations, and contorted laminations are all present in various deposits. Commonly, the medium-fine sand units are contained

Figure A4-13: Sedimentary Sequences, Back-Bar Channels, Braided/Meandering Reaches

A -- Snake River, 65°28' N, 132°26' W

B -- HH 76-3 82-d, 65°48' N, 133°18.5' W

See text for discussion


A
B
Aggregated Clay Balls
Planar cross-laminated Silt
Structureless Clayey Silt
Rippled Sandy Silt
Laminated Silty Clay
Horizontally Laminated Sand
Horizontally laminated Silt
Medium grained sand,
ripples and trough cross-laminations
Medium grained Sand
Gravel
Large
Clast
0
cm
10

[illegible]n complexes also incorporating finely [illegible]izontally-laminated sandy and clayey silt, rippled and planar cross-laminated sandy silt, microlaminated silty clay, and structureless clayey silt and clay (Figure A4-13). Aggregated clay balls are rarely present as thin (less than 3 cm) horizontal strata. Isolated large clasts underlain by contorted sediments are also rarely present. These clasts may have been dropped on to the sediments by ablating seasonal river ice.

Organic detritus is present as thin horizontal laminations and as draped laminations above ripple and trough cross-laminations. The contacts between subunits are generally erosional where horizontal or trough cross-laminated sand is overlain by silt or clay, or where medium sand overlies finer units. Gradational contacts occur between clayey silt and clay units, and between rippled sand and sandy silt.

The sands, silts, and clays are seldom arranged in consistant fining-upwards or coarsening-upwards sequences. Instead, the vertical succession in most modern channels and in older deposits appears to be random. The units are thin (maximum 40 cm) and are laterally discontinuous. Sand units compose 20 - 50 per cent of the complex sediments, with silts representing 30 - 60 per cent and clays 20 - 30 per cent.

The complexes are interpreted as the results of fluctuating energy levels in the channels, combined with ripple and dune migration and input of fine sediment from bank caving. The laterally and vertically discontinuous deposits indicate that flow and energy levels varied considerably and inconsistently. Once a back bar channel is established, it will be re-occupied during each flood season. High energy events capable of transporting gravel would be relatively uncommon, however, as much of the energy of the flood waters would be dissipated during flow over the bar adjacent to the main channel. Establishment of the bar head jam at the head of the back-bar channel would also serve as a barrier to gravel transport and high-energy flow. Consequently, sediments deposited under lower energy conditions would be expected to quantitatively dominate the stratigraphic sequences. The limited erosion (both vertically and laterally) engendered by lower energy flows would lead to the preservation of irregular bodies of fine sediment.

In the upstream portion of the back-bar channels, seasonally-declining flow leads to the deposition of sequences of silt and clay containing abundant organic drift detritus overlying the complexes. The lower contacts may be gradational with fine sand and silt, or abrupt with coarser sediment. The basal sediment in these sequences is structureless coarse-medium silt, yellowish-grey to olive-brown. The overlying material grades upward to fine, dark, organic-rich silt and clay. These sediments are commonly capped by a thin (5-10 cm) layer of organic-poor yellowish-grey to olive-brown medium to coarse silt. These sequences are interpreted as channel-fill deposits formed during the waning stages of flow, capped by loess deposited after channel abandonment. The palynological assemblages preserved in the organic deposits suggest that most of the drift was locally derived, as is common in fluvial systems (eg. Larimore, 1972; Eckblad et al., 1984). The low organic content of the uppermost strata indicates that aeolian transport of organic matter was not effective.

The downstream portions of the back-bar channels retain some water throughout the summer, either as stagnant sloughs or through influx of minor tributaries. Organic detritus is confined to thin lenses, and massive fine sediments accumulate in the deepest pools. Loading of these saturated muds during the subsequent spring influx may produce plastic deformation structures observed in Pleistocene sequences in the region (cf. Ramos and Friend, 1982).

A bar head drift jam deposit of logs (described further below), and a scour pool are usually developed at the head of a back-bar channel. The scour pool is 2-5 m deep, and usually retains water throughout the summer. Poorly sorted silty sand, either massive or with indistinct mud-draped ripple cross-laminations, is exposed on the margins of scour pools. Irregular mud intraclasts are infrequently present.

The configurations of the scour pools (Figure A4-11) indicates that lateral expansion occurs by undercutting of the attached bank of the back-bar channel, rather than the bank developed in the bar. Flow from the principal channel is directed against the surface facing into the influx, and a plano-convex configuration is developed. Maintenance of the scour pool is aided by the presence of the bar head jam upstream, which necessitates vertically downward flow on the lee side and thus enhances erosion.

During the high flow stages, bank caving occurs along the outer banks of the channels. Caving is promoted by rotational slumping along planes of failure beneath the local thalweg (Turnbull <u>et</u> <u>al.</u>, 1966; Laury, 1971). The channel margin areas, with their fining-upward sequences of sand underlying cohesive silt and clay, are particularly susceptible to slumping and toppling (Pizzuto, 1984a, 1984b). Caving may be accentuated locally by the presence of interstitial ice in exposed underlying beds. The injection of blocks of silt and clay into the back-bar channels by slumping may account for the scattered mud intraclasts present in the sediments. Preservation of these intraclasts is possible in the low energy environment of the seasonal channels (Gibling and Rust, 1984), although any clasts produced along the main channel would be rapidly abraded and disaggregated. Caving persists along the cut banks of the main channel throughout the summer.

Minor channels 0.5 - 1.5 m deep and 1 - 4 m wide are also developed on the larger bars (Figure A4-11). These channels are confined to the sides of the large bars that slope towards the back-bar channels. The minor channels resemble flumes or chutes in form, and are termed "bar-chute channels" (McGowen and Garner, 1970; Bluck, 1976, Blacknell, 1982).

Typical bar-chute channel assemblages consist of erosionally-bounded sequences of granular gravel, sand, and silt (Figure A4-14). The granular gravels are well-sorted, generally of low to moderate sphericity, and highly variable in roundness and Zingg (1935) shape. Fine wood detritus is present in some granule subunits. The deposits are usually structureless, but trough cross-bedding (Theta of Allen, 1963a) and imbrication are rarely visible. The sands are moderately to well-sorted, medium to coarse-grained, and gradationally overlie the gravels. Sand subunits infrequently contain planar ripple cross-laminations (Lambda type of Allen, 1963a) but are normally structureless. The silts are moderately to poorly sorted and structureless, and gradationally overlie the sands. The silt subunits generally fine upwards, but a thin veneer of sandy silt often caps the sequence. Adhesion ripples (Hunter, 1973; Kocurek, 1981) are infrequently present in these strata. The silt strata are invariably overlain along erosional contacts by gravel or sand.

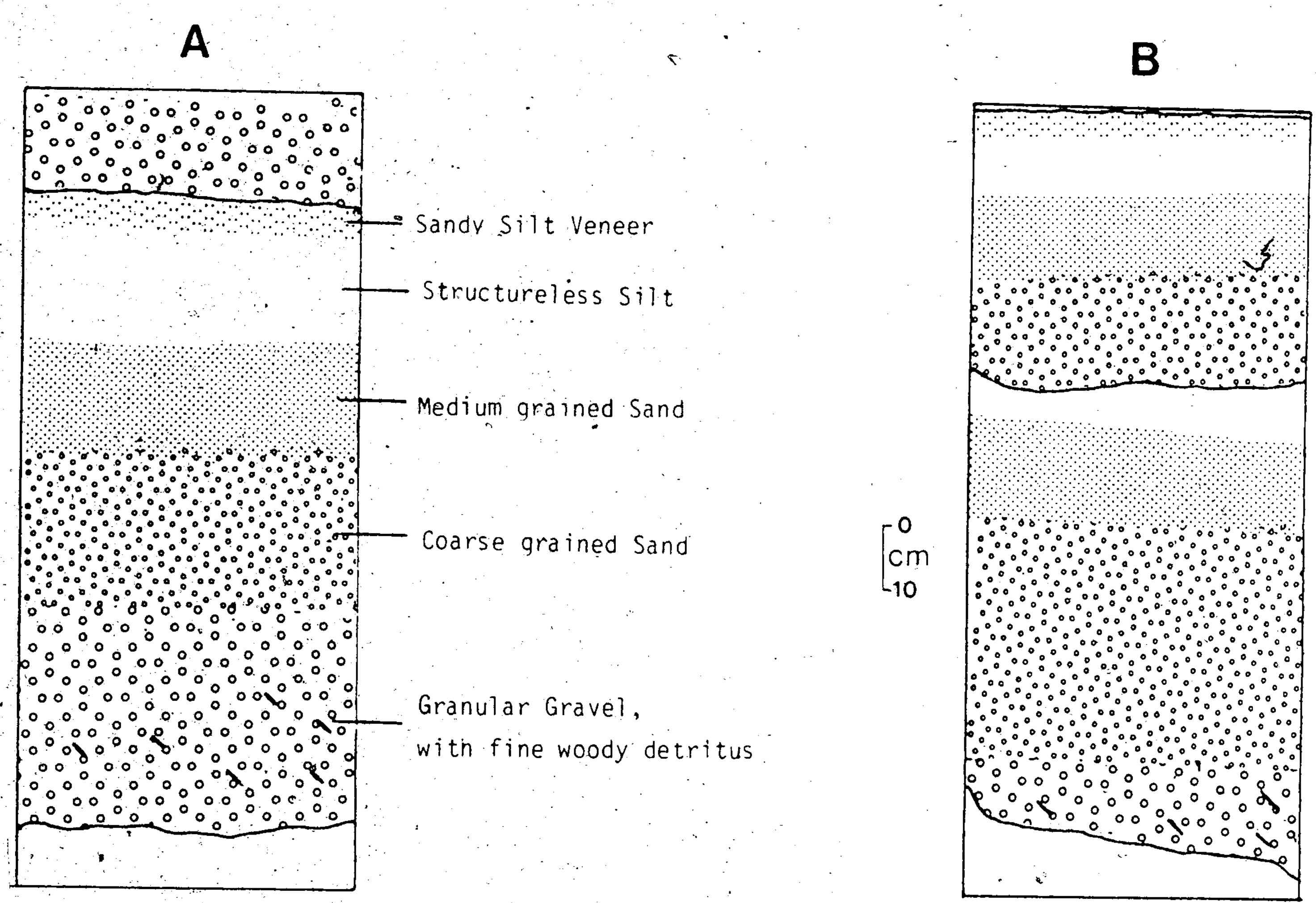
Similar sequences have been interpreted as bar-chute channels developed on major bars by several researchers (eg. Bluck, 1976; Martini, 1977; Levey, 1978; Blacknell, 1982; Kirk, 1983). The variations observed reflect fluctuations in the duration and intensity of flow events. Consistently

Figure A4-14: Sedimentary Sequences, Bar-Chute Channels, Braided/Meandering Reaches

A -- Caribou River, 66°22' N, 134°20' W

B -- HH 62-74 82-1b, 65°53' N, 133°08' W

See text for discussion


A
B
Sandy Silt Veneer
Structureless Silt
Medium grained Sand
Coarse grained Sand
Granular Gravel,
with fine woody detritus
0
cm
10

decreasing flow will produce a gradationally fining-upward sequence, with bedforms constantly being modified and eliminated. Incomplete sequences reflect either erosion or pulsating influx. The regularity of thickness shown by several stacked sequences exposed in active bar-chute channels indicates that channels are stable once established, and that new channels are formed only rarely. Abandonment of the channel, either seasonally or permanently, permits the development of adhesion ripples on the exposed wet-silt surface. The depressed areas also serve as collection sites for sandy silt loess, producing the reverse-graded uppermost silt subunit.

Palynological assemblages from both the back-bar and bar-chute channel sediments are hydrodynamically segregated. The sandy sediments are enriched in the grains which are denser, larger, or more able to resist physical destruction during transport, such as Ericaceae, *Chenopodium*, and *Lycopodium*. Silt strata are enriched in the less dense and more fragile grains, such as *Picea*, *Betula*, and Gramineae. Therefore, the fluctuations in the percentages of these palynomorphs in fluvial sediments may not reflect vegetation changes or climatic alterations, but may be consequences of the hydrodynamic situation.

Confluence of the channels results in deposition of sediment in a variety of features. The shape and internal structure of the confluence deposits is controlled by the relative volumes and persistance of flow of the channels.

Back-bar/bar-chute confluence deposits are lobate, fan-shaped bodies of medium-fine sand and silt, developed from sediments carried in the bar-chute channel (McGowen and Garner, 1970; Ramos and Friend, 1982; Allen, 1983b). The deposit consists of one or more fining-upwards sequences of horizontally laminated sand and ripple cross-laminated fine sand and sandy silt (Figure A4-15), grading laterally into planar cross-laminated sand and silt (Gamma and Epsilon types of Allen, 1963a). In the confluence area, the bar-chute channels are floored with straight-crested and linguoid asymmetric ripples developed in sandy silt, draped with fine organic detritus. Twigs and plant fragments are aligned parallel to the direction of flow.

The development of planar cross-beds and ripple cross-laminations by the bar-chute channel produces structures differing from the thalweg orientation. Although the orientations may differ by as much as 90°, lesser deviations are more common (Smith, 1972). This tendency is accentuated by the fan shape of the deposit. The majority of the ripples in the distal portions of confluence deposits are oriented parallel to the trends of the back-bar channels, however.

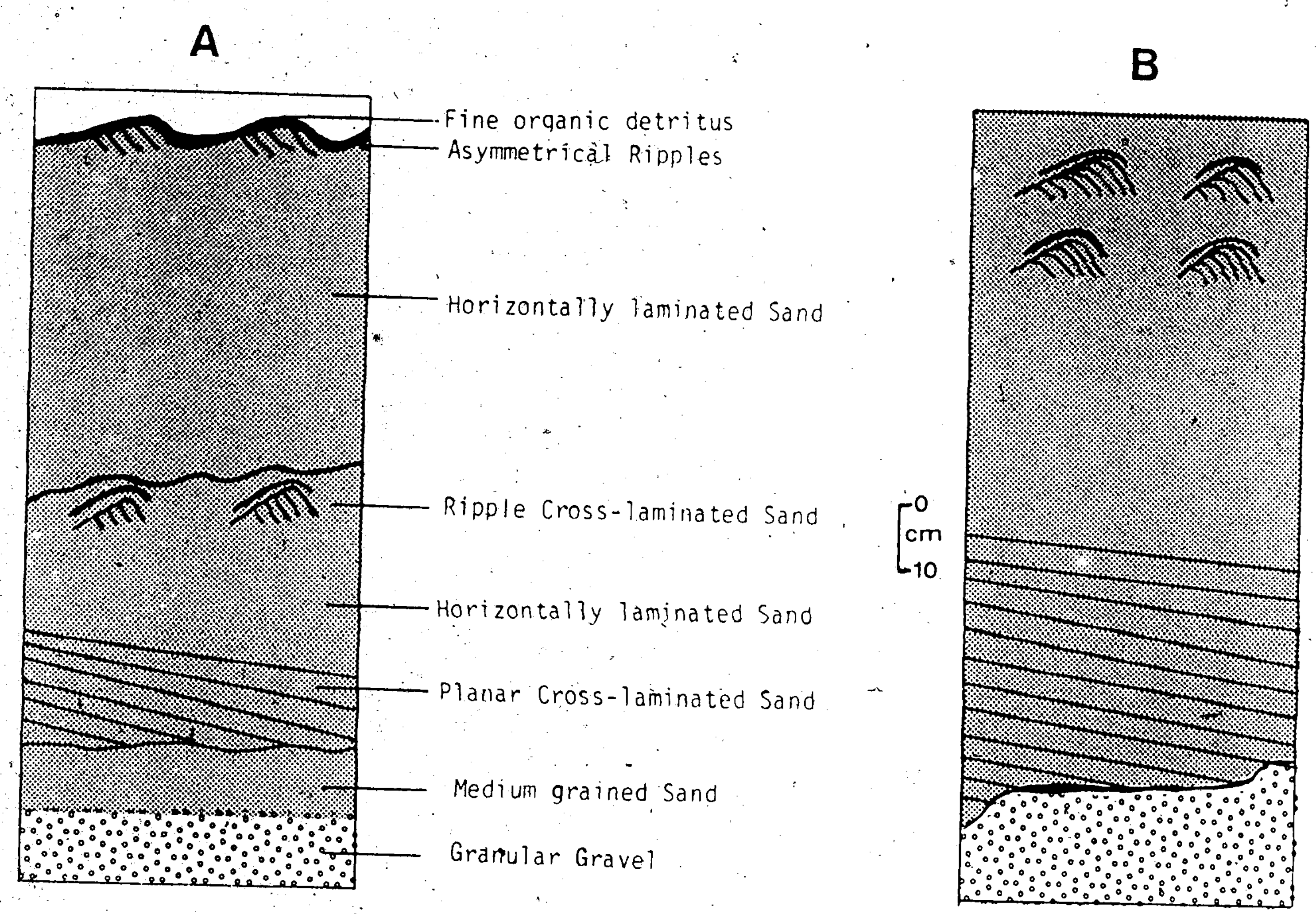
Confluence zones between bar-chute and subsiduary channels are marked by small deposits of medium-fine sand and silt. The forms of the deposits are irregular. Most of the sediment deposited by the bar-chute channel is rapidly transported away from the confluence site by the subsiduary

Figure A4-15: Back-Bar/Bar-Chute Channel Confluence Deposits

A -- Rat River, 67°40'N, 135°29' W

B -- HH 62-74 82-1a, 65°53' N, 133°08' W

See text for discussion


A
B
Fine organic detritus
Asymmetrical Ripples
Horizontally laminated Sand
Ripple Cross-laminated Sand
Horizontally laminated Sand
Planar Cross-laminated Sand
Medium grained Sand
Granular Gravel
0
cm
10

channel. The result is a laterally-truncated body of medium-fine sand, capped with rippled silt. These features are uncommon and the preservation potential is low.

In the confluence zone of the main channel and the back-bar channel, a linguoid, fan-shaped, or crescentic bar that partially or completely blocks the back-bar channel is present. Asymmetric ripples on the surfaces of these features indicate that the bars were constructed by water flowing from the main channel into the back-bar channel. As the main channel waters always possess higher kinetic energy than the back-bar channel waters, no bars are constructed by the back-bar channel in the confluence zone.

The internal structures of the backflow-produced bars consist of gently-inclined planar cross-stratified medium and coarse sand, interbedded with asymmetric ripple cross-laminated fine sand and coarse silt, with minor amounts of draped medium silt. The orientations of the cross-stratification indicate flow towards the back-bar channel (Figure A4-16). Some of these bars serve as traps during the flood season for floating branches and other organic detritus. The resulting bar tail drift jams are aligned parallel to the trend of the bar. Adhesion ripples are developed on the surfaces of some exposed bars.

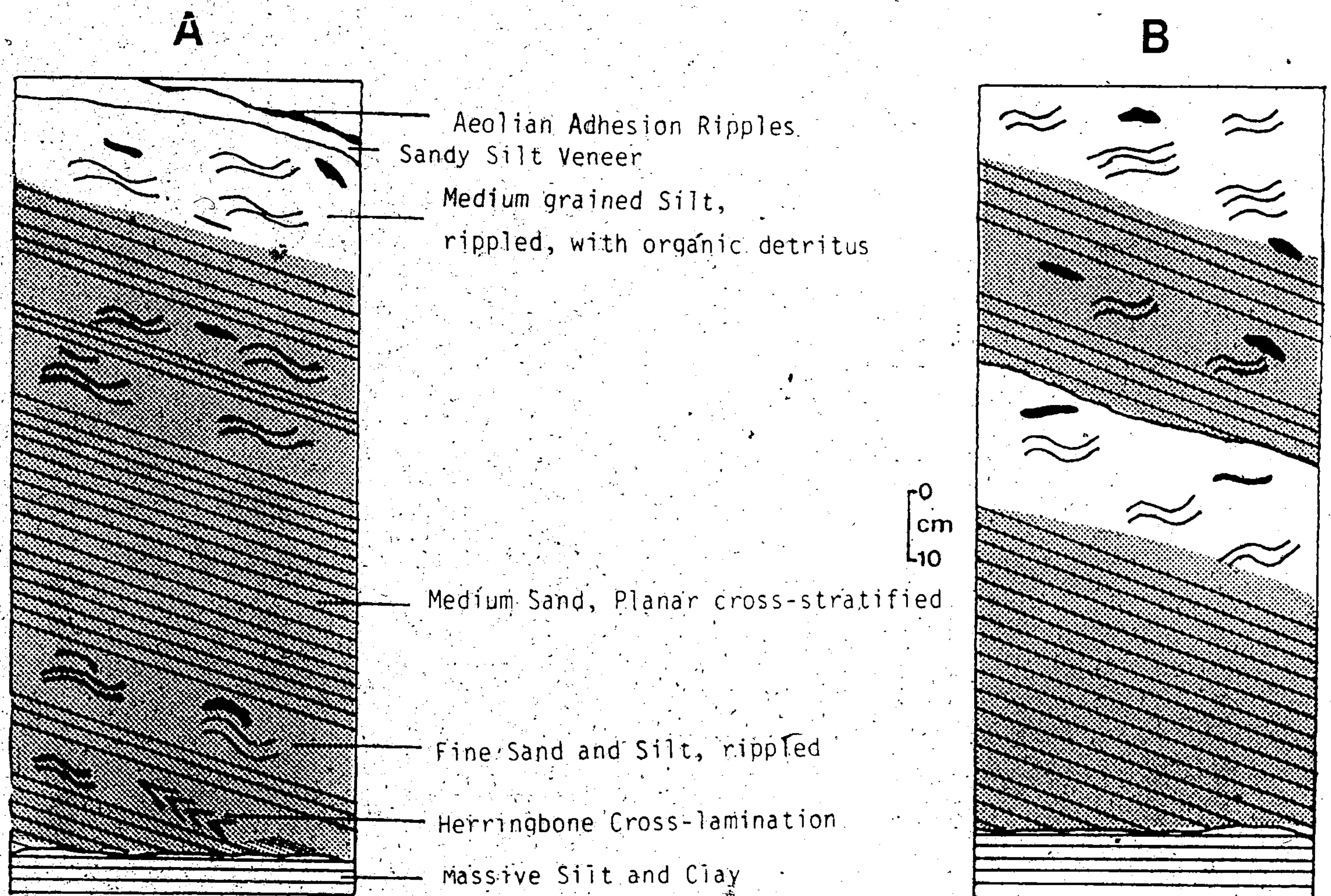
The confluence bars are separated from the main bars by a slough channel 1 - 2 m deep, which serves as a sediment trap for fine sand, silt, clay, and organic detritus. Deposits formed in the slough areas are poorly sorted and

Figure A4-16: Backflow Bar Sedimentary Sequences

A -- Snake River, 65°48' N, 133°18.5' W

B -- HH 76-3 82-c, 65°48' N, 133°18.5' W

See text for discussion


A
B
Aeolian Adhesion Ripples
Sandy Silt Veneer
Medium grained Silt,
rippled, with organic detritus
Medium Sand, Planar cross-stratified
Fine Sand and Silt, rippled
Herringbone Cross-lamination
Massive Silt and Clay
0
cm
10

either horizontally laminated or massive. The slough channel may be deeper than the back-bar channel, or the channels by be of equal depth.

In the zone where the slough channel meets the back-bar channel, the flow direction alternates throughout the year. This backflow zone is characterized by riples of varying orientations, locally producing herringbone cross-lamination. The backflow-produced bars prograde over the slough and backflow deposits, producing coarsening-upward sequences.

Bars formed by backflow from the main channel into subsidiary channels are also developed in the braided-meandering reaches. The bars are modified by flow from the subsidiary channel. Consequently, the features are sinuous and obliquely oriented, with the upstream end dominated by backflow sedimentation and the downstream end by sedimentation from flow of the subsiduary channel. The internal structures of these bars are similar to those of the back-bar/main channel confluence bars.

The dominace of the main channel, and the resultant potential for backflow into the subsiduary and back-bar channels, produce bar deposits that differ somewhat from those produced when channels of equal strength or differing floor elevations meet (eg. Collinson, 1970; Smith, 1970, 1971; Cant, 1978; Allen, 1983b). The backflow currents also produce bedforms and cross-stratification with orientations different from the trends of the channel thalwegs.

Recognition of backflow deposits is important, therefore, if flow direction indicators in ancient deposits are to be used to infer palaeocurrent directions. In laterally extensive sections, the morphology of the units and the progradational sequences developed over slough or subsiduary channel deposits would be sufficient to enable identification of backflow deposits. In isolated small exposures, however, a section through a backflow-produced crescentic bar could not be differentiated from a mid-channel bar exposure or the upper portion of a bar-chute deposit.

The dominant large bar or sand flat type is a diamond- or kite-shaped feature oriented parallel or slightly oblique to the main channel. Similar features have been termed diagonal bars by Church (1972), Smith (1974), and Hein and Walker (1977); lateral bars by Bluck (1976), and kite-shaped emergent sand flats by Allen (1983b). Where bordered by a back-bar channel, the features are effectively attached to the valley margin throughout most of the year. They thus develop similarly to both the point bars described from sand- and gravel-dominated meandering stream reaches (eg. McGowen and Garner 1970; Gustavson 1978; Jackson 1978; Blacknell 1982), and to those developed along gentle meander bends of meandering reaches of Peel Plateau and Peel Plain streams.

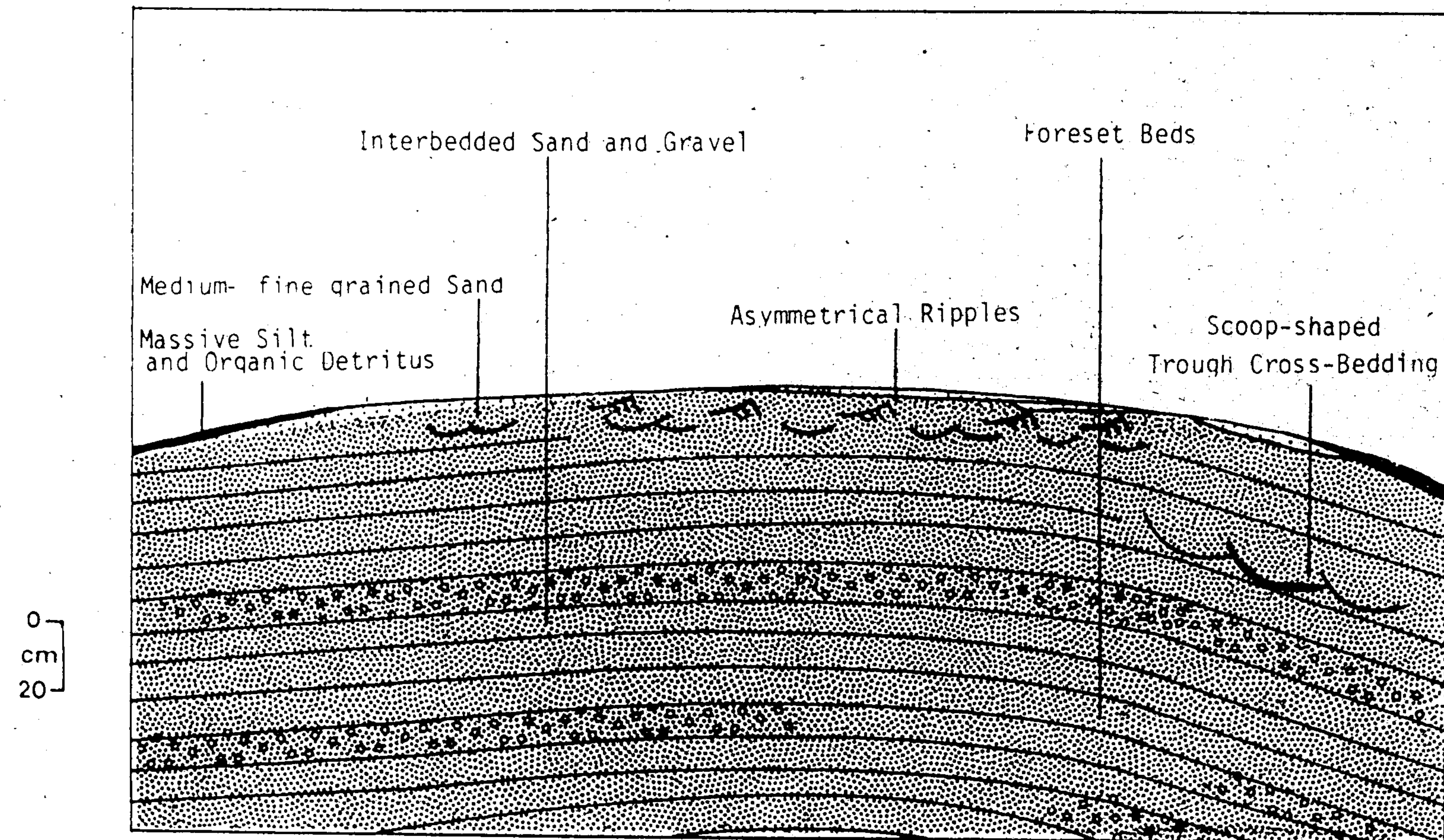
The basal upstream zone of a typical large bar is composed of interbedded horizontal or very gently inclined upstream-dipping foresets of medium-coarse sand with interbedded granule and fine pebble deposits (Figure A4-17). The foresets are aligned at 30° - 45° to the thalweg trend. The sediments fine both downstream and towards the back-bar or subsiduary channel. The central portion of the bar is dominated by horizontally stratified coarse sand, granules, and pebbles. Downstream, the sediments grade laterally into tabular foreset cross-strata (type Gamma of Allen, 1963a) composed of medium-coarse sand, granules, and fine pebbles. Pebbles are aligned transverse to the foresets with intermediate axes oriented down the foreset slope. Alternating coarse and fine beds and laminations present in the foreset sequence were produced by successive avalanching of minor dune bedforms that migrated over the slip face (Smith, 1972; Steel and Thompson, 1983; Rust, 1984). The dip directions of the foresets are aligned at 20° - 50° to the thalweg trend in the downstream areas. Scoop-shaped trough cross-bedding (Pi type of Allen, 1963a) is present in some exposures.

The opposed direction of the upstream and downstream foresets, and the alignment of the intermediate axes of the pebbles parallel to the foreslope trend, produce opposed imbrication patterns at the bar head and tail. Similar imbrication patterns were noted in Oregon stream sediments by Byrne (1963). Pebble imbrication in isolated small lateral bar exposures may not be a wholly reliable guide to palaeocurrent direction. When present with foreset bedding, care must be taken to ascertain whether the deposits were

Figure A4-17: Schematic Cross-Section, Large Lateral Bar,
Braided/Meandering Reach

The cross-section is oriented parallel to the bar axis.
River Flow is assumed to be from left to right.

See text for discussion


Interbedded Sand and Gravel
Foreset Beds
Medium- fine grained Sand
Massive Silt
and Organic Detritus
Asymmetrical Ripples
Scoop-shaped
Trough Cross-Bedding
0
cm
20

formed on the upstream or downstream side of the lateral bar. Erroneous palaeocurrent determinations could result from incorrect identification of stoss-side deposits (Taylor et al., 1971).

The bars are capped by medium to fine sand units 20 - 50 cm thick. The sands are trough cross-bedded (Nu of Allen, 1963a) at the base, and grade upwards to ripple cross-laminated and horizontally-laminated fine sediment (Figure A4-17). The lower foreslope area (#15 on Figure A4-11) and the margins of active channels (#18 on Figure A4-11) remain moist throughout the summer. Straight-crested transverse aeolian ripples and aeolian adhesion ripples (Hunter, 1973; Kocurek, 1981) are present. Massive silt with fine organic detritus is patchily distributed over the lower foreslope, usually occuring on top of small undulations. The upper foreslope and bar crown areas, and dry, exposed channel margin sediments, are devoid of fine organic detritus, silt, and adhesion structures. Aeolian reworking of these areas is indicated by the presence of well-sorted medium sand lag deposits and dreikanter, and by straight-crested asymmetrical ripples.

The bar crown and flank areas rarely display grooves 2 - 10 cm deep and 10 - 30 cm long, aligned parallel to the bar-chute channels. The grooves deepen in the direction of transport. Coarse or medium sand is usually present along the bases of the grooves. Some grooves contain a pebble at the distal end, but most lack coarse clasts. Sediment at the distal end is contorted into ball-like structures. Similar infilled grooves are present in older bar-crown sediments. Associated with the grooves are irregular closed depressions 20-50 cm in diameter and 10-30 cm deep in the bar surface. Coarse sand and granules occupy the floors of these microfeatures.

Both the grooves and depressions are attributed to the actions of transported blocks of river ice during spring breakup (cf. Collinson, 1971). Small ice blocks, dragged along the bar surface during waning flood stages, plough the substrate to form the contorted sediment balls. Tabular blocks of ice become grounded and covered with sediment, forming depressions upon melting. The scarcity of these features reflects their formation at the end of the flood season, when most of the river ice has already melted.

Arboreal vegetation is confined to the areas surrounding the back-bar channels, and to the valley margins. The channel margin vetetation is dominated by _Salix_ and _Alnus_. Vegetated islands are rare, and appear to have been produced by avulsion rather than by colonization of an exposed pre-existing sand flat or bar. Repeated flooding and short growing seasons prevent the rapid colonization of exposed sediment, as has been documented in moderate climates (eg. Haner, 1984).

Drift jam deposits of large logs and smaller fragments of organic drift are developed at the heads of major bars and sand flats (#21 on Figure A4-11), along the flanks of the bars (#22), and on confluence deposits at the mouths of back-bar channels (#23). The flank and tail jam deposits are seldom areally extensive. Logs in the flank jams are aligned at 15°-40° (azimuth), to the thalweg, parallel to the bar-chute channels, and are flat-lying. Logs in the tail jam deposits are oriented parallel to the direction of flow inferred from the underlying confluence deposits, and are also flat-lying.

Bar head drift jam deposits are generally thicker and more extensive than other jams. Large logs in the basal portion of the deposit are oriented parallel to the back-bar channel trend, and the smaller logs are interlocked within this framework. The upper portion of the jam consists of logs oriented irregularly, forming a tight interlocking framework. Fine detritus and clasts of all sizes are present in the interstices. Large jams stand 1-2 m above the surface of the adjacent bars and channel banks, and smaller jams are correspondingly elevated above the adjacent smaller fluvial features. Bar head jams also develop at the heads of subsiduary channels, but usually do not extend across the thalweg. The jam deposits are permeable.

Tributary streams rework the finer-grained deposits on the exposed channel margin surfaces, producing thin lens-shaped deposits of coarse sand. The back-bar channels prevent tributary streams from reworking lateral bar and sand flat deposits.

The braided-meandering stream environments present on the Peel Plateau are characterized by sedimentary successions representing main, subsiduary, back-bar, and bar-chute channels and foreslopes of lateral bars and sand flats. Although the sedimentary record is somewhat biased towards high energy events (Crowley, 1984), thick successions of back-bar and bar-chute sediments are preserved in several locations.

D. Ingrown Meandering Reach

Ingrown meandering streams are characterized by extensive downcutting of both bedrock and previously deposited fluvial sediment. They differ from incised or entrenched meandering streams in that the floodplain is extensive and well-developed, the channel margins are not necessarily composed of downcut material, and the cross valley profile is strongly asymmetric (Lewin, 1978 (type A); Brakenridge, 1984). Downstream migration of meanders and avulsion across point bar surfaces can occur.

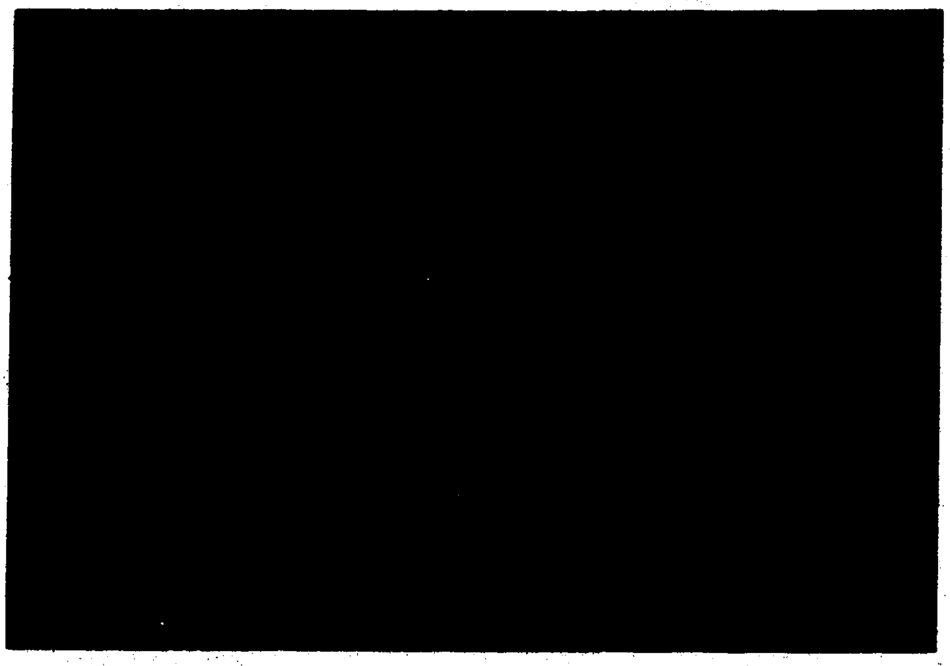
The only ingrown meandering reach in the study region is located along the Peel River, between Nihtal Git Creek and the Snake River (Plates 41 and 42). The river has downcut through lacustrine sediment, Mississippian shale, and the Devonian Imperial and Canol Formations to a depth of 200 m. The gradient in the reach is 1.3 m/river km, and the sinuosity index is 1.5. Cross-valley profiles are strongly asymmetrical.

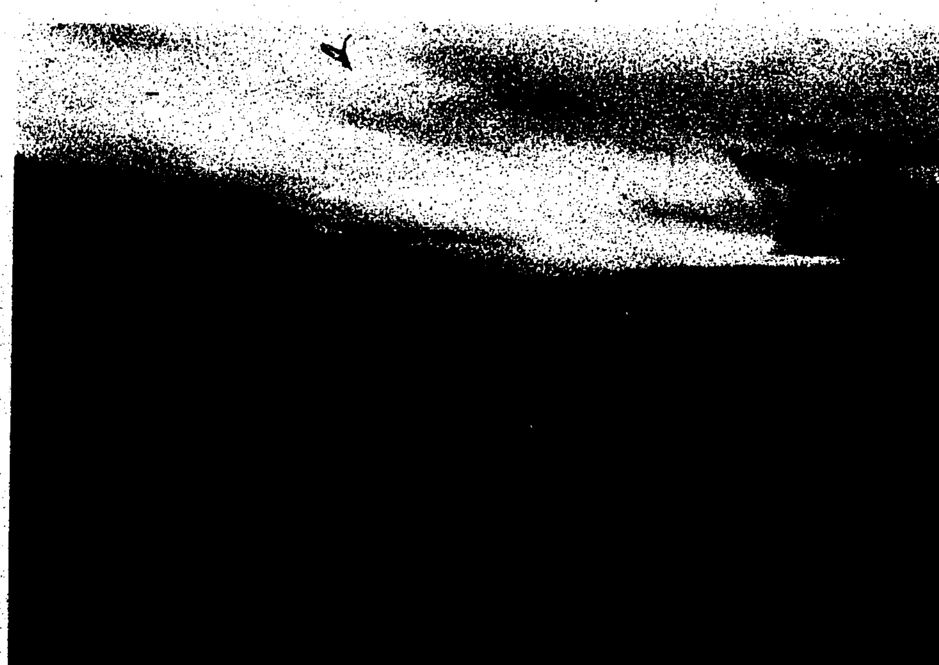
Plate 41 (top)

Ingrown Meandering Reach, Peel River.

Plate 42 (bottom)

Point Bar developed by ingrown meandering reach, Peel River. See text for discussion.





The western portion of the ingrown meandering reach (Figure A4-18) illustrates the features and sediment typical of the river and the Quaternary deposits flanking it. Observations made in August 1981 and July and August 1982 indicate that the main, sinuous channel varies in width from 70 m to 250 m. Although no quantitative assessment of bedload texture was made, shallow riffles in the margins of the main channel were covered with coarse sand and gravel. Suspended material was carried along the thalweg trend, and the channel margins were not turbid.

The relationship of the main channel to the point bar indicates that most of the channel segments were developed by migration around the leading edges of point bars, rather than being products of avulsion. Channel segments along the bar flanks are asymmetric in plan view (Figure A4-18), with the apparent radius of curvature decreasing progressively in the downstream direction. The channel thalweg intersects the cut banks at angles between 50° and 90°, and the steepness of the cut banks at these locations indicates that erosion is most active here. The thalweg length from the upstream inflection point of the meander to the intersection with the cutbank is 40-60% of the length from the intersection point to the downstream inflection point. Avulsion channel segments are straight, and truncate either the entire point bar or the upstream portion only.

The majority of the principal secondary channels are avulsion features. When observed in mid-August 1981, none of

Figure A4-18: Ingrown Meandering Reach, Peel River

1 -- Main Channel

2 -- Primary Avulsion Channel

3 -- Secondary Avulsion Channel

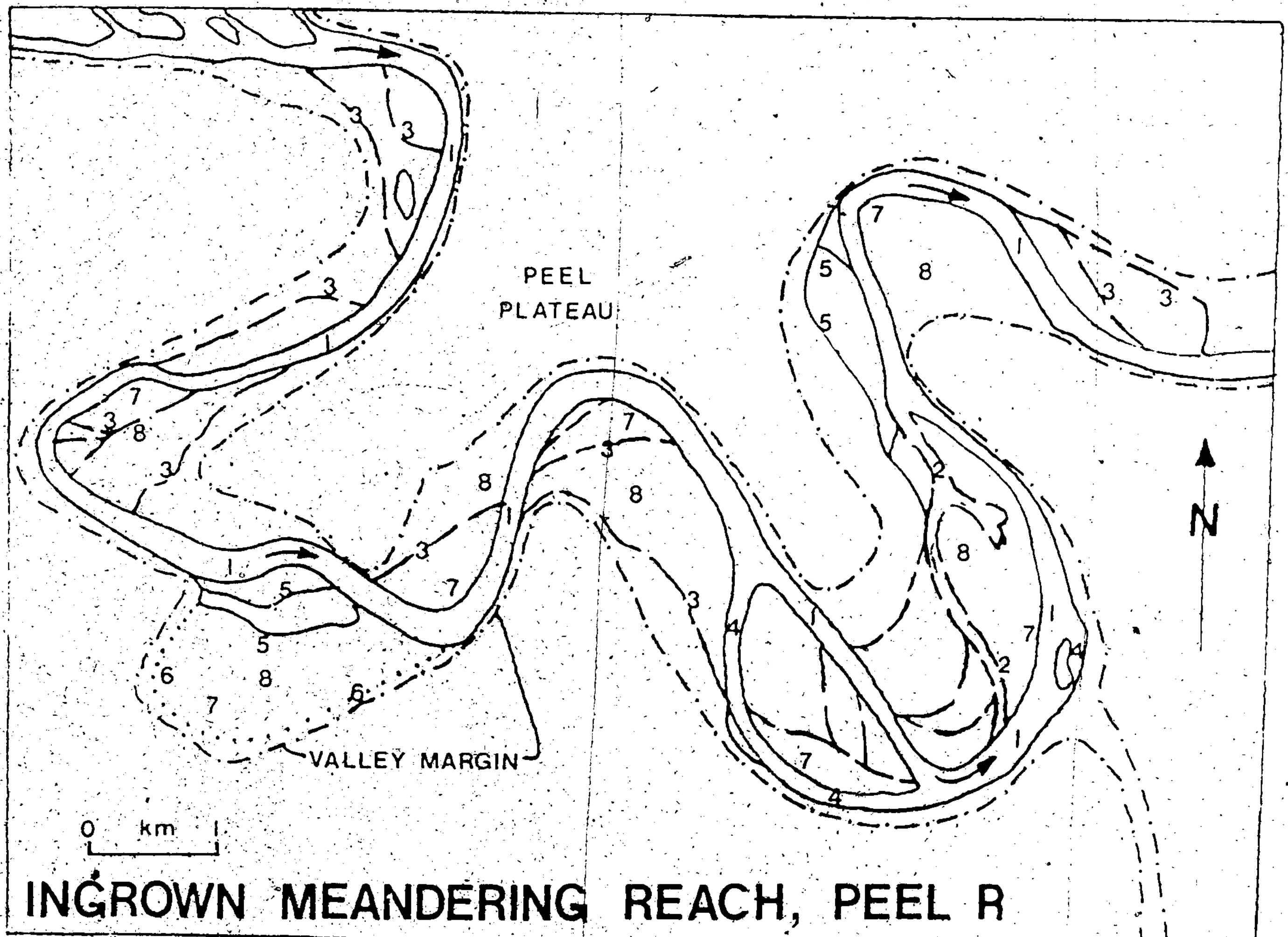
4 -- Primary Abandoned Channel

5 -- Secondary Abandoned Channel

6 -- Abandoned Main Channel

7 -- Point Bar

8 -- Point Bar Crown

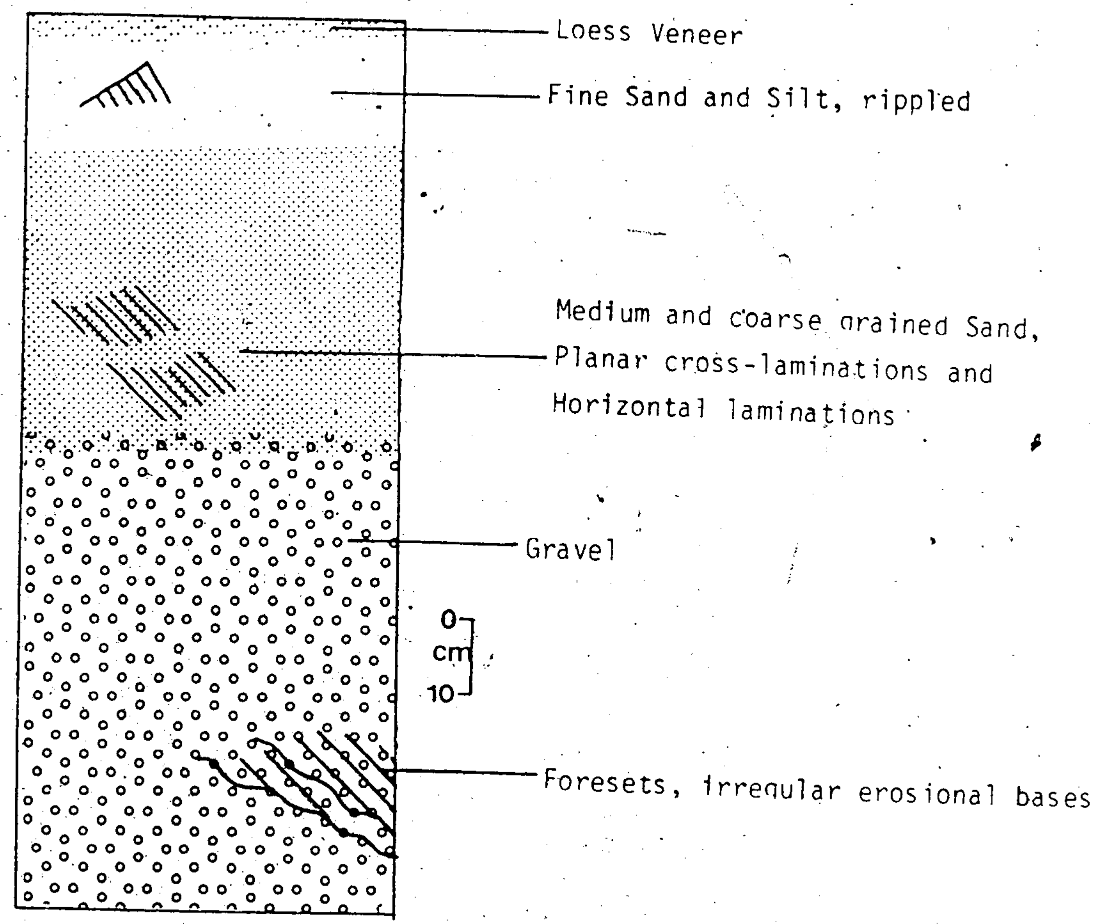

PEEL
PLATEAU
VALLEY MARGIN
N
0 km
INGROWN MEANDERING REACH, PEEL R

these channels contained flowing water, although stagnant ponds were present. The major avulsion channel in the western reach (#2 of Figure A4-18) was active in mid-July 1982, but had ceased to flow in early August.

The basal sediments of the major avulsion channels are moderately to poorly sorted, matrix-supported granule to pebble gravels, with coarse to medium-grained sand matrices (Figure A4-19). Planar foreset bedding is present in the downstream portions of some channels. The foresets have erosional, irregular bases; and are developed normal to the trends of the avulsion channels. No textural grading was observed. The pebble fabrics are dominately transverse to the orientations of the channels. The clasts are generally subrounded and roller-shaped or spherical. These units have a minimum thickness of 50 cm.

Overlying the gravels in the major avulsion channels are horizontally bedded and laminated (0.5 cm to 5 cm thick) medium to coarse sand strata, 10-40 cm thick. Planar cross-laminations (type Lambda of Allen, 1963a) are present in some exposures, and grade laterally into horizontally laminated sediments. The cross-laminations are developed only in medium-grained sands. The contacts between the sands anbd the underlying gravels may be gradational, abrupt and planar, or erosional, although gradational contacts predominate. The minor avulsion channels are floored with similar cross-laminated and horizontally-laminated sand. Both minor and major channels are usually capped with fine

Figure A4-19: Avulsion Channel Sediments, Ingrown Meandering Reach, Peel River


Loess Veneer
Fine Sand and Silt, rippled
Medium and coarse grained Sand, Planar cross-laminations and Horizontal laminations
Gravel
0
cm
10
Foresets, irregular erosional bases

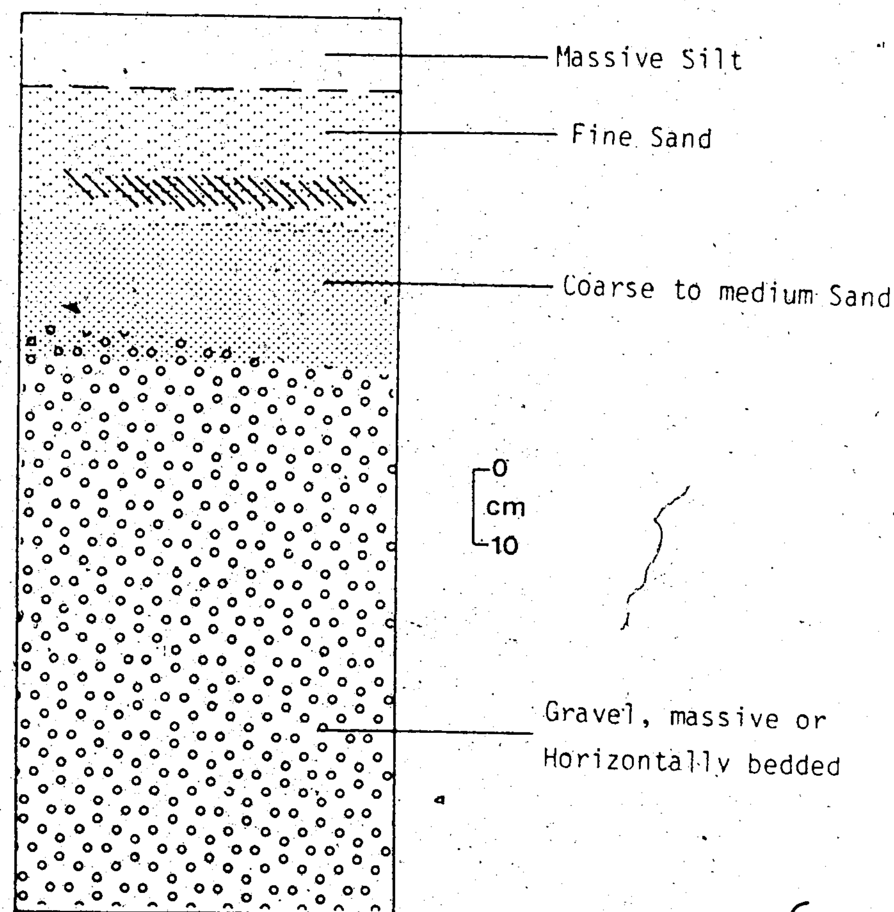
sand or silt, either massive, finely laminated, or rippled. Adhesion-ripple structures (Kocurek, 1981) and thin loess veneers are common.

Similar sequences were observed by McGowen and Garner (1970) and Bridge and Deimer (1983) in channels developed in point bar sequences. The straight channels, without splaying features or diverging foreset bed orientations, indicate that the channels were established, regularly-utilised watercourses, rather than being produced by single avulsion events. The absence of crevasse splay deposits would be expected in mixed gravel and sand load rivers (Jackson, 1978), which characteristically lack levees. Gradual slackening of the flow level is shown by the transition from high-energy, slipface avalanche gravel-dominated cross-beds to sand-dominated planar cross-laminations, produced by straight crested ripples (Allen, 1963a). Rapid declines in energy levels, or sporadic pulses of flow during flood waning, could produce the erosional contacts observed between some fine and coarse cross-laminated units.

The major secondary channels along the flanks of point bars represent portions of the principal channels which have been largely abandoned due to avulsion. Sediments in these channels consist of a basal layer (minimum thickness 1 m) of well to moderately sorted horizontally bedded or massive clast-supported pebble gravel, with granules and coarse sand in the matrix (Figure A4-20). The matrix concentration commonly decreases with depth in the gravel units. The clast

Figure A4-20: Secondary Channel Sediments, Ingrown Meandering Reach, Peel River

See text for discussion


Massive Silt
Fine Sand
Coarse to medium Sand
0
cm
10
Gravel, massive or
Horizontally bedded

fabric is oriented normally to the channel thalweg orientation. Overlying the basal gravel is coarse- to medium-grained horizontally laminated sand, grading upwards to finely cross-laminated medium to fine sand and horizontally laminated or massive silt. These sediment packages are 15 cm to 50 cm thick.

Minor channels developed along the margins of point bars are floored by medium to fine sand, horizontally laminated or ripple cross-laminated (types Kappa and Lambda of Allen, 1963a). The contacts with the underlying sediment are planar or concave upward. The channels are infilled with finely horizontally laminated or massive silt.

The minor channels are usually connected directly to one or more avulsion channels, and cannot be traced around the complete margins of point bars. They thus appear to represent avulsion channels which followed the depressions created by the thalweg of the principal channel, rather than being abandoned segments of the thalweg. Consequently, they lack the coarse sediments characteristic of the major channels.

Abandoned meander lakes (oxbows) are not present along the ingrown reach of the Peel River. The abandoned segments of the principal channel are infilled with massive, organic-poor coarse silt and fine sand, overlain by finely laminated to massive turbated organic-poor fine silt and clay. The coarse silt and fine sand units are discontinuous, and reach a maximum thickness of 10 cm. These units are interpreted as loess, deposited shortly after channel abandonment. The fine silt and clay units are laterally more continuous, and are 5 cm to 20 cm thick. These sediments are interpreted as sheetflow deposits, formed after colonization of the channels by arboreal vegetation. The absence of repetitive cycles of channelling, loess deposition, and sheetflow sediments indicates that avulsion into former meander loops did not occur. The low organic content of the fine sediments suggests that flooding was not sufficiently severe or prolonged either to bring large quantities of organic drift to the site or to regularly drown large amounts of arboreal vegetation.

The foreslopes of the point bars are composed of massive pebble gravels and crudely stratified, low angle planar cross-bedded clast-supported pebble and granule gravel, interbedded with pebbly coarse and medium sand. The contacts between foresets are erosional, and fine sediment is only present dispersed in the sand units. Clusters of imbricated pebbles 20 cm thick and 1 m long are present. The clast fabric is oriented normal to the cross-bedding, and is parallel or oblique to the inferred thalweg. The clasts are subrounded and of moderate sphericity. No fining-upward trends in the succession were observed.

The sediments of point bars of gravel-dominated and sand-and-gravel dominated meandering rivers commonly possess these characteristics (eg. McGowen and Garner, 1970; Gustavson, 1978; Jackson, 1978; Blacknell, 1981, 1982). Flow divergence across the bar foreslope is generally attributed to the influence of helicoidal secondary flow (Hooke, 1975) or to transverse gradients in sediment transport rate (Edwards et al., 1983). The absence of fining-upward sequences, commonly not present in coarse-grained point bars, is attributed to the re-distribution of coarse sediment (initially deposited at the bar apices) during subsequent high flow events (cf. Levey, 1978).

The sediment on the bar crown areas is dominated by horizontally laminated medium and fine sand and structureless silt. These deposits are interpreted as sheetflow and high flood stage events, partially reworked by aeolian activity. Limited vertical sections (maximum 3.5 m thick) exposed in the bar crown and foreslope areas along avulsion channels reveal gradational transitions between the sediments of the bar crown, foreslope, and principal channel environments. The sedimentary successions indicate consistent lateral migration of the bars, with no major avulsion events. The limited thickness of the deposits exposed, however, makes assessment of the frequency of avulsion events impossible.

The total thickness of bar, channel, and floodplain sediments in the reach is unknown. Ingrown meander deposits have not been recognized at any other locations within the Peel Plateau, Peel Plain, Richardson Mountains, or Canyon Ranges. Exposed fluvial deposits bordering the ingrown reach are confined to the basal 10m of the valley sides. Although

Cenozoic sedimentary deposits are present at higher elevations within the valley, extensive flows and slides along the valley margins prevent recognition of the original depositional environment of the sediments.

Ingrown meanders are usually initiated by a combination of regional tectonic activity and a change in base level. In the Peel Plateau, deglaciation would be expected to produce an (indeterminate) amount of isostatic rebound, combined with an increase in regional stream gradient as shorter pathways to the Beaufort Sea gradually were uncovered. The increase in regional stream gradient would thus precipitate downcutting, despite the increase in sea level upon deglaciation estimated at 50-60 m by Forbes (1980) and 60-80 m by P.Hill and S. Blasco (Atlantic Geosciences Centre, Halifax, personal communication, 1984). The simultaneous occurrence of glacioisostatic rebound and fluvial downcutting would act to produce a deep valley, similar to antecedent valleys in active orogenic regions, much more rapidly than would fluvial downcutting in a tectonically stable region.

E. Lower Reach, Peel River

The portion of the Peel River downstream (north) of Satah Creek is straight, with low sinuosity and braiding parameters (Table 4-4). The reach is characterized by low gradients.

Table 4-4

Channel Properties, Lower Reach, Peel River (67°15'-35' N, 134°51'-54' W)

Gradient: Less than 0.1 m/km

Sinuosity (after Leopold and Wolman, 1957): 1.06

Braiding Index (after Brice, 1964): 0.7

Braiding Parameter (after Rust, 1978): 1.4

Date of Photography: April 1954

Sediments exposed along the reach are predominantly finely laminated to massive silts and clays, moderately to poorly sorted, with interbedded minor lenses and laminations of fine organic detritus. Beds of horizontally-laminated medium-fine sand, ripple-laminated fine sand and silt, and aeolian adhesion-rippled fine sand (Kocurek, 1981) are also present. Coarse-grained deposits are rare, and are generally unstratified. The exposures generally show fining-upward sequences.

These exposures are interpreted to represent overbank floodplain sedimentation, with the high energy ~~stages~~ of severe floods represented by the massive coarse deposits, and more usual seasonal flooding represented by the fine sands, silts, and clays. The adhesion ripple sequences may develop either in fluvial sediment or in loess. Channel sequences are not present in any of the exposures, suggesting that the present river thalweg has remained stable for a significant period. Straight-reach deposits are not present at any other locations within the study region.

XVI. Appendix 5: Palynological Analysis

A. Palynological Analysis Techniques

Processing of the peat and sediment samples to recover palynomorphs followed the techniques devised by C. E. Schweger and T. Habgood at the University of Alberta, Department of Anthropology. Peat samples were boiled in sodium hydroxide, acetolysed, rinsed in anhydrous alcohol and acetone, and mounted unstained in 200 ul aliquots of silicon oil. Sediment samples were boiled in sodium hydroxide and treated with hydrochloric acid. The remaining silt and clay was removed through density separation using zinc bromide solution and a hydrofluoric acid treatment. The samples were then acetolysed, rinsed with anhydrous alcohol and acetone, and mounted in silicon oil. Nail enamel was used to fasten the cover slips to the prepared slides. Complete details of the peat and sediment analysis procedures are presented in Appendices 5B and 5C respectively.

Palynological samples mounted in this manner have proven to be susceptible to degradation. The addition of silicon oil to acetone rinsed samples has resulted in the formation of organic molecules capable of attacking the pollen. In addition, the nail enamel used to contain the silicon oil beneath the cover slips of the slides has not been effective in all instances, either because it is permeable or because it is not inert in the presence of silicon oil. The taxa most susceptible to degradation are the legumes, Artêmisia, and Gramineae. Degradation does not appear to begin until several weeks after the preparation of the slides.

Counting was performed utilising the reference collection of C. E. Schweger. For each sample listed as "counted", a minimum of 200 grains were recorded. Although no quantitative calibration of pollen influx was attempted, estimates of pollen concentration were obtained by calculating the total number of grains contained in each aliquot of silicon oil. The number of slides counted varied from 1 to 6, depending upon the richness of the sample. Peat samples proved to be extremely rich in palynomorphs (maximum 3440 grains/aliquot), while the older sediment samples were relatively to extremely poor (35-60 grains/ aliquot). Thirty-eight samples contained small amounts of palynomorphs (less than 25/aliquot), which were insufficient for counting. The taxa observed in these samples were recorded and are reported as trace amounts. Samples considered barren had less than 5 grains/aliquot. Most of these samples contained no organic material of any kind.

Identification of the grains was achieved using the key devised for western Canadian taxa by Habgood (1978, unpublished), with additions from the work of Cwynar (1980, 1982), Bassett et al (1978), Richard (1970), Chanda (1962), and Faegri and Iversen (1975). Diagnostic criteria proposed by these authors were assessed using the reference collection of Schweger. The portion of the key of Habgood which deals with the taxa identified during the study is reproduced as Appendix 5D. The keys of McAndrews et al and Faegri and Iversen (1975) were used as supplementary references. Nomenclature of the taxa follows the system of Porsild and Cody (1980). English equivalents of the scientific names are presented in Appendix 5E.

Pre-Quaternary palynomorphs were not counted, but distinctive types were identified. These grains could theoretically be used to determine the provenance of some sedimentary units. The results of this study are not presented here, but will be presented in a separate report. Although most Quaternary taxa can be reliably identified with experience, three major problems remain. These concern the families Gramineae and Cyperaceae, and the genus *Artemisia*; the genus *Picea*; and the genus *Betula*.

a) Gramineae, Cyperaceae, and *Artemisia*

Colinvaux (1967) noted that the inability of palynologists to subdivide these taxa represented a significant handicap to comprehensive environmental interpretation. This situation has changed little in the intervening years (eg. Ritchie 1977, 1982; Cwynar 1982). Keys devised for the families in other geographical areas (eg. Faegri and Iversen 1975) fail to adequately differentiate reference samples from known taxa currently extant in northeast Beringia. Similarly, attempts to provide criteria for the subdivision of *Artemisia* (Birks 1977) have not proven useful in the northern Yukon (Cwynar 198[illegible]). Consequently, the pollen concentrations from these taxa represent a mixture of ecological situations and can only be used extremely generally for interpretation.

b) Picea

Distinguishing between *Picea glauca* and *P. mariana* would aid considerably in the assessment of soil moisture content and drainage, permafrost development, and age of the stand. The two species have markedly different ecological requirements (Viereck 1975). Various criteria have been proposed, based principally on size. Bassett *et al* (1978) and Habgood (1978 unpublished) both suggest that grains with total breadth greater than 100 um be assigned to *P. glauca*, while those less than 95-100 um breadth can be attributed to *P. mariana*. This criterion was used in this study to obtain semi-quantitative proportions of the species distribution. Since the assignment of grains with breadth measurements of ± 100 um was uncertain, the species are not differentiated on the pollen diagrams. Grains assigned to either species are referred to as *P. glauca* or *P. mariana* type.

c) Betula

Since the inception of quantitative palynology by von Post, the problem of subdividing genus *Betula* into shrub and tree species has occupied a prominent place in boreal and tundra research. Most researchers (eg. Terasmae 1951; Andersen 1961, Praglowski 1966, Ives 1977, and Prentice 1981) use size as a defining criterion, with the tree *Betula* grains considered to be larger than those of the shrubs. Ives (1977) measured the ratio between grain diuameter and pore depth for B. nana, B. glandulosa, and B. papyrifera, and found no clear differentiation among the three species. Grain diameter measurements tended to be larger in arboreal taxa, but considerable overlap existed. Terasmae (1951) suggested that the vestibulae were not as pronounced in the shrub species, and were less deep than the arboreal vestibulae. Terasmae's results are not directly applicable to this study, as he investigated only Scandinavian Betula. However, during the course of this study, the larger grains counted had pores which protruded much more than those of the smaller grains.

Other researchers (eg. Wenner 1953) have argued that the smaller Betula pollen characteristically found in the lower sections of cores and in older sediments have been compressed or shrunken during the fossilization process. Wenner noted that Pinus grains also decreased in size with depth. Nutritional deficiences cannot account for the changes in size,. as all fossil exines devoid of living material fall within the same size range for a given species (Wagenitz 1955).

The approach that has been followed in this study was to differentiate the Betula grains into arboreal and shrub types on a semiquantitative basis. The species are not differentiated on the pollen diagrams. Interpretation of the nature of the population was based on part on grain dimensions and vestibular morphology, and in part on the associated pollen assemblage, following the practice of Ritchie (1982).

B. Processing Peat for Pollen

1. Place peat sample in 100 ml beaker. Add 20 ml of 10 % NaOH, stir well with wooden stick. Boil gently with constant stirring on hot plate for 5 minutes. Allow to cool. Pour off supernatant through fine screen into 150 ml beaker, retaining most of the peat in the original beaker. Wash peat in beaker with several portions of distilled water, pouring off supernatant each time through screen. Transfer the liquid to Nalgene centrifuge tubes, centrifuge, decant. Wash material in bottom of tube with water to which a few drops of 10 % NaOH have been added. Centrifuge, decant. Repeat until necessary until supernatant is straw-coloured.

2. Wash with 10 ml glacial acetic acid.

3. CAUTION. Make up acetolysis mixture. Measure out the desired multiple of 9 ml acetic anhydride intro 125 ml flask. Add cautiously, with swirling, in several portions, the desired multiple of 1 ml concentrated sulphuric acid.

4. Add 10 ml of acetolysis mixture to centrifuge tube. Stir well with clean stick. Heat for 3 to 5 minutes in boiling water bath. Cool. Centrifuge. Decant.

5. Wash with 10 ml glacial acetic acid. Centrifuge, decant.

6. Wash with 10 ml distilled water. Add stain if desired. Centrifuge, decant.

7. Wash with water to which a few drops of 10 % NaOH have been added. Centrifuge, decant.

8. Wash with 10 ml of 98 % ethyl alcohol. Centrifuge, decant.

9. Wash with 10 ml of acetone. Centrifuge, decant.

10. Suspend the pollen in a few ml of acetone, pour into phial. Add 3 drops of silicon oil to phial. Allow acetone to evaporate overnight.

C. Processing Sediments for Pollen

1. Weigh out 10-15 g of sediment into 100 ml beaker. Add 25 ml of 10 % NaOH, stir well with wooden stirring rod. Boil gently with constant stirring on hot plate for at least 5 minutes. Filter immediately through fine screen into 160 ml beaker, retaining sand and coarse particles in original beaker. Add 10 ml distilled water to coarse fraction of sediment, stir, then swirl for 60 seconds. Allow to settle for 30 seconds, pour supernatant solution onto screen. Repeat if necessary. Wash the fine sediment on the screen with distilled water. Transfer contents of beaker to Nalgene centrifuge tubes, centrifuge, decant. Wash sediment in bottom of centrifuge tubes with 10 ml of water to which a few drops of 10 ml of NaOH have been added, centrifuge, decant. Repeat if necessary until supernatant liquid is no longer coloured by humates. Note that most fine sediment

samples contain small amounts of pollen, necessitating the processing of several 10-15 g aliquots. A minimum of 3 is recommended.

2. Add a few drops of concentrated HCl to the sediment, stir well. When the initial reaction has subsided, add a little more HCl, stir again. Continue adding HCl until no further effervescence occurs. Centrifuge, decant.

3. Add 15 ml of zinc bromide solution (specific gravity 1.8-2.0), stir. Stopper tube with rubber stopper, shake gently, and release any accumulated pressure. Stopper again, shake for exactly 3 minutes, then centrifuge for at least 20 minutes. Carefully filter with suction through glass filter paper in Nalgene Buchner funnel. Retain the filtrate for purification and recovery of the zinc bromide. Roll up the filter paper on which the pollen is now deposited, place the wad in clean Nalgene centrifuge tube.

4. CAUTION. USE MASK, GLOVES, and APRON. Cautiously add a few drops of Hydrofluoric acid, swirl. When the violent reaction subsides, add a little more HF, swirl again. Add a total volume of 15-20 ml of HF, stir with CLEAN wooden stirring rod. Loosely stopper tubes with Nalgene stoppers, heat in boiling water bath on hot plate for at least 1 hour. Allow to stand in water bath after boiling for at least 6 hours.

5. Centrifuge, decant. Add 10 ml concentrated HCl, stir. Boil in water bath on hot plate for 5 minutes, cautiously stirring once during the 5 minute period.

Centrifuge while hot, decant. Repeat process at least 6 times.

6. Wash with 10 ml distilled water. Centrifuge, decant.

7. Wash with 10 ml glacial acetic acid. Proceed as for peat samples.

D. Key for the Identification of Quaternary Palynomorphs

This key lists only those taxa which were identified during the present study. The majority of the coding system was devised by Habgood (Department of Anthropology, University of Alberta). Unless otherwise stated, the criteria for identification are hers. Criteria proposed by other authors are referenced following the taxon name. Spores are separated on the bases of the number of laesuri and the nature of the perine. Pollen grains are classified based on the presence or absence and number of pores and colpi, and on the nature of the sculpturing.

Key to Spores

1. Alete

1.1 Psilate

a) Spheroidal. Perine may or may not be present. Equisetum

1.2 Scabrate or verrucate

a) Spore small, 17-22 um. Dicranum

2. Monolete

2.1 Psilate to Scabrate

a) Spore small, approximately 30 um long. Perine in folds. Isoetes

2.2 Scabrate

a) Spore 40 um long. Athyrium

2.3 Rugulate

a) Perine tightly fitting. Spores dimorphic. Cystopteris

aa) Perine loosely fitting. Rugulate-foveolate. Gymnocarpium

2.4 Echinate

a) Echinae up to 6 um high. Polystichum

3. Trilete

3.1 Verrucate Sphagnum

3.2 Fossulate Botrychium

3.3 Reticulate. Muri fine. Lycopodium

Key to Pollen grains

1. Tetrad

a. Components joined tetrahedrally.

b. Components tricolporate.

c. Adjacent grains scarcely delineated (Cwynar 1980, 1982)

d. Tetrad spheriodal, inner walls perforated.

e. Inner walls with numerous perforations throughout, exine psilate Arctostaphylous rubra/alpina ee. Inner walls with few perforations, mostly near juncture with outer walls, exine scabrate *Arctostaphylous uva-ursi*

dd. Tetrad subtriangular, inner walls not perforated

e. Tetrad small, thin-walled, costae linear *Cassiope* (Cwynar 1980) cf. *tetragona* (this work)

ee. Tetrad large, thick walled, costae pennant-shaped *Andromeda* *Polifolia* (Cwynar 1980)

cc. Adjacent grains clearly delineated (Cwynar 1980)

d. Colpae short and irregularly margined, exine coarsely scabrate

e. Colpae with distinct transverse furrows and prominent costae *Rhododendron* *lapponicum*

ee. Colpae without transverse furrows and usually without costae

f. Endexine thicker than ektexine, colpae distinct and not chambered *Empetrum* *nigrum*

ff. Endexine thinner than ektexine, colpus margin indistinct and often chambered *Pyrola*

dd. Colpae medium to long, ektexine psilate-scabrate

e. Exine thick, costae prominent *Ledum* type

ee. Exine medium thickness, costae not prominent or absent *Vaccinium* type.

bb. Components not tricolporate.

c. Components inaperturate. *Juncus*

2. Inaperturate Monads

2.1 Psilate.

a. Grain small, 20-25 um, with scattered small gemmae. Juniperus

aa. Grain large, with narrow annulus, usually non-equatorial. Larix

2.2 Scabrate.

a. Approximately spheroidal. Populus

aa. Not spheroidal. Approximately tetrahedral, with laminae which may be ruptured. Cyperaceae

2.3 Reticulate. Brochi up to 2 um. Potamogeton

3. Vesiculate.

3.1 Sacci 2.

a. Constriction between colpus and sacci. Reticulum of sacci sharply distinct from sculpturing of corpus.

b. Exine of cappa of corpus 1-2 um.

c. Distal portion of corpus without verrucae (Diploxylon) Pinus Banksiana; Pinus contorta

aa. No constriction between corpus and sacci. Reticulum of sacci merging gradually into sculpturing of corpus.

b. Breadth of total grain 80-95 um Picea mariana

bb. Breadth of total grain 100-110 um Picea glauca

4. Monoporate.

4.1. Psilate or scabrate. With Annulus. Gramineae.

5. Triporate.

5.1 Pores vestibulate.

a. Grain and vestibulae small. Betula

aa. Grain large (60 um or greater). Vestibulae large. Epilobium

5.2 Pores not vestibulate.

a. Grain suboblate, semiangular in polar view.

b. Pore aspidate. Endexine dissolved in area of pore. Myrica Gale

bb. Pore not aspidate. Endexine entire in area of pore. Corylus

6. Stephanoporate.

a. Pores weakly vestibulate. Arci present.

b. Polar view diameter 19-26 um, pores strongly aspidate, 1.5-2 um in diameter, wall 3.5-4.5 um thick (Richard 1970, Bassett et al 1978) Alnus crispa

bb. Polar view diameter 20-28 um, pores inaspidate to aspidate, 2-4 um in diameter, wall less than 2 um thick Alnus rugosa/incana

aa. Pores not vestibulate. Arci absent.

b. Pores with heavy collar-like annulus Myriophyllum

7. Periporate.

7.1 Psilate or Psilate-scabrate.

a. Grain spheroidal.

b. Exine 2 um or less.

c. Pores 35-50. Annular thinning. Chenopodiaceae.

bb. Exine thicker, usually 4 um.

c. Pores 8-30.

d. Tectum with scattered distinct perforations, or per-reticulate.

e. Annulus narrow, 17-24 pores. (Chanda 1962) Silene type.

dd. Tectum with indistinct perforations, scabrate. More than 24 pores. (Faegri and Iversen 1975) Sagina type

ddd. Tectum with many minute perforations, pores not in depressions. (Faegri and Iversen 1975) Cerastium type

7.2 Striate.

a. Pores approximately 50. Exine 3-4 um thick. Polemonium

7.3 Reticulate.

a. Exine 3.5 um thick Polygonum

8. Tricolpate.

8.1 Psilate.

a. Intectate.

b. With poroidal area.

c. Poroidal area protruding.

d. Circular in polar view. Grain small, 17-19 um. Fragaria vesca

cc. Poroidal area ruptured.

d. Grain frequently deformed. Viola

dd. Grain well-formed, subprolate. Subangular in polar view. Cornus canadensis

aa. Tectate.

b. Grain large, 70-75 um. Convolvulus

8.2 Scabrate.

a. Tectate.

b. Tectum with scattered microechinae. Tectum undulating. Ranunculus

bb. Tectum not microechinate. Oblate-spheroidal, polar view circular. Anemone canadensis

8.3 Verrucate

a. Grain large, 55-60 um. Verrucae 2 um across. Linum lewisii

aa. Grain not large. Protruding poroidal area. Scattered echinae. Rubus chamaemorus

8.4 Rugulate

a. With poroidal area.

b. Colpus without margo.

c. With operculum. Rosa acicularis

cc. Without operculum.

d. Poroidal area usually not ruptured. Rubus

dd. Poroidal area ruptured. Dryas

8.5 Striate

a. With Poroidal area

b. With operculum. Potentilla

aa. Without Poroidal area.

b. Colpus not equatorially constricted.

c. Striae mostly transverse. Saxifraga alzoides

cc. Striae in whorls. S. tricuspidata type

8.6 Per-reticulate

a. With poroidal area.

b. Heterobrochate, brochi up to 4 um. Salix aa. No poroidal area.

b. Colpus not short.

c. Colpus long, broad, not shallow.

d. Colpus without ragged edge, homobrochate. Cruciferae

9. Tricolporate.

9.1 Psilate.

a. Intectate.

b. Poroidal area sometimes ruptured.

c. Poroidal area protruding. Fragaria vesca

cc. Poroidal area not protruding. Grain frequently distorted. Viola

aa. Tectate.

b. With transverse colpus or equatorially elongated pore.

c. Grain perprolate to prolate. Exine of uneven thickness.

d. Tectum perforate. Heracleum

cc. Grain not perprolate to prolate.

d. Colpus slit-like. Grain approximately 35-40 um in length. Polygonum Bistorta

9.2 Scabrate.

a. Grain subprolate to prolate, colpus not constricted. Shepherdia canadensis

9.3 Verrucate.

a. With poroidal area. Poroidal area protruding, scattered echinae. Rubus chamaemorus aa. With true pore. Grain spheroidal, colpus long. Artemisia

9.4 Rugulate.

a. With poroidal area.

b. With operculum. Rosa

bb. Without operculum.

c. Poroidal area usually not ruptured. Rubus

cc. Poroidal area ruptured. Dryas

9.5 Striate.

a. With Poroidal area.

b. With operculum.

c. Colpus with 2 lip-like projections in polar view. Potentilla anserina, Fragaria virginiana

cc. Colpus without projections. Faintly striate. Potentilla fruticosa type

9.6 Echinate.

a. Microechinate, scattered echinae. Linnaea borealis

aa. Not microechinate. Echinae arranged along meridians.

b. Spines sharp-tipped. High spine. Arnica, Aster types

bb. Spines blunt-tipped. Low spine.

c. Grain 35-40 um. Saussurea

9.7 Suprareticulate.

a. Equatorial limb not rhombohedral, prolate, heterobrochate.

b. Pore indistinct, colpus opening broadly. Lupinus bb. Pore distinct, colpus constricted equatorially. Oxytropis

9.8 Per-reticulate.

a. Heterobrochate, no transverse colpus, with poroidal area. Salix

E. List of taxa, with English names

Taxon	English name
Abies lasiocarpa	Alpine Fir
Alnus crispa	Green Alder
Alnus incana	Grey Alder
Arctostaphylous rubra/alpina	Alpine Bearberry
Arctostaphylous uva-ursi	Common Bearberry, Kinnikinick
Artemisia	Wormwood, Sage
Betula glandulosa	Dwarf Birch
Betula nana	Arctic Birch
Betula papyrifera	White Birch
Botrychium	Moonwort
Celamagrostis canadensis	Reed-Bentgrass
Carex	Sedge
Cassiope tetragona	Arctic White Heather
Cerastium	Mouse-Ear Chickweed
Chenopodium	Goosefoot
Cornus canadensis	Bunchberry
Corylus	Hazel
Cruciferae	Mustard Family
Cyperaceae	Sedge Family
Cystopteris	Fragile Fern
Dryas	Mountain Avens
Dryopteris	Shield Fern
Empetrum nigrum	Curlewberry, Crowberry
Epilobium	Fireweed
Equisetum	Horsetail

Eriophorum	Cottongrass
Fragaria	Strawberry
Gramineae	Grass Family
Hedysarum mackenzii	Liquorice-root
Isoetes	Quillwort
Juncus	Bog-rush
Juniperus communis	Ground Juniper
Larix larcina	Tamarack, larch
Ledum	Labrador Tea
Linnaea borealis	Twinflower
Linum Lewisii	Flax
Lycopodium	Club-moss
Minuartia biflora	Sandwort
Myrica Gale	Sweet Gale
Myriophyllum	Water Milfoil
Oxycoccus microcarpus	Cranberry
Picea glauca	White Spruce, Cat Spruce
Picea mariana	Black Spruce
Pinus Banksiana	Jack Pine
Pinus contorta	Lodgepole Pine
Poa glauca	Bluegrass
Polemonium	Jacob's Ladder
Polygonium	Knotweed
Polystichum	Holly-fern
Potamogeton	Pondweed
Potentilla	Cinquefoil
Pyrola grandiflora	Wintergreen

Ranunculus	Crowfoot, Buttercup
Ribes triste	Wild Red Currant
Rosa acicularis	Wild Rose
Rubus acaulis	Raspberry
Rubus chamaemorus	Cloudberry
Rumex	Dock
Sagina	Pearlwort
Salicornia rubra	Glasswort
Salix	Willow
Selaginella sibirica	Spikemoss
Senecio congestus	Groundsel
Shepherdia canadensis	Soapberry
Silene	Campion
Typha latifolia	Cat-tail
Vaccinium uliginosum	Bilberry
Vaccinium uva-ursi	Mountain Cranberry
Viola	Violet