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PERMAFROST AND GROUND ICE INVESTIGATION MAYO, INTERIOR YUKON

Ottawa-Carleton Centre for
Geoscience Studies

Earth Physics Branch Open File Number 84-24
Dossier public de la Direction de la Physique du Globe No. 84-24

Department of Energy, Mines &
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Earth Physics Branch
Division of Gravity, Geothermics
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ABSTRACT
Stratigraphic, geomorphic and ground ice data is presented for the Mayo area in the interior Yukon. The report is preliminary in nature and meant to provide a geological framework for studies of the thermal and hydrological regime in the same area.

RÉSUMÉ
Des données stratigraphiques et des renseignements sur la glace dans le sol sont rassemblés pour la région de Mayo à l'intérieur du Yukon. Ce rapport de nature préliminaire a pour but d'établir un contexte géologique pour l'étude des régimes thermiques et hydrologiques de la région.

PERMAFROST AND GROUND ICE INVESTIGATIONS, MAYO, INTERIOR YUKON.

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Ottawa-Carleton Centre for
Geoscience Studies

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1. INTRODUCTION

1.1 Objectives

This report presents information relevant to an understanding of permafrost and ground ice conditions in central interior and northern Yukon. By providing stratigraphic, geomorphic and ice crystallographic data the report complements thermal and hydrological studies already in progress in the Mayo area by M. W. Smith (Department of Geography, Carleton University) and F. Michel (Department of Geology, Carleton University). The report is of a preliminary nature and a second report describing other localities in the northern Yukon is anticipated based upon an additional year of support.

The specific aim of permafrost studies in the Mayo area is to provide a mass balance for water in the top 2 m of permafrost at sites of differing thermal and hydrological characteristics. Several processes are thought capable of causing water movement in the surface layers of permafrost. These can be summarised as: (a) upward movement of moisture along the soil temperature gradient in winter, both within permafrost and the active layer (Parmuzina, 1978; Wright, 1981; Mackay, 1983), and

(b) downward movement of water in summer as the active layer thaws, and as water penetrates thermal contraction and/or dessication cracks (Kane, 1980; Mackay, 1983).

It follows that the water which contributes to the ice-rich zone at the base of the active layer and in the upper 1 to 2 m of permafrost originates from either the atmosphere (e.g., Cheng, 1982; Mackay, 1983) or from water moving upwards from and through permafrost lower in the soil profile (Parmuzina, 1978).

In order to measure and quantify the moisture changes that do occur, instrumentation has been installed to maintain the thermal and hydrological conditions of a number of closely related sites in an area of ice-rich silty sediments approximately 2.5 km south of the Mayo townsite. The area is geomorphologically distinctive, since it is characterised by (1) two large, active ground ice slumps (bimodal flows or retrogressive thaw-flow slides), which occur on the south bank of Stewart River and (2) numerous thaw lakes and depressions, some of which are currently experiencing rapid expansion and/or drainage. An account of the thermokarst activity is already available (Burn, 1982), and the field instrumentation which has been installed is described in other reports (e.g., Smith and Burn, 1983).

Within this context, the present report restricts itself to stratigraphic and ice petrographic studies at the Mayo site, in the belief that such information may assist in determining the origin(s) of the ground ice under investigation.

The usefulness of the permafrost stratigraphic approach, widely used in the unglaciated areas of central and eastern Siberia (Katasonov, 1975; Katasonov and Ivanov, 1973; Mackay et al., 1979, pp. 10-11; Sher and Kaplina, 1979) and Alaska (Péwé, 1975; 1977; Sellman and Brown, 1973), rests on the assumption that Quaternary sediment sequences are usually related to the diversity of glacial and post-glacial depositional environments. The pattern of permafrost aggradation in these sediments and, in particular, the nature of the ground ice bodies, is useful in deducing previous geomorphological conditions. In the western Canadian Arctic, stratigraphic studies of permafrost have been undertaken in parts of the Mackenzie Delta (Mackay, 1975; 1976; 1978) and Southern Banks Island (French et al., 1982; Harry, 1982; French and Harry, 1983), but little has yet been accomplished in the central interior Yukon.

Numerous ice fabric studies have been undertaken in the last thirty years, but only a few have analysed ground ice (e.g., Black, 1953; 1978; Gell, 1978a; 1978b; Pollard, 1983; Pollard and French, 1983). The aim of petrofabric analysis is to establish the dimensions, form and orientation of crystals in order to determine growth history. Techniques used to accomplish this aim include surface rubbing of ice (described by Bader, 1951), the partial melting of ice to obtain Tyndall figures (e.g., Péwé, 1978) and utilising the optical properties of ice to determine c-axis orientations (Langway, 1958; Shumskii, 1964, pp. 116-133). The latter is used in this report.

1.2 Work Schedule

The work schedule involved examination of ground ice bodies in other areas of the northern Yukon in addition to the Mayo locality, such as at Hunker Creek, Klondike District (e.g., French et al., 1983), and along the Dempster Highway (e.g., Pollard and French, 1983).

Fieldwork in the northern Yukon was undertaken between September 16 - 26, 1983 by H. M. French and between January 4 - 17th, 1984 by H. M. French and W. H. Pollard. On both occasions, C. Burn assisted for several days while conducting other observations. The objective of the first period of fieldwork was to obtain geomorphological familiarity with the Mayo locality prior to the onset of winter conditions, and to conduct reconnaissance observations upon the occurrence of ground ice exposures in Klondike District prior to sampling the following January. A number of organic and sediment samples were collected for potential ^{14}C and other analyses at a later date. In January ground ice samples were collected from both the Mayo locality and from a placer exposure of muck deposits on Mayes Claim, Hunker Creek, Klondike District. The ice samples, collected using ice axes and chain saw, were transported to Ottawa in refrigerated boxes. At the Mayo locality, detailed stratigraphic and cryotexture observations were undertaken but time and poor weather prevented similar observation at the Hunker Creek exposure.

During February 1984, the universal stage and cold room facilities at the Division of Building Research, National Research Council of Canada, Ottawa, were used to conduct petrofabric analyses. Priority was given to the Mayo samples and only cursory examination had been paid to the Hunker Creek ice samples before use of the cold room facilities by DBR personnel necessitated curtailment of the analysis.

Included within the work schedule was a literature review and air photo interpretation of the Mayo site, with the objective of assembling all known information on the Quaternary history of the region. In the following, the literature review is presented as Section 2, and the regional studies at Mayo as Sections 3 and 4. Work undertaken at Hunker Creek, Klondike District, is not presented in this report.

2. LITERATURE REVIEW: QUATERNARY HISTORY OF INTERIOR AND NORTHERN YUKON

A general outline of glacial, interglacial and interstadial conditions in the area is provided first, followed by discussion of conditions in various parts of the region during the various glacial and interglacial periods.

2.1 Outline of Events and Conditions

During the cold periods of the late Pleistocene, central and northern Yukon can be divided into three regions (Figure 1) (Hughes, 1972). First, there was an area covered by the Laurentide and Cordilleran ice sheets. Cordilleran ice moved north and westwards from the Cassiar and Selwyn Mountains (Bostock, 1966) while Laurentide ice flowed down the Mackenzie Valley and impinged on the northern Yukon coastal plain (Rampton, 1982). Second, alpine glaciers occupied high valleys in the Ogilvie, Wernecke and McArthur Mountains west of the main Cordilleran ice body(ies). Third, lowland areas north and west of the Cordilleran ice lobes remained ice-free (Hughes et al., 1981a; 1983), thereby comprising one of the largest unglaciated areas in Canada

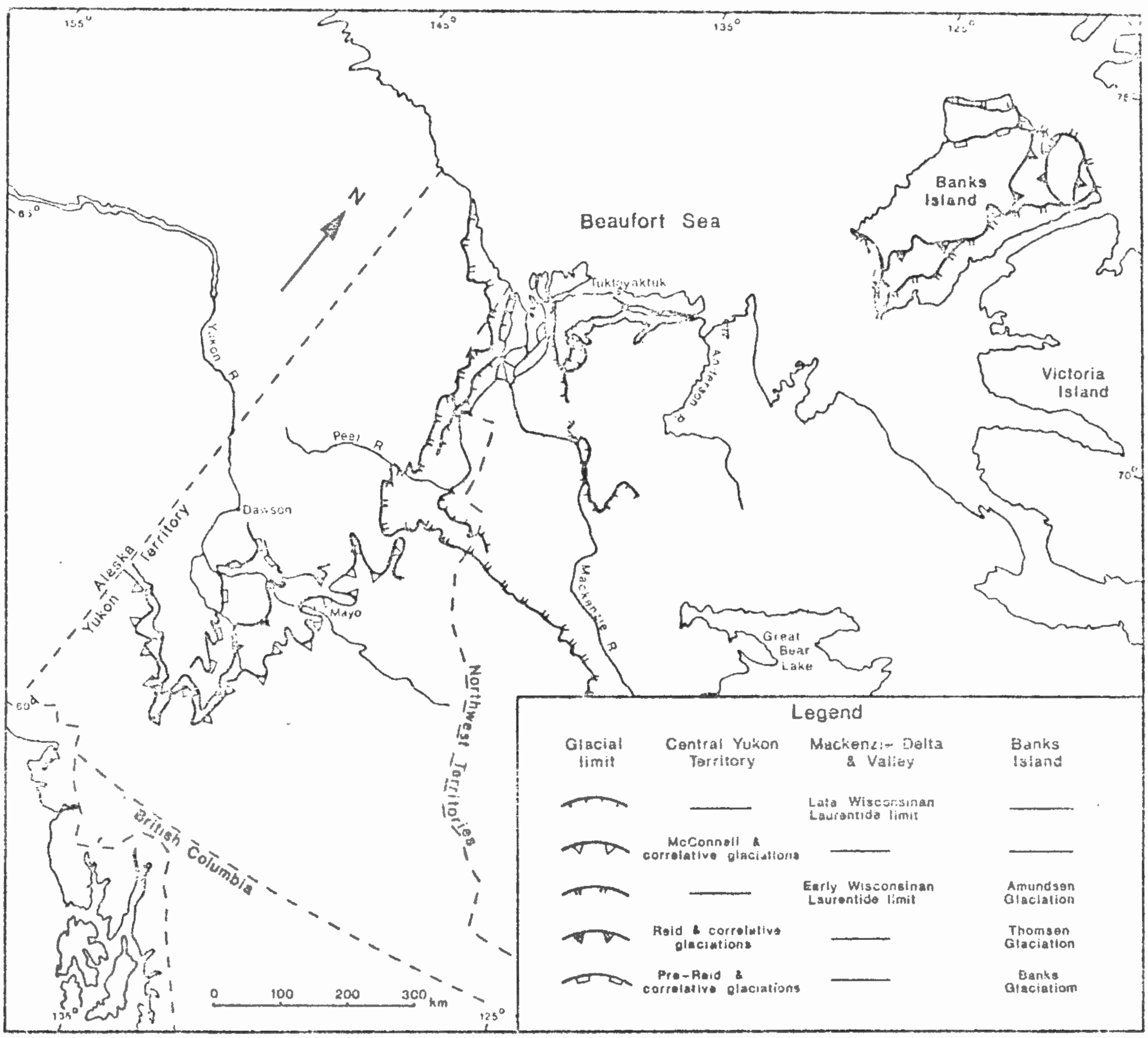

Figure 1. Late Pleistocene glacial limits in the central and northern Yukon and adjacent regions (after Hughes, 1972; Hughes et al., 1981; 1983; Vincent, 1983).

and the eastern portion of the Beringia refugia (Hopkins, 1967; Hughes et al., 1981b; Harington, 1975).

Ice contact deposits and both terminal and recessional moraines indicate the presence of more than one Quaternary glacial advance into the region. Four advances of Cordilleran ice were suggested by Bostock (1966), named, from oldest to youngest, Klaza, Nansen, Reid and McConnell glaciations. It is perhaps fortuitous that the successive advances were progressively less extensive, thereby allowing the terminal limits to be preserved. Based upon surficial geology mapping in part of the southern Ogilvie Mountains, Vernon and Hughes (1966) identified three Pleistocene advances of valley glaciers. Because of their indeterminate age, they were termed 'oldest', 'intermediate' and 'youngest'. It is also clear that several peaks within the margins of valley glaciation by Cordilleran ice also supported cirque glaciers (e.g., McArthur Group, Gustavus Range; see Hughes, 1983).

The dispersal of the Pleistocene ice sheets into the lowlands of central Yukon was controlled to a great degree by topography. Most significant was the confinement of Laurentide ice to the Mackenzie Delta and present Yukon coastal plain by the Richardson and Barn Mountains (Hughes, 1972; Rampton, 1982). In central Yukon in the vicinity of Stewart Crossing, evidence presented by Cairnes (1916), Vernon and Hughes (1966) and Hughes (1983) indicates that at least McConnell, and possibly Reid ice, was confined to valley bottoms as it moved west and northwest. Southeast of Mayo, Bostock (1966) indicates that parts of the Talbot Plateau around Twin Buttes were above the Reid ice limit. Likewise, it is probable that the movement of glaciers north of the Ogilvie and Wernecke Mountains was also restricted to valleys.

The major drainage networks appear to have suffered few dislocations during mid/late Pleistocene times, although the Yukon River in the Klondike area may have flowed in the reverse direction in the late Tertiary/early Quaternary (Hughes and van Everdingen, 1978; Tempelman-Kluit, 1980). One exception is the Porcupine River system which drains Eagle Plain. This system initially flowed eastwards via McDougall Pass into the Mackenzie River system. During at least two periods in the mid/late Quaternary, while McDougall Pass was blocked by Laurentide ice, drainage was to the west. Glacial meltwaters ponded in the Old Crow, Porcupine and Bluefish basins (Jopling et al., 1981; Lichti-Federovich, 1973; 1974). Overflow from these lakes cut a canyon through Cretaceous age limestones known as the Ramparts, a route now followed by the Porcupine River as it traverses the Old Crow Range.

The last advance of the Laurentide, Cordilleran and valley glaciers occurred during 'classical' late Wisconsin time (Hughes et al., 1983). However, it is unclear when the Reid, 'intermediate' valley glaciation, and early Wisconsin Laurentide glaciations occurred, and whether or not they were contemporaneous. Similarly, the age of the lower 'glacial' lake sediments of the Old Crow Basin (Jopling et al., 1981) is unknown.

There are several reasons why this uncertainty exists. The first is that many events in the Quaternary history of the central and northern Yukon occurred at a time beyond the limit of resolution for ^{14}C dating. Moreover, much ^{14}C dating has been of wood fragments, for which there is no guarantee of an autochronous origin. Second, since large areas have experienced extended periglacial (i.e., cold, non-glacial) conditions at different times within the last 200,000 years, cryoturbation and solifluction has disturbed original sedimentary sequences. In the Klondike valleys for example, thick 'muck' deposits (Hughes and van Everdingen, 1978; Naldrett, 1982; French et al., 1983) are testimony to the efficacy of mass wasting processes in the unglaciated Yukon at various times during the Quaternary (Figure 2) and Pleistocene tors are widespread on the uplands. Third, the interpretation of palaeoenvironmental conditions in unglaciated areas from pollen assemblages has been hampered by the lack of data about current assemblages in northern regions (Lichti-Federovich, 1974) although the abundant macrofauna of the region is well documented (e.g., Harington and Clulow, 1973; Harington, 1978).

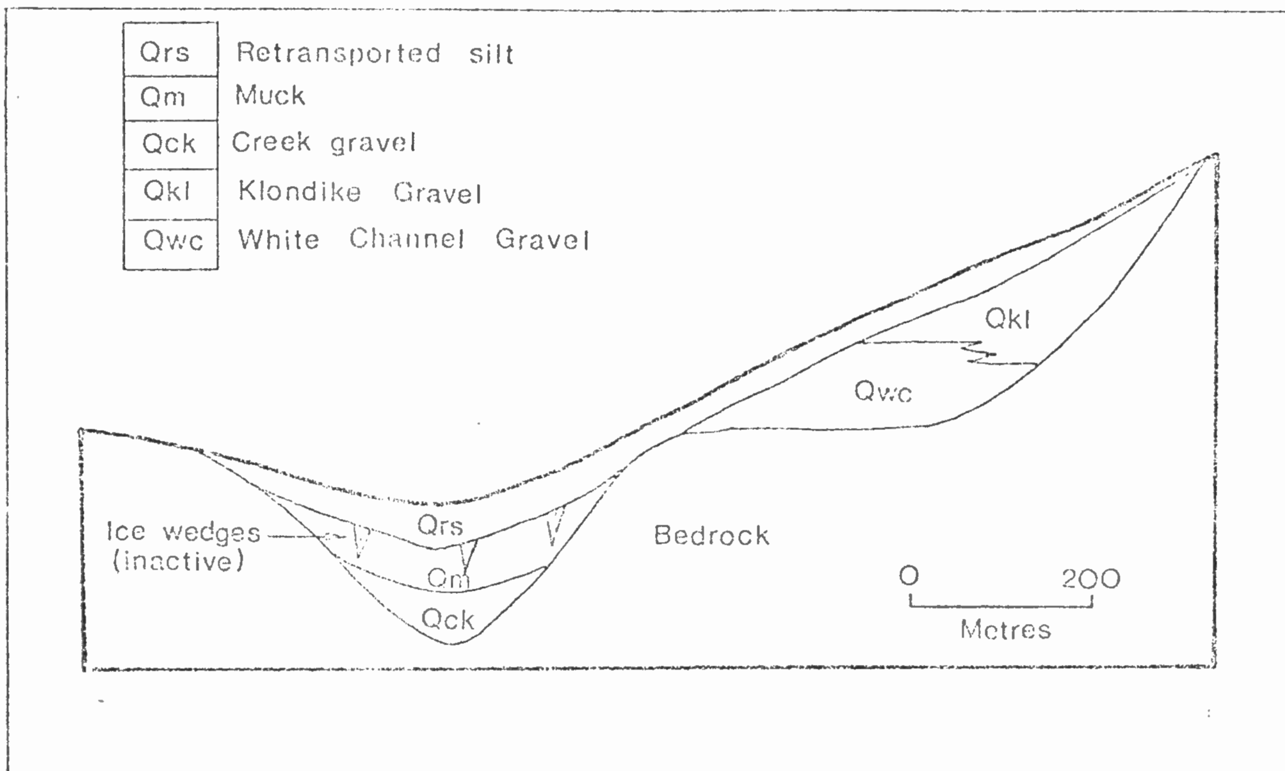

Figure 2. Schematic cross-section through valley deposits in unglaciated Klondike District (after Naldrett, 1982).

2.2 Late Wisconsin Glacial Advances

During the last glacial advance, the extensive southern part of the Laurentide ice sheet reached its southern limit about 20,000 to 25,000 years ago. By contrast, the more restricted northern and eastern part reached its maximum extent only about 10,000 years ago (Dyke, Dredge and Vincent, 1983).

Rampton (1982) concluded that the maximum extent of Pleistocene glaciation along the Yukon Coastal Plain, which he termed the 'Buckland Glaciation', occurred in 'early' Wisconsin times. At that time, ice advanced to within 15 km of Herschel Island and, upon retreat, left recessional moraines along the northern slopes of Buckland Mountains. A major still-stand occurred during the retreat of Buckland ice since an ice-thrust moraine extends from King Point to Kay Point. Rampton (1982) proposed that this stage be called the 'Sabine Phase'.

The precise age of the Buckland Glaciation has not been defined. Most material directly overlying Buckland till postdates 11,000 years B.P. On the other hand, plant fragments in a pond sequence on Buckland till near Stokes Point have yielded a date of 22,400 years B.P. (Rampton, 1982) indicating that the Buckland Glaciation clearly predates the late Wisconsin maximum, estimated by Fyles et al. (1972) to have been approximately 18,000 - 13,000 years

B.P. in the Mackenzie Delta.

Hughes et al. (1981a) have established a maximum age for Laurentide ice in the Bonnet Plume Basin at 36,900 years B.P. This date was obtained for wood found in alluvial sediments underlying till in a section exposed along Hungry Creek. This glacial advance is referred to as the 'Hungry Creek Glaciation' (Hughes et al., 1981a; Hughes et al., 1983) and may be contemporary with the Buckland Glaciation. If this assumption were correct, the Buckland maximum would have been reached in the period 36,900 - 22,400 years B.P. In addition, Hughes (1969; 1972) indicates that the upper lake sediments of the Old Crow Basin were deposited in lakes fed by glacial meltwaters which flowed into this basin from the Bonnet Plume Basin via the Eagle River discharge channel. The maximum age for these sediments of 32,000 years B.P., suggested by Jopling et al. (1981, Figure 3), supports the proposed time frame for the Hungry Creek Glaciation.

However, several ^{14}C ages of approximately 38,000 years B.P. have been obtained from silt overlying till of the late Wisconsin ice limit from the Rat River area, east of McDougall Pass (Hughes, 1972). If the late Wisconsin maximum is to be equated with the Hungry Creek Glaciation then the date of 36,900 years obtained from the Bonnet Plume Basin indicates considerable local variation, both spatially and temporally, in the maximum ice limits.

It is quite conceivable that the Buckland and other Wisconsin maxima were not contemporaneous with the Hungry Creek Glaciation, since topographic effects may have been important in controlling ice flow and/or recession patterns.

Of the four Cordilleran advances identified by Bostock (1966), the first two (Nansen and Klaza) are regarded as ancient, definitely pre-Wisconsin in age (Hughes et al., 1983). The Reid Glaciation is also thought to be pre-Wisconsin in age (Hughes, 1969; Foscoulos et al., 1977) since wood at the base of fill in a channel cut in Reid outwash has given an age of 42,900 years B.P. In addition, wood from beneath McConnell till in a section on the north bank of Stewart River, 2 km downstream from Mayo, has given an age of 46,580 years B.P. Foscoulos et al. (1977) suggest that these dates imply that the Reid-McConnell 'interglacial' existed before 47,000 years B.P. However, both dates are from allochthonous sediments and the map of Bostock (1966) indicates the territory northwest of Mayo was not overriden by Reid ice. Therefore, the wood may have been transported into the lower valleys by mass wasting processes.

O. L. Hughes (Hughes et al., 1983) believes that the pattern of Cordilleran ice expansion of central Yukon followed that of south Yukon more closely than that of the Laurentide ice sheet. In south Yukon for example, Denton and Stuiver (1967) identified glaciation in the northeastern St. Elias Mountains from 49,000 to 37,700 years

B.P., which they term the Icefield Glaciation. This may be contemporaneous to the Reid Glaciation since they also identify a subsequent Cordilleran advance (the Kluane Glaciation) from 30,100 to 12,500 years B.P. which would correlate with the McConnell Glaciation.

Usually, the McConnell Glaciation is assigned to the 'classical' Wisconsin maxima (22,000 to 18,000 years B.P.) (e.g., Bostock, 1966; Hughes, 1969; Foscoulos et al., 1977; Hughes et al., 1983). Hughes et al. (1983) suggest a date of 24,000 years B.P. for the onset of glaciation in the Liard Plain of southeast Yukon while Vernon and Hughes (1966) provide a minimum date for the end of glaciation of 12,900 years B.P., from marl deposits at Hart Lake, north of Mayo.

The chronology of the alpine glaciations is poorly defined. Vernon and Hughes (1966) indicate that the last glaciation is almost certainly correlated with the McConnell advance, but give no indication of the age of the 'intermediate' or 'early' glaciations. It is possible that the 'intermediate' glaciation corresponds to the Reid Glaciation.

2.3 Conditions in Non-Glaciated Areas

Large areas of central and northern Yukon were never glaciated during the Pleistocene. Instead, these areas experienced extended periods of periglacial, or cold, non-glacial conditions. Even in areas where ice was present on some occasions, there were long periods of non-glacial conditions.

There is evidence to suggest that conditions in northwest North America were considerably warmer during the Sangamon interglacial than today. For example, the pollen spectra of pre-Wisconsin deposits at several sites in the region includes pine (Ritchie, 1980). At the same time, Rampton (1982) reports that pre-Buckland sediments on the northern Yukon coastal plain include arboreal pollen indicating that the treeline was north of the Yukon coast during part of the Sangamon.

Weathering profiles in soils of pre-Reid deposits south of Dawson City (Foscoulos et al., 1977; Rutter et al., 1978) suggest two distinct climates for the pre-Reid period. The first, warm and subhumid with grassland/shrub vegetation, was followed by a temperate and humid climate immediately preceding the climatic deterioration which accompanied the onset of Reid Glaciation. Pedogenic evidence for these changes is the presence of a red Luvisol B horizon, up to 2.0 m thick.

During the Wisconsin and earlier glacial advances, a cold and dry climate is inferred for the ice-free parts of the northern and central Yukon. The presence of ice and sand wedge casts is central to this interpretation.

The most convincing evidence is provided by Rampton (1982) who describes an ice-wedge cast in pre-Buckland material, while Jopling et al. (1981, e.g., Figures 7 and 11) note several in their stratigraphic studies at Old Crow, as does Hughes (personal communications, 1982 and 1983) from the Klondike and McQuesten areas. However, the exact time, and the environmental conditions associated with the degradation of the ice wedges to cast forms, are not fully understood.

Palynological studies from the Old Crow Basin (Lichti-Federovich, 1973; 1974) indicate three major stages between the onset of early Wisconsin glaciation to the east and south, and the deposition of sediments in Glacial Lake Kutchin (Jopling et al., 1981). A tundra vegetation assemblage was followed by one more similar to the present, consisting of spruce, birch and herbaceous species. Then, tundra conditions re-established themselves prior to the formation of Lake Kutchin.

Although Jopling et al. (1981) assign an Illinoian age (i.e., pre-400,000 years B.P.) to the lower lake sediments in the Old Crow Basin (the sediments of 'Glacial Lake Old Crow') Hughes et al. (1981) suggest that these sediments are not glacio-lacustrine in origin. The possibility exists that these sediments are more recent than suggested by Jopling et al. (1981); for example, Lichti-Federovich (1974) reports an age of 41,300 years B.P. for allochthonous wood in sediments along the Porcupine River which are close in elevation to the lower lake deposits.

The Hungry Creek type section (Hughes et al., 1981) provides a record of mid-Wisconsin environmental conditions which contains considerable palynological, entomological and faunal information. Unit 2B, lying below the Hungry Creek till and above laminated silt and clay, of possible glacio-lacustrine origin, yielded a pollen spectra which indicated that a climate similar to the present may have existed immediately prior to the Hungry Creek Glaciation.

Vertebrate fossils have been discovered frequently by placer miners working 'muck' deposits in central Yukon (e.g., Harington and Clulow, 1973), and at the Hungry Creek type section (Hughes et al., 1981). For example, the Hungry Creek deposits contain bones of the oldest Yukon specimen of the now extinct Yukon wild ass, and several specimens of ground squirrel and collared lemming. Similar assemblages occur on the Klondike Plateau near Dawson, in Gold Run Creek (Harington and Clulow, 1973) and Hunker Creek (Hughes et al., 1978; Naldrett, 1982; French et al., 1983). Eight of the thirteen species identified by Harington and Clulow (1973) are now extinct. Their presence suggests that a cool, steppe-like grassland existed in much of central and northern Yukon during the late Wisconsin. Unfortunately, because these faunal remains have been transported to their present positions by mass wasting and fluvial/colluvial processes, there is little stratigraphic control. Mammoth bones have been dated at 32,250 ± 1,750 years B.P. while a bison bone has been dated at 22,200 ± 1,400 years B.P.

2.4 Late and Post-Wisconsin Environmental Changes

The chronology of deglaciation in central and northern Yukon is not well described, especially for the period between 25,000 - 20,000 years B.P. (the maximum advance of late Wisconsin ice) and 13,000 years B.P. (the time of formation of marl deposits in Hart Lake, near Mayo). Several recent maps by Hughes (1983) summarise the Quaternary geology and ice limits for the Mayo/McQuesten regions, and Vernon and Hughes (1966) have mapped surficial deposits in the Nash Creek area further north.

Vernon and Hughes (1966) concluded that the retreat of McConnell ice was rapid and continuous, while the alpine glaciers were more erratic. Several large pro-glacial lakes formed in valleys in front of the receding ice margins. Lacustrine silts and clays are found in the South McQuesten, Keno-Ladue and Stewart River valleys (Hughes, 1983). For example, Green (1971) describes the lacustrine deposits of the Stewart Valley as being "up to several hundred feet thick". Since the shorelines of these lakes are relatively subdued, Vernon and Hughes (1966) suggest that the lakes were short-lived.

According to Hughes et al. (1981) Laurentide ice in the Mackenzie Valley and Delta retreated from its maximum by 16,000 years B.P. After a brief advance associated with the Sitidgi Lake moraine (Hughes et al., 1983), it left the area near Fort Good Hope by 11,530 years B.P. at the latest (Mackay and Matthews, 1973). Glacial Lake Kutchin drained by 12,300 years B.P. (Jopling et al., 1981).

The amelioration in climate that occurred about 12,000 years B.P. led to widespread thermokarst in north-western North America. Rampton (1973) concluded that the period 10,000 to 9,000 B.P. was one of active thermokarst in the Mackenzie Delta and French and Harry (1983) conclude that many of the thaw lakes on Southern Banks Island formed between 8,000 and 9,000 years B.P. Indeed, Delorme et al. (1977), using ostracod assemblages, suggest that during the period 14,410 to 6,820 B.P., mean annual air temperatures rose to between 0.8 and 1.5°C. Although there is a consensus that this period was warmer than the preceding late Wisconsin period, Mackay (1978) points out that such a large increase in air temperature would have led to the thaw and disappearance of massive icy beds exposed along the Pleistocene Mackenzie Delta coast.

Warming also coincided with the invasion of tree species such as poplar from more southerly parts of Canada, and culminated in the hypsithermal period, the warmest since glaciation, from 8,000 to 4,000 years B.P. Ritchie and Hare (1971) suggest, from an examination of pollen sequences in the Tuktoyaktuk Peninsula, that temperatures were 5°C warmer than at present. However, during the hypsithermal a lower sea level would have led to the location of the pollen sites being about 50 km inland. Hence, 3 or 4°C of the suggested figure might be accounted for by increased continentality.

After the mid-Holocene hypsithermal, climatic conditions deteriorated to those of today (a decline of 2°C, Hopkins et al., 1981). Rampton (1982) provides evidence that supports this hypothesis, as do Pissart and French (1976) who conclude that many of the pingos on Banks Island are no longer forming. Finally, the Old Crow pollen record of Lichti-Federovich (1973) indicates a similar transition for the area north of the Ogilvie Mountains.

Within this general framework of late Quaternary/Holocene climatic change, any thaw lake topography associated with ice-rich glacio-lacustrine sediments, as found in the Stewart Valley near Mayo, can be easily accommodated. In all probability, thermokarst activity has been ongoing, at varying levels, throughout much of the Holocene.

3. THE MAYO AREA, CENTRAL YUKON

A brief description of the regional geomorphology and Quaternary geology of the Mayo area is presented first, prior to a more detailed description of the thermokarst terrain and permafrost conditions at the study site 1.5 km south of Mayo.

3.1 Quaternary Geology and Geomorphology

The difficulty of access to extensive areas without helicopter support, the hilly terrain and the lack of roads seriously hinders detailed regional geomorphological mapping. The recent publication of four surficial geology maps compiled by O. L. Hughes of the Geological Survey of Canada at a scale of 1:100,000 (Hughes, 1982) summarises much of what is currently known about the area.

The Mayo area lies within the limit of McConnell ice which moved westwards from its source area in the Selwyn Mountains. Toward the western limits of the ice sheet, as it thinned, flow became increasingly concentrated in valleys, leaving interfluves and uplands free of ice. Figure 3 summarises the generalised ice flow directions and McConnell limits for the area.

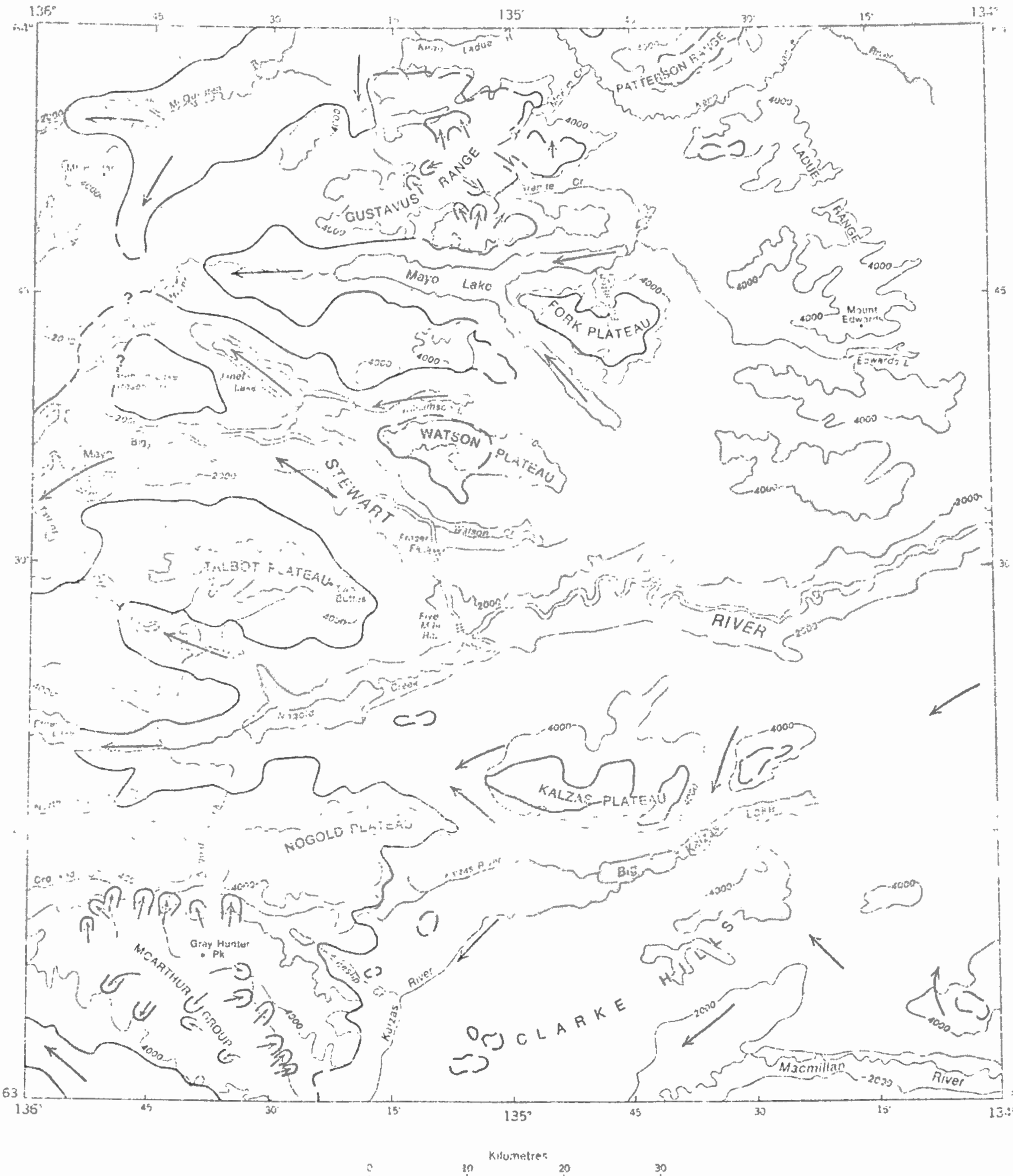

Figure 3. Generalised McConnell limit and ice flow directions (after Hughes, 1982).

According to Hughes (1982), the limits of McConnell glaciation can be inferred from the presence of moraine ridges flanking the middle and upper valleyside walls, and the direction of flow by the existence of drumlinoid ridges and glacial flutings at lower elevations. In places, the earlier Reid glacial limit is inferred from the presence of moraine ridges at higher elevations, especially on Scheelite Dome to the northwest of Mayo. A number of meltwater channels are also mapped, both above and below the McConnell limit, and are clearly associated with different stages of ice retreat.

3.2 Stewart Valley Near Mayo

The surficial geology and geomorphology of the Mayo area is illustrated in Figure 4. Summary notes by Hughes (1982) state that: "During retreat of the Cordilleran Ice Sheet following the McConnell maximum, parts of several major valleys were occupied by glacial lakes in which thick glaciolacustrine sediments were deposited. . . . A glacial lake in Stewart Valley at Mayo was impounded behind massive morainic and glaciofluvial deposits that extend 15 kms downstream. The lake may have extended upstream to Fraser Falls and thence up Watson Creek. If so, the alluvial plain (Ap) and thermokarst alluvial terraces (Atk) on the valley floor from Big Island above Mayo to Fraser Falls

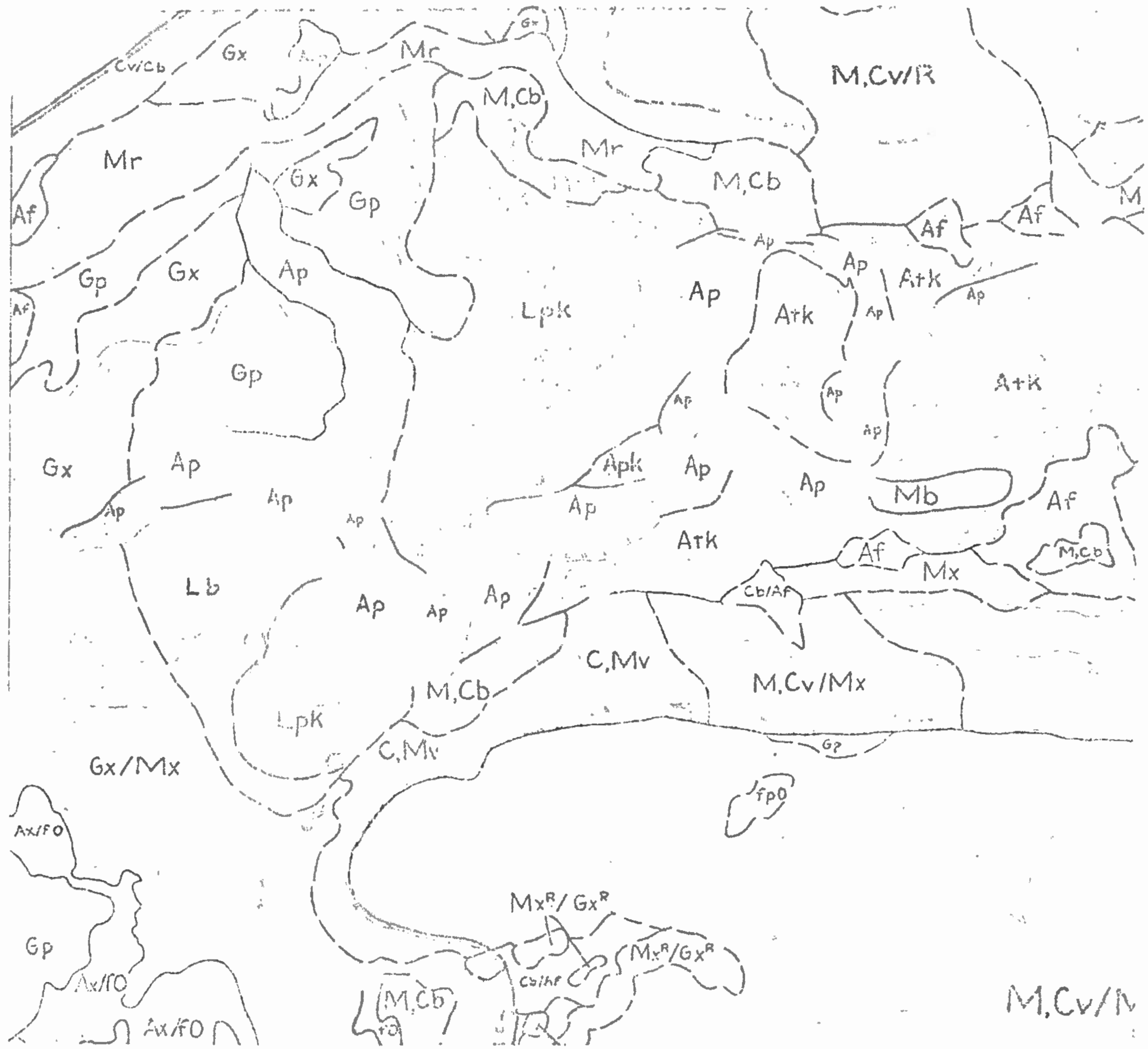

Figure 4. Surficial geology and geomorphology of the Mayo area.

may overlie glaciolacustrine sediments. Remnants of glaciolacustrine deposits extend up Stewart River. A glacial lake . . . may have been impounded by drift that filled the reach between Five Mile Rapids and Fraser Falls."

The events postulated by Hughes (1982) for Stewart Valley were almost certainly duplicated in adjacent valleys. For example, in Nogold Creek east of Ethel Lake, the pattern of glaciolacustrine deposits, thermokarst forms and moraine ridges, clearly supports a similar interpretation.

3.3 Upland Conditions Near Mayo

During both Reid and McConnell glaciations, the ice was restricted to the valleys in the Mayo area, and adjacent upland terrain was exposed to cold, subaerial (i.e., periglacial) conditions. Relatively little is known of these nunatek environments which may have been either tundra or rock desert in appearance.

During McConnell glaciation, land above 3,200 - 3,500' a.s.l. was probably ice-free; during the earlier Reid glaciation, land above 4,000' a.s.l. was probably ice-free. Typical of upland areas are (a) Mount Haldane (Figure 5) and (b) the Gustavson Range north of Mayo Lake (Figure 6). Both are briefly described to illustrate the nature of periglacial conditions during the periods of ice advance in this part of the interior Yukon in middle

Figure 5. Air photograph of area adjacent to Mount Haldane, north of Mayo, showing cryoplanation terraces (after Hughes, 1982) (N.A.P.L. A-19980-47).

Figure 6. Air photograph of part of the Gustavson Range, north of Mayo Lake showing cryoplanation terraces, cirques and probable extent of cirque glaciation (after Hughes, 1982) (N.A.P.L. A-20687-110).

Major cirques

End moraines

Probable ice movement direction

and late Pleistocene times. Mount Haldane is a smooth rounded bedrock massif rising to over 6,000' a.s.l. Tree-line occurs at approximately 4,700' a.s.l. The summit is covered with a veneer of glacial and colluvial material from earlier pre-Reid glaciations. Large flat bedrock surfaces are mapped by Hughes (1982) as being cryoplanation terraces. No glacial erratics are found on these surfaces, hence they appear to be good presumptive evidence that the surfaces lie above the all-time limit for glaciation.

The Gustavson Range rises to over 6,000' a.s.l. and apparently supported a number of small cirque glaciers and/or glacierettes, to judge from the highly glaciated nature of the higher elevations. Cryoplanation terraces are also mapped by Hughes (1982). Extensive cirque glaciation also characterises the McArthur Range to the south (see Hughes, 1982). However, on the Talbot Plateau, immediately south of Mayo, with an elevation of approximately 4,000' a.s.l., neither cirques nor cryoplanation terraces are evident.

On the basis of their relatively sharp morphology, the cirques in the Gustavson and McArthur Ranges were probably occupied by ice during the most recent (i.e., McConnell) glaciation, when the upper limit of valley glaciation was approximately 3,200 - 3,500' a.s.l. Using the lower lim t of cirque elevations as the approximate firn line, an ice-free periglacial zone would have existed between 3,500' and 4,500' a.s.l. in elevation. Cryoplanation terraces would have preferentially developed in the upper part of this zone, since it is generally assumed that intense frost action is responsible for their formation (Reger and Péwé, 1976). Support for this is provided by Hughes (1982) who states that, the lowest local occurrence of cryoplanation terraces on Mount Haldane is 4,200' a.s.l. This may be a critical elevation in the Mayo area. For example, on the adjacent Talbot Plateau, having a maximum elevation of 4,000' a.s.l., neither cirques nor cryoplanation terraces occur. This may mean that the summit elevations of Talbot Plateau were insufficient for the intense frost action conditions thought necessary for cryoplanation.

3.4 Timing of Events

The chronology of events in the Mayo area is only known in general terms. A ^{14}C date from wood in volcanic ash overlying Reid sediments on Stewart River (60° 30.2'N; 137° 16'W) has provided an age of >42,000 years B.P. (GSC-524, Lowdon and Blake, 1968) indicating that the maximum of the Reid advance is at least that age, or older. Several ^{14}C dates may relate to the McConnell glaciation. In the Mayo area, two dates (GSC-180: >35,000 years; GSC-331: >46,580 years) from the base of McConnell till and subsequent silt respectively are inconclusive. Hughes (1982) concludes that the time of the McConnell ice advance was probably comparable to that for other localities in the Yukon; after 30,100 years B.P. in southwestern Yukon and after 24,000 years B.P. in southeastern Yukon.

3.5 Thermokarst Topography and Ground Ice Conditions

The glaciolacustrine sediments south of Mayo are ice-rich and characterised by two rapidly degrading retrogressive thaw flow slides, commonly termed ground ice slumps, and numerous thermokarst lakes and depressions (Figure 7). Sequential air photo coverage since 1950 enabled Burn (1982) to determine the rate of headwall retreat of the slumps, and the rate of expansion of the thermokarst basins. Air photo interpretation indicates the presence of an open system pingo growing in one of the draining lake complexes (see Figure 7).

The largest exposure of ice-rich permafrost is in the headscarps of the two retrogressive thaw flow slides. The larger (A in Figure 7) was subject to detailed stratigraphic study and ground ice analysis.

3.5.1 Stratigraphy

A photomosaic of the complete headwall, as seen in January 1984, is illustrated in Figure 8. The entire face is composed of blue-grey silt and silty clay.

Figure 7. Air photo A 22201-9 showing thermokarst terrain developed in ice-rich glaciolacustrine sediments near Mayo, Yukon.

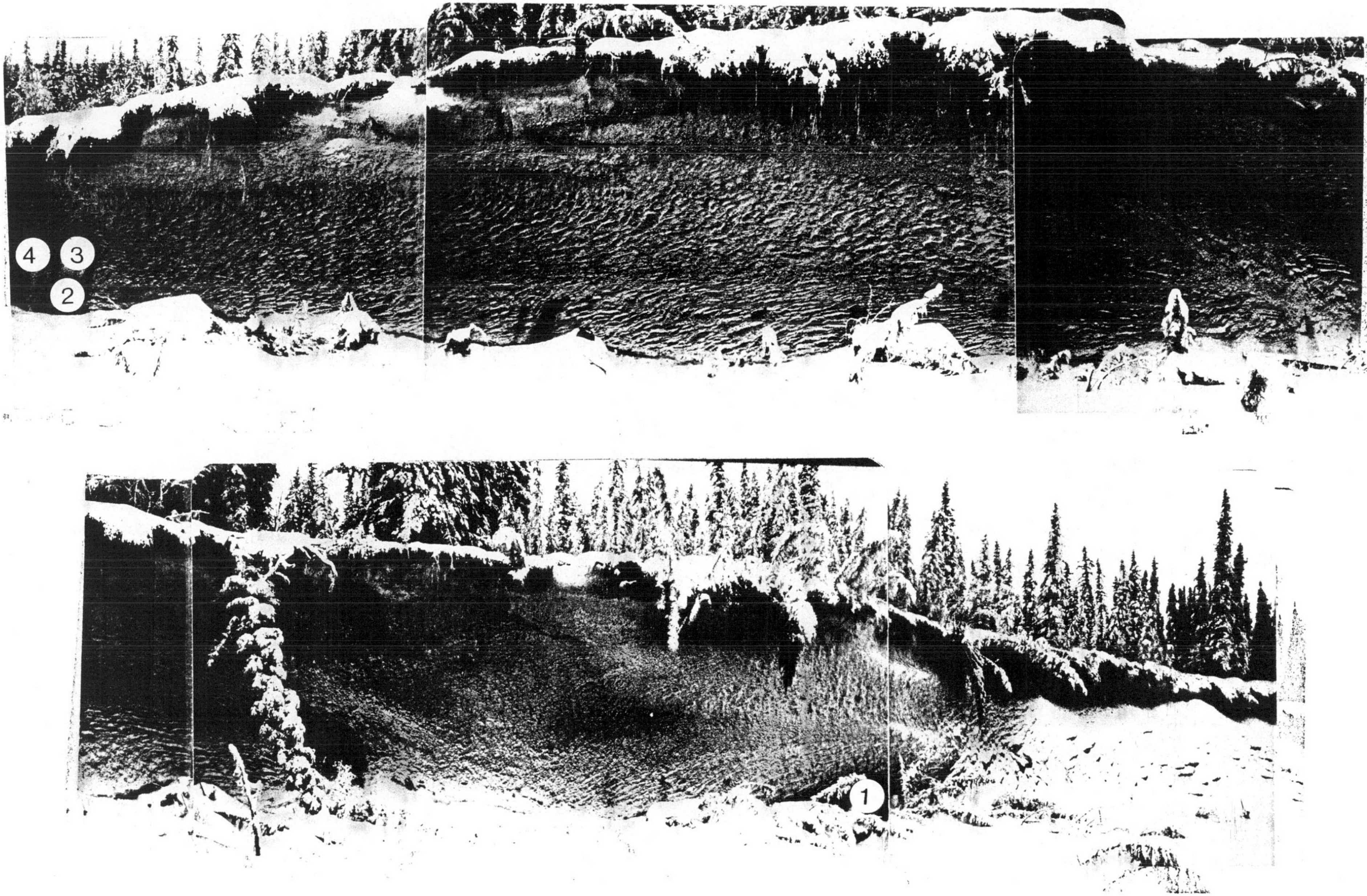

Figure 8. Photomosaic of headwall of retrogressive thaw flow slide near Mayo, in January 1984. Note location of Ice Samples One, Two, Three and Four.

Random determinations of ice contents indicated that excessive ice values of between 10 - 70% characterised the majority of the exposure.

From a stratigraphic viewpoint, two important aspects of the headwall exposure are (a) the presence of organic beds and (b) the highly variable ground ice texture (cryotexture) observed at various parts of the exposure.

The buried organic layers are between 5 - 30 cm thick. They contain an assemblage of non-marine mollusc and pelecypod shells, fragments of willow and birch, together with occasional larger spruce trunks up to 15 cm in diameter. The organic beds can be traced laterally for several metres, especially throughout the southern part of the headwall exposure. Since they separate sediments with distinct and different ground ice cryotextures (see below), they are interpreted as thaw unconformities probably related to the former existence of thermokarst lakes. The latter have been subsequently infilled by reworked, rethawed and subsequently refrozen silt and silty clay. These sediments would have been made available by the initial expansion of the thermokarst lake, only to refreeze again following abandonment of the lake basin.

Relatively few studies have examined the permafrost and stratigraphic conditions associated with thermokarst lake basins. A number of generalised models of thaw lake evolution and drainage are available (e.g., Billings and

Peterson, 1980; Britton, 1966; Everett, 1981; Tedrow, 1979) consisting of sequential stages of initiation, expansion and drainage often associated with migration across the tundra.

Recently, D. G. Harry (e.g., Harry, 1982; Harry and French, 1983; French and Harry, 1983) has concluded that gradual infilling following catastrophic drainage is the dominant mode of evolution for thaw lakes on Southern Banks Island. Typically the sub-basin stratigraphy is (a) 'compressed' and (b) the sediments possess lower ice contents than the undisturbed frozen sediment on either side of the basin.

Material collected from the organic layers has yielded ^{14}C dates of 8,560 ± 130 years B.P. (BGS-841) and 8,520 ± 120 years B.P. (BGS-843) (Burn, 1982, and personal communication, September 1983). In all probability, a major period of thaw lake formation occurred during the early part of the post-glacial climatic optimum, between approximately 8,000 - 5,000 years B.P. in the Mayo area. Given the overwhelming evidence in the western Canadian Arctic for a progressive deterioration of climate following the post-glacial climatic optimum to that of today (e.g., Rampton, 1974; Pissart and French, 1976), any subsequent periods of thaw lake formation would have been initiated by local factors rather than regional climatic changes. Since ice wedges on the Mackenzie Delta coast indicate renewed growth in the last 30 years (Mackay, 1974), the apparent recent activity of thermokarst at Mayo (Burns, 1982) is puzzling. Possible site-specific trigger mechanisms include forest fire, man's activities and rapid bank erosion of Stewart River at this point.

3.5.2 Cryotextures

The stratigraphic interpretation is complicated by the presence of a number of distinct ground ice cryotextures and structures observable in the headwall. During fieldwork in January 1984, segregated ice lenses protruded from the headwall by approximately 5 - 10 cm from the enclosing sediments. Under the cold winter conditions experienced at Mayo, the fine-grained silts backweather at a faster rate than the sublimation of the ice, resulting in a scalloped appearance of the thaw face. The resulting cryotextures, commonly mentioned in the Soviet literature (e.g., Kudryatsev, 1978, pp. 301-323) are rarely described and little understood in North America. During the course of fieldwork in January 1984, the opportunity was taken to examine the frozen headwall in safety, and to qualitatively describe the cryotextures. At least three different cryotextures were recognized. To avoid genetic implications, they are described below as A, B and C.

Cryotexture A (Figure 9):--This was a crude lattice consisted of irregular, broken lenses of ice branching out and thinning at the ends, commonly 30 - 80 cm long

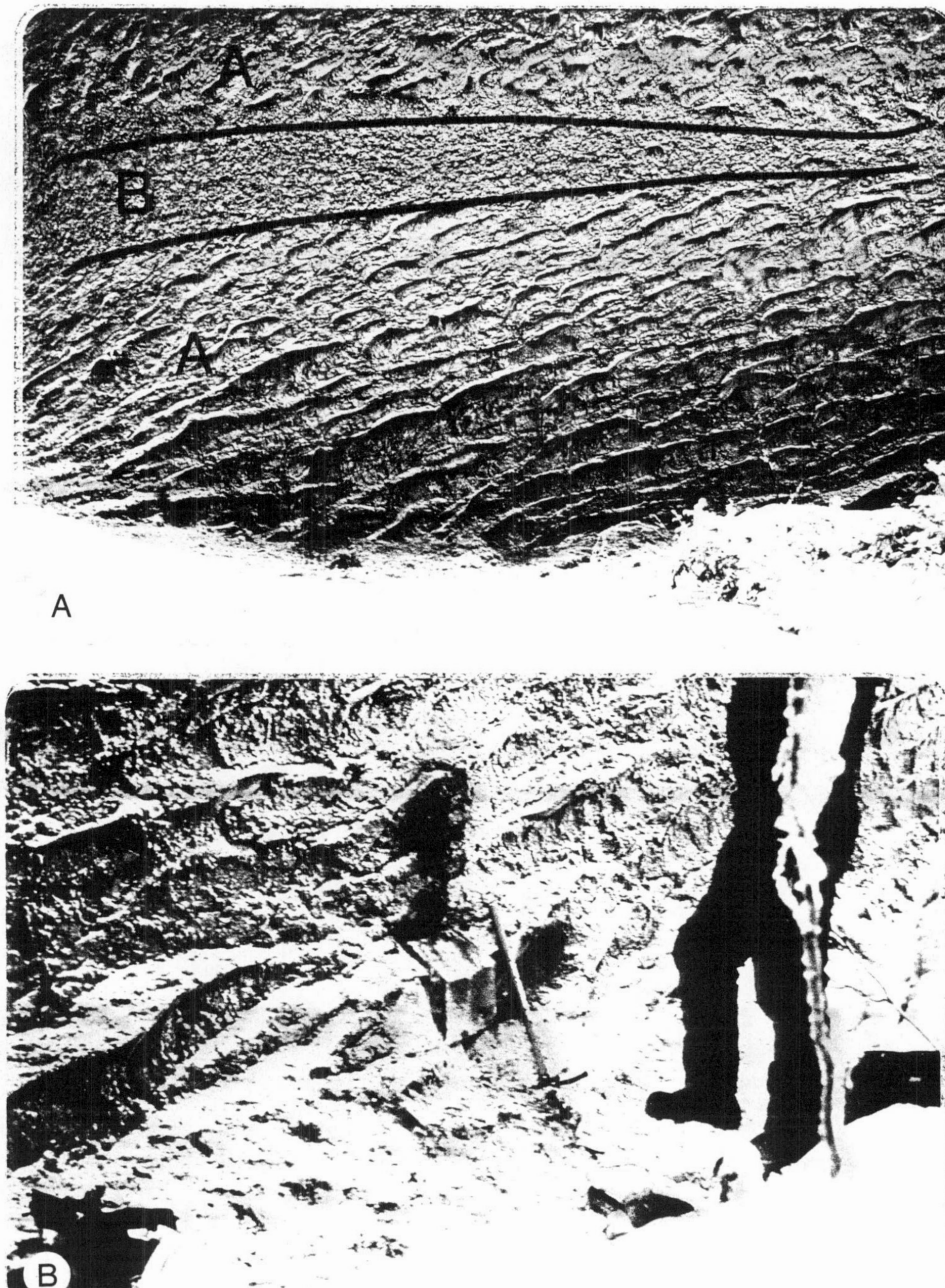

Figure 9. Cryotextures observed in headwall of retrogressive thaw flow slide, as seen in January, 1984. a) Cryotexture A and B with interfingering of cryotexture. b) Ice Sample One in place prior to extraction, January 1984.

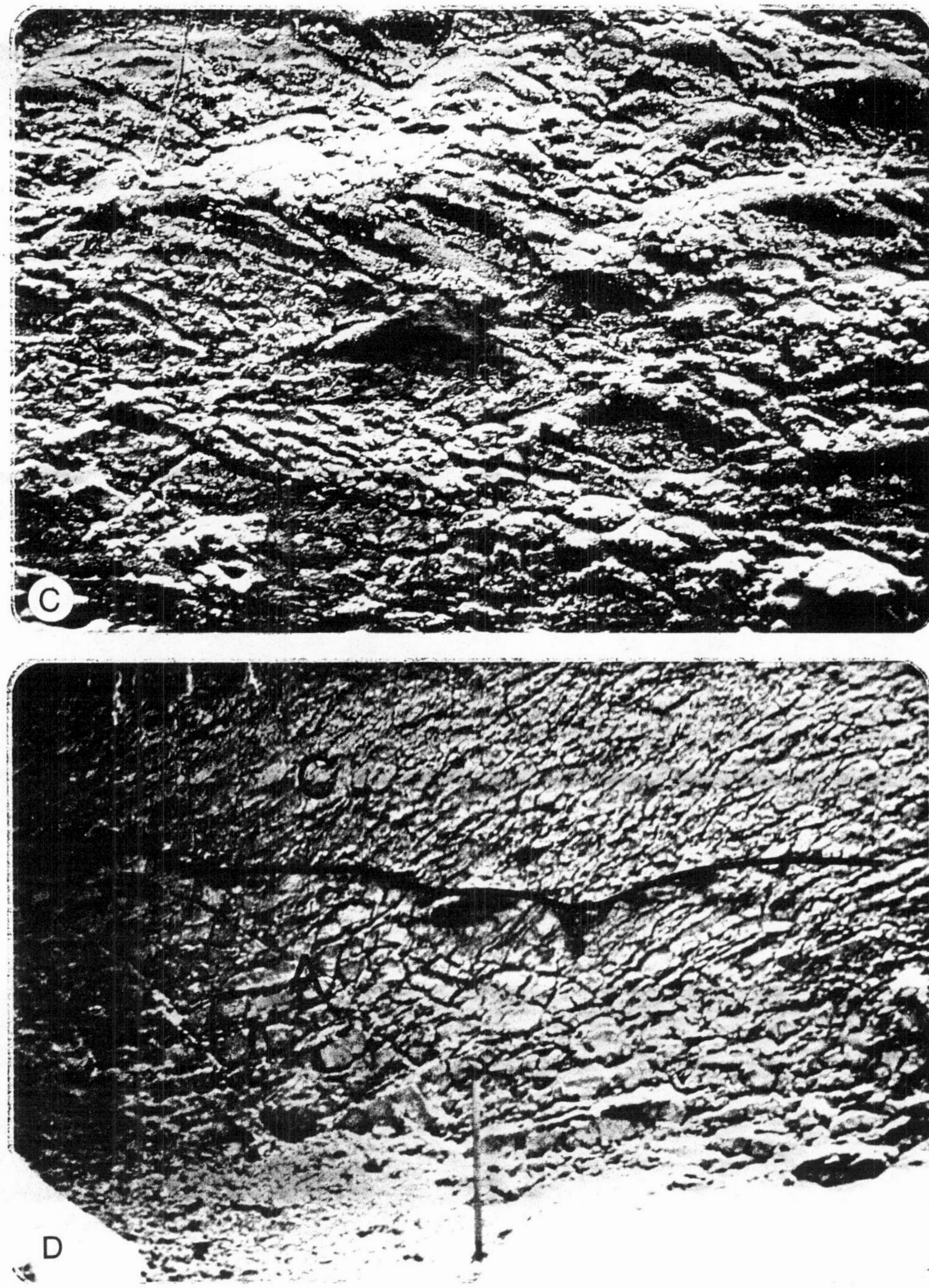

Figure 9. Continued. Cryotextures observed in headwall of retrogressive thaw flow slide, as seen in January, 1984. c) Close-up of cryotexture A. d) View of cryotexture A with schlieres, overlain by cryotexture C and separated by the organic rich thaw unconformity.

and 5 - 10 cm thick. Typically, the lattice possessed a preferred orientation which varied along the face (see Figure 8). Detailed study of a 160 cm profile through this cryotexture showed that the cryotexture actually consisted of a number of different ice-silt layers. However, the silt layers appear to have been forced apart by ice growth and many lithified fragments appear suspended in the enclosing ice bodies. Often ice schlieres cut through the silty layers, further reinforcing the lattice network.

Cryotexture A dominates the exposed headwall in extent. It occurred below the buried organic layer described above and is the texture which occurs at the lowest stratigraphic position in the headwall. It is therefore referred to informally as the 'primary' cryotexture. Following E. M. Katasonov (in Kudryatsev, 1978, Figure 158, pp. 315-316) this cryotexture is interpreted as one associated with subacqueous syngenetic freezing of lake sediments, and typical of non-through taliks beneath lakes. The lenses are oriented with long axes essentially parallel to the freezing front and typically, lie horizontally, inclined at an angle, or vertically (see Figure 9). According to Katasonov, the ice schlieres indicate that the silty clay was only lightly compacted when it froze. The schlieres form along cracks that develop as the result of subacqueous leaching. It is also argued (Kudryatsev, 1978, p. 316) that freezing in this manner can only take place under extreme geocryological conditions, with ground temperatures below -5 to -7°C.

Cryotexture B (Figure 10):--This texture is characterised by an ice matrix enclosing randomly arranged, angular and/or platy silt fragments >1.5 - 2.0 mm in size. An interesting feature is that this texture commonly was observed in narrow zones extending downwards from the buried organic layers into the primary cryotexture (e.g., see Figure 8; also Figure 9a).

The origin of this cryotexture is not clear. One possible interpretation is that it represents a former seepage zone or talik which existed at the time when the thaw lake, represented by the organic layers, was present.

Cryotexture C:--The frozen silts lying stratigraphically above the buried organic layer were not examined closely, since access was not possible. However, examination of the exposure (Figure 11) indicates a lack of massive icy lenses and a lower ice content which is more uniformly dispersed throughout the sediment. Burns (1982) reports that typical excess ice values from this zone are approximately 50% in contrast to those of cryotexture A which may be as high as 90%.

The lower ice content, and the lack of a clearly recognizable and distinct cryotexture in these sediments is perhaps consistent with their origin as being refrozen lake sediments.

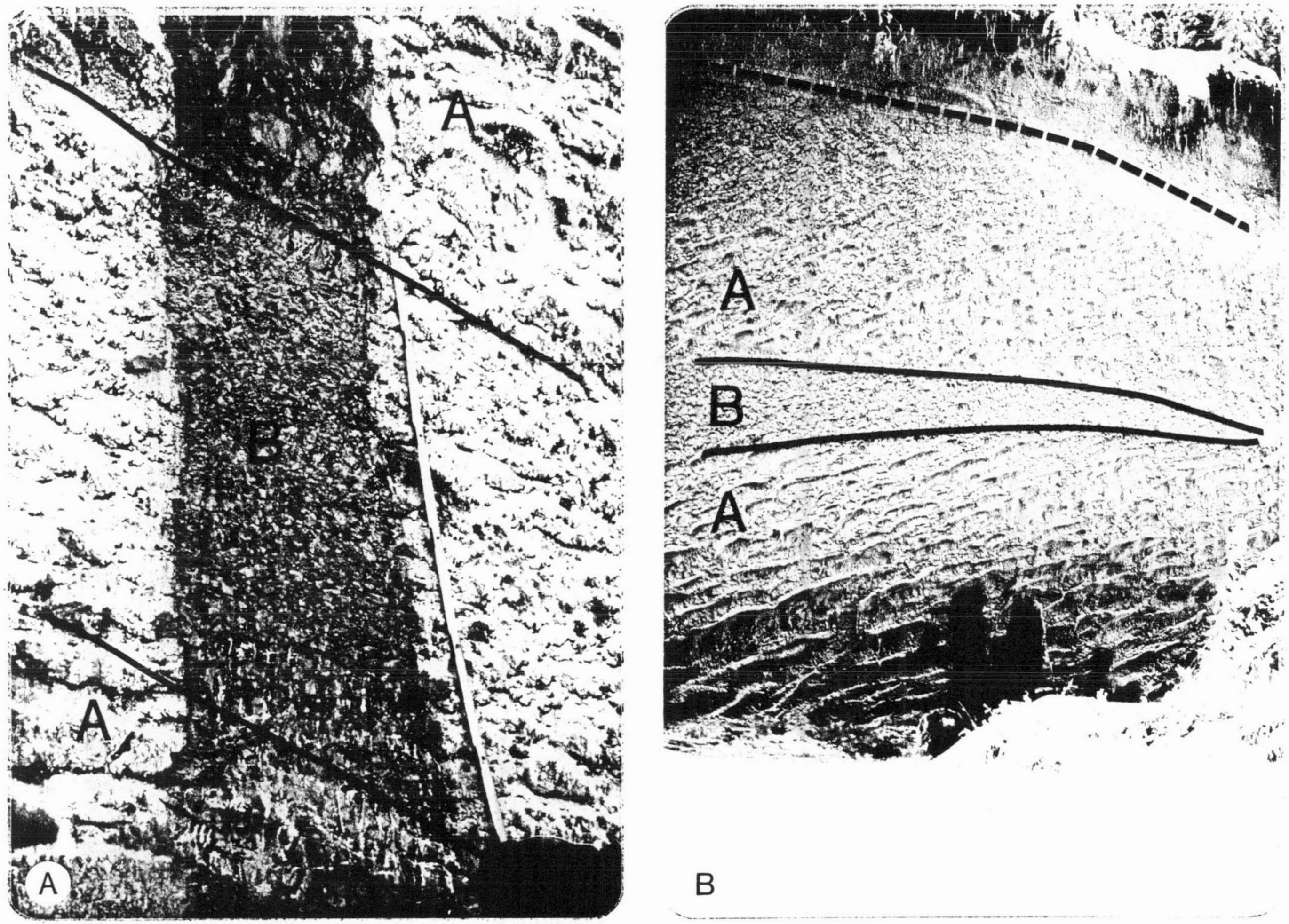

Figure 10. Cryotextures present in headwall of retrogressive thaw flow slide. a) Cryotexture type B. b) View of cryotexture B interfingering with cryotexture type A.

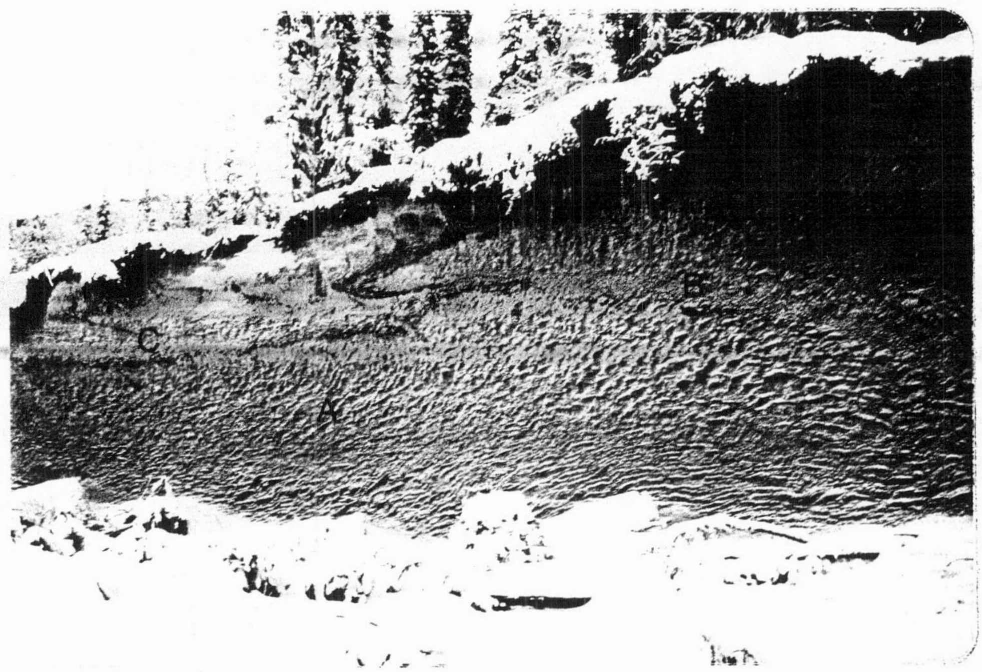

Figure 11. View of southern part of headwall exposed in January, 1984, illustrating cryotextures A, B and C, together with buried organic layer.

A number of X.R.D. analyses, performed on several sediment samples taken from this zone indicate a decrease in the illite:quartz and chlorite:quartz ratios with increasing depth to approximately 2.0 m, after which the ratios remain constant. These data, albeit limited and requiring further analysis, provide additional justification for regarding the sediments lying above the buried organic layer as being a distinct stratigraphic and cryotextive unit. However, our limited understanding of cryotextures at this locality permits neither a more detailed interpretation of this zone nor any further rationalisation for our interpretation.

3.5.3 Sampling Design

During fieldwork in January 1984, four oriented blocks of ice were extracted from the headwall. The locations of these samples is indicated on Figure 8. Their significance is summarised below.

Sample One was taken from a large segregated ice lens of cryotexture A (see Figure 9b). Sample Two was also taken from ice of cryotexture A at the same absolute elevation in the face, but stratigraphically below a zone of cryotexture B and at approximately 75 metres along the face from Sample One. By contrast, Samples Three and Four were taken from ice-rich silty sediments lying stratigraphically above the buried organic layer and are thought representative, therefore, of cryotexture C. Because of the angular fragment matrix of cryotexture B, it was impossible to obtain ice samples of sufficient size and purity for meaningful ice fabric analyses.

4. GROUND ICE PETROGRAPHY

4.1 Introduction

Segregation ice formation is a unique process of crystallisation differentiation that takes place in a complex polyphase system in which moisture is drawn to a stationary freezing front under conditions of high pore water pressure and capillary attraction (Mackay, 1971; 1972; Shumskii, 1964). The inclusion characteristics of segregation ice as well as the size and structural characteristics of ice crystals are controlled by soil type, moisture supply and temperature and heat flow direction at the time of ice formation. Thus, petrography can provide useful information concerning ice genesis and post-freezing history.

Fine-grained sediments, like the glaciolacustrine silts and clays found in Stewart Valley near Mayo, are highly susceptible to segregated ice formation. In general, the morphology of segregation ice bodies, particularly lens thickness, is defined by the relationship between the rate of heat loss and the rate of water influx. For example, as the cooling rate increases, the rate at which water is supplied to the freezing front increases in order to permit continuous growth and the formation of thick ice lenses. If the cooling rate exceeds the rate of water supply (either by increased cooling or decreased water supply) new centres of crystallisation will appear as the freezing front advances. Likewise, if the rate of cooling increases, ice bodies will become smaller and more closely spaced because there will be less time for water migration.

Because of these variables, segregation ice is difficult to interpret, both structurally and genetically. Except for general descriptions by Corte (1962), Gell (1976) and Penner (1971) there is an absence of information on segregation ice petrography in the ground ice literature.

4.2 Ice Characteristics

During fieldwork in January 1984, the exposed ice lenses were covered by a thin layer of hoar frost giving an opaque appearance which obscured the internal ice structure. In some of the large ice lenses (e.g., location of Sample One), sublimation has etched crystal outlines exposing a generally coarse texture characterised by large irregularly shaped crystals. In fresh sections, the ice appeared clear and dark. Vertical bubble trains, composed of vertically oriented, long, tubular bubbles and the occasional sediment inclusion, were visible near upper and lower sediment contacts. The latter were abrupt and uneven in nature. There was an absence of visible ice in the enclosing sediments indicating that moisture has been drawn from the surrounding material to support ice growth.

Under plain light in the laboratory, the ice samples appear to be clear and transparent. Although sediment and gas inclusions are present, they do not obscure the transparency of the ice (Figure 12). Immediately adjacent to the sediment contact is a thin suspended layer of fine sediment and gas bubbles. A similar band of gas and sediment inclusion occurs in Samples Three and Four. In large ice lenses, as represented by Sample One, gas inclusions are concentrated in zones immediately adjacent to the sediment contact. The bubbles nearest (i.e., within 1 cm) to the sediment contact are small (less than 1 mm in diameter) and spherical in shape. Further from the sediment contact (i.e., approximately 1 - 4 cm) gas bubbles are larger (8 - 15 mm long) and range from elliptical to tubular in shape, sometimes forming short bubble trains. Within the main body of the ice lens, both sediment and elongated gas inclusions are randomly distributed. However, thin sheets of small, flattened and oriented gas bubbles occur along fracture planes and strain shadows. Such fractures were noted in the field prior to sampling and are not due therefore to sampling or thermal shock during handling. Fracture planes, visible in Figure 13, and strain shadows (small discontinuous fracture planes subparallel to the

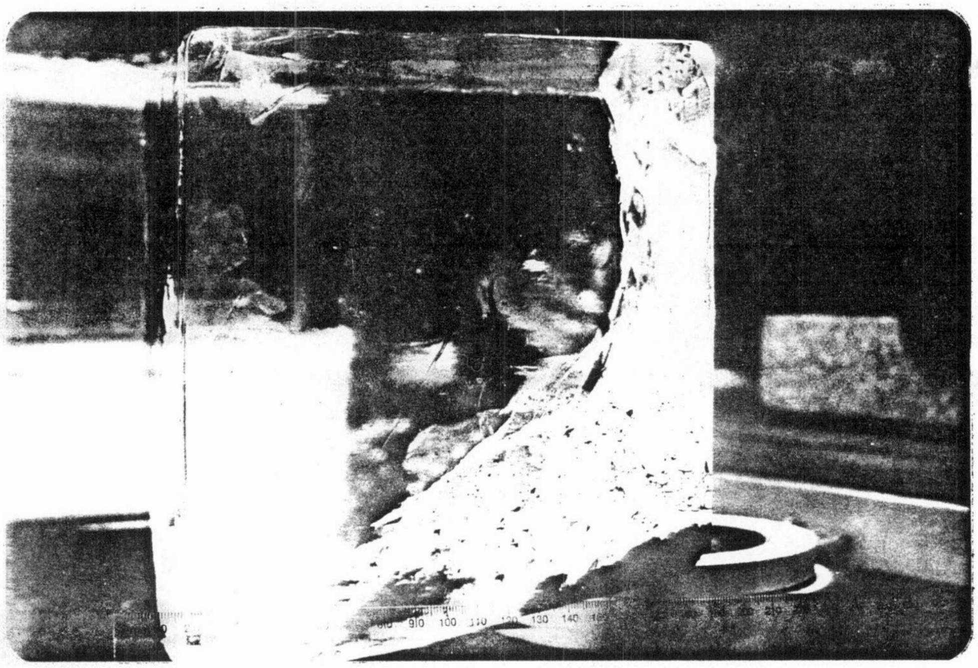

Figure 12. Photograph under plain light of ice Sample One (cryotexture A). Note that the sample includes part of the lower sediment content which is abrupt and unconformable. A large piece of sediment and numerous small sediment fragments or grains can be seen in the lower part of the sample.

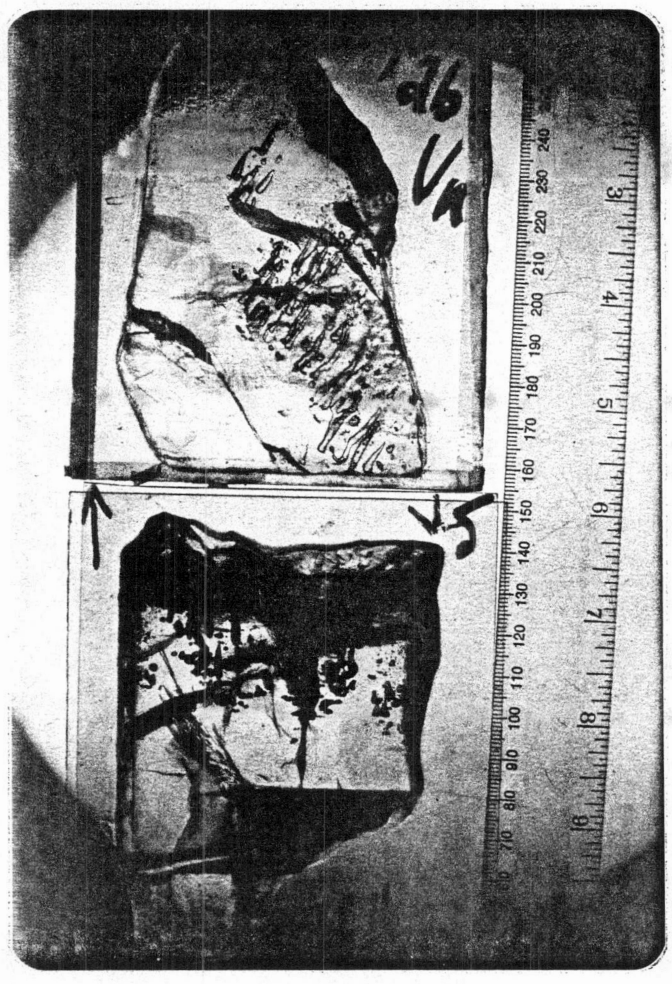

Figure 13. Thick sections, viewed under plain polarised light, of ice Samples Two (cryotexture A) and Three (cryotexture C) showing strain shadows. Fractures appear as distinct planes through the ice while strain shadows are ghost-like and discontinuous.

main fractures) visible in Figure 13, were observed in all four samples.

In ice Samples Two and Three, fracture planes occurred mainly in vertical and horizontal directions. In one case, a fracture completely bisected a long (25 mm) tubular bubble with no apparent displacement along the fracture planes. There was no evidence of displacement where either fractures or strain shadows intersect each other, or where fractures meet sediment contacts. The general absence of displacement associated with fractures or strain shadows may indicate they were derived from thermally generated stresses rather than shear stress.

Samples Two, Three and Four taken from ice bodies considerably smaller than Sample One, are thought representative of small ice lenses. As in Sample One, the ice in these samples is clear and transparent. However, the gas and sediment inclusion content is higher and more evenly distributed through the samples. Gas inclusions constitute approximately 5 - 10% of the total lens volume, while sediment makes up roughly 1 - 2%. In addition to the 1 cm layer of high sediment and gas inclusion content immediately adjacent to the upper and lower sediment contacts, as observed in Sample One, small lenses are also characterised by thick (3 - 5 cm) discontinuous bands of large oval, and tubular bubbles and bubble trains. As a result, both Samples Three and Four have a layered appearance. In Samples Two and Three long tubular gas inclusions closely parallel vertical fracture plains. In general, gas inclusions near the sediment boundaries are small while large, elliptical and tubular bubbles characterise the body of the ice lens.

Deviations from typical gas inclusions were observed in thick sections from Samples Two and Three. In one case, long tubular bubbles were curved or bent while neighbouring inclusions remained straight. In another, the lower end of long tubular bubbles terminated abruptly forming flat or squared ends, and others ended in a thin tail bent normal to the bubble long-axis. Each of these variations may be indicative of changing freezing condition during lens formation; equally, they could be post-formational and represent thermo-migration. The most abnormal bubble shapes were ringed or spiral-shaped bubbles observed near the upper sediment contact in Sample Three. It is thought that these are a post-freezing phenomena formed by thermo-migration when a large bubble broke into smaller forms.

In summary, the occurrence of large spherical and tubular gas inclusions is characteristic of slow to intermediate ice growth. The small spherical bubbles observed near sediment contacts may suggest initially rapid ice growth. The vertical, tubular orientation of the bubbles and bubble trains reflect vertical growth as does the layered nature of some of the ice. Finally, the presence of post-freezing phenomena (e.g., fractures, stress shadows, thermo-migration of bubbles) indicate that the ice has undergone thermal stress after formation.

4.3 Crystal Characteristics

A number of vertical and/or horizontal thin sections were prepared from each of the ice samples. The orientation and size of the thin sections were controlled by the size and shape of the ice samples and the distribution of sediment within them. Examination of thin sections under cross-polarized light provides the basis for texture analysis, while fabric orientations can be performed using a universal stage, also under cross-polarized light. Sample One provided the most information while Samples Two, Three and Four, due to their small size and high sediment content, provided considerably less.

Ice Sample One (cryotexture A) contained large crystals, frequently exceeding 12 - 14 cm in long-axis length. Figure 14a shows two horizontal thin sections, under cross-polarized light, from the upper part of Sample One. The long-axis dimensions of ice grains in all sections ranged between 1.5 and 7.1 cm for whole crystals, and exceeded 10.0 cm for partial crystals. Considerable variation was observed through the section, for example; Figure 14b shows four horizontal thin sections taken from the middle of the sample. The grain dimensions are considerably

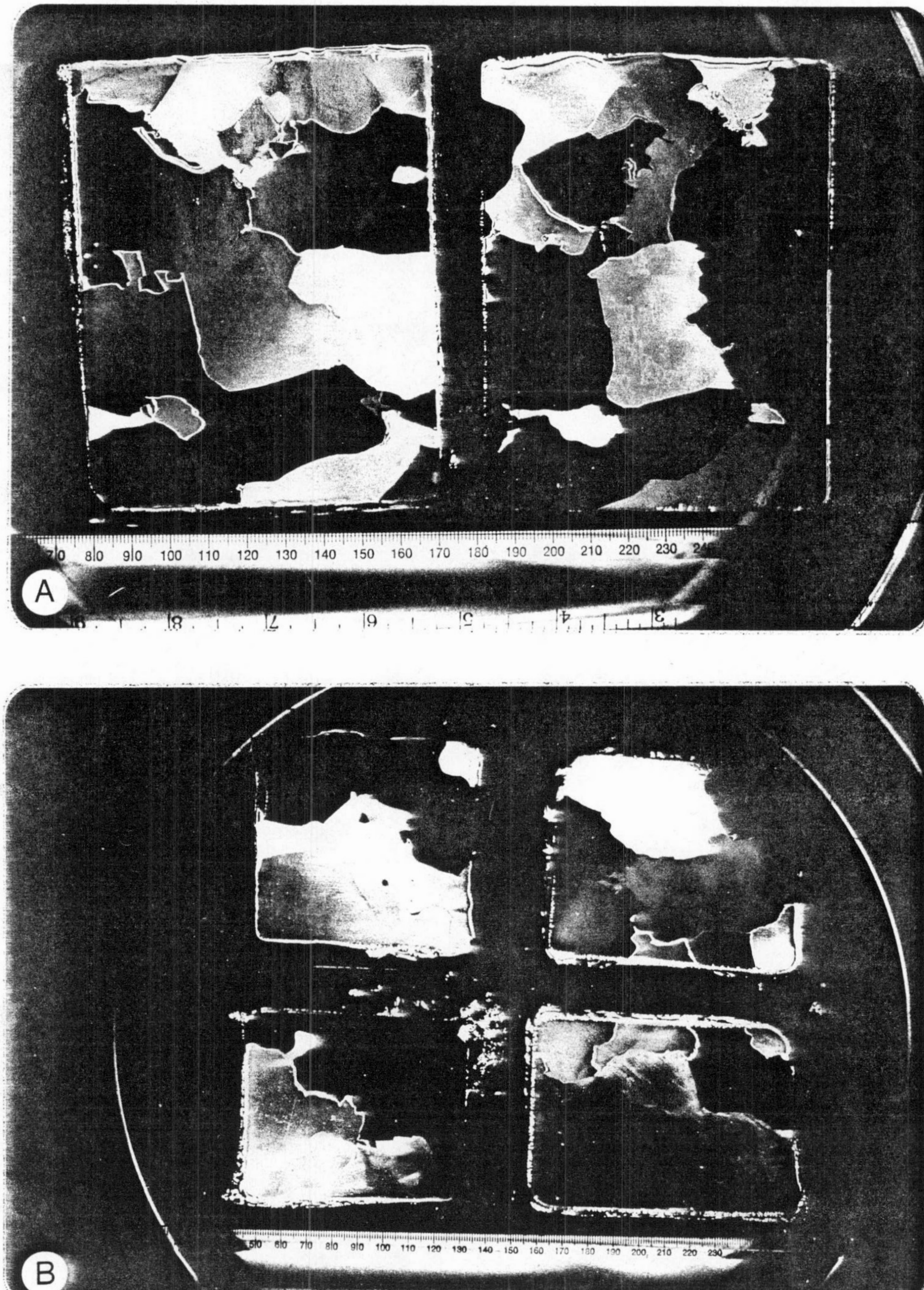

Figure 14. Horizontal thin sections viewed under cross-polarised light from ice Sample One (cryotexture A). a) Two sections from the upper part of the sample. b) Four sections from the middle of the sample.

larger with average crystal size varying between 9 and 22 cm² for ten horizontal thin sections. There was no preferred orientation of crystal long-axis direction.

The dimensions of crystals in vertical thin sections were consistently greater than those observed in horizontal thin sections. Figure 15 shows two vertical thin sections oriented parallel to the thaw face from roughly the same location in Sample One as Figure 14b. The long-axis dimensions of crystals ranged from less than 1.0 cm to 13.4 cm in length. The average crystal size, based on nine vertical thin sections, ranged from 20 to 32 cm². The smallest was 3 cm² and the largest was 45.7 cm².

These large crystal sizes, with a weak preferred vertical orientation of crystal long axes, are indicative of a slow freezing rate and a roughly vertical freezing or heat flow direction.

The texture of the ice in Samples Two, Three and Four was similar to that observed in Sample One, although grain size was not as large. Ice crystals varied from 1.5 cm² to 24 cm² for vertical thin sections from Samples Two and Three and up to 30 cm² for Sample Four. Long-axis dimensions varied from 1.2 cm for small crystals adjacent to lens/sediment contacts, to 8.8 cm for large crystals in the middle of the lens. In small ice lenses, it was not uncommon for large crystals to extend across the entire lens thickness. Similar observations were made by

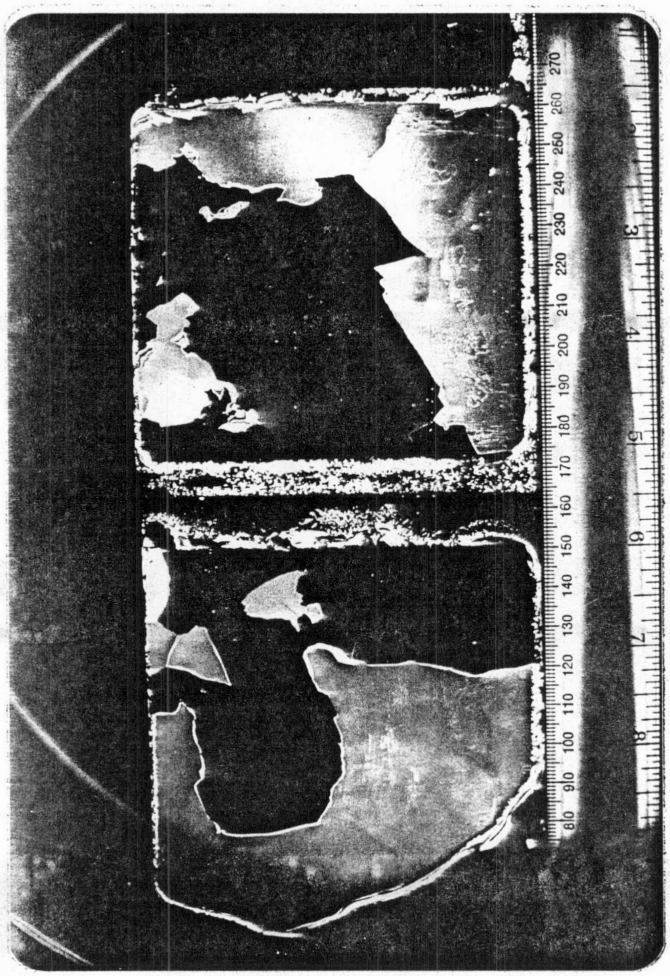

Figure 15. Two vertical thin sections oriented parallel to the thaw face, ice Sample One (cryotexture A). Note the very large crystal dimensions.

Penner (1971) and Gell (1976) in considerably thinner ice lenses.

4.4 Sediment and Gas Inclusions

The influence of gas and sediment inclusions on ice texture was briefly investigated. In Sample One, for example, sediment and gas inclusions are both inter- and intragranular, depending on their location in the sample. In layers of small crystals near sediment contacts large sediment inclusions, and to a lesser degree gas inclusions, are intergranular and located along vertically oriented grain boundaries. As a result, various loc l irregularities in boundary shape are directly associated with gas or sediment inclusions. In the middle of the ice lens, where grain size is coarser, both types of inclusions are exclusively intragranular and have no influence on the crystal texture.

4.5 Fractures and Strain Shadows

Fracture plane occurrence is generally intracrystalline in nature except where bands of small ice grains occur, where they are both inter- and intracrystalline. In contrast, strain shadows were only intracrystalline. In the lower portion of the ice sample in Figure 16 for example,

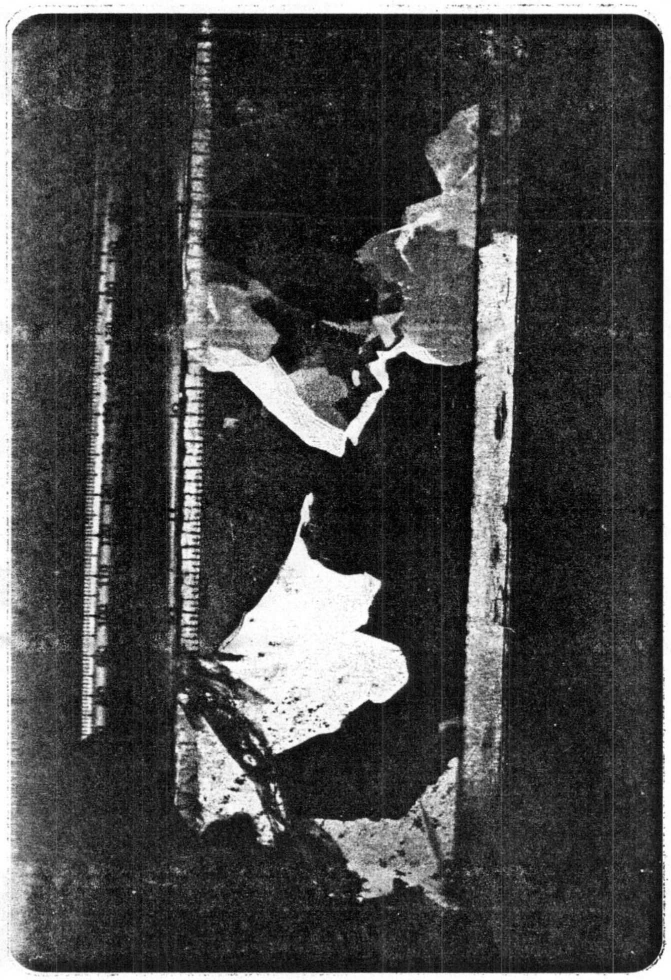

Figure 16. Vertical thick section (5 cm) from Sample One, under cross-polarised light. Note the fractures and strain shadows in the lower right of the sample.

both fractures and strain shadows can be seen. Minor recrystallisation has occurred along some of the larger fracture planes and there is some evidence of minor displacement (approximately 1 mm). Although not apparent in the photographs, crystalline substructure occurs in the form of slightly different extinction angles for different parts of the same crystal.

Fractures and strain shadows occur in greater concentration in Samples Three and Four (cryotexture C) than in Sample One (cryotexture A). They are primarily intragranular in occurrence. Along major fracture planes a small amount of displacement has taken place (up to 1 mm), which may indicate boundary migration or shear displacement.

4.6 Petrofabrics

Petrofabric analysis was performed on a number of vertically oriented thin sections cut from the ice samples in planes either parallel or normal to the thaw face of the Mayo ground ice slump. Sample One was sufficiently large and free of sediment to allow vertical thin sections both normal and parallel to the thaw face to be analysed, as well as several horizontally oriented thin sections. Since the other ice samples were smaller and contained blocks and layers of sediment, the geometry of the ice lens combined with the shape of the sample ultimately controlled the plane through which thin sections could be prepared. Consequently, vertical normal and vertical parallel sections were analysed for Sample Two, while only vertical normal sections for Samples Three and Four.

C-axis orientation measurements are depicted graphically by the Schmidt method (Langway, 1959, pp. 8-12; Schmidt, 1925, p. 392) as a diagram of the density distribution of traces of the optic axes on the surface of a sphere.

Fabric diagrams for both vertical and horizontal thin sections from Sample One are presented in Figure 17. Photo 17a shows a tendency for a preferred pattern of c-axes concentrated in four fairly strong maxima inclined 30 - 40° to the vertical. A dispersed pattern of c-axis orientations surround each pair of maxima. Figure 17b, based on vertical thin sections oriented normal to the thaw face, shows a fabric orientation characterised by two isolated primary maxima inclined 40 - 50° to the vertical. Secondary maxima occur in the opposite quadrants inclined 20 - 30° to the vertical. Both are in planes almost normal to either the upper or lower contacts. Figure 17c shows c-axis orientations of six horizontal thin sections at various levels in the sample. Again, four strong orientation maxima inclined 30 - 50° to the vertical are apparent.

Fabric diagrams for thin sections prepared from Samples Two, Three and Four are presented in Figure 18.

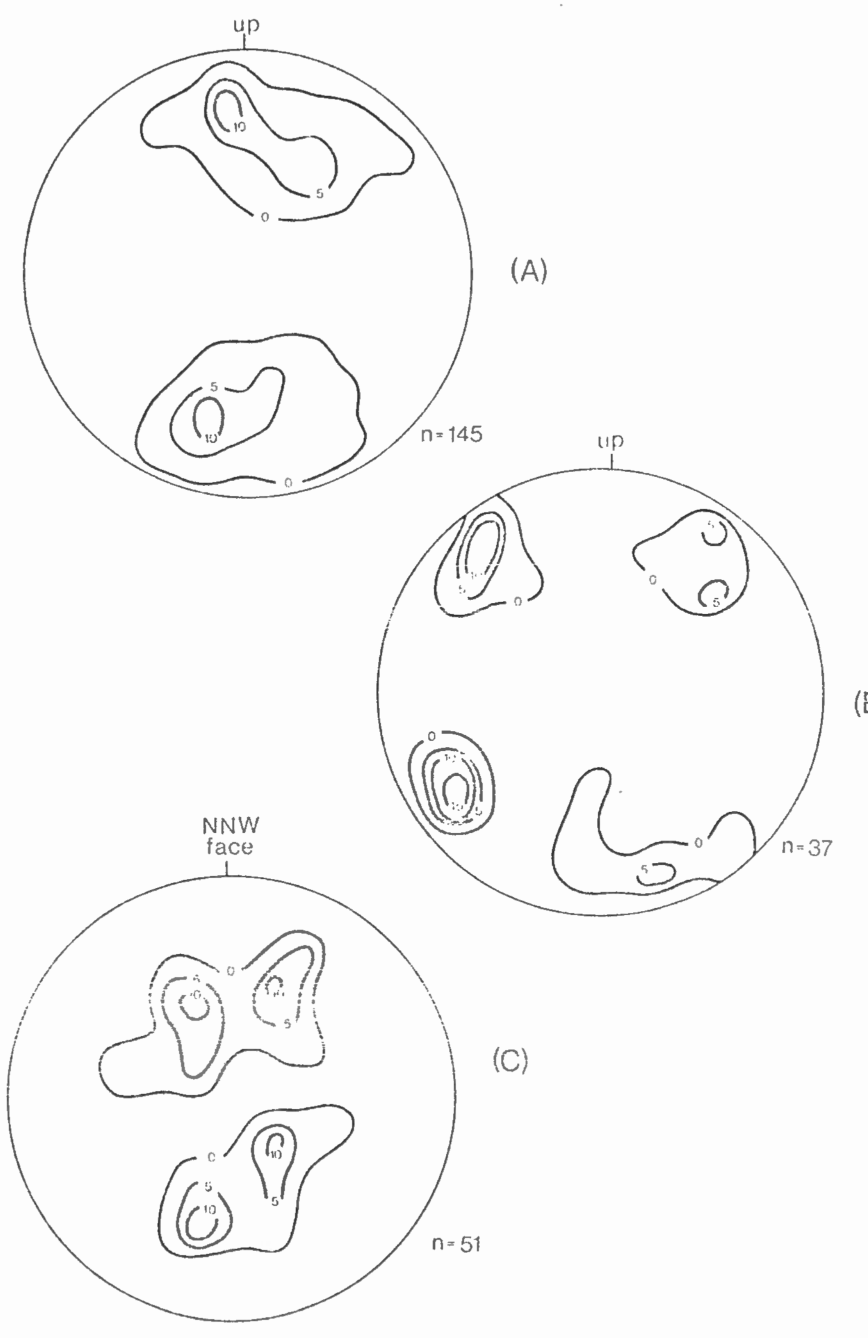

Figure 17. Fabric diagrams of c-axis orientations, Sample One (cryotexture A). a) Vertical thin section parallel to thaw face. b) Vertical thin section oriented normal to thaw face. c) Horizontal thin section parallel to lens axis.

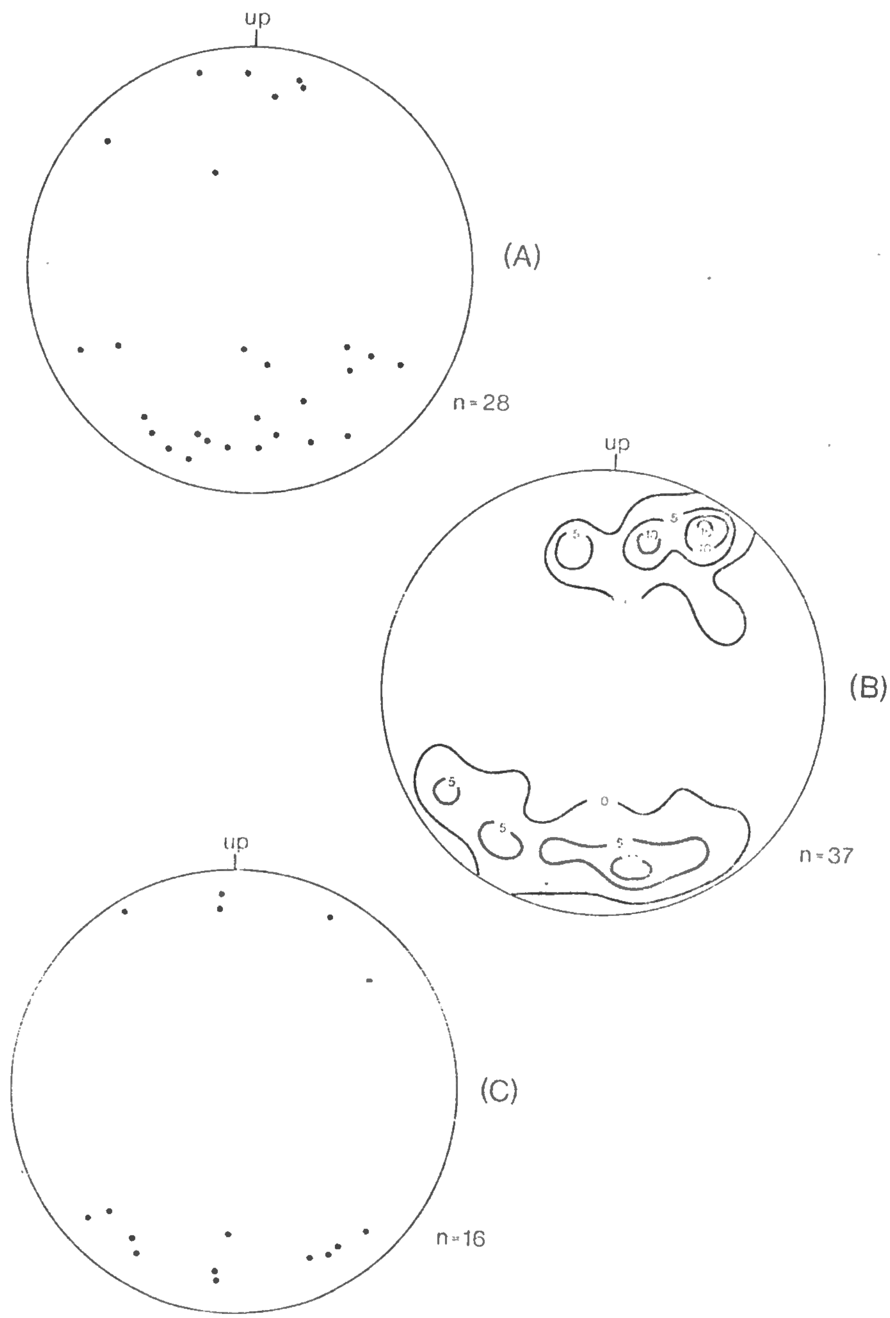

Figure 18. Fabric diagrams for ice Samples Two (cryotexture A) and Three and Four (cryotexture C). a) Ice Sample Number Two. b) Ice Sample Number Three. c) Ice Sample Number Four.

In Figure 18a, the dispersed pattern falls within the same range observed in Sample One. Crystal c-axes are inclined roughly 5 - 45° to the vertical, but there is a general lack of concentrated maxima. Figure 18b shows a weak pattern of preferred c-axis orientations with maxima inclined 10, 15 and 30° to the vertical and a general diffuse pattern inclined 5 - 45° to the vertical. Figure 18c shows a loose, dispersed pattern of c-axes inclined 5 - 35° to the vertical. The distribution of points is similar to the previous diagrams. However, clusters of points or concentration maxima, indicating a preferred orientation of crystal c-axes, are lacking.

In summary, the fabric diagrams indicate that c-axis orientations range from weak and dispersed with a maximum of 5% to a concentration maxima of 15%. The dispersed patterns contain c-axes inclined 5 - 45° to the vertical. Preferred patterns have strong maxima occurring in all quadrants, inclined 10 - 40° to the vertical. The fabric diagrams also indicate an absence of c-axes in either polar or equatorial orientations.

5. DISCUSSION AND INTERPRETATION

The morphology, distribution and setting of the ground ice exposed at the Mayo study site clearly support an ice segregation origin. In all probability, the ice formed as permafrost aggraded into saturated glaciolacustrine sediments of Stewart River valley. Although the physical character of the ice offers little information concerning its age, it does provide an indication of the conditions under which freezing occurred and the post-freezing history.

The large ice lens size, together with the coarse texture of the ice, indicates that ice segregation occurred slowly and at relatively high freezing temperatures. This implies that moisture supply and heat flow conditions were suitable for the formation of large ice bodies. However, the shape of some of the ice bodies, particularly the large lenses that taper horizontally, and the variable lens size, indicate that heat flow and/or moisture supply were spatially and temporally variable. As a result, the shape and orientation of ice lenses varies considerably across the thaw face and provides the cryotextures which are superimposed on the silty sediments.

Lens orientations vary across the face from horizontal to gently dipping. In general, the axes of the ice lenses are inclined 10 - 30° toward the northeast. However, numerous lenses are arched or bent, generating locally steep dip angles. The inclined nature of the ice lenses, as well as the variation in lens shape, may have resulted from a number of factors. These include the litho-stratigraphy of the sediments, and either inclined heat flow or nonlinear heat flow. An uneven supply of moisture, or differential freezing associated with different surface conditions may also have been important. While horizontal variation in ice character probably reflect spatial variations in heat flow and moisture supply, vertical variation indicates temporal differences. Where silt fragments are included in the ice, or where ice lenses and veins are closely spaced, blocks of sediment appear to have been rotated. This is interpreted as the result of either differential growth rates or adjacent crystals or local variation in moisture supply.

Gas inclusion characteristics, particularly the large bubble size and vertical orientation of tubular bubbles and bubble trains, are indicative of vertical heat flow and slow freezing. The generally large crystal size and irregular inequigranular shape also support a slow freezing interpretation. The range in bubble size, and to a lesser degree grain size, suggest that a range of freezing temperatures may have occurred. The low particle inclusion content of the ice is also indicative of slow freezing.

The presence of fractures, strain shadows and thermo-migration phenomena indicate that post-formational modification of the ice has taken place. However, there is no way of determining how much of the observed petrography is the result of post-freezing stresses. The nature of the post-freezing and the lack of displacement and recrystallization associated with fractures suggest thermal post-formational stress rather than shear stress.

In conclusion, the texture and fabric patterns of segregation ice observed at the Mayo site are clearly different from other ground ice types observed elsewhere in the Yukon (e.g., Pollard, 1983; Pollard and French, 1983). For example, the large inequi-granular and weakly oriented grain pattern of the ice lenses is in marked contrast to the small grained and dimensionally oriented pattern of icing ice observed in the North Fork Pass and buried icing ice from the Klondike areas of the Yukon Territory. It is also quite different from the columnar texture of frost blister ice, and that of pingo ice and wedge ice, as described in the literature (e.g., Pollard and French, 1983; Gell, 1978a;b; and Black, 1978).

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