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CURRENT RESEARCH, PART B
INTERIOR PLAINS AND ARCTIC CANADA

RECHERCHES EN COURS, PARTIE B
PLAINES INTÉRIEURES ET RÉGION ARCTIQUE DU CANADA

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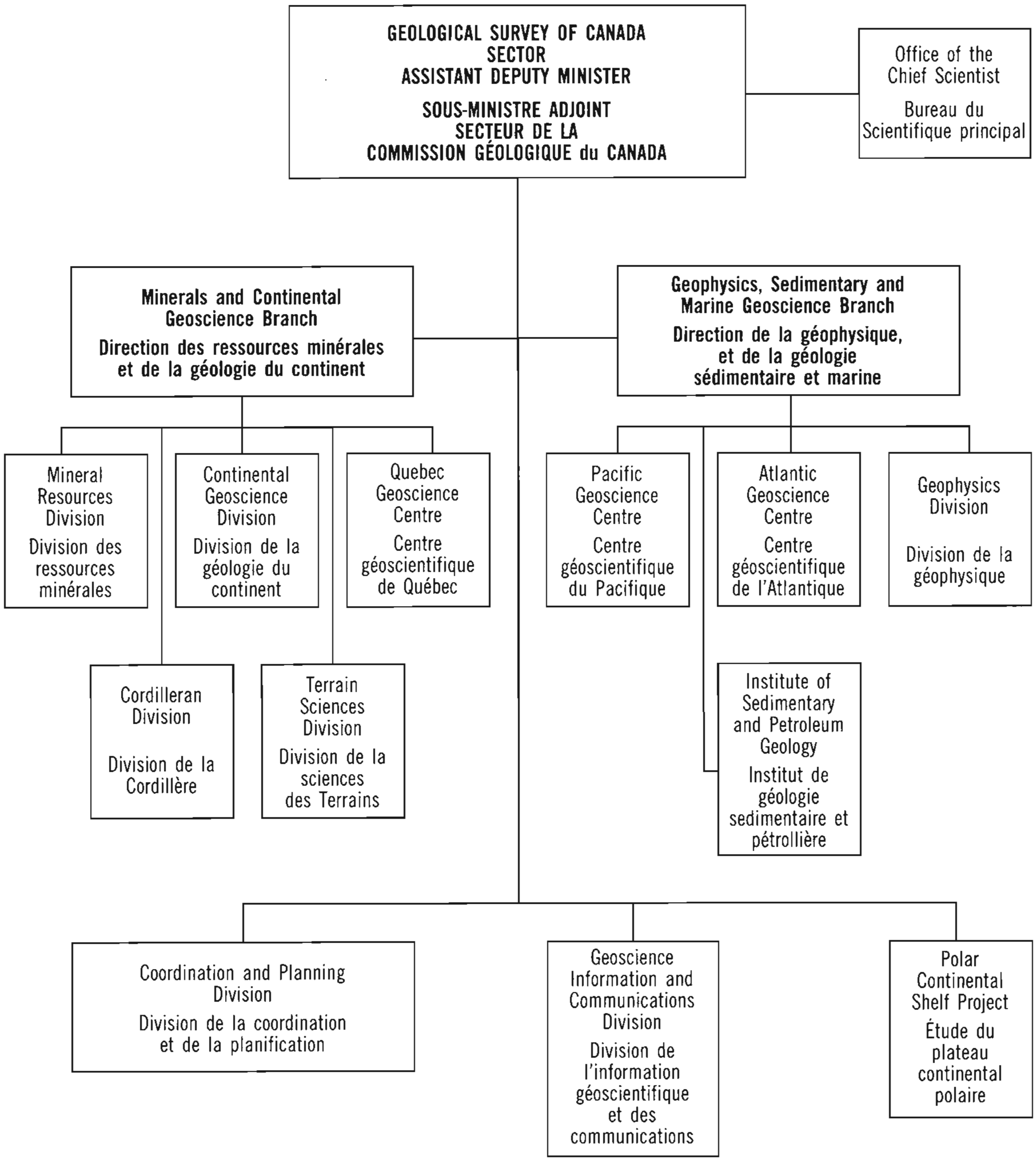
Cover description

View to the northeast along northern Makinson Inlet, south-central Ellesmere Island. Cliff exposes Early Devonian redbeds (Vendom Fiord Formation) and overlying yellowish dolostone (Blue Fiord Formation). A downfaulted slab of Blue Fiord dolostone, about 100 m wide, is in the central part of the cliff. Photo by T. de Freitas.

Description de la photo couverture

Vue vers le nord-est le long de la partie nord de l'inlet Makinson, dans le centre sud de l'île d'Ellesmere. On remarque, dans la falaise, des couches rouges (Formation de Vendom Fiord) du Dévonien précoce recouvertes par des dolomies jaunâtres (Formation de Blue Fiord). Un morceau déplacé vers le bas de la dolomie de Blue Fiord, large d'environ 100 m, occupe le centre de la falaise. Photo prise par T. de Freitas.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CONTENTS

New field observations on the geology of Bathurst Island, Arctic Canada: Part A, stratigraphy and sedimentology of the Phanerozoic succession

T. de Freitas, J.C. Harrison, and R. Thorsteinsson
Institute of Sedimentary and Petroleum Geology, Calgary

de Freitas, T., Harrison, J.C., and Thorsteinsson, R., 1993: New field observations on the geology of Bathurst Island, Arctic Canada: Part A, stratigraphy and sedimentology of the Phanerozoic succession; in Current Research, Part B; Geological Survey of Canada, Paper 93-1B, p. 1-10.

Abstract: Recent field investigations on Bathurst Island have necessitated considerable revision of formational boundaries and nomenclature of Silurian and Devonian rocks. In particular, the names Eids and Blue Fiord will have to be abandoned, and this will probably also be necessary for the Bathurst Island and Stuart Bay formations. In the eastern parts of the island, a unit of conglomerate, redbeds, and dolostone, representing syntectonic deposits of the Boothia Uplift, lies with angular unconformity on various older formations. In earlier studies, these deposits were assigned variously to the Bathurst Island, Stuart Bay, and Disappointment Bay formations, but are here referred to as the Prince Alfred Formation, defined and widely distributed on northern Devon Island. The Prince Alfred grades abruptly, basinward into conglomerate and siltstone of the Stuart Bay Formation, in which Pragian to earliest Emsian age graptolites occur. The Cape De Bray, Awingak, Canyon Fiord, and Beaufort formations, previously unknown from Bathurst Island, are also described.

Résumé : De récentes études de prospection réalisées dans l'île Bathurst ont mené à une révision considérable des limites des formations et de la nomenclature des roches siluriennes et dévoniennes. En particulier, les noms d'Eids et de Blue Fiord devront être abandonnés, et il en sera probablement de même pour les formations de Bathurst Island et de Stuart Bay. Dans les secteurs est de l'île, une unité composée de conglomérat, de couches rouges et de dolomie, qui représente des gisements syntectoniques du soulèvement de Boothia, repose en discordance angulaire sur diverses formations plus anciennes. Lors d'études antérieures, on a diversement placé ces gisements dans les formations de Bathurst Island, de Stuart Bay et de Disappointment Bay, mais on les regroupe ici sous le nom de Formation de Prince Alfred, définie et très répandue dans le nord de l'île Devon. La Formation de Prince Alfred passe abruptement en direction du bassin au conglomérat et au siltstone de la Formation de Stuart Bay où apparaissent des graptolites couvrant l'intervalle du Praguien au tout début de l'Emsien. Les formations de Cape De Bray, d'Emma Fiord, de Canyon Fiord et de Beaufort, autrefois inconnues dans l'île Bathurst, sont également décrites.

INTRODUCTION

A substantial contribution to the geological understanding of Bathurst Island was made during Operation Franklin (McLaren, 1963a; McMillan, 1963; Thorsteinsson and Glenister, 1963). Later, Kerr (1974) and Temple (1965), from a base camp at Polar Bear Pass, carried out more detailed geological investigations in 1963 and 1964. Conodont collections from Lower Devonian rocks (McGregor and Uyeno, 1972) indicated inconsistencies in Kerr's (1974) earlier stratigraphic correlations. Extensive, industry sponsored, oil and base metal exploration resulted in a producing oil well from Devonian carbonates beneath nearby Cameron Island (Fig. 1). Much smaller field operations, mainly by workers from Canadian universities, contributed to several aspects of the geology of Bathurst Island. Polan and Stearn (1984) studied the olistoliths in what they believed to be the Stuart Bay Formation, but most of these blocks are older and belong to the Bathurst Island beds, as defined herein. Aspects of the Devonian clastic wedge were described by Embry and Klovan (1976), while Smith (1980), in search of brachiopods, measured a section in Cut Through Creek.

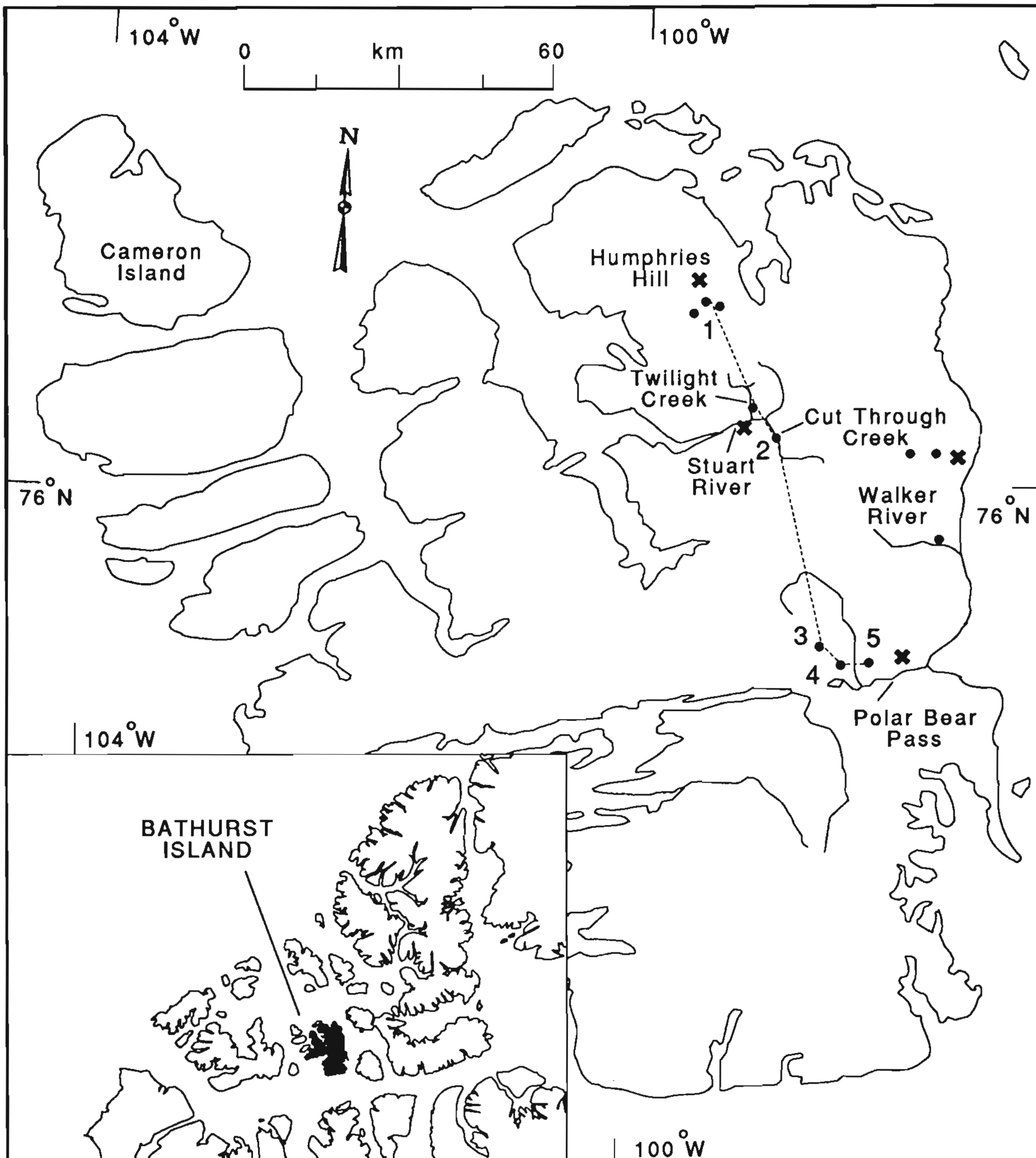


Figure 1. Location of studied stratigraphic sections. Line of section is shown in Figure 2. "•" and "X" show measured stratigraphic sections and base-camp locations respectively for the 1992 work.

Present field investigations were carried out during the summer of 1992 from four main base camps (Fig. 1). Four-wheel, all-terrain vehicles were used for local transportation and helicopter and fixed-winged aircraft were utilized for base camp relocation. The field party consisted of three people, with two main responsibilities: TdF and RT were involved primarily in the stratigraphy and sedimentology of the succession (Table 1), while JCH was concerned more with the mapping and deformational history of the area, which comprises the second part of this two part report (Harrison et al., 1993). This is a preliminary report based mainly on a detailed description and measurement of 11 stratigraphic sections (Fig. 1) and on field identification of graptolites by the authors.

BAY FIORD FORMATION

The formation near Walker River (Fig. 1) is 132 m thick and consists of thin bedded, greyish yellow green to pale yellowish green weathering dolostone that is finely crystalline, thin stylobedded, parallel laminated, and locally bitumen stained. The basal part of the section contains several thick bedded, mottled, light medium grey weathering, petroliferous dolostone beds, while the top 6 m consists of fissile, greenish grey mudrock that is conformably overlain by the Thumb Mountain Formation.

Near Humphries Hill, a well exposed carbonate mosaic packbreccia and underlying evaporite unit form the core of a large anticline (Table 1; Harrison et al., 1993, Fig. 3). The evaporite unit is about 100 m thick and consists of interbedded satin spar, selenite, and thin bedded dolostone; however, the unit is structurally complex, and its base is not exposed. It is overlain conformably by an approximately 20 m thick, thick bedded, mosaic packbreccia unit. Where not preserved as breccia, this unit consists of thick bedded, vaguely mottled, locally chert nodule rich, dolomitic limestone with vague thrombolitic and microbial textures and microbial mounds. This unit is perhaps a correlative of Member B of the Bay Fiord Formation on Ellesmere Island and with Unit 3 of the Bathurst-Caledonian River well of southeastern Bathurst Island (Mayr, 1980). The Canyon Fiord Formation rests with angular discordance on this unit near Humphries Hill. In most other areas, the Bay Fiord is conformably overlain by Thumb Mountain Formation (Harrison et al., 1993). The Thumb Mountain and Irene Bay formations and the Allen Bay Tongue were not studied in detail.

CAPE PHILLIPS FORMATION

At Cut Through Creek, a 113 m thick partial section of the Cape Phillips Formation was examined. The basal part is late Llandovery in age and characterized by thin bedded to fissile, petroliferous, interbedded calcareous mudrock and argillaceous lime mudstone and wackestone. Uncompressed *Monograptus* cf. *priodon* are abundant in some beds, along with late Llandovery graptolites. The overlying Wenlock succession is a thin bedded, yellowish grey weathering, well indurated, calcareous mudrock, containing well preserved cyrtograptids. At Twilight Creek, older beds of the Cape Phillips Formation have yielded abundant Ashgill age trilobites (*Pseudogygites*) and graptolites (including *Orthograptus* cf. *fastigatus*).

A greenish weathering, well indurated mudrock unit occurs at the top of the formation throughout the study area. It is 6 to 32 m thick, light olive grey to yellowish grey weathering, and less calcareous than the underlying Wenlock part of the Cape Phillips Formation. However, because weathering characteristics are broadly similar, the unit is not mappable. Abundant ?*Lobograptus progenitor* occur immediately below and within the unit suggesting an early Ludlow age.

Near Polar Bear Pass, the Cape Phillips Formation consists of interbedded, thin bedded mudrock, argillaceous limestone, and bedded and nodular black chert, conformably overlying a thin argillaceous limestone unit assigned to a tongue of the Allen Bay Formation. The upper contact is abrupt and marked by the appearance of very thin bedded, medium grey, recessive weathering mudrocks of the Devon Island Formation.

DEVON ISLAND FORMATION

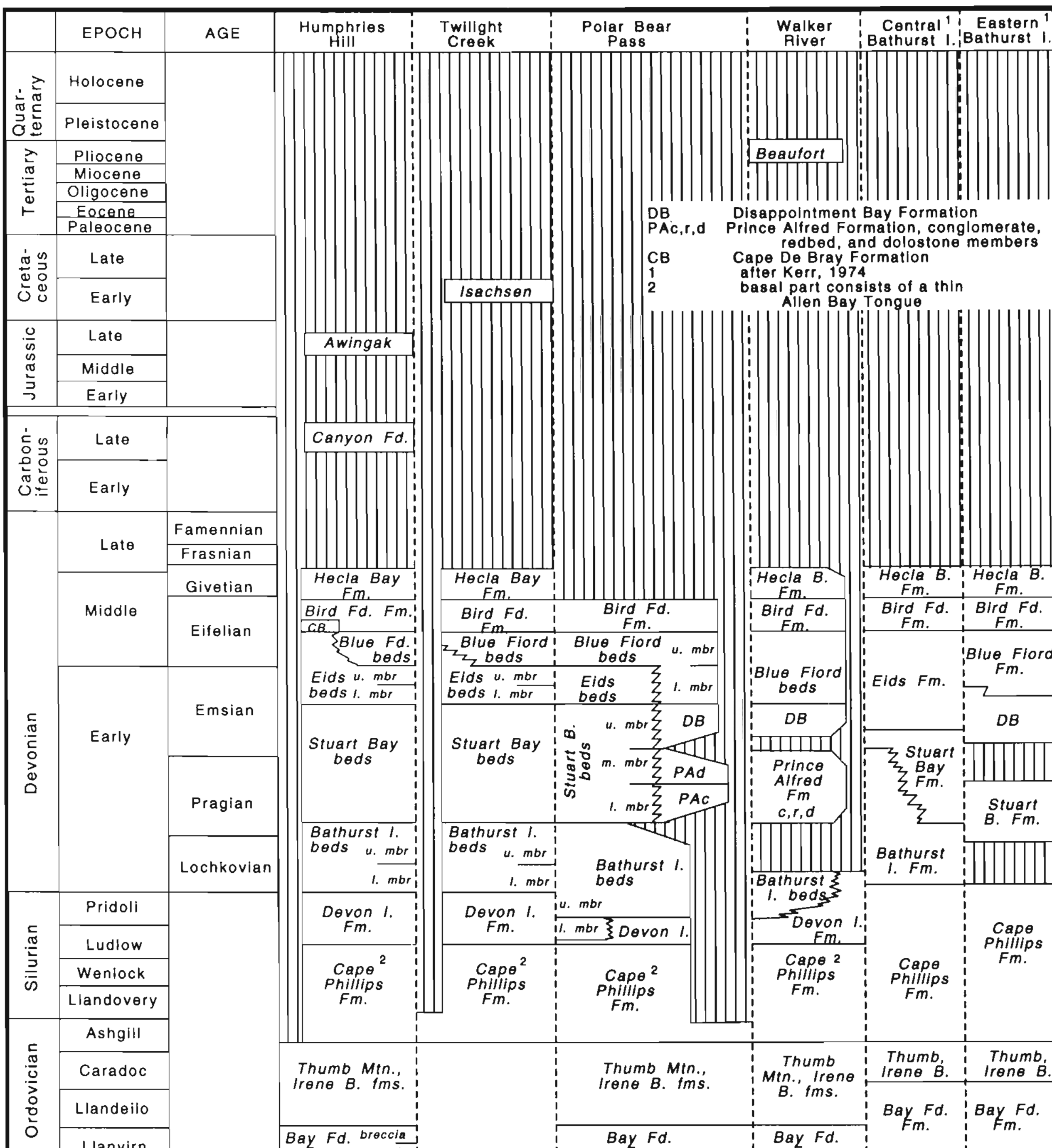
On northern Devon Island, the type Devon Island Formation (Thorsteinsson, 1963) is interposed between platformal carbonates of the underlying Douro and overlying Goose Fiord formations and represents significant platform retreat during late Ludlow time. These platformal carbonates are absent on Bathurst Island, and Kerr (1974) mapped all mudrocks stratigraphically below the Bathurst Island Formation as the Cape Phillips Formation (Table 1). However, late Ludlow platform retreat was manifested in the basin by a substantial reduction in carbonate sediment input, and thus, equivalent beds on Bathurst Island contain little, if any, carbonate, and are darker weathering than older calcareous or younger silty units. These beds appear to constitute a unit, herein referred to as the Devon Island Formation, throughout a large area of Bathurst Island.

The Devon Island Formation is 114 to 233 m thick, and well exposed in sections at Humphries Hill and Stuart River (Fig. 2, Sec. 1, 2). Its lower contact with the Cape Phillips Formation is abrupt, though conformable and marked by an abrupt change in weathering character, from yellowish, resistant weathering, calcareous mudrocks below, to grey, recessive weathering mudrocks above. The Devon Island Formation is predominantly a very thin bedded and fissile, medium grey weathering (greyish black on fresh surfaces) mudrock. Other noteworthy features include abundant *Bohemograptus bohemicus* cf. *tenuis* in the basal part of the formation; dusky red stained, carbonate concretions; and numerous, scattered, pale yellowish orange weathering argillaceous dolostone beds. The upper contact with the Bathurst Island Formation is gradational over about 30 m, and its top is placed below the lowest resistant, thin bedded, yellowish weathering, laminated siltstone or fine sandstone, characteristic of much of the Bathurst Island beds. In the vicinity of Walker River and Polar Bear Pass, the upper contact is marked by the abrupt appearance of huge carbonate olistoliths that form the basal part of the Bathurst Island beds (Fig. 2; Harrison et al., 1993, Fig. 5). The unit is Ludlow to early Lochkovian in age, based on the occurrence of *Bohemograptus bohemicus* ssp. and other Ludlow graptolites above and below the lower contact, and *Monograptus* cf. *uniformis* and *M.* cf. *birchensis* in beds that are transitional with the overlying Bathurst Island beds.

BATHURST ISLAND BEDS

McLaren (1963a) subdivided a 1560 m thick flyschoid succession exposed along Twilight Creek into the Bathurst Island and Stuart River formations on the basis of three thin (23-61 cm) limestone and chert pebble conglomerates that he considered to represent the basal Stuart Bay Formation. He also noted that conglomerate beds pass laterally into limestone. McLaren's pebble beds are readily identifiable

Table 1. Stratigraphic correlation chart



along Twilight Creek, where they constitute the only occurrence in the type section. The lithological characteristics of the Stuart Bay Formation are essentially identical to those of the Bathurst Island, except for the sporadic occurrences of these discontinuous chert pebble conglomerates. Furthermore, it is clear that the conglomerates occur at different stratigraphic levels. In sections that lack conglomerates, such as Cut Through Creek, the base of the Stuart Bay has been placed at lowest stratigraphic occurrence of thin, chert-granule-rich, fossiliferous calciturbidites in an otherwise thick flyschoid succession that includes the Bathurst Island Formation. However, in some areas there is no readily discernable contact between the Bathurst Island and Stuart Bay formations, as recognized by Kerr (1974), who was forced to map these formations as a single unit over much of Bathurst Island. Further field studies will be required before a reasonable judgement can be made between practically separating the beds now included in these two units, or abandoning both names and naming a single formation. For this reason, the two units are referred to as the Bathurst Island and Stuart Bay beds.

The Bathurst Island beds, at Cut Through Creek, are divisible into two locally mappable members. The lower member is a 565 m thick, monotonous, thin bedded siltstone, fine sandstone, and silty mudrock. Fine sandstone forms about 30 to 40 per cent of the unit, and it is consistently more resistant and thick bedded than the finer grained lithologies. Flutes and tool marks were only observed on a few bedding surfaces; however, these were in talus, and transport directions could not be determined. The upper member is 440 m thick and is remarkably similar to the previous unit, except that 5 to 10 m thick, mudrock-siltstone/fine sandstone hemicycles occur, primarily in the basal part. The upper part of this member contains abundant styliolinids, well preserved trace fossils, and several conspicuous, pale yellowish orange weathering, tentaculitid-rich, argillaceous limestone beds. The upper member contains abundant *Monograptus* cf. *yukonensis*, indicating a Pragian or perhaps earliest Emsian age (Fig. 2).

Near Polar Bear Pass, two members can be differentiated based on the occurrence of carbonate olistostromes (Sec. 3, Fig. 2; also see Harrison et al., 1993, Fig. 4, 5). The lower member is at least 190 m thick and consists of greyish yellow to yellowish grey weathering, thin bedded, laminated, calcareous siltstone, fine sandstone, and silty mudrock, a lithology which is pervasive throughout the Bathurst Island and Stuart Bay beds. Poorly preserved *Bohemograptus bohemicus* ssp. occurs near the base of Section 3 (Fig. 2) and well above the base of the member, suggesting that the basal contact of the Bathurst Island beds is diachronous, spanning late Ludlow to Lochkovian time (Fig. 2). The upper contact of this member is early Pridoli in age, based on the occurrence of *Monograptus* cf. *formosus* several metres below and above the upper contact. The member is 271.8 m thick, more widely distributed than the lower member (Fig. 2), and lithologically similar to the underlying member, except for the notable addition of carbonate olistostromes. Blocks are up to 150 m in diameter, clearly visible on many air photos (Harrison et al., 1993, Fig. 4, 5), and composed of well preserved, shallow water reef lithologies, including, from most to least common, stromatoporoid boundstone, coral-stromatoporoid rudstone, skeletal grainstone, oncoidal rudstone, skeletal wackestone, and cementstone. Kerr (1974) and Polan and Stearn (1984) originally included these within the Stuart Bay Formation, but most, if not all, of these blocks occur in the Bathurst Island Formation, and olistoliths of this size are unknown in the type section. Numerous other olistostromes occur throughout this member and in the overlying Stuart Bay beds; however, the largest blocks appear to be confined to the upper member of the Bathurst Island beds (Harrison et al., 1993, Fig. 4, 5). The upper contact is placed below the first occurrence of chert pebble to granule conglomerate, a common component of the Stuart Bay beds. The occurrence of abundant *Monograptus* cf. *yukonensis* graptolites indicates a Pragian age for the upper contact (Fig. 2).

STUART BAY BEDS

Because of reasons discussed above, McLaren's (1963a) original name for this unit is used informally, pending further field work in the area. At the type section, along Twilight Creek, the unit is 597.8 m thick and consists mainly of thin bedded, yellowish grey to greyish yellow, calcareous siltstone, similar lithologically to the underlying Bathurst Island beds, but with the addition of coarse grained, chert-rich sandstone, and minor pebble beds. Uncommon limestone olistostromes, nodular chert-rich limestone, and well preserved plant fragments are common throughout the formation; the latter are particularly abundant where siltstone and mudrock are interbedded with chert-pebble conglomerates. Where these conglomerates do not occur, as in Cut Through Creek, the base of the Stuart Bay beds is placed below the occurrence of thin, chert-granule-rich, fossiliferous calciturbidites.

Near Polar Bear Pass, three members are recognized. All three members contain a large proportion of yellowish weathering siltstone, fine sandstone, and mudrock that characterizes coeval rocks at Twilight Creek. The lower member is 231.1 m thick and contains numerous, laterally discontinuous chert pebble and some boulder orthoconglomerates that comprise channel-fill units up to 40 m thick and 250 m wide. Conglomerates are stratified, poorly to well sorted, and clasts are well rounded and commonly imbricated with angles greater than 20° (Fig. 3). Large accretionary surfaces, internal diastems, and slump structures occur in some channel fills near Polar Bear Pass, but channel fills are less numerous and thinner and wider in profile at Stuart River and Humphries Hill. Conglomerate lithoclasts consist of, from most to least common, dark grey or black laminated chert, graptolitic limestone concretions, laminated limestone, skeletal wackestone, bioturbated lime mudstone, skeletal grainstone, rudstone, yellowish orange weathering dolostone, and stromatoporoid boundstone. Thin to medium bedded, chert-granule-rich calciturbidites and well preserved plant fossils are also notable characteristics of this member.

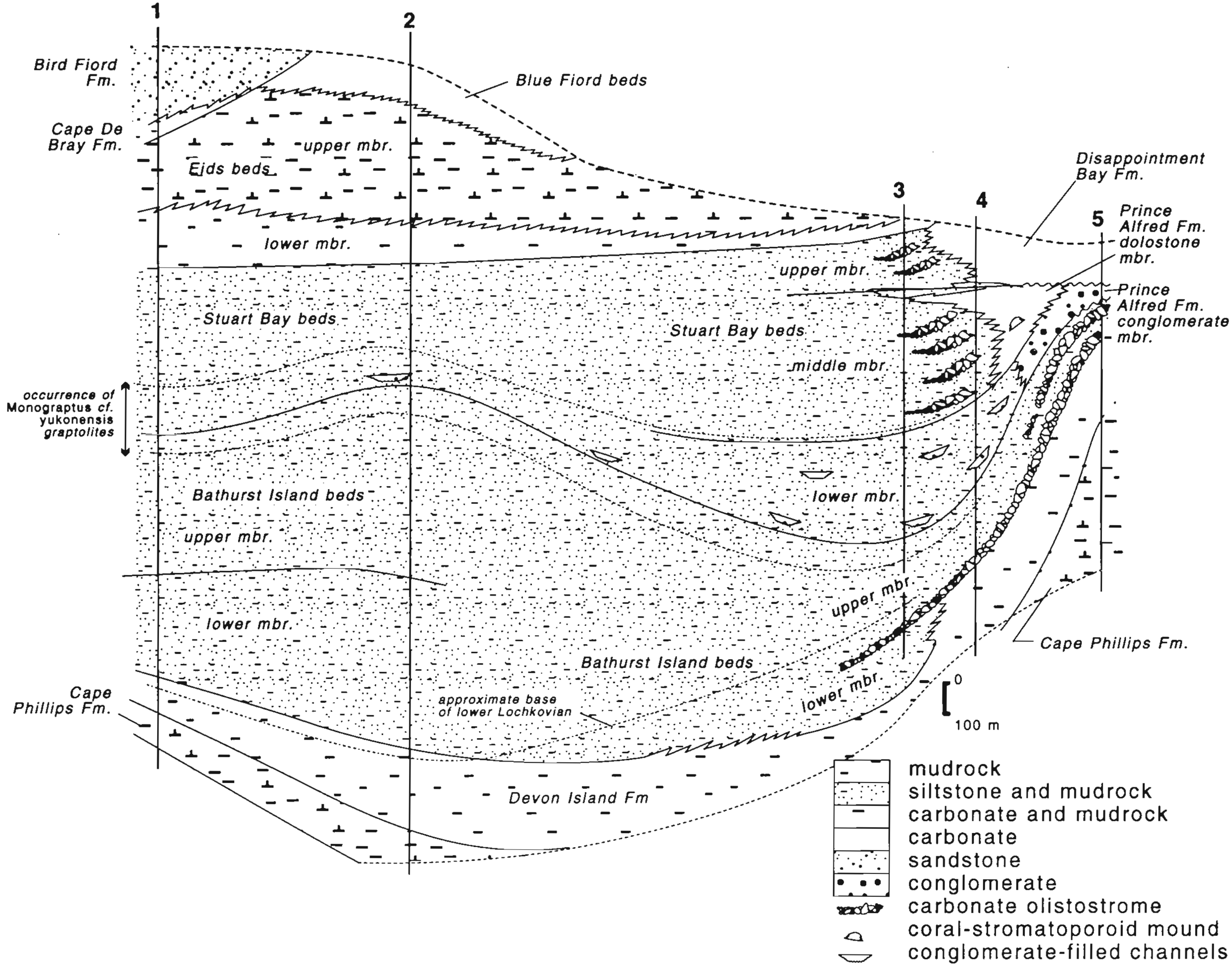


Figure 2. Summary cross-section of the Ashgill to Eifelian succession of central and eastern Bathurst Island. Section locations are shown in Figure 1. Thicknesses of the lower part of Section 5 is based on Kerr (1974).

The middle member is 473.7 m thick and is much like the lower member; however, limestone beds with well preserved brachiopods, corals, and stromatoporoids occur instead of chert pebble conglomerates. Some fossils appear to be in place, particularly the brachiopods, in contrast to the lower member, where fossil material is essentially all allochthonous. The top 10 m of this member consists of medium bedded, mature, medium grained, quartz arenite. The two-holed crinoid, ?*Gasterocoma bicaula*, in the upper part of this member, suggests an Emsian or early Eifelian age.

The upper member is 158.3 m thick and consists of yellowish weathering siltstone, fine sandstone, silty mudrock, and numerous thick carbonate olistostromes, with lithoclasts composed of fossiliferous rudstone and floatstone. Some beds contain well rounded, matrix supported, medium sand sized, quartz grains; and slump structures are very common throughout the member. The upper contact is abrupt and placed below pale yellowish brown, petroliferous, black, nodular chert-rich dolostone and mudrock, forming the lower part of the Eids beds.

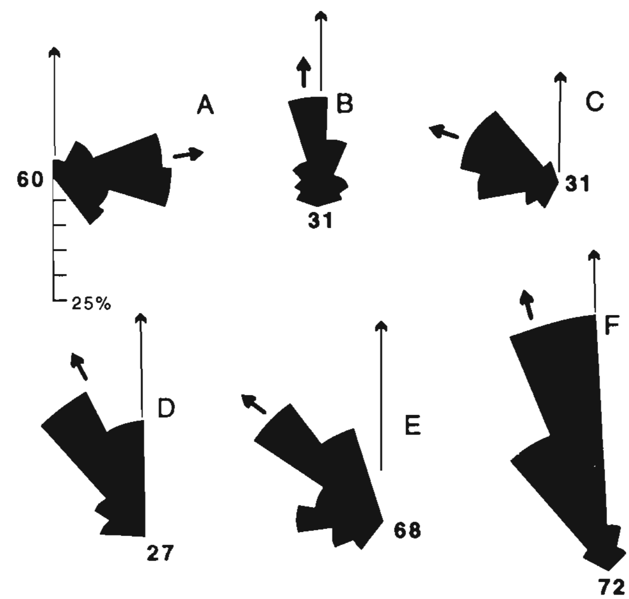


Figure 3. Paleocurrent circular histograms of the Stuart Bay beds and the Canyon Fiord and Prince Alfred Bay formations. **A.** Clast imbrication in the lowest Canyon Fiord Formation, Humphries Hill. **B.** Clast imbrication in the Stuart Bay beds, 3 km due south of Section 3, Figure 2. **C.** Clast imbrication in the Stuart Bay beds, 4 km due south of Section 3, Figure 2. **D.** Clast imbrication in the Stuart Bay beds, 3 km north-northwest of Section 4, Figure 2. **E.** Clast imbrication in the Prince Alfred Formation, conglomerate member, Section 5, Figure 2. **F.** Clast imbrication in Stuart Bay beds, Section 4, Figure 2. Numbers below the plots are total readings for each locality. Short arrows show average transport direction, while long arrows show true north.

PRINCE ALFRED FORMATION

The Prince Alfred Formation was originally named on northern Devon Island (Thorsteinsson, 1963) and is here applied to a syntectonic carbonate and clastic succession deposited along the Boothia Uplift in eastern Bathurst Island. Rocks herein included in the Prince Alfred Formation were assigned by Kerr (1974) variously to Disappointment Bay, Stuart Bay, and Bathurst Island formations. Conglomerate, redbed, and dolostone members are differentiated, and these facies are equivalent to one another.

The conglomerate member, measured only near Polar Bear Pass (Sec. 5), is a 30 m thick, moderately well rounded, carbonate and chert conglomerate, consisting of, in order of decreasing abundance, lithoclasts of laminated carbonate, light yellow brown weathering dolostone, laminated chert, and fossiliferous dolostone. Clasts range from granule to boulder size and show clast imbrication (Fig. 3). Less than 1 km to the west, conglomerates are finer grained, more thinly bedded, and interbedded with laminated silty, microbialite-rich, mudcracked dolostone. North of Walker River, boulder and carbonate pebble conglomerates grade northward into the red siltstone member, locally conglomerate- and dolostone-rich, and then to conglomeratic dolostone of the carbonate member. These facies transitions occur within 8 km.

The dolostone member at Polar Bear Pass (Sec. 4, 5 and vicinity) consists predominantly of medium bedded, light grey weathering dolostone, medium grey weathering fossiliferous limestone, minor fine grained, quartzose sandstone, and chert-rich conglomerate. Stromatoporoid-coral mounds occur locally, although they are poorly exposed. Farther west, these carbonates grade to interbedded carbonate olistostromes, styliolinid mudrock, calcareous sandstone, and cherty paraconglomerates, representing slightly deeper water facies. Farther west, at Section 3 (Fig. 2), coeval units consist of intercalated siltstone, silty mudrock, limestone, and carbonate olistostromes of the middle member of the Stuart Bay beds. Near Walker River the carbonate member is 265 m thick and consists of interbedded, thin to medium bedded, microbialite-rich, finely crystalline, light grey weathering, mudcracked dolostone and well sorted, horizontally and low angle cross-stratified, granule- and chert-pebble-rich dolostone. Conglomerates are particularly common toward the top of the unit, in contrast to the succession at Polar Bear Pass, where conglomerates occur primarily near the base.

The redbed member is poorly exposed near Walker River and to the north (Harrison et al., 1993, Fig. 4). It is thin bedded, flat and cross laminated, and is interbedded with light grey to pink, silty dolostone, and poorly sorted, angular to moderately well rounded chert granule conglomerate. Similar lithologies are exposed 15 km to the north of Walker River, where they are overlain by the conglomerate member. The Disappointment Bay in many places rests with angular discordance on all members of the Prince Alfred Formation.

Two-holed crinoids, ?*Gasterocoma bicaula*, have been collected from the upper part of the Prince Alfred Formation near Polar Bear Pass suggesting an Emsian or early Eifelian age for the upper part of this formation. The facies change of the lower conglomerate member with *Monograptus* cf. *yukonensis*-bearing mudrocks and carbonates of the lower and middle members of the Stuart Bay beds suggests a Pragian age for the lower part of the Prince Alfred Formation.

DISAPPOINTMENT BAY FORMATION

This formation was not examined in detail. It consists of light grey weathering, thick bedded, vuggy, fenestral, medium crystalline, locally bitumen-rich dolostone. Tepee structures, with primary isopachous, fibrous, dolostone-filled cavities, and oncolite rudstone occur locally. In the Polar Bear Pass area, this unit can be shown to pass laterally into the upper member of the Stuart Bay beds (Harrison et al., 1993, Fig. 5), which has yielded abundant two-holed crinoid ossicles, ?*Gasterocoma bicaula*, suggesting an Emsian or early Eiofelianage for this unit.

EIDS BEDS

The Eids Formation was originally established on southern Ellesmere Island for a unit of thin bedded, calcareous shale and siltstone (McLaren, 1963b). Its age was later established as early Emsian (Uyeno, 1990), indicating that the type Eids is older than the Eids Formation on Bathurst Island, which was dated by McGregor and Uyeno (1972). We have thus referred to this unit as the ''Eids beds'', pending redefinition.

Two members occur in the western part of the study area. The lower member is 17.2 to 26.2 m thick in the vicinity of Stuart River, and consists of thin to very thin bedded, light bluish grey to yellowish grey, calcareous, petroliferous, tentaculitid- rich mudrock. Farther west, the unit thickens to about 50 m and consists of interbedded black mudrock, chert, limestone concretions, and thin bedded, laminated limestone. At Humphries Hill, 67.5 m of phosphatic, tentaculitid, fissile, dark grey, petroliferous mudrock occurs. Thin carbonate olistostromes and plant fragments are also notable occurrences.

The upper member consists of apparent westward facing, sigmoidal and oblique clinoforms that, in section, comprise coarsening-upward hemicycles 5 to 15 m thick. Thin bedded argillaceous, tentaculitid, blue-grey weathering limestone and mudrock occur in the lower part, while medium to thick bedded, parallel and swaley laminated limestone occurs in the upper part of these hemicycles. Soft sediment deformation and pull-aparts are well developed throughout all parts of the hemicycles. In the upper part of this member, hemicycles are thinner, more thickly bedded, and contain a greater proportion of benthic invertebrate fossils than in the lower part. The upper contact is gradational and diachronous with the overlying Blue Fiord beds (Fig. 2). This unit ranges in age from late Emsian to early Eifelian, based on conodont work (McGregor and Uyeno, 1972).

BLUE FIORD BEDS

The Blue Fiord Formation was originally named by McLaren (1963b) on southern Ellesmere Island. However, this unit is entirely older (Uyeno and McGregor, 1972; Uyeno, 1990) than the Blue Fiord Formation as mapped by Kerr (1974) on Bathurst Island. Kerr's Blue Fiord Formation is, in fact, a correlative of the lower two members of the Bird Fiord Formation on northern Devon and southern Ellesmere islands (de Freitas and Mayr, 1992). We have thus used the name informally, i.e., "Blue Fiord beds", pending formal redefinition.

The unit is 179.0 to 241.0 m thick near Stuart River and composed of 5 to 20 m thick, shallowing upward hemicycles (Fig. 2, Sec. 2). The lower parts of these cycles are petroliferous, bioturbated, finely crystalline, thick bedded lime mudstone and wackestone, and the upper parts are thick bedded, locally bituminous-rich, fossiliferous rudstone and grainstone.

North of Polar Bear Pass, the Blue Fiord beds are divisible into two, locally mappable units of unknown thickness (Harrison et al., 1993; Fig. 5). The lower member consists of thick bedded, pale yellowish brown weathering, petroliferous dolostone, containing abundant two holed crinoid ossicles. It is overlain by a thick bedded, light grey weathering, arenaceous dolostone, that is a probable correlative of the Blue Fiord beds near Stuart River. Northwest of Polar Bear Pass, the Blue Fiord beds contain substantial thick bedded, fine to coarse grained, dolomitic quartz arenite. The Blue Fiord beds are late Emsian to early Eifelian in age (McGregor and Uyeno, 1972; Mayr, 1980).

CAPE DE BRAY FORMATION

This unit represents the basinal correlative of the Bird Fiord Formation described below. It rests with apparent conformity on the calcareous mudrock and argillaceous limestone of the Eids beds and represents the oldest part of the orogen-derived clastic wedge on Bathurst Island. It consists of sparsely fossiliferous to unfossiliferous, medium grey, sandy, laminated, poorly consolidated mudrock with interbeds of sandstone, similar to those of the Bird Fiord Formation described below. The unit is very poorly exposed, and has only been mapped in the Humphries Hill area, west of the depositional limit of the Blue Fiord beds (Harrison et al., 1993; Fig.3). There, it is about 49 m thick and gradationally overlain by resistant sandstone of the Bird Fiord Formation. The unit thickens substantially westward, at the expense of the underlying Eids beds. The unit is probably early Eifelian in age (McGregor and Uyeno, 1972).

BIRD FIORD FORMATION

The Bird Fiord Formation was examined in detail only at Humphries Hill, where 732 m of the unit, although poorly exposed, was described in detail. The unit is predominantly a thin to medium bedded, well indurated, light olive brown to dusky yellow weathering, micaceous, medium grained, fossiliferous, bioturbated, calcareous sandstone. Thin bedded, sandy calcareous mudrock is common, particularly in the middle part of the formation, where it forms the basal part of numerous, 5 to 10 m thick, mudrock-sandstone hemicycles. These grade upward into unfossiliferous, mature, cross-stratified, sandstone with locally abundant plant fossils. The unit is early to late Eifelian in age (McGregor and Uyeno, 1972; Uyeno, 1990).

CANYON FIORD FORMATION

The unit is exposed extensively in the vicinity of Humphries Hill. Although no paleontological information is available, the unit is lithologically very similar to the Canyon Fiord Formation on nearby northern Devon Island. It rests with angular discordance on Ordovician and Lower and Middle Devonian rocks. The basal 30 to 80 cm, overlying Ordovician carbonates near Humphries Hill (Harrison et al., 1993, Fig. 3), consists of well rounded, monomict, pebble to cobble orthoconglomerate. Lithoclasts show an eastward transport direction (Fig. 3) and appear to have been derived from the Devonian clastic wedge, perhaps from the resistant sandstones of the Givetian Hecla Bay Formation. Elsewhere, up to 10 m of normally graded, basal conglomerate beds contain carbonate lithoclasts that suggest derivation from the Blue Fiord and Eids beds and from the Bird Fiord Formation. These conglomerates grade upward into a poorly consolidated and poorly exposed moderate red siltstone and light bluish grey, sandy mudrock, estimated to be 200 m thick. The upper contact is the present day erosion surface.

AWINGAK FIORD FORMATION?

A small outlier occurs in the vicinity of Humphries Hill (Harrison et al., 1993, Fig. 3). The flat-lying beds unconformably overlie Eids beds and include about 20 m of interbedded, poorly consolidated, brown, parallel laminated mudrock and calcareous, quartz arenite. A limonitic paleosol and calcified sandstone concretions are notable occurrences in this unit. A preliminary palynological study indicates that these outcrops are Late Jurassic or Early Cretaceous in age (D. McIntyre, pers. comm., 1992).

ISACHSEN FORMATION

This unit, only mapped along Stuart River (Harrison et al., 1993), is 60 m thick and consists of mature, poorly consolidated, fine grained, light to very light grey weathering, horizontally and vaguely cross-stratified quartz arenite. Two lignite-grade, approximately 1 m thick, coal seams occur in the upper part, while grey weathering mudstone and limonitic sandstone concretions occur in the basal part of the exposure.

BEAUFORT FORMATION

A substantial area southwest of Walker River is covered by about 20 m of poorly consolidated sands that are tentatively referred to as the Beaufort Formation of Pliocene age (Harrison et al., 1993, Fig. 4). These sands are medium grade, brown weathering, cross-stratified, locally peat bearing, and contain logs up to 2 m long and 15 cm in diameter.

DISCUSSION

Late Ordovician carbonate platform step back initiated deep water sedimentation through most of Bathurst Island. This event is marked by deposition of graptolitic and cherty mudrock of the Cape Phillips Formation over relatively shallow water carbonate of the Allen Bay Formation. Flyschoid siliciclastics began to accumulate in Ludlow time near Polar Bear Pass, but not until Lochkovian time in the vicinity of Stuart River and Twilight Creek, suggesting that the clastics prograded northward or that there was some contemporaneous submarine topography. The main depocentre migrated westward toward central Bathurst Island during Lochkovian time, where more than 1 km of fine grained siliciclastics of the Bathurst Island beds accumulated. Siliciclastics continued to accumulate through the Early Devonian, but were punctuated in the Pragian and Emsian by episodes of coarse clastic and carbonate sediment influx derived from the rising Boothia Uplift or from carbonate platforms attached to this uplifted terrane. The most spectacular of these mass-flow deposits occurred during Pridoli and Lochkovian time, during deposition of the upper member of the Bathurst Island Formation. Some Pridoli blocks are up to 150 m in diameter, indicating derivation from a platform with at least this relief over the contiguous basinal deposits. The lack of exposed, coeval, shallow water, platformal carbonates on Devon Island and the lack of similar blocks along the Barlow Inlet carbonate platform on Cornwallis Island preclude derivation from these areas. Carbonate mass flow beds may have been derived more locally, from an isolated carbonate platform that developed along the rising Boothia Uplift, but was subsequently bevelled during main Pragian movement of the Boothia Uplift on easternmost Bathurst and northern Devon islands. Abundant carbonate lithoclasts that do not have an origin in the Cape Phillips or older formations were likely derived from this eroded carbonate platform succession.

Pragian diastrophism, perhaps, caused a third shift in the depocentre, this time toward the eastern part of Bathurst Island so that 863 m of flyschoid siliciclastics accumulated together with transported shallow water carbonates and continental conglomerates (Stuart Bay beds). Uplifted carbonate and fine grained, terrigenous mudrocks and cherts were shed from northeast or west? trending fault scarps and formed an apron of fanglomerates (Prince Alfred Formation, conglomerate member). At Polar Bear Pass, the continental-to marine-facies transition is extremely abrupt and probably northwest facing, while it is somewhat less abrupt and north? facing near Walker River. In the latter area, nearshore reworking of syntectonic clastics was greater, and a somewhat lower gradient fluvial system resulted in the deposition of nearshore, conglomeratic dolostones and redbeds of the Prince Alfred Formation. Clasts of chert and conglomerate were transported basinward in high concentration flows or density currents to form lenticular, inversely graded and imbricated orthoconglomerates and normally graded, sheet-like, cherty calciturbidites, respectively (Stuart Bay beds).

During the younger stages of uplift, supply of syntectonic clastics lessened, and a carbonate platform formed about the uplift (carbonate member of the Prince Alfred Formation). This facies belt was narrow, rimmed locally by coral stromatoporoid mounds, and shed mostly carbonate material to the basin.

Further movement of the uplift occurred during Emsian time, prior to deposition of the Disappointment Bay Formation. However, the amount of deformation and uplift was probably much less than that which occurred during the Pragian. The Disappointment Bay Formation was deposited with angular discordance on the Prince Alfred and older formations; the hiatus at the base of the formation is probably represented by the mature quartz arenite at the top of the middle member of the Stuart Bay beds, Section 3 (Fig. 2). Flyschoid sediments continued to accumulate basinward, and the depocentre, which had developed during the early part of Stuart Bay deposition, appears to have had little influence on sediment thicknesses during this time.

The Blue Fiord-Eids succession represents a rapidly prograding carbonate platform succession, showing particulary well developed bedforms which characterize regressive system tracts elsewhere, including downlap surface (lower member of the Eids beds), sigmoidal to oblique progradational clinoforms (upper member of the Eids beds), and concordant topset beds (Blue Fiord beds). This sequence is overlain by the Cape De Bray-Bird Fiord Formation sequence, representing similar shelf to basin depositional environments, although clastic supply during this time was substantial.

The Canyon Fiord Formation represents deposition within a small, fault bounded basin, whose location was controlled by pre-existing thrust faults. Sandstone lithoclasts were probably derived from the Hecla Bay and Bird Fiord formations and transported eastward, while carbonate lithoclasts were derived from the Blue Fiord beds and transported basinward in mass flows.

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New field observations on the geology of Bathurst Island, Arctic Canada: Part B, structure and tectonic history

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Abstract: Structural and stratigraphic studies and geological mapping were conducted in four key areas of the salt-based Parry Islands Fold Belt and adjacent Boothia Uplift on Bathurst Island in 1992. Newly discovered exposures of Carboniferous, Jurassic(?) and Lower Cretaceous strata have been located within the lower Paleozoic fold belt in the Stuart River and Humphries Hill areas. These faulted outliers provide new constraints on the timing of salt-involved conjugate thrusting and new evidence for a possible mid-Carboniferous phase of extension involving reactivated slip on pre-existing thrusts. Elsewhere, Pragian, nonmarine conglomerates within the Walker River and Polar Bear Pass areas of eastern Bathurst Island provide part of the new evidence for the earlier of two phases of uplift and related eastward-directed, thin-skinned thrusting along the western margin of Boothia Uplift. North trending folds and interference structures of the second phase may be mid-Tertiary in age.

Résumé : On a réalisé en 1992, dans l'île Bathurst, des études structurales et stratigraphiques et des travaux de cartographie géologique dans quatre secteurs clés de la zone de plissement de Parry Islands, dont la base repose sur des couches salifères, et du soulèvement de Boothia adjacent. Des affleurements nouvellement découverts de strates du Carbonifère, du Jurassique(?) et du Crétacé inférieur ont été localisés dans la zone de plissement du Paléozoïque inférieur, située dans les régions de la rivière Stuart et de la colline Humphries. Les nouveaux renseignements apportés par ces lambeaux de recouvrement faillés permettent de mieux définir l'âge des chevauchements conjugués impliquant des couches salifères, et fournissent de nouveaux indices relatifs à une nouvelle phase de distension peut-être survenue au Carbonifère moyen, qui aurait réactivé les déplacements suivant des failles chevauchantes préexistantes. Ailleurs, des conglomérats non marins d'âge Praguien contenus dans les régions de la rivière Walker et du col Polar Bear (Polar Bear Pass) dans l'est de l'île Bathurst fournissent aux chercheurs une partie des indices relatifs à la plus ancienne des deux phases de soulèvement et à un charriage connexe de couverture dirigé vers l'est, suivant la marge occidentale du soulèvement de Boothia. Des plis de direction générale nord et des structures d'interférence appartenant à la seconde phase pourraient dater du Tertiaire moyen.

INTRODUCTION

A program of stratigraphic and structural studies was initiated by the Geological Survey of Canada on Bathurst Island in 1992 (Fig. 1). The tectonic and structural aspects of this program have three principal objectives: to provide an improved understanding of the deformational style of all stratigraphic units above the sub-Bay Fiord evaporite; to determine the age and kinematic relationships between the Boothia Uplift in eastern parts of the island and the intersection of this belt with the salt-based Parry Islands Fold Belt to the west; and to provide a revised 1:250 000 scale geological map of the entire island. The field component of the program is estimated to require three summer seasons. The present account is preliminary. It summarizes the principal findings obtained in four key areas of the island that were investigated in July and August of 1992. The areas are: two located within the Parry Islands Fold Belt (Stuart River and Humphries Hill areas), one within the Boothia Uplift (Walker River area), and the fourth along the intersection between the two structural provinces (Polar Bear Pass area). The results of stratigraphic and sedimentological studies in these areas are summarized in our companion paper (de Freitas et al., 1993).

The perpendicular intersection of the eastward trending Parry Islands Fold Belt with the northward trending, cratonic Boothia Uplift on eastern Bathurst Island is one of the most spectacular geological features of its size in all of Arctic Canada. Its discovery and original description by Fortier and Thorsteinsson (1953) and McNair (1961) followed directly from the availability of complete air photographic coverage of the area. The region was a significant objective of geological field activities during Operation Franklin in 1955 (Fortier, 1963) and of Operation Bathurst Island in 1963-64 (Temple, 1965; Kerr 1974). Recent reviews, including a synthesis of available seismic data, are provided by Fox (1985), Okulitch et al. (1986, 1991), and Harrison et al. (1991).

RESULTS

Stuart River area

The Stuart River area embraces a 215 km^2 area lying along the hinge of the Stuart Bay Anticline east of the head of Stuart Bay (Fig. 2; Table 1). The general structural pattern established by Kerr (1974) consists of two parallel anticlinal hinges, each cored by the Cape Phillips Formation, with a narrow faulted axial syncline mostly covered by Quaternary drift. Strata on the north limb of the northern anticline are mirrored in width and thickness by strata lying on the south limb of the southern anticline. These panels comprise a continuous conformable succession 3000 m thick embracing all units from the Cape Phillips Formation to the Hecla Bay Formation. Stratigraphic nomenclature follows that of Kerr (1974) with new descriptions and recommended revisions provided in de Freitas et al. (1993).

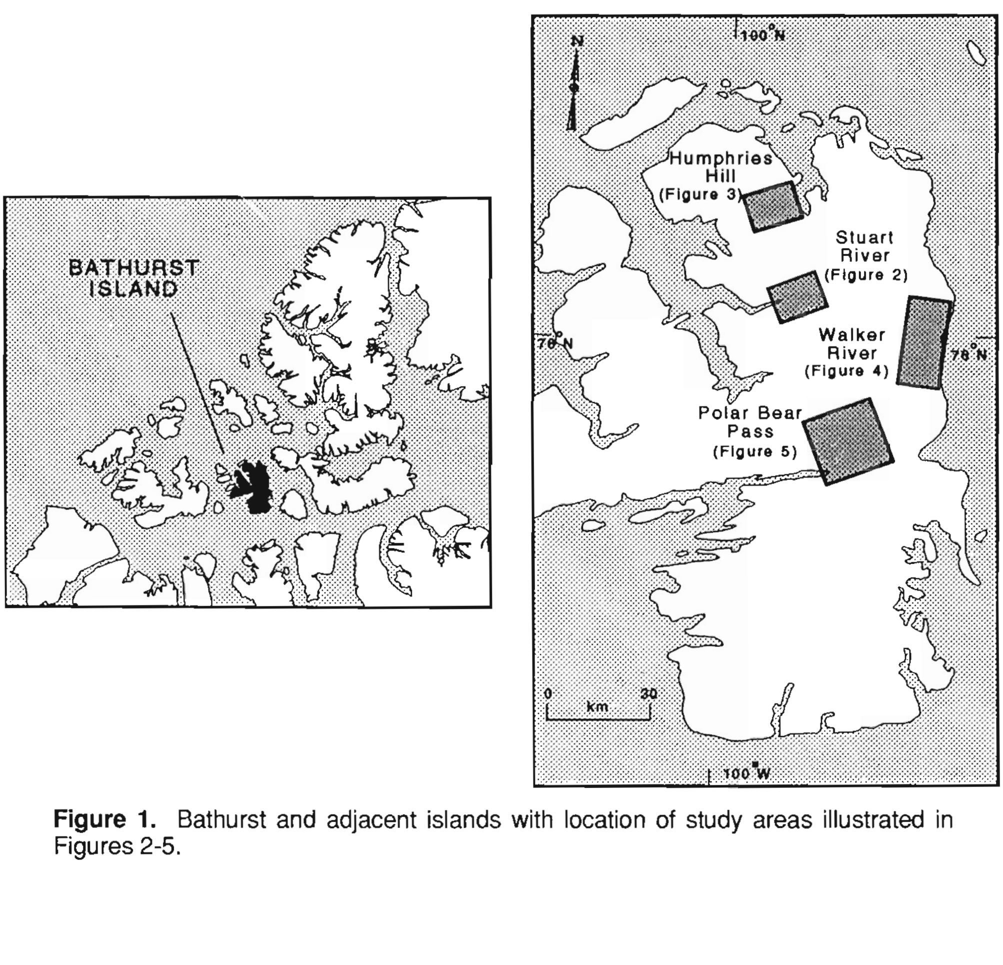

Figure 1. Bathurst and adjacent islands with location of study areas illustrated in Figures 2-5.

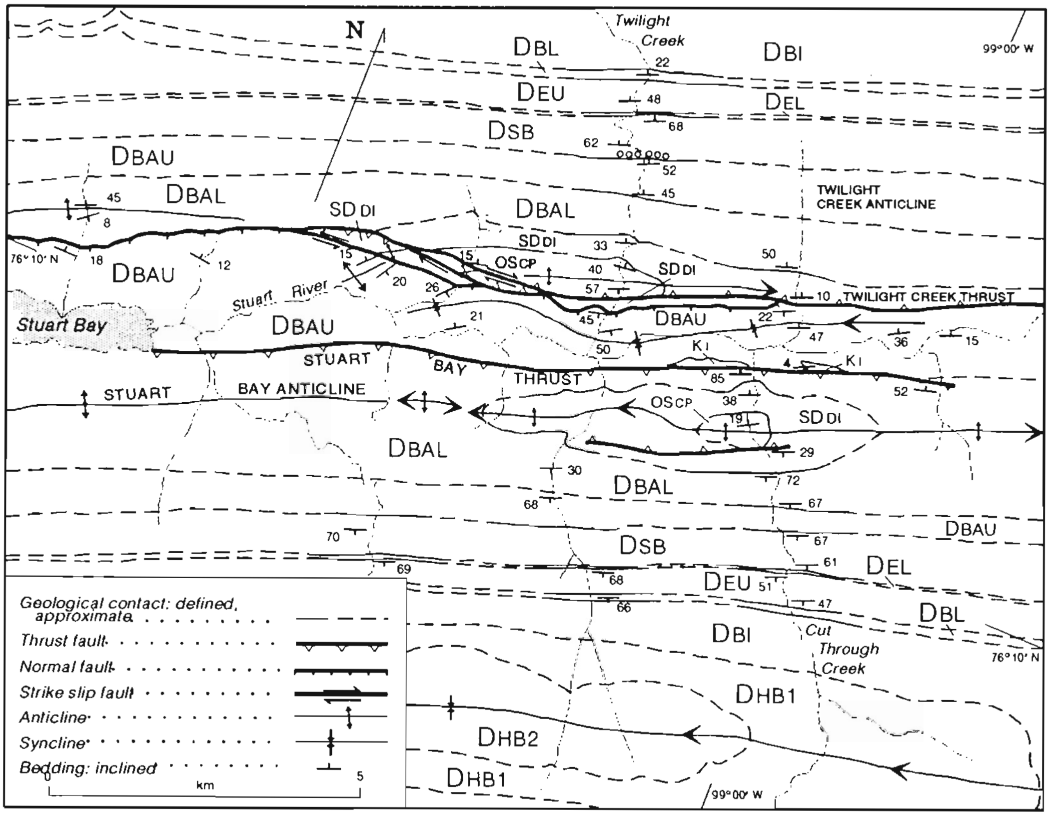

Figure 2. Geology of the Stuart Bay structure, Stuart River area, north-central Bathurst Island (NAPL air photos A16203-28, 56).

The structural style of the region is relatively simple overall but is kinematically complex in detail. Two inward-vergent thrust anticlines located north and south of Stuart River are separated by an intervening upright syncline cored by the exposed upper Bathurst Island beds. Exposure of the Cape Phillips Formation, at the base, is limited to the apex region of two faulted periclinal culminations. The southern culmination point on the Stuart Bay Anticline is located 400 m west of Cut Through Creek and exposes upper Llandovery and younger strata. The anticline is oversteepened and locally overturned on the north-facing limb. This narrow limb lies in the hanging wall of the steep north-vergent Stuart Bay Thrust which, traceable as far west as Stuart Bay, places the lower member of the Bathurst Island on gently dipping beds of the upper member[1]. In two places, the hanging wall Bathurst Island beds are in apparent thrust contact with tilted Isachsen Formation. The south-facing back limb features a small, south-vergent thrust and thousands of metres of beds with upright bedding dips falling in a narrow range between 61 and 74°.[2] However, above the base of the Eids unit (a local detachment surface where it is exposed), measured dips progressively decrease to 47°.

[1] Like most faults on Bathurst Island, the surface trace is covered. Sense of motion must be inferred from map relationships and mesoscopic structures located in the footwall and hanging wall.

[2] Out of the total number measured, only a selection of representative bedding attitudes are shown in Figures 2-5.

An important feature of the back limb strata are kink bands at all scales from hand specimens to map scale kinks involving hundreds of metres of section and nearly uniform dips between kink band boundaries. Outcrop and hand specimen scale kinks are especially well developed in platy limestones, siltstones and shales of the Devon Island, Stuart Bay and Eids formations. Closely spaced fractures along kink band boundaries in these rocks have produced a characteristic talus comprising pencil- and dagger-shaped rock fragments.

To the north, the culmination on the Twilight Creek Anticline features Upper Ordovician (Ashgill) and younger strata. The southward-vergent Twilight Creek Thrust emerges at surface on the narrow southward-facing limb. On Twilight Creek, this fault places the Wenlockian part of the Cape Phillips Formation on medial Devon Island beds. The footwall shales possess a northward-dipping axial planar cleavage, an array of outcrop-scale southward-vergent minor folds and contractional slip planes, many lying on bedding surfaces, with slip lineations parallel to the southerly direction of thrust transport. The northward-dipping back limb includes thousands of metres of beds with dips of 45 to 69° and shallower dips above the middle of the Eids.

The Twilight Creek Anticline is offset to the west by three oblique faults. The two easterly fault strands appear to be dextral linkages between the Twilight Creek Thrust and an unnamed northward-vergent thrust located to the west on the north limb of the Twilight Creek Anticline. The most western

Table 1. Table of formations (to accompany Figures 2-5)

QUATERNARY	
Q	*undivided diamictite, gravel, mud, peat*
PLIOCENE(?)	
$\mathbf{T}_{B}$	*BEAUFORT FORMATION: compacted sand, wood, peat*
LOWER CRETACEOUS	
$\mathbf{K}_{I}$	*ISACHSEN FORMATION: quartz sand, coal*
UPPER JURASSIC(?)	
$\mathbf{J}_{A}$	*AWINGAK FORMATION(?): quartz sandstone, brown shale, concretions*
CARBONIFEROUS(?)	
$\mathbf{C}_{C}$	*CANYON FIORD FORMATION: redbeds, sandstone, conglomerate, minor shale*
MIDDLE DEVONIAN	
$\mathbf{D}_{HB}$	*HECLA BAY FORMATION:* $\mathbf{D}_{HBL}$*: quartz sandstone, siltstone;* $\mathbf{D}_{HBU}$*: quartz sandstone*
$\mathbf{D}_{BI}$	*BIRD FIORD FORMATION: fossiliferous quartz sandstone, siltstone, shale, sandy limestone*
$\mathbf{D}_{CB}$	*CAPE DE BRAY FORMATION: grey shale; fossiliferous sandstone*
LOWER AND MIDDLE DEVONIAN	
$\mathbf{D}_{BL}$	*BLUE FIORD beds:* $\mathbf{D}_{BLU}$, $\mathbf{D}_{BL}$*: fossiliferous limestone; (arenaceous dolostone, quartz sandstone NW of Polar Bear Pass);* $\mathbf{D}_{BLL}$*: petroliferous dolostone*
$\mathbf{D}_{E}$	*EIDS beds:* $\mathbf{D}_{EU}$*: clinoformed platy argillaceous limestone, shale, minor siltstone;* $\mathbf{D}_{EL}$*: petroliferous shale, argillaceous limestone, chert*
LOWER DEVONIAN	
$\mathbf{D}_{DB}$	*DISAPPOINTMENT BAY FORMATION: vuggy fenestral dolostone*
$\mathbf{D}_{PA}$	*PRINCE ALFRED FORMATION:* $\mathbf{D}_{PAC}$*: conglomerate member;* $\mathbf{D}_{PAR}$*: redbed member;* $\mathbf{D}_{PAD}$*: dolostone member*
$\mathbf{D}_{SB}$	*STUART BAY beds: siltstone, shale, argillaceous limestone common tentaculitids and plant fragments;* $\mathbf{D}_{SBL}$, $\mathbf{D}_{SBM}$, $\mathbf{D}_{SBU}$*: lower, middle and upper members*
ooo	*conglomerate*
$\mathbf{D}_{BA}$	*BATHURST ISLAND beds: siltstone, very fine sandstone, shale, common tentaculitids and plant fossils;* $\mathbf{D}_{BAL}$, $\mathbf{D}_{BAU}$*: lower and upper members*
UPPER SILURIAN AND LOWER DEVONIAN	
$\mathbf{SD}_{BA}$	*BATHURST ISLAND beds:* $\mathbf{SD}_{BAU}$*: shale, siltstone, olistoliths;* $\mathbf{S}_{BAL}$*: siltstone, shale*
●	*olistolith*
$\mathbf{SD}_{DI}$, $\mathbf{S}_{DI}$	*DEVON ISLAND FORMATION: black graptolitic shale, common concretions*
UPPER ORDOVICIAN AND LOWER SILURIAN	
$\mathbf{OS}_{CP}$	*CAPE PHILLIPS FORMATION (includes tongue of ALLEN BAY FORMATION): argillaceous limestone and dolostone; graptolitic shale, bedded chert*
MIDDLE AND UPPER ORDOVICIAN	
$\mathbf{O}_{TI}$	*THUMB MOUNTAIN AND IRENE BAY FORMATIONS (undivided): limestone, argillaceous limestone, chert nodules*
LOWER AND MIDDLE ORDOVICIAN	
$\mathbf{O}_{BF}$	*BAY FIORD FORMATION:* $\mathbf{O}_{BFC}$*: limestone, dolostone;* $\mathbf{O}_{BFT}$*: carbonate tectonic breccia, chert concretions;* $\mathbf{O}_{BFE}$*: gypsum*

oblique fault, with a sinistral sense of offset, merges to the west with a southward-dipping sinuous normal fault that places the hanging wall uppermost Bathurst Island in tectonic contact with the lowermost Bathurst Island. To the east this same oblique fault is linked to a separate sinuous normal fault which, near the mouth of Twilight Creek, places hanging wall upper Bathurst Island beds over the footwall lower Devon Island Formation.

In conclusion, the principal phase of compressive deformation and erosion of the entire Stuart Bay structure appears to predate deposition of the Lower Cretaceous Isachsen Formation. Overall tectonic style is reminiscent of the inward-vergent salt-involved contractional structures imaged on seismic profiles of Melville Island (Harrison, 1991, in press). However, oblique strike slip faulting, not recognized on the Melville Island seismic data, provides a newly documented mechanism for the lateral transfer of slip on oppositely-vergent thrusts. The tectonic history of the Stuart Bay structure is complicated by a system of shallow-dipping extension faults. Apparent thrust faulting and tilting of the footwall Lower Cretaceous outliers may be related to the mid-Tertiary Eurekan Orogeny.

Humphries Hill area

The Humphries Hill area (200 km^2) lies along the complex faulted hinge of the Purcell Bay structure between Young Inlet and Purcell Bay on northern Bathurst Island (Fig. 3). The structural style established by Kerr (1974) consists of two inward-vergent thrust panels exposing the Cape Phillips and younger beds, an axial syncline and fold trend-parallel normal faults, two of which envelope "diapiric" Bay Fiord evaporites. Kerr's (1974) cross-section of this structure indicates that the evaporites may have reached the surface in a horst-like structure bound by outward-dipping extension faults that detach on the evaporites beneath the adjacent inward-transported thrust panels to the north and south.

Particular attention in this area was devoted to geological relations within the axial evaporite belt, the potential subdivision of undivided Devonian beds, and the nature of slip along the various map scale faults. A significant early discovery was the occurrence of previously unreported Carboniferous(?) and Jurassic(?) exposures including the Canyon Fiord Formation redbeds, sandstone and conglomerate (in three areas) and shale and quartz sandstone possibly assignable to the Awingak Formation (Jurassic) in a fourth locality (Fig. 3). These units are briefly described in de Freitas et al. (1993). The Canyon Fiord Formation in the easternmost exposure is preserved in a shallow dish-shaped syncline along Young Inlet and inland for more than 3 km.

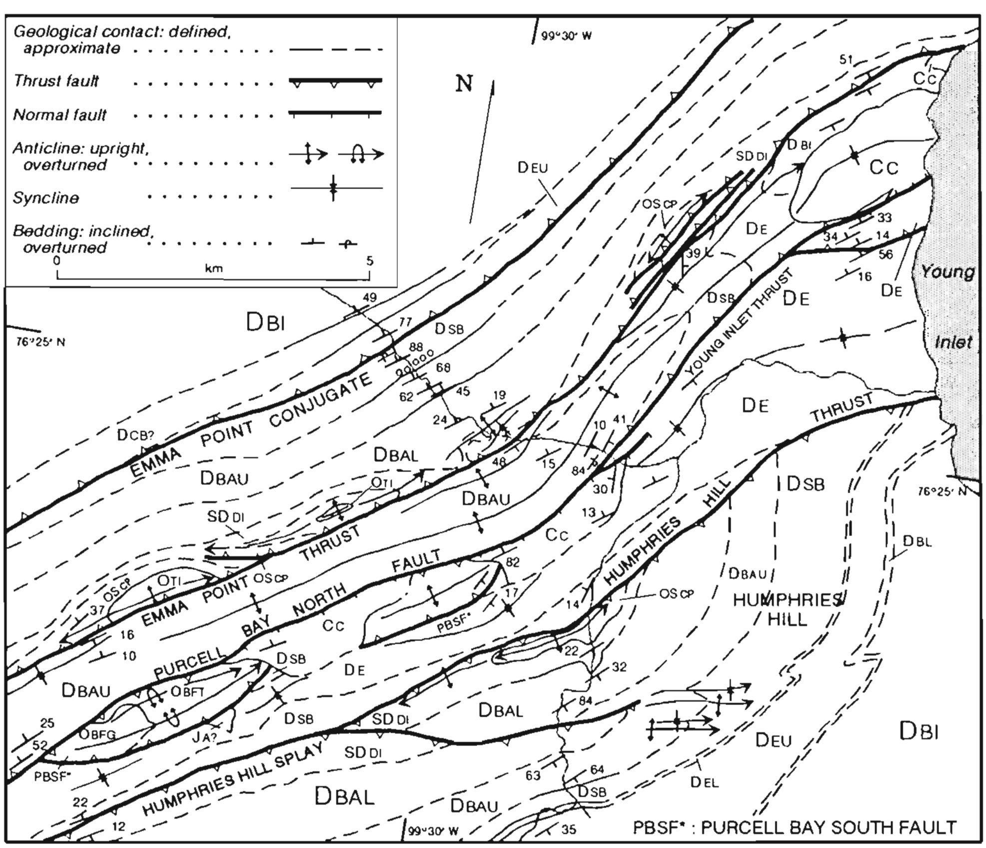

Figure 3. Geology of the Purcell Bay structure, Humphries Hill area, northern Bathurst Island (NAPL air photos A16203-33, 51).

The formation lies with pronounced angular unconformity on the previously folded Eids and Bird Fiord formations. The second and third outcrop belts of the Canyon Fiord Formation are located 7 and 12 km west of the closest part of Young Inlet. A channelized clast-supported conglomerate occurs at the base where the formation lies with angular unconformity on uppermost Stuart Bay and Eids beds and nonconformably on unstratified dolostone tectonic breccias (tectonized upper Bay Fiord Formation). The Canyon Fiord Formation dips to the northwest at 8 to 30° and is truncated on the northwest side by the Purcell Bay North Fault which places the formation in tectonic contact with the upper Bathurst Island beds. The possible Awingak Formation exposure lies 15.5 km southwest of Young Inlet. About 20 m of beds here disconformably overlie poorly exposed Eids strata and, in the northwest, are faulted against Bay Fiord dolostone breccias.

Field studies also identified previously unreported exposures of the Thumb Mountain and Irene Bay formations as well as the Devon Island Formation, upper and lower Bathurst Island beds, Stuart Bay, upper and lower members of the Eids, and the probable northeastern depositional limit of Cape De Bray Formation as more fully described by de Freitas et al. (1993). Seven major sub-Carboniferous tectonic elements exposed in the region are considered from northwest to southeast in the following paragraphs.

A northwestward-vergent thrust, the Emma Point conjugate, places medial Stuart Bay beds over medial Eids strata indicating 300 to 400 m of stratigraphic throw. Hanging wall beds are progressively steeper approaching the talus-covered fault and are locally overturned to the southeast in a manner consistent with the assumed direction of thrust vergence. The fault trace is low in the Stuart Bay to the southwest and northeast and rises into the upper part of the formation near the middle of the map.

A stratigraphic throw of 1000 to 1800 m is recorded on the southeastward-vergent Emma Point Thrust. Observed along strike within each of three faulted hanging wall periclines are: the medial Thumb Mountain Formation (Caradoc?) placed over the upper Bathurst Island beds (Pragian); the Wenlockian part of the Cape Phillips Formation also placed on Pragian beds; and the upper Devon Island Formation (Lochkovian) placed on uppermost Stuart Bay and Eids beds (Emsian) and, farther to the northeast, on Bird Fiord and Canyon Fiord formations.

Continuing across strike to the southeast is an unnamed syncline/anticline pair that plunges to the northeast. Both folds are truncated by erosion on the sub-Carboniferous uncon- formity. The anticline possesses a southeastward-facing asymmetry consistent with the assumed vergence on the underlying Young Inlet Thrust which, in the northeast, places uppermost Bathurst Island and Stuart Bay beds on the footwall Eids. The fault bifurcates to the east in the Eids. The northern strand merges with a normal fault with downthrown Canyon Fiord Formation on the north side. At least 10 m of source proximal conglomerate occur in the lower part of the Canyon Fiord in the immediate fault hanging wall (de Freitas et al., 1993). Farther to the southwest, the Young Inlet Thrust must lie hidden in the footwall of the Purcell Bay North Fault which carries the Bay Fiord and Canyon Fiord formations at surface.

The Bay Fiord beds are exposed in a plunging box fold with dolostone tectonic breccias conformably overlying tectonized gypsum in the faulted culmination. Both northward- and southward-facing limbs are overturned on the two underlying Purcell Bay faults. In three places, thrust-faulted segments of the northwestward-vergent Purcell Bay North Fault pass laterally into normal faulted segments that define the preserved limit of Canyon Fiord Formation outliers. In contrast, the southward-vergent Purcell Bay South Fault (PBSF on Fig. 3), with up to 2200 m of stratigraphic throw, features hanging wall Ordovician breccias and footwall Emsian beds all of which are erosionally truncated by the same Canyon Fiord cover. The allochthonous Bay Fiord evaporites and breccias, together with the throughgoing footwall syncline in the Eids unit, represent the axial line of the entire Purcell Bay structure.

The remaining half of the Purcell Bay structure features the northwestward-transported Humphries Hill Thrust. Wenlockian Cape Phillips Formation in the faulted hanging wall culmination is thrust over intensely deformed uppermost Stuart Bay beds. To the northeast, the Humphries Hill Thrust cuts upsection as far as the upper Eids unit on a regional scale lateral ramp and, to the southwest, the north-vergent Humphries Hill splay dies out upsection in a zone of chevron folding.

In conclusion, the Purcell Bay structure provides an excellent example of salt-involved folding and thrusting which elsewhere is only suspected from industry seismic profiles. Mapping points to significant detachments below the Bay Fiord evaporites, and within the Devon Island Formation, Stuart Bay and Eids units. The remaining units have behaved as fundamentally rigid layers. The high level of emplacement of Bay Fiord evaporites has been facilitated by contractional slip on an underlying conjugate outward-vergent thrust pair rather than by gravitational inversion and diapirism. The overlap of one of these faults by mid-Carboniferous(?) redbeds provides a reliable upper age limit on salt-involved thrusting. Two other thrusts have apparently been reactived as younger normal faults. Block rotation and possible syndepositional slip provides a mechanism for the preservation of the same Carboniferous deposits.

Walker River area

Field work in this area embraces a 300 km^2 area on eastern Bathurst Island between the headwaters of Walker River in the south and Airstrip Point in the north (Fig. 4). Limits of mapping have been defined by Kerr's (1974) northward trending and refolded Queen's Channel and Driftwood Bay anticlines. Kerr's mapping indicated a widespread angular unconformity below the Disappointment Bay Formation and possibly significant local disconformities below the Stuart Bay and Griper Bay formations.

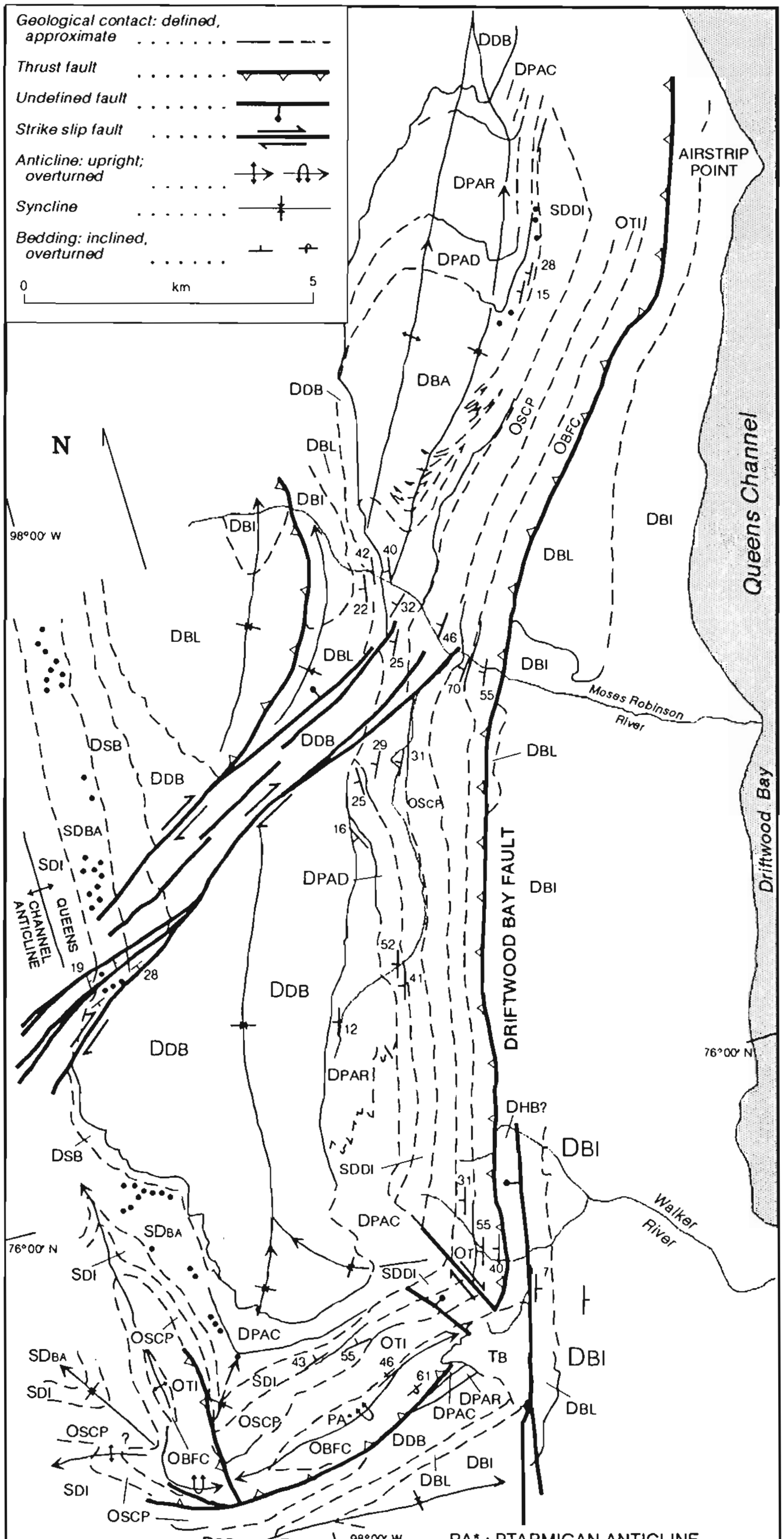

Figure 4. Geology of the Driftwood Bay structure and general Walker River area, eastern Bathurst Island (NAPL air photos A16202-15, 17, 50, 52).

The existence of a disconformity below Griper Bay strata could not be confirmed during the five days of field work in this area. Revision mapping has however confirmed the unconformity below the Disappointment Bay Formation and found these strata to rest variably on beds ranging from the Llandovery (lower Cape Phillips Formation) to the Emsian (Stuart Bay beds). There is no known unconformity below the Stuart Bay beds. Instead these basin facies strata, exposed only on the eastward-dipping limb of the Queen's Channel Anticline, are traceable to the south on airphotos (and presumably also to the east beneath younger cover) into shallow marine dolostones, nonmarine redbeds and conglomerates of the Prince Alfred Formation (de Freitas et al., 1993). The Prince Alfred Formation is exposed in two large outcrop areas on the westward-dipping, hanging wall panel of the Driftwood Bay Fault. In these areas, the formation lies disconformably on either the Bathurst Island beds or the Devon Island Formation. The large southern outcrop belt includes boulder grade conglomerate and a mappable facies change to the northeast into dolostone and redbeds. Separate conglomerate, dolostone and redbed members have also been mapped in the northern exposures.

Large parts of the Walker River area, previously assigned by Kerr (1974) to either the Stuart Bay Formation or an undivided Lower Devonian unit, are actually age equivalent (and also partly older) than the Bathurst Island beds in the type section on Twilight Creek. In consequence these various exposures are also mapped as Bathurst Island strata in spite of significant differences in depositional facies as described by de Freitas et al. (1993). A significant feature of the unit is erosionally resistant house-sized carbonate olistoliths. All new and previously identified occurrences are plotted on the accompanying map (Fig. 4). The greatest concentration lies along the eastward-facing limb of the Queens Channel Anticline. However, additional blocks occur 7 km north of Moses Robinson River and west of the Driftwood Bay Fault. In this same general area, a facies transition between lower Bathurst Island beds in the west and the upper Devon Island Formation in the east can be inferred from air photographs.

Also significant, for tectonic interpretation is the recognition of a belt of upper Bay Fiord Formation lying conformably beneath the Thumb Mountain Formation and traceable for at least 20 km at the lowest stratigraphic levels above the Driftwood Bay Fault. About 130 m of upper Bay Fiord beds are exposed. It is assumed that the top of the evaporite member of the Bay Fiord Formation represents the basal detachment surface occupied by the Driftwood Bay Fault. The southernmost exposures of the Bay Fiord Formation lie within an eastward trending part of the Driftwood Bay structure. Here the Bay Fiord Formation, exposed in the hinge of the doubly-plunging Ptarmigan Anticline, is overturned to the south on a southeastward-vergent thrust that places these Middle Ordovician strata over lower Cape Phillips, Disappointment Bay and Prince Alfred formations. The entire structure is overlapped by sub-horizontal and undisturbed Beaufort Formation (Pliocene).

The Queen's Channel Anticline and Driftwood Bay sheet are bisected north to south by a broad flat syncline, cored over a wide area by the Disappointment Bay Formation. Regional scale uplift in the south of this structure has caused erosion of the Griper Bay, Hecla Bay, and Bird Fiord formations and the Blue Fiord beds-about 1200 m of strata. Near the centre of the map (Fig. 4), formational contacts are offset to the east by a system of northeastward-striking faults. The sense of motion on this fault array is uncertain but would appear to involve components of right-lateral displacement and eastward- or southeastward-directed shortening. A separate north-northwestward-striking fault, located near Walker River, may be a sinistral conjugate. Considering these oblique faults as a kinematically linked system, together with the northward-striking regional syncline and the Queen's Channel Anticline, the implied motion is contractional and eastward-directed on the Driftwood Bay Fault. Stratigraphic throw on this probable thrust is at least 1500 m.

The Driftwood Bay structure appears to have been created by three phases of shortening. The kinematics of the two earlier phases described in the present model are similar to those described by Kerr (1974) but differ in that all the faults (including the Driftwood Bay Fault) are now considered to flatten downward on the sub-evaporite detachment. First phase uplift and exposure of the Driftwood Bay thrust sheet in the Pragian provides a suitably emergent source area for proximal facies Prince Alfred Formation conglomerates shed in a northward and westward direction (de Freitas et al., 1993) into a coeval deep water realm. The second, southward-directed, contractional phase, like most of the mid-Paleozoic salt-based fold belt to the west, is linked to the origin of the westward trending Ptarmigan Anticline and its related underlying thrust and parallel footwall syncline. Components of left-lateral slip on the Driftwood Bay Fault would appear likely at this time. The northward trending Queen's Channel Anticline and the various unnamed structures within the regional syncline to the east were previously considered first phase folds by Kerr (1974). However, these structures display no evidence of shortening in the Pragian and clearly must have been formed by a separate phase of folding; coaxial with the first and perpendicular to the second but, like the second, also post-Eifelian in age. Some reactivated compressive slip on the Driftwood Bay Fault is also probable during phase three.

Deep-seated regional uplift in post-Eifelian time is also indicated by the erosional unroofing of approximately 1200 m of Middle and ?Upper Devonian strata from the syncline. This uplift must originate below the Bay Fiord Formation and, if synchronous with any of the thin-skinned phases of shortening, may point to the existence at depth of a complex linkage between surface faults and other deep-seated and undetected faults that flatten upward onto the sub-evaporite décollement.

Polar Bear Pass area

Investigations north of Polar Bear Pass embrace a 500 km^2 area lying along the intersection between the Parry Islands Fold Belt and the Boothia Uplift. Structures identified by Kerr (1974) include the eastward trending Bracebridge Inlet and Polar Bear Pass anticlines and the intersection of these folds and the intervening unnamed synclines with the northward

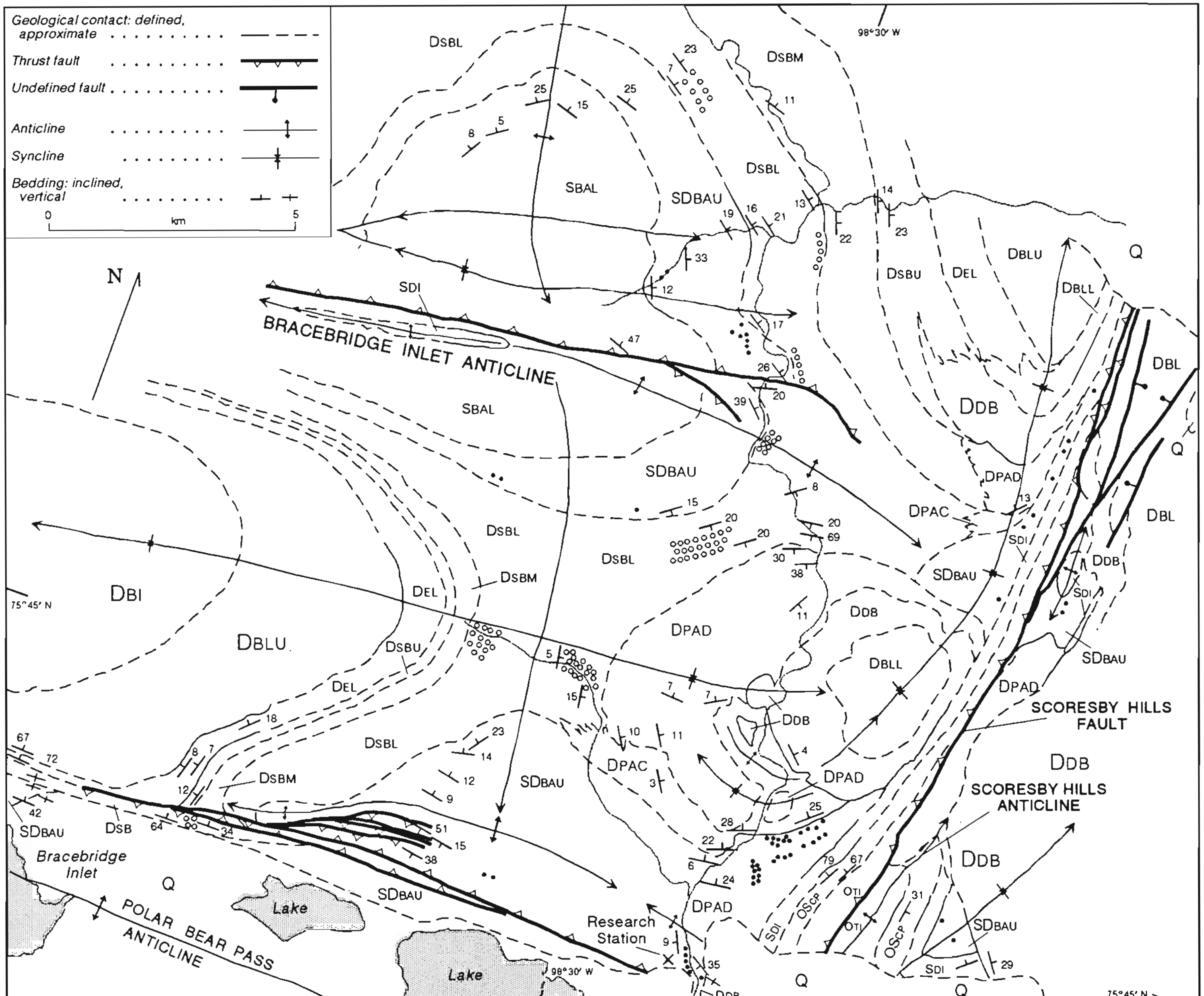

Figure 5. Geology of Scoresby Hill structure and surrounding Polar Bear Pass area, central Bathurst Island (NAPL air photos A16202-154, A16203-19, 63, 65, A16761-179).

trending Scoresby Hills Anticline. Particular interest in the area is attached to Kerr's (1974) mapping of stratigraphic overstep relationships in Lower Devonian strata on the west side of the Scoresby Hills Anticline, a potential facies change between Kerr's Bathurst Island and Stuart Bay formations at the eastern end of the Bracebridge Inlet Anticline, and an uncertain stratigraphic relationship between the Eids and Disappointment Bay formations. This region also features spectacular olistoliths (Polan and Stearn, 1984) in strata previously mapped by Kerr as Stuart Bay Formation.

The present field studies (Fig. 5) have demonstrated that the Polar Bear Pass region provides critical information concerning the depositional relationships between the slope and basin facies Devon Island Formation, Bathurst Island, Stuart Bay and Eids beds deposited over much of central and western Bathurst Island, and the coeval shelf-related strata of the Prince Alfred, Disappointment Bay and Blue Fiord formations. Preliminary biostratigraphic and lithostratigraphic evidence is provided in de Freitas et al. (1993).

The oldest strata occur at the south end of the Scoresby Hills structure where the Thumb Mountain Formation is exposed in the hinge of the Scoresby Hills Anticline, and also to the west where the same beds have been unroofed in the hanging wall of the southeastward-vergent Scoresby Hills Fault. Northward, this probable thrust features progressively younger strata of the Cape Phillips and Devon Island formations and Bathurst Island beds, preserved in the immediate hanging wall. An Early Devonian (Pragian?) phase of north-south folding is implied by stratigraphic relationships in the footwall of this thrust. The dolostone member of the Prince Alfred Formation lies with pronounced angular unconformity on the folded and eroded roots of the Scoresby Hills Anticline and on beds of ?Late Ordovician and younger age. The Prince Alfred Formation, in turn, is overstepped to the north and south by the Disappointment Bay Formation which rests on various Upper Silurian and Lower Devonian strata. The Prince Alfred Formation has also been subsequently folded around the same north-south axes and all formations, as high as the Blue Fiord Formation, have been involved in the shortening.

Farther west, the eastward-transported Scoresby Hills thrust panel is parallel to an asymmetric syncline with a steep eastern limb (to 79°) and a western limb featuring dips of not more than 30°. A plunge depression in the south is probably an interference structure produced by separate phases of shortening involving intersecting perpendicular fold axes. Important facies fronts in the Lower Devonian are evident on air photographs of the western limb of this syncline. At two separate localities, the nonmarine conglomeratic member of the lower Prince Alfred Formation grades westward into the lower part of the much thicker Stuart Bay beds which, in this area, contain numerous submarine conglomerate horizons and well preserved plant fossils. These conglomerates are cut out to the northwest by a dolostone member of the Prince Alfred which, in the north, grades and thickens into a middle member of the Stuart Bay beds. The Disappointment Bay Formation, in turn, oversteps the Prince Alfred dolostone member and rests variably on the Devon Island Formation or Bathurst Island beds on the eastern limb of the syncline. Basinward, the shelf facies Disappointment Bay Formation extends up to 2.5 km beyond the Prince Alfred shelf edge where it grades into an upper member of the Stuart Bay strata. Two facies of the Blue Fiord unit are recognized. The lower petroliferous dolostone member grades into a platy and comparably petroliferous facies of the Eids. The upper Blue Fiord is much thicker and as such has prograded westward out of the Polar Bear Pass region.

The Bathurst Island unit within the Polar Bear Pass area is divisible into two informal members. The lower member is exposed only in two culminations, one within the Bracebridge Inlet Anticline (Fig. 5), where it overlies the Devon Island Formation. In spite of similarities to the type section Bathurst Island beds, this member is entirely Late Silurian in age and therefore an age equivalent facies of the Devon Island Formation (de Freitas et al., 1993). An eastward facies transition of the lower Bathurst Island into the upper Devon Island Formation is inferred as the Bathurst Island upper member lies conformably on the Devon Island Formation at several localities on both limbs of the Scoresby Hills Anticline. These Bathurst Island beds were previously assigned to the Stuart Bay Formation by Kerr (1974). In fact, these beds are age equivalent to, and also partly older than, the entire type section Bathurst Island beds of the Stuart River area (de Freitas et al., 1993) and, despite major differences in depositional facies, are mapped as such on Figure 5. Large olistolith clusters are exposed in the immediate vicinity of the Polar Bear Pass research station and are scattered throughout the upper Bathurst Island member in a northward trending facies belt at least 14 km wide. Plant fossils occur in both the Silurian and Lower Devonian parts of the unit.

In conclusion, the Scoresby Hills structure represents the west side of a periodically emergent crustal block which persisted from the Ludlow to the Emsian. This structure defines the shelfward (eastern) depositional limit of the lower Bathurst Island beds in the Late Silurian, and the basinward extent of the Prince Alfred, Disappointment Bay and lower Blue Fiord shelf carbonates in the Pragian and Emsian. The uplift was submerged in the Pridolian and Lochkovian. Olistoliths of these ages are common in the upper Bathurst Island beds and appear to have been transported from a separate source area east of Polar Bear Pass.

Like the Driftwood Bay structure near Walker River, the Scoresby Hills structure bears evidence of three phases of compressive uplift and erosion. During the first phase of deformation in the Pragian, the hanging wall of the Scoresby Bay Fault was uplifted on an eastward-vergent thrust ramp detached on the sub-Bay Fiord evaporites. This linear uplift was part of an eastern provenance region for proximal facies Prince Alfred Formation conglomerates shed into coeval deep water realms to the west. The Bracebridge Inlet and Polar Bear Pass anticlines and the related thrusts have been created during a second shortening phase. This phase was perpendicular to the first and probably coeval with the rest of the Parry Islands Fold Belt. The Scoresby Hills Fault was reactivated, and other northward trending folds were created by a third phase of folding. This final phase was coaxial with the first and perpendicular to the second, but was also post-Eifelian in age like the second. Culminations and depressions throughout the region have been created by the interference of second and third phase fold axes.

Post-Eifelian northward trending folds and thrusts are common both in the Polar Bear Pass and Walker River areas. Folds of similar trend are common to the north within pre-Neogene and older strata of the Sverdrup Basin (Okulitch and Trettin, 1991). Because there is no other fold system known to account for the young northward trending folds on eastern Bathurst Island, it is reasonable to conclude that they are also mid-Tertiary in age.

ACKNOWLEDGMENTS

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Geological Survey of Canada Project 860006

The characterization of massive ground ice at Yaya Lake, Northwest Territories using radar stratigraphy techniques

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Robinson, S.D., Moorman, B.J., Judge, A.S., and Dallimore, S.R., 1993: The characterization of massive ground ice at Yaya Lake, Northwest Territories using radar stratigraphy techniques; in Current Research, Part B; Geological Survey of Canada, Paper 93-1B, p. 23-32.

Abstract: A series of ground penetrating radar (GPR) surveys were conducted in the vicinity of a massive ground ice body near Yaya Lake, Richards Island, Northwest Territories. The objective of this research was to delineate the lateral extent, thickness, and internal character of the massive ground ice underlying the region. A 16.5 m borehole log was used as point source verification of the GPR survey interpretations.

The base of the ice body was indicated on all of the profiles by a strong reflection varying in depth between 10 and 16 m. This corresponds to the contact observed in the borehole at 12.2 m depth. The top of the ice body was delineated at a depth varying between 2.0 and 13.0 m, overlain by a bedded coarse glaciofluvial layer. The massive ice body was found to be 420 x 210 m with a maximum thickness of 10 m.

Résumé : On a réalisé une série de levés au moyen de géoradars (GPR), à proximité d'un corps massif de glace dans le sol proche du lac Yaya, dans l'île Richards (Territoires du Nord-Ouest). L'objectif de ces recherches était de définir dans cette région l'extension latérale, l'épaisseur et le caractère interne de la glace massive dans le sol. On a employé un diagramme de forage de 16.5 m de long pour vérifier depuis une source ponctuelle les interprétations des levés par géoradar.

La base du corps massif de glace est indiquée sur tous les profils par une forte réflexion à des profondeurs variant entre 10 et 16 m. Elle correspond au contact observé dans le sondage à 12,2 m de profondeur. Le sommet de ce corps a été délimité à une profondeur variant entre 2 et 13 m, et il était recouvert par une couche de sédiments fluvio-glaciaires lités de granulométrie grossière. On a constaté que ce corps de glace massive mesurait 420 x 210 m et avait une épaisseur maximum de 10 m.

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INTRODUCTION

Massive ice bodies are commonly observed in near-surface materials of the Richards Island-Tuktoyaktuk Peninsula region (Mackay, 1971; Rampton and Mackay, 1971; Pollard and French, 1980; Mackay and Dallimore, 1992), generally reported to be overlain by poorly sorted fine grained sediments and underlain by coarser material. The occurrence of massive ground ice bodies overlain by coarse grained material is less common (Mackay, 1971; Mackay and Dallimore, 1992). As the Quaternary history of this region and the origin of much of the massive ice are still the topic of some debate, information concerning the character and genesis of massive ice bodies is required.

Although geotechnical drilling has been utilized successfully to study ground ice distribution, this method is expensive, time-consuming, and only provides limited point specific information. Ground penetrating radar (GPR) provides a relatively inexpensive, nondestructive, quick, and detailed method of subsurface geophysical profiling. When used in conjunction with limited drilling, ground penetrating radar aids in interpolation of geological contacts between boreholes, and allows detailed mapping of the extent, thickness, and internal characteristics of permafrost and massive ice bodies (Dallimore and Davis, 1987, 1992; Dallimore and Wolfe, 1988).

SITE LOCATION AND SURFICIAL GEOLOGY

The Yaya Lake region of south-central Richards Island (Fig. 1) is characterized by rolling topography between 1 and 76 m above sea level, with a median elevation of approximately 30 m. Thick deposits of unconsolidated glacial and deltaic sediments, and the remnants of the fluvial sediments from the Pleistocene Mackenzie Delta (Rampton, 1988) are overlain by variable thicknesses of glacial, lacustrine, and glaciofluvial sediments (Fig. 2). A number of steep-sided, flat-topped kame ridges rise up to 45 m above the shores of Yaya Lake and surrounding terrain, trending northeast-southwest (Ripley, Klohn and Leonoff International Ltd., 1973). These well drained ridges belong to the Early Wisconsinan Toker Point Member (Rampton, 1988).

Geotechnical drilling and geophysical studies in the region have shown many of the ridges to be cored with massive ice up to 20 m thick (EBA Engineering Consultants Ltd., 1975; Dallimore and Wolfe, 1988). The coarse grained glaciofluvial materials have been investigated by Canada Department of Indian and Northern Affairs as a potential source of aggregate material, as high quality granular resources in the region are rare. The presence of discrete ice bodies or lenses within gravel often limits the economic use of a deposit (Dallimore and Davis, 1992). Preliminary studies conducted by EBA Engineering Consultants Ltd. (1986) and Dallimore and Davis (1992) indicated the presence of appreciable amounts of aggregates with identified but unknown amounts of associated ground ice. The exploitation of similar deposits closer to Tuktoyaktuk produced dramatic thermokarst depressions and serious erosion near river channels (Thompson, 1992). Thus, prior to any exploitation of these aggregate deposits, a detailed knowledge of the amount and type of ground ice is required.

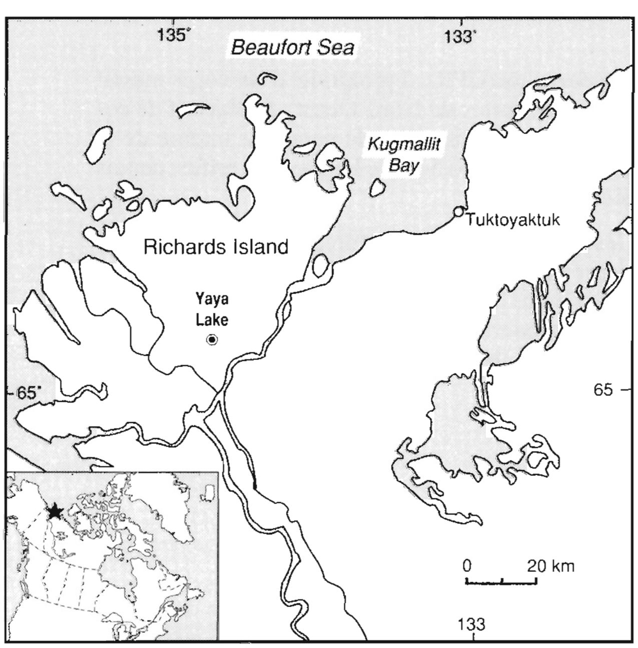

Figure 1. The location of Yaya lake within the Mackenzie Delta, Northwest Territories.

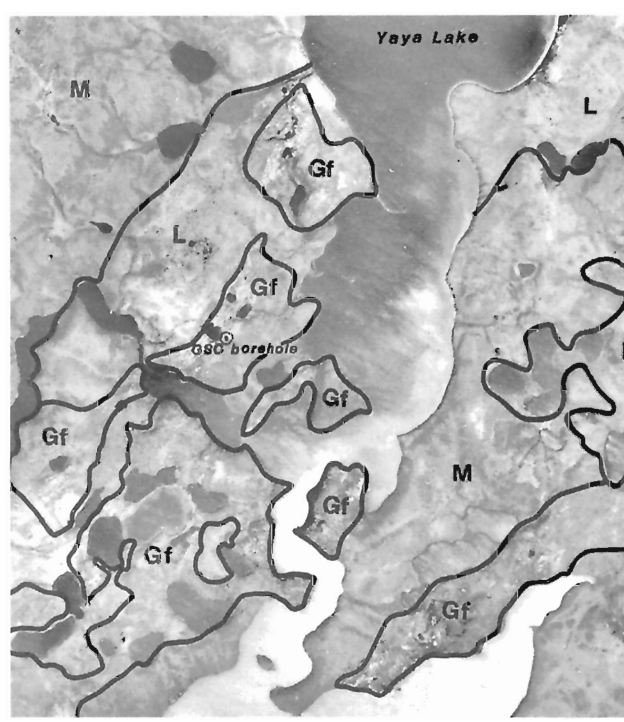

Figure 2. Yaya Lakes region aerial photograph with interpreted surficial geology: M - moraine; l - Lacustrine; Gf - glaciofluvial

METHODOLOGY

Ground penetrating radar technique involves the transmission of short pulses of high frequency electromagnetic (EM) energy into the ground. Measurement of the two-way travel time of returns reflecting from various interfaces within the ground allows the interpretation of subsurface characteristics. Massive ice and frozen coarse grained materials represent ideal transmission media due to the low electrical conductivities causing very little attenuation and enabling deep signal penetration. More complete details of survey methodology and GPR theory are given in Ulriksen (1982) Davis and Annan (1989) and Moorman (1990). Annan and Davis (1976), Dallimore and Davis (1987, 1992), Judge et al. (1991), and Robinson et al. (in press) discuss radar techniques used to map areas of massive ground ice and to study the internal structure of permafrost.

A total of eleven ground penetrating radar profiles were surveyed along Lines A, B, and C at the Yaya Lake study site in April and August 1990, and in September 1992 (Fig. 3). Robinson et al. (in press) studied the effect of increasing antenna frequency on the character of the profiles conducted in 1990, concluding that the surveys using the 50 MHz antennas were the most successful to observe the morphology and internal structure of the ice body. This report presents detailed interpretations of those earlier 50 MHz surveys and in addition, the new profiles surveyed in 1992.

The surveys utilized the pulseEKKO III and pulseEKKO IV GPR systems with 50 MHz antennas, a 400 volt transmitter, a constant 1 m step size between stations, and an antenna separation of 2 m. Topographic surveys allowed elevation corrections to be applied to the radar profiles. A 16.5 m borehole was drilled in April 1990 at the intersection of the two lines and is used for correlation between radar reflectors and geological contacts observed in the drill core.

RESULTS

Simple point averaging noise reduction filters and automatic gain control were the only processing enhancements applied to the profiles. The profiles are presented with vertical axes showing both two-way travel time and reflector depth. The depth scales were calculated using a simple single-layer velocity model with estimates of propagation velocities determined from known properties of sands and gravels under the given conditions (Annan and Davis, 1976), a velocity sounding conducted during the 1992 surveys, and by correlating depth of distinct reflections to horizons measured in the borehole. As a result propagation velocities of 0.14 m/ns and 0.12 m/ns were used to create the depth scales on the winter and summer profiles, respectively. Distances indicated on the horizontal axes of all profiles are in metres along the survey line.

The first return from all the radar profiles presented in this report represents the direct air wave, the signal that travels directly between the transmitter and receiver through the air, with a constant velocity of 0.3 m/ns. The second return represents the direct ground wave, the signal that travels between the antennas through the upper soil layer. The ground wave return time is dependent on the propagation velocity of the upper soil layer and the antenna separation.

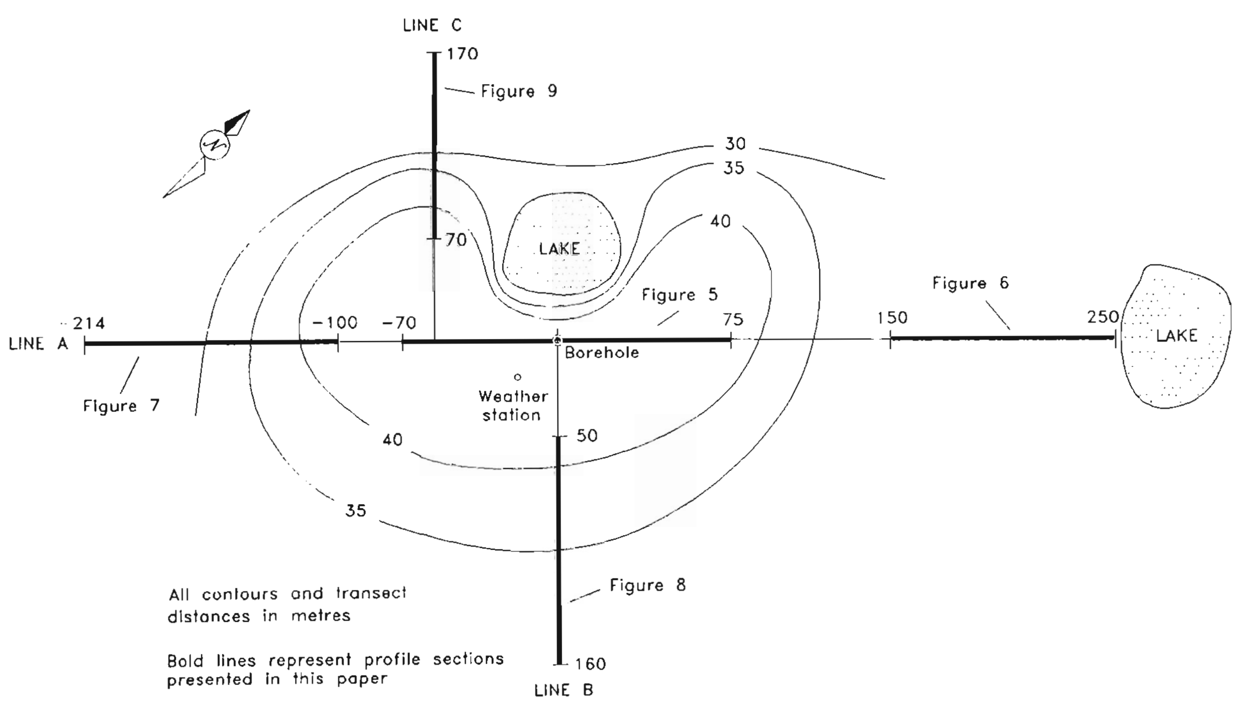

Figure 3. Yaya Lake study site map showing survey line and borehole locations.

Energy returned following these two initial returns generally represents reflections from interfaces with dielectric contrast at depth.

A borehole, located at at the 0 m position in Figure 5, revealed a complex stratigraphy beneath the ridge (Fig. 4). The upper 3.5 m, designated as Layer A, contains bedded sands and sandy silts with pebbles. Four distinct ice layers and an ice-rich silty clay are also found within this upper section. The entire section above the massive ice contact is considered here as one layer for ease of interpretation. Layer B extends from 3.5 to 12.2 m and consists of generally clean massive ice, which in some sections contains minor sediment bands or inclusions. Below 12.2 m, Layer C is a sandy-silty-clay till with clasts up to 1 cm in diameter and some ice-rich sections and ice lenses. Drilling was suspended at a depth of 16.5 m.

Selected sections of the ground penetrating radar surveys conducted along Lines A, B, and C are presented as Figures 6 to 10, below. These profiles best represent the internal characteristics of the massive ice body and the relationships between the ice and enclosing sediments. The 50 MHz antennas provided sufficient resolution for detection of the major contacts without the complication of reflections from minor interfaces as seen in 100 MHz surveys also conducted at the site (Robinson et al., in press).

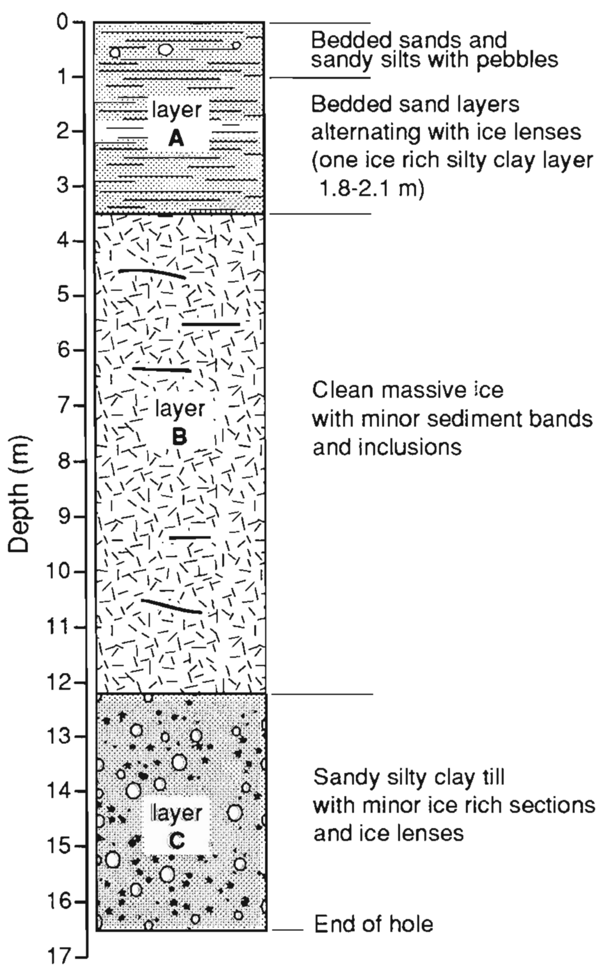

Figure 4. Simplified borehole log from the hole drilled at the intersection of lines A and B.

Surveys in the central portion of the ice body

The profile conducted along Line A in the vicinity of the borehole (Fig. 5) shows the internal structure of the ice body away from its edges. The inclusion of sediment bands within the ice body (layer B) provides continuous reflectors roughly parallel to the top and bottom of the ice, and indicates the wavy nature of the internal structure of the ice. The reflectors within the glaciofluvial overburden (layer A) are less continuous and more chaotic than those from within the ice. As the thickness of overburden increases towards the northeast (Fig. 5, +10 to +75 m), the strength of the returns from the ice below decreases as the signal attenuation in the sand and silt is greater than that in ice.

Reflections observed within the till (layer C) below the ice are of lower frequency and are considerably less continuous, running for less than 10 m, while those within the ice are 30-80 m long. The returned signal strength quickly decreases within a few metres below the base of the ice. This could be due to the fine grained nature of the underlying till attenuating the signal or an increase in the homogeneity of the material producing weaker reflections. Interfaces representing the greatest contrast in dielectric properties (i.e., between the base of the ice and the underlying till) produced the strongest returns. The interface between the ice and overlying coarse grained sediments produced a much weaker reflection, suggesting a relatively minor dielectric contrast.

Figure 5 is the only profile presented in this report conducted under winter conditions, thus lacking the active layer reflectors seen on other profiles. In this profile, the depth to the top of the ice body varies between 2.0 and 8.5 m below ground surface, with the base of the ice varying between 10 and 16 m in depth. These results correspond well with drill core information.

Figure 6 shows another section of Line A starting 150 m northeast of the borehole. In this profile it is seen that the ice body thins to between 2 and 4 m thickness in the northeast. Glaciofluvial overburden thickness reaches a maximum of 13 m near the 220 m position. Some sections show reflectors within the overburden intersecting the upper ice contact at oblique angles (i.e., near 205 m and between 230 and 240 m), suggesting that the sediments above are separated from the ice by an unconformity produced through thermal erosion.

Surveying was not continued along Line A past the 250 m position due to the steepness of a slope leading down to a thermokarst pond. It appears that the meltout of the ice body was responsible for the formation of this pond. The topographic relationship of the pond to the known location of the ice body at its side suggests that most or all of the ice below the lake has melted out. The surface appearance of the shore suggests that the thermokarst activity has abated and the adjacent slopes have stabilized.

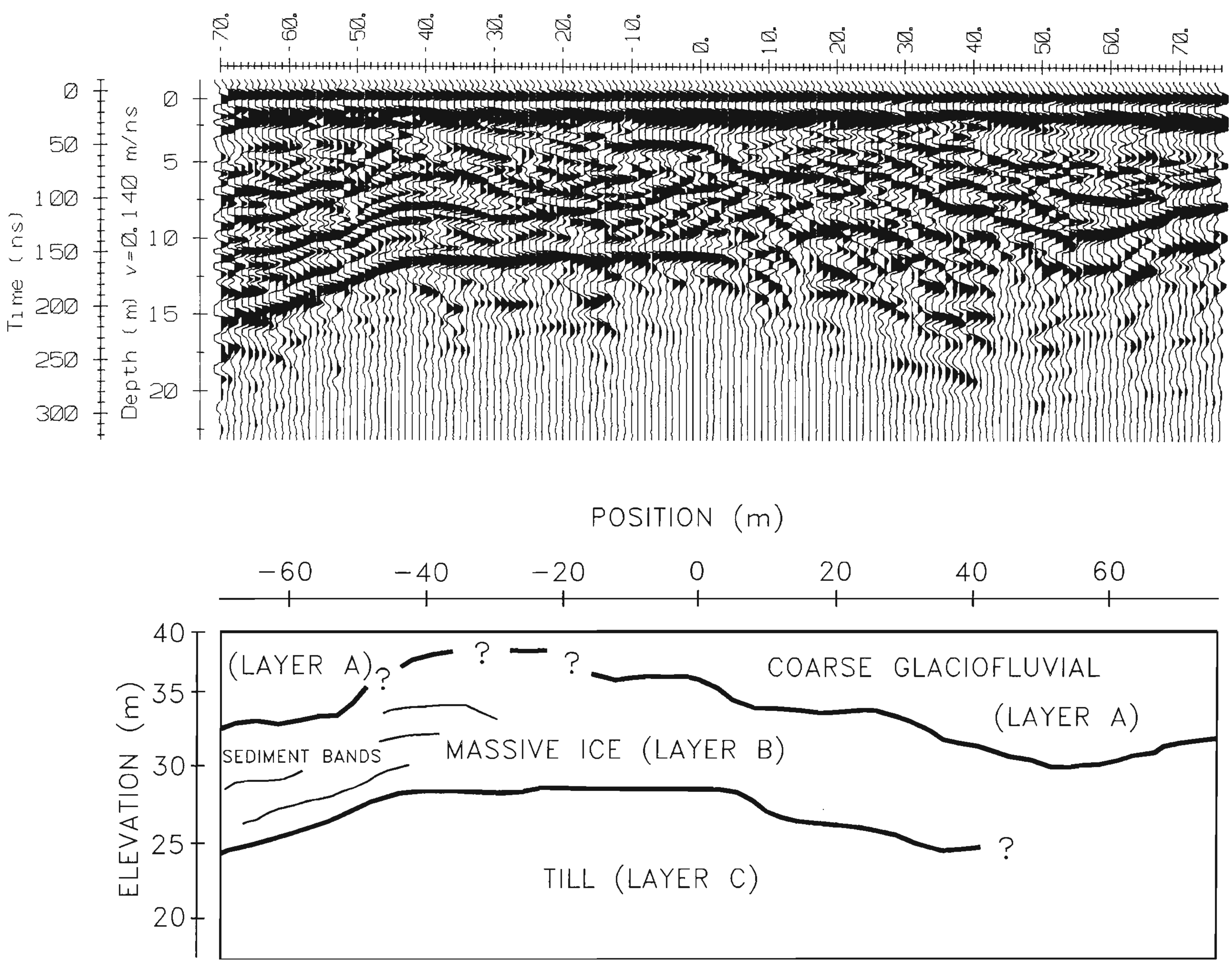

Figure 5. Line A - GPR survey (upper) and interpretation (lower) conducted in April 1990 showing internal structure of massive ice body. Borehole located at 0 m along profile. Note the lack of an active layer.

Surveying at the edge of the ice body

On the slope near the southwestern edge of the ridge (-100 to -150 m in Fig. 7), the ice body pinches out and reflectors are less continuous and appear to rise towards the surface to unconformably contact the upper boundary, with very weak returns coming from the underlying region. The frozen sediment beyond the edge of the ice body (-170 to -214 m) displays reflection characteristics very different from those observed within the ice. The returns from within these sediments are discontinuous and of varying frequency with a single strong flat-lying reflector at depth, suggesting about 10 m of glaciofluvial sand and silt overlying the fine grained till, with no massive ice present.

Figure 8 represents the 50 MHz surveys conducted along Line B, perpendicular to Line A and starting 50 m southeast of the borehole. The glaciofluvial overburden on top of the ridge is approximately 2 to 3 m thick and not as variable as it is to the northeast. The base of the ice body has been

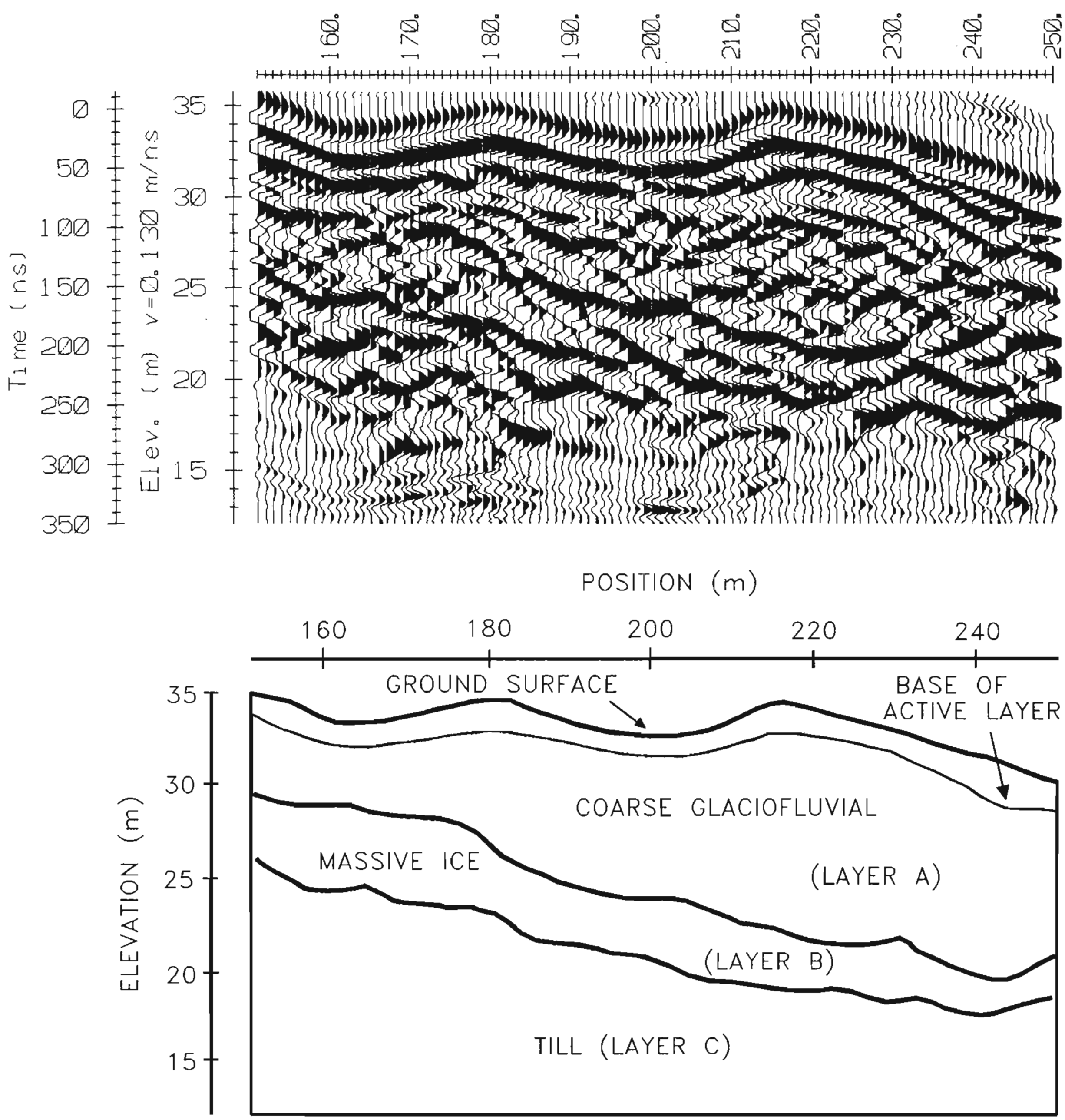

Figure 6. Line A - GPR survey (upper) with interpretation (lower) conducted in September 1992 northeast of the borehole showing the thinning of the ice body and the thickening of the overlying glaciofluvial sediments.

delineated as varying between 10 and 13 m below the ground surface. A sudden change in reflector pattern near +85 m, from the continuous wavy returns typical of ice, to the chaotic reflectors more typical of frozen sediment, suggests an abrupt vertical contact between the ice and glaciofluvial deposits. Reflections from a sharp contact are not evident at this location, perhaps due to the steepness of the interface. Reflector patterns between the +90 and +160 m positions suggest 9 to 12 m of glaciofluvial materials overlying till, similar to that seen in Figure 7.

A profile along Line C was run to study the ice and sediment morphology at the northwest edge of the ice body. Figure 9 shows the ice body overlain by 7 to 8 m of glaciofluvial sediment, pinching out near the 120 m position as the profile continues down slope and onto a lacustrine plain. Steeply dipping reflectors on the lower slope (125 to 145 m) suggest the possible presence of onlapping lacustrine sediments from a higher past lake level. Chaotic returns from within the lacustrine sediments and the penetration of the

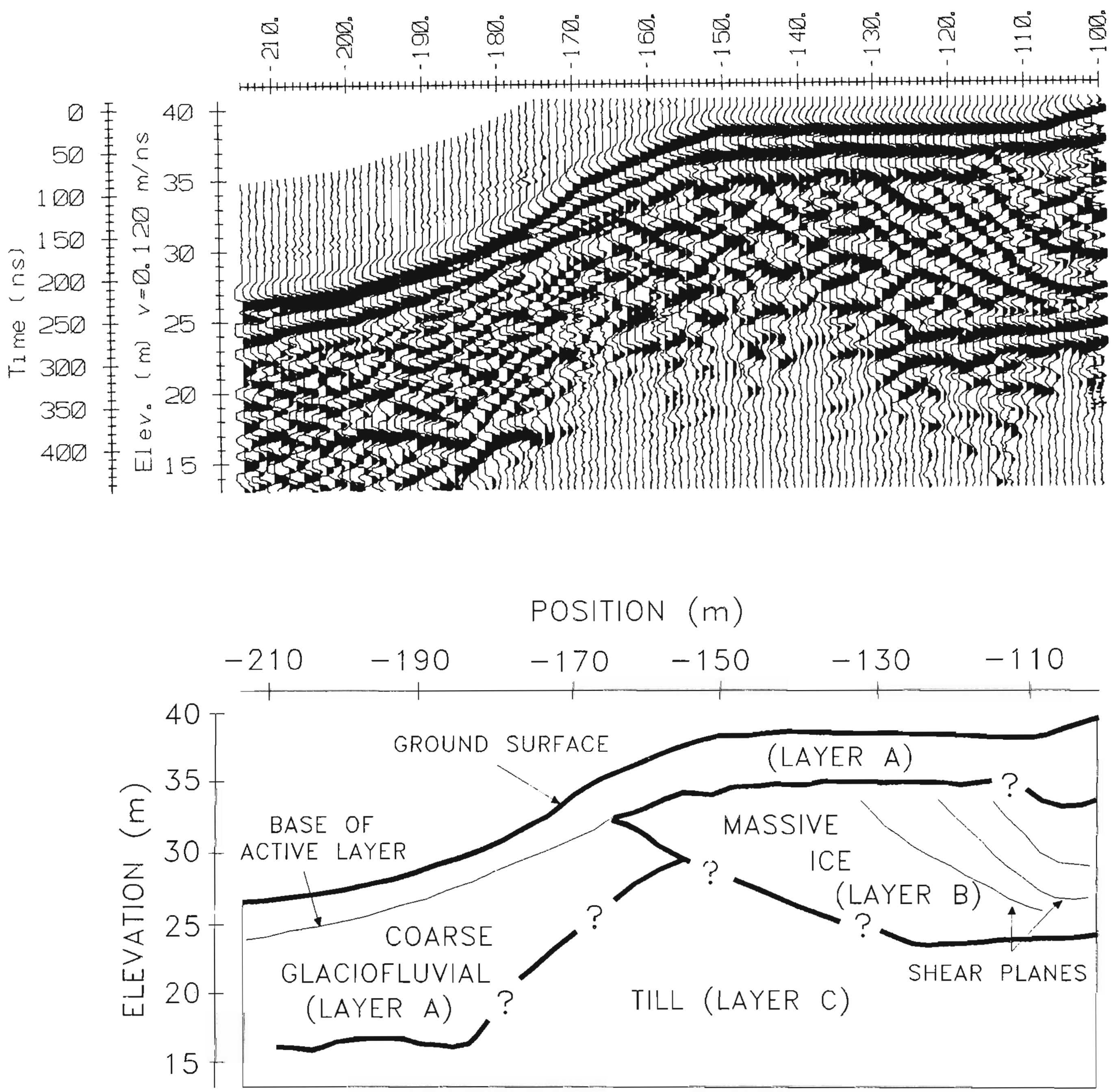

Figure 7. Line A - GPR survey (upper) and interpretation (lower) conducted in August 1990 showing the southwest edge of the massive ice body and the enclosing till and glaciofluvial sediments.

signal to depths of over 20 m suggests a high ice content. The presence of till beneath the lacustrine sediments is not apparent in this profile.

DISCUSSION

Considerable stratigraphic and structural information is evident from reflections within the ice body, from each sedimentary unit and from the delineated boundaries of the ice body.

Within the ice body, two distinct radar stratal patterns can be observed. There are continuous, wavy, subparallel reflections produced near the centre portion of the ice body, best shown in Figure 5 (-20 m to -70 m). This type of reflector appears to be the result of sediment banding or ice flow. Near the southwest edge (-110 m to -150 m in Fig. 7) the reflectors curve upward, become less continuous and unconformably contact the upper boundary, suggesting ice that may have undergone compressional flow (Drewry, 1986). The thinning of the ice body near its edges (Fig. 6 and 9) presents less detailed internal information, however the wavy and continuous nature of the reflectors seen in other profiles is still observed.

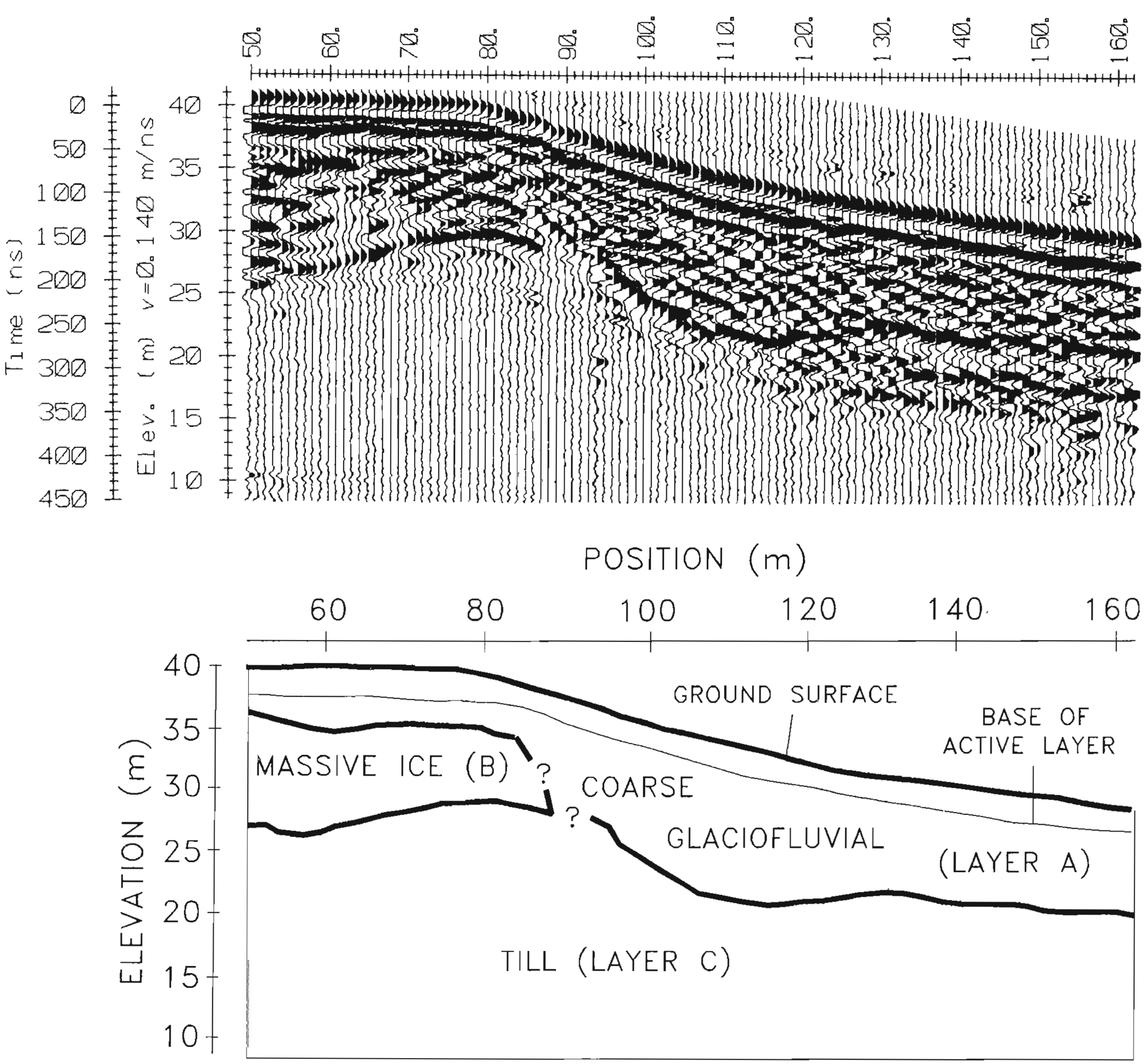

Figure 8. Line B - GPR survey (upper) with interpretation (lower) conducted in August 1990 showing the southeast edge of the massive ice body and the enclosing till and glaciofluvial sediments.

Reflections observed within the till below the ice are subcontinuous to chaotic and are quickly attenuated with depth. Similar radar stratal patterns are also observed in other till deposits (A.S. Judge, unpub. data, 1991). The ice-poor glaciofluvial sediments overlying the ice body produce continuous complex reflection patterns to chaotic reflections. These reflection patterns were also observed by Moorman et al. (1991) in modern fluvial deposits. The lacustrine sediments are characterized by deep signal penetration, high signal velocity, and a chaotic structure. This is similar to reflector patterns seen in ice-rich fine grained sediments by Judge et al. (1991).

Dallimore and Wolfe (1988) concluded that the ice bodies associated with glaciofluvial sediments near Yaya Lake were most likely of glacial origin. This interpretation was based upon the inferred glacial derivation of the sediment bands within the ice, the abrupt nature of the upper ice-sediment contact, and the variable ice fabric. These ice bodies also had lower electrical conductivity and lower cation concentrations than bodies of segregated ice in the region. Additionally, oxygen-18 isotope values from samples taken within a similar ice body suggest isotopic fractionation common to segregated ice did not occur.

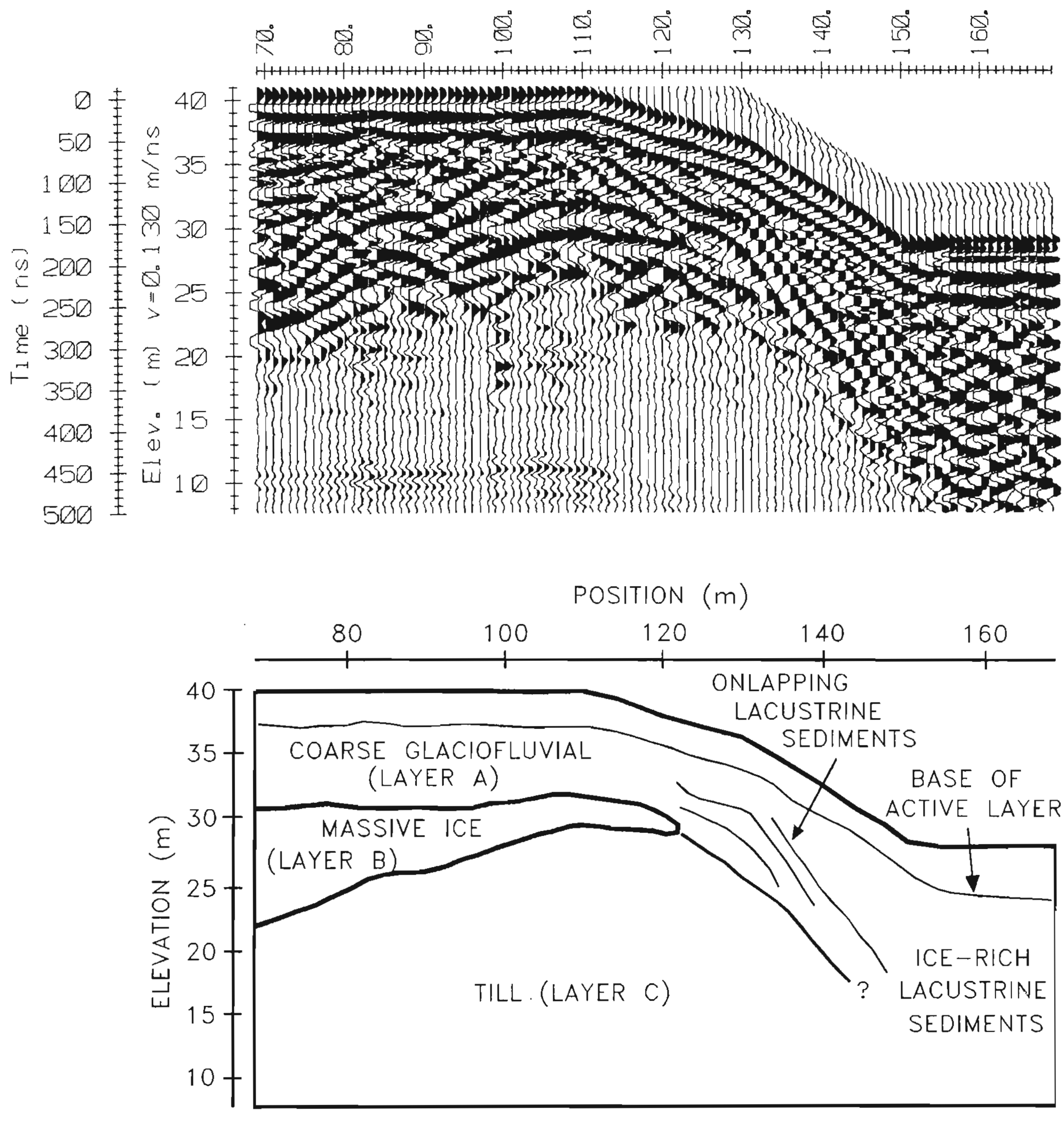

Figure 9. Line C - GPR survey (upper) with interpretation (lower) conducted in September 1992 showing the northwest edge of the massive ice body and the associated ice-rich lacustrine sediments.

Furthermore, if the ice body was of segregation origin the considerable variation in thickness of the sediment layer covering the ice would not be expected. The morphology of the ice body as mapped using ground penetrating radar is highly irregular while the ridge top is flat, probably indicating ice deposition was followed by burial by glaciofluvial sediments. The oblique contacts observed between the ice and reflectors in the overburden support this conclusion.

The patterns of glaciofluvial deposition visible in the radar profiles may suggest meltwater channels cut on the ice surface producing differential sediment deposition. This evidence and the steep edges of the ice body, especially to the southeast (Fig. 8, +80 to +90 m), suggest a morphology consistent with a large block of glacially derived ice as opposed to ice segregated in situ. However, the ice bands found in the overlying glaciofluvial sediment are likely of an epigenetic, segregated origin.

CONCLUSIONS

Ground penetrating radar proves to be an extremely useful tool for mapping buried ice bodies. The series of profiles presented in this report illustrate the striking change in the character of the ice body and the enclosing sediments from one end to the other.

Along the surveyed lines, the buried ice body was found to have approximate horizontal dimensions of 420 X 210 m and a maximum thickness of about 10 m. It is likely that more ice is present beyond the thermokarst pond at the end of line A, as the ridge continues for several hundred metres to the northeast.

The morphology of the ice body and the enclosing sediments revealed through radar profiling suggests the massive ground ice at Yaya Lake is of buried glacial origin. This conclusion is supported by the results of a previous study by Dallimore and Wolfe (1988).

ACKNOWLEDGMENTS

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Revised bedrock geology of Belot Ridge, District of Mackenzie, Northwest Territories

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Abstract: Ordovician and Silurian strata of the Franklin Mountain and Mount Kindle formations are exposed in two structural culminations along Belot Ridge in the Colville Hills, Northwest Territories. Geological Survey of Canada Map 12-1970 is inaccurate because it shows Devonian strata as the oldest exposed rocks in these localities. The culminations occur at deflections in the ridge. If there was a component of strike slip during Laramide tectonism, as suggested by MacLean and Cook (1992), the deflections would have constituted restraining bends and caused the observed structural culminations.

Résumé : Des strates ordoviciennes et siluriennes des formations de Franklin Mountain et de Mount Kindle affleurent dans deux culminations de la crête de Belot, située dans la région des collines Colville (T.N.-O.). La carte 12-1970 préparée par la Commission géologique du Canada est inexacte, parce qu'elle représente les strates dévoniennes comme les strates les plus anciennes affleurant dans ces localités. Les culminations apparaissent en des points de déflexion de la crête. S'il y avait eu une composante du décrochement au cours de la tectonique laramienne, comme l'ont suggéré MacLean et Cook (1992), les déflexions auraient provoqué des gauchissements, et généré les culminations observées.

INTRODUCTION

Belot Ridge was mapped during the 1968 field season (Cook and Aitken, 1971) as part of a large helicopter-supported surface mapping operation. Considering the huge area covered, and the reconnaissance nature of the project, in which most observations and formation identifications were made from the helicopter, some errors in identification were inevitable. Two such errors (see Fig. 1 for locations) were discovered along Belot Ridge during field work in 1992. This note is published as an erratum to Geological Survey of Canada Map 12-1970 (Aitken and Cook, 1970). Important stratigraphic data are taken from Pugh's (1985) description of the Belot Hills M-63 well which is closely adjacent to Belot Ridge in one of the remapped areas.

STRATIGRAPHY

Franklin Mountain Formation, cherty unit

The Lower Ordovician cherty unit (Norford and Macqueen, 1975, p. 12) is the oldest stratigraphic unit exposed in the remapped areas (Fig. 2, 3). The unit comprises fine and medium crystalline, grey and grey weathering dolomite. Its most distinctive characteristic is the presence of abundant silicification, particularly in the form of white silicified stromatolites. The cherty unit is about 380 m thick in the Belot Hills M-63 well (Pugh, 1985).

Mount Kindle Formation

The Mount Kindle Formation is Late Ordovician to Early Silurian (Norford and Macqueen, 1975, p. 17). It consists of generally thick bedded, medium to dark brownish grey, fine to medium crystalline dolomite. Its most distinguishing characteristic is the presence of halysitid, favositid, and horn corals, and orthoconic cephalopods. Regionally, the Mount Kindle is discontinuous due to erosion at the sub-Devonian unconformity. It appears to be absent in the area of Figure 3, is absent in the Mobil Belot Hills M-63 well on the east flank of Belot Ridge (Pugh, 1985), and is discontinuous at the surface in the area of Figure 2, where it is estimated to be less than 50 m thick.

Devonian, undivided

Along most of Belot ridge, the exposed rocks are Devonian, consisting of laminated, fine crystalline, brown dolomites (commonly brecciated), and thin bedded, grey weathering limestones. These strata were mapped by Aitken and Cook (1970) as the undivided Bear Rock Formation. Equivalent

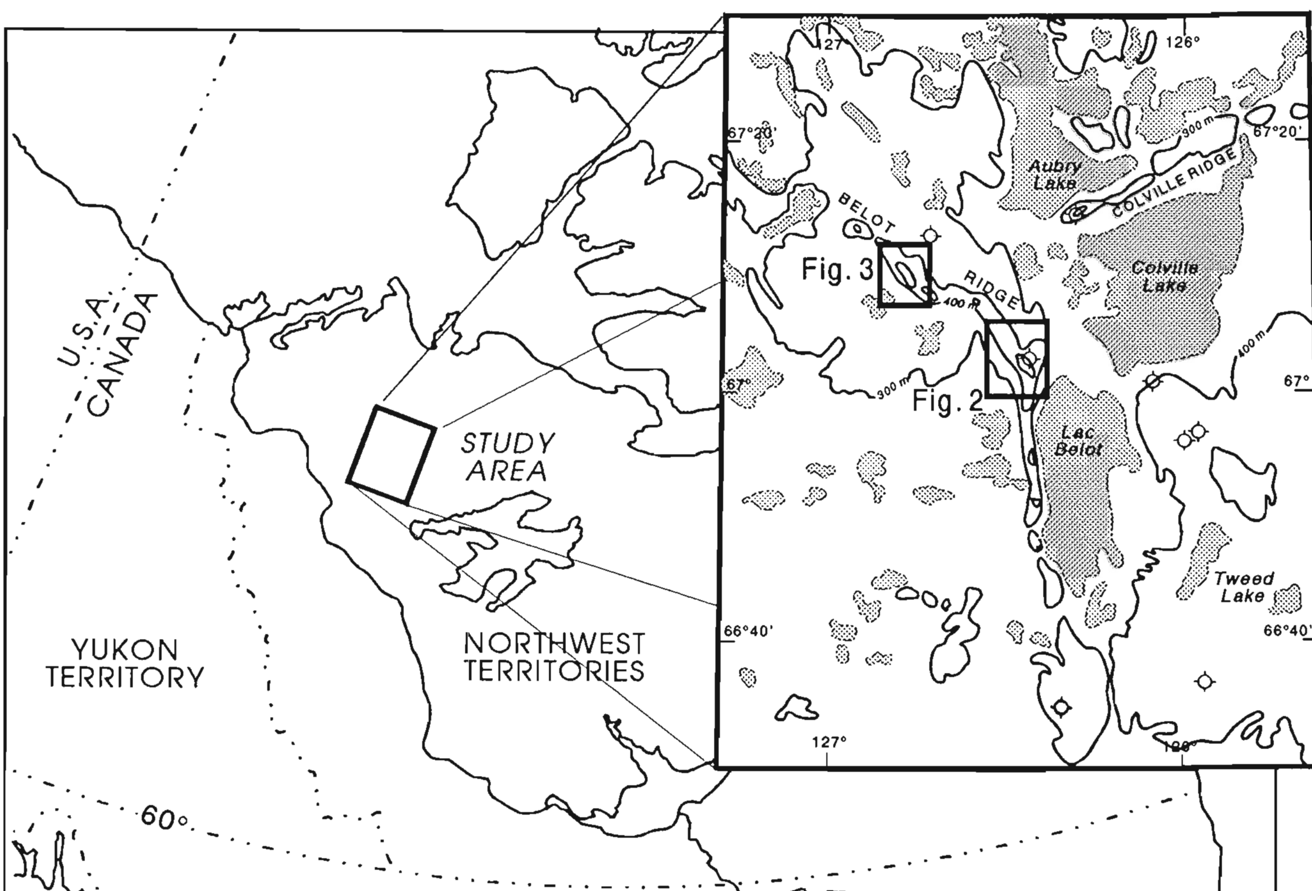

Figure 1. Location map showing Belot Ridge and boundaries of Figures 2 and 3.

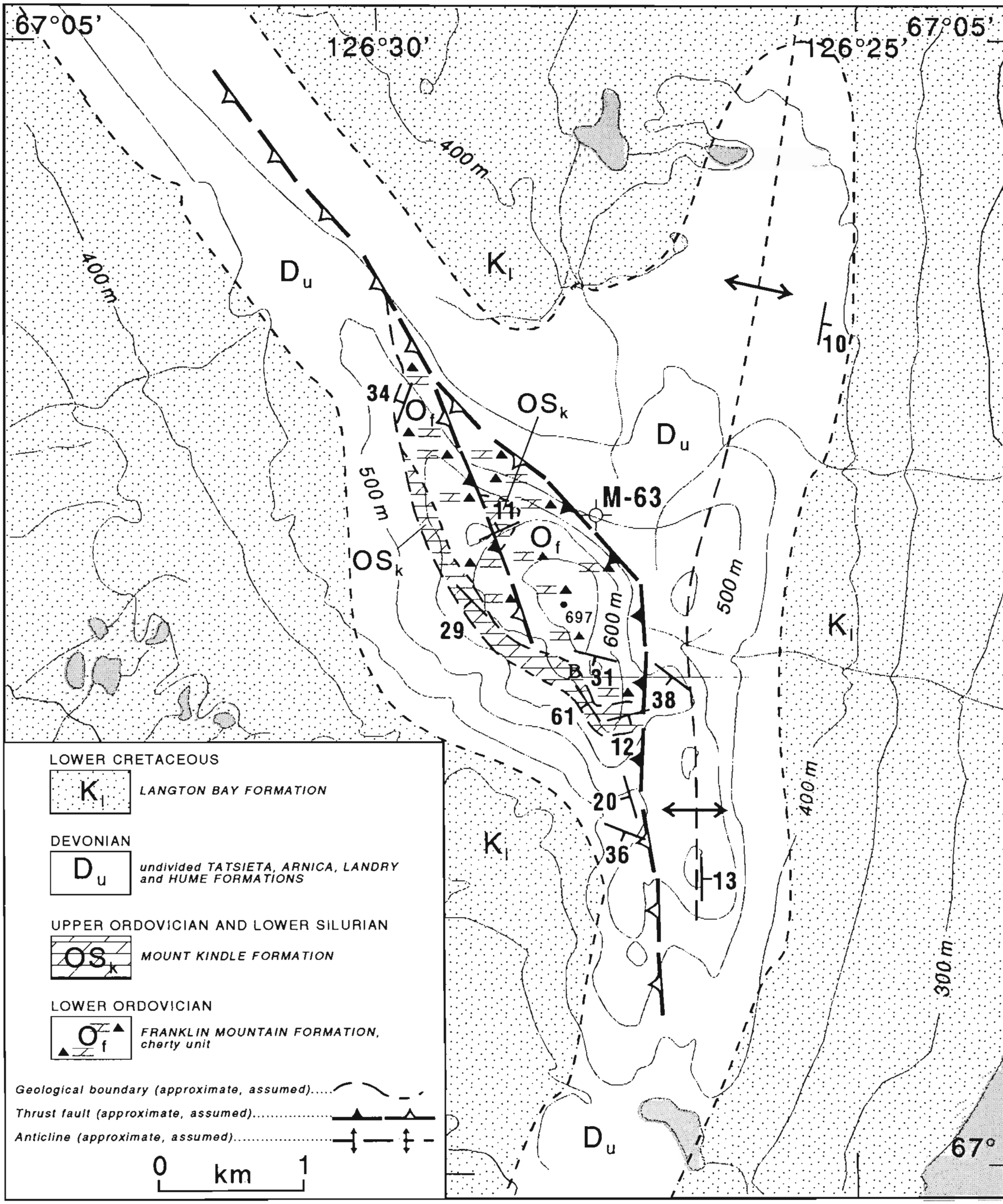

Figure 2. Structural culmination exposing the Franklin Mountain and Mount Kindle formations at the prominent deflection point in Belot Ridge. See Figure 1 for location.

strata in the subsurface have been assigned by Pugh (1985), in upward succession, to the Tatsieta Formation (limestone and shale), Arnica Formation (dolomite), and Landry Formation (limestone). Moreover, Pugh identified the fossiliferous Hume Formation overlying Landry limestones in the Belot Ridge M-63 well adjacent to Belot Ridge (Fig. 2). In the current work, poor exposure precluded reliable mapping of the Devonian carbonate units, and the entire sequence, Tatsieta to Hume inclusive, is mapped as undivided Devonian.

Langton Bay Formation

Basal Cretaceous sands in the Colville Hills region were assigned by Yorath (1981) to the Lower Cretaceous Langton Bay Formation. The sandstones are very poorly exposed but do outcrop in river and stream valleys and locally along the flanks of the uplifted Colville Hills ridges. Although no Cretaceous strata were observed in the areas studied, they are assumed to flank Belot Ridge (Fig. 2, 3) as suggested by Aitken and Cook (1970).

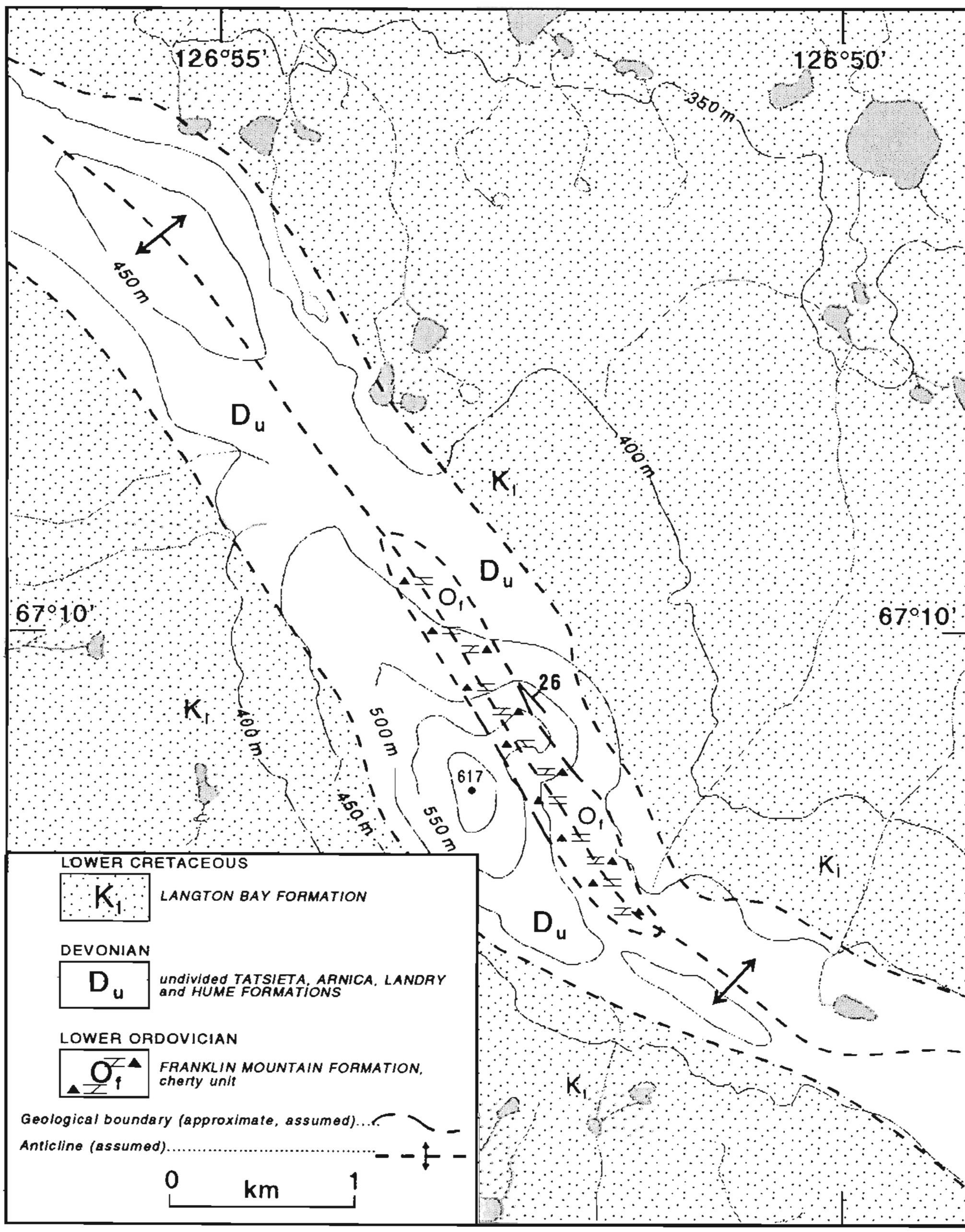

Figure 3. Structural culmination exposing the Franklin Mountain Formation in Belot Ridge. See Figure 1 for location.

STRUCTURE

Belot Ridge is a generally north trending anticlinal range along the west side of Belot Lake. Northwest of the lake it bifurcates into a short north trending ridge and a longer northwest trending one (Fig. 1, 2). A structural and topographic culmination occurs at the point of bifurcation (Fig. 2). An unnamed mountain at the culmination is the highest peak (697 m) along Belot Ridge. At its summit, grey felsenmeer rubble consists of fine and medium crystalline, grey and grey weathering dolomite, with abundant white silicified stromatolites, characteristic of the cherty unit of the Franklin Mountain Formation. The peak is at the core of a half-dome, which generally dips southwestward and plunges to the northwest and to the south. On the east side of the half-dome, a major thrust fault (Fig. 2), whose trace is covered in an intervening valley, is interpreted to place Franklin Mountain strata over west dipping Devonian strata found in the small anticlinal ridge to the east.

In traverses downplunge to the south, and downdip to the west, successively younger strata of the fossiliferous Ordovician-Silurian Mount Kindle and undivided Devonian formations are encountered. Downplunge to the northwest, a small outlier of the Mount Kindle necessitates the interpretation of a small thrust which merges northward with the main thrust. Farther downplunge Devonian strata occur in close proximity to the Franklin Mountain Formation cherty unit (dolomite with silicified stromatolites) and we interpret that, at that location, the Mount Kindle has been removed by erosion at the sub-Devonian unconformity. This is complimentary to Pugh's (1985) interpretation of the Mobil Belot Hills M-63 well, in the immediate footwall of the major thrust, where he reported Devonian strata directly overlying the Franklin Mountain Formation cherty unit.

Cherty unit rocks occur at another topographic culmination (617 m) farther northwest along Belot Ridge (Fig. 3). There, no bedrock is exposed along the main ridge, but a felsenmeer of thin bedded limestones attests to Devonian bedrock, and the morphology of the ridge implies that strata dip southwestward. Downslope to the northeast, a prominent bench has abundant felsenmeer blocks of dolomite with silicified stromatolites. These cherty unit strata were not seen in place, but are overlain by northeast dipping, thin bedded, brown, grey weathering limestones of the undivided Devonian. The limestone beds are in close proximity to cherty unit rubble, and we conclude that the Mount Kindle Formation has been removed by pre-Devonian erosion. The simplest structural interpretation is that the cherty unit rocks occur in the core of an anticline as shown in Figure 3.

TECTONIC IMPLICATIONS

Colville Hills structures are widely separated, and occur as linear, locally faulted, anticlinal ridges that display a variety of structural trends and inconsistent vergences. Their origin has been, and continues to be, equivocal. Cook and Aitken (1971) suggested that they were the consequence of tectonic shortening above a regional detachment level in salt of the Cambrian Saline River Formation. Davis and Willott (1978), using drilling, magnetic, and seismic data, disproved a regional detachment in the Saline River by establishing that Precambrian rocks were involved in these structures. They argued that the Colville Hills were probably a consequence of right-lateral strike-slip reactivation of steep Precambrian faults, a concept that was expanded by Cook (1983). Recently, MacLean and Cook (1992), showed a close spatial relationship between Colville Hills structures and underlying steeply dipping Precambrian faults, and suggested that the Laramide Colville Hills were due to right-lateral strike-slip reactivation of the Precambrian thrusts. Conversely, Clark and Cook (1991) interpreted a regional thin-skin detachment in the Proterozoic, and attributed the Colville Hills (Clark and Cook, 1992) to compressive reactivation of Proterozoic structures related to that detachment.

It is noteworthy that the northwest trending arm of Belot Ridge directly overlies a large Precambrian thrust (MacLean and Cook, 1992, Fig. 4, 7). Moreover, the large culmination shown in Figure 2 occurs very close to the southern termination of the Precambrian fault. Thus it is evident that the Laramide structure is related to the large earlier fault, although the specific structural linkages are not clear. The culmination does not appear to be due to direct compressive reactivation of the underlying Precambrian thrust because the Laramide fault verges to the east, whereas the Precambrian fault verges to the west. On the other hand, the culminations outlined here, because they occur at deflections in the range, are compatible with the strike-slip models of Davis and Willott (1978) and MacLean and Cook (1992). If there was a component of strike slip in the compressive displacement along the Belot Ridge structure, culminations would be expected at structural deflections which would have constituted restraining bends. In particular, the larger culmination (Fig. 1, 2) occurs at a pronounced northwest deflection point in the north trending Belot Ridge, and could be due to a component of right-lateral strike-slip acting along the long north trending segment of the structure.

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Geological Survey of Canada Project 890038

Characteristics of an ice-scoured river bank near Keele River confluence, Mackenzie Valley, Northwest Territories

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Brooks, G.R., 1993: Characteristics of an ice-scoured river bank near Keele River confluence, Mackenzie Valley, Northwest Territories; in Current Research, Part B; Geological Survey of Canada, Paper 93-1B, p. 39-43.

Abstract: The shearing of river ice during break-up along the eastern bank of Mackenzie River near Keele River confluence produced a scoured zone 4-13 m above the river surface. This ice-scoured zone was sloped 40-50° with a concave-upwards profile. Its surface was a boulder pavement consisting of a massive matrix-supported diamicton with clasts generally flush to the surface. The ice-scoured zone was etched with lineations running parallel to the slope of the river surface. The development and preservation of this ice-scoured zone primarily relates to the cohesiveness of the silt/clay-rich sediment forming the lower part of the river bank. Ice-scouring may represent an efficient mechanism in both the erosion of the lower river bank and in the removal of debris generated by slope failures along the bank face.

Résumé : Le cisaillement de la glace de rivière, qui a lieu au cours de la débâcle le long des rives du fleuve Mackenzie près de son confluent avec la rivière Keele, dans les Territoires du Nord-Ouest, a produit une zone d'affouillement entre 4 et 13 m au-dessus de la surface du cours d'eau. Cette zone affouillée par les glaces avait une pente de 40 à 50° et un profil concave vers le haut. Sa surface était un dallage de pierres composé d'un diamicton massif à fabrique à éléments supportés par la matrice, contenant des clastes situés directement en surface. La zone affouillée par les glaces a été marquée de linéations parallèles à la pente du cours d'eau. Le développement et la conservation de cette zone affouillée par les glaces sont principalement liés à la cohésion du sédiment riche en silt et argile qui constitue la partie inférieure de la rive du cours d'eau. L'affouillement par les glaces représente peut-être un mécanisme efficace à la fois du point de vue de l'érosion de la partie inférieure des rives du cours d'eau et du retrait des débris générés par l'effondrement des pentes bordant la surface de la rive.

INTRODUCTION

The annual break-up of river ice is an important geomorphic process along Mackenzie River, Northwest Territories. Break-up generally begins in the shallower reaches with ice being transported downstream, often becoming lodged behind intact ice to form an ice jam (Kamphuis and Moir, 1983). Large jams can obstruct the river causing localized flooding upstream (MacKay and MacKay, 1973a, b; MacKay et al., 1974) which may extend into the lower reaches of tributaries (Egginton and Day, 1977; Egginton 1980). As described by MacKay and MacKay (1977), ice movement and jamming creates 'ice-push' features which dominate the morphology of the banks along Mackenzie River. A distinct trimline situated above the level of high flood commonly marks the upper limit of ice-push. Along bouldery banks, ice movement will form 'boulder pavements', 'loose boulder pavements', and clusters of loose boulders; all subject to varying degrees of annual modification. The bull-dozing of the heads of islands by ice will form 'ice-push island buttresses'. The shearing actions along converging, bouldery river banks can produce a succession of 'boulder ridges' which resemble short, stubby bars that project downstream into the channel.

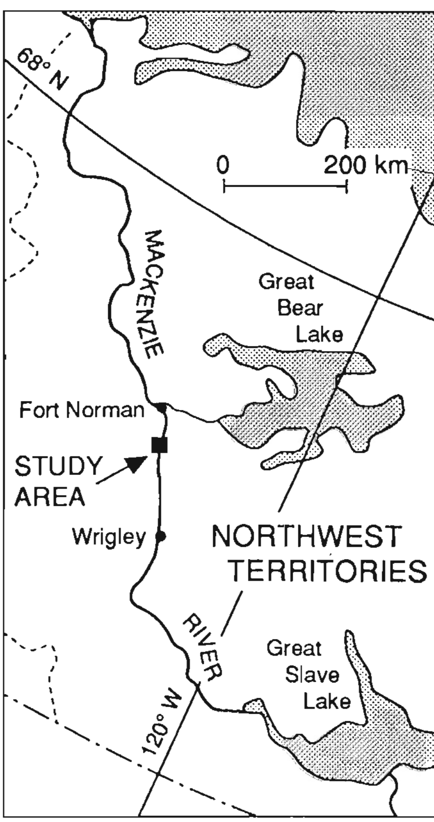

Figure 1. Location of the study area in the Northwest Territories.

The report describes a striking example of an ice-scoured river bank, a less common feature along Mackenzie River that is associated with ice break-up. This type of bank morphology clearly illustrates the ability of river ice to directly erode bank materials and produce a distinctive morphology.

STUDY AREA

The ice-scoured cutbank described here occurs along the eastern side of Mackenzie River opposite Keele River confluence between Wrigley and Fort Norman, Northwest Territories (Fig. 1, 2) and falls well within the zone of discontinuous permafrost. Specifically, the eastern bank

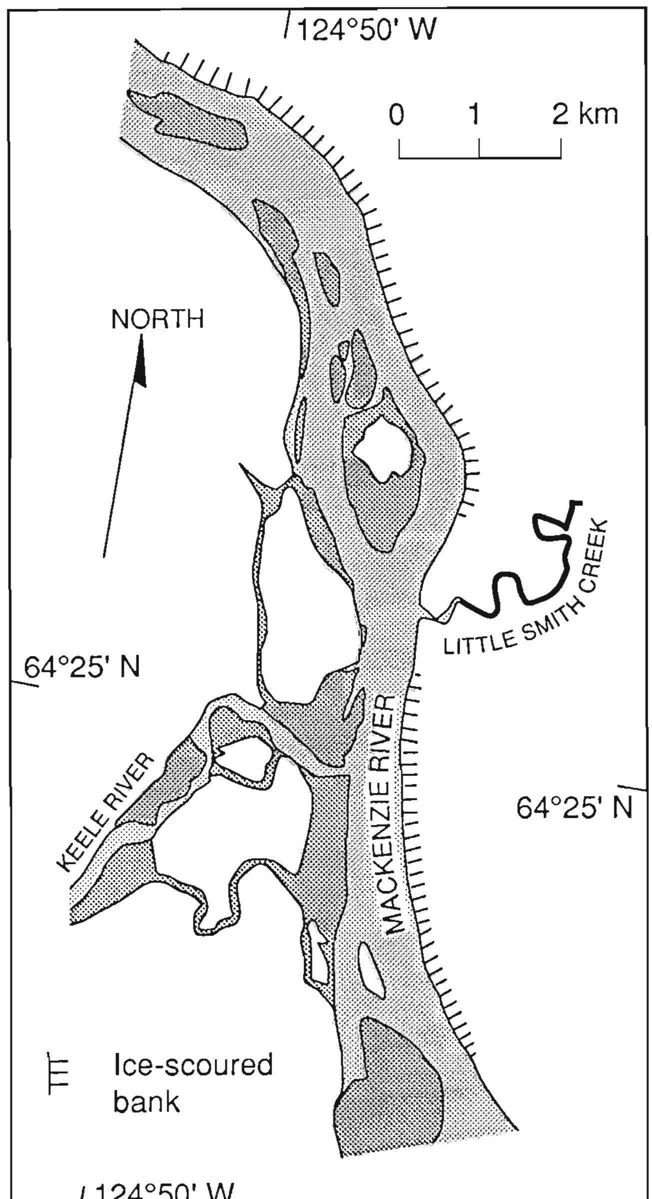

Figure 2. The eastern bank along Mackenzie River exhibiting the ice-scoured morphology in the area of Keele River confluence.

exhibiting this morphology is located between river distances 732-736 km and 738-746 km (see Canadian Hydrographic Service, 1991), being subdivided by the confluence of Little Smith Creek (Fig. 2).

The river banks in the study area generally are 50-60 m high with the lower 13 m exhibiting an ice-scoured morphology (Fig. 3). Above this height, the bank is subject to varying types of slope instability. Code (1973) mapped this reach as unstable being subject to large-scale retrogressive failures (translational slides, slumps and flows) and to a lesser extent, shallow earthflows, detachment slides, solifluction, gully erosion and slope wash. At the time of observation (August 7, 1992), minor slope failures that occurred since the spring break-up have destroyed or buried some of the ice-scoured morphology (Fig. 3). Because of this, the morphology was somewhat discontinuous along parts of the river bank, depending upon the intensity of the failures.

DESCRIPTIONS AND INTERPRETATIONS OF THE BANK MORPHOLOGY

The zone of bank interpreted as being ice-scoured was situated 4-13 m above the river surface. (All height references are with respect to the river surface on August 7, 1992.) Below, the lower zone of the bank (0-4 m) had an average slope of 22° and a surface morphology consisting of loose boulder pavement grading upwards into a boulder pavement (Fig. 3). Discontinuous lines of loose boulders and pebbles ran parallel to the river surface across this zone and represent poorly-defined 'beaches' formed at moderate discharges by wave action. The loose boulder pavement and boulder pavement morphologies with poorly defined 'beaches' superimposed, are common on similarly-angled banks elsewhere along Mackenzie River.

Along the zone 4-13 m above the river (Fig. 3), the surface slope of the bank averaged about 42° from 4-11 m then steepened to about 52° over the upper 2 m, producing a concave-upwards profile (Fig. 4). Along the 42° slope, the generally smooth surface of the face consisted of a massive matrix-supported diamicton. The matrix of the diamicton was composed of a compact sandy silt/clay. Clasts within the diamicton typically ranged in diameter from about 20 cm to very coarse sand and generally were flush with the bank face although some larger clasts projected several centimetres outwards (Fig. 4). Clast concentration of the smooth surface was greater than in situ diamicton exposed at minor slope failure scars occurring laterally along the bank. The surface of many clasts were etched with striations that commonly ran parallel to the slope of the river surface. Over the entire zone, shallow grooves also parallel to the slope of the river surface, formed lineations that were carved several centimetres into the diamicton, some running tens of metres along the bank (Fig. 3, 4). Less frequent lineations running across the bank were furrows up to 10 cm deep and several metres in length, and 'tails' consisting of a slightly raised surface tens of metres long (Fig. 3) that occurred to the lee of larger clasts projecting several centimetres from the bank surface. Occasionally, short gouges cut obliquely across the other lineations. The more steeply sloped part of the zone (11-13 m) generally was similar to the 4-11 m portion, but on close inspection clast concentration was lower, clasts generally projected from the bank surface, and the lineations were deeper and better defined.

The lineations and the smooth surface of the bank must have formed by scouring as rigid debris was transported by the river when the 4-13 m high zone was at least partially submerged. The obvious explanation for this scouring is ice being sheared along the bank during break-up. The height of the scoured zone above the river surface is consistent with ice-push features common to other reaches of Mackenzie

Figure 3. Typical well-developed ice-scoured zone along the lower ~13 m of the eastern river bank; note the lineations running across the lower bank and the tails to the lee of boulders on the bank surface; the direction of ice shearing was from right to left. Above ~13 m, the bank is actively failing with some fresh failures extending into or across the ice-scoured zone. (GSC 1992-245B)

Figure 4. Close-up of the ice-scoured zone showing the concave-upwards profile of the bank surface which is etched with shallow lineations. Note that clasts within the zone are flush with the bank surface. (GSC 1992-245A)

River. The only other reasonable hypothesis is that the scouring may have been caused by floating logs during a high discharge, but fresh logs and log jams are not prolific along the local river banks.

The ice debris hypothesis for the scoured zone produces the following explanations for the various morphometric characteristics of the bank. The lineations formed by the grooves and furrows represent the shearing of individual ice fragments along the river bank. The general shallowness of these lineations and the smooth surface of the ice-scoured zone reflects the intense scraping by river ice. Ice thrusting generated by stresses within the ice flow as it sheared along the bank probably formed the deeper furrows and oblique gouges. The general movement of scoured debris across the bank by the ice would have re-deposited some of the clasts by pressing them into the diamicton forming a boulder pavement thereby increasing the clast concentration. The formation of the boulder pavement and the presence of deeper furrows and gouges suggests that the surface of the bank was unfrozen when it was sheared by the ice. Scouring of the bank gradually would have caused some of the larger clasts to project from the surface as the surrounding, more easily eroded material was removed. These projecting clasts appear to represent local points of resistance to scouring as indicated by the raised surface of the 'tails' that extend to their lee. The grinding of detritus being carried across the bank face by the ice created the striations upon the clasts. River discharge or ice-push levels may only have been sufficient to affect the upper 11-13 m portion of the scoured zone for a comparatively short duration; the face is etched with lineations, but a pavement is not developed. The deeper lineations here also may reflect a less intense scouring.

Above 13 m, the remaining area of the bank was above the direct effects of fluvial and river ice processes. It experiences slope instability typical of many banks along Mackenzie River; fresh detachment slide scars average ~45° and extend up to 18-33 m (Fig. 3). Above these scars, the entire slope to 50-60 m in height has experienced creep as indicated by the bent and leaning trees that are scattered across its face. Gullying into the face occurs where small ephemeral streams and ground water seeps flow downward to the river.

DISCUSSION

Ice-scour mechanism

The presence of an ice-scoured morphology in the study area is related to several variables. In order to form, ice must be sheared across the river bank. At the time of observation, the position of the main flow in the upper part of the reach was aligned slightly towards the eastern bank because of the position of bars and islands within the channel just upstream. While in the middle and lower part of the reach, the main flow was along the outside of a river bend. During break-up when discharge is high, the main flow presumably is similarly located. This current would direct ice towards and along the eastern bank promoting shearing.

Undoubtedly, aspects of the break-up itself are important. The ice-scoured morphology could be affected by, for example, the thickness of the ice flow, the speed of the ice flow, the occurrence of an ice jam, and the influence of an ice damming causing a backwater effect. Unfortunately, details of the 1992 break-up are not known because the study reach is not located near a permanent settlement nor is it monitored regularly. For stations upstream and downstream of the study reach, break-up at Fort Simpson began on May 4 with the ice running by May 8 (J. Wright, pers. comm., September 8, 1992), whereas at Norman Wells it started on May 23 with the ice running on May 24 (J. Symes, pers. comm., September 15, 1992); break-up at the study site would have happened between these dates. While it is not known if the break-up involved an ice jam situated at or immediately downstream of the study reach, the island and bars of this reach are favourable for ice jam formation (M. Hansen, pers. comm., September 8, 1992).

Most importantly, the ice-scoured morphology relates to the strength of the diamicton forming the lower part of the river bank since the shearing appears to have occurred to an unfrozen bank surface. The sandy silt-clay matrix of the diamicton forms a cohesive bank. This material is able to be deformed and scoured by the shearing ice, producing the various types of lineations where a less- or non-cohesive material might be subject to disintegration (e.g., sand). The diamicton also is capable of supporting ~40-50° slopes both under submerged (where the ice-scouring occurs) and subaerial conditions thereby preserving the ice-scoured morphology. It is not known if the variation in development (moderate to well) of the ice-scoured morphology along the river bank reflects compositional changes to the diamicton. Ice-scouring of the river bank undoubtedly is not unique to the study area, but weaker bank materials may not allow the formation and/or preservation of the ice-scoured morphology.

Influence on slope development

The removal of failed materials from the base of the slope is an important control on the stability of a river bank (Thorne, 1982). The accumulation of failed materials tends to protect the intact bank from basal retreat thereby promoting the stability of the slope above against gravitational failure. With the exception of several bank failures having widths of tens or hundreds of metres, all of the observed bank failures crossing the ice-scoured zone were very fresh having been formed since the 1992 spring break-up (see Fig. 3). Ice-scour, thus, appears to be an efficient mechanism in the removal of failed material generated prior to the river ice break-up of 1992. The ice-scoured morphology certainly indicates that erosion of the lower bank is occurring and, therefore, probably destabilizing the lower slope of the bank immediately above the ice-scoured zone. If the removal of these failed materials and erosion into the river bank occur on a annual basis this suggests that the slope instability of the upper bank is partially controlled by the ice-scour processes.

Future observations of the river bank near Keele River confluence would reveal the longer-term behaviour of the ice-scoured zone. It is not known if the ice-scoured morphology is a transitory feature, the product of 'unique' circumstances during the 1992 break-up. Since instability to the slopes of the river bank gradually is destroying or burying the ice-scoured zone, it must be renewed annually in order to exist from year to year. Persistence of the ice-scoured zone would indicate that bank shearing during break-up is an important process in the study area. If this were the case, more detailed monitoring would indicate the rate of scouring into the bank. Annual rates of erosion may reveal relationships with the thickness of the river ice and characteristics of the break-up. An interesting question concerns whether patterns of instability to the upper river bank can be directly associated to the scouring of the lower slopes.

CONCLUSIONS

1. The shearing of a river bank by ice produced a zone of boulder pavement exhibiting a concave-upwards profile that is etched with lineations parallel to the river surface.
2. While ice-scouring of river banks is common along Mackenzie River, the development and preservation of the distinctive ice-scoured morphology reported here likely relates to the strength of the materials in the lower area of the river bank.
3. Direct bank erosion by ice-scouring and removal of debris originating from slope failures may have an important role in the instability of the river bank above the scoured zone.

ACKNOWLEDGMENTS

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Geological Survey of Canada Project 920046

Study of the Triangle Zone and Foothills structures in the Jumpingpound Creek area of Alberta

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Olchowy, B.M. and Spratt, D.A., 1993: Study of the Triangle Zone and Foothills structures in the Jumpingpound Creek area of Alberta; in Current Research, Part B; Geological Survey of Canada, Paper 93-1B, p. 45-50.

Abstract: Foothills structures and stratigraphy in southwestern Alberta between the Bow and Elbow rivers were mapped at a scale of 1:20 000 during the summer and fall of 1992. The surface expression of the Triangle Zone structure at the leading edge of the Foothills is an anticline with east-dipping panels of Upper Cretaceous and Paleocene strata on its east flank, and east-verging folds and thrust sheets developed in Cretaceous strata on its west flank. The dominant structural style observed in outcrop is folding; thrust faults are interpreted based on mapped repetitions of strata and indicators of the stratigraphic top. A rotated to slightly overturned, west-verging thrust sheet is exposed in the core of the Triangle Zone. Shortening at the Mesozoic level is approximately 50 per cent across the Triangle Zone. The Cardium Formation, telescoped by thrust faulting, is progressively coarser grained toward the west.

Résumé : Les structures et la stratigraphie des Foothills dans le sud-ouest de l'Alberta, entre les rivières Bow et Elbow, ont été cartographiées à l'échelle de 1/20 000 au cours de l'été et l'automne 1992. Au centre de la zone Triangle à la limite frontale des Foothills, un anticlinal est observable à la surface. A l'est, les couches à pendage vers l'est appartiennent à la stratigraphie Crétacé supérieur et Paléocene. Dans la stratigraphie Crétacé à l'ouest de la zone Triangle, le régard des plis et des chevauchements est vers l'est. La plupart des structures qui affleurent sont les plis; les chevauchements ont été cartographiées s'ils répètent la stratigraphie et la polarité des couches n'est pas changée. Les couches verticaux sont récoupées par une faille chevauchante, au régard vers l'ouest, qui été renversé par la plissement au centre de la zone Triangle. Le raccourcissement est à peu près 50 % au niveau Mésozoïque. Le Cardium, raccourci à cause du chevauchement, devient à plus gros grain vers l'ouest.

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INTRODUCTION

The Triangle Zone comprises a structural complex, found at the eastern edge of the deformed belt of the southern Canadian Rocky Mountains, where foreland dipping strata are juxtaposed with hinterland dipping strata. The term Triangle Zone was introduced by Gordy et al. (1977) to describe the foreland edge of the orogenic wedge. The ideal Triangle Zone comprises a lower detachment surface that separates deformed foreland-verging allochthonous strata in its hanging wall from autochthonous strata in its footwall, and an upper detachment surface that separates foreland-verging allochthonous strata in its footwall from hinterland-verging autochthonous strata in its hanging wall. The upper and lower detachments define the top and base of an intercutaneous wedge. Simultaneous motion along these two surfaces allows the wedge to advance toward the foreland as it is emplaced between layers of undeformed strata in the Plains (MacKay, 1991).

Two simple two-dimensional models have been proposed by Charlesworth and Gagnon (1985) and by Jones (1982) to explain the geometry and formation of the Triangle Zone. Both models provide a good starting point for analysis of the development of the Triangle Zone, but the geometry of this structural complex does not conform to either of the models. Neither model addresses the problem of changes in geometry and deformation mechanics along strike.

The study area is located approximately 30 km west of Calgary, Alberta. It extends from the Bow River in the north to the Elbow River 25 km farther south (Fig. 1). Jumpingpound Creek cuts through the centre of the area; the geology along the creek is shown in Figure 1. Geological mapping was conducted during the summer and fall of 1992. Observations were plotted on air photographs (1:20 000 and 1:24 000 scale) and all data were compiled on current 1:20 000 scale Alberta digital topographic base maps.

The Jumpingpound Creek area was chosen for this study because it lies between two areas with significantly different structural geometries. South of the study area, at Turner Valley, the upper and lower detachments merge with a taper angle of approximately 30°; both detachments have ramp-flat geometries (MacKay, 1991). North of the study area, at Wildcat Hills (Spratt and Lawton, 1990) and Grease Creek (Soule and Spratt, 1992), seismic data indicate that the upper and lower detachments do not merge within the Triangle Zone; they continue beneath the Plains as flats separated by a 300 to 500 m wide deformation zone. Study of the Jumpingpound Creek area should provide critical information on how the upper detachment develops and transfers displacement along strike. Abundant well data and several seismic sections will be used in conjunction with detailed surface mapping to delineate changes in structural style with depth and along strike.

STRATIGRAPHY

Rocks exposed at the surface in the study area belong to the Upper Cretaceous Alberta Group (Blackstone, Cardium and Wapiabi formations), Brazeau Group, and the base of the Tertiary Paskapoo Formation.

The Blackstone Formation consists of dark grey to black, highly fractured and bioturbated shale that is thin bedded and contains abundant sideritic concretions. The formation is 265 m thick in the study area. The shale contains considerable amounts of fibrous calcite from *Inoceramus* shell fragments.

The Cardium Formation comprises dark grey shale with orange weathering, sideritic concretions and several sandstone and conglomerate interbeds. The Cardium sandstone at the base of the formation is a 2 m thick, highly sideritized, bioturbated, thin to medium bedded sandstone. In the western part of the Jumpingpound Creek area, there are two 3 to 4 m thick conglomerate interbeds that consist almost exclusively of chert pebbles with rust coloured, silty to sandy matrix. The pebbles are subrounded to rounded and range in colour from black to light grey, some with a yellowish tinge. In contrast, the Cardium Formation in the northeast part of the study area consists mostly of dark grey to black, thin bedded shale, interbedded with rippled siltstone, sideritic concretions and concretionary layers. The Cardium Formation is repeated in ten thrust sheets in the Jumpingpound Creek area; it is 60 m thick on average, and is progressively coarser grained toward the west.

Wapiabi Formation overlies Cardium Formation. It contains grey to dark grey, platy shale interbedded with siltstone that is usually rippled or internally very finely laminated. A few discontinuous, yellow weathering sandstone layers or lenses are observed near the top of Wapiabi Formation. In undeformed sections along Jumpingpound Creek, Wapiabi Formation is 445 m thick.

The Brazeau Group comprises medium to coarse grained sandstone and shale, containing abundant plant remains. The sandstone generally forms resistant ridges and contains channels and large scale trough crossbedding. The shale is usually olive-green or dark grey while the sandstone is buff coloured. Brazeau Group is approximately 700 m thick (Jerzykiewicz and Sweet, 1986). Following Ollerenshaw (1976), the Brazeau Group is subdivided into lower and upper parts wherever a conglomerate unit exists between them. In the study area, the conglomerate was encountered only in the westernmost part of the Elbow River, therefore the Brazeau Group was not subdivided everywhere.

The basal part of Paskapoo Formation consists of sandstones, siltstones and mudstones that are similar in appearance to those of Brazeau Group, but Paskapoo sandstones tend to be thicker, coarser grained, and weather a lighter colour. Previous studies in the Foothills have used the presence of fossilized leaves to recognize Tertiary rocks (Jerzykiewicz, pers. comm., 1991). In the study area, the base of the Paskapoo Formation was mapped at the base of a thick (510 m), coarse grained sandstone above which fossilized leaves have been found.

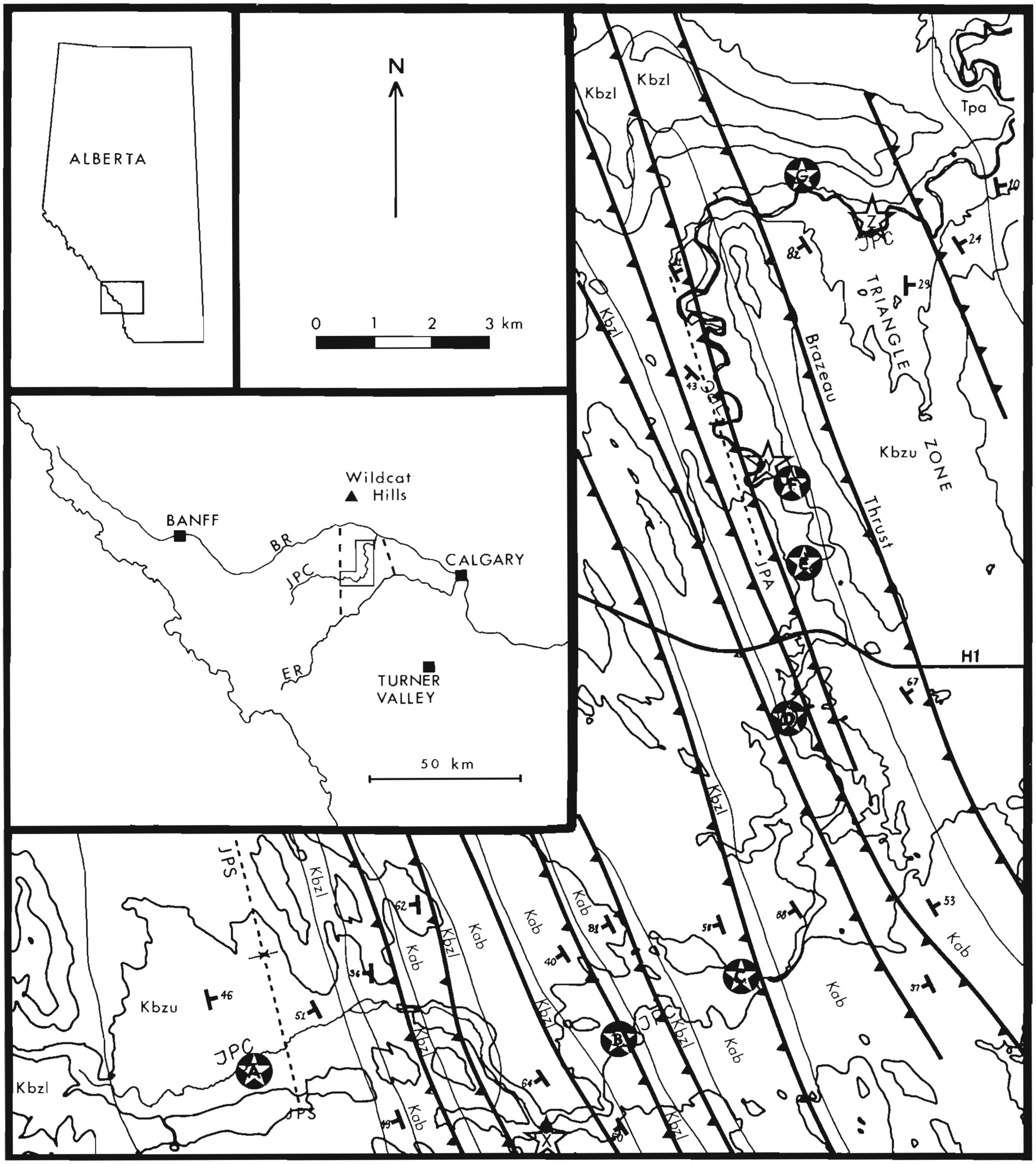

Figure 1. Geological map of a portion of the study area along Jumpingpound Creek (JPC). Inset map shows location of Figure 1 and the rest of the study area (dashed outline). Points A through G are locations of structures depicted in Figure 2. Points X, Y and Z are discussed in the text. BR = Bow River; ER = Elbow River; H1 = TransCanada Highway 1; JPA = Jumpingpound Anticline; JPS = Jumpingpound Syncline; Kab = Alberta Group; Kbzl = Lower Brazeau Group; Kbzu = Upper Brazeau Group; Tpa = Paskapoo Formation.

STRUCTURE

In the study area, the surface expression of the Triangle Zone area is an anticline with east-dipping panels of Upper Cretaceous and Paleocene strata on its east flank, and east-verging folds and thrust sheets, developed in Cretaceous strata, on its west flank. The general geometry of the Triangle Zone at Jumpingpound Creek is similar to that at Wildcat Hills; however, a deeper structural level (Cardium and Blackstone formations) is exposed in Jumpingpound Creek. Tight folds in these Alberta Group strata are similar to those observed in the core of the Triangle Zone at Turner Valley, but the higher level structures in Brazeau Group strata are very different. At Jumpingpound Creek, east-verging thrust sheets of Brazeau Group strata are folded over antiformal stacks of east-verging Alberta Group horses. At Turner Valley, west-verging thrust sheets of Brazeau Group strata are folded over antiformal stacks of east-verging Alberta Group horses (MacKay, 1991).

The earliest maps and interpretations of the Triangle Zone near Jumpingpound Creek do not disclose its true complexity (Beach, 1942; Hume, 1942). Access to seismic and well data along with knowledge of the surface stratigraphy has allowed a better understanding of the structures found on the eastern edge of the deformed belt. Fox (1959), Bally et al. (1966), Gordy et al. (1975), and Ollerenshaw (1976) developed more realistic cross-sections through the triangle zone; however,

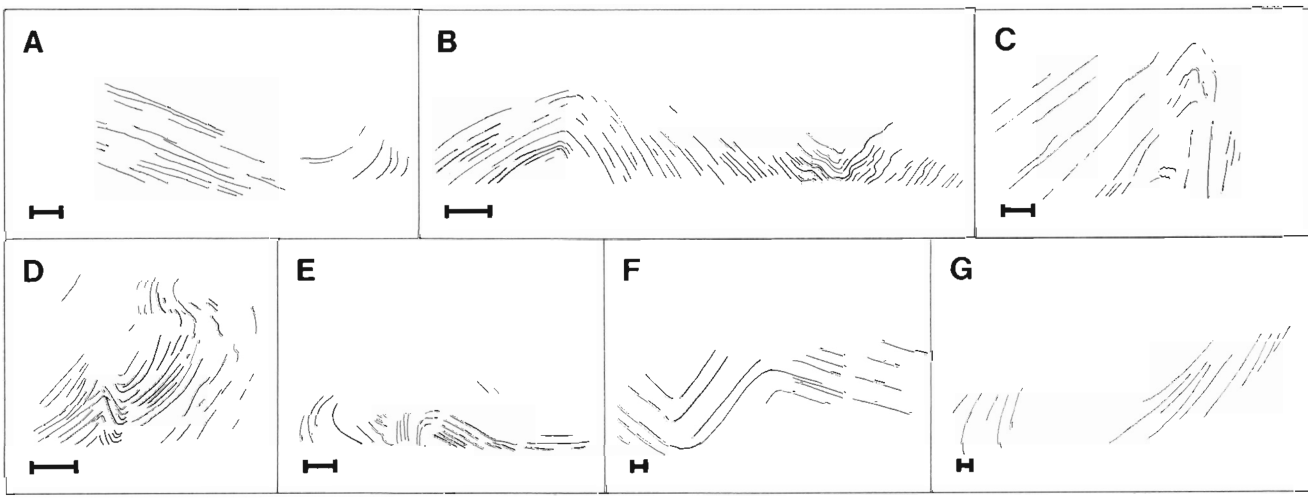

Figure 2. Line drawings (from photographs) of fold styles exposed along Jumpingpound Creek. Labels A-G correspond to Points A-G on Figure 1. Figures 2A and 2G show parallel folds in Brazeau Group sandstones with shale interbeds. Figures 2B,C,D,F depict fold styles in Wapiabi Formation shales. Figure 2E depicts hinge thickening in silty shales of the Cardium Formation. All views are looking northwest. Scale bars are 2 m long.

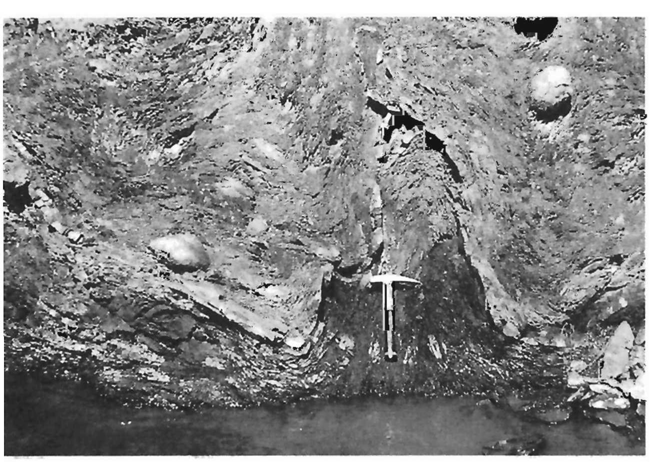

Figure 3. Tight fold in the Wapiabi Formation within the Brazeau Thrust sheet at point Y in Figure 1. The 2 cm thick siltstone layer exhibits 49 per cent shortening due to folding and contraction faulting; the shale is crenulated. View looking northwest from point Y in Figure 1. Hammer is 30 cm long.

Figure 4. Steeply dipping contraction fault (1 m displacement) and horse in vertical to slightly overturned Brazeau Group strata in the core of the Triangle Zone at point Z in Figure 1. View looking northwest from point Z in Figure 1. Stratigraphic top is up to the southwest (left). Person is 1.88 m tall.

the seismic data available to them targeted Paleozoic rocks, so structures they depicted in Mesozoic rocks were oversimplified and their cross-sections were not balanced.

Work by MacKay (1991) in the Turner Valley area has demonstrated that thrust faults at the surface and in the subsurface can only be linked if the structures in exposed Mesozoic rocks are mapped in great detail, with particular attention given to the geometry and vergence directions of folds. Published maps of the Jumpingpound Creek area (Beach, 1942; Hume, 1942; Ollerenshaw, 1976) give dip values rounded off to the nearest 5°; and information on fold styles and orientations is not published. Dozens of fold hinges are exposed in the area. In this study, fold styles are documented photographically and the positions of structural measurements are recorded on field sketches. Hundreds of strike and dip measurements have been collected in the study area for statistically valid structural analysis of fold geometries. A few of these folds are shown in Figures 2 and 3. Styles include upright and inclined, broad to tight, chevron, kink, and box folds (Fig. 2), and tight to isoclinal folds, in shales exhibiting pronounced thickening of their hinge regions (Fig. 2E, 3).

Few thrust fault surfaces are actually exposed in the study area. Thrusts have been mapped between outcrops that repeat strata, their dip directions estimated on the basis of axial plane orientations and vergence directions of nearby asymmetric folds. In areas where fold hinges are not exposed, outcrops have been carefully examined for evidence of the stratigraphic top. In several localities the younging directions, implied by Hume's (1942) and Ollerenshaw's (1976) maps, are incorrect. For example, the Cardium Formation is repeated three times at point X (near the southern border of Fig. 1). Hume (1942) and Ollerenshaw (1976) interpreted the middle one of these three outcrops (lat. 51°01'N; long. 114°36.7'W) as being the overturned limb of an anticline – syncline pair. However, crossbedding, scours, and graded bedding all indicate that the rocks are right way up, so two minor thrusts (that cannot be shown at the scale of Fig. 1) must separate the three Cardium outcrops. In other parts of the map area, panels of rock shown as right way up by Hume (1942) and Ollerenshaw (1976) are, in fact, overturned.

Folds and faults are interrelated in the study area, but many of them do not exhibit the classic geometries of Foothills structures described by Dahlstrom (1970) and Jamison (1987). Figure 4 shows vertical to slightly overturned beds of the Brazeau Group cut by a west-side-down contraction fault displaying a ramp-flat geometry. If a fault with this trajectory were a late feature, one would expect it to be either a west-side-up contraction (reverse) fault or perhaps a west-side-down extension (normal) fault on the limb of the anticline, instead of displaying the observed west-side-down contraction of bedding. If the beds in Figure 4 are rotated clockwise until they are horizontal and right way up, the fault geometry is like that of an early west-verging thrust sheet. Alternatively, the fault could be a late stage accommodation feature with the eastern half of the area wedged up vertically.

Seismic sections reprocessed by Slotboom (1992), examined in conjunction with the tight folds and numerous thrust faults mapped along Jumpingpound Creek, indicate that the shortening of Upper Cretaceous strata has been underestimated by previous studies. Cross-sections by Hume (1942), Bally et al. (1966), Gordy et al. (1975), and Ollerenshaw (1976) depict only 25 to 35 per cent shortening of the Cardium Formation between the Brazeau Thrust footwall cutoff and a pinline in the undeformed foreland (9 km northeast of the surface trace of the Brazeau Thrust). Our preliminary cross-sections indicate that shortening of the Cardium Formation is closer to 50 per cent in this easternmost portion of the Foothills, since subsurface deformation extends farther east beneath the Plains than previously thought.

FUTURE WORK

Balanced 1:20 000 scale cross-sections and palinspastically restored sections will be constructed during 1992-93, incorporating, via downplunge projection, the detailed surface mapping as well as subsurface data from the numerous wells and seismic sections made available for this study. Slotboom (1992) has reprocessed several of these seismic sections to more clearly show the shallow Mesozoic reflectors. His work will provide key information for tying subsurface structures to those we have delineated at the surface.

FIELDLOG software (Brodaric and Fyon, 1989) interfaced with AutoCAD[1] makes it possible to digitize and compile geological field data on the digital base maps each day. X, Y, and Z co-ordinates of drill hole data and depth-converted seismic reflectors will be entered into this database so that cross-sections in any orientation can be drawn at any scale, and data can be selected for further statistical and geometric analysis. Use of AutoCAD, FIELDLOG, and GeoKit[2] software will allow for updates and aid in the three-dimensional determination of positions of deeper lateral ramps. The software will also help with the three-dimensional representation of the geometry of the Triangle Zone in the Jumpingpound Creek area; the geometry will then be tied to structures along strike at Wildcat Hills (Spratt and Lawton, 1990) and Turner Valley (MacKay, 1991). Three-dimensional representations of palinspastically restored sections will aid in the visualization of duplex geometries and fault trajectories.

ACKNOWLEDGMENTS

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[1] AutoCAD is a registered trademark of Autodesk Inc.
[2] GeoKit is a registered trademark of Schreiber Instruments Inc.

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Geological Survey of Canada Project 870007

Analcite-bearing igneous rocks from the Crowsnest Formation, southwestern Alberta[1]

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Abstract: Analcite-bearing igneous rocks occur as small domes and/or flows associated with laharic and other debris flow deposits which contain large blocks of analcite-bearing rocks. Characteristic phenocryst assemblages are analcite+sanidine or sanidine+melanite+clinopyroxene. Physical character of the analcite weighs strongly against hypotheses that it represents an alteration product of leucite, although small amounts of easily recognized secondary analcite are locally present. Mafic inclusions in tuff, including a clinopyroxene megacryst, offer information on the igneous history and emplacement of the volcanics prior to eruption. The most plausible scenario involves derivation from an alkali basaltic parent via trachyte, with rapid ascent from magma chambers at depths greater than 10 km. This model can be tested and quantified using geothermobarometry. The economic potential of the rocks appears to be generally low.

Résumé : Des roches ignées contenant de l'analcime se présentent sous forme de petits dômes ou de coulées associées à des lahars ou à d'autres coulées de débris à grands blocs de roche contenant de l'analcime. Les assemblages caractéristiques de phénocristaux sont les suivants: analcime+sanidine ou sanidine+mélanite+clinopyroxène. Le caractère physique de l'analcime contredit nettement l'hypothèse selon laquelle celle-ci serait un produit d'altération de la leucite, bien qu'il existe par endroits de petites quantités d'analcime secondaire facilement identifiable. Les inclusions mafiques dans les tufs, notamment un mégacristal à clinopyroxène, fournissent des renseignements précieux sur l'évolution magmatique et la mise en place des roches volcaniques avant l'éruption. Selon le scénario le plus plausible, la roche serait issue d'un basalte alcalin parental qui aurait fait place à un trachyte en remontée rapide des chambres magmatiques dont la profondeur aurait dépassé 10 km. On peut mettre ce modèle à l'épreuve et le quantifier au moyen de la géothermobarométrie. Le potentiel exploitable des roches paraît généralement faible.

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INTRODUCTION

The Crowsnest Formation of Albian (Lower Cretaceous) age (Norris, 1964) consists mainly of lithic and crystal tuff and tuffaceous sandstone. Trachyte, analcite phonolite, and analcite-rich phonolite (blairmoreite) occur rarely as domes and/or flows and commonly as blocks in the tuff. The analcite-bearing rocks have attracted the attention of petrologists since their discovery more than a century ago (Dawson, 1885). Petrologists who examined the rocks in the field concluded that analcite formed as primary phenocrysts (MacKenzie, 1914; Crook, 1962; Pearce, 1967). Theoretical petrologists have generally concluded the opposite – that the analcite was not primary but secondary after leucite (Pirsson, 1915; Johannsen, 1938; Karlsson and Clayton, 1991).

A stability field for analcite in equilibrium with melt has been demonstrated by the experimental work of Peters et al. (1966) but this field is quite restricted in P-T space. If some or all of the analcite in the Crowsnest Formation is of igneous origin, then the rocks must have passed through this P-T field during their evolution and emplacement. If they did not pass through this field, then the liquid and subsolidus history must be unusual. We re-examined classic localities of igneous rocks in the Crowsnest Formation and possibly related rocks to establish the physical and chemical evolution of these unusual rocks. The economic potential of the Crowsnest Formation depends on the concentration and/or dispersion of elements during differentiation and emplacement. Our study therefore assesses this potential.

DISTRIBUTION AND STRATIGRAPHY OF THE CROWSNEST FORMATION

The Crowsnest Formation outcrops as a series of narrow belts over an area 70 km long from north to south and 20 km wide from east to west (Fig. 1). The formation appears in at least three major west-over-east thrust sheets, and is truncated on the west by the Lewis thrust which brings Paleozoic over Cretaceous rocks.

The Crowsnest Formation lies between nonmarine sandstones of the Lower Cretaceous Blairmore Group and marine black shale of the Upper Cretaceous Blackstone Formation of the Alberta Group. The lower contact is gradational, with pale grey feldspathic sandstone, locally glauconitic, at the contact. The Cadomin conglomerate, with distinctive pebbles of white chert and coal, commonly underlies the Crowsnest Formation and forms a convenient marker horizon. The upper contact is a disconformity with relief of up to 35 m on the pre-Blackstone surface (Pearce, 1970).

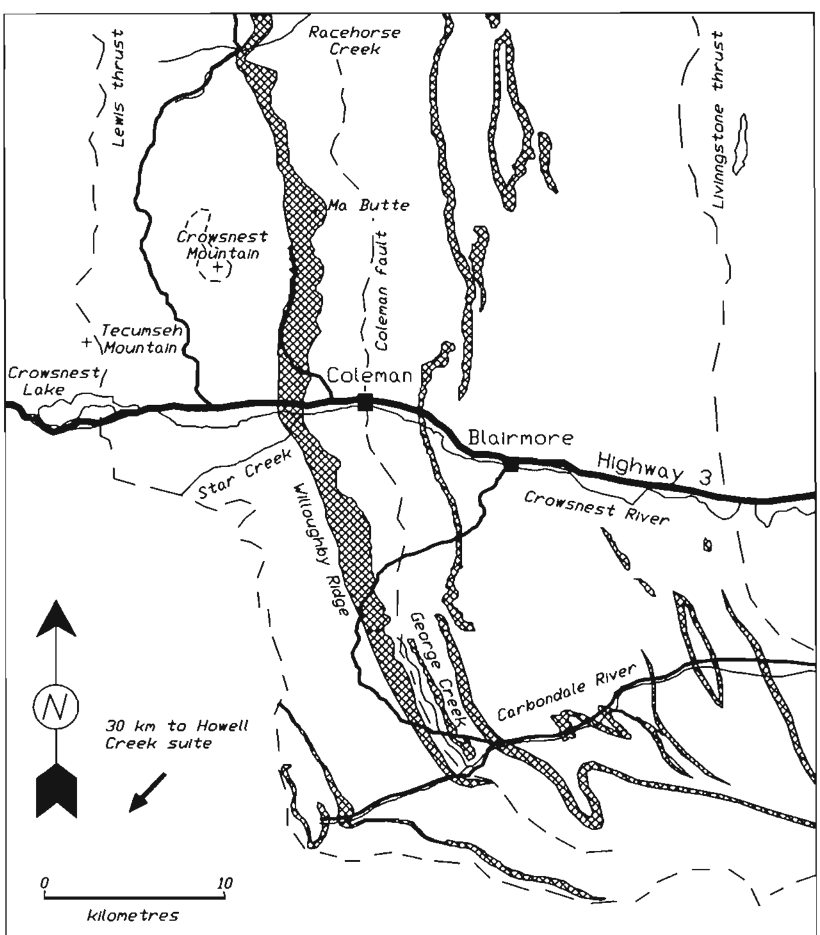

Figure 1.

Distribution of the Crowsnest Formation in the Coleman-Blairmore area, southern Alberta. The shaded areas are underlain by the Crowsnest Formation.

Many small faults of varying orientation and motion offset the strata making accurate determination of the thickness of formations difficult. However the thickest part of the Crowsnest Formation extends about 40 km from Racehorse Creek to the Carbondale River, and Pearce (1967) estimated a maximum thickness for the formation of about 425 m in the vicinity of Coleman. Thick sections occur around four localities of domes and/or flows on Racehorse Creek (49°49'N, 114°34'W), Ma Butte (49°43'N, 114°32'W), Star Creek (49°37'N, 114°32'W), and George Creek (49°30'N, 114°26'W). The distribution of material is compatible with aprons of lahars and other mass-flow deposits around domes, and more extensive, but thin, blankets of tuff, in part reworked by water. Thickness of Crowsnest Formation decreases rapidly eastward away from the domes and/or flows, from several hundred metres to less than 25 m within 10 km. Similar thinning probably occurs to the west, but the evidence is concealed beneath the Lewis thrust. In areas remote from the igneous centres there is a crude stratigraphy to the deposits, with a basal tuffaceous sandstone rich in white feldspar, and locally glauconitic, overlain by a coarse pink sandstone containing sanidine and melanite. Both of these units commonly display crossbedding. The upper part of the formation contains two or more units of crystal lithic tuff with fine grained, fragment-poor bases interpreted as air-fall material.

FIELD RELATIONS OF PRIMARY IGNEOUS ROCKS

Primary igneous rocks occur within the Crowsnest Formation as small domes and/or flows, and as multitudes of blocks within tuff and volcanic conglomerate. Some of these blocks exceed 1 m in size. The primary igneous rocks can be divided into three types based on megascopic phenocryst content, namely those containing sanidine+melanite (trachyte), those containing only analcite (blairmoreite), and those containing sanidine+analcite (analcite phonolite). There is complete gradation between various types marked by varying size distributions of the crystals. Clinopyroxene phenocrysts occur in all three varieties. Pearce (1967) mentions magnetite and titanite as phenocryst phases, but we have not recognized these minerals as megascopic phenocrysts, although titanite forms sparse microphenocrysts. We found no evidence of consistent stratigraphic order to phenocryst assemblages. All assemblages occur within a very fine grained matrix which appears to have been originally glassy. This matrix is jade green in the freshest specimens, but more commonly red or brown due to oxidation.

The most spectacular analcite-rich rocks contain analcite euhedra up to 4 cm across which may form 40% of the volume of the rock. Analcite crystals display both trapezohedral and icosahedral faces. Many of the crystals exhibit a homogeneous, resinous brown colour and are transparent, without cracks or obvious alteration, although a thin, white, altered rind is commonly present. Large analcite euhedra can be collected by breaking out the crystals with the fingers. These crystals exhibit delicate oscillatory zoning with thin brown bands rich in microphenocrysts. Such rocks contain abundant groundmass alkali feldspar but few feldspar phenocrysts and sparse mafics. We interpret these rocks to represent analcite cumulates entrained by residual liquid. Some analcite-rich blocks in tuff contain white analcite euhedra less than 5 mm across scattered evenly through an aphanitic matrix. These crystals are firmly welded to the matrix and cannot easily be recovered unbroken. The matrix contains microphenocrysts of analcite, melanite, clinopyroxene, sanidine, apatite, and rare titanite as well as syenitic clots and possible magma inclusions. We interpret these rocks to sample magma which crystallized primary analcite.

Fresh analcite is either resinous brown or white. However much analcite is rusty red, and looks altered presumably due to oxidation of a small iron content. Some large phenocrysts have recrystallized into mosaic analcite, and zoning is progressively erased. On Star Creek flinty hydrothermal alteration zones (termed "felsite dykes" by Pearce, 1967) exhibit haloes up to 1 m wide in which the analcite is opaque jade green and altered. Pearce (1967) reported this material gave an analcite X-ray pattern with an extra (200) reflection. We interpret it to be a low temperature phase probably produced by hydrothermal alteration.

Sanidine, like analcite, occurs in two distinct types of phenocryst. Spectacular, large phenocrysts of sanidine up to 8 cm long have length-to-breadth ratios of 2.5:1 or less. Commonly they lie in crude trachytic alignment. Many crystals exhibit colour layers with pink or red margins and paler cores. Large, tabular sanidine phenocrysts are consistently associated with melanite euhedra up to 5 mm in diameter. Smaller sanidine phenocrysts, up to 3 cm long, are more acicular, with length-to-breadth ratios greater than 3:1, and exhibit white or pale pink uniform colour. Analcite- rich rocks do not contain large sanidine crystals. Contrasting size distributions of sanidine presumably result from the interplay of analcite flotation and sanidine fractionation.

Clinopyroxene is sparsely present in all rock types, forming greenish black tabular euhedra with length-to-breadth ratios of 2:1 to 5:1. Considering only phenocryst minerals, the proportion of dark minerals (clinopyroxene and melanite) is small, commonly less than 5%. The igneous rocks of the Crowsnest Formation were all highly differentiated prior to eruption.

MAFIC AND OTHER INCLUSIONS

Although no mafic flows or dykes occur as part of the Crowsnest Formation, we found a variety of mafic inclusions, some of which may be cognate. Such inclusions tend to occur in clusters. A particularly rich location lies along the road from Blairmore to the Carbondale River on the west flank of the pass over Willoughby Ridge. This locality yielded a single pale yellow clinopyroxene megacryst 1x2x3 cm which contains inclusions of other minerals. A similar pale green clinopyroxene occurs in several melanite-bearing inclusions together with apatite. This material appears to be a cumulate. Pyroxenite xenoliths (grain size <2 mm) averaging 2 cm in length consist of black pyroxene, biotite, and magnetite euhedra, a mineral assemblage of alkaline affinity. We have also found single examples of greenish black ultramafic schist and an unusual coarse-grained, strongly deformed amphibolite.

In addition to mafic inclusions, a variety of other inclusions occur in the tuffs. Medium grained pink granitoid rocks, presumably derived from local crystalline basement are abundant. Medium grained pink syenite, possibly either metasomatized granite, or a plutonic equivalent to some of the volcanic rocks, is also abundant locally. Blocks of older cognate tuff up to 1 m across occur in all tuff beds. Autoliths of tuff coated by fine grained (air-fall?) material occur just west of Coleman. A tuff unit at this locality contains numerous melanite-bearing inclusions which are also rich in apatite, clinopyroxene, and pyrite. The euhedral pyrite contains fresh inclusions of clinopyroxene and apatite and appears to be an igneous phenocryst preserved in a cumulate assemblage. Blocks of the surrounding sedimentary strata are notably rare and small.

POSSIBLE CORRELATIVES OF THE CROWSNEST FORMATION

A number of igneous rocks west of the Lewis thrust have been proposed as correlatives of the Crowsnest Formation, including dykes on Crowsnest and Tecumseh mountains and at Crowsnest Lake, and the Howell Creek intrusive suite of southeastern British Columbia (Pearce, 1967; Carey, 1991; Price, 1962). We examined the dyke at Crowsnest Lake and the Howell Creek suite. The former is a fine grained grey trachyte with aligned sanidine phenocrysts up to 1.5 cm long. It contains no analcite or melanite and is strongly replaced by carbonate. Pearce (1967) reported the Tecumseh Mountain occurrence to be similar. Pearce (1967) reported the Crowsnest Mountain occurrence to be tinguaite. Nepheline has not been reported from the Crowsnest Formation. The Howell Creek suite is poorly exposed, but quite diverse, ranging from trachytic rocks like Crowsnest Lake, through medium grained syenitic rocks with sanidine laths and melanite, to pegmatitic biotite syenite. All of the igneous rocks contain traces of quartz. Substantial amounts of sedimentary rocks outcrop within the area shown by Price (1962) as Howell Creek suite. The field relations suggest a dyke complex, possibly involving several suites of dykes.

Our observations suggest that none of the suggested hypabbysal correlatives of the Crowsnest Formation exhibit similar composition or phenocryst assemblages. Crowsnest Formation igneous rocks are undersaturated (Pearce, 1967) but did not lie in a nepheline-stable field. All the suggested correlatives lie west of the Lewis thrust. If they are the same age as the Crowsnest Formation, they were transported tens to hundreds of kilometres eastward subsequent to emplacement. The rapid decrease in thickness of the Crowsnest Formation away from a central axis makes it very improbable that cognate magmatism extended so far. We conclude that none of the proposed correlatives west of the Lewis thrust is closely related to the Crowsnest Formation.

GEOTHERMOBAROMETRY

The stability field of analcite in equilibrium with melt lies between 600 and 650°C and 6-14 kbars water pressure (Peters et al., 1966) although the addition of potassium to the system markedly lowers the water pressure required (Peters et al., 1966). If analcite is a primary phenocryst in the Crowsnest Formation, the melt must have been low-temperature, high pressure, and water-rich. Conditions under which the phenocrysts formed can, in principle, be determined from compositions of coexisting minerals. The required analytical data do not presently exist, but will be collected in the coming months. The following reactions are suggested by reported phenocryst assemblages:

$$NaAlSi_3O_8 \text{ (sanidine 1)} = NaAlSi_3O_8 \text{ (sanidine 2)}$$

$$6Fe_2TiO_4 + O_2 = 2Fe_3O_4 + 6FeTiO_3$$

$$NaAlSi_2O_6 + H_2O = NaAlSi_2O_6{\cdot}H_2O$$

$$18CaTiSiO_5 + 16Fe_3O_4 = 6Ca_3Fe_2Si_3O_{12} + 18Fe_2TiO_4 + 5O_2$$

$$3Ca_3Fe_2Si_3O_{12} + Fe_3O_4 + 9NaAlSi_3O_8 + 9H_2O = 9CaFeSi_2O_6 + 9NaAlSi_2O_6{\cdot}H_2O + 2O_2$$

$$2KAlSi_3O_8 + 2Fe_3O_4 + H_2O = 2KAlFe_3Si_3O_{10}(OH)_2 + O_2$$

$$2NaAlSi_2O_6{\cdot}H_2O = NaAlSiO_4 + NaAlSi_3O_8 + H_2O$$

The first five reactions have reached equilibrium within the Crowsnest volcanics, while the latter two represent limiting cases since neither biotite nor nepheline are present. From these seven reactions the parameters P, T, water fugacity, and oxygen fugacity need to be determined. Internally consistent thermodynamic data for all phases were tabulated by Berman (1988), and adequate solution models already exist for all phases so that determination should be straightforward.

ECONOMIC CONSIDERATIONS

Elevated gold values have been reported from the Crowsnest Formation near Coleman (Johnson, 1989). The tuff in this region contains a significant pyrite content associated with mafic inclusions. Other parts of the formation contain negligible amounts of pyrite, and negligible gold values. We consider it probable that the gold is associated with pyrite. The mafic inclusions probably represent cumulates from magma parental to the Crowsnest Formation but if so, the source is several kilometres below the surface.

Differentiation of water-rich peralkaline rocks commonly leads to extreme concentration of rare-earth elements (Miller, 1989). Rare-earth values for the Crowsnest Formation have been reported to be erratic (Lambert et al., 1987) but there is little systematic information.

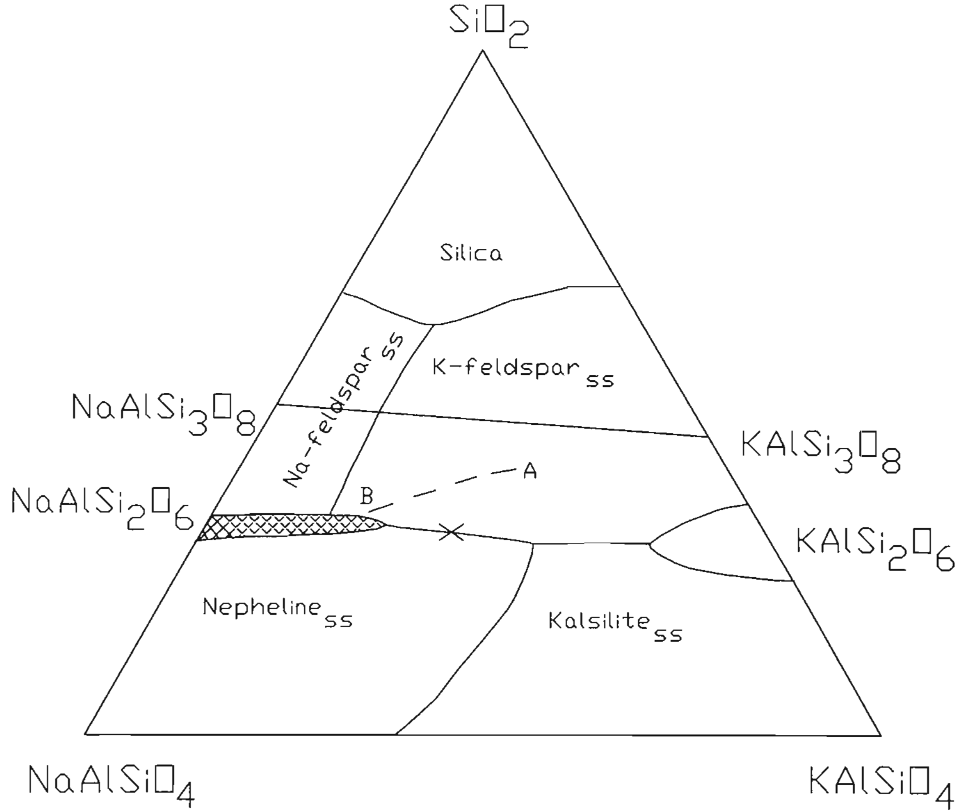

Figure 2. Schematic equilibrium diagram for $NaAlSiO_4$-$KAlSiO_4$-SiO_2 at P_{H_2O}=6 kbars (>15 weight per cent H_2O) projected onto the anhydrous base. Lavas of the Crowsnest Formation are assumed to project near point A and initially crystallize K-rich feldspar, with residual liquid progressing to B at which point analcite+sanidine begins to crystallize. A pressure drop at this point forces liquid into an analcite-only region (Peters et al., 1966). Phase boundaries as shown are compatible with the experimental work of Zeng and MacKenzie (1984) and Peters et al. (1966), but not required by them. In the configuration shown, nepheline is not stable with Na-feldspar. The "X" marks the approximate position of the thermal maximum on the nepheline-K-feldspar join. The shaded area crystallizes analcime only.

The Howell Creek suite is weakly mineralized with copper and lead (Price, 1962), which would imply some potential in the Crowsnest Formation if the igneous rocks are correlative. We believe that the two are not correlative, and that the base metal potential of the Crowsnest Formation is low due to the rarity of sulphides.

Some of the blairmoreite has a striking appearance and attractive colour, and could have some potential as decorative stone. It would be difficult to quarry large blocks of this material due to pervasive fracturing. Analcite has a number of industrial uses due to its ion-exchange properties. We have not attempted to assess the possibilities of an industrial mineral operation, but believe them to be low due to iron contamination.

DISCUSSION

Field examination of the Crowsnest Formation strongly supports the view that it contains primary igneous analcite phenocrysts. The analcite euhedra are fresh and unaltered, do not display cracks, and contain no zonal inclusions. If they were originally leucite, altered at subsolidus temperatures to analcite, alteration would lead to an 11% increase in volume, producing cracks in the analcite and surrounding host. Such a phenomenon is not generally present. Formation of low temperature analcite should result in expulsion of trace elements admitted at higher temperatures, particularly iron. This phenomenon can be seen locally in altered analcite, but is absent in fresh material. Leucite characteristically exhibits zonal inclusions, particularly in large crystals. Such inclusions are absent from the large, fresh analcite euhedra. Given the freshness of many of the igneous rocks, and the large size of many analcite euhedra, some relics of any primary leucite should surely be preserved.

Karlsson and Clayton (1991) have recently argued in favour of the primary leucite hypothesis, citing isotopic evidence that the analcite presently contains meteoric water. Isotope exchange with meteoric sources is common in igneous rocks, even in such classic igneous complexes as the

Skaergaard massif (Norton and Taylor, 1979) but commonly does not destroy primary mineralogy. We see no field evidence that it did so in the Crowsnest Formation.

We suggest that field evidence favours an igneous descent of the analcite-rich rocks from a sanidine-phenocrystic parent. This is the reverse of the succession proposed by Pearce (1967), who thought that the crystallization of analcite led to subsequent crystallization of sanidine. However sanidine is well known as a phenocryst at temperatures much higher than the stability of analcite, and would permit a reasonable line of liquid descent from an ultimate alkali basaltic parent via trachyte. Such a line of descent should be witnessed in trace element signatures, and this line of investigation will be pursued. We assume that the trachytic parent crystallized sanidine with about 25% $NaAlSi_3O_8$. In the projection $NaAlSiO_4$-$KAlSiO_4$-SiO_2 (Fig. 2) the residual liquid is driven away from this composition, becoming progressively enriched in $NaAlSi_2O_6$ component. A stability field for primary analcite is demonstrated by the experiments of Peters et al. (1966). Liquid reaching the boundary of this field begins crystallizing sanidine+analcite. Peters et al. (1966) showed that the liquid coexisting with this assemblage becomes enriched in albite component with decrease of pressure. A decrease in pressure, for example due to upwelling of magma, will cause the liquid to pass into the analcite-only field. This sequence could explain the observed assemblages.

If it is assumed that analcite forms a primary phenocryst, the emplacement of the igneous rocks must also be explained. For pure Na-analcite the magma would have to rise from a minimum 14 km (4.75 kbars) without either disrupting or resorbing the phenocrysts. According to Peters et al. (1966), the rise could be less in a potassic system, but the minimum would be 7 km (2.3 kbars). Since the magma must be extremely water-rich (15 weight per cent H_2O) in order for analcite to be stable, decompression should lead to explosive degassing. This model is compatible with the tuffaceous character of most of the Crowsnest Formation, and the lack of local sedimentary fragments. The local domes of igneous rock presumably represent rise of small amounts of degassed magma that was not totally disrupted by outgassing. Analcite could be preserved in such magma by armouring with a thin reaction rim, such as that observed in even the freshest phenocrysts. The water-rich, highly fractionated character of the magma requires that it solidify on loss of water, but small amounts could still move on lubricating films of vapour. We believe that this model can be quantified, and hope to do so in further work.

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Geological Survey of Canada Project 680071

Preliminary study of the surficial geology of Virden area, southwestern Manitoba

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Sun, S., 1993: Preliminary study of the surficial geology of Virden area, southwestern Manitoba; in Current Research, Part B; Geological Survey of Canada, Paper 93-1B, p. 57-61.

Abstract: Preliminary study of the surficial geology of the Virden area leads to the following interpretation of the deglacial history: (1) Glacial Lake Hind developed as an ice marginal lake impounded by ice to the north and east and by topography to the south and west. (2) At one time the ice margin stood along the modern Assiniboine River. (3) Large volumes of meltwater flowed into glacial Lake Hind through channels and ice tunnels. (4) Glacial Lake Hind was drained after excavation of the Assiniboine channel, through Brandon, to Lake Agassiz.

Résumé : L'étude préliminaire de la géologie des formations en surface dans la région de Virden a permis l'interprétation suivante des phases de la déglaciation dans la région de Virden: 1) le lac glaciaire Hind se serait constitué sous forme de lac proglaciaire, dont les eaux étaient retenues par les glaces au nord et à l'est et par la topographie au sud et à l'ouest. 2) À une certaine époque, la marge des glaces se serait située en bordure de la rivière Assiniboine actuelle. 3) De vastes volumes d'eaux de fonte se seraient écoulés dans le lac glaciaire Hind par des chenaux et par des tunnels sous-glaciaires. 4) Les eaux du lac glaciaire Hind se seraient vidées, après le creusement du chenal Assiniboine, en passant par Brandon, dans le lac Agassiz.

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INTRODUCTION

Mapping of the Virden map area (62F; Fig. 1) was started the summer of 1992 as part of the Prairie NATMAP Project. The objectives of this project are to produce surficial materials maps of the area at a scale of 1:100 000, to prepare a GIS-compatible database of all surficial geology information, and to prepare a comprehensive final geological report. Farming is the dominant industry in the Virden region but this is also the heart of Manitoba's oil industry, and the area is crossed by important transportation links such as the Trans-Canada Highway, and interprovincial oil and gas pipelines. The information gathered through this project will aid in exploitation and protection of aquifers, in identification of soil quality problems related to surficial sediments, in location and exploitation of granular resources, and in mitigation of environmental problems effecting surface materials.

The summer's work was restricted to the northeast quarter of the map area (62 F/9, 10, 15, 16). This area makes up most of the basin of glacial Lake Hind which was named by Elson (1956). It is planned that the history of glacial Lake Hind will form the basis of a Ph.D. thesis at the University of Manitoba.

PHYSIOGRAPHY AND DRAINAGE

The study area is entirely within the Saskatchewan Plains and can be subdivided into 3 physiographic areas as defined by Klassen (1979): the Souris Basin, the Assiniboine River Plain, and the Souris River Plain.

There are two main rivers in this area. Assiniboine River enters the area from the northwest and then flows eastward. Souris River enters from the southwest and then flows eastward to leave the area via the Pembina Spillway. Another major stream, Pipestone Creek, flows into the area from the west and ends in Oak Lake, a shallow body of water which lies near the western edge of what was once the basin of glacial Lake Hind (Fig. 2).

PREVIOUS STUDY

Bedrock geology in the Virden area was studied by Wickenden (1945). Bedrock topography was mapped by Klassen and Wyder (1970) and Betcher (1983). Surficial sediments in the Virden area were first studied by Upham (1890), and were mapped by Elson (1962) from 1953 to 1955. The work of Elson was largely followed in recompilations of the surficial geology by the Aggregate Resources Division (1980) of the Manitoba Department of Energy and Mines, and by the Water Resources Branch of the Manitoba Department of Natural Resources (Betcher, 1983). Soil survey has been done by Ehrlich et al. (1956). Subsurface sediments have been studied by several workers including Klassen and Wyder (1970), Betcher (1983), and Manitoba Water Resources Division (1976). These studies show that subsurface sediments in most of the area consist of thick till overlain by thick clay and silt, which in turn is overlain by relatively thin sand. Few radiocarbon dates are available for the area. Those that are available are mainly from the alluvial fill of Assiniboine Valley and are reported by Klassen (1972). The oldest of these, on a piece of wood from Assiniboine Valley fill near Virden, yielded a radiocarbon age of 11 600 ± 430 BP (GSC-1081, Lowdon et al., 1971) and is a minimum age for final deglaciation and drainage of glacial lakes from the area.

BEDROCK GEOLOGY

The study area is underlain by Upper Cretaceous shale which was originally assigned to the Fort Pierre Group (recognized in Nebraska) by Tyrrell (1890). Tyrrell divided the Fort Pierre into two "series", the Millwood, soft dark grey clays and shales, and the Odanah, light grey hard clays and shales. In his influential publication on the Mesozoic stratigraphy of the eastern plains, Wickenden (1945) used the term Riding Mountain Formation for the unit which included the Millwood and Odanah. McNeil and Caldwell (1981), in their extensive work on the Cretaceous of the Manitoba Escarpment, proposed returning to the original terminology and referred to the Upper Cretaceous shale as the Pierre Formation. They recognized the Millwood and Odanah members and inserted an unnamed member at the top to include largely unstudied soft shales and clays which are poorly exposed in the Turtle Mountain area.

From the standpoint of surficial units in the study area, the shales are rarely seen at the surface. The Millwood Member is a soft greenish-brown, bentonitic, silty shale which outcrops along Assiniboine Valley and underlies thick drift in the area east and north of Oak Lake. The Odanah Member, which is a hard grey siliceous shale that outcrops along

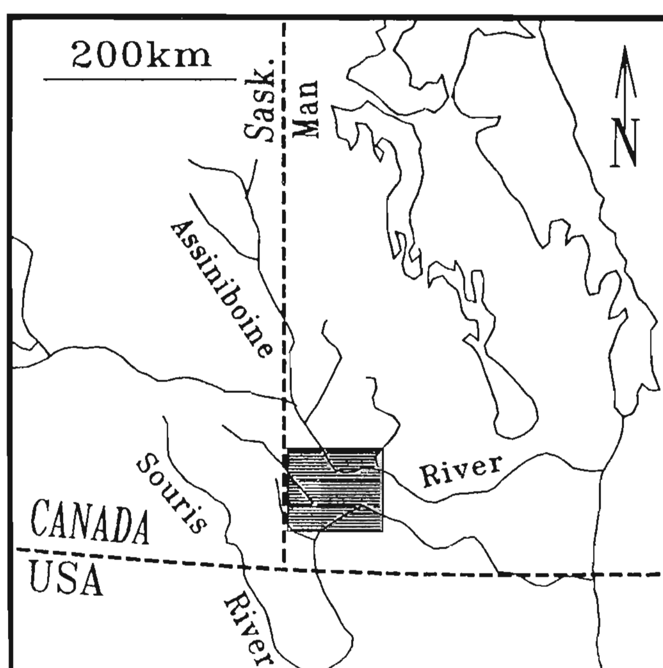

Figure 1. Location of the study area.

Assiniboine Valley west of Griswald, is overlain by thin till or glacial fluvial sediments in the area north of the Assiniboine and west of Kenton, and underlies thick drift west and south of Oak Lake (Klassen and Wyder, 1970; Betcher, 1983).

SURFICIAL SEDIMENTS

Most of the first season's field work was concentrated in map areas 62F/10 and 15. Three types of surficial sediments predominate in this area.

Till

Lennard and pre-Lennard till

Four till units have been identified in a section exposed in roadcuts on the east side of Pipestone Creek (sec. 3, tp. 10, rge. 29W). Based on stratigraphic position, these four units are correlated to the Tee Lake, Shell, Minnedosa, and Lennard formations described by Klassen (1979). In fresh cuts these tills vary from dark olive grey to olive grey, while oxidized faces are lighter. The oxidized Shell till is olive with a reddish tone, while oxidized tills in the Tee Lake Formation and Minnedosa Formation are pale olive. The Lennard Formation is a darker olive. Further studies will be conducted on this, the only exposure of multiple till units found this summer.

The lithology of the till exposed at the surface in the area studied is identical to the Lennard till in the Pipestone Creek section. In general, it contains more shale fragments and carbonate clasts than are found in the underlying tills. In a narrow area from Assiniboine River eastward to the town of Kenton this surficial till is thin or absent and consists of more than 80% shale. The high shale content is due to the large amount of shale that was incorporated into ice as it overrode the easily eroded shale exposed at the surface of this area. The morphology of Lennard till is dominated by low linear ridges (corrugated moraine, Klassen, 1979; 1989, p. 149) west of Virden, and is flat to undulating near Assiniboine River.

End moraine

An end moraine, which is about 2 km long north to south and 0.3 km wide from west to east, occurs west of Little Saskatchewan River at sec. 28, tp. 11, rge. 21W. Sediments in the end moraine consist of Arran till (Klassen, pers. comm., 1992). The end moraine stands 1 to 2 m above the area to the

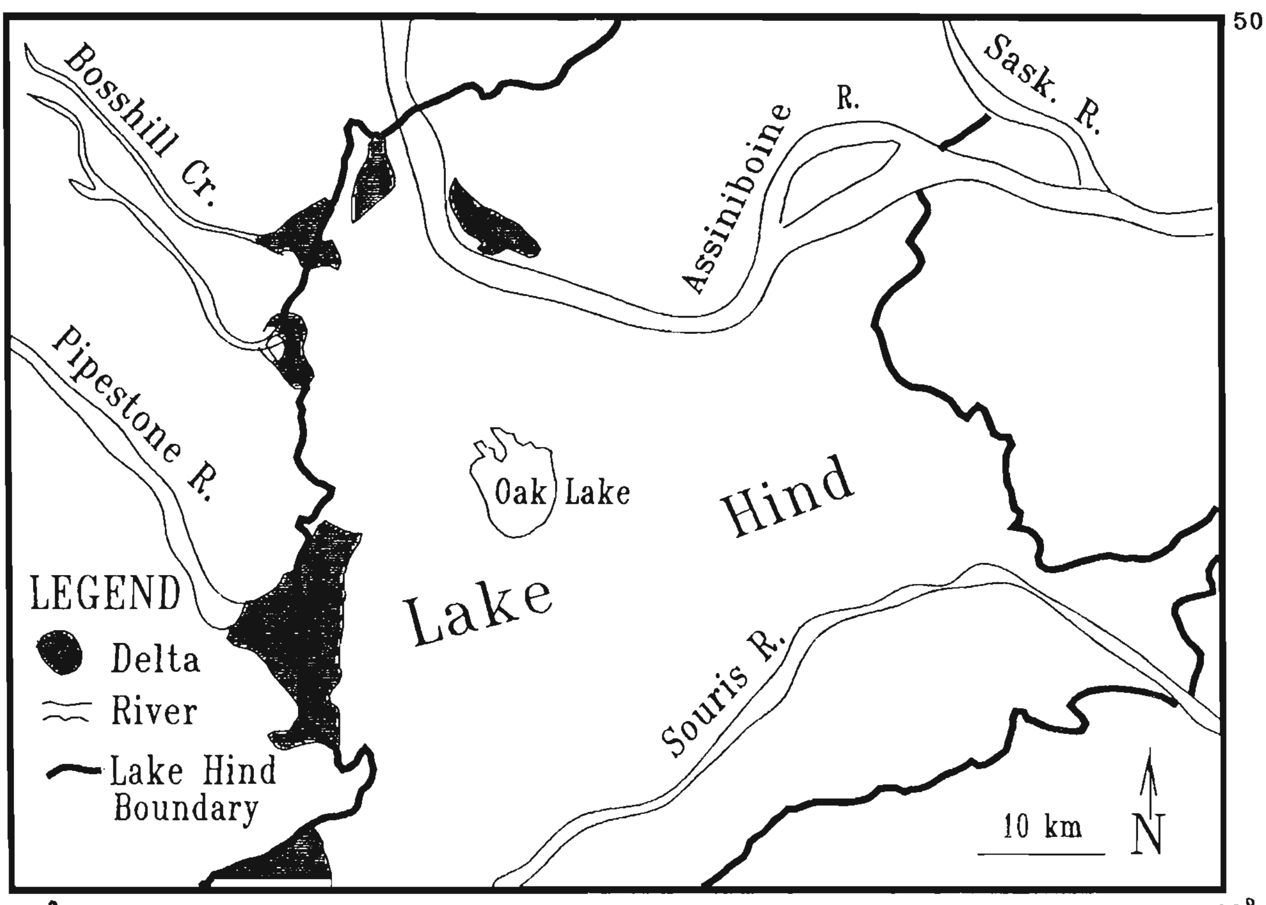

Figure 2. Drainage system and deltas in the glacial Lake Hind basin.

west and 3 to 4 m above the area to the east. To the west, the till interfingers with stratified outwash sand/gravel or laminated lacustrine clayey silt. To the east, surficial sediments are dominated by compact till overlain by thin lacustrine sand or clay. This end moraine may indicate the ice marginal position of the Red River Ice Lobe which late during the last glaciation advanced westward from the basin of glacial Lake Agassiz (Elson, 1956, p. 283).

Glaciofluvial sediments and features

Esker

A major esker complex lies north of Assiniboine River, north of Oak Lake (Fig. 3). This complex consists of a series of anastomosing ridges that are up to 0.5 km in width near Assiniboine River and 3 km in width at the northern end near Lenore. Ridges of the esker complex are more continuous in the southern half than the northern half. Sediments in the esker are dominated by shale fragments with well preserved crossbeds, which are commonly overlain by horizontally laminated sandy silt. Two low ridges at the eastern side of the esker merge with the main esker to the north (Fig. 3) and disappear southward under lacustrine silt. Sediments in the two low ridges consist of sand and fine shale fragments with no visible sedimentary structures.

Delta

Deltas were deposited along the western margin of glacial Lake Hind by Assiniboine River, Pipestone Creek, Bosshill Creek, and Gopher Creek (Fig. 2). The Souris delta near Melita is south of the area studied this summer. Sediments in the deltas consist of sandy gravel near the river mouths, grading to fairly well sorted sand near delta fronts. The fronts of all these deltas are at an elevation of 440 m.

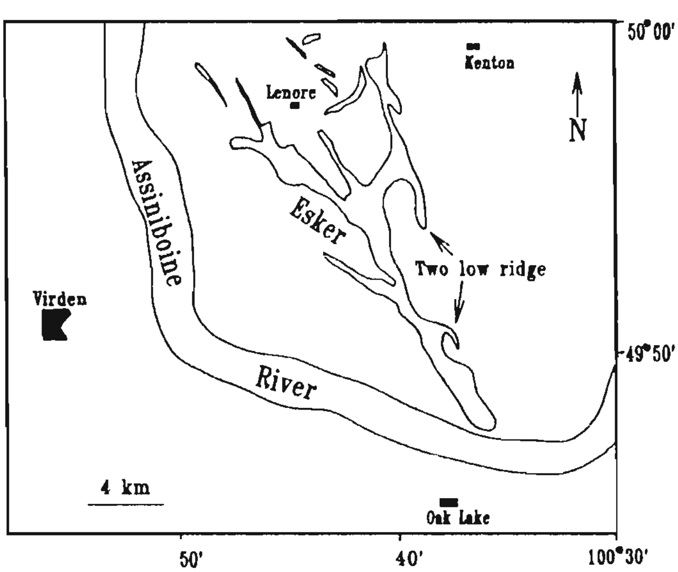

Figure 3. An esker in the northern part of glacial Lake Hind.

Lacustrine sediments and morphology

Lacustrine sand

Lacustrine sand occurs in the area between Assiniboine River and Souris River. Sediments are dominated by well sorted to moderately well sorted fine to medium sand. The surface of the lacustrine sand is generally flat except in areas that have been reworked by the wind, where it is undulating to rolling.

Lacustrine silt/clay

Lacustrine silt and clay occur throughout the glacial Lake Hind basin. In the area between Assiniboine and Souris rivers, lacustrine silt and clay beds are thick (>40 m, R.W. Klassen, unpub. information) and overlain by lacustrine sand. North of Assiniboine River lacustrine silt and clay beds are thinner and are exposed at the surface. They have a flat depositional surface but in a number of areas display a highly rolling morphology, which is inherited from the underlying till. The content of dropstones in sediments generally increases eastward. Sedimentary structures in outcrops are dominated by horizontally bedded clay and silt couplets.

PRELIMINARY LATE QUATERNARY HISTORY

At the onset of Lennard Glaciation, de4uring the Late Wisconsinan, the Assiniboine Ice Lobe advanced toward the south and southeast (Klassen, 1979). As this lobe retreated, meltwater was impounded in glacial Lake Hind by the Assiniboine Ice Lobe to the north and the Red River Ice Lobe to the east. Early during the lake phase, the Assiniboine ice margin stood along the modern Assiniboine valley, as indicated by the termination of the major esker at Assiniboine River (Fig. 3). Sediments deposited along the ice margin were either transported by meltwater or reworked by lake water, evidenced by laminated lacustrine sediments overlying bedrock in several exposures along Assiniboine River.

Lacustrine silt and clay sediments occur from an elevation of 420 m to 450 m, suggesting that water level of Lake Hind once reached as high as 450 m. This probably occurred when the Pembina spillway was blocked by advance of the Red River Ice Lobe at about 11 200 BP during the Lockhart Phase of glacial Lake Agassiz [as suggested by Conley (1986) and Fenton et al. (1983, p. 66)]. At that time lake water was forced to flow southward through the Sheyenne spillway to glacial Lake Agassiz. There is little evidence of this high phase in Souris Valley south of the study area, other than lacustrine sand overlying gravel in a few locations.

Re-opening of the Pembina spillway lowered the level of the glacial lake and most of the deltas in the study area were then built at an elevation of 440 m. During that time lacustrine sand was deposited over lacustrine silt and clay because of the shallow water depth and abundant sediment supply. A readvance of the Assiniboine ice apparently occurred after these events as till overlies deltaic and lacustrine silt sediments 10 km north of Virden.

After the Red River Ice Lobe retreated from the Brandon area, Lake Hind drained into Lake Agassiz through Assiniboine Valley.

ACKNOWLEDGMENTS

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Geological Survey of Canada Project 910024

Quaternary geology of the Bissett area, southeastern Manitoba: applications to drift prospecting[1]

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Henderson, P.J., 1993: Quaternary geology of the Bissett area, southeastern Manitoba: applications to drift prospecting; in Current Research, Part B; Geological Survey of Canada, Paper 93-1B, p. 63-69.

Abstract: Quaternary geological studies in the Bissett area, Manitoba are designed to aid mineral exploration by establishing a geological framework for interpreting glacial dispersal of geochemical and mineralogical components of till associated with bedrock mineralization. The area is underlain by the Archean Rice Lake greenstone belt, a volcano-sedimentary sequence with associated intrusions, that has known gold and base metal deposits. Ice flow indicators and distribution patterns of subglacial and ice-marginal sediments indicate that the region was glaciated by ice flowing southwest during the last glaciation. Drift prospecting within the Wanipigow River valley, the major valley underlain by mineralized rocks of the greenstone belt, is complicated by thick deposits of glaciolacustrine sediments. Uplands areas are characterized by a single till unit occurring as a thin veneer over bedrock. Regional till sampling concentrated in uplands south of the mineralized zone should indicate areas for follow-up studies.

Résumé : Les études géologiques du Quaternaire réalisées dans la région de Bissett au Manitoba sont conçues pour venir en aide aux travaux de prospection minérale en établissant un cadre géologique permettant d'interpréter la dispersion glaciaire des composantes géochimiques et minéralogiques du till qui sont associées à la minéralisation du substratum rocheux. Le sous-sol de la région se compose de la zone archéenne de roches vertes de Rice Lake, séquence volcano-sédimentaire accompagnée d'intrusions, qui renferme des gisements connus d'or et de métaux communs. Les indicateurs de l'écoulement des glaces et les schémas de répartition des sédiments sous-glaciaires et de contact glaciaire montrent que la région a été recouverte par des glaces qui s'écoulaient vers le sud-ouest au cours de la dernière glaciation. La prospection glacio-sédimentaire de la vallée de la rivière Wanipigow, à savoir la principale vallée dont le sous-sol contienne des roches minéralisées appartenant à la zone de roches vertes, est rendue difficile par la présence d'épais dépôts glaciolacustres. Les régions de plateaux sont caractérisées par une seule unité de till qui se présente sous forme d'un mince placage recouvrant le substratum rocheux. L'échantillonnage régional du till, qui a principalement porté sur les plateaux situés au sud de la zone minéralisée, pourrait indiquer les secteurs se prêtant à des études complémentaires.

[1] Contribution to Canada-Manitoba Partnership Agreement on Mineral Development 1990-1995, a subsidiary agreement under the Economic and Regional Development Agreement. Project funded by the Geological Survey of Canada.

INTRODUCTION

Surficial geology mapping at the 1:50 000 scale and till sampling were initiated in the Bissett area, Manitoba (NTS 62P/1 and 52M/4) as part of the joint Canada-Manitoba Partnership Agreement on Mineral Development (Fig. 1). This field season marked the first in a three-year program designed to aid mineral exploration by studying the Quaternary geology and glacial history of the area in order to interpret dispersal patterns of those chemical and mineralogical components of till associated with bedrock mineralization, by establishing a regional geochemical data base, and by determining drift prospecting techniques suitable to the region. The area has a high potential for gold and base metal mineralization with many reported showings and occurrences dating back from the first gold discovery in 1911 (Theyer, in press). Over 1.5 million oz. gold have been produced from the region with 80% coming from the San Antonio mine in Bissett, which is no longer in production (Stephenson, 1971).

Prior to this study, the Quaternary geology of the area was mapped at 1:100 000 based on air photo interpretation with limited ground checking (Nielsen, 1980). Although no regional till sampling has been done, local studies relating to aggregate resources (Berssenbrugge, 1986) and placer gold deposits of the Manigotagan area (NTS 62P/1) have been published (Nielsen, 1986; Nielsen and DiLabio, 1987).

During the 1992 field season, investigations were concentrated within the eastern half of the study area (NTS 52M/4 Bissett). They included the examination of surficial deposits and stratigraphic sections, the measurement of striae and other ice flow indicators, and till and humus sampling. Over 120 till samples were collected 1-5 km apart, from hand-dug pits approximately 1 m deep, and from natural and man-made exposures. Surface samples of humus were also collected at most sites. All samples are presently being analyzed for trace metal concentrations. In addition, tills are being analyzed for carbonate content and texture.

The sampling was focused along a 5-6 km wide corridor paralleling the main road (Highway 304) through the area and roughly coincident with the zone of mineralization. Access was by truck along major roads, all-terrain vehicle along trails and abandoned logging roads and by boat on the major lakes (Rice, Wanipigow, Kaneesho) and along Beaver Creek. This paper will outline the major observations from this field work and summarize preliminary conclusions with implications for drift prospecting.

REGIONAL SETTING

The study area is located in southeastern Manitoba near the margin of the Superior structural province of the Canadian Shield (Fig. 1). It is underlain primarily by Archean supracrustal and intrusive rocks of the southeasterly-trending Rice Lake greenstone belt which is in fault contact to the north and south with plutonic and gneissic rock types (McRitchie and Weber, 1971). The overlying Paleozoic quartz sandstone and limestone outcrops along the western edge of the study area (NTS 62P/1 English Brook), on the shore and islands in Lake Winnipeg.

Glaciation occurred by ice flowing from the northeast (Fig. 1), presumably from a Labradoran dispersal centre (Manitoba Mineral Resources Division, 1981). The confluence between this ice flow and southeasterly flowing ice in the interlake area (the Red River Lobe) is preserved in a major moraine south of Lake Winnipeg which may extend as far north as the western margin of the study area (Prest et al., 1968). During ice retreat, the entire area was inundated by glacial Lake Agassiz resulting in extensive deposition of fine

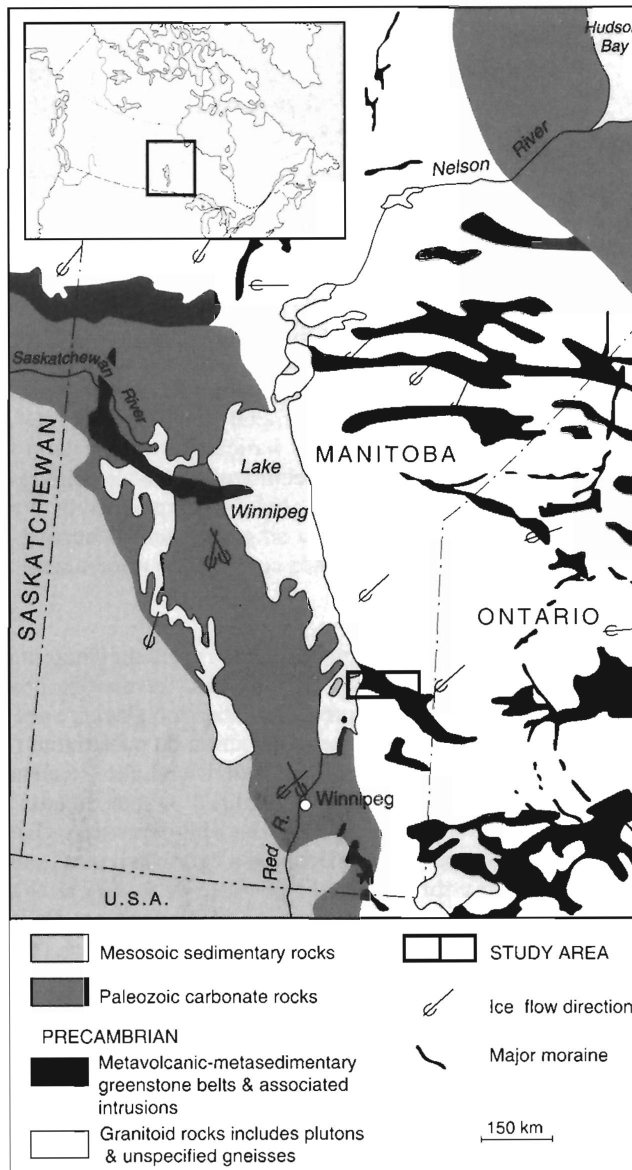

Figure 1. Location map for the study area in southeastern Manitoba showing major moraines and regional ice flow trends (Prest et al., 1968).

grained glaciolacustrine sediments within the major valleys (Nielsen, 1980). Models for the deglaciation of southern Manitoba and the probable extent of Lake Agassiz are presented in Elson (1967), Teller and Clayton (1983) and Dyke and Prest (1987) among others.

The physiography of the Bissett area is typical of Shield terrane. It is characterized by smooth, bare to thinly drift covered outcrops separated by low, generally muskeg-covered ground. Relief is low with few outcrop ridges over 40 m high. Drainage is westward to Lake Winnipeg. The major river valleys are wide and drift-filled. The Wanipigow River valley, underlain by mineralized rocks of the greenstone belt, has fairly steep margins which coincide with major structural lineaments. The area supports a dense forest, however, some parts have been extensively logged or burned by forest fires allowing access with all-terrain vehicles.

BEDROCK GEOLOGY AND MINERALIZATION

The Rice Lake greenstone belt trends east-southeast across the area (Stockwell, 1938; Russell, 1949; Davies, 1950; McRitchie and Weber, 1971). It is one of several greenstone belts within the Red Lake Archean fault block which extends across northern Ontario and consists of folded and fractured metavolcanic and metasedimentary rocks of the Rice Lake Group intruded by quartz diorite plutons, mafic sills and dykes and ultramafic plugs, and overlain by metasedimentary rocks of the San Antonio Formation. Margins of the greenstone belt are fault-bounded. The Manigotagen gneissic belt, to the south, consists of a suite of paragneiss, schist, tonalite, and monzonite while the Wanipigow River plutonic complex to the north and east is composed of quartz diorite, granodiorite, and gneissic rocks. Ultramafic intrusions occur as discontinuous serpentinized lenses within the Wanipigow River plutonic complex (Scoates, 1971).

Mineralization in the area is structurally controlled. It is confined predominantly to auriferous quartz veins in shear zones hosted by mafic intrusions and directly associated with carbonate-chlorite-sericite alteration of veins and adjacent sheared country rock. Gold occurs as particles in microcrystalline quartz or disseminated in sulphide minerals, particularly pyrite, arsenopyrite, and chalcopyrite, occurring as small inclusions or brecciated veinlets. Gold content of the veins varies widely depending on the chemical and mineralogical composition of the enclosing wall rock (Amukun and Turnock, 1971). Occurrences are erratically distributed and, in several places, have exhibited high concentrations with small tonnage.

Placer gold in low concentrations has also been reported from Quaternary sand and gravel deposits in the Manigotagan area. Nielsen (1986) suggests that these gold grains were incorporated into till through glacial erosion of a local bedrock source, reworked from till deposits, and concentrated through nearshore littoral processes associated with Lake Agassiz.

Within the area, mafic and ultramafic bodies have been examined as potential environments for massive sulphide and/or platinum group mineralization (Theyer, 1987; Weber, 1991). In Manitoba, there is a relationship between nickel sulphide deposits and discrete ultramafic intrusions and differentiated layered bodies (e.g. Lynn Lake gabbro, Bird River Sill), however these minerals are rare in the Rice Lake area (Scoates, 1971). Similarily, only trace amounts of sulphides have been found in the mafic and ultramafic bodies of the English Brook area (Young and Theyer, 1990); however, hydrothermal alteration in mafic metavolcanic rocks indicates a potential for volcanogenic massive sulphide deposits (Weber, 1991).

ICE FLOW INDICATORS

Throughout the area exposed bedrock surfaces are striated, polished, and glacially moulded. The dominant striation trend ranges from 240° to 250° indicating regional ice flow toward the southwest (Fig. 2). There is a continuum, however, in striation observations ranging from 227° to 260°, which appears to be reflected regionally by a slight westerly swing in apparent ice flow associated with the Wanipigow River valley. Several ice flow trends have been recognized by striations preserved on protected outcrop faces or crosscutting relationships (Fig. 3) and suggest that ice flow at 260° and 236° postdated the dominant regional flow at approximately 248°. The significance of these relative ages is unknown.

SURFICIAL SEDIMENTS

In topographically higher areas dominated predominantly by rocks of the Wanipigow Plutonic complex and the Manigotagen Gneissic Belt, drift cover is thin and discontinuous. Thick deposits up to 10 m are confined to the Wanipigow and Manigotagen River valleys and coastal areas of Lake Winnipeg. These consist predominately of glaciolacustrine rhythmically bedded sand, silt, and clay and glaciofluvial ice-contact deposits modified by lacustrine processes.

Till

Till is characteristically a grey to grey-brown sandy diamicton, commonly stoney, massive to poorly stratified with sandy lenses and/or stringers, and generally loose. It is non-calcareous with the exception of several samples in the western map sheet.

In bedrock-dominated terrain of highland areas till forms a single discontinuous veneer with thicker accumulations occurring in depressions or as tails on the down-ice side of bedrock knobs. It is commonly capped by a bouldery mantle of unsorted debris, or a veneer of fine grained to sandy glaciolacustrine sediment. In places, the material appears to be extensively reworked and approaches a poorly sorted gravel, lacking the fine grain sizes characteristic of till.

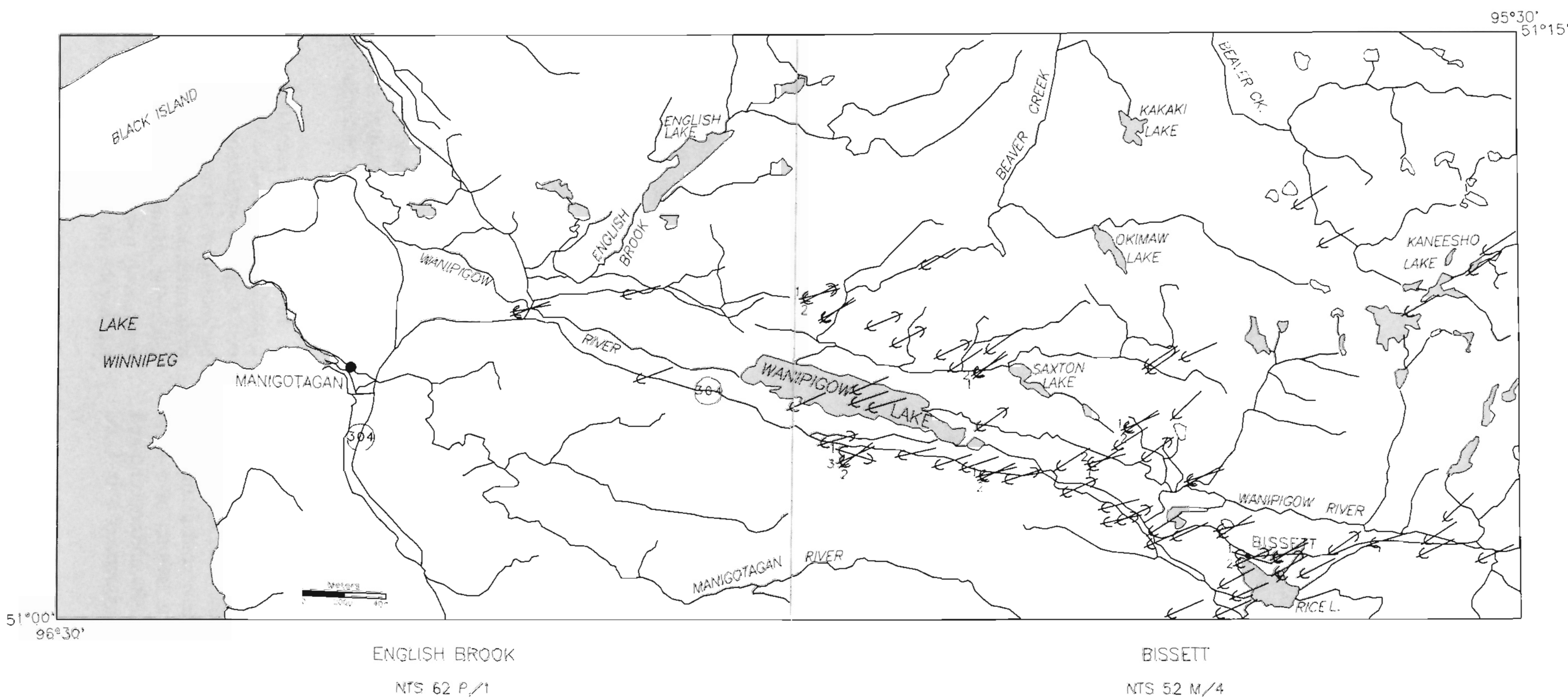

Figure 2. Map of the study area showing striation trends and relative ages (1 = oldest). Double-sided striae signify ice flow direction unknown.

In the major valleys, till is more difficult to find under the thick glaciolacustrine cover. It occurs commonly on the lee side of bedrock-controlled topographic highs, where the overlying glaciolacustrine sediments thin. These lee-side deposits may be complex, comprising interbedded diamicton, sand, and gravel, however there is no evidence to suggest that they represent stratigraphic sequences deposited by multiple flow events. At several sites within the central part of the Wanipigow River valley, diamictons overlie fine grained sorted sediment. The contact between the two units is sharp, dips southwest (down-ice) and may exhibit flame and load structures which suggests that the diamicton is a till flow, presumably deposited subglacially or at the ice margin. The thickest deposits of this type are preserved along the north side of this valley. At one site, a diamicton over 2 m thick overlies an undetermined thickness of fine sand with clasts. The basal 20 cm of the diamicton is sandy, characterized by sand stringers and laminations and thick sand pockets underlying clasts, while the overlying material appears homogeneous and unstratified.

Glaciolacustrine/glaciofluvial sediments

Glaciolacustrine sediment is the dominant surficial material in the area. It forms deposits exceeding 10 m thickness in the major valleys which blanket the bedrock and previously deposited glacially derived material. The texture and sedimentology of these deposits varies.

Discontinuous sand and gravel deposits occur along the length of the Wanipigow River valley. They are characterized by a sequence which generally fines upward from gravel, to crossbedded fine- to medium-grained sand, ripple drift crosslaminated fine sand, and laminated silt and clay. Current directions within the rippled and crossbedded units indicate flow to the west. The presence of isolated clasts interpreted as dropstones, interbedded diamictons, and massive fine sand or sand/silt rhythmites within the glaciofluvial sequence support the interpretation of these deposits as subaqueous outwash deposited at or near the ice margin from subglacial meltwater streams (Rust and Romanelli, 1975). The sequence is commonly overlain by poorly sorted, matrix-supported gravel or well sorted sand likely formed through reworking of previously deposited material by rivers or beaches as Lake Agassiz levels fell (Fig. 4).

Figure 3. Striated top surface of dioritic outcrop showing two directions of ice flow. The dominant direction of 233° is aligned with the compass and presumably postdates the deep groove and associated striae oriented at 242°. GSC 1992-266A

Figure 4. Poorly sorted matrix-supported gravel, likely representing reworked deposits formed as Lake Agassiz levels fell, overlie interbedded ripple-drift crosslaminated fine sand and thinly bedded sand interpreted as subaqueous outwash. GSC-1992 266C

Thick deposits of offshore glaciolacustrine sediments consisting of rhythmically bedded fine sand, silt, and clay overlying till or subglacial outwash sequences are present in depressions within the Wanipigow River valley. These sequences become finer grained up-section grading from predominantly normally graded sand/silt rhythmites to relatively thinner silt/clay rhythmites to massive clay, and may be calcareous.

DISCUSSION

During glaciation bed material (either bedrock or previously deposited sediment) is eroded and deposited down-ice from the source. This glacial dispersal and the ability to recognize bedrock mineralization through drift prospecting is affected by many factors among which are:

1. The glacial history of the area, particularly the direction or directions of ice flow and the number of recorded events.
2. Ice dynamics as it relates to the glaciological response to local and regional variables such as topography and bed composition during glaciation and deglaciation. This response affects ice flow and the manner in which debris is eroded, entrained, transported, and deposited by the glacier.
3. The type, size, and erodibility of bedrock units associated with mineralization.

The stratigraphy and ice flow indicators observed within the Bissett area appear to be related to glacial and deglacial events associated with the last (Late Wisconsinian) glaciation. Striations indicate a consistent direction of glacial transport toward the southwest. Minor variations in trends may be more a response of the glacier to local topographic irregularities, particularly during deglaciation, than an indication of several ice flow events. No definitive multiple till units have been observed in exposed sections within the area. Interbedded diamicton and sand deposits occurring on the lee-side of bedrock knobs are interpreted as subglacial deposits. The significance of calcareous till in areas east (up-ice) of the Paleozoic/Precambrian contact must be examined, however.

The expression of ice dynamics is preserved in the patterns of subglacial and ice marginal sedimentation. In general, only one till is present as a thin veneer on bedrock throughout the upland areas. In this setting, till composition will generally be independent of till facies. In lee-side and thicker till deposits such as those preserved on the northern margin of the Wanipigow River valley, there may be a significant variation in till composition between and within units which must be addressed. Presumably basal facies will reflect more local sources and mineralization.

Ice marginal sedimentation, particularly subaqueous outwash, and glaciolacustrine deposits inhibit drift prospecting since they mask basal till units. In the Wanipigow River valley, these deposits cover mineralized rocks of the Rice Lake greenstone belt. Till sampling, therefore, has been concentrated in upland areas south of the valley in an attempt to recognize bedrock mineralization associated with the greenstone belt. Anomalous concentrations of gold or base metals within the glaciolacustrine deposits may be more a reflection of secondary processes associated with reworking of glacial sediments than primary bedrock mineralization.

Mineralization within the study area occurs within shear zones associated with mafic or ultramafic intrusions that tend to be small, with limited surface exposure, and erratically distributed. Because the source area for glacial erosion is restricted, it is unlikely that dispersal trains associated with specific mineral deposits will be sampled or recognized within the sample density of the project. Geochemical trends identified will most likely reflect bedrock units with potential for hosting gold and base metal deposits and areas where detailed follow-up work is indicated.

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