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PLAINES INTÉRIEURES ET RÉGION ARCTIQUE DU CANADA

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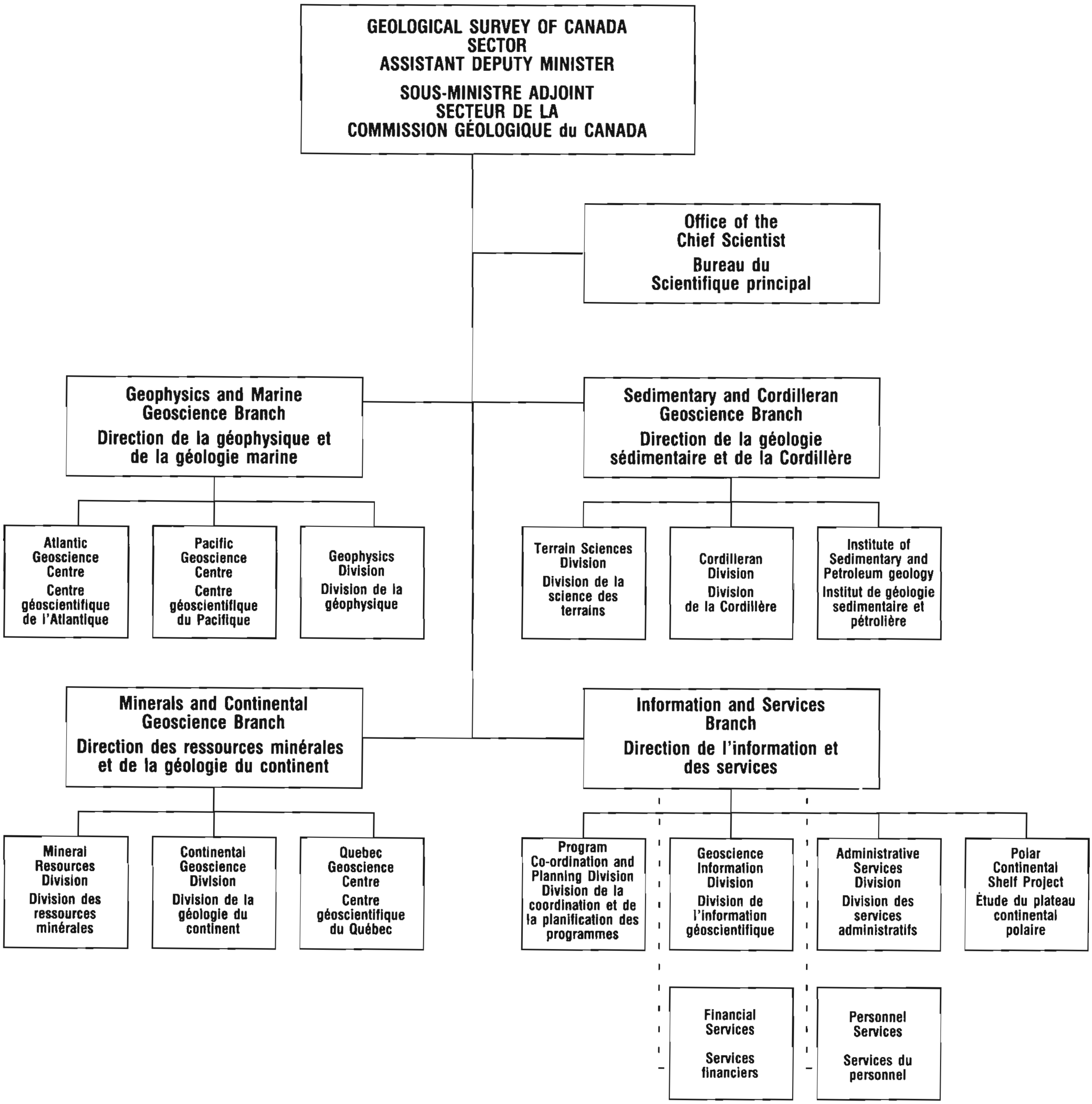
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CONTENTS

Quaternary investigations in the Expedition Fiord area, west-central Axel Heiberg Island, Northwest Territories

Donald S. Lemmen, Robert Gilbert[1] and Alec E. Aitken[1]
Terrain Sciences Division, Calgary

Lemmen, D.S., Gilbert, R., and Aitken, A.E., Quaternary investigations in the Expedition Fiord area, west-central Axel Heiberg Island, Northwest Territories; in Current Research, Part B, Geological Survey of Canada, Paper 91-1B, p. 1-7, 1991.

Abstract

Terrestrial and marine Quaternary deposits suggest a significant expansion of local glaciers in the Expedition Fiord area during the (Late ?) Wisconsinan. Ice marginal landforms record a glacier advance at least 10 km beyond the fiord head. The limited sediment accumulation in the inner fiord (<20 m) places some constraints on the timing of this advance, with the most probable scenario suggesting that deglaciation occurred during the Late Wisconsinan/early Holocene. All fossil marine fauna discovered in raised marine deposits are extant within the fiord, and unique contemporary faunal assemblages may provide important analogues for the paleoecological interpretation of the ancient deposits.

Résumé

Les dépôts terrestres et marins du Quaternaire semblent indiquer que durant le Wisconsinien (tardif?), a eu lieu une expansion importante des glaciers régionaux dans la zone du fjord Expedition. En marge des glaces, la topographie témoigne d'une avancée des glaciers sur au moins 10 km au-delà du fond du fjord. L'accumulation sédimentaire limitée qui a eu lieu dans le fjord interne (<20 m) permet de délimiter dans une certaine mesure la chronologie de cette avancée; selon le scénario le plus vraisemblable, cette déglaciation aurait eu lieu au cours du Wisconsinien tardif ou du début de l'Holocène. Toute la faune marine fossile découverte dans ces sédiments marins soulevés existe à l'intérieur du fjord, et des associations fauniques exceptionnelles du même âge pourraient représenter d'importants éléments analogues à l'appui de l'interprétation paléoécologique des anciens sédiments.

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INTRODUCTION

The nature of late Quaternary glacier fluctuations in the Queen Elizabeth Islands has been a controversial topic for the last two decades (for recent review see Hodgson, 1989). This ongoing debate is fuelled by the palimpsest nature of the terrestrial glacial record, the general scarcity of glacial landforms over much of the landmass, and the near complete absence of a stratigraphic record spanning more than the last 10 ka. The marine record of the adjacent fiords and channels may well be the key to resolving this controversy (Lemmen, 1990), yet it remains almost completely uninvestigated. Surveys within the large channels of the southern Queen Elizabeth Islands (MacLean et al., 1989) yielded important data, however, correlation with the terrestrial record from the adjacent islands is difficult. While Lemmen (1990) discussed the first sediment cores collected from a fiord in the high Arctic, the absence of detailed bathymetric and seismic data limits the interpretation with respect to glacial history. Therefore, the present study represents the first interdisciplinary investigation of terrestrial and marine Quaternary geology conducted in a high Arctic fiord environment.

Study area

Expedition Fiord on west-central Axel Heiberg Island (Fig. 1) was selected as the focus for this study because it is a high arctic site with considerable data on contemporary glacial and climatic conditions as a result of long-term and ongoing studies in glaciology centred at the Jacobsen-McGill Research Station (Adams, 1987). The fiord is 31.5 km long to the confluence with Strand Bay and has an average width of 5.3 km. The upper reaches of the 1575 km^2 drainage basin are heavily glacierized, with the majority of inflow entering at the fiord head via Expedition River which is fed by two outlet glaciers. Gilbert (1990), determined that the rate of delta progradation (since 1959) at the mouth of Expedition River was the highest reported in the Canadian Arctic, suggesting a relatively high energy marine depositional environment. Building upon the bathymetric and oceanographic data collected in this preliminary investigation of contemporary sedimentary environments, objectives of the 1990 field season were to conduct a subbottom acoustic survey of the fiord, and to collect benthic samples for sedimentological and biological analysis (Fig. 2). The final phase of the marine investigations involving collection of long sediment cores is planned for spring 1991.

No surficial geology maps exist for western Axel Heiberg Island (Hodgson, 1989) and little is known about the local or regional glacial history. Boesch (1963) described two zones of glacial features near the head of Expedition Fiord which he attributed to different ages of glacial advance. The lower zone, which extends to about 280 m asl, was interpreted to be of Wisconsinan age while the upper zone was tentatively interpreted as being pre-Wisconsinan. Marine limit upvalley of the fiord head was reported to be about 80 m

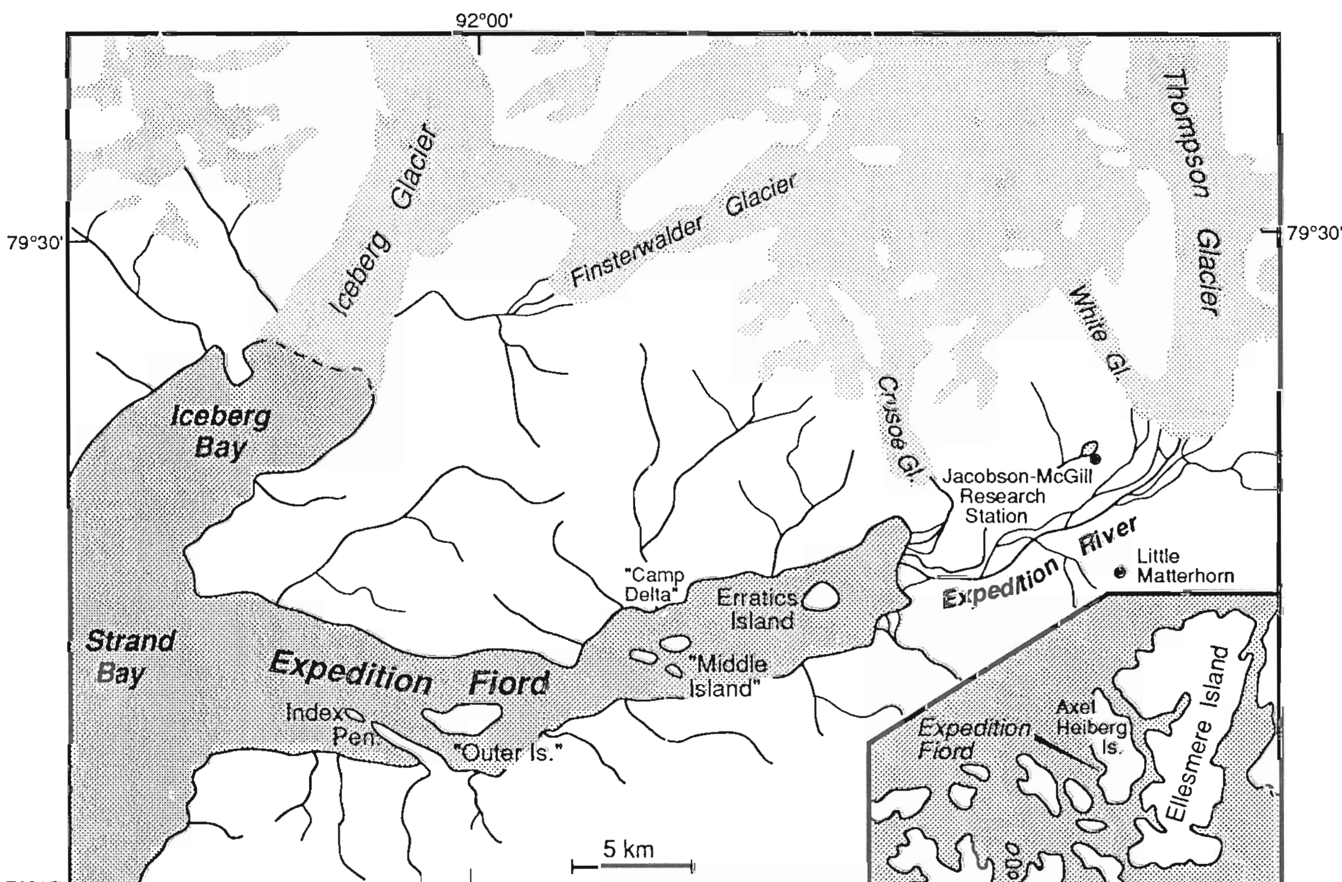

Figure 1. Expedition Fiord area. Glaciers are shaded light, marine waters shaded dark.

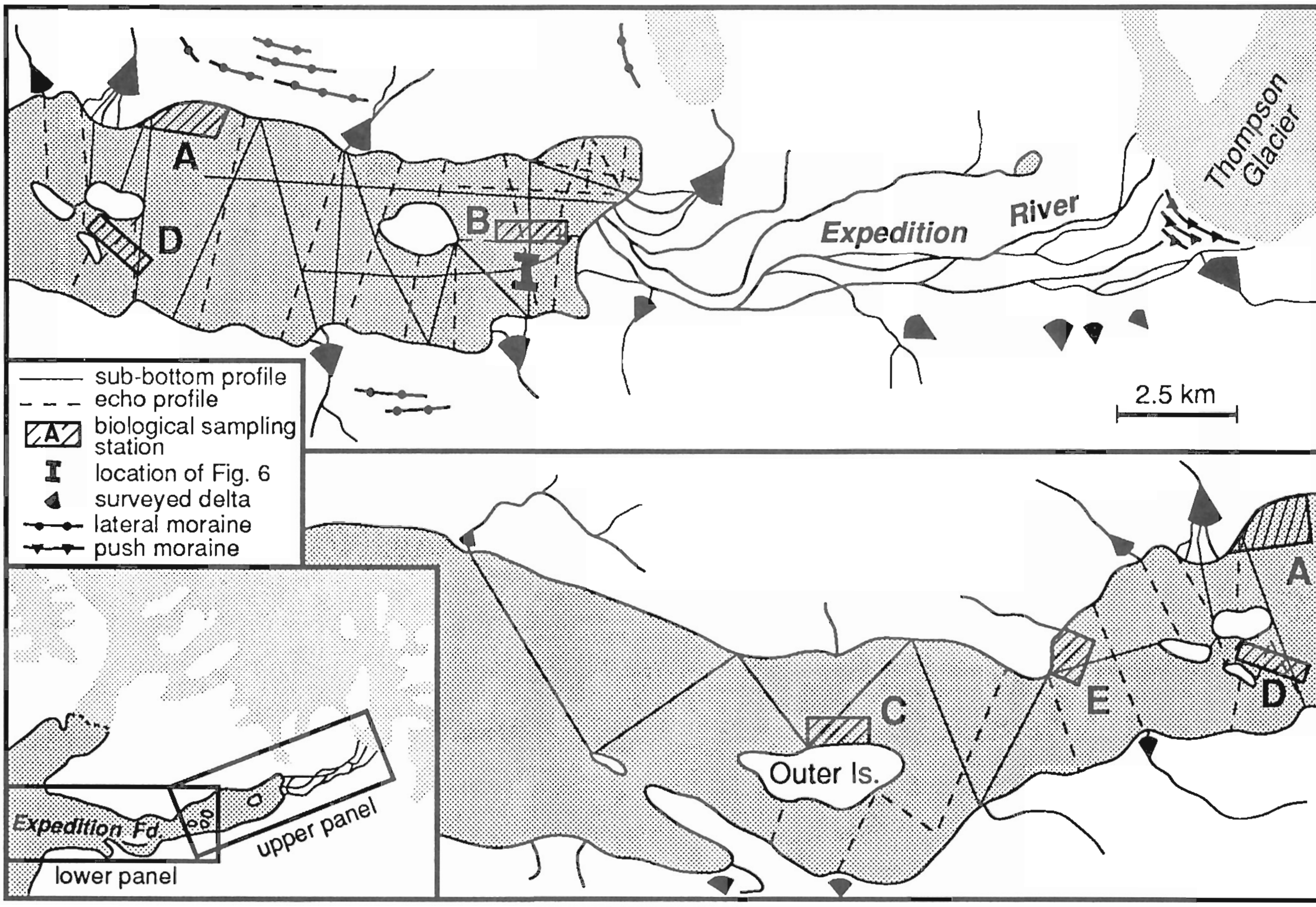

Figure 2. Areas of field sampling during the 1990 season. Glaciers are shaded light, marine waters shaded dark.

a.s.l. (Boesch, 1963), and shells collected from that elevation within 5 km of the terminus of Thompson Glacier (Fig. 1) dated ca. 9 ka (Müller, 1963). Field work on terrestrial surficial geology during 1990 focused on the coastal fringe adjacent to Expedition Fiord. Emphasis was placed upon surficial mapping and deciphering relative sea level history. Future work will expand the area of investigation to produce a 1:100 000 surficial geology map for the western half of the Strand Fiord map sheet (NTS59H).

SURFICIAL GEOLOGY - TERRESTRIAL

The dominant surficial materials within the map area are bedrock and weathered rock, as is common throughout the Queen Elizabeth Islands (Hodgson, 1989). Physical weathering of the poorly lithified Mesozoic and Cenozoic bedrock (Thorsteinsson, 1971) produces large volumes of material that on slopes are reworked by colluvial processes. Significant glacigenic sediments are largely restricted to the lower reaches of major valley systems and some higher plateau surfaces. Till is generally <1 m thick, although greater thicknesses do occur, particularly around the margins of modern outlet glaciers. Elsewhere, scattered erratics are found to at least 650 m a.s.l. in the fiord head region (Boesch, 1963) and suggest that most of the region was covered by (local?) glacier ice at some time. No erratics that unequivocally relate to source areas beyond Axel Heiberg Island were found. The sandur of Expedition River comprises the most extensive fluvial deposit in the area; most other rivers are incised into bedrock in their lower reaches. The mouths of most major valleys are occupied by glaciomarine deltas with initial grade to local marine limit. Fine grained marine sediments thinly mantle most slopes below marine limit. However, extensive deposits of thick glaciomarine silt that are common in many fiords in the high Arctic (Bednarski, 1988) are rare in Expedition Fiord.

Ice-cored push moraines are found at the margins of outlet glaciers that presently terminate on sandur plains (Fig. 3). Beyond these spectacular moraines, however, glacial landforms are relatively rare. The few resistant lithologies present preserve striae and some show minor ice-moulding. Along central Expedition Fiord, ice marginal features include subtle lateral moraines and meltwater channels that crosscut the bedrock structure (Fig. 2). On the north side of the fiord the lowest, and most prominent of these features can be traced more than 4 km between major valleys (Fig. 4). Airphoto interpretation suggested that this feature was a product of wave erosion; however, a relatively steep gradient of 8 to 9 m/km toward the west (descending from 140 m a.s.l.) precludes the possibility that it records a former shoreline, and hence it is interpreted as an ice marginal feature. The gradient decreases and reverses slightly at its western limit. This reversal in gradient likely relates to the coalescence of a tongue of ice in the fiord and a glacier within the tributary valley. These ice marginal features record a tongue of ice in Expedition Fiord, extending at least 10 km down from the fiord head and 20 km beyond the present terminus of Thompson Glacier.

Marine deltas were surveyed in 16 tributary valleys to Expedition Fiord and Expedition River valley (Fig. 2). Elevation of the upper delta terraces, recording local marine limit, ranges from 102 m a.s.l. west of the Little Matterhorn to 78 m a.s.l. south of Index Peninsula (Fig. 1). Lower marine limits in some valleys reflect the late retreat of local valley glaciers, or simply indicate that small basins did not generate sufficient sediment to grade to marine limit. Beaches and washing limits are not well developed along slopes between major valleys, although breaks in slope which commonly extend above the highest delta surfaces may be of marine origin. Nowhere along the fiord is it possible to trace a continuous former shoreline for any significant distance, while an irregular pattern of marine limit elevations suggests diachronous formation.

Figure 3. Contemporary push moraine at terminus of Finsterwalder Glacier. Moraine is composed primarily of ice-thrust alluvium and is at least partially ice cored.

SURFICIAL GEOLOGY - MARINE

A subbottom acoustic survey was conducted from a small boat using a 3.5 kHz Datasoncis SBP-5000/EPC-1650 system. A total of 73 km of line were run with reference to features recognizable on air photographs. These records, combined with additional conventional echo sounding, were used to prepare a bathymetric map of the fiord (Fig. 5) which shows that the eastern portion is a collection of shallow basins (maximum depth, 132 m) separated from the western portion by a sill at 70 m depth north of a group of three islands in mid-fiord. West of these islands the depth increases to a maximum measured value of over 300 m. As the bathymetry of Strand Bay is unknown, it cannot be shown whether a sill separates the waters of Expedition Fiord from Peary Channel.

The floor of the fiord is mantled with a veneer of soft, well stratified sediment. At the fiord head where Expedition River contributes large loads of suspended and bed material (Gilbert 1990), these materials reach maximum thickness in excess of 30 m toward the north (right) side of the fiord, reflecting the Coriolis effect on overflowing, sediment-laden fresh water. Toward the sides of the floor of the fiord and at the base of the sediment column, the stratified sediment lies conformably on the acoustically opaque substrate, while in mid fiord the effect of ponding of sediment, at least partly derived from gravity flows can be seen (Fig. 6).

West of Erratics Island, still in the eastern basins, the mantle of glaciomarine sediment has a maximum thickness of about 20 m. Above about 70 m depth (the depth of the sill that controls the size of icebergs admitted from Iceberg Bay to eastern Expedition Fiord) iceberg scouring has disturbed, and in places removed, much of this material. West of the sill isolated pockets contain up to 40 m of glaciomarine sediment, but elsewhere berg scouring to depths greater than 300 m has altered much of this material.

Analysis of short cores obtained in 1990 and long cores to be recovered in 1991, in combination with oceanographic data obtained in 1988 (Gilbert 1990) and 1990, will be used with the acoustic evidence to corroborate interpretations and link to the terrestrial component of this study.

Figure 4. Lateral moraines and meltwater channels along north coast of central Expedition Fiord. Elevation of lowest moraine at solid arrow is about 140 m a.s.l., local marine limit in adjacent valley to east (open arrow) is 93 m a.s.l.

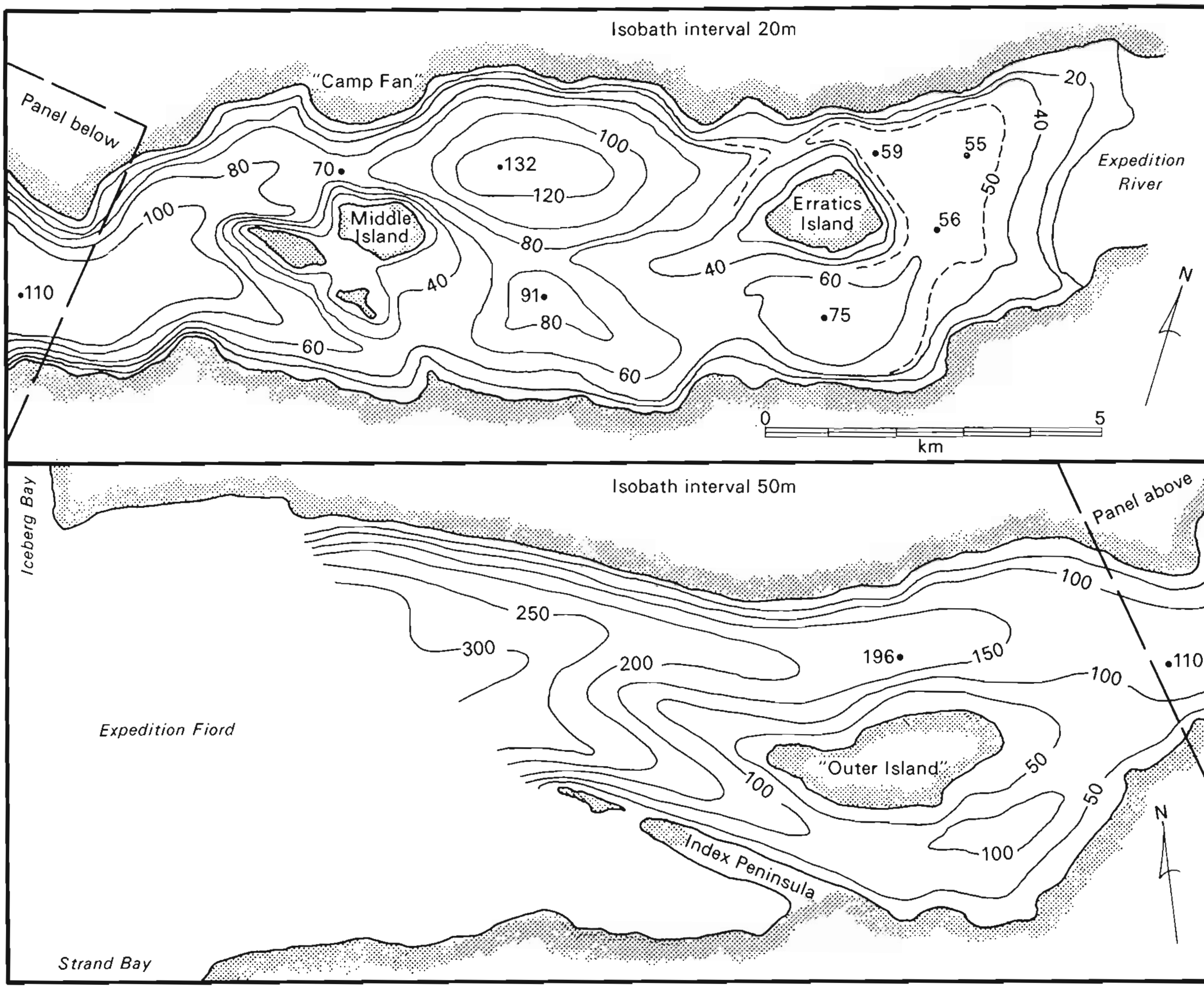

Figure 5. Bathymetry of Expedition Fiord interpreted from echo sounding and subbottom profiling. Isobaths are interpolated between runs (see Fig. 2 for locations).

MARINE ECOLOGY AND PALEOECOLOGY

The present study represents the most complete record of the macrobenthos inhabiting a Canadian arctic fiord north of 70° latitude. Thirty-eight dredging tows (using a Kolquitz dredge) were attempted at depths of 5-80 m at 5 sites within the fiord (Fig. 2). Material recovered includes representatives of the following invertebrate taxa: bivalves, gastropods, polychaetes, echinoderms (brittlestars, sea urchins, holothurians), and a variety of crustaceans including amphipods, isopods, cumaceans, tanaidaceans, pycnogonids, and shrimp.

Among the invertebrate taxa, the molluscs and echinoderms possess the greatest preservation potential in Quaternary marine sediments, hence they are of considerable interest in understanding the paleoecology of Quaternary marine depositional environments. Presence/absence data recorded from the dredge samples suggest 3 macrofauna associations within Expedition Fiord: 1) an association consisting of the bivalves *Portlandia arctica, Thyasira* sp., and small brittlestars occurring at 50-80 m depth on the front of the Expedition River delta (Site B); 2) an association consisting of the bivalves *Astarte borealis, Macoma calcarea, Mya truncata, Hiatella arctica, Clinocardium ciliatum,* the gastropod *Cylichna* sp., the sea urchin *Strongylocentrotus* sp., and a variety of brittlestars occurring at 5-20 m depth at Sites A, C, D, and E; and 3) an association consisting of the bivalves *Portlandia arctica, Thyasira* sp., *Delectopecten* sp., the gastropod *Cylichna* sp., and a variety of brittlestars occurring at 40-60 m depth at Sites A, C, D, and E. The presence of *Portlandia arctica* at depths of 5-20 m at Site D in association with an Arctic *Macoma* community is unusual. Syvitski et al. (1989) described a similar association inhabiting depths of 5-20 m on the fiord-head delta of McBeth Fiord, Baffin Island and they suggested that it is typical of shallow water environments characterized by low sedimentation rates in Baffin Island fiords. Observations in Expedition Fiord support this view.

The molluscs, *Astarte borealis, Hiatella arctica, Mya truncata,* and *Cylichna* sp., are the only marine invertebrates to be recovered from raised marine sediments bordering

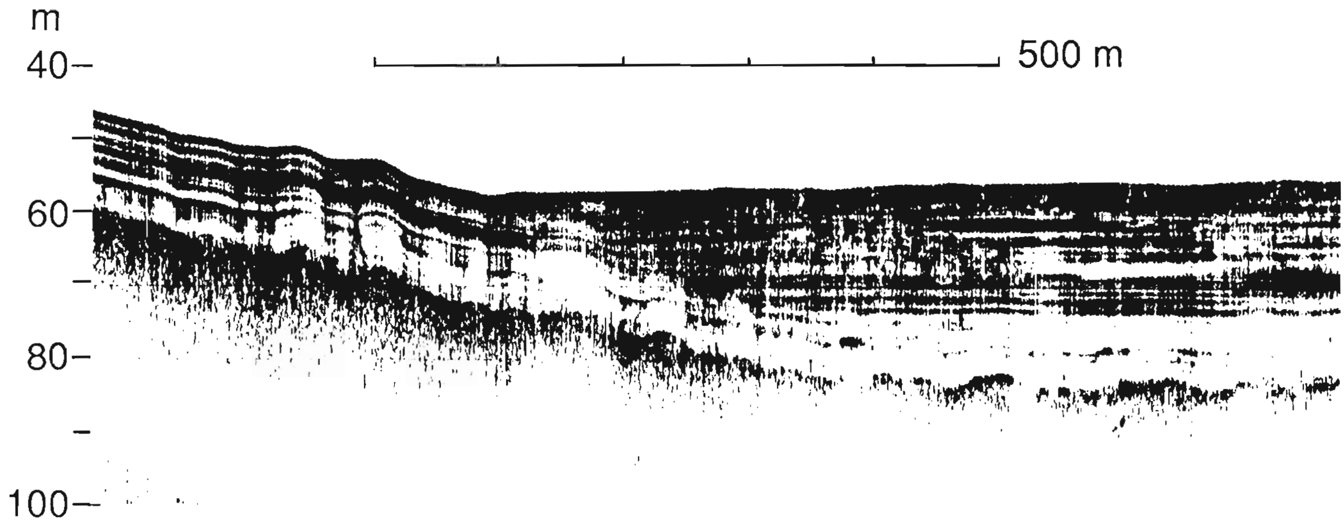

Figure 6. Portion of acoustic subbottom record showing the accumulation of glaciomarine sediment near the fiord head (see Fig. 2 for location).

Expedition Fiord. *Astarte, Hiatella,* and *Mya* are always found in association with one another, suggesting that a community similar to the shallow water association (5-20 m) observed at Sites C, D, and E in the fiord, inhabited these marine sediments. An ice-distal environment, characterized by low sedimentation rates is suggested by the fossil mollusc association.

DISCUSSION

The present study is unique, for it is the first time that a reconstruction of glacial history in the high Arctic based upon traditional geomorphological data is augmented by the adjacent marine sedimentary record. Evidence from both records suggests a significant expansion of local glaciers in the Expedition Fiord area during the Late Wisconsinan. The gradient of the lateral moraines along the central fiord indicates that they record a glacier grounded in the fiord. The associated relative sea level at this time could not have been greater than about 100 m a.s.l., otherwise the glacier would have floated. Measurements of marine limit in this area are all <102 m a.s.l. Therefore, the sea level record is compatible with a Late Wisconsinan/early Holocene age for the moraines. The limited sediment accumulation in the inner fiord (less than about 20 m) also suggests the presence of grounded ice in the fiord during the Late Wisconsinan. Using a minimum date on deglaciation of 9 ka (Müller, 1963), average sedimentation rates for the inner fiord have been about 1-2 m/ka. Sedimentation rates were likely highest at the time of active glacier retreat (cf. Lemmen, 1990), therefore actual accumulation rates were likely somewhat lower for much of the Holocene. Although high in comparison to some Arctic marine environments (see Lemmen, 1990), these sedimentation rates are consistent with data on the contemporary sedimentary environment and the rapid rate of progradation of the Expedition River into the fiord. The limited sediment accumulation in the fiord suggests that ice retreat occurred rapidly, for raised marine sediments elsewhere in the high Arctic commonly feature many tens of metres of ice-proximal glaciomarine sediment (Bednarski, 1988).

Evidence was not found to suggest that the moraines along central Expedition Fiord record the limit of the last glaciation in this area. While Boesch (1963) suggested the presence of two weathering zones in the fiord head area, observations supporting this proposal were not made during the past field season. Boesch's (1963) observation that glacial features are more common in the lower valleys may well be attributed to glacier dynamics as opposed to reflecting different durations of exposure to subaerial weathering (e.g., Fyles, in Jenness, 1962). Hence the moraines along central Expedition Fiord provide only a minimum estimate of the limit of the last glaciation. Although no major moraines are visible in the acoustic records from the fiord, the slightly thicker sediment accumulations (more than 40 m) in the western portion of the fiord suggest a longer period of deposition than in the inner fiord, although some of this material may originate from Iceberg Bay. Nevertheless, accumulation of glaciomarine sediment everywhere in the fiord is minimal compared to the fiords of Baffin Island, where accumulations of up to several hundred metres are present (Gilbert, 1985). This suggests that sedimentation may have occurred during a relatively short period after glaciers reached beyond the mouth of Expedition Fiord.

The macrofauna of high Arctic fiords remain poorly documented (cf. Curtis, 1972; Dale, 1985). Fossil assemblages from the raised marine sediments investigated thus far along Expedition Fiord all suggest ice-distal environments, and the scarcity of fossiliferous ice-proximal sediments in the raised marine record is consistent with the model of an extensive former ice cover and rapid glacier retreat. The study of the autecology of marine invertebrates inhabiting Expedition Fiord offers considerable promise for expanding our paleoenvironmental interpretations of raised marine deposits.

ACKNOWLEDGMENTS

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Occurrence of early snow-free zones on Fosheim Peninsula, Ellesmere Island, Northwest Territories

Ming-ko Woo[1], Sylvia A. Edlund, and Kathy L. Young[1]
Terrain Sciences Division

Woo, M-K., Edlund, S.A., and Young, K.L., Occurrence of early snow-free zones on Fosheim Peninsula, Ellesmere Island, Northwest Territories; in Current Research, Part B, Geological Survey of Canada, Paper 91-1B, p. 9-14, 1991.

Abstract

In May 1990 on Fosheim Peninsula snow disappeared from some areas earlier than from the rest of the lowland areas. Comparisons of microclimate and soil hydrology data from three sites, one within the early snow-free zone, and two just outside the snow-free zone, show that the frost table deepened early in the snow-free zone and remained deeper than those areas that emerged later. The spatial variability in microclimate and soils on different slope aspects at a site, however, often created differences of a magnitude as great as the inter-site comparisons; this makes extrapolation of single point data difficult. Vegetation studies suggest that the increase in length of growing season within the early melt zone is not sufficient to enrich the diversity of the flora. The mosaic distribution patterns of plant communities at all three sites, however, reflect the variability of microclimate and soil hydrology.

Résumé

En mai 1990, dans la péninsule de Fosheim, la neige a disparu de quelques régions plus rapidement que dans le reste des zones des basses terres. En comparant les données relatives au microclimat et à l'hydrologie du sol à trois endroits, l'un situé dans la zone précocement libre de neige, les deux autres immédiatement à l'extérieur de la zone libre de neige, on a noté que la profondeur du mollisol augmentait précocement dans la zone libre de neige, et restait plus importante à cet endroit que dans les régions émergeant plus tardivement des neiges. La variabilité spatiale du microclimat et des sols selon diverses orientations et versants à un endroit donné a toutefois souvent donné lieu à des différences aussi importantes que celles découlant de comparaisons établies entre les localités étudiées; cette situation rend difficile une extrapolation des données ponctuelles isolées. Les études de la flore semblent indiquer que l'augmentation de durée de la saison de croissance dans la zone de fonte précoce ne suffit pas pour augmenter la diversité de la flore. Les schémas de répartition en mosaïque des communautés végétales aux trois endroits reflètent cependant la variabilité du climat et de l'hydrologie du sol.

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INTRODUCTION

Since 1988 Terrain Sciences Division has been conducting multidisciplinary research on the interrelationships of climate, vegetation, soil hydrology, and geomorphic processes on Fosheim Peninsula, an intermontane region of the northeastern Queen Elizabeth Islands (Edlund and Alt, 1989; Edlund et al., 1989). Initial studies of climate, hydrology, and vegetation focused on the base camp area along lower Hot Weather Creek (Woo et al., 1990).

When the base camp was selected in June 1988, the site was snow-free while the area around the Eureka weather station and areas above 120 m a.s.l. had a thin and discontinuous snow cover. In 1990, this general pattern recurred, and observations were made to enable comparisons of the microclimate, slope hydrology, and vegetation in areas within and outside the early melt zone. This report explores some of the implications of early melt in this region, and its effects on hydrology and vegetation.

STUDY AREA AND INSTRUMENTATION

In 1990 the observation network established the previous year (Woo et al., 1990) was expanded from the slopes and plateau at lower Hot Weather Creek (60 m a.s.l.) to include two other locations. One site is 0.5 km north of the air strip at Eureka at approximately 15 m a.s.l., 22 km west of the Hot Weather Creek base camp. The other is located in the upper part of the Hot Weather Creek basin (approximately 137 m a.s.l.), 10 km north of the base camp (Fig. 1).

Snow and vegetation surveys and the progressive positions of the frost table were studied at all three sites on north, south, east, and west slopes. Air temperature, solar radiation, and rainfall were recorded on the study slopes in the vicinity of the base camp and at a nearby plateau site, as described in Woo et al.(1990). At the upper basin site, air temperature, solar radiation, and rainfall were monitored on a knoll above the monitored slopes. The weather station at Eureka provided comparable meteorological data for the Eureka site. Snow ablation and groundwater levels were also measured at the base camp.

A regional mosaic of snow melt on Fosheim Peninsula was provided by several NOAA satellite images from 17 May 1990. Several other images in May and June 1990 were examined as well. Snow-free areas appear as dark patches on an otherwise white landscape.

RESULTS

Spring air temperatures

At the base camp, daily maximum temperatures greater than 0°C occurred on May 20-22, with a maximum of 2°C (Fig. 2). Daily maximum temperatures stayed above 0°C after May 28 and the minimum daily temperature stayed above 0°C after June 6. At Eureka the mid May maximum daily temperatures continued to be below 0°C. Maximum daily temperatures stayed higher than 0°C after June 3, and the minimum daily temperatures from June 8. The values of daily maxima at Eureka and Hot Weather Creek automatic weather station were similar; however, the minimum daily temperatures at Eureka were consistently lower, thus reducing the mean daily temperatures at Eureka to values lower than those of the automatic weather station. Daily maxima and minima were lowest at the upper basin site. Maximum daily temperatures rose above 0°C only after June 8 and the minimum daily temperature remained above 0°C after June 15. Daily air temperatures were consistently lower at this site throughout spring.

Snow surveys

Initial surveys of snow accumulation were carried out between 10 and 13 May, 1990. At that time some snow on west and south slopes had already sublimated. Table 1 shows that all north slopes accumulated large quantities of snow, followed by east slopes. The exception was at upper basin where the south slope collected the maximum amount of snow; this may reflect a local departure from the general accumulation pattern, exaggerated by early sublimation on south slopes at the other two locations. While all the north slopes surveyed had large snow drifts, the drifts at the base camp study slopes were particularly pronounced, causing accumulation to be twice as large as that of the flat plateau site.

Snow ablation

Daily measurement of snow ablation was repeated at the four study slopes and plateau site near Hot Weather Creek base camp after 14 May (Fig. 3). The west slope, with a dust cover on the snow, was the first site to experience significant melting after mid May. Melt was arrested by several snowfall events between 19-27 May, but with early melt and a thin snow cover, the west slope was bare by early June. The south slope was covered by cleaner snow than the west slope and melt did not begin until late May. By 10 June most of the snow had disappeared. Melt initiation was further delayed on the east and north slopes and the plateau. Most of the thin snow cover on the plateau and east slope was gone in less than a week after melt began. Snow stayed past mid June at the north slope because of the presence of a large drift, but daily ablation rates were high because of the large radiation energy it received and because of the advection of heat from the adjacent bare terrain.

Although detailed measurements were not made at the other locations, cursory observations indicated that the timing

Table 1. Snow Surveys, 10-13 May 1990

	WATER EQUIVALENTS (mm) AT THREE STUDY SITES		
	Eureka	Base Camp	Upper Basin
Plateau	—	118	—
North Slope	144	247	158
East Slope	84	72	104
South Slope	47	63	186
West Slope	62	62	76

Number of depth samples range from 14-81 per slope.
Number of density samples range from 4-14 per slope.

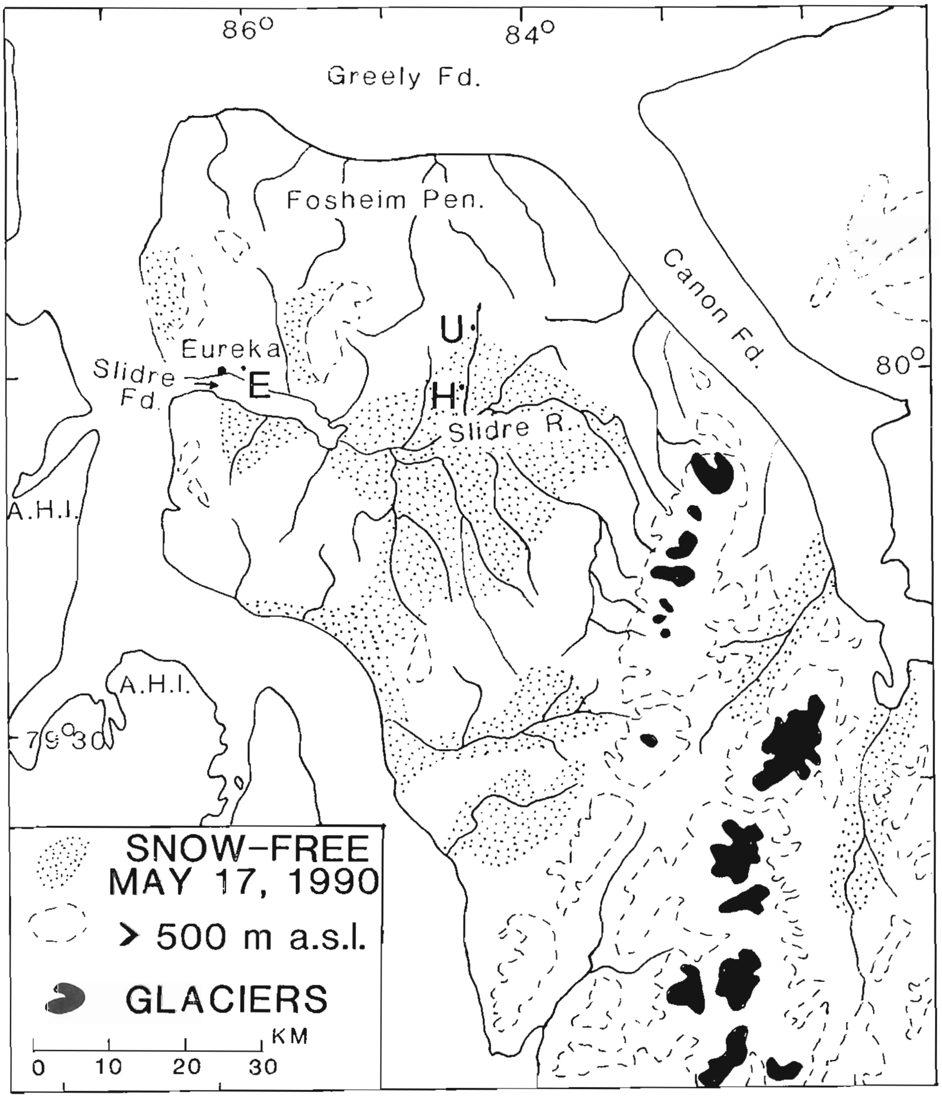

Figure 1. Location of field sites on Fosheim Peninsula, Ellesmere Island. U = Hot Weather Creek upper basin sites: H = Hot Weather Creek study slopes near base camp; and E = Eureka sites. The shaded areas indicate dark spots that appeared on NOAA satellite imagery of 17 May 1990, representing areas where most of the snow has melted.

of ablation followed a similar sequence, with earliest melting at the west slope, and snow remaining for the longest period on the slope with the most snow.

On a regional basis, intense melting took place first in the vicinity of the base camp. A traverse 16-17 May along the length of Hot Weather Creek, the lower reaches of Slidre River, and the north shore of Slidre Fiord to Eureka, showed considerable melt in the vicinity of lower Hot Weather Creek, while snow persisted in the upper part of Hot Weather Creek basin. This melted zone, and other, smaller early melt areas on the Fosheim Peninsula, showed up as large dark patches on the NOAA satellite imagery of 17 May (Fig. 1). The largest patch measured roughtly 40 × 30 km and included the Hot Weather Creek base camp and study sites; Eureka and upper basin sites were located outside the snow-free area.

Frost tables

Thawing of the active layer began as soon as snow disappeared from each site: first on the west slope at the base camp, followed by thawing on the south and east slopes. The north was the last to begin thawing because of its deep snow cover. Similar trends with a delay in timing were observed at upper basin where the first set of frost table depths were measured on 7 June (Fig. 4).

The frost table on the north slope at base camp deepened quickly to depths comparable to that of the west slope which had a much longer duration of thaw. This trend was also seen in 1989 (Woo et al, 1990) and can be attributed to the sandy soil of the north slope with a higher thermal conductivity than the other soils. Rapid deepening of the frost table on the south slope of upper basin could have been caused by its higher thermal conductivity, but no soil sample has yet been collected to confirm this.

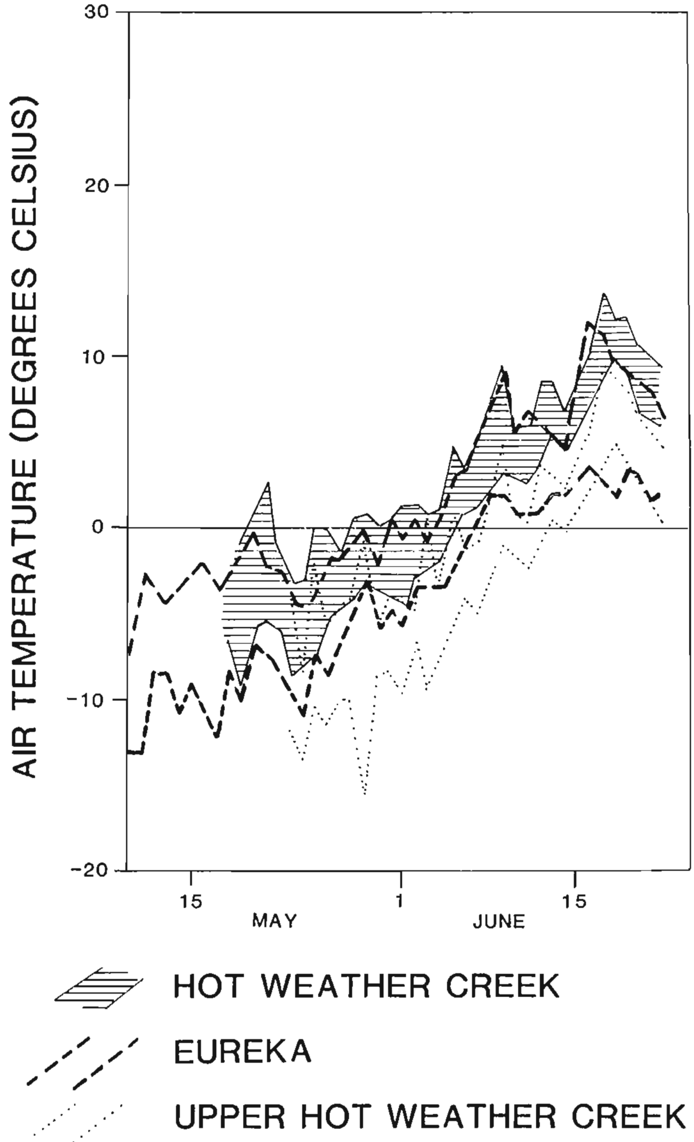

Figure 2. Comparisons of the maximum and minimum temperatures at Eureka weather station, Hot Weather Creek automatic weather station on the plateau, and the upper basin plateau site in spring 1990.

A comparison of the frost table depths at all three sites revealed that during summer the thawed zone was usually deeper at the study slopes near Hot Weather Creek base camp. Other than the anomaly of the upper basin south slope, both Eureka and upper basin had comparable frost table depths at any particular time during the summer, which is related to the similarity of their air temperatures.

Vegetation

The vegetation at the two additional study sites contain largely the same flora as was reported in Woo et al. (1990). The slopes are covered by a patchy mosaic pattern of several different communities. Woody species, particularly *Dryas integrifolia* and *Salix arctica* dominate most slopes and the hummocky plateau on both sandy and silty soils. The presence of a cryptogamic (moss and lichen) ground cover depends on presence of snow cover in winter, some surface moisture in summer, and minimal deposition or erosion at the soil surface. *Cassiope tetragona* heath tundra, commonly associated with sheltered sites having a persisting snow cover and ample moisture throughout the summer, occurs on the lower aspects of the south and north slopes of upper basin, and to a lesser extent on the lower north study slope. No heath community occurs at the Eureka study slopes, although it does occur close by.

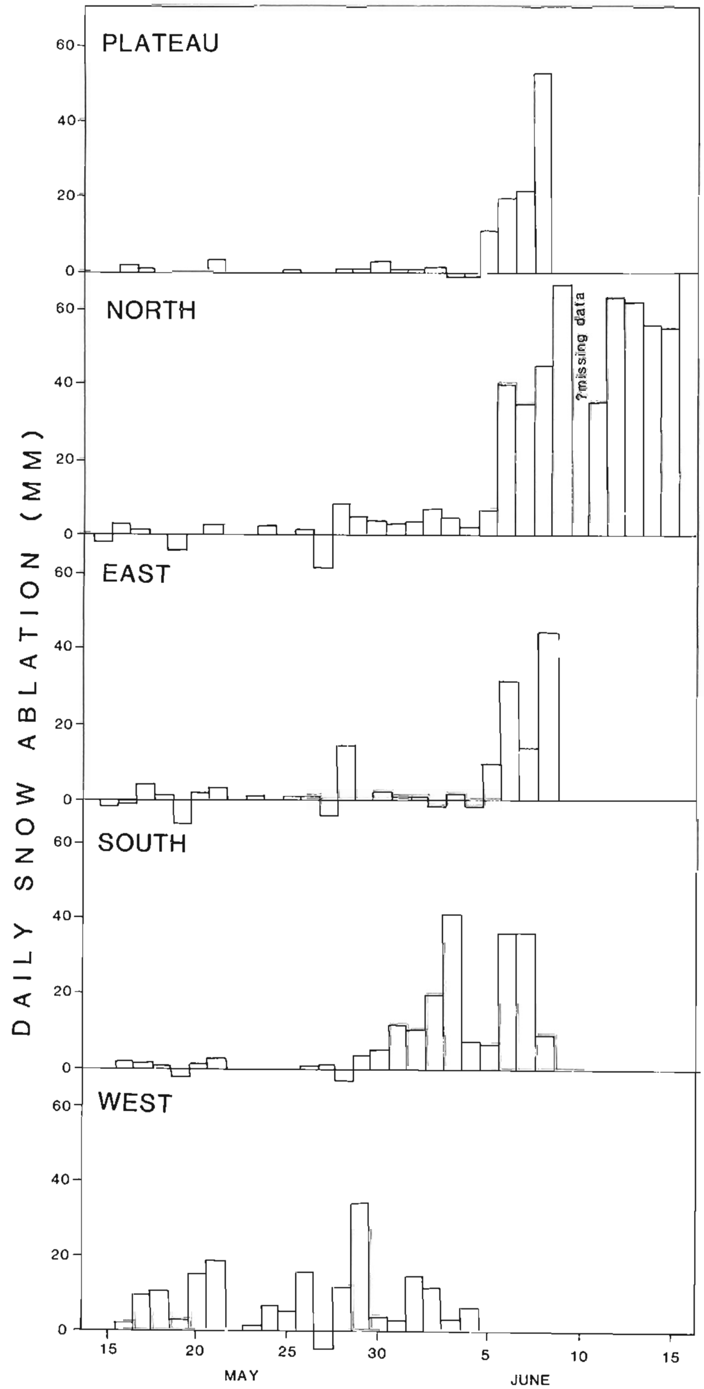

Figure 3. Daily snow ablation measured at five sites: the plateau, north, south, east and west study slopes near the Hot Weather Creek base camp.

The biggest difference between the slopes of the two new sites and those at the Hot Weather Creek is that mid to upper slopes at Eureka and the upper slopes at upper basin have patches of a sparse grass-dominated, saline-tolerant herbaceous community. This community consists of varying amounts of *Puccinellia angustata, P. langeana, P. poacea, Poa hartzii, P. glauca, Agropyron violaceum, Potentilla nivea, P. pulchella, Melandrium affine, Braya purpurescens, and B. thorild-wulffii*, and *Oxyria digyna*, with less than 5-10 total per cent cover. It usually occurs in warm, arid regions where soils are persistently dry throughout most of the thaw period. A thin salt crust on the soil surface is also common wherever this community occurs.

Although some of these species are present in association with woody plant communities on the Hot Weather Creek study slopes, the complete community and salt crusts do not occur. This community, however, does occur extensively in the vicinity of the Hot Weather Creek base camp, particularly on remnants of marine sediments and the slope wash from such sediments.

DISCUSSION

Certain areas on the Fosheim Peninsula became snow-free at an early date. The 17 May, 1990 NOAA satellite imagery revealed dark patches which delineate snow-free areas. This was confirmed by ground observations during snow surveys undertaken at the same time. The base camp site was the only study site located in the early snow-free zone, whereas the other two sites were still covered by snow at this time.

The largest snow-free zone on 17 May occurred in the centre of the lowlands on Fosheim Peninsula, which has a dissected, rolling topography. Other smaller, snow-free zones are associated with steep-sided slopes such as the west slope off of Black Top Ridge, 5 km east of Eureka, and another steep scarp facing Eureka Sound, 15 km west of Eureka. The bottoms and lower slopes of steep-walled valleys between the mountain ranges of eastern Fosheim Peninsula also have similar early snow-free zones. Although the extent and timing of these snow-free areas no doubt vary each year, such early snow-free areas developed in 1988 and 1990 in similar locations.

The development of the largest snow-free zone on central Fosheim Peninsula cannot be fully explained at this time. Preliminary observations at Hot Weather Creek base camp suggest that thinner snow cover, higher air temperatures, and lower albedo may contribute to earlier snow melt and ground exposure. Wind and sublimation of the dust-covered snow may also play an important part.

The presence of snow at Eureka in mid May and the accompanying cooler conditions may be attributed to its proximity to Slidre Fiord which was ice bound. Air temperatures at the coastal station were reduced by the advection of cooler air from the Fiord. An elevational difference of about 70 m between the base camp and upper basin sites may have partially contributed to the lower temperature at the upper basin, but lapse rate alone (between 0.6° and 1.0°C per 100 m elevation) cannot account for their large temperature differences. This will be studied further.

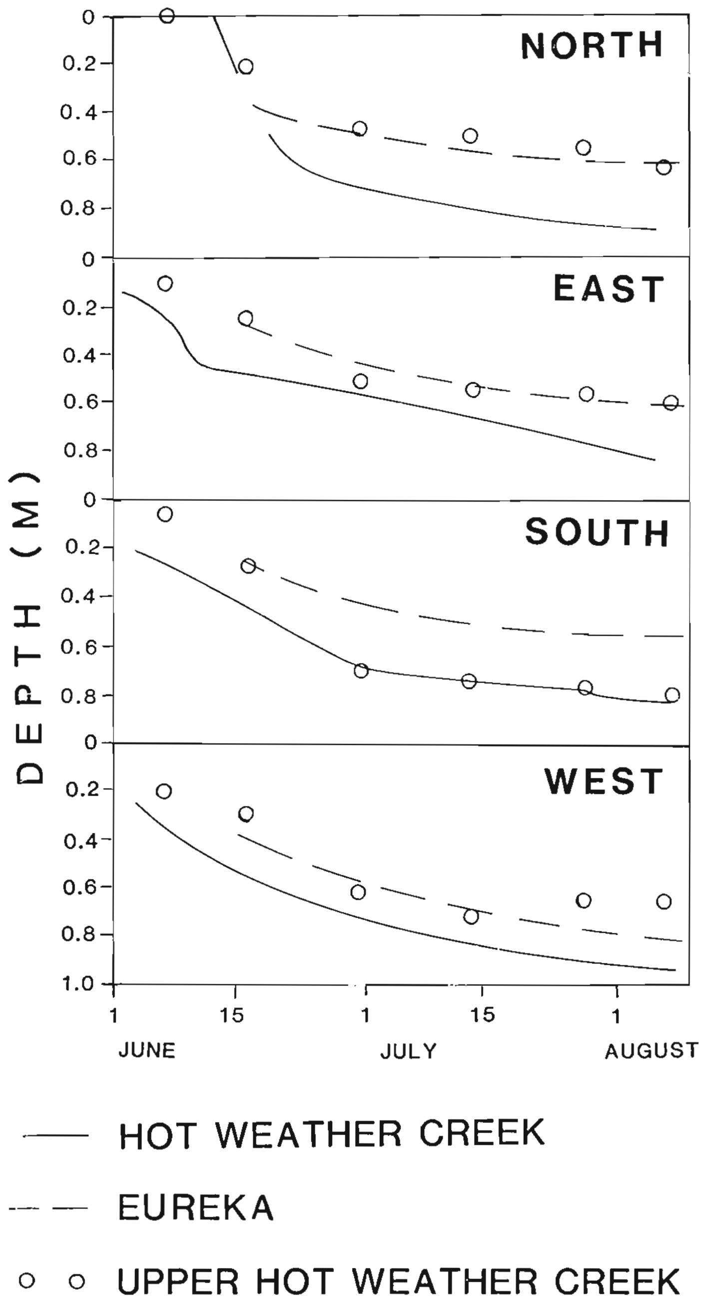

Figure 4. Comparison of frost table development at the middle of various slopes at the three study sites on Fosheim Peninsula.

Regional differences in microclimate within and outside the early snow-free zones produced considerable variations in ground thermal regime. Frost table development was affected noticeably. Given slopes of similar orientation and materials, those that were located within the early snow-free zones had deeper frost tables because of earlier ground thaw and because of warmer conditions that prevailed during the summer.

The spatial differences in the microclimate and the snowmelt regime at the three sites were not large enough to modify the floral diversity or dominants within their plant communities. Variability on a local scale between the slopes at each site overwhelms the regional differences so that the effect of early snow-free areas were not apparent from the vegetation surveys undertaken at the three study sites.

The patterns, instead, testify to the strong control that summer soil moisture and depth and local persistence of snow cover has over vegetation. Heath communities roughly correspond to areas where snow drifts persist several days to weeks after snow has generally disappeared from the area. They also have ample summer groundwater. The saline-tolerant herbaceous community reflects soils that probably never become saturated and experience prolonged summer drought conditions. Such soils do not experience runoff or rainfall in sufficient amounts to dissolve and remove the salts accumulated at the surface. This suggests that these upper slopes have only a thin snow cover in winter. Woody plant-dominated communities, the most common ones, reflect conditions where snow cover is continuous in winter, and some soil moisture is present throughout the thaw period.

The floras of the three study sites on Fosheim Peninsula represents the greatest diversity in the area, essentially the greatest in this entire warm, intermontane region. The thawing degree days and the length of the growing season at all sites are than adequate for the vascular plants to complete their life cycles. In 1988 and 1990 many vascular plants completed their life cycles and started into dormancy while air temperatures were still warm, and no climatic cue such as snowfall or freezing temperatures had occurred.

Differences between climate records from Eureka and from the other Fosheim locations raise concern regarding the representativeness of climatic records from coastal stations. The scarcity of long records in the Arctic necessitates heavy reliance upon the few coastal weather stations for baseline temporal data against which future climatic changes will be gauged. Results of this investigation demonstrate that point data cannot represent the entire surrounding landscape. Thus much caution must be exercised when extrapolating the coastal station data to a region.

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Preliminary analysis of data from the Lincoln Sea aeromagnetic surveys 1989-1990

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Abstract

An aeromagnetic survey of the Lincoln Sea was initiated in 1989, continued in 1990, and is expected to be completed in 1991. With a line spacing of 4 km, this is the most detailed aeromagnetic survey of the Canadian Arctic north of 82°N. General features in the data have been identified with regional geological structures of the Pearya terrane within the Northern Ellesmere Magmatic Belt and the Judge Daly Fault Zone between Ellesmere Island and Greenland. An interesting zone of high frequency anomalies has been mapped at the shelf-slope break north of Ellesmere Island. The proposed Wegener Fault Zone crossing the Lincoln Sea is not strongly represented in the data. Depth-to-source estimates have been made from the raw profile data assuming two-dimensional structures perpendicular to the flight path.

Résumé

Un levé aéromagnétique de la mer de Lincoln, entrepris en 1989 et poursuivi en 1990, devrait se terminer en 1991. Avec un espacement des parcours de 4 km, il s'agit du levé aéromagnétique le plus détaillé réalisé dans l'Arctique canadien, au nord de 82°N. Les données ont permis de relever les traits généraux des structures géologiques régionales du terrane de la Pearya au sein de la zone magmatique du nord d'Ellesmere et de la zone de failles de Judge Daly entre l'île d'Ellesmere et le Groenland. Une zone intéressante d'anomalies de haute fréquence a été cartographiée au rebord de la plate-forme continentale et du talus, au nord de l'île d'Ellesmere. La zone de failles de Wegener proposée traversant la mer de Lincoln ne ressort pas nettement des données. Des estimations de la profondeur à la source ont été faites à partir des données de profil brutes, supposant des structures bidimensionnelles perpendiculaires à la trajectoire de vol.

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INTRODUCTION

An aeromagnetic survey of the Lincoln Sea was initiated in 1989 (Hardwick et al., 1990) continued in 1990, and is expected to be completed in 1991. Funding was provided by the Department of National Defence (DND), the Defense Research Establishment Pacific (DREP), the Geological Survey of Canada (GSC), and the Institute for Aerospace Research (IAR). The Convair 580 aircraft operated by IAR was used to collect the data. The Convair 580 is instrumented for total-field, vertical and horizontal gradient aeromagnetic measurements. The Arctic survey missions provide an opportunity for further development of a platform for potential field research and enable the GSC and DREP to benefit from a highly developed airborne system. In the Lincoln Sea a total of approximately 20000 line kilometres has already been flown, and a further 2500 line kilometres remains to be completed.

Tectonically, the Lincoln Sea area is poorly understood. From the north, the area contains possible effects of the Alpha Ridge, the Lomonosov Ridge, and early developments of the opening of the Fram Basin. From the south, clues to the tectonic adjustments between Ellesmere Island and Greenland, including the area of Nares Strait, are contained in the crustal structure of the continental shelf that is reflected in the magnetic anomaly pattern. The aim of the Lincoln Sea surveys has been to provide relatively detailed magnetic data that may be used to clarify crustal features related to the opening of this part of the Arctic Ocean.

Previous aeromagnetic data in this area had a line spacing of approximately 20 km. The present survey represents the first relatively detailed survey this far north, with flight line spacing of 3-4 km and tie lines approximately 50 km apart. In addition, it is the first aeromagnetic survey that covers part of the land masses of northern Ellesmere Island and Greenland.

The sampling rate of the magnetics system is 8 Hz, with a bandwidth of 0-1.6 Hz. For an average aircraft speed of 110 m/s, this translates to an along-track resolution of approximately 66 m. For the 1989 survey, after compensation, the residual magnetic noise for all manoeuvres of the aircraft including 30° bank turns can be characterized as follows:

Total field 0.03 nT (rms)
Lateral gradient 8 pT/m (rms)
Vertical gradient 44 pT/m (rms).

In addition to its magnetic capabilities, the aircraft is used as a platform to conduct studies into precise navigation. Thus, relative to other Arctic aeromagnetic surveys, the Lincoln Sea program will clearly benefit from this auxilliary program at no extra cost. The aircraft has a full compliment of navigation systems including:

LTN90 Strapdown Inertial System (INS) *
CMA-786 C/A Code GPS Receiver (GPS) *
GNS 500 VLF-Omega Receiver (GNS) *
Decca Doppler'72 *
Video On-Top System *
Arnav-40 Loran-C Receiver
Del Norte 540 Transponder System
Norstar 1000 GPS P-Code Receiver o

* used for flight guidance in 1989 and 1990
o used for track recovery in 1990.

The survey elevation was specified to be 300 m over water or 300 m in a draped mode over land. Flight lines were laid out as north-south great circles equally spaced from the meridian running through the centre of the area and parallel to it. Positioning for the guidance was calculated by a Kalman filter that modeled the errors in the inertial navigation system using measurements of the data from the VLF-Omega receiver, the Doppler, occasional visual updates via the video system and, when available, GPS data. GPS coverage was available for about half of the survey.

For the post-flight track recovery, a different positioning technique was used. This consisted of weighted least-squares fitting a relatively simple error model of the following functions:

(VLF-Omega) minus (INS),
(Video On-Top) minus (INS),
(GPS) minus (INS).

The INS error model fitted the observed data quite well. When GPS was available for at least part of a flight, the final positioning was to within about 50 m (rms). For flights where no GPS was available, final positioning could be in error by as much as 800 m.

A cesium vapour magnetometer was set up as a base station magnetometer in order to measure and record the diurnal changes to the external field. These data were sampled and recorded at high sampling rates (3.6 Hz in 1989 and 8.0 Hz in 1990) and the time synchronization with the aircraft system clock was within 1 second. The base station was established at the seismic vault maintained by the GSC at Alert, away from the cultural noise of the operations area.

This report presents the results of preliminary analysis of the 1989 and 1990 portions of the Lincoln Sea aeromagnetic survey. A more complete description of the data acquisition systems and processing methods is outlined in Nelson et al. (1991).

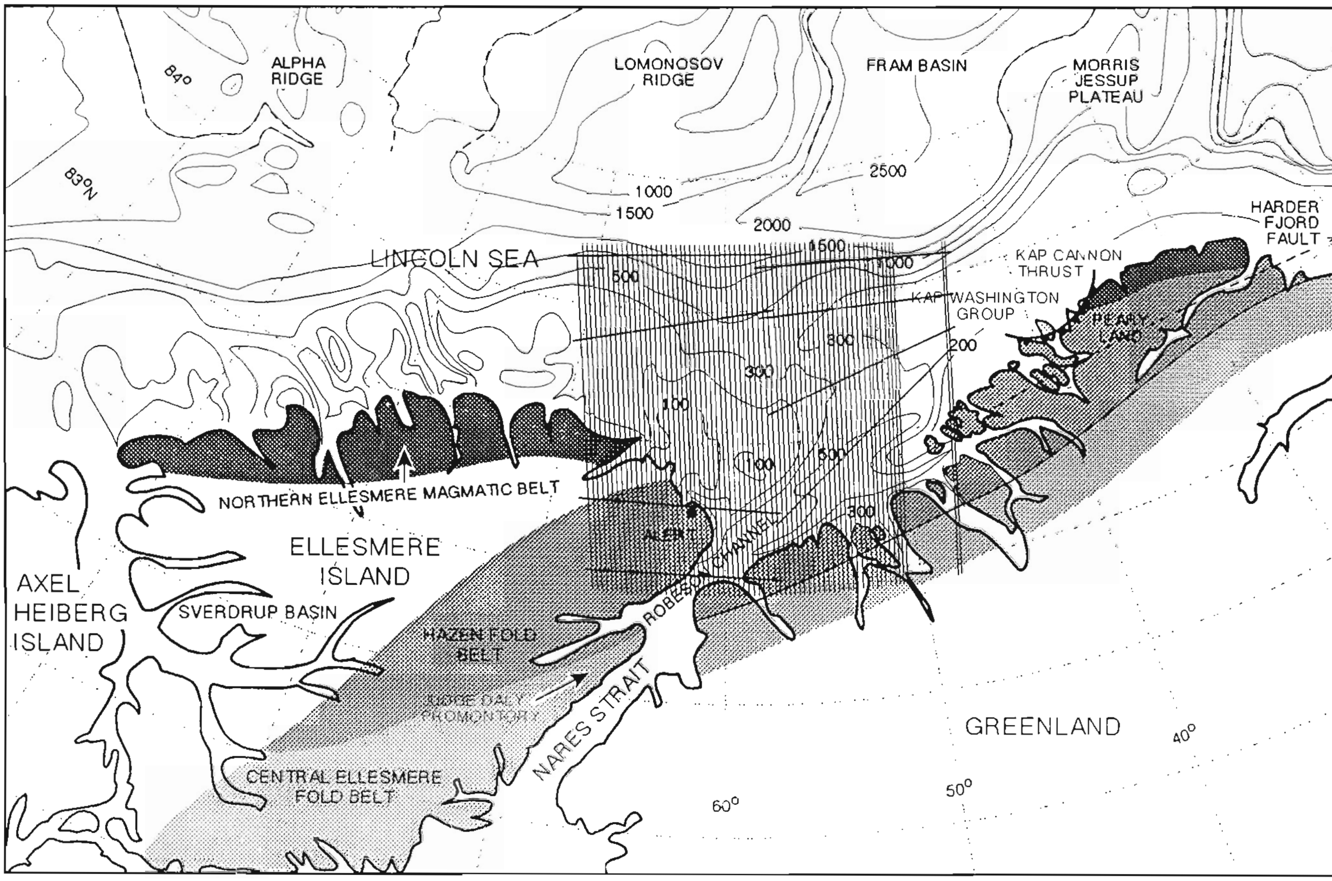

Figure 1. Generalized geological map of northern Ellesmere Island, northern Greenland, and the Lincoln Sea showing the north-south survey and tie line coverage.

REGIONAL GEOLOGY

Recent summaries of the regional geological and tectonic framework may be found in Trettin and Okulitch (1989), Trettin (1989) and Okulitch et al. (1990). The onshore geology of northeast Ellesmere Island can be divided into four major belts (Fig. 1). The Northern Ellesmere Magmatic Belt (NEMB) is composed mainly of the composite Pearya terrane (Trettin, 1989). Pearya terrane consists of four major successions ranging in age from Middle Proterozoic to Late Silurian. The two successions most relevant to the Lincoln Sea area are: (a) a coastal crystalline basement region of granitoid gneiss with lesser amounts of amphibolite, schist, marble and quartzite (metamorphism and intrusions are dated at 1.0-1.1 Ga) and (b) an Upper Proterozoic to possibly Middle Ordovician succession that includes sedimentary and volcanic rocks of greenschist to amphibolite metamorphic grade. The sources of magnetic anomalies are likely to be found in the crystalline basement rocks and in the mafic volcanics and dark clastics of succession (b) (Trettin, 1989). The terrain is rugged with a relative relief of 1000 to 2000 m.

A northeast tapering wedge of Sverdrup Basin sediments of the Clements-Markham Fold Belt lie southeast of the NEMB. Structural trends are generally NE. In the map area, thick shale and sandstone sequences form the basis for mountainous plateaus with elevations of over 2 km. Diabase dykes and sills have been found in the Sverdrup Basin region of Axel Heiberg Island and on the northwest corner of Ellesmere Island (Ricketts et al., 1985). Similar intrusive rocks may also be present beneath the Sverdrup Basin region of the survey area.

Immediately south of the Sverdrup Basin is the Hazen Fold Belt. This area is similar in age (Higgins et al., 1982), composition, and topography to the NEMB. It contains many tight folds and a number of major faults.

The Central Ellesmere Fold Belt (CEFB) trends NE from the southwest tip of Ellesmere Island to the Judge Daly Promontory. It is characterized by extensive folds and the Judge Daly Fault Zone that runs along Judge Daly Promontory and extends into Robeson Channel (Okulitch et al., 1990; Hood et al., 1985). The CEFB is underlain by limestone carbonates and covered by a layer of turbidites (Hurst and Kerr, 1982). The thickness of this sedimentary layer increases markedly from southeast to northwest (Kerr, 1977).

Northern Greenland contains three of the four geological belts trending sub-parallel to its northern coast (Fig. 1). The Greenland counterpart of the NEMB is found on the northern tip of Peary Land between 40°W and 30°W. This area has rough, mountainous terrain, and comprises folded Devonian carbonates and sandstone. It contains a few narrow dykes, predominantly oriented north-south (Soper et al., 1982). The northeastern extent of Sverdrup Basin rocks may underlie the southern Lincoln Sea area.

To the south of Peary Land lies the counterpart of the Hazen Fold Belt. The folds in this region are numerous and complex, and the NE-SW Harder Fjord Fault crosses the entire region. There are a great many dolerite dykes in Peary Land, ranging from 25-200 m in thickness, and predominantly oriented north-south. Chemical analysis (Soper et al., 1982) indicates that these dykes contain approximately 10% by weight FeO, and 3% Fe_2O_3. Thus, the dykes are extremely magnetic and the magnetic map of this geological province and its extension on Ellesmere Island should provide clues to the region's tectonic history.

South of the counterpart of the Hazen Fold Belt is a 20-30 km wide terrane similar in geology to the CEFB. The folds are not as complex as those to the north, and the thickness of the Devonian sediments decreases to the south. The region is still mountainous with peaks reaching elevations of 1000 m.

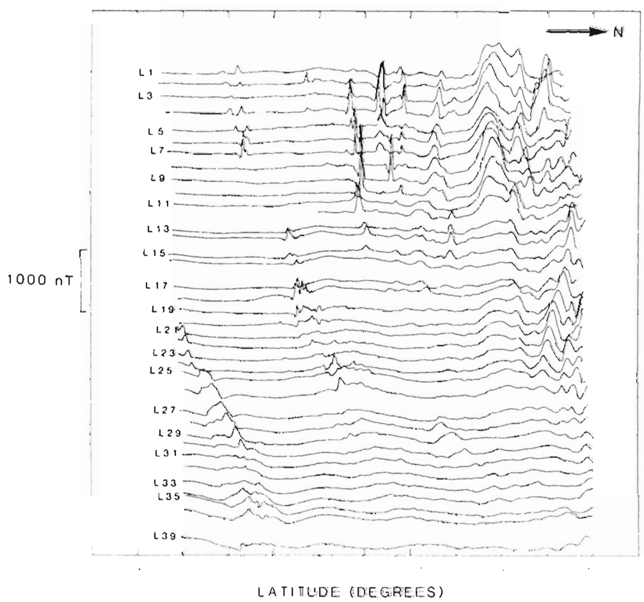

Figure 2a. Raw total-field profiles vs latitude for the 1989 data.

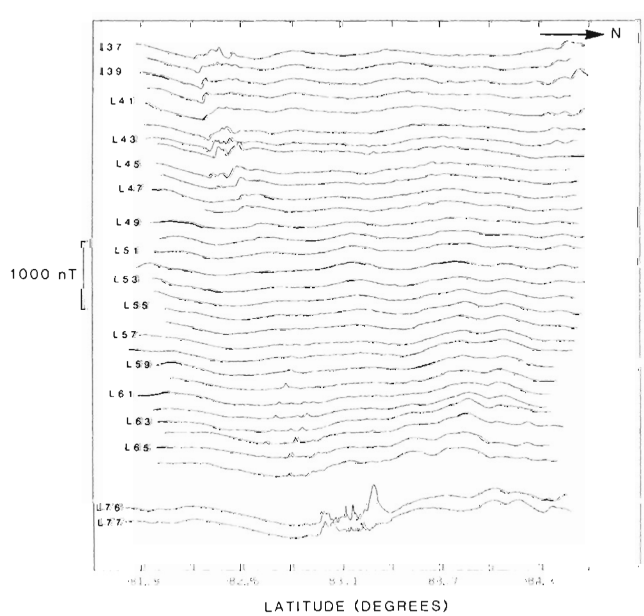

Figure 2b. Raw total-field profiles vs latitude for the 1990 data.

In addition to these three geological belts, volcanic rocks known as the Kap Washington Group occur near 40°W on the northwest tip of Peary Land. The age of the Kap Washington Group is estimated at 63 Ma. (Larsen et al., 1978). The Kap Cannon Thrust is a south dipping feature separating the Kap Washington Group from the surrounding fold belts.

The Lincoln Sea shelf is less than 200 m in depth from Ellesmere Island out to 80 km offshore. The shelf-slope break occurs at a depth of about 500 m near 84°N, and the continental slope extends to 1500 m at 84.25°N. The seafloor then rises slightly to about 1000 m at 85°N. Robeson Channel is typically 300 m deep at the entrance to the Lincoln Sea.

ANALYSIS OF THE TOTAL-FIELD DATA

Only the total-field data are analyzed in this report. The 1989 data set was analyzed independently of the 1990 data set, but the gridded data sets were merged to produce the contour maps and surface grids shown in Figures 3 and 4.

Figure 1 shows the flight line and tie line coverage. The raw total-field profile data are shown in Figures 2(a) (1989) and 2(b) (1990).

Total-field maps were produced from 1 Hz (subsampled) aeromagnetic and 1 Hz (interpolated) navigation data. Gridding was accomplished using the program "MAGIC" at DND. Figure 3 is a surface plot of the entire data set (1989 and 1990) produced by merging the two gridded data sets, and Figure 4 is a grey-scale contour plot of the same data.

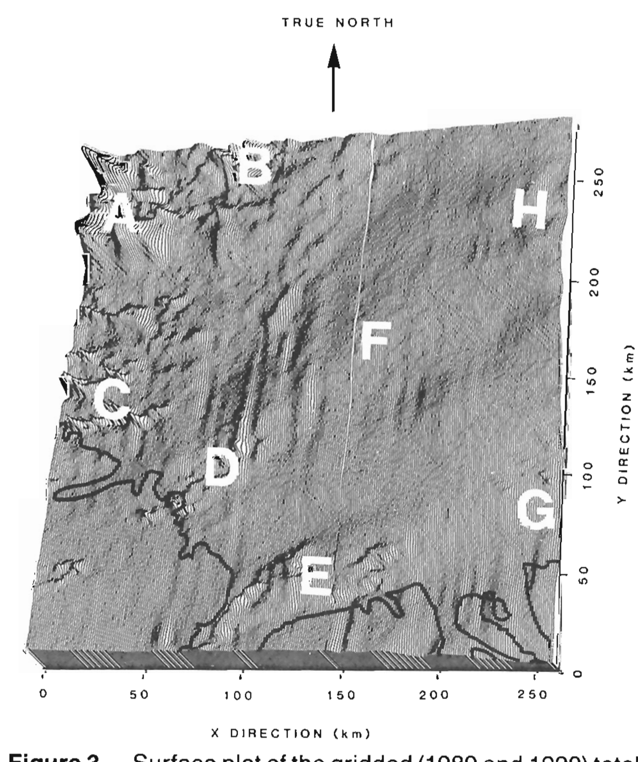

Figure 3. Surface plot of the gridded (1989 and 1990) total-field data.

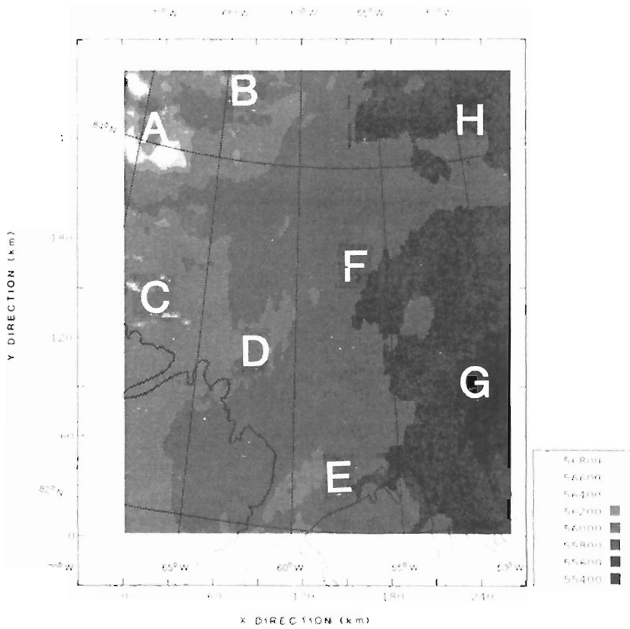

Figure 4. Grey-scale contour plot of the gridded (1989 + 1990) total-field data. Units are nT. Altitude = 300 m.

Figure 5 is a map of the survey area showing estimated depths to sources at the anomaly locations studied. For depth-to-source analysis, the total-field profiles were subsampled at 1 Hz (approximately 110 m sample spacing). Individual anomalies from the raw profile data were analyzed with "MAGMOD3", a commercial software package from Geosoft Inc. This program calculates the magnetic and geometric parameters of a thin dyke, an infinite extent tabular body, or a finite extent tabular body, by a least-squares comparison to the measured anomaly. The program assumes that the aircraft track was perpendicular to the strike and directly over the centre of the causative body. Some anomalies were also analyzed with the half-slope method (Rao and Babu, 1984) to determine depth-to-source.

DISCUSSION AND CONCLUSIONS

There is a significant variation in the regional anomaly pattern from the complex field north of Ellesmere Island to the relatively flat field north of Greenland (Fig. 2, 3). Higher frequency anomaly trends also change from northeast between Ellesmere Island and Greenland to east-west north of Ellesmere Island to northwest in the northwest corner of the area.

In the northwest corner (area A, Fig. 3), the large-amplitude, low-frequency anomaly near x=40 km, y=220 km has an estimated source depth of approximately 5 km below the sea surface. In area B to the northeast, however, magnetic source depths of 1-2 km appear close to the bathymetric contours (Fig. 5), so sources are covered by a few hundred metres of sediment or less. These anomalies occur at the southern corner of a larger region of high frequency anomalies that extends to the north of the survey area (Kovacs, 1982; Forsyth et al., 1990). Bathymetrically (Fig. 1), this region to the north appears to be a plateau. Kovacs (1982) suggested the plateau may have been connected to the break-up of a region of mafic material that also produced the Cap Washington volcanics and a region of high frequency anomalies immediately west of the Morris Jesup Plateau (Fig. 1). Irrespective of the postulated tectonic origins of anomalies in areas A and B (geological constraints are few), the following observations may be made:

(a) Area B anomalies have a wavelength of 5-10 km, and peak amplitudes of 300-400 nT. These anomalies are much higher frequency than those of the Lomonosov Ridge to the north, and the Lincoln Sea to the south (Forsyth et al., 1990) and appear to be higher frequency and more linear than anomalies over the Alpha Ridge to the west. As a fragment of the Barents shelf, generally underlain by sedimentary sequences (Okulitch et al., 1989), the Lomonosov Ridge should be characterized by a generally long wavelength, low amplitude magnetic anomaly pattern. The high frequency, lineated anomaly pattern of area B is not typical of a sediment-laden shelf environment. Instead, it may be an area characterized by many intrusives.

(b) Area A features a high amplitude anomaly with a wavelength of about 25 km amidst a field of anomalies similar to those in area B (Fig. 3, Fig. 5). Assuming the bathymetry is correct, these high frequency anomalies reflect relatively shallow sources beneath the continental shelf-slope and, like area B, are not characteristic of a sediment-laden shelf area. Although areas A and B appear to be separated by the continental slope, the similarity in magnetic anomaly pattern may indicate that similar tectonic activity has affected both regions. Depth-to-source estimates and bathymetry indicate the sources are covered with approximately 1.5 km or less of sediment.

The high-frequency, large-amplitude, east trending anomalies labelled C closely resemble the structural trends in the Pearya terrane nearby (Trettin, 1987) and suggest an extension of these trends beneath the continental shelf as far north as about 83°30'N. The estimated source depths are 300-800 m, indicating possible intrusive sources that almost reach the seafloor.

Area D contains a band of low-amplitude, medium-frequency anomalies trending ENE. These anomalies extend offshore, suggesting a continuation of the Clements Markham Fold belt that truncates the Pearya structures beneath the shelf. Most of the estimated source depths lie within 600 m of the seafloor.

Area E features two northeast-trending series of high-frequency anomalies. The western series is aligned with the centre of Robeson Channel as it enters the Lincoln Sea and appears to be a continuation of the Judge Daly Fault Zone (Okulitch et al., 1990). The anomalies to the east form an en echelon trend off the Greenland coast. Both anomaly trends appear to end about 50 km north of the coasts of Ellesmere Island and Greenland, but do not continue in a straight line out into the Lincoln Sea. The proposed Wegener Fault Zone (Kovacs, 1982) separating Greenland and Ellesmere Island is not a prominent magnetic feature across the survey area. Magnetic source depths suggest that sediment cover over the sources of these anomalies is only a few hundred metres.

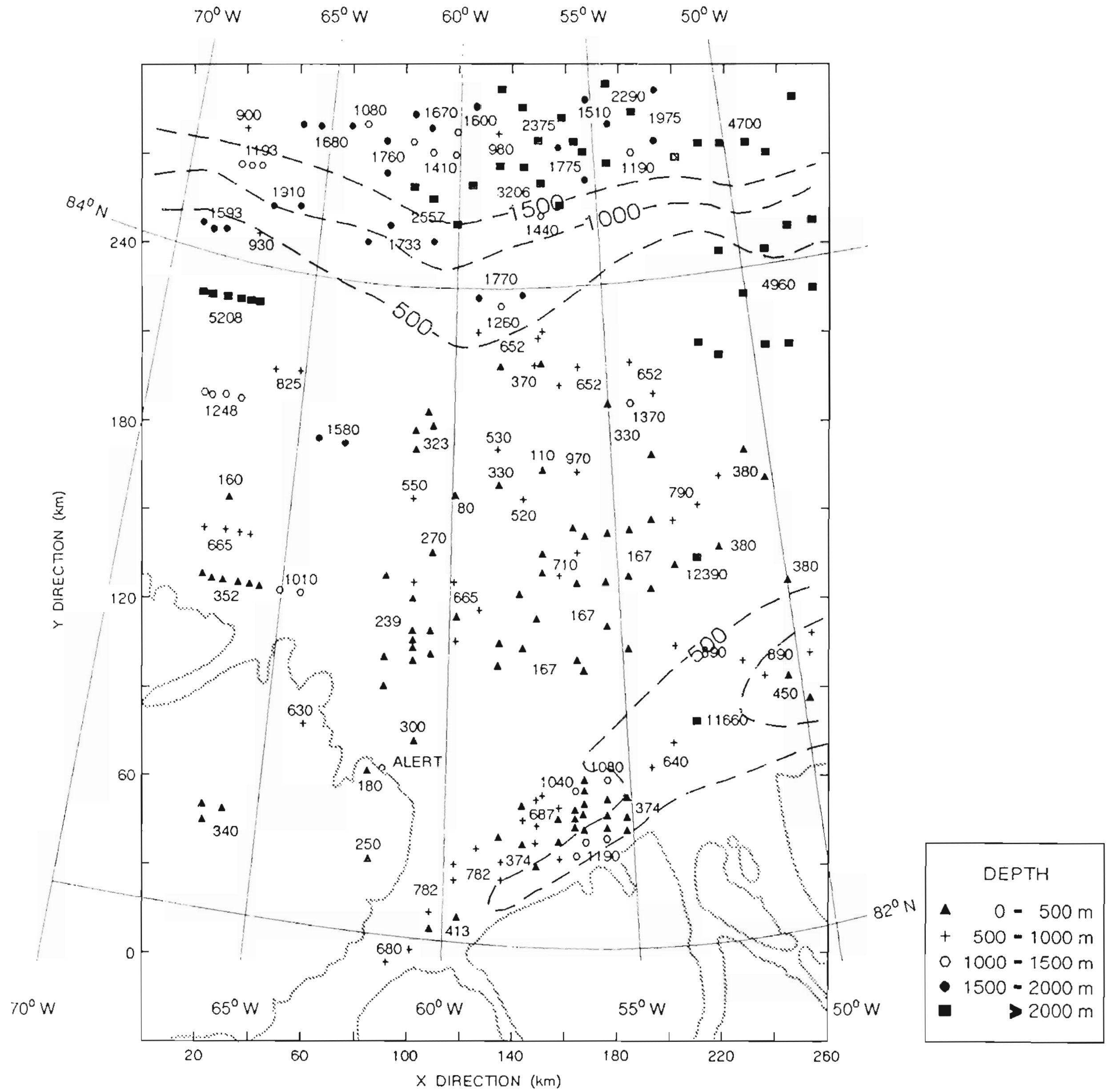

Figure 5. Estimated depth-to-source. Bathymetry in metres.

The central and eastern areas (F) are characterized by small-amplitude, high-frequency anomalies on top of small-amplitude, low-frequency anomalies (see also Fig. 2(b)). The sources of the high-frequency anomalies are near the seafloor. Because the amplitudes of these anomalies are about 50-100 times smaller than the intrusive anomalies in area C, either the magnetizations are 50-100 times less than those of the intrusive material, or the physical sizes of the sources are 50-100 times smaller. These high-frequency, shallow source anomalies have not been described in previous work from the area.

Depth-to-source analysis was difficult to perform on the low-frequency anomalies in area F. However, two longer wavelength anomalies on line "L51" yielded depth estimates of approximately 12000 m.

Area G contains several, high-frequency anomalies trending roughly E-W (see also Fig. 2(b), L60-66). These sharp anomalies are likely due to intrusive sources and may be covered by about 500 m of sediment.

The northeast corner of the survey area (H) contains low-frequency, low-amplitude anomalies trending roughly E-W, parallel to the north Greenland shore. The anomaly sources are beneath the continental shelf and may indicate the transition between oceanic and continental crust. It is estimated that about 3-4 km of sediment overlie the sources of these anomalies.

Many new magnetic features with unknown geological and tectonic significance have been identified on the preliminary magnetic map. Although many of the depth-to-source estimates generally agree with those of Kovacs (1982), the greater precision of these data allow definition of anomaly wavelengths, directions and strike length that have not been quantified before. The east-west features beneath the shelf in the western Lincoln Sea, the high frequency anomalies beneath the outer shelf north of Ellesmere Island, the termination of anomaly trends near the end of Robeson Channel and the high frequency trends from the central Lincoln Sea (area F) must be considered in future tectonic reconstructions for the area.

The 1989 and 1990 Lincoln Sea surveys have demonstrated that relatively detailed aeromagnetic surveying can be achieved in the high Arctic with the available navigational systems. In addition to acquiring high quality data, the 1989 and 1990 surveys included testing and developmental components with the aim of improving future missions. Navigational precision will increase, and post-processing requirements will decrease as the high latitude GPS coverage improves.

FUTURE WORK

It is hoped that the nine survey lines remaining in the Lincoln sea survey may be completed in April-May 1991. In addition, an aeromagnetic survey extending from Ellef Ringes Island to Ellesmere Island (79°-83°N, 85°-106°W) is proposed for 1991-1993. This coincides with the area of the 1985, 1986 and 1990-1990 seismic refraction surveys.

ACKNOWLEDGMENTS

The authors are indebted to the flight crew, ground crew, and technical support staff of the Convair 580 program at IAR for collecting these data under difficult conditions. Discussions with A. Okulitch and H. Trettin have added significantly to our geological appreciation of the anomaly trends. Constructive critiques of the manuscript and significant improvements were received from A.G. Green, M. Pilkington and P. Keating. We are grateful for the support and encouragement received from Robin Riddihough, GSC Chief Scientist.

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A newly recognized Lower Permian reef tract, west-central Ellesmere Island, Arctic Archipelago

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Abstract

A northeasterly trending belt of Lower Permian reefs, on the north shore of Greely Fiord, west-central Ellesmere Island, is documented for the first time. Many of the buildups are partly or completely dolomitized, yielding a very porous fabric that points to their economic potential. The Greely Fiord buildups are in the Mount Bayley Formation, most of them within the uppermost part. This reef tract occurs near the northwestern boundary of a subbasin that restricts the distribution of the deep water Mount Bayley evaporites. The stratigraphic relationships of several of the more than twenty reefs within this reef tract are discussed. The internal development of one of these reefs has been investigated in more detail. The reef originated as a fenestellid bryozoan-Tubiphytes core with subsequent construction characterized by biotic diversity in the flanks and crest, including phylloid algae, ramose bryozoans and fusulinaceans.

Résumé

La découverte d'une ceinture récifale, orientée nord-est/sud-ouest, et située sur la côte nord du fjord Greely, est ici rapportée pour la première fois. Plusieurs des ces récifs ont été dolomitisés, de façon partielle ou totale, résultant en une texture très poreuse qui pourrait présenter une importance économique. Les récifs du fjord Greely appartiennent à la Formation de Mount Bayley; la plupart se manifestent dans sa partie supérieure. Cette ceinture de récifs est située près de la marge nord-ouest d'un sous-bassin qui contient les évaporites d'eaux profondes de la Formation de Mount Bayley. Les relations stratigraphiques de plusieurs des quelques vingt récifs formant la ceinture récifale sont analysées. Le développement interne d'un récif en particulier a fait l'objet d'une analyse plus poussée. La croissance de ce récif s'est faite autour d'un noyau à Bryozoaires (fénestellés) et Tubiphytes, pour ensuite passer à une construction à plus grande diversité biologique tel qu'en atteste la présence d'Algues phylloïdes, de Bryozoaires rameux et de Fusulinidés dans les flancs et la crête.

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INTRODUCTION

A wide variety of upper Paleozoic carbonate buildups are known to occur in Sverdrup Basin (Davies et. al., 1989b). These buildups range in age from Moscovian (Late Carboniferous) to Artinskian (Early Permian), and are located in various areas on Ellesmere and Axel Heiberg islands in the Canadian Arctic Archipelago. Different reef-building organisms have contributed to the erection of these biogenic constructions. These organisms include: donezellid and beresellid tubular algae (Davies and Nassichuk, 1989), fenestellid and ramose bryozoans (Davies et al., 1989a), crinoids, phylloid algae and *Palaeoaplysina* (Beauchamp et al., 1989a), *Tubiphytes* (Beauchamp, 1989a), calcareous sponges (Beauchamp, 1989b), and a vast array of encrusting foraminifers and algae (Mamet et al., 1987). Morphologically, the buildups comprise small patch reefs, wide tabular banks and large reef-mounds, ranging in thickness from a few metres to more than 200 m. In some areas close to the paleoshelf edge, stacked and coalesced reef-mounds form nearly continuous tracts that rim much deeper-water depocentres. Spectacular examples of reef tracts are known from the Blind Fiord and northern Bjorne Peninsula areas of southwestern Ellesmere Island (Beauchamp et al., 1989a), and from the Blue Mountains of northwestern Ellesmere Island (Davies et al., 1989a). Such reef tracts, if discovered in the subsurface, could prove to be prime targets for oil and gas exploration.

This paper documents the recent discovery (summer of 1990) of another reef tract in Sverdrup Basin. This northeasterly trending belt of reef-mounds occurs on the north shore of Greely Fiord, immediately west of the mouth of Tanquary Fiord (Fig. 1). More than twenty reefs, ranging in thickness from a few tens of metres to more than 125 m, have been observed over a distance of 6 km in a series of sea cliffs overlooking Greely Fiord. These buildups are part of the Mount Bayley Formation. In addition to adding to the inventory of known carbonate reefs in Sverdrup Basin, this newly recognized occurrence may facilitate the interpretation of the relationships between the Mount Bayley evaporite and the surrounding carbonates. Furthermore, some of the buildups appear to be fairly porous, as a result of dolomitization. The origin and extent of this porosity-enhancing secondary process has yet to be determined. However, it is reasonable to infer that the combination of dolomitized buildups and subaqueous evaporites is worth investigating for its hydrocarbon potential. The discovery of organic-rich dark shale within the Mount Bayley Formation and underlying Antoinette Formation (J. Utting, pers. comm., 1990) makes the above prospect even more enticing.

A preliminary assessment of the Greely Fiord reef tract, based mostly on field observations, is provided in this report and, in addition to documenting these new buildups, some hypotheses are presented to explain the occurrence of such a reef tract in this part of Ellesmere Island. Various possibilities concerning the relationships of these buildups to the adjacent Mount Bayley evaporite are also examined. Additional biostratigraphic and detailed petrographic work on one buildup are currently under way (Olchowy, work in progress) and will provide the basis for a more comprehensive publication.

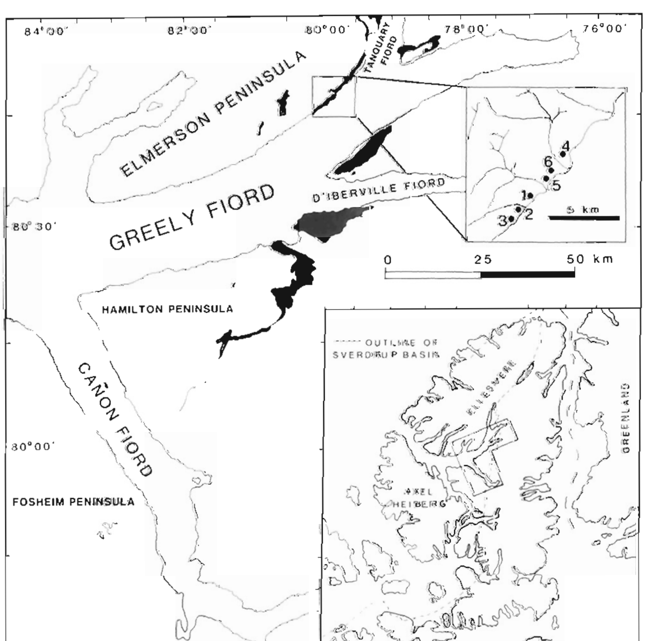


Figure 1. Map, showing the location of the Greely Fiord reef tract on west-central Ellesmere Island. Black areas represent outcrops of Mount Bayley Formation. Numbers in upper inset refer to localities described in text.

REGIONAL SETTING

The buildups form part of the Mount Bayley Formation, which outcrops widely on west-central Ellesmere Island. The Mount Bayley Formation, together with the Antoinette and Tanquary formations (Fig. 2), form a stratigraphic succession in Sverdrup Basin, ranging in age from Early Moscovian (Late Carboniferous) to Late Sakmarian (Early Permian) (Thorsteinsson, 1974; Beauchamp et al., 1989b, c). This succession is exposed on Fosheim and Hamilton peninsulas, and on the north shore of Greely Fiord, east and west of the mouth of Tanquary Fiord. The Antoinette/Mount Bayley/Tanquary succession is correlative with the Nansen Formation, a very thick sequence of shelf carbonate rocks that rims the central and deeper part of the Late Carboniferous to Early Permian Sverdrup Basin. These rocks are believed to have been deposited as a series of high-order glacio-eustatic(?) cycles superimposed on two long-term transgressive-regressive cycles (Beauchamp et al., 1989c).

The occurrence of the Mount Bayley evaporite in west-central Ellesmere Island suggests that this area was once cut off from the main Sverdrup Basin, where open marine carbonates (Nansen Formation) and deeper water muddy sediments (Hare Fiord Formation) are the coeval deposits. Several additional lines of evidence suggest that the area around Fosheim and Hamilton peninsulas was a subbasin of Sverdrup Basin. This subbasin, referred to here as the Fosheim-Hamilton subbasin (see also Morin et al., 1991; Scott et al., 1991; and Thériault and Beauchamp, 1991) appears to have been bounded to the northwest by a poorly defined "high", informally referred to as the "Elmerson High". This element, which is defined solely on the basis of thickness variations in various Permian and Triassic

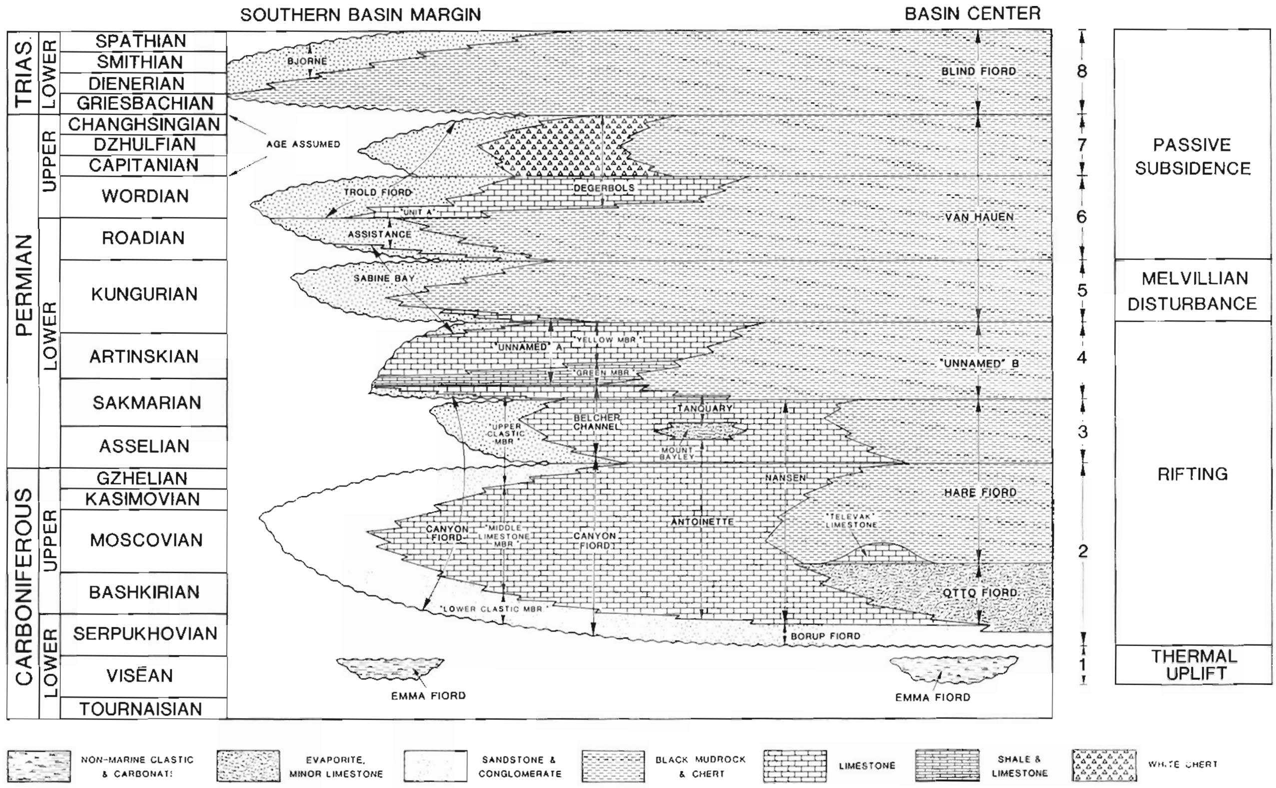


Figure 2. Upper Paleozoic second-and third-order transgressive-regressive sequences (1 to 7) in Sverdrup Basin along a schematic proximal-to-distal line of cross-section. Note stratigraphic position of Mount Bayley and Tanquary formations in Sequence 3. (From Beauchamp et al., 1989c.)

formations (J. Devaney, pers. comm., 1990), appears to have separated two highly different depositional provinces in the late Paleozoic, represented by quite different facies assemblages north and south of the presumed "high". The presence of the Mount Bayley evaporite south of the presumed "high" and its absence north of it provide additional evidence for inferring the presence of such a "high" in the environs of Elmerson Peninsula (Fig. 1).

Based on the knowledge that Sverdrup Basin was in a rift phase throughout its entire Late Carboniferous, and part of its Early Permian, history (Beauchamp et al., 1989b), a suggestion can be made that the "Elmerson High" might be a fault block that remained elevated throughout late Paleozoic to Early Triassic deposition. Regardless of its origin, this high was the most likely barrier isolating the Fosheim-Hamilton evaporite subbasin to the south while normal marine carbonates were being deposited immediately to the north. Further isolation of the Fosheim-Hamilton subbasin was likely provided by the North Greely Fiord reef tract, which grew in the vicinity of the "Elmerson High". This is indicated by the fact that tongues of Mount Bayley evaporites can be observed abutting against the buildups. These relationships are described in the next section.

The Greely Fiord reef tract occurs near the northwestern boundary of the Fosheim-Hamilton subbasin. As far as we know, this subbasin was asymmetrical with its thickest deposits lying immediately south of the "Elmerson High". From this locality the Mount Bayley Formation thins southeastward and southwestward through a complex diachronous intertonguing with the underlying Antoinette Formation (Morin et al., 1991). Wallace and Beauchamp (1990) have provided evidence to suggest that most of the Mount Bayley evaporite was deposited in relatively deep, marine water; that is, below the reach of wave generated currents. The interfingering of these evaporites with buildups that attain thicknesses of up to 125 m further supports the interpretation of a deep-water origin for the formation.

Based on their stratigraphic position, the age of most buildups in the Greely Fiord reef tract is likely to be middle to late Sakmarian (Early Permian). This conclusion is based on the occurrence of the conodonts *Streptognathodus constrictus* and *S. elongatus* in the first cycle of the Tanquary Formation, immediately above the Mount Bayley Formation south of D'Iberville Fiord. These conodonts belong to Zone P5 of Henderson (in Beauchamp et al., 1989b; see also Henderson, 1988), which is correlated with the Sterlitamakian sub-stage (Late Sakmarian) of the USSR. Wallace and Beauchamp (1990) presented some evidence, based on the number of shelf cycles in the overlying Tanquary Formation, to suggest that the top of the Mount Bayley Formation is synchronous across west-central Ellesmere Island. An additional piece of evidence may be the presence of a horizon dominated by a *Pseudoschwagerina*-like spherical fusulinacean within the first Tanquary cycle throughout the outcrop belt (Morin et al., 1991).

GREELY FIORD REEF TRACT

The Greely Fiord reef tract (centered about lat. 80°47'N and long. 79°57'W) comprises at least twenty buildups, ranging in thickness from less than 30 m to more than 125 m (estimated). Most of the buildups occur in the uppermost Mount Bayley Formation immediately below the contact with the overlying Tanquary Formation. Field relationships clearly indicate that the buildups grew during Mount Bayley deposition and prior to Tanquary cyclic sedimentation (see below). At least five buildups are totally encased within the Mount Bayley Formation. All buildups, regardless of their size, display a classic reef-mound morphology; that is, they are symmetrical mounds flanked on both sides by steeply inclined beds. Each buildup can be subdivided into three zones: 1) the core, which, in most cases, appears massive and structureless; 2) the flanks, which display well defined, inclined beds dipping between 25 and 45° off-mound; and 3) the crest, which is an area of bedded deposits linking the flanks. Some three-dimensional exposures show reef-mounds that are nearly circular, although some of the buildups may have a more oblong geometry. This cannot be recognized in two-dimensional exposures.

Most buildups form nearly vertical limestone walls perched up to 300 m above the surrounding surface. Helicopter "fly-bys" are the best way to observe the general geometry of the buildups and their relationships with the surrounding strata. Adventurous and, needless to say, fearless geologists might want to consider rappelling down the cliff faces to get additional information. Some buildups, however, are more easily accessible, and lend themselves to easier data gathering (see below). A shallow diamond-drilling program would probably be the best way to obtain three-dimensional information on the composition of the buildups.

A complex array of relationships is displayed between the buildups and the enclosing Mount Bayley and Tanquary strata. Intertongues are characterized by complex pinchouts of evaporites into the flanks of the buildups, and by onlapping, offlapping and toplapping relationships with Tanquary strata. The whole spectrum of relationships is displayed in Figures 2 to 7, and summarized in Figure 8. Each case is briefly described and interpreted.

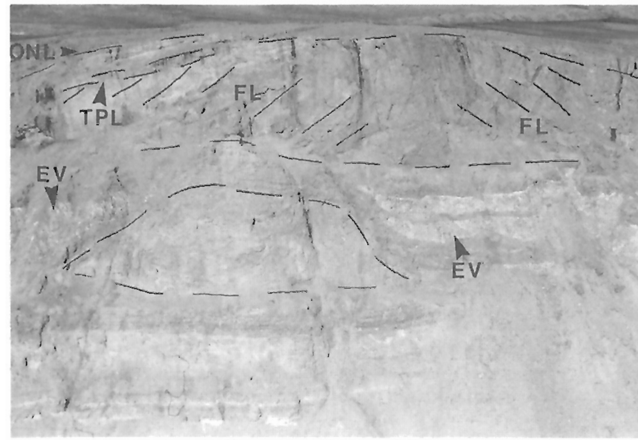


Figure 3. Two large buildups (80 m and 125 m thick) at Locality 1. Note: evaporite strata (EV) abutting against lower buildup flanks; inclined buildup flanks (FL) in the upper buildup; erosional toplap surface (TPL) in the upper buildup; and onlap (ONL) of Tanquary bedded carbonates.

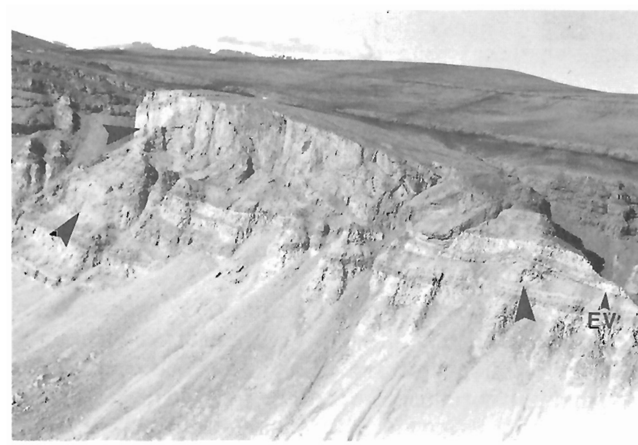


Figure 4. Small buildup (right arrow) encased within Mount Bayley evaporite strata, 0.5 km northeast of Locality 1. Note: evaporite strata (EV) abutting against buildup flanks, and lower and upper buildups at Locality 1 (left arrows).

Locality 1

Two large buildups occur at Locality 1 (Fig. 3). One buildup, approximately 80 m thick, is completely encased within the Mount Bayley Formation. A much larger buildup (125 m thick) occurs at the contact between the Mount Bayley evaporites and the Tanquary shelf carbonates. Two prominent tongues of Mount Bayley evaporites abut and pinch out against the flank of the lower buildup, clearly indicating that evaporite deposition was later than buildup growth. A similar relationship occurs around a smaller buildup that occurs 0.5 km to the northeast (Fig. 4).

The larger buildup that marks the Mount Bayley/Tanquary contact is offset by about 150 m relative to the lower buildup (Fig. 3). Two sets of inclined beds dipping between 35 and 40° flank this buildup. These flanking strata are wrapped around the buildup core, which is about 40 m high and 120 m wide. The contact between the flanks and the overlying Tanquary carbonate rocks is characterized by an erosional toplap surface. The Tanquary cycles appear to onlap that surface toward the buildup crest. The most plausible sequence of events for explaining these relationships is: 1) buildup growth, possibly during a sea level rise; 2) subaerial exposure and erosion of the buildup as a result of a sea-level drop; and 3) subsequent onlap of cyclic shelf carbonate as a result of a sea-level rise and marine transgression.

Locality 2

A reef-mound, 100 m thick, occurs at Locality 2 (Fig. 5). Both flanks are well developed, dipping between 35 and 40° off mound. Thin evaporite tongues of the Mount Bayley Formation appear to dissappear laterally into the buildup flanks. The exact nature of this relationship is uncertain because of a lack of exposure. An erosional toplap surface marks the top of the buildup. This is clearly indicated by the truncation of the flanks beneath the exposed Tanquary carbonate rocks. The latter appear to drape around the topography created by the buildup. Like the buildup at Locality 1, this buildup appears to have been subaerially exposed and eroded prior to Tanquary deposition. Some of the older Tanquary cycles may onlap the buildup, but this relationship remains speculative because of a lack of exposure.

Locality 3

The buildup at Locality 3 appears to be composite as it can be divided into at least two separate entities characterized by different dip attitudes (Fig. 6). The two masses apparently coalesced to form a single body, more than 100 m thick, the top of which was eroded (toplap surface). One Tanquary bed clearly onlaps this erosion surface, whereas the younger beds appear to be wrapped around the buildup.

Figure 5. Large buildup (100 m thick) at Locality 2. Note: evaporite strata (EV) abutting against buildup flank; inclined buildup flanks (FL); erosional toplap surface (TPL); and Tanquary bedded carbonates (TQ) wrapping around the buildup.

Figure 6. Large buildup (100 m thick) at Locality 3. Note: two coalesced masses (arrows); erosional toplap surface (TPL); and onlap (ONL) and Tanquary bedded carbonates (TQ) wrapping around the buildup.

Locality 4

The buildup (100 m thick) at Locality 4 is relatively large (Fig. 7), and displays well defined flanks wrapping around a more massive core. The flanks dip between 35 and 45°. The buildup appears to be onlapped by Mount Bayley evaporites, although this relationship is obscured by poor exposure.

Locality 5

A partially preserved buildup, displaying only its northeastern flank and part of its core, is exposed at Locality 5 (Fig. 8). At this locality the cliff wall is 80 m thick, but the actual buildup must have been much larger than that. The northeastern flank is onlapped by carbonate and evaporite strata.

Relationships

More than a dozen additional buildups occur in the Greely Fiord reef tract. All of them display the kind of relationships

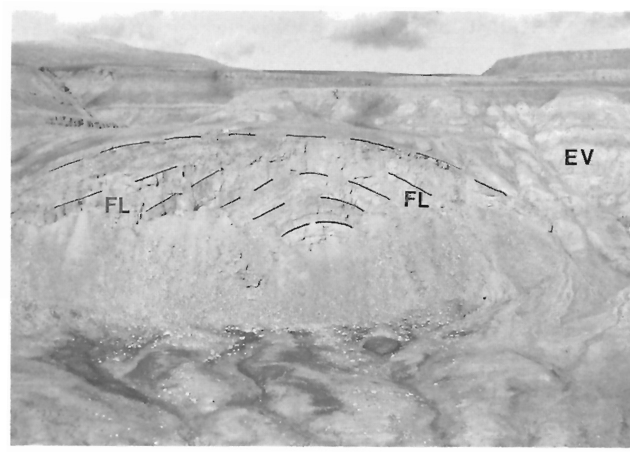


Figure 7. Large buildup (100 m thick) at Locality 4. Note: evaporite strata (EV) abutting(?) against buildup flanks, and inclined buildup flanks (FL).

Figure 8. Partly eroded buildup at Locality 5. Note: evaporite strata (EV) abutting against buildup flanks; and inclined buildup flank (FL) wrapping around buildup core (CO).

described above. The nature of the relationships for some of the largest buildups marking the contact between the Mount Bayley and Tanquary formations is summarized in Figure 9. These relationships suggest the following scenario:

1. The buildups grew in open, relatively deep, marine water. Buildup thickness represents the minimum bathymetry of the surrounding seafloor in the latest stage of buildup growth. The true bathymetry is in fact represented by buildup thickness plus the depth of the water body on top of the buildups. The largest buildups are around 125 m thick, which suggests a bathymetry in that order or slightly greater. Evidence derived from the study of one specific mound (see below) suggests that buildup growth was initiated during a sea-level rise, and carried on during a sea-level highstand.
2. Subsequently, the buildups were subaerially exposed, as a result of a substantial sea-level drop. This led to the erosion of the buildup tops. The exposed reef tract must have created a series of barrier islands that induced a restriction of circulation in the adjacent subbasin, as suggested by the fact that the Mount Bayley evaporites abut against the buildup flanks.
3. Renewed transgression led to the progressive onlap and covering of the buildups by cyclic shelf carbonates of the Tanquary Formation. Thus, onlapping relationships are observed before the Tanquary carbonates ultimately flatten out.

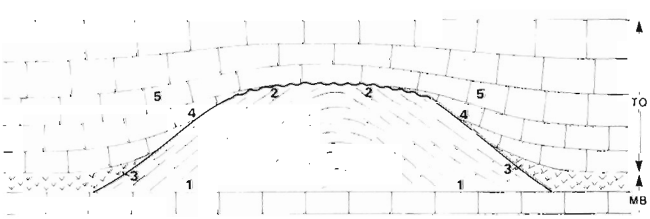


Figure 9. Summary of relationships between the large buildups in the upper Mount Bayley Formation (MB) and overlying Tanquary Formation (TQ). Note: 1) downlap surface between buildup flanks and carbonate sole; 2) erosional toplap surface; 3) evaporite strata abutting against buildup flanks; 4) onlap of Tanquary carbonates; and 5) Tanquary carbonates wrapping around the buildup.

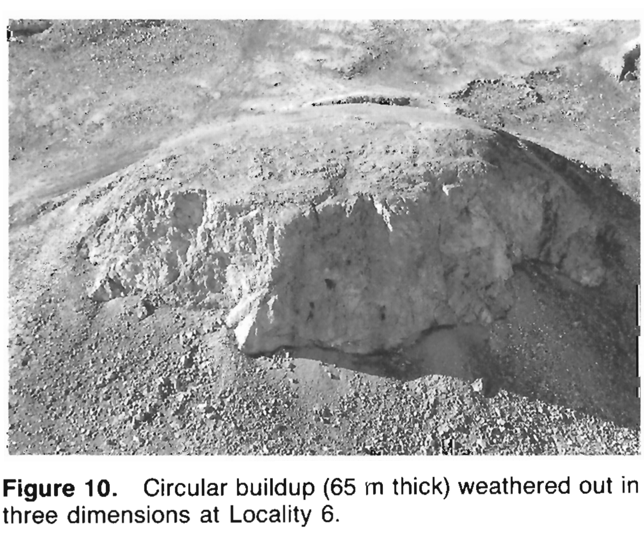


Figure 10. Circular buildup (65 m thick) weathered out in three dimensions at Locality 6.

INTERNAL BUILDUP GEOMETRY

One accessible buildup was investigated in detail (Locality 6). This reef has weathered out in such a way that it can be observed in three dimensions (Fig. 10), although its relationships with the surrounding strata are partly obscured. Mount Bayley evaporite strata outcrop a few tens of metres to the southwest, at approximately the same stratigraphic level as the lowest portion of the buildup. It is likely that these evaporites abut against the buildup flank. Tanquary cyclic shelf carbonate rocks overlie the buildup.

The reef is roughly circular in plan. It is 65 m thick and approximately 200 m in diameter. Well defined symmetrical flanks, dipping at 39 to 45°, are wrapped around a massive core, 30 m thick and 70 m in diameter (Fig. 11A). A nearly flat area, referred to as the buildup crest, marks the top of the reef. More than 100 samples were gathered from these three distinctive units (core, flank, crest), each of which is characterized by a very different suite of rocks.

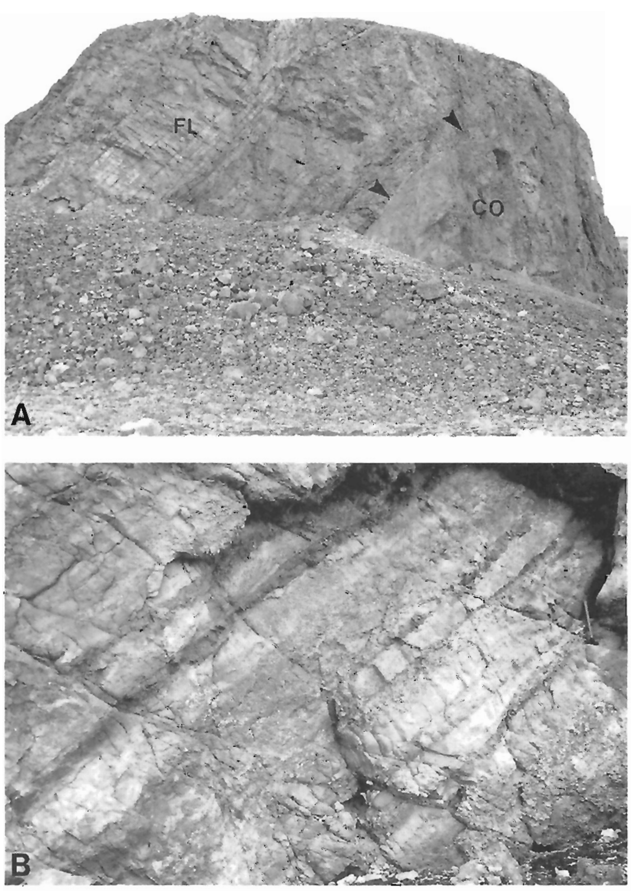
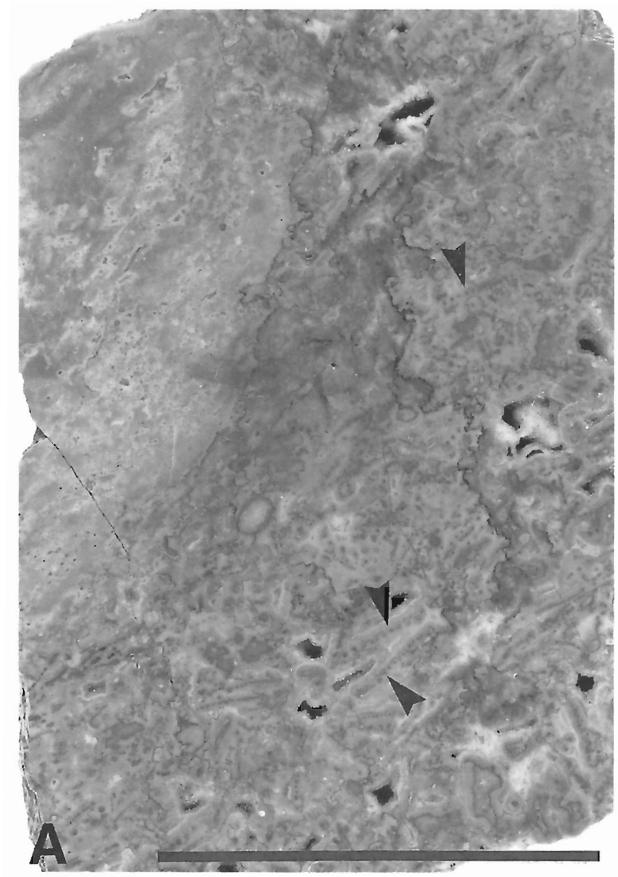
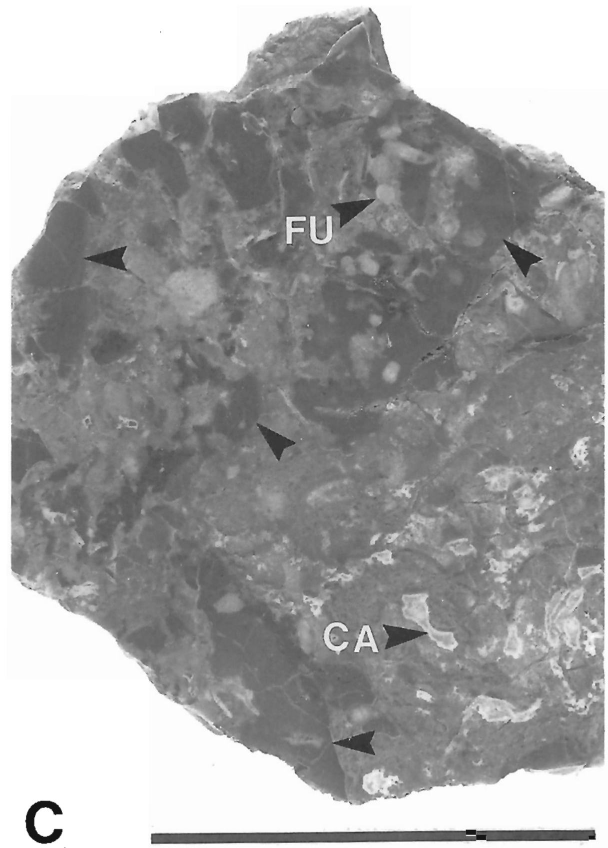


Figure 11. Details of the buildup at Locality 6.
A. Contact (arrows) between massive core (CO) and inclined southwest flank (FL).
B. Alternating light and dark beds on inclined flank. Hammer for scale.





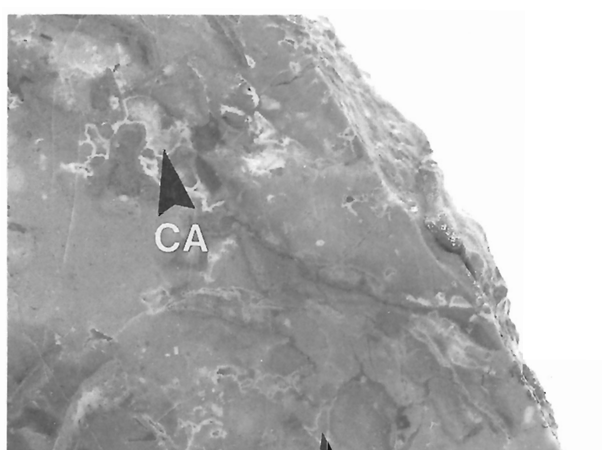
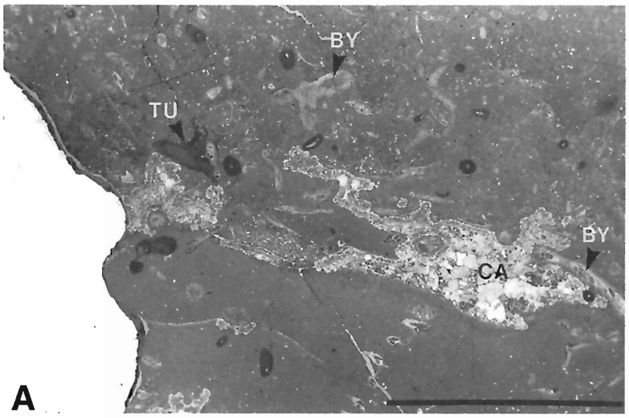


Figure 12. Polished slabs of buildup lithofacies from Locality 6.
A. Fenestellid bryozoan "cementstone" from core. Note fenestellid bryozoans in cross-sections (arrows). Scale bar is 5 cm.
B. Bryozoan-*Tubiphytes* wackestone from core. Note open-space cavities (CA) of unknown origin. Scale bar is 5 cm.
C. Carbonate breccia from flank. Note angular debris (arrows), fusulinaceans (FU), and open-space cavities (CA). Scale bar is 5 cm.

Core

The core is formed by a massive body of light grey limestone represented by two different facies. One facies consists of a bryozoan-*Tubiphytes* "cementstone"; that is, a meshwork of in situ fenestellid bryozoans encrusted by the enigmatic organism *Tubiphytes*, and cemented by multi-generations of isopachous cements (Fig. 12A). The other facies consists of a bryozoan-*Tubiphytes* wackestone in which well preserved fenestellid and ramose bryozoans fronds are associated with common *Tubiphytes* in a very pure micrite (Figs. 12B, 13). Variably shaped open-space cavities, now thoroughly cemented, also occur in association with this facies. These cavities, of unknown origin, range in size from a few millimetres to a few centimetres. Other accessory organisms include: brachiopods, sponge spicules, and the small foraminifer *Eolasiodiscus*. The bryozoan-*Tubiphytes* wackestone facies is locally brecciated and infilled by a darker micritic material, which may constitute an early diagenetic phase.



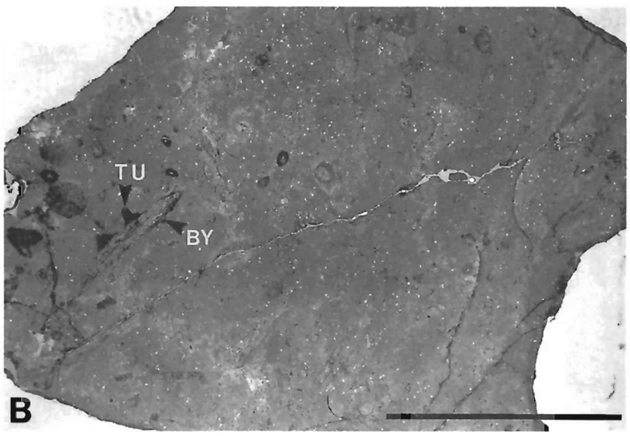
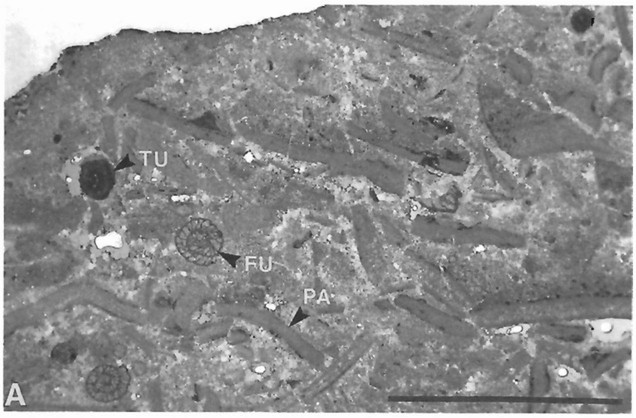


Figure 13. Buildup core lithofacies at Locality 6.
A. and B. Bryozoan-*Tubiphytes* wackestone. Note: bryozoans (BY), *Tubiphytes* (TU) and open-space cavities (CA) of unknown origin. Scale bar is 1 cm.



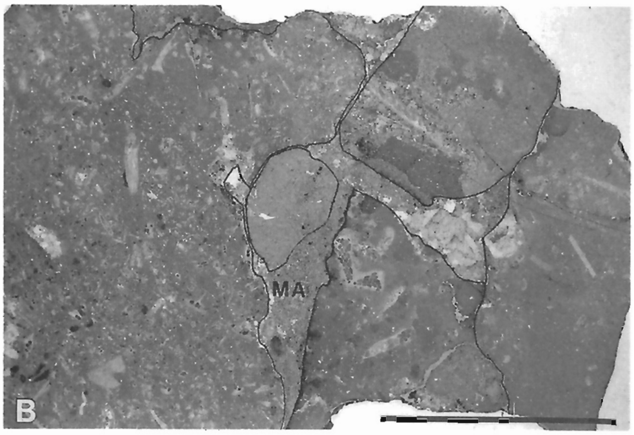


Figure 14. Buildup flank lithofacies at Locality 6.
A. *Tubiphytes*-fusulinacean-phylloid algal grainstone (rudstone). Note: phylloid algae (PA), *Tubiphytes* (TU), and fusulinaceans (FU). Scale bar is 1 cm.
B. Carbonate breccia. Note: diversity of carbonate debris (black outlines); and finely crystalline dolomitic matrix (MA). Scale bar is 1 cm.

Flanks

The reef flanks comprise relatively well bedded, inclined, carbonate strata that dip between 39 and 45°. A wide and random spectrum of facies is displayed in the flank beds. The most obvious feature of the flanks is an irregular alternation of dark and light bands. This feature is common to all buildups encountered in the north Greely Fiord reef tract (Figs. 3-8). The dark bands comprise a finely crystalline, vuggy, silty dolostone. Some vugs clearly result from dissolved fossils. The light bands form three distinct lithofacies, the most common of which is a variably dolomitic, coarse to very coarse, fossiliferous grainstone with abundant bryozoans and *Tubiphytes*, and a variable content of echinoderms, brachiopods, fusulinaceans, and phylloid algae (Fig. 14A). A less common light lithofacies comprises finer grained, silty packstone and grainstone that are generally dolomitized. Fossils include common to abundant ramose and fenestellid bryozoans and *Tubiphytes*, associated with fusulinaceans, brachiopods, and echinoderms. A third light-coloured lithofacies, which is far less abundant, consists of a carbonate breccia with angular debris ranging in size from 2 mm to 3 cm (Figs. 12C, 14B). This debris occurs in a finely crystalline, silty, dolomitic matrix. The debris represents a variety of facies, the most important of which is bryozoan-*Tubiphytes* wackestone that is characteristic of the core. Debris from the fossiliferous packstone and grainstone that typically occur in the flanks is also present. Some of the debris is resedimented breccia, which suggests a complex depositional history.

Each flank bed appears to be the product of a rather short-lived sedimentary event. In addition to the presence of carbonate breccias that can be interpreted as debris or grain flows, most beds are either graded or inversely graded, generally poorly sorted, and commonly scoured — all of these features suggest rapid, if not catastrophic, deposition in some sort of slope environment (the buildup flank is undoubtedly a slope environment). It is likely that all this material came from the crest of the buildup. Storm-and/or gravity-induced transport must have relocated this material on the buildup flanks. The high angle of repose (up to 45°), however, suggests that the flank slopes must have somehow been stabilized during buildup growth. Early marine cementation, and perhaps the binding action of some encrusting organisms are the most likely factors that led to an early stabilization of the slopes. An in-depth petrographic analysis of the flank facies will be required to ascertain the actual stabilization factors involved.

Crest

The crest is the more or less horizontal area that links the flanks. It could only be accessed by walking on top of the exposed reef. This led to a very incomplete observation of its vertical succession, since the crest was continuously rising while the buildup was expanding laterally. Facies observed on top of the reef included a well bedded, medium grained, dolomitic packstone with abundant bryozoans, *Tubiphytes* and fusulinaceans, in association with minor palaeotextularid foraminifers, echinoderms, and brachiopods.

Buildup growth

Based on field observations and a brief preliminary petrographic assessment, the following simple model for buildup growth is proposed. Firstly, the buildup core was formed on the seafloor. The abundance of fenestellid bryozoan "cementstones" and wackestones in the core suggests a rather deep water origin. Indeed, identical facies are characteristic of Waulsortian-like constructions that occurred at different times in Sverdrup Basin. These include the large Moscovian buildups encased within the Hare Fiord Formation in the Blue Mountains (Davies et al., 1989a), a Sakmarian buildup on the west side of Blind Fiord (Beauchamp, 1989a), and the Artinskian buildups found east and west of Blind Fiord in an unnamed formation (Beauchamp, 1989b). In all three cases, the absence of phylloid algae, and for that matter any kind of algae, has been interpreted as indicating fairly deep water during buildup growth. A slope to deep ramp environment is also indicated by the study of adjacent sediments in these occurrences. A minimum water depth of at least 65 m (buildup thickness) is suggested for the Lower Permian buildups under study. As there is no evidence that the buildup reached fairweather wave base, a minimum of 20 to 30 additional metres can be added to the estimate of a minimum bathymetry. These values are in agreement with the conclusions reached by Wallace and Beauchamp (1990) that the bulk of the surrounding Mount Bayley evaporite was deposited in relatively deep water, below storm wave base. It should be pointed out, however, that evaporite precipitation probably occurred while the reefs were subaerially exposed.

Once the core was formed on the seafloor, the buildup grew somewhat like an onion shell; that is, outward as a series of successive layers (flank and crest) wrapping around the original core. As discussed above, it appears that the flank sediments were derived from the crest during a series of short-lived events. A combination of gravity sliding and storm generated washover is probably responsible for the relocation of these sediments. Based on this interpretation, one can speculate that the contact between the massive core and the bedded flank/crest shell may represent the point in time when the buildup reached the storm wave base. This would in turn suggest that the buildup shallows upward, a hypothesis that is supported by the presence of the reworked phylloid algae and shallower water biota, such as fusulinaceans, that are found in some of the flank beds.

Thus, it is likely that buildup growth started during a transgressive phase and continued during the subsequent sea-level highstand. The following sea-level drop must have led to the cessation of buildup growth as the crest impinged upon fairweather wave base. Whether or not the buildup was later subaerially exposed and eroded has yet to be determined, but, based on the observed relationships in the other buildups, this remains a genuine possibility.

ECONOMIC POTENTIAL

Some of the buildups in the Greely Fiord reef tract appear to be partly to completely dolomitized, yielding a very porous fabric that may have turned many of the reefs into quality reservoirs. The origin of the dolomitizing fluids is presently unknown, but it is well known that reefal carbonates lying in close proximity to a sequence of evaporites are often dolomitized. The reef/evaporite combination has been tested for its hydrocarbon potential in many areas of the world, and very lucrative discoveries have been made. These include well documented discoveries of oil and gas in the Michigan (Silurian) and Delaware (Upper Permian) basins in the USA, and in the Ural Mountains (Lower Permian) of the USSR.

The numerous so-called pinnacle reefs of the Michigan Basin, many which are of the same size and shape as the Greely Fiord buildups, are relevant to this study. Pinnacles in the Michigan Basin are dolomitized and completely encased in evaporite strata that provided both the source of dolomitizing fluids and a sealing mechanism.

CONCLUSIONS

The newly discovered reef tract on the north shore of Greely Fiord comprises more than twenty buildups, ranging in thickness from less than 30 to more than 125 m. At least one buildup was found to have evolved vertically from a fenestellid bryozoan-*Tubiphytes* reef into a more complex construction that contains phylloid algae, ramose bryozoans and *Tubiphytes* among other organisms. All buildups appear to have been initiated in fairly deep water (probably exceeding 125 m in some cases), and then grew during a sea-level highstand. A subsequent sea-level drop likely led to the termination of buildup growth, and ultimately to the subaerial exposure and erosion of buildup crests and flanks. Based on observed relationships with the surrounding Mount Bayley Formation, it is likely that the exposed reefs became a tract of barrier islands that contributed to isolating the rather deep subbasin (of Sverdrup Basin) to the south (the Fosheim-Hamilton subbasin), in which relatively deep water, subaqueous evaporites were deposited.

ACKNOWLEDGMENTS

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Sedimentological highlights of the Lower Triassic Bjorne Formation, Ellesmere Island, Arctic Archipelago

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Devaney, J.R., Sedimentological highlights of the Lower Triassic Bjorne Formation, Ellesmere Island, Arctic Archipelago; in Current Research, Part B, Geological Survey of Canada, Paper 91-1B, p. 33-40, 1991.

Abstract

The Bjorne Formation of Ellesmere Island was deposited on the eastern margin of Sverdrup Basin during three Early Triassic transgression-regression cycles. Well exposed sandy braided-river deposits are dominated by sheetlike planar cross-sets and locally unimodal paleocurrent indicators, with only minor thin shale lenses. In contrast, meandering-river deposits consist of fining-upward sequences with thicker and much more laterally persistent shale units containing a variety of overbank features (splays, worm burrows, paleosols), including the first vertebrate tracks reported from Sverdrup Basin. Marginal marine, prograded, shoreface parasequences with Skolithos ichnofacies burrows are gradational to storm-dominated shelf facies of the more distal and shaly Blind Fiord Formation. Vertical persistence of facies assemblages, paleocurrent data (which show a radial, deltaic pattern), and subregional facies changes suggest the repeated progradation of fluvial/wave interaction deltaic complexes in the central Ellesmere Island area.

Résumé

La Formation de Bjorne de l'île d'Ellesmere constituait la marge orientale du bassin de Sverdrup au cours des trois cycles de transgression et de régression du Trias précoce. Les dépôts sableux de rivière anastomosée, bien exposés, sont dominés par des ensembles de couches obliques, plans et stratiformes. On y observe également quelques rares lentilles minces de shale et, par endroits, des indicateurs de paléocourants unimodaux. Par contre, les dépôts de rivière à méandres sont constitués de séquences positives avec des unités de shale plus épaisses et beaucoup plus persistantes latéralement contenant une variété de structures de dépôt de plaine d'inondation (cônes secondaires, galeries d'annélides, paléosols) et comportant les premières traces de vertébrés observées dans le bassin de Sverdrup. Aux paraséquences progradantes de l'avant-plage, de milieu margino-marin, dans lesquelles sont reconnus des ichnofaciès à Skolithos, succèdent, de façon graduelle, les faciès de plate-forme, où l'action des tempêtes prédomine, de la formation plus distale et plus shaleuse de Blind Fiord. Dans la partie centrale de l'île d'Ellesmere, la persistance verticale des assemblages de faciès, les données sur les paléocourants (qui indiquent une structure radiale, deltaïque) et les changements de faciès subrégionaux dénotent la progradation répétée de complexes deltaïques caractérisés par l'interaction des processus fluviaux et des vagues.

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INTRODUCTION

Forming part of the Carboniferous to Tertiary Sverdrup Basin succession (Trettin, 1989) in Canada's northern Arctic Islands, the Bjorne and Blind Fiord formations are the proximal and distal portions, respectively, of three major Early Triassic transgression-regression cycles (Embry, 1986, 1988). The sandstone-dominant Bjorne Formation is of fluvial and marginal marine origin, and is both laterally and vertically gradational to the more shaly basinal facies (shelf and slope assemblages) of the Blind Fiord Formation.

Exposures on Ellesmere and Axel Heiberg islands (Fig. 1) offer an opportunity to study Lower Triassic sandstone and shale sequences on various scales in a mountainous and nearly unvegetated Arctic environment. Aside from improving our knowledge regarding the geological history of Sverdrup Basin, working with excellent exposures (such as laterally extensive cliffs) allows more complete observations and (perhaps) more reliable interpretations than those based on poorly exposed sediments or drill cores.

Spatial variations of facies within the various paleoenvironmental assemblages of the Bjorne Formation will be of interest to petroleum reservoir geologists (e.g., Miall, 1988); particularly because the fluvial deposits of this formation are similar in age and style to those of the Ivishak Formation of northern Alaska, which forms most of the largest oil reservoir in North America, at Prudhoe Bay (Jones and Speers, 1976). Both the Ivishak and Bjorne formations were likely deposited in delta and coastal plain environments along the northern margin of North America during the Early Triassic (see Embry, 1989, Figures 1, 11).

Methods and exposure

This preliminary account is based on reconnaissance work done in 1985 and more detailed sedimentological studies performed during the 1988, 1989, and 1990 field seasons. More than 50 stratigraphic cross-sections of the Bjorne and Blind Fiord formations have been examined at the 29 localities shown in Figure 1. Most of the localities provide only partial sections of the formations.

The emphasis in this study is on paleoenvironmental analysis. The level of sedimentological observation has generally been between that of reconnaissance and bed-by-bed detail. Quality of exposure varies from excellent to average, from spectacular sea cliffs with sandstone ridges and buttresses, to rubbly hillsides and plateaux with few outcrop ledges. Reconstruction of outcrop-scale facies architecture — the three-dimensional geometry of surfaces, beds, and sequences — was often not possible, but gullied cliffs allowed localized (up to 100 m) three-dimensional views of some features.

The nature of the exposures limited the amount of paleocurrent data collected. Very few good quality paleocurrent measurements have been made because so many of the directional structures are either in two-dimensional outcrop planes or are inaccessible. Therefore, two types of paleocurrent data have been recorded: quantitative compass measurements of accessible, well exposed features, and qualitative estimates of inaccessible or partly exposed features. The latter are probably accurate to only about +/- 45°; for example, a planar-tabular crossbed with a paleocurrent orientation estimated as being to the northwest lies within the 270-360° quadrant. Such paleocurrent estimates have proven to be both adequate and useful in other studies (e.g., Allen, 1983).

PALEOENVIRONMENTS OF THE BJORNE FORMATION

Following the biostratigraphic work of Tozer, map compilation by Thorsteinsson (1974), and brief reports by Roy (1972) and Moore (1981), reconnaissance-scale descriptive aspects of the Bjorne and Blind Fiord formations (including the naming of six members) were summarized by Embry (1986) and will not be repeated here.

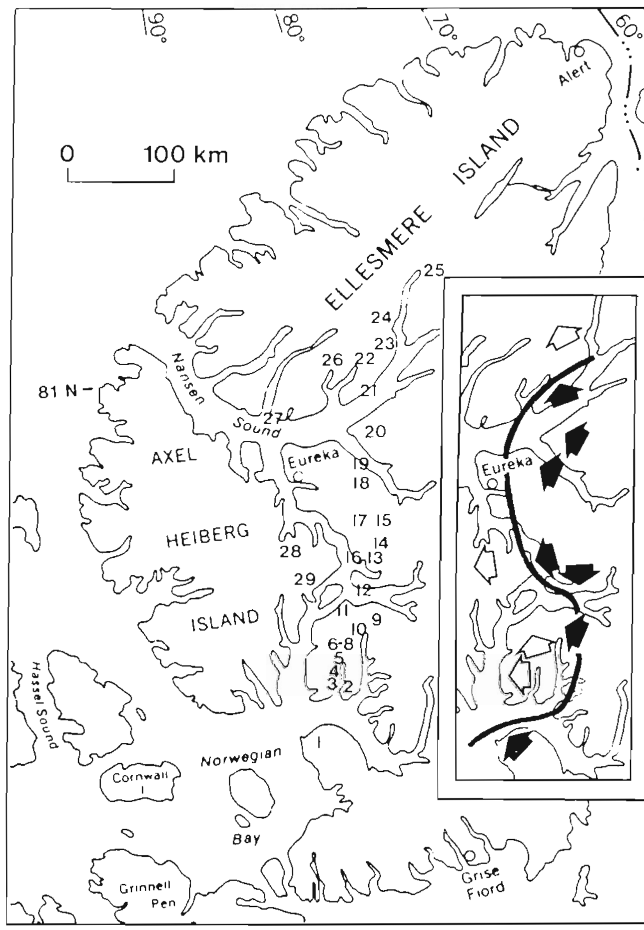

Figure 1. Map showing the locations of the Lower Triassic exposures studied. (See the geological maps of Thorsteinsson, 1974, for details.) Localities are: 1, Bjorne Peninsula; 2-8, Blind Fiord area of Raanes Peninsula; 9, 10, north of Trold Fiord; 11, 12, Bay Fiord; 13, Vesle Fiord; 14, 15, Fosheim Peninsula east of Sawtooth Range; 16-18, Sawtooth Range; 19, Canyon Fiord ("Big Cliff"); 20, Hamilton Peninsula; 21, Greely Fiord; 22, Mt. Leith (Esayoo Bay); 23-25, Tanquary Fiord (24: McKinley Bay); 26, Krieger Mtns.; 27, Blaa Mtn.; 28, Mokka Fiord (Buchanan Lake); 29, north of Whitsunday Bay. Inset at right is a generalized paleogeographic sketch map emphasizing: a deltaic bulge, sandy braided-river paleocurrent trends (black arrows), the approximate position of coastlines during regressive phases (heavy black line), and marine paleocurrent trends (white arrows point distally).

The following synopsis of paleoenvironments is offered in the form of an extended abstract. Each facies assemblage has similar characteristics in all three stratigraphic members of the formation, so the members do not need to be treated separately. Trace fossils are described in more detail in a separate section below. Localities referred to in the text are shown in Figure 1.

Sandy braided-river assemblage

Sandy braided-river deposits constitute significant parts of the stratigraphic sections at Localities 13, 16 to 21, and the Bjorne Formation type section (very poor exposures at Locality 1). This lithofacies assemblage is dominated by crossbedded sandstone and forms highly resistant cliffs, including parts of the best exposures of the Bjorne Formation — sea cliffs up to 600 m high at Localities 19 and 21.

Cross-sets are typically tens of centimetres thick (up to 1 to 3 m thick) with minor associated ripples and plane laminae. Features such as reactivation surfaces, descending tabular cross-sets (Haszeldine, 1983), overturned foresets, and sigmoidal foresets are uncommon. Trough and planar crossbeds cannot always be reliably distinguished in two-dimensional cliff walls. Poorly to well defined fining- and thinning-upward sequences, a few metres thick, are rich in planar cross-sets and have a very high sandstone: shale ratio (Fig. 2A, B), expressed as thin (10 cm) and laterally impersistent shaly sequence caps (wedging out over a few metres). Shale intraclast lag bands are far more common than shale beds. Both the individual cross-sets and sequences are sheetlike in laterally extensive cliffs (Fig. 2C); clearly visible channels and channel-like (concave-up) forms are small, shallow and subtle. Paleocurrents, measured or estimated mostly from planar cross-sets, are locally unimodal and tend to be vertically persistent in direction at each ''Locality''.

The abundance of sheetlike cross-sets, unimodal paleocurrents, and the very small amount of thin shale lenses in the sections indicate deposition by broad, shallow, sandy,

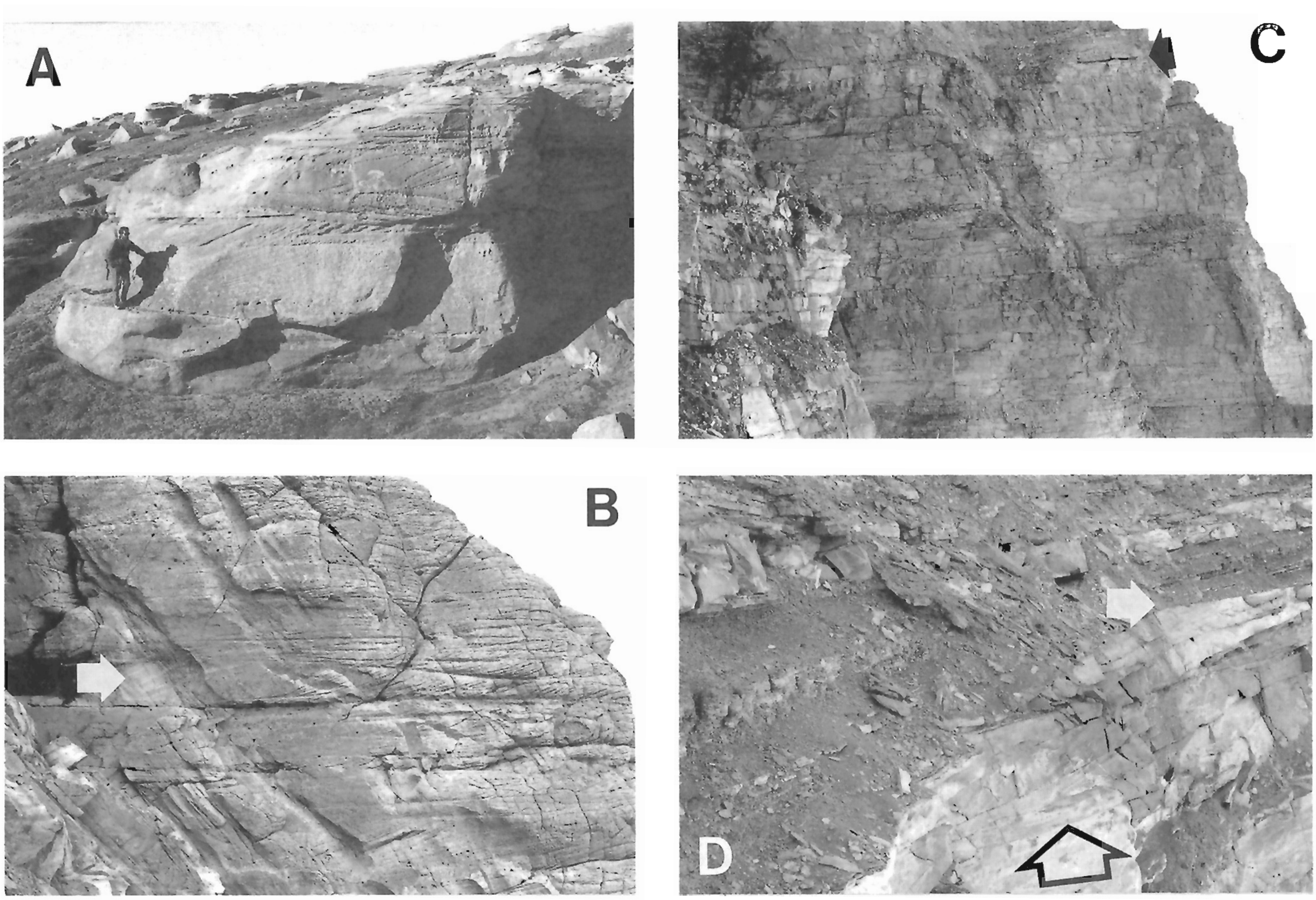

Figure 2. Fluvial deposits in the Bjorne Formation. (A — C are braided-river deposits.)
A. Stacked tabular cross-sets indicating unimodal (to left) paleocurrents.
B. Thin sheet- and wedge-shaped cross-sets — all paleocurrents directions are to the left. Thick cross-set (arrow) shows reactivation surfaces via changes in foreset angles, and, at the far right, a thinning-upward series of wedge-shaped cross-sets suggests vertical aggradation of the upstream flank of a sandbar.
C. Sandstone cliff (about 50 m of section) showing sheetlike sequences or units, with some low-angle truncation surfaces and subtle channel-like forms. Note the near absence of shale interbeds (arrow shows one).
D. Sandstone-shale fining-upward sequences of meandering-river origin. Channel sandstone is trough crossbedded (black arrow), capped by overbank shales with sandstone splay beds; abandoned channel fill to right of white arrow.

braided rivers (Allen, 1983; Collinson, 1986). A fractal network of accreting sandbars and highly mobile, low-sinuosity channels, eroded to shallow depths, produced thick accumulations of thin crossbedded sheets. Muds settled in abandoned bar-top channels, most of which were eroded away during subsequent channel migration, leaving only a few small shale beds and numerous mudclasts. The large volume of sandy braided-river deposits cannot be distilled into one simple local facies model; sequences vary from those rich in planar-tabular cross-sets to others with mixed trough and planar crossbeds, and may be thin (2-4 m common) or thick (5-6 m). Large portions of the sections display few or no orderly sequence patterns, possibly evidence of more random accumulations of eroded bar and channel fragments.

Accessible parts of cliffs allow examination of the geometry (architecture) of beds and erosion surfaces at a variety of scales. At corners, where rock walls meet at near right angles, architectural details can be viewed in three dimensions. Where the paleocurrent directions are locally unimodal and vertically persistent, as is common in these deposits, features such as downstream-inclined accretion surfaces (Miall, 1988) may be identified. At Locality 21, one good example of these features shows mud drapes on such surfaces, indicating incremental downstream bar growth. This particular site offers a view in the "across-paleocurrent" direction. Around a corner in the outcrop, views down-paleocurrent show subtle channel shapes, illustrating how the identification of architectural details can depend on the angle from which they are viewed.

At Locality 20, large plateau areas of flat-lying strata offer a poor stratigraphic section, but display good bedding plane exposures of trough crossbeds. Large blocks (e.g., 6 m high by 15 m long) that have fallen from cliff ledges provide numerous small-scale, three-dimensional views of the stratification within (e.g., sloping bounding surfaces, and wedging out of cross-sets).

At a subregional scale, that of central Ellesmere Island, a radial paleocurrent pattern can be discerned (Fig. 1, inset). This pattern suggests a large delta plain. Most of the paleocurrent data were obtained from the upper part of the Bjorne Formation, but the radial pattern is consistent with other proximal/distal indicators and is thought to be representative of all the major regressive phases of Bjorne deposition.

Meandering-river assemblage

In contrast to the sandy braided-river deposits (above), Bjorne Formation strata deposited by meandering rivers occur as well defined fining-upward sequences, 5 to 13 m thick, which contain more trough crossbeds, have a more equal sandstone : shale ratio, and have thicker (a few metres) and much more laterally persistent shaly sequence caps revealing a variety of thin bedded overbank facies (Fig. 2D). These deposits are exposed at Localities 10, 12, 13, 16 to 19, 21, 22, and 24; a more widespread distribution than that of the braided fluvial strata.

The lower, sandy horizons of the fining-upward sequences are generally sheetlike, thin bedded (plane laminae and ripples) in their upper parts, and may contain inclined stratification (epsilon cross-stratification) and small channels. The lateral tapering of both single story and multistoried sandstone bodies can be seen in some of the better exposures.

The thick, shaly, overbank units provide information not available from the thin and impersistent shales in the braided fluvial assemblage. The thicker shales contain sandstone interbeds (sharp based, laminated or rippled beds; climbing ripples; and one metre thick coarsening-upward units) interpreted as splays (Fig. 2D). Horizontal worm burrows are rare. Evidence of subaerial exposure includes desiccation cracks, paleosols, and extremely rare vertebrate tracks (Fig. 4A). Paleosols hosted in sandstone are uncommon and are indicated by massive horizons with rhizoliths (root casts) and a sparse amount of small (1 cm) carbonate nodules; the latter suggesting the poorly developed caliches of a semiarid paleoclimate. Transported and scattered plant debris is uncommon and is the only coaly organic material in the Bjorne Formation; no coal beds have been observed.

Although the standard meandering-river facies model is currently thought to be overly simplistic (Collinson, 1986; Miall, 1988), it explains well the nonmarine fining-upward sandstone/shale sequences in the Bjorne Formation.

The braided versus meandering distinction is also thought to be simplistic by some fluvial specialists, but it is the fundamental natural division of nonmarine strata in the Bjorne Formation and is very important paleogeographically. Both cross-sectional views of outcrops, and the wider regional distribution of meandering-river strata relative to braided-river deposits, suggest that the meandering facies are normally more distal than the braided-river assemblage.

Marginal marine assemblage

The stratigraphic sections at Localities 10 to 12, 21, 22, and 24 are relatively rich in sandstone and shale deposited in marginal marine settings. On a small (outcrop) scale, the influence of waves is generally more obvious than that of tides or fluvial input. Trace fossils of the *Skolithos* ichnofacies (Ekdale et al., 1984) are characteristic.

Foreshore deposits consist of horizontally to low-angle laminated sandstone (swash layers, with some convergence and truncation of laminae) with a low degree of bioturbation (*Skolithos, Diplocraterion*; Fig. 3A). Within these laminated beds, rare examples of ripples (beach runnel fills), scours, and regularly spaced clusters of granules (antidune deposits ?) suggest the rare preservation of minor beach features.

Trough crossbeds (upper shoreface dunes), rippled and bioturbated sandstone — including sharp-based storm beds (more thinly layered lower shoreface strata) and more shaly units with sandy interbeds (laminated storm beds and ripples, representing the transition zone between the shoreface and inner shelf/prodelta) — are most commonly arranged in coarsening- and thickening-upward parasequences (cf. Van Wagoner et al., 1990). These parasequences are several metres or more thick, products of upward shallowing during sandy shoreface (or stream mouth bar) progradation. Coarsening is mostly via loss of shale rather than an upward increase in sand grain size (Bjorne Formation sandstone is

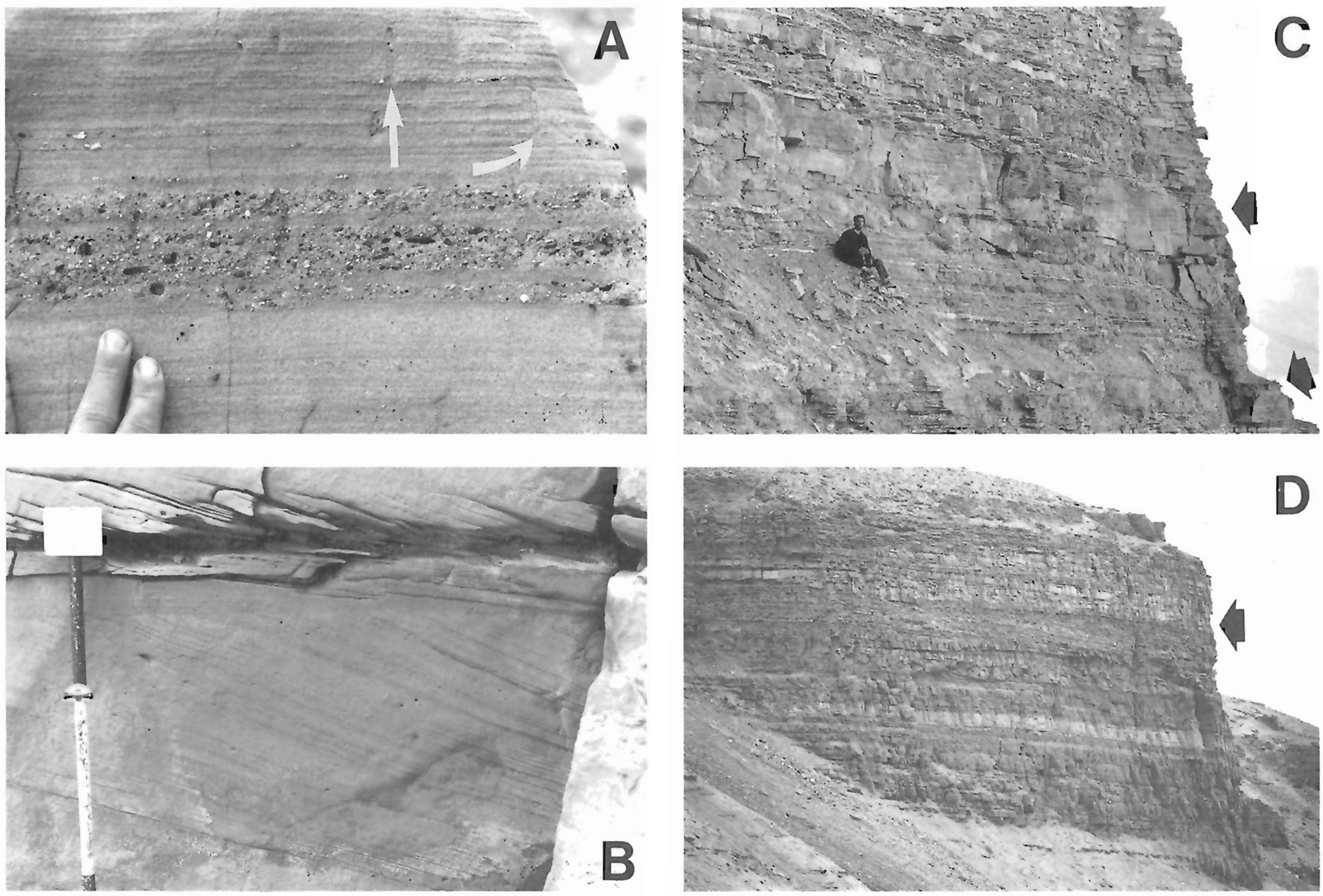

Figure 3. Marginal marine deposits in the Bjorne Formation.
A. Horizontally laminated sandstone with well segregated pebble beds and a few *Skolithos* burrows (arrows); a beach deposit.
B. Cross-set with regularly spaced mud-draped (dark) foresets (tidal bundles, likely ebb oriented). Sigmoidal foresets at top of staff.
C. Coarsening- and thickening-upward parasequence (between arrows). Shaly base, overlain by thin bedded sandstone (storm beds, ripples) and shale, ball-and-pillow horizon (by man's head), and crossbedded sandstone cap.
D. Sandstone cliff, 45 m thick, with internally crossbedded channel deposits (lighter-toned units) more obvious in the upper half (see text). Some channels are offlapping to the left (arrow).

generally fine grained and well sorted), and the thickening trends are displayed more in the lower parts of the parasequences.

Deposits of unequivocal tidal origin are uncommon. At Localities 10 to 12, 21, and 24, units a few metres thick contain one or more of the following: herringbone (bimodal-bipolar) crossbedding, tidal bundles (regularly spaced muddy foreset laminae; Fig. 3B), sigmoidal foresets, flaser and wavy bedding (mud-draped ripples), and thinning-upward sequences (shallowing of a channel or sandbar). In the herringbone cross-strata, the onshore-directed component is opposite to the locally dominant paleocurrent direction of fluvial input; for example, herringbone paleocurrent patterns are north-south at Localities 12 and 21, with the fluvial input direction to the south at Locality 12 and to the north at Locality 21.

At Locality 24, a progradational parasequence set (term of Van Wagoner et al., 1990) outcrops as a cliff band, 50 m thick. The best defined sandy parasequences are about 4 m thick (for example, see Figure 3C), with large ball-and-pillow structures commonly localized at the bases of thicker bedded units. Parasequences become less defined higher in the section as the sandstone: shale ratio increases and channels become more prevalent (Fig. 3D). Such channels are typically up to 2 m deep and 30 m wide, and are most easily recognized in outcrop views to the north and northwest, which are likely down-channel, down-paleocurrent views; the dominant paleocurrent direction in the area is to the north and northwest (few measurements at this cliff dominated "Locality", mostly estimates from ripples).

The parasequences are thought to record settings ranging from inner shelf (or transition zone) up to barred shoreface. As progradation continued, sands of shallower subenvironments constituted an increasingly greater percentage of individual parasequences, the section becoming sandier overall. The upper part of the parasequence set is interpreted as being rich in marginal marine facies, with the small but broad and shallow channels representing inter-bar rip current channels incised into lower shoreface strata. Such channels would have funnelled storm sands toward the shelf (Aigner, 1985), and help to account for the abundant event (storm) beds in inner shelf to lower shoreface deposits in the vicinity. Note also that the presumed orientation of the paleoshoreline in this area was approximately northeast-southwest (Maurel, 1989, Fig. 3), so that the channels would have been oriented approximately perpendicular to the paleoshoreline and the dominant paleocurrent trend would have been an offshore-directed one.

PALEOENVIRONMENTS OF THE BLIND FIORD FORMATION

Sections of the Blind Fiord Formation, totalling about 800 m thick, have been examined at Localities 2 to 8. These storm-dominated shelf deposits have been briefly summarized elsewhere (Devaney and Embry, 1990). Distinguishing tempestites from turbidites is often not possible in the shale-rich units in this area, but is less of a problem in the sandier and more bioturbated (more proximal) units. Regional stratigraphic context and a comparison with the different facies assemblage at Localities 27 to 29 — where monotonous sections of shale with unbioturbated event beds (interpreted as turbidites) are interpreted as slope and/or submarine fan deposits — suggest that many of the event beds at Localities 2 to 8 are indeed shelfal storm beds.

Compared to Localities 2 to 8, storm-dominated shelf deposits on northern Ellesmere Island (Localities 22, 24, 26, 27) suggest more proximal shelf settings; the sections are generally sandier, hummocky cross-stratification is more common, thicker and coarser shell beds are present (as both tempestite layers and minor bioclastic limestone beds) and *Diplocraterion/Rhizocorallium* ichnofossils are more steeply oblique to bedding. Moore (1981) briefly described exposures in the area between Localities 24 and 27 and offered a similar interpretation.

TRACE FOSSILS

Most of the trace fossils in the Bjorne and Blind Fiord formations are common shallow water, marine types (Ekdale et al., 1984) and were identified in the field at the ichnogenus level.

Nonmarine trace fossils are very rare. The few examples of horizontal worm burrows, some with meniscate backfill structures, have been found in the overbank facies of meandering-river deposits. Desiccation cracks crosscut some of the burrows.

Vertebrate tracks have been found on two bedding planes in the lower Bjorne Formation (Cape Butler Member) at Locality 22. These are the first vertebrate tracks reported from Sverdrup Basin. Figure 4A shows parts of two trackways of tridigitate undertracks on the bottom of a sandstone slab. About 50 m higher in the same outcrop section, convex hyporeliefs of curved digit or claw marks suggest forward locomotion and a grasping motion indicative of stiff mud.

Well defined vertical burrows predominate in the sandy marginal marine deposits. *Skolithos* (Fig. 4B), *Diplocraterion, Arenicolites*, and horizontal worm burrows (*Planolites/ Paleophycus*) are common. Three types of *Skolithos*, usually mutually exclusive, are found: i) thin (2-3 mm wide) vertical shafts; ii) steeply inclined shafts, varying from thin to fat (1 cm wide); and iii) 45° diagonal shafts. *Diplocraterion* is usually protrusive and oriented vertically, less commonly diagonally. *Skolithos* and *Diplocraterion* can be tens of centimetres deep. Most examples of *Arenicolites* are large, about 10 cm deep and wide. Less common burrows in the coastal facies include *Aulichnites, Rosselia* (Fig. 4C), escape burrows (Fig. 4D), and rare *Cylindrichnus* and *Monocraterion*.

Trace fossils in shelf deposits have been identified mostly from sandstones, not the interbedded shales. Vague mottling of unknown classification, probably the result of intense bioturbation by horizontal burrowers, is far more common than the ichnogenera listed below. Horizontal worm burrows (*Planolites, Paleophycus*) are the most common trace fossil in the shelf assemblage. Horizontal to low-angle *Rhizocorallium* is more abundant in the upper Blind Fiord Formation (Svartfjeld Member) than in the older strata. Good examples of *Teichichnus* burrows are uncommon. *Thalassinoides* is very rare. In the upper parts of event (storm) beds, *Skolithos* is common, plus some *Catenichnus* (McCarthy, 1979) and escape burrows.

Thus, the shelf strata vary from a *Cruziana* ichnofacies (Ekdale et al., 1984) to the more proximal mixed *Skolithos-Cruziana* ichnofacies of a storm-dominated shelf.

Three specific ichnogenera also illustrate proximal-to-distal trends. In shoreface sandstones, abundant shafts of *Skolithos* are small to large (wide or deep), versus inner shelf sandstones in which the *Skolithos* pipes are small and preferentially located in the upper parts of some event beds. *Diplocraterion* is usually vertical in marginal marine settings, compared to the inclined forms in sand-rich inner shelf sections. The latter are in turn gradational to low-angle to horizontal *Rhizocorallium* in sandstone beds in shaly, distal shelf strata, illustrating a distal flattening of U-shaped spreite-bearing burrows (cf. Fursich, 1975).

STRATIGRAPHY AND PALEOGEOGRAPHY

The Lower Triassic succession of Sverdrup Basin, consisting of the proximal Bjorne Formation (fluvial, marginal marine) and distal Blind Fiord Formation (shelf and slope/submarine fan facies), has been subdivided into three transgression-regression (T-R) cycles by Embry (1986). The general nature of these T-R cycles, or third-order depositional sequences, has been described by Embry (1988), Embry and Podruski (1988), and Galloway (1989), and differs from the use of the term "sequence" by those of the Exxon school (Van Wagoner et al., 1990).

The three Bjorne/Blind Fiord T-R cycles or sequences are each hundreds of metres thick and have been given stratigraphic member status comprising one or two members each (Embry, 1986). Because the regressive deposits are very much thicker than the transgressive, the former dominate the stratigraphic section at any single locality. Very similar paleoenvironmental assemblages are repeated vertically throughout the formations; for example, sandy braided-river deposits in one member are in most respects identical to those in other members. This vertical persistence of the dominantly regressive facies leads to paleogeographic generalizations that are biased toward the most regressive facies assemblages, which

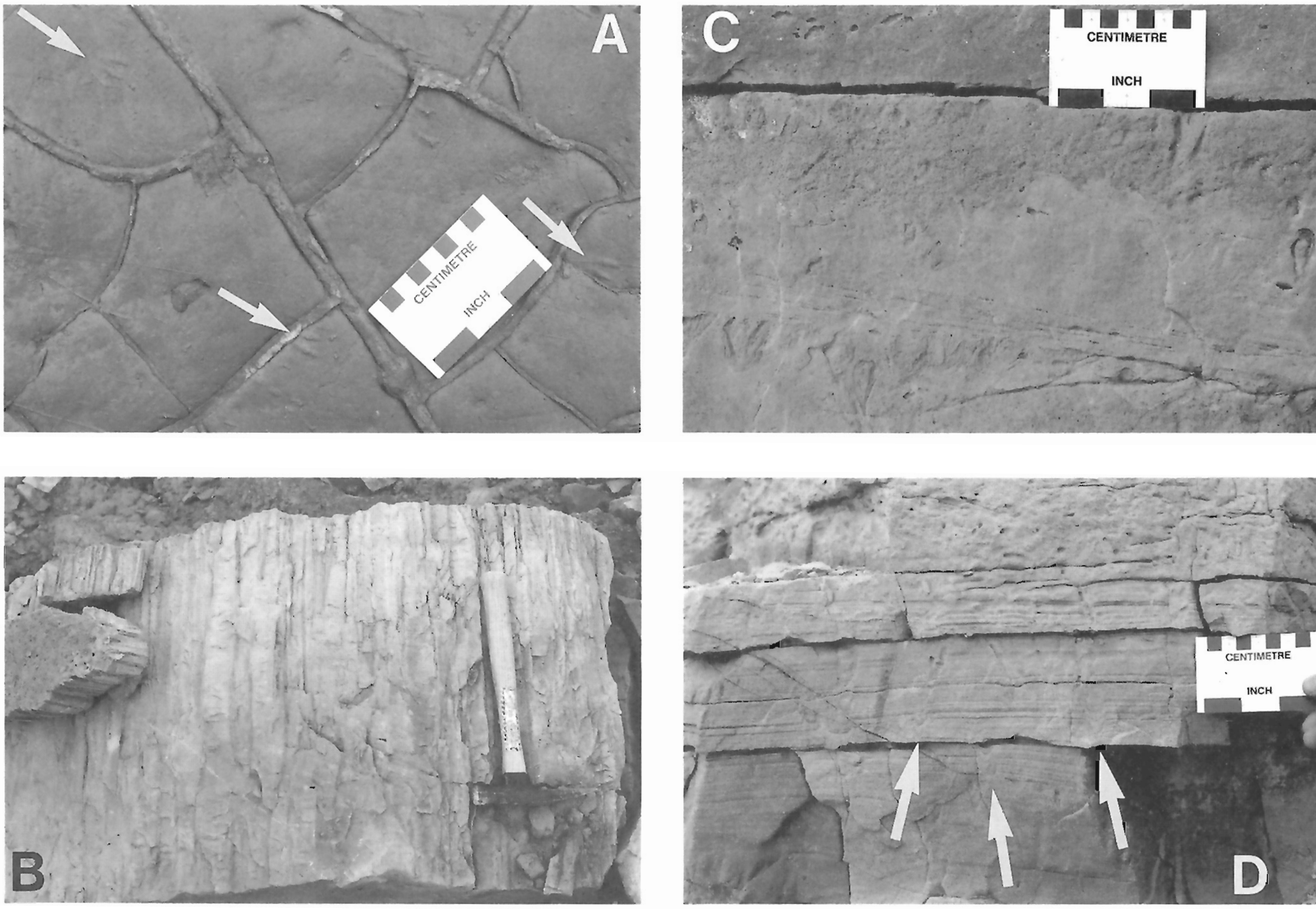

Figure 4. Trace fossils in the Bjorne Formation. (B-D are in shoreface deposits.)
A. Underside of a sandstone slab with vertebrate tracks (arrows) crosscut by desiccation cracks; meandering-river overbank facies.
B. *Skolithos* shafts in cross-section (small fragment at far left shows bedding plane and cross-section views).
C. *Rosselia socialis* burrows. Note the differences in bioturbation intensity in different layers (none/low/high).
D. Escape burrows (arrows) in planar laminated sandstone (storm bed). Note the upward increase in bioturbation.

occur high in the T-R cycles and usually constitute a large part of the cycle thickness. For example, strata of all three members of the Bjorne Formation at Localities 16 to 21 are rich in fluvial deposits, in contrast to those at Localities 10, 12, 22 and 24, which are rich in marginal marine and shelf facies and contain only minor amounts of fluvial deposits (Fig. 1). This pattern continues and the Bjorne Formation is distally transitional to the more shaly shelf deposits of the Blind Fiord Formation (maps of Thorsteinsson, 1974; Embry, 1986, Fig. 36.1) at Localities 2 to 8, 26 and 27 and to the most distal (deepest water) slope and/or submarine fan deposits at Localities 27 to 29.

The vertical persistence of the paleoenvironmental assemblages, paleocurrent data (Fig. 1, inset), and subregional facies changes, outward to more distal, finer grained and deeper water deposits, reflect a paleoenvironmental gradient away from a fixed deltaic input centre. During the deposition of each of the three T-R cycles, a major river system that probably drained northern Greenland supplied sand and mud, but almost no gravel, to a deltaic complex located in the central Ellesmere Island area. The radius of these delta systems was about 100 km (Fig. 1, inset). Paleocurrents from sandy braided-river deposits (which tend to be locally unimodal) outline the radial drainage pattern of the ancient delta plains that were repeatedly depositional sites along the east margin of Sverdrup Basin (at Localities 13 to 21) during the regressive phases of the Early Triassic. Roy (1972) noted that paleocurrents in the Bjorne Formation on Ellesmere Island were generally to the west, but grouped the data from all his localities together and missed the strong radial component that is present. A similar radial paleocurrent pattern has been found in the Bjorne Formation on Melville Island, in western Sverdrup Basin (Agterberg et al., 1967).

Poorly exposed fluvial deposits at Locality 1 represent different coastal plain or delta systems that were derived from the south to southeast (Fig. 1, inset). The paleocurrent directions at Locality 9, a section containing fine, gravelly, braided-river conglomerate (a 9 m thick unit) and pebbly sandstone, are very different (to the west and northwest) from those of the sandy braided-river deposits to the north and northwest (at Localities 13, and 16). This is indicative of a relatively small, local, fluvial basin near Locality 9, perhaps a separate inter-deltaic drainage system.

The large sediment bulge that was present along the eastern Sverdrup Basin margin during the Early Triassic shows that fluvial input repeatedly outpaced the effects of waves and tides. Tidalites are rare, compared to the prograded shoreface parasequences and well developed beach deposits locally common at paleogeographically appropriate localities (e.g., Localities 12 and 21; heavy line in Figure 1 inset), suggesting that the shallow marine facies assemblages record a fluvial/wave interaction type of delta margin. Stratification in the more distal prodeltaic shelf, perhaps an inter-deltaic embayment at Localities 2 to 8 (Fig. 1), was produced predominantly by background suspension muds alternating with storm sand beds derived from the delta front slopes.

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Study of the Triangle Zone and Foothills structures near Pincher Creek, Alberta

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Abstract

Foothills structures and stratigraphy in the Pincher Creek and Beaver Mines areas of southwestern Alberta were mapped at a scale of 1:20 000 during the summer of 1990. West dipping beds are cut by thrusts that repeat Cretaceous stratigraphy at the surface. A series of structures, previously mapped as vertical transverse faults, occurs southwest of the town of Pincher Creek. Our fieldwork indicates that these transverse features can be interpreted as several duplexes occurring within a single major thrust sheet. Because their dips are oblique to dominant trends, both roof and floor thrusts of the duplexes are represented at the surface. A local interpretation of the Triangle Zone at the leading edge of the Foothills belt is given. The relationship between geometries of the Triangle Zone, the duplexes, and their proximity to the Crowsnest Deflection is currently being investigated.

Résumé

Les structures et la stratigraphie des Foothills dans les régions de Pincher Creek et de Beaver Mines dans le sud-ouest de l'Alberta ont été cartographiées à l'échelle de 1:20 000 au cours de l'été 1990. Les couches à pendage vers l'ouest sont recoupées par des failles chevauchantes qui répètent la stratigraphie du Crétacé en surface. Une série de structures, auparavant cartographiées en tant que failles transversales, se trouvent au sud-ouest de la ville de Pincher Creek. Les travaux sur le terrain des auteurs indiquent que ces failles transversales peuvent être interprétées comme plusieurs duplex occupant une même grande nappe de charriage. Comme leur pendage est oblique par rapport aux directions dominantes, les chevauchements inférieur et supérieur des duplex sont observables à la surface. Une interprétation locale de la zone Triangle à la charnière frontale des Foothills est présentée. La relation entre la géométrie de la zone Triangle, les duplex et leur proximité de la Déflexion de Crowsnest est présentement à l'étude.

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INTRODUCTION

The study area, depicted in Figure 1, is located in the "Crowsnest Deflection" (Dahlstrom, 1970), where the fold and thrust belt of the southern Canadian Cordillera bends sharply to the southeast. The area extends north of the Waterton area, where Gordy et al. (1977) first described the Triangle Zone, and east of the Lewis Thrust trace to the town of Pincher Creek. Deformation in the study area is dominated by closely spaced, northeast verging, low-angle thrust faults that repeat the Jurassic and Cretaceous Kootenay through Belly River formations (Fig. 1). In addition to the Lewis Thrust, the following four major faults (see Price, 1962) have been labelled on Figure 1: the Tetley Fault, the Mill Creek Thrust, the Livingstone/Harland Lakes Thrust, and the Twin Butte Thrust. A series of transverse features, which our fieldwork indicates can be interpreted as several oblique duplexes, occur near Fish Lake, in the footwall of the Lewis Thrust (Fig. 1).

This particular area was chosen for study of the Triangle Zone and associated Foothills structures because, at this locality, the dramatic change in regional structural trend accentuates the en echelon, oblique, and transverse structures, so that they are more readily recognized here than elsewhere along strike in Alberta. The area was last mapped by Hage (1941, 1943, 1945) and Douglas (1951) at a scale of 1:63 360, with portions mapped at 1:31 680, and is included in Price's (1962) 1:126 720 scale compilation. Mapping on a scale of 1:20 000 was required in this study to delineate the duplex structures and the areas of reversal in vergence in the Triangle Zone. Careful delineation of these zones is critical to the construction of palinspastic restorations, balanced cross-sections, and models of the three-dimensional sequential development of the observed structures, which we will prepare in 1991.

STRATIGRAPHY

The following is a brief summary of significant new field observations to date. A corresponding stratigraphic column, with brief descriptions of all formations exposed in the study area, is shown in Figure 2.

Since the Kootenay Group always occurs in the immediate hanging wall of thrusts, no complete section of this formation is exposed anywhere in the study area. However, a section thicker (approximately 30 m thick) than any described by previous workers in the area has been identified northwest of the town of Beaver Mines. The formation thins across the area; it does not outcrop and presumably thins to its zero edge east of Highway 6 (Fig. 1) in the subsurface.

The Gladstone Formation, the basal formation of the Blairmore Group, is remarkably consistent in lithology across the area. The lower, coarse grained quartzose sandstone unit, 15 to 20 m thick, lacks any evidence of the basal chert-pebble conglomerate commonly referred to elsewhere as the Cadomin Conglomerate. The uppermost unit (50 cm) of the Gladstone Formation is laterally consistent, composed of a thin limestone bed overlain by a calcareous unit that hosts a fauna of various bivalves and gastropods. The upper part of the Blairmore Group is quite consistent in lithology across the area. An exception is the presence of an igneous-pebble conglomerate whose thickness and stratigraphic position vary considerably along strike and from one thrust sheet to the next. The conglomerate is 30 m thick at one locality west of Beaver Mines, but has not been observed east of Mill Creek.

The Crowsnest Formation changes dramatically across the map area. Near the western boundary, the unit is 100 m thick and contains evidence of its flow and volcaniclastic origins. Agglomerate clasts include trachyte, some hosting alkali feldspar crystals larger than two centimetres. Eastward, and with each successive thrust sheet away from its source, the unit increases in ash content while decreasing in grain size and thickness. The easternmost outcrop, along Pincher Creek, consists of 6 to 7 m of poorly consolidated ash beds.

A sharp contact marks the boundary between the Crowsnest Formation and the overlying Alberta Group, which consists of the Blackstone, Cardium, and Wapiabi formations. The basal Blackstone Formation consists of a widespread chert-pebble conglomerate, up to two metres thick at some localities.

The Bearpaw Formation is only exposed at one location along the Castle River. It was found to be highly deformed, hosting various fauna including part of a vertebrate bone approximately 40 cm in length.

The St. Mary River Formation, exposed only in the northeast part of the study area, contains a 70 cm thick coal bed overlain by a one metre thick oyster bed; together they serve as an excellent marker approximately 100 m above the base of the formation (Fig. 2).

STRUCTURE

Two previous interpretations of structural style in the area need to be re-evaluated. The first concerns the geometry of the leading edge of the deformed belt. Published maps of the region, including those for the Beaver Mines (Hage, 1941), Cowley (Hage, 1945), and Pincher Creek (Douglas, 1951) map areas, all depict a simple anticlinal structure adjacent to the undisturbed Interior Plains. This is a characteristic surface expression of the Triangle Zone. A preliminary geometric interpretation of the Triangle Zone, based on surface mapping and data from four wells in the study area, is included in Figure 3. Recent work elsewhere in the Foothills belt, by Charlesworth et al. (1985, 1987), Jones (1982, 1989) and others, suggests that the Triangle Zone can be simply modelled as either an intercutaneous wedge or a duplex. We plan to combine our field data with well information and seismic data not available to previous workers, and analyse them to determine whether the geometry of the study area complies with these models or not.

The second structural style that needs to be re-evaluated is the nature of faults with trends that are oblique to the strike of bedding. Hage (1941) and Douglas (1951) interpreted the area as being sporadically cut by minor vertical transverse faults, with a significant concentration occurring between Pincher and Drywood creeks. Their maps depict north-trending transverse faults truncating northwest-trending beds, but we have mapped the faults as duplexes in Figure 1.

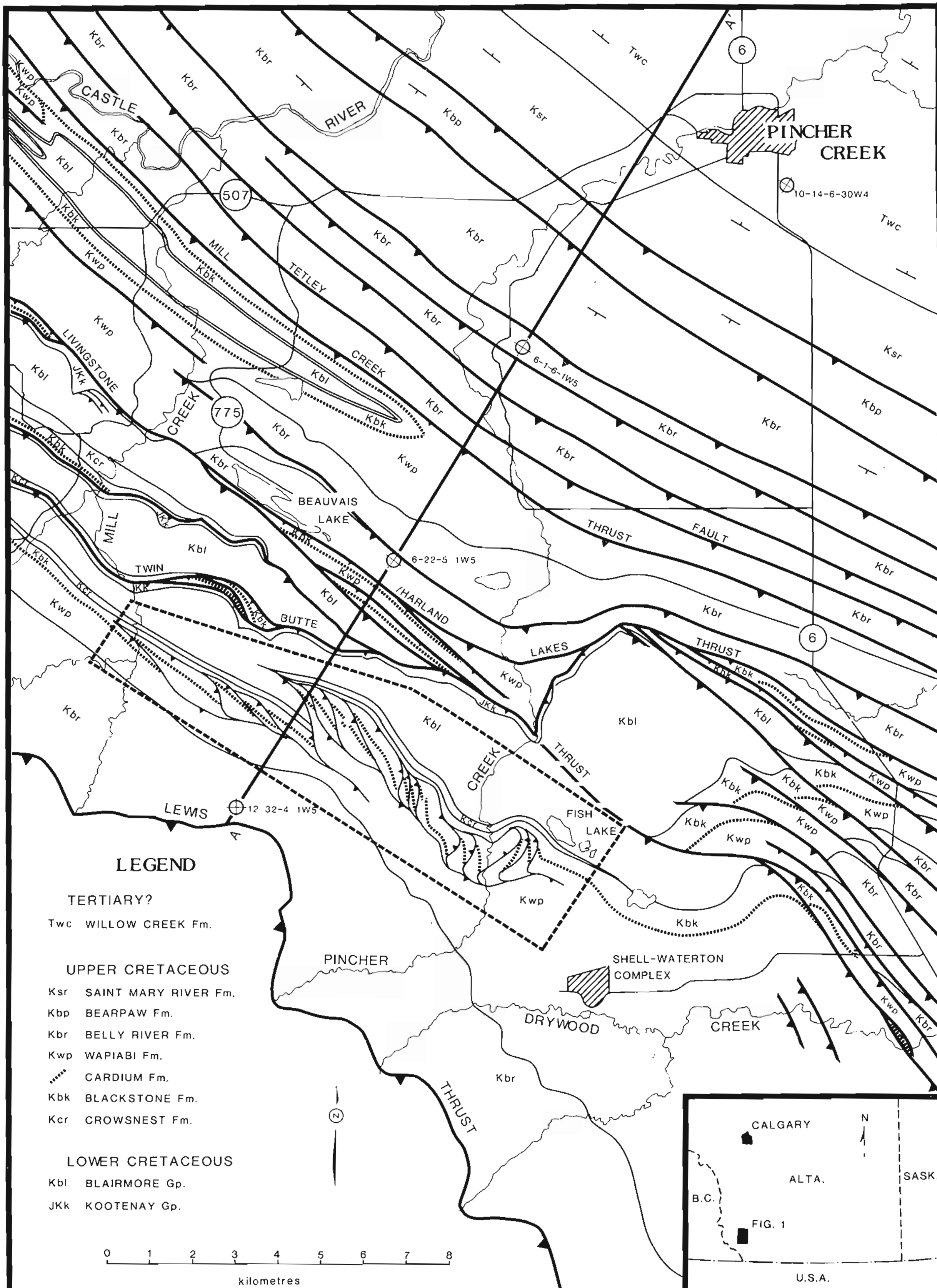

Figure 1. 1:125 000 scale geological map of the study area. The dashed line encloses the oblique duplexes discussed in the text. Line of cross-section (Fig. 3) is labelled A-A'.

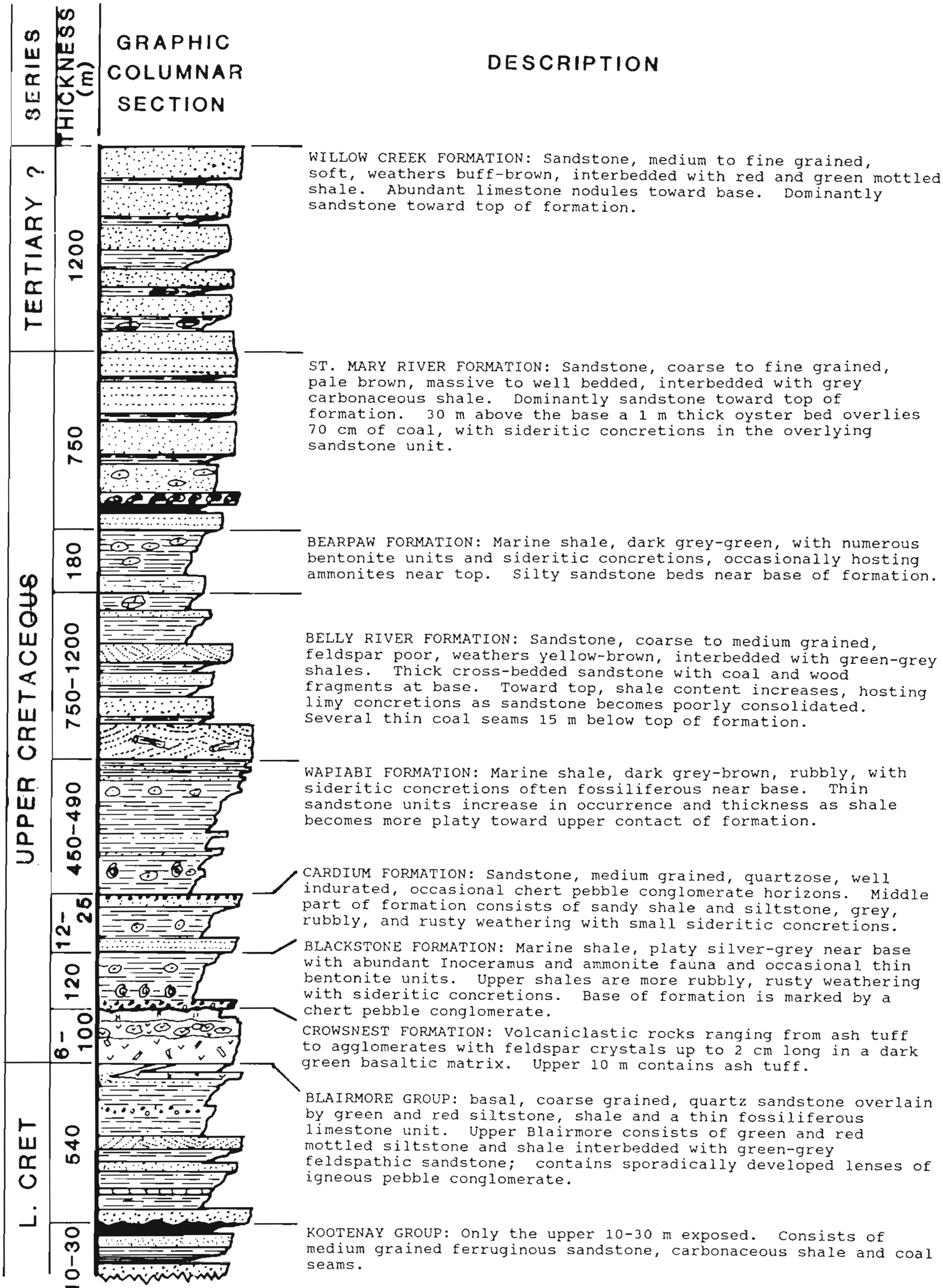

Figure 2. Stratigraphic section for the study area (based on our fieldwork). The column is not to scale. The thicknesses given were measured in the field or calculated from representative map intervals.

Although transverse faults have been mapped elsewhere in the Foothills belt (refer to work in the Bragg Creek area, 12-22-6W5, by Hume and Beach, 1941), their occurrence is rare and commonly associated with some later structural event (refer to Ollerenshaw's 1976 work at Jumpingpound Creek, 25-6W5); they are, therefore, anomalous with respect to the general structural style of the region. Horizontal separation across the faults, as originally mapped by Hage (1941) and Douglas (1951), ranges from 100 m to 1200 m with a predominantly sinistral offset. These faults may be classified into two groups. The first group has been mapped with transverse displacement of the Kootenay/Gladstone contact, individual faults occurring sporadically across the study area. The second group, forming a tight cluster of structures, appears to offset formation contacts of the Alberta Group only. The upper sandstone of the Cardium Formation serves as the only mappable marker. The atypical nature of these structures prompts us to ask several questions. If the transverse faults illustrated by Hage (1941) and Douglas (1951) exist, why are they restricted to certain stratigraphic intervals? Is it merely due to the lack of mappable markers in the over- and underlying formations? How do these structures relate to the larger thrust sheet in which they occur and to the major deflection of structural trend in the region?

DISCUSSION OF FIELD OBSERVATIONS

Our recent field observations suggest that, of the seven transverse faults that cut the Kootenay/Gladstone contact, as mapped by Hage (1941), three are merely lithostratigraphic inaccuracies. For example, a sandstone unit within the upper part of the Kootenay Group has been misrepresented as the younger Gladstone Formation and hence a transverse fault has been mapped (refer to Hage's 1941 Beaver Mines Map 739A, 10-1-5W5). At one locality, Hage (1941) mapped a transverse fault where a tight fold hinge is well exposed. The nature of the remaining four faults of this group is yet to be determined. They may be true transverse structures that cut across the Kootenay/Gladstone contact and then either die out or join previously unrecognized bedding-parallel faults in the over- and underlying formations.

The second, tighter, group of transverse faults mapped by Hage (1941) and Douglas (1951) is marked in the field almost exclusively by offset of the upper sandstone of the Cardium Formation. Apart from this marker, evidence of the lateral extent of displacement is conjectural. Our field observations concur with much of Hage's (1941) mapping, except that the thrust slices are more numerous and more closely spaced than previously mapped. Hage (1941) also had a tendency to link outcrops into series of folds even in areas where all beds are parallel and no hinges or reversals in "way-up" directions are observed. We have interpreted these structures as duplexes with horses dipping obliquely relative to the surface topography, so that, as shown in Figure 1, both the roof and floor thrusts are represented at the surface. A smaller version of this duplex geometry, developed within the Wapiabi Formation, is approximately 3 m high by 6 m long and is completely exposed on both sides of a meander in Mill Creek in the footwall of the Livingstone Thrust; the roof and floor thrusts and over a dozen horses can be traced out.

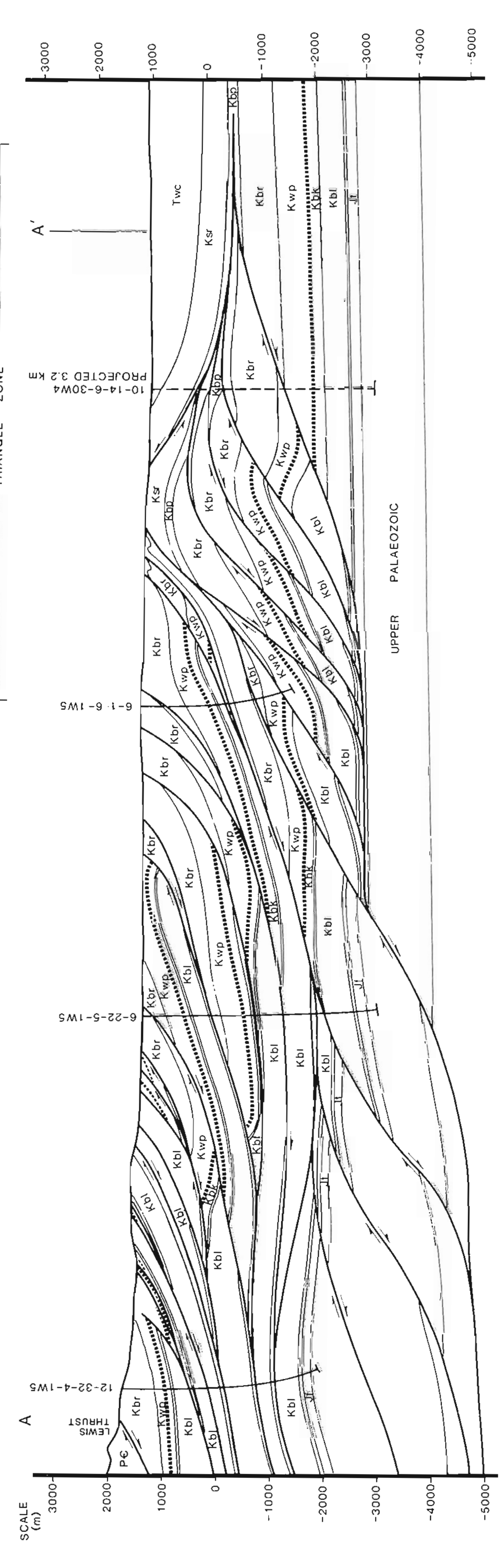

Figure 3. Geological cross-section through the study area (1:125 000 scale). Location of section line (A—A') is given in Figure 1. Section is based on surface mapping by S.N. Hiebert and data from four wells (labelled). Symbols for rock units as in Figure 1 legend. PC — PreCambrian rocks in the hanging wall of the Lewis thrust.

PLANS FOR FUTURE WORK

The recognition of duplexes and other structures that are not continuous along strike, along with the compilation of more recent well and seismic data, will constrain and aid us during 1991 in the construction of palinspastic restorations, balanced cross-sections, and sequential reconstructions of the development of the Foothills and Triangle Zone structures in the Pincher Creek and Beaver Mines areas.

Shortening across the Triangle Zone, depicted in Figure 3 and by Price (1962, 1981), suggests that this portion of the Triangle Zone is not as internally deformed as portions described by Charlesworth et al. (1985, 1987), Jones (1982, 1989), MacKay (1990), and Spratt and Lawton (1990). Hence, the Pincher Creek area may represent an early stage in the development of triangle zones; it will form an element in the comprehensive model of forward and longitudinal development of triangle zones that we are generating.

ACKNOWLEDGMENTS

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Representation and statistical analysis of directional sedimentary structures in the uppermost Cretaceous-Paleocene of the Alberta Foreland Basin

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Jerzykiewicz, T. and Labonté, M., Representation and statistical analysis of directional sedimentary structures in the uppermost Cretaceous-Paleocene of the Alberta Foreland Basin; in Current Research, Part B, Geological Survey of Canada, Paper 91-1B, p. 47-49, 1991.

Abstract

Over 800 measurements of the directional sedimentary structures in channels of the post-Wapiabi fluvial sediments have been analysed. Rose diagrams, plotted separately for the five stratigraphic sequences of the uppermost Cretaceous-Paleocene, are applied to show changes in drainage systems through time. The circular means were superimposed onto each rose diagram as vectors showing the preferred flow directions of the fluvial channels. Preferred directions of the channels are different in each of the five stratigraphic sequences analysed. Sequences III and V are characterized by unimodal downslope-directed azimuthal paleocurrent patterns, indicating northeasterly drainage systems (perpendicular to the mountain front). Paleocurrent patterns for Sequences I, II, and IV include bimodal and polymodal distributions, indicating much more diversified paleodrainage systems with a predominant southeasterly direction (parallel to the mountain front). Changing directions of the paleodrainage systems are interpreted in terms of response to tectonics in the Alberta Foreland Basin thrust belt.

Résumé

Plus de 800 mesures de structures sédimentaires directionnelles relevées dans des chenaux des sédiments fluviatiles plus récents que ceux de la Formation de Wapiabi ont été analysées. Des diagrammes en rosette, construits pour chacune des cinq séquences stratigraphiques du Crétacé terminal-Paléocène, sont utilisés pour montrer l'évolution temporelle des systèmes de drainage. Les moyennes circulaires ont été superposées sur chaque diagramme sous forme de vecteurs, indiquant les directions préférentielles d'écoulement dans les chenaux fluviaux. Les directions préférentielles des chenaux sont différentes dans chacune des cinq séquences stratigraphiques analysées. Les séquences III et V sont caractérisées par des réseaux de paléocourants unimodaux dirigés vers le bas des pentes, révélant des systèmes de drainage d'orientation nord-est (perpendiculaires au front des montagnes). Les réseaux de paléocourants dans les séquences I, II et IV présentent des distributions bimodales et polymodales, révélant des systèmes de paléodrainage beaucoup plus diversifiés de direction générale sud-est (parallèles au front des montagnes). Les changements de direction des systèmes de paléodrainage sont interprétés comme une réponse à l'activité tectonique de la zone de chevauchement du bassin de l'avant-pays de l'Alberta.

INTRODUCTION

Paleocurrent readings from ancient fluviatile sediments are particularly useful in basin analysis, because often they can be directly related to the depositional strike. Although significant variance of fluviatile paleocurrents may be common in some fluvial systems (especially meandering), the directions of the main fluvial channels tend to be unimodal. The paleocurrent patterns described in most studies of ancient fluviatile deposits are largely unimodal, downslope-directed, azimuthal patterns (Potter and Pettijohn, 1977). Paleocurrents in fluvial and deltaic channels, if measured over a sufficiently large area, reflect the regional paleoslope (Selley, 1969; Tanner, 1971).

Published data on paleocurrents in the uppermost Cretaceous/Paleocene of the Alberta Foothills are from relatively small and restricted areas, and are limited to a portion of the post-Wapiabi stratigraphic interval (Rahmani and Schmidt, 1975; McLean and Jerzykiewicz, 1978). No paleocurrent maps based on directional sedimentary structures have been published for the entire basin. Existing sketch maps are based on the regional dispersal of framework components and heavy minerals in the post-Wapiabi sandstones (Carrigy, 1971; Rahmani and Lerbekmo, 1975). These sketch maps indicate southeasterly oriented longitudinal transport for the entire basin, which has led to the conclusion that the Alberta Foreland Basin received sediment from the northern part of the Columbian Orogen (Eisbacher, et al., 1974). This view requires some corrections in the light of data presented in this paper on the orientation of fluviatile channels in post-Wapiabi strata.

SCOPE AND METHODS

Orientations of the following sedimentary structures have been measured in surface sections and isolated outcrops of post-Wapiabi strata: 1) large-scale trough and tabular-planar crossbedding, 2) tree trunks and stems in channel lag deposits, 3) groove marks at the bottom of channels, and 4) pebble imbrication. Only well exposed structures have been measured. In the case of trough crossbedding, the most common type of the structure measured, the methods described by DeCelles et al. (1983) were used. Most of the readings were taken at sections situated in the outer Foothills, on the west limb of the Alberta Syncline; that is, east of the Triangle Zone where tectonic deformation is minor. Tectonic dip was less than 25°, so that correction for this component was not difficult.

The measurements were grouped stratigraphically and areally and are presented in the form of rose diagrams on five partly overlapping map segments, each segment corresponding with a post-Wapiabi stratigraphic sequence (Fig. 1). The stratigraphic subdivisions of the post-Wapiabi strata and correlation between the formations in the northern and southern parts of the basin are based on the work of Jerzykiewicz and Sweet (1988).

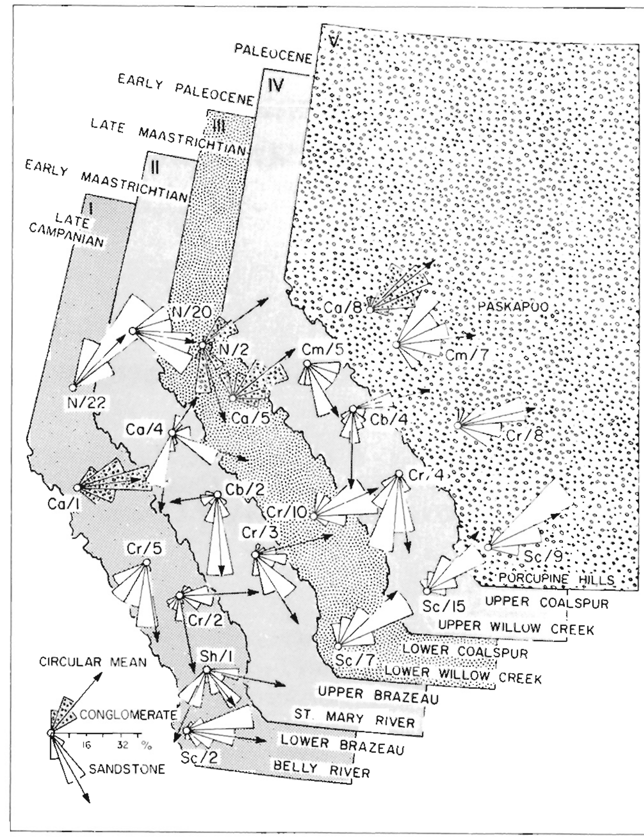

Figure 1. Orientation of directional sedimentary structures in fluviatile channels from five stratigraphic sequences within the uppermost Cretaceous/Paleocene of the Alberta Foothills. The identification numbers, locations, and measurement information for each rose diagram are as follows: N/22 — Bald Mountain Creek (Tp. 69, R. 5), 25 readings; Ca/1 — Yellowhead Highway (Tp. 50, R. 26), 17 readings; Cr/5 — Cripple Creek (Tp. 37, R. 14), 23 readings; Cr/2 — Coalcamp road along the Red Deer River, and adjacent outcrops (Tp. 31, R. 7), 96 readings; Sh/1 — Highwood River (Tp. 18, R. 2), 33 readings; Sc/2 — Railway tracks west of Lundbreck Falls (Tp. 7, R. 2), 22 readings; N/20 — Redwillow River (Tp. 69, R. 10), 33 readings; Ca/4 — Highway 40 near a bridge over the Athabasca River, and ajacent outcrops (Tp. 51, R. 26), 35 readings; Cb/2 — Blackstone River above Brown Creek mouth (Tp. 43, R. 16), 37 readings; Cr/3 — Coalcamp road along the Red Deer River, and adjacent outcrops (Tp. 32, R. 6), 41 readings; N/2 — Highway 40, (Tp. 59, R. 6), 46 readings; Ca/5 — Entrance quarry, by railway tracks, (Tp. 51, R. 26); Cr/10 — Silver Creek, (Tp. 29, R. 6), 27 readings; Sc/7 — Crowsnest River north of Cowley (Tp. 7, R. 1), 29 readings; Cm/5 — Highway 40 at Coalspur (Tp. 48, R. 21), 20 readings; Cb/4 — Chungo Creek (Tp. 43, R. 17), 52 readings; Cr/4 — Coalcamp road along the Red Deer River (Tp. 31, R. 8), 11 readings; Sc/15 — Castle River (Tp. 7, R.1), 57 readings; Ca/8 — Gravel pit (Tp. 51, R. 24), 33 readings; Cm/7 — Embarras River at Robb, and adjacent outcrops (Tp. 49, R. 21), 37 readings; Cr/8 — Fallentimber Creek (Tp. 31, R. 6), 25 readings; Sc/9 — Oldman River Dam site (Tp. 7, R. 29), 76 readings.

The rose diagrams were obtained using Williams' (1980) "ROSENET" plotting program. All angular measurements for each location (or group of outcrops) were compiled in a file compatible, as input data, with this program. Each location has a rose diagram in which the sectors of the circular histograms correspond to 20°. Each histogram was plotted utilising a linear scale for the frequencies. The circular mean value (as defined by Rock, 1988) was computed, utilising the method outlined by Till (1974) in all the cases where there was only one obvious preferred direction. Where there was more than one preferred direction, the angular values were sorted by value. The values around each mode or peak in the histogram were treated as separate files, corresponding to each mode and its immediately adjacent angle values. For each location, as many subfiles were produced as there were separate modes or peaks in the circular histogram. For each of these subfiles, a circular mean was computed in a manner similar to that explained above. The circular means were superimposed onto each rose diagram as vectors (arrows), showing the preferred direction(s).

PRELIMINARY INTERPRETATION

The rose diagrams indicate the directions of fluviatile channels in the post-Wapiabi strata of the outer Foothills belt, where the thickness of these strata is greatest (in excess of 3 000 m); that is, in the foredeep of the Alberta Foreland Basin. The maps presented (Fig. 1) show paleocurrent directions only in the proximal part the basin, and should not be generalized for the distal eastern areas. Significant differences in paleocurrent pattern between the distal and proximal portions of the basin may be expected as a result of a distance from the uplifted region at the mountain front. Northeastward, unimodal, downslope-directed, azimuthal paleocurrents, clearly related to the uplift in the foreland basin thrust belt, are pronounced in Sequences III (lower Coalspur Formation and lower Willow Creek Formation) and V (Paskapoo and Porcupine Hills formations).

The southeasterly paleocurrent direction, postulated by Eisbacher et al. (1974), and by Rahmani and Lerbekmo (1975) as the main paleoflow in the entire basin, is reflected by the directional sedimentary structures discussed in this paper, but only in sequences I, II and IV (lower Brazeau/Belly River, upper Brazeau/St. Mary River, and upper Coalspur/upper Willow Creek). However, even during deposition of these sequences, paleodrainage patterns were much more diversified that previously expected. This is reflected by the polymodal distributions of the paleocurrent vectors (Fig. 1). Some of the modes are southwesterly oriented (Figure 1, Sequences II and IV, roses: Sh/1, Ca/4, Cb/4 and Cr/4) and others are even westerly (rose Cb/2) toward the area presently covered by the thrust belt sequence. This may indicate that the drainage area extended much farther westward than the present boundaries of the deformed belt.

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Preliminary observations on Lower Cretaceous (Albian) paleosols in the Mill Creek Formation of southwestern Alberta

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McCarthy, P.J. and Leckie, D.A., Preliminary observations on Lower Cretaceous (Albian) paleosols in the Mill Creek Formation of southwestern Alberta; in *Current Research, Part B, Geological Survey of Canada, Paper 91-1B, p. 51-58, 1991.*

Abstract

Multiple, lithified paleosols occur in a well exposed section of the Lower Cretaceous (Albian) Mill Creek Formation in the Foothills of southwestern Alberta. The paleosols are well developed and are characterized by their red and olive colour, distinct mottling, vertical root traces, peds, clay films, gradual boundaries, absence of sedimentary structures, and profiles 0.2 to 2.4 m thick.

The paleosols formed during a time when one or more basinwide hiatuses occurred as a result of either eustatic sea-level fluctuations or local tectonic events. These paleosols represent the terrestrial record of base level fluctuations which resulted in reduced rates of sedimentation on the floodplain. The paleosols probably developed under a herbaceous open woodland in a warm, sub-humid climate, on a floodplain of low relief and subject to little erosion. Topographic position, distance from stream channel, and depth to water table appear to have been the major factors controlling pedogenesis. Cumulatively, these paleosols appear to represent a considerable period of time.

Résumé

De nombreux paléosols lithifiés sont observables dans une section bien exposée de la Formation de Mill Creek du Crétacé inférieur (Albien), dans les Foothills du sud-ouest de l'Alberta. Les paléosols sont bien développés et se distinguent par leur couleur rouge et olive, une marbrure particulière, des traces de racines verticales, des agrégats de particules, des pellicules d'argile, des limites graduelles, l'absence de structures sédimentaires et des profils de 0.2 à 2.4 m d'épaisseur.

Les paléosols se sont formés durant une période marquée par un ou plusieurs hiatus étendus à l'échelle du bassin, résultant de fluctuations eustatiques du niveau marin ou d'événements tectoniques locaux. Ces paléosols représentent la signature sur le continent des fluctuations du niveau de base qui se sont traduites par une réduction des taux de sédimentation sur la plaine d'inondation. Les paléosols se sont probablement formés dans une zone de forêt claire herbacée dans un climat subhumide chaud sur une plaine d'inondation peu accidentée et soumise à une érosion légère. La position topographique, la distance des cours d'eau et la profondeur de la nappe phréatique semblent avoir été les principaux facteurs agissant sur la pédogenèse. Considérés dans leur ensemble, ces paléosols semblent représenter une période de temps très longue.

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INTRODUCTION

The purpose of this paper is to document the presence of Lower Cretaceous (Albian) paleosols in the Mill Creek Formation of southwestern Alberta. To date, the description of Cretaceous paleosols in the Western Canada Sedimentary Basin has been sparse (Leckie and Foscolos, 1986; Leckie et al., 1989), although nonmarine sequences in which fossil soils should be preserved are widespread. Recently, paleosols have gained in appeal owing to their ability to provide detailed resolution to paleoenvironmental and paleogeographic questions (Retallack, 1983, 1990; Bown and Kraus, 1987; Kraus, 1987; Allen and Wright, 1989). As a result, more comprehensive studies of paleosols are required in order to critically assess the value of ancient soils in the reconstruction of past landscapes and climates, and to more clearly define the criteria for the recognition of paleosols in sediments. This paper will contribute to the paleosol data base in the Western Canada Sedimentary Basin.

Paleosols are old soils formed under environmental conditions that differed, particularly with respect to climate and vegetation, from those of the present day (Duchaufour, 1982). Unfortunately, paleosols of the geological record are rarely simple buried soils. Compound soils, in which two or more phases of pedogenesis have acted independently, and complex soils, in which more recent pedogenesis affects both the upper paleosol and one or more paleosols below it, are probably very common. As well, the question of diagenesis and its effect on fossil soils must be considered.

Jenny (1941) considered soil development as a function of climate, organisms, topography, parent material, and time. Detailed studies of fossil soils permit inferences as to the operation of these controls within the paleolandscape. The paleosols presented here are from a 30 m thick sequence outcropping along Livingstone River (Twp. 12-Rge. 4W5; 50°00'25"N, 114°20'54"W) in southwestern Alberta (Fig. 1). Macroscopic field evidence documented here allows some preliminary inferences regarding past climate, relief, vegetation, parent material, and time. More detailed petrographic and geochemical analyses will be carried out in the future.

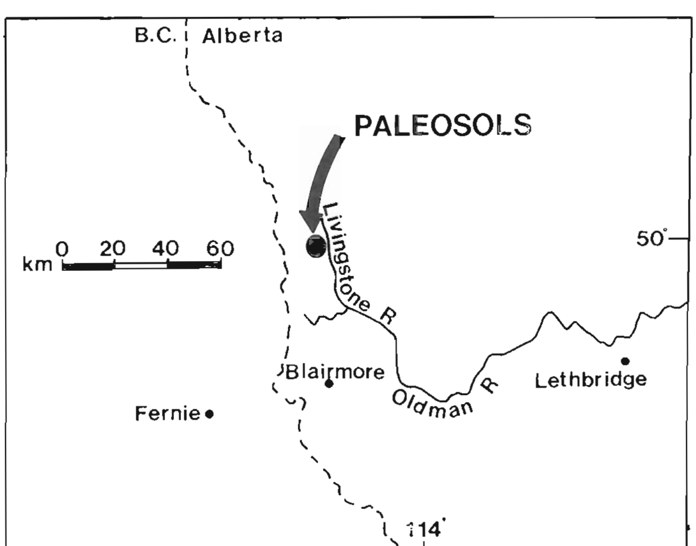

Figure 1. Location of Mill Creek Formation palesols in the Rocky Mountain Foothills of southwestern Alberta.

REGIONAL GEOLOGY

The Mill Creek Formation forms part of the Lower Cretaceous Blairmore Group in southwestern Alberta and is unconformably bounded below by the Beaver Mines Formation and above by the Crowsnest Formation. The Mill Creek Formation is 120 m thick at its type section on Ma Butte, but thins to the north and east (Norris, 1964; McLean, 1982). The bulk of Blairmore detritus was derived from highlands situated to the south and west of the present outcrop areas in mid-Cretaceous time (Mellon, 1967). A depositional hiatus exists between the Beaver Mines and Mill Creek formations; after this hiatus, renewed uplift to the west resulted in the deposition of a thick wedge of fluviatile sediments (Mellon, 1967). Near the eastern edge of the Foothills, upper Blairmore beds interfinger with marine beds that thicken and extend eastward to form the Bow Island Formation (Mellon, 1967).

Mellon (1967) distinguished Mill Creek strata from Beaver Mines strata on the basis of sandstone composition and floral content. The sandstone compositional change is from feldspathic in the Beaver Mines Formation to quartzose in the lower Mill Creek Formation. Two distinct floral groups, non-dicotyledonous and dicotyledonous, were recognized by Dawson (1886) and subsequent workers and particularly emphasized by Mellon (1967). Bell (1956) assigned an Albian age to the Mill Creek strata based upon the floral assemblage, while Mellon (1967) suggested an age of late Middle to Late Albian. Mill Creek strata interfinger with strata of the Crowsnest Formation but are usually quite easily distinguished on the basis of lithological differences (Mellon, 1967).

Sediments of the Mill Creek Formation are composed largely of interbedded mudstones, siltstones and very fine grained sandstones. Conglomerates are present locally. These lithologies are commonly gradational from one to the other, mostly in coarsening-upward sequences (McLean, 1982). The thick, sharp based, fining-upward sandstone beds in the formation have been interpreted by other workers as fluvial channel deposits, and the finer grained beds between have been interpreted as overbank deposits of alluvial plain deposition (Glaister, 1959; Norris, 1964; Mellon, 1967; Vincent, 1977; McLean, 1982).

DESCRIPTION

A well exposed section of the Mill Creek Formation was measured and described along Livingstone River (Fig. 2). At least 17 paleosols are present, ranging in thickness from 0.2 to 2.4 m over a 30 m interval (Fig. 3). The Mill Creek Formation paleosols described below are not unique to this section and are commonly present within Mill Creek deposits in southwestern Alberta. A number of macroscopic features provide field evidence establishing the presence of well developed paleosols within the Mill Creek Formation (Fig. 4).

Figure 2. Stratigraphic section, including paleosols, of the Mill Creek Formation at Livingstone River. The resistant beds are sandstone and the less resistant ones are siltstone and shale. Note the gradual boundaries and even transitions between beds. Person for scale.

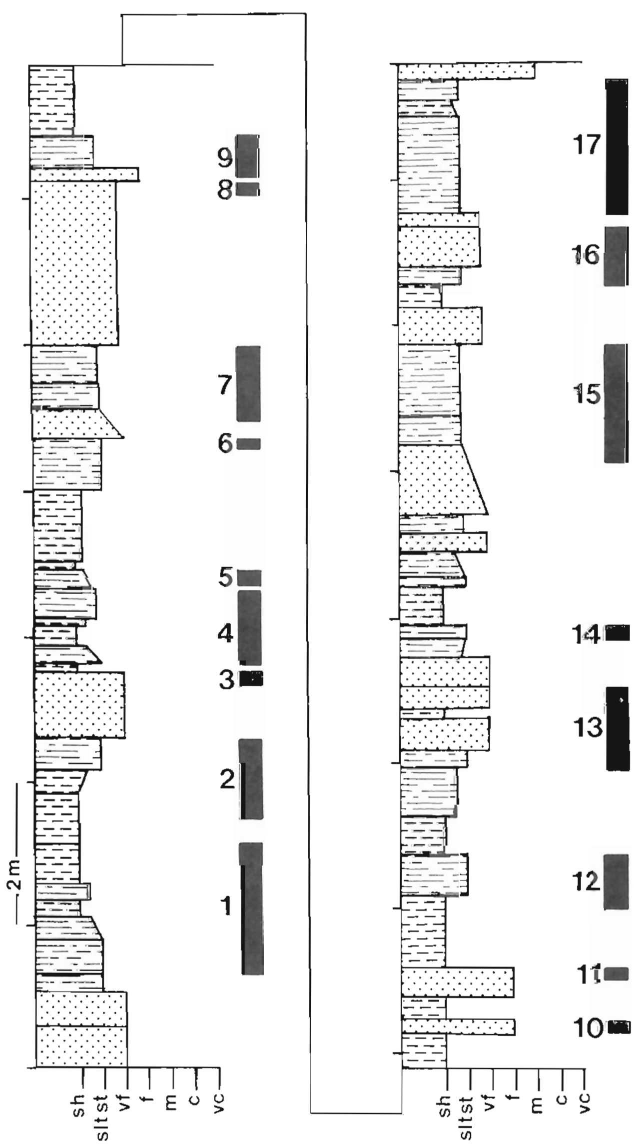

Figure 3. Measured section of the Mill Creek Formation at Livingstone River. The numbered black bars indicate the locations of the paleosols. sh—shale, sltst— siltstone, vf—very fine grained sandstone, f—fine grained sandstone, m—medium grained sandstone, c—coarse grained sandstone, vc—very coarse grained sandstone.

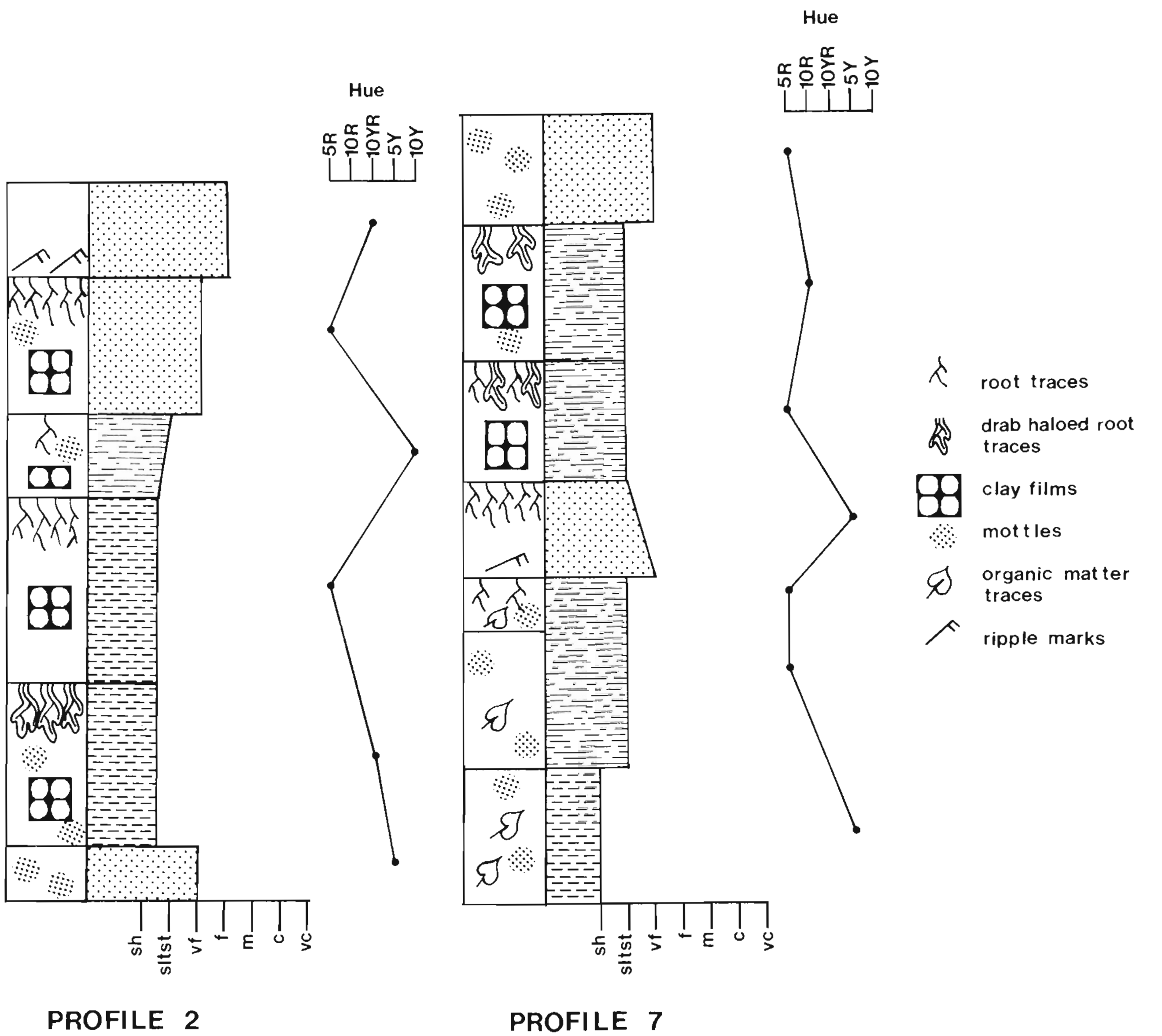

Figure 4. Detailed view of macroscopic features present in paleosol profiles 2 and 7. These features are characteristic of well developed paleosols in the Mill Creek Formation.

Root Traces

In situ fossil roots and root traces are one of the best criteria for recognition of paleosols in sequences of sedimentary rocks younger than Silurian (Retallack, 1990). Fossil root traces are common in Mill Creek Formation paleosols. Fine, carbonaceous root traces (2-5 mm diameter, 5-20 mm long), commonly with pale olive (10 Y 6/2) halos and greyish red (5 R 4/2) matrix, are most common (Fig. 5). Retallack (1990, p. 29, 30) presents a number of explanations for these drab-coloured root traces. The most plausible explanation for the root traces of this type observed during the present study is that they represent organic matter buried within the paleosols which has been subjected to reduction by anaerobic bacterial decay (i.e., a "burial gley" origin).

A second type of root trace found within pale olive (10 Y 6/2) horizons in Mill Creek Formation paleosols consists of carbonaceous root traces surrounded by pale olive (10 Y 6/2) halos that in turn are surrounded by very dusky red (10 R 2/2) halos (Fig. 6). These dusky red halos are believed to form when iron mobilized in the ferrous state from the rhizosphere is oxidized near the roots (Retallack, 1990). Generally, the root traces are most abundant in the top 10 to 50 cm of the paleosols and branch downward.

Medium and coarse root traces (5 to > 15 mm in greatest dimension) are present but are uncommon. Some of the large carbonaceous root traces have characteristic "concertina" outlines, suggesting that some compaction and consolidation of the sediments has occurred (Retallack, 1988). Root traces also are commonly found in clusters within the paleosols rather than as laterally continuous bodies.

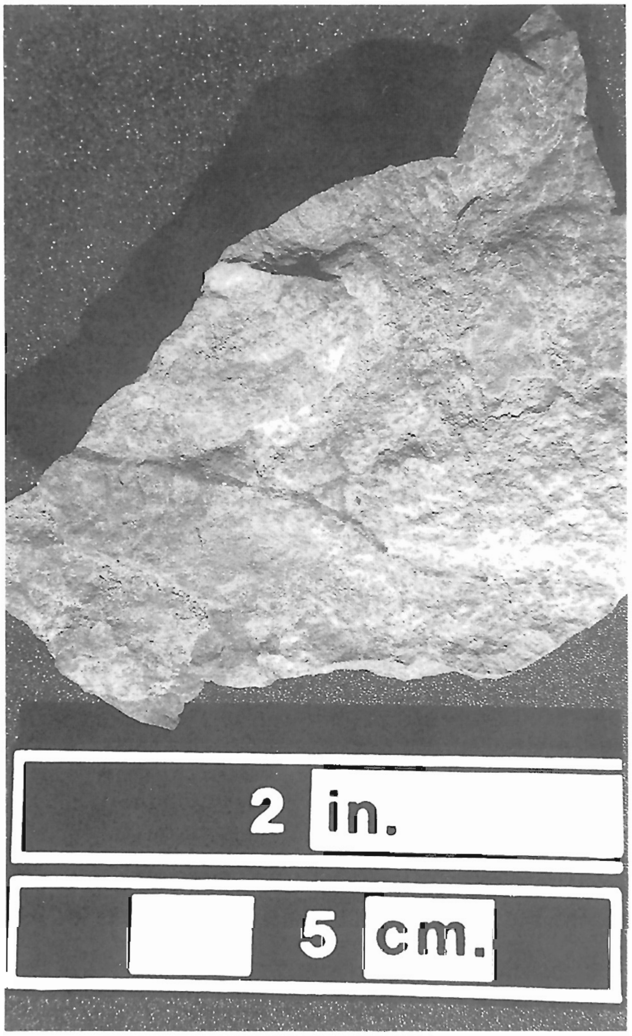

Figure 5. A well preserved carbonaceous root trace. Note the branching nature of the root hairs near the top.

Soil structure

Paleosols may appear to be fragmented, featureless or massive compared to other geological structures; however, soils develop characteristic structures of their own which progressively overwhelm preexisting structures in the parent material. A number of macroscopic soil structural elements are found in the Mill Creek Formation paleosols including peds, cutans, glaebules, and pedotubules.

Peds are naturally occurring aggregates of soil material separated from adjacent peds by cutans or natural voids (Brewer, 1976). Compaction typically destroys most of the original soil peds although some are still visible in the Mill Creek Formation paleosols. Where present, the peds are angular blocky to subangular blocky and prismatic and are 0.5 to 3.5 cm in diameter and are usually accentuated by waxy-looking clay films (Fig. 7). Paleosol horizons tend to weather in a distinctive blocky pattern, which further suggests the presence of an inherent structural control.

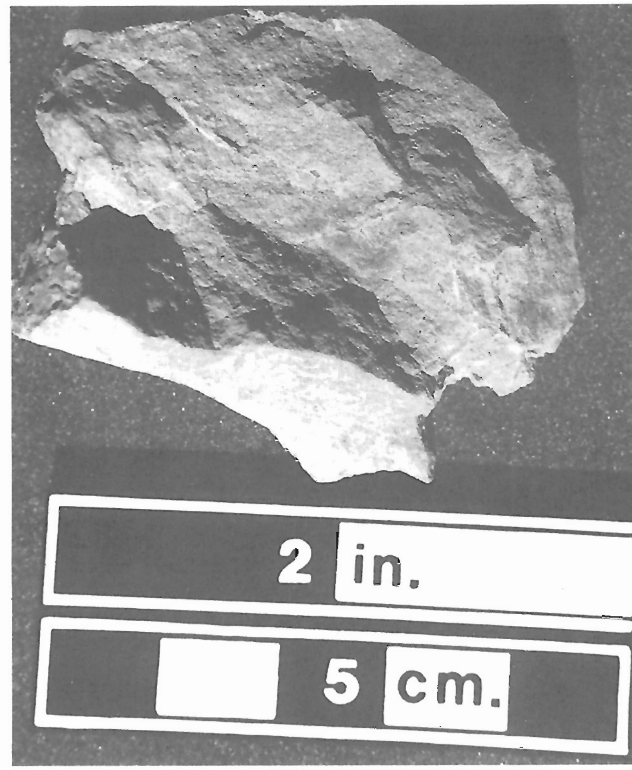

Figure 6. A fine carbonaceous root trace in a pale olive (10 Y 6/2) matrix, surrounded by a very dusky red (10 R 2/2) halo.

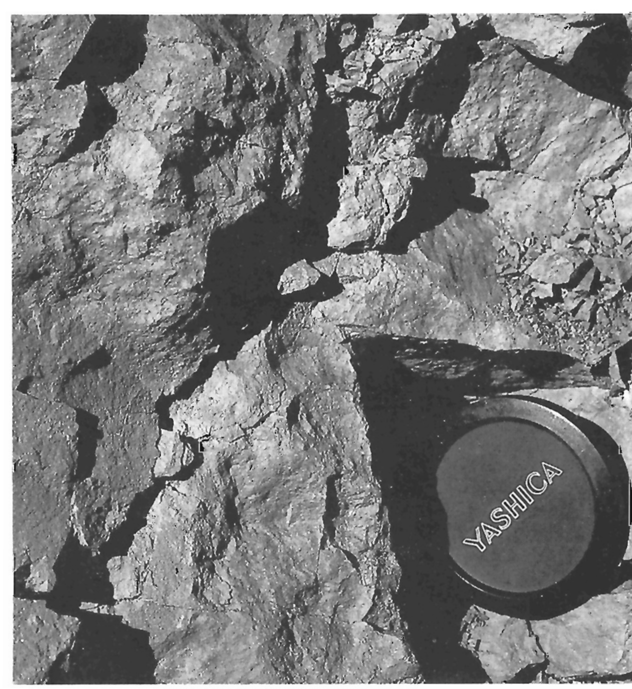

Figure 7. Well developed clay films surrounding peds. Note the waxy lustre, polished surfaces, and random orientation on all sides of the peds. The lens cap is 6 cm in diameter.

Cutans are modified surfaces of peds (Brewer, 1976). Clay films aligned along ped faces occur within the Mill Creek Formation paleosols and can be recognized by their polished and finely grooved surfaces, their waxy lustre, and their random orientation on all sides of the peds. These clay films are present within most of the well developed paleosols, although they vary from very few (< 5 % thin, discontinuous clay films to common (25-50 %), moderately thick clay films. The clay films are not tectonic in origin and probably formed as a result of the dispersion of clays in suspension and their translocation by water moving slowly through pores and cracks in the soil (Birkeland, 1984). A dispersion origin is favoured by several features, including a low electrolyte content in the soil solution, and the absence of positively charged colloids (Barshad, 1964). Clay films form in soils in which the water table is below the surface for some part of the year (Retallack, 1988).

Glaebules are naturally segregated lumps of soil material (Brewer, 1976). The most typical kinds of glaebules are nodules, concretions, and mottles. Only mottles are present in the Mill Creek Formation paleosols at Livingstone River, and these occur as diffuse patches of coloured material. Mottles are very common and vary from a few (< 2 %, fine to medium (< 5 - 15 mm), distinct mottles to common (2 - 20 %), fine to medium, prominent mottles (Fig. 8). Typical mottle colours include greyish red (5 R 4/2), very dusky red (5 R 2/6), blackish red (5 R 2/2), moderate reddish brown (10 R 4/6), dark yellowish orange (10 YR 6/6), and pale olive (10 Y 6/2). Mottles typically form under a fluctuating water table regime, where conditions fluctuate between reducing and oxidizing and the Fe and Mn present alternate from mobile, as $Fe2^+$ and $Mn2^+$, to precipitated, in the $Fe3^+$ and $Mn3^+$ or $Mn4^+$ form (Birkeland, 1984). The result can be a net loss of Fe and Mn, but with local enrichment of these elements in the brightly coloured mottles. The position of mottles helps indicate the position of the water table (Simonson and Boersma, 1972).

"Pedotubule" is a convenient nongenetic term for all tubular features such as burrows and root traces in soils and paleosols (Brewer, 1976). Root traces have been discussed above. Clearly recognizable burrows are uncommon in the Mill Creek Formation paleosols; however, some tubular, pale olive (10 Y 6/2) to light olive (10 Y 5/4), horizontal and subhorizontal mottles, 2 to 5 mm in diameter, may be burrows (Sigleo and Reinhardt, 1988).

Soil Horizons

Boundaries between soil horizons and the underlying parent material are commonly gradational. In the Mill Creek Formation paleosols, the transition from one horizon to the next commonly occurs over an interval of 2 to 15 cm (clear to gradual boundaries), and the horizon boundaries are smooth to wavy laterally. Ordinarily, the top of the uppermost horizon of a paleosol is truncated sharply by an erosion surface (Retallack, 1988). This is not usually the case in the Mill Creek paleosols and the gradual passage from one paleosol horizon to the next may reflect the upward growth of the paleosol as small amounts (10 to 20 cm thick) of alluvium were deposited on the existing soil to form cumulative profiles. Typical horizon thicknesses range from 10 to 60 cm.

Figure 8. A typical mottled horizon containing in situ root traces. The mottles are pale olive (10 Y 6/2) and the matrix is greyish red (5 R 4/2). Fine, haloed root traces are present just above the lens cap (6 cm in diameter).

One of the most striking features of the Mill Creek Formation paleosols is their colour. Pale red (5 R 6/2) to greyish red (10 R 4/2) beds alternate with greyish olive (10 Y 4/2), pale olive (10 Y 6/2), and yellowish grey (5 Y 7/2) in a distinctive pattern. The reddish colouration in the paleosols is probably the result of diagenetic alteration of yellow and brown ferric oxyhydrates during burial (Walker, 1967). This diagenetic change would have enhanced the colour difference between the red horizons and the rest of the paleosol; however, it likely reflects iron enhancement within these horizons in the original soils. The red colouration of these horizons, along with bright mottling in others, suggests that they were dry for at least part of the year, but that drainage was merely impeded at other times. The greyish olive colouration in other horizons probably reflects the zone of poor drainage at depth within the paleosols, or, in some cases, the light colouration may be caused by leaching of humus from surface horizons (Retallack, 1983).

Organic matter is present within the Mill Creek Formation paleosols as comminuted carbonaceous debris rather than as fossil plant remains or as discrete organic horizons. The widespread occurrence of this organic debris in the paleosols suggests that the parent material for the soils may have come, at least in part, from the erosion of existing soils along the banks of stream channels. The absence of organic horizons in the paleosols suggests that the latter were well drained and that oxidation of surfical organic matter was a fairly rapid process. Carbonaceous impressions of fossil gymnosperms and angiosperms have been documented at other locations in the Mill Creek Formation in shales and siltstones overlying the paleosols (Mellon, 1967). These carbonaceous impressions may be representative of plant material that formed the organic A horizons in the original soils.

Absence of sedimentary structures

Primary sedimentary structures are generally absent, indicating that the paleosols are well developed. Destruction of the original sedimentary structures is attributed to bioturbation by roots, organisms, and other soil processes (Retallack, 1990). In a few of the less well developed paleosols there is a decrease in the amount of pedogenic alteration with depth, until original sedimentary structures, commonly parallel lamination and ripple crosslamination, become apparent in the parent material.

INTERPRETATION

Parent Material

The original parent material consists of alluvial floodplain deposits. The paleosols are best developed within siltstones and mudstones, although some weakly developed paleosols occur within sandstone units. Thin, tabular sandstone units (20 to 80 cm thick) are interpreted as crevasse splay deposits. The siltstone and mudstone units are interpreted as deposits of the more distal floodplain of a meandering stream (Vincent, 1977), and shale units are interpreted as small lake deposits. These lithologies are typically gradational from one to the other, mostly as coarsening-upward sequences, or are thinly interbedded. The Mill Creek Formation deposits contain variable proportions of detrital constituents (quartz, feldspar, chert, and nonvolcanic rock fragments), small amounts of volcanic detritus, and a relatively high proportion of metasedimentary, micaceous and chloritic rock fragments (Mellon, 1967).

Climate

The presence of red oxidized horizons throughout the paleosols, the absence of organic horizons and abundance of root traces, and the presence of clay films indicate that the climate was subhumid, probably with a well defined dry season. The general absence of caliche or calcareous horizons precludes an interpretation that the climate was drier. Diagenetic reddening of originally yellow and brown horizons suggests that the soils probably formed under a warm-temperate temperature regime rather than a tropical one. The irregular nature of root distribution within the paleosols and the presence of root traces with drab halos suggest that the soils supported an open woodland type of vegetation, consisting of scattered trees with a herbaceous ground cover (Retallack, 1983).

Topography

Most of the paleosols appear to have been moderately well drained for at least part of the year. Red colouration as well as bright red, orange and purple mottles indicate that water table fluctuations were common within these soils. Fluctuations of this type could be generated within floodplain soils that were elevated slightly above stream level. The drab nature of some lower horizons suggests that the water table was never more than 0.5 to 1 m from the surface of the soils, and that lower horizons may have been permanently gleyed. Low relief, and consequently little erosion, might also account for the presence of numerous superposed paleosols (Leckie and Foscolos, 1986).

Time

A number of characteristics of the Mill Creek Formation paleosols indicate that substantial periods of time were required for their formation. Well differentiated soil horizons, profiles up to 2 m thick, well developed peds with clay films, and an absence of sedimentary structures indicate that some of the paleosols took a long time to develop (Retallack, 1976). Some of the paleosols are much less well developed and probably required substantially less time to develop although, cumulatively, the paleosols represent a long period of time when deposition was slow and there was little erosion. The fact that the soils developed on weathered and reworked alluvial material suggests that they would have required less time to form, but the well developed soils probably still represent time intervals in the order of several thousands of years. A further complication in assessing the time period these paleosols represent is that some of the thick paleosols probably represent compound and complex cumulative profiles in which soil development and sediment influxes occurred concurrently (Duchaufour, 1982). Soil profiles of this type are commonly found in floodplain environments.

CONCLUSIONS

At least 15 paleosols are present within a 30 m thick section of the Mill Creek Formation along Livingstone River in southwestern Alberta. A number of macroscopic features are useful in the identification of the paleosols, including: in situ vertical root traces; angular blocky and subangular blocky peds; clay films coating the outer surfaces of some peds; the striking greyish red (5 R 4/2), greyish olive (10 Y 4/2) and pale yellowish brown (10 YR 6/2) colouration of some horizons; intensely mottled horizons with gradual, wavy boundaries; and the virtual absence of sedimentary structures.

The paleosols appear to have formed in a warm subhumid climate with a fluctuating water table. The absence of organic horizons, despite the abundance of root traces, suggests that organic matter was rapidly oxidized and incorporated into the mineral material of the soils. The bright red horizons, although diagenetically enhanced, are believed to reflect accumulations of iron and aluminum in the yellow or brown B horizons of the original soil. The rooting pattern, as well as the nature of the drab-haloed root traces, suggest that the original vegetation was a type of herbaceous open woodland. The fact that numerous superposed paleosols exist suggests that the rate of sediment influx was low and that very little erosion took place, probably as a result of low relief. The time required to develop such a suite of paleosols suggests that landscapes of the Mill Creek Formation were generally stable. The stability of the landscape and resulting long time period available for pedogenesis may have been related to a major Late Albian hiatus, reflecting a regional sea-level fluctuation during this time, caused by either eustatic or tectonic mechanisms.

It is premature to attempt to classify the types of soils that developed within the Mill Creek Formation; however, it is possible to deduce some of the possible pedogenic processes. The clay films identified in some of the paleosols probably formed as the result of the translocation of fine clay particles in suspension from the surface horizons to the B horizons. It also appears that some translocation of iron, and probably aluminum, into the B horizons has occurred. The limited data available at present suggest that the soils formed under warm temperate, subhumid conditions, with topography, distance from stream channel, and depth to water table being the main factors controlling soil genesis. Less well developed paleosols appear to have been influenced by the same genetic mechanisms as the well developed soils but over much shorter periods of time. The well drained nature of the Mill Creek Formation paleosols and the absence of coal seams indicate that these paleosols are unique relative to any previously described from the Western Canada Sedimentary Basin. Ongoing data analysis of whole rock chemistry, clay mineralogy, and micromorphology may reinforce or modify these conclusions.

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The western margin of the Upper Devonian Fairholme Reef Complex, Banff-Kananaskis area, southwestern Alberta

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Abstract

The Fairholme Reef Complex is the southernmost and largest in an extensive reef domain developed in "Alberta" during the Frasnian. Its western margin differs significantly from all other reef margins within the domain. The early and late stages of the margin are typically dominated by stromatoporoids and amphiporids, but the middle stage (upper Cairn) is dominated by corals; a change associated with a northeastward backstepping of the coral reef facies.

This change in reef community may have been due to a temporary increase in nutrient supply, which inhibited stromatoporoid growth and provided an ecological niche for corals and bioeroders. A later return to lower nutrient levels caused a strong southwestward progradation of a stromatoporoid dominated reef front (Peechee), changing the configuration of the upper reef margin significantly.

An abruptly deepening basin to the west was partly infilled by anoxic sediments (Perdrix), reef-derived debris beds and oxygenated sediments (Mt. Hawk), and was finally aggraded by westerly derived Fammenian siliciclastics (Sassenach).

Résumé

Le complexe récifal de Fairholme est l'entité la plus volumineuse et la plus méridionale d'un domaine de récifs étendu qui s'est formé en « Alberta » pendant le Frasnien. Sa marge occidentale diffère beaucoup de toutes les autres marges récifales du domaine. Les stades anciens et récents de la marge sont en général dominés par des stromatoporoïdés et des amphiporidés, mais le stade intermédiaire (Cairn supérieur) est dominé par des coraux, ce qui reflète un retrait vers le nord-est du faciès de récif corallien.

Ce changement dans la communauté récifale a pu être causé par une augmentation temporaire de la quantité d'éléments nutritifs disponibles, inhibant la croissance des stromatoporoïdés et constituant une niche écologique pour les coraux et les bioérodeurs. Une raréfaction ultérieure d'éléments nutritifs a causé une forte avancée vers le sud-ouest d'un front récifal (Peechee) dominé par les stromatoporoïdés, changeant considérablement la configuration de la marge récifale supérieure.

L'ouest, un bassin qui s'est approfondi brusquement a été partiellement envahi par des sédiments anoxiques (Perdrix), des couches de débris d'origine récifale ainsi que de sédiments oxygénés (Mt. Hawk) et a été finalement alluvionné par des roches silicoclastiques du Famennien en provenance de l'ouest (Sassenach).

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INTRODUCTION

During Frasnian (early Late Devonian) time, central and western Alberta was the site of an extensive reef domain (Fig. 1). The eastern and central parts of this reef domain are now entirely in the subsurface, but the western part is well exposed in a series of thrust sheets along the eastern side of the Cordillera. Four large reef complexes are exposed in the Alberta/British Columbia Rocky Mountains: Ancient Wall, Miette, Southesk-Cairn, and Fairholme (Fig. 1). This paper outlines the paleogeography, stratigraphy, and depositional history of the western margin of the Fairholme Reef Complex and its associated basinal strata.

Study area and previous work

Field studies were carried out in 1989 and 1990 in the Front Ranges of the Canadian Rocky Mountains south of the Bow Valley, encompassing part of Banff National Park, Kananaskis Country (including Peter Lougheed Provincial Park), and Mt. Assiniboine Provincial Park (Fig. 2).

Paleogeographic reconstructions of the Fairholme Reef Complex show only dashed lines or question marks along its western margin (Mountjoy, 1980; Geldsetzer, 1987). Moore's (1989) map shows the western margin more precisely, but is not accompanied by a discussion of evidence

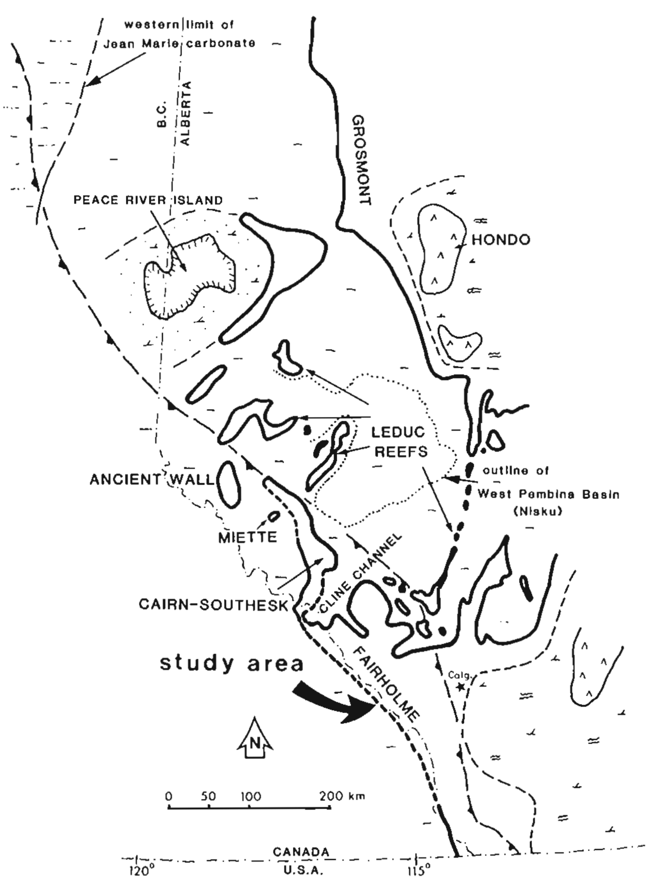

Figure 1. Distribution of Upper Devonian reef complexes and location of study area. (After Geldsetzer, 1987.)

documenting this margin. Detailed geological maps of the Banff and Canmore areas have been published as part of the Bow/Athabasca project (Price and Mountjoy, 1970a,b, 1972a,b), and recently of Peter Lougheed Provincial Park (McMechan, 1988). Usher (1959) and Bielenstein et al. (1971) have published large-scale geological maps of portions of the study area.

Geological accounts of Fairholme stratigraphy and sedimentology in the study area include Belyea and McLaren (1956), Usher (1959), Weihmann (1979), Beales and Brown (1963), Workum and Hedinger (1988), and Bloy et al. (1988).

LITHOSTRATIGRAPHY

The stratigraphic units of the Fairholme Group, comprising both the carbonate buildup and basinal successions, are summarized in Figure 3. The carbonate buildup succession was subdivided by McLaren (1956) into the lower ("black reef") Cairn Formation and the upper ("white reef") Southesk Formation (Fig. 4). The basinal succession comprises a lower carbonate platform unit (Flume Formation) and overlying basin-fill shale and limestone of the Perdrix and Mount Hawk formations.

Carbonate buildup succession

The Cairn Formation is generally a grey to dark grey fossiliferous dolomite, and has been subdivided into two members: Flume and "Upper Cairn" (informal). The Flume Member continues laterally beyond the buildup as the Flume Formation, below the basinal succession. This regional carbonate platform unit onlaps the West Alberta Ridge, overstepping Ordovician carbonate rocks along the western slope of the ridge (western part of the study area) and Cambrian carbonates toward the ridge crest at Fisher Peak and Mt. McDougall (eastern part of the study area). Thin lenses of Middle Devonian sediments of the Yahatinda Formation occur locally below the Flume carbonate platform.

The basal beds of the Flume Member contain brachiopod-crinoid-gastropod mudstones and wackestones and, locally, a thin brown sandstone. The main portion of the Flume is biostromal, consisting of stromatoporoid- and *Amphipora*-bearing rudstone and floatstone, with thin cryptalgal laminite beds.

The thickness of the Flume Member commonly ranges from 21 to 42 m, but exceeds 94 m in the Sundance Range. In outcrop sections, the contact with the overlying Upper Cairn member is often difficult to recognize. In the area studied, the Flume Member lacks the characteristic cherty beds present in the Ancient Wall and Miette buildups (Mountjoy and MacKenzie, 1974). The stromatoporoid lithofacies of the Flume continues into the Upper Cairn member, but is locally separated by a brachiopod/crinoid dolomudstone lithofacies. This lithofacies probably correlates with the basal Perdrix Formation of the basinal succession and may be the local depositional response to an event that inhibited further growth of the Flume carbonate platform and led to buildup inception and localization (Mountjoy, 1980; Geldsetzer, 1988).

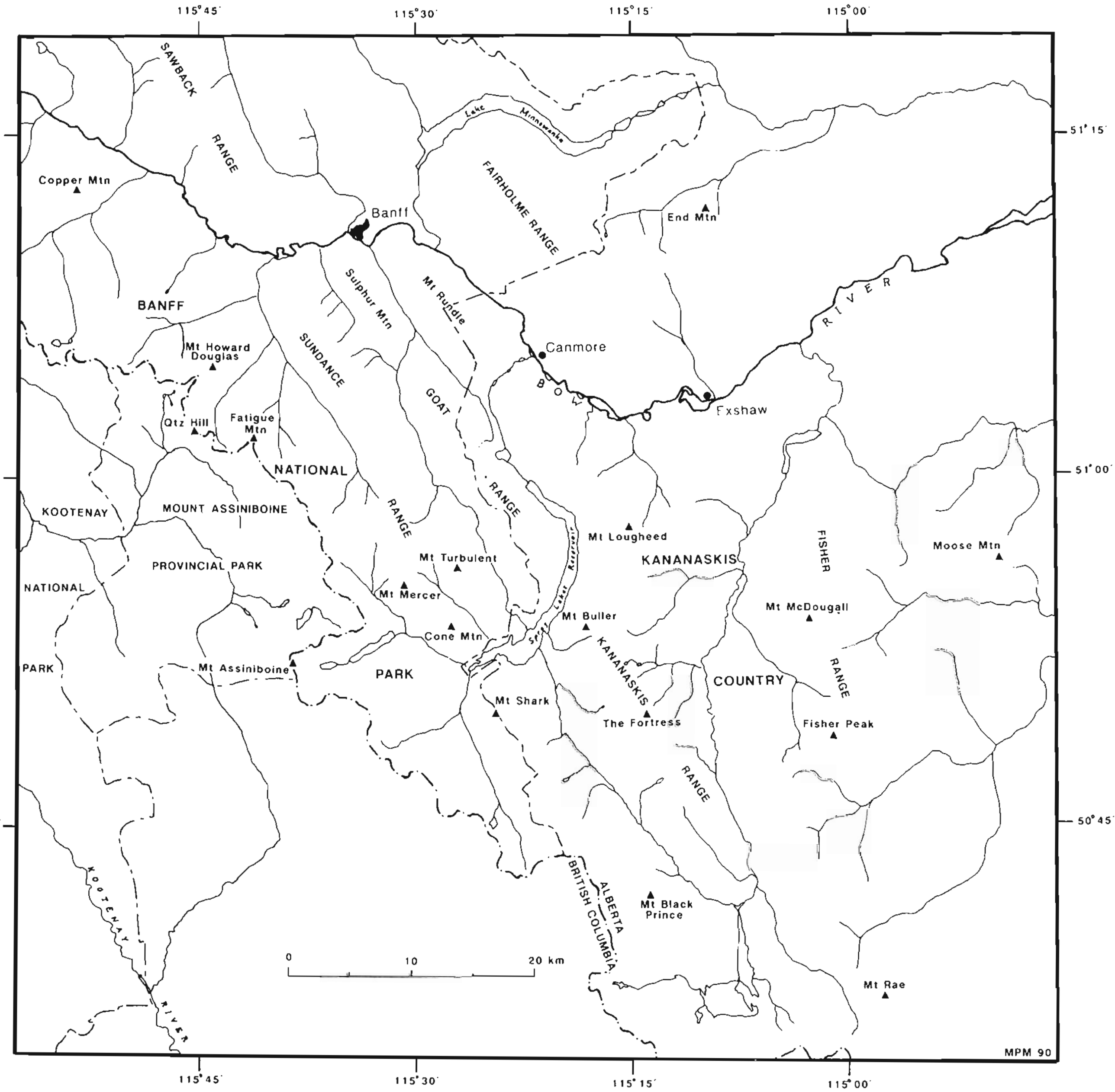

Figure 2. Detailed map of study area, Banff — Kananaskis region.

The lower part of the Upper Cairn member invariably consists of alternating beds of three main lithofacies: i) a medium to thick bedded, stromatoporoid biostromal boundstone; ii) a thin to medium bedded *Amphipora* floatstone; and iii) a thin bedded cryptalgal laminite. Commonly, these three lithofacies occur in the order (from the base) listed above (Fig. 5) and form cycles 2 to 6 m in thickness. The three lithofacies commonly persist into the upper part of the Upper Cairn member at other localities within the Fairholme Reef Complex and at buildups such as Southesk-Cairn and Ancient Wall (Geldsetzer, 1988; Shields and Hedinger, 1990). In the study area, however, the upper part of the Upper Cairn member is characterized by a unique coral lithofacies; no stromatoporoids are present. The coral lithofacies is dominated by large colonies of robust branching corals (Fig. 6); *Thamnopora*, solitary and colonial rugose, and tabulate corals are also common. These corals built biostromes and local patch reefs 30 to 40 m thick (Workum and Hedinger, 1988). Commonly associated with the coral beds are skeletal dolowackestones containing brachiopods, crinoids, and gastropods. The coral lithofacies persists up to the Cairn/Southesk contact; its thickness ranges from 43 m in the west to 154 m in the eastern part of the study area at Fisher Peak. The total thickness of the Cairn Formation ranges from 186 m (in the east) to 340 m (toward the west).

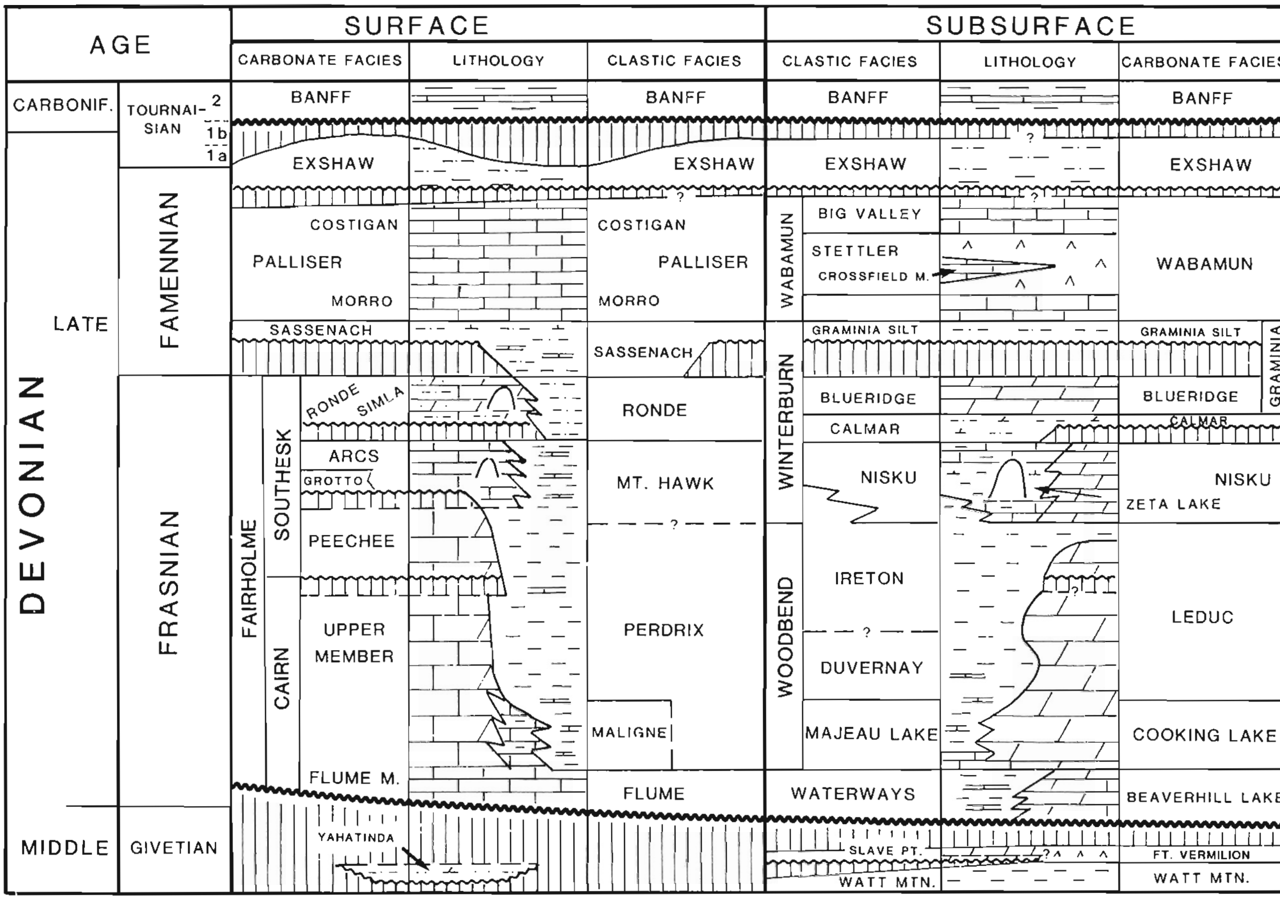

Figure 3. Upper Devonian stratigraphy and correlation chart, west-central Alberta. (From Geldsetzer, 1987.)

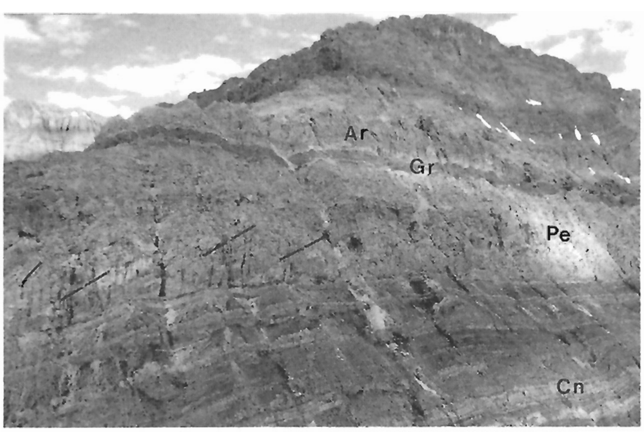

Figure 4. Buildup margin section, showing the Fairholme carbonate succession, 1.5 km northeast of Mt. Mercer. Note the basinward (southward) progradation of sloping Peechee beds (at left of photo). View toward the west. Cn = Cairn, Pe = Peechee, Gr = Grotto, Ar = Arcs.

The overlying Southesk Formation is divided into four members: Peechee, Grotto, Arcs, and Ronde. Only the Peechee Member is part of the reef complex proper, whereas the other three members represent post-reef carbonate deposition with rare local patch reefs. The type sections for the Peechee, Grotto, and Arcs members were located by Belyea and McLaren (1956) at Mt. Rundle near Canmore.

The Peechee Member is invariably a light grey, coarse crystalline, resistant weathering dolomite. Although primary textures are commonly obscured by pervasive dolomitization, the three lithofacies described in the lower part of the Upper Cairn member can also be recognized within parts of the Peechee. Sharp, irregular bedding contacts and local unconformities are present, as are trough crossbeds and channel structures infilled with dolograinstone. Also preserved within the Peechee at two localities near the buildup margin is a coral lithofacies 5 to 7 m thick. The lower contact with the Cairn Formation is locally sharp, and in places irregular, but is commonly characterized by alternating light and dark grey weathering beds, a typical "transitional zone" (Mountjoy and Mackenzie, 1974; Shields and Hedinger, 1990). The thickness of the Peechee Member increases from east to west, from 51 to 165 m; an anomalous thickness of 359 m was measured at the buildup margin near Fatigue Mountain and Mt. Howard Douglas. Repetition of strata due to bedding-parallel thrust faults is a possible explanation, but extremely difficult to document in outcrop.

Figure 5. A "condensed" sequence of stromatoporoid (S), *Amphipora* (A), and algal laminite (L) lithofacies of the Cairn Formation, Sundance Range, Banff National Park. Lens cap is 5 cm in diameter.

Figure 6. A large, robust branching, colonial rugose coral; coral lithofacies of the Upper Cairn member at Fortress Mountain in Kananaskis Country.

The Peechee/Grotto contact is sharp and irregular, and probably represents a regional unconformity (Mountjoy, 1980; Workum, 1983). The Grotto Member is a massive, dark grey, slightly recessive unit consisting mainly of colonial and rugose corals, brachiopods and crinoids. At buildup interior sections exposed in the eastern part of the study area, the Grotto Member is difficult to recognize. The Grotto "interval" is dark weathering, and consists of alternating, light and dark grey, medium bedded dolomite with *Amphipora* skeletal fragments and algal(?) laminae. The thickness of the Grotto Member ranges from 5 to 36 m.

The Arcs Member conformably overlies the Grotto. The Arcs Member is typically a light grey, thick bedded, resistant dolomite unit, with local massive, vuggy, stromatoporoid-rich biostromes and bioherms near the buildup margin. Laminated dolomites, local unconformities, and dolograinstones (within tidal? channel structures and crossbeds) are more common in surface sections of the buildup interior. The Arcs Member ranges from 22 to 128 m in thickness.

The Ronde Member can be easily recognized only near the buildup margin, at Mt. Buller and along the southern portion of Sundance Range, where it is approximately 75 m thick. The Ronde can be subdivided into two units: a distinctive, light brown weathering, basal siltstone (Calmar-equivalent), and an upper carbonate unit. The "Calmar" siltstone unit (3 to 10 m thick) disconformably overlies the Arcs Member, and probably represents a new depositional phase following a brief regional hiatus (Morrow and Geldsetzer, 1988). The upper carbonate unit consists of medium to thick bedded, light grey dolomite, and rare silty dolomite. Common textures include planar- and crosslaminated dolograinstones, cryptalgal laminites, and colonial rugose coral beds. At Mt. Buller, stromatoporoid mounds, 21 m thick, are present at the base of the carbonate unit.

The Sassenach Formation is a siliciclastic unit which disconformably overlies the Fairholme Group strata of the carbonate buildup (Fig. 7). It is a light brown to grey brown, medium to thick bedded siltstone and silty dolomite, with laminated and trough crosslaminated beds. The coarse silt-sized quartz grains are typically subangular and well sorted. The Sassenach Formation is 16 to 30 m thick above the carbonate buildup succession. At the buildup margin, the Sassenach thickens basinward to a thickness of 272 m where it conformably overlies basinal strata of the Mt. Hawk Formation. Just northwest of Fatigue Mountain near the buildup margin, the Sassenach pinches out completely (Fig. 7), and then thickens again northward from this point toward the buildup interior (Price and Mountjoy, 1972a).

In the interior of the carbonate buildup, both the Ronde Member and the Sassenach Formation were deposited in very shallow water, and are difficult to differentiate. The term Alexo Formation is used to include both units in these areas. The Alexo Formation is generally recessive, consisting of thin, light brown to brownish grey siltstone, dolomitic siltstone, and silty dolomite. Shallow water textures are common, including mudcracks, algal laminites, desiccated algal mats, and breccias (with laminated clasts). The Alexo Formation is 25 to 50 m thick; a thickness of 101 m was measured at Mt. Rundle near Canmore but, in this section, the upper part of the formation is disturbed by thrust faulting.

Basinal succession

The lowermost unit of the basinal succession is the carbonate platform of the Flume Formation (de Wit and McLaren, 1950), which overlies lenses of Middle Devonian sediments and Ordovician carbonate rocks unconformably along the western slope of the West Alberta Ridge. Owing to its time-transgressive nature the Flume Formation seems to correlate

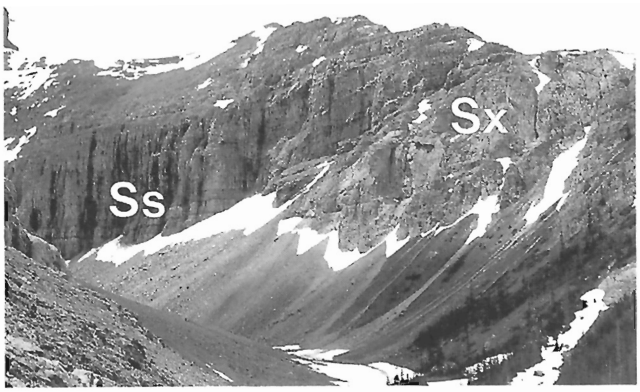

Figure 7. Sassenach Formation (Ss), thinning and onlapping the upper Southesk (Sx) Formation (Ronde Member) near the Fatigue Mountain buildup margin, 3 km southeast of Mt. Howard Douglas, Banff National Park.

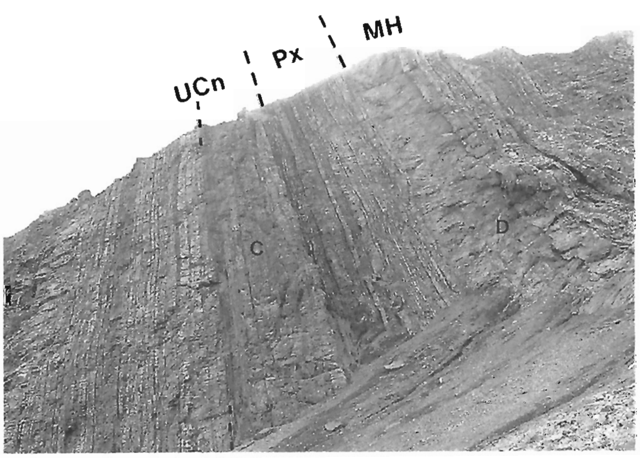

Figure 8. Upper Cairn (UCn), Perdrix (Px), and basal Mount Hawk (MH) strata at Cone Mountain, Banff National Park. The coral lithofacies (c) within the Upper Cairn member is 28 m thick. Note the light coloured debris interbeds within the Perdrix Formation, and the 9 m thick debris flow unit (D) at the base of the Mount Hawk Formation.

only with the basal, non-stromatoporoidal part of the Flume Member, below the buildup margin, and may not be represented at all along the highest part of the ridge crest.

The Flume Formation consists of grey to dark grey, argillaceous limestone and dolomitic limestone, replete with well preserved atrypid and spiriferid brachiopods, and rare crinoids and *Stachyodes*. The Flume Formation is overlain by carbonate rocks of the Upper Cairn member at the buildup margin, or by shales of the Perdrix Formation in the basinal setting.

The Perdrix Formation typically comprises a dark grey to black, euxinic shale, but may also contain brachiopods and crinoid skeletal fragments, tentaculitids, chert nodules, and interbeds of argillaceous limestone and dolomite. Scattered debris beds are present at Quartz Hill, where the total thickness of the Perdrix Formation reaches 130 m. The buildup margin sections at Mt. Turbulent, Cone Mountain and Copper Mountain exhibit Perdrix "tongues" (20 to 42 m thick) overlying the Cairn Formation of the carbonate buildup (Fig. 8). Thick bedded, dolomitic debris beds (Fig. 8) occur within the dark shale and nodular limestone of the Perdrix incursions.

The overlying Mt. Hawk Formation consists of greyish brown weathering, nodular, argillaceous limestone with interbeds of black shale and light brown silty dolomite. The Perdrix/Mount Hawk contact is commonly a zone of variable thickness across which weathering colour changes from the dark grey/black of the Perdrix shale and shaly limestone to the grey-brown of the Mount Hawk nodular limestone. The Mount Hawk Formation ranges in thickness from 123 to 193 m and locally contains fossiliferous beds; tentaculitids, brachiopods, tabulate and rugose corals, crinoids, and fish fragments are commonly present. In addition, preliminary examination of residues from samples digested in acid indicate the presence of pelecypods, ostracodes, sponge spicules, and conodonts.

At localities close to the buildup margin, the Mount Hawk Formation contains a basal debris flow unit, up to 40 m thick (Figs. 8, 9). Debris flow deposits at this stratigraphic level have been described from the Ancient Wall and Miette buildups, where they form excellent local stratigraphic markers (Cook et al., 1972; Mountjoy and MacKenzie, 1974; Geldsetzer, 1988). These debris beds indicate that the base of the Mount Hawk correlates with the top of the Peechee (Mountjoy and MacKenzie, 1974; Mountjoy, 1980).

The debris beds in the study area contain angular to subrounded lithoclasts that range from less than 1 cm to 144 cm in largest dimension, and float in a skeletal lime mud matrix (Fig. 10). The clasts originate from peloidal grainstones, skeletal packstones, and wackestones. Tabulate and rugose coral debris dominates the skeletal component; brachiopods, crinoids, and stromatoporoids are subordinate. The lithoclasts and surrounding matrix are limestone, except at one outcrop where both constituents are dolomitized. The debris beds at Fatigue Mountain directly overlie

Figure 9. Debris flow unit in the basal Mount Hawk Formation at Fatigue Mountain. Arrows point to the lobe-shaped debris units.

Figure 10. Close-up view of the limestone debris flow unit shown in Figure 9. Fatigue Mountain, Banff National Park.

Figure 11. Buildup-margin slope at Fatigue Mountain, showing the Upper Cairn member (Cn) overlain by bioherms (bh) and a dolomitized debris flow unit at the base of the Mount Hawk (MH) Formation. The bioherms are discontinuous (arrow).

Figure 12. Megabreccia clast within a debris flow unit at Copper Mountain in Banff National Park.

stromatoporoid bioherms on the reef slope (Fig. 11). These debris deposits are lobe shaped; the "lobes" are 5 to 10 m thick, separated laterally by dolomitic, crinoidal mudstone beds. At Copper Mountain, light grey megabreccia blocks (as large as 2 m thick and 4 m wide) float in a dark grey matrix containing debris of corals, stromatoporoids, crinoids, brachiopods, and lithoclasts (Fig. 12).

The Mount Hawk is overlain by the Sassenach Formation. The contact generally coincides with a change to thick bedded, light brown, silty dolomites and siltstones. The disappearance below the contact of tentaculitids, robust branching and thamnoporid corals, and brachiopods is quite apparent. At Cone Mountain, black calcareous shale and argillaceous lime-mudstone beds occur at the base of the Sassenach Formation, as in the outcrop section at Medicine Lake in Jasper Basin where the base of the black sediment represents the Frasnian/Famennian boundary (Geldsetzer et al., 1987). The upper part of the Sassenach Formation commonly consists of light grey weathering sandstone, with interbeds of light brown siltstone and silty dolomite. The contact with the overlying grey mottled limestone of the Palliser Formation is sharp.

BIOSTRATIGRAPHY

Studies of Frasnian biostratigraphy in Western Canada have documented and utilized zonal schemes of brachiopods (McLaren 1962; Maurin and Raasch, 1972; Norris 1983); conodonts (Pollock, 1968; Uyeno, 1974; Weissenberger, 1988; Klapper and Lane, 1985, 1988); ostracodes, brachiopods and conodonts (Braun et al., 1988); stromatoporoids (Stearn, 1979); and corals (Sorauf and Pedder, 1986). Unfortunately, there is a general lack of regional markers for time correlation in the Alberta Basin (Mountjoy, 1980). Brachiopod zones can be too broad within certain formations, and conodonts may be limited to certain biofacies and are commonly absent from carbonate buildups. This complicates detailed correlation between the buildup and basinal formations.

Conodont and brachiopod biostratigraphy

The Flume Formation yielded brachiopods and conodonts indicative of the Lowermost to Lower *asymmetrica* Zone. At an isolated outcrop near Sunshine Village, the brachiopods *Ladogioides kakwaensis* and *Schizophoria* cf. *S. allani* were collected from beds together with the conodont *Polygnathus* cf. *P.* sp. B of Uyeno (1974). The forms suggestive of *S. allani* and *P.* sp. B occur typically in the Calumet Member of the Waterways Formation of northeastern Alberta (Uyeno, 1974; Norris, 1983). The Middle/Upper Devonian boundary between the Lowermost and Lower *asymmetrica* Zones occurs within the lower part of the Calumet Member (Uyeno, 1974; Uyeno, in Braun et al. 1988). *L. kakwaensis* occurs typically in the Flume Formation in the Kakwa Lake area of British Columbia, and in the lower Cairn Formation at Ancient Wall. Maurin and Raasch (1972) suggested that *L. kakwaensis* occurs at a slightly higher level than *L. pax*. The latter form occurs in the Firebag Member of the Waterways Formation, where it is associated with conodonts of the Lowermost *asymmetrica* Zone (Norris, in Braun et al., 1988).

The lower and middle parts of the Flume Formation at Quartz Hill contain the brachiopods *Pseudogruenewaldtia multicostellata, Allanella minutilla, Desquamatia clarkei, Eleutherokomma jasperensis, Schizophoria* sp., *Cranaena* sp., and *Cyrtina* sp., all of the *E. jasperensis* Zone, within conodont Lower *asymmetrica* Zone (Norris, *in* Braun et al. 1988). Associated with these brachiopods at Quartz Hill is the conodont *Polygnathus cf. P. dubius*, found in Lowermost to Lower *asymmetrica* zones of the Micritic limestone beds, Souris River Formation of Manitoba (Uyeno, pers. comm., 1990).

Ancyrodella rotundiloba occurs in the uppermost beds of the Flume Formation, within three metres of the contact with the overlying Perdrix Formation at Quartz Hill. *A. rotundiloba* occurs in the first three zones of the Montagne Noire sequence in France (Klapper, 1988). It also occurs in the Calumet, Christina and Moberly members (Lower *asymmetrica* Zone) of the Waterways Formation (Uyeno, 1974).

Few attempts have been made to correlate buildup and basinal formations using conodont biostratigraphic zones, because of the difficulties noted earlier. Weissenberger's (1988) preliminary study documented conodont biostratigraphy of the Southesk-Cairn and Fairholme reef complexes in the Nordegg area. Ziegler's (1971) standard conodont zonation was used by Weissenberger in an effort to correlate the Frasnian succession.

Klapper and Lane (1988, p. 469) argued that the Frasnian standard zonation is not applicable to the Alberta sequence, especially from the *A. triangularis* Zone upward, because of inconsistency in taxonomy and correlation between different biofacies. They outlined an alternative Frasnian conodont sequence for the Alberta basin, recognizing five informal zones (from the top of Flume Formation), succeeded by three faunal intervals (Fig. 13). This Alberta sequence includes nearshore forms such as *Polygnathus* and *Icriodus*, as well as deep-water forms such as *Palmatolepis* and *Ancyrodella*, thus enabling more precise correlations across biofacies boundaries.

The results of a preliminary analysis of conodont biostratigraphy for the study area, using the Alberta sequence of Klapper and Lane (1988), are outlined below and summarized in Figure 13.

1. At the thickest occurrence of the Cairn coral lithofacies (Fisher Peak and Mount McDougall) the basal beds belong to zone 2. This is based on the occurrence of *Polygnathus* aff. *P. angustidiscus* (zones 2-4a) and *Mesotaxis* n. sp. Q (Klapper and Lane, 1985)? (zone 2).
2. The Cairn/Peechee boundary is no older than zone 4b, and probably occurs at or near the 4b/5a boundary. The species *Polygnathus* cf. *P. elegantulus* (zones 4b-6), *P.* cf. *P. evidens* (zones 4a-5b), *P.* cf. *P. morgani* (zones 4b-6), and *Ozarkodina* cf. *O. postera* (zones 4b-7/8) occur in the uppermost Cairn Formation at various localities.
3. The Perdrix/Mount Hawk contact falls within the 5a zone; probably in the middle of the zone as at Mt. Haultain (Klapper and Lane, 1988). The age of this contact is based on the occurrence of *Palmatolepis* cf. *P. domanicensis* (zones 4a-5a, at Mt. Turbulent), and *P. semichatovae* (zone 5a,b, at Fatigue Mountain) within the lowermost beds of the Mount Hawk Formation.
4. The Grotto Member is no younger than zone 5b, based on the occurrence of *Ancyrodella curvata* (zones 4a-5b) at Sundance Range, and, since Weissenberger (1988) noted the occurrence of *Polygnathus evidens* in the upper Peechee Member, the Peechee/Grotto contact must lie below the top of the 5b zone.

The top of the Cairn Formation is interpreted as being constrained within the 4b zone, because it is overlain at buildup margin surface sections by 30 to 40 m of Perdrix (Mt. Turbulent) or Peechee (Fatigue Mountain) strata, in turn overlain by basal beds (zone 5a) of the Mount Hawk Formation (Fig. 13). The zonal assignment for the top of the Peechee

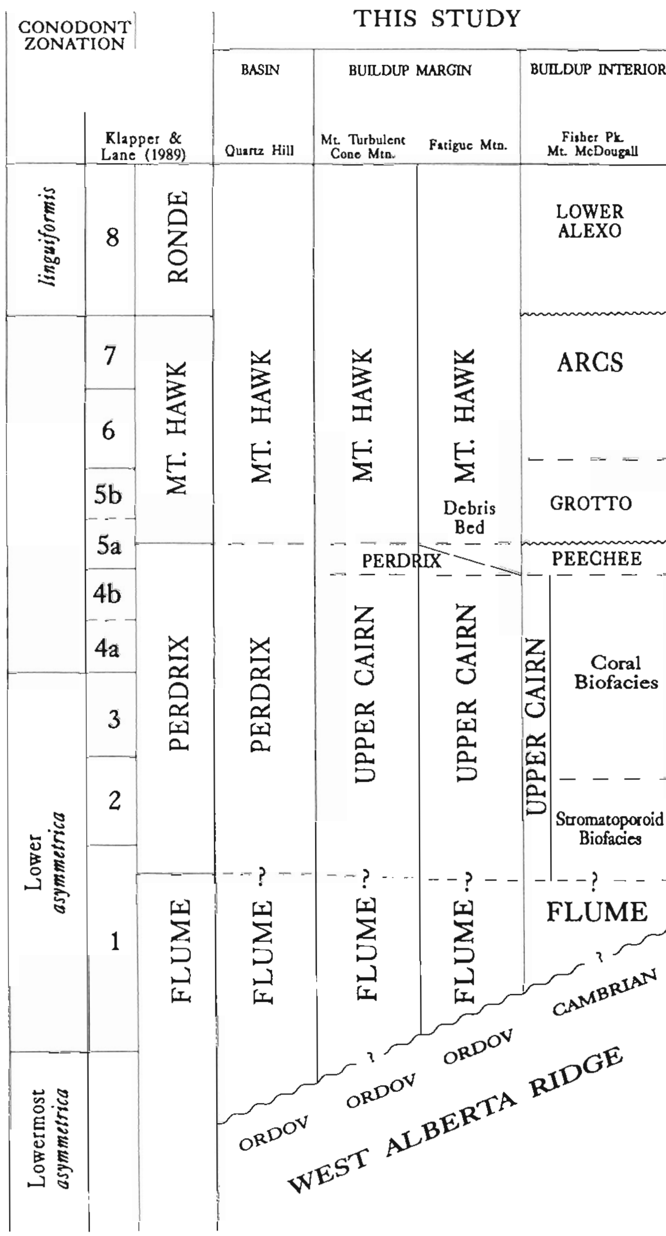

Figure 13. Frasnian conodont biostratigraphy and preliminary correlation with Upper Devonian lithostratigraphic units, western Alberta.

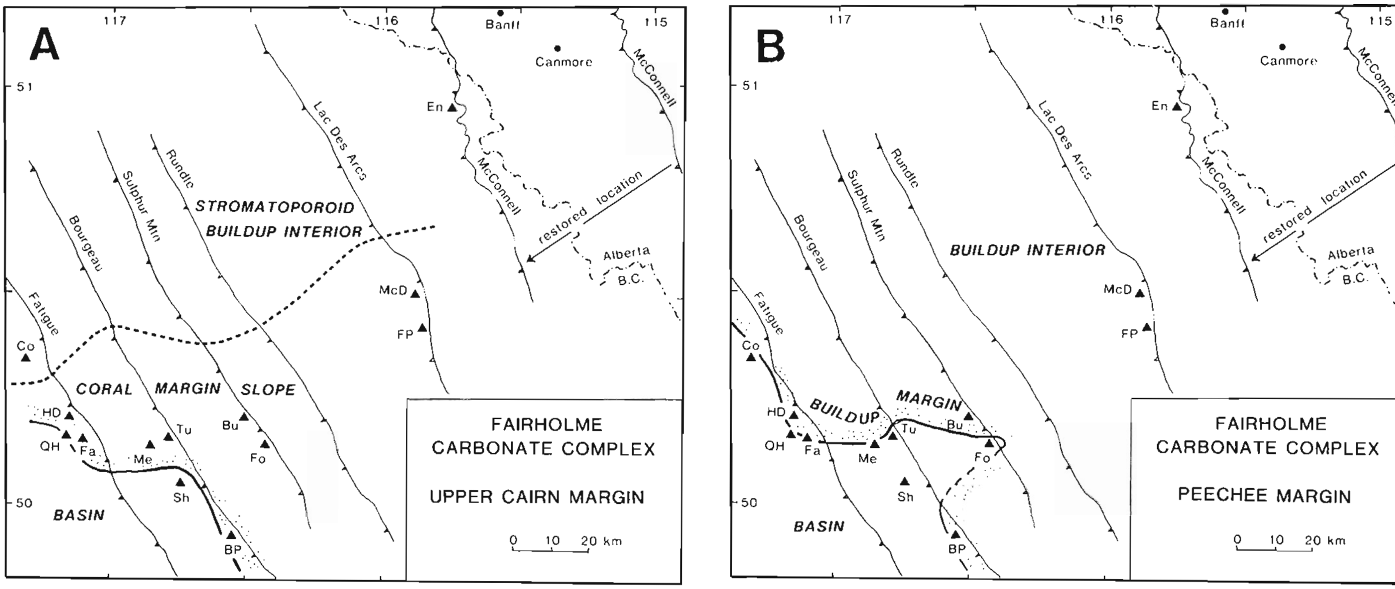

Figure 14. Paleogeography of the western margin of the Fairholme Reef Complex.
A. Early Frasnian (Upper Cairn member) paleogeography.
B. Mid-Frasnian (Peechee Member) paleogeography.
The thrust faults have been palinspastically restored, based on Price and Fermor's (1985) restoration. Only the present location of the McConnell thrust fault is shown. Co = Copper Mtn., En = End Mtn., Fa = Fatigue Mtn., FP = Fisher Peak, Fo = Fortress Mtn., BP = Mt. Black Prince, Bu = Mt. Buller, HD = Mt. Howard Douglas, McD = Mt. McDougall, Me = Mt. Mercer, Sh = Mt. Shark, QH = Quartz Hill, Tu = Turbulent Mt.

Member is uncertain, but the stratigraphic position of debris beds at Ancient Wall indicates that the basal part of the Mount Hawk can be correlated with the top of the Peechee Member (Mountjoy, 1980).

DISCUSSION

Detailed litho- and biostratigraphy of the western margin of the Fairholme Reef Complex reveals surprising changes in reef communities and margin configuration.

Biostratigraphic data show that the Flume carbonate platform started to onlap the western slope of the West Alberta Ridge as early as latest Givetian (Lowermost *asymmetrica* Zone). Carbonate production on the Flume platform was halted by a transgressive event that flooded low-lying regions with dysaerobic water, whereas elevated areas were either not affected at all or responded by a temporary facies change to brachiopod/crinoid mudstones. These elevated areas were sites of reef inception while the low-lying areas were covered with black anoxic sediment of the Perdrix Formation. A prominent pre-Devonian topographic high existed around Fatigue Mountain, influencing the configuration of the buildup margin throughout the growth of the carbonate complex (Fig. 14).

An unexpected result of this study was the identification of the change in biofacies from a typical stromatoporoid/*Amphipora* community in the lower reef phase (lower part of the Upper Cairn member) to a coral dominated community above (upper part of the Upper Cairn member), and a return to a stromatoporoid/*Amphipora* biofacies in the highest reef phase (Peechee Member). The change to a coral biofacies affected a broad northeasterly trending area and is associated with a northeastward backstepping of the coral biofacies (Fig. 14A). What caused the appearance of a new biofacies and its northeasterly retreat? The authors speculate that an increase of nutrient supply, due to upwelling, provided a new ecological niche for corals. A further gradual increase in nutrients allowed additional life forms, such as bioeroders, to enter the new habitat. Gradually, bioerosion surpassed bioconstruction preventing further accumulation of the coral biofacies in the west. However, bioconstruction continued in the eastern area where the nutrient supply may have remained at a lower level. The relationship between nutrient availability and reefs has been well documented for modern reefs by Hallock (1988). A return to nutrient-poor conditions during Peechee time re-established the habitat for the low-diversity reef community of stromatoporoids and amphiporids. The rapid southwestward progradation of the Peechee was probably caused by a decreasing rate of relative sea-level rise.

The new configuration of the buildup margin (Fig. 14B) was determined by a relatively steep slope to a rapidly deepening basin to the west. Anoxic sediments of the Perdrix, reef-derived debris beds, and oxygenated limestone and shale of the Mt. Hawk Formation had only partly infilled the basin, because the basin was located on the lee-side of the Frasnian reef domain. The reefs had effectively blocked the westward transport of Mt. Hawk sediments, the major infill component within and east of the reef domain. The water depth in this western basin probably exceeded 300 m since westerly derived siliciclastics of the Sassenach Formation completely infilled the basin in early Famennian time, and are up to 285 m thick just off the buildup margin near Fatigue Mountain. One could speculate that this massive influx of detrital material was the response to an early pulse of the Antler Orogeny.

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Preliminary results and interpretations of Early Permian cyclic shelf sedimentation, west-central Ellesmere Island, Arctic Archipelago

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Abstract

Eight stratigraphic sections of Asselian and Sakmarian (Lower Permian) strata were measured on west-central Ellesmere Island. These rocks belong to the Canyon Fiord, Belcher Channel, Antoinette, Mount Bayley, Tanquary, and Nansen formations. Eight different types of high-order (100 000 to 250 000 years) cycles were identified: 1) conglomerate, 2) sandstone-grainstone, 3) sandstone, 4) grainstone, 5) grainstone-buildup, 6) packstone-buildup, 7) packstone, and 8) gypsum. Three stratigraphic packages are recognized: a pre-evaporite (early to middle Asselian), a syn-evaporite (middle Asselian? to middle Sakmarian) and a post-evaporite (middle to late Sakmarian). The cyclicity in the syn-evaporite package displays both glacio-eustatic and tectonic affinities; cycles record deposition during an active phase of rifting. Cycles in the pre- and post-evaporite packages are only of glacio-eustatic affinity; they record deposition during more passive phases of subsidence.

Résumé

Huit sections stratigraphiques de couches asséliennes et sakmariennes (Permien inférieur) ont été mesurées dans la partie occidentale du centre de l'île d'Ellesmere. Ces roches appartiennent aux formations de Canyon Fiord, de Belcher, Channel, d'Antoinette, de Mount Bayley, de Tanquary et de Nansen. Huit types différents de cycles d'ordre élevé (100 000 à 250 000 ans) ont été identifiés: 1) conglomérat, 2) grès-grainstone, 3) grès, 4) grainstone, 5) grainstone-récif, 6) packstone-récif, 7) packstone, et 8) gypse. Trois assemblages stratigraphiques sont reconnus: un assemblage pré-évaporitique (Assélien inférieur à moyen ?), un assemblage syn-évaporitique (Assélien moyen ? à Sakmarien moyen ?), et un assemblage post-évaporitique (Sakmarien moyen à supérieur). Le caractère cyclique des assemblages pré- et post-évaporitiques témoigne à la fois des affinités glacio-eustatiques et tectoniques; les cycles se sont accumulés lors d'une phase active de distension (riftin). Les cycles de l'assemblage post-évaporitiques ne montrent qu'une affinité glacio-eustatique; ils se sont formés lors d'une phase de subsidence plus passive.

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INTRODUCTION

Asselian and Sakmarian (Lower Permian) rocks outcrop over a large area in west-central Ellesmere Island. Nearly 800 m of clastic rocks, carbonate rocks, and evaporites occur in mountainous exposures on Fosheim and Hamilton peninsulas, and east and west of Tanquary Fiord. These deposits belong to the Canyon Fiord, Belcher Channel, Antoinette, Mount Bayley, Tanquary, and Nansen formations, which are part of the Sverdrup Basin succession (Thorsteinsson, 1974). Sediments in all these formations are stacked in a series of recurring packages, or cycles, each representing an episode of sea-level rise and fall that lasted for 100 000 to 250 000 years (5th- to 6th-order cycles). These high-order cycles are superimposed on a much longer transgressive-regressive cycle that encompasses most of the Asselian-Sakmarian time interval (5 to 10 Ma duration; 3rd-order) (Beauchamp et al., 1989b). Between these two extremes an intermediate level of cyclicity probably exists, recording periodic sedimentation every 1 to 2 Ma (4th-order). These 4th-order cycles are very difficult to observe because of the 5th- to 6th-order overprint. They are, however, believed by some to be synchronous and worldwide, and to parallel major biostratigraphic zonal boundaries (Ross and Ross, 1985). For more than two decades, the origin of 2nd- to 6th-order cycles in the sedimentary record has been, and still is, a hot topic that often triggers passionate debates. A whole spectrum of interpretations has been proposed with glacio-eustacy and tectonism probably representing end members. No consensus has been reached, and there is none in sight.

This study was undertaken in order to understand the various mechanisms that led to 3rd- to 6th-order, Early Permian, cyclic sedimention in west-central Ellesmere Island, in an effort to evaluate which order of cyclicity may be useful for intra- and extrabasinal correlations. West-central Ellesmere Island was chosen as the study area (Fig. 1), because this region is known to have been tectonically active, to a variable extent, throughout the Asselian and Sakmarian (Beauchamp et al., 1989a,b). In this area, Lower Permian rocks comprise a complex, interfingering succession of clastic rocks, carbonates, and evaporites which were deposited in a narrow subbasin, the Fosheim-Hamilton subbasin, at the margin of the main Sverdrup Basin. Parallel to the understanding of cyclicity, this project will facilitate an interpretation of the depositional environments and paleogeography of the Fosheim-Hamilton subbasin. The data for this study came from eight stratigraphic sections, and hundreds of petrographic and biostratigraphic samples acquired during the summer of 1990. This study constitutes Morin's M.Sc. research project at the University of Ottawa. The present paper is a summary of field observations, with some preliminary interpretations. More definitive conclusions will be reached with the incorporation of new petrographic and biostratigraphic data.

GEOLOGICAL SETTING

The formations studied are part of Sverdrup Basin, an intracontinental rift basin containing more than 12 km of Lower Carboniferous to mid-Tertiary sedimentary and minor volcanic rocks (Trettin, 1989; Davies and Nassichuk, in press). Rifting and fault controlled subsidence in Sverdrup Basin, took place from the Viséan (Early Carboniferous) to the earliest Kungurian (late Early Permian), following which a regime of passive subsidence was established until the end of the Permian and beyond (Beauchamp et al., 1989a). Three main phases of active rifting occurred during the Early Carboniferous (Serpukhovian), Late Carboniferous (Bashkirian and Moscovian) and the Early Permian (Asselian and Sakmarian), as suggested by abundant syntectonic deposits in rocks of these ages. The syn-rift paleogeography of Sverdrup Basin basin must have been characterized by a series of parallel subbasins, separated by linear highs reflecting a tectonic landscape dominated by extensional half-grabens.

Several lines of evidence point to the existence of such a subbasin at the margin of the main Sverdrup Basin, in west-central Ellesmere Island. The best evidence comes from the presence of up to 400 m of Lower Permian subaqueous evaporites (upper Antoinette and Mount Bayley formations), over an area of more than 7 500 km^2, immediately south of an area where penecontemporaneous open marine carbonates (Nansen Formation) and deeper water mudrocks (Hare Fiord Formation) are known to occur. Other evidence includes the presence of westerly derived conglomerates and deposits with half-graben affinities in the Upper Carboniferous Canyon Fiord Formation (Thériault and Beauchamp, 1991), and the distribution of Artinskian strata on Fosheim and Hamilton peninsulas. These strata thin in southwesterly, northeasterly and, possibly, northwesterly directions, reflecting closure of the subbasin (Scott et al., 1991). This subbasin, informally called the Fosheim-Hamilton subbasin, clearly must have been isolated from the main Sverdrup Basin in Asselian-Sakmarian time in order for the Mount Bayley evaporites to precipitate. This very fact points to the existence of an undefined linear high between the restricted Fosheim-Hamilton subbasin and the main, open marine Sverdrup Basin. Such a linear high is believed to have existed in the environs of Elmerson Peninsula and extended for some distance in a southwesterly direction. This high is here informally referred to as the "Elmerson High" (see also Beauchamp et al., 1991; Scott et al., 1991; and Thériault and Beauchamp, 1991).

The upper Paleozoic succession of Sverdrup Basin is characterized by seven, long term (2nd- and 3rd-order), transgressive-regressive sequences, bounded at the basin margin by major unconformities that pass basinward into their equivalent conformities (Beauchamp et al., 1989b). One of these long term sequences is recorded in Asselian and Sakmarian (Lower Permian) deposits. The lower sequence boundary is characterized by an intraformational unconformity near the Carboniferous/Permian boundary and within the Canyon Fiord, Belcher Channel, Antoinette, and Nansen formations; the upper sequence boundary is represented by an unconformity between the Artinskian "lower green unit" (Scott et al., 1991) and strata of the Canyon Fiord, Belcher Channel, Tanquary, or Nansen formations (Beauchamp et al., 1989b). This study focuses on rocks of the Asselian-Sakmarian sequence. All but one (McKinley Bay) stratigraphic sections were measured in the Fosheim-Hamilton subbasin. The measured and sampled formations include: Canyon Fiord, Belcher Channel, Antoinette, Mount Bayley, Tanquary, and Nansen.

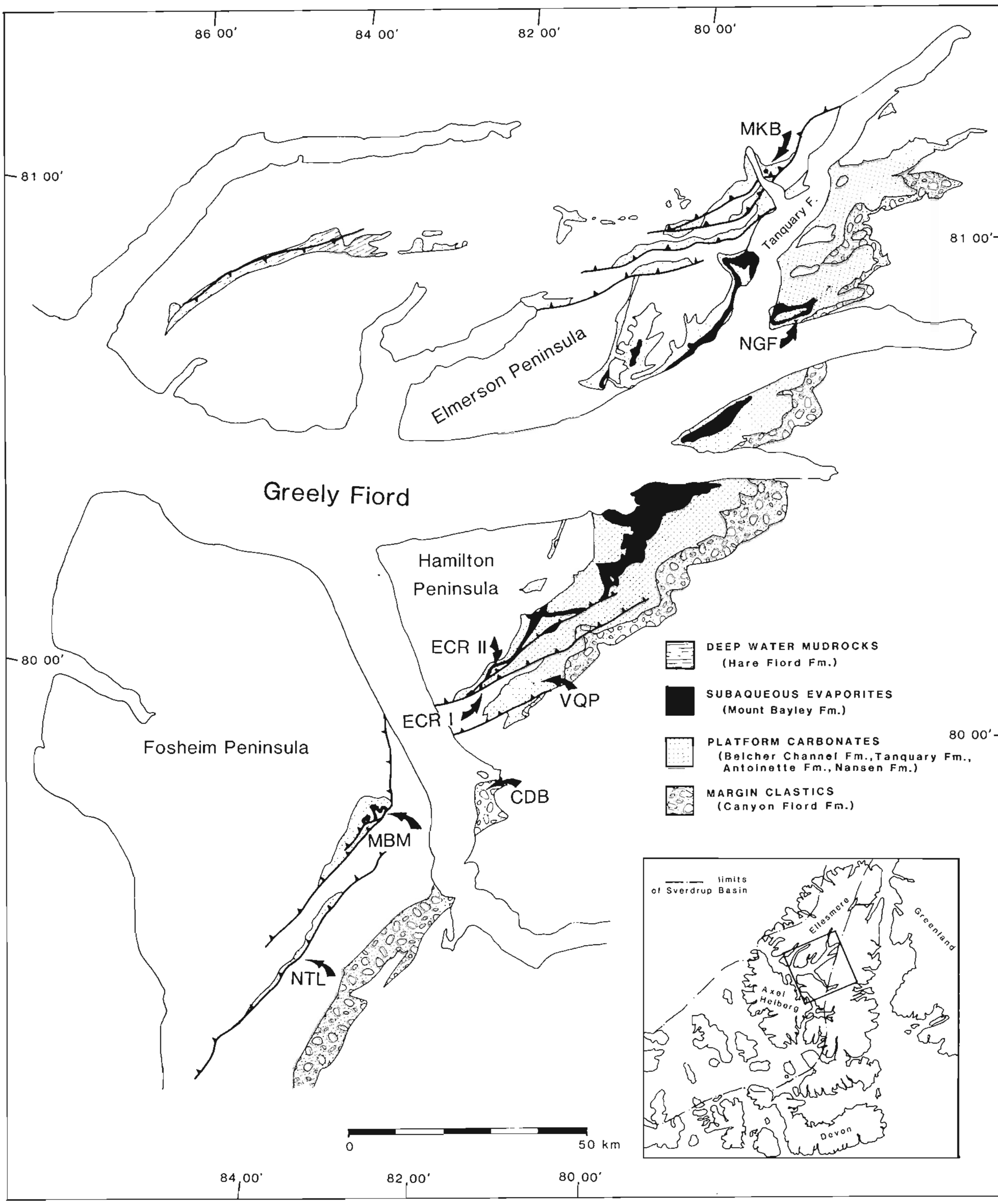

Figure 1. Map of west-central Ellesmere Island, showing section locations and the outcrop distribution of Lower Permian formations. Measured sections are: "Notch Lake" (NTL), Mount Bridgman (MBM), Caledonian Bay (CDB), East Cape River I (ECR I), East Cape River II (ECR II), "Vache-qui-pleure" (VQP), North Greely Fiord (NGF), and McKinley Bay (MKB). (Geology modified from GSC maps 1308A, 1311A, 1309A, 1306A and 1348A.)

SUMMARY OF FIELD OBSERVATIONS

Ideal cycles

The lithological succession observed in the measured sections is very complex, owing to the large number of high-order cycles. No two cycles are identical, and each contains a vast spectrum of facies. In order to facilitate section descriptions, the measured succession has been broken into eight different types of cycles (Fig. 2). The succession of type 1 to type 8 cycles shown in Figure 2 roughly corresponds to a proximal-to-distal section across the axis of the depositional system. The cycles were defined on the basis of their megascopic features: thickness, composition, and recurrence. Petrographic information based on the study of thin sections will be added to the cycle definitions later. The cycles were named after their main lithological constituents: 1) conglomerate, 2) sandstone-grainstone, 3) sandstone, 4) grainstone, 5) grainstone-buildup, 6) packstone-buildup, 7) packstone, and 8) gypsum. The distribution of the various cycle types is shown in Figures 3 and 4.

The *conglomerate cycle* (1) consists of an assemblage of conglomerate, sandstone and grainstone. The conglomerate unit usually displays an erosional base; it is composed of polymict rounded cobbles in a sand-sized reddish matrix. The conglomerate is overlain by a pale grey, quartz-rich sandstone displaying large-scale, low-angle crossbedding. A light coloured, fine grained grainstone usually caps the sandstone.

The *sandstone-grainstone cycle* (2) is composed of sandstone overlain by grainstone of variable thickness. The sandstone consists of light greenish grey, fine grained sandstone with small and medium scale crossbedding. Massive fine grained grainstone, commonly capped by bioturbated, greenish or reddish mudrock, occurs at the top.

The *sandstone cycle* (3) is characterized by variable thicknesses of light greenish to reddish grey, fine to medium grained sandstone, underlain and overlain by layers of greenish to reddish siltstone and shale. The sandstone displays large and medium scale crossbedding and its top is commonly bioturbated.

The *grainstone cycle* (4) comprises numerous undifferentiated grainstone types. The ideal cycle consists of grey mudrock overlain by packstone that passes progressively upward into grainstone. The grainstone is characteristically medium grained and massive; large scale, low-angle crossbedding sometimes occur. Colonial rugose corals of various sizes occur at the top of the cycle.

The *grainstone-buildup cycle* (5) is very similar to the grainstone cycle with the addition of Palaeoaplysina-phylloid algal buildups of various sizes and shapes.

The *packstone-buildup cycle* (6) consists of grey mudstone at the base, followed by dark grey, nodular wackestone that is generaly thin bedded and bioturbated. The wackestone is overlain by dark grey, commonly massive packstone containing few solitary corals and fusulinaceans. Metre sized colonial Rugosa are common in packstones beneath the buildup. The buildup is overlain by relatively thin wackestone followed by mudrock.

The *packstone cycle* (7) comprises thick bedded, dark grey lime mudstone and fusulinacean-rich wackestone at the base. The bulk of the cycle is composed of dark grey to black, organic-matter-rich, massive fusulinacean packstone with some large colonial Rugosa in the upper part. Wackestone followed by mudstone mark the top of the cycle.

The *gypsum cycle* (8) consists of greyish to greenish mudstone interstratified with fossil-rich to unfossiliferous, dolomitized packstone or grainstone, and overlain by a thick gypsum or anhydrite unit.

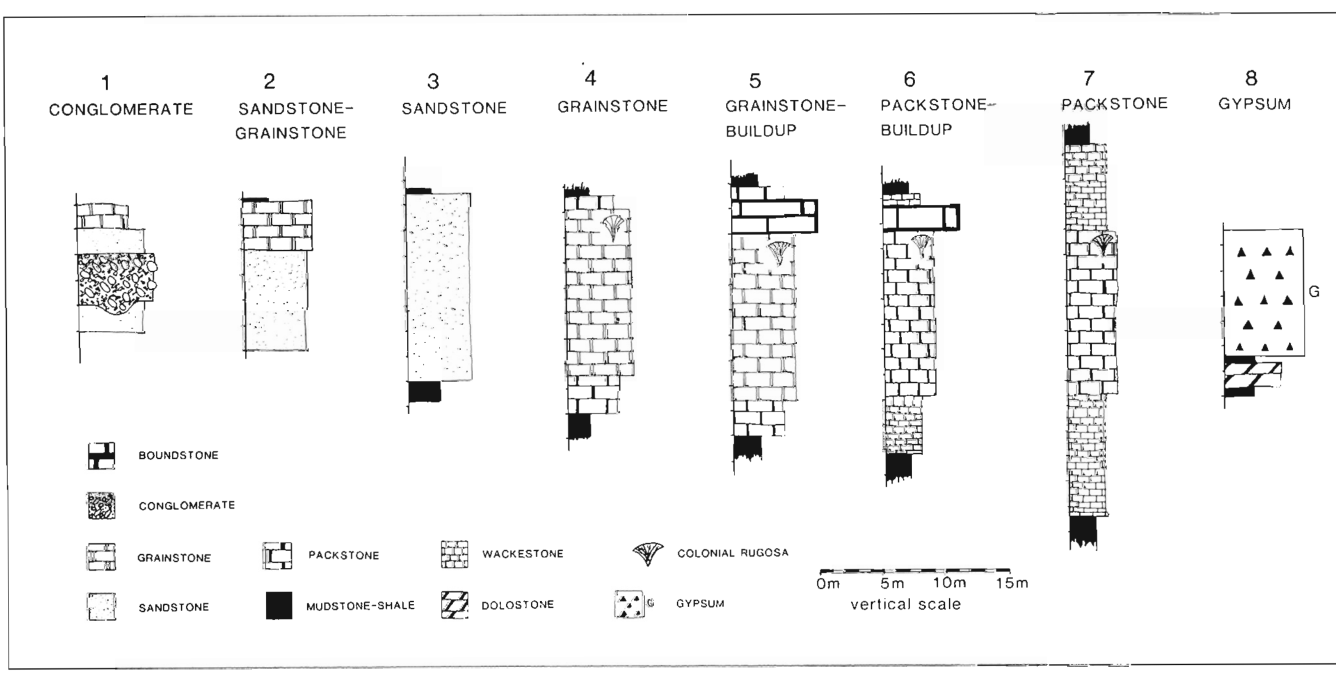

Figure 2. Ideal cycle types observed in the Lower Permian succession on west-central Ellesmere Island.

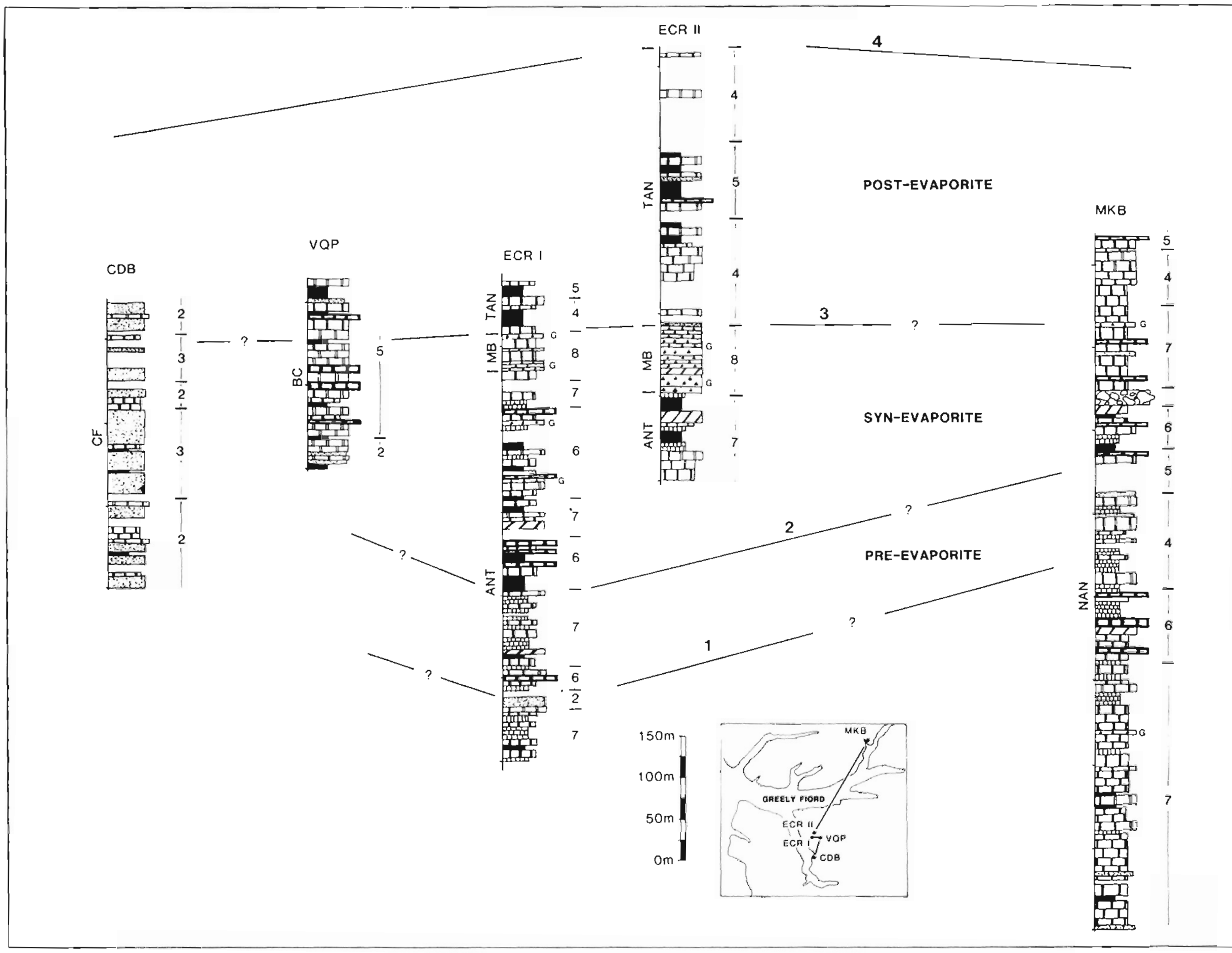

Figure 3. Cross-section showing ideal cycle types at Caledonian Bay (CDB), "Vache-qui-pleure" (VQP), East Cape River I (ECR I), East Cape River II (ECR II), and McKinley Bay (MKB). Lithology symbols as in Figure 2. Correlation lines 1 to 4 are discussed in the text. Formations are Canyon Fiord (CF), Belcher Channel (BC), Antoinette (ANT), Mount Bayley (MB), Tanquary (TAN), and Nansen (NAN).

Section descriptions

"Notch Lake" section (NTL)

The "Notch Lake" section (NTL) is located at the southwestern limit of the study area (Fig. 1). The measured section is 375 m thick, but the base is marked by a thrust fault and the top by a normal fault. Approximately twenty high-order transgressive-regressive cycles were identified (Fig. 4). The base of the section is characterized by a series of conglomerate cycles (type 1) passing upward for the next 100 m into sandstone-grainstone cycles (type 2), followed by a succession of sandstone cycles (type 3) up to 275 m, above which grainstone (type 4) and grainstone-buildup (type 5) cycles dominate. The top of the section is represented by sandstone cycles (type 3).

Thorsteinsson (1974) assigned the Lower Permian carbonate succession that outcrops in the hanging wall of Vesle Fiord Thrust in the Sawtooth Range to the Tanquary Formation. In doing so, he assumed that these carbonates are underlain in the subsurface by evaporites of the Mount Bayley Formation, even though no evaporites are exposed in the hanging wall of Vesle Fiord Thrust on Fosheim Peninsula. We are of the opinion that this is a very unlikely possibility, and that the carbonate succession in question should be assigned to the Belcher Channel Formation instead. The facies in this succession are far more characteristic of the Belcher Channel Formation with its predominant very shallow-water cycles (conglomerate, sandstone-grainstone and sandstone). Also, the "Notch Lake" section (375 m thick) is thicker than the thickest known Tanquary exposures elsewhere, in areas closer to the Fosheim-Hamilton subbasin depocentre (Mount Bridgman, East Cape River, North Greely Fiord). If anything, the Tanquary Formation should thin substantially in a southwesterly direction toward the Bay Fiord area, as do the younger Artinskian strata (Scott et al., 1991). For both lithological and stratigraphic reasons, these rocks would be better included in the Belcher Channel Formation.

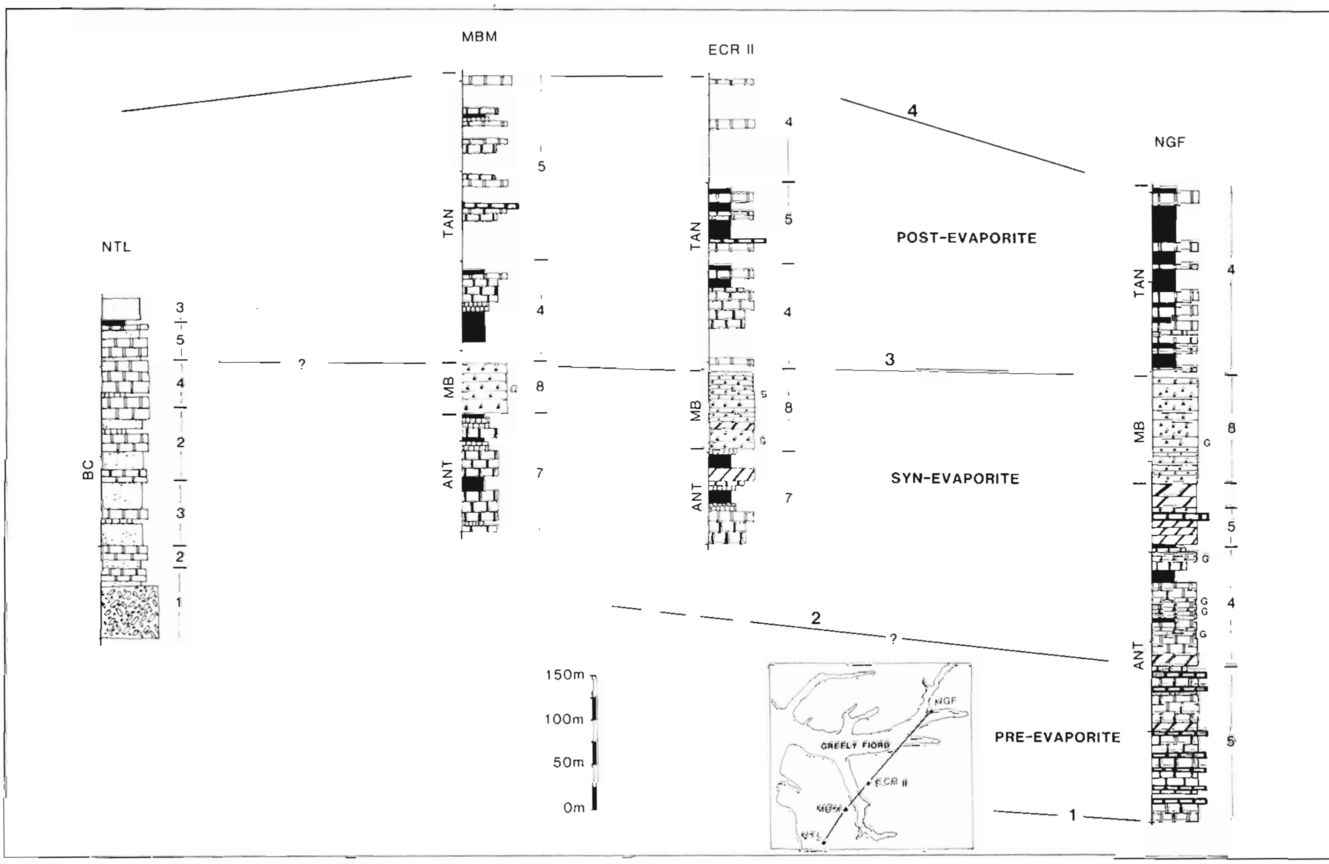

Figure 4. Cross-section showing ideal cycle types at "Notch Lake" (NTL), Mount Bridgman (MBM), East Cape River II (ECR II), and North Greely Fiord (NGF). Lithology symbols as in Figure 2. Correlation lines 1 to 4 are discussed in the text. Formations are Belcher Channel (BC), Antoinette (ANT), Mount Bayley (MB), and Tanquary (TAN).

Caledonian Bay section (CDB)

The Caledonian Bay section is an incomplete section of the Canyon Fiord Formation, 525 m thick (Fig. 3). Twenty-five sandstone-grainstone (type 2) and grainstone (type 4) cycles were counted. The entire succession is almost exclusively composed of sandstone, slightly argillaceous in the lowest 100 m, but much purer above, with common large- and small-scale crosslamination. At 150 m above the base there is a carbonate bed that appears to be heavily calichified. The upper 100 m of the section are characterized by strata forming cycles, separated by broad recessive intervals.

"Vache-qui-pleure" section (VQP)

This section was measured in a narrow creek, 23 km east of Canyon Fiord; it is informally referred to as the "Vache-qui-pleure" section, because of a large glacier tongue that feeds the creek upstream. The section is 225 m thick, and includes rocks of the Belcher Channel Formation. About 11 high-order cycles are recognized. The basal part of the section is characterized by sandstone-grainstone cycles (type 2), whereas the remainder contains grainstone-buildup cycles (type 5). Cycle thickness is very variable. Above 150 m all cycles are characterized by the occurrence of significant units of recessive shale at their bases. A grainstone bed at 155 m is noteworthy for its content of abundant, large, spherical fusulinaceans resembling *Pseudoschwagerina*.

East Cape River I section (ECR I)

The East Cape River I and East Cape River II sections are little more than 100 m apart. A fault separates the two sections. Originally interpreted as a normal fault (Thorsteinsson, 1974), this structure is now believed to be the continuation of an unnamed thrust fault that has been recognized farther to the northeast (Fig. 1). This interpretation is based on kinematic indicators along the fault plane and on the occurrence of very different evaporite assemblages across it. For example, 77 m of typical and continuous Mount Bayley evaporites are present in the East Cape River II section, whereas only a few evaporite units, separated by important carbonate units, occur at the East Cape River I section (Fig. 3). The Mount Bayley Formation cannot be mapped as such, therefore, at East Cape River I, the entire Antoinette/Mount Bayley/Tanquary succession would be better included in the correlative Belcher Channel Formation. East Cape River I clearly lies near the lateral transition between these two correlative successions. The original formational assignments are temporarily retained in Figure 3.

Thirty-one high-order cycles, spread over 580 m, occur in the East Cape River I section. The first 470 m of the section are assigned to the Antoinette Formation, the overlying 50 m to the Mount Bayley Formation, and the remaining 60 m to the Tanquary Formation. The lowest part of the section comprises packstone cycles (type 7). Of note is the occurrence, at 60 m, of conglomerate with rounded limestone debris, overlain by a cross-stratified sandstone cycle (type 2). From that level up to 200 m, the cycles are of more or less the same thickness, whereas a much wider range of cycle thicknesses was observed in the interval between 200 and 470 m, where both packstone-buildup and packstone cycles (types 6, 7) were observed. A bed characterized by massive Microcodium, associated with some caliche development, ocurs at 230 m. The lowest levels of gypsum and anhydrite occur at 350 m, whereas a few gypsum cycles (type 8) are present above 470 m. Grainstone and grainstone-buildup cycles (types 4, 5), characterized by very thick recessive units at their bases, characterize the remainder of the section, which is assigned to the Tanquary Formation.

East Cape River II section (ECR II)

The East Cape River II section is 520 m thick and contains 22 high-order cycles (Figs. 3, 4). The Antoinette Formation, which is represented by the lowest 100 m, comprises typical packstone cycles (type 7). Gypsum cycles (type 8) characterize the overlying Mount Bayley Formation, which is 77 m thick. The remaining 240 m constitute a complete section of the Tanquary Formation, which is represented by grainstone and grainstone-buildup cycles (types 4, 5), each characterized by thick basal units of recessive shale.

Mount Bridgman section (MBM)

The Mount Bridgman section is, in many respects, very similar to the East Cape River II section (Fig. 4). About 24 high-order cycles were recognized in the Antoinette, Mount Bayley, and Tanquary formations. Packstone cycles (type 6) characterize the lower 130 m, which belong to the Antoinette Formation. Gypsum cycles (type 8) of the Mount Bayley Formation occur between 130 and 190 m. The complete Tanquary Formation is represented by grainstone and grainstone-buildup cycles, characterized by thick, recessive, basal shale units. Pseudoschwagerina-like fusulinaceans are abundant in the first cycle above the Mount Bayley Formation. The presence of a fault that terminates at the base of the Mount Bayley Formation is worthy of mention; it may be a synsedimentary structure.

North Greely Fiord section (NGF)

The North Greely Fiord section is the type section of the Antoinette, Mount Bayley, and Tanquary formations (Fig. 4). Approximately 40 high-order cycles were identified in 700 m. The Antoinette Formation can be divided into two parts: a lower part (0-275 m) characterized by grainstone-buildup cycles (type 5) of similar composition and thickness; and an upper part (275-375 m) of similar grainstone and grainstone-buildup cycles (types 4, 5) associated with gypsum units of variable thickness. The uppermost 30 m of Antoinette strata comprise a series of thin cycles interpreted as the product of tidal flat sedimentation.

The Mount Bayley Formation is 125 m thick and is represented by a series of gypsum cycles (type 8). A synsedimentary normal fault that terminates in the Mount Bayley Formation was observed in the vicinity of the measured section (Fig. 5). Grainstone cycles (type 4), characterized by thick recessive shale units at the base, occur in the 200 m thick Tanquary Formation. Abundant Pseudoschwagerina-like fusulinaceans occur in the first cycle at the base of this formation.

McKinley Bay section (MKB)

An incomplete section of the Nansen Formation, 840 m thick, was measured at McKinley Bay (Fig. 3). As many as 55 high-order cycles were identified. The basal part of the section comprises packstone cycles (type 7) passing upward into packstone-buildup cycles (type 6). A thin unit of gypsum found at 240 m above the base may be correlative with the Upper Carboniferous Otto Fiord Formation. A thick succession of grainstone cycles (type 4) follows, overlain by a series of grainstone-buildup (type 5), packstone-buildup (type 6) and packstone (type 7) cycles. At 630 m there is a spectacular olistostrome with blocks that are more than 5 m in diameter and of the same composition as the immediately underlying rocks. Some blocks are folded and injected by an unidentified matrix. The overlying 50 m consist of several thin packstone cycles (type 7) characterized by a black shale basal unit. A unit of gypsum and anhydrite at 730 m is likely contemporaneous with the Mount Bayley Formation farther south. The upper part of the measured section comprises a series of relatively thick grainstone and grainstone-buildup cycles (types 4, 5).

PRELIMINARY INTERPRETATION

Without adequate biostratigraphic and petrographic data, only speculations can be advanced regarding correlations between the sections, the various mechanisms that led to cyclicity, and the paleogeography of the study area. The following is a brief introduction to some of the concepts that will be dealt with later in this project when data from fossil and thin section analyses become available. Conclusions reached from the field study should be viewed as working hypotheses to be tested later.

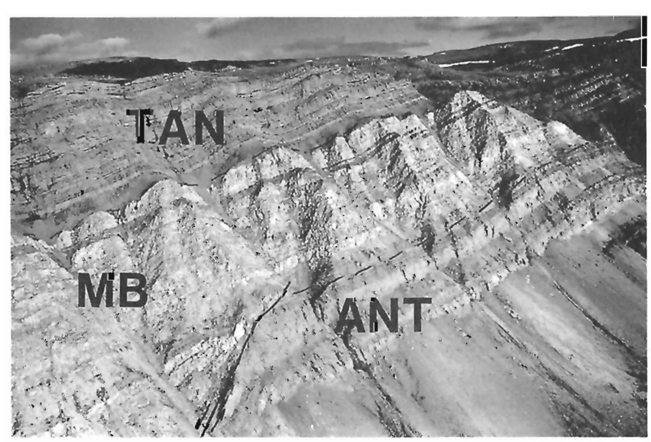

Figure 5. Synsedimentary normal fault juxtaposing evaporites of the Mount Bayley Formation (MB) and carbonates of the Antoinette Formation (ANT), approximately 1 km west of North Greely Fiord section (NGF in Figure 1).

Correlations

Only four correlation lines were drawn on the cross-sections in Figures 3 and 4; two of these lines are considered to be relatively valid (correlations 3 and 4), the other two are far less certain. These correlation lines separate the succession into three assemblages, characterized by different tectono-sedimentary signatures: a pre-, a syn-, and a post-evaporite assemblage.

The pre-evaporite assemblage lies between correlation lines 1 and 2. Correlation 1 is believed to closely correspond with the boundary between the long term sequences 2 and 3 of Beauchamp et al. (1989b). In many areas of Sverdrup Basin, this boundary is an unconformity that straddles the Carboniferous/Permian boundary. No unconformity was observed in the study area at that level. Nevertheless, the sequence boundary is believed to occur near a conglomerate and sandstone cycle (type 2) at the East Cape River I section, and near the base of shallow water grainstone cycles (type 4) at McKinley Bay (Fig. 3). The thick conglomerate at the base of the "Notch Lake" section could correlate with this sequence boundary.

Correlation line 2 marks a level (observed in several sections) at which there is a vertical change from a succession of cycles of similar composition and thickness (pre-evaporite assemblage), to a succession in which the cycles vary greatly in composition and thickness (syn-evaporite assemblage). The pre-evaporite assemblage comprises eleven evaporite-free cycles.

The syn-evaporite assemblage lies between correlation lines 2 and 3. Correlation line 3 is taken as the top of the Mount Bayley Formation. Wallace and Beauchamp (1990) expressed the opinion that this surface probably represents a time line, since an equal number of cycles were counted in the overlying Tanquary Formation at various localities, and because the same conodont zone (P5 of Henderson, 1988) was found in the lowest Tanquary strata at several localities. The presence of abundant *Pseudoschwagerina*-like fusulinaceans was also used to correlate the basal Tanquary with a level high in the Belcher Channel Formation at the "Vache-qui-pleure" section. The *Pseudoschwagerina*(?) level is always associated with a change in the style of cyclicity, characterized by the presence of thick, recessive intervals at the base of each cycle. This criterion was used to correlate the top contact of the Mount Bayley Formation with a horizon in the Canyon Fiord Formation at Caledonian Bay, with one in the Nansen Formation at McKinley Bay (Fig. 3), and one in the Belcher Channel Formation at "Notch Lake" (Fig. 4). The syn-evaporite assemblage thus contains a series of cycles of varying composition and thickness. Some cycles display prominent, recessive, shaly bases. The assemblage is also characterized by the presence of tidal flat sediments, caliche and *Microcodium* "horizons", and a thick olistostrome unit at McKinley Bay. All of these features indicate complex episodes of subaerial exposure and erosion that may have been related to synsedimentary tectonic movements, as suggested by the observation of a synsedimentary growth fault in the Mount Bayley Formation (Fig. 5).

The post-evaporite assemblage lies between correlation lines 3 and 4. Correlation line 4 is the contact between Sakmarian and Artinskian strata, the latter being represented by a greenish weathering unit of recessive shale and mixed clastic and carbonate rocks ("lower green unit" of Scott et al., 1990). In many places, this contact is an unconformity, which corresponds to the boundary between the long term sequences 3 and 4 of Beauchamp et al. (1989b). The post-evaporite assemblage comprises eleven high-order cycles that are relatively similar in composition and thickness and are all characterized by a thick, basal, recessive shale unit.

Cyclicity

The succession of assumed Asselian and Sakmarian age studied comprises approximately 40 high-order cycles, which is reasonably close to the number of cycles counted in the correlative succession in the Blind Fiord area (between 40 and 50; Beauchamp, 1987). Beauchamp (op. cit.) expressed the opinion that these cycles are of glacio-eustatic origin; the reasons which he invoked include: 1) the high amplitude of sea-level fluctuation involved in each cycle; 2) the existence of a similar number of cycles at this level in other parts of the world, such as the midcontinent in the United States; 3) the knowledge that a major ice sheet existed in the southern hemisphere at the same time; and 4) the existence of a similar number of glacial/interglacial cycles in coeval glacial deposits of Gondwana (Veevers and Powell, 1987). Upper Paleozoic high-order cycles, perhaps more than any other cycles in the geological record, show strong glacio-eustatic affinities, and this interpretation has been put forward by many authors (see Wilson, 1975).

In Sverdrup Basin, however, an important tectonic component appears to have been superimposed on the glacio-eustatic control. This is indicated by the highly variable thickness of individual cycles vertically and, in many cases, laterally as well. Based on the knowledge that Sverdrup Basin was in an extensional rift phase during the time interval studied, one can assume that fault-related subsidence must have played an important role in shaping individual cycles. Indeed, one can envision a setting in which subsidence rates varied greatly from time to time, both within individual fault blocks and across fault blocks. This would yield high-order cycles of varying thickness and composition. In contrast, glacio-eustatic cycles superimposed on a regime of passive thermal subsidence would result in a succession of cycles of uniform thickness and composition. As indicated above, the pattern of cycles in the study area has varied through time; one observes a syn-evaporite succession of highly irregular cycles set between far more regular cycles in both pre- and post-evaporite successions. This very fact suggests greater rift-related tectonic activity during sedimentation of the syn-evaporite succession than before and after. The observation of synsedimentary normal faults in the syn-evaporite assemblage fits this conclusion (Fig. 5). Also relevant is the inference of major syntectonic movements in the coeval succession west of Blind Fiord in southwestern Ellesmere Island (Beauchamp, 1987).

Another effect of synsedimentary rifting superimposed on high-order glacio-eustatic fluctuations is likely to be the partial to complete masking of overprinted lower-order fluctuations. In a regime of passive subsidence, high-order cycles (100 000 to 250 000 years duration) should form regular packages of 5 to 10 cycles, reflecting the next overprinted lower-order fluctuations (1-2 Ma). Theoretically, the lower half of each package should thicken upward, reflecting the low-order transgression, and the upper half should thin upward, reflecting the low-order regression. In a regime of differential fault-controlled subsidence, however, this theoretical pattern may or may not be visible because of the significant, and rapidly changing, intercyclic thickness variations. Lower-order grouping of cycles was impossible to achieve in the study area, suggesting that the effect of tectonism masked the effect of 4th-order (1-2 Ma) sea-level fluctuations. These are the fluctuations that Ross and Ross (1985) believed to be synchronous and worldwide. One exception may occur in the Tanquary Formation, in which lower-order fluctuations appear to exist (Fig. 6). This would point to a far more passive regime of subsidence for the Tanquary Formation than for the underlying units.

Figure 6. High-order carbonate cycles in the Tanquary Formation at East Cape River (ECRII in Figure 1). Thinning-upward nature of successive cycles (arrows) may reflect lower-order of cyclicity.

Paleogeography

Although this study is still in a preliminary stage, some broad generalizations can be drawn regarding the paleogeography of the study area. With the exception of the McKinley Bay section, all sections lie within the Fosheim-Hamilton subbasin, which is the northeasterly trending depression containing the Mount Bayley Formation, and older and younger upper Paleozoic sediments. At all times, the subbasin appears to have been bounded to the southwest by a high of unknown origin (referred to as the "Bay Fiord High"), to the northeast by the well documented Tanquary High (Nassichuk and Christie, 1969; Maurel, 1989), and to the northwest by a feature of unknown origin or magnitude (referred to as the "Elmerson High"). This last high is believed to exist in the subsurface of Elmerson Peninsula.

The cross-sections in Figures 3 and 4 illustrate some of the Lower Permian facies variations in the Fosheim-Hamilton subbasin. In very general terms, Figure 3 (pre- and syn-evaporite assemblages) shows sandstone-grainstone and sandstone cycles at Caledonian Bay passing laterally into a succession dominated by grainstone-buildup cycles at "Vache-qui-pleure", and thence into packstone-buildup, packstone and gypsum cycles in the East Cape River area. This south/north transition from the Caledonian Bay section to the East Cape River II section is clearly indicative of increasing water depth in a proximal to distal transition. The same sort of transition is illustrated in Figure 4 (syn-evaporite assemblage), which shows shallow water conglomerate, sandstone-grainstone, sandstone, and grainstone cycles at "Notch Lake" passing laterally into deeper water packstone and increasingly thick gypsum cycles toward North Greely Fiord. Clearly, the syn-evaporite and probably the pre-evaporite depocentre must have been located near the north shore of Greely Fiord. The discovery of large carbonate buildups in the Mount Bayley Formation immediately west of the mouth of Tanquary Fiord (Beauchamp et al., 1991) provides additional evidence as to the maximum water depth in that area. The presumably coeval succession in the McKinley Bay section is much thinner and likely represents sedimentation outside the Fosheim-Hamilton subbasin, probably on the other side of the "Elmerson High".

The geometry of the Fosheim-Hamilton subbasin in pre- and syn-evaporite time (early Asselian to middle Sakmarian) appears to be asymmetrical, with its thickest deposits lying against its northwest margin. The subbasin was probably formed much earlier, in Late Carboniferous time, as one or a series of extensional half-grabens bounded by a submerged rift shoulder ("Elmerson High") to the northwest. Based on available data, one can surmise that evaporite sedimentation was caused by a renewal of rifting in the Fosheim-Hamilton subbasin, which may have led to the emergence of its northwestern shoulder ("Elmerson High"), and to the onset of restricted evaporitic conditions to the south. It is clear however, that episodic sea-level rises associated with each high-order cycle replenished the Fosheim-Hamilton subbasin, resulting in the reciprocal deposition of carbonate, evaporite, and clastic sediments in a cyclic fashion (see also Wallace and Beauchamp, 1990).

Rifting activity clearly decreased during deposition of the post-evaporite assemblage, while a regime of more passive subsidence was established. At that time, the depocentre of the Fosheim-Hamilton subbasin appears to have shifted from the environs of north Greely Fiord to the East Cape River area, as suggested by thickness variations between these two areas (Fig. 4). This is no surprise, considering that, in Artinskian time, the depocentre was located near Mount Bridgman (Scott et al., 1991), and near "Notch Lake" in Late Permian time. These observations suggest that the Fosheim-Hamilton depocentre migrated in a southwesterly direction, from an asymmetrical configuration in Early Permian and possibly Late Carboniferous time while rifting occurred, to a more symmetrical configuration in the remainder of the Permian, while a regime of passive thermal subsidence was established.

CONCLUSIONS

The Asselian and Sakmarian succession of Fosheim and Hamilton peninsulas is characterized by more than 40 high-order cycles, which display important facies variations across the depositional axis of the Fosheim-Hamilton subbasin.

These cycles are grouped into three assemblages: a pre-evaporite, a syn-evaporite, and a post-evaporite. The syn-evaporite assemblage was deposited during an active phase of rifting in the Fosheim-Hamilton subbasin, whereas the other two, especially the post-evaporite assemblage, were deposited during a more passive phase of thermal subsidence.

ACKNOWLEDGMENTS

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Field investigations of Artinksian (Lower Permian) strata, west-central Ellesmere Island, Arctic Archipelago

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Abstract

Seven stratigraphic sections of Artinskian (Lower Permian) strata were measured on west-central Ellesmere Island. Two major units characterize the Artinskian sequence: 1) a "lower green unit" of recessive shales and mixed clastics and carbonates; and 2) an "upper yellow unit" of resistant, fossiliferous limestones and minor sandstones. These rocks were deposited in a large subbasin at the margin of Sverdrup Basin. The maximum thickness (478 m) was observed at Mount Bridgman. From there, the succession thins in both southwesterly and northeasterly directions, reflecting closure of the subbasin.

Résumé

Sept sections stratigraphiques de lits artinskiens (Permien inférieur) ont été mesurées dans la partie occidentale du centre de l'île d'Ellesmere. La séquence artinskienne est représentée par deux unités importantes: 1) une « unité verte inférieure » récessive de schistes argileux et de roches carbonatées et clastiques mixtes; 2) une «unité jaune supérieure» résistante de calcaires fossilifères et de grès secondaires. Ces roches se sont formées dans un grand sous-bassin à la marge du bassin de Sverdrup. L'épaisseur maximale (478 m) a été observée à Mount Bridgman. De là, la succession s'amincit tant vers le sud-ouest que vers le nord-est, ce qui reflète la fermeture du sous-bassin.

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INTRODUCTION

Artinskian (Lower Permian) carbonate strata are remarkably well exposed along the eastern margin of Sverdrup Basin, a major depocentre of the Canadian Arctic Archipelago. These rocks form an unconformity-bounded, 5 to 10 Ma duration, transgressive-regressive sequence (Beauchamp et al., 1989b), the transgressive and regressive portions of which are represented by two distinct stratigraphic units (Beauchamp and Henderson, work in progress). These units comprise a lower, deepening-upward, greenish weathering, mixed clastic/carbonate succession, overlain by an upper, shallowing-upward, yellowish weathering carbonate succession with minor sandstone. Inconsistently assigned to different formations, these two Artinskian units will soon be given formal formation names (Beauchamp and Henderson, work in progress). A whole series of higher-order cycles of unknown origins are superimposed upon the broad transgressive-regressive package, the sediments of which were deposited contemporaneously with a major climatic shift in the Canadian Arctic.

Artinskian strata were the focus of a detailed investigation in west-central Ellesmere Island (Fig. 1) during the summer of 1990. This research project was carried out to define the environments of deposition, understand the higher-order intraformational cyclicity, and assess the magnitude of the climatic shift in Artinskian time. In addition, this study was undertaken to help delineate the geometry of a northeasterly trending subbasin of Sverdrup Basin, the Fosheim-Hamilton subbasin, which extends from the environs of Bay Fiord in the southwest to the mouth of Tanquary Fiord in the northeast.

Seven stratigraphic sections from a variety of depositional settings within the Fosheim-Hamilton subbasin were measured. More than 550 lithological samples were collected for petrographic analysis, and 70 paleontological samples were gathered for conodont biostratigraphic analysis. Field descriptions and some broad generalizations regarding the architecture of the Fosheim-Hamilton subbasin are provided in this report. This study, still in a preliminary stage, constitutes Scott's M.Sc. research project at the University of Calgary, and will result in the publication of more definitive articles over the next two years.

PREVIOUS WORK

Per Shei, a geologist attached to the "Second Norwegian Expedition in the Fram, 1898-1902", was the first to examine Artinskian strata in the Canadian Arctic. He gathered an important fossil collection from Great Bear Cape on Bjorne Peninsula, Ellesmere Island, which was later examined by several paleontologists. Although the Permian affinity of the sampled fauna was recognized, no agreement on a more precise age could be achieved [A Wolfcampian Series age was proposed by Tschernyshew and Stepanow (1916), and agreed upon by Troelson (1950); an Early Permian age was suggested by Dunbar (1955) and Gobbet (1963); and an early Late Permian age was recommended by Minato (1960)]. Field parties of the Geological Survey of Canada later studied and sampled strata from Raanes and Bjorne peninsulas on Ellesmere Island (Thorsteinsson and Tozer, 1957; Harker and Thorsteinsson, 1960; McLaren, 1963; Thorsteinsson, 1963; and Tozer, 1963). Following these investigations, Thorsteinsson (1974) published a stratigraphic framework for the upper Paleozoic of Sverdrup Basin in which Artinskian strata were assimilated into various Upper Carboniferous to Upper Permian formations.

A few years later, Nassichuk (1975) and Nassichuk and Wilde (1977) assigned these strata to an "unnamed" formation of late Early Permian age, based on their distinct lithological and faunal content. Subsequent field investigations by B. Beauchamp and C.M. Henderson from 1983 to 1989 confirmed that this "unnamed" formation represents a separate stratigraphic entity of early to late Artinskian age, with the uppermost part possibly being equivalent to the earliest Kungurian (Henderson, 1988; Beauchamp et al., 1989a). Moreover, these authors recognized that the "unnamed" formation can be broken into two very distinct and mappable units: a lower transgressive unit and an upper regressive unit. These two units (referred to as the "green" and "yellow" members of the "unnamed" A formation by Beauchamp et al. (1989a,b), and as the "lower green" and "upper yellow" units in the present article) will be given formational names (Beauchamp and Henderson, work in progress). A third formation, the basinal equivalent of the two new formations referred to as "unnamed" B by Beauchamp et al. (1989a,b), will also be named.

REGIONAL SETTING

Artinskian sediments were deposited in the northeasterly-trending Sverdrup Basin, a rift that originated in the Early Carboniferous (Thorsteinsson, 1974; Trettin, 1989; Davies and Nassichuk, in press). A thick upper Paleozoic to Tertiary succession of sedimentary rocks with minor volcanics was deposited within the basin. The succession was folded and faulted in the Tertiary, during the Eurekan Orogeny, and mountain ranges were formed in the eastern Arctic where the Artinskian rocks are exposed. Seven long-term (5-50 Ma) transgressive-regressive sequences characterize the upper Paleozoic succession of the basin. These sequences are bounded at the basin margin by significant unconformities and basinward by their equivalent conformities (Beauchamp et al., 1989b). Each transgressive-regressive package is characterized by marginal clastics that pass basinward into platform carbonates and/or clastics, which in turn grade into basinal shales, evaporites, or cherts. Facies assemblages in each sequence vary as a result of low- to high-order sea-level fluctuations, ongoing tectonic activities, and climatic fluctuations (Beauchamp et al., 1989b).

One of the seven long-term transgressive-regressive sequences was deposited during the early Artinskian/earliest Kungurian time interval (5 to 10 Ma duration). The transgressive phase of the sequence is represented by generally recessive, greenish weathering, deepening-upward shales and mixed siliciclastic-carbonate facies ("lower green unit"); the regressive phase of the sequence comprises cliff forming, yellowish weathering, shallowing-upward carbonates and minor sandstones ("upper yellow unit"). The "lower green unit" is underlain by either sandstones of the Canyon Fiord Formation or carbonates of the Belcher Channel, Tanquary,

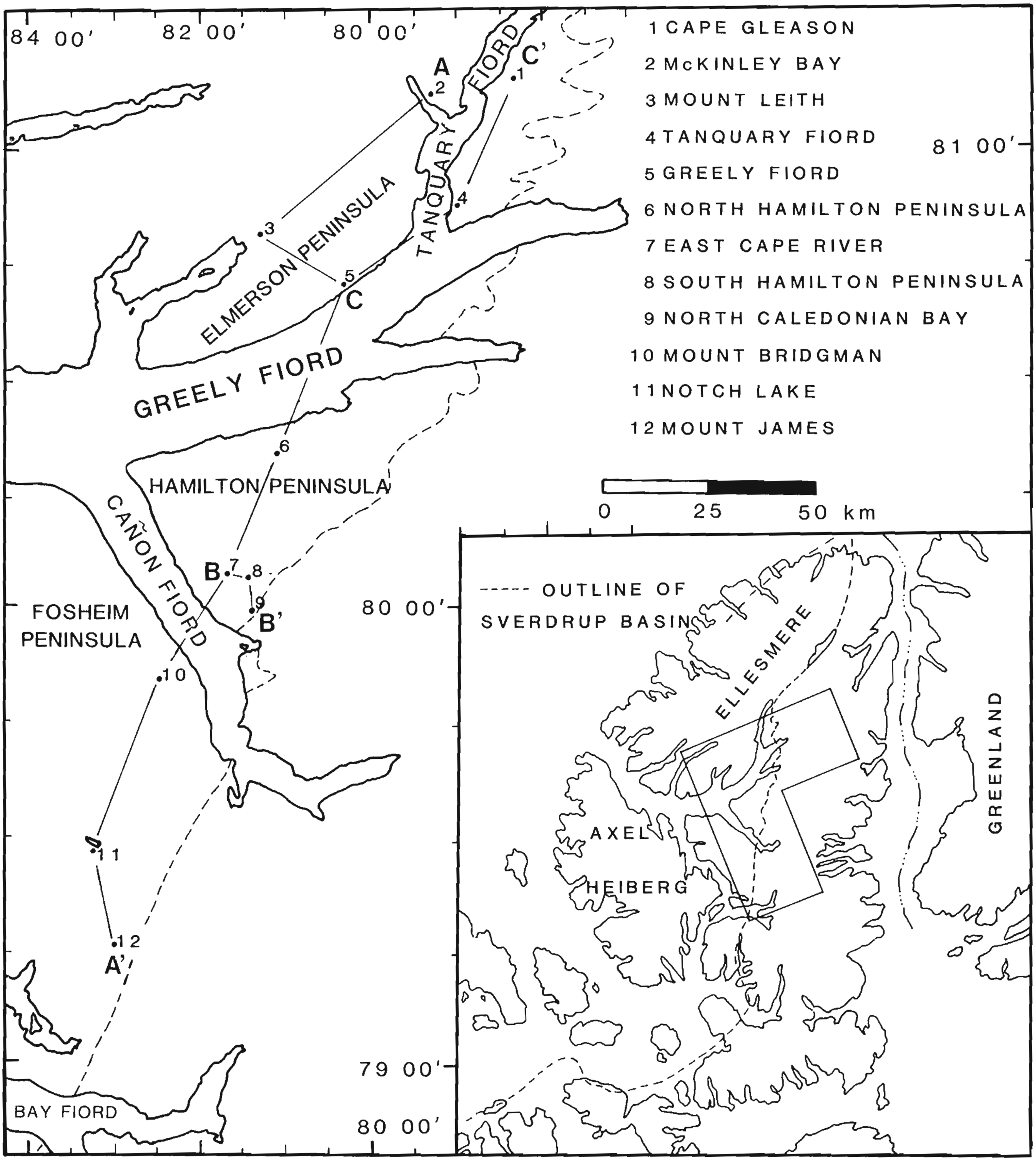

Figure 1. Map, showing location of measured sections and cross-sections AA', BB' and CC' on west-central Ellesmere Island.

and Nansen formations. The "upper yellow unit" is overlain either by marginal clastics of the Sabine Bay Formation, marine clastics of the Assistance or Trold Fiord formations, or slope to basinal shales and chert of the van Hauen Formation.

Artinskian strata are well exposed on Fosheim and Hamilton peninsulas, where both the "lower green unit" and the "upper yellow unit" have been mapped as one or other of the upper Tanquary, Belcher Channel, and Canyon Fiord formations. The same units have been assigned, north of Greely Fiord and east of Tanquary Fiord, to the upper parts of the Tanquary, Belcher Channel, and Nansen formations. Two sections were measured on Fosheim Peninsula (Mount James and Mount Bridgman), three sections on Hamilton Peninsula (East Cape River, South Hamilton Peninsula, and North Hamilton Peninsula), one section west of the mouth of Tanquary Fiord (Greely Fiord), and one section east of the mouth of Tanquary Fiord (Tanquary Fiord) (see Figure 1). These sections are briefly described below and illustrated in three cross-sections (Figs. 2-4). Five additional sections were added to the cross-sections. These sections ("Notch Lake": 79°29'N, 83°18'W; North Caledonian Bay: 80°01'N, 81°31'W; Mount Leith: 80°01'N, 81°45'W; McKinley Bay: 81°09'N, 81°45'W; and Cape Gleason: 81°11'N, 77°57'W) were not described thoroughly and are added only to improve the assessment of the thickness variations.

SUMMARY OF FIELD OBSERVATIONS

Mount James (Section 12)

The Mount James section (Fig. 5) is located within the Sawtooth Range at latitude 79°18'N and longitude 83°05'W. The total thickness of the Artinskian sequence is 129.5 m — 79 m of the "lower green unit" and 50.5 m of the "upper yellow unit". The basal contact of the "green unit" with

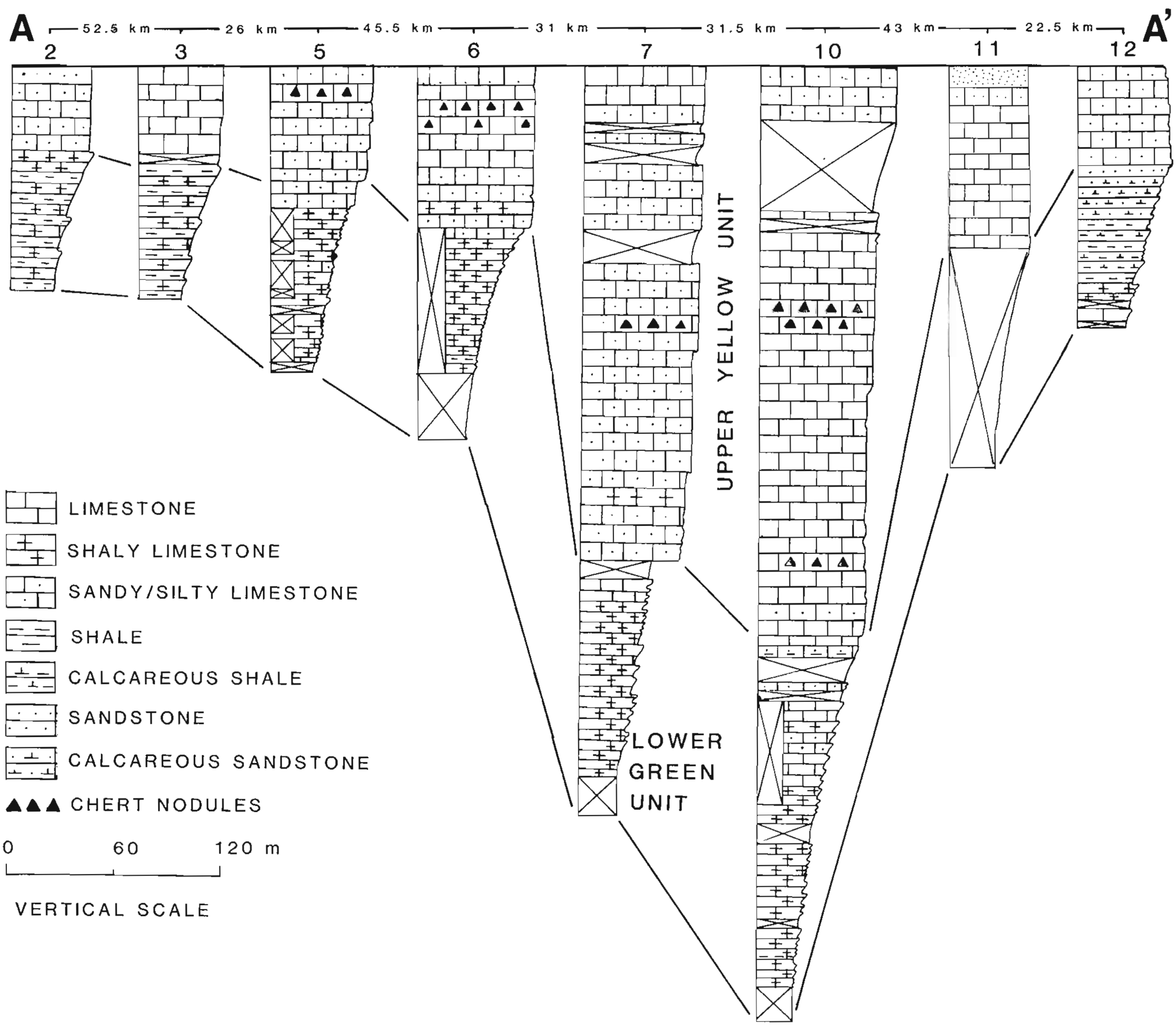

Figure 2. Cross-section AA', showing thinning of Artinskian strata in southwesterly and northeasterly directions.

the Canyon Fiord Formation is covered, as is the contact of the "yellow unit" with the overlying Trold Fiord Formation. The "green unit" at Mount James is divisible into seven packages on the basis of recessive-resistant outcrop patterns. These packages may represent fourth-order (1-2 Ma duration; Vail et al., 1977) relative sea-level fluctuations. Each package is characterized by a higher order (100 000 to 500 000 years) of cyclicity with 3 to 6 cycles within each package.

The lower part of the "green unit" is dominated by laminated, wavy bedded shales which become sandy upward. Toward the top of the "green unit", thin, shaly limestone (wackestone) beds alternate with sandstone beds. Megafossils are scarce in the "green unit", but gastropods are present as well as a few brachiopods, crinoids, and bryozoans. The fossil content is difficult to discern at the outcrop and will be better defined once slabbed and thin-sectioned samples have been examined.

The "upper yellow unit" is exposed in continuous cliffs. The contact between the lower and the upper units is gradual, as alternating limestones and sandstones pass into silty limestones. The limestone becomes coarser grained and more fossiliferous upward; crinoids and brachiopods are the dominant megafossils, with less common bryozoans. Beds of colonial rugose corals in growth position occur in the uppermost 25 m of the section.

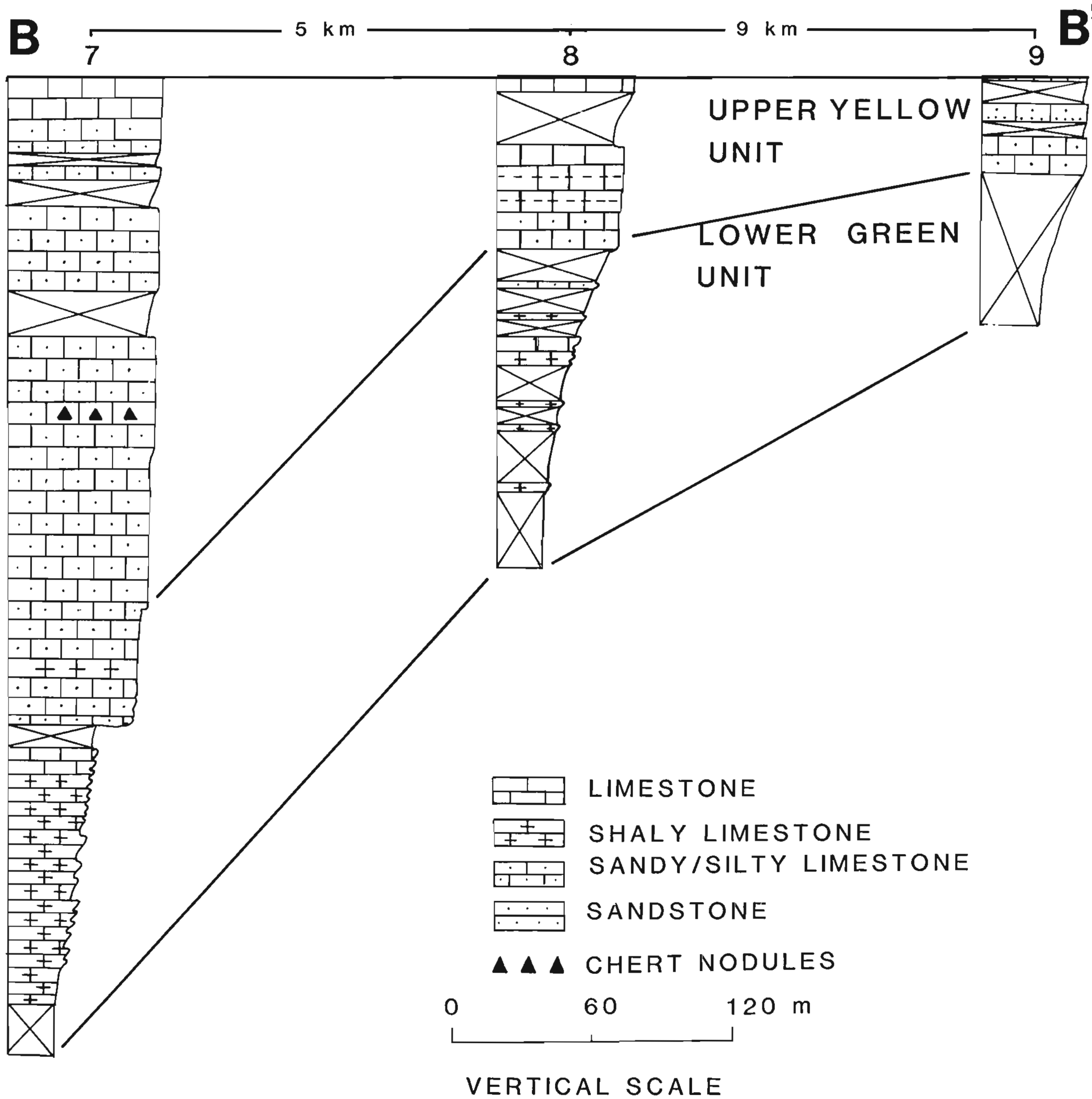

Figure 3. Cross-section BB', showing thickening of Artinskian strata in a northwesterly direction.

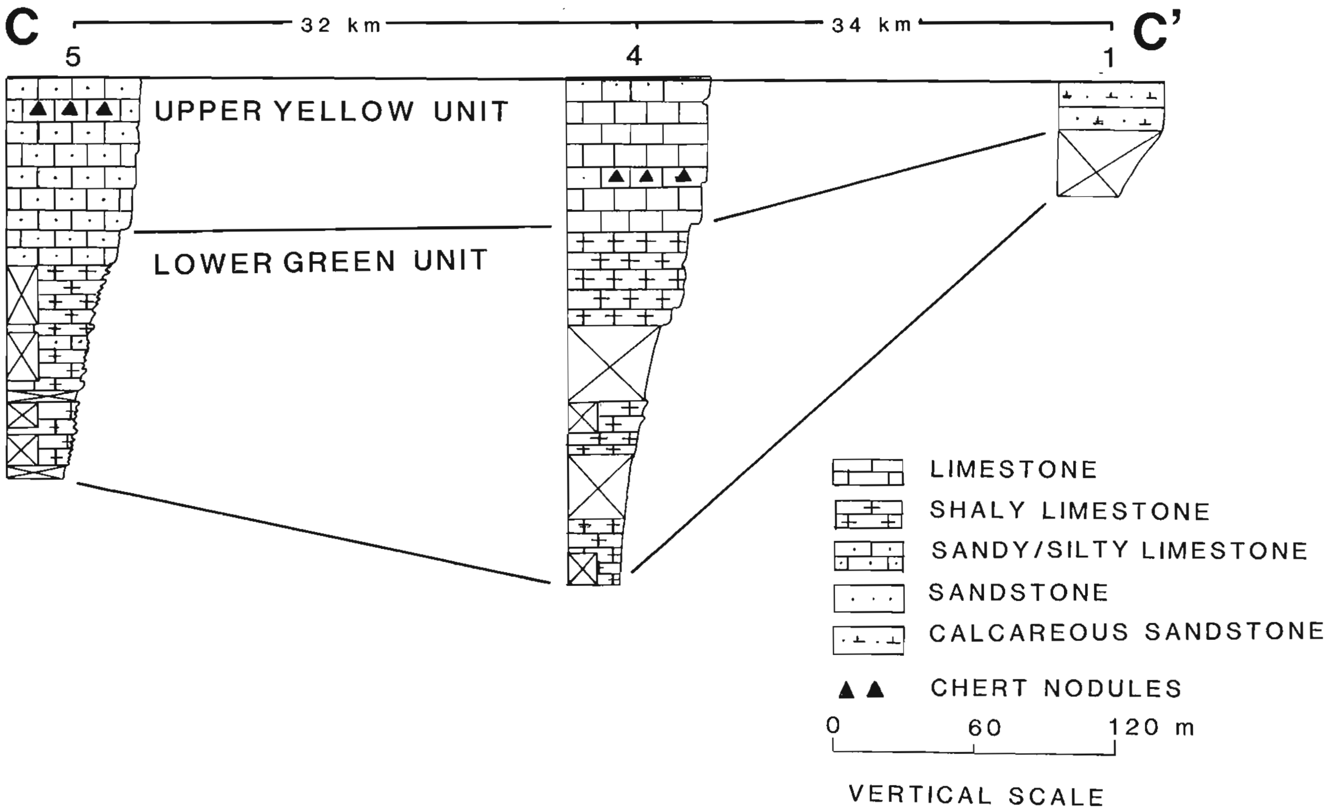

Figure 4. Cross-section CC', showing thickness variations of Artinskian strata across Tanquary Fiord.

Figure 5. Mount James section. (Section 12 in Figure 1.) Note contacts between Canyon Fiord Formation (C), "lower green unit" (lg) and "upper yellow unit" (uy).

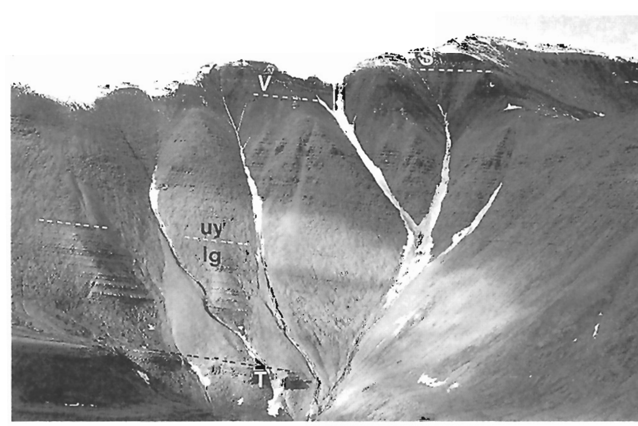

Figure 6. Mount Bridgman section. (Section 10 in Figure 1.) Note contacts between Tanquary Formation (T), "lower green unit" (lg), "upper yellow unit" (uy), van Hauen Formation (V), and Sabine Bay Formation (S).

Mount Bridgman (Section 10)

The Mount Bridgman section (Fig. 6), at latitude 79°51'N and longitude 82°38'W, represents the thickest accumulation of Artinskian sediments in the study area with 195 m of the "lower green unit" and 283 m of the "upper yellow unit" (total thickness of 478 m). The "lower green unit" overlies limestones of the Tanquary Formation, but the contact between the two formations is covered.

The "lower green unit" is easily separated into at least 7 packages with 4 to 9 higher-order cycles within each package. The lithology is dominated by greenish grey, shaly limestones interbedded with shales, particularly toward the base of the section. The limestone beds become thicker and less shaly toward the top of the "green unit". The rocks are very fossiliferous with abundant fenestrate and ramose bryozoans, crinoids, and brachiopods. Rugose horn corals are scattered throughout the unit, and large specimens (average 5 cm diameter) are found concentrated within a single 30 cm thick bed in the first package of the "green unit". The contact between the "lower green unit" and the ''upper yellow unit'' is marked by a gradual change into less shaly limestones that lack common shale interbeds.

The rocks of the ''upper yellow unit'' are exposed in nearly continuous outcrops and cliff faces. The limestones are predominantly packstones with a buff-yellow to orange weathering colour that changes to a reddish-rust colour higher in the section. The rocks are fossiliferous, containing abundant bryozoans, crinoids and brachiopods, and occasional horn corals, sponges, and fusulinaceans. The top of the section consists of a white sandstone, characterized by parallel laminae passing into large-scale trough crossbedded units, capped by a crinoidal grainstone. The division of the ''upper yellow unit'' into depositional cycles was not obvious from field observations; this is often the case elsewhere in Sverdrup Basin (Beauchamp and Henderson, work in progress). Dark siliceous shales and chert of the van Hauen Formation rest on top of the ''yellow unit''. These dark lithologies in turn pass upward into clean sandstones of the Sabine Bay Formation.

Figure 7. East Cape River section. (Section 7 in Figure 1.) Note erosional contact between ''upper yellow unit'' (uy) and tongue of van Hauen Formation (V), and conformable contact between van Hauen mudrocks and Sabine Bay sandstones (S).

East Cape River (Section 7)

The section at East Cape River (lat. 80°06'N, long. 81°48'W), is 372 m thick, and both the ''lower green'' and ''upper yellow'' units are well exposed. The contact of the ''lower green unit'' with Tanquary limestones is covered — as in all other sections measured within the study area.

The 200 m thick ''lower green unit'' comprises highly bioturbated, shaly to silty limestone. Vertical and horizontal burrows are abundant and give the greenish grey limestone a mottled appearance. The unit is divisible into 5 packages with 3 to 9 higher-order cycles within each package. Megafossils are present throughout the section, but are more abundant toward the top of the unit. Bryozoans are the most frequently observed fossils, and crinoids also are common. Brachiopods and small (< 2.5 cm diameter) rugose horn corals are scattered within the unit, and fusulinaceans were observed within one bed.

The contact between the ''lower green unit'' and the ''upper yellow unit'' is marked by a lithological change from shaly limestone to silty-sandy(?) limestone lacking shale. The ''upper yellow unit'' is 172 m thick and consists of bioturbated sandy limestone that contains four shallowing-upward cycles, each capped by a bioclastic or intensely bioturbated zone. In addition, parallel laminated beds, averaging 8 cm in thickness, are occasionally interspersed within some of the bioturbated intervals. A sill, approximately 60 m thick, cuts the upper portion of the unit. The sequence is capped by a crystalline, unfossiliferous limestone and a nonmarine sandstone that marks the contact with the overlying Sabine Bay Formation (Fig. 7). A 5.5 m thick sequence of dark, siliceous shales and limestones occurs at the base of the Sabine Formation. It represents a tongue of the deep water van Hauen Formation, indicating a major drowning of the carbonate platform after deposition of the ''upper yellow unit''.

South Hamilton Peninsula (Section 8)

The South Hamilton Section is located five kilometres east-southeast of the East Cape River Section at latitude 80°05'N and longitude 81°36'W. The total thickness of Artinskian strata at this locality is 186.5 m (125.5 m in the ''lower green unit'' and 61 m in the ''upper yellow unit''). The contact between the ''lower green unit'' and the underlying Belcher Channel Formation is covered.

The ''lower green unit'' is poorly exposed with sporadic outcrops on a recessive, talus covered slope. The outcrop pattern makes accurate division of the formation into depositional cycles difficult. The unit is predominantly a shaly to silty limestone which becomes fossiliferous and bioturbated toward the top. Bryozoans, crinoids and brachiopods are the most common megafossils, with some small (< 2.5 cm diameter) rugose horn corals. The shaly limestones at the base of the unit are interbedded with very thin shale beds. The contact between the ''lower green unit'' and the ''upper yellow unit'' is a gradual transition into resistant, less shaly, more fossiliferous limestone.

In contrast to the ''lower green unit'', the ''upper yellow unit'' is well exposed. The basal part of the section consists of 0.5 to 1 m thick beds of wackestone to packstone, which alternate with very thin shale beds. Abundant colonial corals, not in growth position, are present in the shaly interbeds. The wackestone and packstone grade upward into a silty limestone with abundant megafossils, including bryozoans, brachiopods and crinoids, and some colonial corals, horn corals and fusulinaceans. Shale interbeds occur in the top part of the unit. The contact between the ''upper yellow unit'' and the overlying Sabine Bay Formation was not observed.

North Hamilton Peninsula (Section 6)

The North Hamilton Peninsula section at latitude 80°21'N and longitude 81°16'W is 185.5 m thick and is divisible into a 117 m thick ''lower green unit'' and a 68.5 m thick ''upper yellow unit''. The contact between the ''lower green unit'' and the Tanquary Formation is covered, as is the contact of the ''upper yellow unit'' with sandstones of the Sabine Bay Formation.

The ''lower green unit'' is sporadically exposed on a recessive slope. Outcrops average 0.5 to 1 m in thickness in the lower part of the section, and increase to a maximum thickness of 3 m at the top. The lithology is primarily shaly wackestone, which becomes silty wackestone to packstone toward the top of the unit. The section is heavily bioturbated with numerous horizontal and vertical burrows visible throughout the section. Thin shale beds are interbedded with the limestone in the basal portion of the unit. The contact between the "lower green unit" and the "upper yellow unit" is transitional.

The "upper yellow unit" forms a resistant, nearly continuous outcrop of silty limestone (packstone). The lower half of the unit is heavily bioturbated (primarily with horizontal burrows) giving the unit a mottled appearance. Chert nodules are present in the upper portion of the unit, and the top of the section is capped by a crystalline, unfossiliferous(?) limestone (caliche?). Bryozoans, brachiopods, and crinoids are the primary faunal constituents throughout the section, with some horn corals and sponges. Fusulinaceans are present only within the upper unit.

Greely Fiord (Section 5)

The Greely Fiord section (Fig. 8) is located on the north shore of Greely Fiord at latitude 80°44'N and longitude 80°15'W. A fault through the section allows access to a nearly continuous cliff exposure of the "upper yellow unit". The "lower green unit" is only partly exposed, and the contact with the underlying Belcher Channel Formation is covered.

The "lower green unit" is easily divisible into five major packages with 3 to 9 cycles evident within each package. The outcrops are 0.3 to 1 m thick at the base of the packages, increasing to up to 3 m at the top. The rocks are primarily shaly limestones with some silty limestone beds. Bryozoans, brachiopods, and crinoids are the predominant megafossils, with fusulinaceans, and rugose horn corals present in lesser amounts. Bioturbation is evident in the lower part of the section, and becomes pervasive in the upper part of the "lower green unit". The contact between the lower and upper units is sharp and is marked by a change from alternating silty limestone and argillaceous beds to continuous silty limestone.

The "upper yellow unit" is a silty packstone/grainstone that is heavily bioturbated at the base, but bioturbation decreases in the upper part. Chert nodules are present toward the top of the unit, which is capped by silty, laminated limestone. The dominant megafossils are bryozoans, brachiopods, and crinoids. The contact of the "upper yellow unit" with the Sabine Bay Formation is covered.

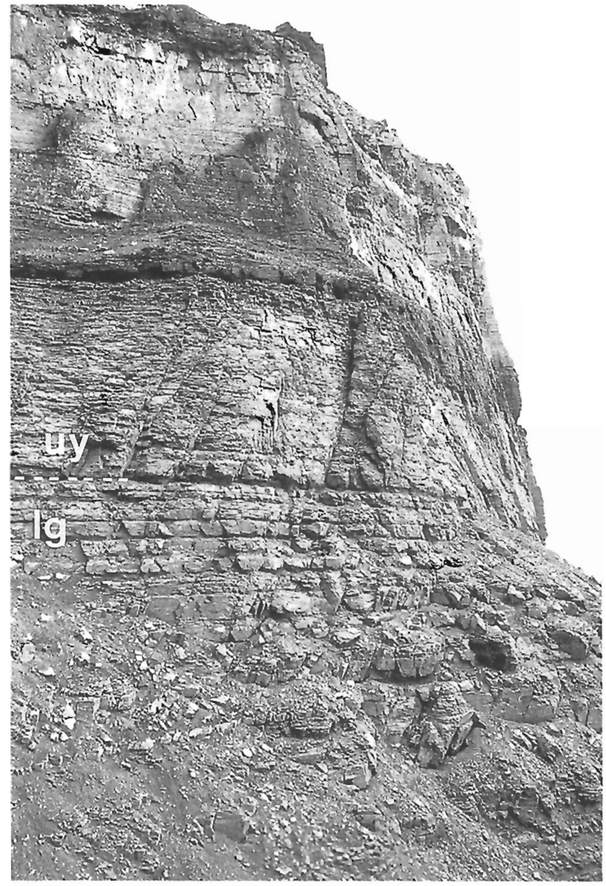

Figure 8. Greely Fiord section. (Section 5 in Figure 1.) Note sharp contact between "lower green unit" (lg) and "upper yellow unit" (uy).

Tanquary Fiord (Section 4)

The Tanquary Fiord section (Fig. 9) is located at latitude 80°55'N and longitude 78°49'W, on the east side of Tanquary Fiord, approximately 7 km northeast of the tip of the peninsula between Greely and Tanquary fiords. The section is 191.5 m thick (136.5 m in the "lower green unit", and 55 m in the "upper yellow unit"). The "lower green unit" overlies the Tanquary Formation at this locality.

The "lower green unit" was measured from discontinuous outcrops on a recessive slope. The formation is characterized by greenish grey, bioturbated, shaly wackestone to packstone interbedded with calcareous shale. The shale interbeds are most common toward the base of the section, although they are occasionally present in the upper parts of the unit also. Bryozoans, brachiopods, and crinoids are the predominant megafossils present with less common rugose horn corals. The contact between the lower and upper units is marked by an upward lithological change to limestone (predominantly packstone) which lacks the shale content of the underlying rocks.

The basal portion of the "upper yellow unit" is fossiliferous; brachiopods, crinoids and bryozoans are the most common fossils with minor rugose horn corals, sponges, and fusulinaceans. The unit is divisible into three packages on the basis of outcrop pattern, weathering colour, and lithology. The upper unit section forms a resistant cliff, but each package is defined by a definite break in the resistance profile. The first package is a yellow weathering succession that is fossiliferous and highly bioturbated at the base. Layers of abundant shell fragments are common. The package is capped by a 5 cm thick layer of fusulinaceans. Strata of this package are red weathering and show evidence of heavy bioturbation; however, megafossils are less abundant than in the first unit. Chert nodules are present toward the base, where sponges were also observed. Fossils are commonly silicified near the

Figure 9. Tanquary Fiord section. (Section 4 in Figure 1.) Note onlap of Artinskian strata (arrows) on uppermost Tanquary bed (T); and the contact between "lower green unit" (lg) and "upper yellow unit" (uy).

base. Overlying the second package is a yellow weathering limestone devoid of megafossils other than occasional brachiopod shell fragments. A glauconitic sand is present at the top of the unit; it may belong to the Assistance Formation or the Trold Fiord Formation. Otherwise, the contact between the upper unit and the overlying Triassic Bjorne Formation is covered.

DISCUSSION

Thickness variations and facies trends

Important thickness variations are depicted in the cross-sections in Figures 2 to 4, and are illustrated in an isopachous map for the study area (Fig. 10). These variations indicate that Artinskian sediments were deposited in an important subbasin at the margin of Sverdrup Basin. This subbasin, referred to as the Fosheim-Hamilton subbasin (see also Beauchamp et al., 1991; Morin et al., 1991; and Thériault and Beauchamp, 1991), had its Artinskian depocentre located in the environs of Mount Bridgman. From there, the Artinskian sequence thins in a southwesterly direction toward Bay Fiord, in a northeasterly direction toward the head of Tanquary Fiord, and in a north or northwesterly direction toward the head of Hare Fiord. The actual and extrapolated contours in Figure 10 delineate the Artinskian Fosheim-Hamilton subbasin. The contours reflect the effect of Tertiary thrusting superimposed on the primary depositional patterns. This is most evident in the B-B' cross-section which shows a dramatic thickening of the Artinskian succession over a distance of 14 km. This cross-section represents a proximal to distal line across the depositional axis of the Fosheim-Hamilton subbasin. The cross-section is short as the result of Tertiary telescoping along two major thrust faults.

The northeasterly and southwesterly thinning, as depicted in the A-A' cross-section (Fig. 2), appears to reflect syndepositional patterns. Indeed, the succession thins in a southwesterly direction toward a known paleogeographic element, referred to as the Bay Fiord high by Beauchamp (1987). This element separates two areas of reverse depositional polarities: Raanes Peninsula to the south, where all

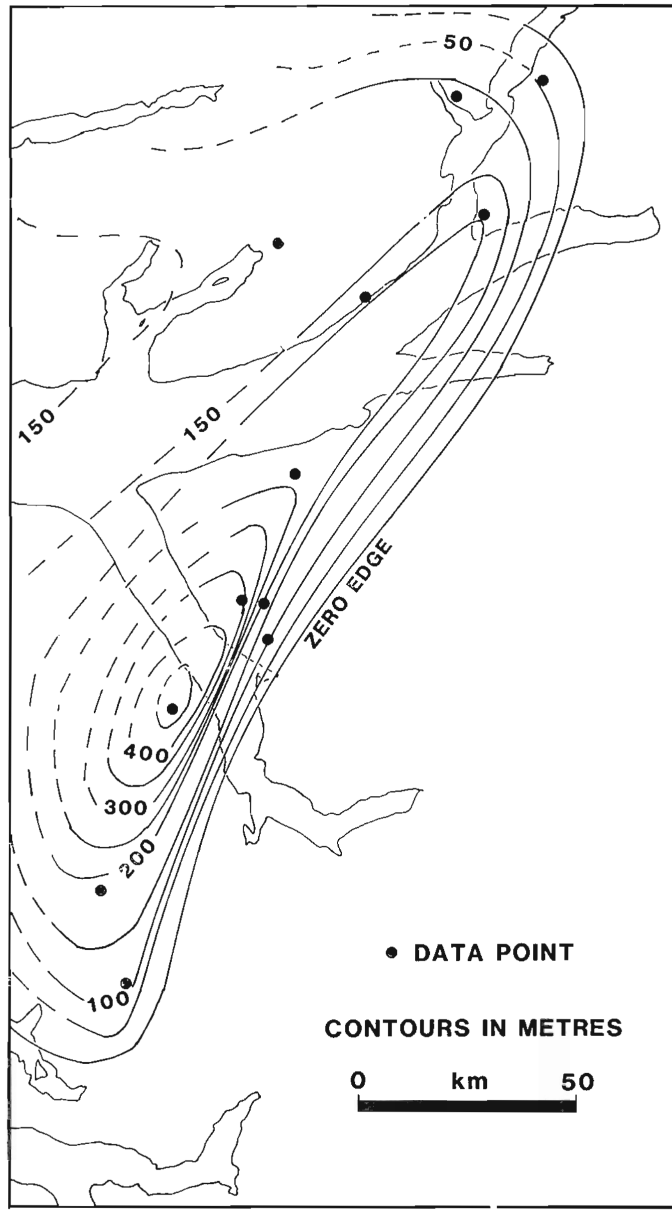

Figure 10. Isopach map of Artinskian strata on west-central Ellesmere Island. Actual (solid) and extrapolated (dashed) contours at 50 m intervals.

upper Paleozoic formations thicken, and their contents indicate increasing water depth in a southwesterly direction; and Fosheim Peninsula to the north, where the opposite stratigraphic and sedimentological trends can be observed. The origin of the "Bay Fiord High" is at present unknown, but it seems to have acted as an important source of clastic sediments on Raanes Peninsula, and possibly on Fosheim Peninsula — as indicated by the presence of thick crossbedded sandstones in the upper part of the "Notch Lake" section (Fig. 11).

The northeasterly thinning of the Artinskian succession may be attributed to the presence of the Tanquary High to the north. The Tanquary High is an east-west oriented structural element of Sverdrup Basin, which influenced surrounding sedimentation from the Late Carboniferous to the Jurassic (Nassichuk and Christie, 1969). According to Maurel (1989), the high was formed during the Carboniferous rifting of Sverdrup Basin, and it must have been active, and episodically emergent, throughout the Carboniferous and Permian. For example, the high must have played a major role in restricting circulation in the Fosheim-Hamilton subbasin in Asselian and Sakmarian time, which led to the accumulation of up to 400 m of evaporite strata in the Antoinette and Mount Bayley formations. The high must also have been emergent during the Artinskian, as suggested by the presence of nearshore and perhaps fluvial sandstones in the upper part of the "upper yellow unit" at Section 1. In addition, it is likely that syn-Sabine Bay movements affected the high, leading to the translocation of the zero edge in a westerly direction, toward the head of Hare Fiord. This is indicated by the presence of cobble-sized limestone clasts of Early Permian age in conglomerate-filled channels of the Sabine Bay Formation at Section 1.

The origin of the Fosheim-Hamilton subbasin remains a mystery, although its origin has been the subject of much speculation (see Morin et al. 1991, and Thériault and Beauchamp, 1991). Morin et al. (1991) postulate that, in addition to the Tanquary High, another positive linear element of some sort (the "Elmerson High") must have been present in the environs of Elmerson Peninsula in Asselian-Sakmarian time, in order to restrict circulation between the main Sverdrup Basin and the Fosheim-Hamilton subbasin. The so-called "Elmerson High" does not appear to have been a very active element in Artinskian time, as indicated by the open marine biota in sediments to the south. Accordingly, Artinskian sediments were deposited during a long-term episode of tectonic quiescence, apparently characterized by passive subsidence, whereas Sakmarian and older sediments appears to have been deposited during an active phase of rifting. In Artinskian time, the Elmerson Peninsula area north of Greely Fiord appears to have been a very shallow and stable area bordering the Fosheim-Hamilton subbasin.

The isopach map in Figure 10 is based on a number of data points, but also on great deal of speculation. Indeed, no one can tell at this stage whether the contours form tight loops around the Mount Bridgman area (as shown in Figure 10) or whether they carry without interuption to the north, where great thicknesses of Artinskian strata are known to occur (Blue Mountains, Van Hauen Pass). As the key to the solution lies in the subsurface north and northwest of Greely Fiord, only speculations can be advanced. The tight loop interpretation shown in Figure 10 was favoured for two main reasons. Firstly, based on the great thickness of Lower Permian strata on both sides of the presumed "Elmerson High", and based on the magnitude (up to halite facies) and the duration (up to 5 Ma) of the restriction involved, one can postulate that the "Elmerson High" must have been a very important and relatively stable structure separating two different subsidence provinces of Sverdrup Basin. It likely remained that way throughout the entire Permian, even though a better connection with the main Sverdrup Basin was clearly established in post-Sakmarian time. Secondly, the tight loop interpretation is favoured because Upper Permian deposits (Trold Fiord Formation) show a thinning in a northwesterly direction, from a depocentre in the environs of "Notch Lake" to the subsurface Fosheim N-21 well (van Hauen Formation) on northern Fosheim Peninsula. The Upper Permian succession then thickens again toward the Blue Mountain/Hare Fiord area (van Hauen and Degerbols formations).

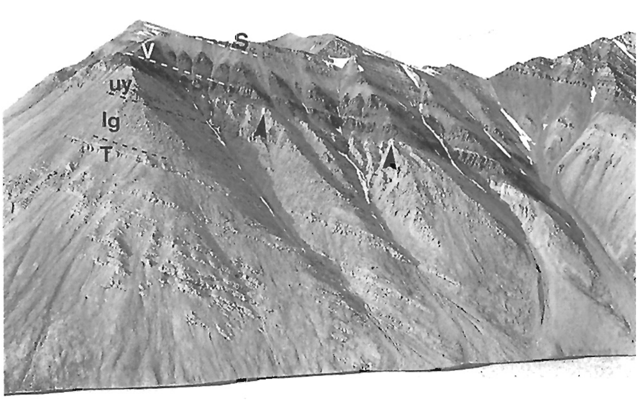

Figure 11. Notch Lake section. (Section 11 in Figure 1.) Note contacts between Tanquary Formation (T), "lower green unit" (lg), "upper yellow unit" (uy), van Hauen Formation (V), and Sabine Bay Formation (S); and the thick sandstone unit in the uppermost "upper yellow unit" (arrows).

Correlations and cyclicity

Correlations derived from field measurements are speculative, owing to the large distances between the sections, and the absence of good regional marker beds, other than the contact between the "lower green unit" and the "upper yellow unit". Petrographic analysis, in addition to biostratigraphic information derived from conodont and fusulinacean samples, will help with future correlations. The datum chosen for the cross-sections in Figures 2 to 4 is the top of the "upper yellow member". This surface most likely represents an erosional unconformity throughout the study area. Indeed, an erosional contact was observed between the "upper yellow unit" and the Sabine Bay Formation at East Cape River (Fig. 7), which, with the exception of the Mount Bridgman section, is the most distal section. It is almost certain that this unconformity is of regional extent, as suggested by the occurrence of an equivalent unconformity elsewhere in Sverdrup Basin (Beauchamp and Henderson, work in progress). In the other sections, limestones or sandstones of the uppermost "upper yellow unit" are overlain by red weathering, poorly consolidated shales (with caliche nodules) of the Sabine Bay Formation. These sediments, interpreted as floodplain deposits, quite likely rest unconformably on limestones of the "upper yellow unit".

Most of the measured sections display five major third- or fourth-order depositional cycles (Vail et al., 1977) within the "lower green unit". The Mount Bridgman and Mount James sections contain seven cycles. The Mount James section is characterized by nearshore (tidal flat?) sediments, and does not correlate well with the other sections. In all sections, depositional cycles are more readily apparent in the "lower green unit" than in the "upper yellow unit" owing to the recessive-resistant, cyclic, outcrop profile of the former. Although repeated shallowing-upward depositional packages were observed within the "upper yellow unit" in many of the sections, correlation of these cycles between sections was not established.

Fauna and climate

The fauna in both the "lower green" and "upper yellow" units is dominated by bryozoans, crinoids, and brachiopods. Less common rugose horn corals were found within both units, whereas colonial corals were only found in the upper unit in the Mount James and South Hamilton sections. Sponges are also occasionally present in the lower unit, but are abundant within the upper part of the upper unit. Fusulinaceans occur within both units, but are far more abundant in the "upper yellow unit".

The dominance of bryozoans, echinoderms, and brachiopods in Artinskian carbonate rocks indicates that environmental conditions were much cooler during that time interval than ever before in Sverdrup Basin (Beauchamp et al, 1989a). Indeed, Sakmarian and older fossiliferous deposits contain a wide variety of organisms, which include bryozoans, echinoderms and brachiopods, in addition to several families of foraminifers, phylloid, dasycladacean and red algae, and other exotic reef-builders such as *Palaeoaplysina* and *Tubiphytes*. These organisms, in addition to the presence of red beds, thick caliche profiles and evaporites, indicate that the pre-Artinskian was a period characterized by a rather dry and relatively warm tropical type climate. In contrast, the faunal impoverishment in Artinskian deposits suggests that the climate shifted to cooler, and perhaps more humid, temperate-like conditions, in a relatively short time interval.

CONCLUSIONS

Artinskian rocks on west-central Ellesmere Island comprise two distinct units: a "lower green unit" of shale and mixed clastics and carbonate rocks deposited as a series of multi-order cycles; and an "upper yellow unit" of fossiliferous carbonate rocks and subordinate sandstones deposited as poorly defined cycles. These rocks were deposited in a relatively cool, temperate climate in a northeasterly trending subbasin (Fosheim-Hamilton subbasin) at the margin of the main Sverdrup Basin. The current subbasin geometry reflects: 1) the occurrence of a depocentre in the environs of Mount Bridgman; 2) the presence of two major paleogeographic elements, "Bay Fiord High" to the southwest and Tanquary High to the northeast; 3) the possible occurrence of a linear high ("Elmerson High") to the northwest; and 4) an important, Tertiary, thrust-related telescoping of strata on Fosheim and southern Hamilton peninsulas.

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Preliminary interpretation of the depositional history and tectonic significance of the Carboniferous to Permian Canyon Fiord Formation, west-central Ellesmere Island, Arctic Archipelago

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Abstract

The depositional and tectonic settings of the Canyon Fiord Formation in central Ellesmere Island are interpreted. Three distinct facies assemblages have been recognized: 1) a conglomerate assemblage, deposited in eastward-flowing, proximal braided streams during an extensional faulting event; 2) a sandstone assemblage, deposited in a westerly direction in the braided stream and floodplain environment, during or following the extensional faulting event; and 3) a sandstone/limestone assemblage, deposited as high-order ($< 2 \times 10^5$ yrs) sea-level cycles in the paralic environment. The sandstone/limestone assemblage is further related to a second-order ($\sim 3 \times 10^7$ yrs) marine transgression. Patterns of sediment dispersal strongly suggest that the succession was deposited within an asymmetric half-graben — a predominant structure in several modern and ancient rift basins.

Résumé

Les environnements sédimentaires et le contexte tectonique de la Formation de Canyon Fiord sont interprétés pour la région du centre de l'île d'Ellesmere. Trois assemblages de faciès sédimentaires ont été identifiés: 1) un faciès à conglomérats, formé dans un milieu de rivières anastomosées proximales s'écoulant vers l'est, lors d'un épisode d'extension le long de failles normales;2) un faciès à grès, formé durant ou après l'épisode de formation de failles, dans des milieux de rivières anastomosées et de plaines d'inondation s'écoulant vers l'ouest; et 3) un faciès à grès et calcaires, formé dans le milieu paralique sous forme de cycles marins d'ordre élevé ($< 2 \times 10^5$ années). Le faciès à grès et calcaires est de plus associé à une transgression marine de deuxième ordre ($\sim 3 \times 10^7$ années). Les réseaux de dispersion sédimentaire semblent indiquer que la séquence de Canyon Fiord s'est accumulée à l'intérieur d'un demi-graben, reconnu comme la structure prédominante de plusieurs rifts modernes et anciens.

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INTRODUCTION

The Carboniferous Canyon Fiord Formation is a clastic dominated succession that outcrops as a narrow band along the margin of Sverdrup Basin, in the Canadian Arctic. Exposures are known from northern Melville Island (Tozer and Thorsteinsson, 1964), northwestern Devon Island (Nassichuk and Davies, 1975; Kerr, 1976), and from many areas of Ellesmere Island (Thorsteinsson, 1974; Mayr, in press). The formation was deposited in a series of small subbasins during an active phase of continental extension associated with Late Carboniferous to Early Permian rifting (Beauchamp et al., 1989b). One of these subbasins, on northeastern Melville Island (Weatherall Depression) has been shown, through field and seismic analysis, to be a classical half-graben, with its master listric fault dipping cratonward (Harrison et al., 1988).

Half-grabens are now considered by most authors to be the dominant feature of all extensional basins (A.W. Bally, pers. comm., 1990), a conclusion reached through the acquisition of unequivocal evidence from seismic data over the last two decades, and through the study of modern analogues such as the East African Rift. The sedimentary fill of many continental and marine half-grabens has already been the focus of several classic studies and a number of models have been proposed (e.g., Leeder and Gawthorpe, 1987). These models are such that it is possible to interpret an ancient sedimentary package as the product of half-graben sedimentation without observing the actual half-graben structure. The models are especially useful for uplifted areas, where structural inversions occurred on the master listric fault (and, for that matter, any related synthetic or antithetic faults) that bounded the early extensional basins.

The present study was undertaken in order to understand the early depositional history of the Canyon Fiord Formation in west-central Ellesmere Island, an area where half-graben sedimentation is likely to have taken place in Late Carboniferous time. Extensive Tertiary deformation associated with the Eurekan Orogeny overprinted the synsedimentary structural style and, therefore, sedimentological data must be employed to interpret half-graben sedimentation. The style of sedimentation observed in the Canyon Fiord Formation appears to compare reasonably well with various sedimentary models of half-grabens, although the complexity of the depositional system is still under study.

This preliminary report focuses on the distribution, paleocurrents, and interpretation of facies assemblages in the Canyon Fiord Formation in an effort to reconstruct the synrift paleogeography throughout the area of its occurrence. Indispensable biostratigraphic data will be added later in order to further constrain some of the proposed interpretations.

GEOLOGICAL SETTING

Sverdrup Basin is an upper Paleozoic to middle Cenozoic successor basin underlying the Canadian Arctic Archipelago (Fig. 1, inset). It was initiated in the Early Carboniferous, with the faulting and collapse of the Franklinian Mobile Belt, and terminated in the middle Tertiary with the onset of the Eurekan Orogeny. Sverdrup Basin, which is about 1 000 km long and less than 400 km wide, is interpreted as a rift basin, based primarily on its intracontinental position, the nature of its thick and nearly continuous Viséan (Early Carboniferous) to Kungurian (Early Permian) succession, and the presence of associated volcanic rocks of extensional origin (Thorsteinsson, 1974; Beauchamp et al., 1989a,b; Cameron, 1989). Recently acquired direct and indirect evidence has shown that the basin margin was segmented longitudinally into a series of asymmetric half-grabens (Harrison and Riediger, in press; Thériault, work in progress), which is consistent with the findings of recent studies of modern and ancient rifts (Rosendahl et al., 1986; Frostick et al., 1988; Hamblin, 1989).

The Canyon Fiord Formation was deposited at the margin of Sverdrup Basin during the active phase of continental extension. The age of the formation is known to range from early Bashkirian (early Late Carboniferous) to middle Sakmarian (middle Early Permian), based on ammonoid, conodont, and foraminifer identifications (Thorsteinsson, 1974; Nassichuk, 1975; Beauchamp et al., 1989a; Henderson, 1989; Pinard, 1990). Three informal members have been recognized on Raanes Peninsula in southwestern Ellesmere Island (Beauchamp, 1987): i) a Lower clastic member, comprising fluvial conglomerate and sandstone; ii) a Middle limestone member, dominated by shelf limestone; and iii) an Upper clastic member, mostly consisting of shallow water, marine sandstone and limestone. Similar subdivisions have been identified on Melville Island (Harrison and Riediger, in press). Detailed sedimentological studies (Harrison and Riediger, in press; Thériault, work in progress) demonstrate that the Lower clastic member is, both on southwestern Ellesmere Island and northeastern Melville Island, derived from intrabasinal sources, its deposition being closely related to extensional faulting at the basin margin. The Middle limestone member accumulated during a second-order Moscovian (Late Carboniferous) marine transgression, probably as the result of fault related basin-floor collapse and subsidence. The Upper clastic member is believed to have originated from both intrabasinal and extrabasinal sources, likely during and following an active phase of growth fault development.

The study area (Fig. 1) extends from southern Fosheim Peninsula to east Tanquary Fiord, in west-central Ellesmere Island. Several lines of evidence suggest that this area was part of a rather large subbasin located at the margin of the main Sverdrup Basin. This northeasterly trending subbasin, herein referred to as the Fosheim-Hamilton subbasin, was bounded to the southeast by the craton; to the northeast by the Tanquary High (Nassichuk and Christie, 1969; Maurel, 1989); to the southwest by a positive feature of unknown origin, referred to as the "Bay Fiord High" (Beauchamp, 1987; Beauchamp et al., 1989a); and to the northwest by a poorly defined feature here referred to as the "Elmerson High". The existence of a structural high to the northwest is suggested by a significant thinning of exposed upper Paleozoic (Beauchamp, in progress) and basal lower Triassic (J. Devaney, pers. comm., 1990) formations that occurs toward Elmerson Peninsula, and beyond which these units thicken again northward.

The depocentre of the Fosheim-Hamilton subbasin migrated in a southwesterly direction through time, from the environs of north Greely Fiord in the Asselian-Sakmarian (Early Permian; Morin et al., 1991), to Mount Bridgman in the Artinskian (Early Permian; Scott et al., 1991), and

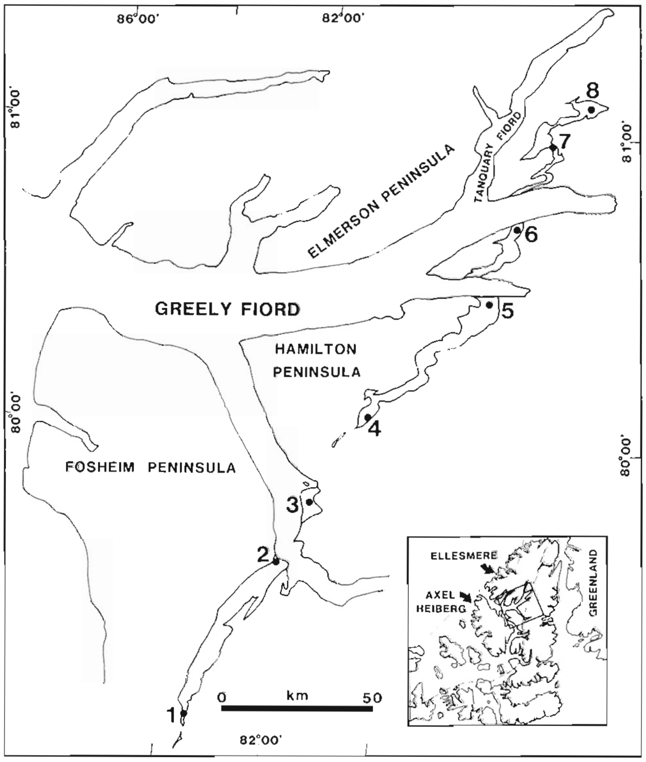

Figure 1. Location of study area and measured sections of the Canyon Fiord Formation on central Ellesmere Island. Dashed line in inset indicates the margin of Sverdrup Basin. Shaded area marks outcrop distribution of the Canyon Fiord Formation.

to "Notch Lake" in the Kungurian-Wordian (Early to Late Permian). Episodic closures of the Fosheim-Hamilton subbasin occurred in the Asselian and Sakmarian, resulting in the deposition of up to 400 m of relatively deep-water evaporites interfingering with fossiliferous carbonates (Antoinette and Mount Bayley formations; Thorsteinsson, 1974; Wallace and Beauchamp, 1990). The growth of rather large carbonate buildups in the vicinity of the "Elmerson High" is likely to have contributed to the restriction of the Fosheim-Hamilton subbasin (Beauchamp et al., 1991). The structural geometry of the Fosheim-Hamilton subbasin is at present unknown, and only speculations can be advanced as to whether the subbasin comprises one or more half-grabens.

STRATIGRAPHY

The upper Paleozoic succession of Sverdrup Basin consists of seven transgressive-regressive sequences, ranging in duration from 5 to 50 Ma (Beauchamp et al., 1989b). Each sequence consists of basin marginal facies that pass basinward into shelf and deep water facies. The Canyon Fiord Formation, located at the basin margin, encompasses the second and third sequences; its adjacent basinward correlatives include the Belcher Channel, Antoinette, Mount Bayley, and Tanquary formations. The marginal Canyon Fiord Formation represents the oldest Sverdrup Basin deposits of west-central Ellesmere Island, as it unconformably overlies deformed strata of the Franklinian Mobile Belt. The last orogenic event to affect Franklinian rocks was in the Late Devonian, when Pangaea accreted onto the northeastern margin of North America (Trettin, 1989).

Seven stratigraphic sections were measured during the 1990 field season, ranging in thickness from 290 to 1 010 m. Section 8 and the tops of Sections 4 and 6 were taken from Thorsteinsson (1974) and added to the stratigraphic profile for the sake of completeness and to facilitate correlations (Fig. 2). The total stratigraphic thickness decreases markedly toward the "Bay Fiord High" to the south, and toward the Tanquary High to the north.

In this study, the Canyon Fiord Formation is subdivided into three distinct facies assemblages:

1. A conglomerate assemblage (CF_1 in Figure 2), equivalent to the basal part of the Lower clastic member of Beauchamp (1987)
2. A sandstone assemblage (CF_2), corresponding to the upper part of the Lower clastic member
3. A sandstone/limestone assemblage (CF_3), encompassing both the Middle limestone member and Upper clastic member.

The Middle limestone member in west-central Ellesmere Island is not as distinct from the Upper clastic member as it is in other parts of the basin (e.g., Raanes Peninsula; Beauchamp, 1987), owing to active siliciclastic input throughout the Late Carboniferous. Accordingly, these two members have been combined into one sandstone/limestone assemblage, representing the intercalation of fluvial/marine sandstone and limestone.

As this preliminary study lacks biostratigraphic data, it is difficult to determine the nature of the stratigraphic relationships among the three facies assemblages. However, the vertical arrangement of these assemblages is similar at each locality, characterized by CF_1 overlain by CF_2, in turn overlain by CF_3 (Fig. 2). This succession appears to indicate that three distinct tectono-sedimentary phases affected the Canyon Fiord subbasin of deposition in west-central Ellesmere Island. However, it is at present impossible to know whether the boundary between each facies assemblage is time correlative across the study area (as suggested by the vertical succesion), or if these boundaries are diachronous, in which case complex facies relationships would instead exist. Other notable relationships displayed in Figure 2 are: conglomerate units are almost entirely restricted to CF_1, with only minor occurrences in CF_2 and CF_3; the entire Canyon Fiord succession is thickest at Caledonian Bay (Section 3); and this succession is markedly thin at the southern (Section 1) and northern (Section 8) limits of the study area. This bidirectional thinning reflects the southeasterly and northeasterly closure of the Fosheim-Hamilton subbasin.

The upper part of the succession consists of either marginal facies of the Canyon Fiord Formation, or of coeval deeper water facies of the Belcher Channel, Antoinette, Mount Bayley, and Tanquary formations (Fig. 2). Therefore, the current erosional margin in the areas south and north of D'Iberville Fiord and immediately east of Tanquary Fiord is not represented by basin marginal sediments. This implies that significant erosion of the Canyon Fiord Formation must

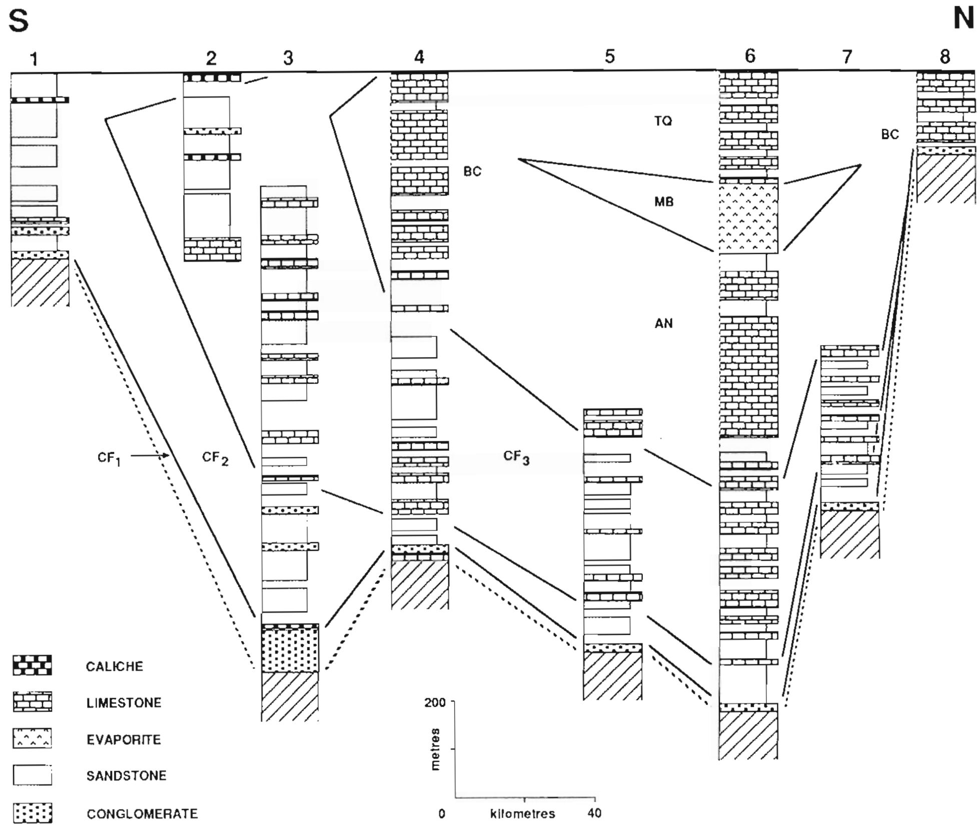

Figure 2. Stratigraphic cross-sections 1 to 8 of the Canyon Fiord Formation. See Figure 1 for locations. CF_1 — conglomerate assemblage; CF_2 — sandstone assemblage; CF_3 — sandstone/limestone assemblage). Correlative Belcher Channel (BC), Antoinette (AN), Mount Bayley (MB), and Tanquary (TQ) formations are also represented.

have taken place east of the current erosional margin in that area, resulting in a very incomplete sedimentary record. As a result, Sections 5 and 6 are considered distal compared to Sections 2 and 3.

SEDIMENTOLOGY

Conglomerate assemblage (CF_1)

The conglomerate assemblage occurs at the base of the Canyon Fiord Formation, and unconformably overlies deformed sandstones, mudstones, and limestones of the Franklinian Mobile Belt. Its thickness ranges from 3 to 82 m, but is commonly less than 10 m.

The assemblage consists predominantly of clast supported, pebble to cobble conglomerate (Fig. 3A). Beds are typically 30 to 60 cm thick, and display non-erosional to channelled basal contacts; individual channel structures are up to 1 m in depth and 5 m across. Normal grading is common, characterized by a well sorted, openwork base grading into a matrix-rich and finer grained upper part (Fig. 3B). The conglomerates are mostly crudely bedded, with local trough crossbedding. Several beds display a strong imbricated clast fabric, characterized by an a-axis orientation transverse to flow, and an imbricate b-axis dipping upstream *(a(t)b(i)* fabric of Harms et al., 1982). This type of imbrication indicates bedload transport by rolling, and has been observed in many modern rivers (Rust, 1972). Groove casts are common on well exposed bedding planes.

The maximum particle size (MPS), representing an average of the ten largest clasts, was systematically measured at several stratigraphic levels. MPS values range from 5 to 30 cm, and no fining- or coarsening-upward trends are apparent. Individual clasts are angular to rounded, composed of varicoloured chert, limestone, conglomerate, sandstone, and mudstone. Their compositions generally reflect that of the underlying basement rocks. The matrix is poorly sorted and sand rich, yet the presence of mud was sufficient to promote hematization, which imparted a reddish colour to the rock. Conglomerate intervals that display an openwork fabric are devoid of matrix, and blocky sparite represents the sole interstitial component.

Massive and laminar caliche fabrics are ubiquitous, and are found hosted by the conglomerate. They are commonly a few metres thick. These pedogenic horizons were developed during periods of low sedimentation rates, possibly under a strong control from channel avulsion (Kraus, 1987).

Interlayers of pebbly sandstone and sandstone are also common in the conglomerate assemblage. Their thickness is highly variable, ranging from a few centimetres to five metres. These intervals generally host nodular and massive beds of limestone, interpreted as caliche, again suggesting lower rates of sedimentation. Calcified rootlets are locally abundant, especially within nodular caliche beds.

The abundance of caliche beds, along with pervasive reddish colouration indicate a continental setting for the conglomerate assemblage. Furthermore, the common occurrence of sandstone interlayers, channel structures, normal grading, and an a(t)b(i) clast fabric is indicative of streamflow processes in the proximal to middle reaches of the braided stream environment. Debris flow deposits are absent, as indicated by the lack of matrix supported framework and the relatively good sorting of the conglomerates. Such deposits are common in modern alluvial fans in semiarid settings, and the lack of such deposits in the Canyon Fiord conglomerates may indicate the absence of alluvial fans at the exposed basin margin (Bull, 1977). Alternatively, the absence of debris flow deposits may reflect the two-dimensional nature of the Canyon Fiord outcrop belt. Such deposits could theoretically have existed to the southeast, and/or they may still be present in the subsurface to the northwest. Paleocurrent measurements of imbricate clasts and gutter casts consistently indicate a northeastward to southeastward flow direction, which is toward the cratonic margin and perpendicular to the basin's longitudinal axis (Fig. 4A). This pattern strongly suggests that sedimentation was controlled by one or more northeast/southwest trending structures, lying northwest or west of the outcrop area.

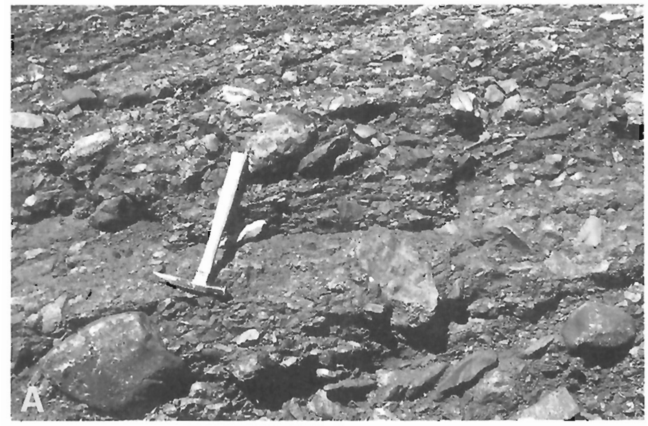

Figure 3. Conglomerate assemblage. **A.** Clast-supported pebble/cobble conglomerate with poorly sorted matrix. Hammer is 25 cm long. **B.** Normally graded bed displaying openwork fabric. Primary voids have been infilled with coarse sparite. Hammer is 25 cm long.

Sandstone assemblage (CF_2)

The sandstone assemblage overlies the conglomerate assemblage at every locality in the study area — except for Section 8 where it is absent. The thickness of this sandstone assemblage varies from 70 m at Section 4 to 370 m at Section 1. Two distinct lithofacies associations are present: i) interbedded sandstone/pebbly sandstone facies; and ii) interbedded sandstone/mudstone/caliche facies.

Interbedded sandstone/pebbly sandstone facies

This facies is dominated by medium to coarse sandstone, with subordinate granule to pebble sandstone and rare interbeds of conglomerate and mudstone. Colour variations are from deep brownish red on weathered surfaces to light reddish grey on fresh surfaces.

The sandstones are thin to thick bedded, and commonly display broadly channelled basal contacts. Pebbly sandstone generally infills such channel structures, which are up to 17 m wide and 2.5 m deep. Trough and tabular crossbedding and parallel stratification are very common throughout (Fig. 5A). Fining-upward sequences occur above scoured bases, and some are capped with discontinuous mudstone interlayers. The sediments are well sorted and composed essentially of rounded chert and quartz, indicating that they are texturally and mineralogically highly mature. Desiccation cracks and raindrop imprints have been observed on bedding plane surfaces, indicating periodic subaerial exposure.

The interbedded sandstone/pebbly sandstone facies is interpreted as braided-stream deposits. This setting is indicated by the common occurrence of broad and shallow channel structures, fining-upward sequences, and both scarcity and discontinuity of mudstone interlayers. Sediment texture and composition also suggest deposition in the distal portion of braided streams, with possible input from an extrabasinal source, as suggested by the high maturity of the deposits. A bimodal paleocurrent distribution was determined from measurements of primary current lineations (Fig. 4B). East-west orientations are predominant, indicating sediment transport across a very gently inclined topography. The poorly defined north-south trend may be related to the presence of an axial fluvial system, a common characteristic of continental half-grabens (Leeder and Gawthorpe, 1987). Other half-graben structures in Sverdrup Basin display similar fluvial drainage patterns (Harrison and Riediger, in press; Thériault, work in progress).

Interbedded mudstone/caliche/sandstone facies

This facies association is restricted to parts of Sections 1, 2, and 6. It is dominated by reddish mudstone, seldom well exposed, and resistant caliche beds. Reddish sandstone interlayers are commonly present at the base of fining-upward sequences.

The sandstones are medium grained, thin to thick bedded, and are 1 to 6 m thick. They either display crossbedding and parallel stratification, or are massive with the occasional presence of incipient caliche nodules. Such massive

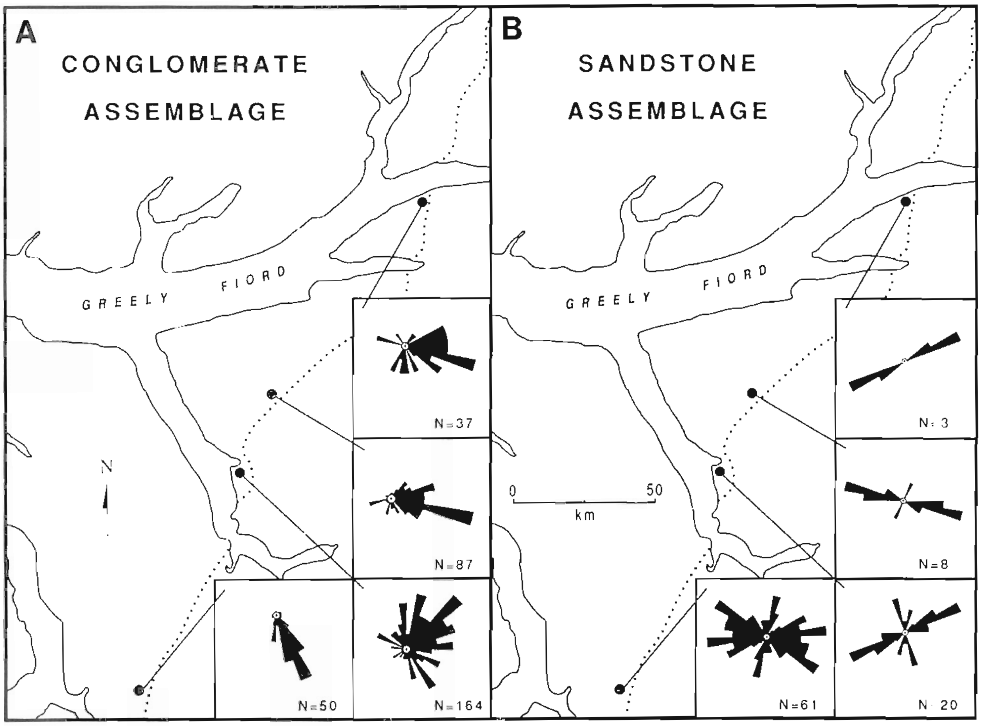

Figure 4. Paleocurrent orientations. **A.** Conglomerate assemblage; directional measurements are from imbricate clasts and gutter casts. **B.** Sandstone assemblage; bidirectional measurements are from primary current lineations.

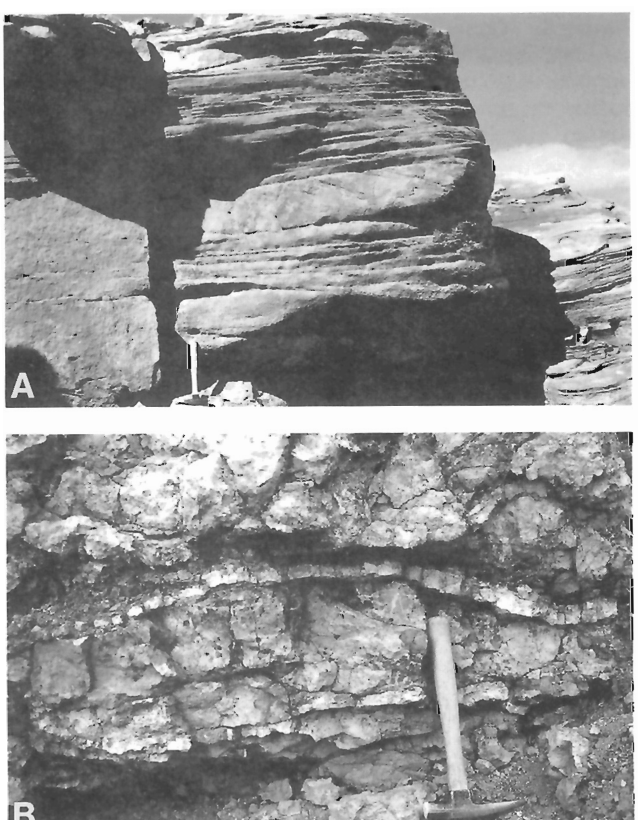

Figure 5. Sandstone assemblage. **A.** Braided stream sandstone, displaying planar-tabular crossbedding and parallel stratification. **B.** Massive and laminar caliche beds hosted by floodplain mudstone/sandstone. Hammer is 25 cm long.

sandstones are commonly the result of destratification and homogenization of the sediment by pedoturbation (Wright, 1989).

Mudstones gradationally overlie the sandstone intervals, and consist of interlayered shale and siltstone units that are commonly 5 m or more thick. They are thickly laminated to thin bedded, and may host thin beds of nodular to massive caliche. Vertical transitions into thick beds of caliche are common, the fabrics of which include chalky, nodular, massive, and laminar varieties (Fig. 5B). Nodular and massive beds predominate, with nodular beds typically grading upward into nodular-massive and massive beds. Such an upward increase in soil maturity is expected to develop on distal floodplains, where sedimentation rates are minimal and there are long periods of nondeposition (Allen, 1989). Some caliches are locally complex, arranged in composite profiles interpreted as reflecting periodic variations in sedimentation rates, or alternating states of deposition, nondeposition, and erosion. A composite caliche profile at Section 2 attains a maximum thickness of 15 m. Mudstone is believed to have been the original host for these caliches, but it is now limited to the nodular beds, having been completely displaced in the more mature massive and laminar beds.

The interbedded mudstone/caliche/sandstone facies was deposited in the floodplain environment of high sinuosity streams. This is shown by the low sandstone/mudstone ratio (average of 1:2), the arrangement of the facies in fining-upward sequences, and the ubiquitous and thick development of caliche beds.

Sandstone/limestone assemblage (CF_3)

The sandstone/limestone assemblage is up to 840 m thick, and overlies the sandstone assemblage. Its basal part is overall transgressive, and related to a low-order ($\sim 5 \times 10^7$ yrs) sequence of strong tectonic affinity (Beauchamp et al., 1989b). At a smaller scale, the assemblage consists of several high-order cycles (probably $< 2 \times 10^5$ yrs), averaging 5 to 8 m in thickness and characterized by intercalated sandstone, limestone, and mudstone. The basal sandstones are of two types: reddish and unfossiliferous sandstone, interpreted as fluvial in origin; and greyish sandstone with body and trace fossils, interpreted as shallow marine or coastal deposits. The two types seldom occur together within the same cycle.

The fluvial sandstones are red weathering, thin to thick bedded, and display crossbedding and parallel stratification. Desiccation cracks are locally common. The trace fossil *Beaconites*, an endichnial burrow, was observed at Section 4. It is interpreted as an escape burrow of the fluvial floodplain environment, as documented from the Old Red Sandstone of Ireland (Bamford et al., 1986). These fluvial sandstones were deposited during periods of low sea level.

The shallow water, marine, coastal sandstones are generally grey weathering, medium to very thick bedded, and commonly contain a wide variety of marine fossils. The most common body fossils are fusulinids, gastropods, nautiloids, brachiopods, bivalves, and solitary rugose corals (Fig. 6A). Trace fossils include *Planolites, Palaeophycus, Olivellites*, and *Macaronichnus*. Whereas *Planolites* and *Palaeophycus* have no environmental significance, *Olivellites* is found in shallow water, subtidal sandstones (Yochelson and Schindel, 1978), and *Macaronichnus* in intertidal to shallow water, subtidal sandstones (Clifton and Thompson, 1978).

Grainstones and packstones commonly overlie the fluvial and shallow water, marine sandstones gradationally. This sequence is marked by a gradual transition from reddish sandstone to greyish calcareous sandstone to sandy grainstone to relatively pure grainstone/packstone. These transitional facies occur over an interval of 20 to 50 cm. The grainstones and packstones are predominantly composed of fusulinids, brachiopods, echinoids, solitary and colonial rugose corals, and chaetetid sponges (Fig. 6B). They were deposited in the inner shelf environment, during periods of high relative sea level.

Many limestones are sharply overlain by mottled to reddish, poorly consolidated, terrigenous mudstones. Small nodules of caliche are locally common. These mudstones are interpreted as terra rossa paleosols, developed in part from the residual accumulation of insoluble material at the surface of subaerially exposed limestones (James and Choquette, 1984). The paleosol intervals are 0.3 to 1 m thick, and were developed during periods of low sea level. They are generally overlain by fluvial sandstones.

The sandstone/limestone assemblage consists predominantly of transgressive hemicycles formed in the fluvial to shallow water marine, paralic environment. These transgressive deposits are characterized by either one of the following two upward sequences: i) fluvial sandstone/coastal

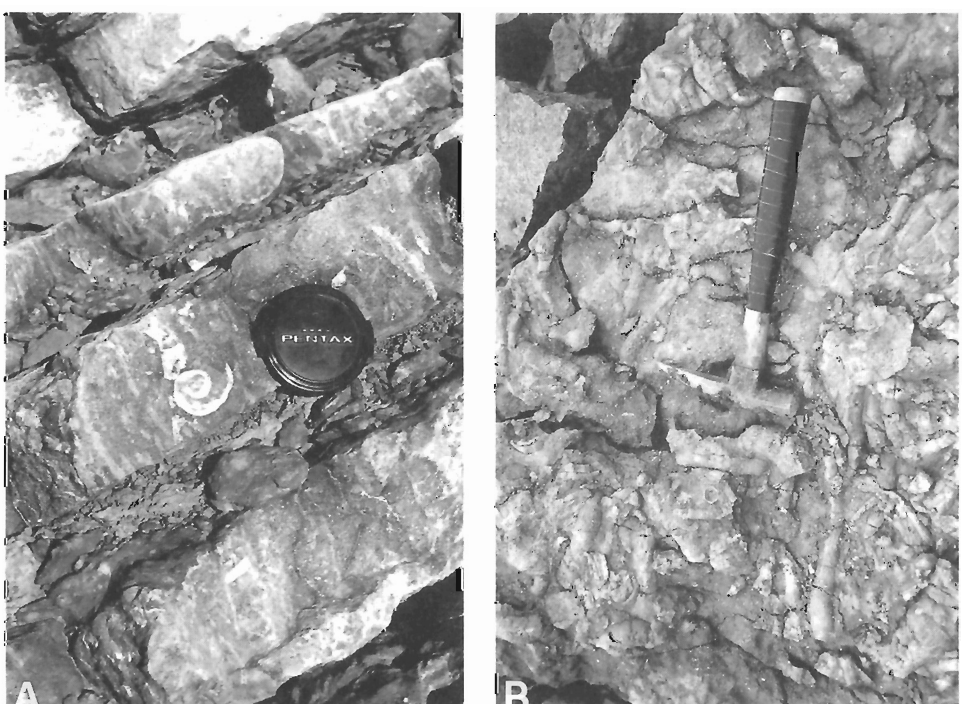

Figure 6. Sandstone/limestone assemblage. **A.** Mottled sandy grainstone with gastropods, interlayered with thin, red weathering, sandstone layers. Lens cap is 5 cm in diameter. **B.** Sandy grainstone with abundant solitary rugose corals.

sandstone/inner shelf limestone; or ii) coastal sandstone/inner shelf limestone. Individual hemicycles generally range from 4 to 10 m in thickness. The paleosols overlying most limestone intervals (< 1 m thick) probably developed during intervening regressive conditions.

Tectonism does not appear to have been an important factor in controlling cyclicity, in that the sediments in cycles are repetitive and transgressive, with each cycle forming a unit of constant thickness and rarely exceeding 15 m. Furthermore, coarse terrigenous sediments are essentially absent, although this may be due to the lack of proximal exposures. Delta-lobe switching is also an unlikely mechanism, as, under this scenario, fluvial sandstones would not be as widespread, and prograding deltaic sequences have not been identified. Moreover, cyclicity could have been controlled by glacio-eustatic sea-level fluctuations. A similar predominance of transgressive deposits over regressive deposits has been reported from nearshore sequences and areas of coastal onlap controlled by glacio-eustacy (Vail et al., 1977; Beauchamp, 1987).

Primary current lineations and planar-tabular crossbeds were measured from the fluvial sandstones to determine paleocurrent orientations. A few measurements (not enough to be illustrated) suggest an east-west orientation, similar to that of the underlying sandstone assemblage. Sedimentation patterns may have been controlled by a paleoslope of regional extent at the eastern margin of the basin, and the source of the sediments was partly or totally extrabasinal (as suggested by the high maturity level of the sandstones).

DISCUSSION

The Canyon Fiord Formation of west-central Ellesmere Island is characterized by three distinct facies assemblages:

1. A basal conglomerate assemblage, deposited eastward in the proximal braided stream environment
2. A sandstone assemblage, deposited in east-west and possibly north-south directions in braided stream and floodplain environments
3. A sandstone/limestone assemblage, deposited in the paralic environment as high-order relative sea-level cycles.

The distribution and dispersal patterns displayed by the three facies assemblages suggest that sedimentation could have occurred within one or a series of tectonically active, asymmetric half-grabens, which is the setting for the formation elsewhere in Sverdrup Basin.

Unfortunately, Tertiary deformation in the study area has led to a considerable modification of Carboniferous structural patterns, to the extent that half-graben structures of that age are no longer readily recognized. In the absence of structural data demonstrating the occurrence of a half-graben, the most important sedimentary features that suggest half-graben sedimentation are (Fig. 7):

1. Paleocurrents: sediments enter half-grabens from various directions that, in general, are either perpendicular or parallel to the master listric fault
2. The distribution of facies assemblages across the sedimentary wedge — one should see well defined zones characterized by very different assemblages
3. The geometry of the sedimentary package — this package should form a wedge thickening toward the presumed master listric fault.

Some, but not all, of these features are present in the Canyon Fiord succession in the study area. Firstly, the paleocurrents are in accordance with a half-graben setting: the conglomerates were clearly derived from the west and transported cratonward perpendicular to the presumed rift axis (Fig. 4); likewise, most sandstones were transported along an east-west axis, some along a north-south axis. Secondly, both the conglomerate and sandstone assemblages can be viewed as typical half-graben fills, although their lateral distribution is impossible to assess due to poor three-dimensional control on stratigraphic relationships. Thirdly, thickness variations are evident in the formation across the study area from north to south, and the possibilty that the thicker southern part is more proximal to the axis of subsidence must not be ruled out. Unfortunately, the narrow

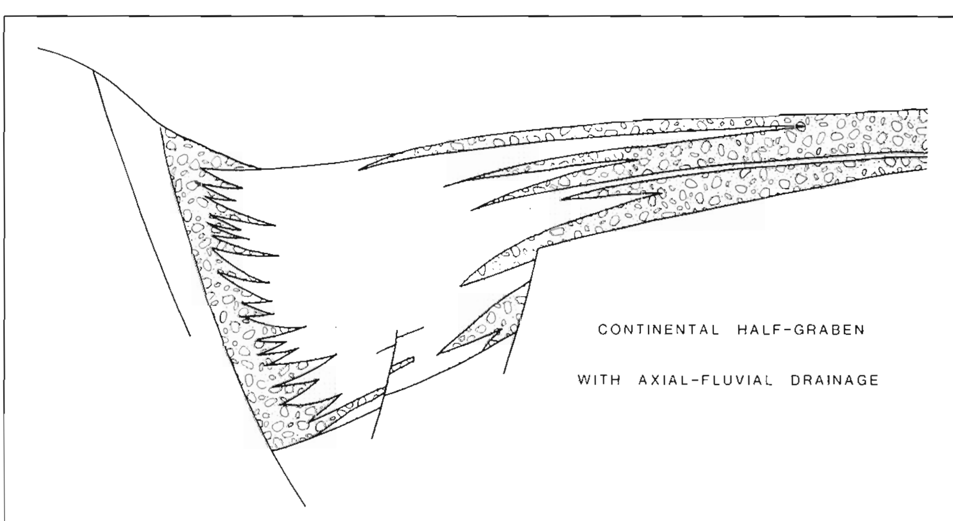

Figure 7. Model of half-graben sedimentation. (Modified after Leeder and Gawthorpe, 1987.) Stippled pattern indicates sandstone; other pattern is conglomerate.

outcrop belt limits our understanding of east-west facies relationships, and the key for assessing the geometry of the sedimentary package remains concealed in the subsurface.

The lack of stratigraphic control also prevents the determination of the true relationship of the conglomerate assemblage to the overlying sandstone assemblage. Based on half-graben models, these two assemblages should, in part, be contemporaneous. It can be suggested that the conglomerate assemblage passes laterally into the sandstone assemblage, but this suggestion cannot be verified without biostratigraphic data. It is equally plausible that these two assemblages were deposited during two separate episodes; for example, an early syn-faulting episode (conglomerate assemblage) followed by a post-faulting episode (sandstone assemblage).

Most continental half-graben fills are characterized by both coarse and fine grained clastic facies assemblages. It was proposed in early models that the bulk of the coarse fraction enters the graben from its footwall side, through a series of fans across the listric fault escarpment. More recent studies have shown that, in most cases, the graben fill, including a substantial coarse fraction, is derived from the hanging wall ramp side (Fig. 7) (Leeder and Gawthorpe, 1987). These studies have emphasized the importance of the actual depositional relief (as opposed to the fault throw) on the fault escarpment, and the rather short drainage area associated with the footwall block. The lack of very coarse alluvial fan deposits in the study area could indicate a low depositional relief near the various fault escarpments. Alternatively, coarse deposits may have existed to the east but subsequently have been removed by erosion, or they may still exist in the subsurface west of the outcrop belt.

Sedimentological observations can be accounted for by both ''down-to-the-craton'' and ''down-to-the-basin'' half-grabens. Dispersal patterns agree with a down-to-the-craton geometry, consequently implying that the eastward-flowing conglomerate assemblage is derived from a footwall escarpment, and that the sandstone assemblage is predominantly derived from the shallow, hanging wall ramp. In contrast, these same dispersal patterns do not preclude a down-to-the-basin geometry, whereby both the conglomerate and sandstone assemblages would have been mainly derived from the hanging wall ramp. It is thought that the sandstone assemblage, as well as the sandstone portion of the sandstone/limestone assemblage, are, regardless of half-graben geometry, predominantly derived from an extrabasinal source to the east, as evidenced by the relatively high compositional and textural maturity of these sediments. This last scenario reflects the possible interplay between intrabasinal versus extrabasinal sources of terrigenous detritus in Sverdrup Basin during its late Paleozoic history.

Whether the Canyon Fiord Formation sediments were, in the study area, deposited within a single half-graben or a whole series of such depressions has yet to be determined. Although it is now fairly clear that the area surrounding Fosheim and Hamilton peninsulas was the locus of a rather large subbasin in late Paleozoic time, the structural configuration of this subbasin is still nebulous. It appears unlikely, however, that this subbasin was created through the growth of a single half-graben. This conclusion can be reached by mere comparisons between the size of the Hamilton-Fosheim subbasin (75 km wide and 250 km long) and the size of the known half-graben on Melville Island (Weatherall Depression: less than 20 km wide and 60 km long). A dozen Weatherall Depression-like half-grabens could fit in the Fosheim-Hamilton subbasin, and even more if Tertiary telescoping is restored.

CONCLUSIONS

The Canyon Fiord Formation consists, in west-central Ellesmere Island, of three distinct facies assemblages, genetically related to the early rifting history of Sverdrup Basin:

1. A conglomerate assemblage, deposited eastwardly in proximal braided streams
2. A sandstone assemblage, deposited in east-west and north-south orientations in braided stream and floodplain environments
3. A sandstone/limestone assemblage, deposited in the paralic environment as high-order sea-level cycles. This assemblage can be correlated to a basinwide second-order transgression.

The distribution and dispersal patterns displayed by the three facies assemblages suggest that sedimentation could have occurred within one or a series of asymmetric half-grabens, which is the tectono-sedimentary setting of the Canyon Fiord Formation elsewhere in Sverdrup Basin. The relief on the fault escarpment must have been fairly minor at any given time, as evidenced by the relatively small thickness of the conglomerate assemblage and apparent absence of alluvial fan deposits. An alternative explanation may be that alluvial fan deposits and thicker conglomerates either still occur in the subsurface to the west or have been eroded away to the east. Because of outcrop limitations, it is at present impossible to determine whether one or more half-grabens were involved in Canyon Fiord sedimentation, and whether the master listric fault bounding these structures was oriented toward the craton or the basin.

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Unique mollusc find in the Beaufort Formation (Pliocene) on Meighen Island, Arctic Canada

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Abstract

Marine bivalve species of Pacific origin occur together with a bivalve of Arctic-North Atlantic affinity at one locality on Meighen Island in the Canadian Arctic Archipelago. This faunule occurs in sediment with reversed magnetic polarity in the Pliocene Beaufort Formation. Overlying fluvial sands contain plant and insect macrofossils indicating forest-tundra boundary conditions. This first known co-occurrence of Pacific and Arctic-Atlantic marine organisms in late Cenozoic strata of the Arctic Ocean region resulted from and closely followed opening of Bering Strait. The same event is recorded in the well known Tjörnes sequence in Iceland, where it is marked by the first appearance of many molluscan taxa of Pacific origin. Paleomagnetic and age data at Tjörnes would place the event either at about 3 Ma, in one of the two reversed subchrons of the Gauss Normal Polarity Chron, or in the Gilbert Reversed Polarity Chron (i.e., older that 3.4 Ma). Chronological and paleomagnetic information presently available for Meighen Island favours the first of these age alternatives.

Résumé

Une localité située dans l'île Meighen de l'archipel arctique canadien, des espèces de bivalves marins provenant du Pacifique coexistent avec un bivalve d'affinité arctique-nord-atlantique. Cette faune locale loge dans des sédiments à polarité magnétique inverse dans la Formation de Beaufort du Pliocène. Les sables fluviatiles sus-jacents contiennent des macrofossiles de plantes et d'insectes indiquant des conditions de limite forêt-toundra. Cette coexistence d'organismes marins du Pacifique et de l'Arctique-Atlantique, établie pour la première fois dans des couches du Cézonoïque tardif de la région de l'océan Arctique, a été causée par l'ouverture du détroit de Bering qu'elle a suivie de près. Le même événement a laissé sa trace dans la séquence bien connue de Tjörnes en Islande où il se manifeste par l'apparition de nombreux taxons de mollusques du Pacifique. Selon les données paléomagnétiques et chronologiques recueillies à Tjörnes, l'événement se situerait soit à environ 3 Ma, dans l'un des deux sous-chrons inverses du chron de polarité normale de Gauss, soit dans le chron de polarité inverse de Gilbert (c'est-à-dire avant 3.4 Ma). Les données chronologiques et paléomagnétiques recueillies à ce jour dans l'île Meighen favorisent la première de ces deux possibilités.

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INTRODUCTION

This preliminary report, arising from brief field investigations of the Pliocene Beaufort Formation on Meighen Island in the Canadian Arctic Archipelago, deals mainly with the implications of a single site where Marincovich has recognized co-occurrence of Pacific and Arctic-Atlantic molluscs. This co-occurrence of Pacific and Arctic-Atlantic bivalves is presumed to have occurred as an exceedingly brief event directly following opening of Bering Strait.

Occurrences of the Beaufort Formation on Meighen Island are unique because they include marine sediments containing marine fossils believed to indicate Pliocene age. In contrast, all other exposed occurrences of the Beaufort Formation on the Arctic Islands (Fig. 1), including its type locality on Prince Patrick Island (Fyles, 1990; Matthews et al., 1990) consist only of fluvial sands containing wood and plant and insect fossils.

Meighen Island is underlain by horizontal, unlithified sandy and clayey sediments originally assigned to the Beaufort Formation by Thorsteinsson (1961). These strata are approximately 200 m thick above sea level and are presumed to form the uppermost part of a 3000 m succession of Tertiary clastic sediments penetrated by a petroleum exploration well on the island (Asudeh et al., 1989). Plant macrofossils, investigated by Kuc (1974), Hills (1975), and Matthews (1987), indicate sparse coniferous forest and forest-tundra boundary conditions. Fossils of insects (Matthews, 1977) support this conclusion. Clay and silt beds have yielded marine shells (Fyles, 1962; Hills and Matthews, 1974) of the Atlantic bivalve *Arctica* (L. Marincovich, Jr. and K. McDougall, personal communication, 1980), as well as benthic foraminifers including *Cibicides grossus* (McNeil, 1990), which is Pliocene in age.

Earlier reports concerning the Beaufort Formation on Meighen Island, based on sites bordering Bjaere Bay (Fig. 2), refer to fluvial sand interbedded with concentrations of plant material and underlain by marine clay. Matthews (1987, p. 74) reported that "the clay is concentrated in a marine tongue that approaches 100 m above sea level in the southern part of the island", and Hills and Matthews (1974, p. 90) noted that this unit "is underlain and overlain by cross-stratified sands and gravels typical of the Beaufort Formation." Investigations during 1990 in the same part of the island amplify and generally support the foregoing, as indicated in the tentative stratigraphic succession outlined in Table 1. Figure 3 shows the strata listed in this table as they are exposed at several sites that have been investigated (or reinvestigated) in 1990, including the paleomagnetic sample sites. Although the "muddy beds" in these and other localities have a number of features in common and in several places include marine shells, they have not been proved to make up a single stratigraphic unit or to record a single event. In this context, the term "a marine tongue", cited above, is not used in this report.

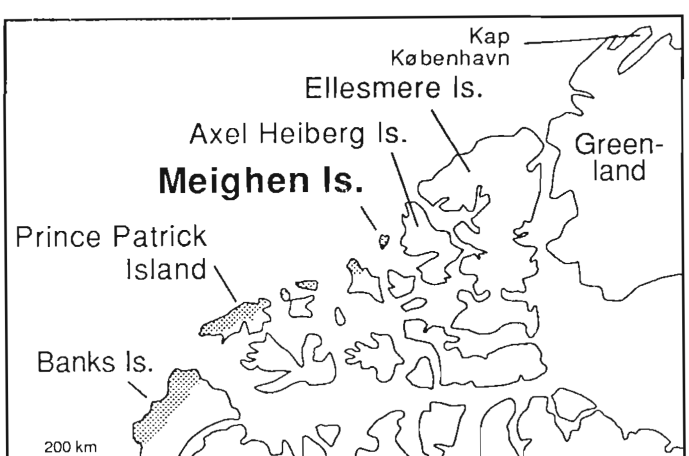

Figure 1. Location of the Beaufort Formation (stippled areas) and of Meighen Island.

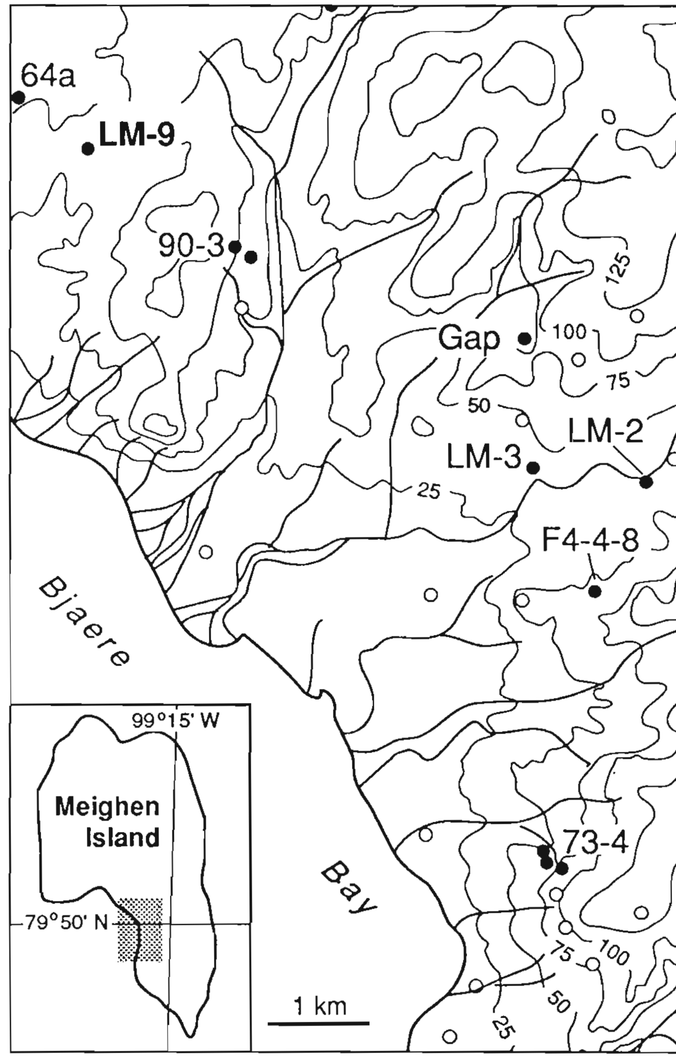

Figure 2. The Bjaere Bay area of Meighen Island. Black dots mark sites referred to in this report, particularly Figure 3. Open circles locate other sites investigated in 1990 and earlier. Contour interval 25 m.

LM - 9 SITE

At the LM - 9 site Marincovich discovered marine bivalve species of Pacific origin occurring together with a bivalve of Arctic-Atlantic affinity. This site (USGS Menlo Park

Table 1. Inferred stratigraphic succession for the exposed part of the Beaufort Formation in the Bjaere Bay region of Meighen Island.

Thickness (m)	Section	Description
100	mainly sand ? limited exposure	sand horizontally stratified, crossbedded, contains wood and other plant remains; a few beds of clay are interbedded with the sand
20	sand	sand quartz-rich, horizontally stratified, mainly medium to coarse grained, crossbedded; contains pebbles, abundant wood and beds of varied plant material
10-20	muddy beds	clay and silt, commonly with interbeds of quartz-rich sand; at some sites includes beds with deformed stratification and irregular, deformed bodies of mixed sediment; contains bivalve shells at some sites; conformable with overlying and underlying sand
10	sand	at one site only, beneath muddy beds; sand medium grained, quartz-rich, crossbedded with wood and other plant material
50	mainly covered; isolated sand, clay	sand (as isolated exposures of a few metres) similar to that higher in the section; clay (in one valley only) at least 30 m thick with top about 10 m lower than base of muddy beds nearby; this clay yielded the *Arctica* shells and foraminfers reported in earlier papers

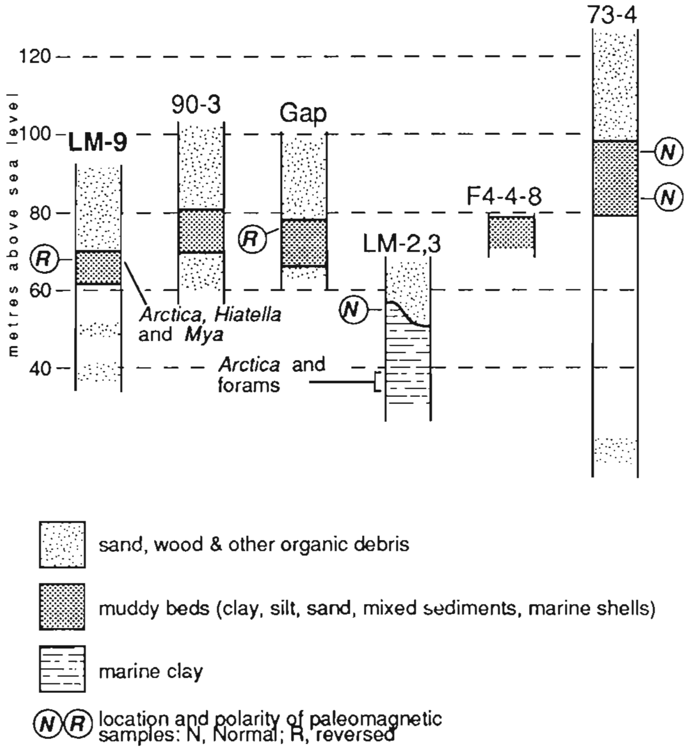

Figure 3. Stratigraphic sections at several sites investigated (or reinvestigated) in 1990; locations of these sites are marked on Figure 2.

Cenozoic Locality M9297), is located at 79°53.6'N and 91°25.7'W as shown in Figure 2. The shells occur approximately 65 m above sea level in the "muddy beds" of the tentative Meighen Island stratigraphic succession (Table 1). Figure 4 is a general view of the site; the colour change extending horizontally across the upper part of the photograph marks the top of the muddy beds which are exposed in the gully between the person and the right margin of the photograph. In this gully (Fig. 5) about 1 m of massive dark clayey silt is underlain by 2 m of medium grained quartz sand with deformed crossbedding and by 3 m (base not exposed) of mixed sediments. This succession is shown diagrammatically in the lower part of Figure 6. The mixed sediments are dark brown to dark grey and consist of irregularly layered to massive pebbly, sandy, silty, clayey materials (Fig. 5B, C) and include shell-rich layers and lenses (Fig. 5C). The shells are abundant and include whole valves and paired shells but break into fragments when extracted from the enclosing material. These varied deposits may have resulted from slumping during and following sediment accumulation at the seaward margin of a delta.

The upper part of the LM-9 section, above the muddy beds described in the preceding paragraph, consists of sand containing much plant material. This sand unit extends upward at least to the ridge top 20 m above the clay, as shown diagrammatically in Figure 6. The sand is well sorted, quartz-rich, medium to coarse grained and some layers include pebbles. Large-scale crossbedding is common. A bed of moss 25 cm thick lies in the sand 2 m above the clay. A metre higher, a discontinuous bed of sticks (maximum diameter 10 cm) is followed by a 20 cm bed of fine plant detritus. This layer has been traced more-or-less horizontally along the hill face for several tens of metres. Plant macrofossils contained in a sample from this bed are reported below. Both horizontal and inclined (crossbedded) plant layers in the sand higher on the face are poorly exposed and have not been investigated.

Fossils - muddy beds

Samples from the mixed sediments near the bottom of the exposures at LM-9 have yielded pelecypod fossils as well as foraminifera, ostracoda, amber, hydrozoan thecae, rare seeds of *Hippuris, Potamogeton,* and *Silene* and a single fragment of a beetle elytron *(Pterostichus)*.

Dominating the pelecypod shells are the extant species *Hiatella arctica* Linnaeus and *Mya truncata* Linnaeus which migrated through the Arctic Ocean and into the north Atlantic following the opening of Bering Strait. Also present is the bivalve *Arctica* sp. that is of Arctic-Atlantic affinity. All specimens from Meighen Island, except some *Hiatella arctica*, are fragmented, presumably from the annual freeze-thaw cycle, but all are unambiguously identifiable and common at the locality. The *Arctica* species is presumably *A. islandica* (Linnaeus), the only living species of the genus, or a closely related species or subspecies. The subcircular outline, dimensions, and moderate inflation of *Arctica* valves observed in the field and the preserved hinge fragments on hand indicate no differences from *A. islandica*, although conspecificity is not certain. Living *A. islandica* dwell in temperate waters. In the western Atlantic it ranges geographically from Cape Hatteras, North Carolina (35°N) to southern Newfoundland (47°N), but is infrequent north of Northumberland Strait, Nova Scotia (46°N) (Nicol, 1951). In the eastern Atlantic the species is abundant around Iceland (65°N), and from Britain and northern France to northern Norway, with occasional records off the Kola Peninsula (70°N) and in the White Sea (65°N), due to the warming influence of the Gulf Stream, and south to the Gulf of Cadiz, Spain (37°N) (Nicol, 1951). *Arctica islandica* is evidently unable to live in water at 0°C, which excludes it from the Arctic Ocean and cold parts of the North Atlantic (Arcisz et al., 1945), but it can tolerate temperatures up to about 19°C (Nicol, 1951). The species is always or nearly always found on substrates of mud or sandy mud (Turner, 1949; Nicol, 1951). *Arctica islandica* has an aggregate depth range from the lowest intertidal zone to 500 m, but is most common in depths of 10 to 280 m. Arcisz et al. (1945) observed the densest populations at depths of 25 to 45 m off of Rhode Island, with no specimens living in less than 18 m depth. According to Nicol (1951), in colder habitats the species is more abundant in shallower water.

Hiatella arctica and *Mya truncata* are well known, geographically widespread species with North Pacific histories extending back to the late Oligocene or earliest Miocene for the former (Allison and Marincovich, 1981) and to the middle Miocene for the latter (MacNeil, 1965). Both species also are circumarctic, so show a broader thermal tolerance than *Arctica islandica*. *Mya truncata* is an infaunal species in substrates of sandy mud at depths of 0-50 m (Keen and Coan, 1974). *Hiatella arctica* is an epifaunal nestler, with a maximum depth range of 0-120 m (Keen and Coan, 1974), but only 5-50 m in northern regions such as eastern Greenland (Ockelmann, 1959) and northern Alaska (Foster, 1981). It is common, however, in shallow-water faunas (0-50 m) in countless fossil and modern faunas in the North Pacific and Arctic oceans.

Fossils - sand unit

The sands overlying the marine sediments at LM-9 contain layers of sticks, mosses, and fine plant detritus. Another exposure near the LM-9 site also contains fine plant detritus within a few metres of the contact with the marine sediments. There is no evidence to suggest that the sands are greatly different in age from the underlying marine sediments; therefore, the fossils from the organic zone provide information on the environment on land shortly after the deposition of marine sediments.

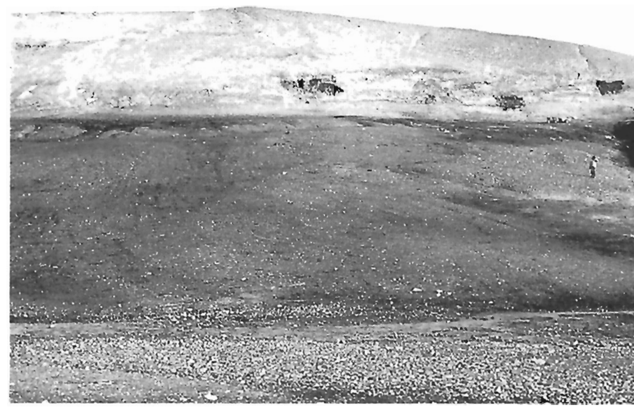

Figure 4. General view of the LM-9 site. The pale upper part of the hill face (about 20 m high above the horizontal colour change) consists of sand containing plant fossils. The dark material forming the lower part of the face, exposed in the small gully at the right, comprises the muddy beds, containing marine shells. 205318D

The organic debris includes fossils of both plants and insects. Many of the tree stumps are small and have extremely narrow rings indicating slow growth, and the other plant remains represent shrubs and herbs that grow in open sites. Similar assemblages of plant fossils have been found at other Meighen Island sites (Matthews, 1987) and also represent mixed coniferous-deciduous forests such as might be found near treeline. The major difference between the fossil flora and a modern treeline flora is that the former was richer; for example, it included up to six different types of conifers, one of them, a species (probably extinct) of pine in the five-needle subsection *Cembrae* (Critchfield, 1986). *Cembrae*-type white pines have not been recorded previously from Meighen Island (Matthews and Ovenden, 1990).

Insect fossils are well preserved and abundant. They also include many of the forms typical of the Beaufort Formation at other sites on Meighen Island, and exposures of the Beaufort Formation on other islands (Matthews, 1977). One of them is an extinct species of the ground beetle genus *Diacheila* (Matthews, 1979). Though not yet described, this species may eventually have chronological significance. For example, it is the only species of *Diacheila* from the Beaufort Formation on Meighen Island, whereas at the 2-2.5 Ma Kap København site on northern Greenland, only the extant species *Diacheila polita* has been found (Bennike and Böcher, 1990). The Kap København fauna and flora also represent a treeline forest, but unlike Meighen Island, the insect fauna contains bark beetles and several other taxa that are associated with trees. Bark beetle fossils are rare in Meighen Island assemblages and the only fossils from the sands immediately above the LM-9 marine sediments that imply presence of forests are ant mandibles and rare fragments of the ground beetle *Dromius*.

PALEOMAGNETIC DATA

Paleomagnetic measurements are being carried out on 163 samples collected from 7 sites on Meighen Island. So far, 30 samples have been selected for a pilot study and have been treated in detail. The natural remanent magnetization of these samples, which is a measure of all magnetization components present in the sample before treatment, reveals acceptable internal coherence (Fisher precision, K) and an intensity of magnetization well within measurement range of the Schonstedt spinner magnetometer used for this study.

The pilot samples were also measured after alternating field demagnetization in fields of 5, 10, 20, 40, 60, 80, and 100 millitesla (mT). Most specimens had less than 20 % of their magnetization remaining after 100 mT demagnetization. Initial inspection of declination, inclination, intensity, and K values after stepwise demagnetization would suggest that while polarity data can be derived from these samples, reliable directional data probably will not be forthcoming. The data so far fall into two groups: five sites with normal polarity and two with reversed polarity. The polarities marked in Figure 3 indicate the two sites with reversed polarity and four of the five sites with normal polarity (three sites are grouped at 73-4).

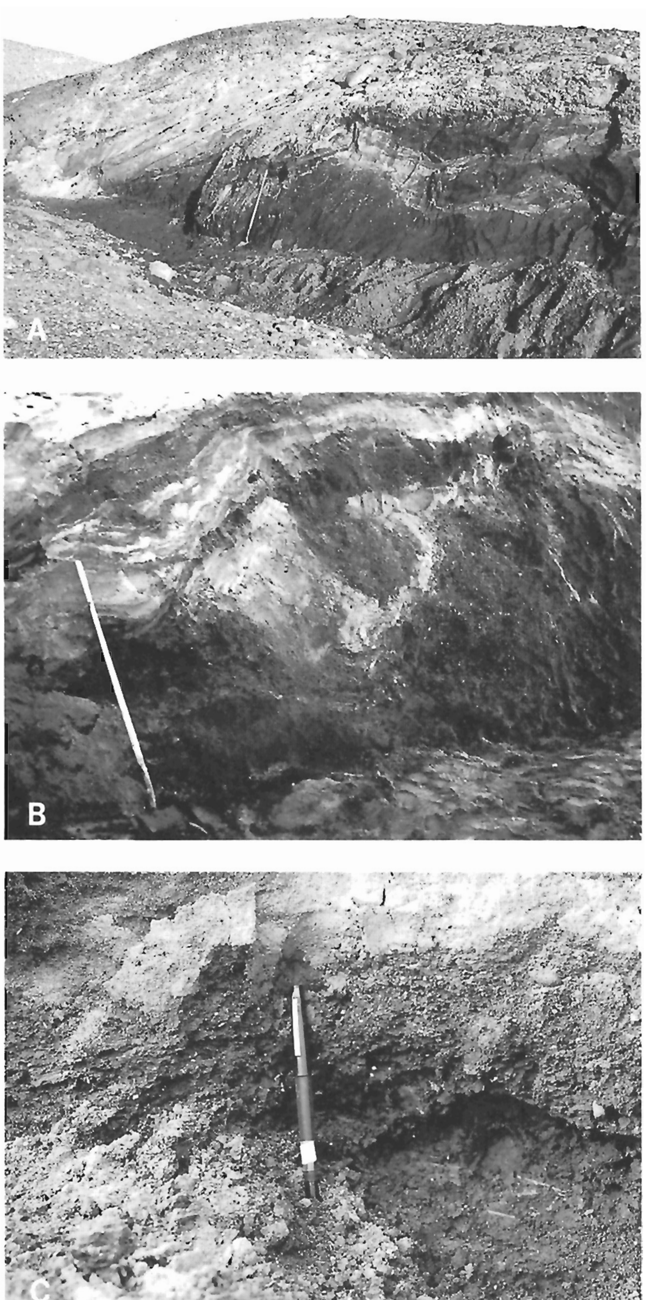

Figure 5. The muddy beds at LM-9 site: three views of the exposure illustrated diagrammatically in the lower part of Figure 6. Bivalve shells occur in the dark material (mixed sediment) near the shovel handle in 5A and 5B and surrounding the pen in 5C. 205318A, B, 205361B

DISCUSSION

Prior to the opening of Bering Strait, the Pacific and Arctic oceans had been separated by a land barrier, produced by the accretion of microplates, since about the late early Albian (about 110-105 Ma) or somewhat later (Marincovich et al., 1990). The formation of Bering Strait in the Pliocene produced a dramatic change in the composition of Arctic Ocean shallow-water marine faunas, because the migration of taxa was predominantly northward owing to prevailing surface currents. Durham and MacNeil (1967) noted that after Bering Strait opened 125 North Pacific invertebrates, mostly molluscs, invaded the Arctic-North Atlantic region but that only 16 species or species-groups of northern origin entered the North Pacific.

When Bering Strait opened, the indigenous Arctic Ocean molluscan fauna was not instantly (except in a geological sense) replaced by North Pacific taxa. For a short while after the opening of Bering Strait, species from both faunal realms probably comingled in the Arctic Ocean, before Arctic/North Atlantic species contracted their geographic ranges southward. The presence together of *Arctica* sp., *Hiatella arctica*, and *Mya truncata* on Meighen Island is the first example of this assumed faunal comingling and shows that this event did occur. In view of the short time during which Pacific and Arctic-Atlantic molluscs are presumed to have lived together in the Arctic Ocean after Bering Strait opened, the Meighen Island faunule is an exceedingly fortuitous discovery.

Dating of this and other sites associated with the opening of Bering Strait is of particular interest in inter-regional correlation and for establishing a chronology for the Pliocene. So far, the event is best dated in the Tjörnes beds of the Icelandic Tjörnes Sequence. In the Tjörnes beds, the majority of Pacific molluscs first appear at the base of the *Serripes* Zone (Einarsson et al., 1967; Eiríksson, 1981). Despite considerable research on this site, dating of the Tjörnes beds, and particularly the base of the *Serripes* Zone, is still uncertain. Magnetically reversed sediments and pillow lava at about the level of the base of the *Serripes* Zone could represent either one of the two reversed subchrons within the Gauss Normal Polarity Chron or part of the Gilbert Reversed Polarity Chron (Eiríksson et al., 1990). The first alternative would place the opening of Bering Strait (and the LM-9 faunule on

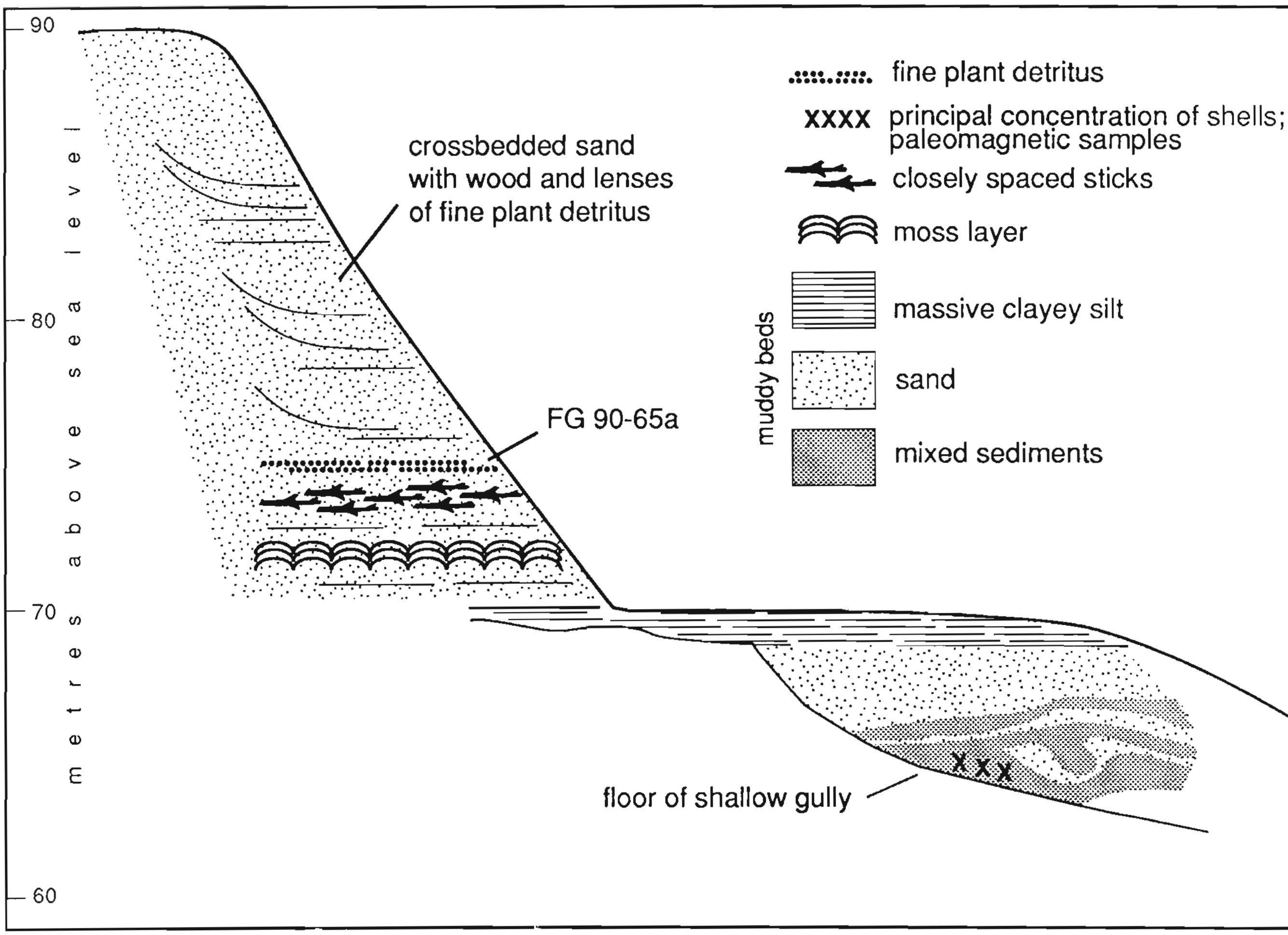

Figure 6. Diagrammatic sketch of the section at the LM-9 site.

Meighen Island) in the Kaena Subchron (2.91-3.00 Ma) or the Mammoth Subchron (3.03-3.17 Ma). The second alternative would make the opening of Bering Strait (and the LM-9 faunule) older than 3.40 Ma, which is the age of the Gauss/Gilbert boundary (McDougall, 1979). From the chronological and paleomagnetic information summarized below for Meighen Island, we favour the first of these alternatives as the more plausible age for the LM-9 faunule on Meighen Island.

Age and chronology, Meighen Island

Discovery of the LM-9 faunule coeval with the opening of Bering Strait leaves no doubt regarding the Pliocene age of the Beaufort Formation exposed on Meighen Island, confirming earlier age estimates from foraminifers (McNeil, 1990), strontium ratios (K. Miller, personal communication to J Fyles, 1989; Kaufmann et al, 1990), and amino acid D/L ratios (Brigham-Grette et al., 1987). Several lines of interpretation bear further upon the absolute age of the marine strata recognized in the Beaufort Formation on the Island.

1. Based on amino acid ratios for *Arctica* shells from LM-3, Brigham-Grette et al. (1987, p.3) inferred ''that ground temperatures dropped well below 0°C shortly after deposition and subsequent emergence of the marine sediments'' and that ''regional climate has remained as cold or colder since that time''. On this basis the forest-tundra boundary conditions recorded in the Beaufort Formation on Meighen Island could represent the last warm interval of the Pliocene, now dated at approximately 3.1 Ma (Dowsett and Poore, 1990: sea-surface temperature data from a North Atlantic core).

2. The marine deposits exposed on Meighen Island lie within a stratigraphic interval of less than 100 m in a sequence of completely unlithified and uncompressed sediments (Fig. 3). Thus, in the absence of specific information on the amount of time involved in accumulation of the exposed marine beds, we infer that they could have accumulated in a geologically short interval, perhaps during a single eustatic high. Coastal regions of the North Atlantic record such a high stand of sea level at 30-35 m above present between 3 and 3.5 Ma (Cronin, 1990).

3. The paleomagnetic polarity data from marine beds within the 100 m stratigraphic interval cited above include both normal and reversed polarity. Two sites with reversed polarity are stratigraphically above one normal site and are lower than other normal sites. Thus these initial data seem to indicate that the short (?) interval of time represented by the marine beds was characterized by magnetic reversal(s).

In summary, the presence of both reversed and normally magnetized marine sediments on Meighen Island; the reversed polarity of LM-9 faunule sediments; the reversed polarity lava and sediments near the base of the *Serripes* Zone in Iceland, and the circumstantial amino acid, climatic, and eustatic correlations combine to make 3 Ma a plausible age estimate for the LM-9 faunule and the marine sediments in the Beaufort Formation on Meighen Island.

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