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GEOLOGICAL SURVEY OF CANADA
MEMOIR 439

BASIN ANALYSIS, EUREKA SOUND GROUP, AXEL HEIBERG AND ELLESMERE ISLANDS, CANADIAN ARCTIC ARCHIPELAGO

Brian D. Ricketts

1994

Natural Resources Canada
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Appendix 1: Palynology
D.J. McIntyre

Appendix 2: Micropaleontology
J.H. Wall and D.H. McNeil

1994

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Cover description

Tidal flat and shallow shelf strata of the Cape Pillsbury Member, Iceberg Bay Formation, in a recumbent syncline, near the south end of Cañon Fiord (South Bay), Ellesmere Island. GSC 1993-054

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Preface

Upper Cretaceous and Paleogene strata of the Eureka Sound Group comprise an important tectonostratigraphic assemblage that records the demise of the long-lived Sverdrup Basin. This memoir discusses the evolution of three main types of sedimentary basins and provides details of their stratigraphic architecture, linking the major depositional packages, or sequences, to Arctic plate tectonics.

In addition to providing a sound geological framework of the eastern Arctic for this period of time, this study has important implications concerning the potential trapping mechanisms of hydrocarbons in older rocks of Sverdrup Basin, both onshore and beneath the modern Polar Continental Shelf. Significant coal reserves have also been identified in the Eureka Sound Group, the details of which have been published in a number of other articles.

Elkanah A. Babcock
Assistant Deputy Minister
Geological Survey of Canada

Préface

Les couches du Crétacé supérieur et du Paléogène du Groupe d'Eureka Sound comportent un important assemblage tectonostratigraphique qui témoigne de la disparition du bassin de Sverdrup dont l'existence a été de longue durée. Le présent mémoire traite de l'évolution de trois principaux types de bassins sédimentaires et présente des données détaillées sur leur architecture stratigraphique, en établissant un lien entre les principaux ensembles sédimentaires, ou séquences, et la tectonique des plaques dans l'Arctique.

En plus d'établir un cadre géologique solide de l'Arctique oriental pendant cette période de temps, l'étude a des incidences importantes sur les mécanismes de piégeage possibles des hydrocarbures dans les anciennes roches du bassin de Sverdrup, à la fois sur le continent et sous la plate-forme continentale polaire moderne. On a découvert des réserves de charbon d'une certaine importance dans le Groupe d'Eureka Sound. Les détails de ces découvertes sont publiés dans d'autres articles.

Elkanah A. Babcock
Sous-ministre adjoint
Commission géologique du Canada

CONTENTS

Appendixes

Illustrations

Tables

BASIN ANALYSIS, EUREKA SOUND GROUP, AXEL HEIBERG AND ELLESMERE ISLANDS, CANADIAN ARCTIC ARCHIPELAGO

Abstract

The Eureka Sound Group of the eastern Arctic Archipelago records, in its 40-45 Ma stratigraphic history, the death throes of Sverdrup Basin. Definitive ages for the >4000 m succession range from middle Campanian to middle Eocene, but may extend into the Late Eocene. Eureka Sound Group strata preserve the transition from thermally-dominated subsidence that characterized much of Sverdrup Basin's earlier history, to foredeep subsidence beginning in the Paleocene. The ultimate demise of the contiguous Sverdrup Basin during Early Eocene to Middle Eocene time resulted in discrete, synorogenic intermontane basins.

Five third-order depositional sequences have been extracted from the lithostratigraphic and lithofacies framework. Middle Campanian to Maastrichtian deltas in Sequence 1 reflect a continuation of typical Sverdrup Basin sedimentation (3 cm/ka) and subsidence. Throughout eastern Sverdrup Basin, Upper Maastrichtian, and in some places all Maastrichtian deposits, were eroded, marking the base of Sequence 2. The sub-Paleocene unconformity also signals the transition to significantly greater sedimentation rates (15 cm/ka) in the rapidly subsiding Sverdrup foredeep. The mid-Paleocene transgression initiating Sequence 3 appears to have extended over much of the Arctic; paleontological evidence further suggests a tentative connection with ancestral Baffin Bay at this time. In addition to the foredeep succession, Sequence 4 also contains synorogenic deposits in small, fault-bounded half grabens signalling the earliest effects of crustal failure within the foredeep. Sequence 5 (Middle Eocene), is entirely synorogenic and signifies the complete destruction of Sverdrup Basin. Sedimentation rates in at least six intermontane basins were as high as 20 cm/ka.

Résumé

Dans son profil stratigraphique de 40 à 45 ma, le Groupe d'Eureka Sound dans l'est de l'archipel arctique atteste la disparition du bassin de Sverdrup. Les âges définitifs établis pour la succession de >4 000 m s'échelonnent du Campanien moyen à l'Éocène moyen, mais pourraient se prolonger jusqu'à l'Éocène tardif. Les couches du Groupe d'Eureka Sound témoignent du passage d'une subsidence principalement thermique, qui a caractérisé presque tout le début de l'histoire du bassin de Sverdrup, à une subsidence d'avant-fosse qui a débuté au Paléocène. La disparition finale du bassin de Sverdrup au cours de l'Éocène précoce-moyen a entraîné la formation de bassins intramontagneux syntectoniques distincts.

Cinq séquences sédimentaires de troisième ordre se reconnaissent à partir de la lithostratigraphie et des lithofaciès. Les deltas formés entre le Campanien moyen et le Maastrichtien dans la séquence n^o 1 reflètent une prolongation de la sédimentation (3 cm/ka) et de la subsidence typiques du bassin de Sverdrup. Dans tout l'est du bassin de Sverdrup, les sédiments du Maastrichtien supérieur et, par endroits, tous les dépôts maastrichtiens, ont été érodés, marquant la base de la séquence n^o 2. La discordance sub-paléocène indique en outre une nette augmentation des vitesses de sédimentation (15 cm/ka) dans l'avant-fosse de Sverdrup qui s'enfonçait rapidement. La transgression du milieu du Paléocène qui marque le début de la séquence n^o 3 semble avoir touché à la grande partie de l'Arctique; des indices paléontologiques révèlent en outre un lien possible avec la proto-baie de Baffin de l'époque. En plus de la succession d'avant-fosse, la séquence n^o 4 contient aussi des sédiments syntectoniques dans de petits demi-grabens limités par des failles qui indiquent les premiers effets de la rupture crustale au sein de l'avant-fosse. La séquence n^o 5 (Éocène moyen) est entièrement syntectonique et témoigne de la destruction complète du bassin de Sverdrup. Les vitesses de sédimentation dans au moins six bassins intramontagneux ont atteint un maximum de 20 cm/ka.

Summary

The Eureka Sound Group in the Queen Elizabeth Islands, Canadian Arctic Archipelago, contains the final 40-45 million year record of the long-lived (Late Carboniferous to Middle Eocene) Sverdrup Basin. Eureka Sound Group deposition began about middle Campanian time and ended in the Middle Eocene resulting in a composite thickness of more than 4000 m. Clastic sediment deposited during the middle Campanian and Maastrichtian resulted in a continuation of the stratigraphic style which characterized much of Sverdrup Basin. Regional uplift at the end of the Cretaceous and development of a basin-wide sub-Paleocene unconformity, signalled the end of this era and subsequent formation of a foredeep-like basin which survived until the Middle Eocene; sedimentation rates increased markedly at this time. Crustal shortening, beginning in the Early Eocene and culminating in Middle Eocene time (the Eurekan Orogeny) destroyed the contiguous Sverdrup Basin and in its place produced several small intermontane basins. Thus the Eureka Sound Group spans the gamut of tectonosedimentary regimes from pre-orogeny to late-orogeny.

Analysis of the Eureka Sound Group lithofacies, biostratigraphy, and unconformities provides the basis for recognition of five third-order sequences. Sequence 1 began with the Kanguk transgression during Turonian to early Santonian time, followed by an extended period of relative highstand deposition until the Maastrichtian. Wave-dominated deltas characterize the west Axel Heiberg region, with a thin remnant of coastal plain sediment preserved on north Fosheim Peninsula (Lower member, Expedition Formation). Contact between sequences 1 and 2 is the sub-Paleocene unconformity – a disconformity where it overlies older Sverdrup Basin strata, and an angular discordance over deformed Lower Paleozoic bedrock. This contact also marks the transition to foreland basin style deposition. Sedimentation rates during Sequence 1 averaged 3 cm/ka; rates during deposition of sequences 2 to 4 increased dramatically to average 15 cm/ka. Some or all of the Maastrichtian component of Sequence 1 was removed prior to deposition of Sequence 2.

A distinctive volcanic pebble conglomerate overlies the sub-Paleocene unconformity in several places and is inferred to have accumulated during the earliest Paleocene or latest Maastrichtian relative sea level lowstand (base of Sequence 2, Upper member, Expedition Formation). Subsequent transgression is manifested as valley-fill estuarine deposits. Overlying delta, coastal plain, and sandy shelf lithofacies mark the transition to progradation and a relative sea level highstand.

Contact between sequences 2 and 3 is placed at the base of a transgressive sandstone facies in the upper few metres of the Expedition Formation. The abrupt contact between the Expedition Formation and Strand Bay Formation shale represents a surface of maximum flooding. This flooding event in mid-Paleocene time seems to have been Arctic-wide, with counterparts recognized as far afield as Mackenzie Delta and Spitzbergen (Barents Shelf). The regressive part of Sequence 3

Sommaire

Le Groupe d'Eureka Sound dans les îles de la Reine-Élisabeth, dans l'archipel arctique canadien, correspond à la dernière tranche de 40-45 millions d'années du bassin de Sverdrup qui a existé pendant une longue période allant du Carbonifère tardif à l'Éocène moyen. La sédimentation du Groupe d'Eureka Sound a débuté vers le Campanien moyen pour se terminer à l'Éocène moyen, donnant une épaisseur composite de plus de 4 000 m. Les sédiments clastiques déposés durant le Campanien moyen et le Maastrichtien ont prolongé le style stratigraphique qui avait caractérisé la grande partie du bassin de Sverdrup. Le soulèvement régional à la fin du Crétacé et la formation d'une discordance sub-paléocène à l'échelle du bassin ont marqué la fin de cette ère et la formation subséquente d'un bassin d'avant-fosse qui a existé jusqu'à l'Éocène moyen; les vitesses de sédimentation ont nettement augmenté à cette époque. Le raccourcissement crustal, débutant à l'Éocène précoce et culminant à l'Éocène moyen (orogenèse eurékienne), a détruit le bassin de Sverdrup contigu et a produit à sa place plusieurs petits bassins intramontagneux. Par conséquent, le Groupe d'Eureka Sound représente la gamme des régimes tectonosédimentaires depuis le régime qui a existé avant l'orogenèse jusqu'à celui de la fin de l'orogenèse.

L'analyse des lithofaciès, de la biostratigraphie et des discordances caractérisant le Groupe d'Eureka Sound sert de base à l'établissement de cinq séquences de troisième ordre. La séquence n^{o} 1 a débuté par la transgression de Kanguk du Turonien au Santonien précoce, suivie par une longue période de sédimentation de haut niveau qui a duré jusqu'au Maastrichtien. La région occidentale de l'île Axel Heiberg se caractérise par des deltas de tempête et un reste mince de sédiments de plaine littorale dans le nord de la péninsule Fosheim (membre inférieur, Formation d'Expedition). Le contact entres les séquences n^{o} 1 et n^{o} 2 est la discordance sub-paléocène – une disconformité là où elle surmonte les plus anciennes couches du bassin de Sverdrup – et une discordance angulaire au-dessus du substratum rocheux déformé du Paléozoïque inférieur. Ce contact marque également la transition vers une sédimentation de bassin d'avant-pays. Les vitesses de sédimentation de la séquence n^{o} 1 ont atteint en moyenne 3 cm/ka; celles des séquences n^{os} 2 à 4 ont augmenté considérablement pour se situer en moyenne à 15 cm/ka. Une partie ou toute la composante maastrichtienne de la séquence n^{o} 1 a été érodée avant le dépôt de la séquence n^{o} 2.

Un conglomérat distinct à cailloux volcaniques repose sur la discordance sub-paléocène à plusieurs endroits, et l'on suppose par inférence qu'il s'est accumulé durant la période de bas niveau au Paléocène initial ou au Maastrichtien terminal (base de la séquence n^{o} 2, membre supérieur, Formation d'Expedition). La transgression subséquente a laissé des sédiments estuariens remblayant une vallée. Les lithofaciès sus-jacents de delta, de plaine littorale et de plate-forme sableuse marquent la transition vers une progradation et un haut niveau.

La contact entre les séquences n^{o} 2 et n^{o} 3 se situe à la base d'un faciès de grès transgressif dans les quelques mètres supérieurs de la Formation d'Expedition. Le contact abrupt entre la Formation d'Expedition et le shale de la Formation de Strand Bay représente une surface de crue maximale. Cette crue du milieu du Paléocène semble avoir recouvert tout l'Arctique, et des dépôts équivalents se rencontrent aussi loin que dans le delta du Mackenzie consists of fluvial-dominated delta facies on western Axel Heiberg Island (Strand Bay and Lower member, Iceberg Bay Formation), and coastal plain, muddy shelf, and calcareous tidal flat facies (Cape Pillsbury Member - new) in the eastern and southern parts of Sverdrup Basin. At Strathcona Fiord a unique calcareous foraminifera assemblage in the lower Cape Pillsbury Member has some affinities with Atlantic faunal elements, suggesting a tentative link with an ancestral Baffin Bay.

Differentiation between sequences 3 and 4 can be made at Strathcona Fiord area where the Cape Pillsbury Member is abruptly overlain by transgressive sandstone and shale of the Braskeruds Member (new), Iceberg Bay Formation. A significant subaerial unconformity is also present between oldest units of the Buchanan Lake Formation and the Iceberg Bay Formation at Otto Fiord and Emma Fiord. Elsewhere Sequence 3 and Sequence 4 are undivided. Most of Sequence 4 is represented by coal-bearing strata of the Iceberg Bay Formation which range in age from Early to Middle Eocene.

A key element of Sequence 4 is the presence of syntectonic conglomerate, belonging to the Buchanan Lake Formation, in small fault-bound basins at Otto Fiord and Emma Fiord, northwest Ellesmere Island, and possibly in thrust-related basins on eastern Ellesmere Island. This is the first indication of Eurekan deformation in the Sverdrup foredeep, that eventually led to the demise of Sverdrup Basin as a contiguous entity.

Crustal shortening that fragmented Sverdrup Basin reached a climax during the Middle Eocene. Deposition of coarse, commonly conglomeratic sediment within relatively small intermontane basins, was in several cases associated with major thrust systems (e.g. Stolz Thrust, Parrish Glacier Thrust, Lake Hazen Thrust). This period of intense tectonic activity and rapid sedimentation (rates up to 20 cm/ka) was short-lived, with most structural shortening over by the Middle Eocene. It is possible that tectonic activity continued into the Late Eocene but there is no unequivocal evidence for this in the onshore Arctic geology.

The transitions in basin form and subsidence correspond reasonably well to Arctic seafloor spreading events. Sverdrup Basin-style thermal subsidence persisted until the end of the Cretaceous, the transition to foreland subsidence coinciding with the inception of seafloor spreading in northern Labrador Sea. A significant increase in spreading rates and a concomitant change in spreading direction between 59 Ma and 56 Ma may have resulted in increased horizontal intra-plate stress and accelerated foredeep subsidence. This also coincides with possible initial compression on eastern Ellesmere Island (e.g. Lake Hazen Basin and Lake Hazen Thrust). Deposition of syntectonic Buchanan Lake conglomerate in Otto Fiord and Emma Fiord subbasins reflects local extension in areas that were remote from the main eastern Ellesmere deformation front. Rotation of Greenland relative to North America, underway by

et le Spitzbergen (plate-forme de Barents). La partie régressive de la séquence n^{o} 3 se compose d'un faciès de delta surtout fluvial dans l'ouest de l'île Axel Heiberg (Formation de Strand Bay et membre inférieur, Formation d'Iceberg Bay) et de faciès de plaine littorale, de plate-forme boueuse et de wadden calcaire (Membre de Cape Pillsbury – nouveau) dans l'est et le sud du bassin de Sverdrup. Au fjord Strathcona, un assemblage unique de foraminifères calcaires dans la partie inférieure du Membre de Cape Pillsbury présente certaines affinités avec des éléments fauniques de l'Atlantique, indiquant un lien possible avec une proto-baie de Baffin.

On peut différencier les séquences n^{o} 3 et n^{o} 4 au fjord Strathcona, là où du grès et du shale transgressifs du Membre de Braskeruds (nouveau) de la Formation d'Iceberg Bay surmontent nettement le Membre de Cape Pillsbury. Une importante discordance subaérienne se rencontre également entre les plus anciennes unités de la Formation de Buchanan Lake et la Formation d'Iceberg Bay au fjord Otto et au fjord Emma. Ailleurs, les séquences n^{os} 3 et 4 ne sont pas divisées. La grande partie de la séquence n^{o} 4 est représentée par des couches houillères de la Formation d'Iceberg Bay dont l'âge varie de l'Éocène précoce à moyen.

L'un des éléments clés de la séquence n^{o} 4 est la présence d'un conglomérat syntectonique de la Formation de Buchanan Lake dans de petits bassins limités par des failles au fjord Otto et au fjord Emma, dans le nord-ouest de l'île d'Ellesmere, et peut-être dans des bassins associés à des chevauchements dans l'est de l'île d'Ellesmere. C'est là la première indication d'une déformation eurékienne dans l'avant-fosse de Sverdrup, déformation qui a fini par faire disparaître la bassin de Sverdrup comme entité contiguë.

Le raccourcissement crustal qui a fragmenté le bassin de Sverdrup a atteint son point culminant durant l'Éocène moyen. Le dépôt de sédiments grossiers généralement conglomératiques au sein de bassins intramontagneux relativement petits a été, dans plusieurs cas, associé à d'importants systèmes de chevauchements (par ex., le chevauchement de Stolz, le chevauchement de Parrish Glacier, le chevauchement de Lake Hazen). Cette période d'activité tectonique intense et de sédimentation rapide (jusqu'à 20 cm/ka) a été de courte durée, la plus grande partie du raccourcissement structural ayant eu lieu avant l'Éocène moyen. Il est possible que l'activité tectonique s'est poursuivie au cours de l'Éocène tardif, mais il n'existe aucun indice non équivoque de cette activité dans la géologie de l'Arctique continental.

Les variations dans la forme et la subsidence du bassin correspondent raisonnablement bien aux événements liés à l'expansion du fond de l'océan Arctique. La subsidence thermique du style du bassin de Sverdrup a persisté jusqu'à la fin du Crétacé, le passage à une subsidence d'avant-pays coïncidant avec le début de l'expansion du fond océanique dans le nord de la mer du Labrador. Une augmentation significative des vitesses d'expansion et un changement concomitant dans la direction de l'expansion entre 59 ma et 56 ma ont pu accroître la contrainte horizontale intraplaque et accélérer la subsidence de l'avant-fosse. Cela a en outre coïncidé avec une compression initiale possible dans l'est de l'île d'Ellesmere (par ex., le bassin de Lake Hazen et le chevauchement de Lake Hazen). La sédimentation du conglomérat syntectonique du lac Buchanan

49 Ma, provided further impetus for significant crustal shortening during the Middle Eocene, and the complete fragmentation of Sverdrup Basin. Seafloor spreading had mostly ceased by the end of the Eocene (35 Ma), consistent with the geological constraints on deformation and sedimentation.

dans les sous-bassins du fjord Otto et du fjord Emma reflète une extension locale dans des zones qui étaient éloignées du front de déformation principal dans l'est de l'île d'Ellesmere. La rotation du Groenland par rapport à l'Amérique du Nord, commencée avant 49 ma, a contribué de façon importante au raccourcissement de la croûte durant l'Éocène moyen et à la fragmentation complète du bassin de Sverdrup. L'expansion du fond océanique avait en grande partie cessé avant la fin de l'Éocène (35 ma), ce qui concorde avec les données géologiques sur la déformation et la sédimentation.

INTRODUCTION

Intent

Herein is an analysis of the Upper Cretaceous-Paleogene Eureka Sound Group in the eastern Arctic Islands, primarily Ellesmere and Axel Heiberg islands (Fig. 1a, b). Strata in the Upper Cretaceous and Paleogene basins record the transition from thermally-dominated subsidence that characterized most of Sverdrup Basin history, to foredeep subsidence and eventual fragmentation into much smaller basins. The latter phase of basin development is usually regarded as the ultimate response to Eurekan deformation. Thus, the Eureka Sound Group spans the gamut of tectonosedimentary regimes, from pre-orogenic to late-orogenic phases, over a period of about 40-45 million years.

The database

Field studies of the Eureka Sound Group began in 1983 and were mostly complete by 1988. Surveys were conducted out of fly-camps and larger Geological Survey of Canada base camps, and for brief periods out of Eureka Weather Station (Atmospheric Environmental Services). Helicopter and fixed-wing support was provided by the Polar Continental Shelf Project out of Resolute Bay.

The stratigraphic database consists of 74 measured sections for a cumulative measured thickness of 22 km. Additional stratigraphic sections also were 'walked through'. Most of these sections are illustrated schematically in Figures 2a (1984), 2b (1985), 2c (1987) and 2d (1988) (in pocket); measured sections at Strand Fiord (1983) are illustrated in Ricketts (1991a). Open file maps of the Eureka Sound Group have been published for the Strand Bay (Ricketts, 1984, 1991a), and Strathcona Fiord/Bay Fiord areas (Ricketts, 1985). Map units outlined in the open files were subsequently formalized by Ricketts (1986). Updated maps for Strathcona Fiord and additional maps of the Cañon Fiord area are presented in Figure 3 (in pocket).

The paleontological database has three components. Macrofossils occur in the Upper Cretaceous Lower member of the Expedition Formation, and include a sparse Inoceramid and ammonite fauna (identified by J.A. Jeletzky, summarized in Ricketts, 1991a).

Palynology provides the critical biostratigraphic database that permits elucidation of the time stratigraphic framework. More than 400 samples were examined by D.J. McIntyre from strata ranging in age from Campanian to Middle Eocene. Published reports include Ricketts and McIntyre (1986); McIntyre and Ricketts (1989); McIntyre (1991a, b). Details of the palynoflora not treated in these papers are presented by D.J. McIntyre in Appendix 1.

Approximately 100 samples were analyzed for microfossils from the Eureka Sound Group. Arenaceous foraminifera (identified by J.H. Wall, (see Appendix 2) occur in the Upper Cretaceous part of the succession and provide an important adjunct to the palynological database. Where Upper Cretaceous Eureka Sound Group strata are preserved they gradationally overlie the Kanguk Formation and contain similar microfaunal elements (e.g. Wall, 1983). A sparse but distinctive assemblage has also been found in the Middle to Upper Paleocene Strand Bay Formation (Wall, 1991). This mid- to Upper Paleocene assemblage contains species similar to those found in the Ministicoog Formation (Mackenzie Delta; Wall et al., 1988). A distinctive calcareous foraminifera assemblage – the first of its kind in the Eureka Sound Group – has been discovered in Upper Paleocene strata in the Bay Fiord area (Cape Pillsbury Member – new, Iceberg Bay Formation – see Wall et al., 1988).

Historical perspectives

The exploits of intrepid eighteenth, nineteenth, and twentieth century Arctic Island explorers have been recounted in several essays, the most recent and comprehensive being that of Christie and Kerr (1981), and the DNAG compilation by Christie and Dawes (1991). Additional accounts dealing with more specific geographic attributes and geological problems relevant to this study include contributions to the Decade of North American Geology by Miall (1991 – Late Cretaceous and Tertiary basin development and sedimentation, Arctic Islands), Embry (1991 – Mesozoic history of the Arctic Islands), and Bustin and Miall (1991, Coal geology). The discoveries of Tertiary fossil forests have been summarized by Christie and McMillan (1991). Geological investigations of the Eureka Sound Group in the Strand Fiord/Expedition Fiord map area are detailed by (Ricketts, 1991a). Sediment composition, diagenesis, and thermal maturation of Eureka Sound Group sediments have been studied by Allen (1986).

An impressive paper by Balkwill (1978) is one of the first modern syntheses that discusses Sverdrup Basin stratigraphy, structure, and magmatism, including the Eureka Sound Group (then Formation) in the context of broad depositional regimes, sedimentation rates, and the more regional aspects of plate tectonics. Despite the fact that the timing and details of some events have changed in light of more recent studies, the present study owes a significant debt to the insights provided by Balkwill's analysis.

Lithostratigraphic schemes

The name Eureka Sound 'group' was first used by Troelsen (1950) for the predominantly nonmarine, coal-bearing Cenozoic strata underlying Fosheim Peninsula. This succession was originally interpreted as postdating the last period of orogeny (Troelsen, 1952) but was later discovered by Thorsteinsson and Tozer (1957) to have been included in the deformation. Following the more extensive investigations of Operation Franklin, Tozer (1963) redefined the unit as a formation, and suggested that exposures on Fosheim Peninsula were typical of the formation. However, a type section was also erected by Souther (1963) at Strand Fiord and it is this designation that has continued in the literature.

The Eureka Sound Group and its constituents were defined simultaneously by Miall (1986) and Ricketts (1986), superseding Troelsen's (1950) informal designation of group, and Tozer's (1963) designation of formation. The relative merits of each nomenclatural scheme have been discussed by Miall (1988) and Ricketts (1988), from which the following discussion is extracted.

Miall's scheme is based on (interpreted) lithofacies assemblages, whereas my scheme relies on the more traditional approach of mapping relatively homogeneous rock units. My scheme (Fig. 4) does not deny the existence of the facies assemblages but considers them unimportant in the definition of stratigraphic units having formation status.

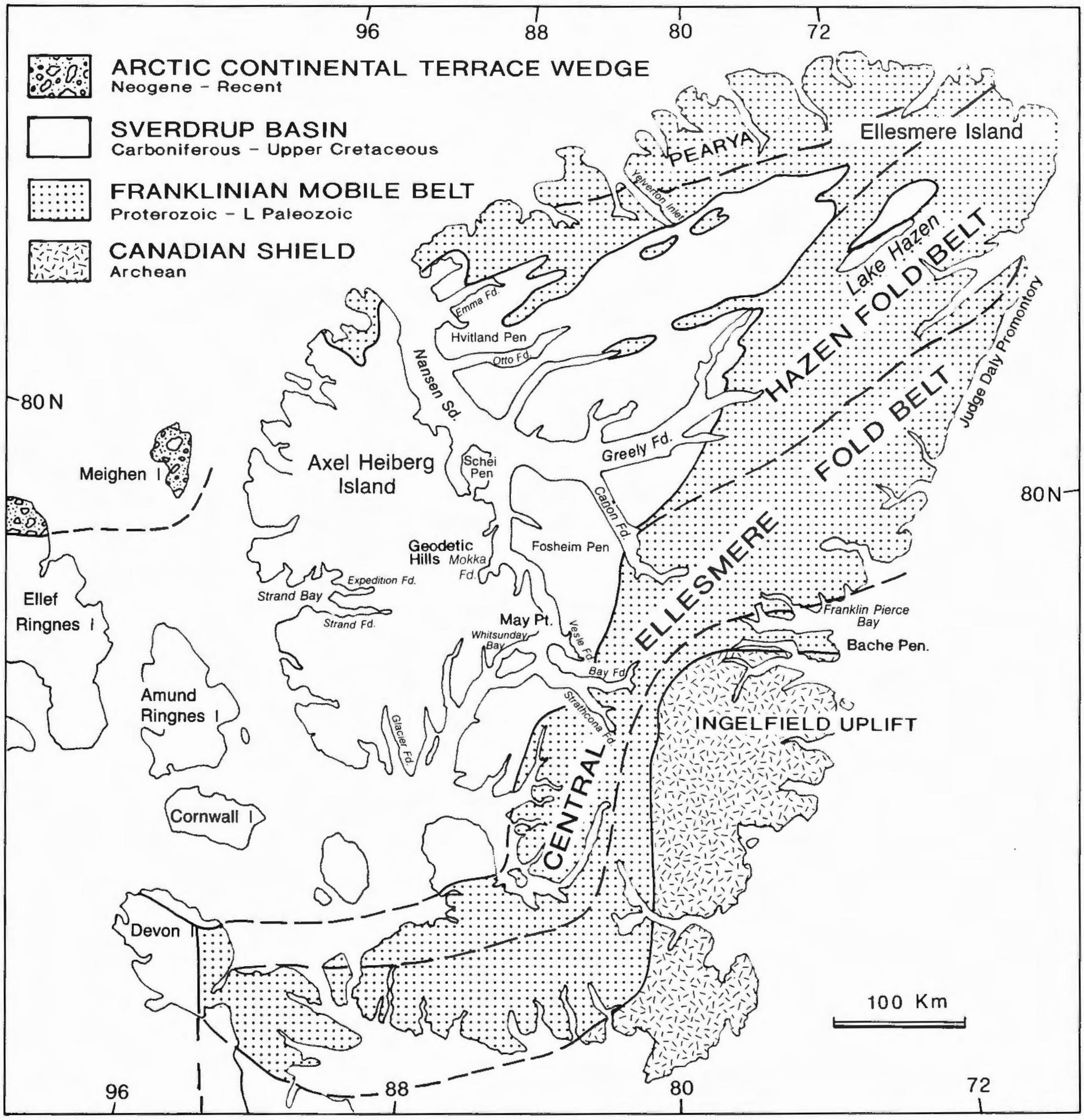

Figure 1a. Location map for the Queen Elizabeth Islands, depicting principal tectonostratigraphic components (from Trettin, 1989).

The definition of formal lithostratigraphic units and that of facies assemblages serves two different functions. Formations (and members) do little more than 'label' rock units, providing a mechanism for communicating and identifying the physical and geographical limits of the rock body; formations as such provide few clues to the evolution of sedimentary basins. Sedimentary facies and facies assemblages on the other hand, while defined in part on supposedly objective criteria, also possess a large dose of interpretive input – these rock bodies, the boundaries of which may or may not coincide with formation boundaries, provide a fundamental ingredient for creative interpretation of basin evolution. Another important ingredient is the identification of lacunas and the regional effects of base-level change, and the recognition of depositional sequences.

Why define formations at all, especially considering their relatively low level of importance in the overall interpretation of basin analysis? The most compelling reason, and one that provides a good argument against Miall's formal scheme, is one of terminological stability: the interpretations of specific sedimentary facies and facies assemblages, as well as the significance of unconformities, have the potential to change depending on who is studying the rock units. Thus, reinterpretation of a lithofacies may necessitate redefinition of its conjoined formation, particularly if the boundaries of the facies assemblage also change. Specific examples of this problem have been cited by Ricketts (1988).

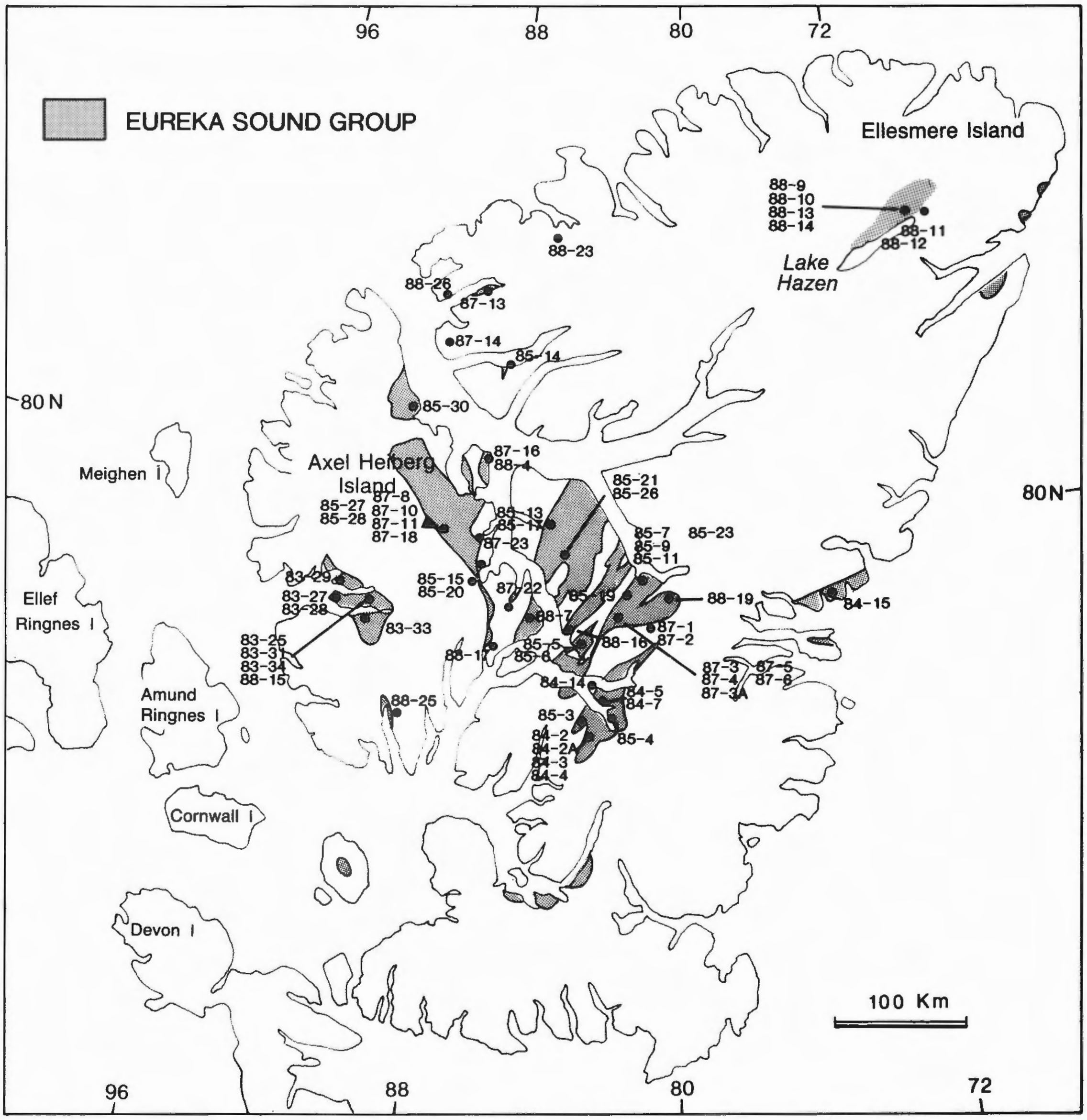

Figure 1b. General distribution of Eureka Sound Group outcrop and location of key stratigraphic sections referred to in the text, Figures 2a-d, and Appendices 1 and 2. Sections are denoted by year and number (the prefix RAK has been omitted).

Competing hypotheses of basin evolution

Interpretations of Late Cretaceous-Paleogene basin evolution by Balkwill and Bustin (1980), and Miall (1984a, 1985, 1986), have posited hypotheses of Eureka Sound strata filling multiple basins. According to these hypotheses the basins were separated by arches and uplifts generated during the early stages of Eurekan tectonism at the end of the Cretaceous or Early Tertiary (Balkwill, 1978). Examples include the ancestral Princess Margaret Arch and Cornwall Arch separating Remus, Strand Fiord (formerly Meighan Basin), and West Sverdrup basins (Balkwill et al., 1975; Bustin, 1977; Miall, 1985, 1986).

The opposing hypothesis (espoused here) is that basin development took place in three stages: an early stage that essentially represents the continuation of typical Sverdrup Basin; a stage of rapid foredeep-like subsidence; and a stage of syntectonic, intermontane basins that fragmented Sverdrup Basin.

Clearly, the critical factor in resolving this dispute over basin configuration is the timing of arch formation, which is established on the basis of stratigraphy and facies relationships. Criteria that support the single basin hypothesis have been discussed by Ricketts and McIntyre (1986), and Ricketts (1987a, 1991a), the main arguments being:

1. There is no demonstrable stratigraphic thinning of Eureka Sound Group strata against the flanks of Princess Margaret Arch;
2. Sedimentary facies and sediment composition do not exhibit proximal/distal relationships with respect to the arch;
3. The age of sediment undoubtedly derived from the Princess Margaret Arch (Buchanan Lake Formation) is no older than Middle Eocene;
4. Paleocene and early Eocene strata of the Iceberg Bay Formation have been found in the Geodetic Hills and Whitsunday Bay, both areas being very close to the axis of Princess Margaret Arch.
5. Distinctive pebbles and cobbles of acid volcanic composition derived from the northeast margin of the basin, occur immediately above a regional sub-Paleocene unconformity on both sides of Princess Margaret Arch.

Tectonostratigraphic framework

Deposition of the Eureka Sound Group, beginning about middle Campanian time, represents the first major influx of sand into the Arctic-wide Kanguk sea. From the middle Campanian until some time in the late Maastrichtian, sedimentation was essentially a continuation of that occurring in Sverdrup Basin with a depocentre located on west-central Axel Heiberg Island (Fig. 1). From a regional perspective, the renewal of clastic sedimentation coincides with initial seafloor spreading in Labrador Sea (Srivastava, 1985; Srivastava and Tapscott, 1986).

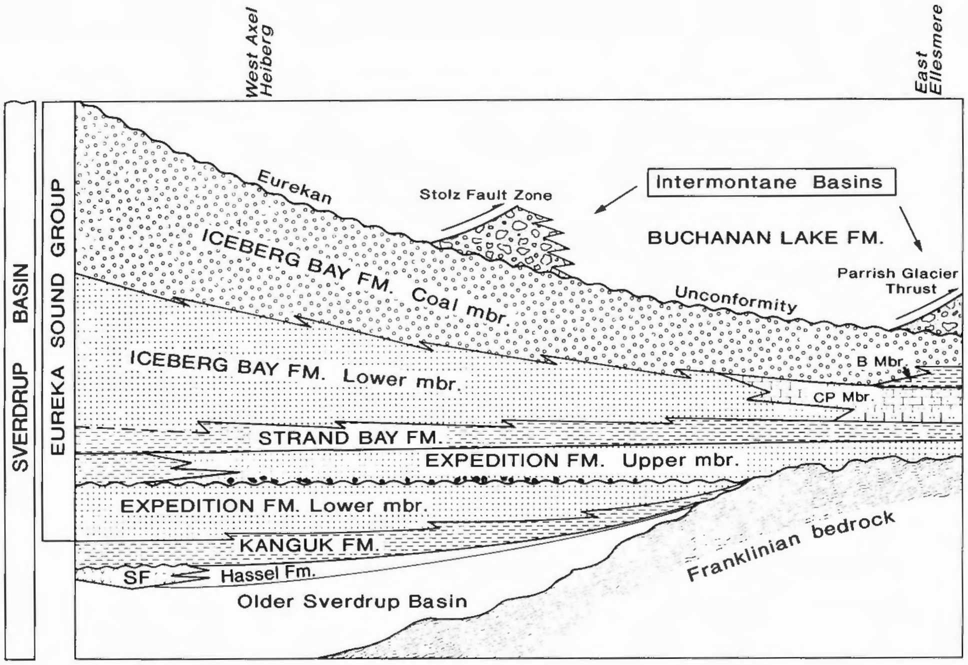

Figure 4. A schematic summary of the Eureka Sound Group lithostratigraphy along an approximate west-east transect from western Axel Heiberg Island to eastern Ellesmere Island. B = Braskeruds Member; CP = Cape Pillsbury Member; SF = Strand Fiord volcanics.

A significant lowering of base level towards the end of the Cretaceous, attributable to regional uplift, resulted in a basin-wide sub-Paleocene unconformity (see Time stratigraphy section); an event that has correlatives as far afield as West Greenland and Spitzbergen (western Barents shelf) (see Synthesis section). Increased subsidence, a changing thermal regime, and the increasing influence of Eurekan compression in the eastern Arctic Islands were the hallmarks of the Paleogene foredeep. Depositional limits were greatly expanded from the Late Cretaceous Sverdrup Basin stage, such that Paleocene and younger strata encroached upon the Lower Paleozoic Franklinian Fold Belt. Further encroachment of the Eureka Sound Group east and south is evidenced by fault-bounded outliers in the Lower Paleozoic Arctic Platform, for example at Bache Peninsula (Christie, 1967), Viks Fiord (Devon Island, Thorsteinsson and Mayr, 1987) and northern Somerset Island (McIntyre, 1989). However, the former continuity of these scattered outliers is difficult to assess.

The Eocene was witness to a tectonic revolution wherein small, syntectonic alluvial basins were superposed upon a fragmented Sverdrup Basin, possibly beginning in the Early Eocene and reaching a climax in Middle Eocene time. Some of the basins are associated with major arches (crustal scale folds of Stephenson and Ricketts, 1990), large frontal thrusts, or reverse faults. Many of the faults may be reactivated Ellesmerian or Sverdrup Basin rift structures (Balkwill, 1983a, b; Miall, 1986; Trettin, 1987). At least six basins and subbasins that are preserved onshore, contain variable proportions of sediment derived from the unroofing of Mesozoic and Upper Paleozoic Sverdrup Basin rocks, and Lower Paleozoic Franklinian rocks.

The timing of events identified within the Eureka Sound Group is now reasonably well established and can be compared to known plate kinematic data. Igneous activity continued sporadically in Sverdrup Basin to the end of the Paleocene (Trettin and Parrish, 1987; Embry and Osadetz, 1988; Muecke et al., 1990). Volcanism and intrusion can be related to certain specific stratigraphic units as well as to the broader characteristics of basin thermal regimes.

Goals

This memoir has three goals:

i. To provide details of the primary data base for the Eureka Sound Group. Where the details of sedimentology, structure, and paleontology have been published elsewhere, summaries only will be provided here. Readers are directed towards the relevant papers. The sedimentology will be discussed in terms of facies and facies assemblages within the lithostratigraphic units.

ii. To incorporate the stratigraphic architecture of the Eureka Sound Group into a sequence stratigraphic framework.

iii. To present a basin synthesis in terms of the sequence stratigraphic, structural, magmatic, and plate kinematic constraints.

TIME STRATIGRAPHY

The time-stratigraphic organization of a sedimentary basin, furnished primarily by biostratigraphy and radiometric dating, is a key element of any basin synthesis. In the previous section it was noted that the biostratigraphic database consists of palynology, microfossil, and sparse molluscan fossil assemblages. Radiometric dating of some felsic volcanics on northwestern Ellesmere Island (Trettin and Parrish, 1987; Mueck et al., 1990) also constrains the timing and position of the northern margin of Sverdrup basin as well as key depositional events early in the Paleocene.

Four time-stratigraphic profiles have been constructed, centred about the Remus Creek/north Fosheim Peninsula area (Fig. 5), and radiate to fundamentally different segments of the Eureka Sound Group basins (Figs. 6 to 9).

Regional unconformities

Sedimentation early in the history of the Eureka Sound Group (Upper Cretaceous, Expedition Formation) was essentially a continuation of that typical of Sverdrup Basin, namely the development of large deltas. The Paleogene Sverdrup Basin, on the other hand, expanded beyond its Upper Cretaceous limits and also overlies a regional sub-Paleocene unconformity. Over much of the area Maastrichtian rocks either are partly represented or completely missing (Ricketts, 1991a, c). Where Lower Maastrichtian strata are missing it is clear that the unconformity is mostly erosional. Whether the Late Maastrichtian was characterized by erosion, nondeposition, or a combination of these two effects is unknown. The sub-Paleocene unconformity is most profound where it occurs on Lower Paleozoic rocks of the Franklinian Fold Belt (Figs. 7, 8).

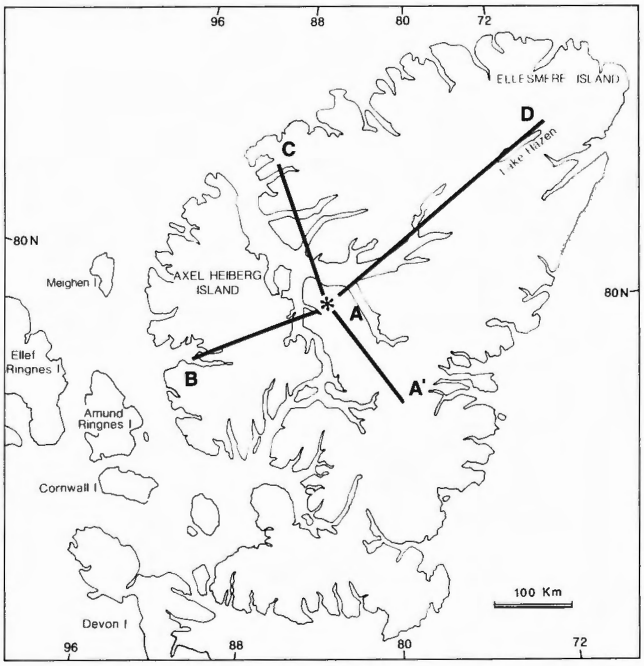

Figure 5. Location of time-stratigraphic profiles illustrated in Figures 6-9.

Major unconformities are also present below the Buchanan Lake Formation. Considering the highly variable degree of stratigraphic discordance, for example along eastern Axel Heiberg Island where Buchanan Lake conglomerate overlies strata ranging in age from Triassic to Middle Eocene (Ricketts, 1991b), it is reasonable to infer that erosion accounts for most of the missing stratigraphy. At Otto Fiord and Lake Hazen the unconformities are overlain by conglomerate as old as Early Eocene (see Buchanan Lake Formation section), whereas on east Axel Heiberg Island it is confined to the Middle Eocene, because the oldest Buchanan Lake beds and the underlying Iceberg Bay Formation are both Middle Eocene (Ricketts, 1991b; Ricketts and McIntyre, 1986). This does not preclude the possibility that Eurekan deformation continued later in the Eocene, but to date no onshore stratigraphic evidence has been found that would verify this assertion. In fact the only evidence that deformation might have continued into the Late Eocene or earliest Oligocene is derived from completely independent plate kinematic data (see Synthesis). Regardless of the ultimate duration of Eurekan tectonism, most of the 4-6 km of pre-Buchanan Lake erosion on eastern Axel Heiberg Island must have taken place during a brief period within the Middle Eocene.

THE EXPEDITION FORMATION

The Expedition Formation is the oldest unit in the Eureka Sound Group, ranging in age from about middle Campanian to Early Paleocene. Its formal status was assigned by Ricketts (1986) with a type section at Kanguk River, north of Strand Fiord. Two members, the Upper and Lower members, were defined in the Strand Fiord area by Ricketts (1991a), and subsequently these have been extended to the remainder of the basin (Ricketts, 1989a, 1991c). In the early phase of Eureka Sound Group mapping, the separation of the Upper and Lower members by the sub-Paleocene unconformity was not recognized. Admittedly, the presence of a major unconformity within a formation is far from ideal, but there is no point in creating the potential for nomenclatural confusion by changing the formal status of these units.

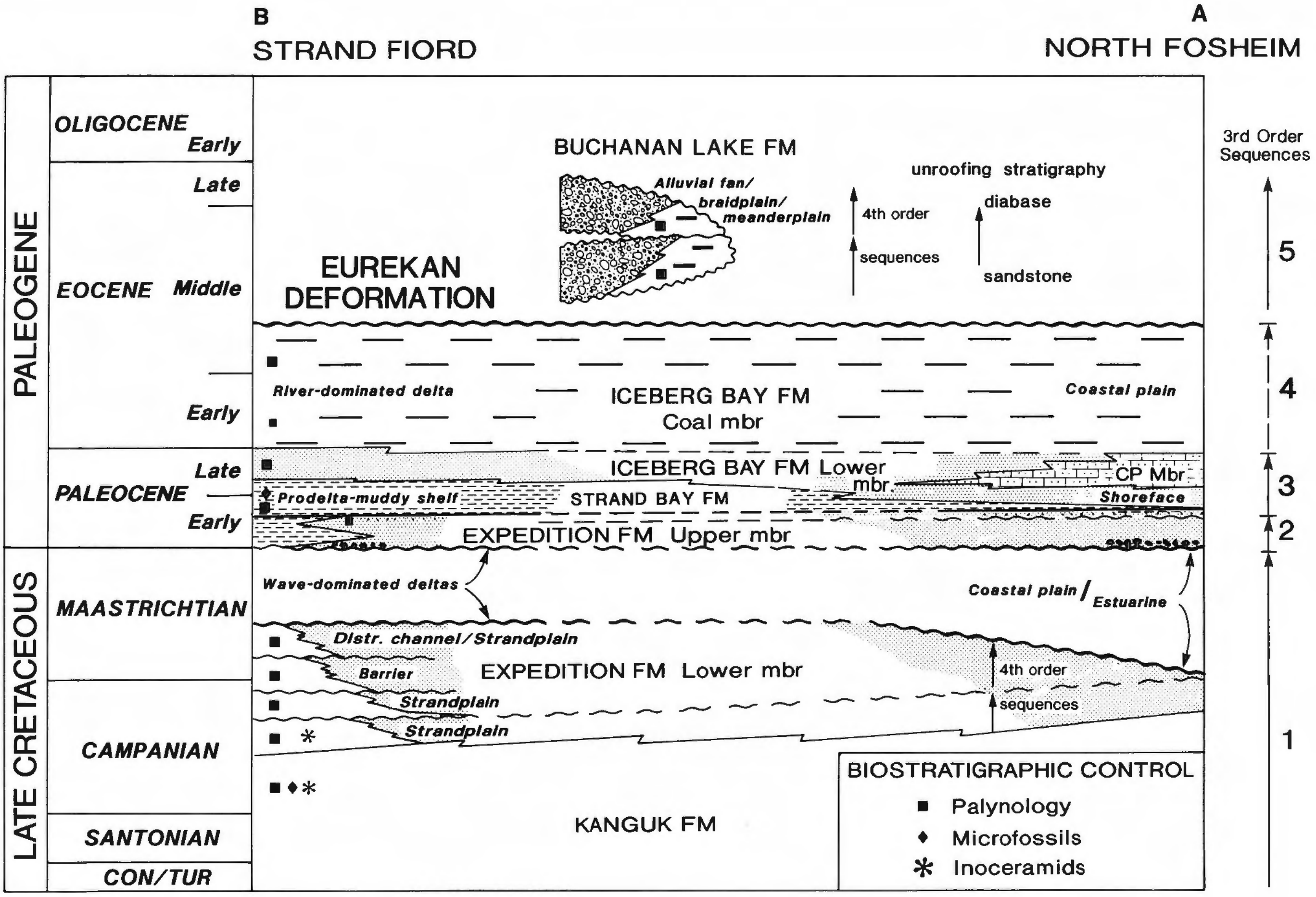

Figure 6. Time stratigraphic relationships, third- and fourth-order stratigraphic sequences, lithostratigraphy and principal facies, and generalized biostratigraphic control in the Eureka Sound Group along a west Axel Heiberg Island-north Fosheim Peninsula profile (see Fig. 5). CP Mbr = Cape Pillsbury Member (Iceberg Bay Formation). The unroofing stratigraphy refers to Axel Heiberg Basin (Buchanan Lake Formation).

Lower member rocks are well exposed in the Strand Fiord-Expedition Fiord area (Ricketts, 1991a), but their preservation elsewhere is limited to northern Fosheim Peninsula, May Point, and scattered localities around Emma Fiord. Upper member strata, on the other hand, are exposed over a much greater part of the basin, especially in the eastern and southeastern regions where they onlap Lower Paleozoic bedrock. Measured sections and their locations are portrayed in Figures 2a-d.

Age

The best biostratigraphic control is at Strand Fiord, and includes molluscan (Lower member only, Ricketts, 1991a), micropaleontological, and palynological assemblages (McIntyre, 1991a; McIntyre and Ricketts, 1989; Wall, 1991). Relevant sample locations are indicated in Figures 2a-d and Appendices 1 and 2. The Upper member biostratigraphy is based primarily on palynology.

Lower member strata of the Expedition Formation range in age from about middle Campanian to Maastrichtian. The age of the Upper member was quoted in Ricketts (1991a) as ranging from Maastrichtian to Early Paleocene. However, identification of the sub-Paleocene unconformity, together with an expanded palynology and microfossil database, necessitates revision of the age. Upper member rocks are now known to be restricted to the Lower Paleocene.

Lower member, Expedition Formation

General characteristics

With the exception of the northern Sverdrup Basin margin, basal strata of the Lower member are in gradational contact with the Kanguk Formation. The upper contact of the Lower member is a regional erosion surface, to the extent that in some places the Upper member directly overlies Kanguk shale (e.g. south of Slidre Fiord, and the western parts of Bay and Strathcona fiords).

A minor modification is made to the Lower member/Upper member contact at the Kanguk River section (RAK 31-83); it is located at the base of a thin but prominent conglomerate at 492 m (Fig. 10a). This increases the maximum thickness of the Lower member to 347 m. Elsewhere the contact is unchanged. The conglomerate bed itself forms the base of

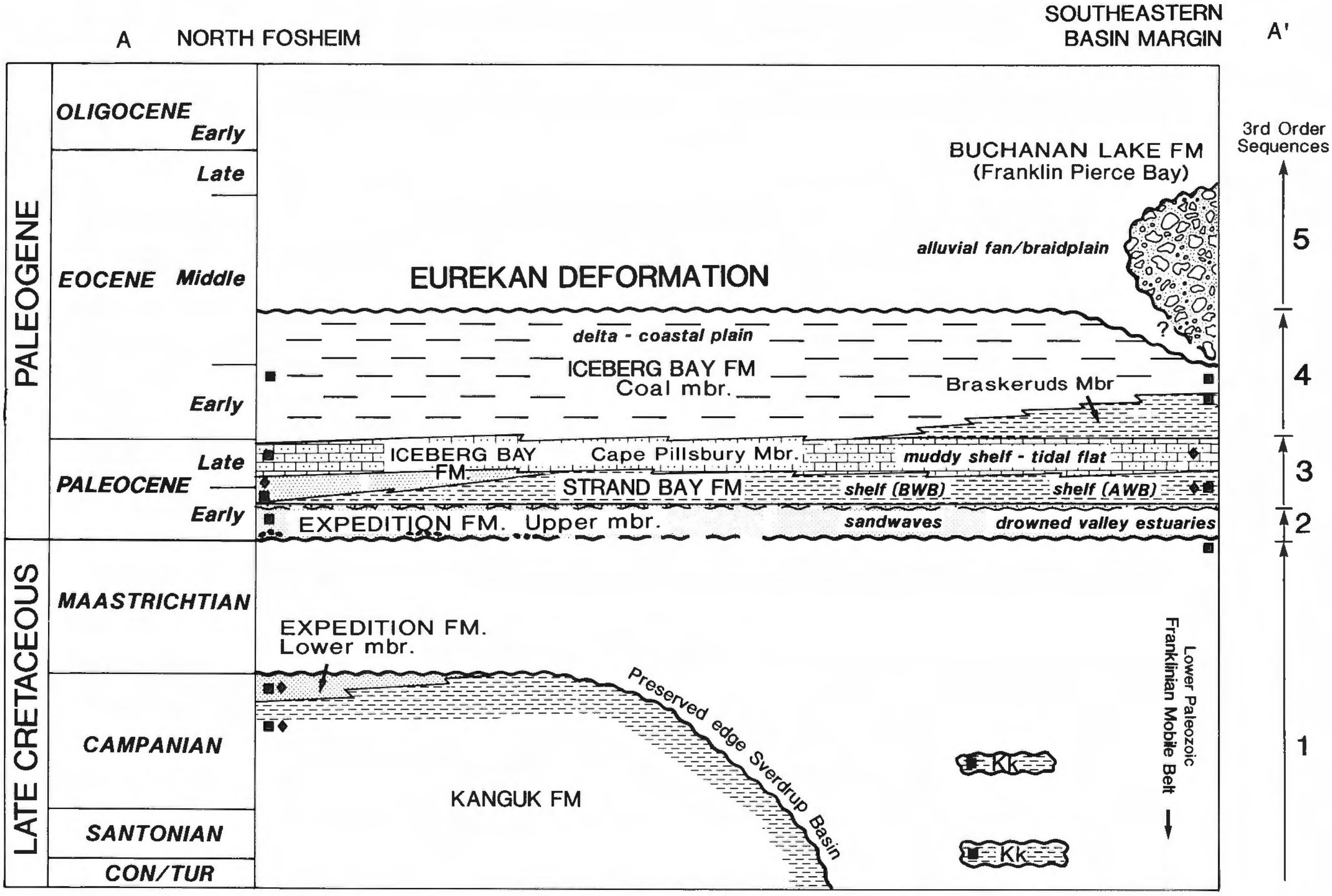

Figure 7. Time stratigraphic relationships, stratigraphic sequences, lithostratigraphy and principal facies, and general biostratigraphic control in the Eureka Sound Group along the Fosheim Peninsula-Cañon Fiord profile. Fossil group symbols are indicated in Figure 6. BWB and AWB refer to below wave base and above wave base respectively.

the Upper member, as it does in other areas of Fosheim Peninsula and May Point. Distinctive pebbles of dacite in the conglomerate are not found at any other stratigraphic level in the Eureka Sound Group.

At Strand Fiord and environs the Lower member ranges from 170 to 347 m thick, whereas 50 km due south at Glacier Fiord only 5 to 10 m of interbedded sandstone and mudstone occur, underlying a distinctive white quartz sandstone assigned to the Upper member (Fig. 10b). Thicknesses up to 100 m are recorded on north Fosheim Peninsula (Remus Creek – sections RAK 13-85 and 17-85, and Fosheim anticline – RAK 21-85, Fig. 2b), and May Point (RAK 7-88, Fig. 2d).

At Strand Fiord the Lower member ranges in age from early or late Campanian to Early Maastrichtian. One of the best sections of the Lower member on Fosheim Peninsula is found at Remus Creek (RAK 13-85). Here, both palynomorphs and microfossils indicate a middle to late Campanian age (Appendices 1 and 2 respectively). Here too the Maastrichtian is completely missing.

Lower member representation along the northern margin of the basin is restricted to a few tens of metres of interbedded sandstone, shale, and coal of early to middle Maastrichtian age along the north shore of Emma Fiord (Fig. 9; also McIntyre, Appendix 1, GSC sample numbers C-153180 to C-153182). Strata exposed along the north shore of Emma Fiord overlie, and in part interfinger with, bimodal alkalic volcanics which have whole rock $^{40}Ar/^{39}Ar$ ages of 80 ± 2 Ma (i.e., middle to upper Campanian; MacRae et al., 1990; Muecke et al., 1990). These authors have postulated that the discrepancy between the radiometric and palynological ages (some 5 million years and more) is a product of biostratigraphic diachroneity.

Delta facies assemblage (wave dominated)

Lower member strata in the Strand Fiord area can be subdivided into three principal facies (for details of stratigraphic organization and sedimentology see Ricketts, 1991a). The lowest coarsening-upward shale-sandstone facies occurs in units up to 100 m thick (units A and B in Fig. 10a). Resistant and locally fossiliferous (e.g., *Inoceramus*) ironstone beds

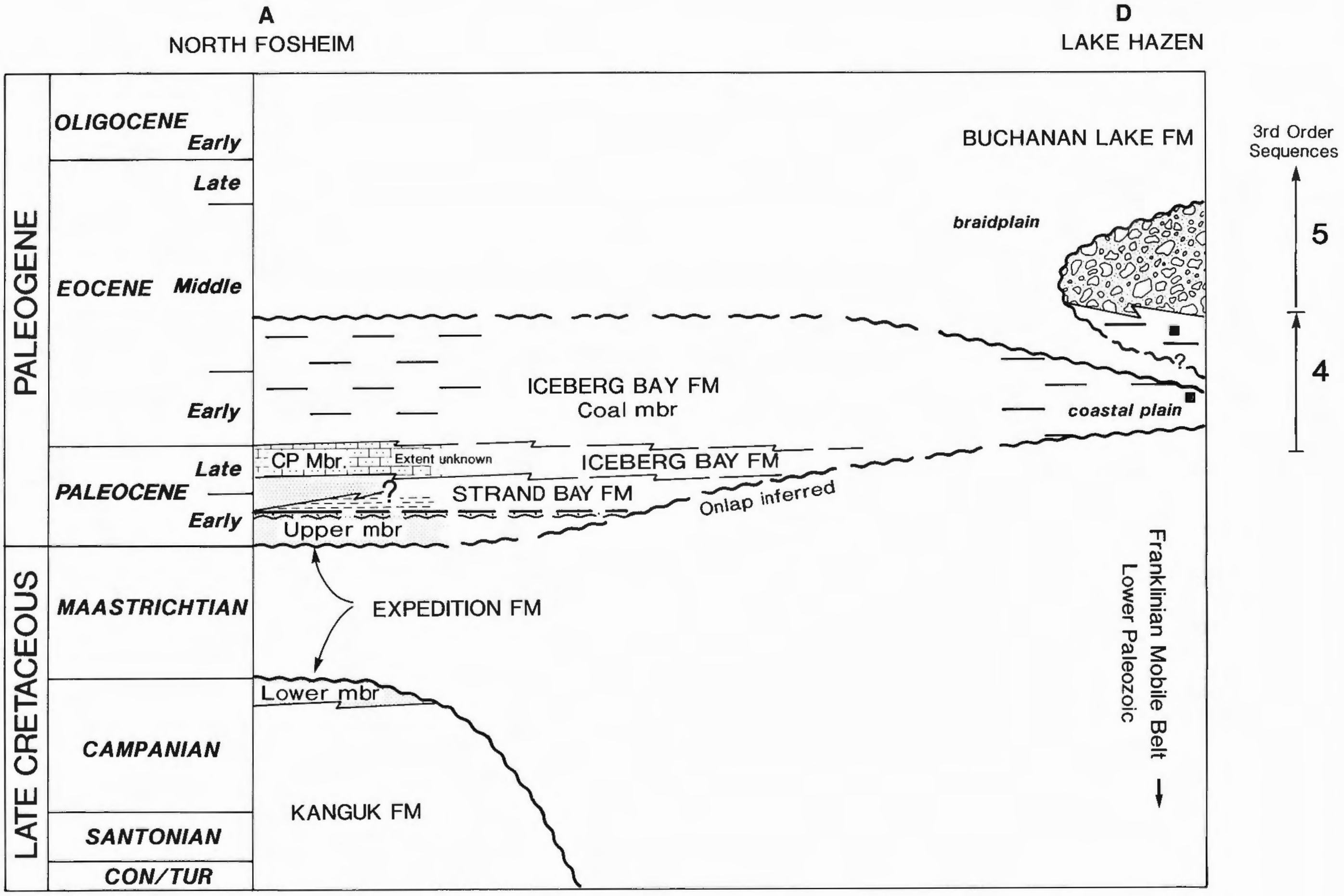

Figure 8. Time stratigraphic relationships, stratigraphic sequences, lithostratigraphy and principal facies, and general biostratigraphic control in the Eureka Sound Group along the north Fosheim Peninsula-Lake Hazen profile. Only sequences 4 and 5 are present in the Lake Hazen area. Palynology symbols are indicated in Figure 6. The onlapping configuration is inferred.

occur near the base of each shale component. Sandstone at the top of each unit is capped by a bed of greenish hue and contact with the overlying shale is abrupt. Each coarsening-upward unit consists of many small-scale coarsening-upward sandstone cycles 5 to 15 m thick that become progressively sandier towards the top of each unit. Typically, shale at the base of each cycle passes into thin, interbedded sandstone and shale that is thoroughly bioturbated (e.g., *Chondrites, Teichichnus, Terebellina*). Mature quartz sandstone at the top of each cycle contains low-angle planar crossbedding, parallel lamination, ripples, planar-tabular crossbedding, coarse sandstone/fine sandstone couplets, and reactivation surfaces, all burrowed by *Skolithos*.

A substantial change in facies organization occurs above coarsening-upward units A and B with the appearance of a more symmetrical, coarsening- then fining-upward facies (unit C, Fig. 10a). The coarsening-upward component (about 80% of unit thickness) is similar to that seen in units A and B, with an additional component of channel-filling trough crossbeds. *Skolithos* and associated escape burrows abound. Fining- and thinning-upward trends in interbedded sandstone and mudstone are accompanied by decreasing bedform size and increasing carbonaceous content. Capping the facies is carbonaceous and coaly shale with thin rippled sandy layers and root structures.

Massive cliff-forming sandstone (unit D in Fig. 10a), up to 30 m thick, constitutes the thick bedded sandstone facies. The texturally and compositionally mature sandstone consists of stacked, large planar-tabular crossbed sets up to 3 m thick separated by pebble veneers. Only at the top of the facies is there any indication of a fining-upward trend. *Skolithos* burrows are common.

The top of the Lower member at Strand Fiord consists of an additional 80 m coarsening-upward shale-sandstone unit (unit E, Fig. 10a) similar to units A and B.

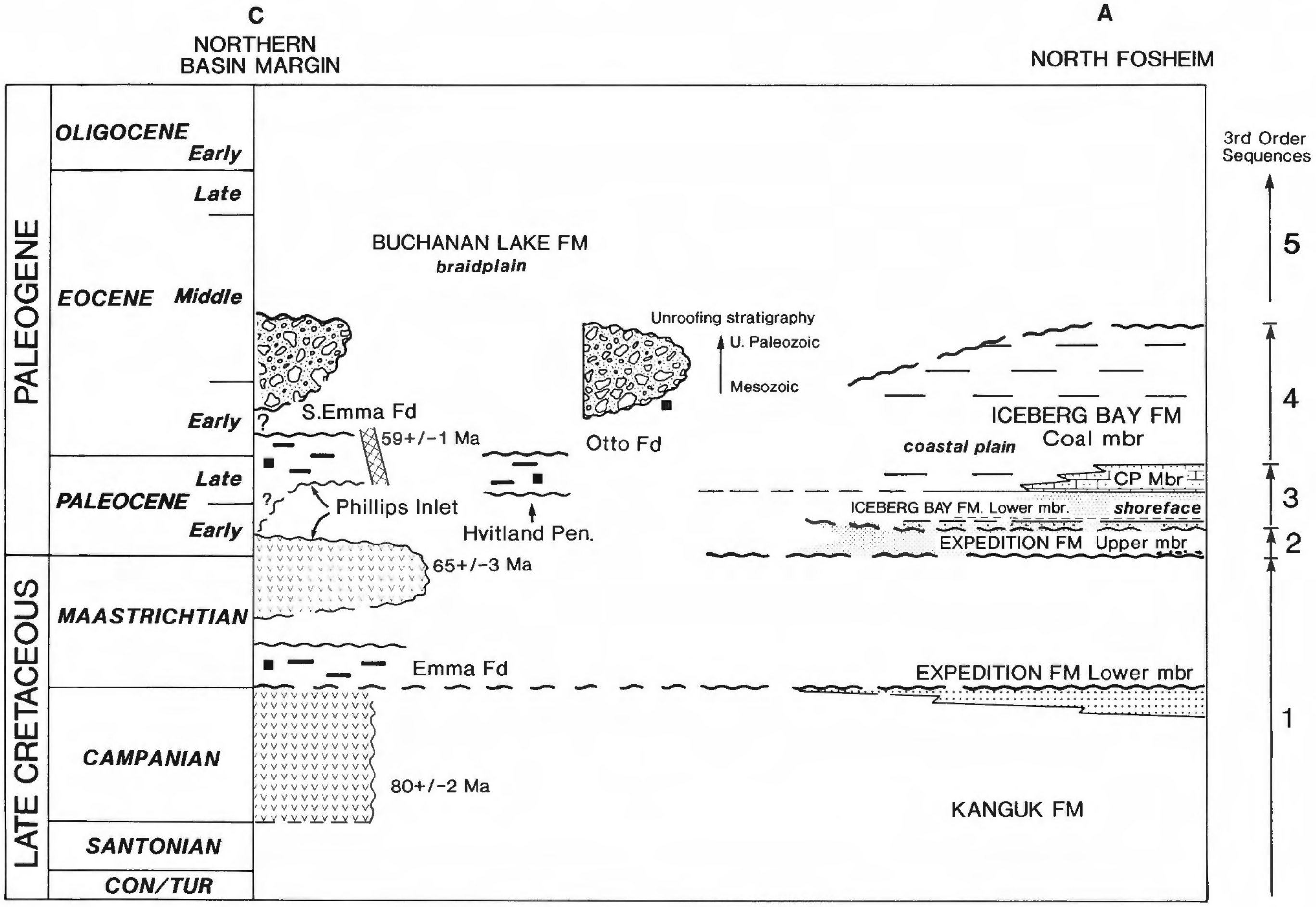

Figure 9. Time stratigraphic relationships, stratigraphic sequences, lithostratigraphy and principal facies, and general biostratigraphic control in the Eureka Sound Group along an approximate north Fosheim Peninsula-Emma Fiord/Phillips Inlet profile. Palynology symbols are indicated in Figure 6. Radiometric dates from Muecke et al. (1990). The unroofing stratigraphy refers to Otto Fiord Basin. Note that the Strand Bay Formation is replaced by a shallow water facies included in the Iceberg Bay Formation Lower member.

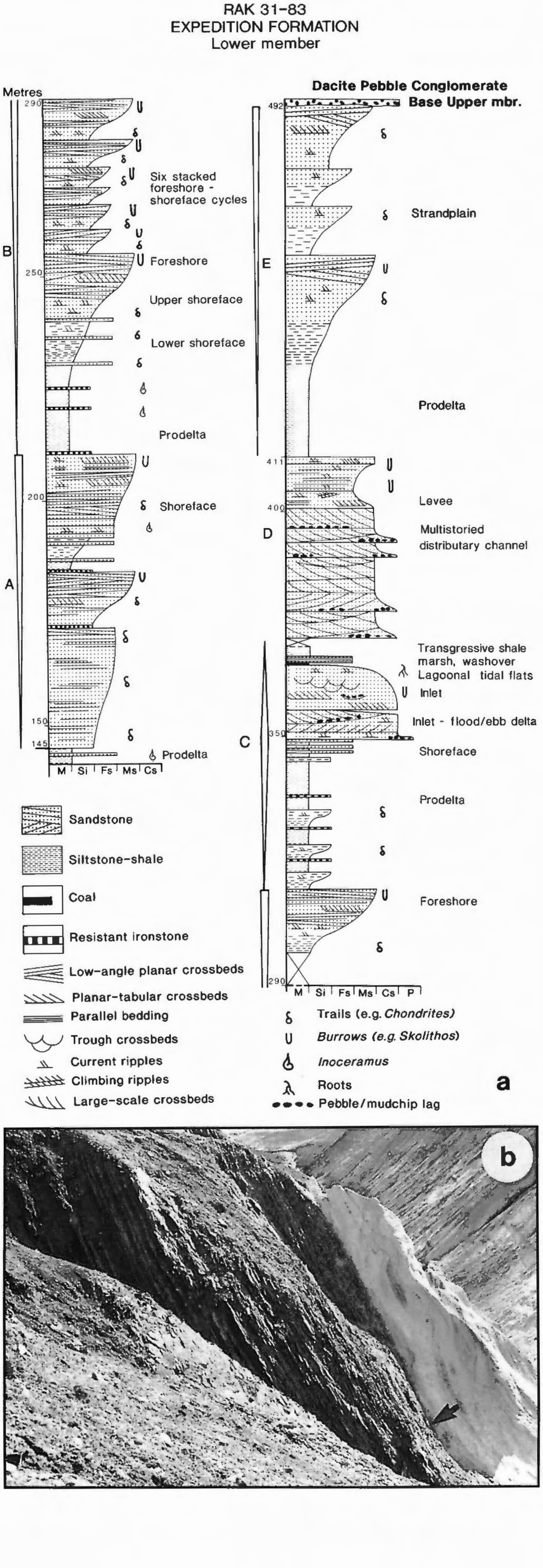

Figure 10. ***a****) Schematic stratigraphic column of the Lower member, Expedition Formation, along Kanguk River (RAK 31-83; 79°16′N; 90°40′W) showing principal sedimentation units A to E. Detailed sedimentology is illustrated in Ricketts (1991a, modified from Fig. 6). M = mudstone; Si = siltstone; Fs = fine grained sandstone; Ms = medium grained sandstone; Cs = coarse grained sandstone; P = pebble conglomerate (Wentworth scale);* ***b****) Abrupt contact (arrow) between the Expedition Formation Upper member (white sandstone) and a 5 m thick sliver of the Lower member, near the head of Glacier Fiord (78°36.5′N; 89°51′W). The Lower member conformably overlies Kanguk Formation shale. GSC 1993-039C*

When traced westwards along Kanguk Peninsula, the three facies described above become finer grained, texturally less mature, and more bioturbated. At the western limit of exposure on Kanguk Peninsula the three facies merge into a single coarsening-upwards unit where the most prominent features are small metre-thick coarsening-upward packages. Mudrock decreases from 60% of the succession at the base, to 30-40% at the top. Limy, fossiliferous mudstone beds cap several of the packages. Bioturbation by *Chondrites* is intense.

Lower member strata on western Axel Heiberg Island represent a series of stacked, wave-dominated deltas and associated facies (Fig. 10a), that prograded approximately west to southwest over a period of about eight million years (details in Ricketts, 1991a). Coarsening-upwards cycles (units A and B) in the proximal part of the delta each exhibit the transition from prodelta to strandplain, within which there are many higher order shoreface-strandplain cycles. Green sandstone that caps each cycle represents the transgressive component of the succeeding cycle, whereas contact with the overlying shale approximates the surface of maximum flooding during this transgression. Resistant sideritic ironstone beds near the base of each coarsening-upward cycle reflect a marked decrease in clastic input during transgression and constitute condensed stratigraphic sections.

The assemblage of sedimentary structures in the upper part of each strandplain cycle generally indicates high energy, upper shoreface settings, occasionally traversed by small two-dimensional ripples; channelling or confined flow was not common. Infaunal activity in the upper shoreface deposits was restricted to opportunistic genera such as *Skolithos*, reflecting continual shifting of the shoreface sands.

Delta construction was interrupted in the Strand Fiord area at the time of Unit C deposition because of a major lateral shift in the distributary channel. Overall, unit C is analogous to Recent barrier-inlet, ramp, back barrier, and washover deposits in many microtidal, wave-dominated barrier island settings (Hubbard et al., 1979; Kraft and John, 1979; Hayes, 1975, 1976; Hennessey and Zarillo, 1987). The delta strandplain of unit B was abandoned, resulting in an overall rise in base level and landward migration of the strandline. Sand on the strandplain platform was subsequently reworked into a series of barrier islands, tidal inlets, and lagoons, now recorded in the coarsening- then fining-upward facies (unit C). The shoreface was cut by a tidal inlet or tidal delta ramp consisting of stacked planar-tabular and trough crossbedded sandstone that contains evidence of tidal current reversal (reactivation surfaces, sandstone couplets). Furthermore, the fining-upward component indicates progressively shallower conditions and eventual exposure in a restricted lagoonal setting. Thin, laminated sandstone beds that interfinger with the coaly lagoonal facies are interpreted as storm washover deposits.

Thick, cliff-forming sandstone composing unit D, with its large two- and three-dimensional subaqueous dune bedforms, is interpreted as a major distributary channel and mouth bar complex. Levee deposits cap the channel and mouth bar sandstone indicating some degree of lateral migration within the delta. That this is the only distributary channel recognized in the area is consistent with observations of modern wave-dominated deltas. Unlike river-dominated systems, the number of active channels is generally much less. Thus the sudden appearance of this facies suggests an important event in the delta, where, in concert with continued subsidence, the delta distributary switched to the area formerly occupied by the barrier island facies.

Delta plain-coastal plain facies assemblage

Upper Campanian strata correlative with the wave-dominated delta complex at Strand Fiord are best exposed on Fosheim Peninsula, where they constitute the 'thin end' of the Upper Cretaceous sedimentary wedge (most of the Maastrichtian component has been eroded). Typical facies associations are shown in Figure 11. The succession is characterized by coarsening, or sandier-upward units 7 to 12 m thick, beginning with silty shale. Concomitantly, ripples and parallel laminae are succeeded by lenticular and flaser bedding, small planar-tabular crossbeds, and less commonly hummocky cross-stratification. Some of the better sorted sandstone beds contain both low-angle planar crossbedding and abundant *Skolithos*.

Strikingly regular interbedding of very thin, brown carbonaceous, and clean white sandstone composes the uppermost unit in Figure 11, having an overall thinning-upward trend, and also containing many *Skolithos* burrows (Fig. 12). White sandstone layers 3-5 cm thick contain parallel laminae and a few ripples; thinner carbonaceous layers that complete each sandstone-mudstone couplet, become more prominent towards the top of the unit. Other coarsening-upward units also are succeeded by shale and coaly beds (containing root structures), and starved or lenticular sandstone ripples.

Correlation of the above facies with specific components of the western Axel Heiberg wave-dominated delta complex is not possible. However, the general association of facies here characterizes a delta plain-coastal plain setting ranging from subtidal and relatively high energy shoreface, to perhaps more localized muddy tidal flats and marshes cut by meandering tidal channels (e.g. Fletcher et al., 1990). As such they logically represent the landward extension of, or perhaps the

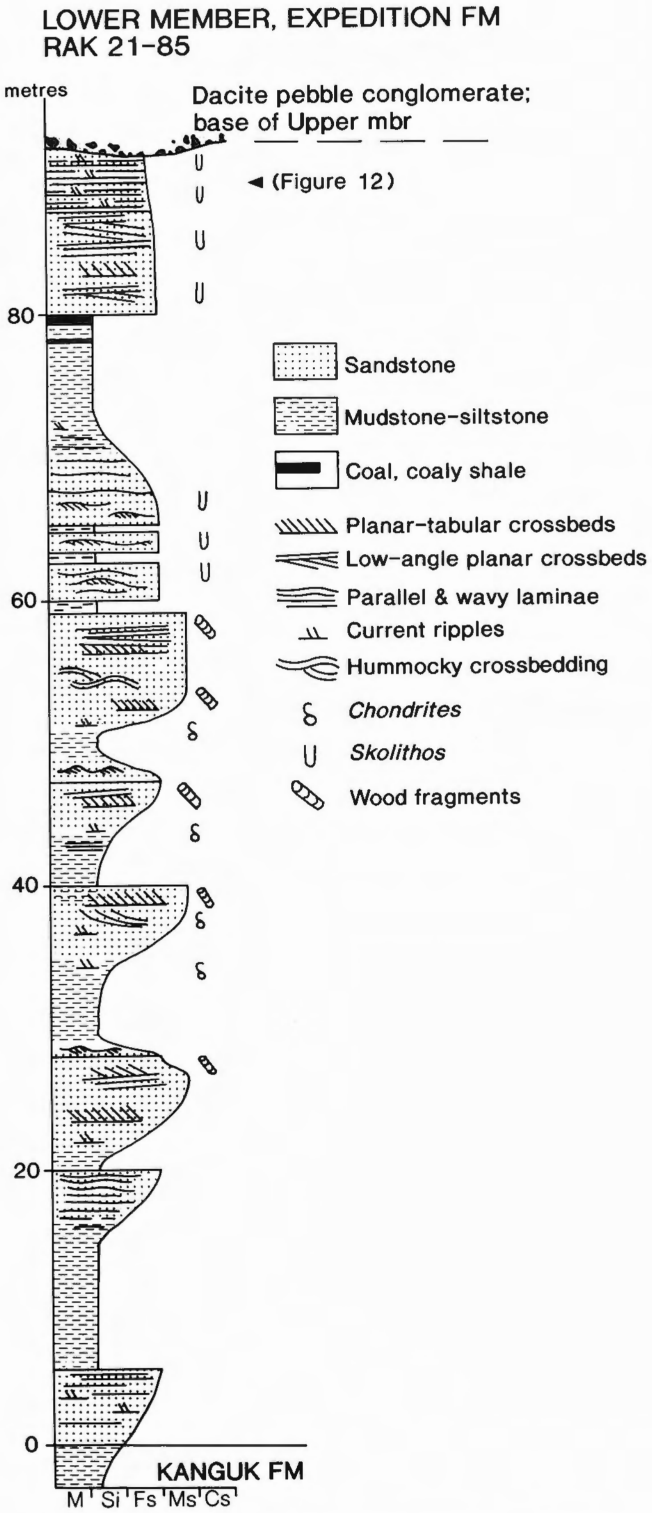

Figure 11. Schematic representation of tidal flat, shoreface, and shallow shelf facies in the Lower member, Expedition Formation, west flank of Fosheim Anticline. Contact with the Kanguk Formation is conformable; that with the Upper member is an erosional (sub-Paleocene) disconformity. Total thickness of the Lower member is 88.5 m.

lateral (adjacent) equivalent to the main delta complex. Regularly bedded successions of sandstone-mudstone couplets are the predicted result of laterally accreting point bars in tidal channels (e.g., Smith, 1988; Nio and Yang, 1991), either in tidal flats or estuaries. The couplets here attest to significant asymmetry with respect to flood or ebb tides and periods of slack water.

Fluvial facies assemblage

Fining-upward units of sandstone, mudstone, and coal on the north shore of Emma Fiord range in age from middle or late Campanian to Early Maastrichtian (MacRae et al., 1990; D.J. McIntyre, pers. comm., 1990). Strata on the north shore interfinger with and overlie bimodal volcanics, including feldspathic pyroclastic rocks. Despite rather poor exposure of the coarse grained components, the overall appearance is of classic fining-upward cycles, of trough crossbedded sandstone, rippled sandstone, mudstone, and coaly beds with root structures (e.g. Allen, 1970). There is no bioturbation. The fine grained component accounts for at least 50% of unit thickness. Although limited outcrop precludes three-dimensional analysis of the units, they are comparable to other ancient, high sinuosity fluvial channel deposits. Braided channel origins are tentatively precluded because of the high ratio of muddy deposits (Cant and Walker, 1976). There are no indications of channel stacking as might be expected in an anastomosing fluvial system.

Summary of Lower member

The depositional limits of Lower member strata of the Expedition Formation are similar to many of the older coarse grained units in Sverdrup Basin; there are no indications that the deposits encroached directly on Franklinian bedrock. Lower member facies assemblages are interpreted as the products of a large, west- or southwest-prograding, wave-dominated delta system – not necessarily as a single delta lobe, but as a series of coalescing lobes. The delta complex accumulated over a period of about eight million years. Relative rise in base levels over this period resulted in significant shifts in the shoreline during successive periods of delta construction and abandonment. Representatives of the delta at Strand Fiord constitute the prodelta and strandplain

Figure 12. Clean sandstone/carbonaceous sandstone couplets comprising an estuarine or tidal flat point bar facies at the top of the Lower member, Expedition Formation, Fossheim Anticline (RAK 21-85, see Fig. 11). Thicker sandstone beds containing dacite pebble conglomerate occur above the sub-Paleocene unconformity (dashed line). Hammer (circled) is 33 cm long. ISPG 2408-283

(delta front) components, whereas those further east on Fosheim Peninsula and May Point are the delta plain and adjacent coastal plain components. Interbedded sandstone, mudstone, and coal at Emma Fiord represent the greatest fluvial influence in the depositional system. A fitting modern analogue for the Lower member delta might be segments of the Texas coast, where rivers such as the Brazos and Rio Grande are presently building wave-dominated deltas with broad strand plains, which in turn supply sand to some of the world's largest barrier islands (Shepard and Wanless, 1971).

Upper member, Expedition Formation

General characteristics

Like the Lower member, the Lower Paleocene Upper member is dominated by sandstone, but differs from the former unit in two fundamental ways: the Upper member overlies a basin-wide sub-Paleocene unconformity; and Upper member sedimentation extended well beyond the depositional limits of the Lower member. Both of these factors played an important role in the ensuing Lower Paleocene sedimentation patterns, and reflect fundamental changes in the subsidence characteristics of Sverdrup Basin.

At Strand Fiord (Ricketts, 1991a, Fig. 49) the Upper member thickness ranges from zero at a depositional pinchout on the western end of Kanguk Peninsula, to 200 m near Kanguk River. Up to 540 m are recorded on Fosheim Peninsula (RAK 21-85, Fig. 2b), about 515 m along the north arm of Vesle Fiord (RAK 16-88, Fig. 2d), and at least 500 m south of Strathcona Fiord (RAK 2A-84, Fig. 2a).

The age of the Upper member is restricted to the Early Paleocene, contrary to an earlier designated age range of Maastrichtian to Early Paleocene (Ricketts, 1986). A suite of closely spaced palynology samples at Remus Creek (RAK 13-85) indicate that Lower Paleocene strata disconformably overlie upper Campanian rocks (Appendix 1). Foraminiferal representation in the Upper member is poor.

A distinctive conglomerate bed that contains dacite pebbles forms the base of the Upper member at Strand Fiord, various localities on Fosheim Peninsula, and May Point. In other areas, a significantly different basal contact is present. Where the Upper member unconformably overlies Lower member or older Sverdrup Basin strata, for example at Glacier, Strathcona, Bay, and Vesle fiords, the contact is at or close to the base of distinctive and easily mapped white sandstone beds. Similar white sandstone, in places underlain by thin carbonaceous shale, also marks the contact of the Upper member with deformed Franklinian bedrock, from Cañon Fiord to south of Strathcona Fiord.

Over most of the region, the contact between the Expedition Formation Upper member and the overlying, shale-dominated Strand Bay Formation is abrupt. However, the upper contact is less well defined where sandy, shoal-water facies of the basal Iceberg Bay Formation occur between Remus Creek and Hot Weather Creek on north Fosheim Peninsula (RAK 17-85 and RAK 21-85, Fig. 2b). Here, the contact is placed at the transition from (nonmarine) fining-upward to (marginal marine) coarsening-upward cycles, the latter containing some foraminifera.

Deposits of the Expedition Formation Upper member can best be described in terms of five basic facies assemblages: a delta assemblage, which like the Lower member is best represented on western Axel Heiberg Island (Ricketts, 1991a); an estuarine assemblage along the east and southeast basin margins (Ricketts, 1991c); a widespread assemblage that is loosely referred to as delta plain/coastal plain; a sandy shelf assemblage equivalent in part to the estuarine and delta/coastal plain assemblage; and a thin transgressive assemblage capping the Upper member.

The sub-Paleocene unconformity

Stratigraphic relationships between the basal few tens of metres of the Upper member and subjacent strata indicate considerable paleotopographic relief on the sub-Paleocene surface. Topographic relief is best demonstrated in the eastern and southern regions of Sverdrup Basin – these areas corresponding to marginal marine settings.

Two types of paleodrainage are observed on the unconformity (Ricketts, 1991c):

1. Shallow valleys: Broad, shallow valleys developed where the sub-Paleocene unconformity overlies older, relatively undeformed and weakly indurated Sverdrup Basin strata; primarily Expedition Formation Lower member and Kanguk Formation. Relief on the unconformity is identified where distinctive, white quartz sandstone present in the basal Paleocene succession are in contact with different stratigraphic levels of the subjacent units. For example on north Fosheim Peninsula the basal Paleocene sandstone beds overlie different thicknesses of the Expedition Formation Lower member (Fig. 13), and in places directly overlie Kanguk Formation shale (Fig. 14). The paleovalleys are typically kilometres to tens of kilometres wide and at least 100 m deep based on the thickness of missing Lower member rock. The valleys represent the product of relatively young, consequent drainage. Such drainage patterns were probably controlled by antecedent channels or distributaries in the older formations. The sub-Paleocene unconformity is mostly erosional. Where the lacuna is greatest, the basal Paleocene sandstone beds are thickest, and although along-strike exposures of this interval frequently are discontinuous, the basal beds appear to thicken and thin in concert with the different levels of sub-Paleocene erosion (Ricketts, 1991c).

2. Deep valleys: During the Early Paleocene expansion of the Sverdrup Basin limits, presumably in concert with increased subsidence and concomitant transgression (Ricketts, 1991c) resulted in Upper member sediments onlapping Lower Paleozoic bedrock of the Franklinian Mobile Belt (Fig. 15). The Paleozoic rocks were deformed during the Devonian and early Carboniferous, and pre-Paleocene erosion patterns clearly were controlled by the antecedent structures. The existence of considerable topographic relief on the sub-Paleocene surface can be observed at many localities, where basal Paleocene

strata infill depressions in the bedrock and progressively overstep valley walls. Distinctive white-weathering sandstone units (the white crossbedded sandstone facies), useful as stratigraphic markers, can be traced across paleovalleys to positions where they directly overlie bedrock. Excellent examples of these stratigraphic relationships are found south of Cañon Fiord where basal Paleocene sandstone overlies folded rocks ranging in age from Devonian to Ordovician, mostly in synclinal paleovalleys (Fig. 15). One such antecedent synclinal paleovalley at south Cañon Fiord is 8-10 km wide but topographic relief within the valley itself resulted in considerable local geomorphic complexity (Fig. 16a, b). White sandstone at the base of the valley fill, 60-70 m thick near the main valley axis, thins to less than 12 m against the

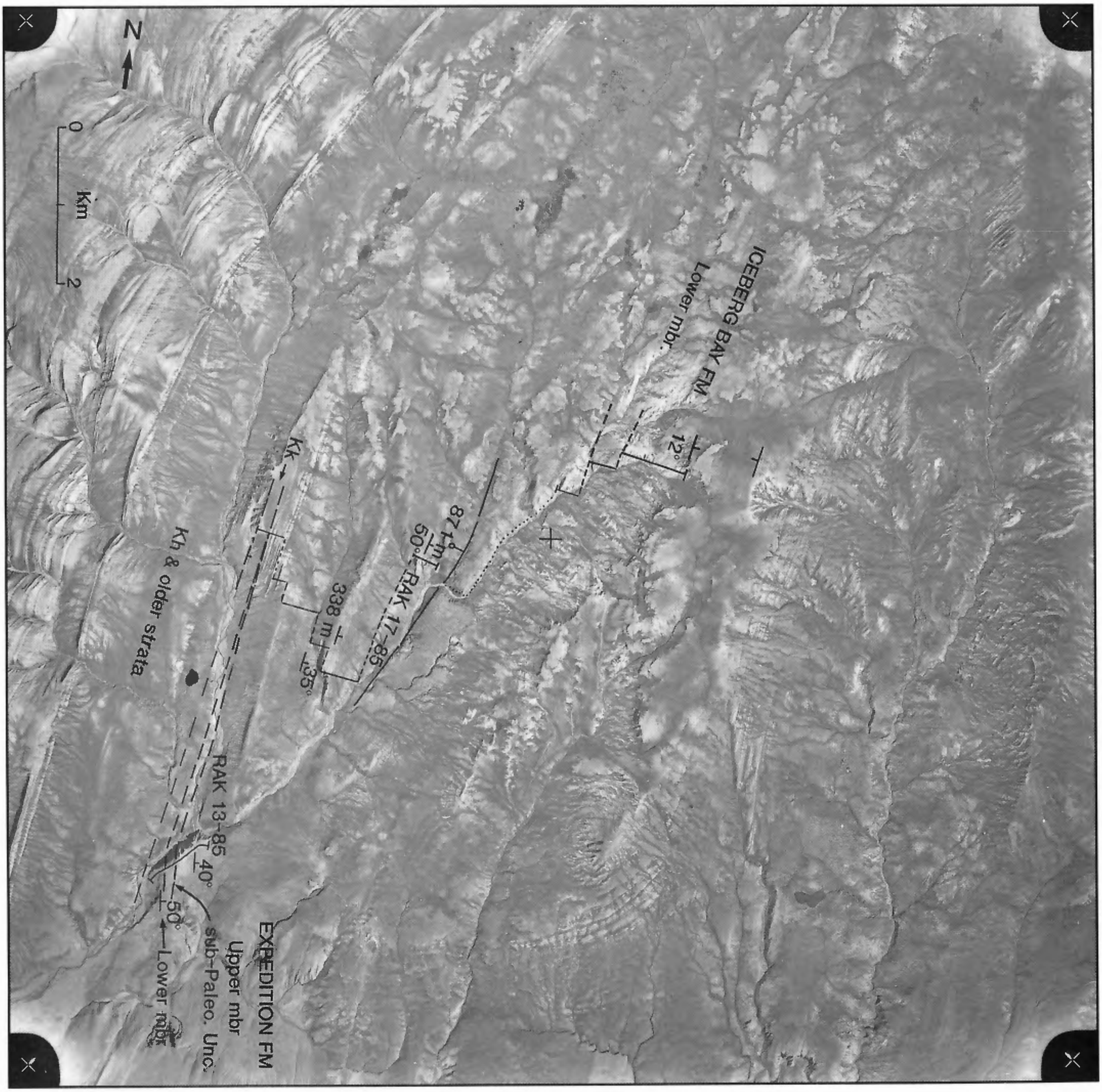

Figure 13. Aerial photograph (A 16686-16) of Remus Creek sections, north Fosheim Peninsula. Note the position of the sub-Paleocene unconformity and erosional pinch out of the Lower member, Expedition Formation between sections RAK 13-85 and RAK 17-85. Kk = Kanguk Formation; Kh = Hassel Formation. Photo centre — 80°00′N; 84°56′W.

bedrock 'high', or valley wall. Angular discordances of 15° between the bedrock and Upper member strata are not uncommon. Overstepping of bedrock highs by the basal Paleocene succession is again illustrated in an example near Mount Moore (Fig. 17). Local microtopography is also visible here.

Dissection of the Lower Paleozoic bedrock into relatively narrow (hundreds of metres to kilometres wide), deep valleys took place over an protracted hiatus encompassing most of the Mesozoic and probably involved several episodes of erosion and nondeposition. The drainage is therefore a mixture of much older consequent and subsequent patterns.

Figure 14. Aerial photograph (A 16605-38) of the northwest shore of Fosheim Peninsula, Eureka Sound, showing the Upper member of the Expedition Formation disconformably overlying shale of the Kanguk Formation (Kk). Photo centre – 79°44'N; 85°44'W.

Where basal Paleocene deposits overlie Paleozoic carbonate, limestone regoliths and shallow karst features (shallow dissolution joints) developed (Fig. 18). Some of these weathering features were responsible for the micro-relief developed in the more deeply incised valleys (Fig. 17), a feature not observed in the broader, shallow valleys.

Felsic volcanic clasts

Pebbles and cobbles of dacite and trachyte in lag-concentrates further distinguish the base of the Upper member at Remus Creek, Fosheim Anticline, May Point, and as far west as Kanguk River at Strand Fiord. Maximum clast diameters range up to 15 cm. These highly distinctive clasts are not found at any other stratigraphic level in the Eureka Sound Group, and because they abruptly overlie the sub-Paleocene unconformity, provide important clues to the nature of the early Paleocene margin of Sverdrup Basin. Bimodal, alkali basalt and rhyolite (including trachytes) that have ^{40}Ar/^{39}Ar ages of 65 ± 3 Ma occur in the Phillips Inlet area (northwest Ellesmere Island, Muecke et al., 1990). No other felsic volcanics of like composition are known in Sverdrup Basin. Silurian keratophyres and related rocks are known locally on northeastern Axel Heiberg and northwestern Ellesmere islands (Thorsteinsson and Trettin, 1972a, c), but apparently these can be distinguished from the Upper Cretaceous-Paleogene examples (G. Muecke, pers. comm., 1990). Therefore the most likely sources for the Upper member pebbles are the isolated felsic volcanic centres between Emma Fiord and Phillips Inlet.

It is possible that the volcanic clasts were reworked from older Maastrichtian regressive deposits (Lower member) but this hypothesis is not favoured here because of the following points:

i. the susceptibility of the clasts to mechanical abrasion, and,

ii. their distance from the putative source area near Emma Fiord – more than 250 km in the case of the Strand Fiord and May Point localities. It is more likely that the clasts are the products of first cycle erosion and sedimentation.

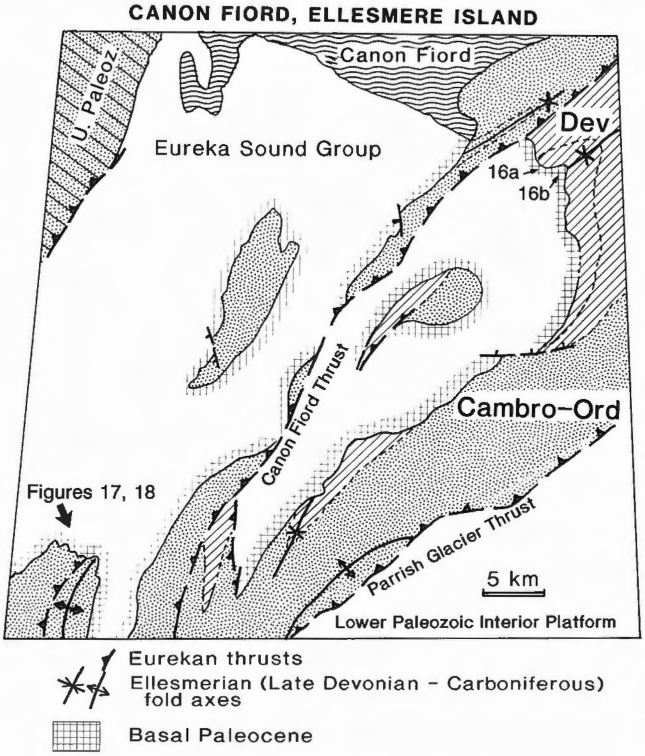

Figure 15. *Simplified map of the south Cañon Fiord area showing onlapping relationship of the Upper member, Expedition Formation, with various Cambrian to Upper Devonian units in the Franklinian Fold Belt. Bedrock geology modified from Thorsteinsson (1972b). See Figure 3 for map details of Eureka Sound Group geology.*

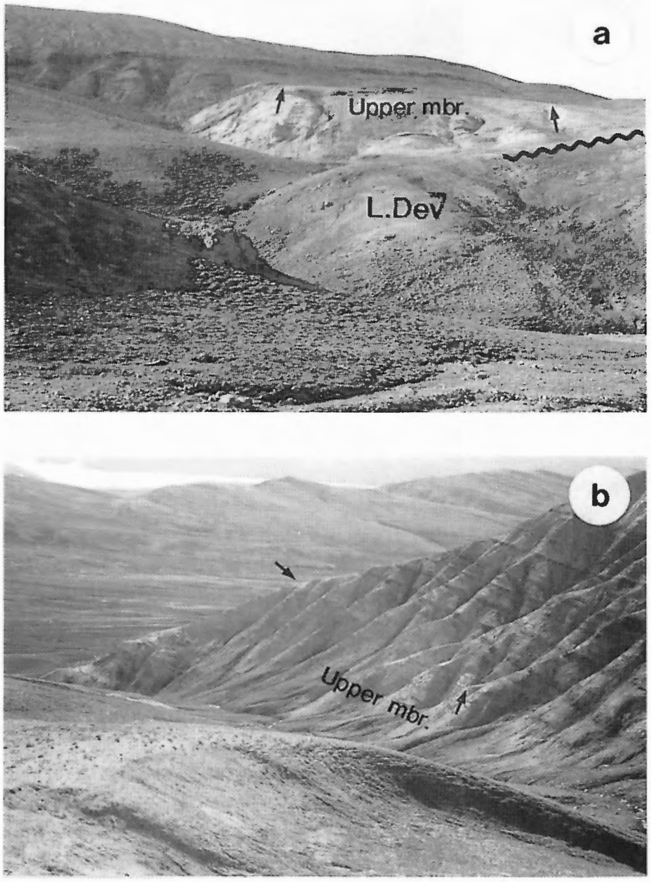

Figure 16. ***a)*** *Basal white sandstone of the Expedition Formation Upper member oversteps Lower Devonian bedrock with an angular discordance up to 25°, south of Cañon Fiord (see Fig. 15). The basal sandstone here is about 12 m thick, and is overlain by interbedded sandstone and coal of the coastal plain facies (above arrows; compare Fig. 16b). ISPG 3070-149;* ***b)*** *Basal white sandstone about 75 m thick, 2 km southwest of the site in Figure 16a, corresponding to infill of antecedent valleys in the Franklinian bedrock. Arrows indicate the contact between coaly coastal plain deposits and the white sandstone (location indicated in Fig. 15). Basal unconformity indicated at left centre. ISPG 3070-153*

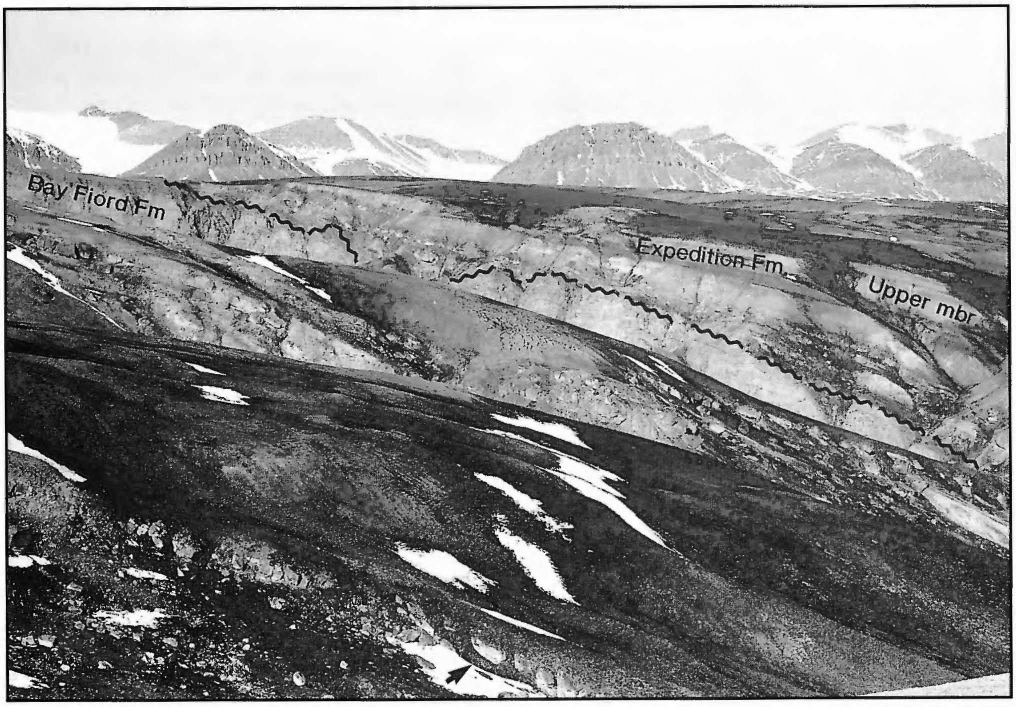

Figure 17. Sub-Paleocene unconformity exposed in the Mount Moore area (see Fig. 15). Basal estuarine strata of the Upper member, Expedition Formation onlap eroded Ordovician limestone (Bay Fiord Formation). Some of the microtopography on the unconformity resulted from shallow karsting. In the foreground (arrow) blocky weathering sandstone directly overlies bedrock. Ordovician rocks also underlie the rugged peaks in the background and form the hanging wall panel of a major (unnamed) Eurekan thrust. ISPG 2835-62

Wave-dominated delta facies assemblage

Upper member facies at Strand Fiord are similar to those in the Expedition Formation Lower member (for details see Ricketts, 1991a). Coarsening-upward cycles contain a lower shale unit typically of prodelta origin. Resistant ironstone-sandstone beds at the base of the shale components probably reflect reduced sedimentation rates, i.e. condensed stratigraphy. The coarsening- and thickening-upward cycles are several tens of metres thick and contain many small-scale cycles that mimic the larger cycle characteristics. Wave-generated bedforms predominate (parallel laminae, low-angle planar, some planar-tabular and rippled crossbeds). The opportunistic trace fossil *Skolithos* abounds, with as many as 3000 to 5000 burrows per square metre. Like their Lower member counterparts, the Upper member cycles on western Axel Heiberg Island are interpreted as prograding delta-strandplain deposits, where wave processes predominated.

Cliff-forming sandstone units up to 10 m thick, like those in the Lower member, exhibit the hallmarks of distributary channels. Large two- and three-dimensional subaqueous dunes are commonly veneered by lags of mudchips and plant fragments. Stratigraphic terminations in the channel deposits are of two kinds. In the first, the upper sandstone contact is abrupt, indicating sudden abandoning of the channel. In contrast, sandstone units that contain a fining-upward component and a concomitant upward decrease in bedform size, are indicative of laterally migrating channels.

Figure 18. The sub-Paleocene unconformity represented by a regolith (arrow) developed in Ordovician carbonate rocks. Mount Moore area (see Fig. 15). Carbonaceous shale and sandstone compose the estuarine facies in the onlapping Upper member, Expedition Formation. ISPG 2835-71

Estuarine facies assemblage – shallow paleovalleys

The following measured sections contain representatives of the shallow valley estuarine fill: RAK 3-85 (south Strathcona Fiord); RAK 13-85, RAK 17-85, RAK 21-85 (Fosheim Peninsula; Fig. 2b); RAK 7-88 (May Point; Fig. 2d).

Estuarine facies occupy the lower 50 to 100 m of the Upper member in the central and eastern part of Sverdrup Basin (summarized in Table 1, Fig. 19a; for details see Ricketts, 1991c). The white crossbedded sandstone facies forms a distinctive stratigraphic marker (Fig. 19b). Crossbed azimuths are generally towards the east or southeast. In

Table 1. Lithological characteristics of estuarine facies in the Upper member, Expedition Formation.

Facies	Thickness	Sedimentary structures	Bioturbation
SHALLOW VALLEY ESTUARINE FILL			
White crossbedded sandstone	18 to 20 m	- clean, well sorted - stacked planar-tabular crossbeds, pebble lags, reactivation surfaces, sand-mud couplets - trough crossbeds, within larger channel scours, ball-and-pillow - very large planar sets up to 6.5 m thick	
Coaly mudstone	5 m	- carbonaceous, roots - thin laminated sandstone, few current ripples - Metasequoia leaves and cones	indeterminate
Sandstone-mudstone	to 23 m	- coarsening, thickening upward - flaser, lenticular, wavy crossbeds, reactivation surfaces - crossbed dimensions increase up - low-angle planar crossbedding at top	*Skolithos, Chondrites Thalassinoides,* Planolites
Sandstone wedge	to 6 m	- clean, well sorted - sandstone beds thin and converge to southeast, replaced by coaly mudstone - stacked planar-tabular crossbeds within sandstone wedges	
Fining upwards	to 15 m	- sandstone to coaly mudstone - trough, ripple crossbeds - sandstone-mudstone couplets in large accretionary foresets	rare *Skolithos*
DEEP VALLEY ESTUARINE FILL			
Trough crossbedded	to 10 m	- clean, well sorted - overstep or abut valley walls - isolated channels, or laterally associated with foreset bed facies - some planar-tabular crossbeds	
Foreset bedded	to 6 m	- large planar-tabular foresets up to 6 m thick, individual or stacked sets - reactivation surfaces - some top-set beds	*Skolithos* *Paleophycus* *Planolites*
Fining upwards	to 5 m	- sandstone-mudstone couplets - sandstone bundles inclined as foresets	indeterminate
Sandstone-mudstone coaly mudstone		- as above	*Skolithos* *Muensteria*

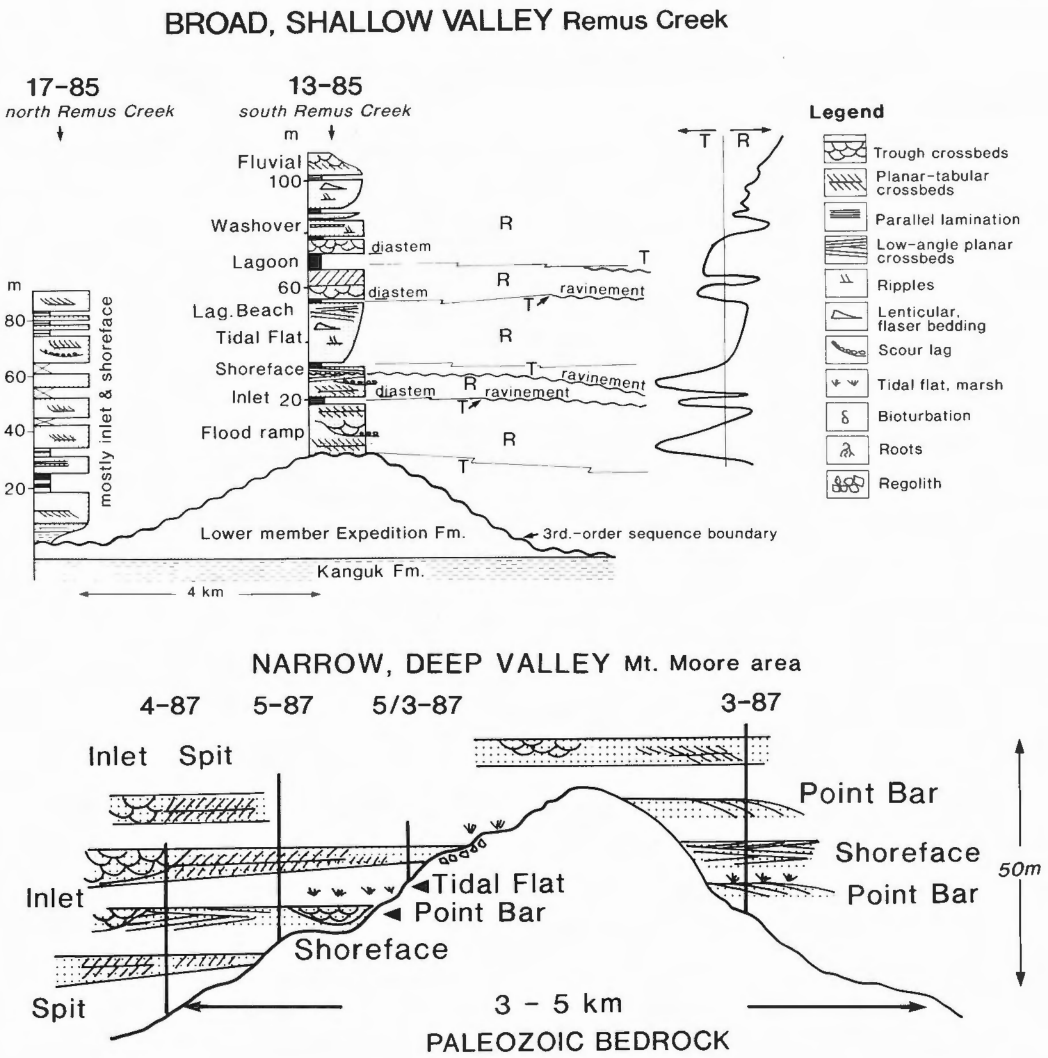

Figure 19. a) A comparison of stratigraphic relationships and facies in the two types of estuarine deposits: at Remus Creek overlying older Sverdrup Basin, and at Mount Moore overlying Lower Paleozoic bedrock. Detailed sections at Remus Creek are illustrated in Figure 2b; those at Mount Moore in Figure 2c. Modified from Ricketts (1991c).

places, tabular bedded sandstone containing low-angle planar crossbeds abruptly overlie the facies. Coarsening- or fining-upward trends are not present.

The coaly mudstone facies contains interbedded mudstone, shaly coal, and thin brown sandstone. Contact between the coaly mudstone facies and crossbedded sandstone facies is abrupt, but with other facies it is gradational. The coarsening-upwards sandstone-mudstone facies contains a distinctive array of relatively low energy bedforms indicative of tidal current asymmetry (flaser, lenticular, and wavy bedding), in addition to the trace fossils *Thalassinoides* and *Skolithos*. Low-angle planar crossbedded sandstone also caps some coarsening-upward cycles.

Beds of clean, well sorted sandstone making up the sandstone wedge facies are generally less than 25 cm thick and separated by veneers of carbonaceous mudstone (Fig. 20). The beds are stacked in an east- to southeast-thinning wedge, where sandstone is replaced by mudstone containing abundant plant fragments. Most sandstone wedges overlie the white crossbedded sandstone facies. They dip southeast towards the paleoshoreline.

Fining-upwards facies sandstone, mudstone, and shaly coal only occur at the top of the estuarine successions. A characteristic feature is the stacking of sand-mud couplets into large accretionary foresets, similar to the heterolithic stratification of Thomas et al. (1987).

Interpretation of the facies and their significance with respect to sea level fluctuations is detailed by Ricketts (1991c) and summarized in Table 2 and Figure 21. Broad, antecedent valleys along the eastern and southeastern margins of Sverdrup Basin were flooded during the earliest Paleocene transgression. Flooding, with an abundant supply of quartz sand that probably was relict on adjacent shelves and deltas, gave rise to a series of progradational cycles. These cycles

Figure 19. *b) Large, channel-fill trough crossbedded sandstone capped by stacked, planar-tabular crossbeds of the tidal inlet delta/channel facies, in basal sandstone of the Upper member, Expedition Formation; May Point, Section RAK 7A-88. Carbonaceous beds that cap the inlet facies are of probable lagoonal origin. GSC 1993-039A*

record successive transitions from barrier island, tidal inlet (channel and delta), and various back barrier environments such as lagoon, beach, storm wash-over lobes, and tidal estuarine channels with point bars. In most cases, the wash-over deposits are all that remain of the subaerial portion of the barrier islands.

Relative sea level fluctuations, on the scale of individual cycles, produced significant facies changes ranging from barrier shoreface to lagoon. With only a single exception (e.g. 227 m, RAK 13-85, Fig. 2b) ravinement surfaces are not preserved. In most cycles the transition from regression to transgression occurs within the lagoonal units, the corresponding ravinement presumably being developed seaward of the lagoon/barrier systems. Strata which have open, sandy shelf affinities that possibly are equivalent to the estuarine deposits, occur in south Wolf Valley (discussed below – RAK 16-88, Fig. 2d). Fluctuations in sea level, on the scale of third-order sequences, record general progradation where successive cycles become increasingly 'lagoonal' in character, and where the entire estuarine succession is overlain by delta plain/fluvial deposits.

Estuarine facies assemblage – deep paleovalleys

Measured sections containing relevant information are RAK 9-85 (south of Cañon Fiord) and RAK 11-85 (South Bay) (Fig. 2b); RAK 2-87 (below Schei Summit), RAK 3-87, RAK 4-87, RAK 5-87 (west of Mount Moore) (Fig. 2c); and RAK 19-88 (southeast Cañon Fiord) (Fig. 2d).

Five principal facies are recognized (Table 1, details in Ricketts, 1991c). Diagnostic of the facies are the abrupt lateral changes within and between paleovalleys (Fig. 19a). Although similar in some respects to the white crossbedded sandstone facies, the trough crossbedded facies differs significantly in its geometry. Trough crossbedded sandstone forms isolated channels up to several tens of metres wide, or channels laterally associated with the foreset bedded facies. These channels are commonly located in topographic depressions in the bedrock.

Table 2. Lithofacies interpretations for the estuarine deposits in the Upper member, Expedition Formation.

Facies	Interpretation
BROAD SHALLOW PALEOVALLEYS: Low relief, nondeformed bedrock, no microtopography, young consequent drainage controlled by antecedent fluvial or distributary channels.	
White crossbedded sandstone	Barrier island and spit, tidal inlet channel and flood delta, shoreface
Coaly mudstone facies	Lagoon, marsh
Sandstone-mudstone	Tidal flat, back-barrier beach
Sandstone wedge	Storm wash-over-lagoon
Fining upwards	Tidal estuarine channel, point bars
NARROW DEEP PALEOVALLEYS: High relief, deformed and indurated bedrock, pronounced local topography, old consequent and subsequent drainage.	
Trough crossbedded	Tidal inlet and tidal delta
Foreset bedded	Spits (attached to bedrock headlands)
Fining upwards	Tidal, estuary channel, point bars
Sandstone-mudstone	Tidal flat, lagoonal beach
Coaly mudstone	Lagoon, marsh

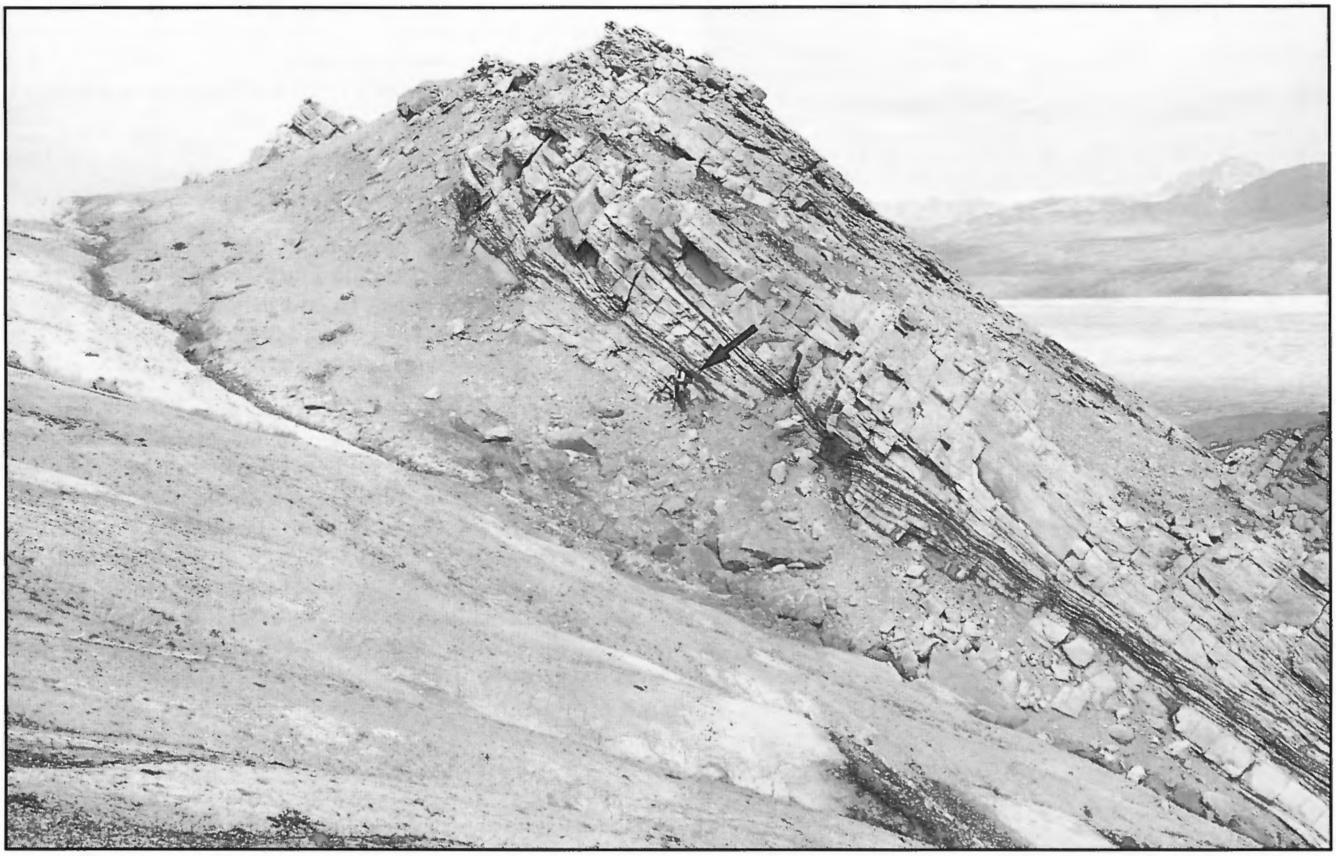

Figure 20. *Large, lagoonward-dipping sandstone wedge sets of the storm washover facies overlying tidal inlet sandstone. The wedge set interfingers with, and progrades over, carbonaceous, back barrier mudstone. South shore of Bay Fiord, due south of Marie Island. Geologist (centre) is 1.85 m high. ISPG 2238-91*

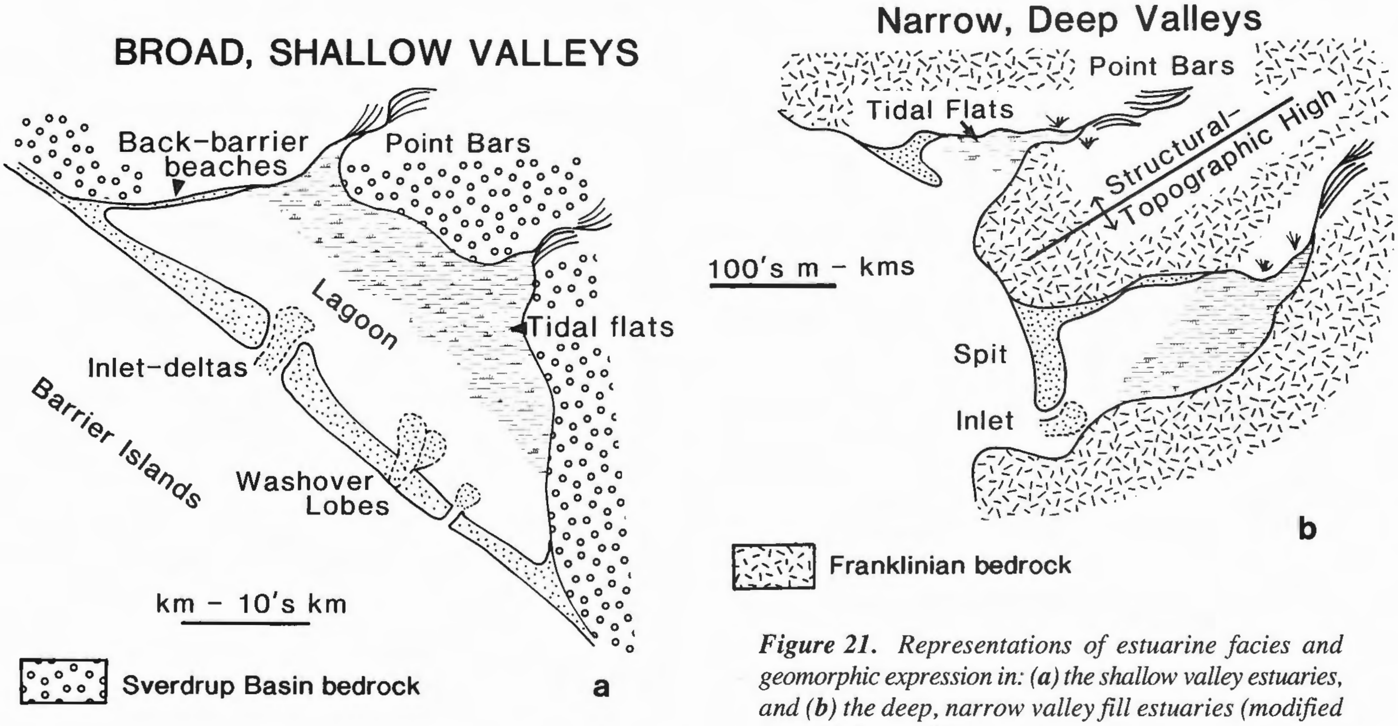

Figure 21. *Representations of estuarine facies and geomorphic expression in:* ***(a)*** *the shallow valley estuaries, and* ***(b)*** *the deep, narrow valley fill estuaries (modified from Ricketts, 1991c).*

The conspicuous foreset bedded facies contains individual planar-tabular sets up to 6 m thick. Contact with other facies is usually abrupt. The fining-upwards facies, like that in the shallow valley successions, is characterized by large, shallow-dipping foresets having sandstone-mudstone couplets and tidal bundles (e.g. Smith, 1988). Strata composing the sandstone-mudstone and coaly mudstone facies also are similar to those present in the shallow valleys.

Paleoenvironmental interpretation of the deep valley deposits is illustrated in Fig. 21 (and Table 2). The transition from estuary to shallow shelf is delimited by spits and bars attached to headlands along the highly irregular Early Paleocene coast. Components of the estuary include narrow tidal flats and lagoonal beaches, cut by small tidal channels. Tidal channels containing point bars reflect the more sinuous aspect of channel thalwegs and the landward transition to more fluvial-dominated settings. Like the shallow valley estuaries, the deep valley estuarine complex is overlain by extensive delta plain deposits.

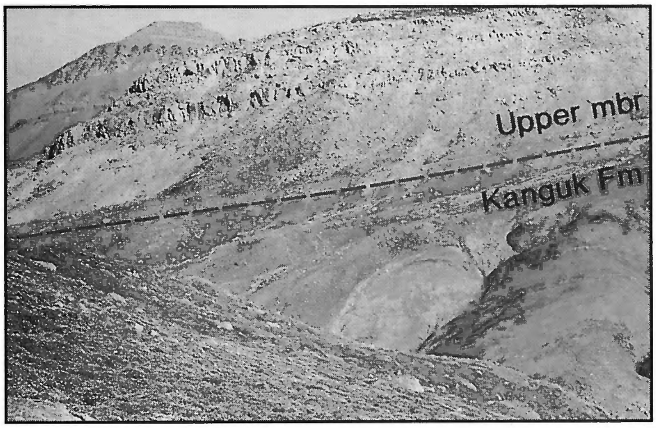

Figure 22. Thick, tabular bedded sandstone at the base of the Expedition Formation Upper member, north arm of Vesle Fiord (Section RAK 16-88). Contact with the Kanguk Formation appears to be disconformable. Each sandstone unit contains stacked shallow shelf (two dimensional) dunes (see Fig. 25). ISPG 3070-82

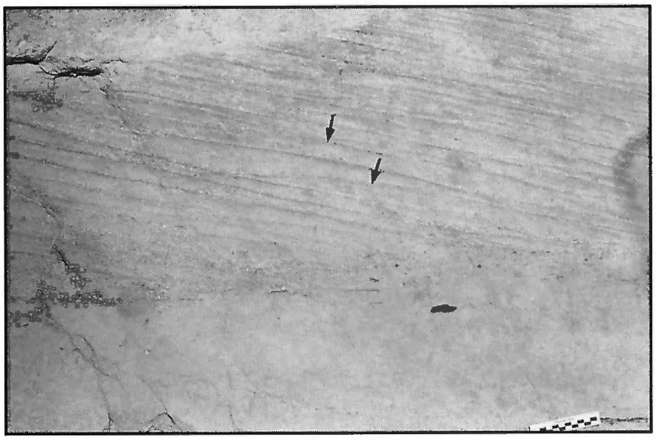

Figure 23. Bundles (couplets) of clean light coloured sandstone and a veneer of fine carbonaceous sandstone, occur in foresets of small, two dimensional dunes (tabular-planar crossbeds). Reactivation surfaces also are present (arrows). Same location as Figure 22. ISPG 3070-86

Sandy shelf facies assemblage

The only observed outcrops of the sandy shelf facies are located along the north arm of Vesle Fiord, south-southwest of Wolf Valley; section RAK 16-88 (Fig. 2d). Here the basal Paleocene succession lies below the footwall of Vesle Fiord Thrust. The Upper member disconformably overlies the Kanguk Formation.

The lower 260 m of the succession contains stacked, tabular bedded sandstone units each up to 20 m thick, separated by recessive intervals of about the same thickness (Fig. 22). Some units show crude coarsening-upward trends. Contact between successive units is abrupt and planar. None of the cycles contains evidence of subaerial exposure or shoreface deposition.

Crossbedding is dominated by stacked, planar-tabular sets up to 1.5 m thick, and a few trough crossbeds. Bundles of alternating light (medium grained) and dark (slightly carbonaceous and finer grained) foreset layers, 1-2 cm thick, many having reactivation surfaces, are common in the planar-tabular varieties (Fig. 23). The dark layers are generally only a few millimetres thick. Fine pebble (chert, vein quartz, ferruginous mudstone) lags are also common. Wood fragments are strewn across bedding. Dish structures and water escape pillars created local distortion of foreset layers. Foreset azimuths tend to be bimodal with primary flow directions to the east-northeast and west-southwest (Fig. 24). Bioturbation is rare in the crossbedded sandstone units. There are no mudstone drapes.

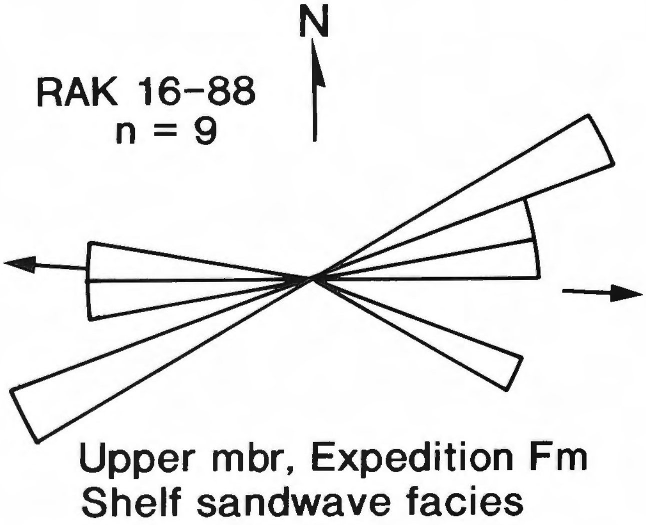

Figure 24. A bimodal-bipolar distribution of planar-tabular crossbed azimuths in the shelf sandwave facies along the north arm of Vesle Fiord. (RAK 16-88)

The large planar-tabular crossbeds are contained within large scale wedge-shaped units bound by planar master surfaces (Fig. 25). The wedge sets thicken from a pinchout up to 4 m, frequently towards the southwest. Wedge boundaries tend to dip towards the northeast at 5-8°. Individual wedges have been traced more than 60 m along bedding (limited by outcrop). Some wedge sets consist of a single crossbed having foresets dipping 15°-20° to the northeast, whereas others contain stacked crossbeds. Crossbed migration in successive wedge sets was down-dip on the master surface. Successive wedge sets are stacked within each sandstone unit.

Exposure of the interbedded fine grained sandstone and mudstone in the intervening recessive units is poor. Parallel lamination and ripple crossbedding are common. A few trails were observed in talus blocks.

Bimodal paleocurrent directions, sandstone bundles, and reactivation surfaces, plus the absence of any shoreface or subaerial indicators, all suggest a tide-dominated, subtidal shelf setting. Each sandstone unit is interpreted as a complex of amalgamated sand waves (after the definition of Allen, 1980; more recently referred to as large, two dimensional subaqueous dunes – Ashley et al., 1990). A three-fold hierarchy of bedform bounding surfaces reflects the development of the sand wave complexes (Fig. 26). Third-order sandstone bundles in planar-tabular crossbeds reflect foreset accretion during successive tidal cycles. The limited amount of paleocurrent data suggests approximately equal flood and ebb tidal flow energy. Second-order surfaces, that define the larger wedge sets, represent the active surfaces of sand waves having amplitudes up to 4 m and wave lengths in excess of 60 m. Northeasterly dips on these surfaces indicate that flood-directed flow may have dominated over the long term development of the sand waves. First-order surfaces, which define the upper and lower bedding of entire sandstone units, formed by the accretion of successive sand waves.

The internal organization of these sand waves corresponds to Allen's (1980) categories IVA and V, which in terms of hydrodynamic regime, represent the transition from flow separation to unseparated flow at the dune lee face. They are also similar to modern (Berné et al., 1988) and some ancient sand wave complexes (e.g. Allen and Homewood, 1984; Uhlir et al., 1988), although differences do exist. For example, tidal bundles described from several ancient analogues are commonly thicker and contain mud drapes. Morphologic differences such as these may be attributed to the paleogeographic position of sandwave accumulation; sandwaves in estuarine channels or tidal inlets are more susceptible to

Figure 25. *Large wedge sets bounded by second-order surfaces typical of sandstone units in the sandy shelf facies, Upper member, Expedition Formation. Each second-order set is made up of smaller scale (third-order) crossbeds (Fig. 23). Jacob's staff (bottom right) is 1.5 m long. North arm Vesle Fiord, RAK 16-88 (see also Figure 26). ISPG 3070-92*

tidal velocity asymmetry, and hence slack water conditions, than sandwaves that accumulate on more open shelves (Terwindt and Brouwer, 1986). Tidal bundles in trough crossbeds (3-D dunes) also tend to be thicker than those in 2-D dunes. Bundle thickness distribution in Figure 27 does not exhibit a categorical distinction between a diurnal or semi-diurnal tidal regime; perhaps it was a mixed regime on this Early Paleocene shelf. Flood-tide dominance observed in the estuarine tidal inlets also appears to be reflected in the second-order surfaces of individual sandwaves on the shelf, although this dominance is less pronounced.

The basal portion of the sandwave complex at Vesle Fiord may have been the offshore equivalent to both the broad, shallow estuarine valley and narrow, deep estuarine valley successions. However, the bulk of the sandwave succession was probably related to the subsequent regressive coastal plain depositional phase (discussed below). Successive sandstone units (bound by first-order surfaces) low in the Vesle Fiord succession can, in a general way, be equated with the cyclicity observed in the estuarine deposits. Contact between each sandstone unit and the overlying fine grained lithologies might correspond to the transition from progradation to transgression (ravinement) in the estuarine deposits. However, the sandwaves developed seawards of the ravinement surface and the sandstone-mudstone contact in this case therefore represents a combination of initial transgression and maximum flooding.

The thickness and repetition of subtidal shelf sandstone units attests to the continued availability of sand, the supply of which presumably was linked to the coastal/estuarine regime. Sandwave complexes are known in some modern wide-mouthed estuaries which appear to be transitional with adjacent shelves, for example Bristol Channel and Thames Estuary (Harris, 1988). It is also possible that sand derived from the adjacent wave-dominated delta (west Axel Heiberg area) was transported to the eastern and southeastern margins by longshore currents although there is no direct proof of this.

How was the sand transported from the estuaries and overlying coastal plain system onto the adjacent shelf? Clear evidence of storm deposition on the Lower Paleocene shelf is lacking, and in shoreface deposits attached to the estuarine systems hummocky cross-stratification is uncommon. In the present scenario, it is conjectured that sand transported onto the shoreface (of barriers and spits) during each phase of progradation became palimpsest during subsequent transgressions. During the succeeding stage of progradation this sand was reworked into sandwaves on the adjacent shelf. A consequence of this interpretation is that each cycle of sandwave formation would have been out of phase with each cycle of barrier and spit progradation.

Delta plain/coastal plain facies assemblage

This loosely defined assemblage constitutes a large part of the Expedition Formation Upper member, but is the least exposed. It gradationally overlies the basal estuarine and sandy shelf facies in eastern and southeastern Sverdrup Basin (Fig. 28). The bulk of the Upper member can thus be viewed as a shoaling- upward succession of major proportions. The assemblage is laterally equivalent to the wave-dominated delta facies on west Axel Heiberg Island (Fig. 6, 7). Relevant measured sections are: RAK 2-84 and RAK 2A-84 (Strathcona Fiord) (Fig. 2a); RAK 9-85 (south Cañon Fiord), RAK 17-85 and RAK 21-85 (Fosheim Peninsula; Fig. 2b); RAK 3A-87 (west of Mount Moore; Fig. 2c); RAK 16-88 (Vesle Fiord) and RAK 19-88 (southeast Cañon Fiord; Fig. 2d).

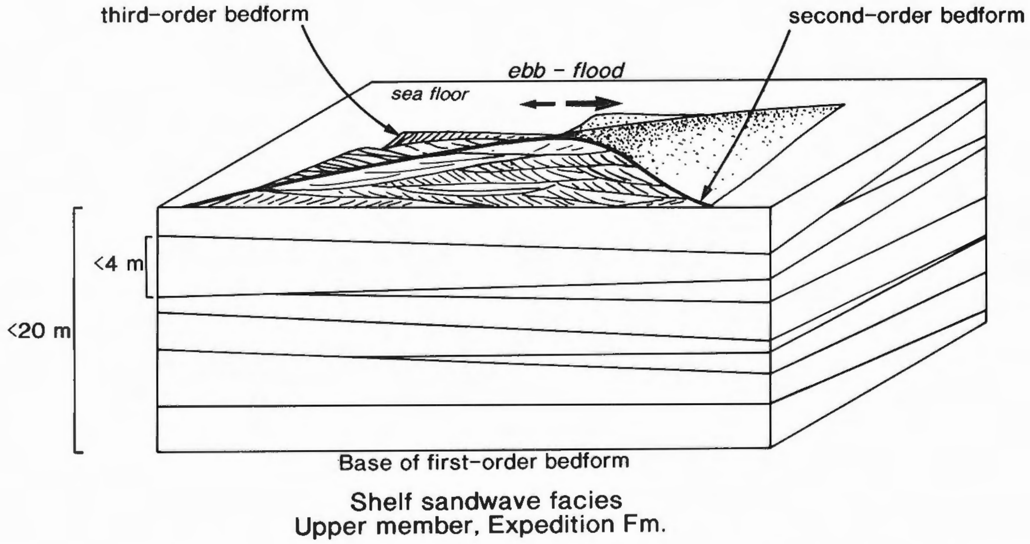

Figure 26. A schematic representation of bedform hierarchy in the shelf sandwave facies, showing first-, second-, and third-order bounding surfaces.

Most of the lithofacies consists of interbedded buff- and rusty-weathering, fine grained sandstone, mudstone, and coal, arranged into coarsening- and fining-upward cycles, usually less than 10 m thick. Fining-upwards cycles become more common towards the top of the Upper member. Bioturbation and other indicators of marine deposition also decrease upwards.

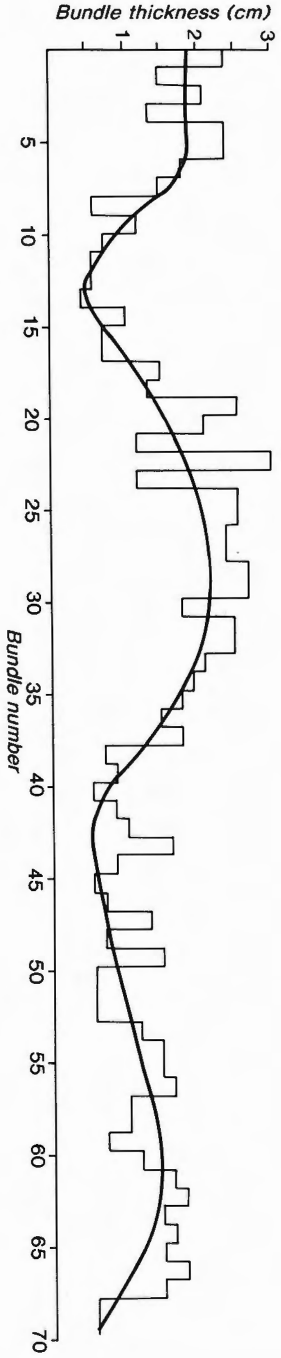

Figure 27. Plot of bundle thickness/bundle number for a single two-dimensional dune at the 27 m level in RAK 16-88; Upper member, Expedition Formation, shelf facies. Only one fully developed cycle of 28 days preserved here. The succeeding cycle is less well preserved.

Fining-upward cycles low in the succession are, in many cases similar to those encountered in the estuarine deposits and are characterized by sandstone-mudstone couplets, bioturbation (*Skolithos, Planolites*), and indicators of tidally influenced deposition such as flaser, lenticular, and wavy bedding, and reactivation surfaces on ripples. Deposition on tidal, or estuarine point bars also is indicated by accretionary foreset geometry in some units. The tops of the point bars were capped by muddy coal, some containing upright tree stumps, which resulted from encroaching marshes and coastal woodlands. Intervals of mudstone and coal a few metres thick are interrupted by thin, abrupt-based sandstone beds containing abundant current ripple and ripple drift crossbeds (Fig. 29), similar in organization to overbank or crevasse splay sands associated with delta plain channels. *Chondrites* is present in the muddy lithologies.

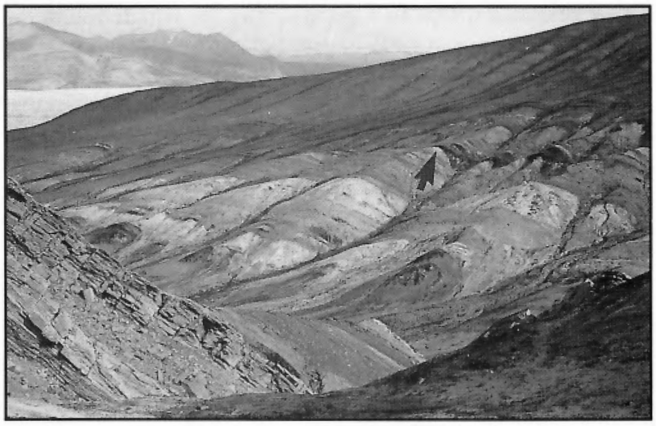

***Figure 28.** The transition (arrow) from basal estuarine deposits of the Upper member, Expedition Formation, to coastal plain facies is gradational and reflects blanketing of the antecedent topographic relief; south shore, Bay Fiord. ISPG 2238-90*

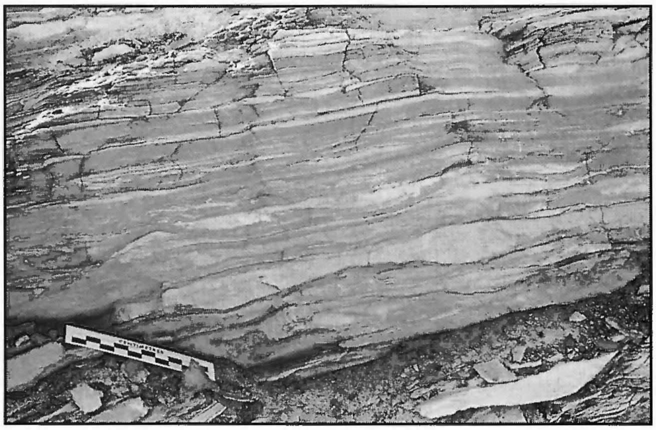

***Figure 29.** Lenticular, flaser, and wavy bedding in tidal deposits on estuarine point bars, delta plain/coastal plain facies; Upper member, Expedition Formation. Level 430 m, section RAK 16-88; scale in centimetres. ISPG 3070-107*

Coarsening-upward cycles of mudstone to buff-weathering sandstone display a similar array of sedimentary structures indicative of tidal influence (Fig. 30). However, low-angle planar and small planar-tabular crossbeds also are present in buff sandstone in the upper part of the cycles. Clean, white-weathering quartz sandstone beds occur locally. Some coarsening-upward cycles are capped by muddy coal beds containing abundant root structures. The coarsening-upward cycles developed along successive, low wave- and current-energy shorelines, perhaps with narrow beaches of clean sand which formed in slightly more open settings. Many modern delta plains and distributary bays possess similar facies associations (see Coleman, 1981; Elliot, 1974). The low-energy shoreface deposits interfinger with, and alternate stratigraphically with, other low energy coastal-plain tidal channel and marsh facies. The coastal plain strata in turn interfinger with delta plain and crevasse splay deposits as a result of frequent incursions from the major delta complex centred on western Axel Heiberg Island.

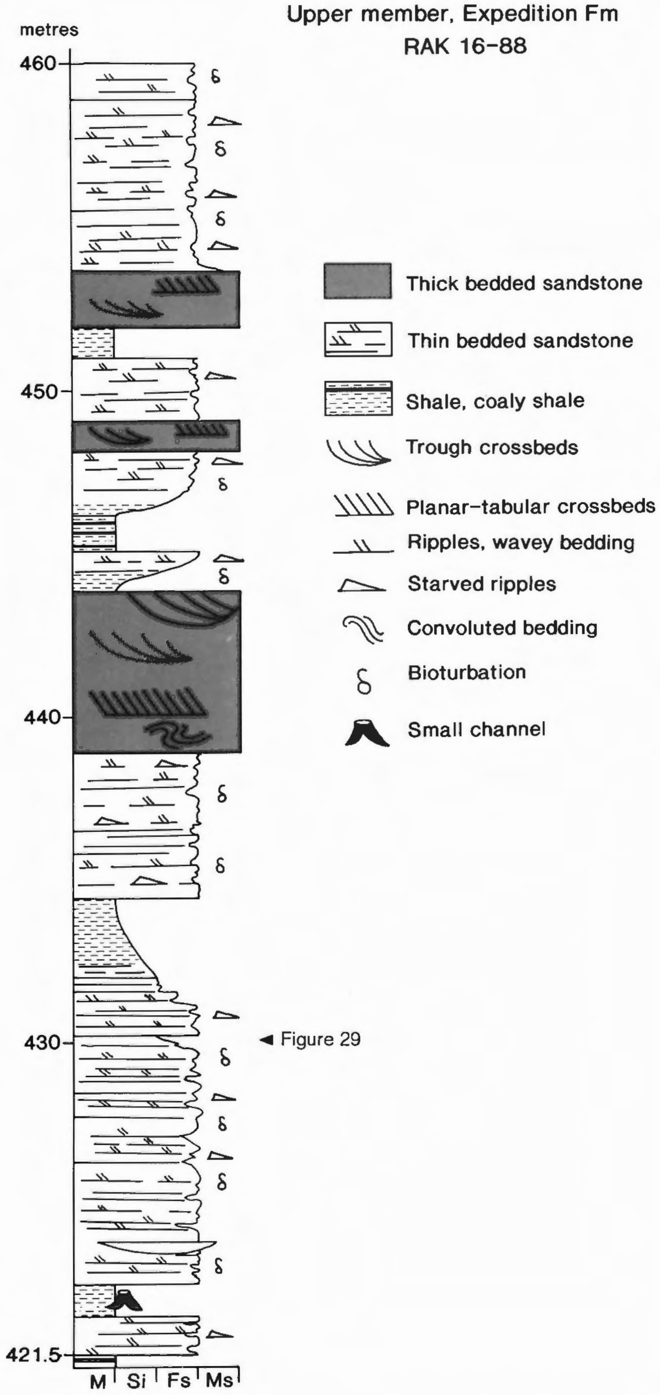

Figure 30. Expanded portion of section RAK 16-88 (Fig. 2d) shows typical, prograding, mixed sand/mud tidal flat and vegetated marsh facies, cut by tidal channels and developed in a back-barrier setting; Upper member, Expedition Formation.

Cycles a few metres thick in the coastal plain/delta plain facies assemblage reflect frequent migration of the paleoshoreline. Towards the top of the Upper member, fining- upward cycles of greater fluvial aspect become more common (except at RAK 16-88, Vesle Fiord area). Coal seams are thicker and less muddy. Flaser and lenticular crossbeds are absent, and bioturbation is rare. Trough crossbeds also are more common in the sandstone beds. Fining-upward cycles of this type are usually equated with high sinuosity fluvial channel and associated overbank sedimentation. Therefore, the Expedition Formation Upper member displays an overall progradational trend, which in the eastern and southeastern regions of Sverdrup Basin, culminated in fluvial dominated channel systems in the upper reaches of the coastal plain and adjacent delta plain.

Transgressive facies assemblage

Capping the Expedition Formation Upper member in several areas is a sandstone unit up to 55 m thick, for which the following observations are made (e.g. Strathcona Fiord, RAK 2A-84, Fig. 2a at 484-501 m; north Vesle Fiord, RAK 16-88, Fig. 2d, at 460-515 m; Kanguk Peninsula, RAK 25-83 at 717-723 m, Ricketts, 1991a, Fig. 49). Contact between the sandstone unit and underlying coastal plain/delta plain deposits is abrupt and locally erosional. For example, in section RAK 25-85 (Strand Fiord) the sandstone overlies and erodes a coaly mudstone that contains root structures. Contact with the overlying prodelta shale (Strand Bay Formation) is equally abrupt (Fig. 31). The sandstone consists of clean, well sorted quartz sand, and low-angle planar, stacked tabular-planar and some trough crossbeds. In section RAK 16-88, large accretionary foresets draped with mud veneers and burrowed by *Skolithos* occur low in the sandstone unit. In section RAK 2A-84 the sandstone is gritty. All of these features, when compared to similar facies in the Expedition Formation suggest that sandstone deposition took place in a middle to upper shoreface setting.

Regionally the Expedition/Strand Bay formation contact reflects a major change in base level. Considering all of the above factors, the uppermost Expedition Formation sandstone is interpreted to be the product of regional transgression. Sand, derived from the underlying coastal plain, was reworked as relative sea level rose, and the shoreface migrated landward. The lower contact of the sandstone unit corresponds to a ravinement surface; on a regional scale this surface is identified as a third order sequence boundary (discussed more fully in Stratigraphic sequences and paleogeography section). Ravinement probably removed any evidence of subaerial exposure except at Strand Fiord where the surface overlies a coaly paleosol. Contact with the overlying prodelta shale corresponds to the maximum flooding surface during transgression.

Summary of the Upper member

Regional uplift and erosion of Sverdrup Basin at the end of Cretaceous time gave rise to a sub-Paleocene disconformity of basin-wide extent, and a new phase of basin evolution. Along the eastern and southern parts of Sverdrup Basin, initial sedimentation on the eroded sub-Paleocene surface during Early Paleocene transgression, produced valley-fill estuaries (shallow and deep valley facies assemblages). Part of the shelf sandwave succession (sandy shelf facies) may have been coeval with the estuarine fill, however the bulk of the shelf facies probably accumulated during subsequent regional progradation. Farther north and west, large quantities of sand accumulated on a wave-dominated delta, similar to the delta which formed during Lower member deposition (middle Campanian to Maastrichtian). Creation of significant accommodation space, coupled with an abundant supply of sediment during the Early Paleocene resulted in accumulation of a thick west- to southwest-prograding coastal plain succession in the eastern and southern regions of Sverdrup Basin, that interfingered with delta plain deposits to the north and west. Regional marine onlap at the end of the Early Paleocene, signifying the transition to the thick Strand Bay Formation shale succession, is recorded by thin transgressive shoreface sandstone that forms the uppermost unit of the Expedition Formation. The base of the transgressive facies is a ravinement surface, which from a regional point of view, corresponds to a third-order sequence boundary. The Expedition Formation-Strand Bay Formation contact is a surface of maximum flooding.

STRAND BAY FORMATION

Thick Paleocene shale that can be mapped throughout eastern Sverdrup Basin on Axel Heiberg and Ellesmere islands and correlated beyond, has been assigned to the Strand Bay Formation (Ricketts, 1986). Contact with the underlying Expedition Formation is abrupt; that with the overlying Iceberg Bay Formation is gradational. Because the upper contact is gradational, the positioning of the map boundary is somewhat arbitrary, but generally is located at the point where sandstone predominates, coinciding with more resistant outcrop.

The type section on Kanguk Peninsula is 287 m thick, thinning basinward to 141 m at the west end of the peninsula (RAK 25-83 and 27-83 respectively in Fig. 49, Ricketts, 1991a). Strand Bay Formation shale is widely distributed around west-central Ellesmere Island although complete or well exposed sections are uncommon; about 200 m are present at north Vesle Fiord and Fosheim Peninsula, and between 276 m and 350 m in the Mount Moore – south Cañon Fiord area (e.g. RAK 6-87, Fig. 2c).

Figure 31. The sandy, transgressive facies that was a precursor to the Strand Bay Formation shale, exposed in tabular sandstone beds at the top of the Expedition Formation. Note abrupt eroded base; south shore of Kanguk Peninsula, section RAK 25-83. GSC 1993-039B

Age

Based on palynomorphs and foraminifera at the type section, a mid-Paleocene age is assigned, perhaps extending into the Late Paleocene (Ricketts, 1991a). The foraminiferal assemblage at Strand Fiord (RAK 25-83, Fig. 49, Ricketts, 1991a) and in the Strathcona Fiord area (RAK 2A-84, Fig. 2a) contains index species similar to those in the Ministicoog Member (Moose Channel Formation) in the Beaufort-Mackenzie Basin (*Reticulophragmium* sp. and *Cibicidoides* sp., Wall et al. (1988) Appendix 2). A *Saccammina-Hippocrepina* dominated assemblage, although sparse, occurs in many sections of the Strand Bay Formation, and is similar to Late Paleocene/Early Eocene assemblages in the Aklak Member of the Reindeer Formation. However, the Strand Bay Formation is no younger than Late Paleocene because it is overlain everywhere by Iceberg Bay Formation that is as old as Late Paleocene.

Palynomorph representation in this fully marine formation is generally poorer than in other units of the Eureka Sound Group. However, in areas other than Strand Fiord a mid- to Late Paleocene age is usually indicated by forms such as *Momipites wyomingensis* (e.g. RAK 2A-84, GSC sample number C-122568, Appendix 1).

Facies assemblages

Despite the predominance of shale in the Strand Bay Formation, varying amounts of sandstone exist, permitting the characterization of three depositional facies assemblages: prodelta, muddy shelf-below wave base, and a muddy shelf-above wave base. It should be emphasized that all three facies were the products of generally regressive/progradational conditions that followed the mid-Paleocene regional transgression.

Prodelta facies assemblage

Shale that accumulated in a prodelta setting is generally confined to the west Axel Heiberg Island region (details in Ricketts, 1991a). The succession on Kanguk Peninsula, up to 287 m thick, contains only 10% sandstone of a kind completely different from that in Strand Bay successions on Ellesmere Island (Ricketts, 1991a). Furthermore, Strand Bay shale on west Axel Heiberg Island grades into delta front sandstone of the overlying Iceberg Bay Formation, whereas in other areas the stratigraphic transition is to shelf deposits.

Sandstone units within the prodelta facies at Strand Fiord, up to 10.2 m thick, have abrupt tops and bases and do not form part of any fining or coarsening upward trend (see Fig. 18, Ricketts, 1991a). Basal contacts are locally erosive. Small planar-tabular and trough crossbeds are common. The sandstone is compositionally and texturally mature. The number and thickness of sandstone beds decreases westwards along Kanguk Peninsula. At one locality (RAK 27-85) a sandstone unit is capped by thin, sulphurous, coaly mudstone with root structures, which is overlain by almost 100 m of marine shale.

The sandstone units clearly represent high energy sediment transport, and perhaps even confined flow where three-dimensional bedforms predominate. Given the prodelta setting for much of the shale, these units present something of a dilemma for interpretation. In an earlier discussion I outlined two possible scenarios (Ricketts, 1991a): the sandstone bodies represent offshore sand bars, encased in shale; or alternatively they represent an abandonment facies of barrier island/inlet deposits, analogous to the modern Chandeleur Island system in Mississippi Delta (Frazier, 1967; Penland et al., 1985). Because the sandstone bodies occur within the progradational component of the Strand Bay Formation, the analogy with abandonment facies seems unlikely. More appropriate is the comparison to offshore bars, or perhaps a brief period of delta growth, in response to small scale changes in relative sea level superposed on the regional transgressive/regressive regime. A relative sea level drop could have resulted in new sand being added to the delta platform. Subsequent sea level rise would have terminated this supply, the sand then being moved landward in concert with a shoreface ravinement. The only component of the regressive barrier island system to be preserved would be the inlet facies which would survive if deposited lower than the ravinement (e.g. Demarest and Kraft, 1987). If this interpretation is correct, the ravinement surface corresponds to the abrupt upper contact of the sandstone bodies; the lower sandstone contacts being channel diastems.

Muddy shelf facies assemblage 1 – below wave base

Up to 200 m of shale having less than 10% sandstone interbeds can be mapped around Strathcona Fiord (RAK 2A-84, Fig. 2a), east Vesle Fiord (RAK 5-85, Fig. 2b), and the north arm of Vesle Fiord (RAK 16-88, Fig. 2d). Monotonous units of grey, blocky weathering shale are interrupted by scattered, thin, poorly sorted, fine- to medium-grained sandstone beds. Many of the sandstone beds, less than 10-20 cm thick, have abrupt tops and bases. Some are normally graded Bouma Tb units. Current ripples are rare. There are no indications of shoaling or of wave interactions with the substrate.

In some respects shale deposits of the muddy shelf facies resemble the prodelta facies but lack sandstone sheets. Also, basal Iceberg Bay Formation strata (next section) in this part of Sverdrup Basin, into which the Strand Bay shale grades, are of shelf origin rather than delta origin. Therefore, this component of the Strand Bay Formation is regarded as the outer, muddy portion of a broad shelf, albeit one that was adjacent to the large Axel Heiberg delta complex to the west and northwest. Much of the outer shelf was below storm wave base.

Muddy shelf facies assemblage 2 – above wave base

Facies of the Strand Bay Formation which contain up to 20% sandstone (per cent thickness) are mapped east and northeast of the mud dominated shelf facies, in the area between Cañon Fiord and Mount Moore. Two representative sections are illustrated in Figures 2b (RAK 19-85) and 2c (RAK 6-87).

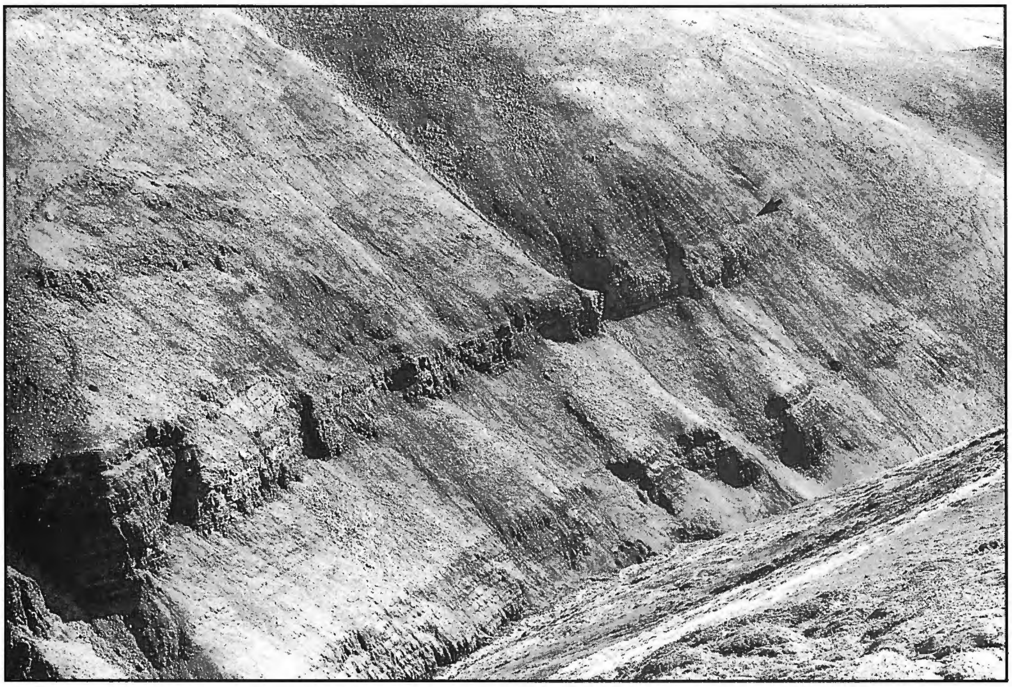

Figure 32. Local angular discordances of 5°-8° (arrow) or downlap surface developed above a coarsening-upward cycle in the muddy shelf facies, Strand Bay Formation; Section RAK 6-87; (Fig. 2c), Mount Moore area; muskox hoofprints for scale. ISPG 2835-80

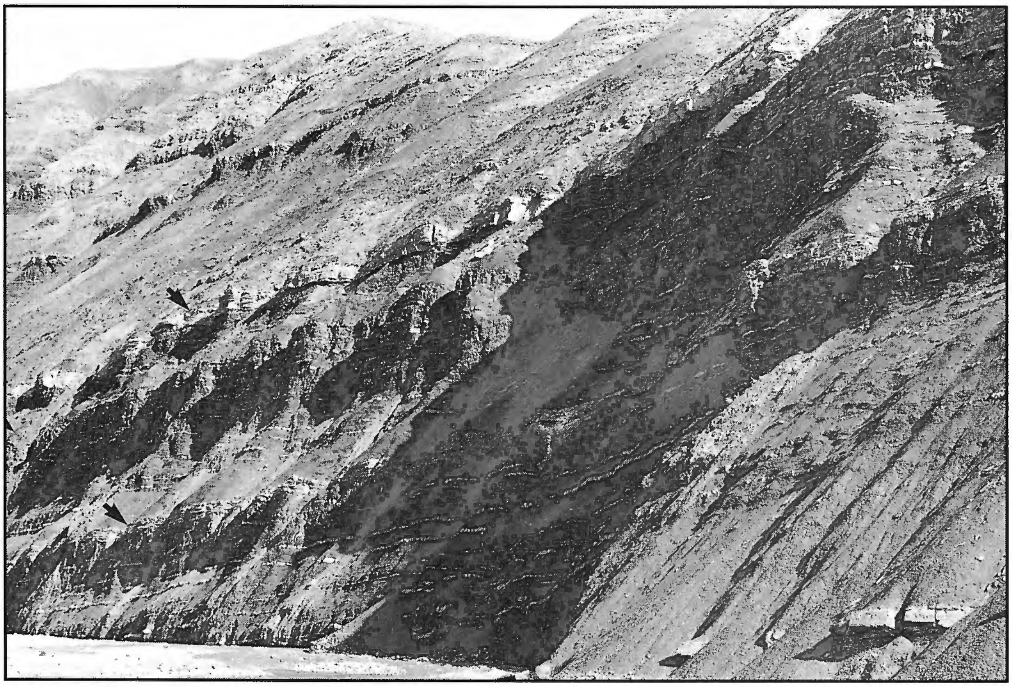

Figure 33. Coarsening-upward cycles in the muddy shelf facies (tops indicated by arrows), Strand Bay Formation, about 12 km southwest of South Bay (RAK 19-85; Fig. 2b). Contact with the Cape Pillsbury Member (Iceberg Bay Formation) at the ridge top is gradational. The main cliff section in shadow is 52 m thick. ISPG 2408-243

The formation in these areas is generally thicker than its western (basinward) counterparts – a minimum 276 m in RAK 19-85, and 350 m in RAK 6-87.

The distinguishing feature of this facies is the presence of at least twenty coarsening- and thickening-upward sandstone cycles (Fig. 32). Individual cycles reach 30-40 m in thickness. Interbedded laminated siltstone and mudstone grade up into thin bedded sandstone. Parallel lamination and ripple crossbedding is common. Hummocky and swaley crossbeds capped by ripples generally occur in the upper, sandier part of the cycles (Fig. 33).

Lower-order stratigraphic trends, on a scale of 70-100 m and including several of the small scale cycles, reflect progressive shoaling wherein the uppermost cycles in the trends are capped by clean, quartz sandstone beds that contain ripples and low-angle planar crossbeds. Some layers are thoroughly bioturbated by *Skolithos* and *Thalassinoides*. An additional feature of the shoaling packages is the presence of thin coaly shale beds which overlie a few cycles. Marine shale in the succeeding cycles abruptly overlies the coaly beds. Two and possibly three large scale cycles are present in section RAK 19-85.

Low-angle stratigraphic discordances between some coarsening-upward cycles are exposed in section RAK 6-87 (Fig. 34). Discordance angles are less than 8°. There are no indications of erosion or slumping at the contacts.

Individual coarsening-upward cycles in this facies represent successive episodes of sedimentation on a shallow, sandy part of the shelf, with frequent aggradation above storm wave base. Large scale cycles, up to 100 m thick, record longer term shoaling trends which culminated in shoreface and perhaps beach deposition. Thin beds of coaly shale overlying the shoreface/beach deposits indicate that coastal plain conditions may have been established locally. However, it is possible that the coaly beds are detrital; no root structures or soil profiles occur, and in this case the coal detritus may have accumulated through a period of sediment starvation during transgression. The angular discordant surfaces are interpreted as local downlap surfaces over which the shelf deposits prograded during relative sea level high stand.

The sandy component of this shelf facies represents the most landward extension of the Strand Bay Formation in the eastern part of Sverdrup Basin. The paleogeographic significance of this will be discussed below.

Summary of the Strand Bay Formation

Transgressing seas in mid-Paleocene time inundated both the major delta complex on west Axel Heiberg Island and the broad coastal plain that had developed along the eastern and southern margin of Sverdrup Basin. The stratigraphic indicator of maximum flooding is the abrupt sandstone/shale contact between the Expedition and Strand Bay formations. However the actual transgressive sandstone facies occurs at the top of the Expedition Formation.

Most of the Strand Bay Formation represents regional progradation. In the west this is manifested as prodelta shale (west Axel Heiberg Island). Elsewhere in Sverdrup Basin muddy shelf conditions prevailed. Two broad facies belts on the shelf are recognized. The deepest, outer shelf setting which was below the effects of storm wave base, is preserved in the Vesle Fiord-Bay Fiord area; this region is characterized by mudstone and is virtually devoid of sandy deposits. The Strand Bay shelf became shallower towards the east and north where sand deposited on shorefaces and perhaps in mid-shelf regions, was reworked by storms (muddy shelf facies, above storm wave-base).

The regional extent of Strand Bay deposition is not clear. A sandy coastal facies (Fig. 35), equivalent to the Strand Bay Formation but part of the Iceberg Bay Formation (see section below), defines the northern limit of the Strand Bay sea. Although shoaling trends also are recognized in the Cañon Fiord area the easternmost and southernmost limits of the

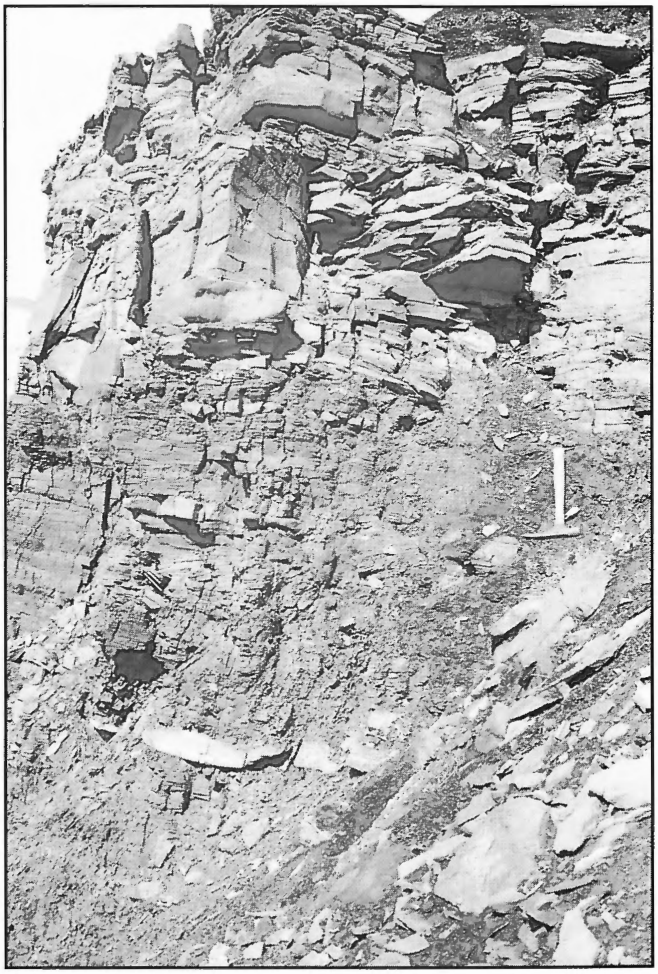

Figure 34. *Laminated, hummocky, and swaley bedded sandstone characterize the upper part of coarsening upward cycles in the Strand Bay Formation. Same locality as Figure 33. Hammer is 33 cm long. ISPG 2408-247*

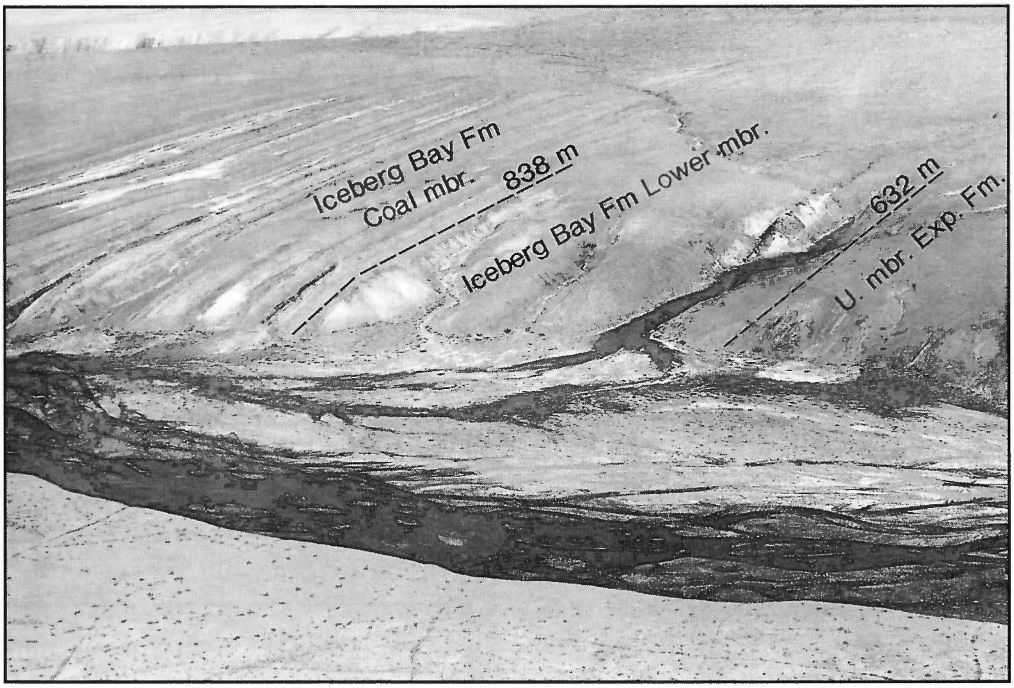

Figure 35. Panorama of the transition on north Fosheim Peninsula, from the Expedition Formation to the Lower member of the Iceberg Bay Formation, through a sandy coastal facies correlative with the Strand Bay shale, west flank of Fosheim Anticline; Section RAK 21-85 (Fig. 2b). ISPG 2408-310

shoreline are no longer preserved. The possibility that a tentative link was established between the Strand Bay sea and the proto-Baffin Bay has been suggested by Wall et al. (1988), based on the Atlantic affinities of elements of the foraminifera assemblage. There also may have been a link with the Beaufort Mackenzie Basin several hundred kilometres to the southwest.

ICEBERG BAY FORMATION

The Iceberg Bay Formation is the youngest stratigraphic unit in Sverdrup Basin and the penultimate unit in the Eureka Sound Group. Originally defined in the Strand Fiord area (Ricketts, 1986, 1991a) the formation has since been mapped throughout west-central Ellesmere Island, in outliers on northwestern Ellesmere Island (Hvitland Peninsula, Emma Fiord), and north of Lake Hazen. The type section on Kanguk Peninsula consists of interbedded sandstone, mudstone, and coal.

Few complete sections of the Iceberg Bay Formation exist because the stratigraphic top commonly coincides with the modern erosion surface. Four localities that do contain a preserved stratigraphic contact with the overlying Buchanan Lake Formation are (Fig. 36):

- Mokka Fiord (east Axel Heiberg Island), 1500 m recorded by Bustin (1977); see also RAK 15-85, Fig. 2b. However the base of the formation is covered here;
- Between the head of Whitsunday Bay and the Whitsunday Bay Diapir (Thorsteinsson, 1972a), approximately 200 m of interbedded, weakly indurated sandstone, mudstone, and coal that are correlated with the Coal member, occur east of the Stolz Thrust. The beds disconformably overlie Christopher Formation shale (the contact is poorly exposed), and in turn are overlain by diabase-rich conglomerate of the Buchanan Lake Formation;
- The south shore of Emma Fiord (northwest Ellesmere Island), where less than 100 m of Iceberg Bay Formation, overlain by Buchanan Lake conglomerate, are faulted against Silurian volcanics, and;
- Turnabout River near the northeast end of Lake Hazen, where Miall (1979) recorded up to 450 m of section (see also RAK 11-88, RAK 12-88, Fig. 2d); It is possible that some of the crossbedded sandstone included in Miall's map unit actually belongs to the Buchanan Lake Formation. At this locality the Iceberg Bay Formation is in angular unconformable contact with Lower Paleozoic rocks.

A substantial thickness of Iceberg Bay Formation occurs at the type section on Kanguk Peninsula – 1950 m, but here too the top is eroded (see RAK 25-83, Fig. 49, Ricketts, 1991a). More than 1200 m of section is exposed, albeit poorly, between Strathcona and Bay fiords (RAK 3-84, 5-84, 7-84, Fig. 2a), and at least 600 m of the lower part of the formation is well exposed along south Cañon Fiord (RAK 23-85, Fig. 2b).

Except on northern Fosheim Peninsula, the Iceberg Bay Formation conformably and gradationally overlies the Strand Bay Formation on Axel Heiberg and Ellesmere islands. On Fosheim Peninsula, between Remus Creek and Fosheim Anticline, up to 200 m of the basal Iceberg Bay Formation interfingers with the Strand Bay Formation (RAK 17-85, 21-85; Fig. 2b). At a few localities on east Axel Heiberg Island (Fig. 36; Depot Point syncline, Gibs Fiord, north Whitsunday Bay, north Mokka Fiord and the west side of Flat Sound), strata that are correlated with Iceberg Bay rocks are in disconformable contact with middle Cretaceous and older rocks of Sverdrup Basin (see also Balkwill et al., 1975). This correlation is based on lithology and palynology (see Appendix 1). These disconformities appear to be of local significance and are spatially associated with anhydrite/gypsum diapirs. Ricketts (1987a) has hypothesized that the local disconformities were a product of movement on the diapirs prior to Iceberg Bay deposition, such that up to 3000 m of pre-Iceberg Bay stratigraphy was removed.

Profound unconformity also exists between the Iceberg Bay Formation and Lower Paleozoic Franklinian rocks north of Lake Hazen (RAK 11-88, 12-88; also Miall, 1979). Several other northern outliers with only a few tens of metres of Iceberg Bay strata are faulted against Upper and Lower Paleozoic rocks, for example Hvitland Peninsula, south Emma Fiord, and Phillips Inlet (Fig. 36).

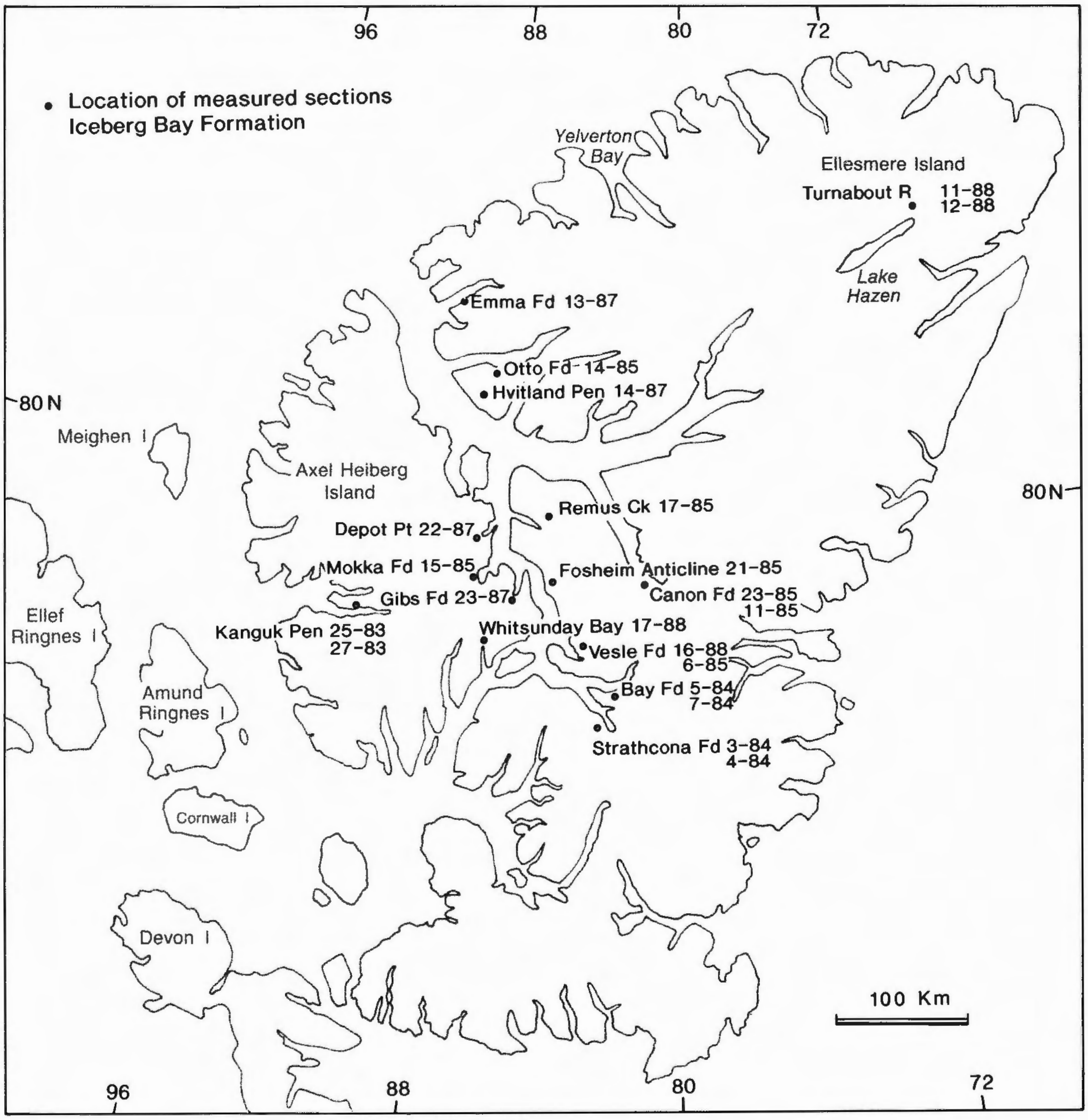

Figure 36. Map of measured section localities for the Iceberg Bay Formation. For summary sections see Figures 2a-d.

Two members have been defined at Strand Fiord (Ricketts, 1991a):

- the Lower member consists of sandstone and mudstone and minor coal and is arranged into numerous coarsening-upward cycles, and
- the overlying Coal member consisting of fining-upward sandstone-mudstone-coal cycles. The Coal member, 1060 m thick at Strand Fiord, has been mapped over most of Sverdrup Basin, from Strathcona Fiord in the south (at least 600 m), and as far north as Emma Fiord and Phillips Inlet and northeast to Lake Hazen. It is the most widespread of all formations in the Eureka Sound Group.

Elsewhere, rocks that are stratigraphically equivalent to the Lower member at Strand Fiord have a more restricted distribution and are generally found only in the Fosheim Peninsula-Strathcona Fiord area. Sufficient differences in lithology also exist in these areas to warrant the definition of two new members: the Cape Pillsbury Member, and the Braskeruds Member (see Fig. 6 to 9).

Cape Pillsbury Member (new)

At least 600 m of calcareous, very fine grained sandstone, siltstone, and mudstone, with a smattering of coal and scattered lenses of coarse sandstone, form precipitous cliffs fringing the modern drainage basin that empties into South Bay (Fig. 37). The carbonate content, unique in the Eureka Sound Group, is locally high enough to justify the term 'silty limestone', the rocks commonly exhibit flaggy weathering. More than 50 fining-upward cycles occur at the type section. This distinctive facies is mapped west and south to Vesle Fiord, and between Strathcona Fiord and Bay Fiord where more than 500 m are poorly exposed (RAK 5-84, Fig. 2a). South of Strathcona Fiord only 200 m or so of the flaggy facies is present, overlain by an additional 165 m of shale (Braskeruds Member). The Cape Pillsbury Member thins toward Remus Creek in the north where equivalent strata (Lower member) consist of thinly bedded, light brown, noncalcareous, fine grained sandstone and mudstone, and minor coal (Fig. 38).

Contacts

Contact with the underlying Strand Bay Formation and overlying Coal member is gradational, in each case over a few metres. With the Strand Bay Formation the contact corresponds to a change from recessive to resistant beds. With the Coal member there is a substantial increase in coarser sandstone and thick coal beds. At Strathcona Fiord contact with the overlying Braskeruds Member shale is abrupt (see below).

Type section

More than 600 m of section (RAK 23-85, Fig. 2b) is found in steep cliffs on the south side of the entrance to South Bay. Co-ordinates: 79°38′N; 81°26′W. Grid reference: 49H, N8841000, E491200.

Name

Cape Pillsbury is the westernmost point on the peninsula between Strathcona and Bay fiords.

Content

Fining-upward cycles of flaggy weathering, calcareous, fine grained sandstone and siltstone are unique to the Eureka Sound Group. Other than small ripples, crossbedding is rare, in keeping with the consistently fine grain size. Parallel lamination predominates and soft sediment deformation is common. Coaly beds, are generally less than 10 cm thick, muddy, and associated with abundant root structures. Bivalves and gastropods are rare. Some trace fossils occur.

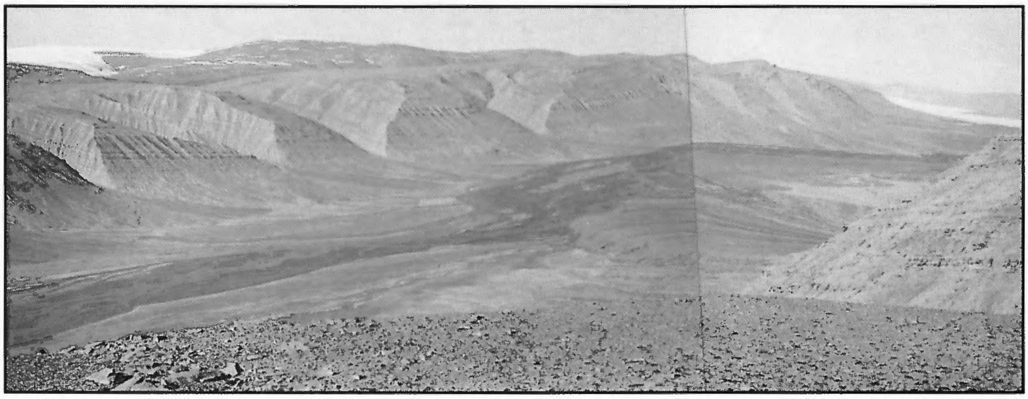

Figure 37. *Flaggy siltstone and mudstone sequences of the Cape Pillsbury Member, exposed in prominent cliffs up to 450 m high, adjacent an unnamed trunk river draining into South Bay (Cañon Fiord). The distance to Cañon Fiord is about 15 km. ISPG 2408-363; ISPG 2408-364*

Braskeruds Member (new)

Thick shale abruptly overlies fine grained calcareous sandstone and coal of the Cape Pillsbury Member along a ridge which intersects the south shore of Strathcona Fiord. In section RAK 4-84 (Fig. 2a), at least 165 m of shale are more or less continuously exposed; the stratigraphic top has been eroded. Stratigraphically equivalent shale has been traced to the north side of Strathcona Fiord (Fig. 3). However the unit appears to pinch out towards Bay Fiord.

Contacts

Contact between the Braskeruds shale and underlying Cape Pillsbury Member is abrupt and marked by the transition from relatively thick coal containing distinctive orange-weathering, permineralized tree stumps, to a green-grey silty sandstone 50-100 cm thick, and thence to grey shale (Fig. 39; also Ricketts, 1989b). The basal contact appears to be nonerosional. Nowhere is the upper contact exposed, but from stratigraphic and structural considerations the shale unit is presumed to underlie the Coal member.

Between Strathcona and Bay fiords the Braskeruds shale pinches out. In the same area a pebble conglomerate bed, first noted by West et al. (1981) occurs at the base of their Member IV, equivalent to the Coal member.

Type section

Best exposure is afforded by gullies incised into a north-trending ridge, 4.5 km due west of the head of Strathcona Fiord (Fig. 3; section RAK 4-84, Fig. 2a). Coordinates: 78°33.2′N; 82°35′W. Grid reference: 49E, N8722000, E468200.

Figure 38. Thinly bedded, locally calcareous buff sandstone, mudstone, and thin coal, exposed along Remus Creek (RAK 17-85; see Fig. 13) correspond to the northern pinch-out of the Cape Pillsbury Member. Jacob's staff is 1.5 m long (lower left). ISPG 3070-39

Name

Braskeruds Plain is a featureless region underlain primarily by Quaternary sediments, between Strathcona and Vendom fiords. The location derives its name after Ove Braskerud, a stoker who died on Otto Sverdrup's second polar expedition (1898-1902).

Content

Sandstone and silty sandstone with green-grey hues forms the basal few centimetres, grading into monotonous grey shale with less than 5% fine grained sandstone interbeds.

Age of the Iceberg Bay Formation

Palynological analysis of more than 80 samples from the Iceberg Bay Formation on Kanguk Peninsula indicates an age range from Late Paleocene to Middle Eocene (McIntyre, 1991a).

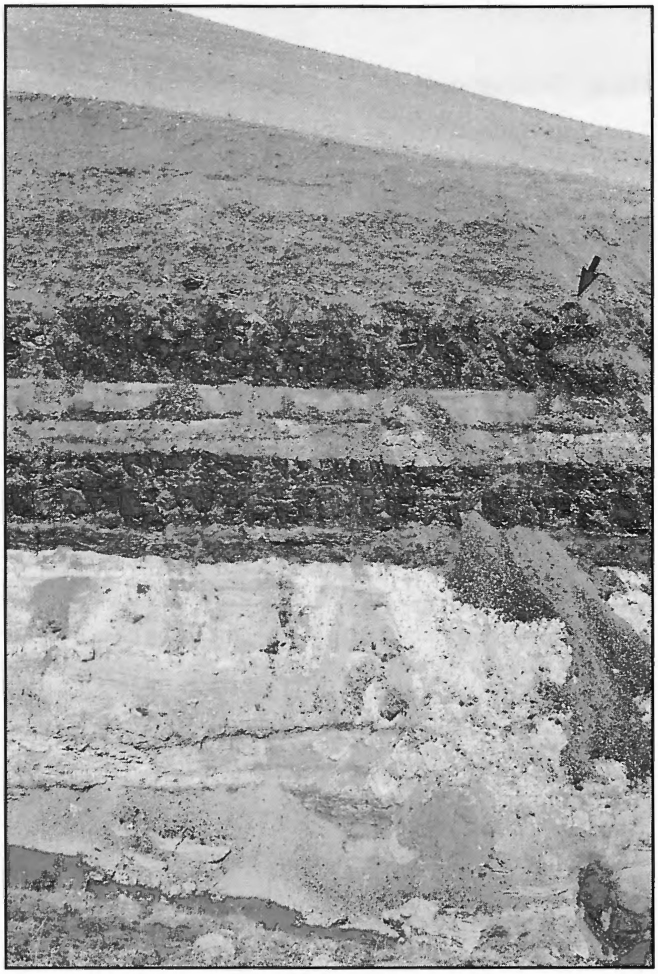

Figure 39. Abrupt contact between the Cape Pillsbury Member and Braskeruds Member (arrow; RAK 4-84, south Strathcona Fiord). Basal sandy beds in the Braskeruds Member above the contact have a greenish hue and are probably of transgressive origin. Coal seam below contact is 1.2 m thick. ISPG 2238-36

Lower member strata are restricted to the Upper Paleocene whereas the Coal member ranges into the Middle Eocene. The Paleocene/Eocene boundary occurs within the Lower member.

Age determinations of the Coal, Cape Pillsbury, and Braskeruds members on Ellesmere Island generally coincide with the Strand Fiord ages (Appendices 1 and 2). Pollen and microfossil assemblages indicate that the Cape Pillsbury Member is predominantly Late Paleocene. A diverse assemblage of calcareous foraminifera, unique to the Eureka Sound Group, has been discovered in lower units of the Cape Pillsbury Member (Wall et al., 1988). Characterized by *Melonis* and *Cibicides*, the assemblage is comparable to Late Paleocene or possible Eocene faunas having Atlantic affinities; this latter feature has particular significance with respect to paleogeographic reconstructions for the end of Paleocene time (see Stratigraphic sequences and paleogeography). A few foraminifera also occur in basal Lower member mudstone units at Fosheim Anticline (RAK 21-85, Fig. 2b; Appendix 2 – GSC sample number 134481), but they are not age-diagnostic.

Other fossil groups

Vertebrate and invertebrate fossils are important components of the paleontology in the Iceberg Bay Formation, particularly the Coal Member. Bivalves, small gastropods, and scaphopods of marine and brackish water origin (Dawson et al., 1975) are scattered throughout the Cape Pillsbury Member, but are absent in the Braskeruds shale, and scarce in the Coal Member; nonmarine bivalves occur in the latter. Dawson et al. (1975) also report crinoid stems from the Vesle Fiord area in probable Cape Pillsbury beds, which is interesting in view of the trace fossil *Scolicia* (discussed below) also found in nearby Cape Pillsbury strata (RAK 6-85, Fig. 2b).

Otoliths representing at least four families of marine fishes also have been found in probable Cape Pillsbury rocks on the south side of Strathcona Fiord (West et al., 1975). All osteichthyan families here were regarded as probably Eocene in age. Shark teeth are also present.

The most spectacular fossils discovered in the Coal member are vertebrates, of which a remarkably diverse assemblage indicates an Early Eocene age. The following elements are present (West et al., 1977; Estes and Hutchison, 1980):

- bony fishes (fresh water) including *Amia* and *Lepisosteus*;
- the large salamander *Piceoerpton*;
- reptiles, including abundant turtles and tortoises, lizards, boid snakes, and the Paleogene alligator *Allognathosuchus*;
- rare birds;
- a diverse assemblage of mammals with primates (Paromomyidae), rodents (Ischyromidae), carnivores (Hyaenodontidae, Miacidae), ungulates (Hyrachyidae, Brontotheriidae, Equidae), and representatives of the orders Pantodonta, Dermoptera, Taeniodonta, Proteutheria, and Multituberculata.

Facies assemblages-Lower member

Fluvial dominated delta-front facies assemblage

At least 42 coarsening-upward cycles make up a succession almost 900 m thick in the Iceberg Bay Lower member on Kanguk Peninsula, west Axel Heiberg Island. The diagnostic features and their interpretation have been detailed in Ricketts (1991a). The succession gradationally overlies prodelta shale of the Strand Bay Formation. Individual cycles, up to 45 m thick, differ considerably from deltaic units in the Expedition Formation in the preponderance of regular, tabular bedding and generally smaller bedforms, and a different trace fossil assemblage. Crossbedding includes only small tabular-planar varieties, small wave and current ripples, and interference ripples. Sole structures such as flute and groove casts are common. The trace fossil assemblage is dominated by *Planolites, Paleophycus*, and *Gyrochorte*, and only minor *Skolithos* (see Ricketts, 1991a, Table 3).

Lateral changes in facies organization exist locally, where thin fining-upward cycles contain abundant flaser and lenticular crossbedding, current and wave ripples, mud-chip lags, and coaly and carbonaceous mudstone containing root structures and tree stumps in growth position.

Sequential changes in internal organization of the coarsening-upward cycles occur towards the top of the succession. The proportion of sandstone increases to 60%, thin coaly mudstones cap many cycles, and low-angle planar crossbed sets become more common. Lateral, downslope (westward) changes also exist wherein cycle thicknesses decrease to an average of 3-4 m, and where *Chondrites* burrows predominate.

The content of the coarsening-upward cycles, and where outcrop permits, reconstruction of the overall facies geometries are fundamentally different to the wave-dominated delta facies in the Expedition Formation. Depositional energies were significantly lower, and there is no evidence for barrier islands, tidal inlets, or extensive strand plains. Nevertheless sedimentation must have been rapid because of the great thickness (900 m) of stacked cycles for which the overall stratigraphic variation is relatively minor; in other words, the relative position of the strand line changed little throughout the 900 m succession. The Lower member succession at Strand Fiord has been interpreted as a river-dominated delta front assemblage, containing a series of stacked, interdistributary bay and delta lobes. The paucity of major distributary channel sandstone bodies in the Strand Fiord rock record means that direct comparison with the classic Mississippi Delta analogue is not entirely suitable, although many aspects of the Mississippi interdistributary environments are appropriate. Nevertheless, a thick sandstone facies interpreted as distributary channel, does occur at the base of the Coal member, a stratigraphic position that marks the transition to delta plain. An additional analogy in the modern Tabasco lower delta plain has been invoked, with its smaller number of radially oriented distributary channels, although even here the level of wave influence at the Tabasco delta front is probably greater than that in the Lower member facies.

A succession of coarsening-upward mudstone to sandstone cycles, with minor coal, which is equivalent to the deeper water shale of the Strand Bay Formation, is discontinuously exposed on north Fosheim Peninsula (e.g. RAK 21-85, Fig. 2b). Coarsening trends in the cycles begin with thin interbeds of fine grained sandstone and mudstone, mostly laminated or rippled. Wavy, lenticular, and some flaser bedding are also common. Planar-tabular crossbeds up to 10 cm thick commonly have mud-draped reactivation surfaces. Sandstone at the top of the cycles is generally well sorted, and characterized by low-angle planar crossbeds. *Skolithos* burrows are common in these upper levels. The uppermost sandstone beds are capped by carbonaceous mudstone and coal beds up to a metre thick. Root structures extend into the sandstone below. Woody debris, Metasequoia, and hardwood leaves abound.

Each cycle in the north Fosheim Peninsula sandy facies represents a period of shoreface and beach progradation over muddy shelf deposits, and culminated in the development of coastal plain marsh or woodlands. Indications of storm influence in the facies are not preserved, but this may be a function of section location with respect to the paleogeographic position of wave base on the shelf.

Facies assemblages-Cape Pillsbury Member

Flaggy siltstone facies assemblage

Calcareous mudstone, siltstone, and subordinate very fine grained sandstone compose a facies that is unique to the Eureka Sound Group. It consists of many cycles, generally less than 5-6 m thick, which in the south Cañon Fiord area (type section) are stacked into a succession more than 600 m thick (Fig. 37). Some of the cycles exhibit fining-upward trends from fine grained sandstone to mudstone or coal, whereas in other cycles the apparent fining-upward weathering profile seems partly to be a function of an upward decrease in carbonate content (Fig. 40). Cycles are capped by thin (few centimetres) carbonaceous mudstone and shaly coal.

Mudstone and siltstone beds are thin, tabular, and as the facies name suggests, commonly display flaggy weathering. Weathered surfaces are cream to pale buff. An intriguing aspect of the beds is the preservation of remarkably delicate, parallel to slightly wavy millimetre-thick laminae and layers 1-2 cm thick (Fig. 41). Asymmetric current ripples and starved ripples are scarce. Large scale crossbedding is absent.

Laminated intervals are interrupted by beds exhibiting intense synsedimentary disruption (Fig. 41, 42); proportionally these beds account for 5-20% of cycle thickness. Disrupted beds contain poorly sorted, mud-chip conglomerate, convoluted laminae, small detached folds and pull-apart structures. Basal contacts of the beds are eroded and commonly contain load, flute, and groove casts (Fig. 43). The tops of disrupted beds are draped by undeformed laminae; ripples formed locally on the topographically rough-bed surface.

Synsedimentary microfaulting is also a characteristic feature of the laminated beds (Fig. 42). In many cases faulting is most intense immediately below disrupted beds, but also occurs elsewhere in the cycles apparently not associated with bed disruption. Such faulting imparts a 'wispy' texture to these rocks. Fault planes range through a few millimetres to 30 cm of strata, and terminate in continuous, nonfaulted layers. Both normal and reverse fault displacements occur.

Coaly beds that cap the calcareous mudstone-siltstone cycles are usually a few centimetres thick, rarely reaching 70 cm in thickness, and contain abundant muddy layers. Root structures abound, including impressions of vertically oriented, fluted reed stalks. On bedding planes, reed and broadleaf hardwood impressions are also common although fossil tree trunks were not found.

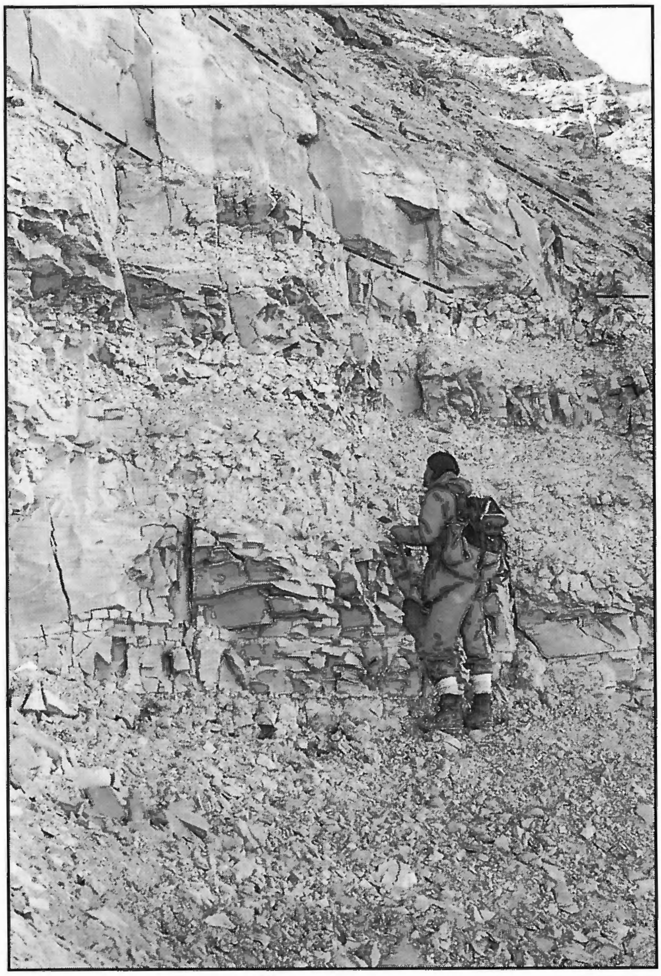

Figure 40. *Typical cycles of flaggy siltstone and fine grained sandstone, Cape Pillsbury Member, showing subtle fining upwards trends, capped by a coaly veneer and associated root structures. Cycles in this view are overlain and partly eroded by channel sandstone (dashed line). Section RAK 23-85, south Cañon Fiord. ISPG 2408-317*

Evidence of desiccation is present in the upper portion of the Cape Pillsbury succession. Desiccation affecting only one or two mudstone layers is uncommon. However, desiccation prisms up to 40 cm thick are so prevalent in some cycles that they completely disrupt primary layering (Fig 44). Polygon spacing is variable, ranging from 8-45 cm, and may in fact represent a bimodal distribution of mudcrack sizes. The margins of polygons are diffuse and locally contain an infill of mud-chips. A few polygon margins coincide with reed stems. Desiccated beds are overlain by nondeformed layers.

The flaggy siltstone facies contains a unique trace fossil assemblage that is best developed in the basal few cycles. Horizontal burrows such as *Planolites* (Fig. 45a) and *Thalassinoides* are the most common forms. *Ophiomorpha* is much less common. Two taxa not recognized anywhere else in the Eureka Sound Group include *Cosmorhaphe* (distinctly meandering, unbranched sand-filled trails – Fig. 45b), and *Scolicia* (unbranched, straight to sinuous, bilobed burrows with conspicuous meniscate structure and a central canal; Fig. 45c).

Sandstone channel facies assemblage

Within the flaggy siltstone-mudstone facies are medium- to coarse-grained sandstone bodies that truncate flaggy units and have distinct channel or lenticular geometry. Proportionally the sandstone channels make up 10-15% of the total Cape Pillsbury Member thickness. Channel widths range from 5 m to 80 m. Maximum observed thickness is 15 m, but most channels are less than 6 m at their thickest point. Basal contacts are eroded with up to 50 cm of local relief. Channel tops are either abrupt (Fig. 40) or gradational into overlying flaggy mudstone and siltstone, the latter type forming fining-upward units.

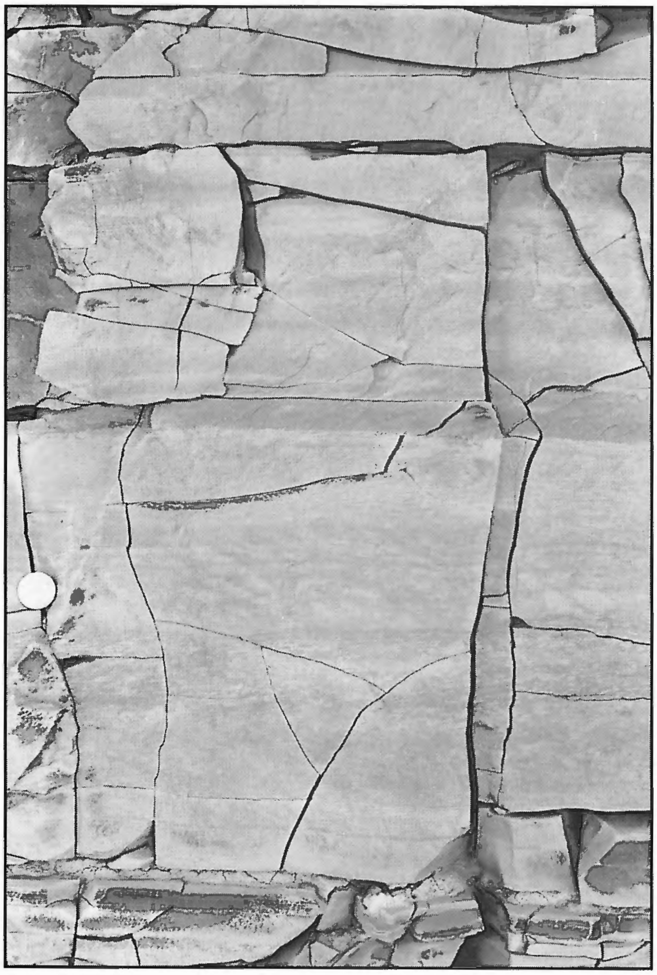

Figure 41. *Thinly bedded and delicately laminated, calcareous, fine grained sandstone, Cape Pillsbury Member (Section RAK 23-85). Note the bed of intraformational pebble conglomerate just below coin (24 mm diameter), and the absence of crossbedding. ISPG 2408-319*

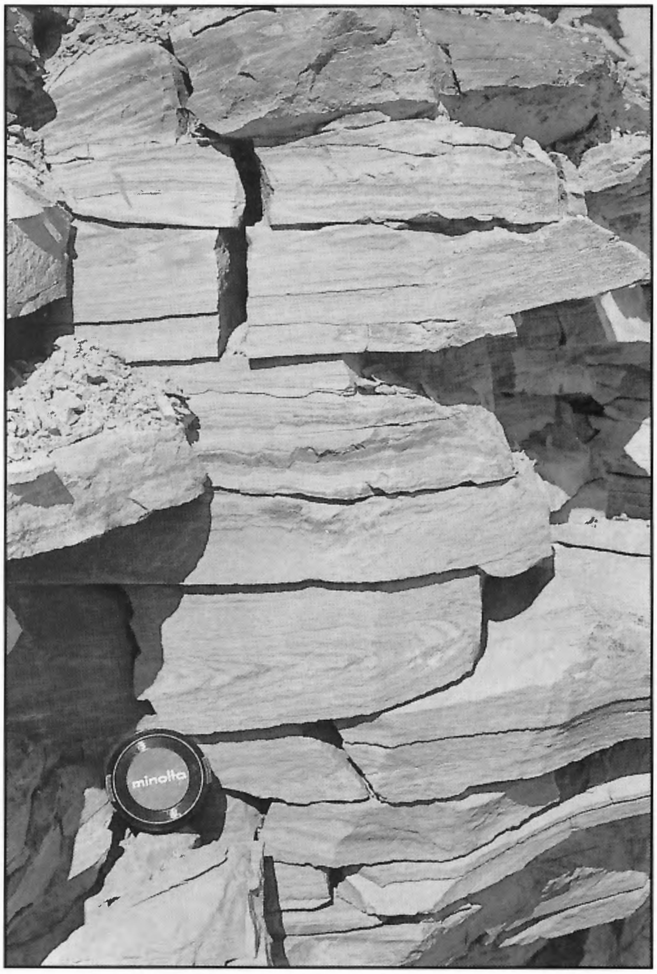

Figure 42. *Convoluted bedding, detached folds, and synsedimentary faults in laminated, calcareous siltstone and fine grained sandstone, Cape Pillsbury Member; section RAK 6-85, Vesle Fiord. ISPG 2408-35*

Common bedforms in the channels are trough and planar-tabular crossbeds. Basal lags contain mud-chips, plant fragments, comminuted bivalves, and gastropods. Some crossbed set contacts and foresets are veneered by mudstone. Herringbone crossbeds, ripples, and climbing ripples also are present. Crossbed foresets and mudstone veneers are commonly over-steepened or convoluted.

Shallow shelf facies assemblage

This assemblage, a variation of the flaggy facies, is best developed in sections west of South Bay (RAK 11-85; Fig. 2b) and in the north Vesle Fiord area. Here, flaggy facies cycles interfinger with coarsening- and bed thickening-upward sandstone cycles of 5-6 m thickness (Fig. 46a, b). Shale and siltstone (noncalcareous) low in the cycles, although poorly exposed, contain parallel lamination, current ripples and a few *Chondrites* and *Planolites*. What makes these cycles interesting is the frequency of hummocky cross-stratification in beds up to a metre thick, which are associated with laminated and low-angle planar crossbedded sandstone in the upper portions of cycles (Fig. 47). Individual hummocks range in wave length from 80-120 cm, and 2-4 cm amplitude. Beds containing the hummocks are capped by a veneer of current ripples. Larger scale planar-tabular crossbeds (sets 15 cm) are usually restricted to the top of cycles. A few *Skolithos* occur.

Figure 43. *Groove and flute casts at the base of fining upward cycles low in the Cape Pillsbury Member; section RAK 6-85, Vesle Fiord. ISPG 2408-12*

Figure 44. *Large desiccation polygons in fining upwards cycles near the top of the Cape Pillsbury Member; section RAK 23-85, south Cañon Fiord. ISPG 2835-353*

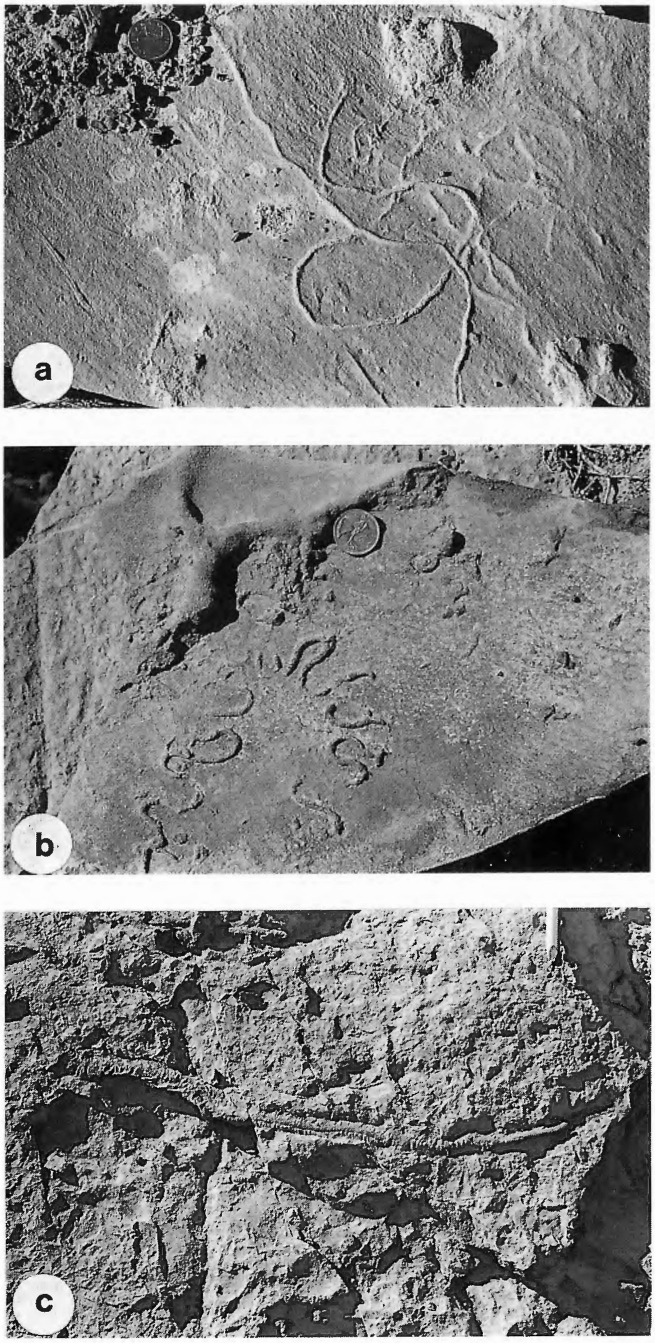

Figure 45. *Trace fossils in the basal few tens of metres of the Cape Pillsbury Member. Three distinctive taxa include: (**a**) Planolites, note short Gyrochorte trace on lower left; ISPG 2408-9 (**b**) Cosmorhaphe; ISPG 2408-16 (**c**) Scolicia. ISPG 2408-67*

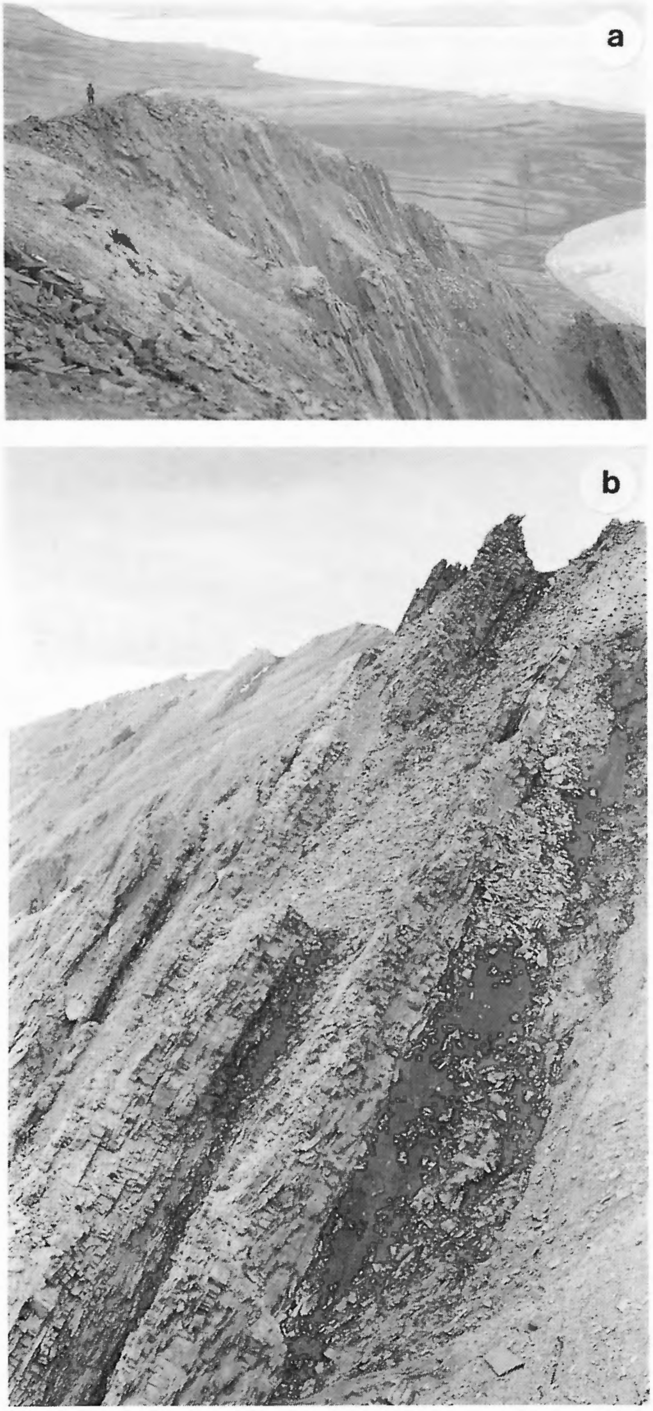

Figure 46. ***a)*** *Distinctive coarsening upward units of the shallow shelf facies, Cape Pillsbury Member, west side of South Bay (RAK 11-85). View is approximately north towards Cañon Fiord. ISPG 2408-109* ***b)*** *Coarsening upward sandstone units in the shallow shelf facies (Cape Pillsbury Member). The sandstone units are interlayered with the fining upward flaggy facies. Same locality as Figure 46a. ISPG 2408-112*

Evidence of subaerial exposure, abundant in the flaggy facies, is generally absent in the shallow shelf facies. Coaly beds with recognizable root structures are rare.

Summary, Cape Pillsbury Member

An intriguing lithological aspect of the Cape Pillsbury Member is the thickness of mud-dominated rocks of shallow water character, primarily in the flaggy facies. The following inferences are made:

- The fact that all flaggy facies cycles terminate with an autochthonous coaly bed indicates proximity to a strand line throughout the succession. This inference is augmented by the presence of desiccation polygons in the upper part of the member.
- The relative stability of the strandline through at least 600 m of strata, confined mostly to the Upper Paleocene, indicates a reasonable rate of sediment supply with respect to creation of accommodation space in the basin; notably, the same scenario is observed in the correlative Iceberg Bay Formation Lower member (stacked delta-front cycles) on west Axel Heiberg Island.
- Fully open marine conditions existed in at least the southern extent of the Cape Pillsbury Member as indicated by the calcareous foraminifera assemblage near Strathcona Fiord.

Until recently I have found it difficult to apply an appropriate modern analogue to the flaggy facies. Within the mud-dominated facies itself there is little indication of bedload or traction deposition, except for discrete pulses of high energy bed disruption and erosion. There is no evidence in the flaggy facies for tidal current asymmetry despite clear evidence of shallow marine conditions. However, high energy depositional conditions did occur on the shelf, as

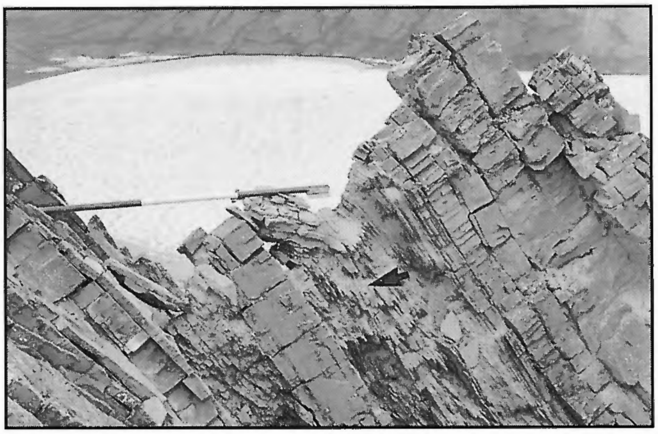

Figure 47. *Laminated and hummocky cross-stratified sandstone (arrow) typical of the shallow shelf facies, Cape Pillsbury Member, South Bay; section RAK 11-85. Jacob's staff is 1.5 m long. ISPG 2408-120*

evidenced by hummocky cross-stratification in the laterally associated coarsening- upward sandstone cycles. These cycles bear the hallmarks of shelf sedimentation which in section RAK 11-85 probably ranged from lower shoreface to storm wave base. It is tempting to link the evidence of storms on the shelf with the discrete, disrupted, mud intraclast beds in the flaggy facies.

Coarse grained lenticular sandstone bodies within the flaggy facies have obvious channel form. The size and distribution of the channels further indicates an intricate pattern of incision into the muddy facies, analogous to dendritic patterns developed on broad tidal flats, some classic examples being the North Sea coasts of Netherlands and Germany (Reineck and Singh, 1975). Additional modern analogues that are found on the French Guiana/Surinam and west South Korea coasts, in microtidal and macrotidal settings respectively, contain dendritic, braided, and anastomosing tidal and subtidal channels; both regions are characterized by high sedimentation rates (Froidefond et al., 1988; Wells et al., 1990). Some evidence of tidal current asymmetry (admittedly equivocal) is present in the Cape Pillsbury channel sandstone facies in the form of herringbone crossbeds and mud drapes on other bedforms. No accretionary point bar deposits or sandstone-mudstone couplets were observed.

Direct analogy of sedimentary structures characteristic of the flaggy facies with modern analogues is possible:

- The delicate nature of mudstone and siltstone lamination in the flaggy facies is very similar to that recorded in many intertidal and shallow subtidal flats (although it is by no means restricted to these environments).
- Desiccation polygons and prisms compare favourably to open fracture networks in mudflats bordering Severn Estuary (Allen, 1987). Fractures up to a metre deep form when prolonged exposure during warm weather coincides with low tides. Mud-chip lags are generated and these are available for reworking during higher energy events.
- Intertidal and subtidal mudflats on the Surinam coast are periodically subjected to flow and slide deformation that produces shear zones with intense microfaulting, extension cracks, convolution, and folding of muddy layers (Wells et al., 1980). All of the features, which in the modern analogue were observed to form during fair weather conditions, are present in the discrete, disrupted beds in the flaggy facies. Soft sediment deformation and shearing in the modern examples seems to be related to the build up of excess pore water pressure during tidal exchange. However, mud-chip lags and conglomerate that are commonly associated with this array of deformation structures may indicate that shearing can also form during storm wave activity.

Based on their stratigraphic disposition and internal organization, the Cape Pillsbury flaggy mudstone-siltstone cycles are interpreted as successive episodes of broad, mud-dominated intertidal and subtidal flat accretion and progradation (Fig. 48). A network of intertidal and subtidal channels dissected the mud flats. The tidal flats passed seaward into a shallow shelf of mixed sand and mud that was influenced by storms.

Sedimentation was dominated by suspension processes and perhaps low-energy bed load transport; the paucity of bedforms indicates that the threshold from no bed movement to ripples was rarely attained under normal conditions. High energy flow did occur however, perhaps during storm surges that also left their mark on the adjacent shelf. Soft sediment disruption of the laminated muds probably took place during the storms, but the formation of mudslides and associated structures during regular tidal events is not precluded.

The entire Cape Pillsbury succession has a shoaling-upward character. Flaggy mudrock cycles near the top of the member contain increasing evidence of desiccation and thin coal beds become more abundant. Cycles here record transitions to supratidal-marsh settings, vegetated primarily by reeds and some broad-leafed hardwood plants. The succession continues into coastal plain facies of the Coal member in the Cañon and Vesle fiord areas. At Strathcona Fiord, coal seams locally have cumulative thicknesses of 18 m (Ricketts, 1989b), and represent the more stable components of an encroaching coastal plain. However, deposition at Strathcona and Bay fiords was abruptly terminated by the Braskeruds Member shale.

Source of mud in the Cape Pillsbury Member

The predominance of mudrocks over a considerable thickness of the Cape Pillsbury succession (>600 m), many of them limy and even limestone, poses an interesting problem with respect to Late Paleocene sediment supply and the creation of accommodation space in Sverdrup Basin. Where did the mud come from, especially the carbonate component, and how was the mud entrained in such a thick, repetitive setting?

Petrographic examination of the fine grained facies shows that calcite grain size is consistent within specific layers – layers ranging from submillimetre to one centimetre in thickness. In the thinnest layers there is calcite as fine as 5-10μ m. Furthermore, calcite grain size within any layer is consistent, in terms of hydraulic equivalence, with the accompanying terrigenous component (quartz, feldspar, chert), although the carbonate to siliciclastic ratio is difficult to determine because of the fine grain size. There are no indications of advanced recrystallization discordant with primary depositional layering. Sandstone in the channel facies also contains 30-60% well rounded, sand-sized lime mud clasts in addition to siliciclastic types. About 5% of the framework includes single crystal grains that exhibit some abrasion. Incipient recrystallization of calcite does occur in the coarser lithologies but is confined mostly to contacts with quartz and feldspar grains.

Petrographic evidence indicates that most of the carbonate mud, silt, and sand components are detrital in origin. The sand sized mud clasts have crystal size ranges and fabrics identical to the autochthonous muddy layers in the flaggy facies and most likely are of intraclast origin. Given the overall humid and warm-temperate climatic conditions that have been deduced from the Late Paleocene and Eocene floral and faunal assemblages (West et al., 1977), it is unlikely that the carbonate mud and silt components originated either as in situ chemical precipitates, or as the products of biologically influenced precipitation, such as Codiacean algae. Fabric criteria indicate a detrital origin, a logical source being the Lower Paleozoic bedrock over which eastern Sverdrup Basin onlapped. Primary candidates for source rocks include several Devonian, Silurian, and Ordovician carbonate platforms now exposed in the Franklinian Fold Belt (Thorsteinsson, 1972b).

The terrigenous component of the flaggy and channel facies could also have been derived from the Lower Paleozoic rocks. An alternative source however, is the large river-dominated delta system on west Axel Heiberg Island. Considerable volumes of terrigenous mud are derived from large delta systems, for example the Mississippi and Amazon river deltas. Some of the mud is redistributed into adjacent coastal storage systems; huge volumes of mud derived from the Amazon are presently accumulating in the coastal zone from French Guiana to Surinam, and perhaps as far as the Orinoco River delta (Venezuela), several hundred kilometres from their source (Wells et al., 1980).

Deposition of mud on the Cape Pillsbury tidal flat/shelf took place under very low-energy conditions. This poses further problems in terms of Late Paleocene paleogeography. The eastern part of Sverdrup Basin must have been open enough to receive the sediment, (the calcareous foraminifera assemblage also indicates open marine conditions), and yet been sufficiently restricted such that fairweather wave and current influence were minimal.

Restricted conditions in a relatively active basin like Sverdrup Basin might occur if prevailing winds were offshore or if some physical submarine or land barrier were present. In the first case, constructing testable consequences is difficult and at least for the time being the idea of climate controlled basin restriction remains speculative. In the second case, there is some evidence for Eurekan tectonism beginning locally in the Early Eocene and possibly latest Paleocene (see Buchanan Lake Formation section) that may have influence the restricted conditions of eastern Sverdrup Basin. For example, incipient vertical uplift producing minor relief on ancestral Princess Margaret Arch could have provided a partial barrier to this part of the basin. However this hypothesis too is difficult to test because the onlap relationship of strata onto the putative arch would be very subtle and almost impossible to demonstrate. A third possibility is that the major delta complex on west Axel Heiberg Island was itself a major barrier that attenuated wave and current activity in the basin to the southeast.

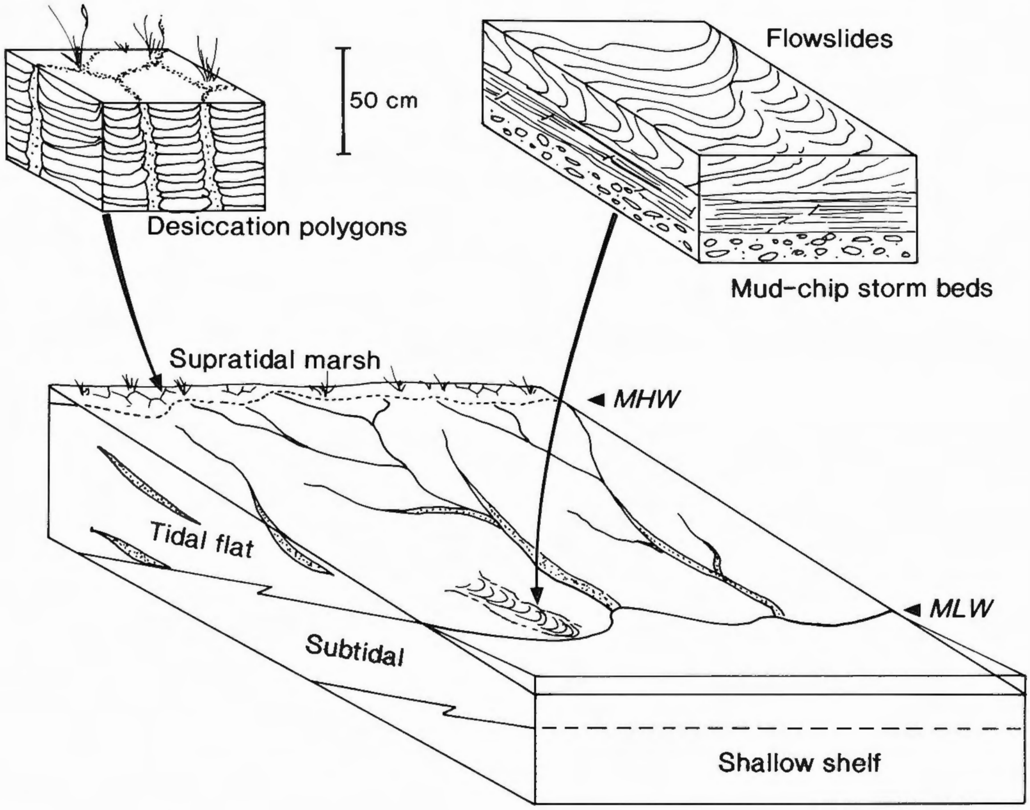

Figure 48. Schematic reconstruction of prograding tidal flats incised by sandy tidal channels, Cape Pillsbury Member facies. Insets show details of desiccation polygons in the upper intertidal/supratidal setting, and flow slides in the intertidal and possible subtidal setting.

Facies assemblage – Braskeruds Member

The Braskeruds Member can be adequately described by two facies: a shale facies, and a basal sandstone facies. Monotonous, medium grey, friable shale containing 5% or less interbedded sandstone makes up the bulk of the member. The thin (<5 cm) sandstone beds that are present are graded and have sharp bases, and resemble the Tb and Td divisions of Bouma units. Current ripples are rare.

The single variation on this theme is located at the base of the member, where grey-green, medium- to fine-grained, weakly laminated sandstone abruptly overlies coal seams of the Cape Pillsbury Member (Fig. 39).

Marine incursion into the Strathcona-Bay fiord area terminated tidal flat/coastal plain sedimentation of the Cape Pillsbury Member. Record of the transgression is seen in the basal grey-green sandstone at the base of the Braskeruds Member. The bulk of the shale accumulated in a prodelta setting as part of the succeeding progradational sequence. Although the upper contact of the Braskeruds Member is not exposed it is presumed to be gradational into coastal plain/delta plain strata of the Coal member.

Facies assemblages – Coal member

The Coal member, although the most widespread of any unit in Sverdrup Basin (Fig. 36), is also the least exposed. Over most of the basin the unit consists of interbedded sandstone, mudstone, and coal and it is the coal beds that tend to be the most resistant to weathering. Two of the most prominent facies assemblages were deposited in delta plain and coastal plain settings, although in many places the distinction between the two is not clear.

Distributary channel facies assemblage

At Strand Fiord, the distributary channel facies occurs at the contact between the Lower and Coal members (1920 m in RAK 25-83, details in Ricketts, 1991a). Trough and planar-tabular crossbedded sandstone, in units up to 4.5 m thick, are stacked into a bluff-forming succession 26 m thick. Eroded contacts contain basal lags of mud-chips and wood debris. The upper part of the facies fines upward into carbonaceous, laminated sandstone, that is capped by coaly mudstone.

The facies has been interpreted as a multistoried, delta distributary channel, that presumably was one of the feeders to the delta front deposits constituting the Iceberg Bay Formation Lower member on western Axel Heiberg Island (see previous discussion).

Delta plain-coastal plain facies assemblage

Throughout most of Sverdrup Basin, the Coal member is characterized by fining-upward cycles of moderately to poorly indurated sandstone, mudstone, and coal. Details of this facies at Strand Fiord (Ricketts, 1991a) apply to most areas of the basin. Fining-upward cycles, 3 to 10 m thick, contain features typical of fluvial channels: basal pebble lags, abundant trough crossbeds which commonly are oversteepened, bedforms that decrease in size upwards, and a gradual transition into coaly beds with root structures. Sandstone/shale ratios in each cycle range from 40 to 60%. Coal seams are commonly 1 to 2 m thick and locally 6 m thick (Ricketts and Embry, 1984). Other cycles pass abruptly from sandstone to shale or coal. Plant fragments abound and in situ tree stumps up to a metre in diameter are common.

Cycles low in the member are locally capped by thin, abrupt-based, calcareous, white-weathering sandstone that contains abundant ripple crossbedding, and *Planolites* and *Gyrochorte* burrows. In comparison, rusty weathering paleosols are present capping many cycles in the upper part of the member.

Only scattered coarsening-upward, or coarsening- then fining-upward cycles are present in the Coal member, the best example occurring on Kanguk Peninsula (RAK 25-83, Ricketts, 1991a). One 25 m thick unit near the top of the Coal member, consists of thinly bedded sandstone having low-angle planar and tabular-planar crossbedding. Bioturbation is common. The unit is capped by an interval of root structures and coal. In three dimensions, the beds are arranged into a single lateral accretion package.

As indicated above (Fig. 36), strata correlated with the Coal member are found in several outliers on Axel Heiberg and Ellesmere islands. Their palynology is described in Appendix 1.

Whitsunday Bay, Depot Point Syncline, Gibs Fiord, Mokka Fiord

Five hundred seventeen metres of the Coal member are recorded in one section at Mokka Fiord (RAK 15-85, Fig. 2b); Bustin (1977) reports 1500 m elsewhere in the area. At the other three localities only 100-200 m are preserved overlying the Christopher Formation. All localities are characterized by lithological associations similar to fining-upward cycles described elsewhere in the Coal member. Permineralized wood fragments are common and some tree stumps are in growth position. Vertebrae of fresh water fish have been found at scattered localities near Mokka Fiord.

The floral assemblages from Gibs Fiord and Depot Point, although sparse, indicate an Early Eocene age in the former, and (questionably) Paleocene in the latter (Appendix 1). The Mokka Fiord assemblages are Early to Middle Eocene (see Ricketts and McIntyre, 1986).

Southwest Hvitland Peninsula

An outlier 6 km north of Otto Fiord contains about 100 m of interbedded white sandstone, mudstone, and coal (Fig. 49). It is fault-bounded along its eastern and western margins, and in apparent stratigraphic contact with Lower Triassic strata towards the south (Thorsteinsson and Trettin, 1972a). Concretionary ironstone bands are common and contain excellent impressions of *Metasequoia* and broad leafed plants. The palynology indicates a probable Late Paleocene age (Appendix 1).

Figure 49. *Interbedded sandstone, mudstone, ironstone, and coal of the Coal member, Iceberg Bay Formation, in a fault-bounded outlier on southwestern Hvitland Peninsula. The coal seam at bottom-centre is about 1.5 m thick (arrow); section RAK 14-87. ISPG 2835-247*

South shore, Emma Fiord

About 15 m of slumped white sandstone and coaly beds is exposed on the southern flank of a prominent ridge, the ridge itself being underlain by Buchanan Lake Formation (Fig. 50). Tree trunks up to 30 cm diameter are common. Ironstone nodules are infused with plant fragments. The floral assemblage indicates a Late Paleocene to Early Eocene age.

South of Phillips Inlet

Thin interbedded, dark grey carbonaceous shale, thin coal beds, and buff-weathering sandstone are exposed in a tributary about 1 km east of the trunk river that drains into the middle arm of Phillips Inlet (and 16 km south-southwest of the Inlet; see National Air Photo Library A16605-81). The beds are deformed into tight folds and penetrative cleavage. Permineralized tree trunks abound. Some of the sandstone beds are distinctly feldspathic, presumably derived from volcanic flows nearby. Alkali basalts mapped in the area have an ^{40}Ar/^{39}Ar age of 65 ± 3 Ma and a basaltic dyke has an age of 59 ± 1 Ma (Muecke et al., 1990), both of which could have supplied volcanic debris to the Coal member sediments. A palynological age of Late Paleocene confirms the older of these volcanic groups as the most likely source.

North of Lake Hazen

Excellent exposures of Eureka Sound Group beds (Coal member, Iceberg Bay Formation) are located in the Turnabout River area, about 13 km northeast of Lake Hazen. Although

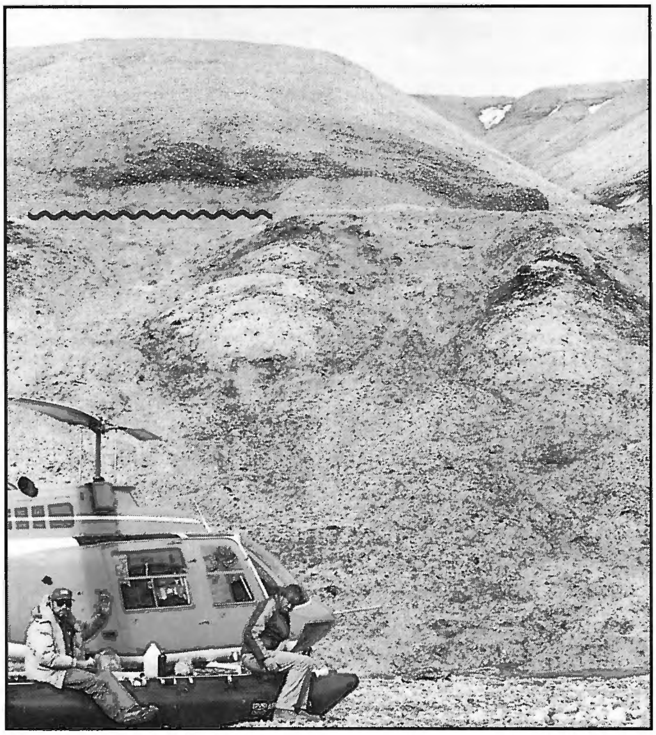

Figure 50. *The stratigraphic contact (corrugated line) between coal-bearing strata of the Iceberg Bay Formation and the conglomeratic Buchanan Lake Formation, near the south shore of Emma Fiord (section RAK 13-87). ISPG 2835-223*

the stratigraphy here is typical of the Coal member, there are abrupt local changes because of relief on the sub-Eureka Sound Group unconformity (below which are deformed Lower Paleozoic rocks). In places clear onlap of the unconformity can be demonstrated (Fig. 51); for example compare sections RAK 11-88 and 12-88 (Fig. 2d). Palynological analyses indicate an Early or possible Middle Eocene age, and McIntyre (Appendix 1) suggests that the flora is comparable to the upper part of the Iceberg Bay succession at Strand Fiord (e.g. RAK 25-83, Ricketts, 1991a). There are blocky weathering coal seams up to 1.7 m thick.

Summary, Coal member

The fining upward cycles and the bedforms contained therein are typical of those developed in high sinuosity fluvial environments. Proportionally high mudstone/coal components are consistent with channels being separated by broad floodplains. Substantial coal deposits accumulated in stable floodplains and swamps that supported fully developed *Metasequoia* and hardwood forests and woodlands. Wooded lands in eastern Sverdrup Basin were a haven to various reptiles and mammals; the rivers were stocked with fish. For the most part, the lithofacies and biofacies are representatives of upper delta plain and the landward portions of coastal plains. The Late Paleocene to about Middle Eocene period was one where the vegetated regions were the most extensive, reaching from western Axel Heiberg to Bay Fiord and south, to Emma Fiord and north, and extending northeast of Lake Hazen. Western regions of the delta plain were periodically inundated by the sea, as evidenced by the white, bioturbated (elevated salinities) sandstone beds capping many cycles at Strand Fiord. More substantial marine flooding resulted in occasional coarsening- then fining-upward cycles. These transgressions were probably of local extent, perhaps related to abandonment and reworking of subdelta lobes.

Downslope equivalent delta-front or shelf deposits of the Coal member are no longer preserved onshore in the Arctic Archipelago; presumably they underlie the modern polar continental shelf.

BUCHANAN LAKE FORMATION

Strata belonging to the Buchanan Lake Formation, the youngest unit in the Eureka Sound Group, record fundamental changes in the tectonostratigraphic organization of Sverdrup Basin. Deposits in the Geodetic Hills area of east Axel Heiberg Island have been examined most in detail and their relationship to predeformation Sverdrup Basin, and to major Eurekan structural elements (e.g., Stolz Thrust) has been demonstrated (Ricketts and McIntyre, 1986; Ricketts, 1987a, 1991b). From these studies it is clear that the conglomeratic Buchanan Lake deposits are a direct consequence of regional deformation and fragmentation of Sverdrup Basin; deformation that also involved the Lower Paleozoic Franklinian

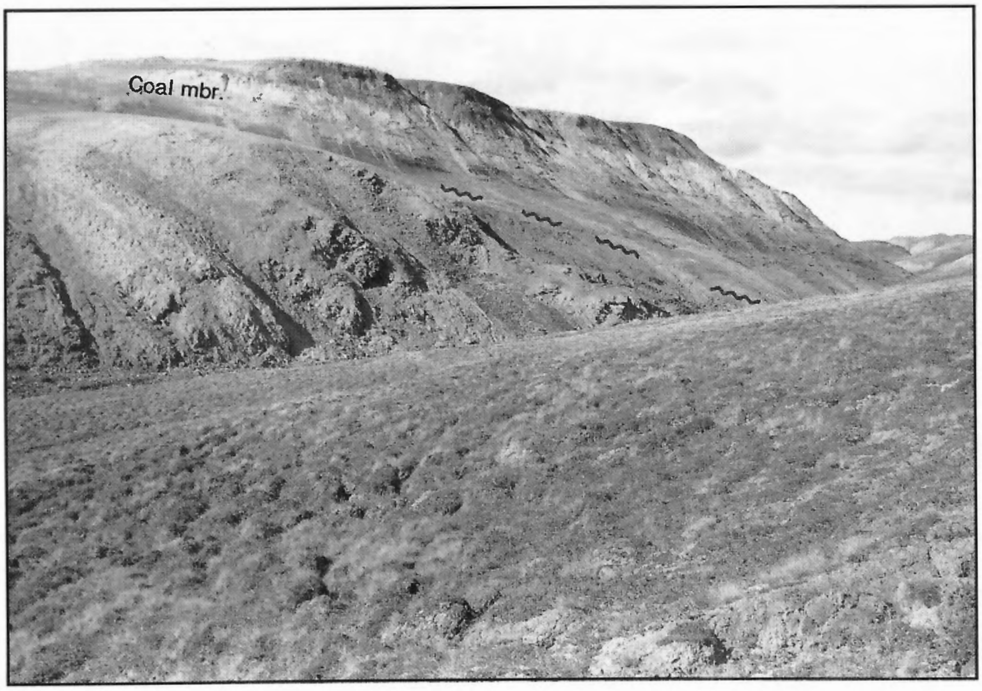

Figure 51. Coal member beds onlap topographic highs of deformed Lower Paleozoic rocks, above the sub-Iceberg Bay Formation unconformity, northeast of Lake Hazen; sections RAK 11-88, 12-88. ISPG 3070-47

Mobile Belt. Furthermore, based on lithofacies associations and distinct differences in clast composition from one area of outcrop to another, Buchanan Lake deposition is seen to have taken place in at least six (perhaps more) separate basins or subbasins having intermontane character (Fig. 52):

Axel Heiberg Basin
Otto Fiord subbasin
Emma Fiord subbasin
Franklin Pierce Basin
Lake Hazen Basin
Judge Daly subbasin

The name Lake Hazen intermontane basin also was used by Miall (1979) but his definition included rocks assigned to the Iceberg Bay Formation and therefore part of Sverdrup Basin. Lake Hazen Basin herein refers only to the synorogenic Buchanan Lake rocks dated as Early and possibly Middle Eocene (Appendix 1). Likewise, Judge Daly basin was originally defined by Miall (1982) to include a number of map units, the youngest of which is a thick boulder conglomerate. As discussed below, clast compositions are similar to those encountered in the Lake Hazen Basin and it is not clear whether the two basins were originally connected, or were separated by faults with similar Lower Paleozoic rocks in their hanging walls.

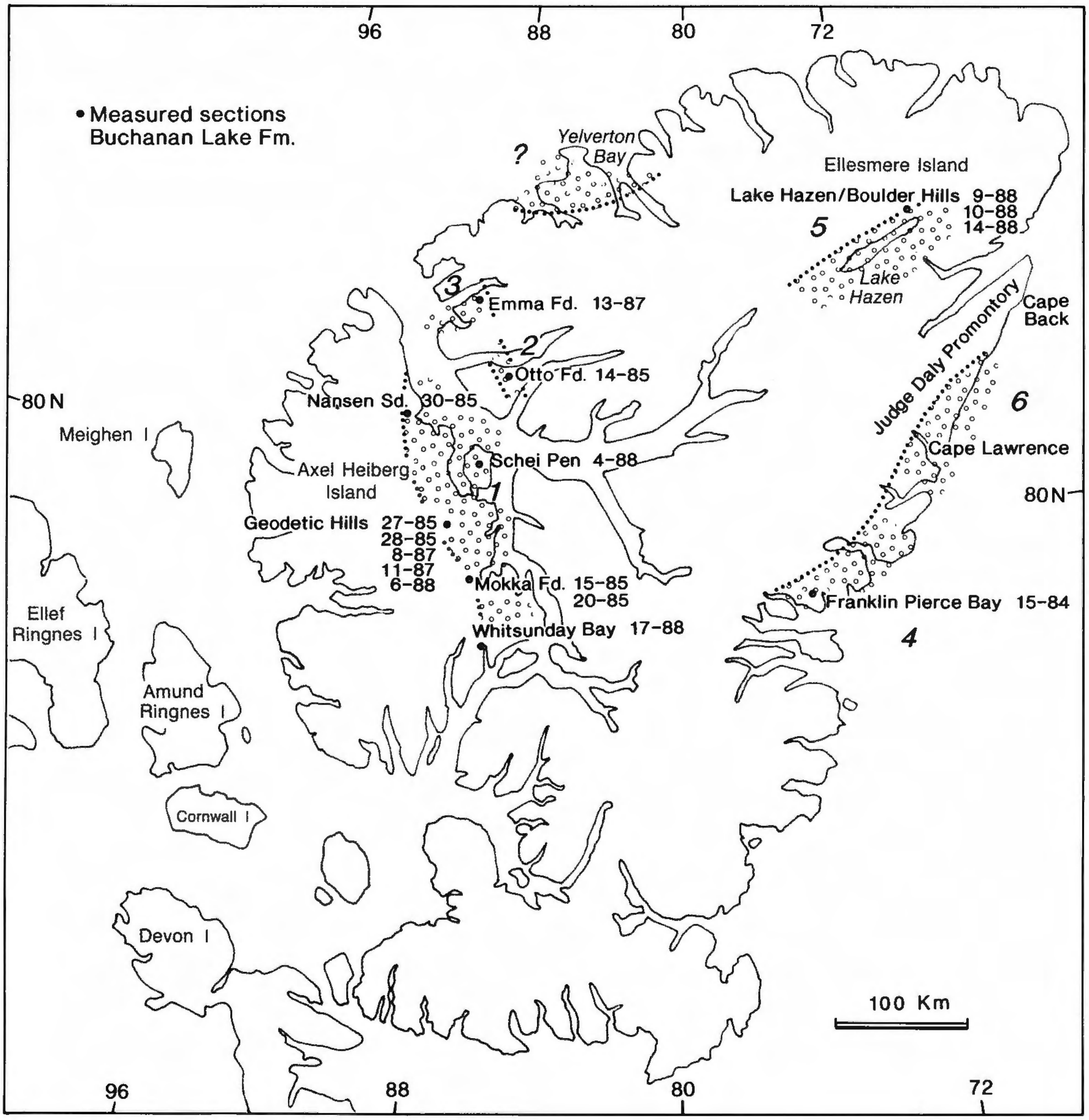

Figure 52. Location map of basins (open circles), measured sections, and other important localities for the Buchanan Lake Formation. 1 = Axel Heiberg Basin, 2 = Otto Fiord subbasin, 3 = Emma Fiord subbasin, 4 = Franklin Pierce Basin, 5 = Lake Hazen Basin, 6 = Judge Daly subbasin. A possible seventh basin is indicated at Yelverton Bay.

Strata previously identified as Beaufort Formation at Yelverton Inlet (Wilson, 1976) have so far not produced a conclusive palynological age. They too may belong to the Buchanan Lake Formation and constitute a possible seventh basin.

An unroofing stratigraphy defined by variations in clast composition is recognized in Axel Heiberg Basin and Otto Fiord subbasin (Table 3).

General characteristics

Representatives of the Buchanan Lake Formation are widely scattered in the eastern Arctic Islands. Originally defined along the south shore of Mokka Fiord (Ricketts, 1986), the formation is characterized by conglomerate, coarse lithic sandstone and minor coal. Buchanan Lake, for which the formation is named, spills into Mokka Fiord. At the type section, clast composition is predominantly diabase with variable amounts of indurated quartz sandstone and shale; compositions which are distinctly different from older mature and supermature Eureka Sound Group deposits. Clast compositions vary significantly between basins, but in each basin the compositional character can be related directly to source rock compositions in the structurally juxtaposed strata.

Contact between the Buchanan Lake and Iceberg Bay formations is exposed at Mokka Fiord (RAK 15-85, Fig. 2b); between Boulder Hills and Turnabout River north of Lake Hazen (see also Miall, 1979); Emma Fiord (Fig. 50); and discontinuously at Whitsunday Bay (RAK 17-88, Fig. 2d).

At only one locality, east of Geodetic Hills, is there a possible upper stratigraphic contact with peat and gravel beds that may be similar in age to the Beaufort Formation on Prince Patrick Island (J. Fyles, pers. comm., 1991; see RAK 28-85, Fig. 2b; also Ricketts, 1991b). However, Fyles (1990) has suggested that the name 'Beaufort Formation' be restricted to western Arctic Island occurrences.

The thickest succession of Buchanan Lake Formation is preserved in Axel Heiberg Basin, in a graben west of Stolz Thrust (possibly up to 1000 m thick according to Bustin, 1982), and in a composite section 850 m thick east of Stolz Thrust (Ricketts, 1991b). Sections elsewhere on Axel Heiberg Island include: Mokka Fiord (type section 370 m); Whitsunday Bay (350 m) and Nansen Sound north of Stang Bay (300 m, Ricketts and McIntyre, 1986). About 300 m occur at both Emma Fiord and Franklin Pierce Bay, and at least 350 m at Otto Fiord. The composite thickness north of Lake Hazen is 500 m (RAK 14-88, Fig. 2d).

Each of the Buchanan Lake basins listed above will be discussed in terms of their stratigraphy, age, lithofacies (summarized in Table 4), provenance, and association with major tectonic structures.

Axel Heiberg Basin

Axel Heiberg Basin is the most extensively exposed and best preserved of all the Buchanan Lake basins. Covering a wide expanse of east Axel Heiberg Island, the basin extends from Nansen Sound in the north to Whitsunday Bay – a strike

Table 3. Sedimentary Facies in the Buchanan Lake Formation basins.

	Basins and subbasins containing Buchanan Lake Formation				
Facies	**Axel Heiberg**	**Otto Fiord**	**Emma Fiord**	**Franklin Pierce (Judge Daly)**	**Lake Hazen**
Matrix-supported conglomerate	X			X	
Thick conglomerate	X	X	X	X	X
Conglomerate-sandstone	X	X	X	X	X
Thick sandstone	X				X
Conglomerate-mudstone (coal)	X		X		X
Sandstone-coal (Paleosol)	X	X			X
Siltstone-coal (Paleosol)	X				
ADDITIONAL STRATIGRAPHIC FEATURES					
Basal Formation contact exposed	X		X		X
Red beds					X
Unroofing stratigraphy	X	X			

length of at least 200 km, and a preserved width of 70 km or more from Geodetic Hills to Schei Peninsula. Buchanan Lake strata outcrop in three principal areas, the largest being an expanse of dissected alluvial plain on east-central Axel Heiberg Island (Fig. 52). Conglomerate, sandstone, and coal beds that include the well known 'fossil forests', have been studied in considerable detail between Geodetic Hills and Flat Sound (Ricketts, 1991b, and other papers in Christie and McMillan, 1991).

Table 4. Facies associations and principal bedforms, Buchanan Lake Formation.

Facies Associations	Principal Bedforms
Matrix-supported Conglomerate	- very poor sorting, mud matrix - beds up to 10 m thick - clasts up to 80 cm across - lacking internal layering - generally nongraded
Thick Conglomerate	- tabular bedded units, 3-10 m - sandstone interbeds <25% - clasts to 45 cm, avg. 8-10 cm - crude stratification - large wedge sets - clast imbrication, wood - rare crossbedding
Conglomerate-Sandstone	- tabular bedded units - clasts to 50 cm, avg. 10 cm - horizontal stratification - planar-tabular sets to 3.5 m - channel fill - clast imbrication, wood - sandstone wedges, crossbeds - rare peaty paleosols
Thick Sandstone	- stacked planar-tabular and trough crossbedding - ripples - scour and fill - rare podzols, peaty paleosols
Conglomerate-Mudstone (coal)	- fining upwards units - basal pebble lag or conglomerate bed - planar-tabular and trough crossbedding - thin woody lignite
Sandstone-Coal	- fining upwards units - channel fill - point bar foresets - mostly trough crossbeds - lignite seams to 1.5 m - leaves, cones, seeds - tree stumps, logs - red-brown podzols
Siltstone-Coal (paleosol)	- fining upwards units - prominent lignite seams - laminated, rippled - fossil forests - leaves, cones, seeds - phytolitic beds, gleysols

Geodetic Hills/fossil forest area

The Buchanan Lake succession occurs in the hanging wall and footwall of the major reverse fault system – the Stolz Thrust. Four informal members have been mapped in the area (Fig. 53; for maps see Fig. 2 in Ricketts, 1991b). Massive cliff-forming diabase conglomerate making up the conglomerate member, and more than 1000 m thick, is confined to the hanging wall of Stolz Thrust. Stratigraphic contact with the Lower Cretaceous, coal-bearing Isachsen Formation is exposed in two places, about 13 km northeast of Geodetic Hills, adjacent to a major ice lobe. The unconformity is probably subaerial.

Interbedded fine grained sandstone, siltstone, and coal comprise the lower coal member. A sliver of this unit interfingers with the conglomerate member about 1.5 km west of Stolz Thrust, but the most extensive exposure occurs in the footwall of the thrust. The lower coal member also disconformably overlies the Isachsen Formation. Best exposure is located along the right bank of an incised river immediately adjacent the ice-cap lobe. About 350 m total thickness is estimated for this member.

The conglomerate-sandstone member is characterized by a cyclic repetition of these two lithologies. Contact with the underlying lower coal member is poorly exposed but appears to be conformable and abrupt, whereas that with the overlying upper coal member is gradational. The dissected plateau between Stolz Thrust and the fossil forest site contains the best exposures of this member, but in Axel Heiberg Basin it is seen as far afield as north Hansen Sound.

Fossil forest-bearing coal beds and associated fine grained sandstone, siltstone, and mudstone, comprise the upper coal member. This, the youngest member in the Buchanan Lake Formation (of Axel Heiberg Basin) is at least 200 m thick, and is best represented in sections 25 km northeast of Geodetic Hills.

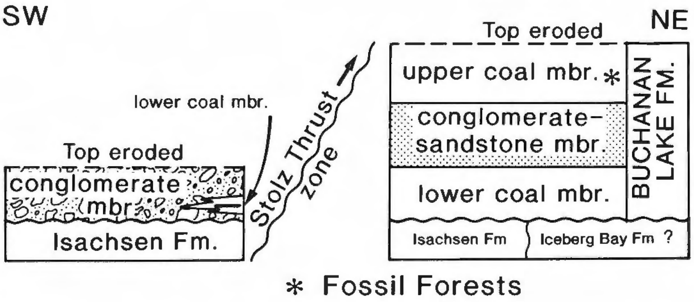

Figure 53. Summary of Buchanan Lake Formation lithostratigraphy in Axel Heiberg Basin. Modified from Ricketts (1991b, Fig. 3).

Lithological and facies characteristics of the members are summarized in Table 4. In a general way, the most proximal facies are positioned west of Stolz Thrust, and more distal facies east of the fault. The most distal of these units extends as far east as Schei Peninsula where fine grained sandstone, mudstone, and coal overlie Lower Triassic rocks with angular unconformity.

Paleoenvironmental domains that have been teased from the facies assemblages include the following (for details see Ricketts, 1991b):

1. Middle and outer alluvial fan debris-flow deposits are recorded in the matrix supported conglomerate facies at the western limits of exposure. Debris flows were capable of transporting clasts as large as 80 cm across. Further east, towards the Stolz Thrust, thick debris flow units (up to 10 m) are increasingly intercalated with thick conglomerate facies that are characterized by crude tabular layering and imbrication. This latter facies indicates the increasing influence of bed load transport in stream flood conditions, mostly as longitudinal bars. There is a distinct paucity of transverse bedforms. This trend marks an important transition from the leading edge of the alluvial fan (*sensu stricto*), interfingering with the most proximal braided river or braidplain realm. Notably, the extensive conifer and hardwood forests that developed on more distal floodplains attest to the humid conditions in which the debris-flow alluvial fans developed.

2. Braidplains, which according to Rust (1984) merge upslope with alluvial fans, are represented by the conglomerate-sandstone facies. Unlike the thick conglomerate association, crossbeds are common, crudely stratified tabular beds being subordinate. Clearly, transverse flow conditions prevailed with construction of characteristic gravel braid-bars during stream floods. Many lenticular sandstone bodies developed during waning flood stages, some as bar-edge wedges.

3. The transition from braidplain to meanderplain is represented by two facies: thick sandstone, containing abundant trough and planar crossbedding, is interpreted as the product of sandy braided stream deposition, or possibly stacked anastomosing streams; and a fining-upward conglomerate-mudstone (coal) association that accumulated in mixed load (Schumm, 1981), gravelly meandering stream channels.

4. Definitive high sinuosity stream and associated deposits are found in the fining-upward sandstone-coal and siltstone-coal facies. Meander channel-fill and point bars indicate migrating thalwegs and accretion of adjacent floodplains, although some channels were abruptly abandoned. Individual channel widths were many tens of metres, and meander belts were 5 km and more wide. Paleoslopes were at their lowest in these settings. Extensive floodplain and crevasse splay environments are represented by the siltstone-coal association. Mixed conifer and hardwood forest cover is preserved in various stages of development, from newly vegetated floodplains to full forest conditions – for example the Geodetic Hills fossil forests, preserved in at least twenty separate stratigraphic levels. Paleosols are common and have been described in detail by Tarnocai and Smith (1991). Red-brown and grey-brown podzols and gleyed podzols up to 1.4 m thick formed in well drained, near-channel settings. The leached, lighter coloured horizon, typical of many podzols, is thin in most Geodetic Hills paleosols. Organic layers are also thin and plant material is permineralized. Gleysols a little more than a metre thick, formed in loamy sandstone in the poorer drained, more distal areas; they have organic-rich upper horizons.

Paleocurrent indicators in the braided, meandering stream and floodplain facies (mostly azimuths of trough and planar-tabular crossbeds) indicate a consistent shift in sediment transport from eastwards in the proximal facies, to southeast-directed in the more distal facies (Ricketts, 1991b, Fig. 6). Bustin (1982) has also recorded southeasterly directed sediment transport in the fossil forest area.

Two principal clast types occur in the Buchanan Lake Formation on west Axel Heiberg Island: diabase and indurated quartz sandstone. Subordinate amounts of shale clasts also occur, some of which contain small pelecypods typical of the Lower Triassic Heiberg Group. In a measured section due west of the fossil forest site (Ricketts and McIntyre, 1986, section B), there is a distinct stratigraphic trend in clast composition, where diabase comprises about 20-30% of lithologies near the base, and 80-90% near the top. There is also an increase in clast size. Clasts of indurated sandstone are typical of the Heiberg, Awingak, and Isachsen formations. Diabase sills, although present in most pre-Kanguk Formation units, most commonly intrude the Blaa Mountain Formation shale. Spectacular examples are exposed at Buchanan Lake and several places in the hanging wall of the Stolz Thrust. The trend in Buchanan Lake clast composition reflects an unroofing sequence during uplift along the thrust system, wherein the younger Mesozoic sandstones were eroded first, and most of the diabase sills in Triassic strata later.

A schematic portrayal of depositional environments for the Buchanan Lake Formation at Geodetic Hills is shown in Figure 54. Illustrated here is uplift of ancestral Princess Margaret Arch supplying debris to the east-southeast prograding, Middle Eocene alluvial fan/meanderplain system. Paleocurrents indicate east to southeast sediment transport in the meanderplain settings, a shift from the more easterly directed flow closer to the source region. If these trends reflect regional paleoslopes, it is inferred that the basin axis was oriented approximately southeast such that the basin was closed to the north and open to the south.

Mokka Fiord area

The Buchanan Lake Formation type section is illustrated in Ricketts (1986), and redrawn in Figure 2b (RAK 15-85; RAK 20-85). Ninety per cent of the succession consists of the conglomerate-sandstone and the thick sandstone facies (Table 4). About 40 m of massive, dark grey thick conglomerate is located in the hanging wall of the frontal thrust in the Stolz fault system. Details of local structural complexities in the fault zone are discussed by van Berkel (1986).

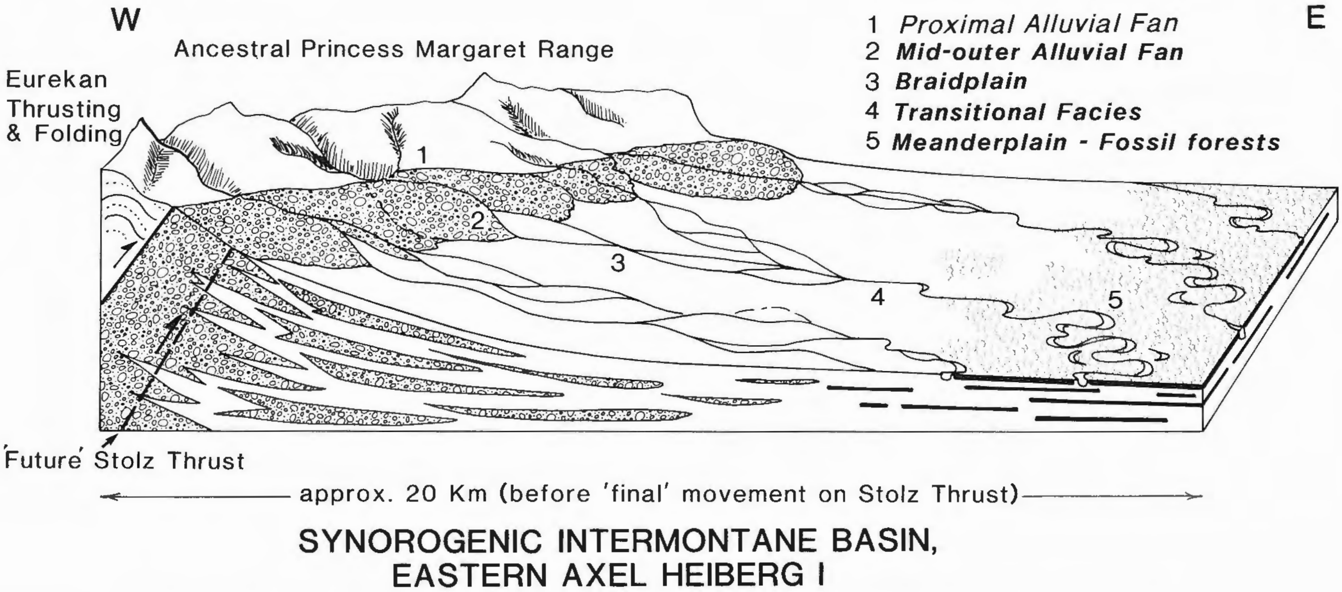

Figure 54. *A reconstruction of paleoenvironments for the Buchanan Lake Formation along eastern Axel Heiberg Island (from Ricketts, 1991b).*

Figure 55. *Interbedded sandstone and crossbedded conglomerate in the conglomerate-sandstone member, interpreted as stacked transverse braid bars; Mokka Fiord (section RAK 20-85). ISPG 2408-269*

Figure 56. *Ripple and ripple-drift bedding in sandy facies associated with channel or overbank development. Ripples above the pencil are mostly in-phase drift and are overlain by parallel laminated, plane-bed deposits; section RAK 20-85, Mokka Fiord. Scale is 18 cm long. ISPG 2408-210*

Contact with the Iceberg Bay Formation is probably a subaerial disconformity because:

i. beds above and below the disconformity are of fluvial origin, and

ii. no detectable hiatus can be resolved by palynology, and both formations at this locality are Middle Eocene in age.

Beds immediately overlying the disconformity belong to the thick sandstone facies. None of the finer grained facies of meandering stream/floodplain origin are preserved. Abundant trough crossbeds are littered with plant debris. Fining-upward cycles are rare. The few mudrocks that do occur fill small channels. Coaly layers are also rare. Channel and transverse bar structures are common in both sandstone and conglomerate (Fig. 55). Gravel bars are no thicker than 1.5 m. Sandstone transverse bars up to 60 cm thick are commonly stacked into units 5-8 m thick.

These facies, like their counterparts at Geodetic Hills, accumulated in braided streams having mixed sand and gravel loads. Rapid accretion of within-channel bars was followed in places by a veneer of mudstone followed by ripple-drift sandstone composing beds up to 2 m thick. Ripple sets with lee-side preservation are most abundant, and as shown in Figure 56, these grade up to in-phase sets (both lee and stoss faces preserved; Jopling and Walker, 1968). Formation of ripple drift is usually attributed to mixed bedload and rapid suspension fallout. Separation of within-channel transverse bars from ripple drift beds by thin mudstone suggests that the channel migrated and that rapid sedimentation took place in the adjacent floodplain during waning flood conditions.

Thick conglomerate capping the Mokka Fiord succession bears many of the features seen in the Geodetic Hills examples, and represents proximal braided stream or braidplain deposits. Overall, the stratigraphic progression here is consistent with a prograding clastic wedge. However the alluvial fan component was presumably cannibalized during the latest stage of thrusting.

Clast composition is uniform throughout, with diabase ranging from 75% to 90%, and the remainder indurated sandstone and shale.

Whitsunday Bay area

At least 350 m of Buchanan Lake strata are exposed in cliffs eroded in the footwall of Stolz Thrust between the head of Whitsunday Bay and Whitsunday Bay Diapir (Fig. 57). An additional 300 m (for a total of 700 m) is estimated in an area of recessive weathering between the cliffs and the outcrop limit of Iceberg Bay Formation east of the modern trunk river. Buchanan Lake conglomerate is overthrust by the Carboniferous-Permian Hare Fiord Formation, except at the southern limit of exposure where a sliver of black fossiliferous shale (probably Blind Fiord Formation) and diapiric anhydrite (Otto Fiord Formation) have been structurally inserted between the Permian rocks and the conglomerate (Fig. 58). Structural dips on Buchanan Lake beds as high as 75°E occur in the footwall.

Stratigraphic trends that can be discerned in section RAK 17-88 (Fig. 2d) include:

- a change from fining-upward cycles of conglomerate-mudstone (lacking coal) to thicker bedded units of conglomerate-sandstone high in the section.
- Concomitantly, there is an upward decrease in planar tabular crossbedded gravels; the largest sets are 3.5 m thick near the base.
- Maximum clast size increases upwards from 12-15 cm near the base, to 30 cm at the top.
- Many conglomerate units in the middle of RAK 17-88 are clearly wedge shaped, extending laterally 50-80 m. The overall stratigraphic pattern is a series of offset, stacked wedges. Thicker conglomerate units towards the top of the section tend to have greater lateral extent.
- Beds containing root structures and paleosols (gleysol type) were located at only two sites midway through the section.

The entire succession at Whitsunday Bay is generally indicative of mixed load, braided stream deposition, not unlike that at Mokka Fiord. Alluvial fans are not preserved, presumably because of thrusting, and/or cannibalizing. At the opposite depositional extreme, meandering stream and forested floodplain components did not develop in the area.

Clast composition varies from about 75% diabase at the base of the section to 90% at the top, the remainder being sandstone and shale. However there is considerable local variation between these two values and therefore a clear unroofing trend is equivocal.

West Nansen Sound

Buchanan Lake rocks underlie an incised coastal plain some 15 km northwest of Stang Bay (Fig. 52). In a small outlier due west of Stang Bay, conglomerate and sandstone unconformably overlie the Heiberg Formation with an angular discordance of 5°-8° (Fig. 59). Elsewhere the Buchanan Lake Formation overlies diabase sills that intrude the Blaa Mountain Formation.

The 300 m of section recorded is a minimum value for the area; an additional 200-300 m may occur stratigraphically below this section (Fig. 59). Structural dips here are locally as high as 40°.

Four facies are present: thick conglomerate (30%), conglomerate-sandstone (10%), conglomerate-mudstone (coal) (15%), and sandstone-coal (45%). Their internal organization is similar to facies in other parts of Axel Heiberg Basin, with minor variations. Lignite seams and associated root intervals are only a few centimetres thick; abraded wood debris abounds. Spruce and *Metasequoia* cones are common in basal lags. Log jams, apparently confined to small channel structures, consist of a chaotic mixture of logs, finer woody debris, leaves, cones and pebbles.

Unlike other localities in Axel Heiberg Basin, the stratigraphic variation in facies is not clear. Conglomeratic facies tend to be more common in the central part of the succession but it is not known whether this trend is of local or greater significance because of limited exposure.

The proportion of diabase clasts varies between 50% and 75%, the remainder of the framework consisting of indurated sandstone. There is no discernible stratigraphic trend in composition.

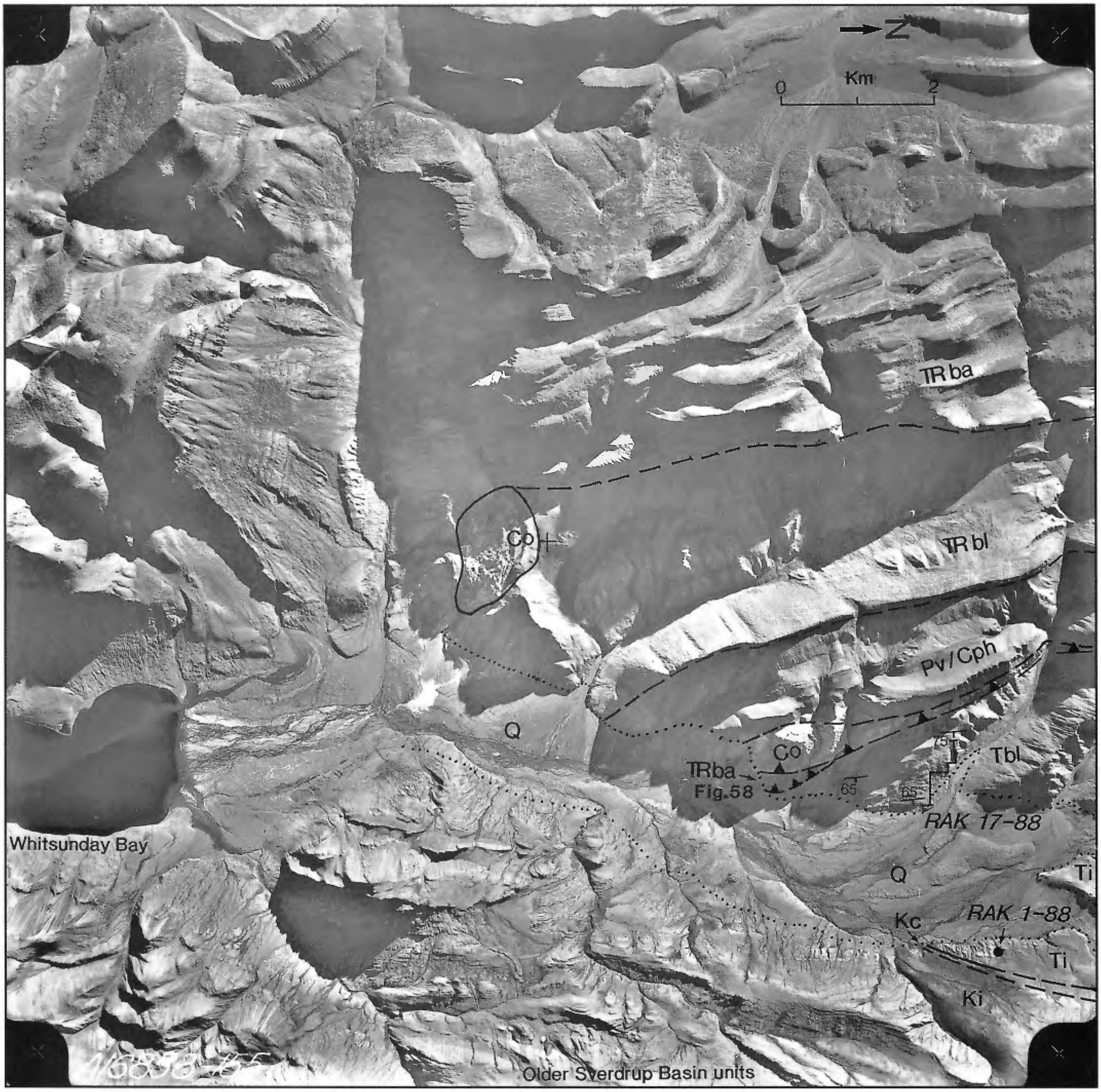

Figure 57. Aerial photograph of mapped Buchanan Lake (Tbl) and Iceberg Bay (Ti) formations, just north of Whitsunday Bay. Buchanan Lake Formation conglomerates are in thrust contact with Otto Fiord evaporites (Co) and Lower Triassic Blaa Mountain Formation shale (TRba). Pv/Cph = Van Hauen and Hare Fiord formations respectively; TRbl = Blind Fiord Formation; Q = Quaternary. NAPL vertical photograph A16858-155. Photo centre 79°06'N; 87°06'W. The thrust is the southern extent of Stolz Fault zone. Older bedrock geology from Thorsteinsson (1972a).

According to facies interpretations elsewhere in Axel Heiberg Basin, the above associations correspond to proximal braidplain or braided stream, mixed load fining upward facies that are transitional between fully braided regimes and high sinuosity regimes. More distal sand-dominated meandering streams and vegetated floodplains also are represented.

Eastern Schei Peninsula

Flat-lying Buchanan Lake strata occupying the eastern part of Schei Peninsula unconformably overlie resistant Blaa Mountain Formation shale and indurated sandstone of the Blind Fiord Formation with an angular discordance up to 50°. One measured section (RAK 4-88; Fig. 2d) records a stratigraphic thickness of 58 m; the total thickness is unknown.

Although superficially resembling the Iceberg Bay Formation Coal member, the Schei Peninsula succession lacks the typical 'striped' appearance. In addition the sandstone is lithic, like all Buchanan Lake sandstone, rather than quartz rich. The palynofloral assemblage is also comparable to that found in the fossil forest lignites on Axel Heiberg Island (Appendix 1).

Two lithofacies associations are present – the sandstone-coal and siltstone-coal associations, both in fining-upward cycles 1-4 m thick. Lignite seams that cap the cycles range in thickness from a few millimetres to a metre, and are generally muddy, woody, and locally herbaceous, leafy, and resinous. Large wood fragments have been compressed between 2:1 and 3:1. Seat earths underlie some seams. Ripples and trough crossbeds, some lined with small diabase pebbles, are common in the fine grained sandstone beds.

Sedimentologically, the Schei Peninsula facies resemble the relatively low energy, high sinuosity streams and adjacent vegetated floodplains elsewhere in Axel Heiberg Basin. Because of their fine grain size and distal character they can be regarded as distal equivalents of the Geodetic Hills fossil forests. Alternatively, the deposits may have accumulated in a small subbasin, possibly separated from Axel Heiberg Basin by the ridge of Triassic rocks that form the backbone of Schei Peninsula. Distinction between these two options is not possible given the limited exposure and dearth of paleocurrent indicators. If the deposits are part of Axel Heiberg Basin, then a minimum basin width of about 70 km can be inferred.

Summary of Axel Heiberg Basin

The Middle Eocene Axel Heiberg Basin has a strike-parallel length of more than 200 km. The Buchanan Lake Formation overlies with profound unconformity rocks as old as Early Triassic, and as young as Middle Eocene (Iceberg Bay Formation). The unconformity is probably subaerial in origin. Therefore the principal phase of deformation leading to formation of Axel Heiberg Basin must have occurred during the Middle Eocene. However, it is possible that some deformation continued into the Late Eocene (McIntyre, 1991b) and even earliest Oligocene, although independent verification of this is lacking. Only the most proximal Buchanan Lake rocks were involved in Eurekan thrusting. Deformation of strata outboard of the Stolz frontal thrust system produced only gentle, open warps and minor extension faults. Most of the alluvial fan deposits were removed, possibly cannibalized; surviving alluvial fan facies are preserved only in the western most limits of exposure at Geodetic Hills. Furthermore, with coalification ranks as low as 0.14 (Goodarzi et al., 1991) and preservation of some woody material, it is unlikely that the Buchanan Lake deposits were ever deeply buried; only a veneer of probable Neogene-age sediment overlies the formation. From these stratigraphic and structural considerations it is inferred that most of the Buchanan Lake deposits in Axel Heiberg Basin represent a relatively late stage of Eurekan deformation, unroofing, and sedimentation. The Middle Eocene period, lasting 8 to 10 Ma, witnessed the conclusion of Iceberg Bay sedimentation, most of the Eurekan deformation, and deposition of the syntectonic Buchanan Lake Formation in Axel Heiberg Basin.

Figure 58.

Details of hanging wall units thrust (Stolz system) over Buchanan Lake conglomerate (BL), north of Whitsunday Bay (near section RAK 17-88), viewed towards the southeast (see Fig. 57). TRba = Blaa Mountain Formation. ISPG 3070-146

Otto Fiord subbasin

General characteristics

A sliver of Buchanan Lake Formation, fault-bounded on its eastern margin and in apparent stratigraphic contact with Triassic strata on the west side, is exposed along the south shore of Otto Fiord, 18 km southwest of Van Hauen Pass (Fig. 60). The stratigraphic, structural, and compositional relationships at this locality are sufficiently different to any other Buchanan Lake locality to warrant designation as a separate basin or subbasin.

A composite of measured sections indicates at least 350 m of strata in the basin (RAK 14a-85, RAK 14c-85, Fig. 2b) which are folded into a north-plunging syncline. And here an

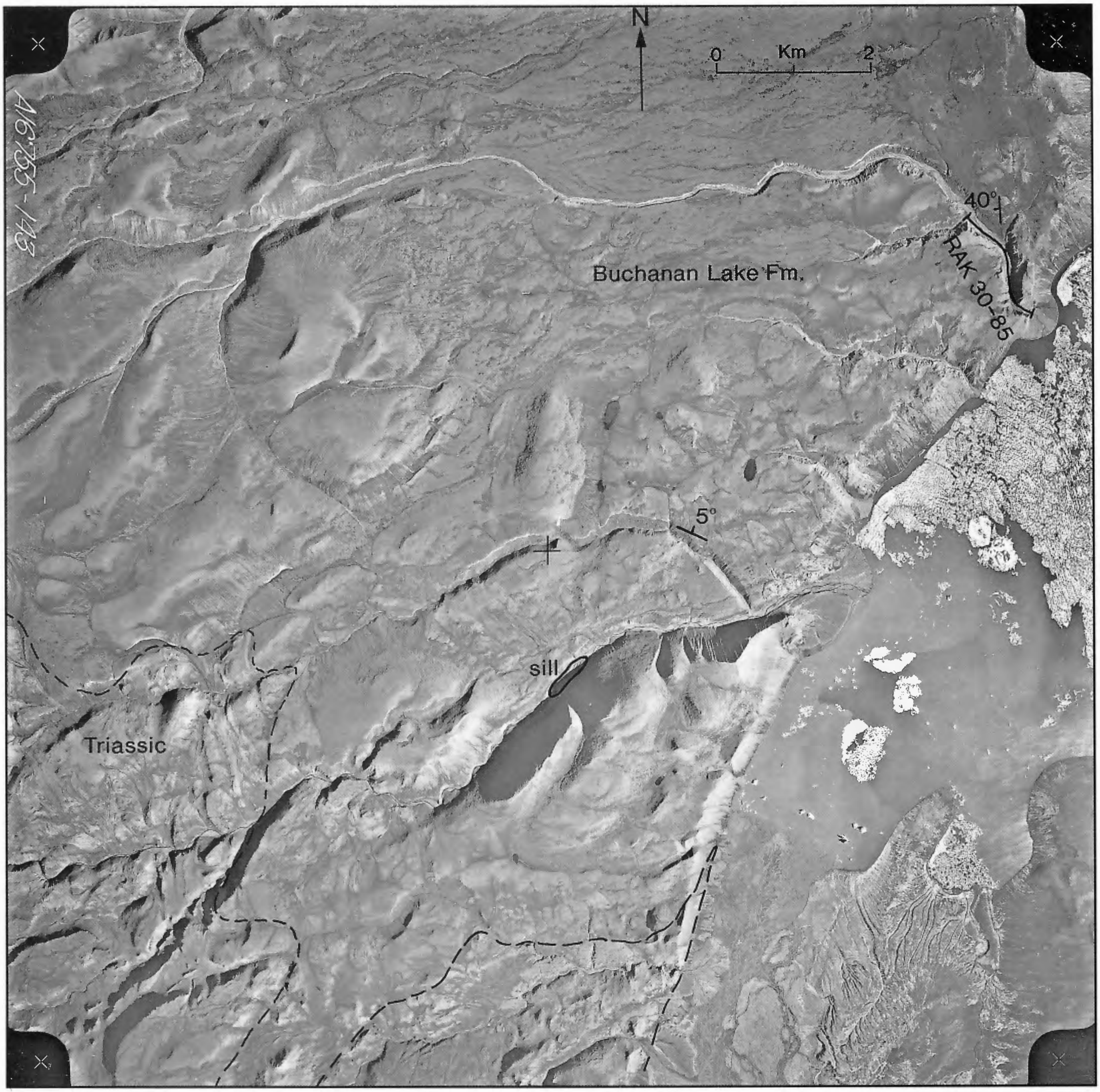

***Figure 59.** The northern limit of Buchanan Lake Formation, primarily the conglomerate-sandstone member, located on the west shore of Nansen Sound, about 15 km northwest of Stang Bay. Exposure is poor except along a stream cut at section RAK 30-85. Dips in small fault blocks are up to 40°. NAPL vertical photograph A16755-143. Photo centre 80°36.5'N; 90°53'W.*

intriguing stratigraphic problem exists. West of the fold axis, conglomerate beds consisting predominantly of siliceous, fossiliferous limestone and intercalated with recessive sandstone and mudstone, are in stratigraphic contact with basal sandstone beds of the Triassic Heiberg Formation. East of the fold axis the same conglomerate beds are underlain by a thick succession of lithic sandstone and diabase-rich conglomerate, which is faulted against a Blaa Mountain Formation succession replete with thick diabase sills.

The diabase-rich lithologies are very similar to those encountered in Axel Heiberg Basin. Sandstone beds, mostly recessive, contain trough and planar-tabular crossbeds and

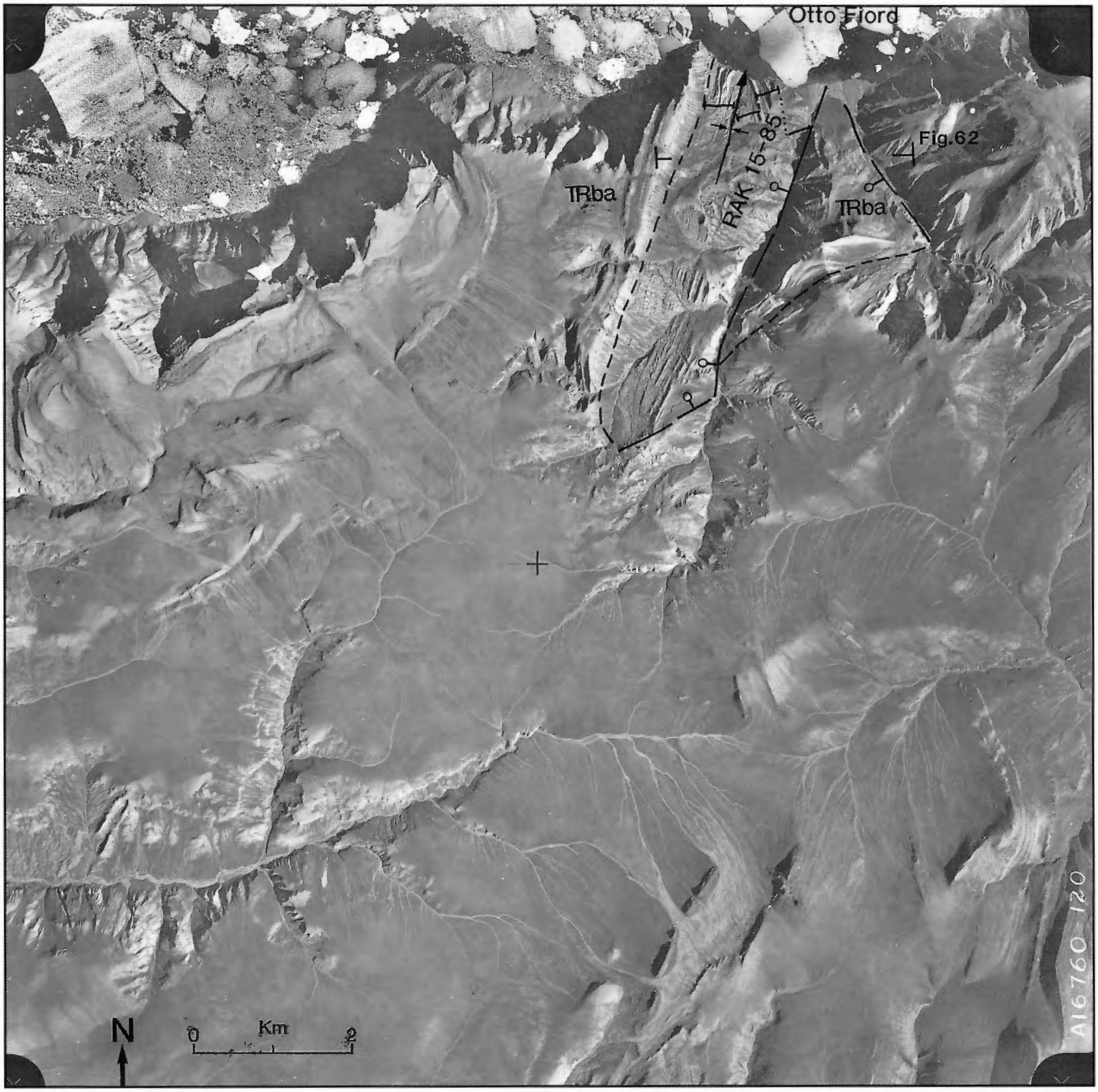

Figure 60. *Buchanan Lake Formation exposed in a fault block south of Otto Fiord. Pebble conglomerate of limestone and siliceous limestone are exposed in the core of the syncline; thick diabase conglomerate on the eastern flank. The line of section for Figure 62 is indicated. TRba = Blaa Mountain Formation. NAPL vertical photograph A 16760-120. Photo centre 80°57′N; 86°55′W.*

thin pebble layers. Conglomerate beds tend to be massive and crudely layered, with maximum clast sizes up to 15 cm. Most appear to have clast supported frameworks. These lithologies belong to the thick conglomerate and conglomerate-sandstone facies (Tables 3 and 4). Compositionally they are distinct from the Axel Heiberg Basin varieties with the presence of 10-15% limestone and 20-25% chert clasts, in addition to 45-50% diabase, 10-15% indurated sandstone, and 3-5% shale clasts (Fig. 61).

Age

Representatives of the sandstone-coal facies are present near the top of the succession although it is only the lignite beds that are well exposed. Lignites are woody and contain abundant conifer and hardwood leaf impressions. A probable Early Eocene age has been determined by D.J. McIntyre from the pollen assemblage (Appendix 1). The Buchanan Lake Formation at Otto Fiord is therefore older than that in Axel Heiberg Basin.

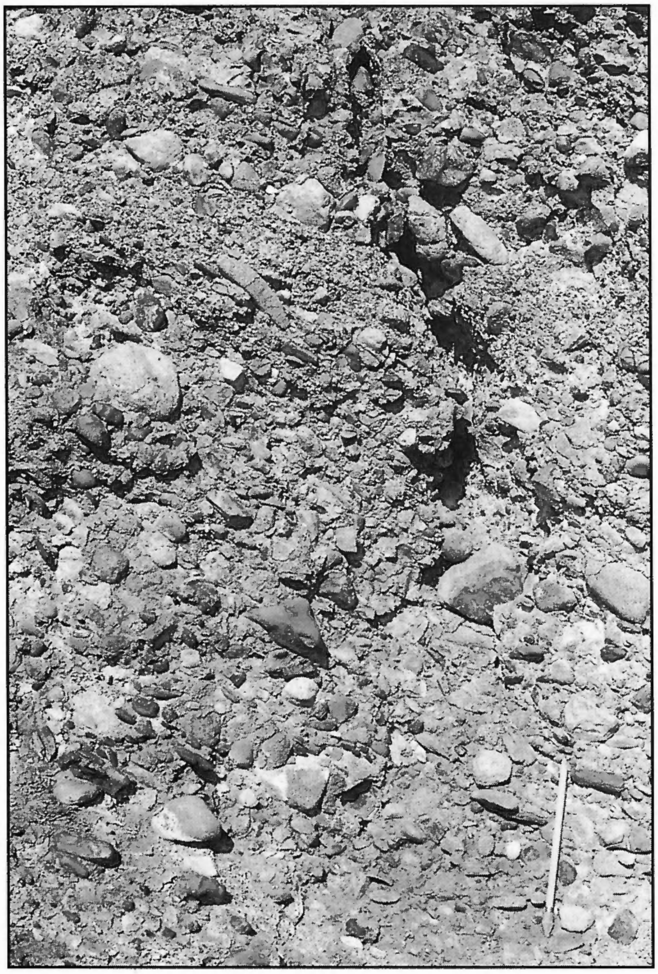

Figure 61. *Typical diabase-rich conglomerate of the Buchanan Lake Formation, east flank of the Otto Fiord fault block; section RAK 15-85. Pencil is 18 cm long. ISPG 2408-197*

Provenance and unroofing

Conglomerate units high in the Otto Fiord succession contain a distinctive clast assemblage of 56% limestone and siliceous limestone, commonly bearing corals, bryozoa, and brachiopods; 34% grey, white, and rarely black chert; and 10% diabase pebbles. Three main conglomerate units occur – 10 m, 40.5 m, and 18 m thick. Only in the thickest unit was trough crossbedding visible, in sets up to 50 cm thick.

As in the Axel Heiberg Basin, clasts of diabase were probably derived from the great thickness of Cretaceous diabase sills that invade the Blaa Mountain Formation. The most obvious source of fossiliferous limestone, siliceous limestone, and chert is the Upper Carboniferous and Permian Nansen Formation, a large swath of which underlies Hvitland Peninsula north of Otto Fiord (Thorsteinsson and Trettin, 1972a, b); Nansen Formation also occurs in faulted inliers on Svartfjeld Peninsula (Thorsteinsson and Trettin, 1972c). The Permian Degerböls Formation is another possible source of chert.

Unroofing trends analogous to those observed at Geodetic Hills are evident in the eastern limb of the syncline. Diabase clasts are positioned low in the Buchanan Lake succession, whereas carbonate and chert clasts derived from Upper Paleozoic strata are located near the top of the section. This unroofing trend is less apparent on the western limb of the syncline, where the succession is considerably thinner, does not contain significant diabase-rich conglomerate, and is in stratigraphic contact with Triassic bedrock. An explanation of this basin architecture is schematically portrayed in Figure 62: A half-graben was bound on its eastern margin by a north-striking, down-to-the-west fault (also the present fault displacement), such that the deepest part of the basin was adjacent to the fault. An uplifted shoulder was situated on the west side of the half-graben. The unroofing sequence would result in diabase-rich conglomerate that was deposited first in the down-faulted eastern portion, and progressively onlapped the uplifted shoulder toward the west. Subsequent erosion of Upper Paleozoic rocks resulted in substantially different conglomerate units that onlapped Triassic bedrock farther to the west. Later contraction produced minor folding. If this scenario is correct the Otto Fiord Basin clearly had a different origin than the Axel Heiberg, Franklin Pierce, and Lake Hazen basins, all of which are associated with major thrusts. Notably, the normal fault that displaces the eastern basin margin is strike-parallel with Black Top Fault on Fosheim Peninsula.

Emma Fiord subbasin

Steep, bluff-forming conglomerate rising above the south shore of Emma Fiord (Fig. 63) comprises the most northerly occurrence of Buchanan Lake Formation on west Ellesmere Island. The north-dipping panel, down-faulted on the south side against Silurian volcanic rocks, contains a 300 m succession which at its base is in disconformable contact with the Iceberg Bay Formation (Coal member; see Fig. 50). The disconformity is probably subaerial, similar to the contact in Axel Heiberg Basin.

Two facies associations are present: the thick conglomerate and conglomerate-sandstone associations, the latter containing prominent tabular-planar crossbed sets 1-2 m thick. All lithologies are weakly consolidated and detailed examination of the cliff exposures is difficult. Maximum clast diameter is about 40 cm in beds near the basal contact, decreasing higher in the succession; average clast size is about 10 cm. Imbrication in the conglomerate-sandstone facies generally indicates southwest stream flow. Coal beds are not present, although a few rafts of coaly shale do occur. There are no discernible stratigraphic trends in the facies.

Clast compositions are similar to those in the carbonate- and chert-rich beds at Otto Fiord subbasin: crinoidal and bryozoan limestone (including corals; 50-55%), siliceous and dolomitic limestone (3-5%), grey and green chert (10-12%), diabase (12-15%), and indurated sandstone (20-25%). No stratigraphic changes in clast composition were observed.

A short distance west of the Emma Fiord cliffs, interbedded conglomerate, lithic sandstone, and shale are exposed in a stream bank. Conglomerate beds up to 4 m thick contain clasts up to 12 cm across, averaging 3-4 cm; sandstone beds are 1-2 m thick; shale intervals 5-6 m thick. One shaly lignite bed 10 cm thick is present. Few sedimentary structures are visible. The overall appearance of these lithologies suggests comparison with the conglomerate-mudstone facies. Clast composition is identical to conglomerate in the adjacent cliffs, however, the exact stratigraphic relationship of the stream exposures to those in the cliffs is not clear. Projection of the beds along strike indicates that they correspond to the interval containing the Buchanan Lake/Iceberg Bay formation contact, but there is no equivalent sand and shale unit where the contact is exposed. D.J. McIntyre has determined an Early Eocene age from palynomorphs in the shale and lignite beds, an age that corresponds to that obtained from the Otto Fiord subbasin deposits. Thus, the Emma Fiord subbasin is as old as Early Eocene, but it is not known whether the main body of conglomerate here is the same age or younger.

Nevertheless, it is significant that both the Emma Fiord and Otto Fiord subbasins are of similar age, and contain rocks of like composition; both are significantly older than either Axel Heiberg or Lake Hazen basins. Similarity in clast types

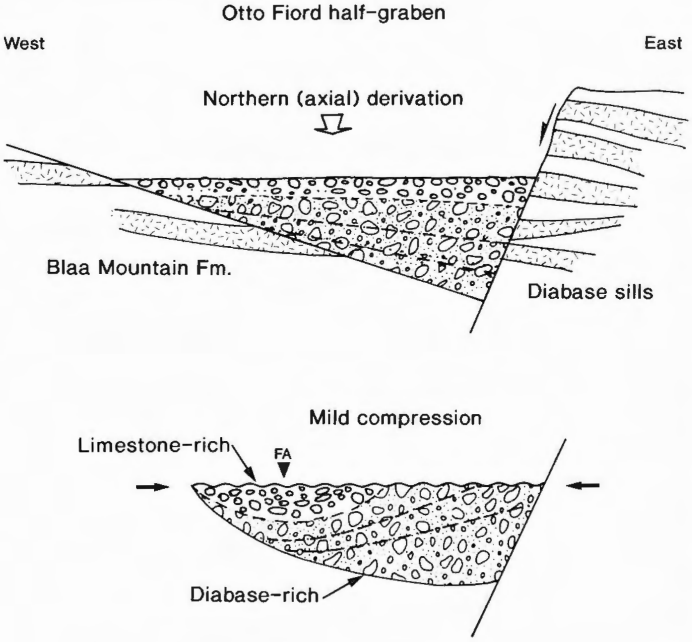

Figure 62. *Deposition of Buchanan Lake conglomerate in the Otto Fiord fault block interpreted as half-graben fill during the earliest stages of Eurekan tectonism. Later folding took place during mild compression. The unroofing stratigraphy is indicated by Upper Paleozoic-derived, limestone-rich conglomerate in the core of the syncline. FA = syncline axis; the line of section is indicated on Figure 60.*

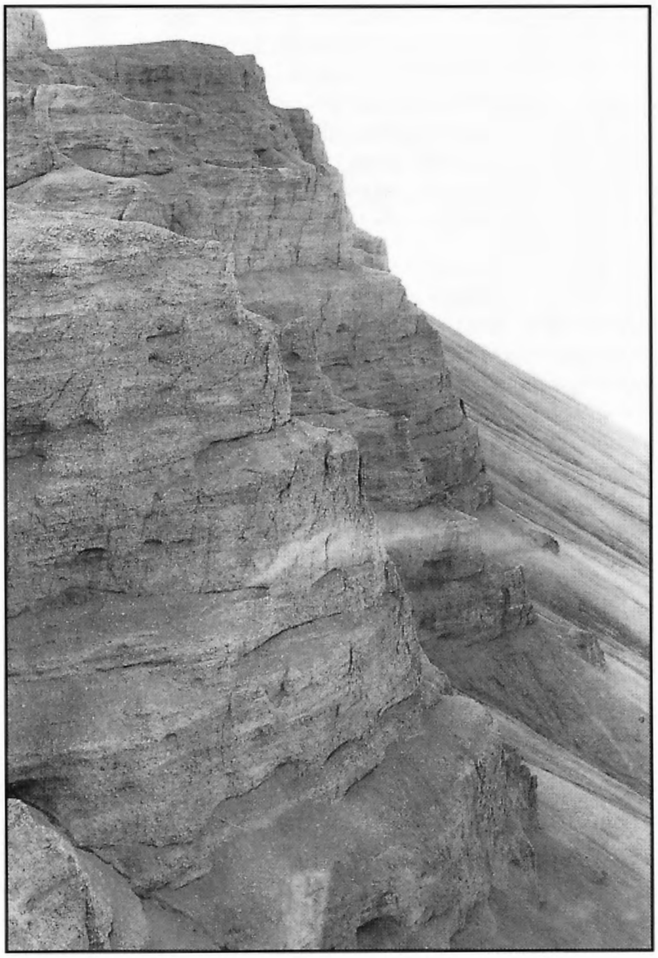

Figure 63. *Buchanan Lake conglomerate exposed in cliffs along the south shore of Emma Fiord. View (about 80 m thick) shows high lateral variability in bedding, and some stacking of planar-tabular crossbed sets; section RAK 13-87. ISPG 2835-224*

further indicates a common source type, which presumably was being eroded at about the same time. But unlike the Otto Fiord basin, no unroofing stratigraphy is evident at Emma Fiord.

The range of sedimentary environments preserved at Emma Fiord is limited to proximal braided river facies and possibly channels that were transitional with gravelly meandering reaches.

Franklin Pierce Basin (Judge Daly Basin)

Conglomerate assigned to the Buchanan Lake Formation is exposed discontinuously along the east-central coast of Ellesmere Island, from Franklin Pierce Bay and adjacent areas in the south, to Cape Back (Judge Daly Promontory) in the north. Previous descriptions by Miall (1982), Mayr and de Vries (1982), and Ricketts (1986) identified the general Tertiary nature of the deposits. In the Franklin Pierce Bay area they were previously mapped as Ordovician to Lower Devonian (Kerr, 1973). Clues to their Tertiary age come primarily from fossil broad-leaf impressions, commonly of birch and oak, and Mayr and de Vries (1982) report a Paleocene age based on the form taxon *Crednaria*. Because it is uncertain whether the age range of this form extends into the Eocene, the Paleocene designation is equivocal.

In his earlier analysis, Miall (1984a, b) inferred the existence of separate basins in the area of Judge Daly Promontory and Lake Hazen, however the stratigraphic and structural criteria in support of this contention can be questioned. Critical information provided by the depositional facies and paleocurrents is ambiguous because outcrops are widely scattered. It cannot be stated categorically whether the Buchanan Lake outcrops were once part of a continuous belt or basin along eastern Ellesmere Island, or if they constituted isolated basins. Presently, the areas of exposure are juxtaposed against different structural elements of the Eurekan Orogen, although all the sediments apparently originated from Lower Paleozoic bedrock. Two broad areas will be considered in this report: Franklin Pierce Bay, and the expanse between Dobbin Bay and Cape Back (Fig. 52).

Franklin Pierce Bay

The swath of Buchanan Lake Formation conglomerate extending from Dobbin Bay in the north to Copes Bay, lies outboard of the Parrish Glacier Thrust (Fig. 64). Parrish Glacier Thrust dips north at 30° to 50° (Mayr and de Vries, 1982) and carries Lower Paleozoic rocks of the Franklinian Fold Belt in its hanging wall. Small imbricate thrusts also disrupt footwall strata, including the Buchanan Lake beds. Mayr and de Vries (1982) report a klippe between Franklin Pierce and Allman bays. A shallow-dipping klippe of limestone was also observed in 1984 east of the mouth of Franklin Pierce Bay (Fig. 64 and 65). Limestone samples submitted for conodont analysis proved barren of fossils. The position of the klippe indicates that a thrust flat may extend up to 8 km from the principal Parrish Glacier Thrust trace. This is one of the few surface expressions of ramp/flat geometry to be seen in the Eurekan Orogen.

More than 300 m of distinctive red-weathering conglomerate is exposed in spectacular (but highly unstable) canyons and sea cliffs. Structural dips are very low. The red colouration is unique to the eastern Ellesmere Island exposures. Conglomerate predominates, with sandstone comprising less than 5% of the succession. Three lithofacies associations are represented: matrix-supported conglomerate facies (about 10%), thick conglomerate facies (45%), and conglomerate-sandstone facies (45%).

Units of matrix-supported conglomerate up to 10 m thick are present in a few places. Sorting is characteristically poor, with components ranging in size from 35 cm to sand-size (Fig. 66a). Intercalated fine granule layers tend to have clast-supported frameworks and some crossbedding; the layers further indicate that flow units were composite. Crude inverse grading is present in some beds (Fig. 66b). Also illustrated in Figure 66b is a possible example of sieve deposits (e.g. Gloppen and Steel, 1981) which contain a strongly bimodal clast size distribution and an apparently open framework, which in this case has been filled with sparry calcite cement.

Most of the conglomerate beds in the area are thick and relatively massive, having tabular or wedge geometries (Fig. 67). Beds belonging to the thick conglomerate facies contain crude internal layering and some clast imbrication in clast-supported frameworks. Representatives of the conglomerate-sandstone facies have similar thicknesses, up to 1-2 m, and contain large tabular-planar crossbed sets up to 1.5 m thick. There are no fining- or coarsening-upward trends. Thin, highly irregular sandstone veneers separate a few beds and crossbed sets. Asymmetric current ripples are common and broad-leaf impressions occur sporadically (Fig. 68). Original carbonaceous material associated with the leaves and other plant debris has been completely oxidized.

Clast composition is distinctly different from deposits in the other synorogenic basins. Most abundant is grey, fossiliferous limestone and mottled argillaceous limestone (55-65%); tabulate corals (e.g. *Halysites*) and colonial rugose corals are present. Subordinate varieties include white quartz sandstone (12-15%), buff fine sandstone (10-15%), laminated argillite (5%), black chert (1%), and pink recrystallized limestone (1-2%). Clast shape is highly variable, even in clasts of similar composition, ranging from well rounded to angular. All of the clast types are common in the adjacent Lower Paleozoic terrane.

Interpretation of the lithofacies follows the same lines of reasoning applied to the Axel Heiberg Basin, with some significant differences. Intercalation of matrix-supported, thick conglomerate, and conglomerate-sandstone facies indicates that deposition took place at the distal portion of alluvial fans which fed the proximal reaches of braided streams or braidplains. There appears to have been little vegetation cover. Pervasive oxidation and reddening of sediments indicates relatively arid conditions, and hence the debris flows were generated on arid-type alluvial fans rather than the humid varieties inferred for Axel Heiberg Basin, Otto Fiord, Emma Fiord, and Lake Hazen basins. The difference possibly resulted from rain-shadow effects created by elevated topography in the thrusted terrain. Sieve deposits may also have been more common in this type of setting. Like the proximal settings in the other Eurekan basins, within-channel bedforms most commonly consisted of longitudinal gravel bars in the higher reaches, and transverse bars in the lower reaches.

Textural differences compared to the other basins also are important. Highly variable clast shape may reflect different degrees of thrust-related sediment reworking, or cannibalization; rounded clasts are multicycle, reworked from thrusted panels of Buchanan Lake strata, whereas angular varieties are first cycle products. Indeed, Buchanan Lake rocks were involved in thrust deformation although there is too little data in the present study to relate specific examples of textural inversion to individual thrusts. The argument that degrees of angularity were related to different original source rock compositions (e.g. carbonate versus sandstone) is discounted because textural variation exists within specific rock types.

Dobbin Bay to Cape Back (Carl Ritter Bay)

Tertiary conglomerate, exposed in isolated patches at Cape Back, Cape Lawrence, and a few other places has been described by Mayr and de Vries (1982) and Miall (1982). Up to 1000 m is recorded at Cape Lawrence. Christie (1974) has described steeply dipping conglomerate beds in unconformable contact with Franklinian bedrock. In contrast to lithologies at Franklin Pierce Bay, these conglomerates are oligomictic, being composed of carbonate clasts. Lower Cambrian sandstone clasts are rare, even though these rocks are in structural contact with the conglomerate. Lateral changes in rock types in the immediate footwall of Rawlings Bay Thrust, from conglomerate to Lower Paleozoic strata, were interpreted to reflect prethrusting erosion of Buchanan Lake Formation.

At least 630 m of conglomerate exposed at Cape Back (Carl Ritter Bay locality of Miall, 1982) is interpreted as being in unconformable contact with subjacent rocks, here correlated

Figure 64. Oblique aerial photograph of Franklin Pierce Bay, eastern Ellesmere Island, viewed to the west, showing Buchanan Lake Formation in relation to Parrish Glacier Thrust and a klippe of Lower Paleozoic limestone thrust over the conglomerate (see Fig. 65). Figure 67 is also located. Photo NAPL T399L-190.

with the Iceberg Bay Formation. Boulders up to 55 cm across form a clast-supported framework. Conglomerate lithofacies were interpreted as proximal braided river and alluvial fan assemblages (Miall, 1982, 1984b). No paleocurrent data are available from this area. Miall describes spectacular internal stratigraphic discordances and minor folding, all of which indicate syndepositional deformation. The Buchanan Lake beds at Carl Ritter Bay are down-faulted against Ordovician carbonate units by a steeply dipping reverse fault, where the hanging wall carbonate beds dip almost vertically.

The present disposition of map units is a function of younger Eurekan deformation and bears little relationship to original basin paleogeography. It is possible, as Miall (1984b) suggests, that Buchanan Lake deposits on Judge Daly Promontory accumulated in a small extensional basin during the Paleocene, but given the isolated nature of exposures here this hypothesis is difficult to test. What is clear however, is that the Buchanan Lake conglomerate is associated with a major phase of tectonism. If the Paleocene age for the conglomerate is correct (Mayr and de Vries, 1982) then the deformation is considerably older than that determined (from much more substantial data bases) in Axel Heiberg, Otto Fiord, Emma Fiord, and Lake Hazen synorogenic basins.

Lake Hazen Basin

In the footwall of, and southeast of the Hazen Fault (thrust) zone, a swath of sandstone and conglomerate is included in the Buchanan Lake Formation. An earlier definition of Lake Hazen Basin by Miall (1979) includes interbedded sandstone, mudstone, and coal beds here equated with the Iceberg Bay Formation (see section above). In this discussion the basin name applies only to the syntectonic Buchanan Lake strata. The Lake Hazen 'depression' that preserves Buchanan Lake and Sverdrup Basin strata is structural in origin and is associated with Eurekan faulting and folding.

Trettin (1971), Christie (1974), and Miall (1979) recognized that Tertiary sediments at Lake Hazen overlie Mesozoic rocks between North Boulder Hill and South Boulder Hill. The same rocks onlap probable Paleocene Iceberg Bay Formation east of Lake Hazen, which in turn onlaps deformed Lower Paleozoic strata underlying Hazen Plateau. Palynomorph assemblages from the lignite beds contain sparse floras which McIntyre (Appendix 1) considers to be Early or possibly Middle Eocene.

Almost 500 m of Buchanan Lake Formation are exposed north and west of Lake Hazen in the Boulder Hills area. Detailed measured sections have been described by Miall (1979); additional sections are shown in Figure 2d (RAK 10-88, RAK 14-88). Strata low in the formation, possibly 160 m and more, consist of interbedded lithic sandstone, mudstone, and coal, and include the thick sandstone and sandstone-coal, and subordinate conglomerate-sandstone and conglomerate-mudstone facies associations. Overlying units consist mostly of the thick conglomerate association, although exposure of this facies is generally poor and is best seen at the top of section RAK 14-88.

Details of the individual facies are very similar to those found in Axel Heiberg Basin and will not be discussed further. General stratigraphic features include the following:

- The lower 80-100 m is dominated by sandstone and mudstone, frequently in fining-upward cycles.
- The first appearance of pebble and cobble conglomerate in section RAK 14-88 occurs at the 84 m level, as mixed gravel and sand, locally in channel structures (Fig. 69).
- Massive cobble and boulder conglomerate at the top of section RAK 14-88 appears abruptly but probably conformably (Fig. 69). Clasts greater than a metre in diameter are present. Percussion marks are common on clast surfaces.

Clast composition is as follows: white indurated sandstone 45%; grey fossiliferous sandstone (containing brachiopods, corals, bryozoa) 40%; dark chert 10%; and gabbro 5%. Buchanan Lake sandstone is lithic in contrast to the quartz-rich sandstone typical of the underlying Iceberg Bay Formation.

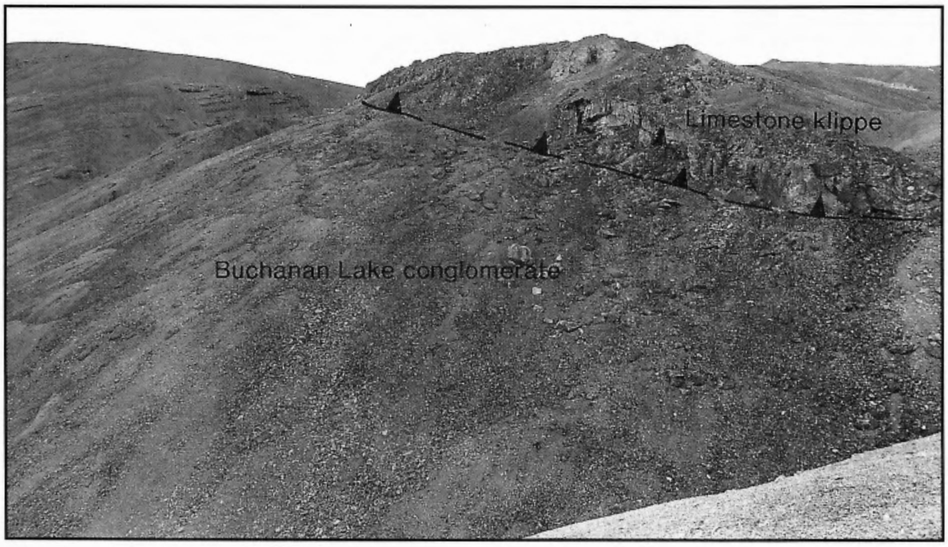

Figure 65.

Dense, crystalline limestone (probably Lower Paleozoic) thrust over rusty weathering conglomerate of the Buchanan Lake Formation, in a klippe approximately 10 km south of Parrish Glacier Thrust (see Fig. 64). ISPG 2238-167

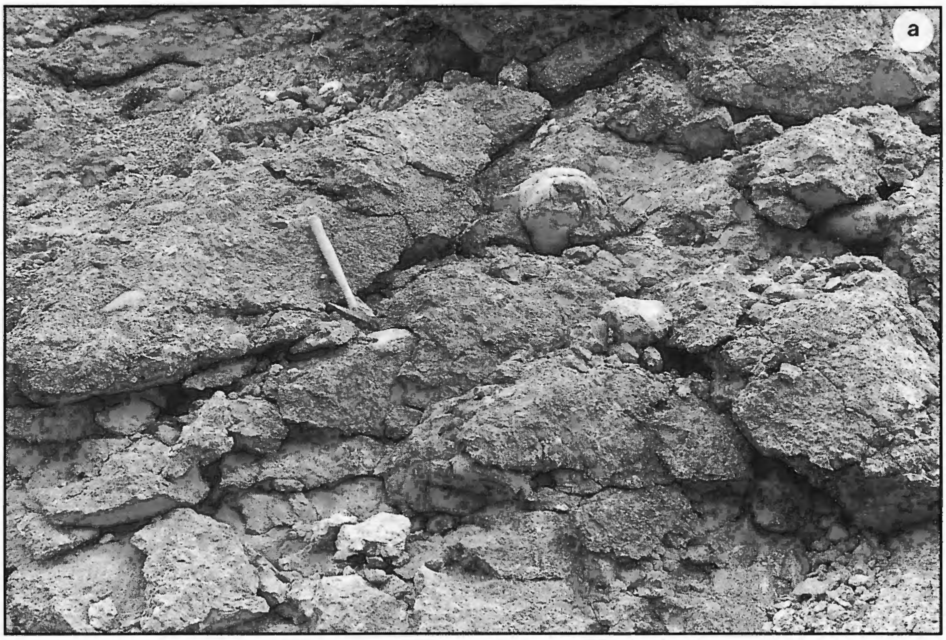

Figure 66.

a) *Red, rubbly weathering, matrix-supported conglomerate of probable debris flow origin. Boulders up to 40 cm across. Buchanan Lake Formation, Franklin Pierce Bay. All plant material has been oxidized. Hammer is 33 cm long. ISPG 2238-151.* ***b)*** *Alternating layers of fine and coarse pebble conglomerate in large planar-tabular crossbeds. Pebbles consist of Lower Paleozoic limestone, dolostone, sandstone, shale, and frequently coral fragments. White patches (arrows) are open-framework foresets that subsequently filled with coarse calcite. Buchanan Lake Formation, Franklin Pierce Bay. Scale is 3 cm long (lower right). ISPG 2238-160*

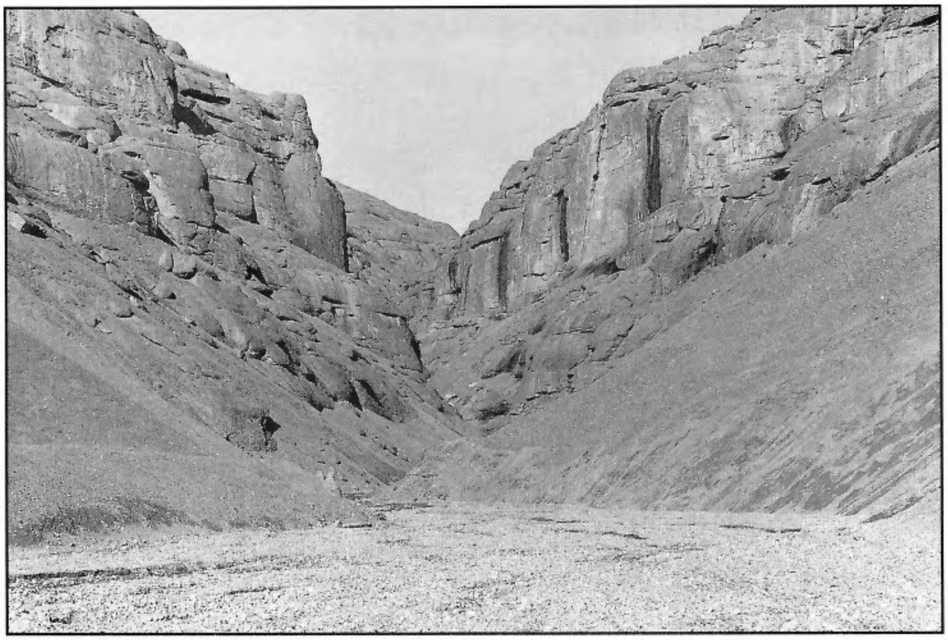

Figure 67.

Steep (and unstable) cliffs characterize Buchanan Lake conglomerate exposure in the Franklin Pierce Bay area (about 300 m of strata here – see Fig. 64). Bedding style is substantially different from conglomerate facies elsewhere in the Buchanan Lake Formation. ISPG 2238-142

Paleocurrent data is plotted on stratigraphic sections RAK 10-88, and 14-88 (Figure 2d; data are also presented by Miall, 1979). Most crossbed azimuths were taken from sandstone units below the massive conglomerate. Considerable variation in sediment transport directions is indicated, even at closely spaced localities. Miall (1979) interpreted the distribution of vector means to reflect both transverse and longitudinal paleoflow into the basin. This interpretation is reasonable (note that, unlike Miall, I consider the basin to have evolved during, and not prior to, Buchanan Lake deposition). The principal source area appears to have been Lower Paleozoic rocks in the hanging wall of the Lake Hazen thrust zone. No major structural element, having dimensions similar to that of the Lake Hazen structure, is known on the southeast margin of the basin and it is likely that Lake Hazen Basin had an asymmetric profile, like that of many foredeeps-deepest and thickest below the Lake Hazen thrust panel.

Paleoenvironmental attributes of strata equivalent to the Buchanan Lake Formation have been well illustrated by Miall (1984b). Miall portrayed a system of alluvial fans and braidplains which represent the transverse sediment transport components of Lake Hazen Basin, that were transitional into axial fluvial channel systems. The latter may be represented by the fining-upward sandstone-coal and conglomerate-mudstone facies.

Summary of the Buchanan Lake Formation

The Buchanan Lake Formation is contained within at least six syntectonic, intermontane basins. Based on palynology, there appear to be two age groups: an older group, as old as Early Eocene and possibly even Late Paleocene (Otto Fiord, Emma Fiord, Franklin Pierce, Judge Daly, and Lake Hazen basins), and the younger Axel Heiberg Basin of Middle Eocene age. Note that McIntyre (1991b) does not discount extending the age of Axel Heiberg Basin into the Late Eocene.

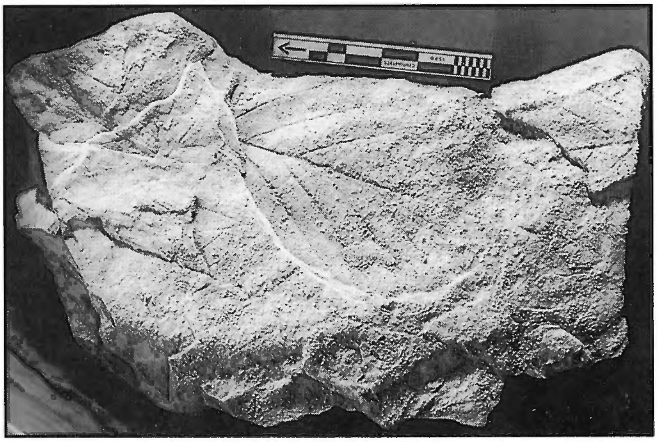

Figure 68. *Leaf impressions in thin sandstone lenses, Buchanan Lake Formation, Franklin Pierce Bay. Birch and oak forms are the most common. All organic material has been oxidized. ISPG 2293-3*

Two types of synorogenic basins are also present: basins that developed in front of major thrusts and high-angle reverse faults (Axel Heiberg, Franklin Pierce-Judge Daly, Lake Hazen), and small basins that are apparently associated with extensional grabens or half grabens (Otto Fiord, and possibly Emma Fiord basins).

The Buchanan Lake Formation on Axel Heiberg and Ellesmere islands is entirely nonmarine. Styles of sedimentation are broadly similar in each basin, wherein alluvial fan, braidplain, and meanderplain facies of various kinds are recognized. Facies that are most proximal are conglomeratic. Facies types, cyclicity, grain size, and compositional unroofing trends all indicate rapid erosion of faulted, high relief terrains, and concomitant rapid sedimentation.

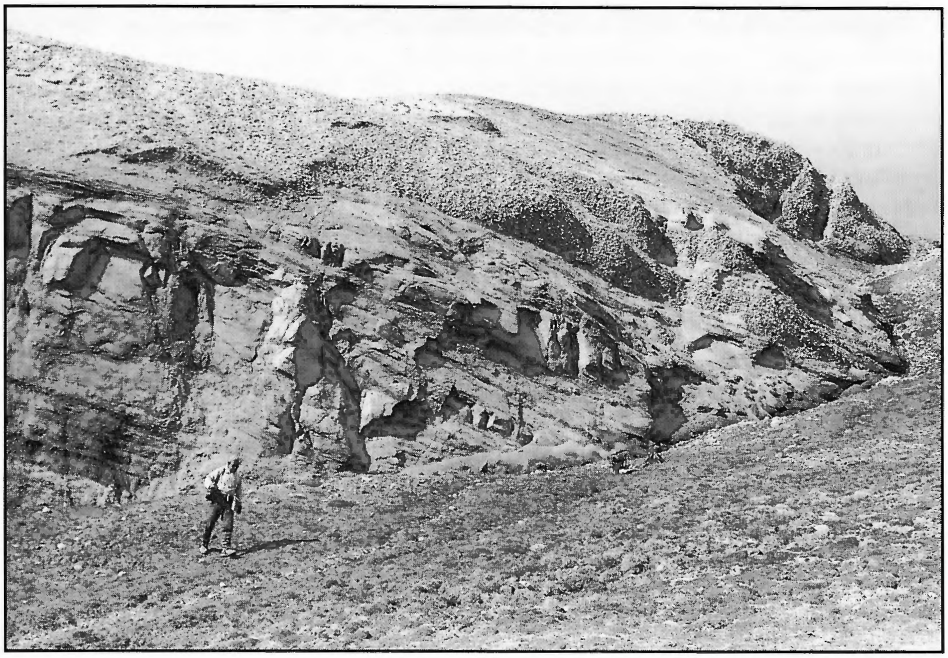

Figure 69.

Interfingering of sandstone and conglomerate in the south Boulder Hills area located between the 87-105 m levels in section RAK 14-88. This corresponds to the transition from sandy braided to gravelly braided channels. ISPG 3070-69

All of the basins were probably separate entities. All developed in geologically short time spans, in less than 10 million years.

STRATIGRAPHIC SEQUENCES AND PALEOGEOGRAPHY

Preamble

Stratigraphic sequences are the building blocks of sedimentary basins. How we define sequence boundaries, whether by unconformities and their correlative conformities (e.g. Vail et al., 1977), or using other significant boundaries such as maximum flood surfaces (e.g. Galloway, 1989), depends on utility and an *a priori* perception of stratigraphic succession. The importance of sequences in stratigraphic analysis is not new. Blackwelder (1909), recognizing the significance of unconformities and their correlative "...continuous piles of marine strata..." presaged later definitions of Sloss (1963) and Vail et al. (1977). Others have referred to sequences variously as 'cycles', or 'transgressive-regressive (T-R) cycles' (e.g. Embry and Osadetz, 1988).

In this presentation, (third-order) sequence boundaries have been placed at major unconformities, subaerial where sufficient data permits such an interpretation, and in the case of marine successions, at the base of regionally extensive transgressive units (identified by regional mapping). Third-order sequences, equivalent to T-R cycles elsewhere in Sverdrup Basin (Embry and Osadetz, 1988) are delineated in Figures 6 through 9 and reconstructed in Figure 70a. Five third-order sequences are recognized. Figure 70b shows the relationship among sequences, lithofacies assemblages, lithostratigraphy and relative sea level. I have generally avoided the use of the term 'systems tract' because of the inherent implication that such depositional entities owe their existence to eustatic changes in sea level, whereas in most geological situations the clear separation of eustasy, tectonism, and sediment flux is not possible.

Sequence 1: Cenomanian to Maastrichtian

This, the oldest depositional sequence in the Eureka Sound Group includes all of the Kanguk Formation and the Lower member, Expedition Formation. At Strand Fiord the lower boundary is a disconformable contact with the Strand Fiord volcanics which in this area have a significant subaerial volcaniclastic component, (Ricketts et al., 1985) and therefore the disconformity is conceivably subaerial. The upper boundary is the sub-Paleocene (subaerial) unconformity.

Sequence 1 is also well exposed on north Fosheim Peninsula, but differs in its thickness and facies. At Remus Creek an unknown thickness of the sequence has been eroded below the sub-Paleocene unconformity. East and north of Fosheim Peninsula, Sequence 1 either never accumulated or was completely removed by erosion below the sub-Paleocene unconformity.

The paleogeographic limits of Sequence 1 are those of Sverdrup Basin (Fig. 71). The transgressive component occurs near the base of the Kanguk Formation where, on Kanguk Peninsula Wall (1991) describes a diverse shelf assemblage of foraminifera. The age range of the transgressive shale is probably Turonian to early Santonian (see also Embry and Osadetz, 1988). Prodelta shale in the remainder of the Kanguk Formation, together with wave-dominated deltas that accumulated in the northwest part of Sverdrup Basin during the middle and upper Campanian, comprise the progradational lithofacies assemblage. Relative highstand delta plain and coastal plain deposits also accumulated along the northern, southern, and eastern basin margins. Delta progradation was to the west and southwest.

Sequence 1 at Strand Fiord can be divided into two higher-order sequences (fourth-order). In the Lower member, contact between the fourth-order sequences is placed at the base of a green-weathering sandstone interpreted as the transgressive component of the succeeding fourth-order sequence. A significant unit of shale overlies the transgressive sandstone and corresponds to a flooding surface (between units A and B, Fig. 10a). Basal strata immediately below the prodelta shale contain concretionary ironstone beds that commonly have *Inoceramus* fragments and leaf impressions and comprise the condensed component of the sequence.

Sequence 2: Early Paleocene

Over most of the region, erosion below the sub-Paleocene unconformity removed all or some of the Maastrichtian and possibly some upper Campanian deposits of Sequence 1. This unconformity is the lower boundary of Sequence 2 and represents a significant change in sedimentation patterns and basin limits. Pebbles of dacite, probably derived from centres situated between Emma Fiord and Phillips Inlet, were widely distributed throughout Fosheim Peninsula, May Point, and even Strand Fiord, and occur immediately above the sub-Paleocene unconformity; the dacite pebbles are not found at any other stratigraphic level. Subaerial erosion of Sequence 1 strata produced a low-relief valley and ridge topography, onlapped by transgressive estuarine deposits (Fig. 72a). Distribution of the volcanic pebbles is inferred to have taken place during the relative sea level lowstand, in earliest Paleocene and perhaps latest Maastrichtian time. Sequence 2 (Upper member, Expedition Formation) onlapped several tens of kilometres beyond the depositional limits of Sequence 1. These changes in basin configuration also correspond to changes in subsidence rate and plate kinematics (see Synthesis). From Cañon Fiord to south of Strathcona Fiord the base of Sequence 2 is a profound unconformity, below which is Lower Paleozoic bedrock.

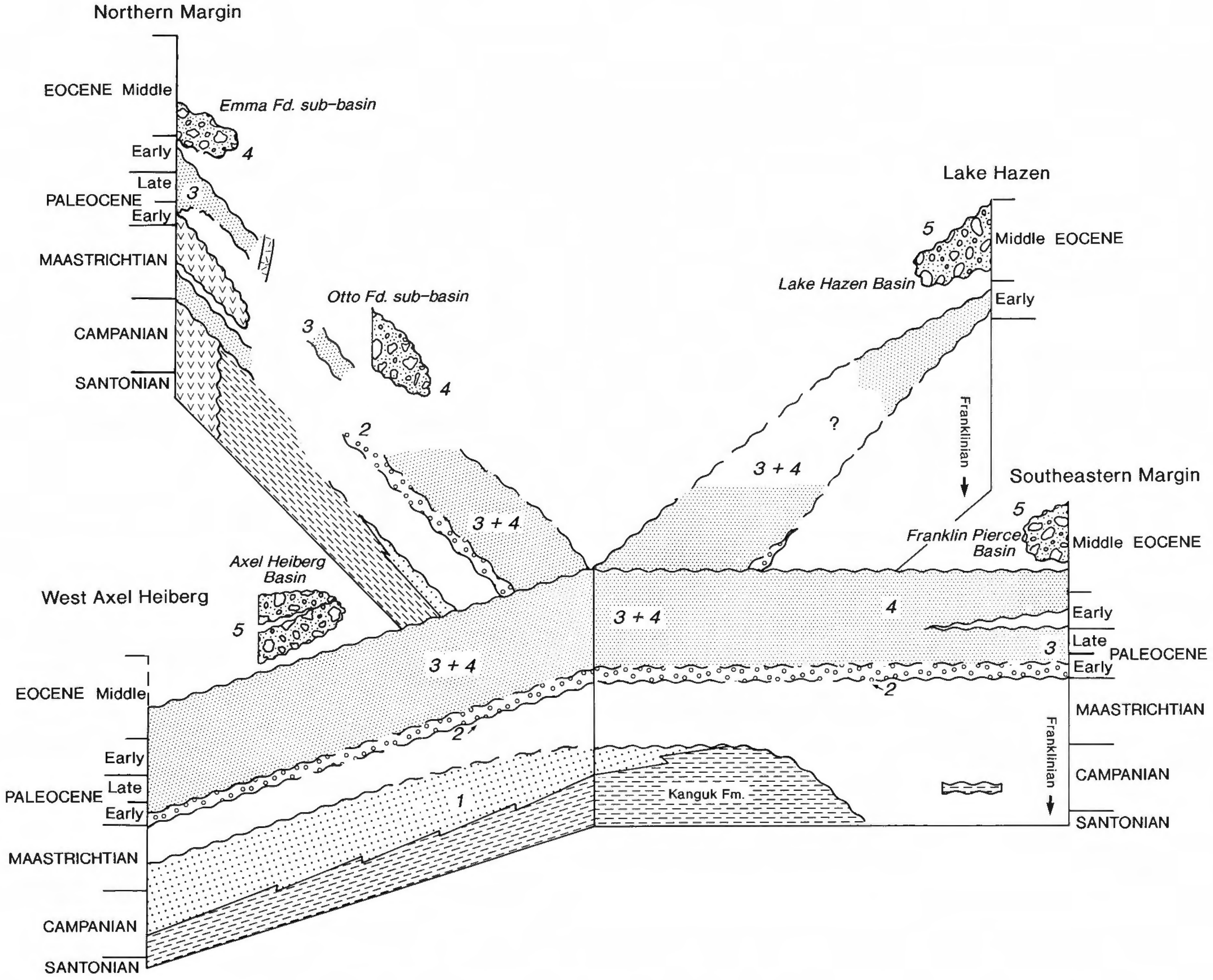

Figure 70. ***a)*** *Third-order sequence stratigraphy of the Eureka Sound Group in the Sverdrup and intermontane basins, summarized from Figures 6 to 9. Numbers refer to sequences in the text.*

In all areas the lower sequence boundary and the basal contact of the Upper member coincide. This is not the case for the Sequence 2 upper boundary, which on west Axel Heiberg and Ellesmere islands is placed at the base of a transgressive sandstone unit within the Expedition Formation Upper member itself.

Sequence 2 signifies a substantial metamorphosis in basin configuration and paleogeography compared to Sequence 1 (Fig. 72a, b). Coastal plain, drowned valley estuary, barred estuary, and associated sandy shelf environments (Fig. 70b) record the initial response to Early Paleocene transgression in central Ellesmere Island and eastern Axel Heiberg Island. Wave-dominated deltas persisted during the subsequent relative highstand in the west Axel Heiberg region. Volcanism continued along the northern margin of Sverdrup Basin, but magmas were much more differentiated (alkaline felsic and trachytic) than during previous volcanic episodes (Mueck et al., 1990).

Sequence 3: mid- to Late Paleocene

Sverdrup Basin continued to expand and deepen during the Paleocene. Regional transgression in the mid-Paleocene is recorded in several places by retrogradational facies at the top of the Expedition Formation (Fig. 70b, 73). The disconformity at the base of Sequence 3 is probably subaerial in most places. If any relative sea level lowstand wedge exists it is probably offshore beneath the modern polar continental shelf. Subsequent regression and progradation of muddy shelf and delta bodies during the early relative highstand stage produced the Strand Bay Formation, a shale unit of possible Arctic-wide extent, plus the Iceberg Bay Formation Lower member (river-dominated deltas, west Axel Heiberg Island) and Cape Pillsbury Member shelf deposits (Cañon Fiord to Bay Fiord). Contact between the Expedition and Strand Bay formations corresponds to the surface of maximum flooding in Sequence 3.

Contact between sequences 3 and 4 is found in three main areas. In the Bay Fiord area the contact is placed at the Cape Pillsbury/Braskeruds member contact. A significant unconformity also occurs between Buchanan Lake Formation strata of probable Early Eocene age at Otto and Emma fiords, and Iceberg Bay beds, the latter being included in Sequence 4. Elsewhere in Sverdrup Basin, no clear distinction can be made between sequences 3 and 4 (most of the region between Strand Fiord and Fosheim Peninsula), and in Figure 70a are designated '3 & 4 undivided'. The paleogeographic limits to the late stage of the Sequence 3 relative highstand are shown on (Fig. 74). Sedimentation on west Axel Heiberg Island produced the thickest succession of delta facies known in the Eureka Sound Group. In comparison, relatively shallow shelf conditions prevailed in the eastern sector of the Basin (Cape Pillsbury Member).

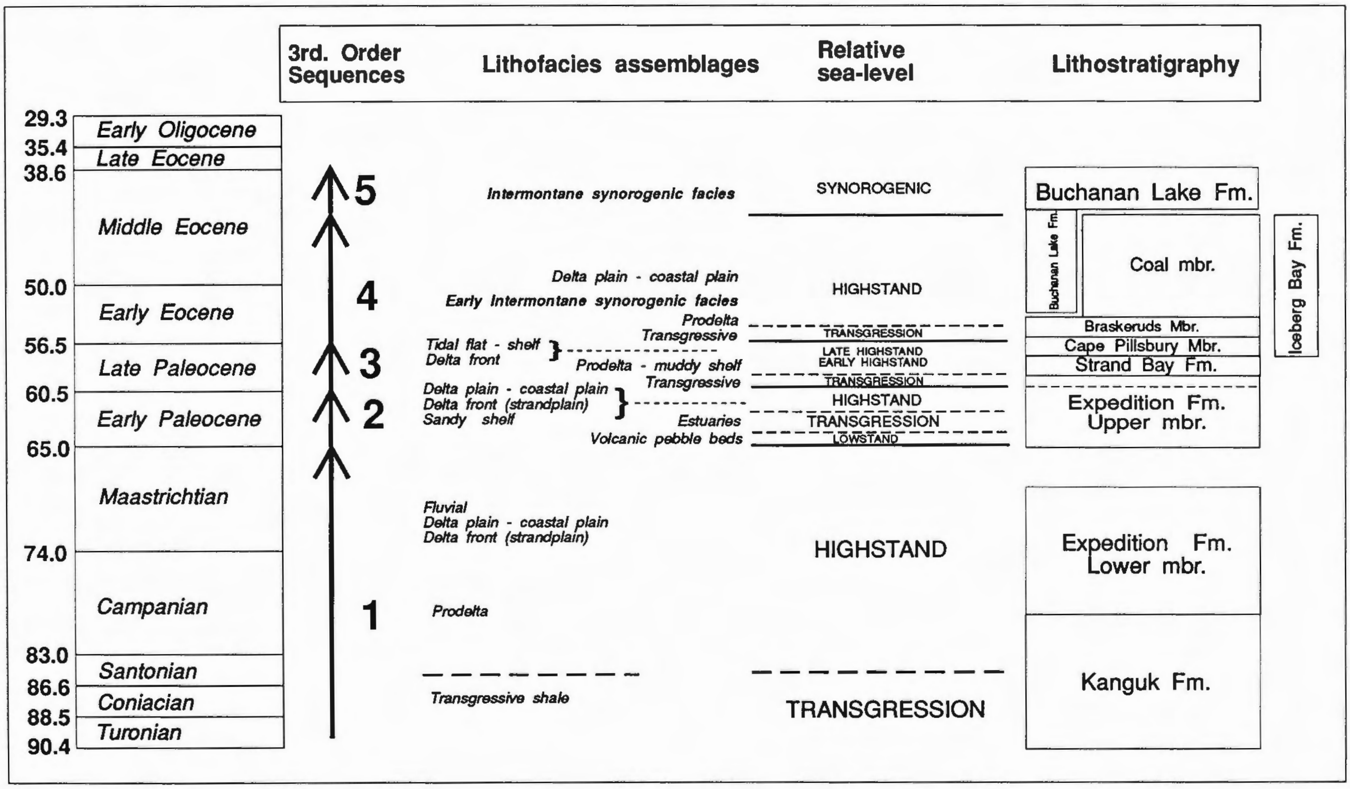

Figure 70. b) Relationship between third-order sequences, depositional facies assemblages, relative sea level and formal lithostratigraphy in the Eureka Sound Group and Kanguk Formation.

The transgression that gave rise to Sequence 3 was widespread. Definitive coastal facies are known only in sandy deposits at the base of the Iceberg Bay Formation near Remus Creek and Fosheim Anticline. Proximity to a shoreline on the eastern margin of Sverdrup Basin is suggested by the presence of one or two thin coaly beds in the upper part of the Strand Bay Formation near Mount Moore. However, during early Strand Bay deposition (early highstand stage) it is possible that a temporary sea link existed between Sverdrup Basin and ancestral Baffin Bay from which the earliest marine incursions originated. Evidence in support of this hypothesis is derived from the presence of a unique calcareous, benthic foraminifera assemblage discovered in basal Cape Pillsbury strata. In addition to indicating open marine conditions, the foraminifera assemblage exhibits some affinities with Atlantic assemblages. Nevertheless, this tentative connection was short-lived because strata in the Cape Pillsbury succession show progressive shoaling trends.

Sequence 4: Late Paleocene to Middle Eocene

Sequence 4 is best defined in the Bay Fiord area, and at Emma and Otto fiords. At Bay Fiord, the lower boundary is an abrupt contact between the Cape Pillsbury Member and Braskeruds Member (Iceberg Bay Formation). A basal, fine grained, green-grey sandstone at the base of the Braskeruds

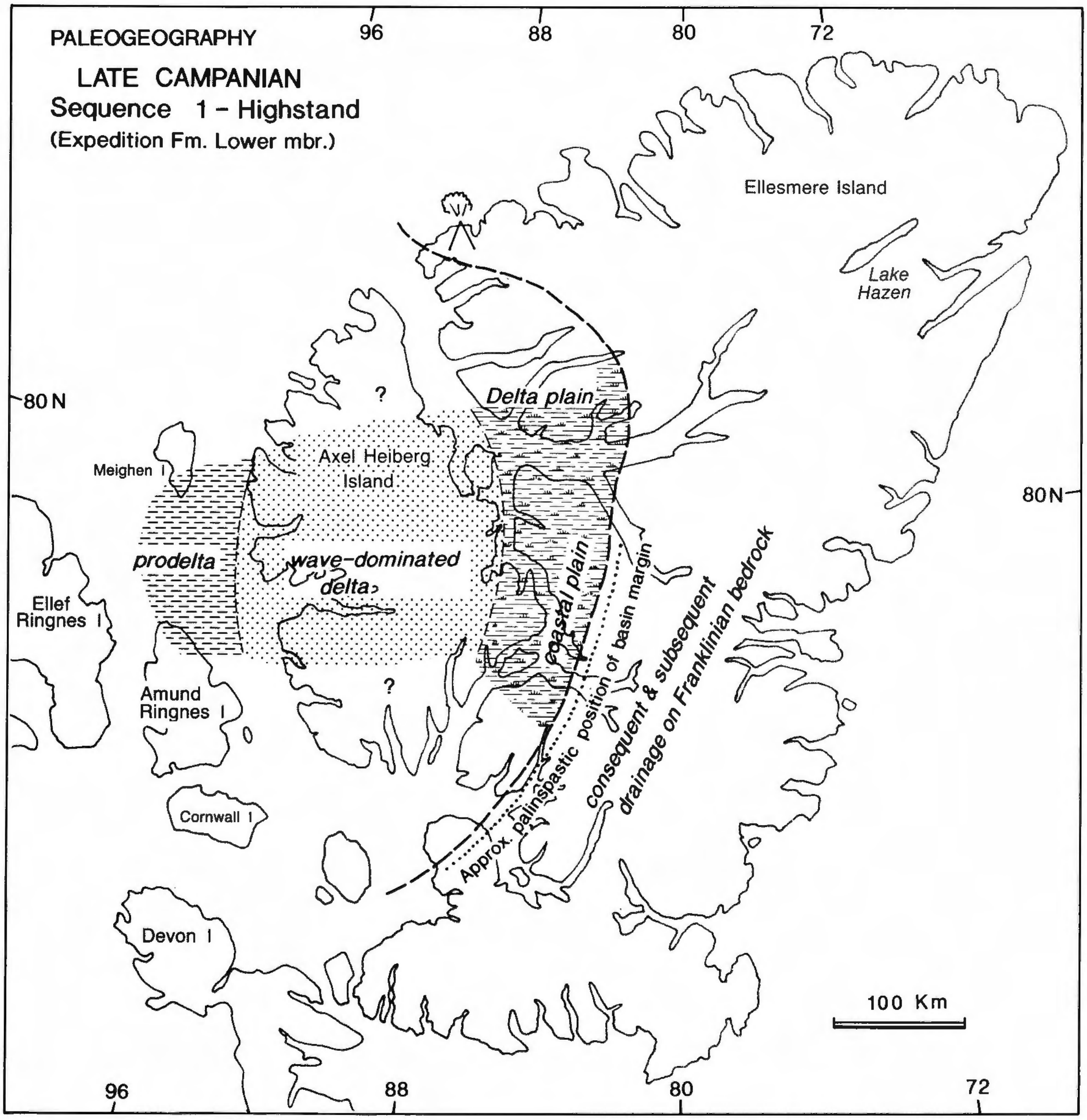

Figure 71. Paleogeography map for the upper Campanian (Sequence 1), Eureka Sound Group. The southern and northern limits to the relative highstand wave-dominated delta system are unknown. The approximate palinspastic position of the eastern basin margin in this and subsequent figures is based on preliminary reconstruction along thrust faults on Fosheim Peninsula and east Axel Heiberg Island (Ricketts, 1987b).

Member is the transgressive component of the sequence (exposed on the north and south shores of Strathcona Fiord). Between Strathcona and Bay fiords this contact is probably equivalent to a conglomerate bed below the Coal member. Shale making up most of the Braskeruds Member is of prodelta origin, and was succeeded by delta front and delta plain sandstone, mudstone, and coal of the Coal member, all components making up the highstand succession. Elsewhere on Axel Heiberg and central Ellesmere islands the bulk of Sequence 4 consists of the Coal member (Iceberg Bay Formation), but cannot be distinguished with confidence from Sequence 3.

Sequence 4 corresponds to a time of substantial change in Sverdrup Basin. Much of the basin continued to subside and expand rapidly during the Eocene (Fig. 75). In most areas except Otto and Emma fiords, Sequence 4 is represented by the Coal member (Iceberg Bay Formation; including outliers at Hvitland Peninsula, Emma Fiord, Phillips Inlet, Lake Hazen, and Judge Daly Promontory). By Middle Eocene time Sverdrup Basin had attained its greatest areal extent, the foredeep having filled to sea level. Upper delta plain and coastal plain environments persisted throughout the area; delta front and prodelta settings are presumed to be preserved offshore beneath the modern polar continental shelf.

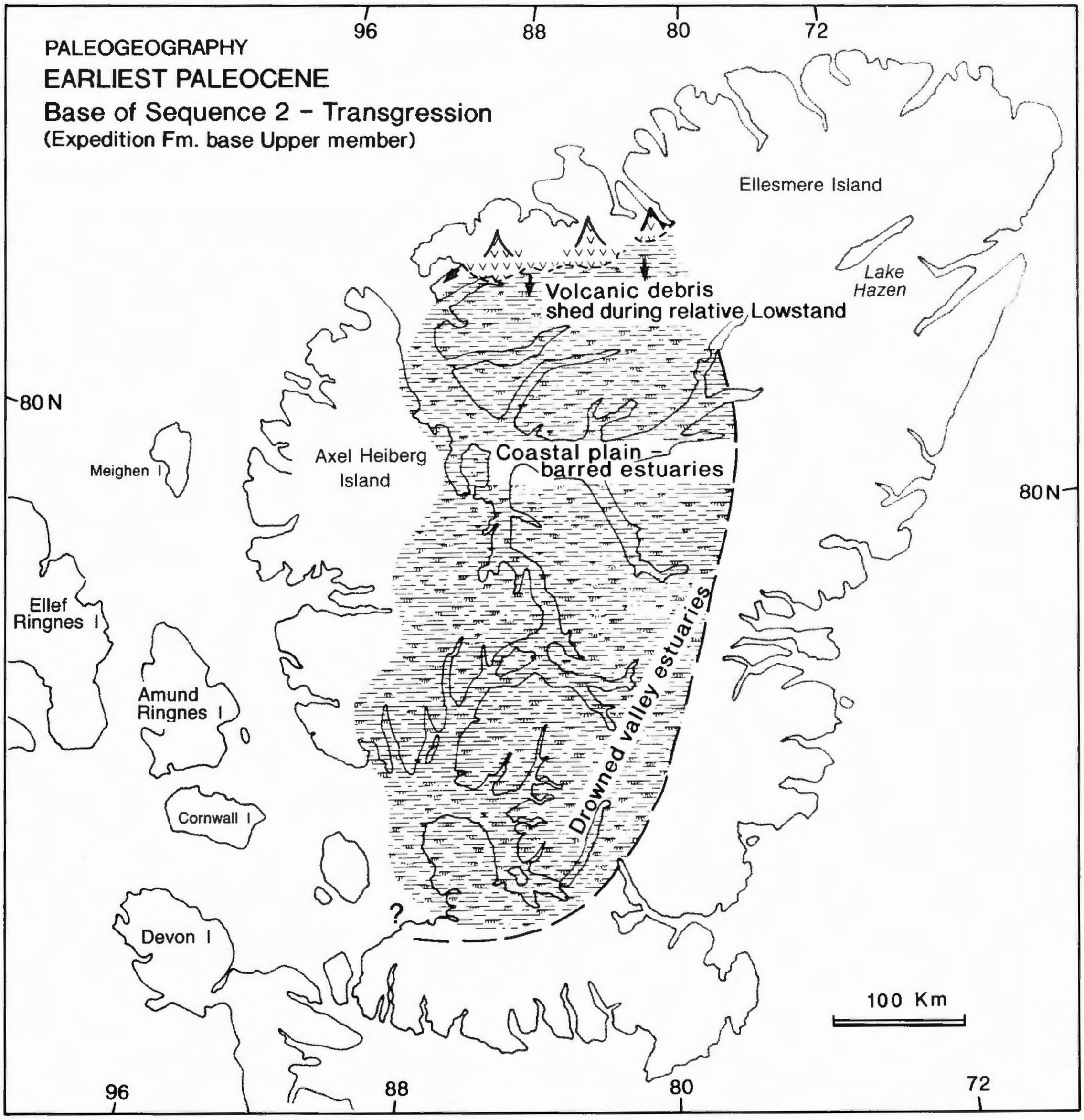

Figure 72. a) Paleogeography map for the earliest Paleocene (base of Sequence 2), Eureka Sound Group. The estuarine facies comprises a transgressive succession. Volcanic debris above the Sequence 2 boundary was probably shed from the northern basin margin during the relative lowstand stage of Sequence 2.

Nevertheless, while the foredeep had apparently attained maximum levels of subsidence, a profoundly different style of basin formation had emerged at Otto and Emma fiords and the Lake Hazen area (Fig. 75). Here, contact between the Buchanan Lake and Iceberg Bay formations is a subaerial unconformity. Coarse grained Buchanan Lake Formation deposits accumulated in the relatively small Otto Fiord and Emma Fiord fault-bounded subbasins, and initially in Lake Hazen Basin, and these are the first indications of crustal failure of the antecedent Sverdrup Basin. Otto Fiord subbasin in particular is interpreted as a small half graben, suggesting local extension within an overall contractional orogen. The eastern margin of Otto Fiord subbasin is defined by a normal fault that appears to be the continuation of Black Top Fault on Fosheim Peninsula. Deposition in Lake Hazen Basin, and more equivocally Franklin Pierce Basin, indicates that tectonism had also begun on eastern Ellesmere Island in Early Eocene time; this region was to become the main locus of deformation later in the Eocene.

Sequence 5: Middle Eocene

A profound unconformity exists at the top of the Iceberg Bay Formation, which records the demise of Sverdrup Basin and the main phase of sedimentation in at least six synorogenic intermontane basins (Fig. 76). Three of these basins (Otto Fiord, Emma Fiord, and possibly Lake Hazen basins) were initiated during Sequence 4. This, the main phase of Eurekan deformation, produced major thrusts and folds that fragmented the contiguous Paleocene to Middle Eocene foredeep of Sverdrup Basin. It is represented by the conglomeratic Buchanan Lake Formation.

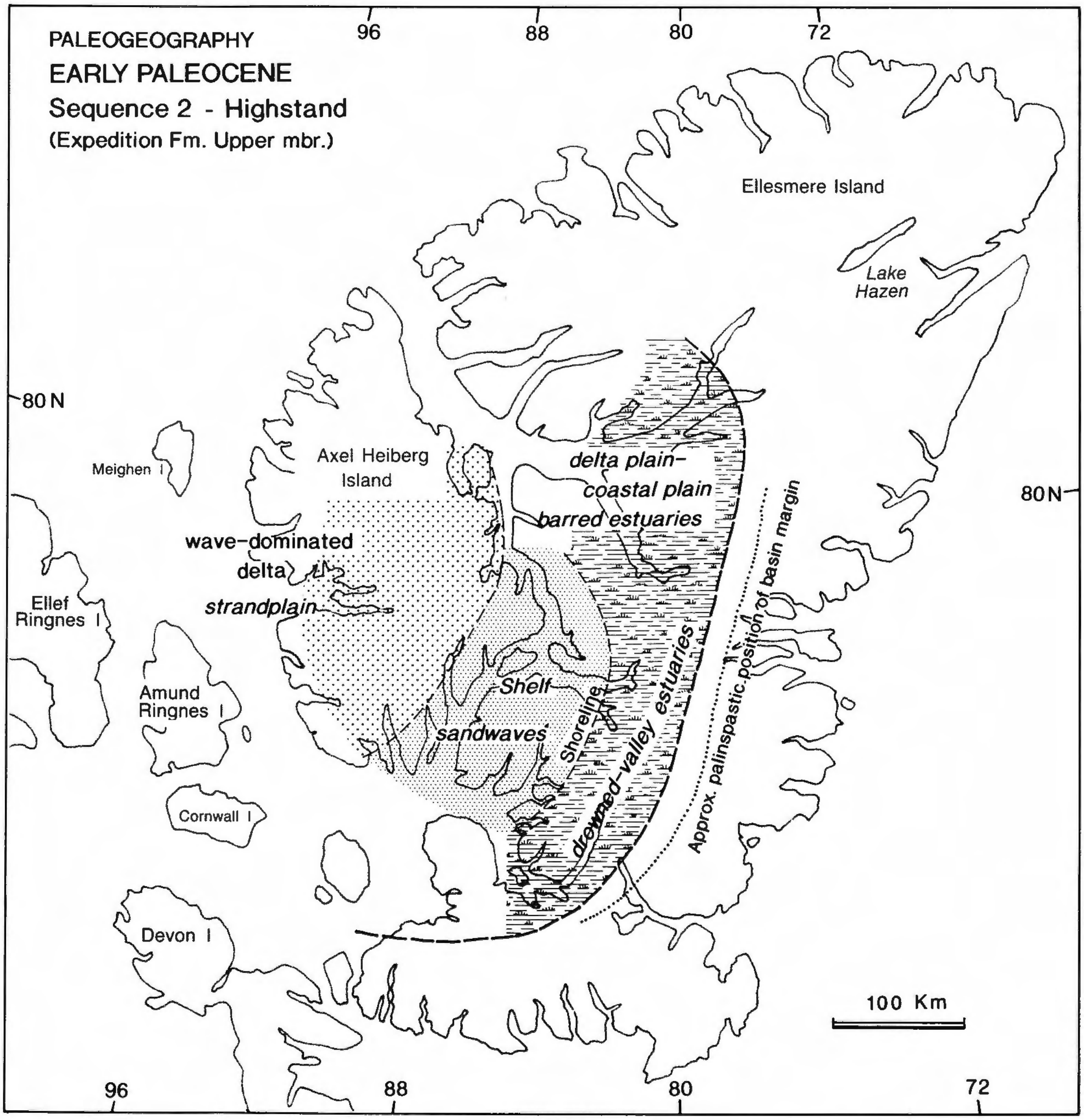

Figure 72. b) *Paleogeography map for the Early Paleocene (Sequence 2), Eureka Sound Group, with facies comprising a relative highstand succession.*

The lower sequence boundary ranges from a disconformity to an angular unconformity with discordances up to 50° (e.g., Schei Peninsula). Stratigraphic units below the unconformity are also highly variable, ranging from the Iceberg Bay Formation (as young as Middle Eocene) to the Lower Triassic Blaa Mountain Formation and Blind Fiord Formation. In most areas the upper sequence boundary has been eroded; at only one site – east of Geodetic Hills, are Buchanan Lake strata overlain by possible Neogene deposits (J. Fyles, pers. comm., 1991).

Analysis of each of the basins reveals a common suite of facies representing humid and arid alluvial fans, braidplains, and forested meanderplains. In Axel Heiberg, Lake Hazen, and Franklin Pierce/Judge Daly basins, sediment was derived from large, steeply dipping thrusts. Axel Heiberg Basin was initiated in the Middle Eocene. Lake Hazen Basin may have begun earlier in the Eocene although the timing, based on palynology, is less precise than that in Axel Heiberg Basin. Dating of Buchanan Lake rocks in both Franklin Pierce and Judge Daly basins is also equivocal, although a Paleocene age is often quoted by other workers (Mayr and de Vries, 1982; Miall, 1982). Clearly, the Buchanan Lake Formation is diachronous and the synorogenic intermontane basins formed at different stages of Eurekan tectonism, beginning in the Early Eocene or possibly latest Paleocene of Sequence 4, and ending in the Middle Eocene Sequence 5 (compare Fig. 75 and 76).

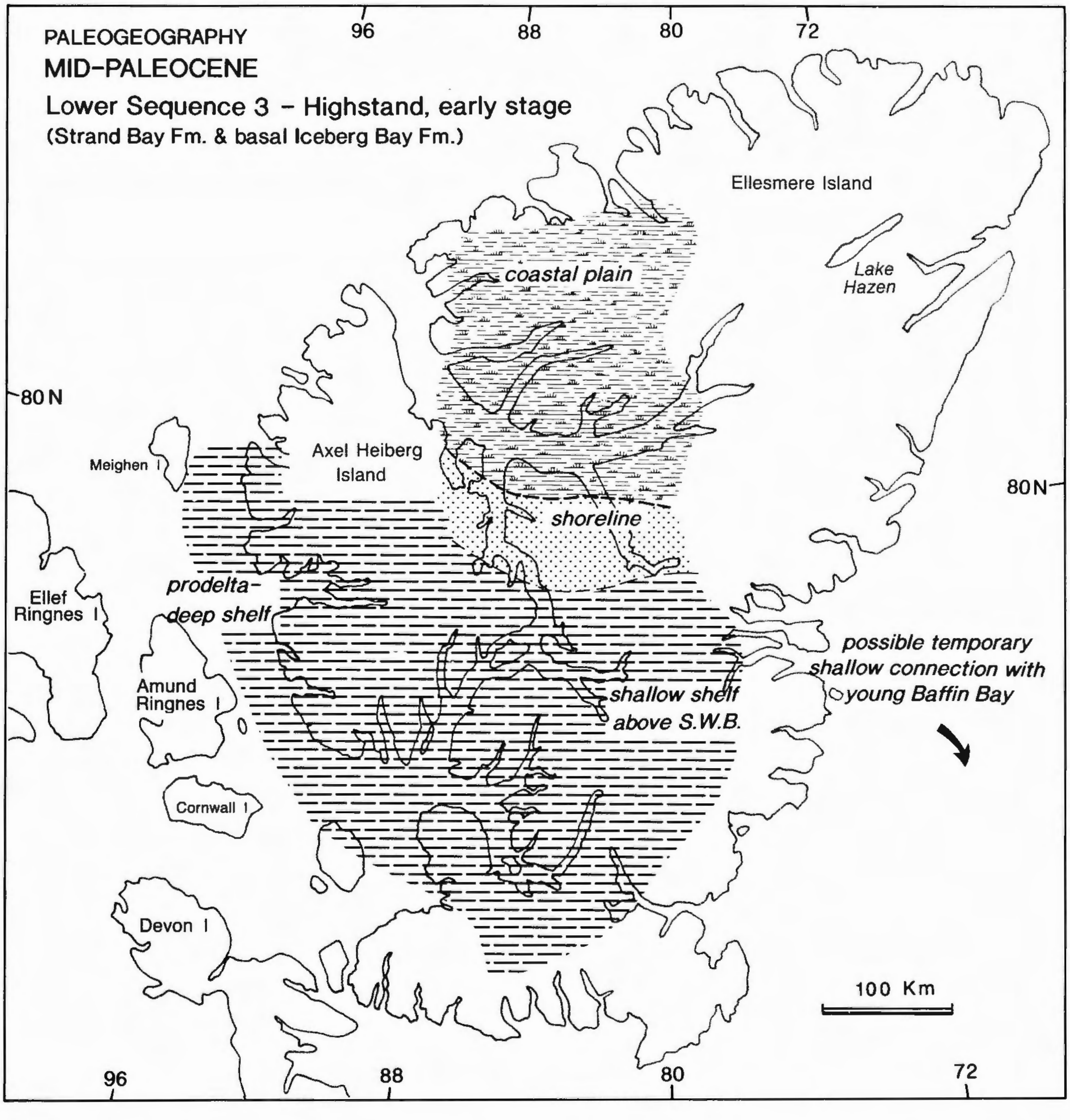

Figure 73. Paleogeography map for the mid-Paleocene (lower part of Sequence 3), Eureka Sound Group, during the early stage of relative highstand progradation. The preceding transgressive unit is identified at the top of the Expedition Formation. A possible connection with proto-Baffin Bay is inferred.

Sequence 5 can be subdivided into two higher-order sequences in Axel Heiberg Basin (Fig. 6, 70a; also Ricketts, 1991b). Each subsequence records a major phase of fault uplift, presumably thrusting, and subsequent alluvial fan to meanderplain progradation. Overall fining-upward trends in each subsequence indicate erosional retreat of the source rocks in the hanging wall.

EUREKAN STRUCTURAL GEOLOGY

Regional evaluation of Eurekan structures by Balkwill (1978) and Balkwill and Bustin (1980) emphasized the character of three principal structural types (see Fig. 76): (a) folds and thrusts that deformed Sverdrup Basin rocks, reactivated Ellesmerian structures, and possibly involved basement (e.g. Balkwill, 1983a, b); (b) large crustal-scale arches; and (c) halokinetic structures.

Faults and folds

Structures in the Axel Heiberg/Ellesmere island region are characterized by steeply dipping thrust faults and associated folds. On Fosheim Peninsula prominent mountain ranges formed in the upper panel of the southeast-verging Vesle Fiord and Cañon Fiord thrusts. North of Greely Fiord, northwest-verging (Blue Mountain Thrust) and southeast-verging

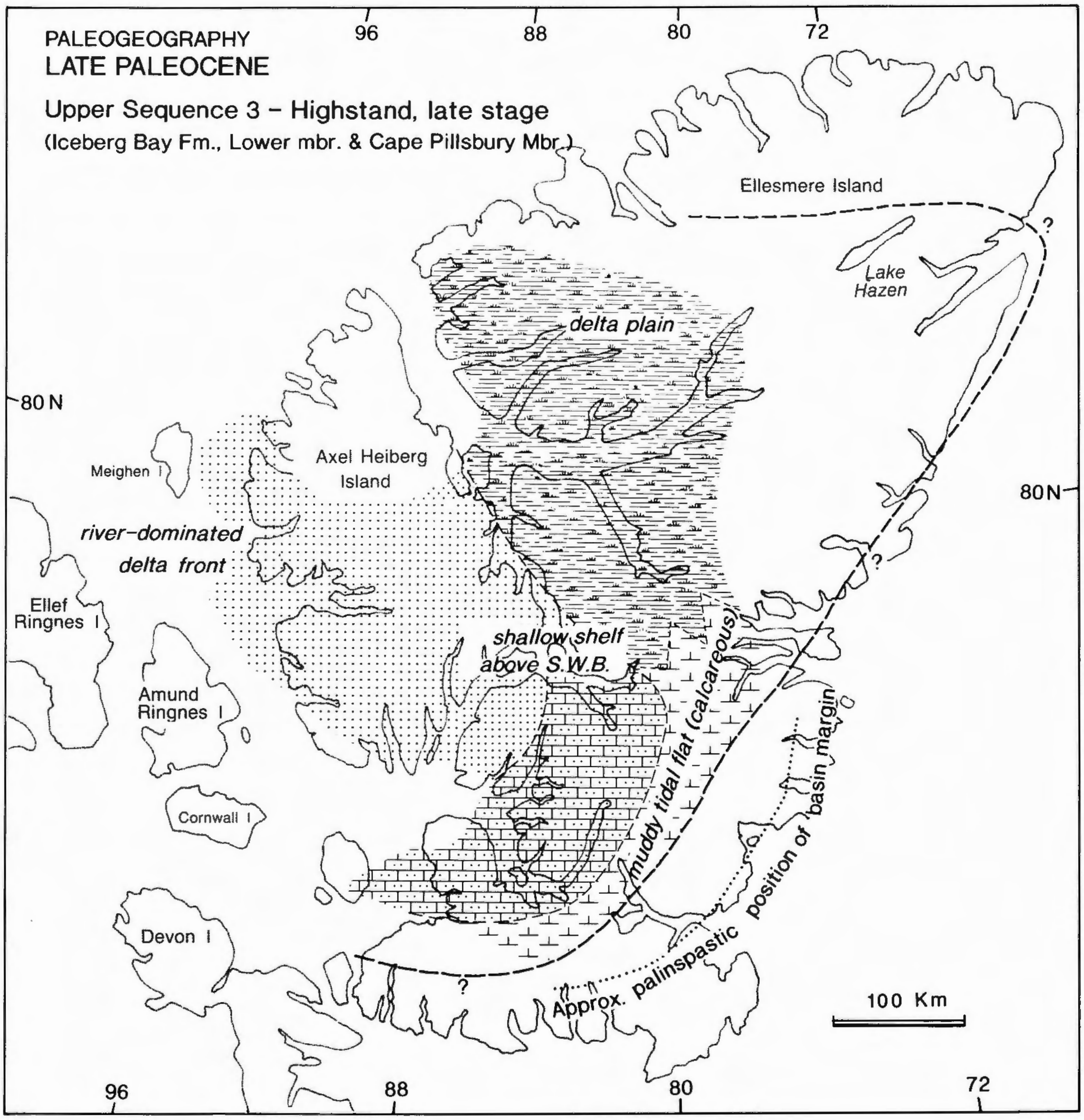

Figure 74. Paleogeography map for the Late Paleocene Eureka Sound Group (upper part of Sequence 3) during the late stage of relative highstand progradation. Extension of the basin limits to northern Ellesmere Island is based on isolated outcrops of Iceberg Bay Formation along Judge Daly Promontory (Miall, 1979), and at Lake Hazen. SWB = storm wave-base.

thrusts are present in addition to along-strike changes in thrust vergence. Pop-up anticlines resulting from the forward- and backward-verging thrusts may be similar to structures described by Harrison and Bally (1988) on Melville Island. Such structures usually indicate a shallow detachment. Potential detachments for the thrusts include Upper Paleozoic (Sverdrup Basin) evaporite units (Mount Bayley Formation – Lower Permian; Otto Fiord Formation – Upper Carboniferous; Balkwill, 1978; Ricketts, 1987b). Thrusts propagated through thick but mechanically weak Lower Triassic shale formations. Major detachments were also located in evaporites of the Lower Ordovician Baumann Fiord Formation. However, the geometry of the detachments in the subsurface is unknown.

Three major high-angle thrusts in the Axel Heiberg/ Ellesmere island region coincide with significant tectonostratigraphic boundaries. Parrish Glacier Thrust, with a strike length greater than 200 km, extends from southern Fosheim Peninsula north to Dobbin Bay. This major structure has the distinction of marking the boundary between the Central Ellesmere Fold Belt and the relatively undeformed (Lower Paleozoic) Interior Platform to the east (Thorsteinsson, 1972b; Trettin, 1989). It contains one of the few known examples of Eurekan thrust flats, determined from klippe of Paleozoic limestone that overlie Paleogene conglomerate (Buchanan Lake Formation) in the Franklin Pierce Bay area.

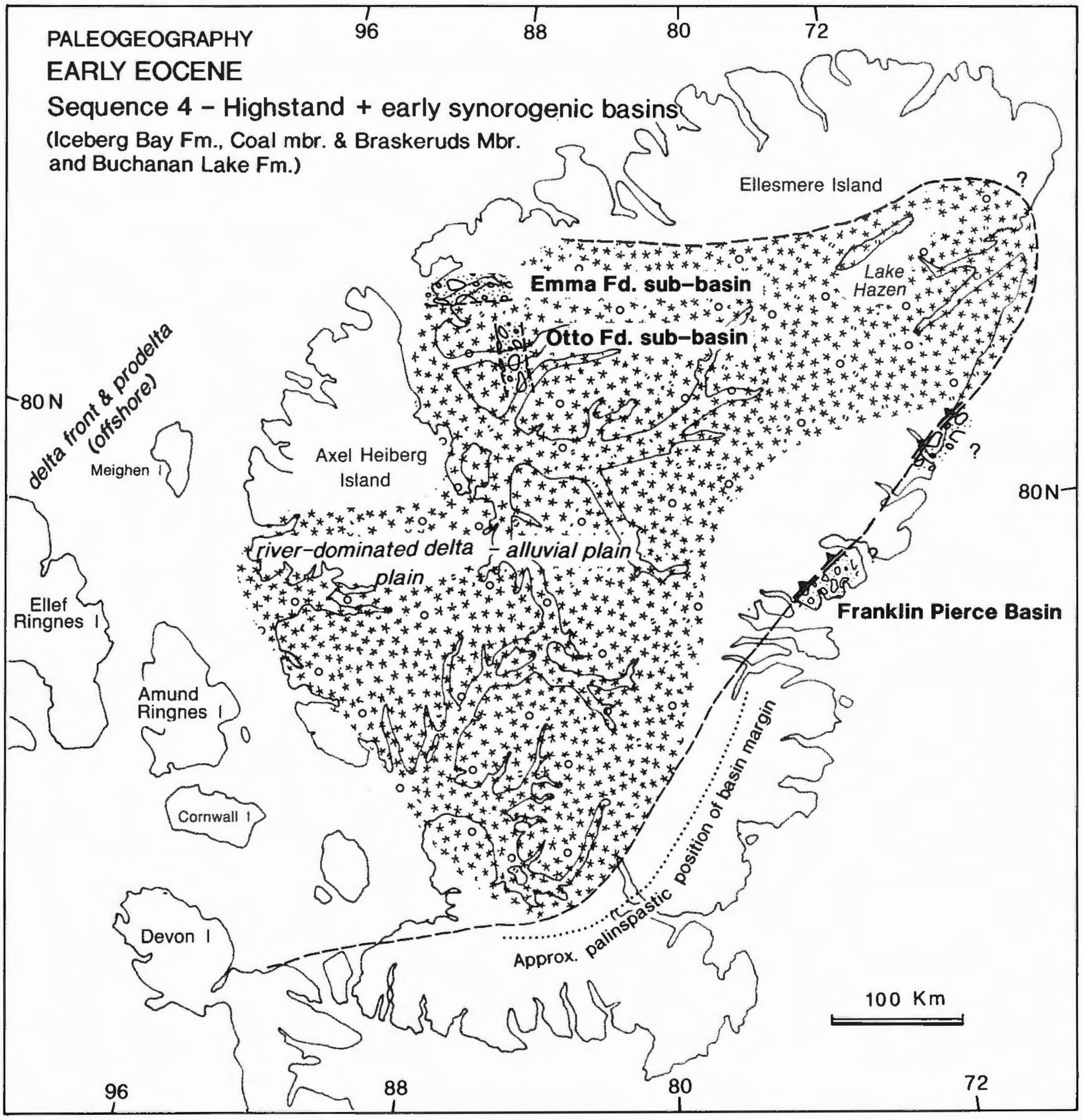

Figure 75. Paleogeography map for the Early Eocene Eureka Sound Group (Sequence 4). This, the final phase of deposition in Sverdrup Basin, represents the relative highstand stage of Sequence 4 and is coeval with the earliest stage of syntectonic, intermontane basin formation; a precursor to the main phase of Eurekan contractional deformation (Sequence 5).

Lake Hazen Fault Zone has a similar strike length to Parrish Glacier Thrust, and contains thrusts that in the Lake Hazen area carry strata of the Franklinian Mobile Belt over rocks as young as the Buchanan Lake Formation. Klaper (1990) recommends that the thrusts originated in the Paleozoic, and were reactivated during Eurekan tectonism but with little additional shortening. Controversy exists over the geometry of Lake Hazen Thrust at depth. Based on fold and fault geometry determined in outcrop, Osadetz (1982) argued that the thrust and associated Grantland Uplift has a thin-skinned structural style with a detachment that dips toward the orogen at a shallow angle. An alternative proposal by Higgins and Soper (1983) portrays the fault planes dipping steeply in most places, and lacking shallow detachments. They infer that the steep upthrusts also involve basement and possibly a component of transpression. This hypothesis requires very little horizontal displacement compared to the thin-skinned thrust hypothesis.

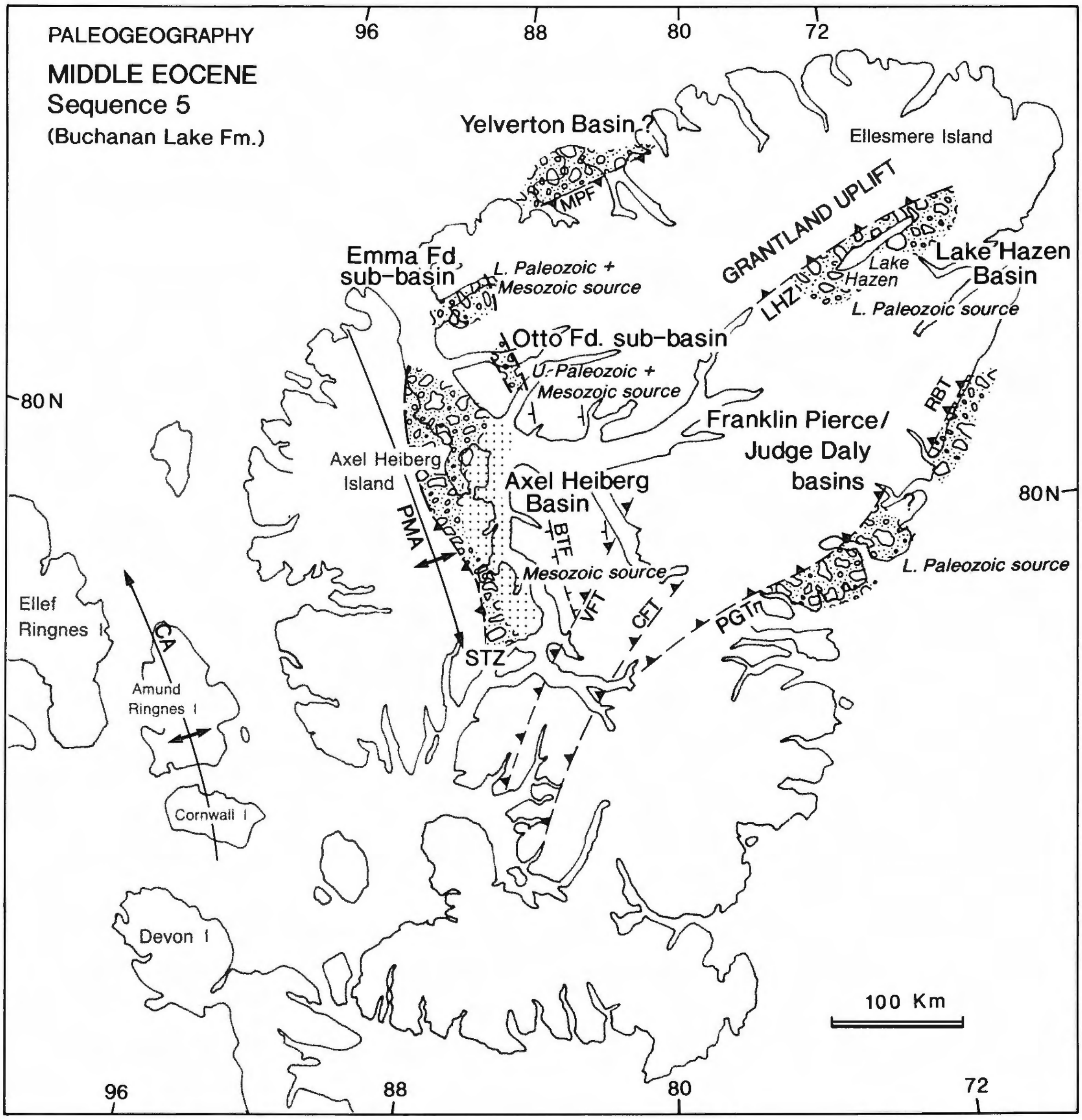

Figure 76. Paleogeography map for the Middle Eocene Eureka Sound Group (Sequence 5). For each basin/subbasin, the predominant source rocks and associated major structures are indicated. For the Axel Heiberg, Lake Hazen, and Franklin Pierce/Judge Daly basins the source rocks are located in the hanging walls of major thrusts: STZ = Stolz Thrust Zone, LHZ = Lake Hazen Thrust zone, PGT = Parrish Glacier Thrust. Other major Eurekan structures include VFT = Vesle Fiord Thrust, CFT = Cañon Fiord Thrust, RBT and MPF = Rawlings Bay Thrust and Mitchell Point Fault respectively (Trettin, 1989), BTF = Black Top Fault. PMA and CA refer to the ancestral Princess Margaret Arch and Cornwall Arch respectively. The existence of Yelverton basin is speculative.

The problems of Lake Hazen Fault geometry in fact apply to most of the larger faults in the Eurekan Orogen – are they steep dipping faults rooted somehow in the basement, or do they belong to a thin-skinned, foreland-style regime? Stolz Fault Zone, which extends along the backbone of Axel Heiberg Island for at least 100 km, is no different in this regard. In most places the Stolz fault is steep, dipping up to 65°. Ricketts (1987b) constructed, in simplistic fashion, a foreland thrust and fold style of geometry with flat detachments in Upper and Lower Paleozoic evaporite units; Upper Paleozoic evaporites are exposed in the hanging walls of several prominent thrusts. However, even if the detachments do have this kind of geometry their lateral extent remains unknown. The structural cross-sections of Ricketts (1987b) are approximately balanced, but only to the inferred detachment in Otto Fiord evaporites. Shortening across the fold belt is estimated at 25-30%. Shortening at the Stolz Fault Zone itself is about 10 km. Depth to crystalline basement determined from stratigraphic thicknesses is 15-18 km and this corresponds reasonably to the boundary determined from gravity data along a transect from Grinnell Peninsula (Devon Island) and Ellef Ringnes Island to the polar continental shelf (Sobczak et al. 1986; Sweeney et al., 1990).

Balkwill (1983a) has employed the analogy of large listric-style faults possibly rooted in basement, and originating during early Sverdrup Basin rifting. These faults, reactivated during the Eurekan Orogeny, were linked to uplift on Princess Margaret and Cornwall arches. It is possible that shallow dipping detachments of limited lateral extent exist within this regime, as depicted schematically in Stephenson et al. (1990, Fig. 3).

Large, north-striking normal fault segments that may be kinematically linked to the thrusts are present on Fosheim Peninsula and the area north of Greely Fiord, an area that coincides with a significant gravity low (see Fig. 76; also Stephenson and Ricketts, 1990, Fig. 3). The Black Top Fault and its lateral equivalents apparently truncate the Blue Mountain Thrust near its termination, whereas about 185 km farther south the fault is truncated by the Vesle Fiord Thrust. In places (e.g., Fosheim Anticline) 300-400 m of stratigraphic displacement is estimated. The normal faults are generally orthogonal to the thrusts. Like the thrusts, the normal faults may have been reactivated from structures that formed during early Sverdrup Basin rifting. Interestingly, one set of normal faults, paralleling the Black Top Fault and extending from Atwood Point (Borup Fiord) north to Hare Fiord, is on trend with a major facies change in the Permian carbonate platform-to-basin transition (Nansen Formation and Hare Fiord Formation), possibly reflecting antecedent structural control (this fault system may also extend south, across Greely Fiord towards Mount Bridgman on northeast Fosheim Peninsula). Furthermore, the normal fault bounding the eastern margin of the syntectonic Otto Fiord subbasin also appears to be the along-strike extension, and has the same sense of displacement as the Black Top Fault.

Arches

Three structural elements having crustal proportions are the Grantland Uplift, Cornwall Arch, and Princess Margaret Arch (see Fig. 76). Grantland Uplift (Trettin, 1971) is a topographic and structural high, whose southern boundary is delineated by Lake Hazen Fault Zone which places rocks as old as Cambrian over Early or Middle Eocene syntectonic conglomerate. Grantland Uplift has a complex history, with uplift during the Early Paleozoic (Grant Land Anticlinorium of Maurel, 1989), and overprinting by Eurekan thrusts and folds (Maurel, 1989; Klaper, 1990).

Cornwall Arch straddles Cornwall and Amund Ringnes islands. The axis of this large wavelength, structurally simple anticline is about 200 km long; Balkwill (1983a) estimated 4-5 km of structural relief. Early investigations bracketed the timing of arch inception as Maastrichtian to possibly Paleocene (Balkwill, 1978). Recently, McIntyre and Ricketts (1989), in reprocessing palynology samples, showed that the age bracket can be extended into the Early Eocene. Although the timing of arch formation remains imprecise, the Eocene age at least permits comparison with the timing established for the Princess Margaret Arch (Ricketts, 1987a; Ricketts and McIntyre, 1986).

Princess Margaret Arch is a more complex structure than its parallel counterpart in the Cornwall Arch. It is some 300 km in axial length and possesses similar structural relief, but has been disrupted by many faults and folds, including the Stolz Fault Zone, the trace of which parallels, but lies east of, the arch axis. The principal period of arch formation was coincident with thrusting and folding during the main period of Eurekan tectonism, which on Axel Heiberg Island was Middle Eocene (Ricketts, 1987a, 1991b).

Acquisition of gravity data from several thousand stations, offshore (on sea ice) and onshore, including in 1987 two detailed transects across the Eurekan Orogen, provides the basis for modelling of large scale crustal features in the central and eastern Arctic Islands (Stephenson and Ricketts, 1989, 1990). Modelling of the three major arches demonstrates:

- A significant positive gravity anomaly coincides with the axis of Cornwall Arch and indicates that the arch is underlain by a crustal/mantle upwarp.
- At the other end of the spectrum, Grantland Uplift is isostatically compensated at depth, indicating tectonic thickening, possibly by thrusting. However, as noted above, the main components of tectonic thickening are probably much older than the Eurekan event.
- Princess Margaret Arch is transitional between these two end members; it exhibits partial isostatic compensation, resulting from tectonic thickening along major thrusts during the Eurekan event (e.g. Stolz Fault Zone), but is cored by a broad, deep crustal upwarp, also of Eurekan origin.

Given the wavelength of Cornwall and Princess Margaret arches at about 200 km, and assuming typical flexural rigidities for continental crust, Stephenson and Ricketts (1990) and Stephenson et al. (1990) have hypothesized that the two structures developed during Eurekan tectonism as crustal-scale folds. The westward decrease in degree of isostatic compensation from Grantland Uplift in the east to Cornwall Arch also corresponds to increasing distance from the zone of maximum Eurekan compression.

Evaporite diapirs

Tectonically and halokinetically emplaced evaporite diapirs are common on the southern half of Axel Heiberg Island and west-central Ellesmere Island. Their origin has been discussed in considerable detail by Gould and de Mille (1964), Balkwill (1978), Schwerdtner and Osadetz (1983), van Berkel (1986), and van Berkel et al. (1984). Most diapirs contain anhydrite and some gypsum; one example north of Whitsunday Bay is cored by halite (Hugon and Schwerdtner, 1982). Halite was also encountered in the Hoodoo Dome on Ellef Ringnes Island (Davies, 1975). At Strand Fiord, skinny evaporite tongues have breached anticlines, forming what van Berkel et al. (1984) have called wall-and-basin structures. These structures probably developed during the main phase of Eurekan folding and faulting (Middle Eocene on Axel Heiberg Island). Diapiric slivers also occur in the surfaces of some Stolz Fault Zone thrusts (e.g., at Whitsunday Bay and Buchanan Lake), but it is not clear whether they were emplaced during thrusting or later squeezed along zones of weakness created by the fault planes. A number of domal structures are also present that are probably cored by evaporite, for example Fosheim Anticline, wherein flanking Expedition Formation beds are cemented by gypsum.

Evaporite rock in the diapirs was derived from the Carboniferous Otto Fiord Formation. There is mounting evidence that evaporite mobilization took place as early as the Upper Triassic, demonstrated by stratigraphic thinning over the flanks of diapirs (Embry, 1982). Movement continued sporadically at least into the post-Middle Eocene because Buchanan Lake conglomerate at Geodetic Hills has been pierced by small diapirs (Ricketts, 1991b).

SYNTHESIS

Strata of the Upper Cretaceous to Paleogene Eureka Sound Group record the break up of Sverdrup Basin, and the transition to syntectonic, intermontane basins. The transitions in basin type in terms of third-order sequences, tectonic events, plate kinematics and subsidence regimes is summarized in Figure 77.

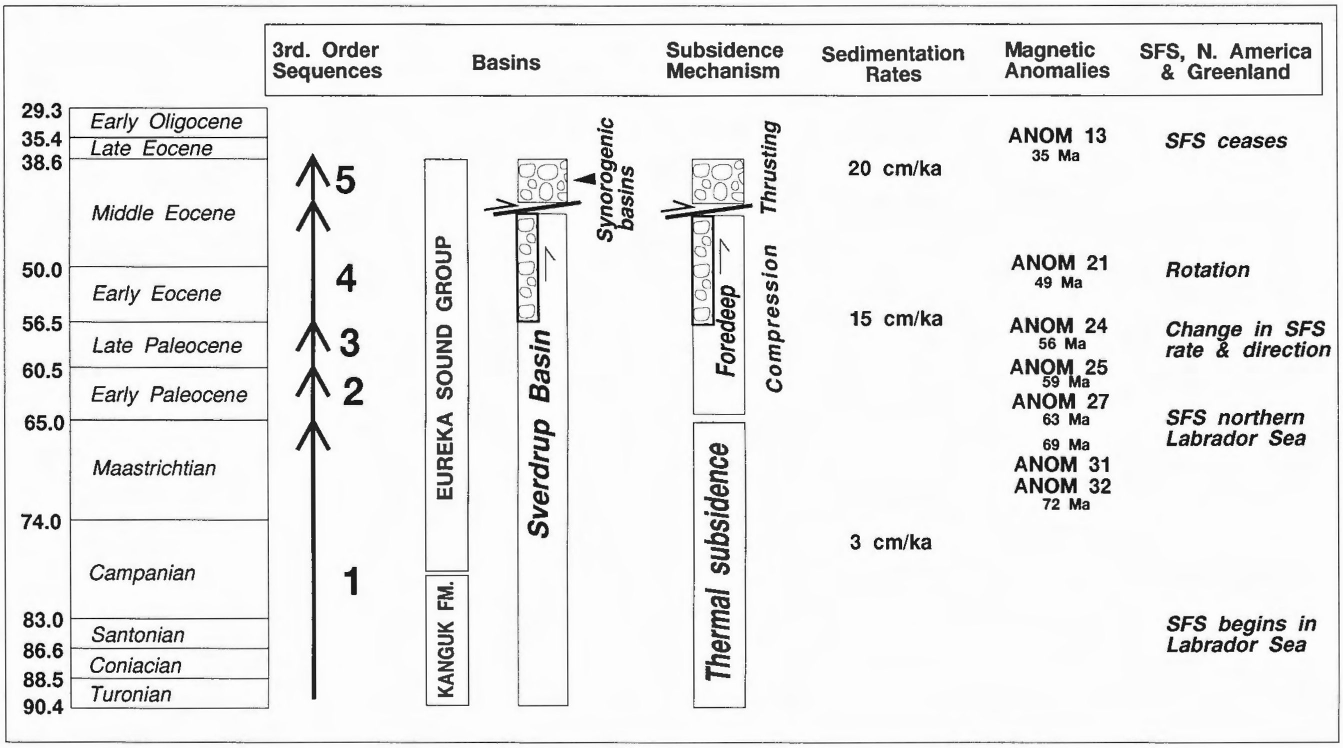

Figure 77. Comparison of third-order sequences, subsidence mechanisms, sedimentation rates, magnetic anomalies and plate tectonic events, for the Upper Cretaceous-Paleogene Sverdrup and intermontane basins (Eureka Sound Group), showing transitions in basin type. Time scale from Harland et al. (1990).

Sverdrup Basin

Thermal subsidence phase

Modelling of the thermomechanical properties of Sverdrup Basin by Stephenson et al. (1987) suggests that subsidence was thermally driven following the Upper Paleozoic rifting event which initiated the basin. A later event, which spanned much of the Early Cretaceous, corresponded to the main rifting phase that initiated the Amerasian Basin. The actual break-up unconformity of this phase (corresponding to seafloor spreading) is considered by Embry and Dixon (1990) to span the Albian through Cenomanian. Residual thermal subsidence is inferred to have continued into the Campanian, although the thermal regime may have been perturbed by Late Cretaceous volcanism and intrusion (Stephenson et al., 1987).

The Eurekan story begins with an Arctic-wide, and perhaps North American-wide transgression in the early Turonian (basal Kanguk Formation, base of Sequence 1). On west Axel Heiberg Island, Kanguk seas transgressed the subaerial component of the Strand Fiord Formation volcanic edifice that was probably the continental extension of the Alpha Ridge (Ricketts et al., 1985; Embry and Osadetz, 1988).

The first hint of sandy detritus that signalled encroaching deltas and coastal plains is seen in the mid-Campanian, corresponding to the base of the Eureka Sound Group. Along the eastern margin of Sverdrup Basin most of Sequence 1 was removed by pre-Paleocene erosion.

Based on decompacted stratigraphic thicknesses for the mid-Campanian to end-Maastrichtian on western Axel Heiberg Island, sedimentation rates were approximately 3 cm/ka. The overall rate of sediment supply not only kept pace with the creation of stratigraphic accommodation space but permitted the delta systems to prograde, notwithstanding a high degree of wave modification.

The sub-Paleocene unconformity forms the upper limit to Sequence 1. Volcanism persisted from the Late Cretaceous to the Early Paleocene but unlike previous magmatic episodes, the younger volcanic and intrusive rocks became increasingly alkalic and were confined to the northern margin of the basin.

The foredeep phase

The Paleogene Sverdrup Basin was characterized by expansion and a drastic increase in sedimentation rates. Uplift and subaerial erosion during late Maastrichtian time was followed by Early Paleocene transgression. Pebbles derived from sanidine-bearing, acid volcanic centres along the northern basin margin, were distributed over the central and western part of Sverdrup Basin during the early Paleocene or latest Maastrichtian sea level lowstand. There is no evidence that ancestral Princess Margaret Arch or Cornwall Arch were active at this time (Ricketts, 1987a). Estuarine deposits infilled antecedent drainage on the pre-Paleocene surface along the eastern and southern basin margins (Ricketts, 1991c), and represent the initial response to Early Paleocene transgression. Gradual blanketing of the pre-existing bedrock topography during the remainder of the Early Paleocene resulted in accumulation of widespread coastal plain and wave-dominated delta facies.

Sequence 2 was terminated some time in the mid-Paleocene by a basin-wide, and probably Arctic-wide transgression; a similar event is seen in the Yukon/Mackenzie Delta region (Dixon, 1986) and Spitzbergen (Helland-Hansen, 1990). The boundary between sequences 2 and 3 is placed at the base of a transgressive sandstone unit in the uppermost Expedition Formation. Thick shale (Strand Bay Formation) overlying the Expedition Formation is, for the most part, a component of the succeeding regressive package that also includes the lower Iceberg Bay Formation.

Sequence 3 in the north and west is again characterized by deltas, but in this case fluvial-dominated systems. East and south in Sverdrup Basin calcareous, shallow shelf and tidal flat mudrock (Cape Pillsbury Member) contain calcareous foraminifera including genera that have affinities with North Atlantic forms. Thus it is possible that a tentative connection between Sverdrup Basin and ancestral Baffin Bay was established during mid- Paleocene time, perhaps initiated during the Strand Bay transgression. However, the Baffin Bay connection was short-lived.

Sequence 4 contains forested (coal-bearing) delta plains and coastal plains (Coal member, Early to Middle Eocene). At Strathcona Fiord this setting was home to a diverse assortment of vertebrates. The Coal member extended from at least Vendom Fiord on south Ellesmere Island (Riediger and Bustin, 1987), to Lake Hazen in the north, west to Strand Fiord, and northwest to Emma Fiord. Laterally equivalent delta front and prodelta facies are presumably present offshore beneath the polar continental shelf.

The Paleogene succession of Sverdrup Basin records the onset of rapid subsidence and increased sedimentation, in addition to sporadic volcanism along the northern margin (northern Ellesmere Island). Sedimentation rates, based on decompaction of the Early Paleocene to Early Eocene (65 Ma to ca. 50 Ma) stratigraphic column on western Axel Heiberg Island are 15-17 cm/ka, five times the rate estimated for the Upper Cretaceous phase. Clearly the production of accommodation space by subsidence is a major factor. Thermal decay presumably played some role here, but the most significant influence was increasing Eurekan crustal shortening. Perhaps the best evidence for this is seen in the highly diachronous, synorogenic Buchanan Lake Formation which is probably as old as Early Eocene (Sequence 4). In eastern Sverdrup Basin significant crustal failure is first manifested in small Early Eocene half-grabens at Otto Fiord and Emma Fiord (Fig. 75). The eastern margins of these two basins are delineated by Black Top Fault, a normal fault that appears to be the extension of a major fault on Fosheim Peninsula. If this is true, then significant extension was taking place locally in early Eocene time. Farther east, reverse faulting, or thrusting may have been initiated at about the same time along Judge

Daly Peninsula, Franklin Pierce Bay, and Lake Hazen. Furthermore, Sverdrup Basin had expanded as far north as Lake Hazen by the end of the Paleocene and Early Eocene, so that the hiatus between the end of foredeep sedimentation and creation of Lake Hazen Basin must have been brief.

The foredeep phase of Sverdrup Basin, from its inception early in the Paleocene to its eventual demise in the Eocene, underwent a continual metamorphosis during its approximately 20 million year existence. Preliminary examination of tectonic subsidence curves suggest that subsidence accelerated during the early Paleogene. Drastically increased sedimentation rates, gradual extinguishing of volcanism, and evidence for contemporaneous, albeit local tectonism late in its history are indicative of increasing crustal shortening and the eventual disintegration of Sverdrup Basin.

Results of seismic reflection and refraction surveys of the polar continental shelf conducted from the Ice Island have been summarized by Forsyth et al. (1990). Velocity profiles between Nansen Sound and Ellef Ringnes Island indicate at least 12 km of probable post-Early Cretaceous strata, i.e., strata that accumulated since the opening of Amerasian Basin, and inception of Sequence 1. The sub-Buchanan Lake Formation unconformity, so prominent onshore, is inferred offshore from a significant change in velocity gradient between velocity units 4 and 5. Up to 10 km of Upper Cretaceous-Tertiary strata occur beneath the polar shelf in a basin that was probably contiguous with Sverdrup Basin. Velocity unit 5 would therefore be equivalent to the synorogenic Buchanan Lake Formation and perhaps the postorogenic Neogene clastic wedge.

Nevertheless, the Sverdrup foredeep is not typical of other well known foredeeps or foreland basins. Unlike the Alberta Basin and its deformed foreland in Western Canada, there is no obvious vertical crustal load associated with a major thrust system. The thrusts that do occur in the eastern Arctic are younger than the inception of the foredeep, and are primarily responsible for subsidence of the much smaller, syntectonic intermontane basins. Some of the faults are no older than Early Eocene or possibly latest Paleocene. Likewise, uplift of the major arches began about the same time and they cannot be considered as crustal loads earlier in the Paleogene. One possible mechanism of foredeep subsidence here involves horizontal loads, for example gradually increasing compression during convergence of Greenland and Ellesmere Island, but prior to significant crustal failure. The response of the crust to horizontal stresses in terms of subsidence or uplift has been analyzed by Cloetingh (1988) and Stephenson and Cloetingh (1991). It is conceivable that rapidly changing Arctic plate interactions during the Paleogene resulted in horizontal loads, coupled with remnant thermal effects and sediment loading, that were sufficient to produce the abrupt increase in subsidence rate beginning in the Early Paleocene.

The synorogenic phase: the death of Sverdrup Basin

Sporadic tectonism, beginning in the Early Eocene (perhaps earlier in places) and confined primarily to the northern part of Sverdrup Basin, reached a climax in the Middle Eocene. The result was the formation of at least three additional intermontane basins that were superimposed on the antecedent contiguous Sverdrup Basin. The Buchanan Lake Formation in Sequence 5 and the upper part of Sequence 4, provides the stratigraphic record of these events.

Each synorogenic basin was mechanically linked to major structures. Axel Heiberg Basin, Lake Hazen Basin, and Franklin Pierce/Judge Daly basins occur outboard of major frontal thrusts or reverse faults. Otto Fiord (half graben) and possibly Emma Fiord subbasins appear to be extensional in origin, within what is regionally a contractional orogen. The largest basin, Axel Heiberg Basin, is spatially associated with the Princess Margaret Arch, forming along the fractured (Stolz Fault Zone) eastern limb. Princess Margaret Arch and its western companion Cornwall Arch have been modelled as large, continental crustal flexures that formed during the main phase of Eurekan compression (Stephenson and Ricketts, 1990; Stephenson and Cloetingh, 1991). Compositional trends in Axel Heiberg Basin conglomerate are consistent with the general unroofing stratigraphy that would be expected during erosion of a hanging wall above the Stolz system of thrusts.

Sediment infill of Axel Heiberg Basin is reasonably constrained as Middle Eocene, by palynology and the stratigraphic relationship with Middle Eocene strata of the Iceberg Bay Formation (note that McIntyre, 1991b, does not preclude the possibility of extending the age of Axel Heiberg Basin into the Late Eocene). Nevertheless, because the Buchanan Lake Formation accumulated as a direct response to uplift and faulting, and itself experienced only minor deformation, it is reasonable to infer that most of the Eurekan deformation associated with this particular basin was over by the end of the Middle Eocene or the Late Eocene. However, as already noted, deformation may have taken place earlier in basins closer to the principal region of compression on northern and eastern Ellesmere Island.

Sedimentation rates estimated for Axel Heiberg Basin are about 20 cm/ka; more than six times the rate for the Upper Cretaceous Sverdrup Basin, and 25-30% greater than that for the Paleogene foredeep phase of Sverdrup Basin.

The intermontane basins represent the culmination of a trend from early thermal-dominated and compressional subsidence throughout much of Sverdrup Basin history, to complete crustal failure during the climax of Eurekan deformation.

Basin transitions and plate kinematics

Figure 77 illustrates the correspondence, and in some cases lack of correspondence, between important stratigraphic events marked by depositional sequence boundaries, magnetic anomalies, and changes in plate motions. Most of the plate kinematic data utilized is from Srivastava (1985), Srivastava and Tapscott (1986), Rowley and Lottes (1988), Scotese et al. (1988), and Roest and Srivastava (1989).

Continental rifting that eventually gave rise to the opening of Labrador Sea probably began about Barremian time (Rowley and Lottes, 1988), and partly overlaps the main phase of rifting that eventually gave rise to the Amerasian

Basin (Embry and Osadetz, 1988; Embry and Dixon, 1990). The actual production of oceanic crust in southern Labrador Sea began about 90 Ma, again corresponding reasonably well to the timing of continental breakup in Amerasian Basin, the cessation of Strand Fiord volcanism on west Axel Heiberg Island, and the transgression that initiated earliest deposition of Sequence 1 (Kanguk Formation).

Thermal subsidence during the Upper Cretaceous phase of Sverdrup Basin and deposition of Sequence 1 continued until the end of the Cretaceous. During this period Canada Basin continued to open and the axis of spreading in Labrador Sea migrated north towards Baffin Bay. Greenland had begun its counterclockwise journey of rotation away from North America in the south and there continued to be a large separation (underlain by continental crust) between northern Greenland and Ellesmere Island.

Uplift of Sverdrup Basin, probably in the latter part of the Maastrichtian, and development of the sub-Paleocene unconformity is a signal event in the eastern Arctic. Alkalic continental magmatism, although present during the Cretaceous, predominated in the Paleogene. Sedimentation rates increased dramatically in the Paleogene foredeep phase of Sverdrup Basin (sequences 2 to 4). Seafloor spreading had effectively ceased in Amerasian Basin according to Embry and Dixon (1990), whereas spreading had reached the present position of Baffin Bay resulting in an outpouring of thick tholeiitic basalt flows. Early Paleocene olivine tholeiites at Cape Dyer are considered to have been contiguous with similar rocks on West Greenland and record the opening of Baffin Bay-Davis Strait (Clarke and Upton, 1971; Upton, 1988). Upton (1988) indicates that most of the Baffin Bay-Davis Strait volcanism took place between 63 and 56 Ma (sequences 2 and 3), but continued at a lesser pace into the Oligocene. Northern Greenland had begun its approach towards Ellesmere Island and by Early Eocene the effects of crustal shortening were felt in Sverdrup Basin (Sequence 4). By magnetic anomaly 25 (59 Ma, beginning of Sequence 3) seafloor spreading in the North Atlantic had also begun. In the Norwegian-Greenland sea this event is manifested in the Central Tertiary Basin on Spitsbergen (Helland-Hansen, 1990; Müller and Spielhagen, 1990). Re-evaluation of magnetic anomalies in Labrador Sea by Roest and Srivastava (1989) further indicates that between anomalies 24 and 25 there was a significant change in the direction and increase in the rate of seafloor spreading. An implication stated in their work is that Greenland must also have been about 100 km farther south of earlier estimates at the time of initial seafloor spreading. If this is the case, some dextral motion of Greenland relative to Ellesmere Islands is also implied.

West of the Queen Elizabeth Islands, in northern Yukon and Mackenzie Delta (Dixon, 1986) a regional unconformity separates Late Maastrichtian and younger rocks from mid-Cretaceous and older depositional sequences as a consequence of widespread Late Cretaceous to Early Tertiary tectonism. A major sub-Paleocene unconformity, called the Bylot unconformity by McWhae (1981), also extends from the Labrador Shelf to north Baffin Island. On Bylot Island this unconformity occurs below Miall's (1986) Mount Lawson Formation, and on southeast Baffin Island, below the Cape Searle Formation (Burden and Langille, 1990). Even farther afield, the Lower Tertiary succession on Spitsbergen overlies with profound unconformity Lower Cretaceous strata; the product of Late Cretaceous uplift (Nottvedt, 1985). Clearly the latest Cretaceous event was widespread across the North American Arctic and Barents Shelf region, prior to opening of the North Atlantic.

By anomaly 21 (about 49 Ma) motion between northeastern Ellesmere Island and northwest Greenland was almost orthogonal to the trend of Nares Strait. Significant crustal failure and synorogenic sedimentation was initiated in Sequence 4. Substantial crustal shortening occurred between anomalies 24 and 21 and by the Middle Eocene was manifested in major fold and thrust domains, and complete tectonic disintegration of Sverdrup Basin. On a global scale there was general re-organization of plates, including the dramatic collision of India with Asia (anomaly 22; Scotese et al., 1988). Thrusting generated by transpression along the western margin of the Central Tertiary Basin on Spitsbergen probably began at anomaly 25, and continued until anomaly 21 (after which strike slip displacement predominated; Müller and Spielhagen, 1990). Deformation and subsequent eastwards migration of the foredeep in Spitsbergen thus corresponds closely to similar events in the Paleogene of Sverdrup Basin and the synorogenic intermontane basins.

From the earliest Oligocene on (anomaly 13), the relative motion of Ellesmere Island and northwest Greenland perpendicular to Nares Strait was almost complete. This is consistent with the onshore geology of Queen Elizabeth Islands where Eureka Sound Group rocks younger than Middle Eocene are unknown (the younger rocks may have been eroded during the Neogene; they may also occur beneath the polar continental shelf). In addition, deformation associated with Axel Heiberg Basin, the youngest synorogenic basin encountered to date, appears to have been almost complete by the Middle Eocene.

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APPENDIX 1

Palynology

D.J. McIntyre

Palynology and age determinations of samples are arranged according to formation and section number. Sample positions are in metres from the base of measured sections and where appropriate are plotted on Figures 2a to 2d. For details of biostratigraphy on western Axel Heiberg Island see McIntyre (1991a); for the fossil forest beds at Geodetic Hills (Buchanan Lake Formation) see Ricketts and McIntyre (1986), McIntyre (1991b); and Cornwall and Amund Ringnes islands (McIntyre and Ricketts, 1989).

Expedition Formation
Lower member

Section RAK-85-13, NTS 49G, 79°57′N, 85°10′W, Slidre Fiord, Ellesmere Island. Figure 2b

C-131384, P2793-53, 13-6 (155.5-156.2 m).

The limited pollen assemblage contains *Fibulapollis* sp. This suggests a possible Campanian age.

C-131385, P2793-54, 13-7 (156.2-162.2 m).

The palynoflora contains numerous dinoflagellates of the genera *Alterbidinium*, *Isabelidinium*, and *Chatangiella* which indicate a Turonian to Campanian age. Pollen present includes *Fibulapollis* sp., *Azonia sufflata* Wiggins, *Aquilapollenites quadrilobus* Rouse, *A. formosus* Srivastava and Rouse, *A.* sp. cf. *A. rigidis* Tschudy and Leopold and *A. trialatus* var. *uniformis* Tschudy and Leopold. This pollen assemblage indicates a late middle to late Campanian age.

C-131387, P2793-55, 13-9 (192.6 m).

Pollen species present here are *Fibulapollis* sp., *Aquilapollenites trialatus* var. *variabilis* Tschudy and Leopold and *A. reticulatus* (Mtchedlishvili) Tschudy and Leopold. A late Campanian age is indicated by these species.

Section RAK-88-26, NTS 560D, 81°25′N, 89°46′W, North shore of Emma Fiord, due north of Cape Coastguard, Ellesmere Island.

C-153180, P3370-14, 26-1 (6 m).

The pollen flora includes *Wodehouseia gracile* (Samoilovitch) Pokrovskaya, *Aquilapollenites unicus* (Chlonova) Chlonova and *A. augustus* Srivastava.

C-153181, P3370-15, 26-2 (10 m).

The palynological assemblage, richer than that of sample 1, includes *A. unicus*, *A. augustus*, *W. gracile*, (abundant) *A. quadrilobus* Rouse, *Cranwellia rumseyensis* Srivastava, and *Pseudointegricorpus protrusum* Takahashi and Shimono.

C-154182, P3370-16, 26-3 (20 m).

Pollen species in this assemblage include *Aquilapollenites augustus*, *A. unicus*, *Wodehouseia gracile* (abundant), and *Pseudointegricorpus protrusum* (abundant).

The abundant pollen of the 'aquiloid' and 'oculata' groups in these three samples indicate an early Maastrichtian age for this section.

Expedition Formation
Upper member

Section RAK-85-13, NTS 49G, 79°57′N, 85°10′W, Slidre Fiord, Ellesmere Island. Figure 2b

C-131391, P2793-56, 13-13 (215.1-217.1 m). C-131392, P2793-57, 13-14 (215.1-217.1 m). C-131393, P2793-58, 13-15 (228.5-229.4 m). C-131394, P2793-59, 13-16 (251.7-252.2 m). C-131395, P2793-60, 13-17 (263.7-268.7 m). C-131396, P2793-61, 13-18 (263.7-268.7 m). C-131397, P2793-61A, 13-19 (285 m). C-131399, P2793-62, 13-21 (298.7-304.2 m).

The pollen floras from these eight samples are similar and contain few species or specimens. Pollen present includes Alnus, Betula, Ulmus (*Ulmipollenites undulosus* Wolff), *Paraalnipollenites alterniporus* (Simpson) Srivastava, *Triporopollenites mullensis* (Simpson) Rouse and Srivastava, and Ericaceae (both *Ericipites* and *Ericaceoipollenites rallus* Stanley). The limited assemblages of these samples indicate a Paleocene age, and the absence of the genera *Momipites* and *Caryapollenites* suggests a possible early Paleocene age

There is no indication in the samples examined from section RAK- 85-13 of the presence of Maastrichtian strata.

Section RAK-85-17, NTS 340B, 80°07′N, 84°30′W, Hot Weather Creek, Ellesmere Island. Figure 2b

C-131331, P2793-80, 17-1 (106.5-113.5 m).

The pollen flora consists mainly of Tertiary forms with long ranges. Some *Paraalnipollenites alterniporus* (Simpson) Srivastava and abundant *Hazaria sheoparii* Srivastava is also present.

The age can be determined no closer than Paleocene.

C-131332, P2793-81, 17-2 (368.5-374.5 m).

The assemblage contains *P. alterniporus*, *Caryapollenites inelegans* Nichols and Ott, *C. imparalis* Nichols and Ott, and *C. wodehousei* Nichols and Ott. Pollen of Ericaceae (*Ericipites*) is abundant.

The presence of the *Caryapollenites* species indicates a late Paleocene age.

Section RAK-88-3, NTS 49G, 79°38′30″N, 87°08″W, north side of Mokka Fiord, Axel Heiberg Island.

C-153167, P3370-1, 3-1.

Black organic material; no pollen seen.

C-153168, P3370-2, 3-3.

The sparse assemblage includes *Alnus* sp. and Ericaceae (*Ericipites*). This sample is no older than Paleocene. The limited pollen flora suggests that it is likely no younger than Paleocene.

Section RAK-88-7, NTS 49G, 79°20′N, 85°28′W, 14 km NW of May Point, Axel Heiberg Island. Figure 2d

C-153172, P3370-6, 7A-1. C-153173, P3370-7, 7B-3.

The sparse microfloras contain *Triporopollenites* spp., *Alnus* sp., and Ericaceae (*Ericaceoipollenites rallus* Stanley and *Ericipites*). A Paleocene age is indicated.

Strand Bay Formation

Section RAK-84-2A, NTS 49E, 78°28′N, 82°40′W, Strathcona Fiord, Memoir Figure 2a

C-122568, P2715-20, 2A-6

The presence of *Momipites wyomingensis* Nichols and Ott indicates a middle to late Paleocene age for this low diversity microflora.

For additional biostratigraphic details on the Strand Bay Formation see McIntyre (1991a), and Wall, Appendix 2 this volume.

Iceberg Bay Formation
Cape Pillsbury Member

Section RAK-84-3, NTS 49E, 78°31′N, 82°40′W, Strathcona Fiord, Figure 2a

C-122522, P2716-1, 3-1

The sparse microflora contains rare *Caryapollenites imparalis* Nichols and Ott, *C. wodehousei* Nichols and Ott, and *Momipites wyomingensis* Nichols and Ott. These species indicate a middle or late Paleocene age for this sample.

Section RAK-85-2, NTS 49H, 79°07′N, 83°30′W, Vesle Fiord

C-122637, P2793-8A, 2-1

The very limited pollen flora recovered from this sample contains rare *Caryapollenites imparalis* Nichols and Ott which indicates a middle or late Paleocene age.

Section RAK-85-6, NTS 49H, 79°07′N, 83°30′W, Vesle Fiord

C-122677, P2793-31A, 6-2, Figure 2b

The microflora is similar to that of the other samples (C122522, C122737, C122568, C131363) of this report and is probably also middle or late Paleocene.

Section RAK-85-11, NTS 49H, 79°35′N, 81°50′W, South Bay.

C-131363, P2793-49A, 11-5, Figure 2b

The presence of *Momipites wyomingensis* Nichols and Ott and *M. ventifluminis* Nichols and Ott shows that this sample also is middle or late Paleocene.

The limited pollen assemblages of the four Cape Pillsbury Member samples consist mainly of bisaccate conifer pollen of *Picea* and *Pinus* type, Taxodiaceae/Cupressaceae, *Alnus* sp., *Betula* sp., and *Paraalnipollenites alterniporus* (Simpson) Srivastava. Such pollen floras are typically present in the Paleocene in this area but conclusive Paleocene determinations are provided by the species of *Momipites* and *Caryapollenites* recorded

Iceberg Bay Formation
Coal member

Section RAK-83-34, NTS 59H, 79°19′N, 91°43′W, North side of Kanguk Peninsula near head of Expedition Fiord.

For sample locations see Ricketts (1991, Fig. 49).

C-112107, P2614-148, 34-1, 2.0 m; C-112108, P2614-149, 34-2, 15.0 m; C-112112, P2614-150, 34-6, 12.5-13.5 m; C-112115, P2614-151, 34-9, 27.5 m; C-112116, P2614-152, 34-10, 51.0 m; C-112117, P2614-153, 34-11, 52.5 m;

C-112118, P2614-154, 34-12, 51.5 m; C-112119, P2614-155, 34-13, 54.0 m; C-112120, P2614-156, 34-14, 62.0 m; C-112121, P2614-157, 34-15, 64.0 m; C-112122, P2614-158, 34-16, 66.0 m; C-112123, P2614-159, 34-17, 77.0 m; C-112124, P2614-160, 34-18, 77.5 m; C-112125, P2614-161, 34-19, 80.0-80.5 m; C-112126, P2614-162, 34-20, 104.0 m; C-112127, P2614-163, 34-21, 113.0 m; C-112131, P2614-164, 34-26, 130.0-130.8 m; C-112137, P2614-165, 34-30, 159.5-162.0 m.

Most pollen assemblages from the predominantly coaly samples of Section RAK-83-34 consist of only a few species and small total floras. A few samples, particularly at the top of the section, contain more diverse palynofloras. Conifer pollen (*Picea* type, *Pinus* type, and Taxodiaceae) is abundant in most samples, including those with limited palynofloras. Each taxon in the following list occurs in only a few of the samples.

Sphagnum sp.
Osmunda sp.
Deltoidospora sp.
Laevigatosporites sp.
Radialisporis radiatus (Krutzsch) Krutzsch
Alnus sp.
Betula sp.
Caryapollenites wodehousei Nichols and Ott
Carya sp.
Corylus sp.
Diervilla sp.
Ericaceae
Liliacidites sp.
Liquidambar sp.
Monocolpopollenites sp.
Nudopollis sp.
Paraalnipollenites alterniporus (Simpson) Srivastava
Pistillipollenites mcgregorii Rouse
Sparganium sp.
Triporopollenites spp.
tricolpate pollen
tricolporate pollen
Ulmus sp.
Pesavis tagluensis Elsik and Jansonius

Age and comments

The occurrence of *Pistillipollenites mcgregorii* clearly indicates that this section is not older than Middle Paleocene and not younger than Middle Eocene. The presence of rare *Paraalnipollenites alterniporus* suggests that it is no younger than Early Eocene. The upper part of the section contains *Tilia* sp., *Diervilla* sp., and *Carya* sp. (Eocene type, see section RAK-83-25, McIntyre, 1991a). The presence of these three pollen types shows that the upper part of section RAK-83-34 is of Eocene age. The pollen floras of the section suggest correlation with that part of section RAK-83-25 (C-111844-C-112308) where the Paleocene-Eocene boundary is considered to be present. It is possible that the first appearance of *Tilia* sp. in section RAK-83-34 (at 77 m-C-112123) marks the base of the Eocene but the limited pollen floras in many samples suggests that such a determination is somewhat speculative because of the lack of definite evidence.

Section RAK-87-13, NTS 560D, 81°30′N, 88°W, 5 km north of Fire Bay, Emma Fiord, Ellesmere Island.

C-153105, P3123-1, 13-1; C-153107, P3123-3, 13-3.

Extremely limited microflora of nondiagnostic spores and pollen in both samples. The presence of very rare *Aquilapollenites tumanganicus* Bolotnikova in sample 13-1 clearly indicates a late Paleocene or Early Eocene age.

Section RAK-87-14, NTS 560D, 81°04′N, 89°25′W, 20 km SE of White Point, Hvitland Peninsula, Ellesmere Island.

C-153114, P3124-1, 14-1; C-153115, P3124-2, 14-2;
C-153116, P3124-3, 14-3; C-153117, P3124-4, 14-4.

Pollen and spores present in the palynofloras of these samples include *Cicatricosisporites cicatricosoides* Krutzsch, *Paraalnipollenites alterniporus* (Simpson) Srivastava, Ericaceae, *Alnus* sp., *Betula* sp., *Triporopollenites* spp., *Ulmus* sp. (*Ulmipollenites undulosus* Wolff), and *Caryapollenites wodehousei* Nichols and Ott. The assemblages are of Paleocene age and the rare specimens of *C. wodehousei* indicate late Paleocene.

Section RAK-87-18, NTS 59H, 79°53′N, 89°30′W, 10 km NE of Geodetic Hills, Axel Heiberg Island.

C-153134, P3128-1,2, 18-1; C-153135, P3128-3, 18-2;
C-153136, P3128-4, 18-3; C-153137, P3128-5, 18-4
(79°53′N, 89°25′W).

The pollen floras from this section include *Juglans* sp., *Tilia vescipites* Wodehouse, *T. crassipites* Wodehouse, *Tricolporopollenites kruschii* (Potonie) Thompson and Pflug, *Paraalnipollenites alterniporus* (Simpson) Srivastava (rare), *Caryapollenites wodehousei* Nicholls and Ott, *C. viridifluminipites* Wodehouse, and *Caryapollenites* sp. (undescribed Eocene form). The presence of *Juglans*, *Tilia* (abundant), *Tricolporopollenites kruschii* and the younger types of *Caryapollenites*, and rare *Paraalnipollenites alterniporus* indicates a probable early Eocene age for this interval. The pollen floras from the overlying fossil forest interval contain species not recorded in section RAK-87-18, for which a middle Eocene age cannot be completely eliminated.

C-153138, P3128-6, 18-5. (79°54′N, 89°18′W,
12.5 km NE of Geodetic Hills).

This sample is probably the same age as the rest of 87-18 but contains few pollen grains.

Section RAK-87-22, NTS 49G, 79°26′N, 86°08′W, in axis of syncline, 20 km south of Depot Point, Axel Heiberg Island.

C-153152, P3132-1, 22-1; C-153153, P3132-2, 22-2.

The palynofloras of both samples contain *Alnus* sp., *Triporopollenites* spp., Ericaceae (*Ericipites*), *Liliacidites* sp., and *Paraalnipollenites alterniporus* (Simpson) Srivastava (abundant). A Paleocene age is indicated by this limited assemblage.

Section RAK-87-23, NTS 49C, 79°48′N, 87°37′W, Gibs Fiord, Axel Heiberg Island.

C-153154, P3133-1, 23-1; C-153155, P3133-2, 23-2.

The pollen assemblages include *Alnus* sp., *Ulmus* sp., *Tilia vescipites* Wodehouse, *Carya viridifluminipites* Wodehouse, and *Caryapollenites* sp. The occurrence of *T. vescipites*, *C. viridifluminipites*, and *C.* sp.(distinct from Paleocene forms) suggests an Eocene (probably early) age.

Section RAK-88-11, NTS 120F and G, 82°03′N, 68°04′W, Near small lake, 2 km north of hairpin bend, left bank on Turnabout River, Ellesmere Island. Figure 2d

C-153174, P3370-8, 11-5 (13 m); C-153175, P3370-9, 11-6 (24.3-25.3 m).

Both samples contain limited pollen assemblages which include Ericaceae (*Ericipites*), *Betula*, *Alnus*, *Ulmus*, and *Paraalnipollenites alterniporus* (Simpson) Srivastava.

This assemblage suggests a probable Paleocene age.

Section RAK-88-12, NTS 120F and G, 82°03′N, 68°04′W, Near small lake, 2 km north of hairpin bend, right bank on Turnabout River, Ellesmere Island. Figure 2d

C-153176, P3370-10, 12-1 (0-3.5 m).

Details of section location, pollen flora and age determination as for section RAK-88-11.

Section RAK-88-23, NTS 340C, 81°44′N, 85°54′W, West Phillips Inlet, Ellesmere Island.

C-153160, P3368-1, 23-1.

Pollen flora includes Ericaceae (*Ericipites*), *Alnus*, *Betula*, *Caryapollenites* sp. cf. C. wodehousei Nichols and Ott, Onagraceous sp. of Ioannides and McIntyre, and *Paraalnipollenites alterniporus* (Simpson) Srivastava.

C-153161, P3368-2, 23-2.

The sparse microflora contains Ericaceae, *Caryapollenites* sp., and *Paraalnipollenites alterniporus*.

C-153163, P3368-4, 23-4.

Sparse pollen assemblage which includes *P. alterniporus*. The pollen floras recovered from these samples indicate a late Paleocene age.

Buchanan Lake Formation

Section RAK-85-14, NTS 340B, 80°59′N, 86°45′W, Otto Fiord, Ellesmere Island. Figure 2b

C-131302, P2793-64, 14-2.

The palynoflora contains abundant pollen including *Paraalnipollenites alterniporus* (Simpson) Srivastava, *Alnus* sp., Ericaceae, *Nudopollis* sp., *Tilia vescipites* Wodehouse, *Saxonipollis* sp. A of Ioannides and McIntyre, *Aquilapollenites tumanganicus* Bolotnikova, *Pterocarya* sp., *Caryapollenites wodehousei* Nichols and Ott, *C. inelegans* Nichols and Ott, and undescribed younger forms of *Caryapollenites*.

The presence of *Aquilapollenites tumanganicus* indicates a late Paleocene to early Eocene age. The abundance of *Paraalnipollenites alterniporus*, the presence of *Saxonipollis*, the rarity of *Tilia vescipites* and the presence of *Caryapollenites inelegans* and *C. wodehousei* favours a late Paleocene age.

C-131303, P2793-65, 14-3.

Pollen flora similar to 14-2 but fewer grains present. Species include *C. wodehousei*, *C. inelegans*, and *Caryapollenites* sp. (undescribed). The presence of *Ilex* sp. and *Tricolporopollenites kruschii* (Potonie) Thomson and Pflug suggest that this sample may be early Eocene.

C-131308, P2793-66, 14-8.

This sample contains *Caryapollenites wodehousei* as well as abundant specimens of *Carya pollenites* spp. commonly present in the Eocene of the Arctic. *Tilia vescipites* occurs commonly and *T. crassipites* is also present. The pollen assemblage indicates an early Eocene age.

Section RAK-87-13, NTS 560D, 81°29′N, 88°50′W, Fire Bay, Emma Fiord, Ellesmere Island.

C-153112, P3123-3, 13-8; C-153113, P3123-4, 13-9.

Both samples contain sparse microfloras in which reworked Carboniferous spores are common. Tertiary pollen present includes *Tilia* spp. (both *T. vescipites* Wodehouse and *T. crassipites* Wodehouse) and *Caryapollenites* sp. (undescribed types recorded elsewhere in the Arctic from Eocene strata). The presence of these pollen species indicates a probable early Eocene age for this interval.

Section RAK-88-4, NTS 340B, 80°21′N, 87°50′W, NE Schei Peninsula, Axel Heiberg Island. Figure 2d

C-153169, P3370-3, 4-1 (15.7 m); C-153170, P3370-4, 4-2 (41.1 m); C-153171, P3370-5, 4-3 (47.0 m).

The palynofloras contain abundant pollen which includes *Juglans* spp., *Alnus* sp., *Betula* sp. *Tsuga* sp., Ericaceae, *Tilia vescipites* Wodehouse, *Pterocarya* sp., *Engelhardtia* sp., *Intratriporopollenites* sp., *Lonicera* sp., cf. *Viburnum, Pistillipollenites mcgregorii* Rouse. The pollen assemblages from this section are similar to assemblages from the Geodetic Hills Fossil Forest interval and also indicate a middle Eocene age (Ricketts and McIntyre, 1986; McIntyre, 1991b).

Section RAK-88-14, NTS 120F and G, 82°02′30″N, 69°10′W, South Boulder Hills, 8 km due west of Turnabout River, Ellesmere Island. Figure 2d

C-153177, P3370-11, 14-2 (44 m).

The sparse microflora includes pollen of *Alnus* sp., *Betula* sp., *Caryapollenites* spp., (Eocene forms) and *Novemprojectus traversii* Choi.

C-153178, P3370-12, 14-5 (69 m).

This sample yielded few pollen grains. Present are *Alnus* sp., *Betula* sp., *Tilia* sp., and *Tsuga* sp.

C-153179, P3370-13, 14-8 (100.8 m).

Only a few grains which have long ranges were found among the abundant detrital material.

The presence of *Tilia*, *Caryapollenites*, *Tsuga*, and *Novemprojectus traversii* indicate an early or possibly middle Eocene age. This limited flora suggests correlation with the uppermost part of section RAK-83-25 at Strand Fiord, Axel Heiberg Island (McIntyre, 1991a).

Re-examination of three samples from Amund Ringnes Island and three samples from Cornwall Island, District of Franklin.

The six samples had previously been examined and reported on by W.S. Hopkins, Jr. After extra processing of the residues and new preparations of some of the samples, palynofloras that were significantly better than those of the original preparations were obtained. More precise age determinations are available from these richer palynofloras.

Cornwall Island samples, NTS 59C, 77°37′N, 94°42′W.

C-19605, P832-1, HFA-72-42; C-19606, P832-2, HFA-72-43; C-19607, P832-3, HFA-72-44

These three samples, collected from an outlier of a Tertiary yellow-buff sandstone, lying unconformably on Blaa Mountain Formation, in the centre of Cornwall Island, yielded similar pollen floras. All preparations, from both the silty or muddy portions and the organic or coaly portions, contain pollen assemblages which do not vary significantly in composition except for the abundance of reworked Late Cretaceous dinoflagellates in the silty samples and their absence in the coaly fractions. Conifer pollen, *Picea* and *Pinus* types, is abundant in all preparations, and may be dominant, while pollen of the Taxodiaceae-Cupressaceae complex is abundant in all preparations. Simple tricolpate angiosperm pollen, not further differentiated or identified, is abundant in all preparations, but has little significance at present for determination of age. Other angiosperm pollen, some of which are known to have greater age significance, are commonly present but none of the types listed below is abundant and some are not present in all preparations.

Alnus sp.
Betula sp.
Corylus sp.
other betulaceous pollen (indeterminate)
Carya spp.
Ericaceae
Engelhardtia sp.
Ilex sp.
Pterocarya sp.
Tilia sp.
Ulmus spp. (both verrucate and rugulate forms)
Quercus sp.
Pachysandra type
Sparganium sp.
Liliacidites sp.
Monocolpopollenites sp.
Aquilapollenites tumanganicus Bolotnikova
Pistillipollenites mcgregorii Rouse
Paraalnipollenites alterniporus (Simpson) Srivastava
Tricolporopollenites kruschii (Potonie) Thomson and Pflug
Trudopollis sp.
Triporopollenites bituitus (Potonie) Elsik
Diervilla sp. and *Novemprojectus traversii* Choi

The pollen floras from the three samples indicate an early Eocene age for the Tertiary strata, from which they were collected, on Cornwall Island. The occurrence of *Pistillipollenites mcgregorii* clearly shows that these floras are not younger than middle Eocene and the presence of rare *Paraalnipollenites alterniporus* indicates that they are no younger than early Eocene. Rouse (1977) determined that *Pistillipollenites mcgregorii* does not occur later than middle Eocene in the Arctic Islands and Ioannides and McIntyre (1980) indicated that *Paraalnipollenites alterniporus*, a primarily Paleocene species, occurs rarely as late as early Eocene in the Mackenzie Delta area. Pollen of *Tilia* spp. occurs rarely in the pollen assemblages. This is mainly *T. vescipites* Wodehouse but a few grains may be *T. crassipites* Wodehouse. The occurrence of *Tilia* spp. indicates an early Eocene, or younger, age. This pollen type was recorded in the Eocene, but not the Paleocene, of the Arctic by Rouse (1977), Doerenkamp et al. (1976) and Ioannides and McIntyre (1980). It does appear in the middle Paleocene of Wyoming (Nichols and Ott, 1978) but Pocknall (1987) noted that *Intratriporopollenites* sp.

(*Tilia crassipites*) did not appear until the Eocene. *Tilia* sp. (*vescipites* type) does occur rarely in Paleocene strata at Strand Fiord, Axel Heiberg Island (McIntyre, 1991a) and Spitsbergen (M.J. Head, pers. comm., 1987). The types of *Carya* pollen in the samples from this outlier are the same as those recorded from samples considered to be Eocene by Choi (1983) and McIntyre (1991a) from Strand Fiord and from outcrops of Eocene strata in the Mackenzie Delta (Ioannides and McIntyre, 1980) and Banks Island (Doerenkamp et al., 1976). These grains differ in size and type of polar thinning from the *Caryapollenites* species described from the Paleocene of Wyoming by Nichols and Ott (1978) and recorded from Paleocene strata at Strand Fiord by McIntyre (1991a). At Strand Fiord the *Carya* species, which are the same as those here recorded from Cornwall Island, immediately succeed the latest Paleocene *Caryapollenites* species. Further evidence of early Eocene age is provided by the occurrence of *Tricolporopollenites kruschii* which was not recorded until the Eocene in the Arctic (Rouse, 1977; Doerenkamp et al., 1976) and first occurs in the Strand Fiord section in strata of Eocene age (McIntyre, 1991a). Pollen of *Ilex* sp. and *Engelhardtia* sp., not known before the Eocene elsewhere, occurs very rarely in the Cornwall Island samples. The presence of *Aquilapollenites tumanganicus* indicates that this material is not younger than early Eocene. This species first appears in the latest Paleocene at Strand Fiord and last appears in early Eocene and is succeeded by the related form *Novemprojectus traversii* which is not known to occur prior to the Eocene (McIntyre, 1991a).

The evidence from the pollen species and flora discussed above gives a clear indication that the palynological assemblages from the Cornwall Island outlier are of early Eocene age and are more precise determinations than the Late Paleocene to Early Eocene ages determined previously. These assemblages are closely comparable with early Eocene assemblages from the upper part (2425-2805 m, section RAK-25-83) of the Iceberg Bay Formation (Eureka Sound Group) at Strand Fiord, Axel Heiberg Island (Ricketts, 1991a), where younger strata are also present.

Amund Ringnes Island, NTS 59F, 78°11′N, 95°55′W.

C-19608, P832-4, BAA-72-118

One sample from an outlier of yellow-buff sandstone, lying unconformably on Deer Bay Formation contains pollen flora similar to those from Cornwall Island (C-19605, 6, 7) but is less varied. It contains abundant pollen of *Pinus* and *Picea* and Taxodiaceae-Cupressaceae. Tricolpate angiosperm pollen is also abundant. Other angiosperm present are:

Alnus sp.
Betula sp.
Corylus sp.
Carya sp.
Ulmus spp. (both verrucate and rugulate forms)
Ericaceae
Sparganium sp.
Liliacidites sp.
Monocolpopollenites sp.
Pistillipollenites mcgregorii Rouse
Aquilapollenites tumanganicus Bolotnikova

The presence of *Pistillipollenites mcgregorii* shows that the age of this sample is within the range of late Paleocene to middle Eocene. The *Carya* pollen identified is similar to that seen in the Cornwall Island samples and suggests that this sample may also be early Eocene in age. The rest of the pollen flora provides no evidence to confirm or deny this suggestion.

In Balkwill's (1983) study, W.S. Hopkins, Jr. (Appendix, p. 74) indicated a Paleocene age for this sample but the limited evidence available from the new preparations suggests that a slightly younger age is possible. The pollen listed by Hopkins as the Paleocene *Aquilapollenites* cf. *A. spinulosus* Funkhouser is here considered to be the Paleocene-Eocene *Aquilapollenites tumanganicus*, of which a specimen was seen in the original preparation.

Slime Peninsula, southwestern Amund Ringnes Island, NTS 69D, 78°00′N, 97°25′W.

C-21974, P832-48, BAA-72-204

Sample from about 210 m above base of Eureka Sound Group.

The sample contains abundant Late Cretaceous dinoflagellates, which are certainly reworked from Kanguk Formation strata of Turonian to Santonian age. A rich assemblage of angiosperm pollen including the following is present.

Aquilapollenites quadrilobus Rouse
Aquilapollenites reticulatus (Mtchedlishvili) Tschudy and Leopold
Azonia strictiparva Frederiksen
Expressipollis barbatus Chlonova
Expressipollis ocliferius Chlonova emend. Bondarenko
Orbiculapollis globosus (Chlonova) Chlonova
Porosipollis porosus (Mtchedlishvili) Krutzsch
Tumidulipollis accuratus (Chlonova) Bondarenko
Tumidulipollis sibiricus (Bondarenko) Bondarenko
Wodehouseia edmontonicola Wiggins

This pollen assemblage clearly indicates an early Maastrichtian age. It is similar to early Maastrichtian palynofloras from the Expedition Formation at Strand Fiord (642-749 m, section RAK-83-25) which were discussed by McIntyre (1991a). The preparation obtained from extra processing of the residue is much better than that available to W.S. Hopkins, Jr. who reported only a Maastrichtian age (Balkwill, 1983) for this sample (wrongly recorded as C-21784).

Slime Peninsula, southwestern Amund Ringnes Island, NTS 69D, 77°59′N, 97°05′W.

C-22929, P917-2, BAA-72-202

Sample from several metres above base of Kanguk Formation.

The abundant dinoflagellates in this sample include:

Alterbidinium minor (Alberti) Lentin and Williams
Chatangiella ditissima (McIntyre) Lentin and Williams
Chatangiella granulifera (Manum) Lentin and Williams
Chatangiella scheii (Manum) Lentin and Williams
Chatangiella verrucosa (Manum) Lentin and Williams
Elytrocysta druggii Stover and Evitt
Fromea chytra (Drugg) Stover and Evitt
Heterosphaeridium difficile (Manum and Cookson) Ioannides
Isabelidinium acuminatum (Cookson and Eisenack) Stover and Evitt
Isabelidinium cooksoniae (Alberti) Lentin and Williams
Laciniadinium arcticum (Manum and Cookson) Lentin and Williams
Spinidinium sverdrupianum (Manum) Lentin and Williams
Trithyrodinium suspectum (Manum and Cookson) Davey
Wallodinium luna (Cookson and Eisenack) Lentin and Williams

This dinoflagellate assemblage indicates an age in the range of Turonian to Santonian. W.S. Hopkins, Jr. (p. 73 in Balkwill, 1983) reported a sparse terrestrial palynoflora and a Late Cretaceous age. He noted the presence of miscellaneous phytoplankton but in both the old and new preparations microplankton are completely dominant and spores and pollen comprise only a very minor proportion of the palynological assemblage.

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Illustrations of selected pollen and spores

Plates 1 to 4

Slides containing the figure specimens are curated in the type collection of the Geological Survey of Canada, 601 Booth Street, Ottawa, Ontario K1A 0E8. They are at present in temporary storage at the Institute of Sedimentary and Petroleum Geology, 3303 33rd Street NW, Calgary, Alberta T2L 2A7, where all duplicate slides are permanently stored. In the descriptions for Plates 1 to 4 the species name is followed by the GSC locality number (prefixed C), the slide number (prefixed P) and the Geological Survey of Canada type number (prefixed GSC). Stage co-ordinates and England Finder readings for Reichert-Jung Polyvar microscope 392166 at the Institute of Sedimentary and Petroleum Geology, Calgary, Alberta are on file with the curated specimens. All figures on the four plates are shown at a magnification of x1000.

The figures on the four plates appear approximately in order of stratigraphic appearance. Campanian and Maastrichtian pollen are shown on Plates 1 and 2. Maastrichtian and Paleocene pollen and spores and longer ranging pollen are shown on Plate 2. Paleocene and early to middle Eocene pollen are illustrated on Plates 3 and 4.

PLATE 1

Figure 1. *Aquilapollenites trialatus* var. *uniformis* Tschudy & Leopold; C-131385, P2793-54c, GSC 103442

Figure 2. *Aquilapollenites trialatus* var. *variabilis* Tschudy & Leopold; C-131387, P2793-55b, GSC 103443

Figure 3. *Aquilapollenites formosus* Srivastava & Rouse; C-131385, P3793-54c, GSC 103444

Figure 4. *Aquilapollenites unicus* (Chlonova) Mtchedlishvili; C-153181, P3370-15e, GSC 103445

Figure 5. *Aquilapollenites augustus* Srivastava; C-153181, P3370-15e, GSC 103446

Figure 6. *Fibulapollis sp.*;C-131387, P2793-55b, GSC 103447

Figure 7. *Tumidulipollis accuratus* (Chlonova) Bondarenko; C-21794, P832-48e, GSC 103448

Figure 8. *Aquilapollenites reticulatus* (Mtchedlishvili) Tschudy and Leopold; C-131387, P2793-55b, GSC103449

Figure 9. *Wodehouseia gracile* (Samoilovitch) Pokrovskaya; C-153181, P3370-15g, GSC 103450

Figure 10. *Cranwellia rumseyensis* Srivastava; C-153181, P3370-15c, GSC 103451

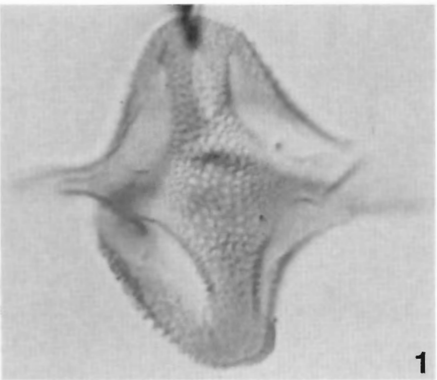
Figure 11. *Pseudointegricorpus protrusum* Takahashi and Shimono; C-153182, P3370-16b, GSC 103452

Figure 12. *Expressipollis sibiricus* (Bondarenko) Bondarenko; C-21794, P832-48e, GSC 103453

Figure 13. *Azonia strictiparva* Frederiksen; C-21794, P832-48e, GSC 103454

Figure 14. *Wodehouseia edmontonicola* Wiggins; C-21794, P832-48e, GSC 103455

Plate 1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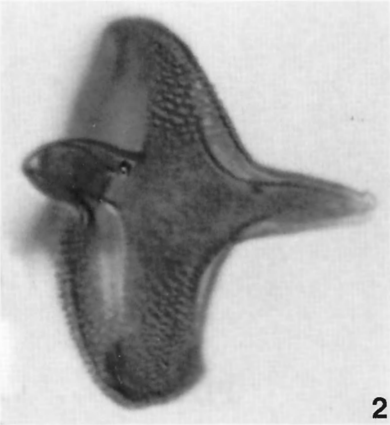
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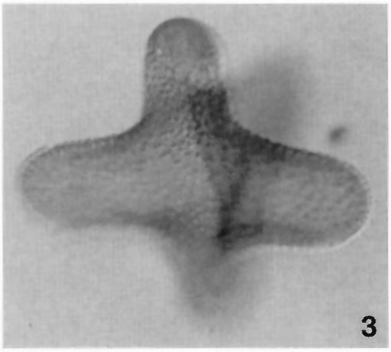
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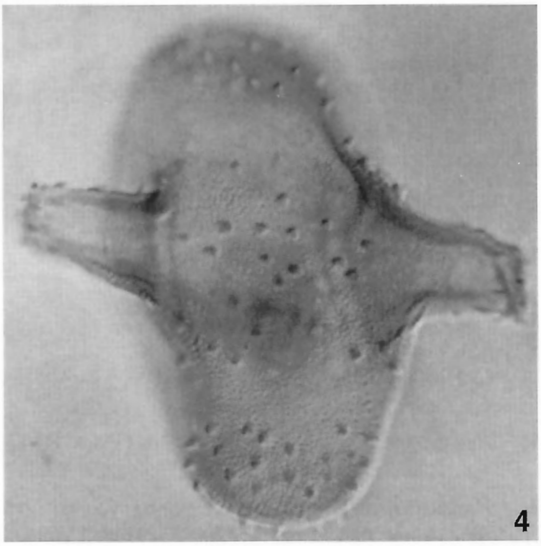
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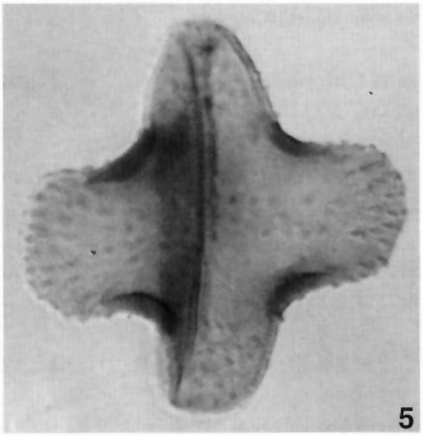
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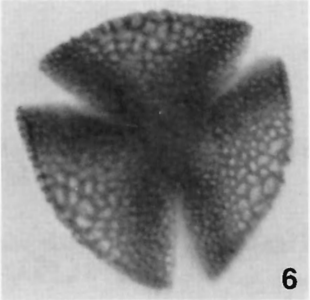
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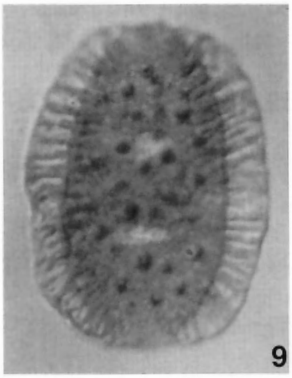
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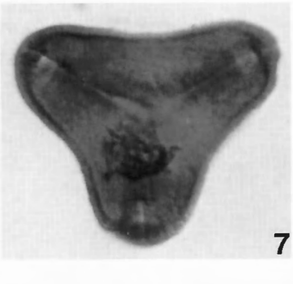
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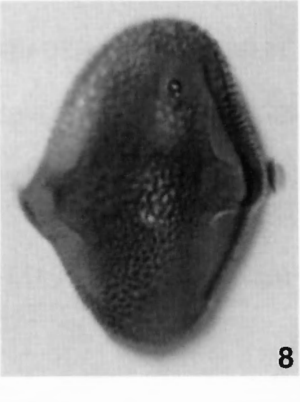
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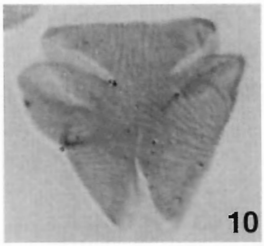
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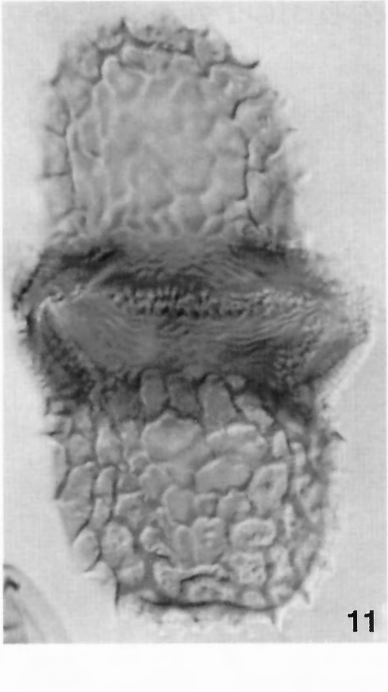
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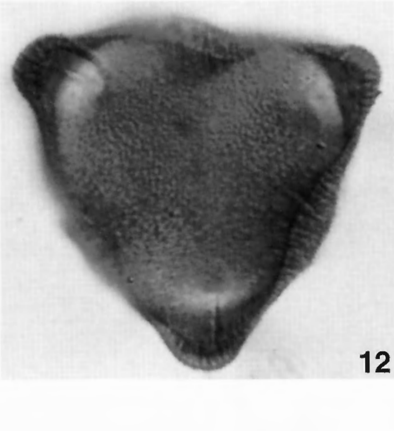
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PLATE 2

Figure 1. *Orbiculapollis globosus* (Chlonova) Chlonova; C-21794, P832-48e, GSC 103456

Figure 2. *Expressipollis barbatus* Chlonova; C-21794, P832-48e, GSC 103457

Figure 3. *Porosipollis porosus* (Mtchedlishvili) Krutzsch; C-21794, P832-48e, GSC 103458

Figure 4. *Hazaria sheoparii* Srivastava; C-131331, P2793-80a, GSC 103459

Figure 5. *Pinus* sp.; C-153117, P3124-4b, GSC 103460

Figure 6. *Cicatricosisporites cicatricosoides* Krutzsch; C-153115, P3124-2c, GSC 103461

Figure 7. *Osmunda* sp.; C-153171, P3370-5c, GSC 103462

Figure 8. *Picea* sp.; C-19607, P832-3g, GSC 103463

Figure 9. Taxodiaceae (*Metasequoia* sp.); C-153135, P3128-3c, GSC 103464

Figure 10. *Laevigatosporites* sp.; C-10607, P832-3g, GSC 103465

Figure 11. Taxodiaceae (*Sequoiapollenites* sp.); C-153117, P3124-4c, GSC 103466

Figure 12. Taxodiaceae (cf. *Taxodium* sp.); C-112125, P2614-161d, GSC 103467

Figure 13. Taxodiaceae (*Metasequoia* sp.); C-112123, P2614-159d, GSC 103468

Figure 14. *Ericaceoipollenites rallus* Stanley; C-112123, P2614-159c, GSC 103469

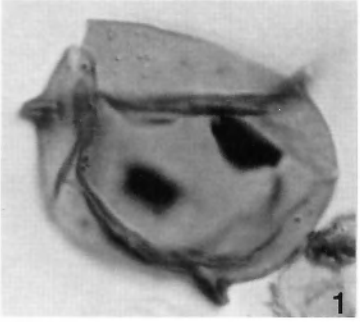
Figure 15. Ericaceae (*Ericipites* sp.); C-19606, P832-2j, GSC 103470

Figure 16. *Betula* sp.; C-153115, P3124-2b, GSC 103471

Figure 17. *Paraalnipollenites alterniporus* (Simpson) Srivastava; C-131392, P2793-57c, GSC 103472

Figure 18. *Triporopollenites mullensis* (Simpson) Rouse & Srivastava; C-131331, P2793-80a, GSC 103473

Plate 2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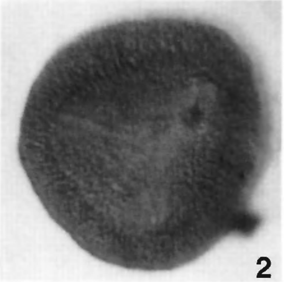
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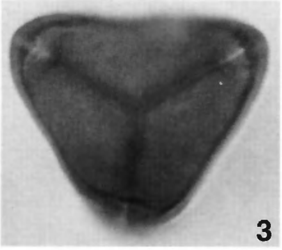
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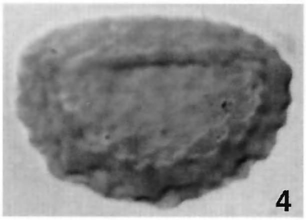
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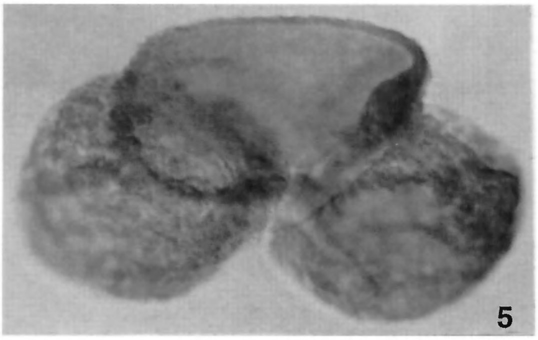
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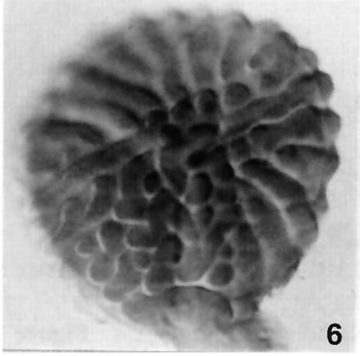
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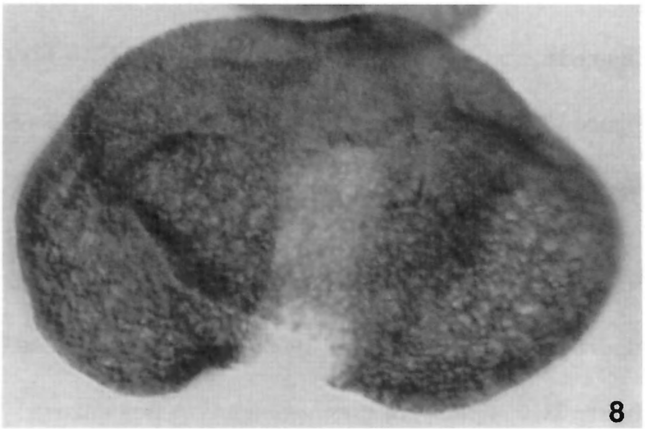
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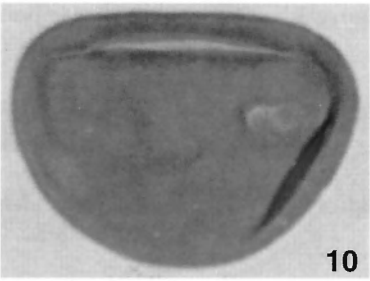
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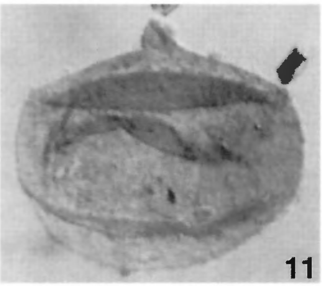
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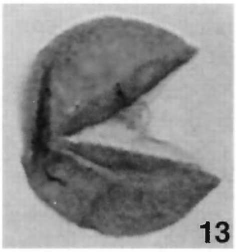
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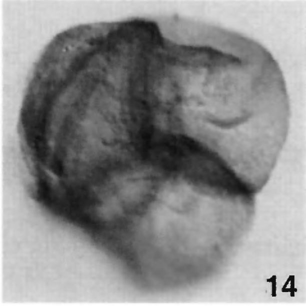
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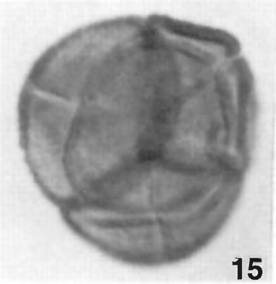
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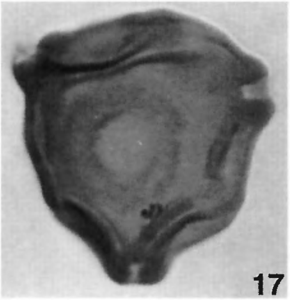
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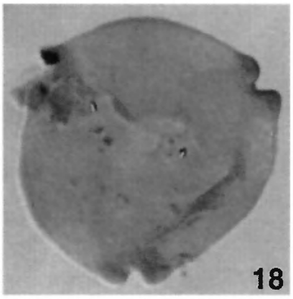
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PLATE 3

Figure 1. *Alnus* sp.; C-19607, P832-3g, GSC 103474

Figure 2. *Alnus* sp.; C-153171, P3370-5c, GSC 103475

Figure 3. *Momipites wyomingensis* Nichols and Ott; C-122568, P2715-20c, GSC 103476

Figure 4. *Caryapollenites inelegans* Nichols and Ott; C-131302, P2793-64c, GSC 103477

Figure 5. *Caryapollenites imparalis* Nichols and Ott; C-122522, P2716-1b, GSC 103478

Figure 6. *Caryapollenites wodehousei* Nichols and Ott; C-153134, P3128-2c, GSC 103479

Figure 7. *Caryapollenites wodehousei* Nichols and Ott; C-153134, P3128-2c, GSC 103480

Figure 8. *Ulmipollenites undulosus* Wolff (*Ulmus* sp.); C-134896, P2793-126b, GSC 103481

Figure 9. *Ulmoideipites krempii* Anderson (*Ulmus* sp); C-19608, P832-4e, GSC 103482

Figure 10. *Momipites ventifluminis* Nichols and Ott; C-131363, P2793-49Ac, GSC 103483

Figure 11. *Paraalnipollenites alterniporus* (Simpson) Srivastava; C-131391, P2793-56c, GSC 103484

Figure 12. *Liquidambar* sp.; C-153171, P3370-5c, GSC 103485

Figure 13. *Monocolpopollenites* sp. (cf. *Gingko* sp.); C-19607, P832-3g, GSC 103486

Figure 14. *Pterocarya* sp.; C-19606, P832-2j, GSC 103487

Figure 15. *Pistillipollenites mcgregorii* Rouse; C-19606, P832-2j, GSC 103488

Figure 16. *Quercus* sp.; C-153154, P3133-1c, GSC 103489

Figure 17. *Nudopollis* sp.; C-131302, P2793-64c, GSC 103490

Figure 18. *Pistillipollenites mcgregorii* Rouse; C-19606, P832-2j, GSC 103491

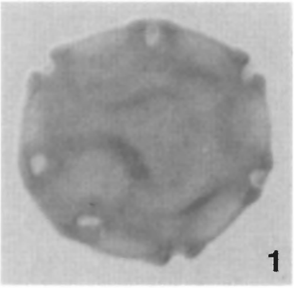
Figure 19. *Sparganium* sp.; C-19608, P832-4e, GSC 103492

Figure 20. *Liliacidites* sp.; C-19606, P832-2k, GSC 103493

Figure 21. *Aquilapollenites tumanganicus* Bolotnikova; C-19607, P832-3g, GSC 103494

Figure 22. *Diervilla* sp.; C-153170, P3370-4b, GSC 103495

Plate 3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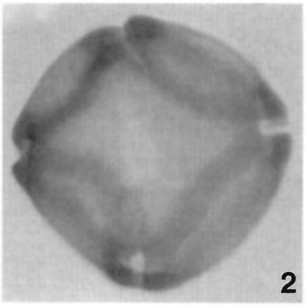
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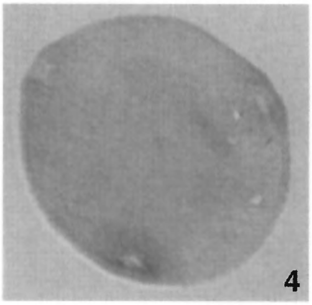
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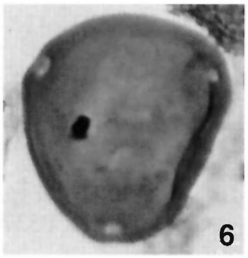
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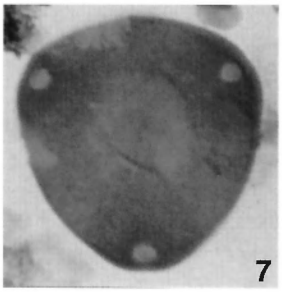
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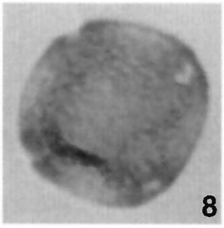
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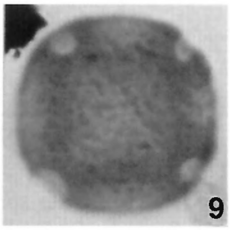
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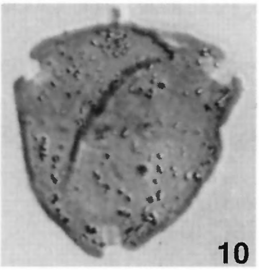
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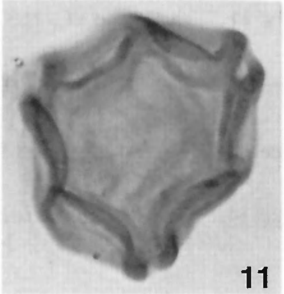
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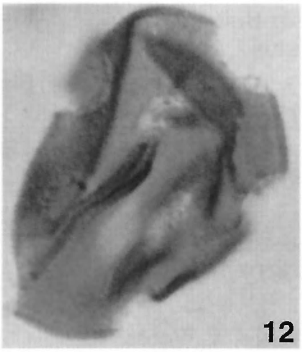
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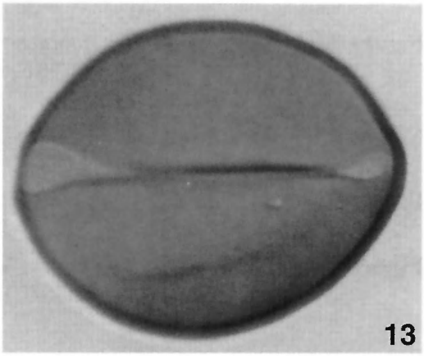
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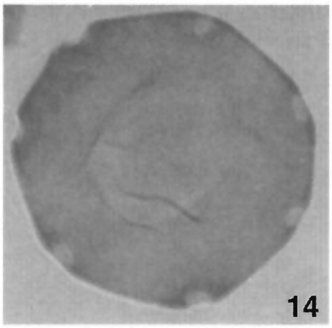
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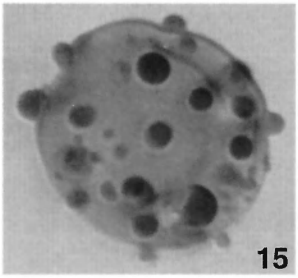
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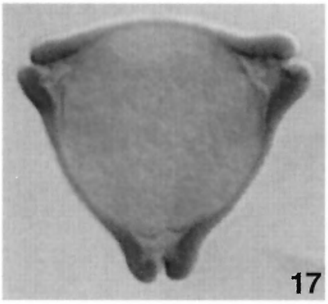
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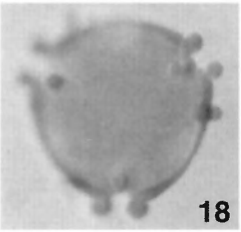
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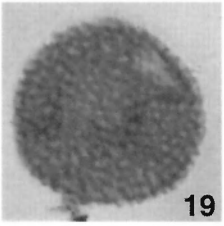
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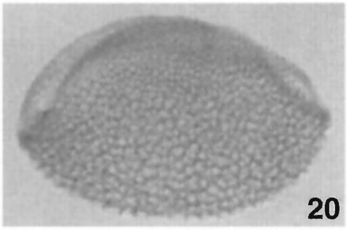
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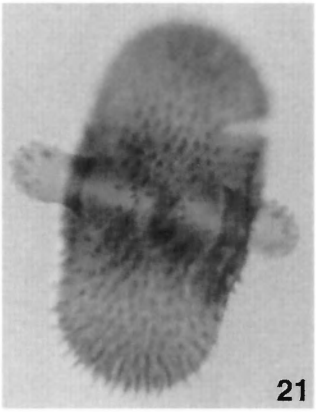
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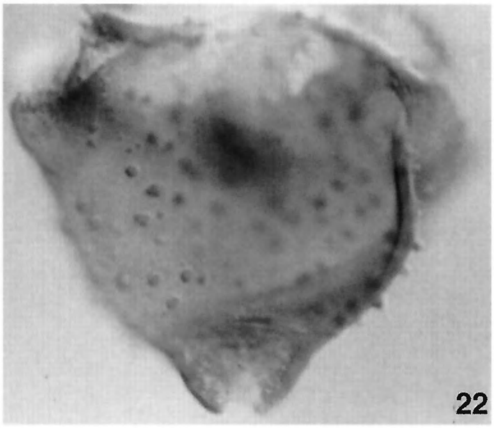
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PLATE 4

Figure 1. *Tilia* sp. (*T. crassipites* Wodehouse); C-19607, P832-3i, GSC 103496

Figure 2. *Tilia* sp. (*T. vescipites* Wodehouse); C-131302, P2793-64c, GSC 103497

Figure 3. *Aquilapollenites tumanganicus* Bolotnikova; C-19607, P832-3g, GSC 103498

Figure 4. *Pachysandra* sp.; C-134888, P2793-5a, GSC 103499

Figure 5. *Saxonipollis* sp.; C-131302, P2793-64b, GSC 103500

Figure 6. *Carya viridifluminipites* Wodehouse; C-19606, P832-2j, GSC 103501

Figure 7. *Caryapollenites veripites* (Wilson & Webster) Nichols & Ott; C-131302, P2793-64c, GSC 103502

Figure 8. *Juglans* sp.; C-153170, P3370-4c, GSC 103503

Figure 9. *Intratriporopollenites* sp.; C-134896, P2793-126b, GSC 103504

Figure 10. *Engelhardtia* sp.; C-153171, P3370-5d, GSC 103505

Figure 11. *Carya* sp.; C-153134, P3128-2c, GSC 103506

Figure 12. *Ilex* sp.; C-131303, P2793-65a, GSC 103507

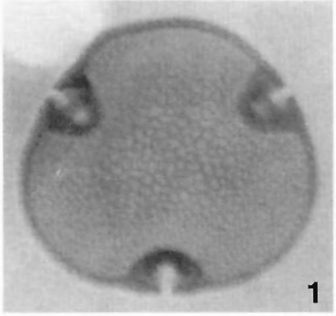
Figure 13. *Viburnum* sp.; C-134896, P2793-126a, GSC 103508

Figure 14. *Novemprojectus traversii* Choi; C-19607, P832-3g, GSC 103509

Figure 15. *Tricolporopollenites kruschii* (Potonie) Thomson & Pflug;C-134917, P2793-124a, GSC 103510

Figure 16. *Tsuga* sp.; C-153178, P3370-12b, GSC 103511

Plate 4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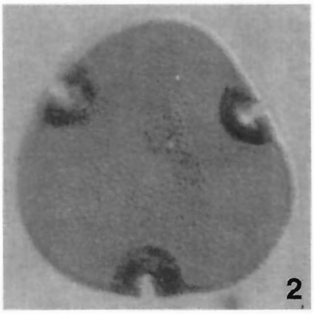
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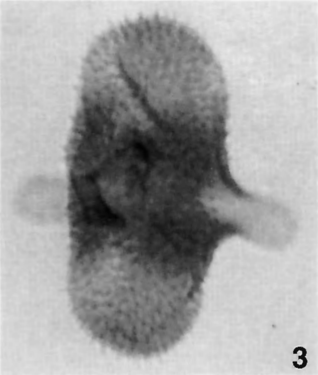
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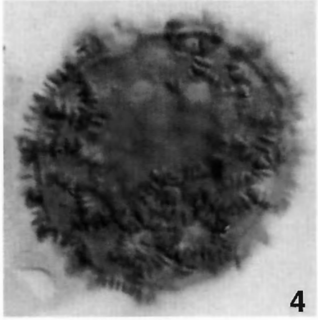
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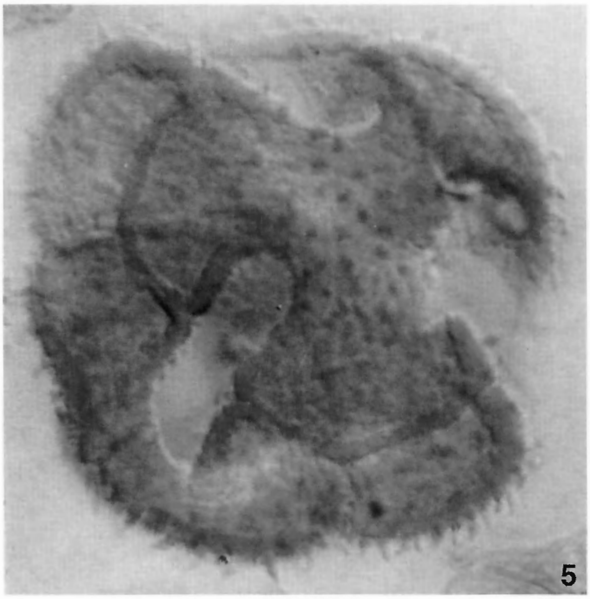
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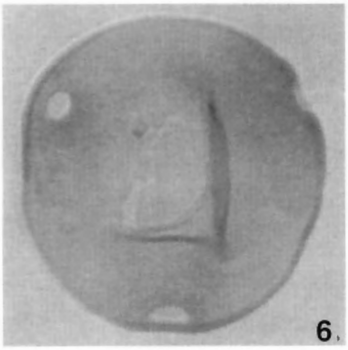
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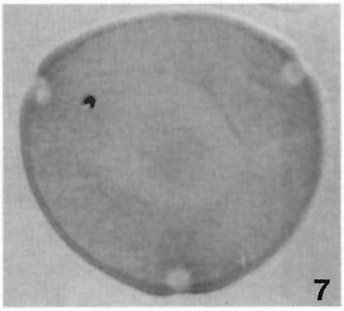
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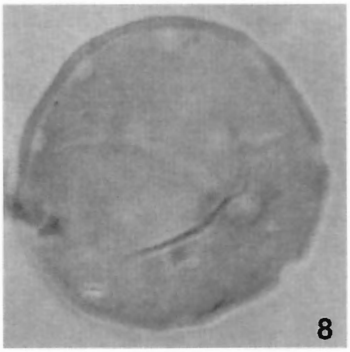
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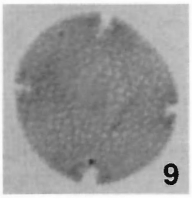
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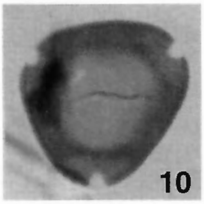
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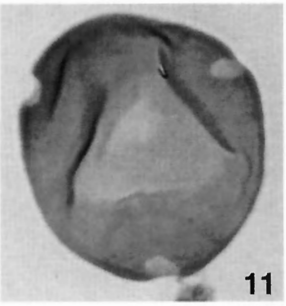
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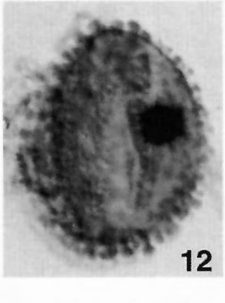
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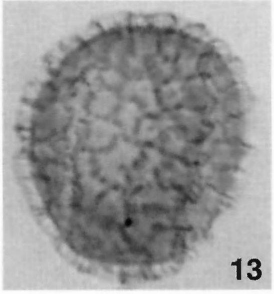
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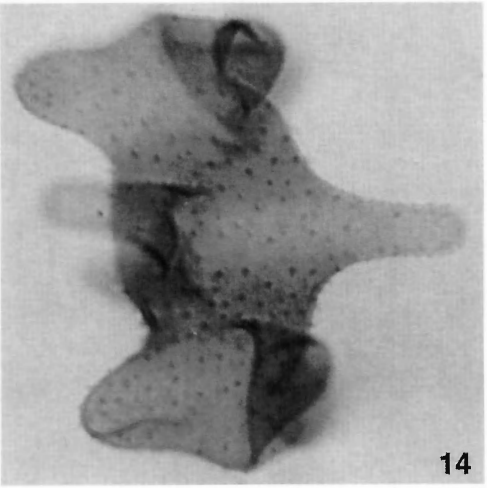
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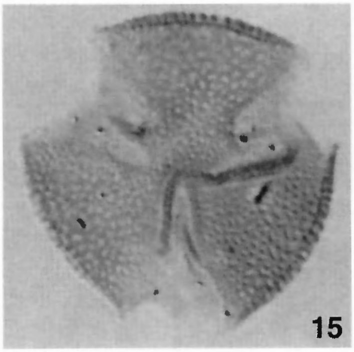
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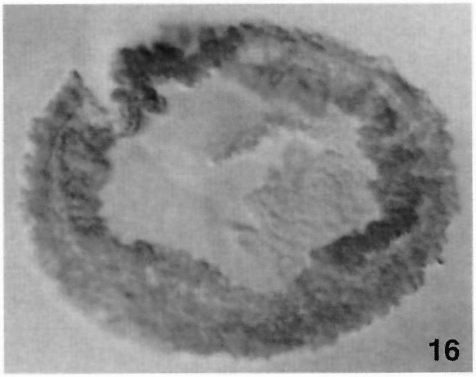
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APPENDIX 2

Micropaleontology

J.H. Wall and D.H. McNeil

Introduction

Microfossils, chiefly foraminifera, are present in outcrops of the Cretaceous Kanguk Formation at Slidre Fiord and of the Paleocene Strand Bay and/or Iceberg Bay formations of the Eureka Sound Group in the vicinity of Vesle, Canon, and Strathcona fiords in west-central Ellesmere Island (Fig. 6). A subsurface section of what is probably the Strand Bay Formation in the Panarctic Union Talemen J-34 well has also yielded foraminifera. The first published record of foraminifera in the Eureka Sound Group of the area is by Hornaday in West et al. (1975, p. 575) who reported species of *Cyclammina* and *Cornuspira* at Vesle and Strathcona fiords in close proximity to the localities discussed herein. The following faunal lists and accompanying comments on age and environment of selected assemblaes were extracted from internal reports compiled for B.D. Ricketts by J.H. Wall in 1985-86, with subsequent updating of the taxonomy and correlation of the Paleocene suites by D.H. McNeil. Many of the Kanguk Formation foraminifera have been illustrated previously by Wall (1983, pl. 7), while those from the Eureka Sound Group, originally assembled for a poster session (Wall et al., 1988), are reproduced herein (Pls. 1 to 5). The Cretaceous-Paleocene foraminiferal biostratigraphy of the region is summarized in conclusion.

APPENDIX 2. (cont.)

Surface Sections

Field No. & Stratigraphy	**Locality, Microfossils, Age and Environment**	**GSC Loc. No.**
85 RAK-13-1 Kanguk Fm., 21-27 m from base section (composite sample	Slidre Fiord 79°57′N, 85°10′W; NTS 49G (=same as A.F. Embry's South Remus Creek GSC section 79 EL-12, as illustrated by Wall (1983, fig. 10) Foraminifera (numerous, residue mostly of foraminiferal tests): *Saccammina* sp. 2 of Wall (1983) *S.* sp., flask-shaped with neck *Reophax* sp. *Miliammina* sp. cf. *M.* sp. 2 of Wall (1983) – two *Haplophragmoides* sp. cf. *H. bonanzaense* Stelck and Wall *H. fraseri* Wickenden *H. howardense* Stelck and Wall *H.* sp. cf. *H. rota* Nauss *H.* spp. *Recurvoides* sp. *Ammobaculites* spp. – most specimens with a 3 to 5-chambered, straight, uncoiled portion suggesting affinity with *A. fragmentarius* Cushman, a long-ranging species *Pseudobolivina rollaensis* (Stelck and Wall) *Trochammina albertensis* Wickenden *T.* sp. 2 of Wall (1967) *Verneuilinoides bearpawensis* (Wickenden) *Arenobulimina* sp. cf. *A. torula* Tappan *Dorothia smokyensis* Wall – smaller than typical specimens Age: Late Cretaceous, Late Santonian to Late Campanian, probably about mid-Campanian. The fauna contains components of both the *D. smokyensis* assemblage of Turonian to Early Campanian age and the *V. bearpawensis* assemblage of Late Campanian age, as defined by Wall (1983, p. 264-266). Thus, it is difficult to be more definitive in dating this fauna. Environment: Marine, of shallow to moderate depth in shelf zone.	C-131379
85 RAK-13-2 Kanguk Fm., 34.5-42 m above base section (composite sample)	Same locality Foraminifera (less common): *Ammodiscus* sp. – one incomplete specimen *Reophax* sp., very small – one *Miliammina* sp. cf. *M.* sp. 2 of Wall (1983) - one *Haplophragmoides* sp. cf. *H. howardense* Stelck and Wall *H.* sp. *Spiroplectammina* sp. 2 of Wall (1967) *Trochammina* sp., small – one *Dorothia smokyensis* Wall(?) – a terminal portion *Serovaina orbicella* (Bandy)(?) – two very small specimens *Anomalina*(?) sp., poorly preserved - two Diatomacea: *Lepidodiscus* sp. cf. *L. elegans* Witt *Coscinodiscus* sp., thin frustule morphotype F (drum type) of Wall (1975) – one Bivalvia (Pelecypoda): *Inoceramus* sp .– presence indicated by clusters of aragonite Coprolites(?): ovoid pellets of unknown origin – common	C-131380

Surface Sections (cont.)

Field No. & Stratigraphy	**Locality, Microfossils, Age and Environment**	**GSC Loc. No.**
	Age: Late Cretaceous, probably Late Campanian, based on the presence of the diatom *Lepidodiscus* cf. *elegans* and the few calcareous foraminifera associated with the *V. bearpawensis* assemblage of Wall (1983). As *Spiroplectammina* sp. 2 is a component of the older (Turonian-Early Campanian) *D. smokyensis* assemblage, it is possible that this sample, like the preceding C-131379, may be close to the boundary of the *D. smokyensis* and *V. bearpawensis* assemblage intervals, and perhaps is of mid-Campanian age. Environment: Marine, of shallow to moderate depth in shelf zone.	
85 RAK-13-3 Kanguk Fm., 50-71 m above base section (composite sample)	Same locality Foraminifera (common): *Saccammina* sp. 2 of Wall (1983) – two *S.* sp., flask-shaped with neck *Ammodiscus* spp., two species represented, one with distinct suture (one specimen), the other very thin-walled with a less distinct suture (four specimens) *Reophax* spp. - two specimens *Haplophragmoides howardense* Stelck and Wall *H.* sp. cf. *H. kirki* Wickenden - two *H.* sp. cf. *H. rota* Nauss *H.* spp. *Evolutinella* or *Trochamminoides* sp., indistinct *Ammobaculites* spp. – most specimens with a 3 to 4-chambered, straight, uniserial portion suggesting affinity with *A. fragmentarius* Cushman, a long-ranging species *Trochammina albertensis* Wickenden *T.* sp. *Verneuilinoides bearpawensis* (Wickenden) *V.* sp. - two *Arenobulimina* sp. cf. *A.* torula Tappan *Dorothia smokyensis* Wall – smaller than typical specimens Bivalvia (Pelecypoda): *Inoceramus* sp. – presence indicated by clusters of aragonite Age: Late Cretaceous, probably Late Campanian, as most of the forms present are associated with the *V. bearpawensis* assemblage of Wall (1983) considered to be this age. The presence of *D. smokyensis* again may indicate a position near the boundary of the *D. smokyensis* and *V. bearpawensis* assemblage intervals, i.e. an approximate mid-Campanian age, or it may represent a slightly extended range of this species into younger strata. Environment: Marine, relatively shallow depth in the shelf zone.	C-131381
85 RAK-13-5 Kanguk Fm./Expedition Fm. (Lower member) gradational contact, 88-110.5 m above base section (composite sample)	Same locality Foraminifera: *Saccammina* spp., poorly preserved – a circular species and a flask-shaped species with neck *Ammodiscus* sp., thin-walled, as in previous entry GSC loc. C-131381 *Haplophragmoides* sp. cf. *H. rota* Nauss *H.* spp. *Ammomarginulin*a sp., small, with very short uncoiled portion – prominent *Trochammina albertensis* Wickenden *T.* spp.	C-131383

Surface Sections (cont.)

Field No. & Stratigraphy	Locality, Microfossils, Age and Environment	GSC Loc. No.
	Verneuilinoides bearpawensis (Wickenden) *Marssonella(?)* sp., small – one *Heterolepa(?)* or *Eponides(?)* sp., poorly preserved and eroded, possibly reworked – two Diatomacea: *Coscinodiscus* sp. – two morphotype F (drum type) of Wall (1975) Seeds: *Costatheca* sp. – one (may not be indigenous) Pisces: traces of scales Age: Late Cretaceous, Campanian or possibly younger. The fauna, although not as well developed as in the preceding sample, does not differ greatly in composition. As the small species of *Ammomarginula*, prominent in this assemblage, has not been previously observed in either the Kanguk or Eureka Sound, its biostratigraphic significance, if any, is not known. Environment: Marine, shallow depth in shelf zone.	
85 RAK-2-1 Iceberg Bay Fm., Cape Pillsbury Member, 30-35 m above base (composite sample)	Vesle Fiord 70°07′N, 83°30′W; NTS 49H Foraminifera: *Bathysiphon* spp. – two species represented, one very fine grained, the other coarse grained *Hyperammina* sp., very fine grained *Ammodiscus* sp., poorly preserved *Haplophragmoides* spp. *Recurvoides* spp. *Reticulophragmium borealis* (Petracca) – has been recorded previously from the Paleocene of the Mackenzie Delta in the Ministicoog Member of the Moose Channel Formation (Price et al., 1980, p. 183 wherein this species was identified as *Cyclammina coksuvorovae* Ushakova; McNeil, 1989, p. 219 and fig. 5) *Trochammina* sp., small – one *Tritaxia* sp. – one *Verneuilinoides* spp. – apparently two species represented, one slender with nearly parallel sides, the other more tapered *Cibicidoides* sp. A (=*Cibicidoides* sp. 3450 of McNeil, 1989). This species was recognized previously in the subsurface Paleocene of the Mackenzie Delta area (McNeil, 1989, p. 211-212, pl. 4, fig. 8) Age: Mid-Paleocene, based on similarity of assemblage to that of the *R. borealis* Zone recognized in the Paleocene of the Mackenzie Delta (McNeil, 1989). In addition to the two key species listed above, others in the Ellesmere assemblage identical to Mackenzie Delta components include the species of *Bathysiphon, Haplophragmoides*, and *Verneuilinoides.* Environment: Marine, likely of moderate depth in the shelf zone. Note: This locality is near Vesle Fiord locality 6 from which Hornaday in West et al. (1975, p. 575) reported a species of *Cyclammina*. Although we have not seen this collection, there is a reasonable basis for believing the species may be *Reticulophragmium borealis*, formerly described under *Cyclammina* (Petracca, 1972).	C-122637

Surface Sections (cont.)

Field No. & Stratigraphy	**Locality, Microfossils, Age and Environment**	**GSC Loc. No.**
85 RAK-6-2 Iceberg Bay Fm., Pillsbury Member, 13 m above base	Same locality, offset section Foraminifera: *Bathysiphon* spp., fine- to medium-grained *Saccammina* sp., coarse grained – two *Ammodiscus* sp. – one *Haplophragmoides* spp. *Reticulophragmium borealis* (Petracca) *Verneuilinoides* sp. miliolid genus indeterminate, poorly preserved – one specimen *Cibicidoides* sp. A (=*C.* sp. 3450 of McNeil, 1989) Age: Mid-Paleocene, as in previous entry, i.e. GSC loc. C-122637. Environment: Marine, likely of moderate depth in shelf zone.	C-122677
85 RAK-11-5 Uppermost Strand Bay Fm., or basal Iceberg Bay Fm. (Cape Pillsbury Member), 825 m above base section	South Bay, Canon Fiord area 79°35′N, 81°50′W; NTS 49H Foraminifera: *Saccammina(?)* sp., siliceous, elongate flask-shaped, may also be an altered calcareous unilocular form such as *Lagena* – one specimen *Ammodiscus* sp. – two *Ammodiscus* or *Miliammina* sp., poorly preserved – one *Haplophragmoides* sp., as at GSC locs. C-122637 and 122677, Vesle Fiord *Reticulophragmium borealis* (Petracca) *R.* sp. cf. *R. arctica* (Petracca) *Verneuilinoides* sp., as at Vesle Fiord *Age: Mid-Paleocene, as at GSC locs. C-122637 and 122677, Vesle Fiord.* Environment: Marine, shallow to moderate depth in shelf zone.	C-131363
84 RAK-2A-6 Strand Bay Fm., 50 m above base	Strathcona Fiord 78°28′N, 82°40′W; NTS 49E Foraminifera: *Haplophragmoides* sp. *Reticulophragmium borealis* (Petracca) – one Age: Mid-Paleocene, based on the occurrence of *R. borealis* at this level in the Mackenzie Delta (McNeil, 1989). Environment: Marine, shallow.	C-122568
84 RAK-2A-8 Strand Bay Fm., 83 m above base	Same locality Foraminifera: *Haplophragmoides* sp. *Reticulophragmium borealis* (Petracca) Age and Environment: As for previous entry. Note: This locality is near Strathcona Fiord locality 19 from which Hornaday in West et al. (1975, p. 575) reported the same species of *Cyclammina* as he had at Vesle Fiord. Our commentary on the Vesle Fiord locality is applicable here as well.	C-122570

Surface Sections (cont.)

Field No. & Stratigraphy	**Locality, Microfossils, Age and Environment**	**GSC Loc. No.**
84 RAK-3-1 Iceberg Bay Fm., Cape Pillsbury Member, 140 m above base	Strathcona Fiord 78°31′N, 82°40′W; NTS 49E Foraminifera: *Tritaxia(?)* sp. – one *Miliolinella* spp. *Nodosaria* spp. *Dentalina(?)* sp. – one incomplete *Marginulina(?)* sp. – one incomplete *Lagena acuticostata* Reuss – one *L.* spp. *Favulina* sp. – one nodosariinid indeterminate – internal fragments *Globulina* sp. – one *Cibicidoides* sp. B – prominent *Pullenia americana* Cushman – Serova (1966, pl. 9, figs. 1-3) recorded this species from the Paleocene of eastern Kamchatka *P.* spp. *Melonia* sp. cf. *M. nobilis* (Brotzen) — prominent. This form seems very close to *M. nobilis* s.s. as illustrated by Hansen (1970, pl. 14, figs. 8-10) from the Danian of West Greenland. genus indeterminate, possibly new – a trochospiral, coarsely perforate form with an interio-marginal aperture, probably belonging to subfamily *Pallaimorphininae* Loeblich and Tappan (1988) – fairly prominent in sample. Scaphopoda: *Dentalium* sp. – one Age: Mid-Paleocene, based on limited comparisons with faunas elsewhere and stratigraphic position above the assemblage characterized by *Reticulophragmium* borealis and *Cibicidoides* sp. A. Environment: Marine, moderate depth in outer portion of shelf.	 C-122522

Subsurface Section

Well: Panarctic Union Arco Talemen J-34
Location: 79°53′45"N, 83°47′00"W; NTS 49H
GSC Loc. No. C-39239
Rock unit: Eureka Sound Group, probably Strand Bay Formation.

Depth	Foraminifera, Age and Environment
213.4-243.8 m (700-800 ft.)	*Saccammina(?)* sp. – one *Haplophragmoides* sp.-spp., identical or closely related to those in Vesle Fiord, Section 85 RAK-2, sample no. 1, C-122637.
259.1-265.2 m (850-870 ft.)	*Saccammina* sp. *Haplophragmoides* spp., as above – common Age (213.4-265.2 m/700-870 ft): Paleocene indicated but not confirmed as the guide fossil *Reticulophragmium*

Surface Sections (cont.)

Field No. & Stratigraphy	Locality, Microfossils, Age and Environment	GSC Loc. No.
	borealis was not positively identified. Environment: Marine, shallow.	
268.2-310.9 m	*Bathysiphon* spp. – two specimens	
Depth (880-1020 ft)	Foraminifera, Age and Environment *Saccammina* sp. *Haplophragmoides* spp. *Reticulophragmium borealis* (Petracca) Age: Mid-Paleocene, based on the occurrence of *R. borealis* at this level in the Mackenzie Delta (McNeil, 1989). Environment: Marine, shallow to moderate depth in shelf zone.	

Summary

The foraminifera from the upper part of the Kanguk Formation at Slidre Fiord appear to be of mid to Late Campanian age. The lowest sample (C-131379) carries elements of both the *Dorothia smokyensis* and younger *Verneuilinoides bearpawensis* assemblages as defined by Wall (1983, p. 264-265) with an estimated age of mid-Campanian. Successively higher samples reveal a stronger affinity with the Late Campanian *V. bearpawensis* assemblage, supported by the occurrence of an associated diatom *Lepidodiscus* sp. cf. *L. elegans* in sample C-131380. The content of the highest sample C-131383, within the Kanguk/Expedition Fiord transitional interval, does not differ greatly from its predecessor and there is no evidence to support a younger age designation.

Three Paleocene assemblages were recognized in the general area by Wall et al. (1988), two of which are present in the Strand Bay and Iceberg Bay formations of west-central Ellesmere Island. Firstly, the *Reticulophragmium borealis* assemblage (Pls. 1, 2), with *Cibicidoides* sp. A as key associate species, is developed within the Strand Bay Formation at Strathcona Fiord (Section 84 RAK-2A), in the lower part of the Iceberg Bay Formation at Vesle Fiord, and in either the uppermost Strand Bay or basal Iceberg Bay beds at South Bay, Canon Fiord area. It also is present in what is likely the Strand Bay Formation in the Talemen J-34 well. The assemblage has been known for a decade or longer from the Mackenzie Delta area, where it represents a marine transgression in the Ministicoog Member of the Moose Channel Formation and is dated as mid-Paleocene, i.e. early Late Paleocene by McNeil (1989, p. 219)

Secondly, the *Melonis* sp. cf. *M. nobilis* assemblage (Pls. 3, 4), with *Cibicidoides* sp. B as key associate species, was recorded only at Strathcona Fiord (section 84 RAK-3), where it occurs in the Iceberg Bay Formation about 170 m above the preceding *R. borealis* assemblage in the Strand Bay Formation. The composition of this almost entirely calcareous fauna, which is also thought to be of mid-Paleocene age, suggests possible Atlantic affinities.

Although not as yet observed on Ellesmere Island, a third assemblage of brackish-water arenaceous foraminifera dominated by species of *Placentammina* (Pl. 5) is developed in the Strand Bay Formation at Strand Fiord (section 83 RAK-25), western Axel Heiberg Island. Previously (Wall et al. 1988; Wall in Ricketts 1991, Appendix 2, Part B), the fauna was reported to be characterized by species of *Saccammina* and *Hippocrepina* which are now assigned to the genus *Placentammina*. The composition of this assemblage partially resembles that of the *Portatrochammina* Assemblage Zone of McNeil (1989, p. 217), formerly described as the *Saccammina-Trochammina* assemblage by Young and McNeil (1984), from the Aklak Creek Member and higher strata in the Reindeer Formation of the Beaufort-Mackenzie Basin. The age of the *Portatrochammina* assemblage extends from late Late Paleocene to mid-Middle Eocene (McNeil, 1989), whereas the age of the *Placentammina* assemblage on Axel Heiberg Island is no younger than Late Paleocene from the associated palynomorphs.

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PLATE 1

Foraminifera of the *Reticulophragmium borealis* Assemblage
Iceberg Bay and Strand Bay Formations
West-central Ellesmere Island

Figure 1-3. *Bathysiphon* spp., Vesle Fiord. 1, 3 – GSC loc. C-122637, GSC 100150, 100151. 2 – GSC loc. C-122677, GSC 100152.

Figure 4. *Saccammina* sp., Vesle Fiord, GSC loc. C-122677, GSC 100153.

Figure 5. *Hyperammina* sp., Vesle Fiord, GSC loc. C-122637, GSC 100154.

Figure 6, 7. *Ammodiscus* sp., Vesle Fiord, GSC loc. C-122677, side and apertural views of GSC 100155.

Figure 8-15. *Haplophragmoides* spp. 8, 9 – South Bay, GSC loc. C-131363, side and peripheral views of GSC 100156. 10 – Talemen J-34 Well, GSC loc. C-39239/213.4-243.8 m (700-800 ft.), GSC 100157. 11– Vesle Fiord, GSC loc. C-122677, GSC 100158. 12-15 – Vesle Fiord, GSC loc. C-122637, side and peripheral views of GSC 100159, 100160.

Figure 16-21. *Recurvoides* spp., Vesle Fiord, GSC loc. C-122637, side, peripheral and opposite side views of GSC 100161, 100162.

Plate 1

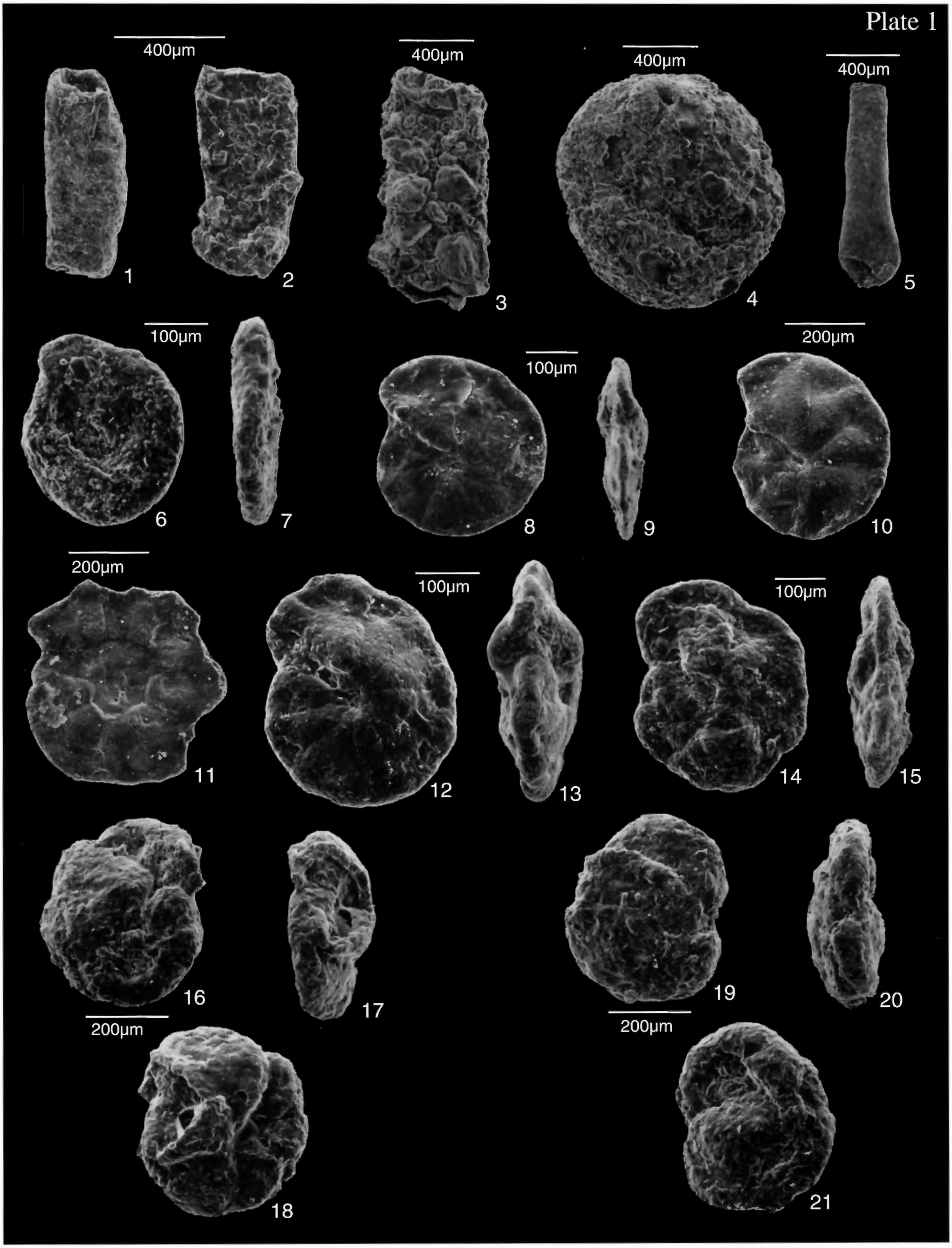
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PLATE 2

Foraminifera of the *Reticulophragmium borealis* Assemblage
Iceberg Bay and Strand Bay Formations
West-central Ellesmere Island

Figure 1-4, 7, 8.	*Reticulophragmium borealis* (Petracca). 1-4 – Vesle Fiord, GSC loc. C-122637, side and peripheral views of GSC 100163, 100164. 7, 8 – Talemen J-34 Well, GSC loc. C-39239/286.5-292.6 m (940-960 ft.), side and peripheral views of GSC 100165 (note partial removal of outer wall revealing alveolar structure).
Figure 5, 6.	*Reticulophragmium* sp. cf. *R. arctica* (Petracca), South Bay, GSC loc. C-131363, side and peripheral views of GSC 100166.
Figure 9-13.	*Verneuilinoides* spp., Vesle Fiord, GSC loc. C-122637. 9, 10 – opposite side views of GSC 100167, 11 – GSC 100168; 12, 13 – opposite side views of GSC 100169.
Figure 14-16.	*Tritaxia* sp., Vesle Fiord, GSC loc. C-122637, opposite side and apertural views of GSC 100170.
Figure 17-22.	*Cibicidoides* sp. A, Vesle Fiord, GSC loc. C-122637, opposite side and peripheral views of GSC 100171, 100172.

Plate 2

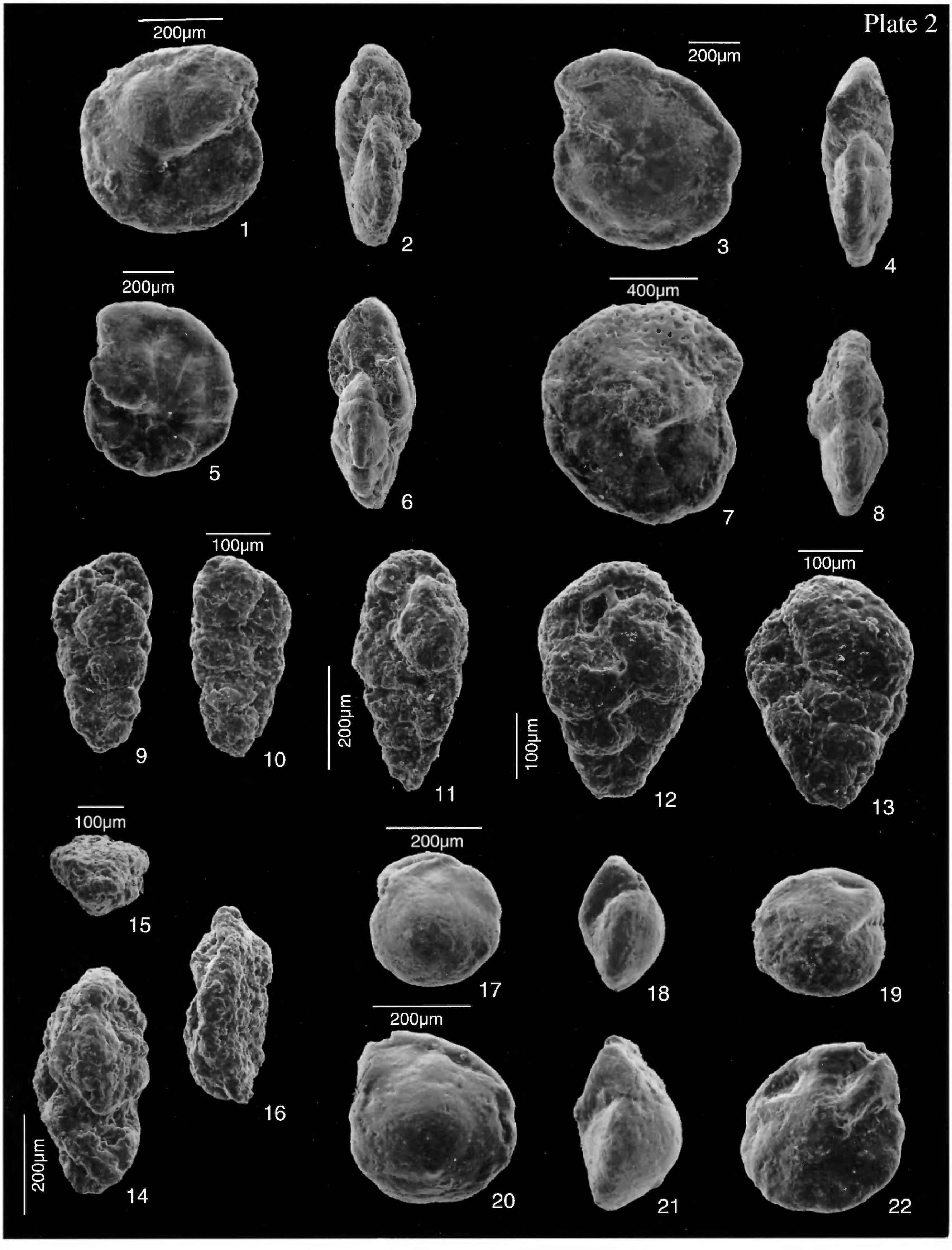
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PLATE 3

Foraminifera of the *Melonis* sp. cf. *M. nobilis* Assemblage
Iceberg Bay Formation
Strathcona Fiord, GSC loc. C-122522
West-central Ellesmere Island

Figure 1-6. *Miliolinella* spp., opposite side and peripheral views of GSC 100173, 100174.

Figure 7-9. *Nodosaria* spp. 7, 8— side and apertural views of GSC 100175; 9 – GSC 100176.

Figure 10. *Lagena* sp., GSC 100177.

Figure 11. *Lagena acuticostata* Reuss, GSC 100178.

Figure 12, 13. *Favulina* sp., side and apertural views of GSC 100179.

Figure 14, 15. *Oolina* sp., side and apertural views of GSC 100180.

Figure 16-19. *Melonis* sp. cf. *M. nobilis* (Brotzen), side and apertural views of GSC 100181, 100182.

Figure 20, 21. *Pullenia americana* Cushman, side and peripheral views of GSC 100183.

Figure 22, 23. *Pullenia* sp., side and peripheral views of GSC 100184.

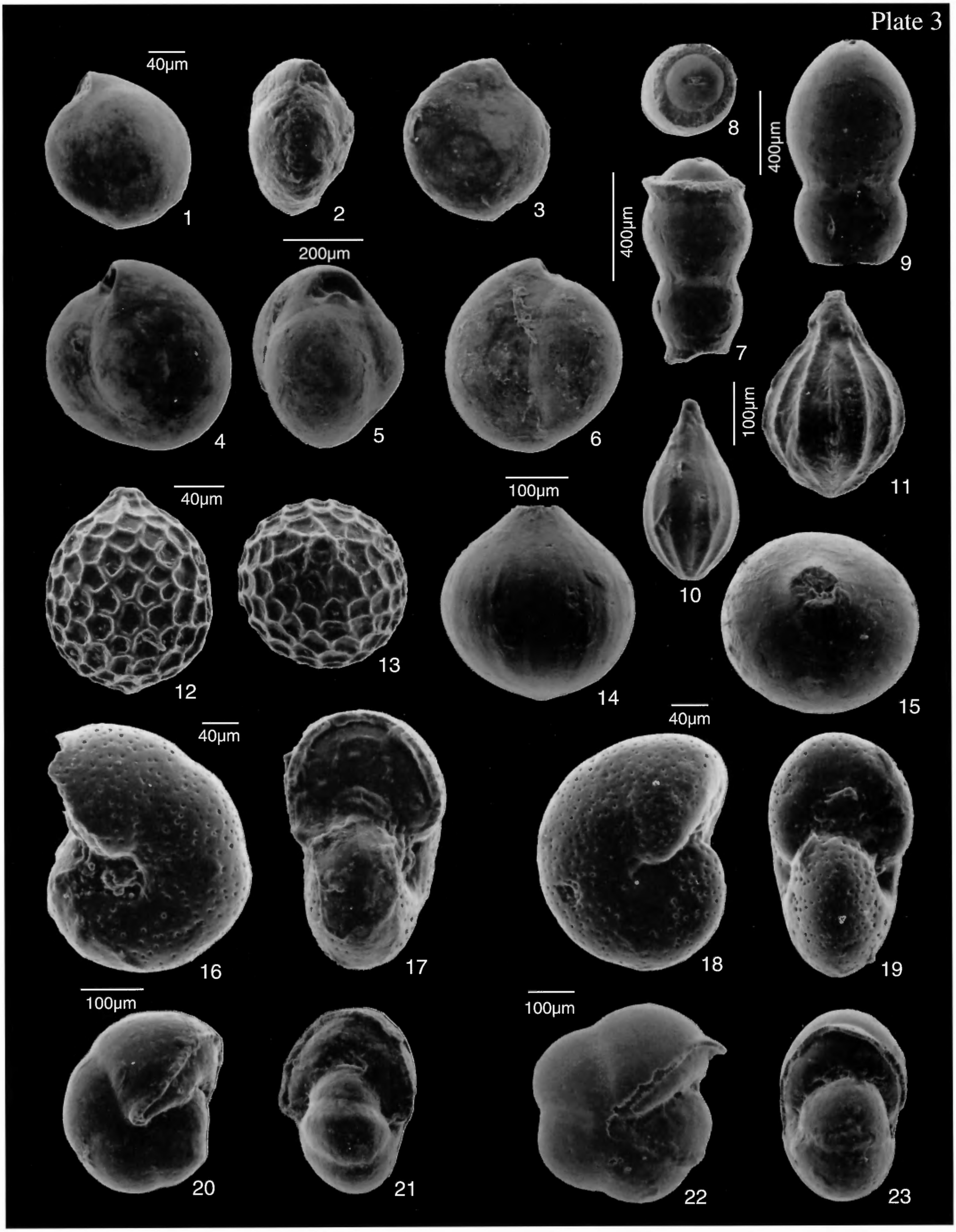
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PLATE 4

Foraminifera of the *Melonis* sp. cf. *M. nobilis* Assemblage
Iceberg Bay Formation
Strathcona Fiord, GSC loc. C-122522
West-central Ellesmere Island

Figure 1-10. *Cibicidoides* sp. B. 1-3 – opposite side and peripheral views of GSC 100185; 4-6 – opposite side and peripheral views of GSC 100186; 7-9 – opposite side and peripheral views of GSC 100187; 10 – interior view of GSC 100188.

Figure 11-19. Genus indeterminate, Subfamily Pallai- morphininae(?) Loeblich and Tappan (1988). 11-13 – opposite side and peripheral views of GSC 100189; 14-16 – opposite side and peripheral views of GSC 100190; 17-19 – opposite side and peripheral views of GSC 100191.

Plate 4

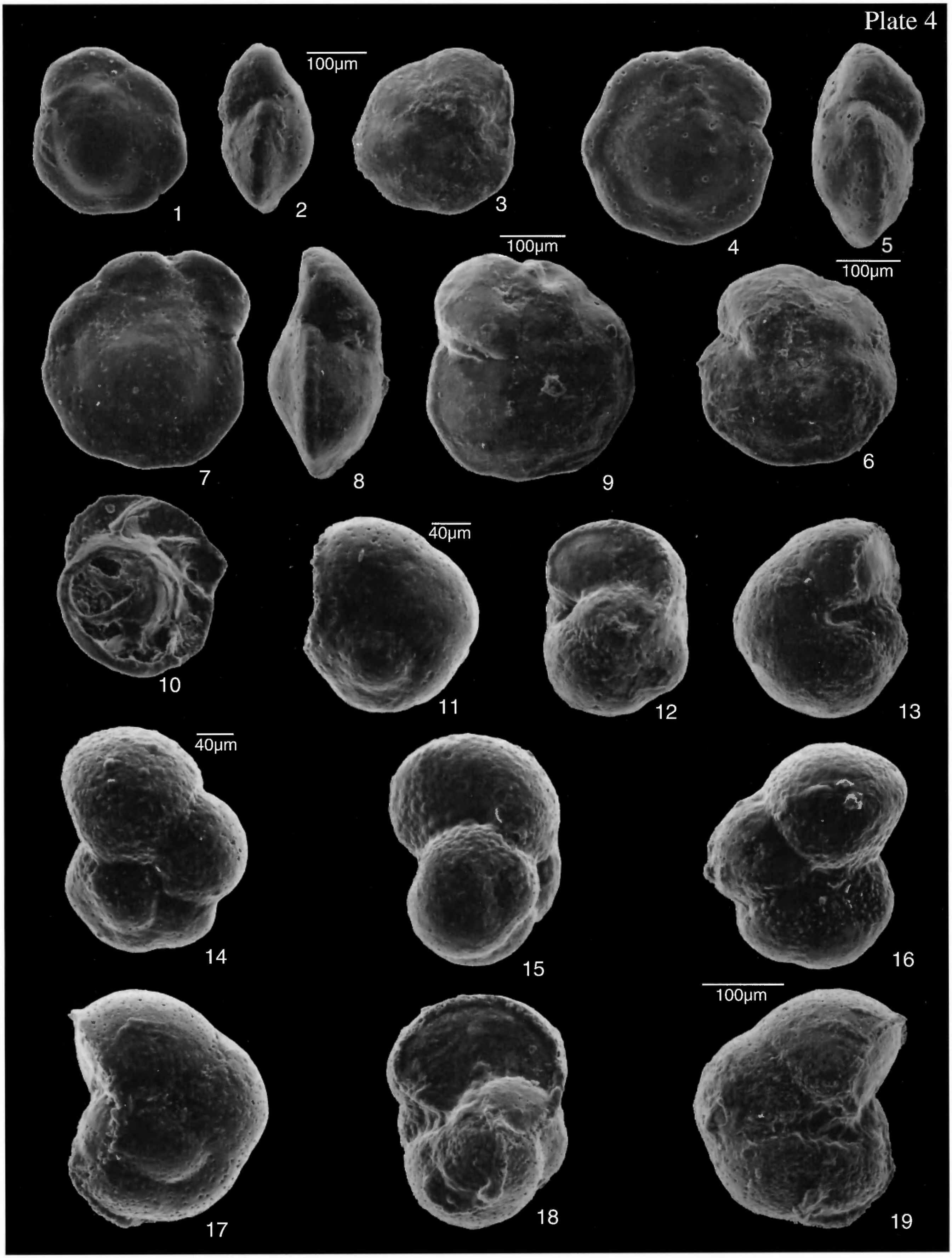
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PLATE 5

Foraminifera of the *Placentammina* spp. Assemblage Strand Bay Formation, Strand Fiord Western Axel Heiberg Island (Metres in brackets after GSC locality numbers are above base of formation)

Figure. 1-5, 7-13, 19. *Placentammina* spp. 1, 2 – GSC loc. C-111789 (27.3 m), GSC 100192, 100193. 3, 9, 11, 12 – GSC loc. C-111792 (48.3 m), GSC 100194, 100195, 100196, and 100197. 4 – GSC loc. C-111806 (267.9 m), GSC 100198. 5, 8, 10, 13 – GSC loc. C-111788 (21.3 m), GSC 100199, 100200, 100201 and 100202. 7 – GSC loc. C-111791 (39.3 m), GSC 100203. 19 – GSC loc. C-111801 (199 m), GSC 100204.

Figure 6. *Saccammina* sp., GSC loc. C-111791 (39.3 m), GSC 100205.

Figure 14, 17, 18. *Hippocrepina*(?) spp. 14 – GSC loc. 111792 (48.3 m), GSC 100206; 17 – GSC loc. 111789 (27.3 m), GSC 100207; 18 – GSC loc. 111788 (21.3 m), GSC 100208.

Figure 15, 16. *Hippocrepina* spp., GSC loc. C-111788 (21.3 m), GSC 100209, 100210.

Figure 20-23.3. *Miliammina*(?) sp. 20, 21 – GSC loc. C-111801 (199 m), opposite side views of GSC 100211; 22, 23 – GSC loc. C-111806 (267.9 m), opposite side views of GSC 100212.

Plate 5


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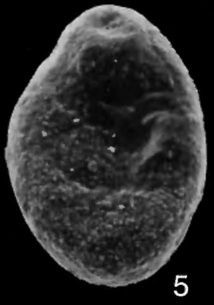
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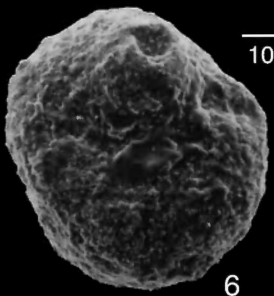
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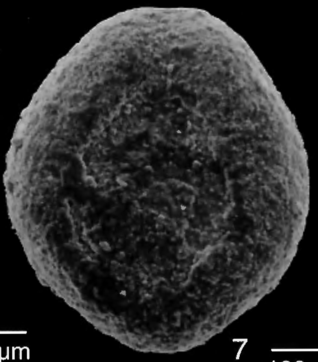
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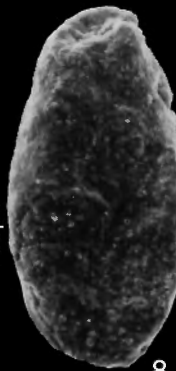
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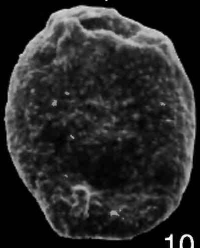
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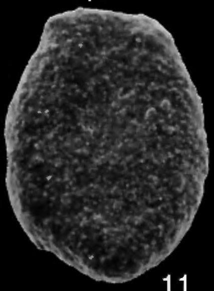
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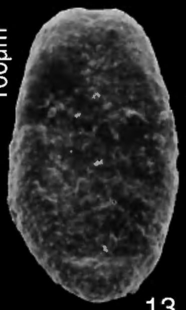
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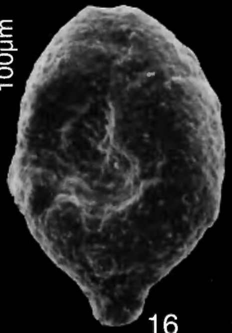
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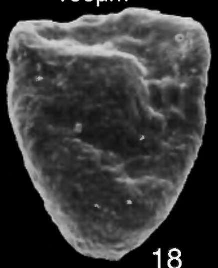
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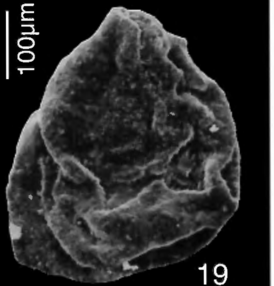
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