Geological Survey of Canada
Commission géologique du Canada

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CURRENT RESEARCH PART G
FRONTIER GEOSCIENCE PROGRAM, ARCTIC CANADA

RECHERCHES EN COURS PARTIE G
PROGRAMME GÉOSCIENTIFIQUE DES RÉGIONS PIONNIÈRES, RÉGION ARCTIQUE DU CANADA

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COMMISSION GÉOLOGIQUE DU CANADA
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PROGRAMME GÉOSCIENTIFIQUE DES RÉGIONS PIONNIÈRES, RÉGION ARCTIQUE DU CANADA

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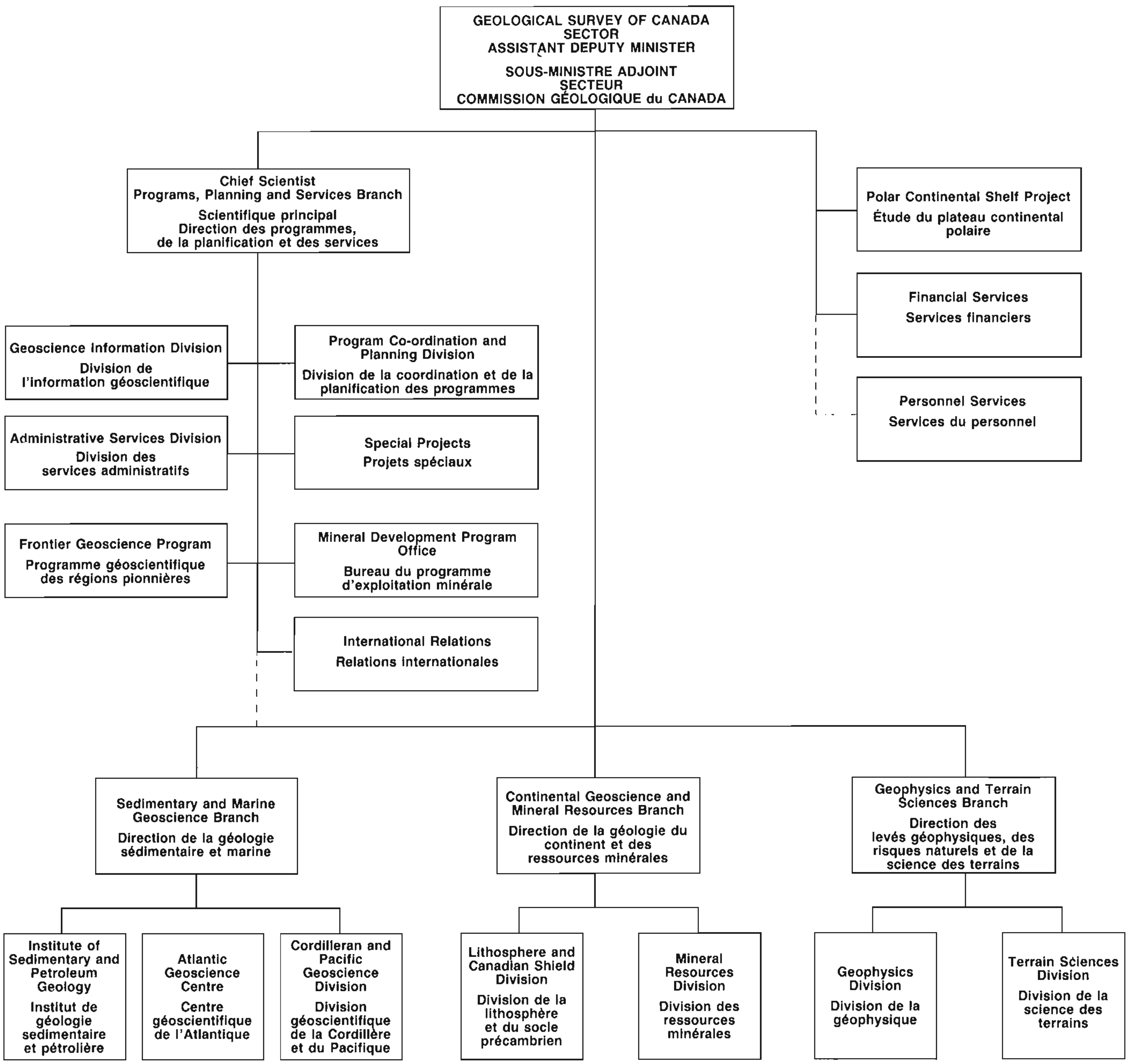
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INTRODUCTION

The Frontier Geoscience Program (FGP) was initiated in 1984 to respond to two events that placed major new demands on the Geological Survey of Canada (GSC). These were the discovery of oil and gas in the north and the east coast offshore, and the extension of Canadian boundaries offshore to 200 miles, or where wider, to the edge of the continental shelf. This latter event resulted in an expansion of Canadian jurisdiction and commercial interest over an ocean area larger than all the provinces of Canada.

The Frontier Geoscience Program is a long-term effort to address information needs likely to extend well into the next century. Recently reviewed by Cabinet after four years in operation, the program has now been made a permanent part of the Geological Survey's ongoing funding base. The present level of funding is $15.3 million annually.

The challenge presented to the GSC in developing the Frontier Geoscience Program was to provide to both government and industry the same fundamental geoscience information for areas offshore and in the north as that available for areas where oil and gas exploration is already established, such as in the Western Canada Sedimentary Basin. Such data are deemed essential for orderly and responsible policy formulation, exploration, development and protection of the environment.

In general terms, the objective of the program is to ascertain the geological history and development of all sedimentary basins offshore and in the north where oil and gas resources might occur. More specifically, the goals of the Frontier Geoscience Program are to:

1. Establish the deeper geological controls affecting the development of the sedimentary basins in the north and offshore.
2. Outline the internal geology and evolution of the basins.
3. Elucidate the processes governing the generation, accumulation and preservation of hydrocarbon resources.
4. Identify and analyze hazards and constraints to development.
5. Provide the essential supporting research, development, analysis and synthesis.

INTRODUCTION

Le Programme géoscientifique dans les régions pionnières (PGRP) a été établi en 1984 en réponse à deux événements qui ont créé des nouvelles demandes importantes auprès de la Commission géologique du Canada (CGC). Il s'agit de la découverte de pétrole et de gaz extra-côtiers dans l'Est et dans le Nord ainsi que l'élargissement des frontières canadiennes jusqu'à 200 milles au large des côtes, ou, à certains endroits, jusqu'à l'extrémité de la plate-forme continentale. Ce dernier événement s'est traduit par l'expansion de la juridiction et des intérêts commerciaux canadiens sur une superficie marine plus grande que toutes les provinces du Canada.

Le Programme géoscientifique dans les régions pionnières vise, à long terme, à combler les besoins d'information qui demeureront probablement pendant une bonne partie du siècle prochain. Examiné récemment par le Cabinet après quatre ans de mise en œuvre, le Programme fait maintenant partie du budget permanent de la Commission géologique du Canada. Actuellement, 15,3 millions de dollars y sont consacrés par année.

Lors de l'élaboration du Programme géoscientifique dans les régions pionnières, le personnel de la CGC relevait le défi de fournir au gouvernement et à l'industrie, pour les régions extra-côtières et celles du Nord, la même information géoscientifique de base que celle qui est disponible pour les régions où l'exploration du pétrole et du gaz est déjà établie, comme dans le bassin sédimentaire de l'Ouest du Canada. Cette information est jugée essentielle à l'exécution méthodique et responsable des activités de formulation de politiques, d'exploration, de mise en valeur et de protection de l'environnement.

En gros, l'objectif du Programme est d'établir l'histoire géologique et les antécédents de la formation de tous les bassins sédimentaires situés en mer et dans le Nord, susceptibles de contenir des ressources pétrolières et gazières. Plus particulièrement, le Programme géoscientifique dans les régions pionnières vise à:

1. établir les contrôles géologiques profonds qui influent sur la formation des bassins sédimentaires dans le Nord et en mer;
2. définir la géologie interne et l'évolution des bassins;

6. Provide a database to support assessment of the nature and distribution of the hydrocarbon potential of a particular basin.

An important feature of the program is the emphasis on dissemination of information to users. This volume is a selected compendium of the most recent results from FGP field activities in the Arctic, and complements the talks and poster sessions presented at the Oil and Gas Forum in February, 1989 in Calgary. Results from recent field activities on the West and East coasts are showcased at the Vancouver Round-up and the Ottawa Forum, respectively.

In addition to this presentation of results immediately following the field season, several hundred papers in GSC publications and scientific and technical journals have been produced since the inception of the program. It is expected that most of the significant geoscience information resulting from the program will also be presented in an atlas format, with a separate atlas for each basin or group of basins. All data available for each basin will be compiled and synthesized into a separate volume. These will provide a complete package for each of the significant sedimentary basins and permit easier updating as additional data become available.

In their most complete form, each atlas will contain crustal maps, geophysical maps, cross-sections, plate reconstructions, maps of tectonic elements, a complete suite of stratigraphic maps and sections, hydrocarbon potential assessments, bathymetric maps, surficial maps, as well as development hazard maps. In addition, an annotated bibliography will be assembled for each atlas. The first atlas, for the Labrador Sea, is scheduled for release in 1988-1989.

To facilitate management of the FGP, the program was divided into four regional tasks and assigned to the Geological Survey's regional centres.

Western Arctic

Studies are focused on extending and enhancing the knowledge of presently known oil and gas areas in the MacKenzie Delta, Beaufort Sea, and adjacent areas by gathering and analyzing seismic, well-log, paleontological, structural and other geoscience information, in order to provide a better regional framework for further exploration and development. However, the Western Arctic poses unique challenges due to its harsh environment. Sea ice, icebergs, permafrost, frozen gas hydrates and poorly consolidated rocks are but a few of the problems with major geoscience implications. As plans for exploitation of the resources of the MacKenzie Delta and Beaufort Sea develop, geoscience information will hold the key to the design of production and transportation facilities in the region.

The long range objective of this task is to produce comprehensive geological syntheses covering the various sedimentary basins of the MacKenzie Delta, Beaufort Sea and northern interior regions.

The level of information obtained over such a large and relatively unknown area will vary greatly, but the syntheses will cover basic geological structure, origin, and evolution of the sedimentary basins, and the nature and distribution of oil and gas resources that they contain.

3. décrire les processus qui régissent la formation, l'accumulation et la préservation des ressources en hydrocarbures;
4. déterminer et analyser les dangers et les contraintes liés à la mise en valeur;
5. effectuer la recherche, le développement, les analyses et les synthèses essentiels à la mise en valeur;
6. constituer une base de données permettant d'évaluer la nature et la répartition des ressources potentielles en hydrocarbures d'un bassin donné.

La communication d'information aux utilisateurs représente un élément important du Programme. Le présent volume constitue un condensé des plus récents résultats choisis qui ont découlé des activités du PGRP réalisées dans l'Arctique. Il complète les exposés ainsi que les expositions d'affiches présentés au Forum sur les activités gazières et pétrolières, en février 1989, à Calgary. Les résultats des récentes activités réalisées sur la côte ouest et la côte est sont exposés au public au cours de forums qui se tiennent à Vancouver, et à Ottawa, respectivement.

En plus des ces présentations de résultats, qui suivent immédiatement la saison des travaux sur le terrain, plusieurs centaines d'articles ont paru dans les publications de la CGC et dans des revues scientifiques et techniques depuis le début du Programme. On s'attend à ce que la plupart des informations géoscientifiques importantes résultant du Programme soient également présentées sous forme d'atlas, soit un atlas pour chaque bassin ou groupe de bassins. Toutes les données disponibles pour chaque bassin seront compilées et rassemblées en un volume distinct. L'ensemble formera un ouvrage complet pour chacun des bassins sédimentaires importants et facilitera la mise à jour à mesure que des données supplémentaires seront recueillies.

Dans sa présentation la plus complète, chaque atlas comprendra des cartes lithosphériques, des cartes géophysiques, des coupes transversales, des reconstructions de plaques lithosphériques, des cartes d'éléments tectoniques, une série complète de cartes et de coupes structurales, des évaluations des ressources en hydrocarbures non découvertes, des cartes bathymétriques, des cartes des formations en surface ainsi que des cartes indiquant les risques pour la mise en valeur des ressources. En outre, chaque atlas comprendra une bibliographie annotée. La publication du premier atlas, portant sur la plate-forme du Labrador et de l'île Baffin, est prévue pour 1989.

Pour faciliter la gestion du PGRP, le Programme a été divisé en quatre volets régionaux et attribué aux centres régionaux de la Commission géologique du Canada.

Arctique occidental

Ce volet comprend des études qui portent principalement sur l'élargissement et l'amélioration de la connaissance des régions pétrolières et gazières connues dans le delta du MacKenzie, la mer de Beaufort et les régions adjacentes par la collecte et l'analyse de données sur la sismologie, sur la diagraphie de forage, sur la paléontologie, sur la structure et sur d'autres domaines géoscientifiques en vue de constituer

Arctic Islands

The Arctic Islands represent one of the least explored of the frontier areas, but 169 wells have been drilled to date. Knowledge of the stratigraphy and deep structural controls of the sedimentary basins is available on only a reconnaissance scale for most of this region. A number of relatively large oil and gas discoveries make this an area of considerable future interest. Basic geological work will provide the foundation for estimating resource potential and planning exploration activities.

Activities for the Western Arctic and Arctic Islands tasks are conducted from the Institute of Sedimentary and Petroleum Geology, Calgary.

East Coast

In order to provide a better regional framework for further exploration and development, the East Coast task concentrates on the analysis and synthesis of information relating to presently known oil and gas areas, and on the gathering of seismic, well log, paleontological, structural and other geoscience information. Research on the structure, rock types and deformation of the East Coast continental margins will aid in assembling a regional overview, as a necessary precursor to exploration in the deep water of the continental slopes. Activities required in support of this program range from seismic surveys for both deep structures and new surface detail, to geochemical analyses required to determine the origins of oil and gas. Surveys of the type and distribution of ocean bottom sediments and shore and nearshore material will be necessary for production engineering, design, regulation and environmental control.

The long range objective of this task is the production of a comprehensive geological synthesis covering the various sedimentary basins of the East Coast offshore, from the Canada-U.S.A. border to the continental shelf off Baffin Island. While the level of information required over such a large area will obviously vary greatly, the syntheses will cover basic geological structure, origin and evolution of the sedimentary basins and the oil and gas resources they contain.

Activities for the East Coast task are conducted from the Atlantic Geoscience Centre, Dartmouth.

West Coast

Petroleum exploration on the West Coast has been limited during the past decade by a moratorium on drilling west of Vancouver Island and in the Queen Charlotte Sound area. Regional studies and seismic surveys have recently been conducted in the Queen Charlotte Basin, however, and data from wells already drilled will be included in the regional syntheses necessary to estimate the hydrocarbon potential in the area. In anticipation of the possibility of increased drilling activity in the next few years, the program will also focus on problems related to ocean bottom stability, seismic risk, and other potential environmental problems. Surveys of seabed geological conditions potentially constituting constraints or hazards to resource development will also be included.

un meilleur cadre pour d'autres explorations et mises en valeur dans la région. Cependant, l'Arctique occidental pose un défi unique en raison des conditions environnementales qui y règnent. Les glaces de mer, les icebergs, le pergélison, les hydrates de gaz gelés et la faible consolidation des roches ne sont que quelques-uns des problèmes qui ont des conséquences géoscientifiques importantes. À mesure que les plans d'exploitation des ressources du delta du MacKenzie et de la mer de Beaufort seront élaborés, les données géoscientifiques joueront un rôle clé en ce qui a trait à la conception des installations de production et de transport dans la région.

L'objectif à long terme de ce volet est de produire des synthèses géologiques globales portant sur les divers bassins sédimentaires du delta du MacKenzie, de la mer de Beaufort et des régions intérieures du Nord.

Les quantités d'informations obtenues au sujet d'une région aussi vaste et relativement inconnue varieront beaucoup, mais les synthèses comprendront la structure géologique fondamentale, l'origine et l'évolution des bassins sédimentaires ainsi que la nature et la répartition des ressources pétrolières et gazières qu'ils contiennent.

Archipel Arctique

L'archipel Arctique constitue l'une des régions pionnières les moins explorées, même si 169 puits y ont déjà été forés jusqu'à présent. Dans la majeure partie de cette région, seule une reconnaissance de la stratigraphie et des contrôles structuraux profonds des bassins sédimentaires a été effectuée. La découverte d'un certain nombre de gisements de pétrole et de gaz relativement importants rendent cette région très intéressante pour l'avenir. Des travaux géologiques de base permettront d'estimer les ressources non découvertes et de planifier les activités d'exploration.

En ce qui concerne l'Arctique occidental et l'archipel Arctique, les activités sont réalisées à partir de l'Institut de géologie sédimentaire et pétrolière de Calgary.

Côte est

Afin de fournir un meilleur cadre pour les activités futures d'exploration et de mise en valeur sur la côte est, ce volet met l'accent sur l'analyse et la synthèse de données concernant les régions pétrolières et gazières connues actuellement ainsi que sur la collecte de données liées à la sismologie, à la diagraphie de forage, à la paléontologie, à la structure, et à d'autres domaines géoscientifiques. Les recherches sur la structure, le type de roches et la déformation de la marge continentale de la côte est aideront à constituer un aperçu régional, outil nécessaire à l'exploration dans les eaux profondes du talus continental. Les activités servant à appuyer ce programme comprennent des relevés sismiques portant sur des structures profondes et de nouveaux détails superficiels ainsi que les analyses géochimiques requises pour déterminer l'origine du pétrole et du gaz. Il sera nécessaire d'obtenir des relevés sur le type et la répartition des sédiments du fond de l'océan ainsi que sur les matériaux qui constituent le littoral et la zone périlittorale pour les travaux techniques, la conception, la réglementation et le contrôle environmental de la production.

Activities for the West Coast task are conducted from the Cordilleran and Pacific Geoscience Division offices in Vancouver, and Sidney, British Columbia.

L'objectif à long terme de ce volet est de préparer une synthèse géologique globale sur les divers bassins sédimentaires de la côte est allant de la frontière Canada-États-Unis jusqu'à la plate-forme continentale située au large de l'île Baffin. Bien qu'il soit évident que les quantités d'informations concernant une région si vaste varieront beaucoup, les synthèses comprendront la structure géologique fondamentale, l'origine et l'évolution des bassins sédimentaires ainsi que les ressources pétrolières et gazières qu'ils renferment.

Pour ce qui est de la côte est, les activités sont réalisées à partir du Centre géoscientifique de l'Atlantique, de Dartmouth.

Côte ouest

Au cours de la dernière décennie, l'exploration pétrolière sur la côte ouest a été limitée par un moratoire sur le forage à l'ouest de l'île de Vancouver et dans la région du détroit de la Reine-Charlotte. Toutefois, des études régionales et des levés sismiques ont été effectués récemment dans le bassin de la Reine-Charlotte, et des données sur les puits déjà forés seront incluses dans les synthèses régionales nécessaires pour estimer les ressources d'hydrocarbures non découvertes de cette région. En prévision de la possibilité que les activités de forage augmentent au cours des prochaines années, le Programme mettra également l'accent sur les problèmes liés à la stabilité des fonds marins, aux risques sismiques et à d'autres problèmes environnementaux en puissance. Il comprendra également des relevés des conditions géologiques du fonds marin qui représentent des contraintes ou des dangers possibles pour la mise en valeur des ressources.

En ce qui concerne la côte ouest, les activités sont effectuées à partir des bureaux de la Division géoscientifique de la Cordillière et du Pacifique, situés à Vancouver et à Sydney (Colombie-Britannique).

Massive ice: some field criteria for the identification of ice types[†]

J. Ross Mackay[1]
Terrain Sciences Division

Mackay, J.R., Massive ice: some field criteria for the identification of ice types; in Current Research, Part G, Geological Survey of Canada, Paper 89-1G, p. 5-11, 1989.

Abstract

There are many field criteria that can be used to identify different types of massive ice. Important criteria are whether the ice contacts are gradational or sharp, conformable or unconformable, thaw or erosional, and primary or secondary. A close-up examination of the bubbles in the ice, just below a contact, can help to differentiate between buried and segregated ice. Dykes and sills of intrusive ice transecting massive ice or the overburden appear associated with segregated ice. The orientation of reticulate ice veins above a contact can help in the identification of ice types.

Résumé

De nombreux critères peuvent être utilisés sur le terrain pour identifier différents types de masses de glace. La nature des contacts entre les types de glaces sont des critères importants : s'agit-il de contacts progressifs ou nets, concordants ou discordants, dus au gel ou à l'érosion, primaires ou secondaires. Un examen détaillé des bulles dans la glace, juste sous un contact, peut aider à distinguer entre de la glace enfouie et de la glace de ségrégation. Les dykes et les filons-couches de glace intrusive recoupant de la glace massive ou des morts-terrains anastomosés semblent associés à la glace de ségrégation. L'orientation des filons de glace au-dessus d'un contract peut aider à identifier les types de glace.

[†] Contribution to Frontier Geoscience Program

[1] Department of Geography, University of British Columbia, Vancouver, British Columbia V6T 1W5.

INTRODUCTION

Massive ice (Fig. 1), which can be defined as a large mass of ground ice with a gravimetric ice content exceeding 250 % (Permafrost Subcommittee, 1988), is widespread in many permafrost areas of the world. In Canada, massive ice is present in many of the Arctic Islands such as Axel Heiberg, Banks, Ellesmere, King Christian, Melville, and Victoria; it is abundant along the western arctic coast; and massive ice is present in interior Yukon Territory (Carter et al., 1987; Dallimore and Wolfe, 1988; Dallimore et al., 1988; French, 1987; French and Pollard, 1986; Fujino et al., 1988; Harry et al., 1985; Harry and French, 1988; Mackay, 1958, 1963, 1966, 1971, 1973; Mackay and Stager, 1966; Pollard and Dallimore, 1988; Rampton, 1974, 1982, 1988a, 1988b; Rampton and Mackay, 1971; Rampton and Walcott, 1974; Shah, 1978). If pingo ice is excluded, little massive ice has been reported from the coastal plain of northern Alaska (Lawson, 1986). Massive ice is widespread in the U.S.S.R. (e.g., Danilov, 1987; Dubikov, 1982; Popov, 1982; Solomatin, 1986); and it occurs in China and elsewhere. This paper discusses some field criteria that may be helpful in the identification of different types of massive ice with specific reference to the western Arctic coast (Fig. 2).

TYPES OF MASSIVE ICE

Most massive ice can be divided into two main types. First, buried ice can be river, lake, sea, snowbank, and glacier ice. Burial of areally large bodies of river, lake, and sea ice is very rare; snowbanks of small size can be buried. The literature on buried glacier ice in the U.S.S.R. is voluminous (e.g., Baulin, 1972; Danilov, 1987; Dubikov, 1982; Karpov, 1981; Koreisha et al., 1981; Popov, 1982; Solomatin, 1986); but that for Canada is sparse (e.g., Dallimore and Wolfe, 1988; Lorrain and Demeur, 1985; Rampton, 1988b). In general, no clear distinction is usually made between buried glacier ice derived from snow and buried subglacier regelation ice, which, genetically, is segregated ice.

The second main type of massive ice forms by the in situ freezing of water within sediments to form massive segregated ice. If bulk water is intruded under pressure and then freezes, the ice is massive intrusive ice. If the water pressure fluctuates between the lower pressure required to form massive segregated ice and the higher pressure required to uplift the overburden and form massive intrusive ice, then segregated-intrusive ice forms.

MASSIVE ICE CONTACTS

A close-up examination of the contact of massive ice with the enclosing material provides many clues to the identification of ice type. Usually, only the upper contact is visible. Lower contacts, although rarely seen, are commonly more diagnostic than upper contacts. A close-up study can rarely be carried out in summer because of problems associated with thaw and accessibility.

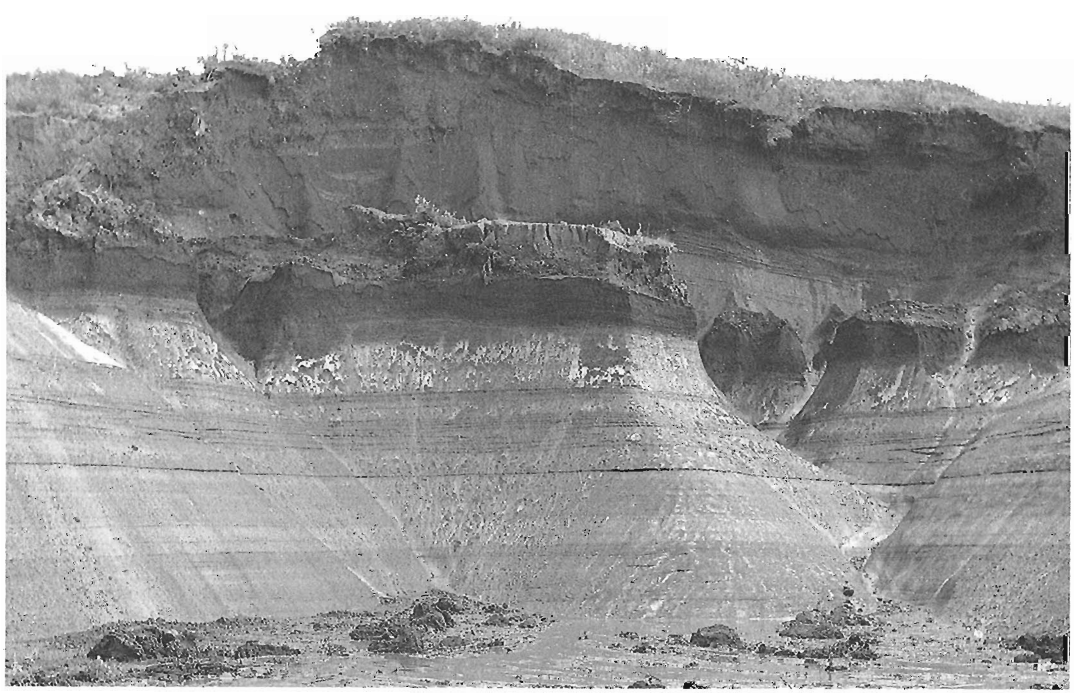

Figure 1. Massive segregated ice more than 6 m thick on the mainland coast to the east of Nicholson Peninsula (cf. Fig. 2).

Gradational contacts

Some massive ice exposures have a gradational contact between the ice and adjacent material (Fig. 1, 3a). The gradational zone can range from a few centimetres to several metres. A gradational contact can only develop where the ice is approximately of the same age (i.e., syngenetic) or is younger (i.e., epigenetic) than the enclosing material. Gradational contacts should rarely exist with buried ice, because the ice must predate the material that buried it. Gradational contacts are typical of massive segregated ice. Gradational contacts may also occur with intrusive ice where suspended fragments below the contact are common (Fig. 3b). An example is the gradational contact sometimes present between the overburden of a pingo and the ice core where segregated ice grades downward into intrusive ice.

Conformable contacts

Conformable contacts have an undisturbed relationship between the massive ice and the adjacent sediments (Fig. 1). Conformable contacts are usually found with massive segregated ice when the freezing plane parallels the bedding of the sediments. Conformable contacts in segregated ice commonly show bubbles extending downward from the soil/ice contact (Fig. 3c). If water is intruded along a bedding plane and then freezes, the resulting intrusive ice can have conformable contacts. In some cases water may be intruded along a fracture in permafrost (Fig. 4) with both conformable and unconformable relations. Lake and river ice buried by sedimentation could have either a conformable or an unconformable upper contact; the lower contact should be unconformable. Buried glacier ice should have unconformable upper and lower contacts.

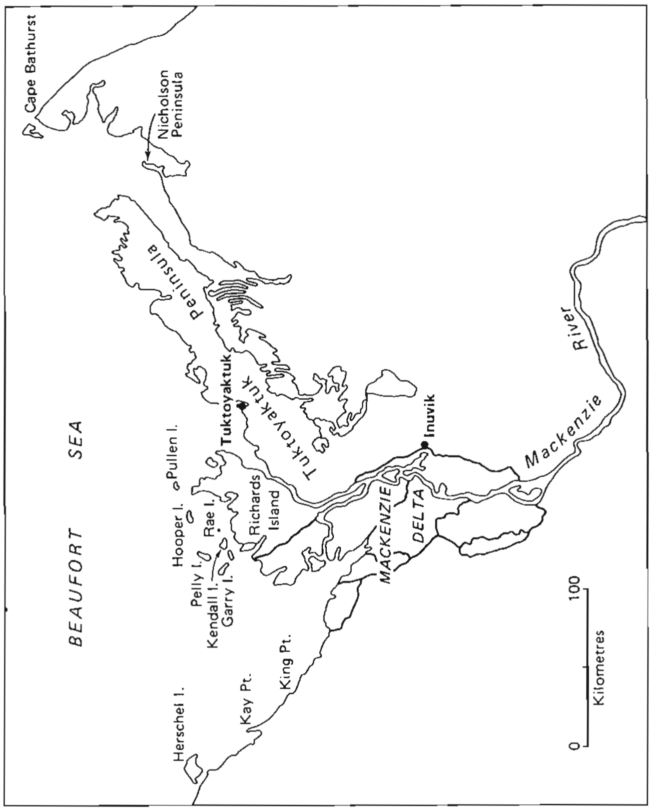

Figure 2. Location map.

Thaw and erosional contacts

The initial contact formed between the ice and the overlying material is here referred to as a primary contact.

Primary thaw or erosional contacts between an overlying material and underlying ice indicate that the ice predates

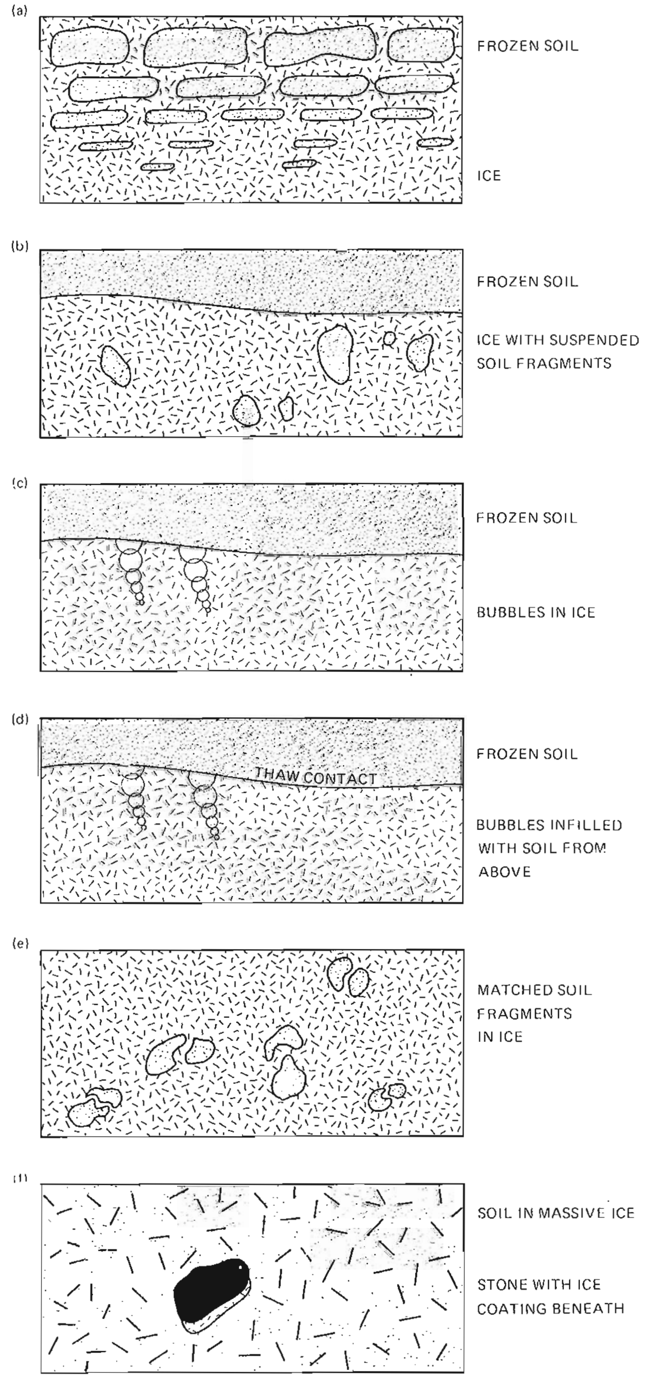

Figure 3. Features commonly seen in close-up studies of massive ice. See text.

burial; therefore, the ice must be buried ice. Because most ice has bubbles, a primary thaw or erosional contact will often have the bubbles, just below the contact, infilled with sediment from above, especially if the sediment is sandy (Fig. 3d). Care must be taken, however, to ensure that the bubble infilling is not a near surface secondary feature resulting from recent summer thaw. Examination of a fresh section well back from the ice face is required for the use of the infilled bubble criterion. In addition, discolouration from the downward movement of ''dirty'' water may sometimes be seen in some of the intercrystalline ice boundaries just below a thaw contact. The discolored zone is usually no more than a few centimetres thick. If the buried ice has cracks, such as in river, lake, and glacier ice, infilling of these cracks at the time of burial would be expected. The infilling of bubbles beneath secondary thaw or erosional contacts, however, provides no indication of ice type. For example, secondary thaw contacts are abundant where retrogressive thaw slides produce thaw unconformities (Fig. 5).

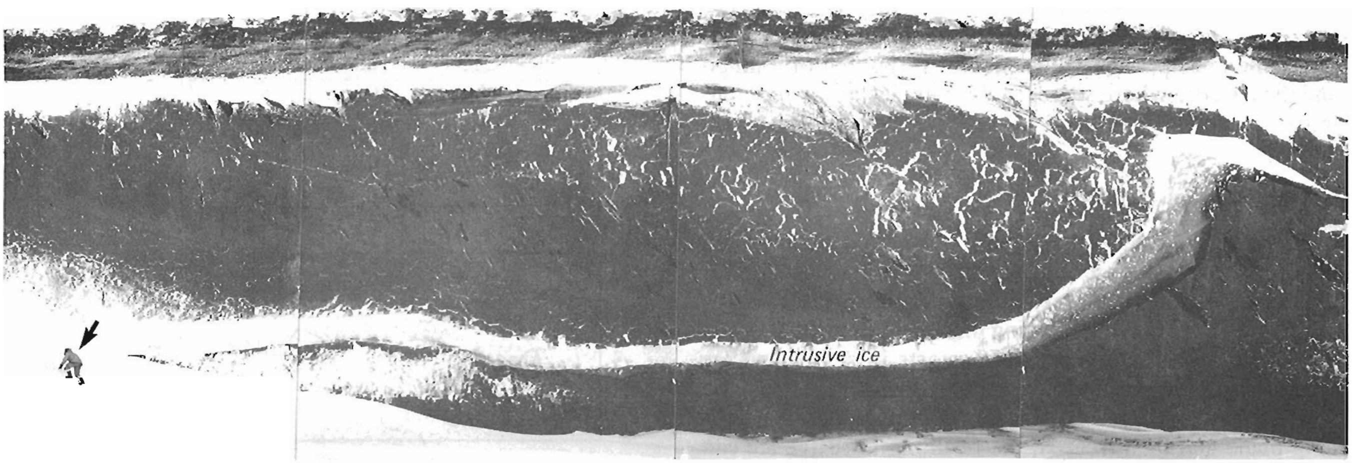

Figure 4. Intrusive ice in glacially deformed sediments, Pelly Island (cf. Fig. 2); note the person (marked by an arrow) for scale.

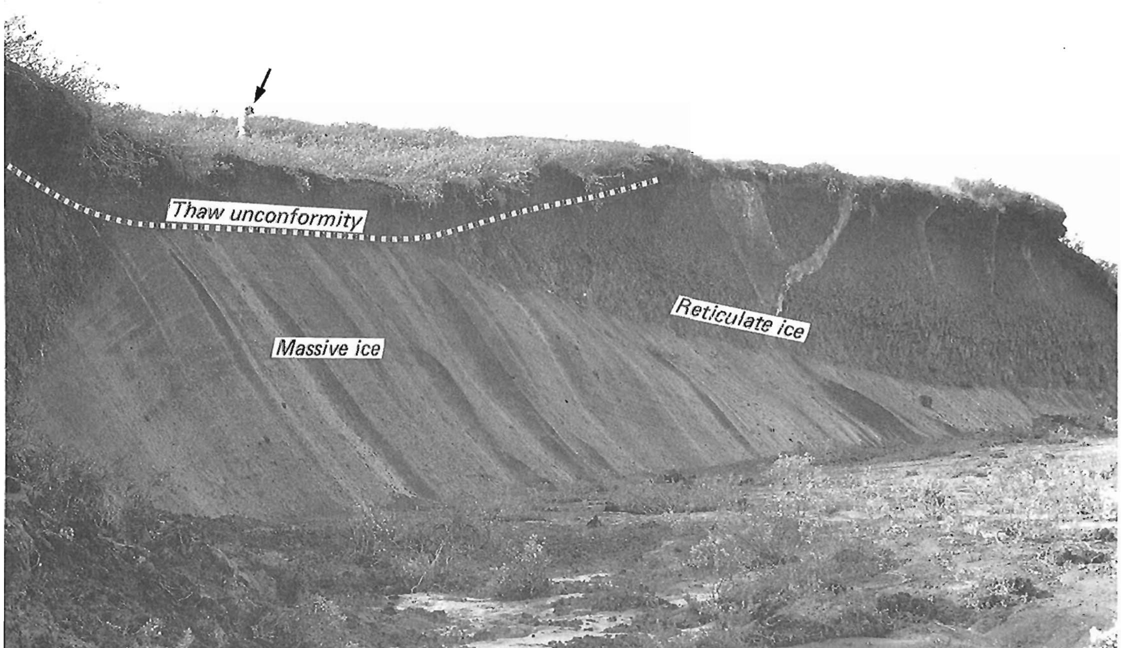

Figure 5. Massive ice, believed to be segregated ice, southeast of Tuktoyaktuk (cf. Fig. 2). Note person for scale. The main system of reticulate ice veins is perpendicular to the contact. The anticline has probably resulted because of unloading associated with mass wasting. See text.

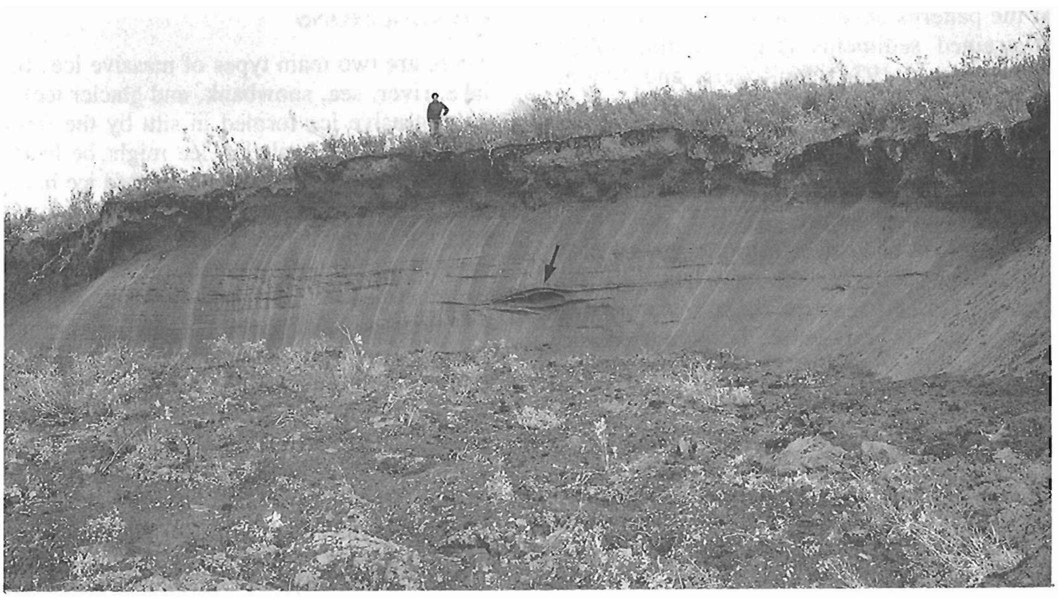

Figure 6. A clay pod (marked by an arrow), in massive segregated ice to the southeast of Tuktoyaktuk; note person for scale.

Figure 7. An intrusive ice dyke cutting through massive segregated ice, Garry Island. The top of the dyke has been truncated by a secondary thaw unconformity. The ice dyke has matching sides and chill zones on the sides. The water source was from below the bottom of the exposure.

FEATURES FOUND IN MASSIVE ICE

A variety of features suggestive of ice origin may be found within the ice. Matched soil fragments within the ice (Fig. 3e) usually indicate segregated ice and less commonly intrusive ice. If the massive ice is stony, some stones may have a clear ice coating on one side (Fig. 3f). Such ice coatings may be found in contemporary areas of ice lens growth and indicate a segregated origin. Soil pods (Fig. 6) in massive ice also suggest a segregated origin.

Dykes and sills of intrusive ice

Water is sometimes intruded from depth under high pressure into permafrost along fractures. The intruded water may freeze to form dykes, sills, and other features similar to those formed by the intrusion and solidification of magma in bedrock. Fracture contacts commonly have: straight to angular sections; matched "walls"; chill zones with small ice crystals at the contact; detached fragments of the host material "suspended" in the ice; candle ice normal to the two cooling surfaces; indications of two-sided freezing with a central bubbly seam like the central seam of wedge ice; and discoloured, typically brownish to yellowish, ice in the last-to-freeze zone of solute concentration. A dyke of intrusive ice cutting through massive ice at Garry Island, Northwest Territories (Fig. 2) is show in Figure 7. Figure 8 shows an ice dyke extending upward from the well known Peninsula Point massive ice exposure, about 6 km southwest of Tuktoyaktuk, Northwest Territories. Because the ice dyke of Figure 8 originates at the top of the massive ice and extends upward 8 m into the overburden, the ice dyke must therefore be younger than the overburden. Inasmuch as the ice dyke appears continuous with the massive ice, by inference the massive ice must also be younger thant the overburden. Similar dykes have been observed in the U.S.S.R. (Karpov, 1981). Dykes of intrusive ice show, conclusively, that the dykes are younger than the enclosing material and that water movement must have been upward from a pressurized water source at depth. Ice dykes are to be expected with massive segregated ice and segregated-intrusive ice, because the growth of such ice types requires water under pressure.

Reticulate ice

A three-dimensional reticulate (lattice) pattern of ice veins is commonly present in fine grained sediments, irrespective of the presence or absence of massive ice. Although opinions differ as to the origin of reticulate ice, there is general agreement that the patterns develop during the freezing of saturated fine grained sediments (e.g., Baulin, 1972; Dubikov, 1982; Mackay, 1974; McRoberts and Nixon, 1975; Vtyurin, 1975). Reticulate ice veins are common in fine grained sediments above massive ice. The best developed ice veins tend to be perpendicular to the upper contact of the massive ice irrespective of whether the contact of the massive ice is horizontal or inclined (Fig. 5, 8). Three important conclusions can be drawn from the above: 1) If the massive ice is buried ice, then the reticulate ice veins are younger than the buried ice so the contact is a thaw or erosional contact. 2) If the massive ice is segregated ice, then the massive ice is younger than the reticulate ice and the contact is a downward freezing contact. 3) If reticulate ice veins are approximately perpendicular to an inclined conformable contact, then the ice is segregated ice that has been deformed (Fig. 5, 8).

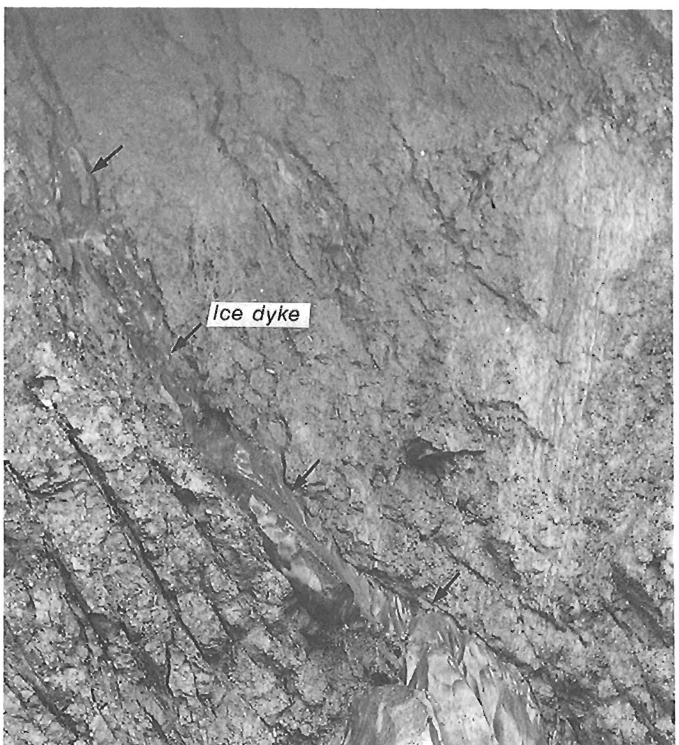

Figure 8. An ice dyke that originates at the top of massive ice, Peninsula Point exposure, 6 km southwest of Tuktoyaktuk. The dyke extends upwards about 8 m into the overburden. The ice is clear with a chill zone on the two sides and with bubbles extending inward, normal to the two sides, a sure sign of two-sided inward freezing. Detached wall fragments occur in the ice. The central part of the dyke is discoloured yellowish in the last-to-freeze portion. The dyke also branched upward, a sign of upward crack propagation.

CONCLUSIONS

There are two main types of massive ice: buried ice (e.g., lake, river, sea, snowbank, and glacier ice) and segregated and intrusive ice formed in situ by the freezing of water. Buried glacier regelation ice might be hard to distinguish from some in situ grown segregated ice because both types involve ice lensing. If a contact between massive ice and the adjacent material can be examined close-up, a study best done in winter, many diagnostic criteria can be helpful in ice identification. Gradational contacts are to be expected with segregated ice but not buried ice. Primary thaw and/or erosional contacts are expected with buried ice but not with segregated and intrusive ice. Secondary thaw and/or erosional contacts can occur with all ice types. Intrusive ice, in the form of ice dykes and ice sills, indicates a pressurized water source at depth and is to be expected in association with some segregated and intrusive ice. Ice dykes and sills should be rare with buried ice. A close-up study of bubbles in the ice at a soil-ice contact and the identification of features within the ice, such as soil pods or matching soil fragments, can provide valuable clues to ice type.

ACKNOWLEDGMENTS

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A summary of Carboniferous and Permian biostratigraphy, northern Yukon Territory and northwest District of Mackenzie[†]

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Bamber, E.W., Henderson, C.M., Jerzykiewicz, J., Mamet, B.L., and Utting, J., A summary of Carboniferous and Permian biostratigraphy, northern Yukon Territory and northwest District of Mackenzie; in Current Research, Part G, Geological Survey of Canada, Paper 89-1G, p. 13-21, 1989.

Abstract

Abundant invertebrate fossils and palynomorphs occur in (?)Tournaisian to Wordian strata in the Richardson and British mountains, the Barn Range, and the Ogilvie Mountains - Eagle Plain area. The strata comprise continental, and shelf to basinal marine facies. Regional and international correlation of this succession is based on biostratigraphic information combining age data from brachiopods, foraminifers, palynomorphs, corals, ammonoids, and conodonts. New information from recent studies in the Ogilvie Mountains - Eagle Plain area includes a preliminary account of the Viséan to Asselian conodont succession and a preliminary palynological zonation dated by associated marine faunas. Throughout most of the succession, the faunas and palynofloras show close relationships with correlative fossil assemblages from the Sverdrup Basin and with those from more southerly parts of the Eastern Cordillera and Plains of Western Canada.

Résumé

Un grand nombre de palynomorphes et de fossiles d'invertébrés sont présents dans des couches (?) tournaisiennes à wordiennes dans les monts Richardson et British, dans les chaînons Barn et la région des monts Ogilvie et de la plaine d'Eagle. Les couches comprennent un faciès continental et un faciès marin dont le type varie de celui de plateau à celui de bassin. La corrélation régionale et internationale de cette série est basée sur des données biostratigraphiques combinant des données d'âge provenant de brachiopodes, de foraminifères, de palynomorphes, de coraux, d'ammonoïdes et de conodontes. De nouvelles données provenant d'études récentes sur la région des monts Ogilvie et de la plaine d'Eagle comprennent un relevé préliminaire de la série des conodontes de la période s'étendant du Viséen à l'Assélien, et une zonation palynologique préliminaire datée par des faunes marines associées. Dans la plus grande partie de la série, les faunes et les palynoflores présentent des rapports étroits avec des associations de fossiles corrélatifs provenant du bassin de Sverdup et avec ceux des parties plus méridionales de l'est de la Cordillère et des plaines de l'Ouest canadien.

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INTRODUCTION

Carboniferous and Permian rocks of the study area (Fig. 1) contain well preserved invertebrate marine fossils and palynomorphs ranging in age from (?)late Tournaisian to Wordian. The succession outcrops mainly in the Ogilvie Mountains - Eagle Plain area, the northern Richardson Mountains and the British Mountains (Fig. 1). It is continuous with a similar succession found in adjacent Alaska (Wood and Armstrong, 1975), and consists of shallow shelf to basinal marine carbonates and siliciclastics, with subordinate nonmarine deposits in the British and Richardson mountains and Snake River area.

Carboniferous rocks in the study area are separated from older strata by a regional unconformity that resulted from uplift and erosion during the Ellesmerian Orogeny. A second regional unconformity separates Permian from underlying Upper Carboniferous strata throughout the area. Latest Carboniferous to Early Permian tectonism is indicated by regional and local truncation of Carboniferous and older strata beneath the disconformity (Bamber and Waterhouse, 1971). Extensive latest Carboniferous and Early Permian erosion and redeposition at this level resulted in the widespread occurrence of abundant, reworked Carboniferous and older palynomorphs and marine invertebrate fossils in Lower Permian units. An essentially continuous Carboniferous succession of Viséan to Moscovian carbonates and siliciclastics is preserved beneath the sub-Permian disconformity. Younger Carboniferous deposits occur only in the Ogilvie Mountains, where Kasimovian to Gzhelian strata are present, and sedimentation may have been locally continuous across the Carboniferous/Permian boundary (Fig. 1, 2; Waterhouse and Waddington, 1982). Tournaisian strata have been definitely identified only in the southeast, near the Snake River (Bamber and Waterhouse, 1971, p. 50). The Permian succession ranges from Asselian to Wordian in age and is separated from Triassic and younger strata by a regional disconformity. For an outline of upper Paleozoic stratigraphy and depositional and tectonic history of the area, the reader is referred to recent summary papers by Richards et al. (in press) and Henderson et al. (in press).

BIOSTRATIGRAPHY

Formal upper Paleozoic lithostratigraphic nomenclature and biostratigraphic zonation for the study area were established by Bamber and Waterhouse (1971). Their biostratigraphic scheme was based mainly on brachiopods and foraminifers, supplemented by age determinations from isolated occurrences of ammonoids, palynomorphs and corals. Subsequent research, funded by the Geological Survey of Canada Frontier Geoscience Program, has been centred mainly in the Ogilvie Mountains, southern Eagle Plain, and British Mountains. It has resulted in a preliminary zonation of the Carboniferous and Lower Permian palynofloras, recognition of Lower Carboniferous coral zones previously established in the southern Cordillera (Sando and Bamber, 1985), further detailed investigation of the Upper Carboniferous and Lower Permian non- fusulinacean foraminifers, and the first investigations of upper Paleozoic conodont faunas in the area. The age relationships of these faunas and floras and their distribution within the lithostratigraphic units recognized in the study area are discussed below and shown in Figures 2 and 3. Biostratigraphic correlation within the area and with standard chronostratigraphic units are generally supported by data from all phyla studied. Minor discrepancies, yet to be resolved, are discussed under the appropriate fossil groups.

Brachiopods

Brachiopod zones and faunas recognized in the northern Yukon and northwest District of Mackenzie are discussed in Bamber and Waterhouse (1971), Sarytcheva and Waterhouse (1972), Waterhouse and Waddington (1982), Nelson (1961, 1962), and Nelson and Johnson (1968).

Waterhouse (in Bamber and Waterhouse, 1971) established several Lower and Upper Carboniferous brachiopod assemblage zones ranging in age from Viséan to Gzhelian. This zonal scheme was modified (Fig. 2) by Waterhouse and Waddington (1982). In the Eagle Plain and the Ogilvie Mountains, the Viséan to Serpukhovian *Quadratia* brachiopod zone occurs in the Hart River Formation and has been tentatively identified in the lowest part of the Blackie Formation. Bashkirian brachiopods have been recognized in the Wahoo Formation of the British Mountains, and Bashkirian to Gzhelian brachiopods are present in the Blackie and Ettrain formations of the Eagle Plain and Ogilvie Mountains. Zones proposed for these post-Serpukhovian faunas include the upper Bashkirian to lower Moscovian *Orthotichia, Composita*, and *Martiniopsis* zones; the middle to upper Moscovian (Desmoinesian) *Buxtonia* and *Praehorridonia* (=*Reticulatia*, =*Gemmulicosta*) zones; the upper Moscovian(?) (Desmoinesian) *Purdonella* and *Gibbospirifer* zones; the Kasimovian or lower Gzhelian (Missourian or lower Virgilian) *Kozlowskia* Zone; and the (?)Gzhelian *Orthotichia-Septospirifer* and overlying, unnamed zones.

Seven Permian brachiopod assemblage zones have been recognized in the JungleCreek Formation of the northern Ogilvie Mountains (Fig. 3; Waterhouse and Waddington, 1982). These include three Asselian zones (*Kochiproductus-Attenuatella*; *Orthotichia*; and *Attenuatella-Tomiopsis*) and four Sakmarian (Tastubian to Aktastinian) zones (*Yakovlevia; Attenuatella; Tornquistia;* and *Jakutoproductus*). Deposition in this area was probably continuous across the Carboniferous/Permian boundary, but to the east and south of the northern Ogilvie Mountains, several brachiopod zones are missing at the boundary (Bamber and Waterhouse, 1971; Waterhouse and Waddington, 1982). The lower Tahkandit Formation of the northern Ogilvie Mountains includes two zones (*Antiquatonia* and *Sowerbina*) that can be correlated with the Baigendzhinian (Artinskian). The middle and upper Tahkandit Formation has been subdivided into the *Pseudosyrinx* and *Thuleproductus* zones (Roadian) and the overlying *Cancrinelloides* zone (Wordian). In the unnamed Permian succession of the northern Richardson Mountains, four Permian zones have been recorded — the *Yakovlevia, Neochonetes, Lissochonetes*, and *Cancrinelloides* zones. The *Neochonetes* and *Lissochonetes* zones are correlative with the *Pseudosyrinx* and *Thuleproductus* zones, respectively, of the Ogilvie Mountains. Several of

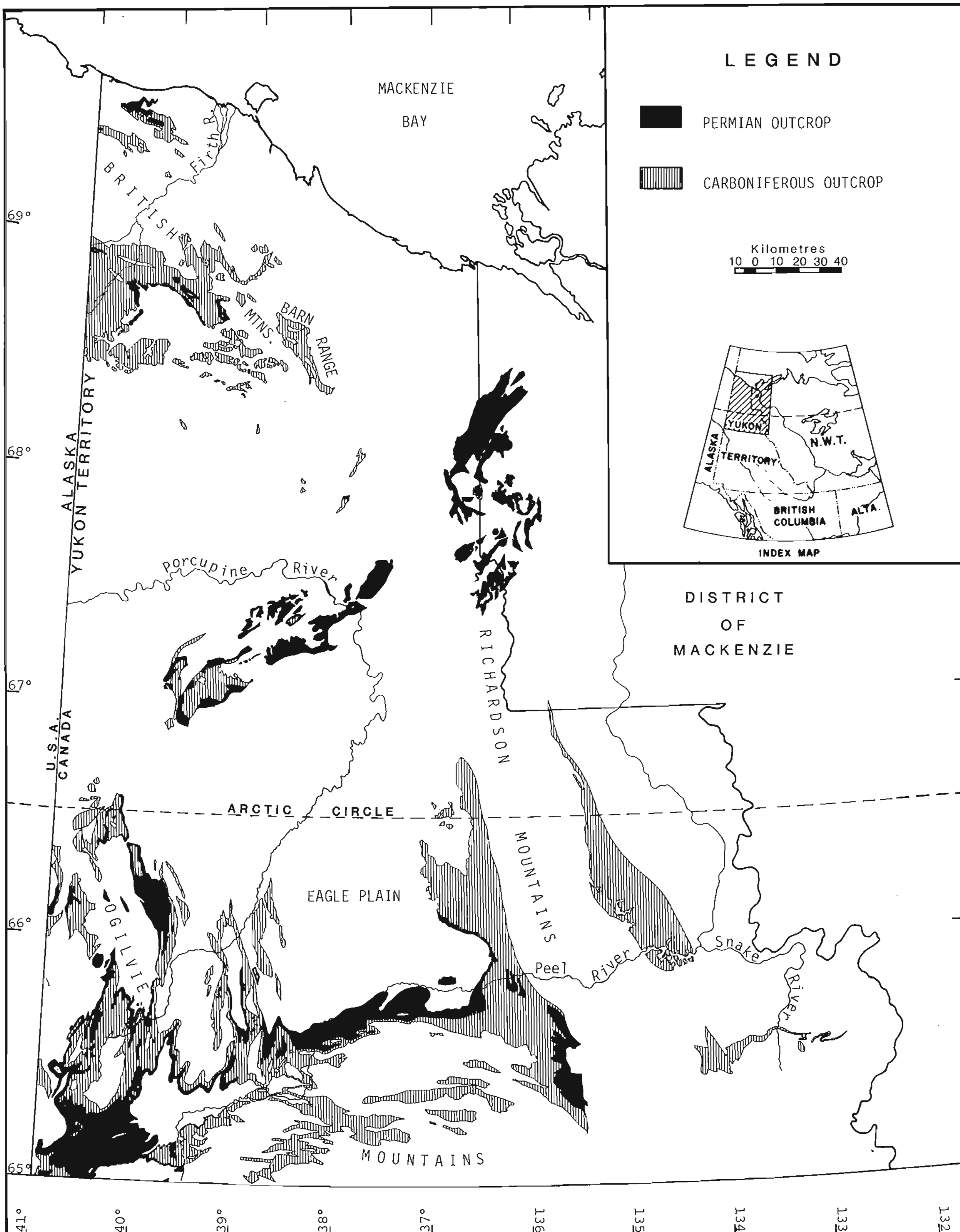

Figure 1. Distribution of Carboniferous and Permian deposits, northern Yukon Territory and northwestern District of Mackenzie.

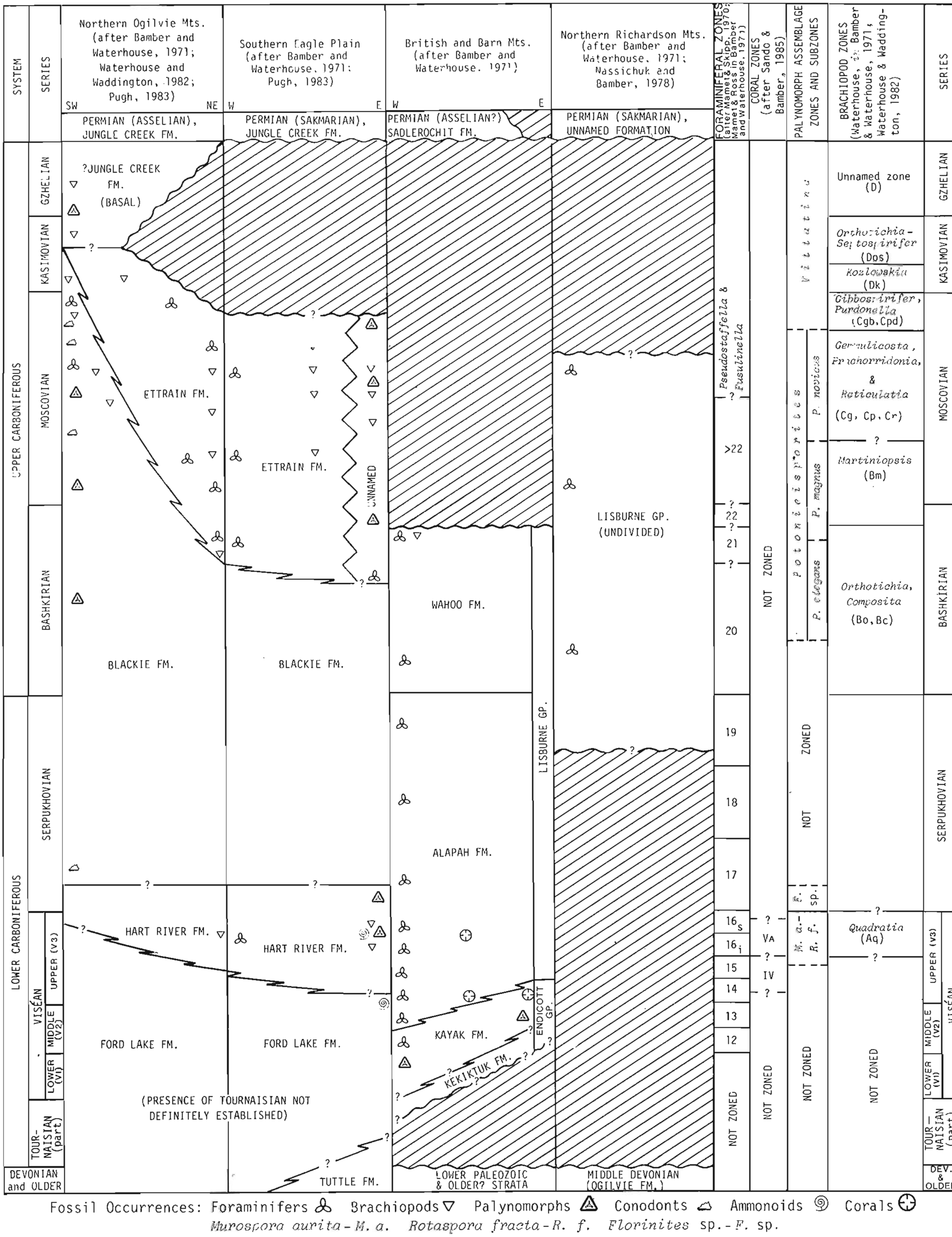

Figure 2. Carboniferous correlations and biostratigraphy, northern Yukon Territory and northwestern District of Mackenzie. (After Richards et al., in press.)

the Permian brachiopod faunas occurring in the study area show strong similarity to faunas from correlative strata of the southern Cordillera and the Sverdrup Basin in the Arctic Archipelago (Bamber and Waterhouse, 1971, p. 146, 172). The correlation with standard Permian chronostratigraphic units used here (Fig. 3) is based mainly on data from brachiopod zones and differs slightly from that based on detailed conodont studies in the Canadian Arctic Archipelago (Beauchamp et al., 1989).

Foraminifers

In the British Mountains and Barn Range, Mamet (in Bamber and Waterhouse, 1971) identified a continuous succession of Carboniferous foraminifer zones (Fig. 2) within shelf carbonates of the Endicott and Lisburne groups. These assemblage zones range in age from middle Viséan to late Bashkirian and are part of a comprehensive, international zonal scheme (Mamet and Skipp, 1970) used for correlation of Carboniferous rocks throughout the Cordillera of Western Canada and in the Sverdrup Basin of the Canadian Arctic Archipelago. Zone 12 (middle Viséan) occurs in the upper part of the Kayak Formation and Zones 13 to 21 (middle Viséan to late Bashkirian) are present in the overlying Alapah and Wahoo formations. The regional distribution of these zones indicates that the Kayak - Lisburne contact is strongly diachronous on a regional scale, becoming younger toward the north and east (Fig. 2; Mamet, 1984).

In the Ogilvie Mountains and southern Eagle Plain, Mamet (ibid.) reported upper Viséan foraminifer Zone 16 from the Hart River Formation, and Serpukhovian (lower Namurian) to lower Moscovian Zones 17 to >22 from shallow shelf and slope carbonates of the Blackie and Ettrain formations. Recent investigations by Mamet have demonstrated the presence of Moscovian or younger Carboniferous foraminifer faunas (younger than Zone 22) in the Blackie Formation of the northern Ogilvie Mountains (B.L. Mamet, unpublished data). These faunas had previously been known only from the upper Ettrain Formation in the study area.

Upper Bashkirian fusulinaceans were identified by Ross (1967; in Bamber and Waterhouse, 1971) in foraminifer Zone 21 within the Wahoo Formation of the British Mountains. Equivalent and younger fusulinaceans also occur at

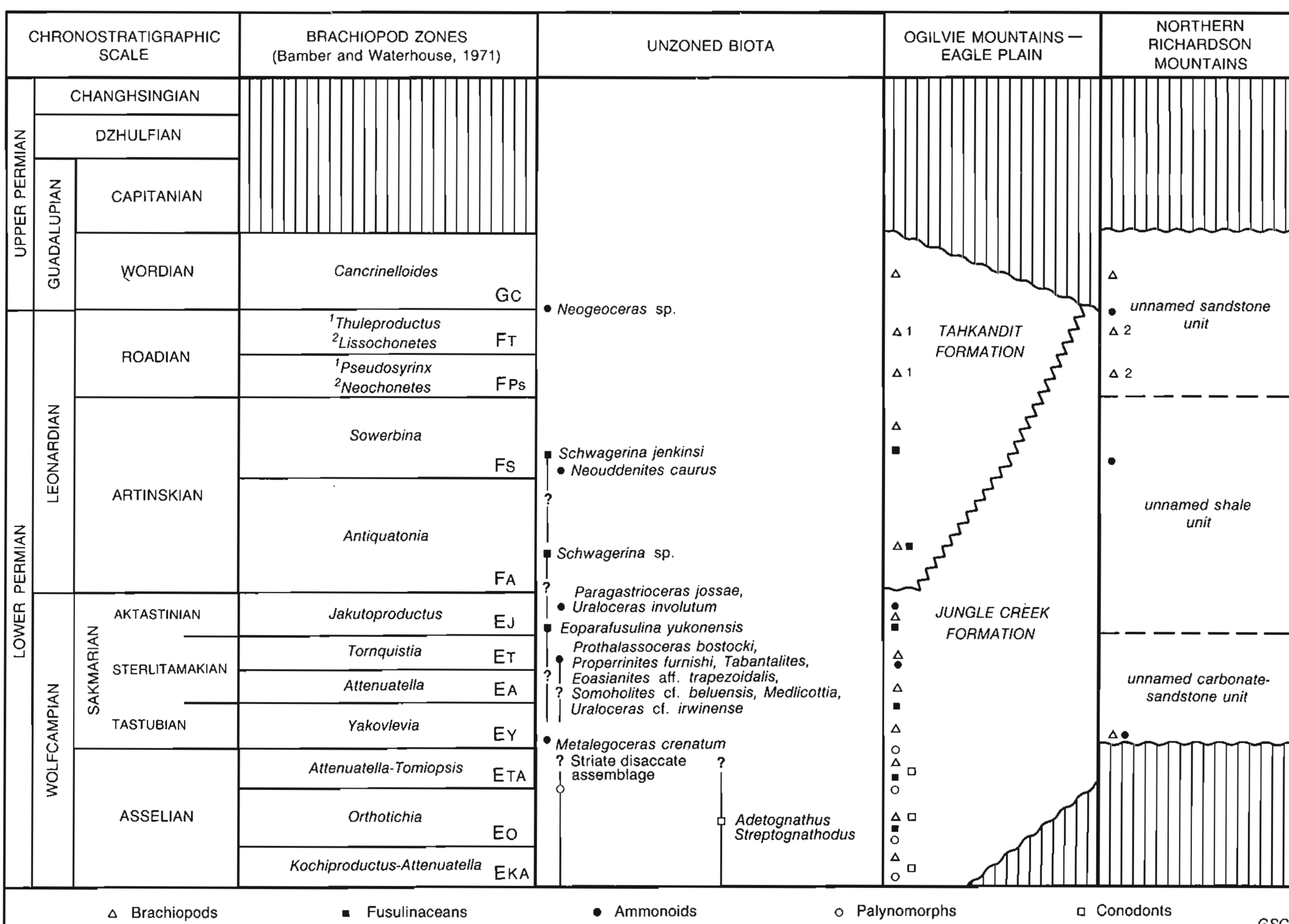

Figure 3. Permian correlations and biostratigraphy, Ogilvie Mountains, Eagle Plain, and northern Richardson Mountains. (After Henderson et al., in press.)

several levels in the Ettrain Formation of the Ogilvie Mountains (Ross, ibid.). These include the genera *Profusulinella, Pseudostaffella, Fusulinella,* and *Ozawainella,* and occur in association with and above microfossils of foraminifer Zones 21 and 22.

In the northern Richardson Mountains, Upper Carboniferous carbonates preserved as erosional remnants beneath the regional sub-Permian disconformity have yielded abundant Bashkirian and Moscovian foraminifers (Nassichuk and Bamber, 1978). The youngest Carboniferous microfaunas known from this area are upper Moscovian (Desmoinesian) fusulinaceans of the genera *Eowaeringella* and *Wedekindellina*, reported by Ross (1969, p. 129-134).

Four Permian fusulinacean assemblages were defined by Ross (1967; in Bamber and Waterhouse, 1971) from the Ogilvie Mountains. The Jungle Creek Formation contains the *Pseudofusulinella-Schubertella* and *Eoparafusulina yukonensis* assemblages, dated by Ross (ibid.) as Asselian and Early Sakmarian, respectively. Brachiopods occurring with the *E. yukonensis* assemblage belong to the *Jakutoproductus* Zone, dated as Late Sakmarian (Aktastinian) by Waterhouse and Waddington (1982). These are succeeded in the overlying Tahkandit Formation by the *Schwagerina* sp. B and *Schwagerina jenkinsi* assemblages, which have been assigned a Late Sakmarian to Artinskian age by Ross (ibid.). Preliminary results from a study by Mamet of recently collected, nonfusulinacean foraminifer faunas from the Ogilvie Mountains and southern Eagle Plain indicate an abundance of *Protonodosaria* in the lower part of the Jungle Creek Formation (B.L. Mamet, unpublished data). This confirms an Asselian age at both localities.

Conodonts

No published data are available on upper Paleozoic conodonts from the study area. Rich Carboniferous and Permian faunas, recently collected from the northern Ogilvie Mountains in the southwestern corner of the study area and from southeastern Eagle Plain (Fig. 1), are presently being studied by C.M. Henderson. Preliminary results of this study indicate that several biostratigraphically restricted assemblages can be recognized. These show a strong similarity to correlative, Bashkirian to Asselian assemblages found in the Sverdrup Basin of the Arctic Islands (Beauchamp et al., 1989). A notable exception is the presence in the Yukon of rachistognathid species, which do not occur in the Sverdrup Basin. In southeastern Eagle Plain (Peel River section), upper Viséan conodont assemblages including *Gnathodus texanus* Roundy, G. *pseudosemiglaber* Thompson and Fellows? (juvenile specimens), *"Bispathodus stabilis"* (Branson and Mehl), and *Rachistognathus prolixus* Baesemann and Lane, have been identified from the Hart River Formation. The overlying lower and middle Ettrain Formation yielded *Rachistognathus muricatus* (Dunn), *R. minutus* (Higgins and Bouckaert), and *Adetognathus spathus* (Dunn), indicating an early Morrowan to early Atokan (Bashkirian to Early Moscovian) age range. This Ettrain fauna and correlative conodont faunas of the middle Blackie Formation of the northern Ogilvie Mountains occur with brachiopods of the *Praehorridonia* and *Gemmulicosta* zones and with foraminifers younger than Zone 22. Both the brachiopods and foraminifers have been assigned a mid-Moscovian or younger age. Samples from uppermost Ettrain strata on the Peel River are barren of conodonts. In this section, the lower part of the Permian Jungle Creek Formation can be dated as Asselian to early Sakmarian from palynomorph, brachiopod, and foraminifer data. Conodont faunas from this interval include only Carboniferous forms that are similar to assemblages from the underlying Ettrain Formation and presumably were derived from Carboniferous strata that was eroded during Permian time.

In the northern Ogilvie Mountains the lowest samples collected from the lower Blackie Formation contain uppermost Viséan or Serpukhovian conodont assemblages, including *Gnathodus bilineatus* (Roundy), *Gnathodus girtyi collinsoni* Roades, Austin and Druce, *G. girtyi simplex* Dunn?, and *Paragnathodus commutata* (Branson and Mehl). In the middle Blackie Formation, an assemblage including *Idiognathodus sinuosis* Ellison and Graves, *Idiognathoides sulcatus sulcatus* Higgins and Bouckaert, *Streptognathodus suberectus* Dunn, and *Diplognathodus coloradoensis* Murray and Chronic correlates with the upper Morrowan or lowermost Atokan (Upper Bashkirian to Lower Moscovian). These are overlain by faunas containing *Idiognathoides* sp. cf. *I. sulcatus parva* Higgins and Bouckaert, which may be referred to *Idiognathoides marginodosus* Grayson (morphotype C of Grayson, 1984), of Atokan (Early Moscovian) age. Conodont samples from the upper Blackie are either barren or contain sparse, nondiagnostic faunas. Permian assemblages from the overlying Jungle Creek Formation include species of *Adetognathus* and *Streptognathodus*, indicative of an Asselian age. Younger faunas have not yet been recovered.

Ammonoids

Within the study area, Carboniferous ammonoids have been found almost exclusively in Viséan strata along the Peel River in southeastern Eagle Plain. Upper middle Viséan and lower upper Viséan occurrences were reported from the upper Ford Lake Formation by Bamber and Waterhouse (1971, p. 45). The overlying Hart River Formation contains the upper Viséan species *Goniatites crenistria* Phillips and *G. granosus* Portlock (Bamber and Waterhouse, 1971, p. 51; Mamet and Bamber, 1979, p. 49). The only other Carboniferous occurrence in the area was reported by Nassichuk (1971a), who described *Metapronorites stelcki* Nassichuk, of (?)Kasimovian age, from an unnamed shale succession in the southern Ogilvie Mountains.

Permian ammonoids of Sakmarian and Artinskian age are known from southern Eagle Plain, Porcupine River area, and northern Richardson Mountains. An assemblage from the Jungle Creek Formation on the Peel River in southeastern Eagle Plain, including *Prothalassoceras bostocki* Nassichuk, *Properrinites furnishi* Nassichuk, *Tabantalites bifurcatus* Ruzhencev, *Eoasianites* sp. aff. *E. trapezoidalis* Maximova, *Somoholites* sp. cf. *S. beluensis* (Haniel), *Medlicottia* sp. and *Uraloceras* sp. cf. *U. irwinense* Theichert and Glenister, was assigned an early Sakmarian age by Nassichuk (1971b). Brachiopods of the *Tornquistia* Zone from the same horizon indicate a Middle Sakmarian (Sterlitamakian) age (Waterhouse and Waddington, 1982). Nassichuk (ibid.) described a younger, Baigendzhinian (Artinskian) assemblage, which includes *Paragastrioceras jossae subtrapezoidale* Maximova and Tchernov, *P. jossae exile* Ruzhencev and *Uraloceras involutum* (Voinova), from the upper Jungle Creek or lower Tahkandit Formation farther west on the Peel River. A similar fauna of approximately the same age occurs on Porcupine River on the northern margin of Eagle Plain (Nassichuk, 1971c). A Wordian age was indicated by Nassichuk et al. (1965) for *Neogeoceras* sp., recovered from an unnamed sandstone unit in the northern Richardson Mountains. According to Bamber and Waterhouse (1971), this locality lies within or above the Roadian *Lissochonetes* brachiopod zone. Other ammonoid faunas from the northern Richardson Mountains (Nassichuk et al., 1965) include *Metalegoceras crenatum* Nassichuk, Furnish and Glenister, which occurs with Tastubian (Lower Sakmarian) brachiopods of the *Yakovlevia* Zone in an unnamed carbonate formation at the base of the Permian succession, and *Neouddenites caurus* Nassichuk, Furnish and Glenister from an overlying unnamed Artinskian shale unit.

Palynomorphs

Diverse Carboniferous and Permian palynomorph assemblages have been recovered from outcrop sections in the British Mountains, Barn Range, and Ogilvie Mountains, and from subsurface cores in the Eagle Plain. These assemblages are presently being studied by J. Utting (Lower Carboniferous) and J. Jerzykiewicz (Upper Carboniferous and Lower Permian).

In the British Mountains and Barn Range, spores from the Kayak Formation are thermally altered (Utting, 1989a) to the extent that detailed age determinations are difficult. Correlative assemblages of similar aspect obtained from coal measures approximately 15 km southeast of the Barn Range contain sufficiently well preserved palynomorphs to indicate an age within the range of early to middle Viséan (V1 to V2) based on correlations with the zonal scheme of Western Europe (Clayton et al., 1977). These assemblages include *Cingulizonates bialatus* (Waltz) Smith and Butterworth, *Cirratriradites elegans* (Waltz) Potonié and Kremp, 1956, *Densosporites duplicatus* (Naumova) Potonié and Kremp, 1956, *D. spitsbergensis* Playford, 1963, *Tripartites incisotrilobus* (Naumova) Potonié and Kremp, 1956, *Monilospora dignata* Playford, 1963, *Lophozonotriletes rarituberculatus* (Luber) Kedo, 1957 and *Lycospora pusilla* (Ibrahim) Schopf, Wilson and Bentall, 1944.

In the upper Paleozoic of the Eagle Plain and Ogilvie Mountains, five preliminary assemblage zones (introduced below) range in age from late Viséan to Early Permian. Ages assigned to the Upper Carboniferous and Lower Permian zones are those determined from brachiopod zones (Fig. 2).

1. *Murospora aurita - Rotaspora fracta* Assemblage Zone
The zone contains the following stratigraphically important species: *Cingulizonates bialatus* (Waltz) Smith and Butterworth, 1967, *Colatisporites decorus* (Bharadwaj and Venkatachala) Williams, 1979, *Acanthotriletes multisetus* (Luber) Potonié and Kremp, 1955, *Densosporites aculeatus* Playford, 1963, *D. anulatus* (Loose) Smith and Butterworth, 1967, *D. rarispinosus* Playford, 1963, *D. spitsbergensis* Playford, 1963, *Diatomozonotriletes rarus* Playford, 1963, *D. saetosus* (Haquebard and Barss) Hughes and Playford, 1961, *Lycospora pusilla* (Ibrahim) Schopf, Wilson and Bentall, 1944, *Monilospora triungensis* Playford, 1963, *Murospora aurita* (Waltz) Playford, 1962, *M. dupla* (Isachenko) Playford, 1962, *Reticulatisporites peltatus* Playford, 1962, *Rotaspora* sp., *Spelaeotriletes arenaceus* Neves and Owens, 1966, *Tetraporina horologia* (Staplin) Playford, 1962, *Tripartites incisotrilobus* (Naumova) Potonié and Kremp, 1956, and *Waltzispora albertensis* Staplin, 1960. A late Viséan (V3) age was determined for this zone, which occurs in the upper part of the Ford Lake Formation and throughout all but the uppermost part of the Hart River Formation in southern Eagle Plain. The assemblage in this zone resembles those from the upper Viséan Emma Fiord Formation of the Sverdrup Basin (Utting et al., 1987; in press) and from correlative strata in the Mattson Formation and the Stoddart Group of the central Canadian Cordillera and adjacent Plains. Abundant spinose acritarchs occur in the uppermost part of the Ford Lake Formation in southeastern Eagle Plain. The lower and middle part of the Hart River Formation contains abundant amorphous organic matter and few palynomorphs.

2. *Florinites* sp. Assemblage Zone
Many of the species in the *Murospora aurita — Rotaspora fracta* Zone persist into this zone, but in addition *Florinites* sp. appears, indicating an early Serpukhovian age. The zone occurs in the uppermost part of the Hart River Formation in southern Eagle Plain and is also found in the uppermost Mattson Formation of the central Canadian Cordillera and Plains.

3. *Potonieisporites* Assemblage Zone
This zone contains three subzones, all of which are characterized by abundant monosaccate pollen.

a) *Potonieisporites elegans* Assemblage Subzone. In this subzone there is a dominance of trilete spores, especially *Lycospora* (*L. noctuina* Butterworth and Williams, 1958, *L. granulata* Kosanke, 1950), *Calamospora* (*C. pedata* Kosanke, 1950, *C. obscura* Peppers, 1964, *C. parva* Guennel, 1958), *Densosporites, Cingulizonates, Cirratriradites* and *Convolutispora*. Monolete spores (*Laevigatosporites vulgaris* Ibrahim, 1933, *L. ovalis* Kosanke, 1950) are rare, and monosaccate pollen are abundant. The latter include *Florinites* (*F. antiquus* Schopf, Wilson, and Bentall, 1944, *F. millotti* Butterworth and Williams, 1954, *F. visendus* (Ibrahim) Schopf, Wilson, and Bentall, 1944), *Potonieisporites elegans* (Wilson and Kosanke) Wilson and Venkatachala emend. Habib, 1968, *Costatascyclus, Wilsonites* and *Cannanoropollis*. Rare disaccate pollen grains of the following genera occur: *Plicatipollenites, Vesicaspora, Illinites*, cf. *Pseudoillinites, Pityosporites* and *Limitisporites*. This subzone, which is of Bashkirian age, occurs in the lower part of the Blackie Formation in the northern Ogilvie Mountains. Similar assemblages occur in the upper part of the upper Bashkirian to lower Moscovian Otto Fiord

Formation in the Sverdrup Basin (Utting, 1985), although the latter also contains striate, disaccate pollen.

b) *Potonieisporites magnus* Assemblage Subzone. The lower part of this subzone is dominated by *Potonieisporites* (*P. elegans*) (Wilson and Kosanke) Wilson and Venkatachala emend. Habib, 1968, *P. magnus* Lele and Karim, 1971), *Florinites* (*F.pumicosus* (Ibrahim) Schopf, Wilson, and Bentall, 1944, *F. visendus* (Ibrahim) Schopf, Wilson, and Bentall, 1964, *F. eremus* Balme and Hennelly, 1955, *F. mediapudens* (Loose) Potonié and Kremp, 1956), *Cordaitina, Costatascyclus* and *Cannanoropollis*. Disaccate pollen (nonstriate and striate) are rare. This subzone contains the oldest occurrence of the monolete spores *Thymospora* and *Torispora*. Trilete spores (e.g., *Lycospora, Calamospora, Punctatisporites, Granulatisporites*) are common at some levels. In the upper part of the subzone *Vittatina* is rare and occurs sporadically.

This subzone is of late Bashkirian to Moscovian age. It occurs low in the unnamed equivalents of the Ettrain Formation in southern Eagle Plain, and in the lower Blackie Formation of the Ogilvie Mountains. Similar assemblages are found in Upper Moscovian to Kasimovian strata within the lower part of the Canyon Fiord Formation of the Sverdrup Basin, although the latter has yielded no specimens of *Vittatina* (Utting, this volume, b).

c) *Potonieisporites novicus* Assemblage Subzone. As in the previous subzone, monosaccate pollen grains continue to be abundant, but the most conspicuous species is *Potonieisporites novicus* Bhardwaj, 1954. Disaccate pollen are also relatively abundant (e.g., *Pityosporites, Illinites, Limitisporites, Protohaploxypinus*). *Vittatina* is rare, but occurs more frequently in the uppermost part of the subzone. Trilete and monolete spores are rare. The zone, which is Moscovian in age, occurs in the middle and upper parts of the unnamed equivalents of the Ettrain Formation (southern Eagle Plain) and the middle and upper parts of the Blackie Formation (northern Ogilvie Mountains).

4. *Vittatina* sp. Assemblage Zone

This zone is characterized by abundance of the *Potonieisporites-Cordaitina-Guthoerlisporites-Nuskoisporites* complex. *Vittatina* is common and disaccate, nonstriate and striate pollen are abundant. Trilete and monolete spores are rare. The zone is late Moscovian to Gzhelian in age and occurs in the uppermost part of the unnamed Ettrain equivalents (southern Eagle Plain), in the upper Blackie Formation (northern Ogilvie Mountains), and in the lower part of the overlying Jungle Creek Formation (southern Ogilvie Mountains). The assemblage in this zone resembles that in the Gzhelian to Asselian *Potonieisporites* sp. — *Vittatina* sp. Assemblage Zone, found in the middle to upper parts of the Canyon Fiord Formation of the Sverdrup Basin (Utting, this volume, b). The assemblages in the Canyon Fiord, however, contain species of *Vittatina* not yet seen in the Yukon, and are probably slightly younger.

5. Striate Disaccate Assemblage Zone

This zone is characterized by poorly diversified assemblages, dominated by the genus *Vittatina* and disaccate striate pollen (e.g., *Protohaploxypinus latissimus* (Luber) Samoilovich, 1953, *P. longelinus* Naumova, 1964, *Striatoabieites multistriatus* (Balme and Hennelly) Hart, 1965, *Striatopodocarpites* and cf. *Hamiapollenites*). Nonstriate disaccate pollen are present. Monosaccate pollen are represented by *Potonieisporites, Cordaitina, Nuskoisporites* and *Florinites*. Trilete and monolete spores decrease upward, and monocolpate pollen occur sporadically. The zone is of Early Permian (Asselian to Sakmarian?) age and occurs in the lower Jungle Creek Formation. It contains palynomorphs similar to those in the *Weylandites striatus - Protohaploxypinus perfectus* Assemblage Zone of the Sverdrup Basin, which is of late Asselian to Sakmarian age (Utting, 1989b).

Corals

Lower Carboniferous (upper Viséan) coral zones IV and VA of Sando and Bamber (1985) have recently been recognized by Bamber (work in progress) in the upper Kayak and lower Alapah formations of the British Mountains. Preliminary identifications indicate the presence of the genera *Faberophyllum, Stelechophyllum*, and *Sciophyllum* from Zone IV and *Schoenophyllum* from Zone VA. Upper Carboniferous and Permian corals are rare in the study area. Those present include durhaminids and bothrophyllids with close affinities to contemporaneous North American-Uralian, Boreal coral faunas from the Canadian Arctic Archipelago and Alaska.

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Fossils: thermal maturation indicators, northwestern mainland Canada

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Progressive and irreversible changes in organic material caused by rising temperatures provide a means of detecting the degree of alteration in sedimentary rocks after diagenesis and low grade metamorphism (e.g., Hunt, 1979; Staplin et al., 1982; Robert, 1985; Tissot et al., 1987). In addition to widely used geochemical and optical techniques such as vitrinite reflectance and Rock-Eval analysis are those directly related to paleontological studies. Some techniques, such as determination of the temperature- dependent variation in colour of conodonts (e.g., Harris, 1979; Nowlan and Barnes, 1987) or palynomorphs (e.g., Utting, 1987; Utting et al., in press) are inexpensive, easily acquired by-products of taxonomic and biostratigraphic research. Others, such as the optical characteristics of graptolites (e.g., Goodarzi and Norford, 1985, 1987), or scolecodonts (Goodarzi and Higgins, 1987; Bertrand and Heroux, 1987) require special preparation and study. In recent years, the Geological Survey of Canada has initiated multidisciplinary programs to compile and synthesize data from all these types of investigations. The following three reports in this issue describe preliminary results on Devonian, Carboniferous, and Jurassic fossils from northern Yukon and adjacent Northwest Territories.

There are now considerable data from vitrinite reflectance, Rock-Eval and other organic geochemical techniques, which yield consistent, reproducable results. The comparison of results from different methods applied to the same samples, or samples from the same localities, has resulted in the publication of "standard" calibration scales for other methods, which have been compared mainly with vitrinite reflectance data. In addition to providing regional data relevant to oil and gas generation, and the occurrence of economic minerals, current studies within the Geological Survey of Canada are providing further insight into the relative maturation rates of different types of fossil material. The goal is to compare and refine the relative calibration scales for different techniques. Many fossils are useful as indicators of age, and being able to determine the age of individual specimens that are also thermal maturation indicators has a powerful advantage over techniques that rely solely on analysis by machine, or that statistically generalize data derived from the total organic content of a sample. For example, the Thermal Alteration Indices (TAI) and Optical Reflectance characteristics identified from palynological preparations of Jurassic rocks in northern Yukon yield results that, while generally consistent, contain some anomalies (Davies and Poulton, 1989). Some of these anomalies can be explained by the difficulty in recognizing reworked material in polished preparations of dispersed "vitrinite". The palynologist's ability to interpret both age and thermal characteristics of each palynomorph in a preparation may help to resolve the problem of reworked material.

Thermal maturation studies contribute to regional geological and tectonic interpretations. Questions arise from studies of regional trends as well as local anomalies in both thermal maturity map patterns and stratigraphic successions. Among the questions are those relating to the significance of unconformities in the histories of basins; thickness of overburden, which is in turn related to burial/subsidence characteristics of a basin or orogenic terrane; and variations in heat flow characteristics, which relate to rate of burial, fluid flow, local intrusive features, etc.

In summary, paleontological techniques include powerful tools for the interpretation of ancient thermal regimes. The studies being currently conducted in the northwestern mainland are directed to:

1. Oil and gas exploration
2. Mineral exploration
3. Comparison of the results of different techniques, interpretation of the differences, and refinement of relative calibration scales, and
4. Contributions to studies of regional geology, basin analysis and tectonic interpretations.

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Current status of Jurassic biostratigraphy and stratigraphy, northern Yukon and adjacent Mackenzie Delta[†]

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Poulton, T.P. Current status of Jurassic biostratigraphy and stratigraphy, northern Yukon and adjacent Mackenzie Delta; in Current Research, Part G, Geological Survey of Canada, Paper 89-1G, p. 25-30, 1989.

Abstract

There have been great advances recently in our knowledge of the Jurassic stratigraphy of northern Yukon and adjacent Northwest Territories. The Jurassic strata comprise an interbedded series of sandstone and shale formations in the east, included in the Bug Creek Group, and the Husky, Porcupine River and North Branch formations. These formations pass westward and northwestward into a thicker, and more complete shelf facies of shale-siltstone, the Kingak Formation. Facies and thickness trends indicate a series of point sources for the sediments, from which the sands were dispersed by marine shelf mechanisms. Throughout the Jurassic, there was a southward migration of the sediment source associated with the progressive transgression of the craton.

Detailed study of ammonites and other guide fossils has provided the means of resolving the stratigraphic relationships of these strata. The ammonites and bivalves are a mixture of species with affinities closest to those of Europe and the northern Atlantic in some cases, and northeastern Siberia and the northeast Pacific in others. Microfossils and palynomorphs have also been used for zonation.

Résumé

Les couches du Jurassique dans le nord du Yukon et dans les parties adjacentes des Territoires du Nord-Ouest renferment des séries interstratifiées de formations de grès et de schiste argileux d'origine marine à l'est: il s'agit du groupe de Bug Creek et des formations de Husky, de Porcupine River et de North Branch. Vers l'ouest, elles deviennent une séquence plus épaisse et complète composée de schiste argileux et microgrès, soit la formation de Kingak. Les tendances observées au niveau du faciès et de l'épaisseur indiquent un décalage vers le sud des sources de sédiments et de la dispersion sur une plateforme à l'angle nord-ouest du continent jurassique.

L'étude détaillée des ammonites et des autres fossiles guides a permis d'éclairer les relations stratigraphiques entre ces couches. Les ammonites et les bivales constituent un mélange d'espèces dans certains cas plus apparentées à des espèces d'Europe et de l'Atlantique nord et dans d'autres, à des espèces du nord-est de la Sibérie et du nord-est du Pacifique. Des microfossiles et des palymorphes ont servi à établir la zonation.

[†] Contribution to Frontier Geoscience Program

INTRODUCTION

Since the last biostratigraphic summary and bibliography of Jurassic ammonites from northern Yukon (Poulton, 1978), there have been major advances in our knowledge and understanding of the macropaleontology, micropaleontology and palynology of the Jurassic of northwestern mainland Canada. Also during this time, the lithostratigraphy and its tectonic and paleogeographic implications have become better understood (Fig. 1). The most recent advances have coincided with the emplacement of the Frontier Geoscience Program of the Geological Survey of Canada. Much of the new information has been published in references cited in the following summary; some major reports are still not generally available, however (Poulton, in press a, b, c), and it is an objective of this report to present a preview of the most significant results from those reports. This report is a brief geological overview and provides a summary of current literature, which might be particularly useful to a newcomer to the area.

PALEONTOLOGY AND BIOSTRATIGRAPHY

Ammonites are the primary means of dating and correlating marine epicontinental Mesozoic rocks. With an average time span of about 720 000 years (Callomon, 1984), the Jurassic ammonite zones are unrivalled chronological tools. They occur with sufficient frequency in the Lower, Middle and lower Upper Jurassic outcrops of northern Yukon and adjacent Northwest Territories to place the stratigraphic sequences and lateral relationships on firm ground. Many of them have been described and their biostratigraphic usefulness evaluated (Frebold, 1975; Poulton, 1987, in press, b). In general, they can be easily correlated with similar species in the standard Northwest European succession, and they provide new insight into the paleogeographic distribution and faunal provincialism of the taxa represented. For example, the lower Sinemurian ammonite *Arnioceras*, and many other Lower and Middle Jurassic genera, are documented for the first time from northern areas. Occurring at the northwest corner of the Jurassic continent, the ammonite faunas are of intermixed southern (Pacific and Cosmopolitan) and northern (Boreal) origin, permitting the long-standing correlation problems of the major faunal realms to be further understood and partially resolved.

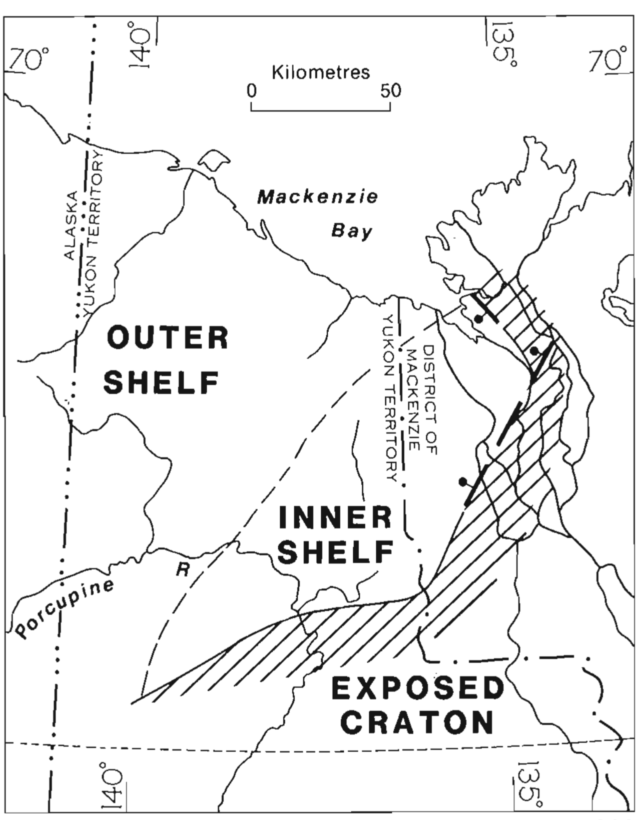


Figure 1. Major facies belts of study area.

The shallow marine faunas are rich in bivalves, and some beds contain significant assemblages of belemnites, crinoids, brachiopods, and other invertebrates. Those of the Middle Jurassic have not yet been studied in detail, but can be readily used as guides to the Middle Jurassic and some, such as species of the genus *Retroceramus*, can be correlated in more detail with described species from Siberia. In contrast, the Lower Jurassic and Aalenian bivalves and other invertebrates have been described and illustrated in detail, and a sequence of distinctive bivalve faunas closely associated stratigraphically with the ammonites has been recognized (Poulton, in press, b). Many of these fossils are identical at the species level to those of the same ages in Siberia and Europe, so that a common zonation for the northern hemisphere, with precision to about the stage level, is a reasonable expectation.

Ammonites are rare in the upper Oxfordian and younger Upper Jurassic and the principal guide fossils are an evolutionary sequence of species of the bivalve genus *Buchia*, also well established as zonal indices throughout the northern part of the globe (Jeletzky, 1984, and references there; Poulton, 1978, in press a, c).

Microfossils have been studied from most of the Jurassic formations of the Richardson Mountains and adjacent Mackenzie Delta, and distinctive species have been described (Leskiw, in Poulton et al., 1982; Hedinger, in press). The same is true for palynomorphs (Audretsch, in Poulton et al., 1982; Fensome, 1987). Figure 2 is the first compilation of ammonite as well as palynomorph zones. The refining and extending of these locally, regionally, and internationally significant interdisciplinary syntheses is continuing.

LITHOSTRATIGRAPHY, TECTONICS, AND PALEOGEOGRAPHY

Major recent advances include the description and distribution of the Jurassic formations (Poulton et al., 1982; Dixon, 1982 a, b; Braman, 1985). Because of the relative monotony of the interbedded sandstone and shale, the sequence can only be resolved by documenting the succession of time diagnostic ammonites and the bivalve *Buchia*.

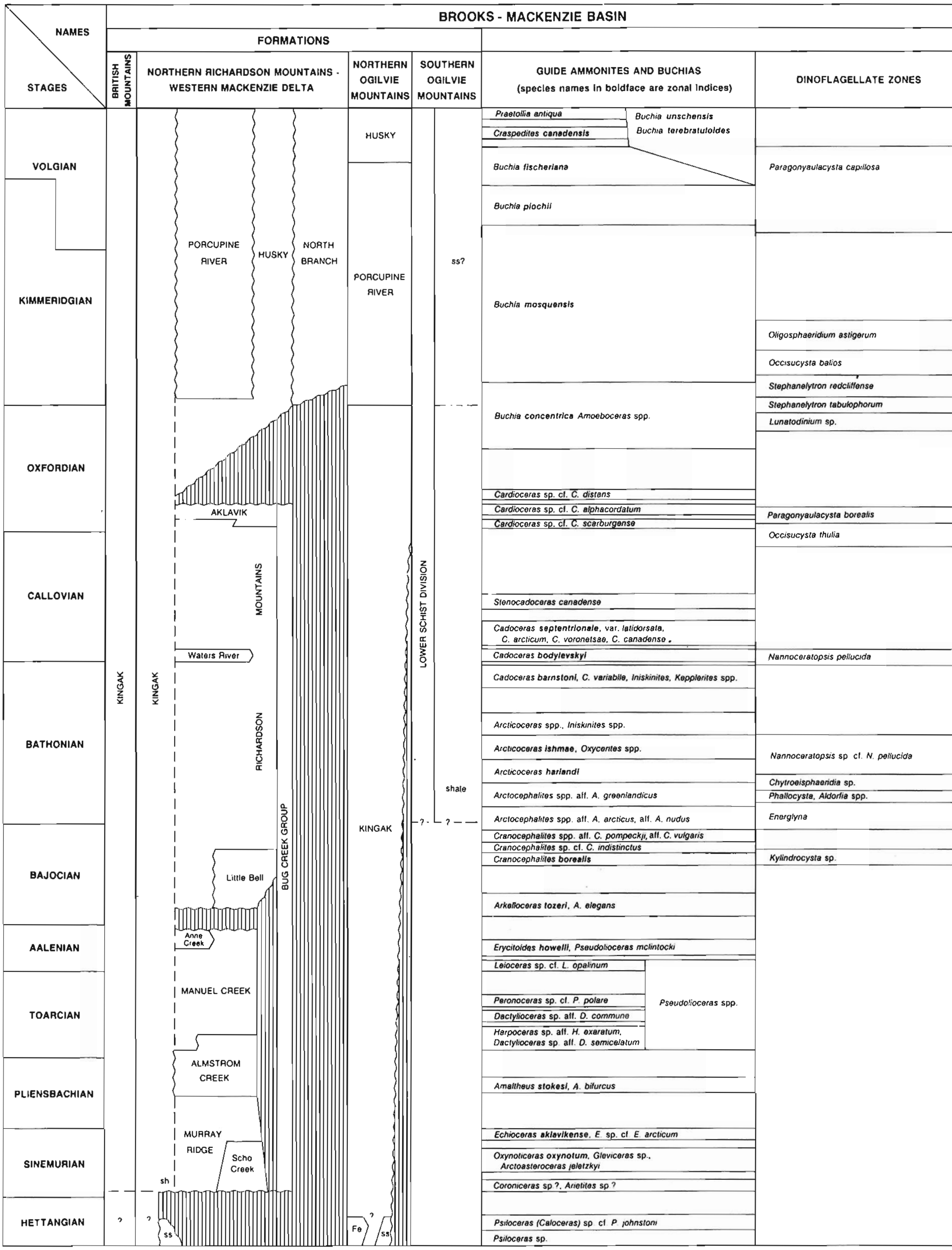
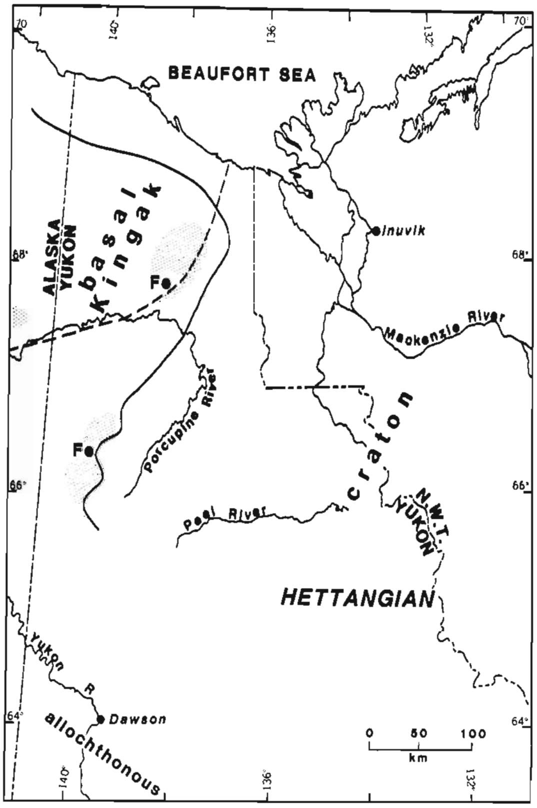


Figure 2. Jurassic ammonite and dinoflagellate zonation, the latter compiled by E.H. Davies (from Poulton, in press, a). The wavy vertical lines indicate facies changes and the dashed lines indicate arbitrary nomenclatural changes.

Figures 3-7. Facies distribution maps for selected intervals of the Jurassic. Note the progressive southward shift of the sediment source and of the sandstone belt (from Poulton, in press, c). Dashed line in all diagrams indicates position of so-called "Kaltag Fault" for reference. "Fe" in Figure 3 indicates ironstone deposit.



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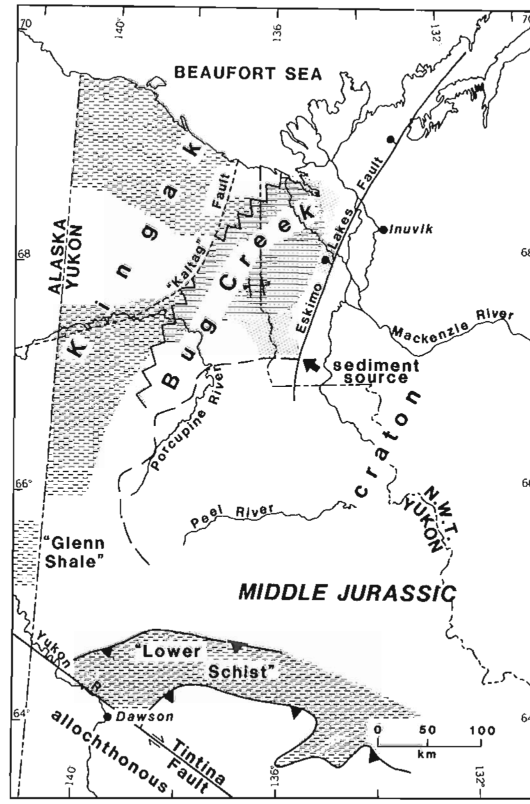

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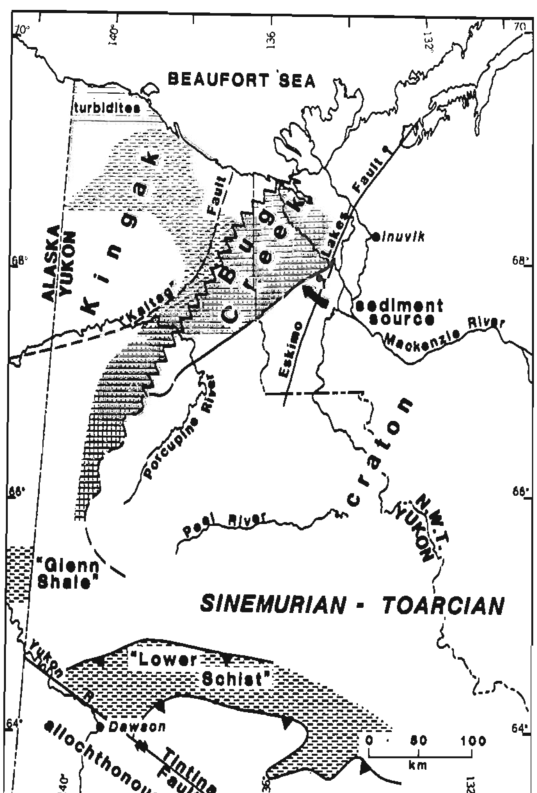

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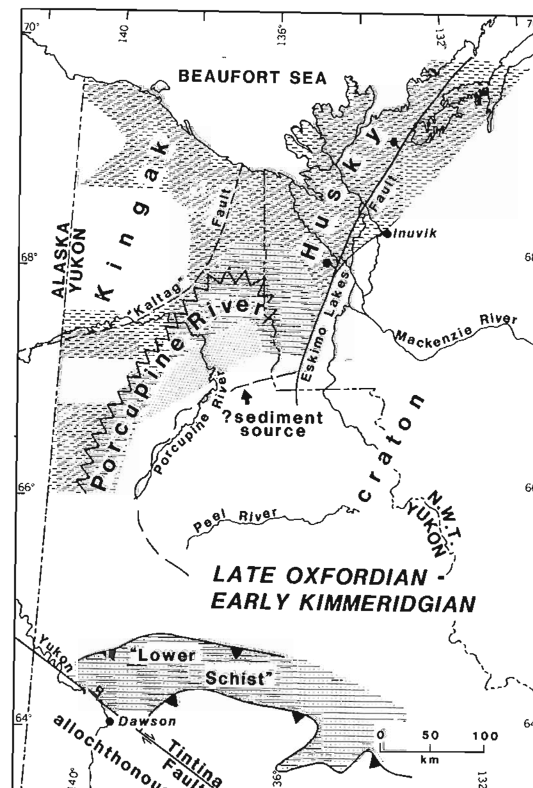

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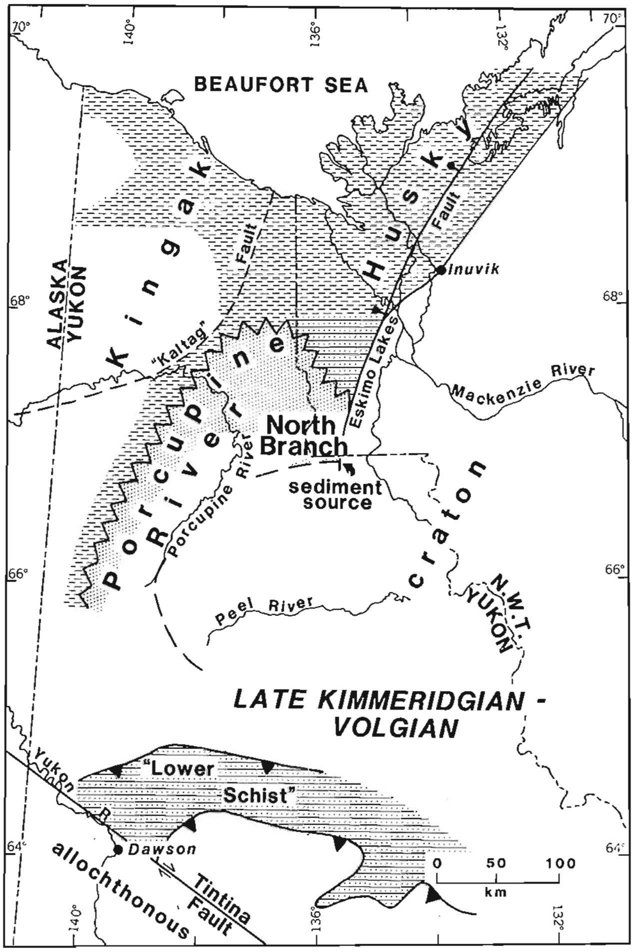

Figure 7.

The Jurassic formations are a series of alternating sandstone and shale in the eastern part of the area, i.e. in the subsurface of western onland parts of Mackenzie Delta and in the northern Richardson Mountains. The sequence becomes thicker, more complete, and generally more finely grained toward the northwest, with a change to shale-siltstone facies in the Kingak Formation near the western edge of the northern Richardson Mountains. The details and distribution of the facies suggest marine reworking of sands that were introduced from a point source that migrated southward during the Jurassic, in association with progressive but intermittent transgression of the craton (Figs. 3-7). These general trends were complicated by minor syn-Jurassic intracratonic uplifts, including an enigmatic series of uplifts along the present north coast of the area, suggesting that the continental margin was somewhat unstable. The sequence of formations has been interpreted in terms of sea level fluctuations and cycles (Fig. 8) and comparison of the cycles with those of other basins suggests there were four major events of interregional significance (Poulton, 1988). These events are well controlled biostratigraphically, are approximately early Hettangian, late Pliensbachian, earliest Bajocian and early Oxfordian in age, and are primarily major regressive events. The basal Jurassic unit (basal Kingak Formation), of Hettangian age, is a scattered, heterogeneous assemblage characterized by ironstones and *Cardinia*-bearing sandstones (Fig. 3). The next major transgressive unit, of Bajocian age, is also characterized by ironstones with abundant phosphatic nodules — a typical, starved shelf transgressive facies. Paleontological discoveries in the Old Crow Basin area, together with clarification of the stratigraphic relationships in the northern Richardson Mountains (Poulton, 1982), confirm that no significant

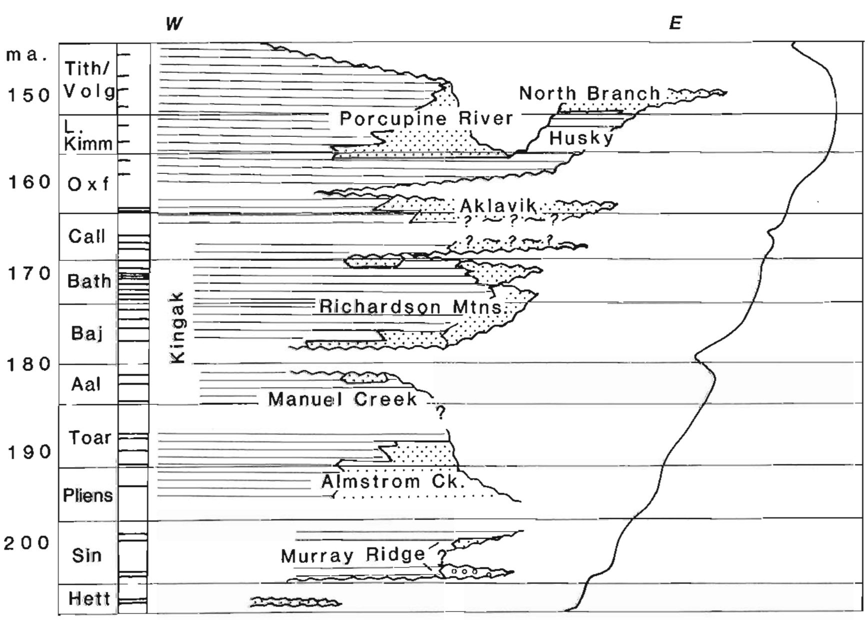

Figure 8. Time-stratigraphic diagram of Jurassic formations of northern Yukon and adjacent Northwest Territories, and subsidence versus sedimentation curve (from Poulton, 1988). High rate of slope of the curve to the right indicates a high rate of subsidence compared with rate of deposition; inflections to the left are major regional regressive events. Bars in the column next to the stage names at the left indicate diagnostic ammonite occurrences; shorter bars in the upper part of the same column indicate diagnostic Buchia occurrences. The Bug Creek Group comprises the Murray Ridge through Aklavik formations, shown in Figures 3-6.

source highland existed to the west in Jurassic time, as had been previously suggested. Indeed the mixed Boreal/Pacific character of the Middle Jurassic ammonite fauna along Porcupine River (Poulton, 1987), together with considerations of the facies relationships, indicate that during the Jurassic, the northern Yukon area was the site of a broad, shallow shelf facing the contemporary ocean that bordered the northwestern corner of the continent, and that most parts of Alaska were not yet in their present relative positions.

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Thermal maturation data from the Jurassic rocks of northern Yukon and adjacent Northwest Territories[†]

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Davies, E.H. and Poulton, T.P., Thermal maturation data from the Jurassic rocks of northern Yukon and adjacent Northwest Territories; in Current Research, Part G, Geological Survey of Canada, Paper 89-1G, p. 31-36, 1989.

Abstract

Thermal Alteration Index (spore colouration) and vitrinite reflectance measurements are reported for Jurassic argillaceous formations from 22 localities in northern Yukon and adjacent N.W.T. They indicate a broad range of thermal alteration levels. Within the northern Richardson Mountains, these levels range from marginally mature in the northeast to highly thermally altered in the southwest. The orientation of this distribution pattern coincides with that of local structural trends. Samples from farther west are generally strongly altered, but do not yield consistent trends.

Résumé

Un indice d'altération thermique (coloration des spores) et des mesures de réflectance de la vitrinite sont présentés pour des formations argileuses du Jurassique de 22 régions du nord du Yukon et des Territoires du Nord-Ouest. Il en ressort toute une gamme de niveaux d'altération thermique. Dans le nord des monts Richardson, cette gamme va d'une altération parvenue à un stade de maturité marginal dans le nord-est à un degré d'altération thermique prononcé dans le sud-ouest. Cette répartition est orientée suivant les directions structurales locales. Les échantillons prélevés plus à l'ouest sont en général fortement altérés, mais ne présentent pas des tendances uniformes.

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[1] Bujak Davies Group, Calgary
[†] Contribution to Frontier Geoscience Program.

INTRODUCTION

Jurassic sedimentary rocks are widespread in northern Yukon and adjacent Northwest Territories, where they are found in different geological settings. Until now, their thermal alteration levels have been relatively unknown, and the only published reflectance data for the area are for Paleozoic rocks (Cameron et al., 1986).

Latour (1970) illustrated the regional distribution of lignitic and bituminous coals but did not specify their ages. The samples reported here were collected mainly in the northern Richardson Mountains by Poulton in 1975, 1976 and 1978. They are all from shale or siltstone outcrops, and with scattered samples from other areas to the west, they provide the best possible regional coverage to date. Within the northern Richardson Mountains (Fig. 1), stratigraphic sections of the Bug Creek Group contain shale/siltstone intervals, which had been sampled and analyzed for micropaleontological and palynological purposes earlier (Poulton et al., 1982). For this report, several samples were selected from each of these sections, to provide checks on each locality and to evaluate variation between different units within a section. Localities west of the northern Richardson Mountains are not well suited for the measurement of stratigraphic sections because of tectonic complications within a dominantly shale package, the Kingak Formation. From these western localities, single samples were analyzed in many cases.

All samples were measured for both vitrinite reflectance and spore colouration (Thermal Alteration Index), in order to provide comparative data, and to provide control for accuracy of measurements. It was thus largely possible to avoid conclusions based on reworked organic matter.

It has been stated previously (e.g., Poulton et al., 1982) that carbonization becomes a strong negative factor for palynological studies west of the northern Richardson Mountains and in their western parts. It has not been possible previously to obtain a reasonable recovery for even the most preliminary of studies in these western parts. The present study suggests that only in the southwest of the area, in northern Ogilvie Mountains, are TAI values so consistently high that palynological samples would not be useful.

LABORATORY PROCEDURES

The samples were crushed and macerated with cold hydrochloric acid and hydrofluoric acid. Kerogen was subsequently removed from undigested inorganic material through a heavy liquid separation utilizing zinc bromide solution with a specific gravity of 2.0. The kerogen residue was divided in two, one part for the preparation of slide mounts for transmitted light microscopy, and the other for the preparation of stub mounts for reflected microscopy.

The procedures for the mounting and polishing of the vitrinite samples follows the methodology described by Davies and Avery (1984), using a triple mount stub and standard 600 grit, 0.3 micron and 0.05 micron polishing procedures, in accordance with the procedures approved by the American Society for Testing and Materials (1979). This methodology results in a polished mount of unoxidized kerogen.

Fifty vitrinite reflectance measurements for each sample were taken wherever possible. Both reworked and in situ vitrinite particles were measured, but only the in situ particles were used to calculate the mean maximum reflectance (Romax) and standard deviation values reported (Tables 1-4).

COMPARISON OF OPTICAL REFLECTANCE AND TAI VALUES

The colouration of spores and pollen as well as other plant debris present in the kerogen residues changes, primarily with increased heating and with the passage of time. This phenomenon has been utilized by the petroleum industry over the past 20 years as an indicator of the thermal maturity (Staplin, 1969). Numerous Thermal Alteration Index (TAI) scales have been introduced during this time. The one employed in this study is the scale published by Staplin (1969). The manner in which the types of kerogen react to the various levels of maturation has been reviewed by Burgess (1974), Bayliss (1975, 1977) and Bujak et al. (1977a, b, 1978). Figure 2 illustrates the approximate maximum and minimum Ro correlatives for TAI values (after Staplin, 1969).

Figure 3 illustrates the actual correlation of vitrinite reflectance values with Thermal Alteration Indices, as determined by this study. In general, there is a close match between this correlation and that predicted in the generalized chart (Fig. 2). Two variations require comment. In the actual correlation chart (Fig. 3) two samples have unusually high Ro values compared to TAI. This probably suggests that the vitrinite particles, which were machine-read for these samples, were predominantly reworked. Secondly, TAI readings between 2 and 3- to 3 tend to have slightly low Ro values. This may be due to reworked or slightly oxidized miospores. The suggestions regarding reworking and oxidation of these samples are based on observations of the palynomorph preparations, and are extrapolated to the vitrinite stub mounts.

GEOLOGICAL INTERPRETATION

The TAI values are plotted on the map (Fig. 1) and contoured for the eastern part of the northern Richardson Mountains. The Ro values for each sample are shown in Tables 1 to 4 and would produce similar map patterns.

The samples come from shaly horizons that represent all levels within the Jurassic column. In particular, wherever possible, samples were chosen from several levels at each locality. Within the sections, which reach 800 to 1000 m in thickness in western areas, no significant relationship between thermal alteration or vitrinite reflectance values and position within the sections was detected. Therefore, thicknesses of this order alone are not significant causes of variation in thermal alteration, and it is not considered necessary to plot separately nor to discuss in detail the relative ages and stratigraphic position of each sample.

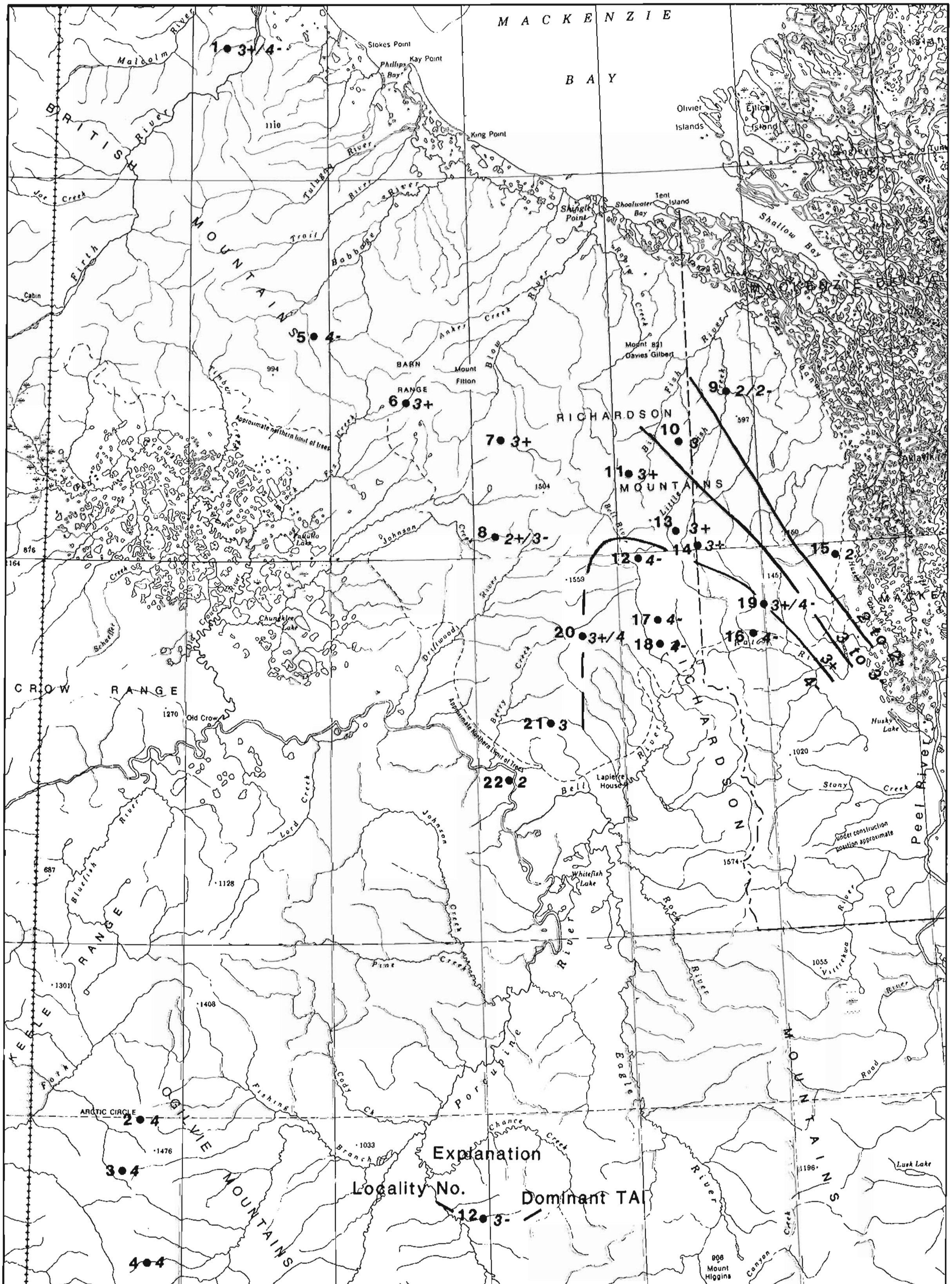

Figure 1. TAI values for Jurassic samples from localities in northern Yukon and adjacent Northwest Territories. The data are simplified, to show only the dominant value for each locality. Supporting data, and correlative Ro values for each sample are shown in Tables 1-4.

Firth River (Table 1)

Two samples from Lower Jurassic beds of the Kingak shale have Ro values between 2.8 and 3 and TAI values of 3+ to 4- and 4- respectively. Little can be inferred from these intermediate to high values because of the absence in the area of other data with which to compare them. Cameron et al. (1986) also reported reflectance values from 3.5 to 4.0 for the anthracites of the Carboniferous Kayak Formation of this area and suggested that their high values were due either to deeper burial or the thermal effects of intrusions. They suggested that the bireflectance data for one of their samples (Station 5) were the result of more rapid thermal effects than those that could be achieved by burial alone, and they assumed that local rather than regional factors were the cause of the maturity levels.

Northern Ogilvie Mountains (Table 2)

Seven samples from three localities have consistently high values of Ro (from 5 to greater than 6) and TAI (4). These values, the highest in the report area conform with the southwestern location of the northern Ogilvie Mountains within the area, closest to the Tintina Trench and the tectonically and plutonically active inner belts of the Cordillera. Whether this area also was characterized by an overburden load greater than those areas farther north and east is not known, and cannot be determined by the present knowledge of Mesozoic and Tertiary stratigraphy of northern Yukon.

Table 1. Ro and TAI values, Jurassic samples from northwestern Yukon (Loney Creek- Firth River area).

Ro and TAI Analysis Northwestern Yukon

	SAMPLE #	%Ro	TAI
1	C-53469	2.838+/-0.273	4-
	C-53479	2.949+/-0.410	3+ to 4-

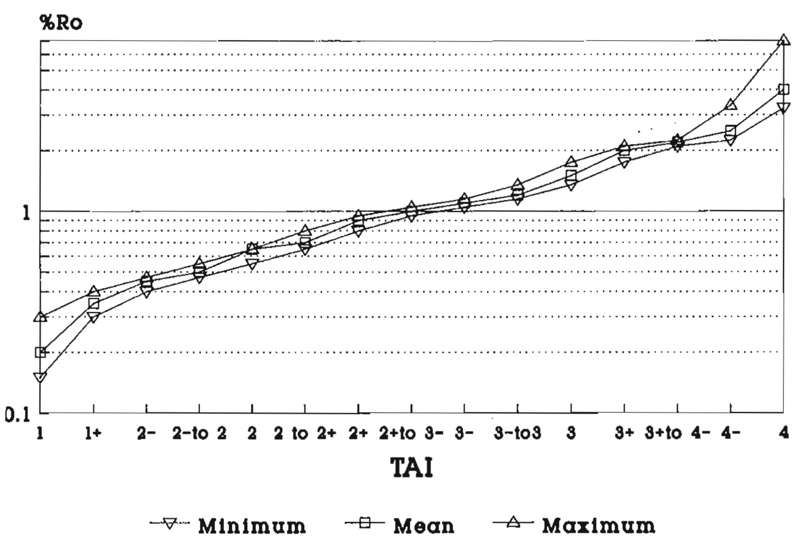

Figure 2. Idealized correlation of Ro and TAI values, based on Staplin (1979).

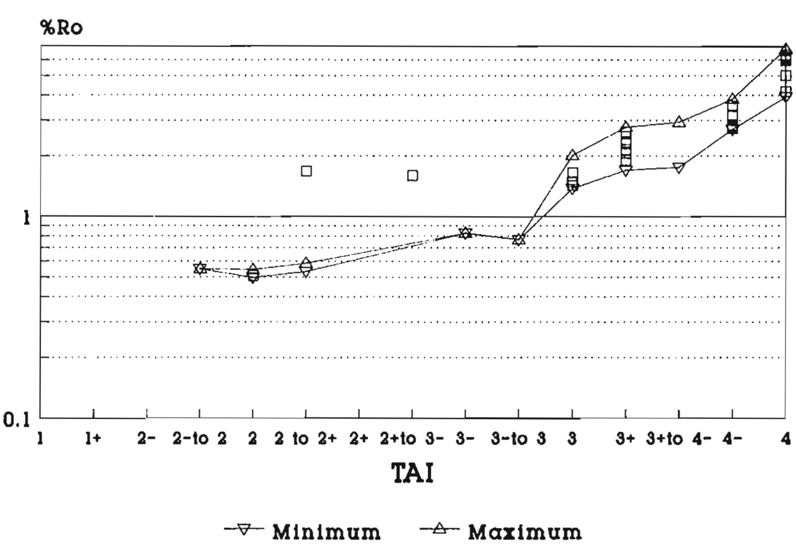

Figure 3. Correlation of Ro and TAI values, from Jurassic samples within northern Yukon and adjacent Northwest Territories.

Northern Richardson Mountains - northeastern part (Table 3)

Contours of the Ro values show no relationship to the Jurassic basin configuration, but rather cut across the facies and thickness trends in the northern Richardson Mountains almost at right angles.

Twenty-six samples from 10 stratigraphic sections show consistent trends increasing from low values (Ro 0.5 to 0.6; TAI 2, 2 to 2-) in the northeast, bordering northwestern Mackenzie Delta, to higher values (Ro 2.5 to 3; TAI 4-) in the southwest, in the McDougall Pass-Summit Lake area. Given the limited number of data points, the trends are relatively smooth. In general, greater depth of burial (and subsequent stripping) and/or heat flow are indicated toward the southwest. However, the trend does not continue westward into northern Yukon, and seems to be reversed there (see next section). The trend of the contours is the same as that of a series of northwest-trending faults in the northern Richardson Mountains, including the Bell River and Cache Creek faults (see Norris, 1975, 1984).

The area with the highest values in this region lies immediately west of the Bell River Fault, in the area of Summit Lake and the head of Rat River (Fig. 1, locs. 12, 16-20). This fault is the southwestern boundary of a fault-bounded block that includes the small uplift of lower Paleozoic rocks, the White, or White Mountains, Uplift. This uplift had been active intermittently since at least the Middle Jurassic, but other parts of the fault-bounded block had not been, judging by the surrounding Jurassic stratigraphy (Poulton and Callomon, 1976; Poulton et al., 1982). Presumably a long history of activity pertained, perhaps associated with local abnormally high heat flow values along the faults.

Northern Richardson Mountains - western part (Table 4)

This area includes the west side of the northern Richardson Mountains, from the Barn Range-Mount Fitton area southward through the headwaters of Driftwood River and Johnson Creek toward the mouth of Bell River and Whitefish Lake. Although the values from the western parts of the northern Richardson Mountains remain high (TAI 2-4),

Table 2. Ro and TAI values, Jurassic samples from northern Ogilvie Mountains.

Ro and TAI Analysis
Northern Ogilvie Mtns

	SAMPLE #	%Ro	TAI
2	C-81387	5.024+/-0.577	4
	C-81390	6.148+/-0.436	4
3	C-81373	6.338+/-0.201	4
	C-81375	6.783+/-0.505	4
	C-81378	5.978+/-2.899	4
	C-81384	6.185+/-0.393	4
4	C-81357	6.098+/-0.543	4

Table 3. Ro and TAI values, Jurassic samples from eastern parts of the northern Richardson Mountains.

Ro and TAI Analysis
Eastern Northern Richardson Mtns

	SAMPLE #	%Ro	TAI
9	C-80273	0.587+/-0.084	2 to 2-
	C-80256	0.546+/-0.053	2
	C-80252	0.534+/-0.022	2 to 2-
10	C-163780	1.429+/-0.282	3
	C-163781	1.489+/-0.286	3
	C-163782	1.652+/-0.314	3
11	C-163777	2.066+/-0.435	3+
	C-163778	2.153+/-0.272	3+
	C-163779	2.302+/-0.45	3+
12	C-163794	2.707+/-0.260	4-
	C-163795	---------	4-
13	C-53569	1.764+/-0.257	3+ to 4-
	C-163791	2.087+/-0.358	3+
	C-163792	2.780+/-0.284	3+
	C-163793	1.702+/-0.202	3+
14	C-53582	0.828+/-0.117	3+
	C-53583	0.766+/-0.082	3+
	C-53590	1.380+/-0.388	3
15	C-163789	0.509+/-0.031	2
	C-163790	0.550+/-0.044	2- to 2
16	C-163787	3.926+/-0.564	4-
17	C-163783	2.747+/-0.312	4-
	C-163784	2.743+/-0.380	4-
	C-163785	3.534+/-0.554	4-
18	C-53421	3.834+/-0.670	4-
19	C-80288	2.848+/-0.531	3+ to 4-

Table 4. Ro and TAI values, Jurassic samples from western parts ofthe northern Richardson Mountains.

Ro and TAI Analysis
Western Northern Richardson Mtns

	SAMPLE #	%Ro	TAI
5	C-53435	3.191+/-0.383	4-
6	44176	2.335+/-0.352	3+
7	86821	--------	3+
8	C-81335	1.599+/-0.237	2+ to 3-
20	52777	2.629+/-0.170	3+
	C-53383	4.177+/-0.510	4
21	C-81295	2.015+/-0.202	3
22	C-80879	1.684+/-0.393	2 to 2+
	C-80880	0.498+/-0.059	2

they generally tend to be lower than those in the Summit Lake — White Mountains area, suggesting a generally lower heat flow/overburden load. These results are consistent with previous (unpublished) observations on the relatively low thermal alteration characteristics of a larger number of palynomorph samples from Salmon Cache Canyon along Porcupine River just north of the mouth of Bell River, and the head of Johnson Creek. The values from the western parts of the northern Richardson Mountains are highly variable and not amenable to drawing contours, although there may be a tendency to lower values toward the south, toward Eagle Plain and the Cretaceous Eagle "Basin". These data indicate that a simple high heat flow or thick overburden model for the entire northern Yukon area is probably not appropriate and that local causes should be considered.

ACKNOWLEDGMENTS

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A conodont-based thermal maturation study of some Lower and Middle Devonian rocks, northwestern District of Mackenzie and Yukon Territory†

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Dougherty, B.J. and Uyeno, T.T., A conodont-based thermal maturation study of some Lower and Middle Devonian rocks, northwestern District of Mackenzie and Yukon Territory; in Current Research, Part G, Geological Survey of Canada, Paper 89-1G, p. 37-42, 1989.

Abstract

Thermal maturity maps, based on conodont colour alteration indices (CAI), of some Devonian rocks in northwestern Canada are presented. Samples from five Lower Devonian formations outcropping in 19 separate localities in northern Yukon Territory (NTS 116 F-K), and from four Middle Devonian formations from 26 localities in northwestern District of Mackenzie (NTS 96 E, L, M; 106 H, I, P), are plotted.

Conodonts from the Lower Devonian samples are dark, with CAI readings of 3.0 or greater. Values of this level suggest that the enclosing rocks are thermally overmature. Trends in CAI readings are not immediately apparent in this structurally complex area.

Middle Devonian conodonts are considerably lighter. Their CAI readings can be aligned northwest-southeast, the lightest being in the northeast. Samples in the vicinity of the Norman Wells oilfield have a reading of 1.5 to 2.0, which is within the oil window.

Résumé

Des cartes de maturité thermique basées sur des indices d'altération des couleurs (IAC) des conodontes sont présentées pour certaines roches dévoniennes du nord-ouest du Canada. Des échantillons de cinq formations du Dévonien inférieur affleurant dans 19 zones distinctes du nord du Yukon (SNRC 116 F-K), et de quatre formations du Dévonien moyen provenant de 26 zones du nord-ouest du district de Mackenzie (SNRC 96 E, L, M; 106 H,I,P) ont été portés sur les cartes.

Les conodontes des échantillons du Dévonien inférieur sont foncés, les IAC atteignant 4,0 et plus. Un tel niveau indique que les roches encaissantes ont atteint un stade d'hypermaturité thermique. Les IAC ne présentent pas de tendance évidente dans cette région de structure complexe.

Les conodontes du Dévonien moyen sont de couleur beaucoup plus claire, et leurs IAC peuvent être alignés dans l'axe nord-ouest - sud-est, les teintes les plus claires se trouvant dans le nord-est. Les échantillons prélevés à proximité du champ pétrolifère de Norman Wells présentent des valeurs de 1,5 à 2,0, à l'intérieur de la fenêtre de pétrole.

† Contribution to Frontier Geoscience Program.

INTRODUCTION

The purpose of this study is to outline the thermal maturation trends of Lower and Middle Devonian rocks of the northwestern part of the District of Mackenzie and of northern Yukon Territory (Fig. 1). The biostratigraphic ages and the levels of thermal maturity of the rocks were determined by evaluating the conodont faunas in each sample.

The thermal history of a basin is an important aspect of basin development. Many thermal maturation indices have been developed, tested, and proven useful in evaluating the thermal history of a given area (Héroux et al., 1979). Many of these indices were specifically developed to aid in hydrocarbon exploration, and are based on determining progressive low temperature changes in the organic material of potential or known source rocks (Bustin et al., 1985). It is necessary to have several thermal maturity indices available, as no single index can be used in every circumstance. Since organic material tends to be sparse in carbonate rocks, the need for a thermal maturation index that could be used within carbonate sequences became apparent. One such index was proposed by Epstein et al. (1977) in their paper dealing with conodont colour change.

Conodonts are the phosphatic remains of an extinct group of marine animals that existed from Late Cambrian to Triassic times. Owing to their rapid evolution, widespread distribution, and relative abundance, they are considered to be excellent biostratigraphic index fossils. Conodonts contain trace amounts of organic material. When subjected to heat, this organic material undergoes irreversible chemical changes, which results in visible colour changes in the conodonts. The conodont Colour Alteration Index (CAI) quantifies the colour change in such a way that the maximum temperature to which the conodonts were subjected can be determined (Epstein et al., 1977).

The CAI scale has eight divisions ranging from 1.0 (pale yellow), through 5.0 (black), to 8.0 (recrystallized and clear). From CAI 1.0 to 5.0 the organic matter is altered by carbon-fixing processes (Epstein et al., 1977; Rejebian et al., 1987). CAI readings greater than 5.0 are a result of progressive destruction of the organic molecules until no organic material remains. The changes can be observed as a lightening of colour until no colour is apparent at CAI 8.0 (Rejebian et al., 1987). The thermal window for commercial oil production generally corresponds with CAI 1.5 to 2.5. The gas window occurs within a CAI range of 1.0 to 4.5, while the thermal cutoff for most hydrocarbon production is CAI 4.5 (Harris et al., 1980).

CAI values of Lower and Middle Devonian conodonts were determined for over 500 samples located on the mainland, north of latitude 65° and between longitude 125° west and the Yukon-Alaska border. The sample localities were plotted on a map and any locality which was isolated from others was removed from the database. The resulting maps are shown in Figures 2 and 3. Tables 1 and 2 contain the geographic information and CAI values for each sample locality.

LOWER DEVONIAN

Lower Devonian samples tended to be restricted to between latitudes 65° and 67°N and longitude 135°W, and the Yukon-Alaska border (Fig. 2 and Table 1; NTS 116 F-K). Samples were from the Ogilvie, Michelle, Dempster, and Vittrekwa formations and an unnamed unit. All samples were from isolated outcrops or continuous surface sections located in the Ogilvie and Richardson mountains.

CAI values are high, mostly 3.0 or greater. Readings of this level suggest that the enclosing strata are thermally

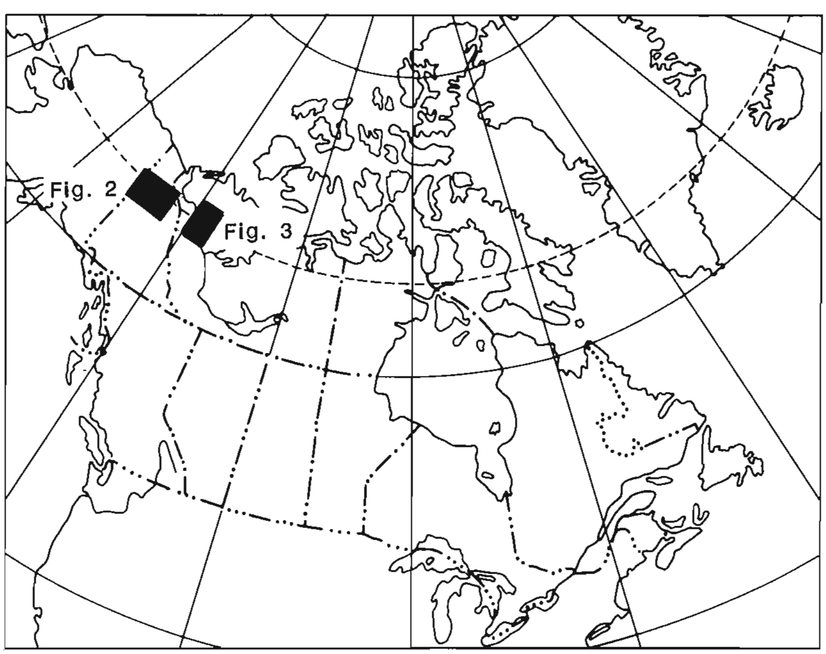

Figure 1. Index map showing study areas of Figures 2 and 3.

overmature (Epstein et al., 1977). Trends in CAI are not immediately apparent, and more samples are needed to determine whether there is a pattern in thermal maturation within the study area.

The rocks in the study area underwent a complex history of deposition and deformation (Norris, 1985). In post-Devonian time, thick upper Paleozoic and Mesozoic clastic sediments accumulated, followed by intense deformation during the Columbian and Laramide orogenies. The combination of thick overburden and deep burial resulted in the high level of thermal maturity (M.P. Cecile, pers. comm., 1988).

MIDDLE DEVONIAN

Middle Devonian samples were found to be restricted to latitude 65° and 68°N, and longitude 126° and 130°W (Fig. 3 and Table 2; NTS 96 E, L, M; 106 H, I, P). Samples were from the Gossage, Hume, Hare Indian, and Ramparts formations, the last also including the Kee Scarp Formation.

The CAI values define a set of linear northwest-southeast trends through the study area. Maturity is lowest in the northeast. There, the CAI average is 1.5 throughout the region, suggesting little alteration in the organic matter. Geochemical analyses, performed for determining potential source rocks in this region, indicate that levels of thermal maturity are immature to marginally mature (Snowdon et al., 1987).

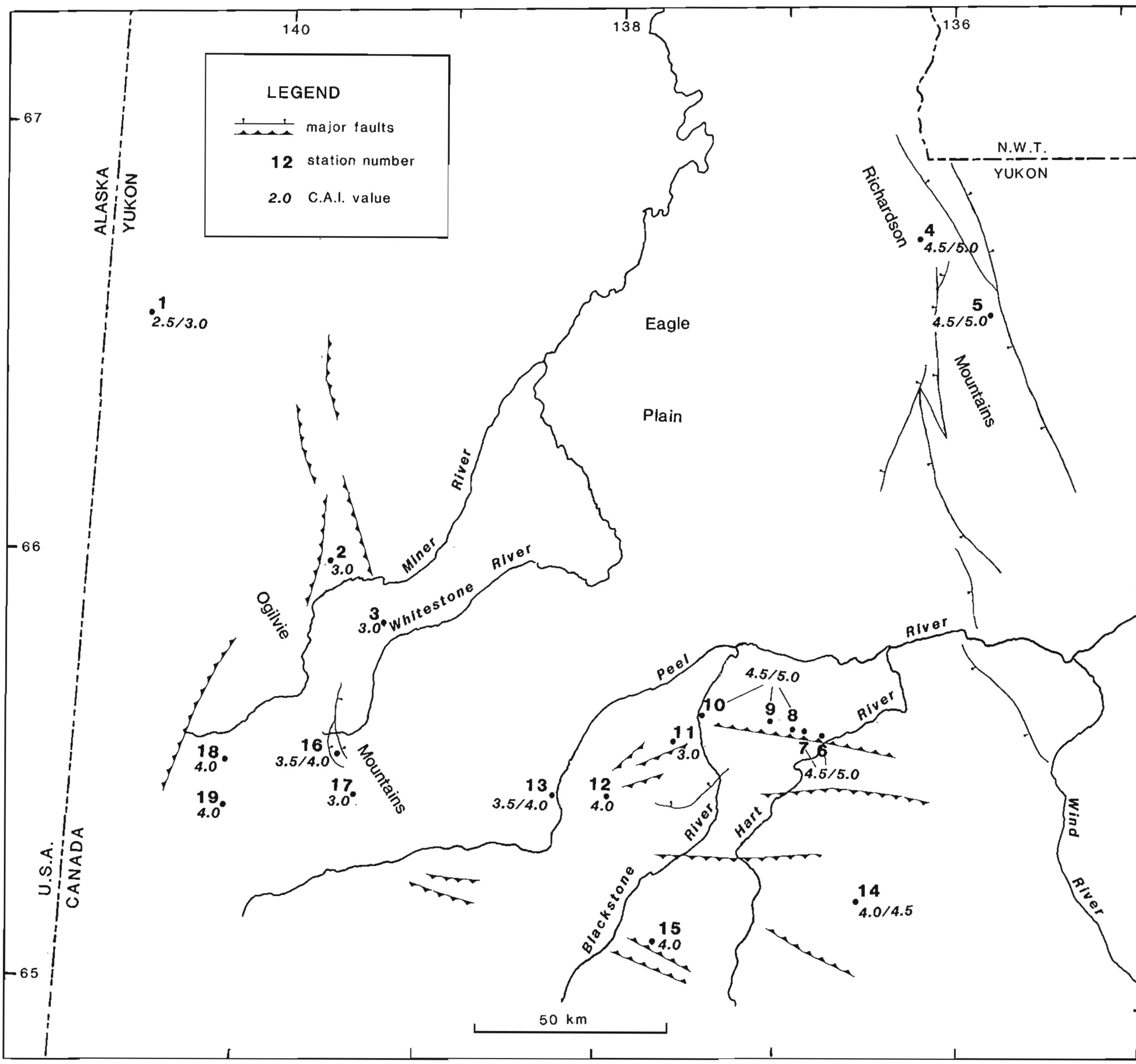

Figure 2. Thermal maturation map (based on conodont CAI values) of Lower Devonian rocks in northern Yukon Territory. (Refer to Table 1 for detailed geographic and stratigraphic information.)

Thermally mature rocks capable of producing hydrocarbons are found within areas of CAI values of 1.5/2.0 and 2.5. According to Snowdon et al. (1987), it is possible that the hydrocarbons from the Norman Wells oil field were generated from source rocks located within this thermal regime.

Areas with CAI values ranging from 2.5 to 4.5 are only capable of producing gas, as the strata have been heated beyond the range of oil production (Epstein et al., 1977). Areas with a CAI of 3.5 are therefore considered overmature.

The CAI trends reflect the structural trends of the Franklin Mountains, the major tectonic freature within the study area. Deformation during the Cretaceous-Tertiary and the loading which resulted in the development of the Franklin Mountains, also contributed to the maturation pattern.

SUMMARY

This paper presents the initial compilation of conodont CAI values from northwestern Canada. It shows that within the Devonian System, differing tectonic terranes have varying levels of thermal maturity.

Hydrocarbon exploration can be aided by using CAI to delineate thermally mature, potential source rocks, particularly within carbonate sequences. The relatively undeformed Middle Devonian rocks in the central part of the Mackenzie Corridor are shown to be thermally mature.

The use of CAI values in a regional geological study can aid in establishing and delineating structural trends within the area. Also, areas with anomalously high thermal readings may allow tracking of hot spots and detecting of igneous intrusions. In northern Yukon Territory, more frequent

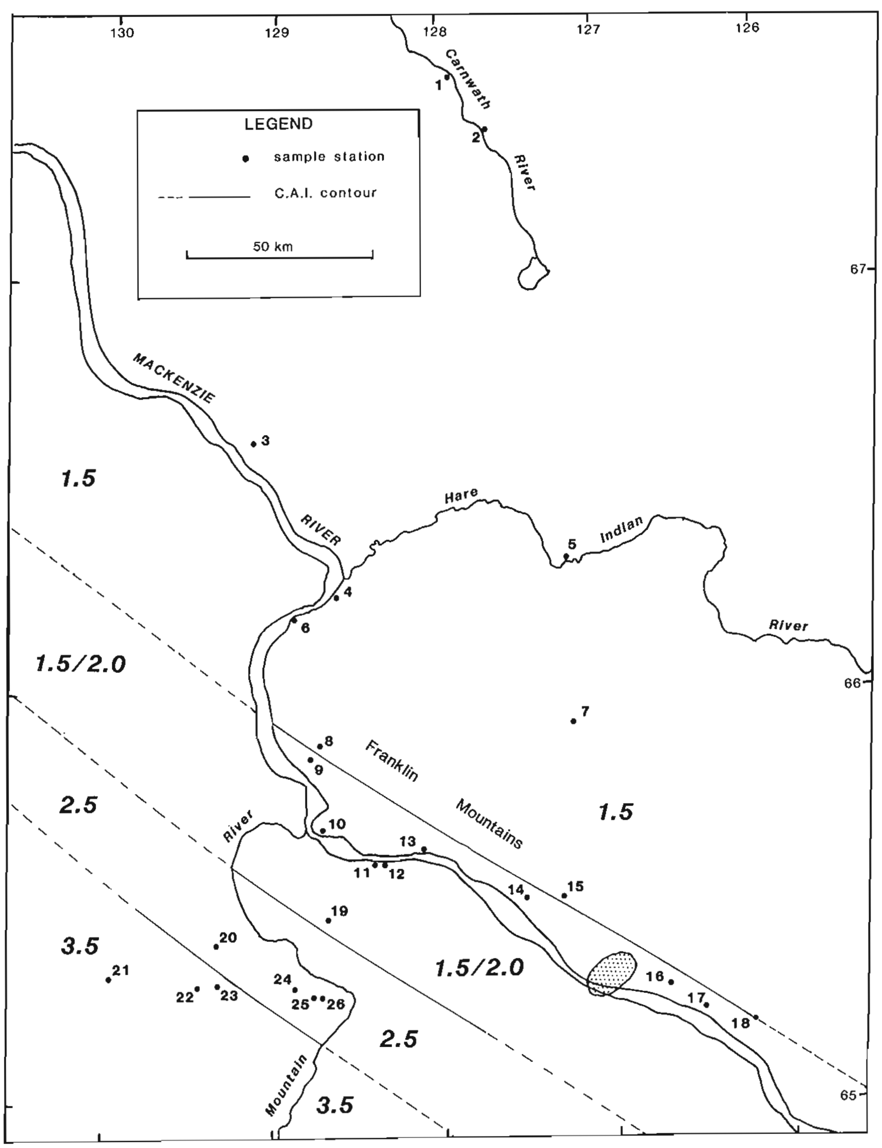

Figure 3. Thermal maturation map (based on conodont CAI values) of Middle Devonian rocks in northwestern District of Mackenzie. Stippled area represents the Norman Wells oilfield. (Refer to Table 2 for detailed geographic and stratigraphic information.)

Table 1. Locations and CAI values of Lower Devonian Samples

Station number	Latitude (north)	Longitude (west)	Formation(s)	CAI value
1	66°34'30"	140°47'	unnamed	2.5 - 3.0
2	66°01'	139°36'	Ogilvie	3.0
3	65°54'45"	139°14'	Ogilvie	3.0
4	66°50'	136°15'	Dempster	4.5 - 5.0
5	66°40'	135°48'	Dempster, Vittrekwa	4.5 - 5.0
6	65°38'	136°45'	Michelle, Ogilvie	4.5 - 5.0
7	65°39'	136°48'	Ogilvie	4.5 - 5.0
8	65°39'30"	136°50'	Ogilive	4.5 - 5.0
9	65°41'30"	137°01'	Ogilvie	4.5 - 5.0
10	65°41'30°	137°26'	Michelle	4.5 - 5.0
11	65°37'30"	137°38'	Michelle	3.0
12	65°29'	137°57'30"	Michell, Ogilvie	4.0
13	65°28'	138°13'30	Michelle	3.5 - 4.0
14	65°15'25"	136°31'20"	Ogilvie	4.0 - 4.5
15	65°10'	137°43'	Ogilvie	4.0
16	65°34'30"	139°27'30"	Ogilvie	3.5 - 4.0
17	65°27'	139°25'	Ogilvie	3.0
18	65°32'	140°07'	Ogilvie	4.0
19	65°24'	140°06'	Ogilvie	4.0

Table 2. Locations and CAI values of Middle Devonian Samples

Station number	Latitude (north)	Longitude (west)	Formation(s)	CAI value
1	67°32'	127°56'30"	not listed	1.5
2	67°23°	127°44'	not listed	1.5
3	66°37'	129°08'	Hare Indian, Ramparts	1.5 - 2.0
4	66°14'00"	128°41'50"	Hare Indian, Ramparts	1.5 - 2.0
5	66°20'	127°16'30"		1.5
6	66°11'36"	128°53'40"	Ramparts	1.0 - 1.5
7	65°57'	127°15'	not given	1.5 - 2.0
8	65°52'38"	128°44'31"	Ramparts	1.0 - 1.5
9	65°51'29"	128°47'06"	Hare Indian, Ramparts	1.5 - 2.0
10	65°41'30"	128°43'	Ramparts	1.5
11	65°36'45"	128°23'	Ramparts	1.5 - 2.0
12	65°37'45"	128°14'30"	Hare Indian, Ramparts	1.5 - 2.0
13	65°38'15"	128°08'	Ramparts	1.5 - 2.0
14	65°30'45"	127°32'30"	Ramparts	1.5 - 2.0
15	65°30'	127°21'	Ramparts	1.0 - 1.5
16	65°18'	126°43'	Ramparts	1.5 - 2.0
17	65°14'20"	126°23'40"	Hare Indian	1.5 - 2.0
18	65°12'30"	126°13'	Gossage, Hume	1.5 - 2.0
19	65°27'	128°40'	Ramparts	1.5 - 2.0
20	65°24'45"	129°21'15"	Hume	2.5 - 3.0
21	65°19'45"	129'59'	Gossage, Hume, Hare Indian	3.5 - 4.0
22	65°17'30"	129°24'	Ramparts	3.0 - 3.5
23	65°17'30"	129°21'15"	Hume, Hare Indian	3.5 - 4.0
24	65°17'30"	128°52'30"	Hare Indian, Ramparts	2.5 - 3.0
25	65°16'30"	128°47'	Hume, Hare Indian, Ramparts	2.5
26	65°16'30"	128°46'	Hume, Hare Indian, Ramparts	2.0 - 2.5

sampling of Lower Devonian rocks may show trends in the CAI values that may aid in unravelling the geological history of a basin.

The study area will be extended southward, and this should add to our understanding of the overall geological history. Also planned for the future is the addition of CAI values from other systems. Showing how CAI values vary through time should add another dimension to our understanding of basin development.

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Preliminary organic maturation studies of Horn River strata in the Tathlina High area, Northwest Territories†

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Feinstein, S., Goodarzi, F., Gentzis, T., Snowdon, L.R., and Williams, G.K., Preliminary organic maturation studies of Horn River strata in the Tathlina High area, Northwest Territories; in Current Research, Part G, Geological Survey of Canada, Paper 89-1G, p.43-50, 1989.

† Contribution to Frontier Geoscience Program.

As part of a regional study of thermal maturation in the Mackenzie Corridor (Feinstein et al., 1988); vitrinte reflectance levels have been both measured and calculated, and Rock-Eval pyrolysis Tmax determinations (Table 1) have been carried out for samples of Horn River and equivalent strata (Williams, 1983 and references therein) from a number of wells in the Trout Lake/Great Slave Lake area of the southern Northwest Territories. These data (Feinstein et al., 1988, Tables 2-4) have been used along with other results (Geochem/AGAT, 1977; Gunther and Meijer Drees, 1977; Snowdon et al., 1987) to interpret local maps of thermal maturation (Fig. 1, 2).

The isopleths, or lines of equal maturity (reflectance in Figure 1 and Tmax in Figure 2), indicate a general increase in maturity to the west-southwest in response to increasing burial of the Horn River and equivalent units and/or increasing heat flow toward the present mountain front. Significant deflections of the isopleths do occur in the central portion of the area, and these are possibly related to northeast trending fault zones; for example, the Liard and Celibeta fault zones (Williams, 1977; 1981). Higher than expected thermal maturity between these fault zones may be the result of structural relief and concomitant increased burial or, more likely, anomalous heat flow associated with subsurface fluid movement focussed by the fault system. Present-day temperatures (Majorowicz et al., 1988) are consistent with relatively high geothermal gradients in this area.

The level of thermal maturation of the Horn River Formation ranges from less than that normally used as the beginning of the oil generation window, through the conventional oil window, to overmature (0.41 to >2% Ro). Significant shows of oil and gas have been reported in this area (Fig. 4). Both oil and gas occur within and updip from the inferred moderately mature to mature area, but only gas has been reported for the area indicated to be beyond the oil window.

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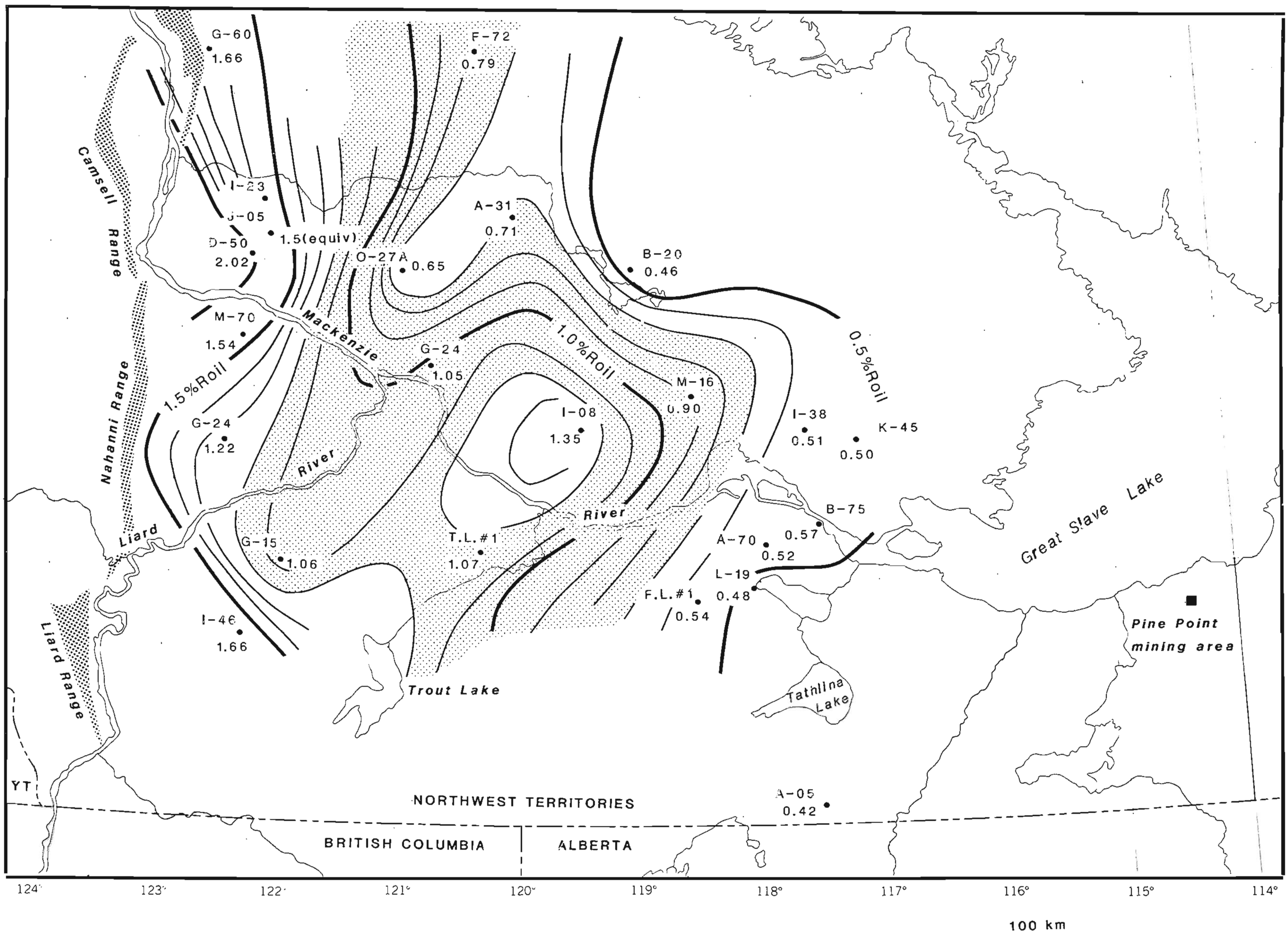

Figure 1. Maturity isopleths interpreted from vitirinite reflectance values. Lightly stippled area indicates conventional oil window (0.7 to 1.2% Ro).

Table 1. Reflected light microscopy and rock-eval pyrolysis data

SAMPLE	VITRINITE REFLECTANCE					PYROBITUMEN		ROCK-EVAL	
		%Ro RANGE		MEAN	STD.		VRo		
DEPTH: m (ft) FORMATION	n	MIN.	MAX.	%Ro	DEV.	MEAN	EQUI.	TMAX	PI
ARROWHEAD I-46[1]									
60°45′N, 122°22′W									
1858.4 (6097.0) Muskwa	5	1.42	1.65	1.56	0.09	2.13	1.71	560	.04
1870.4 (6136.5) Muskwa	35	2.05	2.66	2.27	0.12	2.98	2.23		
1870.4 (6136.5) Muskwa	3	1.72	1.85	1.77		2.16	1.73	565	.01
1980.3 (6497.0) Slave Pt.-Sulphur Pt.	5	1.45	1.59	1.54		1.74	1.47		
1980.3 (6497.0) Slave Pt.-Sulphur Pt.						2.52	1.95		
CAMERON HILLS A-05[1]									
60°04′N, 117°30′W									
1283.2 (4210.0) Hay River	3	.41	.42	.41				433	.20
CARTRIDGE F-72[1]									
63°11′N, 120°29′W									
202.7 (665.0) Horn River (B)	7	.75	.80	.78	0.02			446	.14
207.0 (679.0) Horn River (B)	2	.77	.83	.80				445	.12
CORM LAKE G-15[1]									
61°04′N, 122°02′W									
1185.7 (3890.0) Horn River (B)	4	1.01	1.17	1.06		1.14	1.10	509	.23
EBBUTT D-50[1]									
62°20′N, 122°15′W									
445.3 (1461.0) Horn River (B)	8(R)	1.84	2.22	2.00	0.11	2.59	2.00	530	.09
451.1 (1480.0) Horn River (B)[2]				3.09					
460.3 (1510.0) Horn River (B)[2]				2.83	0.11			524	.05
466.3 (1530.0) Horn River (B)	8(R)	1.91	2.20	2.04	.113			568	.17
EBBUTT J-05[1]									
62°24′N, 122°15′W									
655.3 (2150.0) Horn River (B)						1.49	1.31	518	.36
FISH LAKE G-60[1]									
62°09′N, 122°55′W									
457.2 (1500.0) Canol (E)	5	1.50	1.63	1.57	0.05	2.19	1.74	512	.49
499.9 (1640.0) Hare Indian	4	1.57	1.83	1.75	0.11				1.00
FOETUS LAKE No. 1[1]									
60°55′N, 118°31′W									
791.6 (2597.0) Pine Point	9	.50	.56	.54	0.02			438	.06
FORT SIMPSON M-70[1]									
61°59′N, 122°27′W									
527.3 (1730.0) Horn River (B)	7	1.40	1.60	1.54	0.07			524	.82
HARRIS RIVER A-31[1]									
62°30′N, 120°06′W									
164.6 (540.0) Horn River (B)	3	.68	.72	.71				447	.15
173.7 (570.0) Horn River (B)	3	.68	.72	.71				447	.19
207.3 (680.0) Horn River (B)	5	.70	.74	.72	.015			446	.17
HIGHLAND LAKE I-23[1]									
62°33′N, 122°20′W									
560.8 (1840.0) Horn River (B)						1.50	1.32	555	.24
KAKISKA LAKE L-19[1]									
60°58′N, 118°03′W									
660.8 (2168.0) Evie (like)	12	.44	.50	.48	0.02			441	.06

Table 1. Continued

SAMPLE	VITRINITE REFLECTANCE					PYROBITUMEN		ROCK-EVAL	
DEPTH: m (ft) FORMATION		%Ro RANGE		MEAN	STD.		VRo		
	n	MIN.	MAX.	%Ro	DEV.	MEAN	EQUI.	TMAX	PI
LAFERTE RIVER M-16[1]									
61°45′N, 118°33′W									
314.3 (1031.0) Horn River (A)	7	.81	.85	.83	0.013			443	.18
359.7 (1180.0) Evie	6	.87	.94	.92	0.03	1.07	1.06	441	.11
MILLS LAKE A-70[1]									
61°09′N, 117°56′W									
647.7 (2125.0) Lonely Bay	2	.51	.52	.52				442	.07
640.4 (2101.0) Evie	3	.66	.73	.71				438	.10
MILLS LAKE B-75[1]									
61°14′N, 117°28′W									
448.3 (1470.7) Evie	3	.57	.59	.58				438	.09
443.6 (1455.5) Evie	9	.54	.59	.56	0.02			438	.04
MINK LAKE I-38[1]									
61°37′N, 117°35′W									
266.7 (875.0) Horn River (A)	21	.44	.56	.49	0.03			440	.08
277.7 (911.0) Horn River (A)	7	.49	.57	.52	0.03			444	.10
PROVIDENCE K-45[1]									
61°34′N, 117°47′W									
142.8 (468.5) Horn River A[2]				.57					
251.9 (826.5) Evie				.52					
252.2 (827.5) Evie[2]				.54					
210.6 (691.0) Horn River[2]	3	.48	.52	.50				440	.13
254.5 (835.0) Evie	4	.49	.53	.50	0.02			439	.12
RABBITSKIN I-08									
61°37′N, 119°30′W									
421.2 (1382.0) Horn River B[2]				1.32					
429.5 (1409.0) Horn River B[1]	5	1.35	1.43	1.38	0.02			463	.44
SIBB LAKE G-24[1]									
61°33′N, 122°34′W									
487.7 (1600.0) Horn River B	5	1.17	1.27	1.22	0.04	1.54	1.34	442	.27
STRONG POINT G-24[1]									
61°53′N, 120°49′W									
414.5 (1360.0) Horn River B	10	.96	1.14	1.05		1.33	1.22	454	.26
TURKEY LAKE No. 1[1]									
61°07′N, 120°22′W									
905.3 (2970.0) Horn River	3	1.04	1.08	1.07		1.31	1.20	456	.35
WILLOW LAKE B-20[1]									
62°17′N, 119°04′W									
621.8 (2040.0) Horn River B	3	.44	.46	.46				437	.15
577.6 (1895.0) Horn River B	4	.39	.43	.42	0.02			438	.12
WILLOW LAKE O-27A[1]									
62°16′N, 121°04′W									
362.7 (1190.0) Horn River B	5	.61	.68	.65	0.03			453	.25

[1]Feinstein, S., Brooks, P.W., Gentzis, T., Goodarzi, F., Snowdon, L.R., and Williams, G.K., 1988, Thermal maturity in the Mackenzie Corridor, Northwest and Yukon Territories, Canada; Geological Survey of Canada, Open File Report 1944.
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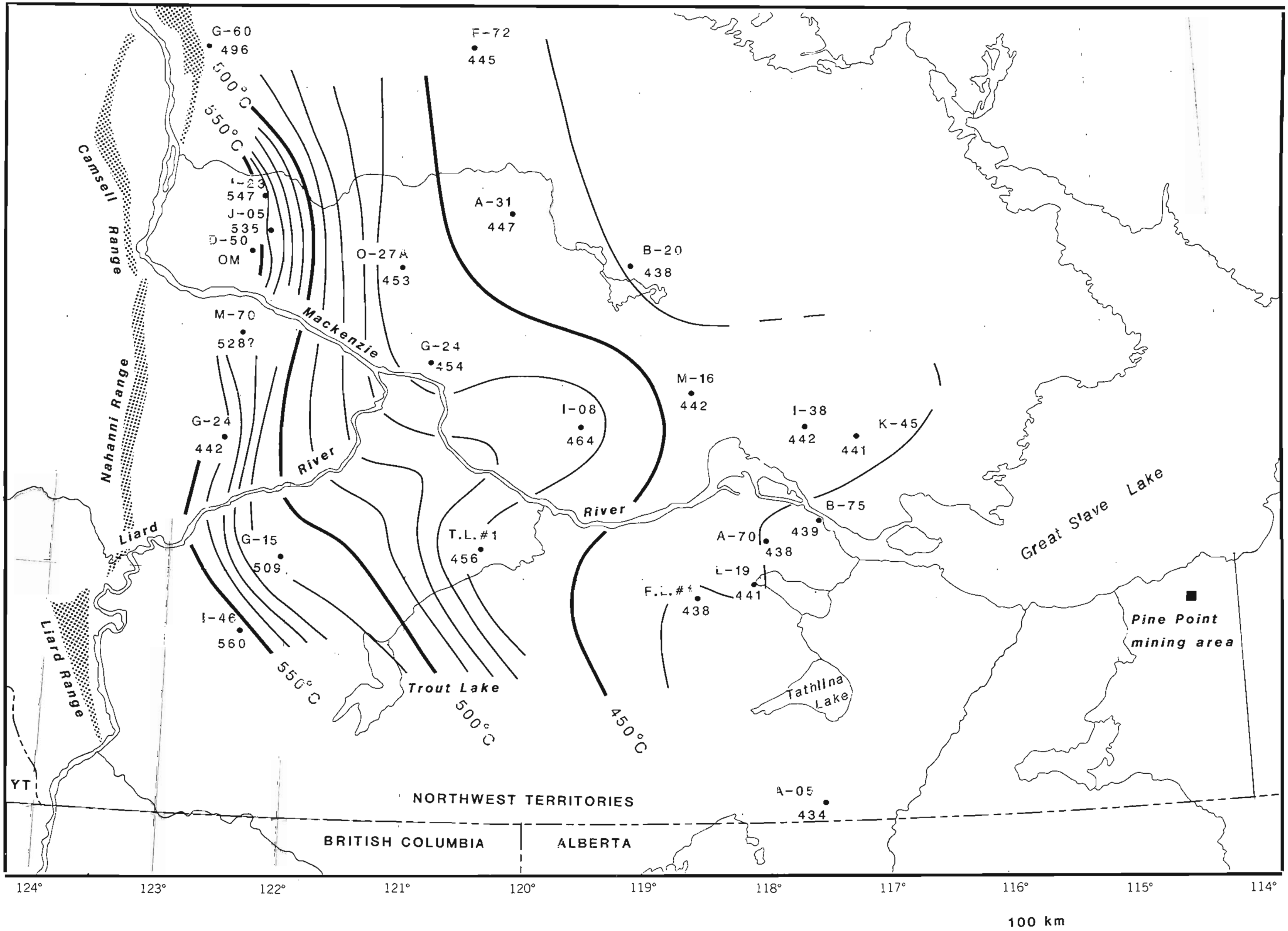

Figure 2. Maturity isopleths interpreted from Rock-Eval Tmax measurements. OM = overmature.

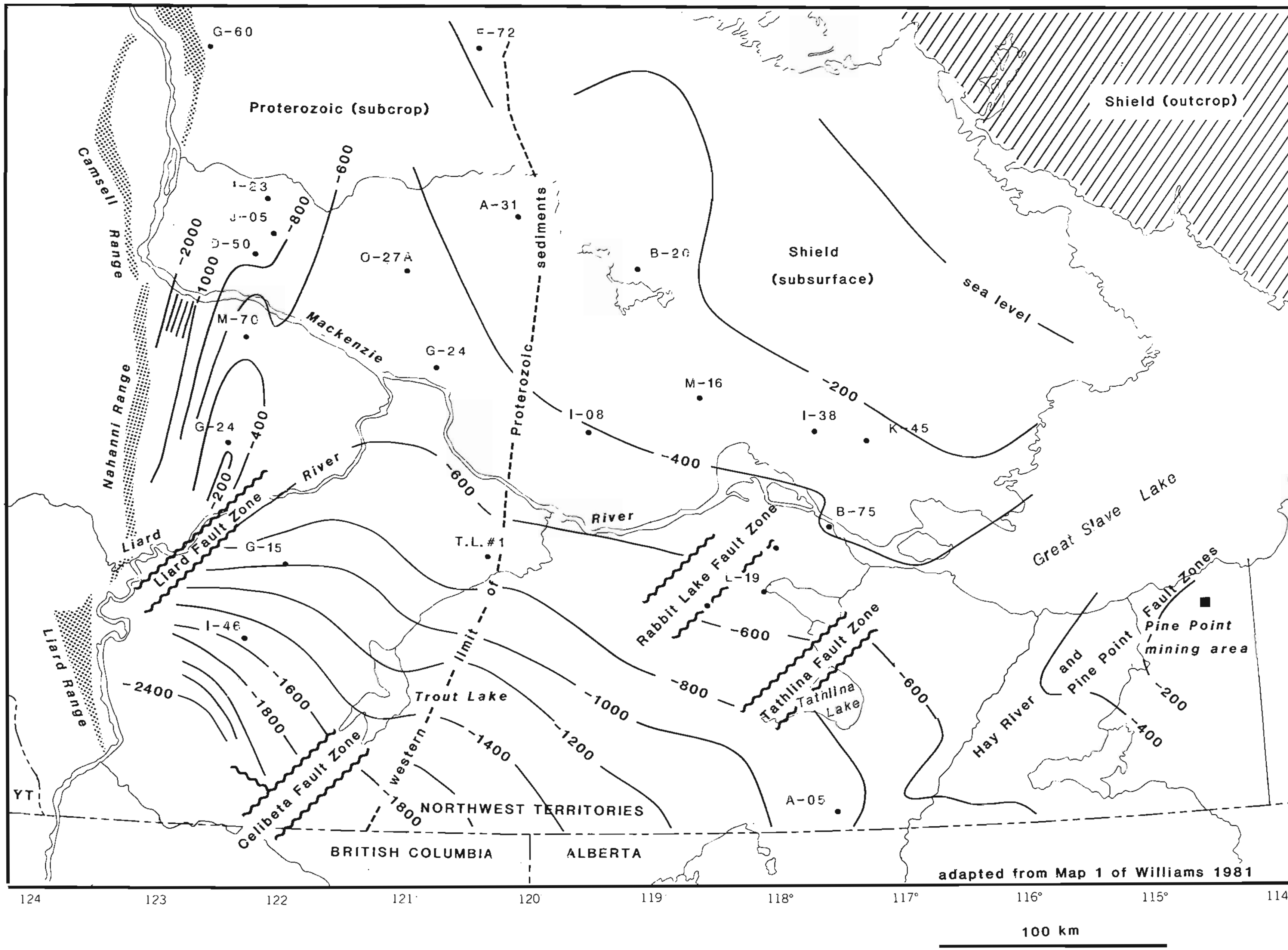

Figure 3. Structure of the Precambrian; contour in metres, sea level as datum (adapted from Williams, 1981).

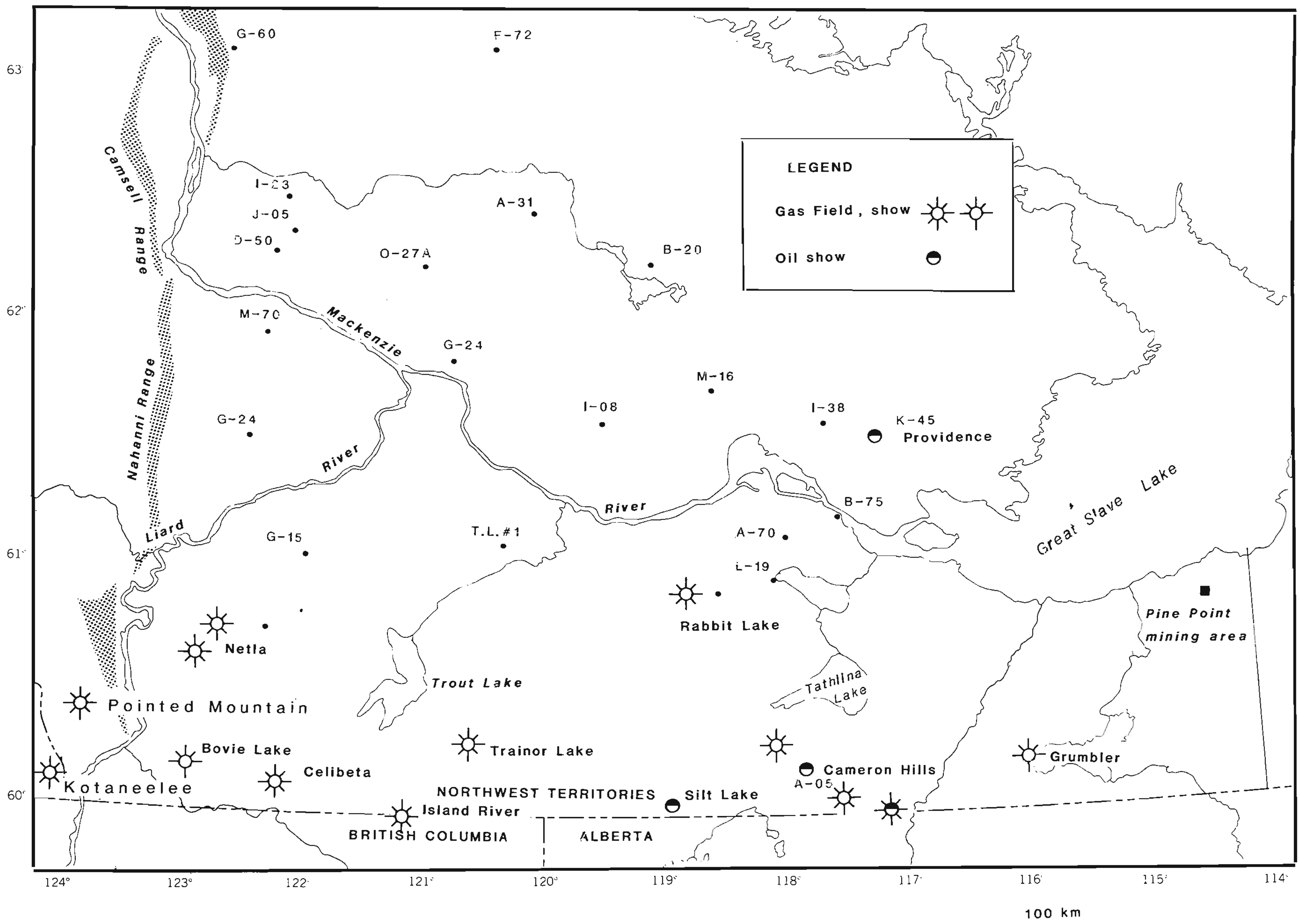

Figure 4. Location map of significant gas and oil finds in the Trout Lake/Slave Lake area.

Optical and compositional characters and paleothermal implications of a diverse suite of natural bitumens from Middle Devonian carbonate rocks, Pine Point, Northwest Territories[†]

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Kirste, D., Fowler, M.G., Goodarzi, F., and Macqueen, R.W., Optical and compositional characters and paleothermal implications of a diverse suite of natural bitumens from Middle Devonian carbonate rocks, Pine Point, Northwest Territories; in Current Research, Part G, Geological Survey of Canada, Paper 89-1G, p. 51-55, 1989.

Abstract

The Pine Point lead-zinc property offers a unique opportunity to study the variations and origins of a diverse suite of natural bitumens derived from a single source facies and occurring in a geographically and geologically limited area — features unknown elsewhere. Reflectance, fluorescence, and solubility studies indicate that the Pine Point property hosts natural bitumen species ranging from asphalt to gilsonite, glance pitch, grahamite and epi-impsonite, as well as an unclassified species. Biomarker distributors from bitumen extracts display non-biodegraded to severely biodegraded fingerprints and low to moderate maturity, and confirm the F facies as the source facies, as earlier concluded by conventional organic geochemistry and stratigraphic studies. The Pine Point bitumen suite verifies that the optical character of natural bitumen combined with chemical composition may be a very sensitive index of thermal and chemical environments that were present in the evolution and alteration of these natural bitumens.

Résumé

La propriété plombo-zincifère de Pine Point offre une occasion unique d'étudier les variations et les origines d'une succession variée de bitumes naturels provenant d'un même faciès et se manifestant dans une région limitée sur les plans géographique et géologique, soit des caractéristiques inconnues ailleurs. Des études de la réflectance, de la fluorescence et de la solubilité indiquent que la propriété de Pine Point recèle des espèces naturelles de bitume allant de l'asphalte à la gilsonite, l'asphaltite, la grahamite et l'épi-impsonite, ainsi que des espèces non classées. La distribution de biomarqueurs dans des extraits de bitume présente des indices de biodégradation nulle à très forte et un degré de maturité faible à moyen, et confirme que le faciès F est le faciès d'origine, tel qu'établi antérieurement à la suite d'études de géochimie organique et de stratigraphie classiques. La succession des bitumes de Pine Point confirme que le caractère optique du bitume naturel, combiné à sa composition chimique, peut être un indice très juste des milieux thermique et chimique qui ont présidé à l'évolution et à l'altération de ces bitumes naturels.

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INTRODUCTION

Natural bitumens are secondary organic matter derived from the heavier components of crude oils. They may be produced during early stages of maturation of suitable organic matter, or by biodegradation, water-wasting, oxidation/weathering, and/or by advanced thermal maturation (Rogers et al., 1974). Natural bitumens occur throughout the sedimentary record, and are greater in proportion to the other organic matter in lower Paleozoic rocks, making them of particular interest in assessing the maturation of rocks devoid of the land-plant-derived maceral vitrinite, the most common maturation indicator for Devonian and younger sediments.

Comprehensive studies of bitumens include those of Curiale (1986), Jacob (1967, 1973), and Jacob and Wehner (1981). Jacob (ibid.) and Jacob and Wehner (ibid.), following a study of a wide-ranging suite of bitumens from various geographic locations, devised a bitumen classification scheme based on reflectance, fluorescence, solubility in organic solvents, and melting temperature. Although this scheme is useful for assessing maturation, the classification parameters that were used minimize important genetic factors related to source and mode of origin, including level of biodegradation.

One important means of identifying the origin of natural bitumens is through the study of the biomarkers that occur in bitumens. Biomarkers are compounds whose carbon structures are closely related to those produced by living eukaryotic organisms. As a single example, steroids, which occur in many species of eukaryotes, are transformed by diagenetic reactions in sedimentary environments to steranes, which are polycyclic alkanes widespread in source rocks, petroleum and natural bitumens. The value of the biomarker signature of natural bitumens was discussed first by Douglas and Grantham (1974) and more recently by Williams and Goodarzi (1981) and Curiale (1986). No study has yet been reported, however, that involves the petrography and geochemistry (including biomarkers) of a range of bitumen species that have been derived from a single source facies. Pine Point, a recently operating zinc-lead mine, located on the south shore of Great Slave Lake in the southern Northwest Territories, provides such an opportunity.

PINE POINT SETTING

As originally outlined by Skall (1975) and recently refined by Rhodes et al. (1984), zinc-lead orebodies at Pine Point, numbering close to 100, are located within a nearly flat-lying Middle Devonian carbonate complex. These host rocks have been uplifted, subjected to widespread karst development, dolomitized and gently flexed and/or fractured by tectonic movements associated with underlying Precambrian rocks. Krebs and Macqueen (1984) discussed the sequence of depositional and diagenetic events that affected the host rocks and led to the development of the orebodies.

BITUMENS AT PINE POINT

Associated with the host rocks and the orebodies are a variety of natural bitumens, noted by early workers, but first studied systematically by Macqueen and Powell (1983), Powell and Macqueen (1984) and Powell (1984). Macqueen and Powell (1983) recognized two main varieties of bitumen: a soft, flowing variety, mostly to completely soluble in organic solvents and with atomic H/C ratios of $\cong 1.4$, and sulphur contents averaging 7.8 %; and a second variety, termed "altered bitumen", which is brittle and glassy, largely to completely insoluble in organic solvents, and has atomic H/C ratios of $\cong 1.02$ and sulphur contents averaging 22 %.

Powell and Macqueen (1984) and Powell (1984) proposed a source facies for the Pine Point bitumens on stratigraphic and organic geochemical evidence, and advocated a thermochemical sulphate-reducing reaction to explain the origin of the altered bitumens and the zinc-lead orebodies. The reaction proposed would yield H_2S, S° and altered bitumens at temperatures of $\cong 100\,^\circ C$.

Bitumens at Pine Point are sourced by an organic-rich (up to 30 % total organic carbon), hypersaline, sulphur-rich, euxinic, argillaceous carbonate unit termed the F-facies (Skall, 1975; Macqueen and Powell, 1983). Migration appears to be very local, on the order of metres to hundreds of metres. The bitumens have undergone various levels of biodegradation and thermal alteration, resulting in a diverse population of bitumen species (Macqueen and Powell, 1983; Powell, 1984; Powell and Macqueen, 1984). The present study is an extension and refinement of this work.

SAMPLE SUITE

The Pine Point property contains core from more than 12,000 cored diamond drill holes, and more than 40 previously active open pits. Some 360 samples of bitumens and host rocks were collected from 5 open pits and 32 diamond drill cores. Of these, about 180 bitumen and source facies samples were selected for detailed study.

ANALYTICAL METHODS

Microscopic methods used are those described by Jacob and Wehner (1981). Reflectances of bitumen samples were determined under water and converted into oil reflectances using Jacob's Formula (Jacob, 1975). Elemental analyses for C, H and N were performed on a Perkin Elmer elemental analyzer and S was determined on a Leco SC32 sulphur and carbon analyzer. Selected samples of bitumens and source facies were subjected to soxhlet extraction, column chromatrography and gas chromatography. A limited number of samples were examined by gas chromatography-mass spectrometry for biomarker content, following methods described by Snowdon et al. (1987).

RESULTS AND DISCUSSION

Bitumen reflectance values range from Ro = 0.01 to 1.61 %, and fluorescence intensity ranges from FI = 12.0 to 0.01 % (Table 1). These values and the results of solubility tests indicate that Pine Point bitumen species range from asphalt to gilsonite, glance pitch and grahamite, and, finally, to epi-impsonite (Tables 1, 2). Most samples appear to be homogeneous and isotropic in nature. A common

Table 1. Classification of optical and chemical properties of natural bitumen species at Pine Point

Bitumen species*	R_o range*	FI range*	Observed number of Pine Point specimens	H/C	S/C	% Sulphur
Asphalt	0.02-0.07%	0.4-4.0%	4	1.4	0.03-0.04	≃8%
Gilsonite	0.07-0.11%	0.05-0.4%	11	1.5-1.3	0.04-0.05	8-10%
Glance Pitch	0.11-0.30%	0.05-0.22%	42	1.5-1.2	0.03-0.07	7-14%
Grahamite	0.30-0.70%	≤0.05%	30	1.0-1.1	0.05-0.11	9-22%
Epi-impsonite	0.70-2.0%	≤0.02%	71	1.1-0.6	0.08-0.29	17-43%

*From Jacob and Wehner, 1981

Table 2. Optical properties of Pine Point bitumens

Type	Number observed	Description
Homogeneous	80	Homogeneous, isotropic, variable reflectance (0.03-1.5% R_o).
Heterogeneous		
Type A[1]	20	Homogeneous, isotropic, low reflecting (<0.30 R_o); main mass with small, rounded bodies of low reflecting (<0.10 R_o), high fluorescing (>2% FI) bitumen; bright yellow under ultraviolet light (UV); may or may not flow upon introduction of heat from ultraviolet source.
Type B[2]	15	Inhomogeneous mixture of low reflecting (<0.20% R_o) moderately fluorescing (<0.35% FI) (red in UV) bitumen; and low reflecting (<0.10% R_o), high fluorescing (0.35<F1<1.00%) (yellow brown in UV) bitumen that flows upon introduction of heat.
Type C[3]	20	Inhomogeneous mixture showing insoluble phase mixing commonly with higher reflecting (>0.30% R_o) and low reflecting (at least 0.20% R_o lower) with slightly anisotropic (<0.60% R_o) bands; fluorescence (0.02%→12%) bright yellow to black in UV.
Type D[4]	24	High reflecting (>0.5% R_o) isotropic bitumen containing veins of lower reflecting (0.10-0.5% R_o lower) bitumen.

[1] Includes gilsonite or glance pitch, and unclassified bitumen.
[2] Includes asphalt and/or gilsonite and/ or glance pitch.
[3] Includes unclassified bitumen.
[4] Includes grahamite and/ or ep-impsonite.

observation for natural bitumens is the presence of textures indicating more than one generation, apparently the result of the mobility of these materials in their liquid phase (Jacob and Wehner, 1981). At Pine Point this phenomenon was observed to occur in four separate forms (Table 2). The presence of high reflecting dual and triple bitumen generations supports the proposed episodic pulsating flow of hot waters, some metal-bearing, as suggested by Krebs and Macqueen (1984).

There is no obvious correlation between depth and reflectance, which is not unexpected for this region, in which hot waters were channelled through fracture conduits and karst structures.

Although not presented here, the biomarker distributions are characteristic of oils sourced by hypersaline, reducing, carbonate environments (e.g. even-odd predominance in normal alkanes, low pristane/phytane ratios, and high abundance of extended hopanes (C_{31}-C_{35})). High sulphur content and early migration from the F facies source rocks indicate a sulphur-rich depositional environment (Orr, 1986). A sulphur-rich depositional environment is further supported by the presence of highly distinctive alkyl benzenes (aryl isoprenoids), which are aromatic compounds considered to be biomarkers of sulphur oxidizing bacteria (Powell, 1984; Summons and Powell, 1987). The triterpane and sterane biomarker distributions also suggest a diverse biota as primary source input, that has undergone extensive bacterial reworking.

The level of maturation of those bitumens examined for biomarker content is of early to moderate maturity as indicated by stereochemical isomerization:

$$(\mathrm{C29}\ \alpha\alpha\alpha S/S+R\ \text{sterane} = <0.5;\ \mathrm{C29}\ \frac{\alpha\beta\beta}{\alpha\alpha\alpha + \alpha\beta\beta}\ \text{sterane} = <0.6)$$

It must be noted, however, that once a reflectance of approximately 0.50% Ro is reached, the bitumens are essentially insoluble, ruling out the study of biomarkers, which must be obtained through extracting soluble portions of bitumens in organic solvents. A reflectance of 0.50% Ro for bitumen is approximately equivalent to 0.80% Ro vitrinite (Jacob and Hiltmann, 1988), approximately the peak of oil generation for most materials. Of particular interest is the separation of Pine Point biomarker results into two main groups, one corresponding to type C bitumens (Table 2), and a second comprising the remainder of the samples analyzed. The biomarker ratios used to recognize these groups include the C_{27} $18\alpha/17?$ trisnorhopane ratio (Ts/Tm), C_{23} tricyclic/C_{30} hopane ratio, and C_{29} norhopane/C_{30} hopane ratio, all of which are high for type C samples (Table 2) and low for all other samples. Sterane ratios also support the recognition of two groups — type C bitumens, and all others. The meaning of these groups (variable source facies? differing degrees of biodegradation? acid-catalyzed reactions?) is being investigated.

The sensitivity of natural bitumens at Pine Point to differing temperatures is obvious from the data appearing in Tables 1 and 2. In the case of the two bitumen groups, as defined by biomarkers, there are no significant changes in the biomarker standard maturity ratios (Mackenzie, 1984). Although this observation may relate to the relatively narrow range of maturity (reflectance range of 0.17-0.50% R_o; approximately equivalent to 0.50-0.80% Ro vitrinite), it may also reflect the likelihood that the time of exposure of the bitumens to elevated temperatures was short on a geological time scale (Macqueen and Powell, 1983; Krebs and Macqueen, 1984). Thus reflectance of these bitumens appears to be a more sensitive measure of past temperature than the biomarker isomerization reactions chosen in this study.

Another question raised by these data (Tables 1, 2) lies in estimation of the maximum temperatures achieved. The highest bitumen reflectance measured when converted to its vitrinite equivalent (Rb = 1.61% = ≅ R vitrinite = 1.4%; Jacob and Hiltmann, 1988) suggests paleotemperatures between 120 and 170°C (Héroux et al., 1977) and possibly higher because of the short duration of heating. Temperatures at the higher end of this range are further indicated by the presence of pyrolitic carbon (Goodarzi, 1985) in the single sample showing a reflectance of 1.61%. Temperatures in the range of 120 to 170°C exceed those estimated by fluid inclusion temperatures in sphalerite and dolomite (Roedder, 1968). Typically, however, minerals showing fluid inclusions are relatively late in the mineralization or diagenetic sequence (Krebs and Macqueen, 1984), and it may be that bitumens at Pine Point provide a more accurate, although indirect, measure of paleotemperatures.

Meanwhile, the Pine Point setting offers a unique opportunity to study the natural variation and origin of a diverse suite of bitumens derived from a single source facies and occurring in a geographically and geologically limited area — features unknown elsewhere.

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Stratigraphy and structure of the Neruokpuk Formation, northern Yukon[†]

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Lane, L.S. and Cecile, M.P., Stratigraphy and structure of the Neruokpuk Formation, northern Yukon; in Current Research, Part G, Geological Survey of Canada, Paper 89-1G, p. 57-62, 1989.

Abstract

The results of six traverses across northeastern exposures of the Neruokpuk Formation near Firth River, northern Yukon, indicate that the local strata are lower Paleozoic. Eleven new macrofossil collections include three identifiable graptolite collections, one of Early Ordovician age, and two of Early Silurian age, six trace fossil collections (including Oldhamia *and* Planolites*) of probable Early Cambrian age, and one sample of possible vascular plant material. The stratigraphic succession is very similar to those of the nearby Barn Mountains and the distant Selwyn Basin.*

The rocks are tightly folded and thrust faulted, with predominant northeast vergence. The principal décollement zone is a maroon and green argillite horizon in the Lower Cambrian unit. Previous work indicates that most of the deformation occurred during the Paleozoic, with some reactivation during the Mesozoic-Tertiary.

Résumé

Les résultats de six cheminements recoupant des affleurements nord-est de la formation de Neruokpuk près de la rivière Firth dans le nord du Yukon, indiquent que les couches locales datent du Paléozoïque inférieur. Parmi les onze nouvelles collections de macrofossiles, il y a trois collections de graptolites identifiables, une de l'Ordovicien inférieur et deux du Silurien inférieur, six collections de fossiles en traces (notamment Oldhamia *et* Planolites*) probablement du Cambrien inférieur et un échantillon de matière végétale probablement vasculaire. La succession stratigraphique est très semblable à celles des chaînons Barn à proximité et du bassin éloigné de Selwyn.*

Les roches présentent des réseaux denses de plis et de chevauchements, de direction dominante nord-est. La zone de décollement principale est un horizon d'argilite bordeaux et vert au sein de l'unité du Cambrien inférieur. Des travaux antérieurs indiquent que la déformation s'est surtout produite pendant le Paléozoïque, avec un peu de réactivation durant le Mésozoïque-Tertiaire.

[†] Contribution to Frontier Geoscience Program

INTRODUCTION

Six traverses across a well exposed part of the Neruokpuk Formation east of the Firth River, North Yukon National Park, outlined the structure and stratigraphy over 14 km across strike (Figs. 1, 2). The area is underlain by tightly folded and thrust faulted repetitions of lower Paleozoic basinal clastics and cherts with thin units of carbonate. These preliminary results form part of an ongoing mapping and structural analysis project intended to define the structural evolution of the Beaufort Sea continental margin, funded through the Frontier Geoscience Program (FGP).

STRATIGRAPHY

Reconnaissance mapping during Operation Porcupine (Norris, 1972, 1981, 1985) identified seven gross lithological packages (Pn0 - Pn6) in the Neruokpuk, and interpreted their ages as Proterozoic. Total thickness of the entire succession was estimated to be 13.4 km (Norris, 1985).

Detailed mapping at a scale of 1:50 000 during the1988 field season, has revealed four stratigraphic units. The oldest consists of Lower Cambrian quartzites, argillites and carbonates. The upper part of this unit is dominated by distinctive maroon and green argillites with thin, ripple cross-laminated siltstone layers (**€a**). *Planolites* and *Oldhamia* trace fossils are preserved at the argillite/siltstone contacts (see Hofmann and Cecile, 1981). White weathering, dark grey quartzites and maroon sandstones (**€q**) interlayered with carbonates (**€l**) and maroon argillites, are dominant in the southwest part of the study area (Figs. 2b, 3). Overlying the Cambrian maroon and green succession are dark grey, black, and, locally, distinctive "bottle green" cherts interlayered with dark grey argillites (**Oac**) that yielded one graptolite collection of Early Ordovician age; rusty weathering green argillite (**Sa**) that locally contains Early Silurian graptolites; and, finally, a sequence of grey argillites interbedded with impure sandstones and chert-pebble conglomerates (**?Da**), locally displaying trails and burrows of indeterminate age, and possible vascular plant fragments (see Table 1). The contact of this uppermost unit with underlying rocks appears to be discordant; however, due to insufficient exposure, we are presently unable to distinguish unequivocally whether the contact is unconformable. This unit overlies the lower Silurian argillite, but appears to be involved in the intense deformation that has affected the underlying lower Paleozoic rocks.

The Cambrian to Silurian succession is very similar to that of the Barn Mountains 80 km along strike to the southeast (Cecile, 1988). Furthermore, the entire succession is remarkably similar to that of the Selwyn Basin 900 km to the southeast (Gordey, 1980). The lower Paleozoic stratigraphy in the Firth River area also continues along strike into northeastern Alaska (Reiser et al., 1980). Within the area outlined in Figure 2a, the Paleozoic succession comprises parts of units Pn0 to Pn2 of Norris (1981, 1985). Mapping westward in adjacent areas during 1987 and 1988 has also

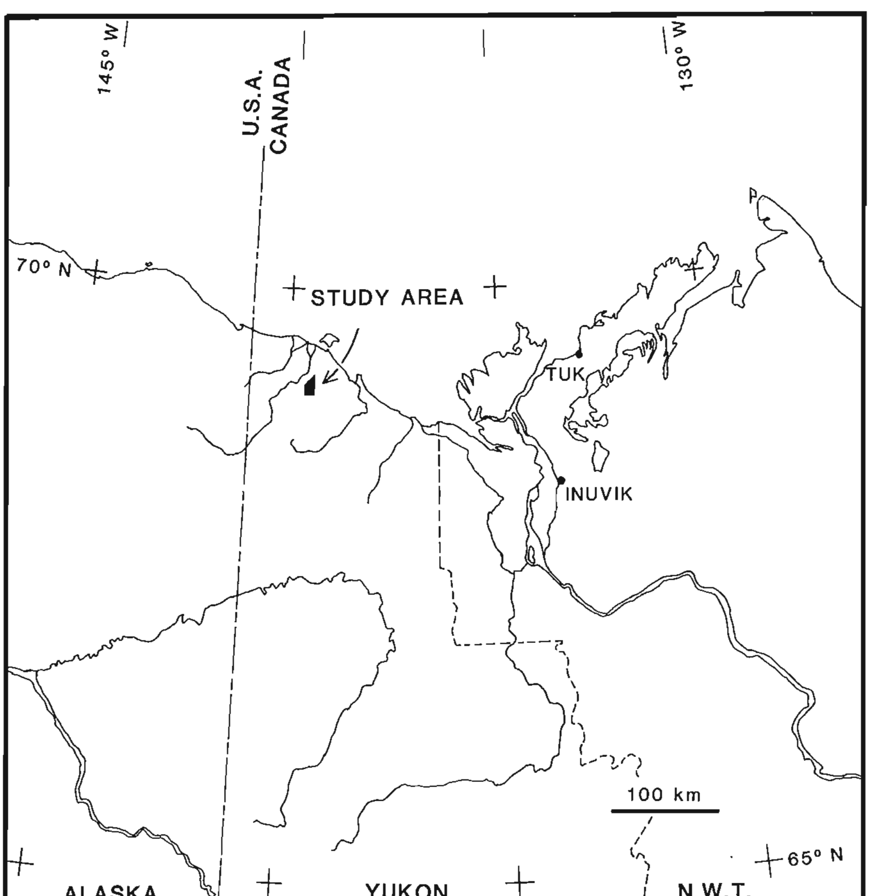


Figure 1. Location map of the Beaufort Sea margin showing the study area.

identified maroon and green argillite and grey chert successions in map units Pn3 and Pn4 (Norris, op. cit.), which may be equivalent to the Cambro- Ordovician strata identified here. Two trace fossil collections (not reported here) from Pn3 west of Malcolm River, contain *Planolites*. Five additional trace fossil collections from strata northwest of the Firth River area may also contain *Planolites*.

The similarity of the (?)unconformable argillite-psammite-conglomerate unit (Fig. 2), to the Devono-Mississippian Earn Group of the Selwyn Basin (Gordey et al., 1982); and to the Imperial and Tuttle formations of the northern Richardson Mountains (Pugh, 1983) is notable. A unit of apparently similar age and contact relationships, though perhaps somewhat different facies, has been identified in adjacent northeast Alaska (Unit Ds, Reiser et al., 1980). On the bases of the strata's similarity to published lithological descriptions of strata in adjacent areas, the occurrence of possible vascular plant fragments, and its stratigraphic position above Lower Silurian argillites, we have tentatively assigned a Devonian age to this unit.

STRUCTURE

The stratigraphic units are tightly folded and thrust faulted into panels in the order of 0.5 to 2 km thick (Fig. 2b). The scale and style of deformation are the same as those of the Barn Mountains farther southeast; the maroon and green argillite unit representing a décollement horizon (Cecile, 1988). The common occurrence of thin siltstones with well preserved ripple crosslaminae suitable for stratigraphic top determinations, together with faunal control, allow a generally well-constrained structural interpretation (Fig. 2b). A change in structural style across strike, from mainly faulting to mainly folding, occurs near the area labelled B in Figure 2, and reflects the northeastern limit of exposure of the thick Cambrian quartzites. The lithological change reflects either a major step in the basal detachment horizon or a facies change in the Cambrian succession. This ambiguity may be resolved with future mapping; however, in this paper, the cross-sections are drawn without a step in the basal detachment.

The relative contributions of Paleozoic and Mesozoic-Tertiary deformations are uncertain. Reconnaissance work in adjacent areas, together with the dramatic angular unconformity beneath the Carboniferous and younger succession (Norris, 1981) confirms that at least some of the deformation is Paleozoic in age. Although we have not mapped Mesozoic rocks in the area described, the stratigraphy and structures continue for 10 km northwestward below the Loney Syncline, a prominent structural and physiographic feature underlain by well dated Triassic and Jurassic sediments deposited unconformably on the lower Paleozoic succession. The syncline and unconformity were affected by south-directed reverse faulting during Cretaceous-Tertiary Brookian deformation, but a major component of the strain observed in the lower Paleozoic rocks clearly predates Triassic sedimentation (Norris, 1986). The dominant sense of tectonic transport is northeastward in the study area, and is consistent with the eastward tectonic transport direction observed in the Barn Mountains 80 km along strike to the southeast. There, east directed folds and thrust faults were unconformably overlain by Mississippian strata, and then reactivated by Tertiary Brookian deformation (Cecile 1988).

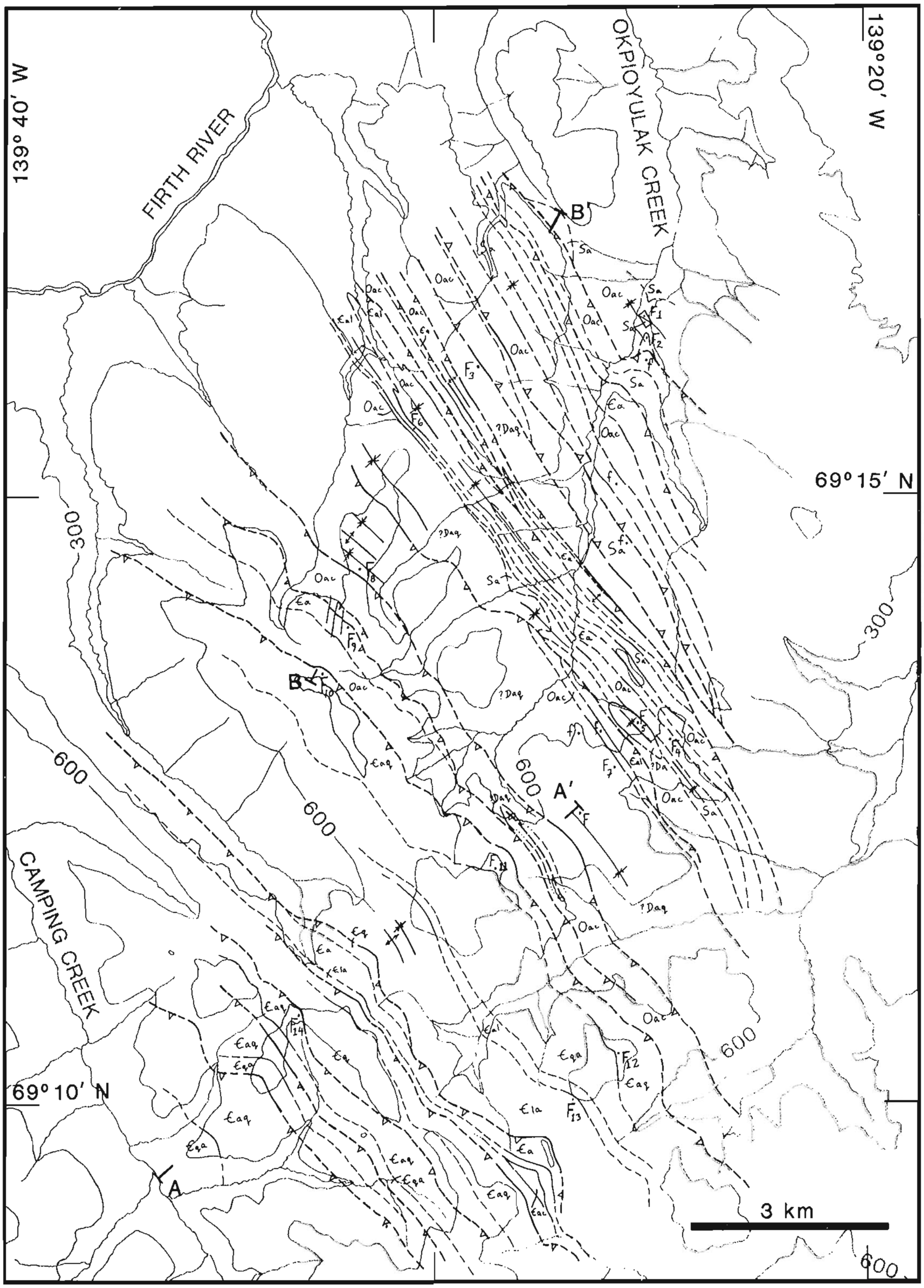
DISCUSSION

The Neruokpuk Formation was originally defined as a sequence of predominantly "quartzite schists" of pre-Carboniferous age (Leffingwell, 1919, p. 103-105). Later work resulted in an expanded definition, which included nearly all of the pre-Carboniferous rocks in northeastern Alaska; however, the discovery of lower Paleozoic fossils in several units (Dutro et al., 1972) led to the recommendation that usage be restricted to rocks correlative with the type Neruokpuk as defined by Leffingwell (Reiser et al., 1978). Assignment of the Neruokpuk to the Proterozoic was based on the interpretaion of Dutro et al. (1972) in which the quartzite unit was placed low in the stratigraphic succession.

Some of the problems of correlation between exposures in Alaska and the Yukon were discussed by Norris (1985). Although three lower Paleozoic fossil localities were excluded from the Neruokpuk (Norris, 1981, 1985, 1986) the interpretation of a Proterozoic age for the Neruokpuk in Yukon Territory apparently conflicts with interpreted Paleozoic ages of adjacent strata in Alaska (Reiser et al., 1980). Although the Canadian fossil localities, including those resulting from current work, are restricted to a 25 km wide belt along the northeast limit of exposure, three additional localities in the Demarcation Point Quadrangle, Alaska, are more broadly distributed (Reiser et al., 1980). In Canada, the numbering of units Pn0 to Pn6 does not necessarily imply stratigraphic order (Norris, 1981), and our recent work indicates that at least units Pn0 to Pn4 contain intensely imbricated repetitions of the lower Paleozoic strata reported here. However, we cannot rule out the possibility that Proterozoic rocks are exposed in the area.

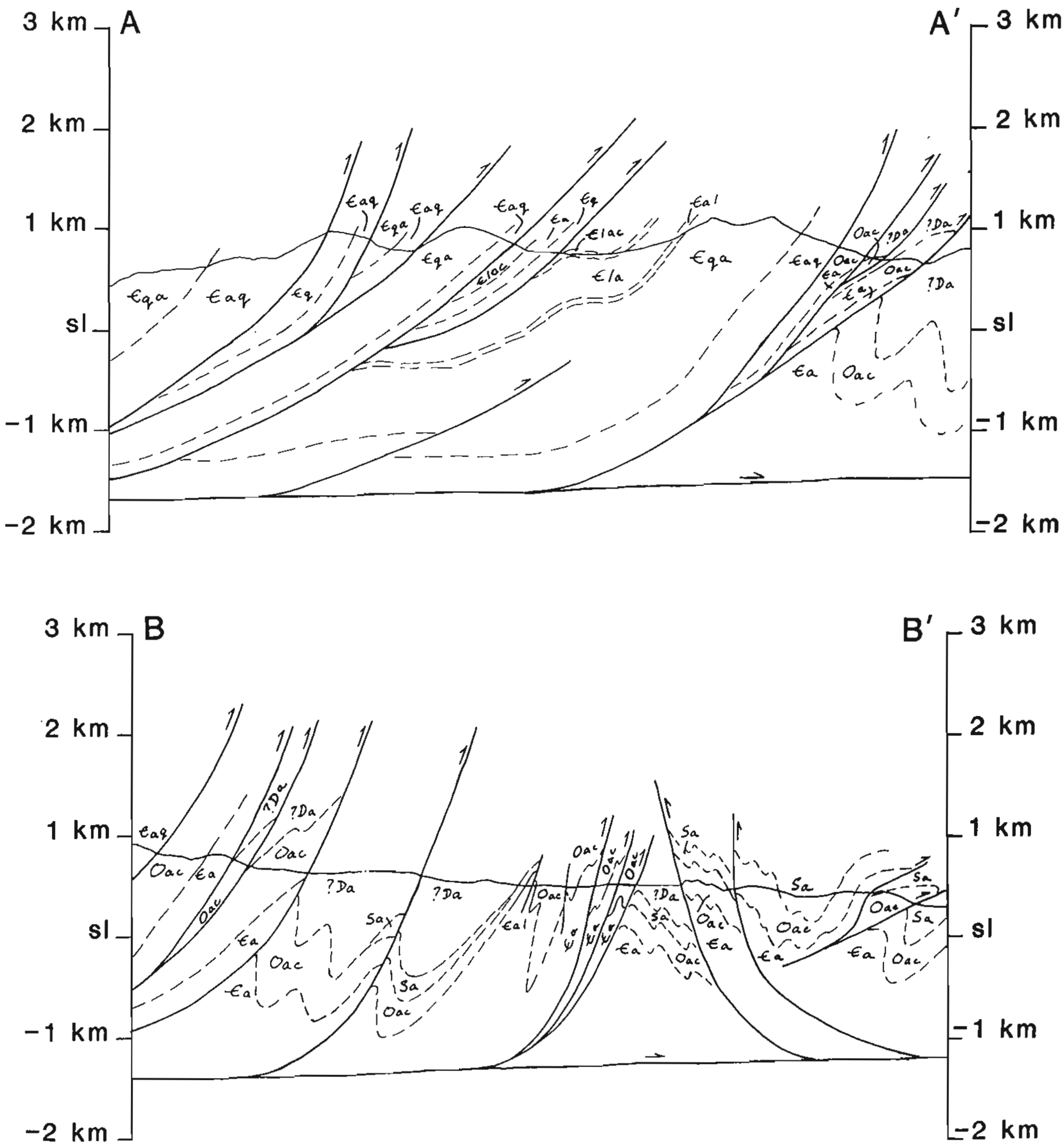
As part of a reciprocal arrangement for cross-border correlations, one of us (LSL) studied a section of Neruokpuk strata (Reiser et al., 1980) near the Egaksrak River in northeastern Alaska, as the guest of C. Hanks and W. Wallace. The strata closely resemble the quartzite-dominant "southern facies" of the fossiliferous Lower Cambrian unit mapped in the Firth River area (Fig. 3). Dutro et al. (1972) inferred a Precambrian age for the unit in Alaska, but, if the lithological correlation noted here is appropriate, the Neruokpuk (as defined in Reiser et al., 1978) may include rocks of Early Cambrian age.

The region is known to be part of the Brooks Range tectonic block, as demonstrated by stratigraphic and structural continuity. The remarkable correlation of the early to middle Paleozoic stratigraphy with that of the Selwyn Basin strengthens the hypothesis that there was a stratigraphic link between the Brooks Range block and the Cordillera (i.e., both lay along an ancestral Pacific-Arctic continental margin) in the early Paleozoic. In the Beaufort-Mackenzie region, the Ordovician-Silurian carbonate-shale facies boundary lies to the east of the study area, with an additional carbonate bank promontory extending westward across the Porcupine Platform to the south (Tipper et al., 1981). If we



2a

Figure 2. a. Sketch map of study area east of Firth River, North Yukon National Park. Contour interval is 300 m. b. Sketch cross-sections A-A',B-B' (see Figure 2a for locations). Horizontal scale = vertical scale.



2b

?Da	(?)Devonian dark grey to black argillite, interbedded with foliated micaceous quartz sandstone, quartzite, and chert-pebble conglomerate.
Sa	Silurian dark grey to green, siliceous, light- or rusty-weathering argillite.
Oac	Ordovician dark grey to black argillite interbedded with grey to black chert, locally green; upper parts weather a dark blue-grey.
€a	Cambrian argillite, pale green, buff, red or maroon, interbedded with ripple crosslaminated grey, green or maroon siltstone and minor quartzite (€aq); or locally, chert (€ac), or limestone (€al).
€l	Cambrian limestone, light to dark grey, usually argillaceous and platy, weathers brown or buff colour, interbedded with argillite (€la) and, locally, chert (€lac).
€q	Cambrian quartzite, red, grey, black or olive green, usually weathers white or pink; interbedded with grey, green or maroon argillite (€qa).

Geological contact; known, inferred.

Thrust fault; known, inferred; teeth on upper plate.

Axial surface trace of anticline, syncline.

Fossil collection locality (listed in Table 1). F_6

Fossil fragments or traces observed but not collected. f conclude that the study area was part of an early Paleozoic basin, the existence of an ancestral Arctic-Pacific ocean of early Paleozoic age in the present position of the Beaufort Sea, including at least the northeast part of the Brooks Range block, is supported. Such a paleogeography does not require large tectonic displacements of the Brooks Range block since Paleozoic time, neither does it rule them out.

Table 1. Fossil Collections (Fig. 2a)

Field no.	GSC locality no.	Fossils recovered	Age
F1	C-142239	biserial graptolite *Monograptus* 2 spp	Early Silurian (Llandovery)[1]
F2	C-142238	biserial graptolite *Monograptus* spp. *M. exiguus primulus* Boucek and Pribyl *M. turriculatus* (Barrande)	Early Silurian (Late Llandovery)[1]
F3	C-170255	bulk sample for microfossil extraction	
F4	C-142237	*Didymograptus* sp. (extensiform) *Goniograptus* sp. *Tetragraptus* sp.	Early Ordovician (Arenig)[1]
F5	C-170259	indeterminate fossil fragments[1], possible plant fragment	
F6	C-170254	*Planolites*? sp. *Oldhamia* sp	Early Cambrian[2]
F7	C-170380	bulk sample for microfossil extraction	
F8	C-154900 C-170260	indeterminate burrows[1] indeterminate trails	
F9	C-154898	*Planolites* sp *Gordia*? sp *Oldhamia radiata* Forbes	Early Cambrian[2]
F10	C-154897	*Oldhamia* sp	Early Cambrian[2]
F11	C-142236	*Oldhamia*? sp	Early Cambrian[2]
F12	C-170272	*Oldhamia radiata* Forbes *Planolites*? sp	Early Cambrian[2]
F13	C-170277	bulk sample for microfossil extraction	
F14	C-142242	*Oldhamia radiata* Forbes	Early Cambrian[2]

[1]Identified by B. S. Norford, 1988
[2]Identified by A. W. Norris, 1988

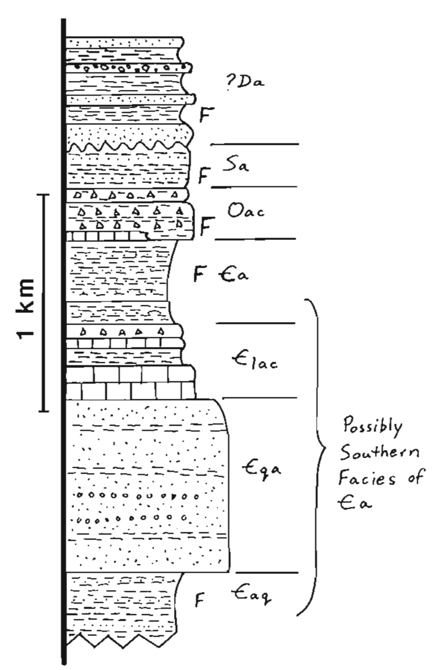


Figure 3. Preliminary stratigraphic column (see Figure 2 for legend). "Southern Facies" may be stratigraphically equivalent to Ꞓa unit exposed in the northern part of the map area. Scale bar = approximately 1 km.

SUMMARY

Preliminary results of 1:50 000 scale mapping of the Neruokpuk Formation in the Firth River area of northern Yukon indicates a Cambrian to (?) Devonian age for the Neruokpuk in that area. An implication is that much of the area mapped as Neruokpuk in Canada is underlain by rocks that are Paleozoic in age. Cross-border lithological correlation suggests that the Neruokpuk as currently defined in Alaska may include rocks of Early Cambrian age.

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Comparison of the geothermal and organic maturation gradients of the central and southwestern Beaufort-Mackenzie Basin, Yukon and Northwest Territories†

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Abstract

The geothermal regime of the Beaufort-Mackenzie Basin is being studied through the analysis of bottom-hole temperature and net rock data from 34 petroleum exploration wells in the basin. The heat flow and geothermal gradient calculations from this study have been compared with organic maturation data (vitrinite reflectance measurements) from 8 of the 34 wells.

Data from the Yukon coastal plain and central Mackenzie Delta indicate that both areas are characterized by low geothermal gradients and heat flow, and near-identical organic maturation gradients, but substantially different absolute maturation levels. The maturation level difference between the two areas is believed to be the result of erosional processes rather than intrabasinal paleogeothermal variations. In particular, maturation data and those resulting from regional geological studies indicate that 6 to 7 km of strata have been eroded from the Yukon coastal plain area.

Résumé

Le régime géothermique du bassin de la mer de Beaufort et du Mackenzie fait l'objet d'une étude basée sur l'analyse de données sur les roches et la température de fond de puits provenant de 34 puits de recherche de pétrole dans le bassin. Les calculs de flux thermique et de gradient géothermique réalisés grâce à cette étude ont été comparés à des données de maturation organique (mesures de réflectance de la vitrinite) provenant de huit des 34 puits.

Des données relevées dans la plaine côtière du Yukon et le centre du delta du Mackenzie indiquent que les deux régions sont caractérisées par des gradients géothermiques et des flux thermiques faibles, et par des gradients de maturation organique presque identiques, mais par des niveaux de maturation absolue très différents. La différence de niveau de maturation entre les deux régions serait due aux processus d'érosion plutôt qu'aux variations paléogéothermiques intrabassinales. En particulier, les données de maturation et celles découlant d'études géologiques régionales indiquant que de 6 à 7 km de couches ont été détruits par l'érosion dans la région de la plaine côtière du Yukon.

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† Contribution to Frontier Geoscience Program.

INTRODUCTION

The Beaufort-Mackenzie Basin in northwest Arctic Canada contains up to 12 km of Mesozoic-Cenozoic sedimentary strata. The geothermal regime of the basin is currently being studied through the analysis of bottom-hole temperatures and net rock data from some 34 petroleum exploration wells in the area. The geothermal gradient and heat flow calculations from the study have been compared with existing organic maturation data for 8 of the 34 wells. The purpose of this paper is to present some preliminary results and discussion on the geothermal and organic maturation gradients within the Mesozoic-Cenozoic basin fill. In particular, organic maturation variations between the central and southwestern portions of the basin are discussed in relation to geothermal gradients and basin margin tectonics.

REGIONAL GEOTHERMAL SETTING

The geothermal regime of northwestern Arctic Canada, including the Beaufort Sea - Mackenzie Delta area, has been described by Majorowicz, et al. (1988) and Jones et al. (in press a). These studies identified the Beaufort-Mackenzie Basin as an area of low geothermal gradients (20-30 mK/m) and low heat flow (40-55 mW/m^2; mean of 49 mW/m^2). These values are equal to or lower than the gradient and heat flow values for many parts of the Arctic Platform, and less than the values in the Cordilleran Foldbelt south of the Mackenzie Delta area. The Beaufort-Mackenzie Basin heat flow calculations can also be compared with heat flow data from more northerly, circum-Arctic Ocean locations (Fig. 1). Heat flow values of 54 mW/m^2 have been measured from the Prudhoe Bay area of northern Alaska (Lachenbruch et al., 1982), and low heat flow values of 40 to 50 mW/m^2 have been reported from a study of Sverdrup Basin wells (Jones et al., in press b). Heat flow values from the Canada Basin and Chukchi Plateau regions are generally higher than adjacent sedimentary basins, with values of 56-62 mW/m^2 reported by Lachenbruch and Marshall (1969) and Langseth et al. (1988). The Mendeleev and Alpha ridges have mean heat flow values of 52 and 56 mW/m^2 (Taylor et al., 1986) and the Lomonosov Ridge has heat flow values of 60 to 65 mW/m^2 (Judge, 1980).

GEOTHERMAL GRADIENTS: BEAUFORT-MACKENZIE BASIN

Calculations of geothermal gradients from corrected bottom-hole temperature data have been made for 34 wells along 4 profiles crossing different portions of the Beaufort-Mackenzie Basin (A-A' to D-D', Fig. 2). With the exception of the easternmost wells (along D-D' and C-C'), all of the calculated geothermal gradients are low (23-33 mK/m), with small variations along and between profiles. The gradients in the three Yukon coastal plain wells (Roland Bay, Spring River and Blow River) along A-A' are 24 to 26 mK/m. These values are slightly lower than the average 30 ± 3 mK/m for the central Mackenzie Delta area wells along B-B' (Ikhil to Immerk). This small difference in geothermal

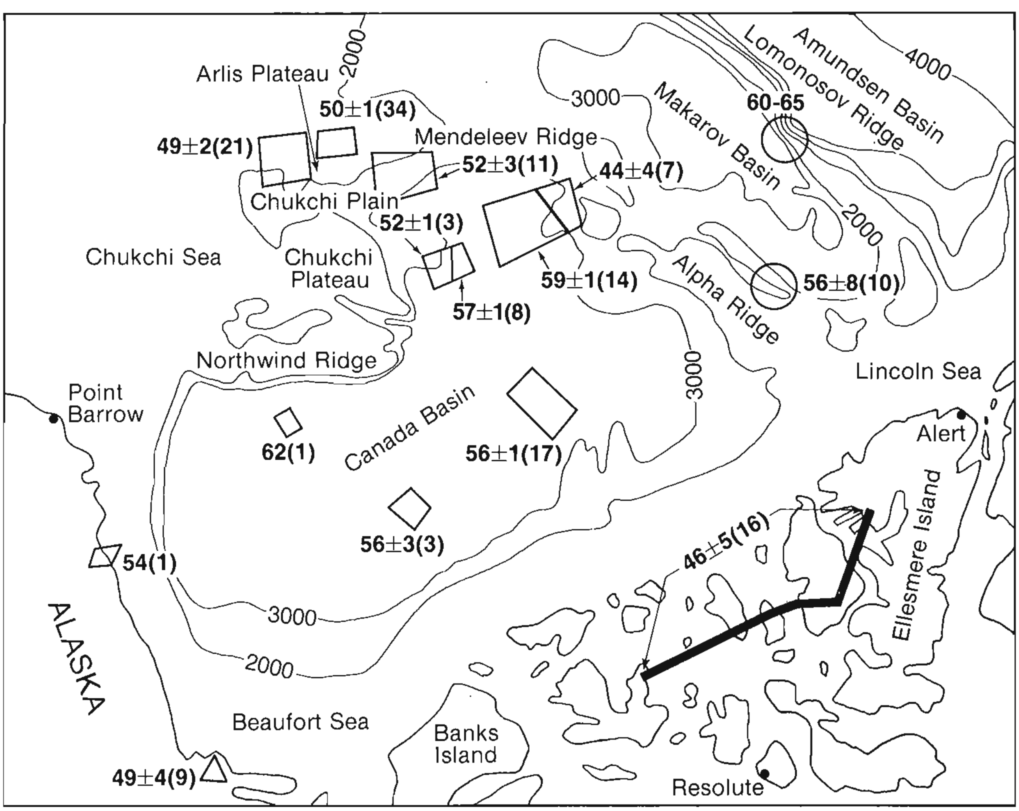

Figure 1. Summary of heat flow data for the North American sectorof the Arctic Basin. Heat flow values given are in mW/m^2 with standard error values (±); number in parenthesis represents number of individual measurements. See text for references.

gradients appears to be the result of differences in heat conductivity in the strata penetrated in the two areas. The Yukon coastal plain wells penetrated highly deformed and compacted Mesozoic-lower Cenozoic sediments with higher thermal conductivities than the more porous Mesozoic-Cenozoic sediments in the Mackenzie Delta area wells. Although detailed quantitative analyses of heat conductivity variations are still in progress, it seems likely that two areas have nearly-identical present-day heat flow characteristics.

ORGANIC MATURATION GRADIENTS

Previous studies of organic maturation data from the Beaufort-Mackenzie Basin have shown that most of the Mesozoic-Cenozoic section penetrated by wells is characterized by low levels of thermal maturation (Gunther, 1976; Powell and Snowdon, 1983; Dixon et al., 1985; Snowdon, 1987). A significant exception to this general trend is seen in the Yukon coastal plain area, where the strata contain organic matter at very high levels of thermal maturation. The organic maturation gradients in both the Yukon coastal plain and central Mackenzie Delta areas are presented as plots of vitrinite reflectance versus depth (Fig. 3). Dow (1977) observed that plots of the logarithms of reflectance (%Ro) against depth produced a straight line. This linear relationship has been confirmed in studies of the Sydney Basin (Middleton and Schmidt, 1982) and the Alberta Basin (Majorowicz and Jessop, 1981; Majorowicz et al., 1985; England and Bustin, 1986). Logarithmic plots of vitrinite reflectance versus depth for four Mackenzie Delta wells along profile B-B' (Ikhil, Reindeer, Ya Ya, Immerk) and the 3 Yukon coastal plain wells (along profile A-A') are illustrated in Figure 3. Absolute reflectance values for the Mackenzie Delta wells (Fig. 3a) increase from 0.2 % Ro at shallow depths to 0.7 % Ro at a depth of 4000 m. In contrast, the reflectance values for the Yukon coastal plain wells (Fig. 3b) are much higher, increasing from 2 % Ro at 500 m to 4 % Ro at 3000 m. Although there is a dramatic difference in absolute reflectance values (and hence maturation level) between the two areas, our calculations show that the two reflectance gradients are nearly identical at 0.13 log Ro %/km (compare 3a and 3b). This maturation gradient is comparable to other basins in Canada. The maturation gradient in the Jeanne d'Arc Basin has been reported to fall withina range of 0.07 to 0.24 log Ro %/km (Avery et al., 1986) and the gradient in the southern Alberta Basin has been calculated at 0.1 to 0.2 log Ro %/km (Majorowicz, Jones and Osadetz, unpublished data).

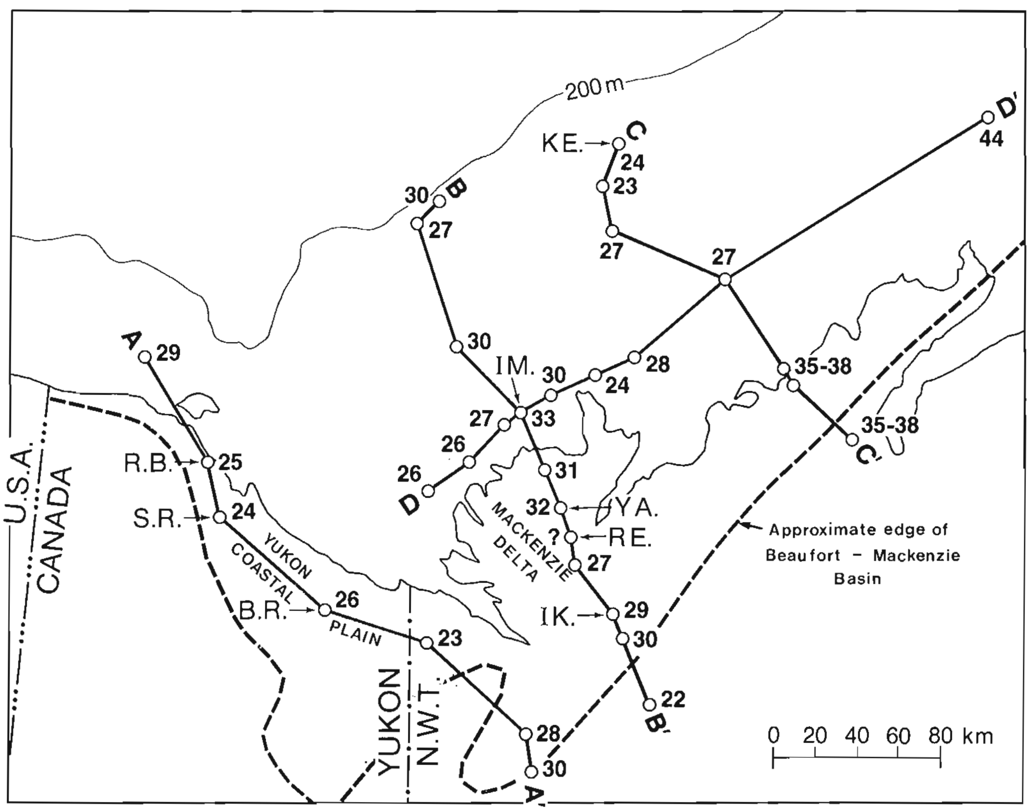

Figure 2. Geothermal gradient estimates (in mK/m = °C/km) for the Beaufort-MacKenzie Basin, using the base of frozen ground (~0°C) according to Judge (1986) and the deepest corrected bottom-hole temperatures. (Geotech, 1983). The locations and initials of the names of the wells for which vitrinite reflectance data have been used are shown and indicated by arrows: Ikhil I-37 (IK.); Immerk B-48 (IM.); Kenalooak J-94 (KE.); Reindeer D-27 (RE.); Ya-Ya P- 53 (YA.); Blow River E-47 (BR.); Spring River N-58 (SR.; Roland Bay L-41 (RB.).

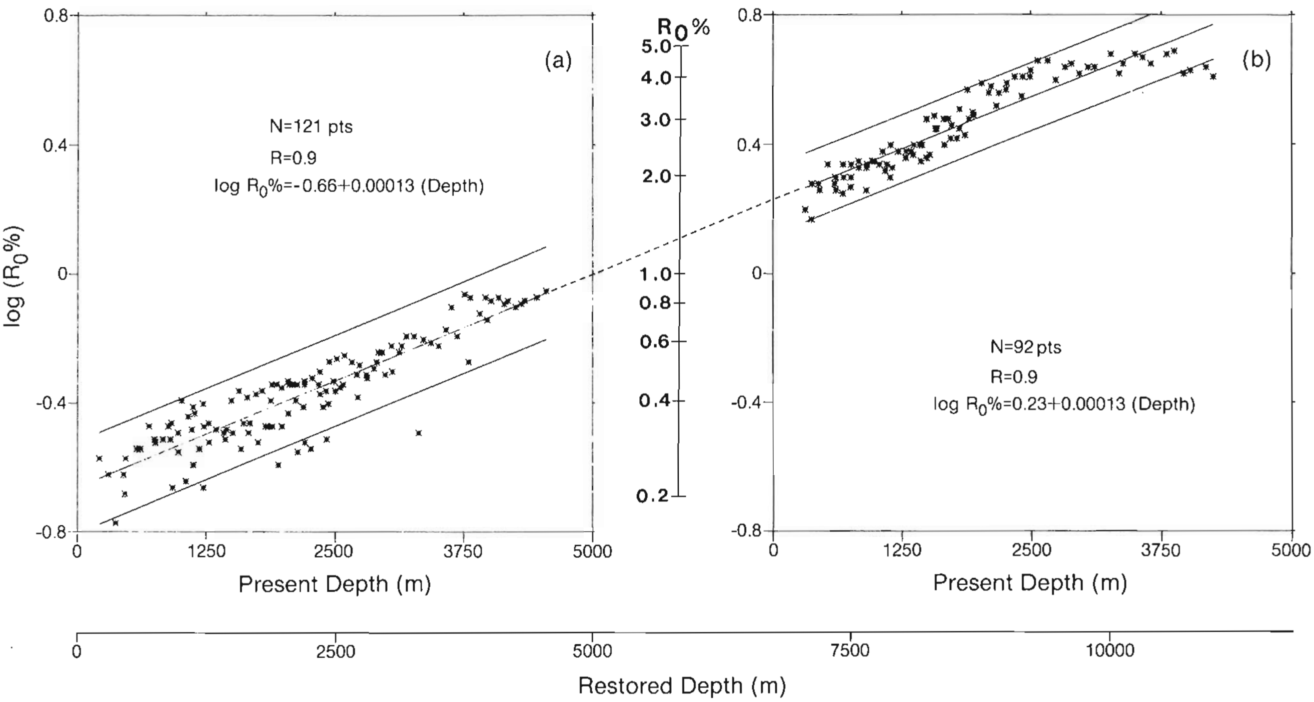

Figure 3. Plots of log Ro% versus depth and maturation profile fitted by linear regressions for two groups of wells (a) for the Mackenzie Delta (N.W.T.) (b) Yukon coastal plain; wells. Maturation and gradient for both groups of wells is 0.13 (log Ro%/km).

DISCUSSION

Our analysis and comparison of geothermal and organic maturation data for wells along profiles A-A' and B-B' indicate that the Yukon coastal plain and central Mackenzie delta areas have very similar heat flow characteristics and geothermal gradients and nearly identical organic maturation gradients. The present-day absolute maturation level of strata at similar depths is, however, dramatically different between the two areas (overmature versus immature to marginally mature). We suggest that the maturation level variations between the two areas are not the result of differences in paleo-geothermal history, but rather are the result of significant erosion of strata in the Yukon coastal plain area. Analysis of reflection seismic and well data from the Beaufort Sea shelf offshore of northern Yukon has led to the identification of two, major erosional unconformities within the Tertiary section (one of Middle Eocene age and another of Late Miocene age; Dietrich et al., 1989). It appears that most of the Yukon coastal plain erosion is the result of the Middle Eocene phase of tectonism and uplift. The amount of missing section in the Yukon coastal plain wells can be estimated by extrapolating the maturation gradient from these wells (Fig. 3b) to the gradient line associated with the lower maturation levels in the Mackenzie Delta wells (Fig. 3a). This 'restored' depth indicates that 6 to 7 km of strata have been eroded from the Yukon coastal plain area. This estimate is in general agreement with a projection of seismically mapped erosional trends from offshore to onshore.

CONCLUSIONS

Comparison of geothermal and organic maturation levels and gradients of the Mackenzie Delta and northernmost Yukon areas reveals similar geothermal regimes, but dramatically different tectonic/erosional histories. The maturation levels of the strata in the Yukon coastal plain area were "locked in" during early Tertiary periods of maximum burial, some 6 to 7 km below present depths. In contrast, the central Mackenzie Delta part of the basin has been subjected to considerably less Tertiary erosion, and present burial depths are closer to maximum paleodepths. The paleogeothermal regime of the basin (at least for Cenozoic time) may have been very similar to present-day conditions. Further study is required to evaluate the significance of more subtle, intra-basin variations in heat flow, gradients and organic maturation history.

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Preliminary report on Ordovician-Devonian conodont collections from carbonate and fine grained clastic facies of northern Yukon Territory and northwest District of Mackenzie, N.W.T.[†]

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Abstract

In 1986-88, 361 conodont samples were collected from 30 sections of the fine grained clastic (Road River Group) and generally unnamed carbonate facies of northern northwestern Canada in order to study conodont evolution in trough and platform settings, to refine conodont biostratigraphy, and integrate the conodont and graptolite zonal schemes.

Only one third of the data has been processed, yielding over 23 076 diverse conodont elements ranging in age from Ordovician through Devonian. Ordovician conodonts range in age from Early Arenig to late Richmondian. Ordovician faunas from Blackstone River, Prongs Creek and Churchward Anticline are of particular interest. Some of the faunas and zones include: Paracordylodus *fauna,* O. sesquipedalis *fauna,* P. serra *Zone,* G. ensifer *Zone. Silurian faunas and zones recognized at Prongs Creek, Illtyd Range, and Pat Lake tentatively include the* D. kentuckyensis, P. celloni, *and* P. amorphognathoides *zones, and the O. n. sp. A –* I. *sp. B, and* O.? fluegeli *faunas. Ludlow-Lochkovian conodonts are recognized at Illtyd Range. Beds at Royal Creek have yielded Upper Silurian – Lower Devonian conodonts.*

Résumé

Trois cent soixante et un échantillons de conodontes ont été prélevés de 1986 à 1988 dans 30 coupes du faciès de roches clastiques à grain fin (groupe de Road River) et de carbonates (en général innommé) dans la partie nord du nord-ouest du Canada. Les échantillons ont servi à étudier l'évolution des conodontes dans des milieux de dépressions et de plates-formes, à détailler la biostratigraphie des conodontes et à intégrer les répartitions zonales des conodontes et des graptolites.

Un tiers seulement des échantillons ont été traités, mais ils contenaient des conodontes variés et abondants (23 076 éléments) provenant de la période s'étendant de l'Ordovicien au Dévonien. Les conodontes de l'Ordovicien vont de l'Arenig inférieur au Richmondien supérieur. Les faunes ordoviciennes de la rivière Blackstone, du ruisseau Prongs et de l'anticlinal Churchward présentent un intérêt particulier. Parmi les faunes et zones relevées on compte: faune à Paracordylodus, *faune à* O. sesquipedalis, *zone à* P. serra, *zone à* G. ensifer. *Les faunes et les zones siluriennes identifiées au ruisseau Prongs, dans les chaînons Illtyd et au lac Pat incluraient les zones à* D. kentuckyensis, P. celloni *et* P. amorphognatoides *et les faunes à* O. *n. sp. B et* O. ? fluegeli. *Des conodontes du Ludlow et du Lochkovien sont identifiés dans les chaînons Illtyd. Les couches au ruisseau Royal ont produit des conodontes datant de la période s'étendant du Silurien supérieur au Dévonien inférieur.*

[†] Contribution to Frontier Geoscience Program.

INTRODUCTION

Conodont collections provide the data necessary for detailed taxonomic studies, which will improve the accurate dating and correlation of strata in Canadian sedimentary basins and thus contribute to the assessment of strata for hydrocarbon and mineral exploration. The strata of northwestern Canada, because of their diverse paleophysiography, have the potential for contributing to the integration of the biostratigraphic zonal schemes of two major index fossil groups – the graptolites and conodonts.

Lenz and McCracken (1982, 1988) and McCracken and Lenz (1987) studied strata from the fine grained clastic facies of northern Yukon Territory. Other workers such as Tipnis et al. (1978) and Nowlan et al. (1988c) have studied faunas from facies that are predominantly carbonate. Studies based on data from the carbonate facies provide the detailed evolutionary framework needed in refining a biostratigraphic scheme; those from the clastic facies, where conodont-bearing carbonates are relatively rare, allow integration of the conodont scheme with that of the graptolites.

A total of 361 conodont samples (Table 1) were collected in 1986-88 from 30 sections in northern Yukon Territory and northwestern District of Mackenzie (Fig. 1). One of the goals of the study was to have closely spaced, and large samples from the clastic facies. Collections made at closely spaced levels in the more nearshore carbonate facies may provide the necessary framework of conodont evolution and thus the basis for a detailed conodont biostratigraphic scheme for northwestern Canada. Large (5-6 kg) samples were taken to increase the chance of finding conodonts in environments that may have had relatively small populations, and to provide additional material for lithological and geochemical analyses. Accurate paleontological dating will eventually provide the basis for the continuing study of the geochemical evolution of western Canadian sedimentary basins (cf. Goodfellow and Jonasson, 1984; Nowlan et al., 1988a, b). So far, only samples collected in 1986 have been processed for conodonts; these have yielded a total of more than 23 076 conodont elements. Only the 1986 collections are discussed below.

Collections were made from strata representing three principal depositional environments, from a number of paleophysiographic subdivisions (Norris, 1985, Fig. 9) in northwestern Canada. The fine grained clastic facies is typically the Road River Group, comprising shale, mudstone, chert, and lesser amounts of generally dark coloured limestone. The lack of numerous carbonate beds does not allow a detailed biostratigraphic assessment of the conodonts from this facies. Graptolites provide the accuracy needed for geochemical study, and the rare conodont biohorizons can usually be linked with the graptolite biostratigraphy. The fine grained clastic facies is represented in this study by the strata at Blackstone River, Pat Lake South (Blackstone Trough), Barn Mountains, Driftwood Hills (Rapid Trough), Lower Porcupine River, and Mount Lang (Richardson Trough). Some carbonate beds within this facies are undoubtedly allochthonous, having been derived from the more shallow substrates of the carbonate "platform" facies. The carbonates of the South Illtyd Range (Royal Mountain Carbonate Platform), Bluefish and Driftwood basins in the Old Crow area (Yukon Stable Block), Vunta and Fish creeks (White Mountains Platform), Campbell Lake area (Mackenzie Platform), and possibly the carbonates of the Devonian at Little Fish Creek represent the carbonate platform facies. Graptolites are rare in this depositional environment, whereas conodonts are more common, and more readily recoverable using acid dissolution techniques. Strata representing an intercalation of these facies extremes were not encountered. Instead, the third depositional environment is one of vertical facies change. Strata at Royal and Prongs creeks (area of Royal Mountain Platform) and Churchward Anticline (Blackstone Trough area) comprise carbonates succeeded by fine grained clastics. An interval of lithic transition may be present between.

FINE GRAINED CLASTIC FACIES

Blackstone River

The Ordovician-Silurian succession of the Road River Group at Blackstone River (Fig. 1, loc. 2) is a dominantly fine grained clastic sequence with a 14 m thick unit of predominantly limestone. Only one of the 17 samples processed (from GSC loc. C-150522) yielded conodonts (439 elements). A few favositid corals, brachiopods, and a trilobite were also collected from this level 30 cm below the top of the limestone unit, which is overlain by thin "paper" shales. It was from the uppermost beds of this unit that Lenz and McCracken (1982) reported a Late Ordovician conodont fauna (BR:50.3 m), which was collected in1977. These writers assigned a graptolite fauna from the fine grained clastics 13.7 m below the conodonts to the *P. pacificus* Zone, and tentatively identified the *G. persculptus* Zone at a level 3 m above this conodont fauna. The conodont fauna collected in 1986 includes *Amorphognathus ordovicicus* Branson and Mehl, *Belodina confluens* Sweet?, *Drepanoistodus suberectus* (Branson and Mehl), *Panderodus gracilis* (Branson and Mehl), P. spp., *Paroistodus* sp. A Nowlan and McCracken (in Nowlan et al., 1988c), *Plectodina*? sp., *Protopanderodus liripipus* Kennedy et al., *Pseudobelodina*? *dispansa* (Glenister), *Pseudooneotodus* cf. *P. beckmanni* (Bischoff and Sannemann), *Scabbardella altipes* (Henningsmoen) subsp. B Orchard and, significantly, *Gamachignathus ensifer* McCracken et al., *Noixodontus girardeauensis* (Satterfield), and *Icriodella*? sp. This fauna can be compared to that of Lenz and McCracken (1982), which was recently described in detail by McCracken (1987).

The presence of several elements of *G. ensifer* and at least one element of *N. girardeauensis* confirms the unique co-occurrence of these two species. McCracken (1987, Pl. 1, fig. 31) reported a single sagittodontiform element of *Icriodella superba* Rhodes? from the original BR:50.3 m sample. The single fragment of *I.*? sp. from GSC locality C-150522 is part of a platform bearing three pairs of discrete denticles, and may be a fragment of *I. superba, I. prominens* Orchard, or I. n. sp. A Nowlan (1983). Alternatively, both the sagittodontiform and the platform fragments may be part of *Birksfeldia* or *Noixodontus* (see McCracken, 1987, for a discussion of these genera).

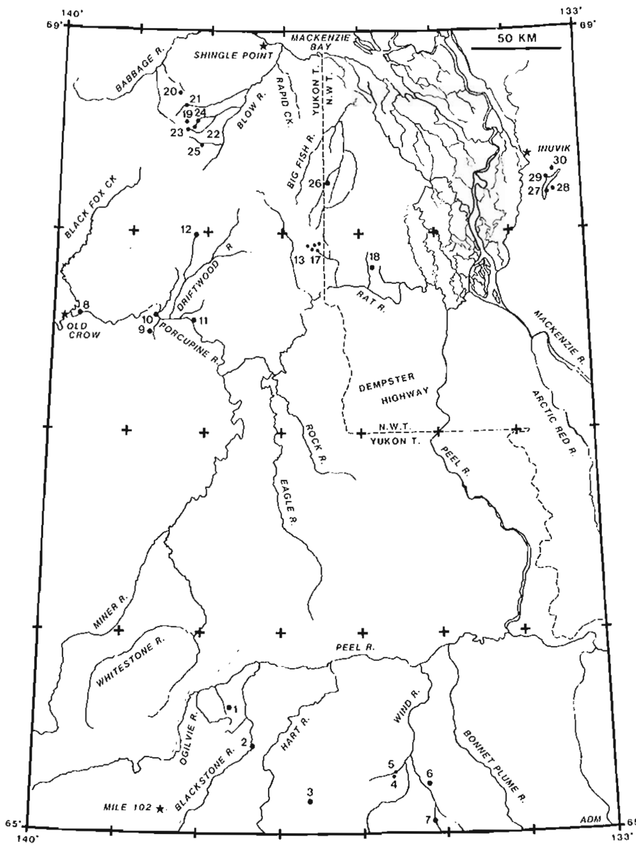

Figure 1. Localities of 1986-88 conodont collections, northern Yukon Territory and northwestern District of Mackenzie, N.W.T.

1 — Churchward Anticline: 65°36′N, 137°42′W
2 — Blackstone River: 65°26′N, 137°20′W
3 — Pat Lake South: 65°08′N, 136°38′W
4 — Upper Prongs Creek: 65°17′N, 135°42′W
5 — Lower Prongs Creek: 65°17′30″N, 135°41′10″W
6 — South Illtyd Range: 65°15′N, 135°14′W
7 — Royal Creek: 65°04′N, 135°09′W
8 — Bluefish Basin: 67°36′N, 139°37′W
9 — Driftwood Basin East I: 67°30′N, 138°42′W
10 — Driftwood Basin East II: 67°36′N, 138°39′W
11 — Lower Porcupine River (Rat Indian Creek): 67°33′N, 138°12′W
12 — Driftwood Hills: 68°58′N, 138°10′W
13 — Vunta Creek West: 67°57′N, 136°41′W
14 — Vunta Creek: 67°56′N, 136°36′W to 67°56′N, 136°34′W
15 — Fish Creek: 67°56′N, 136°33′W
16 — Fish Creek North: 67°56′N, 136°33′W
17 — Fish Creek East: 67°56′N, 136°33′W
18 — Mount Lang West: 67°51′N, 135°48′W
19 — Barn Mountains I: 68°34′N, 138°19′W
20 — Barn Mountains II: 68°43′N, 138°25′W
21 — Barn Mountains III: 68°38′N, 138°20′W
22 — Barn Mountains IV: 68°32′N, 138°13′W
23 — Barn Mountains V: 68°31′N, 138°18′W
24 — Barn Mountains VI: 68°32′N, 138°10′W
25 — Barn Mountains VII: 68°24′N, 138°07′W
26 — Little Fish Creek: 68°15′09″N, 136°25′20″W
27 — Campbell Lake Southwest: 68°12′N, 133°26′W
28 — Campbell Lake Southwest Quarry: 68°13′N, 133°25′W
29 — Campbell Lake Northwest: 68°15′N, 133°26′W
30 — Campbell Lake Northeast: 68°19′N, 133°19′W

Table 1. Distribution of samples collected from northern Yukon Territory and northwestern District of Mackenzie, N.W.T.

Location	Conodont	Macrofossil	Lithological	Geochemistry	No. of Samples
Blackstone River	20	8	2		30
Prongs Creek	44	5	2		51
Illtyd Range	20	5	1		26
Pat Lake	19	11			30
Royal Creek	31	2		10	43
Churchward Anticline	26	3	3		32
Little Fish Creek	28	9			37
White Mountains, Mount Lang	116	18			134
Barn Mountains	23			38	61
Campbell Lake	23	7			30
Driftwood Hills, Porcupine River	11	3		7	21
Total	**361**	**71**	**8**	**55**	**495**

The G. ensifer Zone was used by McCracken and Lenz (1987) and McCracken (1987) to identify the Upper Ordovician conodont fauna at Blackstone River (the term was formally introduced as a first appearance zone by McCracken and Nowlan, 1988). As noted by Lenz and McCracken (1982), their uppermost Ordovician conodont fauna is equivalent to the upper part (i.e. late Richmondian) of the range of the *A. ordovicicus* Zone, and these authors also noted that it probably was not Gamachian because of the presence of species of *Plectodina* (see McCracken and Nowlan, 1988 for details of conodont ranges in the *G. ensifer* Zone). Thus, the limestone interval on Blackstone River is regarded as Ashgill and late Richmondian in age, based on both the graptolites and conodonts. The recently announced decision on the position of the base of the Silurian (Cocks, 1985) does not affect the conodont age determinations, but of course means that the overlying, tentative *G. persculptus* Zone is now latest Ordovician, not earliest Silurian in age.

Pat Lake South

Upper Ordovician-Lower Silurian strata north of Pat Lake are dominantly recessively weathered, fine grained clastics. This succession is broken by a few metres of macrofossil rich, light weathering and very finely crystalline limestone, which forms a prominent scarp in the rounded hills of Road River Group clastics. The limestone interval measures about 8 and 10.5 m at two nearby exposures at Pat Lake South (Fig. 1, loc. 3), and about 8.5 m at the Pat Lake North Section 9 of Norford (1964), about 8 km to the north. In their report on a Pat Lake section about 3 km to the west of Pat Lake South, Lenz and McCracken (1982) identified the Upper Ordovician *P. pacificus* Zone in the clastics beneath a 9 to 10 m thick limestone unit, and (using the traditional systemic boundary definition) the Llandovery *G. persculptus* (tentative) and *P. acuminatus* zones in the clastics above the limestones. Both the graptolitic and lithic succession in the Pat Lake area seem comparable to those of Blackstone River about 45 km to the northwest. As for the strata at Blackstone River, the base of the ("traditional") Silurian could be at the base, top or within the carbonate interval. Conodonts from the limestone unit at Blackstone River are well known Upper Ordovician species. Unfortunately, in the Pat Lake area the conodonts are not as indicative of the limestones' systemic position.

The Pat Lake Section of Lenz and McCracken (1982, p. 1318) yielded poorly preserved (corroded, detritus-covered) conodonts that are for the most part undiagnostic: *Dapsilodus obliquicostatus* (Branson and Mehl), *Panderodus gracilis*, and *Walliserodus curvatus* (Branson and Branson). All three Silurian species are very similar morphologically to Upper Ordovician species. Also present is a species of *Ozarkodina* (O. n. sp. A McCracken and Lenz, 1987), a genus that is rare in the Ordovician but common in the Silurian. McCracken and Lenz (1987) assigned these conodonts to an informal *Ozarkodina* n. sp. A – *Icriodella* sp. B association. This new species of *Ozarkodina* has a Silurian aspect but must be regarded as Ordovician because of the recent systemic boundary decision (see McCracken and Lenz, 1987, p. 649, 650 for more discussion on this species and the boundary).

The new conodont collections from Pat Lake South are better preserved than those illustrated by Lenz and McCracken (1982) from their section to the west. Their locality was in the proximity of a major fault and this may have affected the conodonts. Only 8 of 19 conodont samples from Pat Lake South have been processed; these have yielded 262 conodont elements (GSC loc. C-150632-150636 have not been processed – they represent a nearby exposure of the limestone interval). In the eight metres of limestones (GSC locs. C-150613-150623; sample levels from 0.7-7.9 m) *Dapsilodus* is rare; the coniform component of the fauna is instead represented by *Panderodus* and *Walliserodus*. *Ozarkodina* n. sp. A is present throughout the limestone interval, and two samples contain robust hyaline elements suggestive of *Distomodus* or *Icriodella*. One of these is a distomodiform element (GSC loc. C-150617; 3.5 m) and thus suggests the presence of *Distomodus*, a genus not known from Ordovician strata. A sagittodontiform element occurs at 3.7 m (GSC loc. C-150618) and could be assigned to an Ordovician *Icriodella* species, or the Silurian species *Icriodella deflecta* Aldridge, or *I. discreta* Pollock et al. There is nothing in these conodont collections to change my opinion that the fauna has a "Silurian aspect".

T.E. Bolton (GSC, Ottawa) has studied the silicified macrofauna from the acid residues of samples from 2.2 to 4.7 m (GSC locs. C-150616-150620) and 7.9 m (GSC loc. C-150623), and has identified the genera *Clathrodictyon, Conocardium, Heliolites, Palaeofavosites, Palaeophyllum, Leptaena*, and a cameratoechid brachiopod. He commented that although *Heliolites* ranges from Middle Ordovician in Australia to Middle Devonian in Europe and North America, it appears first in mid-Llandovery and reaches a peak in the late Llandovery of North America.

CARBONATE-CLASTIC FACIES SUCCESSION

Prongs Creek

Norford (1964) recorded over 583 m (1914 ft.) of strata at Prongs Creek. The lowest 25 m (82 ft.) of these were measured on the southwest limb of a syncline on a tributary to Prongs Creek, the remainder were measured on Prongs Creek itself. Conodont samples were collected in 1986 from two major sections in the Prongs Creek area. The first, Upper Prongs Creek Section (Fig. 1, loc. 4) is on the same tributary mentioned above; the second (Lower Prongs Creek Section, Fig. 1, loc. 5) is at the Upper Silurian brachiopod beds discussed by Lenz (1970). Samples from the former have yielded at least 16 235 conodont elements (a number of abundant samples are only partially picked). The tributary flows through a narrow gorge formed of unnamed, resistant, light coloured limestones assigned to map unit CDb (Norris, 1982). The upper part of these were collected for conodonts, as were the overlying dark argillaceous limestones that Norford (1964) assigned to his Unit 1. Beds of Norford's Unit 1 comprise aphanitic to finely crystalline, greyish black limestones and minor shales, and contain a Lower Silurian brachiopod and trilobite fauna (Raasch et al., 1961). These are transitional between the lighter coloured CDb limestones and the darker coloured fine grained clastics of the Road River Group. This lithological change presumably represents a change from platformal to more offshore deposition.

Samples were collected from sixty-five metres of platformal limestone and 1 m of transitional limestone in the gorge of this tributary (GSC locs. C-150532-150551; sample levels 1.5-66 m); and, just upstream at another exposure, the transitional limestone (GSC locs. C-150552-150559; 0.0-3.7 m), and succeeding fine grained clastics and limestone interbeds of the Road River Group were sampled (GSC locs. C-150560-150574; 6.2-20.0 m). Within the Road River Group of this latter exposure, Norford (1964) identified the *M. turriculatus* Zone in collections taken between 6 to 8 m (20-26 ft.) and the *M. spiralis* Zone between 13.7 and 18.3 m (45-60 ft.). Norford (1964) recorded 5.1 m (18 ft.) of his Unit 1 (transitional unit herein), and 19.5 m (64 ft.) of his Unit 2 (the Road River Group). My estimates for these two units are 3.7 and 19.3 m, respectively. The slight discrepancy in measurement of the Road River Group at this locality means a possible error of at least 20 cm between Norford's and my data. Based on the height above the base of the Road River Group, I estimate that in my section, Norford's graptolite intervals would be approximately at 69.8 to 71.8 m for the *M. turriculatus* Zone, and 77.3 to 81.9 m for the *M. spiralis* Zone.

Ordovician limestones contain a fauna characterized by species of *Aphelognathus* and *Juanognathus*, and *Pseudobelodina* cf. *P. adentata* Sweet. Forms similar to the conodonts from these beds (GSC locs. C-150532-150539; 1.5-33.5 m) are found in the transitional Whittaker Formation at Avalanche Lake, N.W.T. *Pseudobelodina adentata* in the Western Midcontinent Province occurs in the mid-Maysvillian to Richmondian (Sweet, 1979), and thus indicates an age range for this interval.

The beds between 46.0 m and 61.0 m (GSC locs. C-150540-150545) contain an undiagnostic collection of *Panderodus* elements and can only be interpreted as Late Ordovician to Early Silurian in age. The Llandovery *Oulodus? fluegeli* (Walliser) fauna occurs first in the platformal limestones at 64.5 m (GSC loc. C-150546). The top of the platformal limestone is at the 65 m level; detailed sampling of the beds immediately below (GSC loc. C-150547; 65 m) and above (GSC loc. C-150548; 65.05-65.15 m) the base of the transitional unit produced *O?. fluegeli* faunas. Immediately above are beds bearing trilobites and brachiopods of the types described by Raasch et al. (1961). It is interesting to note that, in the three samples from the 65 cm of strata contiguous with this lithofacies change, *Panderodus* and *O?. fluegeli* are the major faunal components from the platformal carbonate bed at 64.5 m (GSC loc. C-150546), and *Dapsilodus obliquicostatus*, and *Decoriconus* sp. are more common in the fauna from the last platformal and first transitional beds (GSC locs. C-150547-150548). The lithological change may reflect a relative increase in water depth and, if so, supports the interpretation for a more offshore habitat for *D. obliquicostatus* (Aldridge and Mabillard, 1981; Barrick, 1983), and possibly for *Decoriconus* sp. as well, if not also a pelagic mode of life for these conodonts.

Distomodus staurognathoides (Walliser) may be represented by two platform fragments at 64.5 m (GSC loc. C-150546), and is present at 65 m (GSC loc. C-150547). Two samples from levels 65 m through 65.15 m (GSC locs. C-150547, 150548) contain ozarkodiniform elements of what may be *Pterospathodus amorphognathoides* Walliser? *s.l.*, or a closely related form (diagnostic platform element not found). Just upstream on this tributary, *Oulodus? fluegeli* faunas are found at the nearby exposure of transitional limestone and the overlying *Monograptus*-bearing fine grained clastics of the Road River Group. Here, the conodont faunas include *O?. fluegeli, Dapsilodus obliquicostatus, Walliserodus curvatus* and rare *Panderodus* sp. (a fused cluster of several *Panderodus* elements was found at 67.6 m; GSC loc. C-150557) and *Ozarkodina* sp. *Oulodus? fluegeli* occurs throughout the Llandovery part of the Prongs Creek strata, but new elements are introduced at higher levels. For example, *Astropentagnathus irregularis* Mostler first occurs at 72.7 m (GSC loc. C-150561), *Carniodus carnulus* Walliser, appears first at 77.8 m (GSC loc. C-150567), *Aulacognathus nelsoni* Over and Chatterton? at 78.5 m (GSC loc. C-150569), and rare *Pterospathodus a. amorphognathoides?*, *P. pennatus angulatus* (Walliser) and *A. nelsoni?* are found at 80.4 m (GSC loc. C-150571). Higher in the section, the fauna of GSC loc. C-150574 (85.0 m) is diverse and includes *Apsidognathus tuberculatus* Walliser, *C. carnulus, D. staurognathoides, P.* cf. *P. a. rhodesi* Savage, and *P. p. procerus* (Walliser).

The faunas from the uppermost platform beds at 64.5 to 65 m (GSC locs. C-150546, C-150547), which contain *Distomodus staurognathoides*, are probably not as old as the middle Llandovery *D. staurognathoides* Zone because they also contain *Oulodus? fluegeli*, a species that occurs in the *P. celloni* and *P. amorphognathoides* zones (upper Llandovery-lower Wenlock) of the Carnic Alps conodont succession. Secondly, it must be noted that *D. staurognathoides* ranges upward through both the late Llandovery *P. celloni* and *P. amorphognathoides* zones, and possibly higher (Barrick and Klapper, 1976).

If *Pterospathodus a. amorphognathoides* is indeed present in the uppermost platform bed and lowest transitional bed at 65 to 65.15 m (GSC locs. C-150547, C-150548) then its occurrence indicates a late Llandovery (C_6 Subdivision) to early Wenlock age range for these beds, and those of the succeeding trilobite and brachiopod fauna – this age would be anomalously young with respect to the graptolites and conodonts from higher levels (see below).

The occurrence of *Astropentagnathus irregularis* at 72.7 m suggests assignment of this level to the *P. celloni* Zone. This zone may extend upward to include the bed at 78.0 m, but it must be noted that *Carniodus carnulus*, which is found at 77.8 m, is not too diagnostic of age since it occurs in both the *P. celloni* and the succeeding *P. amorphognathoides* zones. *Aulacognathus nelsoni?*, found at levels 78.5 m and 80.4 m is a poorly known species; Over and Chatterton (1987) recorded it from the middle *P. amorphognathoides* Zone at Avalanche Lake, N.W.T. The occurrence at 80.4 m of *Pterospathodus pennatus angulatus*, if correctly identified, is puzzling since it is found in the lower

P. celloni Zone of the Carnic Alps. The other species from this level, *P. a. amorphognathoides*? and *A. nelsoni*? indicate the upper Llandovery-lower Wenlock *P. amorphognathoides* Zone. At 85.0 m, the species are indicative of this zone (especially *Pterospathodus pennatus procerus* and *P. amorphognathoides rhodesi*), although some (*A. tuberculatus, C. carnulus, D. staurognathoides*) are also found in the older *P. celloni* Zone. In spite of some anomalies, the conodont evidence suggests that the *P. celloni* Zone extends from at least 64.5 to 72.7 m, and possibly to 78.0 m, and the tentative *P. amorphognathoides* Zone (upper Llandovery-lower Wenlock) is found at 80.4 to 85 m, and possibly as low as 78.5 m.

If most of these tentative identifications withstand additional study, they present a new refinement in the integration of the conodont and graptolite zonal schemes. Unpublished data from northern Yukon show that, at Blackstone River, conodonts from a level that is within the *M. turriculatus* Zone (base of the middle, or Telychian Stage, *sensu* Cocks, 1985) and 5.1 m below the *M. spiralis* Zone, represent the *P. celloni* Zone. The tentative *P. amorphognathoides* Zone at Tetlit Creek (66°44′N, 135°47′W) is 20 m above the highest occurrence of the *M. spiralis* Zone (i.e., above interval barren of diagnostic forms) and just 1 m below the highest Llandovery *C. sakmaricus*–*C. laqueus* Zone (unpublished data; graptolite identifications by A.C. Lenz, London).

At Prongs Creek, the *P. celloni* Zone faunas are below the known occurrence of, and within the *M. turriculatus* Zone (in my estimation between 69.8 and 71.8 m). This conodont fauna extends above the upper known range of this zone at Prongs Creek, and possibly up to 78.0 m. This is within Norford's (1964) *M. spiralis* Zone based on my estimate of 77.3 m for the lower limit of the zone (bearing in mind that he measured 19.5 m, while I recorded 19.3 m). That is, the upper limit of the *P. celloni* Zone may extend into the *M. spiralis* Zone. The tentative upper Llandovery - lower Wenlock *P. amorphognathoides* Zone may be present from the 78.0 m level, that is, within the *M. spiralis* Zone.

Most conodont samples from the Lower Prongs Creek Section have not been processed; those that have been processed have yielded 96 elements. Lenz (1970) reported late Ludlow or early Pridoli conodonts (identified by L.E. Fahraeus, St. John's) from the beds containing a brachiopod fauna.

Royal Creek

Strata sampled for conodonts at Royal Creek (Fig. 1, loc. 7) comprise an interval of unnamed limestone and dolostone (map unit CDb), followed by fine grained clastics and minor carbonates of the Road River Group. A total of 4103 conodonts have been recovered from Royal Creek samples. The carbonate unit (Unit 1 of Norford, 1964) was measured in 1986 as being over 717 m thick, but this undoubtedly includes fault-repeated strata. Norford (1964) reported 369 m (1209 ft.) of carbonates measured from above a fault within this riverside outcrop. These carbonates contain Ordovician conodonts, although those from the upper part are not diagnostic and could be either Ordovician or Silurian. The Ordovician collections generally are sparse.

The few more diagnostic conodont species (and collection locality numbers and levels) are: *Pygodus serra* (Hadding) (GSC loc. C-150644; 150.0 m), ?*Acodus* sp., ?*Oepikodus* sp., *Paracordylodus gracilis* (GSC loc. C-150646; 205.0 m), *Phragmodus undatus* Branson and Mehl (GSC loc. C-150654; 632.0 m), and *Juanognathus*? sp., *P. undatus*, and *Pseudobelodina adentata* Sweet? (GSC loc. C-150656; 656 m). *Pygodus serra* is the nominal species of a Llanvirn zone, and its stratigraphic position below the level bearing the lower Arenig *Paracordylodus gracilis* fauna must be due to faulting. *Phragmodus undatus* has a range from Trentonian to Gamachian. The single phragmodiform element at 656 m (GSC loc. C-150656) is unusually robust, but similar forms are known in unpublished collections from the Canadian Arctic Archipelago. A Late Ordovician age of mid-Maysvillian to Richmondian for this sample level is based on the range of *Pseudobelodina adentata*. The overlying 61 m of carbonate contain the undiagnostic *Panderodus* and thus could be either Upper Ordovician or Silurian. Unit 2 of Norford (1964) is 133 m (437 ft.) thick and was not sampled owing to its recessive, covered nature.

Units 3 (96 m; 315 ft.) and 4 (32 m; 105 ft.) of Norford (1964) have yielded very abundant conodont faunas of Lochkovian to Pragian age. *Belodella* sp., *Eognathodus sulcatus* Phillip, *Pandorinellina optima* (Moskalenko), and *Panderodus* sp. are common at GSC locality C-150666, which is about 30 to 33 m above the base of Unit 3 of Norford (1964), and associated with monograptid graptolites, possibly *Monograptus yukonensis* Lenz (GSC loc. C-150665). The next productive interval is regarded as Unit 4 of Norford (1964). *Eognathodus sulcatus* and *Panderodus* sp. occur in the lower two samples (GSC locs. C-150675, C-150677), these are about 68 to 75 m above the level of GSC loc. C-150666, and about 9 to 13 m above the base of Unit 4. Also found in the sample at GSC loc. C-150677 are *Belodella* sp. and *Pelekysgnathus serratus* Jentzsch. Thirteen metres above this is a fauna that includes *Belodella* sp., *Icriodus pesavis* Bischoff and Sannemann?, *Icriodus* sp., *Panderodus* sp., and *Pandorinellina optima* (GSC loc. C-150679). Klapper (1969) illustrated the non-coniform component of similar faunas from other nearby Royal Creek sections.

CHURCHWARD ANTICLINE

Nearly 500 m of strata were measured on the north limb of the Churchward Anticline (Fig. 1, loc. 1), which is west of the Blackstone River and south of the Ogilvie River. The strata comprise unnamed limestone and dolostone (map unit CDb) with a thin recessive interval of shale, mudstone and rare limestone at the base of the measured section. At the top of the section are massive limestones containing brachiopods, gastropods, crinoid debris, corals and stromatoporoids. Above this biostromal unit are beds transitional to the fine grained clastics of the Road River Group.

All 19 of the productive samples (1002 conodonts) from the dominantly carbonate sequence studied in the Churchward Anticline are Ordovician in age. The stratigraphically lower collections (from GSC loc. C-150685, spot sample; and GSC locs. C-150686-150691, sample levels 26.0-151.0 m) are indicative of the Middle Ordovician. Samples from 26.0 m and 36.0 m (from GSC locs. C-150686 and C-150687, respectively) are from felsenmeer limestone within the recessive interval at the base of the section. The fauna from 36.0 m is abundant and rich, and includes *Ansella* sp., *Drepanoistodus* sp., *Paroistodus*? sp. A, *Periodon grandis* (Ethington), *Protopanderodus robustus*? (Hadding), *P. varicostatus* (Sweet and Bergström), *Pygodus serra*, and two platform elements of the *Cahabagnathus, Eoplacognathus* or *Polyplacognathus* variety. This fauna is indicative of the *P. serra* Zone of middle to late Llanvirn age. *Periodon grandis* is found between 26.0 and 261.0 m (GSC locs. C-150686 to C-150697) at Churchward Anticline; *Protopanderodus varicostatus* ranges from the *P. serra* bed at 36.0 m to the level of 151.0 m (GSC loc. C-150691), where it occurs with *Drepanodus*? sp., *Paroistodus*? sp. A, and elements of *Panderodus*.

The next distinctive fauna is from GSC locality C-150697 (261.0 m), which contains *Belodina confluens*?, *Besselodus* sp., *Panderodus* sp., *Paroistodus?* sp. *A*, *Pseudooneotodus* cf. *P. beckmanni*, and *Strachanognathus parvus* Rhodes, in addition to *Periodon*. *Belodina confluens* is known to range from the middle Shermanian substage to the Richmondian. *Phragmodus undatus* Branson and Mehl is another noteworthy species, occurring in the sample from 311.0 m (GSC loc. C-150700). This species appears first in the Rocklandian substage (lower "Trentonian" of Sweet, 1984). *Belodina confluens* ranges upward from 261.0 m (GSC loc. C-150697) to 397.0 m (GSC loc. C-151057), where it occurs with *Ozarkodina sesquipedalis* Nowlan and McCracken (in Nowlan et al., 1988c), among other species. This highest level is probably Richmondian in age.

The topmost bed of the biostrome mentioned above is a lime mudstone at 432.0 m that yielded an unusual conodont fauna (GSC loc. C-151061). Conodonts of this fauna are found in typical Richmondian, and Ashgill, faunas in northwestern Canada. However, this fauna differs from the others in that there is no trace of ramiform species, the fauna comprising only coniform species, including *B. confluens, Besselodus* sp., *Drepanoistodus suberectus, Panderodus* spp., *Paroistodus*? sp. A, *Pseudobelodina*? *dispansa*, and *Scabbardella altipes* subsp. B, as well as a few phosphatic inarticulate brachiopod valves. Transitional carbonates immediately overlie this bed but the only sample from this unit is from 478.0 m and is barren (GSC loc. C-151062).

CARBONATE PLATFORM FACIES

South Illtyd Range

Norford (1964) measured 1092 m (3582 ft.) of resistant and massive, light coloured, fine grained limestones at this section on the southern edge of the Illtyd Range (Fig. 1, loc. 6). He described it as including pelletoid, rarer conglomeratic mudlump, and organic-fragmental lime mudstones, and finely crystalline limestones, all of which are suggestive of a relatively shallow carbonate bank. Norris (1985) assigned the upper part of the section to the Kutchin Formation.

Conodont yields from the 16 productive samples are generally low (939 elements), except for the interval of the *Oulodus*? *fluegeli* fauna. Most conodonts are Silurian in age. Samples collected near the base of the section contain common Llandovery conodonts such as *Dapsilodus obliquicostatus, Panderodus gracilis*, and *Walliserodus curvatus*. Early to middle Llandovery species such as *Oulodus*? *kentuckyensis* (Branson and Branson)? and *Ozarkodina oldhamensis* (Rexroad)? are rare. Rare elements assignable to *Distomodus*? and *Icriodella* are also present. These faunas at 0.2 and 5.0 m (GSC locs. C-150583, C-150584, respectively) may represent the lower to middle Llandovery *D. kentuckyensis* Zone. Higher in the section, the fauna of the ?*D. kentuckyensis* Zone at 160.0 m (GSC loc. C-150594) is succeeded by a fauna containing *O.*? *fluegeli* and abundant *W. curvatus* and *P. gracilis* (GSC locs. C-150596 and C-150597; 168.0 m and 192.0 m, respectively). In the more abundant collection (GSC loc. C-150596) they occur with *Ozarkodina* sp., and rare *D. obliquicostatus* and *Oulodus*? sp. The units from which these two samples were taken correspond to the interval of the *P. celloni* to *P. amorphognathoides* zones (upper Llandovery-lower Wenlock). Higher samples yielded undiagnostic collections of *Panderodus*. Two spot samples taken near the top of the mountain include *Ozarkodina* cf. *O. remscheidensis* (Zeigler)? and *Pelekysgnathus*? sp. (GSC loc. C-150607), and *O. confluens* (Branson and Mehl) (GSC loc. C-150608). The former corresponds to Unit 11 of Norford (1964), the latter is from his Unit 12. These species suggest assignment of the beds to the uppermost part of the *O. eosteinhornensis* to *I. woschmidti* zones (upper Ludlow to Lochkovian).

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Lower Paleozoic stratigraphy of the White Mountains, Yukon and Northwest Territories, and sedimentological evidence for the existence of a "White Mountains platform"[†]

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Abstract

The Ordovician to Devonian stratigraphic sequence exposed in the White Mountains of northern Yukon Territory indicates the possibility of a major change in regional paleogeography following deposition of the Ordovician to Silurian Vunta Formation. The undolomitized, high energy, carbonate grainstone deposits of the Vunta Formation are consistent with deposition on a small White Mountains Platform separated from the coeval, dolomitized, eastern carbonate shelf sequence. In contrast, the overlying unnamed Upper Silurian and the Devonian Kutchin and Ogilvie formations are very similar to their coeval counterparts on the eastern carbonate shelf. Regressive cycles in the Ogilvie Formation resemble similar cycles in the Arnica and Landry formations farther east. Unconformities in this part of the White Mountains lower Paleozoic sequence are present in both of these regions. These similarities may indicate that from Late Silurian to Devonian time, the White Mountains Platform was in reality a peninsula and part of the eastern, lower Paleozoic carbonate shelf.

Résumé

La séquence stratigraphique d'âge ordovicien à dévonien affleurant dans les monts White dans le nord du Yukon, indique que la paléogéographie régionale a pu subir un changement important après la mise en place de la formation de Vunta d'âge ordovicien à silurien. Les gisements de calcaire carbonaté à débris jointifs de la formation de Vunta, caractérisés par leur nature non dolomitisée et mis en place dans un milieu de haute énergie, concordent avec le processus de sédimentation sur un petit plateau des monts White distinct de la séquence contemporaine de plate-forme de roches carbonatées dolomitisées à l'est. Par contre, la formation innommée sus-jacente du Silurien supérieur et les formations dévoniennes de Kutchin et d'Ogilvie s'apparentent à leurs pendants contemporains sur la plate-forme de roches carbonatées à l'est. Les cycles régressifs dans la formation d'Ogilvie ressemblent à des cycles similaires dans les formations d'Arnice et de Landry plus à l'est. Dans les deux régions, il y a des discordances dans cette partie de la séquence du Paléozoïque inférieur des monts White. Ces ressemblances pourraient indiquer que, du Silurien supérieur au Dévonien, le plateau des monts White était en réalité une péninsule et faisait partie de la plateforme est de roches carbonatées du Paléozoïque inférieur.

[†] Contribution to Frontier Geoscience Program.

INTRODUCTION

This investigation of the White Mountains forms a part of a Frontier Geoscience Program (Project 212-8531) devoted to a stratigraphic analysis of lower Paleozoic strata of the northern Yukon Territory and contiguous parts of the District of Mackenzie of the Northwest Territories. More specifically, Project 212-8531 is devoted to the investigation of shelf or platform deposits in this region and to the delineation of the positions of paleo-shelf transitions to argillaceous and shaly basinal strata. The White Mountains represent the northernmost occurrence of significant lower Paleozoic platform carbonate deposition in the Yukon. This study presents some preliminary sedimentological and stratigraphic data that may be pertinent to the understanding of the relationship between the lower Paleozoic platform carbonates of the White Mountains and the main mass of shelf carbonates east and south of the White Mountains.

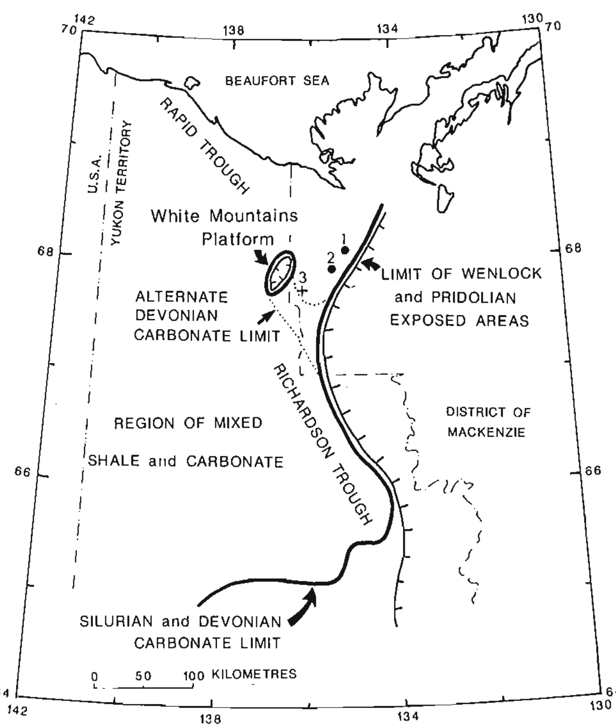


Figure 1. Location of study area in the White Mountains. Numbered localities indicate occurrences of Road River Formation shale between the White Mountains and the eastern lower Paleozoic carbonate shelf toward the craton. Locality 1 is the Union Aklavik F-38 well, locality 2 indicates the presence of Road River shale in two closely spaced wells, the Banff Rat Pass K-35 well and the Banff Treeless Creek I-51 well (see Pugh, 1983, Fig. 25a). Locality 3 is the Road River sequence exposed in Rat Uplift in the northeast part of the Bell River map area (Norris, 1981a). The configurations of the western limits of this shelf sequence and of areas of exposure on this shelf are derived from Norford (1964), Norris (1985), and unpublished information (Norford pers. comm.). There is a possibility that in Late Silurian to Devonian time the White Mountains region was part of the eastern carbonate shelf.

The White Mountains were recognized early as one of several inliers of lower Paleozoic strata surrounded by upper Paleozoic and Mesozoic strata. The White Mountain inlier (Fig. 1) is unusual with respect to the other lower Paleozoic inliers, in that it is composed of shallow water carbonate strata rather than shales of the Road River Formation, which are more typical of lower Paleozoic strata in this area (Norford, 1964). One of these Road River shale inliers, the Rat Uplift, which occupies the area between the White Mountains and the mass of lower Paleozoic shelf carbonates farther east, displays a complete Road River sequence spanning Cambrian to Devonian time (locality 3 in Figure 1). Other, smaller areas of Road River exposure occur west, north and south of the White Mountains. This isolation of the White Mountains carbonates within Road River shales has led to the conclusion that these carbonates accumulated on a carbonate platform separate from the shelf carbonates farther east (Fig. 1; Norford, 1964). The focus of this study concerns the documentation and interpretation of some sedimentological aspects of the stratigraphic units exposed in the White Mountains with regard to their inferred platformal paleogeographic setting.

STRATIGRAPHY AND SEDIMENTOLOGY - LOWER PALEOZOIC SUCCESSION

Vunta and sub-Vunta strata

Figure 2 shows a composite, columnar section of the lower Paleozoic sequence exposed in the White Mountains. This section is incomplete in its lower part because structural complications precluded accurate thickness measurements of strata underlying the Vunta Formation and the lower part of the Vunta Formation itself. The Vunta Formation was measured along Vunta Creek, which is its type section (Norford, 1964), along a line of section coincident with that shown for stratigraphic section 7b on the recent 1:250,000 geological map of the Bell River map area (Norris, 1981a). Similarly, sections of the post-Vunta strata were measured east of Vunta Creek near stratigraphic Section 7a, also on Norris's geological map.

There are two map units beneath the Vunta — the C1 and C2 map units of Norris (1981a). These were examined briefly, and 306 m of the uppermost part of map unit C1 were measured up to the contact with the overlying map unit C2 along a line of section coincident with Section 16 in the Bell River map area (Norris, 1981a). Map unit C1 is composed of a variegated recessive sequence of olive-green and lavender shale interbedded with resistant intervals of grey quartzite and orange, brecciated dolostone. An occurrence of trilobites of the genus *Olenellus* (Norford, pers. comm.), 16 m below the top of map unit C1, indicates an Early Cambrian age for the upper part of C1. The lower part of the overlying C2 carbonates is a remarkably uniform sequence of grey, peritidal dolostone that is predominantly cyclically bedded. Grey, subtidal intervals grade upward to light grey, laminated dolostone displaying fenestral fabric. Yellow-stained, brecciated beds occur in the upper part of the C2 map unit (Norford, 1964). This map unit was not measured

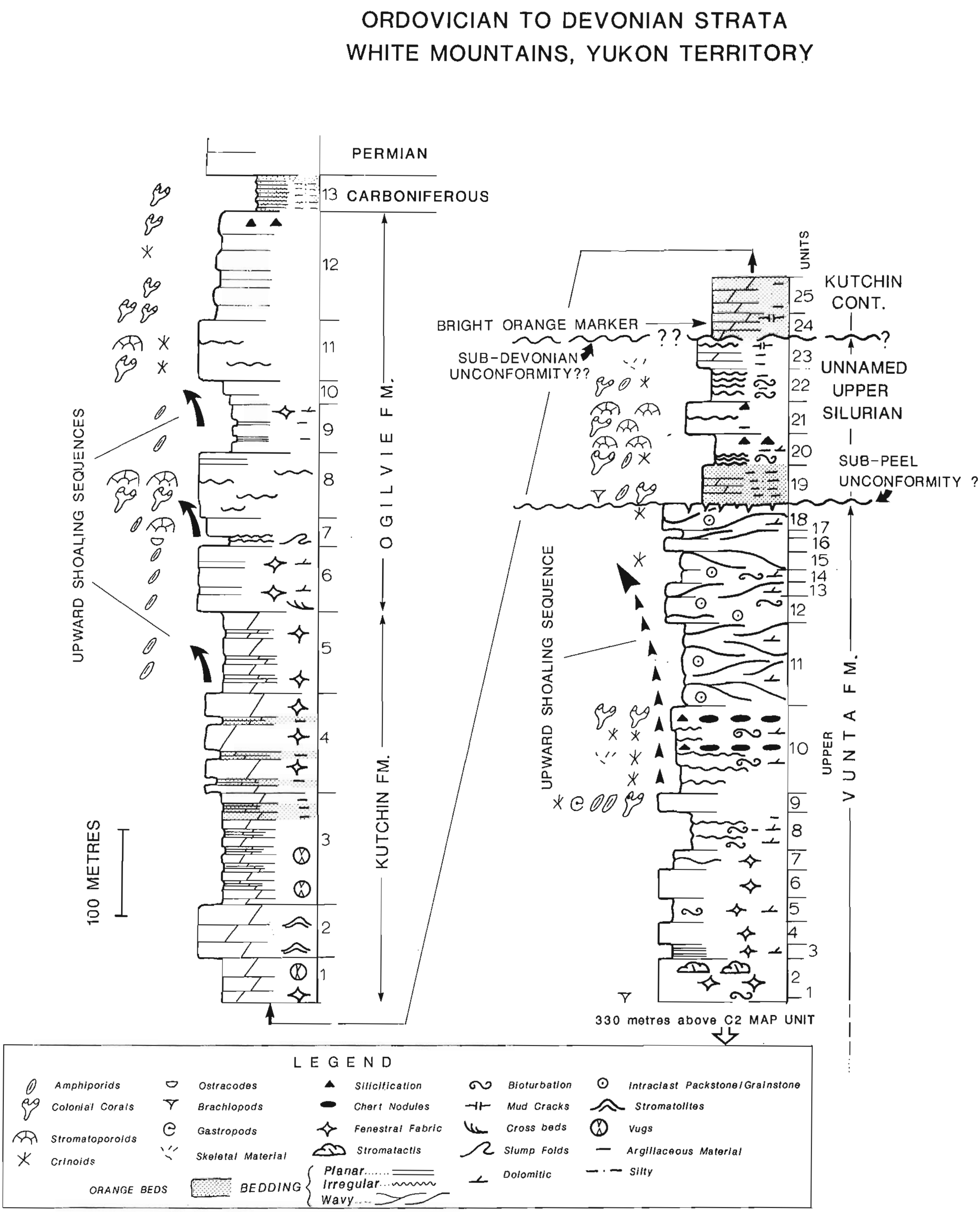


Figure 2. Composite stratigraphic sequence of the Ordovician to Devonian sequence exposed along and near Vunta Creek in the White Mountains. Two separate sections have been amalgamated. Units are numbered consecutively upsection. Relative spacing of bedding planes is indicative of bedding thicknesses.

or examined in detail in this study because of structural complications along the line of section along Vunta Creek. However, this map unit is estimated to be about 750 m ($\pm$100 m) thick as measured photogrammetrically from air photos. The upper contact of the C2 map unit with the overlying Vunta Formation occurs in a poorly exposed interval. The base of the Vunta Formation is estimated to be about 330 m below the base of unit 1 in Figure 2 in order to conform to the total thickness of 870 m (2853 ft.) assigned by Norford (1964) to the Vunta Formation in this section.

The combined middle and upper parts of the Vunta Formation, with a measured thickness of 543.5 m, comprise entirely limestone and dolomitic limestone. This Vunta sequence may be subdivided into three major depositional intervals. The lowermost interval includes units 1 to 7. This interval is characterized by calcite-cemented fenestral fabric (Fig. 2) and thick, continuous, and well developed bedding (Fig. 3A). Most of this interval is light to medium grey weathering, light grey to tan coloured pelletal wackestone and mudstone, but there are a few light greyish yellow dololaminite interbeds. Stromatactis fabric alternates intimately with coarse fenestral fabrics in beds in the lower part of this interval (Fig. 2). These lower units probably form the upper part of unit 5 of the previously measured Section 23 of Norford (1964, p.10).

The middle Vunta interval, composed of units 8 to 10, is darker grey with thick, irregular, and discontinuous bedding. Much of this interval is composed of vaguely mottled and bioturbated skeletal wackestone, particularly crinoidal wackestone. Colonial corals, such as *Cystihalysites, Favosites* and *Paleofavosites* are very abundant near the top of the interval along with scattered solitary corals. The finely comminuted skeletal wackestone matrix contains fragments of ostracodes, brachiopods and trilobites.

Silicification of fossil material, particularly of the common colonial corals, is pervasive, and several prominent zones of black, potato-shaped chert nodules occur in unit 10. This interval of the Vunta Formation is similar in lithology to unit 6 of Section 23 of Norford (1964) but also must include the uppermost beds of his unit 5.

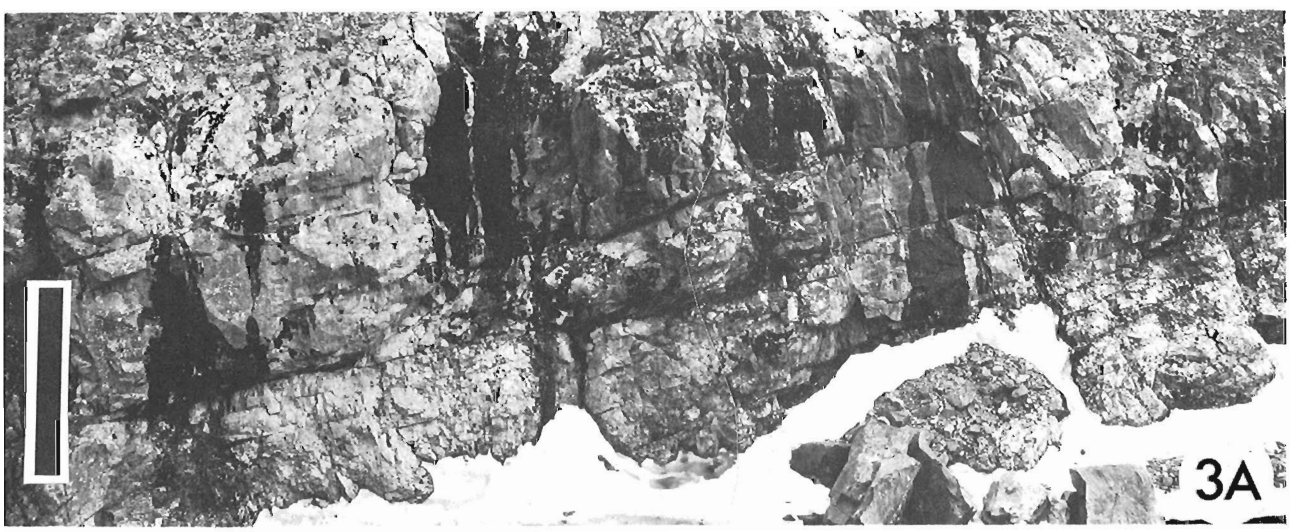


Figure 3. Outcrops of the Vunta Formation. 3A. Thick, continuous bedding characteristic of the lower part of the Vunta (Unit 3). Bar length is 1 m. 3B. Wavy bedding and sigmoidal bedding packages characteristic of the sand wave type of grainstone deposits in the upper part of the Vunta Formation. Bar length is 30 cm.

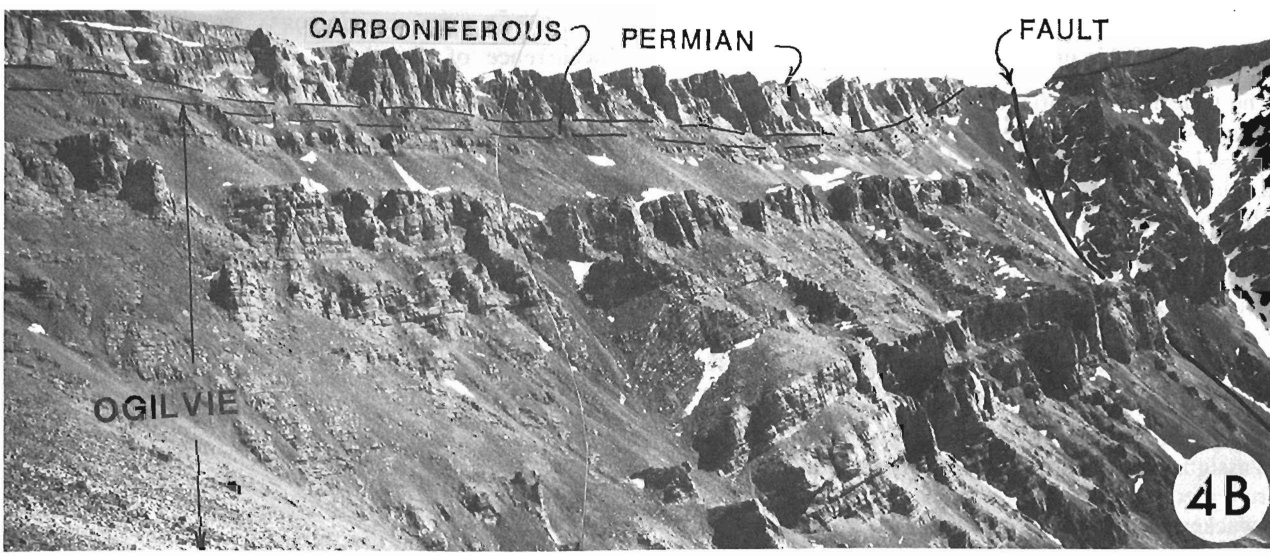
The uppermost Vunta interval extends from unit 11 to unit 18. This interval is characterized by thick, wavy to lenticular, medium to thick beds, commonly displaying discontinuous wavy partings (Figs. 2, 3B). Thicker beds tend to be defined or bounded by sigmoidal bed partings and are more resistant. Intraclast packstone and grainstone form most of this interval. Individual intraclasts appear to be formed of aggregates of pellets, and thus resemble the 'bahamite' (Purdy, 1963) of the modern Great Bahama Bank. Many beds are slightly dolomitized. This interval is largely equivalent to units 7 and 8 of Norford's (1964) Section 23.

At the top of the Vunta Formation is a sharp erosional unconformity with an irregular, fissured and corroded upper surface. Yellow, argillaceous material from the overlying unnamed Upper Silurian beds has infiltrated fissures and irregularities of the uppermost bed of the Vunta Formation.

The age of the Vunta Formation probably ranges from Middle Ordovician to mid-Silurian (Norford, 1964). This is consistent with the probable Early Silurian age assignment for the colonial coral faunas collected from unit 10 in this study (Norford, pers. comm.). The Vunta Formation itself is the approximate age equivalent of the Mount Kindle Formation developed farther east throughout the District of Mackenzie (see Norford and Macqueen, 1975). This indicates that the strata of map unit C2 underlying the Vunta Formation probably extend from upper Lower Cambrian or Middle Cambrian to the Upper Cambrian or possibly even to the Lower Ordovician, and are therefore approximately correlative with the Franklin Mountain Formation of the District of Mackenzie (Norford and Macqueen, 1975).

Devonian and Upper Silurian post-Vunta strata

Strata of the unnamed Upper Silurian (Norris, 1985) unconformably overlie the Vunta Formation and consist of 188 m of relatively recessive and fossiliferous, yellow and grey weathering argillaceous limestone and dolostone (Figs. 2, 3A). Most of this unnamed unit consists of medium to dark grey, thin to medium bedded, argillaceous, fossiliferous



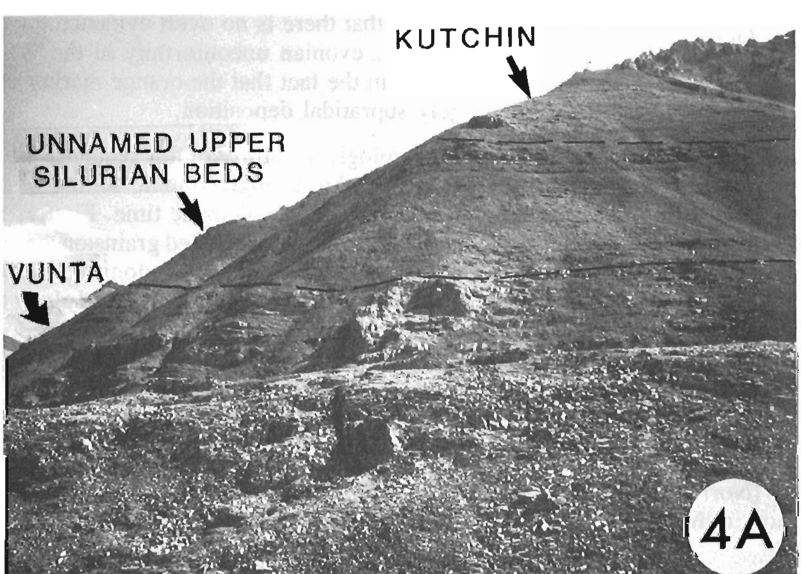


Figure 4. Views of the Upper Silurian and Devonian sequence above the Vunta Formation. 4A. Dark orange-grey weathering, argillaceous limestones of the unnamed Upper Silurian unconformably overlying the resistant Vunta carbonates. This unit and the overlying Kutchin Formation are similar to coeval sequences that outcrop in the northern Mackenzie Mountains. 4B. Almost the entire Ogilvie Formation exposed along a cirque face. This sequence displays well developed, large-scale bedding cycles of regressive sedimentation that are capped by prominent cliff-forming units.

lime mudstone and wackestone with interspersed, yellow laminated, dolostone beds. Small, ramose, thamnoporid-like corals and amphiporids are very abundant in some beds. Yellow and green argillaceous material forms partings between beds and occurs within beds as small, discontinuous partings. Unit 21, in the middle of the unnamed Upper Silurian, is a well developed, stromatoporoidal boundstone containing abundant, densely packed, hemispheroidal stromatoporoids that have been partly silicified. The upper contact of the unnamed Upper Silurian strata with the overlying Kutchin Formation occurs at a well defined break in the slope of the ridge along the line of section (Fig. 4A).

The Kutchin Formation (Norris, 1985), overlying the unnamed Upper Silurian, has been assigned a thickness of 522 m. It is a light greyish yellow weathering, thin to medium bedded, unfossiliferous, commonly laminated dolostone. Fenestral fabric is common and stromatolites occur infrequently. The basal unit of the Kutchin (unit 24) is a bright orange weathering, mudcracked and faintly laminated, medium bedded dolostone that has been inferred to overlie the 'sub-Devonian unconformity' that is well developed farther east (Pugh, 1983, p. 22), although the degree of exposure of the sequence discussed here did not permit identification of this unconformity.

The Delorme Formation, mapped by Norris (1981a) in the White Mountains, includes strata that comprise all of the unnamed Upper Silurian and probably most of the Kutchin Formation. The exact relationship between these stratigraphic units remains uncertain. The base of the unnamed Upper Silurian appears to coincide with the base of the Delorme Formation of Norris (1981a) but the top of the Delorme lies somewhere within the Kutchin Formation. The top of Unit 4 in the Kutchin Formation may coincide with the top of the Delorme Formation mapped by Norris (1981a) at this locality.

Above the Kutchin lies the Ogilvie Formation, which has been assigned a thickness of 459 m. These strata are almost entirely medium to dark grey, fossiliferous lime mudstone and wackestone. Medium bedded, dark grey, amphiporid wackestone alternates with lighter grey, slightly laminated lime mudstone in many intervals of small-scale repetitive bedding cycles. The most striking feature of this Ogilvie sequence, however, is the development of large-scale bedding cycles that are slightly more than a hundred metres thick individually (Figs. 2, 4B). These appear to be upward-shoaling sequences, in which dark grey amphiporid- and ostracode-bearing, thin to medium beds grade upward into medium grey, very resistant cliffs of thick bedded coralline and stromatoporoidal wackestone and boundstone. Argillaceous material occurs in some of the more recessive beds and crinoids occur throughout, including the two-holed crinoid *Gasterocoma bicaula*, which is abundant in unit 11 near the top of the Ogilvie (Fig. 2; Norford, 1964; Norris, 1985).

The Ogilvie Formation is overlain abruptly by 32 m of medium brown, recessive, dolomitic siltstone that have been assigned to the Carboniferous Lisburne Group (Norris, 1981a). Large, rugose, solitary corals are common in this siltstone. Ogilvie beds immediately beneath the Carboniferous siltstone are strongly silicified and are orange stained. The Carboniferous siltstone is sharply overlain by unnamed Permian limestone that forms a prominent, well defined ridge (Fig. 4B).

The thicknesses of the unnamed Upper Silurian beds (188 m) and the Kutchin Formation (522 m) recorded in this study are similar to those recorded earlier by Norris (1985) for these units. However, the thickness of the Ogilvie Formation recorded in this study (459 m) is much greater than the thickness recorded by Norris (1985) for Ogilvie strata at this locality near Vunta Creek. Structural complications, including some high angle faulting near the top of the Ogilvie Formation, may account for part of this discrepancy in measured thicknesses.

DISCUSSION

The nomenclature of the post-Vunta, lower Paleozoic strata of the White Mountains and other parts of the northern Yukon Territory has been debated vigorously (see Norris, 1985; Pugh, 1983). The nomenclature illustrated in Figure 2 is that of A.W. Norris (1985). The previously recorded occurrence of abundant atrypellid brachiopods in the unnamed Upper Silurian, indicating a Late Silurian age for these strata (Norris, 1985; Norford, 1964), may indicate that the unnamed Upper Silurian correlates partly or completely with the Upper Silurian Peel Formation of Pugh (1983), developed throughout the subsurface of the entire northern Yukon Territory and northern District of Mackenzie. This correlation implies that the unconformity at the base of the unnamed Upper Silurian is continuous with the sub-Peel unconformity (Pugh, 1983). These correlations are tentative and more definitive ages are required for these strata. To this end, collections for conodont identification were made immediately above and below the unconformity surface.

If the correlations suggested above are correct, then it is possible that the bright orange marker (Unit 24) at the base of the Kutchin Formation correlates with the Lower Devonian Tatsieta Formation of Pugh (1983). The reader is cautioned, however, that there is no overt evidence for the existence of the sub-Devonian unconformity in the White Mountains, apart from the fact that the orange marker unit represents largely supratidal deposition.

Several sedimentological features of this sequence are of interest with regard to the paleogeographic setting of the White Mountains during early Paleozoic time. Firstly, the thick upper interval of sigmoidally bedded grainstones in the Vunta Formation is evidence of deposition along well developed marine sand belts, similar to those developed along the present-day Great Bahama Banks (Ball, 1967). The predominance of such high-energy marine sand belt deposition is more consistent with the hypothesis that the White Mountains area was an isolated carbonate platform during Vunta deposition. It is less likely that such deposits would be as common in this sequence if it were really physically continuous with the correlative Mount Kindle Formation farther east. The fact that the Vunta Formation is largely undolomitized is also an argument in favour of its deposition in a region physically separated from the nearby, but totally dolomitized, Mount Kindle sequence.

Conversely, the strata above the Vunta Formation bear a closer resemblance to their counterparts farther east. The bright orange marker of this study may correlate with the orange dolostone unit that contains fish remains of possible Early Devonian age near Snake River in the northern Mackenzie Mountains (see Norris, 1985). At that locality, the orange marker overlies a thin, grey, brachiopod-bearing limestone unit, 21 m thick, that is Late Silurian in age (Norris, 1985). This limestone rests directly and unconformably on the Mount Kindle Formation. Tentatively, this limestone may be correlated with the unnamed Upper Silurian of the White Mountains. The Kutchin Formation is similar to the coeval Arnica and Sombre formations farther east. This similarity caused Pugh (1983) to extend the name Arnica Formation westward toward the Ogilvie Mountains.

The Ogilvie Formation of the White Mountains is atypical in its development of large-scale, regressive or upward-shoaling cycles (Fig. 2). Pronounced, large-scale, upward-shoaling cycles are absent in the Ogilvie Formation in the Ogilvie Mountains farther west and south. Instead, Ogilvie strata of these areas were deposited under dominantly transgressive conditions, and the upper parts of these Ogilvie sequences are largely deeper water, basin slope facies (Dubord et al., 1986). Unlike the Ogilvie Formation of the Ogilvie Mountains, the Ogilvie of the White Mountains resembles age equivalent strata developed farther east across the carbonate shelf that occupied the District of Mackenzie throughout early Paleozoic time (Fig. 1). The Hume sequence comprises three to five, well developed, large-scale regressive subsequences (Tassonyi, 1969). However, the occurrence of *Gasterocoma bicaula*, of Emsian to earliest Eifelian age (Norris, 1985), near the top of the Ogilvie probably indicates that most of this Ogilvie sequence is slightly older than Hume strata farther east. It is more likely that this sequence is correlative with strata equivalent to the Arnica Formation and to the Landry Formation of the Snake River map area (Norris, 1982). A section (measured by the author but not published) near the headwaters of Cranswick River in the Snake River map area contains an entire Silurian to Devonian sequence unconformably overlying the Mount Kindle Formation and underlying argillaceous limestone of the Hume Formation. This sequence closely resembles the Silurian-Devonian sequence above the Vunta Formation in the White Mountains. In the Cranswick River section, the strata immediately beneath the Hume Formation are composed of thin to thick bedded regressive bedding cycles with abundant fenestral fabric and amphiporids similar to the Ogilvie strata of the White Mountains.

It is evident that the post-Vunta depositional history of the White Mountains closely paralleled that of the Silurian-Devonian carbonate shelf immediately east of the White Mountains, but is unlike that of the Silurian-Devonian farther southwest in the Ogilvie Mountains. Both the White Mountains and the eastern carbonate shelf have contemporaneous unconformities, indicated as areas of Wenlock and Pridolian exposure in Figure 1. These unconformities are the sub-Devonian and sub-Peel unconformities indicated on Figure 2.

These comparisons raise the possibility that in Late Silurian to Devonian time, the White Mountains Platform was part of the eastern carbonate shelf as a peninsula-like extension rather than a separate, isolated platform. This possibility is shown on Figure 1 along with the occurrences of Road River shale between the White Mountains and the eastern carbonate shelf. These localities are separated structurally from the White Mountains by the Cache Creek Fault. There is an indication that some right lateral movement has occurred along this fault because its southern limit trends east-west and terminates in a northward dipping thrust fault; the east side of the Cache Creek Fault forms the upper or overthrust plate (Norris, 1981b). This would permit the reconstruction of a broader area than that shown in Figure 1 for an alternative Devonian carbonate limit.

CONCLUSIONS

The dominance of high energy carbonate grainstone facies and the absence of pervasive dolomitization in the Vunta Formation of the White Mountains lends support to the concept that this sequence was deposited on a relatively small and separate "White Mountains Platform". However, the similarities between the Late Silurian to Devonian sequence of the White Mountains and the Silurian-Devonian of the coeval carbonate shelf farther east may indicate that these regions were joined in Late Silurian to Devonian time (Fig. 1). This raises the possibility of the existence of Silurian-Devonian carbonate strata in the subsurface in the region between the White Mountains and the eastern lower Paleozoic carbonate shelf, which might provide an additional potential exploration target for hydrocarbon reservoirs in the northern Yukon Territory.

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An integrated analysis of the Brackett Coal Basin, Northwest Territories[†]

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Abstract

The Little Bear, East Fork and Summit Creek formations, which constitute the Brackett Basin, outcrop along the Mackenzie River and adjacent areas between latitudes 64° and 65° N. The Santonian to Campanian Little Bear Formation represents the first pulse of coarse clastic deposition into the Cretaceous seaway, which produced a coastal plain with coal-forming swamps on the western side of Brackett Basin. The overlying grey mudstones of the uppermost Campanian to lower Maastrichtian East Fork Formation signalled the re-establishment of the last fully marine phase before the final infilling of the basin by the nonmarine, coal-bearing, mid-Maastrichtian to Paleocene Summit Creek Formation - a syntectonic, largely conglomeratic sequence derived from the rising Mackenzie Mountains during Laramide tectonism.

Four informal pollen and spore zones within the Little Bear and East Fork formations, and seven others within a revised Summit Creek Formation, are recognized to support intrabasinal correlations directed toward establishing the potential coal reserves within the Brackett Basin.

Résumé

Les formations de Little Bear, East Fork et Summit Creek, qui renferment le bassin Brackett, affleurent le long du fleuve Mackenzie et des environs entre les 64e et 65e parallèles nord. La formation d'âge santonien à campanien de Little Bear constitue la première poussée de roches clastiques à gros grain dans la voie maritime du Crétacé qui, du côté ouest du bassin Brackett, a mené à l'apparition d'une plaine côtière parsemée de marais propices à la formation de charbon. Les pélites grises sus-jacentes datant du Campanien supérieur au Maastrichtien inférieur (formation d'East Fork) ont marqué la rétablissement de la dernière phase entièrement marine avant le remplissage final du bassin par la formation non marine carbonifère et de Summit Creek, au cours de la période s'étendant du Maastrichtien moyen au Paléocène : il s'agit d'une séquence syntectonique principalement congloméretique dérivée des monts Mackenzie en soulèvement durant la phase tectonique de Laramide.

Quatre zones de spores et de pollens informelles dans les formations de Little Bear et d'East Fork et sept autres dans une formation révisée de Summit Creek corroborent des corrélations voulant que le bassin Brackett recèle des réserves possibles de charbon.

[†] Contribution to Frontier Geoscience Program.
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INTRODUCTION

Brackett Basin extends from near the flanks of the Mackenzie Mountains to east of the Mackenzie River in the vicinity of Fort Norman (C in Fig. 1; Fig. 3). A comprehensive framework for the Cretaceous and Tertiary geology of the region was established by Yorath and Cook (1981). The present report summarizes results of a more detailed multidisciplinary study of Upper Cretaceous and Paleocene strata in the Brackett Basin. The complex structure imposed on this depositional basin and the facies control on coal distribution and quality, necessitated the establishment of a detailed biostratigraphic framework to elucidate otherwise obscure intrabasinal relationships. In this report, emphasis will be on the intra- and interbasinal correlation of lithostratigraphic units, and brief summaries will be given of the sedimentological and coal petrographic results.

This study is based on about 400 samples. The stratigraphic and geographic position of samples referred to in the Plates and Figures will be given in a more detailed text in preparation.

REGIONAL STRATIGRAPHY

Figure 1 shows the location of the study area (C) and the localities for which lithostratigraphic columns (Fig. 2) were prepared. Time relationships for the various lithostratigraphic units of Figure 2C, D and E are based on the palynological zonation given in this paper in combination with those published for more southern parts of the Western Interior Basin by Srivastava (1970), Norris, et al. (1975), Sweet (1978a), Sweet and Hills (1984), and Nichols and Sweet (in press). The East Fork and Bearpaw formations are shown as correlative, as are the underlying Little Bear and Judith River formations. This differs from the relative stratigraphic position shown for the East Fork Formation in Kalgutkar and Sweet (1988, Fig. 6.1) as they accepted the inclusion of a sandstone member in the upper part of the formation; in this paper, the sandstone member is placed in the overlying Summit Creek Formation.

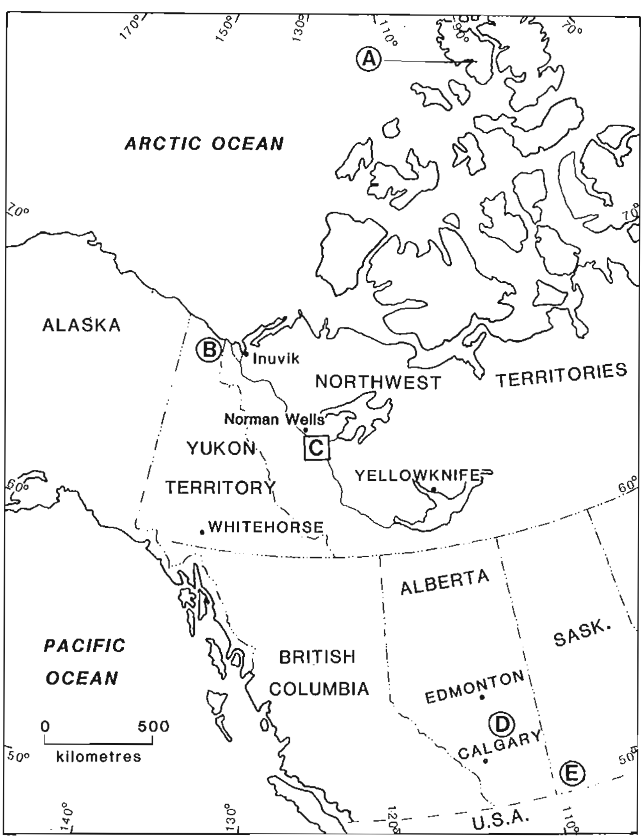


Figure 1. Regional map showing locations of lithostratigraphic columns in Figure 2.

The columns in Figure 2A and B fall within depositional complexes 8 and 9 of Dixon (1986) from northern Yukon and northwestern District of Mackenzie. There is no evidence in the Brackett Basin for the lower Paleocene marine embayment, which Dixon (1986, p. 64) inferred as probably extending into the Norman Wells area. Correlation with formations in the region of the Yukon Coastal Plain utilizes the assemblage illustrated by Sweet (1978b), and unpublished data. The palynoflora from the basal part of the Tent Island Formation (Sweet, 1978b) compares closely with the assemblages found within the late Maastrichtian, *Porosipollis porosus* and *Myrtipites scabratus/Aquilapollenites delicatus* var. *collaris* zones of this paper, confirming a late Maastrichtian age for the Tent Island Formation. The ages of the formations within the Eureka Sound Group (Arctic Islands) and the Kanguk Formation are based on the work of Ricketts (1986b).

The Summit Creek Formation as outlined in this paper (Fig. 2) spans the Maastrichtian and Paleocene, post-Bearpaw interval of Alberta and Saskatchewan. As in its Alberta counterpart, the Cretaceous-Tertiary boundary in the Police Island section occurs in association with a coal-bearing horizon. However, its exact position is within the lowest coal seam (No. 1) in the Police Island section, whereas in Alberta and Saskatchewan it occurs at or immediately below the base of a coal seam. The Summit Creek Formation spans the time interval represented by the disconformity marking the base of the Tent Island, probably the Scollard and, in places, the Frenchman formations. In the Tertiary Hills the stratigraphic position of this disconformity is probably correlative with the base of the first major conglomeratic unit, although continuity within the palynofloras suggests no major disconformity. In the Police Island section also there is no evidence of a major time break within the section; the time span of the disconformity at the base of the Scollard Formation is probably represented by part of the *Porosipollis porosus* and *Myrtipites scabratus/Aquilapollenites delicatus* var. *collaris* zones discussed below.

SENONIAN AND PALEOCENE STRATIGRAPHY AND SEDIMENTOLOGY OF THE BRACKETT BASIN

Coal occurs in both the Little Bear and Summit Creek formations. The dominantly grey mudstone of the East Fork Formation intervenes between these two primarily coarse clastic facies.

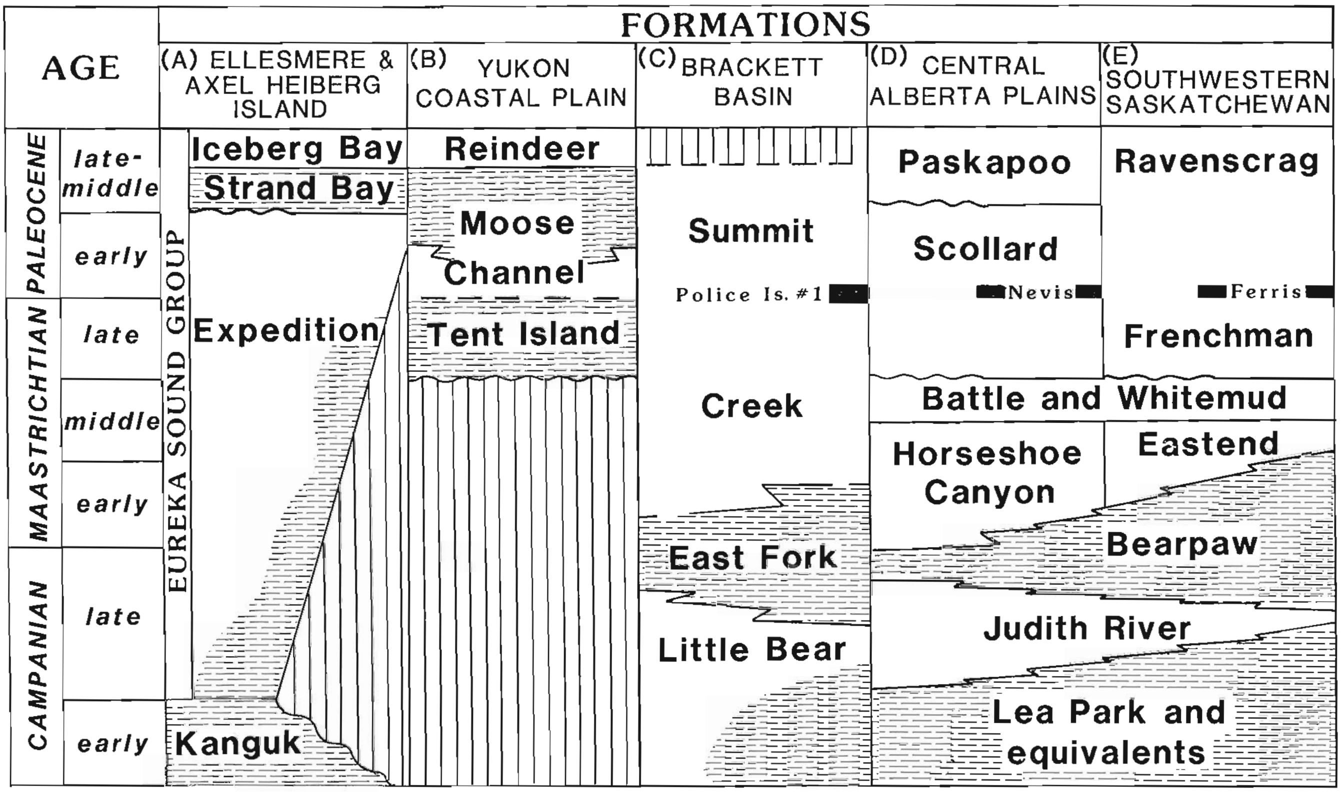


Figure 2. Table of Upper Cretaceous and lower Tertiary formations for five selected regions. Shading indicates marine strata, and vertical lines indicate missing time.

Little Bear Formation

This formation contains both nonmarine and marine facies. It is differentiated from the shales of the underlying Slater River Formation and overlying East Fork Formation in comprising mostly sandstone and siltstone. The type section is composite (Fig. 4, NC-83-35 and NC-83-36); Yorath and Cook (1981) considered that its two parts formed a stratigraphically continuous section. The two outcrops forming the type section are about 2 km apart. Approximately 230 m of strata outcrop in the composite type section, which Yorath and Cook (1981) assumed were capped by shales of the East Fork Formation. The basal Little Bear-Slater River contact is not exposed. Brideaux (1981) assigned a Campanian age to a sample from the upper part of the type section and a Santonian age to samples from its lower part.

The Little Bear Formation is at least 482 m thick in the nearest well (Candel Little Bear I-70) with a complete section (Yorath and Cook, 1981); therefore, less than half its local thickness is exposed in the type section. Given the physical separation and palynological dissimilarity of the two type section outcrops (discussed below), it would now seem probable that there is a stratigraphic separation between the upper and lower parts (as shown in Fig. 4) as opposed to the entire missing interval being covered at the base of the section. As a cautionary note, the palynological argument for the separation of the two parts is weakened by lack of biostratigraphic control for almost all the upper half of the lower part and for the lower 40 m of the upper part of the type section.

Based on outcrop information, at least two coal-bearing intervals occur within the Little Bear Formation (Fig. 4). The only known outcrop of the oldest of these is in the footwall of the Gambill Fault on the west side of Brackett Basin (NC-83-2; Figs. 3, 4). Here there is an outcrop of an approximately two metre thick seam (top not exposed) of possibly high volatile bituminous coal (Ro random = 0.52). The upper coal-bearing interval outcrops along the Little Bear River in the upper part of the type section (Fig. 4, NC-83-35) and at localities in the vicinity of Coal Creek (Fig. 4, SLA-88-2, 3) over a lateral distance of at least four kilometres. In the vicinity of Coal Creek, two coal seams occur within a 10 m thick interval bounded by sandstone. The lower seam is 0.6 m thick and occurs at the top of a 2.3 m interval of interbedded mudstone, coaly shale and coal. The upper seam maintains a thickness of about 2.0 m over the lateral extent of the outcrops. Two thin (0.3 m) seams occur at the top of the type section within a 9 m recessive interval overlying a sandstone unit.

There is an additional coal-bearing but otherwise dominantly sandstone outcrop located east of the MacKay Range (Fig. 4, NC- 83-31). At the top of the section a partially exposed, 5 m thick coal seam occurs at the base of an 11 m thick otherwise covered recessive interval bounded by sandstone. This coal is correlative with the second coal-bearing interval within the Little Bear Formation. No coal occurs in the MacKay Range outcrop in strata correlative with the lower coal-bearing interval. Yorath and Cook (1981) included this late Campanian outcrop in their discussion of the East Fork Formation. However, its age, as discussed in Yorath and Cook (1981) and re-evaluated in this paper, precludes it from being correlative with the Maastrichtian sandstones they included in their description of the East Fork Formation. The basal portion of the MacKay Range section is correlative with two exposures located near the Gambill Fault (Fig. 3, NC-83-2 and NC-83-3) and its upper part with the upper part of the type section (NC-83-35) of the Little Bear Formation. This implies that the Gambill Fault localities occur stratigraphically between the two parts of the type section (Fig. 4).

East Fork Formation

The dominant lithology of the East Fork Formation is a dark grey, marine shale as described by Yorath and Cook (1981) for the 'type locality' of the formation (Fig. 4, SLA-85-16). It is underlain by the coal-bearing beds and sandstone of the Little Bear Formation. The upper limit of this formation is open to question as, in the type area, its contact with the Summit Creek Formation is not present. In the subsurface,

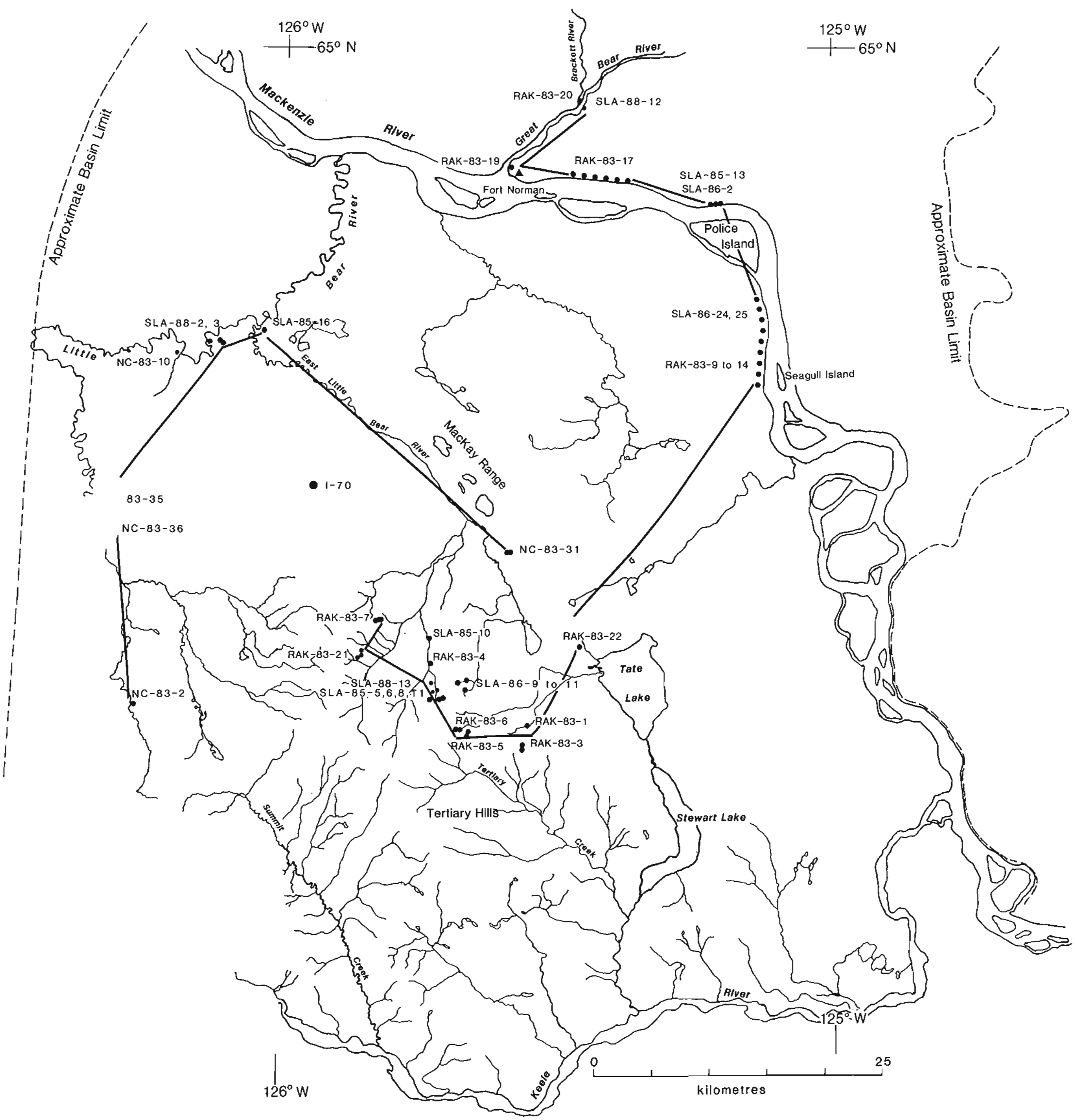


Figure 3. Detailed locality map for the Fort Norman — Tertiary Hills region. Heavy lines indicate the sections on the correlation diagrams (Figs. 4, 5).

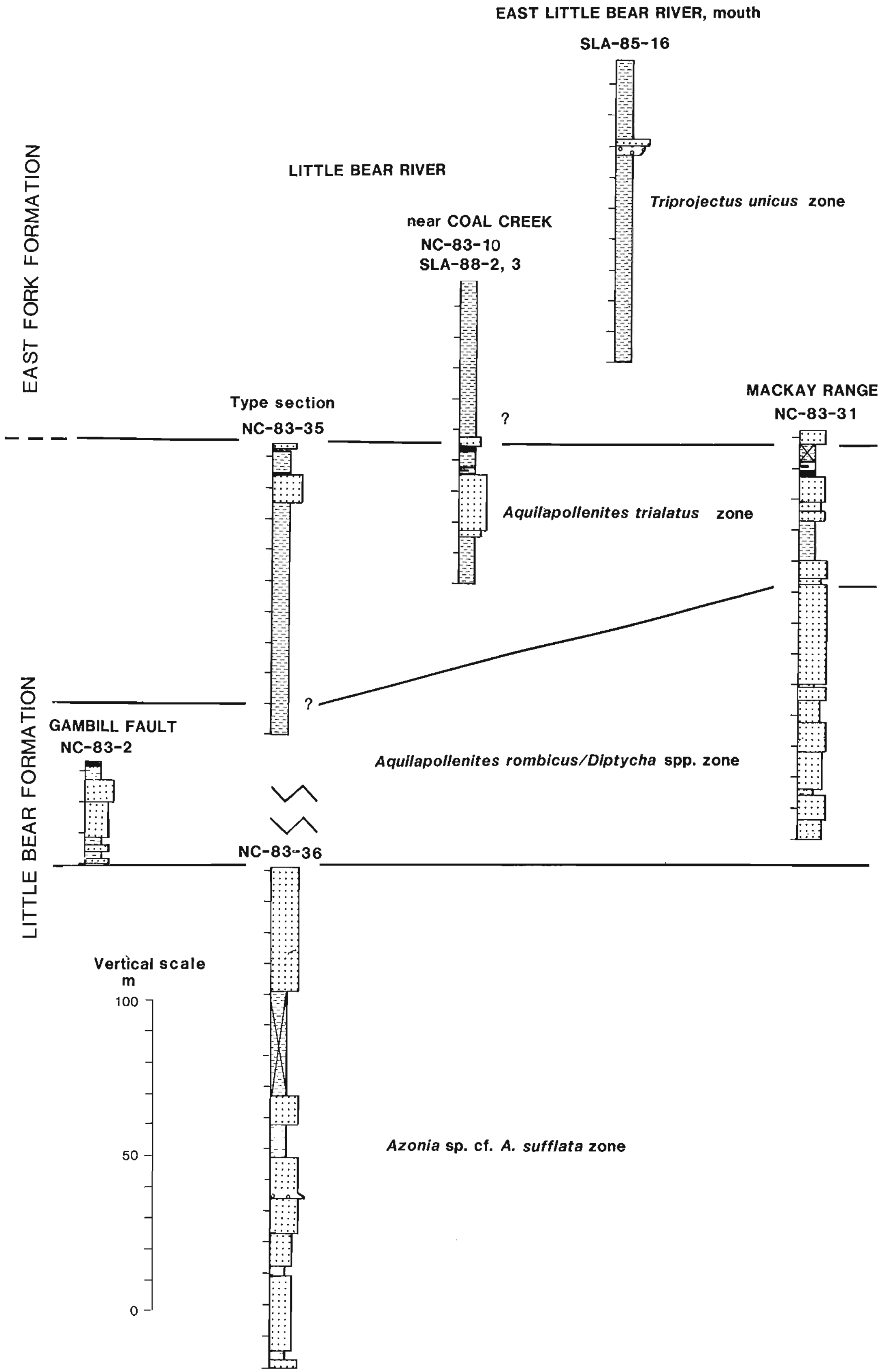


Figure 4. Correlation diagram for outcrop sections of the Little Bear and East Fork formations, based on palynostratigraphy. Zig-zag lines indicate a missing interval between sections.

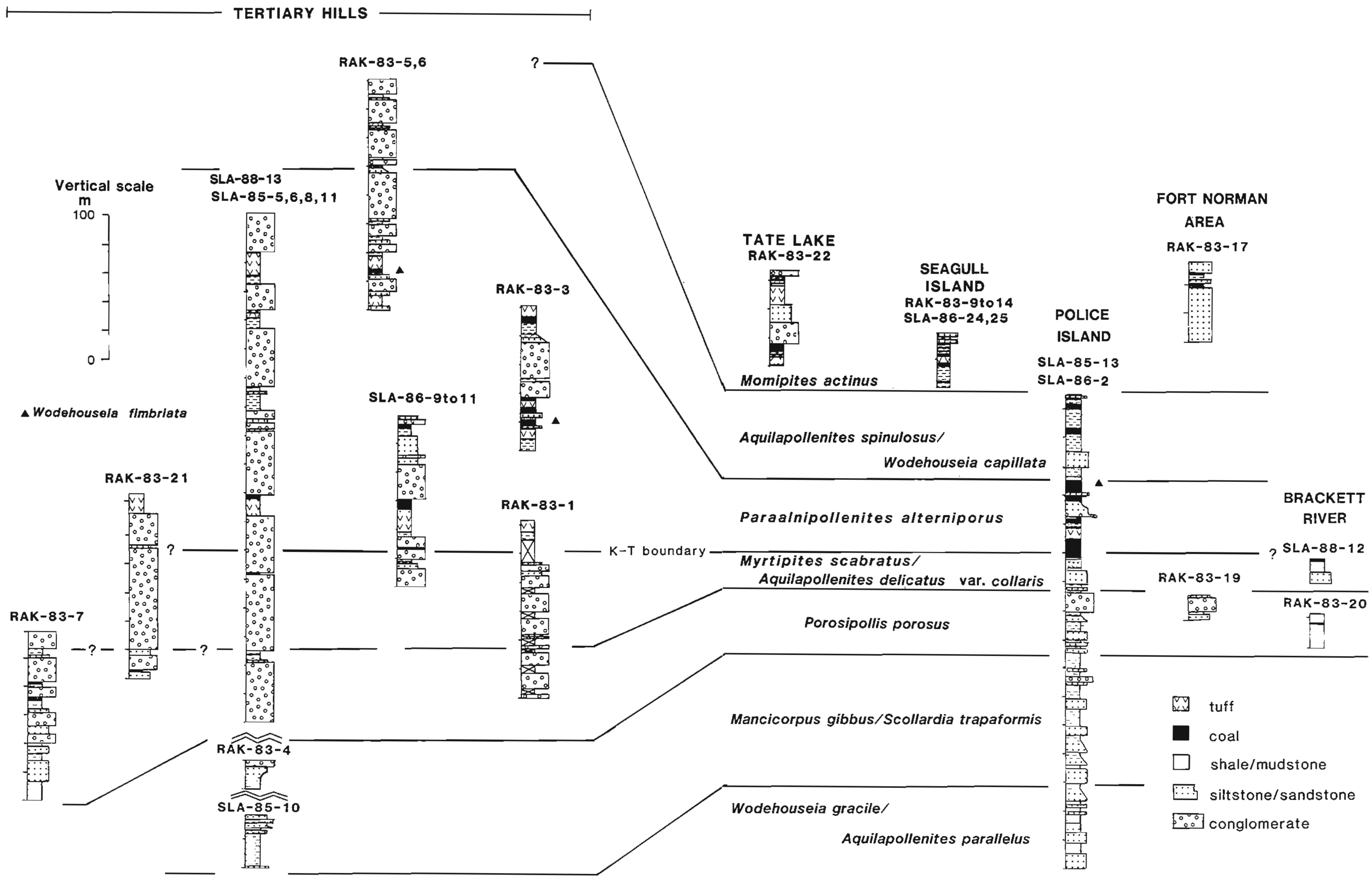


Figure 5. Correlation diagram for outcrop sections of the Summit Creek Formation, based on palynostratigraphy. Zig-zag lines indicate a missing interval between sections.

Yorath and Cook (1981) included a sandstone member in the upper part of the East Fork Formation. This member is here treated (informally) as part of the overlying Summit Creek Formation, a choice based on the mappability of the contact on the flanks of the Tertiary Hills as well as in the subsurface (K. Williams, pers. comm.). The East Fork Formation ranges in age from late Campanian to possibly early Maastrichtian (Brideaux, 1981). Within the region of the "type locality" subsurface thickness ranges from 442 to 845 m (Yorath and Cook, 1981, but excluding the sandstone member).

Summit Creek Formation

The Maastrichtian and Paleocene, nonmarine, coal bearing Summit Creek Formation has an areal extent of about 3000 sq. km (1200 sq. miles). Yorath and Cook estimated its thickness to be 488 m in the Tertiary Hills. In this paper the total thickness is projected to be in excess of 600 m in the Tertiary Hills and greater than 325 m along the Mackenzie River. A composite type section based on four outcrop localities on the east flank of the Tertiary Hills was selected by Yorath and Cook (1981). These localities are the same as sections RAK-83-1, RAK-83-3, RAK-83-5 and RAK-83-6 of this study (Fig. 5) and span approximately the upper two thirds, mostly Paleocene portion, of the formation. As noted above, Yorath and Cook (1981) included a sandstone member, separating the more conglomeratic upper part of the Summit Creek Formation from the underlying East Fork mudstones, in the East Fork Formation at subsurface localities. However, they appear to have mapped the lower contact of the Summit Creek Formation at the base of this sandstone member, hence their reference to the lower beds of the Summit Creek Formation being recessive. This latter definition of the Summit Creek Formation is followed in this paper.

In the Tertiary Hills, outcrops of the sandstone member at the base of the Summit Creek Formation occur in an upper distributary of the East Little Bear River (Figs. 3, and 5; RAK-83-4, SLA-86-10) and form the lower 180 m of the Police Island exposure (Figs. 3, 5). In the Police Island section, below the exposed base of the Summit Creek Formation and underlying a concealed interval of about 20 m, the wash line along the Mackenzie River comprises grey mud. These presumably indicate the presence of the East Fork Formation. The mudstones within the sandstone member of the Summit Creek Formation differ from those of the East Fork Formation in usually being greenish grey and, at some localities, carbonaceous. There is no conclusive palynological evidence for a direct marine influence during the deposition of any of the Summit Creek Formation, whereas marine microplankton are abundant in the underlying East Fork Formation.

In the Tertiary Hills, exposures of the more proximal deposits can be divided in ascending order into: 1) a dominantly sandstone interval transitional with the East Fork; 2) a lower conglomeratic interval with thin coal seams; 3) a middle coal and tuff bearing interval; and 4) an upper conglomeratic interval with thin coals. The more distal exposures of the Summit Creek Formation along the Mackenzie and Great Bear rivers contain a higher percentage of coal and mudstone units and fewer intervals of conglomerate. This is exemplified by the cumulative total thickness of the more than five separate coal zones in the Tertiary Hills being about 15 m, whereas in outcrops along the Mackenzie River a cumulative thickness in excess of 30 m is reached. Individual seams on the Mackenzie River are up to 12.1 m thick. Evidence of local variations in the thickness of seams within individual coal zones is provided by the direct field observation of channels cutting out coal beds and from the intrabasinal correlation of the coal zones.

The Summit Creek Formation is a stacked braidplain/meanderplain system (Ricketts, 1986a). In the Tertiary Hills, the conglomerates accumulated as transverse and longitudinal bars in moderate to low sinuosity channels with braided characteristics. Fining-upward sequences (several metres thick) developed as a result of channel aggradation, and decreasing slope and stream competence, producing an upper sandstone channel-bar facies. Floodplain deposits of coal and mudstone were of limited lateral extent. Along the Mackenzie River, the more distal facies are dominated by sandstone that was also deposited as channel and bar facies within a braided river system, although the presence of point bars in some channels suggests that sinuosity may have increased. This transition to more distal facies occurs over a paleogeographic distance of 50 to 55 km.

Time lines, based on palynological data, define a northward-thinning cone or wedge of strata that accumulated during the Early Paleocene; this geometrical configuration can logically be inferred for the entire formation.

The combination of lithofacies types, their lateral variation and the geometry of both small- and large-scale sediment bodies, suggest that the Summit Creek Formation accumulated on an alluvial braidplain/meanderplain, outboard of the presumed alluvial fan(s) (no longer preserved). In particular, a humid setting is inferred because of the presence of coal seams and evidence of flourishing vegetation. Both the palynomorph assemblage and macroflora (especially *Metasequoia*) are indicative of humid temperate conditions.

Paleocurrent indicators and lateral facies associations indicate that the general trend of progradation was approximately to the north and northeast. Alluvial plain sediments of the Summit Creek complex prograded over the last vestige of an Upper Cretaceous marine environment that is preserved as shoreface sandstones in the lower part of the formation. Clastic sediment was supplied by the rising Mackenzie Mountains during the Laramide Orogeny. Thus, Summit Creek deposition was essentially syntectonic, with sediment accumulating in the Brackett Basin foredeep east of the uplifted Cordilleran chain. The sediments were subsequently folded and faulted by a later phase of Laramide tectonism.

Seven informal palynological zones are utilized to establish the intrabasinal relationships of the following localities: Summit Creek Formation sections located in the Tertiary Hills; the Tate Lake section on the east flank of the Tertiary

Hills; the sections located along the west bank of the Mackenzie River opposite Seagull Island; those located on its north bank near Fort Norman and opposite Police Island; the outcrop along the Great Bear River at Fort Norman (RAK-83-19); and those near the mouth of the Brackett River (Fig. 5). The conclusions arising from the intrabasinal correlation of strata include:

1. Correlations established palynologically provide evidence that the coal and tuff horizons are laterally discontinuous, thus refuting the simple lithological correlation of these horizons (compare Figures 3 and 5 with Ricketts, 1985, Figures 1 and 2).
2. The clastic wedge thinned from west to east as might be expected within the sedimentological framework established for the basin.
3. Coal deposition commenced earlier in the west than in the east, reflecting an easterly to northeasterly prograding system.
4. The basal and thickest coal in the Police Island section is unrepresented in the Tertiary Hills.
5. The youngest coal seams occur opposite Seagull Island, upstream from Fort Norman and in outcrops near Tate Lake. Slightly older strata in the Tertiary Hills contain only thin coaly shales.

COAL COMPOSITION WITHIN THE SUMMIT CREEK FORMATION

Reflectance values for these coals of 0.39 to 0.59 (Ro random) indicate ranks ranging from lignite to high volatile C bituminous. The lower seams in the Police Island section on the Mackenzie River appear more attrital with higher inertinite content than seams in the upper part of the coal-bearing interval. Coals in the type section (RAK-83-3) in the Tertiary Hills also show attrital layers interbanded with more woody zones. The lower seams in the Police Island section as well as the seams in the RAK-83-3 section belong to the *Paraalnipollenites alterniporus* zone (Fig. 5). The higher seams in the Police Island section are brighter, that is they contain higher amounts of the eu-ulminite component of the group maceral huminite. These seams belong to the *Aquilapollenites spinulosus/Wodehouseia capillata* zone (Fig. 5). The coals with the highest huminite content and least amount of inertinite occur in the Seagull Island sections sampled on the west bank of the Mackenzie River (Fig. 3). These coals are the youngest in the sequence and belong to the *Momipites actinus* zone (Fig. 5). Abundance of the various huminite macerals, particularly eu- ulminite with tissue structure in various states of retention suggests a forested swamp environment and/or good conditions of preservation. A dark reflecting variety of eu- ulminite (eu- ulminite A) is abundant in some coal layers and appears to bear some relationship to the abundance of gymnosperm pollen. About 70 per cent of the samples of coal from the basin show ash contents of 10 to 30 % (dry basis). The average sulphur content is about 0.6 per cent.

PALYNOSTRATIGRAPHY

In addition to the reports included in Yorath and Cook (1981), three previous studies have dealt with some aspect of the palynology of the Brackett Basin. Palynomorphs from the upper part of the Little Bear Formation and the lower beds of the East Fork Formation were dated as late Campanian or early Maastrichtian by Brideaux (1971). Bihl (1973) described palynological assemblages from the Summit Creek and East Fork formations from outcrops within the Tertiary Hills and correctly inferred a Maastrichtian to Paleocene age for these sediments. In a regional study of sediments spanning the Cretaceous-Tertiary boundary, Wilson (1978) included three samples from the Police Island section of the Summit Creek Formation, two of which he found contained a Maastrichtian assemblage, and one a Paleocene assemblage.

In the present study, a palynological zonation has been developed for the Little Bear to East Fork formations and the overlying Summit Creek Formation. These will be discussed separately. The informal zones described below are conceived as being closest to concurrent range zones although sometimes importance is placed on the relative abundance of particular species or groups. Taxa used to label the zones may be supplanted in the future by as yet undescribed species, which may have greater utility as zonal index species.

Informal zones of the Little Bear and East Fork formations

Three zones are recognized in outcrops of the Little Bear Formation and one in the overlying East Fork Formation (Fig. 4). As no single continuous outcrop spans this interval, the palynomorph assemblages are based on several localities. Given the sporadic nature of the outcrop coverage it is possible that additional assemblages occur within the Little Bear and East Fork formations. Index and selected characteristic species of the zones are illustrated in Plate 1, figures 1 to 18.

Azonia sp. cf. *A. sufflata* zone

Little Bear Formation; Little Bear River; lower part of type section (NC-83-36); Figures 3 and 4; Plate 1, figures 1, 2.

Definition

The presence of a finely reticulate form of *Azonia sufflata* Wiggins, 1976 (seen only within the 40 to 80 m interval of the section) is definitive for this zone. Only a sparse and nondescript assemblage was recovered from the lower 40 m of section NC-83-36.

Characteristic species

In addition to *Azonia, Umbosporites callosus* Newman, 1965 occurs in an otherwise sparse and nondescript pollen and spore assemblage, the dominant form of angiosperm pollen being simple, prolate, tricolpate pollen. The presence of dinoflagellates supports a marine environment of deposition for this interval.

Age

Santonian. The absence of *Aquilapollenites* in combination with the microplankton reported by Brideaux (1981) support this age assignment.

Aquilapollenites rombicus/Diptycha spp. zone

Little Bear Formation; MacKay Range section (NC-83-31), 0 to 75 m; Figures 3 and 4; Plate 1, figures 3 to 7.

Definition

Aquilapollenites rombicus Samoilovich, 1965 and closely allied species including forms similar to *A. rigidus* Tschudy & Leopold, 1971 occur throughout this zone. Its top is marked by the last occurrence of *Diptycha* spp.

Characteristic species

In addition to the above species this zone is characterized by a diverse suite of angiosperm pollen including species of *Azonia, Echiperiporites, Fibulapollis, Jacutiana* and *Weylandipollis*. The miospore *Umbosporites callosus* is also consistently present. This diversity contrasts sharply with that found in the *Azonia* sp. cf. *A. sufflata* zone. The presence of dinoflagellates supports the interpretation of a marine environment of deposition for the interval containing this assemblage in the MacKay Range section.

Age

Early Campanian. Species in this zone occur within assemblage 4 of Nichols and Sweet (in press) inferred to be of early Campanian age, based on comparisons with assemblages from Montana described by Nichols et al. (1982).

Aquilapollenites trialatus zone

Little Bear Formation; MacKay Range section (NC-83-31), 106 to 122 m; Figures 3 and 4; Plate 1, figures 8 to 17.

Definition

The first records of *Aquilapollenites trialatus* Rouse, 1957 mark the base of this zone, and the last records of *Mancicorpus calvus* (Tschudy & Leopold) Tschudy, 1973 should prove useful in marking its top. In the MacKay Range section there is an intervening, 31 m thick, dominantly sandstone interval between this and the underlying zone. The few mudstone samples from the intervening 31 m interval are of low diversity and appear to lack species characteristic of either zone.

Characteristic species

This zone is characterized by the presence of *Aquilapollenites insignis* Mchedlishvili, 1961; *A. trialatus* var. *uniformis* Tschudy & Leopold, 1971; *A. trialatus* var. *variabilis* Tschudy & Leopold, 1971; *A. turbidus* Tschudy, 1973; *Atopospora cancellata* (Burbridge & Felix) Jansonius & Hills, 1978; *Azonia parva* Wiggins, 1976; *A. recta* (Bolkhovitina) Samoilovich, 1961; *A. sufflata* and *Trudopollis* sp. Together with the above species, *Expressipollis accuratus* Chlonova, 1961 occurs in the upper 2 m of the type section of the Little Bear Formation in beds assigned to the East Fork Formation by Yorath and Cook (1981). Both marine mudstone yielding rich dinoflagellate assemblages and nonmarine coal-bearing strata occur within the interval spanned by this zone.

Age

Late Campanian. This zone contains some of the same species as assemblage 5 of Nichols and Sweet (in press), which is considered to be of late Campanian age based on its similarity to Judith River Formation assemblages. The distinctive species *Atopospora cancellata* and *Expressipollis accuratus* also occur in the Eureka Sound Group of Ellef Ringnes Island, called Maastrichtian by Felix and Burbridge (1973), an age that possibly should be questioned.

Triprojectus unicus zone

East Fork Formation; type 'locality' (SLA-85-16); Figures 3 and 4; Plate 1, figure 18.

Definition

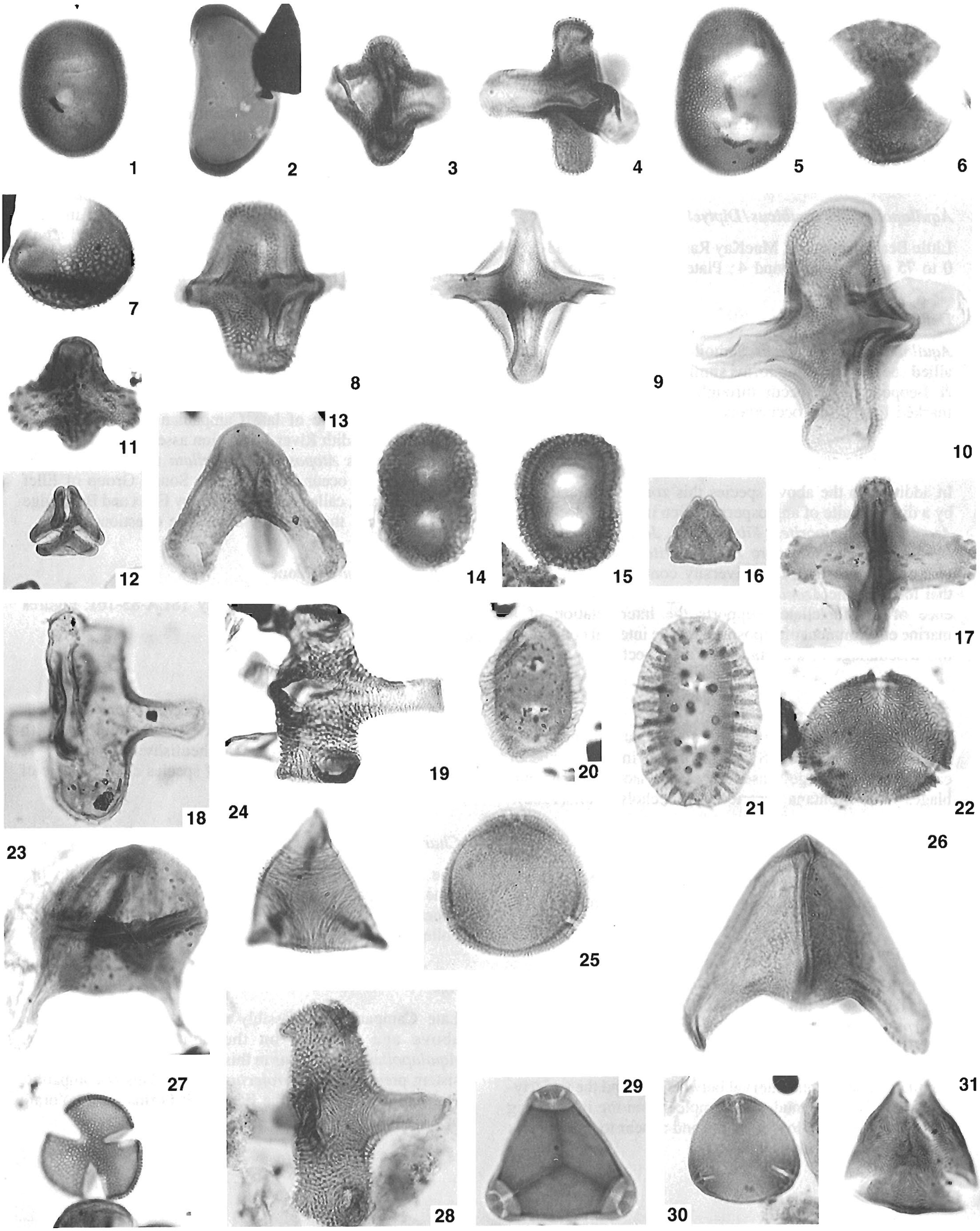
This zone is identified by the presence of *Triprojectus unicus* (Chlonova) Mchedlishvili, 1961 together with *Aquilapollenites senonicus* (Mchedlishvili) Tschudy & Leopold, 1971 and the absence of species characteristic of the over- and underlying zones.

Characteristic species

The range of *Aquilapollenites trialatus* extends into this zone. Most palynomorphs within this fully marine facies are either reworked Carboniferous and younger spores and pollen or marine microplankton.

Age

Late Campanian to possibly earliest Maastrichtian. The above age is based on the continued presence of *Aquilapollenites trialatus* in this zone together with the consistent presence of *Triprojectus unicus*. This is compatible with the age given for the East Fork Formation in Yorath and Cook (1981).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Informal zones of the Summit Creek Formation

Four zones are described from the Maastrichtian portion and three from the Paleocene portion of the Summit Creek Formation (Fig. 5). As the near vertically dipping beds in the Police Island section (SLA-85-13; SLA-86-2) span the greatest stratigraphic interval and provide the most continuous control of any of the outcrop sections, it is used as the main reference section. It is supplemented with the sections opposite Seagull Island for the description of the youngest zone. Index and selected characteristic species for the zones are illustrated in Plate 1, figures 19 to 31 and Plate 2, figures 1 to 16.

Wodehouseia gracile/Aquilapollenites parallelus zone

Police Island section, 0 to 60 m; Figures 3 and 5; Plate 1, figure 20.

PLATE 1

All figures x750, unless otherwise stated.

Figures 1, 2. *Azonia* sp. cf. *A. sufflata* zone.

1. *Azonia* sp. cf. *sufflata* Wiggins, 1976; GSC 93631, P2605-5a, 131.2 x 9.2, GSC locality C-107811.
2. *Umbosporites callosus* Newman, 1965; GSC 93632, P2605-5c, 134.6 x 14.6, GSC locality C-107811.

Figures 3-7. *Aquilapollenites rombicus/Diptycha* spp. zone.

3. *Aquilapollenites rombicus* Samoilovich, 1965; GSC 93633, P26071b, 132.0 x 16.3, GSC locality C-107830.
4. *A. rigidus* Tschudy & Leopold, 1971; GSC 93634, P2607-8c, 116.3 x 16.0, GSC locality C-107837.
5. *Azonia* sp. cf. *A. parva* Wiggins, 1976; GSC 93635, P2607-4b, 126.9 x 13.6, GSC locality C-107833.
6. *Diptycha* sp.; GSC 93636, P2607-6b, 128.0 x 18.8, GSC locality C-107835.
7. *Weylandipollis retiformis* Takahashi, 1964; GSC 93637, P2607-8a, 108.3 x 9.4, GSC locality C-107937.

Figures 8-17. *Aquilapollenites trialatus* zone.

8. *A. trialatus* var. *variabilis* Tschudy & Leopold, 1971; GSC 93638, P2605-15b, 134.1 x 2.6, GSC locality C-107821.
9. *Aquilapollenites trialatus* var. *uniformis* Tschudy & Leopold, 1971; GSC 93639, P2607-12c, 108.4 x 8.6, GSC locality C-107841.
10. *Aquilapollenites trialatus* Rouse, 1957; GSC 93640, P2605-22b, 112.0 x 19.2, GSC locality C-107828.
11. *A. turbidus* Tschudy, 1973; GSC 93641, P2605-18b, 113.2 x 8.9, GSC locality C-107824.
12. *Atopospora cancellata* (Burbridge & Felix) Jansonius and Hills, 1978; GSC 93642, P2607-14d, 136.6 x 10.3, GSC locality C-107843.
13. *Mancicorpus calvus* (Tschudy & Leopold) Tschudy, 1973; GSC 93643, P2605-22b, 111.8 x 13.7, GSC locality C-107828.
14. *Azonia recta* (Bolkhovitina) Samoilovich, 1961; GSC 93644, P2605-17e, 131.6 x 17.3, GSC locality C-107823.
15. *A. sufflata* Wiggins, 1976; GSC 93645, P2605-15b, 108.7 x 19.8, GSC locality C-107821.
16. *Trudopollis* sp.; GSC 93646, P2607-14b, 130.6 x 9.3, GSC locality C-107843.
17. *Aquilapollenites insignis* Mchedlishvili, 1961; GSC 93647, P2605-23b, 116.6 x 9.6, GSC locality C-107829.

Figure 18. *Triprojectus unicus* zone.

18. *Triprojectus unicus* (Chlonova) Mchedlishvili, 1961; GSC 93648, P3029-13b, 114.0 x 13.6, GSC locality C-135668.

Figure 19. See *Myrtipites scabratus/Aquilapollenites delicatus* var. *collaris* zone below.

Figure 20. *Wodehouseia gracile/Aquilapollenites parallelus* zone.

20. *Wodehouseia gracile* (Samoilovich) Pokrovaskaya 1966; GSC 93650, P2980-30c, 130.4 x 11.2, GSC locality C-150383.

Figures 21-27. *Mancicorpus gibbus/Scollardia trapaformis* zone.

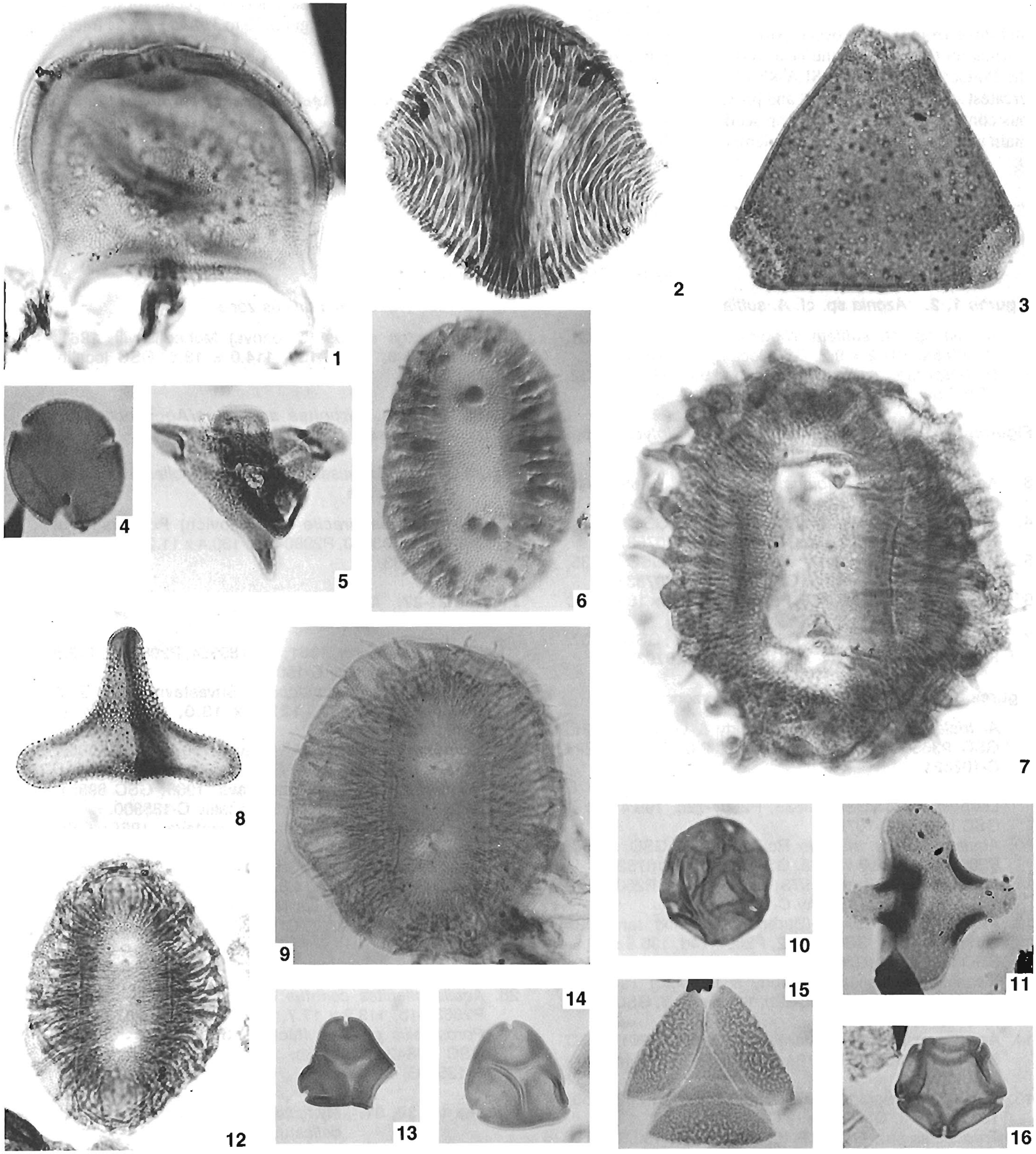
21. *W. spinata* Stanley, 1961; GSC 89854, P2980-1b, 113.6 x 20.3,GSC locality C-150354.
22. *Callistopollenites tumidoporus* Srivastava, 1969; GSC 93651, P2980-23b, 121.6 x 13.0, GSC locality C-150376.
23. *Mancicorpus gibbus* Srivastava, 1968; GSC 93652, P2980-8b, 134.6 x 11.6, GSC locality C-150361.
24. *Scollardia trapaformis* Srivastava, 1966; GSC 89851, P2863-1b, 127.1 x 4.6, GSC locality C-135600.
25. *Pulcheripollenites inrasus* Srivastava, 1969; GSC 93653, P2863-1b, 116.2 x 11.3, GSC locality C-135600.
26. *Mancicorpus rostratus* Srivastava, 1968; GSC 93654, P2599-26b, 107.6 x 14.0, GSC locality C-111625.
27. *Tricolpites interangulus* Newman, 1965; GSC 89847, P2980-2c, 108.8x11.7, GSC locality C-150355, x1000.

Figures 28, 29. *Porosipollis porosus* zone.

28. *Aquilapollenites conatus* Norton, 1965; GSC 89857, P2863-31b, 113.2 x 17.7, GSC locality C-99873A.
29. *Porosipollis porosus* (Mchedlishvili) Krutzsch, 1969; GSC 89842, P2863-6b, 116.8 x 18.3, GSC locality C-135605, x1000.

Figures 19, 30, 31. *Myrtipites scabratus/Aquilapollenites delicatus* var. *collaris* zone.

19. *Aquilapollenites conatus* Tschudy, 1969; GSC 93649, P2863-15b, 124.3 x 5.1, GSC locality C-135612.
30. *Myrtipites scabratus* Norton in Norton and Hall, 1969; GSC 89856, P2863-33Ad, 132.6 x 5.7, GSC locality C-98765.
31. *Striatellipollis radiata* Sweet, 1986; GSC 89855, P2863-22b, 118.9 x 3.3, GSC locality C-135699.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Definition

The top of this zone is defined by the upper limit of *Aquilapollenites parallelus* Tschudy, 1969. It is separated from the overlying zone by a 13 m thick sandstone unit. *Wodehouseia gracile* (Samoilovich) Pokrovaskaya, 1966 occurs to within 25 m of the top of this zone. The base of this zone does not outcrop in the Police Island section.

PLATE 2

All figures x750 unless otherwise stated.

Figures 1-8. *Myrtipites scabratus/Aquilapollenites delicatus* var. *collaris* zone.

1. *Mancicorpus* sp. cf. *M. gibbus* Srivastava, 1968; GSC 89865, P2863-9b, 135.6 x 17.3, GSC locality C-135693.
2. *Jiangupollis major* Song in Song Zhi-chen, Zheng Ya-hui et al., 1980; GSC 89849, P2980-16b, 131.8 x 10.0, GSC locality C-150369.
3. *Proteacidites* sp. cf. *P. globosiporus* Samoilovich, 1961; GSC 93655, P2599-20b, 114.4 x 20.8, GSC locality C-111619.
4. *Tricolpites parvistriatus* Norton in Norton & Hall, 1967; GSC 89862, P2863-33Cd, 128.2 x 12.8, GSC locality C-98765C, x1000.
5. *Aquilapollenites quadricretaeus* Chlonova, 1961; GSC 93656, P2965-14b, 127.2 x 14.7, GSC locality C-150308.
6. *Wodehouseia quadrispina* Wiggins, 1976; GSC 89863, P2863-33Cb, 119.2 x 20.7, GSC locality C-98765C.
7. *Wodehouseia fimbriata* var. *constricta* Wiggins, 1976; GSC 89866, P2599-26e, 124.2 x 19.0, GSC locality C-111625.
8. *Aquilapollenites delicatus* var. *collaris* Tschudy & Leopold, 1971; GSC 93657, P2863-22b, 117.6 x 12.4, GSC locality C-135699.

Figures 9, 10. *Paraalnipollenites alterniporus* zone.

9. *Wodehouseia fimbriata* Stanley, 1961; GSC 93658, P2863-51b, 123.6 x 16.7, GSC locality C-99893G.
10. *Paraalnipollenites alterniporus* (Simpson) Srivastava, 1977; GSC 93659, P2863-36c, 114.2 x 13.4, GSC locality C-98765K.

Figures 11, 12. *Aquilapollenites spinulosus/Wodehouseia capillata* zone.

11. *Aquilapollenites spinulosus* Funkhouser, 1961; GSC 93660, P2984-15a, 114.8 x 6.6, GSC locality C-147177.
12. *Wodehouseia capillata* Wiggins, 1976; GSC 93661, P2696-4b, 123.4 x 14.1, GSC locality C-111636.

Figures 13-16. *Momipites actinus* zone.

13, 14. *Momipites actinus* Leffingwell, 1971; (13) GSC 93662, P2699-3a, 129.4 x 11.6, GSC locality C-111655; (14) GSC 93663, P2699-4d, 132.2 x 3.8, GSC locality C-111656.

15. *Insulapollenites rugulatus* Leffingwell, 1971; GSC 93664, P2699-4a, 131.3 x 11.6, GSC locality C-111656.

16. *Alnipollenites verus* Potonié, 1931; GSC 93665, P2699-4d, 130.6 x 6.2, GSC locality C-111656.

Characteristic species

Triprojectus unicus (Chlonova) Mchedlishvili, 1961 is abundant throughout the interval. Commonly occurring taxa include *Callistopollenites* spp.; *Orbiculapollis lucida* Chlonova, 1961; *Pulcheripollenites inrasus* Srivastava, 1969; *Siberiapollis* spp.; and *Wodehouseia spinata* Stanley, 1961. Of the above species only *Orbiculapollis lucida* and *Wodehouseia spinata* range into the zone immediately below the Cretaceous-Tertiary boundary.

Age

Mid(?) Maastrichtian based on the age of the overlying zone.

Mancicorpus gibbus/Scollardia trapaformis zone

Police Island section, 72 to 150 m; Figures 3 and 5; Plate 1, figures 21 to 27.

Definition

The top of this zone corresponds with the stratigraphically highest range of *Mancicorpus gibbus* Srivastava, 1968 and *Scollardia trapaformis* Srivastava, 1966. It is separated from the overlying zone by a 2.5 m thick sandstone bed with an erosional base. The base of this zone approximates the first occurrences of *Kurtzipites trispissatus* Anderson, 1960; *Mancicorpus rostratus* Srivastava, 1968; *M. gibbus; Striatellipollis striatella* (Mchedlishvili) Krutzsch, 1969; and *Wodehouseia octospina* Wiggins, 1976.

Characteristic species

Triprojectus unicus continues to be abundant in this interval and *Cranwellia rumseyensis* Srivastava, 1966 becomes very abundant within its upper part. Those taxa listed as commonly occurring in the previous zone extend through this assemblage. Rare records of *Tricolpites interangulus* Newman, 1965 were made and *Aquilapollenites conatus* Norton, 1965 was recorded in the upper 1 m of this interval.

Age

Mid-Maastrichtian. This zone is probably correlative with the *Scollardia trapaformis* Zone (VI) of Srivastava (1970), which spans the interval of the Thompson coal seam in the upper part of the Horseshoe Canyon Formation of Alberta.

Porosipollis porosus zone

Police Island section, 150 to 180 m; Figures 3 and 5; Plate 1, figures 28, 29.

Definition

The top and base of this zone are defined by the local range of *Porosipollis porosus* (Mchedlishvili) Krutzsch, 1969. In the Police Island section it is separated from the overlying zone by a 13 m thick conglomerate unit. In addition to

P. porosus, its top is also defined by the last occurrence of *Callistopollenites* spp. and its base by the the first occurrence of *Leptopecopites pocockii* (Srivastava) Srivastava, 1978.

Characteristic species

This zone is the first in which *Aquilapollenites conatus* is consistently present.

Age

Mid or late Maastrichtian. This and the lower part of the following zone do not appear to be represented in the Red Deer Valley of central Alberta.

Myrtipites scabratus/Aquilapollenites delicatus **var.** ***collaris*** **zone**

Police Island section, 193 to about 217 m; Figures 3 and 5; Plate 1, figures 19, 30, 31 and Plate 2, figures 1 to 8.

Definition

The top of this zone is defined by the last occurrence of *Wodehouseia quadrispina* Wiggins, 1976 and its base by the first occurrences of *Myrtipites scabratus* Norton in Norton & Hall, 1969 and *Striatellipollis radiata* Sweet, 1986. It appears to be in conformable contact with the overlying zone, the boundary occurring within a 12 m thick coal seam.

Characteristic species

Several large and morphologically exotic species occur in the lower, approximately 14 m, of this zone. These include *Mancicorpus* sp. cf. *M. gibbus*, which is restricted to the interval, and the longer ranging species *Siberiapollis* spp., *Jiangupollis major* Song in Song Zhi-chen, Zheng Ya-hui et al., 1980 and *Triprojectus unicus*. *Wodehouseia fimbriata* var. *constricta* Wiggins, 1976 is present in all but the uppermost part of this zone as well as the underlying zone. *Proteacidites* sp. cf. *P. globosiporus* Samoilovich, 1961 and *Tricolpites parvistriatus* Norton in Norton & Hall, 1967 are prominent throughout this zone. *Aquilapollenites delicatus* var. *delicatus* is common in the lower part, whereas abundant specimens of *A. delicatus* var. *collaris* Tschudy & Leopold, 1971 characterize the upper (but not uppermost) part of the zone. *Aquilapollenites quadricretaeus* Chlonova, 1961 and *Wodehousiea quadrispina* occur frequently in the upper coal portion of this interval.

Age

Latest Maastrichtian. This zone is at least in part correlative with the *Wodehouseia spinata* zone of Srivastava (1970), lower part of the Scollard Formation, central Alberta. It has many species in common with those of the Frenchman Formation of Saskatchewan and the Hell Creek Formation of the northern mid-continental United States. Its upper limit marks the Cretaceous-Tertiary boundary. A presumably correlative assemblage with a significantly reduced diversity occurs directly below the Paleocene in the Tertiary Hills (Sections SLA-88-13 and SLA-86-9). This allows the position of the Cretaceous-Tertiary boundary to be projected into sections within the Tertiary Hills as the prime datum for intrabasinal correlations (Fig. 5).

Paraalnipollenites alterniporus **zone**

Police Island section, about 217 m to 268 m; Figures 3 and 5; Plate 2, figures 9, 10.

Definition

Paraalnipollenites alterniporus (Simpson) Srivastava, 1977 ranges both higher and lower than the perceived zonal limits. The upper limit of this zone conforms approximately to the stratigraphically highest occurrence of *Wodehouseia fimbriata* Stanley, 1961 — the margin of uncertainty being an overlap of less than one metre in the upper limit of *W. fimbriata* and the lower limit of *Aquilapollenites spinulosus* Funkhouser, 1961, the index species for the base of the next higher zone. Its base is the last occurrence of most typically Maastrichtian species.

Characteristic species

Paraalnipollenites alterniporus is conspicuous throughout this zone, which is otherwise characterized by the near absence of morphologically distinct species. Exceptions are the presence of new species of *Wodehouseia* and of *Striatellipollis* about 4 m above the assumed position of the Cretaceous-Tertiary boundary and the occurrence of *Wodehouseia fimbriata* starting at a horizon about 22 m above the Cretaceous- Tertiary boundary. The palynomorph assemblage is dominated by *Laevigatosporites* and Taxodiaceae-Cupressaceae pollen. Simple, prolate-reticulate and less commonly psilate, tricolpate species are the most abundant forms of angiosperm pollen present. *Ericaceoipollenites rallus* Stanley, 1965 occurs sporadically.

Age

Earliest Paleocene. Brideaux (1981) and Wilson (1978) placed the Cretaceous-Tertiary boundary about 20 m above the base of the *Paraalnipollenites alterniporus* zone based on the presence of well developed mid to upper Maastrichtian assemblages in mudstone within its lower part. These specimens are herein considered to be reworked.

Aquilapollenites spinulosus/Wodehouseia capillata **zone**

Police Island section, 241 to 302 m; Figures 3 and 5; Plate 2, figures 11, 12.

Definition

The lowest occurrences of *Aquilapollenites spinulosus* mark the base of this zone. *Wodehouseia capillata* Wiggins, 1976 occurs in the interval between 275 to 302 m, which is the top of the Police Island section. In the Seagull Island sections, *W. capillata* occurs in the stratigraphically lowest beds providing a link between the two localities and with the overlying *Momipites actinus* zone.

Characteristic species

Alnipollenites verus Potonié, 1931 first appears at 288 m and is consistently present until the top of the section at 302 m. Otherwise, with the exception of *W. fimbriata* not being present, the species are as for the underlying assemblage.

Age

Mid-Paleocene, based on the presence of *Aquilapollenites spinulosus* (see Demchuk, 1987).

Momipites actinus zone

Seagull Island sections (RAK-83-9 to 14; SLA-86-24); Figures 3 and 5; Plate 2, figures 13 to 16.

Definition

Momipites actinus Nichols and Ott, 1978 occurs sporadically but persistently throughout the zone. Both *Insulapollenites rugulatus* Leffingwell, 1971 and *Wodehouseia capillata* occur in the lowest part of this zone, otherwise these species appear to have mutually exclusive ranges.

Characteristic species

Alnipollenites verus continues to be common to abundant and triporate pollen referable to the Betulaceae-Myricaceae complex are abundant as they commonly form 1 to 10 per cent of the total assemblage. *Alnipollenites trina* (Stanley) Norton, 1969, *Aquilapollenites spinulosus, Ericaeoipollenites rallus*, and *Paraalnipollenites alterniporus* continue their range throughout this zone. New entries include *Ericipites longisulcatus* Wodehouse, 1933, *Insulapollenites rugulatus, Phaseoliidites stanleyi* Elsik, 1968, *Ulmipollenites undulosus* Wolff, 1934, and a new species of *Wodehouseia* with a narrow flange. All these species occur sporadically.

Age

Mid-Paleocene. Because of the presence of *Momipites actinus*, it appears that the age range of this zone should be restricted to the P3 and P4 zones of Nichols and Ott (1978). Additionally, the absence of *Pistillipollenites mcgregorii* Rouse, 1962 indicates a middle Paleocene (P3 or P4), rather than a Late Paleocene age. It is of note that Nichols and Ott did not record *Insulapollenites rugulatus* until their P5 zone, which conforms with the generally accepted Late Paleocene range for *I. rugulatus* (McIntyre, this volume).

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Thermal maturity of Lower Carboniferous rocks in northern Yukon Territory[†]

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Utting, J., Thermal maturity of Lower Carboniferous rocks in northern Yukon Territory; in Current Research, Part G, Geological Survey of Canada, Paper 89-1G, p. 101-104, 1989.

Abstract

Analyses of palynomorphs and organic matter indicate that the Thermal Alteration Index is high (TAI 3+ to 5) in the Lower Carboniferous Kayak Formation of the British Mountains and Barn Range; low (TAI 2 to 2+) in the Ford Lake and Hart River formations of the southeastern part of Eagle Plain, where many samples contain amorphous kerogen; medium (TAI 3+) in the southern subsurface of Eagle Plain (Hart River Formation); and low to medium (TAI 2- to 3+) in the western Ogilvie Mountains (Hart River and Blackie formations).

Résumé

Des analyses de palynomorphes et de matière organique indiquent que l'indice d'altération thermique est élevé (IAT de 3+ à 5) dans la formation du Carbonifère inférieur de Kayak dans les monts British et les chaînons Barn; faible (IAT de 2 à 2+) dans les formations de Ford Lake et de Hart River dans la partie sud-est de la plaine d'Eagle, où de nombreux échantillons contiennent du kérogène amorphe; moyen (IAT de 3+) dans la subsurface sud de la plaine d'Eagle (formation de Hart River); et faible à moyen (IAT de 2– à 3+) dans la partie ouest des monts Ogilvie (formations de Hart River et de Blackie).

[†] Contribution to Frontier Geoscience Program.

INTRODUCTION

The Thermal Alteration Index (TAI) of Lower Carboniferous rocks from northern Yukon Territory was assessed from the colour of the spores, and a value estimated using a five point scale (Table 1) summarized by Hunt (1979) and modified by Utting (1987) and Utting et al. (in press). On this scale, pale yellow spores have a TAI of 1, and vitreous, black, brittle spores have a TAI of 5. Those between 1 and 5 are assigned a value according to the indices shown in Table 1. The assessments are somewhat subjective and vary according to differences in exine thickness (Utting et al., in press), but the data obtained are nevertheless of considerable value for determining the thermal maturation history and hydrocarbon potential.

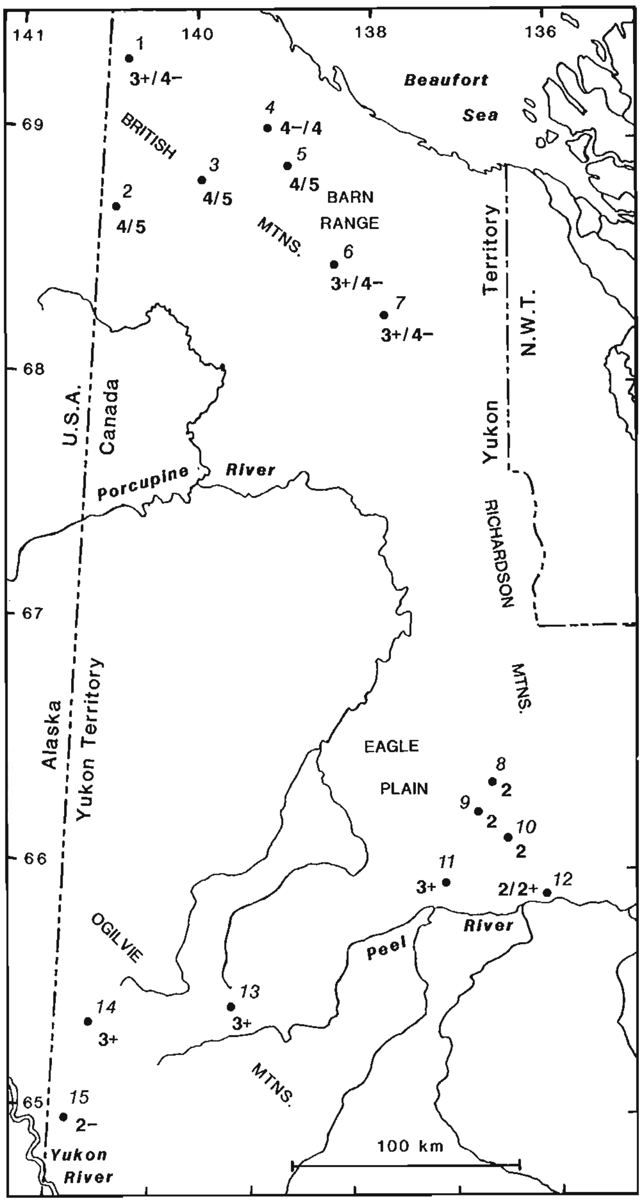

Figure 1. Sample localities and Thermal Alteration Indices.

In the northern part of the area, (British Mountains and Barn Range) the thermal maturity of the Kayak Formation (Viséan) was studied at seven outcrop localities (Fig. 1 and Table 2, stations 1-7). At these localities, the Kayak Formation consists of calcareous and carbonaceous shale, coal seams, limestone, and minor amounts of sandstone and conglomerate. Biostratigraphic data from foraminifers and corals indicate that the upper Kayak Formation is diachronous within the British Mountains (Bamber and Waterhouse, 1971), and ranges in age from early middle Viséan in the west, to early late Viséan in the east. Palynological work in progress confirms a Viséan age. The ranks of coals and carbonaceous shale were assessed by Cameron et al. (1986). Their work indicates a high level of thermal maturity, with vitrinite reflectance values varying from 2.33 in the Crow River area (Fig. 1 and Table 2, station 4), to a high of 4.03 at the Alaskan border. Coal exposures at Hoidahl Dome (Fig. 1 and Table 2, station 7), including samples from two exposures of a 5.5 m thick seam, were also reported on by Cameron et al. (1988); the rank of this coal, as indicated by vitrinite reflectance (2.89 to 3.27% Ro_{max}), spans the semianthracite/anthracite boundary.

In the southeastern and southern part of Eagle Plain (Fig. 1), the upper part of the Ford Lake Formation and the Hart River Formation were studied at four outcrop localities (Fig. 1 and Table 2, stations 8-10, 12) and one subsurface locality (Fig. 1 and Table 2, station 11). The Ford Lake Formation consists of dark grey and reddish brown weathering shale. Results of goniatite studies (Bamber and Waterhouse, 1971), indicate that the upper part (91.5 m below the top of the formation) is of late middle Viséan age. Palynomorphs from the uppermost part of the formation (8.0 m below the top of the exposed section) indicate a late Viséan (V3) age (Utting, personal observation). The Hart River Formation, proposed by Bamber and Waterhouse (1971), consists of light and dark brownish grey weathering, laminated, silty, skeletal-micritic and micritic limestone, silty, calcareous dolomite, and chert. The age of the formation in this area, based on the marine fauna and palynomorphs

Table 1. Thermal Alteration Indices of spores and pollen

Colour of spores and pollen in transmitted light	Thermal Alteration Index
Transparent to pale yellow	1
Yellow	1+
Yellow to light orange	2−
Medium orange	2
Dark orange to light brown	2+
Medium brown	3−
Dark brown	3
Thinner parts of spore exine dark brown, thicker parts brownish black	3+
Thinner parts of exine brownish black, thicker parts black	4−
Dull black, haptotypic mark barely visible	4
Vitreous black, fossils brittle, often fractured, species rarely identifiable (haptotypic mark may be visible with reflected light).	5

(Utting, personal observation), is late Viséan (V3) and early Namurian. The Blackie Formation, proposed by Pugh (1983), is the deeper water, basinal equivalent of the Ettrain Formation of Bamber and Waterhouse (1971). Its lowest part may be a facies equivalent of the uppermost part of the Hart River Formation. The Blackie Formation (Fig. 1 and Table 2, station 15) and the Hart River Formation (Fig. 1 and Table 2, stations 13 and 14), were studied in the Ogilvie Mountains.

RESULTS

In the British Mountains and Barn Range (Fig. 1 and Table 2, stations 1-7) the thermal maturity of the Kayak Formation is high (TAI 3+ to 5); it lies within the dry gas generation zone, or, at some localities, beyond the dry gas preservation limit. The TAI is lower (3+ to 4−) in the vicinity of Hoidahl Dome and the Barn Range (Fig. 1 and Table 2, stations 6 and 7), and increases northwestward to 4− to 5 along the British Mountains, only to decrease again slightly to 3+ to 4− in the Malcolm River area, close to the Alaskan border (Fig. 1 and Table 2, station 1). This decrease differs from the trend suggested by vitrinite reflectance values; maximum values were recorded close to the Alaskan border (Cameron et al., 1986).

In the southeastern part of Eagle Plain (Fig. 1 and Table 2, stations 8-10, 12), in the upper part of the Ford Lake Formation and the overlying Hart River Formation (including the type section of the latter on the Peel River; Fig. 1 and Table 2, station 12), the thermal maturity is low (TAI 2 to 2+), and thus within the oil window. Conodonts from the Hart River Formation type section (GSC locs. C-149017-149029, C-149031-149039, C-149042-149047) have a Colour Alteration Index (CAI) of 2 to 2.5 (C.M. Henderson, pers. comm., 1988), again indicating a thermal maturity within the oil window. In many samples from the Hart River Formation, at the two localities studied (Peel River and Palmer Lake; Fig. 1 and Table 2, stations 12 and 10, respectively), there is an abundance of amorphous kerogen, which suggests that this formation merits further study as a possible source rock for liquid hydrocarbons. In the subsurface, for example, the Socony Mobil — W.Minerals Blackie YT M-59 well (Fig. 1 and Table 2, station 11),

Table 2. Sample localities and Thermal Alteration Indices

Station	Location (lat.; long.)	Locality name	GSC loc. number	Formation	TAI
1	69°19′18″N; 140°42′12″W	Malcolm River	C-161106 to C-161115	Kayak	3+/4−
2	68°42′09″N; 140°49′36″W	Firth River	C-161227 to C-161243	Kayak	4/5
3	68°50′N; 140°00′W	Trail River Headwaters	C-161206 to C-161216	Kayak	4/5
4	69°00′N; 139°28′W	Crow River	C-153031 to C-153050 C-161434 to C-161436	Kayak	4−/4
5	68°51′36″N; 139°00′W	Mount Sedgwick	C-161136 to C-161155	Kayak	4/5
6	68°28′30″N; 138°25′W	Barn Range	C-161437 to C-161444	Kayak	3+/4−
7	68°17′N; 137°53′W	Hoidahl Dome	C-161116 to C-161133	Kayak	3+/4−
8	66°20′30″N; 136°44′W	Eagle Plain	C-92525	Ford Lake	2
9	66°15′30″N; 137°48′30″W	Eagle Plain	C-149127 to C-149131	Ford Lake	2
10	66°09′N; 136°26′W	Palmer Lake	C-130343 to C-130355	Hart River	2
			C-130354	Ford Lake	2
11	66°58′55″N; 137°11′11″W	Blackie M-59	C-39387 (1601.4 to 1931.8 m)	Hart River	3+
12	65°53′20″N; 136°06′W	Peel River	C-149021 to C-149061	Hart River	2/2+
			C-148951 to C-149015	Ford Lake	2/2+
13	65°27′45″N; 139°10′30″W	Mount Chief Isaac	C-150945	Hart River	3+
14	65°24′N; 140°40′W	Jungle Creek	C-105403 to C-105418	Hart River	3+
15	64°58′N; 140°54′W	Tatonduk River	C-149137 to C-149243	Blackie	2−

located in the southern part of Eagle Plain, the thermal maturity is medium (TAI 3+).

In the Ogilvie Mountains, the thermal maturity of the Hart River Formation is medium (TAI 3+) in the Mount Chief Isaac area (Fig. 1 and Table 2, station 13), and similar values occur in the west part of the Ogilvie Mountains close to the Alaskan border on Jungle Creek (Fig. 1 and Table 2, station 14). However, farther south (Fig. 1 and Table 2, station 15), the lower part of the Blackie Formation on the Tatonduk River, has a TAI of 2-; the conodont Colour Alteration Index of two samples (from GSC locs. C-149141-149142) is 1.5 (C.M. Henderson, pers. comm., 1988).

CONCLUSIONS

Lower Carboniferous strata in the British Mountains and Barn Range have a high thermal maturity consistent with considerable depth of burial; it is possible that tectonic loading may have increased the thermal maturity in some localities. The influence of intrusions is not significant in this area. This high thermal maturity, along with the presence of anthracitic coal seams up to 5.5 m thick in the Hoidahl Dome, suggests a good potential for gas generation. To the south of Eagle Plain the thermal maturity, at least in some areas, is very low, suggesting that in places there was no significant depth of burial. These areas may well contain liquid hydrocarbon source rocks, especially in view of the abundant amorphous organic matter in the Hart River Formation.

In the Ogilvie Mountains there is medium thermal maturity (TAI 3+) at Jungle Creek and at Mount Chief Isaac, yet low maturity (TAI 2-) in the Tatonduk River area. There are not yet sufficient data to explain these differences.

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Upper Paleozoic stratigraphy and basin analysis of the Sverdrup Basin, Canadian Arctic Archipelago: Part 1, time frame and tectonic evolution†

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Abstract

The Carboniferous and Permian paleogeography of the Sverdrup Basin, Canadian Arctic Archipelago, was characterized by a peripheral shallow water platform surrounding a series of deep water basins. During that time interval, climatic conditions evolved from warm and arid to cold and more humid. Twenty-three conodont zones have been recognized in the upper Paleozoic succession of the Sverdrup Basin, providing a basic time framework which has been used to date the various depositional and tectonic events that affected the basin. Upper Paleozoic sediments were laid down during four phases of abortive continental rifting characterized by: 1. thermal uplift, 2. rifting, 3. strike-slip faulting (Melvillian Disturbance), and 4. passive subsidence.

Résumé

La paléogéographie carbonifère et permienne du bassin de Sverdrup de l'archipel Arctique canadien se distinguait par la présence, en eaux peu profondes, d'une plate-forme périphérique entourant plusieurs bassins profonds. Pendant ce temps, le climat évoluait de chaud et aride, à froid et plus humide. Vingt-trois zones de conodontes ont été reconnues dans la succession du Paléozoïque supérieur du bassin de Sverdrup. Ces zones ont permis d'établir un canevas stratigraphique servant à dater les divers événements tectoniques et sédimentaires qui ont affecté le bassin. La sédimentation datant du Paléozoïque supérieur a été accompagnée de quatre grandes phases tectoniques qui témoignent d'un rift continental qui a avorté: 1) soulèvement thermique, 2) création d'un fossé, 3) formation de failles de décrochement (accident du Melvillien), et 4) subsidence passive.

† Contribution to Frontier Geoscience Program.

INTRODUCTION

This series of two articles summarizes our current stratigraphic, sedimentological, and tectonic knowledge of the upper Paleozoic succession of the Sverdrup Basin. More specifically, it focuses on the interpretation of data acquired by the authors through six field seasons (1983 to 1988) on Axel Heiberg, Melville, Devon and Ellesmere islands. This data formed part of Ph.D. research projects with the University of Calgary (Beauchamp and Henderson) and Rice University, Houston, (Harrison). This research was carried out in the context of ongoing, long-term projects with the Geological Survey of Canada, financed by the Frontier Geoscience Program, and logistically supported in the field by the Polar Continental Shelf Project.

In this article, the overall upper Paleozoic geological setting is reviewed with emphasis on paleogeography, paleoclimatology, biostratigraphic time frame, and tectonic evolution.

GEOLOGICAL SETTING

Sverdrup Basin

The Sverdrup Basin is a 1000 km long and 400 km wide rift basin underlying the northernmost islands of the Canadian Arctic Archipelago (Fig. 1). It formed in Early Carboniferous time through extension, faulting and collapse of the Precambrian to Devonian rocks of the Franklinian Mobile Belt (Fig. 2). These had previously been deformed by at least two orogenies, the more recent and important being the Ellesmerian Orogeny of latest Devonian or earliest Carboniferous time (Thorsteinsson and Tozer, 1970). Rocks of the Sverdrup Basin were deformed in the Early Tertiary by the Eurekan Orogeny, a tectonic event probably associated with the creation of the Baffin Bay and Atlantic spreading centres (Kerr, 1981), which led to extensive faulting and folding in the eastern Arctic, and minor deformation in the western Arctic.

Paleogeography

Throughout its late Paleozoic history, the Sverdrup Basin was surrounded by a continental land mass to the north and south (Embry, 1988). Only narrow and probably shallow connections existed with the nearby oceans, permitting exchange with the global ocean system (Beauchamp et al., 1987). As a result of sea level fluctuations, there were times when the Sverdrup Basin was nearly landlocked, such as early in the Late Carboniferous, and possibly in latest Permian time.

Both the southern (Thorsteinsson, 1974) and northern (Trettin, 1969; Mayr, in press) basin margins were rimmed by a belt of pericontinental, coarse clastic sediments. Because of the opening of the Arctic Ocean, the northern clastic belt is preserved only in some areas of Axel Heiberg and Ellesmere islands. The preserved marginal clastics pass basinward into platform carbonates and clastics, which in turn grade into basinal, probably euxinic shale, evaporite, and chert. The deep water deposits accumulated in three

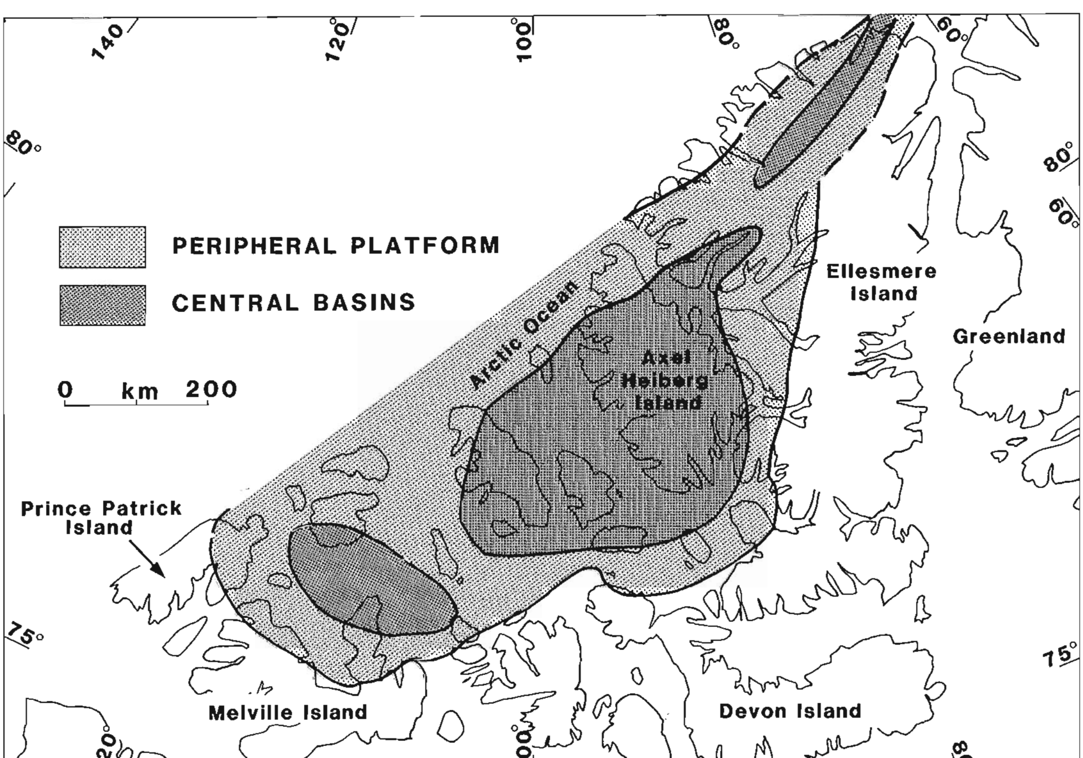

Figure 1. Known (solid) and inferred (dashed) outline of Sverdrup Basin and late Paleozoic paleogeographic elements.

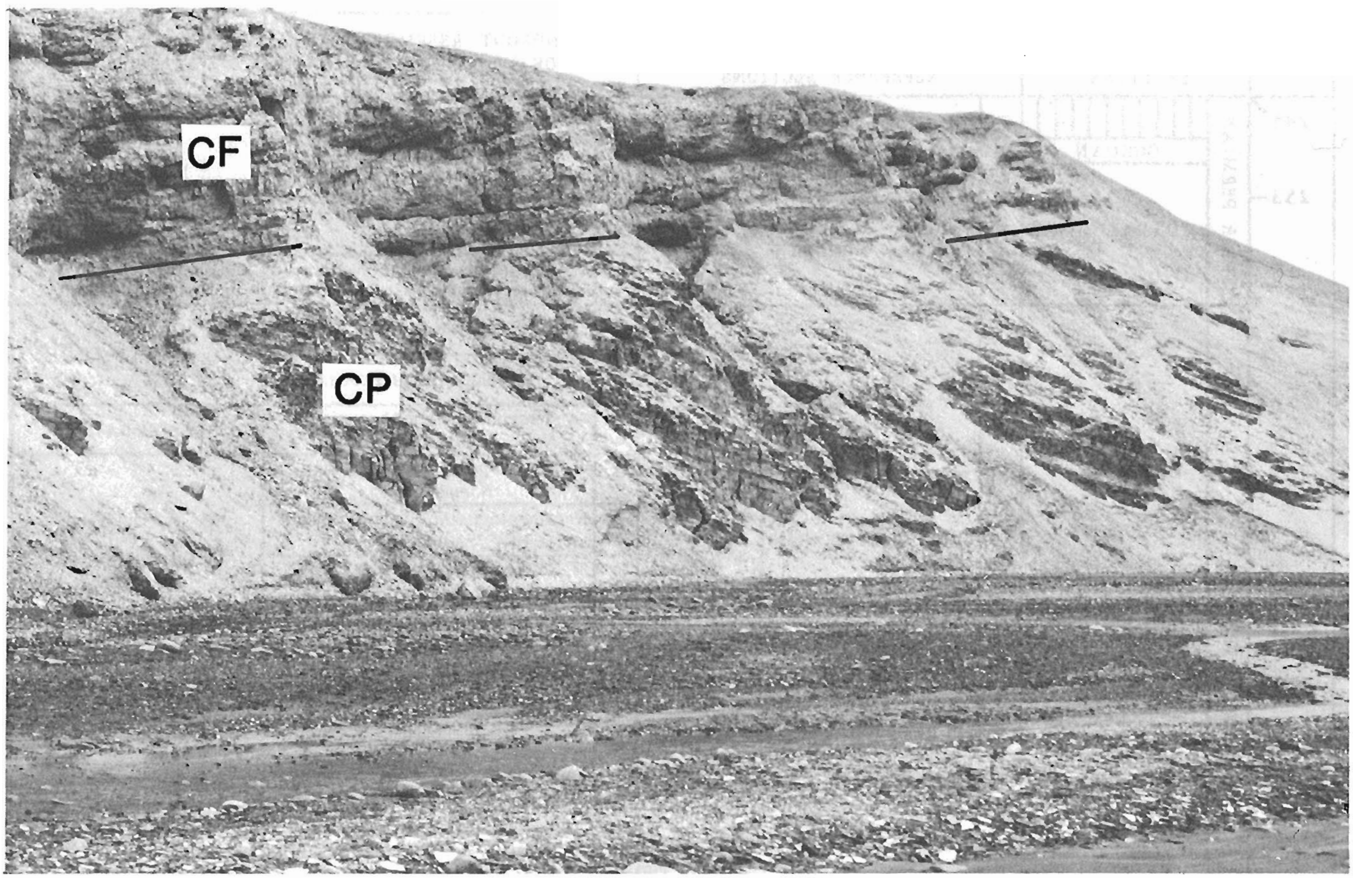

Figure 2. Angular unconformity between fine clastic sediments of the Siluro-Devonian Cape Phillips (CP) Formation (Franklinian Mobile Belt), and conglomerates of the Upper Carboniferous Canyon Fiord (CF) Formation (Sverdrup Basin). Location is 5 km southwest of McCormick Inlet, Melville Island.

large depressions, centred north of Sabine Peninsula, Melville Island (Barrow Basin), west-central Axel Heiberg Island (Axel Heiberg Basin), and north of Lake Hazen, northern Ellesmere Island (Meneley et al., 1975; Mayr, in press) (Fig. 1). The boundaries between the marginal, platformal, and basinal belts migrated considerably through time, and their respective facies content changed substantially as a result of low- to high-order sea level fluctuations, ongoing tectonic activity, and climatic variations.

Late Paleozoic paleotectonic plate reconstructions of North America place the Sverdrup Basin within the Northern Hemisphere. During the Early Carboniferous to Late Permian time interval, the Sverdrup Basin wandered from less than 10° N to more than 50° N (Smith et al., 1981).

Paleoclimatology

The northward continental migration was paralleled by a substantial drop in ambient temperatures, which led to a net decline in the diversity of biotic and abiotic assemblages. Highly diversified tropical assemblages dominated by calcareous algae, foraminifers, ooids and oncoids, prevailed throughout the Late Carboniferous and early part of the Early Permian, whereas poorly diversified temperate assemblages, dominated by sponges, bryozoans, echinoderms and brachiopods, characterized the later part of the Early Permian and the early part of the Late Permian (Beauchamp, 1987). This trend was marked by three significant events: 1. one event in latest Sakmarian (Early Permian) time, which signalled the definitive establishment of impoverished temperate associations, with rare tropical-like leftovers such as colonial corals, fusulinaceans and some small foraminifers; 2. one event in late Artinskian (Early Permian) time, which led to the eradication of all fusulinids and colonial corals, and most families of small foraminifers; and 3. one event in latest Wordian time (Late Permian), which resulted in the disappearance of all calcareous shelly animals.

During the same time interval, the Sverdrup Basin climate evolved from semi-arid to relatively humid (Beauchamp, 1987). This is evident from the presence of abundant redbeds, caliche-like crusts, and evaporites in Lower Permian (Asselian-Sakmarian) and older deposits, and the lack thereof, in addition to the occurrence of coal layers, in the younger deposits. Palynomorph assemblages show a similar trend of decreasing aridity through time (Utting, 1989).

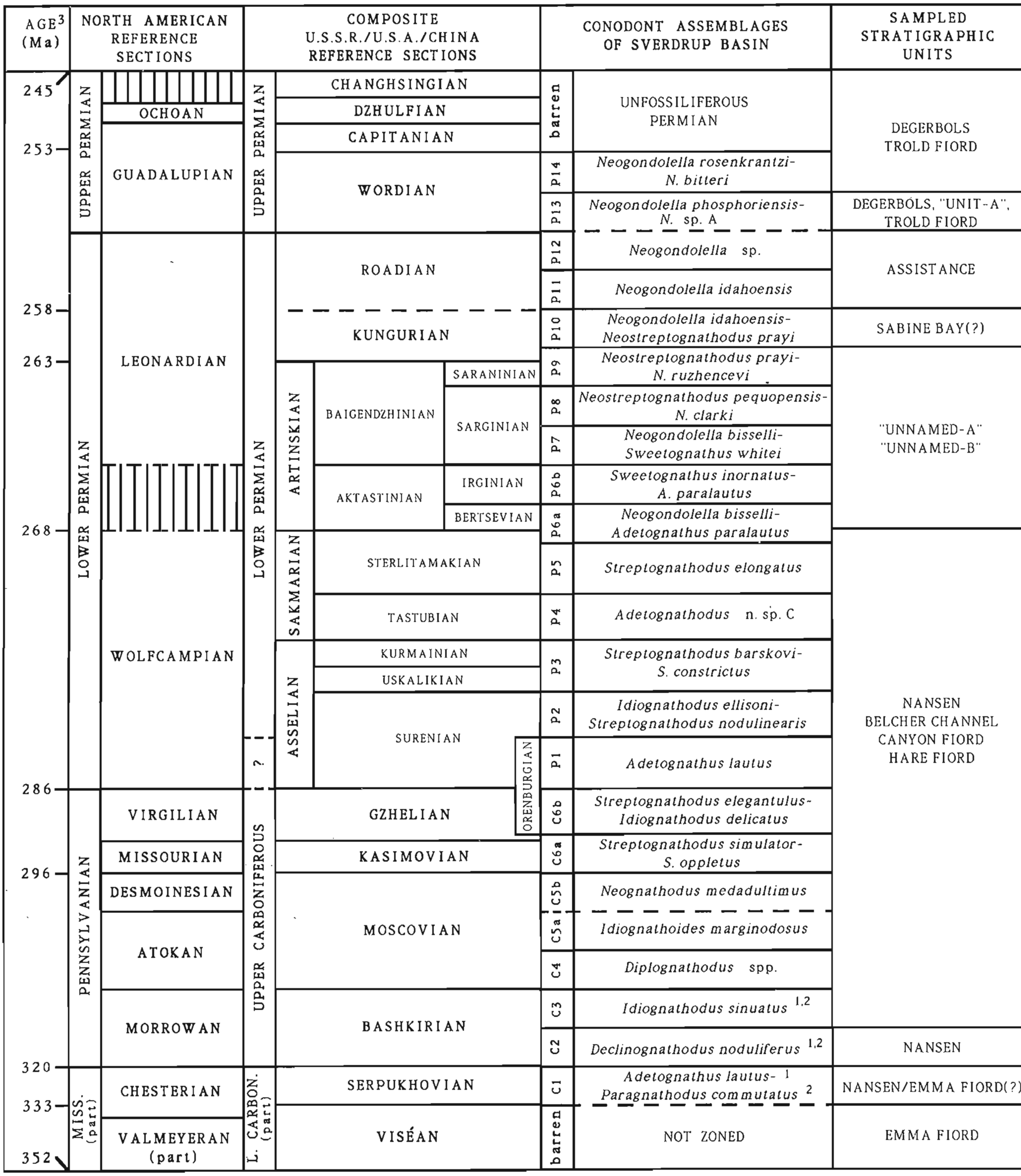

Figure 3. Preliminary interpretation of the age ranges for conodont assemblages defined in the Sverdrup Basin. Zones C1 to C3 are based on (1) Bender (1980) and (2) A.C. Higgins (pers. comm., 1987). Zones C4 to P9 are based on Henderson (1988). Zones P10 to P14 are based on Henderson (1981). (3) Age dates are from Palmer (1983). Except for Zone C4, which is an assemblage zone, all zones are interval range zones (partial range or concurrent range).

TIME FRAME

Many fossil groups have been shown to be useful for dating parts of the upper Paleozoic succession of the Sverdrup Basin; these include ammonoids (Nassichuk 1975a, b; Nassichuk et al., 1965), brachiopods (Waterhouse and Waddington, 1982; Liao and Henderson, work in progress), conodonts (Bender, 1980; Nassichuk and Henderson, 1986), corals (Bamber in Nassichuk and Wilde, 1977), small foraminifers (Mamet in Thorsteinsson, 1974), fusulinaceans (Nassichuk and Wilde, 1977), and palynomorphs (Utting, 1989; Utting et al., in press). The age assignments outlined in this report are based largely on conodonts, as recent developments in upper Paleozoic conodont biostratigraphy have led to the definition of 23 zones, the faunal content of which can be correlated with precision, at the stage and sub-stage level, with chronostratigraphic stratotypes of North America and Europe. These zones, shown in Figure 3, provide a basic time framework that has been used to bracket and correlate the various depositional sequences and tectonic events that affected the Sverdrup Basin (Fig. 4). To date, conodonts have been recovered from clastic and carbonate rocks ranging in age from Serpukhovian (late Early Carboniferous) to Wordian (early Late Permian), and representing a vast array of deep to shallow marine depositional settings. Viséan strata, which in the Sverdrup Basin are nonmarine (Emma Fiord Formation), have been precisely dated by palynology (Utting et al., in press).

Chronostratigraphic correlation of Carboniferous and Permian strata necessitates comparison with "standard" successions in the Soviet Union and the United States. Assemblages C1 to C5 (Fig. 3) are largely correlated with North American sequences of Chesterian to Desmoinesian age (Lane and Straka, 1974; Merrill, 1975; Grayson, 1984). Assemblages C6 to P10 are correlated with stratotype sequences in the Ural Mountains (Movshovich et al., 1979; Movshovich, 1984; Akhmetshina et al., 1984). Some of these latter assemblages do not correspond exactly between the two regions, a situation which may be related to local facies variations and possible endemism of the conodont faunas. However, where differences do occur, the assemblages can be bracketed between other assemblages that are directly correlatable. Assemblages P11 to P14 are compared to sequences in West Texas and the Great Basin region of the U.S.A. (Wardlaw and Collinson, 1979; 1986; Wardlaw and Grant, 1987).

The chronostratigraphic scheme for the Permian presented in the Canadian Volume of DNAG (Henderson et al., in press) indicates that the Artinskian is succeeded by the Roadian. The synchroneity of this correlation is doubtful in view of conodont faunal comparisons between the two type sections — a procedure that is further complicated by the wide separation of the type regions (Glass Mountains, West Texas for Roadian, and Ural Mountains, U.S.S.R. for Artinskian). Faunas in zones P9 and P10 in the Sverdrup Basin (Henderson, 1988) seem to compare well to both post-Artinskian (Kungurian; Movshovich et al., 1979) and pre-Roadian (upper Cathedral Mountain Formation; Wardlaw and Grant, 1987) faunas; as a result, the Kungurian is inserted between the Artinskian and the Roadian.

TECTONIC EVOLUTION

Upper Paleozoic sediments of the Sverdrup Basin were laid down during four phases of abortive continental rifting characterized by: 1. Early Carboniferous thermal uplift, 2. Early Carboniferous to Early Permian rifting, 3. Early Permian strike-slip faulting (Melvillian Disturbance), and 4. late Early Permian to latest Permian passive subsidence (Fig. 4).

Thermal uplift

In this initial phase of rifting (Fig. 4A), spanning Viséan to earliest Serpukhovian time, a thermal plume produced symmetrical thermal uplifts thoughout the areal extent of the Sverdrup Basin. Based on the observation of three distinct depositional trends in the subsequent marine sequences, it is suggested that one major rift arm extended northeast to northern Ellesmere Island, a second major arm extended west to northern Prince Patrick Island, and a third possible rift arm trend extended to the southeast in the direction of Bjorne Peninsula, southwestern Ellesmere Island. These rift basin systems, now almost completely covered by younger strata, preserve the earliest record of rift-related sedimentation and are characterized by deposits resulting from internal drainage. These include lacustrine marlstones of the Emma Fiord Formation (Davies and Nassichuk, 1988), and proximal alluvial fan deposits of the Borup Fiord Formation. The shoulders of these axial rift basins, although presumably fractured, remained high during the period of thermal uplift and thus, were probably in a continuous state of erosion.

The thermal maximum is marked by a phase of Serpukhovian alkali basaltic volcanism now represented by the flows preserved in the Audhild and Borup Fiord formations (Trettin, 1988; Nassichuk and Davies, 1980). Regional uplift created by this thermal maximum has locally produced an unconformity above the Emma Fiord Formation.

Rifting

The beginning of thermal collapse and the main phase of rifting began in the latest Early Carboniferous (Serpukhovian) (Fig. 4B). Growth faulting was most active in the Late Carboniferous but may have continued as late as the middle Early Permian (Artinskian) (Fig. 4C). The onset of rifting is marked by a net preservation of sediments above the post-Emma Fiord unconformity. The previously uplifted shoulders of the rift system experienced rapid subsidence during this main phase of extension.

An array of sub-basins and genetically associated uplifts has been produced in the rift marginal belt by block rotation and extensional slip on rift parallel listric growth faults that root into Franklinian basement. Some of these sub-basins have been exhumed by later uplift and erosion, and are now exposed at the surface. Examples include the McCormick Depression of northwestern Melville Island, and the Weatherall Depression of Sabine Peninsula (Harrison and Riediger, in press). Other upper Paleozoic depressions have

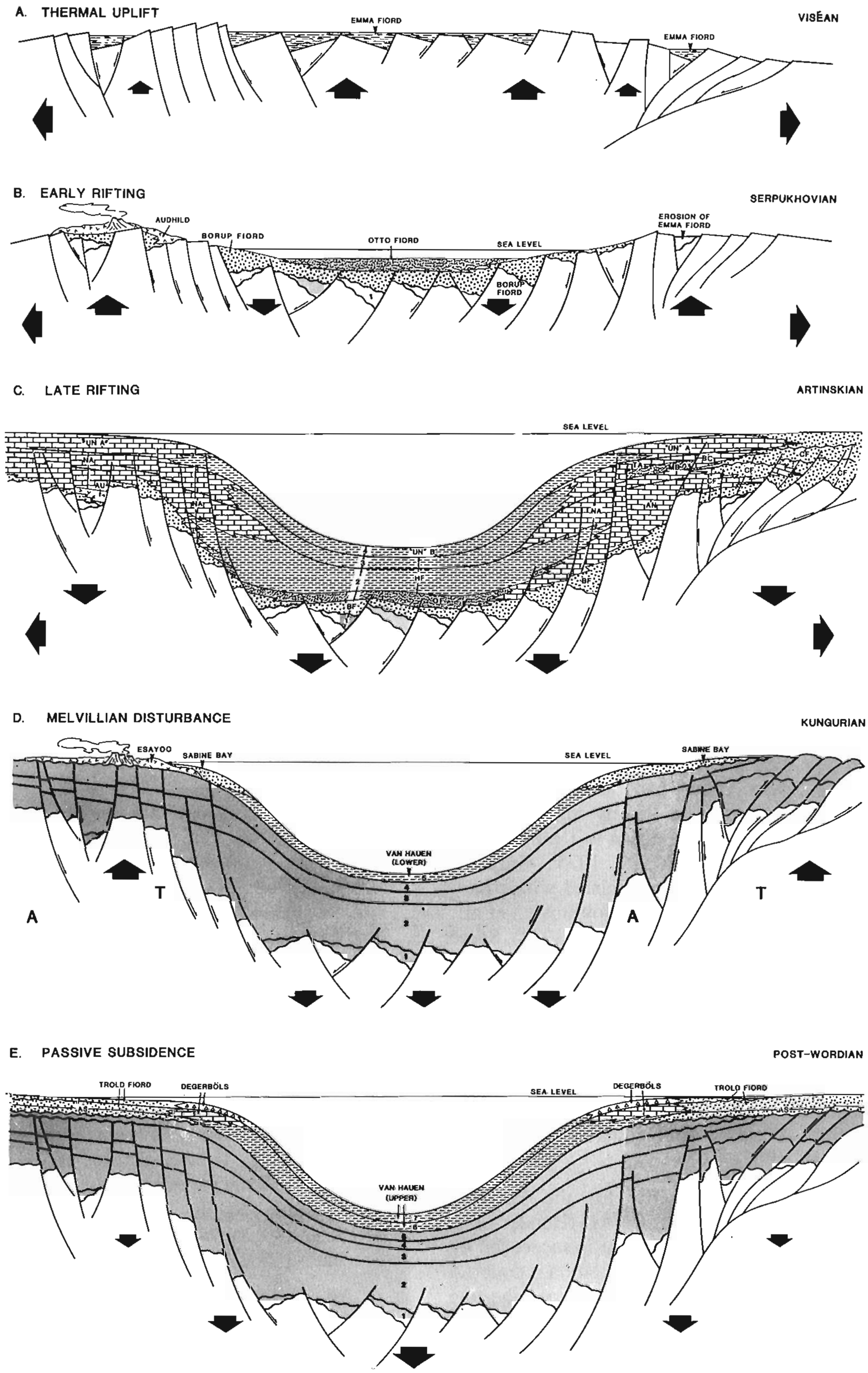

Figure 4. Schematic north-south cross-sections depicting the late Paleozoic tectonic evolution of the Sverdrup Basin. All faults are conjectural. A. Initial thermal uplift. B. Early rifting. C. Late rifting. D. Melvillian Disturbance. E. Passive subsidence. Numbers 1 to 7 refer to transgressive-regressive sequences described in the second article of this series. Lithology symbols as in Figure 1 of second article. NA, Nansen; AU, Audhild; BF, Borup Fiord; OF, Otto Fiord; HF, Hare Fiord; AN, Antoinette; MB, Mount Bayley; TA, Tanquary; BC, Belcher Channel; CF, Canyon Fiord; "UN" A, "Unnamed" A; "UN" B, "Unnamed" B; AS, Assistance.

been identified on reflection seismic records of central Sabine Peninsula, Cameron Island, and offshore subsurface areas southwest of Graham Island.

Rift sub-basins also formed farther north, in the marginal areas of the Sverdrup Basin, as is indicated by: 1. the sudden collapse, in Early Permian time, of the Nansen carbonate platform in the Blind Fiord area, southwestern Ellesmere Island, leading to the formation of a narrow, reef-bounded, deep water trough (Beauchamp et al., in press); and 2. the sudden collapse of the Antoinette carbonate platform on Fosheim and Hamilton peninsulas, leading first to sedimentation of deep water Hare Fiord-like argillaceous limestone and shale, and then to deposition of the Mount Bayley subaqueous evaporite. Evidence of contemporaneous tectonic activity in those areas includes the presence of thick and coarse fan conglomerates in the Sakmarian part of the Canyon Fiord Formation on southeastern Fosheim Peninsula, and the presence of Lower Permian, probably extensional, basaltic flows interfingering with limestones of the Tanquary Formation near Notch Lake in the Sawtooth Range.

These volcanic rocks indicate that thermal activity resumed from time to time throughout the main rifting phase of the Sverdrup Basin. As will be seen later, the rifting phase was accompanied by the deposition of three broad sequences resulting from long-term (5 to 50 Ma) relative sea level fluctuations. It is possible that these sea level fluctuations were induced by corresponding long-term cycles of thermal uplift and collapse, the former leading to sea level withdrawal and the latter leading to transgression. This suggestion is in accord with current models of sedimentary basin formation (Vail et al., 1977; McKenzie, 1978; Watts, 1982).

Melvillian Disturbance

The Melvillian Disturbance was the name given by Thorsteinsson and Tozer (1970) to the episode of folding and faulting that affected the Upper Carboniferous to Lower Permian Canyon Fiord Formation and older strata, and was responsible for a widespread angular unconformity beneath Kungurian or Roadian strata of northwestern Melville Island (Fig. 5). The Melvillian Disturbance and coeval deformation is now thought to be associated with a phase of wrench faulting throughout the Sverdrup Basin during the Kungurian (Harrison, work in progress). In this strike-slip setting, transpressive uplift and folding is believed to have been produced by restraining bends on strike-slip faults. In the same time frame, transtensive subsidence and sedimentation would have occurred on fault-releasing bends (Fig. 4D).

Figure 5. Evidence of Melvillian uplift: angular unconformity between flat-lying Roadian Assistance and Wordian Trold Fiord formations (AS-TF) and inclined Moscovian to Sakmarian Canyon Fiord (CF) Formation. Lower Triassic Bjorne (BJ) Formation occurs in the background. Location is 30 km southwest of Cape Grassy, Melville Island.

On northwestern Melville Island, dextral transpressive wrench deformation has resulted in the structural inversion of the previously deposited rift-related redbeds of the Canyon Fiord Formation. The detrital products of this inversion are found in the Sabine Bay Formation: a progradational clastic wedge that is discontinuously preserved around the Sverdrup Basin margin. On Melville Island, the Sabine Bay Formation fringes the Melvillian uplifted area on the northeast, east and southeast sides. In other parts of the Sverdrup Basin, the implied tectonic regime was one of simultaneous transtensional growth faulting and net basin subsidence. Such is the case in the subsurface of the Sabine Peninsula area, as indicated on seismic reflection data.

Local uplifts can be attributed to the Melvillian Disturbance in many areas of the Sverdrup Basin outside the Melville Island area. These areas include: 1. the Bjorne Peninsula area, where Artinskian basinal rocks are unconformably overlain by Triassic strata; 2. the Bunde Fiord area, northwestern Axel Heiberg Island, where Triassic sediments rest upon Artinskian platform carbonates; 3. the Bay Fiord area, west-central Ellesmere Island, where a probably emergent high was active in Kungurian time; and 4. the Tanquary High area, northwestern Ellesmere Island, where substantial crustal movements occurred during the same time span (L. Maurel, pers. comm., 1988). Simultaneous transtensive tectonism is also implied by the extrusion of alkali basalt flows now preserved beneath the Sabine Bay Formation in the Esayoo Formation in many areas of northwestern Ellesmere Island and Axel Heiberg Island (Cameron and Muecke, 1988).

Passive subsidence

The Melvillian Disturbance marks the end of widespread late Paleozoic faulting in the Sverdrup Basin area. Reflection seismic data from the Sabine Peninsula area indicate that many growth faults die out in the seismic interval that correlates with the Sabine Bay Formation on the basin margin and with the lowest part of the Van Hauen Formation toward the basin centre (Fig. 4E).

A few basin marginal and basin parallel faults continued to be active during later Permian, and Mesozoic time. In general, however, the sub-Roadian unconformity (sub-Assistance Formation) marks the onset of passive subsidence (Fig. 4E). This is the unconformity that on a trailing continental margin would be labelled the "break-up" unconformity (Falvey, 1974). In the case of the Sverdrup Basin, rifting and subsequent tectonic activity was abortive and did not lead to complete separation of lithospheric plates. In fact, the transition from the rifting phase of tectonism to strike-slip deformation may have been part of the process that terminated late Paleozoic plate separation in the Sverdrup Basin.

CONCLUSIONS

Twenty-three conodont zones have been recognized so far in the Carboniferous and Permian succession of the Sverdrup Basin. These zones have been used to bracket the climatic and tectonic history of the basin. The tectonic history was marked by: 1. a Viséan phase of thermal uplift, 2. a Serpukhovian to early Kungurian phase of rifting, 3. a Kungurian phase of strike-slip deformation (Melvillian Disturbance), and 4. a post-Kungurian phase of passive subsidence.

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Upper Paleozoic stratigraphy and basin analysis of the Sverdrup Basin, Canadian Arctic Archipelago: Part 2, transgressive-regressive sequences†

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Abstract

The Carboniferous and Permian succession of the Sverdrup Basin, Canadian Arctic, is represented by seven, long-term, transgressive-regressive sequences, bounded by major unconformities at the basin margin, which pass basinward into equivalent conformities. The time span represented by each sequence ranges from five to fifty million years. The seven sequences encompass the following chronostratigraphic stages: 1. Viséan, 2. Serpukhovian-Asselian, 3. Asselian-Sakmarian, 4. Sakmarian-Kungurian, 5. Kungurian, 6. Roadian-Wordian and 7. post-Wordian(?).

Résumé

La succession carbonifère et permienne du bassin de Sverdrup de l'Arctique canadien est représentée par sept grandes séquences transgressives et régressives, séparées, à la marge du bassin, par des discordances majeures, passant, vers le centre du bassin, à une sédimentation continue. Chaque séquence représente un intervalle allant de 5 à 50 millions d'années. Les sept séquences sont comprises dans les étages chronostratigraphiques suivants: 1) Viséan, 2) Serpukhovien-Asselien, 3) Asselien-Sakmarien, 4) Sakmarien-Kungurien, 5) Kungurien, 6) Roadien-Wordien et, 7) post-Wordien(?).

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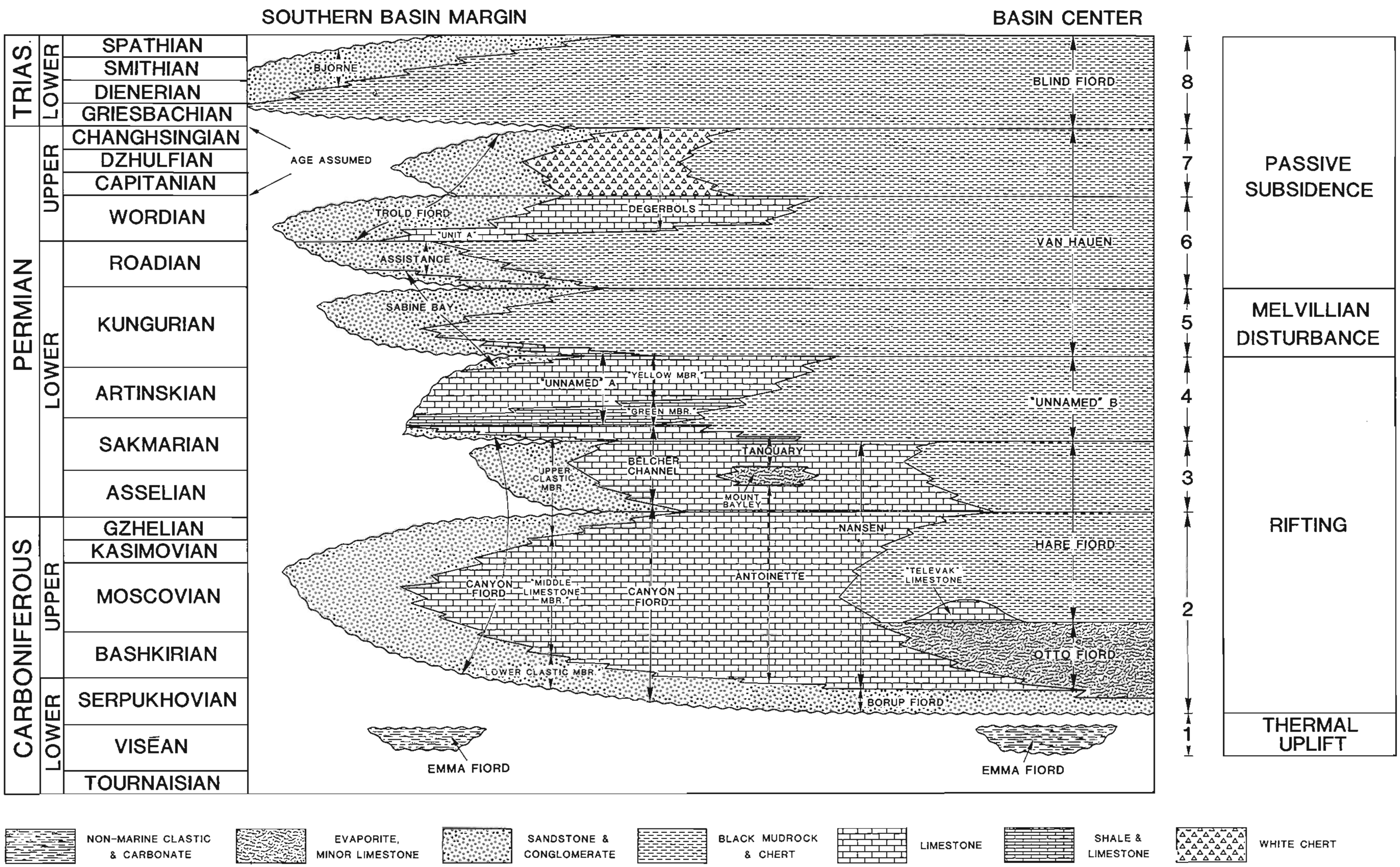

Figure 1. Upper Paleozoic second- and third-order transgressive-regressive sequences (1 to 7) in the Sverdrup Basin along schematic proximal-to-distal cross-section. Sequence 8 is Triassic and is not discussed in the text.

INTRODUCTION

The upper Paleozoic of the Sverdrup Basin is represented by seven, long term transgressive-regressive sequences, separated by major unconformities at the basin margin that pass basinward into equivalent conformities (Fig. 1). The seven sequences encompass the following time intervals: 1. Viséan, 2. Serpukhovian-Asselian, 3. Asselian-Sakmarian, 4. Sakmarian-Kungurian, 5. Kungurian, 6. Roadian-Wordian, and 7. post-Wordian(?). The time represented by each sequence ranges from five to fifty million years. They can therefore be viewed as the product of second- and third-order relative sea level fluctuations (Vail et al., 1977).

This article, the second of two, outlines the facies content of each of these sequences, with emphasis on the sequence boundaries, and intervening deepening-upward and shallowing-upward trends.

CONTROLS ON SEA LEVEL FLUCTUATIONS

The seven sequences appear to be closely related to the four phases of tectonic evolution outlined in the first paper. Sequence 1 was deposited during the initial phase of thermal uplift, Sequences 2, 3 and 4 were contemporaneous with rifting, deposition of Sequence 5 took place during the Melvillian Disturbance, and Sequences 6 and 7 accumulated during the phase of passive subsidence. These associations suggest that the sequences were, to a large extent, genetically linked to local tectonism rather than global eustasy. However, the occurrence of similar and coeval sequences in the Canadian Rocky Mountains, from southeast British Columbia to northern Yukon (Henderson et al., in press), suggests that most of the tectonic factors controlling relative sea level fluctuations not only affected the Sverdrup Basin, but probably the whole western North American cratonic margin.

Higher-order cycles superimposed upon the main sequences also occur thoughout the upper Paleozoic succession. Between 200 and 300 well-defined shelf cycles are characteristic of Upper Carboniferous to Lower Permian Sequences 2 and 3. They probably resulted from high-amplitude, glacio-eustatic, fifth- to sixth-order sea level fluctuations (100 000 to 250 000 years), as suggested by the occurrence of identical cycles in contemporaneous deposits in the Northern Hemisphere (Wilson, 1975), and their synchroneity with a widespread glaciation in the Southern Hemisphere (Veevers and Powell, 1987). High-order cycles also occur within some of the younger sequences. These are far less common and display only mild facies variations in vertical succession compared to their older counterparts. Seven such cycles are recognized within Sequence 4, and between ten and twelve within Sequence 6. They likely formed in response to low-amplitude fourth-order to fifth-order (500 000 to 1 000 000 years) sea level fluctuations, of unknown origin.

SEQUENCE 1: VISÉAN

Sequence 1 comprises fine to coarse clastic sediments associated in certain areas with minor limestones. These rocks are late Viséan in age (Utting et al., in press) and belong to the Emma Fiord Formation. A maximum of 400 m of Emma Fiord sediments are known to occur in the Sverdrup Basin, representing a time interval of approximately 15 Ma. The base of the Emma Fiord Formation is an unconformity; the Emma Fiord rests on deformed and peneplaned Devonian or older Franklinian basement rocks (Fig. 1).

Most of the Emma Fiord Formation is nonmarine, and was deposited within variably spaced lakes (Davies and Nassichuk, 1988). Sparse exposures are known from Grinnell Peninsula on Devon Island, Kleybolte Peninsula on northwestern Ellesmere Island, Svartevaeg Cliff on northern Axel Heiberg Island, and Feilden Peninsula on northeastern Ellesmere Island. The Feilden Peninsula exposure is unique in that marine sediments are believed to occur in its upper part (Mayr, in press). It is likely that many, and probably quite thick, additional occurrences of Emma Fiord sediments are present in the axial region of Sverdrup Basin, and in the subsurface of the Axel Heiberg and Barrow basins.

Emma Fiord sedimentation took place after the peneplanation of the Ellesmerian Orogen, which occurred in Late Devonian or earliest Carboniferous time. As discussed earlier, deposition of the Emma Fiord Formation is believed to have been contemporaneous with the thermal uplift that preceded the rifting of Sverdrup Basin. As a result, a series of half-grabens developed, in which variably deep, stratified lakes accumulated clastic sediments shed from the surrounding exposures through internal drainage. Organic-rich oil shales accumulated in some of the better stratified lakes (Davies and Nassichuk, 1988). In areas where clastic sedimentation was reduced to a minimum, carbonate precipitation took place in the form of oolites and algae-induced tufa crusts. Other environments represented by Emma Fiord sediments include fluvial and marsh environments (Utting et al., in press).

SEQUENCE 2: SERPUKHOVIAN-ASSELIAN

Sequence 2 is a broad transgressive-regressive stratigraphic couplet encompassing more than 50 Ma of depositional history. It includes up to 1800 m of Serpukhovian (Lower Carboniferous) to lower Asselian (Lower Permian) clastics, carbonates, evaporites and minor volcanics belonging to the Borup Fiord, Audhild, Otto Fiord, Hare Fiord, Nansen, Canyon Fiord, and Antoinette formations (Fig. 1). Basal rocks of Sequence 2 either unconformably overlie the Viséan Emma Fiord Formation (Sequence 1) or rest, with angular unconformity, directly on the Devonian or older Franklinian basement.

Rocks directly above the basal unconformity comprise red weathering sandstone and conglomerate, interfingering with minor limestone. In the central areas of Sverdrup Basin, the redbeds are Serpukhovian in age (Mamet, *in* Thorsteinsson, 1974) and belong to the Borup Fiord Formation. In the southern marginal areas, the redbeds are known to be no older than early Bashkirian (Mamet, in Thorsteinsson, 1974), and form the "lower clastic member" of the Canyon Fiord Formation (Beauchamp, 1987). Redbeds of the Borup Fiord Formation and lower Canyon Fiord Formation are genetically related to the initial rifting phase of Sverdrup Basin, and represent sedimentation within fault-bounded grabens and half-grabens. A wide variety of sedimentary environments are represented by these formations, ranging from nonmarine (alluvial fans, braided rivers, lakes, etc.) to marginally marine (estuaries, beaches, lagoons, and tidal flats).

Above the redbeds there is a marine succession that records a substantial transgression that culminated in middle Moscovian time, followed by a regression that reached a maximum during the early Asselian.

Transgression

In the axial regions of the basin, the transgression first led to the deposition of alternating subaqueous evaporite (Nassichuk and Davies, 1980) and biogenic, in places biohermal (Davies and Nassichuk, in press), limestone of relatively shallow water origin (Otto Fiord Formation), followed by deepening-upward, slope to basinal, argillaceous limestone and shale (Hare Fiord Formation). The Otto Fiord-Hare Fiord transition corresponds to the time when a connection between the Sverdrup Basin and adjacent open oceans was fully established, ending evaporite precipitation in the central areas. This transition was also characterized by the local development of Waulsortian-like carbonate buildups, that attained considerable size in keeping pace with rising sea level (Davies et al., in press). These buildups, referred to as the ''Televak'' limestone (Bonham-Carter, 1966) occur at the base of the Hare Fiord Formation in a number of localities (e.g., Blue Mountains, Ellesmere Island).

The Serpukhovian-Moscovian transgression can also be monitored within the platform succession, where it is represented by the Nansen and Antoinette formations. The basal parts of these two units are transitional with the underlying redbeds; increasingly marine calcareous sediments lie above the coarse clastics. Then follows a succession of highly fossiliferous, deepening-upward carbonates, deposited in inner to outer shelf environments. These ultimately grade into thin bedded, Hare Fiord-like, slope to basinal, deep water argillaceous limestone, marking the maximum extent of the transgression.

A similar deepening-upward succession occurs at the basin margin, where the ''lower clastic member'' of the Canyon Fiord Formation is progressively replaced by the ''middle limestone member'', successively represented by inner to outer shelf, fossiliferous limestones (Beauchamp, 1987). These limestones document the maximum advance of the transgressive sea in Moscovian time.

Regression

The subsequent regression led to a shallowing-upward succession in all areas of the Sverdrup Basin. In the central regions, the upper Moscovian-uppermost Gzhelian interval is condensed, as is evident from the narrow interval between lower Moscovian and Asselian ammonoid-bearing strata in the Hare Fiord Formation (Nassichuk, 1975a). The overlying argillaceous limestones are increasingly fossiliferous upward, indicating a net decrease in bathymetry.

In the platform areas, a shallowing-upward succession is displayed within the Nansen and Antoinette formations, as Hare Fiord-like, slope to basinal, argillaceous limestones grade upward into outer to inner shelf carbonates, and ultimately into nearshore carbonates and minor sandstones. Similarly, the basin margin succession is characterized by a return to shallower water facies, as outer shelf carbonates of the ''middle limestone member'' of the Canyon Fiord Formation give way successively to inner shelf carbonates, nearshore limestones and sandstones, and pericontinental to fluvial sandstones. These sandstones form the ''upper clastic member'' of the Canyon Fiord Formation, which prograded basinward as sea level fell (Beauchamp, 1987).

SEQUENCE 3: ASSELIAN-SAKMARIAN

Sequence 3 reaches a maximum thickness of 800 m. It comprises carbonates, clastics, and evaporites of early Asselian to late Sakmarian age (Early Permian), spanning 15 Ma of depositional history. Rocks of Sequence 3 are included in the Hare Fiord, Nansen, Antoinette, Mount Bayley, Tanquary, Canyon Fiord and Belcher Channel formations (Fig. 1). In the marginal regions of Sverdrup Basin, the basal sequence boundary is an unconformity, as demonstrated by the absence of Kasimovian to Asselian biota from certain areas. The time span represented by this unconformity decreases rapidly basinward, and in the central areas, the sequence boundary occurs within a conformable succession. Throughout the Sverdrup Basin, most of Sequence 3 was deepening-upward and transgressive, followed by a rather abrupt regression in late Sakmarian time.

Transgression

In vertical sections, higher-order carbonate cycles of the Nansen and correlative Antoinette and Tanquary formations progressively pass from grainstone-dominated, nearshore to shallow inner shelf sediments, to wackestone- and packstone-dominated, outer shelf carbonate facies associated with *Palaeoaplysina*-phylloid algal buildups of various sizes and shapes (Beauchamp et al., in press).

This succession passes landward into the Canyon Fiord - Belcher Channel couplet, which in vertical section shows a transition from pericontinental sandstones and minor conglomerates (''upper clastic member'') to shallow, inner shelf, grainstone-dominated carbonates and mixed sandstone-carbonates (Belcher Channel Formation). The Canyon Fiord - Belcher Channel contact becomes progressively younger in a landward direction, reflecting the marine transgression that was taking place in the marginal areas.

Regression

The regression in late Sakmarian time was associated with the progradation of sandstone-rich, high-order cycles in the upper part of the Belcher Channel and Tanquary formations. In the more distal Nansen Formation, the regression is monitored by the return to shallower shelf cycles, commonly associated with diagenetic features indicative of subaerial exposure, such as meteoric leaching and vadose silts.

SEQUENCE 4: SAKMARIAN-KUNGURIAN

Sequence 4 is up to 700 m thick. Its basal, deepening-upward succession is displayed in the uppermost Nansen, Tanquary and Belcher Channel formations. The peak of the transgression (maximum flooding surface) in earliest Artinskian time, is represented in the lowest part of an "unnamed" formation ("unnamed" A in Fig. 1; Nassichuk, 1975b; Nassichuk and Wilde, 1977), and correlative basinal deposits ("unnamed" B in Fig. 1). The greater part of these two units displays broad, shallowing-upward successions. Upper Sakmarian to lowest Kungurian fossils were recovered from Sequence 4, which represents a time span of about 5 Ma. At the basin margin, the basal sequence boundary is a mild unconformity, substantiated by paleontological evidence. A conformable transition with Sequence 3 occurs basinward.

The pericontinental clastic equivalent of Sequence 4 has not yet been observed in the marginal areas of Sverdrup Basin. It probably was eroded following the Artinskian sea level drop, and subsequent uplifts related to the Melvillian Disturbance.

Transgression

The last high-order cycles of the Nansen, Tanquary and Belcher Channel formations possess a deeper water character than the immediately underlying cycles (Sequence 3), indicating a substantial rise in sea level. These cycles are also much thicker, reflecting a change in the periodicity of high-order sea level fluctuations. *Tubiphytes*-bryozoan buildups are known to occur within these transgressive deposits (Beauchamp, in press a). A maximum of three cycles are known to occur below the maximum flooding surface, which occurs within the lowest "unnamed" A formation. However, that number decreases toward the basin margin, reflecting the stratigraphic onlap of uppermost Sakmarian strata on top of Sequence 3.

Regression

The "unnamed" A formation is a broad, shallowing-upward succession, comprising a greenish weathering, deep water, mixed shale-limestone, recessive basal member, passing upward into an increasingly fossiliferous, yellowish weathering, cliff forming, fossiliferous limestone member (Fig. 2). This limestone contains a temperate-type biogenic association characterized by crinoids, bryozoans, brachiopods and sponges, contrasting sharply with the underlying formations, which contain highly diversified tropical-type associations. Large, sponge-bryozoan buildups locally occur in the middle part of the "unnamed" A formation (Beauchamp, in press b). The uppermost part of the formation generally comprises variably sandy, crossbedded, nearshore limestone, capped by a conglomerate and/or sandstone, that marks the regression minimum, and defines the boundary with Sequence 5 (Fig. 3).

The "unnamed" A formation grades basinward into a much deeper succession ("unnamed" B), which retains its general shallowing-upward aspect. It comprises thin bedded, black, siliceous shale at the base passing upward into increasingly thick, spiculitic chert, turbiditic in origin. Near the transition with the platform carbonates of the "unnamed" A formation, these black rock types interfinger with minor, bioclastic limestone tongues, which become increasingly abundant upward. The "unnamed"-equivalent basinal strata have been inconsistently mapped as either the Van Hauen or the Hare Fiord formation, but should be defined as a different stratigraphic unit, that has yet to be erected ("unnamed" B in Fig. 1).

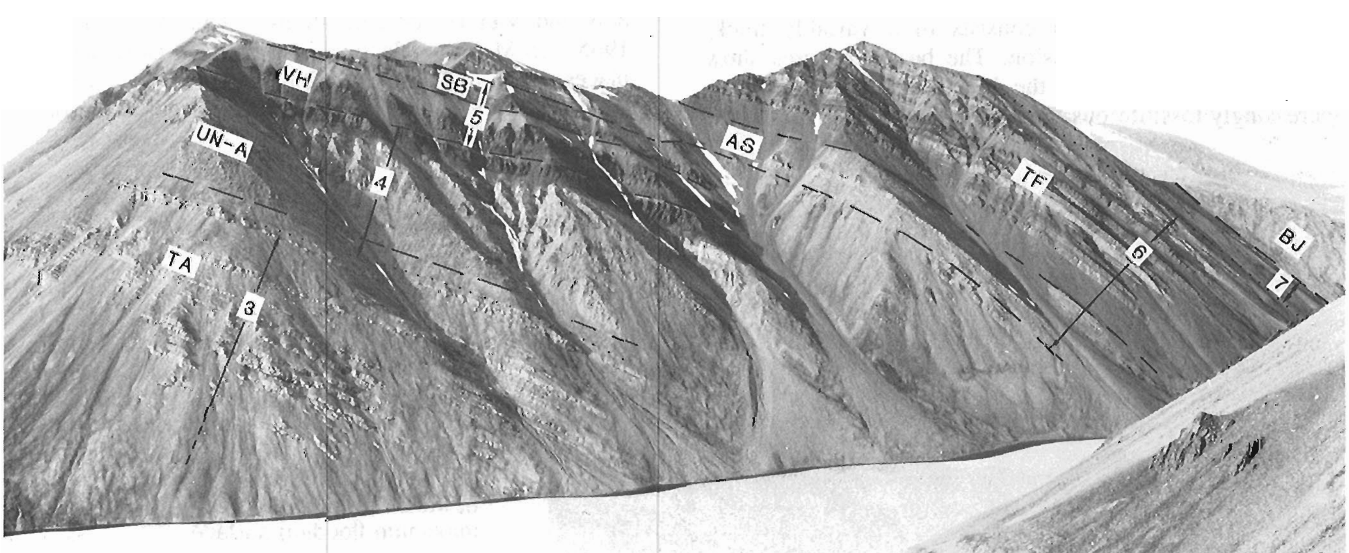

Figure 2. Uppermost Asselian(?) to Triassic succession on south side of Notch Lake, Sawtooth Range, west-central Ellesmere Island. More than 1200 m of strata are shown. Formations (dashed lines) exposed are: Tanquary (TA), "unnamed" A (UN-A) Van Hauen (VH), Sabine Bay (SB), Assistance (AS), Trold Fiord (TF), and Bjorne (BJ) formations. Sequences 3 (part) to 7 are shown. Two, thick, diabase sills appearing as dark horizons occur at the "unnamed" A-Van Hauen contact. Sequence boundaries 4-5 and 5-6 are shown to correspond to formation boundaries, but minor differences do occur (see Figs. 3 and 4).

SEQUENCE 5: KUNGURIAN

Sequence 5 contains strata of Kungurian age. It is up to 350 m thick, and probably represents a depositional interval not exceeding 5 Ma in duration. Clastic sediments characterize Sequence 5, which is represented by the Sabine Bay Formation and its basinal-equivalent, the lower Van Hauen Formation. Sequence 5 can be subdivided into two broad successions: a lower deepening-upward succession, which is rather thin, but probably represents a long-term depositional interval, and an upper shallowing-upward succession, which is thick, but likely represents a short-lived depositional interval. A significant, in places angular, unconformity, confirmed by paleontological data, marks the basal sequence boundary in the marginal regions of Sverdrup Basin. None of the stratigraphic record is missing basinward.

Transgression

In the deep basin, small volumes of sediment were deposited during the transgression, resulting in a condensed interval. In the marginal areas, the transgression led to deposition of a thin, deepening-upward succession; impure, biogenic carbonates overlie the sandstone or conglomerate that marks the basal sequence boundary (Fig. 3). The carbonate is succeeded by increasingly dark and siliceous shales, the fossil content of which decreases rapidly upward. The shale belongs to the Van Hauen Formation. At the basin margin, where parts of the underlying "unnamed" A formation were eroded, the succession, which ranges from a few centimetres to a few metres in thickness, is represented by a green, glauconitic and phosphatic, locally conglomeratic sand passing upward into a shale.

Regression

The bulk of Sequence 5 consists of a variably thick, shallowing-upward succession. The best exposures show black, siliceous shale at the base passing upward into increasingly fossiliferous, calcareous and sandy rock types, the weathering colours of which successively change from dark grey, dark brown, and tan, to pale yellow (Fig. 3). Most of the succession can be mapped as Van Hauen Formation. It ultimately grades into a clean, pale yellow to white sandstone, which belongs to the Sabine Bay Formation. Significant facies variations are displayed throughout the lateral and vertical extent of the Sabine Bay Formation, with at one end of the spectrum, *Zoophycos*-bearing marine sand, and at the other end, unconsolidated nonmarine sand interfingering with thin, coaly horizons. Intermediate facies include trough to tabular cross-stratified sandstones, channelled sandstones and conglomerates, and epsilon cross-stratified sandstones. In vertical sections, the Sabine Bay sediments coarsen and the beds thicken as the succession shallows upward. The Sabine Bay grades basinward into the lower part of the Van Hauen Formation, which consists mostly of siliceous shale.

The facies spectrum displayed by the Sabine Bay clastic wedges suggests that these were deposited as independent deltaic complexes in various areas of the Sverdrup Basin. These deltas likely formed in response to local uplifts associated with the Melvillian Disturbance, which shed vast amounts of detritus into the basin over a relatively short time interval. Attesting for these inferred high depositional rates is the absence of glauconite within the Sabine Bay marine sandstones, a feature that contrasts markedly with the high glauconitic content of the overlying Assistance and Trold Fiord formations.

SEQUENCE 6: ROADIAN-WORDIAN

Sequence 6 comprises sediments ranging in age from early Roadian to late Wordian, and encompasses more than 5 Ma of depositional history. These sediments are contained within the uppermost Sabine Bay Formation, the whole Assistance Formation, and parts of the Trold Fiord, Degerböls and Van Hauen formations. "Unit-A" (Nassichuk, 1965), on Melville Island, is also part of that sequence. In places, Sequence 6 is represented by more than 700 m of strata, which can be subdivided into two successions: a

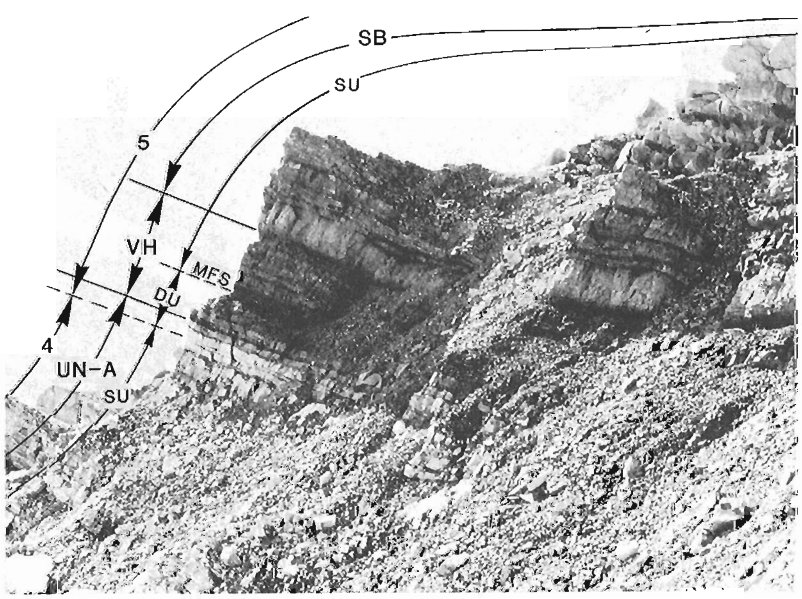

Figure 3. Boundary between Sequences 4 and 5, on north side of East Cape River, Hamilton Peninsula, west-central Ellesmere Island. Uppermost "unnamed" A (UN-A), Van Hauen (VH), and lower Sabine Bay (SB) formations are shown. SU is shallowing-upward, DU is deepening-upward, MFS is maximum flooding surface.

basal, deepening-upward succession comprising the Assistance Formation and correlative part of the Van Hauen Formation, and an upper, shallowing-upward succession containing the bulk of the Trold Fiord and Degerböls formations, and correlative parts of the Van Hauen Formation. The contact between the Assistance and Trold Fiord formations is, in vertical sections, placed at the base of a prominent shale unit. This unit corresponds approximately to the maximum of the transgression, below which the succession is deepening-upward (Assistance Formation), and above which it is substantially more glauconitic and shallowing-upward (Trold Fiord Formation).

In the marginal areas, the base of Sequence 6 is an unconformity, which has not yet been substantiated by fossil evidence, because the underlying Sabine Bay sandstones are, in most areas, devoid of any biogenic components. However, exposures in the Trold Fiord area show gently inclined Sabine Bay sand sheets truncated by erosion, and then onlapped by the transgressive uppermost Sabine Bay Formation and Assistance Formation (Fig. 4). Toward the basin centre, the transition between Sequences 5 and 6 is conformable.

Transgression

The transgression is represented by the uppermost Sabine Bay Formation and the Assistance Formation, which display a number of higher-order cycles. In vertical sections, the lithological contents of these cycles show features successively indicative of nearshore (Sabine Bay), inner shelf and outer shelf (Assistance) environments. Overall, the Assistance Formation comprises variably glauconitic, argillaceous, calcareous sandstones that are extremely fossiliferous at the base, with a profusion of brachiopods of various kinds; ramose, tabular and fenestellid bryozoans; and subordinate crinoids. The fossil content decreases upward, reflecting the increase in water depth. The upper part of the formation is dominated by intensively bioturbated sandstones and siltstones displaying abundant *Zoophycos*. Approximately seven higher-order cycles can be recognized in the Assistance Formation in certain areas, but this number decreases toward the basin margin, reflecting the onlap of successively younger strata over Sequence 5. In the deep basin, Assistance-equivalent strata belong to the Van Hauen Formation, and are probably represented by a condensed section, as clastic sediments were trapped near the basin margin by the rising sea level.

Regression

Above the maximum flooding surface occurs a glauconitic, calcareous shale that progressively passes into a coarsening-, thickening- and shallowing-upward, highly glauconitic sandstone, which also becomes more fossiliferous upward. These sediments belong to the Trold Fiord Formation, which is well developed in the marginal areas of Sverdrup Basin. The Trold Fiord Formation comprises five or six higher-order cycles, the thickness of which decreases upward. Each cycle comprises outer to inner shelf sediments. Some cycles are capped by laterally extensive beach conglomerates and conglomeratic sandstones. Biota of the Trold Fiord Formation include a wide variety of brachiopods, and tabular, fenestellid, bulbous and ramose bryozoans, associated with minor echinoderms. Glauconite is pervasive in the Trold Fiord Formation, either in the form of small oval pellets, or cements within various organic cavities. The presence of abundant glauconite suggests a slow rate of sedimentation, associated with cold water conditions.

On Melville Island, the first high-order cycle above the maximum flooding surface of Sequence 6 is carbonate-dominated, and has been referred to as "Unit-A" (Nassichuk, 1965). Sedimentation of "Unit-A" reflects a time when clastic sediments were trapped near the shoreline by the transgressing sea and possible shoaling structures, leading to carbonate deposition on the shelf. "Unit-A" is stratigraphically equivalent to the lower Trold Fiord Formation elsewhere in Sverdrup Basin.

The Trold Fiord Formation grades basinward into a succession comprising black, siliceous shale at the base, passing upward into spiculitic chert and then into a fossiliferous

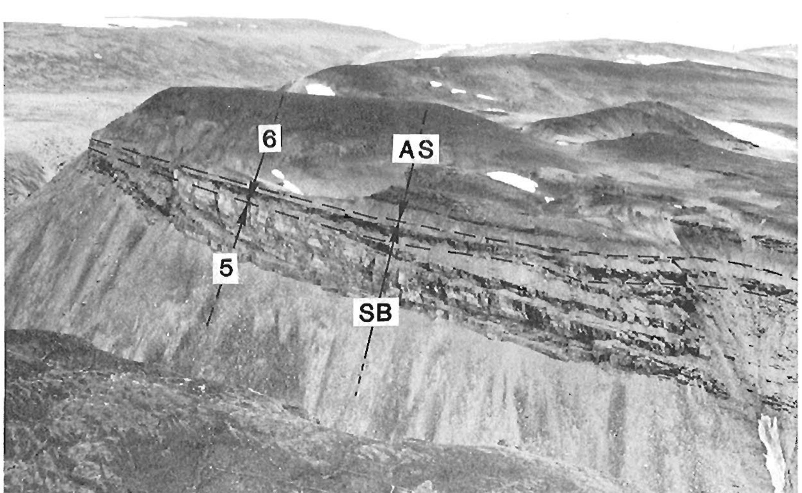

Figure 4. Apparent toplap surface marking the boundary between Sequences 5 and 6, west of head of Trold Fiord, Raanes Peninsula, southwestern Ellesmere Island. Note dip difference between beds of Sequences 5 and 6. SB is Sabine Bay Formation and AS is Assistance Formation.

limestone. The shale and chert belong to the Van Hauen Formation, the limestone to the Degerböls Formation. Crinoids, brachiopods and bryozoans are the main biotic components of the Degerböls, which represents shelf sedimentation in areas of the Sverdrup Basin where the clastic influx was reduced to a minimum. The Degerböls Formation grades basinward, first into a spiculitic chert, and then into a siliceous shale. Both the shale and the chert belong to the upper part of the Van Hauen Formation. No carbonate buildups are known to occur in the Degerböls Formation, providing additional evidence for a cold climate in Late Permian time.

SEQUENCE 7: POST-WORDIAN(?)

At the margin of the Sverdrup Basin, Griesbachian strata (lowest Triassic) rest disconformably on Wordian strata (lowest Upper Permian), indicating a major sub-Triassic erosional interval. However, in the more distal areas, an unfossiliferous succession occurs between dated Triassic and dated Upper Permian rocks, raising the possibility that parts of the post-Wordian succession may be preserved (Fig. 2, 5). In the axial regions of the basin, there is evidence of continuous sedimentation as deep water fissile shales of the Triassic Blind Fiord Formation rest upon siliceous shale of the Upper Permian Van Hauen Formation. This subtle lithological difference is reflected in the presence of siliceous sponge spicules in the Van Hauen shale, and the lack of spicules in the Blind Fiord shale, but it does not indicate a break in sedimentation. This contradicts the widely accepted belief that the Permian-Triassic boundary is a major erosional unconformity throughout most of the world, except in China and Iran (Tozer, 1979).

The condensed succession, referred to as Sequence 7, is no more than 70 m thick, but may represent as much as 8 Ma of depositional history. A post-Wordian age (Capitanian, Dzhulfian or Changhsingian) has not yet been been substantiated by fossil evidence, as these rocks are characteristically unfossiliferous, except for siliceous sponge spicules that occur in large quantities. Because of the lack of age control, it is impossible to demonstrate a major unconformity between these sediments and rocks of Sequence 6. Nevertheless, we infer that the post-Wordian(?) succession is an independant third-order sequence of the Sverdrup Basin, because of: 1. the time span that may be involved, as suggested by the lack of a sedimentary break in the deep basin; 2. the deepening-upward nature of its basal sediments, and 3. the increase in thickness represented by the transgressive-regressive package (accommodation), indicating that post-Wordian(?) sediments are not part of a higher-order cycle of Sequence 6.

Near the basin margin, post-Wordian(?) sediments belong to the uppermost Trold Fiord Formation (Fig. 2). They comprise bright green, highly glauconitic, poorly consolidated, unfossiliferous sand, associated with rare *Zoophycos*. These pass laterally into a light-coloured chert, which contains a profusion of sponge spicules, with unidentified trace fossils. The chert is contained in the upper part of the Degerböls Formation (Fig. 5). It grades basinward into the uppermost Van Hauen Formation, represented by dark-coloured chert near the transition with the Degerböls Formation, and by siliceous black shale in the more distal areas.

The absence of fossils other than sponge spicules in Sequence 7 and the condensed nature of the succession cannot be readily explained. The post-Wordian(?) time interval was undoubtedly characterized by extremely low clastic influx in the basin, and by environmental conditions unfavourable to the development of any shelly fossils. As seen earlier, the Permian biological associations of the Sverdrup Basin indicate a trend toward colder climatic conditions, but this alone cannot explain the disappearance of shelly fossils. Indeed, fossil-rich deposits are known to interfinger with glaciogenic sediments that formed at a paleolatitude as high as 80° during the Early Permian in the Southern Hemisphere (Rao and Green, 1982).

It is possible that, in addition to the stress caused by the cold climatic setting, the well documented latest Permian worldwide sea level drop (Hallam, 1984; Schopf, 1974) might have led to the partial closure of Sverdrup Basin, and

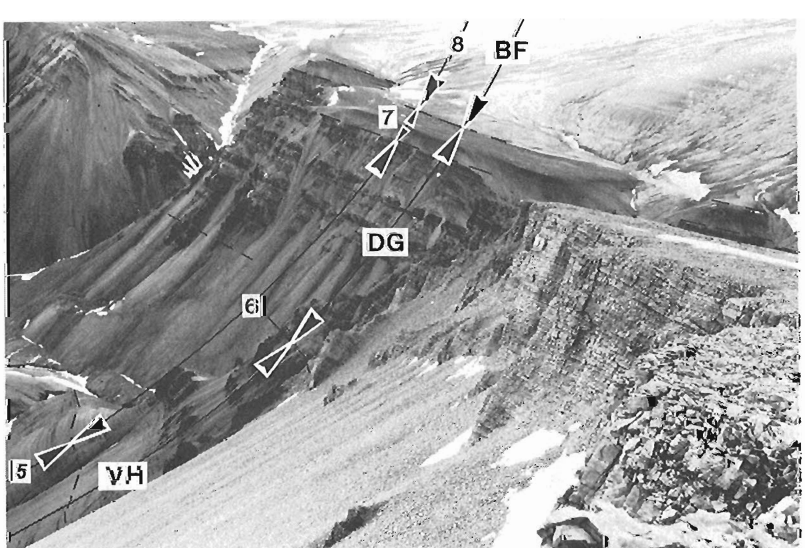

Figure 5. Permian-Triassic transition on west side of Blind Fiord, Raanes Peninsula, southwestern Ellesmere Island. Formations shown are Van Hauen (VH), Degerböls (DG), and Blind Fiord (BF). Sequences 5 (part), 6, 7 and 8 (part) are shown. Note condensed nature of Sequence 7. Sequence 8 is Triassic.

development of brackish water conditions. Episodic closure of Sverdrup Basin occurred early, during, the Late Carboniferous, as attested by the development of Otto Fiord basinal evaporites. It is plausible that similar events affected the Sverdrup Basin in latest Permian time, but because of an environmental setting in which precipitation exceeded evaporation, a brackish lake, instead of an evaporative basin, could have formed. During that time, the Sverdrup Basin was never completely nonmarine, as is evident from the high glauconite content and occasional *Zoophycos*, but salinity might have dropped substantially, leading to the eradication of all shelly animals, whose development was already stressed by cold climatic conditions. Accordingly, the last shelly animals present in these Upper Permian rocks are dominated by *Lingula*, an inarticulate brachiopod with greater tolerance to reduced salinities than the productid and spiriferid articulate brachiopods that dominate most assemblages in Sequence 6 (McKerrow, 1978; Dodd and Stanton, 1981; Emig, 1981).

CONCLUSIONS

Seven long-term (5 to 50 Ma) transgressive-regressive sequences characterize the Carboniferous and Permian succession of Sverdrup Basin. These resulted from second- and third-order relative sea level fluctuations of probable tectonic origin. Each sequence is bounded at the basin margin by significant unconformities passing basinward into equivalent conformities. The Viséan Sequence 1 is nonmarine, representing lacustrine sedimentation. The Serpukhovian to post-Wordian(?) Sequences 2 to 7 display deepening-upward/shallowing-upward couplets, deposited in various shelf to basin environments.

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Structural transects in northwestern Ellesmere Island, Canadian Arctic Archipelago[†]

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Abstract

Recent field studies near Yelverton Bay, Ellesmere Island, have produced new information about the geometry and relative ages of meso- and macroscopic structures in the Proterozoic through lower Paleozoic rocks of a possible accreted terrane (Pearya). In two of the areas studied — one south of Kulutingwak Fiord, the other east of Yelverton Inlet — Upper Proterozoic through Lower Silurian rocks are incorporated in south directed folds and thrusts. Large- and small-scale structures record several approximately coaxial phases of compressional deformation. The ages of these phases are not well constrained, and it is not yet clear whether the overprinted structures record a single period of progressive deformation or represent distinct tectonic events. Although the first-order structures in the region probably developed during middle to late Paleozoic time, deformed Cretaceous and lower Tertiary rocks on Wotton Peninsula indicate that the Tertiary Eurekan Orogeny did affect northernmost Ellesmere. Remarkable similarities between Proterozoic metasedimentary rocks in the Yelverton area and a coeval sequence in southwest Spitsbergen are consistent with the hypothesis that Pearya was associated with the northern Caledonides.

Résumé

De récentes études menées sur le terrain près de la baie Yelverton, dans l'île d'Ellesmere, ont fourni de nouveaux renseignements sur la géométrie et les âges relatifs de structures méso- et macroscopiques dans les roches du Protérozoïque au Paléozoïque inférieur d'un terrain (Pearya) peut-être formé par alluvionnement. Dans deux des zones étudiées, une au sud du fiord Kulutingwak, l'autre à l' est de l'inlet Yelverton, des roches du Protérozoïque supérieur au Silurien inférieur sont incorporées dans des failles et un chevauchement de direction sud. Des structures de petite et grande envergure témoignent de plusieurs phases de déformation par compression à peu près coaxiales. Les âges de ces phases ne sont pas bien définis, et il n'est pas encore certain si les structures en surimpression témoignent d'une seule période de déformation progressive ou si elles représentent des événements tectoniques distincts. Même si dans la région les structures de premier ordre se sont probablement formées pendant le Paléozoïque moyen à supérieur, des roches déformées du Crétacé et du Tertiaire inférieur de la péninsule de Watton indiquent que l'orogenèse d'Eureka au Tertiaire n'a pas touché l'extrémité nord de l'île d'Ellesmere. D'étonnantes ressemblances entre les roches métasédimentaires du Protérozoïque dans la région d'Yelverton et une séquence contemporaine dans le sud-ouest du Spitzberg confirment l'hypothèse selon laquelle Pearya est associé à la région nord des Calédonides.

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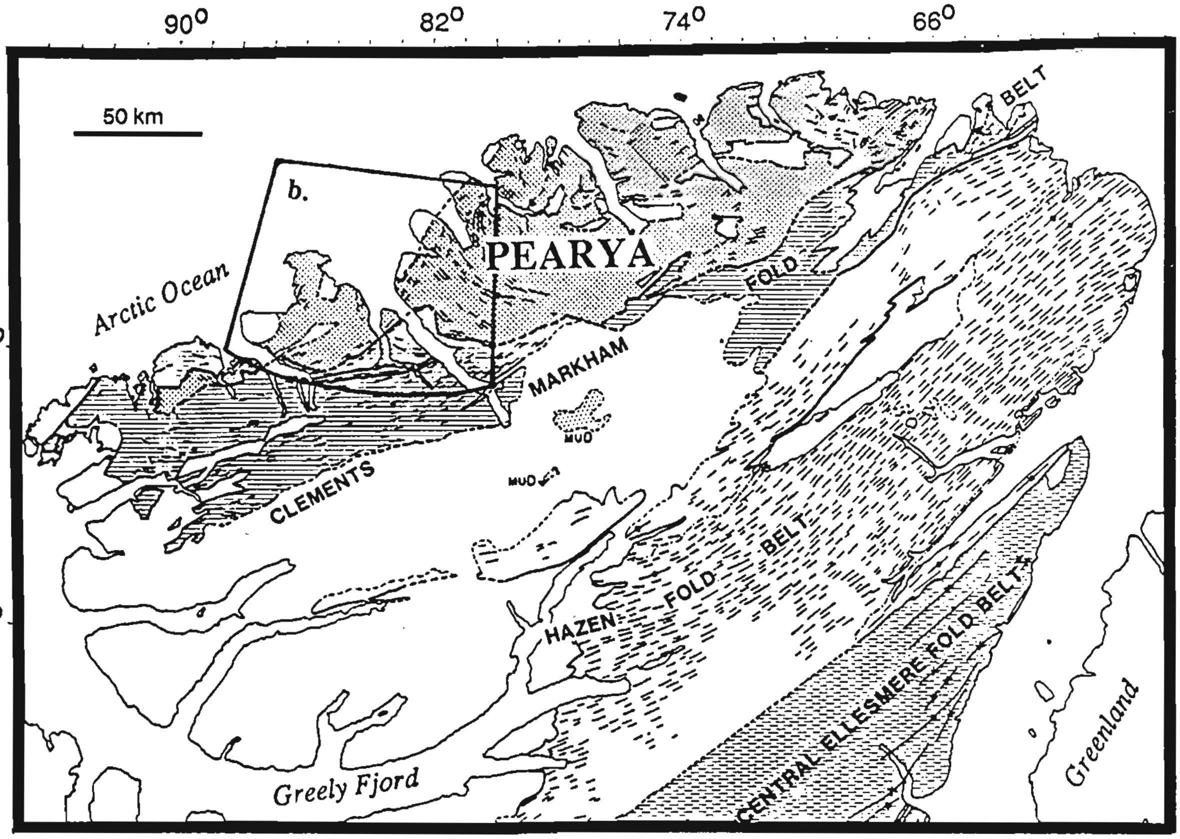
INTRODUCTION

This report presents preliminary results of 1988 structural field studies in the Yelverton Inlet-Kulutingwak Fiord area of northwestern Ellesmere Island (Fig. 1). These investigations focused on structurally problematic areas within "Pearya", a possible exotic terrane of Upper Proterozoic and lower Paleozoic rocks, identified by Churkin and Trexler (1980) and Trettin (1987).

The region defined as Pearya (Fig. 1) appears to have evolved separately from the rest of the Canadian Arctic prior to Late Silurian time. According to Trettin (1987), characteristics that set Pearya apart include:

1. Basement-complex metamorphic ages of 1.0 to 1.1 Ga, in contrast to Archean and Early Proterozoic ages elsewhere in the Arctic Archipelago.
2. A distinctive uppermost Proterozoic-lowermost Cambrian sequence that includes glaciogenic (?) diamictites and that has no clear equivalent in neighbouring parts of the Arctic Islands.
3. Evidence for a Middle Ordovician tectonic event, the M'Clintock Orogeny, recorded by an angular unconformity between Lower and Upper Ordovician strata in Pearya, but not elsewhere on Ellesmere Island or the adjacent islands.

a.



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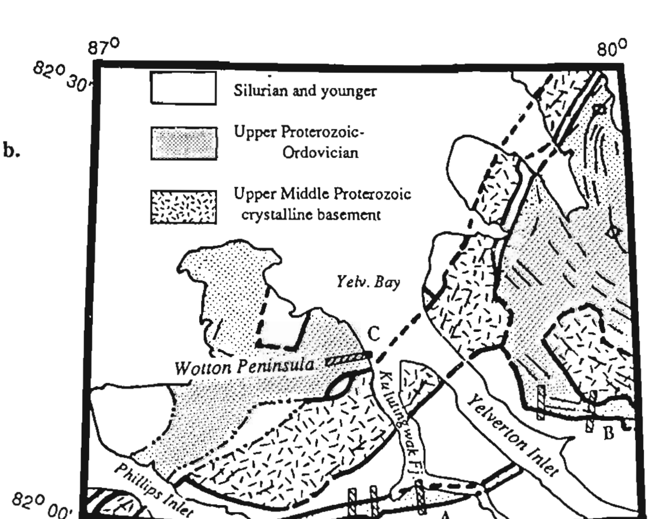

Figure 1. Geological setting of the study areas. (from Trettin, 1987.)

a. Structural provinces of northern Ellesmere Island. Pearya: Middle Proterozoic through Silurian rocks distinct from those to the south. Clements Markham, Hazen, and Central Ellesmere fold belts: Cambrian through Lower Devonian rocks. Blank areas indicate Carboniferous and younger rocks; muD signifies Middle-Upper Devonian.

b. Locations of transects studied (shaded rectangles).

These characteristics and others suggest that Pearya had close pre-Silurian ties with the northern Caledonides, particularly the Svalbard Archipelago (Trettin, 1987). Based on his detailed stratigraphic studies in northern Ellesmere Island and on a tectonic hypothesis for Caledonian Svalbard, Trettin (1987) proposed that the structurally complex Pearya Terrane was accreted to North America in Late Silurian (Ludlow) time by large-scale, sinistral, strike-slip motion along east to northeast trending faults.

Trettin's model has provided a conceptual framework for geological studies in northern Ellesmere Island. Testing this model was one of the principal purposes of the field investigations described in this paper.

OBSERVATIONS

Detailed structural observations were made in three areas identified by Trettin (pers. comm., 1988) as potential keys to understanding the geological evolution of northern Ellesmere Island. These areas, previously mapped at a scale of 1:125 000 (Trettin and Frisch, 1987) were:

1. An east-west trending zone south of Kulutingwak Fiord (Fig. 1b, area A).
2. An area on the east side of Yelverton Inlet (area B).
3. The eastern part of Wootton Peninsula (area C).

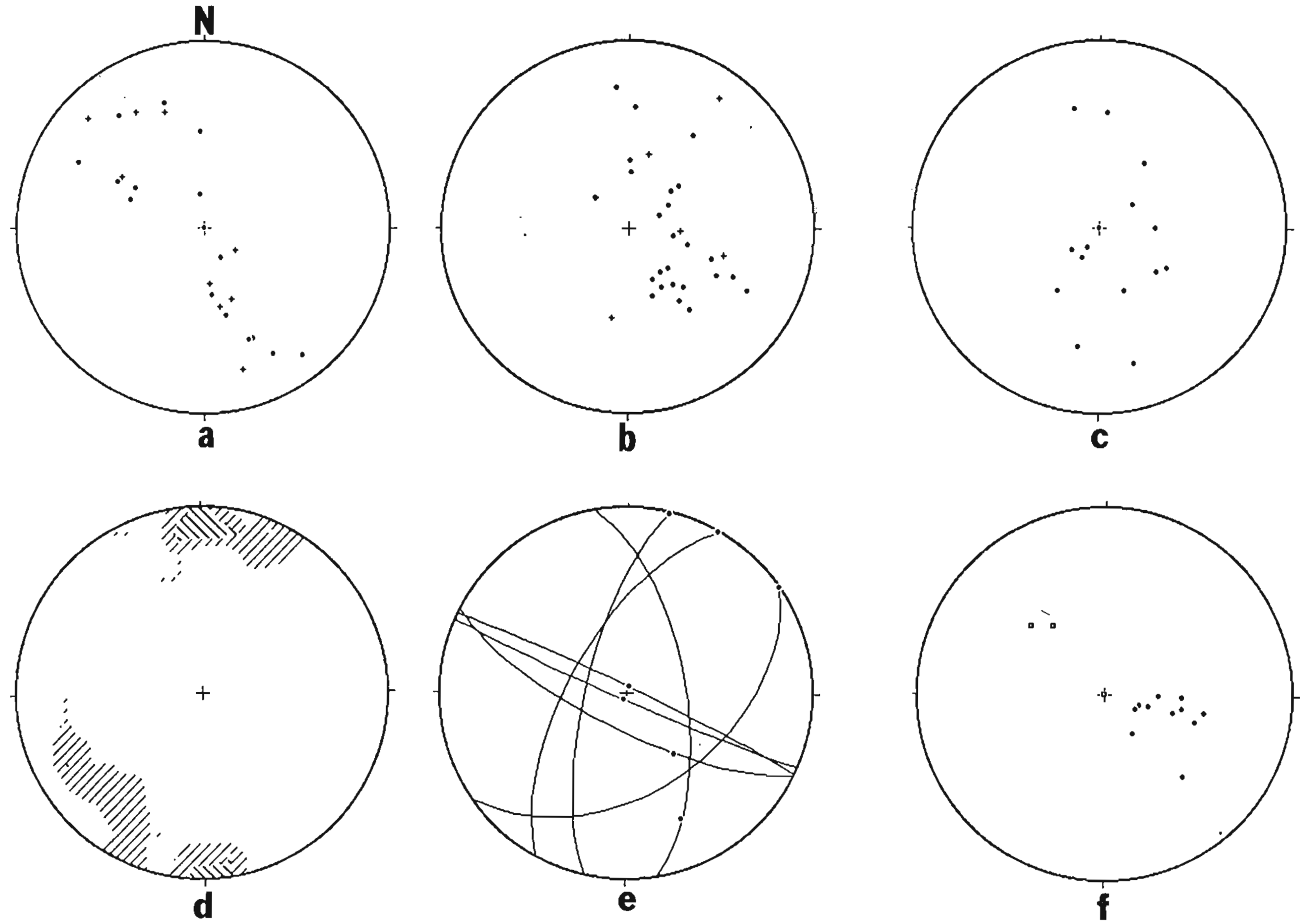

Figure 2. Selected structural data from the three study areas (lower hemisphere equal area projections).

a, b. Poles to bedding planes measured in the western and central Kulutingwak transects, respectively. Dots: carbonate/schist and diamictite; crosses: Imina Formation.

c. Poles to spaced cleavage in the Kulutingwak belt.

d. Contoured plot of poles to bedding planes, east side of Yelverton Inlet (54 observations; 2% contour interval). Most of the northwest striking beds were measured near the north-northwest striking fault immediately east of the inlet.

e. Slip surfaces (great circles) and slickensides (dots) measured east of Yelverton Inlet.

f. Poles to bedding planes, Wootton Peninsula. Dots. Upper Proterozoic(?) metasedimentary rocks; boxes: lower Tertiary Eureka Sound Group.

The 1988 work in the first two areas (Kulutingwak and Yelverton) produced new information about the Late Proterozoic-early Paleozoic history of the region, and observations from the third area (Wootton) shed light on Tertiary (Eurekan) deformation in northern Ellesmere.

Kulutingwak structural zone

The Kulutingwak structure is an east to east-northeast trending zone that extends approximately 55 km from the east side of Phillips Inlet to the west side of Yelverton Inlet (Fig. 1b). Although the Kulutingwak zone is less than 8 km wide, at least four distinct rock sequences occur within it in a complex array of thrust-bounded slices:

1. The Lower Silurian (Llandovery) Imina Formation, a calcareous turbidite sequence with clearly visible cross-beds and graded beds.
2. An undated carbonate and schist sequence, older than the Imina Formation. Crinoids found in the carbonate in 1988 established for the first time that the sequence is post-Proterozoic.
3. Diamictite, possibly Upper Proterozoic.
4. Serpentinite, perhaps an ophiolite fragment, of unknown age.

The metamorphic grade of the sedimentary rocks is probably no higher than middle greenschist facies. With the possible exception of the Imina Formation and the crinoidal carbonate sequence, which appear to be in stratigraphic contact, the four rock sequences are juxtaposed by faults. Most of these faults dip moderately to steeply northward, and the asymmetry of slickensides and small-scale folds within the fault zones consistently indicates south directed (reverse) displacement. Poles to bedding planes in the various sequences are distributed about subhorizontal east to east-northeast trending axes (Fig. 2a, b).

The geometry of the Kulutingwak structure seems to be as complex along strike as in cross-section. Individual faults are difficult to trace longitudinally, indicating that thrust slices are probably bounded by lateral ramps. Serpentinite is exposed only in the central part of the structure, where it occurs as laterally discontinuous bodies in apparent fault contact with the diamictites. Locally, the serpentinite has a distinctive fragmental texture, in which millimetre- to decimetre-sized pieces of serpentinite — ranging from rounded to angular — are enclosed in a serpentinite matrix. While such afabric could be sedimentary (Trettin, 1987), a tectonic origin seems more plausible considering the localized occurrence of the serpentinite along fault zones and the extremely low yield strength of the rock. In any case, the juxtaposition of possibly oceanic serpentinite with the other rock types in the Kulutingwak structure suggests that the area is the site of deep-seated thrusts, which may mark a fundamental tectonic boundary.

Despite along-strike structural variations in the Kulutingwak zone, several general statements can be made about three north-south transects studied within the zone. First, macroscopic structural relationships across the zone point to two distinct generations of south directed faults: 1. an early set of low-angle thrusts marked by bedding-parallel zones of mylonitization; and 2. a later set of steeply dipping faults that offset the older slip surfaces. The dip of the later faults was probably increased by subsequent folding. A natural cross-section through the western end of the Kulutingwak belt shows the relationship between these two sets of faults (Fig. 3).

The mesoscopic structural fabric of the rocks exposed in the Kulutingwak zone also indicates multiple phases of tectonism. A mylonitic, bedding-subparallel foliation occurs in the crinoid-bearing carbonate/schist sequence and the Upper Proterozoic(?) diamictites. The foliation surfaces are clearly shear planes. In the diamictite, recrystallized tails around clasts are asymmetric with respect to bounding foliation planes, and in the carbonate, the original layering is locally transposed along the foliation surfaces (Fig. 4a). Although the foliations in the two rock sequences are geometrically and mechanically similar, it is possible that the diamictite foliation is an older, perhaps reactivated fabric. A comparable bedding-parallel foliation occurs only in sand-free intervals of the turbiditic Imina Formation, even though the Imina seems to be in stratigraphic continuity with the more pervasively foliated carbonate/schist sequence. The Imina Formation, the carbonate/schist sequence, and

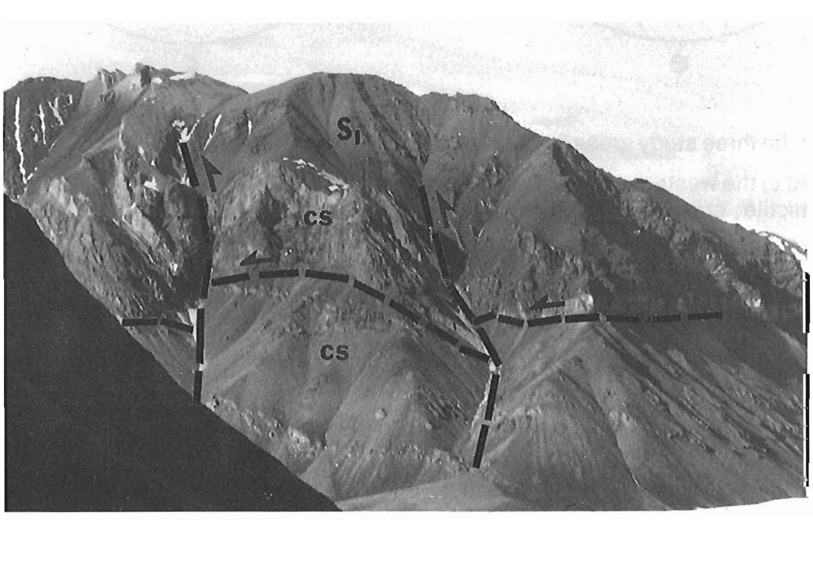

Figure 3. View to west of central part of Kulutingwak structure, showing relationships between early bedding-subparallel faults and later steeply dipping faults. Rocks exposed in this transect are the Imina Formation (Sl) and the lower Paleozoic carbonate-schist sequence (cs).

the diamictites all exhibit a spaced cleavage that crosscuts the shear foliation and typically lies at a large angle to bedding (Fig. 4b). This cleavage, in turn, appears to have been folded about a horizontal, east-west trending axis (Fig. 2c).

In summary, meso- and macroscopic structural evidence from the Kulutingwak zone indicates that the area experienced at least three, nearly coaxial phases of deformation: two phases of south directed thrusting, followed by broad, east-west folding. Obviously, tectonic interpretation of these phases is not possible until their absolute ages are better constrained. The mesoscopic fabric of the diamictites may be unrelated to the present macroscopic structural geometry. The large-scale imbricate thrusts are clearly post-Early Silurian, since they involve the Imina Formation, but the absence of upper Paleozoic strata in this part of northern Ellesmere makes it difficult to place an upper age limit on the deformation. Stratigraphic evidence from outside the immediate area points to a period of tectonism in Ludlow time, which Trettin (1987) interpreted as the accretion of Pearya to the craton. The possibility of Late Devonian Ellesmerian deformation in northernmost Ellesmere should not be dismissed, however.

Tight, east-northeast trending folds in the eastern Hazen fold belt (Fig. 1a) involve Lower Devonian strata and appear to be continuous with structures to the west and north (E. Klaper, pers. comm., 1988). Penetrative deformation during the Tertiary Eurekan Orogeny can probably be ruled out, since Cretaceous(?) mafic dikes observed in the Kulutingwak zone show no secondary tectonic fabric. Conodont analysis of the crinoidal carbonate and $^{39}Ar/^{40}Ar$ age determinations from foliated specimens should provide much-needed geochronological information.

East side of Yelverton Inlet

Like the Kulutingwak belt, the area east of Yelverton Inlet (Fig. 1b) is a zone of south directed thrusts and folds. Three ''packages'' of rock occur as a north dipping imbricate stack. From south to north, these are:

1. The Lower Silurian Imina Formation.
2. An Upper Proterozoic(?) metasedimentary sequence of siliceous carbonates, quartzite, diamicite, and conglomerate.

Figure 4. Secondary foliations in rocks of the Kulutingwak zone.

a. Shear foliation (horizontal in photograph) in lower Paleozoic dolomite. Arrow points to transposed bedding.

b. Spaced cleavage, refracted in graded bedding of Lower Silurian Imina Formation.

3. Feldspathic gneiss that has yielded U-Pb and Rb-Sr metamorphic ages between 1.0 and 1.1 Ga (Trettin et al., 1987).

The metasedimentary sequence is assigned to the Upper Proterozoic on the basis of its diamictite strata and its apparently unconformable contact with the gneiss basement rocks (Trettin, 1987). The Yelverton sequence begins with gritty, siliceous dolomite, overlain by dark, silty limestone, diamictite and conglomerate with brownish to greenish matrix, green quartzite, sandy and cherty dolomite, and, finally, sedimentary carbonate breccia. This stratigraphic sequence is virtually identical to a metasedimentary succession that lies above a profound mid- to Late Proterozoic unconformity in Wedel Jarlsberg Land, southwest Spitsbergen (Bjornerud, 1987). A distinctive feature of the Spitsbergen diamictites is the abundance of oolitic, oncolitic, and stromatolitic dolomite clasts in certain strata. On Wootton Peninsula, where fragments of the Yelverton sequence occur (see below), many such clasts were found in brown-matrix diamictite. This apparent Proterozoic link between northern Ellesmere Island and southwest Spitsbergen supports Trettin's (1987) concept of Pearya as a Caledonian terrane, but the full significance of the Ellesmere-Svalbard link is not yet clear.

Bedding planes in the Imina Formation and the Proterozoic metasedimentary sequence east of Yelverton Inlet generally dip steeply to the north or to the south (Fig. 2d). Along the north dipping thrust contact between the Proterozoic sequence and the Imina Formation, Imina strata are generally overturned. The west trending trace is offset by steep, north striking fault zones — apparent "tear" faults that accommodated differential movement within the thrust sheet. These secondary strike-slip faults have a consistent spacing of 2 to 3 km, and they control the positions of topographic valleys in the area. A major north-northwest striking fault between the Imina Formation and the Proterozoic sequence immediately east of Yelverton Inlet (Fig. 1b) may be another of these strike-slip tear faults. The anomalous northwesterly strike of bedding planes close to this fault (Fig. 2d) is consistent with drag along a dextral strike-slip fault. Mesoscopic observations of slickensides bear out the relationship between dip- and strike-slip faults in this area. West to west-northwest striking slip surfaces generally have dip-direction slickensides, while north to north-northeast striking surfaces have strike-parallel slickensides (Fig. 2e).

Although the macroscopic structural geometries of the Imina Formation and the Proterozoic sequence are similar, the mesoscopic fabrics of the two sequences are quite different. The Imina Formation, which is coarse grained and locally conglomeratic in this area, is foliated only close to its thrust contact with the Proterozoic rocks. Small- scale folds were rarely observed in the Imina Formation, but younging-direction reversals within almost homoclinally dipping strata indicate the presence of tight mesoscopic folds. Vertical kink bands are common in the Imina Formation, especially in the vicinity of north striking strike-slip faults.

The Proterozoic sequence has a penetrative foliation that is typically more visible than bedding. The foliation is axial planar to isoclinal folds with similar-fold geometry, and clasts in the diamictite and conglomerate, strongly oblate, lie in the plane of this foliation. Long (1 to 3 cm), acicular crystals of a tremolitic amphibole are common in sandy and cherty dolomite strata in the upper part of the Proterozoic sequence. Interestingly, these crystals have no systematic orientation, even where the secondary foliation is well developed. On the west side of Yelverton Inlet, Y. Ohta (pers. comm., 1988) has observed the same phenomenon in a lower Paleozoic(?) metasedimentary sequence called the Peterson Bay assemblage (Trettin, 1987). There, randomly oriented crystals of mafic amphibole occur in a strongly foliated biotite schist. These observations point to a protracted tectonic history in which thermal equilibration occurred long after deformation. As in the Kulutingwak belt, however, reconstructing this tectonic history is not possible without better geochronological constraints.

Wootton Peninsula

In contrast to the dominantly west trending and south verging folds and thrusts in the Yelverton and Kulutingwak areas, first-order structures on eastern Wootton Peninsula (Fig. 1b) strike north-northeast and verge east-southeast. Explaining this anomalous orientation was a principal objective of the 1988 field studies on the peninsula, and the new work led to some unexpected conclusions.

Most of the area is underlain by metasedimentary rocks (dolomite, diamictite, quartzite) which are probably part of the same Upper Proterozoic(?) sequence exposed east of Yelverton Inlet. On the extreme eastern side of the peninsula, however, a thick felsic volcanic sequence is exposed beneath a steeply west dipping contact with metasedimentary rocks. Previously, the relationship between the two groups of rocks was unclear, and the volcanic rocks were thought to be interstratified with the Proterozoic dolomite/diamicite/quartzite sequence (Trettin and Frisch, 1987). Now it has been observed that, in places, volcanic material crosscuts foliation surfaces in the dolomite, indicating that the volcanic rocks are younger, and possibly much younger, than the metasedimentary rocks. It seems most plausible, in fact, that the volcanic rocks on eastern Wootton Peninsula are part of the Cretaceous Hansen Point volcanics, exposed to the south and to the east across Yelverton Bay (Trettin, pers. comm., 1988).

If the volcanic rocks on Wootton Peninsula are Cretaceous, then the anomalous north-northeast trending structures there must be Eurekan, since the volcanics are either overthrust by or folded with the metasedimentary sequence. On the northern end of Wootton Peninsula, moreover, steeply dipping beds of the lower Tertiary Eureka Sound Group provide evidence for some folding or tilting during the Eurekan Orogeny (Figs. 2f, 5; cf. Ricketts, 1987).

These observations raise questions about the extent to which Eurekan deformation affected the rest of northern Ellesmere Island. Even if the Tertiary event did not cause penetrative deformation of the Silurian and older rocks, Tertiary reactivation of old fault zones seems probable. Tertiary tectonism must be at least partly responsible for present outcrop patterns. New analysis of stratigraphic trends in the Carboniferous- Cretaceous Sverdrup Basin indicate that

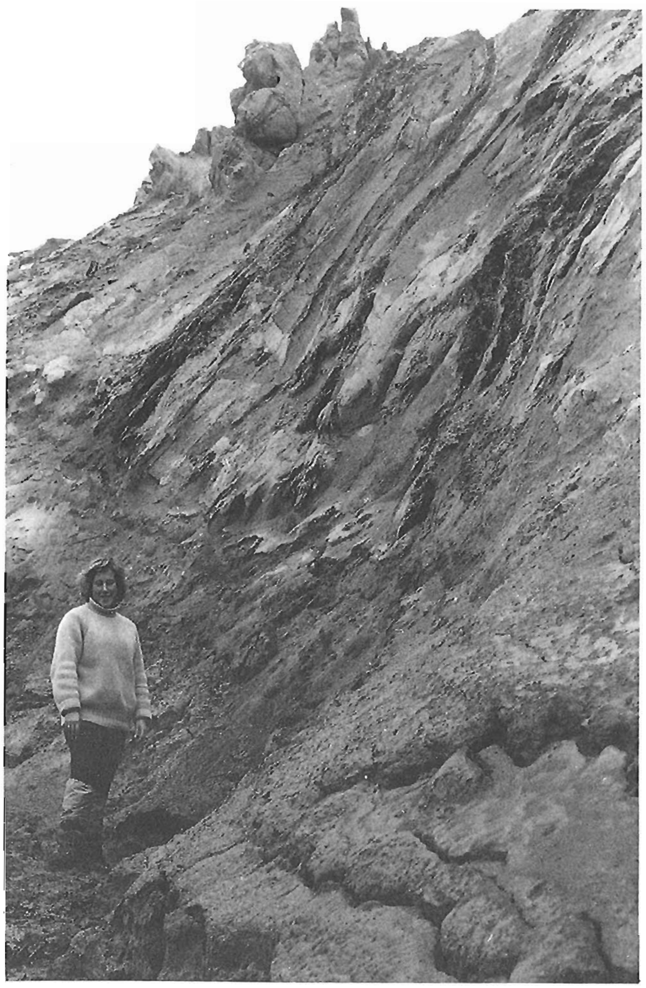

Figure 5. Steeply dipping beds of sandstone and shale, lower Tertiary Eureka Sound Group, northern Wootton Peninsula.

Sverdrup Basin strata — perhaps 7 km thick — may have covered northern Ellesmere Island (L. Maurel, pers. comm., 1988). If so, the present exposure of pre-Devonian rocks in northern Ellesmere requires that the Sverdrup Basin sediments were somehow uplifted and removed during Tertiary time.

CONCLUSIONS

Although it is too early to make comprehensive tectonic interpretations of the field observations presented here, the following conclusions can be drawn from the fragmental information now available:

1. The Kulutingwak structural belt and the east side of Yelverton Inlet can be characterized as zones of south directed folds and imbricate thrusts involving Upper Proterozoic and lower Paleozoic (pre-upper Silurian) rocks. No direct structural evidence for large-scale strike-slip movement (e.g., steeply plunging folds, folds and thrusts arranged *en echelon*) has been observed in the Kulutingwak or Yelverton areas. Overprinted structural features in both areas indicate multiple phases of compressional deformation, but the ages of the various phases are poorly constrained. The problem of linking particular structural features with tectonic events is not trivial. Conceivably, rocks in these areas could record the Middle Ordovician M'Clintock Orogeny, the postulated Late Silurian accretion of the Pearya Terrane (Trettin, 1987), Devonian- Carboniferous Ellesmerian deformation, and Tertiary Eurekan tectonism.
2. Upper Proterozoic(?) metasedimentary rocks, including diamictites, exposed east and west of Yelverton Bay, are strikingly similar to a sequence overlying a major mid- to Late Proterozoic unconformity in southwest Spitsbergen. This apparent correlation supports the hypothesis that the Pearya Terrane of northernmost Ellesmere had Caledonian origins (Trettin, 1987).
3. Folded and faulted Cretaceous and Tertiary rocks on Wootton Peninsula indicate that some Tertiary deformation occurred in northern Ellesmere. The effect of Tertiary tectonism on Silurian and older rocks in the region remains unclear.

This report has been intended only as a summary of field observations. Geochronological, petrographic and microstructural analyses will augment these observations and allow more complete interpretation of them.

ACKNOWLEDGMENTS

I wish to thank H. Trettin, Y. Ohta, E. Klaper and L. Maurel for enlightening discussions in the field. I am especially grateful to Trettin for inviting me to participate in the 1988 expedition to northern Ellesmere.

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The geology, biostratigraphy and organic geochemistry of the Natsek E-56 and Edlok N-56 wells, western Beaufort Sea[†]

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Abstract

The Dome Pacific et al. PEX Natsek E-56 and Dome et al. Edlok N-56 wells, drilled in the southwestern part of the Canadian Beaufort-Mackenzie Basin, penetrated thick sections of Tertiary sedimentary strata. The lithology, microfossil content, geochemical characteristics and reflection seismic expressions of the stratigraphic units penetrated by both wells are described in this report. Important results and interpretations include: the recognition and stratigraphic identification of major unconformities of Early and Middle Eocene age, the introduction of the Taglu and Aklak sequence names (as revisions of previous descriptions of the Beaufort-Mackenzie Basin sequence stratigraphy), descriptions of the nature of the sedimentary fill of the Demarcation sub-basin, and identification of submarine fan deposits in Eocene strata.

Résumé

Les puits PEX Natsek E-56 de Dome Pacific et coll. et Edlok N-56 de Dome *et coll.* forés dans la partie sud-ouest du bassin canadien de la mer de Beaufort et du Mackenzie, ont traversé d'épaisses sections de couches sédimentaires du Tertiaire. La lithologie, le contenu en microfossiles, les caractéristiques géochimiques et la représentation par sismique-réflexion des unités stratigraphiques recoupées par les deux puits sont décrits dans le présent rapport. Ce dernier contient des résultats et des interprétations importants ayant trait à la reconnaissance et l'identification stratigraphique des grandes discordances de l'Éocène, inférieur et moyen, à l'introduction des appellations de séquence Tagl et Aklak (en tant que révision de descriptions antérieures de la stratigraphie des séquences du bassin de la mer de Beaufort et du Mackenzie), aux descriptions de la nature des sédiments de remplissage du sous-bassin de Démarcation et à l'identification de dépôts de cône d'éboulis sous-marins dans les couches de l'Éocène.

[†] Contribution to Frontier Geoscience Program.

INTRODUCTION

The Dome Pacific et al. PEX Natsek E-56 well was drilled during the summers of 1978 and 1979 in the western Beaufort Sea (Lat. 69°45′21.46″N, Long. 139°44′34.58″W) to a total depth of 3520 m below kelly bushing (KB). Water depth at the Natsek location was 33.8 m. The Dome et al. Edlok N-56 well was drilled during the summer of 1985 to a total depth of 2530 m below KB, at a location some 20 km west of the Natsek site (69°45′50.73″N, 140°14′23.43″W) (Fig. 1). Water depth at the Edlok location was 31.5 m. Both wells were abandoned after failing to encounter any hydrocarbon accumulations.

The Natsek and Edlok wells are located in the westernmost portion of the Canadian Beaufort-Mackenzie Basin. The two wells provide important stratigraphic tie points for the seismically-based regional geological interpretations of the Beaufort Sea continental margin, offshore of northern Yukon. The Edlok well also provides the first direct geological information on the nature of the sedimentary fill of the Demarcation sub- basin, a major Upper Eocene to Miocene depocentre located in the Canada-U.S.A. offshore boundary area.

This report presents descriptions of the lithology, microfossil content, geochemical characteristics, and reflection seismic expressions of the stratigraphic units penetrated by both wells. Interpretations of the ages, depositional environments and regional stratigraphic correlations of the various units are also discussed. In addition, a brief description of the structural setting of the Natsek-Edlok area is included.

Cuttings from both wells were analyzed for lithology, microfossil content (foraminifers and palynomorphs), organic matter content and thermal maturation level. The geochemical observations were made from vitrinite reflectance measurements from both wells and Rock-Eval pyrolysis measurements from the Edlok samples. Geophysical well logs (primarily the gamma ray and sonic log curves) were used for identification of lithology, depositional environments (for certain intervals) and unconformities. Seismic reflection data were used to identify sequence boundaries, structural configurations and depositional environments. Several thousand kilometres of reflection seismic profiles from the offshore of northern Yukon were interpreted. The seismic profiles were acquired by various

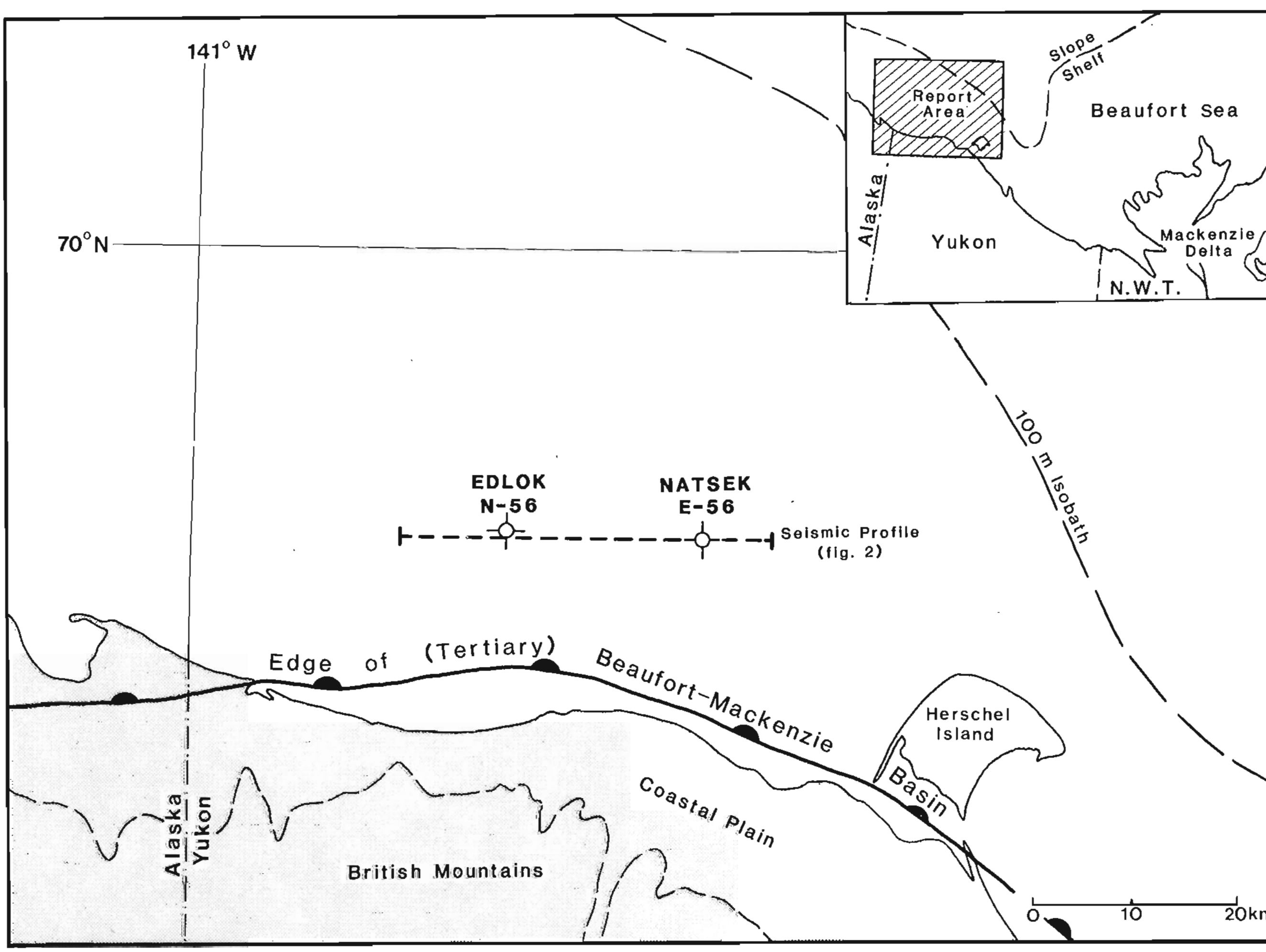


Figure 1. Location of Edlok N-56 and Natsek E-56 wells, western Beaufort Sea, Arctic Canada.

petroleum exploration companies in the 1970s and early 1980s. One east-west oriented seismic section is included in this report to illustrate the sequence identifications and correlations from Natsek to Edlok.

Previous work

The regional geological setting of the western Beaufort Sea area of Arctic Canada has been briefly discussed in numerous papers, including the works of Yorath and Norris (1975), Young et al. (1976), Hea et al. (1980), and McWhae (1986). More detailed accounts of the Cenozoic geology of the Beaufort Sea region include the works of Willumsen and Coté (1982), Young and McNeil (1984), Dietrich et al. (1985) and Dixon et al. (1985). The last two publications include stratigraphic interpretations for the Natsek E-56 well. The Natsek stratigraphy was discussed in terms of the depositional sequence divisions introduced informally in the Dietrich et al. (1985) paper. The depositional/seismic sequence concept is maintained in this report, but the sequence identifications in the Natsek well have been modified from these earlier papers. The sequences identified in the Natsek-Edlok area are discussed in terms of their correlations with the sequences of Dietrich et al. (1985) and the formations of Young and McNeil (1984) from the central Beaufort Sea-Mackenzie Delta area.

STRATIGRAPHY

Six major, unconformity-bounded sequences of Tertiary age have been identified in the Natsek-Edlok area. The sequence identifications are based on regional seismic correlations, biostratigraphy and relative stratigraphic position. A reflection seismic profile across the Natsek and Edlok well locations (Fig. 2) illustrates the seismic expression and correlation of these sequences. The Edlok N-56 well penetrated the Iperk, Kugmallit and Richards sequences and the upper half of the Taglu sequence (Figs. 2, 3). The Natsek E-56 well penetrated the Iperk, Taglu and Aklak sequences and the upper portion of the Fish River(?) sequence (Figs. 2, 4). The Kugmallit and Richards sequences are not preserved at the Natsek location due to sub-Iperk truncation. The Fish River(?) sequence is underlain by highly deformed strata below an unconformity at reflection times of 3.0 to 3.5 seconds (Fig. 2). Strata below this major structural discontinuity and detachment surface may be Boundary Creek (Cenomanian-Turonian) and/or Lower Cretaceous clastics.

The following discussion describes each of the six Tertiary sequences, with particular focus on the geology at the two well locations.

Iperk sequence

The Iperk sequence is a relatively thin, shallow unit extending from the seafloor to a depth of 164 m (below KB) in the Edlok well (Fig. 3) and 215 m in the Natsek well (Fig. 4). The sequence consists of interbedded, unconsolidated gravels, sands, and muds with abundant pyrite and woody debris.

Seismically, the Iperk sequence appears as a thin interval of high amplitude, horizontal reflections truncating inclined reflections in the underlying sequences (Fig 2). The precise location of the base-Iperk truncation surface is commonly obscured by short period multiples in the upper half-second of the section.

In the Natsek well, foraminifers recovered from the sampled portion of the Iperk sequence (Fig. 4) include *Buccella frigida* (Cushman), *Cassidulina teretis* Tappan, *Cibicides grossus* ten Dam and Reinhold (see Plate 4), *Cribroelphidium clavatum* (Cushman), *C. ustulatum* (Todd), *Islandiella norcrossi*(?) (Cushman) and *Haynesina orbiculare* (Brady). The assemblage indicates a Pliocene age and is characteristic of a high latitude, inner shelf environment. Sparse palynomorph assemblages in the Iperk sequence are dominated by pollen of *Picea* and *Pinus*. Pollen of *Alnus, Betula, Salix, Ulmus*, Gramineae and Ericaceae is present, and rare specimens of Caryophyllaceae and Chenopodiaceae pollen occur (Fig. 5). Recycled palynomorphs, especially Upper Cretaceous dinoflagellates, are abundant. The palynofloras suggest a Miocene or Pliocene age and are comparable to those described by Norris (1986).

Kugmallit sequence

The Kugmallit sequence was penetrated by the Edlok well in the interval from 164 to 1115 m (Fig. 3). The sequence is a mudstone-dominant succession containing scattered thin beds of siltstone and fine grained sandstone. Shell fragments and rod- shaped pyrite are abundant, particularly in the upper part of the sequence. Coal-bearing zones were identified in the cuttings from the 620 to 650 m and 1010 to 1045 m intervals. The sonic log across these intervals does not show the typical low velocity signature of coal seams, suggesting that the coal probably occurs in thin beds (<1 m thick).

The Kugmallit sequence is characterized seismically by low to moderate amplitude, parallel reflections that dip 8 to 10° to the northwest, in marked contrast to the horizontal reflections in the Iperk sequence (Fig. 2). West of the Edlok location, low- angle hummocky clinoforms occur in the lowermost portion of the sequence. In the vicinity of the Edlok location, the Kugmallit sequence base is marked by a high amplitude reflection that is generally conformable with underlying and overlying reflections.

Kugmallit strata contain organic matter at very low levels of thermal maturation (Fig. 3). Vitrinite reflectance measurements of .1 to .35 % Ro and Rock-Eval Tmax values of less than 425°C for the entire sequence indicate the immature state of the organic material. A Tmax of 435°C or a vitrinite reflectance of .7 % Ro are values commonly used to mark the maturation level needed for the onset of significant hydrocarbon generation. Isolated Tmax readings of 470+°C in the Kugmallit sequence indicate reworked organic material. Organic carbon content (TOC) in the Kugmallit mudstones averages 1 to 2 %. Samples from the coal-bearing zones have high TOC values, and individual measurements exceed 20 %. Organic matter in the Kugmallit sequence (and indeed in all of the sequences

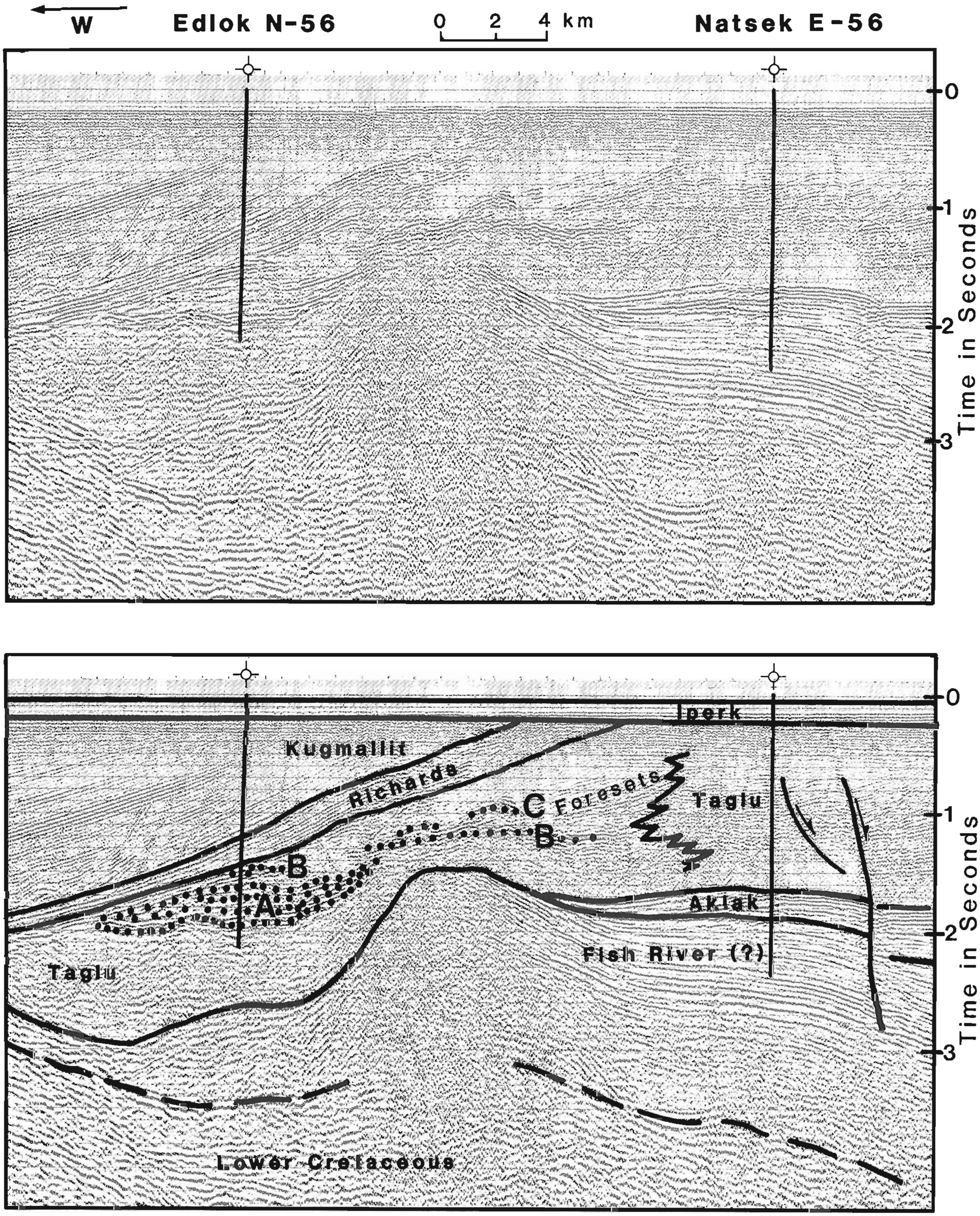


Figure 2. Uninterpreted and interpreted reflection seismic profile across the Edlok and Natsek well locations.

penetrated by the Edlok well) is totally dominated by terrestrial debris and hence Type III kerogen. Hydrogen index values of the measured Edlok samples are low (<200), typical of terrestrial organic material. The nature of the organic matter suggests that (at appropriate maturation levels) Tertiary strata in the Edlok area would have much more gas than oil source potential.

Microfossils in the Kugmallit sequence include a palynomorph assemblage, which spans the 162 to 1329 m interval, and has abundant pollen of *Picea, Pinus*, and Taxodiaceae/Cupressaceae, and lesser amounts of *Tsuga, Alnus, Betula*, Ericaceae, *Ulmus, Pterocarya, Quercus, Fagus, Juglans, Ilex* and *Lonicera* pollen. The presence of rare specimens of Gramineae, Chenopodiaceae, and *Corsinipollenites* sp. (Plate 3), and the absence of *Tilia* and *Carya* pollen suggest that the age is not older than Oligocene. The 162 to 432 m interval of the Kugmallit sequence in the Edlok well (Fig. 3) contains a rich and diverse assemblage of calcareous benthic foraminifers diagnostic of the lower part (*Turrilina alsatica*) of the *Cibicides* spp. assemblage zone (Young and McNeil, 1984) dated as Oligocene (probably Late). The assemblage occurs in the lower Mackenzie Bay and upper Kugmallit formations of the Mackenzie Delta region (Young and McNeil, op. cit.). In the 432 to 770 m interval (Fig. 3), the calcareous, benthic foraminifers are still abundant but diversity is reduced, and *E. brunnescens, Globobulimina* sp. 1300 and several of the miliolids have disappeared. The 432 to 770 m interval is characterized by the occurrence of *Cancris subconicus* (Terquem)(see Plate 4), the first recognition of this species in the Beaufort-Mackenzie Basin. In the North Sea basin, *C. subconicus* occurs in Lower Oligocene to Middle Eocene strata (King, 1983). Its association with *T. alsatica* in the 432 to 770 m interval suggests an Early Oligocene age. Following a general downhole trend, the diversity of the foraminiferal assemblages is reduced in the 770 to 1008 m interval. The first downhole occurrence of *Brizalina* sp. 1435 distinguishes this unit, which is probably Early Oligocene in age. Below a barren zone from 1008 to 1044 m, the lowermost portion of the Kugmallit sequence (1044-1115 m) is poorly fossiliferous. *Bulimina* sp. 1450 and *Rectobolivina* sp. 1464 make their first downhole occurrence but their age significance is unknown.

The mudstone-dominant nature of Kugmallit strata, their abundant foraminiferal content, and seismic expression indicate a marine shelf depositional environment. The presence of two coal-bearing zones in the Edlok well indicate, however, that at least certain sections of the Kugmallit sequence were deposited in marsh-lagoonal environments.

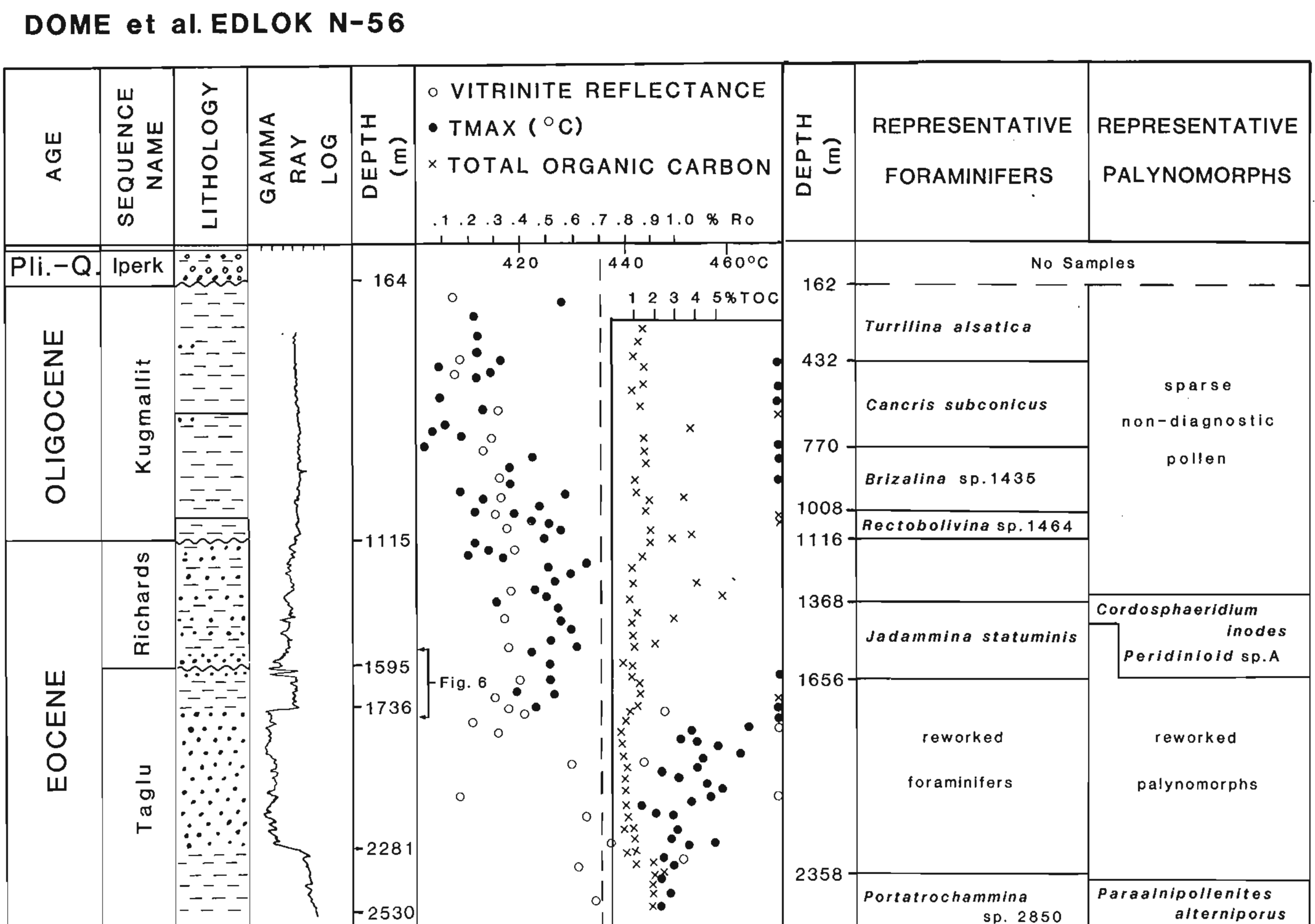


Figure 3. Geological, geochemical and biostratigraphic summary of Edlok N-56 well.

Paleoshorelines may have been close to the Edlok area during Kugmallit deposition. West of the Edlok location, the seismic clinoforms indicate the presence of deeper-water slope deposits in the lower portion of the sequence. The clinoform orientations indicate progradation of paleoshelf margins in a northwesterly direction.

Richards sequence

The Edlok well penetrated Richards sequence strata in the 1115 to 1595 m interval (Fig. 3). The sequence consists of interbedded mudstone and fine to medium grained sandstone. The sandstones are weakly cemented, and porosity averages 20 to 30%. The upper half of the Richards sequence contains a thin bedded mudstone-sandstone succession with no apparent fining- or coarsening-upward trends. The lower half of the sequence (below 1326 m) contains more thick bedded sandstones within small-scale fining-upward and, less commonly, coarsening-upward cycles. The base-Richards unconformity at 1595 m is directly overlain by a 40 m thick fining-upward sandstone succession (Fig. 6).

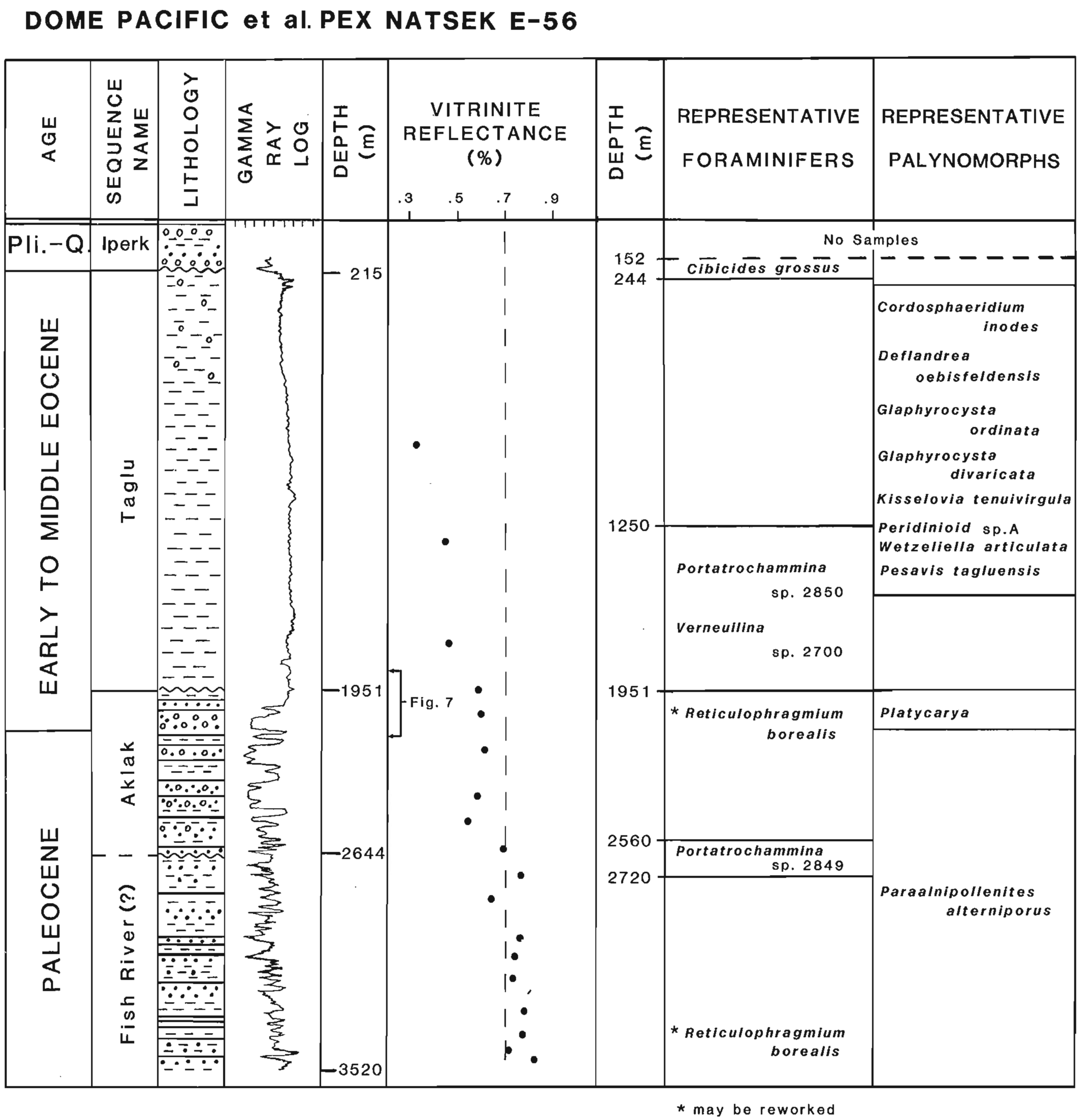


Figure 4. Geological, geochemical and biostratigraphic summary of the Natsek E-56 well.

The Richards sequence appears seismically as a westward tapering wedge of moderate amplitude, subparallel reflections, with higher amplitude reflections developed locally at the sequence top and base (Fig. 2). The sequence base is a prominent angular unconformity occurring at 1.45 seconds at the Edlok location. Progressively older strata are truncated below the sequence base in a westerly direction. Taglu sequence strata subcrop below the unconformity along the length of the illustrated profile (Fig. 2); pre-Taglu and pre-Tertiary strata subcrop some 10 to 15 km west of the Edlok site.

Organic matter in Richards strata is thermally immature, as indicated by low vitrinite reflectance values (.3 to .4 % Ro) and low Tmax measurements (<430°C). Organic carbon content is very low (< 1 %) in the sandy intervals but increases to 1 to 2 % in the mudstone sections (Fig. 3). Isolated TOC values of 4 to 5 % are probably representative of coaly debris.

The upper portion of the Richards sequence (from 1115-1326 m) is barren of foraminifers and contains only sparse palynofloras similar to those in the overlying sequence (Fig. 3). The lower half of the sequence is characterized by the presence of the dinoflagellates *Cordosphaeridium inodes* (Klumpp) Eisenack emend. Morgenroth, and Peridinioid sp. A, and the appearance of rare pollen of *Tilia* and *Carya*. The palynological assemblage indicates an Eocene age (possibly Middle). A foraminiferal assemblage in the 1368 to 1656 m interval (Fig. 3) is distinguished by abundant *Jadammina statuminis* (Plate 5) a characteristic marker that occurs in the inner shelf facies of the Richards Formation in the Mackenzie Delta area (McNeil, 1983). *J. statuminis* suggests a Middle to Late Eocene age for the interval.

The sandstone-bearing Richards sequence strata in the Edlok well are unlike the typical mudstone-dominant shelf/prodelta deposits in the Richards sequence of the Mackenzie Delta area. In the Edlok well, the fining-upward sandstone cycles in the lower portion of the sequence are probably marginal marine transgressive deposits that are overlain by thin bedded inner shelf strata in the upper part of the sequence.

Taglu sequence

The Taglu sequence was penetrated by both the Natsek and Edlok wells, in the intervals from 215 to 1951 m and 1595 to 2530 m, respectively (Figs. 3, 4). Correlation of the sequence between the two wells is primarily seismically based (Fig. 2), but is supported by foraminiferal correlations. The sequence unconformably underlies Iperk strata in the Natsek area and Richards strata in the Edlok area. In the Natsek E-56 well (Fig. 4), the sequence consists entirely of weakly consolidated, silty to pebbly mudstone. Traces of carbonaceous debris and pyrite occur in the mudstones along with some thin beds of ironstone concretions. The upper 400 m of the sequence contains abundant chert pebbles. The lithology of Taglu strata in the Edlok well is distinctly different; the sequence contains sandstone-dominant sections in the 1595 to 1618 m and 1736 to 2281 m intervals (Fig. 3). The sandstones are fine to medium grained and consist of rounded to subrounded quartz and chert grains. The sandstones are cemented to varying degrees with silica and, less commonly, pyrite. The variability in degree of cementation within Taglu sandstones results in porosity values ranging from lows of 10 % up to 25 % and more. The higher porosity values occur in numerous thin zones or streaks encased by thicker, less porous (average 10-15 %) sandstone sections. The Taglu sequence top (at 1595 m) was picked on the basis of the distinct sonic velocity differentiation between sandstone units (Fig. 6). The lower sandstone section in the Edlok well (1736-2281 m) is a nearly massive, 545 m thick interval containing scattered, thin, mudstone interbeds. No significant fining- or coarsening-upward trends are apparent within this interval. The 1736 to 2281 m sandstone section is abruptly overlain and underlain by mudstones that span the 1618 to 1736 m and 2281 to 2530 m (TD) intervals. The mudstones are dark grey to black and silty. The upper mudstone section contains coal fragments between 1715 and 1720 m (Fig. 6).

Organic carbon content in the Edlok Taglu strata varies from less than .6 % in the sandstone-dominant sections to 1 to 2 % in the mudstone sections (Fig. 3). A single sample from the coal-bearing zone at 1720 m measured 11 % TOC. Vitrinite reflectance measurements from the Edlok samples increase from .3 to .4 % Ro (immature) in the 1618 to 1736 m interval, up to .6 to .7 % Ro (marginally mature) in the 2281 to 2530 m interval toward the bottom of the well. In the thick sandstone section from 1736 to 2281 m, the reflectance measurements are widely scattered, indicative of a mixed, reworked assemblage of organic material. Tmax measurements of the Edlok samples show similar maturation characteristics. The Tmax values increase from 420 to 430°C in the uppermost part of the sequence, to 440 to 450°C near the well bottom. The top of the sandstone section (at 1736 m) is marked by an abrupt increase in Tmax values (from 425°C average up to 450+°C). This abrupt increase in maturation rank (not so readily seen in the %Ro values) probably reflects the abrupt downhole increase in the amount of reworked, higher-rank organic material across the mudstone-sandstone contact. Below the 1736 m level there is also a very subtle indication of an inverse (decreasing downhole) maturation trend within the sandstone section. One possible interpretation of such a trend is that the sandstones were deposited from source terrains in which progressively higher-rank material was exposed as the erosion-deposition cycle evolved through time. Taglu sequence samples from the Natsek E-56 well display a much simpler maturation trend with reflectance measurements increasing from .3 % Ro at 900 m to .45 % Ro at 1800 m (Fig. 4).

In the Natsek well, the 251 to 1250 m interval of the Taglu sequence is barren of foraminifers but contains a rich and diverse assemblage of pollen and dinoflagellates (Figs. 4, 5 and Plates 1, 2, 3). Pollens of *Picea, Pinus, Carya, Pterocarya* and Taxodiaceae/Cupressaceae (including *Metasequoia* type) are common to abundant. *Tsuga* and *Juglans* occur in the top of the interval. Other pollens present in the sequence include those of *Tilia, Liquidambar, Ulmus*, Ericaceae, *Nyssapollenites* sp. and *Tricolporopollenites kruschii* (Potonie) Thomson and Pflug. Massulae of the freshwater fern *Azolla* occur commonly from

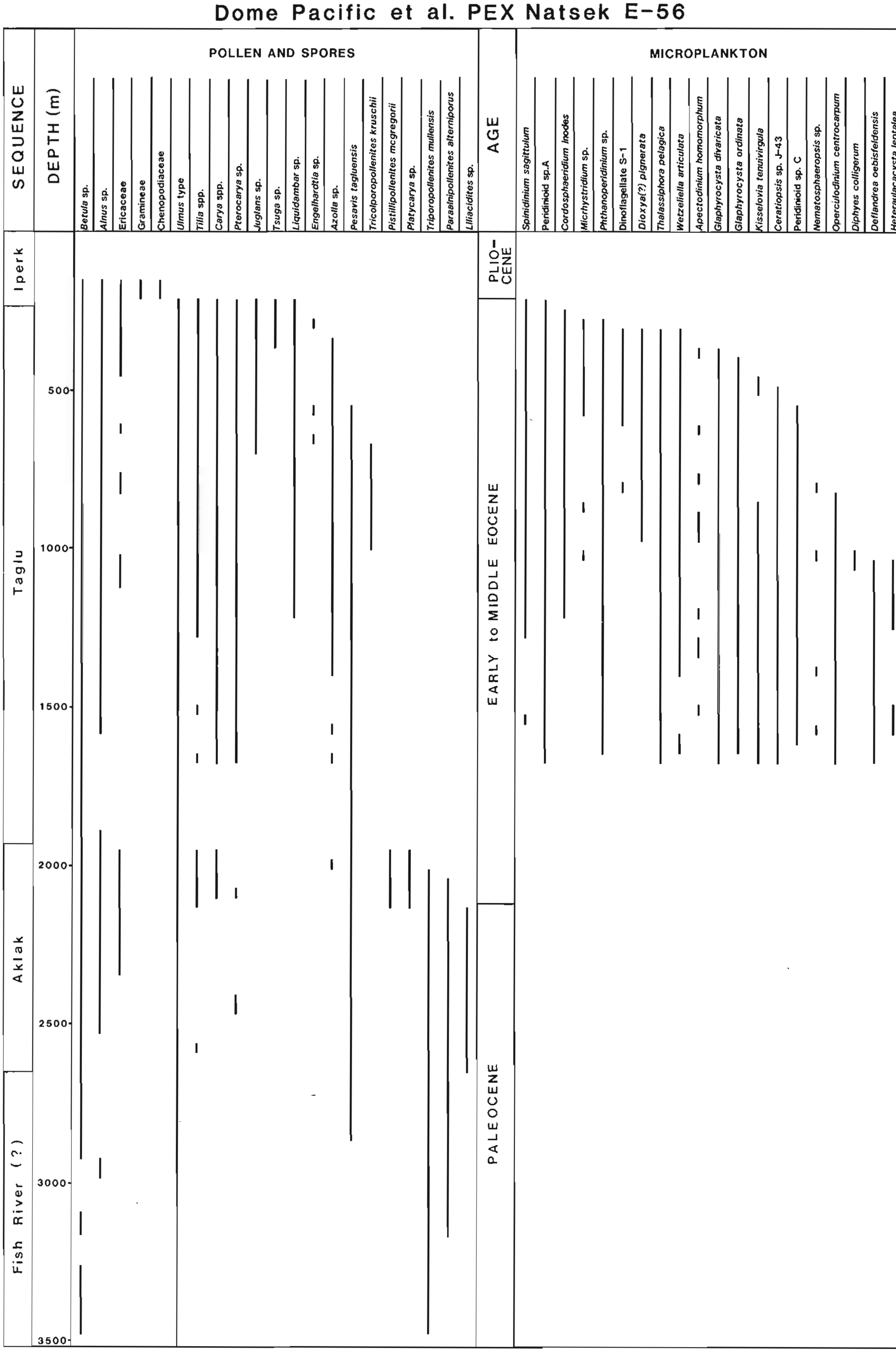


Figure 5. Distribution of selected palynomorphs in the Natsek E-56 well.

335 to 914 m and rarely below. The habitat of *Azolla* is quiet water, such as in ponds, pools and still areas of streams, rivers and lakes. Fungal spores are common below 305 m and include species of *Multicellaesporites, Brachysporisporites, Diporicellaesporites* and, rarely, *Striadiporites*. The stratigraphically important species *Pesavis tagluensis* Elsik and Jansonius occurs below 700 m. The palynomorph assemblages of the sequence are characterized by the presence of abundant dinoflagellates, most prominent of which are *Wetzeliella articulata* Eisenack and *Kisselovia tenuivirgula* (Williams and Downie) Lentin and Williams. Other dinoflagellates present include *Spinidinium sagittulum* (Drugg) Lentin and and Williams, *Phthanoperidinium* sp., *Cordosphaeridium inodes* (Klumpp) Eisenack emend. Morgenroth, Peridinioid species A, B, C, *Ceratiopsis* sp. J-43 of Staplin, *Palaeoperidinium* sp. A of Norris (Dinoflagellate sp. S-1 in Staplin), *Thalassiphora pelagica* (Eisenack) Eisenack and Gocht, *Glaphyrocysta divaricata* (Williams and Downie) Stover and Evitt, *G. ordinata* (Williams and Downie) Stover and Evitt and *Deflandrea oebisfeldensis* Alberti. The abundance of specimens of the *Wetzeliella* complex (especially *W. articulata* and *K. tenuivirgula*) indicates an Early to Middle Eocene age. The *Wetzeliella* zone, prominent in the Natsek well, is commonly present in the upper Reindeer and lower Richards formations of the Mackenzie Delta area. This zone correlates well with Early to Middle Eocene dinoflagellate floras in Europe where there is a greater variety of members of the *Wetzeliella* complex. The presence of *Deflandrea oebisfeldensis, Glaphyrocysta divaricata*, and *G. ordinata* suggests that the lower part of this interval may be of Early Eocene age. *D. oebisfeldensis* is not known to occur later than Early Eocene, and the *Glaphyrocysta* species are not common in the Middle Eocene in Europe. The palynomorph assemblages of the Taglu sequence in the Natsek well have been recorded from the Mackenzie Delta area, with some differences in species, by Doerenkamp et al. (1976), Staplin (1976), Brideaux and Myhr (1976), Ioannides and McIntyre (1980) and Norris (1986). These assemblages suggest a nearshore shallow marine environment. The palynomorph assemblages in the Taglu sequence extend, with no significant change in species or abundances, to a depth of about 1676 m.

The 1250 to 1555 m interval in the Natsek well (Fig. 4) contains abundant agglutinated foraminifers, including

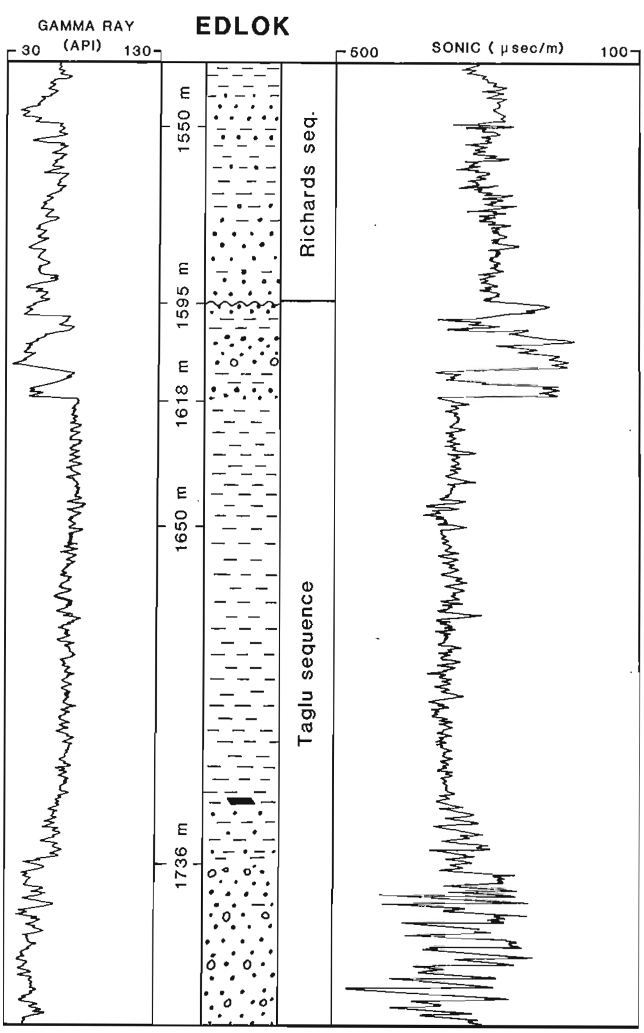


Figure 6. Gamma ray and sonic log curves across the base-Richards unconformity in the Edlok well.

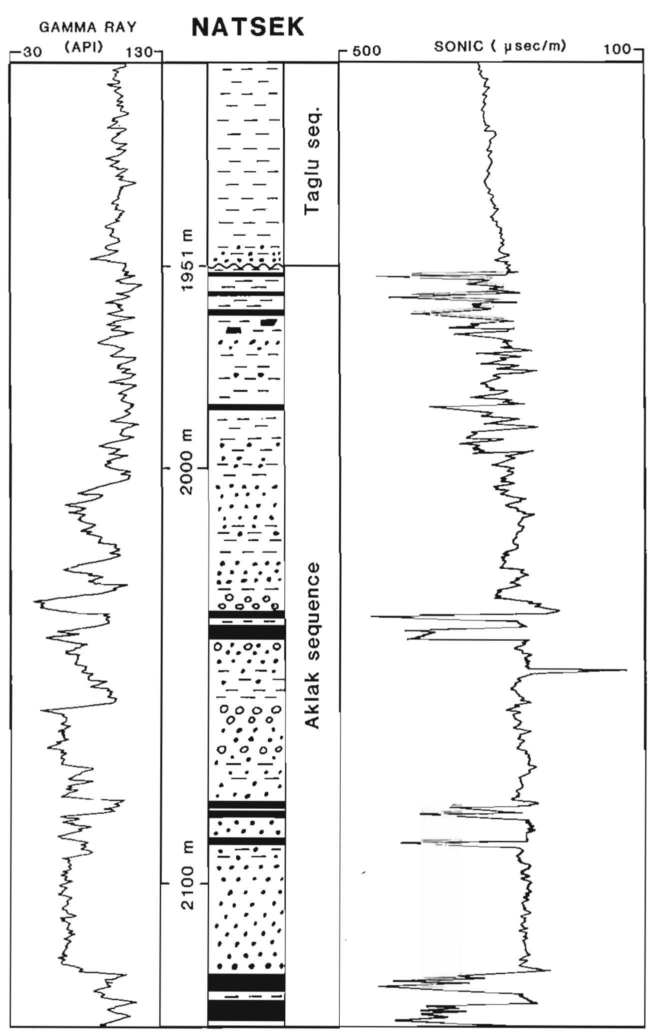


Figure 7. Gamma ray and sonic log curves across the base-Taglu unconformity in the Natsek well.

Ammodiscus sp., *Arenobulimina* sp., *Brachysiphon* sp., *Haplophragmoides* sp., *Pelosina* sp., *Saccamina* sp., *Portatrochammina* sp. 2850, and *Verneuilinia* sp. 2700 (Plate 5). This assemblage has been recognized in the Reindeer Formation of several Mackenzie Delta area wells. The age of the assemblage is Early Eocene and the depositional environment was likely that of an inner shelf with less than normal marine salinity. The foraminiferal assemblage becomes less diverse below 1555 m. Reworked siliceous foraminifers of Early Cretaceous age are common below 1250 m.

Taglu sequence strata in the Natsek well contain other microfossils, including *Azolla* megaspores and the algal cyst *Leiospheridea* (particularily abundant in the upper part of the sequence). These microfossils have been identified in the lower Richards and upper Reindeer sequences in the Tarsiut A-25 and Netserk F-40 wells of the central Beaufort Sea area. Another characteristic feature of the upper half of the Taglu sequence in the Natsek well is the consistent occurrence of fish bones and fragments. A similar fish-bone-yielding unit occurs in the Reindeer Formation in the Adgo F-28 well.

Microfossil and palynological assemblages in the Taglu sequence of the Edlok N-56 well are dominated by abundant reworked Early Cretaceous foraminifers, and Cretaceous spores and dinoflagellates, respectively (Fig. 3). The palynofloras contain few grains considered to be indigenous, but the presence of pollen of *Tsuga, Betula, Alnus, Ulmus, Tilia* and Ericaceae, mainly near the top of the interval, suggests an Eocene age. The presence of *Paraalnipollenites alterniporus* (Simpson) Srivastava in the 2412 to 2530 m (TD) interval indicates a Paleocene to Early Eocene age, but does not provide a good correlation with the Natsek E-56 well. The lower mudstone section (below 2358 m) in the Edlok well yields the Lower Eocene foraminifer *Portatrochammina* sp. 2850 (Fig. 3 and Plate 5) suggesting a correlation with Taglu sequence mudstones in Natsek E-56 (1250-1951 m).

Seismically, the Taglu sequence is characterized by variable reflection amplitudes and geometry. In the vicinity of the Natsek well, the sequence appears as an interval of low amplitude, subhorizontal reflections (Fig. 2). The reflections are locally discontinuous and slightly wavy. West of the Natsek location, the subhorizontal reflections merge with inclined, west dipping reflections. The inclined reflection zones in turn pass downdip into higher amplitude reflections. The high amplitude reflections occur in at least three tabular or lens-shaped bundles (identified as units A, B and C in Figure 2). The lowest unit (A) is a 300 to 400 millisecond interval that can be described seismically as a mounded onlap-fill succession. The upper units (B and C) are thinner zones of hummocky to subparallel reflections. The high amplitude units are locally downlapped by overlying reflections. West of the Edlok location, seismic units A and B are truncated by the base-Richards unconformity. The Edlok well penetrated both units A and B in the 1.5 to 2.0 second interval. Unit A corresponds to the 1736 to 2281 m sandstone section in the well. The higher amplitude reflections at the top and base of unit A are the result of the acoustic impedance contrasts across the top and base of the sandstone section. Seismic unit B is a two-cycle reflection zone that appears to have been generated (at least in part) by the high velocity sandstone at 1595 to 1618 m (Fig. 6).

Figure 8 shows the seismically mapped areal extent of intra-Taglu units A, B and C and outlines their lobate to irregular shaped distributions. Depositional edges occur on all but their northwest margins. Updip (southeast) of units A, B and C, the transition from inclined to subhorizontal reflections is interpreted as marking an intra-sequence boundary between foreset and topset deposits. The seismically complex foreset zones extend close to, but apparently not through, the Natsek location (Fig. 2). The foreset reflection zones are interpreted as deep water, slope deposits. The Taglu sandstone sections (represented by the high amplitude seismic units) are thought to be submarine fan complexes. The interpretation of submarine fan deposition in the Edlok area is consistent with the sand-rich nature of the units, their cross-sectional and areal geometries, and their locations within and downdip of complex slope deposits. The mudstone sections in the upper part of the Taglu sequence in the Edlok area are interpreted as deep marine slope or inter-fan deposits (Fig. 9). The lowermost mudstones penetrated by the Edlok well (containing the neritic foraminifer *Portatrochammina* sp. 2850) may be shelf deposits. The shallow marine shelf facies become more predominant to the east in the vicinity of the Natsek well (Fig. 9). The silty and pebbly mudstones in the upper part of the Taglu sequence at the Natsek location may be nearshore and/or estuarine deposits, as indicated by the palynomorph content.

The Lower to Middle Eocene Taglu sequence in the Natsek-Edlok area is equivalent to the upper part of the Reindeer Formation or sequence of the Mackenzie Delta area. The division of the ''Reindeer'' succession into the Taglu and Aklak sequences (in this report) is an important revision of the original sequence stratigraphy of the Beaufort-Mackenzie Basin (Dietrich et al., 1985).

Aklak sequence

In the Natsek area, the Taglu sequence base is marked by a prominent unconformity, evident in both the seismic (Fig. 2) and well data (Fig. 4). Below the unconformity, which was penetrated at 1951 m in the Natsek well, is a thick succession of interbedded sandstone, conglomerate, shale and coal (Figs. 4, 7). The sandstone and conglomerate beds contain quartz and chert grains and pebbles with minor amounts of feldspar, muscovite and lithic fragments. Calcite and silica cementation is pervasive throughout the sandstones, resulting in low average porosity values (typically 10-15 %). The Aklak sequence extends from 1951 to (?)2644 m in the Natsek well and is characterized by an abundance of thick coal seams (individually up to 7 m thick). The coal seams have distinctive low interval velocities (2000-2500 m/sec.; Fig. 7). The interbedding of low velocity coal seams with high velocity sandstones/conglomerates produces a high amplitude seismic facies within the Aklak and underlying Fish River(?) sequence (Fig. 2). The seismic character change across the Taglu-Aklak contact produces one of the most prominent seismic sequence boundaries in the region. The base-Taglu unconformity is also a major

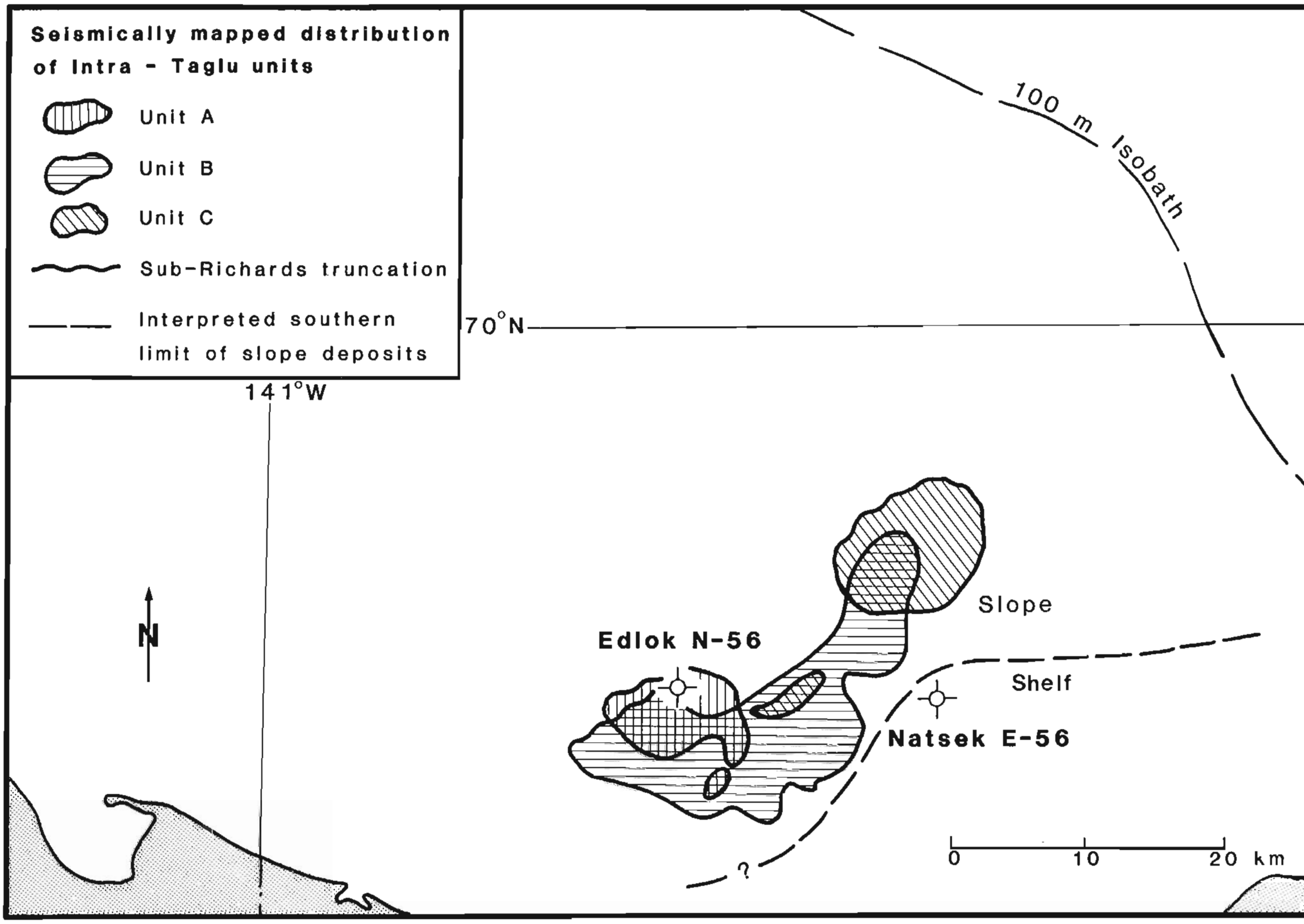


Figure 8. Areal distribution of seismic units A, B and C within the Taglu sequence.

erosional surface with substantial thicknesses of strata truncated below the unconformity. The Aklak sequence reflections are continuous and subparallel to divergent in the Natsek area.

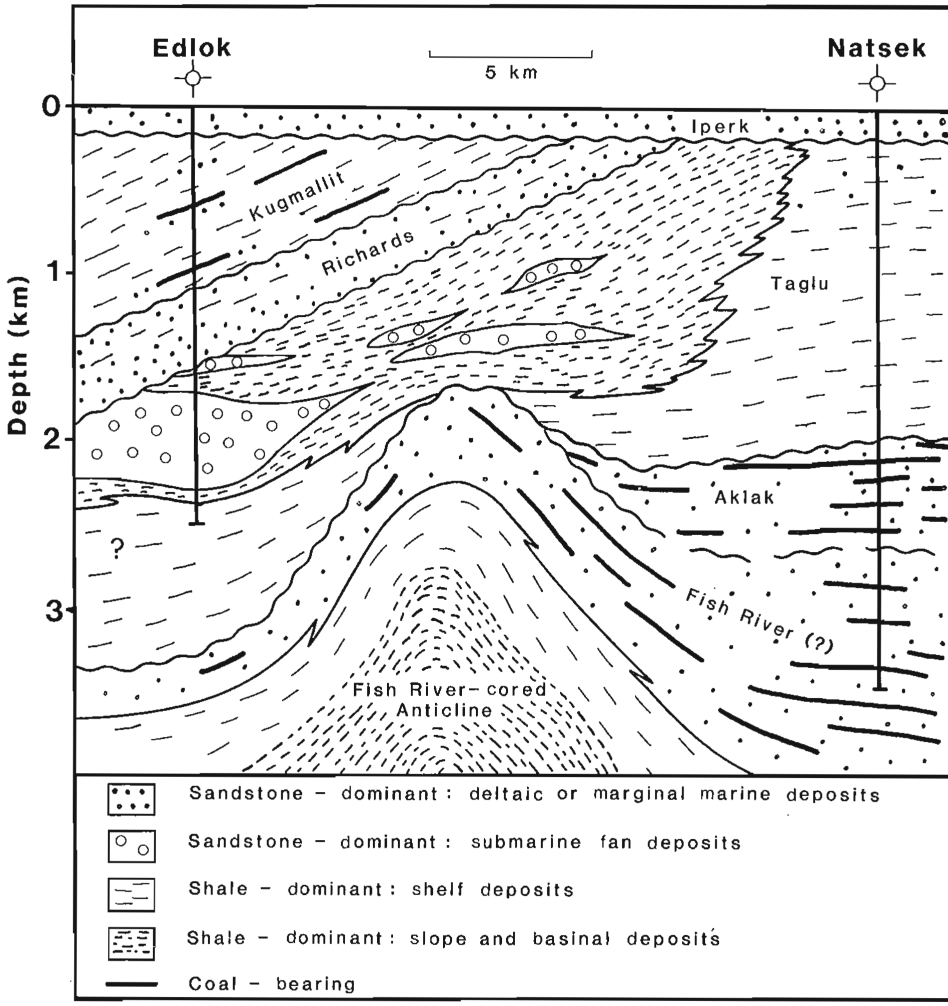
The depositional setting of Aklak strata is clearly a nonmarine, upper delta plain environment. Stacked fluvial channel deposits are common (blocky gamma-ray log responses; Fig. 7). Coarsening-upward, crevasse-splay deposits (7-10 m thick, funnel shaped log curves) are commonly capped by coal seams. Shaly intervals within the Aklak sequence are probably floodplain or levee deposits.

Organic matter in the Aklak sequence is marginally mature with vitrinite reflectance values averaging .6 % Ro (Fig. 4). A maturation discontinuity occurs across the Taglu-Aklak contact, where reflectance values increase abruptly from .45 % to .6 % Ro. This maturation trend break is probably the result of substantial pre-Taglu erosion and removal of Aklak strata.

Aklak strata are barren of foraminifera with the exception of two thin intervals yielding sparse *Reticulophragmium borealis* (2012-2042 m and 2195-2225 m) and an interval yielding *Portatrochammina* sp. 2849 (Plate 5) at the sequence base (2560- 2720 m; Fig. 4). *Portatrochammina* sp. 2849 occurs also in a marine shale unit in the Aklak Member of the Reindeer Formation in the Yukon Coastal Plain, and appears to be in situ in the Natsek well. *R. borealis*, however, is apparently reworked, for the species has been known previously only from Paleocene strata of the lowermost Reindeer sequence (Ministicoog Member) in the Mackenzie Delta area. Reworked, silicified foraminifers of Early Cretaceous age are common in the Aklak and underlying strata.

Palynomorphs in the Aklak sequence are most abundant in the 1951 to 2134 m interval, which yielded pollen of *Picea, Pinus*, Taxodiaceae, *Carya, Alnus, Betula, Ulmus, Tilia, Platycarya, Paraalnipollenites alterniporus* and *Pistillipollenites mcgregorii* Rouse (Figs. 4, 5 and Plates 2, 3). The fungal body, Pesavis tagluensis is also present. The presence of *Paraalnipollenites alterniporus*, which occurs as high as the Lower Eocene in the Mackenzie Delta area (Ioannides and McIntyre, 1980), and *Platycarya*, abundant in this interval, indicates an Early Eocene age. The occurrence of *Platycarya* has been noted by some authors (e.g., Wing, 1984), who conclude that it is an important Lower Eocene taxon.

Figure 9. Geological cross-section illustrating major lithotypes and depositional environments in the Natsek-Edlok area.



Below 2134 m in the Natsek well, only impoverished pollen assemblages occur. The absence of typical Maastrichtian and Eocene species, the presence of *Paraalnipollenites alterniporus*, the sparse nature of the pollen assemblages, and the Early Eocene age for the interval above suggest a Paleocene age for this interval.

The limited microfossil assemblages indicate a Late Paleocene-earliest Eocene age for Aklak strata. The sequence appears to be equivalent to the lower Reindeer sequence or Aklak Member of the Reindeer Formation of the Mackenzie Delta area. The major unconformity separating the Aklak and Taglu sequences probably developed in the Early Eocene. This sequence boundary had not been recognized in earlier studies of the Mackenzie Delta area succession. Recent work, however, has indicated that the Natsek area Taglu-Aklak unconformity may correlate with the base of the "Ellice shale" interval of the Reindeer Formation (Dixon et al., 1985).

Fish River(?) sequence

The sixth Tertiary sequence has been identified, in part, by the recognition of a seismic discontinuity within the high amplitude seismic facies (Fig. 2). The seismic discontinuity appears as a surface of onlap separating divergent Aklak sequence reflections from more parallel underlying reflections. In the Natsek well, this contact is picked at 2644 m, at the base of a 20 m thick fining-upward conglomeratic sandstone (Fig. 4).

The stratigraphic identification of the 2644 to 3520 m (TD) interval in the Natsek well is questionable. The interval may comprise Aklak sequence strata, and the seismic discontinuity may be interpreted as a local, structurally related intra-sequence contact. Alternatively (and our preferred interpretation at present), the interval may be composed of the upper part of the Fish River sequence. This latter interpretation appears to fit well with the seismically-defined position of the section in relation to underlying, highly deformed Cretaceous strata (Fig. 2).

The Fish River(?) strata below 2644 m in the Natsek well (Fig. 4) consist of interbedded sandstone, shale and coal. In contrast to the Aklak sequence, the Fish River(?) sequence contains a higher percentage of shale, fewer conglomeratic beds and a greater number of small- and medium-scale coarsening-upward cycles. In general, Fish River(?) strata appear to be dominated by outer delta-plain deposits with at least one delta-front cycle occurring in the 3200 to 3249 m interval. Fluvial channel deposits are noticeably less abundant in the Fish River(?) sequence. Sandstone composition and cement are similar in both sequences.

The Natsek well penetrated only the upper portion of the sequence (3520 m TD occurring at about 2.4 seconds). The high amplitude seismic facies extends below the Natsek well TD to reflection times of about 3.0 seconds. Gradationally underlying the high amplitude facies is an interval of low amplitude, discontinuous reflections extending to the interpreted Fish River(?) sequence base at reflection times of 3.5 to 3.7 seconds. Based on the seismic data, the total thickness of coal-bearing deltaic strata in the Fish River(?) sequence is estimated to be about 1800 m at the Natsek location.

Organic matter in Fish River(?) strata is mature, with vitrinite reflectance values averaging .7 to .8 % Ro (Fig. 4). The coaly, Type III organic matter is expected to be more gas- than oil-prone. A subtle maturation discontinuity is apparent across the Aklak-Fish River(?) boundary, where average reflectance values increase from .6 % to .7- .75 % Ro.

Microfossils in the sequence include *Reticulophragmium borealis* (probably reworked) and an impoverished palynomorph assemblage similar to that seen in the 2134 to 2644 m interval of the Aklak sequence (Fig. 6). The age of Fish River(?) strata penetrated by the Natsek well (to 3520 m TD) is considered to be Paleocene. There is no indication of any Maastrichtian (Late Cretaceous) fossils in the 2644 to 3520 m interval.

STRATIGRAPHIC SUMMARY

The Natsek and Edlok wells penetrated portions of six Tertiary sequences containing lithotypes from a spectrum of depositional environments (Fig. 9). The Iperk sequence, penetrated by both wells, contains unconsolidated fluviodeltaic to nearshore sands and muds of Pliocene to Pleistocene age. The shallower portions of the sequence contain glaciomarine deposits. The Kugmallit and Richards sequences were penetrated by the Edlok well. The Kugmallit sequence contains mudstone-dominant inner shelf and lagoonal deposits of Oligocene age. The Richards sequence

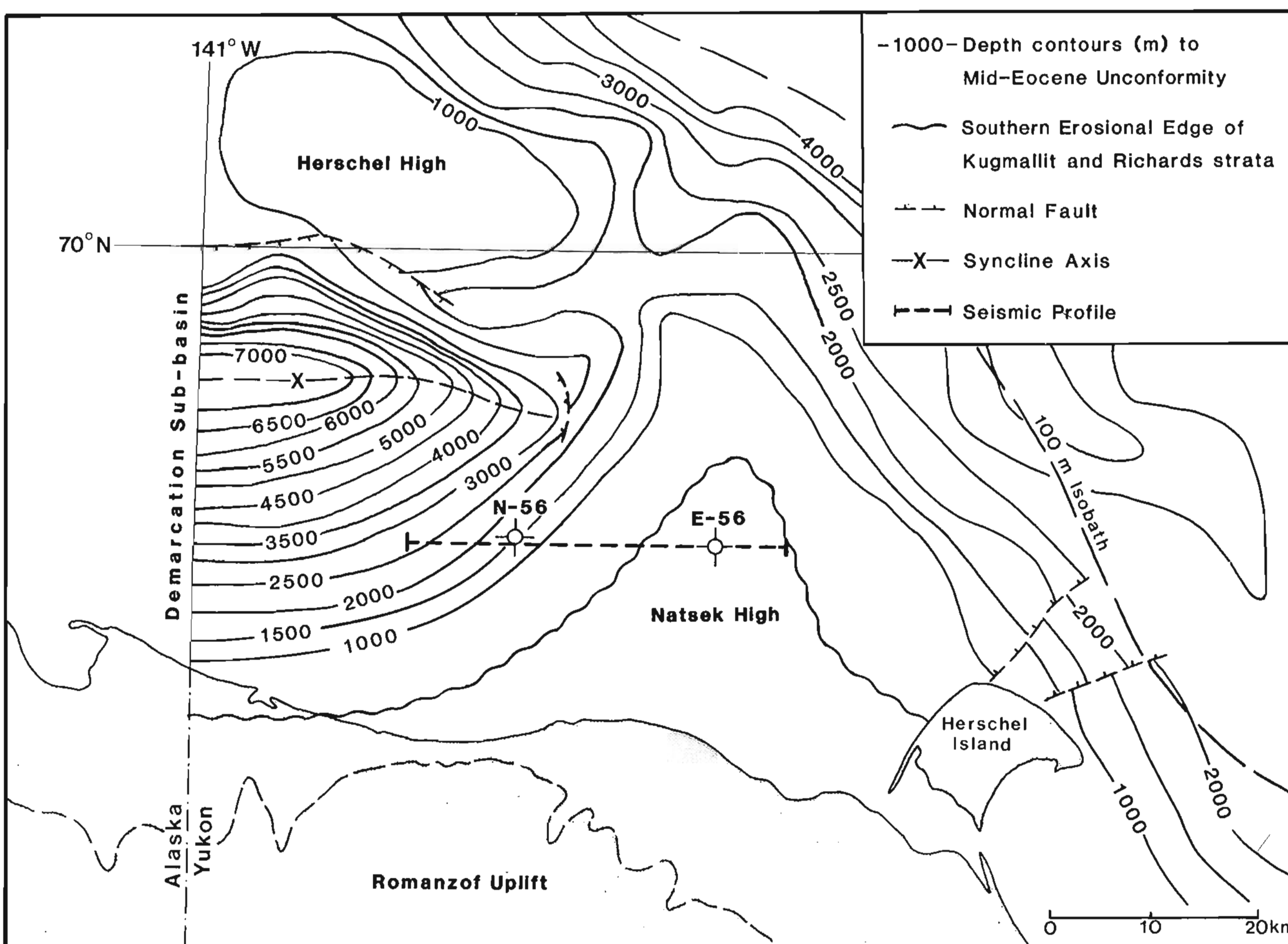


Figure 10. Depth converted seismic structure map of the Middle Eocene unconformity.

is an interbedded sandstone- mudstone succession of nearshore deposits of Middle to Late Eocene age. The Taglu sequence was penetrated by both the Natsek and Edlok wells. The sequence is pre-dominantly a mudstone-dominant shelf to slope succession, but also contains several sandstone-rich, submarine-fan deposits in the Edlok area. The age of the sequence is Early to Middle Eocene. The Natsek well penetrated the Aklak and Fish River(?) sequences. Both sequences contain coal-bearing fluviodeltaic strata. The ages of the these sequences are Late Paleocene to Early Eocene for the Aklak sequence, and Paleocene for the Fish River(?) sequence. The lower portion of the Fish River(?) sequence, not penetrated by the Natsek well, undoubtedly contains mudstone-dominant shelf and basinal strata that form the core of the anticline between the two wells.

STRUCTURAL SETTING

The Natsek-Edlok region of the Canadian Beaufort-Mackenzie Basin has had a complex history of deposition and subsidence, relative sea level variations and tectonic uplift and compression. The major structural patterns and elements within the Tertiary sequences can be identified at two separate stratigraphic levels: the Middle Eocene unconformity and the Early Eocene unconformity. The Middle Eocene unconformity (base-Richards sequence) can be mapped seismically across most of the Beaufort Sea shelf, offshore of northern Yukon. The Early Eocene unconformity (base-Taglu sequence) can only be mapped locally, due to sub-Richards truncation and complex structural deformation. Depth structure contours at the level of the Middle Eocene unconformity outline the Demarcation sub-basin and Natsek and Herschel highs (Fig. 10). The Demarcation sub-basin is an asymmetric, east-west trending depocentre containing up to 7000 m of upper Eocene to (?)Miocene strata. The sub-basin contains progressively younger strata toward its axis northwest of the Edlok location. The Demarcation sub-basin is bounded to the north and east by the Herschel and Natsek highs. The Herschel High is a structurally positive area cored by highly deformed Aklak and Fish River(?) strata. The Natsek High is a structurally positive area that separates the northwesterly dipping Demarcation sub-basin strata from the northeasterly dipping strata of the main Beaufort-Mackenzie basin-fill. Unlike the Herschel High, a large portion of the Natsek High area is underlain by less intensely deformed Taglu, Aklak and Fish River(?)

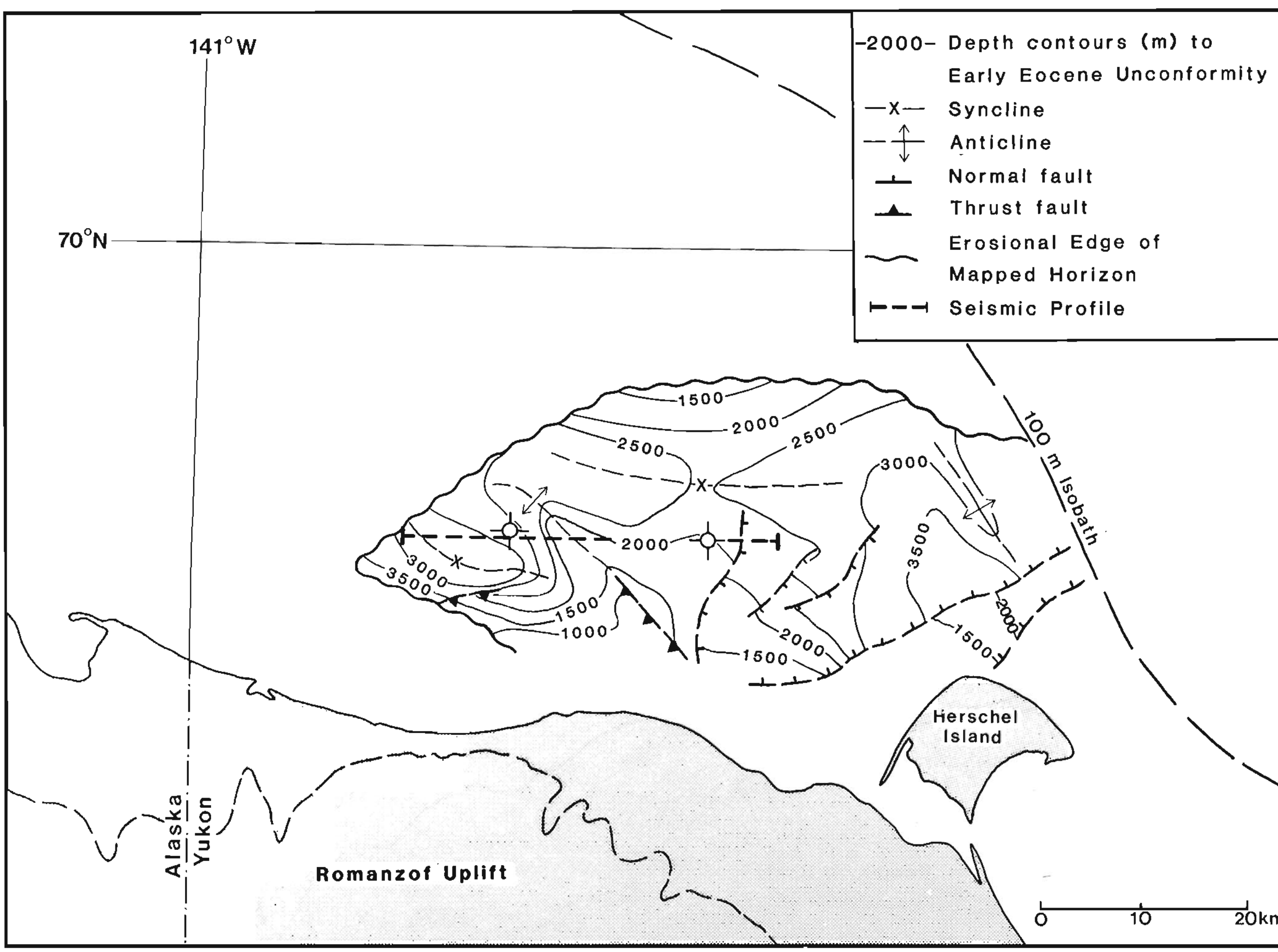


Figure 11. Depth converted seismic structure map of the Early Eocene unconformity.

strata. The deeper structural patterns beneath the Natsek High area are outlined by depth structure contours at the level of the Lower Eocene unconformity (Fig. 11). These older structural trends include an east-west oriented syncline, southeast-northwest oriented detachment folds and reverse faults, and southwest- northeast oriented normal faults. The Natsek well is located on the north limb of the main syncline, and there appears to be no structural closure at the unconformity level. The well was undoubtedly targeted to test potential traps associated with the subcrop truncation of Aklak strata. The Edlok well is located on the southwest flank of one of the southeast-northwest oriented folds. This well appears to have been targeted to test a potential stratigraphic trap involving the subcrop truncation and updip pinchout of Taglu strata. The absence of hydrocarbons in both wells may be more the result of the lack of reservoir seals than any other factor.

The development of the Lower and Middle Eocene unconformities is clearly related to periods of widespread basin margin uplift, probably associated with north directed pulses of compression along the Cordilleran front in northern Yukon/northeastern Alaska. Eocene compression also was intermittently active during Taglu deposition, resulting in the generation of local folds and faults and, perhaps, uplifting local basin margins that were the source of the intra-Taglu coarse grained submarine-fan deposits. The base- Richards unconformity truncates most of the older structural features, indicating that early Tertiary tectonism in this area ended in the Middle Eocene. Upper Eocene to Miocene strata were subsequently deposited during periods dominated by subsidence and gravity tectonics. The third major unconformity present in this area (and indeed present throughout the entire Beaufort-Mackenzie Basin) is the Upper Miocene, base-Iperk unconformity. The relatively thin section of Plio-Pleistocene strata above this unconformity is flat lying and undeformed. The Upper Miocene sequence boundary marks a long period of emergence and erosion of the entire southern Beaufort Sea region. The sequence boundary appears to have developed in response to widespread uplift and a major eustatic sea level fall.

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PLATES 1-5

Plates 1 to 3 illustrate representative dinoflagellates, pollen, and spores from the Natsek E-56 well; Plates 4 and 5 illustrate representative foraminifers from the Natsek E-56 and Edlok N-56 wells. All specimens recovered from well cuttings.

In the plate captions, palynomorph and foraminifer species names are followed by the GSC locality number (C-80502 for Natsek E-56; C-130896 for Edlok N-56) and the GSC type number (prefixed GSC). Palynomorph captions also include the slide number (prefixed P) and stage coordinates. The stage coordinates are for the Reichert-Jung Polyvar microscope (number 392166) at the Institute of Sedimentary and Petroleum Geology, Geological Survey of Canada, Calgary, Alberta.

All samples used in this study, for both the Natsek E-56 and Edlok N-56 wells, including the slides that do not contain figured specimens, are stored at the Institute of Sedimentary and Petroleum Geology, Geological Survey of Canada, 3303 - 33rd Street N.W., Calgary, Alberta, T2L 2A7. The slides containing the figured specimens are curated in the type collection of the Geological Survey of Canada, 601 Booth Street, Ottawa, Ontario, K1A 0E8. They are at present on loan to the Institute of Sedimentary and Petroleum Geology, Calgary, Alberta.

PLATE 1

All figures x500

Figure 1. *Wetzeliella articulata* Eisenack
C-80502/3600-3690, P2204-32e, 663 x 1140, GSC 91997.

Figure 2. *Nematosphaeropsis* sp.
C-80502/5100-5190, P2204-47e, 641 x 1281, GSC 91998.

Figure 3. *Dioxya(?) pignerata* Norris
C-80502/1200-1290, P2204-8e, 595 x 1210, GSC 91999.

Figure 4. *Kisselovia tenuivirgula* (Williams and Downie) Lentin and Williams
C-80502/5000-5090, P2204-46e, 631 x 1094, GSC 92000.

Figure 5. *Wetzeliella articulata* Eisenack (cf. *Wetzeliella hampdenensis* Wilson)
C-80502/1200-1290, P2204-8e, 582 x 1103, GSC 92001.

Figure 6. *Apectodinium homomorphum* (Deflandre and Cookson) Lentin and Williams emend. Harland
C-80502/4900-4990, P2204-45e, 701 x 1261, GSC 92002.

Figure 7. *Heteraulacacysta leptalea* Eaton
C-80502/5000-5090, P2204-46e, 574 x 1145, GSC 92003.

Figure 8. *Cordosphaeridium inodes* (Klumpp) Eisenack emend. Morgenroth
C-80502/1200-1290, P2204-8e, 497 x 1226, GSC 92004.

Figure 9. *Glaphyrocysta divaricata* (Williams and Downie) Stover and Evitt
C-80502/3900-3990, P2204-35e, 646 x 1217, GSC 92005.

Figure 10. *Phthanoperidium* sp.
C-80502/2300-2390, P2204-19e, 626 x 1193, GSC 92006.

Figure 11. *Phthanoperidium* sp.
C-80502/2100-2190, P2204-17e, 636 x 1243, GSC 92007.

Figure 12. *Glaphyrocysta ordinata* (Williams and Downie) Stover and Evitt
C-80502/2000-2090, P2204-16e, 380 x 1219, GSC 92008.

Plate 1

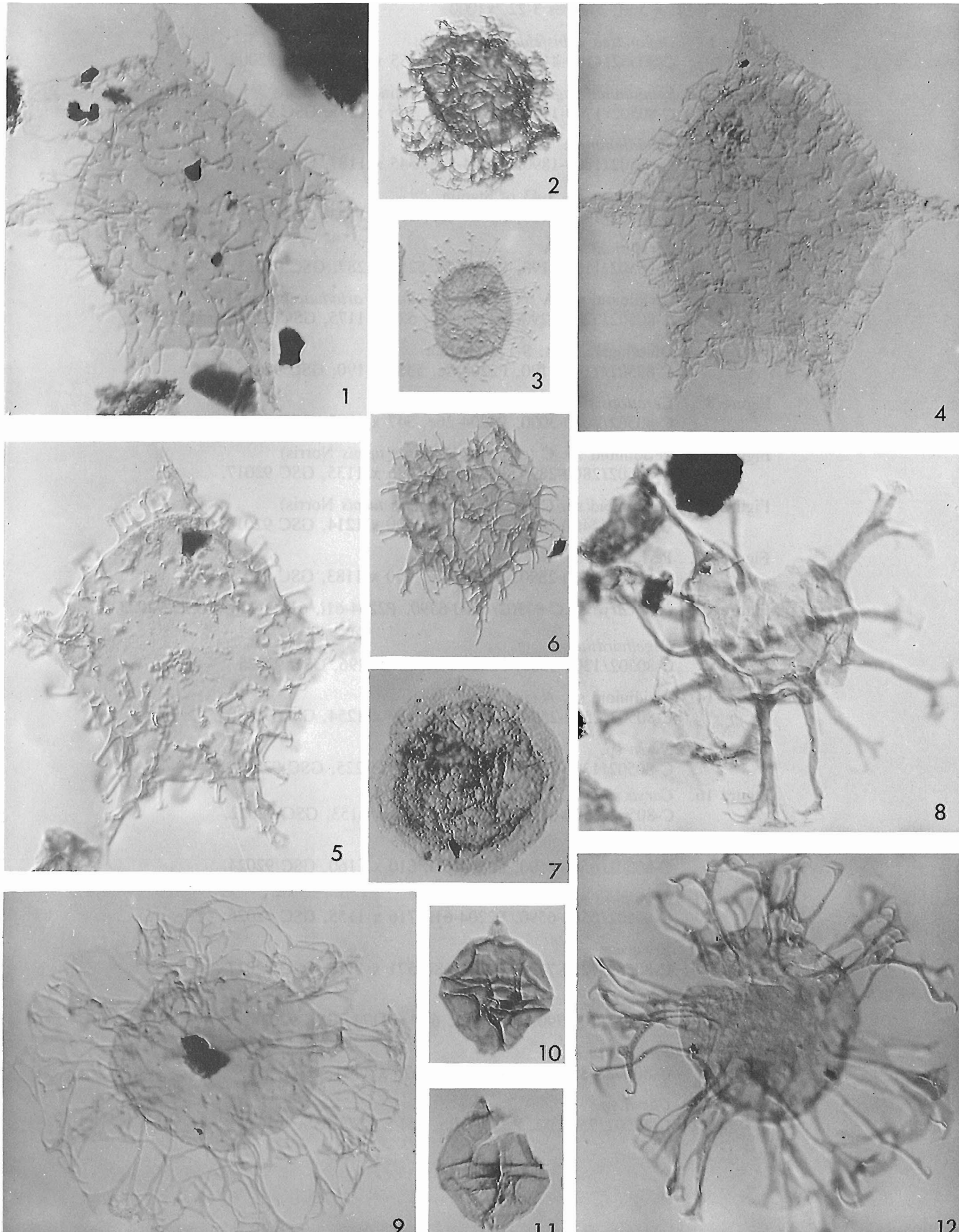


PLATE 2

Figures 1-12, x 500; Figures 3-22, x1000

Figure 1. *Deflandrea oebisfeldensis* Alberti
C-80502/4100-4190, P2204-37e, 555 x 1174, GSC 92009.

Figure 2. *Spinidinium sagittulum* (Drugg) Lentin and Williams
C-80502/1700-1790, P2204-13e, 706 x 1295, GSC 92010.

Figure 3. *Spinidinium* sp. cf. *S. sagittulum*
C-80502/1800-1890, P2204-14e, 545 x 1187, GSC 92011.

Figure 4. *Ceratiopsis* sp. J-43 of Staplin
C-80502/4900-4990, P2204-45e, 570 x 1113, GSC 92012.

Figure 5. Peridinioid sp. A
C-80502/1100-1190, P2204-7e, 525 x 1287, GSC 92013.

Figure 6. Peridinioid sp. A (cf. *Palaeoperidinium ariadnae* Norris)
C-80502/3200-3290, P2204-28e, 522 x 1175, GSC 92014.

Figure 7. Dinoflagellate sp. S-1 of Staplin
C-80502/1100-1190, P2204-7e, 538 x 1190, GSC 92015.

Figure 8. *Ceratiopsis* sp. J-43 of Staplin
C-80502/3000-3090, P2204-26e, 547 x 1282, GSC 92016.

Figure 9. Peridinioid sp. C (cf. *Maduradinium turpis* Norris)
C-80502/2800-2890, P2204-24e, 525 x 1135, GSC 92017.

Figure 10. Peridinioid sp. C (cf. *Maduradinium turpis* Norris)
C-80502/2400-2490, P2204-20e, 590 x 1214, GSC 92018.

Figure 11. Peridinioid sp. B
C-80502/2500-2590, P2204-21e, 700 x 1183, GSC 92019.

Figure 12. *Platycarya* sp.C-80502/6500-6590, P2204-61i, 616 x 1143, GSC 92023.

Figure 13. *Engelhardtia* sp.
C-80502/1300-1390, P2204-9f, 512 x 1196, GSC 92025.

Figure 14. Peridinioid sp. A
C-80502/2500-2590, P2204-21e, 579 x 1254, GSC 92020.

Figure 15. *Nyssa* sp.
C-80502/1300-1390, P2204-9e, 605 x 1225, GSC 92021.

Figure 16. *Carya* sp.
C-80502/6500-6590, P2204-61i, 601 x 1153, GSC 92022.

Figure 17. *Platycarya* sp.
C-80502/6500-6590, P2204-61i, 610 x 1160, GSC 92024.

Figure 18. *Momipites* sp.
C-80502/6500-6590, P2204-61i, 716 x 1155, GSC 92026.

Figure 19. Ericaceae
C-80502/7200-7290, P2204-68i, 571 x 1110, GSC 92027.

Figure 20. *Juglans* sp.
C-80502/700-790, P2204-3e, 677 x 1174, GSC 92028.

Figure 21. *Pterocarya* sp.
C-80502/1300-1390, P2204-9e, 675 x 1191, GSC 92029.

Figure 22. *Carya* sp.
C-80502/750-790, P2204-3Bb, 544 x 1278, GSC 92030.

Plate 2

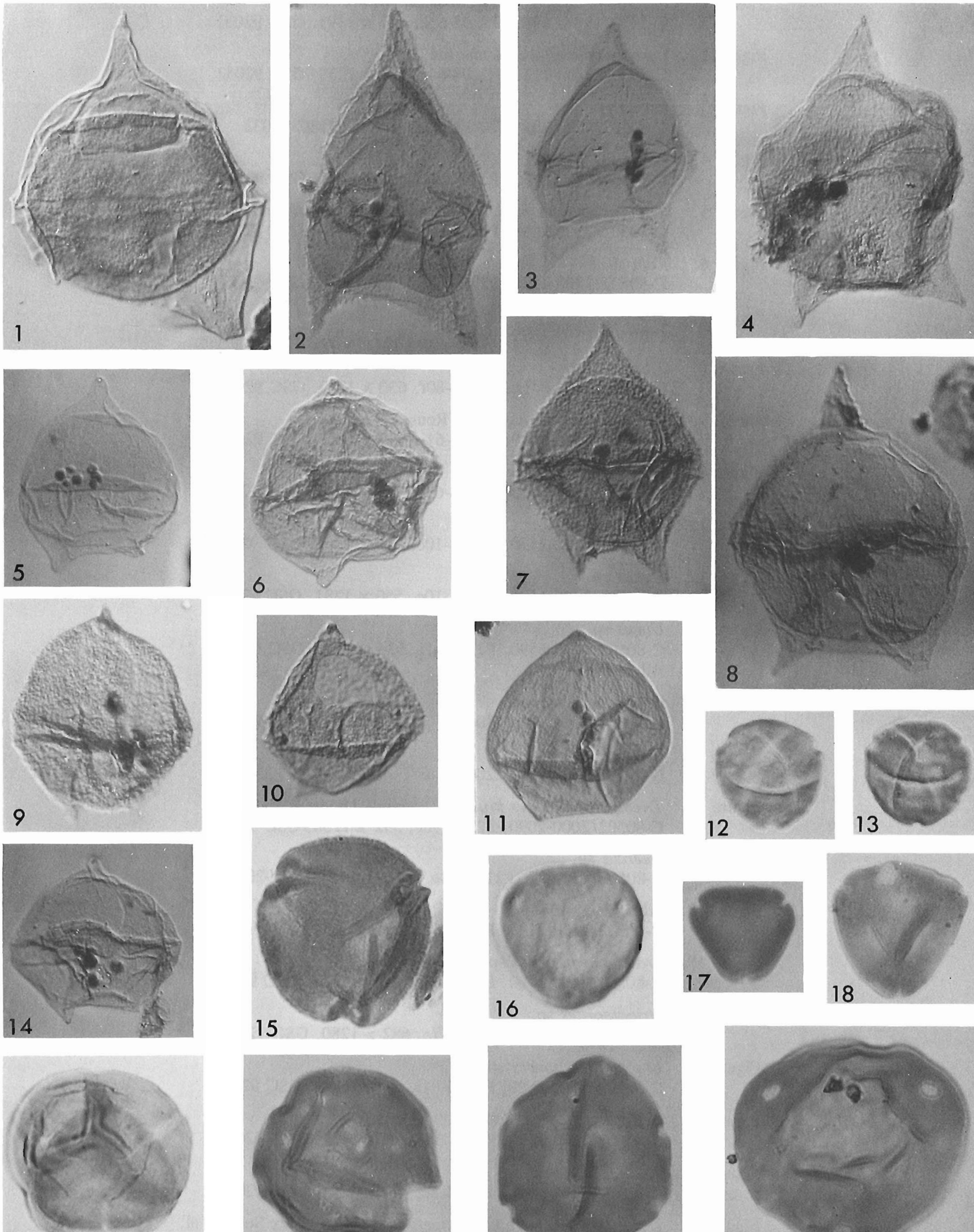


PLATE 3

All figures x1000

Figure 1. *Paraalnipollenites alterniporus* (Simpson) Srivastava
C-80502/6700-6790, P2204-63k, 457 x 1191, GSC 92031.

Figure 2. *Paraalnipollenites alterniporus*
C-80502/2750-2770, P2204-86c, 641 x 1233, GSC 92032.

Figure 3. *Betula* sp.
C-80502/600-690, P2204-2c, 625 x 1257, GSC 92033.

Figure 4. *Alnus* sp.
C-80502/8200-8290, P2204-78f, 601 x 1156, GSC 92034.

Figure 5. *Triporopollenites mullensis* (Simpson) Rouse and Srivastava
C-80502/7000-7090, P2204-66k, 511 x 1164, GSC 92035.

Figure 6. *Tilia* sp.
C-80502/1300-1390, P2204-9e, 412 x 1258, GSC 92036.

Figure 7. *Tilia* sp.
C-80502/1300-1390, P2204-9e, 547 x 1294, GSC 92037.

Figure 8. *Tilia* sp.
C-80502/8400-8490, P2204-80f, 630 x 1283, GSC 92038.

Figure 9. *Pistillipollenites mcgregorii* Rouse
C-80502/6500-6590, P2204-61i, 630 x 1202, GSC 92039.

Figure 10. *Pistillipollenites mcgregorii*
C-80502/6700-6790, P2204-63k, 527 x 1187, GSC 92040.

Figure 11. *Liquidambar* sp.
C-80502/1400-1490, P2204-10e, 615 x 1264, GSC 92041.

Figure 12. *Ulmus* sp.
C-80502/1400-1490, P2204-10e, 550 x 1281, GSC 92042.

Figure 13. *Ulmus* sp.
C-80502/6500-6590, P2204-61i, 631 x 1174, GSC 92043.
All figures x1000

Figure 14. Chenopodiaceae
C-80502/500-590, P2204-1c, 561 x 1159, GSC 92044.

Figure 15. Gramineae
C-80502/500-590, P2204-1c, 677 x 1169, GSC 92045.

Figure 16. *Tricolporopollenites kruschii* (Potonie) Thomson and Pflug
C-80502/2000-2090, P2204-16e, 592 x 1250, GSC 92046.

Figure 17. *Quercus* sp.
C-80502/750-790, P2204-3Bc, 586 x 1140, GSC 92047.

Figure 18. *Liliacidites* sp.
C-80502/7100-7190, P2204-67k, 586 x 1181, GSC 92048.

Figure 19. *Brachysporisporites* sp.
C-80502/3400-3490, P2204-30e, 644 x 1276, GSC 92049.

Figure 20. *Striadiporites* sp.
C-80502/1100-1190, P2204-7e, 682 x 1280, GSC 92050.

Figure 21. *Brachysporisporites* sp.
C-80502/6500-6590, P2204-61i, 654 x 1140, GSC 92051.

Figure 22. *Pesavis tagluensis* Elsik and Jansonius
C-80502/3500-3590, P2204-31e, 519 x 1131, GSC 92052.

Figure 23. *Azolla* sp.
C-80502/1200-1290, P2204-8e, 712 x 1141, GSC 92053.

Figure 24. *Diporicellaesporites laevigataeformis* Ke et Shi ex Sung et al.
C-80502/3400-3490, P2204-30e, 589 x 1130, GSC 92054.

Figure 25. *Ctenosporites wolfei* Elsik and Jansonius
C-80502/4500-4590, P2204-41e, 660 x 1237, GSC 92055.

Plate 3

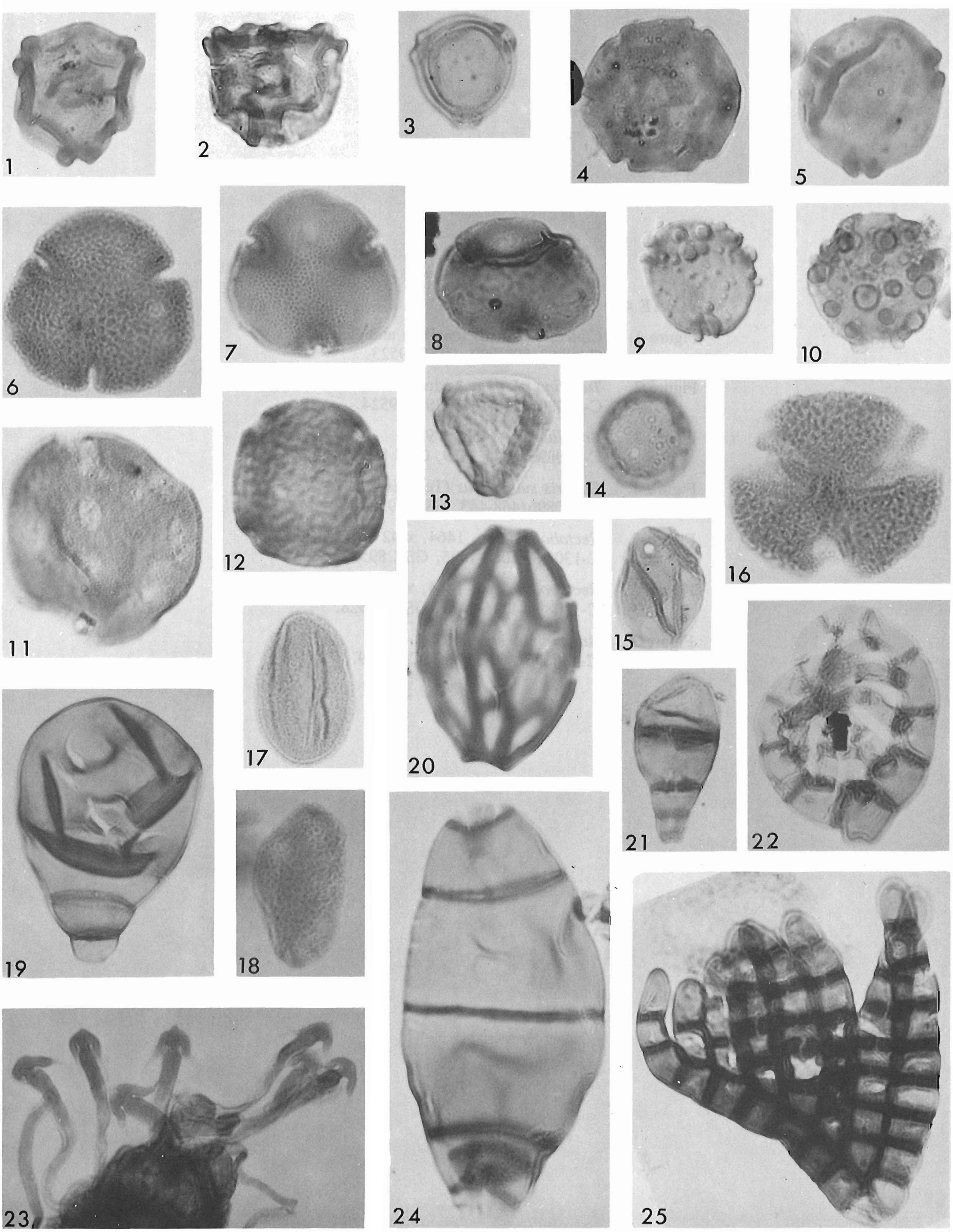


PLATE 4

Figure 1. *Cibicides grossus* ten Dam and Reinhold, x 46
C-80502/500-600, GSC 89523.

Figure 2. *Turrilina alsatica* Andreae, x 65
C-130896/216-231, GSC 89524.

Figure 3. *Brizalina* sp. 1435, x 93
C-130896/1044-1059, GSC 89526.

Figure 4. *Cancris subconicus* (Terquem), x 74
C-130896/468-483, GSC 89525.

Figure 5. *Rectobolivina* sp. 1464, x 42
C-130896/1080-1095, GSC 89527.

Figure 6. *Rectobolivina* sp. 1464, x 34
C-130896/1116-1131, GSC 89528.

Figure 7. *Anomalinoides* sp. 1600, x 52
C-130896/360-375, GSC 89529.

PLATE 4

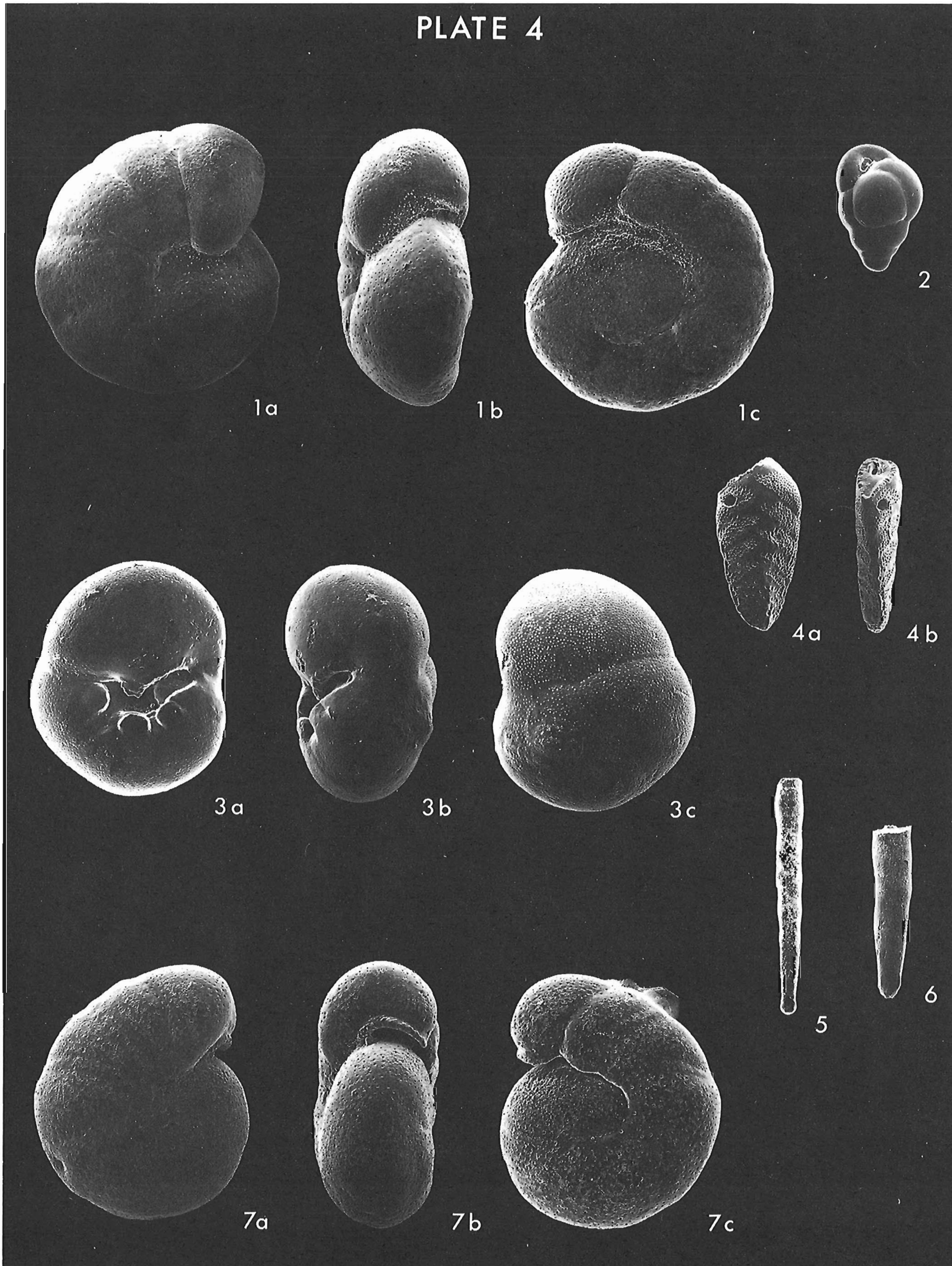
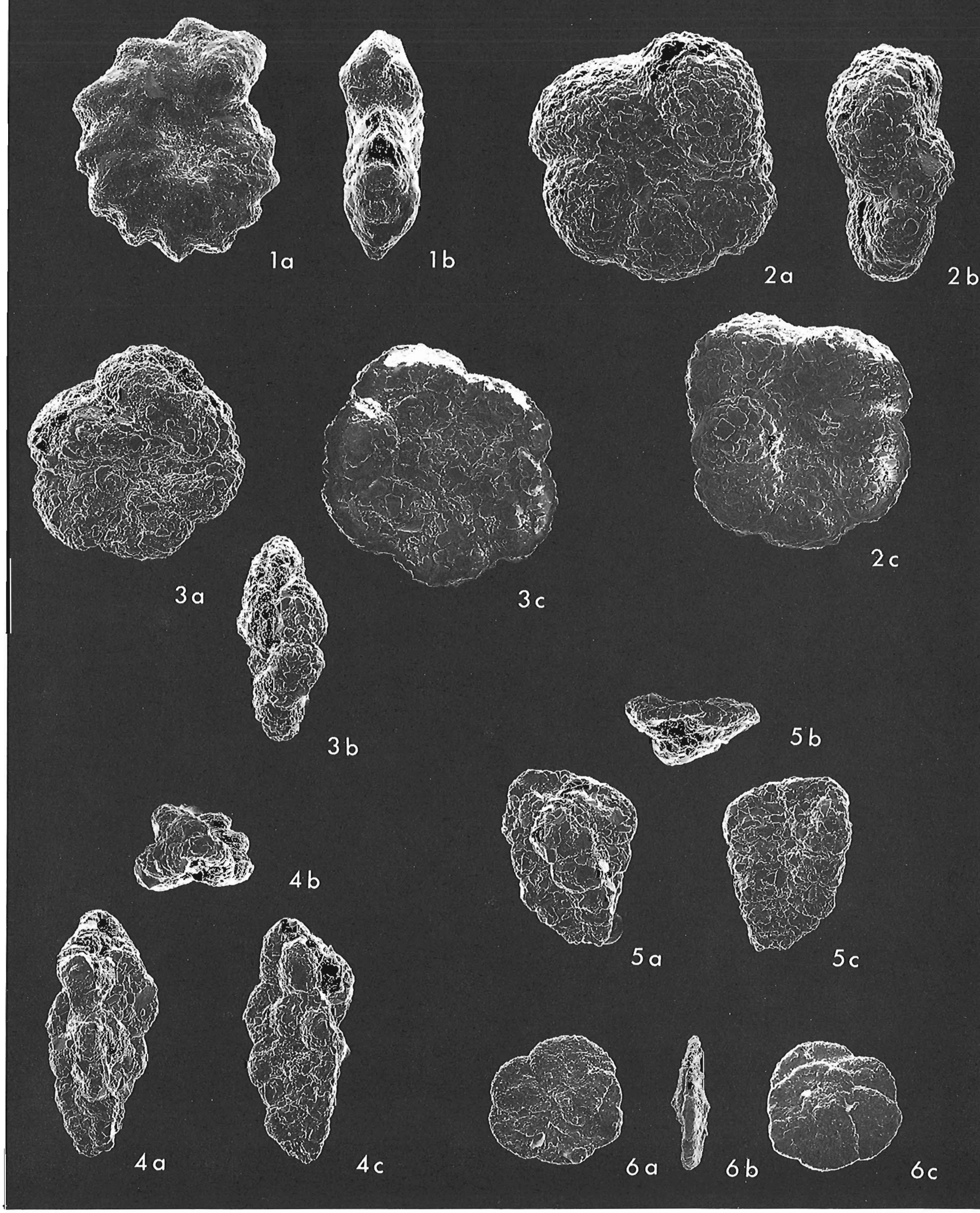


PLATE 5

Figure 1.	*Jadammina statuminis* McNeil, x 68 C-130896/1476-1491, GSC 89530.
Figure 2.	*Trochammina* sp. 2850, x 87 C-130896/2529-2531, GSC 89531.
Figure 3.	*Portatrochammina* sp. 2850, x 75 C-80502/4700-4800, GSC 89532.
Figure 4.	*Verneuilina* sp. 2700, x 68 C-80502/4200-4300, GSC 89533.
Figure 5.	*Verneuilina* sp. 2700, x 71 C-80502/4200-4300, GSC 89534.
Figure 6.	*Portatrochammina* sp. 2849, x 95 C-80502/8610-8640, GSC 89535.

PLATE 5



A brief assessment of potential hydrocarbon source rocks of the Canadian Arctic Archipelago[†]

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Goodarzi, F. and Stewart, K.R., A brief assessment of potential hydrocarbon source rocks of the Canadian Arctic Archipelago; in Current Research, Part G, Geological Survey of Canada, Paper 89-1G, p. 159-161, 1989.

Abstract

Vitrinite reflectance versus depth profiles have been compiled for a total of 124 boreholes in the Sverdrup and Franklinian basins. Petrology and maturity of 200 Devonian coal samples has also been determined.

Good to excellent quality petroleum source rocks occur in four stratigraphic units within the Schei Point Group. Good quality, oil prone source rocks also have been identified in the Jameson Bay Formation (Lower Jurassic) and in the Ringnes Formation (Upper Jurassic).

The bituminous shales of the Cape Phillips Formation (Upper Ordovician to Lower Devonian) and Eids Formation (Middle Devonian) are the best quality potential hydrocarbon source rocks in the Franklinian Basin.

The liptinite rich coals of Devonian age have a potential to generate hydrocarbons.

Résumé

Des profils de réflectance de la vitrinite en fonction de la profondeur ont été compilés pour un total de 124 trous de sonde dans les bassins de Sverdrup et de Franklin. La pétrologie et la maturité de 200 échantillons de charbon du Dévonien ont aussi été déterminées.

On trouve des roches mères pétrolifères de qualité bonne à excellente dans quatre unités stratigraphiques du groupe de Schei Point. D'autres roches mères de bonne qualité ont aussi été identifiées dans la formation de Jameson Bay (Jurassique inférieur) et dans la formation de Ringnes (Jurassique supérieur).

Les schistes bitumineux de la formation de Cape Phillips (Ordovicien supérieur à Dévonien inférieur) et de la formation d'Edis (Dévonien moyen) sont les meilleures roches mères possibles du bassin de Franklin.

Les charbons riches en liptinite d'âge dévonien pourraient se prêter à la production d'hydrocarbures.

[†] Contribution to Frontier Geoscience Program.

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MATURATION AND SOURCE ROCK QUALITY STUDY

Sverdrup Basin

Substantial information has been assembled on oil-prone source rocks in both the Sverdrup Basin and the Franklinian Geosyncline in the Canadian Arctic Archipelago (Fig. 1). Significant progress has been made in assessing local and regional variations in the distribution, thickness, organic-richness, kerogen type or quality, and level of organic maturity (see the references) of the source rocks in the basins. The degree of organic maturity attained by potential hydrocarbon source rocks has been evaluated primarily from vitrinite reflectance measurements. To date, vitrinite reflectance versus depth profiles have been compiled for a total of 89 out of 112 boreholes in the Sverdrup Basin and for a total of 35 out of 59 boreholes in the Franklinian Geosyncline. Interpretation of this data indicates that there is ample evidence from both basins for discontinuous, segmented log linear and curvilinear Ro gradients that can be related to specific geological causes such as:

1. Significant changes in the paleo heat flow associated with periods of tectonism and differences in the thermal conductivity of the various rock types in the stratigraphic sequence.
2. Marked changes in subsidence and sedimentation rates.
3. Igneous intrusives.
4. Hydrothermal anomalies due to the movement of superheated waters through porous aquifers.
5. Major erosional unconformities.
6. Possible effects of thrust faults.

Samples from all formations containing possible hydrocarbon source rocks in the Sverdrup Basin have been evaluated by Rock-Eval/TOC analyses of hand-picked samples of drill cuttings to assess their organic richness (TOC content), type or quality of organic matter (HI, S1 + S2/S3 values), and state of thermal maturity (Tmax and PI values).

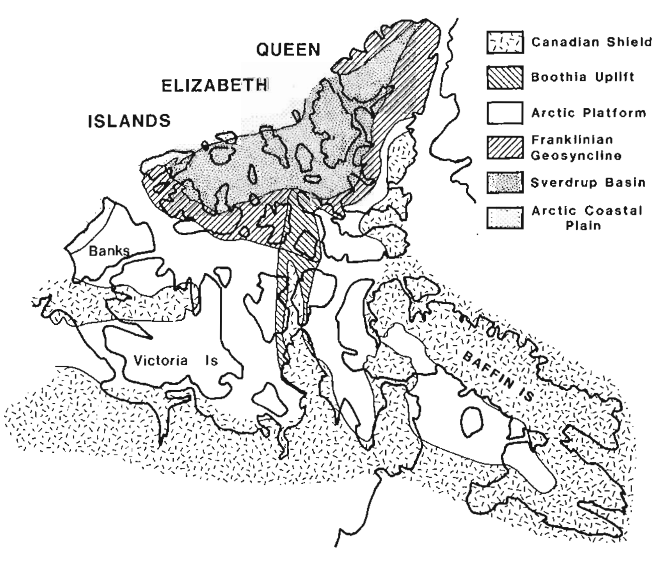

Figure 1. Geological provinces of the Canadian Arctic Archipelago (after Thorsteinsson and Tozer, 1970).

Good to excellent petroleum source rocks, comprising of Type II or an admixture of Types II and I kerogens, occur in four stratigraphic units within the Schei Point Group (Middle-Upper Triassic) in the Sverdrup Basin. For a general description of this Group see Embry (1984a). Oil-prone sources within these four stratigraphic units are widely distributed in the Sverdrup Basin, and the quality, maturity and thickness of the source rocks vary considerably. The maturity of these oil-prone source rocks ranges from immature to marginally mature along a relatively narrow belt parallel to the basin margins, to mature and overmature in the basin centre.

Other moderate to good quality oil-prone source rocks have been identified in the Jameson Bay Formation (Lower Jurassic) and in the Ringnes Formation (Upper Jurassic). For a general description of these formations see Balkwill et al. (1977) and Embry (1984b). The main limiting factor for hydrocarbon generation from these formations is that they have not reached a sufficient level of maturity over most of the western portion of the basin.

Results obtained from detailed geochemical and organic petrological studies of hand-picked samples from the Schei Point Group have contributed to the publication of two papers (Brooks et al., in press; and Goodarzi et al., in press a). A regional maturation and source rock quality study of the sedimentary strata penetrated by boreholes in the environs of Lougheed Island is in progress and the results will be published in a future paper.

Franklinian Geosyncline — maturation and source rock quality study

In the Franklinian Geosyncline, the best quality potential oil-source rocks are bituminous shales in the Cape Phillips Formation (Upper Ordovician to Lower Devonian) and the Eids Formation (Middle Devonian). For a general description of these formations see Christie et al. (1981). These shales are deep water basinal equivalents of the Allen Bay, Read Bay and Blue Fiord formations, which comprise mainly shelf carbonates and reefs. The bituminous nature and dark colour exhibited by these shales suggest that they were deposited, at least in part, under highly anoxic conditions, favorable for the development of oil-prone source rocks. An evaluation of vitrinite reflectance data indicates that the Ro gradients change from log linear to curvilinear with depth. This suggests that the Ro gradients have been affected by either significant changes in paleo heat flow or by differences in the thermal conductivity of the various rock types in the stratigraphic column, or possibly a combination of both factors. Oil-prone source rocks in the Cape Phillips and Eids formations are post-mature, with respect to oil generation, at most of the borehole locations evaluated. It is postulated that the high maturation levels attained by these oil-prone source rocks are mainly due to differences in the thermal conductivity of the various rock types in the stratigraphic column. The Ro gradients are generally low through the underlying thermally conductive evaporites and carbonates of the Bay Fiord, Thumb Mountain, Read

Bay/Allen Bay formations, and high through the Cape Phillips, Eids and Cape de Bray shales, which have a relatively low thermal conductivity and a corresponding high Ro gradient. The Ro gradients, above the source rock intervals, through the Middle-Upper Devonian sandstones of the Weatherall, Hecla, Beverly Inlet, and Parry Islands formations, are low due to their relatively high thermal conductivity.

There is ample evidence in the form of abundant bituminoids and live oil-staining that considerable quantities of oil were once present in reservoirs and/or migrated through rocks of the lower Paleozoic sequence of the Franklinian Geosyncline. Detailed organic petrological and geochemical studies will be performed on these bituminoids in an attempt to correlate them with suspected oil source beds such as those of the Cape Phillips and Eids formations.

DEVONIAN COAL — A POSSIBLE HYDROCARBON SOURCE

Devonian coals from Melville Island occur in Givetian, Frasnian and Famennian strata of the Weatherall, Hecla Bay and Beverley Inlet formations.

Most of the coals are finely banded and are semi-bright. The dominant lithotype is clarain; vitrain is less common. Fusain and durain are very rare. The vitrinite content is low to moderate, consisting of telo-and desmocollinite, whereas the most conspicuous liptinite maceral is sporinite, megaspores in particular. Cutinite, alginite, resinite, exsudatinite and fluorinite are present in small amounts.

The microlithotype composition of the coals (liptinite-rich clarite, vitrinertoliptite) indicates deposition under water, in bays and swamps of the interfluvial areas near the coastline.

Boron content (4-69 ppm), petrology (cannel and canneloid) and sedimentology indicate that these coals were deposited in fresh to semi-brackish water in a lower delta plain setting.

The reflectance (0.57-0.92 R_0max) and spectral fluorescence of the Devonian coals indicate a rank of high-volatile bituminous C-A, which is within the mature zone of hydrocarbons for type II and III kerogen (exinitic-vitrinitic, Brooks et al., 1987). Due to the high liptinite content of these coals, they have considerable potential to generate hydrocarbon.

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A summary report on mass transfer of elements in Middle Triassic shale-sandstone successions, Sverdrup Basin, Canadian Arctic[†]

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Foscolos, A.E., A summary report on mass transfer of elements in Middle Triassic shale-sandstone successions, Sverdrup Basin, Canadian Arctic; in Current Research, Part G, Geological Survey of Canada, Paper 89-1G, p. 163-166, 1989.

Abstract

In this study, diagenetic changes in the fine mineral fraction of shales have been observed by examination of X-ray diffractograms, and by thermal and chemical analysis. These changes may be observed over vertical intervals of as little as 2 m, provided that a drainage system, such as a permeable sandstone, occurs adjacent to the shale. This work contradicts previous research, which implied that chemical and mineralogical changes, that is catagenetic alterations, cannot be recognized over intervals of less than several hundred metres thickness. The quantitative aspects of these changes were not evaluated owing to lack of sufficient samples from both shales and sandstones.

It appears that in this sequence of Triassic calcareous sandstones and limestones, the formation of authigenic interstratified 2:1 hydrous layer silicates, illite-smectite-vermiculite, is a result of the affects of organic matter. It is suspected that chelation of divalent and trivalent ions (Ca, Mg, Al, Fe) by organic matter is responsible for the transport of elements from the shales to the sandstones, since the pH values of the carbonate rocks are around 8.0.

Résumé

Dans la présente étude, on a observé des changements diagénétiques dans la fraction minérale fine de schistes argileux en examinant des diffractogrammes de rayons X et en procédant à des analyses thermique et chimique. Ces changements peuvent s'observer dans des intervalles verticaux d'à peine 2 m, pourvu qu'un système de drainage, comme un grès perméable, juxte les schistes argileux. Cette étude infirme une recherche antérieure qui indiquait que des changements chimiques et minéralogiques, c'est-à-dire des altérations catagénétiques, ne peuvent être décelés dans des intervalles inférieurs à plusieurs centaines de mètres d'épaisseur. Les aspects quantitatifs de ces changements n'ont pas été évalués à cause du nombre insuffisant d'échantillon de schistes argileux et de grès.

Il semble que dans cette séquence de grès calcaires et de calcaires du Trias, la formation de phyllosilicates hydratés (illite, smectite et vermiculite) authigènes et interstratifiés de type 2:1 soit due aux effets de la matière organique. On pense que la chelation des ions bivalents et trivalents (Ca, Mg, Al, Fe) par la matière organique serait responsable du transport d'éléments depuis les schistes argileux vers les grès, car le pH des roches carbonatées se situe autour de 8,0.

[1] Technical University of Crete, Greece

[†] Contribution to Frontier Geoscience Program.

INTRODUCTION

To date, the main thrust in the study of catagenesis in shales and sandstones has been to observe mineralogical changes as detected by X-ray diffraction studies, and to relate these transformations to burial depth and/or temperature. Additional work has been carried out relating the generation of hydrocarbons from source rocks to the removal of certain discrete layer silicates, such as kaolinite, and the stepwise transformation of smectite to an interstratified 2:1 layer silicate, such as an illite/smectite. An important aim of this study was to define the oil generating window of organic-rich, fine grained source rocks. A vitrinite reflectance (Ro) of 0.5, which usually is related to the onset of oil generation, was found to coincide with a d_{001} spacing of 12.5 Å of the interstratified clays (Foscolos et al., 1976; Foscolos and Powell, 1980). This "sloughing-off" of the water, as observed from the d_{001} spacing migration from 15.8 Å to 12.5 Å, takes place around 100°C (Foscolos and Kodama, 1974).

The effect of organically derived, organic acids from thermal (<80°C) and bacterial decarboxylation of organic matter during the early stages of diagenesis of organic matter (eodiagenesis) has been discussed recently (Surdham and Crossey, 1985, 1987; Gauthier et al., 1985), not only in relation to the dissolution of feldspars and the ensuing chelation and transport of aluminum from source rocks to sandstones, but also in connection with the creation of secondary porosity.

Since copious amounts of water and water-soluble acids evolve during diagenesis of organic matter and catagenesis of minerals of fine grained, organic-rich source rocks, substantial concentrations of elements may be transported to reservoir sandstones. Very little pertinent data concerning the influence of diagenesis and catagenesis on the elemental composition of shales and adjacent sandstones exists, however. The mass transfer of elements and the mechanism of transport from source rocks to reservoir rocks, and its implication in understanding the compositional variation within reservoir rocks, has not been investigated thoroughly.

As a result, this research study was undertaken in an attempt to understand the ultimate destination of the elements mobilized as a function of mineral dissolution (including feldspars), and mineral transformations such as smectite converting to 1 Md illite through the intermediate step of an illite/smectite interstratification process.

SVERDRUP BASIN STUDY

Setting and samples

For this mineralogical and geochemical study, cores from a portion of the Middle Triassic of the Sverdrup Basin were retrieved from two wells drilled in the Arctic Islands during 1985. The Eldridge Bay (Anisian) and Cape Caledonia (Ladinian) members of the Murray Harbour Formation were cored in Panarctic et al. East Drake L-06 and Panarctic et al. Skybattle Bay M-11 (Fig. 1). Two continuous, full diameter, 18 m cores were cut in the East Drake L-06 well from 1122.6 to 1159.0 m. This well is a shut-in Jurassic gas well located at 76°25′35″ north latitude and 107°33′11″ west longitude, immediately east of the Drake gas field on Sabine Peninsula, Melville Island. A second, 18 m, full diameter core was cut in the Skybattle Bay M-11 well from 2520.0 to 2538.0 m. This well is located at 77°10′56″ north latitude and 105°06′44″ west longitude, on the southern tip of Lougheed Island (Fig. 1 and 2). The geological and stratigraphic setting of these cores, and a brief description, are given by Embry and Podruski (1988).

Fifty-one samples of calcareous shales, limestones, and calcareous sandstones were collected from the cores of the East Drake L-06 well and forty-four samples of similar lithology were collected from the core of the Skybattle Bay M-11 well.

Results

Data on the mineralogy, inorganic and organic geochemistry of whole rock samples as well as scanning electron microscopy on chips obtained from the whole rock samples of both wells have been released by Foscolos et al. (1988). These results do not provide information concerning mineral transformations within shales or sandstones that may indicate transfer of elements, a fact attributed to the overabundance of carbonate minerals in all rock types encountered. Carbonate minerals in quantity mask the identification of layer silicates produced in mineral transformations.

Further research work was undertaken to study the fine fractions of shales and sandstones following removal of carbonates. Hydrous layer silicates are concentrated in this fine fraction, and the effect of diagenesis should be discernible. Physiochemical changes affect hydrous layer silicates more completely than any other mineral of the earth's crust because these silicates are the by products of equilibrium between the hydrobiosphere and rocks. Changes in equilibrium can be detected by the transformation of hydrous layer silicates, as identified by X-ray diffraction, thermal analysis, chemical techniques and scanning electron microscopy. Following this approach, fifty-one carbonate-free $<.2\mu$ fractions were collected by centrifuging the samples of calcareous shales, limestones and calcareous sandstones from the cores of the East Drake L-06 well, and forty-four similar fractions from the Skybattle Bay M-11 well samples.

X-ray diffraction results on the fine fractions show d_{001} peak movement of the hydrous 2:1 interstratified layer silicates upon descending from the shale to the shale/sandstone interface, to the sandstone. The d_{001} spacing is expanded initially, then collapsed, and finally re-expanded, indicating that water has been progressively expelled from the layer silicates downward from the shale toward the shale/sandstone interface. Finally, water has been reabsorbed on the layer silicates within the sandstones. X-ray diffraction data are substantiated by thermogravimetric analysis results (Foscolos, in press). The reason for these transformations appears to be related to elemental changes that have taken place within the crystal lattices of the layer silicates. Chemical elements including K, Al, S, and Mg have been transferred from shales to sandstones, resulting in the construction of new hydrous layer silicates. This pattern of interstratified layer silicates has been identified three times

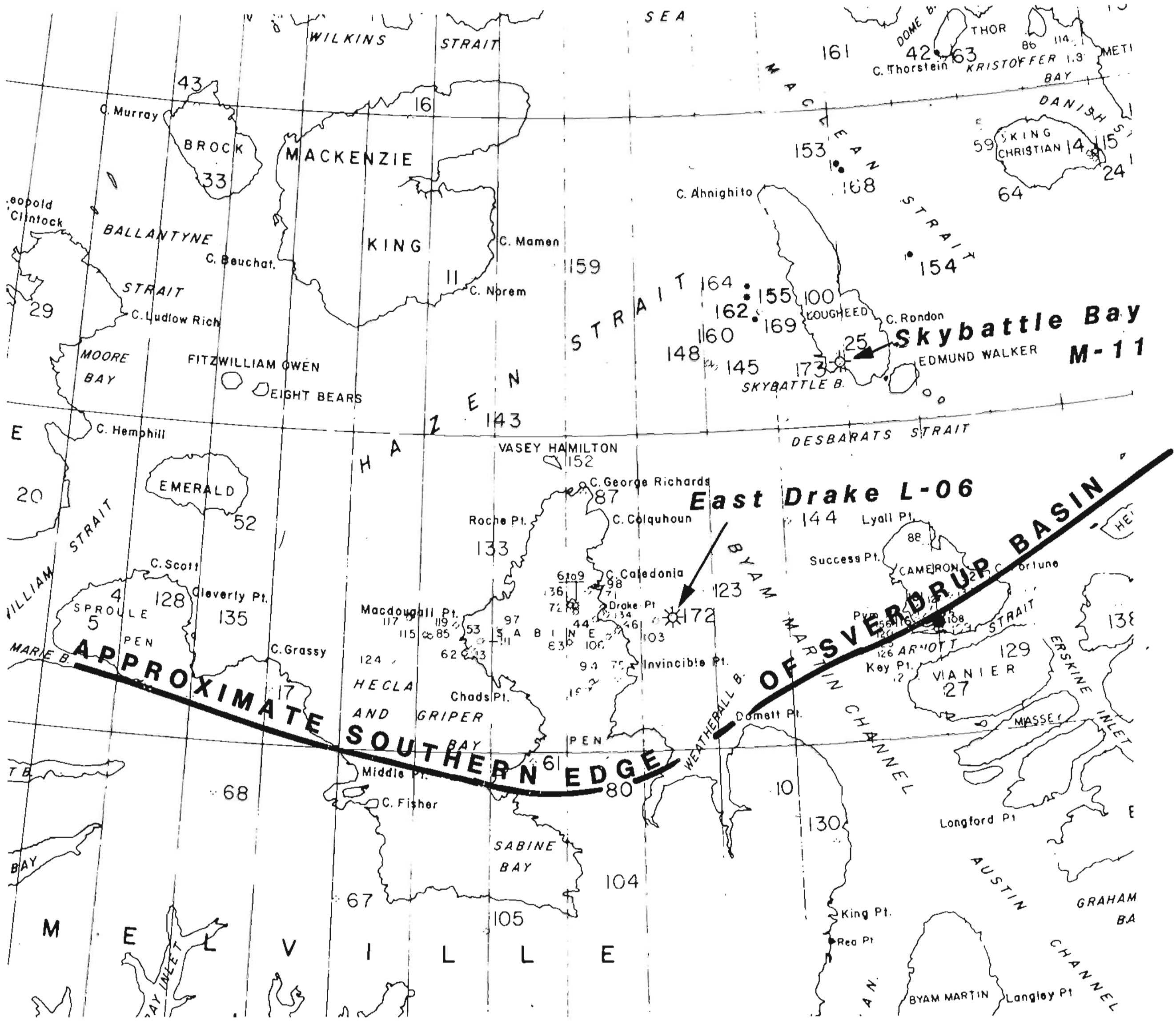

Figure 1. Southwestern Sverdrup Basin, Arctic Islands, showing locations of the East Drake L-06 and Skybattle Bay M-11 wells.

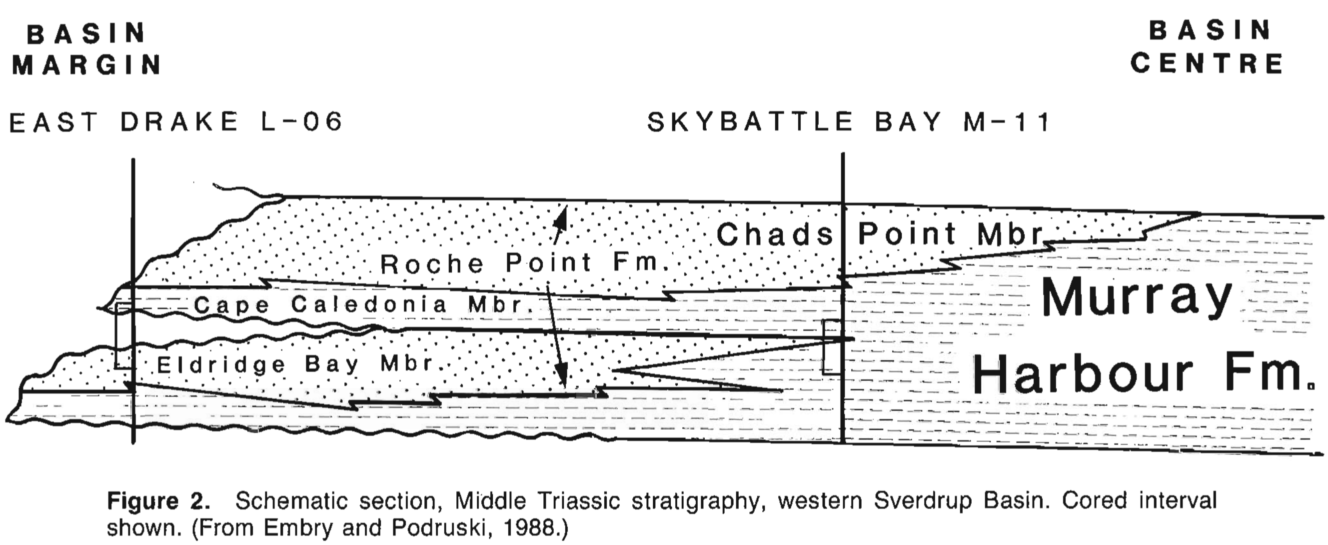

Figure 2. Schematic section, Middle Triassic stratigraphy, western Sverdrup Basin. Cored interval shown. (From Embry and Podruski, 1988.)

within the shale-sandstone succession studied from the East Drake L-06 well and once within the Skybattle Bay M-11 well core. Data from chemical analysis of the $< .2\mu$ fraction substantiate the above explanation. However, SEM work on thin sections was not helpful in identifying authigenic clays in the pore throats of the sandstones, because of the abundance of carbonate minerals. Results of this second phase of the project are being released by Foscolos (in press).

In conclusion, diagenetic changes in the fine mineral fraction have been observed by X-ray diffraction, thermal analysis and chemical analysis within 2 m intervals of sediments, where a drainage system, such as permeable sandstone, exists adjacent to the shales. Quantitative aspects of these changes were not evaluated owing to lack of sufficient samples from both shales and sandstones. Thus we were unable to proceed with detailed chemical work, such as differential dissolution techniques, in order to quantify changes in the amounts of quartz and various feldspars, and in the amounts of amorphous (non-crystalline) material, which are difficult to detect by X-ray diffraction work.

To quantitatively analyze diagenetic changes, carbonate-poor or carbonate-free samples from an area with detailed well control (abundant cores) should be used for comparative analyses. Such a research project in a well-drilled oil or gas field should promote our understanding of how enhanced hydrocarbon recovery techniques will affect the transfer of elements from source rocks to sandstones. Upon precipitation, these elements, carbonates, silicates etc. create new minerals that plug the pores of the reservoir rocks. This process alters reservoir properties, a problem of immense significance to reservoir engineers.

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Sequence stratigraphy, lithostratigraphy, and hydrocarbon potential of the subsurface upper Paleozoic section of Sabine Peninsula, Melville Island, Canadian Arctic Archipelago†

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Majid, A.H., Sequence stratigraphy, lithostratigraphy, and hydrocarbon potential of the subsurface upper Paleozoic section of Sabine Peninsula, Melville Island, Canadian Arctic Archipelago; in Current Research, Part G, Geological Survey of Canada, Paper 89-1G, p. 167-176, 1989.

Abstract

Five major depositional sequences are recognized in the subsurface Carboniferous and Permian of southwestern Sverdrup Basin on Sabine Peninsula. The database incorporates descriptions of cores and cuttings from seven wells supplemented by biostratigraphic, log, and seismic data to build a stratigraphic and depositional sequence framework extending from a shallow basin margin in the south of Sabine Peninsula to a proximal deeper basin in the central and northern peninsula. Some subsurface refinements are proposed relating to basinal facies in the Trold Fiord and Degerböls formations.

Although geochemical data show that the van Hauen and Hare Fiord formations are gas-prone source rocks and are overmature, the possibility of finding oil-prone source rocks still exists. The Canyon Fiord, Assistance and Sabine Bay formations are the most promising reservoir units. The area between the Eldridge and Weatherall wells in the upper and lower parts of the half-graben structures should be investigated to define reservoirs and closures in the upper Paleozoic section.

Résumé

Cinq séquences de sédimentation importantes sont reconnues pour le Carbonifère et le Permien sur la péninsule de Sabine, dans la partie sud-ouest du bassin de Sverdrup. La base de données comprend des descriptions de carottes et de déblais provenant de sept puits qui sont complétées par des données biostratigraphiques, des diagraphies et des données sismiques afin de permettre l'établissement d'un cadre stratigraphique et de sédimentation depuis la marge d'un bassin peu profond au sud de la péninsule de Sabine jusqu'à un bassin proximal plus profond dans les parties du centre et du nord de la même péninsule. Des raffinements du faciès sédimentaire profond sont proposés pour les formations de Trold Fiord et de Degerbols.

Quoique les données géochimiques indiquent que les formations de van Hauen et de Hare sont composées de roches mères susceptibles de renfermer du gaz et parvenues au stade d'hypermaturité, la possibilité de découverte de roches mères susceptibles de renfermer du pétrole persiste. Les formations de Canyon Fiord, d'Assistance et de Sabine Bay sont les plus prometteuses quant à la présence d'éventuels réservoirs. La région comprise entre les puits Eldridge et Weatherall dans les parties supérieure et inférieure des structures en forme de demi-graben devrait être étudiée afin d'établir la présence de réservoirs et de fermetures dans le profil d'âge paléozoïque supérieur.

† Contribution to Frontier Geoscience Project

INTRODUCTION

This paper summarizes the results of a study of the sequence stratigraphy, sedimentology and hydrocarbon potential of the Permian and Carboniferous succession in the subsurface of Sabine Peninsula, northern Melville Island, Canadian Arctic Archipelago (Fig. 1). Upper Paleozoic strata of the Sverdrup Basin are well exposed on Axel Heiberg and Ellesmere islands (Fig. 1) and have been studied extensively by several authors. Thorsteinsson (1974) established a stratigraphic framework for the Carboniferous and Permian sequence in the Sverdrup Basin, which was based partly on studies conducted by Troelsen (1950), Harker and Thorsteinsson (1960), Tozer and Thorsteinsson (1964) and Nassichuk(1965, 1967, 1969, 1972). Nassichuk (1975a, b) and Nassichuk and Wilde (1977) examined the upper Paleozoic sequence in many parts of the Sverdrup Basin, and introduced several new rock units. Beauchamp (1987) conducted an extensive stratigraphic and facies analysis of the Canyon Fiord, Belcher Channel and Nansen formations on Raanes Peninsula, southwestern Ellesmere Island.

There are approximately 170 exploratory and development wells in the Canadian Arctic Archipelago. Only 38 of these wells penetrate Carboniferous and/or Permian rocks. Seven of these wells are located in the study area (Fig. 1 and 2), where the upper Paleozoic section shows a facies change from shelf to basin. These wells are: Weatherall O-10, Eldridge Bay E-79, Sherard Bay F-34, Marryatt K-71, Drake Point D-68, Hecla J-60 and Chads Creek B-64. The main objectives of the project were:

1. To interprete the subsurface section in terms of sequence stratigraphy, based on lithology, well logs and seismic profiles. Thorsteinsson's (1974) surface stratigraphic nomenclature was used as a guide.
2. To provide a correlation between lithostratigraphic, biostratigraphic and sequence stratigraphic units.
3. To assess the hydrocarbon potential of the Permian and Carboniferous section on Sabine Peninsula and provide recommendations for future exploration.

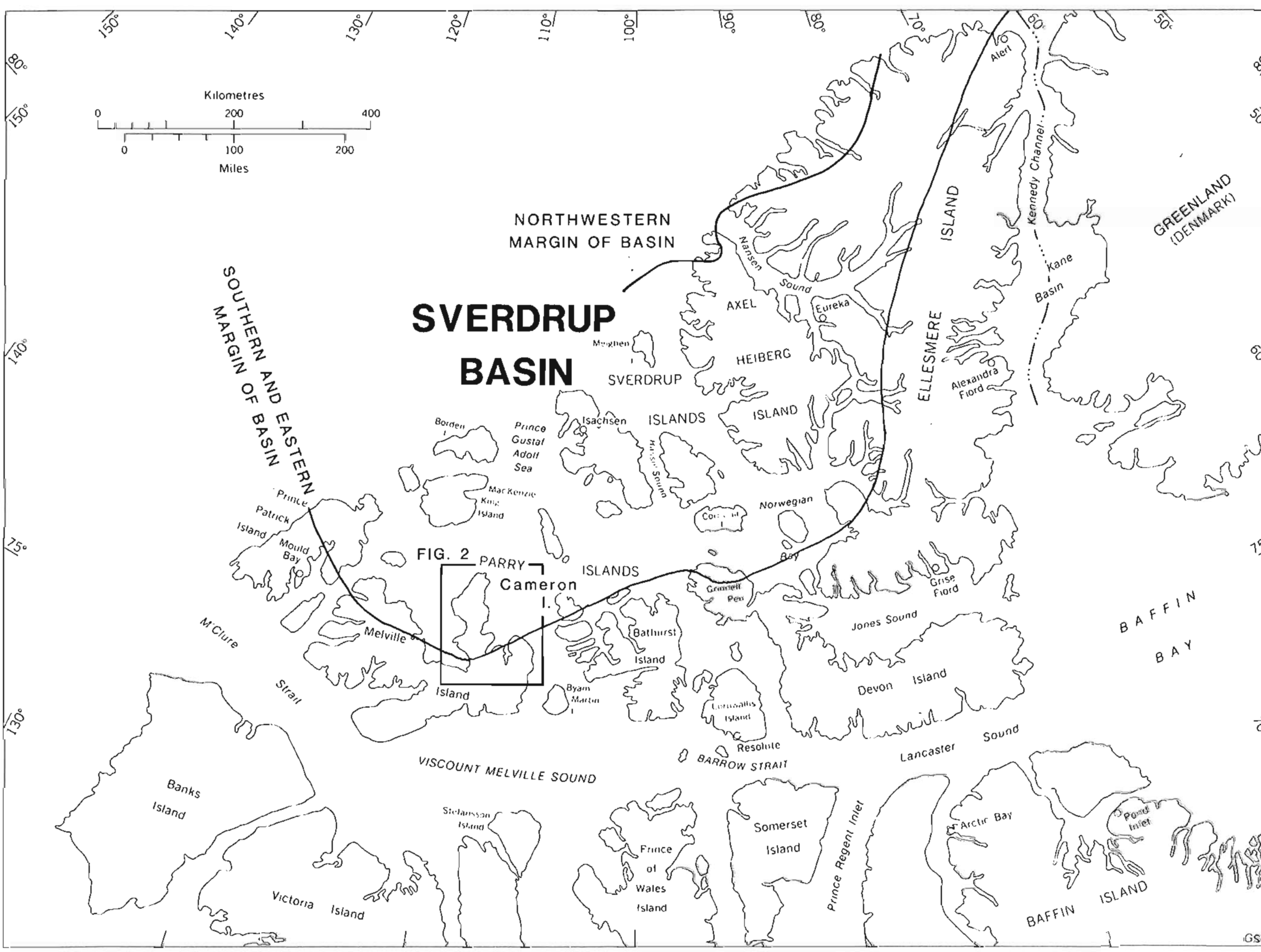

Figure 1. Index map of the northernmost islands of the Canadian Arctic Archipelago showing the upper Paleozoic Sverdrup Basin.

Methodology

A total of 300 m of core from the upper Paleozoic section was examined and described in terms of rock type and texture, porosity, oil shows, diagenetic features, and facies subdivisions. Additionally, 16 000 m of cuttings were examined and described. For detailed lithological identification and diagenetic studies, about 130 thin-sections from cuttings and 130 thin-sections from cores were prepared.

The biostratigraphic framework is based on existing palynological, micro- and macro-fossil reports enhanced by newly acquired palynological data from 250 samples of cuttings and cores examined by Utting (1989). Rock units were correlated between wells and with outcrop sections using lithological and log characteristics, and then interpreted in terms of depositional sequences. These sequences were then compared to sequences defined from seismic profiles of the study area by other authors (Harrison, pers. comm., 1988), and from exposed sections on Melville and Ellesmere islands (Beauchamp et al., 1989).

STRATIGRAPHY

The regional stratigraphic framework for the upper Paleozoic succession in the Sverdrup Basin (Table 1) was established by Thorsteinsson (1974) and was based on surface exposures in the northern and eastern sectors of the basin. The succession reflects deposition on shallow platforms surrounding several deeper basinal areas. Coarse clastics were deposited near the depositional edge, and shallow-water carbonates on the peripheral platform. Deeper-water carbonates, shale, and sulphate and halite evaporites were deposited in the depocentres. Thorsteinsson's stratigraphic framework was based on the recognition of five stratigraphic sequences, the boundaries of which were believed to represent basinwide erosional unconformities. He assigned a specific set of formations to each sequence. Beauchamp et al. (1989) have recently refined the stratigraphy and depositional sequences of the upper Paleozoic section. These refinements are based on surface sections using the sequence stratigraphy concepts documented by Vail et al. (1977).

In this paper, the established surface stratigraphic terminology of Thorsteinsson (1974) is applied to the subsurface without significant modification, and closely follows the stratigraphic framework established by the geological staff of Panarctic Oils Limited, as operators of the exploratory wells used in this project. Thorsteinsson's surface stratigraphic framework is not summarized here; the reader is referred to his 1974 paper, to Beauchamp (1987), and to Davies and Nassichuk (in press).

TECTONIC STYLE OF SVERDRUP BASIN

The Sverdrup Basin is a southwest to northeast trending, elongate, pericratonic depression approximately 1300 km long and 400 km in width, infilled by at least 13 000 m of Lower Carboniferous to Tertiary rocks (Balkwill, 1978). Initial subsidence of the basin was accompanied by crustal attenuation (thinning) and development of a complex continental rift system in Carboniferous time, followed by post-Early Permian thermo- isostatic subsidence (Stephenson et al., 1987).

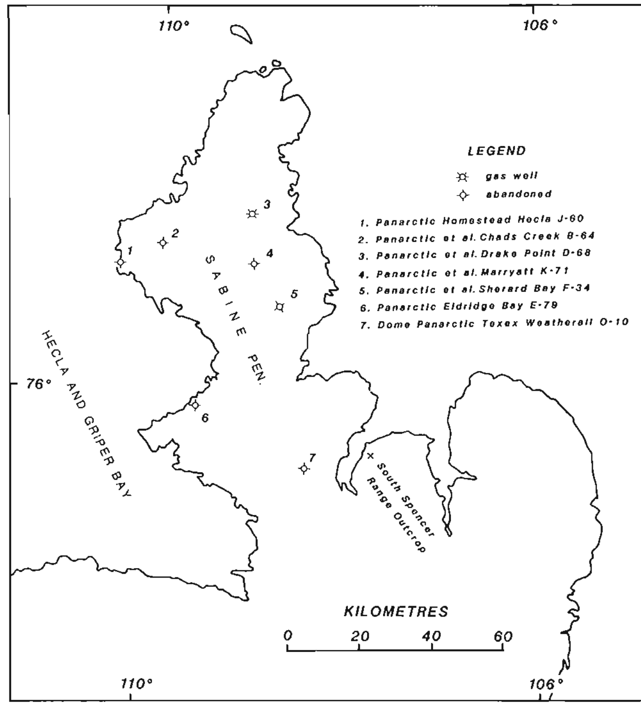

Figure 2. Map of Sabine Peninsula showing the location of the exploratory wells and outcrop section.

The oldest rocks in the Sverdrup Basin are Viséan, lacustrine marlstones that were associated with early synrift redbed conglomerates and finer siliciclastics that are interpreted as recording the onset of crustal thinning and thermal uplift (Beauchamp et al., 1989). The subsequent thermal collapse and extensional component of basin formation is recorded by Namurian to Sakmarian syntectonic conglomerates and clastics along the rift margin, a broad belt of marine shelf carbonates, the basal rift evaporites, and an overlying, deep-water, rift-centre shale and limestone facies. An episode of minor folding and faulting is evident along the southern and eastern margins of the basin, termed the Melvillian Disturbance by Thorsteinsson and Tozer (1970). The Melvillian Disturbance occurred from late Sakmarian to early Artinskian time and marked a change to strike-slip tectonic style from extension (Beauchamp et al., 1989). This phase was followed, in turn, by a long period of passive thermal subsidence that extended from Roadian time (uppermost Lower Permian) through to Early Cretaceous time.

Increased tectonic activity and rifting in the Amerasian Basin to the west and northwest of the Sverdrup Basin in Cretaceous time resulted in renewed rifting and increased subsidence in the Sverdrup Basin. Extrusion of tholeiitic basalts in the northwestern part of the Sverdrup Basin accompanied this renewed subsidence, with widespread intrusion of dykes and sills of the same age. These intrusives are also present in the Carboniferous and Permian section in the subsurface of Sabine Peninsula.

The southern and southwestern boundary of the Sverdrup Basin, in the area of northern Melville Island to

Table 1. Table of formations (after Nassichuk and Wilde, 1977). Vertically ruled areas denote disconformities; diagonally ruled areas denote absence of strata due to faulting and erosion.

SYSTEM		NORTH AMERICAN SERIES	STAGES		DEVON ISLAND Grinnell Peninsula	ELLESMERE ISLAND Bjorne Peninsula (South)	Blind Fiord (East Side)	Blind Fiord (West Side)	Blue Mountains (South)	Van Hauen Pass	Hare Fiord (Head)
TRIAS								BLIND FIORD	BLIND FIORD	BLIND FIORD	
PERMIAN	UPPER	Guadalupian	Capitan	Kazanian							
PERMIAN	UPPER	Guadalupian	Wordian	Kazanian		DEGERBOLS		DEGERBOLS		DEGERBOLS	
PERMIAN	LOWER	Leonardian	Roadian	Kungurian	ASSISTANCE	? ? ?	VAN HAUEN ? ?	VAN HAUEN	VAN HAUEN	VAN HAUEN	
PERMIAN	LOWER	Leonardian	Artinskian		BELCHER CHANNEL	UNNAMED FORMATION	UNNAMED FORMATION		HARE	HARE	
PERMIAN	LOWER	Wolfcampian	Sakmarian						? ? ?	? ? ?	
PERMIAN	LOWER	Wolfcampian	Asselian		BELCHER CHANNEL	BELCHER CHANNEL	BELCHER CHANNEL	BELCHER CHANNEL	FIORD	FIORD	NANSEN (Upper)
CARBONIFEROUS	PENNSYLVANIAN	Virgilian	Gzhelian			? ? ?	? ? ?				
CARBONIFEROUS	PENNSYLVANIAN	Missourian	Kasimovian			CANYON FIORD	CANYON FIORD				? ? ?
CARBONIFEROUS	PENNSYLVANIAN	Desmoines	Moscovian		CANYON FIORD				"Tellevak Limestone Member" HARE FIORD	HARE FIORD (Lower)	NANSEN (Lower)
CARBONIFEROUS	PENNSYLVANIAN	Atokan	Moscovian								
CARBONIFEROUS	PENNSYLVANIAN	Morrowan	Bashkirian						OTTO FIORD	OTTO FIORD	
CARBONIFEROUS	MISSISSIPPIAN	Chesteran	Namurian								BORUP FIORD ? ?
CARBONIFEROUS	MISSISSIPPIAN	Chesteran	Visean		EMMA FIORD						

GSC

Sabine Peninsula (Fig. 1), is characterized by a series of normal, listric and strike-slip faults related to initial rifting and to the later Melvillian Disturbance (Harrison, pers. comm., 1988). From south to north along the axis of Sabine Peninsula, Carboniferous and Permian syntectonic redbeds and associated finer clastics and marine limestones in half-grabens are overlain by younger, Carboniferous and Permian siliciclastics and carbonates, and ultimately by Mesozoic and Tertiary rocks. South-to-north seismic sections along the length of the Peninsula demonstrate the presence of southward-dipping listric faults bounding a half-graben at the southern end of the Peninsula, and major down-to-the-basin normal and probable listric faults in the central part of the peninsula, marking the boundary of the main basin depocentre.

The seven wells drilled on Sabine Peninsula (Fig. 2) record the sedimentary response to rifting and thermal subsidence along the southwestern margin of the Sverdrup Basin, and preserve a record of deeper-water environments in the rift trough.

SEQUENCE STRATIGRAPHY

Despite major progress in other exploratory techniques, stratigraphic analysis continues to be a fundamental tool for oil exploration. It provides the context for basin analysis, prediction of reservoirs, source rocks and seals, the interpretation of migration pathways and dating of structural deformation (Vail, 1987). The traditional way to obtain stratigraphic data for analysis is by geological mapping of lithostratigraphic units and their dating through biostratigraphy, using classification and nomenclature established by national or international codes. The most effective basin studies are those that involve mapping of depositional facies and systems rather than simple lithofacies mapping.

Several attempts to relate parachronostratigraphic units to depositional episodes have been made in the last thirty years (Forgoston, 1957; Sloss, 1963; and Bush, 1971). However, the depositional sequence (Vail et al., 1977), a parachronostratigraphic unit with different magnitudes, appears to be the most satisfactory tool to be used in basin analysis. This is because the depositional sequence has important geological implications and using it stands out from classification or nomenclatural procedures. The depositional sequence concept is derived mainly from seismic stratigraphy, and has been documented by Vail et al. (1977) and in associated papers. A depositional sequence is defined as a stratigraphic unit composed of a relatively comparable succession of genetically related strata, bounded at its top and base by unconformities or their correlative conformities (Vail et al., 1977). The fundamental control on a depositional sequence is short-term eustatic changes of sea level superimposed on longer-term tectonic changes (Vail et al., 1977; Vail, 1987).

The following four steps were carried out to establish the stratigraphic sequences:

1. Identification of sequence boundaries in the study area from seismic sections. This was conducted by T.A. Brent and J.C. Harrison of the Geological Survey of Canada, as part of an independent project.
2. Well-log sequence analysis and age determination (this paper); preliminary estimates of sequence and system tracts were conducted by first interpreting the depositional lithofacies on wireline logs using cores and cuttings to calibrate the log. Following this, sequence interpretations were carried out from interpreted lithofacies.
3. Utilization of synthetic seismograms for five wells, conducted by T. Brent. This facilitated correlation of well data with seismic traces, and determined "time lines" at wells that were correlated by means of continuous seismic reflections on data between those wells.
4. Seismic facies analysis to determine as objectively as possible the implications of seismic patterns within individual seismic sequences.

Given 10 to 80 Ma as the duration of second order sequences (Vail et al., 1977), five major depositional sequences were defined in the upper Paleozoic section from seismic profiles and well logs supported by biostratigraphic data and supplemented by surface data (Harrison, pers. comm., 1988) in the vicinity of the Weatherall area (South Spencer Range) on Sabine Peninsula (Fig. 1). Sharp lithological contacts record transgressive events near the intersequence boundaries. These broad sequences are (Figs. 3, 4; Table 4):

Sequence 1. Late Serpukhovian — late Asselian: Canyon Fiord and Hare Fiord formations.

Table 2. Formation tops and sequence boundaries in project wells, Sabine Peninsula.

Formation names and units	J-60	B-64	D-68	K-71	F-34	E-79	0-10
Degerbols, Trold Fiord or equivalent	2557	2990 eq.	3127 eq.	3313 eq.	2640	424	-
Unit A or equivalent	3197	3709 eq.	4197 eq.	4045 eq.	3002 eq.	518	-
Assistance Fm. or equivalent	3277 eq.	3822 eq.	4212 eq.	4150 eq.	3210 eq.	655	75
Sabine Bay Fm. or equivalent	3612	4267 ?	?	?	3345 eq.	744	152
van Hauen Fm.	3054	3604	3313 or 3688	3313 or 3753	2955	Unit A	-
Upper Belcher Ch. Fm. or unnamed fm.	NP	4574	4599 eq. or H.F. Fm.	4705	3695	823	213
Lower Belcher Ch. Fm.	NP	4685	4865 eq.	4900	3883	-	-
Canyon Fiord Fm.	NP	NP	NP	-	4410	853	235
Silur. or Dev.	NP	NP	NP	NP	5270	1323	1186
Sequence Boundaries							
Sequence 5	2557 3277	2990 3822	3127 4212	3313 4150	2640 3210	423 655	? 75
Sequence 4	3277 3277	3822 4574	4212 4599	4150 4705	3210 3695	655 823	75 213
Sequence 3	NP	4574 4685	4599 4865	4705 4900	3695 3883	823 853	213 235
Sequence 2	NP	4685 NP	4865 NP	4900 NP	3883 4530	853 948	235 344
Sequence 1	NP	NP	NP	NP	4530 5270	948 1323	344 1186

NP: Not penetrated
HF: Hare Fiord

Sequence 2. Late Asselian — late Sakmarian: Canyon Fiord, lower Belcher Channel and Hare Fiord formations.

Sequence 3. Late Sakmarian — late Artinskian: upper Belcher Channel and Hare Fiord formations.

Sequence 4. Late Artinskian — late Roadian: Sabine Bay and van Hauen formations.

Sequence 5. Late Roadian? — late Wordian?: Assistance, Unit A, Trold Fiord, Degerböls and van Hauen formations.

Each of these major sequences, particularly the last two, can be subdivided into several third-order (1-10 Ma) parasequences. In the late Sakmarian-late Artinskian sequence, two third-order parasequences are recognized, whereas in the late Roadian — late Wordian sequence, three second-order parasequences are recognized. It is believed that a formation should not extend the depositional sequence boundary (Vail et al., 1977). Since such a task would require major changes in stratigraphic and biostratigraphic nomenclature, no major attempts were carried out to change, modify, or propose new nomenclature. Hence, the major contribution of this paper in terms of sequence stratigraphy will be the recognition of sequences and parasequences. However, it is recommended that future stratigraphic nomenclature in the Permian and Carboniferous section of the Sverdrup Basin should take into consideration these subsurface data.

LITHOSTRATIGRAPHIC REFINEMENTS

Trold Fiord and Degerböls formations

The Trold Fiord Formation is recognized in Eldridge Bay E-79, Weatherall O-10 and Hecla J-60 wells. Regionally, in subsurface sequences, the Trold Fiord Formation correlates laterally with the Degerböls Formation in the shelf area (Thorsteinsson, 1974). In the Sabine Peninsula wells, the Trold Fiord Formation can be correlated with the Degerböls Formation in the Sherard Bay F-34 well. According to Beauchamp (pers. comm., 1988), the basinal equivalent of the Trold Fiord Formation in outcrop on southwestern Ellesmere Island is the van Hauen Formation. However,

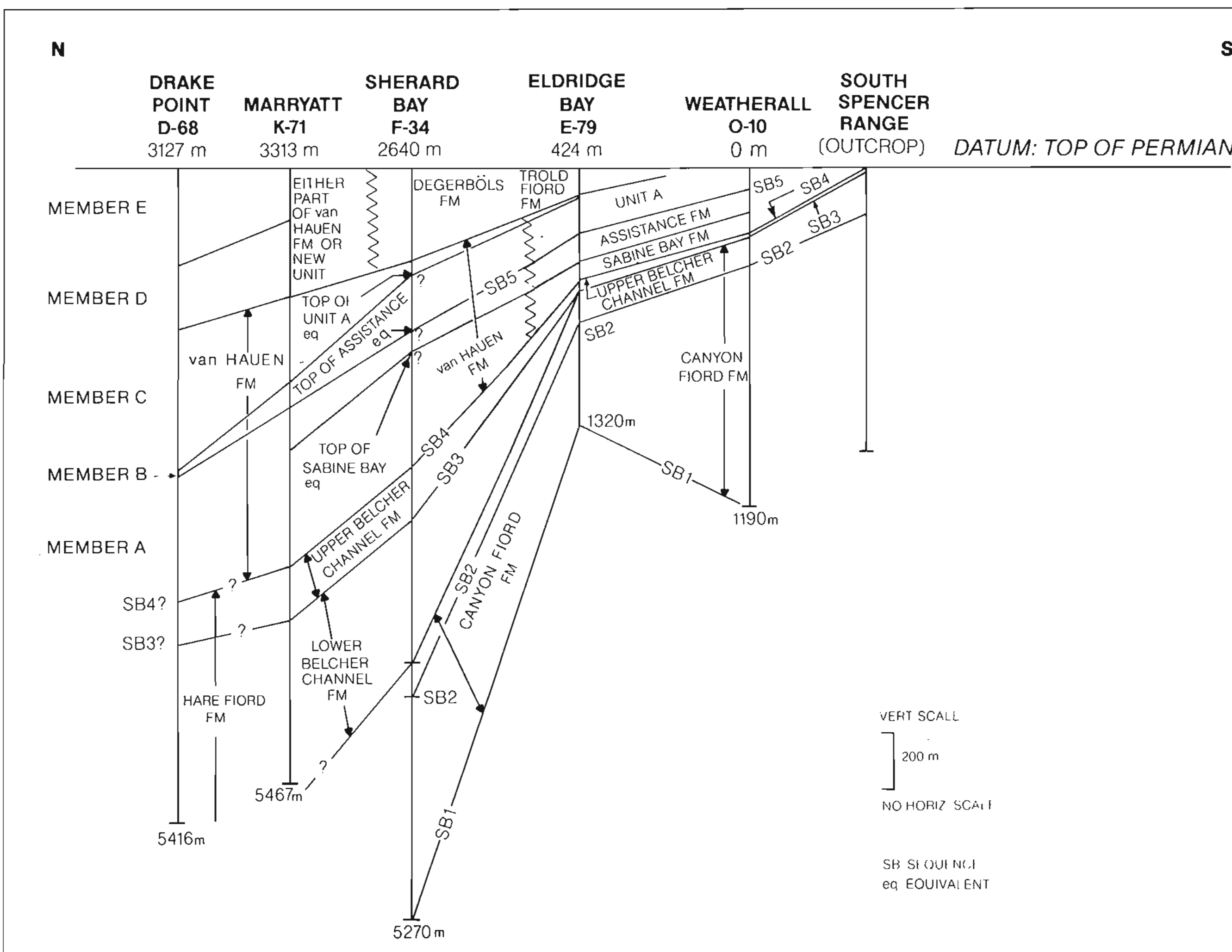

Figure 3. North-south cross-section showing formation and depositional sequence boundaries.

Thorsteinsson (1974) gave the name van Hauen to a sequence of dark coloured shale, siltstone and chert that is bounded below and above by regional disconformities. He stated that the "van Hauen Formation is generally overlain by either the Degerböls Formation or the assumed facies equivalent of the Degerböls, the Trold Fiord Formation". The Trold Fiord Formation in the outcrop sections and in some of the subsurface sections is of a nearshore depositional environment. However, Trold Fiord-equivalent rocks consisting mainly of glauconitic shale, siltstone and chert occur in the Marryatt K-71 (440 m thick), Chads Creek B-64 (614 m thick) and Drake Point D-68 (620 m thick) wells (Figs. 3 and 4; Table 2). Lithological and palynological characteristics indicate that these rocks were deposited in a basinal environment and are of Wordian (Kazanian) age. Thus, it is believed that this sequence is the basinal equivalent of the Trold Fiord and the Degerböls formations. Since the lithological separation of this sequence from the underlying van Hauen Formation is difficult, this succession is not considered to be a separate (new) stratigraphic entity, but is treated as part of the van Hauen Formation. This basinal unit may require renaming in the future, should more information be acquired on its distribution, lithofacies and thickness.

Van Hauen Formation

The van Hauen Formation is recognized in the following Sabine Peninsula wells: Drake Point D-68, Marryatt K-71, Sherard Bay F-34, Hecla J-60, and Chads Creek B-64. The van Hauen Formation includes basinal equivalents of the Trold Fiord and Degerböls formations. Log signatures show that the van Hauen consists of five units (Fig. 3). In this report, they are designated as members A (base) to E (top); these members are recognizable in seismic sections.

Unnamed formation

The "unnamed" formation in the Lower Permian succession of the Sverdrup Basin was introduced by Nassichuk and Wilde (1977) who suggested that "rocks mapped as 'Assistance Formation' on Bjorne Peninsula and adjacent areas are sufficiently distinct from typical Assistance rocks both lithologically and faunally that they might be accommodated in a new, as yet unnamed formation" (Nassichuk and Wilde, 1977, p. 3). Since the so-called "unnamed" formation is not formally defined, it is not used in this paper.

Belcher Channel Formation

The Belcher Channel Formation is recognized in several wells and, in this report, is subdivided into two members, the upper Belcher Channel and the lower Belcher Channel. The subdivision is based on the identification of a specific log signature within the Belcher Channel Formation of Artinskian age in each of the project wells. The upper Belcher Channel Formation may be the equivalent of the "unnamed" formation, but this is not confirmed.

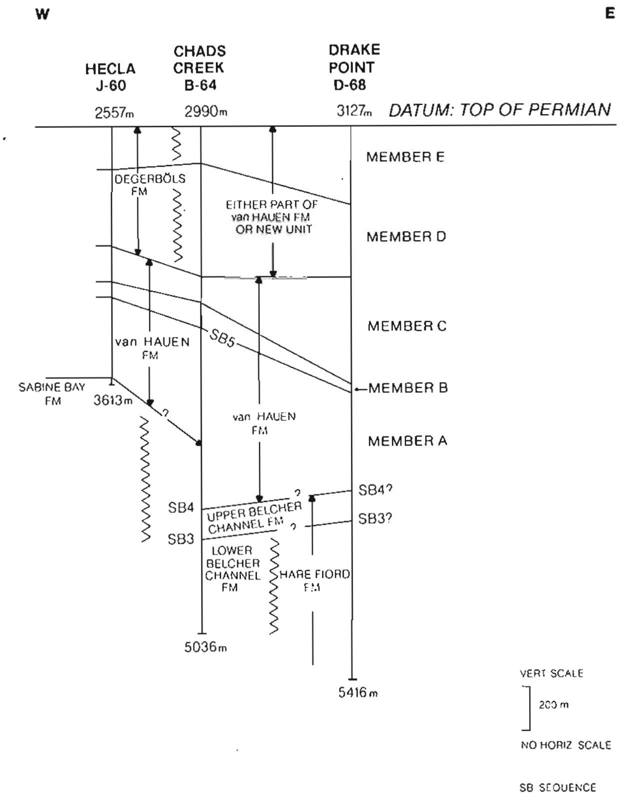

Figure 4. East-west cross-section showing formation and depositional sequence boundaries.

HYDROCARBON ASSESSMENT

Introduction

One of the objectives of this report is to assess the hydrocarbon potential of the Carboniferous and Permian section on Sabine Peninsula through published and newly acquired geochemical data (Gentzis, pers. comm., 1988) and Stewart (1988), and to identify areas that have the stratigraphic and/or structural elements necessary to trap hydrocarbons. To achieve this, it is necessary to discuss the occurrences of the following:

1. Potential upper Paleozoic source rocks of units in terms of Rock-Eval/TOC data.
2. Levels of maturation in terms of vitrinite reflectance and some paleontological indicators.
3. Porous and permeable carrier beds.
4. Traps.

Previously published hydrocarbon assessments of the Permian and Carboniferous section of Sabine Peninsula concentrated on source rocks and levels of maturation (Henao-Londoño, 1977; Powell, 1978). None of the published reports has so far addressed the question of reservoir and trap assessment. In this report, the assessment of upper

Paleozoic reservoirs is based on petrographic examination of clastic and carbonate rocks from Sabine Peninsula wells and the results of sedimentological studies conducted by Beauchamp (1987) of outcrop sections on Ellesmere Island. The existence of traps is based on interpretation of well and seismic data from Sabine Peninsula.

Hydrocarbons have not yet been discovered in the upper Paleozoic section of the Canadian Arctic Archipelago; however, oil seepages are known to occur about 25 km southeast of the Weatherall O-10 well (Fig. 2). Almost all of the hydrocarbon fields in this area are in the Mesozoic section, where most of the previous geochemical studies were concentrated; little attention has been paid to the Permian-Carboniferous sequence (Snowdon and Roy, 1975; Henao-Londoño, 1977; Stuart-Smith and Wennekers, 1977; Powell, 1978). Henao-Londoño (1977) concluded that the Sverdrup Basin is essentially a gas-prone area and that oil could be expected only as a subsidiary product in the Permian-Carboniferous section. Powell (1978) concluded that the upper Paleozoic rocks in the Sverdrup Basin would likely yield only gas.

In this report, the potential source rocks and maturation levels of the Permian-Carboniferous will be discussed using geochemical data newly acquired by Gentzis (pers. comm., 1988) and Stewart (1988). Maturation levels of a few samples from Sabine Peninsula are also assessed using paleontological indicators (spores and conodonts) (Utting et al., in press).

Maturation levels and source rocks

Chads Creek B-64

Vitrinite reflectance (% Ro) at 3600 m depth is 0.90 % indicating conditions within the "oil window". This value increases to 1.10 % at 4200 m and to 1.30 % at 4800 m depth (Stewart, 1988 and Gentzis, pers. comm., 1988). In addition to vitrinite, two distinct types of pyrobitumen present at depths of 4600 and 4800 m have reflectance values of 1.50 and 2.50 %, respectively (*ibid.*). All these values indicate overmaturity. Gentzis (pers. comm., 1988) does not anticipate that heat generated by intrusive igneous rock at depths of about 3400 m and 4400 m has significantly influenced the reflectance values. Thermal alteration indices of the van Hauen Formation vary from 3- to 4, indicating an oil to gas window (Utting et al., in press). The average HI (hydrogen index) value in the van Hauen Formation (9 shale samples) is less than 20 mg HC/g TOC (Stewart, 1988).

Drake Point D-68

Vitrinite reflectance values of the van Hauen Formation range from 0.80 at a depth of 3200 m to 0.95 % at a depth of 4100 m (Gentzis, pers. comm., 1988). Pyrobitumens have reflectance values of 1.0 and 1.12 at a depth of 4200 and 4350 m, respectively. The overall high reflectance values indicate that the van Hauen and Hare Fiord formations are in the mature to overmature stage of hydrocarbon generation and that only gas (possibly dry gas) could be generated (Gentzis, pers. comm., 1988). In contrast, the thermal alteration index in the Hare Fiord Formation is 3+, indicating that the strata are oil prone, and the average HI values in the van Hauen Formation (8 samples) is less than 38 mg HC/g TOC (Stewart, 1988).

Hecla J-60

Vitrinite reflectance analysis of one Degerböls Formation sample gives a value of 0.90 %. The average reflectance value of pyrobitumen of the Degerböls Formation is 1.45 %. The vitrinite equivalent of this value would be 1.29 %, indicating overmaturity. The average HI value in the Degerböls and van Hauen formations (5 samples) is less than 39 mg HC/g TOC.

Discussion

Vitrinite reflectance values from the Permian-Carboniferous section of the Sabine Peninsula area indicate that these rocks range from mature to overmature. In addition, conodont and palynomorph colour alteration indices show that almost all the upper Paleozoic section of the Sverdrup Basin lies within the "oil window" (Utting et al., in press). On the other hand, Rock-eval/TOC data of these samples show none of them to be oil-prone source rocks. It has been demonstrated that oil has been discovered in sedimentary basins at a depth of more than 6 km. Therefore, it appears that source rock type rather than the level of maturation in the upper Paleozoic section is the major factor in generating oil. It should be noted that the van Hauen and Hare Fiord shale samples studied for this report are from wells located at the margin of the Sverdrup Basin. However, it appears that reducing conditions were most likely attained when the van Hauen and Hare Fiord sediments were being deposited in the depocentre of the basin (Axel Heiberg Island) in view of new evidence of the existence of $\delta^{13}C$ enriched water in the Permian-Carboniferous sediments in the Sverdrup Basin (Beauchamp et al., 1987). The $\delta^{13}C$ enrichment in the Sverdrup Basin is explained in terms of stagnation and thermohaline stratification in a partly closed basin, with preferential preservation of organic matter in oxygen-depleted bottom waters(*ibid.*). This means that Type II and perhaps Type I organic matter could have accumulated during the deposition of the van Hauen and Hare Fiord formations.

It has to be pointed out that the widespread presence of hard bitumen in the Permian outcrops near the South Spencer Range section shows that source rocks are present and that some maturation has taken place. In conclusion, the author believes that the possibility of finding oil-prone source rocks is not remote, despite the high vitrinite reflectance values and the depth of burial. In addition, it should be noted that the Hare Fiord Formation at a depth of 2438 to 2453 m in the Robert Harbour K-07 well on Cameron Island (Fig. 1) has an atomic H/C (Hydrogen and Carbon ratio) value of 1.11 (Powell, 1978). The Hare Fiord is the basinal facies of the upper Paleozoic in the Sverdrup Basin; and the data mean that Type II, oil-prone, organic matter may be present in the upper Paleozoic section of the Sverdrup Basin. This result should be confirmed by conducting Rock-Eval/TOC procedures to measure the HI values.

Reservoir assessment

It has been demonstrated by geochemical analyses that the Permian-Carboniferous section of Sabine Peninsula is gas-prone and perhaps oil-prone. It is conceivable that these hydrocarbons may have migrated into nearby reservoirs. Lithological examination of potential reservoir rocks of the shelf carbonate regime of the Belcher Channel Formation and of some units within the Canyon Fiord Formation have shown them to lack porosity because of early and late diagenetic cementation. Furthermore, crinoids are the major constituents of the Permian-Carboniferous carbonates and syntaxial overgrowths on crinoids are a major force in destroying porosity within these units. Where carbonate sediments have been affected by meteoric leaching, the secondary porosity is infilled by late-diagenetic cements. Furthermore, dolomite is very rare (except one unit, 15 m thick in Marryatt K-71) in the upper Paleozoic subsurface section, although dolomitization may have played a role in porosity development elsewhere in the Peninsula. Porous dolomite units from the Nansen Formation have been observed in outcrop at Blind Fiord, but dolomitization appears to be restricted to that area, probably related to the proximity of synsedimentary growth faults (Beauchamp, 1987).

Primary intergranular porosity may have been preserved within carbonates of the middle units of the Canyon Fiord and Belcher Channel formations, particularly in areas farther shoreward. Such porosity occurs in areas where skeletal sand shoal carbonates are developed. These shoals may be controlled by previously high structures (for example, horsts) where high energy wave action removed the lime mud constituent during deposition, and porous grainstone units were thus developed.

In contrast, clastic marine sequences show a high percentage of intergranular porosity (more than 15 %), particularly in the Assistance and Sabine Bay formations, and to some extent in the Canyon Fiord and Trold Fiord formations (Weatherall O-10 and Eldridge Bay E-79). This is because of incomplete cementation processes in which the sandstones are poorly indurated, as well as possible secondary leaching processes. These marine and nonmarine clastic sequences may have played a role as carrier beds. In addition, dead and live oil residues have been detected within these clastic units (Sabine Bay and Canyon Fiord formations), particularly in the Eldridge Bay and Weatherall wells. Hydrocarbon seepages from the Permian rocks are present in areas ~25 km southeast of the Weatherall O-10 well. It therefore appears that the clastic units of the Canyon Fiord, Assistance and Sabine Bay formations are the most promising potential reservoirs within the upper Paleozoic section, particularly in the southern part of the peninsula.

Trap and seal assessment

Another factor controlling the occurrence of hydrocarbons is trap formation and its timing relative to migration. During Carboniferous and Permian time, several pulses of rifting affected the Sverdrup Basin. Surface and subsurface (well data and seismic) studies of Melville Island show that a series of half-graben structures formed in response to this rifting (Harrison, pers. comm., 1988). It is possible that graben-associated faulting and/or associated unconformities provided a trapping mechanism and assisted in dolomite development, particularly in the area between the Weatherall and Sherard Bay wells, or more precisely between the Weatherall and Eldridge Bay wells, in the upper and lower parts of the graben structures. A trap must have a seal, normally a shale or other impervious rocks. As the Sabine Peninsula area was subjected to various late Paleozoic changes in sea level, and a corresponding marine transgression, seals may have been formed by shale and impermeable argillaceous mudstone deposited at that time. The best seal probably is the black shale of the Blind Fiord Formation of Triassic age and/or tight (well cemented) sandstone of the Assistance and/or Trold Fiord formations.

Stratigraphic traps may occur in the southern part of the Hecla and Griper Bay area where the Sabine Bay Formation acts as an effective reservoir and the van Hauen Formation as a seal. This trap is suggested by the Assistance Formation resting on Devonian rocks at the Green Creek outcrop section (Tozer and Thorsteinsson, 1964) and the Sabine Bay Formation resting under the van Hauen Formation at the Hecla J-60 well.

Exploration implications and recommendations

1. Geochemical data indicate that the van Hauen and Hare Fiord formations are overmature, gas-prone source rocks. However, the possibilities of finding oil-prone source rocks are not totally remote, as is shown by the geochemical data from the shale of the Permian-Carboniferous section in the Robert Harbour K-07 well, by oil stains in subsurface samples of the Canyon Fiord, Assistance and Sabine Bay formations, by oil seepages from the Canyon Fiord Formation in the vicinity of the Weatherall O-10 well and by palynomorph colour alteration indices, indicating that the samples lie within the oil window.
2. It is unlikely that a porous limestone reservoir will be found in the study area.
3. The Canyon Fiord, Assistance and Sabine Bay formations are the most promising reservoir units within the Permian-Carboniferous section of Sabine Peninsula. It is possible that very porous zones within the Canyon Fiord Formation can be located in areas close to former hingelines, where hydrocarbon-filled sandstones might be expected.
4. The area between the Eldridge Bay and Weatherall wells, particularly near the half-graben structures, should be investigated for reservoirs and traps. More seismic surveys are required in this area. It must be noted that the quality of seismic data in the vicinity of the half-graben structures on Sabine Peninsula is very poor, and that these were shot in the early 1970's. Most of the high-quality seismic reflection data is presently concentrated in the vicinity of the Sherard Bay F-34 well and farther north.
5. Subtle stratigraphic traps could be present in the southern part of the Hecla and Griper Bay area, where the Sabine Bay and van Hauen formations may contain reservoir and seal rocks, respectively. The hydrocarbon source in this area may be Devonian.

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Geometry and evolution of the Tanquary Structural High and its effects on the paleogeography of the Sverdrup Basin, northern Ellesmere Island, Canadian Arctic[†]

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Abstract

The Tanquary High is a fault controlled, structural-topographic feature about 50 km wide, which trends for about 100 km east-west across the southern margin of the Sverdrup Basin. Thinning or erosion of Carboniferous to Triassic formations, as well as lateral variations of facies from marine to fluviatile environments are the main stratigraphic and sedimentological manifestations of the high in the Tanquary Fiord map area. Regionally, the development of the Tanquary High is coeval with the initial formation and subsequent subsidence of the Sverdrup Basin. Reconstruction of the pre- Carboniferous and pre-Triassic subcrop map of the Tanquary Fiord map area in northern Ellesmere Island led to the structural definition of the Tanquary High.

Tertiary faulting as a result of the Eurekan Orogeny has not significantly modified the shape of the Tanquary High, thus supporting the thesis for little shortening and even less transpression in Tertiary time in the region.

Résumé

La hauteur de Tanquary est un élément structural et topographique contrôlé par des failles; large de 50 km environ, la hauteur poursuit une direction est-ouest sur 100 km environ et traverse la marge sud du bassin de Sverdrup. L'amenuisement ou l'érosion des formations du Carbonifère au Trias, ainsi que les variations du faciès d'un milieu marin à un milieu fluviatile, sont les principales manifestations stratigraphiques et sédimentologiques de la hauteur dans la région cartographique du fjord Tanquary. Régionalement, la formation de la hauteur de Tanquary est contemporaine à l'ouverture et à la subsidence ultérieure du bassin de Sverdrup. La reconstruction de la carte des sous-affleurements du pré-Carbonifère et du pré-Trias de la région cartographique du fjord de Tanquary dans le nord de l'île d'Ellesmere a mené à la définition de la hauteur de Tanquary.

Le fait que la formation de failles au Tertiaire à la suite de l'orogenèse de l'Eureka n'a pas beaucoup modifié la géométrie de la hauteur de Tanquary, vient appuyer la thèse selon laquelle un rétrécissement faible et une transpression encore moindre ont eu lieu pendant le Tertiaire dans la région.

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INTRODUCTION

Regional setting

The study area is located in northwestern Ellesmere Island, north and northeast of the head of Tanquary Fiord (Fig. 1). The rocks underlying this area are divisible into two, first-order stratigraphic successions that are separated by an angular unconformity of regional extent:

1. Hadrynian or Lower Cambrian to Upper Devonian strata of the Franklinian mobile belt, deformed by the Paleogene Eurekan Orogeny, the Late Devonian — Early Carboniferous Ellesmerian Orogeny and earlier events.
2. Carboniferous to Cretaceous strata, deformed only during the Paleogene Eurekan Orogeny.

The Precambrian to Silurian rocks of northern Ellesmere Island were deposited in a shelf province and in an adjacent deep water basin, the latter divisible into a southeasterly sedimentary subprovince and a northwesterly sedimentary-volcanic subprovince (Fig. 1; Trettin, 1987a, b). The sedimentary subprovince is linked with the shelf (coincident to some extent with the Central Ellesmere Fold Belt) by interlocking facies changes, but the sedimentary-volcanic subprovince may include allochthonous elements (Trettin, 1987c). The latter is bordered on the north by Pearya, a composite terrane that is Middle Proterozoic or older to Late Silurian in age. Available limited evidence suggests that Pearya was accreted to the Franklinian mobile belt by sinistral transpression in latest Silurian time, a process that presumably resulted in the deformation of the sedimentary-volcanic subprovince (termed Clements Markham Fold Belt in northern Ellesmere Island) and adjacent parts of the sedimentary subprovince (northern Hazen Fold Belt) (Trettin, 1987a). The study area straddles parts of these two belts. Middle-Upper Devonian (?Givetian and Frasnian) clastic sediments, preserved locally in the Clements Markham Fold Belt, probably postdate the Late Silurian event because they appear markedly less deformed than the Silurian and older rocks, but the base of the Devonian strata is concealed. In northeastern Ellesmere Island, the Clements Markham and Hazen fold belts are separated by a transcurrent fault (Porter Bay Fault of the Feilden Fault Zone) that had dextral motion in post-Early Permian, probably Paleogene time, but may have originated in Late Silurian time as a sinistral fault (Fig. 1; Okulitch and Trettin, in press). In the present area (Fig. 2), the contact between the two belts is covered by younger strata of the Sverdrup Basin.

The Sverdrup Basin strata are well exposed along a corridor of cliffs between the head of Tanquary Fiord and Yelverton Inlet (Fig. 2). Topographic relief and continuous exposures of upper Paleozoic formations along the corridor permitted the reconstruction of structural, paleostructural and stratigraphic profiles of the basin.

The Sverdrup Basin was initiated by the development of an intracontinental rift system in Early Carboniferous time (Thorsteinsson, 1974; Beauchamp et al., 1989). Continuing subsidence, intermittently interrupted by minor tectonic uplift and eustatic regressions, led to the deposition of up to 12 km of clastics, carbonates and evaporites, Carboniferous to Cretaceous in age, in the basin (Trettin and Balkwill, 1971; Plauchut, 1971; Davies and Nassichuk, in press; Embry, in press). Only the southern margin of the basin is

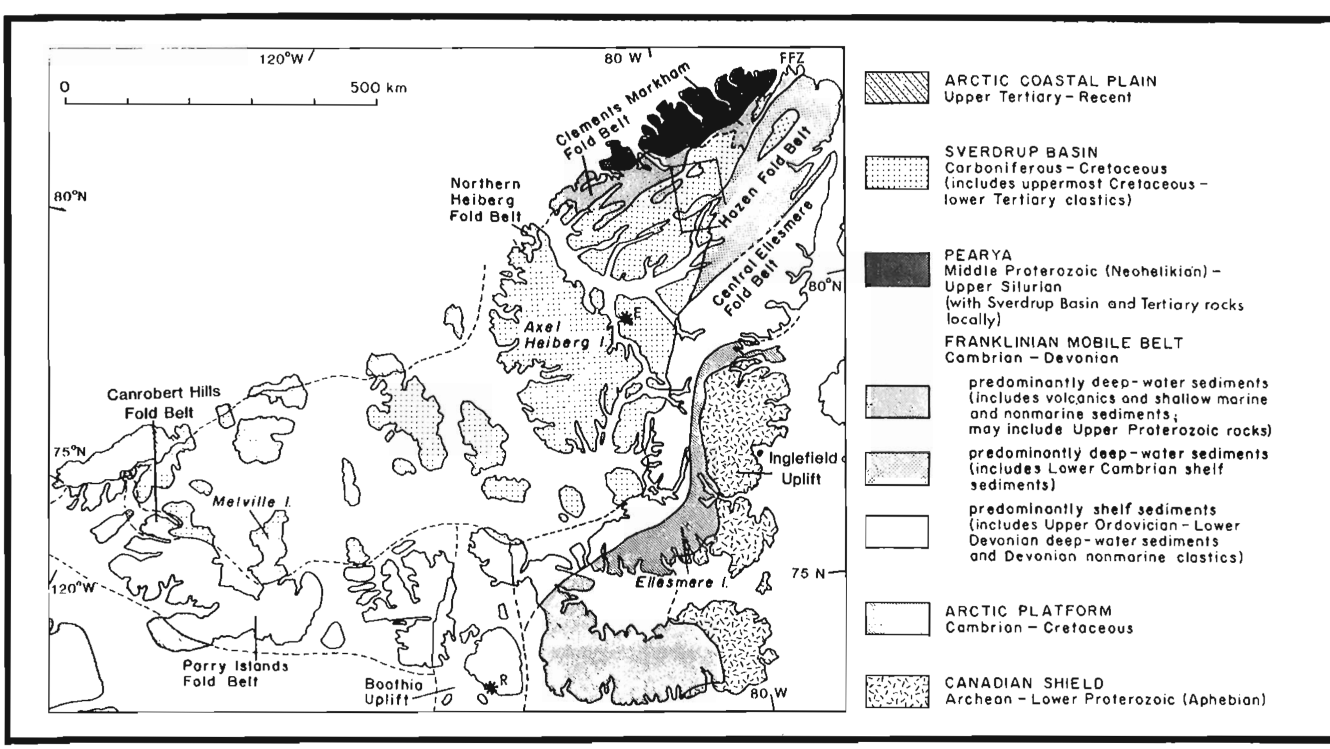
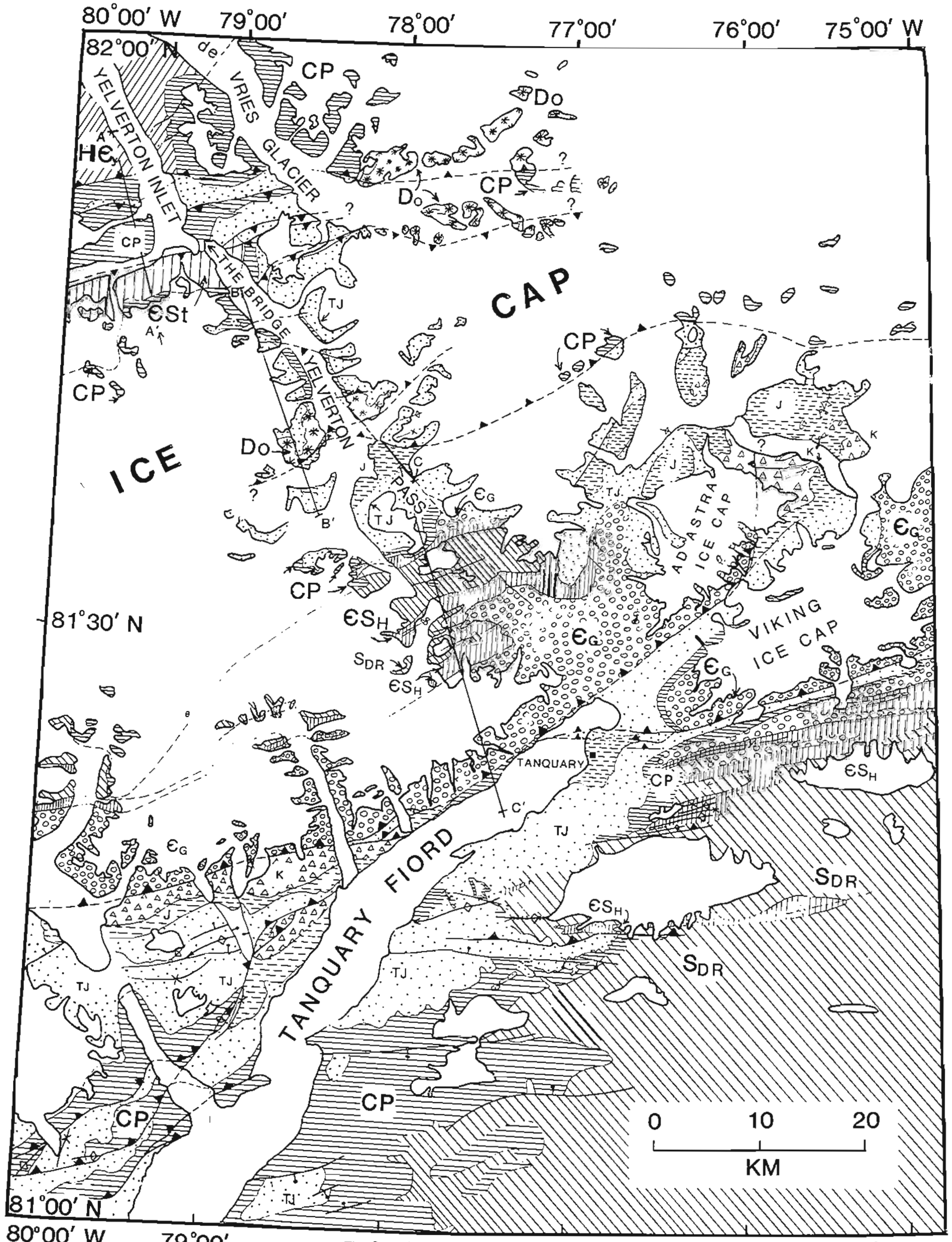


Figure 1. Index map of the structural framework of the Canadian Arctic Islands and legend (from Trettin, 1987a); Map area outlined in box; FFZ, Feilden Fault Zone; E, Eureka; R, Resolute.



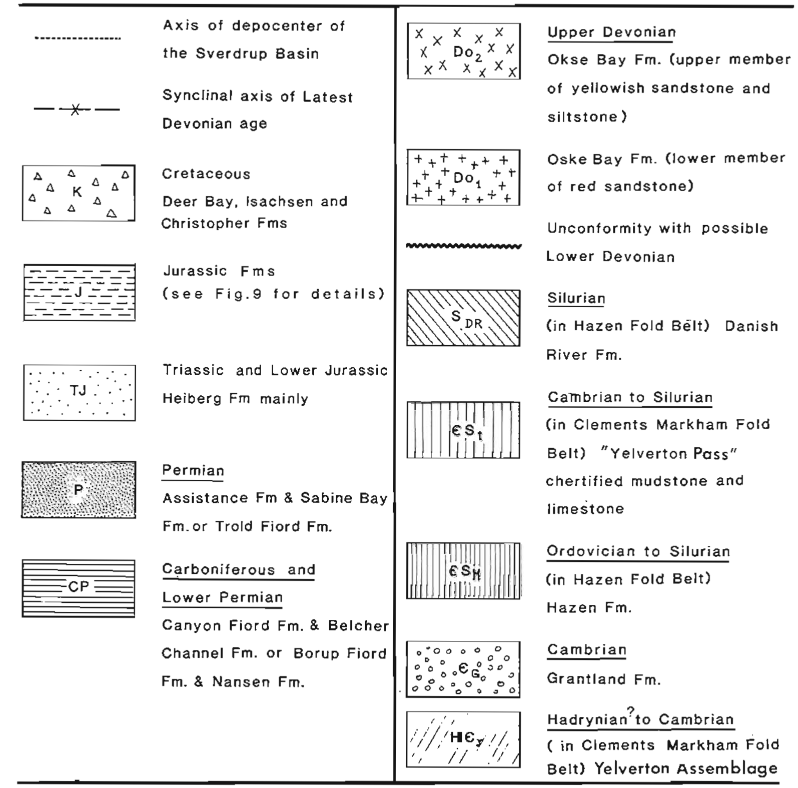


Figure 2. Simplified geological map of the Tanquary Fiord-Yelverton Inlet area, showing lines of cross-section AA', BB', and CC'.

completely exposed across the Arctic Islands. The northern margin is reconstructed from sporadic outcrops in northern Axel Heiberg Island and northern Ellesmere Island (Fig. 3). Each margin can be subdivided into a marginal facies belt, proximal to shoreline, and a more distal, carbonate belt (Mayr, work in progress). However, both belts were mobile through time, producing interlocking facies changes (Beauchamp et al., 1989).

For broader inter-regional comparisons, it is important to note that the Svalbard archipelago (including Spitsbergen) lay adjacent to northern Ellesmere Island prior to Eocene sea floor spreading (Le Pichon et al., 1979).

THE TANQUARY HIGH: PREVIOUS WORK

The Tanquary High is a structural-topographic feature of the Sverdrup Basin that was exhumed during the late Paleozoic and remained active, to a lesser extent, during the Triassic (Nassichuk and Christie, 1969) and the Early Jurassic (Embry, in press). It is located in northern Ellesmere Island at the head of Tanquary Fiord. Thinning or erosion of Carboniferous to Triassic formations, as well as lateral variations of facies from marine to fluviatile environments, are the main stratigraphic and sedimentological manifestations of the high in the Tanquary Fiord-Yelverton Pass region (Nassichuk and Christie, 1969; Wilson, 1976; Mayr, work in progress). The high affected the southern margin of the basin and deflects the isopachs of the Carboniferous to Triassic formations toward the depocenter axis of the basin (Embry, in press). The high also isolated two sub-parallel basins on its flanks from early Bashkirian to early Artinskian time (Figure 3; and Thorsteinsson, 1974; Balkwill, 1978; Christie, 1964).

Whereas the sedimentological effects of the Tanquary High are well established, its geometry and origin have been a matter of contention. Nassichuk and Christie (1969) and Mayr (work in progress) postulated and described pre-Triassic normal faults bounding the Tanquary High. Nassichuk and Christie (op. cit.) discussed the possible origins of the high — whether it might have begun as a tectonically active arch, a passive, non- subsiding high, or a fault-bounded basement horst. However, the superposition of the reverse faults of the Tertiary Eurekan Orogeny led to confusion between pre-Triassic structures and Tertiary ones. As pointed out by Osadetz (1982), the Grant Land High, located in the same area, is a Tertiary compressional structure that was partially overprinted onto the Tanquary High.

In northern Ellesmere Island, a major uplift occurred in Late Cretaceous time (Embry and Osadetz, in press). It was followed by faulting and folding in Paleocene to Late Eocene or earliest Oligocene time (Eurekan Orogeny). Localized intramontane basins formed in the forelands of thrust highlands (Miall, 1979). These deposits were subsequently remobilized by folding and faulting in the latest stages of Eurekan tectonism. In Pearya Terrane, where the northern marginal facies of the basin were deposited (Mayr, work in progress), only small remnants of the Sverdrup Basin deposits are preserved because of pronounced uplift during and after the Eurekan Orogeny. However, in the Tanquary Fiord map area, Eurekan uplift and reverse faulting did not counterbalance the preceding subsidence of the Sverdrup Basin. Thus, a large part of the basin is preserved there. The erosional surface of the Late Cretaceous unconformity limits the extent of Tertiary structures above the present-day erosional surface, and is projected in the structure section of Figure 4a.

The Eurekan compressional structures have altered the original morphology of the high, but not significantly. Shortening through folding is limited to the vicinity of fault zones, which are rare across the 100 km section through Yelverton Pass (Fig. 2 and 5; and Maurel, work in progress). Less than 10 per cent shortening is inferred for the area of the Tanquary High (Fig. 4a). Strike-slip is also minimal in the region (Higgins and Soper, 1983) and seems to be related to localized oblique and transverse hanging wall ramps of steep reverse faults (Maurel, work in progress). Consequently, no palinspastic restoration of the high was attempted, since its modification is insignificant and since it is partly concealed along strike under the surrounding ice caps.

CURRENT INVESTIGATIONS

The purpose of this paper is to clarify the geometry and mode of origin of the Tanquary High, primarily by means of subcrop maps that reveal its configuration during different time intervals (Fig. 6, 7). The effects of Paleogene Eurekan deformation are shown in a preliminary structural cross-section (Fig. 4a). The basic information was obtained by precise remapping, during the summer of 1988, of areas mapped in reconnaissance fashion by Nassichuk and Christie (1969), Thorsteinsson and Trettin (1972), and Mayr et al. (1982).

STRATIGRAPHY, STRUCTURAL CHARACTERISTICS AND PALEOGEOGRAPHIC EVOLUTION

Lower Paleozoic and (?)older units

Hazen Fold Belt

The Hazen Fold Belt, comprising four units of formational rank (Trettin, 1987b), is characterized by tight, upright to moderately inclined complex folds of chevron type (Fig. 8; and Ramsay, 1967; Richard, 1971), the axes of which are subhorizontal or plunge at moderate angles northeast or southwest.

The oldest exposed unit, informally termed the Nesmith beds, consists of impure, calcareous turbidites with a slaty to phyllitic fabric. The base of the unit is concealed and the thickness of the exposed formation is uncertain because of excessive deformation. The unit is not fossiliferous, but regional lithological correlations with northeastern Greenland and the shelf province indicate an Early Cambrian (early Atdabanian) age.

The Grant Land Formation, which overlies the Nesmith beds with gradational contact, consists of perhaps 2 km of quartzite and multicolored slate with minor pebble conglomerate. This unfossiliferous unit also is assigned to the Lower Cambrian on the basis of regional relationships.

The Hazen Formation overlies the Grant Land Formation with a sharp contact. A lower member, consisting of

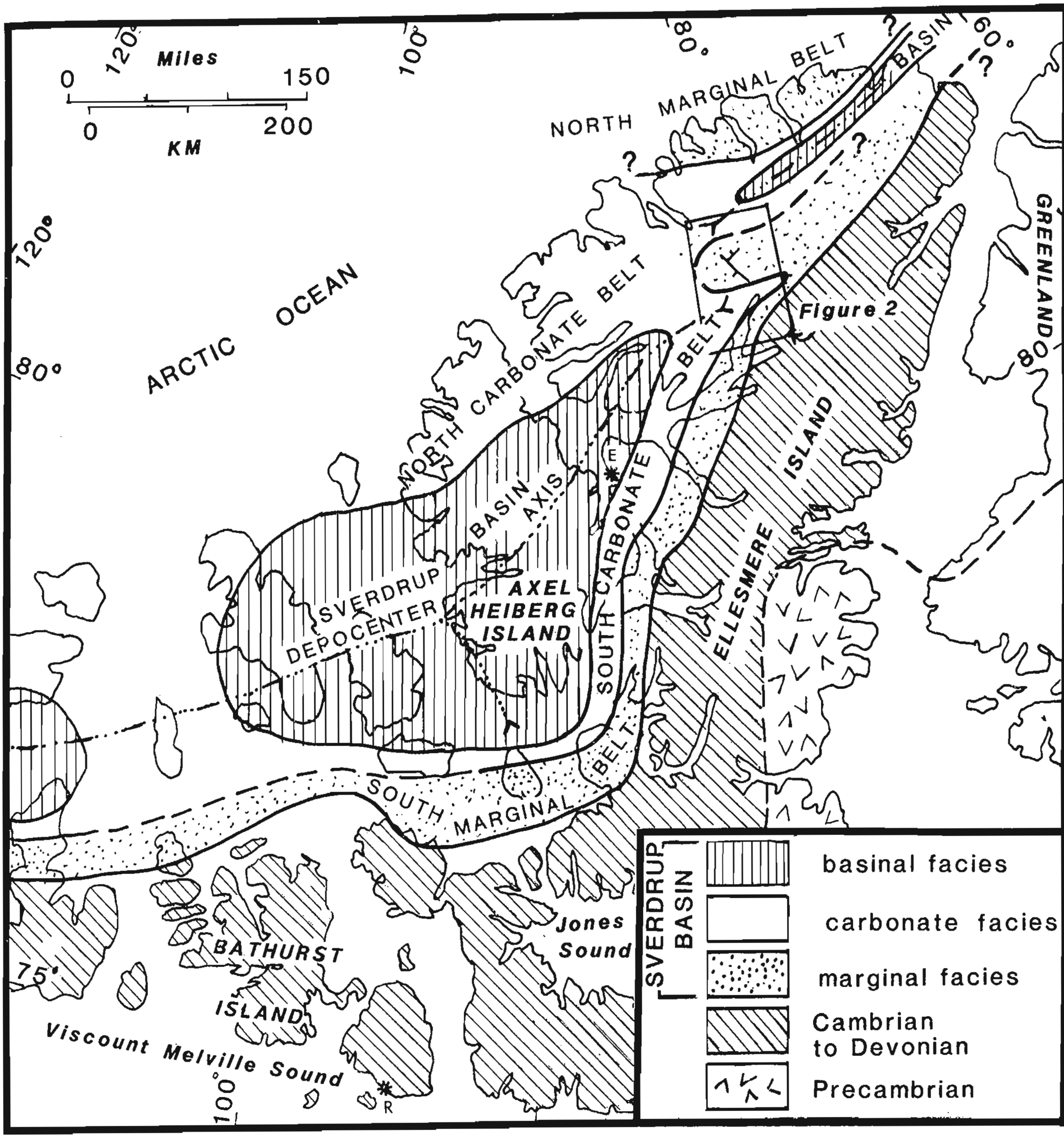


Figure 3. Major paleogeographic provinces of the Sverdrup Basin in the Canadian Arctic Archipelago in Artinskian time and location of the map area (from Mayr, work in progress).

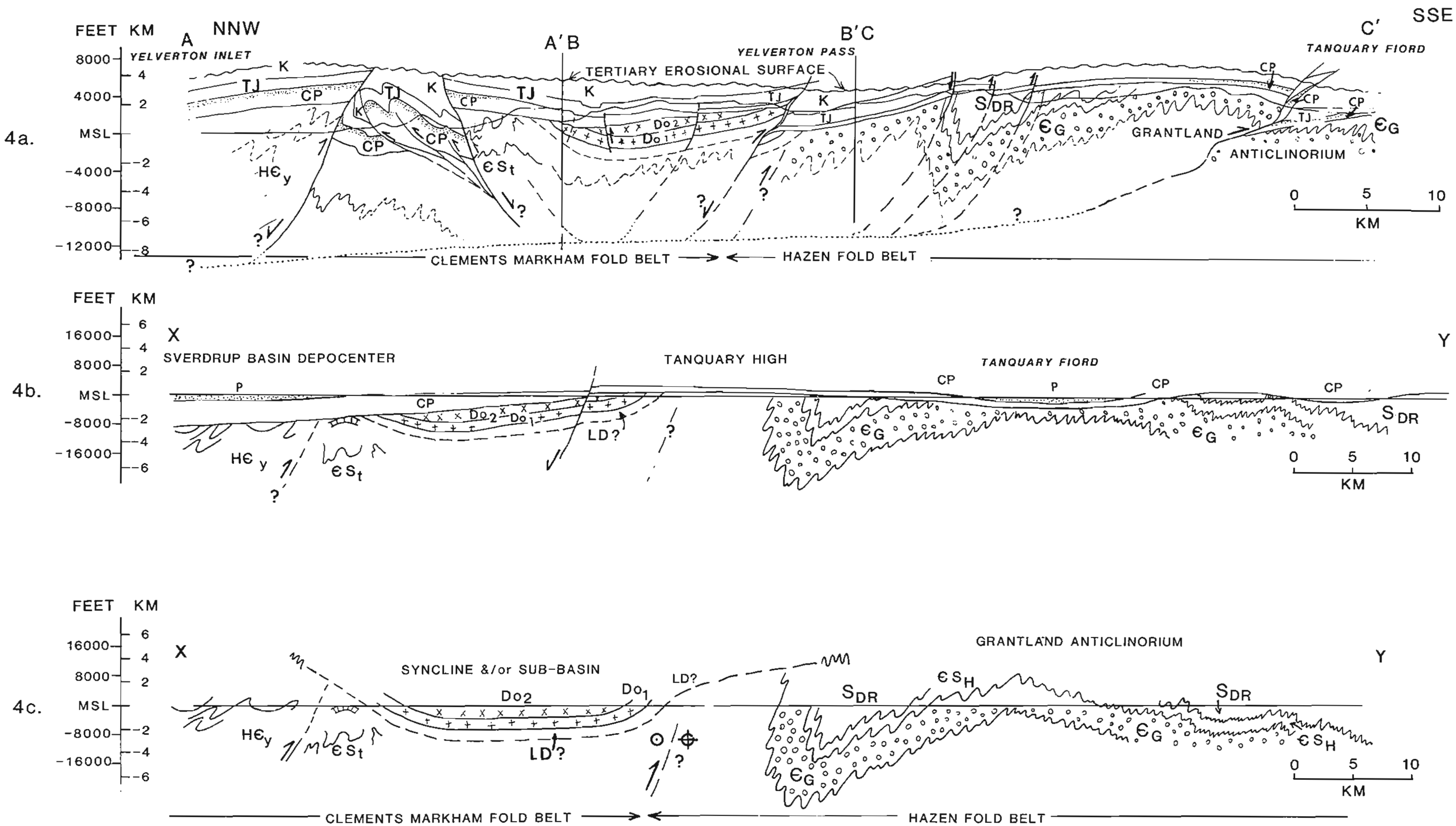


Figure 4. Three cross-sections illustrating the structural evolution of the map area. a: Present-day composite cross-section; b: Pre-Triassic cross-section; c: Pre-Carboniferous cross-section. Legend same as for Figure 2.

resedimented carbonate with mudstone, sandstone and minor chert can be distinguished from an upper, black, radiolarian chert member containing minor mudstone and carbonate (Trettin, 1987b). Average total thickness in the Tanquary Fiord area is about 250 m. The formation is Early Cambrian to Early Silurian in age (Trettin, 1987b).

The conformably overlying Danish River Formation comprises up to 3 km of calcareous sandstone and mudrock with turbiditic characteristics. At the southern margin of the basin, the unit is about 3 km thick and ranges in age from Early Silurian to earliest Devonian, but in the Tanquary Fiord area probably only strata of Early Silurian (Llandovery and Wenlock) ages are preserved.

Clements Markham Fold Belt

The Clements Markham Fold Belt is characterized by severe faulting and intense folding, with chevron-type folds common in Silurian turbidites and recumbent folds common in older units. The belt includes Lower Cambrian to Upper Silurian sedimentary units that are comparable, to varying extents, to formations of the Hazen Fold Belt, as well as sedimentary and volcanic assemblages that are absent from the latter (Trettin, 1987a).

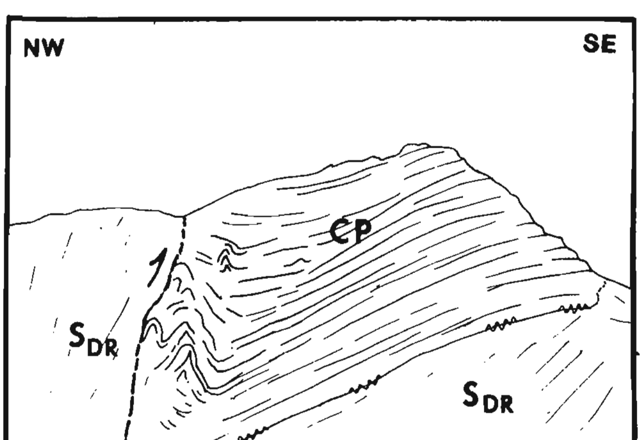


Figure 5. View of Tertiary, steep reverse fault, folding Permo-Carboniferous carbonates in its footwall. Note the unconformity between the dark-coloured slates of the Danish River Formation and the light-coloured planar beds of the Belcher Channel Formation (zig-zag pattern). Field of view is about 500 m to the south of Yelverton Pass. CP, Permo-Carboniferous; SDR, Silurian.

Only a few units of the two categories of this belt are exposed in the study area and they are discussed here only briefly because they are marginal to the topic of this paper. The first category is represented by quartzite and multicoloured phyllite northeast of the head of the Yelverton Inlet that are comparable to the Grant Land Formation; and by a largely chertified unit farther south, comparable to the carbonate member of the Hazen Formation in general, but containing coarser carbonate material and quartz than are found in the exposures of Tanquary Fiord.

The second category includes two units, informally referred to as the Yelverton Assemblage and the Yelverton Pass limestone (Trettin, 1987b). The Yelverton Assemblage is best known from the adjacent Otto Fiord map area, where it has been thrust over the Grant Land equivalents. It comprises metamorphosed, subalkaline basalt (pyroclastics and flows) with some arc characteristics (inferred from stable trace elements), and interbedded metamorphosed carbonate rocks, mudrock and chert (Trettin et al., in press). This unit is probably Hadrynian or earliest Cambrian in age if it belongs to the same stratigraphic succession as the Grant Land equivalents, but of indefinite Late Proterozoic or early Paleozoic age if exotic (Trettin et al., in press).

The Yelverton Pass limestone, which locally overlies the chertified Hazen equivalents with (?)fault contact, has yielded corals and conodonts of early to middle Llandovery age; that is, slightly younger than conodonts from the volcanic (andesitic) and sedimentary Kulutingwak assemblage exposed northwest of the study area (Trettin et al., in press).

Devonian units

Red sandstone and shale of the Okse Bay Formation, Middle or Lake Devonian in age, are exposed with a slight angular unconformity below the Sverdrup Basin strata in the Yelverton Pass area and along de Vries Glacier (Fig. 2; and Mayr, work in progress). Two members of the Okse Bay Formation can be distinguished: a lower red member of conglomeratic sandstone, siltstone and shale, and an upper, light-coloured member of sandstone and conglomerate. Paleocurrent directions measured by Mayr in the southernmost outcrops of the Okse Bay Formation indicate a southeasterly to easterly source area. Chert pebbles predominate in the composition of the conglomerates, and are likely derived from exposed highlands of the Hazen Formation to the south. Total thickness of exposed strata varies between 650 and 1300 m from south to north in the map area (Mayr, work in progress).

The Devonian sedimentary rocks are less intensely deformed than the Cambrian to Silurian succession of the Clements Markham and Hazen fold belts. Consequently, the main tectonic phase is inferred to be older than late Middle Devonian (pre-late Givetian or pre- Frasnian) (Trettin, 1987b).

The angular unconformity with the overlying Sverdrup Basin strata may have resulted from tilting of the Devonian strata because of extensional block faulting. Intramontane

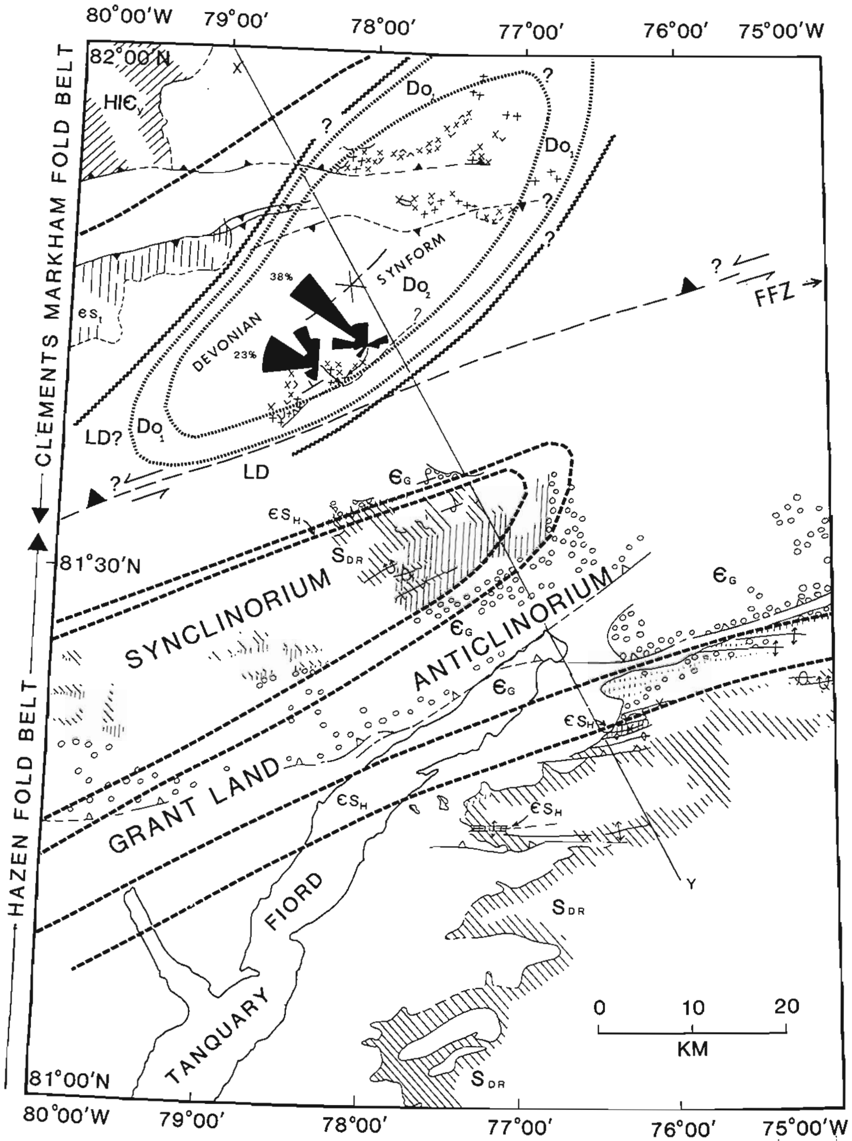


Figure 6. Pre-Carboniferous subcrop map of the Tanquary Fiord-Yelverton Pass area. Legend same as that for Figure 2. Rose diagrams indicate paleocurrent directions for the lower and upper members of the Okse Bay Formation (left and right diagrams, respectively). Dashed and dotted lines mark the subcrop extent of the specific formations. FFZ, Feilden Fault Zone.

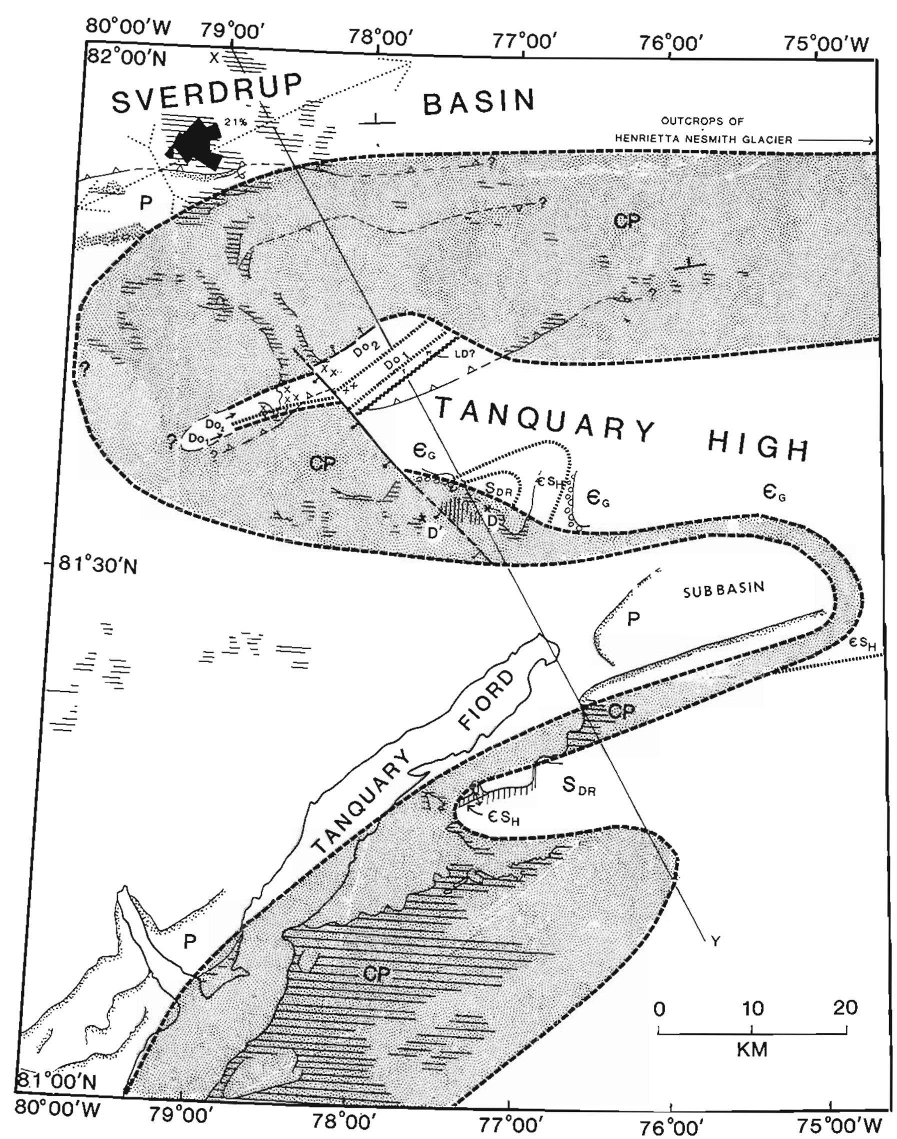


Figure 7. Pre-Triassic subcrop map of the Tanquary Fiord-Yelverton Pass area. Legend same as that for Figure 2. The rose diagram indicates paleocurrent directions for the Borup Fiord Formation with two maxima at 21% to the northeast and southeast. The shaded area marks the subcrop extent of the Carboniferous-Permian Nansen and Belcher Channel formations.

basins such as those observed in Svalbard (Harland, 1969) may have developed in northern Ellesmere Island after the collision of Pearya. Alternatively, the unconformity may have resulted from post-Frasnian, pre-Viséan folding, such as that observed in the Ellesmere-Parry Island Fold Belt to the south and southwest (Fig. 1). Unfortunately, the base of the formation is not exposed, owing to Tertiary faulting.

Sverdrup Basin formations

The folded lower Paleozoic strata of the Hazen and Clements Markham fold belts and the gently dipping Devonian clastic rocks are overlain unconformably by the sediments of the Sverdrup Basin. The oldest strata of the basin, namely the Viséan Emma Fiord Formation, are not exposed in the map area (Mayr, work in progress). However, a nearly continuous paleogeographic profile of the southern margin and axis of the basin is exposed along the Tanquary Fiord-Yelverton Inlet section (Figs. 2, 4b, 7, 9). The red sandstone and conglomerate of the Borup Fiord Formation, Bashkirian to early Moscovian in age, and the shelf dolostone of the Nansen Formation, Moskovian to Sakmarian in age, characterize the southern marginal and carbonate belts of the Sverdrup Basin (Mayr, work in progress) along the XY Section of Figures 4b and 7).

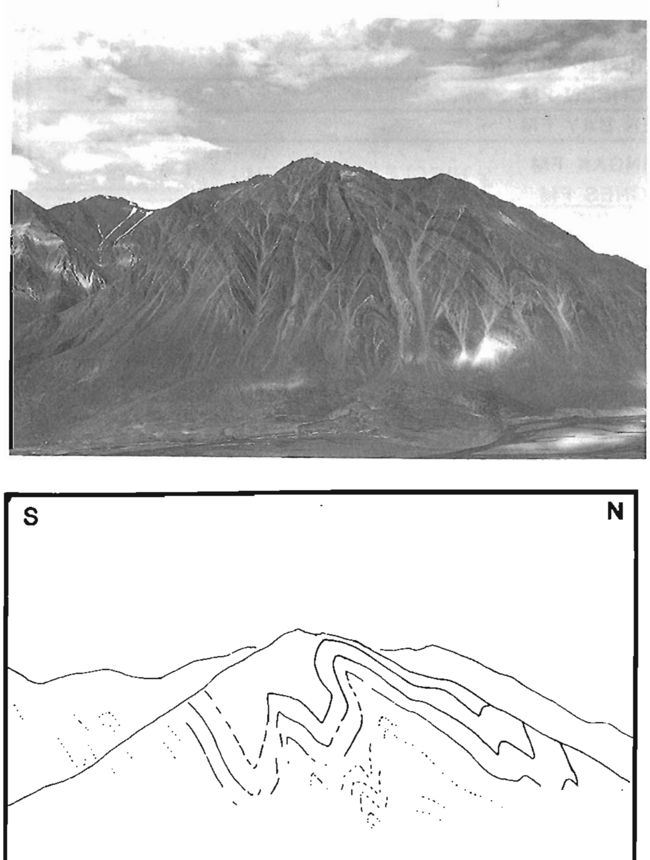


Figure 8. View from Tanquary Fiord of the chevron folds in the Grant Land Formation. Note the steep, northwest dipping axial planes of the folds. Foreground field of view is about 2 km.

In the Tanquary Fiord area, a diachronous lateral transition from the carbonate belt to the marginal facies belt occurs. The Canyon Fiord Formation and Belcher Channel Formation replace the Borup Fiord Formation — Nansen Formation couplet (Fig. 9). The Canyon Fiord Formation consists of basal conglomerate and sandstone, Bashkirian in age, whereas the Belcher Channel Formation consists of fossiliferous limestone, as young as Sakmarian (Mayr, work in progress). The diachroneity of the basal conglomerates of the Borup Fiord and Canyon Fiord formations is interpreted as resulting from the widening of the Sverdrup Basin between Namurian and Sakmarian times.

At the head of Tanquary Fiord and in the Yelverton Inlet area, Permian formations consist of 100 m of sandstone and shale of the Sabine Bay Formation, which are overlain disconformably by up to 150 m of siltstone and limestone of the Assistance Formation. In the area of the Tanquary High, Carboniferous to Permian formations are absent, and Triassic strata lie unconformably over the lower Paleozoic folds.

The Triassic stratigraphy of the map area is characterized by fine grained clastic formations, each separated by unconformities. These unconformities account for the thinning of the Triassic sequence over the Tanquary High. The thickest and most prominent formation is the Heiberg Formation. It consists of quartzarenite, siltstone and shale, and ranges in thickness from 540 m in the Tanquary Fiord area to 700 m in Yelverton Pass. Numerous tholeiitic sills of Cretaceous age thicken the formation. The underlying Blind Fiord, Pat Bay, Hoyle Bay and Barrow formations are thin and more recessive. The combined thickness of these formations does not exceed 200 m.

Above the Heiberg Formation, fine grained sandstone, siltstone and shale of the Sandy Point, McConnel, Hiccles Cove, Awingak and Deer Bay formations represent a conformable sequence of Jurassic age (Fig. 9). Up to 600 m of the section are exposed in the Ekblaw Lake area. In the Tanquary Fiord-Yelverton Inlet section, the Jurassic strata are cut off in the footwall of a Tertiary reverse fault.

The Cretaceous Isachsen and Christopher formations represent the youngest rocks exposed in the map area. Together they consist of up to 400 m of shale and siltstone with minor sandstone and conglomerate. The upper limit of the section is in footwall contact with a Tertiary reverse fault. No Tertiary rocks are exposed in the map area.

STRUCTURAL ANALYSIS OF THE SUBCROP MAPS

Subcrop maps below specific unconformities show the map patterns that result from the tectonic events predating the formation of the unconformities. A subcrop map is defined by the present-day outcrop and inferred subcrop extent of formations older than the unconformity. On the map, an original subcrop contact point between two formations is localized where the unconformity outcrop line overlaps two juxtaposed formations successively. The present-day outcrop pattern is of interest in that it shows the limits of erosion of specific formations, and therefore the limits of past antiforms.

Unconformity surfaces were also warped and faulted during the Eurekan Orogeny. The determination of the restored dip of the pre-unconformity strata is often limited to evaluating the sense of the dip.

The pre-Carboniferous subcrop map

The rugged topography of the Tanquary Fiord map area permits one to observe different structural levels, from the tightly folded structures of the lower Paleozoic strata to the upper part of the less deformed strata of the Sverdrup Basin. The lower Paleozoic structures act as a "basement" for the Sverdrup Basin. The pre- Carboniferous unconformity represents the peneplain following the first major structural modification of the map area (Figs. 4c, 6). The pre-Devonian subcrop map would be of interest, but nowhere are Devonian strata seen in direct contact with lower Paleozoic rocks. Devonian strata are restricted to a synformal trough located in the northern part of the map area. This synform is either fault-bounded, or synclinal, or a combination of both. Only the dip of the limbs of the Devonian synform could be well defined after restoring the dip of the overlying Carboniferous strata. Hence, the pre-Carboniferous subcrop map mainly shows the folded structures of the lower Paleozoic formations.

Carboniferous strata overlie the folded lower Paleozoic slate and quartzite with a sharp angular unconformity, and they overlie the planar strata of the Middle and Upper Devonian sandstone with an angular unconformity of less than 20° (Figs. 4b, 6). No Lower Devonian rocks are exposed in the map area. However, they ought to be considered since they do occur in other parts of the Devonian clastic wedge of the Arctic Islands (Embry and Klovan, 1976), and in the graben-type intramontane basins of Svalbard (Harland, 1969).

The pre-Carboniferous subcrop map shows the juxtaposition of a northeast-southwest trending, upright, subhorizontal anticlinorium of the Grant Land Formation, and a subparallel, inclined synclinorium to the north that involves the Danish River Formation. Hence, shortening increases northward, toward the collision front of the Pearya Terrane (Fig. 4c). These two large structures consist of a myriad of inclined chevron-type folds. Figure 6 does not show the smaller-scale, sawtooth-like outcrop pattern of the individual formations, but rather the maximum zero-edge limit of the extent of the specific formations prior to Carboniferous sedimentation.

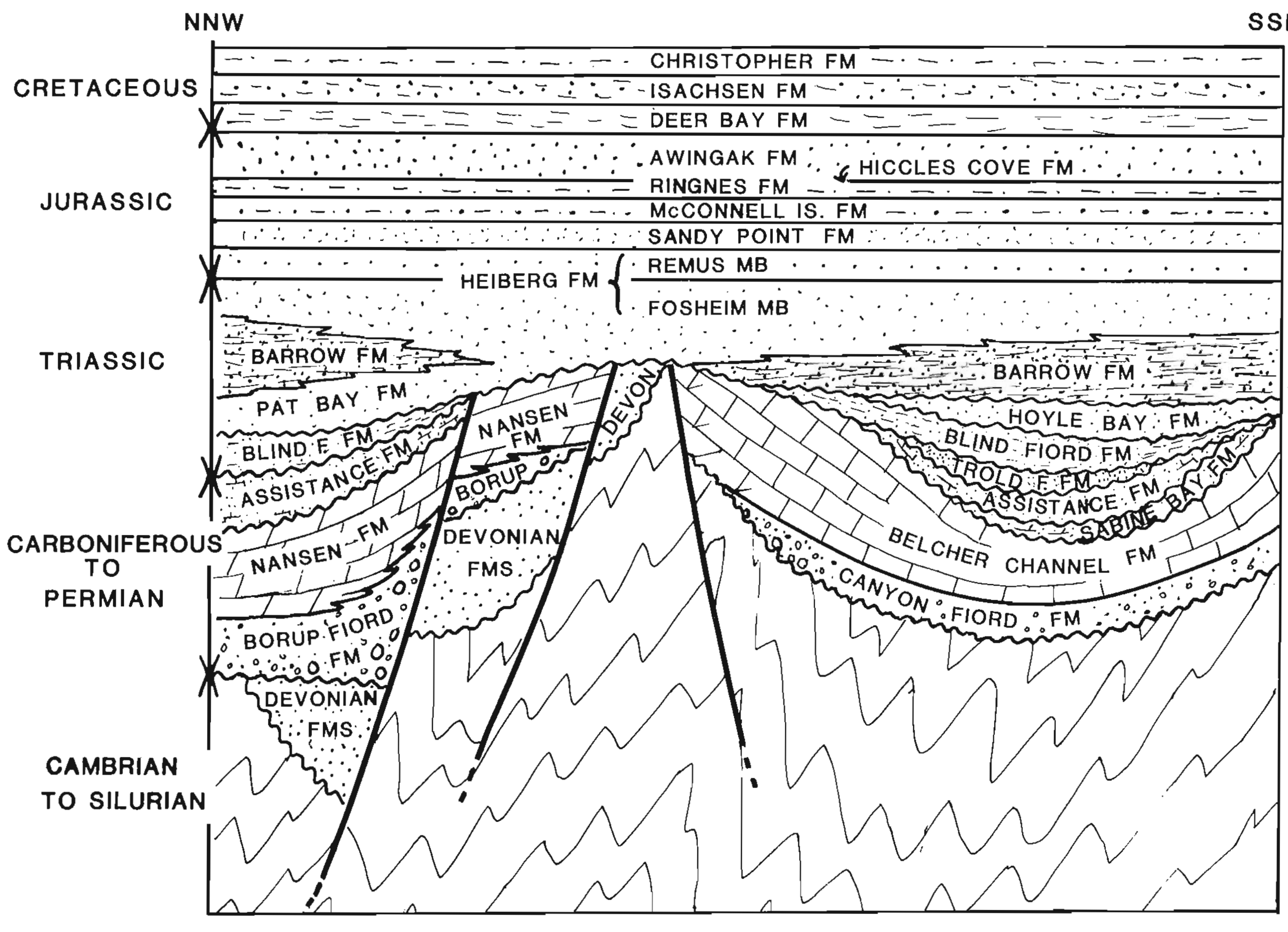


Figure 9. Diagrammatic stratigraphic column for the upper Paleozoic and Mesozoic formations of the map cross-sections (not to scale).

North of the Devonian subcrops are the probably exotic rocks of the Clements Markham Fold Belt. Consequently, the presence of a major fault is inferred between the overturned limb of the synclinorium of the Hazen Fold Belt at Yelverton Pass and the rocks of the Clements Markham Fold Belt at "The Bridge" of Yelverton Inlet (Fig. 6). The compressional nature of the deformation suggests a reverse fault, assumed to be parallel to the steep, regional, axial planar cleavage of the lower Paleozoic folds. However, the tectonic event that produced the folding of the region represents the last phase of the accretion of the Pearya Terrane, which itself was demonstrated farther north to be largely sinistral (Trettin, 1987a). The inferred fault boundary between the Clements Markham and the Hazen fold belts therefore could be of a sinistral nature also (Figs. 4c, 6) and would coincide with the Feilden Fault Zone northeast of the map area (Trettin, 1987b).

The measured paleocurrent directions and the dominantly cherty clast composition of the Devonian Okse Bay Formation indicate that the Hazen Formation of the northern flank of the Grant Land Anticlinorium is the most probable source of clastics. The resistant quartzites of the Grant Land Anticlinorium may have formed a topographic barrier separating the Devonian synform from the rest of the Devonian clastic wedge basin, widespread to the south over the Arctic Islands. The synform can then be interpreted as a Devonian subbasin.

Also, the minor unconformity (less than 20°) with the overlying Carboniferous strata implies that the Devonian strata were slightly tilted during latest Devonian-Early Carboniferous time. This phase of deformation is coeval with the "Ellesmerian Orogeny", which produced the Central Ellesmere and Parry Islands fold belts across the Arctic Islands.

In Moskovian time, the Tanquary High area was defined by the deposition of coarse fluviatile to deltaic (quartzite and chert) conglomerates of the Canyon Fiord Formation, which were shed from the exposed highlands of the Grant Land Anticlinorium. Hence the Tanquary High was not yet exhumed. The rest of the Sverdrup Basin (as represented in Fig. 4b) was defined by the deposition of the Nansen Formation carbonates (Mayr, work in progress; Fig. 9).

Permian unconformities must have preserved the Carboniferous structures, but the pre-Triassic uplift led to erosion of the Permian strata of the Tanquary High, thus leaving little control for the making of a pre-Permian subcrop map.

The pre-Triassic subcrop map

Strata of the Triassic transgression, which preserve beneath them the cumulative effects of all the Paleozoic tectonic events, are still widespread enough at the present level of exposure to permit the reconstruction of the pre-Triassic subcrop map (Figs. 4b, 7).

The Triassic transgression marks a major change in the paleogeography of the Sverdrup Basin with the immersion of the Tanquary High and the covering of the synsedimentary normal faults that were associated with the formation of the high (Fig. 4b). The pre-Triassic subcrop map shows the maximum extent of the Tanquary High after its development during the Carboniferous and Permian. The erosional edge of the Permian Sabine Bay and Trold Fiord formations can be interpreted as a boundary marking the maximum extent of the Tanquary High. As such, the high was an exhumed, fault- controlled structure trending east-west for about 100 km from the southern margin of the Sverdrup Basin (Figs. 3, 7). It was nearly 50 km wide and separated a southern subbasin from the main depocenter of the Sverdrup Basin to the north. The uplands of the high consisted of outcrops of the Grant Land Anticlinorium and of its flanking synclinorium. Therefore, the Grant Land, Hazen and Danish River formations were potential sources of clastics for the Triassic sediments. Part of the southern limb of the Devonian synform was also exposed prior to Triassic transgression. It was affected by normal faults striking either parallel or perpendicular to the Sverdrup Basin trend (Fig. 7). Both types of faults are pre-Triassic since the Triassic strata still remain at the same structural level along and across Yelverton Pass.

One major normal fault can be seen near Yelverton Pass below flat-lying and undisturbed Triassic strata (Fig. 10). It downdrops the carbonates of the Nansen Formation northward beside the sandstones of the Okse Bay Formation. This fault was later reactivated as a minor reverse fault that produced a fold ramp above a low-angle detachment in the Triassic strata of the Blind Fiord and Heiberg formations (Fig. 10). It is possible, therefore, that more Tertiary reverse faulting reactivated pre-Triassic normal faults and obscured the original fault motions.

A transverse normal fault strikes along Yelverton Pass and underneath the glacier corridor leading to Yelverton Inlet. Only the southern termination of the fault is visible in the area in the lower Paleozoic rocks, where it remains steep and straight along its strike. At a higher structural level, near the surrounding summits, the Carboniferous strata outcrop at different elevations; this is caused by their positions with respect to the footwall and the hanging wall of the transverse fault. More than 300 m of upper Paleozoic strata were measured by Mayr (work in progress) in the hanging wall of the transverse normal fault (Point D' of Figure 7) whereas only a few metres of the same strata can be observed along strike below the still undisturbed Triassic unconformity, 7 km from the former section and in the footwall of the transverse fault (Point D of Figure 7). A 50 m wide dyke now marks the strike of this transverse fault across the overlying Jurassic formations of Yelverton Pass (Fig. 2), and indicates its reactivation in Cretaceous time.

The extent of this major cross-basinal fault is poorly known but it may be inferred from the topographic depression of the glacier corridor stretching for more than 20 km between Yelverton Pass and "The Bridge" near Yelverton Inlet.

In the area of Yelverton Inlet, paleocurrent directions from a section of the Borup Fiord Formation located along the depositional axis of the Sverdrup Basin (Mayr, work in progress) indicate a dominantly northwest-east — southeast trend (Fig. 7), thus marking a highland to the west. Consequently, the Tanquary High possibly extends northwestward along a transverse trend to produce the saddle marking a high point along the axis of the Sverdrup Basin.

CONCLUSIONS

The Tanquary High is a positive structure on the southern margin of the Sverdrup Basin in northern Ellesmere Island. It partially reactivates an older positive structure, the Grant Land Anticlinorium, and is locally modified by a steep reverse fault zone of Tertiary age to form parts of the Grant Land High. The Tanquary Structural High affects the paleogeography of deposits as old as Early Carboniferous and as young as Jurassic, and thus is coeval with the growth of the Sverdrup Basin. The Tanquary High is oblique to the strike of the Sverdrup Basin but it is controlled by transverse and strike- parallel normal faults. Transverse faults that are coeval with the growth of the Sverdrup Basin are rare in the Arctic Islands, but they strike for long distances across the Sverdrup Basin (Plauchut, 1972; Kerr, 1975; Maurel, work in progress; Beauchamp et al., this volume). To the south of the Tanquary High, the Tanquary Fiord subbasin is flanked by a minor salient of exposed lower Paleozoic rocks (Fig. 7). This salient outlines a structural high secondary and subparallel to the larger Tanquary High. Hence the Tanquary High may not be the only structure affecting the paleogeography of the southern margin of the Sverdrup Basin.

The offset of the depocenter axes of the Sverdrup Basin and the subsequent deflection of facies and isopachs may result from the resistance to subsidence of the fault block system of the Tanquary Structural High. Alternatively, the Tanquary Structural High may result from the offset development of two, independent depocenter axes of the Sverdrup Basin.

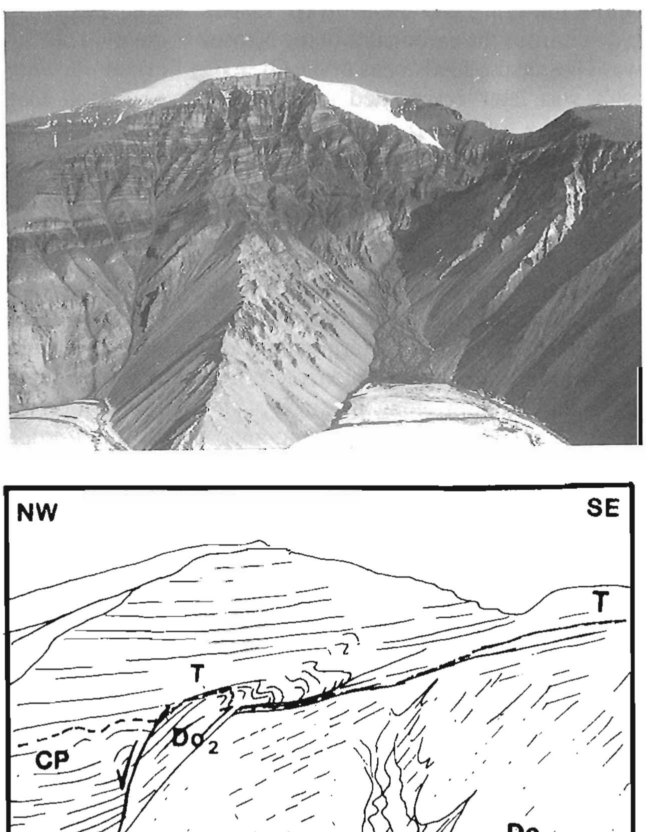


Figure 10. View of a pre-Triassic normal fault reactivated as a reverse fault during the Eurekan Orogeny. Note the slight angular unconformity between the planar strata of the Okse Bay Formation and the overlying beds of Triassic age. Field of view is 500 m, looking northeast from Yelverton Pass. CP, Permo-Carboniferous; T, Triassic; DO_2, Upper Devonian (upper member Okse Bay Formation); DO, Devonian (lower member Okse Bay Formation).

ACKNOWLEDGMENTS

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Paleocene palynoflora from northern Somerset Island, District of Franklin, N.W.T.†

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McIntyre, D.J., Paleocene palynoflora from northern Somerset Island, District of Franklin, N.W.T.; in Current Research, Part G, Geological Survey of Canada, Paper 89-1G, p. 191-197, 1989.

Abstract

Samples from outcrops of the Eureka Sound Group at the northern end of Somerset Island yielded diverse pollen assemblages, which are dated as Late Paleocene, a determination based particularly on members of the Momipites-Caryapollenites *lineage. The pollen floras are similar to other Paleocene assemblages from the Canadian Arctic, and also have characteristics in common with those from both Spitsbergen and the western interior basin of North America.*

Résumé

On a trouvé dans des échantillons provenant d'affleurements du groupe d'Eureka Sound à l'extrémité nord de l'île Somerset diverses associations de pollens qui remontent au Paléocène supérieur, résultat basé notamment sur des membres de la lignée évolutive Momipites-Carvapollenites. *Les flores de pollen s'apparentent à d'autres associations du Paléocène de l'Arctique canadien, ainsi qu'à des flores contemporaines provenant du Spitzberg et du bassin intérieur de l'ouest de l'Amérique du Nord.*

† Contribution to Frontier Geoscience Program.

INTRODUCTION

Stewart (1987) noted that several fault-bounded outliers of the Eureka Sound Group are present on Somerset Island. The outcrops of one outlier, in a narrow, fault-bounded valley southeast of Cunningham Inlet, at the northern end of Somerset Island, were mapped by Stewart and Kerr (1984), who show a series of exposures over a distance of approximately 32 km. Seven spot samples were collected from this outlier in 1970 by W.S. Hopkins. Palynological results from these samples were recorded by Hopkins (1971), who noted that all samples had essentially similar microfloral assemblages. He concluded that uppermost Cretaceous and/or lowermost Tertiary rocks were present along this fault zone.

At the request of W.D. Stewart, these palynology preparations have been reexamined. The supposed uppermost Cretaceous/lowermost Tertiary material of Hopkins (1971) lacks many of the species characteristic of Upper Cretaceous pollen floras from northern Canada, and contains species like those of early Tertiary microfloras recently recorded from the Eureka Sound Group (McIntyre, in Ricketts, in press). Because the determinations from the original preparations were somewhat inconclusive, new preparations using refinements introduced in recent years, were made. These are cleaner than the original ones and contain a greater number and variety of spores and pollen. Because of the richer assemblages, it is possible to present a more precise age determination for the Eureka Sound Group outlier at the north end of Somerset Island.

SAMPLES

All the samples examined were collected by W.S. Hopkins from outcrops in a deep, narrow, straight-sided valley, southeast of Cunningham Inlet, at the north end of Somerset Island. Exact locations for the samples were not recorded by Hopkins (1971) but the information given indicates that they were collected from the valley between 73°54′N, 92°35′W and 74°02′N, 93°33′W (NTS 58 C, F). The location of the outlier is shown by Stewart and Kerr (1984). The dominant lithology of the outcrops is fine to medium grained sandstone, with some carbonaceous siltstone and pebble-conglomerate present. Crossbedding is common, and poorly preserved plant remains, apparently stems, twigs or rootlets, are abundant in many the sandstones, which also contain some carbonized leaves, probably of angiosperms (Hopkins, 1971). The geology of the area from which the samples were collected has been documented by Hopkins (1971), Stewart and Kerr (1984), and Stewart (1987). Details of the carbonaceous siltstone samples examined are as follows:

Field Number	GSC Locality Number	Palynology Number
HFA-70-25	C-7163	P759-25
HFA-70-26	C-7164	P759-26
HFA-70-27	C-7165	P759-27
HFA-70-28	C-7166	P759-28
HFA-70-29	C-7167	P759-29
HFA-70-30	C-7168	P759-30
HFA-70-31	C-7169	P759-31

POLLEN AND SPORE ASSEMBLAGES

The new preparations yielded lower Tertiary pollen and spore assemblages containing the species recorded by Hopkins (1971) as well as others. The assemblages from all seven samples are similar, but vary considerably in the total amount of pollen recovered, species diversity and relative abundance of each species. All samples have significant amounts of reworked dinoflagellates of Campanian age, and pollen, mainly of early Maastrichtian age. The reworked palynomorphs are conspicuous elements of the assemblages, and initially might have indicated a Late Cretaceous age, but they do not constitute more than four per cent of the assemblage in any of the samples. Gymnosperm pollen is dominant in all samples: Pinaceae (*Picea* and *Pinus* types) and the Taxodiaceae/Cupressaceae complex. These groups

PLATES 1, 2

Slides containing the figured 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, Calgary, Alberta. In the descriptions for Plates 1 and 2, the species name is followed by the GSC locality number (prefixed C), the slide number (prefixed P), stage coordinates, and the GSC type number (prefixed GSC). The stage coordinates are for Reichert-Jung Polyvar microscope 392166 at the Institute of Sedimentary and Petroleum Geology, Geological Survey of Canada, Calgary, Alberta. All figures on both plates are magnified at x1000.

PLATE 1

Figure 1. *Schizaea triangula*
C-7164, P759-26f, 525 × 1164, GSC 93592.

Figure 2. *Cicatricosisporites cicatricosoides*
C-7164, P759-26f, 375 × 1108, GSC 93593.

Figure 3. *Sequoiapollenites* sp.
C-7165, P759-27f, 464 × 1180, GSC 93594.

Figure 4. *Liliacidites* sp.
C-7163, P759-25g, 679 × 1153, GSC 93595.

Figure 5. *Sparganium* sp.
C-7163, P759-25g, 405 × 1147, GSC 93596.

Figure 6. *?Quadrapollenites* sp.
C-7163, P759-25f, 498 × 1199, GSC 93597.

Figure 7. *Arecipites tenuiexinous*
C-7165, P759-27g, 675 × 1268, GSC 93598.

Figure 8. Ericaceae
C-7163, P759-25f, 412 × 1134, GSC 93599.

Figure 9. cf. *Acer* sp.
C-7163, P759-25f, 684 × 1249, GSC 93600.

Figure 10. *Ulmus* sp.
C-7169, P759-31g, 586 × 1231, GSC 93601.

Figure 11. *Pachysandra* sp.
C-7164, P759-26f, 487 × 1237, GSC 93602.

Figure 12. *Tilia* sp. (*T. vescipites*)
C-7165, P759-27e, 590 × 1257, GSC 93603.

Figure 13. *Liquidambar* sp.
C-7169, P759-31f, 579 × 1173, GSC 93604.

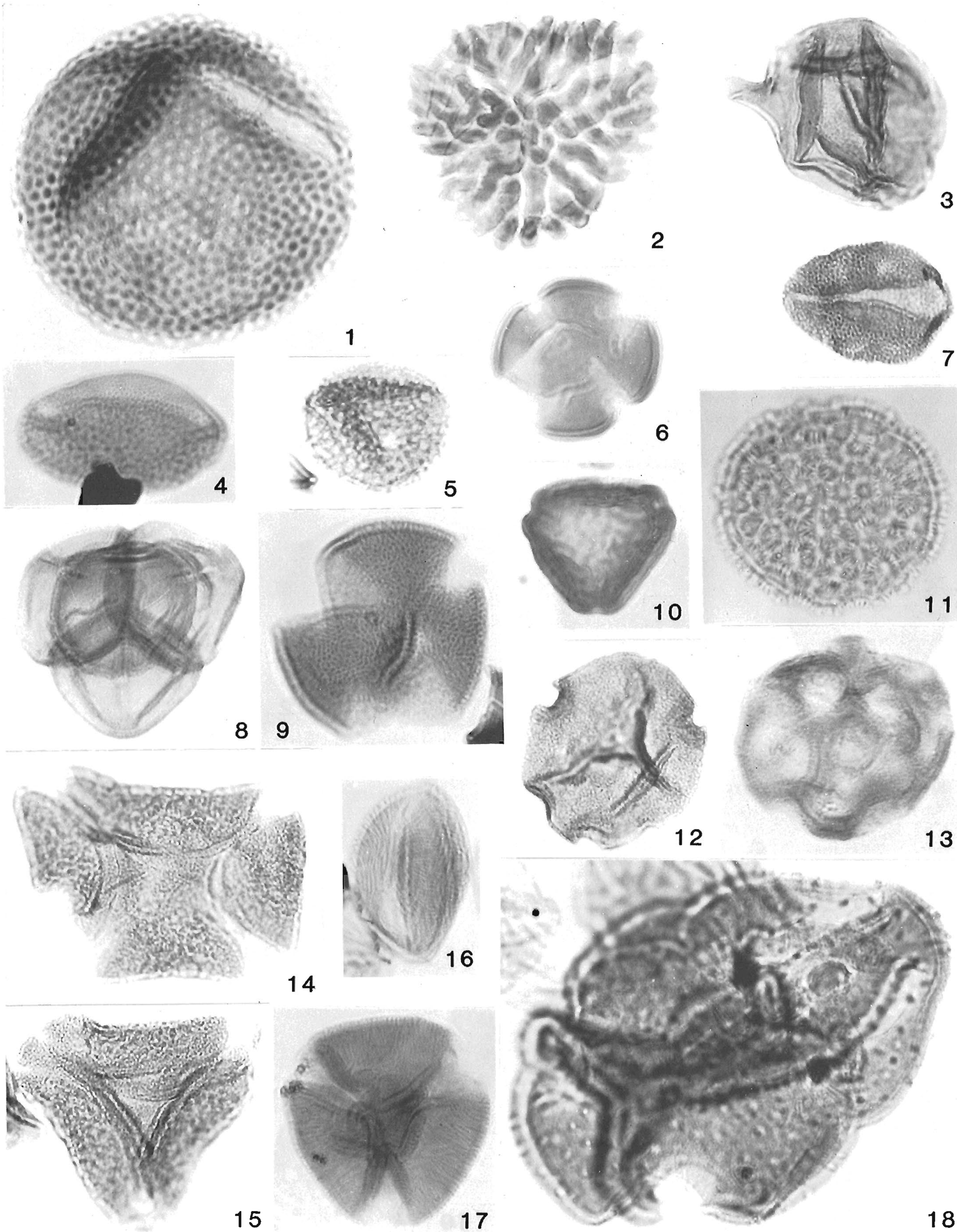
Figure 14. *Insulapollenites rugulatus*
C-7165, P759-27f, 399 × 1235, GSC 93605.

Figure 15. *Insulapollenites rugulatus*
C-7165, P759-27f, 642 × 1220, GSC 93606.

Figure 16. ?Rosaceae
C-7164, P759-26g, 636 × 1196, GSC 93607.

Figure 17. ?Rosaceae
C-7169, P759-31f, 489 × 1245, GSC 93608.

Figure 18. *Saxonipollis* sp. A
C-7169, P759-31f, 624 × 1164, GSC 93609.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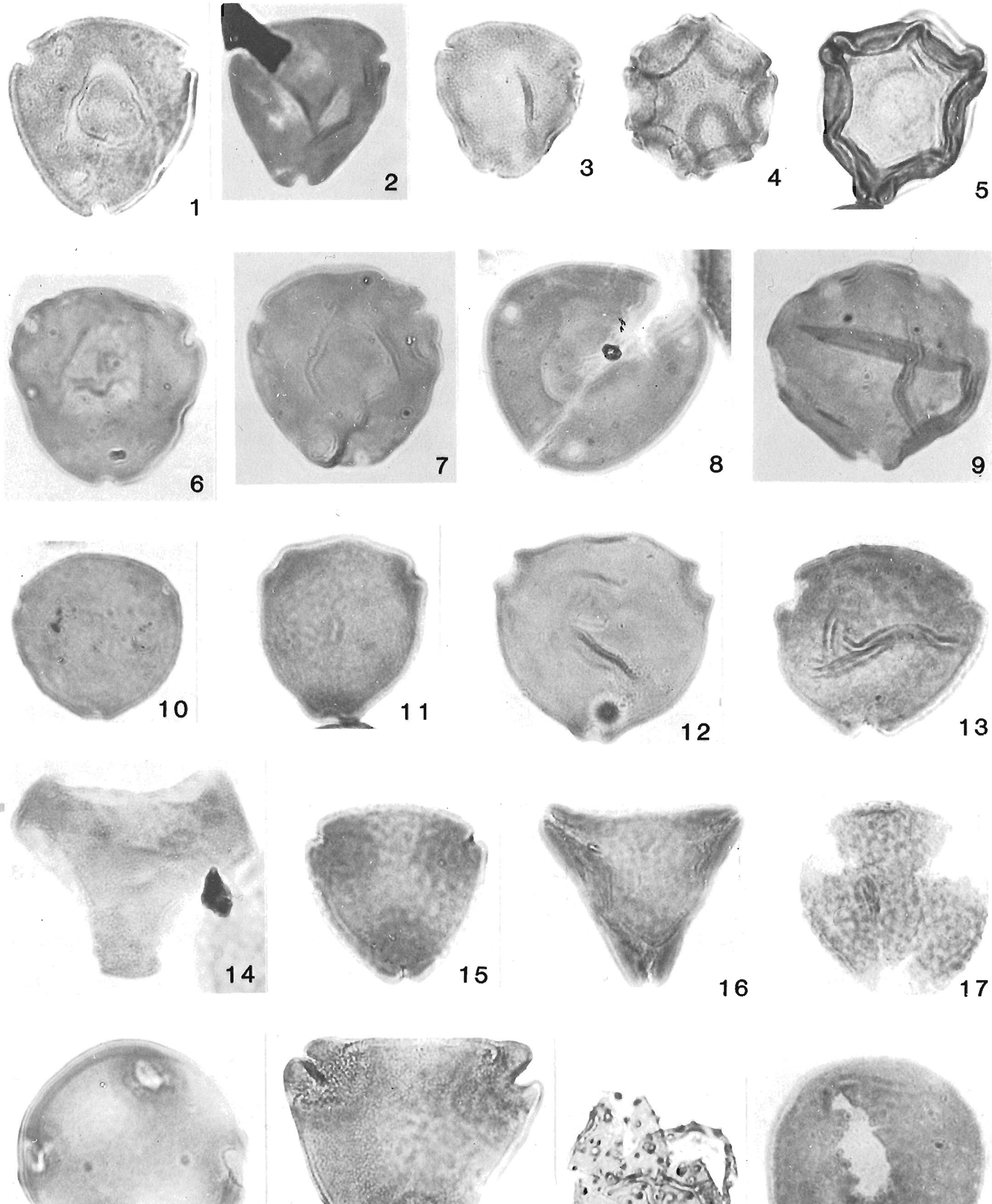
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together constitute over 50 per cent of the total pollen and spore flora in all samples. Preliminary results were presented by McIntyre (1985; and in Stewart, 1987).

The taxa listed below are considered to be indigenous to the Cunningham Inlet samples. See Plates 1 and 2 for illustrations of many of the taxa listed below:

Sphagnum spp. [*Stereisporites antiquasporites* (Wilson and Webster) Dettmann and *Stereisporites regius* (Drozhastichich) Felix and Burbridge]
Lycopodium sp. (*Reticulatisporites incomptus* Manum)
Laevigatosporites spp.
Osmunda spp.
Hamulatisporites hamulatis Krutzsch
Hazaria sheoparii Srivastava
Cicatricosisporites cicatricosoides Krutzsch
Schizaea triangula Stanley
Pinaceae (*Pinus* and *Picea* spp.)
Taxodiaceae/Cupressaceae
Sequoiapollenites sp.
Alnus sp.
Betula sp.
Corylus sp.
Triporopollenites mullensis Simpson
Triporopollenites spp.
Paraalnipollenites alterniporus (Simpson) Rouse and Srivastava
Trudopollis barentsii Manum
Trudopollis sp., Type B of Manum
Nudopollis sp.
?*Quadrapollenites* sp.
Tetrapollis conspectus Manum
Ulmus sp. (*Ulmipollenites undulosus* Wolff)
Momipites anellus Nichols and Ott
Momipites tenuipolus Anderson
Momipites ventifluminis Nichols and Ott
Momipites wyomingensis Nichols and Ott
Caryapollenites imparalis Nichols and Ott
Caryapollenites inelegans Nichols and Ott
Caryapollenites veripites (Wilson and Webster) Nichols and Ott
Caryapollenites wodehousei Nichols and Ott
cf. *Quercus* sp.
Pterocarya sp.
Ericaceae (*Ericipites* sp.
Liquidambar sp.
cf. *Acer* sp.
?Rosaceae
Pachysandra sp. (*Erdtmanipollis procumbentiformis* (Samoilovitch) Krutzsch)
Tilia sp. (*T. vesciptes* Wodehouse)
Insulapollenites rugulatus Leffingwell
Saxonipollis sp. A of Ioannides and McIntyre
Tricolpites spp.
Tricolporopollenites spp.
cf. *Cercidiphyllum* sp.
Sparganium sp.
Pandaniidites radicus Leffingwell
Milfordia sp.
Arecipites tenuiexinous Leffingwell
Arecipites columellus Leffingwell
Liliacidites spp.

AGE AND CORRELATION OF PALYNOFLORAS

The pollen assemblages of the Somerset Island samples clearly show that they are of Paleocene age, and are not as old as latest Cretaceous as suggested by Hopkins (1971). The presence of *Insulapollenites rugulatus*, and species of *Momipites* and *Caryapollenites* provides evidence that these sediments are undoubtedly Paleocene. Nichols and Ott (1978) established a zonation for the Paleocene in Wyoming based on morphological changes in pollen of the *Momipites-Caryapollenites* lineage. This zonation was discussed by Pocknall (1987) who noted its validity and utility. Demchuk (1987) used the lineage, with slight modifications, and established an equivalent zonation in the Paleocene of central Alberta. Some of the *Momipites* and *Caryapollenites* species of Nichols and Ott (1978) appear in the Somerset Island samples, but always very rarely, in contrast to the situation in Alberta and the northern western interior of the U.S.A., where they are usually abundant and sometimes dominant. The presence of *Momipites anellus, M. ventifluminis* (zones P3-P6 of Nichols and Ott), *Caryapollenites*

PLATE 2

Figure 1. *Momipites anellus*
C-7165, P759-27e, 412 × 1254, GSC 93610.
Figure 2. *Momipites ventifluminis*
C-7166, P759-28g, 661 × 1244, GSC 93611.
Figure 3. *Momipites wyomingensis*
C-7165, P759-27g, 710 × 1224, GSC 93612.
Figure 4. *Alnus* sp.
C-7165, P759-27g, 663 × 1230, GSC 93613.
Figure 5. *Paraalnipollenites alterniporus*
C-7165, P759-27g, 556 × 1227, GSC 93614.
Figure 6. *Caryapollenites wodehousei*
C-7166, P759-28f, 376 × 1185, GSC 93615.
Figure 7. *Caryapollenites imparalis*
C-7166, P759-28f, 498 × 1186, GSC 93616.
Figure 8. *Caryapollenites veripites*
C-7169, P759-31f, 542 × 1191, GSC 93617.
Figure 9. *Pterocarya* sp.
C-7169, P759-31f, 642 × 1243, GSC 93618.
Figure 10. *Corylus* sp.
C-7166, P759-28g, 428 × 1278, GSC 93619.
Figure 11. *Triporopollenites mullensis*
C-7163, P759-25f, 412 × 1147, GSC 93620.
Figure 12. *Triporopollenites* sp.
C-7163, P759-25f, 521 × 1224, GSC 93621.
Figure 13. *Triporopollenites* sp.
C-7169, P759-31f, 487 × 1222, GSC 93622.
Figure 14. cf. *Cercidiphyllum* sp.
C-7163, P759-25f, 624 × 1182, GSC 93623.
Figure 15. *Trudopollis* sp., Type B
C-7163, P759-25f, 457 × 1122, GSC 93624.
Figure 16. *Nudopollis* sp.
C-7166, P759-28f, 458 × 1218, GSC 93625.
Figure 17. cf. *Quercus* sp.
C-7163, P759-25f, 539 × 1266, GSC 93626.
Figure 18. *Tetrapollis conspectus*
C-7169, P759-31f, 497 × 1217, GSC 93627.
Figure 19. *Trudopollis barentsii*
C-7165, P759-27f, 652 × 1238, GSC 93628.
Figure 20. *Pandaniidites radicus*
C-7165, P759-27e, 471 × 1133, GSC 93629.
Figure 21. *Milfordia* sp.
C-7163, P759-25f, 378 × 1298, GSC 93630.

imparalis, C. wodehousei (P4-P6), *C. inelegans* and *C. veripites* (both restricted to zones P5-P6) show that the Somerset Island pollen floras are similar to those recorded from farther south in North America by Leffingwell (1971), Nichols and Ott (1978), Pocknall (1987) and Demchuk (1987). As the southern assemblages are of undoubted Late Paleocene age, the assemblages from Somerset Island also are considered to be Late Paleocene. *Caryapollenites veripites* has previously been recorded from the Late Paleocene and Early Eocene of the Arctic by Rouse (1977), Doerenkamp et al. (1976) (*Caryapollenites* sp.), and Ioannides and McIntyre (1980) (*Caryapollenites* sp. B). Further evidence of a Late Paleocene age and correlation with southern assemblages is the presence of *Insulapollenites rugulatus*, a rare pollen species recorded from the late Paleocene by Leffingwell (1971), Nichols and Ott (1978), Pocknall (1987) and Demchuk (1987). This species is now known to occur also in the middle Paleocene (Sweet et al., 1989). The presence in most samples of *Triporopollenites mullensis* and *Paraalnipollenites alterniporus* is also indicative of a Paleocene age. The very rare occurrence of *Tilia vescipites* indicates an age no older than Late Paleocene. Neither Doerenkamp et al. (1976) nor Rouse (1977) recorded *Tilia* pollen from Paleocene strata in the Arctic Archipelago but *T. vescipites* occurs very rarely in the late Paleocene from a section at Strand Fiord, Axel Heiberg Island (McIntyre, in Ricketts, in press). It also occurs in the Late Paleocene of Wyoming, (Nichols and Ott, 1978; Pocknall, 1987), and is very rare in the Late Paleocene of Spitsbergen (M.J.Head, pers. comm., 1987). *Saxonipollis* sp. of the Somerset Island assemblage appears to be the same as *Saxonipollis* sp. A, recorded by Ioannides and McIntyre (1980) from the Mackenzie Delta. *Droseridites spinulosus* Manum from the Paleocene of Spitsbergen is a very similar form.

Trudopollis barentsii and *Trudopollis* sp. Type B occur in most samples, abundantly in some. These species were recorded from probable Paleocene of Spitsbergen by Manum (1962), who did not see them in a sample from Ellesmere Island. Recent unpublished work (D.J.McIntyre) on palynology of sections from Axel Heiberg and Ellesmere islands indicates that these species occur rarely in the Paleocene. The presence of abundant *Trudopollis* spp. in Somerset Island and Spitsbergen samples provides evidence that palynofloras from the southern Arctic islands have affinities with lower Tertiary microfloras, where this type of pollen of the Normapolles group is common. The occurrence together of species of *Trudopollis* and the *Momipites-Caryapollenites* complex (not recorded from Spitsbergen by Manum, 1962) suggests an overlap, in the southern Arctic, of eastern and western floral provinces. Pollen of *Momipites-Caryapollenites* is abundant in the Paleocene of the western interior of North America where pollen of the Normapolles group is rare. Further evidence of affinity of the Somerset Island palynoflora with the European Normapolles pollen flora is the presence, on both Somerset Island and Spitsbergen, of *Tetrapollis conspectus*. *Nudopollis* sp. of Somerset Island is similar to *Nudopollis* sp. from the Paleocene of Banks Island (Doerenkamp et al., 1976) and Caribou Hills, Mackenzie Delta (Ioannides and McIntyre, 1980), and *Extratriporopollenites* sp. from Bonnet Plume Basin (Rouse and Srivastava, 1972).

The northern Somerset Island pollen assemblages are very similar to, and are apparently correlative with, palynofloras previously described from Ellesmere Island (P3 and P4 zones of Rouse, 1977), Banks Island (T1b Zone of Doerenkamp et al., 1976) and Mackenzie Delta (lower part of Interval B of Ioannides and McIntyre, 1980). Recent work shows that closely similar Late Paleocene pollen floras are present in the Strand Bay Formation and the lower part of the Iceberg Bay Formation, Eureka Sound Group, at Strand Fiord on Axel Heiberg Island (McIntyre, in Ricketts, in press). Choi (1983) has also described similar pollen assemblages from sections on Ellesmere and Axel Heiberg islands. It is evident that the pollen floras described here from Somerset Island are widespread in the Arctic Archipelago, and that work in progress will provide further details of these assemblages.

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New palynological data from Cornwall Arch, Cornwall and Amund Ringnes islands, District of Franklin, N.W.T.[†]

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McIntyre, D.J. and Ricketts, B.D., New palynological data from Cornwall Arch, Cornwall and Amund Ringnes islands, District of Franklin, N.W.T.; in Current Research, Part G, Geological Survey of Canada, Paper 89-1G, p. 199-202, 1989.

Abstract

Re-evaluation of palynofloras from outliers of Tertiary sandstones on Cornwall and Amund Ringnes islands shows that the age of these strata can now be defined as Lower Eocene, compared to previous age assignments of Paleocene-Eocene. This determination places the upper limit for uplift and erosion along Cornwall Arch within the Early Eocene, a date that has important paleogeographic implications for the Eureka Sound Group in the eastern Arctic.

Résumé

La réévaluation des palynoflores provenant de lambeaux de recouvrement des grès du Tertiaire dans les îles de Cornwall et Amund Ringnes indique que l'âge de ces couches peut maintenant être situé dans l'Éocène inférieur, contrairement aux évaluations antérieures qui les font remonter au Paléocène-Éocène. Cela signifie que la limite supérieure du soulèvement et de l'érosion le long de l'arche de Cornwall se situerait au début de l'Éocène inférieur, date qui a des répercussions paléogéographiques importantes pour le groupe d'Eureka Sound dans l'est de l'Arctique.

[†] Contribution to Frontier Geoscience Program.

INTRODUCTION

Cornwall Arch, a major anticline that plunges northward through Cornwall and Amund Ringnes islands (Balkwill, 1974, 1983), is an important component of the Late Cretaceous to early Tertiary sedimentary and tectonic history of the eastern Arctic Archipelago (Fig. 1). It has been suggested (Balkwill et al., 1975; Balkwill, 1983) that Cornwall Arch, and perhaps its eastern counterpart, Princess Margaret Arch, were the earliest manifestations of the Eurekan Orogeny. This uplift occurred between Maastrichtian and Paleocene-Eocene time, based on palynological analyses (Hopkins, in Balkwill, 1983). In subsequent paleogeographic reconstructions of Upper Cretaceous- Paleogene sedimentary strata (Eureka Sound Group), in the general area of Axel Heiberg and western Ellesmere islands, Cornwall Arch was viewed as a major topographic feature separating West Sverdrup Basin from Strand Fiord Basin (e.g., Miall, 1981, 1986). Likewise, the ancestral Princess Margaret Arch was visualized as separating Strand Fiord Basin from Remus Basin (Miall, op. cit.).

An alternative proposal for the Late Cretaceous-Eocene paleogeography of the region places the timing of the Princess Margaret Arch uplift at about middle Eocene (Ricketts, 1987; Ricketts and McIntyre, 1986). In this hypothesis, the Eureka Sound Group basins were considered to be a contiguous entity (called Fosheim Foredeep), bounded on the north and east by uplifted Sverdrup Basin, Franklinian and Precambrian Shield rocks, and on the west by Cornwall Arch. Within this framework, it was clearly implied that Cornwall Arch is older than the ancestral Princess Margaret Arch.

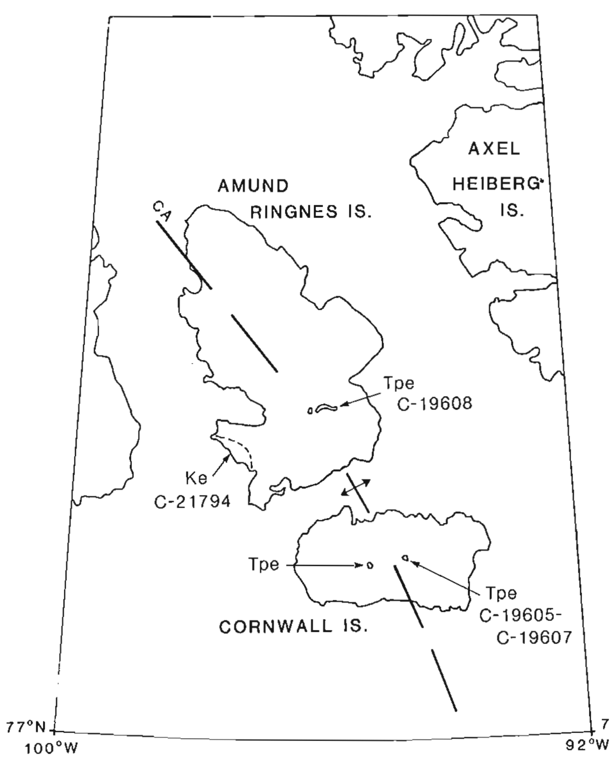

Figure 1. Location of Eureka Sound Group samples (identified by "C" numbers) on Amund Ringnes and Cornwallis islands (from Balkwill, 1983). Map unit Ke is Eureka Sound Group on Slime Peninsula; Tpe are the outliers of unnamed Tertiary strata.

Most interpretations of tectonic evolution for the eastern Arctic Archipelago consider the two arches to be of similar age. They were formed by compression in response to crustal shortening associated with the opening of Baffin Bay and rotation of Greenland away from North America. Gravity modelling using recently acquired data (Stephenson and Ricketts, 1989; Stephenson and Ricketts, work in progress), further suggests some similarities between the deep crustal structure of Cornwall and Princess Margaret arches. Clearly, a more precise definition of the age limits of the Cornwall Arch uplift is important. For this reason, samples that had been collected for palynological analyses from Cornwall and Amund Ringnes islands (Hopkins, in Balkwill, 1983) have been re-examined.

Stratigraphic framework, Amund Ringnes and Cornwall islands

Basal strata of the Eureka Sound Group are poorly exposed on Slime Peninsula, southwestern Amund Ringnes Island, and are at least structurally concordant with the underlying Kanguk Formation (Balkwill, 1983). Poor exposure precludes more concise definition of this contact. The 300 m of sandstone and shale are probably stratigraphically equivalent to the Expedition Formation on Axel Heiberg and Ellesmere islands (Ricketts, 1986).

An unnamed unit of sandstone (the Paleocene-Eocene map unit Tpe of Balkwill, 1983) occurs as erosional outliers on Amund Ringnes and Cornwall islands. A major unconformity exists between these sandstones and underlying, folded and eroded strata that range in age from Triassic to Early Cretaceous.

PALYNOLOGY

Samples that have been reprocessed and re-examined include four from the Tertiary unnamed unit, and one from the Eureka Sound Group. After re-processing there was a significant improvement in the palynofloras recovered, compared with the original preparations (because of greatly improved processing techniques). More precise age determinations of the samples were therefore possible. Detailed stratigraphic information for the samples is given in Balkwill (1983, p.73-74).

Eureka Sound Group

One sample (GSC loc. C-21794, BAA-72-204, P832-48) from about 210 m above the base of the Eureka Sound Group on Slime Peninsula, southwestern Amund Ringnes Island, (NTS 69 D; 78°00′N, 97°25′W) yielded an assemblage of Upper Cretaceous angiosperm pollen that is more diverse than that reported by Hopkins (in Balkwill, 1983). The important elements of this palynoflora include *Aquilapollenites quadrilobus* Rouse, *A. reticulatus* Stanley, *Azonia cribrata* Wiggins, *Expressipollis barbatus*

Chlonova, *E. ocliferius* Chlonova emend. Bondarenko, *Orbiculapollis globosus* (Chlonova) Chlonova, *Porosipollis porosus* (Mtchedlishvili) Krutzsch, *Tumidulipollis accuratus* (Chlonova) Bondarenko, *T. sibiricus* (Bondarenko) Bondarenko and *Wodehouseia edmonticola* Wiggins. This pollen assemblage indicates an early Maastrichtian age for this part of the Eureka Sound Group. *Wodehouseia edmonticola* is restricted to the early Maastrichtian (Wiggins, 1976) and *Azonia cribrata* occurs from late Campanian to early Maastrichtian. Both *Orbiculapollis globosus* and *Porosipollis porosus* first appear in the early Maastrichtian. *Expressipollis barbatus*, *E. ocliferius*, *Tumidulipollis accuratus* and *T. sibiricus* are also important late Campanian to early Maastrichtian species of northern Canada and Siberia. A similar diverse assemblage was recorded from the lower part of the Expedition Formation at Strand Fiord (McIntyre, in Ricketts, in press). Similar assemblages were also recorded from Arctic Canada by McIntyre (1974), Felix and Burbridge (1973) and Doerenkamp et al. (1976). These palynofloras were also determined to be of Maastrichtian age and their affinity with pollen floras from western Siberia was discussed.

Unnamed sandstone (Tpe)

After reprocessing, three samples (GSC locs. C-19605, C-19606, C-19607; HFA-72- 42-44, P832-1-3), from central Cornwall Island (NTS 59 C; 77°37′N, 94°42′W), yielded pollen floras that were more diverse than those of the original preparations, which had formed the basis for the age determinations of Hopkins (in Balkwill, 1983). In the three assemblages, which differ only in abundances and absence or presence of a few species, conifer pollen of Taxodiaceae-Cupressaceae and bisaccate grains referrable to *Picea* and *Pinus* are abundant. Tricolpate angiosperm pollen, consisting of numerous forms not further identified to species or genus level, is common in all samples, but it presently has limited stratigraphic significance. Other angiosperm pollen present, forms which range through the Paleocene and Eocene, includes *Alnus* sp., *Betula* sp., *Corylus* sp., indeterminate betulaceous pollen, Ericaceae, *Pterocarya* sp., *Ulmus* spp. (*Ulmoideipites krempii* Anderson and *Ulmipollenites undulosus* Wolff), *Pachysandra* sp. (*Erdtmanipollis procumbentiformis* (Samoilovitch) Krutzsch) and *Sparganium* sp.

An early Eocene age is indicated by a few other angiosperm pollen species in the Cornwall Island samples. The occurrence of *Pistillipollenites mcgregorii* Rouse clearly shows that the pollen floras are not younger than middle Eocene. Rouse (1977) determined that this species does not occur later than middle Eocene in the Arctic Islands. The presence of rare *Paraalnipollenites alterniporus* (Simpson) Srivastava and *Aquilapollenites tumanganicus* Bolotnikova indicates an upper age limit of early Eocene. Ioannides and McIntyre (1980) showed that *P. alterniporus*, a mainly Paleocene species, occurs rarely as late as early Eocene in the Mackenzie Delta area. *Aquilapollenites tumanganicus* was recorded from the Paleocene of the western coast of the Sea of Japan by Bolotnikova (1973) and the early Eocene of the Bohai region of China (as *A. spinulosus*) by Sung et al. (1978). Staplin (1976) recorded it (as *Aquilapollenites* sp.) from the Paleocene of the Mackenzie Delta. Choi (1983) recorded it as *Triprojectus echinatus* Mtchedlishvili, and McIntyre (in Ricketts, in press) noted its occurrence in upper Paleocene to lower Eocene strata of Axel Heiberg Island. It has not so far been seen in middle Eocene strata of the Eureka Sound Group.

Pollen of *Tilia* species occurs rarely in the Cornwall Island samples, indicating that they are not older than Eocene. Most of these grains are *T. vescipites* Wodehouse, but a few are referrable to *T. crassipites* Wodehouse. *Tilia* was recorded in the Eocene, but not the Paleocene, by Rouse (1977), Doerenkamp et al. (1976) and Ioannides and McIntyre (1980). However, McIntyre (in Ricketts, in press) noted that *T. vescipites* occurred rarely in upper Paleocene strata at Strand Fiord. It also occurs in the Paleocene of Wyoming (Nichols and Ott, 1978). Pocknall (1987) noted, however, that the more coarsely reticulate *T. crassipites* (recorded as *Intratriporopollenites* sp.) did not occur until the Eocene in Wyoming. Pollen of *Carya* species, which occurs commonly in the Cornwall Island samples, is considered here to be the same as that recorded from the lower Eocene by Doerenkamp et al. (1976), Ioannides and McIntyre (1980), Choi (1983) and McIntyre (in Ricketts, in press). These grains are morphologically distinct from the *Caryapollenites* species described from Wyoming by Nichols and Ott (1978) and recorded from Paleocene strata at Strand Fiord by McIntyre (in Ricketts, in press). At Strand Fiord, *Carya* spp. immediately succeed the uppermost Paleocene *Caryapollenites* species. The evidence from the *Tilia* and *Carya* pollen species, therefore, indicates that the Cornwall Island assemblages are no older than Eocene. Further evidence of an Eocene age is provided by the presence of *Tricolporopollenites kruschii* (Potonié) Thomson and Pflug, which was first recorded by Doerenkamp et al. (1976) and Rouse (1977) in Eocene strata in the Arctic. This species also occurs in Eocene strata in the Strand Fiord section, as noted by McIntyre (in Ricketts, in press). The occurrence of *Ilex* sp., *Nudopollis* sp. (of Doerenkamp et al.) and *Engelhardtia* sp. (small form, *Momipites coryloides* form A of Rouse) further indicates that the age of the assemblages is Eocene.

One sample (GSC loc. C-19608, BAA-72-118, P832-4) from southern Amund Ringnes Island (NTS 59 F; 78°11′N, 95°55′W) yielded a pollen flora that is similar to those from Cornwall Island, but less diverse. A Paleocene age for this sample was indicated by Hopkins (in Balkwill, 1983), but the palynological evidence from new preparations indicates an early Eocene age, the same as that of the Cornwall Island samples. The presence of *Pistillipollenites mcgregorii*, *Carya* spp. (the same as the Cornwall Island forms) and *Aquilapollenites tumanganicus* indicates an early Eocene age. The Paleocene *Aquilapollenites* cf. A. *spinulosus* Funkhouser, recorded by Hopkins, is here determined to be *A. tumanganicus*, which was seen in both the original and new preparations.

The palynofloras from the unnamed Paleocene-Eocene sandstone on both Cornwall and Amund Ringnes islands are here determined to be of early Eocene age. This determination, using improved preparations, is more precise than that recorded in Balkwill (1983). The pollen floras are closely comparable to lower Eocene pollen assemblages from the upper part of the Iceberg Bay Formation, Eureka Sound Group, at Strand Fiord, Axel Heiberg Island.

DISCUSSION

The sample from the Eureka Sound Group, 210 m above its contact with the Kanguk Formation and near the limit of exposure on Slime Peninsula, is of early Maastrichtian age. Samples of the unnamed sandstone unit from widely spaced localities have been consistently dated as early Eocene. Clearly, the period of uplift and erosion of strata along Cornwall Arch can only be restricted to between the early Maastrichtian and early Eocene, as suggested by previous workers. Of greater significance, however, is the fact that the upper age limit can be more precisely defined as early Eocene, rather than ranging from Paleocene to middle Eocene as stated by previous workers. Therefore, it is possible that the formation of Cornwall Arch occurred at a much later time than previously thought, for example during the late Paleocene or early Eocene. This permits the hypothesis that the uplift of Cornwall Arch is related to, but slightly earlier than, the uplift along the ancestral Princess Margaret Arch.

The possibility of a younger age for the Cornwall Arch uplift has important paleogeographic implications. For example, during the Late Cretaceous and earliest Tertiary, there would have been no effective barrier to sedimentation between Fosheim Foredeep and the putative basin west of the arch (called West Sverdrup Basin by Miall, 1986). In other words, the early depositional limits of the Eureka Sound Group would have been coextensive with those of the upper Mesozoic Sverdrup Basin.

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Foraminiferal zonation and biofacies analysis of Cenozoic strata in the Beaufort-Mackenzie Basin of Arctic Canada[†]

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Abstract

Twelve interval zones and seven assemblage zones are formally proposed for the Cenozoic strata of the Beaufort-Mackenzie Basin. The interval zonation is based on last appearance datums of benthic foraminifers and each datum is assigned a minimum absolute age. The assemblage zones are defined by discrete associations of benthic foraminifers.

The Paleocene to Eocene zones comprise mainly endemic agglutinated species, reflecting isolation from the world's major oceans. Oligocene to Miocene assemblages comprise agglutinated and calcareous species that indicate well established marine connections between the Arctic and North Atlantic oceans. Pliocene to Holocene assemblages consist almost entirely of calcareous benthic species that reflect climatic deterioration in the late Cenozoic.

Each assemblage zone is divided into a shallow-water biofacies, approximating the inner neritic zone, and a deeper-water biofacies, approximating the outer neritic to bathyal zones.

Résumé

Douze zones d'intervalles et sept zones d'associations sont formellement proposées pour les couches du Cénozoïque du bassin de la mer de Beaufort et du Mackenzie. La zonation des intervalles est basée sur des repères de dernière apparition de foraminifères benthiques, et à chaque repère est assigné un âge absolu minimal. Les zones d'associations sont définies par des associations discrètes de foraminifères benthiques.

Les zones du Paléocène à l'Éocène contiennent surtout des espèces agglutinées endémiques, dénotant leur isolement des grands océans. Les associations de la période s'étendant de l'Oligocène au Miocène contiennent des espèces agglutinées et calcaires, témoignant de l'existence de liaisons marines bien établies entre l'Arctique et l'Atlantique Nord. Les associations datant du Pliocène à l'Holocène se composent presque entièrement d'espèces benthiques calcaires qui font état d'une détérioration climatique au Cénozoïque supérieur.

Chaque zone d'association est divisée en biofaciès d'eau peu profonde, approchant la zone néritique intérieure, et en biofaciès d'eau profonde, approchant les zones néritique à bathyale extérieures.

[†] Contribution to Frontier Geoscience Program.

INTRODUCTION

Cenozoic strata of the Beaufort-Mackenzie Basin consist almost entirely of terrigenous clastic sediments, most of which were deposited at relatively high depositional rates in deltaic-influenced regimes. Total thickness of the sedimentary pile approaches 12 km or more at the basin depocentre, located in the area of the North Issungnak L-86 well (Fig. 1). The succession consists of 10 regionally recognized depositional sequences (Dietrich et al., 1985; Dixon and Dietrich, 1988) that were documented by seismostratigraphic methods integrated with lithostratigraphic and biostratigraphic observations from many of the 200 or more exploration wells drilled in the basin. A description of the depositional sequences will not be presented here (the reader is referred to Dixon et al., 1985), except to emphasize that the hydrocarbon-rich deltaic deposits of the Upper Paleocene-Middle Eocene Reindeer sequence and the Oligocene Kugmallit sequence are conspicuous components of the section, each with depocentres in excess of three kilometres thick.

It is impossible to understand the biostratigraphy of the Beaufort-Mackenzie Basin without resolving the fundamental differences between chronostratigraphic units and facies-controlled units. This paper attempts to resolve these biostratigraphic differences by establishing a two-fold zonation consisting of interval zones and assemblage zones for Cenozoic rocks of the basin coupled with a preliminary biofacies analysis of the assemblages encountered. This dual approach (time units and biofacies) has been described thoroughly by Ludvigsen et al. (1986).

The zonal scheme comprises 12 interval zones defined by the last appearance datums (LADs) for 12 benthic foraminiferal species which collectively range in age from the Early Paleocene to the Holocene. The interval zonation is complemented by a formal assemblage zonation of seven zones, which are further divided into generalized inner neritic and deeper-water biofacies. The foraminiferal database from which this zonation and biofacies analysis has been drawn consists of approximately 100 wells in the Beaufort Sea and Mackenzie Delta, and Paleocene-Eocene outcrop data from the southwestern and southeastern margins of the basin. The distribution of the zones throughout the Beaufort-Mackenzie Basin will be published separately (McNeil, et al., work in progress).

BACKGROUND

A comprehensive record of Cenozoic Arctic history has been documented in previous biostratigraphic studies of the

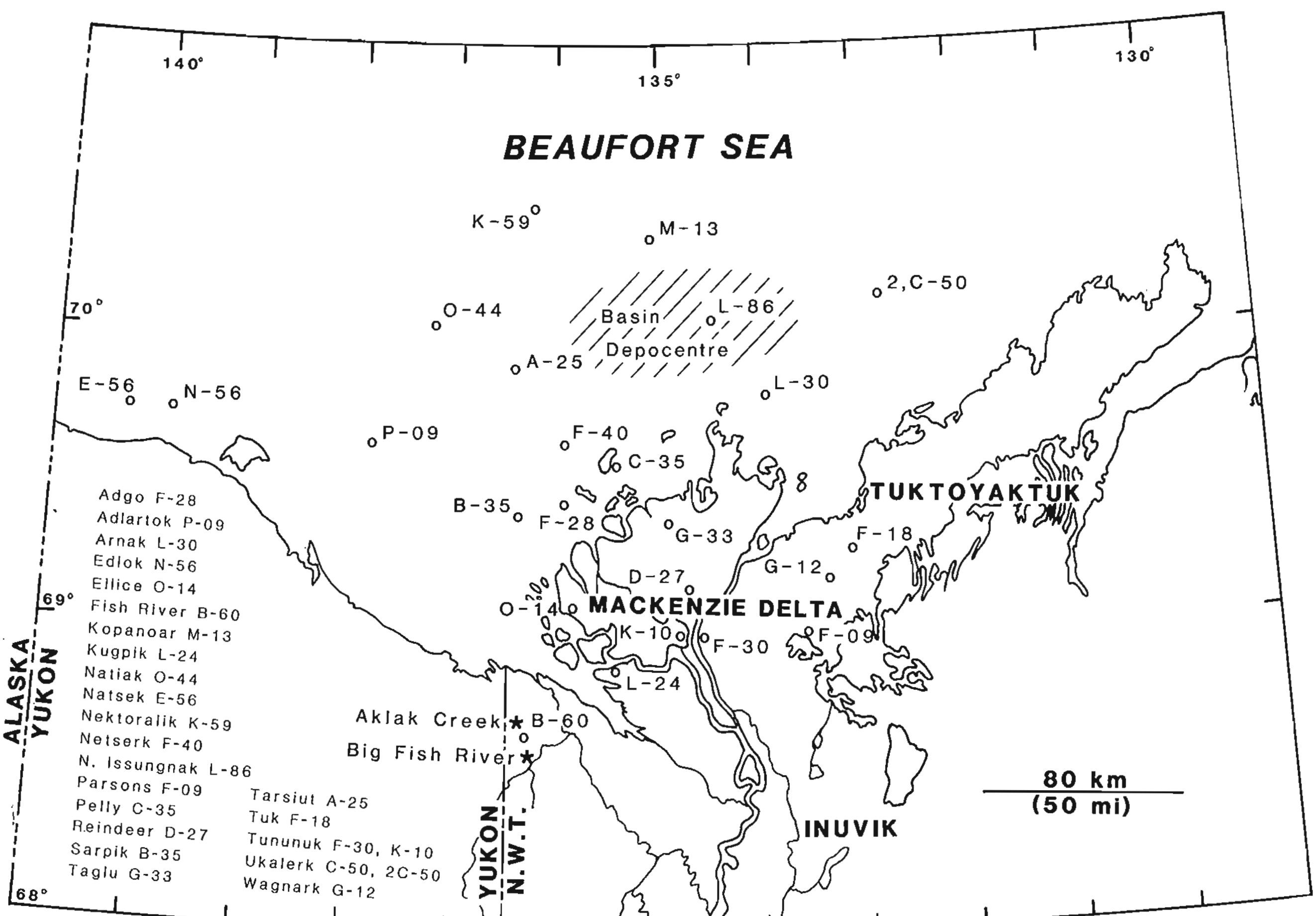


Figure 1. Location of selected exploration wells in the Mackenzie Delta and Beaufort Sea, and Paleocene outcrop localities in the Yukon Coastal Plain. Cretaceous-Cenozoic sedimentary fill in excess of 12 km at basin depocentre.

Beaufort-Mackenzie Basin. Beginning essentially with Chamney (1971) and Petracca (1972), much of the biostratigraphic work focused on foraminifers. Important revisions were achieved by Staplin (1976) as well as Young and McNeil (1984), who provided the first comprehensive treatment of Cenozoic foraminifers in the Mackenzie Delta area. Summaries of numerous previous works on foraminiferal biostratigraphy in the Beaufort-Mackenzie Basin can be found in publications by McNeil (1985, in press).

Equally important, a good deal of attention has been focused on palynomorphs, including pollen, spores, and dinoflagellates. Studies have focused on statistical analyses (Staplin, 1976), specific well studies (Brideaux and Myhr, 1976; Ioannides, in McNeil et al., 1982; Bujak and Davies, in Dixon et al., 1984; Norris, 1986; McIntyre, in Dietrich et al., 1989; and White, 1989), outcrop studies (Doerenkamp et al., 1976; and Ioannides and McIntyre, 1980), regional summaries (Young and McNeil, 1984; and McIntyre, 1985), and climatic interpretation (Norris, 1982).

The Geological Survey of Canada has historically played an important role in establishing the regional stratigraphic and biostratigraphic framework for rocks of the Beaufort-Mackenzie Basin, but an additional impetus for research was provided with the introduction of the Frontier Geoscience Program in 1984. Funds from this program have stimulated many research programs covering the Jurassic to the Holocene. For the Cenozoic, for example, the program has provided resources to establish the foraminiferal database (McNeil, et al., work in progress), which is vital for chronostratigraphic and paleoenvironmental control in the structurally and stratigraphically complex Beaufort-Mackenzie Basin. This regional database forms the basis for the zonal schemes and biofacies analyses presented herein.

THE FORAMINIFERAL SUCCESSION

The Cenozoic foraminiferal record of the Beaufort- Mackenzie Basin is almost entirely benthic; three main phases can be recognized: an initial phase of dominantly agglutinated foraminiferal assemblages in the Paleocene and the Eocene, a subsequent phase of mixed agglutinated-calcareous assemblages in the Oligocene and parts of the Miocene, and a final phase of dominantly calcareous benthic foraminiferal assemblages from the Miocene to Holocene. This overall trend is not unique to the Beaufort-Mackenzie Basin, but is mirrored to a greater or lesser degree in many other high latitude areas such as the North Sea (King, 1983) and the Labrador Shelf (Gradstein and Williams, 1976) and thus reflects widespread similarities in high latitude oceanographic conditions during the Cenozoic.

The specific composition of the Paleocene-Eocene microfaunas is, however,quite unique, and a high degree of endemism is the result of pronounced restrictions between the Arctic regions and the world's major oceans. The paleogeographic reconstruction for this time period by Briggs (1987) indicates a severely restricted Arctic marine basin that McNeil (in press) refers to as the "Arctic Gulf", connected to the southern oceans by the epicontinental Turgai Straits in Siberia and the restricted neritic Greenland-Norwegian Straits in the North Atlantic. Low diversity agglutinated microfaunas characterized by species of *Placentammina* (and other unilocular genera), *Haplophragmoides, Recurvoides, Cyclammina, Reticulophragmium, Trochammina, Portatrochammina, Verneuilina*, and *Verneuilinoides* thrived and evolved during this Arctic Gulf phase. Although the microfaunas are primarily agglutinated, this may be more a reflection of selective preservation than original faunal composition, since species of calcareous benthic foraminifers do occur in places. For example, Wall et al. (1988) have recorded a diverse calcareous microfauna from Paleocene deposits of the Arctic Islands.

The Arctic Gulf phase was terminated at approximately the Eocene/Oligocene boundary when seafloor spreading in the northern North Atlantic provided a sufficient pathway for open circulation between the Arctic and the Atlantic. By the Late Oligocene, deep marine connections were well established between the Arctic and Atlantic oceans (Berggren and Olsson, 1986) and oceanic circulations of a modern aspect became entrenched. Diverse assemblages of both agglutinated and calcareous microfaunas entered the Arctic Ocean at this time and became well established in the Beaufort-Mackenzie area. For example, Dixon et al. (1984) documented agglutinated microfaunas characterized by species of *Bathysiphon, Spirosigmoilinella, Haplophragmoides, Reticulophragmium, Recurvoides, Trochammina*, and *Gravellina* from Oligocene deep-water deposits penetrated by the Kopanoar M-13 well in the Beaufort Sea. In contrast, Dietrich et al. (1989) reported Oligocene assemblages consisting entirely of calcareous species from the neritic facies in the Edlok N-56 well in the western Beaufort Sea. The factors controlling the foraminiferal distributions stem apparently from the influence of the Oligocene Kugmallit Delta, with agglutinated microfaunas being prolific in areas affected by the delta and in deeper waters north of the deltaic depocentre.

From the Miocene to the Holocene, the foraminiferal record for the Beaufort-Mackenzie Basin is dominated by calcareous benthic assemblages, although the Miocene sediments do in places yield mixed assemblages. The Miocene microfaunas are relatively rich and have been documented in part by McNeil et al. (1982) and Young and McNeil (1984). Common genera include *Asterigerina, Cibicidoides, Ehrenbergina, Elphidiella, Melonis, Miliolinella, Pullenia, Sphaeroidina, Trifarina*, and *Valvulineria*. Evidence of pyritic steinkerns of agglutinated genera, such as *Cyclammina*, indicate that the preserved record is probably not truly representative, with preferential preservation of calcareous forms, rather than agglutinated.

Although calcareous benthic microfaunas are characteristic from the Miocene to the Holocene, there was an abrupt and dramatic change in the assemblages near the end of the Miocene, marked by the regionally recognized "Late Miocene unconformity". This event, which McNeil et al. (1982) suggested represented a eustatic lowering of sea level (i.e., the Messinian crisis) left one of the most profound records of change in Arctic Tertiary history. Virtually the entire foraminiferal assemblage changed — diverse, probably relatively warm-water microfaunas of the Miocene were replaced by lower diversity microfaunas reflecting cooler climatic conditions, culminating in an essentially modern Arctic microfauna by the Late Pliocene.

INTERVAL ZONATION

Definition

An interval zone is defined as a body of strata occurring between two biostratigraphic datums, in this case last appearance datums (LADs) of selected benthic foraminifers. The zones presented herein (Fig. 2) are named according to the taxon utilized in drawing the LAD that forms the upper boundary of the zone. Ideally, the interval zone approximates an isochronous unit. In reality, using benthic foraminifers, the zones no doubt contain some degree of diachroneity, but this possiblility is discussed where appropriate (eg., the *Portatrochammina* sp. 2850 Interval Zone). Each zonal boundary is defined by the LAD of a single species. In exceptional cases, an alternate species that has proven to have a more or less identical LAD may be used to define the zonal boundary, where the standard index is obviously anomalously absent. The species utilized in this zonation have been chosen for their widespread distribution, relative abundance, and clearly recognizable last appearance datum. These criteria ensure that the zonation is regionally applicable. Many other useful microfossils that are eliminated will nevertheless provide additional information on paleoenvironments, chronology, and detailed local correlations. Each interval zone is defined by reference to a specific stratotype; relevant stratigraphic data for these stratotypes (mostly in wells) are provided in the Appendix. The presentation of this zonation follows guidelines set forth in the International Stratigraphic Guide (Hedberg, 1976).

Cassidulina reniforme Interval Zone

Stratotype. Adlartok P-09 well, 200 to 380 m below Kelly Bushing (K.B.).

Age. Holocene.

Microfauna typical of the zone. Buccella frigida (Cushman), *Cassidulina reniforme* Nørvang, *C. teretis* Tappan, *Cribroelphidium asklundi* (Brotzen), *C. clavatum* (Cushman), and *Nonion labradoricum* (Dawson).

Discussion. Cassidulina reniforme is extant and the zone is intended to represent strata from the Recent horizon to the LAD for *Cribroelphidium ustulatum.* Because the uppermost sections of hydrocarbon exploration wells were not sampled, the *C. reniforme* Zone is poorly controlled in this study and could potentially contain a diverse modern microfauna. In the stratotype, an unconformity occurs below the zone, and the zone's lower boundary is drawn at the LAD of *Turrilina alsatica.* In a conformable succession, the LAD for *Cribroelphidium ustulatum* would define the lower boundary. In addition to being the zonal index, *C. reniforme* is representative of cold arctic climatic conditions (Sejrup and Guilbault, 1980). It ranges into Pleistocene and Late Pliocene strata, and its detailed distribution is useful for discriminating glacial-interglacial trends.

Cribroelphidium ustulatum Interval Zone

Stratotype. Natiak O-44 well, 260 to 435 m below K.B.

Age. Latest Pliocene to Pleistocene, C. ustulatum LAD >1.6 Ma.

Microfauna typical of zone. Buccella frigida (Cushman), *Cassidulina teretis* Tappan, *Cribroelphidium asklundi* (Brotzen), *C. bartletti* (Cushman), *C. clavatum* (Cushman), *C. excavatum* (Terquem), *C. ustulatum* (Todd), *Elphidiella gorbunovi* (Stschedrina), *E. groenlandica* (Cushman), *E. sibirica* Goës, *Haynesina orbiculare* (d'Orbigny), *Islandiella helenae* Feyling-Hanssen and Buzas, *I. islandica* (Nørvang), *I. norcrossi* (Cushman), and *Stainforthia concava* Höglund.

Discussion. Cribroelphidium ustulatum is widespread in arctic regions, and the zone is well developed in the Beaufort-Mackenzie Basin. Along with the overlying zone, it represents the period of arctic glacial fluctuations. Unfortunately, well cuttings are of limited value in deciphering the details of this dynamic phase of arctic history, so little of the detailed history encompassed by this zone is known.

Cibicides grossus Interval Zone

Stratotype. Natiak O-44, 435 to 943 m.

Age. Early Pliocene to earliest Late Pliocene. Cibicides grossus LAD >2.4 Ma.

Microfauna typical of the zone. Alabamina sp. 750, *Cassidulina teretis* Tappan, *Cibicides grossus* ten Dam and Reinhold, *C. scaldisiensis* ten Dam and Reinhold, *Elphidiella* sp., *Epistominella vitrea* Parker, *Melonis* sp., and miliolids.

Discussion. Cibicides grossus is widespread in the Arctic and in the North Sea and its distribution has been documented by Feyling-Hanssen (1980) and McNeil (in press). Besides being a widespread and useful biostratigraphic marker, *C. grossus* is significant in that its LAD corresponds approximately to the climatic deterioration and onset of continental glaciation (Feyling-Hanssen et al., 1983), which has been dated at 2.4 Ma based on data from DSDP sites in the North Atlantic (Ruddiman and Raymo, 1988). In a conformable sequence, the lower boundary of the zone is the LAD of *Cibicidoides* sp. 800, which approximates the Miocene/Pliocene boundary. The *C. grossus* Interval Zone in the Beaufort-Mackenzie Basin thus includes older strata than that of the partially equivalent *C. grossus* Zone of Greenland and Baffin Island as described by Feyling-Hanssen (1980) and Feyling-Hanssen et al., (1983).

Cibicidoides sp. 800 Interval Zone

Stratotype. Natiak O-44 well, 943 to 1216 m below K.B.

Age. Late Miocene, Cibicidoides sp. 800 LAD >5.3 Ma.

Microfauna typical of the zone. Bulimina elongata d'Orbigny, *Cibicidoides praemundulus* Berggren and Miller, C. sp. 800, C. sp. 782 (previously *Eponides binominatus* Subbotina in Young and McNeil, 1984, Pl. 5, fig. 3 only),

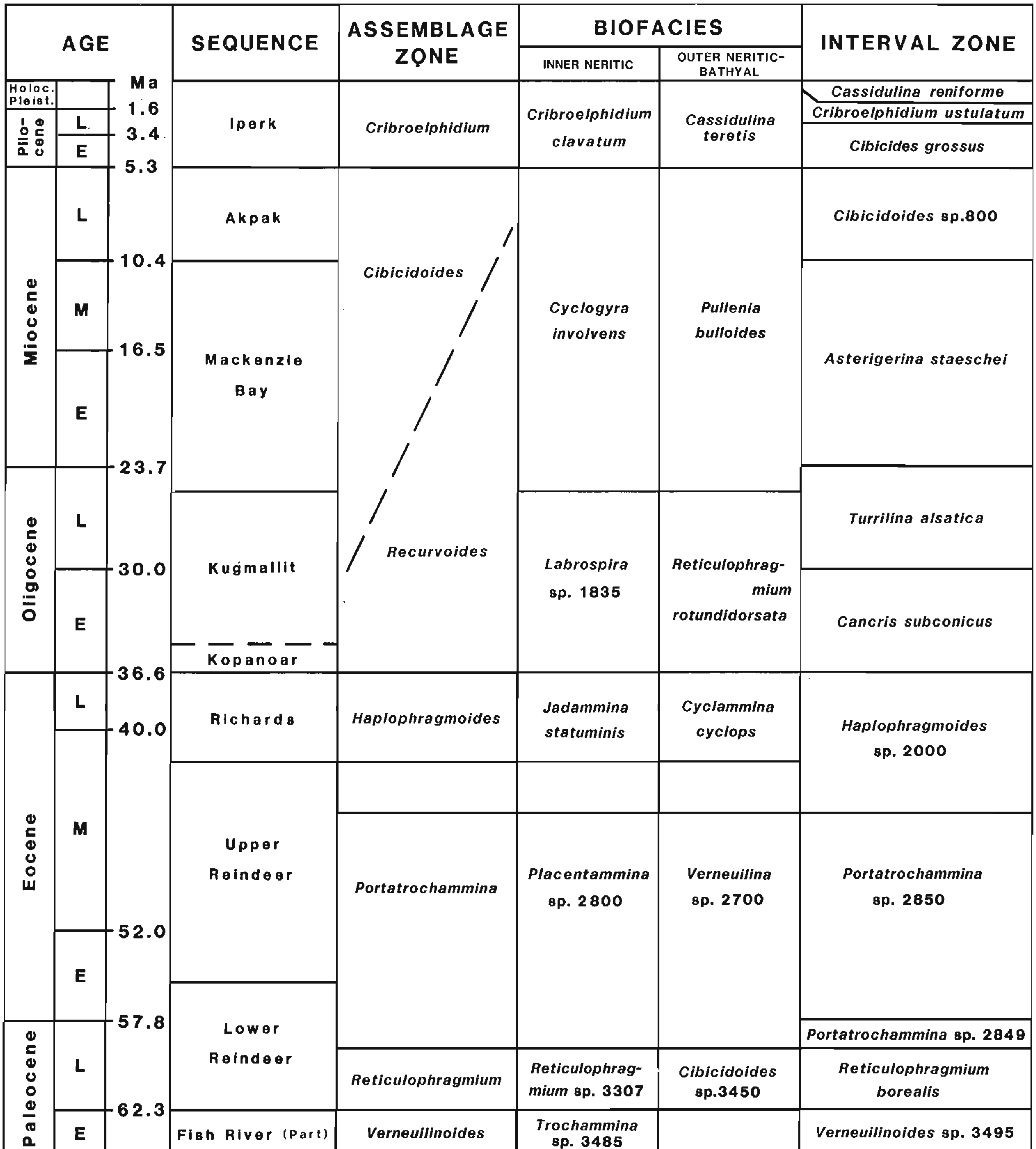


Figure 2. Foraminiferal zonation and biofacies of the Cenozoic sequences of the Beaufort-Mackenzie Basin. Sequences after Dietrich et al. (1985) and Dixon and Dietrich (1988). Assemblage zonation modified from Young and McNeil (1984) and McNeil (1985). Geological time scale from Berggren et al. (1985) and Snelling (1985).

Cribroelphidium sp., *Cyclogyra involvens* (Reuss), *Ehrenbergina praepupa* Spiegler, *Elphidiella* sp., *Globobulimina* sp. 1300, *Globocassidulina* sp., miliolids, *Pullenia bulloides* (d'Orbigny), *P. quinqueloba* (Reuss), *Sphaeroidina bulloides* d'Orbigny, *Trifarina fluens* Todd, unilocular species. In some sections, particularly in deep water, species of the following agglutinated foraminifers are common: *Ammodiscus, Ammomarginulina, Bathysiphon, Cystammina, Haplophragmoides, Recurvoides, Reophax, Spirosigmoilinella*, and *Verneuilinoides*.

Discussion. The *Cibicidoides* sp. 800 Interval Zone is essentially coincident with the Akpak sequence. The upper boundary is marked by the "Late Miocene" unconformity which affects, or in places, eliminates the zone, particularly landward toward the Mackenzie Delta. The lower boundary in a conformable succession is the LAD of *Asterigerina staeschei*.

Cibicidoides sp. 800 is apparently endemic to the Arctic. Its LAD, at approximately 5.3 Ma, is dated by its position relative to other stratigraphic markers and by interpretation of paleoceanographic events. The species occurs below well documented Pliocene deposits of the *Cibicides grossus* Interval Zone and above the well documented LAD of *Asterigerina staeschei* (about 10.4 Ma). The occurrence of *Ehrenbergina praepupa* within the *C.* sp. 800 Interval Zone suggests a Late Miocene age. A minimum age is assessed, following the reasoning of McNeil et al. (1982), who determined that the sub-Iperk unconformity, above which C. sp. 800 does not occur, is interpreted as representing a major eustatic sea level drop (i.e., the Messinian crisis) and is therefore dated at >5.3 Ma.

Asterigerina staeschei Interval Zone

Stratotype. Natiak O-44 well, 1216 to 1867 m below K.B.

Age. Early to Middle Miocene, *Asterigerina staeschei* LAD at approximately >10.4 Ma.

Microfauna typical of the zone. Asterigerina staeschei (Franke), *Cancris* sp. 775, *Chilostomellina* sp., *Cibicidoides* sp. 782 (previously "*Eponides binominatus* Subbotina" of Young and McNeil, 1984, Pl. 5, fig. 3 only), *Cyclogyra involvens* (Reuss), *Ehrenbergina variabilis* Trunkó, *Elphidiella(?) brunnescens* Todd, *E.* sp., *Globobulimina* sp. 1300, *Globocassidulina* sp., *Melonis affine* (Reuss), miliolids, nodosarids, *Pullenia bulloides* (d'Orbigny), *P. quinqueloba* (Reuss), *Robertina declevis* (Reuss), *Sphaeroidina bulloides* d'Orbigny, *Trifarina fluens* Todd, *Valvulineria petrolei* (Andreae). In deltaic or deep-water biofacies, the following agglutinated species may occur commonly: *Ammolagena clavata* (Parker and Jones), *Ammodiscus latus* Grzybowski, *Bathysiphon nodosariaformis* Subbotina, *B. cylindrica* (Glaesner), *Cyclammina* sp., *Cystammina pauciloculata* (Brady), *Eggerella* sp., *Haplophragmoides carinatus* Cushman and Renz, *Martinotiella communis* d'Orbigny, *Reophax pilulifer* Brady, *Reticulophragmium rotundidorsata* (Hantken), *R.* sp., *Spirosigmoilinella* sp., and *Trochammina subvesicularis* Homala and Hanzlíková.

Discussion. Asterigerina staeschei is a widespread and reliable index of Lower to Middle Miocene strata in the Beaufort-Mackenzie Basin. Its LAD corresponds closely to the upper boundary of the Mackenzie Bay sequence (Fig. 3), and it is not known to range below this sequence. Its LAD is confidently placed at approximately the Middle/Late Miocene boundary (10.4 Ma) based on correlation with the foraminiferal succession in the North Sea Basin and adjacent northwestern Europe, as summarized by King (1983).

McNeil (in press) compares the foraminiferal record of the Beaufort-Mackenzie Basin with Wolfe's (1978) evidence of global warm climatic episodes from the Late Oligocene to the Late Miocene with an optimum in the late Early to early Middle Miocene based on leaf records. The presence of *Asterigerina staeschei* substantiates Wolfe's interpretation, as it suggests relatively warm-water conditions during deposition of Mackenzie Bay strata.

Turrilina alsatica Interval Zone

Stratotype. Natiak O-44 well, 1867 to 2560 m below K.B.

Age. Late Oligocene, *Turrilina alsatica* LAD >23.7 Ma.

Microfauna typical of the zone. Anomalinoides sp. 1400, *Charltonina* sp., *Chilostomellina* sp. 790, *Cibicidoides praemundulus* Berggren and Miller, *C. eocaenus* Gümbel, *C. perlucidus* (Nuttall), C. sp., *Cyclogyra involvens* (Reuss), *Elphidiella(?) brunnescens* Todd, *Globobulimina* sp. 1300, *Globocassidulina* sp., *Gyroidina soldanii* d'Orbigny, miliolids, *Melonis affine* (Reuss), *Nuttallides* sp. 1414 [previously "*Eponides binominatus* Subbotina" of Young and McNeil, Pl. 5, fig. 4 only], *Pullenia quinqueloba* (Reuss), *Robertina declevis* (Reuss), *Rotaliatina mexicana* Cushman, *Sphaeroidina bulloides* d'Orbigny, *Trifarina fluens* Todd, and *Valvulineria petrolei* (Andreae). In deltaic and deep-water facies, the following agglutinated foraminifers may dominate the assemblage: *Ammodiscus* spp., *Ammolagena clavata* (Parker and Jones), *Ammomarginulina matchigarica* Voloshinova and Budasheva, *Arenoturrispirillina* sp., *Bathysiphon nodosariaformis* Subbotina, *B. cyclindrica* (Glaesner), *B. pseudoloculus* (Myatliuk), *Cystammina pauciloculata* (Brady), *Eggerella* sp., *Gravellina* sp., *Haplophragmoides carinatus* Cushman and Renz, *H. subtrulisatus* Grzybowski, *Labrospira* sp. 1835, *Psamminopelta* sp. 1530, *Psammosphaera fusca* Schultze, *Recurvoides* spp., *Reophax* spp., *Reticulophragmium rotundisorsata* (Hantken), *R.* spp., *Spirosigmoilinella* sp., *Textularia* sp., *Trochammina(?) subvesicularis* Homala and Hanzlíková, and T. spp.

Discussion. Turrilina alsatica is a widespread and reliable stratigraphic index fossil in the Beaufort-Mackenzie Basin, occurring in the lower Mackenzie Bay sequence and in the underlying Kugmallit sequence. The LAD for *T. alsatica* is reliably dated at >23.7 Ma, based on correlations with well-dated sections in the North Sea and northwestern Europe. The taxonomy and evolution of *T. alsatica* has recently been reviewed by Revets (1987), and King (1983) has summarized its stratigraphic occurrence in the North Sea Basin and neighbouring areas.

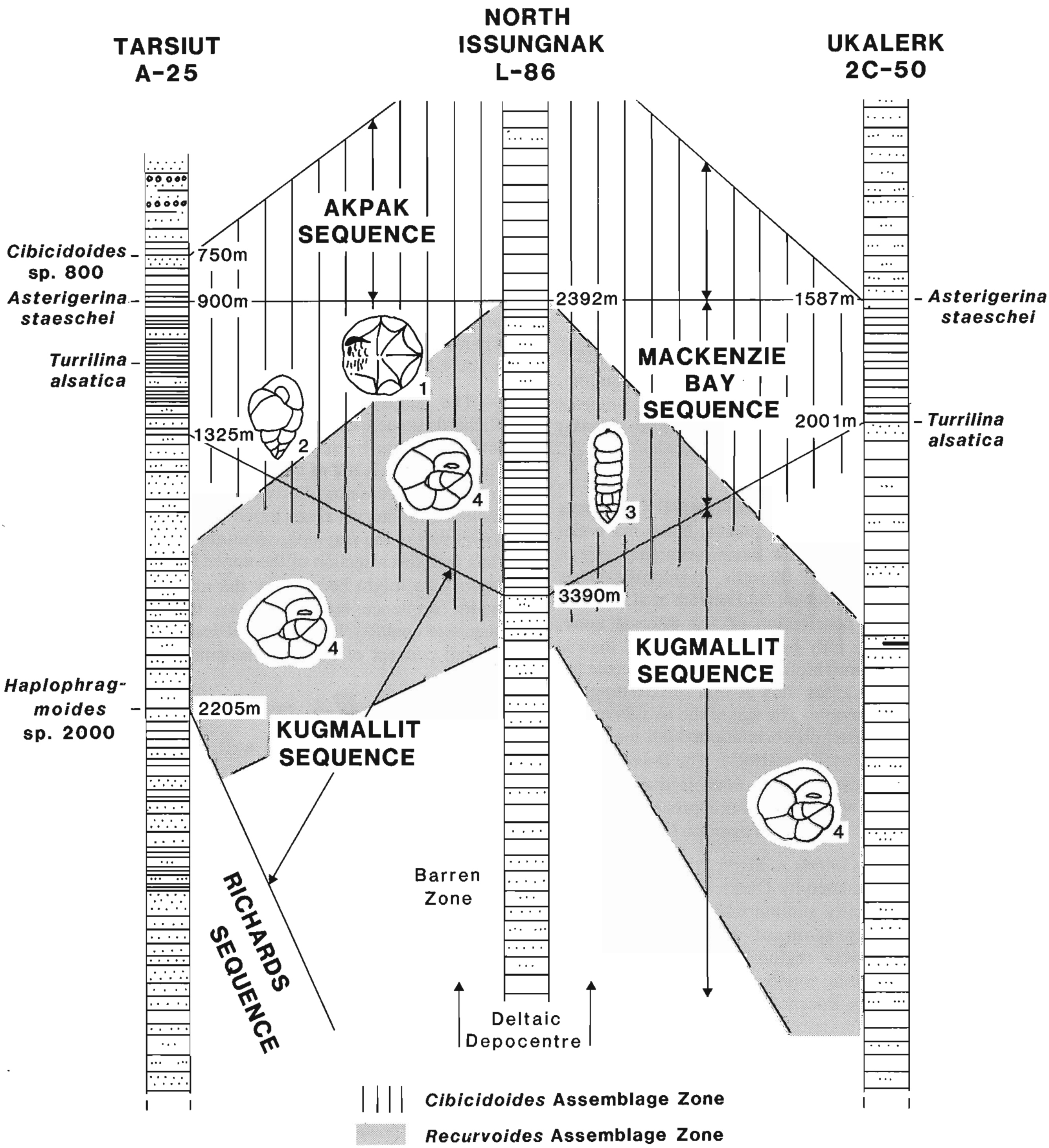


Figure 3. Distribution of the *Recurvoides* and *Cibicidoides* Assemblage Zones in three Beaufort Sea wells (see Figure 1 for locations). Stratigraphic datum is the LAD for *Asterigerina staeschei*. Correlation boundaries of the *Recurvoides* Assemblage Zone are markedly diachronous, rising from the Late Oligocene to the Middle Miocene. LADs of chronostratigraphic markers are indicated on either side of cross-section. Most of the species in the *Recurvoides* Zone have a potential range through the entire zone, but a few, such as *Martinotiella communis* (no. 3), are age diagnostic. Sequence boundaries are indicated in metres below K.B. [1 = *Asterigerina staeschei*, 2 = *Turrilina alsatica*, 3 = *Martinotiella communis*, 4. = *Recurvoides* sp.]

The *T. alsatica* Interval Zone contains a varied foraminiferal assemblage that ranges from the extremes of being calcareous dominant to agglutinated dominant. Agglutinated foraminifers are more dominant in deltaic and in deeper-water facies. Regardless of pronounced paleoenvironmental differences, *T. alsatica* is widely distributed in neritic to bathyal Beaufort-Mackenzie Basin facies, although less abundant in agglutinated dominant biofacies.

Cancris subconicus Interval Zone

Stratotype, Adlartok P-09 well, 440 to 1145 m below K.B.

Age: Early Oligocene, *Cancris subconicus* LAD > 30.0 Ma.

Microfauna typical of the zone. Bulimina sp. 1450, *Brizalina* sp. 1435, *Cancris subconicus* (Terquem), *Chilostomella ovoidea* Reuss, *Elphidiella* sp., *Eoeponidella* sp., *Epistominella(?)* sp., *Melonis* cf. *M. affine* (Reuss) [thick sutured species], miliolids and *Rectobolivina* sp. 1464.

Discussion. The *Cancris subconicus* Interval Zone spans the lower part of the Kugmallit sequence, but is not widespread in the Beaufort-Mackenzie Basin, occurring only in the Edlok N-56 and Adlartok P-09 wells. Its microfauna has been partly illustrated by McNeil (in Dietrich et al., 1989). Despite the restricted distribution of the interval zone, Lower Oligocene strata may be more widespread than is apparent, since pronounced biofacies changes may make the zone unrecognizable in areas such as that surrounding the Kugmallit Delta, for example. The age of the *C. subconicus* LAD is > 30.0 Ma based on correlations with the North Sea Basin, as described by King (1983). The lower boundary of the *C. subconicus* Interval Zone is marked by a regional unconformity, which coincides approximately with the Late Eocene LAD of *Haplophragmoides* sp. 2000.

The significance of *Cancris subconicus* in the Beaufort-Mackenzie Basin is discussed by Dietrich et al. (1989) and McNeil (in press). Briefly summarized, it represents the Early Oligocene migration of calcareous benthic foraminifers into the Arctic regions following the opening of the Arctic-North Atlantic connections by seafloor spreading, which began in the Eocene.

Haplophragmoides sp. 2000 Interval Zone

Stratotype. Netserk F-40 well, 2468 to 3772 m. [Stratotype is arbitrarily truncated at sequence boundary. The zone's lower boundary at LAD of *Portatrochammina* sp. 2850 is not present in Netserk F-40.]

Age. Latest Middle(?) to Late Eocene. *Haplophragmoides* sp. 2000 LAD approximately >36.6 Ma.

Microfauna typical of the zone. Ammodiscus sp., *Ammomarginulina matchigarica* Voloshinova and Budasheva, *A.* sp. 1510, *Bathysiphon pseudoloculus* (Myatliuk), *Cyclammina cyclops* McNeil, *Gravellina* sp., *Haplophragmoides carinatus* Cushman and Renz, *H. subtrulisatus* Grzybowski, *Jadammina statuminis* McNeil, *Recurvoides* sp., *Trochammina(?) subvesicularis* Homala and Hanzlíková, and *Trochammina* spp.

Discussion. The LAD for *Haplophragmoides* sp. 2000 is truncated at the regional unconformity (sea level lowstand) separating the Richards and Kugmallit sequences. The unconformity is crossed by some agglutinated foraminifers (species of *Bathysiphon, Haplophragmoides, Recurvoides,* and *Trochammina*), but *H.* sp. 2000 and the distinctive *C. cyclops* and *J. statuminis* have LADs within the Richards sequence and *H.* sp. 2000 Interval Zone.

Because *H.* sp. 2000 is endemic to the Arctic, the age of its LAD is not well constrained. The lower part of the zone contains dinoflagellates such as *Wetzeliella*, which have been dated as Early to Middle Eocene by McIntyre (1985). *Haplophragmoides* sp. 2000 thus occurs between probable Middle Eocene and Lower Oligocene strata. Therefore, its range is approximately the Late Eocene with a LAD >36.6 Ma.

The sedimentary succession from the base of the Richards sequence to the Paleocene-Maastrichtian boundary is characterized by major alternations of marine and nonmarine strata and is not as amenable to interval zonation as the overlying succession. Barren nonmarine sections are thus included in interval zones based on marine organisms. The lower part of the *Haplophragmoides* sp. 2000 Interval Zone thus includes a section of the upper Reindeer sequence. An alternative might be to draw the interval boundary at the major sequence boundary (i.e., the Reindeer-Richards sequence contact), but this would destroy the objectivity and overall concept of the zonal scheme.

Portatrochammina sp. 2850 Interval Zone

Stratotype. Natsek E-56 well, 1249 to 2560 m above K.B.

Age. Early to mid-Middle Eocene, *Portatrochammina* sp. 2850 LAD probably >45.0 Ma.

Microfauna typical of the zone. Haplophragmoides sp., *Lagenammina* sp., *Pelosina* sp., *Portatrochammina* sp. 2850, *Recurvoides* sp., *Reophax sp., Saccammina* sp., *Trochammina* sp., *Verneuilina* sp., and *Verneuilinoides* sp.

Discussion. Portatrochammina sp. 2850 is distributed in marginal marine and other neritic marine environments. In the stratotype, a fairly diverse assemblage of agglutinated foraminifers suggests a neritic, perhaps mid-shelf paleoenvironment. In other sections, *P.* sp. 2850 occurs in a marginal marine biofacies typically associated with *Placentammina* sp. 2800. Given the complexities of upper Reindeer deltaic sedimentation and the biofacies developed, the LAD for *Portatrochammina* sp. 2850 is markedly diachronous, as illustrated in Figure 4.

In Natsek E-56, the LAD for *Portatrochammina* sp. 2850 occurs just below or possibly overlaps the lower part of a section rich in dinoflagellates (e.g., *Wetzeliella, Kisselovia, Spinidinium,* and numerous peridinioid spp.), which McIntyre (in Dietrich et al., 1989) considered to be of Early to Middle Eocene age. The age of the *P.* sp. 2850 LAD has been determined within this framework and is therefore >45 Ma.

The lower boundary of the *P.* sp. 2850 Interval Zone is drawn at the LAD of *Portatrochammina* sp. 2849, which occurs in latest Paleocene strata. As defined in the stratotype, the lower part of the zone spans the sequence boundary between upper and lower Reindeer sequences. The lowest part of the zone is barren of in situ foraminifers but yields the palynomorph *Platycarya*, which McIntyre (in Dietrich et al., 1989) dates as earliest Eocene.

Portatrochammina sp. 2849 Interval Zone

Stratotype. Outcrop sections within the Aklak Member of the Reindeer Formation, Aklak Creek, Yukon Coastal Plain (Fig. 4).

Age. Late Late Paleocene, *Portatrochammina* sp. 2849 LAD >57.8 Ma. Microfauna typical of the zone: *Portatrochammina* sp. 2849.

Discussion. *Portatrochammina* sp. 2849 has been recovered only from marginal marine sediments of the Aklak Member of the Reindeer Formation. Its stratotype includes a series of outcrops through to the base of the Aklak Member and its lower contact is drawn at the LAD for *Reticulophragmium borealis*, which occurs in the Ministicoog Member of the Moose Channel Formation. *Portatrochammina* sp. 2849 occurs in the Natsek E-56 well between 2560 and 2720 m, but the section there is ambiguous as *R. borealis* has been recovered rarely from both above and below *P.* sp. 2849. Dietrich et al. (1989) considered the Natsek occurrences of *R. borealis* to be reworked since they occurred in atypical associations.

The age of the *P.* sp. 2849 LAD is determined to be >57.8 Ma, based on the palynological determination of the Paleocene-Eocene boundary above the *P.* sp. 2849 LAD in outcrop sections of the Aklak Member on Aklak Creek (A.R. Sweet, pers. comm.).

The *P.* sp. 2849 Interval Zone is of limited extent in the Beaufort-Mackenzie Basin and further data on its distribution will be necessary to test its utility.

Reticulophragmium borealis Interval Zone

Stratotype. Adlartok P-09 well, 1505 to 2647 m (T.D.) below K.B.

Age. Early Late Paleocene, *Reticulophragmium borealis* LAD >60 Ma.

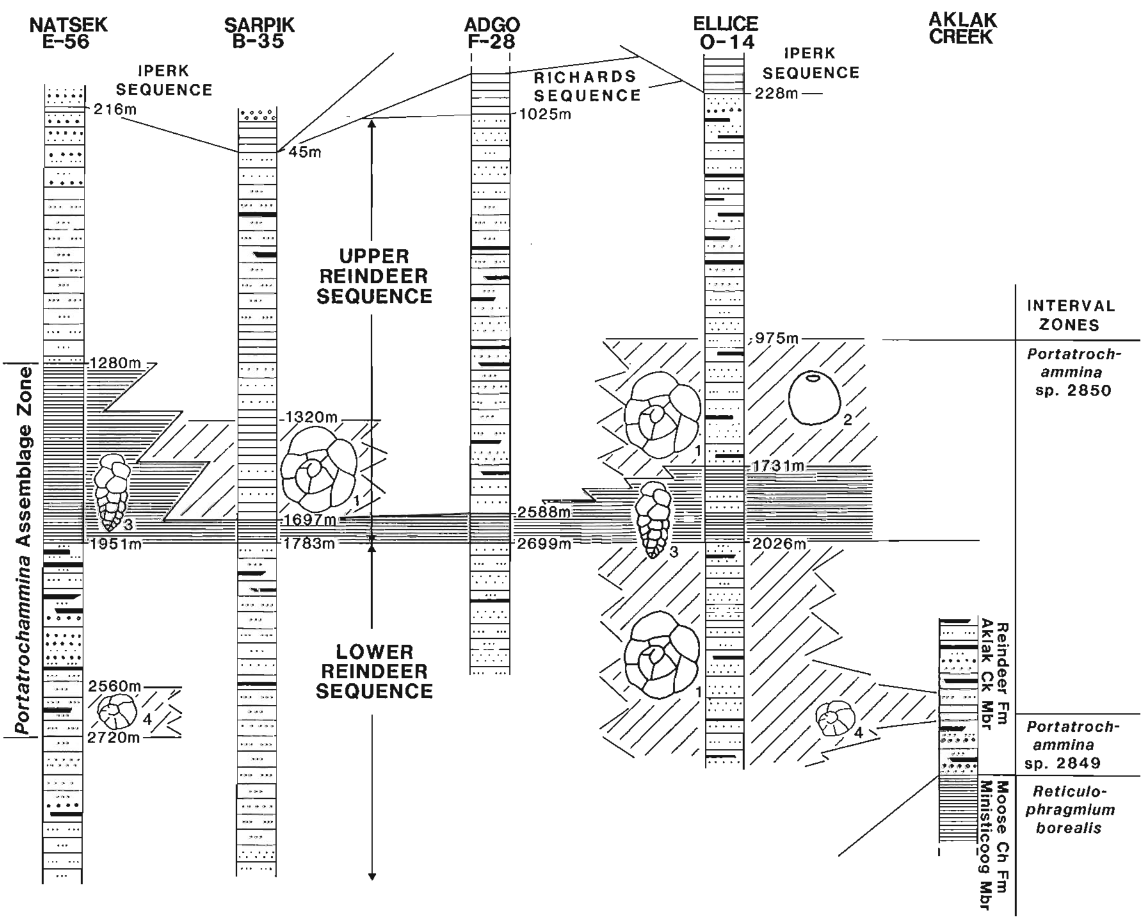


Figure 4. Distribution of *Portatrochammina* Assemblage Zone, and its biofacies, relative to the interval zonation in selected wells and outcrop strata on Aklak Creek, Yukon Coastal Plain (see Figure 1 for *locations*). Sequence boundaries are indicated in metres below K.B. [1 = *Portatrochammina* sp. *2850*, 2 = *Placentammina* sp. *2800*, 3 = *Verneuilina* sp. *2700*, 4 = *Portatrochammina* sp. *2849*]

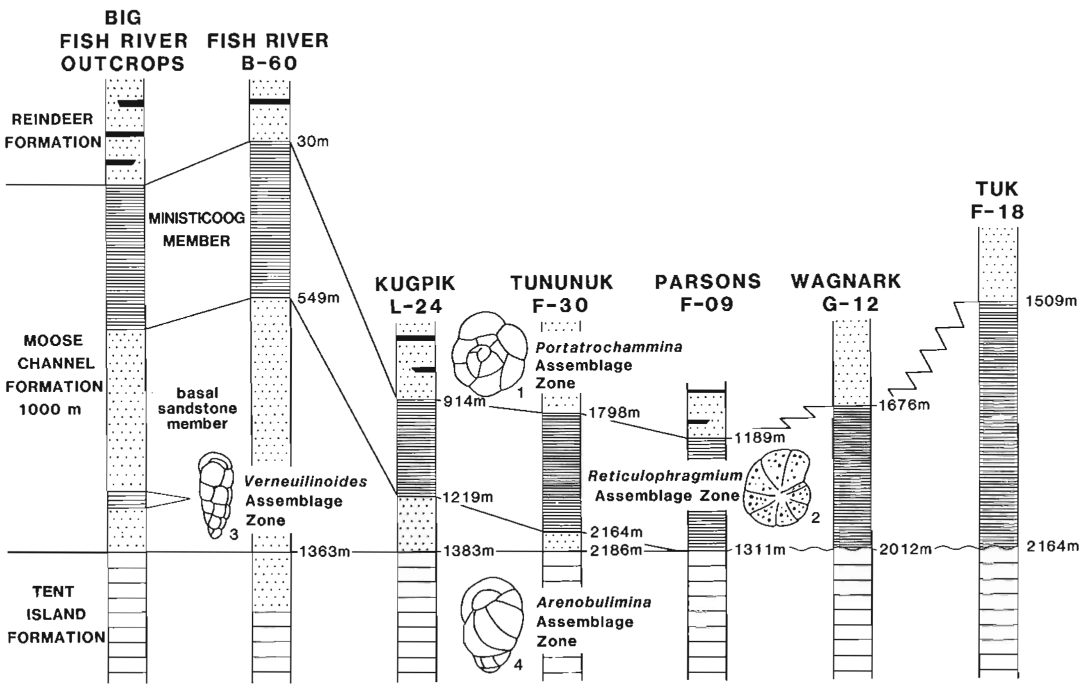


Figure 5. Distribution of *Reticulophragmium* and *Verneuilinoides* Assemblage Zones, illustrating diachronous rise of *Reticulophragmium* Zone in area of Tuk Peninsula (see Figure 1 for locations). LAD of *Reticulophragmium borealis* is approximately coincident with upper boundary of Ministicoog Member. Lithostratigraphic boundaries are indicated in metres below K.B. [1 = *Portatrochammina* sp. 2850, 2 = *Reticulophragmium borealis*, 3 = *Verneuilinoides* sp. 3495, 4 = *Arenobulimina* sp.]

Microfauna typical of the zone. *Arenoturrispirillina* sp. 3218, *Cibicidoides* sp. 3450, *Labrospira* sp. 3375, *Praebulimina* sp., *Recurvoides* sp., *Reophax* sp., *Reticulophragmium borealis, R.* spp., *Placentammina* spp., *Trochammina* spp., *Verneuilina* sp., and *Verneuilinoides* sp.

Discussion. The *Reticulophragmium borealis* Interval Zone is widespread and readily mappable in mid-Paleocene strata of the Beaufort-Mackenzie Basin. It occurs in outcrop on both sides of the Mackenzie Delta and is well documented in the subsurface of the outer Mackenzie Delta (Fig. 5). As in other zones based on endemic agglutinated foraminifers representative of the early Tertiary Arctic Gulfian phase, an age assignment is based on the associated palynological record (the Ministicoog Member being dated as mid-Paleocene by A.R. Sweet, pers. comm.)

As illustrated in Figure 5, the LAD for *R. borealis* is almost certainly diachronous, rising stratigraphically toward the northeastern area of the basin. In the more or less conformable sequence exposed in the Yukon Coastal Plain, the lower boundary of the zone is drawn at the LAD of *Verneuilinoides* sp. 3495. In much of the Mackenzie Delta area, the lower boundary occurs at the Cretaceous-Tertiary unconformity delineated by the Maastrichtian LAD of *Arenobulimina* sp. (Fig. 5).

The *R. borealis* Interval Zone represents a widespread and significant marine transgressive event in the Arctic, and equivalents of the zone have recently been documented from the Arctic Islands by Wall et al. (1988).

Verneuilinoides sp. 3495 Interval Zone

Stratotype. Outcrop sections of the basal sandstone member of the Moose Channel Formation on Big Fish River, Yukon Coastal Plain (Fig. 5).

Age. Early Paleocene, *Verneuilinoides* sp. 3495 LAD approximately >62.3 Ma.

Microfauna typical of the zone. *Placentammina* sp., *Trochammina* sp. 3485, and *Verneuilinoides* sp. 3495.

Discussion. The *Verneuilinoides* sp. 3495 Interval Zone has a distribution limited to the Big Fish River and Eagle Creek areas of the Yukon Coastal Plain. Throughout this area, the zone represents a low diversity marginal marine biofacies. Deeper water equivalents are unknown. In the subsurface of the Mackenzie Delta, the zone is absent due to an unconformity.

An Early Paleocene age is assigned to the *V.* sp. 3495 Interval Zone, based on associated palynomorphs. The lower boundary of the zone is drawn at the LAD of *Arenoturrispirillina* sp. which occurs at the top of the Maastrichtian Tent Island Formation. A.R. Sweet (pers. comm.) has indicated that the Cretaceous-Tertiary boundary in the Yukon Coastal Plain is drawn approximately at the Tent Island-Moose Channel Formation contact, based on evolutionary trends primarily in the angiosperms *Aquilapollenites* and *Wodehouseia*.

DISCUSSION OF INTERVAL ZONATION

Benthic foraminifers from the Cenozoic of the Beaufort-Mackenzie Basin provide an overall level of precision that ranges from 0.8 Ma at best to 13.0 Ma at worst. Surprisingly, the interval zone of longest duration, that of *Asterigerina staeschei* (Early to Middle Miocene), is relatively fossiliferous. Presumably, it represents a stable oceanographic episode during which there was little successional faunal change. Detailed studies of sections such as this, however, will be of priority in further refinements to the scheme. A feature of interval zonation is that additional zones can be added without disrupting the scheme.

The question of diachroneity is important but difficult to resolve precisely because a zone cannot measure its own degree of diachroneity independently. The zonal indices utilized in this study have all been scrutinized against the regional depositional seismic sequence framework of Dietrich et al. (1985) and this has provided the practical proof that the zonation is reliable basinally; i.e., that interval zone boundaries or LAD horizons do not cross the major sequence boundaries. So, the amount of diachroneity is probably insignificant in terms of calibrating the major depositional sequences. In general, however, the calcareous benthic LAD horizons of the Oligocene to Holocene are more precise than their agglutinated counterparts in the Paleocene and Eocene. This appears to result at least partly from a difference in the facies sampled; i.e., the Paleocene-Eocene sections represent inner shelf to terrestrial facies where migrating shorelines and deltaic shifting leads to diachroneity in the associated fossil record. Ideally, the interval zonation would be established in the fossiliferous, relatively stable biofacies of the outer neritic to upper bathyal zones.

The proposed interval zonation for the Beaufort-Mackenzie compares favourably with similar zonations of the Cenozoic at high latitude. For example, King (1983) recognized 17 benthic foraminiferal zones in the Cenozoic of the North Sea Basin and Gradstein et al. (1988) recognized 10, based mostly on benthic foraminifers. In the Labrador Shelf and north Grand Banks, Gradstein and Agterberg (1982) recognized eight zones based mostly on benthic foraminifers.

ASSEMBLAGE ZONATION

Definition

An assemblage zone is defined as a body of strata that is distinguished by a discrete association of fossils. The association, or assemblage, is the fossil representation of a natural biocoenosis or its thanatocoenosis and is intimately connected with of the strata's depositional history. The assemblage zone can usually be divided into biofacies that represent, at least partly time equivalent subassemblages. The scheme of assemblages and preliminary biofacies utilized in this study has been described in detail by Ludvigsen et al. (1986). The presentation of the assemblage zonation follows the guidelines set forth in the International Stratigraphic Guide (Hedberg, 1976).

The assemblage zones described herein (Fig. 2; Plates 3, 4) are purposely broad in scope. They catagorize the major changes in the foraminiferal microfaunas through the Cenozoic in the Beaufort-Mackenzie Basin and provide a framework for biofacies analysis. Most of the assemblage zones can readily be divided into two components — a "shallow-water" biofacies and a "deep-water" biofacies. The biofacies (Fig. 2; Plates 3, 4) are presented here on a preliminary basis. Further study of biofacies and a more thorough documentation are currently in progress. The words "shallow" and "deep" are used qualitatively and refer in a general way to inner neritic and outer neritic-bathyal environments. The inner neritic microfaunas are distinctive because of variable, coastally influenced water (e.g., low salinity) and the deeper biofacies represent substrates covered by more uniform oceanic water masses.

Cribroelphidium Assemblage Zone

Stratotype. Natiak O-44 well, 210 to 943 m below K.B.

Age. Pliocene to Holocene.

Cribroelphidium clavatum biofacies (shallow). *Buccella frigida* Cushman, *Cribroelphidium clavatum* (Cushman), *C. ustulatum* (Todd), and miliolids.

Cassidulina teretis biofacies (deep). *Buccella frigida* Cushman, *Cassidulina reniforme* Nørvang, *C. teretis* Tappan, *Cibicides grossus* ten Dam and Reinhold, *Cribroelphidium* spp., *Elphidiella groenlandica* (Cushman), *E. sibirica* (Goës), glandulinids, *Haynesina orbiculare* (d'Orbigny), *Islandiella helenae* Feyling-Hanssen and Buzas, *I. islandica* (Nørvang), *Melonis* sp., *Nonion labradoricum* (Dawson), polymorphinids, and *Stainforthia concava* Höglund.

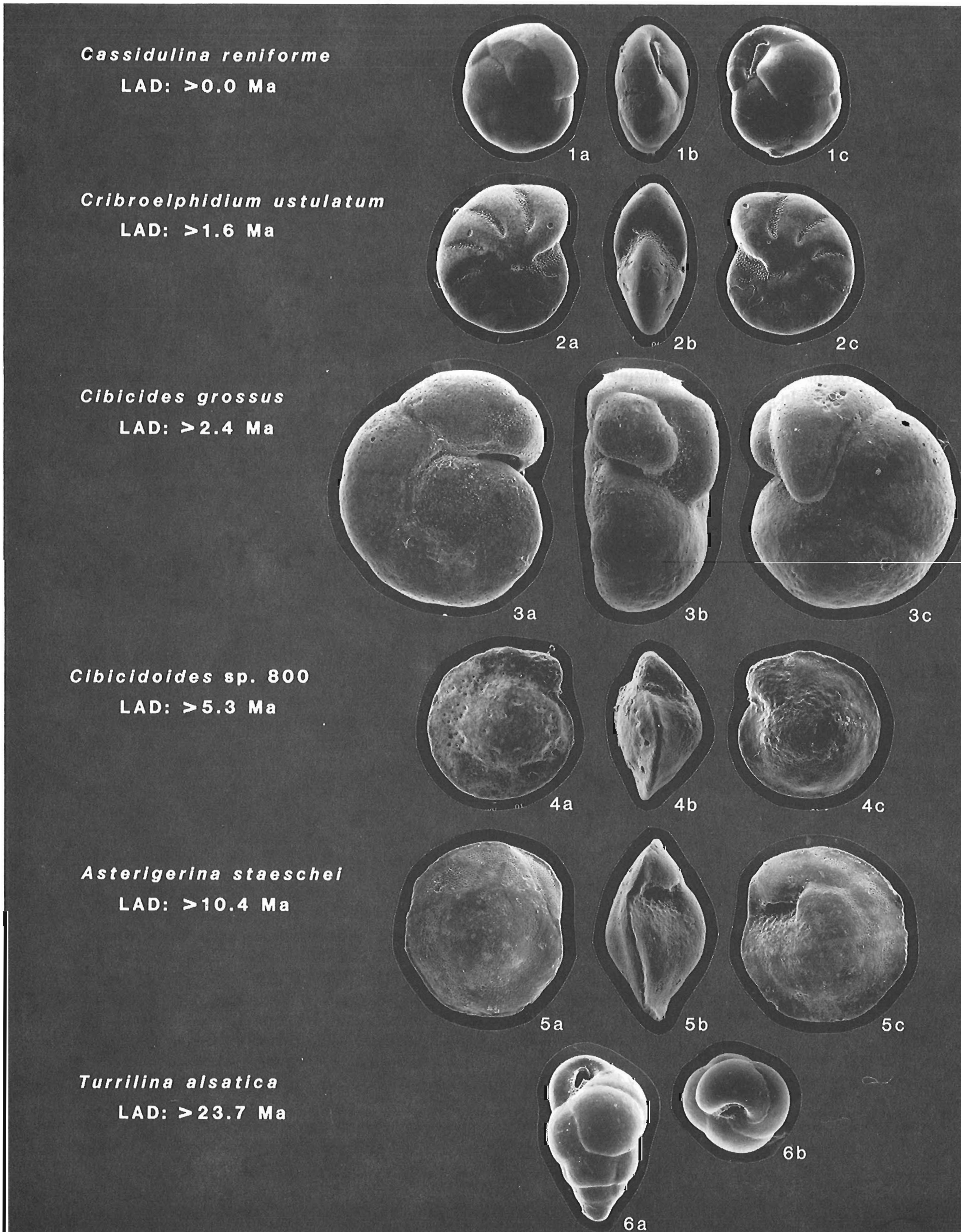
Discussion. The *Cribroelphidium* Assemblage Zone is widely distributed in the Iperk sequence. It consists almost entirely of calcareous benthic foraminifers; agglutinated specimens or the planktonic "*Globigerina*" *pachyderma* (Ehrenberg) are rare.

The shallow-water biofacies of the zone was described as the *Elphidium* assemblage by Young and McNeil (1984) in their study of the Cenozoic of the Mackenzie Delta. Subsequent drilling in the Beaufort Sea has exposed the assemblage from a spectrum of neritic to bathyal environments.

The *Cribroelphidium* assemblage is the only assemblage in the Beaufort Cenozoic section that has a direct analogue in the modern marine environment of the area. Vilks et al. (1979) have studied the modern distribution of the assemblage in the Beaufort Sea, recognizing a *C. clavatum*-dominant biofacies in low-salinity environments and a *C. teretis*-dominant biofacies in normal marine salinity environments. The distributional data of Vilks et al. (op. cit.) indicated that the dividing zone between these two biofacies occurred at a water depth of 50 to 100 m in areas directly north of the Mackenzie Delta, but that salinity, not water depth, was the controlling factor. Hence, in areas not affected by the Mackenzie Delta, these depth boundaries might be distinctly different.

PLATES 1-4

Illustrated specimens in Plates 1 to 4 are stored in the type collections of the Geological Survey of Canada, Ottawa. All specimens derived from exploration wells come from well cuttings. The author's numerical open nomenclature is used for species that have not been named.



Assemblages equivalent to the *Cribroelphidium* assemblage are known from the circum-Arctic region and are well described in works by Feyling-Hanssen (1980 and 1986) and Feyling-Hanssen et al. (1983).

Cibicidoides Assemblage Zone

Stratotype. Natiak O-44 well, 943 to 1531 m below K.B.

Age. Oligocene to Miocene.

Cyclogyra involvens biofacies (shallow). *Asterigerina staeschei* (Franke), *Cyclogyra involvens* (Reuss), *Elphidiella(?) brunnescens* Todd, *Miliolinella* sp., *Trifarina fluens* (Todd), and *Turrilina alsatica* Andreae.

Pullenia bulloides biofacies (deep). *Asterigerina staeschei* (Franke), *Brizalina* sp. 1435, *Bulimina* elongata d'Orbigny, *B.* sp., *Cancris subconicus* (Terquem), *Chilostomella ovoidea* Reuss, *Chilostomellina* sp. 790, *Cibicidoides eocaenus* (Gümbel), *C. perlucidus* (Nuttall), *C. preamundulus* Berggren and Miller, *C.* sp. 800, *C.* sp., *Cyclogyra involvens* (Reuss), *Ehrenbergina praepupa* Spiegler, *E. variabilis* Trunkó, *Elphidiella(?) brunnescens* Todd, *Globocassidulina* sp., *Globobulimina* sp. 1300, *Gyroidina soldanii* d'Orbigny, *Hoeglundina elegans* (d'Orbigny), *Melonis affine* (Reuss), *Melonis* sp., miliolids, nodosarids, *Nuttallides* sp., *Pullenia bulloides* (d'Orbigny), *Rotaliatina mexicana* Cushman, *Sphaeroidina bulloides* d'Orbigny, *Trifarina fluens* Todd, *Turrilina alsatica* Andreae, unilocular species, and *Valvulineria petrolei* (Andreae).

Discussion. The *Cibicidoides* Assemblage Zone contains what was previously referred to as the *Cibicides* spp. assemblage by Young and McNeil (1984), who recognized the microfauna in the Mackenzie Bay and Kugmallit formations of the Mackenzie Delta. Further drilling in the Beaufort Sea has documented a widespread distribution of the assemblage primarily in the Akpak and Mackenzie Bay sequences.

The diverse microfauna of the *Cibicidoides* assemblage no doubt has potential for much finer biofacies discrimination, but such a study is beyond the scope of this report. By way of comparison, Culver (1988) has recently completed a thorough depth distribution analysis of foraminifers in the northwestern Gulf of Mexico, recognizing 12 zones from nearshore to abyssal depths. Despite the tenuous nature of comparisons between such widely separated regions, the generic distribution may be useful for biofacies anlysis in the Beaufort-Mackenzie Basin. Culver's data, supplemented with previous information provided by Albers et al. (1966), Boltovskoy and Wright (1976), and Murray (1973) indicate that the deep-water indices of the *Cibicidoides* assemblage should include *Pullenia, Chilostomella, Hoeglundina, Sphaeroidina, Valvulineria,* and *Gyroidina.* Detailed documentation of species distribution will be necessary to confirm these generalizations, for it is well established that genera may have markedly different ecological ranges in different regions. For example, the generally deep-water foraminifer, *Cyclammina,* inhabits progressively shallower water at higher latitudes along the continental slope and shelf of North America, apparently following the rise of the permanent thermocline (McNeil, 1988). Furthermore, the closely related genus, *Reticulophragmium,* has been recovered abundantly in the inner neritic facies in the Beaufort-Mackenzie Basin.

Recurvoides Assemblage Zone

Stratotype. Natiak O-44 well, 1531 to 2560 m below K.B.

Age. Oligocene to Miocene.

Labrospira sp. 1835 biofacies (shallow). Labrospira sp. 1835 and *Textularia* sp.

Reticulophragmium rotundidorsata biofacies (deep). *Ammodiscus latus* Grzybowski, *A.* sp., *Ammolagena clavata* (Parker and Jones), *Ammomarginulina matchigarica* Voloshinova and Budasheva, *A.* sp., *Arenoturrispirillina* sp., *Bathysiphon cylindrica* (Glaesner), *B. nodosariaformis* Subbotina, *B. pseudoloculus* (Myatliuk), *Cyclammina* sp., *Cystammina pauciloculata* (Brady), *Eggerella* sp., *Glomospira* sp., *Gravellina* sp., *Haplophragmoides carinatus* Cushman and Renz, *H. subtrullisatus* Grzybowski, *Hyperammina* sp., *Martinotiella communis* (d'Orbigny), *Psamminopelta* sp., *Psammosphaera fusca* Schultze, *Recurvoides* spp., *Reophax nodulosus* Brady, *R. pilulifer* Brady, *Reticulophragmium rotundidorsata* (Hantken), *R.* spp., *Spirosigmoilinella* sp., and *Trochammina(?) subvesicularis* Homala and Hanzlíková, and *T.* spp.

Discussion. The *Recurvoides* Assemblage Zone equates to strata yielding the *Recurvoides* assemblage, as described by McNeil (1985). The assemblage is absent in marginal marine to terrestrial Oligocene-Miocene sediments in the Mackenzie Delta and so was not recognized in an earlier study by Young and McNeil (1984). The *Recurvoides* Assemblage Zone is, however, widespread in neritic and bathyal sections beneath the Beaufort Sea and has been illustrated from bathyal sediments of the Kopanoar M-13 well (Dixon et al., 1984).

PLATE 1

Figure 1, a-c *Cassidulina reniforme* Nørvang, x 103, Adlartok P-09, 215-227 m, GSC 89543.

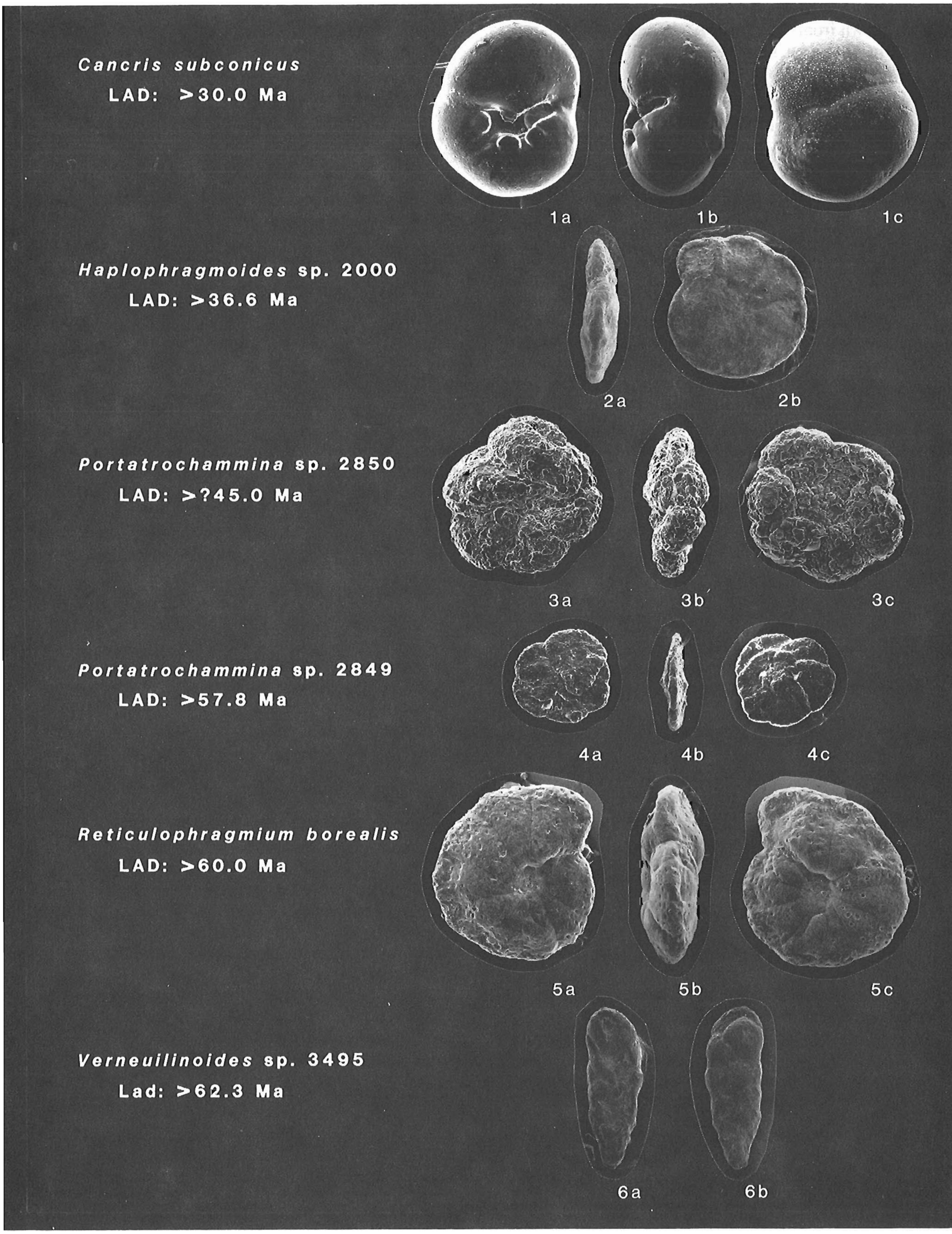
Figure 2, a-c *Cribroelphidium ustulatum* (Todd), x 72, Natiak O-44, 435-455 m, GSC 89544.

Figure 3, a-c *Cibicides grossus* ten Dam and Reinhold, x 48, from unnamed Pliocene strata, GSC Locality O-48222, White Point, Ellesmere Island, 81°05′48″N, 89°58′W, GSC 89536.

Figure 4, a-c *Cibicidoides* sp. 800, x 68, Natiak O-44, 1069-1087 m, GSC 89545.

Figure 5, a-c *Asterigerina staeschei* (Franke), x 55, Natiak O-44, 1489-1507 m, GSC 89538.

Figure 6, a, b *Turrilina alsatica* Andreae, x 59, Nuwok Member, Sagavanirktok Formation, Carter Creek, Alaska, 144°39′42″N, 69°56′45″W, GSC 89540.


Cancris subconicus
LAD: >30.0 Ma
1a
1b
1c
Haplophragmoides sp. 2000
LAD: >36.6 Ma
2a
2b
Portatrochammina sp. 2850
LAD: >?45.0 Ma
3a
3b
3c
Portatrochammina sp. 2849
LAD: >57.8 Ma
4a
4b
4c
Reticulophragmium borealis
LAD: >60.0 Ma
5a
5b
5c
Verneuilinoides sp. 3495
Lad: >62.3 Ma
6a
6b

The assemblage is fairly diverse and abundant in its deeper-water biofacies, but species diversity decreases in the inner neritic zone, where *Labrospira* sp. 1835 (Pl. 4) and *Textularia* sp. distinguish a shallow-water assemblage, which disappears into either low salinity marginal marine facies or terrestrial facies.

The boundaries of the *Recurvoides* Assemblage Zone are markedly diachronous, as the assemblage shares a complex relationship with the calcareous benthic *Cibicidoides* assemblage (Fig. 3). Arbitrarily, the *Recurvoides* Assemblage Zone is applied in sections containing mixed *Recurvoides-Cibicidoides* assemblages, because of the difficulty in reliably separating the assemblages in well cuttings. Alternatively, sections with mixed microfaunas could be assigned to an undivided "*Cibicidoides-Recurvoides* Assemblage Zone". The assemblage is well developed in the area of the Oligocene Kugmallit deltaic depocentre and in Oligocene-Miocene deeper-water facies. It appears in sediments as young as the Upper Miocene Akpak sequence, or, in contrast, may be absent in Miocene and Oligocene sections, for example in the Edlok N-56 well, where Oligocene sediments are devoid of agglutinated foraminifers and the section yields only the *Cibicidoides* microfauna. Precise distributions of these microfaunas are not presented here, but work is in progress on documenting agglutinated foraminifers of the *Recurvoides* Assemblage Zone vis à vis the Kugmallit Delta.

Haplophragmoides Assemblage Zone

Stratotype. Arnak L-30 well, 11,280 to 14,840 ft. (T.D.) (3438-4523 m) below K.B.

Age. Late Middle(?) to Late Eocene.

Jadammina statuminis biofacies (shallow). *Jadammina statuminis* McNeil.

Cyclammina cyclops biofacies (deep). *Ammomarginulina* sp., *Bathysiphon nodosariaformis* Subbotina, *Bathysiphon pseudoloculus* (Myatliuk), *Cyclammina cyclops* McNeil, *Gravellina* sp., *Haplophragmoides carinatus* Cushman and Renz, *Haplophragmoides* sp. 2000, *Recurvoides* spp., *Textularia* sp., *Trochammina* sp., and *Trochamminoides* sp.

Discussion. The *Haplophragmoides* Assemblage Zone is essentially coextensive with the Richards sequence in the outer Mackenzie Delta and shelf of the Beaufort Sea. The assemblage was previously described from the Richards Formation in the Mackenzie Delta area by Young and McNeil (1984), and McNeil (1983) documented the distribution of its "shallow-water" biofacies, which is characterized by the monospecific assemblage of *Jadammina statuminis*.

The *Haplophragmoides* assemblage has a broad generic similarity to the *Recurvoides* assemblage, but has a much lower diversity and is easily distinguished by several endemic species such as *C. cyclops*, *H.* sp. 2000, and *J. statuminis*. As mentioned earlier, the endemic nature of this assemblage is a reflection of the restrictions between the Arctic Gulf and the world's major oceans during the Paleocene-Eocene (McNeil, in press).

The upper boundary of the *Haplophragmoides* Assemblage Zone is typically abrupt, truncated by a Late Eocene(?) regional unconformity that has been documented by Dixon et al. (1985, Fig. 36). The lower boundary is apparently gradational with unfossiliferous strata of the Reindeer sequence. There are no known sections that reveal the *Haplophragmoides* Zone passing gradationally into foraminiferal-yielding older Eocene strata in the Beaufort-Mackenzie Basin.

Portatrochammina Assemblage Zone

Stratotype. Natsek E-56 well, 1250 to 2720 m below K.B.

Age. Late Paleocene to Middle Eocene.

Placentammina sp. 2800 biofacies (shallow). *Portatrochammina* sp. 2849, P. sp. 2850, and *Placentammina* sp. 2800.

Verneuilina sp. 2700 biofacies (deep). *Ammodiscus* sp., *Arenobulimina* sp., *Bathysiphon* sp., *Brachysiphon*, *Haplophragmoides* sp., *Lagenammina* sp., *Pelosina* sp., *Portatrochammina* sp. 2849, *P.* sp. 2850, *Recurvoides* sp., *Reophax* sp., *Placentammina* sp. 2700, *Trochammina* sp., and *Verneuilina* sp. 2700.

Discussion. The *Portatrochammina* Assemblage Zone is characterized by a low diversity assemblage of agglutinated foraminifers, which have previously been described as the *Saccammina-Trochammina* assemblage from the Reindeer Formation in the Beaufort-Mackenzie Basin by Young and McNeil (1984) and McNeil (1984). The zone can be mapped extensively within the Reindeer sequence. Its biofacies provide control for distinguishing marginal marine facies from deeper neritic facies (unfortunately specific water depths cannot be determined), as illustrated in Figure 3. This also illustrates the marked diachroneity of the zonal boundaries in complex deltaic facies of the Reindeer sequence.

PLATE 2

Figure 1, a-c *Cancris subconicus* (Terquem), x 56, Edlok N-56, 468-483 m, GSC 89525.

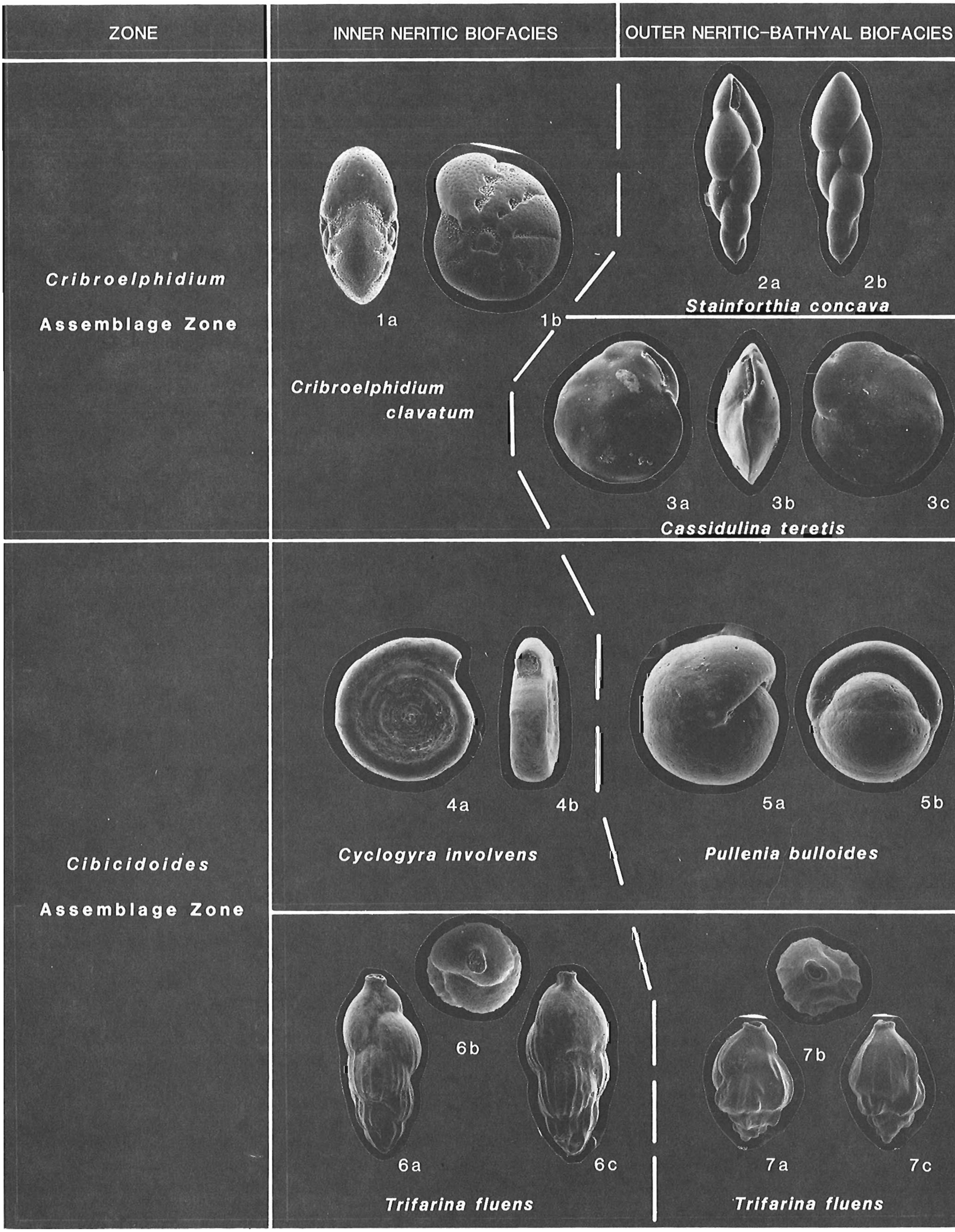
Figure 2, a, b *Haplophragmoides* sp. 2000, x 36, Pelly C-35, 2713-2743 m, GSC 68665.

Figure 3, a-c *Portatrochammina* sp. 2850, x 55, Natsek E-56, 1433-1463 m, GSC 89532.

Figure 4, a-c *Portatrochammina* sp. 2849, x 73, Natsek E-56, 2624-2633 m, GSC 89535.

Figure 5, a-c *Reticulophragmium* borealis (Petracca), x 37, Adlartok P-09, 1535-1547 m, GSC 89546.

Figure 6, a, b *Verneuilinoides* sp. 3495, x 35, GSC Locality C-59648 basal sandstone member, Moose Channel Formation, Big Fish River, northern Yukon, 68°34′03″N, 136°14′12″W, GSC 89547.


ZONE
INNER NERITIC BIOFACIES
OUTER NERITIC-BATHYAL BIOFACIES
Cribroelphidium
Assemblage Zone
1a
1b
Cribroelphidium
clavatum
2a
2b
Stainforthia concava
3a
3b
3c
Cassidulina teretis
Cibicidoides
Assemblage Zone
4a
4b
Cyclogyra involvens
5a
5b
Pullenia bulloides
6a
6b
6c
Trifarina fluens
7a
7b
7c
Trifarina fluens

The *Portatrochammina* assemblage bears little resemblance to younger or older assemblages and, like the other Paleocene-Eocene microfaunas, has a high degree of endemism. Its low diversity is no doubt a function of regionally low marine salinities in the face of the Reindeer deltaic system.

Reticulophragmium Assemblage Zone

Stratotype. Adlartok P-09 well, 1498 to 2647 m (T.D.) below K.B.

Age. Early Late Paleocene.

Reticulophragmium sp. 3307 biofacies (shallow). *Reticulophragmium* sp. 3307, *Reticulophragmium borealis* (Petracca).

Cibicidoides sp. 3450 biofacies (deep). *Ammodiscus* sp., *Arenobulimina* sp., *Arenoturrispirillina* sp., *Bathysiphon* sp., *Cibicidoides* sp. 3450, *Gyroidina* sp., *Haplophragmoides* sp., *Hyperammina* sp., *Labrospira* sp., *Lagenammina* sp., *Praebulimina* sp., *Pseudonodosaria* sp., *Recurvoides* sp., *Reophax* sp., *Placentammina* sp., *Thurammina* sp., *Trochammina* sp., *Verneuilina* sp., and *Verneuilinoides* sp.

Discussion. Foraminifers from the *Reticulophragmium* Assemblage Zone have been recorded in publications by Chamney (1971), Petracca (1972), Staplin (1976) and McNeil (1985). As illustrated in Figure 5, the zone is confined to the Ministicoog Member of the Moose Channel Formation in the Beaufort-Mackenzie Basin. Its age has been inferred from associated palynomorphs in the outcrop sequences of the Yukon Coastal Plain (A.R. Sweet, pers. comm.)

The microfauna of the *Reticulophragmium* Zone represents a widespread mid-Paleocene marine transgression that has been documented both in the Beaufort-Mackenzie Basin and in the Sverdrup Basin of the Arctic Archipelago (Wall et al., 1988). In both areas, mixed agglutinated-calcareous faunas are typical, but in the Beaufort area, agglutinated species are dominant, and in the Sverdrup area, calcareous species dominate.

PLATE 3

Figure 1, a, b *Cribroelphidium clavatum* (Cushman), x 90, Ukalerk C-50, 765-777 m, GSC 64627.

Figure 2, a, b *Stainforthia concava* Höglund, x 60, Nektoralik K-59, 427-457 m, GSC 89548.

Figure 3, a-c *Cassidulina teretis* Tappan, x 65, Natiak O-44, 880-898 m, GSC 89549.

igure 4, a, b *Cyclogyra involvens* (Reuss), x 41, Tarsiut A-25, 830-845 m, GSC 89550.

Figure 5, a, b *Pullenia bulloides* (d'Orbigny), x 82, Natiak O-44, 1069-1087 m, GSC 89551.

Figure 6, a-c *Trifarina fluens* Todd, x 100, Ukalerk C-50, 1615-1630 m, GSC 64656.

Figure 7, a-c *Trifarina fluens* Todd, Kopanoar M-13, 2637-2652 m, GSC 89552.

The shallow-water biofacies is characterized by either a monospecific assemblage of *R. borealis*, particularly in the area of the Tuktoyaktuk Peninsula, or by the distinctive, thin-walled *Reticulophragmium* sp. 3307, which usually occurs in the lower part of the *Reticulophragmium* Assemblage Zone in outcrop sections in the Yukon Coastal Plain and in the subsurface Mackenzie Delta.

Verneuilinoides Assemblage Zone

Stratotype. Outcrop sections within the basal sandstone member of the Moose Channnel Formation on Big Fish River, Yukon Coastal Plain (Fig. 5).

Age. Early Paleocene.

Trochammina sp. 3485 biofacies (shallow). *Trochammina* sp. 3485, *Verneuilinoides* sp. 3495.

Deep-water biofacies. unknown.

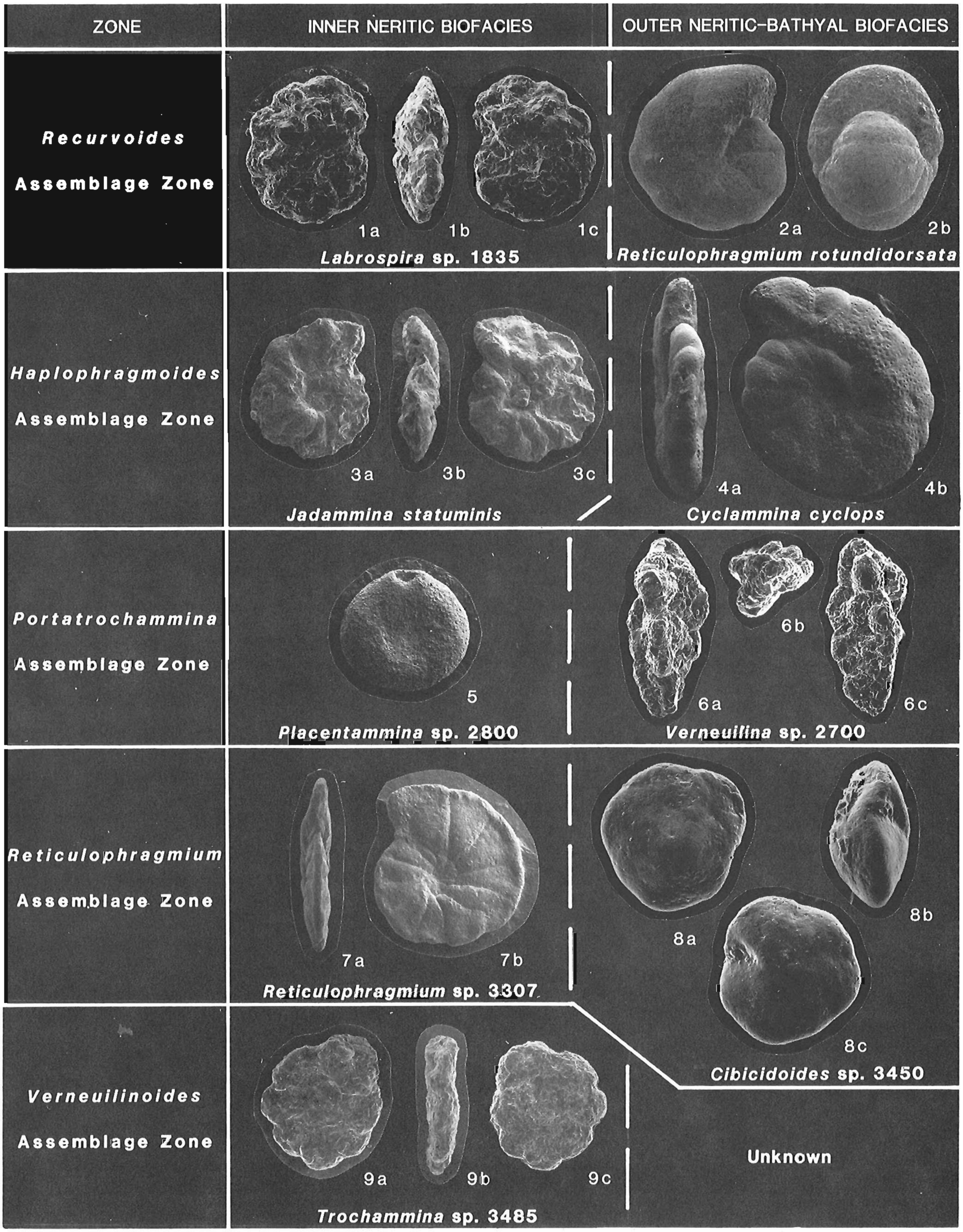
Discussion. The *Verneuilinoides* Assemblage Zone is known only from the Fish River and Eagle Creek areas of the Yukon Coastal Plain. The microfauna was referred to briefly by McNeil (1985), who recognized it as marking a thin, marginal-marine zone within the basal sandstone member of the Moose Channel Formation. The assemblage is composed of abundant specimens of only two species, *Trochammina* sp. 3485 and *Verneuilinoides* sp. 3495, which is typical of a low-salinity, marginal marine environment. The assemblage is dated as Early Paleocene on the basis of associated palynomorphs, as discussed above (*Verneuilinoides* sp. 3495 Interval Zone).

DISCUSSION OF THE ASSEMBLAGES

The Cenozoic benthic foraminiferal assemblage zones of the Beaufort-Mackenzie Basin are an important record of the basin's paleoenvironmental history, and they provide a means of understanding the basin in the context of major paleoceanographic events. Many of these events, such as the mid-Paleocene transgression, or the Early Oligocene establishment of broader connections between the Arctic and North Atlantic oceans, or the eustatic sea level drop near the end of the Miocene, are clearly reflected in the microfaunas of the Beaufort-Mackenzie Basin and are vital to understanding the depositional history of the basin (McNeil, in press).

The assemblage zones and their integral biofacies directly reflect many of the important sedimentation events in Beaufort-Mackenzie Basin history. For example, the marginal marine facies of the Kugmallit deltaic complex is outlined by a transition from unfossiliferous (terrestrial?) sediments into strata containing the low-diversity, agglutinated, *Labrospira* sp. 1835 biofacies. In succession, the *L.* sp. 1835 biofacies can be traced farther offshore, until it is replaced by diverse microfaunas diagnostic of the deeper-water *Reticulophragmium rotundidorsata* biofacies.

Agglutinated microfaunas of the *Recurvoides* assemblage are often combined in complex association with the calcareous *Cibicidoides* assemblage. The interplay between


ZONE
INNER NERITIC BIOFACIES
OUTER NERITIC-BATHYAL BIOFACIES
Recurvoides Assemblage Zone
1a
1b
1c
Labrospira sp. 1835
2a
2b
Reticulophragmium rotundidorsata
Haplophragmoides Assemblage Zone
3a
3b
3c
Jadammina statuminis
4a
4b
Cyclammina cyclops
Portatrochammina Assemblage Zone
5
Placentammina sp. 2800
6a
6b
6c
Verneuilina sp. 2700
Reticulophragmium Assemblage Zone
7a
7b
Reticulophragmium sp. 3307
8a
8b
8c
Cibicidoides sp. 3450
Verneuilinoides Assemblage Zone
9a
9b
9c
Trochammina sp. 3485
Unknown

agglutinated and calcareous microfaunas in the Beaufort-Mackenzie Basin Cenozoic is at present poorly understood and requires further research. There are many variables that could have been operable, such as primary environmental controls on the biotas (i.e., water masses), or secondary controls such as selective preservation (i.e., pH of sediments). Additionally, the record could be complicated by redeposition. The resolution of the controlling factors will, of course, be particularly difficult, given a data set based on well cuttings. Nonetheless, the microfaunal relationships within the Kugmallit sequence are important and are the subject of research currently in progress.

It has been demonstrated that some of the assemblage zone boundaries, such as the Cibicidoides-Recurvoides boundary, can be markedly diachronous (Fig. 3). Diachroneity is not a problem, if it can be recognized and measured (by interval zonation). The diachronous boundaries are, in fact, particularly significant, since they might shed light on the factors controlling the major facies trends in the basin.

SUMMARY

Twelve interval zones and seven assemblage zones are formally proposed for the Cenozoic strata of the Beaufort-Mackenzie Basin. The interval zonation is drawn on the last appearance datums (LADs) of selected benthic foraminifers, and each LAD is assigned a minimum absolute age. Paleocene to Eocene interval zones are based on the following succession of agglutinated species: *Verneuilinoides* sp. 3495 (>62.3 Ma), *Reticulophragmium borealis* (>60.0 Ma), *Portatrochammina* sp. 2849 (>57.8 Ma), *Portatrochammina* sp. 2850 (>?45.0 Ma), and *Haplophragmoides* sp. 2000 (>36.6 Ma). Oligocene to Holocene interval zones are based on the following succession of calcareous species: *Cancris subconicus* (>30.0 Ma), *Turrilina alsatica* (>23.7 Ma), *Asterigerina staeschei* (>10.4 Ma), *Cibicidoides* sp. 800 (>5.3 Ma), *Cibicides grossus* (>2.4 Ma), *Cribroelphidium ustulatum* (>1.6 Ma), and *Cassidulina reniforme* (>0.0 Ma).

The assemblage zones are defined by discrete associations of agglutinated and/or calcareous benthic foraminifers. The Paleocene to Eocene zones comprise mainly agglutinated species. The zones are named *Verneuilinoides* (Lower Paleocene), *Reticulophragmium* (lower Upper Paleocene), *Portatrochammina* (upper Upper Paleocene to mid- Middle Eocene), and *Haplophragmoides* (upper Middle-Upper Eocene). The Paleocene-Eocene assemblages represent the early Tertiary Arctic Gulf phase of Arctic Cenozoic history and are characterized by endemic species that were isolated from the world's major oceans. Oligocene to Miocene assemblages are composed of agglutinated and calcareous benthic foraminifers, many of which also occur in the marginal basins of the northern North Atlantic, a reflection of well established marine connections between the Arctic and Atlantic oceans. Pliocene to Holocene assemblages consist almost entirely of calcareous benthic species. They compare closely with microfaunas of the marginal basins of the northern North Atlantic and provide valuable information for regional correlation.

Each assemblage zone is divided into a shallow-water biofacies, approximating the inner neritic zone, and a deeper-water biofacies, approximating the outer neritic to bathyal zones. The seven shallow-water biofacies are informally named in ascending order: *Trochammina* sp. 3485, *Reticulophragmium* sp. 3307, *Placentammina* sp. 2800, *Jadammina statuminis*, *Labrospira* sp. 2835, *Cyclogyra involvens*, and *Cribroelphidium clavatum*. The Lower Paleocene *Verneuilinoides* Assemblage Zone is not known from a "deep-water" facies, but the remaining six zones are represented in deep-water biofacies, which are informally named in ascending order: *Cibicidoides* sp. 3450, *Verneuilina* sp. 2700, *Cyclammina cyclops*, *Reticulophragmium rotundidorsata*, *Pullenia bulloides*, and *Cassidulina teretis* biofacies.

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PLATE 4

Figure 1, a-c *Labrospira* sp. 1835, x 38, North Issungnak L-86, 3615-3620 m, GSC 89553.

Figure 2, a, b *Reticulophragmium rotundidorsata* (Hantken), x 36, Kopanoar M-13, 3054-3069 m, GSC 68753.

Figure 3, a-c *Jadammina statuminis* McNeil, x 36, Taglu G-33, 2240-2256 m, GSC 68725.

Figure 4, a, b *Cyclammina cyclops* McNeil, x 25, Netserk F-40, 2850-2865 m, GSC 89503.

Figure 5, *Placentammina* sp. 2800, x 60, Reindeer D-27, 1722-1734 m, GSC 68659.

Figure 6, a-c *Verneuilina* sp. 2700, Natsek E-56, 1280-1311 m, GSC 89533.

Figure 7, a, b *Reticulophragmium* sp. 3307, x 33, Ministicoog Member, Moose Channel Formation, Big Fish River, northern Yukon, 68°36′35″N, 136°10′00″W, GSC 89554.

Figure 8, a-c *Cibicidoides* sp. 3450, x 100, Tununuk K-10, 756-765 m, GSC 89555.

Figure 9, a-c *Trochammina* sp. 3485, x 114, GSC Locality C-59648 basal sandstone member, Moose Channel Formation, Big Fish River, northern Yukon, 68°34′03″N, 136°14′12″W, GSC 89556.

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APPENDIX

Depths of tops of sequences in wells containing biostratotypes

Adlartok P-09	Shallow Bay	93 m
	Kugmallit	400 m
	Richards	1189 m
	Lower Reindeer	1498 m
	Well T.D.	2647 m
Arnak L-30	Iperk	75 m
	Mackenzie Bay	944 m
	Kugmallit	1225 m
	Kopanoar	2981 m
	Richards	3447 m
	Reindeer	4511 m
	Well T.D.	4523 m
Natiak O-44	Iperk	55 m
	Akpak	960 m
	Mackenzie Bay	1222 m
	Kugmallit	2026 m
	Richards	2639 m
	Upper Reindeer	2990 m
	Lower Reindeer	3400 m
	Well T.D.	4650 m
Natsek E-56	Iperk	45 m
	Upper Reindeer	216 m
	Lower Reindeer	1951 m
	Fish River (?)	2644 m
	Well T.D.	3520 m
Netserk F-40	Iperk	20 m
	Akpak	719 m
	Mackenzie Bay	817 m
	Kugmallit	1422 m
	Richards	2426 m
	Reindeer	3773 m
	Well T.D.	4370 m

Gravity modelling in the Eurekan Orogen, Canadian Arctic Islands[†]

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Abstract

New gravity data have permitted the construction of gravity and topography profiles crossing three major structural elements of the Eurekan Orogen: Grantland Uplift on northwestern Ellesmere Island, Princess Margaret Arch on north-central Axel Heiberg Island, and Cornwall Arch on Cornwall Island.

Density models of the crust, incorporating near-surface geology and where necessary, inferred deeper structures, are presented for the Cornwall Arch profile and for part of the Princess Margaret Arch profile where it crosses the Stolz Thrust zone. In the latter case, a significantly large gravity gradient that delineates the eastern edge of the fault zone is several kilometres east of the mapped trace of the Stolz Thrust. The gravity signature of Cornwall Arch necessitates the incorporation of basement and/or crustal upwarping.

Analysis of the relationship between gravity and topography along the profiles of the three major uplifts indicates a gradational change from isostatic compensation beneath the Grantland Uplift, to antithetic isostatic compensation beneath Cornwall Arch. Princess Margaret Arch may be transitional between these end-members.

Résumé

De nouvelles données sur la gravité ont permis d'établir des profils de gravité et de topographie qui recoupent trois éléments structuraux importants de l'orogenèse de l'Eureka : le soulèvement de Grantland dans le nord-ouest de l'île Ellesmere, l'arche de Princess Margaret dans le centre-nord de l'île Axel Heiberg et l'arche de Cornwall dans l'île Cornwall.

Des modèles de densité de la croûte, tenant compte de la géologie près de la surface et des structures plus profondes telles que déduites, sont présentés pour le profil de l'arche de Cornwall et pour une partie du profil de l'arche de Princess Margaret là où il traverse la zone de chevauchement de Stolz. Dans ce dernier cas, un important gradient de gravité qui délimite la bordure orientale de la zone de failles se manifeste à plusieurs kilomètres à l'est de la trace cartographiée du chevauchement de Stolz. La signature de la gravité de l'arche de Cornwall exige qu'on tienne compte du bombement du socle ou de la croûte.

L'analyse de la relation entre la gravité et la topographie le long des profils des trois grands soulèvements révèle un passage graduel entre une compensation isostatique sous le soulèvement de Grantland et une compensation isostatique antithétique sous l'arche de Cornwall. L'arche de Princess Margaret constituerait une transition entre ces deux membres extrêmes.

[†] Contribution to Frontier Geoscience Program.

INTRODUCTION

One of the outstanding problems in interpreting the evolution of the Eurekan Orogen in the eastern Arctic Archipelago concerns the nature of structural elements of regional significance, at deep levels in the crust. Large-scale products of Eurekan compression, such as major arches and uplifts several hundred kilometres in length, and large reverse or thrust faults, represent a scale of crustal shortening that must be balanced with the shortening produced during the opening of Baffin Bay and the Labrador Sea (e.g., Balkwill, 1978, 1983a). The inferred style of detachment associated with structures like Cornwall and Princess Margaret arches, Grantland Uplift, and the Stolz and Lake Hazen thrust zones will clearly have an impact on these estimates of shortening. The estimates are in turn important for palinspastic reconstructions of sedimentary basin limits and also their subsidence histories.

Debate has generally been centered around the relative importance of two different styles of fault detachment: 1. subhorizontal detachments along zones of weakness (e.g., the upper and lower Paleozoic evaporite units — Nassichuk and Davies, 1980; Osadetz, 1982; Ricketts, 1987); or, 2. large, steep, reverse listric faults that perhaps were reactivated from Ellesmerian or Sverdrup Basin rift structures (e.g., Balkwill, 1983b; Higgins and Soper, 1983).

Reconciliation of these problems requires careful balancing of structural cross- sections derived from detailed structural analysis. Modelling of gravity data can also provide some constraints on the inferred geometry of large-scale crustal structures. This paper is a progress report on the analysis of new gravity data, collected by the Geological Survey of Canada in 1987, in the Axel Heiberg and western Ellesmere Islands area. The database is augmented by gravity data collected in the adjacent offshore and onshore regions since the establishment of the Polar Continental Shelf Project in 1958. The primary goal is to deal with some of the principal structural elements making up the Eurekan Orogen (Fig. 1) including Cornwall Arch (Cornwall and Amund Ringnes islands), Princess Margaret Arch and the Stolz Thrust (Axel Heiberg Island), and Grantland Uplift (northern Ellesmere Island). This progress report deals

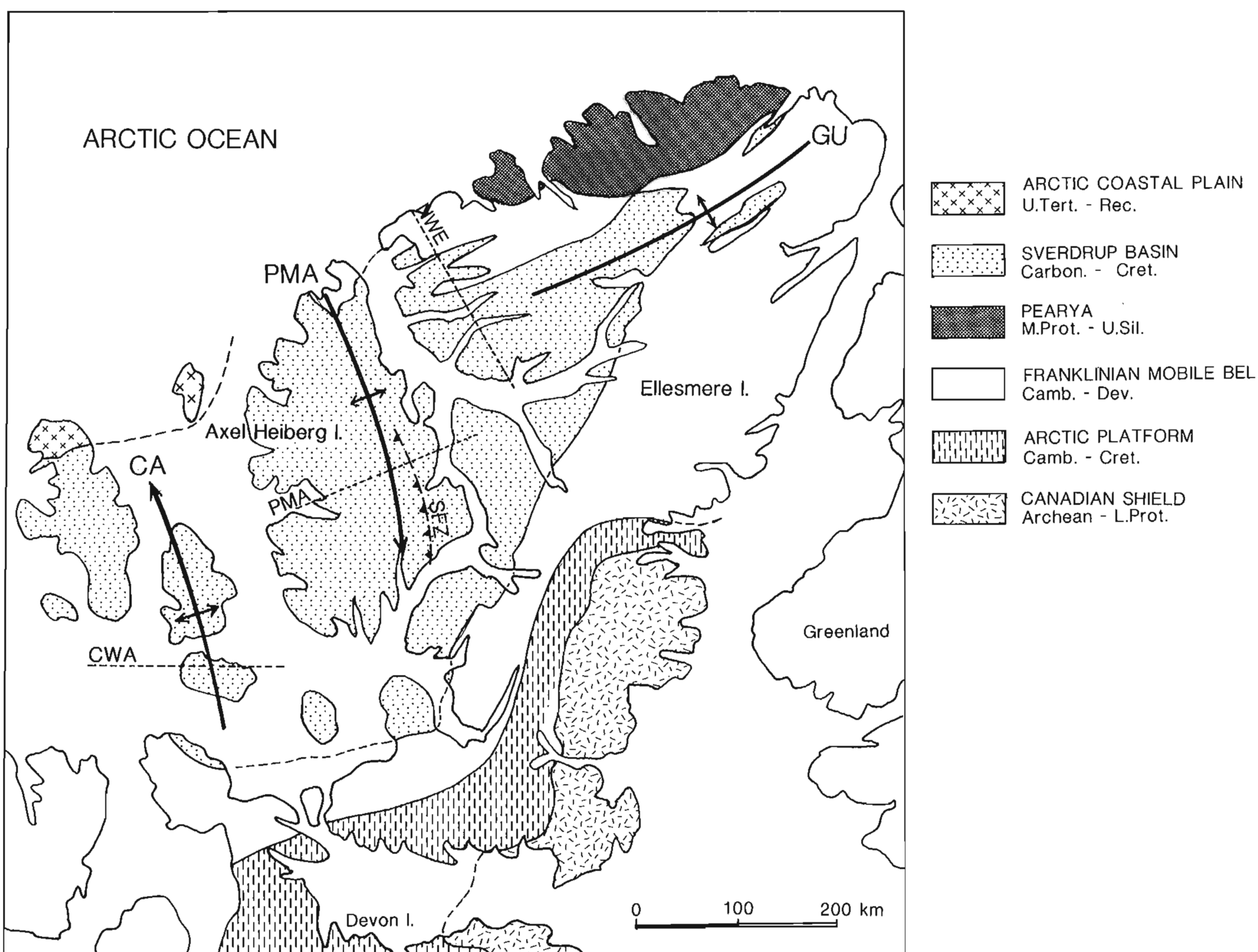

Figure 1. Regional tectonostratigraphic elements of the Canadian Arctic Islands, showing the locations of the Bouguer gravity and topography profiles discussed in the text. GU, Grantland Uplift; PMA (solid line), Princess Margaret Arch; CA, Cornwall Arch; SFZ-Stolz Fault zone; NWE, profile normal to the extension of the Grantland uplift; PMA (dashed line), profile normal to Princess Margaret Arch; CWA, profile across the Cornwall Arch.

mainly with the Stolz Thrust, which generates a gravity anomaly that is a fundamental component of the Princess Margaret Arch gravity signature, and with Cornwall Arch.

GRAVITY DATABASE

The motivation behind the present study was the new gravity data collected by the Geological Survey of Canada in 1987 in the area of onshore northern Axel Heiberg and northwestern Ellesmere islands (Fig. 2). This survey was undertaken in order to provide a new means of elucidating the regional structure of the Eurekan Orogen in an area where the regional stratigraphy was otherwise reasonably well known. The new data comprise nearly 500 regionally spaced stations (about 10 km spacing), supplemented by about 200 stations observed at approximately 2 km intervals along two transects: southwest- northeast across central Axel Heiberg Island, approximately normal to the axis of Princess Margaret Arch; and northwest-southeast across northwesternmost Ellesmere Island, normal to the southern limit of Grantland Uplift (Fig. 2). Elevations of gravity stations in the 1987 survey area were determined using an Inertial Surveying System supplied by the Geodetic Survey of Canada and, for the most part, are estimated to have an accuracy of ±3 m. All of the gravity measurements were also corrected for regional terrain effects (L. Losier and J.B. Boyd, pers. comm.), using a digital terrain model derived from 1:250,000 topographic maps, the largest scale available at the present time. The terrain corrections computed in this fashion, averaging 6 mGal, are believed to be within ±10 per cent of their total values. The accuracy of individual Bouguer anomalies is estimated to be within 3 mGal for all but several of the gravity stations. There are as yet no Geological Survey of Canada gravity data for much of the remainder of Ellesmere and Axel Heiberg islands, as shown in Figure 2; these areas, and the area covered in 1987, are among the last in Canada to be regionally gravity surveyed, mainly because of the logistical difficulties arising from the rugged terrain of the Eurekan Orogen and the remoteness of northern latitudes.

Figure 2. Distribution of new gravity data collected during the 1987 onshore survey.

In this paper, only the 2 km-spacing gravity transects of the 1987 survey are considered: profiles NWE (Northwestern Ellesmere, normal to the Grantland Uplift), and PMA (across Princess Margaret Arch and Stolz Thrust), as well as a third profile, CWA (across the Cornwall Arch along the major axis of Cornwall Island) — constructed from older (mainly 1961) regionally spaced data. The Bouguer anomalies measured along these profiles are shown in Figure 3a, b, and c. Also shown is the topography along each of the profiles, derived from measurements made at the gravity stations. All of the gravity and topography profiles were computed by fitting bicubic spline minimum curvature surfaces (Briggs, 1974) to the observed data within 20 km-wide corridors straddling each profile and then taking the projection of each surface along the profile position, thus allowing for some consideration of the lateral variations in gravity and topography and providing a slightly smoothed version of each. Also shown in Figures 3a-c are low-pass filtered versions of the Bouguer gravity and topography along each profile (dashed lines). These were computed from the first four harmonics of the Fourier transforms (and the mean) of each profile, so that they comprise the spectral energy of wavelengths greater than 50 km only. The local variations in the long wavelength functions most evident near the ends of the CWA topography and both NWE profiles are artifacts resulting from the application of discrete Fourier transforms to data series of finite length. The dotted lines in Figure 3a — profile NWE — display only 200 km wavelength spectral components (and the respective means) of the profile values.

ANALYSIS AND MODELLING OF SELECTED GRAVITY PROFILES

A fundamental difference in gravity signature along the gravity profiles in Figure 3 is observed. A typical isostatic compensation relationship emerges for northwestern Ellesmere Island such that the long-wavelength topographic high is coincident with a regional Bouguer gravity low. This implies that crustal shortening in the area, and concomitant erection of topographic highlands, was likely accompanied by crustal thickening. Cornwall Arch, on the other hand, exhibits an antithetic relationship whereby a major, positive Bouguer anomaly coincides with the topographic high of the arch. A similar antithetic pattern may be present along the Princess Margaret Arch profile, although the relationship there is complicated by the presence of more high frequency content to the gravity and topography, probably due to the greater structural complexity of Princess Margaret Arch compared to Cornwall Arch. In particular, much of the

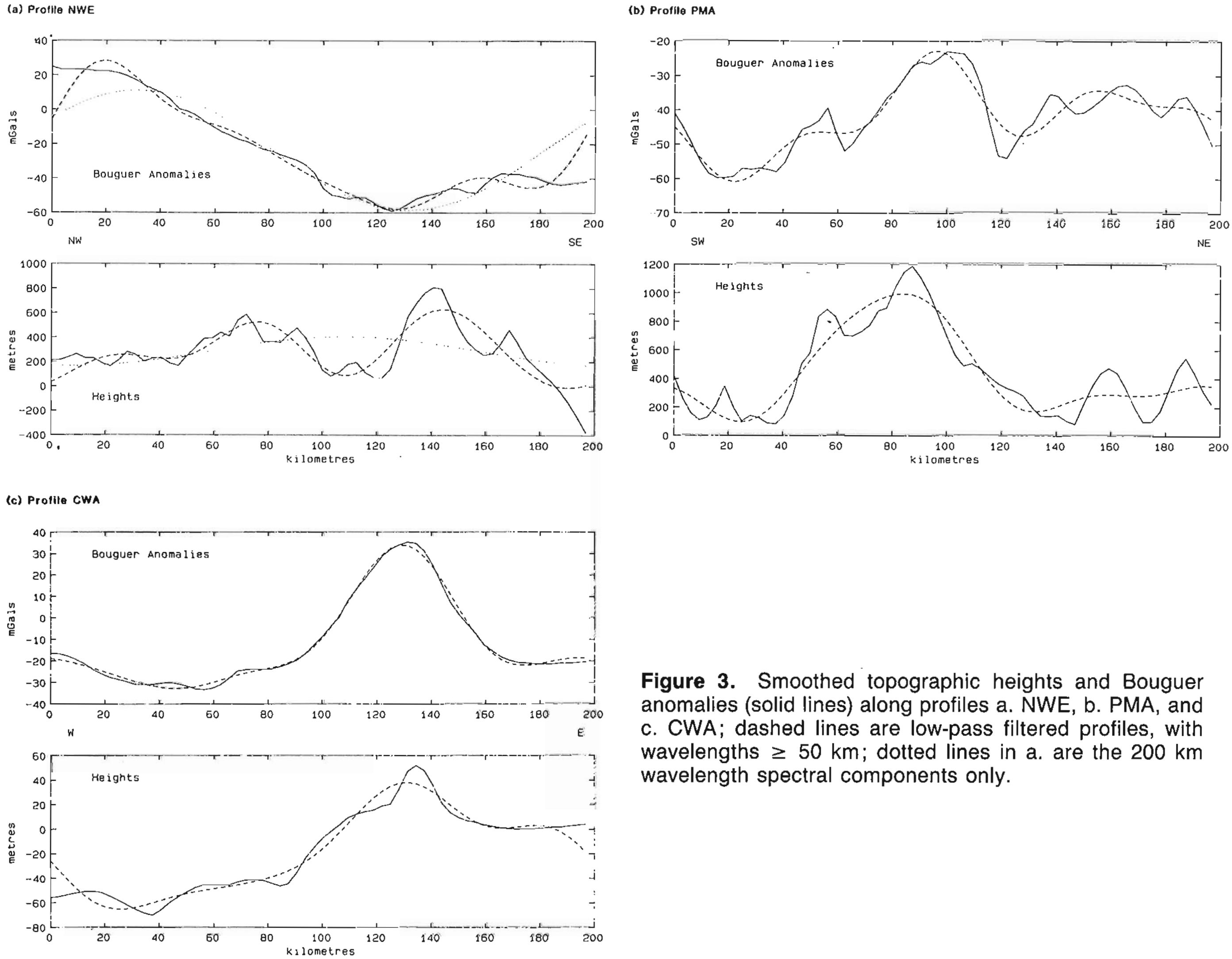

Figure 3. Smoothed topographic heights and Bouguer anomalies (solid lines) along profiles a. NWE, b. PMA, and c. CWA; dashed lines are low-pass filtered profiles, with wavelengths ≥ 50 km; dotted lines in a. are the 200 km wavelength spectral components only.

gravity high in central Axel Heiberg Island, and especially the very high gravity gradient on its eastern flank, appears to be related to the hanging wall of the Stolz Thrust.

The remainder of this paper presents the results of crustal density modelling of the gravity anomaly across Cornwall Arch in order to demonstrate more fully its apparent antithetic isostatic character, as well as an investigation of the near-surface geometry of the Stolz Thrust. One outcome of the latter is an elucidation of the role of the Stolz Thrust in the possibly isostatically antithetic gravity signature of Princess Margaret Arch. A more extensive appraisal of the isostatic nature of these features and of the Eurekan Orogen generally in this area is provided by Stephenson and Ricketts (in press).

CWA density model

A possible crustal and supracrustal density model that reproduces the salient features of the gravity anomaly along CWA is shown in Figure 4. The gravity calculations were made using the program MAGRAV2 (Broome, 1987). The shallow part (≤ 6 km) of the density model is based directly upon the published geological cross-section of Balkwill (1983a) and is intended to approximate the geometry and lateral density variations of a broad anticline forming Cornwall Arch on Cornwall Island. The approach taken has been to maximize, as much as seems reasonable, the gravity effect of these near-surface structures and then to consider what deeper sources, if any, were also necessary to provide a satisfactory model.

For modelling purposes, the Sverdrup Basin and Franklinian stratigraphic successions observed in the vicinity of Cornwall Island have been divided into four layers of different density:

1. A top layer about 2 km thick (Balkwill, 1983a) of Jurassic and Lower Cretaceous sandstone and shale, with an average density of at least 2400 kg/m^3 (Sobczak et al., 1986).
2. A middle layer comprising the early Mesozoic Heiberg and Blaa Mountain formations with average sandstone/shale densities of 2480 kg/m^3 (Sobczak et al., 1986). However, the Blaa Mountain Group, and, to a

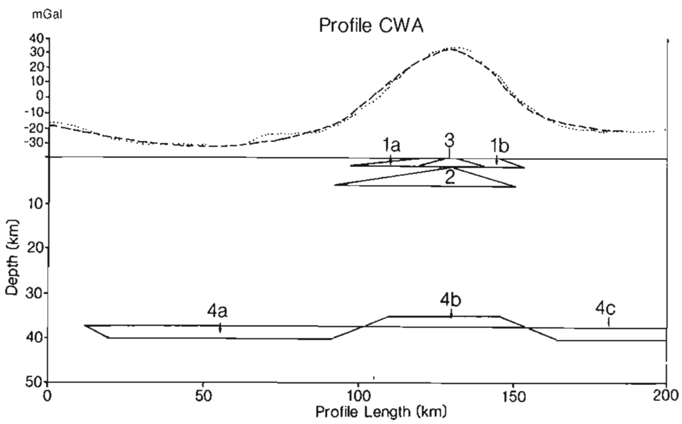

Figure 4. Profile CWA crustal density model, with model bodies numbered, showing smoothed observed (dotted line) and computed (dashed line) Bouguer anomalies.

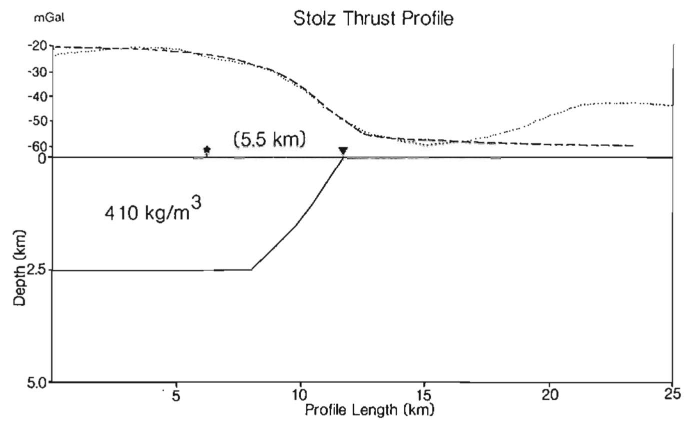

Figure 5. Stolz Thrust zone density model; the mapped trace of the Stolz Thrust (Thorsteinsson, 1971) is marked by a star; the modelled trace of the fault 5.5 kilometres to the east is marked by a solid triangle.

lesser extent, the Heiberg are the main hosts for thick, areally extensive, Cretaceous diabase and gabbro intrusions, having densities in the range of 2900 to 2950 kg/m^3 (Sobczak et al., 1986). Results from drilling on Amund Ringnes Island suggest that they may form about 20 to 25 per cent of this layer in the area of Cornwall Arch (e.g., Sobczak and Overton, 1984).

3. The Carboniferous to Lower Triassic units of the Sverdrup Basin underlying the Blaa Mountain Group, also with a thickness of about 2 km (e.g., Stephenson et al., 1987) and with a maximum density of 2550 kg/m^3 (Sobczak et al., 1986).
4. The lower Paleozoic Franklinian strata, of indeterminate thickness and having a probable average density of at least 2650 kg/m^3 (Sobczak et al., 1986).

The attributes and interpretations of the density bodies making up the model shown in Figure 4 are as follows.

1. Bodies 1a and 1b are representative of the Heiberg-Blaa Mountain layer, intruded by basic igneous rocks, flanking the axis of the Cornwall Arch Anticline. The assigned average density of 2600 kg/m^3 implies an upper limit of 25 per cent of igneous rock within the body. Even having adopted this upper limit, bodies 1a and 1b can account for no more than about 25 per cent of the total anomaly associated with the Cornwall Arch along the profile. Thus, in the absence of an extremely large, homogeneous, basic igneous body intruding the Sverdrup Basin along the axis of Cornwall Arch, and only along the axis of Cornwall Arch (e.g., Sobczak and Overton, 1984), the gravity data imply with some certainty that the décollement level of the Cornwall Anticline lies below the base of the Sverdrup Basin.
2. Body 2 represents Franklinian sediments filling the core of the anticline. To maintain a constant density contrast of $+100$ kg/m^3 with respect to the reference supracrustal section, body 2 must be presumed to increase in density by an additional 50 kg/m^3 in the core of the Cornwall Anticline; the model as shown also implies a similar density increase in strata of the lower Sverdrup Basin layer where they can be inferred to flank the anticline. With these probably overestimated modifications, about 50 per cent of the amplitude and 50 per cent of the wavelength of the Cornwall Arch gravity anomaly is modelled.
3. The near-surface body 3, with a density of 2700 kg/m^3, and representing lower Sverdrup strata consisting of up to 40 per cent basic intrusive material, a situation for which there is no direct geological evidence, is necessary in order to model the horizontal gradients observed along the Cornwall Arch profile. However, even incorporation of this shallow, high density body fails to fully account for the amplitude (and wavelength) of the observed anomaly.
4. Bodies 4a, 4b, and 4c emulate a crust-mantle boundary topography that, together with the supracrustal part of the density model (bodies 1-3), provides a satisfactory fit to the observed Bouguer gravity profile. It should be noted that similar, long wavelength density bodies located within the crust rather than at its base, for example, at a depth that could correspond with the base of the Franklinian succession (8 to 10 km; not shown in Figure 4), can also suitably reproduce the observed data.

Stolz Thrust model

The local variation in Bouguer gravity in the vicinity of the Stolz Thrust where it is crossed by profile PMA is shown in Figure 5. The observed gravity profile in Figure 5 was constructed directly from the values measured at approximately 2 km intervals along the transect. The horizontal Bouguer gravity gradient near the trace of the Stolz Thrust is notably large, reaching a maximum of about 8 mGal/km. The immediate hanging wall of the mapped trace of the fault (Thorsteinsson, 1971) comprises Blaa Mountain Group shales intruded by basic igneous sills, with up to 50 per cent of the succession made up of the latter (personal observation, 1988). A simple investigation of the geometry of the Stolz Thrust zone has been undertaken, making initial assumptions as follows:

1. The observed gradient is due entirely to the juxtaposition of two blocks having different densities: one (the hanging wall) comprising strata of the Blaa Mountain-Heiberg groups, massively intruded by basic igneous material, and the other (the footwall) comprising essentially unintruded Jurassic and Lower Cretaceous strata stratigraphically overlying the Heiberg Group.
2. The boundary between the two blocks represents the Stolz Thrust and is constrained to be vertical or dipping to the west, either as a plane or with a listric-style geometry.
3. The upper boundaries of the blocks lie at the datum surface (sea level) and the lower boundaries lie at a common depth no greater than 2.5 km below sea level. This maximum depth value is based upon (i) estimated initial thicknesses of 2.5 km and 4 km, respectively, for the overlying Jurassic-Lower Cretaceous unit and for the Blaa Mountain-Heiberg unit (with intrusive layers) in the vicinity of north-central Axel Heiberg Island (e.g., Ricketts, 1987), and (ii) an estimated structural relief on Princess Margaret Arch (of which the Stolz Thrust forms part of the eastern limit) of 4 km (Gould and deMille, 1964; Balkwill, 1978). Hence, only 2.5 km of the denser unit would be preserved in the hanging wall, and the underlying units would not be expected to be highly intruded by basic material. These values are not tightly constrained and the implications of adopting a greater thickness for the hanging wall block are discussed briefly.

Given the stated assumptions, the observed Bouguer gravity gradient reasonably constrains the position of the surface trace of the fault and its subsurface dip (initially assumed to be uniform), as shown in Figure 5. The computed dip of the fault at the surface of 38 to 40° compares to a value of 40° reported by van Berkel (1986). (The model also shows a slight flattening of the fault at 1.25 km depth but, given the nature of the adopted assumptions and a number of variables that have not been considered, the presence of this particular geometric style is poorly constrained by these data). On the other hand, the position of the fault trace at the surface is well constrained to be about 5.5 km east of where it has been mapped on the basis of a prominent fault scarp. Thus, the gravity model, even with its simplifying assumptions, appears to indicate unambiguously that the fault zone in this area is composed of a number of imbricate slices. It is noted, in this respect, that Blaa Mountain strata are locally exposed in the apparent footwall of the mapped fault (e.g., Thorsteinsson, 1971) and it is surmised that these strata form part of one or a number of thrust slices east of the fault mostly covered by till and other, younger, materials.

Given an assumed thickness for the hanging wall block of 2.5 km, as shown in Figure 5, the relative density of this block versus the footwall block is fixed by the amplitude of the anomaly on the hanging wall side of the fault, and is +410 kg/m^3. If a "background" density of 2400 kg/m^3 is adopted for the Jurassic-Lower Cretaceous strata (Sobczak et al., 1986) assumed to make up the footwall, then the inferred hanging wall density of 2810 kg/m^3, for a 2.5 km thick body, is considered too large. A density as high as this implies that 70 to 80 per cent of the material in the hanging wall consists of basic sills (adopting a density of 2900-2950 kg/m^3 for this material and a density of 2480 kg/m^3 for the host shales and sandstones; Sobczak et al., 1986), a figure not supported by any geological observations in the area. A related problem is the minor misfit, which cannot be improved within the confines of the given assumptions, between the observed and computed Bouguer anomalies at the foot of the gradient. The misfit in this region is primarily a consequence of the high density contrast required to generate the amplitude of the hanging wall anomaly. These problems can be interpreted as implying that material of density less than 2400 kg/m^3, such as evaporite (2310 kg/m^3; Sobczak et al., 1986), must be present in the footwall block. In this respect, it is noted that, although surface-penetrating footwall diapirs are not observed in the immediate vicinity of the location of the profile shown in Figure 5, they occur elsewhere along the thrust zone (e.g., Thorsteinsson, 1971). Their likely presence in the footwall is related to the probability that at least one décollement level for the thrust exists in the evaporitic units at the base of the Sverdrup Basin (e.g., Ricketts, 1987). A slightly thickened wedge of essentially unconsolidated Eureka Sound Group sediments (density 52040-2320 kg/m^3; Sobczak et al., 1986), perhaps deposited coevally with motion on the Stolz Thrust (e.g. Ricketts, 1987), may also contribute to the inferred negative density body in the footwall of the fault.

However, the presence of negative density element(s) in the footwall is unlikely to resolve completely the enigmatically large amplitude and gradient associated with the hanging wall gravity anomaly. Reasons for this include:

1. The possible countervailing attenuation of the amplitude of the hanging wall anomaly by the negative density bodies.
2. The probable presence of some basic intruded material in the "background" Jurassic-Lower Cretaceous footwall unit.
3. The common occurrence of basic sills in contact with, and possibly related to footwall diapirs.
4. The fact that the entire Blaa Mountain-Heiberg hanging wall sequence is unlikely to be uniformly intruded by the sills as assumed.

It follows, of course, that relaxation of the 2.5 km maximum depth constraint can result in alternative gravity models, with thicker bodies of systematically lower relative density making up the hanging wall block. However, this approach is not consistent with the final point made above; for example, if the hanging wall block consisted entirely of Blaa Mountain shales with about 25 per cent basic intrusions, then it would have to be 6 to 7 km thick. This does not appear geologically likely.

Alternatively, the large gravity high on the hanging wall of the Stolz Thrust zone, coinciding as it does with the trace of Princess Margaret Arch, may in part be due to more deep-seated regional structures, perhaps analogous to the Cornwall Arch, and onto which the Stolz Thrust anomaly has been superimposed. The similarity in the character of the long-wavelength topography and Bouguer gravity signatures of Cornwall and Princess Margaret arches (Fig. 3) is consistent with this possibility.

DISCUSSION

The topography on northwestern Ellesmere Island appears to be isostatically compensated, implying that crustal shortening in this area was accompanied by crustal thickening. This fact in itself does not strongly support either the thin-skinned (i.e., Osadetz, 1982) or the basement involvement (Higgins and Soper, 1983) hypothesis for this part of the Eurekan Orogen. However, recent structural considerations favour the latter (e.g., Maurel, this volume; Okulitch and Trettin, in press).

Upper-limit estimates of the supracrustal density perturbation inferred from the observed geology of Cornwall Arch (i.e., a moderately simple anticlinal form associated with the diabase intrusions), cannot closely approximate its associated gravity signature. On the other hand, if the arch uplift is assumed to also involve crystalline basement lying approximately at or below the base of the Franklinian succession (and possibly as deep as the crust-mantle boundary), then a satisfactory model for the gravity anomaly can be found. Whereas conventional isostatic models typically predict that structures like Cornwall Arch (with a wavelength about 200 km and as old as 50 m.y. or more) would be compensated, these results appear to inescapably imply that not only is the topographic signature of Cornwall Arch not isostatically compensated but in fact, that a negative or antithetic isostatic compensation exists. A possibly strong piece of evidence in favour of this idea is provided by the interpretation of seismic refraction data across Cornwall Arch (Forsyth et al., 1979) which shows a basal crustal upwarp, similar in amplitude to the structure argued here solely on the basis of the gravity and topography signatures.

Investigation of the high gravity gradient associated with the eastern edge of Princess Margaret Arch reveals that it can be accounted for by an idealized (or simplified) form of the geology observed at the Stolz Thrust zone. It is important to note that the analysis establishes that unexposed, imbricated thrust slices are present in the area of the transect. Folded Blaa Mountain strata in the footwall of Stolz Thrust along the transect may, in fact, correspond to the termination of blind thrusts. The analysis also suggests that the Stolz Thrust hanging wall anomaly might be superimposed upon a more regional gravity high associated with Princess Margaret Arch.

Thus, the Princess Margaret Arch profile may be transitional between Cornwall Arch and Grantland Uplift, not only geographically within the Eurekan Orogen, and to some extent in terms of the shallow Eurekan structures that make up Princess Margaret Arch, but also in terms of large-scale crustal characteristics and incumbent isostatic mechanisms.

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Preliminary palynological zonation of surface and subsurface sections of Carboniferous, Permian and lowest Triassic rocks, Sverdrup Basin, Canadian Arctic Archipelago†

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Abstract

Nine preliminary palynological assemblage zones defined in the Sverdrup Basin succession facilitated local correlation between the surface and subsurface, and interregional correlation with strata in the Yukon Territory, Eastern Greenland, Svalbard, Western Europe, and the Pechora Basin of the Soviet Union. Most of the samples used for establishing the zonation were collected from sections in the shallow basin margin facies, which have been accurately dated in terms of standard marine stages on the basis of ammonoids, conodonts, brachiopods, and foraminifers.

In the subsurface of the western Arctic, seven of the nine zones have been documented in wells penetrating shallow platform facies on south Sabine Peninsula, and deeper basin facies farther north.

Quantitative assemblage variations at a suprageneric level in the Carboniferous and Permian palynofloras may reflect a series of climatic changes, from humid to arid, throughout the succession. Thermal Alteration Indices (TAI) of subsurface samples range from 2- to 4; in general, values increase with depth of burial and proximity to igneous intrusions.

Résumé

Neuf zones d'associations palynologiques préliminaires, définies dans la succession du bassin de Sverdrup, ont facilité la corrélation locale entre la surface et la subsurface, et la corrélation interrégionale avec le Yukon, l'est du Groënland, le Svaldard, l'Europe de l'Ouest et le bassin de Pechora en Union soviétique. La plupart des échantillons utilisés pour établir la zonation ont été prélevés dans des coupes du faciès de la marge du bassin peu profond, lesquelles sont bien datées en termes d'étages marins standard par des ammonoïdes, des conodontes, des brachiopodes et des foraminifères.

Dans la subsurface de la partie ouest de l'Arctique, sept des neuf zones ont été documentées dans des puits recouvrant un faciès de plate-forme peu profonde au sud de la péninsule de Sabine et un faciès de bassin plus profond plus au nord.

Des variations quantitatives des associations à un niveau macrogénérique observées chez les palynoflores du Carbonifère et du Permien peuvent traduire une série de changements climatiques, de conditions humides à conditions arides qui se seraient succédés, tout au long de cette succession. Les indices d'altération thermique (IAT) d'échantillons prélevés sous la surface varient de 2- à 2; de façon générale, les valeurs augmentent en fonction de la profondeur d'enfouissement et de la proximité des intrusions ignées.

† Contribution to Frontier Geoscience Program.

INTRODUCTION

The oldest known sediments in the Sverdrup Basin are the lacustrine, fluviatile and coal-bearing deposits of the Emma Fiord Formation deposited on the eroded surface of strata of the Franklinian Geosyncline (Davies and Nassichuk, 1988). Oil shales occur within the formation at some localities, such as Devon Island, where lacustrine sediments predominate (Goodarzi et al., 1987). At other localities, for example at the type section on Kleybolte Peninsula, Ellesmere Island, and at Svartevaeg Cliffs on Axel Heiberg Island,the formation is dominantly fluviatile. Palynological data indicate that the Emma Fiord Formation is of late Viséan (V3) age (Utting et al., 1987; Fig. 1).

The Emma Fiord Formation is overlain disconformably by rift-associated, red, quartzose sandstone and conglomerate of the Borup Fiord and Canyon Fiord formations. These formations are overlain by Serpukhovian to Moscovian marine siltstone, shale, dolomite and limestone, indicating transgression (Thorsteinsson, 1974; Balkwill, 1978; Fig. 1).

Sedimentation during the rest of the Late Carboniferous and Permian took place on shallow water platforms surrounding one, or several, deeper-water basinal areas (Nassichuk and Davies, 1980; Beauchamp et al., 1987). A number of laterally equivalent facies belts developed, including coarse grained marginal silicilastics, shelf carbonates and basinal shale. These belts were diachronous, and they fluctuated and migrated with time; they are summarized by Beauchamp et al., this volume). Formations yielding palynomorphs in the marginal facies include the Canyon Fiord (in part), Sabine Bay, Assistance, and Trold Fiord formations (Fig. 1). Shallow water carbonates and evaporites on the adjacent peripheral platform belong to the Canyon Fiord, Nansen, and Belcher Channel formations, the Unnamed formation of Nassichuk and Wilde (1977), and the Degerböls Formation. Deeper-water shales, cherts, argillaceous limestones, and evaporites in the central areas (Hare Fiord and van Hauen formations) were deposited under poorly oxygenated, probably anoxic conditions (Beauchamp et al., 1987). The Upper Carboniferous Otto Fiord Formation, in the axial part of the basin, consists of thick evaporites, comprising gypsum, anhydrite and halite, and contains thin shale intercalations.

SERIES	STAGE	FORMATION	PALYNOMORPH ZONE
Lower Triassic	Griesbachian	Bjorne	*T. stoschiana - S. richteri*
Upper Permian	Wordian	Trold Fiord	*Taeniaesporites* sp.
Lower Permian	Roadian	Assistance Sabine Bay	*A. insignis - Triadispora* sp.
	Artinskian	Unnamed	*L. monstruosus - V. costabilis*
	Sakmarian	Belcher Channel	*W. striatus - P. perfectus*
	Asselian	U	*Potonieisporites* spp. - *Vittatina* sp.
Upper Carboniferous	Gzhelian	M Canyon Fiord	
	Kasimovian	L	*P. novicus*
	Moscovian		Disaccate
	Bashkirian	Otto Fiord	No Data
Lower Carboniferous	Serpukhovian	Borup Fiord	
	Viséan 3 2 1	Emma Fiord	*M. aurita - D. saetosus*
	Tournaisian		No Data
Upper Devonian	Famennian		

Figure 1. Age of palynomorph zones and rock units.

No upper Permian rocks of Capitanian, Dzhulfian and Changhsingian age have been identified at the margin of the basin, where the Triassic rests disconformably on Wordian rocks. However, in the deeper part of the basin, sedimentological evidence suggests continuous deposition, but paleontological evidence for post-Wordian sediments has not yet been found. Nevertheless, at some localities in the upper part of the Trold Fiord Formation, sediments lacking macrofossils occur and it is possible that these are uppermost Wordian or younger (Beauchamp et al., this volume). The lowest Triassic (Griesbachian) of the Sverdrup Basin consists of marine greenish grey shales of the Blind Fiord Formation in the centre of the basin; laterally equivalent terrestrial sandstones of the Bjorne Formation occur at the margins (Fig. 1).

PALYNOLOGICAL ASSEMBLAGE ZONES

The assemblage zone, as defined in the North American Stratigraphic Code (1983), has been used in this study. Many of the zonal boundaries are not yet fully defined because the sections studied are not continuous. The zones are recognized on the basis of distinctive assemblages of species, and on qualitative and quantitative criteria. The species used to name the zones are common, but may also occur in overlying and underlying zones. The zonal scheme is based on work in progress on approximately 500 outcrop samples collected mainly from the shallow platform facies along a 50 to 100 km wide outcrop belt on the southern depositional margin of the basin. Along this belt, thermal maturity is low (Utting et al., in press) and palynomorphs are generally very well preserved (Utting, 1985). In addition, relatively good biostratigraphic control is provided by marine faunas; for example, ammonoids (Nassichuk, 1975), fusulinaceans (Nassichuk and Wilde, 1977), nonfusulinacean foraminifers (B.L. Mamet, in Nassichuk, 1975), and conodonts (Henderson, in Beauchamp et al., this volume), enabling reliable dating to be made by comparison with standard marine stages. In the proximal, deeper water facies, anoxic conditions led to the growth of pyrite crystals on spore and pollen exines, causing severe damage and making accurate identification difficult.

OCCURRENCE OF PALYNOLOGICAL ASSEMBLAGE ZONES IN THE SUBSURFACE

Subsurface determination of palynological zones was based on analysis of approximately 250 samples of core and cuttings from seven wells drilled on Sabine Peninsula, Melville Island (Fig. 2). In the south, the wells (Weatherall O-10, Eldridge Bay E-79) penetrate platform facies, whereas those farther north (Sherard Bay F-34, Marryatt K-71, Hecla J-60, Chads Creek B-64, and Drake Point D-68) penetrate platform and proximal deeper basin facies (Fig. 2). Formation tops, supplied by Majid (Majid, 1989) were picked by combining information from lithological changes, log characteristics and interpretation of seismic data. The top of each palynological zone is placed at the highest occurrence of the assemblage in the well. Some of the boundaries of the zones are not certain because of caving and barren samples. For example, caving may considerably distort the lower ranges of zones, especially where barren intervals underlie fossiliferous strata. Intervals where no data were obtained are shown in Figure 3.

DESCRIPTION OF ZONES

The newly proposed assemblage zones are summarized below, commencing with the oldest. The distribution of zones is listed for surface sections throughout the Sverdrup Basin and for subsurface sections on Melville Island.

Murospora aurita — Diatomozonotriletes saetosus
Assemblage Zone

This zone is characterized by a number of stratigraphically important species such as *Cingulizonates bialatus* (Waltz) Smith and Butterworth, 1967, *Colatisporites decorus* (Bharadwaj and Venkatachala) Williams, 1979, *Diatomozonotriletes saetosus* (Hacquebard and Barss) Hughes and Playford, 1961, *Murospora aurita* (Waltz) Playford, 1962, *Knoxisporites stephanephorus* Love, 1960, *K. triradiatus* Hoffmeister, Staplin, and Malloy, 1955, *Lycospora pusilla* (Ibrahim) Schopf, Wilson, and Bentall, 1944, *Perotrilites tessellatus* (Staplin) Neville, 1968, *Spelaeotriletes arenaceus* Neves and Owens, 1966, and *Waltzispora planiangulata* Sullivan, 1964. *Rotaspora fracta* (Schemel) Smith and Butterworth, 1967, is rare.

Outcrop occurrences

Emma Fiord Formation, Lyall River, Grinnell Peninsula, Devon Island

Similar assemblages occur in the type section of the Emma Fiord Formation, Kleybolte Peninsula, Ellesmere Island, and at Svartevaeg Cliffs, Axel Heiberg Island. At these two more northern localities the thermal maturity is high (TAI 4- to 5), and specimens are difficult to identify. Despite this bad preservation it is still apparent that the assemblages are dominated by *Cingulizonates bialatus*, *Densosporites* spp., *Murospora aurita* and *Lycospora pusilla*.

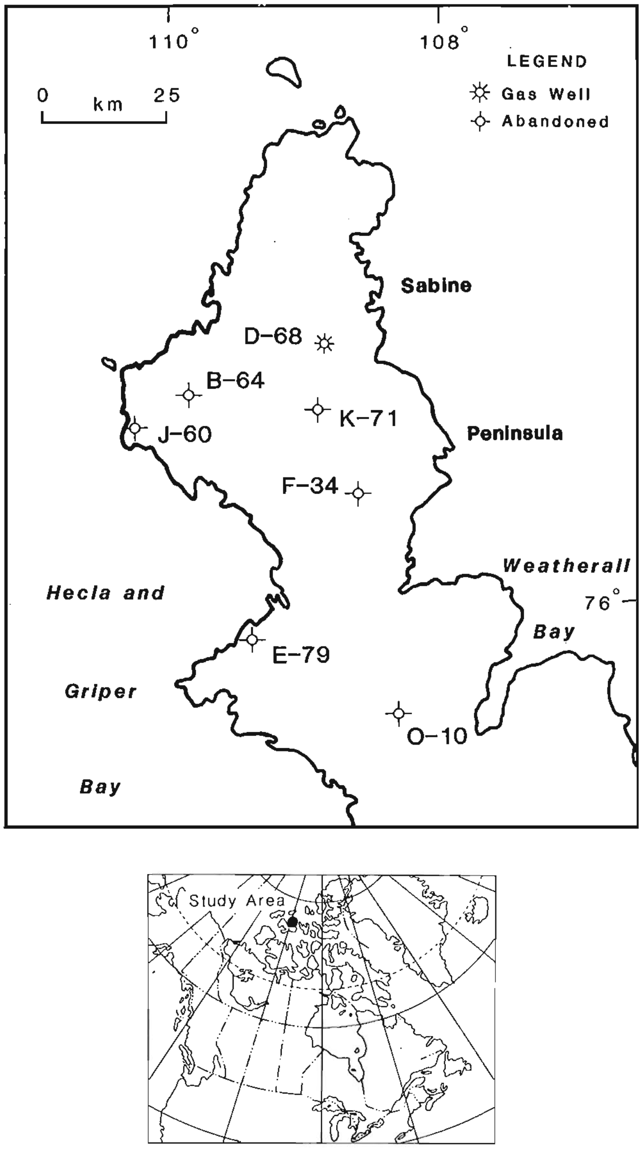
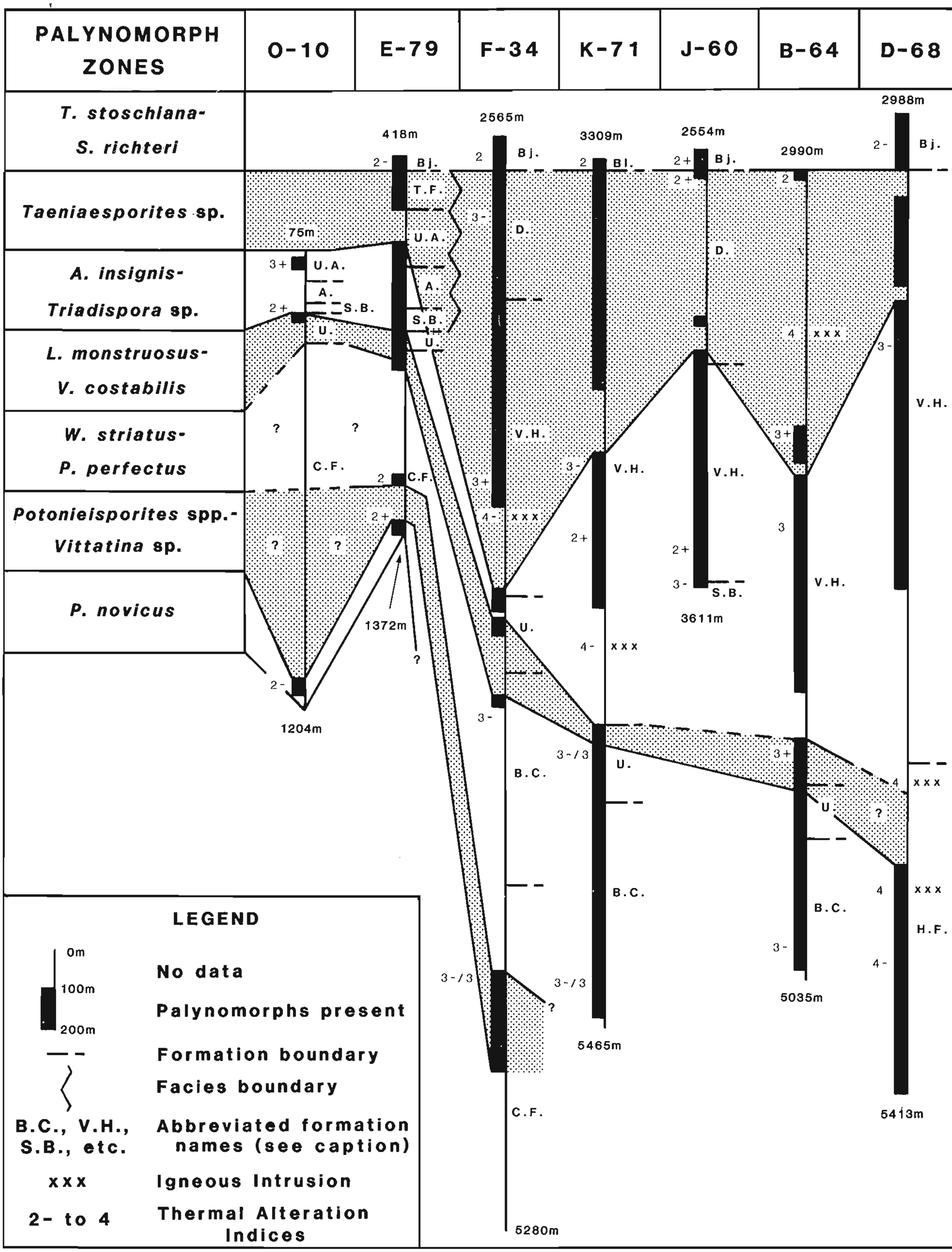


Figure 2. Location of wells studied on Sabine Peninsula, Melville Island.

O-10 = Dome, Panarctic, Texex, Weatherall 0-10; Lat. 75°49′51.9″N, Long. 108°31′50″W, GSC locality C-46886.
E-79 = Panarctic et al. Eldridge Bay E-79; Lat. 75°58′21.02″N, Long. 109°29′38.42″W, GSC locality C-59874.
F-34 = Panarctic et al. Sherard Bay F-34; Lat. 76°13′21.622″N, Long. 108°43′39.423″W, GSC locality C-61559.
K-71 = Panarctic et al. Marryatt K-71; Lat. 76°21′37.48″N, Long. 108°58′25.90″W, GSC locality C-97716.
J-60 = Panarctic et al. Hecla J-60; Lat. 76°19′37.88″N, Long. 110°19′49.07″W, GSC locality C-23415.
B-64 = Panarctic et al. Chads Creek B-64; Lat. 76°23′08″N, Long. 109°54′21″W, GSC locality C-97730.
D-68 = Panarctic et al. Drake Point D-68; Lat. 76°27′36.85″N, Long. 108°55′23.01″W, GSC locality C-45610.



PALYNOMORPH ZONES
O-10
E-79
F-34
K-71
J-60
B-64
D-68
T. stoschiana- S. richteri
Taeniaesporites sp.
A. insignis- Triadispora sp.
L. monstruosus- V. costabilis
W. striatus- P. perfectus
Potonieisporites spp.- Vittatina sp.
P. novicus
75m
1204m
418m
1372m
2565m
5280m
3309m
5465m
2554m
3611m
2990m
5035m
2988m
5413m
LEGEND
0m
100m
200m
No data
Palynomorphs present
Formation boundary
Facies boundary
B.C., V.H., S.B., etc.
Abbreviated formation names (see caption)
xxx
Igneous Intrusion
2- to 4
Thermal Alteration Indices

Subsurface occurrences

The zone has yet to be found in the subsurface of the western Arctic, although rare reworked species of this zone occur in some Permian samples.

Age

Comparison of the Grinnell material with the zonal scheme of Western Europe (Clayton et al., 1977) suggests correlation with the *Raistrickia nigra-Triquitrites marginatus* (NM) Zone, or possibly with the *Tripartites vetustus - Rotaspora fracta* (VF) Zone, of late Viséan V3 age (Fig. 1). Comparison with assemblages from beds in Western Canada (Mattson Formation and Stoddart Group), where marine invertebrate fossils occur, also indicates a late Viséan (V3) age (Richards et al., in press). The Kleybolte and Svartevaeg material may be of slightly different age within the late Viséan V3 , and/or may reflect deposition in different sedimentary environments, the Grinnell rocks representing lacustrine sedimentation and the Kleybolte - Svartevaeg, mostly fluviatile conditions.

Disaccate Assemblage Zone

The assemblage contains species of *?Cyclogranisporites, ?Pityosporites, ?Piceapollenites, ?Potonieisporites, ?Protohaploxypinus,* and *?Striatoabieites.*

Identifications are tentative because of thermal alteration, and severe corrosion of disaccate pollen, monosaccate pollen, and trilete spores. Scolecodonts are rare.

Outcrop occurrences

Upper part of type section of Otto Fiord Formation, Ellesmere Island; Otto Fiord Formation, Barrow Dome, Sabine Peninsula, Melville Island.

Subsurface occurrences

None known.

Figure 3. Palynostratigraphic correlation of wells drilled by Panarctic et al., Thermal Alteration Indices, and formations. Alternate zones shaded for clarity. Depths of intervals studied indicated at top and bottom of columns (see Figure 2 for location).

Bj.	Bjorne Formation	A.	Assistance Formation
Bl.	Blind Formation	S.B.	Sabine Bay Formation
D.	Degerbols Formation	U.	Unnamed formation
T.F.	Trold Fiord Formation	B.C.	Belcher Channel Formation
U.A.	Unit "A"		
V.H.	van Hauen Formation	N.	Nansen Formation
H.F.	Hare Fiord Formation	C.F.	Canyon Fiord Formation

Age

A Morrowan (Bashkirian) to early Atokan (late Bashkirian to early Moscovian) age was determined from the associated marine fauna (conodonts and ammonoids,Nassichuk and Davies, 1980).

In Yukon Territory, samples from beds of Bashkirian age (Bamber and Waterhouse, 1971) in the lower part of the Blackie Formation of the Northern Ogilvie Mountains, are dominated by trilete spores. Monolete spores and rare monosaccate pollen, including *Potonieisporites elegans* (Wilson and Kosanke) Wilson and Venkatachala emend. Habib, 1968, occur (Bamber et al., this volume).

Potonieisporites novicus Assemblage Zone

This zone is dominated by *Potonieisporites novicus* Bharadwaj, 1954, and its morphological variants. In addition, there are representatives of the *Cordaitina - Plicatipollenites - Nuskoisporites* complex and species of *Discernisporites.*

Outcrop occurrences

None known.

Subsurface occurrences

Weatherall O-10 (1192-1242 m) and Eldridge Bay E-79 (1289-1347 m) wells; Canyon Fiord Formation.

Age

In the Eldridge Bay E-79 well, at a depth of 1164-1173 m, fusulinacean foraminifers in a limestone bed are of Desmoinesian, or late Moscovian age (R. Neves, pers. comm., 1975), suggesting that the *Potonieisporites novicus* Assemblage Zone is no older than late Moscovian, although it may extend into the Kasimovian.

Similar palynomorph assemblages occur in the lower part of the unnamed equivalents of the Ettrain Formation in southern Eagle Plain and in the lower Blackie Formation of the northern Ogilvie Mountains, Yukon Territory (Bamber et al., 1989); these beds are late Bashkirian to Moscovian in age (Bamber and Waterhouse, 1971).

Potonieisporites spp. - *Vittatina* sp. Assemblage Zone

This zone contains abundant monosaccate grains (*Potonieisporites* and *Florinites*, and species of the *Cordaitina - Plicatipollenites - Nuskoisporites* complex). Polyplicate grains (*Vittatina* sp. and *Weylandites* sp.) are common.

Outcrop occurrences

Middle Canyon Fiord Formation, Melville Island.

Subsurface occurrences

Panarctic et al. Sherard Bay F-34 well (4617-4875 m).

Age

Conodont data (A.C. Higgins, pers. comm., 1986) suggest that these beds are of Gzhelian to Asselian age.

Similar pollen and spore assemblages in the upper part of the unnamed equivalent of the Ettrain Formation (southern Eagle Plain) and the lower part of the Jungle Creek Formation (northern Ogilvie Mountains) of northern Yukon Territory (Bamber et al., this volume), were assigned a Gzhelian age, based on brachiopods (Waterhouse and Waddington, 1982).

Weylandites striatus — Protohaploxypinus perfectus Assemblage Zone

This zone is dominated by polyplicate and striate disaccate grains, such as *Vittatina* spp., *Weylandites* spp., *W. striatus* (Luber ex Jansonius) comb. nov. and *Protohaploxypinus* spp., and *P. perfectus* (Naumova) Samoilovich, 1953. Broadly taeniate grains of *Vittatina* and distinctive striate disaccate grains of *Hamiapollenites tractiferinus* (Samoilovich) comb. nov. are common.

Outcrop occurrences

Uppermost Canyon Fiord Formation, Melville Island.

Subsurface occurrences

Eldridge Bay E-79 (920-1238 m), Sherard Bay F-34 (3966-4053 m), Marryatt K-71 (4737-5465 m), Chads Creek B-64 (4600-5035 m), and Drake Point D-68 (4843-5413 m).

Age

In the B.P. et al. Graham C-52 well (3062-3067 m), located in the eastern Arctic on Graham Island, fusulinaceans from the Belcher Channel Formation were assigned a late Asselian to early Sakmarian age (Nassichuk and Wilde, 1977). In Sherard Bay F-34, algae and non-fusilinacean foraminifers from a depth of 4071.4-4116.4 m suggest a Sakmarian age (B.L. Mamet, pers. comm., 1985). In Marryatt K-71, algae and non-fusulinacean foraminifers from 4945.6-4954.6 m and 5450.5-5466.39 m suggest a probable Sakmarian age (B.L.Mamet, pers. comm., 1983) and conodonts from 4707-4724 m suggest a late Sakmarian age (A.C. Higgins, pers. comm., 1983).

In Yukon Territory, similar assemblages occur in the middle and upper parts of the Jungle Creek Formation of the northern Ogilvie Mountains (Bamber et al., this volume) and are possibly of Asselian to Sakmarian(?) age.

Limitisporites monstruosus — Vittatina costabilis Assemblage Zone

This zone contains many species in common with the *Weylandites striatus - Protohaploxypinus perfectus* Assemblage Zone. It also contains specimens of *Vittatina costabilis* Wilson, 1962, *V. subsaccata* Samoilovich, 1953, and *Limitisporites monstruosus* (Luber and Waltz) Hart, 1965.

Outcrop occurrences

Unnamed formation, Blind Fiord, Ellesmere Island, and Belcher Channel Formation, Sabine Peninsula, Melville Island.

The Belcher Channel Formation on Sabine Peninsula is a lateral equivalent of the Unnamed formation of the eastern Arctic (C.H. Henderson, pers. comm., 1988).

Subsurface occurrences

Weatherall O-10 (225-245 m), Eldridge Bay E-79 (829-859 m) Sherard Bay F-34 (3765-3783 m), Marryatt K-71 (4409-4459 m), and Chads Creek B-64 (4481-4572 m).

Age

The conodonts indicate a late Artinskian (Baigendzhinian) age (C.M. Henderson, pers. comm., 1988).

Alisporites insignis — Triadispora sp. Assemblage Zone

This zone contains an extremely varied assemblage of trilete spores, striate and nonstriate, disaccate pollen, polyplicate pollen and occasional monosaccate pollen. The trilete spores include *Apiculatisporis, Calamospora, Convolutispora, Densoisporites, Diatomozonotriletes, Foveosporites, Kraeuselisporites, Leiotriletes, Lophotriletes, Lundbladispora, Neoraistrickia, Nevesisporites, Punctatisporites, Raistrickia, Verrucosisporites*, and *Waltzispora*.

The assemblage includes the polyplicate pollen *Weylandites striatus, Vittatina simplex* Jansonius, 1962, and *Vittatina vittifer* (Luber and Waltz) Samoilovich, 1953; the striate disaccate pollen *Protohaploxypinus perfectus*, P. spp., *Striatoabieites* sp., *Hamiapollenites bullaeformis* (Samoilovich) Jansonius, 1962, and *Lueckisporites* sp.; and the nonstriate disaccate pollen *Vitreisporites pallidus* (Reissinger) Nilsson, 1958, *Triadispora* sp., *Alisporites insignis* (Warjuchina, 1970) comb. nov. Colpate pollen include *Cycadopites follicularis* Wilson and Webster, 1946, and *Marsupipollenites retroflexus* (Luber) Varyukhina, 1971. Monosaccate grains are rare, but include *Florinites luberae* Samoilovich, 1953, and *Cordaitina* sp. In addition, there are a variety of spinose acanthomorph acritarch species (*Micrhystridium* spp. and *Veryhachium* spp.).

Outcrop occurrences

Sabine Bay and Assistance formations and lowest beds of Trold Fiord Formation, Melville and Devon islands.

Subsurface occurrences

Weatherall O-10 (121-156 m), Eldridge Bay E-79 (615-823 m), Sherard Bay F-34 (3687-3783 m), Marryatt K-71 (4020-4400 m), Hecla J-60 (3048-3611 m), Chads Creek B-64 (3810-4392 m) and Drake Point D-68 (3458-4178 m).

Age

The age of the Sabine Bay and Assistance formations is, based on ammonoids, late Artinskian to Roadian (W.W. Nassichuk, pers. comm., 1988). The age of the lowest beds of the overlying Trold Fiord Formation is uncertain, although the middle and upper parts are Late Permian.

Pollen and spore assemblages from the Sabine Bay and Assistance formations and lowest Trold Fiord Formation, are similar to those described from the Ufimian of the Pechora Basin of the Soviet Union by Varyukhina (1971) and Molin and Koloda (1972).

Taeniaesporites sp. Assemblage Zone

The assemblage lacks diversity, although some species continue through from the underlying zone. There is a relatively high proportion of acanthomorph acritarchs (*Michrystridium* spp., *Veryhachium* spp.) and a limited pollen and spore assemblage. *Taeniaesporites* sp. is rare. *Weylandites striatus, Vittatina vittifer, Protohaploxypinus perfectus, Florinites luberae, Kraeuselisporites* sp., and *Diatomozonotriletes* sp. are common.

Outcrop occurrences

Middle and upper Trold Fiord Formation, Sabine Peninsula, Melville Island, and Grinnell Peninsula, Devon Island.

Subsurface occurrences

Eldridge Bay E-79 (445-539 m), Sherard Bay F-34 (2646-3486 m), Marryatt K-71 (3318-3850 m), Hecla J-60 (2604-2994 m), Chads Creek B-64 (2990-3749 m), and Drake Point D-68 (3222-3440m).

Age

These beds in surface sections are dated as Guadalupian (Wordian to Capitanian) using brachiopod data (Thorsteinsson, 1974), and Wordian, using conodont data (Beauchamp et al., this volume).

Tympanicysta stoschiana — Striatoabieites richteri Assemblage Zone

This zone is characterized by the presence of *Tympanicysta stoschiana* Balme, 1980, *Striatoabieites richteri* (Klaus) Hart, *Taeniaesporites noviaulensis* Leschik, 1956, *Gnetaceaepollenites steevesi*, Jansonius, 1962, and *Lundbladispora obsoleta* Balme, 1970.

Outcrop occurrences

Lower Bjorne Formation, Sabine Peninsula, Melville Island.

Subsurface occurrences

Eldridge Bay E-79 (418-424 m), Sherard Bay F-34 (2565-2640 m), Marryatt K-71 (3309-3339 m), Hecla J-60 (2554-2560 m) and Drake Point D-68 (2987-3121 m).

Age

The assemblages are very similar to those described from the Griesbachian of east Greenland by Balme (1980) and Piasecki (1983), and from the Vardebukta Formation of Svalbard (Utting et al., 1988).

PALEOCLIMATIC IMPLICATIONS

Paleoclimatic implications from upper Paleozoic palynological assemblages of the Canadian Arctic are uncertain. Diverse assemblages of the late Viséan (*Murospora aurita - Diatomozonotriletes saetosus* Assemblage Zone) may be assigned to the *Lophozonotriletes* region (Van der Zwan, 1981), which enjoyed warm humid conditions, such as may be found in equatorial to low latitudes. The Disaccate Assemblage Zone of late Bashkirian to Moscovian age probably indicates an arid climate, to which striate disaccate pollen were well suited (Foster, 1979). The dominance of monosaccate grains in the *Potonieisporites novicus* Assemblage Zone of late Moscovian to Kasimovian age is of uncertain significance, although a similar dominance of this suprageneric group occurs in Upper Carboniferous(?) glacial and periglacial rocks of parts of Gondwanaland; for example, the Lower Karoo of Central Africa (Utting, 1976). Arid conditions are suggested by the dominance of striate disaccate and polyplicate pollen in the *Weylandites striatus - Protohaploxypinus perfectus* and *Limitisporites monstruosus - Vittatina costabilis* Assemblage zones, of upper Asselian to Artinskian age. Humid conditions are indicated by the diverse assemblages, which include a variety of trilete spores, of the late Artinskian to Roadian *Alisporites insignis - Triadispora* Assemblage Zone. The less diverse nature of the assemblages in the Wordian *Taeniaesporites* sp. Assemblage Zone may reflect a further climatic change, possibly to more arid conditions. A return to more humid conditions is indicated by the relative diversity of the lowest Triassic assemblages.

THERMAL MATURITY

Estimates of thermal maturity were made from core samples. Cuttings were used where core was unavailable, and when it was relatively certain that the contained palynomorphs were not caved or reworked. The scale used is similar to that of Hunt (1979), Utting (1987) and Utting et al. (in press). The range of maturity varies within the seven wells from a TAI of 2- to 4 (Fig. 3), and this is almost certainly related to depth of burial and proximity of igneous intrusions. Many intervals have relatively low to medium maturity, and have the potential for hydrocarbon generation. Where the maturity is higher (TAI 3 to 4-), there may have been a potential for gas generation, although within a few metres of igneous intrusions, as in the Sherard Bay F-34, Marryatt K-71, Chads Creek B-64, and Drake Point D-68 wells, the thermal alteration may be above the limit for dry gas preservation (TAI 4).

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A biostratigraphic summary based primarily on conodonts of Upper Ordovician to Middle Devonian rocks of southwestern Ellesmere Island and northwestern Devon Island, Canadian Arctic Archipelago†

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Abstract

*A biostratigraphic summary, based primarily on conodonts, of 18 stratigraphic units on southwestern Ellesmere and northwestern Devon islands and environs, is presented. The strata studied range in age from Late Ordovician (upper part of Fauna 12, Richmondian) to Middle Devonian (*costatus *Zone to possibly* australis *Zone, Eifelian).*

The study area falls into two main depositional provinces, the Arctic Platform and the Franklinian Miogeocline. The latter province coincides with the Central Ellesmere Fold Belt, a part of the Franklinian Mobile Belt. The correlative strata in these provinces display considerable litho- and biofacies differences.

Seven sections and one general area were studied, and of these, three are continuous surface exposures. The Vendom Fiord area is covered by composites of shorter sections, whereas most of the localities on Grinnell Peninsula and environs are represented by scattered outcrops. Two sections are from wells on Bjorne Peninsula on Ellesmere Island. In total, 298 separate localities were studied, 209 of which are outcrops and 89 are subsurface sections.

Résumé

Un résumé biostratigraphique, principalement basé sur les conodontes, est présenté pour 18 unités stratigraphiques du sud-ouest de l'île d'Ellesmere, du nord-ouest de l'île Devon et des environs. Les couches étudiées datent de la fin de l'Ordovicien (partie supérieure de la faune 12, Richmondien) au Dévonien moyen (zone à costatus *peut-être jusqu'à la zone à* australis*, Eifélien).*

La région à l'étude se situe dans deux principales provinces de sédimentation, la plate-forme arctique et le miogéosynclinal franklinien. Ce dernier coïncide avec la partie centrale de la zone de plissement d'Ellesmere, qui fait partie de la zone orogénique franklinienne. Les couches corrélatives présentent dans ces provinces des différences considérables au niveau de leurs lithofaciès et biofaciès.

Sept coupes et une région générale ont été étudiées et, parmi celles-ci, trois sont des affleurements de surface continus. Pour procéder à l'étude de la région du fjord Vendom, on a regroupé une série de coupes plus courtes, alors que la plupart des emplacements de la presqu'île de Grinnell et des environs sont représentés par des affleurements épars. Deux coupes proviennent de puits forés dans la péninsule de Bjorne dans l'île d'Ellesmere. Au total, 298 emplacements ont été étudiés, dont 209 correspondent à des affleurements et 89 se révèlent des coupes établies en subsurface.

† Contribution to Frontier Geoscience Program.

INTRODUCTION

Ellesmere and Devon islands are situated in the eastern part of the Canadian Arctic Archipelago (Fig. 1). The study area encompasses the southwestern part of Ellesmere Island and northwestern part of Devon Island, including Grinnell Peninsula (NTS 49 C, D, E; 59 B; 69 A). Three small islands, Crescent, Hyde Parker, and Spit (Kate), located immediately west of Grinnell Peninsula, have also been included, since results of preliminary sampling have shown that the Devonian rocks there contain conodont faunas not present in other parts of the study area.

Seven stratigraphic sequences were examined, of which three surface sections are continuous (Strathcona Fiord and Bird Fiord on Ellesmere Island; Sutherland River on Devon Island; see Figure 1). The two sequences in the vicinity of Vendom Fiord are composites of shorter sections. Most of the samples from Area 4, covering the southern part of Grinnell Peninsula and environs, are from scattered outcrops, although some can be placed in an approximate sequential order. Two successions are subsurface, from wells on Bjorne Peninsula on Ellesmere Island.

The stratigraphic units studied here range from Late Ordovician to Middle Devonian in age. The units include the type sections of the Devon Island (Section 3), Sutherland River (Section 3), Prince Alfred (Section 3), and Bird Fiord (Section 2A) formations. The Blue Fiord Formation at the Bird Fiord section (Section 2A) and the Douro Formation at Sutherland River (Section 3) were measured and sampled close to their type sections.

Most of the samples were collected by the author. Important collections were made by R.A. Roblesky from the Vendom Fiord area on Ellesmere Island, and by U. Mayr from Sheills Peninsula, located in the southern part of Grinnell Peninsula on Devon Island. Spot samples from elsewhere on Grinnell Peninsula and from three small islands to the west of the Peninsula, were collected by J.W. Kerr. A single, but highly significant, sample from the upper part of the Devon Island Formation from near the junction of Vendom and Baumann fiords was made available by R. Thorsteinsson.

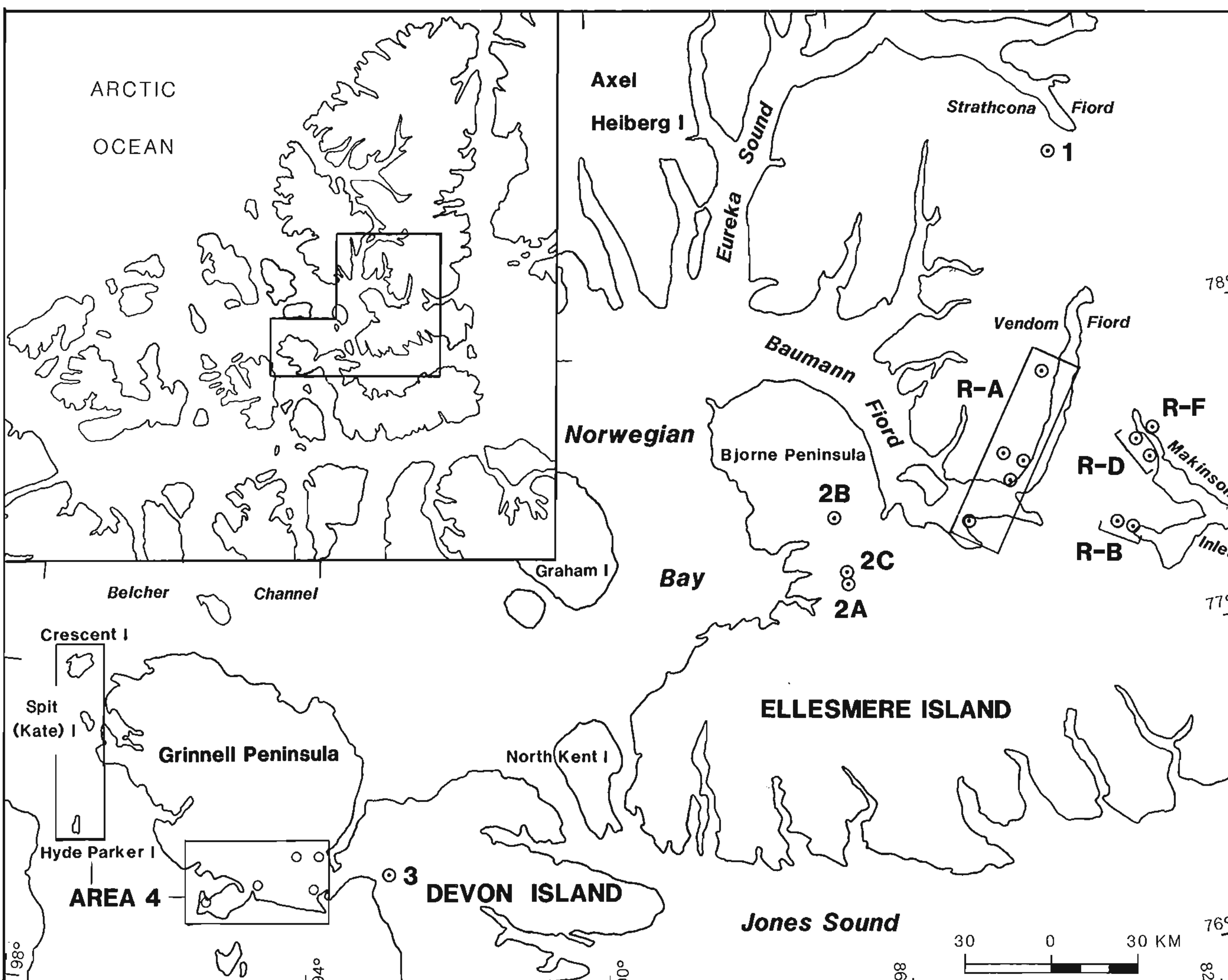

Figure 1. Location map. Circles with dots indicate sections; open circles represent isolated outcrop localities. Note: R-A, R-B, R-D, R-F are composite sections; 2B and 2C are subsurface sections; Area 4 comprises southern Grinnell Peninsula and three small islands.

Conodonts from two wells located on Bjorne Peninsula in southwestern Ellesmere Island (Fig. 1, Sections 2B and 2C), Panaractic Tenneco et al. Eids M-66 and Panarctic ARCO et al. Blue Fiord E-46, were obtained from bulk cuttings. The samples were provided by Panarctic Oils Ltd., Petro-Canada Exploration Inc., and Tenneco Oil and Minerals, Ltd.

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Biostratigraphic analyses from parts of the sequences summarized here were made possible by identifications of palynomorphs by D.C. McGregor, and of megafossils by B.S. Norford, A.W. Norris and R. Thorsteinsson, all of the Geological Survey of Canada. Stratigraphy of the two Bjorne Peninsula wells is by U. Mayr, Geological Survey of Canada, and of the Vendom Fiord area by R.A. Roblesky, now with Shell Canada Ltd., Calgary. I am especially grateful to R. Thorsteinsson for sharing with me his insights into the complexities of Arctic stratigraphy.

REGIONAL GEOLOGICAL SETTING

The sedimentary rocks of southwestern Ellesmere Island and northwestern Devon Island are situated principally within the Franklinian Mobile Belt (Fig. 2). East of Vendom Fiord, Composite Sections R-B, R-D, R-F are located within the Arctic Platform. Preliminary structural cross-sections at and close to the localities studied here were presented by Okulitch (1982).

Of the three major parts of the Franklinian Mobile Belt, the area dealt with in the present study lies entirely within the Central Ellesmere Fold Belt. It is a southeasterly cratonic shelf (miogeocline) consisting predominantly of Cambrian to Ordovician shelf strata, but includes Silurian to Lower Devonian deep-water sediments, and Middle and Upper Devonian nonmarine clastics. The Arctic Platform includes the terrane where the Precambrian basement is overlain by sedimentary strata of early Paleozoic and Mesozoic-Paleogene age. The lower Paleozoic sequence ranges in age from Early Cambrian to Late Devonian in the north, and from Middle Ordovician to Early Silurian in the south. To the north and northwest, the lower Paleozoic sediments are contiguous with folded strata of the Franklinian Mobile Belt and form a westerly and northwesterly dipping homocline (Goodbody, 1987).

BIOSTRATIGRAPHY BY INDIVIDUAL SECTIONS AND AREA

The reader is referred to Table 1 for a graphic presentation of the following summary.

Section 1: Strathcona Fiord area

The section is located 13 km southwest of the head of Strathcona Fiord. The lowest unit sampled, the Irene Bay Formation (84 m) and the lower part of the overlying Allen Bay Formation (293 m) yielded conodonts of Fauna 12 (mid-Maysvillian to Richmondian, Upper Ordovician). Fauna 13 may possibly be represented in the lower Allen Bay with the occurrence of *Panderodus* cf. *P.* n. sp. A of McCracken and Barnes (1981). The highest beds of the Allen Bay Formation contain conodonts of the *celloni* Zone (Telychian, upper Llandovery).

The lower part of the Cape Phillips Formation (547 m) is assignable to the *patula* Zone (Wenlock). The remainder of the formation cannot be accurately dated with conodonts,

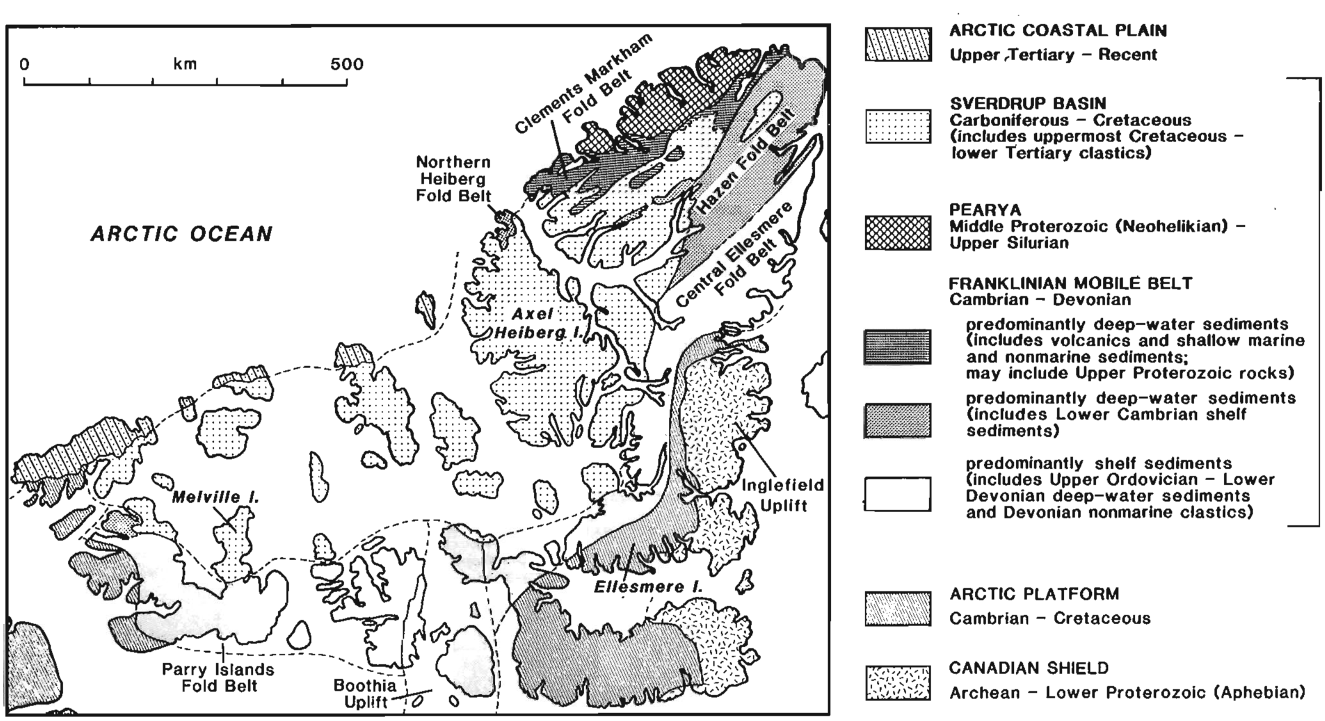

Figure 2. Stratigraphic-structural framework of the Arctic Islands and index. (After Trettin, 1987, Fig. 1.)

Table 1. Correlation charts for Devon and Ellesmere islands.

System	Series		Stages		Conodont zones/faunas		DEVON ISLAND AREA 4: Grinnell Peninsula and environs	DEVON ISLAND SECTION 3: Sutherland River and environs
DEVONIAN	MIDDLE		Givetian		varcus			
			Couvinian	Eifelian	ensensis			
					kockelianus		?	
					australis		Strathcona Fiord	
					costatus		Bird Fiord	
					patulus	partitus	Du5 Undivided Devonian Carbonates	
	LOWER		?	Emsian		patulus		
			Dalejan		serotinus		?	
					inversus		Du1	Blue Fiord
			Zlichovian		gronbergi		?	
					dehiscens		Disappointment Bay	
			Pragian	? Siegenian	kindlei		Cape Phillips?	
					sulcatus		unnamed formation	
			? Lochkovian	Gedinnian	pesavis			? Prince Alfred
					delta			
					eurekaensis			Sutherland River
					hesperius			
SILURIAN	UPPER	PRIDOLI	(stages not yet formally established)		eosteinhornensis			Devon Island
		LUDLOW	Ludfordian		crispa			
					latialata			
					siluricus			Douro
			Gorstian		ploeckensis			
					crassa			
	LOWER	WENLOCK	Homerian		sagitta			
			Sheinwoodian		patula			
					amorphognathoides			
		LLANDOVERY	Telychian		celloni			
			Aeronian		staurognathoides	kentucky-ensis		
					discreta/deflecta			
			Rhuddanian		nathani			
ORDO-VICIAN	UPPER		Gamachian		Fauna 13	?		
			Richmondian		Fauna 12	ordovicicus		

Table 1. (cont.)

ELLESMERE ISLAND

Conodont zones/ faunas	SECTION 2A: Bird Fiord area	SECTION 2C: Bjorne Peninsula, Blue Fiord E-46 well	SECTION 2B: Bjorne Peninsula, Eids M-66 well	COMPOSITE SECTION R-A: west of Vendom Fiord (including Sor Fiord)	COMPOSITE SECTIONS R-B,D,F: east of Vendom Fiord	SECTION 1: Strathcona Fiord
	?	Present day	unconformity	?	?	?
	Strathcona Fiord			Strathcona Fiord	Strathcona Fiord	Strathcona Fiord
	Bird Fiord					
	Upper mbr. / Lower mbr. Blue Fiord			Upper mbr. / Lower mbr. Blue Fiord	Blue Fiord	Blue Fiord
	Eids	Eids	Eids	Eids	Vendom Fiord	Vendom Fiord
					? ?	
	Cape Phillips ↓ ?	Cape Phillips	Cape Phillips	Devon Island	Allen Bay- Read Bay/ Goose Fiord Carbonates ↓ ?	"Eids" (lower limestone mbr.)
				Cape Phillips		Cape Phillips
		Allen Bay- Read Bay Carbonates (fault contact?)				
		Allen ? ? Bay (fault contact?)		Allen Bay- Read Bay Carbonates ↓ ?		Allen Bay
						?
		? Ordovician carbonates	Allen Bay (lower mbr.)			
			Irene Bay		Irene Bay	Irene Bay

although assignments of *siluricus?* and *eosteinhornensis?* zones (Upper Silurian) and *hesperius* to *delta* zones (Lochkovian, Lower Devonian) are possible. The upper part of the formation yielded graptolites of the Zone of *Monograptus praehercynicus* (mid-Lochkovian). Conodonts of the overlying "Eids" Formation (152 m) can only be dated as Lochkovian. The associated brachiopods belong to the *Gypidula pelagica* Fauna of Lenz (1966), of early Lochkovian age. (The stratigraphy of Section 1 is currently being revised by R. Thorsteinsson; the upper part of the Cape Phillips Formation and the lower limestone member of the "Eids" Formation will be incorporated into the Devon Island Formation.)

The lower part of the succeeding Vendom Fiord Formation (550 m) yielded conodonts of early Emsian age (*dehiscens* to *gronbergi* zones); however, the associated brachiopods suggest that the *dehiscens* Zone assignment is more likely. An unconformity representing about three conodont zones thus separates the "Eids" Formation from the Vendom Fiord Formation.

The overlying Blue Fiord Formation (472 m) yielded sparse conodont faunas that lack zonally diagnostic polygnathids, and in this regard are similar to those from the east side of Vendom Fiord, in strata deposited on the Arctic Platform. This is in contrast to the Blue Fiord conodonts from the Bird Fiord area (2A), which inhabited a different biofacies. An Emsian age for the lower part of the formation is indicated by conodont data, whereas spore data from the upper part suggest proximity to the Emsian-Eifelian boundary. Spore flora from the lower part of the succeeding Strathcona Fiord Formation (708 m) of the Okse Bay Group is of early Eifelian age.

Composite sections R-B, R-D, and R-F: east of Vendom Fiord

The upper part of the Allen Bay-Read Bay/Goose Fiord carbonates (unmeasured) at Section R-D yielded only long-ranging species, and may be as old as the *patula* Zone (Wenlock). In some areas, the unit is disconformably overlain by the Vendom Fiord Formation (204 m, R-D; 57 m, R-B), but the exact hiatus represented by this gap is unknown. The upper part of the Vendom Fiord Formation and the lower member (218 m, R-B) of the Blue Fiord Formation can only be dated as Emsian. The upper member of the Blue Fiord Formation (110 m, R-B) and the lowest part of the overlying Strathcona Fiord Formation (287 m, R-D) contain *Pandorinellina expansa* Uyeno and Mason, which ranges from the *inversus* to *costatus* zones (mid-Emsian to early Eifelian).

Composite section R-A: west of Vendom Fiord

The base of the Devon Island Formation (771 m) is late Ludlow in age, as shown by the presence of the graptolite *Monograptus bohemicus tenuis* Boucek. A short distance above it are graptolites of the zone of *M. ultimus*, the basal zone of the Pridoli. The middle part of the formation yielded conodonts of the *hesperius* to *delta* zones, accompanied by graptolites of the *M. uniformis* Zone (basal Lochkovian). The highest beds of the formation from a site near the junction of Vendom and Baumann fiords can be dated as *dehiscens* Zone (lower Emsian).

Only the higher beds of the Eids Formation (766 m) were sampled for conodonts. These beds can be assigned to the *dehiscens* Zone (lower Emsian), and this dating puts narrow time constraints on the age span of the entire formation.

The lowest beds of the lower member (approx. 900 m) of the Blue Fiord Formation yielded brachiopods of Early Devonian age. Conodonts from about the middle of the lower member to the middle of the upper member (358 m) are of the *inversus* Zone (mid- Emsian). This interval is followed immediately by representatives of the *serotinus* Zone in the upper beds of the upper member.

The lower beds of the Strathcona Fiord Formation (550 m) yielded *Pandorinellina expansa*, as did the correlative beds east of Vendom Fiord. These beds may, therefore, be as young as the *costatus* Zone (lower Eifelian).

Sections 2B and 2C: Bjorne Peninsula wells

The two wells analyzed here, Panarctic Tenneco et al. CSP Eids M-66 and Panarctic ARCO et al. Blue Fiord E-46 (hereafter referred to as the Eids (Section 2B) and Blue Fiord (Section 2C) wells, respectively), are located in the southern part of Bjorne Peninsula in southwestern Ellesmere Island. Both wells were spudded in the Eids Formation; the former penetrated a sequence ranging from the Eids through Copes Bay formations (the latter being Lower Ordovician) to a total depth of 11 000 ft (3352.8 m), whereas the Blue Fiord well went down to a depth of 7683 ft (2341.8 m), bottoming in fault repetitions of the Thumb Mountain Formation.

In the Eids well, the upper part of the Allen Bay Formation (lower member; 148.1 m) yielded conodonts of Fauna 12 (Upper Ordovician). This is unconformably overlain by the Cape Phillips Formation (330.7 m), the lower parts of which are assignable to the *staurognathoides* (?) and *celloni* zones (upper Llandovery). The remainder of the Cape Phillips Formation, and all of the Eids Formation (272.8 m) were either barren of conodonts or produced only indeterminate fragments. Palynomorphs from the upper beds of the Eids Formation, however, suggest an early to mid-Emsian age (Uyeno, in press [see page 6]).

In the Blue Fiord well, the Allen Bay Formation (164.3 m) and the lowest part of the overlying Allen Bay-Read Bay carbonates (981.1 m) are questionably assignable to the *staurognathoides* Zone. Within the lower parts of the carbonates, too, conodonts of the *celloni* Zone, *celloni* to *amorphognathoides* zones (upper Llandovery-lower Wenlock) occur, and an interval near the top of the unit can be dated as early Ludlow. No conodonts were recovered from the succeeding Cape Phillips Formation (187.8 m). The Eids Formation (669.0 m) can only be dated as Emsian, but based on the age of the formation at Composite Section R-A, it is probably of early Emsian age.

Section 2A: Bird Fiord area

Conodonts of the Blue Fiord Formation (1261 m) can be assigned to the *dehiscens* through *serotinus* zones (Emsian; Uyeno and Klapper, 1980). The overlying Bird Fiord Formation (853 m; type section), cannot be so confidently dated, since the zonally significant polygnathids are virtually absent. With the shift in lithofacies from carbonates to clastics came a corresponding change in the biofacies. The occurrences of *Pandorinellina expansa* at the level of 157 m below the top of the Bird Fiord Formation, however, indicates an age no younger than the *costatus* Zone. The very top of the formation, therefore, may possibly range to the base of the succeeding *australis* Zone.

Section 3: Sutherland River and environs

(In addition to the Sutherland River succession in northwestern Devon Island, this section includes an area located immediately northwest of Ptarmigan Lake and about 7 km northwest of Sutherland River.)

The Douro Formation (186 m; type section) of the Read Bay Group can be assigned to the *siluricus* Zone (mid-Ludlow). The overlying Devon Island Formation (152 m, type section; 190 m at Sutherland River) ranges in age from late Ludlow (*latialata* Zone) to early Lochkovian (*hesperius* Zone; Thorsteinsson and Uyeno, 1981). Conodonts from the upper half of the Sutherland River Formation (116 m; type section) indicate a range from *hesperius* to *delta* zones (Lochkovian). A single sample from the Prince Alfred Formation (52 m; type section) did not yield any conodonts. In view of the highly clastic lithology, primarily of fluvial quartzose sandstone, this finding is not surprising. The Blue Fiord Formation (139 m exposed) is clearly assignable to the *inversus* Zone (mid-Emsian).

Area 4: southern Grinnell Peninsula and environs

The area includes three small islands: Crescent, Spit (Kate), and Hyde Parker islands, located immediately to the west of the Peninsula.

An unnamed formation was sampled on Hyde Parker and Crescent islands. The unit ranges from the *delta* to *dehiscens* zones (mid-Lochkovian to lower Esmian) and is, therefore, age-equivalent in part to the Disappointment Bay Formation (*dehiscens* to *gronbergi* zones) on Crescent Island and to a unit that is tentatively referred to the Cape Phillips Formation (*kindlei* to *dehiscens* zones) on Spit (Kate) Island.

On southern Grinnell Peninsula, ten samples from the Undivided Devonian Carbonates (of Morrow and Kerr, 1977, 1986) were analyzed for conodonts. The age ranges from the *inversus* Zone (mid-Emsian) for the lowest member, Du1, to *costatus* Zone (lower Eifelian) for the highest, Du5. After the study on which this summary is based was essentially complete (Uyeno, in press), members Du1 and Du2 were assigned to the Blue Fiord Formation and members Du3 to Du5 to the Bird Fiord Formation (Goodbody et al., 1988). The age of the overlying Bird Fiord Formation (sensu stricto) is based on an undescribed collection, which is possibly indicative of the *australis* Zone.

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Palynostratigraphy of the Esso et al. Issungnak O-61 well, Beaufort Sea†

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Abstract

This paper contains a palynological analysis of the Esso et al. Issungnak O-61 well in the Beaufort Sea, Arctic Canada. At a depth of 160 to 180 m, the age of the strata is probably no younger than middle Pliocene. No age diagnostic palynomorphs occur at the base of the well (3583 m), but the presence of the foraminifer Turrilina alsatica *at 2730 m indicates an Oligocene age. This suggests an Oligocene to Recent stratigraphic range for the pollen taxon* Symphoricarpos *sp. cf.* S. occidentalis. *Within the Mackenzie Bay sequence there is a hardwood assemblage that includes several likely temporate arboreal taxa. The upper limits of these taxa occur between 1505 and 1675 m. This possibly represents a Middle Miocene temperature maximum and subsequent marked decline. However, there is a distinct possibility that the assemblage is made up of reworked Paleogene taxa.*

Résumé

La présente étude renferme une analyse palynologique du puits Issungnak O-61 d'Esso et coll. dans la mer de Beaufort, dans l'Arctique canadien. À une profondeur de 160 à 180 m, l'âge des couches remonte au moins au Pliocène moyen. Aucun palynomorphe symptomatique d'un âge ne se trouve à la base du puits (3583 m), mais la présence du foraminifère Turrilina alsatica *à 2730 m situe l'âge à l'Oligocène. Cela situe la distribution stratigraphique du taxon pollinique* Symphoricarpos *sp. cf.* S. occidentalis *dans la période s'étendant de l'Oligocène à l'Holocène. Dans la séquence de la baie du Mackenzie, il y a une association de bois dur qui renferme plusieurs taxons d'espèces arborescentes de zone probablement tempérée. Les limites supérieures de ces taxons se situent entre 1505 et 1675 m. Cela correspondrait à un maximum de température suivi d'une baisse marquée au Miocène moyen. Cependant, il y a une nette possibilité que l'association soit composée de taxons remaniés du Paléogène.*

† Contribution to Frontier Geoscience Program.

INTRODUCTION

The Issungnak O-61 well (latitude 70.01679°N, longitude 134.31331°W, GSC locality C-86839) is an abandoned oil and gas well in the Beaufort Sea (Fig. 1). The well penetrates the Iperk, Akpak, Mackenzie Bay, and Kugmallit sequences (Peach, 1987), and has a total depth of 3583 m.

Sixty-eight palynological samples were analyzed from cuttings representing 20 m intervals, at 50 m increments, from 40-60 m to 3555-3575 m. Palynological preparations were made by standard techniques. In most slides, palynomorphs were too sparse to be statistically useful; the data were plotted only by presence (Fig. 2).

Only palynomorphs that were reasonably well preserved and of similar, relatively low, thermal maturity were included in the tally. Palynomorphs that were badly eroded were ignored. Dinoflagellates were rare and taxa were represented by one or few, mostly indistinct specimens.

Taxonomy of upper Tertiary floras can be based on modern botanical nomenclature, fossil morphological schemes, or some combination of the two. This report, with few exceptions, uses modern botanical nomenclature if a pollen type can be confidently related to a modern taxon. This usage is particularly appropriate for the Neogene. Identifications of modern and Upper Quaternary pollen are commonly limited to family or genus level, and this guides the level of precision in identifications of older palynomorphs.

RESULTS AND INTERPRETATION

Figure 2 shows the stratigraphic distribution of selected taxa, arranged by highest appearance. In the interpretation of these results three major questions were considered: 1) the age at the top of the well; 2) the age at the bottom of the well, and; 3) whether the assemblage with upper range limits between 1505 and 1675 m has chronological and paleoclimatic significance.

Pliocene

Up to five per cent *Tsuga* pollen, plus *Tsuga* macrofossils, occur in the Miocene- Pliocene Beaufort Formation on Banks Island (Matthews et al., 1986). However, above the Beaufort Formation, *Tsuga* pollen is rare and might be recycled, as recycled pollen of temperate hardwoods has been recognized in the Morgan Bluffs Formation (Matthews et al., 1986; Matthews 1987). *Tsuga* pollen is absent from the Upper Pliocene/Lower Pleistocene Gubik Formation of the north slope of Alaska (Nelson and Carter, 1985). Wolfe (1972) indicated that *Tsuga* had disappeared from western Alaska by the Upper Pliocene. Therefore, sample P2511-03 is probably not younger than middle Pliocene. Taxa occurring above the top of the *Tsuga* range in the Issungnak O-61 well could occur in the Pleistocene, but are not limited to it.

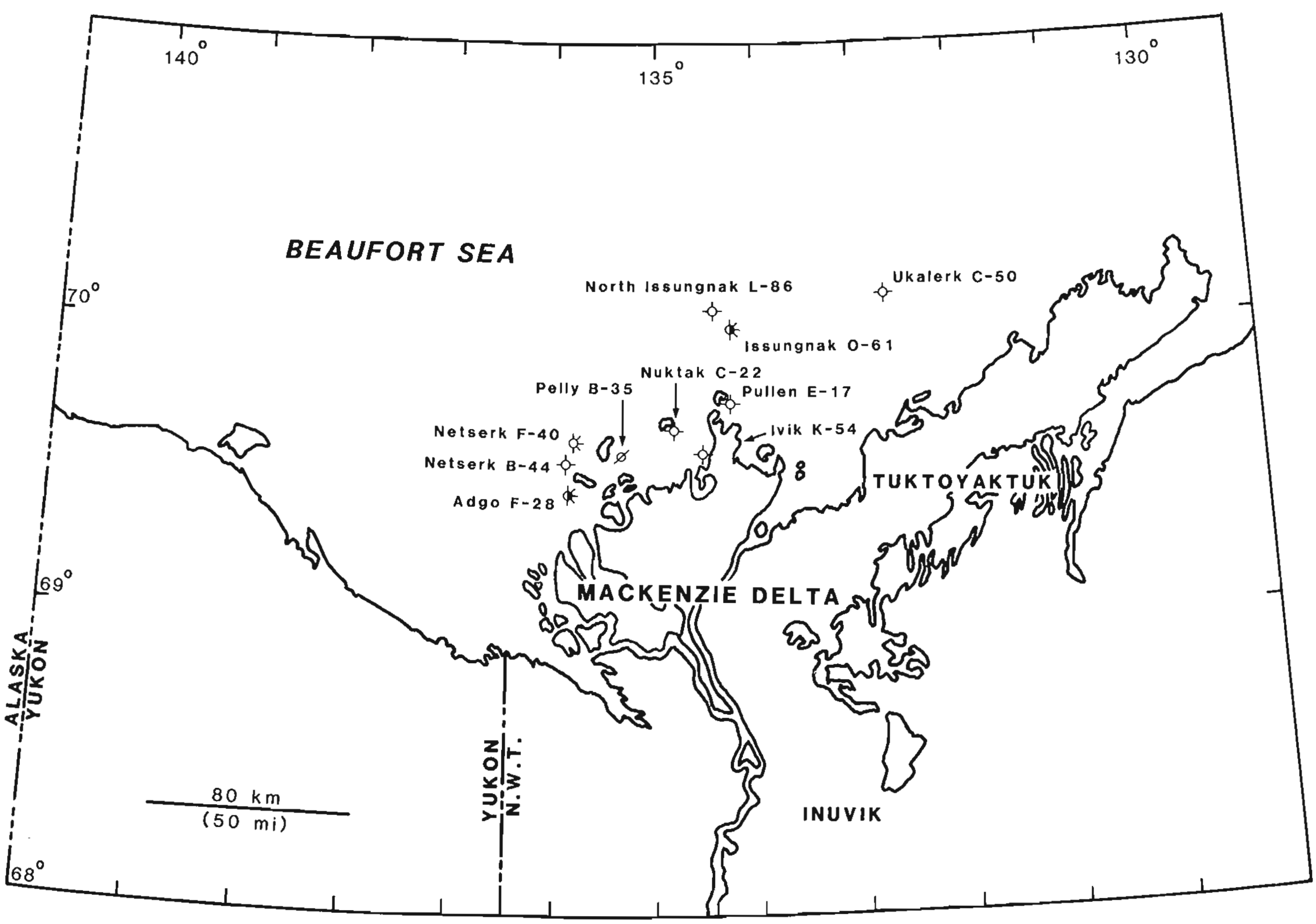

Figure 1. Location of wells in the Beaufort Sea.

Oligocene

Palynomorphs that occur at the base of the Issungnak O-61 well are not age diagnostic for this interval. McNeil (pers. comm., 1987) has identified the foraminifer *Turrilina alsatica* Andreae, an Oligocene index fossil, at 2730 m.

The distinctive pollen of *Symphoricarpos* sp. cf. *S. occidentalis* occurs in samples P2511-49 and 64 (2605-2625 m and 3355-3375 m, respectively), within the upper Kugmallit sequence. A Kremp File search indicates that this taxon has been identified rarely. Leopold (1969) reported this species from the Middle Miocene in the Troublesome Formation, Colorado. Wolfe (1969) reported the one macrofossil *Symphoricarpos* species from the Lower and Middle Miocene flora of Alaska, and noted that *S. albus* may have entered America from Eurasia, via Alaska. Axelrod and Ting (1960) reported *S. albus* from the Pliocene of California. Ioannides (unpublished GSC report) has also reported *Symphoricarpos* sp. cf. *S. occidentalis* in Assemblage B (2000-6840 ft/610-2085 m) in the Netserk F-40 well, within the upper Kugmallit and lower Mackenzie Bay sequences.

The occurrence of *Symphoricarpos* sp. cf. *S. occidentalis* below *T. alsatica* in the Issungnak O-61 well suggests an Oligocene age for the *Symphoricarpos* species. It occurs in the upper Kugmallit sequence in the Netserk F-40 and Issungnak O-61 wells, and in the lower Mackenzie Bay sequence in the Netserk F-40 well. However, *S. occidentalis* is an extant species, so this pollen type could appear in Quaternary or Recent assemblages.

Miocene

In the interval 1505 to 1675 m (P2511-27 to P2511-30) numerous pollen types have their last occurrences (Fig. 2). These include *Carya* sp., Ericaceae (including *Ericipites compactipolliniatus* Norris, 1986), Liliaceae, *Pterocarya* sp., *Ulmus/Zelkova*, Aceraceae, *Castanea* sp., *Juglans* sp., *Liquidambar* sp., *Tilia* sp., unknown tricolporates (several taxa), *Pachysandra/Sarcococca* sp. (*Erdtmanipollis* sp.), *Ilex* sp., and *Potamogeton* sp. All of the above families or genera can be found today in eastern North America (Gray and Sohma, 1964; McAndrews et al., 1973), except *Pterocarya*, which is an Asian taxon (Hora, 1986). Nine of these pollen taxa probably represent arboreal species.

The last occurrence of taxa of this assemblage, at 1505 m, is about 75 m lower than the top of the Mackenzie Bay sequence based on seismic stratigraphy (Peach, 1987). It is possible that the top of the Mackenzie Bay sequence is truncated by an unconformity (D. McNeil, pers. comm., 1987).

Two hypotheses can be entertained to explain the occurrence of the assemblage: 1) the pollen is recycled from Paleogene beds, or; 2) the occurrences are in situ, and represent real stratigraphic ranges of the taxa.

The recycling hypothesis

Supporting the first hypothesis is the fact that several of the taxa may be found in Paleocene, Eocene and Oligocene beds in the Beaufort Sea region (Staplin et al., 1976; Rouse, 1977; Ioannides and McIntyre, 1980; Norris, 1986). *Pistillipollenites macgregorii* Rouse, 1962 is considered an Upper Paleocene to Middle Eocene indicator (Rouse, 1977; McIntyre, 1985), although Wiggins et al. (1988) have reported it from the Lower Oligocene of Alaska. Its occurrence in the Issungnak O-6l well at 1855 to 1875 m, although 200 to 300 m below the upper limit of the deciduous hardwood assemblage, lends support to the hypothesis that much of the assemblage is reworked.

The recycling hypothesis does not adequately explain the cessation in the deposition of sediment containing recycled palynomorphs. J. Dixon (pers. comm., 1987) considers that, for the sediments penetrated by the Issungnak O-61 well, there was no change in the southeast source direction. This alone, however, is not conclusive evidence for rejecting the recycling hypothesis.

The in situ hypothesis

The alternative hypothesis is that the taxa represent an in situ occurrence of Miocene or Pliocene flora. This hypothesis cannot be tested, but can be evaluated by circumstantial evidence; that is, other evidence of distinct climatic or floral transition during this period.

The Issungnak O-61 data should iterate patterns found in other Beaufort Sea wells containing Mackenzie Bay sequence sediments (Young and McNeil, 1984). However, the variations in well-reporting formats, differing taxonomic systems, and the difficulty of implementing a formal system of comparison makes this an imprecise procedure. Similarities may exist, but convincingly close comparisons cannot be made between the assemblage and stratigraphic pattern of last occurrences in the Issungnak O-61 well and the following wells: Adgo F-28 (Austin and Cumming, 1977), Ivik K-54 (Austin and Cumming, 1979b), Netserk B-44 (Austin and Cumming, 1979c), Pelly B-35 (Austin and Cumming, 1979d), and Nuktak C-22 (Norris, 1986).

In the Mackenzie Bay sequence in the Netserk F-40 well, sidewall cores from 1194.8 and 1219.2 m (3920 and 4000 ft), contain *Tsuga, Carya*, Gramineae, *Pterocarya, Tilia, Ilex, Ulmus* and *Symphoricarpos* sp. cf. *S. occidentalis*. *Liquidambar* occurs in a sidewall core from 1225.9 m (4022 ft) (Ioannides, pers. comm., 1979). This assemblage is reasonably similar to that found in the Mackenzie Bay sequence in the Issungnak O-61 well.

A strong similarity to the assemblage with last occurrences in the Mackenzie Bay sequence in Issungnak O-61 is found in the Miocene assemblage immediately below the disconformity in Ukalerk C-50 (McNeil et al., 1982). The palynomorphs *Pterocaryapollenites* sp. (called *Pterocarya* sp. in this report), *Ulmipollenites undulosus* Wolff (*Ulmus* sp.), *Ilexpollenites* sp. (*Ilex* sp.), *Caryapollenites* spp. (*Carya* sp.), *Juglanspollenites* sp. (*Juglans* sp.), and the *Tiliapollenites-Bombacacidites* complex (Tilia sp.) occur just below the disconformity.

The North Issungnak L-86 well is about 10 km north-northwest of Issungnak O-61, and a more detailed comparison of their stratigraphy is merited. Seismic stratigraphy

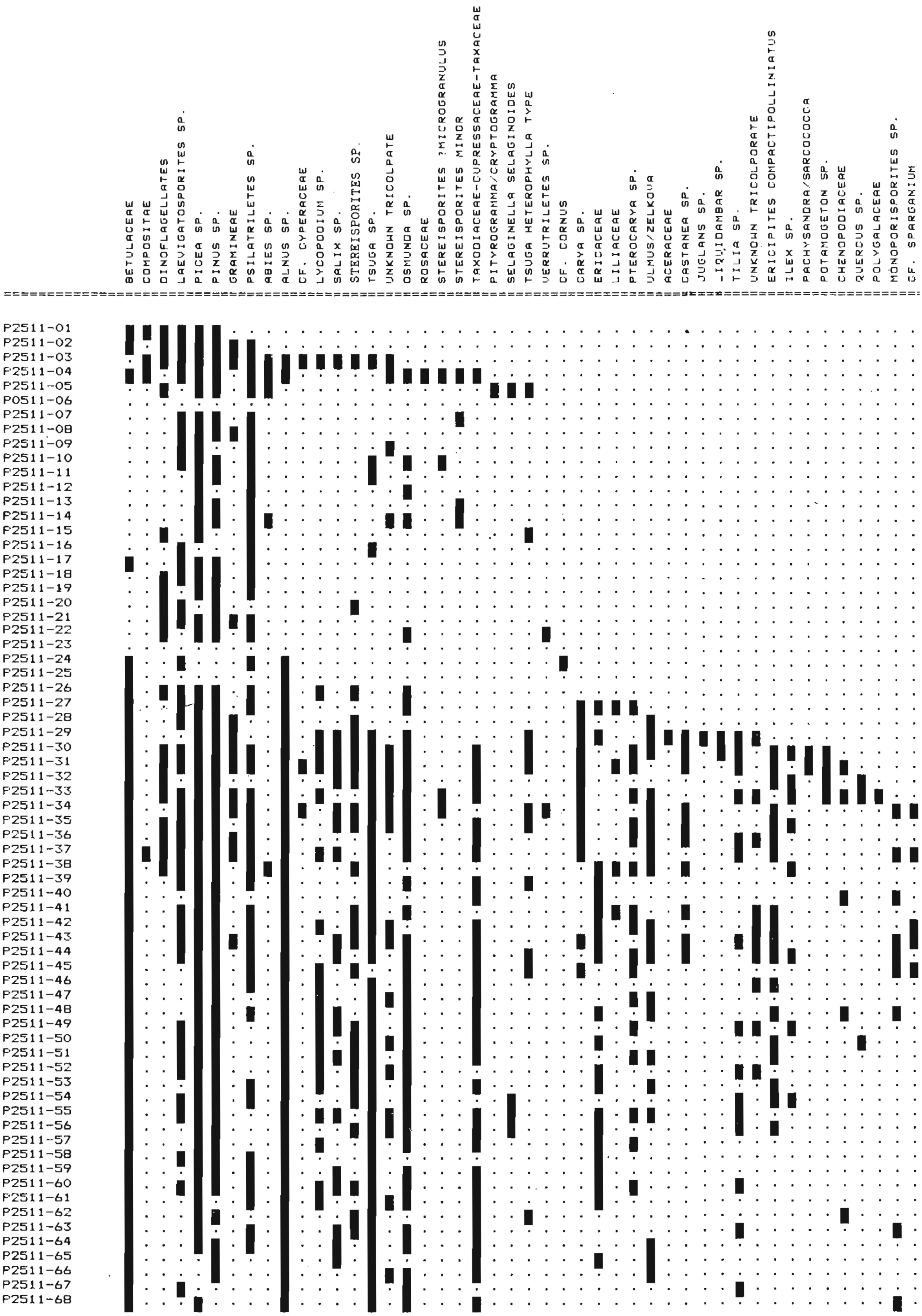

Figure 2. Issungnak O-61 palynostratigraphy; presence/absence of palynomorphs, plotted by stratigraphic tops, with sample numbers, depths and seismic sequence stratigraphy.

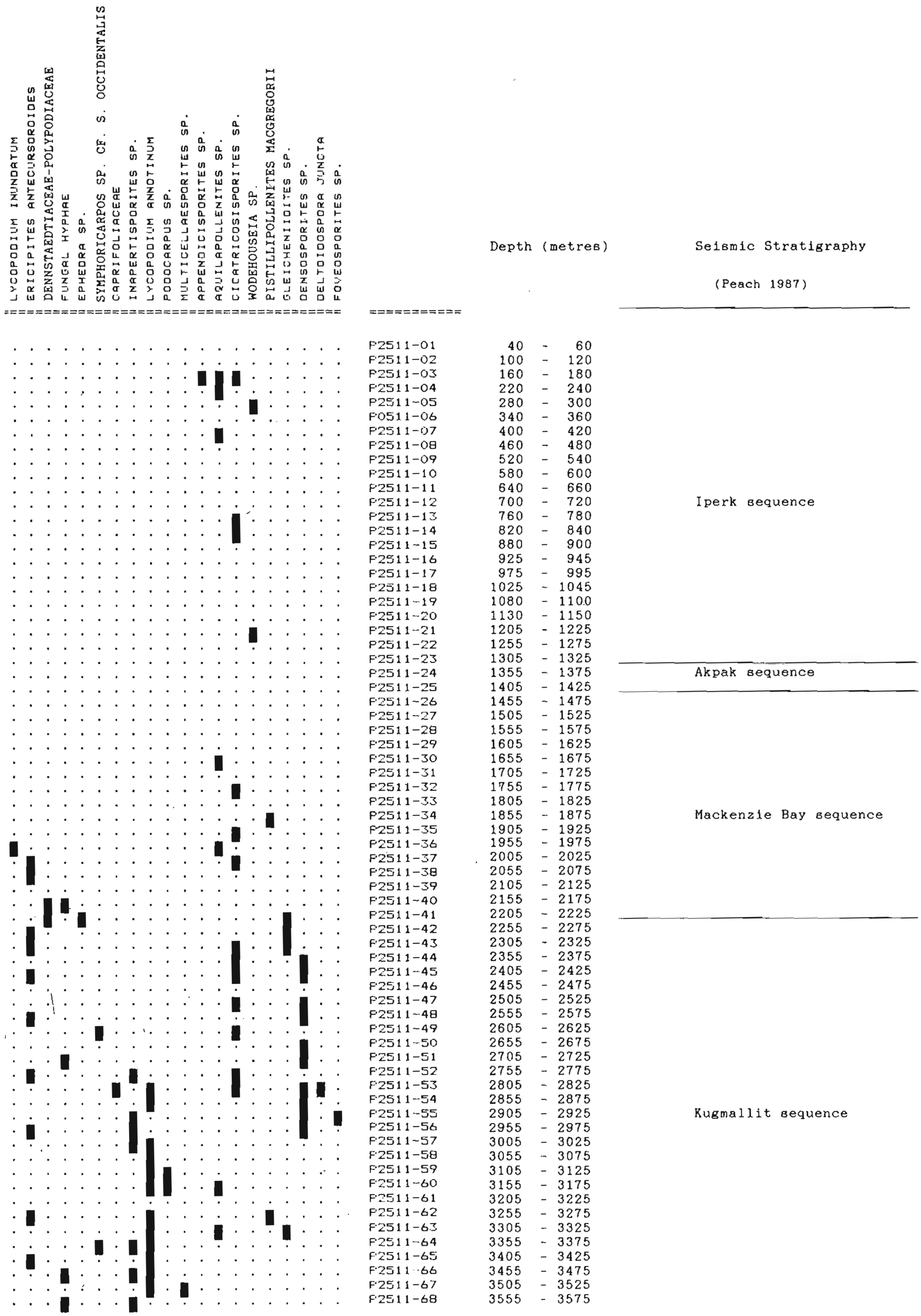

Figure 2. (cont.)

and foraminiferal zonation show that the Akpak/Mackenzie Bay sequence boundary is present, and that the Akpak sequence is much thicker in L-86 than in O-61, possibly due to lack of erosion of the sequence in L-86 (McNeil, pers. comm., 1987). In L-86, the top of the Mackenzie Bay sequence is at a depth of 2392 m (Peach, 1987). This coincides with the boundary between the *Pentadinium laticinctum* and *Systematophora ancyrea* subzones within the *Tsugaepollenites igniculus* Zone, which is interpreted to mark a chronological boundary between the Early Miocene and Middle to Late Miocene (Bujak Davies Group, 1987). However, the details of the palynomorph assemblages are significantly different in the O-61 and L-86 wells.

In L-86, the Mackenzie Bay, Akpak, and Iperk sequences are interpreted from seismic stratigraphy (Dixon and Dietrich, 1985), and from foraminiferal data (McNeil, pers. comm., 1987). Palynological analysis of the well (Bujak DaviesGroup, 1987) shows that at the top of the Mackenzie Bay sequence there is no equivalent to the pattern of last occurrences found in the O-61 well at this boundary, nor is there the diversity of hardwood pollen taxa found in the Mackenzie Bay sequence in the O-61 well. Under either hypothesis for the source of the Mackenzie Bay flora in O-61, recycling or autochthonous occurrence, one would expect greater similarity in the flora of the two wells. Last occurrences in L-86 of taxa that are equivalent to taxa found in the O-61 well range from 2580 m up to 1460 m, or up through the Akpak sequence and into the Iperk; for example: *Polyatriopollenites stellatus* (*Pterocarya* sp.), *Juglanspollenites* sp. A (*Juglans* sp.), *Ilexpollenites margaritatus* (Potonié) Raatz ex Potonié, 1960 (*Ilex* sp.), *Ericipites compactipolliniatus* (same name), *Quercoidites microhenrica* (Potonié) Potonié (*Quercus* sp.), *Caryapollenites simplex* (Potonié) Raatz, 1937 (*Carya* sp.). While erosion or slower deposition of the Mackenzie Bay sequence in the O-61 well could explain stratigraphic compression of the tops, the same diversity of taxa was not observed within the Mackenzie Bay sequence in the L-86 well as in the O-61 well.

Normal faulting has resulted in significant topographic gradients. The top of the Mackenzie Bay sequence is presently approximately 800 m higher in the O-61 well than in the L-86 well, though they are only 10 km apart horizontally (Dixon and Dietrich, 1985). The distribution of hardwoods as far as the Akpak sequence in the L-86 well could have resulted from erosion of sediment from the Mackenzie Bay sequence to the south if fault movement had occurred concurrently with Akpak sequence deposition. Local redeposition of sediment across faults may be as significant as recycling from more distant sources. However, the difference in palynostratigraphic resolution in the two wells remains problematic.

External comparisons and conclusions

The composition of the assemblage and pattern of range tops in the Issungnak O-61 well is similar to two of the other wells considered. It therefore appears conceivable that there is a biostratigraphic pattern and that the in situ hypothesis may be correct. However, the evidence could also indicate similar patterns of sediment recycling, supporting that hypothesis. This lack of a consistent pattern among wells indicates that any interpretation of the composition of this assemblage and its age must be considered tentative.

It is worth considering whether there is any evidence, external to palynology, that would favour either the recycling, or the in situ hypothesis, and assist in assigning an age to the tops.

Plant macrofossil evidence from south and central Alaska may be significant in interpreting Beaufort Sea palynomorphs. A distinct transition occurs from Seldovian to Homerian floral assemblages in the late Middle Miocene (Wolfe, 1972, 1981). Wolfe (1972) described the Seldovian (Lower to Middle Miocene) assemblages as the richest post-Eocene floral assemblages in Alaska. The Seward Peninsula pollen assemblage contains *Carya, Pterocarya, Ulmus/Zelkova*, and *Ilex*, along the dominant Pinaceae. Although Wolfe (1972) included the Pinaceae-dominated Nuwok assemblage of 70°N latitude in the Seldovian flora, it has been determined that this assemblage is Upper Oligocene on the basis of foraminiferal data (McNeil, pers. comm., 1988), and so is not correlative with the assemblage being considered in the Issungnak O-61 well. In contrast to the Seldovian, the Homerian leaf and pollen assemblages are much less diverse (Wolfe et al., 1966). Conifers, especially Pinaceae, dominate Upper Miocene Alaskan vegetation (Wolfe, 1972). With respect to the impoverishment of the temperate deciduous flora, this Seldovian/Homerian floral transition seems closely related to the 1505 to 1675 m last occurrence events in the Issungnak O-61 well.

Independent isotopic paleoenvironmental data are scarce for the northern latitudes in the Pacific. Interpretation of high latitude temperatures relies on the positive correlation between temperatures at high latitudes, where cold water sinks, and the temperature of benthic waters at low latitudes. Ratios of $^{18}O/^{16}O$ in benthic foraminifera indicate that the highest temperatures since the Early Oligocene were in the late EarlyMiocene and early Middle Miocene. This was followed by a sharp temperature drop in the Middle Miocene (Savin, 1977) (Savin et al. 1985). Subsequent research has allowed the recognition of several minor Middle and Late Miocene cooling events (Kennett, 1985; Elstrom and Kennett, 1985). If one accepts the premise that high latitude temperature change is a prime cause of variation in the palynological record, the temperate hardwood assemblage could reflect a relatively warm Early to Middle Miocene climate. Discounting the effects of erosional truncation, it would seem best to attribute the last occurrence of the assemblage to the Middle Miocene high latitude temperature drop, which probably also correlates with Wolfe's (1972) Seldovian/Homerian boundary.

Both the Alaskan macrofossil data and the $^{18}O/^{16}O$ data suggest a distinct floral transition in the Beaufort Sea in the Middle Miocene. Many factors can distort a climatic signal observed through the palynological record and vitiate correlation. No firm conclusions can be reached without further corroborating palynological evidence from other wells, but it appears possible that the assemblage is in place. If so, a Middle Miocene temperature optimum and subsequent decline is the most likely explanation of this palynomorph distribution.

THERMAL MATURITY OBSERVATIONS

Thermal Alteration Index observations were made on the +10 micrometre fraction of kerogen preparations from P2511-22 (1255-1275 m), P2511-26 (1455-1475 m), P2511- 29 (1605-1625 m), P2511-30 (1655-1675 m), P2511-40 (2155-2175 m), P2511-49 (2605- 2625 m), P2511-59 (3105-3125 m), and P2511-68 (3550-3575 m). Fifty-three observations were made on *Triporopollenites, Betula, Alnus, Picea, Lycopodium, Sphagnum*, Compositae, Ericaceae, and Gramineae and exinous fragments. The colour observation standard was that of Pearson (1984). Values obtained were mostly 2 or 2−, with a few 1+ values and values up to 2+ only in thick-walled palynomorphs. Values of 2− or 2 correspond to medium to highly immature values, approximately correlative with vitrinite reflectances just below 0.5 per cent (Pearson, 1984). In comparing the TAI values for each taxon down the well, there appeared to be a slight increase of less than one scale unit in thermal maturity from P2511-22 (Iperk sequence) to P2511-26 (Mackenzie Bay sequence). No trend in thermal maturity was observed within or between the Mackenzie Bay and Kugmallit sequences.

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Seismic reflection profiling from an ice island along the continental shelf of the Canadian Arctic Archipelago†

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Abstract

Seismic reflection profiling has been successfully conducted from a thick tabular ice sheet, an ice island, drifting in the Arctic Ocean. Due to hazards posed by mobile polar pack ice, this method is probably the only practical way to get seismic reflection data for the continental shelf northwest of the Canadian Arctic Archipelago. Field tests have shown that the methods being used by the Geological Survey of Canada are reliable and cost effective and costs may decrease as the techniques are refined.

Seismic reflection data gathered on the inner shelf north of Axel Heiberg Island reveal a thin, presumably Tertiary, succession (about 1 km thick) overlying a basement of presumed Paleozoic age. On the outer shelf, north of Ellef Ringnes Island, the Tertiary strata may be more than 10 km thick.

Résumé

Des profils de sismique réflexion ont été établis avec succès depuis une épaisse nappe de glace tabulaire, une île de glace, à la dérive dans l'océan Arctique. En raison des dangers que pose la banquise polaire mobile, cette méthode est probablement la seule méthode pratique qui permette d'obtenir des données de sismique réflexion pour la plate-forme continentale située au nord-ouest de l'archipel Arctique canadien. Des essais sur place ont montré que les méthodes utilisées par la Commission géologique du Canada sont fiables et rentables et que les coûts pourraient diminuer à mesure que l'on perfectionne les méthodes.

Les données de sismique réflexion recueillies sur la plate-forme intérieure au nord de l'île d'Axel Heiberg indiquent une mince succession (épaisse d'environ 1 km), probablement tertiaire, reposant sur un socle d'âge probablement paléozoïque. Sur la plate-forme extérieure, au nord de l'île d'Ellef Ringes, l'épaisseur des couches du Tertiaire peut dépasser 10 km.

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INTRODUCTION

The geology of the continental shelf to the northwest of the Canadian Arctic Archipelago is poorly known, due to the presence of the permanent but mobile polar ice pack which covers the shelf area and creates formidable logistical problems. In 1983 an excellent opportunity to increase knowledge of the geology of the continental shelf presented itself in the form of an ice island, a tabular block of ice, 7 km by 3 km and 40 m thick. This ice island broke off the Ward Hunt Ice Shelf, on northern Ellesmere Island, near the northeastern end of the continental shelf. By virtue of its size, relatively flat top, and its anticipated southwestward movement along the shelf, the ice island was expected to provide an ideal platform from which a seismic reflection survey and other geological programs could be conducted. This report provides an account of the seismic reflection program of the Geological Survey of Canada on the ice island, which has been in operation since 1984.

Funds for this work were to be provided by the Frontier Geoscience Program and to be managed by the Institute of Sedimentary and Petroleum Geology of theGeological Survey of Canada, Calgary, Alberta.

HISTORICAL REVIEW OF SEISMIC STUDIES OF THE ARCTIC CONTINENTAL SHELF

A brief review of previous attempts to conduct seismic experiments over the Canadian Arctic continental shelf illustrates the importance of conducting this type of work from a drifting ice island. The first known attempts at seismic profiling in the high arctic were undertaken in March 1960 by the newly formed seismic group of the Polar Continental Shelf Project (PCSP). In March, 1961, this group made its first attempt at studying the continental shelf from the shorefast ice off Cape Isachsen, Ellef Ringnes Island, using snowmobiles which had to be manhandled with great difficulty across pressure ridges; instruments and cables similarly had to be manoeuvered into place in rough ice conditions. A single seismic refraction profile resulted from this effort (Hobson and Overton, 1967). In April 1961 an attempt was made to work farther offshore from Cape Isachsen, using a Sikorsky 55 helicopter; this effort also resulted in a single seismic refraction profile (Hobson and Overton, 1967).

At this location the first lessons about the hazards of conducting experiments on mobile pack ice were obtained when, on 30 April, 1961, an emergency signal was intercepted from one of the PCSP single-engine Otter aircraft which had gone down and broken through a freshly frozen lead somewhere offshore north of Meighen Island. The pilot did not have time to radio his position. All available aircraft, including the S-55 helicopter, went on a search and rescue mission. By the morning of May 1 the Otter aircraft had been located and the crew and passengers were evacuated safely to Isachsen; the aircraft was never recovered. At the ocean seismic camp that morning, Overton, returning to the tent housing the other members of the crew, noticed ice cracking under his feet and extending under the tent, and within a large area surrounding the camp. The other members of the crew were awoken, and everyone set about picking up the cables, instruments, and camp gear; some seismometers had already been devoured by pressure ridges which were forming rapidly around the camp. A Mayday signal of the precarious situation was sent before dismantling the radio and antenna. By the time everything was packed and moved to an apparently stable slab of ice, approximately an hour and a half later, there was a large expanse of open water, approximately 100 m diameter where the camp had been; within this short time a large pressure ridge, about 10 m high, had built up nearby. The S-55 helicopter arrived in marginal weather and evacuated all camp personnel and equipment to Cape Isachsen.

Other attempts at seismic profiling on the continental shelf north of Houghton Head, Prince Patrick Island, which were plagued by poor weather and ice conditions, during the spring of 1965 are described by Overton (1970), and in 1966 by Berry and Barr (1971).

Industry interest in hydrocarbon resources resulted in some active seismic programs around 1975. This work was concentrated mainly on the Queen Elizabeth Islands and in the inter-island channels. The only attempts by industry to venture offshore onto the continental shelf were during 1983. This work was done using tracked vehicles, which followed routes marked out by ice reconnaissance crews; they did not venture off the landfast ice north of Ellef Ringnes Island.

Historically, all scientific research on the Arctic Ocean has had to contend with unstable pack ice, except for bathymetric work done by submarines. The pack ice is more or less stable for short periods - enough, for example, for the ice camps for the two months of the Lorex experiment (Weber, 1979), or the less than two months of the CESAR experiment (Weber and Jackson, 1985). The search for such pseudo-stable areas for campsites is in itself a fairly major undertaking. The only really stable features found, namely ice islands, have previously been used for scientific research. Fletcher's ice island, also known as T-3, was one such floating laboratory used by the USA from 1952 to 1974 (Crary et al., 1952; Crary, 1954; Hunkins and Tiemann, 1977) as it circled the Arctic Ocean three times. Another is the present ice island which broke off the Ellesmere ice shelf in 1983 and began its convoluted journey toward the southwest along the Canadian Arctic continental shelf. It has been nicknamed ''Hobson's Choice'' after G.D. Hobson, the then Director of PCSP, whose initiative lead to the establishment of a scientific camp on the platform. Hobson had good reason for his interest; he was party to some of the aforementioned emergencies. Hobson's Choice, abbreviated to H-1 for convenience, is a tabular mass of ice measuring 7 km in length by 3 km in width, and with a variable, undulating thickness of about 40 m. It has moved some 450 km southwest along the continental shelf, approximately parallel to the coastline of the islands, between 13 August 1983 and 4 October 1987 (Fig. 1). En route it has performed some sub-elliptical gyrations. These excursions have added about 50 % to the track, for a total drift path of about 700 km.

THE SEISMIC REFLECTION EXPERIMENT: TECHNIQUES AND ACHIEVEMENTS

Following the discovery of the ice island in 1984, various proposals were considered for scientific research on H-1. The purpose ofthe seismic reflection experiment was to study the structures and stratigraphy of sedimentary rocks in the unexplored regions of Canada's Arctic continental shelf.

For the seismic reflection experiment two proposals appeared viable, one was to use an industry contractor, the other was to use the expertise of the Terrain Geophysics Section of the Geological Survey of Canada (GSC). The latter proposal, submitted by Overton, was to use methods established on previous arctic ocean seismic reflection experiments, but expanded in scale to utilize state-of-the-art, multifold seismic reflection processing techniques. There was some uncertainty as to how well the seismic reflection techniques would work from a thick ice sheet.

Field trials of both methods were conducted on H-1 during September 1984. The most significant difference in method between the two proposals was that the GSC group chose to use the hot water ice drill,developed by the Defense Research Establishment Pacific, to make holes through the ice for hydrophones and seismic sources (Verrall and Baade, 1982). The contract group chose to use a standard, engine driven, shothole auger mounted on rubber tired wheels. The contract group had difficulty clearing the holes of cuttings in drilling through 40 m of ice. The cuttings in the hole above the auger froze together and jammed the cutting head, frequently breaking auger blades. Several times the auger was stuck, frozen deep in the ice. The only way they could free the stem and auger was with the help of the GSC seismic group, using the hot water hole melters. The contract group did finally succeed in augering holes all the way through the ice layer using a common drag bit, but were unable to use the holes for anything but setting hydrophones through the ice into the underlying sea water. This was done with great difficulty using poles to force the hydrophones through the rapidly freezing ice chips in the boreholes. Any attempts to load dynamite were unsuccessful, as the charges disintegrated and the electric detonator leads broke when they were pushed through the freezing ice chips.Also, using

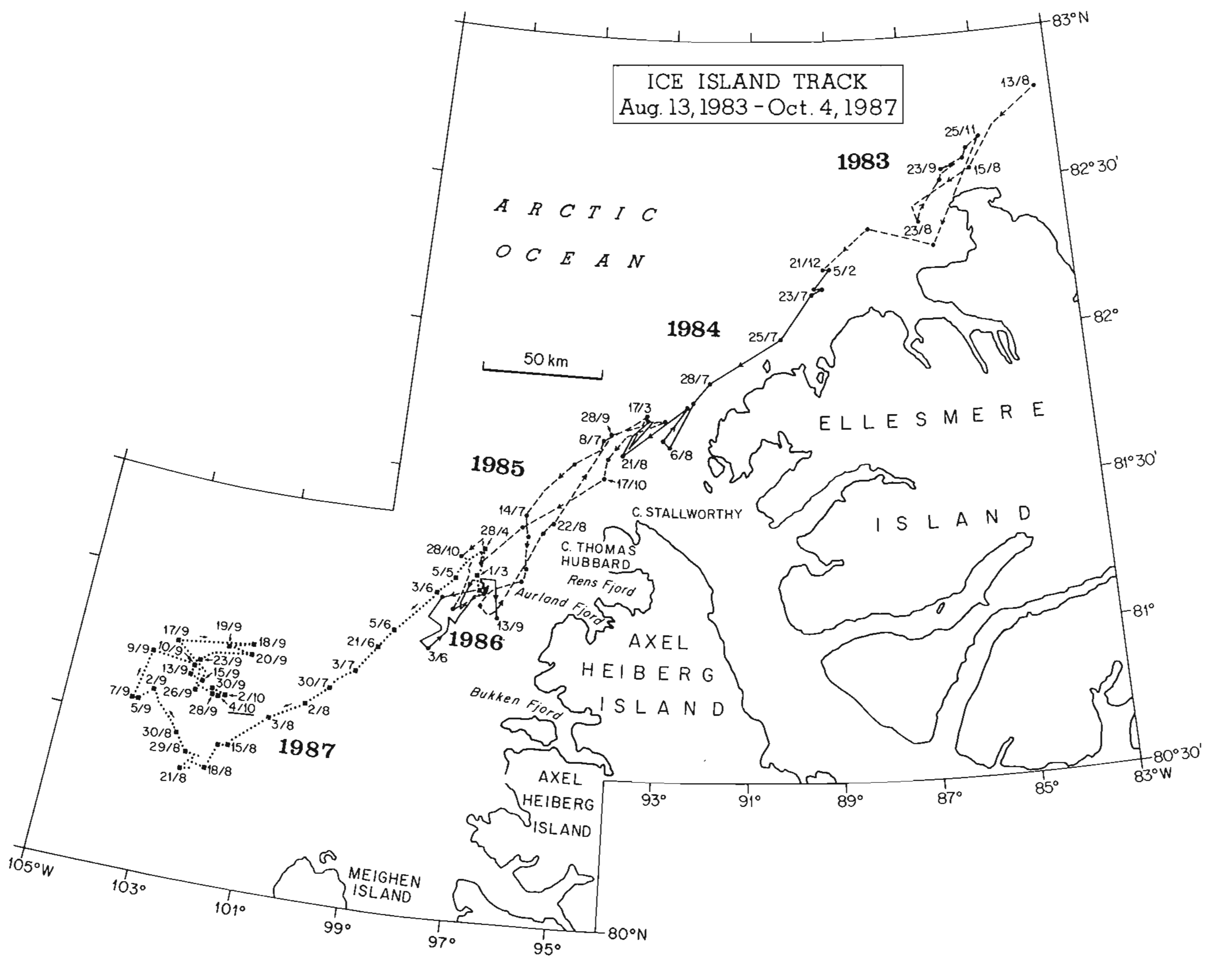

Figure 1. Drift path 1983-1987 (courtesy M. Schmidt, Geo).

these holes for airgun work was out of the question because, at 10 cm diameter, they were too small. Refreezing of the ice cuttings would have prevented airgun work in even larger holes. On the other hand, the melted holes stayed open and usable for several hours, and the contractor used all of the holes melted by the GSC crew, either as hydrophone holes, shot holes, or airgun holes. The hot water drill was proven to be a cost effective, trouble free method of making holes, through 42 m of ice, of diameters ranging from 10 cm to 60 cm. Also, operation of the hot waterdrill does not require the high degree of technical training demanded by mechanical drills. The hot water drill has become the accepted method of making holes through the ice island for runway building as well as for seismic operations.

Finally, both crews succeeded in recording seismic reflection records, establishing that:

1. Seismic reflection was a viable method on a thick icelayer.
2. Seismometers deployed on the ice surface gave superior signals compared with hydrophones in the water under the ice; the hydrophones gave a high frequency, "chattery" response.
3. The GSC approach, using melted holes, was more effective than the contractor's approach, and was also cost effective at an estimated cost of 1/2 to 1/3 that of the contract group.

The drift track for 1984 is shown in Figure 1.

Field season of 1985

The seismic reflection program on H-1 was operated from March 1985 to late August 1985 by the Geological Survey of Canada under the project leadership of Overton, and using a field crew of six students. The crew was rotated, with four in the field and two on leave on an approximate 4 to 6 week schedule.

Under Overton's direction and with the technical assistanceof R.A. Burns the students constructed a small instrument hut to house the seismograph and ancillary equipment. The seismometer array was laid out in the form of a cross, so as to be able to measure reflection dip components, with 107 stations on the main axis of the cross, at 33.3 m spacing. The cross arm of the array was perpendicular to the main arm, and intersected it at station number 76. There were 12 stations on the cross arm spaced at 200m intervals, with six stations on either side of the main axis. Each station consisted of nine Mark Products L-25 D, 4.5 Hz, 60 % damped seismometers, with eight equally spaced on a circle of 10 m diameter and one at the centre. These were frozen into the holes drilled 2 m deep into the ice, under the snow cover.

Thirteen shotholes were used with this array; one at the centre of the cross, and 12 spaced evenly between stations 96 and beyond station 107. These shotholes were cased with 4 inch diameter ABS plastic pipe; the casing was filled with a mixture of diesel fuel weighted with 20 % by volume mixture of trichloroethane. This procedure displaced the water out of the casing and kept the holes open for loading geogel charges as required. The 13 shots were fired in rapid succession for every kilometre of movement of the ice island, thus allowing a 13 fold stack on subsurface reflection points. The signals were received on a 132 channel, Sercel 358 seismograph, and recorded on industry standard nine-track magnetic tape. The drift track of H-1 for 1985 is shown in Figure 1. The program had to be terminated near the end of August to allow the students to return to their respective classes.

Field season of 1986

The seismometer and shothole arrays remained intact over the winter. There were some ice blockages in the shothole fluid, introduced by hauling the firing line, wet with sea water, back up the holes. These blockages were melted with 3 kW electric water heating elements operated by a portable 220 V, AC generator.

The field crew this year, again of six people with four in the field and two on rotation, was obtained under contract with a geophysical company in Calgary, with the intention of operating from March until early October, when the onset of total darkness would force a shutdown. Facilities for the seismic reflection project were upgraded; each of the 13 shotholes had shelters built over them, mounted on stilts embedded in the ice to provide stable footings. These shelters were necessary to allow work to continue during blizzards and throughout extreme cold weather. Also a new, larger instrument shelter was built, and the old instrument hut was increased in size and used to house a 7 kW alternator and tool shed.

The drift track of 1986 is shown in Figure 1. It was of limited areal extent which was a disappointment to all concerned, as a much greater drift path had been expected.

Field season of 1987

As a result of the disappointing drift performance of H-1 during 1986, and because of pressure to assign funds to other priorities, the seismic reflection program on H-1 was suspended for 1987. The drift track for 1987 (Fig. 1), however, was spectacular, crossing areas north of Meighen Island where there is thought to be a transition from thin Tertiary sediments to the east, to thick sequences north of Ellef Ringnes Island. No seismic reflection coverage was obtained over this important transitional region.

Field season of 1988

The favourable drift performance of H-1 during 1987, and its location in the geologically interesting area north of Ellef Ringnes Island, reawakened interest in the reflection program for 1988.

Inspection of the shotpoint and seismometer arrays revealed that neither had fared well through the lapse in operations. The seismic cables were broken in several places and buried deep in refrozen meltwater pools. The shotpoint shelters had collapsed and were also frozen into meltwater pools. The whole operation was redesigned and new techniques investigated. Airgun sources, up to 120 cubic inch capacity, were tried, detonating cord sources and geogel charges of different weights and depths were retested. A new cross array, with three arms of the cross consisting of 24 seismic channels each and the fourth arm consisting of 60 channels, was built. Experience had shown that a single seismometer, well planted by freezing deep into the ice layer, gave as good a response as that of the previous nine seismometer pattern; therefore single seismometers were used to simplify logistics and to reduce costs. In addition, a single shothole, near the centre of the cross array was used; this was a much simpler operation than the previous 13 cased hole pattern. This hole was kept from freezing by circulating sea water up the hole, and discharging it, using a submersible pump through a 50 m hose, at a depth of about 18 m below the bottom of the ice layer. During the cold weather, the shothole froze shut periodically, and had to be reamed occasionally. Also, because the surface water in the hole tended to freeze, heat lamps were suspended above the surface. To prevent the seismometer array cables from being buried in ice over the melt season, a series of short posts was frozen into holes augered into the ice along the length of the array and the cables were suspended from the posts to keep them above the ice surface.

A geogel charge size of 1 kg was found to be the most satisfactory source. A shooting depth of 65 m was chosen to optimize signal signature with respect to ghost and bubble pulses. Shots were recorded on a schedule depending on the drift speed, which was monitored using a Navy Navigational Satellite System which tracks polar orbiting satellites. At speeds of above 300 m/h, shots were recorded every 15 minutes — about the maximum rate possible for reloading and preparing for another shot. At speeds of 100 to 300 m/h, the shooting rate was every half hour; below 100 m/h it was every hour.

During the course of the summer the old shotpoint shelters were recovered and dismantled. They were reconstructed into one new shotpoint shelter, two small and one large explosives magazine, and a storage shelter, all built on substantial stilt footings.

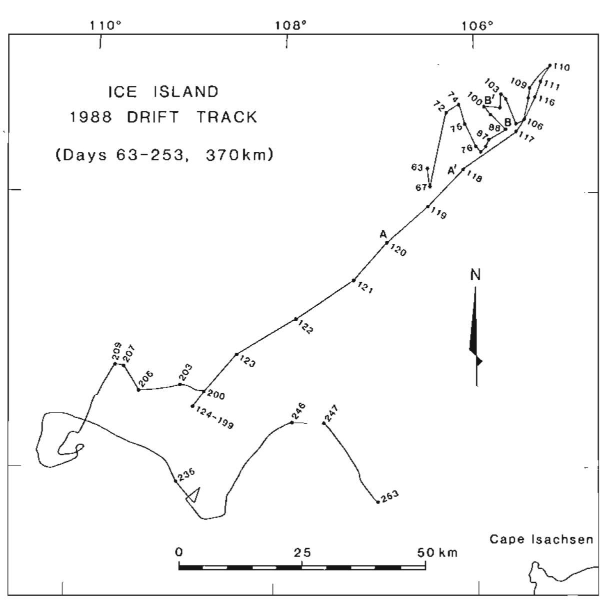

Figure 2. Drift path 1988 (courtesy M. Schmidt, Geo). Dates are given in Julian days.

The field crew of 6 with 4 in the field and 2 on rotation was obtained again under contract with a geophysical company. The season's drift track is shown in Figure 2; the drift track length is about 400 m.

PRELIMINARY INTERPRETATION OF RESULTS

The regional geological framework for the Arctic continental shelf has been interpreted from the geology of the adjacent islands, the assumption that the adjacent Arctic Ocean formed by counterclockwise rotation of Alaska and northeastern Siberia away from the Arctic Islands; the established structural and stratigraphic styles typical of rifted margins; and the established geology of the Beaufort shelf, which is the southward continuation of the Arctic shelf.

The major geological events that shaped the geology of the margin are:

1. Lower Paleozoic sedimentation on the Arctic Platform and Franklinian Geosyncline, with subsequent deformation of the latter, from the Silurian to Early Carboniferous (Ellesmerian Orogeny).
2. Formation of the Sverdrup Basin by rifting, from the Early Carboniferous to Early Permian, followed by thermal subsidence.
3. Opening of the Amerasian portion of the Arctic Ocean by rifting from the Middle Jurassic to the Early Cretaceous, followed by seafloor spreading in the Late Cretaceous.
4. Deformation of the northeastern portion of the Arctic Islands, in the latest Cretaceous to early Tertiary (Eurekan Orogeny).

These events created four major unconformities in the continental margin succession and thus the stratigraphy can be subdivided into five major unconformity-bound megasequences: a) Lower Paleozoic, b) Upper Paleozoic Lower Jurassic, c) Middle Jurassic - Lower Cretaceous, d) Upper Cretaceous Lower Tertiary, e) Upper Tertiary.

Along the inner portions of the shelf, post-rift sequences are relatively thin due to the occurrence of only minor crustal stretching in these areas. Strata underlying the thin rift-drift succession are thought to vary from highly deformed Lower Paleozoic strata in the north, to tilted Upper Paleozoic to Jurassic strata in the central area (between Meighen and northern Prince Patrick Islands), and to tilted Lower Paleozoic to Proterozoic strata in the south (Banks Island to southern Prince Patrick Island). A thick Upper Cretaceous to Tertiary succession is expected on the outer shelf and slope areas where crustal extension during breakup was significant. These areas should also be underlain in part by rift grabens filled with Middle Jurassic to Lower Cretaceous strata which developed during the rifting phase. In the southern and central portions of the shelf the upper Cretaceous to Tertiary succession is likely relatively undeformed. In the north, however, where the effects of the Eurekan Orogeny were felt on the shelf, the Cretaceous to Lower Tertiary strata are probably mildly to moderately deformed and overlain by undeformed Upper Tertiary strata.

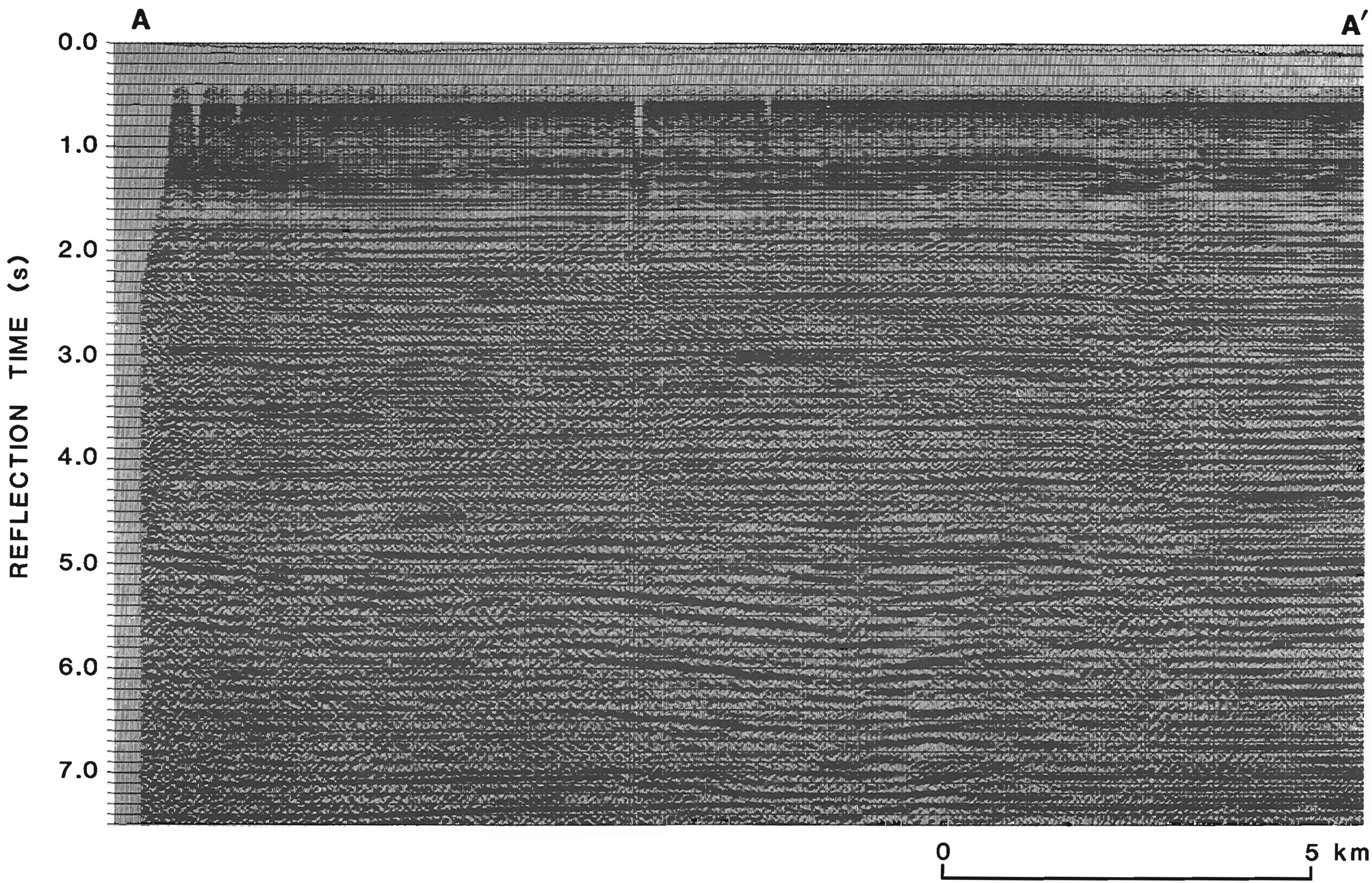

Figure 3. A processed seismic section from 1985, 1986 (after Hajnal and Kesmarky, 1987) showing wedged Tertiary strata, thickening southwest and truncated northeast. Source 1 kg dynamite.

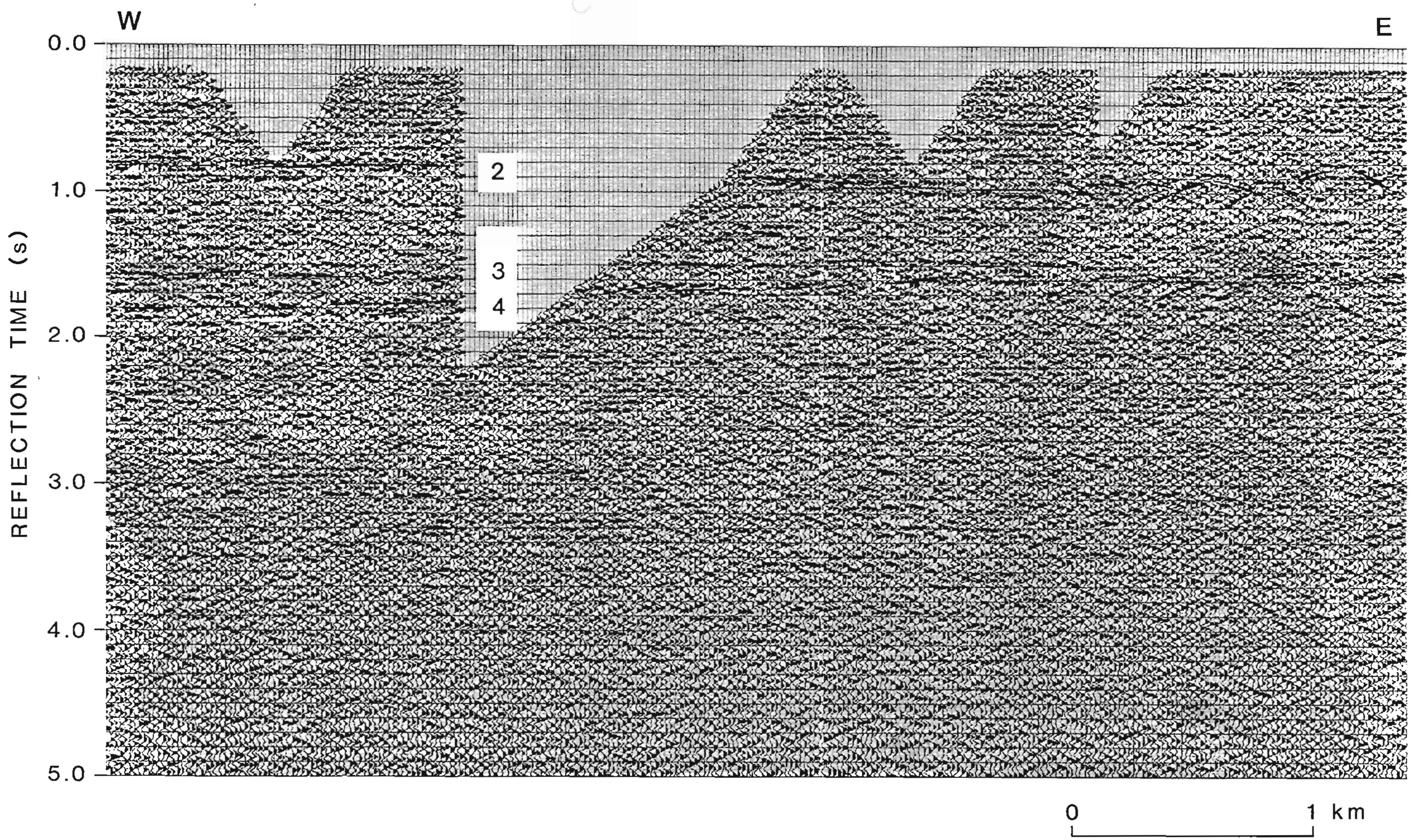

Figure 4. A processed seismic section from 1988 (A-A' in Fig. 2) showing deformed Paleozoic strata overlain by undeformed Tertiary strata. Source 1 kg dynamite.

Sample results from 1985 and 1986 data

The results of the data processing for the 1985 and 1986 field seasons (Fig. 1) are described in detail by Cox (1987), Hajnal and Kesmarky (1987), Burianyk (1988), and Nurse (1988). In summary, limited refraction first-arrival analyses of sedimentary thicknesses in water depths of 200 m or less indicate total thicknesses of less than 1 km of probable Tertiary strata overlying crystalline or deformed Paleozoic basement rocks having velocities of 4.5 to 6.5 km/s. Analysis of reflection profiles is somewhat hampered by the short period reverberations in the shallow water environment. An example is shown in Figure 3. This profile segment is located in Figure 1 near the intersection of the 1985 and 1986 drift tracks northwest of Aurland Fjord. Reflections 2, 3, and 4 are annotated. Reflection 3 is truncated on this profile segment, and also on others cut by a northwest to southeast axis in this area. Interval 3-4 appears to thicken locally toward the southwest and disappears to the northeast. Reflections 2 and 4 are typical for all profiles, and are often evident on the unprocessed field records. Two other reflections, evident on other profiles, represent the seafloor in areas of deeper water; the deepest and least continuous reflection, having an interval velocity in excess of 6 km/s, most likely represents crystalline basement rocks or deformed Paleozoic strata.

Sample results from 1988 data

Figure 4 shows a processed section from the 1988 data set (A-A' in Fig. 2), obtained using 1 kg dynamite charges as the energy source. Coherent reflections are evident as deep as 7.5 seconds reflection time,corresponding to a depth of about 15 km. One of the aims of this work is the identification of primary reflections and their repetitions due to reverberation in the water layer, with the ultimate elimination of the latter phenomena. Some strong reverberations may be seen as the later events of doublets separated by the seafloor reflection interval (the strong event near 0.55 s, for example).

The pronounced trough-like feature near the middle of the section, from 4.8 to 7.5 s no doubt represents real geological structures, possibly associated with deformed Paleozoic-Mesozoic formations. The overlying undeformed section could then represent as much as 10 km of Tertiary strata. These correlations are speculative at this time; however, the proximity of the 1988 drift path to earlier seismic profiling done by industry on the shorefast ice off northern Ellef Ringnes Island, will allow correlations with land based seismic surveys and borehole data.

Other interesting reflections are seen as discontinuous-segments 0.2 s deeper than the seafloor reflection. The best

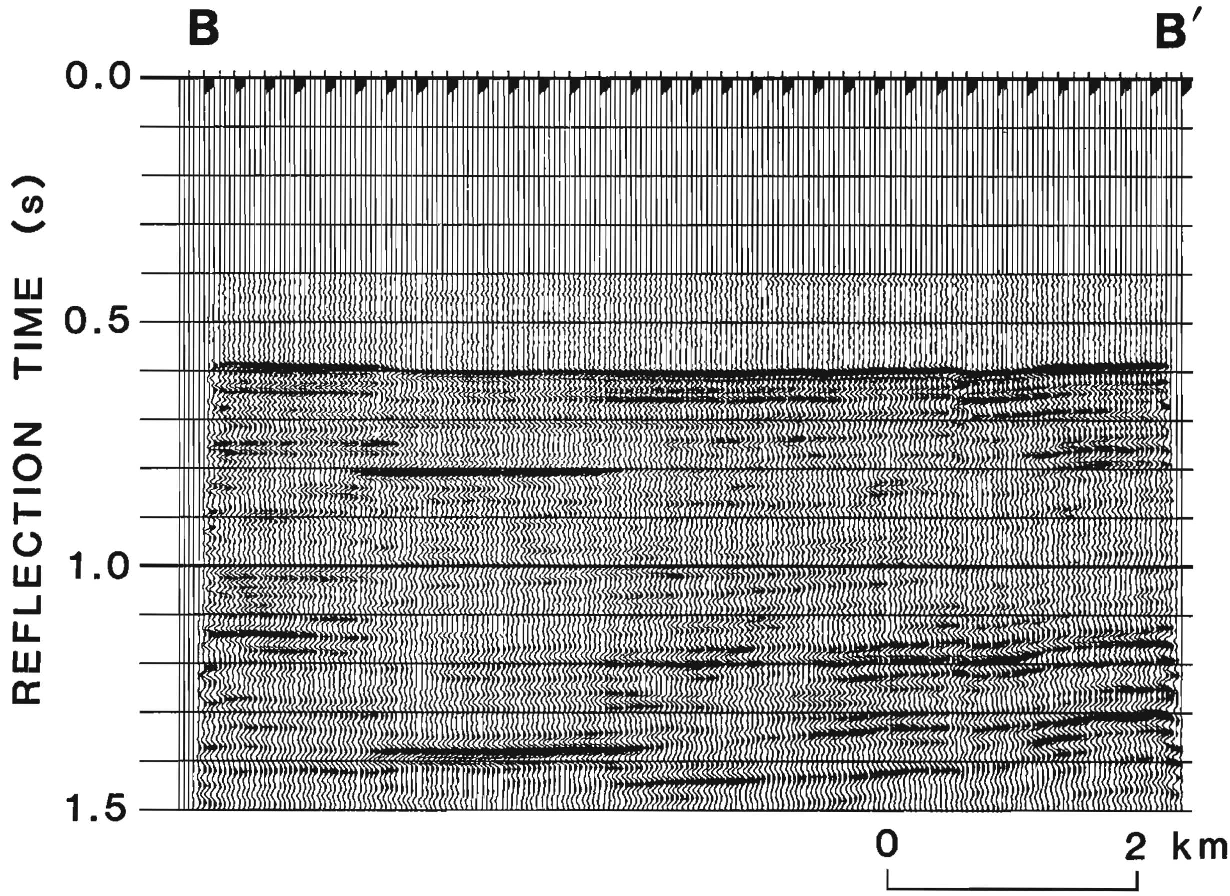

Figure 5. A processed seismic section from 1988 (B-B' in Fig. 2) showing very strong reflections at 0.8 s on the left half of the section. Source 120 cubic inch airgun.

example of this phenomenon is shown in Figure 5 (B-B' in Fig. 2), which was obtained using a120 cubic inch airgun. It is seen as a very strong reflection near the lefthand side of the figure, at a reflection time of 0.8 s; the seafloor is at 0.6 s. These events appear frequently in the 1988 data, and are thought to represent coal seams, or possibly gas hydrate pockets. The strong event at 1.38 sec is a good example of a repetition of the strong primary reflection due to the first reverberation in the water layer. Processing of the 1988 data is still in progress.

CONCLUSION

The drift performance of H-1 ice island has been generally favourable since 1983. It has provided the opportunity to study Canada's continental shelf in this otherwise inaccessible area. The seismic reflection profiling on H-1 is cost effective compared to similar studies in the remote arctic.

ACKNOWLEDGMENTS

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The influence of dissolved sulphate on calcite dissolution: a possible link to the problem of dolomitization†

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Abstract

The dissolution of calcite in sulphate-bearing and sulphate-free solutions that contain no initial calcium has been investigated in order to delimit controls on the rate of calcite dissolution in natural solutions. Calcite dissolution is an essential part of the limestone dolomitization process. This process has commonly occurred in many carbonate-hosted hydrocarbon reservoir rocks.

The data of this study show that the dissolution of calcite is not inhibited in sulphate-bearing solutions that contain no initial dissolved calcium. This corroborates previous assertions that the extent to which dissolved sulphate retards the rate of calcite dissolution is dependent on the initial calcium concentration.

Résumé

On a étudié la dissolution de la calcite dans des solutions sulfatées et non sulfatées ne contenant pas de calcium au départ pour délimiter les paramètres chimiques qui déterminent la vitesse de dissolution de la calcite dans des solutions naturelles. La dissolution de la calcite est un élément important de la dolomitisation du calcaire. Ce dernier phénomène s'est souvent produit dans un grand nombre de roches réservoirs où des hydrocarbures sont logés dans des roches carbonatées.

Les données de cette étude montrent que la dissolution de la calcite n'est pas gênée par des solutions sulfatées ne contenant pas de calcium au départ. Ce résultat confirme des affirmations antérieures voulant que le taux de ralentissement par les sulfates de la vitesse de dissolution de la calcite dépend de la concentration initiale de calcium.

† Contribution to Frontier Geoscience Program.

INTRODUCTION

The dolomitization problem, recognized very early by earth scientists as an important one (e.g., Van Tuyl, 1916), has remained unresolved to the present day (e.g., Machel and Mountjoy, 1986). It has significant economic implications because of the common occurrence of hydrocarbons in reservoirs formed entirely of dolomite, such as, for example, the prolific Devonian Keg River oil fields in the subsurface of northwestern Alberta (Schmidt et al., 1982). Dolomite also hosts the well known Mississippi Valley type lead-zinc deposits that are of common occurrence in the Paleozoic carbonates of Canada (e.g., Macqueen and Thompson, 1978) and elsewhere.

This study deals with the chemical controls on the dolomitization process in general. The investigation of the chemical controls on the dolomitization of calcite (or aragonite) is multifaceted because of the wide range of temperatures under which dolomite apparently forms in nature. Recently, some workers have shown that the concentration of sulphate in solution is an important parameter in determining whether or not dolomite will precipitate. Baker and Kastner (1981) found that at a temperature of 200°C, the dolomitization of calcite was inhibited in solutions with only 5 per cent of the sulphate concentration of seawater. Many workers have been unable to accept the interpretation that sulphate in solution prevents dolomitization because of the common association of dolostones with sulphate-bearingevaporites in ancient rock sequences and because of the Holocene dolomitization that is occurring in settings such as the sabkha flats bordering the Persian Gulf, which are characterized by solutions containing moderate amounts of sulphate (see Hardie, 1987).

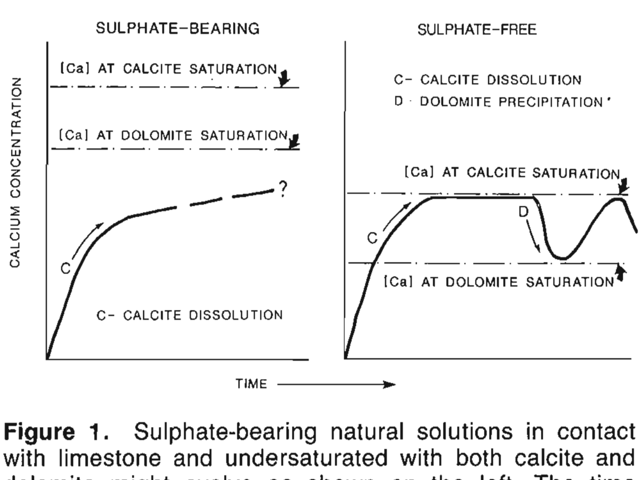

Figure 1. Sulphate-bearing natural solutions in contact with limestone and undersaturated with both calcite and dolomite might evolve as shown on the left. The time required for such solutions to attain dolomite or calcite saturation may be prohibitively long with respect to the time that these solutions are in contact with bodies of limestone, because of the retarding effect dissolved sulphate has on the rate of calcite dissolution. In contrast, sulphate-free natural solutions rapidly attain calcite saturation in contact with limestone bodies. These solutions would then be oversaturated with respect to dolomite. The potential for dolomite precipitation then exists and repetitive cycles of dolomite precipitation and calcite dissolution could cause dolomitization of the entire limestone body. The results of this study indicate that sulphate-bearing groundwaters that have a low calcium activity also have the potential for dolomitization as indicated in the scenario for sulphate-free solutions.

The experimental results of Baker and Kastner (1981) have been verified in part by Morrow and Ricketts (in press) who found that dolomitization was greatly inhibited or totally prevented by the presence of sulphate in solution. However, they also found that dolomite did precipitate from sulphate-bearing solutions that contained no solid calcite, whereas in experiments that utilized calcite as a solid reactant, the rate of calcite dissolution was greatly slowed compared to its rate of disappearance during dolomitization in sulphate-free solutions. They inferred that the presence of sulphate inhibited or slowed the rate of calcite dissolution.

A previous set of experiments involving calcite dissolution were conducted by Sjoberg (1978). Those experiments, conducted at temperatures of 5 to 50°C and at atmospheric pressure, showed conclusively that the dissolution rate of powdered calcite in artificial seawater with a sulphate concentration of 0.028M was less than half the rate for dissolution in a sulphate-free solution of 0.7M KCl under a moderate range of calcite undersaturations. Calcium was present in all these solutions at a concentration of 0.01M and magnesium at a concentration of 0.05M.

The possible geological significance of the dependence of calcite dissolution on the sulphate concentration of the solution is shown in Figure 1. For sulphate-free or sulphate-depleted solutions in contact with limestones, dissolution of calcite proceeds rapidly until the Ca^{2+} concentration rises to a value close to that characteristic of saturation. This value will greatly exceed the concentration of Ca^{2+} characteristic of dolomite saturation in most natural solutions. After a suitable induction period, perhaps dependent on other kinetic factors (Machel and Mountjoy, 1986), dolomite may precipitate, causing the calcium concentration of the solution to fall toward values characteristic of dolomite saturation. Repetitions of this process would cause the dolomitization of entire limestone bodies.

In contrast, dissolution of calcite in limestone bodies in contact with sulphate-bearing solutions may be greatly inhibited. It is not certain how long such solutions must remain in contact with limestone for the concentration of Ca^{2+} to approach that characteristic of calcite saturation or even to marginally exceed that characteristic of dolomite saturation. Given that, in most natural environments of diagenesis, fluids are in constant motion, it seems possible that the inhibiting effect of sulphate might prevent dolomitization because the residence time of these fluids is too short to permit Ca^{2+} concentrations to rise to a level that would permit the precipitation of dolomite (Fig. 1).

The focus of this study is on the dissolution behaviour of calcite in the presence of dissolved sulphate. The set of experiments reported on here involve comparing the dissolution of calcite in pure water with calcite dissolution in a simple solution of 0.01M Na_2SO_4. The purpose of this preliminary investigation is to ascertain the degree to which dissolved sulphate influences the rate of calcite dissolution in the absence of other ions. All other previously discussed studies in which the influence of dissolved sulphate on the rate of calcite dissolution was documented were conducted in solutions containing diverse assemblages of other ions.

METHODS

These experiments were conducted with a sequential sampling technique. Each run consisted of placing 0.500 g analytical grade powdered calcite (Fisher Scientific and Mallinckrodt) into each of six 250 ml laboratory plastic bottles. In some runs, 1.000 g of calcite instead of 0.500 g was used. The bottles were then filled to overflowing with double distilled water in some runs, and 0.01M Na_2SO_4 solutions in other runs. The bottles were then taped with laboratory parafilm and tightly capped. The bottles were placed in a water bath that was maintained at a temperature of 25°C.

The bottles were opened and the contents analyzed at regular time intervals (Fig. 2). Immediately upon opening, the pH of the solution was measured with a Sargent-Welch conbination-type pH electrode attached to a Sargent-Welch pH6000 pH meter. The bottle contents were then filtered into 250 ml capacity Erlenmeyer flasks using micronsep Magna 0.45 micron nylon membrane filters. The total CO_2 equivalent content of 50 ml of the filtered solution was then analyzed with an Orion carbon dioxide gas sensing electrode attached to an Orion 901 ionalyzer potentiometer. This method of analyzing for the total CO_2 equivalent content requires that the test solution be acidified to a pH of 4.5 to 5.0 by means of the addition of an acidic sodium citrate buffer solution. This acidification causes almost all carbonate species in solution to be evolved as CO_2 gas. Additional 5 ml aliquots of the filtered solution were stored in small sealed test tubes. These solutions were later analyzed for their calcium and magnesium concentrations with a Perkin-Elmer 603 Atomic Absorption Spectrophotometer. The calcium concentration in the double distilled water used to form solutions was found to be less than 0.09 ppm. This level of calcium contamination is below the precision of calcium measurements by atomic absorption at the level of the calcium concentrations recorded during this study.

The pH and CO_2 gas measurements are less precise and far less accurate than the solution calcium determinations. The pH measurements have an estimated precision of 0.01 pH units based on multiple measurements of some solutions. Accuracy of pH measurements is estimated to be about 0.02

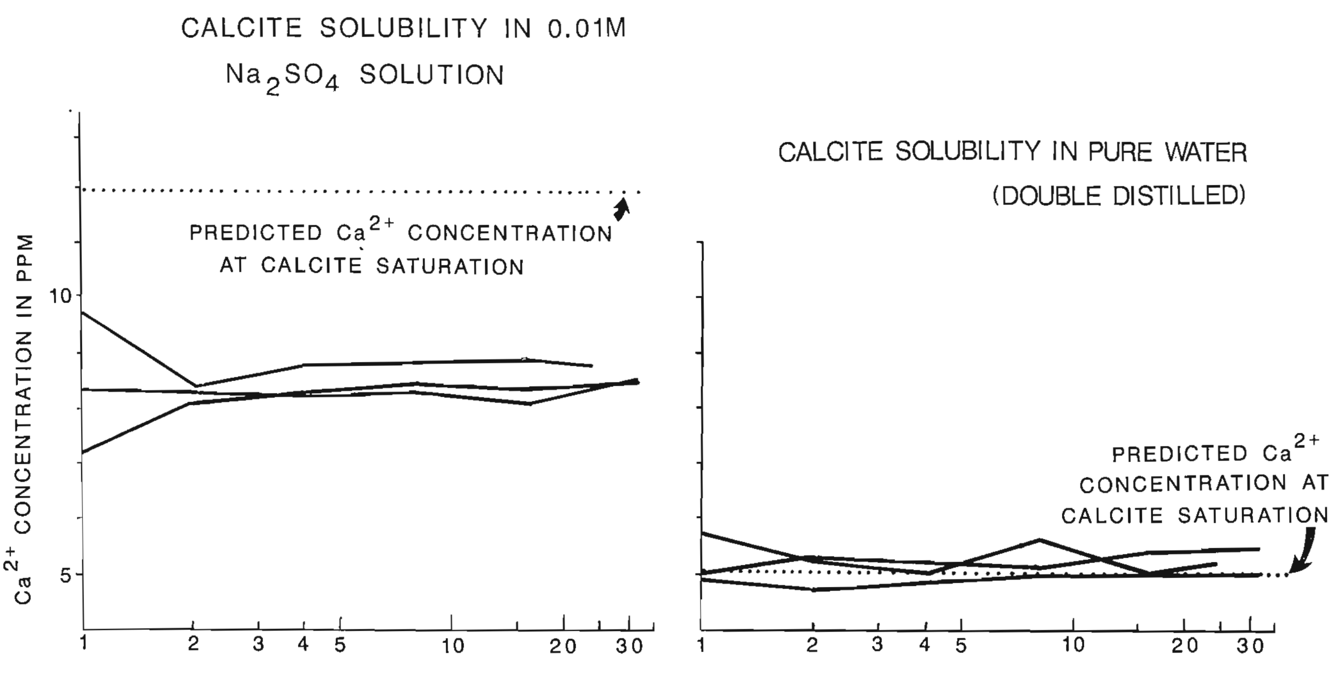

Figure 2. Graphs of solution calcium concentrations for calcite in pure water versus calcite in sulphate-bearing aqueous solution (0.01M Na_2SO_4). The calcium concentrations in the pure water (double distilled) experiments are indistinguishable from equilibrium values. In spite of the fact that the sulphate-bearing solution runs are closer to saturation with respect to calcite than the pure water runs, the concentration of total calcium in solution should rise to close to 12 ppm at calcite saturation. This is because of the formation of the $CaSO_4°$ ion pair.

pH units based on measurements of solutions with known pH values. The CO_2 gas measurements were reproducible within one ppm at the level of the CO_2 gas concentrations recorded in this study. Their accuracy is estimated to be within two ppm as indicated by measurements on solutions with known carbonate contents diluted to the levels of CO_2 observed in this study. Reported calcium concentrations have a precision of less than 0.1 ppm and are accurate to within 0.5 ppm.

RESULTS

The results of this study are shown in Table 1. The pH values are significant only to the first decimal place and the CO_2 values are significant only to one part per million (ppm). However, the CO_2 data have permitted the identification of an aberrant calcium determination from the sample taken on day 4 of run series A. In this sample, and possibly also in the sample taken on day 8 in run series D, it is likely that some calcite passed through the filter system into the filtered solution to yield spuriously high calcium and CO_2 values. Also, it is likely that the pH values for the samples taken on days 16 and 32 in run series E and F are anomalously low because of impaired electrode efficiency during the time of measurement of these runs. Both the pH and CO_2 values exhibit a slight but persistent upward trend. However, this upward trend is less evident in the calcium data (Fig. 2) although there does appear to be a slight positive statistical correlation between calcium and CO_2 values (Table 1).

The results of the time series calcite dissolution experiments are consistent between runs. There does not appear to be any consistent difference between runs conducted with

Table 1. Combined results of solution analyses.

A. 1.000 g calcite with double distilled water

Days	pH	Ca (ppm)	CO_2 (ppm)
1	9.12	5.00	5.72
2	8.92	5.37	5.68
4	9.06	8.41*	11.20*
8	9.21	5.15	6.60
16	8.65	5.41	7.57
32	9.40	5.50	7.64
Av.	9.06	5.28	6.64

B. 1.000 g calcite with 0.01M Na_2SO_4 solution

Days	pH	Ca (ppm)	CO_2 (ppm)
1	9.48	8.37	9.58
2	9.38	8.27	9.25
4	9.53	8.22	10.50
8	9.70	8.29	12.20
16	9.17	8.11	11.20
32	9.68	8.56	14.20
Av.	9.49	8.30	11.16

C. 0.500 g calcite with double distilled water

Days	pH	Ca (ppm)	CO_2 (ppm)
1	8.90	5.67	6.98
2	8.78	5.27	7.48
4	9.61	5.02	5.68
8	9.43	5.65	6.63
16	9.45	5.02	7.26
32	9.49	5.21	8.25
Av.	9.28	5.30	7.05

D. 0.500 g calcite with 0.01M Na_2SO_4 solution

Days	pH	Ca (ppm)	CO_2 (ppm)
1	9.38	9.72	9.84
2	9.33	8.39	10.10
4	9.68	8.75	9.94
8	9.75	10.75*	11.10
16	9.67	8.87	11.60
32	9.76	8.77	10.60
Av.	9.60	8.90	10.53

E. 0.500 g calcite with double distilled water

Days	pH	Ca (ppm)	CO_2 (ppm)
1	9.13	4.87	5.53
2	9.53	4.75	6.63
4	9.78	4.85	6.83
8	9.69	5.00	6.59
16	9.18	4.97	8.60
32	9.06	5.07	7.16
Av.	9.40	4.92	6.89

F. 0.500 g calcite with 0.01M Na_2SO_4 solution

Days	pH	Ca (ppm)	CO_2 (ppm)
1	9.27	7.22	9.39
2	9.75	8.13	12.60
4	9.85	8.27	10.50
8	9.79	8.39	10.30
16	9.44	8.33	12.70
32	9.48	8.46	12.00
Av.	9.60	8.13	11.25

*These are probably aberrant values.

1.000 gm of calcite and those with 0.500 gm calcite (Table 1). Instead, there is a pronounced and consistent difference between runs conducted in pure water versus those in sulphate solution. The average calcium concentration in the pure water runs is 5.17 ppm compared with 8.44 ppm in the sulphate-bearing runs. Similar differences are apparent between average CO_2 and pH values between sulphate-free and sulphate- bearing runs.

The slight variation in calcium values within runs, disregarding aberrant values, is an indication that the solutions were close to equilibrium with the solid calcite reactant. However, sulphate-bearing runs display slightly greater ranges of calcium concentrations and a slightly greater tendency to increase during the course of a run. The analyzed calcium concentrations are the most reliable indicators of the saturation state of the solution with respect to calcite because of the aforementioned problems with regard to precise measurement of solution pH and CO_2 concentration.

DISCUSSION

The analysis of the data of this study depends on the choice of an appropriate, or at least adequate, aqueous model to characterize these systems (e.g., Nordstrom et al., 1979). At standard temperature and pressure (STP) the choice of an aqueous model is dependent primarily upon the ionic strength of the system (Nordstrom et al., 1979). In a system that approaches infinite dilution, such as that in which calcite is in equilibrium with pure water, it is sufficient to regard the activity of ions in solution as equivalent to their concentration or molality (Garrels and Christ, 1965). For more concentrated solutions, however, the activities of ions in solution deviate from their analytical concentrations, and this divergence increases with increasing ionic strength of the solution (Garrels and Christ, 1965).

In dilute solutions that have an ionic strength of up to 0.1, the Debye-Huckel limiting law provides a means of estimating the activity coefficients of free ions in solution and of calculating the degree of saturation of solid mineral phases and aqueous ion complexes in electrolyte solutions (Garrels and Christ, 1965). The saturation states of the experimental solutions with respect to calcite and other mineral phases were calculated with the aid of the solution-mineral equilibrium program SOLMNEQ (Kharakaa and Barnes, 1973). The calculated saturation of calcite ([IAP solution/IAP at equilibrium]; IAP = ion activity product) in the pure double distilled water runs (groups A, C and E in Table 1) ranged from 0.259 to 0.516. The calculated saturation of calcite in the sulphate-bearing solutions (groups B, D and F in Table 1) ranged from 0.508 to 0.632. Those results indicate that all solutions had approached calcite saturation to within less than an order of magnitude. Also, the dissolution of calcite in the sulphate-bearing solutions had proceeded toward calcite saturation to a slightly greater extent than in the pure water solutions. It is uncertain whether this difference in calculated calcite saturations reflects a real distinction in calcite dissolution behaviour or simply slight differences in the electrode responses between these solutions. It is clear, however, that the rate of calcite dissolution in the 0.01M Na_2SO_4 solutions was at least as rapid as that which occurred in pure water.

It is possible to make simple estimates of the concentrations of calcium that would be at equilibrium with calcite based on consideration of the dominant ion complexes or ion pairs that form in these solutions. The formation of ion association complexes in solution causes the total concentration of an ion in solution to be divided between its concentration as a free, unbound ion and the amount of the ion bound to dissolved ion complexes. The fraction of a free metal cation in solution can be determined from

$$[M]f/[M]t = (1 + \sum K^*_{MXi}[Xi]f)^{-1} \qquad (1)$$

where [M] f is the proportion of free cation, [M] t is the total concentration of cation 'M' in solution, [Xi] f is the proportion of free anions that remain in solution after a number (i) of ion associates with the cation 'M' have formed in solution, MXi represents one of these associates and K^*_{MXi} represents the stoichiometric association constant for each ion pair involving the species 'M' (Millero and Schreiber, 1982). Equation (1) is based on the formation of one to one ion complexes according to

$$Mi^+ + Xi^- \rightarrow MiXi^o \qquad (2)$$

Millero and Schreiber (1982, *see* equation 39) have determined the ionic strength dependence of the stoichiometric association constants of ion pairs that occur in seawater and have derived an equation for this dependence based on the parameters for equations used by Pitzer and Mayorga (1973) to estimate the mean activity coefficients of electrolytes.

It is possible to calculate the approximate calcium concentrations that would be attained at equilibrium based on equation 1 if the concentrations of free, unassociated calcium that have been calculated by SOLMNEQ for the test solutions are assumed to be estimates of the concentrations of calcium in these solutions at equilibrium. The major ion pair that will affect total calcium concentrations in the pure water solutions is $CaCO_3°$. However, the calculated contribution of this ion pair to the total calcium concentration in the pure water solutions is only about 0.3 per cent. Consequently, the equilibrium concentration of calcium for the pure water solutions is probably close to the observed values, which are largely in the range 5.0 to 5.5 ppm (Table 1, Fig. 2). These values are consistent with the calcium concentration of 5.05 ppm calculated by Garrels and Christ (1965, p. 79) to be in equilibrium with calcite and pure water.

The concentration of free, unassociated calcium in the sulphate-bearing solutions is governed primarily by the formation of the sulphate ion pairs $NaSO_4-$ and $CaSO_4°$, as well as by the solubility of calcite. In this preliminary evaluation, we have calculated the proportion of calcium retained in the solution in the $CaSO_4°$ ion pair based on the slight reduction in the concentration of free SO_4 in solution through the formation of the $NaSO_4-$ ion pair. The concentrations of free calcium ion calculated by SOLMNEQ for these solutions was again taken to approximate the concentration of free calcium at equilibrium. This approximation indicates a total calcium concentration of 12.0 ppm at equilibrium, a value considerably higher than the observed concentrations (Fig. 2). The preceding calculations, while not exact, do indicate that the calcium concentrations of solutions with variable sulphate contents are not reliable guides to their levels of calcite saturation.

The reader is cautioned that the calculations of this study are approximate only. No consideration has been given to the influence of other ion pairs, such as $NaHCO_3^{\circ}$, $NaCO_3^-$ and $CaHCO_3^+$ at solution equilibrium, or to the recalculations necessary to estimate more closely the ionic strength of the solution following the initial estimate of 0.03 for the ionic strength.

In future, the equilibrium concentrations of Ca^{2+} and other ions will be calculated using programs such as PHREEQE (Parkhurst et al., 1980) or even more exactly by programs based on the ion interaction model (Pitzer, 1973; Nordstrom et al., 1979; Harvie et al., 1984). At this stage, we do not think that the application of these programs will alter the basic conclusion that the sulphate-bearing solutions of this study have approached saturation with respect to calcite at least as rapidly as the pure water solutions.

This conclusion is surprising in view of the previously described results of Sjoberg (1978). However, the results reported here are consistent with Sjoberg's suggestion (Sjoberg, 1978, p. 62) that the effect of sulphate ions increases with increasing calcium concentration. The solutions in Sjoberg's (*ibid.*) study contained a concentration of calcium equal to that of seawater (0.01M), whereas none was present in the solutions employed in this study. Within experimental uncertainty we have shown here that the sulphate inhibition effect, documented by Sjoberg, is undetectable in solutions that are initially calcium-free. Future work is required to document the range of calcite dissolution rates in sulphate-bearing solutions with variable initial concentrations of calcium and magnesium.

CONCLUSIONS

The results of this study indicate that the sulphate inhibition effect on calcite dissolution is dependent on the presence of calcium in the initial solution. Consequently, many sulphate- and magnesium-bearing groundwaters that are depleted in calcium have the potential to cause dolomitization of limestones, in addition to many natural solutions that contain little or no sulphate.

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