STRATIGRAPHY, MUD BUILDUPS, AND CARBONATE PLATFORM DEVELOPMENT OF THE UPPER ORDOVICIAN TO LOWER DEVONIAN SEQUENCE, ELLESMERE, HANS, AND DEVON ISLANDS, ARCTIC CANADA

by

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GENERAL ABSTRACT

The Upper Ordovician to Lower Devonian platform in the Canadian Arctic twice evolved from a ramp to a rimmed shelf profile. Platform backstepping occurred in the *fastigatus*, *acuminatus*, *cyphus*?, *sakmaricus* (in North Greenland only) and *linearis* graptolite zones. Whereas earlier backstepping events appear to have been nearly synchronous and affected a greater proportion of the carbonate platform, younger ones were areally more restricted. Two major phases of pinnacle reef development followed platform backstepping, the first beginning in the lower Llandovery (*cyphus* Zone) and the second in the Ludlow (*linearis* Zone). Pinnacles of the first phase are uncommon, occur in the vicinity of Baumann Fiord, and show a vertical succession of lime mudstone, poorly exposed microbial carbonate, and coralgal biolithite, representing upward shallowing. The last named lithofacies is newly interpreted as representing a high-energy, wave-stressed environment that excluded stromatoporoid growth but favoured a sparse skeletal metazoan fauna, thickly encrusted by microbes. Paleo-surface area of these structures was apparently important for the accumulation of extensive ooids, which are associated with the upper parts of some pinnacle reefs. Coeval stromatoporoid pinnacle reefs occur in Kennedy Channel and North Greenland, but are known in

Canada only on Hans Island and are inferred from olistostromes in northeastern Ellesmere Island. Pinnacle reefs of the second phase are more extensive regionally, but less well known. The reefs have mud cores that were succeeded by stromatoporoid- and coral-rich limestone and dolostone facies in sequences that shallow upward overall.

Three large mud buildups on central Ellesmere Island were established on the shelf margin subsequent to Upper Ordovician (*fastigatus* Zone) platform drowning. These structures show a vertical lithofacies succession: bioturbated lime mudstone is overlain by microbial carbonate then by mudstone-rich stromatoporoid floatstone and bindstone. The succession records overall upward shallowing. An areal restriction of the mud buildups and a latest Llandovery onlapping by foreslope facies may have been related to lithospheric flexure and extensive Caledonian syntectonic flysch deposition in the contiguous Hazen trough. The mud buildups appear to have been drowned in the lower Wenlock, and are overlain by a condensed succession of shale.

The olive green shale unit, well exposed in the vicinity of Trold Fiord, is an areally extensive and mappable middle Ludlow unit of the Cape Phillips Formation that postdates diachronous, middle Ludlow platform backstepping in the vicinity of Baumann Fiord. After backstepping, condensed sequences occurred over paleotopographic highs and expanded sections over lows, and a subsequent second major phase of platform rimming occurred. Evidence of Ludlow backstepping occurs only in northern Devon and southern Ellesmere islands and is diachronous over two graptolite biozones, suggesting flexure related to a renewed phase of Caledonian epeirogenesis.

Stratigraphy known in the southern Arctic Islands is generally applicable to northeastern Ellesmere Island, but local lithological variations occur. The upper part of the Allen Bay Formation at Darling Peninsula, in particular, is unusually thick, and subtidal, perhaps resulting from greater subsidence, related to lithospheric flexure and deep marine clastic sedimentation that drowned the contiguous platform on North Greenland. Other formations recognized in this area include the Cape Storm, Douro, and Goose Fiord formations, although these, too, show minor lithological differences from type sequences. The Douro Formation on Ellesmere Island has an upper resistant limestone member, perhaps indicative of a minor movement of the Inglefield Uplift, and the Cape Storm Formation is considerably thinner, less silty, and contains significantly more subtidal units than in southern Arctic localities.

A thick grey siltstone unit in the vicinity of Bay and Vesle fiords is suggested to be a distal facies of the Red Canyon River Formation. This sequence is a progradational clastic wedge that likely represents the first, largest phase of the tripartite Caledonian Inglefield Uplift which profoundly affected carbonate deposition in the areas of southern and central Ellesmere Island during the late Silurian and early Devonian time. The base of this unit is diachronous and likely late Silurian in age.

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PART I

INTRODUCTION

PREAMBLE

Early explorers endured considerable hardship and spent years at a time in the Arctic gathering various sorts of data. Most used dogs, boats, and skis for much of their travelling, and were considerably resourceful. Troelsen (1952), for example, after losing his dogs to disease, manhandled a sled some 100km over pack ice and rolling terrain to examine rock exposures in the vicinity of Cañon Fiord, a considerable feat, particularly in contrast to the present methods of Arctic travel.

Thorsteinsson and Tozer (1957) investigated rocks of the Franklin and Sverdrup basins by hiking and using dog teams and a freight canoe. They spent some 4½ months gaining valuable field observations of these rocks. These small but effective and economic early field parties benefitted considerably from Inuit assistance, and are in contrast to the expensive, fully air-supported field parties of today's workers. Early workers proved dog sleds to be particularly effective on the ice-covered fiords, which stretch many kilometres inland. Hiking and canoeing were less effective, as distances are great and winters long. Fully helicopter-supported modern field parties clearly have the advantage of greater, more effective mobility, although they bear a much greater expense.

This work was planned on the premise that air support would be available from a large GSC field party working in the study area, but because of changes in scheduling by the GSC and cost constraints, an alternative, less expensive method of data acquisition was sought. In May 1987, with a maximum of 6 hours of Twin-otter air support, Pierre Gravel (my co-worker that summer) and I arrived at Bay Fiord. The study required a reconnaissance of all well-exposed, pertinent outcrops in the vicinity. Cross-country skis provided primary transport and were most effective during June, when rivers were still fully snow clogged and high snow temperatures provided optimum ski glide. We made daily traverses from a base camp and rarely returned before 10 pm; daily excursions averaged about 30km, but some were up to 85km. This method, although effective for most localities near Bay Fiord, was physically taxing. We were particularly loath to continue this method of transport in late June, when food rationing became necessary due to dwindling supplies, a result of our high daily caloric intake through most of June.

Skiing was effective over pack ice, but less so over land. In the vicinity of Irene Bay, there was no more than 10% snow cover in late May, but in areas only 20km west, the snow pack was continuous and some 20-30cm deep. As observed in subsequent field work, the ice capped mountains of eastern Ellesmere Island likely form a precipitation shadow in a narrow band west of the mountain chain. This allows early access to less precipitous rock exposures that are otherwise mostly snow covered until July or August. This condition was important in the cold and snowy summer of 1989, during final field work in conjunction with a GSC field project in the vicinity of Makinson Inlet.

In early July, having skied or hiked to all accessable outcrops in the Bay Fiord area, we were flown to Colin Archer Peninsula where we spent less than a month surveying Allen Bay Formation exposures. We were supplied from a main GSC base camp and were frequently relocated so that minimal time was spent in travelling.

The summer of 1988 presented considerable challenges. As in the previous summer, air support was limited, and a field season was planned that was reminiscent of the early days of Arctic geological work. Although the PCSP (Polar Continental Shelf Project) gave considerable air time to the project, it was not sufficient to accommodate repeated camp relocation in the large study area. Skis were inadequate for the task, and instead, use of a skidoo and Nansen sled was suggested by a Norwegian employee of the PCSP, a man who has had considerable experience and success with this mode of transport. Equipped in this way, we began work at Cape Baird in early April in a light wind, ice fog, and a temperature of -30°C, anticipating a lengthy and unpredictable voyage, especially since niether of us had had more than a single cumulative hour of operating a skidoo! My assistant for this season of field work was Robert Thériault, a then undergraduate student at the University of Ottawa.

Food and gas caches had been set out at Carl Ritter Bay and John Richardson Bay, on our air lift to Cape Baird, at a minimum cost of air time. The food was stored in strong boxes to reduce the possibility off loss in the event of examination by a curious polar bear. We had hoped to exchange rock samples for food and fuel at the caches, which would eventually be recovered during our final evacuation or at some other convenient time, for example, during overflight of a plane returning from some other Arctic work that summer.

At first, field work progressed rapidly during the voyage between Cape Baird and Church Peak, and we examined numerous exposures with minimal snow cover in these areas (Figs.1,2). With very good weather, we were able to travel effortlessly over pack ice devoid of drifts or pressure ridges. Although some of the ice tested our patience, we made a relatively uneventful arrival at our first cache at Carl Ritter Bay, then proceeded southward some 160km to the next outcrop. This was the most perilous part of the trip. The ice was terribly jumbled across all of Kennedy Channel, and we alternated between the ice foot (land-fast ice) and pack ice innumerable times, an operation which required up to two hours of digging and lifting of the skidoo and sled up an ice ridge 1-2m high. It took some 6 days to travel less than 50km, and most of these days began early and ended at 1 or 2am the next day. The weather at this time was also very uncooperative; heavy snow one day in mid-April was followed by a sunny day with a windchill of -90 to -100°C.

We arrived at John Richardson Bay on the 8th day of travel (a distance of about 180km as measured in a straight line which we truly never attained). Although the ice was smooth in the bay, it was blanketed by a metre of snow in places, and it

Figure 1: Section locations in the study area. Stippled lines represent approximate locations of the carbonate platform margin for the times indicated. Sections in Baumann, Bay, and Cañon fiords are located in more detail in figures 3, 4, and 5, respectively.

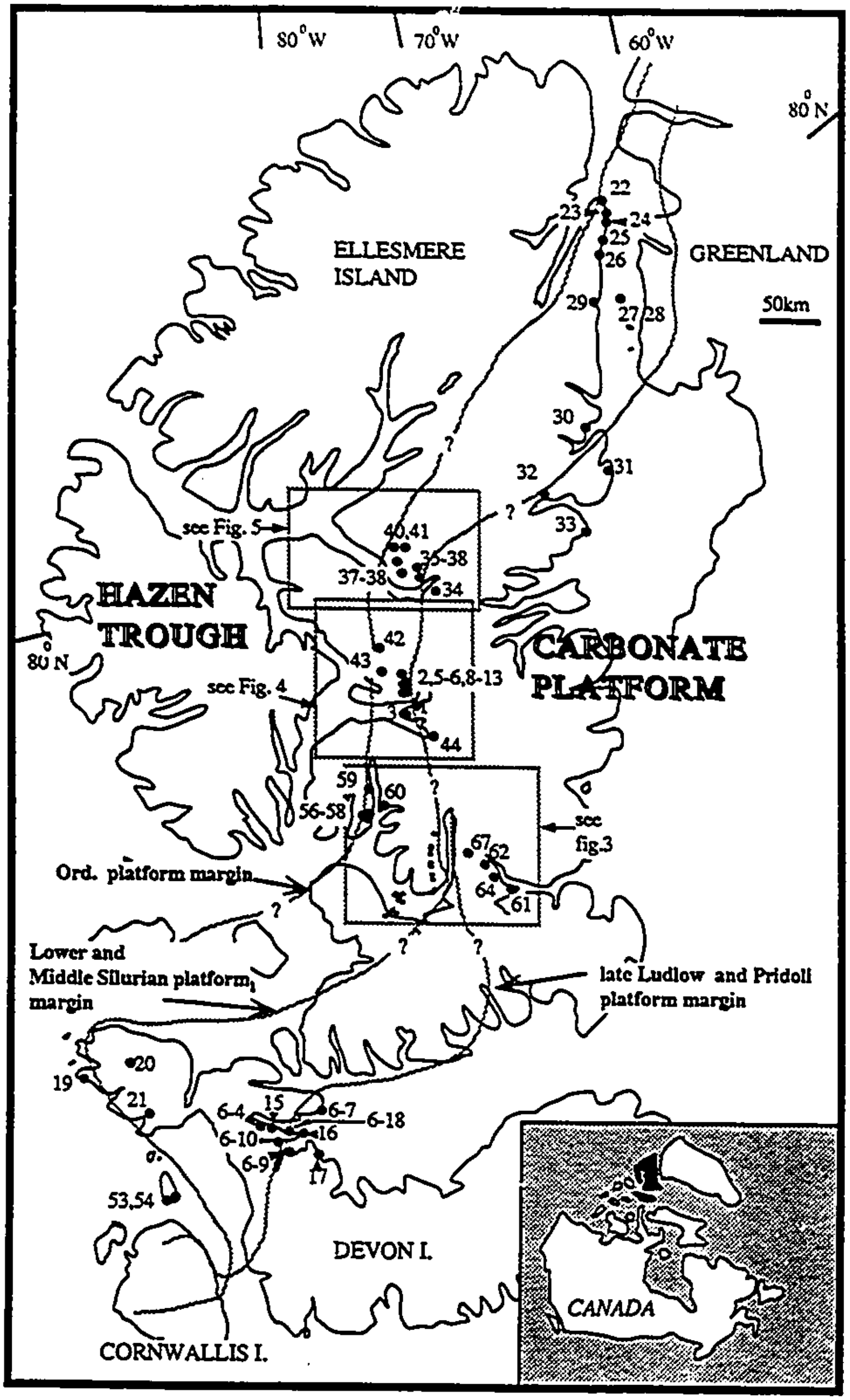
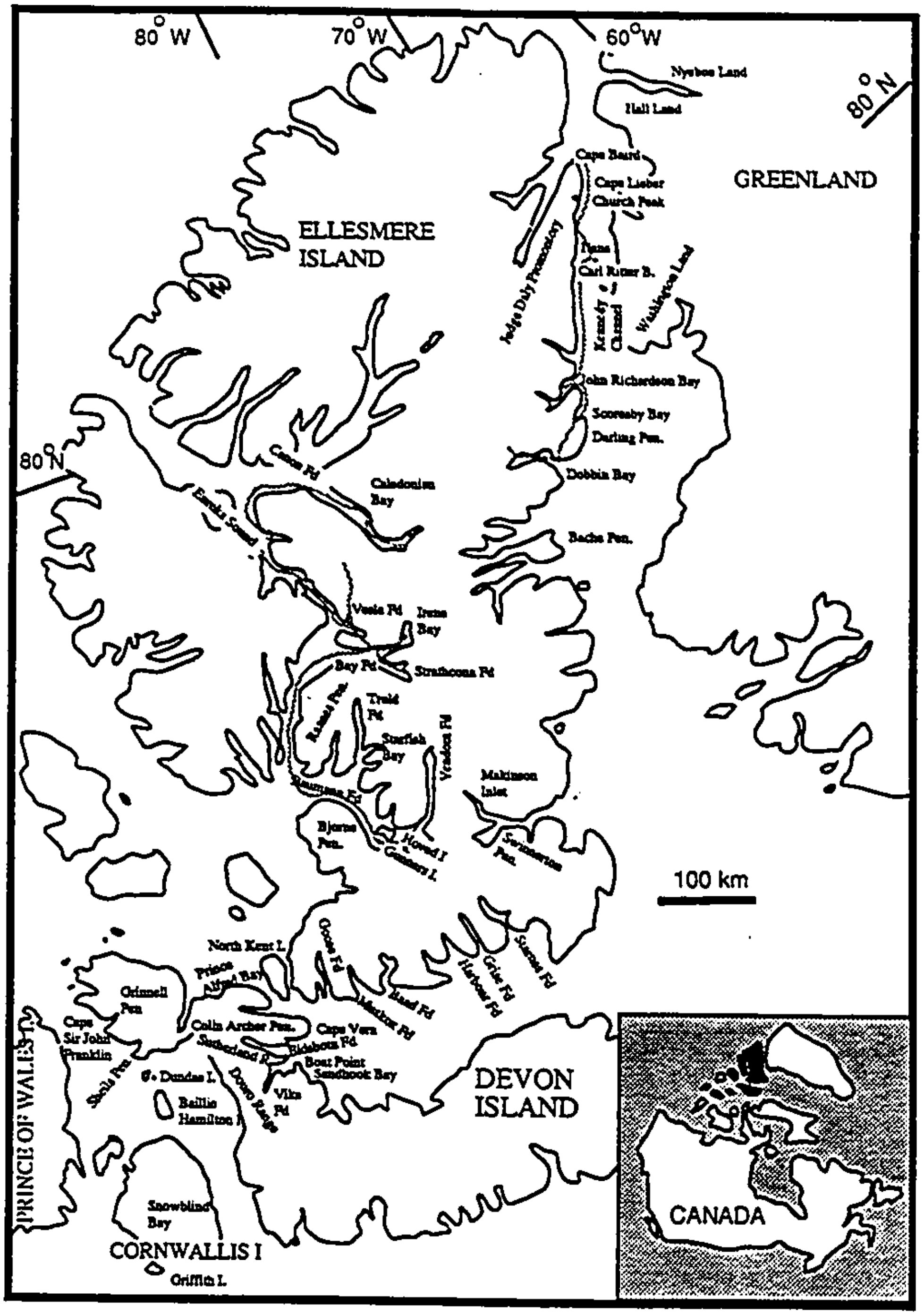

80°W
70°W
60°W
80°N
ELLESMERE ISLAND
GREENLAND
50km
HAZEN TROUGH
CARBONATE PLATFORM
80°N
see Fig. 5
see Fig. 4
see fig.3
Ord. platform margin
Lower and Middle Silurian platform margin
late Ludlow and Pridoli platform margin
DEVON I.
CORNWALLIS I.
CANADA

Figure 2: Locality names referred to in text. Gray line represents route travelled by skidoo and skis during the 1988 field season.


80° W
70° W
60°W
Nyeboe Land
Hall Land
80° N
GREENLAND
Cape Baird
Cape Lieber
Church Peak
ELLESMERE ISLAND
Judge Daly Promontory
Carl Ritter B.
Kennedy Channel
Washington Land
John Richardson Bay
Scoresby Bay
Darling Pen.
80°N
Caledonian Bay
Dobbin Bay
Eureka Sound
Bache Pen.
Vesle Fd
Irene Bay
Bay Fd
Strathcona Fd
Trold Fd
Starfish Bay
Makinson Inlet
Bjorne Pen.
Hoved I
100 km
North Kent I.
Prince Alfred Bay
Grinnell Pen
Colin Archer Pen.
Cape Vera
Boat Point
Sandhook Bay
Cape Sir John Franklin
Dundas I.
Baillie Hamilton I
Viks Fd
DEVON ISLAND
PRINCE OF WALES I
Snowblind Bay
CORNWALLIS I
Griffith I.
CANADA

was as if never disturbed by wind through the winter, a condition encountered in the three other fiords (or bays) visited during the journey down Kennedy Channel. This is apparently typical of many Fiords of eastern Ellesmere Island and North Greenland and was a considerable barrier for dog sled travel in the past (Bob Christie, pers. com., 1987); however, the skidoo and Nansen sled were little hindered, and our progress was only slightly impeded. We had an unhurried look at a well exposed carbonate-shale sequence at this locality then proceeded to Scoresby Bay. Several days in early May were spent there examining a well exposed but extensively snow-covered exposure.

With the prospect of very bad ice in Kane Basin, we sought an alternative overland route. Although an apparently shorter route was available Darling Peninsula, the snow was well over a meter deep and soft under foot, causing us to take more than a day to travel less than 10km. The river bed that we followed was also tortuous and with occasional steep drops required us to unpack, repack, and carry our provisions and samples several times over the distances of the drops. We were greeted at Dobbin Bay by extremely good ice, and we covered ground rapidly. Work progressed quickly on south-facing cliffs, but on north-facing exposure on the southern coast of Dobbin Bay, snow was more than waist-deep and considerably covered the exposure. We unfortunately had to abandon this last potentially very interesting outcrop.

Our difficult route down the coast of Judge Daly Promontory had taken some

50 gallons of gas, and we lacked had none extra to continue the planned travel over Parrish Glacier to Cañon Fiord. It was too expensive to airlift additional fuel, and the rapidly warming air temperatures and diminishing time available for ice travel hastened the decision to take a short airlift to Cañon Fiord.

Travel in the fiords and sounds of central Ellesmere Island, was much easier than in Kennedy Channel. On the smooth ice and hard, wind-packed snow, we found travel up to 150km a day to be relatively effortless. We examined numerous exposures in Cañon Fiord in two weeks, in rapidly warming air temperatures. Although night-time temperatures in mid-May still were below -20°C, by late May, temperatures surpassed the freezing point. We had planned an overland voyage to Bay Fiord, but much newly formed slush in many of the low-lying rivers convinced us that a route on the pack ice of Eureka Sound was simpler, although considerably longer. It took two days to reach the Eureka weather station and an additional day to arrive at Vesle Fiord. Although the time was late May, we were surprised to see many deep puddles forming over the pack ice in Strathcona Fiord, Irene Bay, and some parts of Eureka Sound. Ice leads were becoming numerous and some too wide to cross. Several times we traversed Eureka Sound in search of a safe crossing, or had to "skip" the skidoo across open leads. The latter method was particularly perilous for the one skiing behind the Nansen sled.

In early June we again took to the ice of Eureka Sound after a short stay in Bay Fiord, and travelled the last leg of the trip. Because of innumerable ice leads and puddles, the travel took three long days, but we arrived at Hoved Island, Baumann Fiord, during the fourth blizzard of the trip. A kilometre or two from our destination, at about 2am on the last morning of major travel, in blowing snow and sleet, our skidoo suffered one of its worst breakdowns, but, with the spare parts supplied by the PCSP, we fortunately managed to repair the skidoo, finishing the last segment of the trip in the first week in June. A full two weeks were spent on exposures in the area of Baumann Fiord. We had hoped to venture the length of Vendom Fiord and travel inland to examine some key exposures, but these prospects were wholly impossible with fast-moving rivers and a disintegrating ice pack. Our season ended in late June. Two rock caches were left for subsequent pick up, one at Cañon Fiord and the other at Bay Fiord; the latter was unfortunately never retrieved. We left the Arctic in late June with abundant geological data, attained during more than 1600km of skidooing and skiing.

The following summer (1989), the GSC had a large helicopter-supported field party at Baumann Fiord. I was fortunate enough to be involved with these studies and collected further data for the following report. This third and final portion of the work took 11 weeks of June, July, and August.

PURPOSE AND IMPORTANCE OF THIS STUDY

Because of its remoteness and harsh climatic conditions, the Canadian Arctic remains an area where the sedimentary geology is incompletely known. This thesis represents a major contribution to the Upper Ordovician to Lower Devonian stratigraphy of the Canadian Arctic. It provides substantial new data from areas that were previously unstudied and for which no complete geological maps exist (for example, as on much of Judge Daly Promontory and on Grinnell Peninsula). Detailed imformation was also derived from re-xamination and re-interpretation of the stratigraphy in areas somwhat better known, such as the sections studied by Poey (1988) and McGill (1974) in the vicinity of Baumann Fiord. This study intergrates these data in an updated, comprehensive interpretation of the stratigraphy in the northeastern Arctic Islands and North Greenland. Never before recognized lithostratigraphic units (for example, the Starfish Bay shale) are described and are shown to have been an important part of basin evolution. Most of the lithostratigraphic units in the study area show major regional variations that have not been describe before. Although shale onlapping sequences have been recognized, their regional significance, timing, and sedimentology have never been addressed. With the abundant new biostratigraphic data presented in the following work, many of these sequences are now correlated across the basin or are shown to be of local significance or diachronous. These points are addressed in the overall context of platform evolution and event stratigraphy.

The biostratigraphy integrated with the lithostratigraphic framework, is essential for regional correlation, and is an important aspect of this study. The timing of major events in platform evolution was generally previously unstudied or poorly known. The biostratigraphy has facilitated a more comprehensive regional interpretation of platform evolution. Most of the biozones established before in the Canadian Arctic have been recognized and correlated across the basin in this study, thus providing a vital framework for these regional considerations and a framework within which aspects of the sedimentary sequence can be considered in greater detail. Some of these lithological variations are discussed in the context of global sea-level fluctuation, particularly with respect to the Ordovician-Silurian boundary.

Large and small organic buildups of various types were recognized (some for the first time) in different parts of this sequence. These buildups and associated facies are the second major emphasis of the thesis; some were examined on a reconnaissance scale, and others in more detail. Mudmounds and mud buildups are recognized for the first time on Ellesmere Island, and main facies are described. These are compared with other buildups in this study area and with sponge mounds and other extrabasinal Silurian mud buildups.

Based on these studies, a geological model is proposed that will perhaps enhance future hydrocarbon exploration within and even outside the Franklin Basin. The lower Paleozoic sequence has already yielded economic deposits of zinc and lead (Polaris Mine, on Little Cornwallis Island) and significant accumulations of sweet oil (Bent Horn Well, near Bathurst Island). The latter was an important source of heating oil and a supply for thermoelectric generators in the small Arctic Inuit communities, particularly where much of this oil could be used unrefined in the generators or furnaces. Because little was known of the geological structure from which this oil was extracted, well exploitation was discontinued and other sources of fuel had to be sought for the communities. Although slightly older, geological structures similar to those occurring in the subsurface of the Bent Horn Well are exposed in the report area, and study of these exposures could enhance future oil exploration and development.

PREVIOUS WORK

Our geological knowledge of the Franklinian Basin has progressed stepwise and, for Ellesmere Island, was very scattered until two major air-supported Geological Survey of Canada (GSC) field projects: Operation Franklin of 1955 (Fortier *et al.*, 1963) and Operation Eureka of 1961 and 1962 (Kerr, 1967, 1976)). These investigations established much of the basic stratigraphy for the Canadian Arctic. From work in Operation Franklin, the Douro, Devon Island, Eids, Goose Fiord, and Sutherland River formations, among many other formations, were named. Kerr (1972a,b,c) and Thorsteinsson (1972a,b,c) made geological maps of central Ellesmere Island, and these were used in planning fieldwork of the present investigation. Poey (1988), Mayr (1974), Roblesky (1979), and McGill (1974) gave more detailed accounts of the geology in the vicinity of Baumann Fiord; this area is by far one of the most interesting and geologically complicated localities of the study area. Mayr (1974) was the first to recognize isolated carbonate buildups on Gunnars Island, while Poey (1988) and McGill (1974) studied shelf margin facies predominantly west of Vendom Fiord. J.C. Sproule and Associates, Petrocanada, and several GSC field parties have visited Baumann Fiord, and the slope sequence exposed at the northwest corner of Baumann and Vendom fiords has been of considerable interest, although these workers have never published their data or interpretations.

Geology of Judge Daly Promontory is less well known, primarily because of its remoteness. Christie (1957,1964) was the first to produce geological maps of this area for the GSC, although the first basic geological maps were drawn by Fielden and de Rance (1878). Etheridge (1878) published significant fossil finds of this area, including the collection of *Pentamerus* from Offley Island and "*Monograptus convolutus*" (presently named *Monograptus spiralis*) from an unspecified locality on Judge Daly Promontory. More recently, Kerr (1972b,c), J.C. Sproule and Associates (unpublished maps), Mayr *et el.* (1982), Trettin *et al.* (1982), and Trettin and Mayr (1990) have produced more detailed and accurate geological maps of the structurally complicated promontory. Hurst and Kerr (1982) collaborated on a study of the Ordovician-Silurian shelf margin sequence in connection with the origin of Nares Strait. They discussed a carbonate "horst structure" which was said to extend across the strait and be exposed on Hall Land and Judge Daly Promontory. As demonstrated in this study, this structure is clearly not a horst but a sedimentary structure related to biogenic (microbial) production of carbonate that began following Upper Ordovician platform backstepping.

On southern Ellesmere Island and Grinnell Peninsula, Kerr (1969) produced preliminary though relatively accurate geological maps of the area; although he collected a vast amount data, much of it is unpublished. Preliminary accounts of the geology of Devon Island are given by Fortier (*et al.,* 1963). Morrow (in Morrow and Kerr, 1977), then a graduate student, subsequently produced a comprehensive report on the geology of the Douro Range with a detailed account of the sedimentology of the Upper Ordovican to Lower Devonian carbonate-clastic sequence. Some of their interpretations, however, are disputed by Mayr and others (in prep.). For example, Kerr (1977) recognized a major disconformity beneath the Cape Strom Formation on Grinnell Peninsula, but this has not been confirmed in this investigation or in work by Mayr and others (in prep.). The base of the Cape Storm Formation in this study is placed higher in the sequence, and represents a significant hiatus.

Much morework has been done in the southern Arctic, in contrast. On southern Devon Island, Thorsteinsson and Mayr (1987) presented abundant new stratigraphic data for a large study area and compared it with other areas, mostly in the southern Arctic. Thorsteinsson and Uyeno (1980) produced a most significant report on arctic stratigraphy, including a detailed account of the biostratigraphy and the regional relationships of main Siluro-Devonian stratigraphic units. Mayr (1978; 1980), primarily using subsurface data, correlated the main Cambrian to Devonian stratigraphic units in the southern Arctic Islands. He was also the first to work out main facies relationships of the isolated carbonate bodies that were established subsequent to platform backstepping on Bathurst Island (Mayr, 1980).

Following the extensive field studies on Cornwallis Island by Thorsteinsson, which began with his Ph.D. field studies in the early 50's (Thorsteinsson, 1958; Thorsteinsson and Kerr, 1968; Thorsteinsson and Uyeno, 1980), subsequent workers have investigated more specific aspects of the sedimentology and paleontology of the island. Graptolites of the Cape Phillips Formation of Llandovery age have been reported by Melchin (1987b), of Wenlock age by Lenz and Melchin (1990), of Ludlow and Pridoli ages by Lenz (1990), all from various localities of the Arctic islands, particularly from Cornwallis Island. In the vicinity of Snowblind Creek, Cornwallis Island, Mallamo (1989) investigated aspects of the shelf-margin facies and correlated them with coeval platform facies on Griffith Island. Sodero and Hobson (1979) did a similar study of the Allen Bay and Cape Phillips formations, but looked at a greater age range of strata in many key platform and platform margin exposures on southern Cornwallis Island. These workers recognized remarkable platform facies variations that were later questioned by de Freitas (1987c) and Mallamo (1989). The Allen Bay Formation was originally named by Thorsteinsson and Fortier (1954) and fully described by Thorsteinsson (1958) from a type section near Resolute Bay Airport on southern Cornwallis Island. The Cape Phillips Formation was also originally described by Thorsteinsson (1958), and he established the first graptolite biozones that were later fully studied by Lenz and Melchin (1990) and Melchin (1989).

Work has also been extensive in the southern Arctic, in the vicinity of Boothia Peninsula and Somerset Island. Operation Franklin in 1955 (Thorsteinsson and Tozer, 1963, *In* Fortier *et al.*, 1963), Operation Prince of Wales in 1962, and Operation Boothia in 1975 and 1976 have all examined aspects of the geology of the area. Miall and Kerr (1977,1980), Stewart (1987), Blackadar and Christie (1963) investigated various aspects and established the basic stratigraphic framework of the area. Miall and Kerr (1980) recognized a sub-Allen Bay Formation disconformity, in places, truncating all of the Irene Bay and part of the Thumb Mountain Formation, but subsequent work (Stewart, 1987) demonstrated that the amount of missing strata is somewhat less than previously thought.

Workers at the University of Ottawa have added considerably to the knowledge of Somerset Island and vicinity. Narbonne (1981) investigated sponge mounds and the regional stratigraphy of the Douro Formation, and was able to recognize two major deepening events in the formation. Jones and Dixon (1975) named the Leopold Formation on northeastern Somerset Island, but Stewart (1987) and Miall and Kerr (1977), noted that the "...lithological differences are not great enough to justify the use of separate formational names...", at least at the scale of present GSC geological mapping (Stewart, 1987, p.23). Trettin (1969) named the Cape Crauford Formation for a gypsiferous and (solution) brecciated rock unit exposed on Brodeur Peninsula, but Nentwitch (1987) showed that there were no evaporites or evaporite solution breccias exposed on the peninsula. Neverthless, the formation name was applied to areas on Brodeur Peninsula and Somerset Island (Stewart, 1987; Mayr, 1978). Mortenssen and Jones (1986), from an examination of Middle Silurian exposures on Prince of Wales Island, were able to map several penecontemporaneous sub-basins that affected thicknesses and facies of the Cape Storm, Douro, and Somerset Island formations.

METHOD

This thesis represents the culmination of 3 summers of field work. As described above, most of the work was accomplished from a "mobile camp" in the late winter and early spring of 1988, from GSC field camps in the summers of 1989 and 1987, and from a single base camp in the spring of 1987. Stratigraphic sections were measured using a 1.5m Jacob staff, and samples and notes were taken of pertinent rock units. Due to time and sampling restrictions imposed by the method of ground travel between study localities during 1988 and part of 1987, results and interpretations in some areas are preliminary. The remoteness of the study area and the expense of air travel precluded a second examination of most outcrops. Of the greater than 60 stratigraphic sections measured, many are represented in summary form in figures through this work.

Samples were brought back by sea freight in the early Fall of each field season. All samples were slabbed and stained for dolomite and calcite, most were thin-sectioned, and some were digested in acetic acid for later conodont separation. Thin sections were particularly useful for describing the fine-grained mud buildup facies and for the upper limestone member of the Allen Bay Formation. Some peels were made, but these were largely inadequate for study of many of the samples. Thin sections of fine-grained lithologies or mud buildup strata were polished and examined under the cathodoluminoscope. This tool was important for deciphering main cementation phases in the rock and for delineating the original textures of many of the completely recrystallized carbonates. Stromatoporoids, for example, appear to be one of the fossils most susceptible to recrystallization and are more readily identifiable using cathodoluminescence.

Insoluble residues from rock digestion in acetic acid were thoroughly washed, then separated by the author in carbontetrabromoethane, calibrated at a density of 2.86 g/cm^2. Residual heavy particles, usually including a few conodonts, were thoroughly washed, then picked by the author or by workers at the Geological Survey of Canada. The conodonts were identified by A.D.McCracken or by T.T. Uyeno, both of the Geological Survey of Canada.

Graptolite samples required much less preparation than conodonts. Many

Silurian graptolites have a distinctive morphology and can be readily identified in the field (for example *Monograptus turriculatus turriculatus*, *Monograptus spiralis spiralis*, *Cyrtograptus sakmaricus*, *Bohemograptus bohemicus behemicus*, some samples of *Saetograptus fritschi linearis*, and many others). Although many were identified in the field, over 200 graptolite samples were shipped to the University of Ottawa for more detailed examination. Most samples were immersed in water and examined under medium power (about 20 X) using a Bausch and Lomb binocular microscope. All samples were examined and identified by the author, but key samples and graptolites that were particularly difficult to identify were sent to experts in the field of graptolite taxonomy for verification (including A.C.Lenz, M.J.Melchin, and R.Thorsteinsson).

Drafting was done with a "Macintosh" computer using version 2.1 of the "Canvas" drafting package.

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My greatest appreciation is extended to O.A.Dixon, who supervised this project from it first stages. He was always available for discussion and offered encouragement for my admittedly unusual approach to the field work in the summers of 1987 and 1988. During these two summers, Pierre Gravel and Robert Thériault

offered their tireless and cheerful assistance, even though working conditions were sometimes very harsh.

Discussions with several individuals at the Geological Survey of Canada (GSC) and the Polar Continental Shelf Project (PCSP) added to the content or planning of this work: Barry, Jim, George, Claude, and others at the PCSP in Resolute Bay helped with preparation of the Nansen Sled, skidoo, and itinerary for the 1600km sled journey of 1988. Barry Hough was particularly instrumental in arranging gas and fuel caches during our voyage and was supportive throughout the journey. At the GSC, Ulrich Mayr is to be particularly thanked for logistical support in 1987 and for introducing me to the geology of northern Devon Island in the summer of 1986. He made available unpublished reports, and also advised me on the locations of some key exposures, particularly on northern and central Ellesmere Island. Chris Harrison and Ray Thorsteinsson, in 1989, are to be thanked for many cheerful and enthusiastic discussions on the geology and for hiring me in the summer of 1989 for a major GSC mapping project in the vicinity of Vendom Fiord. Interpretations in the following report were considerably enhanced due to their geological mapping and understanding of the geology of the Franklinian Basin. Jisuo Jin, Alf Lenz, Sandy McCracken, Mike Melchin, Ray Thorsteinsson, and Tom Uyeno helped in the identification of conodonts, graptolites, and brachiopods, and Tom Bolton and Godfrey Nowlan were also available for discussions of various paleontological and other aspects of the geology of the study area. I benefited considerably from a

directed studies course organized by Godfrey Nowlan in the winter of 1987 at the University of Ottawa.

During my employment by the GSC, numerous field assistants helped in measuring and sampling the 8-10km of stratigraphic sections examined and are to be thanked. Discussions with colleagues at the University of Ottawa and at geological conferences enhanced some of my interpretations. Frank Brunton, Andre Desrochers, Noel James, Susan Kidwell, Markes Johnson, Terry Sami, and Martine Savard were particularly receptive to my inquiries. Mike Melchin is to be thanked especially for giving me a personal short course on graptolite taxonomy. I would like to particularly thank my wife Tina Opsal for offering her support and for typing much of this thesis.

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PART II

UPPER ORDOVICIAN TO LOWER DEVONIAN STRATIGRAPHY

ABSTRACT

The mottled dolomitic limestone of the Ordovician Thumb Mountain Formation is laterally homogeneous and likely indicative of regional platform stability and deposition on a carbonate ramp. The overlying Irene Bay Formation represents a regionally significant event, possibly abrupt sea-level rise, or, alternatively, the influx of cool, nutrient-rich waters onto the shelf that caused phosphogenesis, a reduction in carbonate sedimentation, hardgrounds, and affected the entire platform. The overlying limestone of the basal tongue of the Allen Bay Formation was deposited in somewhat deeper water than inferred for the Thumb Mountain Formation. Regional termination of carbonate deposition occurred subsequently in the *fastigatus* Zone, causing hundreds of kilometres of backstepping in the south, in the vicinity of Cornwallis Island, but only tens of kilometres in the north, in the vicinity of Ellesmere Island (not considering tectonic shortening). Pronounced sinuosity of the Silurian platform margin was possibly related to an underlying subsidence control, perhaps influenced by early movement of the Boothia Uplift.

Shallowing at or near the Ordovician-Silurian boundary is evident from disconformities in the platform carbonates and from conspicuous bioturbated lime mudstones straddling the system boundary in the deep-shelf shales. The coincidence of the systems boundary with the boundary between the lower and middle members of the Allen Bay Formation is questioned, and current biostratigraphy shows it to be

inconsistent; shallowing may have occurred several times in the upper-most Ordovician and lowest Silurian. Evidence of platform backstepping is recognized in the *fastigatus*, *acuminatus*, and *cyphus* zones, and very condensed sequences at the bases of onlapping shales occur in northeastern Ellesmere Island. Chert is common in the *cyphus* Zone, and was likely biogenically controlled and related to prolonged episodes of abundant Radiolaria growth that may have had a paleoceanographic influence.

Subsequent to backstepping in *fastigatus* Zone, isolated mud buildups developed over the drowned Ordovician shelf margin. Stratigraphic relationships suggest that these buildups were on older, long ranging crustal features, possibly tectonically influenced. Three Ellesmere Island buildups appear to have been synchronously drowned in the lower Wenlock. They are large structures, up to 1.3km in thickness, and at Cañon Fiord show lower, upward shallowing and upper, upward deepening sequences. The latter deepening phase was approximately coeval with Hazen Trough flysch deposition and concomitant flexure related to lithospheric loading, with platform termination in the area of North Greenland, and with the evolution of a passive pericratonic basin into an active foreland basin. The locally great thickness of predominantly subtidal carbonates in the upper part of the Allen Bay Formation at Darling Peninsula may be additional evidence for this orogenic episode.

Lower and Middle Silurian platform upbuilding on Cornwallis Island is evident from the vertical succession of platform margin and platform interior facies. The latter is represented by an upward replacement of muddy cyclic carbonates by tidally influenced high-frequency carbonate cycles, indicating a decrease in shelf accommodation overall. Inimical platform waters during minimum accommodation is thought to have influenced shelf margin facies by inhibiting reefal growth, but promoting the deposition of extensive carbonate sand shoals.

A major catch-up phase of carbonate deposition, followed the *cyphus* Zone backstepping on southern Ellesmere Island (Baumann Fiord). Over drowned Lower Silurian ramp carbonates, a mudmound and three or possibly four pinnacle reefs developed. The latter were built of coralgal biolithite that is newly interpreted as having been deposited in a wave-stressed environment that excluded stromatoporoid growth but favoured a sparse coral fauna thickly encrusted by calcareous algae and carbonate-secreting microbial colonies.

The Pridoli platform and basin succession is less well known. Pronounced platform progradation is recorded south and north of, but not in the vicinity of, Baumann Fiord. Coeval deep-shelf clastics also show a north-to-south variation. At Bay Fiord, grey siltstone (informally referred to as the lower grey siltstone member of the Red Canyon River Formation) likely represents a distal facies of a progradational clastic wedge, possibly related to the first and most significant phase of the tripartite Caledonian Inglefield Uplift, which likely began in latest Silurian.

INTRODUCTION

The origin of the Franklinian Basin is controversial. Extensional tectonism is thought to have been an important influence at two different times in the Proterozoic: 0.72Ga (Rainbird, 1990) and 1.2Ga (Trettin, 1989). Deposition of a thick clastic-carbonate succession began in the latest Proterozoic and was continuous until uplift of the basin in the Mississippian, although the precise tectonic interpretation of the basin is uncertain (Surlyk and Hurst, 1984). The thick sedimentary package possibly represents deposition over a "normal" passive continental margin, but the pervasive longitudinal transport of deep-water clastic sediment parallel to the continental margin suggests that a barrier of some sort was present to the north. Surlyk and Hurst (1984) suggested that if the basement is oceanic, a mid-oceanic ridge would have been a likely northern barrier; however, this interpretation is not supported by deep seismic surveys (Balkwill, *et al.*, 1986), which suggests an ensialic basement. "Pearya" is conveniently situated and was initially considered a likely candidate for the northern borderland, but it was subsequently reinterpreted as an allochthonous terrane which docked in latest Silurian or Early Devonian time (Trettin, 1987).

A major, thick, shallow-water carbonate succession, and a deep-water turbidite succession derived from the penultimate epeirogenesis of the Franklinian Basin, were deposited during Ordovician and Silurian time. Silurian carbonate platform deposition was mildly to greatly influenced by this major orogenic episode and was also subject to sea-level fluctuations at or near the Ordovician-Silurian boundary, related to Gondwana glaciation and deglaciation. Although often difficult, if not impossible to unravel tectonic, environmental, and eustatic influences on carbonate platform sedimentation, these controls are addressed in the following stratigraphic study. Major trends in deep-shelf and carbonate platform sedimentation are evaluated and compared to trends in the contiguous North Greenland carbonate platform and trough sediments. Some of this work is ongoing, as the study area is large and platform sequence very thick. Eighty two stratigraphic sections were examined in 3 seasons of field work (Figs.1-5), but this work still is considered incomplete, and many areas need re-examination or investigation for the first time.

IRENE BAY FORMATION (Oci)

The Irene Bay Formation was originally described by Kerr (1967) for the uppermost rubbly argillaceous limestone unit of the Cornwallis Group. Its type section is located approximately 15km north of section 12, northwest of Irene Bay (Fig.4). Because it is a recessive weathering unit occurring between two thickly bedded limestone successions, the Allen Bay Formation and the Thumb Mountain Formation, it is very distinctive, easily recognizable, and is used as a datum for the

Figure 3: Section locations for the Baumann Fiord-Makinson Inlet area.

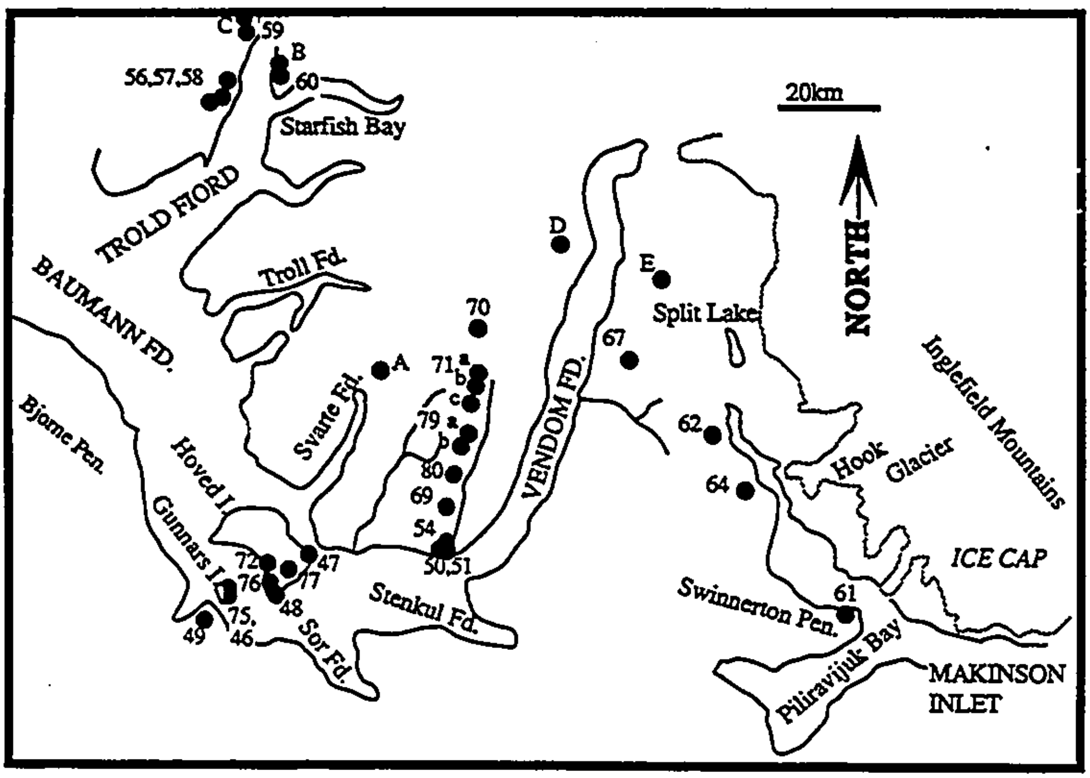

C
59
B
56,57,58
60
20km
Starfish Bay
TROLD FIORD
D
NORTH
Troll Fd.
E
BAUMANN FD.
70
Split Lake
67
A
71 a
b
c
Inglefield Mountains
Bjorne Pen.
Svarte Fd.
79 a
b
VENDOM FD.
62
Hoved I.
Hook Glacier
80
69
64
Gunnars I.
54
ICE CAP
72
47
50,51
76
77
Stenkul Fd.
Swinnerton Pen.
61
75
48
Sor Fd.
49
46
Piliravijuk Bay
MAKINSON INLET

Figure 4: Section locations for the Bay Fiord-Vesle Fiord area.

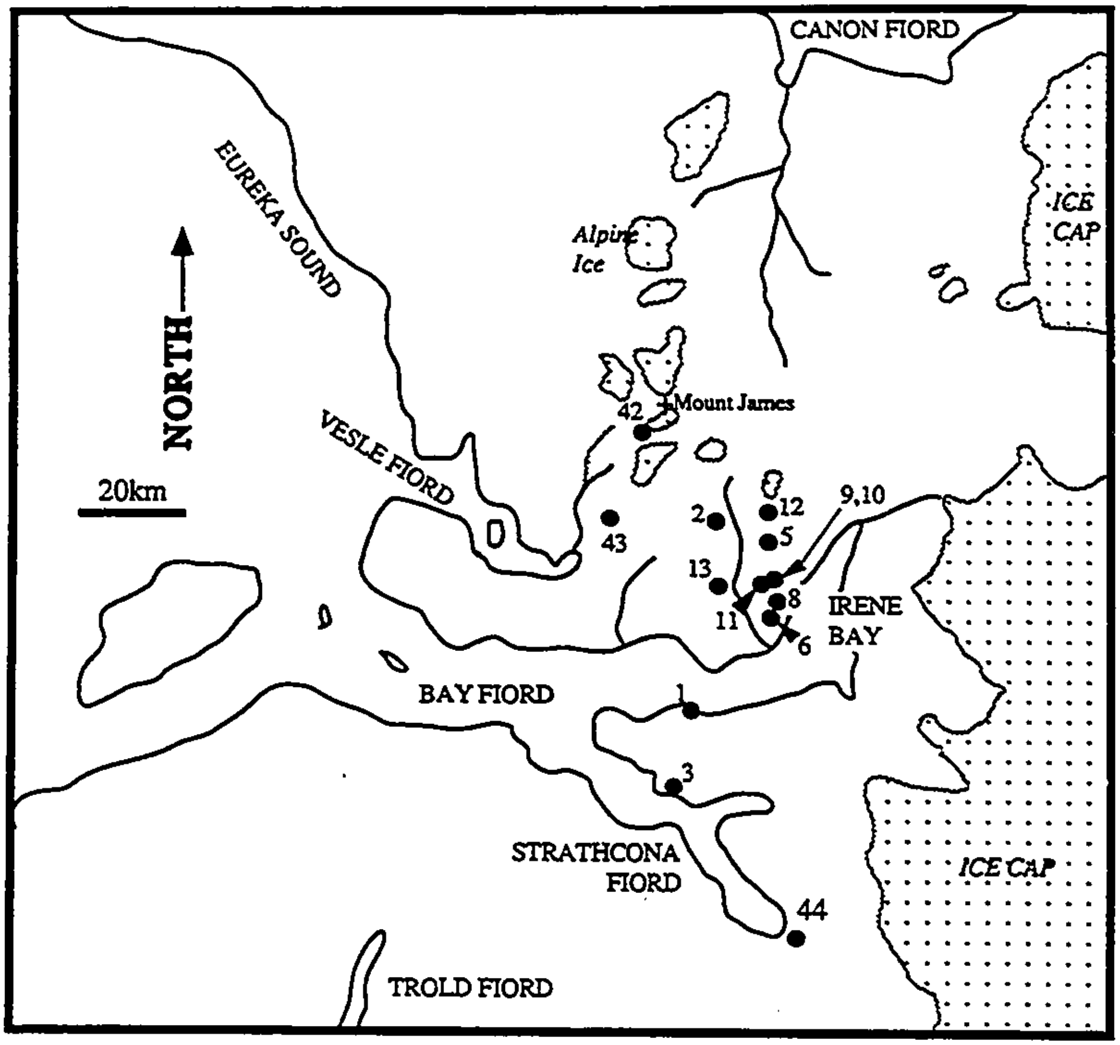
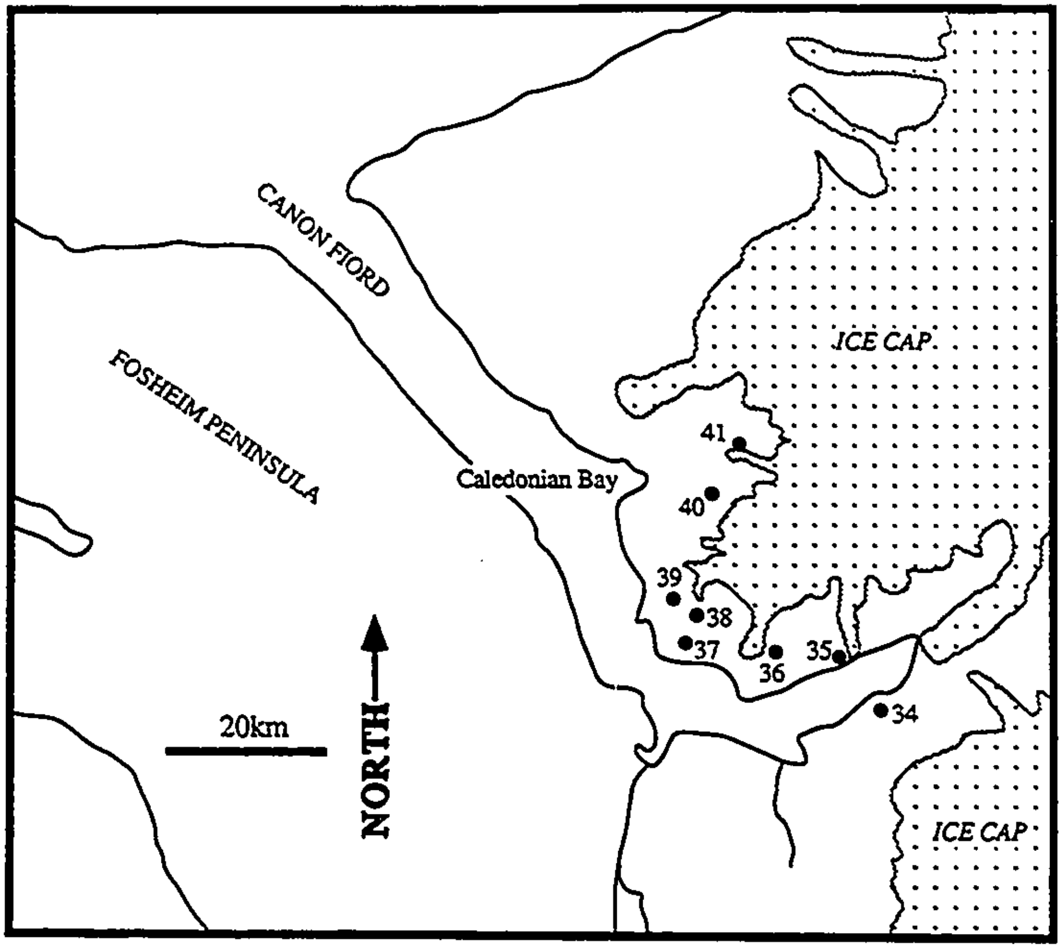

CANON FIORD
EUREKA SOUND
NORTH
Alpine
Ice
ICE
CAP
Mount James
42
VESLE FIORD
20km
2
12
9,10
43
5
13
8
11
6
IRENE
BAY
BAY FIORD
1
3
STRATHCONA
FIORD
ICE CAP
44
TROLD FIORD

Figure 5: Section locations for the Cañon Fiord-Caledonian Bay area.


CANON FIORD
FOSHEIM PENINSULA
ICE CAP
Caledonian Bay
41
40
39
38
37
36
35
34
20km
NORTH
ICE CAP

platform sequence in the following report.

Lithology and distribution

Throughout most of the study area, the formation is characterized by recessive, light green, rubbly weathering, fossiliferous, argillaceous, limestone bounded by massively bedded, resistant, mottled, dolomitic limestone of the Thumb Mountain Formation, below, and the Allen Bay Formation, above. Macrofossils are abundant and include *Maclurites* and other gastropods, *Receptaculites*, corals (including *Calapoecia*, *Paleofavosites*, *Manipora*, *Syringopora*, and other solitary coral forms), breviconic and orthoconic cephalopods, abundant trilobite fragments (?*Pseudogygites arcticus*), brachiopods, ichnofossils, and sponges (*Hindia sphaeroidalis*). Rocks throughout the formation are intensely bioturbated, and contain common phosphate and iron-sulphide-encrusted hardgrounds, particularly in the vicinity of Grinnell Peninsula.

On Ellesmere Island, there is a general westward decrease in thickness and fossil content and a change in characteristic lithology. At Trold Fiord, for example, (Table 1; Plate 8a), the presumably stratigraphically equivalent unit is thin, largely unfossiliferous, although locally cephalopod-rich, and characterized by thin beds of laminated, argillaceous limestone and regularly interbedded fissile black clayshale. At Bay Fiord, in a presumably similar position with respect to the shelf edge as the Trold Fiord occurrence, the Irene Bay Formation is notably and typically thicker.

TABLE 1: Thicknesses of the Irene Bay Fm in study area (numbered sections as in Fig.1, letters as in Fig.3)

Section	Thickness
A	66m
B	7.5
C	4.5
D	42
E	22
20	44
31	81
32	40
36	30
54	63
62	46.4
64	52
N. Somerset[3]	43
S. Devon I.[5]	6-16.5
Cornwallis I.[4]	8.5-10.3
Irene Bay (type)[1]	82.3
Vesle Fd.[1]	60.9
Brodeur Pen.[2] (Baffin I.)	20.1
Strathcona Fiord[1]	45.7
Starfish Bay[1]	121.9-320

[1]Kerr, 1967
[2]Mayr, 1978
[3]Miall & Kerr, 1980
[4]Thorsteinsson and Uyeno, 1980
[5]Thorsteinsson and Mayr, 1987

Receptaculitids are important components and locally form a boundstone. They occur together with abundant and occasionally very large conic cephalopods. These and other macrofossils generally show a westward decline in numbers and eventually disappear.

In addition to the studied localities of this report, the formation, or its equivalent, is recognized on Baffin Island (Member B of the Baillarge Bay Formation, Trettin, 1969), on Somerset and Prince Patrick Islands (Miall and Kerr, 1980; Stewart, 1987), on Cornwallis Island (Thorsteinsson and Uyeno, 1980; Thorsteinsson, 1958, Thorsteinsson and Kerr, 1968); and on Devon Island (Morrow and Kerr, 1977; Thorsteinsson and Mayr, 1987). The formation is synonymous with the Cape Calhoun Formation of North Greenland. It is, however, not recognized as a separate formation on Victoria Island (Tozer and Thorsteinsson, 1964). Kerr (1967) described an anomalously thick section on Trold Fiord, but he undoubtedly included much of the underlying Thumb Mountain Formation, which on the eastern side of Trold Fiord has a thick upper member of fossiliferous, rubbly-weathering Irene Bay-like strata. The thin interbedded limestone-shale succession that overlies the Irene Bay Formation in most drowned shelf localities, for example, as near Trold Fiord, is assignable to the basal tongue of the Allen Bay Formation (*sensu* Plate 8a).

Recognition of the Irene Bay Formation at, or just east of, Trold Fiord, using its typical lithology, is problematical, particularly where it overlies rubbly green-weathering limestones of the upper part of the Thumb Mountain Formation. However, on the western side of the fiord, the Thumb Mountain Formation is somewhat more typically massively bedded and is overlain by a thin shaly unit, then a thin, approximately 2-3m thick limestone bed which is gradationally overlain by the Cape Phillips Formation. This thin shaley unit is predominantly a fissile black clayshale rhythmically interbedded with bioturbated argillaceous limestone which locally contains abundant, small orthoconic cephalopods. Because of its stratigraphic position below a widely correlative graptolite zone and above a massively bedded Thumb Mountain succession, the unit is correlative with the Irene Bay Formation known on central Ellesmere Island. This shaley succession could be considered a distinct lithostratigraphic unit, but, in this investigation, correlation of the shaly and rubbly units is justified based on the inferred depositional history of the formation (discussed further below), in that the Irene Bay Formation likely represents a regionally isochronous unit. Furthermore, the correlation is valid based on art.22(d) in the North American Stratigraphic Code (1985): "...Inferred geological history, depositional environment, and biological sequence have no place in the definition of a lithostratigraphic unit, which must be based on composition and other lithic characteristics; nevertheless, considerations of well documented geological history properly may influence the choice of vertical and lateral boundaries..." (*op. cit.*, p.742).

On Somerset Island, the Irene Bay Formation is partly eroded beneath a sub-Allen Bay disconformity (Miall and Kerr, 1980; Stewart, 1987), and on Devon and southwest Ellesmere Island, the lower member of the Allen Bay Formation contains a thin rubbly limestone unit that is lithologically very similar to the Irene Bay Formation.

The Irene Bay generally shows cratonward *and* basinward thinning, but is thickest approximately in medial shelf areas in a band running parallel to the basement exposure and shelf margin in about the middie of the outcrop belt (in the areas of Baumann and Bay fiords (Table 1); however, as discussed below, it is difficult in some areas to recognize the contact between the upper member of the Thumb Mountain Formation and the Irene Bay Formation, both units locally having a rubbly weathering aspect to their lithology. In more platformward stratigraphic sections the upper member of the Thumb Mountain Formation gives way to a more thickly bedded, less argillaceous limestone that is readily differentiated from the recessive, rubbly weathering Irene Bay Formation.

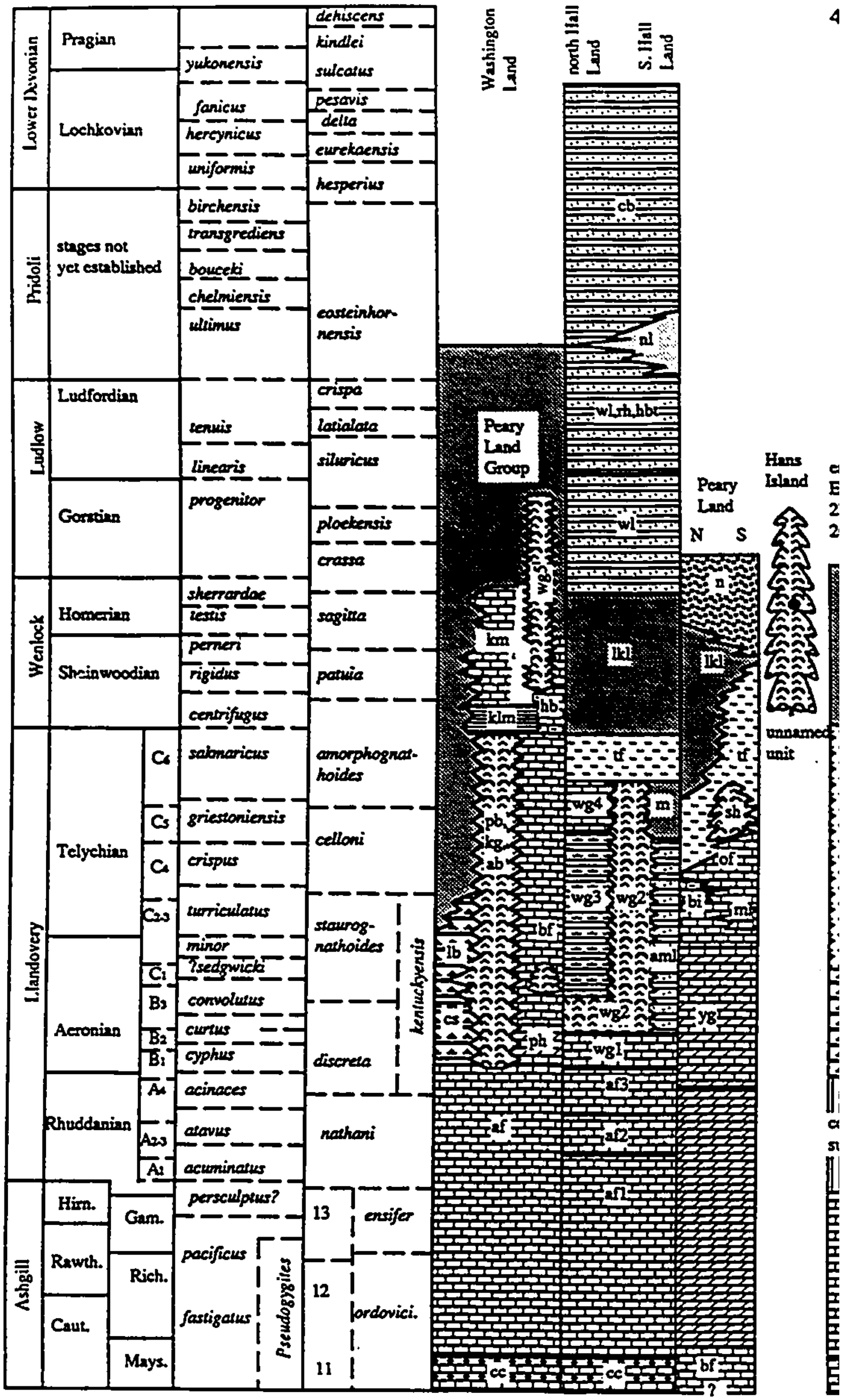
The age of the formation is well established. Conodonts and macrofossils collected in many areas of the Arctic have established the age of the formation as Late Ordovician, faunas 11 and 12, Maysvillian to Richmondian (Barnes, 1974; Barnes, *et al.*,, 1976; Kerr, 1967, Mayr, 1978; Thorsteinsson and Mayr, 1987; also this report). This age appears to be fairly consistent throughout the Arctic and is reported from a number of study areas. Conodonts were collected only in the area consistent with the probable age (Figs.6a-c). The graptolite *Orthograptus fastigatus* was recovered from numerous sections at Trold Fiord (sections 56, 57, 58, 59, 60; Figs.7f, g) several meters above the Irene Bay Formation, indicating an Upper Ordovician Maysvillian to Richmondian age. Similarly, the *fastigatus* Zone was recognized in sections 20 and 21, Grinnell Peninsula, some 24m above the Irene Bay Formation (Fig.8a-b); however, up to 40m of beds, assignable to this graptolite zone were recognized at Trold Fiord, section 58. Although, based entirely on lithological grounds, the formation appears to be a partly condensed sequence, age-diagnostic macrofossils recovered above and below the formation have not indicated a significant hiatus.

Interpretation

The Irene Bay Formation represents deposition in open marine waters below wave base. Mineralized hardgrounds and fossil-rich beds suggest that the sequence is condensed, although biostratigraphically unproven. Common calcareous algae (receptaculitids) indicate deposition in the photic zone, but the relative increase in the abundance of dasyclads in the formations above and below may indicate subtle photic differences during deposition. Whether these differences were related to depth or to other factors, such as turbidity, is unknown.

Several facts suggest that an abrupt sea-level rise was associated with deposition of the Irene Bay Formation:

Figure 6a-c: Stratigraphic chart of the formations discussed in this report. Stratigraphy of North Greenland is based on Escher and Larsen (1987); Hurst (1984); Sønderholm *et al.* (1987); Hurst *et al.* (1983); Hurst and Surlyk (1982, 1983, 1984); Hurst (1979, 1980); Dawes and Peel (1984); and Peel and Hurst (1980). The biostratigraphic scheme is based primarily on discussions with A.C. Lenz, A.D. McCracken, and a written communiqué from G. Nowlan. Other studies used to establish the correlation of graptolites and conodont zonal schemes include Thorsteinsson and Uyeno (1980); Lenz (1978, 1979, 1982, 1988); Cherkesova (1988); Norford, (1988); Barnes and Bergström (1988); Jell and Talent (1989); Lenz and Melchin, (1990); Bjerreskov (1986); Jackson *et al.* (1978); and Berry and Murphy (1975). This correlation is by no means conclusive; many of the biozone boundaries may need readjustment after further biostratigraphic work. The occurrence of *celloni* Zone conodonts in the *turriculatus* Zone in section 47 is an exception to the usual correlation of these two schemes, and several more cases are discussed in the text. Boundaries of the graptolite (left) and conodont (right) biozones are shown on the left side of Figure 6a for convenience. Legend in 6b is used throughout Part II.



Lower Devonian
Pragian
Lochkovian
yukonensis
fanicus
hercynicus
uniformis
dehiscens
kindlei
sulcatus
pesavis
delta
eurekaensis
hesperius
Pridoli
stages not yet established
birchensis
transgrediens
bouceki
chelmiensis
ultimus
eosteinhornensis
Ludlow
Ludfordian
Gorstian
tenuis
linearis
progenitor
crispa
latialata
siluricus
ploekensis
crassa
Wenlock
Homerian
Sheinwoodian
sherrardae
testis
perneri
rigidus
centrifugus
sagitta
patula
Llandovery
Telychian
Aeronian
Rhuddanian
C6
C5
C4
C2-3
C1
B3
B2
B1
A4
A2-3
A1
salmaricus
griestoniensis
crispus
turriculatus
minor
?sedgwicki
convolutus
curtus
cyphus
acinaces
atavus
acuminatus
amorphognathoides
celloni
staurognathoides
kentuckyensis
discreta
nathani
Ashgill
Hirn.
Rawth.
Caut.
Gam.
Rich.
Mays.
persculptus?
pacificus
fastigatus
Pseudogygites
13
12
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ensifer
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Washington Land
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S. Hall Land
Peary Land Group
Peary Land
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Hans Island
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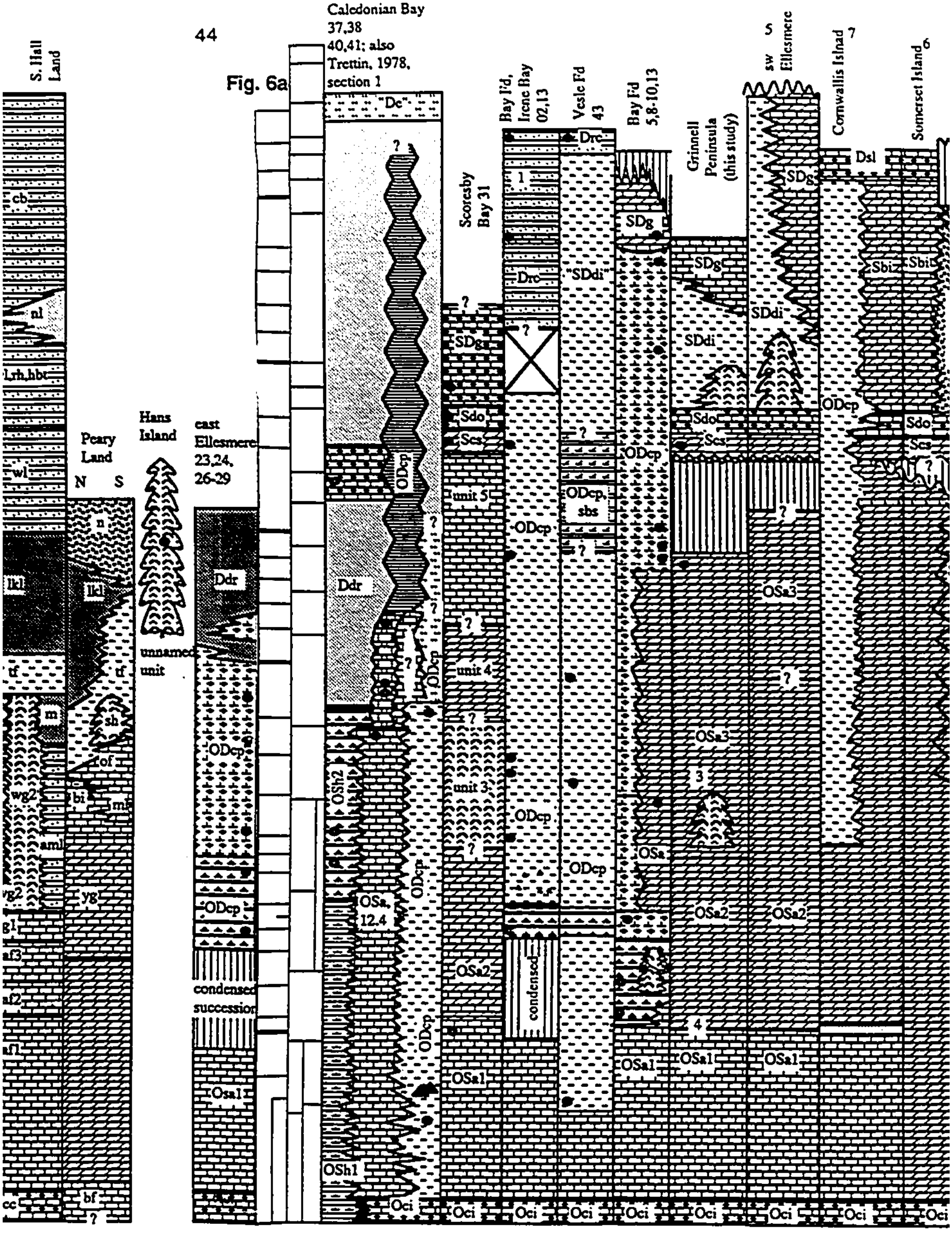
Fig. 6a
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Caledonian Bay 37,38 40,41; also Trettin, 1978, section 1
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Bay Fd, Irene Bay 02,13
Vesle Fd 43
Bay Fd 5,8-10,13
Grinnell Peninsula (this study)
5 sw Ellesmere
Cornwallis Island 7
Somerset Island 6
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Lrh,hbt
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OSh2
OSa, 12,4
OSh1
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Scs
unit 5
unit 4
unit 3
OSa2
OSa1
Drc
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SDg
OSa
SDdi
OSa3
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Dsl
Sbi
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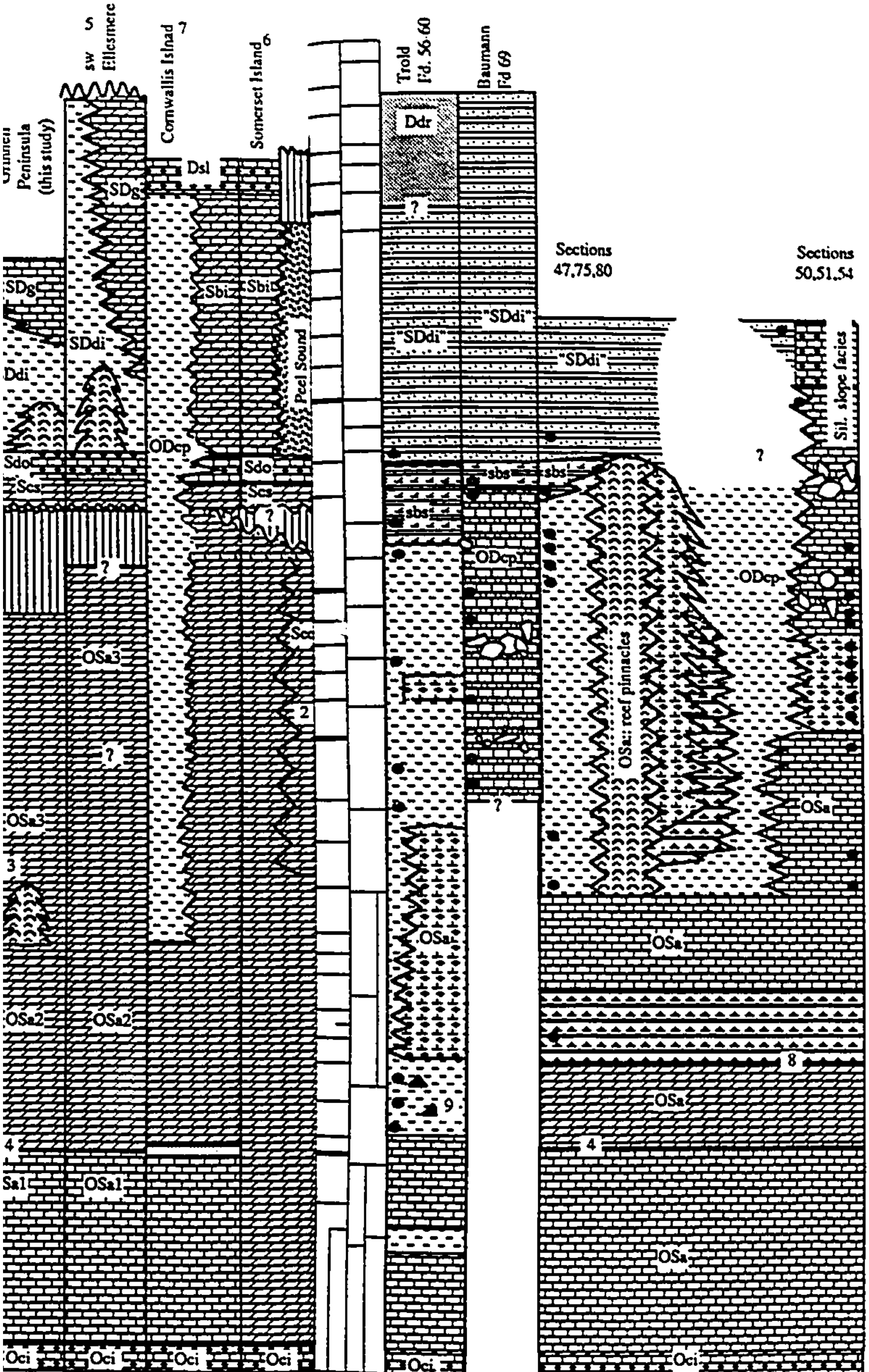

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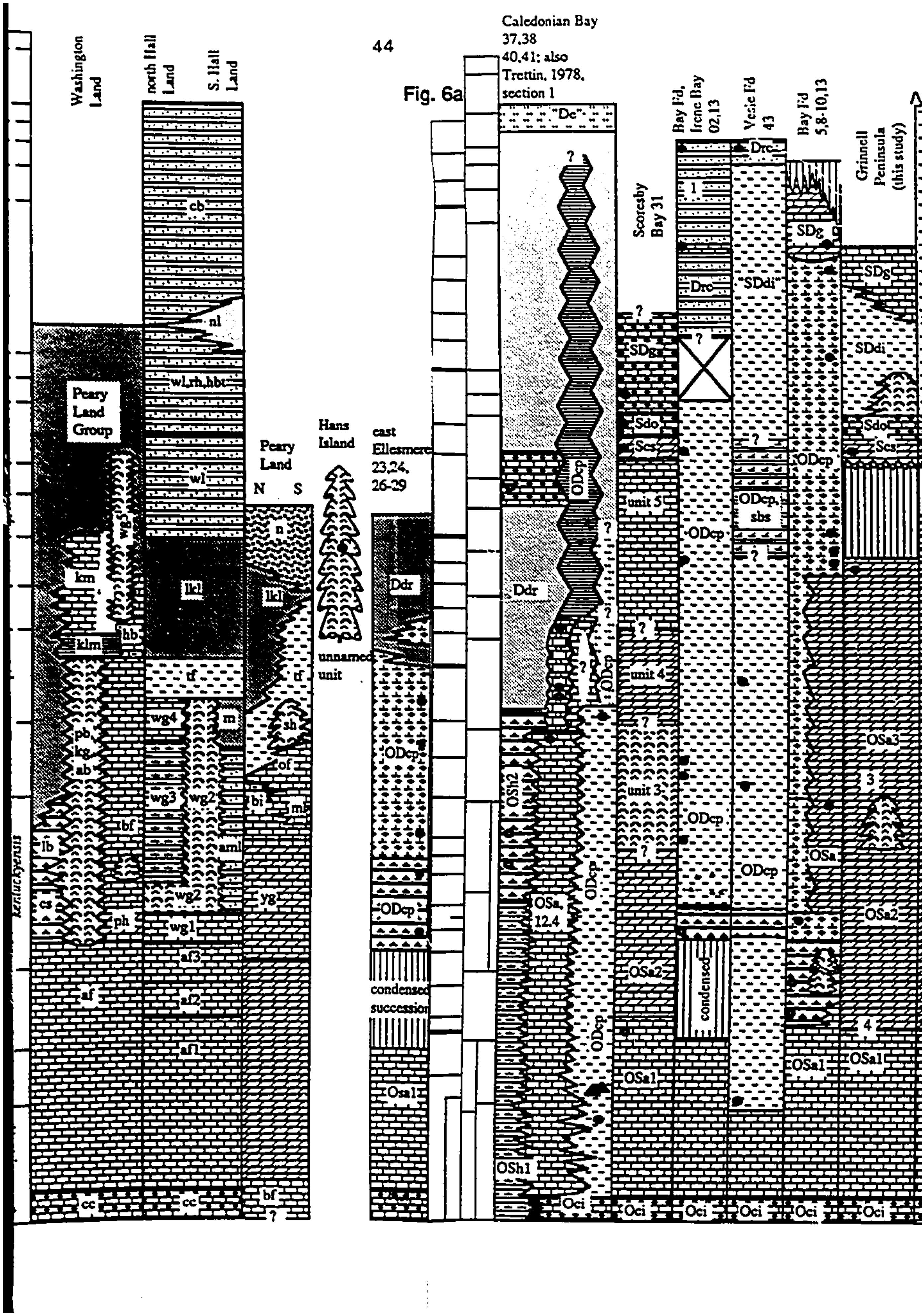
Washington Land
north Hall Land
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east Ellesmere 23,24, 26-29
Caledonian Bay 37,38 40,41; also Trettin, 1978, section 1
Scoresby Bay 31
Bay Fd, Irene Bay 02,13
Vesle Fd 43
Bay Fd 5,8-10,13
Grinnell Peninsula (this study)
Peary Land Group
kentuckyensis
cb
nl
wl,rh,hbt
wl
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wg5
wg4
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wg2
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pb, kg, ab
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unnamed unit
Ddr
ODcp
condensed succession
Osa1
"Dc"
OSh2
OSh1
OSa, 12,4
Oci
SDg
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unit 5
unit 4
unit 3
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SDdi
OSa3
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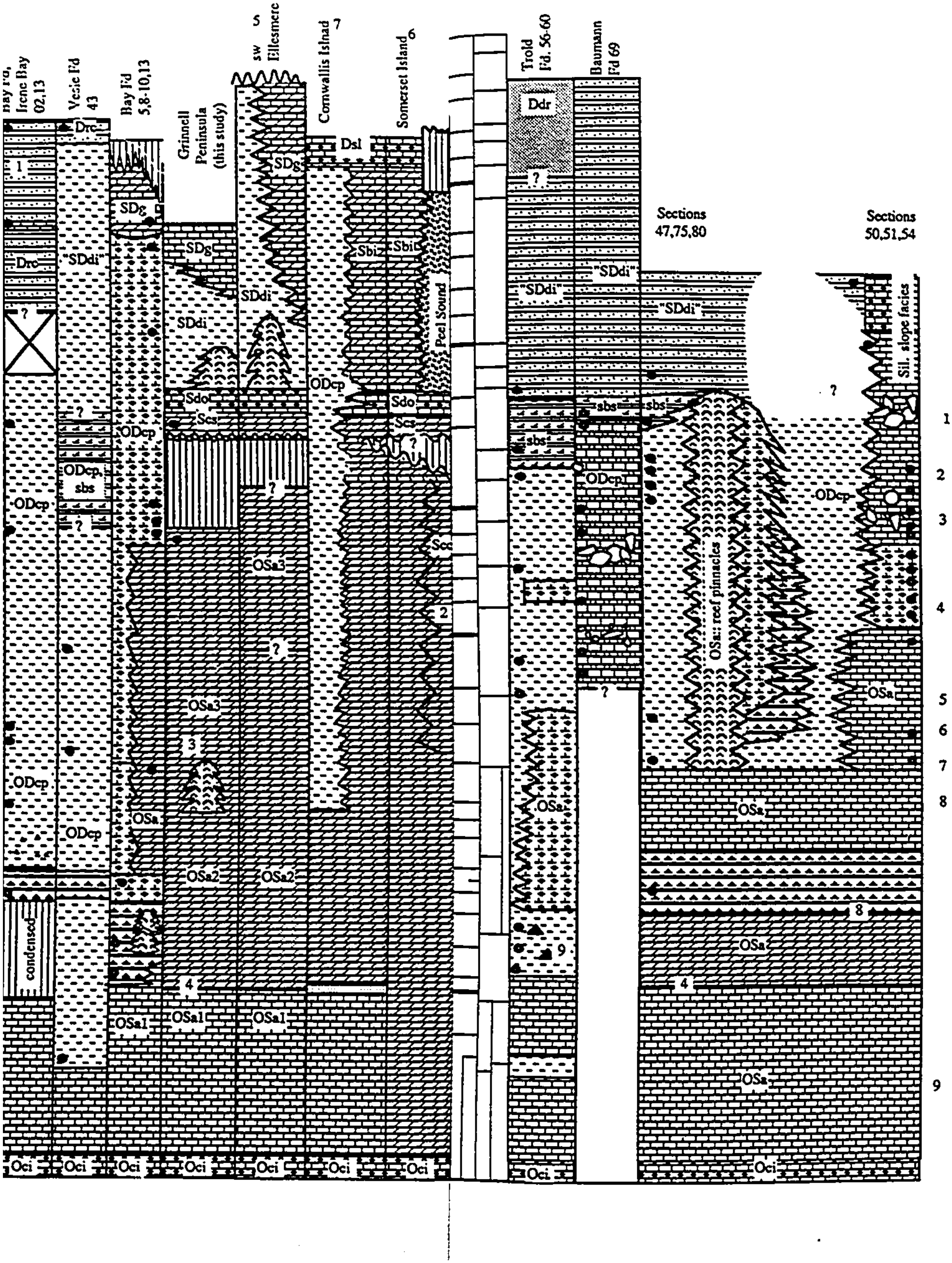
Fig. 6a



Bay Fd, Irene Bay 02,13
Vesle Fd 43
Bay Fd 5,8-10,13
Grinnell Peninsula (this study)
sw Ellesmere 5
Cornwallis Island 7
Somerset Island 6
Peel Sound
Trold Fd. 56-60
Baumann Fd 69
Sections 47,75,80
Sections 50,51,54
Sil. slope facies
OSa: reef pinnacles
Drc
Ddr
Dsl
SDg
"SDdi"
SDdi
Sbi
Sbit
Sdo
Scs
Scc
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ODcp
ODcp, sbs
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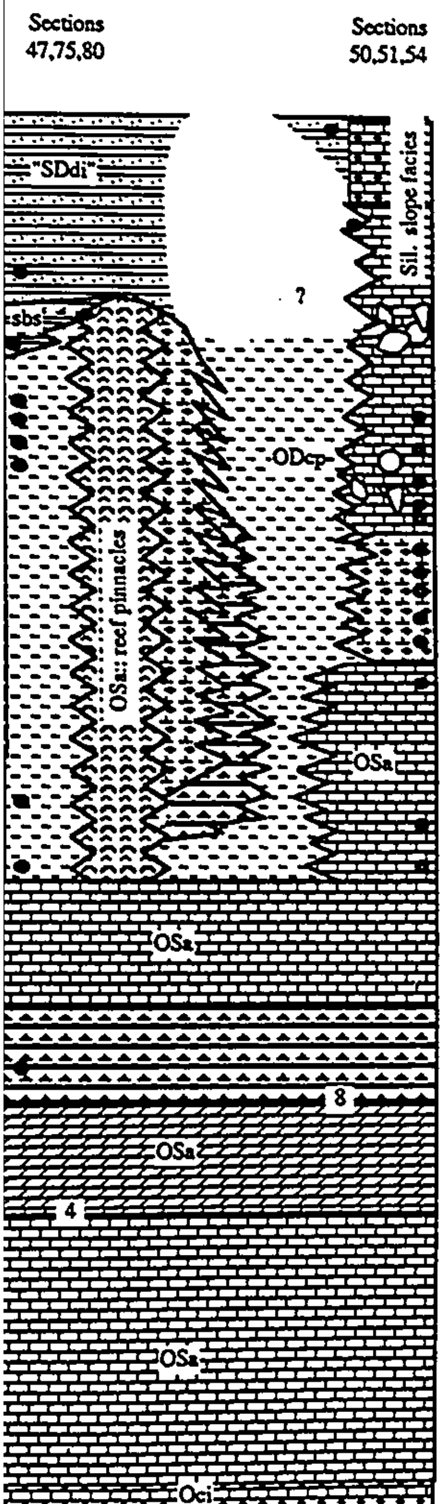


1 Distal siltstone facies of the Red Canyon River Fm

2 The Cape Crauford Fm contains a significant amount of evaporites

3 There is a shelf margin limestone facies on Grinnell Peninsula; however, its biostratigraphy is poor, and the youngest age represented by this facies is unknown.

4 See text for further discussion. The OSa lower limestone mbr-middle dolostone mbr contact does not necessarily correspond to the Ord.-Sil. boundary.

5 from Mayr *et al.*, in prep

6 Thorsteinsson and Uyeno, 1980; Stewart, 1987

7 Thorsteinsson and Uyeno, 1980

8 The age of this contact is uncertain. Poey (1988) intuitively concludes that the boundary coincideswith the Ordovician-Silurian boundary, but Thorsteinsson (pers. com., 1990) believes the boundary is some 100m below at the contact of mottled dolomitic limestone and petroliferous dolostone. Neither argument can irrefutably be proven or disproven, but the thick limestone succession suggests a considerable amount of time is represented by these carbonates, and that perhaps the boundary rests in the carbonates and not in the shales as Poey has suggested.

9 condensed succession

Fig.6b

LEGEND

- limestone
- dolostone
- dolomitic mudrock
- mudrock
- siltstone/mudrock
- dolostone/limestone
- olistostromes and related foreslope facies
- sandstone/siltstone ("flysch")
- nodular limestone
- carbonate buildup
- marlstone/limestone
- clayshale (fissile black shale)
- argillaceous limestone
- conglomerate (other than olistostromes)
- chert-marlstone-clayshale interbeds
- sandstone
- siltstone
- siltstone/carbonate
- oolite
- limestone/siltstone
- marlstone/limestone
- limestone/calcareous mudrock
- limestone/fissile shale partings
- marlstone/shale
- chert
- fossil age
- crinoid material
- fossil debris
- microbialite
- graptolite
- cross beds/lamination
- mottling
- trilobite
- oolite
- lamination
- bioturbation
- synaeresis cracks
- colonial coral
- stromatolite
- patterned carbonate
- cephalopod
- ostracod
- intraclast bed
- wavy or flaser lamination
- trough cross bed
- solitary coral
- stromatoporoid
- desiccation cracks
- bryozoan
- brachiopod
- oncolite

Canada

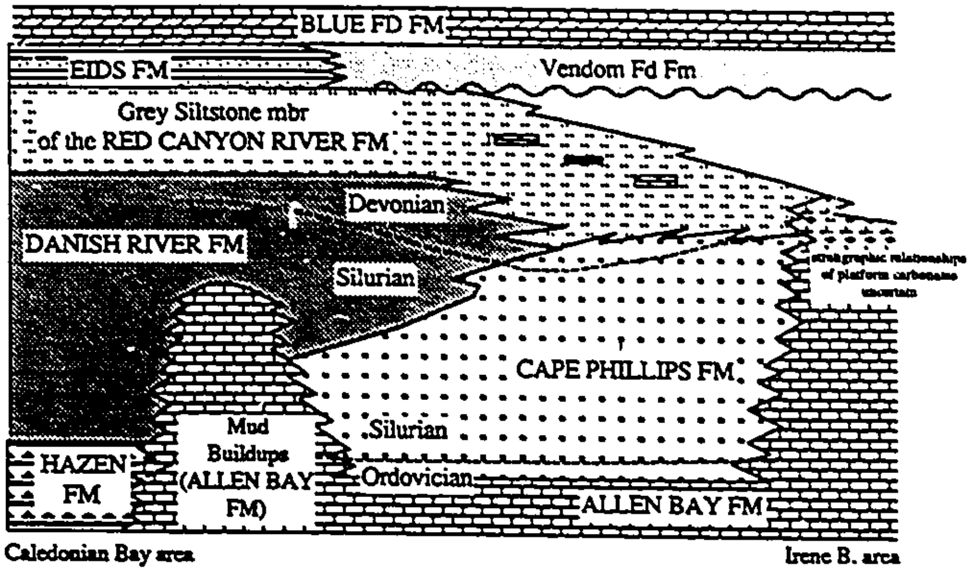
Ddr	Danish River Fm
De	Eids Fm
Dps	Peel Sound Fm
Dsl	Sophia Lake Fm
Drc	Red Canyon River Fm
Dsi	Somerset Island Fm
Dv	Vendom Fiord Fm
Oci	Irene Bay Fm
ODcp	Cape Phillips Fm
OSa	Allen Bay Fm (1: lower limestone member; 2: middle dolostone member etc.)
OSh	Hazen Fm
Sbi	Barlow Inlet Fm
Scc	Cape Crauford Fm
Scs	Cape Storm Fm
Sddi	Devon Island Fm
"SDdi"	Devon Island Fm – provisional assignment
SDg	Goose Fiord Fm
Sdo	Douro Fm
sbs	Starfish Bay shale ("SDdi")

Greenland

ab	Adams Bjerg Fm
af	Aleqatsiaq Fiord Fm
aml	Amundsen Land Gp
b	Borland River Fm
bf	Bessels Fiord Fm
bi	Bare Island Mbr (of)
cc	Cape Calhoun Fm
cb	Chester Bjerg Fm
cs	Cape Schuchert Fm
hb	Hauge Berge Fm
hbt	Hand Bugt Fm
kg	Kap Godfred Hansen Fm
klm	Kap Lucie Marie Fm
km	Kap Morton Fm
lb	Lafayette Bught
lkl	Lauge Koch Land Fm
m	Merqujoq Fm
ml	Melville Land Fm (of)
n	Nordkronen Fm
nl	Nyeboe Land Fm
of	Odins Fiord Fm
pb	Pentamerus Bjerg Fm
ph	Peterman Halvo Fm
rh	Repulse Haven Mbr (wl)
sh	Samuelsen Hag Fm
t	Tureso Fm
tf	Thors Fiord Mbr (wl)
wg	Washington Land Group (wg1: units defined by Sonderholm and others, 1987)
wl	Wulf Land Fm
yg	Ymers Gletscher Fm

Fig.6c

Generalized stratigraphic relationships on central Ellesmere I.



Generalized stratigraphic relationships on southern Ellesmere I.

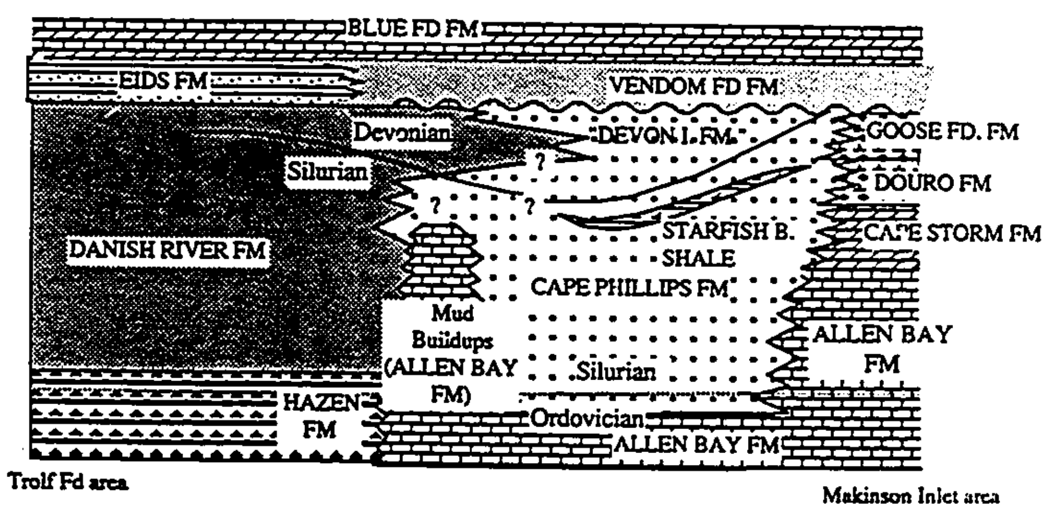
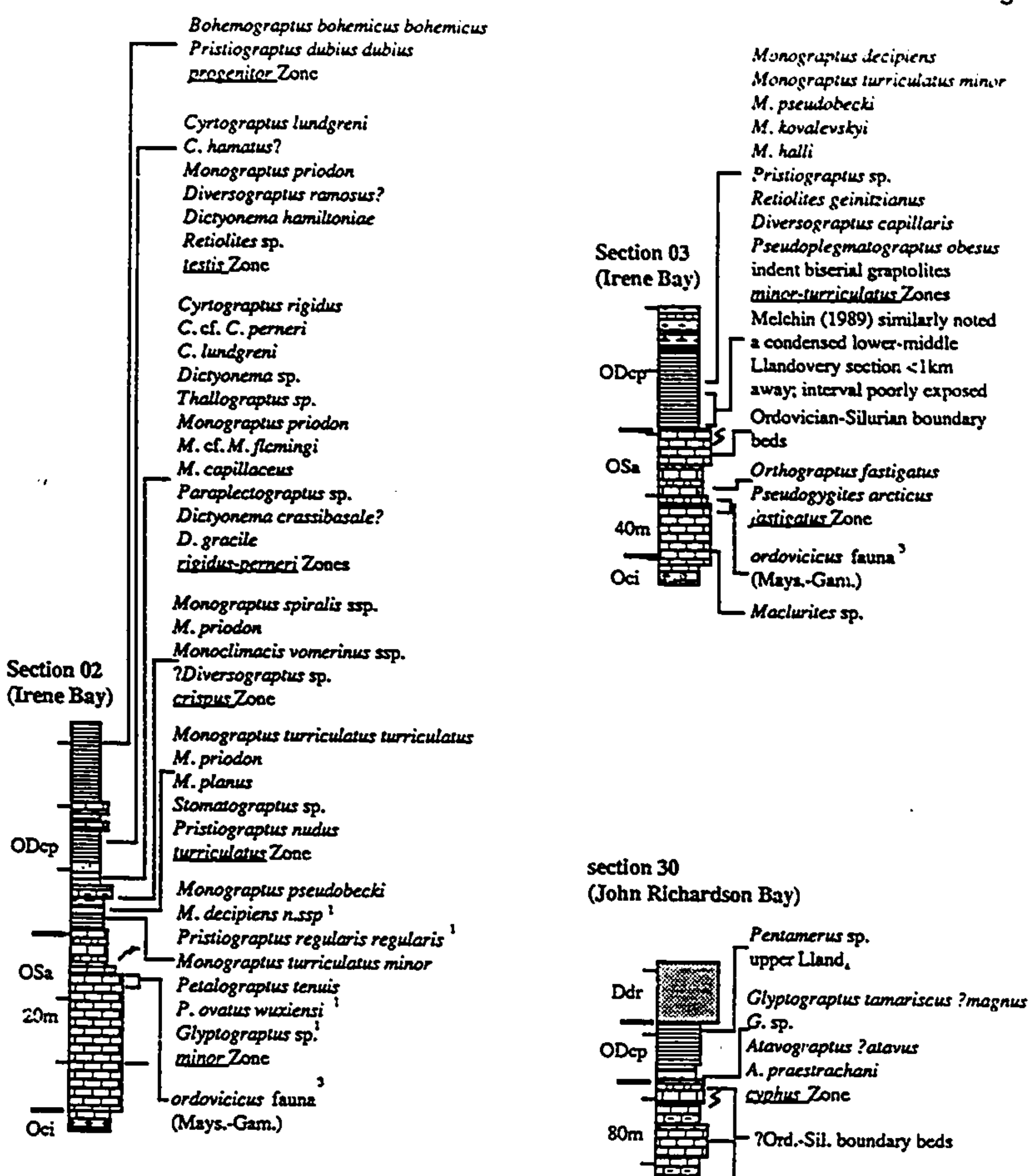


Figure 7a-g: Stratigraphic sections containing the Ordovician-Silurian boundary and the the Irene Bay-Allen Bay, Allen Bay-Cape Phillips, and the Danish River-Cape Phillips formation boundaries. The positions and zonal assignments of biostratigraphically important fossil collections are shown (graptolite assemblages also listed in Appendix two). Lithological and other symbols as in Fig.6d. These sections are correlated diagrammatically in the regional cross sections of Figs. 18, 22, 25, 26).

Fig. 7a



[1] Identification by M.J.Melchin
[2] Identification by A.C.Lenz
[3] Identification by A.D.McCracken
[4] Field identification

Derived graptolite biozone of each sample is underlined.

Fig. 7b

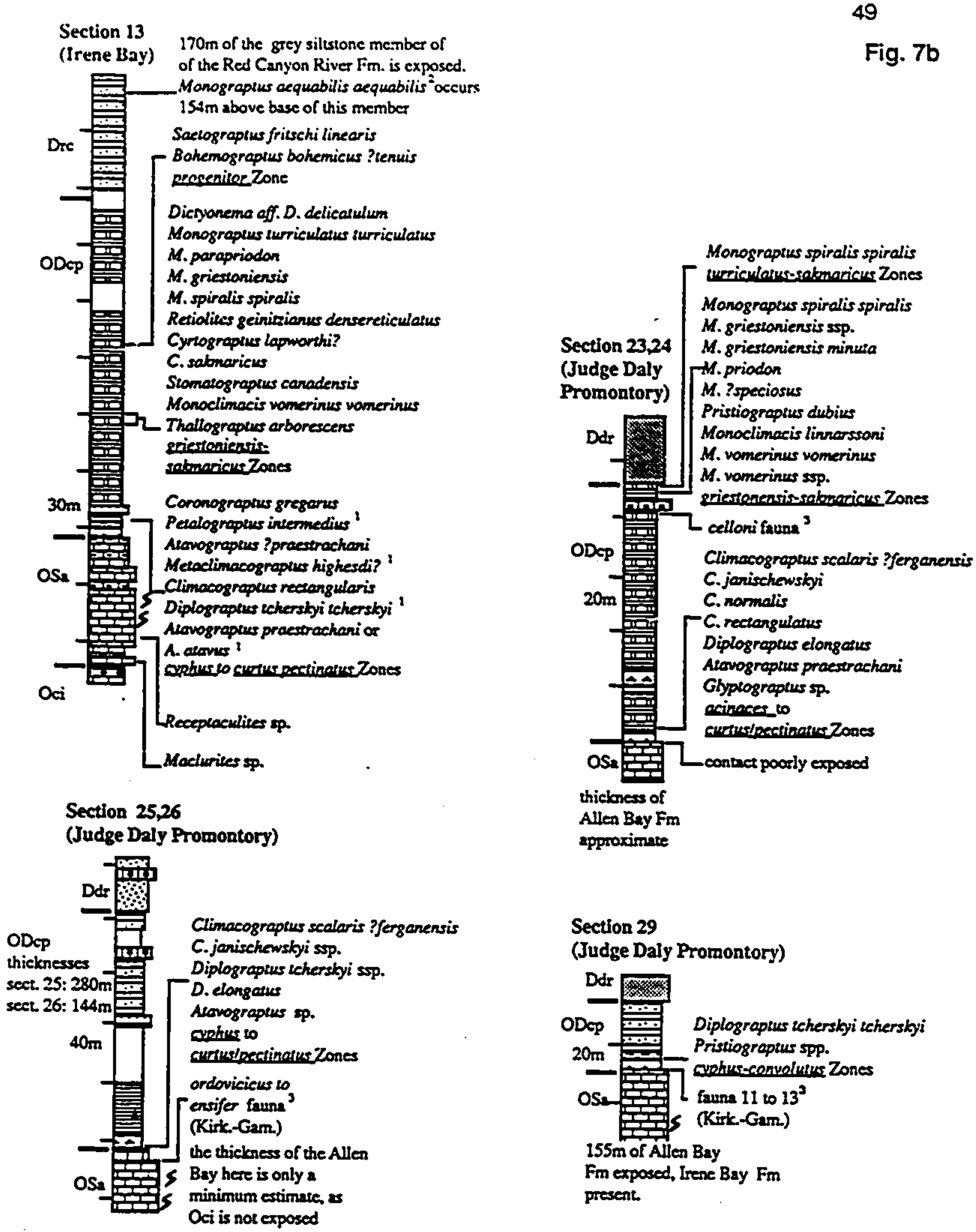
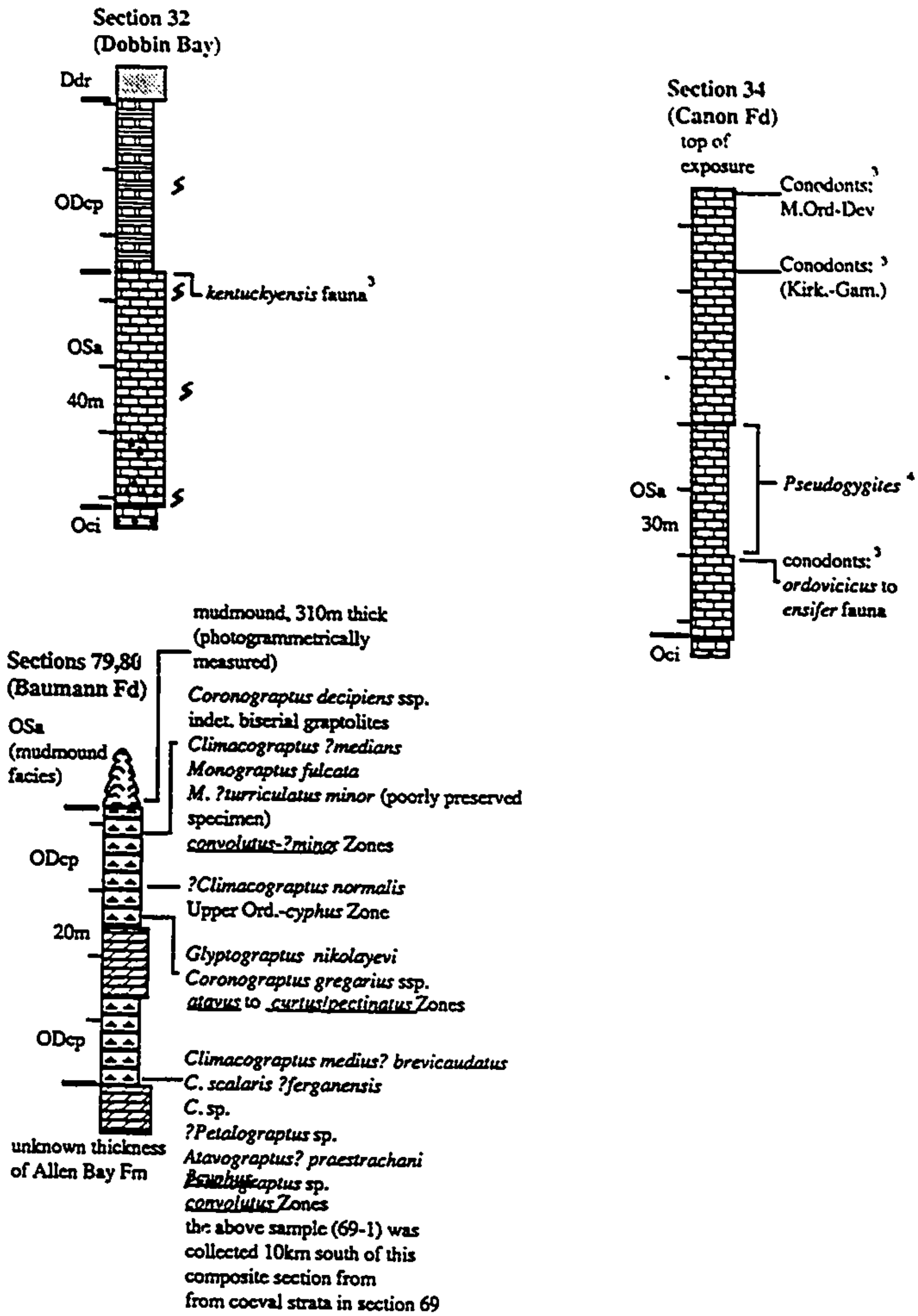

Section 13
(Irene Bay)
170m of the grey siltstone member of
of the Red Canyon River Fm. is exposed.
Monograptus aequabilis aequabilis[2] occurs
154m above base of this member
Drc
Saetograptus fritschi linearis
Bohemograptus bohemicus ?tenuis
progenitor Zone
Dictyonema aff. D. delicatulum
Monograptus turriculatus turriculatus
M. parapriodon
ODcp
M. griestoniensis
M. spiralis spiralis
Retiolites geinitzianus densereticulatus
Cyrtograptus lapworthi?
C. sakmaricus
Stomatograptus canadensis
Monoclimacis vomerinus vomerinus
Thallograptus arborescens
griestoniensis-
sakmaricus Zones
30m
Coronograptus gregarus
Petalograptus intermedius [1]
Atavograptus ?praestrachani
OSa
Metaclimacograptus hughesdi? [1]
Climacograptus rectangularis
Diplograptus tcherskyi tcherskyi [1]
Atavograptus praestrachani or
A. atavus [1]
cyphus to curtus pectinatus Zones
Oci
Receptaculites sp.
Maclurites sp.
Section 23,24
(Judge Daly
Promontory)
Monograptus spiralis spiralis
turriculatus-sakmaricus Zones
Monograptus spiralis spiralis
M. griestoniensis ssp.
M. griestoniensis minuta
M. priodon
M. ?speciosus
Pristiograptus dubius
Ddr
Monoclimacis linnarssoni
M. vomerinus vomerinus
M. vomerinus ssp.
griestonensis-sakmaricus Zones
celloni fauna [3]
ODcp
Climacograptus scalaris ?ferganensis
C. janischewskyi
20m
C. normalis
C. rectangulatus
Diplograptus elongatus
Atavograptus praestrachani
Glyptograptus sp.
acinaces to
curtus/pectinatus Zones
OSa
contact poorly exposed
thickness of
Allen Bay Fm
approximate
Section 25,26
(Judge Daly Promontory)
Ddr
Climacograptus scalaris ?ferganensis
ODcp
C. janischewskyi ssp.
thicknesses
Diplograptus tcherskyi ssp.
sect. 25: 280m
D. elongatus
sect. 26: 144m
Atavograptus sp.
40m
cyphus to
curtus/pectinatus Zones
ordovicicus to
ensifer fauna [3]
(Kirk.-Gam.)
the thickness of the Allen
OSa
Bay here is only a
minimum estimate, as
Oci is not exposed
Section 29
(Judge Daly Promontory)
Ddr
ODcp
Diplograptus tcherskyi tcherskyi
20m
Pristiograptus spp.
cyphus-convolutus Zones
OSa
fauna 11 to 13[3]
(Kirk.-Gam.)
155m of Allen Bay
Fm exposed, Irene Bay Fm
present.

Fig. 7c



Section 32
(Dobbin Bay)
Ddr
ODcp
OSa
40m
Oci
kentuckyensis fauna 3
Section 34
(Canon Fd)
top of
exposure
Conodonts: 3
M.Ord-Dev
Conodonts: 3
(Kirk.-Gam.)
OSa
30m
Oci
Pseudogygites 4
conodonts: 3
ordovicicus to
ensifer fauna
Sections 79,80
(Baumann Fd)
OSa
(mudmound
facies)
ODcp
20m
ODcp
unknown thickness
of Allen Bay Fm
mudmound, 310m thick
(photogrammetrically
measured)
Coronograptus decipiens ssp.
indet. biserial graptolites
Climacograptus ?medians
Monograptus fulcata
M. ?turriculatus minor (poorly preserved
specimen)
convolutus-?minor Zones
?Climacograptus normalis
Upper Ord.-cyphus Zone
Glyptograptus nikolayevi
Coronograptus gregarius ssp.
atavus to curtus/pectinatus Zones
Climacograptus medius? brevicaudatus
C. scalaris ?ferganensis
C. sp.
?Petalograptus sp.
Atavograptus? praestrachani
~~Rastrites~~ graptus sp.
convolutus Zones
the above sample (69-1) was
collected 10km south of this
composite section from
from coeval strata in section 69

Fig. 7d

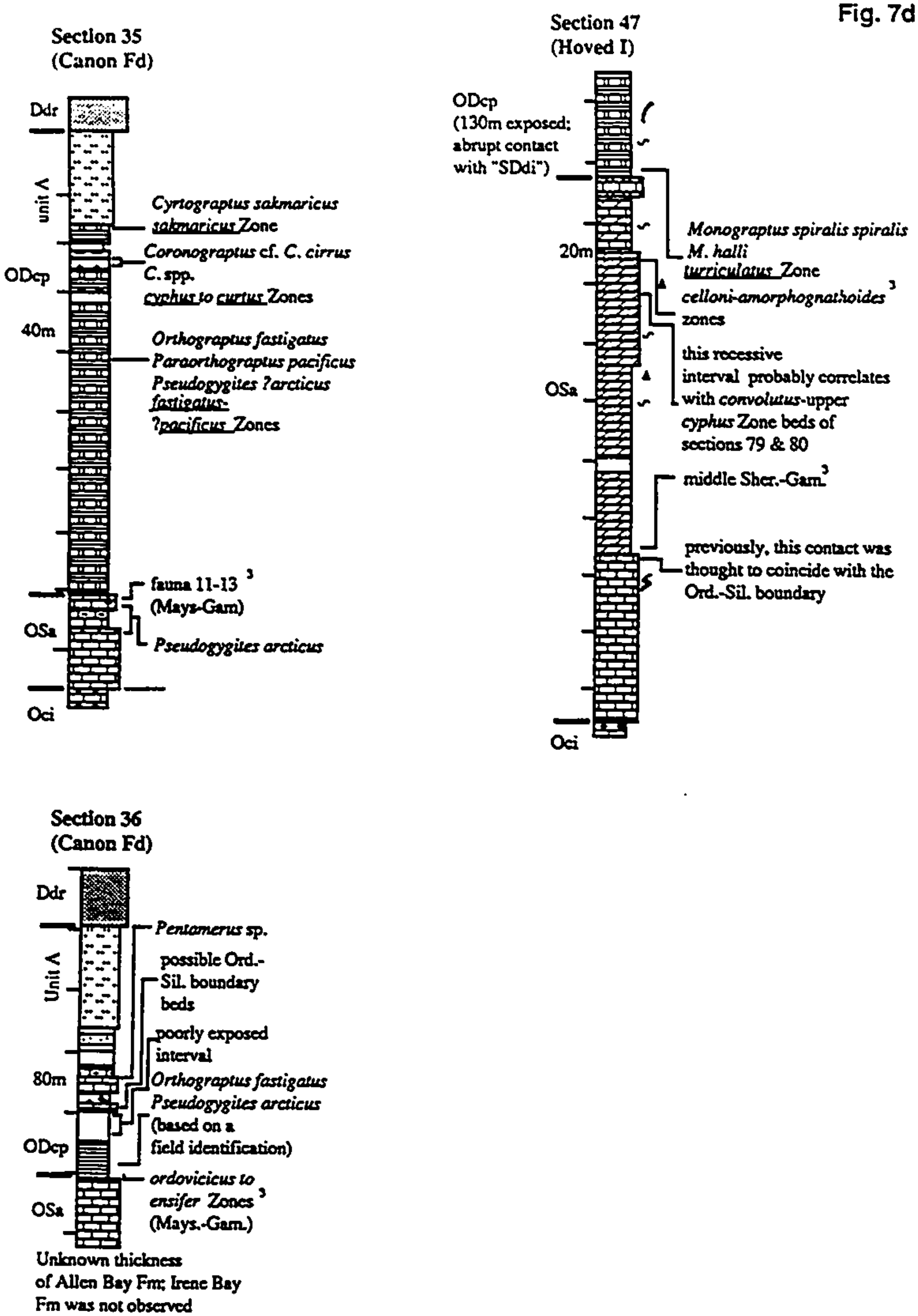

Section 35
(Canon Fd)
Ddr
unit A
Cyrtograptus sakmaricus sakmaricus Zone
Coronograptus cf. C. cirrus
C. spp.
cyphus to curtus Zones
ODcp
40m
Orthograptus fastigatus
Paraorthograptus pacificus
Pseudogygites ?arcticus
fastigatus-?pacificus Zones
fauna 11-13 3
(Mays-Gam)
OSa
Pseudogygites arcticus
Oci
Section 47
(Hoved I)
ODcp
(130m exposed;
abrupt contact
with "SDdi")
Monograptus spiralis spiralis
M. halli
20m
turriculatus Zone
celloni-amorphognathoides 3
zones
this recessive
interval probably correlates
OSa
with convolutus-upper
cyphus Zone beds of
sections 79 & 80
middle Sher.-Gam. 3
previously, this contact was
thought to coincide with the
Ord.-Sil. boundary
Oci
Section 36
(Canon Fd)
Ddr
Pentamerus sp.
Unit A
possible Ord.-Sil. boundary beds
poorly exposed interval
80m
Orthograptus fastigatus
Pseudogygites arcticus
(based on a field identification)
ODcp
ordovicicus to ensifer Zones 3
(Mays.-Gam.)
OSa
Unknown thickness
of Allen Bay Fm; Irene Bay
Fm was not observed

Fig. 7e

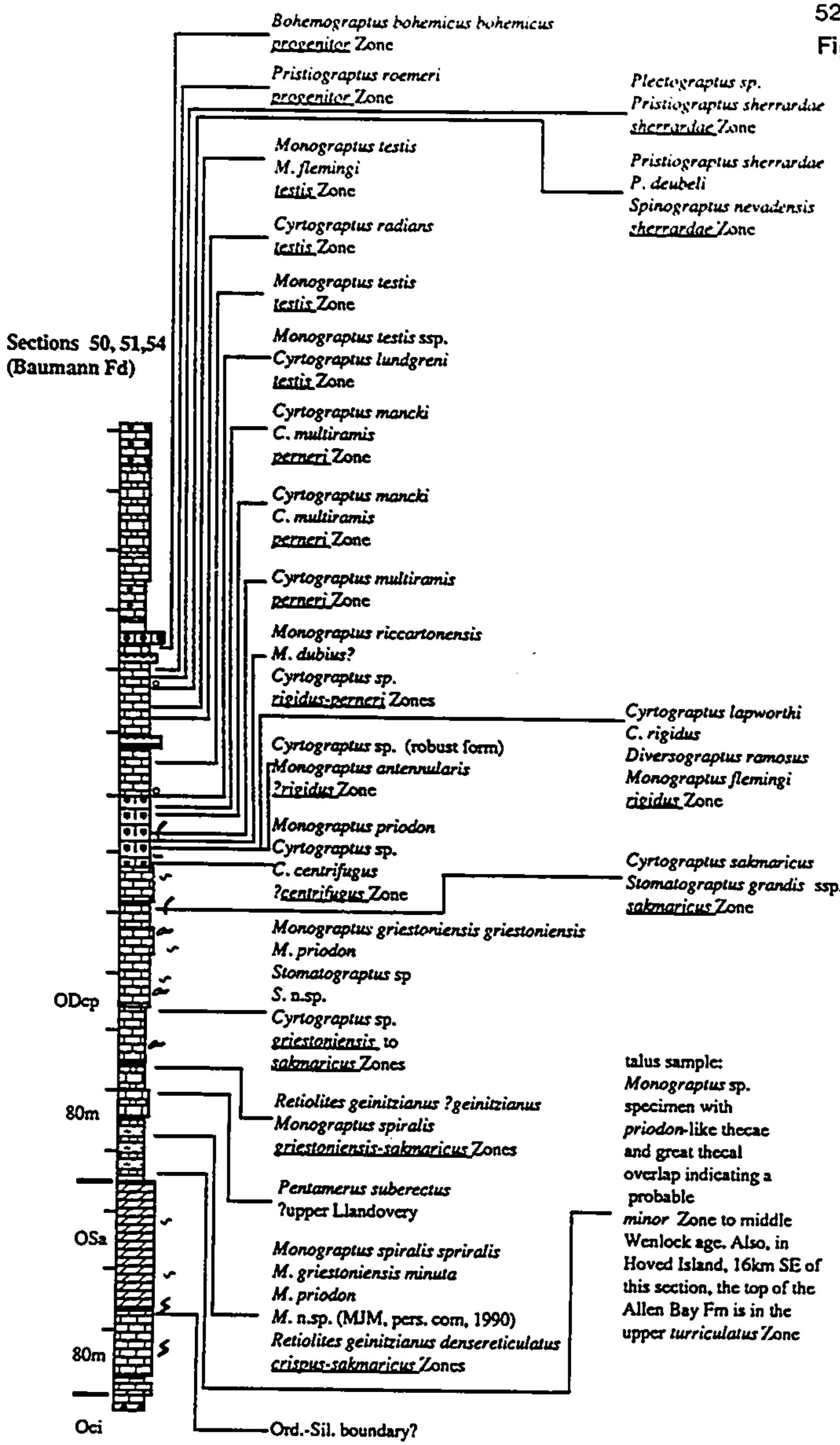
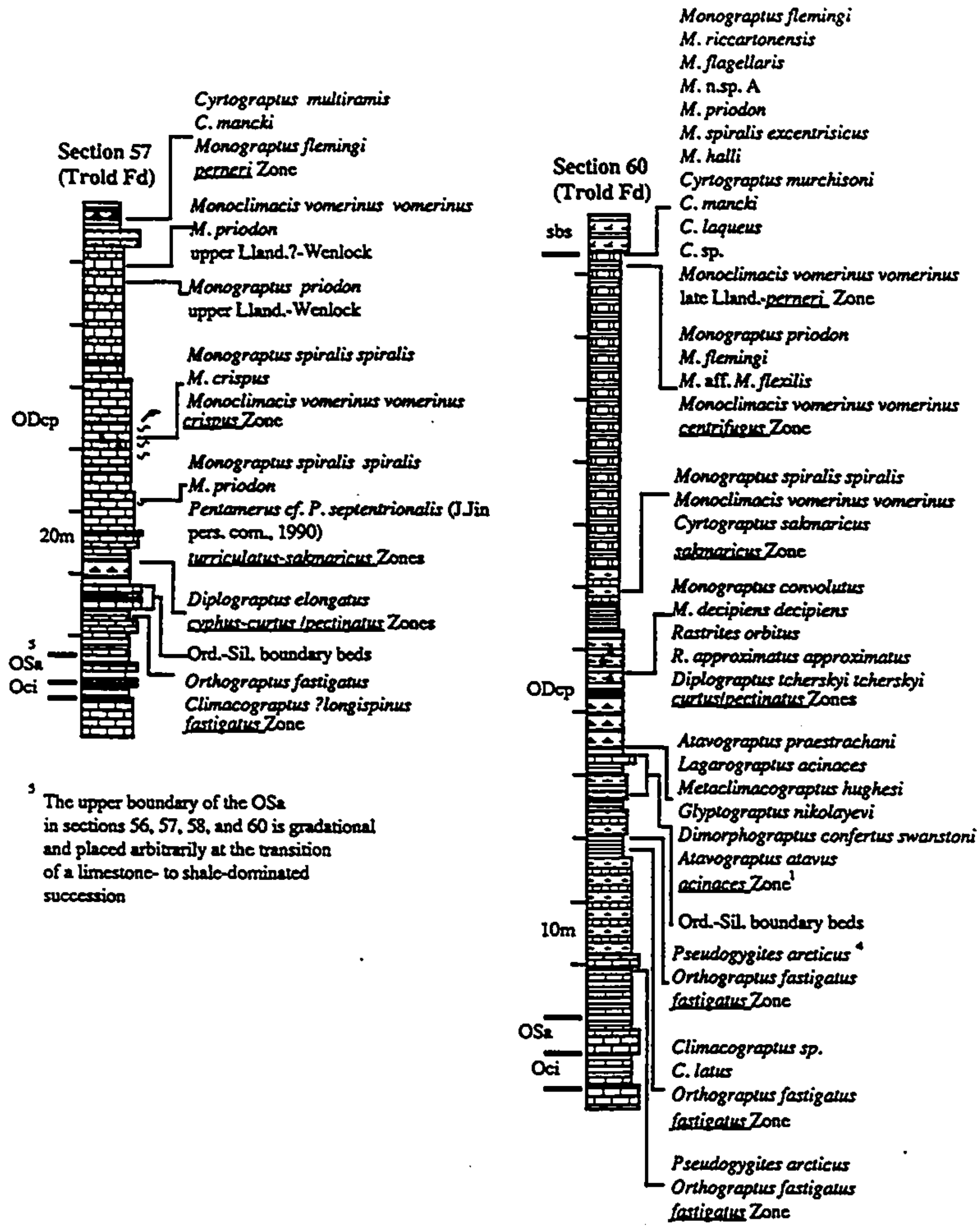
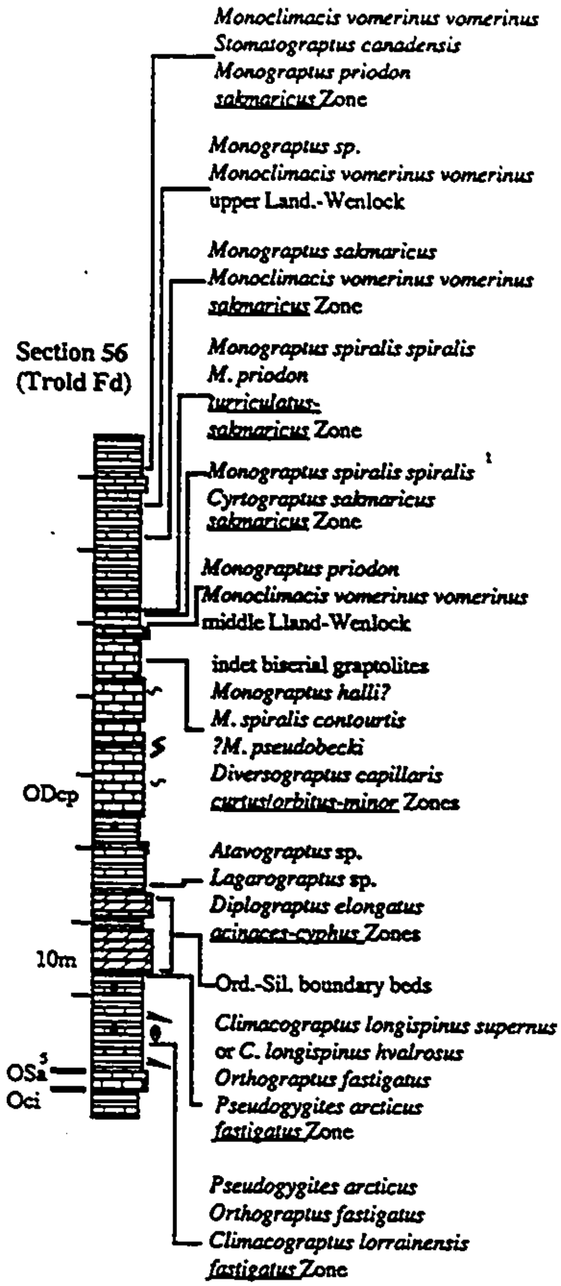

Bohemograptus bohemicus bohemicus
progenitor Zone
Pristiograptus roemeri
progenitor Zone
Plectograptus sp.
Pristiograptus sherrardae
sherrardae Zone
Pristiograptus sherrardae
P. deubeli
Spinograptus nevadensis
sherrardae Zone
Monograptus testis
M. flemingi
testis Zone
Cyrtograptus radians
testis Zone
Monograptus testis
testis Zone
Sections 50, 51, 54
(Baumann Fd)
Monograptus testis ssp.
Cyrtograptus lundgreni
testis Zone
Cyrtograptus mancki
C. multiramis
perneri Zone
Cyrtograptus mancki
C. multiramis
perneri Zone
Cyrtograptus multiramis
perneri Zone
Monograptus riccartonensis
M. dubius?
Cyrtograptus sp.
rigidus-perneri Zones
Cyrtograptus lapworthi
C. rigidus
Diversograptus ramosus
Monograptus flemingi
rigidus Zone
Cyrtograptus sp. (robust form)
Monograptus antennularis
?rigidus Zone
Monograptus priodon
Cyrtograptus sp.
C. centrifugus
?centrifugus Zone
Cyrtograptus sakmaricus
Stomatograptus grandis ssp.
sakmaricus Zone
Monograptus griestoniensis griestoniensis
M. priodon
Stomatograptus sp
S. n.sp.
Cyrtograptus sp.
griestoniensis to
sakmaricus Zones
ODcp
talus sample:
Monograptus sp.
specimen with
priodon-like thecae
and great thecal
overlap indicating a
probable
minor Zone to middle
Wenlock age. Also, in
Hoved Island, 16km SE of
this section, the top of the
Allen Bay Fm is in the
upper turriculatus Zone
Retiolites geinitzianus ?geinitzianus
Monograptus spiralis
griestoniensis-sakmaricus Zones
80m
Pentamerus suberectus
?upper Llandovery
OSa
Monograptus spiralis spiralis
M. griestoniensis minuta
M. priodon
M. n.sp. (MJM, pers. com, 1990)
Retiolites geinitzianus densereticulatus
crispus-sakmaricus Zones
80m
Oci
Ord.-Sil. boundary?

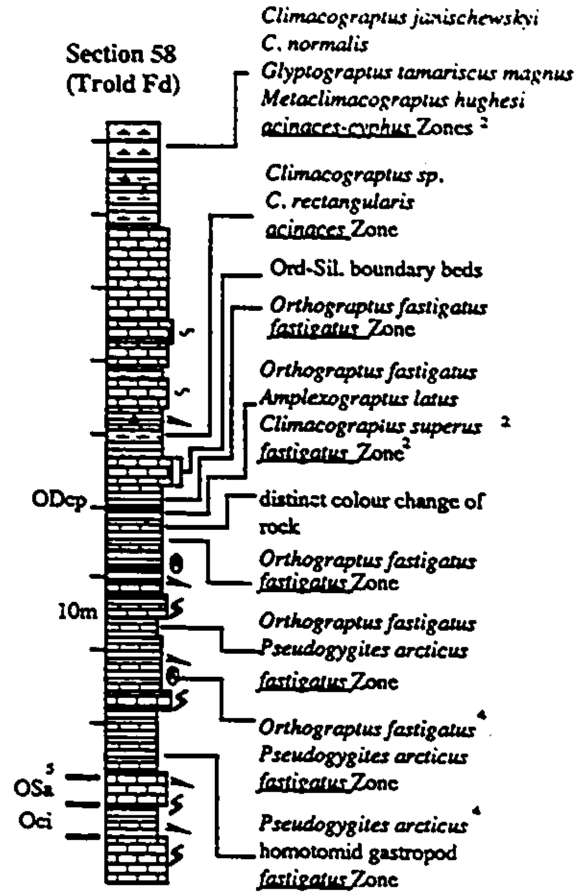
Fig. 7f



[5] The upper boundary of the OSa in sections 56, 57, 58, and 60 is gradational and placed arbitrarily at the transition of a limestone- to shale-dominated succession

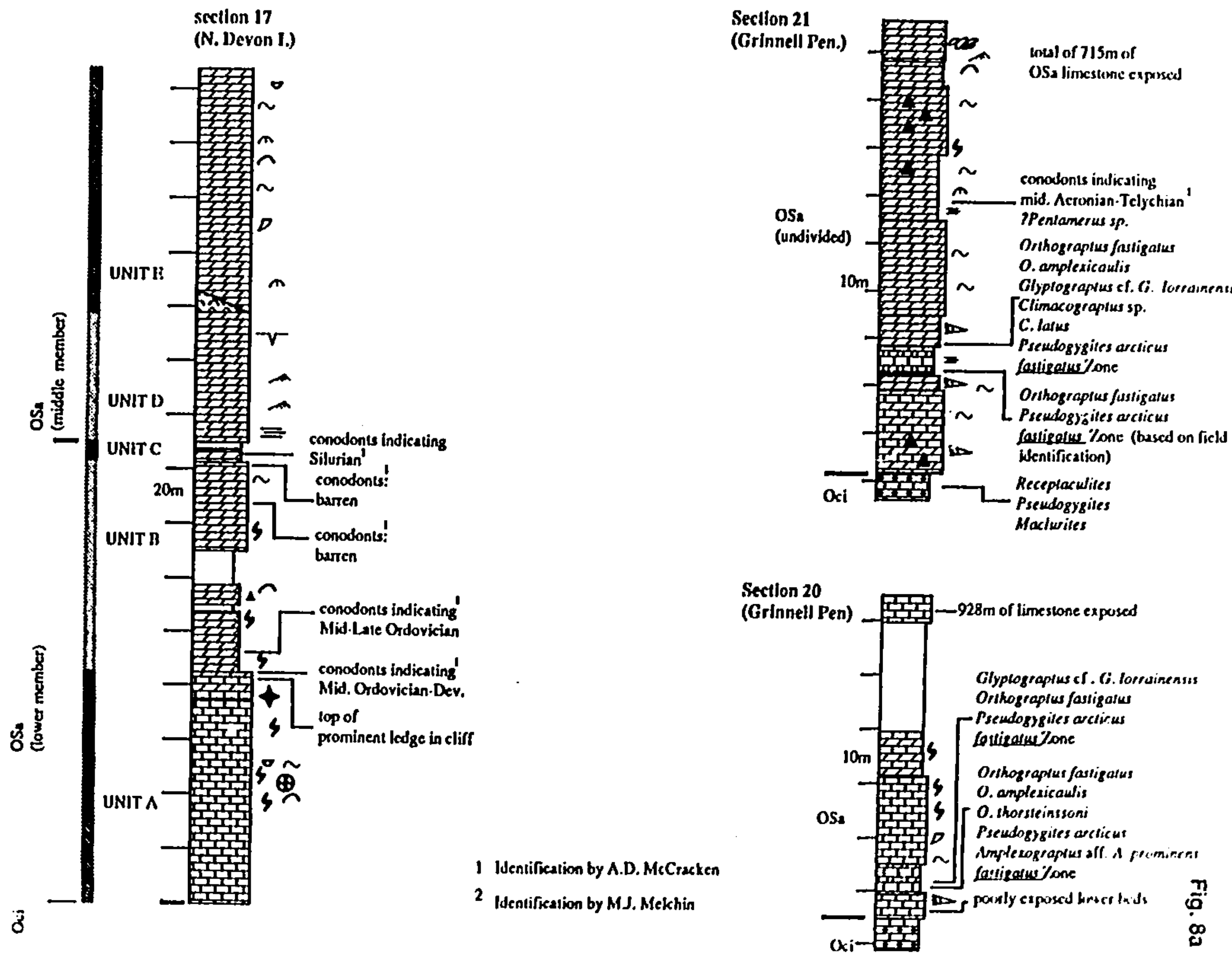
Fig 7g





[5] The upper boundary of the OSa in sections 56, 57, 58, and 60 is gradational and placed arbitrarily at the transition of a limestone- to shale-dominated succession

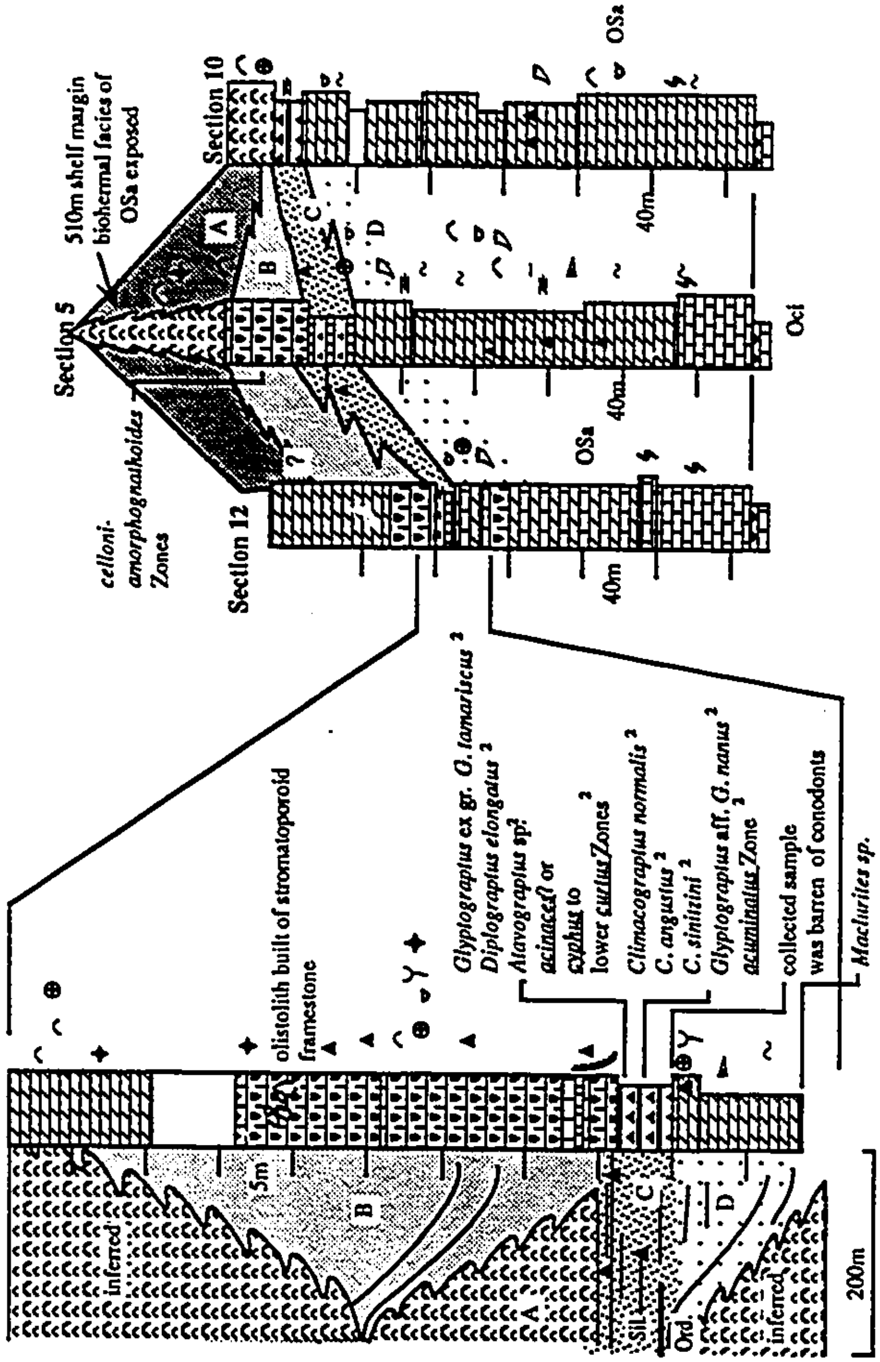
Figure 8: Intraplatform and platform-margin sections straddling the Ordovician-Silurian boundary in the vicinity of Grinnell Peninsula, northern Devon Island, and Ellesmere Island. Units illustrated in section 17 are discussed in the text. In Figure 8b, Ellesmere Island Unit A is biohermal facies, Unit B is allochthonous reef flank facies, Unit C is argillaceous limestone and chert-rich facies, and Unit D is reef flank and bioturbated dolostone facies. Legend as in Figure 6d.



1 Identification by A.D. McCracken
2 Identification by M.J. Melchin

Fig. 8a

Fig. 8b


Section 5
Section 10
Section 12
510m shelf margin biohermal facies of OSa exposed
celloni-amorphognathoides Zones
A
B
C
D
40m
OSa
Oci
olistolith built of stromatoporoid framestone
5m
inferred
Sil.
Ord.
200m
Glyptograptus ex gr. G. tamariscus
Diplograptus elongatus
Atavograptus sp.
acinaces or
cyphus to
lower crispus Zones
Climacograptus normalis
C. angustus
C. sinitzini
Glyptograptus aff. G. nanus
acuminatus Zone
collected sample
was barren of conodonts
Maclurites sp.

(i) The occurrence of Irene Bay-like lithology in the Thumb Mountain at shelf-margin localities (at Trold Fiord) indicates that these argillaceous units are deeper water facies equivalents of the clean, mottled carbonates predominant in the coeval platformward Thumb Mountain succession. These units, as exposed at Trold Fiord, are so much like the Irene Bay Formation that Kerr (1967) originally included them within that formation, but in this report, they are considered to belong to an upper member of the Thumb Mountain Formation, and as such, compare to the platform stratigraphy established on North Greenland.

(ii) The basinward thin, shaly, unfossiliferous characteristics and the platformward carbonate-rich, fossiliferous succession suggest a deep- to shallow-water facies relationship across the shelf.

(iii) The Irene Bay Formation is one of the most widespread and lithologically uniform units in the Canadian Arctic, and also widely similar biostratigraphically, an unlikely relationship if the formation was autogenically controlled. These stratigraphic characteristics of the Irene Bay Formation suggest that sea-level rise accompanied deposition.

Although solely climatic factors may be involved in phosphogenesis, the presence of phosphatic beds provides further evidence for sea-level rise. Phosphogenesis is generally correlated with coastal upwelling and high biological activity (Arthur and Jenkyns, 1981; Baker and Burnett, 1988). Upwelling, although its cause is complex, may be associated with deflection of oceanic currents over sea-floor topography (Riggs, 1984), or it may be associated with oceanic current divergence at continental margins (Sheldon, 1980; Jenkyns, 1976).

Phosphogenesis may also be related to catastrophic killing or to sudden terrigenous input of phosphate; however, there is no evidence for such causes in the Irene Bay Formation. High biological activity was possibly related to upwelling and a highstand which in turn fostered an abundant fauna; however, this event was associated with a bathymetrically depressed carbonate shelf and reduced carbonate deposition. Depths over the shelf were between 50 and 150m: the lower limit is based on the occurrence of calcareous algae while the upper limit is, in part, based on the minimum depth needed to maintain influx of nutrient-rich bottom-water across the shelf (Riggs, 1984) and, in part, related to storm wave base.

Subsequent to deposition of the Irene Bay Formation, basin-wide carbonate deposition was re-established. However, this shelf limestone, assignable to the basal tongue of the Allen Bay Formation, was clearly deposited in deeper water than that of the underlying Caradoc Thumb Mountain Formation. Another deepening event occurred subsequently in the *fastigatus* Zone which caused widespread drowning of the platform. The "Irene Bay event" perhaps marks the first widespread sea-level rise to interrupt an otherwise stable Middle and Upper Ordovician carbonate platform. The event could be considered an early stage of platform drowning, as discussed below.

In the Irene Bay Formation, there is an east-to-west decrease in limestone and fossil content. Thicknesses given in Table 1 show greater values for platform sections in a band that is parallel to the Ordovician shelf margin (compare sections B, 54, and E), but some of these thicknesses are uncertain, and several are based on a reconnaissance survey. It is possible that some of the underlying rubbly weathering upper member of the Thumb Mountain Formation was included within of these Irene Bay Formation thicknesses (possibly in sections A and D), but in other areas sections 12, 62, and 64), the upper and lower contacts of the Irene Bay Formation appear to be relatively well defined.

Furthermore, there is a north-to-south variation of facies along the Ordovician shelf margin. For example, beneath large Silurian mud buildups, the Irene Bay Formation is more or less a typical rubbly weathering fossiliferous limestone; whereas beneath drowned shelf shales and carbonates near the position of the former Ordovician shelf margin, the Irene Bay Formation is an atypically shaly limestone. This relationship suggests that an underlying tectonic control influenced not only the location of large mud buildups over the drowned Ordovician shelf margin, but also the nature of the underlying facies in the Irene Bay and Thumb Mountain formations.

ALLEN BAY FORMATION (AND EQUIVALENTS)

The Allen Bay Formation is a predominantly dolostone unit, originally named by Thorsteinsson and Fortier (1954), with its type section near Resolute Bay, Cornwallis Island. The formation is thickly bedded, locally biohermal, and extensively exposed in precipitous sea cliffs through much of the study area. It comprises three main informal members, that can be recognized on Ellesmere, Cornwallis, and Devon Islands, and all three have the mappable extent of formations. Five other areally restricted members are discussed in the following report.

The lower limestone member, as defined on purely lithological grounds, varies considerably in thickness, primarily due to its diachronous and gradational upper contact with the Cape Phillips Formation. On the platform, however, the contact generally is conformable and has been equated with the Ordovician-Silurian boundary (Thorsteinsson and Mayr, 1987; Mayr *et al.*, in prep.). The boundary between the two overlying units (the middle brown dolostone and upper peritidal dolostone members) is very gradational and arbitrarily established. The contact with the overlying Cape Storm Formation is considered to be disconformable, in agreement with recent work by Mayr and others (in prep.). A summary of the thicknesses of the members of the formation is given in Table 2.

Table 2: Thicknesses of the Allen Bay Formation in the study area.
Numbered sections as in Fig.1

A: upper peritidal member.(OSa-3)

Section	Thickness
15	~320m
16	380[5]
18	370[3]
6-7	140

B: upper limestone member (OSa-4) (southwest and northern Ellesmere Island)

Section	Thickness
61	0m
62	105
64	85
S. Ellesmere I.[3]	100

C: middle dolostone member. (OSa-2)

Section	Thickness
S. Ellesemere I.[3]	500m
62	440
64	485

D: lower limestone member (OSa-1),[4]

Section	Thickness
2	56m
3	100
5	218
10	231
12	161
13	71
17	170
22	7.5
26	46
29	155
30	96
31	120
32	130
34	190
35	65
43	38
57-60	~3
62	95
64	30
Griffith I.[1]	161
Devon I.[2]	93-144

[1] Mallamo, 1989
[2] Thorsteinsson and Mayr, 1987
[3] Mayr *et al.*, in prep.
[4] boundary between the lower and middle members of the Allen Bay Formation not necessarily corresponding to the Ordovician Silurian boundary
[5] incomplete section
[6] DNAG Chart (GSC, in prep.)
[7] Large mud buildup established over the drowned Ordovician shelf margin.

E: total thickness

Section	Thickness
5	728[5]
10	633[5]
12	665[5]
20	985[5]
21	715[5]
31	980
37	1335[7]
39	979[7]
41	1141[7]
42	1310[7]
S.Ellesmere I.[3]	745
S.Cornwallis I.[6]	1290

Lower limestone member(informal name)

Lithology and distribution

The lower member is a lithologically homogeneous unit (Table 2d) characterized by resistant, thickly bedded, mottled, sparsely fossiliferous, moderate brown dolomitic limestone. Fossils include, in decreasing abundance, crinoid debris, dasycladacean algae, cephalopods (mainly orthoconic types), solenoporid algae, solitary rugose corals, trilobites (complete and fragmentary, some identified as *Pseudogygites ? arcticus*), colonial corals (*Halysites*, *Favosites*, *Syringopora*), gastropods (*Maclurites*), stromatoporoids (typically with ragged coenostial margins), and lithistid sponges. Conspicuous tabular favositids occur concentrated along particular horizons at many locations, and chert is also common, particularly in Lower Silurian strata just west of the Silurian shelf margin (as delineated in Fig.1). At Dobbin Bay, an approximately 40m thick white chert-rich limestone unit (section 32, Fig.7c) occurs immediately above the Irene Bay Formation. Nodular, light, tan-brown chert is either concentrated along particular horizons or disseminated through a sparsely fossiliferous to unfossiliferous limestone. Chert is also present at several other localities: at Cañon Fiord, large, resistant, dark grey and brown chert nodules up to a metre in diameter form heads for glacially-shaped inverted flute-like structures. Most of the chert is generally randomly distributed but less commonly forms discontinuous nodular beds in the limestone. Also at Cañon Fiord (section 34: 58m thick, Fig.7c; section 35: 18m thick, Fig.7d) an aerially extensive and mappable, recessive, light brown, mottled dolostone unit occurs and contains abundant intact trilobites, including *Pseudogygites ? arcticus*. Conodonts in this unit belong to faunas of Maysvillian to Gamachian age (section 34, 35 Figs. 7c, d; identification by A.D. McCracken, written com., 1990).

There is a slight basinward decrease in fossil content and mottling, suggesting in a general way that there is a bathymetric control to mottling. In addition, mottling and fossil content show local variations. Tabular favositids, large orthoconic and breviconic cephalopods, skeletal wackestone and packstone, syringoporid corals, and stromatoporoids are locally common. Most of these fossils belong to the well-known Arctic Ordovician fauna (Thorsteinsson, 1958; Sweet and Miller, 1957). Conodonts from the top of the member are Richmondian to Gamachian in age. Because the top of the member is of Late Ordovician age, it in part coincides with the main phase of Hirnantian glaciation (Hambrey, 1985; Brenchley, 1988).

Of note in this member is a thin rubbly limestone unit usually about 15m in thickness and similar lithologically to the Irene Bay Formation. It has been identified in several areas inlcuding in exposures near Baumann Fiord (section 54 and in the Eids M-66 well) and northern Devon Island (section 17). The Irene Bay Formation thickens basinward in exposures about 40km west of section 54, likely at the expense of the limestone beds which separate the Irene Bay Formation and the lower rubbly Irene Bay-like limestone unit in the lower part of the Allen Bay Formation.

Ordovician-Silurian boundary: description of associated strata

The Ordc·᾽ician-Silurian boundary is of considerable interest internationally and still a controvcrsial matter, particularly from a biostratigraphic standpoint. Of note is the abundant and widespread evidence of Upper Ordovician to Middle Silurian glaciation (Hambrey, 1985; Brenchley, 1988), which probably climaxed during the upper Ashgill (Hirnantian). The effects of glacial eustasy on carbonate platforms is a popular topic for study, particularly from the standpoint of computer modelling of allocyclicity and the application of Milankovitch-band orbital perturbations to the observations of platform (and basinal) small-scale cycles (parasequences). The following observations from an extensive outcrop belt nether supports or refutes these ideas, although perhaps elucidating some problems of an approach which does not consider fully the intrinsic complexities of ancient climate modelling and carbonate depositional systems.

The Ordovician-Silurian boundary is recognized within the platform carbonates and deep-shelf shales, although the deep-shelf shales are generally better constrained biostratigraphically. Eight shale-rich sections (Sections 02, 03, 25, 26, 30, 35, 56, 57, 58, 60; Figs. 7a,b,d,f,g) and many of the platform carbonate sections straddle the boundary, but most of the latter sections can only discussed in conjunction with the relevant biostratigraphy.

In the vicinity of Trold Fiord, and stratigraphically above the Irene Bay

Formation, is a thin (less than 3m) carbonate unit. Its contact with the overlying Cape Phillips Formation is gradational and highly diachronous, younging cratonward (Fig. 7d,f; compare the Allen Bay-Cape Phillips formations contact in sections 47 and 60). Beds above the contact contain abundant graptolites (predominantly *Orthograptus fastigatus*), orthoconic and breviconic cephalopods, and trilobites (*Pseudogygites ?arcticus*). These beds consist of rhythmically interbedded black clayshales and laminated and/or bioturbated dark, petroliferous, unfossiliferous to sparsely fossiliferous, finely crystalline limestone. Above this succession, which is up to 20m thick, one or two resistant laminated or bioturbated limestone beds occur (sections 56, 58, 59 and 60 of Figs. 7f,g). Biostratigraphic evidence indicates that these beds, about 10m in thickness, straddle the Ordovician-Silurian boundary, and are overlain by black bedded or nodular chert, and interbedded clayshale, marlstone, and limestone. Abundant samples were collected for conodonts and graptolites, but the conodont samples have not yet been processed. Although fossils representing the *fastigatus* Zone are well represented throughout the study area, those of the *pacificus*, *persculptus*, and *acuminatus* zones are exceedingly rare, absent, or eroded. Similar findings are given by Melchin (1989, 1987a,b) in a detailed study of graptolites from the boundary sequence. He ascribed this absence to erosion, and/or non-deposition, or to deposition in shallow, relatively "restricted conditions" (Melchin, 1987b, p.195). Missing zones, condensed sequences, and a low diversity assemblage characterize the *acuminatus* and *atavus* Zones, but diverse and abundant graptolites characterize the *acinaces* and younger graptolite Zones. However, recently, based on more extensive collection of the sections straddling the boundary and a re-examination of collected samples, these zones were recognized in several localities, and, in addition, the *extraodinarius* Zone was recognized in a bore hole on Truro Island (Melchin, pers. com., 1990). In the Dob's Linn Ordovician-Silurian boundary stratotype, this biozone occurs between the *pacificus* and *persculptus* Zones (Williams, 1988), but other than the rare occurrence recorded by Melchin (pers.com., 1990), this zone is not recognized in the current investigation. Many Llandovery graptolite zones are absent from section 2 (Fig. 7a), and the first zone encountered is the *minor* Zone. Similarly, the lower Llandovery succession is condensed or missing from sections 26, 29, 30, and 36 (Fig. 7a,b,d).

On the platform, the Ordovician-Silurian boundary is considered concurrent with the boundary of the lower and middle members of the Allen Bay Formation. There is generally a sharp lithological break near, or at the system boundary, but west of the Silurian shelf margin, at the transition of platform carbonates to deep-shelf shales, this boundary has not been precisely located on lithological or biostratigraphic grounds. Strata straddling the boundary are considerably impoverished in diagnostic conodonts; Thorsteinsson and Mayr (1987) obtained no diagnostic conodont elements from 20kg of carbonate of probable Lower Silurian age collected near the boundary. Lower Silurian platform carbonates in northeastern Australia are similarly impoverished (Beirschtradt, pers. com., 1990), and a *standard* sample of 20kg is necessary to obtain any elements. In this report, the boundary cannot be precisely located in several sections, and is placed for convenience at a lithological break which is at least broadly faunally bracketed (Fig. 8a,b, sections 5, 17, 20, 21).

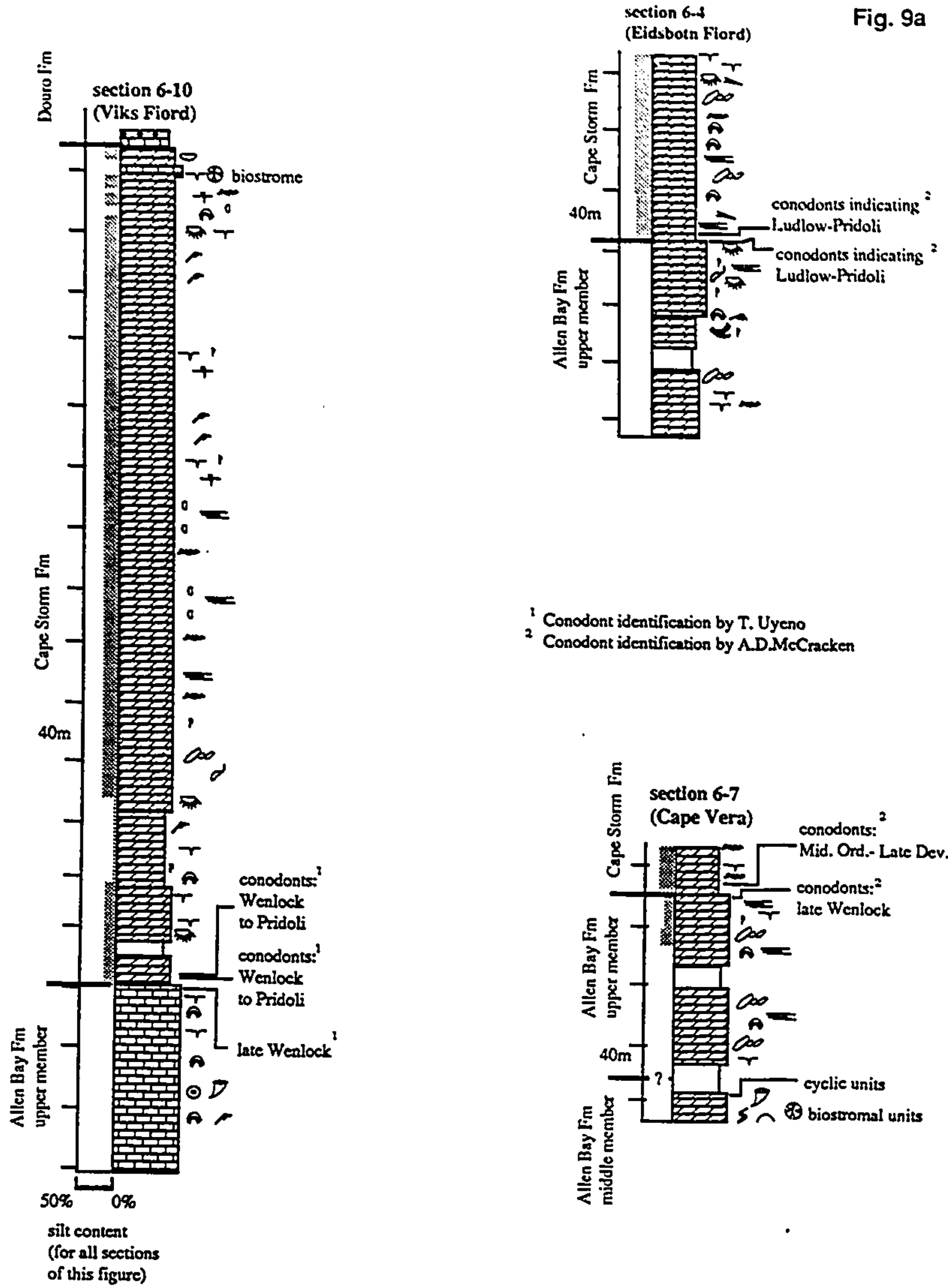
On Grinnell Peninsula (sections 20, 21; Fig. 8a) and near Irene Bay (sections 2 and 13; Fig. 7a,b; Plate 1b), the systems boundary is poorly constrained biostratigraphically; nevertheless, near the systems boundary, graptolitic and cherty shales onlap platform carbonates (Plate 1b), or subtidal, sparsely fossiliferous carbonate is generally sharply overlain by fossiliferous limestone or by strata indicating increased current activity. At Bay Fiord (sections 5, 10, 12; Fig. 8b; Plate 1b), lowest Silurian strata represent shale onlap and subsequent biohermal development and platform slope progradation. Lowest Silurian platform progradation and shallowing are marked by a gradual upward increase in size of olistoliths. These strata are overlain by massive stromatoporoid framestone and bindstone together with dolomitized encrinite grainstone. Abundant silicified macrofossils associated with the graptolitic chert and shale appear to be allochthonous, and the penultimate olistostrome beds contain up to 1m diameter olistoliths built of stromatoporoid framestone. Dolomitization of these younger and overlying biohermal strata has been intense, effectively obscuring much primary texture. In section 12 (Fig. 8b), carbonate beds underlying Silurian graptolitic shales are impoverished in conodonts, although an Ordovician gastropod (*Maclurites*) was recovered 10m below the base of the shaly interval. The Ordovician-Silurian boundary occurs within that 10m interval, perhaps at the base of the shaly unit. The occurrence of the *acuminatus* Zone in the base of the onlapping shaly sequence suggests lowest Silurian platform deepening. The Ordovician carbonates show marked variations in thickness in stratigraphic sections only 1 to 3km apart, and contain reef-derived conglomerates, suggesting that there was perhaps some uppermost Ordovician reef growth, although none of these structures has been observed in outcrop.

On Grinnell Peninsula, a similar relationship has been established (Fig. 8a; sections 20, 21). However, the shaly unit (between 5 and 7m in thickness) lacks chert beds, and is Upper Ordovician in age (*fastigatus* Zone). It overlies a very thin, sparsely siliceous carbonate unit, but basal Silurian units have not been clearly defined on lithostratigraphic or biostratigraphic grounds. The carbonate strata overlying the Ordovician shale generally represent gradual upward shallowing, such that in the middle Llandovery (below *staurognathoides-celloni* Zone), the strata are represented by laminated and mudcracked dolostone, deposited probably in a restricted environment (83m above base, section 21). Above this succession, carbonate was deposited in deepening platform water, and was then succeeded by the deposition of lime muds, microbial lime muds, and ooids (discussed below). This boundary succession is comparable to that recorded in section 17 (Fig. 8a).

On Cornwallis Island, shale onlapping is recorded in platform carbonates at three separate localities and is dated as occurring in the *fastigatus*, *acinaces-curtus*, *minor-turriculatus*, and *griestoniensis* zones (Melchin, 1989; Mallamo, 1989). Strata straddling the boundary were also examined on Hoved Island (Fig. 7d, section 47), and a surprising stratigraphic relationship was established: the strata immediately overlying the contact between the limestone and dolostone contain Ordovician age conodonts. Previously, the Ordovician-Silurian boundary was regarded as generally coincident with the dolostone-limestone contact (Thorsteinsson and Mayr, 1987; Mayr *et al.*, in prep.; Poey, 1988; Plate 4). In section 47, basal dolostone beds are sparsely fossiliferous and contain delicate fasciculate corals, including tabulates and rugosans. Above, onlapping shale at the base of the *turriculatus* Zone marks the beginning of long-ranging shale deposition and the base of the Cape Phillips Formation. At Baumann Fiord (section 54, Fig. 7e) and Makinson Inlet (section 62, Fig. 9c), a similar, sharp lithological break is observed between the lower and middle members of the Allen Bay Formation: fossiliferous wackestone and packstone with complete planispiral gastropods and abundant crinoidal material are abruptly overlain by unfossiliferous dolostone. The contact between these two lithologies is irregular, with up to 10cm of apparent erosional relief; however, conodont samples collected across this probable disconformity have not yet been processed. This lithological break is clearly evident on air photos and in sea cliffs, seen as the pronounced change in weathering colour, from light brown dolostone above to medium brown limestone below.

Figure 9a-d: Stratigraphic sections of platform carbonates from various localities. These are correlated diagrammatically in the regional cross sections in Figs. 18, 22, 25, 26. Conodont identifications are listed in Appendix three. Legend as in Fig. 6d.

Fig. 9a



[1] Conodont identification by T. Uyeno
[2] Conodont identification by A.D.McCracken

Fig. 9b

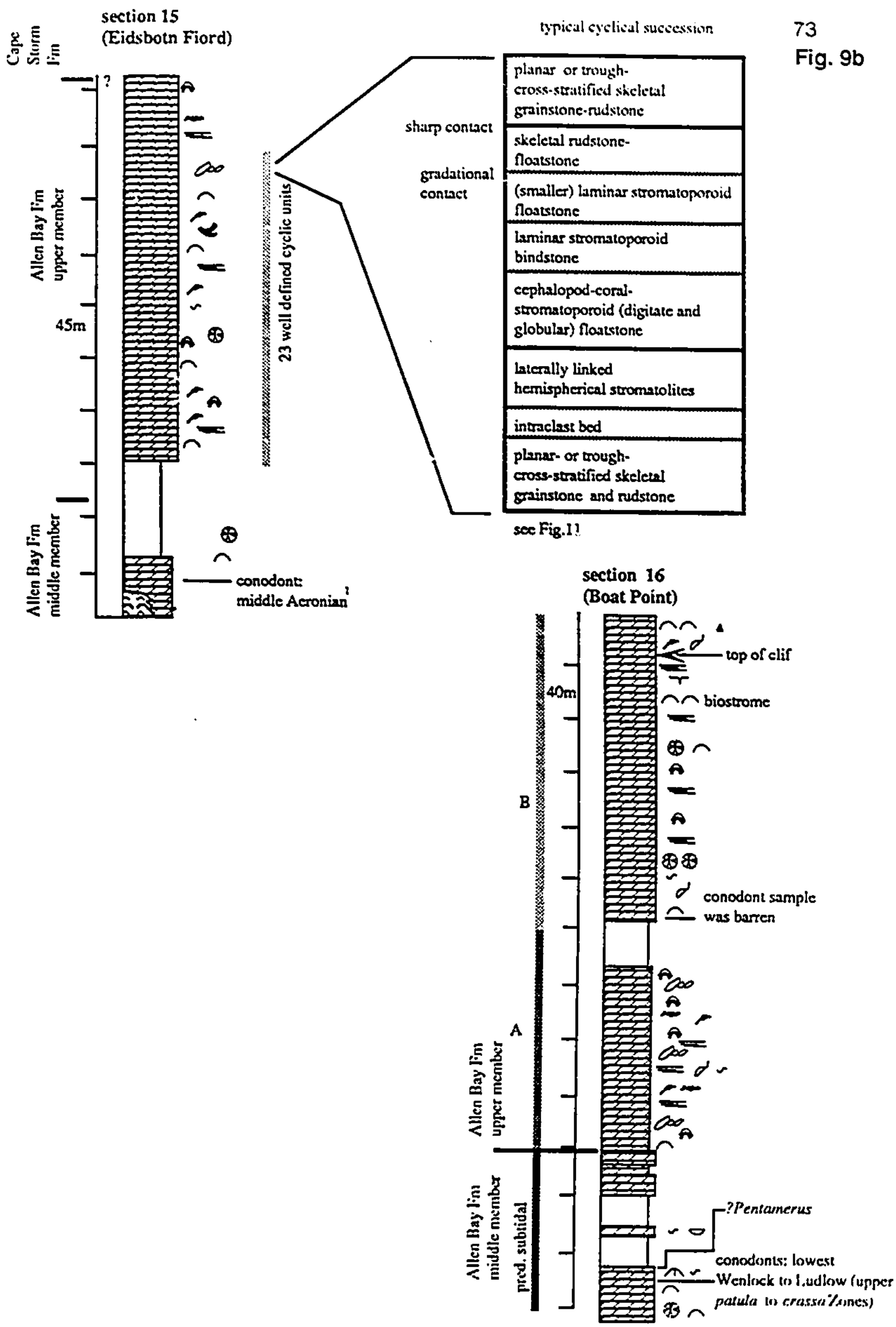
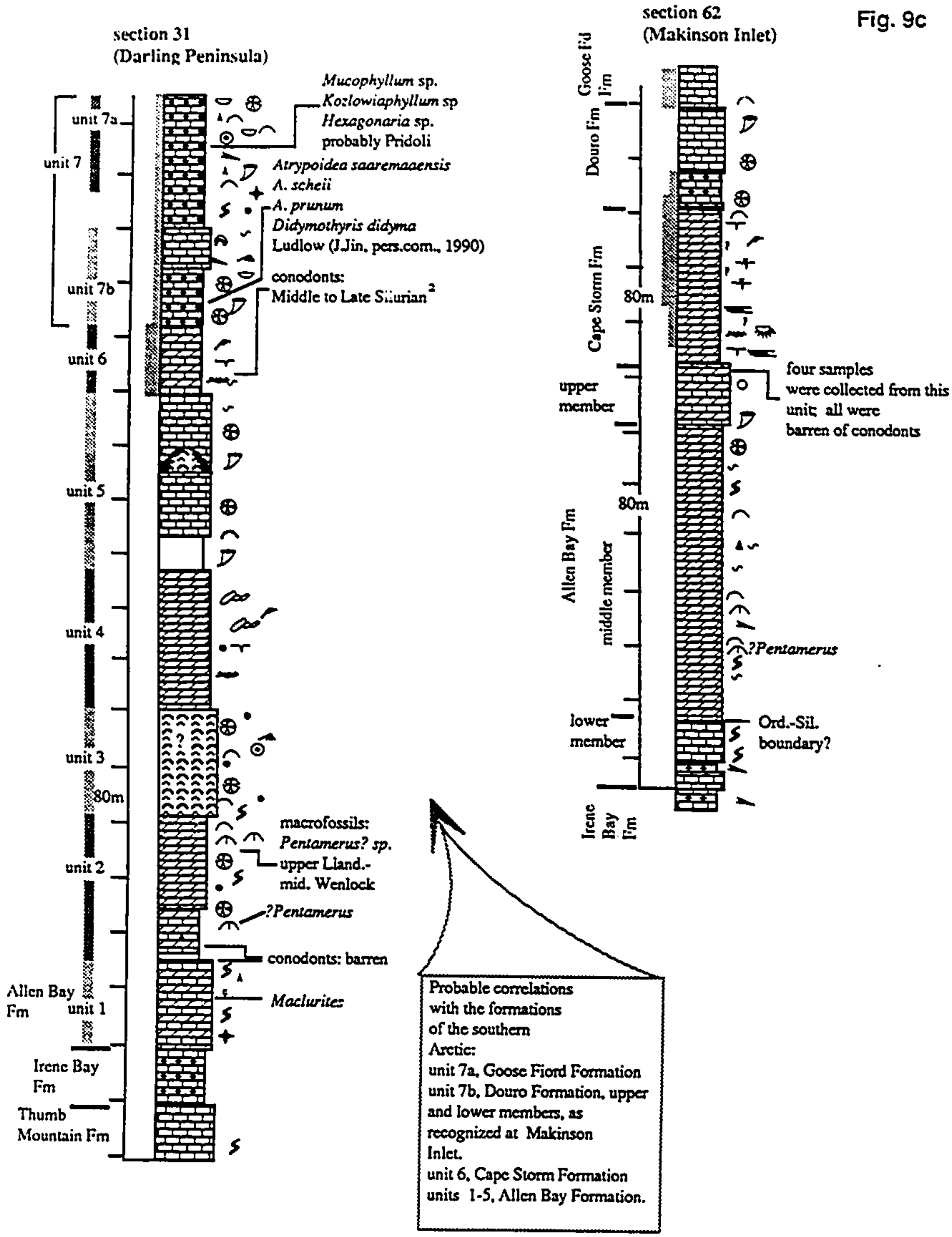
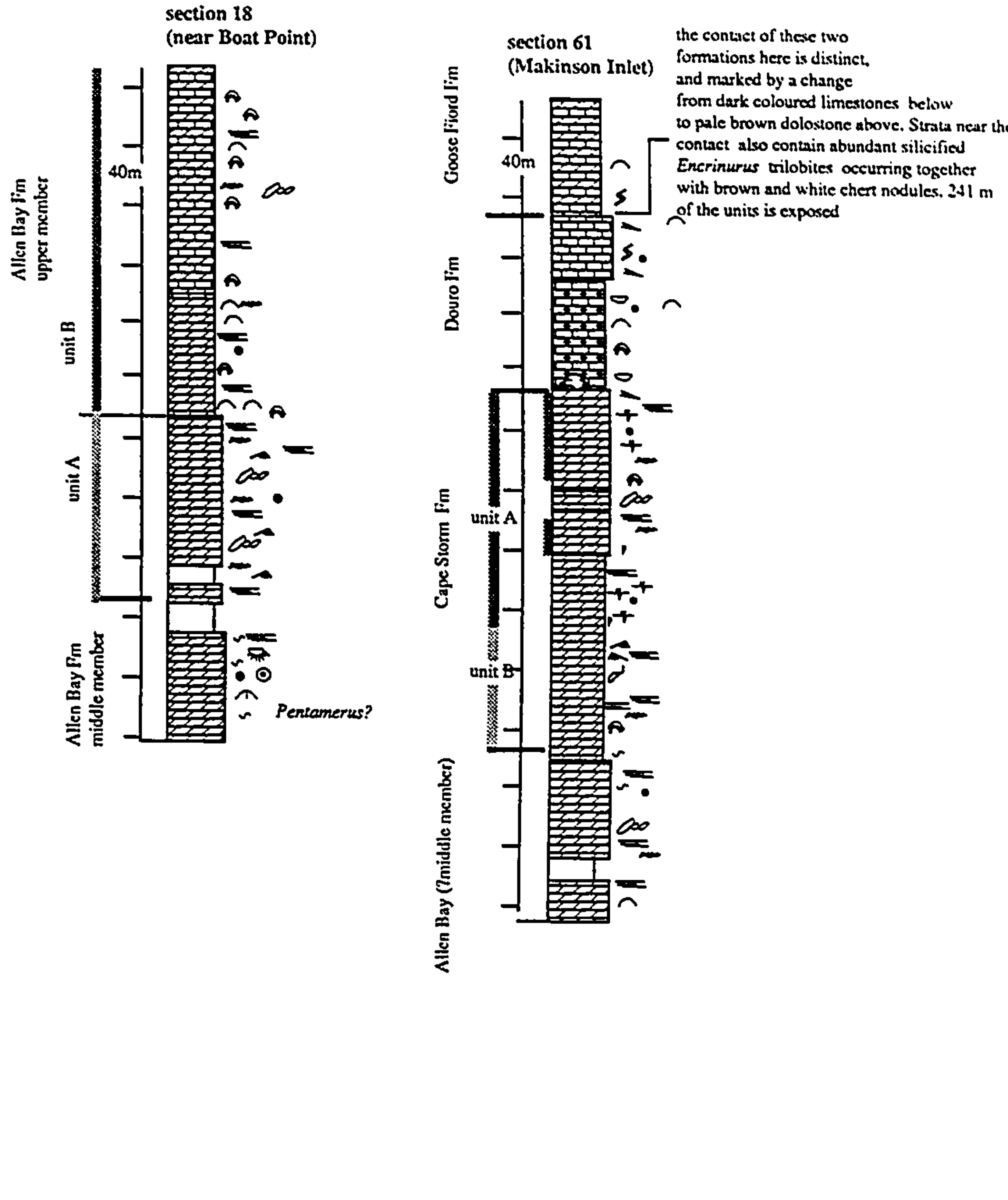

section 15
(Eidsbotn Fiord)
Cape Storm Fm
Allen Bay Fm upper member
45m
Allen Bay Fm middle member
conodont: middle Aeronian
23 well defined cyclic units
typical cyclical succession
sharp contact
gradational contact
planar or trough-cross-stratified skeletal grainstone-rudstone
skeletal rudstone-floatstone
(smaller) laminar stromatoporoid floatstone
laminar stromatoporoid bindstone
cephalopod-coral-stromatoporoid (digitate and globular) floatstone
laterally linked hemispherical stromatolites
intraclast bed
planar- or trough-cross-stratified skeletal grainstone and rudstone
see Fig.11
section 16
(Boat Point)
top of clif
biostrome
40m
B
A
conodont sample was barren
Allen Bay Fm upper member
Allen Bay Fm middle member
pred. subtidal
?Pentamerus
conodonts: lowest Wenlock to Ludlow (upper patula to crassa Zones)

Fig. 9c



section 31
(Darling Peninsula)
unit 7a
unit 7
unit 7b
unit 6
unit 5
unit 4
unit 3
80m
unit 2
Allen Bay Fm
unit 1
Irene Bay Fm
Thumb Mountain Fm
Mucophyllum sp.
Kozlowiaphyllum sp
Hexagonaria sp.
probably Pridoli
Atrypoidea saaremaaensis
A. scheii
A. prunum
Didymothyris didyma
Ludlow (J.Jin, pers.com., 1990)
conodonts:
Middle to Late Silurian[2]
macrofossils:
Pentamerus? sp.
upper Lland.-
mid. Wenlock
?Pentamerus
conodonts: barren
Maclurites
section 62
(Makinson Inlet)
Goose Fd Fm
Douro Fm
Cape Storm Fm
80m
upper member
80m
Allen Bay Fm
middle member
lower member
Irene Bay Fm
four samples
were collected from this
unit; all were
barren of conodonts
?Pentamerus
Ord.-Sil.
boundary?
Probable correlations
with the formations
of the southern
Arctic:
unit 7a, Goose Fiord Formation
unit 7b, Douro Formation, upper
and lower members, as
recognized at Makinson
Inlet.
unit 6, Cape Storm Formation
units 1-5, Allen Bay Formation.

Fig. 9d


section 18
(near Boat Point)
40m
Allen Bay Fm
upper member
unit B
unit A
Allen Bay Fm
middle member
Pentamerus?
section 61
(Makinson Inlet)
40m
Goose Fiord Fm
Douro Fm
Cape Storm Fm
unit A
unit B
Allen Bay (?middle member)
the contact of these two
formations here is distinct,
and marked by a change
from dark coloured limestones below
to pale brown dolostone above. Strata near the
contact also contain abundant silicified
Encrinurus trilobites occurring together
with brown and white chert nodules. 241 m
of the units is exposed

At Sandhook Bay (section 17; Fig. 8a), a contrasting boundary succession is present. At a prominent break in slope (top of unit A, Fig.8a), the contact between the lower and middle members of the Allen Bay Formation is marked by a change in weathering and lithological characteristics, from thickly bedded, mottled, sparsely fossiliferous dolomitic limestone and calcareous dolostone below to more thinly bedded, fossiliferous dolostone above. Unit B, however, is predominantly sparsely fossiliferous bioturbated dolomitized mudstone, wackestone, and packstone. An unfossiliferous, nodular, white chert-rich interval occurs midway in the unit, but is poorly exposed at this section. Conodonts collected from the base of unit B yielded fauna of Ordovician aspect (Fig. 8a); collections above these were barren.

At the top of unit B (Fig. 8a, section 17), a very sharp lithological break is followed by a thin rubbly weathering, dark brown, petroliferous dolostone bed (unit C), yielding conodonts of general Silurian aspect. A 10cm silty dolostone bed at the top of unit C is overlain by a thick unit (D) of even, parallel laminated dolostone that grades upward into a mudcracked and stromatolitic dolostone (unit E). Unit E is abruptly overlain by thickly bedded, biohermal, petroliferous, brown fossiliferous dolostone, containing abundant ?*Pentamerus* floatstone and rudstone and a mudstone-rich stromatoporoid bioherm (13.5m thick). Noteworthy in this section is the evidence of general upward shallowing then upward deepening, apparently entirely of lower Llandovery (pre-C_1) age.

Ordovician-Silurian boundary: discussion and interpretation

It is well established that a major glacial episode occurred in the latest Ordovician. An apparently synchronous fall of sea level in the Hirnantian was followed by transgression which was completed by the end of the Hirnantian (Brenchley, 1988). General environmental stability followed in the Llandovery, and glaciogenic sediments of Wenlock age represent minor, probably isolated areas of alpine glaciation (Hambrey, 1985).

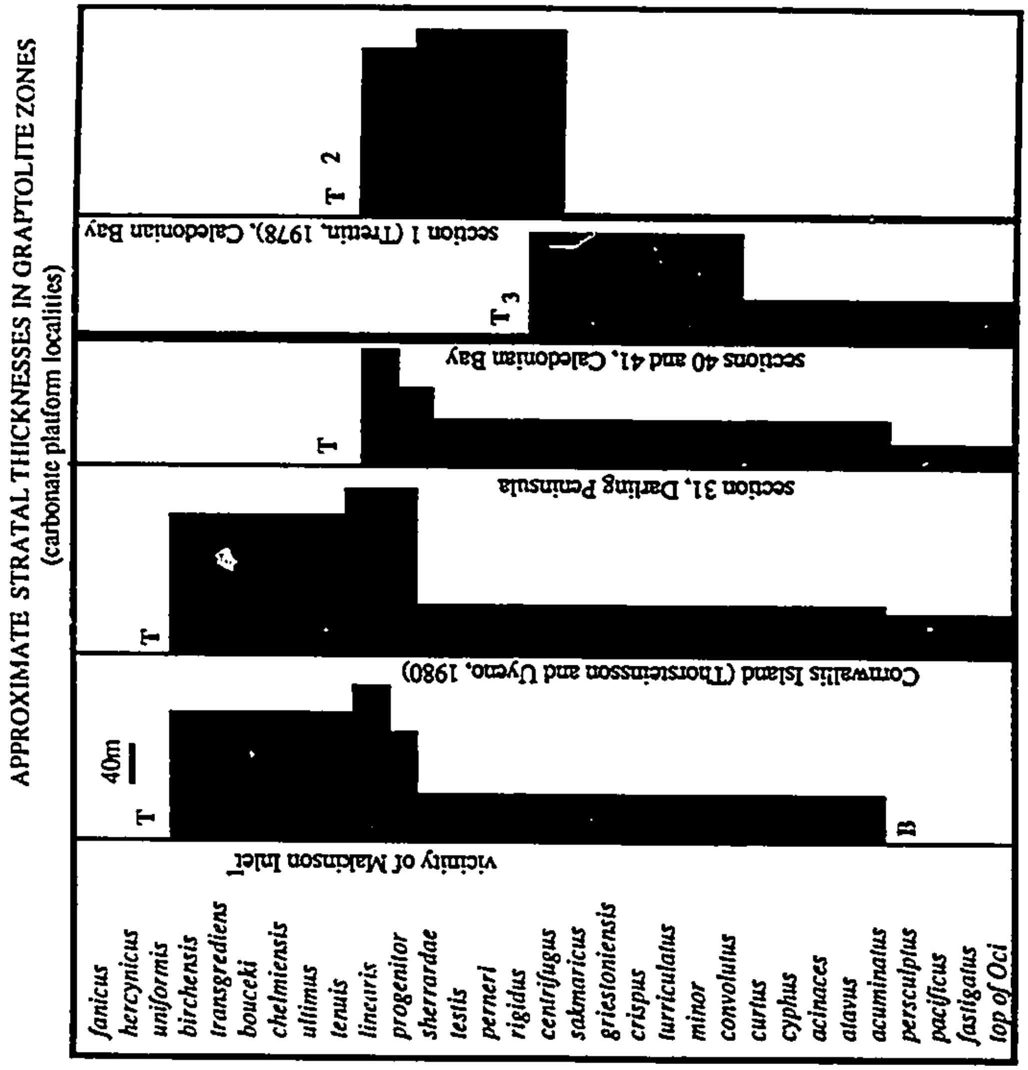
Several sea-level fluctuations are recorded in the platform facies. The first and perhaps most important regionally was the probable rise during deposition of the Irene Bay Formation. Beds that are lithologically similar to the Irene Bay Formation occur in the underlying Thumb Mountain Formation, but only in a north-to-south band at, and parallel to, the Ordovician shelf margin (as delineated in Fig. 1). Distinct rhythms can be observed in these beds, but the cause of this cyclicity, whether allogenic or autogenic, is unclear. The rubbly weathering limestone of the upper member of the Thumb Mountain Formation is considered to be a slightly deeper water facies of the "clean" mottled limestone platform succession. The Irene Bay Formation thus likely represents the maximum development of the deep-water, rubbly weathering limestone facies, hence reflecting platform submergence. Also, based on the stratigraphic relationships described for the area of Trold Fiord, the shale-rich limestone succession below the basal tongue of the Allen Bay Formation is a valid correlative of the Irene Bay Formation, based on art. 22(d) of the North American Stratigraphic Code (1985).

If the proposed bathymetric change explains the observed facies change in the Irene Bay Formation, the change should be reflected by fossil content in, and adjacent to, the Irene Bay Formation. However, dasycladaceans and receptaculatids occur within, above, and below the formation in platform localities, for example on Devon Island and on southwest Ellesmere Island. Although the relationship of depth to the distribution calcareous algae is not clear (Fagerstrom, 1988), the distribution suggests that bathymetry was not greatly altered during deposition of the Irene Bay Formation. Alternatively, climate change, for example flooding of the platform by cool marine waters, may have depressed carbonate production, caused phosphogenesis, while allowing sufficient illumination for benthic calcareous algae. Neither explanation can be substantiated, and the formation cannot be correlated outside the basin with certainty.

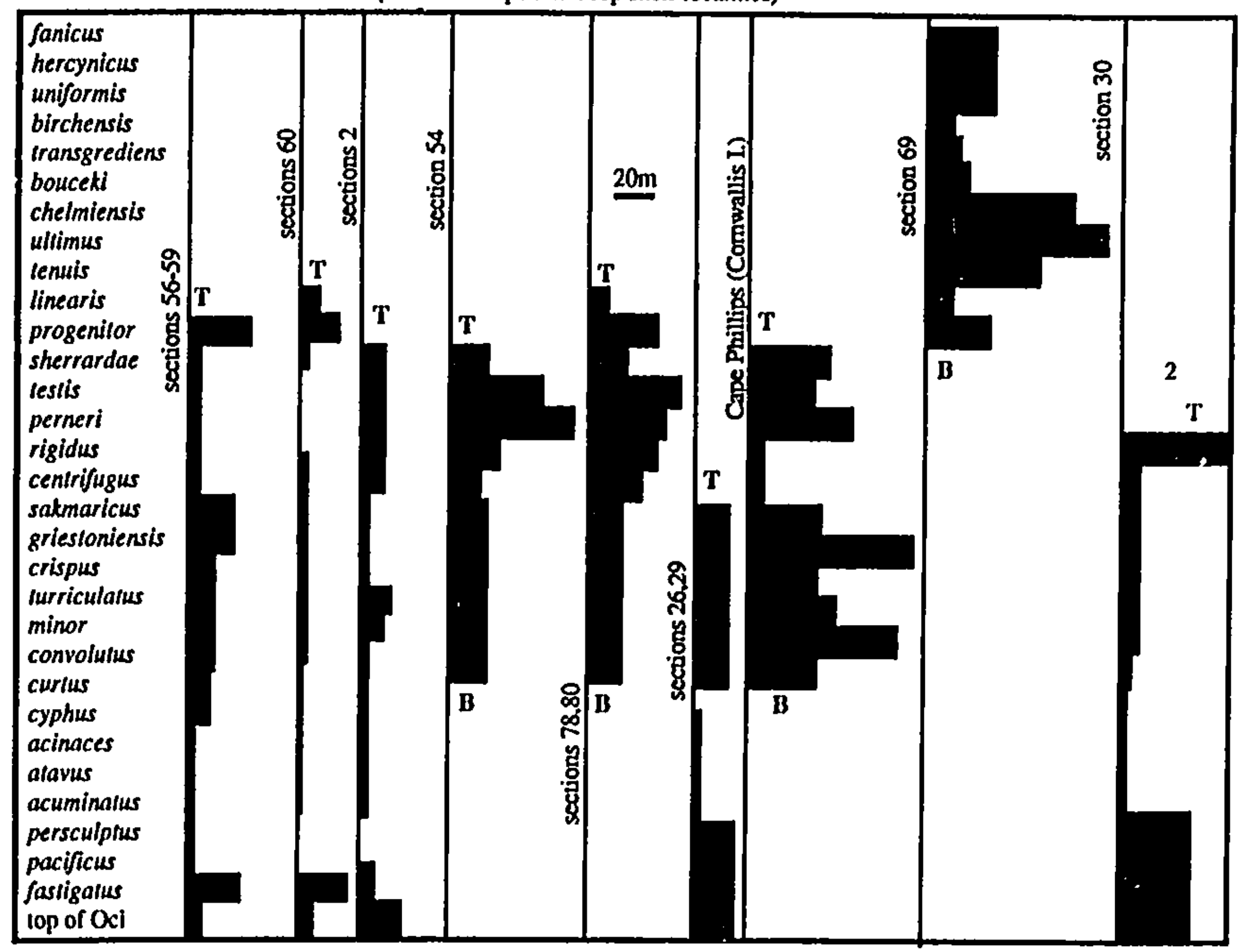
It is also of interest to compare the thicknesses of platform and deep shelf rocks for the 24 recognized Silurian graptolite biozones of the Arctic. This chart (Fig. 10) is to be used with caution, however, as the time connotations of biozones, hence rates of deposition are dependent on several factors, such as missing biozones, the correlation of graptolite and conodont zonal schemes, biozone thickness (or variations in the biozone time-span), compaction, and hiati, particularly in keep-up

Figure 10a-b: Diagrammatic representation of stratal thicknesses for biozones in various localities in the study area. Time connotations are presently subjective, and estimates for the length of time represented by the Silurian vary from 27 to 40Ma (Holland, 1989). Many of the currently recognized graptolite and conodont biozones are globally correlative and are considered to be relatively time-consistent internally (Kleffner, 1990; Koren, 1987). The thicknesses indicated can be converted into rates of sedimentation, although some carbonate platform localities are poorly constrained biostratigraphically. According to Kleffner (1989), the Llandovery and Pridoli were of approximately equal duration, collectively accounting for about 2/3 of the Silurian; they represent 26 to 17Ma, according to different estimates for the length of the Silurian.

Fig. 10b


APPROXIMATE STRATAL THICKNESSES IN GRAPTOLITE ZONES
(carbonate platform localities)
fanicus
hercynicus
uniformis
birchensis
transgrediens
bouceki
chelmiensis
ultimus
tenuis
linearis
progenitor
sherrardae
testis
perneri
rigidus
centrifugus
sakmaricus
griestoniensis
crispus
turriculatus
minor
convolutus
curtus
cyphus
acinaces
atavus
acuminatus
persculptus
pacificus
fastigatus
top of Oci
vicinity of Makinson Inlet
T
40m
B
Cornwallis Island (Thorsteinsson and Uyeno, 1980)
T
section 31, Darling Peninsula
T
sections 40 and 41, Caledonian Bay
T 3
section 1 (Trettin, 1978), Caledonian Bay
T 2

APPROXIMATE STRATAL THICKNESSES IN GRAPTOLITE ZONES
(Platform slope and deep shelf localities)



A disconformity below the Cape Storm Formation may account for significant absence of strata from some areas (Cornwallis Island and southern Ellesmere I.

T Top of section
B Base of section

[1] Biostratigraphy of the platform carbonates is poor, and these thicknesses are averaged.
[2] Flysch sedimentation
[3] Large mud buildups established over the drowned Ordovician shelf margin

carbonate sedimentation, for example, as inferred for the Cape Storm Formation. Kleffner (1990) discussed Silurian chronostratigraphy and demonstrated that the Pridoli and Llandovery are of equal duration, representing about 2/3 of the Silurian, and also, that the Ludlow and Wenlock are of equal duration, each representing about 1/6 of the Silurian. Although these estimates, too, have intrinsic error due to some initial assumptions, Kleffner's (1990) calculations applied to the Arctic biozone scheme indicate that the biozones of the Ludlow and Pridoli are of *approximately* equal duration, but are double the time spans of each Wenlock and Llandovery biozone. Thicker rock sequences in the Ludlow and Pridoli, in part, may reflect variable biozone time spans as worked out by Kleffner (1990), or alternatively suggest a slightly greater rate of carbonate accumulation. The Llandovery platform succession, in contrast, shows a slightly lower rate, perhaps related to several factors, including paleoenvironment, accommodation, or the sub-Cape Storm Formation or other intraformational disconformities. Similarly, the lower Llandovery deep-shelf and mud buildup successions are condensed or thin, respectively, and likely related to hiati or depressed sedimentation during these times. Although there may be some error associated with the preceding assumptions, the comparison of depositional rates is likely most accurate in a comparison within biozones, particularly extrabasinally, where many biozones are globally synchronous.

Clean carbonate deposition occurred in most areas following deposition of the Irene Bay Formation, but the thin carbonate unit assignable to a basal tongue of the

Allen Bay Formation, is unfossiliferous and lacks conspicuous mottling. Based on a regional study of the underlying Thumb Mountain Formation, mottiing, although a predominantly diagenetic structure that is partly controlled by bioturbation, seems to decrease in abundance basinward, possibly suggesting some bathymetric control on its distribution. Perhaps conditions for autochthonous carbonate deposition following deposition of the Irene Bay Formation were marginal, and that carbonate platform sedimentation was in a depressed state following the Irene Bay Formatiom. Subsequent sea-level rise in uppermost Ordovician (*fastigatus* Zone) caused major regional platform backstepping which is recorded in Ellesmere (Plate 8a,b), Cornwallis, and Devon islands. The *fastigatus* Zone deepening event cannot be correlated with certainty outside the basin, and it does not correspond to the peak period of Gondwana continental glaciation (Brenchley, 1988). The *fastigatus* platform backstepping may correlate with the basal Irene Bay-like rubbly limestone of the Allen Bay Formation or with the light brown trilobite-rich unit recorded in the Cañon Fiord stratigraphic sections.

Platform margin drowning in the *fastigatus* Zone was gradual, however, and was marked by a rhythmic alternation of bioturbated and laminated limestones and shales, containing abundant trilobites and graptolites. These fossiliferous beds are replaced by black shales and uncommon bedded chert below the Ordovician-Silurian boundary carbonate beds (as observed in sections 3, 35, 56, 58, 59, 60, Fig. 7a,d,f,g). In carbonate platforms, both recent (Boardman and Neumann, 1984) and ancient

(Grotzinger, 1986; Goldhammer and Harris, 1989) tidalites and rhythmic interbedding of periplatform marlstone and shale have often been ascribed to high-frequency sea-level fluctuations that are ultimately controlled by glacioeustasy and "Milankovitch-band" orbital perturbations. For example, in toe-of-slope and basin-margin settings, off-platform carbonate transport occurs during lowstands and pelagic sedimentation during high stands (Sarg, 1988; Droxler and Schlager, 1985; Schlager and Chermark, 1979). Because of the lack of Silurian carbonate fossils of planktonic origin, this study should be ideal for demonstrating this relationship. These ideas may be simplisticly applied to the rhythmic bedding of the *fastigatus* Zone limestone-shale alternations, but the following observations pose problems for this interpretation: (i) the lack of coeval platform cyclicity; (ii) the lack of platform cyclicity in Hirnantian platform carbonates, a time of maximum continental glaciation and presumably a time of maximum influence of orbital perturbations on sea level; and (iii) the lack of a regional correlation of these beds (at section 19, western Grinnell Peninsula, rhythmites are not present in coeval strata). According to ideas of "Milankovitch band" allogenic control on platform cyclicity, sea-level fluctuations should have been at a maximum during Hirnantian continental ice buildup, but these effects are not recorded in platform carbonates. Perhaps high-frequency sea-level fluctuations were of insufficient magnitude to effect platform bathymetry, but these fluctuations are thought to have been on the order of 100m (Brenchley, 1988; Hambrey, 1985), certainly enough to alter the depositional style of an otherwise lime mudstone-dominated platform succession. It is also interesting to note that in the Upper Silurian, more than 100 high-frequency platform cycles are recorded in the Goose Fiord Formation, presumably a time when continental glaciers were lacking, but during a time of significant tectonic activity related to Caledonian epeirogenesis. These observations suggest that there is no simple relationship of Milankovitch's mathematical equations to ancient sea-level fluctuations (particularly where these calculations are not valid for ancient orbital configurations). Clearly, there are problems with this application, particularly due to the complexity of ancient climatic and oceanic systems. Allogenic influences on platform drowning are complex and have been discussed by several authors (including Schlager, 1981, 1989). Drowning may be ascribed to sea level rise that is related to (i) "stick-slip faulting" (Cisne, 1986); (ii) regional changes of in-plane stresses of the lithosphere (Cloetingh, *et al.*, 1985; Cloetingh, 1988); (iii) flexure, associated with Llandovery "flysch" deposition, as recorded in North Greenland (Hurst and Surlyk, 1984); (iv) increased basin subsidence, as is usually associated with the onset of tectonism (Ingersoll, 1988); (v) geodetic anomaly variations (Mörner, 1976); and (vi) glacioeustasy. Little is known of recent and virtually nothing of past geodetic variations, although this influence on sea level could have been substantial (>100m) over a short time frame, and operative locally (see Mörner, 1976, for review). Upper Llandovery (diachronous) platform drowning was attributed to lithospheric flexure in North Greenland (Hurst and Surlyk, 1984), but coeval strata on Devon, Cornwallis, and southern Ellesmere

Islands show no obvious correlative deepening (or uplift) events. The regional correlation and synchronous nature of backstepping and the lack of recognized Upper Ordovician syntectonic flysch suggests that subsidence related to lithospheric flexure or in-plane stresses probably does not explain the *fastigatus* Zone platform drowning. Perhaps, platform foundering was related to an abrupt, eustatic sea-level rise over a platform that had a relatively slow upbuilding rate, hence was "prone" to drowning (as discussed above with reference to Fig. 10).

Bioturbated limestone apparently straddling the Ordovician-Silurian boundary (sections at Trold Fiord) may represent up to four graptolite zones. Some zones may be missing, or not recognized with present sampling density, or confined to condensed intervals. It is more likely that key graptolite assemblages are restricted to one or two shaly beds below or above the Ordovician-Silurian boundary carbonates. The precise age of these carbonate beds is ambiguous; they could be latest Ordovician, lowest Silurian, or both. These carbonate beds likely represent shallowing and are expanded relative to the adjacent shales. Also, it is enticing to equate them to the Hirnantian event (Brenchley, 1988; Cocks and Rickards, 1988). In either case, shallowing may have occurred up to three times, as recognized at Trold Fiord, and shales overlying these carbonates likely correlate with the *acuminatus* Zone deepening as recorded in sections at Bay Fiord. Lowest Silurian basin expansion is also evident in Judge Daly Promontory, where Silurian *cyphus* Zone shales occur over Ordovician carbonates, but, again, because of the condensed nature of the system boundary sequences, the timing of this flooding is uncertain.

It is noteworthy that the Ordovician-Silurian boundary, based on recent biostratigraphy, does not coincide with the timing of the main rise and fall in sea level. Also the extent and duration of glaciation are not well established; generally, it is considered to have been 1-2Ma in duration and produced less than 100m of sea-level fluctuation (Brenchley, 1988). By Pleistocene standards, glaciation was considerably less extensive and covered only one pole; the northern hemisphere had no land masses, based on present paleogeographic reconstructions. Theoretically, the presence of a huge, ice-covered southern landmass and equatorial and northern glacier-free water masses would have had considerable influence on climate. Storms would have probably been intense at the boundary of continental arctic and oceanic (tropical and subtropical) air masses. A strong temperature gradient associated with a vigorous meridional heat transfer was probably present. Oceanic circulation, which shows a strong dependency on atmospheric circulation, would have been equally vigourous, and is perhaps evident from grey (oxidized) Ordovician (Hirnantian) and black (reduced, organic rich) Silurian shales. Colour change, perhaps associated with this event, is noted in the area of Trold Fiord and is recorded in a number of other localities (Lenz and McCracken, 1988; Koren, 1987; Williams, 1988). Conversely, the base of the Silurian in the study area is commonly associated with black graptolitic shales. Although a simplistic interpretation, widespread anoxia, sluggish atmospheric and oceanic circulation, and a generally equable climate have been used to explain the apparently global extent of Lower Silurian black shales (Leggett *et al.*, 1981).

In summary, there were probably three separate regional relative rises of sea level near the Ordovician-Silurian boundary; these are represented by the Irene Bay Formation, by the onlapping of *fastigatus* Zone shales and by the lowest Silurian post-*acuminatus* Zone beds. The two latest episodes resulted in progressive backstepping of platform carbonates. The "*fastigatus*" and "Irene Bay" events were of greater regional significance, affecting large portions of the Ordovician carbonate platform. Subsequent backstepping in the Lower Silurian was regionally less significant, but was followed by the deposition of extensive black shale, signifying the development of sluggish oceanic circulation and widespread anoxia (Leggett, 1980; Rickards, 1988) during general environmental stability in the Lower Silurian; however, it is difficult to imagine sluggish atmospheric circulation over a large, polar continental land mass that was probably plunged annually into darkness for months at a time.

A lower Aeronian *cyphus* Zone, possibly coeval with a deepening event, is also of regional significance. This event corresponds to the widespread deposition of bedded radiolarian chert, which, at Trold Fiord, is the thickest (12m) occurrence in the report area (Plate 8c,d). Enhanced biological production of silica could have been facilitated by upwelling and flooding of the platform with nutrient-rich waters, and possibly signifies a climatic event and/or sea level rise. In the vicinity of Baumann Fiord, this event coincided with about 500km^2 of platform backstepping. Mudmounds and reef pinnacles were established subsequently and are discussed in

Part III of this report. It is also noteworthy that a global, basal Aeronian highstand is recorded by Johnson (1987), an event consistent with *cyphus* Zone platform backstepping.

The correlation of Ordovician-Silurian boundary sea-level fluctuations or the possible influence of climate change on the carbonate platform succession is tenuous. As discussed above, the boundary between the lower and middle members of the Allen Bay Formation is considered to coincide with the Ordovician-Silurian boundary. This approach may be erroneous, however, because the probable latest Ordovician "Hirnantian event" records a 1-2Ma sea level fluctuation, that was followed by a sea level rise in latest Ordovician. The apparent lithological break between the two formations should then be entirely Ordovician (perhaps Hirnantian) in age. The Ordovician age of the prominent lithological break between the two members is evident in sections 47 (Fig. 7d) and 17 (Fig. 8a). The boundary between the lower and middle members of the Allen Bay Formation is unlikely to be entirely coincident with the Ordovician-Silurian boundary or with the Hirnantian event in all examined stratigraphic sections, but a fairly sharp lithological break generally occurs near the systems boundary and is represented by either dolostone on limestone or grainstone on mudstone. The lithological break is a convenient boundary between the lower and middle members of the Allen Bay Formation.

Middle dolostone member (informal name)

Lithology and distribution

The unit has a lower contact, as discussed above, that is very near to, or coincides with, the Ordovician-Silurian boundary. The upper contact is gradational, and in southern Devon Island coincides with the change from dark brown dolostone below to lighter brown striped dolostone above (Thorsteinsson and Mayr, 1987, p.164). The contact is difficult to locate on cliff exposures but is generally placed where light coloured (predominantly peritidal) deposits become predominant over brown (subtidal) deposits. The entire middle member is cyclic, sparsely to very fossiliferous, and intensely dolomitized.

On Grinnell Peninsula, large portions of the member are well exposed, but all examined outcrops were faulted, and a complete stratigraphic section was never obtained. In section 17 (Fig. 8a), although the Ordovician-Silurian boundary is poorly constrained biostratigraphically and lithologically, the boundary between the lower and middle members is chosen at the base of a shallow peritidal unit (unit D; Fig. 8a). However, this peritidal facies is a local occurrence and is attributed to a near-shore influence, this section being the most eastern (cratonward) section examined. The peritidal unit is abruptly overlain by medium brown, thickly bedded petroliferous dolostone more typical of the middle member.

Reefs have been observed in sections 15 (Fig. 9b) and 17 (Fig. 8a) and reported elsewhere by Thorsteinsson and Mayr (1987) from southern Devon Island. These structures are up to 13.5m thick (section 17 unit E, Fig. 8a) and are dolomitized. They contain various amounts of stromatoporoid framestone and have abundant dololutite that in places predominates. Stromatoporoids commonly weather light brown, are laminar to globular or rarely digitate, and were selectively dissolved, then preferentially silicified, and are lined with coarse, centripetal, euhedral dolomite, partly or completely filling voids. The final infilling appears to be calcite and bituminous residues, probably a reflection of an originally high hydrocarbon content. Some strata underlying these buildups contain small neptunian dikes, and some off-reef strata are conglomeratic, suggesting that the reefs had some primary relief.

Reefs (of sections 17 and 15) are interbedded with medium brown, thickly bedded, petroliferous dolostone containing three types of lithologies that also occur in section 18 (Plate 1a): (i) medium bedded, continuously laminated and microbially laminated, mudcracked, pale brown to almost white, vuggy dolostone; (ii) thickly bedded, sparsely fossiliferous medium brown dolostone with conspicuous stromatoporoid biostromal or floatstone intervals (Plate 1c); (iii) pentamerid (?*Pentamerus*) rudstone and floatstone, commonly 1 or 2m in thickness and rarely up to 4m in thickness. Some of these beds are continuous for several hundred meters, although lateral estimates are usually not possible due to discontinuous exposure and limited accessibility. Some dolostone beds form irregular cycles that consist of three main components: (a) sparsely fossiliferous to unfossiliferous, thoroughly bioturbated dolomitized mudstone; (b) fossiliferous laminar stromatoporoid floatstone and rare, local biostromal framestone; and (c) pentamerid floatstone to rudstone. Some cycles are capped by white laminated and mudcracked dolostone. These three lithologies are irregularly interbedded forming poorly defined cycles as much as 10's of meters thick. Uncommon crinoid holdfast bindstone occurs locally. The cyclic units are similar to platform cycles described by Johnson and Campbell (1980), Johnson and Lescinsky (1986), and Johnson *et al.* (1981, 1989), but a detailed regional correlation of cycles could not be made due to poor biostratigraphic constraints.

Strata assignable to this member occur on south-central and northeastern Ellesmere Island (Table 2c). In **section 61** (Fig. 9d), Swinnerton Peninsula, uppermost Allen Bay Formation beds are characterized by vuggy, tan brown to dark brown, bioturbated to fossiliferous, medium crystalline, very thickly bedded, vuggy dolostone. Globular and laminar stromatoporoids commonly form floatstone units interbedded with bioturbated fossiliferous and sparsely fossiliferous dolostone. Other lithologies observed include light brown, laminated dolostone interbedded with the darker (subtidal) dolostone. Some floatstone beds show an upward succession of digitate and globular, to tabular, to laminar stromatoporoids. Also, an originally mottled texture is inferred, and is represented now by large cm-scale vugs.

In **section 62**, a great thickness of the middle member is exposed (Fig. 9c). The lower contact, as discussed above, is considered to lie near the Ordovician-

Silurian boundary, and the upper contact is gradational and placed where very resistant, finely crystalline limestone of the upper limestone member (discussed below) becomes predominant (Plate 1f). The most common lithology is a medium brown weathering, bioturbated, sparsely fossiliferous, petroliferous dolostone. Chert occurs only sporadically, and faint mottling and some pentamerid beds are present; however, the brachiopod-rich zones (dominated by smooth-shelled ?*Pentamerus*) are distinctly less common than on Devon Island (for example sections 16, Fig. 9b & 17, Fig. 8a). Other conspicuous macrofossils include tabular and globular stromatoporoids, and tabular coral colonies (most commonly *Favosites*). These beds, particularly near the top of the member, are interbedded with lighter coloured, pale brown and grey laminated dolostone, forming cyclic units (tidalites). Cycles comprise alternating bioturbated, variably fossiliferous, thickly bedded, medium brown dolostone and brown and grey laminated dolostone. Laminar stromatoporoids occur uncommonly in these units together with favositid corals. Laminites and intraclast beds are abundant in upper parts of cycle tops.

About 15km northwest of the junction of Irene Bay and Bay Fiord, coeval strata are predominantly biohermal (Plate 1e), hence are unlike the units of the middle member and are described below under a separate heading.

In **section 31** (Fig. 9c), Darling Peninsula, the limited biostratigraphic control suggests that at least 5 units are time-equivalent with the Allen Bay Formation, as established further south. A lower predominantly brown dolostone unit 235m thick (**unit 2**) is probably assignable to the middle member of the Allen Bay Formation recognized in the south. This unit contains abundant *Pentamerus* indicating a middle Llandovery to middle Wenlock age. Other conspicuous macrofossils include laminar stromatoporoids, favositids, and syringoporid corals. These units are interbedded with skeletal packstone, and bioturbated lime mudstone. Also, amorphous white and brown chert nodules occur throughout, but are particularly common at the base of the unit. Pentamerids occur rarely in association with stromatoporoids.

Unit 2 is gradationally overlain by very conspicuous, pale brown-weathering, massive dolostone (**unit 3**; Fig. 9c; Plate 7d). It consists of very coarsely crystalline, vuggy (up to 30% porosity), resistant dolostone with only very poorly preserved stromatoporoids and corals. Favositids are by far the easiest to recognize, and probably have the greatest preservation potential; hence, their frequency is probably overestimated. Remnant stromatoporoid framestone is preserved together with uncommon oncolite-rich rudstone units.

Unit 3 is gradationally overlain by less resistant medium bedded, brown to light brown dolostone (**unit 4**, Fig. 9c) that is predominantly laminated, mudcracked, and fenestral. Intraclast beds and wavy, microbially laminated dolostone are most common. Other sedimentary structures include common cross lamination and flaser and wavy lamination. Cycles display a vertical succession from laminar stromatoporoid floatstone to skeletal calcarenite to wavy and domal microbial laminites interbedded with intraclast beds and calcarenites.

Unit 5 (Fig. 9c) is a thickly bedded, resistant, burrowed, variably fossiliferous, and mottled dolomitic limestone. Two conspicuous laminar stromatoporoid bindstone intervals occur in this unit, but medium brown bioturbated lime mudstone is clearly most common. This limestone rarely contains laminar stromatoporoid floatstone, and the contact of unit 5 with overlying unit 6 is abrupt and is equated with the Cape Storm-Allen Bay contact; however, this conclusion is based partly on lithological grounds. Sparse conodont fauna (Fig. 9c) indicates that this contact is Middle to Late Silurian in age. Beds immediately overlying unit 6 yielded Ludlow age atrypoid brachiopods (J. Jin, written com., 1990), suggesting a correlation with the Douro Formation further south (see Appendix two for brachiopod identifications by J. Jin, 1990). Assignment of unit 7 to the Douro Formation strengthens the case for unit 6 being Cape Storm equivalent.

Upper peritidal dolostone member (informal name)

Lithology and distribution

In cliff exposures on Grinnell Peninsula, this member is typically light brown and grey striped dolostone and limestone and contrasts with the underlying darker brown middle dolostone member and overlying recessive, light yellowish brown dolostone of the Cape Storm Formation. In southern Ellesmere Island and near Makinson Inlet (also as described by Kerr, 1975; Mayr *et al.*, in prep.), the upper limestone member is homogeneous, light brown and discontinuous. The upper contact with the Cape Storm Formation in both areas is likely disconformable. Thicknesses of the member are given Table 2a.

In **sections 16 and 18**, a unit of predominantly laminated and microbially laminated peritidal dolostone is present in the lower part of the upper member (Fig. 9b,d). Above this, the member tends to be unfossiliferous, laminated (dolosiltite to dolarenite), medium bedded, grey to light brown dolostone interbedded with darker brown, bioturbated or laminated dolostone. The amount of limestone is variable, but is particularly common towards the top of the member. Wavy microbialites, laterally linked hemispherical stromatolites, and conspicuous intraclast beds are also present. Although components are obscured by dolomitization, many units appear to be pelleted or have spongiostromate textures (including vermiform, stomatoid, and fenestral microfabrics).

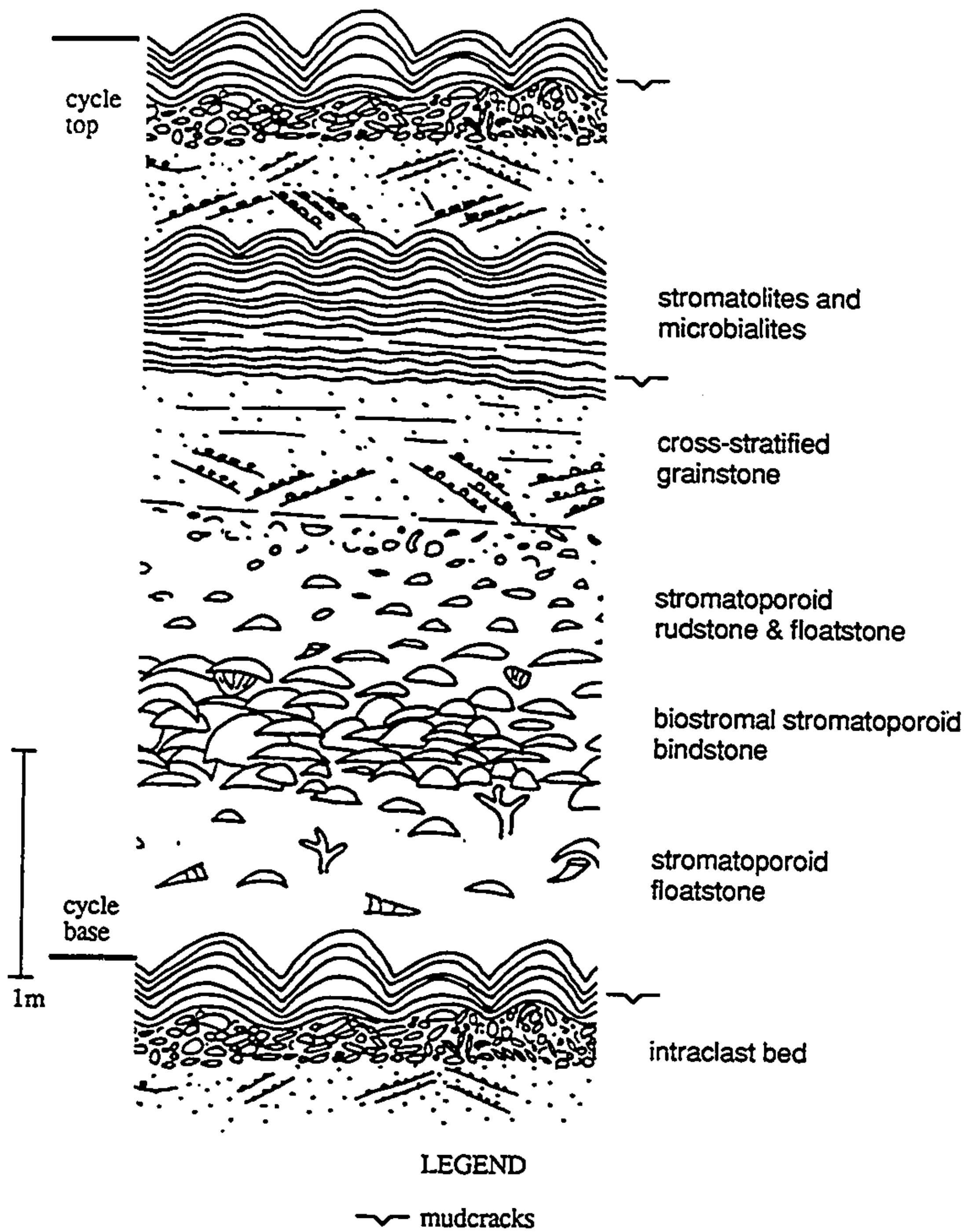
Ninety metres above the base of **section 18** (Fig. 9d), a predominantly peritidal succession begins (unit A, Fig. 9d). It is about 120m thick and predominantly medium bedded, laminated, and microbially laminated grey to light grey dolostone. At **section 16**, the approximately coeval peritidal succession is about 160m thick (unit A, Fig. 9b) and has gradational upper and lower contacts. In both stratigraphic sections, peritidal unit A is gradationally overlain by a cyclic succession (unit B, Fig. 9b) representing overall upward shallowing. These youngest units are laminated, grey and light brown, mudcracked, microbially laminated, medium bedded, medium crystalline dolostone that are interbedded with sparsely fossiliferous, bioturbated, dark brown, petroliferous dolostone with uncommon digitate and laminar stromatoporoids, favositid corals, and uncommon brachiopods. At the top of the succession, stromatoporoids and other bioclasts are pervasively silicified, a distinctive character of the fossils in the upper member of the Allen Bay Formation on Grinnell Peninsula and Devon Island (see Plate 2b). Much of the upper portion of the upper member is cyclic, and is described below with reference to section 15.

Other less common features of the upper member include (i) evaporite pseudomorphs (after gypsum), (ii) breccia beds up to 3m thick, (iii) graded laminae (of dolosiltite, dolarenite, or dolorudite) containing flame and load structures, (iv) laterally linked stromatolites that tend to grade upward into wavy microbialites, (v) large tabular and trough cross-stratified dolostone units, some with erosional bases cutting 10's of centimetres into the underlying laminates, (vi) large spar-filled voids, perhaps replacements of nodular evaporites, and (vii) cephalopod rudstone beds, presumably allochthonous and preserved with domal stromatolites and laminated dolostone. From west to east, there is a conspicuous overall increase in the number and thickness of intraclast beds, fewer peritidal beds, and a greater number of sedimentary structures indicative of highly restricted conditions and/or exposure. Also, subtidal fossiliferous units (in particular, stromatoporoid bindstone) increase westward in abundance. Stromatoporoid bindstone biostromal deposits are particularly abundant at the head of Eidsbotn Fiord and in the northern part of the Douro Range.

Section 4a, about 5km west of section 15 (Fig. 9b), was visited briefly to examine the Cape Storm-Allen Bay Formation contact. The upper peritidal member of the Allen Bay Formation is predominantly a medium bedded, well-indurated, light grey to yellow-grey, medium crystalline, unfossiliferous dolostone. Uncommon limestone is interbedded with dolostone, patterned carbonate, and ubiquitous intraclast beds. The member is disconformably overlain by the Cape Storm Formation. The Allen Bay-Cape Storm formation boundary at this locality (and at most localities throughout Grinnell Peninsula and Devon Island) is placed at (i) the change from resistant limestone and dolostone to more thinly bedded and recessive silty dolostone and (ii) below the first occurrence of thinly bedded, laminated, yellow-brown weathering silty dolostone. The silt content is, however, much less in sections examined at Makinson Inlet and further north (discussed below).

Cyclicity is well developed in **sections 15, 16,** and **18** (Figs. 9b,d, 11), and 28 complete cycles have been recognized in section 15 (Plate 2a,c). The cycles occur within a 220m interval bounded below by a covered interval and above by a poorly exposed, Cape Storm-Allen Bay formation contact. A typical cycle (Fig. 11; Plate 2c) comprises from base to top a (i) discontinuous, subtidal, biostromal, stromatoporoid bindstone (Plate 2b) containing from most to least common, laminar, globular, and digitate stromatoporoids, calcareous algae, colonial tabulate corals (commonly

Figure 11: Schematic representation of a typical high-frequency cycle of the upper middle and upper members of the Allen Bay Formation at Eidsbotn Fiord (section 15), Devon Island.


cycle top
stromatolites and microbialites
cross-stratified grainstone
stromatoporoid rudstone & floatstone
biostromal stromatoporoïd bindstone
stromatoporoid floatstone
cycle base
1m
intraclast bed
LEGEND
mudcracks

Favosites), solitary rugosans, cephalopods, brachiopods, and sponges (including *Hindia sphaeroidalis*); and (ii) an upper supratidal succession with laterally linked stromatolites, wavy microbialites, cross-stratified grainstone and rudstone, and intraclast beds. Cycles near the base of the member are more complete and show well-developed biostromal intervals. Some biostromal units are replaced laterally by or are interbedded with, sparsely fossiliferous bioturbated lime mudstone, rudstone, and floatstone. Commonly, abruptly overlying the lower, dark-brown weathering subtidal lime mudstone parts of the cycles are cross-stratified, poorly to well-sorted rudstone beds up to 2.5m thick. These are interbedded with large domal stromatolites and intraclast beds. Large starved ripples are also present and were rarely the nuclei for subsequent domal stromatolite growth. Laminites, and wavy, flaser, and lenticular bedding are present but are less prevalent than in sections further east (as in sections 16, 18). Laminae in the dolostone appear to be of two types: thin, wavy, discontinuous and thin, even, continuous. The latter consists of interlaminae of graded calcisiltite and calcarenite, with uncommon basal flame and load structures. Some sediment deformation (by density stratification) has created ball-and-pillow-like structures, similar to those observed in some clastic sediments. Mudcracks occur, some as large prism cracks.

Noteworthy east-to-west variations in the above sections include the following: thickness and fossil content of dark-weathering subtidal portions of cycles increase; laminated dolostone becomes less common; and intraclast beds and domal stromatolites become more common. Also, a predominantly tidal-flat succession at the base of the member is present in eastern localities but not to the west.

Upper limestone member

This member has a restricted distribution. It is included in the upper dolostone member as defined on Devon Island by Mayr and others (in prep.). However, although its age is poorly known, it is, in this investigation, likely equivalent to the upper part of the Allen Bay Formation, and is considered distinct from the upper dolostone member of northern Devon Island. Thicknesses of the member are given Table 2b.

Lithology and distribution

The upper limestone member of the Allen Bay Formation (Plate 1f) is a peculiar unit in southwestern Ellesmere Island and is likely coeval with the upper dolostone member described above on Grinnell Peninsula and vicinity. It is known only in the vicinity of Baumann Fiord, Makinson Inlet, and southern Ellesmere Island. Because this unit is a thickly bedded limestone and typically forms a resistant ridge between two relatively recessive dolostone successions, it is an important marker used in geological mapping to separate the Allen Bay and Cape Storm formations.

The limestone is predominantly coarse pelsparite. In hand specimen, individual pellets resemble micritized ooids, but the pellets lack clear concentric laminae, are poorly sorted, and merge with vaguely clotted micrite. These textures are clearly evident in thin sections, but are otherwise obscured by the weathering of fine-grained limestone.

The upper limestone member of the Allen Bay Formation is also well exposed in **sections 61, 62, and 64** (Fig. 9c,d; Plate 1f). In these sections, the thickly bedded limestone member gradationally overlies the middle brown dolostone member and contains basal, poorly sorted rudstone beds interbedded with laminated calcilutite, calcisiltite, and calcarenite. Also near the base, a distinct disconformity separates dark brown brecciated and stromatolitic, dolomitic limestone from underlying irregular, massive cyclic peritidal dolostone. Local relief on the disconformity is up to 2m (Plate 1f); however, samples collected above and below the disconformity as well as throughout the upper limestone member were barren of conodonts.

The characteristic strata above the disconformity are thickly bedded, pelmicritic, microbially laminated, laminated, and oolitic limestone. Precise proportions of these lithologies are difficult to estimate due to limited accessibility and obscurity due to the weathering characteristic of the limestone; however, peloidal micrite is most common, and clotted, fenestral, and stromatolitic micrite, and stromatoporoid-coral-rudstone are less common and were partly affected by dolomitization, particularly near faults.

Peloidal laminae in the limestone are generally wavy, discontinuous to continuous, and are approximately 2-10mm in thickness. Uncommon intraclast beds contain clasts that are well to moderately sorted, and admixed with crinoid debris (uncommon). Dolomitization of the various lithologies was selective such that interspar has been replaced by euhedral, turbid dolostone rhombs, while peloids and intraclasts have remained mostly unaltered. Centripetal inclusion-rich blocky calcite on peloids suggests an early marine acicular (high magnesium?) cement phase. Peloids, and litho- and bioclasts in laminae are poorly to well sorted and some show inverse to normal grading. Some intraclasts are imbricated and contain cortoids, numerous coated grains, and some grapestone. Meniscus cements on peloids were observed in one intraclast, suggesting vadose diagenesis. Calcite coating on intraclastic peloids is typically very thin, although well developed oolites do occur as grainstone, or as packstone preserved in intraclasts.

Interlaminated with pelsparite are thin micrite laminae. These are typically undulatory and are interlaminated with fenestral or homogeneous micrite. Where dolomitization is incipient, laminae, like peloids, tend to be least dolomitized.

The age of the limestone succession in the area of Makinson Inlet has not been established, as conodont identifications were not available at the time of this work; all five samples were barren. Mayr and others (in prep.) similarly commented on the paucity of conodonts in these limestone units, but were nevertheless able to extract conodonts of Wenlock aspect from this member.

Middle and upper dolostone and limestone members of the Allen Bay Formation: age and interpretation

Although samples were collected and processed for conodonts, many yielded non-diagnostic elements or were barren; hence, ages of the members and correlations are, in part, lithostratigraphically based. The presence of Wenlock to Ludlow age conodonts in section 16 and middle Llandovery age conodonts in section 15, both occurring near the top of the middle member, suggests that the contact between the upper and middle members is diachronous, younging eastward. The presence of middle Llandovery conodonts 20m above definite Upper Ordovician argillaceous limestone in western Grinnell Peninsula suggests that the Llandovery sequence is comparatively condensed (sections 20, 21, Fig. 8a).

The age of the Allen Bay-Cape Storm contact is crucial as it represents a major platform disconformity through much of Devon and southern Ellesmere Island. From the ages established on northern Devon Island, it appears that this contact is indeed diachronous, disconformable, and youngs westward. Although the age of the lowest beds of the Cape Storm Formation is poorly known in the study area, a Ludlow age of the formation is established (section 6-4, Fig. 9a). Based on similar lithologies and stratigraphic relationships established on southern Devon Island (Thorsteinsson and Mayr, 1987), the Cape Storm Formation described herein is considered to be essentially isochronous throughout the study area, deposited entirely during the middle Ludlow (*siluricus* Zone). This conclusion is supported by the established age of the Cape Storm Formation (again as *siluricus* Zone) on southern Ellesmere Island (Mayr, *et al.*, in prep.) and on Devon Island (Thorsteinsson and Uyeno, 1980).

Age relationships and detailed facies variations of the shelf margin facies of the Allen Bay Formation on western Grinnell Peninsula are not completely established, and correlations at this point are subjective. These facies relationships are discussed separately below.

The middle and upper members of the Allen Bay Formation on northern Devon Island are an upward shallowing sequence overall, punctuated by numerous high-frequency cyclic units (Fig. 11); however, in western Grinnell Peninsula, the thin lower Llandovery peritidal carbonate package (section 21, Fig. 8a) cannot be correlated across the platform and possibly was autogenically controlled. A similar explanation could be used for the peritidal unit on eastern Devon Island (section 17, Fig. 8a), as its lateral continuity cannot be proven. As discussed below, localized deposition of peritidal carbonates (section 21) in an otherwise predominantly subtidal carbonate shelf succession indicates localized shelf margin conditions, and perhaps the incipient development of islands (and reefs).

The middle member is probably entirely Llandovery in age, and represents deposition on an open marine carbonate ramp. Coeval platform margin units (section 21) are relatively condensed, and lack evidence of allochthonous debris beds and/or well-developed shelf margin facies. The lack of coarse allochthonous debris beds in the proximal deep-shelf facies (the Cape Phillips Formation) at Cape Sir John Franklin and at Sheils Peninsula is further evidence for sedimentation marginal to a ramp rather than to a steep platform slope. In the open marine ramp setting, bioherms, and cyclic units, represented by alternation of laminar stromatoporoid biostromes, pentamerid rudstone, and sparsely fossiliferous bioturbated mudstone, were deposited. Lateral continuity of cycles could not be established, and cycles may have been allogenically and/or autogenically controlled.

The Allen Bay Formation in the study area, as a whole, is a shallowing upward sequence overall. On Cornwallis Island, a more completely exposed platform-to-basin sequence indicates upward shallowing as well, and is perhaps related to the development of an effective barrier, preserved as a shelf-margin patch reef complex (de Freitas and Dixon, 1989a,b; de Freitas *et al.,* 1989). Coeval shelf-margin barrier facies were not examined in detail on Grinnell Peninsula, but the patchy occurrence of oolites in peloidal limestones at examined localities may imply the development of an oolitic complex to the west (of sections 20 and 21), although further work is needed to substantiate this claim. Nevertheless, it appears that shelf shallowing on Grinnell Peninsula was initially abrupt and diachronous, as indicated by the lower peritidal tongue of the upper member. Apparently, shallowing was earlier in the western areas (section 15 versus section 16, Fig. 9b) and may have been related to inherent shoaling at the shelf margin.

Cyclicity in the upper member of the Allen Bay Formation in the study area is represented by metre-scale cycles. The cycles are most numerous at Eidsbotn Fiord (section 15; Plate 2a) where 28 have been recognized. Because the cycles represent abrupt deepening followed by gradual shallowing, they are comparable to some Quaternary and many ancient tidalites. Climatic forcing and orbital variations (glacioeustatic "Milankovitch band" cycles) have commonly been used to explain such cyclicity. This idea is controversial, as glacioeustatic sea-level fluctuations during ice-free periods in earth history are difficult to justify. However, it has been speculated that glacio-eustasy may be caused by fluctuating alpine ice, which may have been present in the Wenlock (Hambrey, 1985). Clearly, one cannot entirely discredit an allogenic mechanism for generating high-frequency cycles. Convincing evidence for allogenic (glacioeustatic) control on ancient shelf tidalites is preserved as either vadose diagenetic cycle tops (Strasser, 1988; Goldhammer *et al.*, 1987), or as lateral continuity of cycles, none of which could be demonstrated in the tectonically disrupted sequence. Hence, the cause of cycles is equivocal.

Several noteworthy conclusions can be drawn from the facies relationships described above for the cyclic peritidal succession. The common association of stratified rudstone and grainstone beds (Fig. 11) and domal stromatolites, both of which abruptly overlie subtidal, fossiliferous units, indicates upward shallowing and increasing energy. Rudstone and grainstone are poorly to moderately sorted, laterally continuous, have low-angle cross (possibly swash) lamination, some of which overlies trough cross-stratified rudstone, indicating deposition on a shoreface or possibly beach. Cross-stratified grainstone perhaps represents shoreface migration of sand waves, that in a progradational sequence scoured underlying offshore lime mudstone units (*sensu* Inden and Moore, 1983). Rudstone clasts are dolomitized, but nevertheless appear to be predominantly lithic, rounded, sorted, and derived probably from pencontemporaneously cemented beach and/or subtidal carbonates. Uncommon fragments of stromatoporoids, crinoids, and brachiopods in the stratified rudstone were perhaps derived from fringing subtidal stromatoporoid biostromes. Some sorted lithic rudstone beds abruptly (erosionally) overlie stromatoporoid bindstone, indicating that the underlying indurated stromatoporoid bindstone may have inhibited down-cutting of the high-energy facies.

Some domal stromatolites are associated with rudstone units. In recent settings, stromatolites occur in inter- or subtidal deposits, and, club-shaped (columnar) forms exist in a higher energy setting than laminar forms (Tucker and Wright, 1990, p.149; Hoffman, 1974); however, there are exceptions to the general trend even within the depositional setting on which this model of stromatolite distribution was based. These fossils are thus equivocal paleoenvironmental indicators (Shinn, 1986; Shinn, 1983b). Their occurrence and distribution can probably be attributed to a variety of factors such as wave stress, grazing, salinity, sedimentation, and hydrology. The tops of recent stromatolites are generally below mean high-water level. Because of a common association of stromatolites in this study with an underlying foreshore and/or possibly beach lithology, they probably represent agitated conditions and occupied a backshore or berm setting. In modern tidal flat settings on the Bahamian platform (Hardie, 1977), a beach levee borders protected tidal flats. Small, incipient, domal stromatolites are present on levee crests, but these are smaller than, and unlike larger linked domes of the Silurian. Moreover, in recent settings that are arid, as inferred for this Silurian example, microbialites are uncommon, and generally restricted to a narrow intertidal zone (Shinn, 1986). These observations of recent settings are further support for a beach-associated intertidal environment for domal stromatolites.

Cratonward (toward sections 17, 18, Figs. 8a, 9d), algal heads were likely replaced by extensive laminar microbialites, laminated dolostone, intraclast beds, and other less common supratidal sediments. These extensive flats were punctuated by ponds, in which ostracods flourished and lime mudstone was deposited. In northeastern Devon Island, beach deposits, domal stromatolites, intraclast beds, and open marine fossiliferous units are less common than further west, and indicate shoaling and the development of more extensive supratidal flats. Exposure was great and hypersaline conditions are indicated locally by evaporite pseudomorphs and breccia beds. Intraclast beds are uncommon in northeastern localities (sections 16, 18, Figs. 9b,d), but storm sedimentation is evident from planar, tabular, and lenticular cross-beds, some erosionally overlying laminites. Although tidal channels have not been identified with certainty, they are generally scarce in ancient and recent progradational tidal flat sequences (Shinn, 1986).

The poor to good sorting, the moderate size of generally labile lithic fragments, the thin laterally continuous nature of the "beach-unit", and the upward gradation of this unit into low-energy supratidal deposits indicates deposition in a relatively low-energy beach environment. There is no evidence of a penecontemporaneous shelf-margin barrier, but perhaps fringing subtidal stromatoporoid biostromes reduced shelf-interior wave stress, and resulted in the observed generally fine-grained tidalites.

Cyclic units of the upper member are partly or mostly truncated by the sub-Cape Storm disconformity, and in southwestern Ellesmere Island, are probably coeval with the upper limestone member. This informal member of the Allen Bay formation was originally included from the lower part of the Cape Storm Formation (Kerr, 1975; Morrow and Kerr, 1977), but because of the abrupt thickness variation over short distances and the omitted conodont zones at its upper boundary, it seems very likely that this contact is disconformable. These arguments are not discussed further here, as Mayr, *et al.*, (in prep.) give a lucid discussion of this contact. However, the lithology and composition of the upper limestone member are distinct in contrast to the coeval microbially laminated upper dolostone member on Grinnell Peninsula.

There is no apparent sedimentological reason for the occurrence of a relatively thin, well-preserved limestone member within an extensively dolomitized platform succession, particularly where the environmental setting was broadly similar during deposition of these members. A peritidal interpretation is generally appropriate for the upper limestone member, for the lower part of the Cape Storm Formation, and for the middle member of the Allen Bay Formation.

Dolomitization in peritidal settings is regarded as diagenetically early, and among other processes, generally related to reflux or downward seepage of hypersaline brines (Land, 1985). Despite the fact that supratidal lime muds are commonly dolomitized preferentially (Murray and Lucia, 1967), the upper limestone member is generally fine grained, clotted, pelleted, and fenestral, and overlies subtidal dololutite of the middle member. Dolomitization of the Allen Bay sequence was likely early (Veizer *et al.*, 1978; Morrow and Kerr, 1977), but the following units were unaffected: (i) the lower limestone member of the Allen Bay Formation, (ii) the Douro Formation, (iii) the lower member of the Goose Fiord Formation, and (iv) the upper limestone member of the Allen Bay Formation. The first three sequences were deposited probably on relatively deep ramps rather than in extensive peritidal environments. However, the upper limestone member of the Allen Bay Formation is a peritidal carbonate, and yet is undolomitized. Clearly, depositional setting was not the only influence, and other factors, such as original mineralogy or porosity may have been important.

Laminae in the upper limestone member are comparable with recent laminites (Wanless *et al.*, 1988; Hardie, 1977). Millimetre-scale couplets of peloidal and homogeneous, fenestral, and finely laminated micrites, suggest ephemeral storm sedimentation on microbe-rich tidal flats. Abundant associated sedimentary structures, including graded laminae, imbricated intraclasts, etc., are further evidence for this interpretation. These strata are generally well indurated, but show incipient dolomitization that appears to have been controlled by permeability variations associated with neospar in pelsparite laminae. Early marine cementation is evident from turbid acicular inclusions on peloids; vadose diagenesis is evident from rare meniscus cements occurring rarely on peloids and cortoids in intraclasts. Microbialites are present on a disconformity and as well are interlaminated with peloid-rich laminites (pelaminae). Fine grained pelaminae and small intraclasts of granule beds indicate sedimentation on a low-energy tidal flat.

The lower member of the Devonian Blue Fiord Formation, a peritidal succession outcropping in the area of Makinson Inlet, lithologically resembles the upper limestone member of the Allen Bay Formation: it is a fenestral limestone succession containing up to 200m of almost entirely microbial and fenestral carbonate (pers. obs.). This type of carbonate could have been deposited by microbial mats (Burne and Moore, 1987; Shinn, 1983b), which trapped and bound micrite and/or induced calcium carbonate precipitation. Moreover, great thicknesses of such sediments are typically associated with humid supratidal "algal marsh" settings (Hardie, 1977; Shinn, 1986, 1983a). In arid climates, algal growth is frequently interrupted by long periods of exposure, allowing wind erosion and desiccation.

However, in humid, protected climates, storms wet the mats, preserving them from subsequent erosion and allowing extended growth without the necessity of a marine flooding. If this occurred in a "protected" landward supratidal setting, mats could accumulate great thicknesses of sediment, largely uninterrupted by ephemeral storm sedimentation. Under these conditions, persistent growth of mats would result in extensive fenestral and microbial carbonate deposition, such as is inferred for the lowest member of the Blue Fiord Formation and for some parts of the upper limestone member of the Allen Bay Formation. It can be speculated that diagenesis in the meteoric phreatic environment was influenced by the predominantly micritic sediment deposition. The microbially controlled calcite precipitation, and frequent wetting would have helped to seal these rocks, preventing early dolomitization of the lower limestone member by seepage or reflux of brines, while strata above and below the member were extensively dolomitized.

UNDIVIDED SILURIAN PLATFORM CARBONATES

Unusual facies equivalents to the Allen Bay Formation-Read Bay Group occur in several areas. These are listed and discussed separately below, and some may, with future stratigraphic work, be ranked as separate formations.

(i) pinnacle reef biolithite facies and coeval mudmound facies,

(ii) mud buildups established over the drowned Ordovician shelf margin,

(iii) shelf-margin limestone facies, Grinnell Peninsula,

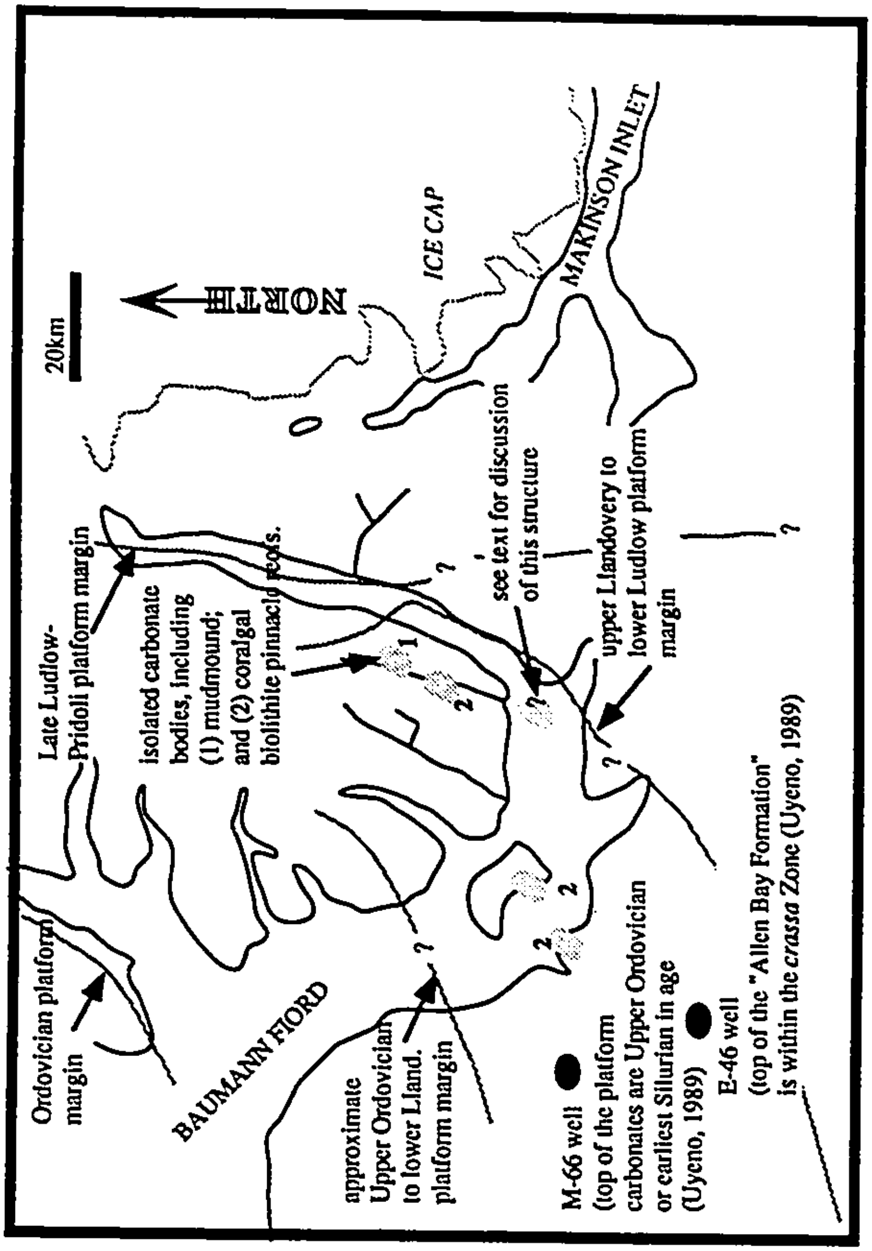
(iv) shelf-margin biohermal facies,

(v) stromatoporoid platform foreslope reef facies.

Pinnacle reef biolithite and coeval mudmound facies

Several pinnacle reefs occur near Baumann Fiord, one in western Hoved Island (Plate 6a-c), another 10km north-northeast of the junction of Baumann and Vendom fiords (Fig. 12). Reef exposures on Gunnars Island and on an unnammed peninsula 1km west of Gunnars Island shows broadly similar facies relationships and biostratigraphy to those of the pinnacle reef on Hoved Island. However, in a stratigraphic section along the northeastern coast of Gunnars Island, a middle Wenlock to middle Ludlow slope sequence suggests that the reefal carbonate exposed at these two localities pinches out toward the northeast, toward Hoved Island. The evidence therefore suggests that these two localities (unnamed peninsula and Gunnars Island) represent separate pinnacle reef occurrences, although the structure in the unnamed peninsula is known only from reconnaissance studies. Another possible pinnacle reef occurs at the junction of Vendom and Baumann fiords, although its presence is indicated only by an exposed slope sequence (Plate 4), as discussed further below.

Figure 12: Unrestored paleogeography of pinnacle reefs and mudmound in relation to Silurian platform margins. Two main episodes of backstepping of the platform are indicated in this area.


20km
NORTH
ICE CAP
MAKINSON INLET
Ordovician platform margin
Late Ludlow-Pridoli platform margin
isolated carbonate bodies, including (1) mudmound; and (2) coralgal biolithite pinnacle reefs.
see text for discussion of this structure
upper Llandovery to lower Ludlow platform margin
BAUMANN FIORD
approximate Upper Ordovician to lower Lland. platform margin
M-66 well (top of the platform carbonates are Upper Ordovician or earliest Silurian in age (Uyeno, 1989)
E-46 well (top of the "Allen Bay Formation" is within the *crassa* Zone (Uyeno, 1989)

None of the pinnacle reefs is completely exposed, and it is not clear whether these structures were originally isolated reefs or are simply erosional remnants of an originally very sinuous shelf margin; however, several facts inconsistent with the latter interpretation tend to favour the pinnacle reef interpretation:

(i) The facies sequence -- coralgal biolithite capped by oolitic shoals -- is unlike coeval platform margin facies occurring north and south of Baumann Fiord;

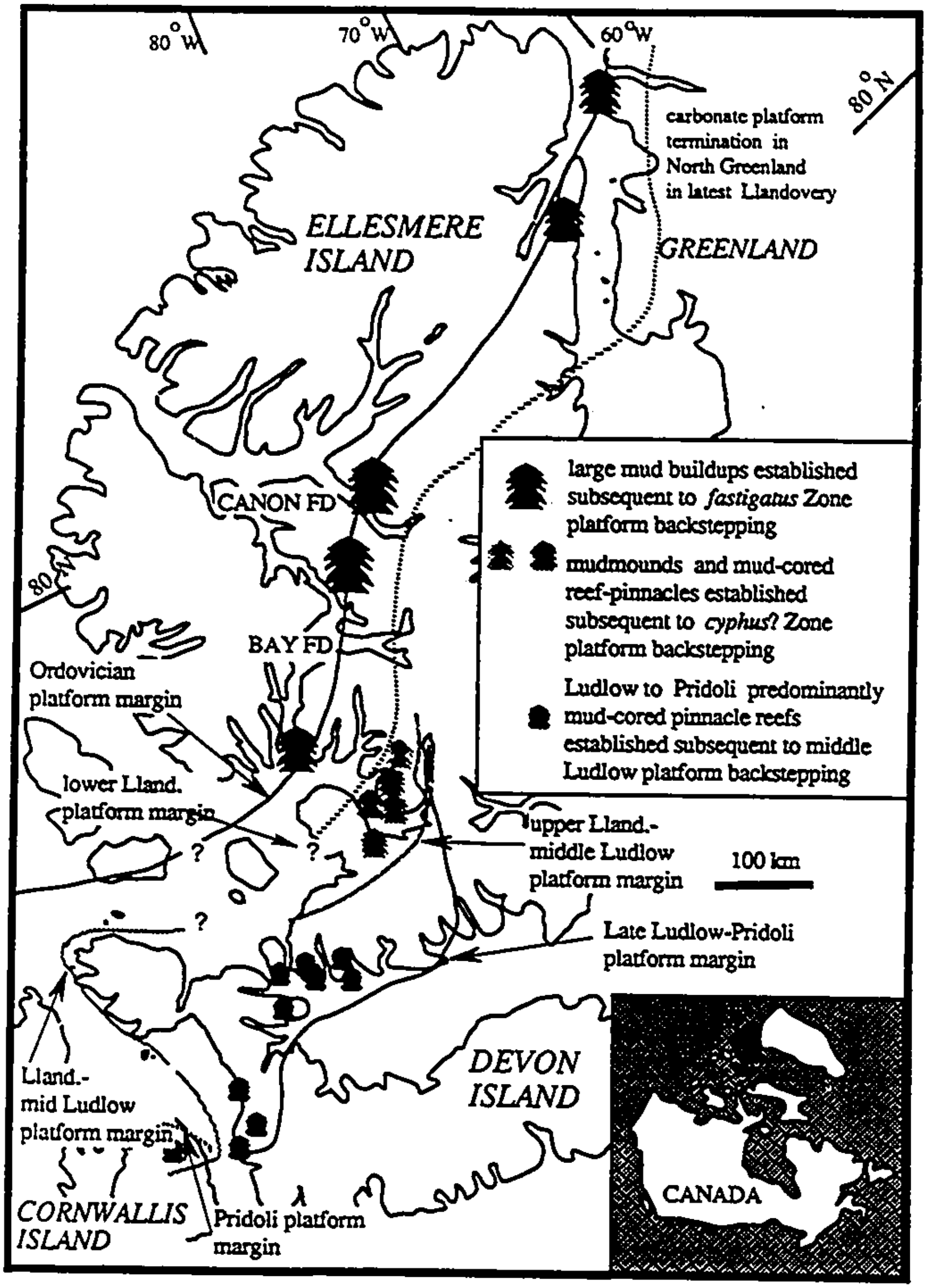
(ii) Re-entrants in the carbonate shelf margin, for example, in the area of Hall Land, North Greenland, indicate platform margin sinuosity of 100's of kilometres of wave length and not the approximately 10's of kilometres necessary if the pinnacle reefs represent platform margin bights.

(iii) To accommodate biostratigraphic relationships established in Bjorne Peninsula by Uyeno (1989), sinuosity of the platform margin would have to have been at least one or two orders of magnitude greater than that known for the Silurian Arctic platform sequence elsewhere (Figs. 12,13).

(iv) One of the buildups, examined on a reconnaissance scale, west of locality 69 (Fig. 3) occurs in the hanging wall of a west-verging thrust, west of coeval shales and platform slope facies; therefore, it is undoubtedly a deep-shelf structure located several kilometres basinward of the Silurian platform margin (as delineated in Figs. 1, 12, and 13).

(v) The slope sequence on Hoved Island adjacent to the pinnacle reef indicates

Figure 13: Distribution of large mud buildups, pinnacle reefs, and mudmounds in relation to main Late Ordovician to Late Silurian shelf-margin positions. Occurrence on on southern Devon Island from Thorsteinsson and Mayr (1987), on northern Devon Island from Mayr *et al.* (1987), on southern Ellesmere Island from Mayr, *et al.* (in prep.), on Judge Daly Promontory from unpublished maps of J.C. Sproule and Associates, and on Hall Land from Sønderholm *et al.* (1987).



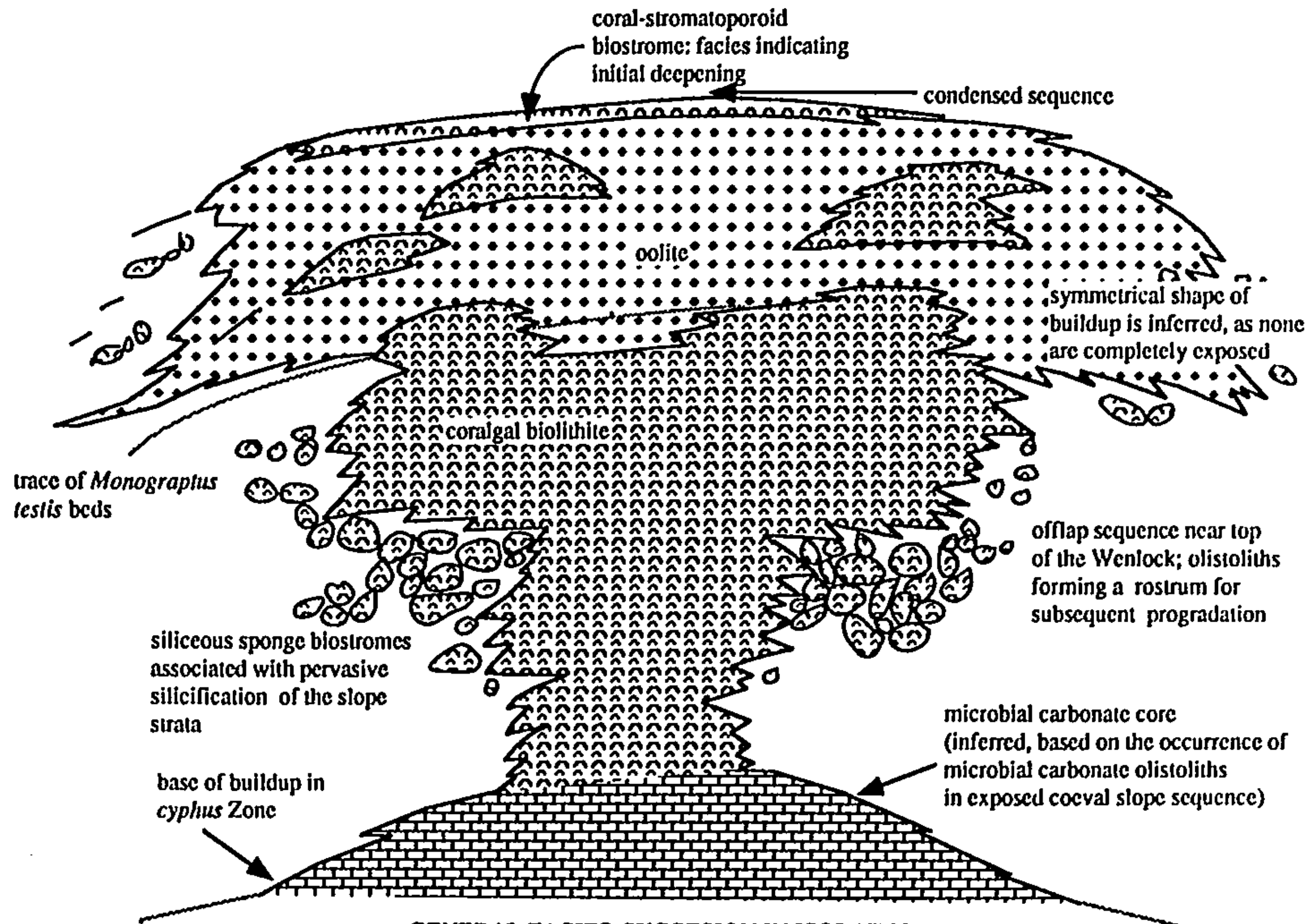
80°W
70°W
60°W
80°N
80°
carbonate platform termination in North Greenland in latest Llandovery
ELLESMERE ISLAND
GREENLAND
CANON FD
BAY FD
Ordovician platform margin
lower Lland. platform margin
upper Lland.- middle Ludlow platform margin
100 km
Late Ludlow-Pridoli platform margin
Lland.- mid Ludlow platform margin
Pridoli platform margin
CORNWALLIS ISLAND
DEVON ISLAND
CANADA
large mud buildups established subsequent to *fastigatus* Zone platform backstepping
mudmounds and mud-cored reef-pinnacles established subsequent to *cyphus*? Zone platform backstepping
Ludlow to Pridoli predominantly mud-cored pinnacle reefs established subsequent to middle Ludlow platform backstepping

allochthonous sediment transport apparently toward the southeast, essentially *toward* the trace of the Silurian shelf margin, in contrast to the regional trends of sediment transport related to the Silurian shelf margin sequence. A similar facies relationship can be seen on the northeastern coast of Gunnars Island, suggesting that, as discussed above, the two reef exposures on Gunnars Island and the adjacent unnamed peninsula were also isolated carbonate bodies.

It is concluded that these carbonate structures were indeed isolated and developed basinward of the Silurian shelf-margin. Because of their thickness, inferred depositional relief, and enclosure in shaly strata, they are termed pinnacle reefs. However, because none of these structures is completely exposed, their precise configuration is unknown. They may, for example, be more laterally extensive than inferred in the schematic reconstruction (Fig. 14), and consequently may more appropriately be considered as isolated carbonate banks.

These reefs are predominantly a coral-stromatoporoid-sponge microbial bindstone, termed coralgal biolithite in this report, as corals are the most abundant skeletal metazoans but are significantly less common than microbial carbonate. The microbial carbonate typically forms thick rinds of dark brown micrite encasing most macrofossils Plate 5b,c,e) and locally forms a "lime mudstone". Macrofossils present include tabular heliolitids, fasciculate rugosans, and tabular favositids, and stromatoporoids. Turbinate and digitate (anthaspidellid) lithistid sponges locally form a framestone fabric (Plate 5c), but these, too, are thickly encrusted by

Figure 14: Schematic representation of the main facies recognized in Baumann Fiord pinnacle reefs. None of the structures is completely exposed, however, and they may be more complex and laterally extensive than suggested by this diagram.



GENERAL FACIES SUCCESION IN ISOLATED CARBONATE BODIES (REEF PINNACLES) IN THE AREA OF BAUMANN FD.

total thickness
about 420m

microbialites. Some bulbous micrite crusts on metazoans are irregular and form large primary cavities that have peloidal infilling (Plate 5b). Some cavities are up to a metre in diameter and are filled with what appear to be large hemispherical cavity-filling, botryoidal calcite hemispheres; however, on slabbing, the hemispheres are simply large micrite-encrusted skeletal metazoans. Other less common components of the coralgal biolithite include locally abundant megalodont bivalves, ooids, and pentamerid brachiopods. Isopachous fibrous cements are locally common. Due to the weathering, details of the lithology and internal structures are difficult to discern in these exposures. The paleogeographic setting and significance of these structures are discussed further in Part III of this report.

The well exposed slope sequence depicted in Plate 4 is an important outcrop in the Canadian Arctic, but its interpretation, however, is not clear. For several years these facies have been interpreted as part of the Silurian platform margin sequence, part of the facies boundary between platform carbonates to the east and deep shelf shales to the west (Mayr, 1974; Poey, 1988; Mayr *et al.*, in prep.; McGill, 1974); however, this interpretation can be questioned for several reasons, as follows:

(i) Allochthonous facies illustrated in Plate 4 are lithologically and chronologically similar to *in situ* facies exposed in the reef pinnacle at Hoved Island (Plate 6a,b) and in the section exposed at the head of Baumann Fiord. For example, the main phase of ooid deposition corresponds to about the base of the *testis* Zone, and drowning occurred in the middle Ludlow.

(ii) The blocks are exceedingly large (Plate 6c), and are *very* similar lithologically: most are coralgal biolithite, typical of the pinnacle reefs.

(iii) The resedimented blocks have a very similar sedimentological interpretation: the olistostromes have little associated allochthonous matrix, suggesting that most boulders simply tumbled or were dispersed downslope. Isolated blocks in shales are very common and are up to 20m in diameter. The sedimentological interpretation of the Lower and Middle Silurian platform slope sequence on Cornwallis Island (de Freitas and Dixon, 1990, 1989a,b; de Freitas *et al.*, 1989) is superficially similar. However, blocks there are predominantly stromatoporoid rudstone and framestone, in contrast to the oolitic and biolithite deposits associated with pinnacle reefs.

(iv) No shelf margin sequence that has been studied in this region shows the lithological characteristics and facies relationships that can be seen in the olistostromes (Plate 4) or in the known pinnacle reefs. For example, coeval shelf margin facies exposed only 10km north of the area shown in Figure 12 comprise the dolomitized equivalents of a very different suite of original facies.

This slope exposure may represent the vestiges of a reef pinnacle flank that was subsequently onlapped by platform carbonates. This interpretation implies that, while some pinnacle reefs were drowned in middle Ludlow time, this structure (Plate 4) persisted and was overstepped by platform carbonates, presently represented by rubbly weathering argillaceous limestone containing abundant calcareous algae, pentamerid brachiopods, and oncoids.

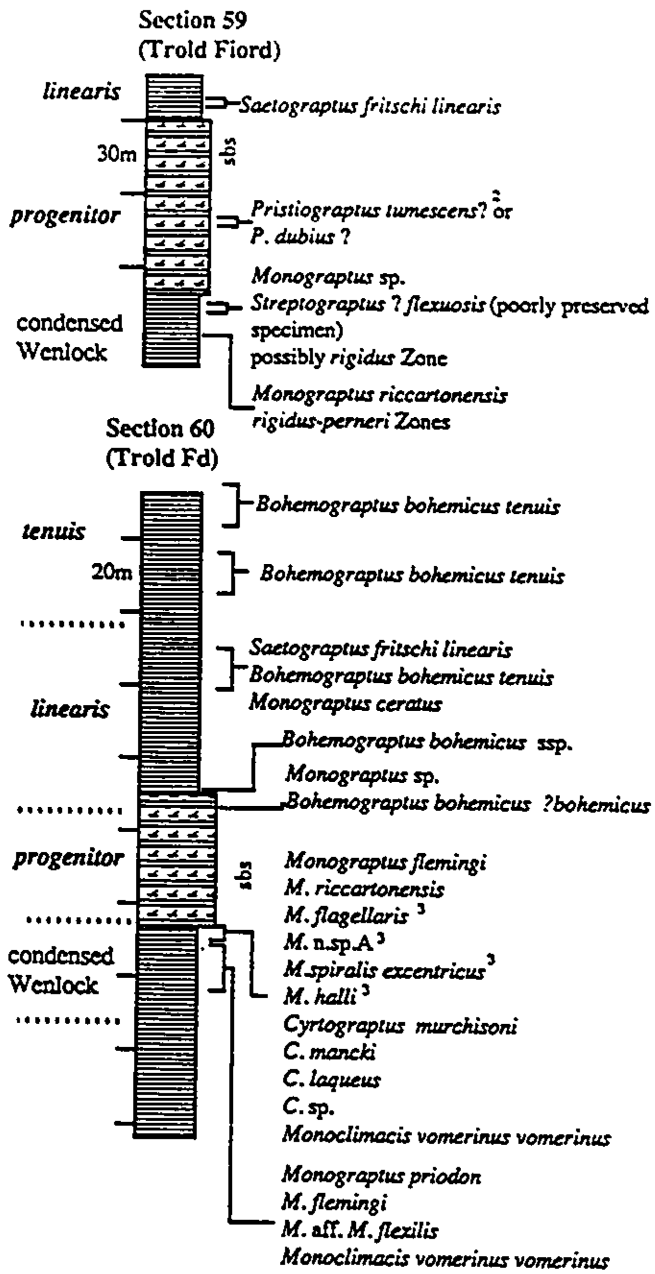
An extensive bioturbated Llandovery carbonate sequence is exposed in the vicinity of Baumann and Vendom fiords (discussed in Part III) and is overlain by pinnacle reefs. The sequence varies in thickness throughout the study area, and has a diachronous upper contact: in sections 54 (Fig. 7e), 48, and 76 (Fig. 15c) it has a gradational upper contact, within the upper *sakmaricus* to lower *centrifugus* graptolite zones, and contains significant thicknesses (up to 2m) of intraclast beds. Many of these intraclast beds are bioturbated, which obscures primary sedimentary textures. These beds probably originated by transport on a submarine slope, and clast buoyancy in the debris flow was likely maintained by matrix yield strength.

Intraclast beds are extensive in the slope facies exposed at Baumann Fiord (Plate 4) and in the mudstone-rich succession underlying the mudmound 16km north of this exposure. Intraclast beds occur together with disarticulated pentamerid (*Pentamerus*) rudstone beds, some up to 2m in thickness, and olistoliths. The presence of microbialite clasts in allochthonous debris beds and the diachronous nature of the upper contact of the Llandovery limestone succession with the overlying shales suggest local upbuilding, perhaps due to microbially influenced precipitation of carbonate. Local relief resulted, and blocks were incorporated in debris flows that moved down submarine slopes.

The pinnacle reefs of coralgal biolithite are overlain by, and are interbedded with, extensive oolites (Fig. 14; Plate 1d). Also, in the vicinity of Gunnars Island, in-place coralgal biolithite is directly associated with oolites, an indication that the two

Figure 15a-d: Stratigraphic sections showing biostratigraphy of Cape Phillips Formation in relation to the Starfish Bay shale. Biozone boundaries are subjective and are included in some diagrams to show the age relationships of the shale in the study area. Biozone boundaries are not shown in the composite section 48, 78 or, in section 69, because these sections were not sampled extensively for graptolites. Legend as in Figure 6d.

Fig. 15a



[1] identification by J. Jin, pers. com., 1990

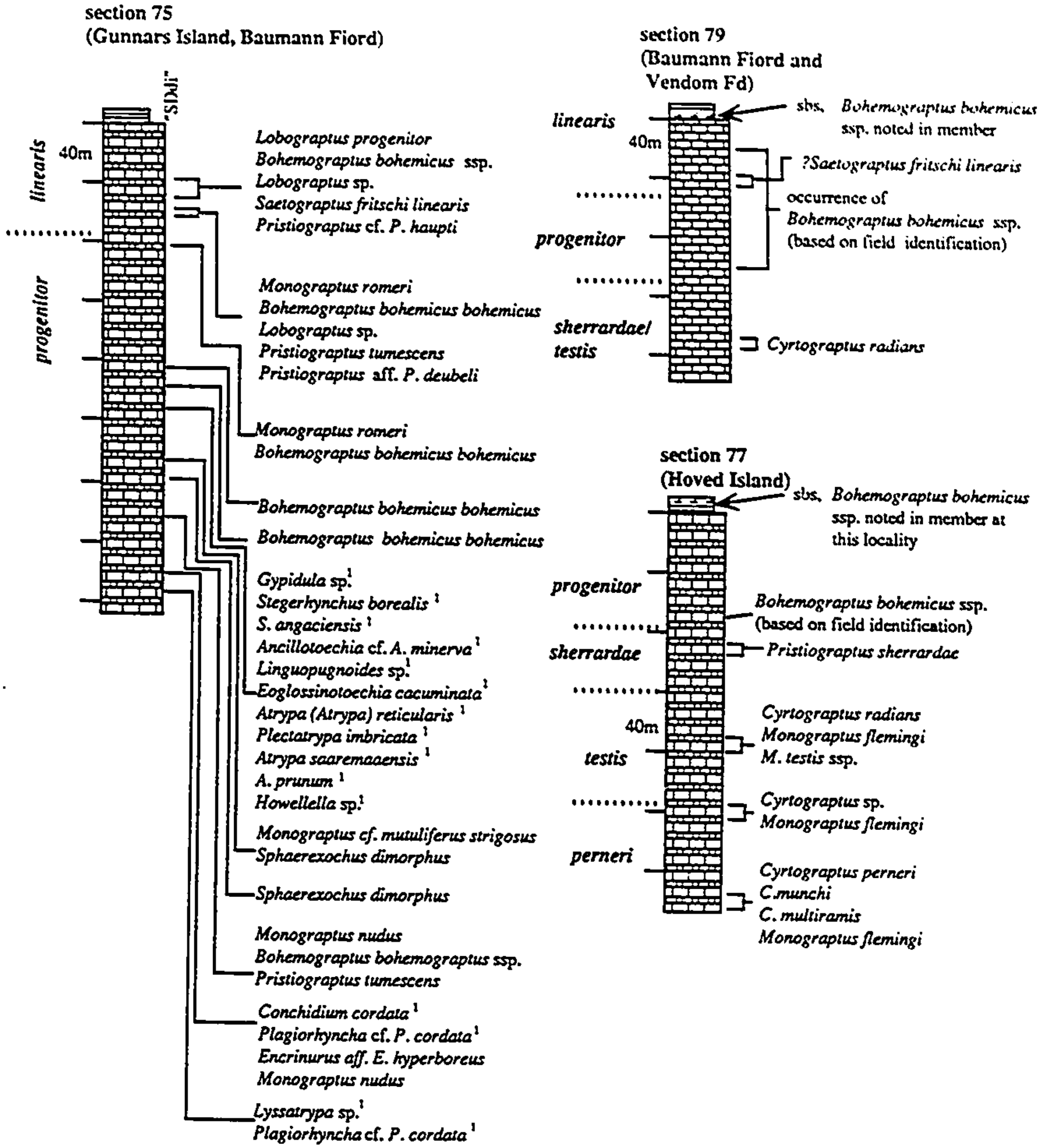
[2] identification by A.C. Lenz, 1989, 1990

[3] graptolites identified by M.J. Melchin, who commented on their poor preservation and their questionable identification. Indeed, these specimens are out of biostratigraphic context.

[4] At this section, the Starfish Bay shale is poorly exposed and is inferred from rubble. However, only 2km to the north (section 80), the shale is well exposed and about 5m thick. From this latter locality J.C. Harrison was able to collect *Bohemograptus bohemicus* ssp. T. Uyeno identified a conodont sample within and below this unit at this locality. He assigned the poorly preserved conodont elements to the *patula* Zone, but based on abundant graptolite samples from the same interval, the conodont identification is considered less reliable.

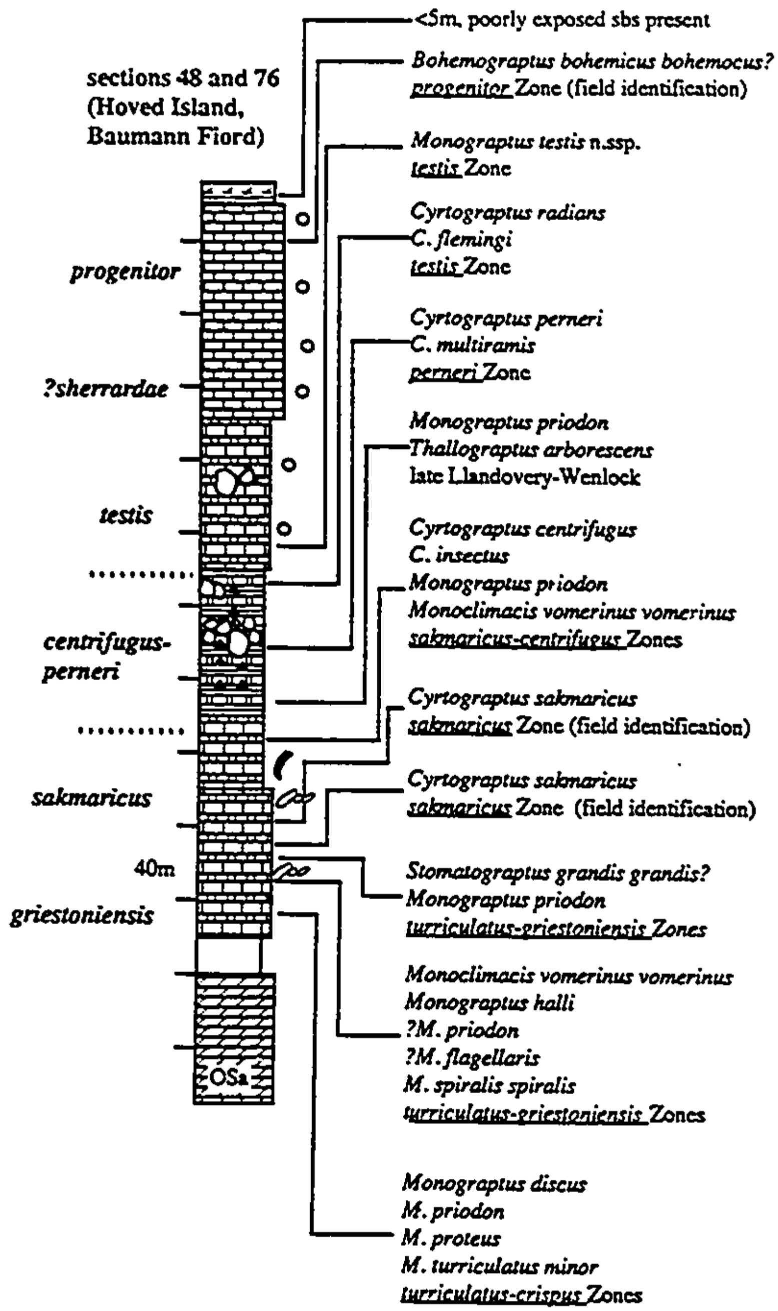
sbs Starfish Bay shale

Fig. 15b



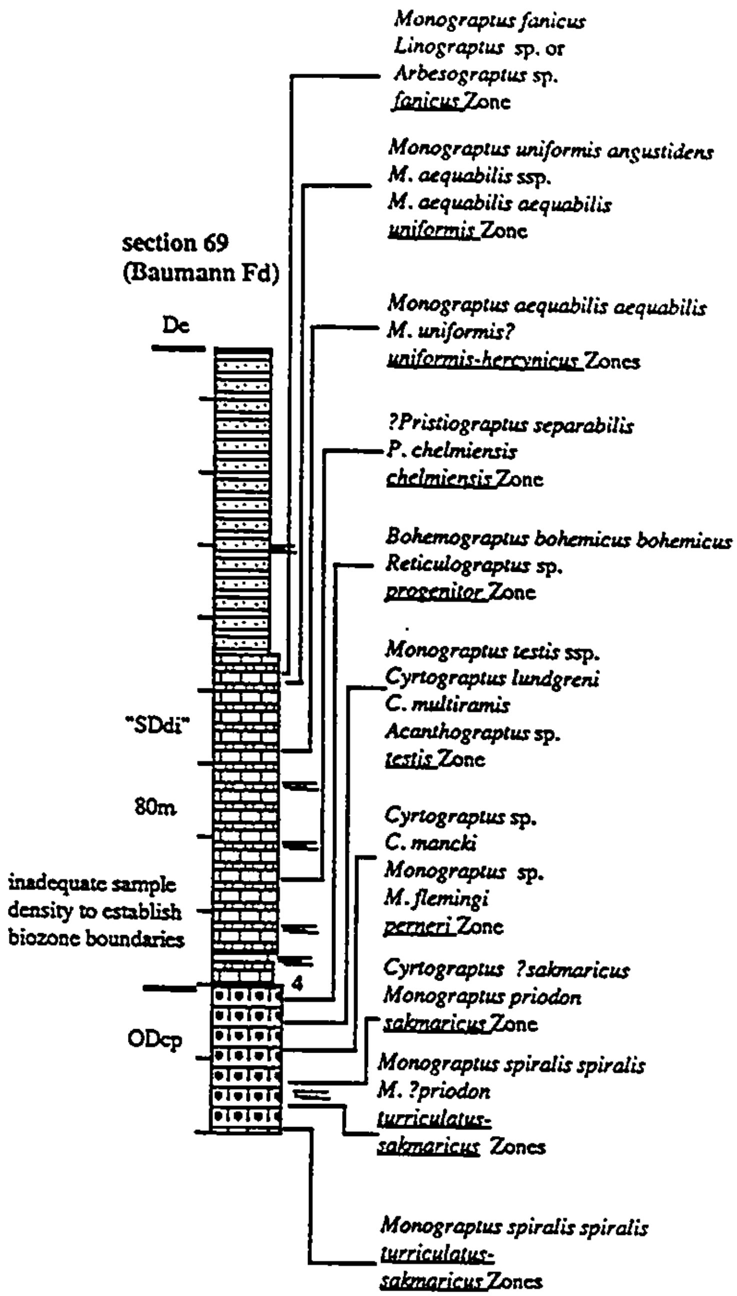
Note: 240m below base of shale limestone contact in section 75 the conodonts *Ozarkodina excavata excavata* *Panderodus* sp. were noted (T.Uyeno, written com., 1990)

Fig. 15c



Wenlock and Ludlow strata of this section were likely deposited on the proximal slope of a pinnacle reef

Fig. 15d



section 69
(Baumann Fd)
De
"SDdi"
80m
inadequate sample density to establish biozone boundaries
ODcp
4
Monograptus fanicus
Linograptus sp. or
Arbesograptus sp.
fanicus Zone
Monograptus uniformis angustidens
M. aequabilis ssp.
M. aequabilis aequabilis
uniformis Zone
Monograptus aequabilis aequabilis
M. uniformis?
uniformis-hercynicus Zones
?Pristiograptus separabilis
P. chelmiensis
chelmiensis Zone
Bohemograptus bohemicus bohemicus
Reticulograptus sp.
progenitor Zone
Monograptus testis ssp.
Cyrtograptus lundgreni
C. multiramis
Acanthograptus sp.
testis Zone
Cyrtograptus sp.
C. mancki
Monograptus sp.
M. flemingi
perneri Zone
Cyrtograptus ?sakmaricus
Monograptus priodon
sakmaricus Zone
Monograptus spiralis spiralis
M. ?priodon
turriculatus-sakmaricus Zones
Monograptus spiralis spiralis
turriculatus-sakmaricus Zones

facies were coeval, although the oolitic deposits were predominant. This association is reflected in the contiguous slope facies (of sections 54, 76, 75) where large reefal blocks are associated with allochthonous oolitic carbonates. Although the contact between oolite-dominated and coralgal biolithite-dominated pinnacle reef facies is not exposed, associated reef flank deposits indicate that ooids became common in the *testis* Zone, and were likely terminated sometime within the *linearis* Zone (Figs.14, 15b; sections 75, 79).

The presence of oolites at the top of the succession puts some constraints on the geometry of these otherwise incompletely exposed reef bodies. Although the term *pinnacle reef* is used, these carbonate bodies were probably flat-topped. Oolites make up a 60m thick succession (exposed on the east coast of Gunnars Island) suggesting prolonged oolith formation and a relatively flat-topped structure instead of a narrow pinnacle. The term "isolated carbonate bank" may be more appropriate to imply this configuration, but the term would be misleading, as associated facies are clearly unlike those of some modern and ancient carbonate banks (*sensu* Bice and Stewart, 1990; García-Mondéjar, 1990; Wilson, 1975), particularly with respect to their lateral association of facies. Although there was probably a coralgal biolithite facies rimming the upper part of the reef pinnacle, it was apparently discontinuous and allowed sufficient wave-energy into the reef interior for ooids to form. The ooids were not completely contained by the discontinuous biolithite facies, and prograded over the slope facies (Fig. 14). Relief during this phase of development was apparently not as great, and olistoliths are less common and substantially smaller. It is also important to note that a small reefal structure in the hanging wall of a west-vergent thrust sheet just west of section 69 (Fig. 15d), although apparently coralgal biolithite, is not associated with oolites. Also, it is coeval with, and relatively close to another mudmound, described in Part III of this study, and also discussed by de Freitas (1990b). These relationships suggest that subsidence in these areas could have been slightly greater than estimated for the Hoved Island exposure and that the associated reef pinnacle did not become progradational and develop the "critical flatness" for ooid formation and containment. Hence, in the vicinity of Baumann Fiord there appears to be a continuous spectrum of structures, from mudmounds, to non-oolitic pinnacle reefs, to oolite-capped pinnacle reefs, that are all inferred to have been generated by differential subsidence, which in turn affected associated carbonate sediments.

Drowning of the pinnacle reef carbonates began with initial deepening (Fig. 14), as indicated by wackestone and muddy stromatoporoid-coral floatstone and bindstone that abruptly overlie oolites and coralgal biolithite. Drowning was completed in the *linearis* Zone, where the carbonates are abruptly overlain by a condensed succession of siltstone and shales. All the buildups in the Baumann Fiord area appear to have been terminated in about the middle of the *linearis* Zone. The youngest reef pinnacle carbonate bed is a corroded hardground containing abundant, thick, metalliferous encrustations, now mainly represented by iron oxides.

The upper part of the slope sequence (section 54, Fig. 7e; Plate 4) shows a slightly different development than in other areas. Whereas on Hoved island, strata overlying reefal carbonates are entirely marlstone and shale, coeval strata in section 54, exposed predominantly as frost-riven rubble, are rubbly weathering, fossiliferous limestone, rich in calcareous algae, oncoids, pentamerid brachiopods, and other fossiliferous beds. This sequence is, in part, similar both lithologically and biostratigraphically to the Douro Formation and perhaps to the lowest part of the Goose Fiord Formation; although no supporting conodont dating is available at present, biostratigraphy reported by Poey (1988) strongly suggests that these limestones are Douro- equivalent. This upper portion of the succession grades into so-called Devon Island Formation shales exposed to the north, according to detailed mapping by R.Thorsteinsson (pers.com., 1989). These rubbly limestones may represent progradation of the carbonate platform over a platform slope pinnacle reef, as discussed above; however, the evidence for the existence of a pinnacle reef at this locality is equivocal.

These carbonate structures, because they occur in a predominantly shaly facies, are potential oil plays, if present in the subsurface. Panarctic Inc. addressed this potential in 1969, with one of two exploratory wells on Bjorne Peninsula penetrating one of these pinnacle reefs. The Panarctic Blue Fiord E-46 well apparently contains 959m of conformable carbonates (as described by Uyeno *et al.,* 1990); however, the stratigraphy of the subsurface reef structure is not readily comparable to that of the exposed pinnacle reefs. Although it is difficult to judge some facies simply based on millimetre-sized well chips, several main differences are present: (i) the subsurface reef is thicker than the exposured pinnacle reefs, (ii) the subsurface reef is more affected by dolomitization; and (iii) oolites are found in three of the exposed pinnacle reefs, but not at all in the subsurface. The overall succession in the subsurface pinnacle reef above the fault at well depth 1837m (Fig.7, Uyeno *et al.*, 1990) represents shallowing upward, then deepening, as follows: from 1837m-1533m, predominantly interbedded, dark-medium grey-brown, argillaceous, cryptocrystalline dolostone, calcareous, medium dark-grey, calcareous shale, and light grey-brown, fossiliferous limestone, dated as Late Llandovery to early Wenlock (Uyeno, *et al.*, 1990); and from 1533-1052m, a light brown, medium crystalline dolostone; and from 1052-2811m, light brown, pyritic, laminated lime mudstone interbedded with medium grey to medium dark grey calcareous shale.

By analogy with the exposure on Hoved Island, the interval 1533-1837m is likely predominantly allochthonous, derived from the expanding pinnacle reef, and presently overlain by the intensely dolomitized core facies. Perhaps, the greatest difference between this subcropping pinnacle reef and those exposed on Hoved Island and vicinity is the facies resting at the top of the succession. Whereas the tops of the surface exposures are generally oolite- or coralgal biolithite-dominated, the top of the subsurface reef appears to be a lime-mudstone unit that was likely deposited below wave-base. The succession, in general, appears to be similar to the platform succession exposed in section 70 (Fig.18), wherein argillaceous, rubbly weathering limestones overlie dolomitized platform carbonates, a facies indicating incomplete drowning and which is a likely correlative of the Douro Formation. The explanation for the two very different facies is not certain, but may be related to local subsidence and/or reef surface area.

The paleoenvironmental interpretation of these structures and their relationships with other mud buildups of the study area are discussed further in Part III.

Mud buildups established over the drowned Ordovician shelf margin.

Large mud buildups are exposed in vicinity of Trold Fiord, Vesle Fiord, and Cañon Fiord. Their internal architecture is probably complex, and the discussion in Part III is based on preliminary data. Additional field studies are needed to elucidate some of the remaining complexities of these large structures, particularly in the vicinity of Vesle Fiord. The details of their sedimentology are presented in Part III and will not be repeated here. Of importance from a stratigraphic standpoint is the timing of several major stratigraphic events in the development of these large buildups:

(i) establishment of the buildups shortly after *fastigatus* Zone platform backstepping.

(ii) platform backstepping and aerial restriction of the buildups at Cañon Fiord in about the *celloni* Zone;

(iii) ending of carbonate production in the lower part of the *patula* Zone (as established at Cañon Fiord);

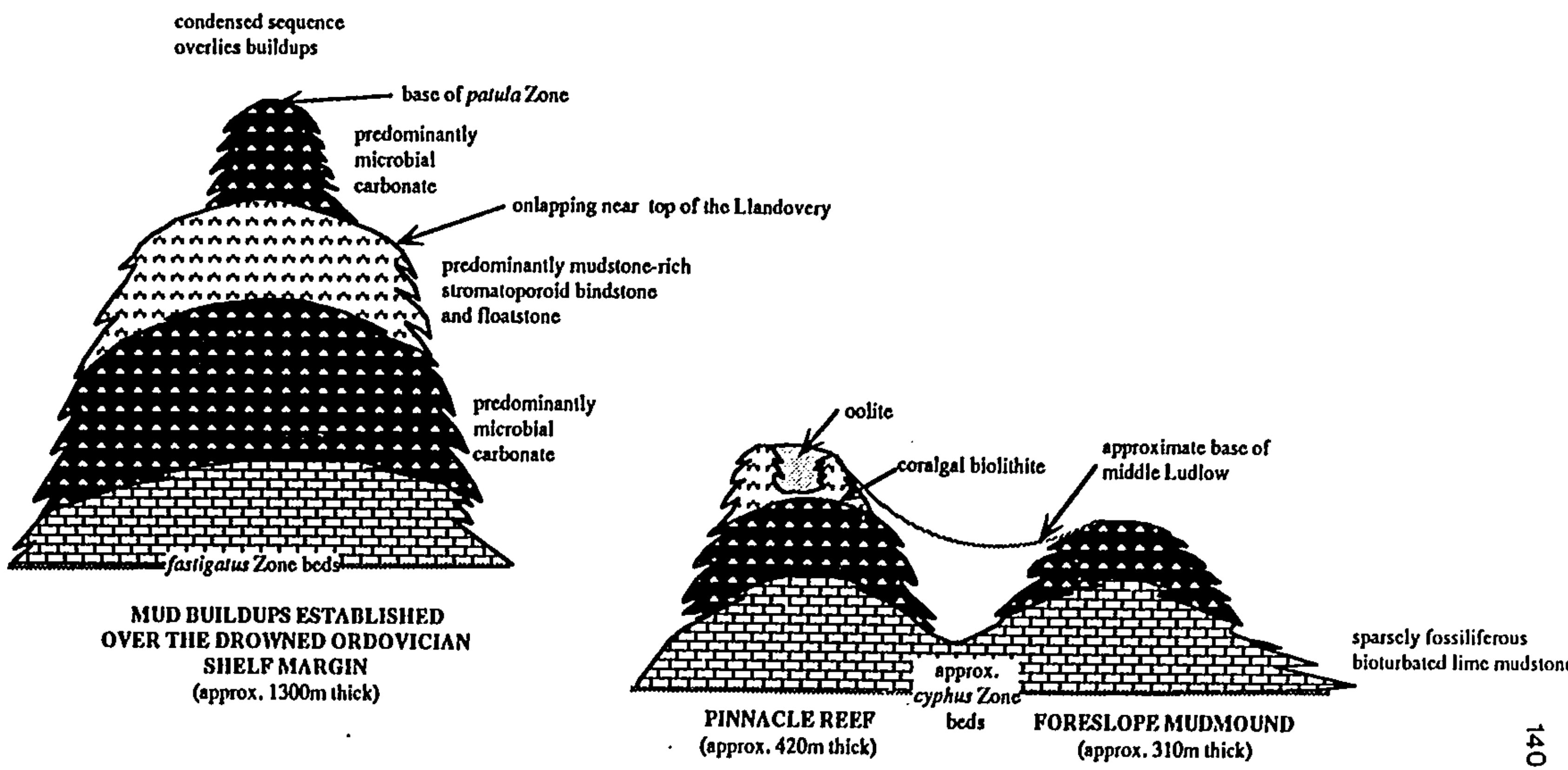
(iv) deposition of a condensed sequence of fissile black shale, resulting from inherent topographic highs over drowned buildups (Plate 5g);

The mud buildups were interpreted as part of a horst by Hurst and Kerr (1982), but clearly, they are isolated structures that developed subsequent to the *fastigatus* Zone platform drowning, discussed above. At least three of these structures have been identified on Ellesmere Island, and their location may have been tectonically controlled. This interpretation is based on facies relationships in the underlying Thumb Mountain and Irene Bay Formations. At the Ordovician shelf margin, it appears that the Irene Bay Formation is less shaly and the Thumb Mountain Formation less rubbly where they underlie mud buildups. Based on the discussion above of the sedimentology of these formations, this stratigraphic relationship suggests that subsidence influenced the location of the mud buildups.

Carbonate accumulation was very rapid (1.3km of carbonate deposited in the interval between the *fastigatus* and *patula* Zones) and was greater than for the coeval platform sequence (672m of peritidal and subtidal dolostone and limestone were deposited near Makinson Inlet) (also see Fig. 10a,b). The basinward increase in platform subsidence is apparently typical for platforms (Pitman, 1978), although subsidence through the Lower Silurian was not uniform. In the *celloni* Zone (approximately equivalent to the *crispus* and *griestoniensis* graptolite biozones), the stromatoporoid-rich unit (Fig. 16) was succeeded by microbial carbonate that coincided with areal restriction of carbonate deposition and pronounced slope onlapping. In the adjacent Hazen Trough, flysch sedimentation abruptly commenced (Fig. 10). The subsidence indicated may have been related to flexural downloading of the lithosphere beneath syntectonic flysch. Of all Silurian sequences examined in the Arctic, these deep-water clastic deposits (the Danish River Formation) show the most rapid deposition (Fig. 10), and were clearly related to a major phase of syntectonic flysch deposition in North Greenland (Hurst and Surlyk, 1984; Hurst, 1984; Hurst *et al.*, 1983). Platform sedimentation was terminated in North Greenland by flysch sedimentation but, in Canada, only minor minor onlapping in the mud buildups is indicated at that time.

Incomplete drowning of the mud buildups in the *celloni* Zone reduced the effective area of carbonate deposition and caused the upward replacement of stromatoporoid-rich limestone by microbial carbonate, possibly reflecting a bathymetric change of several 10's of meters (see Part III of this report; Fig. 16). Primary depositional relief was significant during the latter stages of mud buildup, as large (up to 8m in diameter) olistoliths were shed onto the contiguous slope. However, there are no early marine cements and few frame-building fossil in the mud buildups. This relationship suggests that microbially influenced precipitation of

Figure 16: Depositional sequences of large mud buildups, mudmounds, and pinnacle reefs in the study area.


condensed sequence overlies buildups
base of *patula* Zone
predominantly microbial carbonate
onlapping near top of the Llandovery
predominantly mudstone-rich stromatoporoid bindstone and floatstone
predominantly microbial carbonate
fastigatus Zone beds
MUD BUILDUPS ESTABLISHED OVER THE DROWNED ORDOVICIAN SHELF MARGIN
(approx. 1300m thick)
oolite
coralgal biolithite
approximate base of middle Ludlow
approx. *cyphus* Zone beds
sparsely fossiliferous bioturbated lime mudstone
PINNACLE REEF
(approx. 420m thick)
FORESLOPE MUDMOUND
(approx. 310m thick)

calcite was most important in the accumulation of the more than 1.3km of limestone.

Although some of the mud buildup structures are not continuously exposed, the three distinct buildup types show a similar vertical succession (Fig. 16). The basal two units, in particular, are remarkably similar lithologically, and overall indicate upward shallowing, from below to above storm wave base. However, in the large mud buildups established over the drowned Ordovician shelf margin, shallowing culminated with a mudstone-rich stromatoporoid succession, but with a coralgal biolithite succession in the pinnacle reefs. As discussed in Part III of this report, the pinnacle reefs likely had a greater upbuilding rate than the isochronous mudmound and mud buildups established over the drowned Ordovician shelf margin. Consequently, pinnacle reefs likely existed in a very shallow, wave-stressed marine environment, which facilited the deposition of coralgal biolithite and oolites. This environment generally excluded stromatoporoids but favoured the existence of a sparse skeletal metazoan fauna, mainly corals, thickly encrusted by microbialites.

Shelf margin limestone facies, Grinnell Peninsula

Although the shelf margin facies are an interesting aspect of the stratigraphy of Devon Island, appropriate sections were examined only very briefly due to logistical problems. The data obtained and interpretations are therefore preliminary, and further work is necessary to delineate the major shelf margin lithological variations. Due to the uniform weathering character of the predominantly finely crystalline limestone, thin sections were needed to discriminate several different rock types. Oolitic grainstone, pelsparite, stromatoporoid bindstone (and associated strata), and sparsely fossiliferous lime mudstone predominate and together form a medium bedded, fine-grained limestone.

In section 21 (Fig 8a), oolitic limestone is interbedded with predominant pelsparite and bioturbated lime mudstone. The lithology is varied near the base of the section, but oolites are particularly common. At about 85m above the base of the formation, domal and flat microbialites, laminated and intraclastic dolostone, and synaeresis cracks occur in an areally restricted peritidal unit about 10m in thickness. Conodonts from the base of this unit are middle Llandovery in age (Fig. 8a).

Crinoid-rich stromatoporoid bindstone and floatstone occur above this peritidal succession and show fibrous marine cements on stromatoporoids and other locally abundant bioclasts. These are overlain by a monotonous, finely crystalline limestone that is chiefly peloidal, sparsely fossiliferous, bioturbated micrite, locally grapestone-rich, and laminated.

No shelf-margin barrier facies was examined on Grinnell Peninsula. A lower peritidal unit (section 20, Fig. 8a) is interpreted to represent a restricted marine shelf margin setting, but coeval platform units, at section 15 (Fig. 9b), were deposited in an open marine setting (correlation is based on conodont information). The local association of biohermal and peritidal strata on western Grinnell Peninsula suggests local restriction and the development of islands at the shelf margin, as similarly interpreted for coeval strata on Cornwallis Island (Sodero and Hobson, 1979). Above this middle Llandovery island-reef shelf-margin complex, open marine conditions prevailed. However, the nature of the platform margin, whether it was rimmed or ramp-like, is uncertain, and based on current data, any interpretation at this point is conjectural.

Two features are noteworthy for the limestone facies exposed on Grinnell Peninsula. Firstly, there are numerous grapestone and oolitic units throughout the predominantly peloidal mudstone. Although these allochems could indicate derivation from a shelf-margin barrier (possibly oolitic) facies, no appropriate exposures of barrier facies were identified. Secondly, the deep-shelf margin sequence at Cape Sir John Franklin (section 19, Fig. 1) lacks coarse allochthonous strata, suggesting deposition marginal to a ramp instead of a steep rimmed platform. Based on current data, it appears that open marine carbonate deposition on a ramp prevailed, but more work is needed for a much clearer interpretation of these facies. Future work should address the following points: (i) the precise nature of the shelf margin facies, if any are in fact exposed; (ii) the origin of the limestone, which contrasts to the predominantly dolomitized platform succession exposed to the east; and (iii) the age of the upper part of the limestone succession. The latter point was addressed in this study, but conodont work was not finished in the time for this report.

Shelf margin biohermal facies

On Cornwallis Island, two main shelf-margin facies are recognized: an upper (Allen Bay Formation equivalent) biohermal facies and a thickly bedded dolomitized skeletal grainstone facies (Cape Storm equivalent). The main macrofauna in the biohermal facies includes abundant off-reef fasciculate rugosan colonies, and reefal tabular and globular stromatoporoid framestone and pentamerid (?*Pentamerus*) rudstone containing dolomitized isopachous cements. These reefs appear to form a shelf margin patch reef complex which extends approximately 3km platformward. Based on the thinness of platform slope allodapic beds and the infrequency or absence of olistostromes, the shelf margin relief at this time was probably relatively minor in comparison to Wenlock and Ludlow platform margin facies at this locality and elsewhere in the Canadian Arctic. This reefal shelf margin facies is underlain by an inter-island-bay-lagoon complex as described by Sodero and Hobson (1979) and subsequently re-examined by Mallamo (1989).

Abruptly overlying the biohermal shelf-margin facies is a skeletal grainstone facies (Ludlow, Cape Storm equivalent). The contact between the two is poorly exposed, although disconformable elsewhere, and its precise age and the nature of the contact is uncertain. Predominantly dolomitized, faintly cross-bedded, crinoid-rich skeletal grainstone occurs together with less common, laminated carbonate, skeletal rudstone, and megalodont bivalve rudstone. Biohermal strata occur sporadically, and together with the grainstone facies, grade platformward over a distance of about 2km, into a laminated, low-energy silty peritidal sequence, that is typical of the Cape Storm Formation known on Cornwallis and Devon Islands.

Clearly, the shelf margin grainstone facies on east-central Cornwallis Island represents a high energy, shelf- margin deposit, typical of equivalent deposits of other ancient and recent platforms (Gawthorpe and Gutteridge, 1990; Hine, 1977). During deposition of the Cape Storm Formation, platform accommodation was low, and a continuous keep-up succession of peritidal carbonate was deposited, with little platform progradation. Throughout the Ludlow, platform-margin relief increased gradually, but shelf margin position was relatively static and perhaps had an underlying tectonic (possibly fault) control (*sensu* Wilson, 1990; Davies, *et al.*, 1989).

However, unlike older or younger parts of the Silurian platform sequence on Cornwallis Island, the Cape Storm Formation shows a predominance of peritidal deposits which perhaps influenced the type of coeval shelf-margin facies. Inimical platform waters, possibly suggested by the platform interior facies, could have led to a "default-style" of shelf-margin sedimentation, such as interpreted for the recent Bahamian carbonate sediments (Hine, 1977; Harris, 1979). Generally, where the shallow shelf is open to periodic off-bank flux of hypersaline or brackish water, reefs are uncommon (or are "default"), and carbonate sand shoals occur instead; where there are shelf-margin islands, inimical off-bank water-flux is inhibited, and reef growth encouraged (Harris, 1979). Although there are indeed exceptions, and other factors controlling this general distribution, the facies association on the southwest Florida carbonate platform resembles and may mimic the distribution of facies in the Cape Storm Formation. Clearly, the extensive peritidal facies of the Cape Storm Formation suggests that brackish or hypersaline platform water could have inhibited shelf-margin reefal growth. The top of the sand shoal deposit is coincident with the base of the Douro Formation and a return to open marine, deeper water deposition of fossiliferous lime mudstone.

At Bay Fiord (sections 5, 9, 12, Fig. 8b), the Upper Ordovician to Ludlow shelf-margin facies is exposed on the hanging wall of a west-verging thrust, in a large mountain exposure. The succession is predominantly biohermal, but is interrupted by a thin, lowest Silurian shale tongue. Above the shale, a progradational carbonate platform-margin facies is represented by bedded olistostromes that are overlain by massive, dolomitized biohermal bindstone and rudstone (Fig. 8b; Plate 1b). The biohermal facies, although largely obscured by dolomitization (Plate 1e), contains recognizable laminar stromatoporoid bindstone. Olistoliths of similar lithology occur in the succeeding slope facies, although younger in age.

The contact of the shales and biohermal carbonates is abrupt, occurs at the base of a 2m thick olistostrome bed, and is lower or middle Wenlock in age. An overlying succession of interbedded resedimented carbonates, calcareous mudshale, clayshale, and marlstone is capped by an upper Pridoli encrinite channel-fill sequence (Plate 8f). The channel fills are lensoid, 200-300m in apparent width, typically amalgamated, isolated, or in complexes, and have a basal conglomeratic unit. These distinctive encrinites are interbedded with graptolitic shales and overlain by rubbly weathering carbonates, assignable to the Goose Fiord Formation.

The apparent *cyphus* Zone platform backstepping near Bay Fiord is problematical. If this was a backstepping event, it should be represented on a more regional scale (*sensu*, Schlager, 1981; Sarg, 1988), but is not. Platform backstepping typically results in condensed sequences and regional termination of carbonate supply (Sarg, 1988). The presence of thick conglomerate beds near Bay Fiord at the biohermal carbonate-shale contact and the local backstepping is inconsistent with the characteristics of platform drowning. Furthermore, as described below, the very condensed Llandovery succession of section 3 (Fig. 7a) is anomalous, yet occurs only 1km west of the Bay Fiord biohermal shelf-margin facies. Truncation and slump structures are abundant in the Llandovery part of the deep-shelf succession in section 3, and mass wasting possibly caused sequence condensation. Similarly, the abrupt shale-biohermal carbonate contact possibly represents mass wasting on a larger scale, due to large-scale platform margin collapse. This type of mass wasting appears to be intrinsically related to platform upbuilding and instability related to the juxtaposing of competent, early lithified carbonates and relatively unlithifed toe-of-slope marls and terrigenous muds. Early marine cements are abundant in olistoliths in the study area, suggesting very early lithification of the biohermal carbonates. Moreover, mass wasting at this scale is an exceedingly common phenomenon in ancient and modern carbonate platforms (for example, Playford *et al.*, 1989; Sarg, 1988; Playford and Cockbain, 1989).

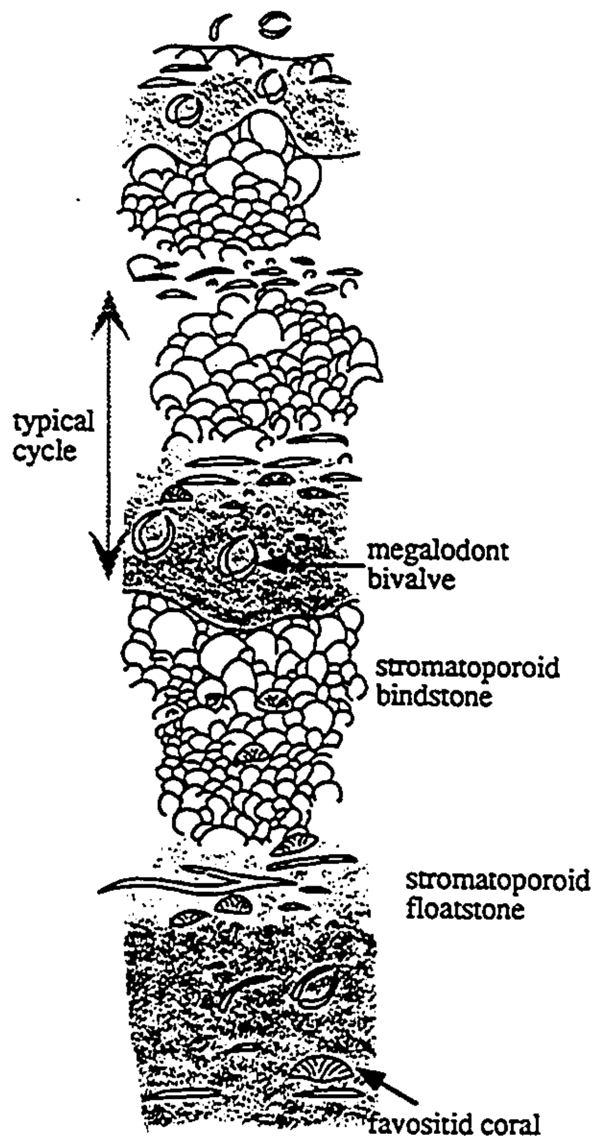
West of Vendom Fiord, another shelf-margin sequence was examined on a reconnaissance scale (in the vicinity of section 70; Fig. 3). Two units are present in this area: a lower unit approximately 1.5km thick (photogrammetrically measured), is intensely dolomitized, and represented mostly by frost-riven rubble. Sporadic outcrops display a variety of lithologies, including megalodont floatstone, pentamerid rudstone and floatstone, stromatoporoid bindstone, and mottled dolostone. An upper unit, 130-150m thick, is a fossiliferous, rubbly-weathering limestone containing poorly exposed stromatoporoid buildups. Coarse olistostromes are associated with these and other buildups. Similar reefal bodies occur in the underlying massive dolostone unit, but are not as well exposed. Biostratigraphic information is sparse, but a single conodont collection from the top of the rubbly limestones yielded *siluricus* Zone conodonts (Uyeno, pers. com., 1990), suggesting either Cape Storm or Douro Formation equivalence; however, the presence of sparse atrypid brachipods in rubbly weathering limestone suggests assignment to the Douro Formation.

Megalodont bivalves are abundant below the rubbly weathering Douro-like unit and probably occurred in a very similar depositional environment to that inferred for the Cape Storm Formation on east-central Cornwallis Island. There, megalodonts associated with skeletal grainstone and patch reefs (as described above) indicate a protected shelf-edge environment.

Stromatoporoid platform foreslope reef facies

Strata on Hans Island are gently dipping and well preserved; they were spared much of the intense deformation and dolomitization that affected much of the strata in northeastern Ellesmere Island. When this area was visited briefly in April of 1988, the accessible, gently dipping beds on the southeast side of the island were snow-covered, but an exposed section was measured on the precipitous east-northeast side of the island. There, the limestone succession is cyclical, and a typical cycle is characterized by (i) basal megalodont floatstone; (ii) laminar stromatoporoid bindstone to floatstone; and (iii) globular and subhemispherical stromatoporoid bindstone and rudstone. The upper part (iii) is abruptly overlain by dark-brown megalodont rudstone. Local relief, up to 3m, on the flooded surface of the stromatoporoid bindstone probably represents incipient biohermal development (Fig. 17). The approximately 143m thick sequence exposed on the island represents upward shallowing overall, with stromatoporoid bindstone and floatstone predominant toward the summit of the island. The strata on Hans Island are an

Figure 17: Main phases of reef growth in Hans Island cyclic deposits. These strata in the southeast cliff exposure on the island are well preserved, but fossil identifications are only general. Of note are the megalodont bivalves, which generally occur exclusive of other macrofossils, the relief on the biostromal deposits prior to burial by lime muds, and the regular, repetetive succession of macrofossils.


typical
cycle
megalodont
bivalve
stromatoporoid
bindstone
stromatoporoid
floatstone
favositid coral

erosional remnant of a presumably larger, more complex structure (de Freitas, 1990a), that could have been a pinnacle reef. Similar intrashelf reefs (Hurst, 1980), established several 10's of kilometres basinward of Silurian shelf carbonates (Surlyk and Hurst, 1984) are exposed in Hall Land, North Greenland. Coeval foreslope bioherms are known along the coast of Hall Land, where they similarly form small channel islands (Hurst, 1980).

Megalodonts bivalves, in this study, appear to be characteristic of muddy, protected shelf-margin environments. The occurrence of laterally continuous megalodont-rich beds in the Hans Island succession, some grading into stromatoporoid bindstone or rudstone, suggests an extensive protected reef-associated environment, perhaps on an atoll-like structure that had a lime mud and bivalve-rich central lagoon. Although the exposure is inadequate to confirm this, the carbonate body is far removed from the main contiguous shelf-margin carbonate complex exposed on North Greenland, indicating that it was indeed isolated and perhaps pinnacle-like during deposition.

Olistostromes of middle to late Llandovery age that occur in sections on eastern Ellesmere Island (Sections 23, 24, 25, 26; Fig. 7b) undoubtedly indicate additional coeval biohermal development over the drowned shelf (Plate 9a,b), although the buildups themselves are not exposed.

The age of the Hans Island structure is uncertain. Samples collected yielded non-diagnostic conodonts, and most macrofossils, mainly stromatoporoids, are long-ranging and not biostratigraphically useful; however, megalodont bivalves are of some use biostratigraphically, as they occur abundantly in biohermal and biostromal strata of Ludlow (and Pridoli) age in the Canadian Arctic and north Australia. Bivalves occur in the Cape Storm Formation and less commonly in Pridoli platform carbonate formations (Goose Fiord and Barlow Inlet formations). Bivalves are uncommon in the Wenlock and exceedingly rare in the Llandovery. A Pridoli age for the Hans Island reef structure is unlikely, because Pridoli clastic deposits are prevalent in North Greenland and succeed extensive "flysch" deposits, which terminated the carbonate platform in North Greenland in latest Llandovery (Hurst and Surlyk, 1984). Also, Wenlock and Ludlow carbonate deposition in North Greenland was restricted to the tops of small reef pinnacles, resembling the Hans Island structure, and none of these structures is younger than Ludlow age (Sønderholm *et al.*, 1987). These data suggest that the buildup is probably Ludlow in age and thus is coeval with the wg5 pinnacle reef described by Sønderholm and others (1987) (Fig. 6a-c).

CAPE STORM FORMATION (Scs)

Distribution and lithology

The Cape Storm Formation is widely recognized through the Arctic Islands and was first described on southern Ellesmere Island by Kerr (1975). On Cornwallis Island and Devon Island, it is generally characterized by thinly to medium bedded, recessive weathering, cliff-forming, yellow-brown to grey dolostone and silty dolostone. It lies, in most areas, disconformably on the Allen Bay Formation. The lower member, described by Kerr (1975) and Morrow and Kerr (1977), is currently included in the underlying Allen Bay Formation (Thorsteinsson and Mayr, 1987; Mayr *et al.*, in prep.). In this study, the Cape Storm Formation was examined on Grinnell and Darling peninsulas and near Baumann Fiord. Thicknesses of the member are given in Table 3.

In section 31 (Darling Peninsula, Fig. 9c) unit 6 is lithologically similar to the Cape Storm Formation as established farther to the south (Fig. 6a-c). Conodonts from the lower part of the unit indicate a Middle to Late Silurian age; however, abundant atrypoid brachiopods in the overlying rubbly weathering fossiliferous units indicate a lower to middle Ludlow age, suggesting that unit 6 is assignable to the Cape Storm Formation. Although its lower contact was largely snow-covered and poorly exposed, unit 6 (Cape Storm Formation, Fig. 9c) apparently rests conformably on thickly bedded subtidal limestone and dolostone of the Allen Bay Formation (unit

TABLE 3: Thicknesses of the Cape Storm Formation in the report area (section locations given in Fig.1)

Section	Thickness
31(unit 6)	160m
6-10	587
61	258
62	220
Cornwallis I.[1]	520-540
Somerset I.[4]	120-260
Prince of Wales I.[6]	52-96
SW.Ellesmere I.[3]	230

TABLE 4: Thicknesses of the Douro Formation in the report area (section locations given in Figs.1 and 3)

Section	Thickness
31(unit 7b)	160m
61	121
62	157
70	50-165
Somerset I.[8]	209-275
Griffith I.[5]	460
Cornwallis I.[1]	340
Sutherland R.[7]	140
S.Ellesmere I.[3]	100
Prince of Wales I.[6]	101-241

TABLE 5: Thicknesses of the "Devon Island Formation " in the report area (section locations given in Figs.1 and 3)

Section	Thickness
59	296m
60	160
69[9]	682
Douro Range[1]	109
Sutherland R.[7]	710
S.Ellesmere I.[3]	116-320

[1] Thorsteinsson and Uyeno, 1980
[2] Mayr *et al.*, in prep.
[3] DNAG Chart (GSC, in prep.)
[4] Miall and Kerr, 1977,1980
[5] Narbonne, 1981
[6] Mortensen and Jones, 1986
[7] Thorsteinsson and Mayr, 1987
[8] Stewart, 1987; Savelle, 1978; Narbonne, 1981
[9] this section may have a component of tectonic thickening
[10] very poorly exposed

TABLE 6: Thicknesses of the Starfish Bay shale (informal name) in the report area (section locations given in Figs.1 and 3)

Section	Thickness
43	~60m
59	69
60	38
79	<5
80[10]	<5

TABLE 7: Thicknesses of the undivided Cape Phillips Formation in the report area (section locations given in Fig.1)

Section	Thickness
5m	140m
9	118
12	108
13	179
19	1420
23	93
26	144
29	26
30	219
32	102
35	288
38	142
40	80
43	632
60	122

5, and underlying units). The equivalent contact is disconformable farther south. Unit 6 (Fig. 9c) is 103m thick and mainly parallel laminated light brown dolostone, containing synaeresis cracks, wavy and flat microbial laminites, uncommon flaser and wavy bedding, and many intercalated beds of burrowed calcareous dolostone and limestone that typically weather dark brown. The unit is variably silty, pale brown weathering, recessive, and contains abundant *Leperditia* valves and very few intraclast beds. The unit is gradationally overlain by a rubbly weathering atrypoid brachiopod-rich unit that is probably equivalent to the Douro Formation.

In section 61 near Makinson Inlet (Fig. 9d), the lower contact with the Allen Bay Formation is poorly exposed and marked by a rather abrupt change from thickly bedded dolostone to less resistant, medium bedded dolostone. The contact also marks an abrupt change in depositional environments, from predominantly subtidal below to restricted and intertidal above. The Cape Storm Formation contains two distinctive divisions (section 61, Fig. 9d): an upper unit A consists of laminated silty dolostone, dolomitic siltstone, with conspicuous silicified synaeresis cracks (Plate 3c,f) and thick intervals of patterned dolostone; and a lower unit B consists predominantly of medium bedded, horizontally laminated, pale brown (Plate 3d), and dark brown dolostone, with rarely bioturbated interbeds. Unit B contains great thicknesses of the laminated dolostone, with laminae that are continuous, parallel, flat, wavy (Plate 3d), flaser, crinkled, or fenestral. Some flat laminae rest on scoured surfaces and grade imperceptibly upward into large domal stromatolites, suggesting that some of the flat laminae have a microbial origin. Interbedded brown and dark brown dolostone are similarly laminated. Stromatolitic units are common approximately 30m above the base, and consist of randomly alternating, laterally linked hemispherical, wavy, and horizontal laminae (microbialites). Some of the domal stromatolites have diameters in excess of 2m (Plate 3a). Bimodal cross lamination in flaser bedding occurs with uncommon bioturbated interbeds having an ichnofabric index (I.I.) of 1-2 (scale after Droser and Bottjer, 1986; Plate 3b). Also, some wavy laminated units occur. Some distinctively striped beds (Plate 3d) consist of interlaminated, light coloured, cross-stratified dolostone and dark, bioturbated and laminated dolostone. Although pervasively recrystallized and dolomitized, some laminae appear to be peloidal, and some are graded.

Unit A (Fig. 9d) in section 61, contains thick intervals of patterned dolostone (Plate 3e) interbedded with microbialite dolostone, and horizontally to cross-laminated dolostone with silicified synaeresis-cracks (Plate 3c,f). Vertical burrows, some up to a meter in length, and leperditid ostracods, are also associated. Other rock types include the following: pale, yellow-brown, laminated, silty dolostones (*very* similar to the Cape Storm lithology in the southern Arctic); gastropod-rich rudstone associated with discontinuously, thickly laminated dolomitic limestone; and less common, recessive, thinly bedded, blue-grey, laminated dolostone (beds about 20cm thick). Small (1.5m high by 2.0m wide) coral-spongiostromate-sponge mounds occur in beds transitional with the overlying Douro Formation, and these strata generally represent deposition in subtidal, more open marine conditions than represented by most of the Cape Storm Formation. The base of the Douro Formation is arbitrarily chosen at the base of a continuous succession of atrypoid-bearing, rubbly weathering limestone to be consistent with the contact established elsewhere in the Canadian Arctic.

In **section 62** (Makinson Inlet, Fig. 9c), the Cape Storm Formation is recessive, medium bedded, pale brown dolostone. The basal contact is sharp, disconformable, and is beneath a thin orange-brown weathering dolomitic siltstone, that rests on irregularly weathered and pitted massive dolomitic limestone of the upper limestone member of the Allen Bay Formation. Beds immediately above the orange-brown siltstone are regularly interbedded light brown, laminated, and less common, dark brown, bioturbated dolostone. As in section 61, several types of lamination are recognized, including flaser or wavy, and flat parallel lamination, and crinkled microbial lamination.

Lower units of the formation in section 62 (Fig. 9c) are commonly pale brown to yellow brown weathering and in marked contrast to the underlying and overlying darker carbonates of the Allen Bay and Douro formations, respectively. Domal stromatolites in the lower portions of the formation are less common than in coeval units at Swinnerton Peninsula (section 61, Fig.9d). Distinctive patterned carbonate, common 75m above the base of the formation, is interbedded with recessive shaly, thin-bedded, blue-grey dolostone (as recorded in unit B, section 61) and brown, recessive, bioturbated dolostone. The top 40m of the formation in section 62 are predominantly dark brown bioturbated dolostone, with amalgamated intraclast beds locally containing clasts encrusted with laminar favositids. Some bioturbated beds (ichnofabric index (I.I.) of 2-3) and uncommon interbedded limestone units contain abundant high-spired gastropods. These upper units are in marked contrast to equivalent units in section 61, which are predominantly synaeresis-cracked dolostone that have few bioturbated interbeds.

The intraclast beds are 1 to 15cm thick and contain rounded to well rounded or thin, fractured lithoclasts. Some clasts are isolated in fenestral carbonate. Cyclic "tidalites" (Hardie, 1977; Ginsburg *et al.*, 1977) are uncommon, but where present are characterized by basal dark brown, bioturbated dolostone overlain by a light brown laminated dolostone.

In central Devon Island, a complete 587m thick section of the Cape Storm Formation was examined on a reconnaissance scale. These rock units are substantially more silty than coeval exposures on Ellesmere Island. The lower part of the formation consists of laminated silty dolostone with ostracods and uncommon synaeresis cracks, and patterned dolostone. The upper part contains a distinctive rock type: 1-5cm thick, locally ostracod- and gastropod-rich, grey-brown, continuous or discontinuous, limestone beds are interbedded with about the equivalent thickness of light yellow-brown laminated silty dolostone and dolomitic siltstone. Some laminae are disrupted by bioturbation (I.I. 2-3), giving these rocks a distinctive, almost reticulate pattern in outcrop (*sensu* Plate 3b of coeval strata on Ellesmere Island). These units are also present, but in lesser amounts, in the upper portions of sections 61 and 62. The uppermost units of the Cape Storm Formation, in all examined exposures are fossiliferous, biostromal, and biohermal.

On Devon Island, the basal Cape Storm Formation of **sections 6-4 and 6-7** (Fig. 9a) was examined to decipher age relationships and the nature of the contact of the Allen Bay and Cape Storm formations. This contact is distinct in cliff exposures, and marked by a change in topography, from the precipitous cliff exposures of the Allen Bay Formation to more recessive, ledge-forming exposures of Cape Storm Formation. The base of the Cape Storm Formation is also considerably more silty than the underlying Allen Bay Formation. Without these characteristics, the two formations would be difficult to distinguish, as both were deposited on extensive carbonate tidal flats.

In **section 6-4** (Fig. 9a), cross-laminated, horizontally laminated, and microbially laminated dolostone are common in the lower part of the Cape Storm Formation. Other notable structures include synaeresis cracks, large, laterally linked hemispherical stromatolites, leperditid ostracods, allochthonous orthoconic cephalopods, and bioturbated calcareous "reticulate" (resembling mottling as observed in coeval strata on Ellesmere Island, Plate 3b) dolostone, as present in other Cape Storm exposures in the study area. Synaeresis cracks are more common in penultimate units, whereas coral-stromatoporoid-spongiostromate-sponge biostromes occur in uppermost units.

Distinctive Cape Storm shelf-margin strata of Cornwallis Island, examined in the summer of 1985 (de Freitas, 1987a,b), is characterized by medium to very thickly bedded, coarsely crystalline, laminated, vuggy dolostone, and is associated with massive biohermal facies. Conspicuous megalodont bivalves are the only recognizable macrofossils in an otherwise intensely dolomitized, faintly cross-bedded calcarenite that is interbedded with coral-stromatoporoid framestone, coarse floatstone, and rudstone. These beds abruptly (possibly disconformably) overlie massive biohermal facies of the Allen Bay Formation. No diagnostic conodonts were present in the processed samples, but the detailed biostratigraphic information for this area is given by Thorsteinsson and Uyeno (1980). The shelf margin strata grade platformward into strata more typical of the Cape Storm Formation, although exposure of this transition is poor.

Age of the Cape Storm Formation and contact with the underlying Allen Bay Formation.

The biostratigraphy of the Cape Storm Formation has been determined by extensive study of materia collected over several years from crucial localities. Detailed work of Thorsteinsson and Uyeno (1980) have demonstrated that the the Cape Storm and Douro formations consist of almost 1000m of strata that fall entirely within the middle Ludlow *siluricus* Zone. Also, the *linearis* graptolite zone, based on correlation across the shelf-margin facies change in east-central Cornwallis Island, lies in the middle of the Cape Storm Formation according to Thorsteinsson and Uyeno (1980). However, Thorsteinsson used an essentially *flat* line for correlating the graptolitic beds to the platform carbonates. Detailed mapping of the shelf margin facies (de Freitas, 1987a; de Freitas, *et al.*, 1989) delineated up to 200m of platform margin relief in the area of central Cornwallis Island. The *linearis* Zone is consequently higher in the succession than previously thought, and probably occurs in the upper and not the middle part of the formation. Although this appears to be a relatively minor discrepancy, the position of the *linearis* Zone in the platform succession is important for interpreting platform margin evolution and in correlating the Starfish Bay shale, discussed below. Moreover, given the thickness of strata represented by the *siluricus* Zone and the unusually high rate of carbonate deposition (some 225m/Ma, Fig. 10a,b), represented by these two formations, it suggests that either the Ludlow was characterized by an *unusually* high rate of subsidence and carbonate production, or that the *siluricus* Zone represents a considerably longer period of time than suggested by the biostratigraphic charts of this and other studies Thorsteinsson and Uyeno (1980), and by Kleffner (1990). Although a comparable rate of sedimentation is known from some modern carbonate settings, the studied strata are compacted and represent settings that would have been optimum for carbonate production, such as tidal flats (Scs) and deep carbonate ramp envrironments (Sdo). In light of these relationships, the time represented by this zone and/or the positions of its boundaries relative to the conodont zonal scheme (for example of Uyeno, 1989) should be re-evaluated.

Several sections of the Cape Storm Formation in this study yielded conodonts of Ludlow aspect (for example, on Grinnell Peninsula), but none of definite *siluricus* Zone age. However, well established ages on southern Ellesmere and Devon islands (Thorsteinsson and Mayr, 1987; Mayr *et al.*, in prep.) show that the formation there rests entirely within the *siluricus* Zone. Samples were also collected on Ellesmere Island, but these have not yet been processed. Conodont samples identified (see Appendix two) indicate that the lower boundary represents a significant time gap, and that some of the Wenlock and the lower Ludlow is missing due to erosion. This relationship has been firmly established on southern Ellesmere Island as well by Mayr, *et al.* (in prep.).

Interpretation and discussion

The Cape Storm Formation was deposited on extensive tidal flats. The strata, in contrast to the underlying Allen Bay Formation, are chiefly laminated dolostone, showing minor regional lithological variations. However, silt content varies considerably, decreasing to the north, and is minor in the vicinity of Makinson Inlet and Darling Peninsula. The silt is mainly angular and rounded quartz and less commonly feldspar (including plagioclase) grains that were concentrated probably by currents and winds on flats to form silt-rich beds in a predominantly carbonate-rich environment. Siltstone beds are generally cross-laminated, suggesting current or possibly wind transport. On modern arid flats, migrating sand dunes leave little evidence of migration across the sabkha (Shinn, 1986), and are represented only by subtidal concentrations of quartz. Similarly, much of the silt in the Ludlow (and possibly younger) deep-shelf succession could have been wind-transported. The lack of silt or sand in the Cape Storm shelf margin facies on east-central Cornwallis Island, suggests winnowing and possibly off-platform transport of terrigenous detritus. In a similar ancient platform margin setting in the Permian of the southern USA (Harris, *et al.*, 1990), clastic detritus transported off-bank was deposited in the coeval basin margin (Harris, *et al.*, 1990). A derivation from a northerly source in the Arctic (Pearya) is unlikely, considering that (i) the silt content increases southward, (ii) across-basin transport of silt by sediment-laden plumes would probably have considerably limited shelf-margin reef growth, and (iii) the only potential northern source for the silt, Pearya, is now considered to be an allochthonous terrain that apparently docked in latest Silurian (Trettin, 1987). According to current paleogeographic reconstructions the north-to-south increase in silt could be explained by a penecontemporaneous easterly (trade wind) circulation across a sparsely vegetated, dry cratonic land mass.

Tidal flat laminites of silt, pellets, bioclasts, and lithoclasts probably represent rapid fall-out of possibly lagoon-derived sediment transported in suspension during storms (Wanless, *et al.*, 1981, 1988). However, a coeval lagoonal facies of the Cape Storm Formation has not been identified. There are, in some localities, interbedded bioturbated lime mudstone units that could have ultimately been a source for the pelleted material in laminites, as Wanless and others (1988) have concluded for a recent Bahamian example, but much of this material likely had a platform margin source, particularly since many of the laminated beds appear to be skeletal, or, at least, non-peloidal. Transported calcisilt was either trapped by cyanobacterial mats or deposited as normally or inversely graded laminae on tidal flats. Crinkled laminae occur with flat laminae and indicate growth of clumped microbial colonies (Monty, 1976). Uncommon fenestral fabrics, in part, may have resulted from trapped gas or uneven growth of microbes, and are generally indicative of prolonged exposure (Ginsburg *et al.*, 1977; Hardie, 1977; Shinn, 1983a,b; Tucker and Wright, 1990). The presence of domal stromatolites suggests intertidal deposition in a mean high-water depth at least as deep as the height of domes (Logan, 1974), and bioturbated, gastropod-rich dolostone indicates deposition perhaps in tidal flat ponds.

Toward the top of the Cape Storm Formation, strata are more varied, and subtidal beds alternate with silty dolostone, patterned carbonate, and abundant synaeresis-cracked dolostone. Patterned dolostone is related to evaporite formation (Kendall, 1977a,b; Dixon, 1976) and replacement, but the great abundance of synaeresis cracked beds is unusual. These occur in intertidal strata that appear to have a large proportion of subtidal intercalations, and it is questionable whether these cracks were formed at the sediment water interface or below. Grotzinger (1986), in his model for "Milankovitch-band" tidalite cyclicity, interpreted synaeresis cracks as substratal structures. Substratal synaeresis has been invoked for platform slope mudcracks (de Freitas and Dixon, 1990), for mudcracks on the bases of load structures (Plumber and Gostin, 1981), and for cracks related to concretions (Calver and Baillie, 1990). Burst (1965) clearly demonstrated that synaeresis-like cracks can develop at the sediment-water interface where hydrophillic clays are present and salinity variations in the overlying water column are great; however, Dangeard *et al.*, (1964) were able to demonstrate that substratal mudcracks could form through compaction. Clearly, there is more than one possible origin of these spindle-shaped cracks, and their occurrence does not necessarily indicate excessive salinity variations or exposure. Moreover, polygonal mudcracks can be formed purely by compaction. Although their occurrence is *generally* attributed to subaqueous dewatering on restricted tidal flats, their presence is equivocal evidence for a peritidal environment, and should only be used to infer environement in association with other peritidal indicators.

The source of silica for preferential silicification of the synaeresis cracks is unclear. Processes of silicification are complex, and its origin appears to be related to mobilization of usually biogenic and less commonly terrestrially derived silica (Maliva and Siever, 1988, 1989; Knauth, 1979). Neither source can be unequivocally demonstrated, but the occurrence of silica at this platform locality appears to have involved only minimal planktonic or benthic-derived silica, and perhaps suggests formation in mixed-meteoric pore waters (*sensu* mechanism of Knauth, 1979).

DOURO FORMATION (Sdo)

This formation was originally described by Thorsteinsson (*In* Fortier *et al.*, 1963) for a rubbly limestone succession in the Douro Range of northern Devon Island. The formation is distinct and widely recognized in the Canadian Arctic Archipelago; its uniform lithology and wide distribution is surpassed only by the Irene Bay Formation. The rubbly weathering limestones have been a focus of study by many workers (Jones and Dixon, 1977; Morrow and Kerr, 1977; Mortenssen and Jones 1986; Narbonne, 1981; Narbonne and Dixon, 1982, 1984, 1989; Savelle, 1978; Stewart, 1987; Thomas and Narbonne, 1979; Thorsteinsson and Mayr, 1987; Thorsteinsson and Uyeno, 1980), perhaps primarily because the formation is one of the most fossiliferous in the Arctic. Thicknesses of the formation are given in Table 4.

Lithology and distribution

On Cornwallis and Devon islands the upper contact, essentially corresponds with the boundary of the *siluricus* and *latialata* conodont Zones (Thorsteinsson and Uyeno, 1980), and is overlain either by encrinite grainstone (base of the Barlow Inlet

Formation, on southwestern Devon Island) or by clastics (lower member of the Barlow Inlet Formation, on eastern Cornwallis and western Devon islands). Its basal contact is gradational and is placed in most areas at the base of a continuous succession of rubbly limestone beds, typical of the Douro Formation. In many areas, atrypoid brachiopods are a conspicuous and abundant component of the formation. Biostromes and mounds occur in the lower boundary sequence several metres thick. Sponge mounds are abundant in two separate intervals on Somerset Island and represent sub-wave base structures, established during two periods of maximum deepening on the Douro platform (Narbonne and Dixon, 1984).

The Douro Formation was briefly examined in only three sections in the study area. In **section 31** (Fig. 9c), unit 7b is likely coeval with the Douro Formation. The base is marked by the appearance of atrypoid brachiopod-rich, rubbly weathering limestone that succeeds an otherwise peritidal Cape Storm succession. Identified atrypoids indicate a lower to middle Ludlow age (J. Jin pers. com., 1990), and a conodont sample 100m below the basal contact yielded wide-ranging Wenlock to Pridoli elements, suggesting that the lower contact may be equivalent to the Cape Storm-Douro contact established in the south. Of the 360m of rubbly weathering strata exposed at this locality, 160m are likely assignable to the Douro Formation. Unit 7b has two distinct lithologies, a lower (83m) recessive, argillaceous and fossiliferous, rubbly weathering limestone and an upper (77m) thickly bedded, sparsely fossiliferous, mottled, dolomitic limestone. The latter unit is similar to the upper unit of the Douro Formation at Makinson Inlet (section 61, (Fig. 9d), and the occurrence of Pridoli corals above the formation at both localities supports this correlation.

In **section 61** (Fig. 9d), 121.2m of carbonates are of typical Douro Formation aspect: they are rubbly weathering, rich in atrypoid brachiopods, and very fossiliferous. The formation in this area is divided into two units similar to those on Darling Peninsula. The lower unit is a typical rubbly weathering atrypoid brachiopod-rich argillaceous limestone, which contains several basal sponge-algal(?) mounds some 3m in height and width. The mud mounds are massive and grade laterally into bedded rubbly weathering strata. They are very small structures that probably had little relief during growth. Sponges are found rarely in the off-mound strata, and beds above and below have abundant solenoporid algae, stromatolites, stromatoporoids, and crinoid debris. A synaeresis-cracked bed above one of the mounds does not necessarily indicate a restricted depositional environment, but may indicate formation through interstratal synaeresis. Also, the paucity of intraclast beds is noteworthy and the apparently cyclic alternation of more resistant, less argillaceous, bioturbated (I.I. 4) units and recessive, bioturbated, argillaceous, fossiliferous units on a 5-10m scale.

The upper unit of the Douro Formation in section 61 is thickly bedded, mottled (I.I. 4-5), dolomitic limestone that contains slightly argillaceous, less resistant intervals. Mottling is conspicuous and very reminiscent of mottling in the Arctic

Ordovician Thumb Mountain Formation. The upper unit is not as fossiliferous as the underlying unit; brachiopods are rare, but orthoconic cephalopods and solitary rugose corals are relatively common. The contact with the overlying Goose Fiord Formation at this locality is marked by a gradual change from thickly bedded, medium brown, mottled dolomitic limestones to recessive, pale-brown weathering, slightly silty dolostone. This contact is conformable. Samples were collected for conodont study, but identifications are not yet available.

In section 62 (Fig. 9c), 157.1m of strata are assignable to the Douro Formation. The basal beds are atrypoid brachiopod-rich and contain a 2m thick *Thamnopora-Syringopora* rudstone bed. Above these basal units, the lower member of the Douro Formation displays irregularly alternating resistant mottled limestone and recessive rubbly limestone. These cyclical patterns are ascribed to variations in argillaceous content. Atrypoid brachiopods are abundant in most beds, particularly in rubbly weathering beds. A distinctive disarticulate megalodont bivalve rudstone bed occurs near the base of the formation, immediately above large trough cross beds, suggesting storm influence on these strata. Paleocurrent measurements indicate north-to-south transport for the cross-bedded calcarenites. The upper unit of the Douro Formation is not well exposed. As in section 61, it is characterized by mottled dolomitic limestone, and uncommon atrypoid brachiopods, orthoconic cephalopods, and mucophyllid, syringoporid and favositid corals. Low domical heliolitid corals, some silicified, occur throughout, and are particularly common in the base of the overlying Goose Fiord Formation. The upper contact with the Goose Fiord Formation is chosen at a silicified interval containing large, white chert nodules, and abundant well-preserved *Encrinurus* trilobites, but the age of the boundary strata is unknown, and biostratigraphic information is not yet established.

Interpretation and discussion

Most workers have interpreted the formation as having been deposited below fair-weather wave base on a carbonate ramp. The ubiquitous rubbly limestone texture, in particular, has been ascribed to burrowing and early diagenetic concretionary cementation and perhaps also to later diagenetic unmixing of the strata (Ricken, 1986; Jones *et al.*, 1979; Möller and Kvingan, 1988; Mullins *et al.*, 1980). The formation represents a major change in the distribution of platform facies, from well differentiated shelf margin and platform interior peritidal dolostone facies of the Cape Storm Formation to a relatively deeper water, undifferentiated shelf limestone facies of the Douro Formation. However, there are several noteworthy lithological variations in the formation recognized in this study.

The upper part of the Douro Formation is an unusually clean, relatively, unfossiliferous, mottled, dolomitic limestone on Somerset Island (Thorsteinsson and Uyeno, 1980) and on Prince of Wales Island (Mortensen and Jones, 1986). This facies occurs on Ellesmere Island, at Baumann Fiord and Darling Peninsula, and was interpreted by the latter authors (*op cit.*, 1986, p.1403) as deposited in shallower water than the associated, rubbly weathering Douro Limestone. On Darling Peninsula, the resistant mottled dolomitic limestone separates rubbly weathering limestones of the overlying Goose Fiord Formation and underlying lower member of the Douro Formation (Fig. 9c). Narbonne and Dixon (1982) suggested that the variability in clay content was ascribed to distribution patterns rather than to bathymetry, and proposed a northerly source for the terrigenous material; however, in more northeasterly areas, clay content diminishes. Based on the effect of the Boothia Uplift on Douro deposition described by Mortensen and Jones (1986) for strata in the vicinity of Prince of Wales Island, the upper mottled dolomitic limestone member of the Douro Formation on Ellesmere Island may be similarly ascribed to an early phase of the Inglefield Uplift and shoaling in this area. Alternatively, regional variation in currents may explain the general lack of clays in the Ellesmere Island localities.

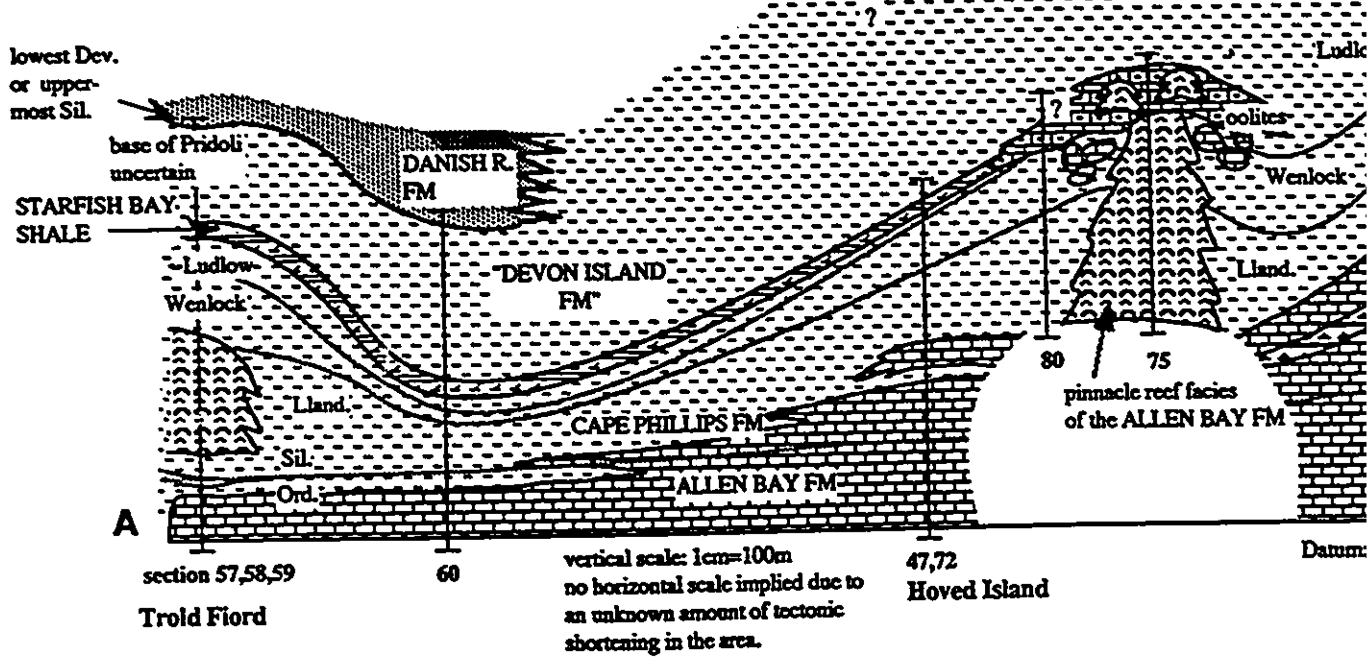
Although the origin of the clay content is inconclusive, the Douro Formation was clearly deposited in a relatively deep, turbid environment, in conditions less than optimal for calcium carbonate production; yet, this formation, based on current biostratigraphy and relevant chronostratigraphy (Kleffner, 1990), together with the Cape Storm Formation (Fig. 10), represents the greatest rate of carbonate accumulation during the Silurian for the Arctic. Clearly, this rate is atypical, particularly when compared with other Silurian formations with very similar inferred depositional environments (for example, middle member of the Allen Bay Formation,

Fig. 10).

A rubbly weathering limestone unit in section 70a, west of Vendom Fiord, with stromatoporoid framestone reefs and associated olistostrome beds has yielded *siluricus* Zone conodonts, strongly suggesting a correlation with the Douro Formation in the south. However, the *siluricus* Zone is apparently long-ranging, as inferred above, and is difficult to correlate accurately with basinal graptolite biozones. Based on the stratigraphic relations established at Baumann Fiord, it appears that the *linearis* Zone is correlative with part of the *siluricus* Zone (Figs. 18, 19), as also concluded by Thorsteinsson and Uyeno (1980). This conclusion is based on the presumption that platform backstepping in this area was synchronous, resulting in areal reduction in carbonate production and transport of allochthonous sediment to the adjacent toe of slope. Abrupt termination of allochthonous slope deposits in the *linearis* Zone and the onlapping of shales on platform carbonates (sections 79, 80; Figs. 7c and 18b) suggest that carbonate supply was terminated due to platform drowning. Based on current biostratigraphic imformation, backstepping was apparently synchronous. This conclusion implies either (i) that the *linearis* Zone is younger than previously thought, and perhaps, in part, correlative with the *latialata* Zone, or more probably (ii) that the rubbly limestone unit is older, suggesting diachronous north-to-south regional backstepping (Fig. 19). The implications of these points are discussed more fully below.

Figure 18a-b: Diagrammatic restored section of the main platform, deep-shelf, and trough stratigraphic units near Trold and Vendom fiords. Legend as Figure 6d. See Figure 3 for section locations. Note that the two pages form a continuous west-to-east section.


100km
B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section 57,58,59
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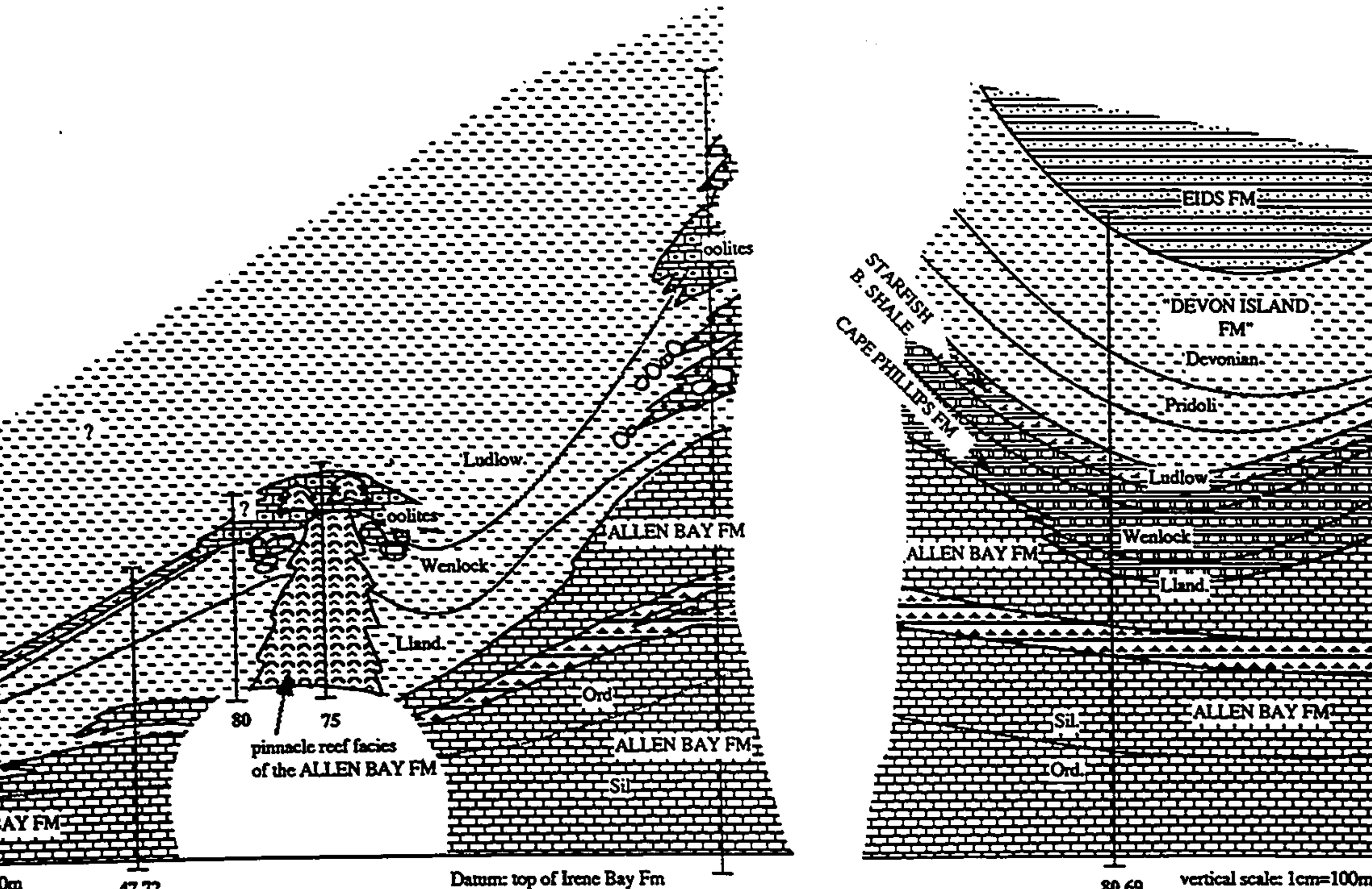
vertical scale: 1cm=100m
no horizontal scale implied due to
an unknown amount of tectonic
shortening in the area.
47,72
Hoved Island
Datum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Devonian
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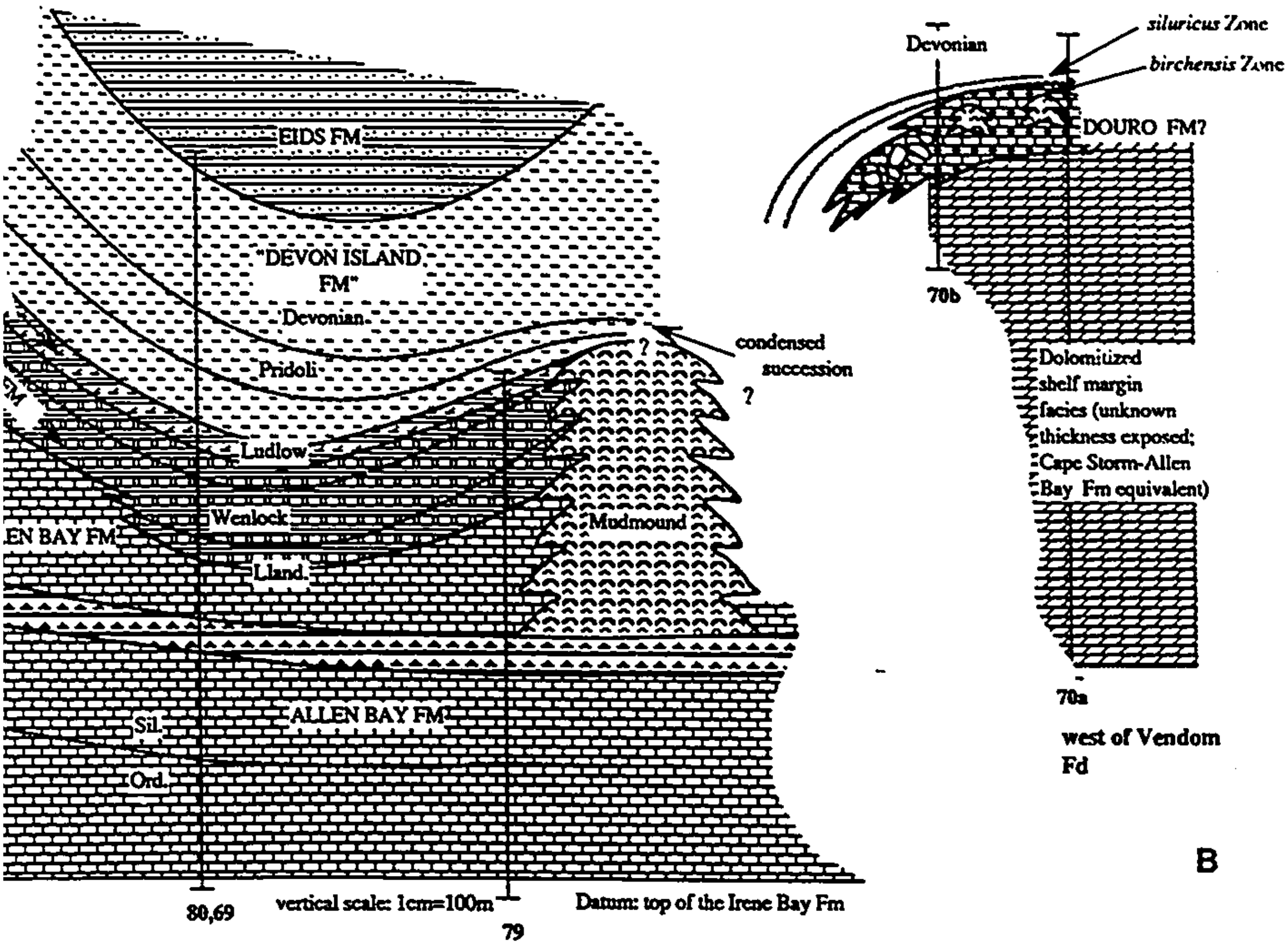
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Sil.
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condensed succession
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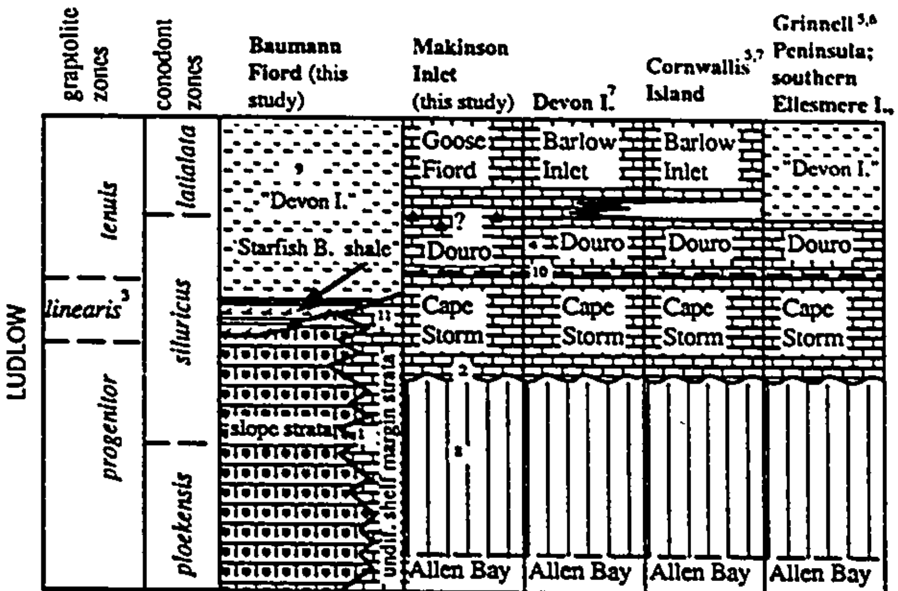
80,69
vertical scale: 1cm=100m
79
Datum: top of the Irene Bay Fm
Devonian
siluricus Zone
birchensis Zone
DOURO FM?
70b
Dolomitized shelf margin facies (unknown thickness exposed; Cape Storm-Allen Bay Fm equivalent)
70a
west of Vendom Fd
B

Figure 19: Age relationships of Ludlow platform and deep-shelf lithostratigraphic units.



[1] Dolomitized shelf margin facies, in part, assignable to the Cape Storm Formation, although unconstrained by conodont biostratigraphy. The facies represented by the upper part of this succession is remarkably similar to the Cape Storm Formation shelf-margin facies on east-central Cornwallis Island.

[2] Placement of the base of the Cape Storm Formation within the *siluricus* Zone i is uncertain

[3] Placement of the *linearis* Zone in relation to the Cape Storm Formation is based on the platform margin stratigraphy described by Thorsteinsson and Uyeno (1980) for east-central Cornwallis Island.

[4] Occurrence of abundant *Bohemograptus bohemicus tenuis* in the Douro Fm (Thorsteinsson and Uyeno, 1980).

[5] Thorsteinsson and Uyeno, 1980)

[6] Mayr *et al.*, (in prep.)

[7] Thorsteinsson and Mayr (1987)

[8] Sub-Cape Storm erosion recognized on southern Ellesmere Island and Grinnell Peninsula by Mayr *et al.* (in prep.), and by Thorsteinsson and Mayr (1987).

[9] See text for discussion on the "Devon Island Formation"

[10] The Douro-Cape Storm Fm contact is gradational and may be diachronous within the very thick *siluricus* Zone. Its placement, other than in Cornwallis Island, in relation to the *linearis* Zone is subjective.

[11] The upper 75-100m of this succession is likely assignable to the Douro Fm, based on the occurrence of *siluricus* Zone conodonts and the rubbly weathering atrypid-bearing limestone. Isolated stromatoporoid reefs, atypical for the Douro Fm, also occur.

OTHER UPPER SILURIAN PLATFORM CARBONATES

Several sections of the Goose Fiord Formation were measured, but for the sake of brevity, and because of incomplete biostratigaphic information, these exceedingly thick sections (greater than 1.6km) are not described in detail here. Between Vendom Fiord and Makinson Inlet, the Goose Fiord Formation is a fossiliferous, biohermal, and cyclical unit, that is informally divided into two members, a fossiliferous lower limestone member and a peritidal, sparsely fossiliferous, cyclical upper dolostone member. The lower contact, as discussed above, is gradational and placed arbitrarily in about the middle of a cherty unit containing abundant, well preserved and silicified *Encrinurus* trilobites. A similar unit was described in detail by Thomas and Narbonne (1979) in about the same stratigraphic interval on Cornwallis Island.

Unit 7a of **section 31** (Fig. 9c) is probably coeval with the lower part of the Goose Fiord Formation. It is characterized by extremely fossiliferous biostromal carbonates containing numerous laminar stromatoporoids, corals, brachiopods, and calcareous algae. Vertical 2-5cm diameter burrows in a calcarenite bed are associated with small mound-like bedding-plane structures that resemble some recent crustacean burrows (Plate 7a). Some beds containing up to 60% bulbous laminar stromatoporoids (Plate 7e) are interspersed with coral-rich beds. Unit 7a has yielded macrofossils of Pridoli aspect (Fig. 9c) and is thus equivalent to the Goose Fiord

Formation.

Discussion and interpretation

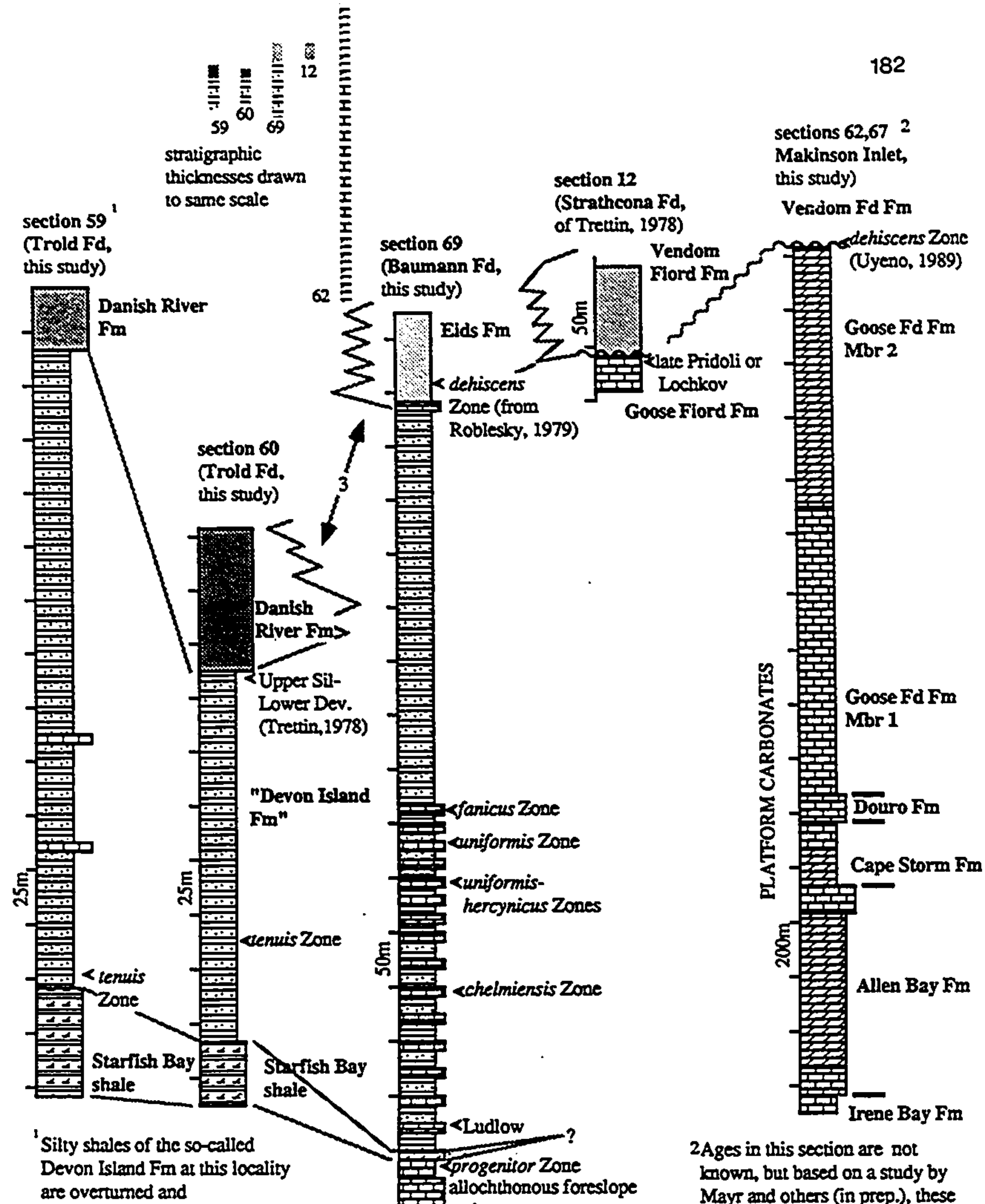
The Goose Fiord Formation is lithologically very diverse and by itself could be treated as an entirely separate study. In general, the fossiliferous and nodular lower unit of the Goose Fiord Formation was deposited on a shallow ramp, in an open marine environment that allowed stromatoporoid biostromes, bioherms, and a diverse assemblage of corals and brachiopods to flourish. These strata show cyclic alternation with mudcracked dolostone, microbialites, and unfossiliferous bioturbated lime mudstone. Cyclicity is very characteristic of the formation. More than 100 "high-frequency" cycles have been identified, and in the upper part of the formation, the cycles contain exposure surfaces and sandstones. Some of these clastics were likely derived from a phase of Caledonian tectonism, the Inglefield Uplift, that strongly influenced this area during the Upper Silurian and Lower Devonian.

While the Goose Fiord Formation was strongly progradational in the areas of Grinnell Peninsula, southern Ellesmere Island (Mayr *et al.*, in prep.), and Bay Fiord, the Goose Fiord shelf-margin remained relatively static through the Pridoli and Lower Devonian, near Vendom and Baumann fiords. The shelf margin itself probably lies beneath Vendom Fiord, and the sequence recorded in section 69 (Fig. 20) is the best representation of coeval slope facies in the study area. However, this sequence lacks coarse allochthonous carbonate detritus, is very silty, particularly in the upper part, and is abruptly but conformably overlain by the Eids Formation (Fig. 20). However, less than 10km eastward, peritidal and, in places, evaporitic platform carbonates imply a shelf-margin barrier. Although the existence of a barrier facies cannot be proven, the platform-to-basin transition appears to have been more-or-less stationary through much of the Pridoli and Lower Devonian, in the vicinity of Vendom and Baumann fiords. To the north, it was highly progradational, and was associated with a prograding wedge of clastics assignable to the Red Canyon River Formation.

Crinoid channel complexes at Bay Fiord (sections 5, 10, 12, Fig. 8b; Fig. 21; Plate 8f) attest to off-bank transport of shelf-margin crinoid detritus, possibly derived from a crinoid bank complex. Crinoidal material, although dolomitized, appears "fresh" (Gawthorpe and Gutteridge, 1990), suggesting high rates of sedimentation and crinoid production. The progradational geometry is consistent with uplift of the platform, decrease in accommodation, or increase in sediment supply.

Coeval deposits showing progradational platform geometry are present on Devon and southern Ellesmere Island (Mayr *et al.*, in prep.) and on Cornwallis Island (Thorsteinsson and Uyeno, 1980). However, the lack of evidence of coeval progradation in the area of Vendom and Baumann fiords is curious and undoubtedly related to an underlying tectonic (probably fault) control. Increased subsidence in this area perhaps impeded platform progradation, and furthermore may explain a regional platform facies change within the Goose Fiord Formation, from

Figure 20: Upper Silurian and Lower Devonian sections of this study near Vendom and Trold fiords and correlation with section 12 described by Trettin (1978). Legend as in Figure 6d.

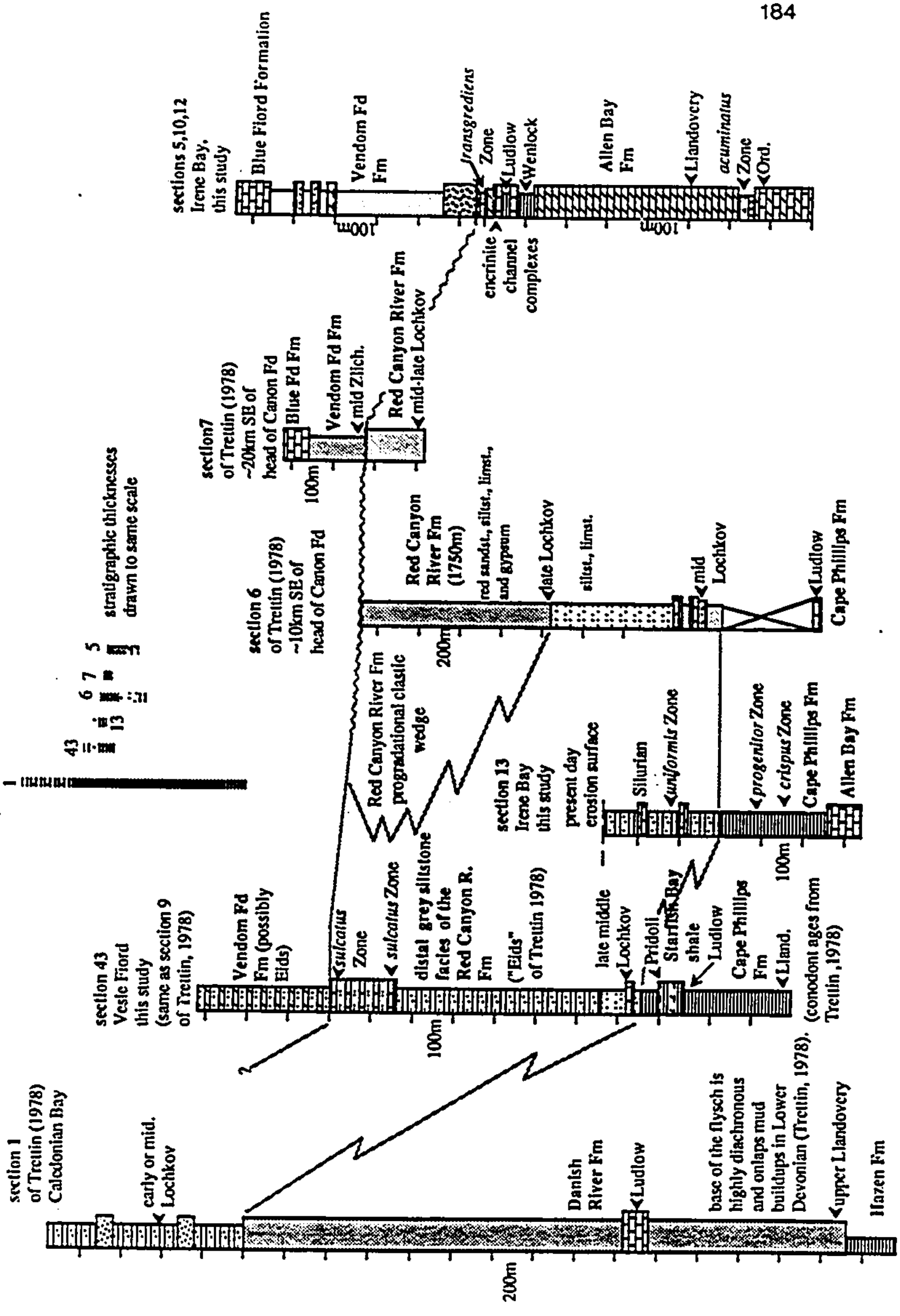


[1] Silty shales of the so-called Devon Island Fm at this locality are overturned and may be tectonically thickened.

[2] Ages in this section are not known, but based on a study by Mayr and others (in prep.), these formations likely have very simiar ages to those established in the southern Arctic (Thorsteinsson and Uyeno, 1980).

[3] It is presently not possible to correlate these two formations.

Figure 21: Upper Silurian and Lower Devonian stratigraphic sections of this study north of Bay Fiord and their correlation with sections described by Trettin (1978). Legend as in Figure 6d.


section 1
of Trettin (1978)
Caledonian Bay
early or mid.
Lochkov
Danish
River Fm
Ludlow
base of the flysch is
highly diachronous
and onlaps mud
buildups in Lower
Devonian (Trettin, 1978).
upper Llandovery
Hazen Fm
200m
section 43
Vesle Fiord
this study
(same as section 9
of Trettin, 1978)
Vendom Fd
Fm (possibly
Eids)
sulcatus
Zone
sulcatus Zone
distal grey siltstone
facies of the
Red Canyon R.
Fm
("Eids"
of Trettin 1978)
late middle
Lochkov
Pridoli
Starfish Bay
shale
Ludlow
Cape Phillips
Fm
Lland.
(conodont ages from
Trettin, 1978)
100m
stratigraphic thicknesses
drawn to same scale
1
43
13
6
7
5
Red Canyon River Fm
progradational clastic
wedge
section 13
Irene Bay
this study
present day
erosion surface
Silurian
uniformis Zone
progenitor Zone
crispus Zone
Cape Phillips Fm
Allen Bay Fm
100m
section 6
of Trettin (1978)
~10km SE of
head of Canon Fd
Red Canyon
River Fm
(1750m)
red sandst., siltst., limst.
and gypsum
late Lochkov
siltst., limst.
mid
Lochkov
Ludlow
Cape Phillips Fm
200m
section7
of Trettin (1978)
~20km SE of
head of Canon Fd
Blue Fd Fm
Vendom Fd Fm
mid Zlich.
Red Canyon River Fm
mid-late Lochkov
100m
sections 5,10,12
Irene Bay,
this study
Blue Fiord Formation
Vendom Fd
Fm
transgrediens
Zone
Ludlow
Wenlock
Allen Bay
Fm
Llandovery
acuminatus
Zone
Ord.
Red Canyon River Fm
encrinite
channel
complexes
100m
100m

predominantly subtidal open marine facies east of Vendom Fiord to restricted and silty facies near southern Makinson Inlet, and southern Devon Island. Some of these relationships are reiterated below.

CAPE PHILLIPS FORMATION (ODcp)

The Cape Phillips Formation was named by Thorsteinsson (1958), and its type section is located on northern Cornwallis Island. It is a widely exposed and thick stratigraphic unit in the Canadian Arctic Islands and is lithologically diverse. This study is limited to consideration of only the broad lithological divisions of the formation and their correlation. For this purpose, about 200 graptolite samples were collected and identified under a binocular microscope (listed in Appendix two). At least as many samples were readily identified in the field, as many Silurian graptolites have very distinct cladi and thecae. These fossils formed a crucial biostratigraphic framework for the study.

Several major lithological subdivisions are recognized in the formation; one at least -- the Starfish Bay shale -- is distinct and mappable and could justifiably be separated as a formal lithostratigraphic unit. Also, the thick grey siltstone unit above this, in central Ellesmere Island, was assigned by Trettin (1978) to the Eids Formation, but it does not now appear to belong, based on its age and stratigraphic relationships. In the following discussion, this grey-weathering siltstone and mudstone unit at Vesle and Bay fiords will be referred to informally as the grey siltstone member of the Red Canyon River Formation. Thicknesses of the formation are given in Table 7.

Pre-Ludlow succession

The base of the Cape Phillips Formation is strongly diachronous. Established ages of this contact indicate that major phases of platform backstepping occurred in the Late Ordovician, early Llandovery, middle Llandovery and middle Ludlow. At Trold Fiord (sections 57, 58, 59, 60, Fig. 7f,g), the contact occurs in the *fastigatus* Zone, above a 2-4m thick tongue of the Allen Bay Formation (Plate 8a). The contact is gradational over about 5m (Plate 8b), and numerous *Pseudogygites* pygidia are associated with this interbedded shale and argillaceous limestone. Units above are varied and include clayshale, mudshale, marlstone, and thick limestone interbeds (Plate 8c,d). These beds are predominantly allochthonous (see part III) and related to deposition on the slope of a platform mud buildup, not exposed at this locality. Numerous slump structures, intraclast beds, graded skeletal packstone beds, and uncommon microbial carbonate olistoliths occur in these predominantly Llandovery carbonates. The youngest major allochthonous carbonate deposits in the Trold Fiord area (section 58; Fig. 7f) occur below the *perneri* Zone, therefore correlating approximately with mud buildup termination at Caledonian Bay, which also occurred in the lower *patula* Zone. Above these allochthonous units, thinly bedded carbonates lacking sedimentary structures are associated with deposition on a carbonate slope.

To the north of Trold Fiord (section 59) Llandovery and lowest Wenlock allochthonous carbonates are less common, and dark grey and black mudshale and silty mudshale occur. The middle and upper Wenlock succession is condensed (*sensu* Plate 9c) and consists predominantly of fissile clayshale and mudshale, with less common limestone and silty limestone beds. These carbonate beds weather a conspicuous orange-brown colour and occur regularly throughout the Wenlock, Ludlow, and Pridoli portions of the succession. The contact with the overlying Starfish Bay shale is gradational and marked by an increase in green-weathering calcareous mudrock and laminated limestone beds. The age of this contact is discussed below.

To the east of Trold Fiord, in **section 60** (Fig. 7f), the Cape Phillips Formation below the Starfish Bay shale is a condensed sequence (122m thick east of Trold Fiord compared to at least 296m west of Trold Fiord, although a complete section of the Cape Phillips Formation was not measured in the west). Section 60 lacks the Llandovery and Wenlock allochthonous carbonates that are present farther west, and contains predominantly thinly laminated, petroliferous limestone beds and significant quantities of bedded black chert: a 10m thick unit consisting of almost entirely of bedded chert and dated as approximately *curtus/pectinatus* Zone (Figs. 7f, Plate 8c,d)). The Wenlock sequence is condensed, and silty limestone, silty mudshale, and siltstone are most common.

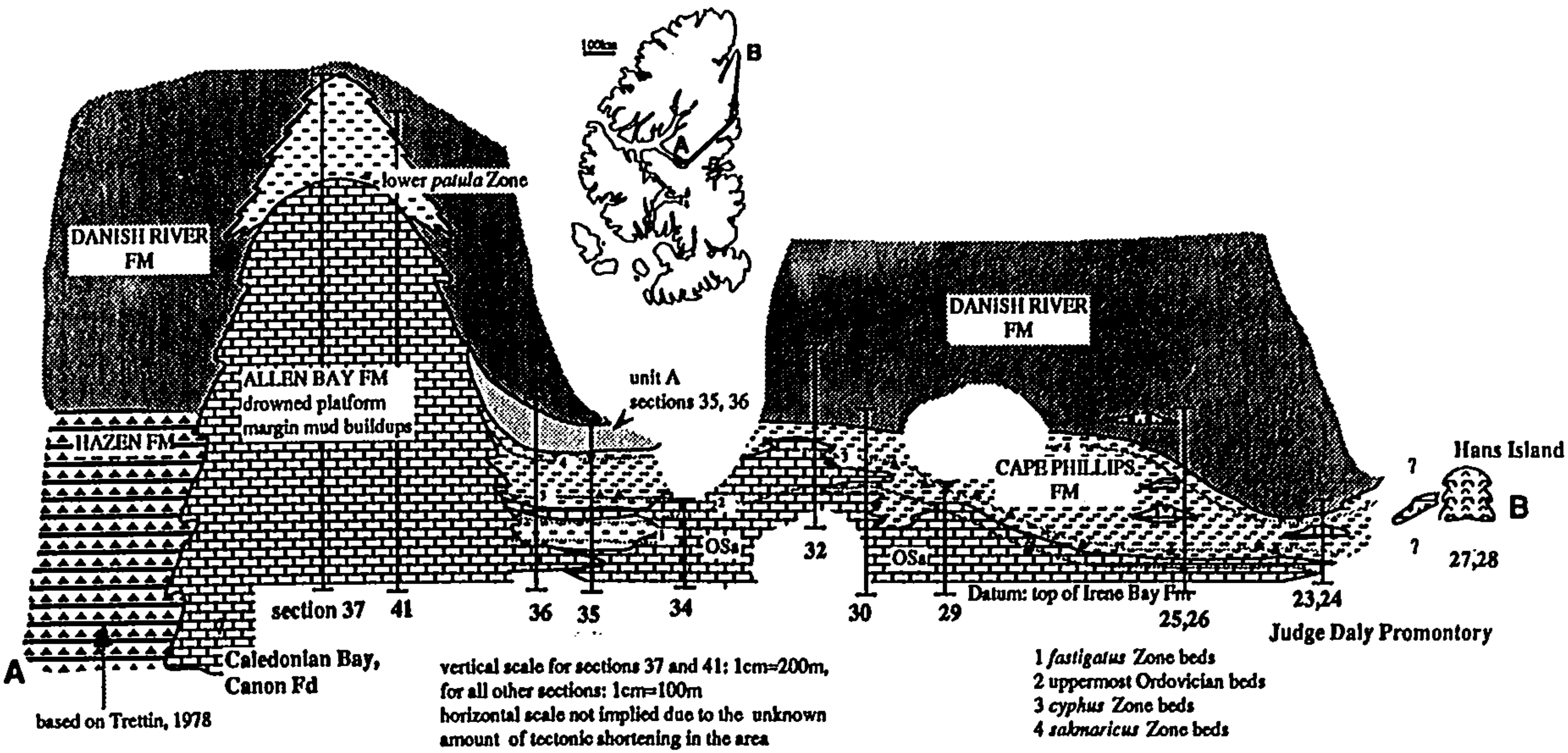
The lower contact with the Allen Bay Formation is also gradational in the Trold Fiord sections and placed at the base of the regularly interbedded, laminated to bioturbated, trilobite-rich petroliferous limestone and fissile black clayshale. The contact occurs within the *fastigatus* Zone. Silt content gradually increases in younger units and becomes abundant in about the upper Llandovery and lower Wenlock part of the section, below the Starfish Bay shale.

At Baumann Fiord, the base of the Cape Phillips is considerably younger (upper *turriculatus* Zone, in **section 47**, Fig. 7d to upper *sakmaricus* Zone in **section 54**, Fig. 7e). There is an extensive lower Llandovery chert unit in the vicinity of Vendom Fiord, below which the base of the Cape Phillips Formation could be placed, as indicated in Figure 7e. However, this boundary placement is uncertain due to the interfingering relationship of carbonate and shale. The chert-rich interval in this area is dated as likely *cyphus* to *convolutus* Zones in age, and probably correlates with the widespread chert unit on northern Ellesmere Island (Fig. 22). In this latter remote study area, the Cape Phillips Formation shales are lithologically very diverse, a characteristic related chiefly to several causes, but most probably due to local allochthonous carbonate slope sedimentation associated with pinnacle reefs inferred to be situated in the Kennedy Channel area (Plate 9a,b).

In **section 54** (Fig. 7e), part of the Llandovery Cape Phillips Formation consists of about 70m of thin bedded black chert and clayshale interbeds, separating lower massive dolostone and limestone from an upper bioturbated, pentamerid-rich

Figure 22: Diagrammatic restored section of the main platform, deep-shelf, and trough stratigraphic units near Cañon and Bay fiords and on Judge Daly Promontory. Legend as Figure 6d. See Figures 1 and 5 for section locations.

Note the reduced scale for sections 37 and 41.


DANISH RIVER FM
lower patula Zone
ALLEN BAY FM
drowned platform
margin mud buildups
HAZEN FM
section 37
41
36
35
34
32
30
29
25,26
23,24
27,28
unit A
sections 35, 36
OSa
DANISH RIVER FM
CAPE PHILLIPS FM
Datum: top of Irene Bay Fm
Hans Island
A
B
Caledonian Bay,
Canon Fd
based on Trettin, 1978
Judge Daly Promontory
vertical scale for sections 37 and 41: 1cm=200m,
for all other sections: 1cm=100m
horizontal scale not implied due to the unknown
amount of tectonic shortening in the area
1 fastigatus Zone beds
2 uppermost Ordovician beds
3 cyphus Zone beds
4 sakmaricus Zone beds

and intraclast-rich lime mudstone; the latter is informally referred to as the "Llandovery carbonate unit" in Part III. A continuously exposed expanded Wenlock succession in section 54 (Fig. 7e; Plate 4) contains numerous lithologies that include the following: from most to least common, marlstone, clayshale, allochthonous slope carbonates including thick megaconglomerate beds (with blanketing allodapic beds), intraclast beds, ribbon limestone, substratally mudcracked argillaceous skeletal wackestone and packstone, bioturbated limestone, slumped ribbon limestone, rare chert beds, skeletal rudstone beds, rare *in situ* sponge biostromes, sponge rudstone beds, oolitic grainstone and packstone, laminated argillaceous limestone, and isolated olistoliths, some as large as 17m in diameter. These Wenlock units are overlain by a Ludlow succession of very fossiliferous rubbly weathering limestones containing abundant oolites, oncoids, pentamerid brachiopods, bryozoans, and other macrofossils.

West-southwest of section 54, a similar stratigraphic section is present in the subsurface of Bjorne Peninsula (Panarctic Eids M-66 well). Logs of this well and the Panarctic Blue Fiord E-46 well were examined by the author and by Uyeno *et al.*, (1990). In the Eids M-66 well, above a 69.9m thick Irene Bay Formation, the lower 148m thick member of the Allen Bay Formation consists of a lower limestone-dominated and an upper dolostone-dominated succession, that is remarkably similar to that in section 54, where 275m of limestone and dolostone are present. In both wells, a chert-rich carbonate-shale unit containing graptolites of *cyphus* Zone age (lower middle Llandovery) overlies these platform carbonates. However, associated with this contact in the Eids well is a considerable hiatus, spanning the Upper Ordovician (fauna 12; Uyeno *et al.*, 1990) to middle Llandovery (*staurognathoides* Zone; Uyeno *et al.*, 1990). Although conodonts have not been collected from the top of the carbonate succession in section 54, by analogy with the bio- and lithostratigraphy in the Eids M-66 well, the Ordovician-Silurian boundary appears to rest in the hiatus associated with the shale-limestone contact instead of in the underlying platform carbonates as suggested by plate 4. This relationship is repeated on Judge Daly Promontory (sections 25 and 26, Fig.7b), suggesting that an Upper Ordovician, post-Richmondian (Hirnantian?), sea-level rise, possibly glacio-eustatic, produced the regional platform drowning surface.

The Cape Phillips Formation is well exposed in the Bay Fiord area (sections 2 and 13, Fig. 7a,b). In section 13, the lower Allen Bay Formation carbonate grades into the Cape Phillips Formation shale over several metres. The lowest cherty argillaceous limestone and limestone beds yielded graptolites assignable to the *cyphus* or *curtus/pectinatus* Zones (Fig. 7a,b). Overlying units are monotonous and form a condensed sequence comprising interbedded silty mudshale, bedded chert, laminated marlstone, and limestone, the latter rarely including conglomerate beds, graded skeletal wackestone and packstone, and intraclast beds. These rock types persist through into the Ludlow part of the section, although becoming less calcareous upwards. Beds above the *linearis* Zone are unfossiliferous clayshale and mudshale, weathering various hues of medium grey and reddish grey. The contact of this unit with the overlying light grey, recessive, monotonous claystone and siltstone (referred to informally as the grey siltstone member of the Red Canyon River Formation) is covered. Neither graptolites nor strata suitable for conodont sampling were present at the contact; hence, the lower age of this contact is unknown. However, 154m above the base of this member, graptolites belonging to the *uniformis* Zone occur. Some allochthonous carbonate beds occur in this member, but these did not yield diagnostic conodonts. Recognition of the "Devon Island Formation" just west of Irene Bay is not possible, based on present observations. The Starfish Bay shale, as discussed below, is not present and the pre- and post-Ludlow shale succession is similar. Exposure is poor, however, and this distinctive shale unit, as established in the vicinity of Baumann Fiord, pinches out platformward and could be obscured in the rubbly exposure.

In **section 2**, 3km north of section 13 (Fig.4, 7b), conodonts from the last thick bedded limestone bed are latest Ordovician in age; 18m above this conodont occurrence, *minor* Zone graptolites are present in the first fissile shale of the section. This 20m interval therefore represents most of the middle and all of the lower Llandovery, and perhaps also part of the Upper Ordovician. It is a condensed sequence of predominantly laminated and cross-laminated silty limestone containing slump, truncation, and flame structures. The upper Llandovery and Wenlock part of the section is also relatively thin, comprising interbedded chert, mudstone, mudshale, and clayshale together with common allochthonous conglomerate beds containing abundant well-preserved macrofossils. Above these units, uppermost Wenlock and Ludlow strata are much less calcareous and are predominantly fissile clayshale and mudshale. It is significant that the very thin Llandovery interval in these sections thickens to 500m in carbonates only 3km platformward. As discussed earlier, this relationship may indicate local penecontemporaneous removal by slumping as is evident by slide scars.

At section 43 (Vesle Fiord), numerous samples were collected from the Cape Phillips Formation in 1988, and , but unfortunately, the cached samples were never retrieved. Much of the following description is, therefore, based on a field record of graptolite identification and hand sample description. This area was also visited by Trettin (1978), and his data was incorporated.

At this locality, the lower contact with the Allen Bay limestone is gradational, and, as in the sections at Trold Fiord, a substantial thickness of the *fastigatus* Zone is present. The succession above this is well exposed, and conspicuously limestone-rich. Slump structures and graded skeletal packstone and wackestone are abundant. Above, a very thick Llandovery and lower Wenlock succession is exposed, and much of the carbonate is probably allochthonous and related to the large shelf-edge buildups just north of this section. A condensed middle and upper Wenlock section is present (Plate 9c), as in the Trold Fiord sections, and is overlain by about 60m of resistant green-weathering, well-indurated calcareous mudrock, assignable to the Starfish Bay shale. About 40m above this are pale grey claystone and siltstone of the Red Canyon River Formation (Plate 9d). The section was terminated at this point, but additional biostratigraphic and lithological data are available in Trettin's (1978; section 9-1, p.21) report. At this locality, Trettin (1978) referred to the thick, grey siltstone unit as the Eids Formation, dividing it into three units, a lower limestone unit, a middle siltstone unit, and an upper limestone unit. He interpreted the limestones as partly allochthonous, and established an age for the formation that spans the *delta* to *sulcatus* conodont Zones.

Several sections of the Cape Phillips Formation were examined on northeastern Ellesmere Island in early April of 1988, when only the most precipitous sections were snow-free. Rubbly recessive intervals were generally under snow (Plates 9a,b; 8e); hence, large covered intervals are present in several of the sections. Nevertheless, main lithological relationships were established, and the following observations are significant (also see Fig. 22):

(i) With the exception of section 32 (Fig.7c), which yielded conodonts equivalent to the *kentuckyensis* Zone, conodonts from the last limestone bed of the upper part of the Allen Bay Formation are invariably latest Ordovician in age (this included sections 26, 35, 36; Fig. 7b,d). Strata immediately overlying these beds were dated as early Llandovery (*curtus* Zone in section 26, Fig. 7b; *cyphus* to *convolutus* Zones in section 29 Fig. 7b; *cyphus* Zone in section 30, Fig.7a), indicating significant hiati associated with these contacts.

(ii) a markedly chert-rich unit is apparently isochronous, coinciding, at least in part, with the *cyphus* Zone. However, the impoverished conodonts in the base of the Llandovery, and occurrence of the *cyphus* Zone beds immediately above the Ordovician carbonates indicate that the gradational contact of the carbonates and the shales is an exceedingly condensed sequence.

(iii) An important observation obtained from sections on northeastern Ellesmere Island is the age of the overlying clastics assignable to the Danish River Formation (Plate 9a,b). The contact is gradational and probably slightly diachronous. The base of this formation, however, is accurately dated in only two sections (30 and the composite section 23/24, Fig.7a,b, and Fig.22), and appears to be late Llandovery, probably within the uppermost part of the *griestoniensis* or lower *sakmaricus* Zones.

(iv) Llandovery carbonate units thicken in proximal shelf margin locations and interfinger with the deep-shelf shales; evidence of major onlapping is recognized in the Upper Ordovician (*fastigatus* Zone); lower Llandovery (*cyphus* Zone); and possibly upper Llandovery (*?griestoniensis* or *sakmaricus* zones). The *cyphus* Zone, in particular, coincides with a chert-rich sequence that is correlative regionally through much of the study area.

Sections 35 and 36 (Fig. 7d) show stratigraphic relationships similar to those recognized on eastern Ellesmere Island. Lowest Cape Phillips shaly units fall within the *fastigatus* graptolite Zone. Above this, a poorly exposed succession of black clayshale, mudshale, and marlstone occurs. Although biostratigraphic information is incomplete, there is a suggestion that the Ordovician/Silurian boundary falls within two thick carbonate units (section 36, Fig. 7d). These carbonate units appear to be bioturbated and, in part, allochthonous. Above them, marlstone and mudshale strata containing *sakmaricus* Zone graptolites and indurated dark-brown-weathering siltstone are present. These strata are laminated, calcareous, rarely graded, and lack significant carbonate and/or graptolitic beds. The unit is 63-140m thick, weathers brown-grey, and breaks into elongate "stick-like" clasts. It is poorly exposed but has a distinctive lithology not recognized elsewhere in the Canadian Arctic. Danish River Formation clastics abruptly but conformably overlie this siltstone, but the age of this contact is presently unknown (see Fig. 22 for the stratigraphic relationship of this unit).

Ludlow to Lower Devonian succession (so-called Devon Island Formation)

The Devon Island Formation was originally defined by Thorsteinsson (*In* Fortier *et al.*, 1963, p.228) for a uniform, medium dark grey to greyish black, calcareous shale and shale with minor argillaceous limestone interbeds. Thorsteinsson and Uyeno (1980) and Thorsteinsson and Mayr (1987) later re-examined the formation on northern Devon Island, in the vicinity of the type section, and recognized two broad subdivisions, a lower 182m thick, remarkably homogeneous medium to dark grey siltstone succession and an upper, 48m thick, grey, vuggy, dolostone succession. More recently J.C.Harrison (1989, unpublished information) assigned sequences in the Vendom and Trold fiord areas to the "Devon Island Formation", and these stratigaphic sections were examined by the author.

For reasons discussed more fully below, and because the Devon Island Formation stratotype is unlike isochronous units in the vicinity of Baumann and Trold fiords, recognition of this Formation in the vicinity of Baumann and Trold fiords is uncertain. Based on findings of this study, establishment of a reference section of the Devon Island Formation (art.8e, North American Stratigraphic Code, 1985) that contains the Starfish Bay shale would clarify this uncertainty and allow the Devon Island Formation to be mapped in the vicinity of Baumann and Trold fiords. Because the stratotye is unlike equivalent units in the Baumann, Trold, and Vendom fiord areas and a reference section has not yet been established, the rocks that Harrison (1989, unpublished imformation) included in the Devon Island Formation in these areas are provisionally included in the Cape Phillips Formation. For clarity, this part of the Cape Phillips formation, pending revision, is therefore referred to as the "Devon Island Formation" or as the so-called Devon Island Formation in the following text and in the relevant diagrams. Thicknesses of this unit are given in Table 5.

Starfish Bay shale (informal name)

The Starfish Bay shale is a newly distinguished unit in the study area and constitutes tha basal part of the "Devon Island Formation", although formal recognition as a formation or member is deferred in this investigation. It is mappable and likely correlative to a similar unit on Melville Island (C.J. Harrison, pers.com., 1990), hence appears to be one of the most regionally extensive Silurian deep shelf units presently known. The unit is best exposed along the coasts of Trold Fiord, but this locality name is preoccupied by a Permian unit (Thorsteinsson and Tozer, 1970; Thorsteinsson, 1974), and for reasons of homonymy (art.7b, North American Stratigraphic Code, 1985), this name cannot used. The unit is also exposed nearby at Starfish Bay, and this name is informally applied to the distinctive sequence of green-weathering mudrock (shale) which is also recognized near Vendom, Vesle, and Baumann fiords. As discussed further below, the Starfish Bay shale is part of the basal portion of the "Devon Island Formation", although not present in the stratotype. Thicknesses of this unit are given in Table 6.

This shale is well exposed and thickest in the Trold Fiord and Bay Fiord areas (Plate 9e). It is entirely Ludlow in age, and has a lower gradational, diachronous contact and an upper gradational, possibly synchronous contact (Fig.15a-d). At Baumann Fiord, a very thin green-weathering clayshale, probably equivalent to the Starfish Bay shale, rests directly on allochthonous slope carbonates (Figs.6a-c,15a-d,18a-b,19). Generally, this shale unit is well indurated, poorly laminated, grey-green weathering, partly bioturbated claystone, mudstone, or mudshale; however, the unit is slightly more recessive and more thinly bedded at Baumann Fiord. Other less common sedimentary features include parallel laminae in carbonate beds and mudrock, pyrite concretions, uncommon pyritized brachiopods, and fish remains. Graptolites are generally absent, in contrast to their abundance above and below the shale unit. *Bohemograptus bohemicus bohemicus* and *B. bohemicus tenuis*, for example, occur in great numbers above and below, but not within the shale unit. At Trold Fiord, parallel laminated, orange-brown-weathering, silty dolomitic limestone beds occur throughout the shale unit at regular 2 to 5m intervals, although they are slightly more abundant near the base, and also occur above in overlying silty mudshale and siltstone of the upper part of the so-called Devon Island Formation. If it were not for this very distinctive green-weathering shale, the "Devon Island Formation" would be essentially indistinguishable from the Cape Phillips Formation and exceedingly difficult to differentiate in geological mapping (Plate 9c), as discussed further below.

At Baumann Fiord, the Starfish Bay shale is discontinuous, and present over allochthonous slope carbonates but not over contiguous autochthonous pinnacle reef and platform carbonates only 1 or 2km away (sections 70, 70b, Fig. 18a-b). Its absence over *in situ* carbonates may be related to their intrinsic topographic relief, a feature which undoubtedly led to condensed sequences at these localities.

This distinctive unit is also present in the subsurface in the Eids M-66 well.

In the depth interval between 454m and 466m, the Starfish Bay shale is readily distinguished by, its medium grey colour, from the surrounding darker coloured silty shales (only in weathered exposures does the unit have the distinctive green-weathering colour). It is relatively thin (12m), as observed in other platform-situated sections and in contrast to the much thicker Trold Fiord occurrences. It has not been recognized in the Blue Fiord E-46 well, where a apparently coeval pinnacle reef structure, probably representing a paleotopographic high during the Ludlow, would have prevented accumulation of the Starfish Bay muds.

Upper siltsone member

On Devon Island, the contact of the Devon Island Formation with the underlying Douro Formation is abrupt, conformable, and coincides approximately with the boundary between the *siluricus* and *latialata* conodont zones (Thorsteinsson and Uyeno, 1980; Thorsteinsson and Mayr, 1987). Its upper contact is distinctly diachronous (Mayr, *et al.*, in prep.), due to the progradational nature of the overlying Goose Fiord platform carbonates on northern Devon Island and southern Ellesmere Island.

Complete sections of the "Devon Island Formation" were examined at Baumann Fiord (section 69) and at Trold Fiord (sections 60 and 59). In section 69 (Fig. 15d, 18), three units are discernable: (i) a lower, poorly exposed thinly bedded green clayshale containing abundant *Bohemograptus bohemcus* ssp. and likely assignable to the Starfish Bay shale, (ii) a middle, Ludlow to Lower Devonian, limestone-rich shale succession (360m thick), and (iii) an upper, monotonous, thinly bedded, laminated siltstone and silty mudshale succession, that is sparsely fossiliferous and contains laminated, bright orange-brown limestone and dolostone interbeds at regular 5-10m intervals. The latter two divisions are assigned to the upper siltstone member of the Devon Island Formation.

In section 59 (Trold Fiord), the upper silstone member of the so-called Devon Island Formation consists of about 296m of dark grey, fissile, laminated siltstone and mudshale and beds of laminated, orange-brown weathering, silty, dolomitic limestone and dolostone that occur at regular 5 to 10m intervals. The latter are very similar to limestone and dolostone beds in section 69 (Fig. 15d), although individual beds could not be correlated regionally. In this section, boudinaged carbonate beds contain polygonal tension gashes with neospar infilling. The formation is conformably but abruptly overlain by "flysch" of the Danish River Formation. The age of this contact is poorly known but is probably lowest Devonian (Trettin, 1978) in age. The succession at this locality is overturned, and because of their intrinsic incompetence, these shales may have provided significant detachment level for west-verging thrusts. As a consequence, the succession may have been significantly thickened tectonically.

In section 60 (Plate 9e), the upper part of the so-called Devon Island

Formation formation is considerably thinner (Fig. 15a). About 160m of mainly dark grey, silty mudshale and siltstone contain interbeds of conspicuous, orange brown-weathering, regularly bedded dolostone, as in section 59. A distinctive carbonate-rich interval, about 30m thick, occurs in the middle part of the formation (Plate 9e), and a conspicuous sequence of green mudrock (Starfish Bay shale) containing minor limestone interbed rests in the basal portion of the "Devon Island Formation" at this locality (Fig.20).

The typical Devon Island Formation in southwestern Ellesmere Island and northeastern Devon Island contains large pinnacle reefs. Examined on a reconnaissance scale, they are well exposed in sea cliffs on southern Ellesmere Island. The reefs are surrounded by siltstone and mudshale, and appear to have mud-rich cores, and abundant stromatactis and crinoidal detritus. The mud cores grade upward into stromatoporoid framestone and rudstone. Lithistid sponges are also common and well preserved in off-reef facies. The reefs rest directly on, and probably extend down into, the Douro Formation. Large "house-sized" olistoliths at Goose Fiord were probably derived from these reef pinnacles (Mayr, pers. com, 1990). On northern Devon Island, the upper contact of the typical Devon Island Formation is strongly diachronous and marked by the appearance of progradational slope carbonates of the Goose Fiord Formation (Mayr *et al.*, in prep.).

Age and implications of the basal contact

The base of the typical Devon Island Formation and of the isochronous sequence in the Baumann Fiord area marks a major change in carbonate platform stratigraphy in the Canadian Arctic. The formation was originally established on northwestern Devon Island by Thorsteinsson (*In* Fortier *et al.*, 1963) for a silty, laminated shale that overlies Douro Formation carbonates; however, the formation is not recognized on Cornwallis Island, only 60km away. The age of the lower contact on Devon Island, as discussed above, is well established and coincides with the boundary of the *latialata* and *siluricus* zones (Thorsteinsson and Uyeno, 1980). According to these workers, the boundary between these two conodont zones corresponds approximately to the upper part of the *tenuis* graptolite Zone, a zone that is well recognized throughout the Canadian Arctic. However, the base of the so-called Devon Island Formation in the vicinity of Baumann Fiord (sections 75, 79, 77; Fig. 15b) lies within the *linearis* Zone. Thorsteinsson and Uyeno (1980) suggested that, in a crucial shelf margin location at Snowblind Creek, Cornwallis Island, a 3.5m interval containing abundant *Saetograptus fritschi linearis* could be traced platformward into about the middle part of the Cape Storm Formation. However, they did not take into account the significant penecontemporaneous platform relief, which, excluding compaction, may have been up to 200m (de Freitas, 1987c). This fact may not be crucial to the biostratigraphy of the area, however, but it does place the *linearis* Zone slightly higher in the correlative Cape Storm Formation.

Figure 15 shows that in the vicinity of Baumann Fiord the base of the Starfish Bay shale lies within the *linearis* Zone, whereas at Trold Fiord the base is well within the *progenitor* Zone. Therefore, the basal contact is diachronous, younging slightly platformward, in an onlapping relationship with the platform foreslope carbonates near Baumann Fiord. As argued above, the *linearis* Zone is likely a correlative of the upper part of the Cape Storm Formation on east-central Cornwallis Island; hence, the Starfish Bay shale appears to be wholly equivalent to the Cape Storm Formation. However, a thick succession of the upper part of the so-called Devon Island Formation limestone and shale succeeds the Starfish Bay shale and underlying allochthonous slope carbonates (section 69, Fig.15d, 18b), while 5 km to the north (sections 70a, 70b, 18b) the Starfish Bay shale is absent and about 175m of Douro-like platform and slope carbonates (yielding *siluricus* Zone conodonts) conformably overlie Cape Storm-Allen Bay Formation shelf margin dolostone. These stratigraphic relationships may be explained as follows: The basal contact of the so-called Devon Island Formation appears to be an "event horizon" that marks abrupt termination of shallow-water carbonate production. As a result, a condensed sequence of fine-grained, terrigenous and carbonate muds was deposited over highs (over shallow water platform carbonates in sections 70a, 70b) while terrigenous muds and marls of the so-called Devon Island Formation (including the Starfish Bay shale) were deposited over intervening lows (Fig. 18). Platform drowning was synchronous regionally and occurred within the *linearis* Zone. The lower limestone-rich portion of the so-called Devon Island Formation (section 69, Fig.15d) is relatively condensed, lacks sedimentary structures indicative of slope transport, and appears to have been deposited predominantly through suspension fallout, perhaps representing off-platform carbonate transport by storms. After drowning, a new platform margin was established 5-10km eastward of its Wenlock and lower Ludlow position. Because the base of the Starfish Bay shale is considered to be coincident with regional platform drowning, the Douro-like beds in sections 70a and 70b must be within or below the *linearis* Zone. This relationship indicates either (i) that the *linearis* Zone is slightly younger than previously thought and, may be, in part, coeval with the *latialata* Zone, or, more likely, (ii) that the contact between the Douro Formation and the so-called Devon Island Formation is diachronous, younging to the south, from Baumann Fiord to northern Devon Island (see Fig. 19).

On Devon Island, the Devon Island Formation consists of a lower 182m thick unit of a "...remarkably homogeneous appearing succession of siltstone that is variably calcareous and dolomitic, with lesser amounts of interbedded dolomite mudstone and wackestone..." and an upper 48m thick unit that is mainly dolomite and subordinate siltstone (Thorsteinsson and Uyeno, 1980, p.16). It is a mappable unit in Devon Island, and recent Geological Survey of Canada mapping in the Trold Fiord, Baumann Fiord, and Vendom Fiord map areas has delineated shale apparently assignable to the the Devon Island Formation. At Trold Fiord, recognition of this so-called "Devon Island Formation" apparently relies on the presence of the Starfish Bay shale, as strata above and below this shale are otherwise essentially indistinguishable. The fact that the Devon Island Formation on Cornwallis Island or in the vicinity of Cape Sir John Franklin (section 19, Fig.1) has not been recognized tends to substantiate this. Despite the fact that the Starfish Bay shale does not occur in the type area of Devon Island, it has been proposed that the base of the shale be used to define the base of the Devon Island Formation on Melville Island and near Trold Fiord, Ellesmere Island (Harrison, pers. com., 1989). The Starfish Bay shale is distinct and mappable, but has not been reported in the Devon Island Formation type area and as such was not a factor in the definition of that formation. Moreover, the two broad lithological subdivisions of the Devon Island Formation in the type area described by Thorsteinsson and Uyeno (1980) are quite unlike the subdivisions of the equivalent deep-shelf units recognized near Vendom Fiord (section 69, Fig. 15d): on Devon Island, siltstone underlies dolostone, but near Vendom Fiord, siltstone and overlies limestone and shale. These relationships raise doubts about extending the Devon Island Formation to the Trold and Vendom fiord map areas. As suggested above, there are two possible solutions to this problem. The first and perhaps most practical soution would be to establish the Starfish Bay shale as a formal or informal member of the Devon Island Formation based on a new reference section (arts. 8e, 19a, and 22c North American Stratigraphic Code, 1985), using the base of the Starfish Bay shale as the base of the Devon Island Formation (art.23c, *op. cit.*). This option would have the greatest utility for geological mapping and for discussions of regional Arctic stratigraphy in that the base of the Devon Island Formation marks a major reorganization of platform deposition in the Canadian Arctic and a significant event in the evolution of the Franklinian basin. The second option, provisionally adopted in this study, is to assign the Starfish Bay shale and overlying silstone and shale ("Devon Island Formation") to the Cape Phillips Formation, until establishment of a reference section containing the Starfish Bay shale. Although the Starfish Bay shale is a distinct and mappable unit, elevation to formation rank would not necessarily clarify stratigraphic relationships (art.25a, *op. cit.*, 1985).

The abrupt Ludlow age contact between the "Devon Island Formation" and underlying carbonates remains an important "event" horizon and signifies pronounced platform backstepping (Fig. 18) that is recognized on southern Ellesmere and northern Devon islands. Following this, several distinct lithostratigraphic units in the southern Arctic signify Caledonian epeirogenesis and concomitant reorganization of platform facies: (i) shallowing, initial deposition of clastics, and subsequent carbonate platform progradation, as represented by the Barlow Inlet Formation on Cornwallis and Devon islands (Thorsteinsson and Mayr, 1987, Thorsteinsson and Uyeno, 1980); (ii) significant epeirogenesis (Boothia Uplift) and syntectonic clastic deposition of the Peel Sound Formation (some 700m thick on Somerset and Prince of Wales islands (Miall, 1986, 1970a, 1970b); (iii) reorganization of an otherwise relatively homogeneous platform facies, and deposition of clastics and carbonates assignable to the Drake Bay and Somerset Island formations on Prince of Wales, Somerset, and Bathurst islands (Thorsteinsson and Uyeno,1980; Mayr, 1978).

Another distinct characteristic of the Devon Island Formation on Devon Island and its equivalent rock units near Vendom Fiord, is the condensed nature of its basal sequence. On northwestern Devon Island in the vicinity of Sutherland River (Thorsteinsson and Uyeno, 1980; Thorsteinsson and Mayr, 1987, p.83), the upper Ludlow and all of the Pridoli are represented by only 30.5m of strata. A similar condensed sequence was recognized north of the mudmound at section 79 (Fig. 18), where *siluricus* Zone conodonts (identification by T.Uyeno) were extracted from the uppermost Douro-like rubbly limestone beds resting on the massive dolomitized Silurian shelf-edge facies. The basal 10m of overlying shale at three separate localities 1-2km apart, yielded graptolites of uppermost Pridoli age (*birchensis* Zone, identification confirmed by A.C. Lenz.). Hence, the upper Ludlow and the entire Pridoli occur within the first ~10m of strata. However, in section 69, some 10km to the south, coeval strata bracketed by the *progenitor* to *uniformis* graptolites zones are more than 150m thick (Fig. 18).

Interpretation and discussion of the Cape Phillips Formation (including the "Devon Island Formation" and Starfish Bay shale)

Rhythmically interbedded shale and bioturbated limestone of the *fastigatus* Zone in the Trold Fiord area represent the oldest record of the Cape Phillips Formation. These were deposited in deep-shelf oxic waters that harboured abundant trilobites, some cephalopods, and brachiopods. The shale beds are regularly interbedded with limestone, suggesting an extrinsic (possibly climatic) influence on carbonate sedimentation, as discussed above.

Subsequent to shallowing and deposition of carbonate beds at or near the Ordovician-Silurian boundary, extensive black chert-rich shale was deposited. In the Trold Fiord area proximal to the drowned Ordovician shelf-margin buildups, the Llandovery part of the sequence is relatively thick and mostly allochthonous; the carbonate was likely derived from the coeval mud buildups through sediment gravity flows and/or through suspension fallout. Llandovery age strata more distal to the mud buildups (section 60) include conspicuous, resistant laminated limestone beds, that were likely bank-derived during major storm events. Associated beds, particularly in the *cyphus* Zone, are very chert rich. This unit likely represents a paleoceanographic significance, related to factors such as upwelling, oxygenation, and increased nutrient levels. Any of these variables may change the physical state of the ocean, leading to enhanced biological activity, the deposition of silicious ooze, and the preservation of bedded (radiolarian) chert (Arthur and Jenkyns, 1981; Maliva *et al.*, 1989; Maliva and Siever, 1988, 1989).

Basin deepening and shale onlapping of platform carbonates near the base of the Silurian were coincident with the widespread deposition of a condensed sequence of black shale and chert (Fig. 22). This appears to have been of global extent, representing an oceanic anoxic event following Upper Ordovician glaciation (Leggett, *et al.*, 1981; Arthur, 1983; Jenkyns, 1980, 1986; Koren, 1987; Schlanger and Jenkyns, 1976). Sluggish paleoceanic circulation coincident with a general ocean warming, decreased dissolved oxygen, and increased surface organic productivity, perhaps resulted in the production of black graptolitic shales (Jenkyns, 1980, 1986; Berry and Wilde, 1978). Minor phosphogenesis during earlier deposition of the Irene Bay Formation possibly represents a converse setting in which enhanced biological productivity was related to more oxygenated bottom conditions and instead caused the release of organic phosphate locked up in the organic remains (Jenkyns, 1976, 1986). The regional time-equivalence of these climatic and/or eustatic events, in contrast to the diachronous nature of tectonically-induced platform flooding events (for example in the upper Llandovery of north Greenland; Hurst and Surlyk, 1984), suggests that climate (and its indirect effects) may have influenced lower Silurian platform evolution.

Silicification in the Cape Phillips Formation was clearly early. Evidence for this is given by (i) the cherty intraclasts in debris-flow deposits; (ii) the uncompressed silicified remains of fossil sponges (de Freitas 1987a, 1989); and (iii) the synsedimentary folds in chert beds. The chert is clearly biogenic, and derived from benthic and possibly pelagic sponges and pelagic Radiolaria (Maliva *et al.*, 1989). Under appropriate paleoceanographic conditions, biological extraction of silica from sea-water by Radiolaria facilitates deposition of silicious (opaline) ooze, of which about 2% remains to form bedded chert (Jenkyns, 1986). Llandovery to Wenlock seas harboured abundant siliceous plankton, particularly during *cyphus* Zone time, which led to widespread formation of bedded chert and also chert in some platform carbonate localities (section 30, for example).

Abundant lithistid sponges and associated pervasive silicification of associated limestone beds have been noted in the Wenlock and Ludlow age pinnacle reef slope facies, and within platform foreslope strata of similar age on Baillie Hamilton Island, Cornwallis Island, and at Bay Fiord. At these localities, sponges representing *in situ* communities and others in gravity flow deposits are well preserved but commonly fragmented in slope strata. Benthic sponges clearly inhabited an oxic platform foreslope environment generally at depths between 50m and 150m, judging from detailed mapping of a well exposed Silurian platform sequence on east-central Cornwallis Island (de Freitas *et al.*, 1989). Rapid burial undoubtedly contributed to the well-preserved condition of the abundant siliceous sponges and to the ubiquitous silicification of associated allochthonous bioclasts. In most examined slope deposits of this age, the close association of sponges and silicification suggests that most silicification was paleobathymetrically constrained to depths of 50-150m, where lithistid sponges flourished.

Silicious sponges in the slope sequences are considered to represent "relict" communities. Whereas sponges were fairly important in Ordovician reef communities (Klappa and James, 1980), their general demise in the Middle Ordovician possibly was related to the appearance of stromatoporoids. Subsequently, in the Silurian, sponges became restricted to relatively deeper water settings where they could effectively compete with stromatoporoids, and, at times successfully build large reefal structures. In the foreslope environment, however, sponges built thin biostromes and led a somewhat opportunistic life style in an environment that was subject to frequent, episodic events of voluminous sediment influx.

Through most of its history, the Cape Phillips Formation was characterized largely by the deposition of dark silty shale; however, in the Trold and Cañon fiord areas, the Wenlock part of the sequence is condensed. Sediment starvation was perhaps associated with abrupt termination of carbonate production on the mud buildups in the Wenlock. The Starfish Bay shale, as discussed below, was deposited relatively rapidly (Fig. 10) and interrupted an otherwise monotonous silty black shale sequence.

The age of the Starfish Bay shale is well established: its base youngs platformward as it onlaps platform carbonates. Although the evidence is equivocal, the age relationships established above, and the greater thickness of the shale toward the west suggest a source in the west for the predominantly terrigenous muds. In contrast to the underlying and overlying siltstone and shale, the Starfish Bay shale is less quartzose, contains more limestone interbeds, and is bioturbated. Furthermore, the Starfish Bay shale does not occur over paleotopographic highs, pinches out toward Grinnell Peninsula, and is not recorded on Cornwallis Island. Perhaps the prominent bight in the platform at Grinnell Peninsula obstructed transport of the terrigenous muds now represented by the Starfish Bay shale.

As discussed above, the base of the so-called Devon Island Formation on Ellesmere Island may eventually be placed at the base of the Starfish Bay shale. This would logically involve formal designation and description of a reference section; however, in the interim, the so-called Devon Island Formation is included provisionally in the Cape Phillips formation. The base of the Devon Island Formation on Devon Island and of the so-called Devon Island Formation near Baumann Fiord is apparently diachronous and signifies progressive southward shale onlapping, suggesting lithospheric flexure such as is commonly associated with tectonism and loading of the lithosphere by sediment and/or thrust sheets (Allen and Allen, 1990). Lithospheric flexure has been inferred for the late Llandovery termination of the North Greenland carbonate platform (Hurst and Surlyk, 1984) and for the areal restriction of the mud buildups at Cañon Fiord during the late Llandovery (as discussed above). In both areas, flexure can be related to lithospheric loading by flysch deposition, but in the case of the diachronous base of the so-called

Devon Island Formation, this can only be inferred. The nearest, oldest flysch sequence, exposed at Trold Fiord, is probably Late Silurian or Early Devonian in age (Trettin, 1978). Flysch deposition could have been earlier in the adjacent Hazen Trough, as eastward onlap of the deep-marine clastics could have been obstructed by relief associated with the Ordovician shelf margin and, in some areas, with the large mud buildups. Based on stratigraphic relations established in this study and by Trettin (1978), the trough-to-drowned-platform relief may have been about 2km and could have been an effective barrier to trough expansion. Based on the available stratigraphic evidence, the lower Ludlow phase of platform backstepping was diachronous, and this may have been was related to loading of the lithosphere during renewed Caledonian tectonism.

The origin of the Starfish Bay shale is uncertain, but probably it was deposited much faster than the underlying Wenlock condensed sequence, and has a contrasting lithology (see Fig.10). The generally low organic content and the evidence of bioturbation, indicate amelioration of bottom environments. The lack of graptolites (which were otherwise extremely abundant during this part of the Ludlow) may indicate increased scavenging and/or degradation of the graptolites, facilitated by improved bottom circulation.

In the overiying so-called Devon Island Formation, fine siltstone laminae deposited primarily by suspension fallout over organic-rich terrigenous muds perhaps signify a return to sluggish bottom circulation and anoxic bottom conditions. The scarcity of abundant organisms having planktonic carbonate tests during the Silurian necessitates an allochthonous platform source for the very regularly interbedded carbonate. Deposition of this sediment from suspension is evident from the even, parallel laminae in these beds, which were perhaps later thickened by diagenetic unmixing. Although off-bank carbonate transport can be facilitated by sea-level lowstands (Sarg, 1988; Droxler and Schlager, 1985), evidence for significant sea-level fluctuations in coeval platform carbonates in the Arctic is lacking. Perhaps variations in climate induced changes in storm intensity which in turn influenced the transport of platform carbonate to the basin. A similar cyclical variation in Triassic periplatform strata was ascribed to fluctuating sea level on the platform (Goldhammer and Harris, 1989). Although Pridoli and Lower Devonian platform cyclicity is pervasive, it shows an upward increase in frequency that is not reflected in the coeval shale-limestone succession. Clearly, sea-level fluctuation is not the only mechanism for generating variations in off-bank transportation, as suggested by some workers (Goldhammer and Harris, 1989; Sarg, 1988).

The so-called Devon Island Formation is coeval with the Goose Fiord Formation on southern Ellesmere Island, the Barlow Inlet and Sophia Lake formations on Cornwallis Island, and with the Red Canyon River Formation on central Ellesmere Island. Pridoli to Devonian carbonate platform deposition occurred during a tectonically active period, when basement blocks were uplifted and fringed by syntectonic clastics. However, the so-called Devon Island Formation silts show relative uniformity throughout the region, suggesting instead stability in the source and/or transport mechanism. Much of this detritus have been wind transported from the craton, or derived from the Caledonian uplift which was the likely source for the thick clastics assignable to the Red Canyon River Formation. This important sequence is discussed separately below.

RED CANYON RIVER FORMATION AND ITS CORRELATIVES

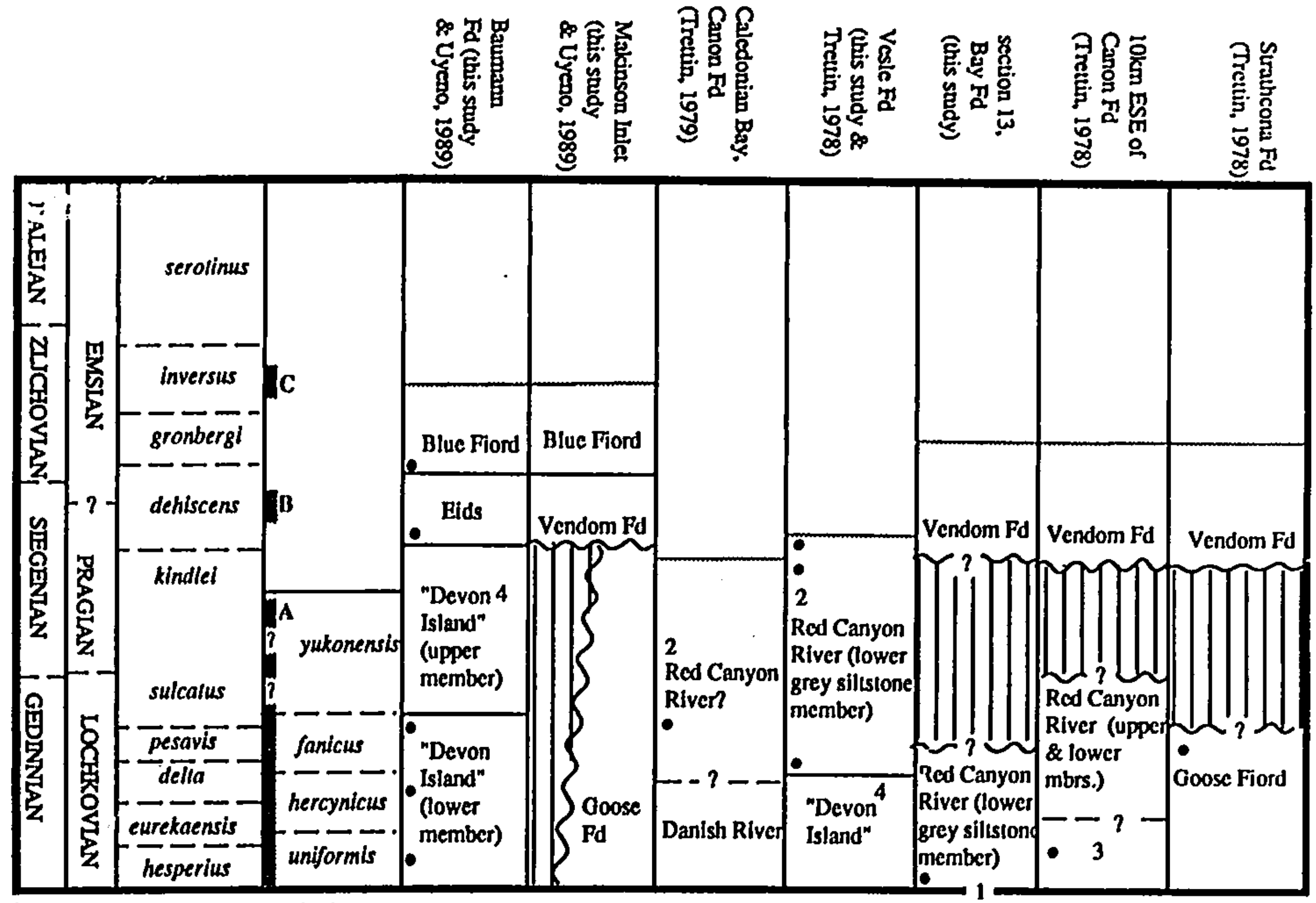
The Red Canyon River Formation in the type section southeast of Canyon Fiord (Trettin, 1978), is 1764m thick and divisible into two members (Figs. 23,24): a lower, predominantly green siltstone and limestone member and an upper, red, gypsiferous sandstone and siltstone member. This clastic sequence is progradational, and represents the evolution of marine to littoral environments during the uppermost Silurian and Lower Devonian (Trettin, 1978). Biostratigraphic information about this formation is sparse, but spores from the type section indicate a Lochkovian age.

The Eids Formation was named by McLaren (In Fortier *et al.*, 1963) on southern Ellesmere Island. Recent work by Roblesky (1979) and Uyeno (1989) have clearly demonstrated that the Eids Formation rests entirely within the *dehiscens* Zone in the vicinity of central and southern Ellesmere Island and is thus a fine-grained correlative of the Vendom Fiord Formation. Based on these observations, and on the stratigraphic relationships indicated in Figures 20 and 21 of this report, it appears that the Eids Formation reported at Vesle Fiord by Trettin (1978) is a significantly older clastic-carbonate sequence than the Eids Formation established by McLaren (*In* Fortier *et al.*, 1963).

Many of Trettin's (1978) sections need to be revisited to confirm the stratigraphic relationships proposed in this study, but it appears that his "Eids Formation" at Vesle Fiord is correlative with the grey siltstone unit at section 13 (Plate 9d). These siltstones are interbedded with autochthonous and allochthonous carbonate beds and have a diachronous lower contact with the so-called Devon Island Formation, a contact that youngs basinward (Figs. 20 and 21). This very distinctively weathering grey siltstone unit pinches out to the south and is not observed at Baumann Fiord. Other so-called "Eids Formation" stratigraphic sections described by Trettin (1978) generally resemble the grey siltstone unit identified at Vesle Fiord and Bay Fiord (this study), but become more carbonate-rich southward, and their established age ranges also are similar (Figs.20,21). These relationships suggest that the Late Silurian to earliest Devonian siltstone sequence at Vesle and Bay Fiords is a distal, fine-grained facies of the Red Canyon River clastic wedge. Directional measurements of sedimentary structures are needed to test this interpretation.

This thick clastic unit interfingers with the coeval platform sequence of the

Figure 23: Chart showing main Upper Silurian and Lower Devonian rock units recognized on Ellesmere Island. Biostratigraphy based on Cherkesova (1988), Thorsteinsson and Uyeno (1980), and Uyeno (1989).



1 Base of this unit at this locality is lowest Devonian or Upper Silurian

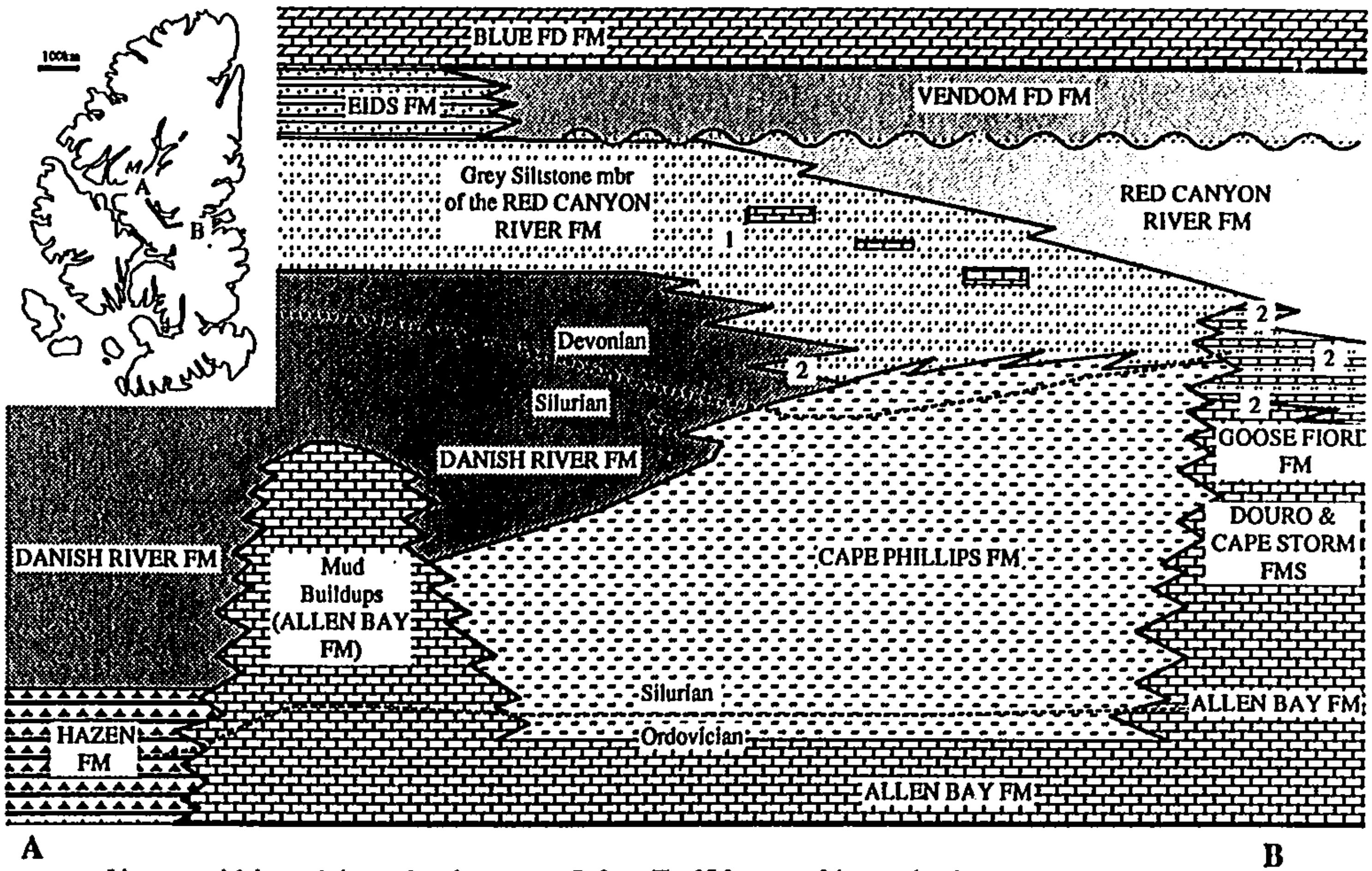
2 This unit was referred to as the Eids Fm by Trettin (1978), but, based on preliminary data of this investigation, it is regarded as belonging to the Red Canyon River Fm..

3 The "Eids Formation, as referred to by Trettin (1978) at this locality, is likely assignable to the lower grey siltstone member of the Red Canyon River Formation, as recognized at Vesle Fiord (also see Fig.15)

4 See discussion in text of this formation

Main phases of the Inglefield Uplift

Figure 24: Schematic reconstruction showing main Upper Silurian and Lower Devonian facies patterns on central Ellesmere Island. Some data are incomplete and correlations, as a result, are speculative. Drawing is not to scale and generally greatly exaggerated vertically.



1 Limestone-rich intervals in an otherwise clastic-dominated succession

2 These parts of the sequence are poorly known, and stratigraphic relationships are conjectural.

Refer to Fig. 25 for ages of these rock units

Legend as in Fig. 6

Goose Fiord Formation on southern Ellesmere Island. Also, some carbonates associated with these probable distal facies of the Red Canyon River Formation may have been local accumulations associated with an otherwise clastic-dominated sequence. This clastic sequence probably represents the first and largest of the three "pulses" that formed the Inglefield Uplift and may have been as early as late Pridoli (Fig. 21, 23, 24). Ultimately, much of the fine clastic rock in the Goose Fiord Formation and the upper part of the so-called Devon Island Formation may be shown to have been derived from this first-phase epeirogenic episode.

PLATFORM EVOLUTION

The general evolution of the Silurian carbonate platform in the Canadian Arctic has been discussed by Trettin and Balkwill (1979), Trettin (1979), Thorsteinsson and Mayr (1987), Thorsteinsson and Uyeno (1980), and Mayr *et al.* (in prep.). These workers established the basic stratigraphic framework for the Silurian in the Canadian Arctic. The following discussion integrates new data presented in this report with their observations and with information from the North Greenland sequence. For convenience, the discussion is divided into six parts, an interpretation of the underlying Ordovician platform, and interpretations of five major events that include episodes of submergence (drowning), upbuilding, backstepping, and reorganization of platform facies.

The Ordovician platform

The Caradoc Thumb Mountain Formation, underlying the Irene Bay Formation, is a thick limestone unit deposited below wave base within the photic zone on a gradually subsiding ramp. The formation is generally spatially and temporally very homogeneous, suggesting stability of depositional environments. Although the facies change to basinal deposits is not known in the study area, the formation grades westward into argillaceous Irene Bay-like strata at Trold Fiord, suggesting that shelf-margin subsidence was greater and that the argillaceous limestone facies is a deep-water equivalent of the clean mottled dolomitic limestone exposed platformward. The Irene Bay Formation possibly represents a phase of maximum submergence, an event that can be correlated through much of the Danish and Canadian Arctic. Because subsidence was greatest basinward (*sensu* Pitman, 1978), argillaceous lime muds grade distally into terrigenous muds. The top of the Irene Bay Formation is thus considered an "event" horizon, and is ideal as a datum in considering the subsequent platform and basinal sequence.

Although a relative change in sea level likely affected deposition of the Irene Bay Formation, the regional lithofacies variation of the latter and of the underlying Thumb Mountain Formation in shelf-margin localities, could be explained by paleoclimatic factors for example, by flooding the platform by relatively cold waters. The associated, distinctive Arctic-Ordovician fauna, in this respect, may represent a cold water fauna (*sensu* Brookfield, 1990). Upwelling in modern oceans is commonly associated with phosphogenesis, faunal abundance, and cold ocean currents (Jenkyns, 1986) and may conveniently explain some of the features observed in the Irene Bay Formation.

The Irene Bay Formation also marks the first, post-Cambrian onlapping of deep-water facies onto platform carbonates. Subsequently, clean lime muds were again deposited (basal tongue of the Allen Bay Formation), but apparently in deeper water than during deposition of the Thumb Mountain Formation. The basal Allen Bay tongue is regionally extensive, from Trold Fiord to western Grinnell Peninsula and to parts of Cornwallis and Bathurst islands. Cratonward, however, the Allen Bay Formation consists more typically of mottled dolomitic limestones that are much more like those of the Thumb Mountain Formation. A subsequent and regionally synchronous deepening during *fastigatus* Zone time gradually terminated the extensive platform as shale onlapped platform carbonates.

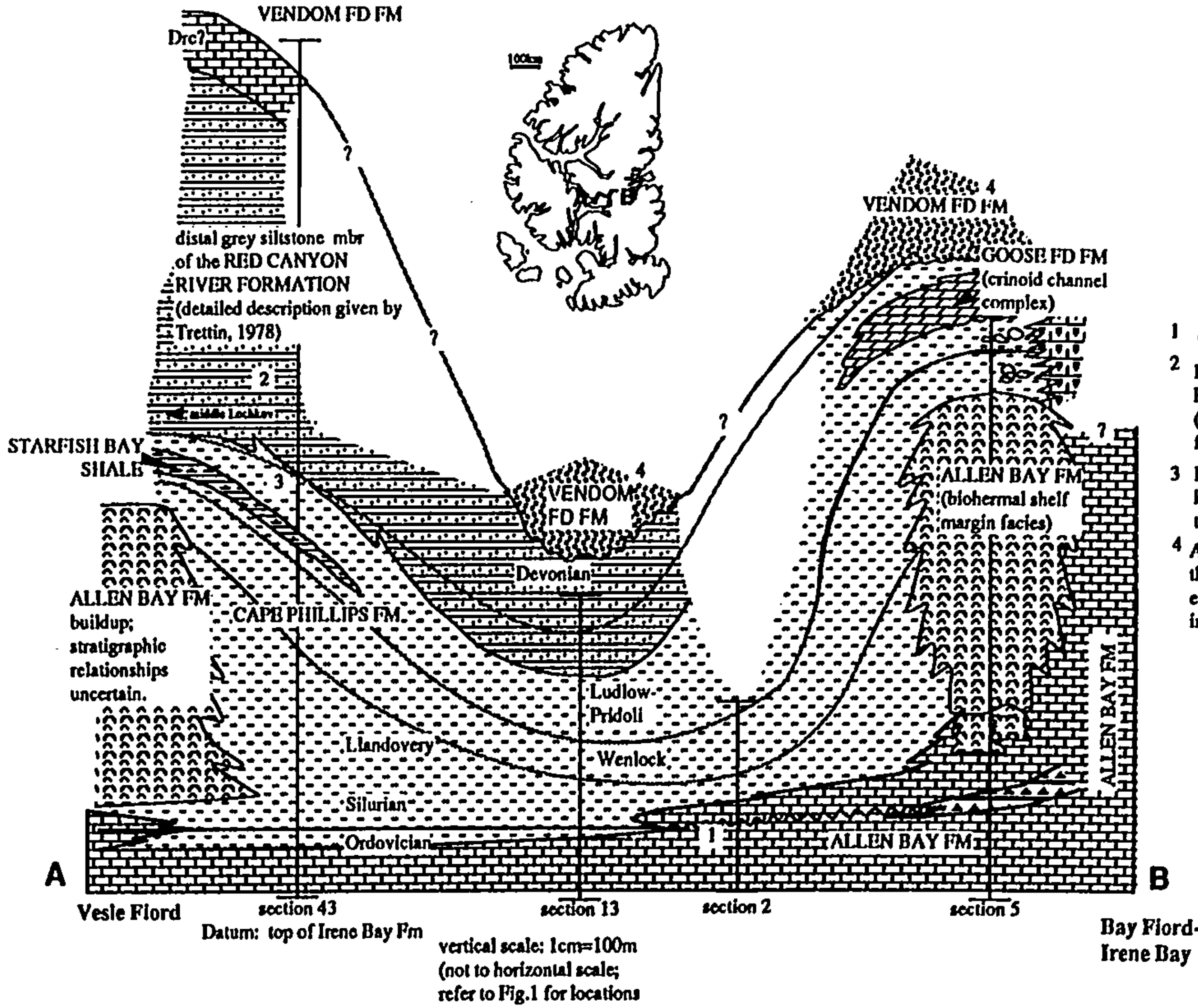
Event 1: *fastigatus* Zone platform drowning

A thin, regionally extensive basal limestone tongue of the Allen Bay Formation overlies the Irene Bay Formation in most deep-shelf localities. This limestone is interbedded with shales, and was likely deposited in deeper water than the underlying mottled dolomitic limestone of the Thumb Mountain Formation, suggesting that conditions were not optimum for carbonate accumulation immediately following deposition of the Irene Bay Formation. In this somewhat "sensitized" state, the carbonate platform was prone to drowning during subsequent sea-level rises, such as is inferred to have occurred in the *fastigatus* Zone. Platform drowning was gradual, and isolated carbonate mud buildups were established over the rapidly subsiding shelf margin. Contemporaneous platform carbonates were re-established cratonward (Figs. 18, 22, 25,).

Reconnaissance studies show that the Irene Bay Formation at Bay Fiord and the Thumb Mountain Formation at southern Trold Fiord are locally less argillaceous and less rubbly weathering where they stratigraphically underlie large drowned shelf-margin mud buildups, in contrast to paleogeographically adjacent localities. This relationship suggests that the large mud buildups were preferentially located where the Irene Bay and upper Thumb Mountain formations were of somewhat shallower shelf character. This possibly reflects underlying control, perhaps related to differential subsidence of discrete basement blocks. Similar basement control of platform morphogenesis has been inferred for other ancient and recent carbonate platforms (Wilson, 1990; Davies *et al.*, 1989) and also apply to this platform sequence in the Canadian Arctic.

Information from sporadic exposures and some seismic surveys, suggests that the Ordovician shelf margin was relatively straight, and its position was probably static through much of the Ordovician and some of the Cambrian. The shelf margin re-established in the Silurian subsequent to the *fastigatus* Zone drowning was

Figure 25: Diagrammatic restored section of the main platform and deep-shelf stratigraphic units near Bay Fiord. Legend as Figure 6d. See Figure 4 for section locations.



1 condensed sequence

2 Biostratigraphy of this member of the Red Canyon R. Fm is based on Trettin (1978); all other biostratigraphy is from this study.

3 Biostratigraphy is not sufficiently known for placement of the base of the Pridoli

4 Although not described in detail, this unit was observed overlying examined stratigraphic units indicated by the section line.

considerably more sinuous than in the Ordovician. The conspicuous promontory near Grinnell Peninsula reflects locally decreased subsidence, perhaps related to the Boothia Uplift, although the major sedimentary effects of this structure were considerably younger.

On eastern Ellesmere Island, Silurian cherty shale overlies Ordovician Allen Bay Formation carbonates, and a considerable hiatus is represented by this contact. The contact clearly represents an abrupt platform submergence, but the precise age of onlapping is uncertain: it may have been coincident with the *fastigatus* Zone drowning, or if slightly younger, coincident with the *acuminatus* Zone onlapping as indicated in central Ellesmere Island. There, cherty *acuminatus* to *cyphus* Zone shale and limestone overlie reefal Ordovician platform carbonates, and were subsequently overlain by Silurian platform-slope and platform-margin carbonates. This Late Ordovician event may have coincided with Gondwana deglaciation, which has a well-established global correlation (Brenchley, 1988). Significantly, graptolite biostratigraphy shows that similar onlapping occurred in North Greenland (Hurst and Surlyk, 1983, p.494, Fig. 20) within the *cyphus* Zone, and it is likely that a similar and considerable hiatus is associated with onlapping in Washington Land. Reefs associated with this backstepping, such as the Hans Island structure, and others described by Hurst (1980), can be considered as representing catch-up carbonate sedimentation approximately correlative with the pinnacle reef structures at Baumann Fiord; however, these more northern reefs are coral-stromatoporoid-dominated whereas at Baumann Fiord they are microbialite-coral dominated.

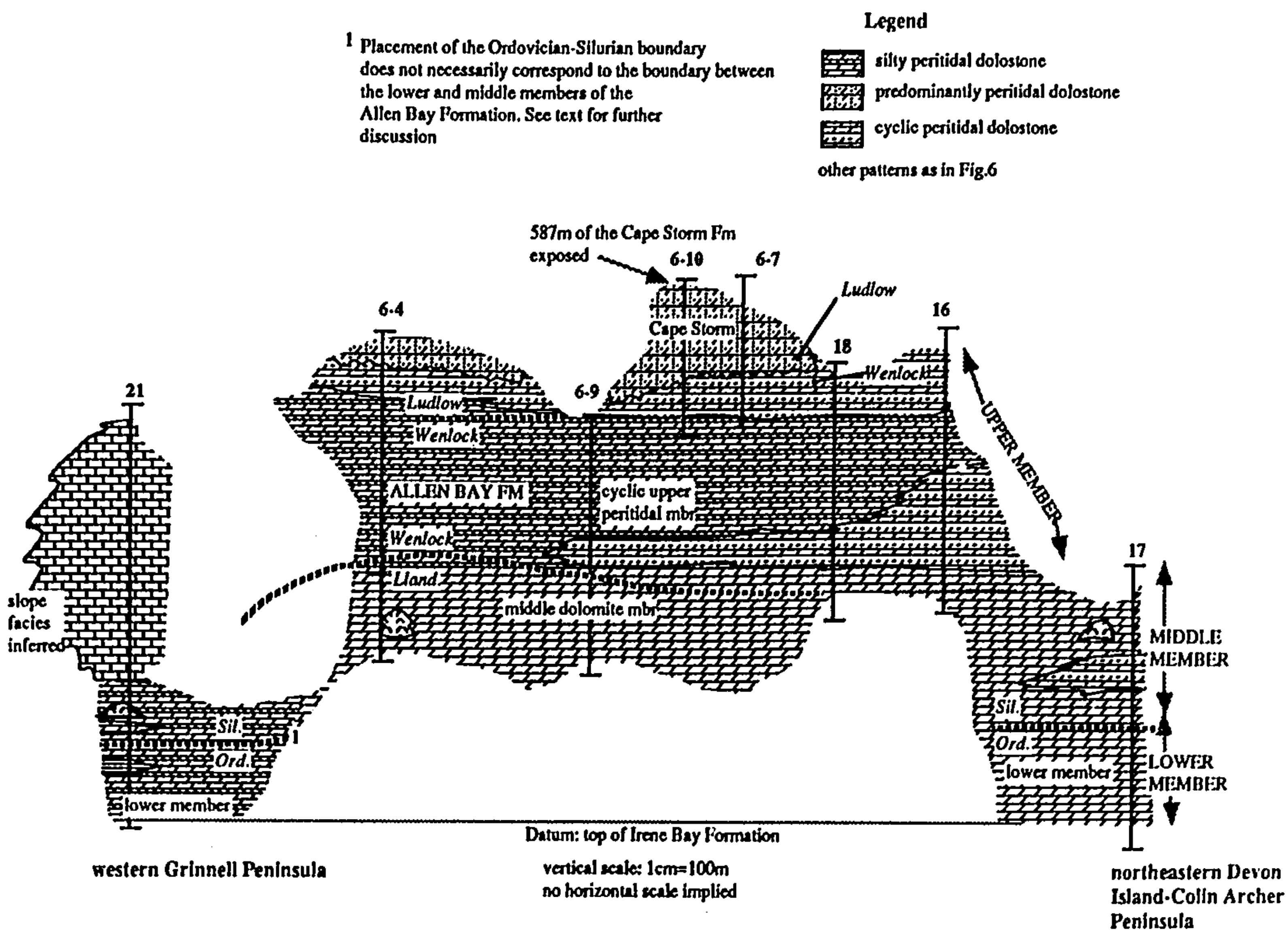
Lowest Silurian cherts, particularly those in the *cyphus* Zone, occur basin-wide and are more abundant toward the position of the drowned Ordovician shelf-margin; however, cherts are uncommon in the Wenlock and are generally unknown in Ludlow and Pridoli shales. Chert deposition generally coincides with shelf-deepening, global high stands of sea level (Johnson, 1987), deglaciation (Brenchley, 1988), and black shale deposition (Leggett, 1980), and may have a regional, if not global paleoceanographic significance.

Event 2: Lower and Middle Silurian platform upbuilding

The lower Llandovery platform margin sequence was marked by sea-level stability. In the Llandovery and Wenlock, platform upbuilding was continuous and generally uninterrupted by major sea-level fluctuations or episodes of tectonism. Initially (lower Llandovery), platform accommodation was minor, and facilitated the development of an island-bay complex on Cornwallis Island (Sodero and Hobson, 1979; Mallamo, 1989) and a similar reef-restricted-lagoon complex on Grinnell Peninsula (this study, sections 20, 21). Restricted shelf-margin environments occurred and are manifested on Cornwallis Island by evaporite pseudomorphs in laminated carbonates and on Grinnell Peninsula by synaeresis-cracked, laminated dolostone, and small stromatoporoid-crinoid reefs. However, shallowing appears to have been limited to these shelf-margin localities, and coeval platform waters remained relatively open (Fig. 26). At Bay Fiord, coeval restricted shelf-margin facies are absent and the shelf-margin was predominantly biohermal, indicating perhaps greater lower Llandovery platform accommodation in this area (Fig. 25).

Platform upbuilding through the Llandovery and Wenlock caused an increase in platform relief, a general restriction of the platform interior marine environments, and an increase in size of allochthonous blocks delivered to the toe-of-slope. On Cornwallis Island, platform margin upbuilding is manifested by an island-lagoon complex which developed into a shelf-margin patch-reef complex on a ramp-to-rimmed platform. On Grinnell Peninsula, although the shelf-margin facies are poorly known, a restricted lagoon-reef complex was succeeded by a lime mudstone-dominated ramp then by an oolitic shelf-margin facies (Fig. 26). At Bay Fiord, accommodation was perhaps greater, and most of the Llandovery and much of the Wenlock sequence appears to have been biohermal (Fig. 25). Llandovery platform interior carbonates on Grinnell Peninsula and southern Ellesmere were deposited on an open marine carbonate ramp and contain ill-defined cycles represented by beds containing zoned brachiopod ecogroups (Zeigler, 1965) interbedded with fossiliferous lime mudstone and tidally influenced laminated carbonates. The regional significance of these units, however, and the allogenic or autogenic influences on their occurrences are unclear.

Figure 26: Diagrammatic restored section of the main platform, deep-shelf, and trough stratigraphic units on Grinnell Peninsula and northern Devon Island. Legend as Figure 6d. See Figure 1 for section locations.



1 Placement of the Ordovician-Silurian boundary does not necessarily correspond to the boundary between the lower and middle members of the Allen Bay Formation. See text for further discussion
Legend
silty peritidal dolostone
predominantly peritidal dolostone
cyclic peritidal dolostone
other patterns as in Fig.6
587m of the Cape Storm Fm exposed
6-10
6-7
Ludlow
16
18
Wenlock
6-4
6-9
21
Cape Storm
Ludlow
Wenlock
ALLEN BAY FM
cyclic upper peritidal mbr
UPPER MEMBER
Wenlock
Lland.
middle dolomite mbr
17
slope facies inferred
MIDDLE MEMBER
Sil.
Ord.
lower member
LOWER MEMBER
Sil.
1
Ord.
lower member
Datum: top of Irene Bay Formation
western Grinnell Peninsula
vertical scale: 1cm=100m
no horizontal scale implied
northeastern Devon Island-Colin Archer Peninsula

Platform carbonates in the area of Baumann Fiord had a slightly different development during the Lower Silurian. The *fastigatus* Zone drowning is not recorded nor is any obvious shallowing evident near the Ordovician-Silurian boundary; however, crucial boundary strata are either inaccessible or poorly exposed, and evidence for sea level variations may be present. Of note in this area is a chert-rich, shale-limestone unit about 40-60m thick. It occurs over an area about 500km^2 and is a distinctive, dark-coloured, recessive marker unit on air photos. Graptolites from this unit , although poorly preserved, belong to the *cyphus* to *convolutus* zones, thus are middle Llandovery in age. The lithology and age of the unit is similar to *cyphus* Zone chert beds recorded in many areas of Ellesmere Island, and this local backstepping event may be related to a climatic change and/or a rise in sea level. Subsequently, a catch-up carbonate sequence was deposited, and pinnacle reefs and downslope mudmounds developed. Three or possibly four of these structures occur in the area of Baumann Fiord, and they are approximately correlative with the Hans Island reefal structure (this study) and the numerous structures termed intrashelf buildups by Hurst (1980) on North Greenland. Where these structures had a great rate of carbonate accumulation, a tripartite facies zonation is evident comprising a basal, sparsely fossiliferous lime mudstone facies, a middle microbial carbonate facies, and an upper coralgal biolithite-oolite facies, representing overall upward shallowing. In areas of subsidence or suppressed carbonate accumulation, a continuous microbial carbonate sequence was deposited prior to middle Ludlow drowning. Some of the intrashelf buildups or pinnacle reefs of North Greenland are similarly zoned, and have mud-rich cores (Hurst, 1980; Sønderholm and Harland, 1989), but have stromatoporoid-coral caps, thus are unlike the mud-cored, coralgal biolithite-capped reef pinnacles near Baumann Fiord.

Coeval shelf and shelf-margin carbonates on Cornwallis Island developed from a restricted island-bay complex to a patch-reef complex, to finally, during the Ludlow (Cape Storm Formation), a stromatoporoid grainstone reef complex. A possible similar development at Baumann Fiord, is less well exposed. Platform relief for the entire Silurian appears to have been at a maximum during deposition of the Cape Storm Formation, as blocks derived from the shelf margin are most abundant and coarsest. Blocks from the coeval pinnacle reefs are also very large and are clearly related to major upbuilding episodes rather than to sea level low stands.

The Cape Storm Formation is a very extensive peritidal platform sequence that concluded an otherwise predominantly open marine sequence as indicated by the lower and middle members of the Allen Bay Formation. The upper dolostone member of the Allen Bay Formation is predominantly peritidal, but it is not as extensive or as thick as the Cape Storm Formation which attains a maximum thickness of 587m thick on northern Devon Island. The thick Cape Storm Formation platform margin facies are characteristically crinoid-skeletal grainstones, representing a shoal complex. The development of this shelf-margin facies may have been intrinsically controlled and partly related to keep-up shelf sedimentation, low platform accommodation, the development of extensive tidal flats, and the consequent accumulation of inimical platform interior waters. The last factor perhaps had the greatest influence on observed facies patterns. Although there may be other influences to consider, shelf-margin reef growth is generally inhibited where there is off-bank movement of (presumably inimical) platform interior waters on the recent Bahamian platform. In these areas, carbonate sand shoals typically develop, and are considered by some workers (Mark Boardman, pers. com. 1988; Harris, 1979) to be a "default-style" of sedimentation. In this Silurian example, a similar facies pattern may have developed due to the predominantly peritidal platform interior setting that developed over a large area.

Event 3: upper Llandovery platform backstepping (North Greenland)

In the area of North Greenland, a major platform drowning event occurred in the latest Llandovery. Platform flexure related to turbidite deposition and Caledonian nappe emplacement in eastern North Greenland caused extensive platform termination (Surlyk and Hurst, 1984). Flysch onlap and contemporaneous pinnacle reef development was a result. However, a thick succession of predominantly subtidal, open marine Allen Bay Formation platform carbonates was deposited in the area of Darling Peninsula, which contrasts to the generally upward shallowing carbonate platform sequence in the southern Arctic Islands. This development was likely due to greater platform accommodation in North Greenland, perhaps related to syntectonic clastic deposition and crustal flexure. Llandovery reefal structures in Kennedy Channel are probably local developments and are correlative with similar reefs on North Greenland. On Hans Island, a probable isolated deep shelf pinnacle reef apparently developed subsequent to major middle or late Llandovery platform drowning in North Greenland, and represents a catch-up phase of carbonate deposition. Coeval reefs in the vicinity of Kennedy Channel were probably still active during early stages of flysch deposition, as suggested by olistostromes interbedded with sandstone turbidites in northeastern Ellesmere Island (Fig. 22).

Other than the occurrence of subtidal carbonates in the upper part of the Allen Bay Formation on Darling Peninsula, there is no evidence of termination of platform sedimentation due to downloading south of Cañon Fiord. However, as discussed in Part III, the constriction of the mud buildups at Cañon Fiord and onlapping of slope carbonates onto microbial carbonate may be a local development related to flexure and voluminous flysch deposition in the contiguous Hazen Trough. The strongly diachronous nature of the lower contact of the flysch sequence, which ranges from upper Llandovery to Lower Devonian over a distance of less than 10km, indicates that the primary depositional relief of the Ordovician platform margin and overlying mud buildups (perhaps up to 2km) was a major influence on deposition until the Lower Devonian, even though carbonate deposition on the mud buildups was terminated in the lower Wenlock.

Llandovery and Wenlock platform carbonates are disconformably overlain by Cape Storm Formation peritidal dolostone on Devon, Cornwallis, and Ellesmere islands. This unit records a minimum Middle Silurian platform accommodation in the Arctic. Coeval upbuilding and huge olistoliths are associated with the Wenlock slope exposure at Baumann Fiord (Plate 4), but it is uncertain whether allochthonous sediments were platform or pinnacle reef derived. Wenlock strata at Bay Fiord contain coarse olistostromes, and significant slumping and contemporaneous erosion of the shelf margin is recorded through the Wenlock, Ludlow, and part of the Pridoli. However, the deep-shelf Wenlock sequence is condensed and carbonate-poor, suggesting effective termination of the allochthonous carbonate supply due to drowning of large mud buildups in the areas of Vesle and Trold fiords. Although platform carbonates were still actively accumulating to the east, they probably had little influence on sedimentation in these areas.

The Wenlock shelf margin, of east-central Cornwallis Island, and the Bay Fiord and Baumann Fiord areas, was evidently rimmed. A carbonate sand shoal-reef complex over the shelf margin on Cornwallis Island and near Baumann Fiord delivered large allochthonous blocks to the contiguous toe-of-slope. Platform relief was at a maximum and may have been as much as 200m (de Freitas *et al.*, 1989).

The predominance of reefal shelf-margin facies at Bay Fiord signifies the persistence of a distinct shelf-margin development during the Lower and Middle Silurian, perhaps related to a greater rate of subsidence.

Event 4: Ludlow carbonate platform reorganization

Deepening through the entire Arctic Archipelago, following deposition of the Cape Storm Formation, is marked by deposition of rubbly limestones on a carbonate ramp (Douro Formation). During this time, numerous reefs and diverse shelly faunas flourished. The Douro Formation is wholly argillaceous on Cornwallis and Devon islands, but is only partly so and partly represented by mottled dolomitic limestone on southern and northeastern Ellesmere Island (this study) and adjacent to the Boothia Uplift on Somerset and Prince of Wales islands (Thorsteinsson and Uyeno, 1980; Mortensen and Jones, 1986; Jones and Dixon, 1977). Mortensen and Jones (1986, p.1403) interpreted this lithology as a shallower water facies of the rubbly weathering limestone; hence, the distribution of the mottled dolomitic limestone facies of the Douro Formation reflects shoaling marginal to the Boothia (Mortensen and Jones, 1986) and Inglefield uplifts (this study). Subsidence was greatest in Cornwallis Island, and deep-shelf shales onlap nodular Douro limestones, and the mottled dolomitic limestone facies is absent.

At Baumann Fiord, a thin succession of Douro-like beds is likely assignable to the Douro Formation known in the southern Arctic, but is probably slightly older, and represents initial deepening on the shelf margin prior to shale onlapping in the middle Ludlow. Shale onlapping on Devon and Ellesmere islands, represented by the base of the Devon Island Formation and equivalent shales, was diachronous, and took place earlier in the Baumann Fiord area than in the vicinity of northern Devon Island. The relationship suggests flexure, perhaps due to downloading and renewed (possibly Caledonian) tectonism; however, any direct evidence for this interpretation is buried beneath the Sverdrup Basin sequence. A major phase of epeirogenesis of the Boothia Uplift occurred to the south at about the same time as Ludlow platform backstepping (Thorsteinsson and Uyeno, 1980; Okulitch, *et al.*, 1986), but the relationship of this uplift to essentially coeval yet slightly diachronous platform drowning is unclear. Emergence of the Boothia Uplift resulted in deposition of the Peel Sound Formation alluvial clastics and pronounced carbonate platform progradation (the Barlow Inlet Formation on Cornwallis Island). Perhaps uplift involved reciprocal tilting of the crust such that subsidence to below the zone of optimal carbonate production in areas more distal to the Boothia Uplift resulted in the diachronous onlapping relationships observed at the base of the Devon Island Formation. However, the latter conjecture does not readily explain the more distal, older onlapping relationship near Vendom Fiord.

Subsequent to drowning, green-grey claystone of the Starfish bay shale onlapped drowned platform carbonates in the Baumann Fiord area. The diachronous lower contact and the onlapping relationship suggests a westerly source area for the mudrock. Although the origin of these strata is enigmatic, they represent a pronounced increase in rate of sedimentation (Fig. 10) and an interruption in deposition of an otherwise monotonous silty shale sequence, in the Trold and Bay fiord areas.

Event 5: Pridoli platform development and the Inglefield Uplift

This part of the sequence on Ellesmere Island is poorly known, and is still under study by the author and several workers at the Geological Survey of Canada. The carbonate platform sequence in the vicinity of Baumann Fiord shows overall upward shallowing. The entire Pridoli and Lower Devonian sequence is cyclic, and at the top shows evidence of exposure, clastic input, and predominantly tidalite deposition. The overall succession represents decreasing platform accommodation, and progradational geometry in central and southern Ellesmere Island; however, in the vicinity of Baumann Fiord the inferred platform margin remained relatively static until the Vendom Fiord alluvial-to-marine clastics and coeval Eids Formation siltstone were introduced by the Inglefield Uplift. At Bay Fiord, latest Pridoli progradational ramp geometry and shelf margin encrinite channel complexes may be a reflection of decreased platform accommodation and platform progradation perhaps related to an earlier phase of the Inglefield Uplift.

Middle Ludlow backstepping was followed by a second phase of pinnacle reef development (Figs. 12, 13) that was much more widespread than the Llandovery and

Wenlock phase near Baumann Fiord. The Ludlow and Pridoli reefs are very similar to the earlier reefs in having mud cores overlain by stromatoporoid-coral caps deposited in shallow water. However, mud cores in the younger reefs contain common stromatactis structures, while in the older phase these structures are absent. Many of these pinnacle reefs have not been studied in detail, but reconnaissance indicates that subsequent to Ludlow drowning, considerably condensed sequences accumulated over these structures. Similarly, condensed sequences occur over drowned Wenlock and Ludlow pinnacle reefs at Baumann Fiord, where significant hiati are recorded, and the top of one of the reefs has been extensively corroded and mineralized. These condensed sequences are associated with expanded sequences in the intervening areas between the reefs (for example expanded sections occur in sections 69, 70a,b, Fig.18). A continuous spectrum of facies appears to be represented in the examined mud buildups, including mudmounds, pinnacle reefs, and large mud buildups established over the Ordovician shelf margin (see Fig. 16). This variation in the vertical lithofacies succession among the three distinct types of mud buildups appears to be reflection of paleoenvironment and different rates of subsidence and/or upbuilding.

Local variations in sediment thickness are particularly apparent in some localities (Fig. 10). For example, in localities proximal to reef pinnacles (sections 78, 80 and 54, Fig. 10) and to mud buildups (section 56 to 59; Fig. 10) sedimentation was relatively rapid (compare sections 56-59 to section 60) and was probably related to rates of upbuilding and concomitant delivery of allochthonous carbonates to the toe-of-slope. Similarly, flysch sedimentation was extremely rapid (section 30 and section 1; Fig. 10) and likely caused lithosphere flexure and platform and mud buildup drowning in North Greenland (Surlyk and Hurst, 1984) and in central Ellesmere Island (this study).

Pridoli platform development is less well known. As discussed above, pronounced progradation occurred in the south, in the vicinity of Grinnell Peninsula and Cornwallis Island, but not in the area of Baumann Fiord. In the latest Silurian, the first phase of the Inglefield Uplift in north-central Ellesmere Island produced a thick prograding clastic wedge (Red Canyon River Formation), and distal fine-grained grey siltstone facies that abruptly overlie relatively starved deep-shelf shales in the areas of Vesle Fiord and Irene Bay. Carbonates likely interfingered from the south or occurred as lenses within the clastic succession. In more distal areas, the Hazen Trough sequence shows evolution from an underfill sequence (Danish River Flysch) to an overfill sequence (distal fine-grained facies of the Red Canyon River Formation), but this area is less well known by the author. To the south and partly coeval with this clastic succession, is a thick carbonate sequence (Goose Fiord Formation). It was spared much of the clastic input, but was uplifted in the Emsian (*dehiscens* Zone) in the penultimate phase of the Caledonian Inglefield Uplift. Syntectonic clastics, represented by the Vendom Fiord and Eids formations then blanketed most of the study area.

CONCLUSIONS

In this study, several distinctive lithostratigraphic units or facies are described for the first time. Several are shown to have the mappability appropriate for formational status, although no new names have been formally proposed. These main units, or facies, as yet informally designated, are as follows:

(1) in the Allen Bay Formation:

(a) lower, middle, and upper members (of regional extent) of Upper Ordovician to about lower Ludlow age;

(b) an upper limestone facies (on southern Elesmere Island), age unknown but likely Wenlock to lower Ludlow;

(c) a limestone "shelf margin" facies (on Grinnell Peninsula) of Upper Ordovician to ?Ludlow age;

(d) biohermal shelf-margin facies (near Baumann Fiord) of Llandovery to lower Wenlock age;

(e) stromatoporoid pinnacle reefs (Hans Island) probably of Ludlow age, but incompletely exposed;

(f) pinnacle reefs and associated mudmound and reef-flank facies (in the Baumann Fiord area) of Llandovery to middle Ludlow age;

(g) mud buildups established over the drowned Ordovician shelf margin and associated slope facies (three areas on central Elesmere Island) of Upper Ordovician to lower Wenlock age.

(2) in the Cape Phillips Formation:

(a) an undifferentiated unit consisting of a condensed sequence of Wenlock to Lochkovian age fissile shale over drowned mud buildups, a Llandovery age chert-rich sequence, and an uppermost Llandovery to Wenlock age siltstone and shale sequence; a fourth monotonous dull grey siltstone sequence occurs locally near Cañon Fiord, but its age is poorly known;

(b) dolomitic or calcareous mudrock unit informally termed Starfish Bay shale;

(3) in the Red Canyon River Formation:

(a) grey siltstone facies (near Vesle and Bay fiords) of latest Silurian to Lochkovian age.

Several regionally correlative "events" influenced Late Ordovician and Silurian carbonate platform sedimentation. The Irene Bay Formation, one of the most widespread formations in the Arctic Islands, represents either a deepening event or possibly a climatic change that influenced conditions on the carbonate platform. Deposition of the Irene Bay Formation was followed by major platform drowning in the *fastigatus* Zone that ended carbonate deposition over a large area and led to the re-establishment of the platform margin several kilometres eastward of its

Ordovician position. Basin-wide sea-level fluctuation(s) occurred at or near the Ordovician-Silurian boundary, but the evidence is obscured by condensed sequences that lack some biozones, by incomplete exposures, and by difficulty of access to some critical sections. Their correlation with well recognized, Hirnantian glacio-eustatic sea level fluctuations (Brenchley, 1988) is unknown. Nevertheless, the apparently synchronous and regional nature of platform flooding events suggests an extrinsic (possibly climatic or eustatic) influence on Early Silurian platform evolution. Chert is relatively common through much of the Llandovery and Wenlock part of the Cape Phillips Formation, but is particularly abundant in the middle Llandovery (Aeronian age, *cyphus* Zone). Chert of this age is regionally extensive, and its formation is coincident with local backstepping in the vicinity of Baumann Fiord that may be related to a global eustatic high stand recognized by Johnson (1987). Platform backstepping in this area was followed by a "catch-up" phase of carbonate deposition. Large reef pinnacles and mudmounds, as a major part of this sequence, were deposited in the vicinity of Baumann Fiord. Three facies are recognized in the pinnacle reefs; a sparsely fossiliferous bioturbated lime mudstone overlain by microbial carbonate then capped by coralgal biolithite-oolite. The last facies is newly interpreted as representing deposition in well-agitated (wave-stressed) environments, while the microbial carbonate represents deposition between storm and fair-weather wave bases. Three distinct types of coeval reef structures are recognized: mudmounds, coralgal biolithite reefs, and coralgal biolithite-oolite reefs. These appear to represent a spectrum of reef development, probably related to differences in upbuilding rate, inferred bathymetry, and the intrinsic form of the buildups during their maximum development.

During the upper Llandovery, profound changes in deposition occurred in response to the evolution of a flexure-controlled foreland basin. The end of platform sedimentation in north Greenland was diachronous with, and related to flexure and syntectonic flysch deposition and nappe onloading (Soper and Higgins, 1987; Hurst and Surlyk, 1984). Late Llandovery (*sakmaricus* Zone) flysch caused initial deepening and local onlapping of slope strata on the microbial carbonate of the Cañon Fiord mud buildup. Also, the great thickness of subtidal carbonate and the development of reefs in the upper part of the Allen Bay Formation at Darling Peninsula, perhaps partly reflects flexure and a greater rate of subsidence in that area. Other than these effects, no other major flexure-controlled influences on carbonate deposition have been recognized in the southern Arctic Islands. The diachronous and areally restricted nature of tectonically induced subsidence contrasts markedly with the probably eustatic, more regional and synchronous events that influenced the Upper Ordovician and Lower Silurian carbonate platform. For example, drowning of the carbonate platform and isolated carbonate structures appears to have been progressive, with initial deepening followed later by complete drowning and deposition of deep-water shale or a condensed sequence. Also of note is the occurrence of cyclic Pridoli platform carbonates during periods of tectonic activity, whereas evidence of platform cyclicity is during glacial periods (as in the Hirnantian) is lacking. This fact is inconsistent with a wholly extrinsic influence (climatic forcing) for carbonate platform cyclicity, as has been hypothesized by many recent workers (Strasser, 1988; Goodwin and Anderson, 1985; Goldhammer, *et al.*, 1987).

After the *fastigatus* Zone platform drowning, isolated areas of lime mud deposition remained over the previous Ordovician shelf margin. Large mud buildups with complex facies associations (see Part III) later developed in these areas. Variation in the Irene Bay and Thumb Mountain formation facies that directly underlie the mud buildups suggests that the buildups above crustal structures had a long-standing significance with respect to their relative rates of subsidence. The large mud buildups on Ellesmere Island were drowned synchronously in the lower Wenlock, and at Cañon Fiord, these buildups remained as prominent topographic features (even though carbonate production had long since ceased) until the onlap of Danish River Formation flysch in the Late Silurian or Early Devonian. A condensed sequence of about 120m of black shale, resting on the mud buildups, includes much of the Wenlock and all of the Ludlow, Pridoli, and Lower Devonian. The reason for drowning is unknown, and no evidence of a coeval influence on the laterally equivalent carbonate platform has yet been recognized. During the Ludlow, a major reorganization of the carbonate platform occurred. The platform flooding that occurred was diachronous and was perhaps related to flexure, due either to the

Boothia Uplift and associated downdropping of basement blocks, or to loading in response to flysch deposition.

Subsequent to local drowning a second phase of pinnacle reef development produced reefs that were more numerous than in the earlier Llandovery to lower Wenlock phase. The younger pinnacle reefs generally show a similar upward development, from a basal stomatactis-rich mudstone core to a stromatoporoid-rich cap.

Upper Ludlow to Pridoli platform evolution is less well known. Platform progradation occurred on Cornwallis and southern Ellesmere islands (Mayr *et al.*, in prep.; Thorsteinsson and Uyeno, 1980) and near Bay Fiord, but not in the vicinity of Vendom Fiord, suggesting an underlying tectonic, possibly fault control. Greater subsidence of the Pridoli-Lochkovian carbonate platform at Vendom Fiord is evident from the change from predominantly subtidal open marine carbonates east of Vendom Fiord to silty peritidal carbonates on southern Ellesmere and northern Devon islands. In central Ellesmere Island, platform progradation in the upper Pridoli was concurrent with a major phase of clastic deposition, producing progradation of the Red Canyon River Formation clastic wedge. This wedge is thickest in the area southeast of Cañon Fiord and interfingers to the south with the Goose Fiord carbonate platform succession which is thickest in the area of Vendom Fiord. The distal grey siltstone facies of the Red Canyon River Formation, in the area of Vesle Fiord, was likely derived from this first phase of the Inglefield Uplift.

A second major phase of uplift essentially terminated carbonate deposition in the study area in the Emsian.

PLATES

Plate 1: Allen Bay Formation exposures. **(A)** Thickly bedded dolostone of middle member of the Allen Bay Formation. Cliff top is about 300m above sea level. Light coloured dolostone in hanging wall of normal fault (arrow) is Cape Storm Formation. Section 6-18 northeastern Devon Island. **(B)** Lowest Silurian (*acuminatus* to approximately *cyphus* Zones) shale interfingering with carbonates. Arrow indicates dolomitized lower Llandovery biohermal strata which are sharply progradational in this area. Section 12, Irene Bay. Also see Figure 8b. About 10m of section is shown. **(C)** Lower beds in stratigraphic section illustrated in 1A, this Plate, consisting of thickly bedded, medium brown dolostone very typical of middle member Allen Bay Formation. Light coloured amorphous patches are globular and tabular stromatoporoids. Notebook (center) is 21cm long. **(D)** Part of oolitic facies of a pinnacle reef near Baumann Fiord. Abundant *Conchidium* brachiopods are typically associated with flank facies of the oolite-coralgal biolithite. Arrow shows bedding parallel, light grey, banded cement, now represented by dolostone; this likely was a neptunian sill filled with early marine cement. Near middle of section 49, Baumann Fiord. Pen knife (top) is about 1.5cm wide. **(E)** Shelf margin biohermal facies, probably late Llandovery or earliest Wenlock in age. Section 12 Irene Bay. Pierre Gravel standing in snow at base of cliff for scale. **(F)** Upper limestone member of Allen Bay Formation, represented by 25m of light-coloured limestone in upper half of precipitous cliff section. Arrow marks minor(?) disconformity near base of member. Note relief associated with this surface. Most of the rock at this

locality is pelmicrite and microbial carbonate, and some domal microbialites in the dark band mimic the topography of underlying disconformity. Darker strata in bottom half of photograph are part of middle member of Allen Bay Formation. Section 62, head of Makinson Inlet.

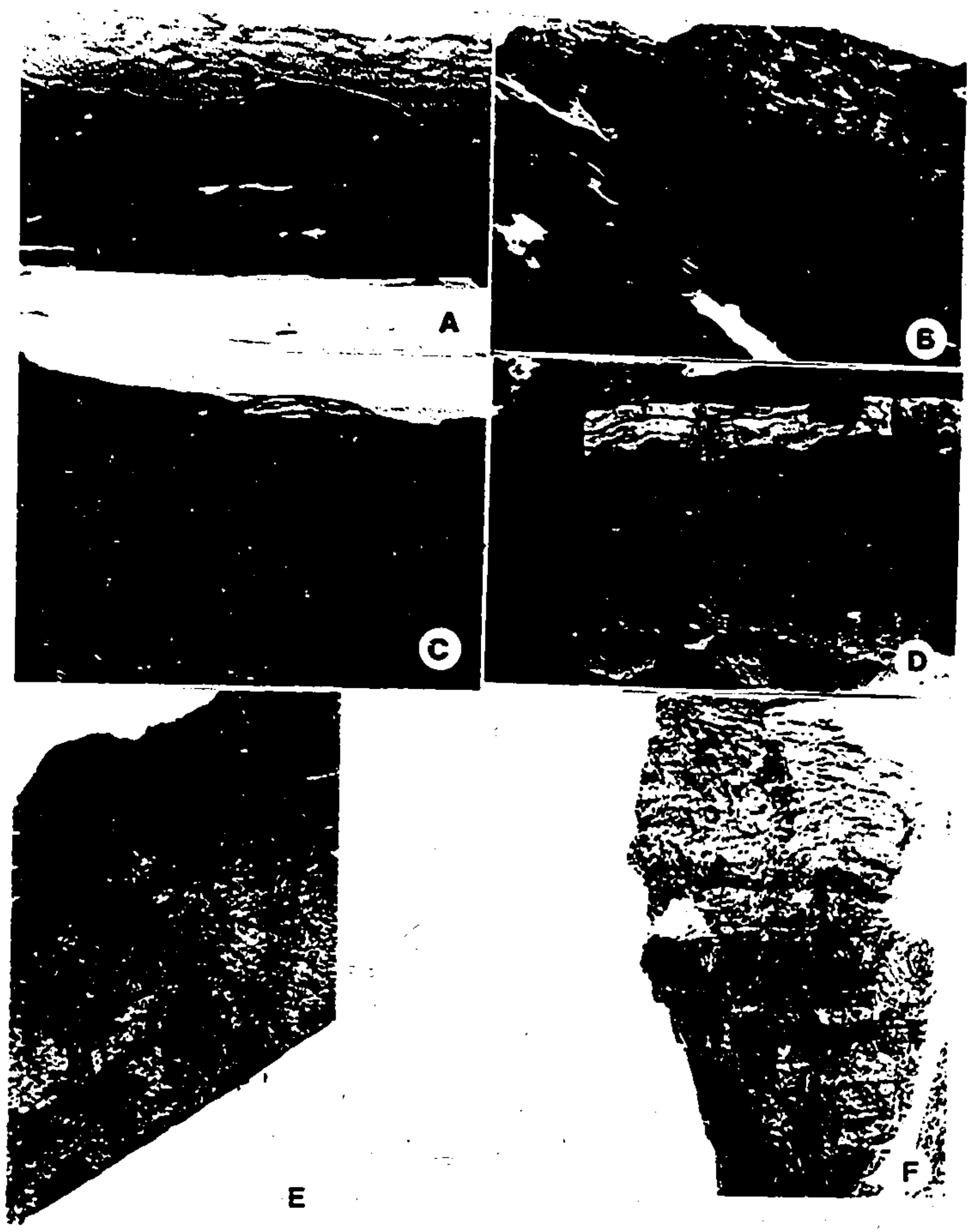

A
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Plate 2: Upper dolostone member of Allen Bay Formation. (A) Typical striped bedding of the member in cliff exposures. Light bands are variable lithologically, and include microbialite, laminite, and calcisiltite. Darker beds are fossiliferous, subtidal dolostone. Light beds near top of photograph are 2-4m thick. Section 15, Eidsbotn Fiord, northern Devon Island. (B) Stromatoporoid bindstone of cycles illustrated in Fig. 11, section 15. Scale card 9cm wide. Extensively silicified fossils are typical of upper member in this area. (C) Typical cycle of upper dolostone member. Arrow 3: stromatoporoid-skeletal bindstone; Arrow 2: sharp contact between well sorted grainstone and underlying rudstone and floatstone units; Arrow 1: base of stromatolite interval not well represented in this cycle. Locality as in 2a. Stick (left-center) is 1.5m long.

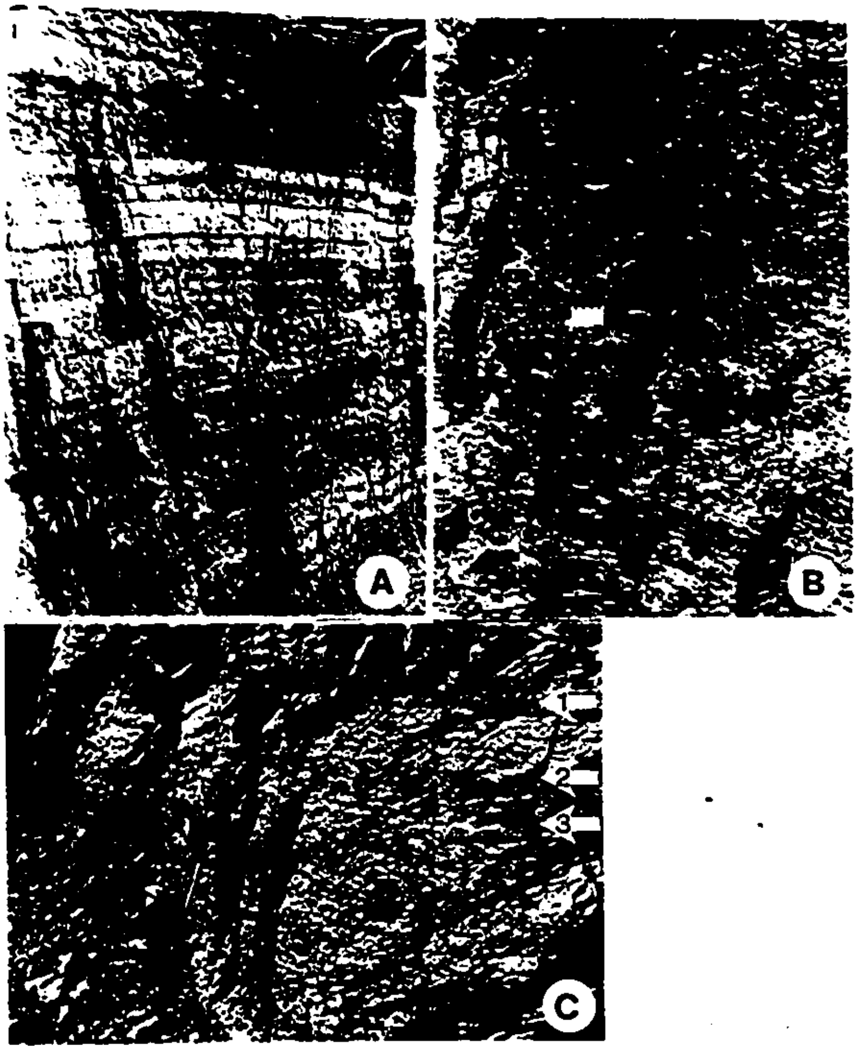
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Plate 3: Cape Storm Formation exposures on southern Ellesmere Island. **(A)** Stromatolites about 2m high near middle of the Cape Storm Formation, section 61, Swinnerton Peninsula. **(B)** Bioturbated dolomitic limestone, characteristic of upper part of Cape Storm Formation in study area. Locality 62, Makinson Inlet. About 3cm of hammer head shown in lower left. **(C)** Silicified synaeresis cracks in silty dolostone, upper part of Cape Storm Formation of Swinnerton Peninsula and Makinson Inlet. Locality 61. End of hammer head (lower left) is 3cm thick. **(D)** Wavy bedding in lower part of Cape Storm Formation, locality 62, Makinson Inlet. About 2cm of length of hammer handle shown (lower right). **(E)** Patterned carbonate of upper Cape Storm Formation, locality 61, Swinnerton Peninsula. **(F)** Silicified synaeresis cracks in the upper Cape Storm Formation, locality 62, Makinson Inlet. Hammer handle (top) is 3cm wide.

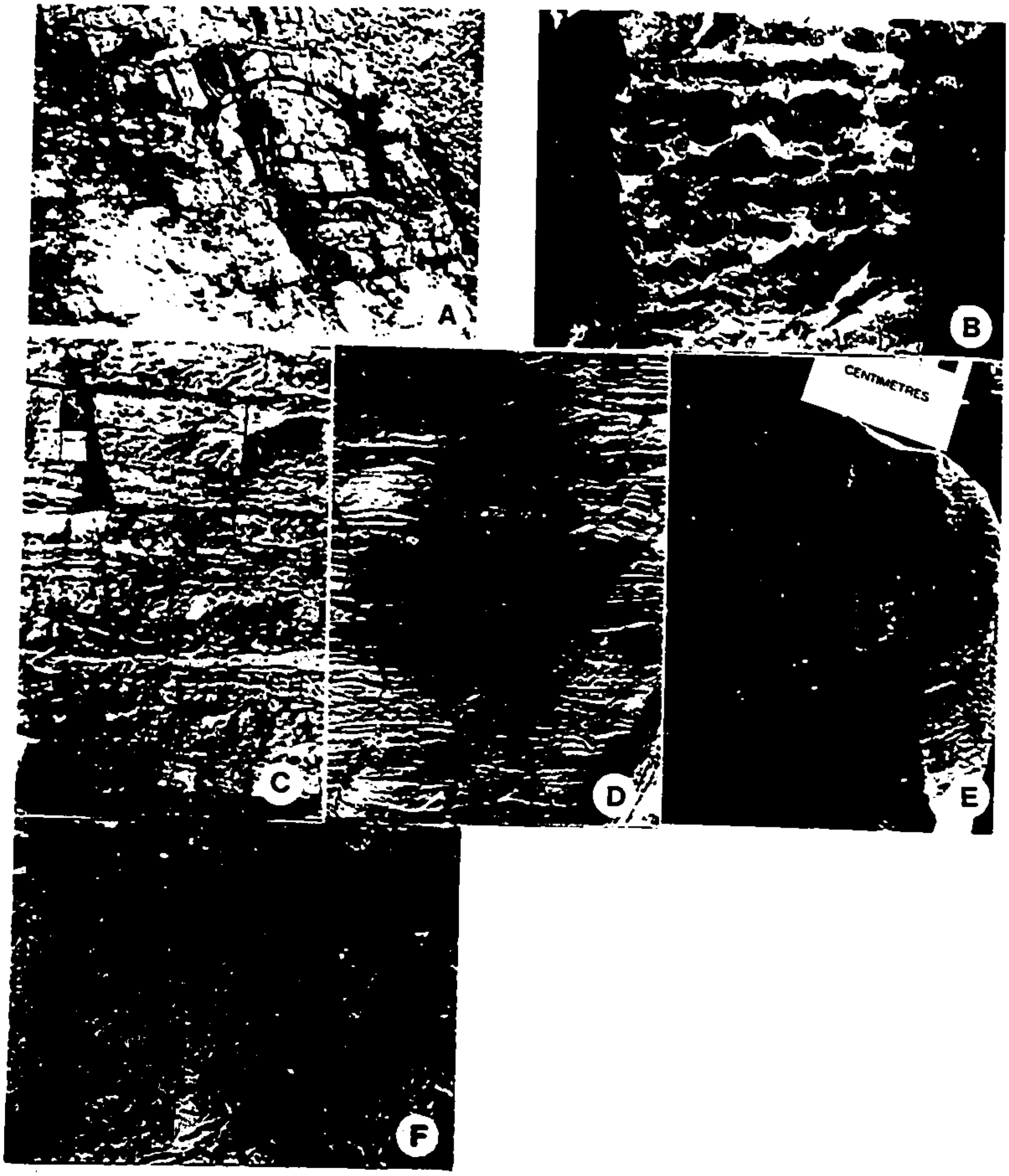
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Plate 4: Biostratigraphy of Silurian slope strata, sections 50, 51, and 54, Baumann Fiord. Black dots mark significant graptolite collections. The Ordovician-Silurian boundary indicated is approximate but consistent with its position in coeval strata on northern Devon Island. Poey (1988) regarded the systems boundary as lying at the shale-limestone contact about 120m above the boundary indicated here. See text for discussion. Height of hill is about 700m. Note olistostromes in the middle Wenlock part of the sequence (*perneri* and *testis* Zones).


permei
rigidus
centrifugus
sakmaricus
estoniensis

Plate 5: Exposures and slabbed samples of various Silurian buildups and associated facies. (A) Pentamerid brachiopods up to 3cm in diameter with isopachous cements now represented by dolostone, form strata flanking a shelf-margin reef (Ludlow) on east-central Cornwallis Island. (B) Approximately 1.5cm thick microbial carbonate crust (dark zones) on a digitate coral colony (lower left), from coralgal biolithite pinnacle reef facies. Note primary void filled with peloidal and skeletal calcisiltite. Slabbed sample from section 76, Hoved Island. (C) Skeletal algae (most light portions) and lithistid sponges (arrows) embedded in microbial carbonate (most dark areas), an uncommon lithology of the pinnacle reef biolithite facies. Sponges indicated by arrows are 0.8 and 1cm in diameter. Slabbed sample from section 76, Hoved Island. (D) Stromatoporoid-rich strata, Hans Island, sections 27, 28. Note the change from laminar forms in floatstone in the lower part to globular or subhemispherical forms in bindstone in upper part. Compare with schematic diagram in Figure 11. Ice axe (arrow) about 1m long. (E) As in 5C, but sponge is 5cm in diameter. (F) Laminar and tabular stromatoporoid floatstone of lower part of a cycle on Hans Island (locality 28). Square at top of stick is 15cm wide. (G) Dark fissile shale overlying resistant limestone of large mud buildups established over the drowned Ordovician shelf margin. Shale-limestone contact in background (arrow) and foreground on two sides of broad anticline. Lighter coloured strata in middle of syncline belongs to the Danish River Formation. Section 41, Cañon Fiord.

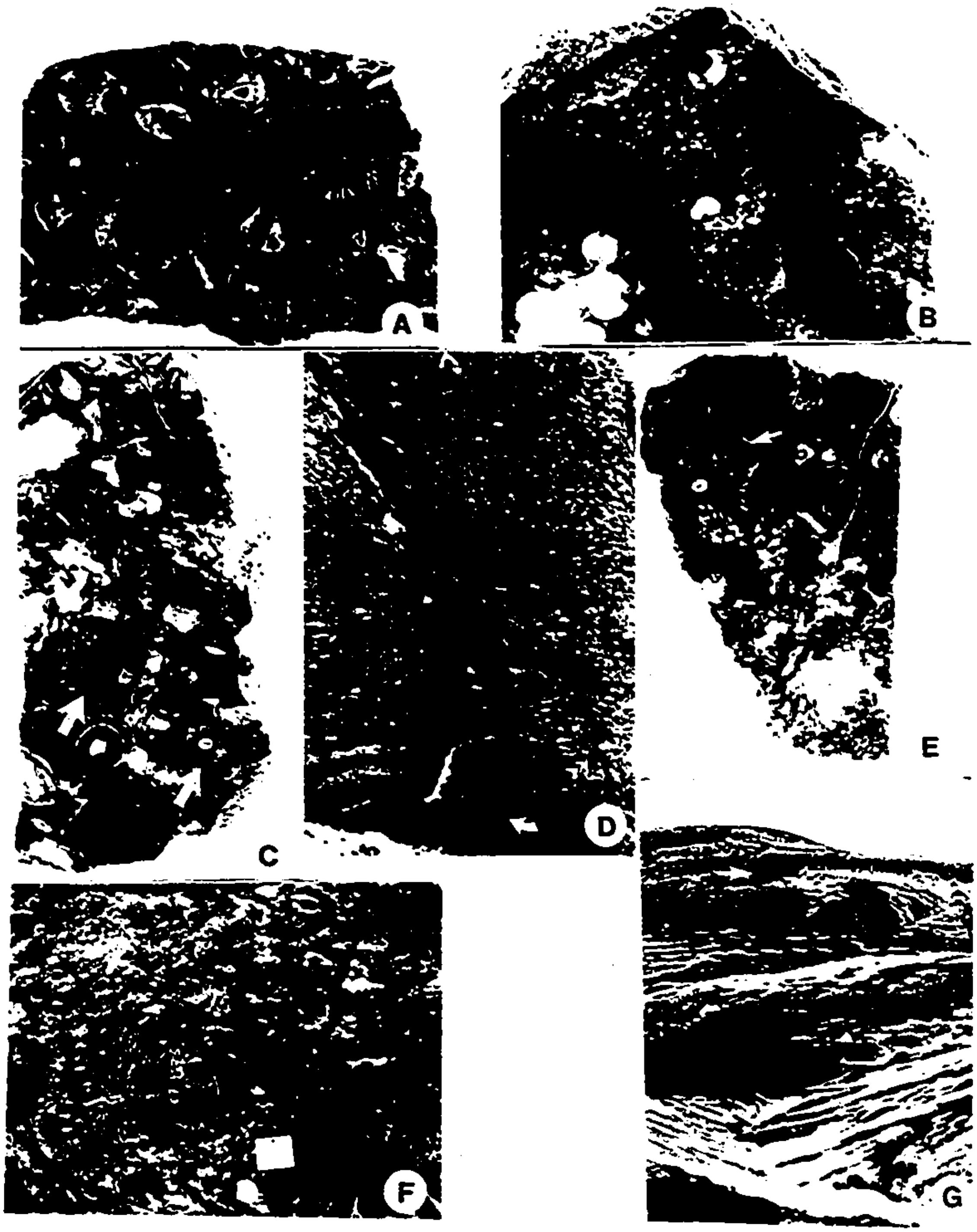
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Plate 6: Pinnacle reef cliff exposure, Hoved Island, Baumann Fiord. Exposure showing *in situ* coralgal biolithite facies, left of dashed line, progradational over slope strata. Some blocks (olistoliths) outlined. Top of structure not exposed, but oolites abundant in slope exposure 2-3km to right of cliff. Total height of cliff is about 380m. Beds dip approximately 10° toward viewer.

Plate 7: Platform carbonates, Darling Peninsula, locality 31 (see Figure 9c, section 31). **(A)** Bedding plane view of large mounded burrows in unit 7a, Goose Fiord Formation equivalent. **(B)** Colonial coral in unit 7a, Goose Fiord Formation equivalent. Diameter of colony 16cm. **(C)** Rubbly weathering argillaceous limestone, Goose Fiord Formation. Lithology also typical of lower Douro Formation in this stratigraphic section. Height of cliff is 14m. **(D)** Main cliff exposure is Unit 3 of the Allen Bay Formation, a distinctive, thickly bedded, vuggy, biohermal(?) unit at this locality. Height of cliff above the snow-covered scree is about 40m. **(E)** Stromatoporoid with weathered-out (commensal) rugosan colony, in Douro Formation equivalent, Unit 7b.

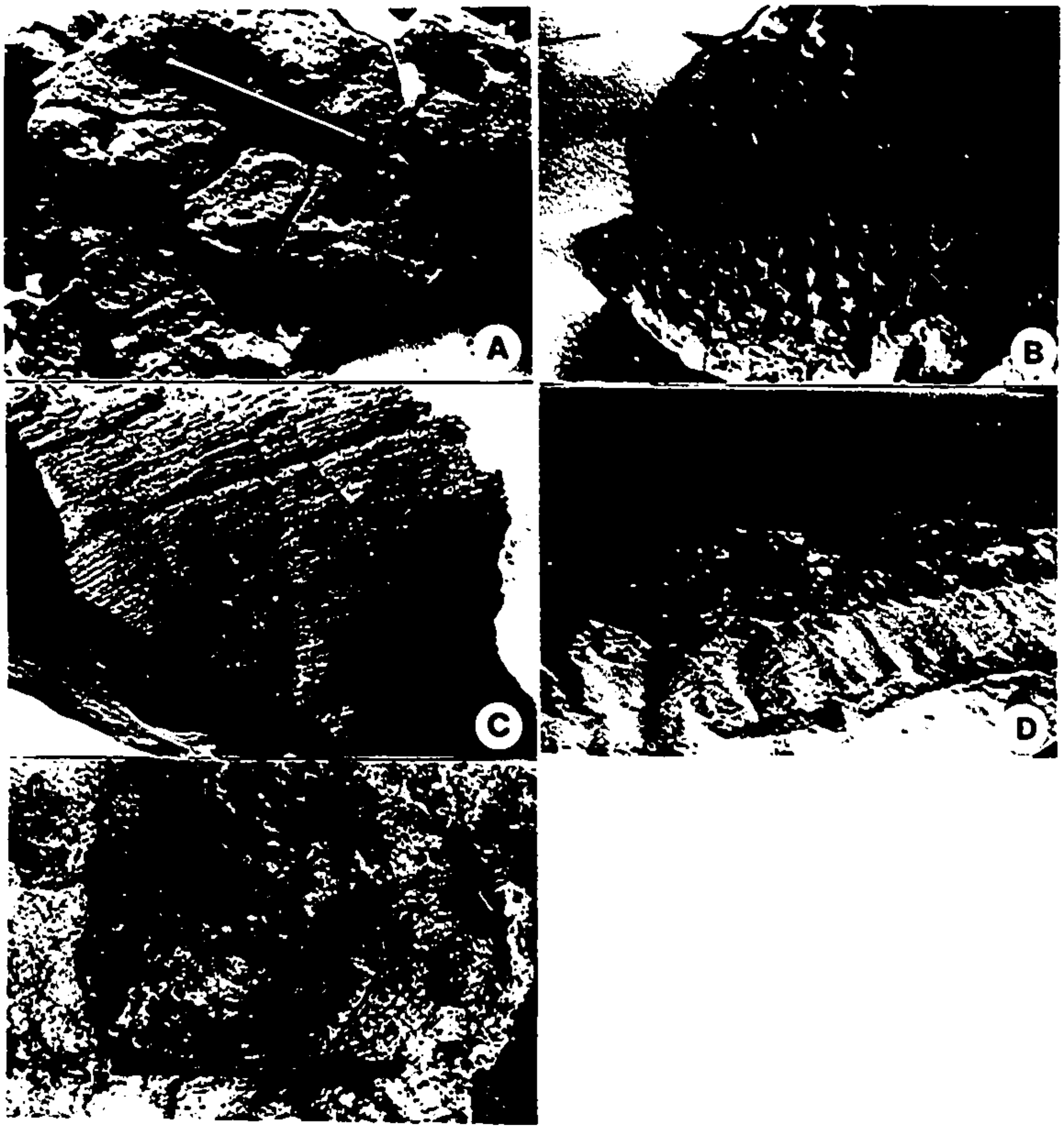
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Plate 8: Deep shelf exposures near Bay and Trold fiords and on Judge Daly Promontory. (A) Upper Ordovician strata, locality 60, Trold Fiord: resistant beds of Thumb Mountain Formation (left of arrow 1); Irene Bay Formation (arrows 1 to 2); basal tongue of the Allen Bay Formation (arrows 2 to 3); and Cape Phillips Formation (right of arrow 3). Major break in slope represents *fastigatus* Zone platform drowning and shale onlapping. Distance between arrows 1 and 2 is 4.5m. (B) Rhythmically interbedded shale and limestone beds in *fastigatus* Zone. Major break in slope represents the *fastigatus* Zone platform drowning event. Total thickness of shaly strata shown is about 60m. Locality 60, Trold Fiord. (C and D) *curtus* to *cyphus* Zone chert beds, Cape Phillips Formation, locality 60, Trold Fiord. Same resistant 2.2m thick limestone bed shown on right of C and left of D. Beds in foreground of C are predominantly chert, whereas in D, chert is more sporadic or locally abundant and interbedded with marlstone and mudrock. (E) Cape Phillips Formation, locality 25. Note the change in attitude of the strata, from near horizontal at arrow 1 to near vertical at arrow 2. Between the two arrows are tightly folded strata typical of the Ellesmere Island Fold belt on Judge Daly Promontory. Height of hill is 300m. (F) Upper Pridoli encrinite channel complex resting on graptolitic shale, locality 12, Irene Bay. Pierre Gravel (arrow) for scale.

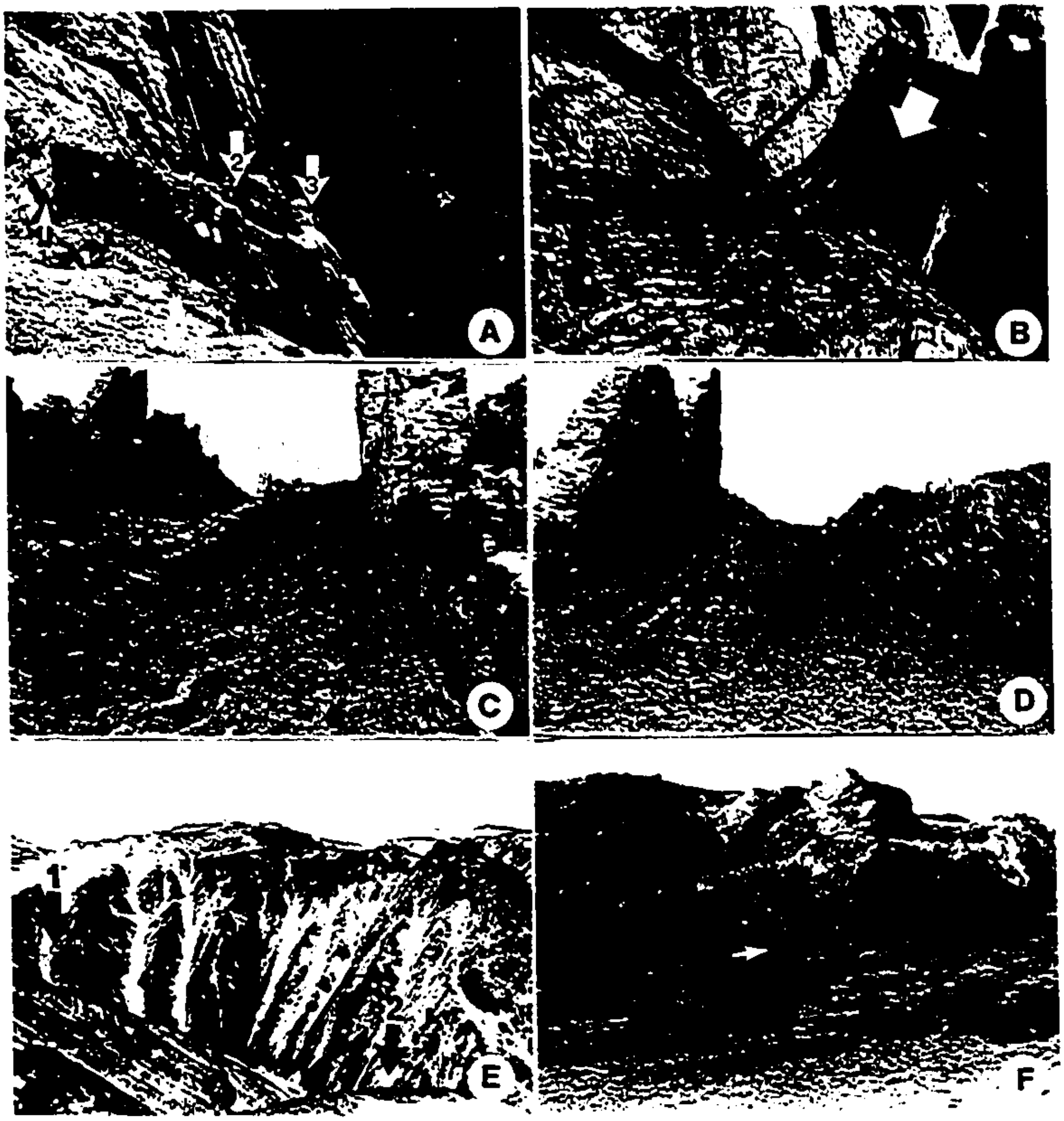
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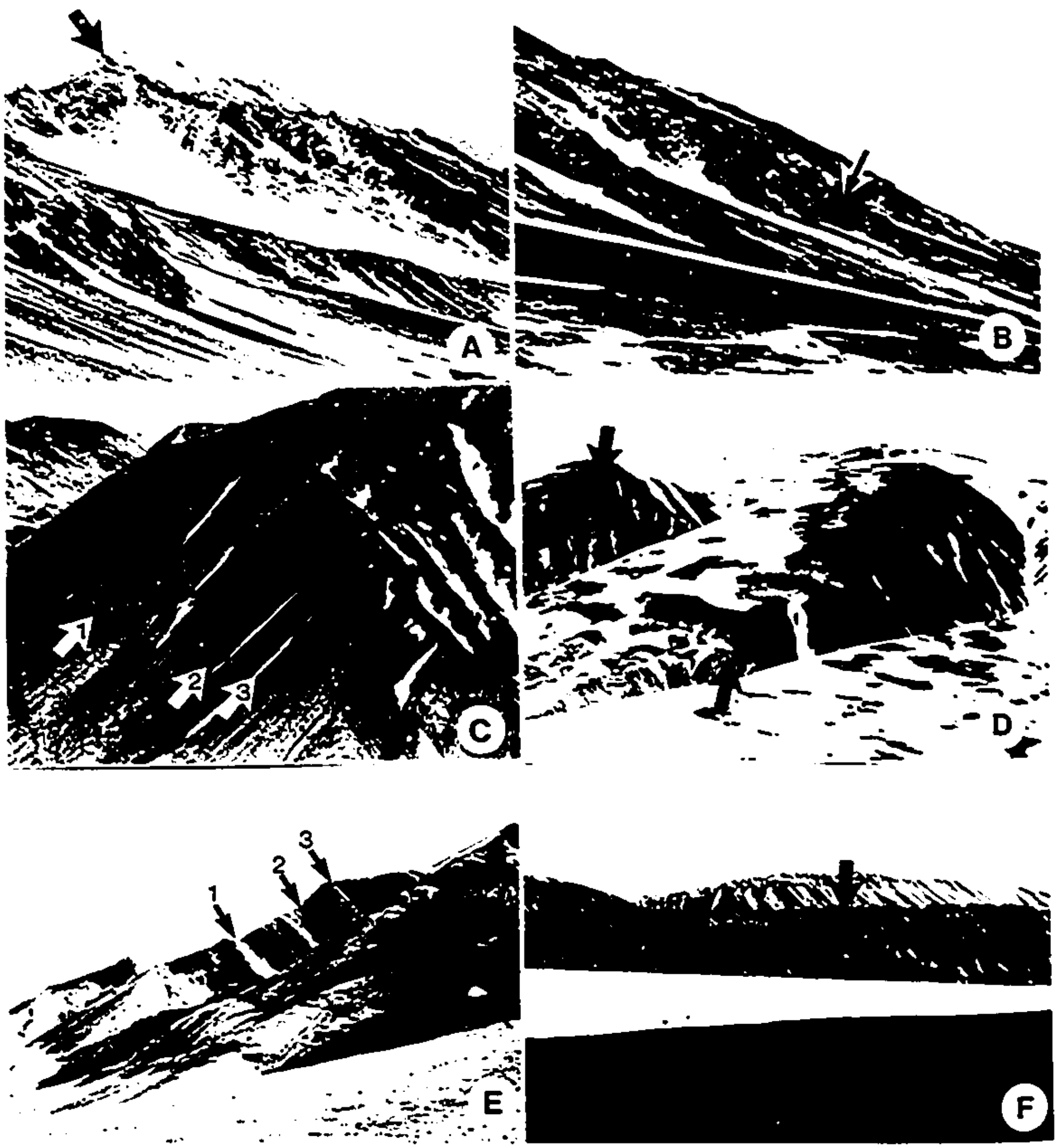
Plate 9: Deep shelf and trough clastic exposures, Trold and Vesle fiords and Judge Daly Promontory. **(A and B)** Cape Phillips Formation, section 26. Arrow in A marks Danish River-Cape Phillips formation contact, placed where brown-weathering siltstone and sandstone upward become predominant over mudrocks. Beds immediately above arrow in A correlate with a resistant >7m thick lenticular olistostrome (below arrow in B). Height of hill in A is about 250m. (C) Cape Phillips Formation, section 43, Vesle Fiord. Arrows 1 and 2 bracket Starfish Bay shale, about 60m thick here. Arrows 2 and 3 bracket strata of middle and upper Wenlock age, a condensed sequence in contrast to the Ludlow and Llandovery parts of the sequence at this locality. Locality 43, Vesle Fiord. (D) Cape Phillips and Red Canyon River formations, same locality as in 9C. Top arrow indicates base of Starfish Bay shale, and lower arrow indicates base of the grey siltstone facies of Red Canyon River Formation, as described in this study. Field of view is about 1.5km. (E) Overturned succession of the Cape Phillips and Danish River formations near locality 60, Trold Fiord. Arrows 2 and 3 indicate top and base respectively, of Starfish Bay shale (about 38m thick) and arrow 1 the base of the Danish River Formation. (F) Overturned Danish River Formation sequence on west side of Trold Fiord, west of locality 60. Wedge shaped structure on this 250m sea cliff likely represents low angle fault.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PART III

SILURIAN MUD BUILDUPS

ABSTRACT

Silurian mud buildups occur at several localities on Ellesmere Island, Canadian Arctic. A mudmound at Baumann Fiord, southern Ellesmere Island, is 310m thick by 1.5km long and was deposited above(?) storm wave base on a platform slope. Off-mound sediments are graptolitic and indicate that the moundrock was deposited at a rate of about 7cm/1000 years, not considering compaction. Basal mound deposits are predominantly burrowed Llandovery lime mudstone, discontinuously present over the study area and locally containing abundant pentamerid brachiopods. This mudmound, although only known from a reconnaissance study, consists of a lower lime mudstone unit, and an upper cryptomicrobial micrite unit, containing rare stromatactoid structures.

Several large mud buildups, established over the drowned Ordovician shelf-margin, are now exposed on western Ellesmere Island. A vertical succession is apparent in a buildup at Cañon Fiord: burrowed, sparsely fossiliferous lime muds are succeeded by cryptomicrobial and fenestral carbonate, then by tabular stromatoporoid bindstone and floatstone, representing an overall shallowing-upward sequence. The stromatoporoid-rich strata are in turn overlain by cryptomicrobial carbonate, signifying a return to slightly deeper water. Microbial carbonate is interpreted to have been deposited above storm wave base, in slightly shallower water than is characterized by the *Pentamerus* benthic community (Zeigler, 1965), but in deeper water than interpreted for the stromatoporoids,

which probably grew at or above fair-weather wave base.

Buildups are up to 1.3km thick and were abruptly drowned in the middle Wenlock. Subsequent to drowning, buildups were local submarine highs and sites of starved sediment deposition, now represented by a condensed, fissile black shale unit. At Bay Fiord, central Ellesmere Island, wackestone and packstone predominate in two 5 to 8 metre thick mounds. These were deposited on the slope of a rimmed shelf and are dated as upper Wenlock, based on an association with beds containing the graptolites *Cyrtograptus ludensis*. Metre-scale wackestone to grainstone beds interbedded with red-stained, cephalopod-, crinoid-, and ostracod-rich lime mudstone and wackestone occur in the mound rock. Allochthonous and/or autochthonous crinoid debris is predominant over lime muds in moundrock and is present together with abundant stromatactoid structures. A crude lithological zonation is interpreted as a succession of mound upbuilding episodes largely as a result of binding and trapping of sediments by microbes and crinoids.

INTRODUCTION

During the Silurian, more than 3km of carbonate strata were deposited over the southern flexure of the Franklinian Basin (Trettin, 1989). A large proportion of this carbonate sequence is reefal, and occurs within an extensive and well-exposed platform complex on Ellesmere, Devon, and Cornwallis islands. Many of the major trends and developments of the platform complex have correlative events in coeval strata of North Greenland (Hurst, 1980; Sønderholm, and Harland, 1989). Mudmounds occur in both areas (Figs. 27, 28), but are known mainly from reconnaissance work, particularly on North Greenland.

Ellesmere Island mudmounds were examined briefly during the summers of 1987, 1988 and 1989, and were part of a study of the Upper Ordovician, Silurian, and Lower Devonian stratigraphy of Ellesmere and Devon islands. Mud-buildups are documented in three areas in this report (Fig.27); however, this work is ongoing and some of the presented data is preliminary. Buildups near Bay Fiord and Cañon Fiord, Ellesmere Island, are better known and were investigated in greater detail. The mudmounds described in this study are the first to be documented in the Middle Paleozoic of Ellesmere Island, although "Waulsortian" reefs have been described from Carboniferous strata of the Sverdrup Basin (Davies *et al.*, 1989). Mound evolution and anatomy is discussed and compared to Silurian mudmounds occurring globally.

Fig.27. Map showing location of studied mud buildups (shown in black). Stippled area indicated on southwestern Ellesmere Island is the inferred extent of the buildup which only has its slope facies exposed. The continuous light grey line is approximately the position of the Silurian shelf margin; shallow-water carbonates were deposited to the southeast (right) of the line and deep-water (shelf and basinal) shales to the left. The three mud buildups to the northwest of the line occur along the former (drowned) Ordovician shelf margin. Platform backstepping occurred in latest Ordovician (see discussion in text). Paleogeography does not take into account northwest-southeast tectonic shortening of the region.

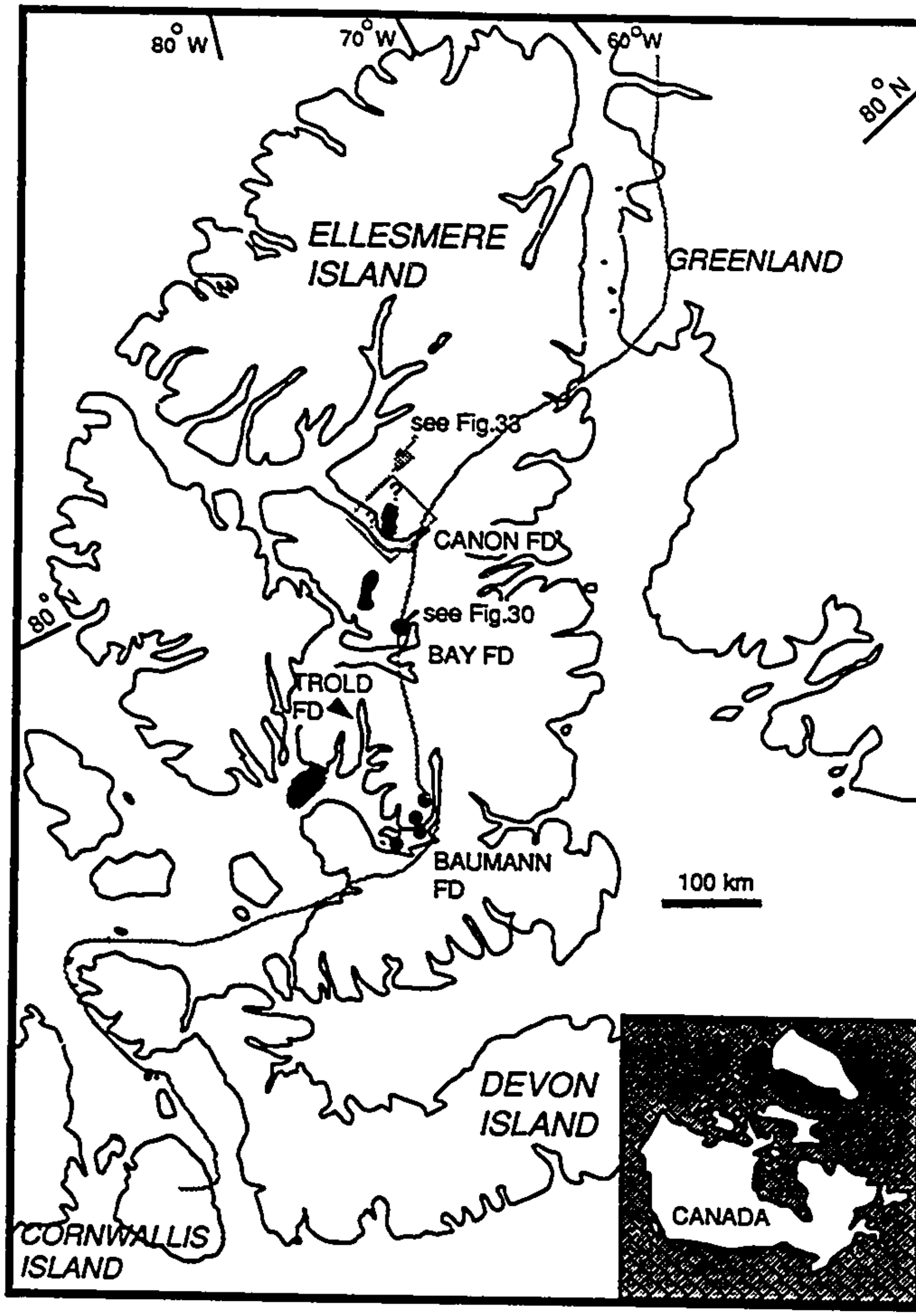
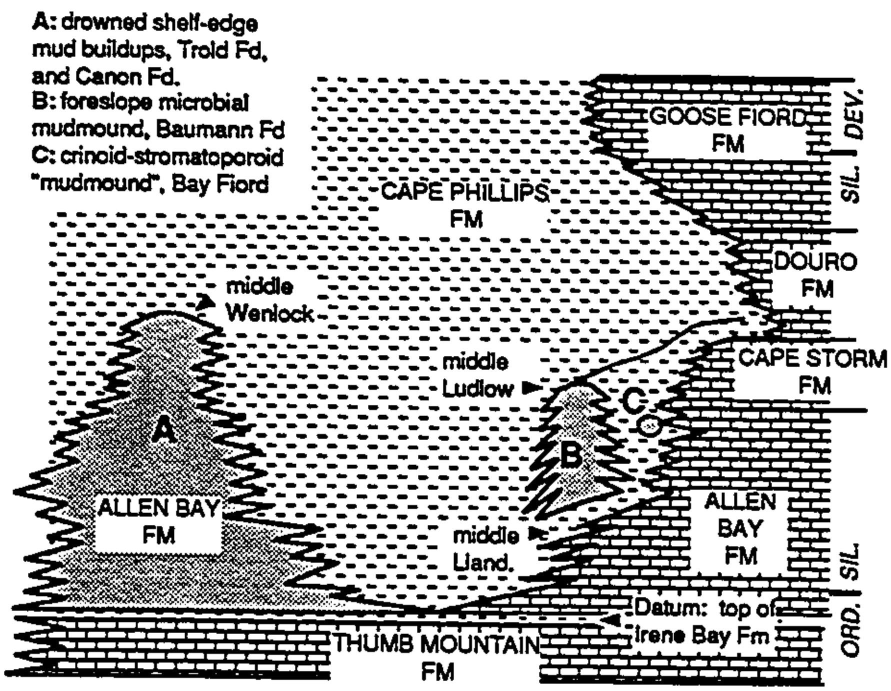

80° W
70° W
60° W
80° N
ELLESMERE ISLAND
GREENLAND
see Fig.33
CANON FD
see Fig.30
BAY FD
80°
TROLD FD
BAUMANN FD
100 km
DEVON ISLAND
CORNWALLIS ISLAND
CANADA

Fig.28. Generalized schematic representation of the main Upper Ordovician and Silurian carbonate units on Ellesmere Island. *Bricks*--shelf carbonates; *dashes*--drowned platform shales; *stipples*--carbonate mud buildups discussed in this report.


A: drowned shelf-edge mud buildups, Trold Fd, and Canon Fd.
B: foreslope microbial mudmound, Baumann Fd
C: crinoid-stromatoporoid "mudmound", Bay Fiord
CAPE PHILLIPS FM
GOOSE FIORD FM
DEV.
SIL.
DOURO FM
middle Wenlock
middle Ludlow
CAPE STORM FM
A
B
C
ALLEN BAY FM
ALLEN BAY FM
middle Lland.
SIL.
Datum: top of Irene Bay Fm
ORD.
THUMB MOUNTAIN FM

SILURIAN MUDMOUNDS: A REVIEW

Mudmounds are abundant in the Silurian platform sequence of the Canadian Arctic. They are distinct from shallow-water stromatoporoid reefal deposits as major frame builders are subordinate and lime mudstone is abundant. Main Silurian mudmound categories present in the Arctic and reported elsewhere, are listed below (also see Table 8 and Figure 29):

(1) Skeletal metazoan-bearing mudmounds including

(a) **crinoid-stromatoporoid "mudmounds"** (this study);

(b) **coral-algal-sponge mudmounds** (Graf and Dixon, 1986; Dixon and Graf, in prep.; also including the "diverse community mudmounds" of Packard, 1985);

(c) **lithistid sponge mudmounds** (Narbonne and Dixon, 1984, 1989).

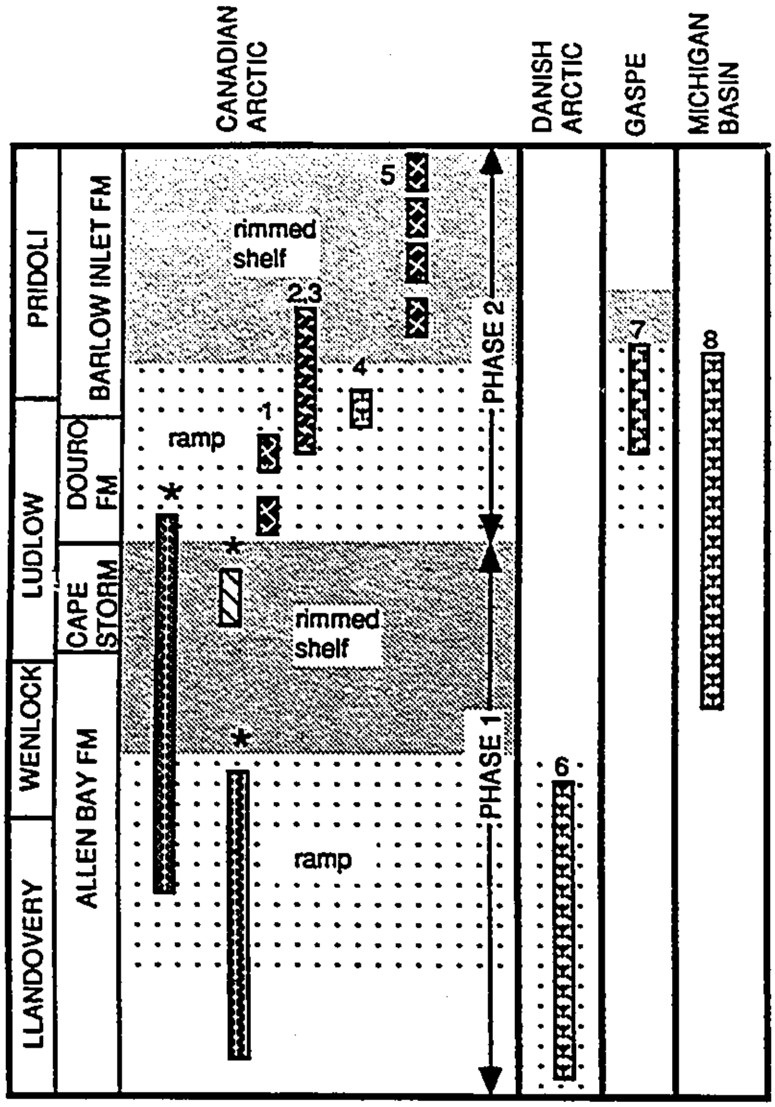
(d) **stromatactis-sponge? mudmounds** (Bourque and Gignac, 1983)

(2) microbial mudmounds including

(a) **stromatactis mudmounds** (Textoris and Carozzi, 1964; Lehmann and Simo, 1989; McGovney, 1989); catagories 1d and 2a are very similar, but see discussion below);

(c) **microbial mudmounds** (this study, as observed at Baumann Fiord, west-central Ellesmere Island, Trold Fiord, and Cañon Fiord; and in this study, excluding "coralgal biolithite reefs", discussed below.

Fig.29. Relationship of Arctic buildups and other mud buildups to platform profile. Also see Table 8. * Buildups of this study; 1. sponge mudmounds, Somerset Island, Canadian Arctic (Narbonne and Dixon, 1984, 1989); 2,3. skeletal metazoan mudmounds (Graf and Dixon, 1986; Dixon and Graf, in prep.; Packard, 1985); 4. Microbial mudmounds (Thorsteinsson and Mayr, 1987); 5. sponge mudmounds (Packard, 1985); 6,7,8. microbial mudmounds (Hurst, 1980); (6) Sønderholm and Harland, 1989; (7) Bourque and Gignac, 1983; (8) Textoris and Carozzi, 1964; Lowenstam, 1950; Shaver and Sundermann, 1989; Shaver, 1974). Reefs in the Michigan basin are either platform-situated or occur on the slope of a subsiding carbonate ramp; the latter are commonly referred to as pinnacle reefs.



lithistid sponge mudmounds; stromatactis-rich mdmnds
coral-algal-sponge mdmnds (diverse community mdmnds)
microbial mudmounds; crinoid-stromatop. mdmnds

Another category not included in this list and, in terms of lithology, probably intergradational with some mudmounds, are "algal" buildups, discussed by Clough and Blodgett (1989, 1985) and Bourque *et al.* (1986). These buildups contain various proportions of skeletal metazoans, mainly tabular corals, that are thickly encrusted by a consortium of microbial carbonate and skeletal algae forming a "coralgal biolithite" (as referred to in this investigation). Although Bourque *et al.* (1986) described Anse à la Barbe reefs as "algal" mud-dominated buildups, they are, in this work, allied to coralgal biolithite buildups of the Arctic and to "algal reefs" described by Clough and Blodgett (1989, 1985), because facies associations, composition of skeletal and non-skeletal algae, and inferred paleoenvironmental setting are very similar. These buildups are distinct from mudmounds listed above in that skeletal algae are abundant and diverse, and skeletal metazoans locally form a framestone fabric. Also, these reefs probably existed in very shallow water (see discussion below), and are unlike the deep-water sponge rich and microbial mounds discussed in this report.

Sponge mudmounds described in the literature are relatively small structures that formed in moderately deep water above storm wave base on platform slopes or ramps. They are relatively common in the Canadian Arctic (Table 8; Fig.29) and typically display a zonation of macrofossils; for example, lithistid sponges and algae are succeeded by corals and sponges (Narbonne and Dixon, 1984).

TABLE 8: SILURIAN MUD BUILDUPS

SIZE/LOCATION	SETTING	MAIN CONSTITUENTS	MINOR CONSTITUENTS
Gros Morbe Mbr, Gaspe, Bourque & Gignac, 1983. *stromatactis mudmounds*, 115m thick.	below wave base at margin of uplifted basement blocks, Gaspe Basin.	lime mudstone, stromatactis, peloidal micrite, sponges (inferred), capped by microbial reef complex forming a rimmed shelf	favositids, stromatoporoids, brachiopods, dasycladaceans, ostracods.
Washington Land Gp, Greenland, Sonderholm and Harland, 1989; Hurst, 1980, zoned *stromatactis mudmounds* 100's m thick.	below wave base(?), intrashelf, & distally steepened ramp.	unknown, ?stromatactis, ?lime mudstone.	unknown.
Several formations, Michigan Basin, Shaver and Sunderman, 1989; Textoris and Carozzi, 1964 zoned *stromatactis mudmounds* >24m thick.	intrashelf in 100's m to shallow water (?), built to wave base.	Core: crinoid, stromatactis, lime mudstone; top: stromatoporoids, crinoid debris.	Core: ostracods, stromatoporoids, bryozoans, sponge spicules; top: stromatactis, sponge spicules, pellets, corals, bryozoans, halysitids, favositids, gastropods. (see Shaver, 1974)
Cape Phillips Fm, Baumann Fd., Ellesmere Island. *mudmound*, 450m thick by 2km long, this study.	below storm wave base, situated on slope of rimmed shelf	lime mud, microbial carbonate.	poorly known; stromatactoid structures.
Cape Phillips Fm, Bay Fd., Ellesmere Island, *crinoid-stromatactoid mudmound*, 15m thick by 30m long, this study.	slope of a rimmed platform.	stromatactoid cavities, pelmatozoan debris, lime mudstone.	ostracods, stromatoporoids trilobites, orthocone cephalopds brachiopods, calcareous algae, microbial carbonate, graptolites, sponges.
Douro Fm, Devon Island, Dixon and Graf, in prep. zoned *coral-algal-sponge mudmound complexes* 52m thick.	ramp, storm wave base to fair weather wave base.	Base: sparsely fossiliferous, lime mudstone, crinoid debris, lithistid sponges; top: corals, stromatoporoids.	Top: stromatoporoids, microbial carbonate, stromatactoids, crinoid holdfasts calcareous algae, brachiopods, ostracods, trilobite fragments, bryozoans, ostracods, stromatoporoids; Base: calcareous algae, cryptalgae, stromatoporoids, corals, stromatactis.

Table 8 cont'd

Douro Fm, Somerset Island, Narbonne and Dixon, 1984, 1989, zoned *lithistid sponge mudmounds*, 35m dia.	calm turbid waters on a ramp.	algal-lithistid sponge consortium, with preserved spongiostromata at base; coral-rich top.	Base: corals, algae, crinoid holdfasts bryozoans; top: bryozoans, algae, cornulitids, trilobites, ostracods, gastropods
Barlow Inlet Fm, Cornwallis Island. *lithistid sponge mudmounds*, Packard, 1985. 2.2m-18m thick.	below wave base on a slope of a rimmed shelf.	Base: sponges, lime mudstone; top: corals, sponges, lime mudstone.	basal crinozoan holdfasts crinoid debris, syringoporids, stromatoporoids, ostracods, bryozoans, rugosans.
Barlow Inlet Fm , Cornwallis Island, Packard, 1985, *diverse community mudmounds* , .4m wide by .2-5.4m thick.	occur within sediments representing shoaling of a platform slope.	Lime mudstone, trepostomate bryozoans, lithistid sponges, calcareous algae, corals (rugosans & tabulates); one mound crudely zoned.	sponge spicules (desmas) crinoid rudstone, spongiostromata, stromatoporoids.
Douro Fm, Devon Island, Thorsteinsson and Mayr, 1987. *stromatactis mudmounds*, 15m wide by 25m thick.	calm conditions on a ramp.	lime mudstone, stromatactis, but poorly known.	corals, brachiopod debris, crinoid debris.
Allen Bay Fm, western Ellesmere Island, this study, *microbial buildups* 10's km long and 1.2km thick.	on margin of drowned Ordovician platform, above storm wave base	zoned: shallowing upward sequence, basal burrowed lime mudstone, middle microbial carbonate, upper stromatoporoid bindstone.	pentamerids, sponges, trilobites, ostracods, favositids, rugose corals, megalodont bivalves
Several Fms., Michigan Basin lime mud-cored pinnacle reefs, (Mesolella et al., 1974; Sears and Lucia, 1979);100's m thick.	ramp foreslope, below wave base?	recrystallized; top: stromatoporoids, corals, cal. algae in reefs with "multiple starts. Base1: algae (microbial carbonate, stromatactis?, corals); Base 2: green & red calcareous algae encrusting corals	Top: brachiopods, gastropods, crinoid debris, & other typical shallow water fossils. Base: sponge spicules, ostracods, brachiopods, stromatoporoids.

Silurian sponge mudmounds are broadly similar to Lower and Middle Ordovician reefs in size and makeup. Sponge genera and even species that inhabited the mounds of the two periods are very similar (de Freitas, 1989); however, Ordovician mounds generally formed in somewhat shallower water than their Silurian counterparts and were generally replaced by stromatoporoid reefs in about Middle Ordovician time (Klappa and James, 1980). This major event in reef evolution, marked by the appearance of polymorphic, carbonate-secreting stromatoporoids, heralded a major decline of reef-building sponges. The difficulty of silica secretion in a carbonate-saturated marine environment undoubtedly was a major factor in this decline.

Main frame builders in Silurian sponge mudmounds are lithistid sponges (predominantly anthaspidellids). Main components of these sponges are desmas (in particular dendroclones, chiastoclones, and other irregular spicules) which are well articulated and form a very regular, ladder-like skeletal net. Their fossilization is distinctive, although they are typically calcitized where preserved in the micrite of the sponge mudmounds in the Canadian Arctic. Stromatactis, although a very minor constituent of the Arctic sponge mudmounds, is associated with core or basal mound lithologies and is preserved together with pervasive cryptomicrobial micrite. The reported composition of sponge mudmounds in the Canadian Arctic is in marked contrast to the dominantly stromatactoid mudmounds which contain "true", abundant stromatactis (Bathurst, 1982, p.167) that can constitute up to 90% of the rock volume (Textoris and Carozzi, 1964).

Bourque and Gignac (1983) described stromatactis mudmounds in Gaspé Peninsula, Québec. They outlined the diagenetic sequence of events that transformed a so-called sponge-rich muddy carbonate to a stromatactis-rich mudstone. By analogy with the diagenetic effects observed in hexactinellids in off-mound strata, they attributed spar-filled stromatactis and enclosing peloidal micrite to decay and calcitization of the soft tissue of a probable hexactinellid sponge. Remnant hexactines and monaxons are, in part, evidence for this process. However, hexactinellids are consistently minor or are not present in described Paleozoic mudmounds. Triassic mound-building lynchniscid hexactinellid sponges are recorded from China (Wendt, *et al.*, 1989), but these sponges have architecturally well-organized hexactine-based skeletons that are quite unlike earlier lyssacine hexactinellids of middle and lower Paleozoic strata, and also did not evolve until the Triassic. Moreover, in sponge communities known from a variety of middle and lower Paleozoic depositional settings, hexactinellids are very minor fossil components. The well articulated desmas and the very rigidly constructed lithistid skeleton is more conducive to reef building, having evolved prior to the stromatoporoids in the Middle Ordovician. This is not irrefutable evidence against a sponge origin for stromatactis and/or the peloidal mud of mudmounds, but it puts in question the importance of hexactinellid sponges as major builders of Silurian mudmounds. In searching for a sponge origin for mudmounds, it is probably better to search for desmas and organized lithistid skeletal nets.

Silurian "mudmounds" are apparently very abundant in the Michigan Basin and occur in two main settings, as platform-situated zoned buildups (Shaver *et al.*, 1978) and as cores to basinal pinnacle reefs (Sears and Lucia, 1979; Mesolella, *et al.*, 1974). In the Michigan Basin, a broad transition zone separates shelf carbonates from basinal shales, and this zone, or foreslope environment, harboured 100's to 1000's of pinnacle reefs that grew on a uniformly subsiding ramp. Cores of pinnacles and platform reefs are dominated by mudstone lithologies, including stromatactis-rich, dense, unfossiliferous lime mudstones (microfacies 3; Textoris and Carozzi, 1964) gradationally overlain by stromatoporoid-rich moundrock. It was suggested that this general succession represents growth of buildups (in both settings) from below to above fair-weather wave base (Shaver and Sunderman, 1989). Stromatactis of the platform mud-cored reefs is very similar to stromatactis described by Bourque and Gignac (1983) for the Gross Morbe mounds and may account for up to 60% of some moundrock.

Mud cores of less well known pinnacles are variable. Sears and Lucia (1979) described a general "carbonate mudmound facies" as being mainly a lime mudstone containing abundant stromatactis and fibrous (marine) cements, which they compared to Permian mudmound cements described by Davies (1977). However, Mesolella (*et al.*, 1974) described pinnacle cores have been described as containing corals that "...are encrusted and bound together with thick rinds of carbonate and incorporated into bulbous lumps" (*ibid.*, p.43). Although recrystallized, the core lithology also rarely contained porostromate algae

Sphaerocodium. They called the unit the "coral-algal facies", which is abruptly overlain by a thin-bedded carbonate unit, then by a massive clotted "algal" unit, both lithologies lacking stromatactis. Although their description was vague, this lithology may be comparable to coralgal biolithite reefs discussed above (Bourque, *et al.*, 1986; Clough and Blodgett, 1989; 1985), and as observed in pinnacle reefs of the Canadian Arctic. Briefly, Arctic pinnacle reefs occur in stratigraphic settings very similar to those of the Michigan Basin; however, coralgal biolithite deposits in the Arctic are overlain by extensive oolites. They are built of a consortium of skeletal and non skeletal algae forming thick encrustations on corals and sponges, which locally form a framestone fabric. Large cavities with fibrous cements are present, and stromatoporoids and megalodont bivalves occur rarely. Slope strata associated with the buildups contain numerous, extremely thick, matrix-poor (orthoconglomeratic) olistostromes, locally containing blocks 20m in diameter. These form a very narrow apron around buildups, suggesting substantial relief of these reefs over the surrounding shales.

Sponge mudmounds occur in three main settings in the Canadian Arctic: on a ramp, on the slope of a rimmed shelf, and on the margin of a drowned platform (carbonate platform terminology after Read, 1982). Platform evolution was complex and is a subject of ongoing study by the author, but is generally divisible into two main phases (Fig.29; discussed below). Mudmounds are associated with shelf margin strata throughout the Silurian but are most common in the early portion of Phase II platform upbuilding.

TECTONIC AND REGIONAL STRATIGRAPHIC SETTING OF ARCTIC MUDMOUNDS

Subsequent to a late Proterozoic rifting event in the Canadian Arctic, a thick Paleozoic carbonate-clastic prism was deposited over a passive continental margin, bordering the peri-cratonic Franklinian Basin (Trettin, 1989). This sequence contains some 3700m of stratified Silurian carbonates that are well exposed in the study area.

The platform shows evidence of two main phases of upbuilding each preceded by a platform drowning event (Figs. 28, 29). During both episodes of rimming of the shelf, coarse allochthonous debris was shed to the adjacent drowned shelf, forming an extensive carbonate apron that contains marlstone, graptolitic shale, and conspicuous coarse olistostromes. Considering the volume of allochthonous sediment present in examined platform-slope exposures, phase II upbuilding must have produced a rimmed shelf with less relief than the earlier upbuilding phase. Two mudmounds discussed in this report are clearly associated with slope strata, and this development was closely related to platform evolution.

The datum used for this study, the Irene Bay Formation, represents a distinctive marker unit that extends through most of the Canadian Arctic: its green, recessive, rubbly weathering argillaceous limestone character and abundant phosphatized fossiliferous hardground horizons are easily distinguishable, in most areas, from bounding massive limestone formations.

Subsequent to the (climatic?, eustatic?) event that caused deposition of the Irene Bay Formation, platform backstepping resulted in a reduced shelf area and in a Silurian shelf-margin positioned some 10's of kilometres eastward of its Ordovician location. Carbonates continued to be deposited locally over the drowned Ordovician platform margin, forming large isolated carbonate buildups, some more than a kilometre thick and forming important oil plays (Mayr, 1980; Fig. 28). These large buildups contain a variety of facies and are discussed in this report.

At Bay Fiord (Fig. 27), two small "mudmounds" occur in carbonate strata containing olistostromes deposited on a platform slope. Although paleorelief is unknown, the size of allochthonous blocks, some 20-25m in diameter, indicates that relief at least equalled olistolith diameter. Similar coeval deposits on Cornwallis Island, where the shelf margin and slope strata are more continuously exposed, suggest a possible (compacted) relief of at least 200m for the Silurian platform at Bay Fiord.

Allochthonous blocks in thick olistostromes are commonly a stromatoporoid bindstone or rudstone, and are distinct in several ways from Bay Fiord mounds: (i) olistolith orientation (based on *in situ* fossils in blocks) is variable; whereas in mudmounds, geopetal and fossil orientation indicates that the moundrock is in place. (ii) Off-mound strata, although discontinuously exposed or represented by frost-riven rubble, indicate that the mound is *in situ* and not simply deposited as a rafted block in a debris flow. (iii) Pelagic graptolites, including one genus characteristic of the Wenlock, occur rarely in the mound rock, an unlikely situation if these were indeed blocks derived from a shallow, high-energy, rimmed shelf. (iv) Lithology of olistoliths is variable, but is commonly stromatoporoid bindstone, megalodont floatstone, or coral-stromatoporoid rudstone, lithologies unlike the moundrock. Although marine cements are common in many of the platform-derived blocks, none of the blocks shows a lithological or faunal zonation like that in the mudmounds. Therefore, the mudmounds are unlikely to be olistoliths.

The mudmound at Baumann Fiord is larger than the Bay Fiord mounds, and partly coeval with them, but built over a greater span of time. It was similarly deposited on the slope of a rimmed shelf during Phase I platform upbuilding. This structure is discussed below, based on reconnaissance studies.

BAY FIORD "MUDMOUNDS"

Enclosing strata

Strata above and below the mounds are similar (Fig.30). Olistostromes contain large blocks up to 25m in diameter and are laterally discontinuous. They are up to 3m thick, but commonly thinner, and are interbedded with thinly bedded, laminated marlstone, calcareous or silty shale, and limestone. The latter are commonly allodapic and consist of bioclasts of crinoids, brachiopods, corals, and stromatoporoids. Slumps, substratal synaeresis cracks, faults, and rarely cross-bedded encrinites are present. Graptolites, conularids, and monaxonous or diaxonous spicules, are also present in some shale or marlstone. Olistoliths rarely protrude above upper olistostrome surfaces and appear to be encrusted by iron oxides and bryozoans and other epibionts.

The contact with the overlying Goose Fiord Formation (Fig.31) is erosional, and is represented by encrinite channel complexes deposited on a progradational ramp. The base of this unit correlates approximately with the base of the Barlow Inlet Formation exposed on Cornwallis Island. This formational contact, although perhaps slightly diachronous regionally, is close to the Ludlow-Pridoli boundary.

Fig.30. Stratigraphic section containing crinoid-stromatoporoid "mudmounds". See Figure 27 for location of section at Bay Fiord, Ellesmere Island. Legend for this figure is the same as for Fig.32.

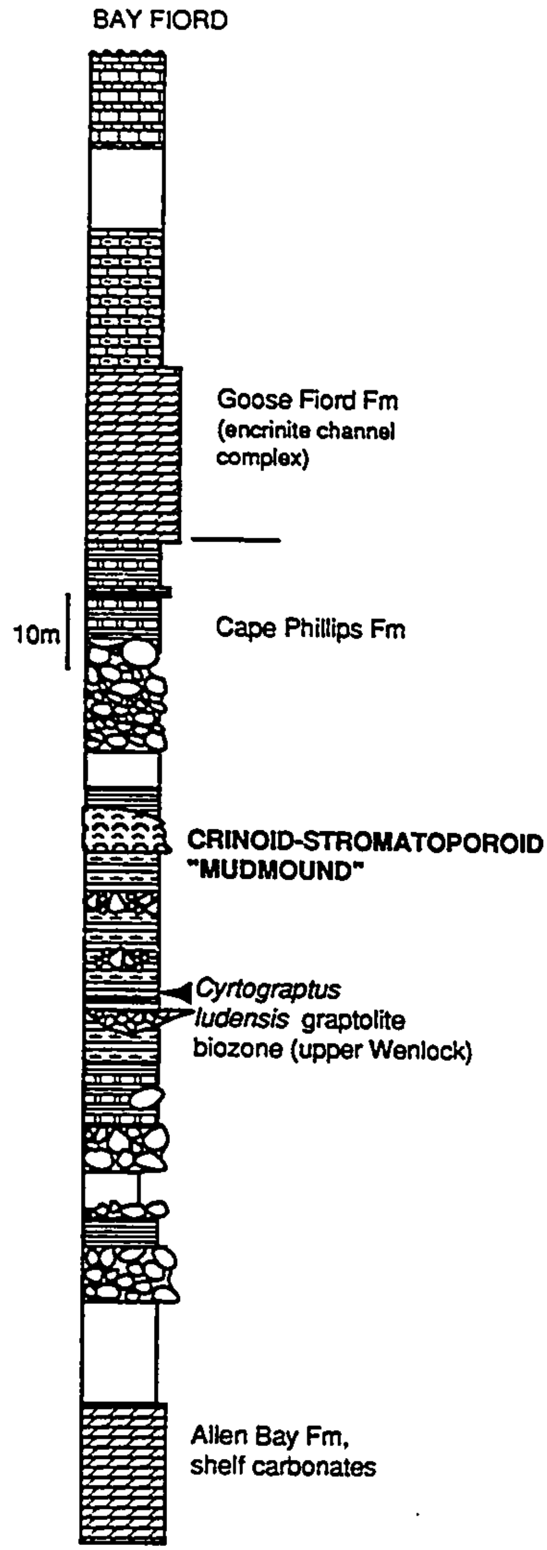
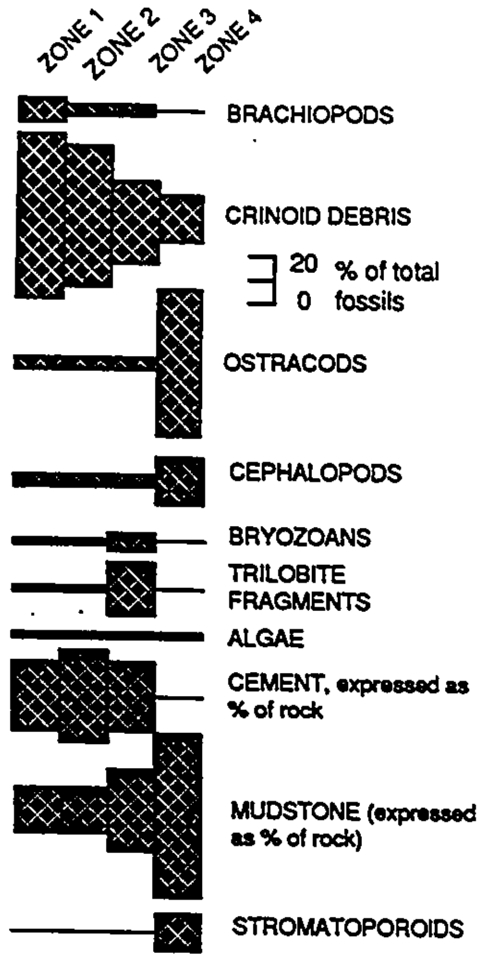
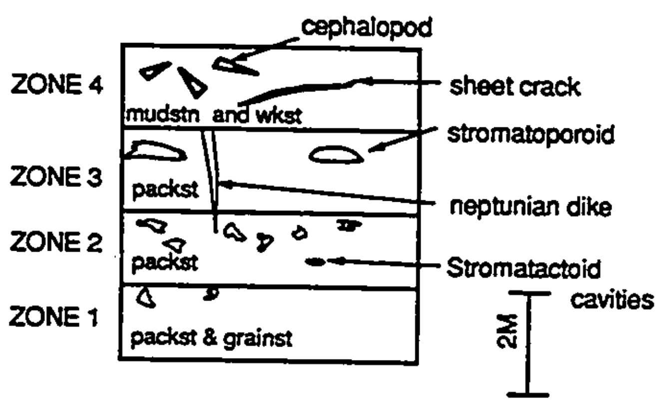

BAY FIORD
Goose Fiord Fm
(encrinite channel complex)
10m
Cape Phillips Fm
CRINOID-STROMATOPOROID "MUDMOUND"
Cyrtograptus ludensis graptolite biozone (upper Wenlock)
Allen Bay Fm, shelf carbonates

Fig.31. Schematic representation of main constituents of crinoid-stromatoporoid "mudmounds" at Bay Fiord.


ZONE 1
ZONE 2
ZONE 3
ZONE 4
BRACHIOPODS
CRINOID DEBRIS
20 % of total
0 fossils
OSTRACODS
CEPHALOPODS
BRYOZOANS
TRILOBITE FRAGMENTS
ALGAE
CEMENT, expressed as % of rock
MUDSTONE (expressed as % of rock)
STROMATOPOROIDS


cephalopod
ZONE 4
sheet crack
mudstn and wkst
stromatoporoid
ZONE 3
packst
neptunian dike
ZONE 2
packst
Stromatactoid cavities
ZONE 1
packst & grainst
2M

Mudmounds occur in strata 20m above beds containing *Cyrtograptus ludensis*, indicating that the mounds are latest Wenlock or earliest Ludlow in age, depending on the biozone thickness, which in this immediate area is presently unknown. The *C. ludensis* biozone has been established on southern Ellesmere Island in rocks of a similar platform slope setting. There, it is estimated to be about 30m thick.

Mound margins are preserved as in place frost-riven rubble; therefore, the estimated mound depositional relief is unknown, but a lateral facies change from moundrock to packstone or wackestone is observed. Strata distal to and immediately above or below the mounds are predominantly thinly bedded, thinly laminated, calcareous shale with various amounts of interbedded marlstone, skeletal wackestone, and olistostromes. Complete brachiosponges and incomplete reticulate, typically pyritized polyactines and monactines are preserved on some shale bedding planes together with uncommon graptolites. Allochthonous bioclastic material in many mass-flow deposits is typically fragmentary, silicified, and comprise bryozoans, stromatoporoids, calcareous algae, and brachiopods.

Zonation in the moundrock

Of the two exposed mudmounds of the slope sequence, the northern most is better exposed. It is 5.5m thick and 30 to 40m wide (Plate 10d). The poorly exposed southern mound is approximately 3 to 8m thick, but much of the outcrop is dislodged, obscuring stratigraphic relationships. The more completely represented mudmound consists of five beds, two of which contain the following distinct fossil and lithological zones (Fig. 31): *zone one* is encrinite packstone to grainstone; *zone two* is encrinite packstone and cementstone; *zone three* is encrinite packstone and stromatoporoid floatstone; and *zone 4* is red-stained mudstone and wackestone. Other mound beds do not contain all four zones, and an admixture of zones 1-3 predominates. Zones 1 to 3 are generally a medium pinkish grey-brown weathering limestone, and contrast to the predominantly red-stained and grey lime mudstones of zone 4. The top-most bed is predominantly zone 4 lithology variably intercalated with crinoidal debris.

Zone 1, basal encrinite packstone and grainstone

Cathodoluminescence petrography was effective in elucidating primary textures in the recrystallized mound rock, particularly in zones 1 to 3. Relative abundances of main allochthonous and autochthonous constituents are given in Figure 31. Crinoid fragments are commonly micritized, or have micrite envelopes (Plate 10e), and syntaxial cements. Stromatactoid cavities locally constitute up to 60% of the moundrock volume (Plate 10f), and perched micrite is present in only a single thin section of encrinite grainstone; otherwise, variably distributed interparticle micrite, in some cases represented as cryptomicrobial and clotted micrite (Plate 11a-d), is predominant.

Complete and disarticulate trilobites including *Encrinurus ?arcticus* are present along with orthoconic cephalopods and broken, disarticulate brachiopod and ostracod valves. Some brachiopods show moderate roundness.

Discontinuous stromatactoid cavities occur at one of the sharp contacts between zone 4 and zone 1 and protrude into overlying zone four mound rock. These marine cement-filled cavities are flat-based structures with digitate upper surfaces and red-stained, ostracod-rich, bioclastic, rarely peloidal, geopetal wackestone bases (Plate 10f).

Zone 2, middle encrinite packstone to cementstone

Zone 2 abruptly overlies zone 1 and is similarly encrinite packstone (Fig.31, Plate 10a); however, it contains abundant stromatactoid cavities. Crinoid ossicles are the predominant bioclast; brachiopods, trilobite fragments, branching trepostomate bryozoans, and solenoporid and dasyclad algae are less common. As in other zones, some crinoidal debris is partially micritized and has micrite envelopes, or is completely micritized and represented by peloids.

Numerous stromatactis-like cavities form a conspicuous fabric in zone 2 (Plate 10f). The term "stromatactoid cavities" is used in this study to differentiate cement-filled cavities from "true" stromatactis, as defined by Bathurst (1982, p.167). Based on a global study of mudmounds, he recognized *five* criteria distinctive to stromatactis. In most of the present study examples, however, cement-filled cavities do not show all five characteristics. Cavities with digitate upper surfaces and flat geopetal bases, although common, seldom form a three-dimensional network in the Bay Fiord moundrock and are herein simply referred to as stromatactoids or stromatactoid structures. So-called "stromatactis" described in the Canadian Arctic by Packard (1985) does not satisfy Bathurst's five criteria (Bathurst, 1982, p.167) and is thus similarly referred to as stromatactoids. This difference is perhaps important in the interpretation of the origin of stromatactis.

Cavities in Bay Fiord mudmounds are 5 to 20cm long by 2 to 3cm high. They are generally irregular, clustered, locally forming a three-dimensional interconnected network, and account for up to 80% of some exposures of the moundrock.

Zone 3, upper encrinite packstone and stromatoporoid floatstone

This zone is lithologically very similar to the two underlying zones, although stromatactoids are uncommon, and large tabular stromatoporoids are present, locally forming floatstone. Identifiable stromatoporoids include *?Clathrodictyon* sp. (identification by T.Bolton), *Densistroma* sp., and *?Actinostroma* sp.; however, intense recrystallization has obscured original stromatoporoid structure, and identification is questionable. *?Clathrodictyon* sp. was the largest stromatoporoid observed, measuring some 80cm in width and 40cm in height. It is a subhemispherical specimen with thick interlaminated sediment and a ragged coenosteum margin. The specimen contains abundant small, 2mm in diameter caunopore tubes, within stromatoporoid laminae or in cavities between laminae. The tubes are commonly entombed in fibrous, inclusion-rich cement and are filled with clear, non-luminescent cement. Subvertical tabulae in tubes, although poorly preserved, are evidence for a commensal syringoporid coral interpretation for these tubes. The coral-sponge relationship is common in Silurian reefs (Mistiaen, 1984).

This zone is predominantly crinoid-rich packstone, with variable and interstitial cryptomicrobial and clotted micrite. Accessory bioclasts include fragmentary dasyclad and solenoporid algae, bryozoans, brachiopods, and peloids.

Zone 4, red-stained, ostracod-rich lime mudstone and wackestone

Orthoconic cephalopods, disarticulate ostracods, algae, crinoids, planispiral and conispiral gastropods, and small brachiopods are the predominant fossils in discontinuously laminated (burrowed?) wackestone. Detrital quartz silt, and clays constitute less than 5% of the moundrock. Small columnar, frondescent crusts resembling *Frutexites* (as described by Playford *et al.*, 1976) occur at several horizons in this zone. Some fronds have opaque to semi-opaque perimeters (Plates 10b, c; 11a) and pellet-like inclusions. They are distinctive in outcrop, weathering dark grey to black and white, which contrasts with the general grey and red weathering of stylolitized zone 4 lime muds.

Red-stained lime mudstone, similar to zone 4 lithology, is present in stromatactoid cavities, sheet cracks (Plate 11a), and neptunian dykes, and is interlaminated with some marine cements. Ostracods and, where fractures were large enough, orthoconic cephalopods are common in lime mud infillings. The top bed of the mound is grey- and red-stained, stylolitized lime mudstone, representing latest and thickest development of zone 4 carbonates.

Diagenetic features

Neptunian dikes and sheet cracks (sills)

Subhorizontal sheet cracks are present in zones 1 to 3, but are most common in zone 4 lime mudstone. They are filled with interlaminated early marine grey- and red-stained lime mudstone, and bladed and fibrous cements. Most cracks are solitary, thus are unlike zebra rocks described in some Cambrian, Ordovician, and Silurian mudmounds, and in peritidal strata (e.g. Myrow, 1990, Fischer, 1964; Ross, 1972). Clotted micrite commonly lines dikes and sheet cracks and is commonly composed of a thin, basal, uniform, undulatory, 0.1mm thick microspar band and an upper 1 to 2mm thick clotted micritic band (Plate 11d). This bipartite zoning is repeated several times within some sills or dykes.

Dikes are filled with predominantly crinoid-rich lime mudstone and small amounts of red and grey lime mudstone and are not observed to cross-cut lime mudstone of zone 4. Orthoconic cephalopods and clasts of moundrock, in places containing fibrous cements, are also present together with the crinoidal infills. Some of the dyke intraclasts have an encrusting clotted micrite fabric, similar to the fabrics of marine cements in moundrock and in sheet cracks (Plate 11d). Radiaxial calcite and bladed spar are common in cracks and are interlaminated locally with micritic sediment (Plate 11c).

Cements

Cements are particularly common in zones 1 to 3 and are an important component of the mounds and stromatactoid infilling (Fig.31; Plate 10f). Together, radiaxial fibrous and fascicular optic calcite are most common and form rinds up to 1 or 2cm thick (Table 9, cement 2). The chemistry of the cements is presently unknown, and interpretations are empirical, based on comparisons with studies of similar reef cements (for example: Kerans, *et al.*, 1986; Walls and Burrowes, 1983). Five broad cement types are recognized, three of which are interpreted as early diagenetic and probably marine (Table 9). Volumetrically, radiaxial fibrous cement is most abundant, accounting for up to 60% of the rock in zone 2 (Table 9, cement 2). This cement is isopachous, turbid, contains micro-dolomite inclusions, is banded and rarely bored, and has formed irrespective of substrate type (Plates 10f; 11c). Some cement bands are distally micritized and some accompany clotted micrite. Bands of fibrous cement are either non-luminescent, or show dull or moderate, mottled (blotchy) luminescence, and some are interlaminated with less turbid, clear, non-luminescent, bladed isopachous calcite (Plate 11c). The clear bands are commonly thinner than turbid cement bands and are distally euhedral forming abrupt scalenohedral or acute rhombic terminations. The clear bands are abruptly overlain and gradationally underlain by turbid radiaxial fibrous cement. Uncommon dasyclad stems and echinoid fragments are rarely entombed in thick isopachous rinds, and some red-stained

TABLE 9
Cements, Bay Fiord crinoid-stromatoporoid mudmounds

*CEMENT IN PLANE POLARIZED LIGHT	LUMINESCENCE PROPERTIES		IRON CONT.	MAX % ROCK	ZONES OBSERVED	INTERP.
clear scalenohedral, bladed	uniform, non-luminescent	1	Fe-poor	<5%	2 4	marine
Turbid radiaxial and fascicular optic	mottled, banded dull, moderate, and non-luminescent	2	Fe-poor	60%	1 2 3 4	marine
turbid syntaxial	slightly mottled, moderate luminescence	3	Fe-poor	60%	1 2 3	marine?
clear, blocky calcite	4a. blockv, euhedral, non-luminescent, distally zoned	4a	Fe-poor	30%	1 2 3	burial
	4b. blocky, moderate to bright luminescence, indistinctly zoned	4b	Fe-rich	20%	1 2 3	burial
turbid, sucrosic, euhedral dolomite	nonluminescent	5	Fe-rich	5%	1 2 3 4	burial

* Cement terminology based on Walker, 1989

marine sediment is interlaminated with cement bands. Where clasts or sediments occur in fibrous cements, the even centripetal bands are disrupted, becoming undulatory or rarely discontinuous.

An earlier bladed calcite and bladed, needle-like calcite is rarely present as a first generation of cement infill (Plate 11c; Table 9, cement 1). This cement is commonly blotchy or uniform, weakly luminescent, and under plane polarized light, the euhedral crystals are encased in calcilutite.

Stromatoporoids of zone 3 contain 2mm diameter caunopore tubes that have isopachous, very turbid, radiaxial blocky to bladed calcite. Abundant inclusions suggest an original acicular texture to the turbid calcite. These cements lack recognizable microdolomite inclusions, are isopachous, and are blotchy dull to moderately luminescent. This type of cement is not recognized other than in the stromatoporoid ?*Clathrodictyon* sp. within zone 3.

Fascicular optic calcite is uncommon and intergradational with radiaxial fibrous calcite cements (Table 9, cement 2). It was noted only in stromatactoid cavities, and, in plane polarized light microscopy, it appears as a series of nested cones (inferred to be cone-shaped in three dimensions). At junctions of cones, sinuous, inclusion-poor, thin spar bands are common (similar to the fascicular optic calcite described by James and Klappa, 1983). In transverse sections to cones, clear bands circumscribe irregular, circular to subpolygonal inclusion-rich structures and are blotchy dully luminescent to non-luminescent.

Syntaxial cement is conspicuous in encrinite grainstone and forms up to

2mm-thick overgrowths on pelmatozoan debris (Table 9, cement 3). The cement is turbid, appears to infill primary space, and is moderately luminescent with various proportions of microdolomite inclusions. Dolomite inclusions, however, tend to be more common in bioclasts than in overgrowths.

Uniformly non-luminescent blocky, iron-free cements are commonly younger than turbid syntaxial, radiaxial fibrous, and fascicular optic cements (Table 9, cement 4a). These cements account for up to 20% of the moundrock and vary in form: between grains, they are granular equant; in primary voids, they are euhedral bladed, or scalenohedral and typically are epitaxial to earlier isopachous fibrous and syntaxial cements. Distal portions of the large euhedral cements of this type are compositionally zoned.

Coarse, equant, blocky, moderately to brightly luminescent, dully zoned, iron-free cement has occluded many of the larger pores (Table 9, cement 4b). Proximal indistinct zoning is very bright, whereas distal zones are moderately to brightly zoned. Coarse equant, blocky, iron-rich calcite and a late-stage, euhedral, coarse, ankeritic dolomite occludes the largest of pores (Table 9, cement 5), but are volumetrically minor compared to other cements.

Minor authigenic microcrystalline quartz is variably present, and is most common in zone 4 lime mudstones. Cements and silicification are cross-cut by at least two sets of fractures. Both are filled with iron-rich, equant, blocky, brightly and dully zoned luminescent calcite (stage 4b cement, Table 9), calcisiltite infill, and iron-rich dolomite and calcite.

Diagenetic history

Marine cements: radiaxial and fascicular optic calcite

Radiaxial fibrous, fascicular optic, and radial calcite cements likely resulted from marine water movement through an interconnecting porous network in the moundrock (James *et al.*, 1976; Harris, *et al.*, 1985). Predominant fibrous radiaxial calcite probably represents a neomorphosed high-magnesium calcite, particularly where cements are associated with microdolomite inclusions (Lohmann and Meyers, 1977). Radiaxial calcite is commonly interlaminated with clear bladed calcite and rarely with calcilutite, and may be ascribed to (i) variation in crystal growth and boring rate as observed in recent radiaxial cements (Saller, 1986), or to (ii) variation in growth rate and crystal splitting rates in ancients carbonates (see Kendall, 1985). Banding morphology is analogous to that described by Kendall (1985) and unlike the dark, abruptly terminated turbid bands that are symmetrical to outer crystal growth surfaces (*sensu* Saller, 1986), suggesting that banding is perhaps related to fluctuating pore waters rather than to crystal growth rate and algal infestations. A similar mechanism has been invoked to explain the juxtapositioning of fascicular optic and radiaxial calcite (Kendall, 1985), and it is commonly observed in many Phanerozoic reef cavities; for example, substrate-selective cements (radiaxial cements) imply low saturation with respect to calcite, whereas fabric selective cements imply the reverse (Mazullo, 1980).

Competency differences facilitated through the early lithification of the mound rock and interbedded unlithified marls caused fractures to develop in the moundrock. Due to a similar cement stratigraphy and history of infilling, sheet cracks (neptunian sills) and dykes formed at about the same time.

Marine cements were precipitated between stromatoporoid laminae and around internal caunopore tubes. Cementation was probably synchronous with sediment winnowing from between laminae. This is evident from the discontinuous nature of the calcilutite and abundant cement-entombed caunopore tubes that cross-cut voids.

Burial cements

Precipitation of non-luminescent "4a" (Table 9) burial cements probably occurred in pore waters deficient in the principal luminescence activator (Mn) in oxidized meteoric phreatic ground waters. These nonluminescent cements were overgrown by brightly luminescent, iron-rich calcite that was precipitated probably from deep(?) burial pore waters with main luminescence activators in the reduced state in waters, and therefore not incorporated into the calcite. This cement phase also fills numerous late stage fractures perhaps produced by the first main episode of deformation of the Franklinian platform, during the Caledonian orogeny. The last cementation phase recognized is iron-rich, non-luminescent dolomite, which infilled some portions of fractures and the centres of largest primary voids, and represents final and deepest? moundrock burial cementation.

Interpretation

Cross stratification, occurring rarely in coeval mound strata, and occasional rounding and moderate clast sorting indicate traction current transport of autochthonous and/or allochthonous mound skeletal clasts. In recent platform settings, episodic bank margin bottom currents extending along the upper portions of slopes have been reported by Mullins, and others (1980), and may have existed in this Silurian setting. These inferred currents could have, in part, controlled the depth of mound growth on the foreslope. In terms of their genesis, buildups can be compared to modern deep-water lithoherms of the Straits of Florida, where sediment driven by strong bottom currents (at depths of 600-700m) is trapped and bound by benthic organisms and cemented by marine carbonates forming large, extensive mound-like structures (Neumann *et al.*, 1977). In these very much smaller Silurian mounds, however, transported bioclastic material was stabilized by microbial communities and rare stromatoporoids, or baffled by crinoids. Currents and sedimentation were episodic, and this is evident from ragged stromatoporoid coenosteum margins (Kano, 1990) and interlaminated fibrous marine cements. Stromatoporoids were rare components, though more active in sediment binding in upper portions of mound beds.

The source of the bioclastic material is unclear. Micritization takes place in clasts down to a depth of 40m, tends to decrease grain durability, and is reported in many shallow-water tropical environments (Swinchatt, 1969). Yet, grain micritization can be abundant down to 800m depth (Hook *et al.*, 1984);

hence, its presence in described bioclasts of this study does not help resolve the problem of clast origin. Clasts can have a shallow-water or an autochthonous source, although pervasive micritization is commonly associated with deposition and stabilization of bioclasts in shallow water settings (Gawthorpe and Gutteridge, 1990). Pelagic test contribution to the moundrock was minor and is simply attributed to the scarcity of Silurian carbonate-secreting pelagic organisms. Graptolites occurring rarely in the moundrock are the only evidence of a pre-existing pelagic bioclastic input, and contrasts with the pelagic carbonate input in modern, probably analogous lithoherms.

Microbe growth in mudmounds has been used to infer an organic origin for stromatactis. Evidence of microbial growth is suggested by the presence of microdigitate ?*Frutexites* crusts (in zone 4 lime mudstone; Plate 10b,c; 11a) or as variably preserved clotted and cryptomicrobial fabric through zones 1 to 3 of the moundrock (Plate 11b,c,d). However, the origin of stromatactis is highly controversial and various biogenic, inorganic, diagenetic, and polygenetic origins have been proposed (see Tsien, 1985, for a review). The following suggestion for the origin of observed stromatactoid structures is hypothetical, and probably not valid for explaining all stromatactis (which in all likelihood are polygenetic), particularly where the form of "stromatactis" in this mound is variable, and generally unlike the form of "true" stromatactis discussed by Bathurst (1982) and by Bourque and Gignac (1983). The observed stromatactoids possibly resulted either from (i) episodic sediment winnowing of variably cemented sediment (Pratt, 1982; Wallace, 1986) or from (ii) replacement of a microbial accretion (Tsien, 1985). However, explanation (i) is unlikely, based on the large size of the stromatactoids in relation to the mud patches in the crinoidal moundrock and on the fact that a mud-rich crinoidal mound sediment would have been essentially impermeable to currents of sufficient magnitude to winnow out sediment and form voids of the observed size. Sediment winnowing was present, however, but perhaps operative on a smaller scale, for example, in winnowing mud-rich stromatoporoid interlaminae and concomitantly depositing marine cements. The concentration of stromatactoid cavities along certain horizons in the mound rock that otherwise show no apparent lithological variation is difficult to explain other than through a biological origin. Stromatactoid cavities are therefore interpreted to be of probable biological origin, and this perhaps supports Tsien's thesis (1985) of a microbial replacement.

Chemical and suspended sediment variations in early (marine) pore-water probably account for the interlamination of marine calcilutite, and radiaxial, fascicular optic, and bladed marine cements in stromatactoids and other primary voids. Early lithification caused a juxtapositioning of competent moundrock and incompetent marls, facilitating the formation of sills and dykes.

Mound growth was interrupted at least four times, each marked by deposition of abundant allochthonous? (zone 4) lime mudstone. Vagrant cephalopods or infaunal ostracods were the dominant benthos during these times, and lime muds were periodically stabilized by microbial ?*Frutexites* crusts, laterally continuous for up to several meters.

Discussion

The origin of sheet cracks is controversial and has been variously attributed to (i) mud shrinkage (Fischer, 1964); (ii) preferential erosion of unconsolidated lime muds trapped between algally bound lime muds (Pratt, 1982); (iii) mechanical (cement?) dilatation, or episodic crustification and sediment erosion (Bathurst, 1982); and (iv) movement of lithified muds downslope over slope curvature (Playford, 1980; Kendall, 1985). Sheet cracks in mudmounds formed subaqueously; hence, known subaqueous shrinkage mechanisms, such as synaeresis through salinity variation or compaction, must be invoked for the shrinkage origin of these sheet cracks. Experiments have shown that salinity variations necessary for clays to dewater subaqueously (Burst, 1965) are probably too great to have occurred in the interpreted deep-water mudmounds. Furthermore, compaction dewatering experiments have demonstrated that fractures are either random and discordant (Dangeard *et al.*, 1964), or horizontal polygonal to discontinuous and, in part, related to the primary sediment textures (Plumber and Gostin, 1981; Calver and Baillie, 1990). All subaqueous mudcracks generated through experiments and observed in nature are quite unlike zebroid rock or sheet cracks in the discussed Silurian mounds or mudmounds of any age; hence, a shrinkage origin for sheet cracks is untenable.

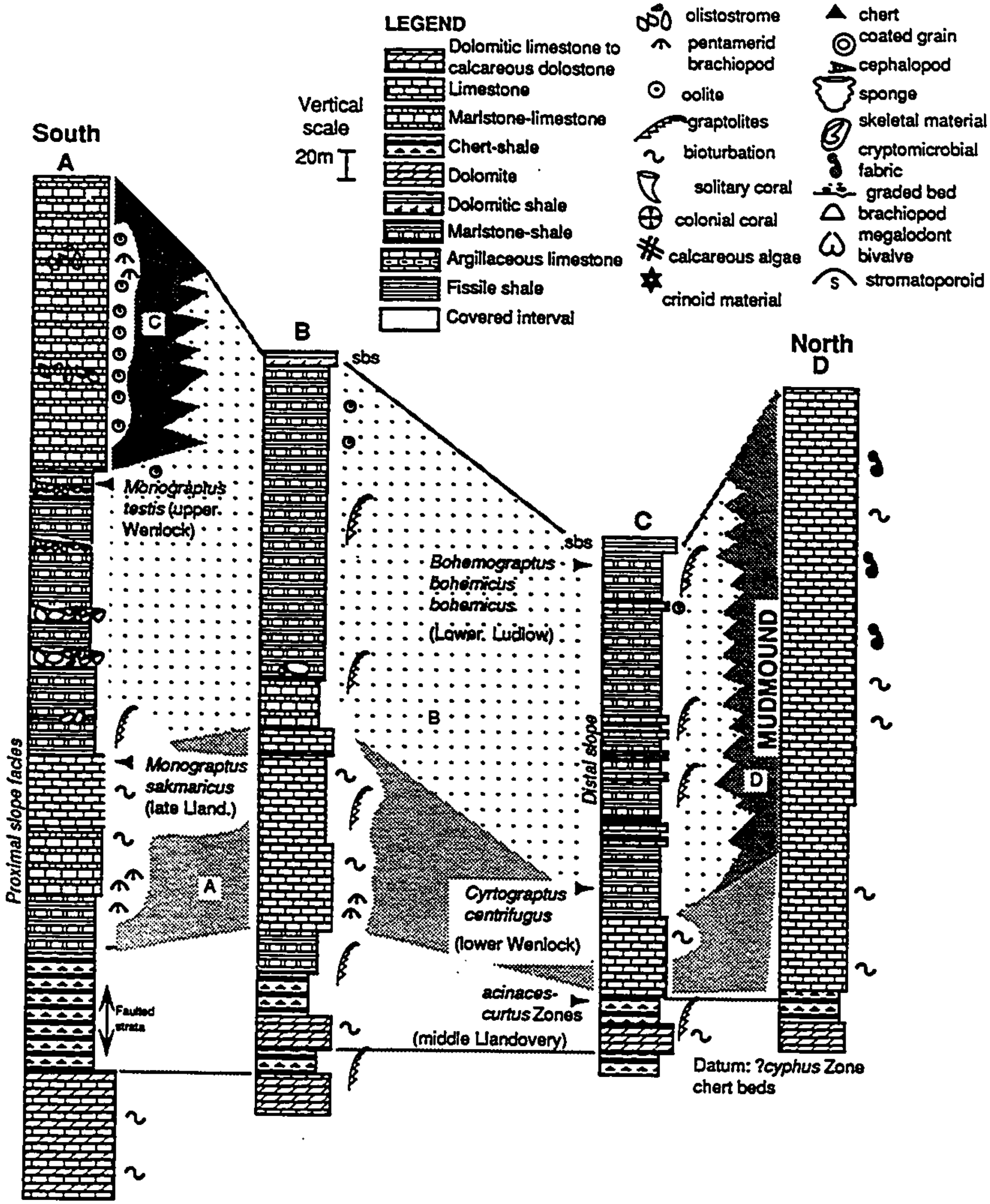
Likewise, mechanical dilatation, perhaps caused by cement emplacement, although a cited process in ancient carbonates (Coniglio, 1989; Braithwaite, 1989), is unlikely to have been a cause of the horizontal fractures. Marine cements interlaminated with calcilutite and bioclasts are difficult to explain if one invokes a displacive origin for the cracks. It appears that sheet cracks were formed through early fracturing, as indicated by grain truncation and microbial encrustations on dyke walls (Plate 11d). Fractures are inferred to have formed in response to competency contrasts and deposition on a slope and were subsequently infilled by sediment, cements, and microbial precipitates (Plate 11a,c). A similar mechanism was invoked to explain abundant marine cement-filled sills in Devonian platform foreslope strata of Australia (Playford and Cockbain, 1989). Dykes probably resulted from similar processes on the foreslope, forming at about the same time as sills, as evident from a similar internal sediment and cement stratigraphy.

BAUMANN FIORD MUDMOUND

Stratigraphic setting and anatomy

The mud buildup at Baumann Fiord (Plate 12, Fig.32) was studied on a reconnaissance scale in the summer of 1989. Future work is planned by the author on some of the remaining faunal and lithological complexities of this large buildup.

Fig.32. Stratigraphic sections of the Baumann Fiord mudmound. Section locations are given in Plate 12. Stippling pattern A is the "Llandovery carbonate" discussed in text. This unit is regionally extensive and formed the foundation for mudmound growth. Stippling pattern B is the slope facies, representing predominantly allochthonous carbonate. Stippling pattern C is the oolitic facies. The datum used for these sections is a cherty dolomite and limestone unit, which in the study area is considered to be an essentially isochronous unit.


LEGEND
Dolomitic limestone to calcareous dolostone
Limestone
Marlstone-limestone
Chert-shale
Dolomite
Dolomitic shale
Marlstone-shale
Argillaceous limestone
Fissile shale
Covered interval
olistostrome
pentamerid brachiopod
oolite
graptolites
bioturbation
solitary coral
colonial coral
calcareous algae
crinoid material
chert
coated grain
cephalopod
sponge
skeletal material
cryptomicrobial fabric
graded bed
brachiopod
megalodont bivalve
stromatoporoid
Vertical scale
20m
South
A
B
C
North
D
sbs
sbs
Monograptus testis (upper. Wenlock)
Bohemograptus bohemicus bohemicus (Lower. Ludlow)
Monograptus sakmaricus (late Lland.)
Cyrtograptus centrifugus (lower Wenlock)
acinaces-curtus Zones (middle Llandovery)
Faulted strata
Proximal slope facies
Distal slope
MUDMOUND
Datum: ?cyphus Zone chert beds

Beds 21m below the mudmound have yielded graptolites of the *Lagarograptus acinaces* (Törnquist) to *Monograptus millepeda curtus* (Obut and Sobolevskaya)-*Monograptus pectinatus pectinatus* Richter graptolite biozones (upper lower Llandovery or lower middle Llandovery), and beds overlying the buildup contain *Saetograptus fritschi linearis*, Bou ek, a biozone indicating middle Ludlow age. Given approximate time divisions for the Silurian (Holland, 1989), the 310m (photogrammetrically measured) of moundrock represents a (compacted) lime mud depositional rate of about 7cm/1000 years.

Underlying the moundrock is a chert-rich shale and laminated, thinly bedded limestone, an essentially isochronous unit recognizable throughout much of southern and central Ellesmere Island. The unit probably represents a major platform drowning event which preceded Lower to Middle Silurian platform upbuilding. Above the cherty unit, sparsely fossiliferous, burrowed, upper Llandovery lime mudstones are discontinuously present in the study area, and are thickest where associated with the position of the Silurian shelf margin. These lime mudstones are succeeded by shales in exposures representing platform foreslope strata, except at the mudmound, where they form the foundation for subsequent mudmound growth.

Coeval beds up paleoslope from the mudmound, some 20km away, contain abundant predominantly allochthonous carbonate debris, in which olistoliths up to 40m in diameter are imbedded. These were derived from shallow-water reefal and oolitic deposits, east and southeast of the mound (Plate 12, Fig.32).

Beds immediately adjacent to the mudmound lack allochthonous shallow-water blocks, but slumped beds and intraclast beds, probably derived from the mudmound, are locally common. Off-mound lithologies recognized include thinly laminated, graptolitic limestone, petroliferous dark shale, laminated marlstone, intraclast beds, and uncommon, poorly graded skeletal, cm-thick wackestone beds and black chert beds and nodules.

Off-mound strata near the top of the buildup contain ooids, uncommon stromatoporoid clasts (?*Actinostroma* sp.), and abundant pentamerids (*Kirkidium* sp. in Ludlow strata). These clasts are common in slope strata proximal to the shelf margin and are considered to have been deposited by mass flows. Pentamerids, ooids, or stromatoporoids were not noted in mound rock, but crinoid packstone or wackestone is relatively common in the proximal off-mound strata.

Preliminary observations indicate that the mound rock is predominantly homogeneous, cryptomicrobial, or clotted micrite. Macrofossils are not apparent, and stromatactoid cavities are uncommon; however, textures in these fine-grained limestones may be obscured by the rubbly exposure. Differential erosion of the more resistant mound lithology has given it a relief of some 40m over surrounding off-mound shales.

Interpretation

A middle Llandovery (?*cyphus* Zone) regional drowning event of the platform is indicated by extensive units of black chert, shale, and limestone. A return to carbonate platform upbuilding is marked by a discontinuous (upper Llandovery) burrowed lime mudstone (lowest stippled unit of Fig.32), which generally thickens and becomes more fossiliferous where underlying the Silurian shelf margin. Coeval intraclast beds present in these units are the first indication of platform depositional relief subsequent to Llandovery platform drowning. Mound growth on the foreslope was far enough away from the source of allochthonous material such that the accumulation of the lime mudstone was essentially uninterrupted from middle Llandovery to lowest Ludlow.

Discussion

The architectural detail of this structure is poorly known, although field examination of the carbonates revealed that the predominant lithology is structureless lime mudstone or clotted or pelleted lime mud; these are discussed below and are considered, in terms of bathymetry, to be similar to those interpreted for the large drowned shelf margin buildups discussed below.

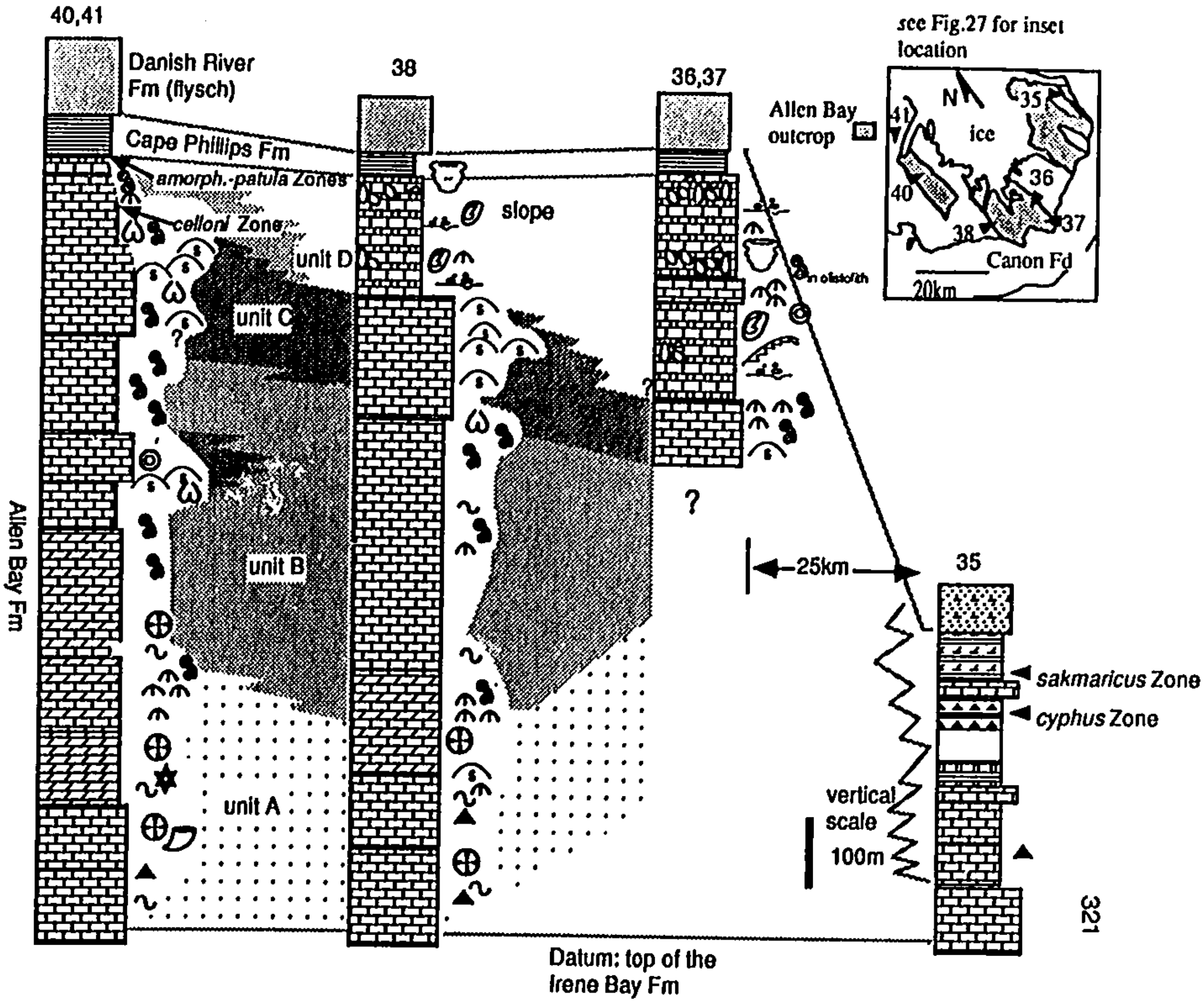
These buildups represent classic type I foreslope buildups as discussed by Wilson (1975), and are comparable to Waulsortian foreslope buildups discussed in the literature (for example Lane, 1982; King, 1986). In the Silurian, however, comparable examples are rare. Mesolella *et al.* (1974) and Sears and Lucia (1979) described Michigan Basin Silurian pinnacle reefs that occupied a similar stratigraphic setting to the Baumann Fiord mudmound. These buildups have variable mudstone cores and stromatoporoid caps, representing an overall shallowing upward sequence. Mudstone cores, however, are either stromatactis rich or coral rich, and are generally unlike the dominant cryptomicrobial and clotted micrite observed in the Baumann Fiord mudmound. However, Mesolella *et al.* (1974) described a structureless algal unit in a pinnacle reef which has similar mound textures, although a detailed description has not been published for the largely recrystallized and dolomitized Michigan Basin examples.

MUD BUILDUPS ON THE DROWNED ORDOVICIAN SHELF EDGE

Description

These large structures were visited briefly in the summers of 1988 and 1989. At Trold Fiord (Fig.27), only the slope sequence of one of these buildups is exposed. Two other buildups are present farther to the north, on the north and south sides of Cañon Fiord (Fig.27,33, Plate 13a). An early geological map of the area (Kerr and Thorsteinsson, 1972) had represented these two buildups as a single structure straddling Cañon Fiord. Although the crucial outcrops are either beneath the fiord or are covered by the younger Sverdrup basin rocks, present

Fig.33. Stratigraphic sections of large mud buildup and associated facies at Cañon Fiord. Unit A represents burrowed lime mudstone, B microbial carbonate, C stromatoporoid ("reefy") bindstone and floatstone, and D allochthonous slope facies. Legend as in Fig.32.



40,41
Danish River Fm (flysch)
Cape Phillips Fm
amorph.-patula Zones
celloni Zone
unit D
unit C
unit B
unit A
Allen Bay Fm
38
slope
36,37
?
Allen Bay outcrop
see Fig.27 for inset location
N
ice
35
36
37
38
40
41
Canon Fd
20km
25km
35
sakmaricus Zone
cyphus Zone
vertical scale
100m
Datum: top of the Irene Bay Fm

data suggest that at least the younger parts of the buildup were separate. Although exposures on both sides of Cañon Fiord were examined, lithological sampling and measurement of stratigraphic sections on the northern buildup were more extensive, hence provided a better basis for interpreting these buildups. Logistical problems limited sampling and measurement of stratigraphic sections in these remote study sites. The structures may be more complex than initially apparent, and additional field work on this important aspect of the Silurian sequence of the Arctic is in progress. The sections represented in Figure 33 are abbreviated schematic representations of these very thick structures. Pertinent data from additional nearby measured sections were used in the interpretation.

The lowest units of these buildups are remarkably similar to coeval cratonward strata throughout the Canadian Arctic and are predominantly sparsely fossiliferous, labyrinthine mottled dolomitic limestone, with rarely occurring fossils including *Maclurites* sp., orthoconic cephalopods, tabular favositids, sponges, calcareous algae, and crinoidal debris (Plate 13b). These carbonates are well developed at the bases of buildups and shelfward, where they form a rostrum for subsequent platform upbuilding. Coeval strata are predominantly a cherty shale, and thinly bedded limestone.

Large amorphous brown chert nodules are present in lower lime mudstone beds (unit A, Fig.33) at Cañon Fiord. These beds are perhaps correlative with extensive basinal cherty deposits, that span the interval from near the Ordovician-Silurian boundary to within the *cyphus* (upper lower Llandovery) graptolite biozone. Above this, about 200-300m above the Irene Bay Formation, *Pentamerus* is abundant and forms rudstone beds up to 3m thick (Plate 13c). These are interbedded rarely with crinoidal grainstone and packstone, tabular stromatoporoid and favositid floatstone, and sparsely fossiliferous bioturbated lime mudstone. The pentamerids are typically robust, disarticulate, randomly oriented, and rarely coated (Plate 13e). Transition of the predominantly burrowed lime mudstone of unit A to the overlying clotted micrite (unit B) is gradational (Fig.33).

Microbial carbonate of unit B forms thick intervals throughout the buildups (Fig.33). The microbial texture is present in several forms (Plates 13f, 14a-f); most common is peloidal, clotted micrite, with fenestrae that are .5mm to 1cm in maximum dimension (Plate 14a,b,d,f). Fenestrae are commonly tabular with digitate or flat roofs and floors and are associated with abundant variably sized, but smaller amorphous fenestrae. These structures are unlike stromatactis in that they contain predominantly burial cements, including an early, brightly luminescent, clear, blocky, calcite and a late-stage, non-luminescent, iron-poor, clear, blocky calcite. Less commonly observed lithologies in the buildups include a clean pelsparite, coated and micritized grainstone (Plate 14c), oncoidal limestone (Plate 13c) (locally abundant), and pentamerid tempestites. The relative proportions of these main lithologies are not precisely known; although the clotted fenestral micrite is clearly most abundant and contains very few fossils or fossiliferous interbeds, but those that do occur include fragmented trilobites, brachiopods, and crinoidal detritus. Some thin intraclast beds also occur. Peloidal or clotted laminae approximately 1cm-thick are abundant and appear to be variably present throughout the buildup.

Slope sediments (Fig.33, unit D equivalent) coeval with the microbial carbonate contain large allochthonous blocks, up to 8m in diameter, composed predominantly of homogeneous clotted micrite. Associated with the olistostromes are oncoids, fasciculate rugosans (*Stylopleura* sp.), pentamerids, rarely occurring stromatoporoids, lithistid sponges (Plate 13d), and articulate crinoid stems. Near the top of the mud buildup, reefal deposits, predominatly tabular stromatoporid bindstone, are most abundant (Fig.33, unit C). These deposits intergrade with clotted carbonate, are richly fossiliferous, and include conspicuous megalodont bivalves. These bivalves are articulated and displaced, or *in situ* in bioturbated lime mudstones, and generally are mutually exclusive of *in situ* pentamerids (*Pentamerus?* sp.). The bioturbated lime mudstone are interbedded with mudstones containing "reefy" taxa, and also form units of disarticulate bivalve floatstone.

In summary, the buildup comprises a basal bioturbated lime mudstone unit (unit A); a middle (unit B) clotted micrite; a upper middle (unit C) "reefy" unit with abundant stromatoporoids; and an topmost unit (unit D) of clotted micrite (Fig.33).

Interpretation

Subsequent to an Upper Ordovician platform drowning event, carbonate buildups were established preferentially in three areas which lie approximately along the drowned Ordovician shelf margin (Fig.27,28). Initially, these buildups were formed of sub-storm wave base(?) burrowed, sparsely fossiliferous carbonates (Fig.33, unit A; Plate 13c), but with continued carbonate accumulation in the Upper Ordovician and lowest Silurian, growth of microbial carbonate replaced the burrowed lime muds. Microbial carbonates (unit "B", Fig.33; Plate 14a-f) accumulated in somewhat shallower water above storm wave base; this interpretation is based on the occurrence of pentamerid rudstone beds, some representing shelly tempestites containing thickened *Pentamerus* sp. The thickness of prismatic calcite in these well-preserved pentamerid valves was an adaptation that increased stability in muddy substrates and apparently reflects an adaptation to a "high-energy bank environment" (Baarl, 1988; Johnson, 1989), or perhaps an environment subject to periodic (high intensity?) storms. The occurrence of intraclast beds, oncoid beds, and, very rarely oolites (some incorporated into the microbialites) also supports an agitated buildup environment. Also, the different types of microbial carbonate, including cryptic structures, spherical structures, and uncommon stromatolitic structures (terminology after Burne and Moore, 1987), suggest that numerous varieties of benthic microbial communities contributed to upbuilding. Furthermore, the lack of evidence for widespread marine cementation (common in several Silurian mudmounds described in the literature) but the establishment of substantial buildup relief and associated coarse allochthonous sediment flux, indicates that biologically influenced calcification was very important for mound upbuilding.

Abundant pentamerids and other rarely occurring invertebrates (isolated stromatoporoids, favositid and rugose corals) inhabited probably slightly deeper water sediments coeval with the microbialites. The general exclusion of *in situ* pentamerids from microbial carbonates and the upward shallowing sequence of the buildups, from bioturbated beds, to pentamerid-rich lime mudstone, to microbial carbonate, to stromatoporoid bindstone, suggest a deeper water interpretation for pentamerids than for microbial carbonate. However, this relationship is not clear, as in other settings, pentamerids are associated with shallow-water coral-stromatoporoid reefal deposits.

The upward transition of microbial carbonates to "reefy" stromatoporoid bindstone suggests shallowing, caused either by carbonate upbuilding or by a relative sea-level fall. During the stromatoporoid stage (unit "C", Fig. 33), faunal diversity increased, and skeletal grainstone and in-place skeletal metazoans (predominantly laminar stromatoporoids) became more abundant. The transition from microbial to stromatoporoidal carbonates was, however, gradational, and microbial carbonate is commonly interbedded with "reefy" strata. Stromatoporoid floatstone and bindstone (unit C; Fig.33) represent the shallowest level attained by the buildup, and were deposited probably within or just below fair-weather wave base, which is generally considered by most authors to be less than about

15m below sea level (Fagerstrom, 1988).

Subsequent to reefal unit C deposition, relative deepening resulted in return to deposition of predominantly microbial carbonates. The transition was again gradual, producing stromatoporoid-rich lithologies that are commonly interbedded with microbial carbonate. This episode was also marked by a general reduction of buildup area and slope strata onlapping. Active upbuilding of microbial carbonates, although areally, generally diminished, is evident from abundant, coarse coeval allochthonous sediments in the form of thick olistostromes containing blocks largely of microbial carbonate blocks, some up to 8m in diameter. This reduction of the area occupied by the buildup occurred during the Wenlock, extending from the *amorphognathoides* through to the *patula* (late Llandovery to middle Wenlock) conodont zones. Relief relative to the contiguous off-buildup shale was probably greater at this time, as suggested by coarse olistostromes preserved in coeval slope strata.

Megalodont bivalves are associated with latest Llandovery and early Wenlock "reefy" strata in this area and represent the oldest known occurrences of this bivalve. They generally occur mutually exclusive of in place pentamerids, but are associated with stromatoporoid bindstone units here and throughout the Arctic (pers. obsrv.). The relationship suggests that these bivalves inhabited a shallower bathymetric position than the pentamerids and were clearly inhabitants of protected reef-associated environments. However, the inferred depths and facies association for these organisms are not necessarily applicable to all situations, such as the occurrence of pentamerids encrusted by thick fibrous marine cements in upper Llandovery shallow-water reefal deposits of the Canadian Arctic (de Freitas and Dixon, 1989a).

Abrupt drowning occurred in middle Wenlock time and resulted in deposition of fissile, petroliferous black shale. This unit is distinctive and condensed compared to coeval "off-buildup" strata. The relationship suggests that the intrinsic topographic highs represented by these buildups promoted local starved sedimentation, quite unlike the setting represented by coeval strata 10km to the south.

Discussion

Although many drowned carbonate platforms have been discussed in the literature (eg. Schlager, 1989; Arthur and Schlanger, 1979; Barnaby and Read, 1990), none of the known Silurian examples are analogous to the setting of these Silurian mud buildups of Ellesmere Island. Late Triassic Dachstein platform carbonates (Schaeffer, 1979) show a drowned platform profile and an incipient development of buildups. These buildups similarly represent upward shallowing, with basal skeletal sand bodies, a middle mudmound unit, and an upper coral-algal bindstone. Playford and Cockbain (1989) and Playford *et al.* (1989) described Devonian pinnacle reefs that developed subsequently to drowning of a carbonate platform. Drowned Devonian platforms were also reported in central Europe (Krebs, 1974), and southwest England (Burchette, 1981; Scrutton, 1977a,b).

In England, stromatactis mudmounds and fossil atolls, and in Europe, isolated stromatoporoid reefs and stromatactis mudmounds were discontinuously established over a drowned Devonian platform.

In the Danish Arctic, Silurian platform deposits contain "intrashelf" stromatactis mudmound-cored buildups (Sønderholm and Harland, 1989; Table 8), capped by stromatoporoid units, suggesting overall upward shallowing. These occur in a drowned platform sequence showing major stratigraphic trends that are similar to those of the Canadian Arctic. In the southwestern US, a drowned Silurian platform sequence is known, but does not show the incipient upbuilding such as is known from the Silurian strata of the Canadian Arctic (Hurst and Sheehan, 1985).

Bourque *et al.* (1986) described large Silurian stromatactis mounds, which developed at the margins of uplifted basin blocks and grew upward into "algal" carbonate. These buildups are unlike the Arctic mud buildups in that they are built of a consortium of corals, stromatoporoids (minor), and skeletal and non-skeletal algae and contain abundant, large marine-cement-filled cavities. The latter reef type, referred to as coralgal biolithite reefs in this study, occur in the vicinity of Baumann Fiord. They existed on the same Silurian platform slope as the mudmound illustrated in Figure 32. The dense, encrusting microbial micrite locally with sponges (in the Arctic only) and corals of the coralgal biolithite pinnacle reefs is, on a microfacies scale, similar to and intergradational with laminar, fenestral, clotted and pelleted, spar-rich micrite of the mud buildups of this report. Calcareous algae (as well as skeletal metazoans) are, however, lacking in the latter buildups. These lithological differences suggest that the two distinct types of micrite-rich buildups occupied different bathymetric settings.

Large olistoliths (up to 20m in diamter) were derived from coralgal biolithite pinnacles in the Canadian Arctic, whereas relatively thin intraclast-rich conglomerate beds are associated with the contiguous mudmound (in Fig.32). The contrasting reef-flank facies indicates that the coralgal biolithite pinnacles had greater relief than for the mudmound, suggesting that the mudmound at Baumann Fiord probably did not attain a growth rate comparable to sea-level rise. Therefore, this mound perhaps formed in deeper water, a conclusion further supported by the presence of skeletal algae in coralgal biolithite, indicating significantly more illuminated conditions during growth. Also, because coralgal biolithite pinnacle reefs and coeval microbial mudmounds developed essentially the same carbonate platform setting, mudmounds may have formed cores for subsequent coralgal reef growth. However, the latter point is hypothetical, because a direct upward replacement of microbial carbonate by coralgal biolithite was not observed. Nevertheless, microbial carbonate olistoliths are present in foreslope facies which stratigraphically underlie coralgal biolithite slope facies, suggesting that microbial carbonate was succeeded by coralgal biolithite.

The relative bathymetry inferred for various "alga"-rich buildups described in this report and in the literature is unclear. Coralgal biolithite reefs (Bourque, *et al*, 1986; Clough and Blodgett 1989, 1985; and pers. obsrv. of coralgal biolithite pinnacle reef in the Canadian Arctic), lithistid sponge mudmounds (Narbonne and Dixon, 1984), stromatactis mudmounds (as discussed by Shaver and Sundermann, 1989), and Ellesmere Island mud buildups described here, all are interpreted to have occupied similar bathymetric positions, that is, between storm and fair-weather wave bases. Although a detailed description is not published for the coral-algal reef rock discussed by Mesolella (*et al.*, 1974), it nevertheless may be comparable in terms of lithology to coralgal biolithite reefs discussed above and to similarly constructed pinnacle reef deposits in the study area. Clough and Blodgett (1985) interpreted coralgal biolithites as having formed in shallow waters, but in a slightly "less agitated" setting than the shallow waters that harbour stromatoporoid reefs. However, the following features discount a "less-agitated" interpretation for coralgal biolithite reefs: (i) the presence of abundant oolites interbedded with the Canadian Arctic coralgal buildups; (ii) the presence of an almost vertical (or, at least, very steep) coralgal reef foreslope; and (iii) the lack of a coralgal facies between the microbial carbonate and stromatoporoid bindstone in the upward-shallowing sequence of the large microbial buildups discussed herein. Bourque (*et al.*, 1986) interpreted Anse à la Barbe "algal" reefs (herein considered to be equivalent to coralgal biolithite reefs) as shallow-water deposits, as are stromatoporoid-rich strata. However, based on the above relationships, algal reefs may signify an environment subject to high wave stress. Silurian reef distribution in this respect may be similar to that interpreted for some modern (mostly Pacific) algal and (Caribbean) coral reefs, that are likely controlled by wave stress.

Coralgal biolithite reefs, although rarely containing sponges, occupied a significantly different setting from that envisaged for lithistid sponge mudmounds and mud buildups described herein. The latter two reefs are both interpreted to have formed in calm waters well(?) below fair-weather wave base, but above storm wave base (Narbonne and Dixon, 1984, 1989; discussed above). Sponges (other than stromatoporoids) have not been recognized in the microbial carbonate buildups, but are abundant in the (deep-water) slope facies (Plate 13d). Although the explanation for this lack of fossil sponges is not clear, it maybe simply related to inadequate sampling, or to paleoenvironmental factors, such as water depth or turbidity.

Coeval buildups are recognized in the western Arctic, on Bathurst and Melville islands (Harrison and Bally, 1988; Trettin, 1989; Mayr, 1980). Some of these structures show complex facies associations (Mayr, 1980) and were built through to the Middle Devonian. In these areas, however, dolomitization and the lack of detailed published descriptions have precluded a comparison with the Ellesmere Island buildups.

The peloidal or clotted microfabric is abundant in the studied mud buildups, and is comparable to the peloidal fabric discussed in the literature (for example Sun and Wright, 1989; Tsien, 1985; and also alternative views of Bourque and Gignac, 1983, p.526). Most workers have concluded that this fabric is produced *in situ* by microbes, either by direct precipitation or trapping, or as a replacement product (Tsien, 1985; Kennard and James, 1986). In the Ellesmere Island buildups, the microfabric is gradational in character among and within buildups, and this suggests a common origin for these micrites. Conspicuous stromatactis is lacking, however.

The origins of stromatactis are diverse, are extensively discussed in the literature, and will not be discussed in detail here. Perhaps the general absence of stromatactis from the Ellesmere Island buildups suggests a biological precursor, adapted to a particular mudmound habitat that simply was not present in these areas (but also see preceding discussion of stromatactoids of crinoid-stromatoporid mounds). Stromatactis is similarly lacking from some pinnacle cores of the Great Lakes area, suggesting that if stromatactis is indeed a biological remnant, perhaps it represents a different setting (deeper water?) than the Ellesmere Island microbial mud buildups. Indeed, Bourque (*et al.*, 1986) suggested that Gross Morbe (truly stromatactis-rich) mudmounds may have been constructed below the photic zone.

CONCLUSIONS

Mud buildups that lack stromatactis occur in four localities on Ellesmere Island; some are large structures established over the drowned Ordovician shelf edge, or, in the example at Baumann Fiord, are somewhat smaller and established on the foreslope of the Silurian shelf margin, located 10's of kilometres cratonward of its Ordovician position. At Bay Fiord, crinoid-rich foreslope mudmounds were built on the re-established Silurian foreslope located 10's of kilometres to the southeast; there, moundrock that is predominantly skeletal packstone and wackestone and minor cryptomicrobial micrite, and is interpreted to have been deposited on a slope subject to episodic bottom currents. Inferred downslope currents transported predominantly allochthonous and/or autochthonous crinoid and other debris that was bound or trapped by microbes or crinoids, respectively. The bioclastic accumulations formed local, low-relief mound structures that, in the latest episodes of upbuilding, rarely harboured laminar stromatoporoids. Stromatactoids are common and probably represent a replacement of an original microbial accretion (*sensu* Tsien, 1985). Early lithification was pervasive and is represented by bladed, radiaxial, and fascicular optic calcite. Competency differences of moundrock and interbedded slope marls facilitated early fracture of the mound, forming neptunian sills and dykes. Three subsequent stages of burial cementation are recognized, and represent continued burial of the strata. Two late-stage fracture episodes are evident and are perhaps related to middle Paleozoic orogenesis which profoundly effected the regional structure of the study area.

During phase I platform upbuilding, two other mudmounds developed. A classic Type I (Wilson, 1975) microbial and lime mudstone mudmound was established on a slope following a middle Llandovery drowning event. This structure accumulated to about 310m in thickness, and had a relief over surrounding marls. Based on reconnaissance work, this mound is apparently predominantly burrowed unfossiliferous lime mudstone in the lower part and microbial carbonate above.

The largest buildups, also formed during phase I upbuilding, occur discontinuously over the drowned Ordovician shelf margin. The structure at Cañon Fiord displays zonation: a basal burrowed, sub-storm wave base, sparsely fossiliferous limestone, is suceeded upward by microbial carbonate, then by stromatoporoid bindstone. In this overall upward-shallowing sequence, the microbial carbonates accumulated as muds above storm wave base, and are closely associated with the *Pentamerus-Pentameroides* ecogroup of Zeigler (1965).

After maximum shallowing, deepening and concomitant onlapping of slope sediments onto the buildup reduced the effective area of the upbuilding. Great relief is associated with this episode as abundant, large blocks of microbial carbonate are associated with upbuilding. However, microbial carbonates generally have few early marine cements suggesting that biologically influenced precipitation of calcite was primary for upbuilding and supplying large olistoliths to buildup flanks. Trapping and binding of carbonate was probably minor in comparison, as the buildups were isolated structures growing in an anoxic deep-shelf setting.

In middle Wenlock time, the structures at Cañon and Trold fiords were abruptly drowned. An intrinsic paleotopographic high associated with the buildup at Cañon Fiord became a site of starved black shale deposition, which contrasts with the coeval, varied "off-buildup" lithologies.

Foreslope mudmounds at Baumann Fiord developed vertically into coralgal biolithite pinnacle reefs. These structure, in comparison to coeval mudmounds, had a greater upbuilding rate and developed steep to escarpment-like flanks. The lithology of the pinnacle reef is unlike that of the coeval mudmound or mud buildups, a difference perhaps ascribed to paleoenvironment. Coralgal biolithite pinnacles perhaps existed in shallow, well-agitated conditions that favoured skeletal metazoans (mainly corals) that were bound by thick microbialite and calcareous algal crusts. Stromatoporoids, although clearly the predominant builders of shallow-water Silurian reefs, are minor components of coralgal biolithite and, due to environmental factors such increased wave stress, may have been generally excluded from this setting.

PLATES

Plate 10. Facies of crinoid-stromatoporoid mounds. **(A)** Lithological zones (1-4) in mounds. See Figure 31 for distribution of components and textures. Hammer handles (arrow) are each 4cm in diameter. **(B)** Photomicrograph of pelsparite that forms continuous and discontinuous horizontal crusts in zone 4 lime muds. Several of these crusts are recognized, and their origin maybe related to a microbial(?) precursor (hand sample of Plate 10c). Scale bar 0.25mm long. **(C)** Hand sample of horizontal microdigitate pelmicrite crust of zone 4 lime mudstone. Shown in photomicrograph indicated in 10b. Sample is actual size. **(D)** Outcrop of crinoid-stromatoporoid mound. Gun in backpack at top of the outcrop approximately 1m long. **(E)** Photomicrograph of encrinite grainstone to packstone of zone 1. This lithology, although variable and commonly with more interstitial mud, is predominant through zones 1-3. Note abundant micrite envelopes on crinoidal debris. Scale bar 1mm long.

(F) Photomicrograph of centre of stromatactoid cavity, showing marginal turbid radiaxial calcite and central (red-stained) pelleted geopetal fill. Scale bar 1mm long.

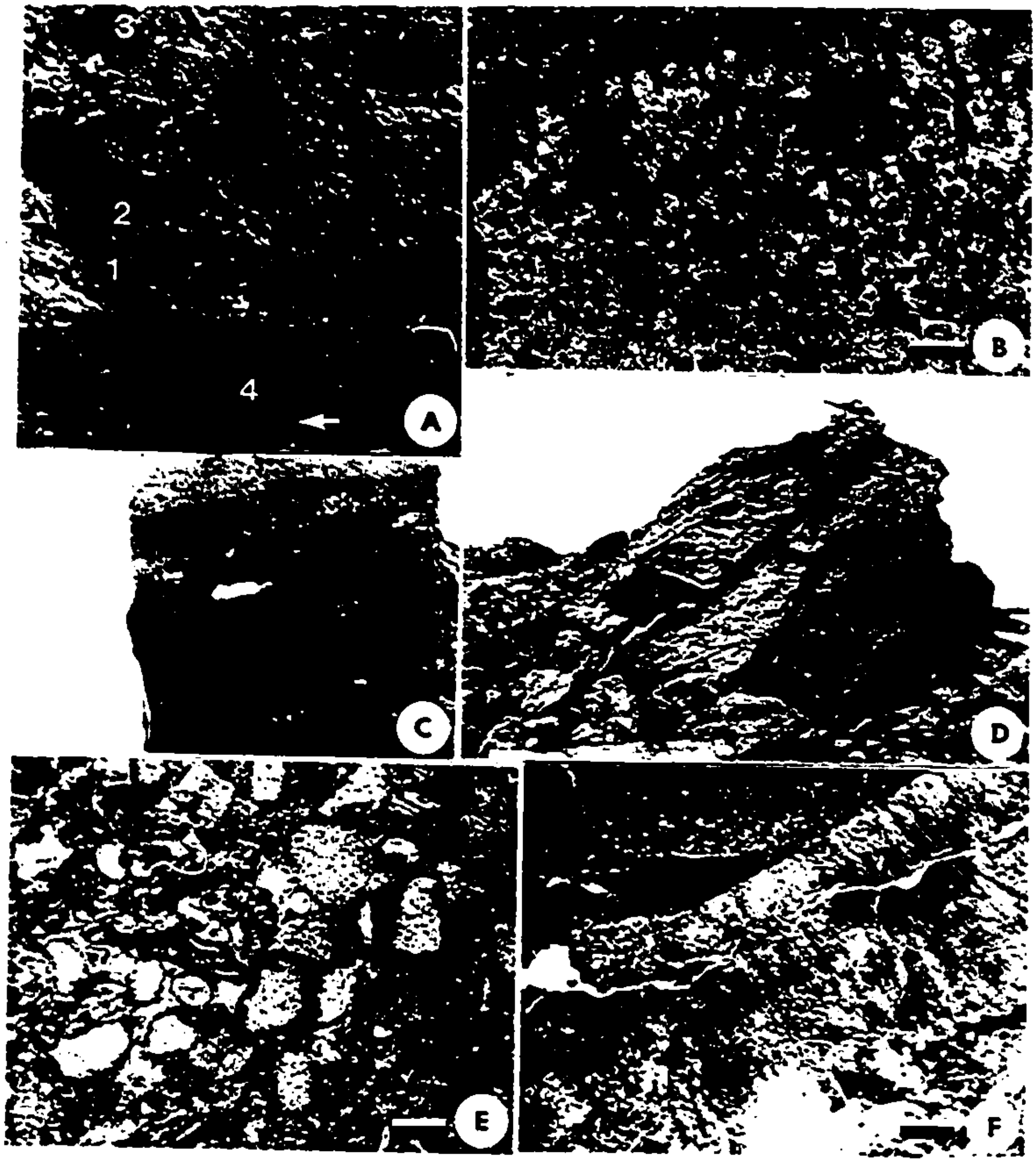

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Plate 11. Microtextures of crinoid-stromatoporoid mounds at Bay Fiord. All photographs right way up. **(A)** Horizontal microbial crust in a sheet crack of zone 4 ostracod- and cephalopod-rich lime mudstone. Spar at the base of the amorphous algal? crust is an early non-luminescent bladed calcite that precipitated on the margins of the sheet crack. Scale bar is .25mm long. **(B)** Clotted micrite (with vermiform fenestrae) intergradational with peloidal micrite. Scale bar 0.25mm long. **(C)** Clotted and peloidal micrite overlain by marine cements: an early bladed (type "1" Table 9) clear calcite, overlain by discontinuous calcilutite, then by turbid and clear radiaxial calcite. Cements above the fracture are late (burial) blocky calcite infilling. Scale bar is 0.5mm long. **(D)** Clotted micrite associated with a dyke wall and with a basal undulatory spar band separating mound rock (and fracture fillings) from clotted micrite. These micritic encrustations, typically with a basal undulatory spar band, are present in several localities in mound dykes and are interpreted as an original cyanophyte? growth. Scale bar 0.3mm long. From a dislodged block.


A
B
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Plate 12. Vertical air photograph of Baumann Fiord mudmound. Letters A-D represent stratigraphic sections of Figure 32. Foreslope mudmound outlined at locality D and equivalent slope facies outlined to north and south. North to south distance in photograph is 16km.

Plate 13. Zoned microbial stromatoporoid buildups and associated rocks, Cañon Fiord. (A) Section 41 (Fig.33) showing black shales and clastic deposits (left) abruptly overlying buildups drowned in about middle Wenlock time. Mountain in background is just under 1000m above glacier and skidoo in foreground. (B) Typical mottled dolomitic limestone characteristic of "unit a", Fig.33. Hammer is 30cm long. (C) Abundant pentamerids (represented mainly by *Pentamerus* sp.) in lower part of "unit B" Fig.33. Ice axe 1m long (lower left) is partly in snow. (D) Slope lime mudstones and interbedded calcareous fissile shales. Basal bioclasts of graded bed are lithistid sponges, mainly fragmentary *Archaeoscyphia* sp. Some sponges are very well preserved, indicating minimal transport, and are common in onlaping slope units in upper part of buildup represented in Fig.33. Other bioclasts in these beds include pentamerids (*Pentamerus?* sp.), fasciculate rugose corals, solitary rugosans, tabulate corals, and oncoids. Pen is 14.5cm long. (E) Coated pentamerid valve in microbial carbonate unit"B", Fig.33. Penknife 8cm long. (F) Photomicrograph of clotted micrite of unit "B", Fig. 33. Scale bar 0.5mm long.

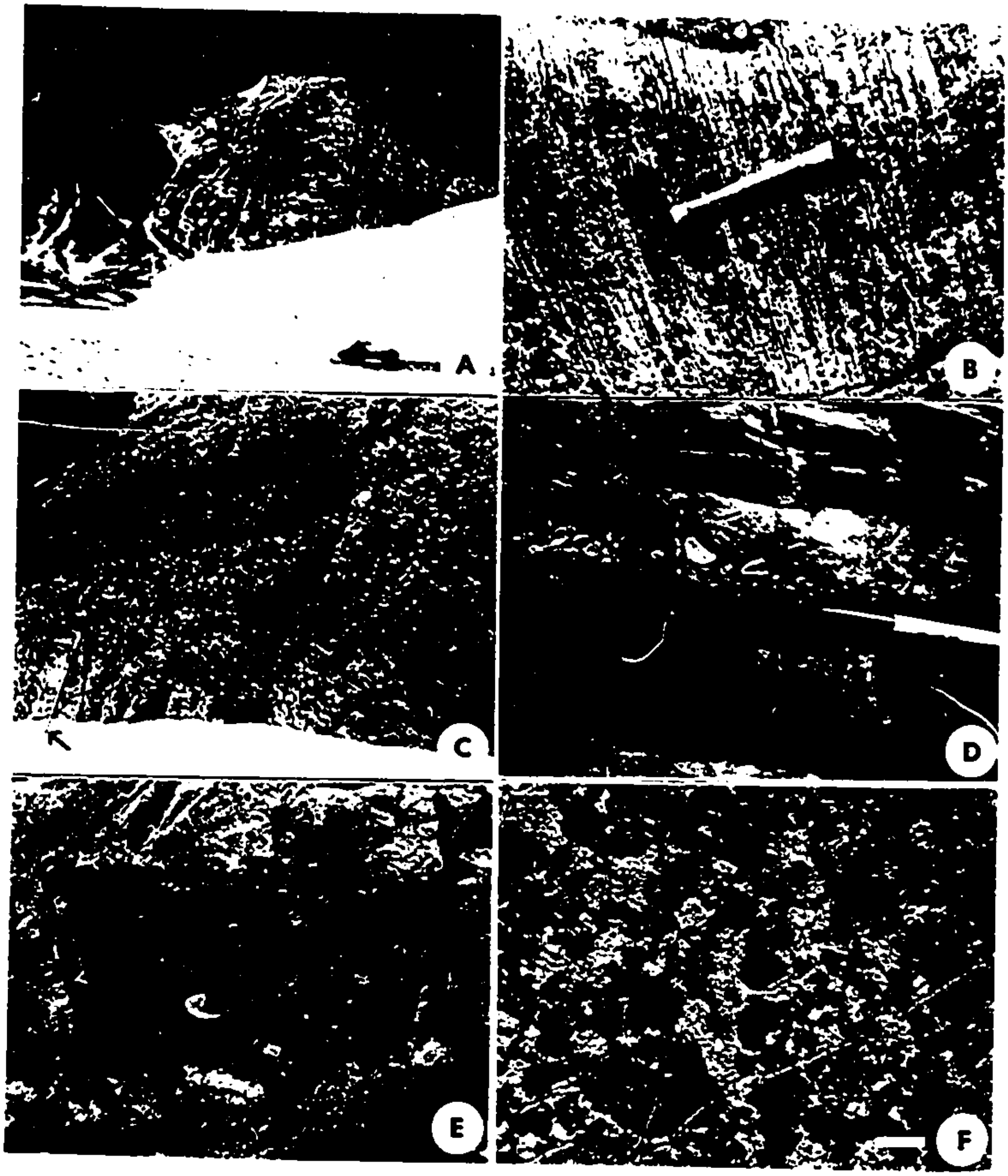
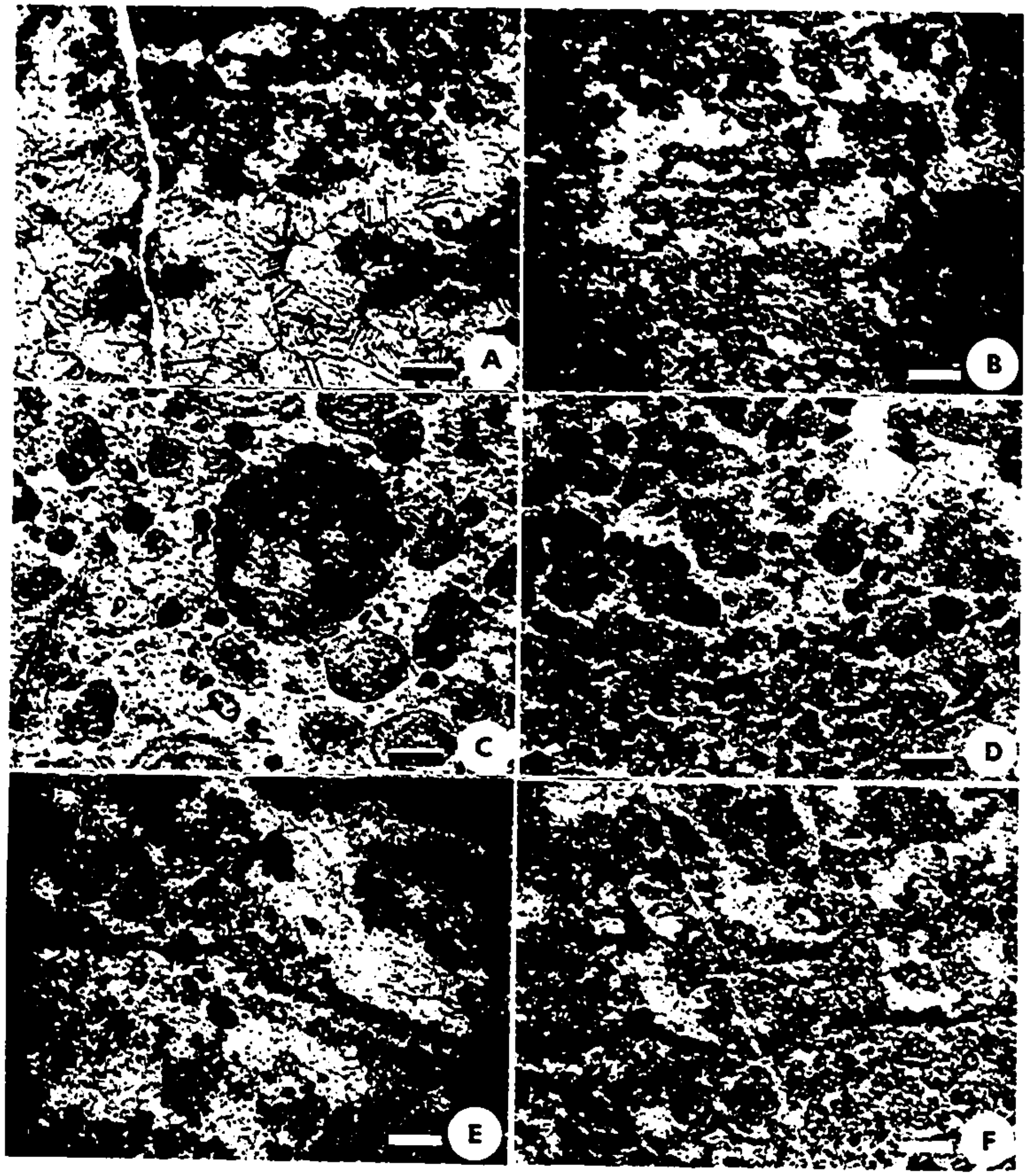


Plate 14. Photomicrographs of unit "B" of microbial build up, Cañon Fiord, Fig.33. (**A**) Clotted micrite lining fenestrae infilled with blocky calcite (probably representing burial cements). Scale bar is 0.5mm long. (**B**) Fenestral clotted micrite. Plate 14a,b,e, and f are dominant textures recognized in unit"b" of mud buildup. These clotted textures are intergradational with a multitude of other micritic textures, but overall, commonly form horizontal laminae 1-10mm thick or form a massive, non-laminated carbonate. Scale bar is 1mm long. (**C**) Poorly sorted coated grainstone. Scale bar is 1mm long. (**D**) Clotted micrite which may be intergradational with other micrites illustrated in this figure. Note even spar which separates individual clots. Scale bar is 0.5mm long. (**E,F**) Laminar fenestral, clotted micrite, one of most common lithologies in mud buildups. Scaie bar is 1mm long. Note thin geopetal bases to fenestrae in (f). Scale bars 1mm long.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PART IV

CONCLUSIONS, REFERENCES, AND APPENDICES

CONCLUSIONS

Much of the Upper Ordovician to Lower Devonian platform-to-basin succession on northern Devon Island and southern, central, and northeastern Ellesmere Island was examined in the field seasons of 1987, 1988, and 1989. Regional facies relationships have been established and the main lithostratigraphic units correlated with those known in southern Arctic. Several distinct and mappable rock units are recognized and described for the first time.

The platform sequence can be divided into two main parts, representing two upbuilding episodes. The first part includes the Cape Storm and Allen Bay formations and the second the Douro, and Goose Fiord-Barlow Inlet formations. The younger part shows laterally variable facies that were influenced by Caledonian orogenesis and uplift of parts of the carbonate shelf. In particular, platform backstepping during this phase was diachronous and locally variable in contrast to more synchronous and regionally extensive older episodes of backstepping.

During Ashgill time, a thick mottled limestone succession, represented by the Thumb Mountain Formation, was deposited on a carbonate ramp within the photic zone but below wave base. The Irene Bay Formation rests stratigraphically above and is one of the most laterally extensive rock units in the Canadian Arctic. It is likely isochronous throughout the study area, based on current biostratigraphy, and represents deposition in deeper water than interpreted for the underlying

Thumb Mountain Formation. It is less fossiliferous, more shale rich, thins basinward, and overlies a rubbly argillaceous limestone upper member of the Thumb Mountain Formation at shelf-margin localities. These stratigraphic relationships suggest that subsidence in the shelf margin was greater, and that the rubbly weathering argillaceous limestone is a slightly deeper water facies of the mottled, sparsely fossiliferous dolomitic limestone. The Irene Bay Formation could therefore represent a eustatic sea-level rise, but the evidence for this is equivocal, as a major change in climate could equally explain the observed facies, phosphatized beds, and fossil make-up of the formation.

After deposition of the Irene Bay Formation, purely carbonate sedimentation returned over most of the study area, including areas over the Thumb Mountain Formation shelf margin, but sedimentation was in slightly deeper water than during deposition of the Thumb Mountain Formation. Limestone of this phase, assignable to the basal tongue of the Allen Bay Formation, is thin, regionally extensive, unfossiliferous, and likely partly correlative with the mottled dolomitic limestone platform succession (lower limestone member of the Allen Bay Formation) exposed in eastern localities. The carbonate platform was subsequently drowned during *fastigatus* Zone platform backstepping. The regionally synchronous and laterally extensive nature of this onlapping sequence suggests a eustatic influence on platform backstepping, although this event cannot be correlated outside the basin with certainty. Platform drowning was gradual in some areas and is associated with a condensed sequence in northeastern Ellesmere Island. In the latter area, *cyphus* Zone beds abruptly overlie Ordovician carbonates, a similar stratigraphic relationship to that of North Greenland, but without an associated condensed sequence. The lack of sequence condensation, however, may be an artifact of inadequate sampling, and the age of the North Greenland platform carbonates is not well known.

Numerous buildups were established subsequent to the Lower Silurian backstepping in North Greenland. Many of these richly fossiliferous reef structures have mud cores (Sønderholm and Harland, 1989) that are overlain by stromatoporoid-rich limestone. Although incompletely exposed, the Hans Island structure is considered to be equivalent to one of the isolated North Greenland (pinnacle reef) carbonate bodies, and several more of these structures likely occur in Kennedy Channel, as inferred from the presence of olistostromes in exposures along the eastern coast of Judge Daly Promontory. The Hans Island pinnacle reef was likely much larger, flat topped, and had laterally variable facies, perhaps including a lagoonal mudstone-rich facies, and a laminar and globular stromatoporoid bindstone barrier facies. Also, it is isochronous with the Baumann Fiord pinnacle reefs; however, its lithology is very different, and it was likely associated with a distinctly different depositional environment, as concluded below.

After the major platform backstepping in the *fastigatus* Zone, a carbonate ramp sequence was re-established 10's of kilometres east and south of its Ordovician position. Platform sinuosity in the area of Grinnell Peninsula may have been due to underlying subsidence control related to early movement of the Boothia Uplift. Subsequent to backstepping, large mud buildups were established in three locations in central and southern Ellesmere Island. These were extremely large structures that had significant topographic relief yet lacked main frame builders. One of these structures, examined on a reconnaissance scale, shows a vertical sequence of bioturbated lime mudstone, microbial carbonate, and mudstone-rich stromatoporoid floatstone or bindstone. The latter is overlain by a microbial carbonate succession, signifying a late Llandovery areal restriction of carbonate deposition.

Pentamerids are common throughout the mud buildup succession, particularly in the lower part, and are associated with thin-shelled megalodont bivalves in younger strata. Their association with shelly tempestites, rare intraclasts, and coated grains suggests that the microbial carbonate was deposited below fair-weather wave base but above storm wave base, and that shallowing to near fair-weather wave base is marked by an upward replacement of microbial carbonate by the mudstone-rich stromatoporoid unit. Overlying microbial carbonate and concomitant onlapping in the Upper Llandovery was related to deepening, perhaps due to lithospheric flexure and flysch deposition in the contiguous Hazen Trough, and to foredeep expansion and Caledonian orogenesis in North Greenland. This event, however, has no platform correlative.

Abrupt termination of the Caledonian Bay mud buildup in the lower Wenlock was likely synchronous with drowning of the two other mud buildups, based on an examination of the associated slope facies. Because these structures accumulated to considerable thicknesses (up to 1335m), they were prominent topographically, and are associated with overlying condensed sequences and contrastingly expanded sequences in the contiguous Hazen Trough or deep-shelf facies. The strongly diachronous nature of the base of the Danish River Formation in the vicinity of Caledonian Bay, as worked out by Trettin (1978), is also attributed to mud buildup paleorelief, which may have been as much as 2km.

However, on Judge Daly Promontory, flysch onlapping was earlier, occurring in the *sakmaricus* Zone, and likely related to the absence of shelf margin buildups or proximity to the evolving Caledonian foreland basin.

The Ordovician-Silurian boundary was identified in the deep-shelf shales and the platform carbonates. In the former, shallowing in the boundary sequence is indicated by carbonate beds occurring in an otherwise shaly succession. These carbonate beds, probably Upper Ordovician in age, directly overlie *fastigatus* Zone shales, indicating possible correlation to the globally synchronous Hirnantian event discussed by Brenchley (1988). A condensed sequence, possibly related to minor basin expansion in the earliest Silurian, overlies the limestone beds. Lowest Silurian basin expansion in the vicinity of Bay Fiord is indicated by onlapping and a subsequent platform progradational sequence that likely coincided with the regional development of black, graptolitic shale, early Silurian climatic amelioration, and sluggish oceanic circulation (Leggett, 1980); however, the interpretation of past climate based on facies patterns is uncertain.

The relationship of carbonate platform sedimentation to glacioeustatic events near the systems boundary is unclear. In the Canadian Arctic, the Ordovician-Silurian boundary has traditionally been associated with the abrupt change from bioturbated lime mudstone below to dolostone above at the boundary of the lower and middle members of the Allen Bay Formation, but reconnaissance biostratigraphy indicates that this relationship is not consistent. Four platform sections of this study are age constrained, and it appears that shallowing could have been latest Ordovician or earliest Silurian in age, and not necessarily coincident with the systems boundary. Also, localized Lower Silurian shallowing on the platform may have been autogenically influenced, confusing a solely lithostratigraphic correlation of the systems boundary. In the present study, the boundary between the lower and middle members is typically abrupt, though not demonstrably equivalent to shallowing during the latest Ordovician Hirnantian event. The timing of glacioeustasy in the platform and deep-shelf sequences is complicated by the lack of conodonts and the condensed sequence of graptolites near the systems boundary.

Little paleorelief and no shelf-margin facies differentiation is recorded in the ramp sequence following *fastigatus* Zone platform backstepping, and it appears that inferred platform paleoenvironment was similar to that of the underlying Thumb Mountain Formation. In the vicinity of Baumann Fiord, carbonate deposition was interupted by middle Llandovery, *cyphus* Zone onlapping and platform backstepping. Backstepping was coeval with the deposition of regionally extensive, black chert beds, suggesting a paleoenvironmental and/or eustatic control on this phase of platform evolution.

Subsequent to *cyphus* Zone platform backstepping, a mudmound and several isolated carbonate bodies (pinnacle reefs) were established basinward of the platform margin. The mudmounds and pinnacle reefs show contrasting upbuilding successions, from bioturbated lime mudstone to coralgal biolithite in the pinnacle reefs and bioturbated lime mudstone to microbial carbonate in the mudmound. The relationships suggest different upbuilding rates for the two lithologically different structures. Also, the coralgal biolithite is newly interpreted to represent a high-energy, wave-stressed environment that excluded the growth of stromatoporoids but favoured the growth of a sparse metazoan fauna thickly encrusted by microbial colonies, forming a coralgal biolithite. In two or possibly three of the pinnacle reef structures, coralgal biolithite is succeeded by and interbedded with oolites, suggesting agitated condition. The apparent lack of oolitic deposits over one of the coralgal biolithite pinnacle reefs suggests a slightly different setting, perhaps related to size and inherent morphology. The smaller pinnacle structures had less surface area thus perhaps could not contain and/or promote extensive ooid formation. At least one of the oolite-rich reef pinnacles shows markedly progradational geometry suggesting that surface area did increase during upbuilding. It appears that the mudmound and pinnacle reefs were abruptly drowned in the middle Ludlow *linearis* Zone. In the vicinity of Baumann Fiord, a thin tongue of the Starfish Bay shale separates a predominantly carbonate succession from a predominantly shaly succession.

The Devon Island Formation was named on Devon Island for a "...uniform series of medium dark grey to greyish black calcareous shale and shale with very minor interbeds of medium dark grey argillaceous limestone..." by Thorsteinsson, (*In:* Fortier *et al.*, 1963, p.228-229). In the Vendom Fiord area, a similar relationship of shales on carbonates is recognized; however, the basal shale units are condensed over paleotopographic highs while expanded over lows. In the latter setting, a limestone succession underlies a siltstone succession, and as such is quite unlike the stratotype. Present GSC mapping, chiefly based on the onlapping relationship of shales on limestone, recognizes the Devon Island Formation in the Vendom Fiord area, and has extended this formation to the Trold Fiord map area. However, in the latter area, platform carbonates are lacking and the base of the so-called Devon Island Formation has been proposed to coincide with the base of the Starfish Bay shale, a distinctive rock unit not present in the type area of the Devon Island Formation and discontinuously present in the Vendom Fiord area. If it were not for the presence of the Starfish Bay shale in deep shelf localities, the so-called Devon Island Formation would not be a mappabie lithostratigraphic unit. Although provisionally included in the Cape Phillips Formation in this study, minor revision of the type lithology through establishment of a reference section (North American Stratigraphic Code, 1985, arts. 8e, 19a, 22c) would be useful for geological mapping and for discussion of the geological history of the Franklinian Basin. As such, the Devon Island Formation would be one of the most widely recognized deep-shelf lithostratigraphic units in the Canadian Arctic.

The base of the Devon Island Formation on Devon Island and the so-called Devon Island Formation near Baumann and Trold fiords is significant in that it represents a major reorganization of carbonate deposition in the study area. Its base, which represents platform drowning, was markedly diachronous, falling within the *linearis* Zone in the vicinity of Baumann Fiord, but approximately on the *siluricus-latialata* conodont biozone boundary in northern Devon Island, suggesting flexure perhaps related to a renewed phase of Caledonian tectonism; however, direct evidence for the latter interpretation is lacking.

In the vicinity of Baumann Fiord, condensed sequences are associated with middle Ludlow platform backstepping over reef pinnacles and platform carbonates, and expanded sections occur over lows, suggesting that paleotopography had a considerable influence on deposition for some time following drowning. The Starfish Bay shale varies similarly, as it is present adjacent to, but not overlying, shallow-water platform carbonates.

Based on a reconnaissance study of platform carbonates in Darling Peninsula, the stratigraphy known in the southern Arctic Islands can be applied to northeastern Ellesmere Island. There are variations within formations, however, and the Allen Bay Formation contains several units that are not recognized in the south: a thick predominantly subtidal sequence in the upper part of the Allen Bay Formation, for example, contrasts with the predominantly peritidal sequence exposed in the south. Its presence suggests that subsidence in the northern part of the study area was greater and likely related to flexure and the extensive deposition of syntectonic flysch in North Greenland and central Ellesmere Island. The Allen Bay Formation is overlain by the Cape Storm, Douro, and Goose Fiord formations, which are similar lithologically to the isochronous formations exposed in the southern Arctic; however, these, too, show some minor lithological variations. For example, between Grinnell Peninsula and Somerset Island, the Cape Storm Formation contains abundant, possibly wind-transported silt, but near Baumann Fiord and further to the north, silt is uncommon. Although present evidence is equivocal, the silt distribution is ascribed to variation in wind transport of cratonic detritus.

The Douro Formation is generally a fossiliferous, rubbly weathering limestone likely deposited below fair-weather wave base; however, it shows some distinct lithological variations in the study area. On Ellesmere Island, from Darling Peninsula to Baumann Fiord, the upper part of this formation is represented by a relatively clean mottled dolomitic limestone, a facies likely inidicative of deposition in slightly shallower water than that interpreted for the argillaceous limestone (Mortensen and Jones, 1986). This facies could indicate shoaling related to early movement of the Inglefield Uplift, and, as such, is very similar to the isochronous limestone succession described from the southern Arctic Islands (*op cit.,*).

Near Irene Bay, two poorly exposed, thin, laterally extensive crinoidal packstone and grainstone mounds were deposited on a platform slope. They are interpreted as sediment zoned drift accumulations, with basal crinoid packstone and grainstone, middle stromatactoid-rich stromatoporoid grainstone and packstone, and upper stromatoporoid-rich floatstone. This succession is abruptly overlain by cephalopod- and ostracod-rich red-stained lime mudstone. Early marine cementation of the sediment was ubiquitous and likely facilitated the development of neptunian dykes and sills. Crinoid-rich mound sediment of an autochthonous and/or allochthonous origin was trapped and bound by microbial colonies and baffled by crinoid growths, and during latter stages of mound growth, was stabilized by laminar stromatoporoids. Microbial carbonate is present in the sediment drift accumulations, but is distinctly less common than in mud buildups and mudmounds. Stromatactoid structures are common, but are not present in the mud buildups or mud mounds, a relationship which suggests an organic origin for these enigmatic structures.

The Allen Bay-Cape Storm sequence represents upward shallowing overall, and its upper boundary is a significant disconformity. On east-central Cornwallis Island, upbuilding is represented by an increase in thickness and frequency of olistostromes, and by an upward development of the shelf margin from an *island-bay complex* to a *patch reef shelf-margin complex* to a *skeletal grainstone-shoal* *complex.* These facies were established subsequent to Lower Silurian platform backstepping, and they represent progressive upbuilding, increase in platform margin relief, and decrease in accommodation through time, shown by the gradual upward replacement of subtidally deposited lime mudstone by tidalites. During earlier stages of upbuilding, the shelf was relatively open, and lime mud-rich stromatoporoid reefs developed perhaps below wave base.

The Cape Storm Formation is an unusual platform facies of the Canadian Arctic in that it represents the most extensive development of Silurian peritidal carbonates but was deposited during minimal platform accommodation. During this time, pervasive inimical platform waters are presumed to have facilitated a "default style" of sedimentation, in which an extensive grainstone shelf-margin complex was deposited instead of reefs, in contrast to the underlying and overlying platform-margin successions. Megalodont bivalves are closely associated with the skeletal grainstone beds, and are interpreted to have inhabited protected shelf-margin environments between migrating carbonate sand shoals. These bivalves reached maximum abundance in the Ludlow in the Canadian Arctic, although they have been recognized in strata as old as uppermost Llandovery, in the mud buildup sequence. However, the older bivalves have thinner shells than those of Ludlow and Pridoli (Packard, 1985) shelf-margin settings. A gradual return to more open platform waters following deposition of the Cape Storm Formation is indicated by the overlying Douro Formation, which represents deposition on an open marine carbonate ramp.

A major platform disconformity, occurring at the base of the Cape Storm Formation, represents erosion of much of the Wenlock and some of the Ludlow platform carbonates. On Swinnerton Peninsula, the entire upper limestone member of the Allen Bay formation is missing, but only 15km to the south, at the head of Makinson Inlet, about 135m of the member is exposed.

A second major phase of carbonate deposition and platform upbuilding is represented by the Barlow Inlet-Goose Fiord Formation. Platform progradation during this time was extensive in the south, in the vicinity of northern Devon Island and southern Devon Island, but no evidence of progradation is observed in the area of Baumann Fiord, possibly suggesting an underlying, fault control on subsidence and the position of the shelf margin. This thick upward shallowing carbonate succession shows an overall decrease in platform accommodation through time. Pronounced high-frequency cyclicity preserved throughout contrasts with the lack of cyclicity in lowest Silurian and uppermost Ordovician platform carbonates, even though the latter represents a time of supposed great sea-level fluctuations due to changes in continental ice volume.

In the deep-shelf succession, dark grey silty clayshale, claystone, and siltstone are abruptly overlain by grey-weathering siltstone and interbedded limestone in the Upper Silurian to Lower Devonian. The base of this unit (referred to the grey siltstone member of the Red Canyon River Formation) is diachronous, youngs basinward, and is not present south of Bay Fiord. It is likely a correlative to the progradational clastic wedge of the Red Canyon River Formation. This sequence is over a kilometre thick south of Cañon Fiord and represents the first and most significant phase of Caledonian orogenesis in the area. Two subsequent phases are recognized, one in the *dehiscens* Zone and the last in the *inversus* Zone. These phases, however, were not examined in detail for this study.

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APPENDIX ONE

LOCATIONS OF STRATIGRAPHIC SECTIONS AND MAIN UNITS MEASURED

Considerations

Section locations are given in Figure 1. Gaps in the numbering scheme are sections which were not pertinent to this study. Many sections are presented diagrammatically in the text; other section that are not presented are on file at the University of Ottawa, the Department of Geology. The following abbreviations are used in the following list and throughout the text:

Ddr -- Danish River Formation
Oci -- Irene Bay Formation
ODcp -- Cape Phillips Formation
$OSa_{l.mbr}$ -- Allen Bay Formation, lower member
$OSa_{m.mbr}$ -- Allen Bay Formation, middle member
$OSa_{u.mbr}$ -- Allen Bay Formation, upper member
sbs -- Starfish Bay shale
Scs -- Cape Storm Formation
"SDdi" -- Devon Island Formation: provisional assignment
SDg -- Goose Fiord Formation
Sdo -- Douro Formation
Drc -- Red Canyon River Formation

Section number	*Location (lat. & long.)*	*Main stratigraphic units*
6-4	Eidsbotn Fiord, Devon Island. 76°12'N, 91°18'W	**OSa**$_{u.mbr}$: 150m (no base, disconformable top) Scs: 120m (no top)
6-7	Cape Vera, Devon Island. 76°12'N, 89°10'W	**OSa**$_{m.mbr}$: 30m (no base) **OSa**$_{u.mbr}$: 140m (disconformable top; gradational base) Scs: 30m (no top)
6-9	near Viks Fiord, Devon I. 76°08'N, 90°50'W	**OSa**$_{l.mbr}$: 108m (gradational top; complete section) **OSa**$_{m.mbr}$: 632m (no top)
6-10	West Fiord, Devon Island. 75°52'N, 90°25'W	**Osa**$_{u.mbr}$: 143m (no base; disconformable top) Scs: 587m (gradational top)
02	Irene Bay, Ellesmere I. 78°59'N, 81°41'W	**OSa**$_{l.mbr}$: 56m (complete) **ODcp**: 95m (no top, gradational lower contact)
03	Strathcona Fiord, Ellesmere I. 78°47'N, 82°45'W	**OSa**$_{l.mbr}$: 100m (poorly exposed; complete) **ODcp**: 220m (no top, gradational base)

Section number	*Location (lat. & long.)*	*Main stratigraphic units*
05	Irene Bay, Ellesmere I. 78°58'N, 82°15'W	**OSa**$_{Lmbr?}$: 218m (Lower Silurian shale tongue at top) **OSa**("shelf margin" facies): 510m (lower Wenlock top) **ODcp**: 140m (disconformable top in *transgrediens* Zone) **SDg?**: 40m (top disconformable with Vendom Fd Fm)
08	Irene Bay, Ellesmere I. 78°57'N, 82°14'W	**ODcp**: 64m (latest Wenlock base on platform carbonates; disconformable top with Sdg) **SDg?**: 37m (disconformable top with Vendom Fiord Fm)
09	Irene Bay, Ellesmere I. 78°58'N, 82°14'W	**ODcp**: 118m (conformable base on platform carbonates, disconformable top with SDg) **SDg?**: 72m (disconformable top with the Vendom Fd Fm)
10	Irene Bay, Ellesmere I. 78°58'N, 82°12'W	**OSa**$_{Lmbr?}$: 241m (Lower Silurian shale tongue at top) **OSa** (shelf margin facies): 402m (top=base of section 9)
11	Irene Bay, Ellesmere I.. 78°58'N, 82°18'W	**ODcp**:42m (no base; disconformable top with SDg) **SDg?**: 61m (disconformable top with the Vendom Fd Fm)
12	Irene Bay, Ellesmere I. 79°01'N, 82°08'W	**OSa**$_{Lmbr?}$: 161m (*acuminatus* Zone shale tongue at top) **OSa** (shelf margin facies): 504m (abrupt upper contact) **ODcp**: 108m (disconformable top with the SDg) **SDg?**: 10m (disconformable top with the Vendom Fd Fm)

Section number	*Location (lat. & long.)*	*Main stratigraphic units*
13	Irene Bay, Ellesmere I. 78°57'N, 82°19'	**$OSa_{l.mbr?}$**: 71m (grad. top) **ODcp**: 179m (top contact covered) **SDrc?**: 170m (no top; *uniformis* Zone 154m above base)
15	Eidsbotn Fiord, Devon Island. 76°12'N, 91°18'W	**$OSa_{m.mbr}$**: ~70m (upper contact very poorly exposed) **$OSa_{u.mbr}$**: ~320m (disconformable, poorly exposed upper contact with the Scs)
16	Boat Point, Devon I. 76°01'N, 90°00'W	**$OSa_{m.mbr}$**: 130m (no base; gradational upper contact) **$OSa_{u.mbr}$**: 380m (no top)
17	Sandhook Bay, northeastern Devon Island. 75°52'N, 89°25'W	**$OSa_{l.mbr}$**: 170m (upper contact marked by abrupt transition from mudstone to x-stratified grainstone) **$OSa_{m.mbr}$**: 140m (no top)
18	West Fd, northern Devon Island. 76°03'N, 90°15'W	**$OSa_{m.mbr}$**: 80m (no base; upper contact gradational) **$OSa_{u.mbr}$**: 370m (no top)
19	Cape Sir John Franklin, Grinnell Peninsula, Devon Island. 76°41'N, 96°50'W	**$OSa_{l.mbr}$**: <10m **ODcp**: 1420m (poorly exposed; upper contaci with the Bathurst Island Fm abrupt)
20	central Grinnell Peninsula. 76°41'N, 96°50'W	**OSa**: (undivided limestone shelf margin facies) 928m
21	Barrow Harbour, Grinnell Peninsula. 76°37'N, 96°18'W	**OSa**: (undivided "shelf margin" limestone 715m

Section number	*Location (lat. & long.)*	*Main stratigraphic units*
22	Cape Baird, Ellesmere I. 81°33'N, 64°32'W	**Oci**: 7.5?m **OSa**: 30m (covered top) **ODcp**: 54m (no top, predominantly snow covered; Ddr rubble noted at top)
23	Cape Lieber, Ellesmere I. 81°29'N, 64°26'W	**ODcp**: 93m (contact with OSa sheared, probably some offset; gradational contact with Ddr).
24	Cape Cracroft, Ellesmere I. 81°24'N, 64°43'W	**ODcp**: 8m (very poorly exposed, snow covered; small outcrop of bedded chert-limestone)
25	Cape Defosse, Ellesmere I. 81°16'N, 65°25'W	**ODcp**: (faulted base) 280m, poorly exposed Ddr contact.
26	Church Peak, Ellesmere I. 81°15'N, 65°40'W	**OSa**: (no Oci) 46m **ODcp**: 144m (gradational contact with overlying Ddr)
27	Hans Island, Kennedy Channel. 80°49'N, 66°26'W	Unnamed unit: 125m (NW side of island, in April, 1988, exposures were predominantly snow covered
28	Hans Island, Kennedy Channel. 80°49'N, 66°26'W	Unnamed unit: 143m
29	Carl Ritter Bay, Ellesmere I. 80°58'N, 67°20'W	**OSa**$_{Lmbr}$: 155m (Oci base) **ODcp**: 26m (gradational contact with Ddr).
30	John Richardson Bay. 80°58'N, 67°20'W	**OSa**$_{Lmbr}$: 96m (Oci base) **ODcp**: 219m (gradational contact with Ddr; 134m ODcp-OSa transitional beds)

Section number	*Location (lat. & long.)*	*Main stratigraphic units*
31	Darling Peninsula, Scoresby Bay, Ellesmere I. 79°50'N, 71°00'W	**Oci**: 81m (well exposed) **OSa**$_{L.mbr}$ (unit 1): 120m **OSa**(unit 2): 235m **OSa**(unit 3): 145m **OSa**(unit 4): 220m **OSa**(unit 5): 270m **Scs**(unit 6): 103m **Sdo**(unit 7b) 160m (placement of upper contact uncertain; two distinct units of the Sdo recognized). **SDg**(unit 7a): 200m (no top)
32	Dobbin Bay, Ellesmere I. 79°52'N, 73°50'W	**OSa**$_{L.mbr}$: 130m (Oci at base) **ODcp**: 102m gradational contact with Ddr)
33	Dobbin Bay, Ellesmere I. 79°46'N, 74°00'W	**OSa**$_{L.mbr}$ (unit 1): 190m **OSa**(unit 2): 280m **OSa** (unit 3): 60m (no top) section incomplete
34	head of Cañon Fd, south shore, Ellesmere I. 79°34'N, 80°05'W	**OSa**$_{L.mbr}$: 190m (no top)
35	head of Cañon Fd, north shore glacier, Ellesmere I. 79°42'N, 80°22'W	**OSa**$_{L.mbr}$: 65m **ODcp**: 288m (63m of siltstone at top of ODcp that correlates with unit A in section 36)
36	Cañon Fd, Ellesmere I. 79°45'N, 80°46'W	**OSa**$_{L.mbr}$: <10m (incomplete) **ODcp**: 325m (140m siltstone unit "A" recognized in section 35; abrupt contact with Ddr)
37	Cañon Fd, Ellesmere I. 79°45'N, 81°30'W	**OSa**: 1335m (undivided mud buildup)

Section number	*Location (lat. & long.)*	*Main stratigraphic units*
38	Cañon Fd., Ellesmere I. 79°47'N, 81°18'W	**OSa**: 460m (mudbuildup facies; no base; sharp contact with ODcp) **ODcp**: 142m (sharp upper contact with Ddr)
39	Cañon Fd., Ellesmere I. 79°48'N, 81°28'W	**OSa**: 979m (undivided mud buildup facies) **ODcp**: ~114m (abrupt contact with the Ddr)
40	Cañon Fd., Ellesmere I. 79°57'N, 81°05'W	**OSa**: 65m (no base; undivided mud buildup facies) **ODcp**: 80m (abrupt contact with Ddr)
41	Cañon Fd., Ellesmere I. 79°58'N, 80°50'W	**OSa**: 1141m (undivided mud buildup facies)
42	Mount James, Ellesmere I. 79°45'N, 81°30'W	**OSa**: 1310m (sharp upper contact with Ddr; undivided mud buildup facies)
43	Vesle Fd., Ellesmere I. 79°04'N, 83°03'W	**OSa**: 38m **ODcp**: 632m (includes ~60m Starfish Bay shale; abrupt upper contact with grey siltstone mbr. of Drc)
46	Gunnars Island, west coast, Baumann Fd. 77°28'N, 85°26'W	**OSa**: (pinnacle facies) 220m (unknown base)
47	southeastern Hoved Island, Baumann Fd. 77°30'N, 84°50'W	**OSa**: 180m **ODcp**: 130m (abrupt contact with "SDdi")

Section number	*Location (lat. & long.)*	*Main stratigraphic units*
48	northwestern Hoved I., Baumann Fd. 77°25'N, 85°30'W	**OSa**: 70m (no base) **ODcp**: 300m (section incomplete; see section 76)
49	unnamed peninsula near Gunnars I., Baumann Fd. 77°26'N, 85°35'W	**OSa**: 140m (pinnacle reef facies: abrupt contact with "SDdi")
50,54	Corner of Vendom and Baumann Fd, top of hill. 77°31'N, 84°06'W	**OSa**: 275m **ODcp**: (slope facies) 960m+ (more section present but poorly exposed)
51	Corner of Vendom and Baumann Fd, base of hill. 77°31'N, 84°06'W	**ODcp**: 840m (poorly exposed proximal slope facies).
52	Corner of Vendom and Baumann Fd, middle of hill. 77°31'N, 84°06'W	**ODcp**: 320m (slope facies)
56	Trold Fiord. 78°09'N, 85°18'W	**OSa**: <5m (Oci present) **ODcp**: 118m (no top)
57	Trold Fiord. 78°09'N, 85°17'W	**OSa**: <5m (Cci present) **ODcp**: 221m (gradational lower contact; no top)
58	Trold Fiord. 78°09'N, 85°19'W	**OSa**: <5m (Oci present) **ODcp**: 125m (gradational lower contact; no top)
59	Trold Fiord. 78°21'N, 84°58'W	**ODcp**: 165m (faulted base) **"SDdi"(sbs)**: 69m (abrupt lower contact; gradational top) **"SDdi"(upper slst. mbr.)**: 296m (gradational lower, abrupt upper contact) **Ddr**: 502m (no top)

Section number	*Location (lat. & long.; all sections on Elesmere I.)*	*Main stratigraphic units*
60	Northeast corner of Starfish Bay and Trold Fiord. 78°21'N, 86°31'W	**OSa**: <5m **ODcp**: 122m **"SDdi"(sbs)**: 38m **"SDdi" (upper slst. mbr.)**: 160m **Ddr**: 412m (no top)
61	Swinnerton Peninsula. 77°24'N, 81°35'W	$\mathbf{OSa}_{m.mbr}$: 103m (no base; disconformable? top) **Scs**: 258m **Sdo**: 121m (grdational lower contact) **SDg**: 241m (no top; gradational lower contact)
62	head of Makinson Inlet, west side. 77°41'N, 81°57'W	$\mathbf{OSa}_{l.mbr}$: 95m (gradational lower contact) $\mathbf{OSa}_{m.mbr}$: 440m (gradational upper and abrupt lower contact) $\mathbf{OSa}_{u.mbr}$: 135m **Scs**: 220m (disconformable lower and gradational upper contact) **Sdo**: 157m (gradational upper and lower contact) **SDg**: 241m (no top)
63	head of Makinson Inlet, west side. 77°43'N, 82°06'W	**SDg**: 1147 (no base, faulted top)
64	head of Makinson Inlet, west side. 77°35'N, 81°52'W	$\mathbf{OSa}_{l.mbr}$: 30m (gradational lower and upper contact) $\mathbf{OSa}_{m.mbr}$: 485m (gradational upper contact) $\mathbf{OSa}_{u.mbr}$: 85m (disconformable? lower contact)

Section number	*Location (lat. & long. all sections on Ellesmere I.)*	*Main stratigraphic units*
65	head of Makinson Inlet, west side. 77°35'N, 81°51'W	**Sdo**: 14m (no base) **SDg**: 49m (disconformable top)
67	Vendom River. 77°47'N, 82°24'W	**OSa**$_{u.\&m.mbr}$: 223m **SDg**: 1688m
69	Vendom Fiord. 77°32'N, 83°55'W	**ODcp**: 166m (incomplete) **"SDdi"**: 682m (abrupt upper and lower contact; v. thin, poorly esposed sbs)
70a	Vendom Fiord. 77°46'N, 83°36'W	undivided shelf-margin carbonates: 375m (no base) **Sdo?**: 90m (poorly exposed lower contact and abrupt, poorly exposed upper contact)
70b	Vendom Fiord. 77°45'N, 83°36'W	undivided shelf-margin carbonates: 50m (no base) **Sdo?**: 165m (poorly exposed lower contact and abrupt, poorly exposed upper contact)
70c	Vendom Fiord. 77°44'00"N, 83°36'W	undivided shelf-margin carbonates: 50m (no base) **Sdo?**: 50m (poorly exposed lower contact and abrupt, poorly exposed upper contact)
71	Vendom Fiord. 77°44'30"N, 83°36'W	**Sdo?**: ~100m (poorly exposed lower and upper contact)

Section number	*Location (lat. & long. all sections on Ellesmere I.)*	*Main stratigraphic units*
72	Hoved Island. 77°32'N, 85°27'W	**OSa**(pinnacle reef facies): 400m (base corresponding to top of **OSa** of section 48, poorly exposed; barometer measured)
76	Hoved Island. 77°25'N, 85°30'W	**OSa** (pinnacle reef foreslope facies) 440m (no top)
79	northwestern Hoved Island. 77°36'N, 83°45'W	**ODcp**: 320m (complete, thin sbs encountered at top; sharp upper contact)
80	Vendom Fiord. 77°34'N, 83°50'W	**ODcp**: 450m (complete; thin sbs abruptly overlying ODcp)

For Oci thicknesses, see Table 1

APPENDIX TWO

FOSSIL IDENTIFICATION

Methodology

The graptolite were identified by the author chiefly using detailed descriptions and plates in publications by Melchin (1987a,b, 1989) Jackson *et al.* (1978); Lenz (1974, 1977, 1978, 1979, 1980, 1982); Bou ek (1933, 1960); Jackson and Ethrington (1969); Lenz and Melchin (1990); Teller (1969); and Jaeger (1959). Specimens were submerged under water and examined with the binocular microscope. Identification was generally limited to only age-diagnostic forms.

Extensive taxonomic discussion is not presented in this report, but a detailed account of graptolite taxonomy is presently in preparation by M.J.Melchin for the Llandovery and by A. Lenz for the Wenlock, and younger strata. Study samples particularly difficult to identify, either because of their preservation or morphology, were confirmed by M. Melchin, A. Lenz, or R. Thorsteinsson. In the following list, these identifications are indicated by MJM, ACL, or RT, respectively.

Conodonts were identified by A.D. McCracken and T.T. Uyeno, brachiopods by Jisuo Jin, and trilobites, by the author. The last two groups of fossils provided supplementary biostratigraphic imformation.

GRAPTOLITE IDENTIFICATIONS

Key

02 05 section number sample number.
LUNDGRENI graptolite biozone
Monograptus priodon graptolite identification made by or confirmed by M.J.Melchin (MJM), A.C. Lenz. (ACL), or Ray Thorsteinsson (RT).

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02 05
MINOR Zone
Monograptus pseudobecki
M. decipiens n.ssp. (Melchin, 89, written com.)
M. turriculatus minor
Pristiograptus regularis regularis (MJM)
Petalograptus tenuis
P. ovatus wuxiensis (MJM)
Glyptograptus sp. (MJM)

02 06
TURRICULATUS Zone
Monograptus turriculatus turriculatus
M. planus
M. priodon
Stomatograptus sp.
Pristiograptus nudus

02 07
CRISPUS Zone
Monograptus spiralis ssp.
M. priodon
Monoclimacis vomerinus ssp.
?*Diversograptus* sp.

02 08
RIGIDUS-PERNERI Zones
Cyrtograptus rigidus
C. cf. *C. perneri*
C. lundgreni
Dictyonema sp.
Thallograptus sp.
Monograptus priodon
M. capillaceus
M. cf. *M. flemingi*
Paraplectograptus sp.
Dictyonema crassibasale?
D. gracile?

02 09
TESTIS Zone
Cyrtograptus lungreni
C. hamatus?
Monograptus priodon
Diversograptus ramosus?
Dictyonema hamiltoniae?
Retiolites sp.

02 10
PROGENITOR Zone
Bohemograptus bohemicus bohemicus
Pristiograptus dubius dubius

03 01
FASTIGATUS Zone
Orthograptus fastigatus
Pseudogygites arcticus

03 03
MINOR-TURRICULATUS Zones
Retiolites geinitzianus
Monograptus decipiens
M. turriculatus minor
M. kovalevskyi
M. halli
M. pseudobecki
Diversograptus capillaris
Pseudoplegmatograptus obesus
Pristiograptus sp.
indet. biserial graptolites

05 08
ULTIMUS-PARULTIMUS Zones
Dendrograptus striatus
"Monograptus" norfordi (ACL)
Palaeodictyon sp.
Pristiograptus spp.

05 09
latest Wenlock?, perhaps *SHERRARDAE* Zone
P. dubius (ACL)
P.deubeli (ACL)
Thallograptus sp.
Carynoides sp.

05 10
perhaps *BIRCHENSIS* or *TRANSGREDIENS* Zones
Pristiograptus dubius
Pseudomonoclimacis cf *P. richardsoni*

05 11
SHERRARDAE Zone
Pristiograptus cf. *P. ludensis* (ACL)

05 12
indet. monograptids

05 14
TRANSGREDIENS Zone
Pristiograptus transgrediens (ACL)

08 01
SHERRARDAE Zone
Pristiograptus nudus
Monograptus cf. *M. formosus* (ACL)

08 06
Late Silurian
Ecclimadictyon fastigiatum (stromatoporoid)
?Clathrodictyon sp. (stromatoporoid)
Stomatopora sp. (stromatoporoid)

08 10
GRIESTONIENSIS-SAKMARICUS Zones
Monograptus priodon (MJM)
M. spiralis spiralis (MJM)
Monoclimacis vomerinus ssp. (MJM)
Stomatograptus sp. (MJM)
This sample and 09-01 were sent to MJM for identification. Both are, however, out of stratigraphic context. An upper Llandovery sample (08-10) occurs stratigraphically above Wenlock (08-01) and Pridoli samples (09-03) identified by ACL. Moreover, this area was revisited by ACL in 1989, and he did not record strata older than Wenlock in age overlying platform carbonates, a finding which is in agreement with the present correlations.

09 01
MINOR or *TURRICULATUS* Zones
Hedrograptus elongatus (*Hedrograptus* is perhaps better refered to as *Climacograptus*, MJM, pers. com., 1988, contrary to what was listed in Melchin's 1987 Ph.D. thesis)
Pristiograptus nudus
P. regularis?
Glyptograptus elegans n.ssp. (MJM, written com.)

09 03
lowest Pridoli
Monograptus cf. *M. formosus* (ACL)
Pseudomonoclimacis tetliensis (Pridolian; ACL)

09 10
CONVOLUTUS or *MINOR* Zones
Pristiograptus nudus
Glyptograptus ex gr. *tamariscus* (MJM)

11 4
Wenlock or later
Atrypa sp.
These brachiopods are considerably compacted; hence, detailed examination was impossible (Jin pers com, 1989). Strata below and above contained numerous graptolites, and stratigraphically below this interval, a large fragment of a probably *Cyrtograptus* sp. was observed in zone 3 strata of the mound rock discussed in part III of this report.

12 05
ACUMINATUS Zone
Climacograptus normalis (MJM)
C. angustus (MJM)
C. sinitzini (MJM). For this graptolite, it is the first noted occurrence outside China (MJM, pers. com., 1989).
Glyptograptus aff. *G. nanus* (MJM)

12 06
(*ACINACES?*) *CYPHUS or* lower *CURTUS* Zones (MJM)
Glyptograptus ex. gr. *G. tamariscus* (MJM)
Diplograptus elongatus
Atavograptus sp.

13 01
CYPHUS-CURTUS/PECTINATUS Zones
Coronograptus gregarus
Petalograptus intermedius (MJM)
Atavograptus ?praestrachani
Metaclimacograptus highesdi? (MJM)
Climacograptus rectangularis
Diplograptus tcherskyi tcherskyi (MJM)
Atavograptus praestrachani and/or *A. atavus* (MJM)

13 03
GRIESTONIENSIS-SAKMARICUS Zones
Monograptus turriculatus turriculatus
M. griestoniensis
M. spiralis spiralis
M. parapriodon
Retiolites geinitzianus densereticularis
Cyrtograptus lapworthi?
C. sakmaricus
Monoclimacis vomerinus ?vomerinus
Stomatograptus canadensis
Thallograptus arborescens
Dictyonema aff. *Dictyonema delicatulum*

13 04
LINEARIS-TENUIS Zones
Saetograptus fritschi linearis
Bohemograptus bohemicus tenuis

13 06
UNIFORMIS Zone
Monograptus aequabilis aequabilis (ACL)

15 12
Silurian
Clathrodictyon lenticula (stromatoporoid)
Stylopleura sp. (coral)
Hindia sphaeroidalis (lithistid sponge)
heliolitid corals

19 20
MINOR Zone
Monograptus turriculatus minor (MJM)
M. marri
M. halli
Pristiograptus nudus (MJM)
Climacograptus cf. *janischewskyi*
Retiolites sp.

19 21
CONVOLUTUS-MINOR Zones
Monograptus turriculatus minor
M. halli
M. marri
Pristiograptus sp.
Petalograptus sp.
Climacograptus sp.

19 22
MINOR Zone
Monograptus turriculatus minor
M. decipiens n.ssp. (MJM)

19 24
TURRICULATUS- to SAKMARICUS Zones
Monograptus spiralis spiralis
M. priodon

19 25
PROGENITOR/BOHEMICUS Zones
Bohemograptus bohemicus bohemicus
?Pseudomonoclimacis sp.
Saetograptus colonus (ACL)

19 28
PROGENITOR Zone
Bohemograptus bohemicus tenuis
B. bohemicus bohemicus

19 31
CHELMIENSIS Zone (ACL)
Spinograptus sp. (ACL)
M. cf *angustidens*
?Lobograptus sp.
?Holoretiolites sp.

19 34
(approx. *TRANSGREDIENS* Zone) Pridoli-Lochkovian or (*YUKONENSIS* Zone) Pragian
M. cf. *telleri* OR
M. uniformis cf. *angustidens*

19 36
Pridoli
Monograptus pridoliensis (ID, ACL)
This section, located at Cape Sir John Franklin, Grinnell Peninsula, and was subsequently visited by ACL in 1989, and recorded preliminary but significant finds of middle and upper Silurian graptolites, in particular, a significant thickness of the *birchensis* biozone (Lenz, 1990)

19 38
Upper Silurian or Lower Devonian
Lacunoporaspis sp. (trilobite)

20 1
FASTIGATUS Zone
Glyptograptus cf. *G. lorrainensis*
Pseudogygites arcticus. This is a common Upper Ordovician trilobite and is abundant on bedding planes containing *Orthograptus fastigatus*.

20 2
FASTIGATUS Zone
Pseudogygites ?arcticus (trilobite)
Orthograptus fastigatus
O. amplexicaulis
O. thorsteinssoni
Amplexograptus aff. *A. prominens*

21 1
FASTIGATUS Zone
Pseudogygites ?arcticus (trilobite)
Orthograptus fastigatus
O. amplexicaulis
Glyptograptus cf. *G. lorrainensis*
Climacograptus sp.
C. latus

23 4
GRIESTONIENSIS-SAKMARICUS Zones
Monograptus spiralis spiralis
M. griestoniensis ssp.
M. griestoniensis minuta
M. priodon
M. ?speciosus
Pristiograptus dubius
Monoclimacis linnarssoni
M. vomerinus vomerinus
M. vomerinus ssp.

23 5
TURRICULATUS-SAKMARICUS Zones
Monograptus spiralis spiralis

24 1
ACINACES-CURTUS/PECTINATUS Zones
Climacograptus scalaris ?ferganensis
C. janischewskyi
C. normalis
C. rectangulatus
Diplograptus elongatus
Atavograptus praestrachani
Glyptograptus? sp.

26 2
CYPHUS-CURTUS/PECTINATUS Zones
Climacograptus scalaris ?ferganensis
C. janischewskyi
Diplograptus tcherskyi ssp.
D. elongatus
Atavograptus sp.

29 2
CYPHUS-CONVOLUTUS Zones
Diplograptus tcherskyi tcherskyi
Pristiograptus spp.

30 6
?CYPHUS Zone
Glyptograptus tamariscus ?magnus
G. sp.
Atavograptus ?atavus
A. praestrachani

35 7
FASTIGATUS-?PACIFICUS Zones
Orthograptus fastigatus
Paraorthograptus pacificus
Pseudogygites ?arcticus (trilobite commonly associated with *Orthograptus fastigatus*)

35 9
CYPHUS-CURTUS Zones
Coronograptus spp.
C. cf. *C. cirrus*

35 10
SAKMARICUS Zone
Cyrtograptus sakmaricus

37 21
CONVOLUTUS-TURRICULATUS Zones
Monograptus halli

47 5
TURRICULATUS Zone
Monograptus spiralis spiralis
M. halli

47 9
(uncertain range, possibly Wenlock)
Michelinoceras spp.? (U. Ord-Mississippian; orthocone cephalopod)
Pristiograptus ?dubius

48 4
TURRICULATUS-CRISPUS Zones
Monograptus discus
M. priodon
M. proteus
M. turriculatus minor

48 5
GRIESTONIENSIS-SAKMARICUS Zones
Monoclimacis vomerinus vomerinus
Monograptus halli
?M. priodon
?M. flagellaris
M. spiralis spiralis

48 6
TURRICULATUS-GRIESTONIENSIS Zones
Stomatograptus grandis grandis?
Monograptus priodon

48 7
SAKMARICUS-CENTRIFUGUS Zones
Cyrtograptus centrifugus
C. insectus
Monograptus priodon
Monoclimacis vomerinus vomerinus

48 9
late Llandovery-Wenlock
Monograptus priodon
Thallograptus arborescens

50 1
MINOR Zone to late Wenlock
only fragments preserved; identification is uncertain, but based on the robust form and the presence of overlapping hooked *priodon*-like thecae, it is likely that these are post- *minor* Zone and pre-Ludlow. However, the age is likely *minor* or slightly younger based on the occurrence of *sakmaricus* Zone graptolites well above this sample.

50 11
SAKMARICUS Zone
Cyrtograptus sakmaricus
Stomatograptus grandis ssp.

50 15
CONVOLUTUS-MINOR Zones
conularids
Monograptus decipiens
M. priodon
Atavograptus spp.
Dendrograptus sp.

50 16
PERNERI Zone
Cyrtograptus perneri

50 19
Wenlock
Pristiograptus dubius

51 1
TURRICULATUS Zone
Monograptus spiralis spiralis
M. halli
Stomatograptus sp.

51 2
TURRICULATUS Zone
Monograptus sartorius
M. priodon
M. turriculatus minor
Dictyonema pentlandica (Sil)
Pristiograptus spp. (Sil)

51 3
TURRICULATUS-GRIESTONIENSIS Zones
Monograptus priodon
M. sartorius
M. kovalevskyi

53 1
CRISPUS-SAKMARICUS Zones
Monograptus spiralis spiralis
M. griestoniensis minuta
M. priodon
M. n.sp.C (Melchin, 1987b)
Retiolites geinitzianus densereticulatus

53 2
RIGIDUS Zone
Cyrtograptus lapworthi
C. rigidus
Diversograptus ramosus
Monograptus flemingi

53 4
TURRICULATUS Zone
Monograptus spiralis spiralis
M. ?sartorius
Retiolites geinitzianus ssp.

54 11
upper Llandovery
Pentamerus subrectus? This pentamerid brachiopod is similar to *P. subrectus* identified from upper Llandovery strata of Norway (J.Jin pers. com., 1990)

54 12
GRIESTONIENSIS-SAKMARICUS Zones
Retiolites geinitzianus ?geinitzianus
Monograptus spiralis spiralis

54 14
GRIESTONIENSIS-SAKMARICUS Zones
Monograptus greistoniensis greistoniensis
M. priodon
Stomatograptus sp.
S. n.sp. (Melchin, 1987b)
Cyrtograptus sp. (fragment)

54 15
CENTRIFUGUS? Zone
Monograptus priodon
Cyrtograptus sp. (fragment)
C. centrifugus?

54 16
RIGIDUS Zone
Cyrtograptus sp. (fragment representing robust cyrtograptid, hence is probably at least early Wenlock in age)
Monograptus antennularius

54 17
RIGIDUS-PERNERI Zones
Monograptus riccartonensis
M. dubius?
Cyrtograptus sp.

54 19
PERNERI Zone
Cyrtograptus multiramis

54 20
PERNERI Zone
Cyrtograptus mancki
C. multiramis (ACL)

54 21
PERNERI
Cyrtograptus mancki
C. multiramis

54 22
TESTIS Zone
Cyrtograptus lundgreni
Monograptus testis
M. flemingi

54 23
TESTIS Zone
Cyrtograptus radians

54 24
TESTIS Zone
Monograptus testis n.ssp. (Lenz and Melchin, 1990)

54 26
TESTIS Zone
Monograptus testis
M. flemingi

54 27
SHERRARDAE Zone
Pristiograptus sherrardae
P. deubeli
Spinograptus nevadensis

54 29
SHERRARDAE Zone (possibly early Ludlow as well)
Plectograptus sp.
Pristiograptus sherrardae

54 30
PROGENITOR Zone
Pristiograptus roemeri

54 31
PROGENITOR-BOHEMICUS Zones
Bohemograptus bohemicus bohemicus

56 1
probably *SAKMARICUS* Zone
Monoclimacis vomerinus vomerinus
Stomatograptus canadensis (biostratigraphically poorly constrained; Lenz (1988) noted association with *C. sakmaricus)*
Monograptus priodon

56 4
upper Llandovery-middle Wenlock
Monograptus sp.
?Monoclimacis vomerinus ssp.

56 5
SAKMARICUS Zone
Cyrtograptus sakmaricus
Monoclimacis vomerinus vomerinus

56 9
TURRICULATUS-SAKMARICUS Zones
Monograptus priodon
M. spiralis spiralis

56 10
SAKMARICUS Zone
Monograptus spiralis spiralis (MJM)
Cyrtograptus sakmaricus

56 11
upper Llandovery-Wenlock
Monograptus priodon
Monoclimacis vomerinus vomerinus
sample sent to R. Thorsteinsson for confirmation.

56 12
CURTUS/ORBITUS-MINOR Zones
"diplograptids" (*minor* biozone or older)
Diversograptus capillaris
Monograptus ?halli
?M. pseudobecki
?M. spiralis contortus

56 14
ACINACES-CYPHUS Zones
Atavograptus sp.
Lagarograptus sp.
Diplograptus elongatus
sample sent to R. Thorsteinsson for confirmation.

56 17
FASTIGATUS Zone
Climacograptus longispinus supernus (more likely *C. long. hvalrosus* due to the thicker, basal spines)
Orthograptus fastigatus
sample sent to R. Thorsteinsson for confirmation.

56 18
FASTIGATUS Zone
Pseudogygites arcticus (trilobite)
Orthograptus fastigatus
Climacograptus lorrainensis

57 1
PERNERI Zone
Cyrtograptus multiramis
C. mancki
M. flemingi

57 2
no graptolites preserved

57 6
upper Llandovery?-Wenlock
Monoclimacis vomerinus vomerinus
Monograptus priodon

57 7
upper Llandovery-Wenlock
Monograptus priodon

57 10
CRISPUS Zone
Monograptus spiralis spiralis
M. crispus
Monoclimacis vomerinus vomerinus

57 11
TURRICULATUS-SAKMARICUS Zones
Monograptus spiralis spiralis
M. priodon
Pentamerus cf. *P. septentrionalis* (brachiopod)

57 12
CYPHUS-CURTUS/PECTINATUS Zones
Diplograptus elongatus

57 13
FASTIGATUS Zone
Orthograptus fastigatus
Climacograptus ?longispinus

58 4
This sample occurs 3.5m below 58 5, and it apparently contains abundant *Pseudogygites* sp. pydigia, suggesting assignment to the *fastigatus* Zone. The sample was sent to Calgary for conodont processing, but identifications were not available for this work.

58 5
FASTIGATUS Zone
Orthograptus fastigatus
Pseudogygites arcticus (Upper Ordovician trilobite, associated with *Orthograptus fastigatus*)

58 10
FASTIGATUS Zone (MJM)
Orthograptus fastigatus
Amplexograptus latus
Climacograptus supernus (MJM)

58 11
FASTIGATUS Zone
Orthograptus fastigatus
Samples sent to R.Thorsteinsson for confirmation.

58 13
ACINACES Zone
Climacograptus sp.
C. rectangularus
Samples sent to R.Thorsteinsson for confirmation.

58 14
ACINACES-CYPHUS Zones (confirmed by MJM)
Climacograptus janischewskyi
C. normalis
Glyptograptus tamariscus magnus
Metaclimacograptus hughesi

58 15
CONVOLUTUS-MINOR Zones
Atavograptus sp.
Monograptus halli?
M. decipiens ssp. (fragment)

59 12
TURRICULATUS-SAKMARICUS Zones
Monograptus priodon
M. spiralis spiralis

59 19
TENUIS Zone
Bohemograptus bohemicus ssp.
Bohemograptus bohemicus tenuis

59 20
Silurian
Pristiograptus? sp. (fragment)

59 21
LINEARIS Zone (ACL)
Monograptus fritschi linearis

59 25
lower Ludlow, probably *PROGENITOR* Zone (ACL)
Pristiograptus tumescens or *P. dubius*

59 26
RIGIDUS? Zone
Monograptus sp.
Streptograptus ?flexuosus (poorly preserved)

59 28
RIGIDUS-PERNERI Zones
Monograptus riccartonensis

59 32
Llandovery-Wenlock
Monograptus priodon (one specimen, poorly preserved)

59 34
SAKMARICUS Zone
Monograptus spiralis spiralis
M. priodon
Cyrtograptus sakmaricus
C. sp.

60 3
FASTIGATUS Zone
Pseudogygites arcticus (trilobite)
Orthograptus fastigatus

60 4
FASTIGATUS Zone
Climacograptus sp.
C. latus
Orthograptus fastigatus
Samples sent to R. Thorsteinsson for confirmation.

60 7
ACINACES Zone (MJM)
Atavograptus praestrachani
Lagarograptus acinaces
Metaclimacograptus hughesi
Glyptpograptus nikolayevi
Dimorphograptus confertus swanstoni
Atavograptus atavus

60 8
CURTUS/PECTINATUS Zone
Monograptus convolutus
Monograptus decipiens decipiens
Rastrites orbitus
R. approximatus approximatus
Diplograptus tcherskyi tcherskyi

60 9
SAKMARICUS Zone
Monograptus spiralis spiralis
Monoclimacis vomerinus vomerinus
Cyrtograptus sakmaricus

60 10
CENTRIFUGUS Zone
Monograptus aff. *M. flexilis*
M. priodon
M. flemingi

60 11
upper Llandovery-middle Wenlock
Monograptus priodon
Monoclimacis vomerinus vomerinus

60 12
middle Pridoli?
sample from drift; exact stratigraphic position uncertain
Monograptus cf *M. mironovi* (ACL)
Saetograptus leintwardensis primus
Monograptus sp.
Zone confirmed by ACL

60 13
PERNERI Zone
Monograptus flemingi
M. riccartonensis
Cyrtograptus mancki

60 14
TURRICULATUS Zone
Monograptus flagellaris
M. n.sp.A (Melchin, pers.com.)
M. priodon
M. spiralis excentricus
M.halli

60 15
lower Wenlock or late Llandovery (biozone confirmed by ACL)
Cyrtograptus murchisoni
Cyrtograptus sp.
C. laqueus
Monograptus flemingi
M. riccartonensis
Monoclimacis vomerinus vomerinus

60 18
PROGENITOR Zone
Bohemograptus bohemicus bohemicus (differentiation from *B. bohemicus tenuis*, in this case, based on thickness of distal end)
Monograptus fragment with *priodon*-like thecae
Monograptus sp.

60 20
probably *PROGENITOR* Zone based on an occurrence below *linearis* Zone.
Bohemograptus bohemicus ssp.

60 22
LINEARIS Zone
Bohemograptus bohemicus tenuis
Saetograptus fritschi linearis
Monograptus ceratus

60 23
TENUIS Zone
Bohemograptus bohemicus tenuis (poorly preserved, identification based on proximal thickness)

60 24
TENUIS Zone
Bohemograptus bohemicus tenuis (poorly preserved, identifiaction based on proximal thickness)

69 1
?*CYPHUS-CONVOLUTUS* Zones
Atavograptus? praestrachani
Pristiograptus sp.
Climacograptus sp.
C. medius ?brevicaudatus
C. scalaria ?ferganensis
?*Petalograptus* sp.
The last specimen, based on a summary investigation of graptolite morphogenesis by Rickards *et al.*, (1977), first appears in the upper part of the *cyphus* Zone. However, these specimens, as with most specimens of this sample, are fragmentary, and this data should be treated with caution.

69 2
TURRICULATUS-SAKMARICUS Zones
Monograptus spiralis spiralis

69 3
TURRICULATUS-SAKMARICUS Zones
Monograptus spiralis spiralis
M. ?priodon

69 4
SAKMARICUS Zone
Cyrtograptus ?sakmaricus (cladium only)
Monograptus priodon

69 5
PERNERI Zone
Monograptus sp.
M. flemingi
Cyrtograptus sp.
C. mancki (possibly *C. lundgreni*, poorly preserved)

69 7
TESTIS Zone
Monograptus testis n.ssp. (Lenz and Melchin. 1990)
Cyrtograptus lundgreni
C. multiramis
Acanthograptus sp. (little biostratigraphic value)

69 12
upper Llandovery (from drift)
?Monograptus veles
?M. cf *M. mutuliferus strigosus*
M. dublei

69 13
PROGENITOR Zone
Bohemograptus bohemicus bohemicus
Reticulograptus sp.
ostracoderm (fossil fish) fragment
conularid fragment

69 19
Silurian
Indet. monograptids

69 20
CHELMIENSIS? Zone
?Pristiograptus separabilis
P. chelmiensis

69 23
UNIFORMIS-HERCYNICUS Zones
Monograptus aequabilis aequabilis
M. uniformis? (fragment)

69 24
UNIFORMIS Zone
Monograptus uniformis angustidens
M. aequabilis ssp. (fragment, possibly from *uniformis* zone)
M. aequabilis aequabilis

69 26
FANICUS Zone
Monograptus fanicus
Linograptus sp. or *Arbesograptus* sp; difference based on disposition of cladium.

71A 1
BIRCHENSIS Zone (ACL)
Monograptus cf. *birchensis*

71A 2
BIRCHENSIS Zone (ACL)
Monograptus cf. *birchensis*

71D 2
BIRCHENSIS Zone (ACL)
Monograptus cf. *birchensis*

72 10
SAKMARICUS Zone
Cyrtograptus sakmaricus
Stomatograptus grandis imperfectus
Retiolites geinitzianus angustidens
Monograptus spiralis spiralis

72 12
GREISTONIENSIS-SAKMARICUS Zones
Monoclimacis vomerinus ssp.
Stomatograptus grandis imperfectus
Monograptus dubius

72 13
?CENTRIFUGUS Zone
Cyrtograptus ?centrifugus
Monograptus flemingi
Dictyonema sp.

72 14
SAKMARICUS Zone
Cyrtograptus centrifugus
Dictyonema sp.

75 1
Atrypoid brachiopods; no graptolites preserved.
see identification of brachiopods by J.Jin at the end of this appendix.

75 2
SILURIAN
Monograptus nudus

75 3
Encrinurus aff. *hyperboreus* (pygidium) much like Wenlock trilobite from Baillie Hamilton Island, described by Perry and Chatterton (1979)

75 4
PROGENITOR Zone
Monograptus nudus
Bohemograptus bohemicus ssp.
Pristiograptus tumescens

75 5
middle Wenlock
Sphaerexochus dimorphus (trilobite, probably mid-Wenlock, much like a specimen described by Perry and Chatterton, 1979)

75 6
middle Wenlock
Monograptus cf. mutuliferus strigosus (approx upper Wenlock)
Sphaerexochus dimorphus (mid-Wenlock trilobite described by Perry and Chatterton (1979)

75 8
PROGENITOR Zone
Bohemograptus bohemicus bohemicus

75 9
PROGENITOR Zone
Bohemograptus bohemicus bohemicus

75 10
PROGENITOR Zone
Monograptus romeri
Bohemograptus bohemicus bohemicus

75 11
PROGENITOR Zone
Monograptus romeri
Bohemograptus bohemicus bohemicus
Lobograptus sp.
Pristiograptus tumescens
P. aff. *deubeli*

75 12
LINEARIS Zone
Lobograptus progenitor
L. sp.
Bohemograptus bohemicus ssp.
Saetograptus fritschi linearis
Pristiograptus cf. *P. haupti*

75 13
Silurian
Monograptus sp.

75 14
Silurian
Monograptus sp.
Bohemograptus bohemicus ssp. noted on bedding planes near this sample location.

75 16
Silurian
Monograptus ssp.

76 2
PERNERI Zone
Cyrtograptus perneri
C. multiramis

76 3
TESTIS Zone
Cyrtograptus radians
Monograptus flemingi

76 5
TESTIS Zone
Monograptus testis n.ssp. (Lenz and Melchin, 1990)

77 1
SAKMARICUS-CENTRIFUGUS Zones
Cyrtograptus insectus
C. sp.
Monograptus flemingi
M. spiralis spiralis
M. speciosus
Retiolites geinitzianus ?augustidens
Retiolites sp.

77 2
SAKMARICUS Zone
Cyrtograptus sakmaricus
Monograptus ?tullbergi spiraloides (fragment)
M. priodon
Retiolites geinitzianus ssp.
Stomatograptus n.sp. (Melchin, 1987b)

77 3
PERNERI Zone
Cyrtograptus perneri
C. munchi
C. multiramis (large fragment)
Monograptus flemingi

77 4
WENLOCK
Cyrtograptus sp.
Monograptus flemingi

77 5
TESTIS Zone
Cyrtograptus radians
Monograptus flemingi
M. testis n.ssp. (Lenz and Melchin, 1990)

77 6
SHERRARDAE Zone
Pristiograptus sherrardae

79 1
ATAVUS-CURTUS/PECTINATUS Zones
Glyptograptus nikolayevi
Coronograptus gregarius ssp.

79 2
GRIESTONIENSIS-SAKMARICUS Zones
Stomatograptus grandis grandis
Monograptus spiralis spiralis
Monoclimacis vomerinus vomerinus

79 3
SAKMARICUS Zone
Cyrtograptus sakmaricus
Monograptus priodon

79 4
TESTIS Zone
Monograptus testis n.ssp. (Lenz and Melchin, 1990)

79 5
TESTIS Zone
Cyrtograptus ?radians

79-8
?LINEARIS Zone
?Saetograptus fritschi linearis

80 1
Upper Ord.-*CYPHUS* Zone
?Climacograptus normalis

80 2
CONVOLUTUS-MINOR Zones
Coronograptus decipiens ssp.
indet. diplograptids
Climacograptus ?medians
Monograptus fulcata
M. ?turriculatus minor (poorly preserved)

80 3
TURRICULATUS-CRISPUS Zones
Monograptus turriculatus ssp.
M. spiralis spiralis

80 4
?CENTRIFUGUS Zone
Cyrtograptus sp.
Monograptus flexosus

80 5
PERNERI Zone
Cyrtograptus sp.
C. perneri (ACL)
Monograptus ex gr. *firmus* (ACL)
M. riccartonensis?
M. priodon

M 1
Encrinurus sp. (trilobite fragment, probably indicating Silurian age).

The following 3 samples were collected from the "brown siltstone member" of the basal Devon Island Formation on Melville Island by J.C. Harrison in 1985. Harrison (1989, pers.com.) believes that this unit is a correlative of the Starfish Bay shale. The Melville Island samples were examined by the author and listed below.

C-133813 (GSC number)
collected 8-10m above base of the brown siltstone member of the Devon Island Formation (J.C. Harrison, pers. com., 1989)
Saetograptus fritschi linearis
S. leintwardensis ?primus
Bohemograptus bohemicus ssp.

C-133814
collected 15-20m below top of the brown siltstone member of the Devon Island Formation
Saetograptus fritschi linearis
S. leintwardensis ?primus
Bohemograptus bohemicus tenuis

C-131174
collected 13.5m below top of the brown siltstone member of the Devon Island Formation
Bohemograptus bohemicus tenuis

CONODONT IDENTIFICATIONS

Identification by A.D. McCracken, GSC, Ottawa, unless indicated by "TU", indicating an identification by Tom Uyeno GSC,ISPG, Calgary. Identifications of conodonts by A.D. McCracken are considered preliminary. A formal report with a complete listing of specimens is in progress by McCracken, and only conodont zones or approximate ages are available for some samples.

86 TF 3 2 (section 6-4) (C-146340)
Ludlow to Pridoli (TU)
Oulodus? spp. (Pa=2; M=2; Sa=2; Sb=2;)
Ozarkodina confluens (Branson & Murphy)
gamma (Pa=3), alpha (Pa=2), and beta(Pa=1) morphotypes of Klapper & Murphy (1975); unassigned element Sa(1).

86 TF 03 4 (section 6-4)(C-146337)
Ludlow to Pridoli (TU)
Oulodus? spp. (Pb=1; M=3; Sa=3; Sb=2; Sc=2)
Ozarkodina confluens (Branson & Murphy)
gamma (Pa=14) and alpha (Pa=10) morphotypes of Klapper & Murphy (1975); unassigned morphotype Pa=1; unassigned elements M(1); Pb(3); Sa(3).

86 TF 7 3 (section 6-7)(C-146344)
Possibly Mid. Ordovician to Late Devonian (TU)
Oulodus? sp. (Sb?=1)
indet. fragments (5)

86 TF 7 7 (section 6-7)(C-146349)
Late Wenlock (TU)
Ctenognathus murchisoni (Pander) (Pa=1),
Ozarkodina sp. (Pa=3; Pb=2; M=1)
Panderodus sp. (11)
Pelekysgnathus sp. (I=3)
unassigned elements

86 TF 8 2 (section 6-8)(C-146341)
Barren

86 TF 9 3 (section 6-9)(C-147003)
Barren

86 TF 10 1 (section 6-10)(C-146349)
Late Wenlock (TU)
Ctenognathodus murchisoni (Pander);

Ozarkodina sp. (Pa=3; Pb=2; M=1);
Panderodus sp. (11);
Pelekysgnathus sp. (I=3);
unassigned elements: M(4); Sa(3); Sb(2);Sc(2).

86 TF 10 2 (section 6-10)(C-146350)
Wenlock to Pridoli (TU)
Oulodus sp. (Sb=1; Sc=4);
Ozarkodina confluens (Branson & Mehl), alpha morphotype of Klappa and Murphy, 1975 (Pa=5), unassigned morphotypes;
unassigned elements Pb(9); M(3); Sa(2); Sb(3); Sc(2).

86 TF 10 3 (section 6-10)(C-147001)
Wenlock to Pridoli (TU)
Ozarkodina confluens (Branson and Mehl);
unassigned morphotypes (Pa=15; Pb=5; M=3; Sa=1; Sb=1; Sc=1)

02-03 (C-104681)
ORDOVICICUS

03 05 (C-104682)
ORDOVICICUS

05 05 (C-104683)
CELLONI-AMORPHOGNATHOIDES

06 01 (C-104684)
lost in processing

07 01 (C-104685)
barren

10 01 (C-104686)
lost in processing

10 04 (C-104687)
Llandovery
Distomodus sp.

11 01 (C-104688)
indet. conodonts

1? 05 (C-104689)
identification not available

13 15 (C-104690)
identification not available

15 05 (C-104691)
Middle Ordovician to Middle Devonian (TU)
Oulodus? sp. [M?-1, Pb-1, Sb-1]
Panderodus sp. [11]

15 07 (C-104692)
Middle Aeronian
Mirza (1976) reported *K? manitoulinensis* from the reefal facies of the Cape Storm Formation and the Cape Phillips Formation on the eastern Arctic Islands. According to Aldridge (1985), *K? manitoulinensis* is middle Aeronian (*discreta-staurognathoides* Zones) (TU)
K? manitoulinensis (Pollock, Rexroad and Nicoll) [Pa-1]
Unassigned element [1]

15 09 (C-104693)
Middle Ordovician to Middle Devonian (TU)
Panderodus sp. [1]
Unassigned Pb element [1]

16 01 (C-104694)
PATULA-CRASSA Zones
based on the age range of *Kockelella absidata* given by Barrick and Klapper (1976), from the *amsdeni* to lower part of the *variabilis* Zone, i.e., approximately upper part of the *patula* Zone to the *crassa* Zone of Walliser (1964, 1971) (TU)
Kockelella absidata Barrick and Klapper [Pa-4, all small]
Ozarkodina sp. [Pa-12, Pb-14, M-4, Sa-1, Sb-1]
Panderodus sp. [109]
cf. "*Spathognathodus*" n.sp. A of Mirza (1976) [Pa-2]
Apparatus A of Uyeno (1981) [3]
Unassigned Sa element [1]

16 09 (C-104695)
Lower Ordovician to Middle Devonian (TU)
Panderodus sp.[38]
"*Spathognathoides*" sp. [Pa-1, small, fragmented]

17 05 (C-104696)
M. Ordovician to Devonian

17 06 (C-104697)
Middle-Upper Ordovician
Plectodina sp.

17 07 (C-104698)
lost in processing

17 08 (C-104699)
Silurian

18 01 (C-104700
Barren

19 49 (C-154922)
Late Silurian to Early Devonian, based on the occurrence of *Ozarkodina steinhornensis* and *Pedavis* sp. (TU)
Belodella sp.
Dapsilodus? sp.
Decoricnus cf. *D. fragilis* (Branson & Mehl).
Pseudooneotodus beckmanni (Bischoff and Sennemann).
spathognathodiform element (Pa of *Ozarkodina steinhornensis* Ziegler?).
simple cones M_2 (of icriodontan?); S_1 of *Pedavis*? sp.

21 04 (C-154928)
Early Silurian; mid Aeronian to Telychian (TU)
Dapsilodus obliquicostatus (Branson and Mehl).
Decriconus fragilis (Branson and Mehl).
Oulodus? fluegeli (Walliser) (Pa=1; Pb=4; M=2; Sa=4; Sb=1; Sc=1).
Panderodus sp.
Pseudooneotodus beckmanni (Bischoff & Sannemann).
Walliserodus? sp.
"Neoprioniodus" planus Walliser.

23 03 (C-104701)
CELLONI Zone
Amorphognathoides irregularis

26 01 (C-104702)
Kirkfield-Gamachian

27 04 (C-104703)
indet. conodont elements

29 01 (C-104704)
Kirkfieldian-Gamachian

30 08 (C-104705)
ORDOVICICUS Zone

31 03 (C-104706)
identification not available

31 08 (C-104733)
possibly Middle to Late Silurian (TU)
?Ozarkodina confluens (Branson & Mehl) [Pa?-2, both badly fragmented]
Apparatus A of Uyeno (1981) [2]
Unassigned elements: M[1], Sc?[1]

31 14 (C-104734)
indet. elements

32 02 (C-104707)
KENTUCKYENSIS Zone

33-02 (C-104708)
Barren

34 01 (C-104709)
ORDOVICICUS-ENSIFER Zones

34 03 (C-104710)
identification not available

34 04 (C-104711)
Kirkfieldian-Gamachian

34 06 (C-104712)
M. Ord.-Dev.
Panderodus gracilis

35 04 (C-104713)
fauna 11-13 (Mays.-Gam.)

36 06 (C-104714)
ORDOVICICUS-ENSIFER Zones

36 15 (C-104715)
Middle Ord.
Glyptoconus quadraplicatus

36 17 (C-104716)
middle Whiterockian

38 02 (C-104717)
Middle Ordovician to Devonian
Panderodus spp.

38 11 (C-104718)
Middle Ordovician to Devonian
Panderodus sp.

39 16 (C-104719)
Barren

39 18 (C-104720)
Middle Ordovician-Devonian
Panderodus gracilis

40 02 (C-104721)
CELLONI Zone

41 01 (C-104722)
Shermanian-Marysvillian

41 02 (C-104723)
Late Edenian-Gamachian

41 15 (C-104724)
AMORPHOGNATHOIDES to *ranuliformis* Zones, based on the range of *Aulacognathus chapini* in Alaska (Savage, 1985) [=*amorphognathoides* to lower part of the *PATULA* Zones of Walliser, 1964, 1971]; the occurrence of *Ozarkodina* aff *O. polinclinata* suggests lower part of this range, i.e., *AMORPHOGNATHOIDES* Zone (Savage, 1985) (TU)
Aulacognathus chapini (Savage) [Pa-1]
Ozarkodina aff *O. polinclinata* (Nicoll & Rexroad)[Pa-1]
Panderodus sp. [3]

46 05 (C-104725)
identification not available

46 17 (C-104726)
identification not available

47 01 (C-104727)
Shermanian-Gamachian

47 04 (C-104728)
CELLONI-AMORPHOGNATHOIDES Zones

47 10 (C-104729)
identification not available

47 12 (C-104730)
mid. Shermanian-Gamachian

49 01 (C-104731)
Barren

49 23 (C-104732)
M.Ord.-Dev.
Panderodus gracilis

62 06 (C-174674)
Barren

62 13 (C-174681)
Mid.Ord.-Dev. (TU)
Panderodus gracilis (Branson and Mehl) a/b element

62 17 (C-174685)
Barren

70C 1 (89-HBB-TdF-70c-1)(C-180554)
Middle Ordovician to Middle Devonian (TU)
Panderodus sp. [9]
Unassigned elements: M [2], Sb [1]

70B-1 (89-HBB-TdF-70B-1)(C-180558)
Middle Silurian *SILURICUS* Zone (TU)
*Oulodus?*sp. [Sb-1, Sc-2]
Ozarkodina douroensis Uyeno [Pa-15, Pb-6, M-1, Sb-1]
Panderodus sp. [42]

71 D3 (C-180560)
Mid.Ord.-Dev. (TU)
Panderodus gracilis (Branson and Mehl) a/b elements

C-152942
Gunnars Island, collected 240m below top of the oolite facies of the pinnacle reef facies of southeastern part coast of the island, section 75.
Middle Silurian to Early Devonian (*PATULA* Zone to Emsian) (TU)
Ozarkodina excavata excavata (Branson and Mehl) [Pa-1, M-1, Sa-2, Sb-1]
Panderodus sp. [7]

C-152944
Cañon Fiord, collected 50m below top of mud buildup facies of the Allen Bay Formation.
Middle Silurian upper part of the *PATULA* Zone of Walliser, 1964, 1971; or *amsdeni* Zone of Barrick and Klapper (1976) (TU)
Dapsilodus obliquicostatus (Branson and Mehl) [18]
Kockelella absidata Barrick and Klapper [Pa-1]
Kockelella patula Walliser [Pa-1]
Ozarkodina excavata excavata (Bronson and Mehl) [Pa-1]
Panderodus sp. [56]
unassigned elements Pb[2], Sa[1], Sc[2]
80 6 (C-152943)
Late Silurian (possibly *patula* Zone to Emsian)
?Ozarcodina excavata (Branson and Mehl) [Pa-1, badly fragmented]
Panderodus sp.

BRACHIOPOD IDENTIFICATIONS

Most of the following report paraphrases a written communication from Jisuo Jin, May, 1990.

31 9
lower Ludlow

Atrypoidea saaremaaensis Copper and Rubel, 1977. 2 shells. The species was separated from the more commonly known *Atrypoidea prunum* (Dalman, 1828) by Copper and Rubel. In Gotland and Estonia (type area), the species is from the Hemse Beds (lower Ludlow).

Didymothyris didyma (Dalman, 1828). 11 complete shells. In Gotland, the species is most common in the Hemse beds (lower Ludlow), and in the Mulde and Klinteberg beds (upper Wenlock; Bassett and Cocks, 1974).

31 11
lower Ludlow

Atrypoidea scheii (Holtedahl, 1914). 5 shells. This species has been described by Fortier *et al.* (1963), Smith (1976) from the Douro Fm, of Devon island and Ellesmere Island.

Atrypoidea saaremaaensis Copper and Rubel, 1977. 5 shells. The species was separated from the more commonly known *Atrypoidea prunum* (Dalman, 1828) by Copper and Rubel. In Gotland and Estonia (type area), the species is from the Hemse Beds (lower Ludlow).

37 1
early Maysvillian-Richmondian
Lepidocyphus sp.

52 5
lower Ludlow

Conchidium cordata (Lindström, 1861). Pedicle valves, mostly broken. Upper Wenlock to lower Ludlow, based on the occurrence in the Klinteberg Beds, Gotland.

74 6
lower Ludlow

Clorinda sp. 2 broken pedicle valves. long-ranging genus: Lower Silurian-Middle Devonian

Atrypoidea saaremaaensis Copper and Rubel, 1977. 2 shells. The species was separated from the more commonly known *Atrypoidea prunum* (Dalman, 1828) by Copper and Rubel. In Gotland and Estonia (type area), the species is from the Hemse Beds (lower Ludlow).

75 1
lower-middle Ludlow

Gypidula sp. 7 broken pedicle valves. Although the genus ranges from Lower Silurian to Upper Devonian, specimens from this sample belong to the ribbed species, which becomes common in post-Wenlock time.

Stegerhynchus borealis (von Buch, 1834). 3 shells. The species is Upper Wenlock age in Gotland. Jones (1981) reported this brachiopod from the Read Bay Formation, Somerset Island. Jin (1989) discovered a specimen in upper Llandovery, Jupiter Formation.

Stegerhynchus angaciensis (Chernyshev, 1937). Two shells, including 1 damaged. In Tuva, USSR, and Mongolia (type area), it is most common in rocks of upper Ludlow-lower Pridoli age, but may occur in Wenlock. Lenz (1970) described the species from the Road River Fm (middle Ludlow to early Pridoli) of northern Yukon.

Ancillotoechia cf. *A. minerva* (Barrande, 1847). 1 shell. In Bohemia (type area), the species occurs in Wenlock rocks. The genus is most common in Wenlock and Ludlow, although is longer ranging from Llandovery to Lower Devonian.

Linguopugnoides sp. 2 specimens. The genus is most common in Upper Silurian and Lower Devonian (Havlicek, 1961).

Eoglossinotoechia cacuminata Havlicek, 1959. 1 shell. In Bohemia (type area), it occurs in Ludlow through Lower Devonian strata.

Atrypa (Atrypa) reticularis (Linnaeus, 1861). 3 specimens. Long ranging, Lower Silurian to Upper Devonian.

Plectatrypa imbricata (Sowerby, 1840). 1 shell. In the type area, Britain, the species occurs in limestones dated as Wenlock. Bassett and Cocks (1974) reported the species from rocks of uppermost Wenlock to lower Ludlow in Gotland.

Atrypoidea saaremaaensis Copper and Rubel, 1977. 1 shell. The species was separated from the more commonly known *Atrypoidea prunum* (Dalman, 1828) by Copper and Rubel. In Gotland and Estonia (type area), the species is from the Hemse Beds (lower Ludlow).

Howellella sp. 1 pedicle valve. Long ranging, Silurian-Devonian.

75 3
upper Wenlock to lower Ludlow

Conchidium cordata (Lindström, 1861). Upper Wenlock to lower Ludlow, based on the occurrence in the Klinteberg Beds, Gotland.

Plagiorhyncha cf. *P. cordata* (Lindström, 1861). 3 shells. In Gotland (type area), the species occurs in the Slite and lower Mulde beds (upper Wenlock).

75 6
upper Wenlock

Plagiorhyncha cf. *P. cordata* (Lindström, 1861). 1 shell. In Gotland (type area), the species occurs in the Slite and lower Mulde beds (upper Wenlock).

Lissatrypa sp. 1 shell. Long ranging, Lower Silurian to Middle Devonian.