SEDIMENTOLOGY AND STRATIGRAPHY OF
EASTERN AND CENTRAL BELCHER ISLANDS,
NORTHWEST TERRITORIES.

by

© BRIAN DAVID RICKETTS

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*Contemplate all this work of Time,
The giant labouring in his youth;*

Tennyson

'In Memoriam A.H.H.'

CXVIII.

ABSTRACT

The middle Precambrian (Aphebian) Belcher Group consists of a thick (7000-9000 m) but remarkably regular succession of distinctive rock units. Collectively, these units represent a complex picture of basin evolution where long-term changes in depositional environments have taken place in response to fundamental changes in tectonic regime. By establishing vertical and lateral facies relationships, paleoenvironments and regional paleoslope changes, four major cycles of sedimentation and volcanism are recognized.

In the first cycle (Kasegalik Formation) 1200 m of carbonates and mudstones represent the transition from supratidal (evaporitic) to shallow subtidal environments that developed on a broad marine platform. Sedimentation was interrupted by emplacement of thick plateau basalts of the Eskimo Formation, thus initiating the second cycle (Eskimo Formation to Laddie Formation). In this cycle, an extended period of subsidence resulted in deposition of 1800-2000 m of carbonate and clastic sediments. The vertical succession from the Fairweather Formation to the Laddie Formation is considered to represent an assemblage of facies that were laterally associated in space and time, and is interpreted as a transgressive platform-slope-basin sequence. The regional paleoslope dipped west to southwest.

The rapid transition from argillites of the Laddie Formation to shallow subtidal and intertidal sandstones of the Lower Rowatt Formation, marks the beginning of the third cycle (Rowatt Formation to Kipalu Formation). Here, as in the two preceding cycles, each formation is considered to be diachronous, representing a shoaling-upwards sequence that formed along a prograding shoreline. Banded ironstones (Kipalu

Formation) were deposited in a restricted or barred basin that was protected from the open ocean by a system of carbonate buildups (Rowatt Formation) and subtidal and intertidal sand bars. The carbonate buildups were positive elements relative to mean sea level.

A second episode of effusive and explosive volcanism (Flaherty Formation) initiated the fourth cycle (Flaherty Formation to Loaf Formation), where accumulation of a thick volcanic edifice resulted in reversal of the paleo^oslope (to east-dipping) and rapid subsidence of the Belcher Basin. Flaherty strata are tentatively compared with emergent segments of an oceanic ridge system, for example Iceland. Subsequent filling of the basin by thick turbidites and fluvial sediments probably took place within a submarine fan-braided river (delta?) complex.

Dolomitization of marine platform-to-slope carbonates (Fairweather to Laddie Formations) occurred within a wide spectrum of sedimentary and diagenetic environments, over a time span that probably ranged from penecontemporaneous to a few million years. It is likely that penecontemporaneous dolomitization of shallow-platform carbonates was widespread during middle Precambrian times.

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INTRODUCTION

Preamble

The highly sinuous outline of the Belcher Islands is a product of the interaction between two fundamentally different and opposing forces: those that build mountains, and those that wear them down. More than 1600 million years separate a tectonic event that led to folding and faulting of the sedimentary and volcanic strata composing the Belcher Group, and continental glaciation that scoured, gouged and plucked this Precambrian terrane during Pleistocene times to produce the unique landscape seen today.

The Belcher Islands are north of the tree-line and exposure of the Precambrian bedrock is excellent. For example the prominent ridges and arcuate peninsulas that characterize the islands are composed largely of volcanic rocks (the Flaherty Formation) and are exposed along more than 2000 km of coastline. In addition, the low grade of metamorphism and scarcity of penetrative deformation have ensured that even the finest details of sedimentary, organo-sedimentary and volcanic structures are well preserved. Thus the Belcher Group provides an opportunity to examine some excellent examples of strata which record the conditions of sedimentation, growth of primitive life forms, and volcanism of almost two billion years ago.

Previous Work

The Belcher Islands (Fig.1) was one of the last groups of islands in Hudson Bay to be systematically explored, although various reports of the islands had been known for some time before the end of the last century. It is uncertain whether Henry Hudson ventured as far south as the Belcher Islands prior to his disappearance in 1610 (Manning, 1947). However, Captain Coats did manage to visit part of the Belcher Islands

Figure 1. Regional map, showing location of the Belcher Islands in eastern Hudson Bay, in relation to the other Proterozoic fold belts (shaded) around the Ungava Craton.

during one of his many expeditions to Hudson Bay from 1727 to 1751.

Before his successful expedition in 1914, Flaherty obtained a map of the Belchers drawn by an Eskimo named Wetalltok, an employee at the Hudson Bay outpost in Richmond Gulf. The map proved to be remarkably accurate and indicated that the islands were far more extensive than previously suspected. It wasn't until September of that year, after two previously abandoned attempts, that the schooner "Laddie" arrived at Tukarak Island (Flaherty, 1918). An earlier attempt by A.P. Low (Low, 1902) also was abandoned because of inclement weather and ice conditions.

Later, Moore (1918) working mostly in the Laddie Harbour area, proposed the name Belcher Series for this sequence. An important discovery at this time was the presence of possible algal structures (stromatolites). Moore is also credited with the first report of microfossils from Precambrian rocks. Subsequent studies by Young (1922), Woodbridge (1922) and the Belcher Mining Corporation Ltd. in the 1950's, centered largely on the ironstones (Kipalu Formation) to determine their economic value. Young also subdivided the sequence into a number of units (Table 1).

The first detailed mapping of the Belcher Islands was undertaken by G.D. Jackson (1960), who afforded the sequence group status and subdivided it into 16 units. These units subsequently were given formation status (Dimroth et al. 1970). K-Ar age dating of volcanic rocks collected by Jackson confirmed the Aphebian age of the Belcher Group. An age of about 1.8 G.a. (Rb-Sr) was subsequently obtained by Fryer (1972) from volcanic rocks and shales of the Flaherty and Omarolluk Formations respectively. Interest in Precambrian microfossils associated with stromatolites in the Kasegalik and McLeary Formations was renewed with

TABLE 1

History of stratigraphic nomenclature of the Belcher Group.

Moore(1918)	Young(1922) (Boundaries approx.)	Jackson(1960)	Dimroth et al.(1970)	This Study
BELCHER SERIES The Tookcarak Diabase and Basalt are probably equivalent to the Eskimo Fm., Flaherty Fm and Haig Intrusion.	Not Mapped	Unit 16	Loaf Fm.	Loaf Fm.
		Unit 15	Omarolluk Fm.	Omarolluk Fm.
	'Effusives'	Unit 13 14	Flaherty Fm.	Flaherty Fm. Haig Int.
	Division 1	Unit 12 (Kipalu Fm.)	Kipalu Fm.	Kipalu Fm.
	Division 2	Unit 11	Mukpollo Fm.	Mukpollo Fm.
	Division 3	Unit 10	Rowatt Fm.	Rowatt Fm.
		Unit 9	Laddie Fm.	Laddie Fm.
	Division 4	Unit 7	Costello Fm.	Costello Fm.
		Unit 6	Mavor Fm.	Mavor Fm.
		Unit 5	Tukarak Fm.	Tukarak Fm.
	Division 5	Unit 4	McLeary Fm.	McLeary Fm.
	Division 6	Unit 3	Fairweather Fm.	Fairweather Fm.
	Lower Diabase	Unit 2	Eskimo Fm.	Eskimo Fm.
	Not Mapped	Unit 1	Kasegalik Fm.	Kasegalik Fm.

studies by Hofmann (Hofmann and Jackson, 1969). A diverse assemblage of filamentous and coccoid cyanophytes, morphologically similar to some Gunflint Formation and Bitter Springs Formation forms, have been reported by Hofmann (detailed systematics in Hofmann, 1976, and several other papers). Stromatolites from the McLeary and Mavor Formations have also received some attention (Donaldson, 1976).

Paleocurrent determinations in pre-Flaherty strata by Jackson (1960) and Dimroth et al. (1970) indicated that the paleoslope dipped west to southwest. This was confirmed in a later study by Barrett (1975).

A geochemical and petrographic study of the Eskimo Formation volcanics was undertaken by Stirbys (1975). Volcanic and volcanoclastic rocks of the Flaherty Formation also have been examined by Schenk (1959), Leggett (1974) and Ware (1978).

Carbonate concretions in greywackes from the Omarolluk Formation were studied by McEwen (1978).

The Present Study

Because of the excellent mapping of the Belcher Islands by Jackson in 1958-59, only minor revisions to the stratigraphy have been introduced here (Table 1 and Fig.2). The present study has concentrated on detailed facies analyses and interpretation of paleoenvironments, which in turn have served as a framework for establishing the overall history of sedimentation and volcanism in the Belcher Basin. This was done by measuring a number of stratigraphic sections in detail (almost 20,000 m were measured with a Jacob's staff, tape and by pacing), by 'walking out' specific units and by making about 2000 paleocurrent measurements. The study concentrated on Tukarak, Mavor and Innetalling Islands and outlying islands, although sections also were measured and examined along

LOAF	220 m ⁺		Red & grey cross-bedded arkoses, mudstones.
OMAROLLUK	2105 m ⁺		Well-bedded greywackes and shales, typical Bouma cycles. Minor flat pebble conglomerates filling channels. Dewatering structures, concretions. Thin tuffs near base.
FLAHERTY	250-1950 m ⁺		Tholeiitic basalts: massive and pillowed flows. Variety of volcanoclastics: thinly bedded tuffs, pyroclastic turbidites, massive pyroclastic beds (predominantly hyalotuffs). Unit 14 = Haig Intrusion.
KIPALU	105-125 m		Banded ironstones, micrites, jaspers.
MUKPOLLO	40-145 m		Cross-bedded quartz arenites, siltstones.
ROWATT	290-390 m		U.Mbr. Brecciated grainstones, dololutes. L.Mbr. Variety of sandstones, mudstones
LADDIE	230-350 m		Red and green argillites and shales.
COSTELLO	240-370 m		Carbonate rhythmites, minor allodapic calcarenites, slump structures.
MAVOR	90-245 m		Extensive stromatolitic dolostones.
TUKARAK	40-93 m		U.Mbr. 'Ribbon rock', stromatolitic dolostones. L.Mbr. Fine sandstones, stone rosettes.
MCLEARY	365-455 m		U.Mbr. Stromatolitic dolostones. M.Mbr. Dolarenites, lutites, beachrock. L.Mbr. Dolostones, abundant stone rosettes. Beachrock Marker Bed at base.
FAIRWEATHER	365-610 m		U.Mbr. Sandstones, siltstones, channels. L.Mbr. Pisolitic dolostones, desiccated mudstones, sandstones.
ESKIMO	0-910 m		Tholeiitic basalts: massive flows, rare pillows, thin volcanoclastics.
KASEGALIK	1220 m ⁺		Stromatolitic dolostones. Red mudstones, halite and sulphate casts.

Figure 2. Generalized stratigraphic column of the Belcher Group,
Belcher Islands, N.W.T.

the Kasegalik anticline (Flaherty Island), eastern Flaherty Island and the Bakers Dozen Islands (Fig.3). A total of 19 weeks was spent in the field during the summers of 1976 and 1977.

The Belcher Group can be subdivided into four major cycles on the basis of changes in sedimentation patterns, for example changes resulting in transgressive-to-regressive sequences, and interruption of sedimentation by periods of volcanism. Each of these cycles is discussed in the first four chapters of this thesis.

The diagenetic histories of a number of carbonate units also are discussed in Chapter 5 and an hypothesis for dolomitization proposed.

A summary of the history of deposition in the Belcher Basin, structure and correlations are given in Chapter 6.

The results of an experimental investigation into carbonate growth in gel media are given in Appendix A, aspects of the late Quarternary geology of the Belcher Islands in Appendix B, and data from modal analyses of sandstones from a number of formations are tabulated in Appendix C.

Studies of paleomagnetism of the Precambrian bedrock were undertaken concurrently by W.A. Morris (in 1976) and P. Schmidt (in 1977). An additional suite of samples was collected by W.A. Robertson in 1973. The results of this study were not available at the time of writing.

Figure 3. Map showing station numbers and locations of sections measured.

CHAPTER 1

A MIDDLE PRECAMBRIAN TRANSGRESSIVE CARBONATE SEQUENCE: PART I. THE KASEGALIK TO MCLEARY FORMATIONS

INTRODUCTION

The first section of this chapter deals with the Kasegalik and Eskimo Formations which, because of time limitations in the field, were not studied in detail. Traverses of the Kasegalik Formation in the central part of Flaherty Island confirm most observations of Jackson (1960).

KASEGALIK FORMATION

This, the lowest stratigraphic unit in the Belcher Group, is discontinuously exposed along the Kasegalik anticline and the Churchill Sound area, where it attains a thickness of more than 1200 m (the base is unexposed). Previously unmapped outcrops of dolostone and red mudstone belonging to this unit are also exposed in a glacially carved valley in central Tukarak Island. In the northern Churchill Sound area, Bell and Jackson (1974) subdivided this unit into five zones:

- (a) lowermost exposure consisting of grey dolostones and about 20% red mudstones;
- (b) red mudstones (60%) and grey dolostones;
- (c) stromatolitic dolostones;
- (d) cherty stromatolitic dolostones;
- (e) pink argillaceous dolostones.

The lower zones containing red mudstones also contain halite casts and probable sulphate moulds. Evaporite pseudomorphs were also observed in Kasegalik Lake sections (central Flaherty Island) and the above subdivision appears to be valid along these sections. Other indicators of subaerial exposure for this formation include desiccation polygons and tepee structures.

Cryptalgal structures in zones c and d exhibit a general upwards progression from small low mounds containing flat and digitate mats, to large domes (2-5 m across) with some furcate and digitate branching. Interbedded clastic dolostones contain flat pebble conglomerates, some of which exhibit edgewise packing in crude radial patterns that in cross-section resemble the stone rosette structures common in several formations higher in the succession.

Jackson (1960) reported tuffs near the contact with the Eskimo volcanics and this is confirmed in the present study.

The Kasegalik Formation appears to be an onlapping marine sequence where supratidal (sabkha?) to shallow subtidal environments are represented. Sedimentation was interrupted by extrusion of thick basalt flows of the Eskimo Formation.

ESKIMO FORMATION

The volcanic succession composing the Eskimo volcanics shows considerable variations in overall thickness, ranging from a maximum 900 m in the Eskimo Harbour region to some 600 m in central Tukarak Island, thinning southward and westward to only a few meters (Jackson, 1960). The formation consists predominantly of basaltic flows up to 20 m thick; the flows are typically massive and commonly exhibit columnar jointing. Stirbys (1975) noted south-directed flow bending of joints in a section on Flaherty Island, but bending towards the northeast was observed in a flow at the southern end of Isaacs Lake on Tukarak Island. Implications of regional flow directions based on this type of structure are therefore premature at this stage. A few pillowed flows were observed, but unlike many examples in the Flaherty volcanics, these are of much more limited lateral extent.

The basalts are largely aphanitic, consisting mainly of ophitic plagioclase (albite-oligoclase) and clinopyroxene (augite). Stirbys (1975) considered the sodic plagioclase to be a product of albitization. The degree of albitization tends to increase towards the top of flows. Porphyritic flows occur locally and make useful stratigraphic markers. Variolitic and spherulitic basalts also occur.

Interflow sediments consist of a few thin red and green argillites and thin hyaloclastite beds.; the latter are made up of angular and non-vesicular blocky fragments that probably are a product of quenching or flow-top brecciation.

Chemical analyses by A.F. Stirbys and R.T. Bell (in Stirbys, 1975) indicate predominantly tholeiitic compositions. On the basis of chemistry, a preponderance of massive flows and evidence of possible subaerial weathering, Stirbys concluded that the Eskimo volcanics were extruded as plateau or flood basalts, possibly as a result of fissure-type eruptions east of the Belcher Islands. The fact that these rocks are sandwiched between subaerial and shallow-water deposits of the Kasegalik and Fairweather Formations supports this suggestion.

Metamorphism attained a grade of prehnite-pumpellyite to subgreen-schist facies. This grade is slightly higher than that of the Flaherty volcanics, probably as a consequence of greater depths of burial.

FAIRWEATHER FORMATION

Introduction:

Excellent outcrops of the Fairweather Formation are exposed in the Fairweather Sound area, particularly at the north end of Mavor Island, where an auxiliary section was designated by Dimroth et al. (1970, p.100),

and the south end of Tukarak Island (Fig.4). At least 338 m of nearly continuous exposure of the upper two-thirds of Fairweather strata can be seen near the southeastern corner of Tukarak Island (Section 39) and this is suggested as a reference section, in preference to that on Mavor Island. The formation is poorly exposed over the remainder of the Belcher Islands, and descriptions and interpretations are based mainly on the Tukarak Island sections.

The thickness of the Fairweather Formation is about 500 m on Tukarak Island, 365 to 610 m along the Kasegalik anticline, and 530 m on Moore Island.

The contact between the Eskimo and Fairweather Formations is poorly exposed except on west Tukarak Island (Section 1) where agglomerates grade rapidly into dolarenites, pisolitic dolostones, thin sandstones and ferruginous argillite lenses. In places the contact appears to have a relief as much as two meters. Lenticular beds of ferruginous argillite and ironstone were encountered only in the lower 20 to 30 m of Fairweather strata, some apparently filling depressions along the contact. Because the argillites form part of a continuous sedimentary sequence with overlying strata, they are included in the Fairweather Formation. An occurrence of granular jasper ironstones about a meter thick at the top of the Eskimo Formation, was reported by Jackson (1960) at the north end of Churchill Sound. A zone of aeromagnetic anomalies associated with the same stratigraphic level in other parts of Churchill Sound and Flaherty Island was attributed to a continuation of the ironstones. Thus the ironstones may extend over much of the Belcher Islands, although not necessarily as a continuous unit.

The contact with the overlying McLeary Formation is conformable

Figure 4. Schematic representation of measured sections and paleocurrents of the Pairweather Formation. All paleocurrent diagrams are plotted relative to the arrow on the locality map. The key to symbols used can be found in the pocket at back.

and is placed at the transition from quartz arenite beds to thick buff dolostones. An additional criterion that proved to be most useful on Tukarak Island is an orange-weathered dolarenite or grainstone bed about 50 to 80 cm thick, that contains zones of beachrock and small digitate stromatolites. The base of this bed is taken as the boundary. The presence of a similar bed on Flaherty Island could not be verified because of poor exposure.

The Fairweather Formation is subdivided into two new members (Fig.5): a Lower Member consisting predominantly of interbedded pisolitic dolostones, sandstones and mudstones, and an Upper Member containing sandstones, siltstones and grit beds.

LOWER MEMBER (310 m)

Scattered outcrops of the lower 160 m of this member consist of pisolitic dolostone beds up to four meters thick, interbedded with thin green and red mudstones, siltstones and sparse beds of purple, immature sandstone. Scattered digitate algal mats occur in some of the dolostones.

Lenticular, banded ironstones at the base of this member consist of hematite-rich layers and silty micaceous layers. There is some colour and grain-size grading, a feature commonly observed in the Kipalu Formation ironstones.

Pisolitic Dolostones and Associated Tidal Flat Sediments:

The upper 150 m of the Lower Member, well exposed on Tukarak Island, is illustrated in Fig.5. Here orange-weathered pisolitic dolostones figure prominently and constitute the upper part of shoaling-upwards cycles which also include sandstones and mudstones. A portion of Section 39 is depicted in Fig.6 and illustrates the important features of the cycles. The cycles, of which at least 18 can be recognized, range

Figure 5. Composite stratigraphic column of the Fairweather Formation (from Sections 1 & 39, Lower Member; Section 39, Upper Member).

Figure 6. Part of Section 39 depicting the rhythmic nature of pisolitic dolostones; P= pisolitic dolostone, M= Mudstone, S= sandstone - siltstone. Fairweather Formation.

in thickness from 1 to 30 m and each can be subdivided into two distinct lithological units.

(1) Clastic Units: In several cycles, basal strata consist of tabular and lensoidal quartz arenites and subarkoses that contain herringbone cross-bedding, channels, scour-and-fill structures, reactivation surfaces, current ripples, flaser bedding and rare dololutite boulder conglomerates. Some thin lutite beds are sharply truncated and eroded. Interbedded mudstones commonly display desiccation structures.

Soft-sediment deformation structures such as load balls are common. In section 4, clastic dykes of dolosiltite intrude dolomitic siltstones and fine sandstones. In cross-section the dykes are similar to examples illustrated by Remnitz and Marshall (1965) that formed in tidal sediments on the South Alaska coast as a result of the 1964 Alaskan earthquake.

The top of each clastic unit is characterized by siltstones and red mudstones containing abundant desiccation cracks. Symmetrical straight-crested ripples and interference ripples are also common. Small pods of coarse sandstone and mudstone rip-ups suggest sudden high-energy sand influxes into an essentially low-energy environment.

Paleocurrent data (Fig.4) show bimodal current directions in the coarse-grained sediments, with transport in a west-southwest direction. This supports earlier suggestions that the paleoslope dipped in this direction (Dimroth et al. 1970; Barrett, 1975). A much broader spread of azimuths was recorded for the symmetrical ripples as might be expected of bedforms that developed under variable wind conditions typical of many modern tidal flats. The bimodal and some quadrimodal current directions, together with a number of diagnostic sedimentary structures (herringbone cross-beds, reactivation surfaces and flaser bedding), are indicative of

a tidal regime and in particular an intertidal environment. Some supratidal environments are also suggested by the abundance of desiccation cracks.

Details of sandstone petrology are given in Ricketts (1978). The sandstones are relatively mature with well-rounded quartz grains and minor amounts of feldspar, chert and volcanic rock fragments, the latter probably being derived from the Eskimo volcanics. Quartz overgrowths are well developed. Authigenic green tourmaline occurs in the sandstones as cavity fillings associated with both primary and secondary (moldic) porosity, as a replacement of boron-enriched silt clasts, and in laminated mudstones. The boron was probably derived from weathering of volcanic rocks of the Eskimo Formation and also as a result of increased salinities in a broad supratidal environment. Petrographic evidence suggests that precipitation of tourmaline was a relatively late diagenetic event.

(2) Pisolitic Units: In each cycle, this division consists of distinctive buff-coloured pisolitic dolostones. A variety of structures are present that compare favourably with examples of modern and ancient vadose pisolites (e.g. Dunham, 1969; Reeves, 1970; Wardlaw and Reinson, 1971; Winchester, 1972). Macroscopic textures include:

(i) pisolites up to 20 mm in diameter exhibit asymmetric and composite growth patterns (Pl.1 Fig.1). Pisolite asymmetry is usually associated with geotropic elongation where gravity has induced thickening of successive growth layers at the base (Pl.14 Fig.1). Evidence of interrupted pisolite growth is seen in composite forms where layers truncated by dissolution, or changes in the rate of carbonate precipitation, have been overgrown by a subsequent generation of isopachous carbonate crystals. In places, adjacent pisolites are enveloped by a single growth layer producing composite forms.

(ii) interference growth patterns occur where pisolite growth was inhibited by clastic detritus or by neighbouring pisolites.

(iii) in some thin laminated dololutites, crust expansion has formed tepee structures. In addition, erosion and truncation of lutite layers has locally resulted in beds rich in mud chips.

(iv) thin, grey dolomitic laminae having a pallisade or micro-digitate structure, overlies some of the pisolite-rich zones. Based on their similarity to structures in other units of the Belcher Group, these are interpreted as algal mats. In places the mats show some silicification, a feature that also is common to stromatolitic units elsewhere. Locally, the mats show evidence of erosion.

(v) there is a fundamental difference in the nature of the upper and lower bounding surfaces of the dolostones. Whereas the lower contacts are highly irregular (non-erosional) and generally diffuse, grading into the underlying mudstones over 5 to 10 cm, the upper contacts are sharp and show evidence of erosion.

(vi) secondary microstylolitic seams abound in these dolostones and contribute to the dirty appearance of the rocks. Similar structures have been described from Cretaceous hardgrounds in chalks of southern England (Kennedy and Garrison, 1975).

A number of petrographic features also provide valuable clues to possible mechanisms that might have been operative during formation of the dolostones. The pisolites are now composed entirely of dolomite where discontinuous and overlapping micritic growth layers alternate with microspar layers. Individual layers are somewhat diffuse, unlike the pisolites documented by Dunham (1969), and are more like caliche pisoliths described by Esteban (1976). There is no evidence of any precursor fibrous fabrics.

Micritic zones between the pisolites appear clotted. Poorly sorted, highly angular fragments of micrite and microspar, some of which are laminated, are cemented by coarse dolomite and appear to have formed by in-situ brecciation of former micrite beds and pisolites. Some pisolites have grown around micrite fragments, thus providing evidence for multiple dissolution-precipitation events.

Interpisolite voids together with voids formed by dissolution, are filled with two or three stages of carbonate cement and occasionally barite: an initial fine-grained rim of brownish scalenohedral dolomite crystals, that may represent relics of an early fibrous calcite rim cement, is followed by a much coarser drusy (rhombohedral) dolomite cement where crystals range from 100 to 200µm in size. The final stage of void infill commonly consists of large calcite crystals up to 2mm across.

Of the features described above, inverse grading, composite growth forms, polygonal fitting and downwards (geotropic) elongation of pisolites all suggest an in situ origin. Support for this is seen in the nature of the bedding contacts. Although some evidence for erosion exists, namely algal mat rip-ups and mud chips, there is no evidence for prolonged bedload transport (eg. cross-bedding). In addition, pisolites are confined to the dolostones and no evidence of pisolite reworking, other than a few angular dolostone cobbles at the upper contact, was seen in sandstones overlying these beds.

Pisolites are known to form in weathering zones, especially where the pedogenic processes are characterized by alternating periods of wetting (dissolution) and drying (precipitation). The macroscopic and microscopic features described above strongly suggest a caliche origin for the Fairweather dolostones.

An example of one of these pisolitic beds is illustrated in Fig.7. Here, as in most of the Fairweather caliche beds, a relatively distinct zonation has developed:

(a) a lower, sand-poor pisolite zone, that contains a number of distinct pisolite bands, each exhibiting inverse grading and culminating in an erosion surface. Each of these bands represents a single period of caliche formation, where carbonate sands and muds accumulated and subsequent leaching-precipitation processes within a vadose diagenetic zone resulted in the growth of pisolites. Erosion and scour of individual caliche beds, accompanied by influxes of sediment, may have been caused by storm-tide surges. Further evidence of high-energy erosion is seen in the truncation of lutite layers.

(b) the pisolite-rich zone grades over several centimeters into a sand-rich, pisolite-poor zone. Here, recognition of discrete caliche horizons is difficult; many of the pisolites are somewhat diffuse. Interruptions of caliche development by influxes of sand (some possibly wind blown) and mud appear to become more abundant towards the top of this zone, possibly heralding the return to intertidal conditions that are represented by the overlying deposits.

To summarize, each cycle in the Lower Member displays a transition from an intertidal, tidal flat environment to a supratidal environment, culminating in the development of a caliche bed.

UPPER MEMBER (180-200 m)

The Upper Member consists predominantly of red and green siltstones and fine sandstones in tabular beds up to 2 to 3 m thick. Cutting these rocks are channels, filled with sublitharenites and quartz arenites, that range in size from one to 80 m across. Structures in the fine-grained

Pisolite Poor
Zone

Mudstones

Sandy Dolostones.
Streaks of Sand
& Silt.

Tepee Structures

Pisolite Zone

Truncated Dololutes

200 cm.

Transition Zone
Mudstones

Figure 7. Detailed sketch of zonation within pisolitic dolostone beds showing irregular, non-erosional basal contact, minor erosional discontinuities within pisolitic zones (P), and a sharp upper boundary. The dolostones are sandwiched between desiccated mudstones.

lithologies include lenticular bedding, symmetrical ripples, load balls (especially in proximity to channels) and desiccation cracks. A few thin laminae and flasers of dololutite also occur.

Tabular-bedded sandstones are uncommon, except in the upper 10 to 15 m of this member. In these lithologies the most common bedforms are planar cross-beds, scour-and-fill, subhorizontal laminae, large ripples with amplitudes of 10 to 15 cm and mud-filled troughs, some herringbone cross-beds and mud or dololutite flasers. Channel lag conglomerates most commonly consist of mud chips. A few thin grit beds are present and contain quartz and feldspar, with subordinate chert, dololutite and volcanic fragments. Paleocurrent directions (Fig.4) are bimodal, similar to those in the Lower Member.

These rocks are considered to represent deposits that accumulated on tidal flats dissected by tidal channels.

SUMMARY:

The transition from the underlying Eskimo volcanics to the Fairweather Formation is relatively sharp, and is characterized by the presence of lenticular beds of ferruginous argillite and mudstone, intercalated with tuffaceous(?) and pisolitic dolostones. The ironstones are interpreted as having formed in small lakes or lagoons that were isolated from the open sea by a broad intertidal platform. Iron-rich waters were supplied to the lagoons by groundwater runoff from the adjacent volcanic plateau (Eskimo Formation).

The intertidal platform can be further subdivided into: (a) a supratidal-intertidal region containing desiccated tidalflat muds and cross-bedded sandstones representing a number of shoaling-upwards cycles, each cycle culminating in the formation of caliche beds (Lower Member);

(b) a lower tidal flat region that was characterized by sand and mud flats, and was dissected by a system of tidal channels (Upper Member).

MCLEARY FORMATION

INTRODUCTION

The stromatolitic dolostones, for which the McLeary Formation has become well known, were first illustrated by Moore (1918), who suggested possible association of the algal structures with ironstones of the Kipalu Formation (Moore's illustrations were probably of Upper McLeary strata). More recently, detailed studies of the stromatolites have been undertaken by Donaldson (1976a) and Golubic and Hofmann (1976): Hofmann (1974, 1975, 1976a) and Hofmann and Jackson (1969) also have identified a diverse and well-preserved microfloral assemblage. The stromatolites are highly diversified, with cryptalgal structures ranging from simple mats to extensive and complex biostromes. Most of these structures occur in the Upper Member of the McLeary Formation (Fig.8) as defined by Jackson (1960) and Dimroth et al. (1970, p.101) from the type section on eastern Tukarak Island. The Middle and Lower Member also contain a number of interesting sedimentological features and these are described below.

A type section was designated by Jackson (1960), 4.8 kilometers west of McLeary Point. The thickness of the formation is relatively constant over most of the Belcher Islands, ranging from 365 to 450 m on Tukarak and Mavor Islands. Contacts with the underlying Fairweather Formation and overlying Tukarak Formation are conformable. The lower boundary is marked by a sandy dolostone or grainstone bed forming a distinctive marker that can be traced about 30 km along eastern Tukarak and Mavor Islands. This bed contains some good examples of beachrock and will henceforth be referred to as the Beachrock Marker Bed (Fig.8).

Figure 8. Composite stratigraphic column of the McLeary Formation.

LOWER MEMBER: (125-140 m)

This lithologically distinctive unit is well exposed on Tukarak and Mavor Islands, but is poorly represented in sections along the Kasegalik anticline; in Flaherty Island sections the basal strata of the McLeary Formation are similar in many respects to Middle Member rocks exposed on Tukarak and Mavor Islands, and hence the possibility exists that the Lower Member pinches out west of Tukarak Island (Fig.9).

The apparent thinning of the stratigraphic section west of Laddie Harbour (Section 35) is probably due to faulting.

The Beachrock Marker Bed:

This bed defines the base of the McLeary Formation and has been traced over most of south Tukarak and Mavor Islands. It consists of sandy dolarenites or grainstones and displays some excellent examples of beachrock. Descriptions and interpretations of beachrock horizons in the McLeary Formation are summarized in Donaldson and Ricketts (1977, and 1979).

A feature characteristic of modern beachrock is the preferential and penecontemporaneous in situ lithification of specific grainstone beds within relatively unlithified beach sands. Recognition of structures indicative of preferential lithification are therefore critical for identification of ancient beachrock; these are summarized below:

(i) a uniform grainstone lithology throughout the grainstone beds under consideration;

(ii) an erosional discontinuity within the grainstone bed that exhibits features such as channel scours or joints enlarged by solution and erosion;

(iii) the lower boundary is typically conformable and diffuse as would be expected from a downwards-migrating cementation front;

(iv) brecciation of specific beds within the grainstones, locally forming flat-boulder conglomerates, whereas the underlying grainstones are undisturbed.

(v) growth of encrusting algal mats on the upper surfaces of lithified beds or blocks. Algal mats tend to conform to surface irregularities and may drape around overhanging ledges;

(vi) juxtaposition of these beds with strata of shallow marine, intertidal and supratidal origin;

(vii) original cement types are mostly isopachous fibrous or micritic high-Mg calcite, or aragonite. Unfortunately the preservation potential of these microfabrics is low, especially in rocks that have undergone extensive dolomitization.

Two additional features are well developed in the Beachrock Marker Bed:

(a) small, finger-like, laterally linked stromatolites locally grew from the surface of the beachrock bed. Algal mats have also grown on the undersides of some boulders that were perched against adjacent beachrock slabs.

(b) edgewise conglomerates have been stacked against the sides of some blocks (Pl.1 Fig.2) in cross-section configurations that resemble stone rosettes, (see next section).

Positive identification of beachrock provides a valuable tool for recognizing ancient strandlines. Modern beachrock is forming predominantly in tropical and subtropical environments and hence is also of considerable importance for paleoclimatic interpretations.

The remainder of the Lower Member consists largely of buff coloured planar and wavy laminated dololutites containing abundant edgewise

conglomerates, and minor intercalated siltstones and sandstones.

Dimroth et al. (1970) note the presence of some quartz arenite beds, but these generally occur near the top of this member.

Stone Rosettes:

On bedding surfaces, edgewise dololutite slabs are commonly arranged in a crude polygonal pattern (Pl.1 Fig.3) and in some places appear to be stacked vertically against lutite cobbles. In cross-section the edgewise clasts form vertically and obliquely stacked clusters, commonly in crude radial patterns, within beds and lenses (Pl.1 Fig.4). Analogous vertical packing structures, variously referred to as stone rosettes, stone roses, stone-centered polygons and vertical stone packings (Gregory, 1930; Rosanski, 1943; Dionne, 1971; Kostyaev, 1973) can be observed on many modern beaches in both marine and lacustrine environments. Two requirements that are common to nearly all reported occurrences are: an adequate supply of platy, bladed or broadly concave-convex clasts, and relatively high-energy wave action particularly within the wave breaking zone of beaches. Thus pavements of these vertically stacked stones commonly form from shale or slate fragments derived from the underlying shore-platform, or from shells. The McLeary examples are also interpreted as stone rosettes that formed pavements in a beach environment (Donaldson and Ricketts, 1977; Ricketts and Donaldson, 1979).

The dololutite slabs rarely display any form of deformation that might be expected if the clasts were in a plastic state. Furthermore, conspicuous grain size differences between the slabs, in which most dolomite grains are less than 10 μ m, and the coarser dolosiltite matrix, suggest the possibility of penecontemporaneous lithification and dolomitization. Dolomitization of this nature is also evident in other supratidal

and intertidal deposits in the Belcher Group, and is discussed in more detail in Chapter 5. Thus the lutite slabs were probably partially lithified at the time of stone rosette formation.

Additional evidence for a supratidal, probably hypersaline environment, is provided by gypsum casts which occur at two different stratigraphic levels in the Lower Member. In the first, gypsum casts are associated with thin, pyrite-bearing jasper layers and lenses, 5 to 10 cm thick. Somewhat higher in the section, a grey dololutite bed contains abundant gypsum casts up to 1.5 cm long. Zones containing gypsum casts have been traced several kilometers parallel to strike on south Tukarak and Mavor Islands.

MIDDLE MEMBER:

The most striking aspect of this member is the number of sandstone-dolarenite cycles that are terminated by beds of beachrock. In some sections up to five different beachrock beds were observed, some representing more than one episode of lithification and erosion.

Tabular beds of sublitharenite, sandy dolarenite (up to 50% quartz) and minor quartz arenites make up the bulk of this member. Most of the sandstones are well sorted and the quartz grains are well rounded. Other lithologies include thin dololutites and some stromatolitic dolostones. The cycles usually begin with sublitharenites (the lithic component is predominantly dolomite) containing planar herringbone cross-beds, current ripples (amplitudes of 12 cm), some trough cross-beds and subhorizontally laminated sets (Fig.10). The vertical transition to dolomitic sandstones and dolarenites is accompanied by a general increase in subparallel laminated sandstones with low-angle truncation of set boundaries. In some instances, steeply dipping planar cross-bedding

(20-25°) abruptly overlies the laminated sandstones. Both of these bedforms are common in beach deposits (eg. Andrews and Van der Lingen, 1969) and it is suggested that both beach foreshore (laminated sandstones) and beach ridge deposits (steeply dipping cross-beds) are represented here.

Beachrock:

Beachrock-bearing grainstones recur at the top of several cycles (Pl.1 Fig.5). A typical section is depicted in Fig.10. In a number of outcrops it is possible, by tracing individual beds laterally for several meters, to observe changes in bedform that can be related to different stages of beachrock formation, from unlithified sands to flat boulder conglomerates (Fig.11). Erosion channels with algal crusts draped around small overhanging ledges occur in some sandy dololutes that overlie beachrock beds (Fig.12), and these may be comparable to small domal stromatolites that have overgrown desiccated and scoured muds in ponded inlets (cf. Gebelein, 1976, Fig.2). The configuration of these laminated micrite crusts argues against them having formed by deposition of muds from suspension.

The grainstones consist of both quartz and dololite clasts in a grain-supported framework that is now cemented by dolomite microspar. Original clast boundaries are well preserved and lutite clasts of different grain sizes sharply abut one another, there being little or no recrystallization gradients across the contacts. This texture is a common feature of other clastic dolostone units in the Belcher Group that are associated with intertidal and supratidal deposits, and is interpreted as representing erosion and deposition of early diagenetic, or contemporaneously dolomitized carbonate sediments.

Figure 10. Representative sections of the Middle Member, McLeary Formation. (a) Sst = sublitharenites and quartz arenites, Gst. = grainstone; note several sandstone-grainstone-beachrock cycles. (b) the transition from sandstones to lutites (lut), dolarenites (Ar) and grainstones (Gst.) in the upper part of this member. Section 31, Mavor Island.

Figure 11. Field sketch depicting lateral changes in the internal organization of a beachrock bed, McLeary Formation. (a) continuous sequence in cross-bedded grainstones, (b) preferentially lithified bed with infilled joints. The eroded upper surface of this bed provides evidence for exposure during or immediately following lithification, (c) dislodged angular blocks of beachrock, and (d) crude imbrication in grainstone boulder conglomerate. Section located at the southeast corner of Tukarak Island.

Figure 12. Eroded dololutes (b) overlying jointed beachrock (a). Note crusts of algal mats (c) lining undercut channels and joints. These in turn are overlain by cross-bedded grainstones (d). Field sketch from southeast Tukarak Island.

The upper few meters of this member exhibit a progressive change in lithology, from the sandstone-dolarenite-beachrock cycles, to beds of dololutite and sandy dolarenite (Fig.10). The increased carbonate content is also accompanied by an increase in the variety of desiccation structures, particularly mudcracks and tepee structures, and also in the appearance of oscillation ripples, pisolitic beds and edgewise conglomerates. Distinct cryptalgal structures are rare, except for a few grey clotted zones. A number of beachrock-bearing dolarenites also contain crystal casts that are rimmed by quartz sand grains and infilled with sparry dolomite that may be pseudomorphic after gypsum.

Some Stromatolites Associated with Beachrock:

In addition to the finger-like, digitate stromatolites in the Beachrock Marker Bed, two different cryptalgal forms were observed in some beachrock-bearing beds in the Middle Member.

(i) Stabilization of flat beachrock-boulder conglomerates (Section 35) appears to have taken place by encrusting algal mats and colonization by closely packed columnar stromatolites (Pl.1 Fig.6). The stromatolites are slightly turbinate, laterally linked and erect with furcate-style branching. However, unlike most other stromatolite-bearing units in the McLeary Formation, most have poorly developed wall structures. Synoptic relief is less than 10 mm.

(ii) Isolated expanding bioherms occur in grainstone lithologies in north Tukarak Island (Sections 27 & 28). Growth of columns took place in two main stages: initial growth of a simple, non-branching bulbous mass, which later gave rise to a dendroid branching pattern of subcylindrical columns. The divergent style of branching bears some resemblance to the Late Precambrian form, Anabaria Komar.

The relatively high-energy environment in which both types of

stromatolite grew is obvious from the association of sedimentary structures, and yet the stromatolites consist of dololutes containing very little coarse clastic debris that might have been trapped by successive layers of actively growing algae. The paucity of even silt-sized detritus is characteristic of many stromatolites throughout the Belcher Group, suggesting that the original carbonate may have been chemically precipitated, rather than being mechanically derived. A well-washed, high-energy environment such as that inferred for most of the Middle Member rocks, would tend to remove much of the finer clastic material.

UPPER MEMBER

The Upper Member is divided into three broad zones and although transitions between these zones are completely gradational over several meters, it provides a convenient means for description and subsequent analysis. This subdivision is not intended to be formal.

The overall pattern of sedimentation can be summarized thus:

(a) a lower zone, some 60-70 m thick, consisting of dololutes, dolarenites and a few sandstone beds. Few recognizable cryptalgal layers occur near the base, but they do become more abundant towards the top of the zone; (b) a middle zone, 80-90 m thick, that displays a marked increase in algal mats and simple domal stromatolites; (c) an upper zone, 20-30 m thick, that is characterized by a rich and diverse stromatolite assemblage.

Lower Zone:

The lower 10 to 20 m consists of thinly bedded buff dololutes and dolarenites containing sedimentary structures indicative of shallow-water deposition alternating with extended periods of subaerial exposure. For example, laminated dololutes exhibiting interference ripples and

superimposed desiccation cracks (Pl.2 Fig.1) commonly are partly eroded and overlain by mud chip-bearing dolarenites. Graded bedding occurs in some of the coarser grained rocks, with mud chips and flat pebble conglomerates near the base. Other structures include large pull-aparts that may have originated as joints or fractures in partially lithified crusts, stone rosettes, a few crust expansion structures (tepees) and some pisolitic or oolitic layers. Although caliche beds similar to those in the Fairweather Formation were not observed here, some pisolites display interference and composite growth patterns suggestive of a vadose origin. Although distinct cryptalgal laminates are sparse, some dolarenite beds display a greyish-clotted texture that may indicate the former presence of sand-binding algae.

A number of tabular and lenticular sandstone and dolarenite beds persist into the lower part of this member. Most tabular beds contain subparallel laminated sandstones, planar cross-beds with low and steep angle set truncations, current ripples and herringbone cross-beds, and resemble many of the sandstone units in the Middle Member that were interpreted as beach deposits. The lenticular bedded sandstones contain abundant trough cross-beds, and probably represent channels.

Laminated algal mats appear in the upper 40 to 50 m of this zone, initially as thin layers coincident with grey clotted dolostones, becoming more prominent higher in the succession. Based on gross morphology, algal mats in the Lower Zone fall into five broad groups (Table 2): parallel laminated, wavy laminated, and tufted laminates that are characterized by a scalloped fabric and occasionally by palisade structures; stratiform digitate mats with very fine fenestrae; and pisolitic laminate associations (Pl.1 Figs.2, 3 & 4). Detailed examination of a number of

		Desiccation Cracks	Tepee Structures	Gypsum Casts	Rip-ups	Edgewise Conglomerates
Upper Zone	Branched Stromatolites	R	-	-	R	-
	<u>Conophyton</u>	R	-	-	R	-
Middle Zone	Elongate Domes (non-branching)	R	-	-	C	C
	Parallel Laminates	C	C	R	C	C
Lower Zone	Wavy Laminates	C	C	-	C	C
	Tufted Laminates	C	-	R	C	C
	Stratiform Digitate Laminates	R	-	-	R	-
	Oncolites	-	-	-	C	C

INCREASING EXPOSURE
↓

TABLE 2. Comparison of crpytalgal laminate types and sedimentary structures, Upper Member, McLeary Formation.

C = common, R = rare.

sections reveals a rather specific association of sedimentary structures with each of the major mat types (Fig.13a-e). Tufted, wavy and parallel laminated varieties are for example commonly desiccated, eroded and in places overturned. Tepee, or crust-expansion structures, and rare gypsum casts are also closely associated. Edgewise conglomerates exhibiting stone rosette patterns on exposed bedding surfaces, are relatively common and many appear to have been stabilized by overlying algal mats (Fig.13d). In contrast to this, digitate mats are generally confined to beds that show little evidence of desiccation, although some are eroded.

The third type of cryptalgal laminate, the pisolitic mats, occur mostly in grey clotted dolarenites that also display erosion structures such as mat rip-ups, or edgewise conglomerates. A number of features help to distinguish these pisolites from forms developed in the caliche deposits (eg. those in the Fairweather Formation):

- they occur in grey cryptalgal laminated and locally stromatolitic dolostones, and are particularly evident in zones that were subjected to high-energy erosion and deposition,
- downwards elongation of pisoliths is uncommon, although composite growth forms do occur, especially in areas where the pisoliths were stabilized by overlying laminar and tufted mats,
- some forms exhibit small 'micro'-digitate colonies that grew on the upper surface of stabilized pisoliths (Pl.2 Fig.4).

A paleoenvironmental interpretation of strata from the Lower Zone, based solely on inorganic sedimentary structures, indicates that deposition probably took place in a supratidal to intertidal environment, that often was subjected to extended periods of subaerial exposure.

Figure 13. Variations in cryptalgal laminates and sedimentary structures, Lower zone, Upper Member, McLeary Formation. (a) Mavor Island, (b) east Eskimo Harbour, (c) Tukarak Island, (d) Tukarak Island, (e) west Eskimo Harbour.

The cryptalgal laminates on the other hand, are morphologically very similar to a variety of forms described from some modern tidal flats, notably Shark Bay (Logan et al. 1974) and Andros Island (Hardie and Ginsburg, 1977); here a distinct zonation of morphotypes has been recognized in relation to such environmental parameters as the degree of desiccation and water depth. With this comparison in mind, the tufted mats in the McLeary Formation would be indicative of upper intertidal to supratidal conditions, whereas digitate forms are more likely to have grown in a slightly deeper intertidal zone that was characterized by less intense desiccation (Table 2). The paleoenvironmental interpretation of lower strata of the Upper Member can thus be established by two independent criteria, namely inorganic sedimentary structures and cryptalgal structures.

Silicification:

An additional feature common to nearly all cryptalgal structures in the McLeary Formation is silicification and formation of black cherts. These occur as pods or irregular layers and generally are confined to the cryptalgal structures, rather than the intercalated sediments. Microstructural details of laminae are well preserved in the cherts, compared to the almost complete destruction of microfabrics by dolomitization and neomorphism. The well-preserved microflora in cherts from stromatolitic units in the Kasegalik and McLeary Formations is, as Hofmann (1975) has suggested, a strong indication of essentially penecontemporaneous preservation by permineralizing solutions.

Middle Zone:

The proportion of grey to buff dolostones increases rapidly in this zone and coincides with an increase in stromatolite-bearing beds,

in particular the appearance of broad domal stromatolites. The ratio of interbedded dololutites to stromatolitic beds decreases from about 3:1 in the lower strata, to 1:1 near the top, where dolarenites and sandstones become sparse. Indications of subaerial exposure such as desiccation cracks still persist, but tepee structures are generally lacking (Table 2). Oscillation and interference ripples are common, and the few remaining sandstones contain structures indicative of current reversals, namely herringbone cross-beds and reactivation surfaces.

The stromatolite domes are simple, non-branching, up to a meter across and 10 to 30 cm high and are made up of smooth or crinkly mats or laminates with a distinct palisade structure. Overall dome morphology is clearly demonstrated in good bedding surface exposures in north Tukarak Island (Pl.3 Fig.1). In plan view, dome elongation may be as high as 4:1 or almost circular. More than 400 measurements of elongation azimuths were made in McLeary strata and these are summarized in Fig.14 along with cross-bed measurements. The general trends of elongations very closely parallel the principle paleocurrent directions indicated by cross-beds and ripples. Hoffman (1967) and Logan et al. (1974) have demonstrated that stromatolite mound elongations generally parallel the direction of wave propagation and hence are normal to the shoreline. The examples illustrated in Pl.3 Fig.1 invite comparison with modern analogues at Shark Bay, in particular the forms that are actively growing in the middle to lower intertidal zone, where wave action and tidal scour are relatively strong. Evidence that similar wave and current conditions existed in the McLeary is provided by small scour-and-fill structures, lutite and algal mat rip-ups and edgewise conglomerates stacked against dome walls.

Figure 14. Paleocurrent trends based on measurements of stromatolite elongations and cross-bedding, from Tukarak, Mavor and Johnny's Islands. Inset shows elongation trends from Eskimo Harbour.

Additional Stromatolites in the Middle Zone:

A single bed containing isolated dendroid stromatolites was encountered in Sections 21 and 22 at 200 m, and in Sections 31 and 33 at about 75 m below the top of the McLeary Formation. In the latter two sections this bed was traced 13 km along strike. The stromatolites consist of strongly divergent columns, each made up of convex laminae, and in some, conical laminae having distinct inflections and crestal zones (Pl.3 Fig. 2). Wall structure also appears to be moderately well developed, and branching occurs in the upper growth stages.

Near the top of this zone, a highly distinctive bed containing densely packed Conophyton stromatolites was observed in the following sections: 27, 28, 1, 2 and possibly 31 (Fig.9), and is thus traceable for at least 32 km along strike. Details of this bed are provided in the discussion on stromatolites (pp. 63-64).

Upper Zone:

The vertical change in stromatolite morphology and abundance in the Upper Member of the McLeary Formation, culminates in the top 20 to 30 m with the appearance of a diverse stromatolite assemblage consisting of bulbous, furcate, digitate, dendroid and conical forms. The few remaining interbedded sediments consist of laminated and clotted dololutes, some showing desiccation cracks, and thin sandy dolarenites that contain ripples, small scours, mud chips and a few stromatolite fragments. The top one to two meters is a conspicuous buff dolostone consisting entirely of closely packed columnar stromatolites, which form part of large, interconnected elongate mounds having amplitudes of up to a meter and wavelengths of 2 to 10 m (Pl.3 Fig.3). This biostrome complex can be traced over most of the Belcher Islands for distances of 50 km and more parallel to strike, and 80 km in an east-west direction.

PALEOCURRENTS

Over 150 cross-bed measurements and some 400 stromatolite elongations (elongations $>2:1$) from McLeary strata are graphically represented in Fig.14. Cross-bed azimuths indicate abundant bimodal and some tri- or quadrimodal current directions, corresponding to structures such as herringbone cross-beds and interference ripples, reactivation surfaces, flaser and lenticular bedding. A few channel-fill sandstones cutting intertidal sediments and containing trough cross-beds display predominantly unimodal azimuths, where currents flowed southwestwards. From these results it can be established that the regional paleoslope dipped west to southwest.

At any one section, stromatolite elongations exhibit a relatively narrow spread of values generally parallel to the major current directions. Locally however, considerable deviations can occur (eg. Section 6, 9, 10, 31 & 35, where elongations trend northerly or even northwesterly). Nevertheless, even here these trends parallel a major current direction. Deviations in elongation trends may be attributed to local changes in shoreline configuration, for example as a result of shoreface progradation or the building out or erosion of headlands, which would result in local deflections in current and wave directions. It should also be noted that stratigraphically, several tens of meters separate most cross-bedded sandstones from the stromatolitic horizons. Thus, although it is likely that the two facies were laterally associated, the shoreline configuration at the time of stromatolite growth may not have closely coincided with shorelines which existed during deposition of the beach and channel sands.

A striking anomaly in the paleocurrent pattern is observed in the

elongations of Conophyton stromatolites, which trend almost at ninety degrees to the regional pattern. Measurements from three widely spaced stations (Fig.14) are consistent and there is little evidence for 'abnormal' wave or current conditions indicated by other sedimentary structures. This phenomenon may be due to some primary attribute of the conophytos, and will be discussed more fully in the following section.

UPPER MCLEARY STROMATOLITES

The usefulness of stromatolites for intrabasin correlation has been clearly established in a number of studies, particularly those undertaken by Russian geologists over the last two decades (eg. Raaben, 1969; Seminkhatov, 1976). Similar studies on other continents, notably Australia, North Africa and North America have also been successful (Donaldson, 1963; Hoffman, 1968; Walter, 1972; Bertrand-Sarfati, 1976). To date, the majority of form taxa have time ranges that appear to be restricted to the Riphean and/or Vendian, especially in the Soviet Union where a biostratigraphic zonal scheme has been erected on the basis of specific stromatolite assemblages. The possibility that these stromatolites could be used for worldwide correlation has been the subject of considerable debate (eg. Cloud, 1942; Hoffman, 1967; Cloud and Semikhatov, 1969; Raaben, 1969). Correlations between upper Precambrian successions in Australia and the U.S.S.R. have also been tentatively suggested (Walter, 1972, pp. 81-82).

Recently, taxa similar to those apparently restricted to the Late Precambrian have been described in strata of Middle Precambrian age, notably Hofmann (1969b), Donaldson and Taylor (1969), Donaldson (1976a) and Hoffman (1976). The test , whether these stromatolites are

similar to those of Riphean age, thus simply extending their time ranges further into the Precambrian, or, whether they represent convergent evolution, will only be realized when more detailed studies are undertaken of stromatolite sequences that are older than Riphean (Aphebian and early Paleohelikian). Such studies are still very much at an early stage. The diverse stromatolite assemblage in the McLeary Formation thus offers an excellent opportunity to at least begin testing the validity of existing biostratigraphic schemes.

Terminology:

Terms used to describe the morphological attributes of stromatolites are those of Hofmann (1969a). This is a taxonomic approach, where a complete range of morphological features is required to define a single taxon, rather than a purely descriptive classification (eg. Logan et al. 1964). The Russian nomenclature of supergroup, group and form-taxa is also used (Raaben, 1969).

General Features:

Stromatolitic dolostones of the McLeary Formation are composed of dololutites and recrystallized dololutites with only sparse detrital quartz and feldspar. Sand grains, ooids or peloids, that in many other Precambrian stromatolitic sequences are concentrated along individual laminae, are absent from the McLeary examples. Sedimentary structures in interbedded dolarenites suggest a relatively high-energy depositional environment. In a few sections, individual beds of columnar stromatolites show evidence of rapid, high-energy erosion, possibly as a result of storms.

A preliminary comparison between McLeary Formation stromatolites and Riphean stromatolites was made by Donaldson (1976a) who observed

that all four supergroups, defined by Raaben (1969), are represented, namely kussiellids, gymnosolenids, tungussids and conophytonids.

Donaldson also suggested comparison with specific groups such as Baicalia, Linella, Inzeria and Jacutophyton. This assemblage occurs in the upper few meters of the McLeary Formation where a broad zonation can be recognized (Fig.15):

- lower strata contain broad, laterally linked, non-branching domes, overlain by beds containing,

- bulbous, dendroid (tungussids), conical (conophytonids) and possibly furcate (kussiellids) stromatolites, followed by,

- digitate, columnar forms (gymnosolenids) at the top of the succession.

In the following section, features of some of these groups are described with detailed discussion on two distinctive forms: Conophyton, and digitate stromatolites from the uppermost buff dolostones.

Type a

Forms belonging to this class are characterized by furcate branching of subcylindrical columns, diagnostic of the supergroup Kussiellida, and is only poorly represented in the McLeary Formation. Most examples have smooth margins, moderately developed wall structure and flat to convex laminae. They bear some resemblance to the group Boxonia Korolyuk (other members of this group usually have fringed margins and lack good walls). Nevertheless the Belcher examples tend to exhibit somewhat tuberous columns, rather than the more typical slender cylindrical columns of Boxonia (see Donaldson, 1976a, Fig.3). In some forms well developed bridges and coalescing of columns also occur. Thus, although kussiellids appear to be represented, there appears to be no direct comparison with established Riphean taxa.

meters

cf. Inzeria

{ cf. Linella, Baicalia
Conophyton cf. garganicus

cf. C. garganicuscf. Linellacf. Anabariacf. C. garganicuscf. Baicalia, Jacutophytoncf. Linellacf. BoxoniaCryptozoon

Figure 15. Stromatolite zonation in the upper 12 meters of the McLeary Formation

The synoptic relief is 1 to 2 cm. Intercolumnar sediments are mostly sand-free dololutites. Terrigenous detritus also is generally absent within the stromatolite laminae.

Comparisons: The combination of subparallel columns, divergent or subparallel branches and wall structure is shown by several illustrated examples of the groups Linella, Gymnosolen Steinmann, Katavia Krylov and Minjaria Krylov. However, most examples of Gymnosolen and Minjaria are generally characterized by more slender, erect and smooth columns, whereas Katavia is conspicuously more bumpy (eg. Hofmann, 1969b, Fig. 20). Development of niche projections is also less common in these groups. In the Belcher examples, the tuberos nature of columns and early branch stages, together with niche projections more closely resemble Linella (see especially Krylov, 1967, Plate 4, Fig. 1; Cloud and Semikhatov, 1969). Some Australian examples, however, tend to be more gnarled (Walter, 1972, Fig.46).

(ii) Baicalia-like branching of some stromatolites has been noted by Donaldson (1976a, Fig.4) whose illustration also shows the somewhat tuberos nature of the columns, with constrictions at points of branching that are typical of this group (Pl.3, Fig.5). However, some of the finer morphological features are not in keeping with this comparison. For example, laminae range in shape from slight to strongly convex, margins are smooth and distinctly enveloped; in most examples illustrated elsewhere, wall structure is patchy or absent and margins are more irregular (Semikhatov, 1962; Komar, 1966; Walter, 1972). The combination of divergent branching, wall structure and laminae

shape appears more like the group Kōtuikania Komar (Komar, 1966, Plate 10, Fig.3 & 6). However, even this comparison falls short of previously described forms which usually exhibit more active branching than that seen in the Belcher specimens. Thus, while it is clear that they are representative of the supergroup Tungussida, comparison with specific groups is difficult. Nevertheless, without more detailed examination of morphological variability and microstructure of this type of stromatolite, erection of a new group is unwarranted.

(iii) Stromatolites having affinities with the group Anabaria have already been illustrated from the Lower and Middle Members of the McLeary Formation (Pl.3 Fig.2). Here, an initially bulbous basal mass gives way to five or six subcylindrical branches with highly divergent axes. A second stage of branching may also occur. Margins are smooth, with few protruberances and wall structure is moderately developed. Laminae are convex and occasionally conical.

The only other group resembling these forms is Tungussia Semikhatov, which is usually characterized by more active branching than that seen here (typical Tungussia occur in the Mavor Formation). The diagnosis of Anabaria also includes stromatolites possessing some digitate branching (Hofmann, 1969b) and possible representatives of this type occur in the Upper McLeary. Here, up to five stages of branching occur. Columns are erect to slightly inclined, subcylindrical or turbinate, and have smooth margins occasionally interrupted by fringes. Walls are patchy. They can be distinguished from other digitate forms such as Gymnosolen or Minjaria by the absence of well developed walls, and from Inzeria or Linella by the scarcity of niche projections.

To summarize, a large proportion of the stromatolites in the Upper

McLeary have the characteristics of the supergroup Tungussida. Possible groups represented are Linella, Baicalia, Kotuikania and Anabaria, all of which supposedly have time ranges that are restricted to the Middle and Late Riphean, and in the case of Linella, the Vendian.

Type c

Distinctive buff dolostones in the uppermost McLeary Formation form an extensive biostromal complex of bioherms that are elongated in a northeast direction (Pl.3, Fig.3). Internally the bioherms consist predominantly of closely packed digitate stromatolites with subcylindrical, erect columns, but locally may be divergent (Fig.16). In plan view columns are circular to ovate, with elongations parallel to that of the parent bioherms (Pl.4, Fig.1). Branching is mostly digitate (Pl.4, Fig.2), with some columns displaying furcate or divergent patterns (Fig.16b, c). Up to five stages of branching are present in some bioherms. A typical example of the final two branching stages has been reconstructed from a block using the serial section technique described by Hofmann (1976b), and is shown in Fig.17a & b.

In the thicker basal columns the laminae are rather flat, becoming increasingly convex and in places conical in the upper branch stages. An increase in the development of wall structure also accompanies the change in laminae shape. Many laminae are crinkled and microunconformities are uncommon.

Margins are relatively smooth, having few bumps or fringes. Fine lateral linkages are also uncommon, but in some instances strong bridging occurs (Fig.16a). Where linkages have been destroyed, either by erosion or recrystallization, small fringes remain. Zones exhibiting divergent branching along with a slight tendency to anastomose are common

Figure 16. Stromatolites from the uppermost biostromal unit, McLeary Formation, illustrating the variation in branching style (a, b and c) and general orientation of columns with respect to the bioherm axis (d).

Figure 17a. Reconstruction from eight serial sections parallel to the column axes of the upper two stages of branching in stromatolites from the uppermost biostromal unit of the McLeary Formation. The style of branching, niche projections and wall structure are typical of Inzeria.

Figure 17b. Details of margin and projections of a single column of cf. Inzeria from Figure 17a above.

on the flanks of the bioherms (Fig.16b). Niche projections and branching from niches also occur. Synoptic relief was probably no more than 25 mm.

Microstructure: On polished slabs, dark macrolaminae range in thickness from 0.5 to 2 mm and alternate with thicker light coloured bands. Dark laminae can be traced up to three centimeters laterally across columns. They are typically ribboned, and some are fragmented in the central part of the column, becoming more continuous towards the margin with a tendency for laminae to coalesce, producing an envelope about 0.5 mm thick. This wall structure becomes better developed in the higher levels of the bioherms. In addition, a more irregular, lumpy fabric frequently occurs, interspersed with zones of regular banding. Thin section examination suggests that this fabric is a result of recrystallization of dolomite, the dark fine-grained remnants appearing to float in a coarser microspar (similar to Cayeux's 'structure grumeleuse'). Some of the 'tabular intraclasts' described by a few workers may in fact be a result of similar recrystallization. In places where recrystallization is not advanced, the darker macrolaminae can be resolved into microlaminae 0.1 to 0.2 mm thick, but can only be traced a few millimeters laterally.

Petrography:

The bulk of the dolostones consists of microspar mosaics with average crystal size about 50 μm and patches of coarser neomorphic spar up to 500 μm . Scattered throughout are floating relics of micritic dolomite that probably represent the original, non-neomorphosed products of early dolomitization. Calcite is rare, but pyrite is dispersed throughout. Microstylolites with relief less than two millimeters are

commonly developed parallel to the laminae and in places resemble crenulated microlaminae. Caution should thus be exercised in distinguishing between these two features, for example in thin section where dissolution surfaces and discordant relationships are best recognized.

Comparisons:

Column form and style of branching invite comparison with the group Inzeria. Specimens illustrated by Walter (1972) compare favourably, not only with the multiple branching modes, but also with the transition from naked or patchy wall structure in basal columns, to well developed walls in the upper branches. In these same examples, coalescing occurs more commonly in the final stages of branching, a feature also observed in McLeary forms. Bioherms from the Southern Urals, illustrated by Krylov (1963; 1967) are also similar.

Comparisons of microstructures with examples of Inzeria described by Krylov (1963) and Walter (1972) reveal some similarities to the Belcher forms. In particular, the dark laminae exhibit sharply defined upper surfaces and grade laterally into lighter coloured microspar zones. Australian examples also demonstrate a tendency for laminae to wedge out, producing discrete lenses, a feature that is not apparent in the Belcher forms. Furthermore, although macrolaminae boundaries tend to be crenulated in the Belcher forms (relief is no more than 0.5 mm), the Australian examples are relatively smooth. Microunconformities are less conspicuous in the Belcher types.

Group Conophyton Maslov

Conophyton displays three distinct modes of occurrence;

- in beds consisting of discrete columns,
- in a mixed assemblage with Linella and Baicalia-like forms,
- and a micro - Conophyton bed in the middle zone of the Upper McLeary.

(i) In this class, columns are 30 to 40 cm high and 10 to 15 cm across, and most are erect or gently inclined to the south (Pl.4, Fig.3). In plan view they are not only elongated as much as 3:1 but have elongation trends almost perpendicular to the regional paleocurrent trend (Pl.4, Fig.4). Even in beds containing a mixed Conophyton - tungussid stromatolite assemblage only the Conophytons display this elongation trend, whereas other forms are either circular or ovate in cross-section.

Typically, the McLeary conophytons possess well-developed crestal zones, about 1 to 2 cm wide, and where individual laminae are preserved in silicified zones, thickening of up to three times occurs. Laminae are characteristically flexed in the crestal zone. Following the classification of Komar et al. (1965) the crestal zones are predominantly type I (Fig.18). Lateral linking by single laminae is common, but extensive bridging is rare. Silicification is also common.

Laminae: Laminae are strongly conical with slopes from the crestal zone to the margin ranging from 50° to 80° . They are relatively flat except in crestal zones and at margins where they become steeply inclined, forming distinct wall structure. Synoptic relief may have been as much as 15 cm.

Microstructure: In non-silicified parts of columns, the finest recognizable macrolaminae are 200 to 300 μm thick. Even at this range of thickness, the distinction between the light and dark laminae is hampered by the high degree of recrystallization. The problem of microstructure

Conophyton Crestal Zones

1000 μm

Figure 18. Crestal zone classification (after Komar et al. 1965) and lamination structure of Conophyton from the Upper Member of the McLeary Formation. Also, a plot of light (L) versus dark (D) lamination thickness in the form Conophyton cf. garganicus n = number of measurements, $\bar{X}$ = mean thickness, S = standard deviation.

identification is alleviated to some extent in zones of chert replacement, where ribboned microlaminae as thin as 25 μm can be resolved. Crenulations and microunconformities are also preserved (Fig.18). Individual microlaminae can rarely be traced more than two centimeters before they coalesce into broader bands. In some cases coalescence of laminae is a result of carbonate recrystallization.

Thickness measurements of dark and light laminae were undertaken in some of the black cherty zones and the results are summarized in Fig.18. The frequency distribution of the two lamination types is almost identical with mean thickness values of 100 μm , and dark-to-light thickness ratios of 1:1. The values of these parameters compare favourably with values determined by Cloud and Semikhatov (1969) and Walter (1972) for some Australian stromatolites that were assigned to the group Conophyton garganicus. In these areas, C. garganicus has a stratigraphic range of Lower to Middle Riphean. Hofmann (1969b) has also compared some of the Helikian Sibley Group conophytons with this taxon. Although the Upper Riphean form C. ressoti exhibits somewhat similar laminae shape and thickness, Cloud and Semikhatov (1969) note that the dark laminae tend to be more wavy or clotted. Other Upper Precambrian representatives of this group are distinguished by much lower ratios of dark-to-light laminae than that observed in the McLeary examples.

Petrography: Original carbonate textures have been almost entirely obliterated by dolomitization and subsequent recrystallization. Even in black cherty zones, advanced dolomite replacement can impart a fragmented or lumpy texture.

A feature common to nearly all Belcher Island conophytons is the abundance of very coarsely crystalline patches of dolomite, imparting

an almost fenestrate appearance to the stromatolites. Crystals up to 1.2 mm in size are not uncommon, and some borders of these patches are outlined by either microstylolites or by black amorphous material, possibly of organic origin. This black material also occurs at crystal contacts in some microspar mosaics. Although these pseudo-fenestrae conform rather crudely to laminae surfaces, they also cut across macrolaminae. Furthermore, there is little textural evidence that would indicate relic void-filling cements, although such relics could have been destroyed by pervasive recrystallization. On the other hand it may be significant that this textural feature is largely confined to Conophyton-like stromatolites, even within a single bioherm where forms transitional from simple domes to more typical conophytons occur. The fenestrae may have originated by gas expansion which resulted in disruption of laminae.

(ii) In this category, Conophyton can take the form of discrete columns, or as a part of bioherms that also display Linella or Baicalia-like characteristics. Some of the latter types somewhat resemble the group Jacutophyton. The overall structure of laminae, crestal zones and margins appears similar to the Conophytonids described above. The isolated, non-branching varieties are comparable to Conophyton gar-ganicus.

(iii) A distinctive one-meter-thick bed of densely packed columns was encountered 45 to 59 m below the top of the formation in a number of sections (Fig.9) and can be traced at least 25 km parallel to strike. Columns are up to 25 cm in height, range from 1 to 5 cm wide, and are erect or inclined to the south (Pl.4, Fig.5). They are typically circular in cross-section and are very closely spaced, with upwards of 300 to 500 columns per square meter. Some groups of three or four

columns are arranged in strings that trend northeast (Pl.4, Fig.6). The elongations displayed by type (i) conophytos are absent in this class.

Laminae: Preservation is similar to that of previously described forms. Macrolaminae, 0.5 to 2 mm thick, are flat, non-crenulated, and have slopes ranging from 70° to 80° . Lateral linking is prominent, and where individual laminae can be traced across two or more columns, the synoptic relief is no more than 15 to 20 mm (an order of magnitude less than type (i) conophytos).

Environmental Interpretation:

The Upper Member of the McLeary Formation represents the transition from an intertidal to shallow subtidal environment; this has been established by independently utilizing sedimentary structures, and by comparing the sequence of morphologic differences in a variety of cryptalgal structures with similar differences observed in modern environments. On this basis it is inferred that the McLeary conophytos grew in intertidal to shallow subtidal conditions where wave and current action were relatively strong. Approximate limits of 1 to 2 m water depth are postulated. It is unlikely that water depths were much greater because desiccated mudstones occur sporadically in the Upper Member. A good example is on Mavor Island (Section 31) where a Conophyton bed 5.8 m below the top of the McLeary is sandwiched between two thin dololutite beds, each containing well-developed desiccation polygons.

The group Conophyton represents one of the few stromatolite forms that, over an extremely long time span (almost 1.5G.a) have retained certain characteristic morphological features, namely distinct conical laminae, distinct crestal zones, and only rare branching. These features

are present in columns ranging in size from a few centimeters to several meters high, and are inferred to have grown in environments ranging from relatively high-energy intertidal (this study) to low-energy subtidal (Donaldson, 1976b).

Recently, Golubic and Focke (1978) presented convincing evidence that certain monospecific Cyanophyte populations can produce similar micro- and macrostructures (in this case small domal stromatolites) under a wide range of growing conditions, for example turbulent subtidal to sheltered intertidal environments. The view that Conophyton may be biologically controlled is not new. Forty years ago, Maslov (1939, in Hofmann, 1969) suggested that Conophyton may be one of the few true 'genera'. That this view should be taken more seriously was also reiterated by Hofmann (op. cit.).

As indicated in the discussion on paleocurrents Conophyton elongation azimuths are almost at right angles to the regional trend. Furthermore, little or no elongation is seen in many of the columnar branching stromatolites associated with beds containing Conophyton. There is no evidence for any local abnormal environmental conditions, such as major changes in wave or current directions, that might explain this phenomenon. It is suggested that biologic rather than environmental control was responsible for both the gross morphology and microstructure of Conophyton garganicus (var. Belcher Islands).

DISCUSSION

Various studies over the last two decades have proven the value of stromatolites for intrabasin correlation, and in Russia especially, they have also been utilized as biostratigraphic markers; the Riphean

was subdivided on the basis of distinctive stromatolite assemblages (Keller et al. 1960, in Walter, 1972). The potential of stromatolites for worldwide correlation was also realized, and more recently this thesis was tested by Walter (1972) who successfully compared a number of Siberian and Australian forms.

The view that stromatolite taxa are temporally restricted is based largely on empirical observations, mostly in rock sequences of Late Precambrian age. An important question thus arises: to what extent has Cyanophyte evolution affected changes in gross morphology and microstructure. A case for biologic control of microstructure has been presented in a number of studies on modern algae (eg. Gebelein, 1974). Thus, any secular biological changes in the morphology of algae should be reflected in the stromatolite microstructure.

One of the most comprehensive studies of microstructural changes within a single taxon, the group Conophyton, is that by Komar et al. (1965). Subdivision of this group is mainly based on the structure of the crestal zone, degree of thickening of laminae, and the microscopic nature of laminae in stromatolites of Early to Late Riphean age. In the present investigation, the form C. garganicus has been recognized, thus extending its time range into the Aphebian. Similar studies of other Aphebian forms are lacking, but the potential for worldwide correlation would be considerably enhanced if evolutionary trends could be convincingly established.

Unfortunately identification of specific microstructures on a millimeter and submillimeter scale is seriously hampered by recrystallization. This has the effect of obliterating original biofabrics, producing an apparent thickening of laminae and at times resulting in

development of lumpy or fragmented fabrics from originally continuous laminae. Thus, subtle differences in microstructure that might be diagnostic of evolutionary changes in the algae may be completely destroyed. These problems are alleviated to some extent by silicification. In most cases this is a very early, almost syngedimentary, diagenetic process that results in the preservation of fine structural details. Although silicification is ubiquitous in the McLeary cyanophytes, it tends to be patchy and interlayered with dolomite microspar and pseudospar. In the present study, however, the spread of thickness values of light and dark laminae is relatively slight and compares favourably with published frequency distributions for C. gargaricus.

It has been reasonably well established that environmental parameters such as wave energy, current energy and sedimentation rates affect the gross morphology of stromatolites to a considerable extent (Hoffman, 1967; Logan et al. 1974; Horodyski, 1977a, b); the extent to which biologic controls affect such differences is not well known. An exception to this is the recent study of Golubic and Focke (1978) who have documented morphological similarities (micro- and macrostructures) in monospecific cyanophyte populations that form domal stromatolites within a wide range of environmental conditions. Awramik (1971) suggested that morphological diversity, particularly in columnar stromatolites, has increased from the pre- to late Riphean, but this is not supported by the present study, where Archean stromatolites representing all four supergroups can be recognized (see also Donaldson, 1976a). This suggestion also carries with it the implication of convergent evolution based on gross morphology. Although this concept is not dismissed, it is more likely to be manifested in stromatolite microfabrics rather than gross morphology.

Comparison of the Belcher Island stromatolites with established groups is based on a wide range of morphological features including column arrangement, ornamentation, style of branching and microscopic details of laminae. As in most biohermal complexes, considerable variation in any one of these features can occur. Thus, as Walter (1972) has noted, complete bioherms are used for taxonomic comparisons. In all, seven groups can be recognized: Linella, Baicalia, Anabaria, Boxonia, Inzeria, Jacutophyton and Conophyton garganicus. All of these groups except the Conophytonids were previously suspected of having time ranges restricted to the Riphean or Vendian. In the present study detailed analysis has been confined to the two groups, Conophyton and Inzeria. Comparison of the microstructure of Inzeria is referred to in the previous section. It would be interesting to speculate that the observed differences are a direct result of specific changes in the stromatolite biota. However, they could equally be due to subtle changes within or between populations of Inzeria, or in the degree of recrystallization and preservation. It is also worth noting that this microstructure is not unlike some illustrated examples of Baicalia and Gymnosolen. In other words, similar microstructures may be present in a number of stromatolites that in terms of gross morphology are quite different. To test the hypothesis that these differences are biologically controlled, it would be useful to examine specimens of stromatolites intermediate in age to those of the Belcher Islands and the numerous studied occurrences from Upper Precambrian strata.

PLATE 1

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PLATE 2

Figure 1. Large desiccation cracks superimposed on symmetric oscillation ripples. Pen (left) is 15 cm long. Upper Member, McLeary Formation, Tukarak Island.

Figure 2. A variety of cryptalgal mats in cherty dolostones. Note the vertical transition from parallel laminates at the base, to tufted and wavy mats in the middle, and digitate-fenestrate mats at the top. Scale is 15 cm long. Located 40 m from the top of the McLeary Formation, Mavor Island (Section 33).

Figure 3. Vertical transition from wavy and fenestrate laminates at the base, to distinct digitate mats with synoptic relief less than 2 mm. Scale is 15 cm long. Upper Member, McLeary Formation, southwest shore of Eskimo Harbour, Flaherty Island (Section 21).

Figure 4. Algal pisolites (oncolites) stabilized by overlying digitate mats. Note tufted mats below main pisolite bed. Width of view is 45 cm. Upper Member, McLeary Formation, Mavor Island (Section 31).

PLATE 3

- Figure 1. Silicified domal stromatolites exhumed on a bedding surface. Note elongation of some domes in the down-dip direction. Hammer for scale (right center). Upper Member, McLeary Formation, north Tukarak Island (Section 27).
- Figure 2. Isolated stromatolite bioherms with strongly divergent columns in the Middle Zone, Upper Member, McLeary Formation. In some columns conical laminae (arrow) are transitional between a basal bulbous mass, and simple convex laminae in the upper parts of the columns. Bioherm is 35 cm high. Mavor Island, Section 31.
- Figure 3. Large elongate stromatolite mounds at the top of the McLeary Formation, trending northeast. Tukarak Formation in the foreground. North shore of the southeast arm of Mavor Island, facing east (grid reference 358179).
- Figure 4. Stromatolites showing dendroid branching of Linella. Note branching from niches (arrows), and a prominent discontinuity in laminae growth (a) that separates the initial bulbous mass from later cylindrical columns. Stylolites are abundant in these upper beds. Upper Zone of Upper Member, McLeary Formation, Tukarak Island.
- Figure 5. Stromatolite displaying divergent branching typical of Baicalia. Conical laminae occur in some columns and points of branching show constrictions. Pen is 15 cm long. Upper Member, McLeary Formation, Mavor Island (Section 31).

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PLATE 4

- Figure 1. Bedding surface showing closely packed columns of Inzeria-like stromatolites, uppermost McLeary Formation, southeast shore of Eskimo Harbour (Section 22).
- Figure 2. Vertical section through digitate stromatolites in the uppermost buff zone, McLeary Formation. Note well developed wall structure and niche projections, cf. Inzeria. High relief stylolites are prominent in this unit. Pen is 15 cm long. North Tukarak Island, Section 28.
- Figure 3. Discrete and laterally linked columns of Conophyton cf. garganicus. The positive relief of conical laminae is a result of selective preservation by black and white chert. Pen is 15 cm long. Upper Member, McLeary Formation, west Tukarak Island (Section 1).
- Figure 4. Bedding exposure of Conophyton columns showing marked elongation. Upper Member, McLeary Formation, west Tukarak Island (Section 1).
- Figure 5. Vertical section of closely packed Conophyton columns, locally inclined, demonstrating the transition from parallel and wavy mats near the base with a few conical initiation zones, to distinct Conophyton columns at the top. Middle Zone of the Upper Member, McLeary Formation, north Tukarak Island (Section 27). Pen is 15 cm long.
- Figure 6. Bedding surface of Figure 5. Differential weathering has removed non-silicified intercolumn dololutes. Some 'strings' of columns are surrounded by a laminated envelope. Pen is 15 cm long.

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CHAPTER 2

A MIDDLE PRECAMBRIAN TRANSGRESSIVE CARBONATE
SEQUENCE: PART II. THE TUKARAK TO LADDIE
FORMATIONS

INTRODUCTION

This chapter is divided into two sections. The first describes and subsequently interprets each of the four remaining formations that compose the transgressive carbonate sequence (Tukarak, Mavor, Costello and Laddie Formations). The second section discusses the overall vertical and lateral facies relationships of these four units, together with the Fairweather and McLeary Formations (from Chapter 1), and formulates a model of deposition to account for this sequence.

All four units (Tukarak to Laddie Formations) are lithologically distinctive, and boundaries between adjacent units are easily mapped. In addition, marker beds in the Tukarak Formation and Mavor Formation proved to be useful for establishing stratigraphic position in these formations, especially in areas where facies changes occur.

TUKARAK FORMATION

The Tukarak Formation (Dimroth et al. 1970, p. 102) is an excellent marker unit in the Belcher Group, and is named after Tukarak Island where it is well exposed. Good exposures also occur on Mavor Island, and also east of the Old Hudson Bay Post on west Tukarak Island, which is suggested here as a reference section. The average thickness at these localities is 60 m, but is 70 m at the type section (west of McLeary Point) and 52 m on Mavor Island. West of Laddie Harbour the formation appears as thick as 93 m, although this could be due to tectonic thickening by displacement along bedding-plane faults (Section 35). Verification of this is difficult because crucial contacts are covered.

There is a general thinning trend along the Kasegalik anticline, where thickness ranges from 40 to 52 m (Sections 19, 21, 22). At these

localities the top of the Tukarak Formation is taken at the base of the first pale buff stromatolitic dolostone. A reliable marker within the Tukarak is a crinkly laminated dolostone bed up to two meters thick, that has been traced over most of the Belcher Islands. In all sections this bed is 14 to 16 m below the Tukarak-Mavor contact.

Contacts with the underlying McLeary Formation and overlying Mavor Formation are conformable and relatively sharp, grading over a meter or less. Differential glacial scouring and unconsolidated cover have obscured much of the lower part of the Tukarak Formation and the basal contact is poorly exposed. This zone appears to be a common site of diabase sill intrusion, particularly along the western side of Tukarak Island. Jackson (1960) has subdivided this unit into two members (Fig.19).

LOWER MEMBER:

Two representative measured sections are illustrated in Fig.19. Fine sandstones, dolomitic sandstones, grey shaly mudstones and some orange-buff weathered stromatolitic dolostones typify this member. Sandstone lithologies contain a variety of cross-beds including herringbone, planar and trough cross-beds and a few reactivation surfaces. Flute and groove casts at the base of one bed exhibit a strong northwest trend (Fig.20). Shaly mudstones are exposed in only a few sections because of their susceptibility to glacial erosion. Starved ripples are common and small vein-like sandstone dykes also occur. No evidence of desiccation was observed in the mudstones.

Isolated stromatolite mounds 1 to 1.5 m across and up to 50 cm thick were observed approximately 3.2 km southwest of Section 33. As is the case for most stromatolitic units in this formation, considerable recrystallization of dolomite and calcite has all but obliterated the

Figure 19. Tukarak Formation; Section 1 from west Tukarak Island, Section 22 from east side of Eskimo Harbour. Crinkly dolomite bed in yellow.

Figure 20 Lateral facies changes in the Tukarak Formation. Stromatolite bioherm elongations are taken from the Upper Member. J.A.D.= sections measured by J.A.Donaldson.

finer structural details of individual columnar stromatolites composing these bioherms. At this locality banded red and buff dololutites and mudstones are draped over the bioherms. Stone rosettes of lutite slabs up to 15 cm across are also common.

UPPER MEMBER

The Upper Member of the Tukarak Formation constitutes one of the most distinctive lithological units in the Belcher Islands. On Tukarak and Mavor Islands it is composed of thin wavy and lenticular bedded ferruginous dolomitic mudstones and dololutites which here are collectively termed 'ribbon rock'. The reddish brown mudstones are composed mainly of quartz, feldspars and abundant micas, although these terrigenous components are largely obscured by microcrystalline hematite and dolomite microspar. The buff dololutites, on the other hand, contain almost no terrigenous admixture. Ribboning results from differential weathering of less resistant lutite beds producing markedly corrugated exposure surfaces (Pl.5, Fig.1). The wavy and lenticular patterns are a result of primary structural differences such as oscillation and current ripple bedding and wavy laminae. Some of the iron-rich mudstones are intimately interlayered with less ferruginous dolostones on a scale of one to two millimeters. In some places mud drapes on the stoss slopes of ripples suggest changes in currents either in direction or velocity, that resulted in a change from bedload transport to deposition by suspension.

Small vermiform dewatering structures are common, and some wedge-shaped varieties resemble desiccation cracks. Stone rosettes are abundant in the lower part of this member.

Towards the top of the Tukarak Formation a number of pale buff

dolostone beds, some coarsely crystalline, are traceable for several tens of kilometers along strike (Figs.19 & 20). Internal structures are generally vague although the outlines of small columnar stromatolites can be recognized in some beds. Good bedding exposures on Mavor Island demonstrate the true 'mound' nature of these structures. Dimensions are similar to those observed in the Lower Member. Mounds are closely spaced (but not linked) and have elongations of 2:1 to 3:1. Elongation trends are highly consistent and show azimuths identical to groove casts from Section 1; they also are similar to stromatolite elongations in the McLeary Formation (Fig.20). Such consistency from station to station strongly suggests a primary origin for these structures, ie. algal bioherms. Synoptic relief was no more than 20 cm. Bedding in the ribbon rock sharply abuts bioherm margins or is draped over the upper surface (Pl.5, Fig.2). Ribbons of mudstone within the bioherm represent minor interruptions in growth resulting from periodic influxes of sediment.

The bioherms are composed almost entirely of dolomite. A few honeycomb-weathered calcitic patches may be a product of dedolomitization.

An additional highly distinctive marker bed that has been traced over most of the Belcher Islands consists of crinkly and wavy laminated dolostone. The bed is consistently 14 to 16 meters below the Mavor Tukarak contact. Subparallel laminae are wavy, crinkled and in places display scalloped structures similar to modern examples of tufted algal mats. Cross-bedding is absent. This crinkly laminated bed is composed entirely of dolomite that has a typical grumeleuse texture, where relics of fine micritic dolomite appear to float in a coarse recrystallized microspar; the transition from micrite to microspar is completely gradational. Stylolites having relief up to two centimeters are also common.

An important facies change observed in Kasegalik anticline sections is the scarcity of ribbon rock lithologies and thus correlation is based more on the presence of the crinkly laminated dolostone bed which, like its eastern counterparts, is sandwiched between dolostones containing some large cryptalgal mounds. Here, the internal structure of the mounds is obscured by dolomite recrystallization and in places dissolution, which produced a vuggy texture infilled with dolomite crystals up to one centimeter across. Locally the margins are brecciated, with coarse spar filling some of the large voids, the breccias bearing some resemblance to solution collapse structures. Intermound sediments show little evidence of advanced recrystallization or brecciation. Nevertheless, like the mounds on Tukarak and Mavor Islands, they are highly elongate; it is unlikely that uniform elongation trends over such a wide area could result from solution collapse, and hence these too are considered to be of biogenic origin.

INTERPRETATION

A shallow-water depositional environment can be inferred from most of the bedforms in strata of the Lower Member, and in particular, herringbone cross-beds, lenticular bedding and reactivation surfaces suggest tidal current asymmetry. Particularly good evidence for shallow-water conditions are pavements of stone rosettes which indicate depths well within wave-base, probably in a swash-backwash zone.

Bioherm elongation trends are similar to those in the McLeary Formation, paralleling major current directions, and are consistent with a west- to southwest-dipping paleoslope. Like bioherms in the McLeary Formation, they are inferred to have formed normal to the ancient shoreline in response to prevailing wave and current directions.

Evidence for prolonged subaerial exposure, such as desiccation cracks, is lacking in the Tukarak Formation and the strata probably represent water depths slightly greater than those of the Upper McLeary Formation. Cryptalgal mound structures show a general increase in synoptic relief towards the top of the formation that probably correspond to increased water depths, although remaining as isolated bioherms (in contrast, Mavor Formation structures constitute an extensive laterally linked complex). The Tukarak Formation can be thought of as predominantly subtidal, with deposition taking place along a broad shallow platform, but well within the zone of effective wave-base (Fig.21).

MAVOR FORMATION

INTRODUCTION

The Mavor Formation takes its name from Mavor Island where Jackson (1960) defined a reference section to supplement the type section (3.2 km southwest of Laddie Harbour). It is a distinctive lithological unit consisting predominantly of coarse crystalline, pale buff stromatolitic dolostones and thin shaly dololutites. A summary of sections is shown in Fig.22. Considerable variations in thickness accompany some interesting facies changes in this unit. The thickest sections occur west of Laddie Harbour (Section 35 = 245 m), thinning southward on Mavor Island to about 190 m and to 153 m on the west side of Tukarak Island. However, a slight thickening is observed in sections along the Kasegalik anticline (Section 22 = 185 m).

Contact with the overlying Costello Formation is poorly exposed, but appears to be gradational over 1 to 2 m. On the west side of Tukarak Island this contact is readily identified by the appearance of black

Figure 21. Paleoenvironmental reconstruction of the Tukarak Formation, uppermost McLeary Formation and basal Mavor Formation. Water depths were probably above effective wave-base.

Figure 22. Stratigraphy of the Mavor Formation:
 Section 35 = east Tukarak Island, 1 & 3 =
 west Tukarak Island, and 22 = Eskimo Harbour.
 Varicoloured unit in green.

shales. The Tukarak-Mavor contact is marked in all places by the sudden appearance of buff stromatolitic dolostone beds up to 25 m thick (Fig.22). This latter unit can be traced throughout the Belcher Islands.

Three principal lithologies occur:

- (i) stromatolitic dolostones;
- (ii) a varicoloured dolostone bed, and
- (iii) shaly dolostones.

STROMATOLITIC DOLOSTONES

Pale buff stromatolitic dolostones are prominent in the lower half of the Mavor Formation. In most sections on Tukarak Island, three main units 12 to 25 m thick can be recognized, with thinner stromatolite-bearing horizons occurring higher in the sequence. However, an exceptionally thick unit west of Laddie Harbour (146 m in Section 35) is almost continuously exposed over the entire length of eastern Tukarak Island and it is in these sections that the nature of what must have been a remarkably extensive biostromal complex becomes apparent.

In steeply dipping beds, particularly those on west Tukarak Island, exposures at high angles to bedding exhibit broad symmetrical mound structures with amplitudes of 30 cm and more and lateral spacing of up to 10 m. On eastern Tukarak Island, dips of 5° to 25° have resulted in good bedding exposures where the elongate nature of these mounds is readily apparent (Pl.5, Fig.3). Sections west of Laddie Harbour also demonstrate an increase in mound amplitude from 30 cm at the base to a maximum of 200 cm at the top of this unit, with a concomitant increase in mound spacing from 2 to 20 m. Aerial photographs indicate that the mounds commonly coalesce 'up section' into single larger mounds. Inheritance of individual mounds reaches 100 m, several times thicker than similar mounds cited by Eriksson (1977) from the Transvaal Supergroup.

Internal Organization of Mounds:

Two varieties of coarse dolostones occur:

(i) dolostones consist mostly of subcylindrical, columnar stromatolites with digitate branching and in places some excellent examples of highly divergent branched forms identical to the Riphean taxon Tungussia. In most dolostones recrystallization has all but obliterated primary cryptalgal fabrics. In general columns are 10 to 20 cm high and have subcircular cross-sections no more than five centimeters in diameter. Laminae appear to be convex. Some of these stromatolites display elongations as much as 10:1 at the crests of mounds, gradually becoming subcircular along the flanks. Intercolumn fill is identical in composition and degree of recrystallization to the stromatolites. Where exposures permit, individual beds can be traced to adjacent mounds and thus mound amplitude can be translated into synoptic relief. A typical mound from the top of this unit is illustrated in Pl.5, Fig.4.

(ii) wavy and crinkly laminated dolostones are interbedded with stromatolitic dolostones (Pl.5, Fig.5). In the lower parts of these units bedding is simple, conforming to the overall mound shape. However, an increase in mound size towards the top is accompanied by the appearance of smaller second-order mounds giving rise to thick laminated dolostones (Pl.5, Fig.6; Fig.23). Elongation directions of these smaller superimposed structures are identical to the larger mounds (ie. east to east-northeast). Towards the top of this unit the large mounds grade rapidly into parallel laminated dolostones, but although the first-order mounds disappear, the second-order structures remain for several meters of section. Ultimately these too die out and eventually coalesce into wavy laminated dolostones (Fig.23).

Figure 23. Vertical changes in bioherm morphology, amplitude and internal structure shown by the main biostrome unit (146m thick), Mavor Formation, east Tukarak Island. Carbonate mud drapes and second-order mounds are more prominent in the upper levels of this complex. The stratigraphic positions of sections a-d are shown in Figure 22.

Parallel laminated and shaly dololutites are interbedded with the two previously described lithologies and in many places are clearly draped over the stromatolitic dolostones; shaly beds thin or pinch out over mound crests and thicken in intermound troughs. Slight warping of laminae in some beds may be due to soft-sediment slumping of muds down the flanks of mounds, especially where there is no apparent updoming of underlying beds. Cross-bedding and desiccation structures appear to be absent.

Petrology:

Stromatolitic and wavy laminated dolostones consist entirely of coarse dolomite microspar with a few quartz grains and floating relics of dolomicrite. Minor authigenic feldspar also occurs. Fine cryptalgal fabrics such as primary laminae are poorly preserved at the microscopic level. Silicification of stromatolites is markedly less prominent than in the McLeary Formation.

The shaly dololutites are composed of much finer microspar mosaic and in contrast to the stromatolitic dolostones their laminae are better preserved. In addition, some laminae contain more than 10% silt-sized quartz, feldspar and rarely micas.

Stylolites are a prominent feature in most beds. Local stylolite relief reaches 2.5 cm, indicating a minimum reduction in thickness of 16-33% for beds 15 to 30 cm thick that are bounded by stylolite seams.

Elongation Trends:

Aerial photographs of eastern Tukarak Island reveal a remarkable regularity in elongation trends of the Mavor mounds. Measurements of elongations on west Tukarak Island indicate similar trends. In both cases, second-order structures mimic these trends and overall there is no more than 10^0 deviation from the regional east-northeast trend.

Origin of the Mavor Mounds:

Three possible origins for these mounds are explored herein: tectonic, soft-sediment and biogenic origins.

(i) A number of features are inconsistent with a tectonic origin. These include:

- bifurcation of mounds has no known structural analogue;
- mound elongation trends do not appreciably change around the noses of doubly plunging anticlines, and trends are the same on both limbs of the Tukarak anticline;
- rocks of equal or less competence above and below the Mavor Formation exhibit no deformation except for local soft-sediment deformation;
- cleavage in the shaly dololutites is associated with the Tukarak anticline and is almost at right angles to trends of mound elongation;
- there is no evidence of flexural slip. In addition, thinning of some beds over mound crests is also incompatible with a flexural flow mechanism.

An interesting structural problem occurs in north Tukarak Island, southeast of Salty Bill Hill, where east-dipping stromatolitic beds of the Mavor Formation have been dragged along a major northeast-trending fault without subsequent rotation of the large mounds. Like most major fault systems in the Belcher Islands, actual shear planes are poorly exposed, but in general they appear to be steeply dipping ($>70^{\circ}$). Where exposed, slickensides plunge 55° to 70° , indicating a large component of vertical movement. Such displacement would not result in any appreciable rotation of pre-existing structures. Thus

down dragging of east-dipping beds and subsequent erosion could result in beds striking almost parallel to mound elongation trends, the individual mounds having undergone very little change.

(ii) Deformation by pre-consolidation slumping can also be discounted. In many sedimentary basins that contain examples of soft-sediment slumping, fold axis trends exhibit considerable variation in response to local differences in paleoslope, although slump vectors generally indicate down-slope movement. However the Mavor mounds display remarkably consistent elongation trends over a wide area and these trends are parallel to the paleoslope. The regularity of coal-escence patterns of mounds over such a wide area is also less amenable to a slump mechanism.

(iii) A biogenic origin seems most likely as previously suggested by Barrett et al. (1976), and the mounds can thus be thought of as large elongate bioherms that form part of a very extensive algal biostrome complex. Synoptic reliefs of up to two meters, remarkably high inheritance, a lack of erosional structures and the scarcity of cross-bedding all suggest that accumulation took place below wave base. At one locality, edgewise conglomerates occur in flat laminated dolostones atop the major biostrome complex, indicating that some parts of the complex may have extended above the zone of effective wave base. Nonetheless it is unlikely that the bioherms were ever totally emergent. Interruption of growth of stromatolites and algal mats by influxes of carbonate mud was intermittent as indicated by shaly dololutites that drape over the bioherms. As previously noted, bioherm elongations are approximately normal to the paleoshoreline (inferred from independent criteria in Chapter 1). Bioherm elongation may therefore have developed in response to offshore tidal currents.

No modern analogues have yet been found to compare with the remarkable size and areal extent of these middle Precambrian algal biostromes. The few ancient examples that have been reported are mostly Precambrian, such as the bioherms in the Transvaal Malmani Dolomite (Eriksson, 1977), and elongate bioherms containing digitate stromatolites and laterally linked domes from a Precambrian succession in Wyoming (Knight, 1968). Coalescing mounds up to 10 m high that possess high inheritance and synoptic relief (up to 2 m) have also been documented from the Aphebian Goulburn Group, Northwest Territories (Cecile and Campbell, 1978). This apparent restriction to successions of Precambrian age may be related to the absence of burrowing and grazing organisms which, as suggested by Garrett (1970), were responsible for the decline of stromatolites during the Phanerozoic.

VARICOLOURED DOLOSTONE BED

So named because of a mixture of buffs, pale greens, mauves and greys, this bed can be traced over most of the Belcher Islands and provides a useful marker for correlating specific stromatolite units within the Mavor Formation. It ranges in thickness from 12 to 25 m on Tukarak Island, but along the Kasegalik anticline it is less than 6 m thick. Like the stromatolitic units it is characterized by coarse dolomite microspar and pseudospar with a few patches of relic micrite. Cryptalgal structures are lacking, but slump folds, bed pull-aparts and brecciated zones abound. The lower meter or so is commonly laminated although this may pass laterally into disrupted zones. Slump folds may be open or tight, upright, recumbent or detached and rotated, and consist of dolomite laminae a few millimeters thick. The plastic nature of deformation is recorded in several complex folds, and some

resemble dewatering structures where beds have been squeezed and reduced almost to wisps. Contact with the underlying stromatolitic dolostones is sharp and locally erosional. A few small-scale cross-beds may also be present near the contact. The upper contact appears to be gradational into stromatolitic dolostones.

Soft-sediment slumping and disruption of bedding is pervasive throughout this unit, and it is tempting to suggest an allochthonous origin. Slumping may have been initiated in layered sediments that accumulated on the seaward-sloping margins of bioherms. Along some parts of Eskimo Harbour, a continuous sequence from stromatolitic dolostones to laminated and finally slumped dolostones can be observed. Petrographic evidence offers some support for this, namely a mineralogy identical to that of adjacent stromatolitic beds where there is almost a complete lack of terrigenous detritus. The presence of relic dolomite also suggests that both sediments were originally carbonate muds.

Two objections to this interpretation are offered:

- whereas the slump horizon is sandwiched between two stromatolite complexes, there is little evidence that debris from these units was incorporated in the slump units. However, deformation and later recrystallization may have obliterated such evidence;

- similar slump deposits do not occur above larger biostromes higher in the Mavor Formation. Two explanations are offered: (i) a single phase of slumping took place being triggered by an event such as an earthquake. This being the case, the varicoloured unit would represent a single time plane, whereas most other units are considered to be diachronous; (ii) the second and more likely explanation is that the

biostrome complex at the base of the Mavor Formation accumulated within effective wave base and was subjected to periodic storms (evidence for storm activity has been documented from the shallow water deposits of the McLeary and Fairweather Formations). By considering these units as being diachronous and as laterally associated facies, biostromes higher in the Mavor succession might then represent deeper water structures that accumulated below wavebase, but within the photic zone.

SHALY DOLOSTONES

The upper 50 to 60 m of the Mavor Formation is made up predominantly of shaly dolostones and laminated dololutites containing a few current ripples and small convolutions and load casts. Chert nodules occur locally. A few thin stromatolitic dolostones are also present, some less than a meter thick. Fine dolomite microspar predominates and silt-sized quartz and feldspar make up 10% of some laminae. There is a sharp contrast in the degree of recrystallization of these rocks compared to the stromatolitic dolostones.

FACIES VARIATIONS WITHIN THE MAVOR FORMATION

In addition to thickness changes, some interesting facies changes occur at the north end of Tukarak Island and along the Kasegalik anticline (Fig.24).

(i) Sandy dolarenites of obvious shallow water origin immediately overlie the thick biostrome complex at the north end of Tukarak Island (Figs.24 & 25). These, the only sandy lithologies in the Mavor, contain mostly low-angle planar stratification with some ripples and trough cross-beds. Pisolites, vertically stacked lutite slabs and mud chips also occur. Some cryptalgal structures not normally encountered in Mavor strata include:

Figure 24. Lateral variations in Mavor Formation facies. Varicoloured unit in green

-hemispherical mounds, about 1.4 m high, that consist of an outer envelope of irregular and wavy grey lutites with small digitate columns scattered throughout, and a core of grey clotted lutites with patches of coarse dolomite spar (Text Fig.25 and Pl.6, Fig.1). The structures lack well defined internal organization more typical of stromatolites and are very similar to thrombolite domes described by Aitken (1967) and Ahr (1971) from subtidal and intertidal sequences;

-the thrombolite zone is overlain by a two meter-thick bed containing columnar and digitate branching stromatolites similar to many Upper McLeary forms.

A return to more typical Mavor stromatolites occurs at the top of this zone.

(ii) A slight thickening of the Mavor Formation occurs in the Eskimo Harbour area (Figs.22 & 24) and is mainly due to a thick section of alternating brown, green and brick-red banded dolostones that superficially resemble the Tukarak Formation ribbon rock. Dolostones and argillaceous dolostones are finely laminated and commonly contain load or convolution structures, but rarely cross-bedding. Irregular calcitic patches are abundant near the base, but decrease upwards. In most places where calcite occurs, primary bedding is obscured and a honeycomb-weathering is characteristic. The calcite probably is a product of dedolomitization.

Sixty meters below the Mavor-Costello contact, a marked lithological change takes place with the appearance of dololutites containing current ripples, lenticular bedding and numerous lenses of stone rosettes. Bedding exposures show that the stone rosettes formed extensive pavements rather than isolated clusters. Higher yet, the dolostones give way to dolomitic limestone - a lithology that is uncommon in the Belcher Group.

Some fine examples of molar tooth structures can be seen in the Eskimo Harbour area (Pl.6, Fig.2). Bedding surface expression usually takes the form of small vermiform mud-filled cracks lacking the triple junctions that are characteristic of most subaerial desiccation cracks. The molar tooth structures probably formed as a result of subaqueous dehydration (syneresis) accompanying salinity changes at the sediment-water interface in a shallow subtidal environment. Such an origin has been suggested for molar tooth structures in general (Smith, 1968; Yeo, 1974).

Large domal stromatolites peculiar to the Mavor Formation occur near the top of these sections. The domes attain widths of 2.5 m. Broadly turbinate columns having furcate-style branching occur between and atop these domes. Both domes and columns have naked walls and fringed margins. This is one of the few stromatolitic beds in which coarse clastic intercolumn debris occurs (lutite chips and slabs). Sandy dolarenite-filled channels locally cut these beds (Fig.26). Zones containing angular beachrock blocks that can be traced laterally into unbrecciated dolarenites are analogous to many of the examples of beachrock in the McLeary Formation.

Occurrences of beachrock and stone rosettes provide positive evidence for at least local intertidal conditions in the western sections of the Mavor Formation.

Summary of the Mavor Formation:

Deposition and accumulation of extensive biostromal complexes that compose a large part of the Mavor Formation largely took place in a subtidal environment that probably extended from wave base into deeper waters. However, shallow-water and intertidal deposits do occur in

Figure 25. Shallow-water dolarenites and thrombolites, Mavor Formation, northeast Tukarak Island, overlain by more typical Mavor stromatolites.

Figure 26. Large domal stromatolites from the Mavor Formation, west Eskimo Harbour, cut by channel-fill dolarenites and overlain by more typical Mavor bioherms containing stromatolites with distinctive Tungussoid-style branching.

north Tukarak Island and in parts of the Kasegalik anticline as indicated by the presence of beachrock and stone rosette pavements. Elongation of the large bioherms is consistent with a northwest-striking paleo-coastline and may be in response to east-directed offshore tidal currents. Slump structures are particularly abundant in a laterally extensive varicoloured dolostone bed near the base of the Mavor. It is tentatively suggested that they represent synsedimentary deformation of layered muds that accumulated on the seaward margin of large bioherms, within or at wave base, the slumping possibly being triggered by storms.

COSTELLO FORMATION

This formation takes its name from Costello Lake, Tukarak Island. The type section, about 2.8 km southwest of Laddie Harbour, contains some 370 m of thinly bedded dololutites and shales, thinning to about 240 m on west Tukarak Island. Dimroth et al. (1970, p.103) estimated that about 310 m occur on Mavor Island, but detailed measurements in the only complete section of Costello strata (33) indicates little more than 200 m. The discrepancy may be due to rapidly decreasing dips in the trough of a small south-plunging syncline.

Contact with the Mavor formation is marked by a distinctive black shale that is well exposed along west Tukarak Island (Fig.27). Locally, this also has been the site of intrusion of a diabase sill, and slates containing relic andalusite porphyroblasts (now completely retrograde) occur within a metamorphic aureole up to 5 m from the diabase margin. Approximately 3 km south of the old Hudson Bay outpost on west Tukarak Island, contact metamorphism of thinly bedded carbonates and shales has formed an almost flawless, grey-green banded soapstone which is mined by the Inuit of Sanikiluaq.

Figure 27. Stratigraphy of the Costello Formation: Sections 1 & 3 from west Tukarak Island, and Section 35 from east Tukarak Island west of Laddie Harbour. Current ripple trends are summarized in the rose diagrams.

Contact with the overlying Laddie Formation is gradational over several meters.

Apart from black shales, the Costello consists predominantly of thin rhythmically bedded dololutites and calcilutites, some with shaly partings in the lower strata (Pl.6, Fig.3). Lower beds up to 15 cm thick are laminated and some are graded. Lunate and straight-crested ripples are common and indicate west-flowing currents (Fig.27). Chert pods are scattered throughout.

The upper part of the Costello Formation consists of intercalated white, grey, red-brown and buff beds. The white and grey lutites are highly calcitic, whereas the buff-white intervals consist of alternating beds of dololutite and calcilutite. Red-brown zones are mainly ferruginous dololutites and argillaceous dolostones. The overall colour zonation of the formation is striking. Differential dolomitization and subsequent weathering of less resistant calcitic beds has resulted in corrugated exposure surfaces. White-grey intervals in particular are characterized by lenticular and lumpy textures due to replacement of wavy and rippled calcilutites (Pl.6, Fig.4). As in many other dolostone units in the Belcher Group, the dolomite is fine grained and thus preserves many primary structures.

A zone about 20 m thick containing a number of thin graded calcarenites, cross-bedded limestones with climbing ripples, intraformational conglomerates and slump folds occurs 120 m above the Mavor-Costello contact and has been traced throughout Tukarak, Mavor and Innetalling Islands. Graded beds contain dololutite grains and dolomitized ooids in a matrix-supported framework which has largely been replaced by calcite cement. Clasts were probably derived from shallow-water deposits,

possibly from McLeary and Fairweather deposits where there is abundant evidence of penecontemporaneous dolomitization having been widespread over a broad supratidal platform. In some places, limestone beds 10 to 20 cm thick that contain climbing ripples overlie and are transitional into graded beds (Pl.6, Fig.4) and thus correspond to typical A, B and C divisions of the Bouma cycle. These beds appear very similar to Meischner's (1964) allodapic limestones (except for the lack of bioclastic debris in the present examples), and thus sediments of shallow-water origin were probably introduced into deeper parts of the basin by periodic turbidity flows. The overlying lutites mark the return to normal 'hemipelagic' deposition from suspension with minor reworking by bottom currents.

Spectacular slump folds occur west of Laddie Harbour in packets that are bound above and below by completely undeformed rocks (Pl.6, Fig.5). In most cases mass transport was to the west or southwest. Unlike slump deposits in the Mavor Formation, these structures show little evidence of liquefaction and remobilization, although fold detachment and rotation did occur. Immediately above the slump packets are pebble-filled channels up to 8 m across (Pl.6, Fig.6). Pebble types are typical of Costello lithologies.

LADDIE FORMATION

A change in the definition of the Rowatt-Laddie Formation boundary leads to a reduction in Laddie Formation thickness from 330 to 230 m at its type section (two kilometers south of the southwest side of Laddie Harbour; Dimroth et al. 1970). Reasons for the change are discussed in Chapter 3. As a result there is little need for subdivision of the

Laddie Formation into members as the redefined formation consists predominantly of red and alternating red-green argillites and shales. Contact with the Costello Formation is gradational over several meters.

The lower red argillites are moderately fissile with a few interbeds of limestone and calcitic pods. Starved ripples also occur at this level but are rare higher in the formation. A marked decrease in carbonate lithologies accompanies this upward trend. Towards the top of the Laddie Formation, interbedded green argillites appear, and as the Rowatt contact is approached, thin dololomite beds and current-bedded lenticular sandstones reappear, heralding a return to much shallower conditions. Some 15 to 20 m below this contact, a zone of soft-sediment slumping 1 to 2 m thick is interpreted as having formed during storm surges.

BASIN MODEL FOR THE FAIRWEATHER - LADDIE SUCCESSION

The vertical succession from the Fairweather Formation to the Laddie Formation (approximately 1800 m thick) records a transition from shallow-marine platform carbonate facies to deeper water carbonate and terrigenous mud facies. This broad transgressive sequence is represented in all sections throughout the Belcher Islands and covers a palinspastically reconstructed area of more than 6000 square km. Determinations of paleoslope in this study confirm earlier suggestions (Dimroth et al. 1970; Barrett, 1975) that the paleoslope dipped west to southwest.

Basis for Interpretation:

All formation contacts are conformable. Several facies exhibit lithological characteristics and vertical relationships similar to some

well-documented modern carbonate environments, and is also in accord with Walther's Law of Correlation of Facies. Thus the vertical succession observed here represents a series of facies that are diachronous, ie. they are laterally associated in space and time. This approach is also taken for interpretation of the overlying Rowatt-Mukpollo-Kipalu succession.

The marine platform can be subdivided into a number of subfacies. Comparisons with Phanerozoic analogues is based primarily on the nature and composition of lithofacies and biofacies. For example supratidal, intertidal and shallow subtidal environments interpreted for McLeary Formation strata, particularly those of the Upper Member, can be determined independently by utilizing sedimentary structures and lithologies, and also a variety of cryptalgal laminate morphologies. The cryptalgal structures compare favourably with modern examples from Shark Bay, Andros Island and the Persian Gulf. However, the extensive biostrome complexes in the Mavor Formation have no known modern analogues although a few broadly similar ancient examples have been reported.

Facies Interpretations:

The characteristic features of each formation are listed in Table 3. The McLeary and Fairweather Formations can be subdivided into a number of subfacies. At least eighteen cycles recording the transition from intertidal or supratidal flat to caliche beds can be recognized in the Fairweather Formation. Each of these cycles is indicative of upwards-shoaling or progradation and is considered to represent a short-term still stand or lowering of sea level, with subsequent progradation of the shoreline. The short-term changes in sea level are superimposed on the overall transgressive regime. An interesting comparison can be

TABLE 3

Diagnostic features of the Fairweather to Laddie succession

FORMATION	CHARACTERISTICS	INTERPRETATION
LADDIE	Red & green argillites & shales.	Deep water, below storm wave base.
COSTELLO	Rhythmically bedded lutites, minor allodapic calcarenites, slumping, channels, intra-formational conglomerates.	Deep water, below storm wave base, turbidites.
MAVOR	Extensive biostromes, slump bed near base, minor cross-bedded dolarenites, stone rosettes, beachrock & thrombolites.	Offshore subtidal to effective storm wave base; Local shallow subtidal above wave base, intertidal.
TUKARAK	Ribbon rock; lenticular & flaser bedding, small bioherms, stone rosettes.	Shallow subtidal, above wave base.
MCLEARY	Complexly branched & simple domal stromatolites, tufted, wavy & digitate mats, oncolites; desiccation structures, evaporite casts, beachrock, stone rosettes.	Shallow subtidal, tidal flat & supratidal, beach, tidal channel.
FAIRWEATHER	Pisolitic caliche, desiccated mudrocks, lenticular sandstones, ferruginous argillites.	Supratidal & upper supratidal, tidal flat, tidal channels, isolated lagoons.

made with eustatic sea level curves established for the last 20,000 years B.P., where overall rise curves have numerous perturbations superimposed on them. These ephemeral sea level changes may be of the order of two meters over a time span of 1000 years or less (eg. Morner, 1971; Schofield, 1973).

In the Middle Member of the McLeary Formation a number of beach-rock beds have been identified which occur at the top of thin upwards-shoaling cycles. Although individual cycles cannot be correlated with those in the Fairweather, they are attributed to similar processes, principally short intervals of shoreface progradation resulting from minor changes in relative sea level. In this case the cycles may also be a result of local changes in carbonate sand influx, where increases in supply exceed changes in sea level rise, resulting in net progradation.

Transitions from the Tukarak to the Laddie Formations correspond to a gradual increase in water depth from shallow subtidal (above effective wave base) to a relatively deep basin environment. An extensive slump horizon in the lower Mavor is inferred to have developed within the range of effective storm wave base. The major biostromal complexes of the Mavor Formation accumulated below this level.

Rhythmically bedded dololutites and calcilutites of the Costello Formation appear very similar to rhythmic limestone lithologies considered by Wilson (1969) to be characteristic of foreslopes of carbonate platforms. Hoffman (1974) has described similar lithologies in his 'platform-basin' facies model. The vertical, conformable relationships between the Costello and Mavor Formations, together with the occurrence of beds having typical Bouma cycle organization and zones of slumping, is compatible with a similar interpretation for the Belcher strata.

Slumping along modern continental margins has been reported from slopes having gradients as low as one degree (Lewis, 1971). The history of deposition of this part of the Belcher basin can therefore be represented in a single model which is illustrated in Figs. 28 & 29. Here a broad continental margin is subdivided into three zones: a shallow marine platform (Fairweather to Mavor Formations), slope (Costello Formation) and basin (Laddie Formation). Transitions from shoal-water platform to deeper slope environments have also been documented in Cambro-Ordovician carbonate sequences in Nevada (Cook and Taylor, 1977) and the central Appalachians (Reinhardt, 1977) and also in the Lower Proterozoic Pethei Group, Great Slave Lake (Hoffman, 1974). Nevertheless, in all of these cases the successions are regressive. The thick transgressive sequence represented by the Fairweather to Laddie strata is therefore unusual in this respect. A closer analogy can be found in the transgressive Silurian to Devonian Helderberg Group, New York State, which records the transition from shallow water intertidal to deeper outer shelf environments (Laporte, 1969).

Paleobathymetry:

The variety and abundance of sedimentary structures in the Fairweather and McLeary Formations leaves little doubt that water depths were very shallow; a maximum of one to two meters is suggested for the uppermost McLeary deposits. The abundance of stone rosette pavements in many parts of the overlying Tukarak Formation supports this. Absolute depths for subtidal deposits are much more difficult to assess. However, the obviously prolific growth of stromatolites to the top of the Mavor Formation does suggest restriction to the photic zone. At present the depth limits of this zone, usually defined as that depth at which

Figure 28. Geometry of formations along an east-west transect through the Belcher Islands, and inferred changes in relative sea level.

Figure 29. Schematic representation of the shallow marine platform-slope-basin model for the Fairweather Formation to Laddie Formation succession. The coastline is depicted as being bordered by plateau basalts of the Eskimo Formation. Vertical scale is exaggerated.

light intensity is 1%, averages about 80 m but may be as much as 110 m as is presently observed in the clear waters of the Caribbean (King, 1975). In more turbid waters this depth is considerably less. The amount of organic and inorganic particulate matter that was present in mid-Pre-cambrian seas is unknown, although microbial forms having planktonic cyanobacterial affinities are reported from the Gunflint Formation (Awramik and Walter, 1978). It is unlikely that the photic zone extended significantly beyond present limits during the period of Belcher Group deposition, and therefore a depth limit for the major Mavor biostromes is tentatively set at about 80 m.

Paleoclimate:

Pisolitic caliche deposits form during periods of prolonged sub-aerial exposure and like all pedogenic horizons, relationships between temperature, precipitation, evaporation, relief and runoff are of fundamental importance. Reeves (1976) noted that the degree of pisolite development increases with time of exposure provided that dissolution, reprecipitation and desiccation persist. Thus the Fairweather caliche beds some up to a meter thick, could represent several thousand years of subaerial exposure. The ideal environment to promote caliche formation appears to be 'neither excessively arid or excessively humid...' (Reeves, 1976, p.84). The existence of a semi-arid climate is supported by the presence of evaporite minerals in the supratidal facies, and the relatively widespread occurrence of beachrock. Modern beachrock formation is generally restricted to the lower latitudes, between 30°N and 30°S.

PLATE 5

- Figure 1. 'Ribbon rock' typical of the Tukarak Formation showing recessive weathering of dolomitic bands relative to the argillaceous dolostones (dark). Scale is 15 cm long. West Tukarak Island (Section 1).
- Figure 2. Stromatolite bioherms, Upper Tukarak Formation, draped by ribbon rock. Stylolites in center of the bioherm conform to bedding. Northwest shore of Mavor Island, grid reference 289158.
- Figure 3. Aerial view of the main biostrome complex, Mavor Formation, 3.2 km west of the Laddie Harbour entrance. The horizontal distance across the mound field is about 800 m. Bedding dips 10° to 15° to the right. Mound spacing and amplitude increase towards the right (up section). Mounds also coalesce in this direction.
- Figure 4. Example of a single large biohermal mound from the main biostrome complex, east Tukarak Island (Section 27). Inheritance observed here is about four meters.
- Figure 5. Wavy and crenulate cryptalgal laminates typical of the large stromatolite mounds in the Mavor Formation. Scale in centimeters. Mavor Island, Section 33.
- Figure 6. Second-order algal mounds superimposed on the major biohermal structures in the Mavor Formation. Inheritance in the second-order structures is up to two meters. Hammer for scale (arrow). Section 35, due west of Laddie Harbour.

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PLATE 6

- Figure 1. Thrombolites (outlined by broken line) in the Mavor Formation, northeast Tukarak Island (Section 27). A crudely laminated envelope can be seen in the mound on the left (arrows). Costello Lake in the background. Thrombolite on the right is 80 cm high.
- Figure 2. Molar-tooth structure in dolomitic limestones of the upper Mavor Formation, east side of Eskimo Harbour. Pen is 15 cm long.
- Figure 3. Rhythmically bedded dark argillaceous dolostones and white calcilutites typical of the Costello Formation, on central Mavor Island (Section 33). Staff is 130 cm long.
- Figure 4. Lumpy white-grey calcilutites and thin cross-bedded calcarenites of the Costello Formation. The bed below the pen contains A, B and C divisions of a Bouma cycle, with climbing ripples in the C division. Central Mavor Island, Section 33.
- Figure 5. Large slump folds in Costello rhythmites. Typical corrugated exposure is due to differential weathering of less resistant calcitic beds. Hammer, left center. Tukarak Island, west of Laddie Harbour (Section 35).
- Figure 6. Flat pebble conglomerate filling channel in Costello rhythmites. Pebble compositions are similar to typical Costello lithologies and are largely calcitic. Same locality as Figure 5.

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CHAPTER 3

THE ROWATT, MUKPOLLO AND KIPALU FORMATIONS: A
PRECAMBRIAN CARBONATE BUILDUP AND BARRED BASIN
COMPLEX.

INTRODUCTION

GENERAL SETTING

The Rowatt Formation-Mukpollo Formation-Kipalu Formation sequence represents an overall shoaling-upwards succession of carbonates, terrigenous clastics and ironstones, and marks important changes in tectonism, depositional environment and paleocoastline configuration from the preceding transgressive regime represented by the Fairweather Formation to Laddie Formation succession. The transition from a transgressive to regressive regime occurs in the upper part of the Laddie Formation.

Analysis of the regional paleoslope by Barrett (1975) confirmed Jackson's (1960) conclusions that pre-Flaherty Formation terrigenous sediments were derived from a cratonic source northeast of the Belcher Basin. In the present study, almost 2000 azimuth measurements of cross-bed foresets, ripple marks and stromatolite elongation azimuth substantiate these earlier conclusions.

STRATIGRAPHY

Brief descriptions of the formations are given by Jackson (1960) and by Jackson in Dimroth et al (1970). Type sections were designated on the south side of Laddie Harbour, eastern Tukarak Island. In Jackson's original definition, the base of the Rowatt Formation was placed at the contact between a unit of well-bedded quartz sandstones intercalated with dolarenites and an overlying unit of dolostones, the dolostones typically forming prominent ridges that are easily traced in the field and on aerial photographs. According to this definition, the Laddie Formation, which is characterized by red and green argillites and shales, also contains cross-bedded sandstones and isolated, lenticular dolostone bodies up to 400 m in length, that have most of the

features attributed to the Rowatt dolostones. Detailed mapping of Tukarak and Flaherty Islands demonstrates that these dolostones pinch out from east to west, and in the central part of Flaherty Island they are replaced by a completely conformable sequence of sandstones and clastic carbonates. The criterion for defining the Laddie-Rowatt boundary therefore becomes untenable in the western part of the Belcher Islands. A new definition is proposed: the base of the Rowatt Formation is placed at the first sequence of tabular cross-bedded sandstones above the typical red and green argillites of the Laddie formation. The Laddie thus contains thinly bedded argillites and shales with only minor lenticular sandstone beds at the top of the formation. The new definition creates lithological uniformity for each of these formations, and, in addition, the criterion can be used throughout the Belcher Islands.

As a consequence of this definition there must be changes in the internal subdivision of the Rowatt Formation. It is divided into two members (Fig.30):

Lower Member: extends from the Laddie-Rowatt contact to the base of the main dolostone unit. It includes isolated dolostone lenses that occur just below the main dolostone unit.

Upper Member: this is a combination of the former Upper and Lower Members defined by Dimroth et al. (1970). The top of this member is at the Rowatt-Mukpollo contact.

THE UPPERMOST PART OF THE LADDIE FORMATION

The transition from thin-bedded red and green argillites to basal Rowatt sandstones is marked by the appearance of small lenticular sandstone

Figure 30. Proposed changes in the Rowatt Formation - Laddie Formation boundary.

lenses, or starved ripples, in the upper 15-20 m of the formation (Fig.31). Symmetric ripples also occur locally, suggesting that deposition probably took place above wave base. This transition zone represents the progressive change from a relatively deep subtidal environment.

A series of beds 15 m thick which displays soft-sediment slump structures, occurs about 30 m below the top of the Laddie Formation in Sections 15 & 16. Each slump packet is 0.5-1.5 m thick, is broadly lens-shaped and contains highly contorted argillites, some in detached blocks that are internally deformed. The slump horizons are bound above and below by completely undeformed rocks although the edges of most slump packets cut steeply across undisturbed beds. The apparent direction of slump transport was to the south. Slumping may have been triggered by earthquakes or by storms which affected sediments within the limit of storm wave base. The latter mechanism is preferred because slumping has not been observed in sections lower in the Laddie Formation

ROWATT FORMATION

The following analysis of the Rowatt Formation is based mainly on studies of sections on Tukarak and outlying islands, where exposure is exceptionally good, particularly in Sections 15 & 35. Exposures on Flaherty Island are generally poor except in Sections 23 & 26, which show important lateral facies changes.

LOWER MEMBER: (105-140 m)

The Lower Member is characterized by terrigenous clastic deposits, although clastic carbonates become increasingly important from about the middle of the sequence towards the dolostones of the Upper Member. This vertical change in the lithology, accompanied by sequential changes in

sedimentary structures, forms the basis for an overall interpretation of lateral facies relationships, and is discussed in more detail later in this chapter.

The Member can be informally subdivided into three zones (Fig.31) on the basis of lithology and sedimentary structures, and although boundaries between zones are gradational, the subdivision provides a convenient means for comparing lateral facies changes over a wide area.

Zone A

This is composed of tabular and lenticular beds, up to 120 cm thick, of cross-bedded sandstone, generally orthoquartzitic, separated by shaly partings or interbedded with thin mudstones (Pl.7, Fig.1). The predominant bed forms include herringbone cross-beds and large ripples having wavelengths up to 100 cm, with smaller current ripples superimposed at 90° (Pl.7, Fig.2). In addition, the sandstones may contain large tabular-planar cross-bed sets in which individual laminae are distinctly graded and make an angular contact with the lower set boundary. In some, small-scale ripples have migrated up the slip face, either as a result of backflow eddy currents, or sudden changes in current directions, and thus they resemble bed forms described from modern intertidal sand bars (e.g. Klein, 1970, 1971).

Interbedded, parallel and wavy-laminated mudstones show a considerable range in thickness, from shaly partings infilling ripple troughs to beds 50 cm thick. Thicker beds also contain a variety of structures such as small sand lenticles, and interference and double crested ripples that occasionally have thin mud-draped reactivation surfaces, all of which indicate rapid changes in current velocity or direction.

Sedimentary structures of this type are typical of modern clastic

Figure 31. Schematic representation of the Lower Member, Rowatt Formation; a composite of Sections 35 (Laddie Harbour) and 15 (south of McLeary Point).

intertidal environments and are indicative of tidal current asymmetry, especially bimodal current reversals, plus late-stage emergence run-off during ebb tides where sudden changes in flow direction can occur in shallow water. There is little evidence of prolonged subaerial exposure.

Paleocurrent directions summarized in Fig.32 show that for Zone A strata, bimodal-bipolar and quadrimodal distributions are relatively well defined. An interesting feature of the large dunes or sand bars is their direction of transport which was predominantly to the northwest. This is almost at right angles to the main direction of tidal bedload transport indicated by a variety of sedimentary structures. Similar structures in Zone C also display these trends. Bedforms such as these may represent sand transport that took place in response to longshore currents flowing approximately north to northwest, parallel to the ancient shoreline. Comparable subaqueous dune forms in Cretaceous sandstones in New Mexico have also been attributed to longshore currents by McCubbin (1965).

Sedimentary structures and paleocurrent azimuth distributions are strikingly similar to those described from many modern and ancient regressive intertidal sequences (Swett et al. 1971; Klein, 1971; and numerous examples in Ginsburg, 1975). Basal strata of the Rowatt Formation are also considered to represent deposition in a tidal flat or shallow subtidal environment that was characterized by migrating dunes and sandbars.

The top of Zone A is generally indicated by the appearance of thick laminated mudstones and siltstones. However, local occurrences of lenticular sandy dolarenite or grainstone beds, 2 to 7 m thick, contain subhorizontal laminae, planar crossbed sets with low-angle

Figure 32. Paleocurrent directions from the Rowatt Formation, eastern Tukarak Island and south of McLeary Point: LR. = large current ripples, IR. = interference ripples, TXB = trough cross-beds, PXB = tabular planar cross-beds, HB = herringbone cross-beds, SR = symmetrical ripples.

truncations (less than 10^0), a few herringbone crossbeds and small scours. The grainstones are typically framework supported and consist of moderately well-sorted dololomite clasts and 10 to 15% rounded quartz grains. This marks the first appearance of clastic dolomite in the lower levels of the Rowatt Formation. An interesting example of one of these sand bodies occurs on Innetalling Island (Section 34) where a single large-scale planar crossbed set has been deposited above dislodged blocks of beachrock. Examples of beachrock occur in a number of formations in the Belcher Islands, and provide a reliable indication of ancient strandlines (Donaldson & Ricketts, 1977; and in press). The association of sedimentary structures and beachrock clearly demonstrates that the grainstones were deposited in an intertidal environment, and in part as beach sands. The relatively restricted occurrence of these grainstones, in relation to the overall tidal flat environment inferred for Zone A, suggests that they may have formed as beaches on small emergent banks or sandbars. The ephemeral nature of such an environment is demonstrated in this example where beachrock has been buried by migrating dune beds.

Zone B

The middle zone is made up predominantly of grey-green mudstones and fine-grained sandstones, and a few coarse tabular sandstone beds, the proportions of which vary from section to section. Coarser lithologies typically contain trough and herringbone cross-beds. The most common bedform in the fine-grained rocks is lenticular and wavy bedding where mud layers alternate with and enclose starved current ripples, small scours and sparse oscillation ripples. Some ripples exhibit mud-draped reactivation surfaces, and flaser bedding is common in the coarser

lithologies. Ripple and crossbed foresets show bimodal directions. The incidence of lenticular bedding increases towards the top of the zone to a 1:1 ratio of mudstone to sandstone, where desiccation polygons first appear.

East-trending channels, 1 to 80 m across and up to 1.5 m deep, cut mudstones about midway through this zone, and are filled with cross-bedded dolarenites. Channel lag conglomerates consist of dololutite cobbles and mudstone rip-ups. Slumping of mudstones that form the channel walls is a common feature (Fig.33a) and some channels display dislodged and deformed blocks, apparently the result of channel undercutting. In other instances, blocks display only minor rotation along small fault planes that are concave upwards. Soft-sediment deformation structures almost identical to these have been described from modern point bar deposits associated with tidal channels, where fracturing occurs in sediments of high water content that are subsequently exposed at low tides (Reineck and Singh, 1975).

Two additional structures in some of the sandstone beds are possible rill marks and dewatering structures. The rill marks (Fig.33b) are similar to meandering and bifurcating rills illustrated by Reineck and Singh (1975) which represent drainage patterns formed in response to exposure during ebb tides. The dewatering structures are closely spaced conical structures in fine-grained laminated sandstone which apparently resulted from expulsion of water during the early stages of compaction (Fig.33c). The cones terminate in a single horizon and some display small bulbous 'caps', 1 to 2 cm across, that in cross-section are similar to pit-and-mound structures described by Shrock (1948, pp.132-134).

Figure 33. (a) Slump structures in mudstones formed along channel walls as a result of dewatering during exposure, (b) bifurcating rills and small current ripples. Note progressive change in ripple amplitude down flow. (c) Pit-and-mound structures in fine sandstones (vertical section) probably resulting from dewatering. All structures from the Lower Member, Rowatt Formation, located south of McLeary Point.

MODERN ANALOGUES

The association of sedimentary structures described above is comparable to many modern intertidal sand and mud flat environments such as the North Sea coast of Europe and the Minas Basin (DeRaaf & Boersma, 1971, Klein, 1970, Reineck 1972). Furthermore, the common occurrence of desiccation cracks in the upper levels of Zone B indicates relatively prolonged subaerial exposure, interrupted by high tides, suggesting transition from lower to upper intertidal and possibly supratidal environments.

The presence of thin cross-bedded sandstones in this interval can be interpreted in two ways: (i) the sandstones represent minor fluctuations in sea level with a return to lower tidal flat or shallow subtidal conditions, or (ii) they represent lateral changes in a dynamic, clastic intertidal environment resulting from sandbar migration along a tidal flat and subsequent local changes in shoreline configuration. Migration of these sandbars may have taken place in response to longshore currents.

In making the above interpretation, comparisons with modern environments is based primarily on sedimentary structures and lithologic associations. There is however, a fundamental difference between Precambrian siliciclastic tidal flat environments and modern analogues in that there is no evidence for the existence of sediment-stabilizing agents comparable to modern benthic fauna and flora such as sea grasses on tidal flats, or Mytilus 'mats' and sabellaria reefs that stabilize sediments on the walls and floors of tidal channels (this is not strictly true for many Precambrian carbonate environments, where growth of algae was often prolific). The degree of bioturbation, although somewhat dependent on the rates of deposition, may also affect sediments to the

point where all primary structures are obliterated. Lack of infaunal and epifaunal invertebrates in early Proterozoic sediments greatly increased the likelihood of preserving even the finest structural details. The absence of slope-stabilizing agents is manifested particularly well in the abundance of small slump structures along the walls and floors of channels. In the example cited below, there is little evidence for the presence of algal mats that might have bound the soft muds. In addition, channel walls rarely cut steeply across or undercut the adjacent mudstones. Under these conditions it is likely that lateral migration of channels could take place at considerably faster rates than occur at present.

Zone C:

This zone is characterized by clastic dolostones, mainly dolarenites, but also includes carbonate conglomerates, dololutites and sandstones. They contain a particularly diverse assemblage of sedimentary structures (Fig.31) and also display the most rapid lateral facies changes of any of the zones described so far (Fig.34).

The top of Zone C marks the lower boundary of the main carbonate unit (Upper Member), a contact that can be traced about 50 km along strike on eastern Tukarak Island. A number of isolated lenticular dolostone bodies, ranging from 100 to 400 m in length, occur below the Upper Member, and details of these will be treated along with this member. A number of features distinguish this zone from the underlying strata:

(i) a strong tidal influence still prevails as indicated by structures such as herringbone cross-beds, interference ripples and re-activation surfaces (Pl.7, Fig.3), and by flaser and lenticular bedding in lutites and mudstones. Paleocurrents are bimodal-bipolar and locally quadrimodal.

Figure 34. Diagrammatic representation of facies changes in Zone C strata, Lower Member, Rowatt Formation. Note rapid pinching out of brecciated dolostone bodies.

(ii) the incidence of subaerial exposure with formation of desiccation cracks and runzel marks is much greater.

(iii) conglomerates consisting mostly of dololutite and dolarenite cobbles and boulders are of particular interest in that clast lithologies are identical to dolostones that compose the Upper Member. This clastic carbonate component increases towards the overlying dolostones. Two types of conglomerate occur.

Channel Lag Conglomerates:

Channels in Zone C are similar in most respects to those described below. In some examples, evidence for channel migration is observed where conglomerates overlies dololutites which contain climbing ripples (Fig.35b). According to Reineck (1975), formation of such bed forms in channels usually occurs at a channel mouth where muds fall rapidly out of suspension.

Bedded Conglomerates:

Lenticular conglomerate beds up to 80 cm thick (Pl.7, Fig.4) grade laterally over 50 m and more into cross-bedded dolarenites and lutites. Some boulders exceed 50 cm in length. Size sorting is extremely poor and clasts are angular, suggesting rapid high-energy deposition with very little reworking. The conglomerates are commonly overlain by planar and herringbone cross-bedded dolarenites which may grade laterally into desiccated mudstones. An example is depicted in Fig.35a, where at least two erosional surfaces can be seen. Here the conglomerates and truncated sandstone bed are thought to be a product of erosion and deposition during storms, possibly as beach ridges, with the overlying strata representing a return to 'normal' intertidal conditions.

(iv) the transition from these intertidal sediments to the

Figure 35. (a) Bedded conglomerates possibly of beach - storm ridge origin. Field sketch. Lower Member, Rowatt Formation, (b) Channel-lag conglomerate (predominantly dolostone boulders) has migrated over dololuites that were likely deposited at the channel mouth. Ripple terminology after Jopling and Walker(1968).

lenticular dolostone bodies and Upper Member dolostones is either gradational or abrupt. The latter type is often marked by desiccated mudstones that underlie folded or brecciated dolostones; the mudstones however exhibit no deformation. Gradational contacts are characterized by grainstone beds up to two meters thick, consisting of clastic dolomite and 10% to 20% rounded quartz grains. In addition to herringbone cross-beds, mud-draped reactivation surfaces and other indicators of current reversals, the grainstones contain subhorizontally laminated intervals and low-angle truncation of planar cross-beds. Beachrock occurs as lateral equivalents to some of these grainstones. A particularly fine example occurs on eastern Flaherty Island (Section 23) where several generations of beachrock can be recognized (Fig.36 & Pl.7, Fig.5); three stages of preferential lithification alternating with periods of erosion and formation of flat boulder conglomerates and encrusting algal mats. By direct analogy with examples of modern eroded beachrock (eg. Emery et al. 1954), the conglomerates provide additional evidence for violent storms along this ancient shoreline. Evidence of early rapid cementation indicative of beachrock is complemented by tidal channel and storm ridge conglomerates that contain cobbles of grainstone and lutite clasts, thus representing several periods of sedimentation, lithification and erosion.

UPPER MEMBER

Descriptions and interpretations of dolostones in this section also apply to the isolated dolostone bodies in the Lower Member. A complete vertical section of this member is exposed on the south shore of Laddie Harbour (Fig.37).

Figure 36. Beachrock beds near the top of the Lower Member, Rowatt Formation, showing several stages of preferential lithification and erosion. 1. Initial beachrock bed, 2. erosion of Stage 1 beachrock, 3. recementation of beachrock boulder conglomerate and subsequent erosion forming small channels, 4. cover of Stage 3 by grainstone followed by lithification, further erosion and subsequent burial by more grainstone. Sketched from photograph. Outcrop is located on eastern Flaherty Island.

Figure 37. Schematic representation of the Upper Member, Rowatt Formation (Laddie Harbour). Insets illustrate typical breccia-bedding relationships: (1) lutite deformation (a) and penecontemporaneous erosion followed by burial beneath cross-bedded grainstones (b) and finally large-scale brecciation (c). (2) chaotic breccia (d) consisting of dololite and grainstone blocks in lutite-sand matrix, overlain by mildly brecciated grainstones (e) and folded dololite beds (f). These sequences are repeated several times in this member.

The main carbonate unit is composed of what were probably well-bedded grainstones and banded dololutites, but which now display different degrees of brecciation and folding. Some tabular sandstone and orthoquartzite beds occur near the top of the member.

Grainstones, generally less than a meter thick, are characterized by subhorizontal laminae, planar cross-beds with graded foreset laminae, and a few thin graded beds. Some beds can be traced 100 m and more before pinching out or before bedding is destroyed by brecciation. In thin section, the grainstones show a clast-supported framework consisting of clastic dololutite and variable proportions of rounded quartz grains. Both carbonate and quartz grains exhibit micritic rinds and possible relic meniscus cements suggesting that cementation was in part vadose. Carbonate clasts are mostly dolomicrites or microspar with different degree of crystallinity; micrite clasts, with average crystal sizes of 5 to 10 μm , sharply abut grains composed of much coarser dolomite mosaics (up to 50 μm). Such variations are considered to reflect differences in the degree of recrystallization of the local source sediments rather than post-depositional changes. The original source sediments probably included penecontemporaneously dolomitized aragonite muds and carbonate crusts (see discussion in Chapter 5). An additional component of the grainstones is large numbers of rounded chert fragments. The proportion of chert in the Rowatt sandstones is greater than in similar lithologies in the Belcher succession and may reflect derivation from laterally related facies of the Kipalu Formation.

Zones of thinly bedded dololutites, in layers up to a meter thick, are interbedded with grainstones in a crude cyclical fashion, and characteristically are highly deformed (Pl.8, Fig.1). Individual beds

range from 1 to 15 mm thick and consist of extremely fine-grained dolomite (3 to 5 μm). Lutite deformation was generally plastic, producing cylindrical folds, although some brittle failure is apparent in isoclinally folded beds (Pl.8, Fig.2). In addition there is a direct correlation between fold wavelength and bed thickness (ie. wavelength decreases with decreasing bed thickness) and suggests a possible buckling mechanism. The dependence of fold wavelength on layer thickness, within a sequence of buckled layers of constant composition, has been verified experimentally (Bell and Currie, 1964; Ramsay, 1967, pp.372-380). Axial planes of folds are subvertical or slightly inclined either to the east or west; although no statistical count was made, measurements at a number of stations indicate no preferred direction of inclination. Fold amplitudes range from 1 to 30 cm.

Crescent-shaped voids have developed in the troughs and crests of folds where cohesion between layers might be expected to be less effective during buckling. The voids are filled with radiaxial fibrous dolomite, presenting a fabric that may be a replacement of an early fibrous aragonite cement (see Chapter 5). The penecontemporaneous nature of lutite deformation is clearly demonstrated in Fig.37, where folds have been truncated by an overlying cross-bedded sandstone.

Brecciation:

The degree of brecciation of grainstones and lutite beds ranges from relatively mild, with little movement of displaced blocks (Pl.8, Fig.3), to zones of chaotic angular blocks. In strata displaying little deformation, fracturing has taken place parallel and normal to bedding (Pl.8, Fig.3). Dololutites that infill some of these fractures show vague layering parallel to fracture walls. In some cases refracturing

of this infill has given rise to fibrous dolomite crystals oriented at right angles to the walls, possibly replacing an earlier cavity-fill cement. Many of the dololutite clasts are truncated at the fracture walls, suggesting some degree of grainstone lithification at the time of brecciation.

Zones of chaotic breccia up to 10 m thick are interspersed throughout the Upper Member. Blocks as much as a meter across are highly angular and consist of dololutite and grainstone. Sorting is extremely poor. The boundaries of these zones appear irregular and some are difficult to trace; in places boundaries cut across bedding in the surrounding dolostones whereas in others they appear to grade laterally into undisturbed beds. The breccias are commonly overlain by dolostones that exhibit very little deformation, except for folding in the lutites, and it is apparent that brecciation was essentially penecontemporaneous.

A few broad open slump folds were observed in areas of more intense deformation, and although axial planes generally strike north, they are inclined both east and west.

Channels:

Upper Member dolostones are cut by channels, some more than 200 m wide, that are infilled with well-sorted dolomitic sandstones and grainstones. Trough cross-bedding is abundant and measurements indicate strongly unimodal paleocurrents that flowed westward down the paleoslope (Pl.8, Fig.4). Locally, large angular blocks of grainstone occur along the floors of the channels, some crudely imbricated and some inverted. These grainstones must have possessed a considerable degree of lithification at the time of erosion and may represent beachrock that formed along channel margins. Channel deposits show only minor deformation.

Sandstones and Dolarenites:

In the upper part of this member, sequences of dolarenite and orthoquartzite are interbedded with brecciated dolostones. Each sandstone sequence, regardless of its thickness, exhibits a distinct lithologic transition from orthoquartzite at the base to dolarenite at the top. The dolarenites in turn grade into brecciated and bedded dolostones. These sandstone-dolostone sequences appear to be minor shoaling upwards cycles; the sandstone-grainstone cycles in the Middle Member of the McLeary Formation are in many ways comparable to these. Bedforms in the sandstones are similar to those in the coarse lithologies of the Lower Member (Rowatt Formation). Paleocurrents are weakly quadrimodal, with a dominant mode in the northwest quadrant that in many places is associated with tabular-planar cross-beds.

Stromatolites:

Subcylindrical non-branching stromatolites occur sporadically throughout the dolostones, mainly in grainstone beds that exhibit a grey-clotted texture; this texture is commonly associated with cryptalgal structures in the McLeary Formation. Algae are thought to have played an important role in binding the carbonate sands.

THE ROWATT FORMATION WEST OF TUKARAK ISLAND

Despite rather poor exposure on Flaherty Island, two vertical sections were measured (Sections 23 & 26) which demonstrate important facies changes (summarized in Fig.38). The most striking lithologic variation is the rapid thinning of Upper Member dolostones which in eastern Flaherty Island are present in two beds, 10 and 2 m thick. Here the boundary between the Upper and Lower Member is placed at the base of the

Figure 38. Facies changes in the Rowatt Formation, in both north-south and east-west directions.

lowest dolostone. The remaining strata in the Upper Member consist of tabular sandstone and dolarenite beds with many sedimentary structures indicative of an intertidal environment. Of particular note are a number of well-preserved beachrock beds near the top, commonly in well-defined shoaling upwards cycles, 3 to 5 meters thick. Each cycle is made up of basal dolarenite containing herringbone and ripple-drift cross-beds, an intermediate zone of subhorizontal laminae and planar cross-beds and an uppermost unit of beachrock that in places exhibits two or three generations of lithification and erosion.

The Lower Member on eastern Flaherty Island is considerably thicker than sections on Tukarak Island. It displays most of the features already described from its more eastern counterparts, and therefore can be readily interpreted as being representative of an intertidal to shallow subtidal environment. Farther west, in Eskimo Harbour (Section 26), the Upper Member dolostones are absent, their place being occupied by sandstones and dolarenites with sedimentary structures similar to those present in Lower Member strata. Desiccation structures are far less common than in any of the sections to the east. In this section there is little to distinguish between the Upper and the Lower Members and thus no subdivision has been made.

SYNTHESIS OF THE ROWATT FORMATION

One of the isolated dolostones bodies (Lower Member) along eastern Tukarak Island was interpreted by Barrett et al. (1976) as an allochthonous debris flow within a calc-flysch sequence that included the Tukarak, Costello and Laddie Formations, but not the intervening Mavor Formation. On the basis of the discussion of these units in the previous

chapter, the term calc-flysch is not considered to be appropriate. Furthermore, observations on this and other dolostones along eastern Tukarak during this study have led the present writer to suggest an alternative origin for the brecciated dolostones of the Lower and Upper Members of the Rowatt Formation. Evidence for this is summarized below:

(i) the contact between the clastic intertidal deposits and dolostones (Lower Member), both laterally and vertically, is either sharp (generally in contact with desiccated mudstones) or gradational; the latter relationship indicates a continuous sedimentary sequence.

(ii) the dolostone bodies and the main carbonate unit (Upper Member) overlie beach and some supratidal deposits that form the upper limits of a shoaling upwards, clastic intertidal wedge. The proportion of clastic dolomite in these intertidal deposits, especially in the conglomerates, also increases upwards towards the overlying dolostones. The cobbles and boulders appear identical to the Upper Member lithologies, and therefore provide evidence for lateral association of these two facies. Furthermore, most of the dolostone bodies are sandwiched between mudstones containing desiccation cracks. If these bodies, or blocks, are allochthonous, then they must have moved downslope into an intertidal - supratidal environment, a situation considered unlikely.

(iii) deformation is confined to the dolostone bodies and Upper Member dolostones. If they were the products of gravity sliding, some deformation might be expected in the underlying deposits, but none has been observed. In addition, deformation has not occurred as a single event. For example, folding of the dololutes can be shown to be penecontemporaneous, having occurred at several stages following deposition and partial lithification of the carbonate sediments.

(iv) the internal organization of the isolated dolostone bodies (crude bedding, folding, brecciation and channels) is identical to that of the Upper Member rocks and thus are likely to be of the same origin. The Upper Member rocks form a thick, continuous wedge that can be traced at least 50 km parallel to strike.

Rocks in the Lower Member of the Rowatt Formation are clearly of intertidal origin, and on eastern Tukarak Island, subdivision into shallow subtidal, lower, middle and upper tidal flat, and beach deposits is possible. It is suggested that the Upper Member dolostones were positive physiographic elements relative to mean sea level, and that they represent an ancient carbonate buildup complex that was bordered on the seaward margin by an intertidal platform. This platform was also characterized by migrating sandbars and small isolated carbonate buildups.

Formation of the dolostones can be regarded as a complex sequence of events involving numerous cycles of carbonate mud and sand deposition, lithification, erosion and deformation, in what was essentially an intertidal-supratidal environment.

Origin of the Carbonate Muds:

The origin of lime mud has long been a subject of discussion, with two major schools of thought evolving: on one side those who propose an inorganic origin by direct precipitation from seawater, and those who favour skeletal breakdown and either direct or indirect biologic control (see discussions in various sections of Bathurst, 1976).

Direct comparison of middle Precambrian carbonate mud sequences with modern examples is at once limited by two factors: the absence of calcium carbonate secreting invertebrates, and the probable lack of codiacean-type algae (coccooid and filamentous algae are the predominant

forms reported from rocks of this age, eg. Hofmann, 1976a). In South Florida and the Bahama Banks, production of carbonate muds is controlled to a large extent by codiacean algae (eg. Halimeda, Penicillus) and also by mechanical breakdown of skeletal organisms. The role played by blue-green algae is precipitating carbonates in the Rowatt lutites is difficult to determine. Cryptalgal structures have not been recognized in the well-bedded dololutites although such structures do exist in some of the grainstones. The possibility of some form of endolithic algae cannot be excluded.

A closer analogy may be the Trucial coast tidal flats of the Persian Gulf, where the influence of Codiacean algae is minimal. Here a possible link between the aragonite of ooids and that composing the lime muds, has been suggested on the basis of almost identical Sr^{2+} concentrations (Kinsman, 1969). However, even here physiological controls on precipitation by (endolithic?) algae are also possible (Kendall and Skipwith, 1969).

An origin by direct (biologically influenced?) precipitation is considered feasible for the Rowatt lutites, possibly by flooding of a supratidal flat during high tides or storm surges. Subsequent evaporation may have resulted in carbonate precipitation, probably as aragonite. On the basis of Sr distributions in penecontemporaneous and early diagenetic dolostones from Paleozoic carbonate rocks, Veizer et al. (1978) suggested an origin whereby aragonite muds originally deposited in a hypersaline environment, or at least an environment having increased salinities, were replaced by dolomite.

Origin of the Layering:

Although now highly deformed, it is clear that the thinly bedded

dololulites in the Rowatt Formation originally were extremely regular. However, as noted above, an algal origin for the layering is unlikely. The most plausible explanation for the degree of preservation is the absence of burrowing and sediment churning organisms during middle Precambrian times. For sediments exposed subaerially for any length of time, desiccation is a further process whereby primary layering can be disrupted; no direct evidence of desiccation was found in the Rowatt lutites. Nevertheless, high-energy erosion accompanied by sudden influxes of sand did produce widespread disruption of these muds.

Ball et al. (1967) and Hardie and Ginsburg (1977) have observed that significant quantities of mud may be stranded on supratidal flats during storms. Thus, within the framework of an intertidal-supratidal paleoenvironment inferred for the Rowatt dolostones, the well-bedded lutites could have formed as a result of flooding by storm surges, with deposition and/or precipitation of aragonite in broad, shallow inland ponds. Continued submergence within these ponds would have prevented intense desiccation. However, muds deposited on the adjacent exposed flats would be highly susceptible to desiccation and erosion, especially in the absence of a sediment-stabilizing flora.

Comparisons can be made with modern ponded flats from the Bahamas (Hardie and Ginsburg, 1977) although important differences include:

- differences in pond-profile, namely a decrease in overall relief, would result from a lack of sediment-stabilizing plants (eg. mangroves);
- in the absence of stabilizing agents the ponded flats would be more ephemeral than their modern analogues;
- lower relief would result in more widespread flooding.

Whereas the lutites were deposited under very quiet conditions, the sudden grainstone influxes may be similar to the spill-over lobes described by Bathurst (1976, p.315). Brecciation and folding of the dolostones occurred at several stages following deposition and lithification of the sediments. A mechanism of penecontemporaneous solution collapse, or possibly karstification, is proposed to account for this deformation

MUKPOLLO FORMATION

Named after Mukpollo Peninsula, this formation has its type section at the southern end of Laddie Harbour (Dimroth et al. 1970). However, a more complete and better exposed section (#38) is exposed 3.5 km south of Laddie Harbour on the southern shore of an unnamed harbour (1.2 km north of McLeary Point) and this is proposed here as a reference section (Fig.39).

Although the Mukpollo Formation is characterized by orthoquartzites, it surprisingly is poorly exposed. Mukpollo rocks crop out over a wide area in the Belcher Islands, including Howard Peninsula of westernmost Flaherty Island. The thickness over this area remains relatively constant and hence the formation has a sheet-like geometry that covers a known palinspastic area of approximately 4700 square km. The upper contact with the Kipalu Formation is known to be exposed at only two sites, on Tukarak and Innetalling Islands (Section 38 & 5 respectively). At both localities the contact is conformable, with a rapid transition from orthoquartzites to banded ironstones within less than 50 cm. Contact with the underlying Rowatt Formation is also sharp and conformable on Tukarak Island. On Flaherty Island, the Mukpollo-Kipalu contact is

Figure 39 . The Mukpollo Formation and measured paleocurrents from a section north of McLeary Point, Tukarak Island.

likely to be sharp (not observed directly), but contact with Rowatt strata is much more difficult to establish. In Sections 23 and 26, the transition from dolarenites and sandstones typical of the Rowatt, to well-bedded orthoquartzites, takes place over several meters.

The lower half of this formation consists of well bedded orthoquartzites, and grey and purple siltstones, with a few dolomitic sandstones near the base. Tabular-planar and trough cross-bed sets up to 40 cm thick are abundant in the coarser lithologies, but herringbone cross-beds, straight-crested and lunate current ripples, and small scours are also common. Measurements of planar cross-bed foresets indicate a dominant north to northeast trend for currents that formed these bed-forms (Fig.39), although the distribution is weakly bimodal. Thin interbedded mudstones commonly display evidence of desiccation.

In the central portion of the Mukpollo Formation, purple-grey siltstones contain lenticular bedding, ripples and desiccation cracks. Sandstone lenticles are commonly deformed by compaction loading. Two other structures are worth noting:

(i) fine-grained dolostone lenses, 2 to 5 m across and less than 50 cm thick, exhibit very little internal structure other than faint, wavy laminations that possibly are of algal origin. Contacts with the surrounding sandstones are sharp.

(ii) channels 4 to 5 m across have been cut into laminated and lenticular siltstones, and filled with cross-bedded quartz arenites. The channeled siltstone beds have warped due to differential compaction. Clasts in the channel lag conglomerates include quartz and chert granules, and dolomite and ironstone pebbles. The most likely source for the ironstone clasts is the overlying Kipalu Formation. Thin sections of

chert grains reveal textures identical to those shown by bedded cherts in the ironstones. The presence of these clasts thus provides evidence for lateral association of the Mukpollo quartz arenite facies and the banded ironstone facies.

The upper 50 to 60 m of the Mukpollo Formation is made up of more massive orthoquartzites with only rare mudstone partings. Sedimentary structures are generally indistinct, although a few bedding surfaces exhibit large ripples 10 cm in amplitude. Festooned planar cross-beds were seen locally on eastern Tukarak Island, and parallel laminated sets with low-angle truncations were observed.

The supermature rounding of grains, extremely good sorting and almost monomineralic composition indicates prolonged abrasion and suggests intense reworking. Conditions such as these exist along coasts possessing active long-shore currents, such currents commonly being manifested in barrier and offshore bar systems (Evans, 1941), as for example along the coast of the Gulf of Mexico (Dickinson et al. 1972) or northern New Zealand (Ricketts, in press). Several of the sedimentary structures are indicative of intertidal conditions, whereas periodic exposure is indicated by desiccation cracks. Thus Mukpollo strata are thought to represent deposition in an intertidal environment, possibly as sand flats and as sand bars or shallow subtidal shoals.

PETROLOGY OF ROWATT AND MUKPOLLO SANDSTONES

Sandy lithologies compose a major part of these two formations, the compositions ranging from almost pure dolarenites to quartz arenites in the Rowatt Formation, and consisting mainly of quartz arenites in the Mukpollo Formation. The majority of allochemical constituents in

the dolarenites are locally derived and details of these are provided in Chapter 5. The following discussion is therefore aimed at the terrigenous components of these rocks (modal analyses in Appendix C).

Mineralogy:

Mineralogically the terrigenous components are similar to sandstones in the Fairweather and McLeary Formations. Well-rounded monocrystalline quartz predominates over polycrystalline varieties by a factor of ten and more. Undulose extinction is common and mostly inherited. Of the feldspars, microcline and orthoclase predominate (usually less than 5%). Lithic components are represented mainly by chert fragments in both formations and have probably been derived from jasper beds in the Kipalu Formation. Heavy minerals are rare. Most of the quartz and feldspar was derived from plutonic and metamorphic terranes.

In the quartz arenites, syntaxial quartz overgrowths are common and represent the first stage of cementation (Pl.8, Fig.5); advanced stages of overgrowth have resulted in almost complete loss of porosity. Some quartz arenites contain minor amounts of late diagenetic calcite or dolomite.

The relative proportions of the various components are illustrated in Figs.40a & b. The terrigenous fraction is both mineralogically and texturally mature, even in samples containing less than 20% of this fraction (eg. high quartz/feldspar ratios, high monocrystalline/polycrystalline quartz ratios, and degree of roundness). Comparisons with other units of the Belcher Group are shown in Fig.40c. Quartz/feldspar and mono-/polycrystalline quartz ratios, sorting and roundness are similar in all groups (except in Omarolluk and Loaf samples). The

(a)

(b)

(c)

Figure 40. Composition plots (a & b) for terrigenous components (quartz, feldspar & lithic fragments) for sandstones from the Rowatt and Mukpollo Formations (n = No. samples; 350 points and more counted for each thin section). In (c) these compositions are compared with sandstones from other units (plotted as envelopes).

major differences in these groups can be attributed to changes in local source rocks/ the Fairweather and to a lesser extent the McLeary Formations contain volcanic rock fragments and some plagioclase derived from the Eskimo volcanic rocks. The Rowatt and Mukpollo Formations on the other hand contain chert fragments as the dominant lithic component. Modal analysis data is tabulated in Appendix C.

KIPALU FORMATION

This unit originally was named the Keepaloo Formation by Moore (1918) after exposures of ferruginous argillites shales and granular jaspers flanking Kipalu Inlet. It is from these rocks that Moore compiled the first report of possible microfossils in Precambrian rocks, although Hofmann (1972) now regards them as dubiofossils. LaBerge (1967) has also illustrated a number of spheroidal structures of possible algal origin from samples of Kipalu chert.

Most of the Kipalu Formation is poorly exposed. Sections in the Kasgalik Lake, Kipalu Inlet, Innetalling Island and Tukarak Island areas have been described in great detail by Young (1922) and also by Jackson (1960). In the present study three relatively complete sections were measured along eastern Tukarak and Innetalling Islands (Fig.41). A number of other areas were examined, but exposure was generally too poor to warrant detailed measurements. Total thicknesses range from 107 to 125 m and like the Mukpollo Formation, the Kipalu has a sheet-like geometry but of much greater known palinspastic extent (at least 8500 square km).

Three major rock types are present: laminated micrites, ferruginous laminated and massive argillites, and granular cherts and jaspers.

Figure 41. Measured sections of the Kipalu Formation from Innetalling Island (5) and Tukarak Island (35 & 38) and correlation of lithologic zones.

Some tuffs occur near the Kipalu-Flaherty contact. On the basis of lithology and the nature of bedding, six zones can be recognized (Fig.41).

Zones 1 and 3:

These consist of well bedded, highly fissile grey and green micrites (siderite, dolomite and serpentine) and micaceous red argillites (iron oxides predominantly as cryptocrystalline hematite). The beds range from 1 to 80 mm in thickness. Size grading in the silt-clay range coincides with a gradation from light to dark colours within individual beds. Rounded sand-sized quartz grains are rare. In some sections small pockets of ooids composed of dolomite and siderite, showing well preserved growth rings, occur within some of the thicker micrite beds.

Isolated lenticular, granule-bearing jasper pods occur near the base of Zone 1. Some of the granules show thin hematite rinds, possibly representing accretionary pisolite growth. They bear some resemblance to oncolites in cherts of the Gunflint Formation (Hofmann, 1969). Most of the granules appear to be texturally primary (there is no evidence of any relic carbonate) and compose a clast-supported framework that in places shows some evidence of microcrystalline chert rim cements and void-filling chalcedony.

Sedimentary structures are rare. Bedding surfaces on some frost-shattered blocks exhibit straight-crested and bifurcating ripples (Pl.8, Fig.6), small scours and one example of symmetrical interference ripples.

Zones 2 and 4:

The transition from well-bedded micrites and argillites of Zone 1 to interbedded jaspers and argillites of Zone 2 is relatively sharp. The argillites and micrites are noticeably less fissile, with somewhat wavy bedding, but compositionally they resemble rocks of Zone 1. A

characteristic feature of both Zones 2 and 4 is the lenticular and wavy bedded granular jaspers, up to 30 cm thick, that contain chert granules or peloids, oncolites, mud chips and a few detrital quartz grains. The framework is clast supported and cemented by chert and chalcedony. Patches of coarse siderite, dolomite, stilpnomelane and small amounts of minnesotaite occur sporadically, commonly enclosed by dark red granule-bearing jasper. Textural evidence points to a secondary origin for the carbonates where they replace chert and hematite.

Deformation structures include small fractures that displace beds one or two centimeters and are abruptly terminated by overlying beds, suggesting synsedimentary fracturing. Unlike later fractures, they have not been filled by quartz or carbonate.

Zone 5:

A predominance of thin bedded red-brown and grey argillites and a concomitant decrease in micrite lithologies make up this zone. In weathered outcrops the rocks appear steel-blue. Bed thicknesses range from 1 to 20 mm except near the top of the zone where they reach 300 mm. Colour and grain size grading are common; granule-bearing jasper is rare.

Zone 6:

The best exposure of this zone is found in Section 5 (Innetalling Island), where the sequence is characterized by bedded cherts, granule-bearing jaspers and thin ferruginous argillites. Magnetite and specular hematite are common. A few laminated siltstone and micrite beds containing intraclasts occur near the top of the Kipalu Formation. The upper 10 m consists of poorly bedded chert containing patches of coarsely crystalline carbonate. Some of the uppermost silty beds are tuffaceous, containing relic glass shards now replaced by a recrystallized quartz mosaic.

General Features:

The presence of grain size grading, planar fabrics imparted by micas and clast-supported frameworks in granule-bearing jaspers, clearly indicate that most of the Kipalu ironstones are of detrital origin. Mudstone rip-ups in the jasper provide evidence for some storm activity. A few low-amplitude ripples and shallow scours in the argillites indicate only rare instances of bed-load transport of sediments.

In addition, sediments of chemical origin are also represented: micrites now composed of siderite, dolomite and minor calcite; and ferruginous cherts. In granular jaspers the presence of sharp clast boundaries, two stages of chert rim cement and void-filling chalcedony, and the occurrence of chert grains in non-cherty lithologies, all suggest an early diagenetic origin for the Kipalu cherts. In the banded jasper zones, thin argillite beds enclosing the cherts show no evidence of chert replacement, the chert being confined solely to granular units. A primary origin for cherts having similar characteristics has been documented in several other iron formations (Gross, 1972; Zajac, 1974; Floran and Papike, 1975).

A different approach has been taken by Dimroth and Chauvel (1973) in their study of the Sokoman Formation. These authors have suggested that original calcareous sediments (probably aragonite) were converted to ferriferous chert during an early phase of diagenesis, in some cases even preceding the conversion of aragonite to calcite (based on oolite morphology). In the Kipalu jasper however, there is little evidence of any precursor carbonate, although many of the granules exhibit concentric hematite-rich rinds. The little carbonate that is

present appears to be a late diagenetic phase that replaced hematite and chert. It is interesting to note that adjacent micrite beds also show no evidence of replacement by silica.

To summarize, jasper lithologies are confined entirely to granule-bearing and peloidal beds. Both chert granules and cements and also cryptocrystalline hematite are texturally primary. Although these criteria strongly suggest a primary origin for the chert, early complete replacement of original carbonate fabrics cannot be ruled out.

Sedimentary Facies in the Kipalu Formation:

Based on the above sedimentological and petrological criteria, three principal depositional subfacies can be recognized:

Banded Argillite Subfacies:

The highly regular nature of bedding, presence of grading and the paucity of structures formed by bed-load transport, strongly suggests a suspension-dominated mechanism of deposition under low energy conditions. Minor reworking by weak bottom currents may also have occurred. The regularity of bed thickness in all sections might also indicate seasonal control of the supply of terrigenous detritus (quartz, feldspars, micas and clays). Deposition probably took place below wavebase, or within a protected embayment.

Micrite Subfacies:

The micrite beds represent periods of low clastic influx and high carbonate production. Isolated pockets of ooids within well-bedded argillites and micrites may have resulted from transport of oolite sands into deeper water. There are no sedimentary structures indicative of high-energy reworking that would normally characterize a shallower oolite-forming environment.

Jasper Subfacies:

In the Kipalu ironstones, chert granules, peloids and possible oncolites are preserved within chert cements, forming a grain-supported framework with good sorting characteristics. In addition, zones of mudstone rip-ups provide evidence of storm activity. A high-energy environment is suggested in which the cherts are compared to oolite-peloid banks or shoals. There is no evidence for subaerial exposure.

COMPARISON WITH OTHER MAJOR IRON FORMATIONS

The principal features of five major iron formations from northwest Australia, South Africa and the Canadian Shield are summarized in Table 4. Perhaps one of the most striking aspects of these and other Proterozoic iron formations is their great areal extent. For example in the Brockman Iron Formation, individual ironstone beds can be traced over many tens of kilometers. Although it is less impressive, the Kipalu Formation can be traced over most of the Belcher Islands, representing an area of about 8500 square km; on Tukarak Island zones a few meters thick have been traced at least 40 km parallel to strike. In most iron formations cross-bedding is rare, with exceptions being the Gunflint and Sokoman Formations. The Sokoman also differs from other successions in the variety of carbonate lithologies, or in many cases chert-after-carbonate. Whereas micrite-type lithologies compose only about 30% of the Sokoman Formation (Dimroth, 1977), the Brockman and Penge Formations are almost exclusively composed of this rock type. The Kipalu Formation contains upwards of 60% banded argillites and micrites, and together with the sedimentological features noted above, appears to most closely resemble the latter two iron formations.

	HAMERSLEY BASIN (Trendall, 1968)	ANIMIKIE BASIN (Goodwin, 1956; Hofmann, 1969)	TRANSVAAL BASIN (Beukes, 1973; Button, 1976)	LABRADOR TROUGH (Dimroth et al., 1970; Chauvel and Dimroth, 1974).	BELCHER BASIN (Young, 1922; Woodbridge, 1922; Jackson, 1960; Present study)
Thickness	915m in the Hamersley Group. About 600m constitutes the Brockman Iron Fm.	Gunflint Fm. 140-170m Other iron formations range from 0-600m. ⁺	Penge Fm. about 460m maximum. Kuruman Fm. about 750m.	Sokoman Fm. 120-210m.	Kipalu Fm. 107-125m.
Stratigraphic Continuity	Can correlate mesobands about one inch thick over 20,000 sq. miles.	Difficult to demonstrate stratigraphic continuity.	Can trace some units 40km and more.	The Southern Labrador Trough equivalent, Wabush Iron Fm. 305m.	Thickness relatively constant over the Belcher Islands (8500 ⁺ sq. km.). Probable correlatives along Hudson Bay, about 400km.
Sedimentary Structures	Cross-beds rare	Graded bedding and some cross-bedding.	Cross-bedding rare. Intraformational breccias.	Cross-bedding, lenticular bedding, scour channels, load casts.	Rare current and interference ripples. Graded bedding in banded argillites.
Microbanding	Conspicuous	Rare	Conspicuous	Present.	Conspicuous.
Stromatolites	—	Variety of stromatolites in Gunflint chert.	—	Pisolites and oncolites.	None observed.
Evaporites	—	—	No evidence for evaporites.	Length-slow chalcedony in intracrystals.	None observed.
Clastics	Not reported	Interbedded tuffs. Locally quartz grains.	Oolitic jaspers in Kuruman Fm.	Terrigenous clastics uncommon. Oolitic and peloidal cherts common.	Banded argillites and siltstones abundant. Rare fine grained sandstones. Oolitic and granular cherts common.
Piebeckite	Massive and fibrous.	Massive and fibrous.	Less well known.	—	Not common
Age	About 2.0 Ga.	Gunflint Fm. about 2.0Ga.	About 2.3 Ga.	1.9 Ga.	About 1.8 Ga.
Basin Size.	Ovoid basin with major axis about 480km.				Arcuate basin with major axis about 400 km. ⁺
Volcanic Association.	Underlain by Fortescue Group volcanics and interbedded with acid volcanic of Woongarra Fm. Shards in the Brockman Fm.	Obvious interstratification of tuffs and flows.	Lower Griquatown Iron Formation overlain by 1100m of Ongeluk Volcanics. Penge Fm. apparently not associated with volcanics.	Overlain by Menihek Slates and volcanics slightly higher in stratigraphic sequence. Underlain by shales and some tuffs of the Ruth Fm.	Overlain by Flaherty Fm. volcanics. Few thin tuffs in the uppermost Kipalu Fm.

TABLE 4. Principal features of some Proterozoic iron formations.

The association of iron formations with volcanics differs considerably from basin to basin, from those containing interstratified flows and volcanoclastics, for example the Gunflint Formation, to sequences having no obvious volcanic associations, as in the Penge Formation. In the case of the Kipalu ironstones, evidence of volcanism appears only at the top of the sequence and possibly heralds the approach of a migrating volcanic front.

ANALYSIS OF THE ROWATT-MUKPOLLO-KIPALU BASIN

Basis for interpretation:

The use of Walther's Law of Correlation of Facies has been discussed in detail by Middleton (1973). The underlying principle implicit in any application of this law is based on an actualistic approach to facies analysis; observation of relationships between modern sedimentary facies is essential. This leads to the main statement of Walther's Law,

'...only those facies and facies areas can be superimposed primarily which can be observed beside each other at the present time.' (Walther, 1894, in Middleton, 1973 p.982).

Application of this law to the interpretation of the Rowatt-Mukpollo-Kipalu succession is discussed briefly.

The Rowatt Formation has been subdivided into two members. The Lower Member, which succeeds relatively deep basin to subtidal deposits of the uppermost Laddie Formation, represents an intertidal clastic sheet with small isolated carbonate buildups near the top; the Upper Member consists of deformed dolostones, interpreted as a major carbonate buildup that accumulated in an upper intertidal to supratidal environment. The intertidal sheet is characterized by actively migrating sand bars or

dunes. The overlying Mukpollo Formation is also intertidal, and thus the interval from the upper Laddie to the Mukpollo represents a shoaling upwards sequence. However, interpretation of the overall nature of the Kipalu basin is hampered by a scarcity of diagnostic features in the ironstones. For example, in the argillite and micrite subfacies there is little indication of depth of deposition, and although it is reasonable to suggest that deposition took place below wave-base, it also could have occurred in much shallower water within a well-protected embayment. Nevertheless, the lateral continuity of this subfacies over several tens of km, together with an almost complete lack of erosional surfaces or cross-bedding, supports the conclusion of a relatively deep-water environment, probably within a restricted basin.

Two lines of field evidence provide support for this interpretation:

(i) Ironstone pebbles and chert fragments derived from the Kipalu Formation occur in channel-fill sandstones and bedded sandstones in the underlying Mukpollo and Rowatt Formations. Although thin lenses of ferruginous argillite occur at the base of the Fairweather Formation, pebbles are rarely observed in strata between the Fairweather and Rowatt Formations and it seems unlikely that the Mukpollo pebbles were derived from these rocks. A similar relationship is seen in the Rowatt Formation, where dolostone clasts up to boulder size (apparently derived from Upper Member rocks) are present in the Lower Member. Both of these examples provide good evidence for a sedimentary model based on the lateral equivalence of facies.

(ii) Conformable contacts exist between the above units; the Rowatt-Mukpollo contact is relatively sharp on eastern Tukarak Island, whereas the same boundary on Flaherty Island is less well defined,

being gradational over several meters. Nowhere is there evidence of an erosional break between these two formations.

Depositional History of the Rowatt-Mukpollo-Kipalu Basin:

The sedimentary model shown in Figs.42 and 43 depicts a shoaling-upwards sequence that is interpreted as having developed along a prograding shoreline, where the Kipalu ironstones represent a barred basin which was protected from the open ocean by a system of carbonate buildups and offshore sand bars (Ricketts, 1978).

The main buildup complex extended for more than 50 kilometers parallel to the ancient north-to northeast-trending shoreline (if palinspastic coordinates are fixed to the present foreland to the east). Probable correlatives of the three Belcher units include ironstones, arkoses, quartzites and stromatolitic dolostones and limestones belonging to the upper part of the Nastapoka Group (Richmond Gulf), the Hopewell Islands, and the upper part of the Long Island-Manitounuk Sound succession (Dimroth et al. 1970). Thus the entire buildup-bar complex may have extended for a distance of some 400 km and can be envisaged as a large arcuate basin that occupied an area approximately the same as that existing at present along the eastern coast of Hudson Bay.

The barred basin model is similar in some respects to one proposed for the Penge Iron Formation by Button (1976). In Button's model the iron formation was considered to be an end member in a series of chemical and sedimentary cycles. The closest modern analogy to the Belcher model, although only in gross physiographic setting rather than sediment types, may be the barrier island-lagoon complex bordering the north and west coasts of the Gulf of Mexico. The padre Island-Laguna Madre system for example, is 175 km long (Dickinson et al. 1972, p.193).

Figure 42. Basin model for the Rowatt, Mukpollo and Kipalu Formations and inferred relative sea level changes.

Figure 43. Reconstruction of paleoenvironments represented by the Rowatt-Mukpollo-Kipalu Formations.

A problem that arises in considering the Kipalu Formation as a barred basin is the source of the large volume of terrigenous detritus in the Rowatt and Mukpollo rocks. There is no evidence of channels or delta systems in the ironstone succession that might have acted as feeders for sands derived from the adjacent craton. This problem is overcome if the quartz-rich sands were reworked along the adjacent marine platform by north-flowing longshore currents rather than being first cycle products of erosion of cratonic rocks. Recycling of sediments from a modern continental shelf into a near-shore barrier island system, and beach and dune sands, has been documented along the southern coast of the United States (Giles and Pilkey, 1965; Pierce, 1969); these conclusions are based on heavy mineral distribution trends and estimates of sediment budgets respectively. Support for this in the present example is provided by high quartz/feldspar ratios in nearly all of the sands, not only in the Rowatt and Mukpollo Formations, but in all pre-Flaherty Formation units (Fig.40c). Even in dolarenite lithologies where the proportion of terrigenous material is 20% or less, modal analyses show the quartz/feldspar ratios to be very similar to ratios in rocks of orthoquartzite composition. In addition, the quartz grains invariably are highly rounded. The proportion of rock fragments is generally low, and detrital heavy minerals are rare in all of the sedimentary units. Mineralogically there is little difference between the terrigenous components in Rowatt and Mukpollo strata and rocks lower in the Belcher succession. Two alternative interpretations are considered on the basis of these results:

- the high quartz/feldspar ratios may reflect a feldspar-deficient source terrane. Because the feldspar is predominantly

microcline and probably is of granitic or metamorphic origin, this seems unlikely.

-the second and more likely explanation is that sands, derived initially from granitic and metamorphic terranes, have been subjected to prolonged and intense reworking. Loss of feldspar by mechanical abrasion and gradual winnowing of heavy minerals would result in relatively quartz-rich sands.

PLATE 7

- Figure 1. Cross-bedded quartz arenites and shaly interbeds in the Lower Member, Rowatt Formation. Section 15, on unnamed island south of McLeary Point (grid reference 425270).
- Figure 2. Large-scale current ripples (crests parallel to hammer) and smaller ripples superimposed at right angles in troughs. This type of structure can form as a result of late-stage run-off during ebb tides. Section 16 on unnamed island south of McLeary Point (grid reference 424264).
- Figure 3. Cross-bedded sandy dolarenites in the Lower Member, Rowatt Formation, showing complex ripple structures (arrow) and reactivation surfaces. Same locality as Figure 1.
- Figure 4. Bedded conglomerate composed of dololomite and grainstone boulders, overlain by dolarenites showing herringbone cross-beds. Same locality as Figure 1, in the upper part of the Lower Member, Rowatt Formation.
- Figure 5. Beachrock in top of the Lower Member, Rowatt Formation, eastern Flaherty Island (Section 23). At least three well-defined stages of beachrock lithification and erosion are visible (see text Figure 36). Pen is 15 cm long.

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PLATE 8

Figure 1. Folded dololutites in the Upper Member, Rowatt Formation.

Note brittle failure in several thick dololutite bands near the top. Weak cleavage occurs near the top of the outcrop. Pen is 15 cm long. Tukarak Island, Section 30.

Figure 2. Detail of the style of deformation in Rowatt dololutites.

Note changes in fold length with bed thickness and cement-filled cavities in folded and faulted zones. Scale is 2 cm wide. Northeast Innetalling Island (Section 34).

Figure 3. Fractured grainstones, Upper Member, Rowatt Formation.

Fractures generally have developed parallel and normal to bedding (here dipping to the right). Outcrop is on unnamed island south of McLeary Point, Section 15.

Figure 4. Channel-fill dolarenites with abundant trough cross-beds.

Upper Member, Rowatt Formation, northeast Tukarak Island (Section 30).

Figure 5. S.E.M. photomicrograph of quartz grain from Mukpollo quartz

arenites, showing advanced stages of secondary quartz overgrowth. Note growth striae on some crystal faces. Bar scale represents 0.1 mm.

Figure 6. Current ripples on the underside of a frost-shattered slab

of banded argillite, Kipalu Formation, Tukarak Island.

Pen is 15 cm long.

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CHAPTER 4

THE FLAHERTY, OMAROLLUK AND LOAF FORMATIONS

THE FLAHERTY FORMATION

INTRODUCTION

The Flaherty Formation, the "backbone" of the Belcher Islands, underlies about 60% of the area exposed in the islands and forms many of the narrow arcuate peninsulas and prominent ridges that are characteristic of the Belcher terrane. Flaherty strata are remarkably well exposed along some 2000 km of coastline (Fig.44).

Moore (1918) first described the rock units in the Belcher Islands, assigning the name Tookarak Diabase and Basalt to what are now called the Flaherty and Eskimo Formations and Haig Intrusion (Dimroth et al. 1970). These rocks were subsequently described by Young (1922), with a more detailed study being undertaken by P.E. Schenk (1959) who was a member of the survey party headed by G.D. Jackson (Jackson, 1960). Schenk inferred these lavas to be spilites.

An almost complete stratigraphic section on the east side of Eskimo Harbour was studied by Leggett (1974), who recognized three phases of eruption based on the relative proportions of volcanoclastic deposits, massive lavas and pillows:

- an initial weak explosive stage;
- an effusive stage;
- a combined effusive and explosive stage.

Chemical analyses of the Flaherty volcanics obtained by R.T. Bell and A.F. Stirbys (in Stirbys, 1975) indicate typical tholeiitic basalt composition rather than a spilitic assemblage. However, the term spilite has recently been revived by Ware (1978), who studied the Flaherty volcanoclastic deposits in detail.

Figure 44. Distribution of Flaherty Formation strata and localities of measured sections.

Age:

Whole rock K-Ar dating of pillow-rim and core samples has revealed two broad age groups (Hofmann and Jackson, 1969): (i) an older group, 1620-1693 m.y., that has been correlated with a metamorphic event during the Hudsonian Orogeny, and (ii) a younger group, 830-1054 m.y., that tentatively has been correlated with a mild thermal or tectonic event during the Grenville Orogeny. Recent whole rock Rb-Sr isochron determinations on a number of volcanic and shale samples from the Flaherty and Omarolluk Formations respectively, give ages (1.35 decay constant) of 1755 ± 69 m.y. and 1790 ± 89 m.y. with a composite isochron defining an age of 1798 ± 38 m.y. (Fryer, 1972).

Measured Sections:

Localities and details of sections measured during this study are shown in Figs. 44 and 45.

General Features of the Flaherty Formation:

Because of considerable lithologic variation, Dimroth et al. (1970) refrained from defining a type section. Nevertheless, it is useful here to define two reference sections:

(i) a continuously exposed 1000 m section seven km east of Sanikiluaq at the Katuk entrance to Eskimo Harbour;

(ii) a less complete but well-exposed 200 m section along the south shore of McLeary Point, east Tukarak Island.

Variations in thickness are considerable; Hofmann and Jackson (1969) quote a range of 290 to 1950 m. The thickest measured section in this study is 1650 m along west Flaherty Island (Fig. 45). There is a general thinning trend from west to east and hence eruption centers or fissures are presumed to have been west of the Belcher Islands. The

Figure 45. Details of stratigraphic sections through the Flaherty Formation. Section localities are shown in Fig. 44.

paleoslope thus imposed by building of a thick volcanic pile is the reverse of that interpreted for strata underlying the Flaherty Formation. Measurements of lava-flow structures and cross-bedding in volcanoclastics confirm an east-dipping paleoslope.

VOLCANIC COMPONENTS

Massive and Jointed Volcanic Flows

Massive and columnar-jointed basalt flows generally make up more than 50% of the Flaherty Formation. Along the western half of Flaherty Island, some individual flows attain a thickness of up to 145 m without any obvious break, but most are less than 30 meters thick. A marked thinning trend is observed eastwards towards Tukarak Island, although the relative proportion of flows remains constant (Table 5). Up to 25 flows have been counted in some sections. Flow-top and flow-base structures are abundant.

Columnar Joints:

These are present in most flows and in cross-section are typically polygonal. Individual columns may reach lengths of 30 m and where curved, inferred flow directions are to the east.

Flow Top Breccias

Most of these monomict breccias, up to five meters thick, grade into the underlying massive flows. Sorting is extremely poor, with clast size up to 80 cm across (Pl.9, Fig.1). There is no evidence of reworking, even though water-laid volcanoclastics overlies some of the breccias. Calcite cement is abundant.

Ropy Lavas

Examples of festooned ropy lavas are particularly well displayed

TABLE 5
Relative proportions of Flaherty flows and volcaniclastics

	20b	20d	26	West Tukarak	Laddie Harbour	27
Thickness (m)	1650	880	983	225	550	206
Flow Number	20	12	16	15	12	14
% Section	60	53	54	48	55	65
Pillows (%)	41	39	33	30	18	18
Massive Pyroclastics	5	8	9	7	7	8
Thinly Bedded Volcaniclastics	1	1	3	15	20	12
Total Volcaniclastics (%)	5	8	12	22	27	20

(West Tukarak and Laddie Harbour data from Ware, 1978)
For location of sections see Figure 44.

in shoreline exposures (Pl.9, Fig.2). These structures are typical of pahoehoe lavas (Wentworth and Macdonald, 1953). As might be expected, flow directions inferred from these structures exhibit considerable variation that is probably a result of local topographic irregularities in the original flow path. However, flow vectors corrected for folding indicate that flow was largely to the east.

Pahoehoe Toes

The structures illustrated in Pl.9, Fig.3 are almost identical to examples from Iceland (Jones, 1968) that developed during the 1961 eruption. Jones attributed such structures to the lobate advance of pahoehoe lavas.

Spiracles

Steam and gas produced when a flow overrides a wet substrate often results in the injection of pipe-like vesicles (Pl.9, Fig.4). Bent spiracles are generally regarded as being reliable indicators of flow direction of lavas (Waters, 1970), although Peterson and Hawkins (1972) have demonstrated that regional interpretations using these structures may not be conclusive, especially where advancing flows are influenced by surface irregularities. Peterson and Hawkins (ibid.) have also suggested that turbulent rather than laminar flow may be responsible for some of the variations observed in spiracle azimuths. Even so, spiracle bending in the Flaherty lavas is consistent with most other indicators of flow direction.

Amygdules

These are common in the upper half of many thick flows. They

reach sizes of 15 cm and are generally infilled with coarse calcite and large prehnite spherulites. In some flows, pipe vesicles project from the amygdule rim into the massive host rock, apparently recording a secondary stage of volcanic gas expansion.

PILLOW LAVAS

The second major constituent of the Flaherty volcanics are pillow lavas ranging in thickness from 5 to 230 m. Thick piles can be traced laterally for several hundred meters. Unlike massive flows, however, there is a regular decrease in the proportion and thickness of pillow lavas towards the east (Table 5).

Typical pillow forms are illustrated in Pl.9, Figs.5 and 6. Budding and constriction structures are common, and detailed examination indicates that isolated pillows are probably quite rare. In exposures revealing the three-dimensional geometries of pillows, the style of bifurcation and the lobate form strongly suggest flow mechanisms similar to those observed in pahoehoe lavas that have flowed into standing bodies of water, as has been described by Jones (1968, Iceland) and Moore et al. (1971, Mount Etna). Within large pillow piles, massive basalt lobes completely surrounded by pillows may represent lava channels that acted as feeders to the pillows. Apart from having vesicular rims, the lobes appear to be structureless.

Cooling structures such as radial fracture patterns and glassy palagonite-rich selvages are common (Pl.9, Fig.6). No systematic changes in vesicularity within pillows or flows was seen, either in individual stratigraphic sections or laterally.

Interpillow voids are commonly filled with graded clastic debris

(hyaloclastites) and coarse sparry calcite (P19, Fig. 6). Some radial fibrous fabrics are preserved in the calcite and there is a hint that botryoidal fibrous carbonate lined these cavities.

Top Determinations

The following structures provide valuable criteria for top determinations in massive and pillowed lavas of the Flaherty Formation:

Flow-top breccias

Ropy lavas

Flow-base spiracles

Graded hyaloclastic sediments interstitial to pillows

Bun-shaped pillows, convex surface up

Large elongate amygdules near the upper margin of pillows

VOLCANICLASTICS

As will be shown in the following pages, most of the volcanic sediments were deposited in a subaqueous environment, although some of the pyroclastics may have been of subaerial origin. Based on modes of deposition, two broad groups of pyroclastics are recognized: those that formed by fallout, and those that formed by flow. Whereas the first group contains only a single category of rock type, namely tuff and lapillistone, the second group can be subdivided further into turbidite flows and massive pyroclastic flows. In addition to pyroclastic lithologies, chemical precipitates also occur. The principal characteristics of each group are summarized in Table 6. Fisher's (1966) classification for indurated tephra is adopted here.

Tuffs and Lapillistones:

This category was termed subaqueous tephra by Ware (1978). These well-bedded pyroclastics consist entirely of ash and lapilli. The

TABLE 6

Principal characteristics of the major groups of volcanoclastic rocks in the Flaherty Formation.

	Tuffs & Lapillistones	Turbidites	Chemical Precipitates	Thick Pyroclastics
Thickness	5-200 cm	10-100 cm	Up To 20 cm	Up To 85 m
Grain Size	<10 mm	Coarse sand to silt	Micrite	<20 mm Few bombs
Grain Type	Ash, Lapilli, Accretionary lapilli, Pumice fragments, Plagioclase crystals	Ash, Pumice, Volcanic fragments, Plagioclase crystals	Calcite	Ash, Pumice, Plagioclase crystals
Sorting	Moderate	Poor	-	Poor to moderate
Framework	Grain supported	Matrix supported	-	Variable; open to intact
Sedimentary structures	Parallel bedding, Grading Tabular cross- bedding, Drape structures	Bouma cycles, Dewatering structures, Concretions, Channels	-	Parallel laminae, Density grading
Contacts	Sharp	Sharp Scoured tops	Sharp Scoured tops	Sharp Non-erosional

presence of small scours and planar cross-beds, some sets as thick as 30 cm (Pl.10, Fig.1), provides evidence for subaqueous deposition and also for some reworking of tephras by bed-load transport. In addition, many of these beds are sandwiched between pillowed lavas. Graded bedding is common and can be distinguished from turbidite deposits by the consistent lack of B to E divisions (of the Bouma cycle) and the presence of a grain-supported framework; the latter structure suggests deposition by fall-out from a vertical ash column (Pl.10, Fig.2).

The subaqueous tuffs are composed entirely of glass shards, pumice fragments and some broken feldspar crystals. The shards are typically highly angular with concave or straight margins and abundant vesicles (Pl.10, Figs.3 & 4). There has been relatively little deformation due to compaction.

Morphological variations in glass shards have been used to distinguish between two fundamentally different pyroclastic rock types, namely hyaloclastites and hyalotuffs; the more recent studies include those of Walker and Croasdale (1972), Heiken (1974) and Honnorez and Kirst (1975). The latter authors have defined these two rock types as follows: hyaloclastites are volcanoclastic rocks generated by non-explosive granulation of volcanic glass that is produced upon quenching of basaltic magmas; and hyalotuffs by phreatomagmatic and phreatic eruptions where ascending magmas form an explosive mixture on contact with ground water. Hyaloclastite debris is typically blocky, with low vesicularity resulting from rapid quenching, whereas hyalotuff fragments are more elongate, having concave surfaces and numerous apophyses resulting from broken vesicle walls. Although these differences are generally valid, the above authors hasten to point out that there is no

definite boundary between the two groups, and that both fragment types may be present in the same deposit.

Whereas explosive phreatic eruptions are probably confined to shallow water (depths less than 500 meters due to confining pressures: McBirney, 1963), the formation of hyaloclastite material is probably independent of depth of origin. Hyaloclastites have been reported from deep-sea deposits in the South Pacific (Bonatti, 1967) and have also been observed to form during the flow of pahoehoe lavas into seawater in Hawaii and Sicily (Moore et al. 1971; Moore et al. 1973). In many of the Flaherty deposits, vesicular varieties predominate and suggest that explosive phreatic-type eruptions, possibly analogous to the Surtseyan eruptions of 1963-64, were common; shallow-water eruptions are also implied.

Accretionary Lapilli:

Leggett (1974) and Ware (1978) have brought attention to well-preserved accretionary lapilli in several beds. In this study they were found in thin graded beds intercalated with well stratified tuffs and lapillistones. These beds are essentially matrix-free (being cemented by clear calcite) and are moderately sorted, and thus were probably deposited by fall-out rather than by turbidity currents. However, the presence of cross-beds in a few layers indicates some re-working in water. Typically, volcanic dust and shards surround a core consisting of a larger palagonite fragment, pumice fragment or devitrified glass (Pl.10, Fig.5). The cortex of a single lapillus may show concentric layering of ash and/or normal size grading of ash, with the outer rim consisting solely of accreted dust particles. These distinctive structures are thought to have formed in ash-charged clouds by accretion

of dust when very moist conditions prevailed, either during light rainfall or in steam-charged clouds (Moore and Peck, 1962; Lowe and Knauth, 1978). Accretionary lapilli generally are regarded as being indicative of subaerial eruptions, although deposition may eventually be either subaerial or subaqueous. Nonetheless, direct observations of subaqueous deposition have not been reported, and little is known about the preservation potential of such lapilli in water. Moore and Peck (1962) have suggested that accretion of ash possibly involves both viscous and electrostatic attraction, although it is likely that such forms would completely disaggregate upon entering water, especially if any form of turbulence exists. However, some lapilli have been observed to possess rigid outer shells and these forms could survive a subaqueous environment. In this case, an alternative mechanism of formation might be one involving accretion and welding of particles in a hot gas cloud: welding would not be so pervasive as to render the entire lapillus plastic, but would affect individual particles only.

In Lowe and Knauth's (1978) description of Archean accretionary lapilli from the Onverwacht Group, evidence for shallow intertidal to subaerial conditions was noted in adjacent strata. Similar supporting evidence is lacking for the Flaherty Formation. Thus while it can be inferred that some ash columns were erupted subaerially during Flaherty times, the probability of subaerial or subaqueous deposition cannot be stated with the same confidence.

Turbidites

As the proportion of volcanoclastic deposits in the Flaherty Formation increases eastwards (Table 6), so too does the proportion of sediments deposited by turbidity currents. Sections up to 20 m thick, consisting of ten or more individual and composite Bouma cycles, occur

on Tukarak Island. In most places only the A, B and C divisions are present, although a few complete cycles were seen. In the latter type the E division is represented by mudstones, some of which have been scoured or drawn out into flame structures by the overlying flow, or by light grey micritic limestones up to 20 cm thick (Pl.10, Fig.6). The association of lime muds with volcanic flows has also been reported by Kania (1929) and Carlisle (1963). Dewatering structures (Pl.11, Fig.1) and concretions present in some Flaherty beds are similar to many occurrences in the Omarolluk Formation. Like the Omarolluk examples, the dewatering structures here provide further evidence for rapid deposition of the volcanic sediments. The writer is not aware of any previously reported dewatering structures in volcanoclastic deposits.

All constituents in the turbidite deposits appear to be of volcanic origin, consisting mostly of shards, pumice and basalt fragments and crystals (Pl.11, Fig.2); craton-derived material is absent. The framework is typically open and is largely matrix supported. Shard- and pumice-clast morphologies are similar to those seen in most other Flaherty tephras, and exhibit a high degree of vesicularity (curved surfaces, broken vesicle walls). However, comminution of shards is more prevalent and is probably a result of abrasion during transport. Some accidental basalt fragments also occur*. Most of this clastic material probably originated from explosive phreatic eruptions. The turbidites invariably are sandwiched between units of massive or pillowed flows.

* 'Accidental' is a term used for volcanic ejecta derived from previously solidified magma, for example from the sides of a vent during an eruption, or from older volcanic or non-volcanic bedrock.

The volcanic ejecta probably accumulated rapidly to form unstable deposits on the submerged flanks of volcanoes. Downslope flow probably was induced by eruption tremors or by gravity. The facies changes observed in these thinly bedded volcanoclastic deposits are considered to represent the eastward transition from relatively shallow-water accumulations to deeper water turbidite deposits.

Thick Pyroclastic Flows

A number of thick, apparently massive, non-welded lapilli tuffs have been observed in sections throughout the Belcher Islands. One unit along west Flaherty Island attains a thickness of 85 m and can be traced more than 30 km along strike.

This same unit is also exposed east of Sanikiluaq (Fig.44). Detailed examination and 'walking out' of this bed has revealed no evidence of depositional or erosional breaks. Parallel laminae occur locally but cross-bedding is rare. Both upper and lower contacts are sharp and apparently non-erosional. A few basalt blocks or bombs occur in the lower few centimeters, although no drape structures were noted. No grain size grading is visible, but a crude form of density grading exists where the proportion of lapilli-sized pumice fragments increases towards the top of the bed. A distinctive wavy banded weathering pattern in the upper few meters is a result of late diagenetic silica replacement of shards, pumice fragments and earlier calcite cement, although the original clast-supported framework is still intact. A single lapilli tuff bed (16.5 m thick) on Bradbury Island (west Tukarak Island) and at McLeary Point, can be traced north into the Laddie Harbour area. Apart from a slight decrease in average grain size, this bed appears similar in all other respects to the thick unit on Flaherty Island, and a tentative correlation is therefore proposed.

Like most other bedded pyroclastic deposits in the Flaherty Formation, clasts show typical hyalotuff characteristics (Pl.11, Figs. 3 & 4). Plagioclase crystals and accidental basalt fragments are more common near the base of these flows. The framework ranges from matrix supported to clast supported with calcite cement.

The absence of depositional breaks suggests that the tuff bed was emplaced during a single event. An ignimbrite-type mechanism is discounted by the lack of welded textures. Pyroclastic flows up to 28 m thick have been described from lower Tertiary sequences in Washington (U.S.A.) and Japan (Fiske, 1963; Fiske and Matsuda, 1964) and have characteristics remarkably similar to the Flaherty examples, including homogeneity, an apparent lack of stratification, non-scoured basal contacts and a density grading of the pumice fragments. Fiske (1963) has attributed the formation of such beds to rapid accumulation of material from a submarine ash column, eventually giving rise to a single highly mobile flow. The predominance of highly vesicular pumice fragments and shards in the Flaherty unit is indicative of explosive phreatic eruptions which, as Honnorez and Kirst (1975) have noted, usually are restricted to shallow subaqueous environments. Here the pyroclastic flows are underlain by pillow lavas and hence these too most likely are of subaqueous origin. The well-defined bedding that is common to many air-fall tuffs is lacking in the Flaherty deposits and thus a depositional mechanism of this type can be ruled out.

The fact that thick beds consisting solely of volcanogenic debris can be traced laterally over considerably greater distances than any single volcanic flow or pillow unit, suggests a volcanic event of enormous magnitude. The area covered by the main pyroclastic bed is

at least 3000 square km. Assuming a wedge-shaped geometry, with thinning to the east, the total volume of erupted material is approximately 110 cubic km (this calculation takes into account only the area where the bed has been observed, plus a reduction correction of about 20% based on the porosity now occupied by calcite cement). This immense volume of material is about six times that produced during the 1883 eruption of Krakatoa (Holmes, 1970, pp.336-339). It also is interesting to note that the Krakatoan eruption produced a 60 m thick, poorly stratified bed of pyroclastic debris in three to four days (Williams, 1941, quoted in Bond, 1973, p.566), although this bed was probably of more limited areal extent than the Flaherty example.

A submarine flow mechanism has been suggested by both Leggett (1974) and Ware (1978), based on comparisons with examples cited by Fiske and Matsuda (1964). The absence of well-developed erosion structures at the base of the bed may indicate a highly mobile flow. This is supported to some extent by the rarity of bed forms such as cross-bedding and scour-and-fill which form under lower flow regime hydrodynamic states. On the other hand, high-energy plane-bed sediment transport is suggested by the presence of parallel laminae (ie. upper flow regime). Shallow-water phreatic eruptions probably supplied most of the fragmental debris in these flows.

PALEOSLOPE DETERMINATIONS

Based primarily on observed eastward thinning, Dimroth et al. (1970) inferred a volcanic source west of the Belcher Islands. This has been confirmed by Ware (1978) and in the present analysis (summarized in Fig.46). The relatively large spread of flow directions obtained from pahoehoe ropes and flow-base spiracles is probably due to local topographic

Figure 46. Summary of cross-bed measurements (CB) and flow top measurements (VF) from the Flaherty volcanics. After Ware(1978) and data from this study.

irregularities to which the flows would have conformed. Especially notable are directions obtained from northeast Innetalling Island. As noted previously, Peterson and Hawkins (1972) have cautioned against the use of flow-base spiracles for regional interpretations. However, cross-bed measurements, mostly from the C division of turbidites, do provide consistent trends and indicate that sediment transport was generally to the east. The regional paleoslope is almost the reverse of that inferred for strata underlying the Flaherty volcanics. This does not necessarily imply that the former southwest-dipping platform was tilted to the east, but rather that the paleoslope may have been imposed by accumulation of a thick volcanic pile. This is supported by the simple disconformable relationship of the volcanics with the underlying strata. Nevertheless, it seems likely that the Belcher basin had to subside fairly rapidly to allow accumulation of the thick succession of volcanic rocks. Appearance of the Flaherty volcanics is thus a harbinger of changes in tectonic environment affecting the Belcher basin.

DEPOSITIONAL ENVIRONMENT

Deposition of volcanics in a subaerial to shallow subaqueous environment was suggested by Leggett (1974) on the basis of a preponderance of massive flows, ropy lavas and accretionary lapilli. The presence of accretionary lapilli in some units provides evidence for subaerial eruption of ash columns, but does not indicate that deposition was subaerial. On the contrary, trough cross-bedding in many tephras suggests subaqueous deposition. The nature of fragmental components in the tephras also suggests relatively shallow-water eruptions; McBirney (1963) considered explosive eruptions at depths greater than

500 m unlikely for K-poor magmas. A general increase in water depths towards the east is inferred from the increased proportion of turbidite beds. There is no evidence for prolonged subaerial exposure, either in the form of desiccation structures or oxidation of ash deposits.

No changes in vesicularity were observed in any of the sections studied (some may be revealed on detailed mapping). A depth of 500 m was tentatively proposed by Ware (1978) by comparing vesicle sizes with examples from Iceland (Jones, 1969). Jones observed a positive correlation between depth of emplacement and vesicle size and abundance (ie. increased confining pressures with depth presumably results in fewer and smaller vesicles). On the other hand, Aumento (1971) has demonstrated that no such correlation exists in some mid-ocean ridge pillow lavas at water depths of 1000 to 3000 m. In this case the vesicle size is commonly in the range of 1 to 20 mm diameter. Flaherty vesicles are generally 1 to 5 mm in diameter.

Possibly water depths early in the history of volcanic accumulation were much shallower than those at the culmination of volcanism, especially in view of the more than 2000 meters of turbidites which overlie the Flaherty Formation.

PETROLOGY AND CHEMISTRY

The flow and pillowed volcanic rocks are composed entirely of medium- to fine-grained, olivine-free basalts and a few massive feldspar porphyries. Most samples examined contain euhedral, relatively fresh subporphyritic pyroxenes and subophitic plagioclase. Pyroxene compositions determined by Schenk (1959) indicate that magnesian augite varieties predominate and that plagioclase has a limited composition

range between Ab_{99} and Ab_{86} , averaging around Ab_{92} . Feldspars are usually cloudy. Chlorite is common, especially in vesicles. Skeletal magnetite and ilmenite replace feldspar and small anhedral sphenes are scattered throughout. Minor pyrite and chalcopyrite occur in some flows and tuffs. Calcite, quartz and anthroxolite (a dense hydrocarbon) are common as fracture- and interpillow-fillings. Calcite and prehnite are common in amygdules.

Chemical analyses undertaken by R.T. Bell and A.F. Stirbys (in Stirbys, 1975) are summarized in Table 7 and Fig.47. Like most of the Eskimo volcanics, the Flaherty samples plot in the tholeiite field on an AFM diagram. Values of Rb and Sr concentrations are typically tholeiitic (Table 8).

METAMORPHISM

Chlorite, sphene, sericite, prehnite and pumpellyite are all indicators of low-grade burial metamorphism. Prehnite is relatively common as spherulitic and bow-tie aggregates in volcanoclastic sediments, amygdules and veins. In some, crystals attain a size of up to 10 mm.

THE HAIG INTRUSIONS

Diabase sills and dykes have intruded all stratigraphic levels below the Omarolluk Formation. They are particularly prominent along west Tukarak Island where they form roches moutonnées. A 40-meter-thick sill occurs throughout the Belcher Islands near the Flaherty-Kipalu contact and often contains columnar jointing.

Petrographically the sills are similar to the Flaherty volcanics and consist of subophitic pyroxenes and plagioclase laths, and in places

TABLE 7

Major element compositions of basalts from the Belcher Islands, Labrador Trough and the Superior Province, before adjustment to 100%. Flaherty and Eskimo Formation samples were analysed by atomic absorption spectroscopy (volatiles not included).

	1	2	3	4
SiO ₂	50.40	49.00	48.80	49.20
Al ₂ O ₃	14.70	14.60	14.05	14.7
FeO	15.60*	13.30*	9.78	8.73
Fe ₂ O ₃	-	-	2.36	2.69
MgO	4.50	5.03	7.28	6.21
CaO	9.60	7.60	10.07	8.90
Na ₂ O	2.40	2.90	2.52	2.48
K ₂ O	0.96	0.30	0.22	0.47
TiO ₂	1.71	1.30	1.08	1.10
MnO	0.25	0.22	0.22	0.21
Total	100.12	94.25	96.38	94.69

* Total iron as FeO

- 1) Average Flaherty Formation basalt, 17 samples.
Volatiles not analysed for. Stirbys(1975).
- 2) Average Eskimo Formation basalt, 22 samples.
Volatiles not analysed for. Stirbys(1975).
- 3) Average normal lavas, Labrador Trough.
Dimroth et al.(1970).
- 4) Weighted mean composition of Superior basalts,
736 samples. Goodwin(1977).

TABLE 8

Rb and Sr contents of Flaherty volcanic rocks. Data from Fryer(1972)

	Range (ppm)	Average	Number of Samples
Rb	0.85-29.8	11.04	9
Sr	37.1-507	191.9	9

- Eskimo Formation
- + Flaherty Formation

Figure 47. AFM plot of Flaherty and Eskimo volcanic rocks

(after Stirbys, 1975). A = $\text{Na}_2\text{O} + \text{K}_2\text{O}$

F = $\text{FeO} + 0.8998\text{Fe}_2\text{O}_3$

M = MgO

Dividing line from Irvine and Baragar (1971). Most samples lie in the tholeiite field (Th).

are porphyritic with feldspar porphyroblasts up to two centimeters across. The metamorphic assemblage is similar to the Flaherty Formation.

Dimroth et al. (1970) have suggested that the diabase intrusions were probably coeval with extrusion of the Flaherty volcanics on the basis of the stratigraphic relationships noted above. However, Rb contents are considerably higher in the intrusions, suggesting alkalic affinities greater than is shown by most Flaherty rocks (see Table 1 in Fryer, 1972; also N.Arndt, pers. comm., 1978). Possibly the diabase represents a later phase of magmatism, being intruded after the bulk of the volcanic edifice had developed but before deposition of the Omarolluk turbidites.

DISCUSSION

The volcanic pile in the Eskimo Harbour area was considered by Leggett (1974) to represent three major eruptive phases: an initial weak explosive stage, an effusive stage, and a combined effusive-explosive stage. However, there is considerable variation in flow-, pillow- and volcanoclastic-facies both parallel and perpendicular to the regional paleoslope, and it is clear that this broad subdivision is not applicable to all sections of the Flaherty Formation. For example, effusive volcanism does predominate for a large part of the western Flaherty Island sections, whereas volcanoclastic deposits become increasingly important in the eastern sections. Nevertheless, even on Flaherty Island explosive-effusive cycles can be recognized, as is illustrated in Fig.48.

Although detailed facies and geochemical data are not available, some interesting, (albeit tentative) comparisons can be made with modern volcanic regimes. The percentage of tephra in island and continental

Figure 48. Sequential development of volcanic sediments and lavas of Flaherty Formation, western Flaherty Island. Bx = flow-top breccia, Hc = hyaloclastite tuff and breccia, Ht = bedded hyalotuffs.

Submarine Effusive Stage I: Pillows and hyaloclastite (glass and blocks that have spalled off pillows).

Effusive Stage II: Massive flows flanking and overlying pillows

Explosive Stage I: Subaqueous, bedded hyalotuffs with cross-beds and graded bedding.

Submarine Effusive Stage III: Pillows and flows.

volcanic arcs characteristically is very high (greater than 90% of the total volcanic products) in rocks ranging from tholeiites to calc-alkalic in composition. However, the tephra content of ocean basin volcanic rocks (tholeiites predominating) generally is less than 10% (Garcia, 1978, p.149; Macdonald, 1972, p.354). In the absence of abundant pyroclastics, paired metamorphic complexes or distinct calcalkaline trends in the major element analyses, a volcanic arc regime is considered unlikely for the Flaherty Formation.

With up to 27% pyroclastics in the Flaherty Formation, the volcanic sequence here compares more favourably with examples from modern oceanic islands such as Iceland and the Azores, both of which are emergent segments of the mid-Atlantic Ridge system. In these two regions both effusive and explosive eruptions have played an important role in the formation of their volcanic edifices (eg. Machado et al. 1962; Thorarinsson, 1967). A typical sequence of events includes: an initial explosive stage (phreatic and phreatomagmatic) which results in island formation; once access of the sea to the ascending magma is blocked off, an effusive stage prevails (Hawaiian- and Strombolian-type eruptions) resulting in massive flows and subsequent protection of the islands. Early in the history of eruption, pillowed lavas may also form a foundation upon which the pyroclastics and massive lavas are laid. In this way all three types of volcanic product can be represented in a vertical column of rock.

OMAROLLUK FORMATION

Regularly bedded greywackes, argillites and shales composing the Omarolluk Formation (2105 m) are exposed on Gilmour Peninsula and Young

Point, east Flaherty Island. The Gilmour Peninsula section has been designated the type section (Dimroth et al. 1970, p.110). Apart from these areas, exposure of the Omarolluk Formation is relatively poor in the Belcher Islands (Fig.49). However, good exposures do occur on the Bakers Dozen and King George Islands north of the Belchers, and have also been reported from several islands between the Belchers and Richmond Gulf (Dimroth et al. op cit.).

Contact with the underlying Flaherty volcanics is a sharp non-erosional disconformity that can be observed along a narrow strip of coastline between Gilmour Peninsula and Young Point. The lowermost fifteen meters of section consists predominantly of black pyritic shales, fine-grained volcanic sandstones and a few thin graded tuffs, the latter being composed almost entirely of volcanic ash, pumice fragments and rare grains of quartz and microcline. Some beds contain angular volcanic clasts up to four centimeters across. The transition from the shales to more typical greywackes is mostly covered.

The original definition of the Omarolluk-Loaf Formation boundary was based largely on the change in colour from grey arkosic wackes to red arkoses (Dimroth et al. 1970). Within this definition the Omarolluk contains strata typical of both turbidite and shallow-water fluviatile facies (facies interpretations are presented in the following paragraphs). Mapping during the present study indicates that a more logical basis for subdivision is the presence or absence of graded sandstone beds. The Omarolluk Formation is thus restricted to lithologies characteristic of turbidite facies. Hence, although the basal Loaf contact is also gradational on the basis of this criterion, it is more easily established in the field, and has more significance in terms of changes in depositional environments.

Figure 49. Distribution of the Omarolluk Formation and generalized stratigraphic column. Solid circles show island exposures mapped only by Jackson (1960); solid triangles represent areas observed in this study and also by Jackson.

Description of Lithofacies

Arenaceous and Pelitic Facies:

The Omarolluk Formation is characterized by extremely regular and tabular bedded greywackes, a few arkosic wackes and minor shales (Pl.12, Fig.1). Bedding in some areas can be traced more than 1000 m parallel to strike with little apparent change in thickness or content. Some individual sandstone beds grade laterally into distinct composite beds, some having shaly partings. The average bed thickness is 40 cm but can be as thick as 160 cm. The sandstone/shale thickness ratio is consistently about 10:1 over most of the exposed sections across Gilmour Peninsula and Young Point. Typical sections are illustrated in Fig.50.

Approximately 1000 m of well-bedded greywackes and shales crop out on Walton and Camp Islands, twenty five kilometers south of Gilmour Peninsula (Fig.49). A brief reconnaissance indicates that bed thickness here averages 20 to 30 cm and that the sandstone/shale ratio may be as low as 3:1.

A characteristic sequence of sedimentary structures can be recognized in most of the greywacke beds corresponding to the classic Bouma cycle (Fig.51; Pl.12, Fig.2):

- a lower graded division (A) consisting of medium to coarse sand-sized detritus. Most basal contacts are flat and non-erosional, but some contain shallow scours infilled with coarse sand and grit. Mudstone rip-ups are abundant. This division grades into:

- a parallel laminated division (B). In some beds individual laminae up to 10 mm thick display internal grading. Similar examples of multi-graded B division laminae have been reported from flysch sequences in the European Alps (Mutti and Ricci Luchi, 1978);

- Composite Bed
- Linguoid Ripples
- Climbing Ripples
- Water-Escape Pillars

(a)

1m

(b)

5m

(c)

Figure 50. Representative measured sections of the Omarolluk Formation, Gilmour Peninsula. Stratigraphic positions of these sections are shown in Fig. 49

Divisions

E

D

C

B

A

No grain
movementCurrent
Ripples

Dunes

Plane Bed

Standing Wave
AntidunesLower
Flow
RegimeUpper
Flow
Regime

Figure 51. Diagram of an ideal Bouma cycle (after Bouma, 1962) and hydrodynamic interpretations of each division (after Walker, 1967).

-a C division consisting of types C and A climbing ripples (terminology from Jopling and Walker, 1968). Some ripple foresets are oversteepened or folded in intricate convolutions (Pl.12, Fig.2);

-parallel laminated fine sandstones and argillites typical of the D division, and finally;

-a shale division representing 'normal' deposition from suspension. Small straight-crested and lunate current ripples are present in some of these layers.

The ubiquity and repetitive nature of Bouma cycles leaves little doubt that the Omarolluk Formation can be regarded as a thick turbidite sequence. Two principal facies are present: an arenaceous facies represented by individual turbidite beds (division A to D) and a pelitic or non-turbidite facies representing a return to normal, relatively deep-water sedimentation.

The complete Bouma sequence is not preserved in all beds. For example the C division may be absent, the resulting sequence being A,B, D,E. In most beds the E division is either absent, or poorly represented as shale partings or thin beds (most are less than five centimeters thick). The paucity of well-developed E divisions is attributed to a rapid succession of relatively coarse-grained turbidite flows. In some thin beds the A division is absent (Pl.12, Fig.2). In comparison, beds as thin as two centimeters may contain well-developed A-to-D or A-to-E transitions where the B and C divisions are missing.

Sole Markings:

Flutes, linguoid flute casts, parting lineations and a variety of groove casts are relatively common on exposed bedding soles and provide a means for establishing the direction of sediment transport of individual

flows (Pl.12, Fig.3). Most groove casts are parallel to flow directions inferred from flute casts. Groove casts may be simple or frondescent and some appear as a linear series of overlapping chatter or tool marks that have produced a chevron effect. Other tool marks such as skip casts are less common.

Paleocurrents

Azimuths of flute casts, groove casts, parting lineations and current ripples are summarized in Fig.52. The sense of flow, and thus sediment dispersal, have been determined from the flute casts to be in a south to southeast direction. Current ripple measurements taken mostly from the pelitic division overlying turbidites indicate flow east to southwest and exhibit, as expected, a wider spread of azimuths.

Conglomerate Facies:

Shallow pebble-filled channels occur at several levels in the lower 600 m of the Gilmour Peninsula section (Fig.49). Channels are rarely more than 120 cm thick, but range from 80 m wide to less than a meter, and in many instances are associated with sequences of thick greywacke beds (Pl.1, Fig.4). Channel fills consist of flat rounded pebbles of siltstone and calcilutite in a coarse quartz-feldspar matrix. An additional component is subspherical calcite concretions that have been reworked from nearby greywacke beds. Some flat lutite pebbles contain remnants of ripple cross-beds and parallel laminations, and in many instances these two resemble elongate concretionary structures that have developed in the C and D intervals of many greywacke beds (see discussion in following section).

Although no large-scale cross-bedding was observed in the channel deposits, a crude imbrication of flat pebbles indicates transport to

Figure 52. Summary of paleocurrent measurements from the Omarolluk Formation. Data grouped in 20° intervals and plotted as number percent.

the east and southeast. Pebble imbrication suggests deposition by traction currents rather than by grain flow or turbidity currents.

Facies Variations:

Although at first glance the Omarolluk Formation appears to be a rather monotonous succession of well-bedded greywackes and shales, a number of broad facies changes do occur both vertically and laterally:

a) Changes in bed thickness: In the Gilmour Peninsula and Young Point sections it is possible to recognize crude cycles of alternating 'thick' and 'thin' beds (Fig.53). Thick bedded intervals generally possess high sandstone/shale ratios and contain greywacke beds as thick as 160 cm. A number of these beds are composite where two or three units containing only A and B division sediments are superimposed, generally in erosional contact. Thin-bedded intervals (not exceeding two meters total thickness) contain mostly thin greywacke and shale beds with structures typical of C, D and E divisions. A few thin graded sandstones may be interbedded and in places beds more than one meter thick suddenly appear. These cycles can be interpreted in broad terms as representing successive periods of low sediment influx with suspension-deposited muds predominating, and intervals during which clastic detritus was rapidly introduced into the sedimentary basin by a series of high-energy turbidite flows.

No systematic change in bed thickness was observed towards the top of these sections. Lateral decreases in bed thickness and sandstone/shale ratio have been noted on Walton and Camp Islands, however. Examination of aerial photographs of islands east of Tukarak Island reveal few changes in overall bed geometry or thickness.

b) Channel Facies: This facies was only observed in the lower 600 m on Gilmour Peninsula.

Figure 53. Schematic representation of bed-thickness variations from Young Point. Total number of beds is 400, covering an exposed stratigraphic thickness of 154 meters. Heavy broken lines indicate breaks due to cover.

c) Representatives of the Omarolluk Formation in the Bakers Dozen Islands are important because they define a transitional zone between turbidite facies and shallow-water facies of the Loaf Formation. On Twin Cairns Island (Fig.49) both graded and massive arkosic wackes are present, intercalated with minor siltstone or shale beds. Intra-formational conglomerates are common. Large-scale festooned trough cross-beds and red mudstones that are typical of the Loaf Formation were not observed at this stratigraphic level.

Facies Interpretations

Studies of turbidite deposits in a number of modern marine basins have shown that the areal distribution of turbidites is largely controlled by a system of submarine slopes, fans and relatively deep basins. Sediments transported across an adjacent marine platform may be funneled down submarine canyons or channels, subsequently dispersed across abroad fans and eventually across a broad, flat deep basin plain. The areas covered and volumes of sediment involved in such complexes are enormous (for example the area covered by the Nile Cone is about 70,000 square km). Although many fan systems occur at depths of several hundreds of meters, bathymetric estimates in most ancient complexes are difficult to determine without the aid of paleontological data. Thus in this thesis, bathymetric terms such as "deep basin" are used in a relative sense only.

Mutti and Ricci Luchi (1978) have developed a model of turbidite facies and facies associations on the basis of the framework noted above in conjunction with studies of a large number of ancient turbidite and flysch successions. Specific associations of facies are used to define slope, fan and basin plain environments.

In comparison with many other turbidite successions, the Omarolluk Formation exhibits a relatively limited range of turbidite facies, the most prominent ones being characterized by parallel and laterally persistent bedding and alternating arenaceous-pelitic lithologies. Bouma cycles are well developed. This corresponds to Mutti and Ricci Luchi's 'C' Facies, or outer fan environment (and probably to Walker's (1967) Proximal Facies). Some intervals of relatively thick bedded greywackes and channel conglomerates may also represent middle to inner fan deposits. Extensive deep basin facies have not been recognized in these rocks although the southward thinning trend at Walton and Camp Islands may indicate that such a plain formed south of the Belcher Islands. In addition, large scale-slumping and thick pelitic intervals supposedly typical of slope deposits are not developed here. However, the channel facies may represent base-of-slope or inner-fan deposits.

To summarize, the Omarolluk Formation is considered to represent a submarine fan complex. Lateral basin equivalents to the fan such as deep plain or slope environments may be established by closer examination of exposures beyond the Belcher Islands.

Post-Depositional Structures in the Turbidites:

(i) Dewatering structures: In many Omarolluk sandstone beds, vertical to obliquely inclined pillars occur in sets that parallel bedding (Pl.12, Fig.5). Individual pillars range up to 15 cm in length and one or two mm wide. In vertical cross-section the pillars are straight to slightly sinuous. The lower boundaries of the pillar sets are somewhat irregular and diffuse, whereas upper boundaries sharply abut fine sandstone or mudstone layers. The structures are accentuated by a clean grain-supported framework resulting in a distinct

colour contrast with the surrounding drab-grey host rock. In plan view the pillars display circular sections and some are interconnected to produce vertical sheet-like structures similar to those reported by Laird (1970). The Omarolluk examples are interpreted as water-escape structures that probably developed during or soon after turbidite deposition. Pillars, or fluidization channels (Lowe, 1975), commonly form in medium- to fine-grained sands deposited at high sedimentation rates, particularly in beds possessing an upwards decrease in permeability. The graded and laminated divisions of turbidite deposits fall into this category. During initial stages of compaction (which begin immediately after deposition) grain settling displaces pore waters which subsequently migrate along paths of greatest permeability. Upward migration in normal graded beds is hampered by the increase in clay and silt content which results in greater cohesion. Development of localized confining pressures may give rise to channelized pore fluids that move fast enough to disturb the sediment framework and remove a large part of the matrix. A return to normal pore pressure allows a framework that is grain-supported to be established. This process probably takes place close to the sediment-water interface. In a few outcrops the bedding surface expression of the vertical water-escape pillars takes the form of miniature mud volcanoes (Pl.12, Fig.6) somewhat similar to structures illustrated by Dionne (1976), or the pit-and-mound structures of Shrock (1948, p.132-134).

(ii) Concretions: The history of compaction and pore-water chemistry of turbidite deposits is preserved in some remarkable carbonate concretions. Calcite concretions have formed in all five divisions of turbidite beds (McEwen, 1978). Sedimentary structures and

permeability differences in these divisions are reflected in the variety of growth forms. For example, large spheroidal concretions up to 20 cm across formed in graded and laminated sandstones where growth was essentially unconstrained by structural barriers or abrupt permeability changes. In this variety, mud chips and calcilutite clasts commonly acted as nuclei for concretion growth (Pl.13, Fig.2). In comparison, concretions that grew in the C,D or E divisions tend to be highly irregular or pod-like, particularly where they mimic convoluted and rippled bedding. In some beds they have coalesced into thin concretionary layers.

The Omarolluk concretions provide an excellent opportunity for establishing the time of diagenetic growth. Parallel laminae around which some of the concretions have formed converge away from the concretion margin, indicating at least 36% post-concretion loss in thickness due to compaction. It is obvious that growth of these concretions took place in relatively unconsolidated sediments. A further constraint on timing is realized by the observation that the water-escape pillars invariably pass through the concretions. Very early growth, probably syngenetic, is indicated in some of the channel conglomerates which contain large numbers of reworked concretions in addition to siltstone and lutite pebbles. Reworked concretions are distinguished from autochthonous varieties by abrupt grain size differences in comparison with the coarser conglomerate matrix, and the fact that they form an integral part of the clast-supported framework. Many calcilutite clasts appear similar to concretions that preserve laminations and ripples in other beds, and they may also be a product of reworking (Pl.13, Fig.2).

Raiswell (1976) has suggested that carbonate cements in pre-consolidation type concretions mimic primary porosity volume for volume. Values quoted by McEwen (1978) suggest porosities ranging from 37% to 42% in the medium-grained Omarolluk greywackes.

McEwen (1978) has attributed Omarolluk concretion growth to supersaturation with respect to calcite arising from diffusion of Ca^{2+} and CO_3^{2-} ions to sites of high pH, and is similar to mechanisms proposed by other workers (eg. Raiswell, 1971). However no satisfactory explanation for the existence of such pH gradients was forthcoming. The presence of calcite and pyrite concentration gradients within some Mesozoic concretions from England led Raiswell (1976) to suggest a correlation between bacterial reduction of SO_4^{2-} and production of HCO_3^- . In Raiswell's model the original distribution of bacterial colonies may have been heterogeneous, giving rise to isolated zones of calcite cementation. Although bacteria undoubtedly were present in middle Precambrian times, it is unlikely that colonies would have preferentially developed within the coarse-grained intervals of turbidite deposits. Furthermore, pyrite is not a common constituent of the Omarolluk concretions.

In most of the coarse-grained Omarolluk lithologies a strong spatial correlation between zones of water-escape structures and zones of concretions is apparent (Pl.13, Fig.1). This correlation also exists between concretions and dewatering structures in pyroclastic turbidite deposits of the Flaherty Formation. The influence of primary sedimentary fabrics on concretion growth has already been noted. In the case of coarse-grained sediments, most of the concretions occur below the B-C division contact which in many cases also corresponds to the upper limit

of the dewatering pillars (Pl.13, Fig.1). This contact, although gradational over a few millimeters, may have acted as a permeability barrier to rapid water expulsion. A situation like this might present an opportunity for initiation of localized concentration gradients as a result of early pore-fluid concentration. Here, the initial stages of Ca^{2+} and CO_3^{2-} concentration would be controlled to a large extent by pore-fluid migration following deposition, rather than by ionic diffusion. However, ionic diffusion resulting from local concentration gradients was probably the dominant process both during and after the early stages of calcite nucleation and precipitation.

LOAF FORMATION

Although the contact between the Omarolluk turbidites and red-grey arkoses that typify the Loaf Formation was not observed, intermediate lithologies such as grey arkosic wackes suggest that this contact is conformable and gradational. As noted in the preceding section, the boundary between these two formations is redefined, being placed at the last appearance of graded bedding. The Loaf thus contains red and grey arkoses containing abundant trough cross-bedding, red and grey siltstones and mudstones. The lowermost strata on Loaf Island (Fig.54) also contain massive beds with a few calcite concretions.

About 200 m of strata are exposed on Loaf Island in a broad syncline. Unfortunately, inclement weather prevented all but a cursory examination of these rocks. Typical sections, sedimentary structures and current direction determinations are summarized in Fig.54.

Tabular and lenticular beds up to three meters thick consist largely of arkoses interbedded with thin red and grey mudstones.

Festooned cross-bedding is particularly common, occurring in sets up to 50 cm thick. On bedding surfaces the largest sets have trough axes up to five meters in length. Some scour and fill structures contain basal pebble lags, consisting of intraformational mud chips and rare pebbles of chert, jasper, vein quartz, quartzite, granite and dacite. Exhumed scour troughs are commonly veneered by current-rippled mudstones. Soft-sediment deformation of cross-bedding is also common (Pl.13, Fig.3). Other forms of cross-bedding include tabular-planar sets up to 10 m in length, low-angle planar sets, herringbone sets and cross-laminated interference ripples. Some of the red mudstones contain large desiccation polygons (Pl.13, Fig.4). Parting lineations occur in parallel laminated sandstones.

Paleocurrent flow directions obtained by measuring trough cross-bed axes indicate that sediment transport was predominantly to the south and southeast. Large planar cross-beds show similar trends, whereas herringbone structures indicate minor transport to the east and west. There is no evidence of major current reversals anywhere in the Loaf succession.

At several localities, red and grey laminated siltstone beds are up to three meters thick and contain numerous lenticular sandstone ripples and mud flasers. Sharp angular discordances between sets of laminated siltstones can be seen in outcrops along the north shore and 1.6 km from the eastern end of Loaf Island. In adjacent beds, structures such as ball and pillow, recumbent folding of cross-stratification and detached load balls all suggest that synsedimentary deformation was relatively widespread. Deformation of this type is common in fluvial sequences (Allen and Banks, 1972).

The abundance of trough cross-bedding and associated synsedimentary deformation structures, together with strongly unimodal paleocurrent directions, all are indicative of a fluvial environment.

Fluvial Facies:

Analysis of specific fluvial environments requires a detailed knowledge of the vertical relationships between different groups of bedforms and lithologies. In the present study only broad groupings could be identified and hence the resulting environmental interpretations are tentative. The principal facies thought to be present are:

a) a trough cross-bedded sand facies. Spoon-shaped scour hollows that develop in front of and are subsequently filled by migrating dunes give rise to trough cross-beds. Some exhumed scour troughs contain small superimposed linguoid ripples in muddy veneers. This sequence represents two different flow regimes, the larger bedform possibly developing during a stage of river flood, and the smaller structures representing low-water conditions;

b) a planar cross-bed facies. This bedform develops upon migration of large-scale ripples or sand waves and is characteristic of linguoid bars in braided river systems (Miall, 1977). The direction of sediment transport indicated by foreset azimuths is similar to that of the trough cross-bed facies. In meandering river systems, large planar cross-bed sets develop in response to lateral migration of point bars during flooding. As a result, foreset azimuths may exhibit a wide range of values. The relatively narrow spread of values in the Loaf examples may indicate that braided river processes predominated;

c) a horizontal laminated sandstone facies, usually with parting lineations.

d) and a desiccated red mud facies. Desiccation prisms up to 20 cm thick occur. This facies typically overlies trough cross-bedded facies sands and may represent overbank deposits.

Miall (1977) has presented some facies criteria for distinguishing between braided and meandering river systems. Of the facies identified here, only the planar cross-bed facies appears to be diagnostic of braided systems. Schumm (1968) has speculated that the lack of plant cover during Precambrian times probably favoured high run-off rates and large floods, resulting in a predominance of braided river systems.

Shoaling Upwards Cycles:

Three crude shoaling upwards cycles have been identified in the Loaf Island sections (Fig.54). The lowermost strata in each cycle consist of medium-grained arkoses containing parallel laminated, low-angle planar and a few herringbone cross-beds. These pass upwards into laminated siltstones and finally trough cross-bedded arkoses and desiccated mudstones. The abundance of trough cross-bedding strongly suggests that shallow-water fluvial sedimentation predominated throughout deposition of the Loaf Formation. However, structures such as herringbone cross-beds and flaser and lenticular bedding may indicate a weak tidal influence in a marginal marine setting. The Loaf cycles are somewhat similar to alternating marine and non-marine regression cycles in the Devonian Catskill Formation, Pennsylvania (termed 'motifs' by Walker, 1971). In the Catskill example, distinction between marine and non-marine environments is based largely on the appearance of rootlets and the disappearance of crinoids and brachiopods. In the Loaf Formation this distinction is based entirely on the vertical association of sedimentary structures. This interpretation is reasonable because the transition

from relatively deep water turbidites to fluvial deposits requires the depositional environment to pass through a strandline intermediate between these facies. The existence of regressive cycles within the Loaf also suggests that the shoreline was prograding.

PETROLOGY AND PROVENANCE OF THE OMAROLLUK AND LOAF FORMATIONS

Because of the transitional nature of the turbidite and fluvial-marginal marine facies, the petrology of these two units is treated under a single heading. In addition, initial examination reveals similar mineralogies, although proportions of the major constituents (quartz, feldspars and rock fragments) differ significantly.

Greywackes and Arkosic Wackes:

These rocks are typically poorly sorted with matrix contents ranging from 15% to 40%. The frameworks consist predominantly of angular to subrounded quartz, feldspar and volcanic fragments. Both textural and compositional criteria indicate a high degree of immaturity.

Arkoses:

Compared to the wackes, the red and grey arkoses exhibit higher degrees of maturity, containing more quartz than feldspar, generally less than 10% matrix and hence possess grain-supported frameworks. However grain size and angularity differ little from the wackes, and according to Folk's (1974) classification, the arkoses also are immature (grain size, angularity parameters and maturity index are those of Folk, 1974).

Modal analyses are summarized in Fig.55 and Appendix C. Eight samples from arkoses of the Pachi Group and Richmond Gulf Formation (Manitounuk Supergroup) are included for comparison.

Figure 55. Triangular composition plots of rocks from the Omarolluk and Loaf Formations. Pachi Group and Richmond Gulf Formation samples are included for comparison. Column shows relative stratigraphic position of each sample. Q = quartz, F = total feldspar, R = total lithic fragments: Ksp = K feldspar, Pl = plagioclase, VR = volcanic rock fragments.

Quartz

Monocrystalline quartz predominates over polycrystalline varieties by a factor of five and greater. Most grains exhibit both undulose extinction and polygonization. Much of this strain probably inherited, there being little evidence of recrystallization in the matrix. Grains are clear and contain sporadic tourmaline and/or zircon inclusions.

Polycrystalline grains consist of equant mosaics and elongate mosaics of subcrystals. Boundaries between subcrystals are commonly sutured and lined with chlorite or iron oxides.

Features such as inherited strain and polycrystallinity suggest that most of the quartz was derived from plutonic and metamorphic terranes. No quartz phenocrysts of obvious volcanic origin were observed and moreover, the most obvious volcanic source, the Flaherty Formation, is not known to contain free quartz other than minor vein quartz.

Feldspars Thin sections were stained with sodium cobaltinitrate to facilitate the distinction among potash feldspar, plagioclase and quartz. Potash Feldspar: Untwinned orthoclase and microcline perthites are common in all samples throughout the entire range of grain sizes from coarse sand to silt. They are generally subordinate to quartz but in one unusual sample (#255) as much as 42% was noted. A few granophyric quartz-feldspar intergrowths occur. There seems to be little doubt that most of these feldspars were derived from a source area similar to that noted above.

Plagioclase: Albite and Carlsbad twinning are common. Untwinned varieties can be distinguished from quartz by the presence of cleavage (accentuated by etching during staining). Although most show advanced sericite alteration, twinning is sufficiently preserved in some grains

to allow determination of extinction angles that suggest sodic compositions similar to the Flaherty volcanics. However, in one or two samples, orthoclase-plagioclase aggregates indicate that some of the plagioclase probably came from a plutonic source.

Rock Fragments Two main types occur. Abundant volcanic fragments containing feldspar laths and microlites, some of which have distinct swallow-tailed quench textures, are similar to many of the Flaherty basalts. Some fragments have undergone silicification and appear chert-like. These can be distinguished from sedimentary cherts by the presence of relic feldspar laths.

Chert and siltstone fragments are subordinate to volcanic varieties. A few quartz-microcline aggregates of probable plutonic origin were also noted.

Heavy Minerals Omarolluk and Loaf deposits rarely contain more than 5% heavy minerals and most samples contain less than 2%. They include muscovite, subrounded zircon, sphene, epidote and pyroxene and rare aggregates of fibrous actinolite. In some Loaf samples, muscovite-quartz aggregates of metamorphic origin occur.

Sphene and pyroxene are both present in the Flaherty volcanics. The remaining mineral species could have been derived from a variety of igneous and metamorphic rocks.

Matrix As Pettijohn et al. (1973) have pointed out, the distinction between framework and matrix in greywackes is quite arbitrary. Henderson (1975) has been followed in this study, placing the boundary at 0.02 mm; no distinction has been made between the primary detrital and diagenetic components of the matrix. These components include quartz, feldspar, iron oxides, chlorite, sericite and minor chert. In concretionary

zones, the calcite largely occupies initial pore space but also has replaced some matrix. Thin calcite reaction rims are present on most framework grains.

To summarize, two distinct source terranes supplied sediment to the turbidite and fluvial depositional basin, namely volcanic and plutonic-metamorphic. The most obvious volcanic terrane is the thick sequence of tholeiitic basalts composing the Flaherty Formation. A similar source was suggested by Jackson (in Dimroth *et al.* 1970), Fryer (1972) and Barrett (1975). The second source was probably cratonic, and based on paleocurrent considerations, it appears to have been north of the Belcher basin.

Stratigraphic Trends:

Triangular plots of quartz-feldspar-rock fragment and K feldspar-plagioclase-volcanic rock fragment compositions illustrate the following broad stratigraphic trends (Fig.55):

- a gradual decrease in the amount of plagioclase and volcanic fragments towards the Loaf Formation arkoses;
- a concomitant increase in potash feldspar;
- a tendency towards cleaner sandstones with less matrix.

In these plots the Loaf Formation samples form a relatively tight cluster compared to those from the Omarolluk. This reflects the two fundamentally different depositional environments that these rock units represent, and to some extent the differences in availability of sediment derived from the two principal source areas. Sediments composing the Loaf Formation were mainly derived from a cratonic source to the north of the Belcher Islands and transported southwards. The turbidite basin on the other hand received sediments from both volcanic and cratonic terranes. In most cases mixing of these two constituents took

place before resedimentation as turbidite flows. However, some turbidite beds contain very little volcanic debris and exhibit distinct Loaf characteristics (eg. sample #255 from the middle of the Omarolluk Formation). The degree of mixing of the two major components appears to have been quite variable and may be related to periodic fluctuations in erosion of, and runoff from, both volcanic and cratonic terranes.

Two samples (#455 and 456) were taken from the top of the Omarolluk Formation at Twin Cairns Island. On sedimentological grounds, this section is inferred to be transitional between the turbidite and fluviatile facies. Petrographically, both samples exhibit distinct Loaf characteristics.

SUMMARY

Almost 2000 m of tholeiitic basalt flows, pillowed lavas and volcanoclastic sediments compose the Flaherty Formation. A general eastwards-thinning trend, flow-base and flow-top structures and cross-bed measurements indicate emplacement of the volcanics and associated sediments from a source somewhere west of the Belcher Islands. An increase in the proportion of turbidite flows in the eastern sections is thought to reflect deepening water in this direction. The Flaherty Formation is disconformably overlain by a 2000 meter-thick turbidite sequence consisting of proximal and distal fan facies but containing little evidence of widespread slope facies. A general southward thinning trend from the Tukarak Island area, concomitant with decreases in sandstone/shale ratios, may reflect lateral transition into a deep-basin facies. The principal sediment transport directions were to the south and southeast. Contact with the overlying Loaf Formation red beds

is gradational. These shallow-water deposits represent mostly fluvial environments with a weak marginal marine influence. River flow was predominantly to the south and southeast.

A Basin Model:

The trend of the turbidite basin axis was approximately south to southeast. Criteria leading to this conclusion are:

- paleocurrent indicators show that sediment transport was largely in this direction;

- a general southward thinning trend of beds and a decrease in the sandstone/shale ratio.

Sediment was supplied to the basin from a volcanic terrane bordering the western margin of the basin, and from a cratonic terrane to the north. In most of the turbidite flows, mixing of components from these two major sources took place before resedimentation. Hence with current directions largely to the southeast, most flows in the Omarolluk probably are representative of longitudinal basin flow. However, some indication of lateral supply can be seen in a few east-trending pebble-filled channels.

Sediment dispersal patterns in some modern marine basins have been illustrated by Stanley and Unrug (1972). The northern Adriatic Sea for example is dominated by longitudinal supply from the Po Delta with only subordinate lateral supply. Stanley and Unrug (op. cit.) also have noted the close relationship between deltas and base-of-slope deposits, a relationship that is easily overlooked in studies of ancient flysch sequences. A number of modern deep basins are in fact fed by such river systems, some particularly prominent examples being the Bengal Fan (Curry and Moore, 1971) and the Nile Cone (Maldonado and Stanley, 1978). A similar depositional regime is envisaged for the Omarolluk-Loaf

sequence. In this model, submarine fan and fluviatile-marginal marine facies are regarded as being laterally associated. Craton-derived sediments were introduced into the basin via a braided river system. Continued supply resulted in southward progradation of a braided fan (delta?) - submarine fan complex, with simultaneous supply of volcanic detritus from a volcanic edifice to the west.

PLATE 9

- Figure 1. Monomict flow-top breccia in contact (arrow) with overlying basalt flow. Basalt clasts are typically angular and poorly sorted. Minor calcite cement occurs below arrow. Pen (left center) is 15 centimeters long. Flaherty Formation, west Flaherty Island (Section 20d).
- Figure 2. Ropy festoons on lava flow top. Flow direction here was from upper left to lower right. Flaherty Formation, McLeary Point, Tukarak Island, Section 17.
- Figure 3. Possible pahoehoe toes formed by the lobate advance of lavas. Lava ropes developed around these structures exhibit highly variable trends. Flow towards observer. Flaherty Formation, south shore of Laddie Harbour, Tukarak Island. Width of area viewed is 150 cm.
- Figure 4. Flow-base spiracles. An amygdule developed near the contact with underlying shales shows radial arrangement of spiracles representing a later phase of gas expansion (arrow). Flaherty Formation, on unnamed island 2.1 km southwest of old Hudson Bay Outpost, Turakak Island. Bar scale represents 4 centimeters.
- Figure 5. Oblique section through pillowed lavas showing numerous constrictions and vesicular selvages. Central pillow is 140 centimeters across. Flaherty Formation, same locality as Fig.4.
- Figure 6. Geopetal structure in interpillow void: graded hyaloclastites are overlain by coarse calcite. Note vesicular selvages. Flaherty Formation, west Flaherty Island, Section 20a.

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PLATE 10

- Figure 1. Cross-bedding in hyalotuffs, Flaherty Formation. Pen is 15 cm long. West Flaherty Island, Section 20b.
- Figure 2. Photomicrograph of suspension deposited, well-bedded hyalotuff showing bimodal grain size distribution of pumice fragments and shards, typical grain-supported framework and sparry calcite cement. Most pumice fragments contain spherical vesicles. One large fragment (arrow) contains stretched vesicles. Bar scale represents 1 mm. Plane polarized light.
- Figure 3 & 4. S.E.M. photomicrographs of glass shards showing characteristic bubble-wall structures. Granular and fibrous material adhering to the surfaces is largely chlorite, serpentine and some prehnite. Specimens obtained by leaching from a calcite-cemented tuff. Bar scale represents 0.1 mm.
- Figure 5. Photomicrographs of accretionary lapilli (arrows) in grain-supported, matrix-free hyalotuff. In places, calcite cement has partially replaced some grain boundaries producing an apparent dispersed framework. Note absence of deformation due to compaction. Bar scale represents 1 mm. Plane polarized light.
- Figure 6. Sequence of pyroclastic turbidite flows in Flaherty Formation, McLeary Point. Each flow is overlain by a thin bed of lime mudstone (white) equivalent to the E division of a Bouma cycle. Hammer (left) for scale.

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PLATE 11

Figure 1. Dewatering pillars up to 15 cm long (in a zone adjacent to pen), in pyroclastic turbidites in the Flaherty Formation. Note convoluted laminae and load structures (arrows) below this zone. Bradbury Island, grid reference 313334.

Figure 2. Photomicrograph of poorly sorted shards and pumice fragments and a few plagioclase crystals (arrows) from a turbidite bed, Flaherty Formation. Compared to suspension-deposited hyalotuffs, these deposits contain a higher proportion of dark matrix and only minor calcite cement. Bar scale represents 0.5 mm. Plane polarized light.

Figure 3. Photomicrograph of a tuff from the massive pyroclastic bed, west Flaherty Island (Section 20d). Note poor sorting and largely matrix-supported framework. Bar scale represents 1 mm. Plane polarized light.

Figure 4. Same as Figure 3 but from a distal section of the massive pyroclastic bed. Framework is matrix supported. The shards are elongate parallel to bedding. Bar scale represents 1 mm. Plane polarized light.

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PLATE 12

- Figure 1. Thick graded greywacke beds typical of the Omarolluk Formation. In this section shales occur only as thin beds or partings. The bed on the right is composite unit more than 3 m thick. North shore of Gilmour Peninsula, Section 18.
- Figure 2. Well-developed Bouma cycles, Omarolluk Formation; (1) A-E, (2) B-E, (3) C-E, (4) B-E. Note flame structures, climbing ripples and convolutions in the C divisions. Bed (2) displays a small deformed scour channel. Scale at left is 6 cm long. North shore of Gilmour Peninsula.
- Figure 3. Flute casts, Omarolluk Formation. Recorded sense of flow is from left to right. North shore of Gilmour Peninsula, Section 18.
- Figure 4. Flat-pebble conglomerate filling channel, overlain by graded greywacke beds up to 120 cm thick. Omarolluk Formation, north shore of Gilmour Peninsula.
- Figure 5. Vertical section through zones of water-escape pillars in greywacke bed. Sharp upper surfaces bounding the sets of pillars probably resulted from abrupt differences in permeability in the parallel laminated deposits. North shore of Gilmour Peninsula, Section 18.
- Figure 6. Bedding surface showing pit-and-mound structures, Omarolluk Formation. Mounds up to 15 mm in height are pock-marked with distinct craters. These craters probably formed by collapse of a fluid-supported framework following cessation of flow of the pore fluids. The mounds formed at the sediment-water interface and were protected from subsequent turbidity flows by a veneer of muds. Scale is 9 cm long. North shore, Young Point.

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PLATE 13

- Figure 1. Oblique section through Omarolluk turbidite beds showing association of concretions (recessive weathering) and zones of water-escape pillars. Traces of dewatering structures pass through the concretions. Note the two principal sets of glacial striae nearly at right angles on surface. Northeast shore of Gilmour Peninsula.
- Figure 2. Water-escape pillars and calcite concretions terminated at the contact between B and D divisions of a turbidite bed, Omarolluk Formation. Concretion on the right (10 cm across) contains a reworked clast of black siltstone containing a truncated calcite concretion.
- North shore of Gilmour Peninsula, Section 18.
- Figure 3. Ball and pillow structures in Loaf Formation arkoses and interbedded mudstones. Northeast shore of Loaf Island.
- Figure 4. Large desiccation polygons in red mudstones, Loaf Formation. Pen is 15 cm long. North-central shore of Loaf Island.

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CHAPTER 5

PETROLOGY OF DOLOSTONES FROM THE BELCHER GROUP

INTRODUCTION

At least 95% of the more than 3000 m total thickness of carbonates in the Belcher Group are composed of dolomite. These rocks represent deposition in a wide variety of paleoenvironments, including shallow-marine platforms ranging from supratidal to below storm wave-base, deeper water slopes and areas of extensive carbonate buildups. This chapter provides an overview of the diagenetic history of these rocks based on fabric criteria and depositional history outlined in previous chapters.

The present discussion concentrates on four major rock types:

- (i) pisolitic caliche,
- (ii) detrital dolostones including beachrock, beach rosettes, and grainstones and dololutes in carbonate buildups of the Rowatt Formation,
- (iii) stromatolitic dolostones,
- (iv) deeper water carbonate mudrocks.

PISOLITIC CALICHE

In the Lower Member of the Fairweather Formation, eighteen shoaling-upwards cycles have been identified, each representing the transition from an intertidal to supratidal environment. The culmination of each cycle is seen in the development of pisolitic dolostones or caliche beds. The caliche beds are thought to have formed in an upper supratidal environment that was periodically flooded by high tides or storm surges.

These rocks have undergone pervasive dolomitization and yet many of the primary textural details are preserved. Examination of these textures provides valuable information for establishing the relative timing of the diagenetic events.

Pisolites:

The Fairweather pisolites are composed entirely of micritic dolomite and microspar (Pl.14, Fig.1). Grain size of the dolomite appears to be laminae-specific; concentric laminae composed of microspar alternate with micritic laminae, although there is a general tendency for grain size to decrease towards the outer margins of the pisolites. Contacts between these two forms of dolomite are sharpest near the margins, becoming more diffuse inwards where the dolomite appears to have grown by aggrading neomorphism. As shown in Pl.14, Fig.1, some of the concentrically arranged laminae overlap to form numerous microdisconformable contacts. Laminae that are continuous around individual pisolites are in fact quite rare. As noted in previous discussions, lamination growth may be asymmetrical, and composite pisolites enveloped in a single layer are also common. Downwards (geotropic) elongation and subsequent thickening of layers towards the base is probably the result of gravity induced cementation. Pisolite cores consist of detrital quartz grains and micritic fragments, the latter being derived by brecciation of early pisolites or micrite beds.

Grain-size differences within pisolites may reflect differences in the original carbonate fabrics. For example in modern cave and mine ooids, layers of radial fibrous calcite alternate with thin dark layers of micrite (Bathurst, 1976, p.306). Dolomitization conceivably could produce fabric changes in the fibrous layers resulting in more equant

microspar mosaics. Scholle and Kinsman (1974) on the other hand described an occurrence of aragonitic pisolites in caliche beds formed in Pleistocene limestone in close proximity to a modern sabkha flat. Although textural evidence here strongly suggests vadose zone cementation, the ubiquity of aragonite and high-Mg calcite cements are more indicative of hypersaline conditions. These authors conclude that hypersaline vadose conditions prevailed. The Fairweather caliche beds are also considered to have formed in a supratidal environment, although there is no evidence to suggest that paleosalinities were elevated to the degree seen in modern sabkha environments.

Brecciated Zones:

The pedogenic processes leading to the formation of caliche deposits are characterized by multiple dissolution-precipitation events. This results in solution brecciation of previously lithified segments of the crust and formation of voids. In the Fairweather examples, the debris resulting from brecciation includes micrite and microspar fragments, some of which are irregularly laminated and probably were derived from earlier pisolites. Grain size in these fragments may be as fine as 5 μm , indicating that recrystallization was relatively minor. Detrital quartz grains are also scattered throughout. Size sorting of this debris is poor. Examples where pisolites have grown around angular micrite fragments illustrate the continuous nature of the dissolution-precipitation process (Pl.14, Fig.1).

Void-Filling Cements:

Probably one of the most important aspects of the brecciated zones is the relationship between fragmental components and void-filling cements; most contacts between cements and grains are sharp. Thus it is

possible that dolomitization was an early event in the diagenetic history of these rocks. Cement-filled cavities also occur between adjacent pisolites.

Examples of void-filling cements are shown in Pl.14, Figs.1 & 2. In the first, an initial dolomite rim cement is probably a replacement of an earlier fibrous calcite, possibly high-Mg calcite. An original aragonite cement is considered unlikely because the aragonite to low-Mg calcite transformation usually results in fabric changes. However, the replacement of high-Mg calcite by low-Mg calcite, and even high-Mg calcite by dolomite without substantial changes in fabric have been documented in Recent and Pleistocene carbonates (eg. Gavish and Friedman, 1969; Muller et al. 1973). In the latter example, original high-Mg calcite cement in beachrock was replaced by dolomite following repeated lowering of sea level and introduction of brines having high Mg/Ca ratios. In the present example, the relic fibrous cement contains abundant brown inclusions whereas the second generation of dolomite cement is inclusion-free and is considered to be primary. Contacts between dolomite crystals are commonly planar and show little evidence of recrystallization. In some samples intercrystalline triple junctions are enfacial (where one of the three angles is 180°) which, as Bathurst (1975, p.425) has noted, is probably one of the more important criteria for distinguishing between cement and neomorphic spar.

In the above examples, three stages of void-filling cement are recognized: (i) fibrous dolomite (relic fibrous calcite); (ii) clear rhombohedral drusy dolomite, and (iii) coarse sparry calcite. There is no evidence of reaction between the third generation calcite and second generation dolomite cements.

Three quite different phases of cements occur where an initial rhombohedral dolomite lining was followed by coarse barite. The barite is poikilotopically enclosed in a later equant quartz mosaic. Several other cavities exhibit similar progressive barite replacement. Less common patches of non-void filling barite typically contain abundant dolomite inclusions. Possibly the formation of voids was partially controlled by the dissolution of earlier barite.

Discussion:

Macrotextural evidence of alternating dissolution-precipitation processes in the caliche beds is complimented by well-preserved multi-generation void-filling cements. The variability in cement type, even within a single hand specimen, indicates that pore water chemistry was continually changing. For example the sequence, fibrous dolomite (relic calcite)-drusy dolomite - calcite could have developed in response to the following changes on pore water chemistry:

- supersaturation with respect to high-Mg calcite, but low Mg/Ca ratios. Experiments conducted by Lippmann (1973, p.109-110) and also during this study (Appendix A) suggest that a Mg/Ca ratio above 2 is sufficient to promote aragonite precipitation and suppress that of calcite.
- precipitation of primary dolomite cement may have taken place concurrently with dolomitization of the earlier fibrous cements. Precipitation might have resulted from increased Mg/Ca ratios subsequent to Ca removal during precipitation of earlier calcite cements, or concomitant pisolite growth in adjacent parts of the caliche.
- a return to meteoric conditions would promote the precipitation of sparry low-Mg calcite.

The above changes would necessitate an open chemical system capable of being flushed periodically by saline and non-saline pore solutions. A possible scheme illustrating such changes is shown in Fig.56. Formation of the caliche deposits took place in a supratidal environment that may have been flooded periodically by high storm tides. The presence of saline pore waters (possibly enhanced by evaporation) could promote precipitation of high-Mg calcite (and aragonite?) and initiate dolomitization. Fresh groundwater run-off would ultimately dilute these brines, resulting in a return to meteoric-vadose conditions with low-Mg calcite precipitation. Any aragonite or high-Mg calcite exposed to the meteoric conditions would rapidly invert to low-Mg calcite; preservation of these early fabrics in dolomite thus supports a hypothesis of penecontemporaneous dolomitization.

Barium for the barite could have been derived by weathering of volcanics of the Eskimo Formation, or from sea water.

DETRITAL DOLOSTONES

Grainstones, dolorudites and intraformational conglomerates are prominent in the McLeary and Rowatt Formations and provide considerable information regarding the timing of lithification and dolomitization. In the McLeary examples, depositional environments range from shallow subtidal to supratidal, having developed on a broad marine platform. Dolostones of the Rowatt Formation on the other hand form part of a carbonate buildup-bar complex. A number of examples from each are discussed.

Beach Rosettes and Beachrock:

Beach rosette pavements composed of vertically stacked dololomite

Figure 56. Diagrammatic representation of possible sequence of events leading to formation of caliche deposits in an upper supratidal environment (Fairweather Formation). The sequence is characterized by repeated episodes of dissolution and carbonate precipitation over a time span of 100's to 1000's of years.

slabs are thought to have formed as a result of wave erosion of preferentially lithified carbonate muds (Ricketts and Donaldson, in press). The presence of gypsum casts in some beds indicates that supratidal marine conditions prevailed; by analogy with modern environments, for example the Persian Gulf, the original mineralogy was probably aragonitic. Distinct grain-size differences between the dolomite composing the slabs (generally less than 10 μm) and the dolosiltite matrix (Pl.14, Fig.3) suggests the possibility of penecontemporaneous dolomitization.

The presence of beachrock in several formations in the Belcher Group provides evidence for rapid penecontemporaneous cementation (Donaldson and Ricketts, in press). In all observed occurrences, both framework and cements are composed entirely of dolomite. Most of the dolomite grains possess sharp boundaries and a few grains exhibit micritic rinds as indicators of a first generation cement. However, unlike cement fabrics in grainstones of the Rowatt Formation, the beachrock cements have been largely destroyed. If, as in modern beachrock, fibrous aragonite cements prevailed, exposure to meteoric conditions would result in rapid inversion of aragonite to calcite with subsequent changes in fabric. Such a process may explain the absence of fibrous fabrics in the ancient examples. Nevertheless, contrasting grain sizes between dololutite clasts and cements does provide evidence for reworking of locally derived dolomite.

Carbonate Buildups of the Rowatt Formation:

An extensive series of carbonate buildups forms part of a system of banks and offshore bars which protected a restricted or lagoonal basin (represented by the Kipalu Formation ironstones) from an open marine platform. The carbonate buildups consist of moderately well bedded

grainstones and banded dololutites which became brecciated and folded, possibly as a result of penecontemporaneous solution collapse within the buildups. This brecciation was part of a complex sequence of events involving numerous cycles of carbonate mud and sand deposition, early lithification and deformation. The buildups were positive elements with respect to adjacent tidal flats and probably were subaerially exposed with periodic submergence during high flood tides or storms.

Grainstones: Relationships between grain-supported frameworks and cement are well preserved in these rocks (Pl.14, Fig.4; Fig.57). For example dololutite clasts composed of 3-5 μm size dolomite crystals sharply abut grains consisting of crystals in the 30-40 μm range with little evidence of any recrystallization gradient between the grains. In addition, inclusion-free dolomite cement is also in sharp contact with these grains. At point contacts this dolomite has the appearance of a meniscus cement (cf. Dunham, 1971) and thus provides some evidence for vadose-zone diagenesis (Fig.57a; Pl.15, Fig.1). The final void-fill sparry dolomite is also inclusion-free with sharp, planar intercrystalline boundaries. Thus the formation of grainstone cements in the Rowatt buildups can be summarized as follows:

- initial micritic rinds (aragonite or Mg calcite),
- meniscus cements at point contacts,
- dolomite spar.

Two variations on this sequence were observed. The first shows an initial drusy dolomite lining, followed by more equant spar. In some cases the later spar is in crystallographic continuity with earlier dolomite rhombohedra, the boundary between the two generations being marked by a zig-zag line of hematite inclusions (Fig.57b).

A second variation involves an initial fibrous dolomite followed by spar (Fig.57c). This sequence resembles some beachrock cements and

Figure 57. Critical cement-framework relationships in Rowatt grainstones. (a) dolomite clasts with micritic rinds, possible relic meniscus cements (M) and void-filling dolomite spar (S). Q = quartz grains. (b) preservation of two generations of cement: an initial drusy dolomite rhombohedral lining, separated from an equant dolomite mosaic (S). Inset shows zig-zag inclusion trails separating these two stages. (c) relic fibrous rim cement (F) followed by a sparry dolomite mosaic (S). The initial fibrous stage may represent relic high-Mg calcite.

may represent early dolomite replacement of a Mg calcite precursor. A similar replacement phenomenon has been documented in recent beachrock from Fuerteventura (Muller et al. 1973).

The thesis that a large proportion of the clastic dololomite was derived from local sources and represents the product of penecontemporaneous dolomitization (Fig.58) is supported by the following criteria:

- sharp contacts between clasts that exhibit varying degrees of crystallinity,
- sharp contacts between clasts and void-filling cements,
- facies control over the distribution of clastic dolomite.

An interesting comparison can be made with allodapic calcarenites from the deeper water Costello Formation. Here lutite clasts and ooids possess varying degrees of micritic dolomite replacement, ranging from completely dolomitized to wholly calcitic, and are set in a sparry calcite cement. Thus, dolomitization of these clasts preceded deposition and calcite cementation.

Thinly Bedded Dololutes:

Well-bedded dololutes consisting of 2-15 mm thick layers are characterized by extremely fine grained size, mostly in the range 3-5 μm . Terrigenous detritus is rare. For the same reasons indicated above, the lutites are thought to represent penecontemporaneous dolomite replacement of aragonite muds. Supporting evidence is provided by the fine grain size and synsedimentary truncation of folded lutite beds that likely produced much of the clastic debris present in the grainstones. Possibly analogous dolomitization of modern aragonite muds has been reported from the Persian Gulf (Illing et al. 1965) and Coorong Lagoon, Australia (von der Borch et al. 1975).

Figure 58. Diagenetic stabilization of originally metastable carbonate assemblage from the Rowatt lutites and grainstones: (a) initial aragonite muds; (b) penecontemporaneous dolomitization; (c) erosion of lutites and formation of grainstones composed of clasts possessing varying degrees of dolomitization. Aragonite and high-Mg calcite probably persist at this stage; (d) early cementation; (e) continued dolomitization and cementation eventually leading to a fully stabilized dolomite assemblage. This stage may be interrupted by renewed erosion and deposition.

The nature and origin of the intricate fold patterns developed in the lutites is discussed in Chapter 3. Of particular interest here is the formation of cavities with crescent-shaped cross-sections that have been filled with a coarse fibrous dolomite mosaic (Fig.59; Pl.15,Figs.2 & 3). Turbid nonferroan dolomite crystals up to a millimeter in length have a length-breadth ratio of 3:1 to 5:1. Crystals display a distinct increase in size from the substrate (dololutite) into the cavity. Closer examination reveals that each crystal is made up of a number of sub-crystals which impart a range of extinction positions of 5° to 15° . In most of these mosaics the extinction shadows converge towards the center of the cavity; the opposite to that expected of normal radial fibrous mosaics. Intercrystalline boundaries are non-planar, suggesting that the dolomite is not a primary cement.

A change in crystal habit from fibrous to equant rhombohedral dolomite takes place across a zig-zag line of inclusions that marks the contact between first and second generation fabrics (Pl.15,Figs.2 & 3). Individual crystals commonly are in crystallographic continuity across this contact. Inclusions are evenly distributed throughout the fibrous dolomite and ghosts of former crystal structures are absent. The rhombohedral crystals on the other hand are essentially inclusion free and staining reveals them to be ferroan dolomite.

The cavity filling dolomite appears identical to radiaxial fibrous mosaics first described by Bathurst (1959) which are characterized by two principal features: single crystals have optic axes that converge towards the cavity, and divergent subcrystals. This fabric has been attributed to both primary void-filling cement and to a later diagenetic replacement fabric. Kendall and Tucker (1973) have presented evidence

Figure 59. Sketch of radiaxial fibrous dolomite fabrics from folded dololutites, Upper Member, Rowatt Formation, illustrating the nature of crystal boundaries, cleavage patterns (continuous over the first (a) and second generation (b) fabrics), and fast directions of subcrystals (arrows).

suggesting that the fabric represents replacement of an early diagenetic acicular cement composed of interfering bundles of radiating crystals. Such a habit is typical of aragonite in marine cements although Lohmann and Meyers (1977) have demonstrated probable fibrous high-Mg calcite precursor cements in some radiaxial fabrics. Nevertheless, most recent accounts of fibrous or platy Mg calcites indicate relatively restricted crystal size (a few tens of microns). Examples include both shallow reef cements (James et al. 1976) and beachrock cements (several examples in Bricker, 1971). Morphologically the Rowatt types are more like the calcite fabrics described by Bathurst (1959, 1976) and by Kendall and Tucker (1973) for which an aragonite precursor is indicated; an aragonite precursor is also proposed here. Furthermore, it is suggested that dolomite replacement proceeded from the aragonite stage rather than through an intermediate low-Mg calcite stage, and was probably concomitant with dolomitization of adjacent aragonite muds. The preferential concentration of iron (Fe^{2+}) in the second generation sparry cements probably reflects inability of the aragonite lattice to accommodate this ionic species. The Fe must have coprecipitated with the sparry cement; if Fe was associated with dolomitization, it might be expected to pervade all of the dolomite fabrics.

STROMATOLITIC DOLOSTONES

Stromatolitic dolostones in the Kasegalik, McLeary, Tukarak and Mavor Formations present quite a different picture compared to dolostones discussed in previous sections. Paleoenvironments interpreted for these units range from intertidal and shallow subtidal to outer platform (below wave base). The dolomite in these rocks has a relatively coarse recrystallized habit where crystals reach 200 μm in size. A characteristic

feature of many of these dolostones is the presence of relic dolomicrite clots and layers that appear to 'float' in the coarser microspar and pseudospar (Pl.15, Fig.4) and thus resemble the structure grumeleuse described by Cayeux (1935). Superficially these clots are similar to peloids, although detailed examination reveals fabrics more typical of neomorphic recrystallization, namely gradation in grain size from micrite to microspar and highly irregular intercrystalline boundaries (cf. Bathurst, 1976, p.485). In addition, there is little evidence of other types of clastic material such as quartz, feldspar or ooids. This is true even for stromatolitic dolostones interbedded with sandstones.

Preservation of Primary Laminae:

The effects of such pervasive recrystallization is the almost complete destruction of primary biofabrics. The most pronounced feature that remains is the light-dark alternation of laminae that is characteristic of most stromatolites in the Belcher Islands. This pattern corresponds to micritic laminae (less than 100 μm thick) alternating with lighter coloured microspar and pseudospar laminae up to 2000 μm thick. Clotted micrite textures developed where advanced recrystallization disrupted the finer laminae. Hofmann (1975) has suggested that the dark laminae represent heavily pigmented layers in the original algal mat, with the pigments possibly inhibiting crystal growth.

A notable feature of these dolostones is that whereas recrystallization has affected the cryptalgal structures, interbedded non-stromatolitic dololutites have remained fine grained even on a scale of a few millimeters.

Silicification:

Cherts are common in stromatolitic dolostones from the Belcher Group. Preservation of microfossils in the cherts suggests that silicification was almost penecontemporaneous where early formation of silica gels prevented degradation of the algal structures (Hofmann, 1975).

Apparently the distribution of these cherts is not random; they are most common in stromatolitic dolostones that formed in intertidal and supratidal environments, particularly in the McLeary Formation. Stromatolites that grew in shallow subtidal environments (uppermost McLeary Formation) and deeper water platform environments (Mavor Formation) exhibit a marked decrease in the degree of silicification. This may indicate some degree of primary facies control on the distribution of chert, for example changes in environmental parameters such as pH, $p\text{CO}_2$ and the concentration of various ionic species (especially H_3SiO_4^-).

Diagenetic Series:

The following series of diagenetic events affecting the carbonates has been recognized in the stromatolitic dolostones:

- a) Chert-free dolostones- dolomicrite (presumably replacing original calcium carbonates),
 - recrystallization to microspar,
 - stylolite formation and growth of late euhedral dolomite rhombs.
- b) Cherty dolostones
 - precipitation of chert and carbonate,
 - dolomicrite replacing precursor carbonate,
 - recrystallization of dolomite to microspar,

- fracturing, late quartz filling and silicification,
- late euhedral dolomite replacing chert and vein quartz.

In most cases two generations of dolomite occur, the first being micrite and accompanying neomorphic spar, and the second being represented by euhedral rhombs which in some cases appear to syntaxially overgrow the earlier generation of microspar.

Discussion:

Although dolomicrite texturally appears to be the primary phase in most stromatolitic dolostones, it is unlikely that this represents the original carbonate mineralogy. However, experiments conducted by Greenfield (1963) and Gebelein and Hoffman (1973) have demonstrated that certain bacteria and cyanophyte algae can preferentially concentrate Mg as organic complexes. This Mg could then be released during organic decay to provide an additional source of this ion for precipitation of Mg calcites and possibly later dolomite. Because algal mat micro-environments are rich in organic matter including bacteria, the physico-chemical conditions within them are highly conducive to carbonate precipitation. Krumbein and Cohen (1977, p.53) have reported a variety of carbonate minerals in recent blue-green algae as coatings and replacement of filaments, including high- and low-Mg calcite, aragonite and even dolomite.

Based on the overall facies analysis and petrology of these dolostones, there is no evidence for the existence of elevated salinities either during growth or after burial of the remarkably extensive sub-tidal biostromal complexes. A penecontemporaneous origin for the

dolomite analogous to the type observed in some modern and ancient supratidal environments is therefore ruled out. Dolomitization of the stromatolitic deposits may have proceeded from initial calcitic precursors at some stage following burial. The pervasive nature of the dolomitization was probably enhanced by preferential concentration of Mg by stromatolite-building organisms. However, an allochthonous source for some of the Mg cannot be ruled out.

'DEEP' WATER CARBONATE MUDROCKS

The upper part of the Mavor Formation and the overlying Costello Formation consist of thin, rhythmically bedded calcilutites, calcisiltites, dololutites and dolosiltites with intercalated shales and a number of calcarenites of probable turbidity current origin. Laterally extensive intraformational slump horizons and beds of flat pebble conglomerates occur in the Costello Formation. The thinly bedded carbonates are typical of Wilson's (1969) rhythmite facies and have been interpreted as slope deposits that accumulated on the basinward side of a shallow marine platform (Fairweather to Mavor Formations).

The Costello rhythmites are characterized by a distinctive lumpy or nodular pattern of dolomitization. In places this pattern mimics primary sedimentary structures such as parallel laminae or ripple cross-bedding, although more commonly the patterns cut across these structures.

The grain size ranges of the calcite and dolomite components of these rocks are similar and normally are less than 40 μm . Patchy lutite zones may be as fine as 5-10 μm . Terrigenous debris generally composes less than 10%, although some thin layers contain up to 60% quartz and feldspar. The graded calcarenites contain dololutite and calcilutite

grains, including grains with intermediate dolomite-calcite compositions, together with dolomitized ooids.

The patchy dolomitization that is characteristic of the Costello Formation is in marked contrast to all other pre-Flaherty Formation carbonate units in the Belcher Group. Any attempt to establish possible mechanisms of dolomitization must take this into account together with the inferred paleogeographic position of the Costello carbonates: as relatively deep water deposits they presumably were remote from conditions of hypersalinity. Before proceeding with this discussion it is useful to examine the possible constraints on the original mineralogy of the Costello rocks.

At present, and probably throughout a large part of the Phanerozoic, pelagic sediments were composed predominantly of skeletal debris. There is no evidence for the existence of a similar source during Precambrian times. However, modern pelagic sediments in slope environments have also been observed to contain significant amounts of platform-derived sediment (Wilson, 1969). The presence of fine terrigenous debris together with reworked ooids attests to a similar source for some of the Costello deposits. Fine-grained, deep water sediments derived from adjacent platforms have been called periplatform ooze by Schlager and James (1978). In the Belcher examples these sediments might have consisted of aragonite and calcite with different Mg contents, although the overall sediment composition was probably not unusually rich in Mg.

A simple calculation of the MgCO_3 budget supports this statement: if the original carbonate mud contained, say 10% by weight MgCO_3 in solid solution with calcite, then the weight percent Mg would be 2.9%; because stoichiometric dolomite contains 13% Mg, then the total possible

amount of dolomite that could be precipitated in the sediments would be 22% by weight. Recalculating to volume percent gives approximately 22%. This calculation assumes that the chemical system is closed to Mg and that all of the Mg available is consumed during dolomitization. Most high-Mg calcites at present occur in invertebrate skeletons and coralline algae, and hence an average value of 10% for the Costello sediments is probably too high. Moreover, aragonite rarely contains more than 1-2% MgCO_3 . By comparison, at least 75% of the Costello sediments have been dolomitized. Thus it is likely that a large proportion of the Mg required for dolomitization was allochthonous.

Dolomitization of Deep Water Carbonates:

Mechanisms of dolomitization involving evaporation, hypersalinities or pore fluids with high Mg/Ca ratios are incapable of explaining many thick dolomitic carbonates that were deposited in deep or shallow water; there is no evidence that such conditions ever existed in the Costello Formation. However, the hypothesis that dolomitization could take place in a mixed meteoric-sea water system has been investigated (Hanshaw et al. 1971; Badiozamani, 1973; von der Borch et al. 1975; Wigley and Plummer, 1976). Mixing of meteoric waters and ocean waters in varying proportions results in non-linear changes in the degree of saturation with respect to calcite and dolomite. At certain levels of mixing such waters may be saturated with respect to dolomite and undersaturated with respect to calcite (Fig.60) and thus replacement of calcite by dolomite is theoretically possible. The application of a mixing system (cf. the Dorag model of Badiozamani, 1973) to outer platform and slope sediments requires the presence of a fresh water aquifer of regional extent with a hydrostatic head being maintained over long periods (Land et al. 1975).

Figure 60. Changes in calcite-dolomite saturation with mixing of samples having different proportions of seawater. (Sample 1, $p\text{CO}_2 = 10^{-2}$ atm.; Sample 2, $p\text{CO}_2 = 10^{-3}$ atm.). Shaded region favours dolomitization by replacement of calcite (after Wigley and Plummer, 1976, Fig. 3, p.994).

Continent-derived waters are known to extend for considerable distances beneath modern continental platforms, for example up to 120 km off the west coast of Florida (Manheim, 1967). The early meteoric influx of continental groundwater with dolomitization taking place along a brackish water mixing zone is therefore postulated for sediments of the Costello Formation. Potential aquifers include sandstones and grainstones of the Fairweather and McLeary Formation, and biostromal complexes in the McLeary, Tukarak and Mavor Formations. Three additional sources of Mg can also be considered: (i) Mg from sea water, (ii) Mg from the adjacent shallow-platform carbonates, and (iii) Mg released by the weathering of cratonic rocks. The existence of such a system within the Fairweather to Laddie transgressive sequence also demands that the mixing zone was moving landwards. By comparison, regressive regimes such as the lagoonal-barrier bar complex at Coorong Lagoon, Australia (von der Borch et al. 1975) exhibit prograding mixing zones.

DISCUSSION

Dolomitization of marine platform-to-slope carbonates (Fairweather to Laddie Formations) has taken place within a wide spectrum of sedimentary and diagenetic environments over a time span that probably ranged from penecontemporaneous to a few tens of millions of years. A schematic representation of salinity zones, mixing zones and possible groundwater flow directions is shown in Fig.61.

There is abundant evidence of early rapid lithification in intertidal and supratidal environments with the precipitation of intertidal and submarine cements (eg. beachrock) and vadose or phreatic cements in supratidal crusts and caliche beds. Penecontemporaneous dolomitization

Figure 61. Hypothetical salinity zones, mixing zones and groundwater flow directions for platform-slope-basin carbonates, Belcher Group. Near-surface landward flow of sea water was probably restricted to the near-shore environment of penecontemporaneous dolomitization (cf. the evaporative pumping mechanism of Hsu and Siegenthaler, 1969). Vertical scale greatly exaggerated.

appears to have been widespread in these environments, initially replacing aragonitic muds and some high-Mg calcite cements. Dolomitization was likely promoted by increased salinities and Mg/Ca ratios as a result of evaporation and subsequent precipitation of calcium carbonates and some gypsum (cf. the evaporative pumping mechanism of Hsu and Siegenthaler, 1969).

It is suggested that dolomitization of intertidal and subtidal stromatolitic carbonates proceeded via a Mg calcite precursor; cyanophyte algae and bacteria are known to be capable of preferentially concentrating Mg as organic complexes. Upon decay the Mg is released and can precipitate as Mg calcite in chemical microenvironments having a high pH (eg. within or attached to algal sheaths). This stage of dolomitization was a relatively late post-burial event.

By contrast, the deeper water slope deposits (below the photic zone) were probably composed of low-Mg calcite and some aragonite. Late dolomite replacement of calcite might have occurred in the brackish water zone resulting from mixing of continental meteoric waters and sea water. Biostromal carbonates may have also come under the influence of this mixing zone (Fig.61); high-Mg calcites are unstable under meteoric ~~sub~~-phreatic conditions and hence could be replaced by dolomite as long as conditions of dolomite saturation were satisfied. The difference in the degree of dolomitization between the stromatolitic carbonates (now almost 100% dolomite) and the deeper water rhythmites of the Costello Formation (approximately 75% dolomite) may reflect differences in the original mineralogy such as the primary deficiencies in Mg content noted above.

- Figure 1. Vadose pisolites in caliche, Fairweather Formation (sample R81, 5 m from base of Section 4). Note asymmetric growth patterns and growth discontinuities (arrows). The lower half of the photograph contains brecciated dolomicrites. The pisolite on the right has grown around three micrite fragments (M) possibly derived from an earlier episode of solution-brecciation. Larger voids are infilled with rhombohedral dolomite cement (D). Bar scale represents 1 millimeter. Plane polarized light.
- Figure 2. Two generations of void-filling cement in pisolitic caliche: an initial rim of fibrous dolomite (arrow) possibly pseudomorphing high-Mg calcite, followed by rhombohedral and blocky dolomite. Note planar intercrystalline boundaries. (Same sample as Fig.1.) Bar scale represents 100 μ m. Crossed nicols.
- Figure 3. Cross-section through dololomite slab from stone rosettes demonstrating the contrast between silt-free lutites and dolosiltite matrix (Sample R125 from top of Section 6). Bar scale represents 1 mm. Plane polarized light.
- Figure 4. Grainstone, Rowatt Formation, with typical grain-supported framework, coarse dolomite cement and clasts having a range of crystallinities (Sample R221, 45 m from top of Section 15). Darkest grains are composed of 3-5 μ m dolomite. Bar scale represents 1 millimeter. Plane polarized light. Region magnified in P1.15, Fig.1 is indicated by an arrow.

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PLATE 15

Figure 1. View of grain boundaries and cements in Rowatt Formation grainstones showing micritic rinds, fine-grained inclusion-rich dolomite at grain contacts (arrows) possibly representing relic meniscus cements, and late sparry inclusion-free dolomite cement (same sample as Pl.14, Fig.4). Bar scale represents 300 μ m. Plane polarized light.

Figure 2. Photomicrograph of radiaxial fibrous dolomite showing sharp contacts with micritic cavity walls. Dolomite crystals have grown normal to margins of the laminae, some of which have been fractured and folded before cementation. A stylolite closely follows the margin of a cavity created by this deformation (right). Sample R442, top of Section 34. Bar scale represents 1 mm. Crossed polars.

Figure 3. Enlarged segment of radiaxial fibrous dolomite. Note increase in crystal size towards center of the cavity and highly irregular intercrystalline boundaries. A zig-zag line of hematite inclusions separates fibrous and equant dolomite mosaics. Lattice continuity across this boundary is indicated by continuous cleavage patterns. Second generation dolomite is relatively free of inclusions. Same sample as Figure 2. Bar scale represents 300 μ m. Plane polarized light.

Figure 4. Photomicrograph of stromatolitic dolostone showing 'floating' relics of dolomicrite (dark) surrounded by microspar and pseudospar. Sample R23, 345 m from the top of Section 1, Tukarak Formation. Bar scale represents 300 μ m. Plane polarized light.

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CHAPTER 6

SUMMARY OF DEPOSITIONAL HISTORY,
STRUCTURE AND CORRELATION
OF THE BELCHER GROUP

DEPOSITIONAL HISTORY OF THE BELCHER BASIN

The Belcher Group (approximately 1.8G a) consists of a thick (7000-9000m) but remarkably regular succession of distinctive rock units. The preceeding chapters have mainly been devoted to establishing the vertical and lateral facies relationships, paleoenvironments and regional paleoslope changes represented by these formations.

Collectively, formations of the Belcher Group present a complex picture of basin evolution, where long-term changes in depositional environments have taken place in response to fundamental changes in the tectonic regime. The history of sedimentation and volcanism in the Belcher Basin is summarized in Figs.62 & 63 and is divided into four major cycles.

First Cycle (Kasegalik Formation)

The earliest record of sedimentation is presented by the Kasegalik Formation, where a vertical sequence of mudstones, sandstones and stromatolitic dolostones represent the lateral transition from supratidal (probably evaporitic) to shallow subtidal environments. It is a 1200-meter thick transgressive sequence that apparently developed on a broad marine platform.

Second Cycle (Eskimo Formation to Laddie Formation)

Sedimentation on the marine platform was abruptly terminated by extrusion of extensive tholeiitic flood basalts of the Eskimo Formation. Emplacement of these basalts probably arose from fissure-type eruptions east or northeast of the Belcher Islands.

An extended period of gradual subsidence followed this phase of volcanism with deposition of some 1800-2000 m of predominantly carbonate sediments. Here too, the vertical stratigraphic succession from the

Figure 62. Idealized stratigraphic column of the Belcher Group, Hudson Bay, showing overall basin evolution in relation to sedimentation and volcanism, prior to regional deformation and metamorphism during the Hudsonian Orogeny.

Figure 63. A generalized stratigraphic cross-section of the central part of the Belcher Group (A-B) illustrating lateral and vertical facies changes.

Fairweather to the Laddie Formation, is considered to represent an assemblage of facies that were laterally associated in space and time, and is therefore interpreted as a transgressive platform-slope-basin sequence. The regional paleoslope dipped west to southwest.

Superimposed on this overall transgressive regime are numerous shoaling-upwards intertidal to supratidal cycles composing the Fairweather Formation, each cycle culminating in pisolitic caliche beds. Tidal flat environments represented by the Fairweather and McLeary Formations are locally evaporitic, and also contain a diverse assemblage of cryptalgal laminates ranging from flat and tufted mats, to broad domal stromatolites. Intervals of relative calm along this ancient shoreline were interrupted by storms as indicated by beachrock boulder conglomerates and disrupted algal mats.

Subtidal environments (McLeary to Mavor Formations) gave rise to a striking array of complexly branched stromatolites, particularly in the Upper McLeary, and to extensive deeper water biostromal mounds of the Mavor Formation. The seaward limit of the Mavor biostromes probably corresponds to the outer margin of the marine platform. Distinctive carbonate rhythmites (Costello Formation), which also contain a laterally persistent zone of soft-sediment slump folds and allodapic calcarenites, are interpreted as slope-facies deposits. Much of the carbonate and terrigenous silt deposited here probably was derived from the adjacent platform. These slope deposits grade into relatively deep-water argillites and shales of the Laddie Formation.

Comparisons with Paleozoic carbonate sequences in North America suggest a possible time span for the Second Cycle equivalent to 50 to 100 million years (comparable to the Devonian or Ordovician Periods), for

example the Cambro-Ordovician sequence of the Central Appalachians (Reinhardt, 1977) or the late Silurian-Devonian Helderberg Group, New York (Laporte, 1969).

Third Cycle (Rowatt Formation to Kipalu Formation)

Argillites of the Laddie Formation grade rapidly into sandstones and dolarenites composing the Lower Rowatt Formation, thus initiating the third cycle. Here, as in the two preceding cycles, each formation is considered to be diachronous, representing a shoaling upwards sequence that developed along a prograding shoreline. Banded ironstones belonging to the Kipalu Formation were deposited in a restricted or barred basin that was protected from the open ocean by a system of carbonate buildups and subtidal and intertidal sand bars. The carbonate buildups were positive elements relative to mean sea level. Possible equivalents to the Kipalu ironstones occur in the Hopewell Islands and Richmond Gulf south to Sutton Lakes. It is possible that this barred-basin system may have extended for some 400 km along the eastern coast of Hudson Bay. Such an environment is comparable, in terms of gross physiographic setting, to the barrier island-lagoon complex bordering the Gulf of Mexico.

The rapid transition from a transgressive (Second Cycle) to prograding regime can be interpreted in several ways, the two options considered here being: relative lowering of sea level, or continued subsidence and subsequent rapid influx of sediments. Although proof for either alternative has not been forthcoming, the second is preferred; the prograding regime follows an extended period of gradual subsidence (Second Cycle) and preceeds a period of very rapid subsidence and subsequent infilling of the basin (Fourth Cycle). The marked change in depositional environment indicated by the Third Cycle is thought to have

taken place in response to an increased rate of subsidence of the marine platform prior to the Fourth Cycle. The supply of sediment by north- to northwest flowing longshore currents, and possibly by reworking from the adjacent platform, was sufficient to keep pace with subsidence resulting in overall progradation.

Fourth Cycle (Flaherty Formation to Loaf Formation)

A few thin tuffaceous siltstone beds at the top of the Kipalu Formation provide the first indications of approaching volcanism. The emplacement of a thick volcanic edifice (almost 2000 m thick) as a result of eruptions from centers west of the Belcher Islands resulted in reversal of the paleoslope (from west-dipping to east-dipping) and rapid subsidence of the Belcher Basin.

Extensive accumulations of tholeiitic pillowed basalts indicate predominantly submarine volcanism, although the presence of ropy lava tops on several flows, and accretionary lapilli and fine stratification in some volcanoclastic beds also suggests the possibility of some sub-aerial eruptions and deposition by fallout. Some of the larger jointed flows are more likely to be subaerial. In addition, most hyalotuffs are composed of highly vesicular fragments indicating shallow-water, largely phreatic eruptions. In general, the proportion of interflow sediments increases towards the east from about 5% to 27%, probably corresponding to increased water depths. Deposits include pyroclastic turbidites consisting entirely of vesicular fragmental debris, and subaqueous non-welded pyroclastic flows, one of which covers a palinspastic area greater than 3000 square km and has a volume of volcanic debris greater than 110 cubic km. Thus the effusive and explosive activity which characterized the Flaherty volcanic episode was occasionally interrupted by explosive events of catastrophic proportions.

Classification of the Flaherty volcanics as to possible tectonomagmatic environments is difficult without detailed geochemical information in addition to the physical attributes of the volcanics and the associated sediments. The limited number of chemical analyses available suggests predominantly tholeiitic compositions; this is based solely on major element analyses. Submarine basalts are highly susceptible to alteration as a result of reactions with sea water, and may lead to spurious results, particularly for the alkali elements. In most Flaherty basalts the plagioclase (largely albite) has undergone some alteration. Prehnite and pumpellyite are common products of regional and/or burial metamorphism.

The proportion of fragmental deposits is an additional criterion useful in distinguishing ocean floor basalts, volcanic arc basalts and oceanic island basalts (Garcia, 1978). For example, volcanic arcs generally contain greater than 90% pyroclastics whereas volcanics associated with deep mid-ocean ridges contain less than 5%.

Based on the variety, nature and abundance of volcanic products the Flaherty strata are tentatively compared to an oceanic ridge volcanic regime, possibly analogous to emergent segments of the mid-Atlantic ridge such as Iceland and the Azores.

Following culmination of Flaherty volcanism, the now rapidly subsiding basin was infilled with some 2100 m of turbidites (Omarolluk Formation), probably as submarine fans. Evidence from paleocurrent indicators demonstrates that sediment transport was largely towards the southeast, generally parallel to the turbidite basin. Thinning of beds and decreasing sandstone/shale ratios in the southern part of the Belcher Islands also suggests a deepening of the turbidite basin in this direction. Sediments were derived from two principal sources: the Flaherty volcanics

probably flanking the western margin of the basin, and a cratonic source to the north. The volcanic component decreases towards the top of the Omarolluk Formation.

Red and grey arkoses of fluviatile, subaerial and possibly marginal marine origin compose the Loaf Formation, the uppermost unit of the Belcher Group. On the basis of mineralogic composition and paleocurrents (sediment transport was largely towards the southeast) these facies are considered to be laterally associated with the Omarolluk turbidites. The Omarolluk-Loaf sequence may therefore be comparable to some modern submarine fan-delta-braided river systems such as occurs in the northern Adriatic Sea, or the Bengal Fan and Nile Cone systems.

STRUCTURE OF THE BELCHER ISLANDS

Folding

The highly sinuous outline of the Belcher Islands is a manifestation of open, predominantly isoclinal, doubly plunging anticlines and synclines. Folds are upright or slightly inclined either to the east or west; the axial plane of the Tukarak anticline dips east at about 70° . Fold axes trend approximately north-south, but as Jackson (1960) has noted, the axes become curved west of Flaherty Island, opening towards the west. Plunges range from 20° to less than 5° (eg. the southern part of Coats Bay syncline). On Tukarak Island the plunges range from 10° - 15° . Axial plane cleavage has developed in argillaceous lithologies and apparently forms a divergent fan in the Tukarak anticline. Crustal shortening ranges from 25% in the Tukarak anticline to more than 40% in the Kasegalik anticline.

Parasitic folds occur principally in fold hinge zones and are

especially prominent in the Fairweather Sound-Mavor Island area, and also in the northern part of Kipalu Inlet. For example in the northeast sector of Mavor Island, up to three orders of parasitic fold that are superimposed on the Tukarak anticline are developed in relatively incompetent mudstones and dololutites belonging to the Tukarak Formation.

Faults

Two major fault types occur. Most prominent are subvertical normal faults, especially in fold hinge zones. Where exposed, slickensides pitch at 70° to 90° , indicating a large component of dip-slip displacement. In the Tukarak Island area most of the large faults are approximately parallel to the principal fold axes, although a few smaller faults are almost perpendicular to these axes. The overall distribution suggests that the main periods of faulting and folding approximately coincide.

The second type of fault is the bedding plane fault where displacements are almost entirely dip-slip (ie. flexural slip). Faults of this type may be easily overlooked in areas where bedding exposure is poor.

Joint Systems

Five major joint sets are recognized:

Set a: subvertical extension joints, perpendicular to fold hinge lines.

Sets b & c: Conjugate shear sets of subvertical to steeply dipping joints (accompanied by development of slickensides), are particularly prominent in Flaherty, Rowatt, McLeary and Fairweather strata on Tukarak Island. Many of the conjugate sets are bisected by Set "a" joints. In addition, all three sets are symmetrically oriented around the Tukarak anticline and in zones of faulting these trends are cut by major faults. Development of these joints was probably an early event in the history of deformation.

Principal stress directions were likely parallel to Set "a" joints and perpendicular to the axial surfaces of folds.

Set d: These steeply dipping joints are parallel to bedding and are analogous to the radial joints of Hobbs et al. (1976, p.294).

Set e: Bedding plane joints, probably developed in response to unloading after retreat of the Pleistocene ice sheets.

The structural style of the Belcher Islands can be explained by a single period of deformation: initial buckling led to formation of extension and conjugate shear joint patterns, followed by predominantly flexural slip folding and concomitant development of major faults. Deformation is considered by most workers to have occurred during the Hudsonian Orogeny (Hofmann and Jackson, 1969; Dimroth et al. 1970; Fryer, 1972).

CORRELATIONS

Early attempts at correlation of Precambrian rocks exposed on numerous islands and along the mainland coast of eastern Hudson Bay were based primarily on lithological attributes (Bell, 1879; Low, 1902). In particular, islands hugging the east coast from Long Island to the Hopewell Islands (Fig.64) are characterized by basaltic cap rocks ("traps") forming prominent cuestas. Distinctive banded ironstones and "concretionary" (algal) dolomites were especially useful as stratigraphic markers in these early studies. So similar are these rocks to types documented in other parts of the Superior Basin that Leith (1910) considered them to be Algonkian equivalents.

Following a survey by Flaherty in 1914 (Flaherty, 1918) the Belcher sequence which contains ironstone, dolostone and volcanic lithologies

CORRELATIONS AROUND EASTERN HUDSON BAY
AND JAMES BAY

Area Shown in Fig.

1	Ottawa Is. - Sleeper Is.	Manning, 1947
		Andrews & Falconer, 1969
2	Hopewell Is.	Kretz, 1960
		Lee, 1965
3	Richmond Gulf	Bell, 1879
	Nastapoka Is.	Low, 1902
	Manitounuk Sound	Parks, 1949
		Kranck, 1951
		Kretz, 1960
4	Long Is. Bear Is. Sunday Is.	Burns, 1952
5	Sutton Lakes	Dowling, 1905
		Hawley, 1926
		Bostock, 1968
		Sanford <i>et al.</i> , 1968

Figure 64. Correlation of Proterozoic strata (red) composing the Belcher Fold Belt.

similar to the above, was tentatively correlated with the sequences bordering eastern Hudson Bay (Moore, 1918; Young, 1922; Woodbridge, 1922). Subsequent studies in other parts of Hudson Bay, including the Ottawa and Sleeper Islands, and Sutton Lake region, resulted in similar, more extensive correlations (summarized in Fig.64).

More recently, detailed correlations have been made with Precambrian fold belts outside the Hudson Bay region, including Cape Smith-Wakeham Bay, the Labrador Trough and the Mistassini area (Wahl, 1953; Bergeron et al. 1962; Dimroth et al. 1970). Correlation of these four elongate troughs around the ancient Ungava nucleus was suggested by Bergeron (1957) and Bergeron et al. (1962). The name Circum-Ungava Geosyncline was proposed to include the Cape Smith Fold Belt, Labrador Trough and Belcher Fold Belt (Dimroth et al. op. cit.).

The structure of the remainder of the Belcher Fold Belt from Ottawa Islands to James Bay has been summarized by Dimroth et al. (1970, pp.88-91). Aerial photographs show that folding in the Belcher Islands dies out north and south of these islands and that the intensity of deformation also decreases eastwards towards Richmond Gulf. The interpretation that volcanic strata (including pillowed basalts) in the Ottawa Islands are a continuation of similar lithologies in the Belcher Islands and James Bay is supported by bathymetric and magnetic data (Hobson, 1969; Hood et al. 1969).

Richmond Gulf and Manitounuk Sound strata dip less than 5° west, and along coastal sections they form prominent cuestas. Strata composing the island chains bordering the east coast of Hudson Bay dip towards the center of the Belcher Basin. The possibility that these rocks mark the preserved margin of an ancient sedimentary basin was suggested by

Hawley (1926). There is little evidence to support Beals' (1958) suggestion that this basin evolved as a result of meteorite impact.

Correlations in the Richmond Gulf Area

Lithologic units in the Richmond Gulf area have been correlated with specific formations in the Belcher Islands; Pachi Group and Nastapoka Group basalts in Richmond Gulf, have been correlated with the Eskimo and Flaherty volcanics respectively. However, the remaining units show few lithological similarities and at best, the Richmond Gulf strata can only be considered as probable continental equivalents of the Belcher Group. An additional problem is the presence of an angular unconformity between the Richmond Gulf Formation and Nastapoka Group which developed during a period of block faulting. According to Chandler (1978), most of this deformation occurred before deposition of Nastapoka sediments and volcanics. An equivalent episode of faulting affecting the Belcher Group is not known. With this in mind, it is possible that both Richmond Gulf Formation sediments and the Pachi volcanics are time equivalents of the Eskimo Formation (the Eskimo attains a thickness of 900 m, whereas the Pachi volcanics are only 50m⁺). North-south thinning of sedimentary and volcanic units in Richmond Gulf support the hypothesis of graben-controlled deposition (Krank, 1951; Woodcock, 1960; Chandler, 1978). Extension of this graben system to the Belcher Islands was also suggested by Kranck (1951).

Age of Richmond Gulf Strata:

Although undoubtedly Proterozoic in age, radiometric age determinations of Richmond Gulf rocks have proved to be inconclusive. Determinations on Pachi and Nastapoka Group basaltic rocks for example cluster around 1300-1400 million years (Stevenson, 1966, K-Ar; Hews, 1976, Rb-Sr), whereas those for argillites and shales sandwiched between these basalts

range from about 1650 to 1800 million years (Esquevin and Menendez, 1973, Rb-Sr; Hews, 1976, 1.47 decay constant used). Discrepancies such as these could have resulted from resetting of ages during the Hudsonian Orogeny or some later thermal event, or inherited source rock ages in the case of the sedimentary rocks (Hews, 1976). An additional problem in the Richmond Gulf area is a lack of detailed information on the stratigraphy of the igneous units (eg. the distinction between flows and sills).

APPENDIX A

EXPERIMENTAL INVESTIGATION OF CARBONATE
PRECIPITATION IN HYDRATED SILICA GELSBackground:

To date, most research on sedimentary carbonates has centered around analysis of primary sedimentary facies, biologic facies and diagenetic facies, and the complex interactions that exist between them. However, two aspects that are fundamental to understanding the origin of calcareous sediments are the chemical and kinetic processes that actually lead to carbonate precipitation (and dissolution) in both organic and inorganic systems.

Experimental investigations of the calcium carbonate polymorphs and mixed Ca-Mg carbonates have largely been carried out in aqueous solutions of compositions ranging from fresh water to sea water, and under a range of temperatures and pressures that approximate conditions near the earth's surface (see Lippmann, 1973, p.104-116; Bathurst, 1976, Chapters 6 & 7). Three principle experimental methods have been employed:

- direct addition of aqueous alkali and carbonate solutions; these usually result in the instantaneous precipitation of very fine-grained, poorly crystallized carbonates (calcite and aragonite),
- outgassing of CO_2 from bicarbonate solutions,
- use of the slow decomposition of potassium cyanate to produce CO_3^{2-} as a soluble ammonium salt. This method provides some control on precipitation rates, and monitoring of effects such as pH and concentration changes is easier than in the two preceding methods.

Attempts to elucidate such problems as preferential aragonite precipitation in seawater, magnesium inhibition of calcite precipitation and effects of various organic compounds on the stability of carbonates, have been approached utilizing the methods outlined above. Perhaps one of the most studied reactions is that of dolomite precipitation. A variety of experimental techniques have been used which have met with varying success, for example: precipitation from modified sea water solutions (Alderman, 1965; Liebermann, 1967); formation of protodolomite via a gel stage by freezing solutions (Muller and Fischbeck, 1973); and at elevated temperatures in hydrothermal systems (Graf and Goldsmith, 1956; Gains, 1974; Katz and Mathews, 1977). The lowest temperature at which dolomite has successfully been precipitated (and accurately documented) is 100°C (Gaines, 1974). Recently Ohde and Kitano (1978) reported protodolomite from aqueous solutions at 25°C, although no information was provided on X-ray diffraction characterization of this material or on experimental conditions such as pH or reaction times.

Most recent investigations indicate that kinetic, rather than purely energetic barriers are the reasons for the inability of dolomite to precipitate instantaneously out of bulk aqueous solutions (even though sea water is saturated several times with respect to dolomite). This has been attributed to problems such as cation ordering (alternating Ca-Mg layers in dolomite), the high dehydration energy of Mg^{2+} relative to Ca^{2+} , and the adsorption and incorporation of Ca^{2+} onto a rhombohedral phase being favoured over Mg^{2+} (Lippmann, 1973, p.114).

The present investigation was initiated to devise a method for controlling carbonate precipitation and nucleation, especially under conditions of high Mg concentration and a range of Mg/Ca ratios. The method utilizes hydrated silica gels as a growth medium.

Experimental Method:

The method of crystal growth in gels dates back to the nineteenth century when keen interest was aroused by a series of experiments conducted by Liesegang (1889, in Henisch, 1970, p.14); hence the phenomenon known as Liesegang Rings. At that time a variety of growth media were used including waterglass, gelatin, agar, and even fruit jam.

The procedure used in the present set of experiments is a modification of one described by Henisch (1970), and is illustrated in Fig.65. The gel is formed by dissolving sodium metasilicate in distilled water and acidifying with acetic acid to the desired pH (gel formation can be considered to be a type of polymerization of monosilicic acid molecules where linkage of Si-O bonds forms a three-dimensional network. The expulsion of water as a result of this reaction is known as syneresis). One of the reactant ions is put into the gel (in this case CO_3^{2-}), the other reactants (Ca and Mg) being put into the supernatant solution. The reaction leading to calcium carbonate precipitation takes place via diffusion into the gel, the rate of diffusion being controlled by the gel density. This in turn controls the rate of precipitation of the carbonate phases.

Preparation of reaction solutions and gels:

The following reagents were dissolved in deionized water: technical grade $\text{Na}_2\text{SiO}_3 \cdot 5\text{H}_2\text{O}$ (B.D.H.), $\text{MgCl}_2 \cdot 6\text{H}_2\text{O}$ (B.D.H. 97%), $\text{CaCl}_2 \cdot 2\text{H}_2\text{O}$ (B.D.H. Analar), NaHCO_3 (Analar). One molar stock solutions of each were prepared.

For all experimental runs, a standard gel solution was made from 10 ml aliquots of each of the stock solutions of sodium metasilicate and sodium bicarbonate and diluted to about 45 mls with

Figure 65. Sketch of experiment apparatus showing typical distribution patterns of the carbonate precipitate. (Liesegang ring structures are well developed in the upper levels of the gel, becoming diffuse towards the base. (sketched from run A-5).

deionized water (Table 9). Bicarbonate was added to the gel in preference to Ca^{2+} because of the tendency for insoluble calcium silicates to form. The metasilicate concentration employed here was found by trial and error to be the most suitable for producing a homogeneous gel. At higher gel concentrations and within the pH range used, gel times increased rapidly; gelling would commonly occur before the liquid could be transferred to the reaction test tubes.

The gel pH was adjusted to the desired value by drop-wise addition of acetic acid with constant stirring (pH was measured with a Corning pH meter and electrodes). The solution was then made up to 50 mls with deionized water and 25 ml portions were poured immediately into two test tubes. Although the gel times were approximately five to ten minutes, the gels were left to stabilize for 24 hours prior to addition of the supernatant liquid.

The concentrations of CaCl_2 and MgCl_2 in the aqueous supernatant solution are summarized in Table 9. A neutral pH was maintained; more basic solutions resulted in spontaneous precipitation of Mg and Ca hydroxides. The solution was added drop-wise on top of the gel, ensuring that the gel surface was not disrupted. The test tubes were sealed to minimize adsorption of atmospheric CO_2 , and placed in a water bath at $20\text{--}22^\circ\text{C}$. Reaction times are given in Table 9. For runs I-1 to I-4, the final pH values were measured (Table 9).

On terminating each run, the precipitate-rich gel was separated and dried at room temperature. The crystals were examined under a binocular microscope and scanning electron microscope (Biology

TABLE 9

Initial reactant concentrations, precipitate mineralogy and run times for selected gel experiments.

Columns 1-3: Cation concentrations and Mg/Ca ratio in the supernatant solution. ([] in molarities).

Column 4: Initial pH of gel. The final pH of runs I-1 to I-4 is in parentheses.

Column 5: Mineralogy of the resultant precipitate, in order of decreasing abundance. A=Aragonite; C=Calcite, subscripts indicate mole % MgCO_3 .

Column 6: Run time in days.

In all runs, the initial sodium metasilicate and sodium bicarbonate concentrations were 0.2M.

RUN	[Mg]	[Ca]	[Mg]/[Ca]	pH	Mineralogy	Time
B-2	0.04	0.2	0.2	8	C ₁₋₂ ,A	14
B-4	0.08	0.2	0.4	8	C ₁₋₂ ,A	39
B-8	0.16	0.2	0.8	8	C ₂₋₄ ,A	14
B-10	0.2	0.2	1	8	A,C ₄₋₆	14
B-20	0.4	0.2	2	8	A,C ₆₋₈	14
B-30	0.6	0.2	3	8	A,C ₆₋₈	14
A-5	-	0.2	-	8	C	25
G-1	0.2	0.04	5	9.5	A,C ₁₃₋₁₅	18
G-2	0.5	0.05	10	9.5	A,C ₁₃₋₁₅	18
G-3	0.2	0.04	5	8	A,C ₁₃₋₁₅	18
G-4	0.5	0.05	10	8	A,C ₁₃₋₁₅	18
I-1	0.2	0.04	5	8.2 (7.0)	A,C ₁₄₋₁₆	18
I-2	0.2	0.06	3.3	8.2 (6.8)	A,C ₆₋₈	18
I-3	0.2	0.1	2	8.0 (6.6)	C ₄₋₅ ,A	18

TABLE 9 (Cont.)

RUN	[Mg]	[Ca]	[Mg]/[Ca]	pH	Mineralogy	Time
Seeded Gels						
C-2	0.01	0.01	1	8	A, C ₂₋₄ on dolomite seed.	14
C-3	0.01	0.01	1	8	A, C ₂₋₄ on dolomite seed.	14
Sodium sulphate (0.1M) in gel.						
E	-	0.10	-	8	Gypsum	43

Department, Carleton University). Mineral phases in the mixture were identified by means of an X-ray diffractometer. MgCO_3 concentrations in calcites were estimated from the shift in the $d(211)$ peak from that of pure calcite $d(211) = 3.035$; A.S.T.M. Catalogue #5-586, and interpolation onto established $d(211)$ versus mole% MgCO_3 curves (Goldsmith and Graf, 1958; summarized in Milliman, 1974, p.28). These values are probably accurate to within 1% to 2%.

In two runs (C1 & C2), small calcite and dolomite seed crystals (cleaved rhombs) were placed in the gel prior to setting.

A single run was also made using Na_2SO_4 in place of bicarbonate (E-1).

Results:

In most runs, fine precipitate appeared in the upper few millimeters of the gel column within two or three days, and Liesegang rings usually developed thereafter (Fig.65). The composition of selected runs are given in Table 9. Run A-5 (Fig.65) serves to illustrate the basic processes of crystal growth gel media. Continuous diffusion of Ca^{2+} into the gel eventually leads to saturation with respect to calcite and crystal nucleation. Nucleation and growth initially is rapid and crystals are small but well formed. The number of crystals growing in the gel decreases, and their size increases with distance from the gel interface. This can be attributed to decreasing diffusion rates and concentration gradients, and hence fewer nucleation sites. Henisch (1970) has also noted that aging of the gel tends towards fewer nucleation sites, the lower parts of the gel column being older than those higher up.

Measurements of pH before and after reaction show a marked decrease

in values (Table 9). A drop of 2-3 pH units is due to hydrogen ion build up from the reaction:

Here CO_3^{2-} is consumed during carbonate precipitation and the equilibrium is pushed to the right.

In mixed Mg-Ca solutions both aragonite and calcite precipitated in different proportions, apparently depending on the Mg/Ca ratio rather than (as Lippmann, 1973, p.110 suggested) on the absolute magnitude of the Mg concentration. However, the proportion of MgCO_3 in solid solution with calcite does appear to be controlled by the concentration of Mg, being highest (15-16 mole %) in more concentrated solutions.

Calcite:

Crystals are typically rhombohedral, ranging in size up to 1 mm (Pl.16, Figs.1 & 2). Crystal terminations appear ragged when viewed at high magnifications, as a result of overlapping calcite growth layers. The vertical steps between successive layers are about 0.5 to 2 μm high in Pl.16, Fig.2.

Unlike most other gel-grown crystals, calcite does not displace the gel during growth, but permeates the gel structure, leaving it intact. This renders gel-grown calcites turbid, although they still retain well-formed crystal faces. X-ray diffraction patterns show that the included gel has little effect on the lattice structure of the calcite. Up to 1.7% SiO_2 has been reported in some calcites grown in this manner (Henisch, 1970).

Aragonite:

Aragonite predominates over calcite in runs with Mg/Ca ratios

greater than 2. In the upper level of the gel column (close to the gel interface), rapid precipitation results in extremely fine aragonite needles. Lower in the gel column, radial spherulites or radial ooids form, containing needles 60 μm and more in length (Pl.16, Figs. 3 & 4). The importance of chemical microenvironments in the gels is illustrated in Fig. 4, where aragonite needles have overgrown earlier-formed calcite crystals. The Mg/Ca ratio of the initial supernatant liquid in this particular run was 0.6, and the first phase of precipitation favoured calcite (with 2-4 mole% MgCO_3), i.e. Mg-poisoning of the actively growing calcite surface was minimal. However, the subsequent removal of calcium during precipitation would have resulted in an increase in the local Mg concentration gradient, to the point where precipitation of calcite was entirely inhibited and aragonite became the stable phase. Continued precipitation of aragonite (which rarely contains more than 1-2% Mg) would have further increased the local Mg/Ca ratio.

Seeded Experiments:

Calcite was found to nucleate on both calcite and dolomite seeds. Epitaxial growth of calcite rhombs onto dolomite surfaces shows that growth patterns are controlled to a large extent by substrate defects, especially cleavage fractures (Pl.16, Fig. 5).

Gypsum:

Large, clear radiating crystals of gypsum grew on the gel interface rather than within the gel column (Pl.16, Fig. 6). This would necessitate upward diffusion of SO_4^{2-} from the gel into the supernatant liquid. Crystal growth rates were subsequently much slower than those observed for calcite and aragonite (43 days). On the other hand, these

slow upward diffusion rates are compensated for by a concomitant decrease in nucleation sites, giving rise to fewer, but larger, crystal clusters.

Discussion:

This brief investigation was not embarked upon in the hope of providing solutions for such long unanswered questions as to the why and wherefore of dolomite formation under conditions of surface temperatures and pressures. Rather, the gel experiments were undertaken to evaluate their possible relevance to problems of carbonate formation, and in particular dolomite formation. The eventual synthesis of dolomite, or protodolomite, has little scientific value unless the reaction system producing this material can be monitored, providing information that might lead to an understanding of reaction mechanisms and kinetics. The use of gels as a growth medium for carbonates has such a potential.

One of the most attractive attributes of this method is the ability to control crystal nucleation, especially under high concentrations and, in the case of the Mg-Ca carbonates, high Mg/Ca ratios. The high MgCO_3 content of some calcites produced in these experiments (up to 16 mole %) is possibly a function of slower nucleation and precipitation rates; if dehydration of the Mg ion in aqueous solution is a rate-determining step; the longer residence time of these cations adsorbed onto the crystal surface will favour this dehydration reaction. By comparison, the MgCO_3 content of calcites is well above that documented by Lippmann (1973, p.114) for experiments that utilized the slow decomposition of potassium cyanate in aqueous solutions (i.e. a maximum of 5 mole % at 20°C).

By quantitatively examining changes in pH together with concentration changes in other ionic species with time, the kinetics of mixed Mg-Ca

carbonate reactions over a range of MgCO_3 contents might be obtained. The highest MgCO_3 content obtained thus far is 16 mole %. With slower diffusion rates (i.e. increased gel densities) and/or use of a buffer within the gel, the MgCO_3 content possibly could be pushed even higher. Another approach might be the addition of organic compounds such as amino acids or enzymes, because organic complexes may also serve to reduce precipitation rates. Calcites with up to 12.5% MgCO_3 have been made in this way (Kitano and Kanamori, 1966).

A further potential use of this method is the determination of Sr partition coefficients in Mg calcites grown at low temperatures, by spiking the solutions with strontium salts. Little data are available at the present time for partition coefficients in the high-Mg calcites, especially those approaching dolomite composition. Accurate determination of this coefficient in these various phases is important to the study of Sr distributions in modern and ancient carbonate rocks, especially with respect to the recognition of primary depositional and diagenetic facies.

PLATE 16

- Figure 1 and 2. S.E.M. photomicrographs of gel-grown rhombohedral low-Mg calcite crystals. Vertical steps between successive growth layers are less than 5 μm . Bar scale represents 40 μm .
- Figure 3. S.E.M. photomicrographs of gel-grown aragonite spherulite. Bar scale represents 40 μm .
- Figure 4. S.E.M. photomicrograph of intergrown calcite crystals (note successive growth layers) overgrown by radiating aragonite needles more than 60 μm in length. Bar scale represents 40 μm .
- Figure 5. S.E.M. photomicrograph of rhombohedral dolomite seed crystal with gel-grown epitaxial calcite overgrowths (calcite has 2-4 mole % MgCO_3). Note control on growth pattern by substrate cleavage. Bar scale represents 300 μm .
- Figure 6. S.E.M. photomicrograph of radiating gel-grown gypsum crystals. Bar scale represents 300 μm .

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APPENDIX B

ASPECTS OF THE LATE QUATERNARY GEOLOGY OF
THE BELCHER ISLANDS

The erosive action of continental glaciers during the late Quaternary followed by rapid post-glacial uplift resulted in the unique Belcher Islands landscape seen today. Ice-moulded structures such as roches moutonnées are common, especially in the igneous units, such as the diabase sills that were intruded into less resistant sedimentary rocks; their steep plucked surfaces face south, indicating that ice movement was largely in this direction.

A veneer of unconsolidated debris covers low-lying areas and fills depressions in several places on the Belchers, although nowhere are original glacial deposits evident. Deposits are composed predominantly of material derived from the local bedrock. Many of the lithologies are distinctive and the source formation can easily be established. However, a large number of erratics of granitic composition (e.g. granite, gabbro) were probably derived from a cratonic source north or east of the Belcher Islands. Clasts of Paleozoic limestone are rare.

Raised beaches are well developed along coastal areas and around lakes now cut off from the sea as a result of uplift. Beach deposits are composed mostly of reworked glacial debris. In some sections over 100 raised beaches have formed within a vertical interval of about 50 m. Boulder ridge structures that are associated with the beaches may represent storm or ice ridges. Stone rosette pavements also occur and are similar to examples along modern beaches in the Belcher Islands.

Glacial Striae and Chatter Marks

Striae and sets of chatter marks (concave in the direction of ice movement) are prominent on polished bedrock surfaces; the latter structures provide evidence for the sense of ice movement. Azimuths of a number of these structures were measured in sections along the west coast of Turkarak Island (opposite the old Hudson Bay outpost), and also on Bradbury Island (summarized in Fig.66). The suggestion by Jackson (1960) that ice movement in this region was largely towards the southwest is confirmed here (approximately 80% of all measurements). However, a number of striae and chatter marks indicate some movement to the west and north-west. In places where the relative 'ages' of striae can be established (by noting cross-cutting relationships), those trending west or north-west usually cut southwest-trending structures. The patterns illustrated in Fig.66 also suggest the possibility that there was a regular clockwise shift in ice-flow, from southwest to northwest. In comparison, Andrews and Falconer (1969) have shown that the most effective direction of ice movement in the Ottawa Islands (although not the oldest direction) was to the north or north-northwest. In addition, several of their sampling stations show a counter-clockwise decrease in relative age, and hence ice-flow (i.e. towards the west).

These differences may reflect close proximity of the Belcher Islands to an ice center which, according to Andrews and Falconer (1969) was located in the Richmond Gulf area. Alternatively, the directions represented by the Belcher structures may reflect movement of ice towards the waterway that cut the ice sheet in two some 7000 to 8000 years B.P. (from Hudson Strait to James Bay).

Figure 66a. Plot of glacial striae azimuths, measured in sections opposite the old Hudson Bay outpost, west Tukarak Island. Note dominant cluster in the southwest-northeast sector.

b. Glacial chatter mark azimuths indicating the most effective direction of ice-flow was towards the southwest (rose diagram shows that 85% of these measurements cluster in this sector). Measurements taken on Bradbury Island.

Faunal Elements of the Tyrrell Sea

Beach sand and boulder beds on the Belcher Islands were deposited in the post-glacial Tyrrell Sea. Fossiliferous sands are locally exposed in stream cuts and also along shores where tidal flats are now exposed as a result of uplift. The faunal content of two samples is given in Tables 10 and 11. Sample G.10 was collected approximately 45 m above mean high water level in a stream cut at the southern end of Isaacs Lake, Tukarak Island. Sample G.5 was collected at present low tide level on the east side of a small island, south of McLeary Point. Foraminifera and Ostracod identifications (Table 11) were made by C. Rodriguez, and samples were prepared by P. Jones.

Both samples are dominated by Mya truncata and Hiatella arctica. Chlamys and Astarte are also common, some of the Chlamys attaining lengths of 10 cm. The molluscan content of G.10 is very similar to that of the much younger G.5, suggesting that salinity and temperature conditions were also similar. At present, the optimum water depths of these forms in Hudson Bay is 35 to 60 m, although they have been found at depths exceeding 100 m (Wagner, 1969). Although sample G.5 was collected at present sea level, the faunal composition indicates that it has probably been uplifted from greater depths.

Carbon-14 age dates are not available for the Belcher material, but a reasonable estimate of sample ages can be made by comparison with uplift-rate curves calculated for the Ottawa Islands by Andrews and Falconer (1969). For example, at 45 m elevation, Sample G.10 is probably 4000 to 5000 years old. Beach deposits at the maximum elevation in the Belcher Islands (137 m on Tukarak Island) could be older than 7000 years B.P. Although these dates seem reasonable, it should be noted that the

proximity of the Belcher Islands to the postulated ice-center in Richmond Gulf area could have resulted in the Belchers having different rates of post-glacial uplift.

TABLE 10

Macrofauna in two samples collected from late Quaternary deposits in the Belcher Islands.

PELECYPODA	Sample	
	G.5	G.10
<u>Astarte crenata</u> Gray	x	x
<u>Chlamys islandica</u> Muller	x	x
<u>Clinocardium ciliatum</u> Fabricus	x	
<u>Hiatella arctica</u> Linne	x	x
<u>Mya pseudoarenaria</u> Schlesch	x	x
<u>Mya truncata</u> Linne	x	x
<u>Mytilus edulis</u> Linne		x
<u>Portlandia arctica</u> Gray	x	
GASTROPODA		
<u>Acirsa borealis</u> Lyell		x
<u>Acmaea testudinalus</u> Muller	x	x
<u>Buccinum tenue</u> Gray	x	
<u>Hydrobia minuta</u> Iotten	x	x
<u>Margarites groenlandicus</u> Gmelin	x	x
<u>Puncturella nochina</u> Linne		x
BRACHIOPODA		
<u>Hemithiris psittacea</u> Gmelin	x	x
CIRRIPEDIA		
<u>Balanus balanus</u> Linne	x	x
ECHINODERMATA		
Spines, indet.	x	x
AMPHINEURA		
Fragments, plates, indet.	x	x

Foraminifera and Ostracoda assemblage of Sample G.5 collected from an unnamed island, due south of McLeary Point, Tukarak Island. Identifications by C. Rodriguez, Geology Dept. Carleton University.

FORAMINIFERA

Astrononion gallowayi Loeblich and Tappan

Brizalina pseudopunctata (Hogglund)

Bucella calida (Cushman and Cole)

Bucella frigida (Cushman)

Bucella hannai (Phleger and Parker) arctica voloshinova

Cassidulina obtusa Williamson

Cibicides lobatulus (Walker and Jacob)

Dentalina frobisherensis Loeblich and Tappan

Dentalina ittai Loeblich and Tappan

Dentalina melvillensis Loeblich and Tappan

Egerella advena (Cushman)

Elphidiella arctica (Parker and Jones)

Elphidium asklundi Brotzen

Elphidium bartletti Cushman

Elphidium excavatum (Terquem)

Elphidium frigidum Cushman

Elphidium subarcticum Cushman

Eoeponidella pulchella (Parker)

Epistominella takayanagii Iwasa

Fissurina immoderata (Buchner)

Fissurina serrata (Schlumberger)

Fissurina trigono-laevigata (Balkwill and Millet)

Florilus atlanticus (Cushman)

Fursenkoina loeblichi (Feyling-Hanssen)

Fursenkoina sp.

Glabratella sp.

Islandiella helenae Feyling-Hanssen and Buzas

Islandiella inflata (Gudina)

Islandiella norcrossi (Cushman)

Lagena deformata Buchner

Lagena gracillima (Seguenza)

Lagena semilineata Wright

TABLE 11 (Cont.)

Neogloboquadrina sp.

Nonionella labradorica (Dawson)

Oolina apiopleura (Loeblich and Tappan)

Oolina borealis Loeblich and Tappan

Oolina caudigera (Wiesner)

Oolina lineata (Williamson)

Oolina melo d'Orbigny

Oolina scalariforme-sulcata (Wiesner)

Oolina squamosa (Montagu)

Oolina striatopunctata (Parker and Jones)

Parafissurina fovigera (Buchner)

Parafissurina himatiostoma Loeblich and Tappan

Patellina corrugata Williamson

Protoelphidium ? orbiculare (Brady)

Quinqueloculina agglutinata Cushman

Quinqueloculina arctica Cushman

Saccamina atlantica (Cushman)

Trifarina hughesi (Galloway and Wissler)

Triloculina trihedra Loeblich and Tappan

(Total Number of Species = 51)

OSTRACODA

Baffinicythere emarginata (Sars)

Baffinicythere howei Hazel

Cythere lutes Muller

Cytheropteron arcticum Neale and Howe

Cytheropteron pyramidale Brady

Cytheropteron nodosoalatum Neale and Howe

Cytheropteron nodosum Brady

Cytheropteron paralatissimum Swain

Cytheropteron sp. cf. C. montrosiense Brady and Robertson

Cytherura granulosa Brady and Robertson

Elofsonella concinna (Jones)

Elofsonella neoconcinna Bassiouni

Eucytheridea bradii (Norman)

Finmarchinella barentzovoensis (Mandelstam)

(TABLE 11 Cont.)

Finmarchinella curvicostata Neale
Finmarchinella finmarchia (Sars)
Hemicythere villosa (Sars)
Hemicytherura clathrata (Sars)
Ilyocypris gibba (Ramdohr)
Munseyella mananensis Hazel and Valentine
Normanicythere leioderma (Norman)
Palmenella limicola (Norman)
Paracytheridea sp.
Semicytherura mananensis (Hazel and Valentine)

(Total Number of Species = 22)

APPENDIX C

Point-count data for sandstones of the Fairweather, McLeary, Mavor, Rowatt, Mukpollo, Omarolluk and Loaf Formations.

Notes :

1. An automatic stage and counter was used, and 300-400 points counted for each sample. All values are quoted as number percent.
2. No distinction is made between undulose and non-undulose monocrystalline quartz (the majority of quartz grains show undulose, or undulose extinction).
3. For the purpose of modal analysis two principal cement types are recognized: carbonate (mainly dolomite) and non-carbonate (quartz as syntaxial overgrowths, feldspar, chlorite and minor hematite). A break-down of carbonate cement types is given in Chapter 5.
4. In some dolostones, the distinction between cement and framework is difficult because of recrystallization. In such cases they are included in a single category; recrystallized framework and cement.
5. The terrigenous component of sandstones is quoted as a percentage of total count.
6. Abbreviations:

Bi = biotite	Ep = epidote
Fs = feldspar	Mu = muscovite
Sp = sphene	Tm = tourmaline
tr.= trace	Zr = zircon
7. Samples located relative to formation boundaries, or, relative to the base or top of the section measured.

FAIRWEATHER FORMATION

Sample Number	R49	R88	R89	R91a	R92	R93	R118	R119	R121	R122
Rock Name	Subarkose	Subarkose	Dolomitic Quartz arenite	Quartz arenite	Dolomitic Quartz arenite	Quartz arenite	Quartz arenite	Dololotharenite	Lithic arenite	Dolomitic Subarkose
Monocrystalline Quartz	48.1	51.9	77.3	63.1	69.0	84.6	75.5	59.8	25.5	31.8
Polycrystalline Quartz	7.2	3.7	6.3	6.9	9.0	1.0	8.8	0.7	15.9	
K Feldspar	13.9	14.6	1.0	2.3	2.2	1.0	0.5	2.4	3.8	8.9
Plagioclase	4.7	3.0						2.2	1.1	2.5
Chert	8.6	4.4	1.0	2.6		1.0		0.2	2.3	0.3
Volcanic Rock Fragments	0.4	0.3							11.3	
Carbonate Clasts	8.6		0.2			1.9			0.2	
Recrystallized Framework & Cement								32.4	35.9	19.1
Carbonate Cement	1.0	0.3	12.4	3.3	12.1	2.9	7.1			
Opakes	1.4	tr.						1.5	0.2	5.7
Other Heavy Minerals	tr. Ep. Mu.	tr. Ep.						tr. Ep. Mu.		0.3 Ep.
Non-carbonate Cement	14.7	21.8	1.0	21.9	7.7	7.1	8.1	0.2	2.8	30.6
Terrigenous Component	82.5	77.6	85.6	74.9	80.0	87.6	84.8	65.3	59.9	44.2
Authigenic Minerals				Tm.	Tm.					
Sample Location	Section 1, 242 m above base of formation.	Section 4, 50 m above base of section.	Section 4, 56 m above base of section.	Section 4, 90 m above base of section.	Section 4, 104 m above base of section.	Section 4, 112 m above base of section.	Section 6, 10 m above base of section.	Section 6, 8 m above base of section.	Section 6, 20 m above base of section.	Section 6, 16 m above base of section.

FAIRWEATHER FORMATION

Sample Number	R123	R419	R519	R523	R526	R527
Rock Name	Subarkose	Dolomitic Quartz arenite	Dololitharenite	Dololitharenite	Quartz arenite	Subarkose
Monocrystalline Quartz	61.0	78.4	38.2	56.6	96.4	69.5
Polycrystalline Quartz	15.5			4.9	1.0	1.9
K Feldspar	4.6	5.0	4.7	1.6	0.2	7.0
Plagioclase	4.4	0.7	1.4	0.5		6.2
Chert	2.4			0.5	tr.	1.7
Volcanic Rock Fragments	1.5					0.2
Carbonate Clasts						
Recrystallized Framework & Cement			47.9	35.8		
Carbonate Cement	1.1	11.0			1.0	6.2
Opakes	0.2			0.5		0.2
Other Heavy Minerals	tr. Zr					
Non-carbonate Cement	9.4	4.8	7.8		1.4	6.0
Terrigenous Component	89.4	84.1	44.3	64.1	97.6	86.3
Authigenic Minerals		Tm.			Tm.	
Sample Location	Section 6, 25 m above base of section.	Section 31, 37 m above base of section.	Section 39, 290 m below top of formation.	Section 39, 232 m below top of formation.	Section 39, 120 m below top of formation.	Section 39, 68 m below top of formation.

MCLEARY FORMATION

MAVOR
FORMATION

Sample Number	R33	R39	R133	R141	R144	R146	R199	R390	R392	R395
Rock Name	Sandy Dolarenite	Sandy Dolarenite	Sandy Dolarenite	Sandy Dolarenite	Dololitharenite	Dololitharenite	Sandy Dolarenite	Quartz arenite	Dololitharenite	Sandy Dolarenite
Monocrystalline Quartz	19.4	24.2	27.8	47.2	60.2	48.1	40.7	88.8	47.8	19.4
Polycrystalline Quartz		0.4		0.5	5.8	4.3	4.2	2.5		1.0
K Feldspar	4.5	2.0	3.7	1.0	1.8	0.8	1.2	2.2	7.1	
Plagioclase								0.2	4.0	
Chert						tr.				
Volcanic Rock Fragments										
Carbonate Clasts		1.0		27.8	15.8	14.9	14.2			57.8
Recrystallized Framework & Cement	75.8	71.3	67.9				39.7		41.0	
Carbonate Cement				22.6	16.4	31.9		6.0		21.8
Opakes	0.1				tr.	tr.		tr.		
Other Heavy Minerals										
Non-carbonate Cement		1.2	0.3	1.0						
Terrigenous Component	23.9	26.6	31.5	48.7	67.8	53.2	46.1	93.7	58.9	20.4
Authigenic Minerals										Fs.
Sample Location	Section 1, 38 m below top of formation.	Section 1, 125 m below top of formation.	Section 8, 5 m below top of formation, grid reference 293154.	Section 9, 30 m below diabase sill.	Section 9, 43 m below diabase sill.	Section 9, 50 m below diabase sill.	Section 6, 127 m below top of section.	Section 27, 314 m below top of formation.	Section 27, 290 m below top of formation.	Section 27, 70 m below top of Mavor Formation.

MCLEARY FORMATION

Sample Number	R420	R426	R427	R457	T41	T41/2
Rock Name	Dololitharenite	Dolomitic Subarkose	Dolomitic Quartz arenite	Quartz arenite	Sandy Dolarenite	Dololitharenite
Monocrystalline Quartz	44.5	75.6	66.4	89.3	48.3	51.3
Polycrystalline Quartz			1.0	1.2		1.4
K Feldspar	1.9	7.2	2.8	1.7	0.3	1.1
Plagioclase		2.9				
Chert	6.9					
Volcanic Rock Fragments						
Carbonate Clasts			2.2	3.2	16.5	13.6
Recrystallized Framework & Cement	46.7				34.9	32.6
Carbonate Cement		14.0	27.8	4.5		
Opaques						
Other Heavy Minerals	tr. Ep.Sp.		tr. Ep.		tr.	tr. Ep.
Non-carbonate Cement						
Terrigenous Component	53.3	85.7	70.2	92.2	48.6	53.8
Authigenic Minerals						
Sample Location	Section 31, 430 m below top of formation.	Section 31, 300 m below top of formation.	Section 31, 270 m below top of formation.	Section 35, 255 m below top of formation.	Southeast Tukarak Island, Collected by J.A.Donaldson	Southeast Tukarak Island, Collected by J.A.Donaldson

ROWATT FORMATION

Sample Number	R224	R232	R345	R405	R406	R440	R441	R464	R465	R466
Rock Name	Sublitharenite	Dolomitic Arenite	Sandy Dolarenite	Dolomitic Quartz Arenite	Sandy Dolarenite	Dololitharenite	Dolomitic Quartz Arenite	Dololitharenite	Quartz arenite	Dololitharenite
Monocrystalline Quartz	80.9	60.6	28.1	87.3	26.6	63.2	78.0	75.2	82.4	62.8
Polycrystalline Quartz	0.3			0.5		0.4	0.3		0.9	
K Feldspar	1.5	0.7	1.0	0.3		0.8	0.9	0.6	2.1	1.5
Plagioclase									0.3	
Chert				0.8	9.6	2.1	0.6	0.9	0.3	6.3
Volcanic Rock Fragments										
Carbonate Clasts	6.2		55.6			22.3	2.5	7.9	0.6	0.6
Recrystallized Framework & Cement		19.2			45.5					20.7
Carbonate Cement	6.2		15.3	10.9		9.9	8.0	11.5	7.8	
Opagues				0.3						
Other Heavy Minerals			0.1 Bi.	tr Ep.	tr Zr.					
Non-carbonate Cement	4.9	19.5			18.3	1.2	9.6	3.9	5.7	8.1
Terrigenous Component	82.7	61.3	29.1	88.9	36.2	66.5	78.3	76.7	86.0	70.6
Authigenic Minerals	Fs.				Fs.	Fs.	Fs.			
Sample Location	Section 16, south of McLeary Point, 110 m below top of section.	Section 16, 42 m below top of section.	Section 23, eastern Flaherty Island. 105 m above base of section.	Section 30, northeastern Tukarak Island. 300 m below top of section.	Section 30, 200 m below top of section.	Section 34, northeastern Innetalling Island. 120 m below top of section.	Section 34, 141 m below top of section.	Section 35, Laddie Harbour. 285 m below top of formation.	Section 35, 250 m below top of formation	Section 35, 190 m below top of formation.

ROWATT
FORMATION

MUKPOLLO FORMATION

Sample Number	R468a	R468b	R94	R95	R97	R98	R99	R474
Rock Name	Quartz Arenite	Quartz Arenite	Quartz Arenite	Quartz Arenite	Quartz Arenite	Quartz Arenite	Sublitharenite	Quartz Arenite
Monocrystalline Quartz	84.2	88.5	71.8	80.9	87.8	88.0	59.9	81.4
Polycrystalline Quartz	7.3	0.3	8.2	8.2	6.8	3.6	16.9	9.4
K Feldspar	2.3	1.3	1.9	2.3	2.8	3.6	tr	5.1
Plagioclase								0.3
Chert	0.6	1.9		1.3		tr	8.0	0.8
Volcanic Rock Fragments								
Carbonate Clasts	}							
Recrystallized Framework & Cement								
Carbonate Cement			0.6			1.2	3.4	
Opakes			0.6					
Other Heavy Minerals								tr Ep.Zr
Non-carbonate Cement	5.6	8.0	16.9	7.2	2.8	3.6	11.9	3.1
Terrigenous Component	94.4	92.0	81.9	92.7	97.4	95.2	84.8	97.0
Authigenic Minerals								
Sample Location	Section 35, Laddie Harbour. 70 m below top of formation.	Section 35, 75 m below top of formation.	Section 5, northeastern Innetalling Island. 60 m below top of formation.	Section 5, 55 m below top of formation.	Section 5, 20 m below top of formation.	Section 5, 1.5 m below top of formation.	Section 5, 0.5 m below top of formation.	Section 35, Laddie Harbour. 100 m below top of formation.

OMAROLLUK FORMATION

Sample Number	R250	R251	R253	R255	R256	R354	R358	R455	R456	R529
Rock Name	Calcite cemented greywacke	Feldspathic greywacke	Volcanic greywacke	Arkose	Feldspathic greywacke	Tuff	Volcanic greywacke	Feldspathic greywacke	Feldspathic greywacke	Feldspathic greywacke
Monocrystalline Quartz	13.1	32.3	22.8	30.4	29.0	3.0	25.2	39.8	47.1	26.0
Polycrystalline Quartz	0.5	0.3	0.9	10.0	0.6		9.6	2.3	1.0	4.7
K Feldspar	0.7	6.6	8.6	42.6	17.2	2.0	13.1	29.4	18.2	11.7
Plagioclase	10.7	7.8	11.1		5.3	1.0	6.0	6.4	3.7	7.7
Chert	1.5	2.1	1.2		3.0			3.0	9.4	2.0
Volcanic Rock Fragments	10.7	3.3	27.2	4.3	16.0	69.0	21.3	0.7		7.3
Carbonate Clasts	45.0	6.3	0.6		12.4		5.7			
Recrystallized Framework & Cement										
Carbonate Cement										
Opakes	0.2		0.6		0.6				0.7	
Other Heavy Minerals	tr. Mu.	0.3	0.3	2.8	1.8		tr. Tm.	1.0		2.2
Matrix	17.5	41.3	26.5	9.7	14.2	25.0	19.1	17.4	19.9	38.0
Terrigenous Component										
Authigenic Minerals										
Sample Location	Section 18, Gilmour Peninsula, Grid reference 282425.	Section 18, Grid reference 283425.	Section 18, grid reference 285424.	Section 18, grid reference 285424.	Section 18, grid reference 285423.	Basal Omarolluk Formation, western shore of Rowatt Harbour, due west of Gilmour Peninsula. Grid reference 272435.	Section 24, Youngs Point, grid reference 284464.	Twin Cairns Island	Twin Cairns Island	Section 18, grid reference 298423.

OMAROLLUK
FORMATION

LOAF FORMATION

Sample Number	R531	R537	R167	R171	R175	R179	R454	R189
Rock Name	Feldspathic greywacke	Feldspathic greywacke	Arkose	Arkose	Arkose	Arkose	Arkose	Arkose
Monocrystalline Quartz	23.2	29.6	42.6	35.8	37.8	45.7	42.0	45.8
Polycrystalline Quartz	4.7	6.4	1.6	0.9	1.9	1.6	2.7	3.1
K Feldspar	20.1	9.6	33.8	34.0	26.0	29.6	33.7	22.7
Plagioclase	5.0	3.5	6.3	3.7	2.5	2.6	4.7	3.1
Chert		2.3	4.7	3.7	4.8	3.5	2.0	7.8
Volcanic Rock Fragments	13.5	6.8	1.9	1.6	1.3	0.3	2.0	0.9
Carbonate Clasts	1.3		1.3	7.5	14.6	7.1	7.7	
Recrystallized Framework & Cement								
Carbonate Cement								
Opakes	0.3		0.6	5.3		0.3		4.0
Other Heavy Minerals	1.9	1.0	tr.	tr.	0.6	0.3		0.2
Matrix	30.1	40.8	7.3	7.5	10.5	9.0	5.7	12.1
Terrigenous Component								
Authigenic Minerals								
Sample Location	Section 18, grid reference 309428. Walton Island.		North shore, Loaf Island.	North shore, Loaf Island.	Northwest shore, Loaf Island.	Eastern shore, Loaf Island.	Northeast shore, Loaf Island.	Eastern shore, Loaf Island.

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*The hills are shadows, and they flow
From form to form, and nothing stands;
They melt like mist, the solid lands,
Like clouds they shape themselves and go.*

Tennyson

'In Memoriam A.H.H.'

CXXIII. St.2.

HUDSON BAY

TUKARAK ISLAND

LEGEND

	LOAF FM.	Red & grey cross-bedded arkoses.
	OMAROLLUK FM.	Greywackes, shales, minor conglomerates.
	FLAHERTY FM.	Tholeiitic basalts: massive & pillowed flows, volcaniclastics.
	HAIG INTRUSION	Intrudes all units below Omarolluk Fm.
	KIPALU FM.	Banded argillites, micrites, jaspers, ironstones.
	MUKPOLLO FM.	Quartz arenites.
	ROWATT FM.	<u>U.Mbr</u> : Brecciated dolostones. <u>L.Mbr</u> : Sandstones, dolostones.
	LADDIE FM.	Red & green argillites & shales.
	COSTELLO FM.	Rhythmically bedded carbonates & shales, minor calcarenites.
	MAVOR FM.	Stromatolitic dolostones, shaley dolostones.
	TUKARAK FM.	<u>U.Mbr</u> : Red mudstones & dololutes. <u>L.Mbr</u> : Fine sandstones.
	MCLEARY FM.	<u>U.Mbr</u> : Stromatolitic dolostones. <u>M.Mbr</u> : Grainstones, algal laminates, beachrock. <u>L.Mbr</u> : Dolostones beach rosettes.
	FAIRWEATHER FM.	<u>U.Mbr</u> : Sandstones <u>L.Mbr</u> : Pisolitic dolostones, desiccated mudstones.
	ESKIMO FM.	Tholeiitic basalts; massive flows.
	KASEGALIK FM.	Stromatolitic dolostones, mudstones, sulphate & halite casts.

SYMBOLS

	Geologic Boundary		Structural dip
	Inferred geologic boundary		Cleavage
	Fault		Parasitic fold
	Inferred fault		Structural trends
	Axial trace of syncline		
	Axial trace of anticline		

GEOLOGY OF TUKARAK & OUTLYING ISLANDS, BELCHER ISLANDS.

Scale: 1:50,000

0 1 2 3 4 5 kilometers

Geology by G.D.Jackson (1960) & B.D.Ricketts (1979).