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LA THÈSE A ÉTÉ
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THE UPPER SILURIAN BARLOW INLET FORMATION,
CORNWALLIS ISLAND, ARCTIC CANADA

by

Jeffrey J. Packard

A thesis
presented to the University of Ottawa
in partial fulfillment of the
requirements for the degree of
Doctor of Philosophy
in
Geology

OTTAWA, Ontario, 1985

Jeffrey J. Packard, Ottawa, Canada, 1985.





Cape Hotham, bearing W. by S.

The latter was a remarkable headland, having at its extremity two small table hills, somewhat resembling boats turned bottom upwards, and was named CAPE HOTHAM, after REAR-ADMIRAL THE HONOURABLE SIR HENRY HOTHAM, one of the Lords Commissioners of the Admiralty.

To the land on which Cape Hotham is situated, and which is the easternmost of the group of islands, (as we found them to be by subsequent discovery,) in the Polar sea, I gave the name of CORNWALLIS ISLAND, after ADMIRAL THE HONOURABLE SIR WILLIAM CORNWALLIS, my first naval friend and patron; and an inlet, seven miles to the northward of Cape Hotham, was called BARLOW INLET, as a testimony of my respect for SIR ROBERT BARLOW, one of the Commissioners of His Majesty's navy.

excerpts from W.E. Parry, 1821



Frontispiece

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COMPOSITE ABSTRACT

Outcropping on Cornwallis and Griffith Islands of the central Canadian Arctic, the lower member of the Upper Silurian Barlow Inlet Formation is a thin (50 m), siliciclastic sequence, stratigraphically entombed in a 2700 m thick succession of relatively pure carbonates. The lower member overlies the Douro Formation conformably, but very abruptly; this contact represents an event horizon signalling a major late Ludlovian pulse of the Cornwallis Disturbance. The lower member itself is composed predominantly of dark mudrocks of 'black shale' aspect, and these mudrocks are interpreted as prodelta sediments deposited in an intermittently anoxic basin. A sandy facies consists of fine-grained, in part bimodally-sorted, subarkose characterized by low-inclined parallel lamination, Runzelmarken, desiccation polygons, asymmetric ripples, primary current lineation, and a Cruziana trace fossil assemblage. The subarkose facies is interpreted as beach shoreface facies.

The lower member resulted from continental mass-wasting of nearby epeirogenic uplands. The detrital source cannot be determined with certainty, but a source area to the south

and east of Cornwallis Island is suggested by the thickness distribution of the lower member. The deltaic sediments that constitute the lower member represent a part of the marine record of Caledonian diastrophism that affected the Boothia Uplift region as well as other parts of the arctic archipelago. The tectono-sedimentary record of Cornwallis Island during this time can be satisfactorily interpreted in terms of pre-epeirogenic, initial epeirogenic, epeirogenic, and immediately post-epeirogenic phases.

The upper member of the Barlow Inlet Formation contains a significant proportion of peritidal lithofacies arranged in hemicyclic series. Two shallowing-upward sequences are recognized: a Restricted Infratidal (RI) Hemicycle and a Tidal Flat (TF) Hemicycle. Each complete cycle contains four members, representing higher energy subtidal lagoon or proximal back-barrier facies through to tidal flat facies.

The RI hemicycles consist of 1) a basal biostromal member, 2) a sparsely fossiliferous mottled dolomitic carbonate member, 3) a massive carbonate mudstone member, and 4) a laminated member. The TF hemicycles consist of 1) a basal oncoidal carbonate member, 2) a mottled carbonate member, 3) a massive carbonate member, and 4) a laminated member. The hemicyclic sediments were deposited in an arid climate on a carbonate platform rimmed by crinoid sand shoals and/or stromatoporoid-dominated biostromes and buildups, subject to minimal wave stress.

The muddy lithofacies and overall setting of the platform compares favourably with present day Florida Bay, as well as ancient back-reef sequences described in the Devonian and Triassic systems.

The TF and RI hemicycles represent co-existing and intergradational environmental regimes. Depths likely did not exceed 7m anywhere on the shelf. Hemicyclicity is best explained in terms of an episodic subsidence and tidal flat progradation model; eustatic sea-level rises need not be invoked.

The Barlow Inlet Formation represents a major carbonate platform complex, over 1.2 km thick. Biocumules, herein defined as variably-shaped concentrations of in-situ organisms and contiguous sediment (collectively embracing biostromes and buildups), occur throughout this complex. They are considered to have developed in depositional environments ranging from medial foreslope to restricted lagoon. Ten biocumule facies are recognized and include: lithistid sponge dominated mud mounds, mucophyllid mud mounds, diverse community mud mounds, fasciculate rugose coral biostromes, stromatoporoid - crinoid rubble skeletal buildups, dolomitized massive skeletal (?crinoid) biocumules, lamellar stromatoporoid biostromes, massive stromatoporoid biostromes, dendroid stromatoporoid biostromes, and Coenites - capped fenestral mud mounds. In addition several biocumules exhibit internal vertical

zonation of biofacies, and are here described under the term 'zoned biocumules'.

Buildups and biostromes of late Silurian-early Devonian age are poorly represented in the literature, and detailed descriptions of complete platform complexes from continental margin settings are unknown. Biocumules of the Platform Margin are most closely comparable to the reef succession of Gotland, Sweden. Platform Interior facies bear close resemblance to Middle and Upper Devonian backreef sequences. Basin Slope biocumules show similarities to late Silurian carbonate ramp buildups in Arctic Canada and foreslope sponge mud mounds of the European Jurassic.

Eight major facies associations, herein called superfacies, are recognized in the Upper Silurian carbonate member of the Barlow Inlet Formation on Cornwallis Island. These superfacies represent environments ranging from distal foreslope (in contact with anoxic basal water mass of the Cape Phillips Basin), through rimmed shelf edge to tidal flat. The environments were associated with a large attached interior platform straddling a flexure between stable craton and miogeocline. Medial foreslope, proximal foreslope, shelf edge and proximal back-barrier superfacies contain diverse buildups and biostromes, that represent biota that apparently flourished in oxygenated near-surface waters of the basin.

The platform (Barlow Inlet Platform) evolved from a precursor ramp (Douro Ramp) by structural modification commensurate with subjacent block faulting of the Boothia Horst in latest Ludlovian time. Continued spasmodic epeirogeny affecting this basement structure, controlled periodic areal enlargement and drowning of the platform. These principal forestepping and backstepping events were superimposed on longer term accretionary outbuilding and regional subsidence. In early-mid Lochkovian time the platform emerged during a third pulse of the Cornwallis Disturbance.

The distal basin slope and cyclic platform interior facies are largely indistinguishable from the C member of the Cape Phillips Formation of northern Cornwallis Island, and the lower member of the Somerset Island Formation of northern Somerset Island, respectively. However mapping contingencies dictate that formational boundaries remain unchanged.

PREFACE

In the summer of 1977, as a neophyte stratigrapher, I commenced work (with considerable apprehension) on what was then the C member of the Read Bay Formation as exposed on the eastern coast of Cornwallis Island. The weather seemed a shade cold, the cliffs somewhat precipitous, and the stratigraphic thicknesses rather overwhelming, but it was apparent even to the apprentice geoscientist that I was, that the rocks themselves had an interesting story to tell. Over 6 years, 6.2 km of section and 600 pages later, as a 'veteran' graduate student I humbly submit what is little more than an outline of the Barlow Inlet Formation story, but a contribution that I hope shall be perceived as both accurate and valuable nonetheless.

The relative merits of this dissertation, I trust, will speak for themselves. I would, however, like to address what I view as the limitations and shortfalls of this project. Prior to this study the Barlow Inlet Formation was considered, in many respects, to be lithologically similar to the Douro Formation. A logical extension of this interpretation, would be that the Barlow Inlet Formation was likewise deposited on a largely undifferentiated subtidal shelf. In fact, as this study demonstrates, the Barlow

Inlet Formation, as presently defined, comprises an extremely heterogeneous assemblage of lithofacies deposited on a highly differentiated shelf. Interpreted depositional environments range from high intertidal zone (possibly supratidal) to distal basin slope. The Barlow Inlet Formation arguably represents one of the best exposed and most extensive Late Silurian carbonate platform complexes in the world. To have been given the opportunity of studying such a broad spectrum of carbonate depositional environments, has been a greatly enriching privilege, but nevertheless it has also been very sobering. The objectives of the project compelled me first 'to see the forest' prior to studying the trees. Unfortunately the forest proved so vast that the trees themselves have been largely ignored. To paraphrase a familiar adage, I have become a jack of all lithofacies, but a master of none. In my own estimation, this lack of comprehensive treatment of any single depositional environment has proven to be an undesirable if unavoidable consequence of this study.

A more serious handicap was lack of biostratigraphic control within the Barlow Inlet Formation. The macrofauna has not provided sufficiently precise age indicators, and a lack of expertise, time and resources has precluded systematic microfossil dating. It is hoped that, in collaboration with the Geological Survey of Canada, meaningful time lines can be constructed through the Barlow

Inlet Platform Complex. Meanwhile the tectono-depositional histories presented herein, remain working hypotheses that must yet stand the 'test of time'.

Overall, I can state of the project and my time spent on it, ... RECISSAMENTUM ERAT.



UNIVERSITÉ D'OTTAWA
UNIVERSITY OF OTTAWA

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I

PAPER I: GENERAL INTRODUCTION

1.1 LOCATION OF FIELD AREA

Field work was carried out principally on eastern Cornwallis Island, in the central part of the Canadian Arctic Archipelago (FIG. 1-1a,b). Ancillary stratigraphic sections were measured on neighbouring Griffith and Somerset Islands.

1.2 SCOPE OF PROJECT

This study encompasses a stratigraphic and sedimentologic analysis of both the lower and upper members (informal) of the Upper Silurian Barlow Inlet Formation (Thorsteinsson, 1958, 1980). The lower member comprises less than 70 m of terrigenous 'black shales' that represent detritus shed from epeirogenic uplands and deposited in a paralic basin during the Cornwallis Disturbance (Thorsteinsson, 1980; Kerr, 1977). The upper member is a thick (almost 1.3 km at the type section) heterogeneous succession with lithofacies representative of environments ranging from distal foreslope to tidal flat developed on, or near, a widespread carbonate platform during a period of relative tectonic quiescence.

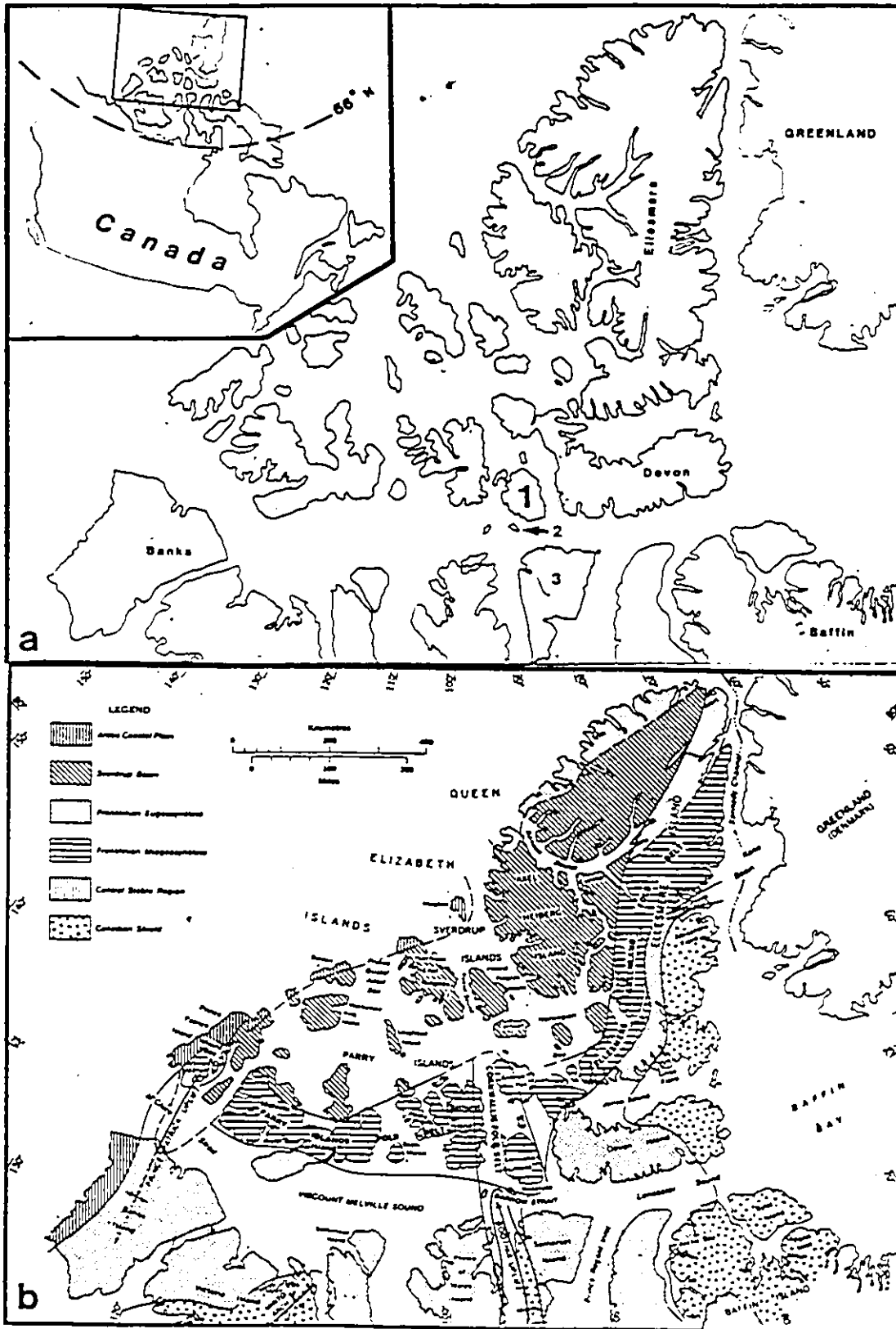


FIG.1-1a Location map of study area. 1 - Cornwallis Island, 2 - Griffith Island, 3 - Somerset Island.

FIG.1-1b Geologic Provinces of the Arctic Archipelago (modified after Morrow and Kerr, 1977)

The project objectives were to establish the basic lithostratigraphic framework for the formation, to determine the paleoenvironmental significance of contained lithofacies, and to construct a depositional model consistent with regional tectono-sedimentary histories.

1.3 PREVIOUS WORK

Due in part to its relative inaccessibility, the Barlow Inlet Formation, and in particular the upper member, has been little studied previously. The Barlow Inlet Formation (née members B and C of the Read Bay Formation) was originally defined and described by Thorsteinsson (1958), based on reconnaissance stratigraphy and mapping conducted between 1950 and 1953. Although a type section for the upper member was chosen, circumstances did not allow for its measurement or description (op. cit., p. 65), until this present study. As a result, the formation's description has remained a generalized one for over 30 years. Only two detailed stratigraphic sections have ever been published on the formation, and these cover only the basal 220 m (op. cit., p. 64-69). Brief descriptions of the Barlow Inlet Formation are found in Thorsteinsson (in Douglas 1970), Gibling and Narbonne (1978) and Thorsteinsson (1980). Mayr (1978) described strata equivalent to the Barlow Inlet Formation from the subsurface of Devon Island (E-45 well).

Works of a biostratigraphic or paleontologic nature dealing in whole or in part with the Barlow Inlet Formation include Bolton (1966), Uyeno (1977, 1980), McGregor and Narbonne (1978), Jones and Packard (1980), Rohr and Packard (1982), Jones (1979a, 1979b, 1981a, 1981b), Parkins (in press) and Narbonne and Packard (1982). More recent studies dealing with stratigraphy and facies analysis of the Barlow Inlet Formation are outlined in Packard and Dixon (1979, 1981, 1982, 1983).

The project forms a part of the ongoing research program conducted by the University of Ottawa Geology Department on Paleozoic carbonate-clastic sequences of the Arctic Archipelago. Relevant theses on Siluro-Devonian strata include Miall (1969), Jones (1975), Gibling (1978), Savelle (1979), Narbonne (1981), and Muir (1982), and research is currently being conducted by W.G. Parkins and J.L. Poey.

1.4 FIELD WORK AND METHODOLOGY

Standard stratigraphic techniques were employed in field work. Pogo sticks (Jacob's staffs) were used to determine thicknesses of most measured units.¹ Some large covered intervals were measured by chain. Confirmation of measured thicknesses was obtained by graphic methods using vertical

¹ a unit, as employed here, is subjectively defined as a stratigraphic interval that is lithologically homogeneous on the scale of individual beds or minimal packages of beds. The boundaries of units are generally sharp, and commonly emphasized by horizontal jointing.

airphotos, and this method was also used to determine the height of the base of some sections above a given datum. Field data were recorded as digital codes on 80 character 'unit cards' modified from one designed by the Geological Survey of Canada. These data were subsequently stored and manipulated on CMS (IBM's Conversational Monitor System) operated by the University of Ottawa Computing Centre's VM370. Fortran IV programs, utilized to facilitate the analysis and graphic presentation of these data, were developed by the author (the software is on file, and available from the Department of Geology, University of Ottawa).

Sampling was comprehensive; representative hand specimens of lithologies and/or fossils were taken from virtually every unit. Peels, stained thin sections and/or polished slabs were used to examine collected samples, appropriate to the size and information desired of each specimen.

The field season in the arctic is short, and there is at best a two month period (generally from June 20th - August 20th) between spring melt and inclement fall weather, when exposures of rocks are relatively snow-free and field work may be pursued profitably (comfortably?). The reader is referred to Jones (1975) for a description of the peculiarities of, and necessities employed in, arctic field geology.

The source data for this study were collected over three field seasons (1977, 1978, 1980). A total of 40 sections were measured, encompassing approximately 1400 units in 6.2 km of collective thickness (FIG. 1-2, Tables 1-1 and 1-2). Sixty one percent of this total thickness represents outcrop, and 39 % is assigned to covered intervals or felsenmeer. The average unit thickness in different sections (with more than 10 units each) ranges from 1.0 to 8.3 m.

1.5 GEOLOGICAL SETTING

The field area is located within the Cornwallis Fold Belt (FIG. 1-1b), a north-south oriented anticlinorium that forms the outer structural shell of the Boothia Uplift. The principal Barlow Inlet outcrop belt occurs on the eastern flank of the Cape Hotham anticline that trends north to northwest, subparallel and proximal to the eastern coast of the island. The exposed strata on the eastern flank have dips ranging from 15 to 50°. Periods of intense deformation affected Cornwallis Island during and after the early Devonian, and therefore the principal structural elements of the Fold belt postdate deposition of the Barlow Inlet Formation.

Although Siluro-Devonian paleogeography of the arctic is not well known in detail, the major features have been delineated (e.g. Trettin in Trettin and Balkwill, 1979). A northerly located land mass (orogenic welt),

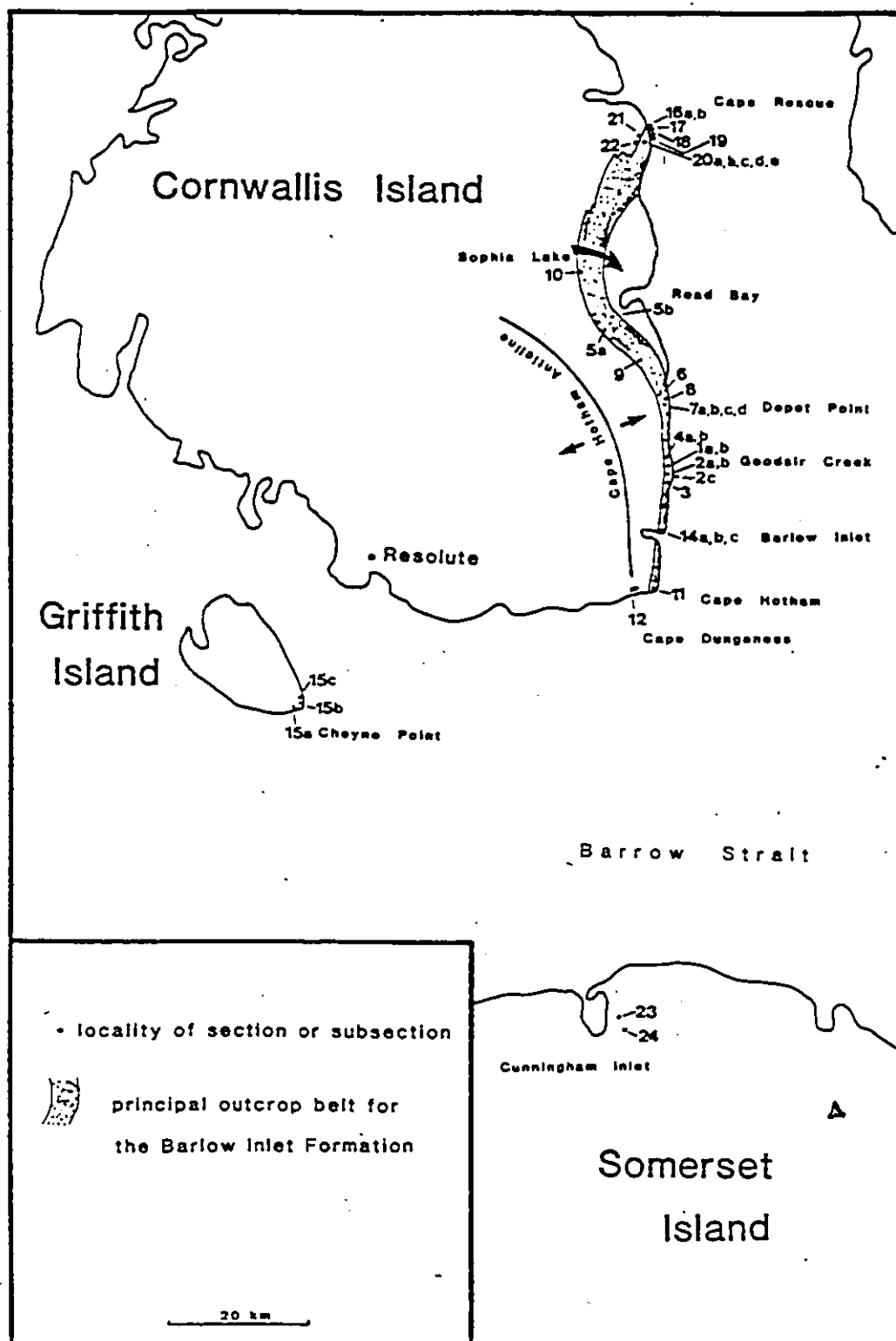


FIG. 1-2 Location of measured sections.

Locality Data for Measured Sections

Cornw.= Cornwallis; Grif.= Griffiths; Som.= Somerset;
 G.C.= Goodsir Creek; R.B.= Read Bay; D.P.= Depot Point;
 C.H.= Cape Hotham; C.D.= Cape Dungeness; B.I.=Barlow Inlet;
 C.P.= Cheyne Point; C.R.= Cape Rescue; A.B.= Advance Bluff;
 C.I.= Cunningham Inlet; S.L.= Sophia Lake
 Units: 1=Barlow Inlet Fm., upper member; 2=Barlow Inlet Fm, lower member; 3,?3=Somerset Island Fm., lower member or equivalent facies; 4=Douro Fm.; 5=Cape Storm Fm.; 6,?6=Cape Phillips Fm., member C or equivalent facies.

Sect.	Sub.	Island	NTS Sheet	Base of Section UTM Coordinates Easting/Northing		Nearest Major Geographic Point of Reference	Units	Photo Reference (NAPL) Vertical/Other
1	a	Cornw.	58F/15	487270	8304950	G.C.	1	Al6175-29
1	b	Cornw.	58F/15	487270	8304950	G.C.	1	Al6175-29
2	a	Cornw.	58F/15	487280	8303550	G.C.	1	Al6175-29
2	b	Cornw.	58F/15	487480	8303430	G.C.	1	Al6175-29
2	c	Cornw.	58F/15	487780	8303350	G.C.	1	Al6175-29
3		Cornw.	58F/15	487350	8302770	G.C.	1	Al6175-29
4	a	Cornw.	58F/15	487300	8305750	G.C.	1	Al6175-29
4	b	Cornw.	58F/15	487300	8305750	G.C.	1	Al6175-29
5	a	Cornw.	58F/15, G/2	479650	8322850	R.B. (type)	1	Al6175-56/ T401c-176
5	b	Cornw.	58G/2	481600	8326650	R.B. (type)	1	Al6175-33/ T401c-176
6		Cornw.	58F/15	486850	8315200	D.P.	1	Al6175-31
7	a	Cornw.	58F/15	487330	8313900	D.P.	1	Al6175-31
7	b	Cornw.	58F/15	487340	8313900	D.P.	1	Al6175-31
7	c	Cornw.	58F/15	487600	8314020	D.P.	1	Al6175-31
7	d	Cornw.	58F/15	487700	8314020	D.P.	1	Al6175-31
8		Cornw.	58F/15	488150	8314950	D.P.	1	Al6175-31
9		Cornw.	58F/15	485800	8317200	D.P.	1	Al6175-31
10		Cornw.	58G/2	476000	8333800	S.L.	1	Al6175-54/ T4011-178

(continued)

Table 1-1(concluded)

10

11		Cornw.	58F/10	476000	8333800	C.H.	4,?3,1	A16175-26
12		Cornw.	58F/10	481200	8284820	C.D.	4,?3,1	A16175-63
14	a	Cornw.	58F/10	486150	8295000	B.I.	4,2,1,?3	A16175-27
14	b	Cornw.	58F/10	486700	8295010	B.I.	1,?3	A16175-27
14	c	Cornw.	58F/10	486700	8295020	B.I.	1,?3	A16175-27
15	a	Grif.	58F/12	432650	8269470	C.P.	5,4	A16194-117
15	b	Grif.	58F/12	433300	8270050	C.P.	4,2,?1	A16194-117
15	c	Grif.	58F/12	433200	8271170	C.P.	?1	A16194-117
16	a	Cornw.	58G/7	484400	835580	C.R.	1	A16175-38
16	b	Cornw.	58G/7	484400	835580	C.R.	1	A16175-38
17		Cornw.	58G/7	484900	835490	C.R.	1	A16175-38
18		Cornw.	58G/7	485600	835370	C.R.	1,?6	A16175-38
19		Cornw.	58G/7	486200	835160	A.B.	1	A16175-38
20	a	Cornw.	58G/2	483800	835130	A.B.	1	A16175-38
20	b	Cornw.	58G/2	483900	835060	A.B.	1	A16175-38
20	c	Cornw.	58G/2	484420	835025	A.B.	1	A16175-38
20	d	Cornw.	58G/2	484300	835030	A.B.	1	A16175-38
20	e	Cornw.	58G/2	484300	835035	A.B.	1	A16175-38
21		Cornw.	58G/7	482500	835200	A.B.	6,?1	A16175-38
22		Cornw.	58G/2	481350	8351100	A.B.	6,?1	A16175-38
23		Som.	58F/2	481000	8220100	C.I.	3,?4	A16837-35
24		Som.	58F/2	479600	8218800	C.I.	3,?4	A16837-35

Table 1-2

11

Summary of Characteristics of Measured Sections.

Sect.	Sub.	Total Thickness	Units	Height of Base of Section from datum (SCOR variable)	Basis	No. of Units	Per- cent Exposure	Average Unit Thick- ness ¹
1	a	218.3	1	0.0	A	44	78	4.5
1	b	58.9	1	+26.8	C	17	88	3.6
2	a	23.4	1	0.0	A	11	75	1.9
2	b	194.8	1	+32.0	A	58	99	3.4
2	c	121.3	1	+240.0	C	34	91	3.6
3		133.3	1	0.0	A	38	83	3.2
4	a	138.3	1	0.0	C	16	70	6.9
4	b	304.5	1	0.0	A	60	82	4.9
5	a	1201.8	1	0.0	A	120	33	4.2
5	b	85.3	1	+1186.5	A	27	76	2.7
6		154.2	1	0.0	B	16	62	8.3
7	a	65.8	1	+30.0	E	23	99	3.0
7	b	140.0	1	+47.8	C	47	98	3.0
7	c	40.5	1	+50.0	E,C	15	88	2.8
7	d	152.7	1	+90.0	C	24	93	6.4
8		103.0	1	+260.0	D	34	72	2.6
9		91.9	1	0.0	E	15	76	6.2
10		779.9	1	0.0	D	71	24	3.4
11		203.7	4, ?3, 1	-126.2, -44.2	B	90	74	1.8
12		150.1	4, ?3, 1	-137.2, -39.8	B	50	58	2.1
14	a	325.8	4, 2, 1, ?3	-280.4	A	70	52	3.6
14	b	29.6	1, ?3	0.0	C	12	64	1.7
14	c	177.6	?3, 1	+29.6	C	100	96	1.8
15	a	456.2	5, 4	-36.9	A	158	79	2.6
15	b	45.5	4, 2, 1	-23.1	A	39	70	1.0
15	c	19.4	?3, 1	+90.0	D	14	100	1.5
16	a	64.4	1	+915.0	E	20	50	2.5
16	b	39.8	1	+942.0	C,E	10	48	3.5
17		23.3	1	+900.0	E	6	94	5.5
18		92.0	1-?6	+925.0	E	26	48	2.6
19		44.3	1	+1250.0	E	14	79	2.7

(continued)

Table 1-2(concluded)

12

20	a	74.0	1	+540.0	E	15	29	3.6
20	b	23.8	1	+628.8	C,E	3	80	10.4
20	c	54.3	1	+730.0	E	8	55	5.4
20	d	8.8	1	+775.0	C,E	5	100	1.8
20	e	4.5	1	+800.0	C,E	1	100	4.5
21		46.6	?6	+270.0	E	10	91	4.7
22		190.3	?6	+125.0	E	20	47	7.5
23		57.0	?3	?0.0		31	82	1.8
24		53.4	?3	?0.0		30	96	1.8
TOTALS 6244.1m (of which 3778.9m is exposed) 1402 units								
MEANS (156.1m) (35) (70.9) (3.7)								
-arithmetic mean of weighted percentage exposure is 61%								

1. Average unit thickness is calculated for exposed units with greater than 50% outcrop only.

Datum is the top of the lower member of the Barlow Inlet Fm.

Units as in Table 1-1.

Basis for the SCOR variable (representing metres between section base and the datum) as follows:

- A scor determined by direct measurement.
- B direct measurement, however contact used as datum is either covered or subject to interpretation.
- C physical correlation with adjacent, superjacent or subjacent sections.
- D estimate based on graphic measurement with good structural and topographic control.
- E "ball-park" estimate (graphic method).

Pearya, bordered a NE.-SW. oriented deep linear basin, known as the Hazen Trough (FIG. 1-3a), and in Silurian time contributed substantial volumes of sediment to it (Trettin, 1973). The southern margin of the Hazen Trough on Ellesmere Island is reefoid, as is a similar trend to the west on Melville and neighbouring islands. The continuity of the reef rim has not been adequately ascertained. Separating this outer carbonate barrier from a widespread inner shelf is a deep back-barrier basin, herein referred to as the Cape Phillips (back-barrier) Basin. The sediments of this basin were predominantly fine-grained and rich in siliciclastic material. To the south, the inner carbonate shelf is divided into separate depo-centers by several intermittently positive basement arches, the principal of which is the Boothia Uplift. In latest Silurian time the Boothia Uplift was emergent (creating the Boothia Axial Landmass). The Barlow Inlet Formation was laid down on an inner carbonate shelf which is transitional northward into the Cape Phillips Basin and southward into the newly emergent continental terrain. The depositional history of the Barlow Inlet Formation records intimate interplay between continued subsidence of the shelf and basinal areas to the north, and episodic uplift of shelf areas to the south.

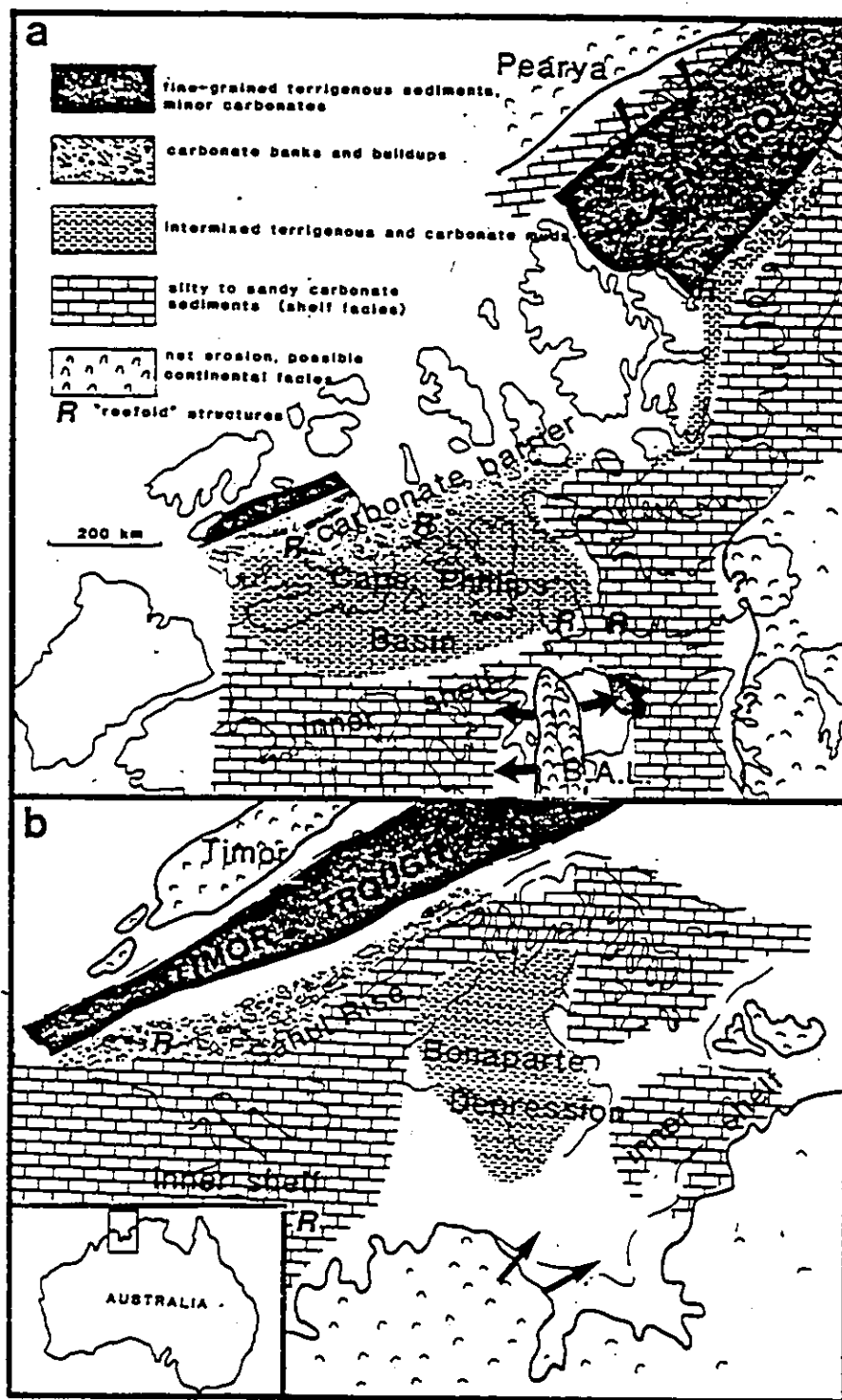


FIG. 1-3a Late Silurian Paleogeography of Arctic Islands Region. Arrows indicate sediment dispersal (modified after Trettin in Trettin and Balkwill, 1979)

FIG. 1-3b Holocene Geography of Sahul Shelf Region, NW Australia (modified from James and Ginsburg, 1978).

To appreciate the interpreted paleogeography described above, it is useful to refer to a Holocene analog present in the area of the Timor Sea off NW. Australia (FIG. 1-3b), whose physiographic elements are developed on a similar scale to those present in the Siluro-Devonian. There is a northerly-located orogene (Timor Island), a deep NE.-SW. trough (the Timor Trough), an external, in part reefoid carbonate shelf (Sahul Rise), a moderately deep (up to 145 m) back-barrier basin filled with siliciclastic-rich muddy sediments (Bonaparte Depression), and an inner carbonate shelf that borders a coastal plain and adjacent uplifted Precambrian fault-blocks.

1.6 PRESENTATION OF FINDINGS

The Barlow Inlet Formation can be considered to be composed of four magnafacies. The lower member is regarded as a TERRIGENOUS CLINOFORM magnafacies, whereas the rocks of the upper member are assigned to either PLATFORM MARGIN (including shelf edge, proximal back-barrier, and proximal foreslope), or PLATFORM INTERIOR (including lagoon and tidal flat), or BASIN SLOPE magnafacies. Born out of an interpretation of these principal divisions of the Barlow Inlet Formation are four independent papers that comprise the bulk of this dissertation. Each paper has been written as a self-contained entity and as a consequence some information is repeated in the separate papers. The four

papers are listed below along with the magnafacies with which they are most concerned.

Title of Paper	Magnafacies	Page No. of Thesis
1. Late Silurian Syntectonic Deltaic Sedimentation, Cornwallis Island Arctic Canada	(TERRIGENOUS CLINOFORM)	p. 25
2. Shallowing-upward Carbonate Hemicycles in an Upper Silurian Miogeoclinal Shelf Succession, Arctic Canada	(PLATFORM INTERIOR)	p. 196
3. Preliminary Survey of Biocumule (Buildup and Biostromal) Facies in an Upper Silurian succession, Arctic Canada	(PLATFORM MARGIN)	p. 345
4. Stratigraphy and Facies Disposition of the Upper Silurian Carbonate Member of the Barlow Inlet Formation, Arctic Canada		p. 513

The last paper presents a facies overview of the upper member, but also pays critical attention to the BASIN SLOPE magnafacies.

1.7. ACKNOWLEDGEMENTS

I am indebted to numerous individuals for their support and assistance rendered throughout the lengthy duration of this project. Foremost I would like to express my gratitude to Dr. O.A. Dixon, my thesis supervisor, who, with the patience of Job, allowed this thesis to 'ferment' at its own (unhurried!) pace. Dr. Dixon's always available and considered insight during the research stages of this project, and his editorial skills during write-up, are most appreciatively acknowledged. Drs. Thorsteinsson, McGregor, Copeland and Powell of the Geological Survey of Canada are thanked for contributing valuable information on fish and graptolite fossils, palynomorphs, ostracodes, and organic geochemistry, respectively. Dr. B. Jones of the University of Alberta identified the brachiopod faunas. My appreciation is extended to Dr. D.P. Eyre for his instructive counsel, and Dr. U. Mayr of the G.S.C. for providing an additional incentive for completing the project. Canadian Superior Exploration Limited kindly exchanged geological data relevant to the study area.

During my tenure as a student at the University of Ottawa, it was both a privilege and a pleasure to share the graduate experience with other similarly afflicted souls. In particular I should like to mention Guy Narbonne and Bill Parkins, whose friendship and advice were and remain greatly appreciated. M. Jackson and the entire

technical staff of the department made the journey that much easier, and were a treasured source of humanity in an inanimate world of rocks and academics. E. Hearn kindly did the calligraphy for the Frontispiece. Much of the drafting was ably and conscientiously executed by K. Venance. H. Hoesny patiently processed corrections on earlier manuscript drafts, and created a bibliography format program. The final completion of this thesis was a long distance affair from several points of view. I am exceptionally indebted (both figuratively and financially) to Geof Burbidge, who very generously offered to act as my surrogate in Ottawa, and oversaw the final assembly of the dissertation.

A field project, such as this, often fails or succeeds on the backs of those unsung heroes, geological assistants. I have been exceptionally fortunate to have had four outstanding individuals work with me, and share their companionship over three field seasons. To Rick Zuran, Jean Dugal, Gary Graf, and Randy Hrynuik thanks!

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Rufus and Fred gave portions of this manuscript their purr-ticular if unsolicited attention. Christopher J., by his much welcomed if somewhat early arrival, has put this entire endeavour into its proper perspective. To my wife Buff, thanks for keeping the home fires burning and believing in me, even when I did not.

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II

PAPER II: LATE SILURIAN SYNTECTONIC DELTAIC
SEDIMENTATION, CORNWALLIS ISLAND, ARCTIC CANADA

Abstract

Outcropping on Cornwallis and Griffith Islands of the central Canadian Arctic, the lower member of the Upper Silurian Barlow Inlet Formation is a thin (generally less than 50m), siliciclastic sequence, stratigraphically enclosed in a 2700 m thick succession of relatively pure carbonates. The lower member overlies the Douro Formation conformably, but very abruptly; this contact represents an event horizon signalling a major late Ludlovian pulse of the Cornwallis Disturbance. The lower member is predominantly composed of dark mudrocks of 'black shale' aspect, and these mudrocks are interpreted as prodelta sediments deposited in an intermittently anoxic basin. A sandy facies consists of fine-grained, in part bimodally-sorted subarkose, characterized by low-inclined parallel lamination, runzel marks, desiccation polygons, asymmetric ripples, primary current lineation, and a Cruziana trace fossil assemblage. The subarkose facies is interpreted as a beach foreshore.

The lower member represents the result of nearby continental mass-wasting of epeirogenic uplands. The detrital source cannot be determined with certainty, but a source area to the south and east of Cornwallis Island is

suggested by the thickness distribution of the lower member. The deltaic sediments that constitute the lower member represent a part of the marine record of Caledonian diastrophism affecting the Boothia Uplift region as well as other parts of the Arctic Archipelago. The tectono-sedimentary record of Cornwallis Island during this time can be interpreted in terms of pre-epeirogenic, initial epeirogenic, epeirogenic, and immediate post-epeirogenic phases.

2.1 INTRODUCTION

The lower member of the Barlow Inlet Formation² on eastern Cornwallis Island (FIG. 2-1), is a thin but significant lithostratigraphic anomaly within the Upper Silurian succession of the central Arctic islands. Approximately 50 m thick at its type section, the siliciclastic lower member is composed principally of unfossiliferous dark mudrocks (of 'black shale' aspect) and texturally mature sandstones. It is distinct in composition and depositional setting from the 2700 m of shelf carbonates that enclose it. The lower member contrasts markedly with the underlying Douro Formation (350 m) and overlying upper member of the Barlow Inlet Formation (1300 m), both of which consist largely of fossiliferous limestones and dolostones relatively free of either arenite-sized siliciclastic detritus, or thick intervals of non-calcareous mudrock (FIG. 2-1).

The purpose of this paper is twofold. Firstly, it provides a detailed description of the lithology and faunal composition of the lower member, reports on its regional variability, and assesses the appropriateness

² The lower and upper members of the Barlow Inlet Formation (Thorsteinsson, 1980) were previously referred to as the B and C members of the Read Bay Formation (Thorsteinsson, 1958). The A member of the Read Bay Formation similarly has been redefined (Thorsteinsson, op.cit.) and is represented largely by the Douro Formation, and in part by the underlying Cape Storm Formation. For brevity, the lower member of the Barlow Inlet Formation shall be referred to hereafter simply as the lower member.

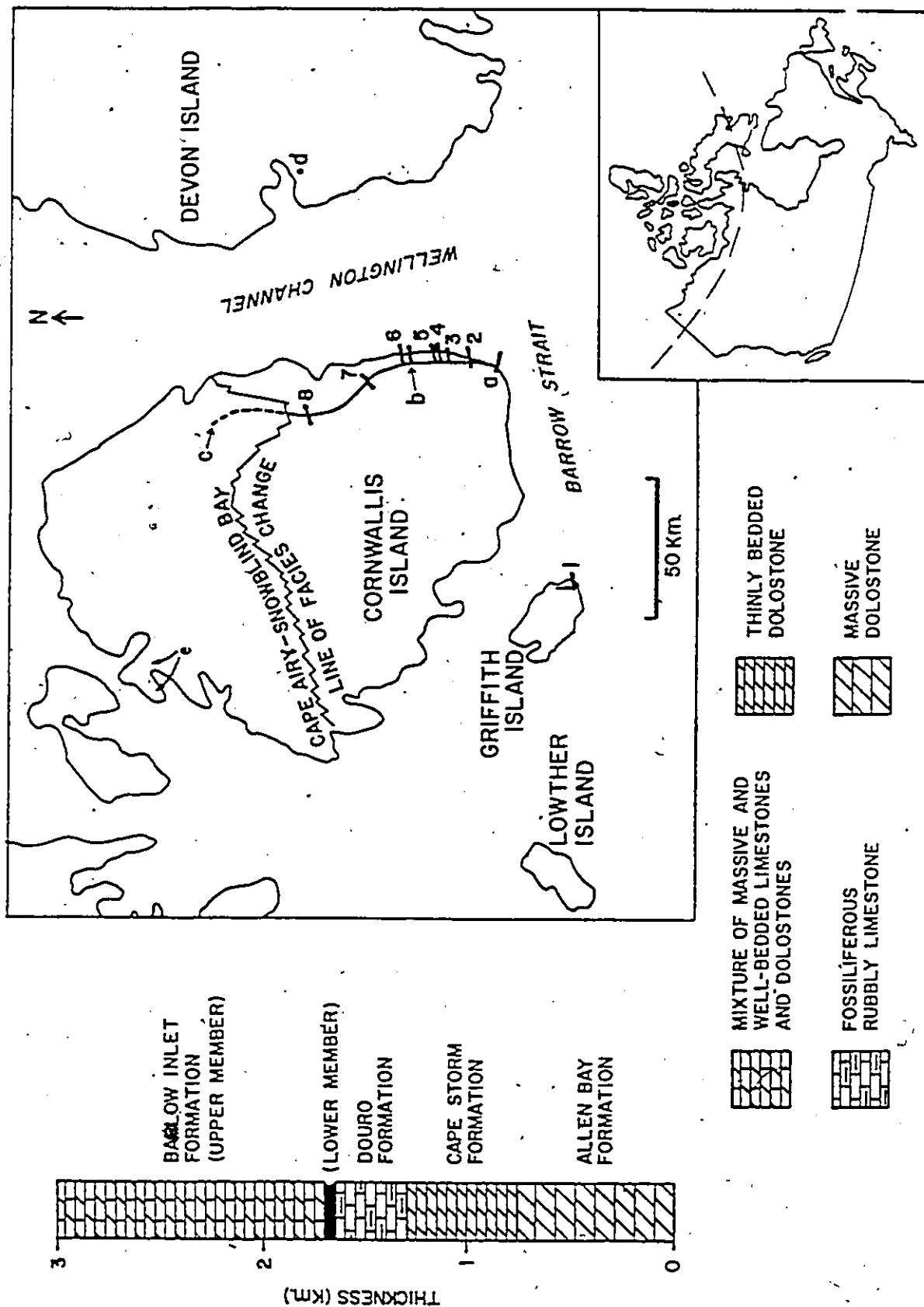


FIG. 2-1 Location map of study area. Numbers represent principal sections. Letters represent localities mentioned in the text. Generalized stratigraphic column depicted at left.

of various depositional models to this particular 'black shale' deposit. Secondly, it considers the tectonic significance of this abrupt influx of concentrated siliciclastic detritus. The latter has previously been only cursorily considered (Thorsteinsson, 1980; Packard and Dixon, 1981), and is herein treated more comprehensively.

The lower member (as member B of the Read Bay Formation) incorporated 60 to 100 feet (18.3-30.5 m) of "greenish black to black, friable shale", and locally, in its upper 20 feet "alternating shale and light grey, very thin-bedded, flaggy, in part cross-bedded, hard, fine-grained sandstone" (Thorsteinsson, 1958, p.64). It lay conformably between the A and C members of the Read Bay Formation. Thorsteinsson chose a type section on the south side of Goodsir Creek on the east coast of Cornwallis Island (FIG. 2-1, loc. 5). In the newly-defined Barlow Inlet Formation (Thorsteinsson, 1980, p.5), the lower member at its type locality was said to be 65 m thick, to consist "mainly of soft-weathering shale that is calcareous and quartzose silty", and, in places, to include "minor amounts of fine-grained quartz sandstone and thin beds of nodular limestone".

Minor references to the lower member (as member B) occur in reports of more regional scope (e.g. Thorsteinsson and Kerr, 1968; Thorsteinsson in Douglas, 1970; and Mayr, 1978) Microfaunal and/or macrofaunal investigations (Uyeno, 1977, 1980; McGregor and Narbonne, 1978) have yielded new insight

into the age and depositional character of the lower member. Papers by Kerr (1977), Gibling and Narbonne (1978), and Thorsteinsson (1980) are important to an understanding of the stratigraphic and tectonic setting of the lower member.

2.2 AGE

The age of the lower member at its type section, and on Griffith Island (FIG. 2-1, locs. 5,1), can be established on the basis of its contained conodonts and position within the stratigraphic succession (Uyeno, 1977; Thorsteinsson, 1980). At the type section, the uppermost beds of the Douro Formation contain the last record of conodonts assigned to the siluricus Zone, and conodonts in the conformably overlying upper member of the Barlow Inlet Formation (approx. 30 m above the top of the lower member), indicate the presence of the latialata Zone (Thorsteinsson, 1980, p.24). On Griffith Island Ozarkodina n. sp. A and Pedavis aff. P. Thorsteinssoni occur 6 m above the base of the lower member; this fauna is also suggestive of the latialata Zone. Conodonts, therefore, suggest that the lower member is of latest Ludlovian age.

The contained macrofauna does not provide as precise an age; at best, it does not negate the age suggested by conodonts (see Table 2-1).

Given the apparent synchronicity and abrupt nature of the upper contact of the Douro Formation (Thorsteinsson, 1980, p.24 and 32), and taking into account the conformable nature of the Upper Silurian succession on Cornwallis Island, it is reasonable to suggest that deposition of the lower member began, for practical purposes, coetaneously throughout the region. No similar evidence is presently available

concerning the onset of deposition of the upper member of the Barlow Inlet Formation.

Table 2-1

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Age Ranges of Selected Fossils in, or just above or below,
the lower member.

MACROFAUNA	STRATIGRAPHIC POSITION	AGE RANGE

ARTHROPOD		
<u>Paraaulacopleura</u> sp. (trilobite)	(10m)	Mid.Ord.-Mid.Dev.
<u>Pterygotus</u> cf. <u>P. pterygotus</u> (eurypterid)	+20m	?Ord., Dev.
<u>Entomozoe</u> sp. (ostracode)	(10m)	Sil.-Perm.
OSTRACODERM		
<u>Ariaspis</u> sp. (cyathaspidid)	(10m), +15m	Early Dev.
GRAPTOLITE		
<u>Monograptus</u> n.sp. (aff. <u>M. ultimus</u>)	(10m)	late Ludlow
BRACHIPOD		
<u>Atrypoides</u> <u>erebus</u>	-1m, +5m	late Ludlow
<u>A. foxi</u>	+12m	late Ludlow
<u>Stegerhynchus</u> <u>angaciensis</u>	+12m	mid.Ludlow-early Pridoli

Note: + above upper contact of lower member.
 - below lower contact of lower member.
 () in lower member, height from base.

2.3 GEOGRAPHIC DISTRIBUTION AND THICKNESS

2.3.1 Surface Exposure

To date, exposures of the lower member have been identified only on Griffith Island and the east coast of Cornwallis Island.

The southern- and westernmost exposure of the lower member examined occurs at Cheyne Point, Griffith Island (FIG. 2-1, loc. 1; PL. 2-1h). The lower member is represented by ~14 m of calcareous mudrock and argillaceous limestone, conformably overlying the Douro Formation and conformably overlain by the Somerset Island Formation. The member is only 9 m thick in the southernmost exposure at Barlow Inlet on eastern Cornwallis Island (FIG. 2-1, loc. 2; PL. 2-1g). A 20 m covered interval above the Douro Formation at Cape Hotham, some 10 km south of Barlow Inlet, conceivably represents the surface expression of the lower member. Although far from conclusive, two factors suggest otherwise:

1. thickness trends suggest rapid southward pinchout of this member, and termination north of Cape Hotham.
2. despite its recessive weathering character, traces of the 'black shale' lithofacies of this member, can generally be detected within overriding talus cover; at Cape Hotham no black shale is apparent.
3. physical correlation suggests that strata immediately below the covered interval be assigned to the upper member of the Barlow Inlet Formation.

North of Barlow Inlet, to the Snowblind Bay - Cape Airy line of facies change (FIG. 2-1), the lower member can easily be traced for some 56 km by virtue of its distinctly recessive character; the lower member is in part responsible for a prominent N-S physiographic feature on the east coast of Cornwallis Island, the Cape Hotham Escarpment (see Thorsteinsson, 1958, p.64; and PL. 2-1a, this paper).

The northernmost exposure of the lower member examined in this study, is west of Sophia Lake (FIG. 2-1, loc. 8), where 20 m of fissile calcareous siltstone and shale outcrop (PL. 2-1f). Thorsteinsson (1980, p.21) reported that it is possible to trace the lower member even north of the facies change for some 20 km into the graptolitic Cape Phillips Formation (FIG. 2-1, loc. C).

Maximum thickness occurs at loc. 7 (FIG. 2-1) midway between Read Bay and Depot Point, where 65 m is exposed. There the upper contact of the lower member is poorly defined, and the thickness must be considered approximate. The lower member at its type section (PL. 2-1b) and newly defined reference section (PL. 2-1e) (FIG. 2-1, locs. 5,4), is 54 and 52 m thick respectively. Thickness variation of surface exposure within the study area is summarized in FIG. 2-2.

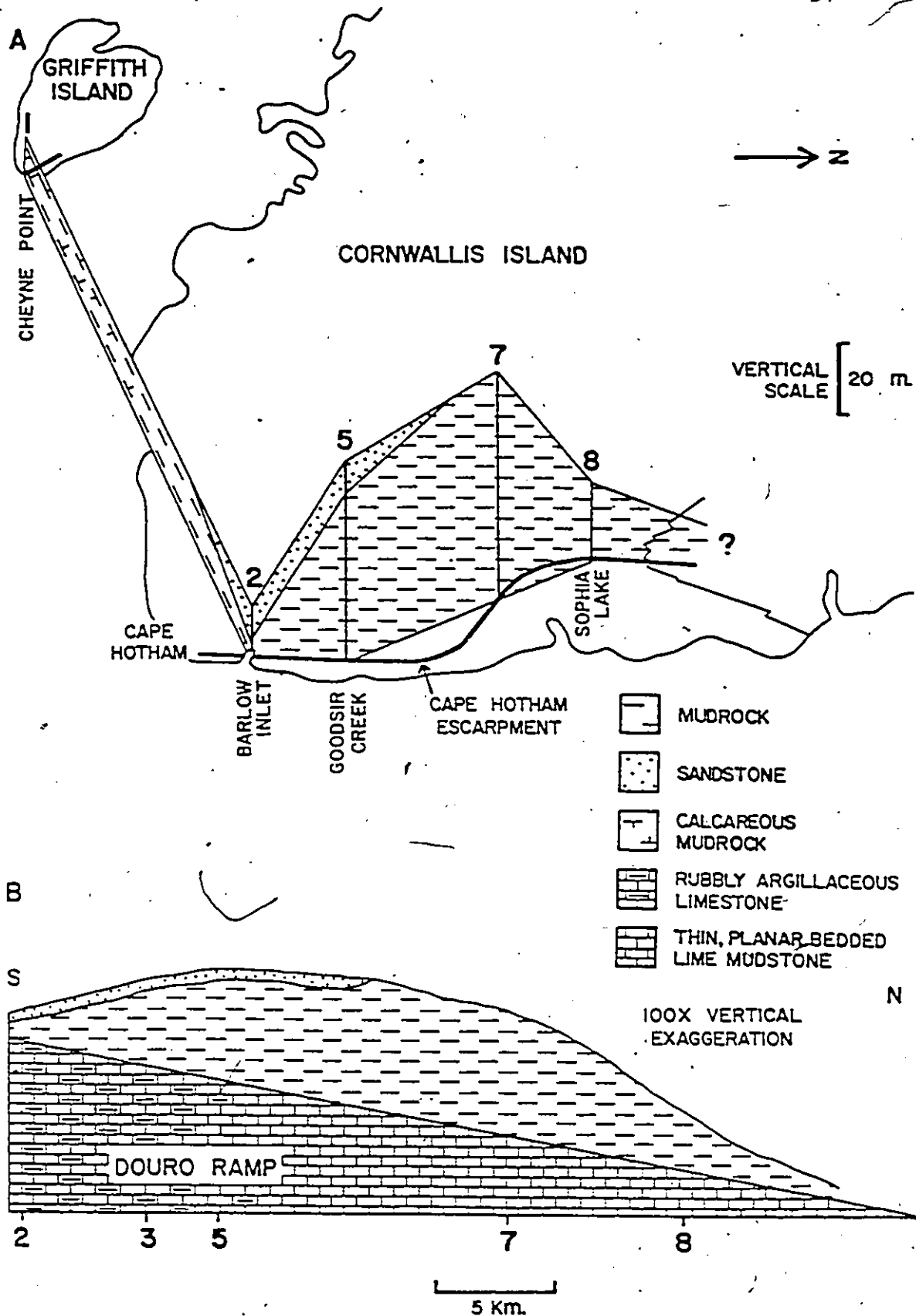


FIG. 2-2 Thickness variation of lower member.

2.3.2 Subsurface Occurrence

Mayr et al. (1980) assigned stratigraphic intervals in two wildcat wells south and west of this study area, to the lower member of the Barlow Inlet Formation. In the Murphy Alminex et al. Victoria Island F-36 well on western Victoria Island; a 344.4 m interval is correlated by Mayr et al. (1980, Fig. 32.1) with the lower member. The Sun Panarctic Russell E-82 well on the south coast of Russell Island, contains a 95.7 m unit of very fossiliferous interbedded limestone and shale, which is also assigned to the lower member.

The present writer feels that regardless of approximate equivalency in age, the lithology of the well log intervals is not sufficiently known to warrant including them in the lower member. More specifically the lower member at its type locality is very sparsely fossiliferous, is overwhelmingly non-carbonate (80-85 % insoluble residue), and is characterized by very dark mudrocks ('black shales'). Also as indicated in FIG. 2-2 the lower member appears to pinch out to the south and the west.

Mayr (1978) assigned to the lower member a 39.4 m interval of equal proportions of, calcareous silty light grey shale, limestone, and dolostone, in the Imperial Panarctic et al. Devon E-45 well on the west coast of Devon Island (FIG. 2-1, loc. d). Conversely, Thorsteinsson (1980, p.6) stated that, "it is of some interest to note here that

the lower shale member is absent in that [Devon] island". This apparent contradiction may be resolved in a forthcoming publication (Thorsteinsson and Mayr, in press), but at present the status of the lower member east of Cornwallis Island is uncertain.

2.3.3 Type and Reference Section

Complementary to the existing type section of the lower member at Goodsir Creek (PL. 2-1b), a reference section is herein designated on the south side of an unnamed rectangular lake, 3 km due south of the type section on the Cape Hotham Escarpment (FIG. 2-1, loc. 4; PL. 2-1e). FIG. 2-3 is a composite section: the uppermost part of the Douro Formation is taken from the reference section, and the lower member is from the type section. The stratigraphic sequence is essentially identical at both localities, and the composite section serves only to maximize the net continuity of exposure.

The lower member is 54 m thick in the type section and 52 m in the reference section. Previous thickness estimates of approximately 65 m (Thorsteinsson and Uyeno, 1978; McGregor and Narbonne, 1978, fig. 3), were based on the inclusion of a covered interval situated above the last occurrence of the lower member's subarkose facies. Excavation of this covered interval at both sections, exposed strata lithologically and faunally related to overlying carbonates

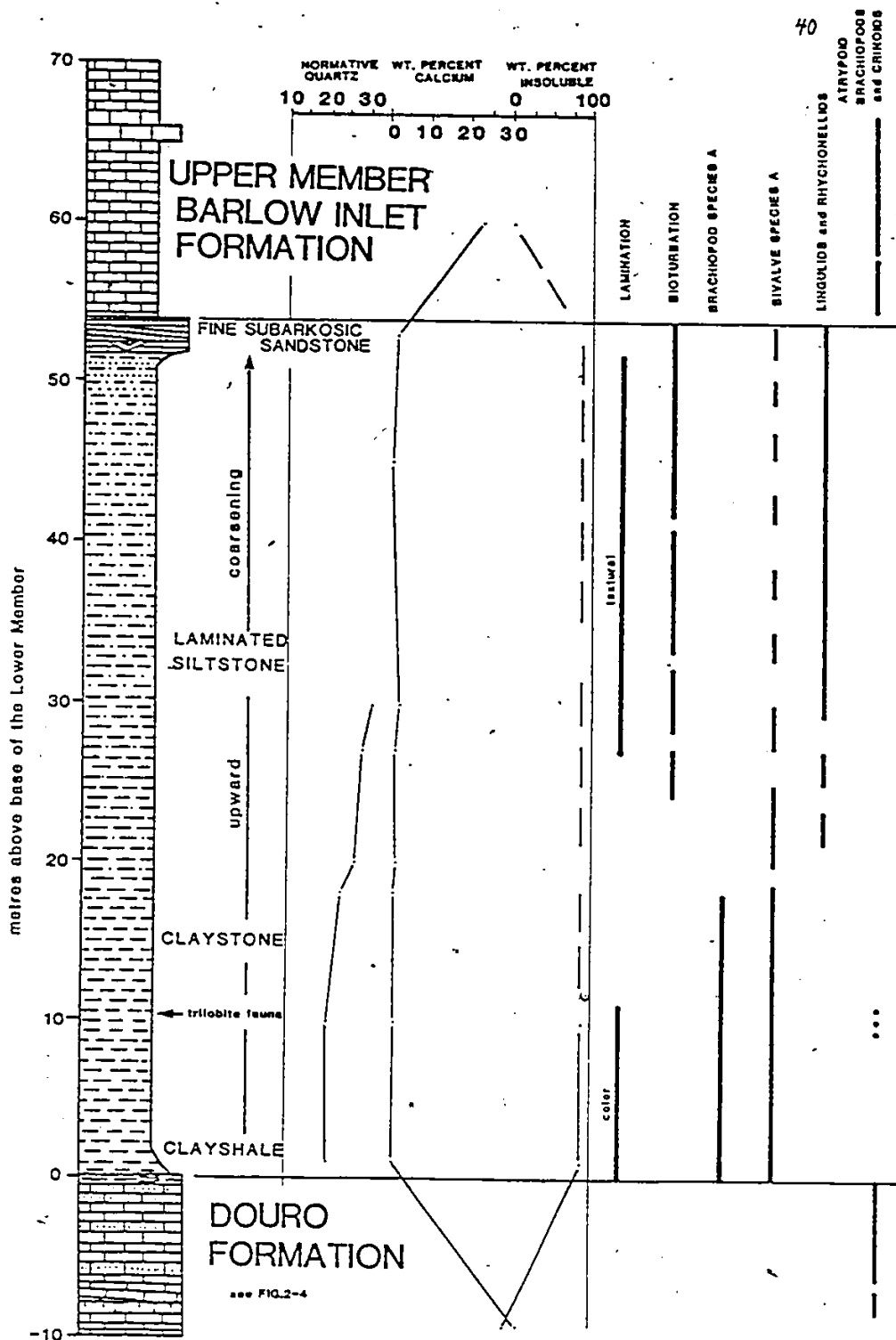


FIG. 2-3 Composite section of lower member from type and reference localities.

of the upper member. These strata are herein assigned to the upper member with the following consequences:

1. the lower member at its type section is almost wholly siliciclastic, and contains no discrete beds of carbonate, and
2. the upper contact for the lower member is drawn at the top of the last occurrence of the subarkose facies.

The basal 52 m of the lower member at its type section consist of dark mudrock (black shale) lithofacies. The uppermost 2 m of both type and reference sections, are a fine-grained subarkose lithofacies. These two lithofacies will be discussed separately.

2.4 METHODS OF STUDY

The extremely friable nature of the principal mudrock facies of the lower member, and hence its very recessive weathering profile, precludes normal bed by bed stratigraphic analysis. Everywhere, the in situ and fresh (albeit frozen) mudrock lies several decimeters beneath scree and/or highly weathered shale. It was found necessary at all study localities, to attempt to excavate at least the upper and lower contacts of the member, and at several localities (e.g. PL. Z-1b) the entire lower member was trenched to obtain representative and sequential lithological samples.

Stratigraphic thicknesses in excess of 20 m were measured using a 'fixed-length chain' method; otherwise a standard Jacob's staff technique was employed. All 'essentially insitu' outcrop of the upper subarkose facies is in fact, at least partially displaced by frost heaving. Nowhere is this facies more than 1.5 m thick, and its most common occurrence is as loose talus blocks (up to 2 X 2 X .5 m). Where the upper contact could not be located through excavation, it was arbitrarily placed at the highest level of blocks of subarkose on the talus slope (given the presence of such blocks, and the absence of black shale residue further upslope).

Apart from standard petrologic techniques (stained thin sections, x-ray diffraction), representative samples of the

mudrock facies were: analyzed for a broad elemental spectrum by X-ray fluorescence, analyzed specifically for Sr, K and Na by means of atomic absorption spectra, submitted for palynomorph analysis (Dr. D.C. McGregor -- Geological Survey of Canada, Ottawa) and organic carbon analysis by means of flame photometry (Dr. T.G. Powell -- Geological Survey of Canada, Calgary). Additionally, polished thin sections of samples of the subarkose facies were subjected to cathodoluminescent photomicrography to determine original textural characteristics.

2.5 UNDERLYING STRATA OF THE DOURO FORMATION

2.5.1 General Sequence

The stratigraphic sequence immediately underlying the lower member has considerable bearing on an understanding of the tectono-sedimentary environment represented by the 'black shales', and is therefore treated here in some detail.

The upper 160 m of the Douro Formation at Goodsir Creek are atypical of the formation as recognized on Somerset and Devon Islands (Narbonne, 1981, p.47). Of this 160 m, the basal 40 m consist of medium grey weathering, fissile, calcareous, silty shale interbedded with unfossiliferous lime mudstone (FIG. 2-4). This particular lithofacies is widely regarded as a southward directed tongue of graptolitic, 'deeper-water basinal' deposits representing either the Devon Island Formation (Mayr, 1978) or the Cape Phillips Formation (Thomas and Narbonne, 1979; Thorsteinsson, 1980). This graptolitic shale can be traced as far south as Cape Hotham (FIG. 2-1, loc. a) (Thorsteinsson, 1958, p.102).

The succeeding 110 m is characterized by thin planar-bedded, virtually unfossiliferous, in part laminated, argillaceous lime mudstone (Thorsteinsson, 1958, p.49; Thomas and Narbonne, 1979, p.3). These sediments are rarely bioturbated, although a few beds contain abundant trace fossils including (commonly)

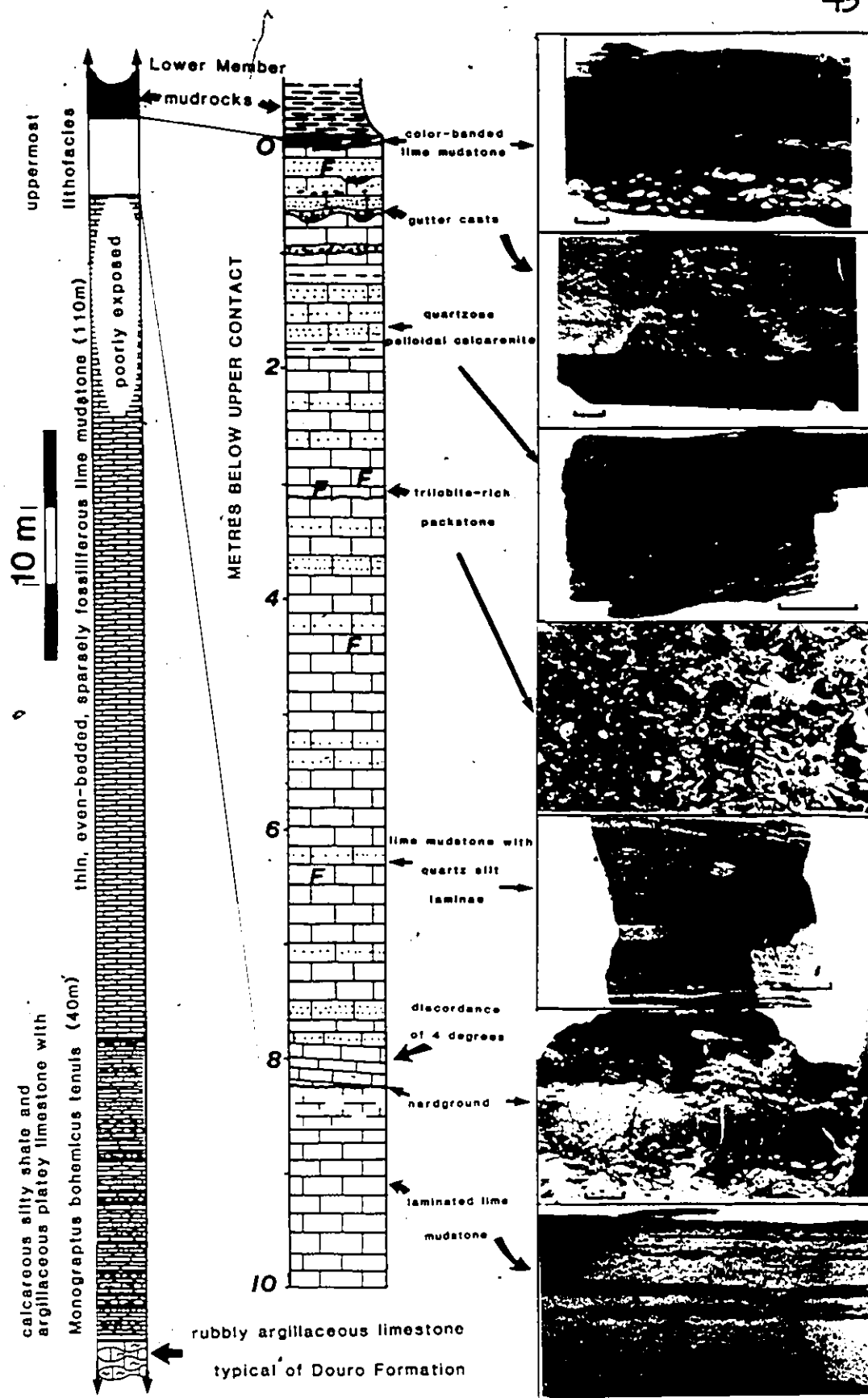


Fig. 2.4 Upper Douro Formation on Eastern Cornwallis Island.

Skólithos, Chondrites, Palaeophycus, Sclarituba, and (rarely) Phycodes, Margaritichnus? and Teichichnus. Narbonne (1981, p.146) termed this ichnocoenose the "Sclarituba association" and considered it to be equivalent to the Zoophycus ichnofacies, which in siliciclastic sequences is interpreted to indicate bathyal depths (see Crimes in Frey, 1975).

Narbonne (1981, p.142) interpreted this 110 m interval as having been "...deposited below storm wave-base...", and further observed that these rocks "...closely resemble peri-platformal slope deposits...". He assigned this interval, as well as the underlying ~40 m of graptolitic shales, to the ?Cape Phillips Formation, but the nomenclature used in this paper follows Thorsteinsson (1980) and refers the sequence in question to the Douro Formation.

2.5.2 Rates of Sedimentation

It is apparent that the 160 m succession beneath the lower member, represents mainly continuous and rapid hemi-pelagic sedimentation although the sedimentation rate is difficult to establish precisely. Only a fraction of the siluricus conodont Zone is represented by this interval. According to Thorsteinsson (1980, p.24), approximately 1000 m of sediment can be assigned to the siluricus Zone in the Cornwallis Island area; this is in contrast to the few metres of beds assigned to this zone in Nevada and Europe.

As stated by Spjeldnaes (1978, p.344), "There is very little reliable isotope dating relevant to the subdivisions of the Silurian". However, a recent compilation by McKerrow et al.(1980) provides estimates of 6.5 to 8Ma for the duration of the Ludlow period. On Cornwallis Island, the Ludlovian shelf sequence is approximately 1300 m thick, which represents an uncorrected sedimentation rate of 16 cm/1000 yrs. (based on 8Ma duration for Ludlow time). This is an extraordinarily high preserved accumulation rate in comparison to estimated rates for other ancient sequences (see Schwab, 1976). If the four European conodont Zones recognized in the Ludlovian succession are considered to be the product of a constant evolutionary rate, then the siluricus Zone could conceivably represent 2Ma. Given the 1000 m represented by this conodont Zone in the vicinity of Cornwallis Island, sedimentation rates as high as 50 cm/1000 yrs. can be entertained. This figure is almost an order of magnitude greater than rates listed for ancient sequences by Wilson (1975, p.15), and is comparable to Holocene carbonate depositional rates. Regardless of the absolute rates involved, it is clear that the upper Douro and lower member constitute a tectono-sedimentary record of high resolution and high sensitivity -- the antithesis of a 'condensed sequence'.

2.5.3 Topmost Strata

Following the lithological monotony and faunal paucity of the underlying 110 m, the uppermost 10 m of the Douro changes abruptly in sediment character (FIG. 2-4). A 10 m (minimum) thick semi-resistant ledge of the Douro immediately underlies dark mudrocks of the lower member at both type and reference sections (PLS. 2-1b, 2-2a,b). Meterages given below, express depth beneath the base of the lower member.

The lowest exposed beds of this resistant ledge are typical of Douro lithofacies described above: thin, even, planar-bedded, laminated, calcilutite mudstone, generally unfossiliferous and with thin calcareous shale interbeds.

A very distinctive hardground 8.2 m below the lower member, exhibits significant microrelief, biotic encrustation of its upper manganiferous surface, and penecontemporaneous micro-fracturing (PL. 2-2c,e,f). Sediment immediately beneath the hardground lacks bedding, and its relatively massive character can be attributed to extensive bioturbation and synsedimentary lithification (PL. 2-2c).

Of particular importance, the beds just above and below the hardground are discordant structurally by about 4 degrees (PL. 2-2c).

Planar, thin-bedded limestone above the hardground is broadly similar to lithofacies at the base of the resistant ledge, but differs in the following critical details:

1. The presence of laminae of coarse siliciclastic silt, and
2. The occurrence of beds that contain a relatively diverse and abundant benthic fauna (Table 2-2), characterized by trilobites, pelmatozoans and brachiopods.

The latter beds occur rarely but with increasing regularity towards the lower member contact. Notably, one such bed, 3 m below the lower member contact, yielded an assemblage of trilobites similar to those described by Thomas and Narbonne (1979) from equivalent strata at Goodsir Creek (loc. 5). The fauna is mostly allochthonous, and concentrated in a packstone shell lag overlying a bored and encrusted hardground (encrusting, demonstrably autochthonous elements include discoid holdfasts, auloporids and bryozoans). No epifaunal encrustation of the loose shell material was evident, suggesting rather rapid burial.

Near 2 m below the lower member, the predominant lithology is plane to wavy inter-laminated quartzose pelloidal calcarenite and calcisiltite mudstone, occurring in planar beds (hardbands) 3 to 7 cm thick. Between beds are thin (<1-3 cm), recessive, more argillaceous and/or thinly interlaminated limestone layers. Toward the base of the lower member, the relative proportion of calcarenite laminae in the Douro increases, as well as the mean size of detrital grains and allochems within the laminae. Laminae exhibit good lateral continuity.

Fauna from uppermost Douro Formation, Locality 4 and 5.
List compiled from Thomas and Narbonne (1979) and
author's own observations.

FAUNA	RELATIVE ABUNDANCE
TRILOBITES	- fragmentary remains - A
<u>Helokybe spio</u>	
<u>Encinurus</u> aff. <u>E. hyperboreus</u>	
<u>Kailia</u> <u>capra</u>	
<u>Hemiarges echinatus</u>	
<u>Illaenimorph</u> <u>scutellid</u>	
BRACHIOPODS	
<u>Atrypa</u> <u>erebus</u>	- A
<u>A. sp.</u> aff. <u>A. phoca</u>	- C
<u>Atrypa</u>	- R
indet. Strophomenids	- C
brachiopod gen. sp. indet.	- A
COMPOUND CORALS	
<u>Aulopora</u>	- C
<u>Favosites</u>	- VR
SOLITARY CORALS	- R
<u>Cystiphyllum siluriense</u> <u>cylindricum</u>	
<u>Plasmophyllum</u>	
<u>Cyathactis</u>	
BRYOZOANS	
encrusting forms	- C
dendroid forms	- C
fenestrate forms	- R
PELMATOZOAN DEBRIS	- A
leperditid ostracodes	- C
sponges and sponge spicules	- C
orthoconic nautiloids	- R
stromatoporoids	- VR
gastropods	- VR
pterinid bivalves	- VR
cyathaspidid fragments	- VR

Note: A - abundant
C - common
R - rare
VR - very rare

At 1 m, interbedded with laminated calcarenite is the first of several brachiopod rudstone coquinas. The predominant brachiopod is Atrypa erebus, and it occurs with numerous specimens of Atrypa sp., strophomenids, and an indeterminate genus (B. Jones, pers. comm., 1982). The brachiopods are not in living position (see Jones, 1981) and less than 50 % are articulated shells, although shell abrasion appears slight suggesting limited transport. Disarticulated pelmatozoan debris, trilobite fragments, gastropods, stacked single valves of leperditid ostracodes, pterinid and other indeterminate bivalves, and sponge spicules are subordinate clast types. The coquinas are irregular tabular deposits and/or shoestring bodies. The latter form results from brachiopods being concentrated in shallow (<4 cm), narrow (1-5 cm), symmetrical, and linear to slightly sinuous channels (FIG. 2-4). Channel fills occur as discrete structures or simply as irregularities at the base of a larger rudstone coquina. In these bodies, single valves assume any orientation, although there is a slight tendency toward concave-up positions. Bioclasts are poorly sorted and ungraded. Umbrella voids are common. Geopetally filled articulated valves show marked consistency in their indication of original horizontality. Intra-allochem sediment corresponds well to the matrix enclosing the bioclasts. Several examples show two clearly defined sediment types reflecting two stages of intra-allochem void

(and channel) filling. Most articulated valves are completely filled by cement (generally Fe-dolomite) suggesting rapid burial and isolation from the sediment/water interface. Paired and parallel channel fills are common. The characteristics of these structures are very similar to those ascribed to "gutter casts" by Whitaker (1973).

Some beds associated with the coquina rudstones, contain only articulated, cement-filled, smooth shelled brachiopods (most probably Atrypa erebus) with their commissure planes uniformly oriented with respect to bedding. No other recognizable skeletal debris is present, and the brachiopods 'float' in a matrix of lime mud. Although controvertible, it is suggested that these floatstone beds represent the biocoenose of the brachiopods, and hence the ultimate source of bioclasts for the coquinas.

The upper 10 cm of the Douro Formation consists of a very distinctively color-banded carbonate mudstone. Individual layers range from less than 5 mm to 3 cm in thickness. They tend to be laterally continuous, although nodular development is locally common. Macroscopically, bands are sharply differentiated from each other by weathering color (FIG. 2-4) which ranges from light tan, to orange-brown, to black. Microscopically, inter-band contacts are also sharp and contact zones do not exceed 200 microns in width. Lensoid subvertical fracture fillings are common along

planes of apparent layer offset. These offsets and fractures are generally restricted to no more than 3 or 4 layers, and commonly only one. The unique nature of the individual layers is seen to be a function not only of color, but also of average grain size, mineral composition and trace element concentration.

At the top of the Douro at Goodsir Creek a 50-100 micron thick layer of phosphorite and iron sulfide, similar to phosphatic "skins" described by Jarvis (1980), is overlain by a 3 to 7 mm thick shell lag consisting of phosphatized leperditid ostracodes, gastropods and ostracoderm fragments, within a partially phosphatized lime mud matrix.

2.5.4 Implications for Paleoenvironments

The presence of hardgrounds in the upper 10 m of the Douro implies a distinctive change in depositional regime, as such hardgrounds are unrecorded in the preceding 110 m. Most authors agree that synsedimentary submarine lithification requires that a sediment layer maintain its position within an optimum cementation zone, some distance below the sediment/water interface (Shinn, 1969; Bromley, 1965). The corollary to this is that sedimentation rates must be extremely low. Therefore, despite very high overall depositional rates for the Douro Formation, the upper 10 m was apparently laid down under conditions of limited or episodic sedimentation.

Fe/Mn encrustation appears to be restricted to lithified or at least partially lithified sediment in direct contact with marine water (see Schlager et al., 1976). Thus an agent is required to re-expose cemented sediment lying beneath the sediment/water interface. The angular discordance at the hardground 8.2 m below the top of the Douro, is best explained in terms of younger sediments infilling a pre-existing depression in older strata. Again a mechanism is required to excavate such a depression. Such an erosional agent could be either turbulent flow, or simply sediment removal by downslope mass transport as a consequence of slope failure (cf. Davies, 1977). The lack of a winnowed shell lag above the hardground argues for the latter, and this is further supported by the common occurrence of slide structures and contorted bedding within the uppermost Douro Formation elsewhere on eastern Cornwallis Island (PL. 2-1f). It is concluded that this particular prominent hardground probably represents an early lithified intraformational truncation surface (Laird, 1976). Nonetheless the structure shows close parallels to the "cuvettes" of the Cretaceous Anglo-Paris basin, which are interpreted by Jarvis (1980) as large, shallow erosional troughs, excavated by currents generated by prevailing winds. Such an origin can not be completely discounted in attempting to explain the discordant attitude of beds above and below the hardground.

The development of hardgrounds is often interpreted to be associated with periods of shoaling (e.g. Bromley, 1965; Shinn, 1969), and the reappearance of a diverse benthos in close stratigraphic conjunction with hardground development, is suggestive of such a shallowing. This conclusion is lent credence by the observations outlined below.

The sequence lying above the hardground (summarized in FIG. 2-4) exhibits an upward increase in both the number and thickness of shell lag deposits, and quartzose pelloidial calcarenite laminae. Similar shell lag packstones with rare gutter casts and plane to wavy laminated siltstones have been described by Kreisa (1981), and interpreted as storm-generated structures on an open shelf (ramp succession of Ordovician age in SW Virginia). Kreisa (1981) cited several properties of these storm deposits that vary with distance from the strandline. With decreasing depth, there is:

1. an increase in thickness and abundance of laminated beds of silt and sand,
2. an overall storm layer frequency increase, with storm layers increasing in thickness from .2 to 3 cm in deeper water to 8 to 28 cm in shallower water. Much of the shallow water record is composed solely of storm layers,
3. the appearance of hummocky-stratified and wave-ripple cross-laminated units, both of which are largely restricted to shallow water.

By analogy, the uppermost Douro beds represent a shallowing-upward sequence. From the first shell lag (approximately 6 m below the top of the Douro), all beds were clearly deposited above storm wave base. A more contentious issue is the shallowest depth achieved during this shoaling cycle. The absence of ripple cross-lamination and the very fine-grained (calcilutite) nature of the beds, suggest strongly that normal wave base was not achieved. Atrypoides erebus is considered to have preferred a very shallow subtidal to intertidal habitat (B. Jones, pers. comm., 1982), but the interpreted in-situ occurrence of these brachiopods in beds associated with the shell lag coquinas, does not commend itself to a peritidal interpretation. Furthermore the Douro Formation as exposed on Somerset Island, was all deposited above storm wave base and typically contains a diverse and abundant shelly benthos (Narbonne, 1981). Diversity is low in the storm shell lags, and fossils are virtually absent in the intervening strata. In particular, certain common Douro fossils are very under-represented such as the rhynchonellids, most smooth spiriferids (apart from A. erebus), gastropods and corals. It is unlikely that a shallowing upward cycle could commence with hemipelagic slope deposits and culminate in ?peritidal sediments without containing a substantial, shallow subtidal record with its normal biotic complement.

Therefore it is concluded that at the end of the Ludlow, in the vicinity of loc. 5, the Douro seafloor was raised above storm wave base (in the modern open ocean this can be as deep as 200 m -- Komar et al., 1972), but remained substantially below normal wave base (which ranges from 3 to 30 m). A similar conclusion was arrived at by Narbonne (1981, p.152) who considered the uppermost beds of the Douro at Goodsir Creek as "transitional between deep subtidal shelf and basin slope".

The vividly banded lime mudstone in the upper 10 cm of the Douro Formation, is the result of at least two processes. Different ferrous/ferric iron ratios, within adjacent sediment layers, have enhanced the color differences between layers. This phenomenon is considered to be primary, caused by alternating oxic and anoxic conditions at or near the sediment/water interface during deposition. The banded to nodular appearance itself is likely the result of syngedimentary lithification, similar to that described by Jones et al. (1978). This conclusion is supported by the presence of intrastratal fracturing, suggesting brittle and plastic response to loading by cemented and unlithified sediment layers, respectively.

By analogy to Recent environments of phosphorite formation, the distinctive phosphatic skin and shell lag in the uppermost Douro Formation at locs. 4 and 5, suggest conditions of limited supply of terrigenous and carbonate

detritus, dysaerobic conditions within the water mass overlying the substrate, and persistent upwelling (Jarvis, 1980). Modern phosphorites are forming off the Pacific coast of Peru and Chile and on the Namibian shelf off southwest Africa, at depths where the oxygen minimum layer impinges upon the continental slope or shelf. Phosphatogenesis does not proceed at depths greater than 400 m (Bentor, 1980), nor generally less than 70 m (Odin and Letolle, 1980). These values, however, are particular to specific oceanic circulation patterns, basin configuration, latitude, and the vertical extent of the oxygen minimum layer. Phosphorites are thus unreliable paleobathymetric indicators. Nonetheless the shallow upper limit is consistent with the envisaged bathymetry of the uppermost Douro beds. If the phosphatic skins are algal in origin as suggested by Jarvis (1980) and Bromley (1965), then the sequence probably developed within the photic zone (<150 m).

In summary, the upper 10 m of the Douro at locs. 4 and 5, are a shallowing upward succession that commences with interpreted deep-water, unfossiliferous basin slope facies and terminates with strata deposited somewhat above storm wave base in the photic zone (perhaps 100 to 150 m below sea level). The banded lime mudstones and thin phosphorite layer suggest low sedimentation rates and intermittently anoxic conditions within the basin at the close of Douro deposition.

2.6 MUDROCK FACIES

2.6.1 Problems of Classification

Although the dark colored, fine-grained siliciclastic facies of the lower member has been referred to as 'black shale' (Thorsteinsson, 1958; Gibling and Narbonne, 1978), it remains a moot point whether or not these rocks are indeed shale, let alone carbonaceous. The arguments that follow are principally semantic, and are pursued here solely for clarity, although it is the writer's conviction that the term 'black shale' is, without a doubt, a most appropriate and descriptive field appellation for these rocks.

The nomenclature adopted herein follows Lundegard and Samuels (1980), wherein fissility is not a primary determinative property, and it is the presence/absence of lamination that distinguishes 'shales' from 'stones'. In this context, the mudrock facies consists of clayshale, mudshale, mudstone, and both laminated and non-laminated siltstones.

If one accepts the criteria proposed by Curtis (1980), black shales have the following essential characteristics:

1. high organic content
2. high pyrite content
3. fissility arising from lamination
4. low carbonate content

Pettijohn (1957, p.622) suggested 5 % as the lower limit of organic content (by weight) necessary to qualify a shale as

black. The lower member mudrocks at the type section contain between .37 and .51 % C org. , a value an order of magnitude smaller than for most black shales (see Vine and Tourtelot, 1971), and more surprisingly, smaller than the mean value of .8 % for Phanerozoic shales (mean value from 78 determinations from Clarke in Pettijohn, 1975, p.274).

Curtis (1980) does not elaborate as to what constitutes a high pyrite content. The lower member mudrocks contain pyrite, both as finely disseminated single spherules, generally 7 to 10 microns in diameter (compound botryoidal forms up to 115 microns) and in the form of replacement mineralization of orthocone and gastropod shells. Visual estimates suggest that pyrite does not exceed 2 %, and in as much as the pyrite 010 peak at 56.3 degrees 2 theta, does not appear in powder diffractograms of the mudrocks, pyrite content is probably less than 5 %. Direct comparison with shales in general and black shales in particular (Pettijohn, 1975), indicates that the lower member mudrocks are average to below average in sulfur content. Although weight % sulfur is not directly proportional to sulfide content (some sulfur is contained within organic complexes), it nonetheless suggests that the lower member mudrocks are probably normal to impoverished in this mineral phase relative to the 'average' shale.

The 'paper' fissility commonly associated with black shales is not exhibited by the mudrocks of the lower member

at the type section. Although these rocks are extremely friable, and show a marked propensity to disaggregate upon wetting, they tend to break by conchoidal fracture across the plane of lamination (PL. 2-3a), and horizontal partings appear subcongruous with the faint lamination.

The mudrocks are carbonate poor, and further exhibit a number of other characteristics considered typical of carbonaceous shales (see Table 2-3). In summation, the lower member consists chiefly of dark to very dark colored mudrocks (olive-gray, showing a general trend toward black with overall diminution of grain size), containing an impoverished, mixed benthic and pelagic fauna. As such they are similar to, but not identical with, typical black shales. This lack of complete congruity requires caution in applying 'black shale' depositional models to the lower member.

2.6.2 Description and Vertical Variation

2.6.2.1 Components

Powder diffractograms indicate that the mudrocks are composed essentially of two minerals, illite-mica and quartz. This interpretation is verified in part by K-Na whole rock geochemistry (atomic absorption standard curve method). K/Na ratios of the mudrocks, and in particular the clayshale, closely approximate values for pure illite samples and are appreciably higher than values for shales

with more heterogeneous clay mineral assemblages (Table 2-4).

Table 2-3

Characters Commonly Exhibited by Black (Carbonaceous) Shales
(from Curtis, 1980; Pettijohn, 1975; Heckel, 1974;
and Twenhofel, 1939)

CHARACTER	STATUS IN LOWER MEMBER		
	PRESENT	DEVELOPED	ABSENT
1. High Organic Carbon Content (>3%)			X
2. High Pyrite Content			X
3. Fissility Arising From Lamination			X
4. Low Carbonate Content	X		
5. Depauperate, Restricted Fauna	X		
6. Predominantly Pelagic Fauna	X		
7. Rare and Micromorphic Benthos	X		
8. Non-Ecological Associations of Fauna	X?		
9. Slight Enrichment of Trace Elements V, U, Ni		?	
10. Phosphates			X
11. Thin-Shelled Mollusks	X		
12. Inarticulate Brachiopods		X	
13. Conodonts, Fishremains, Spores	X		
14. Clay Minerals, Predominantly illite	X		
15. Chert and Siderite Association			X

Table 2-4

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K, Na, K/Na Data from lower member mudrocks and other shales:

	L.M.	a.	b.	c.	d.	e.	f.	g.	h.	i.
% K	4.03	3.	2.58	2.63	1.13	7.38	5.9	5.47	5.07	5.73
% Na	.28	.74	.44	.30	.94	.44	.44	.24	.23	.26
K/Na	14.4	4.1	5.9	8.8	1.2	16.8	13.4	22.8	22.0	22.0

L.M. lower member, Barlow Inlet Formation

a. average Paleozoic shale (Pettijohn, 1975, p. 274)

b.-c. Devonian carbonaceous shale (Ibid, p. 283)

d. Permian carbonaceous shale (Ibid)

e. Precambrian slate - high potassic (Ibid, p. 286)

f. Cambrian high-potassic shale (Ibid)

g.-i. Pure ILLITE samples from Grim (1968, p. 580)

Above 40 m the mudrocks are interlaminated with fine subarkosic sand, and the gross mineralogic complexity of the mudrock facies is consequently increased (see subarkose facies for detrital components). The soluble fraction does not exceed 15 % by weight, and is principally very fine-grained calcite. As discussed previously, pyrite does not exceed 2-5 % of the mudrocks in volume. Organic carbon ranges from .33-.51 wt. % and is generally not observable in thin section, although coarser-grained mudrock towards the top of the lower member exhibits sapropel-rich wisps parallel to primary laminae (PL. 2-3b). Palynomorph macerations are rich in well-preserved spores, cellular sheets, tubes, and acritarchs (McGregor, pers. comm., 1980), and presumably much of the organic carbon was derived from relatively undegraded tissue. Phosphatic and calcareous fossil fragments, chiefly comminuted ostracoderm debris, trilobite elements, and molluscan bivalve shells, are locally common, but on the whole do not contribute substantially to the mudrocks.

The principal components remain unchanged through the basal 30-40 m of the lower member, although the relative percentages of these components vary systematically with stratigraphic height. Normative quartz shows a 48 % increase over the basal 30 m, concomitant with a perceptible textural coarsening (FIG. 2-3). Quartz within the mudrock occurs as subangular silt-sized, monocrystalline grains.

Normative quartz values, determined as excess wt % SiO₂ following illite, range from 20.5 to 31.2 %, and are in excellent agreement (Table 2-5) with calculations based on the TiO₂/Al₂O₃ ratio plot of Spears (1980). Weight percent of Ca shows little variation, suggesting that the carbonate fraction does not significantly change within the mudrocks.

2.6.2.2 Grain Size

Average grain size of the mudrocks shows a systematic increase upsection, reflecting increased amounts of quartz silt (cf. Grout, 1925; Blatt and Schultz, 1976). Grain size analysis of shale is fraught with difficulty (see Spears, 1980) and no attempt was made in this study to determine this textural parameter, apart from visual estimation. In thin sections, some clay minerals attain grain sizes of 40 microns, but the bulk of the clay minerals appear to be in the 3 to 15 micron range. Upsection, the shift toward coarser grain size does not occur to the exclusion of clay-sized detritus, but rather it is affected by the progressive increase in silt (to fine sand) content of the mudrocks (coarse-tail inverse grading on a megascale). By the 30 m mark, the mudrock contains an appreciable ($>2/3$) amount of silt-sized detritus, and is thus considered to be a siltstone. Nonetheless laminae of claystone aspect continue to occur throughout the mudrock facies and are even interlaminated in beds of the basal subarkose sandstone facies.

Table 2-5

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Normative Quartz in lower member Mudrocks at Locality 5.

Sample no.	Metres Above Base	Al O / SiO	Percent K O	Percent SiO	Normative Quartz Percent (SiO in Excess of Illite)	Percent Quartz Based on Spears (1980) Curve	TiO / Al O
7	50m					32.2	.050
6	30m	.27	4.64	55.26	30.4	30.9	.049
5	27m	.29	4.90	54.70	28.1	27.1	.046
4	20m	.30	4.83	54.19	27.1	27.1	.046
3	18m	.34	5.33	52.64	23.0	24.6	.044
2	10m	.36	5.38	50.87	20.5	23.3	.043
1	1m	.36	5.49	50.95	20.6	22.0	.042

Note: 3 Assumptions are made in these calculations

- 1) Only 2 mineral phases are present:
Illite and Quartz.
- 2) Illite has the composition
(OH) Ky (Al Fe Mg) (Si -yAl_y) O O;
where y = 1 to 1.5 and hence Al/Si in Illite
ranges from .714 to .846
- 3) All of Al is taken up in the mineral Illite.

Results from samples 7 may be unreliable, due to assumed higher feldspar content, making assumption 3 above invalid.

2.6.2.3 Color

The mudrocks exhibit considerable variation in color on dry, fresh surfaces. In general, the finest-grained mudrocks at the base are darkest, and colors are slightly but progressively lighter upsection concomitant with the textural coarsening. In terms of the color chart put out by the Geological Society of America (1975 edition), the basal clayshale overall is medium dark to dark olive gray (5Y 4/1 to 5Y 3/1). In detail much of the basal clayshale consists of alternating color laminae of olive gray (5Y 4/1) and dark gray (N3). Samples above 30 m can be described as mixtures of light olive gray and olive gray caused by lamination or bioturbation of coarser silt and clay. The dark olive gray colors are consistent with C org. values, ranging from .37 to .51 % (see Potter et al., 1980, fig. 1.25).

2.6.2.4 Stratification

Clayshale of the basal 10 m exhibits two orders of very faint, fine lamination: thicker (1 to 2.5 mm) laminae distinguished by subtle color differences (PL. 2-3a), and thinner (.2 mm or less), laterally persistent light and dark laminae. It is not clear whether these finer laminae are pervasive in the mudshale. Color bands in the lowest 10 m do not signify macroscopic textural differences.

From 10 to 20 m, the mudrocks appear to lack lamination, but this can be attributed to diminished color difference

between bands, rather than to sediment homogenization by burrowing. Above 25 m, weakly developed textural lamination reappears. Light olive gray siltstone laminae (.15 to 1.0 mm) are interbedded with dark olive gray clay-rich laminae (1.0 to 10. mm). This lamination is commonly disrupted by burrows attributable to the ichnogenus Chondrites (PL. 2-3c). Towards the top of the mudrock facies, lamination is increasingly conspicuous, culminating in intimate interlamination of fine arenaceous, silty, and clay-rich laminae (approx. .1 mm thick). These are grouped into thicker strata (1 to 8 mm) in which laminae of similar textures predominate (PL. 2-8d).

Lamination throughout the sequence is planar, although locally at the top of the mudrock facies laminae are disrupted by various post-depositional phenomena, including vertical burrows (Skolithos sp.) (PL. 2-8a), synaeresis cracks (PL. 2-7c), and contortion due to differential loading.

2.6.3 Macrofauna

2.6.3.1 Faunal Elements

The mudrocks contain a generally well-preserved, low-diversity macrofauna of locally variable abundance (very rare to common) that collectively represents a biofacies of restricted environmental aspect (Table 2-6).

The most common faunal elements are thin-shelled bivalves and brachiopods. Three indeterminate species of bivalves have been distinguished in the lower member (PL. 2-5), although only species A occurs in abundance.

Articulate brachiopods, which are present in prodigious numbers in parts of the underlying Douro Formation (see Jones, 1981), are represented by only four species in the lower member (PL. 2-6a,b), including a rhynchonellid (PL. 2-6g), a ?tannuspiriferid (PL. 2-6i), and two indeterminate species (PL. 2-6a,c,e,f,h,i). Lingulid brachiopods are relatively common but restricted to the upper siltstone beds and subarkose facies. Small, incomplete, and generally compressed shells of orthoconic nautiloids are present at different levels within the mudrock facies, as are comminuted remains of cyathaspidid heterostracans, predominantly Ariaspis sp. (PL. 2-4d). A new species of paraaulacopleurid trilobite (Chatterton and Packard, in prep.) is abundant at a few horizons in the clayshales (PL. 2-4a,b). The rest of the fauna has either a rare and sporadic distribution, or has been detected only as microscopic components in disaggregated mudrock. These fossils include diminutive crinoids (PL. 2-4i,m,n), gastropods (PL. 2-5h), ostracodes and scolecodonts.

Fauna, flora, and lebenspürren of the lower member.

* subarkose only, ** subarkose and mudrock facies.

Data from Uyeno (1977,1980), MacGregor and Narbonne (1978)

MACROFAUNA

Brachiopods

Atrypoides erebus Jones

Lingula sp.**

cyrtinid brachiopod

indet. brachiopod sp.A

indet. brachiopod sp.A

Mollusks

indet. modiomorphid bivalve sp.A

indet. modiomorphid bivalve sp.B**

indet. modiomorphid bivalve sp.C*

indet. pectinacean bivalve sp.D

indet. nuculid bivalve sp.E

orthoconic nautiloid

gastropod sp.A

gastropod sp.B

Graptolites

Monograptus n.sp. (cf. M.ultimus Perner)

Ostracoderms

?Ariaspis sp.

thelodont scales **

acanthodian scales **

Echinoderms

indet. microcrinoid

crinoid ossicles **

?holothurian sieve plates

Arthropods

Paraaulacopleura sp.

entomozoid ostracode

indet. ostracode

(continued)

Fauna, flora, and lebenspürren of the lower member.

MICROFAUNA

Conodonts

Ozarkodina confluens (Branson and Mehl) morphotype

O. excavata excavata (Branson and Mehl)

O.n.sp.A (Klapper and Murphy, 1975)

Kockelella variabilis Walliser

Oulodos sp.

Panderodus sp.

Pedavis sp. aff. P. thorsteinssoni Uyeno

Other

scolecodonts

chitinozoans

graptolite fragments

eurypterid fragments

MICROFLORA

Trilete Spores

Retusotriletes chartulatus McGregor

Punctatisporites sp.

Apiculiretusispora sp.

A. spicula Richardson and Lister

form resembling Synorisporites verrucatus

form resembling Retusotriletes triangulatus

?Emphanisporites protophanus Richardson and Ionnides

Acritarchs

Ammonidium sp.

Gorgonisphaeridium sp.

Moyeria uticaensis Thusu

leiospherid palynomorphs

Others

Tetrads (3 tetrahedral forms, one cruciform type)

tracheid-like fragments

spore-clusters

spirally-thickened tubes

cellular sheets

Hispanaediscus sp.

LEBENSPUERREN

Cruziana *

indet. arthropod scratch marks *

Planolites

Chondrites

Skolithos

2.6.3.2 Preservation

Most of the fossils are preserved as external molds and casts, and details of shell exteriors are commonly compressed onto internal molds. This obscures such features as muscle scars, but leaves external ornamentation sharply defined. Similar preservation has been described for Ludlovian bivalves of the Welsh borderlands by Watkins (1978). Some orthocones and gastropods are also preserved as steinkerns (PL. 2-5f,g) (the gastropods are commonly pyritized). Traces of chitino-phosphatic shell material are common (PL. 2-5f), and of greater significance, calcitic skeletal elements (e.g. crinoid ossicles, rhynchonellid shell material) are also present (PLS. 2-4i, 2-6i).

It is difficult to assess the degree of articulation within populations of bivalved organisms that remain fixed in mudstone matrix. Disaggregation of the mudrock destroys most shelly material greater than 1 mm, but very small articulated ostracodes, bivalves, and smooth-shelled articulate brachiopods can be obtained by this procedure. Most of the larger, matrix-embedded bivalves and brachiopods appear to be disarticulated, although for 3 of 4 species, at least one example of each seems to be an articulated, albeit open, valve pair (e.g. PLS. 2-5b,d; 2-6c). The lower member yielded a solitary example of crinoid which, within the limits of its host mudrock slab, is relatively complete and articulated (PL. 7d). Proetid

trilobites (PL. 4a,b) are exquisitely preserved but rarely complete. Their imperfection can usually be related to the retention of portions of the skeleton by both halves of a split slab of mudrock. In other instances (e.g. PL. 2-4a) the absence of free-cheeks suggests that the remains may in part represent exuviae. There is little evidence of disarticulation by burrowing or current activity.

2.6.3.3 Modes of Life

A large percentage of the organisms were clearly nektic (i.e. orthocones, proetid trilobites, and cyathaspidids) and thus share an indirect and enigmatic relationship with enclosing sediment. The paraaulacopleurid trilobite is small and thin-shelled, with a relatively broad thoracic axis, narrow pleural region, and unspecialized eyes. These attributes suggest a hemi-pelagic (nektobenthic) mode of life (W. Parkins, oral comm., 1982).

Of the shelly fauna, the rhynchonellid brachiopods are the only definitely sessile epifaunal benthos. All lingulids are considered by Chern (1979) to have been infaunal benthic organisms, although several authors including Ruedemann (1934), Bowen et al. (1974), and Watkins and Berry (1977) considered some lingulids to have adopted an epi (or pseudo-) planktic mode of life. Sheehan (1977) suggested that lingulids in deeper water sediments, may have established themselves as epifauna on sunken sea-weeds.

Cited examples of epiplanktic lingulids are generally very diminutive species. Lingula lata, considered by Watkins and Berry (1977) to be epiplanktic, has shells .2 to .6 cm long (see Chern, 1979, fig.3). The few lower member lingulids recovered are 1.3 to 1.6 cm long, and are unlikely to represent an epiplanktic fauna, even if such a mode of life was indeed pursued by other lingulids.

Without precise identification of the bivalves, only speculations can be made about their mode of life. Bivalve species A and B appear to be modiomorphids, a group generally regarded as being semi-infaunal and endobyssate (e.g. Watkins, 1978). Bivalve species C is a member of the Pectinacea (cf. Cornellites sp.). Most pectinoids are epibyssate and sessile in early stages of growth, but some become free-swimming as adults. The inference is commonly made that byssate bivalves in dark gray and black shale sequences, may have been epiplanktic, but incontrovertible evidence of this has not been documented (e.g. Pettijohn, 1957; Heckel, 1972).

It is unclear whether the solitary, diminutive, stalked crinoid represents a juvenile form or an adult 'microcrinoid'. Uniformly small diameters of pluricolumnal elements extracted from disaggregated shale, are consistent with the pelma diameter of the articulated specimen. In the absence of larger ossicles, this feature would favour the identification of the specimen as an adult 'microcrinoid'.

Stalked crinoids are sessile and benthic, although several genera (none of which are microcrinoids) have been described as being pseudoplanktic (Gnass, 1937; Wells, 1941; Seilacher et al., 1968).

Microfossils extracted from a layer 10 m above the base of the lower member (loc. 5), include dextral low-spired involute gastropods, sinistral advolute low-spired gastropods, entomozoid ostracodes, nuculoid bivalves, indeterminate brachiopods, and monograptid stipes. Except for the graptolites, none of the fossils exceeds 1.2 mm in long dimension; probably most represent pelagic organisms.

2.6.4 Environmental Affinities and Autochthony

Available faunal data do not allow for precise delineation of the depositional environments of lower member mudrocks, particularly given the difficulty in determining autochthony.

More vagile members of the fauna (i.e. ostracoderms, eurypterids) have been considered diagnostic in interpreting the depositional environments of the lower member (Narbonne in McGregor and Narbonne, 1978), but their significance is ambiguous without better understanding of the taphonomic processes affecting them. For example, Silurian ostracoderms are considered to have occupied a marginal marine niche (Dennison, 1956), but they also occur in deeper-water basinal carbonates in the Arctic (e.g. Poey,

1982). Diminutive orthocones and proetid trilobites appear to be unique to the lower member mudrocks; in contrast, eurypterids and the ostracoderm, Ariaspis sp., have been identified from the basal carbonate lithofacies of the upper member. Possibly the orthocones and trilobites were particularly adapted to prevailing conditions in the water column of the lower member depositional basin, but they do not define these conditions.

Rhynchonellids are robust, environmentally wide-ranging brachiopods, commonly occurring in black shales which lack a diversified marine fauna (Moore in Moore, Lalicker and Fisher, 1952). (in Moore, Lalicker, and Fisher, 1952, p.241), "They [rhynchonellids] preponderate also in some shaley deposits, especially black shales which lack a diversified marine fauna. These observations indicate that at least some of the rhynchonellids are well-adapted to environments in which other brachiopods could not get along well". Where present, rhynchonellids tend to be clustered in their occurrence on mudrock slabs. However, as they are preserved as internal molds, it is not possible to ascertain the percentage of articulated valves, and hence a measure of their relative autochthony. Nonetheless, it is difficult to reconcile the anticipated current strength necessary to transport for any distance these relatively globose (if articulated) brachiopods, with the calm water conditions implied by the very fine-grained enclosing matrix. They are

therefore considered to be members of an autochthonous benthic community. Specimens similar to (and conceivably representing the same species as) the mudrock rhynchonellids occur in the subarkose facies and basal carbonate facies of the upper member of the Barlow Inlet Formation. It is thus possible, that, this particular brachiopod had wide environmental tolerance.

As individual morphotypes, the bivalves shed little light on depositional environments of the lower member mudrocks. Collectively, their small size and thin shells, are comparable to Ludlow bivalves from deep water siliciclastic sequences (Watkins, 1978; see also FIG. 2-5, this paper). Micromorphic bivalves (commonly Pectinacea) in mollusk-dominated faunas, also characterize dark gray and black

BIOMETRICS OF SHELLY FAUNA

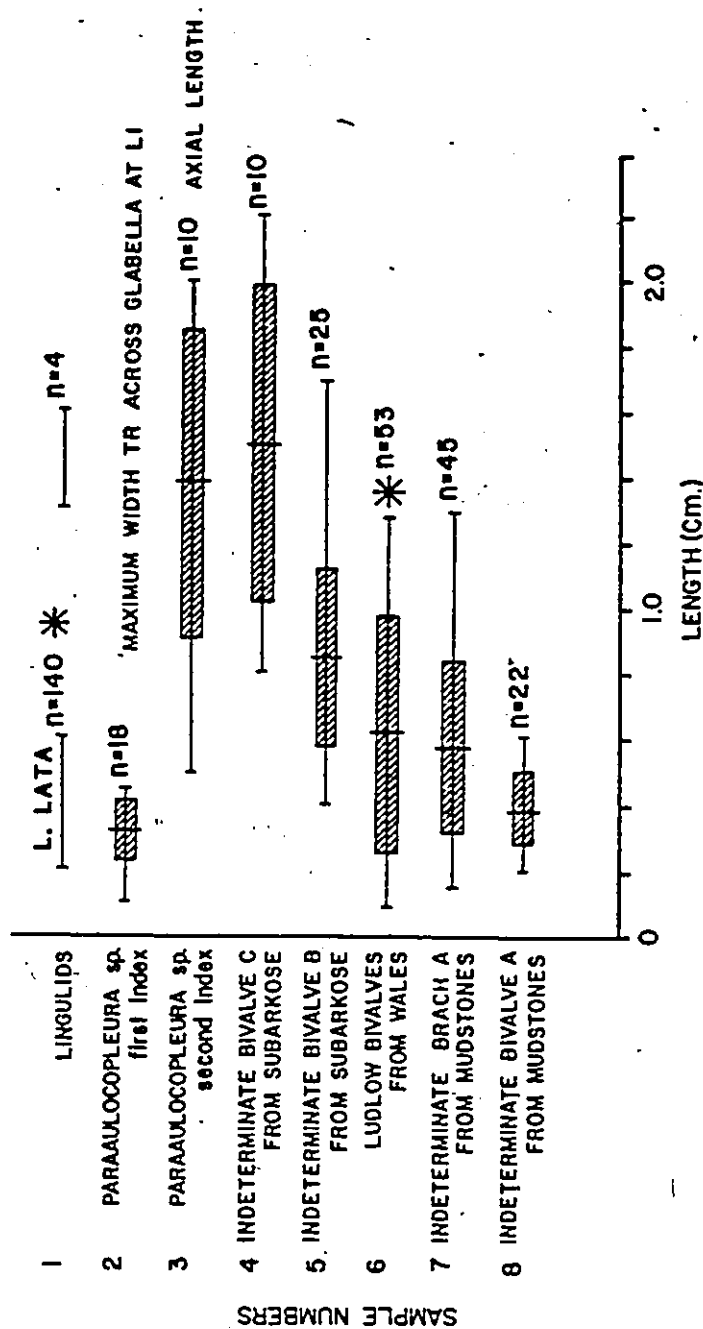


FIG. 2-5 Biometrics of shelly fauna from lower member. * indicates data from Ludlovian basal strata of Wales. Statistics on *Lingula* (L. lata) are from Chern (1980) and those on Ludlow bivalves in laminated shales from Wales are from Watkins (1978).

shale deposits (Twenhofel, 1939; Pettijohn, 1975; Heckel, 1977; Malinky and Mapes, 1982). Bivalve faunas described in the above cited literature appear adapted more to soft substrates and poor oxygenation, than to high turbidity.

An allochthonous origin for the lower member bivalves is suggested by:

1. the apparent high degree of disarticulation, despite general lack of evidence of bioturbation or current activity (at least in the basal 30 m).
2. absence of bivalves in infaunal positions despite inferred rapid depositional rates (this argument holds true only if species B and A are semi-infaunal modiomorphid clams).

Nevertheless there are problems in postulating agents capable of introducing exotic bivalve faunas to the lower member substrates. If some or all of the bivalves were transported initially as epiplankton, what suitable floatant was available in the Upper Silurian that might have acted as an attachment site (large nekctic animals? marine thallophytes?)? If epiplanktic, why are the bivalves absent at loc. 8, a more distal and deeper depositional site for the lower member? If the bivalves were alien, and introduced by storms, their concentrations should be associated with laminae of quartz silt or sand detritus, possibly graded. Such associations do not occur.

Because of the ambiguity of available evidence, the autochthony of the bivalve fauna must remain an open question. The observation that these particular bivalves are preserved solely in the lower member, suggests that they were specifically adapted to conditions in the lower member basin, either on the substrate or in the water column.

Modern lingulid brachiopods generally live in the intertidal zone, but are common to a depth of 40 m, and occur rarely as deep as 80 m (Cherns, 1979). Lingulids are thought to be tolerant of high levels of turbidity (Rudwick, 1965) and low levels of oxygen (Cloud, 1948). In the ancient record, the Lingula community is considered to be the shallowest assemblage (Ziegler et al., 1968), and deeper water occurrences of lingulids have been generally attributed to epiplanktic transport. Cherns (1979) suggested that lingulids as a group are eurybathic. Unfortunately, the lingulids from the lower member occur in bedding parallel positions, and so their relationship to the enclosing sediment is unclear. Their excellent preservation, despite the fragility of their thin shells, suggests a limited transport history, and they are herein considered to be part of the lower member benthic biocoenosis.

2.6.4.1 Distribution

The commoner faunal elements show some indication of vertical zonation within the lower member type and reference section. Brachiopod species A appears to be restricted to the lower 18 m, and bivalve species B is rare above this level (FIG. 2-3). Furthermore, the rare specimens of species B above the 18 m mark are on average larger than those lower in the sequence. Rhynchonellids and lingulids appear to be restricted to the upper 30 m of the mudrock facies. It is doubtful, however, that sampling was adequate to account for rare fossiliferous horizons that substantially bias the data in FIG. 2-3, and thus the poorly-defined apparent zonation, may well be an artifact of sampling.

Evidence of biogenic disruption of sediments begins at, and continues above, the 25 m mark, with the occurrence of the trace-fossil Chondrites sp. (PL. 2-3c). This suggests that although substrates may have been an inhospitable resource for burrowers initially, substrates of the upper 25 m were capable of supporting vagile infauna.

2.6.5 Microflora and Microfauna

The mudrocks yield abundant well-preserved microorganisms or micro-skeletal elements including: conodonts, scolecodonts, ostracodes, fragments of gastropods, sponge spicules, holothurian sieve plates, graptolite stipes, eurypterid cuticle, and a diverse assemblage of palynomorphs.

of both terrestrial and marine origin (Table 2-6). McGregor (pers. comm., 1980) wrote, " [the lower member] is exceptional for strata of late Ludlow age in the diversity and abundance of its organic microfossils, in the abundance of spores and other land-derived palynomorphs at some localities ... and in the excellent preservation of the microfossils at some stratigraphic levels". Palynomorph preparations have yielded trilete spores, tetrads, spore-clusters, acritarchs, scolecodonts, chitinozoans, tracheid-like fragments, spirally-thickened tubes, cellular sheets, and fragments of eurypterids, phyllocarids and graptolites (McGregor and Narbonne, 1978; McGregor, pers. comm., 1980). This suite of organisms is significant in representing a mixture of organisms, some characteristic of marine environments and others continental. This suggests that during latest Ludlow time, localities 4 and 5 were proximal to a strandline, and that terrigenous palynomorphs were dispersed seaward.

Paleozoic black shales commonly contain conodonts, rare fish remains, spores and spore cases (Pettijohn, 1975). Twenhofel (1939) considered spores and spore cases amongst other organic remains to represent a mixture of allochthonous and autochthonous elements, characteristic of his 'bituminous' (as opposed to 'humic') black shales.

2.6.6 Paleoenvironmental Implications

Perhaps these mudrocks are most notable faunally in lacking the fossil assemblages considered typical of Upper Silurian marine shelf facies in the study area. Units immediately above and below the lower member contain atrypoid brachiopod-dominated communities that include sessile colonial filter-feeders (i.e. favositid corals, bryozoans, stromatoporoids). Conditions suitable for colonization of lower member substrates by such organisms evidently did not exist.

Generally the fauna is most concentrated in the basal 15 m. Fossils only occur in abundance at select horizons within the mudrock facies, and many zones are completely barren. This irregular distribution of fossils could be due to one or more of the following processes:

1. pelagic kill -- layers are enriched in chiefly planktic and epiplanktic organisms as the result of short term toxicity (e.g. increased turbidity, decreased dissolved oxygen) in the normally productive surface waters of the basin (see discussion by Pettijohn, 1957, p.623-624).
2. bottom condition amelioration -- prevailing anoxic and/or highly turbid conditions at the substrate/water interface, are altered such that a restricted, tolerant benthic fauna can become established (see Rhoads and Morse, 1971).

3. sediment dilution -- faunal concentration would accompany low rate of sediment influx, and faunal impoverishment high rates. This factor would probably serve to modify the first two processes.
4. storm lag deposition -- allochthonous fossil material derived from nearshore environments can be dumped offshore by storm-generated current action (see Kreisa, 1981, p.825, for a discussion of the limited capability for storms to mix faunas from different benthic assemblages). In the absence of storm-generated structures this model cannot be substantiated.

Because the choice between the first two processes depends on the discrimination of indigenous and foreign faunal constituents, the most probable of these processes cannot be identified.

The fossils are generally diminutive, particularly in the case of the bivalves, orthocones, gastropods and crinoids. This micromorphism, together with the predominance of a molluskan fauna, are attributes of anoxic 'black shale' biofacies (Twenhofel, 1939; Heckel, 1977; Malinky and Mapes, 1982). Conversely, the relatively low organic carbon and pyrite contents of the mudrocks, is unlike values associated with condensed shale sequences deposited under anaerobic conditions. This problem is discussed more fully in the section entitled "Paleoenvironmental Synthesis".

2.7 SUBARKOSE FACIES

2.7.1 Lithology

The uppermost few metres of the lower member at all localities south of, and including Goodsir Creek (loc. 5), consist of a distinctive, substantially coarser-grained siliciclastic facies, that can be characterized as fine-grained subarkosic sandstone. The transition from laminated siltstone to the first bed of relatively homogeneous buff-weathering sandstone is everywhere abrupt, and locally erosive, as evidenced by the incorporation of siltstone intraclasts in some basal sandstone beds. The lower .2 m portion of the subarkose facies typically contains thin, continuous shale partings.

2.7.2 Mineral Components

Five principal mineral phases (excluding opaques and phosphatic skeletal grains) occur within the subarkose facies. These include quartz (73-86 %), K-feldspar (10-25%), dolomite (1-9 %), plagioclase (1-5 %), and calcite (<2 %). Apatite and muscovite are rare accessory minerals. Relative proportions of quartz to feldspar were estimated by point counts following sample preparation with combined Na cobaltinitrite and rhodizonate stain. These modal percentages were in part substantiated by cathodoluminescence photomicrography and x-ray fluorescence. Leaching by HF fumes (without staining) produced sufficient

relief contrast between quartz and the other components, to be most useful in estimating modal composition. The rhodizonate stain was ineffectual in selectively identifying plagioclase in the fine-grained sandstone. Plagioclase, was identified by the presence of polysynthetic twins in cross-polarized light, and by its high luminescence under electron bombardment (PL. 2-3h). Some samples contain up to 15% comminuted heterostracan fragments.

2.7.3 Texture

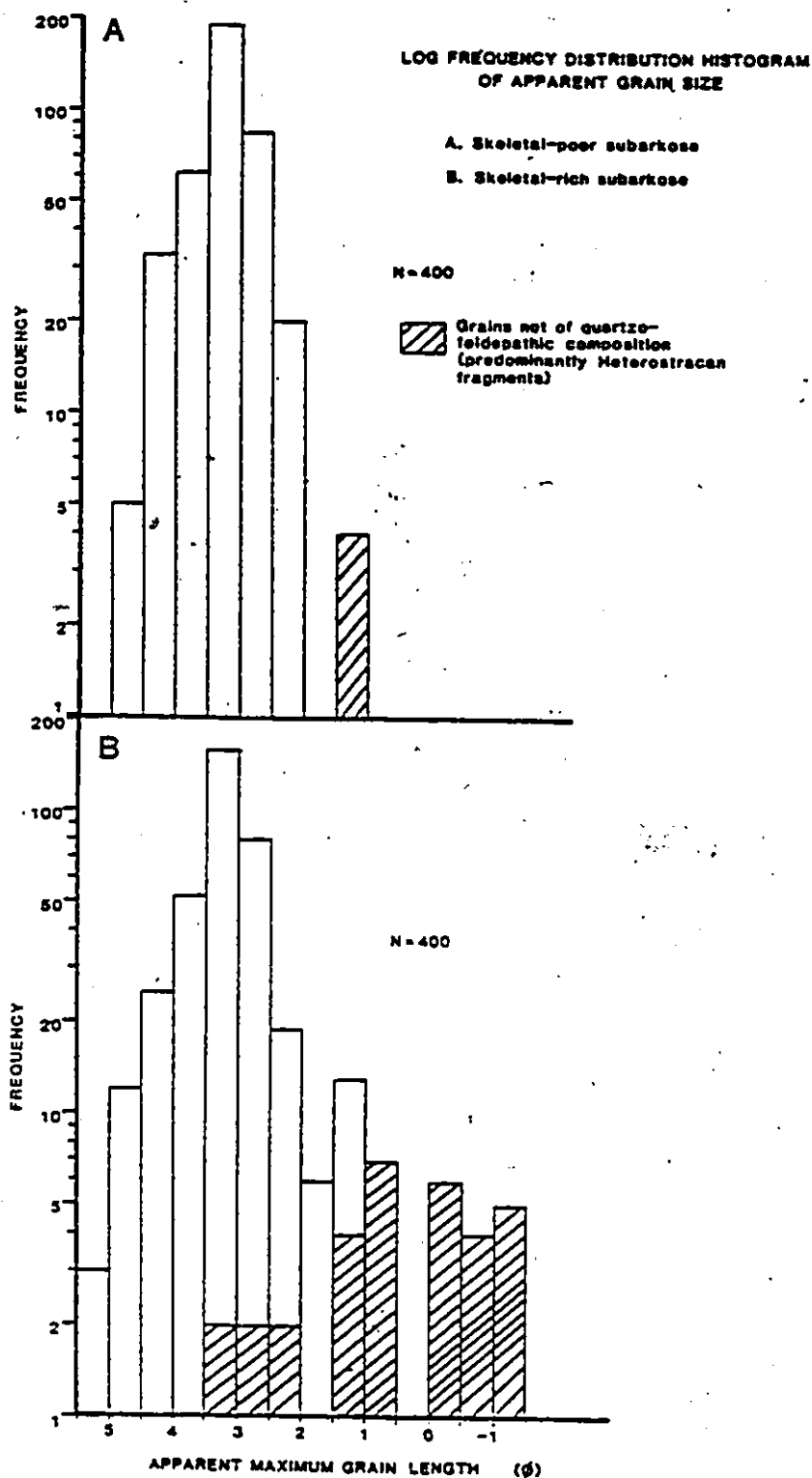
The subarkose has a granular, interlocking mosaic texture (PL. 2-3d,e), reflecting an inordinately high average number (5.6) of grain contacts per grain. According to Taylor's (1950) classification, contacts are approximately 70 % concavo-convex, 20 % long, and 10 % sutural. In plane and polarized light, both feldspar and quartz grains are unaltered, and internally show no dust, Fe-oxide or inclusion-rich rims that might disclose original detrital cores. These observations suggest a high degree of compaction and hence considerable depth of burial. However, apparent pressure solution features are restricted to concavo-convex and sutured grain contacts. Fragments of shell material are undistorted and uncrushed, and detrital muscovites conform only slightly to surrounding quartz grains. Detrital cores can be differentiated from authigenic syntaxial rim cements on many feldspar grains

(PL. 2-3g,h), using cathodoluminescence photomicrography (ELM-1 Nuclide luminoscope). Unfortunately, even with a focused beam at 18kv, the quartz grains did not luminesce. Nonetheless, given the presence of authigenic cements on the feldspars, it is quite possible that the detrital quartz grains have similar overgrowths, in which case most grain contacts can be interpreted as compromise boundaries between competing syntaxial rims.

The original feldspar detrital grains are not markedly rounder than the combined core and overgrowth, which are generally angular to subangular. Based on the assumption outlined above, original grain shapes for quartz are inferred to have been similar to feldspar grain shapes.

Most samples of the subarkose are well-sorted and are normally-distributed (FIGS. 2-6a, 2-7; Table 2-7). Mean grain size (see Friedman, 1967) for undifferentiated 'core and rim' clasts, is 3.7 phi, and although this must be considered maximal, it should be noted that feldspar detrital cores show less than 10 % increase in diagonal length due to authigenic cement, and therefore the figure of 110 microns probably closely approximates original average grain size. Samples of subarkose having a high content of heterostracan fragments, exhibit distinctly bimodal grain size distribution (PL. 2-3e,f). Coarse quartz sand grains are apparently subrounded to rounded, in contrast to co-occurring subangular fine sand.

The coarse-grained fraction, composed of monocrystalline grains of unstrained quartz, is well sorted, with a mean of 1.4 phi and a standard deviation of .4 phi. Associated heterostracan fragments are poorly sorted and range in size from 3.2 phi to .1 phi, having a mean of 1.01 phi. Presumably some degree of hydraulic equivalence promoted simultaneous deposition of ostracoderm debris and coarse quartz grains. •



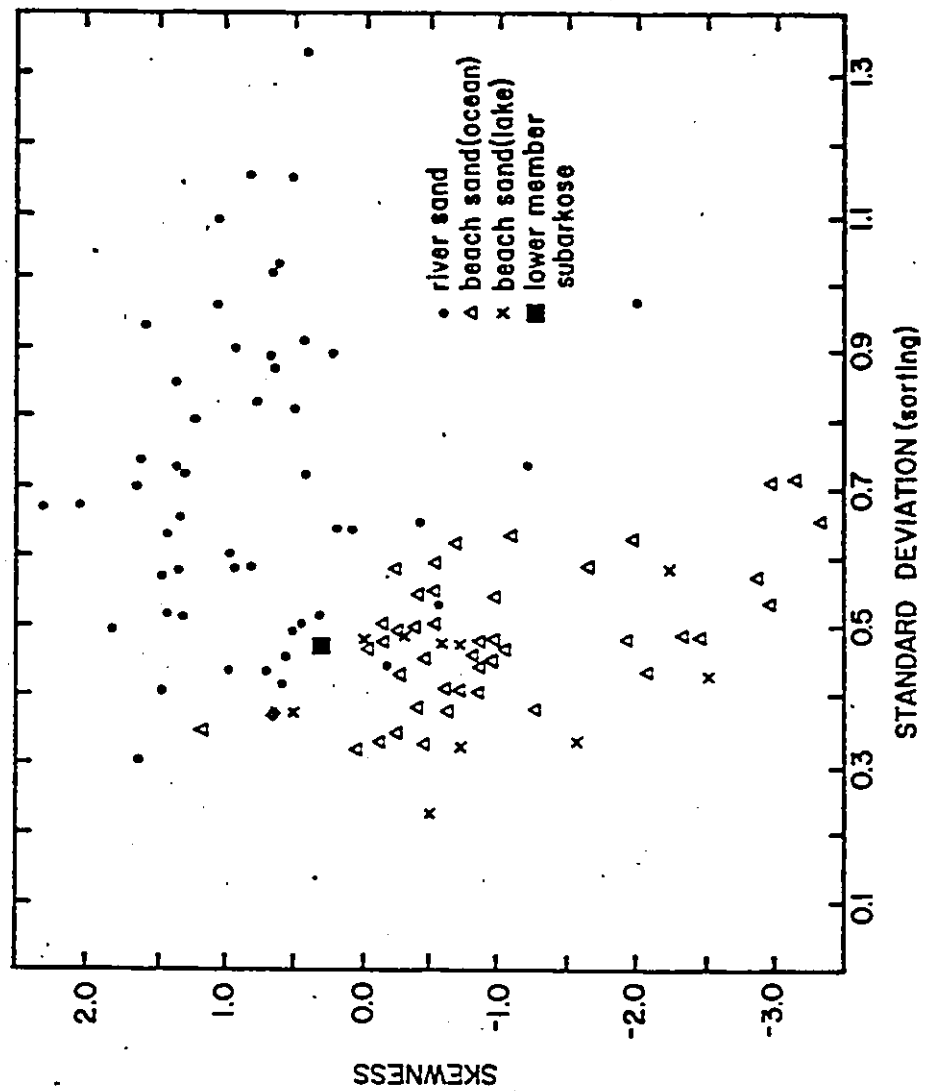


FIG. 2-7 Skewness-sorting plot for subarkose. Lower Member datum from thin section transforms (cf. Friedman, 1958). Skewness and sorting values computed following method of Friedman (1962).

Modal Analysis of Subarkose

Component	Range (in % area)
Quartz	73-86%
Feldspar (total)	11-27%
Plagioclase (est.)	1-5%
K-Feldspar (est.)	10-25%
Calcite	tr-2%
Ferroan dolomite	1-9%
Phosphatic skeletal material (mostly heterostracan)	1-6%
Apatite	tr
Muscovite	tr

- 1) ranges obtained from point counts (300 each) on 10 thin sections of representative samples.
- 2) feldspar was differentiated from quartz on the basis of:
 - a. reduced relief (observed on sections after their treatment with HF fumes.
 - b. staining by Na cobaltinitrite and rhodizonate
 - c. cathodoluminosity.
- 3) plagioclase was differentiated from K-feldspar on the basis of
 - a. the presence of albite twins
 - b. heightened cathodoluminoscity
 - c. slight staining by rhodizonate.
- 4) carbonates were distinguished on the basis of their high birefringence and by staining with Alizarin red S and K ferricyanide.

2.7.4 Sedimentary Structures

Listed in Table 2-8 are structures occurring either in outcrop or in isolated talus blocks of the subarkose facies. Primary structures preserved topographically include straight-crested, weakly asymmetric and symmetric ripples (PL. 2-7a,b), parting lineation, and runzel marks (PL. 2-7d). The ripples tend to have low symmetry indices (1-3) and high ripple indices (11); they lack chevron structures and have unidirectional foreset laminae. Most bedding planes lack primary structures, and this is reflected internally by a predominance of plane and/or gently inclined parallel lamination. In the latter, laminae dip uniformly at no more than 5 degrees with respect to bottom and topset beds. Small-scale channels (PL. 2-7e) and ripple cross-lamination are also common.

Post-depositional but pre-lithification structures include desiccation polygons (PL. 2-7f), and trifurcating, spindle-shaped shrinkage cracks representing either syneresis cracks or incompletely developed subaerial desiccation cracks (PL. 2-7c). Penecontemporaneous structures related to compaction include micro normal faults (PL. 2-8i), load casts (PL. 2-8g), and numerous deformed vertical sand dyklets that may in part represent syndimentary injection phenomena.

Sedimentary Structures Occurring in the lower member
(m) Structure occurs in Mudrock Facies, (Sm) occurs in both facies

PRIMARY BEDDING PLANE STRUCTURES

Asymmetric ripples
Symmetric ripples
Primary current lineation

PRIMARY INTERNAL STRUCTURES

Ripple cross-lamination
Parallel lamination - colour (m)
 - textural (Sm)
Low-angle inclined bedding
Small-scale channel structures

PENECONTEMPORANEOUS BEDDING PLANE STRUCTURES

Desiccation polygons
Synaeresis cracks (Sm)
Runzelmarken

PENECONTEMPORANEOUS INTERNAL STRUCTURES

Load casts
Injection phenomena (sand dyklets) (Sm)
Micro-faults

2.7.5 Biota and Biogenic Structures

Four ichnogenera have been identified in the subarkose facies, of which three, Skolithos, Planolites (PL. 2-8b), and Chondrites (PL. 2-8f), are relatively simple, environmentally non-diagnostic forms that occur but rarely in this facies. Several isolated bedding soles, however, exhibit, in hyporelief, extremely abundant, more morphologically complex, bilobate, crawling traces referable to Cruziana (PL. 2-8e). In addition several indeterminate arthropod scratch marks are present (PL. 2-8c).

In general, the subarkose facies contains few body fossils. Comminuted heterostracan elements predominate in the most fossiliferous strata. External moulds of at least one species of rhynchonellid brachiopod (PL. 2-9a,c), and two species of bivalve (PL. 2-9e,f), are locally abundant. A solitary, well-preserved single lingulid valve was also identified (PL. 2-9b). This lingulid brachiopod is relatively large (15 mm long), and although not preserved in an infaunal position, shows little evidence of transport.

2.7.6 Sandstone Distribution

No outcrop of subarkose has been identified north of loc. 5, although at loc. 6, ripple cross-laminated quartzose calcisiltite occurs in talus resting on the lower member. This sample may be part of a distal northern tongue of the subarkose facies. Lower member sandstone crops out as far

south as loc. 2 (the southernmost exposure of the lower member examined on Cornwallis Island), but does not occur at loc. 1 (Griffith Island). Throughout its distribution, the thickness of the subarkose facies is remarkably consistent, but its proportion to total thickness of the lower member, decreases dramatically to the north (FIG. 2-2). Overall, its uniform thickness and persistent occurrence suggest that the subarkose is a 'sheet-like' sand body, extending at least 10 km in a north-south direction. Its east-west extent remains unknown. It shows no significant changes in sedimentary structures, fauna, or grain size over this distance.

2.7.7 Paleoenvironmental Implications

The large lingulids in the subarkose facies are suggestive of a nearshore shallow environment (Ziegler et al., 1968), and the predominance of Cruziana sp. on some bedding soles suggests that the subarkose was deposited, at least in part, within the Cruziana ichnofacies, typically developed in well-sorted sands and silts laid down in shallow water (Seilacher, 1967).

Preponderance of low-angle inclined bedding is highly suggestive of a beach face environment (therefore high energy), an interpretation consistent with the presence of parting lineation. The ripple forms (symmetric with unidirectional foreset laminae) are suggestive of wave

ripples that develop in the nearshore zone (Newton, 1968), and the relatively high ripple index (most wave ripples are between 6 and 8) can be attributed to the very fine to fine sand size, and may suggest formation of the ripples in the surf zone (see Reineck and Singh, 1980, p.32).

Common occurrence of desiccation polygons and the presence of runzel marks, both indicate periodic emergence of the depositional substrate, reinforcing a 'shallow-water' interpretation for this facies. Small-scale syndimentary gravity faults, may attest as well to a 'wetted' but unsaturated condition of the original sands (see Steidmann, 1974), a situation anticipated in emergent areas of the beach littoral zone.

Micro-faulting, load casts, and deformed vertical burrows all reflect sharp density gradients and compactional differences between the overlying fine sands and underlying mudrocks. Further, these phenomena, together with the intimate interlamination of coarse and fine detritus at the base of the subarkose facies, suggest rather rapid sedimentation within an environment of fluctuating hydraulic energy and sediment supply.

On the basis of biota and structures present, it is reasonable to conclude that the subarkose facies represents principally beach foreshore and shoreface depositional environments. The inferred sheet-like configuration of the sandstone body is consistent with such a shallow marine environment.

Textural evidence is ambiguous, but does not refute the environmental interpretation outlined above. Samples exhibiting unimodal grain size distributions are well sorted (S.D. = .47 phi), as would be anticipated for beach sands. By inspection and calculation, the grain size distribution appears to be unskewed to slightly positively skewed, a feature uncharacteristic of most beach deposits. The accuracy of skewness estimates from thin section data is, however unreliable (see Friedman, 1962), primarily due to difficulty in correctly determining the percentage of very fine detritus (Blatt, Middleton and Murray, 1972, p.62). Thus the skewness value given (+ .425; see also FIG. 2-7) must be regarded with suspicion.

The validity of utilizing thin section transform data in C-M plots has been questioned (see Passega, 1964), nonetheless, median and one percentile values (10 thin sections, 400 counts each) exhibit good reproducibility and are considered sufficiently accurate for gross comparison to data from sieved sediment. In FIG. 2-8, the subarkose facies plots on the edge of the field for beach deposits, suggesting that the relatively small median grain size is atypical with respect to most beach sediment. It is not clear why the sandstone is so fine-grained, although grain size no doubt relates generally to source rock crystallinity and competence of the transport medium.

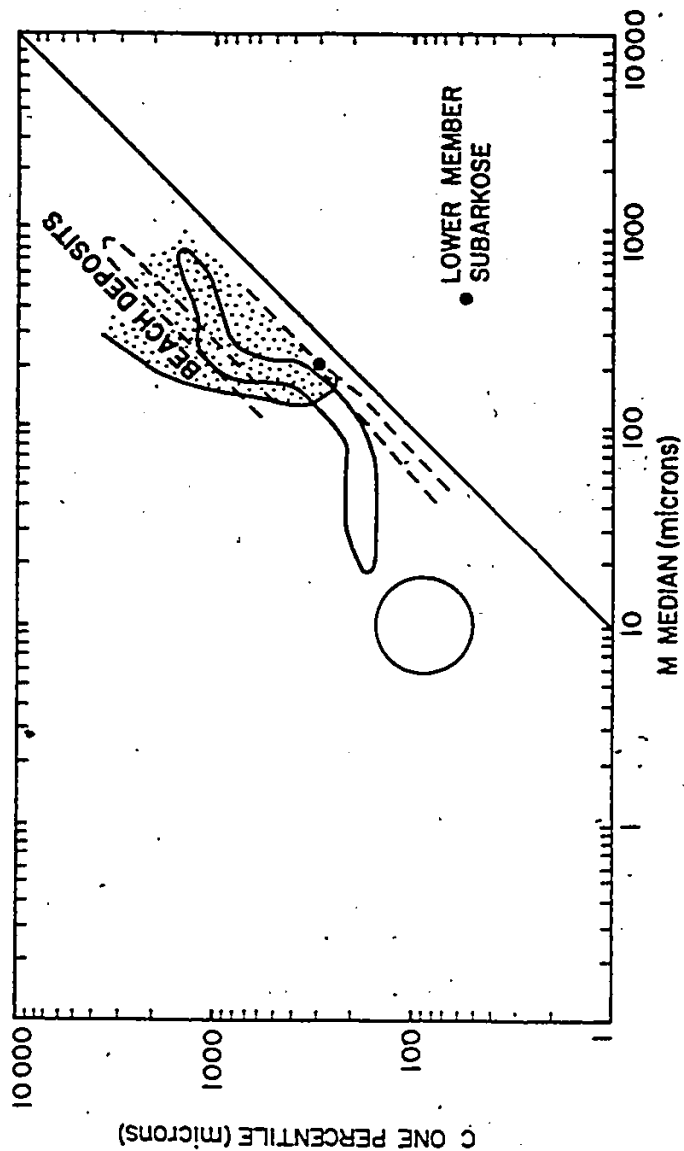


FIG. 2-8 C - M plot for subarkose. Lower Member datum determined from thin section transforms.

The subangularity of feldspar detrital cores (and presumably most other grains), would be anomalous for beach sediment of coarser grade, but is consistent with the fine sand size of the rocks in question (Waskom, 1958).

The most interesting textural characteristic of the subarkose facies is the local bimodality in grain size distribution. Textural inversion phenomena in sandstones have been attributed to a variety of mechanisms (see Scholle, 1979, p.106), including multiple sources of sand, in-situ mixing processes (burrowing), and storm mixing of material in a high energy environment. Folk (1969) promoted sorting by aeolian processes, and production of wind deflation lags, as important in the origin of many bimodal sandstones.

Pertinent to a marine nearshore to shoreline interpretation for the subarkose facies, are studies (e.g. Schiffman, 1965; Visser, 1969) that report markedly bimodal size distribution of beach sands in the transition area between swash and surf zones. The perceived difference in roundness between the two grain size modes (PL. 2-3f) is likely an artifact of pressure solution, as is discussed by Potter (1977).

2.8 OVERLYING STRATA OF THE BARLOW INLET FORMATION

At the type and reference sections, the contact (as redefined herein) between the lower and upper members of the Barlow Inlet Formation is conformable and abrupt, although not evidently erosional. Overlying carbonates consist of gray-green weathering, slightly argillaceous, peloidal lime wackestone to mudstone, with an abundant and diverse open marine macrofauna (see Paper 5). The largely comminuted remains consist of crinoid debris, brachiopod and ostracode shell fragments, sponge spicules, and bryozoans.

The basal carbonates represent a major regional transgression coupled with cessation, or at least reduction of, terrigenous clastic input to Jones-Lancaster basin. The carbonates locally contain up to 10% coarse quartz silt (40-50 micron average), but most samples contain less than 3 % silt and no arenite-sized detritus. Leaching of three samples yielded 13 to 25 % by weight insoluble residue. These values compare favourably with 'typical' Douro limestones that have a mode of 10 to 20 % insoluble residue (Veizer et al., 1978, fig. 6). There is no evidence to indicate extensive reworking of the beach sands, and a ravinement deposit (e.g. Klein, 1974) commonly associated with transgressive sequences, is not evident. This suggests that resubmergence proceeded rapidly (Reinson, 1979, p.61).

2.9 REGIONAL VARIABILITY

2.9.1 Palynomorph distribution

Relative abundance and type of palynomorphs in basin samples has been used to define relative proximity of parts of a basin to an ancient strand line (cf. Cross, 1967; Peltzer, 1975). Semi-qualitative palynomorph analysis on 22 samples from the lower member (FIG. 2-9) indicates a clear increase, from south to north, and east to west, in the proportion of marine to terrestrial-derived palynomorphs. This suggests a source area for the palynomorphs of continental aspect, and hence a shoreline, to the south and east.

2.9.2 Organic Carbon

Six representative samples of lower member mudrocks were analyzed for their C org content. Care was taken to ensure that the three basal samples were from equivalent stratigraphic positions. The limited data suggest that organic carbon is progressively enriched in the lower member to the north (Table 2-9). It also appears that the basal beds contain the highest abundance of organic carbon.

Palynomorph distribution in the lower member, Barlow Inlet Formation, Cornwallis Island. Analyses by D.C. McGregor (Geological Survey of Canada) from samples submitted by the author.

	West<-South<----->North																					
Locality	1	3	4	5					6				7				8					
Height above base(m)	1	5	5	1	50	35	0-1	10	18	18	27	30	4	10	15	25	1	25	55	65	1	5
Spores			x	x	x	x	x	x	x	x	x	x		x	x	x				x		
'Tubes'			x	x	x	x	x	x	x	x	x	x	x		x	x				x		
Euryp- terids	x	x	x	x	x	x	x	x	x	x	x	x		x	x	x				x		
Acrit- archs	x	x	x	x	x	x	x	x	x	x	x	x	x	x	x	x	x	x	x	x	x	x
Scole- codonts	x	x	x	x				x	x	x		x	x	x		x			x	x		x
Chiti- nozoans		x	x	x			x	x		x		x	x	x	x	x	x	x	x	x	x	x
Grap- tolites												x	x	x	x				x			x

Table 2-9

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Weight Percent Organic Carbon in lower member
Analysis by T.G. Powell of the Geological Survey of Canada.

Height Above Base (m)	North<----->South		
	Locality No. (Total Thickness of lower member)		
	8 (20 m)	7 (65 m)	5 (52 m)
65	.57		
30			.39
10			.37
1	3.59	.88	.51

Table 2-10

Weight Percent Insoluble Residue in lower member

Height Above Base (m)	North<----->South			
	Locality No. (Total Thickness of lower member)			
	8 (20 m)	7 (65 m)	5 (52 m)	1 (17 m)
65		46.5		
10	63.0		86.4	
1	20.5	80.8	86.6	53.9

Such a distribution can be accounted for in several ways, as follows;

1. the relative oxidation rate of organic matter may have decreased to the north, in response to more anoxic conditions associated with deeper parts of the basin
2. terrigenous influx was appreciably less to the north, and hence a sediment dilution effect was reduced
3. primary productivity in surface waters was enhanced in areas to the north.

It is not possible to discriminate clearly the relative effects of these three mechanisms, but the present evidence suggests that sediment dilution was not exclusively responsible for the observed organic carbon distribution. This argument is supported by the following rationale:

1. if one assumes that the preserved thickness of the lower member is directly proportional to original sedimentation rates, then in contrasting locs. 8 and 5, one would anticipate at loc. 8 an organic carbon content of 1.3 %, (in contrast to the observed value of 3.6 %). Similarly, if sediment dilution was the exclusive controlling factor in organic carbon distribution, why does the thickest section (loc. 7) have higher values than are obtained at the thinner type section (loc. 7)?

2. The organic carbon is derived from different biota in southern to northern exposures, as suggested by the palynomorph distribution. Gas chromatography on saturated hydrocarbons of the same six samples, corroborated such a shift in carbon source. Both samples from loc. 7, show odd to even predominance in n-alkane distribution pattern, suggesting a contribution from higher (land) plants, whereas further north at loc. 8, the hydrocarbon geochemistry indicates that, "the organic matter is of algal origin and was deposited in a normal marine environment" (personal communication, T.G. Powell, 1980).

It is unlikely that the carbon budget would have remained constant, given the shift from terrestrial to marine organic sources. The sum contribution of terrestrial and marine organic carbon doubtless varies over the basin, dependent upon source proximity and productivity potential of surface waters. The present distribution of organic carbon must reflect in part these initial differences in flux of organic matter to the sea floor.

2.9.3 Texture and Composition

The mudrocks are appreciably more calcareous north and west of the type section (Table 2-10). The fact that carbonate content of the mudrocks increases northward down the regional paleoslope (i.e. basinward) affirms that the calcite compensation depth was not attained within those portions of the Cape Phillips basin affected by lower member sedimentation. The total equivalent thickness of carbonate (average % carbonate X section thickness) for the northernmost locality, is close to the carbonate equivalent thickness for the type section, 8.3 vs. 6.8 m respectively. This fact suggests that carbonate input was relatively uniform over the basin (given approximate time equivalency for the duration of the lower member at all localities); the relatively low soluble content further south merely reflects increased dilution by terrigenous sediment.

Samples from Cheyne Point (loc. 1, PL. 1f) contain the least insoluble residue (53.9 %), and verge on being classified as carbonates, rather than mudrocks. This locality was clearly removed from the locus of terrigenous sedimentation, a conclusion corroborated by the presence of thin interbeds of wackestone containing normal marine benthic organisms.

The fresh mudrocks have a wide range of colors. However, in general the most calcareous mudrocks (loc. 1) are light olive gray (5Y 6/1), and the most carbonaceous shales (loc. 8 at base) are brownish black (5YR 2/1).

A systematic grain size distribution study was not attempted, owing to the inherent problems of interpreting results obtained by disaggregation, resuspension and deflocculation of shales (see discussion by Picard, 1971). With the exception of the Chéyne Point rocks, fine to very fine lamination is nearly ubiquitous in the lower member. The texturally-dependent weathering property of fissility varies markedly, south to north: in the south conchoidally fracturing claystone and clayshale at the type section; in the north, well-developed paper fissility in organic carbon-rich rocks of the basal 1 m at loc. 8. This increase accompanies not only elevated organic content, but curiously, also extremely high carbonate content (79 %). This observation, contrary to the expected relationship (see Alling, 1945), suggests that of the two variables, % C org. is far more significant than % CaCO₃ in determining the fissility of a given mudrock.

2.9.4 Macrofauna

Random samples from localities north of the type section show no major changes in faunal content in that direction. The principal macrofaunal elements remain the brachiopod species A, small orthoconic nautiloids, ostracoderm fragments, and bivalve species B in approximately that order of abundance. No macrofauna has been identified from loc. 8. Graptolite stipes (PL. 2-4k,1) have been collected from

loc. 7. and identified as Monograptus n.sp. (cf. M.ultimus), (R. Thorsteinsson, personal communication, 1979). A single pterinid bivalve (PL. 2-5c) and a microcrinoid have been collected from loc. 7. An orbiculoid brachiopod and indeterminate small smooth ostracodes have been identified from loc. 6. The paleoecologic significance of these isolated occurrences remains at present, uncertain.

South of the type section, the lower member appears to be largely barren of macrofauna, with only scattered brachiopods (species A), fish fragments, scolecodonts, and at loc. 3, the free cheek of an indeterminate trilobite (PL. 2-4c). On Griffith Island (loc. 1), the lower member contains entirely different biota including Atrypoides erebus, indeterminate smooth spiriferids (Protathyris ?), indeterminate bivalves, Atrypa sp., Hemiargus sp., and shell debris of unknown origin. Generally this fauna is restricted to thin interbeds of wackestone within relatively unfossiliferous calcareous mudrocks. However, A. erebus occurs in small numbers within the mudrocks. This low-diversity association of benthic, epifaunal to semi-infaunal organisms probably represents a pioneering community, particularly tolerant of muddy substrates and turbidity.

2.9.5 Contact relationships

The contact relationships of the lower member with the supra- and subjacent stratigraphic units, are summarized in Table 2-11.

The basal contact is everywhere extremely sharp, defined by a marked change in lithology. Toward the north (loc. 1 - loc. 8), lower member mudrocks overlie Douro lithofacies of increasingly deeper-water aspect. On Griffith Island, the uppermost Douro strata contain carbonate buildups composed of stromatoporoids, favositids, cerioid rugosans, Stylopleura sp., and sponges. The irregular upper growing surfaces of the buildups are immediately overlain by lower member mudrocks. At localities 4,5,6 and 7 the uppermost Douro consists of hardgrounds and banded lime mudstones of deep shelf to basin slope aspect. At loc. 8, the uppermost Douro consists of extremely contorted, thin, lime mudstone beds, interpreted as distal basin-slope deposits.

The upper contact is abrupt where that contact is defined by the top of the subarkose facies (locs. 2,3,4,5). North of loc. 5, the contact is poorly exposed and may be gradational vertically over heights of 2-5 m. On Griffith Island, the upper contact is gradational, with recessively-weathering, medium buff-brown, calcareous shale of the lower member underlying more resistant light green buff, poorly fossiliferous, argillaceous limestone of the Somerset Island Formation (see Paper 5).

In general, the inferred bathymetry of the basal upper member lithofacies involves deeper depositional environments northward.

Table 2-11

Regional contact relationships of the lower member.

Locality	Contact Relationships with Douro Formation			Contact Relationships with upper member, Barlow Inlet Formation		
	Nature of Contact	Underlying Lithology	Gross Facies Type	Nature of Contact	Underlying Lithology	Gross Facies Type
1	Abrupt	Boundstone	Reef	Grada- tional	Argilla- laceous lime mudstone	Tidal flat
2	Abrupt	Slightly rubbly argilla- ceous lime mudstone to wackestone	Open marine shelf	Abrupt	Lime wackestone	Open marine shelf
4&5	Abrupt	Hardground, planar- bedded lime mudstone, shelly rudstone, laminated calcarenites	Deep water open shelf	Abrupt	Lime wackestone	Open marine shelf
6	Abrupt	Hardground, planar- bedded lime mudstone	Deep water open shelf to basin slope	Not observed	Planar- bedded lime mudstone	Deep water open shelf to basin slope
7	Abrupt	Hardground, planar- bedded lime mudstone	Deep water open shelf to basin slope	Grada- tional	Planar- bedded lime mudstone	Deep water open shelf to basin slope
8	Abrupt	Contorted lime mudstone, planar- bedded lime mudstone	Basin slope	Not observed	Planar- bedded lime mudstone	Deep water open shelf to basin slope

2.10 PALEOENVIRONMENTAL SYNTHESIS

2.10.1 The Deltaic Model

In general the lower member is a coarsening-upward siliciclastic sequence, that commences with fine-grained dark mudrock containing an impoverished fauna, and terminates in well-sorted to texturally-inverted bimodal fine sandstone, interpreted as a foreshore (beach) facies. Previously, in the absence of more critical data, this simple succession of lithofacies was interpreted as representing a sheltered, shallow, nearshore environment (McGregor and Narbonne, 1978); or a barrier-bar lagoon (Packard and Dixon, 1982). More comprehensive analysis now suggests that the lower member represents a prograding, tide-dominated and wave-influenced delta with both prodelta and delta front beach lithofacies preserved.

Evidence and arguments that serve to eliminate several possible alternative interpretations are listed in Table 2-12, without further discussion. Data supportive of a deltaic interpretation are enumerated below:

1. Thickness variation (FIG. 2-2b) of the lower member is entirely consistent with the typical spoon-shaped wedge of sediments associated with deltas (see Dailly, 1975).

Rejection criteria for alternative depositional models.

CRITERIA AGAINST BARRIER-BAR LAGOON

-
1. lack of associated tidal flat, marsh or wash-over fan facies. The lower member could be most favourably compared with a regressive barrier-bar lagoon sequence, but would require interpreting the uppermost Douro as a nearshore facies, an interpretation inconsistent with the data.
 2. distribution and proportion of subarkose to mudrock facies is atypical of barrier-bar lagoon deposits. Interpretable barrier sands are completely missing north of loc 5.
 3. dimensions (areal extent and stratigraphic thickness) of the deposit are not consistent with known Recent and ancient barrier-bar lagoon sequences.
 4. very uniform upward-coarsening of mudrock facies is unlikely for lagoonal deposits.
 5. the abrupt transition between the Douro Formation and the lower member is inconsistent with progressive barrier and lagoon development.
 6. the underlying lithologies (see Table 2-11) are variable, representative of differing depositional environments.

CRITERIA AGAINST NON-DELTAIC ANOXIC CLINOTHEM

-
1. Available age data do not suggest a condensed, 'starved basin' sequence with low sedimentation rates for the lower member.
 2. regional thickness variation in the lower member is large with abrupt pinchouts, that suggest a localized source for the siliciclastic detritus in the basin. Such thickness variability and distribution would not be anticipated with a non-deltaic anoxic clinothem.
 3. the subarkose facies cannot easily be incorporated into anoxic clinothem model.

2. Mudrocks of the lower member blanket Douro lithofacies of progressively deeper water aspect, from south to north, strongly suggesting that the siliciclastic sediments were laid down on a pre-existing paleoslope descending to the north. The fine-grained terrigenous sediments were distributed in environments ranging from nearshore (possibly intertidal), to reefal, to deep offshore. Lack of bathymetric control on mudrock lithofacies distribution points to allogenic influences on derivation and dispersal of sediment. A deltaic model aptly accounts for the widespread facies-crossing distribution of these fine-grained terrigenous sediments.
3. Mixing of marine and terrestrial palynomorphs with progressive reduction of terrestrial forms northward, is most compatible with a deltaic model.
4. The coarsening-upward nature of the sequence, the thin lamination of silts and clays, the progressive upward change from color to textural lamination, and the fact that bioturbation is restricted to certain zones, are all features that characterize the prodelta environment (see Reineck and Singh, 1982, p.333-334).
5. The trend towards greater carbonate content to the west, north, and east (E-45 well on Devon Island)

suggests increasing dilution of terrigenous clastic material radially away from some point source (e.g. river mouth), a distribution typical of deltaic sediments.

6. All 50 m of the lower member, and at least 80 m of the basal portion of the upper member of the Barlow Inlet Formation at loc. 4, can be assigned to the Pedavis latialata conodont Zone (Uyeno, 1977). The absence of zonal boundaries within the lower member suggests continuous deposition of siliciclastics, a conclusion supported by the lack of physical evidence for either erosional or non-depositional events. The absence of biostratigraphic breaks effectively argues against a 'starved basin (or basin slope)' model, where a condensed faunal zonation would be anticipated. The fact that the lower member consists of a single, uninterrupted, coarsening-upward cycle, strengthens the argument for relatively continuous detrital sedimentation, and is consistent with a prograding delta model.
7. The ubiquitous abrupt change from carbonates of the uppermost Douro Formation to siliciclastic mudrock of the lower member (<20 % insoluble residue to >85-% at the type section), strongly suggests rather precipitous introduction of large volumes of terrigenous material to the Jones-Lancaster Basin.

The well-defined but irregular upper surfaces of buildups at the top of the Douro Formation on Griffith Island, are blanketed by clastic detritus (PL. 2-1h), possibly indicating that reef growth was terminated by increased turbidity. Lack of adaptation of organisms to these conditions on the buildups on Griffith Island, and the absence of marly transition beds elsewhere in sections containing the lower contact, both attest to the apparently rapid influx of continentally-derived material. The Douro-lower member transition is readily explained by advance of a northward prograding marine clastic wedge (i.e. a delta).

2.10.2 Discussion

2.10.2.1 Basin Anoxia

Although most of the lower member mudrock facies is not black shale sensu stricto, it is very dark gray shale of a 'black shale' environmental aspect. Many examples of carbonaceous shale and dark gray shales discussed in the geological literature are closely related, showing intimate stratigraphic and presumed paleogeographic associations (see Heckel, 1977; Baesemann, 1975). For example, the Pennsylvanian dark gray shale facies of the midcontinent U.S. is considered by Malinky and Mapes (1981) to represent offshore deposition in an oxygen-depleted environment, near

the lower limit of the dysaerobic zone, whereas the associated black shales are interpreted as having been deposited beneath anoxic water. By analogy, the lower member mudrocks could represent deltaic sedimentation into a partially (variable with depth) anoxic basin.

It is difficult to evaluate the degree, if any, of oxygen depletion within waters overlying the prodelta muds of the lower member. Although evidence suggests the existence of a stratified basin during deposition of the underlying Douro Formation (see The Status Quo: Pre-epirogenic sedimentation, below), the dysaerobic zone clearly existed only at considerable depth, given the abundance and diversity of epi- and infaunal organisms in units of the Douro interpreted as accumulating at shelf depths approaching storm wave base (see Narbonne, 1981).

Certainly depths near storm wave base or shallower were achieved during deposition of the uppermost Douro beds at Goodsir Creek, and at present there is no conclusive and independent evidence to suggest that the water/sediment interface ever subsided to greater depths during deposition of the lower member mudrocks. It can be argued that the presence of diagenetic pyrite, the dark color of the mudrocks, and the preservation of fine lamination require an anoxic environment, but as has been pointed out by numerous authors (e.g. Curtis, 1980, p.192), anoxic conditions can exist just below a sediment/water interface, even where the

water mass is oxygenated. This is generally explained by the limited efficiency of downward diffusion of dissolved oxygen into fine-grained sediments. In such an instance the rapid isolation of a sediment layer from basin waters is required to prevent significant oxidation of organic matter, and this is most easily achieved by rapid sedimentation, such as would be anticipated in a deltaic environment. Conceivably, the turbid, hyposaline water conditions associated with river debouchment into a marine basin, would, in a manner quite independent from relative oxygen levels in basin waters, effectively exclude burrowing infauna as well as shelly epifaunal benthos by affecting salinity, diluting food resources and interfering with normal metabolic activity (see Heckel, 1972, p.259-261; Lapointe, 1968, p.99). Thus the observed sediment and faunal characteristics of the lower member mudrocks could be explained solely in terms of rates of deltaic sedimentation.

Other factors, however, militate against depositional rates unilaterally determining the overall character of the lower member. At least partial or intermittent anoxia of the Jones-Lancaster basin (or parts thereof), in late Ludlow time, is suggested by the following:

1. The phosphorite layer immediately underlying the lower member, if analogous to Recent deposits, was formed on a substrate exposed to oxygen depleted waters (Bentor, 1980). The absence of distinct

phosphorite layers (such as are typically associated with black shales -- e.g. Phosphoria Fm.) within the lower member mudrocks, can be attributed to dilution effects of rapid sedimentation (e.g. Giresse, 1980). As indicated by Jarvis (1980, p.188), conditions of total anoxia, and normal oxygenation both inhibit phosphatogenesis; dysaerobic, or perhaps intermittently anoxic conditions favour formation of phosphorite.

2. The distinctive light/dark color banding (reflecting ferrous/ferric iron ratios) of the uppermost Douro also suggests alternating anoxic/oxic conditions at the sediment/water interface, or within the diagenetic near-surface environment.
3. As discussed previously, weight per cent organic carbon increases substantially to the north, disproportionately to sedimentation rates inferred from section thickness. Given that the contribution of land-derived organic carbon appears to diminish northward, and assuming that marine sources do not dramatically offset this loss, then an alternate process must account for the increased enrichment of organic carbon at loc. 8. Despite increased residence time both in the water column and on the sea floor (due to decreased sediment influx), the organic matter at loc. 8 has been little oxidized.

This fact suggests that deeper parts of the basin to the north were increasingly anoxic and hence, that basin waters were stratified.

2.10.2.2 Ancient Examples

Although poorly fossiliferous dark gray to black shales are more typically associated with 'starved basin' facies, and hence periods of relatively low sediment influx, several dark shale sequences have been interpreted as representing periods of increased deposition of terrigenous detritus, associated with deltas. Both Walls (1973) and Lewis and Schwietering (1971) described, respectively, Devonian and Carboniferous examples of offshore deltaic black shale facies that accumulated in epeiric euxinic seas. Heckel (1972) provided a paleoenvironmental interpretation for upper Pennsylvanian, olive gray to gray, sparsely fossiliferous shales of the midcontinental U.S.A. The association of these shales with marine limestones, the preserved primary lamination, and the restricted fauna of small thin-shelled bivalves and gastropods, and rarer articulate brachiopods, orbiculoid brachiopods, orthoconic cephalopods, comularids and echinoderm debris, all point to a "restricted marine environment undergoing rapid sedimentation and probably fluctuating ... salinities" (Heckel, 1972, p.260). The same author (op.cit.) noted similarities to Recent prodelta deposits of the Gulf of

Mexico, and chose the modern restricted sea of the Gulf of Chihli (N.W. China), which has a very high sediment influx, as an overall close environmental analogy.

To the author's knowledge, no example of deltaic black to dark gray prodelta muds that were deposited in an ancient mobile belt sequence, has been described in the literature. Nonetheless the intra-cratonic examples cited above, provide useful parallels to the lower member succession.

2.10.2.3 Modern Prodelta Muds

Kolb and Kaufman (1967), in their description of prodelta clays of southeast Louisiana, described several generalized stages for prograding prodeltas. The earliest stage is one of relatively slow deposition, characterized by higher faunal content and the presence of faintly-marked (color) laminae. The second stage is influenced by higher sedimentation rates, with the clay-sized sediment exhibiting no laminae. The shallowest portion of the prodelta exhibits finely spaced silt and clay (textural) laminae. This sequence closely approximates the lower member mudrock succession. However, in modern prodelta sediments, lamination and bedding are commonly disrupted by post-depositional faulting and distortion due to loading. Similar features were not observed in the lower member mudrocks, suggesting that either absence of good exposure precluded their identification, or sedimentation proceeded at such a pace, that excess pore pressure did not develop.

2.10.2.4 Delta Type

In Ludlow time, in the area encompassing the Jones-Lancaster basin, the climate was arid to semi-arid (Miall and Gibling, 1978). In part coeval with the deposition of the lower member, there was development of extensive tidal flats on northern Somerset and southern Cornwallis Islands (Gibling, 1979; Paper 3), implying a significant tidal influence within the basin.

Wind-generated onshore wave activity is difficult to assess, but limited evidence, as follows, would suggest that it was not excessive:

1. in the Lochkovian, paleocurrent data derived from possible aeolian foresets, indicate a northwestward wind direction (Miall and Gibling, 1978), an interpretation supported by paleolatitude determinations (Lapointe and Dankers, 1982) that place Somerset Island in the zone of prevailing south-easterlies. If this wind circulation pattern is correct, and existed as well in the Ludlovian, then predominantly offshore winds would have reduced the influence of waves on the delta front environment.
2. beach shoreline facies are not recorded from approximately time-equivalent facies elsewhere in the Jones-Lancaster basin, suggesting that the presence of such facies in the lower member is a

function of sand availability, and not the basin-wide winnowing capabilities of onshore waves.

3. if the Jones Lancaster basin was indeed a back reef basin, the northerly located barrier (reefs -- Trettin and Balkwill, 1978; miogeanticlinal ridge -- Drummond, 1973) would have damped much of the energy of southerly-directed swells.

Certainly it is not unusual to find beach deposits associated with modern deltas considered to be of low-wave energy type (e.g. Chandeleur barrier beaches, Mississippi delta).

Thus, the lower member appears to have been a wave-influenced but not wave-dominated delta. The absence of bar-finger sands, or indeed any extensive delta front deposits of coarse silt to sand-sized detritus, appears to preclude the lower member being either a river-dominated delta, or any delta with a high bed load/suspended load ratio. In conclusion, considering the inferred tidal influence within the Jones-Lancaster basin, the lower member appears to represent a tide-dominated, wave-influenced, high suspended load delta, prograding into a tectonically unstable restricted basin under the influence of an arid to semi-arid climate. Subsequent marine transgression may have removed and reworked some of the delta's topset deposits, but evidence for such reworking is lacking. No modern analog in the literature is defined by precisely the

characteristics given above. Of the examples outlined by Coleman and Wright (1975), the Burdekin delta in Australia is perhaps most similar. It occurs in a semi-arid climate, and the receiving basin has intermediate wave energy and high tide range. Overall the delta exhibits a net sand distribution comparable to the anticipated three-dimensional distribution of the lower member sheet sands. However the Burdekin delta progrades into a stable, unrestricted basin, and has a high bed load/suspended load ratio.

2.11 REGIONAL TECTONISM AND SEDIMENT PROVENANCE

Within the Upper Silurian exposures in the study area, the abrupt appearance of concentrated siliciclastic detritus (the lower member) within a stratigraphic succession that is otherwise carbonate of open marine aspect, clearly records a significant episode of proximally-located continental erosion. On eastern Cornwallis Island, this period of predominantly detrital sedimentation was brief, occupying as it does but a portion of the latialata Zone. Nonetheless it represents in the sedimentary record, the vanguard of epeirogenic tectonism that was responsible for intensely modifying and ultimately exhuming a significant portion of the Jones-Lancaster Basin during Late Silurian and Early Devonian time.

2.11.1 Tectono-Sedimentary Framework

The pre-eminent physiographic feature within the study area (FIG. 2-10) coincides with a large north-south trending basement high, known as the Boothia Horst. Episodic uplift of the horst has strongly influenced both the nature and disposition of sedimentary facies in this region since the Proterozoic (Kerr and Christie, 1965). The Boothia Horst consists of a series of subparallel, flat-topped blocks of crystalline basement bounded by steep north-south trending faults.

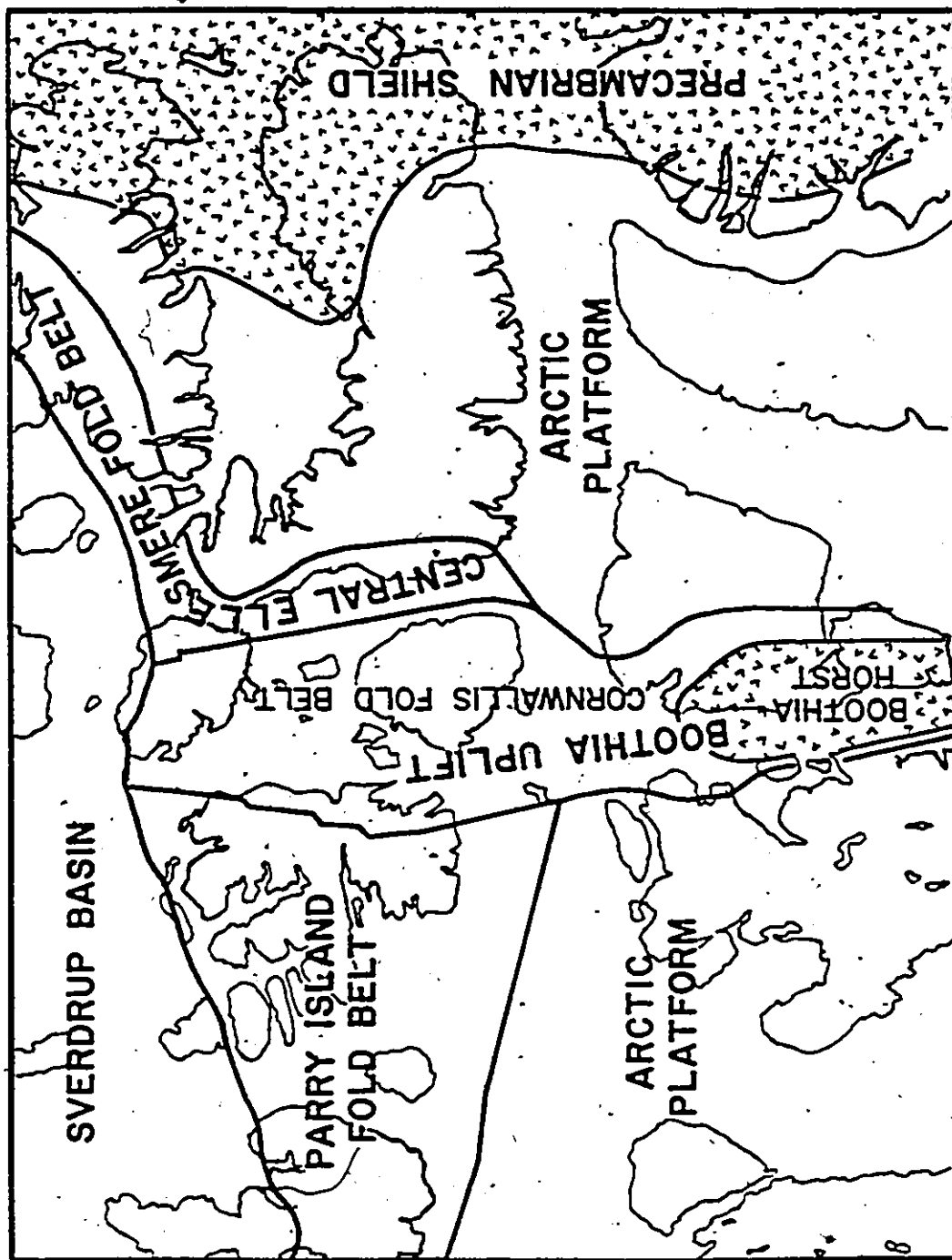


FIG. 2-10 Phanerozoic structural provinces of central arctic archipelago.

Northward the horst plunges beneath a younger sedimentary cover that was deformed by upward movement of the basement blocks at depth. The deformed sedimentary suprastructure composes the Cornwallis Fold Belt, a steep-sided northerly plunging anticlinorium (FIG. 2-11). Together, the Boothia Horst and Cornwallis Fold Belt form part of the Boothia Uplift (Kerr, 1977).³ The principal structural elements of the Boothia Uplift, and the relationship between tectonism and sedimentation in the vicinity of the Uplift, have been described in numerous papers including Thorsteinsson and Tozer (1960), Kerr and Christie (1965), Brown et al. (1969), Miall (1970a, 1970b), Dineley (1971), Gibling and Narbonne (1978), Gibling (1978), Miall and Gibling (1979), Thorsteinsson (1980), Muir and Rust (1983).

The most thorough synthesis to date on the structure and tectonic history of the Boothia Uplift is that by Kerr (1977). In this paper, Kerr elucidated the process considered to be responsible for the structure of the Cornwallis Fold Belt, and outlined the principal periods of tectonism involved in its development.

³ The more informal term, 'Boothia Arch', is used in this paper to refer to both the crystalline horst and the structurally affected sedimentary strata in the vicinity of Peel Sound and the Boothia Peninsula (but not north of Barrow Strait).

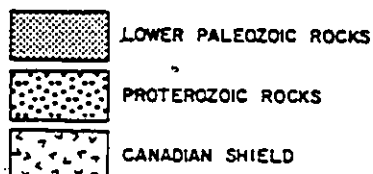
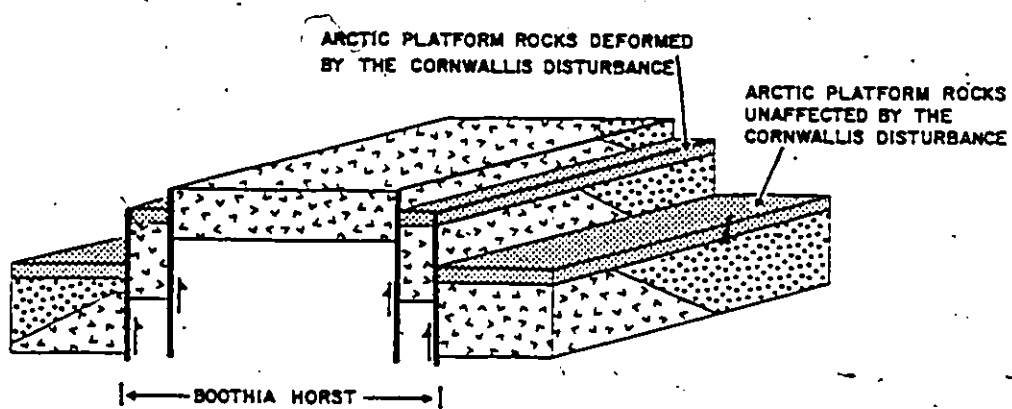
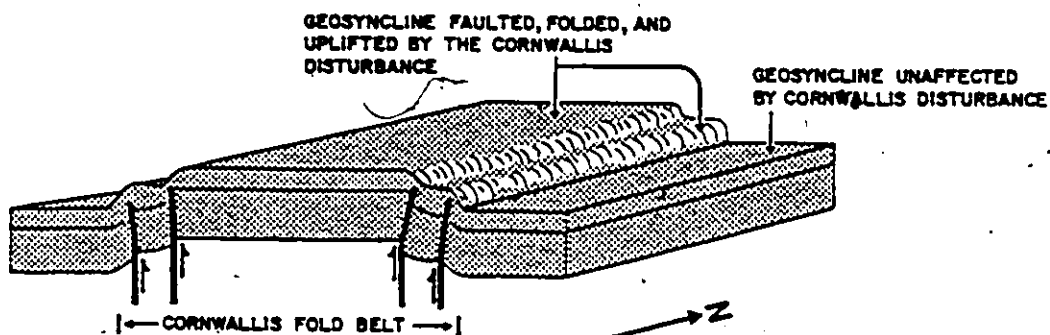
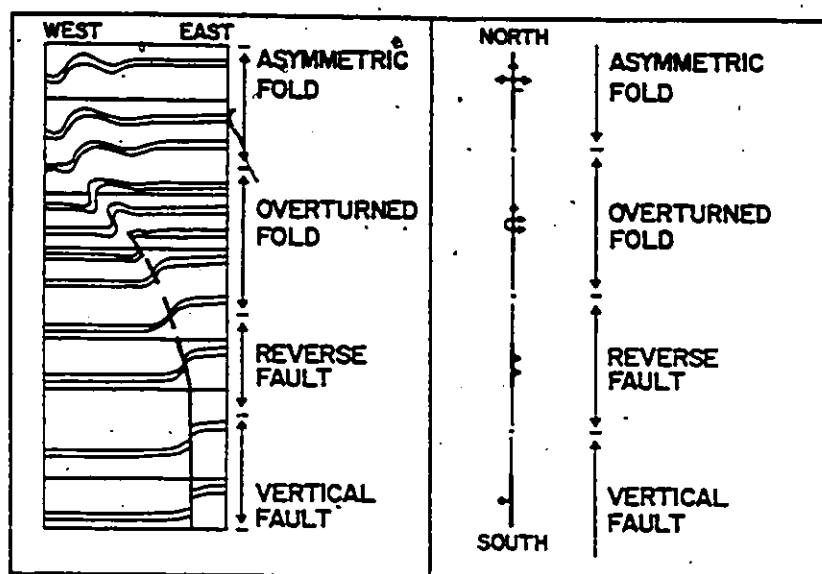


FIG. 2-11 Structure of the Boothia Uplift (after Kerr, 1977, 1981)

Kerr named this spasmodic epeirogenic activity, the Cornwallis Disturbance and identified four discrete pulses of uplift all confined to Siluro-Devonian time. FIG. 2-12 indicates the first three pulses as envisaged by Kerr (op.cit.), here labelled 1, 3 and 4. Miall and Kerr (1980) subsequently suggested the possibility of a fifth pulse of Late Ordovician age. The timing of these pulses has been determined on the basis of 1) the oldest datable beds overlying unconformities, and/or 2) dating of syntectonic sedimentary deposits.

Of significance to the present study, Thorsteinsson (1980) has established that coarse clastic sedimentation (lower member of the Peel Sound Formation on Prince of Wales Island) began in latest Ludlow time. As indicated by Thorsteinsson (1980, p.32), this represents the first evidence of late Ludlovian uplift of the arch, and clearly adds to the original sequence of tectonic pulses of the Cornwallis Disturbance outlined by Kerr (1977, table 1). Thorsteinsson (op.cit., p.7) suggested that the lower member of the Barlow Inlet Formation, being correlative with late Ludlovian basal beds of the Peel Sound Formation, represents "...the distal end of a tongue of detritus derived from uplands developed on the Boothia Uplift south of Barrow Strait." Packard and Dixon (1981) have argued that the lower member reflects a hitherto unrecognized pulse of the Cornwallis Disturbance.

PULSES OF CORNWALLIS DISTURBANCE		DEVONIAN										SILURIAN									
SYSTEM	STAGE	DEVONIAN										SILURIAN									
DEVONIAN	DALEIAN																				
	ZUCKOVIAN																				
	PRACIAN																				
	LOCKKOVIAN																				
	LOCKKOVIAN																				
	LOCKKOVIAN																				
	LOCKKOVIAN																				
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SILURIAN	PRIDOLIAN																				
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FIG. 2-12 Correlation of Siluro-Devonian formations of central arctic archipelago (after Thorsteinsson, 1980). Pulses of Cornwallis Disturbance modified after Kerr (1977). Symbolic representation of intensity and duration is only approximate.

Regardless of the ultimate detrital source area for the lower member, the synchronous initiation of clastic sedimentation both north and south of Barrow Strait can reasonably be related to a common late Ludlovian epeirogenic event of the Cornwallis Disturbance.

2.11.2 Provenance of Detritus

The simplest and consequently most compelling argument for the provenance of lower member detritus, is that newly emergent continental terrane, hereafter referred to as the Boothia Axial Landmass (or B.A.L.; see FIG. 2-13b), that resulted from uplift during the late Ludlovian pulse of the Cornwallis Disturbance, served as a source for the deltaic muds and sands now exposed north of Barrow Strait. However, if this is to be accepted, the following factors require consideration and explanation:

1. The Griffith Island section is anomalous considering its presumed proximity to the B.A.L.
2. No appropriate fluviatile sediments are known to be associated with these deltaic sediments in the region.
3. There is no clear evidence of sufficiently early exhumation of crystalline source rocks of the Boothia Horst.

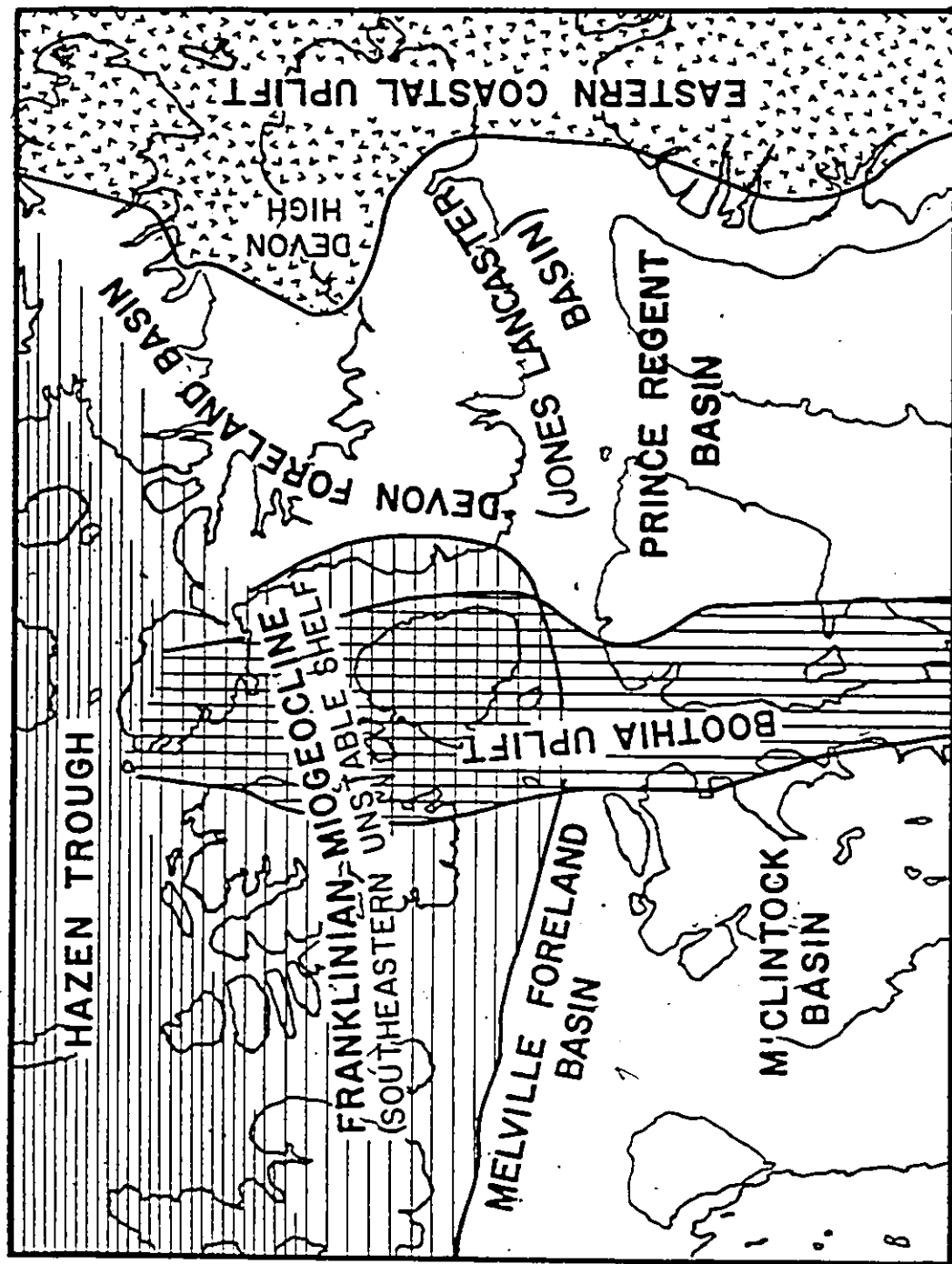


FIG. 2-13a Tectono-depositional elements of central arctic archipelago.

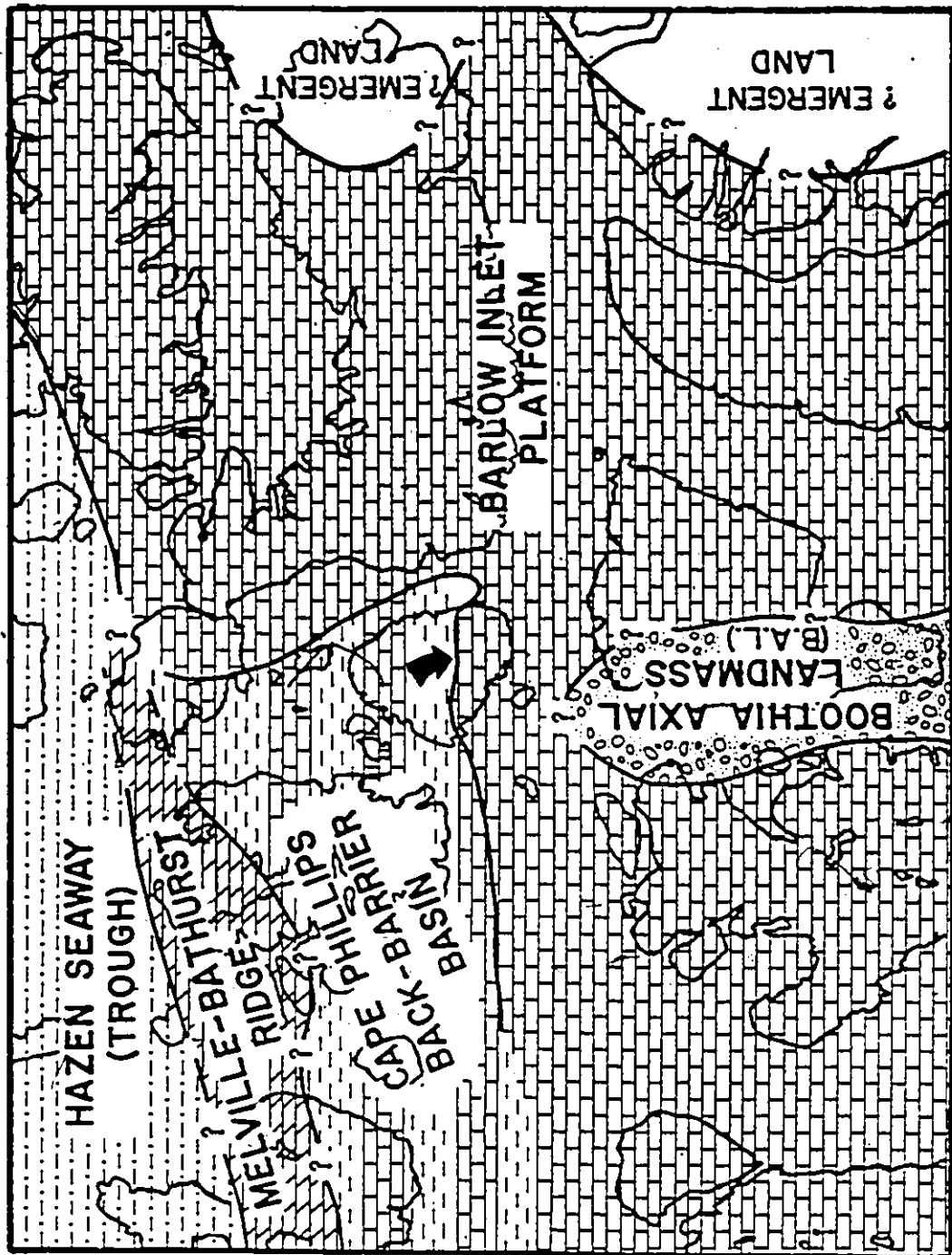


FIG. 2-13b Siluro-Devonian paleogeographical elements of central arctic archipelago.
 ➡ CASBLFC (Cape Airy-Snowblind Bay line of facies change)

4.. The mineral suite in the deltaic rocks is inconsistent with derivation from known Boothia Horst rocks.

5. Other potential source areas are known in the region. Areas of thin Phanerozoic cover over basement (hence high magnetic flux) are widespread in the Arctic, and represent possible source areas (see FIG. 2-14).

These points are discussed in detail in Appendix A.



FIG. 2-14 Magnetic anomaly map of central arctic archipelago. From McGrath and Fraser, 1981, Can. Geol. Surv. Map no. 1512a.

2.11.2.1 Discussion

For reasons cited previously (and detailed in Appendix A), the subarkose facies is interpreted as being derived from erosion of a crystalline source. Although the type and proportion of detrital minerals in the lower member contrasts with Boothia Horst bulk mineralogy, no other proximally-located and presently exposed shield area represents a more suitable source. The Churchill Province is a heterolithic complex, and has been mapped in the archipelago only on a reconnaissance scale. The lower member may well have resulted from erosion of a relatively local but hitherto unreported body of more related mineralogy. Muscovite is commonly reported in Upper Silurian clastic strata (e.g. Gibling, 1979; Jones and Dixon, 1981; Savelle, 1978) deposited in the vicinity of the Boothia Arch. Sericite is reported as a common accessory mineral in Proterozoic and Paleozoic arenaceous rocks elsewhere in the Jones-Lancaster Basin (see Lemon and Blackadar, 1963). It is possible that the muscovite in the lower member subarkose facies is in fact sericite, but its size (over 200 microns) and the fresh appearance of the feldspars, does not support this contention. Alternatively, but less likely, the muscovite could be leached biotite or authigenic illite-mica.

There is positive evidence that the Boothia Horst was breached by erosion within early Pridoli time, and it is

entirely reasonable to assume that at least locally the basement was being eroded by late Ludlovian time.

The existence of a major fluvial system that flowed north on the B.A.L. to feed the lower member delta is required in late Ludlow time (latialata Zone). The studied Ludlovian portion of the clastic wedge preserved on either side of the Boothia Horst, does not indicate extensively developed fluvial deposits. Nonetheless, the outcrops studied are relatively few and scattered. Furthermore post-Ludlovian uplift undoubtedly removed much of the syntectonic record initially developed closer to the axis of the presently exposed Boothia Horst. The seemingly poor development of fluvial facies may simply be an artifact of preservation.

The principal remaining impediment to considering the Boothia Uplift as the source for the deltaic sediments, is the nature of the lower member on Griffith Island. Three arguments have already been advanced (see Appendix A) to reconcile the proximity of the Griffith Island locality to the Boothia Axial Landmass, with the paucity of the lower member record preserved there. A fourth, significantly more speculative and potentially controversial, solution exists: a post-Devonian relative lateral movement of the areas north and south of Barrow Strait.

In light of the substantiation of sea-floor spreading in Baffin Bay (Keen et al., 1972a), Pitman (in Burke and Dewey, 1973) suggested that Lancaster sound is the site of a

transform fault. This was refuted by Kerr (1980), and certainly Keen et al. (1972b) cited no evidence to suggest horizontal motion on the Lancaster graben or half-graben. Nonetheless, dextral strike-slip of 60 to 100km along the Parry Channel fault (or extension thereof) in post-Devonian time would account for the apparent misalignment of clastic facies north and south of Barrow Strait, and make a genetic link between uplift of the arch and lower member sedimentation more tenable.

Any significant post-Devonian strike-slip movement along the Parry Channel fault or similar structure, should have clearly offset the Boothia Uplift. Just such a deflection of approximately 75 km (FIG. 2-15b) is shown on the eastern margin but a complementary offset of the western margin is not indicated. Drummond (1973, p.447) figured a major fault in Lancaster Sound with significant strike-slip movement indicated by offset of both isopach contours and a prominent north-south fault trace (FIG. 2-15c). These offsets however indicate sinistral movement on the fault and if correct would serve further to separate the Boothia arch source from the lower member depocenter.

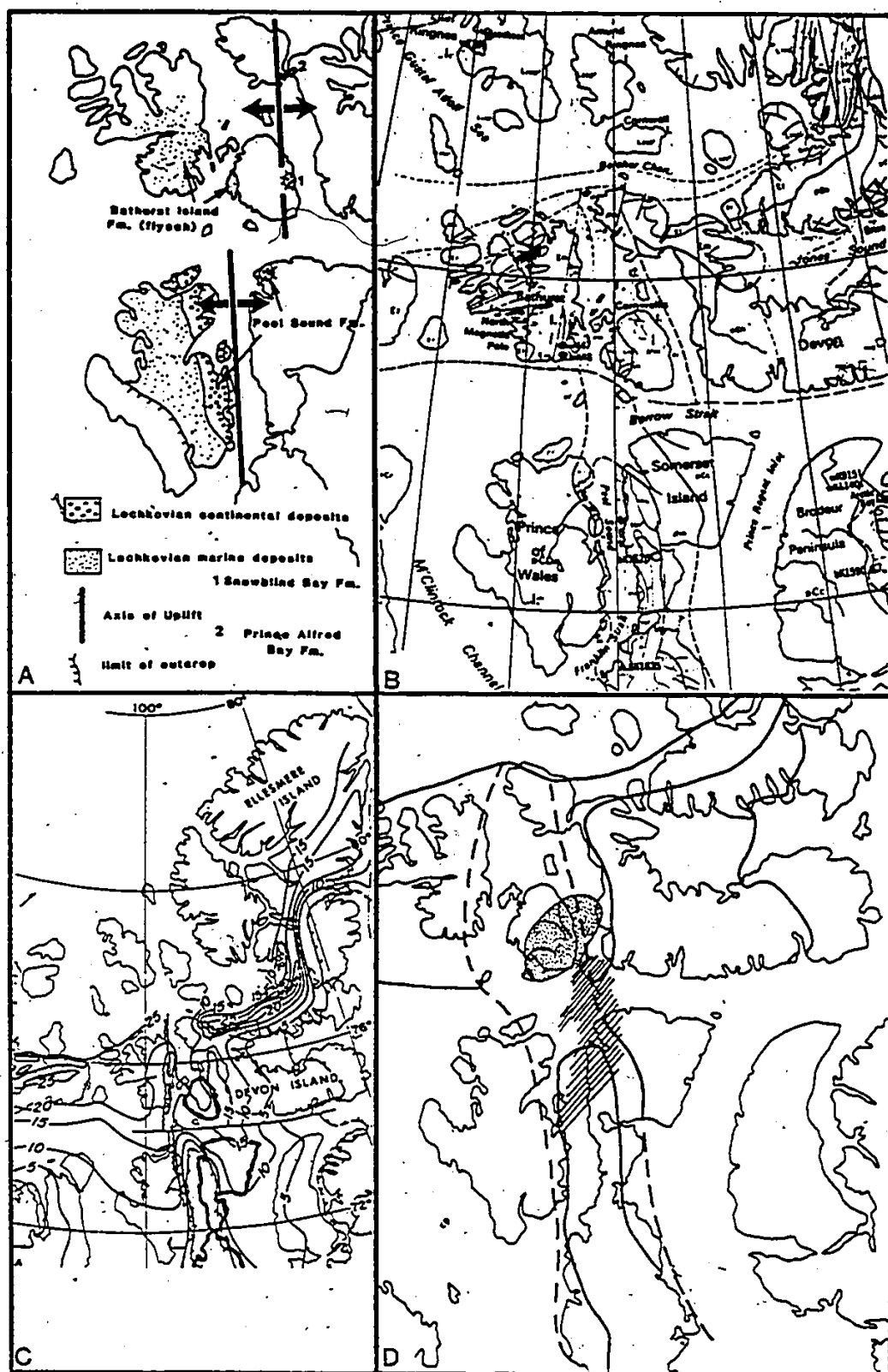


FIG. 2-15 Evidence relating to possible strike-slip motion along Parry Channel: A. Possible displacement of axis of uplift (base map from Thorsteinsson, 1980); B. Tectonic map from Bouglas (1970); C. Sinistral movement along Parry Channel suggested by Drummond (1973). Isopachs in thousands of feet; D. Realigned (pre-faulting) position of islands. B. A.L. shown as cross-hatched area, disposition of lower member shown as dot pattern.

Early to late Lochkovian sediment distribution (FIG. 2-15a) suggests that the axis of uplift north of Barrow Strait was offset to the east. The position of gravity and magnetic anomalies on either side of Barrow Strait argue neither clearly for, nor clearly against offset of basement structures (Douglas, 1970, Bouger Gravity Anomaly Map; Magnetic Anomaly Map of Arctic Canada, McGrath and Fraser, 1981, map 1215A). The writer is unaware of any lower Paleozoic structures that transect the Parry Channel - Barrow Strait - Lancaster Sound line, without clearly showing change in disposition that could be due to post-Devonian strike-slip movement.

The hypothetical reconstruction (FIG. 2-15d), correcting for 100 km of dextral strike-slip movement, not only gives a more plausible facies juxtaposition surrounding the arch, but additionally provides a nearer source (i.e. the Devon Island High) for the siliciclastic detritus in the Leopold Formation on northeast Somerset Island.

In conclusion, the source for the deltaic sediments of the lower member remains enigmatic. The Boothia Uplift, because of its relative proximity and its known coincidental uplift in Late Silurian time, is the most likely source. No evidence definitely links late Ludlovian clastic lithofacies north and south of Barrow Strait, nor are there unassailable arguments against such a linkage. Reliance on the Boothia Uplift as the ultimate source for the lower member, may

reflect the convenience of documented near-contemporaneity of clastic sequences, but this does not exclude the possibility of an alternate, less evident source area not yet recognised.

2.11.3 Marine Record of Late Ludlovian Tectonism

It is inconceivable that the epeirogenic pulse responsible for initial uplift of the Boothia Horst in late Ludlow time, confined its effects to the B.A.L. south of Barrow Strait. Although less manifest than in correlative continental deposits, a sensitive, if subtle, record of complementary tectonic-depositional events can be preserved in synorogenic marine sequences. The usefulness of marine successions in this respect is reduced by the efficiency of marine dispersive agents (waves, currents, mass transport), the susceptibility of such deposits to physical and biogenic reworking, and the dilution of marine clastic units by indigenous carbonate sediments. However, if the cause and effect relationship between tectonism and subtidal shelf sedimentation is somewhat obscured by normal marine processes, the record of the relationship is at least preserved in sequences generally more continuous and more readily dateable than sequences of continental origin.

2.11.3.1 The Status Quo: Pre-epeirogenic marine sedimentation

In the area encompassing Somerset and Cornwallis Islands, pre-epeirogenic sedimentation followed a pattern established in Richmondian (Late Ordovician) time. To the south of Barrow Strait (FIG. 2-13a) lay the stable cratonic platform, which received a substantial, if comparatively thinner, sedimentary record composed predominantly of carbonates.

Separating the cratonic platform from the geocline to the north was an E-W trending broad flexure or hingeline, that is traditionally (and conveniently) drawn within the area presently occupied by Barrow Strait. To the north, the geocline (nee Franklinian Miogeosyncline) accumulated a time-equivalent, predominantly carbonate sequence, 2.4-3.9X thicker than the cratonic sequence.

Within the geocline lay two major depositional realms, a northern deeper-water basin with graptolitic facies, and a southern, shallower, carbonate shelf with shelly facies. Separating these realms is the E-W trending Cape Airy - Snowblind Bay line of facies change. A structural high or ?miogeanticlinal ridge formed the northern margin of the deeper-water basin (Cape Phillips Back-Barrier Basin; FIG. 2-13b), (Drummond, 1973). Presumably this ridge (Melville-Bathurst Ridge) was a synsedimentary topographic high during the Late Silurian, for the Cape Phillips basinal sediments thin northward, and reefs were developed in a linear trend coincident with the ridge. This reef trend "...extended at least intermittently and perhaps continuously from northern Melville Island ... to Ellesmere Island ... to northern Greenland ... a distance of about 1700 km " (Kerr, 1981, p.122). The impact of this ridge on regional sedimentation has yet to be fully evaluated, but it is reasonable to assume that its relative elevation above the contemporaneous Cape Phillips Back-Barrier Basin

sea-floor could well have influenced basin circulation, and hence basin-water stratification and the influx of northerly-derived fine clastic sediment. North of the miogeanticlinal ridge lay the deep elongate Hazen Trough (Trettin, 1972).

The indicated position of the Cape Airy - Snowblind Bay line of facies change (FIG. 2-1) is approximate, and in some instances the paleogeographic demarcation between shelf and basin is inappropriate. Such is the case for the Douro Formation, on which the lower member lies conformably. For the greater part of its depositional history, the Douro Formation apparently was laid down on a carbonate 'ramp', as defined by Ahr (1973). Ramps have no pronounced shelf-slope break, and thus the available hydrodynamic energy is gradually dissipated over a larger, more gently inclined, shoaling surface, with highest energy facies generally closest to the strandline.

From studies conducted by the author, and from detailed research on the Douro by Narbonne (1981), it is evident that no extensive barrier buildup or oolite shoal (i.e. shelf-margin facies) has yet been recorded from stratigraphic intervals considered to be of transitional shelf-slope aspect in the study area. Instead, the lithofacies change gradually (and locally abruptly) from unbioturbated, unfossiliferous planar-bedded lime mudstone, to slightly rubbly, bioturbated but unfossiliferous

limestone, to fossiliferous rubbly argillaceous limestone (see Packard and Dixon, 1983; FIG. 2-16, section d). This facies transition is similar to Byers' (1974) generalized scheme for basins with a stratified water column. By inference, vertical mixing was limited in the late Ludlow Cape Phillips Back-Barrier Basin, leading to increasingly anoxic water conditions with depth. This corroborates Kerr's (1967) interpretation of anaerobic, stagnant basin conditions for the Cape Phillips Formation on Bathurst Island.

Two arguments against the restricted-basin model can be advanced: 1) the relatively light color of the Douro deeper-water carbonates suggests limited preservation of organic carbon within the sediments (.26 % from a single representative sample). The light color, however, could simply reflect the high sedimentation rates envisaged for this facies, as discussed in Part I. 2) the biofacies pattern could simply reflect the limited radiation of then extant burrowing organisms into deeper-water environments (see Seilacher, 1967).

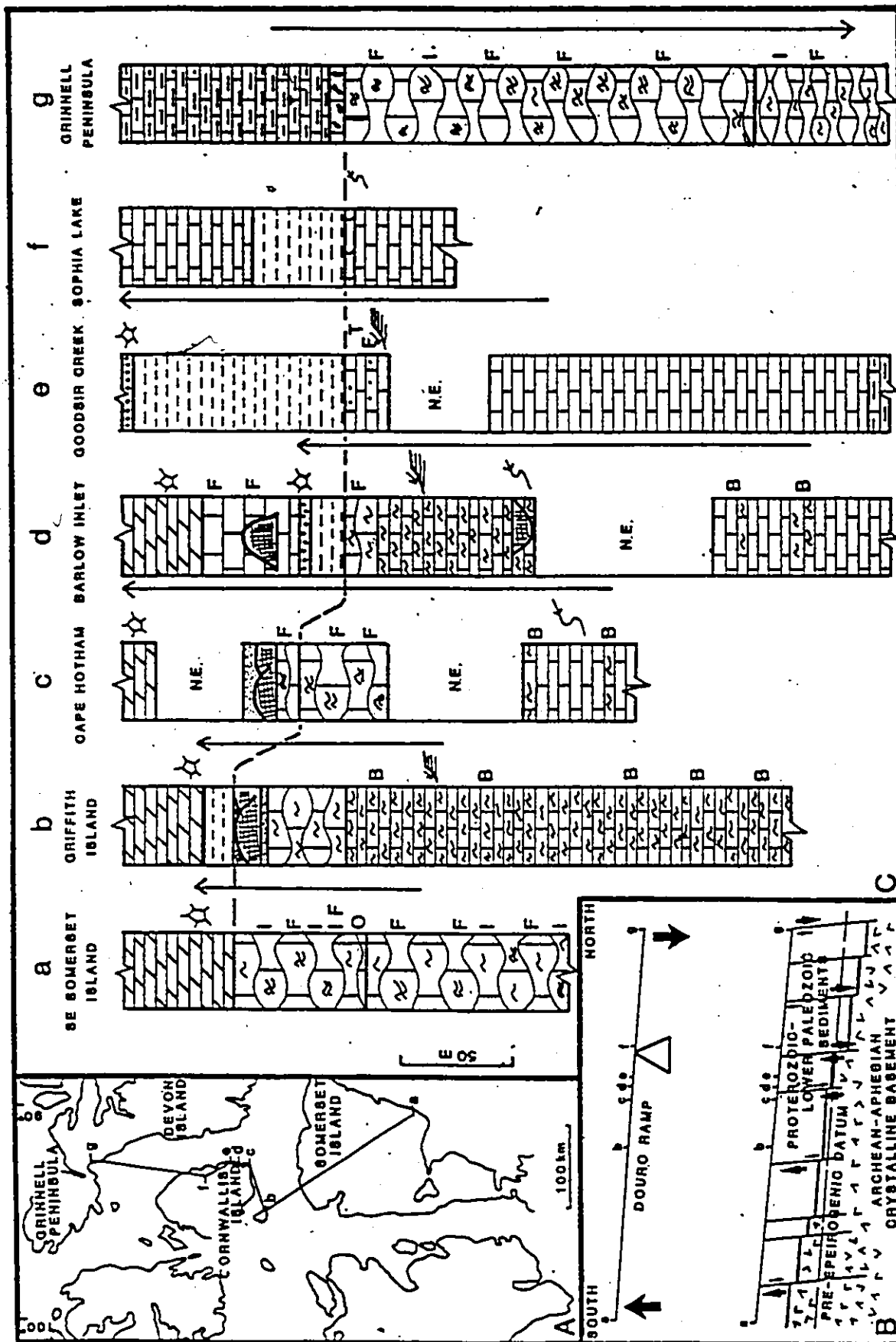


FIG. 2-16 Comparison of upper Douro successions in the vicinity of the Boothia Uplift.

Legend, fig. 2.16



SUBARKOSIC SANDSTONE

SHALE

ARGILLACEOUS LIMESTONE
TO CALCAREOUS MUDSTONE

SLIGHTLY RUBBLY ARGILLACEOUS LIMESTONE

RUBBLY ARGILLACEOUS LIMESTONE

PLANAR, THIN BEDDED LIME MUDSTONE

CRINOIDAL PACKSTONE/GRAINSTONE

THIN-BEDDED DOLOSILTITES

BOUNDSTONE

QUARTZ SILTY LIMESTONE



ALLOCHTHONOUS BLOCKS

F

COMMON BENTHIC FOSSILS

B

BIOTURBATION IN THE ABSENCE
OF ABUNDANT BODY FOSSILS

I

INTRACLASTS COMMON

T

TEMPESTITE STRUCTURES

O

OOIDS COMMON



SHALLOWING TREND



DESICCATION POLYGONS



SLUMP FOLDS AND DISRUPTED BEDDING



INTRAFORMATIONAL TRUNCATION SURFACES

Legend, fig. 2. 16

By Late Silurian time, however, siliciclastic deep-water substrates were exploited by diverse burrowers (Pickerill, 1980) and there is little reason to suspect that similar exploitation of deep water carbonate slope environments was not also achieved. Further, beds within the deeper water Douro facies do rarely contain a diverse ichnofauna, suggesting that depth alone did not account for the distribution of the burrowing organisms.

Strandline facies equivalent in age to the Douro Formation are not exposed in the study area. Such facies are, however, present in the underlying Leopold Formation, and it is reasonable to assume (by Walther's Law), that in this conformable sequence the Leopold Formation is lithologically typical of the nearshore Douro equivalents. The Leopold's intertidal units (e.g. Cape Clarence Member) were clearly deposited in a tidal flat environment (Dixon and Jones, 1980), and are hence strictly-speaking of low-energy type (James, 1978). However the common occurrence of ripples, cross-bedding, channels, intraformational conglomerate and club-shaped stromatolites, and the abundance of sand-sized detritus, all point to prevailing energy conditions considerably greater than those prevailing during deposition of the subtidal Douro Formation. The criteria outlined above suggest a depositional-structural model of ramp rather than platform affinities, for the bulk of the Douro Formation.

The Douro ramp was, in all probability, initially inclined to the north, with paleobathymetric contours parallel to the Ordovician-Silurian line of facies change (FIG. 2-1) that transects present-day Cornwallis Island from east to west (Thorsteinsson, 1958).

2.11.3.2 Vertical Facies Changes Due To Epeirogenesis

Features of the upper portion of the Douro Formation, the nature of the upper contact, and the facies characteristics of the overlying lithostratigraphic units, are summarized in FIG. 2-16. It shows that from section to section there is appreciable lithological variation, both within and immediately above the Douro Formation. This variation can be related directly to the initial paleobathymetric and paleogeographic position of each section on the ramp. In general, sections A to G represent progressively deeper environments of deposition. Irrespective of initial paleobathymetric position, all sections, except F and G, show a distinct shallowing towards the top, as indicated (B to D) by the return of an abundant diverse benthos associated with bioturbated rubbly limestones. The criteria for shallowing at the top of the Douro on southeastern Somerset Island are given by Narbonne (1981), and evidence for shoaling at Goodsir Creek has been documented elsewhere in this paper. Five sections contain intraformational truncation structures and/or slumped bedding; these

structures either just precede or accompany the regressive sequences. Although these structures indicate only slope instability, slope failure is commonly related to earth tremors (see Dott, 1969), and it is reasonable to speculate that earthquakes would have been associated with the initial epeirogenic rise of the Boothia Uplift.

Bioherms and biostromes occur in section B, at the very top of the Douro Formation. Possibly these organic structures developed about an embryonic shoal area overlying the axis of the Boothia Horst. Although speculative, subaqueous doming of the uplift may have substantially altered ramp configuration, with development of an increasingly pronounced shelf-slope break more suitable for organic reef growth. Certainly by latest Ludlovian to Pridolian time, a well-differentiated platform existed (Paper 5). If further research can substantiate these preliminary findings, the Douro Formation may well prove to be an excellent example of ramp to platform evolution, where the end product resulted not from localized autogenic organic accumulation (see Wilson, 1975, p.21) but from external diastrophism.

The presence of phosphorite at the top of the Douro Formation at locs. 4,5,6,7 suggests that the oxygen minimum layer impinged upon the ramp slope substrate. Two possible explanations are advanced:

1. the creation of a fault-angle depression proximal to the Boothia Horst (Walcott, 1970), could have been associated with sudden sea floor subsidence into deeper, anoxic waters, or
2. tectonism could have induced basin topography that further restricted circulation, thereby causing the oxygen minimum layer in the Cape Phillips Back-Barrier Basin to expand upward in the water column. For example, increased elevation of the Melville-Bathurst ridge could have increased basin stagnation. The Melville-Bathurst ridge is coincident with a major east-west geomagnetic anomaly (see McGrath and Fraser, 1981), possibly representing a basement arch. Fault control of the Melville-Bathurst Ridge is speculative, but Trettin (in Trettin and Balkwill, 1979) has suggested "fault-controlled" evolution for reefs bordering the Hazen Trough on Ellesmere Island. These reefs may have developed on a northeast extension of the Ridge. Thus, increased elevation of the Melville-Bathurst Ridge could have been accomplished either by fault block uplift, or by organic rim development, or perhaps by a combination of both.

2.11.3.3 The upper Contact: An Event Horizon

The upper contact of the Douro Formation is remarkably abrupt in all the studied sections, yet nowhere is it erosional. There is local evidence of a limited period of non-deposition (see discussion on the nature of lower contact of the lower member at its type section), although its temporal and spatial extent is not clear. However, there is a remarkable congruity between lithostratigraphic and biostratigraphic contacts (top of Douro Formation, top of the siluricus Zone), as documented by Thorsteinsson (1980). This coincidence of physical and faunal breaks exists despite the clearly differing uppermost Douro facies represented at separate localities.

This implies that a single, substantial event affected the entire depositional basin simultaneously. The upper contact of the Douro Formation is thus an 'event horizon', a conclusion alluded to by Thorsteinsson (1980). This event represents a significant tectonic pulse of the Cornwallis Disturbance. Interestingly, although the magnitude of change resulting from the tectonism appears to be similar in all environments of deposition, the direction of change observed is not consistent. On Devon Island (FIG. 2-16), the Douro Formation is overlain by clinoform deposits of the Devon Island Formation (Morrow and Kerr, 1977), indicating abrupt deepening. In contrast, all other Douro sections except Δ F, which shows no apparent radical change in paleobathymetry, are overlain by deposits that contain,

within 50 m of the Douro upper contact, evidence of subaerial emergence. As indicated on FIG. 2-16b, this disparity in direction of movement may be explained in terms of an imaginary pivotal point (or tilt axis; see discussion by Morrow and Kerr, 1977) separating an area of net crustal uplift from one of crustal subsidence. With block fault movement within the Boothia Uplift (Kerr, 1977), it is reasonable to assume that south of this imaginary pivotal point, uplift of a fault block or series of blocks would have caused net shallowing in the marine environment, and an overall regressive sequence. Complementary isostatic readjustment north of this pivotal point, could have resulted in fault block subsidence, and an apparent transgressive record such as the Upper Silurian succession on northern Devon Island.

The foregoing interpretation requires that the horst act independently north and south of the pivotal point, which in turn necessitates the existence of one or more E-W faults crossing the horst. Muir and Rust (1983) recently postulated that a major E-W synsedimentary fault controlled the long-lived position of the Cape Airy - Snowblind Bay line of facies change. The relative independence of fault block movement along this elongate horst complex, is suggested by the presence of an angular unconformity (representing pulse 1 of the Cornwallis Disturbance) at the base of the Cape Storm Formation on Grinnell Peninsula, NW

Devon Island. To the south on Cornwallis and Somerset islands, the Allen Bay to Cape Storm transition is conformable, and thus the uplift and erosion on Grinnell Peninsula is difficult to explain in terms of progressive south to north uplift of the Boothia Horst, and is more satisfactorily elucidated in terms of independent fault block movement.

2.11.4 Tectonic Synthesis

The epeirogenic phase referred to in the following synthesis, represents the pulse of uplift affecting the study area in late Ludlow time (FIG. 2-12). The post-epeirogenic phase is the period of quiescence and isostatic readjustment in latest Ludlow time that immediately followed this tectonic pulse.

2.11.4.1 Pre-Epeirogenic Phase

For much of siluricus Zone time, the Douro Formation accumulated on a relatively stable, northerly-inclined carbonate ramp that formed the southern and southeastern edge of the Cape Phillips Back-Barrier Basin. Where the ramp lay between storm and normal wave base, sponge mud mounds were abundant (FIG. 2-17). Diverse shelly benthos was associated with this reefal development, and with the shallowest subtidal parts of the ramp (Narbonne, 1981). The waters of the Cape Phillips Back-Barrier Basin were

stratified, with increasingly anoxic conditions with depth, as reflected by the biofacies distribution.

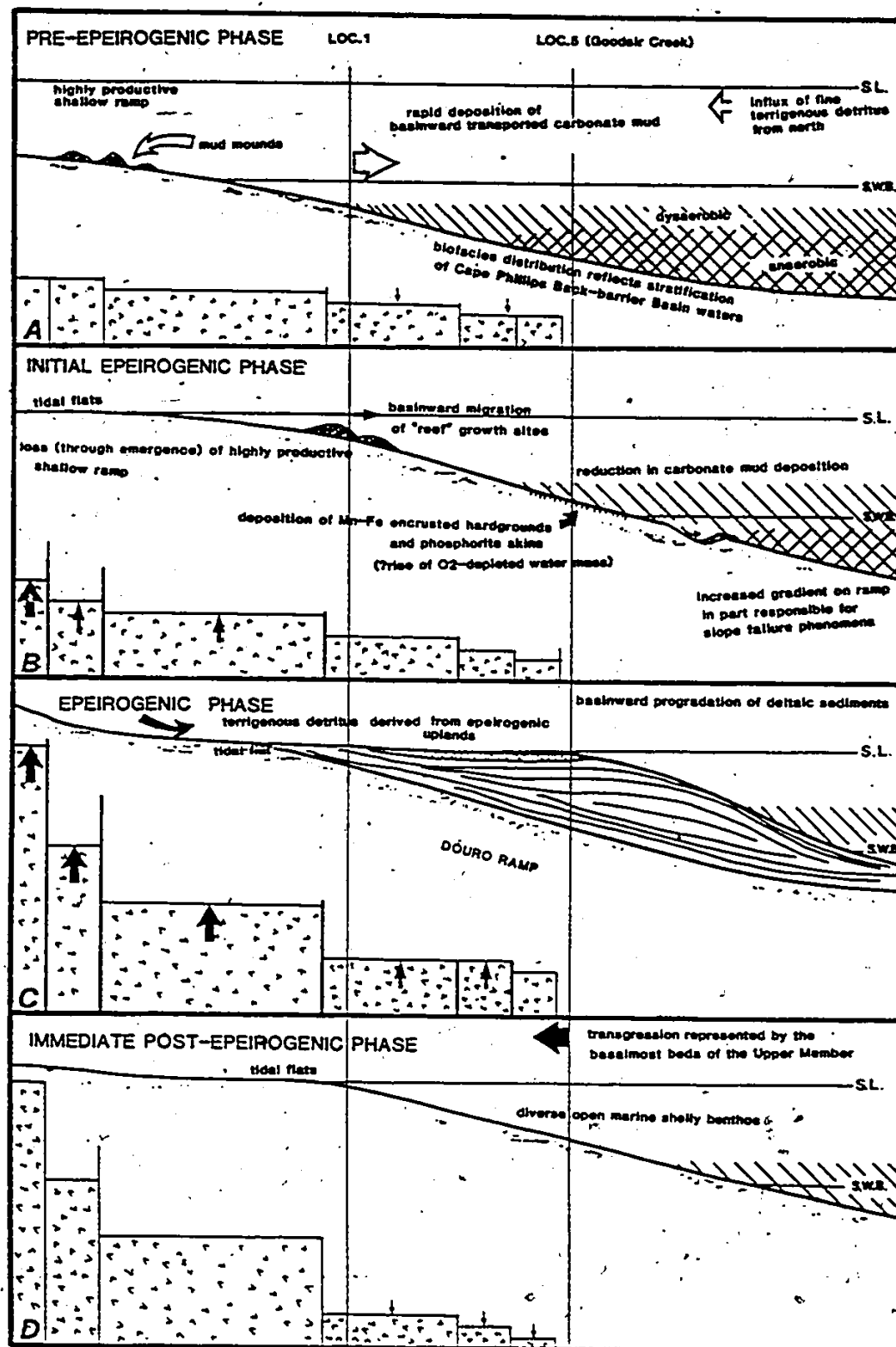


FIG. 2-17 Late Ludlovian tectono-depositional phases.

Rapid subsidence of the geocline accommodated a considerable thickness of Douro sediment. Deeper portions of the ramp received large volumes of lime mud, much of it basinward-transported fine carbonate sediment from the productive shallower portions of the ramp. Fine siliciclastic detritus was introduced from a northerly source (Narbonne, 1981).

2.11.4.2 Initial Epeirogenic Pulse

Although the mechanism remains controversial (Walcott, 1970; Kerr, 1977; Miall, 1983) blocks of Aphebian basement, and suprajacent Proterozoic - Lower Paleozoic sedimentary strata, were strongly uplifted in late Ludlow time in the vicinity of the presently exposed Boothia Horst. To the south of what is now Barrow Strait, this uplift eventually produced an emergent area (B.A.L.), surrounded by extensive tidal flats. On the western margin of the landmass, local areas of higher relief (fault scarps?), resulted in alluvial fan deposits.

The effect of this uplift north of Barrow Strait in the subtidal marine environment was appreciable. The structural-depositional configuration of the original Douro Ramp was substantially altered, with a dramatic reduction in area of the highly productive shallow shelf, and increased depositional gradients at the margins of uplifted basement blocks.

Steeper slopes, in conjunction with elevated seismic activity, caused more frequent slope failure. With loss of a large proportion of the productive shallow shelf, carbonate sedimentation diminished substantially on the slope. This fact, coupled with overall shoaling of deeper subtidal environments, favoured widespread hardground development. The tectonism was responsible for elevating deeper portions of the ramp up to and above storm wave base (e.g. loc. 5). Shoaling south of Barrow Strait eliminated environments suitable for reef growth, and organic buildups were constructed basinward (loc. 1) of their former occurrence, perhaps concentrated at or near a newly evolved shelf-slope break (FIG. 2-17).

At the end of Douro deposition, an expansion of the oxygen minimum layer within the Cape Phillips Back-Barrier Basin, coupled with decreased sedimentation rates, resulted in deposition of phosphorite on the Douro slope. Increased anoxia of the basin waters may have resulted from reduced circulation, in turn arising from substantial alteration of sea-floor topography by ongoing diastrophism (e.g. Smith, 1969). Postulated uplift of the Melville-Bathurst Ridge represents a plausible means of reducing basin circulation.

2.11.4.3 Epeirogenic Phase

Mass-wasting of nearby newly emergent uplands provided the siliciclastic detritus that was then introduced into the

restricted, at least partially anoxic, Cape Phillips Back-Barrier Basin. The resulting marine deposit (the lower member) represents a ?tide-dominated, wave-influenced delta (FIG. 2-17). Prodeltaic sedimentation blanketed a wide range of pre-existing carbonate lithofacies of the Douro Formation, ranging from reef to distal basin slope. Progradation of the delta, and by inference continued uplift and availability of sediment, is evidenced by the pronounced upward coarsening of the deltaic sequence at locs. 4 and 5. Uplift culminated in emergence of depositional surfaces at most localities (1,2,3,4,5,a), within 50 m of the top of the Douro Formation. In contrast, on Devon Island (FIG. 2-16), the Douro - Devon Island Formation transition represents net subsidence, suggesting that uplift in southern Cornwallis Island, was compensated for by a downdrop to the north. A neutral point (hinge line) may have existed in the vicinity of Loc. 8.*

Potential crystalline basement source for the siliciclastics include the Boothia Horst, and shield areas to the east or possibly southeast (eastern Devon and northwestern Baffin Island). If the former, a dextral transcurrent motion along the bounding faults of the Lancaster Aulocogen in post-Silurian time, would explain the

* This accelerated subsidence of a portion of the depositional basin contiguous with, although not proximal to, the rising Boothia Uplift, is a consequence not forecast by the essentially autogenic isostatic model proposed by Walcott (1970).

present facies disposition on either side of Barrow Strait.

2.11.4.4 Immediate Post-Epeirogenic Phase

A major transgression signals the cessation of uplift in latest Ludlow time (mid? latialata Zone). Carbonates of open marine aspect onlap beach facies of the lower member at localities 2,3,4,5, and b (FIG. 2-17). As there is no evidence of global eustatic sea level rise at this time, the transgressive sequence must be explained in terms of accelerated crustal subsidence, perhaps due to post-epeirogenic readjustment of basement fault blocks. This period, tentatively of readjustment, may be reflected south of Barrow Strait in the inclusion of thick units of subtidal, Douro-like lithofacies within the tidal flat succession of the late Ludlovian lower member of the Somerset Island Formation. Such intertonguing is recorded at several localities on Somerset Island at this stratigraphic level (Miall et al., 1978), and is particularly well developed at Cunningham Inlet on the north coast.

Basinward of Loc. b on Cornwallis Island, a facies shift diagnostic of sea-level change is not recognized at the top of the lower member. Instead there is a simple reduction in the amount of fine siliciclastic detritus, and a commensurate increase in carbonate content within lithofacies interpreted as basin slope deposits.

2.11.4.5 Regional Significance

The Boothia Uplift structural province has well-defined boundaries and can be clearly differentiated from either the Arctic Platform, or the Parry Island and Central Ellesmere Fold Belts. Its distinctive structural style, perhaps coupled with the physiographic prominence of the Boothia Horst itself, tends to promote the view that the Siluro-Devonian epeirogenesis (Cornwallis Disturbance) that affected the Uplift was a spatially and genetically isolated event. The disturbance is seen to have been manifested by means of vertical movements along pre-existing north-south lines of structural weakness in the basement (Kerr, 1977), perhaps triggered by simple isostatic adjustment to differential loading (Walcott, 1970). The principal elements involved in the Uplift, at its deepest structural level, are subparallel, contiguous basement blocks of indefinite but apparently uninterrupted north-south length. Previous studies suggested that the Uplift has not undergone lateral displacement, and thus the Cornwallis Fold Belt and Boothia Horst are presently correctly aligned with respect to their original Siluro-Devonian positions.

The perception outlined above, may prove to be too limited. Miall (1983) recently questioned the validity of simple vertical movements as the overriding mechanism producing the Uplift, suggesting as a hypothetical alternative, large-scale thrust faulting. Trettin, as early

as 1963, recognized that Caledonian* age disturbances affected not only Boothia Uplift, but also the Rens Fiord Uplift on Axel Heiberg Island (FIG. 2-18). Trettin (1963, p.113-114) further observed that, "...structural trends of Tertiary age suggest that all of the Sverdrup Islands were involved in the [Caledonian] movements but this evidence is not conclusive". In summary he stated, "...the approximate coincidence in time of the Caledonian disturbance of the Canadian Arctic Islands with important movements in eastern Greenland, Spitsbergen, northern Russia and Siberia, possibly Scandinavia ... and parts of the North American continent is probably not incidental but points to a stress field of large dimensions". More recent syntheses (Trettin and Balkwill, 1979; Christie, 1979; Kerr, 1981) document widespread, contemporaneous episodes of crustal uplift on at least five discrete basement arches in the archipelago (summarized in Table 2-13; see also FIG. 2-18). Apart from the above cited basement uplifts, mention has

* Caledonian is used here in the restricted sense of Late Silurian - Early Devonian.

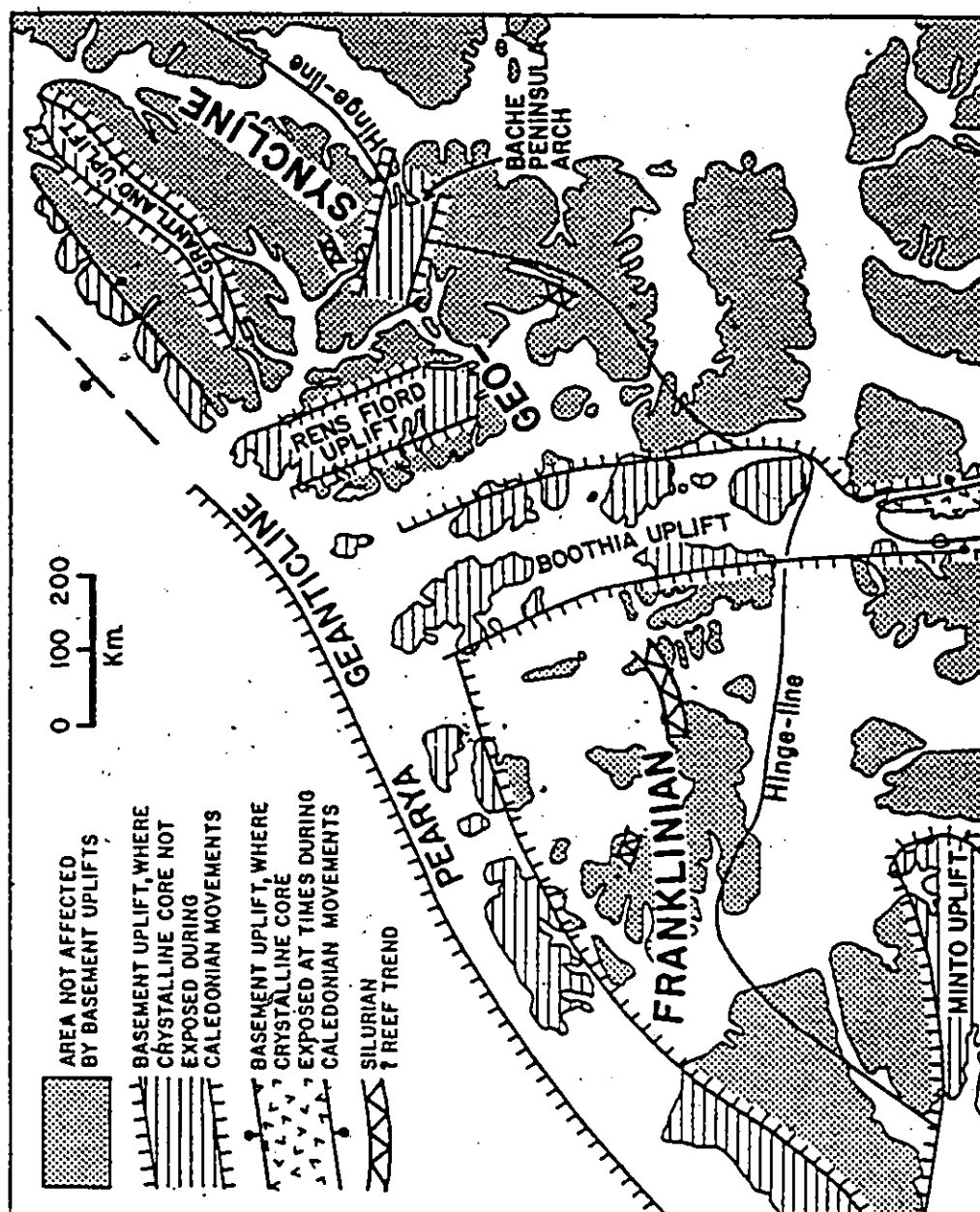


FIG. 2-18 Major structural highs of the Canadian arctic affected by the Caledonian diastrophism. (modified after Kerr, 1981).

Areas affected by Caledonian movements in the Arctic Archipelago

STRUCTURE	TIMING OF DIASTROPHISM	EVIDENCE and/or REFERENCE
1. BOOTHIA UPLIFT (fault- bounded)	early Lochkovian late Ludlow late Wenlock - late Ludlow	-synorogenic non-marine clastic deposits -regressive marine sequences -penecontemporaneous faulting and folding -regional and local unconform- ities
2. RENS FIORD UPLIFT * (fault bounded)	younger than late Wenlock to mid-Devonian	-major unconformity -late Silurian intermediate to acid volcanics
3. BACHE PENINSULA ARCH	Lochkovian	-regional unconformity -synorogenic red beds
4. GRANTLAND UPLIFT	early Devonian	Trettin (1972) Kerr (1981,p.124)
5. PEARYA GEANTI- CLINE	Lochkovian	-plutonism circa 390 ± 20 m.y. -metamorphism K/Ar 403 ± 17 m.y. 389 ± 21 m.y.
6. MINTO UPLIFT (?fault bounded)	?Devonian	Kerr (1981,p.135)
7. FAULT ZONE EAST SIDE OF MELVILLE PENINSULA	?early Devonian	Trettin (1975)

* Trettin (1979,p.63) postulated that uplifts other than the Rens Fiord Uplift, and as yet unidentified, may have provided some of the detritus to the Stallworthy Formation, member C of the Lands Lokk Formation and the Caledonian Bay member of the Imina Formation.

been made in this paper of other areas potentially involved in Siluro-Devonian tectonism, including: 1) northwest Baffin Island, 2) eastern Devon Island, 3) the Melville-Bathurst Ridge and 4) northwest Cornwallis Island. According to Kerr (1981), the Boothia and Rens Fiord Uplifts, and possibly the Minto Uplift, were probably connected with, or were a part of, the Pearya Geanticline. The interrelationship between all these structural elements is evidently complex, but their near contemporaneous deformation suggests a common genetic link to a single plate event. The collision between the North American and Eurasian Plates, resulting in the Caledonian Orogeny, represents one such event. This collision produced severe deformation of the Caledonian Geosyncline of northeastern Greenland, and may have induced sufficient stress in the Platform and Innuitian Mobile Belt to initiate widespread basement uplift. Early Devonian magmatic activity in the Pearya Geanticline has been variously interpreted to indicate incipient rifting (Trettin and Balkwill, 1979), or the initial stage of the Ellesmerian Orogeny (Kerr, 1981, p.126), that may have been caused by collision of the Barents Splinter and North American Plates (Christie, 1979). This second major tectonic event could represent another candidate, or at least a significant modifying influence. The Cornwallis Disturbance and the Ellesmerian Orogeny overlap considerably in time, and thus

* Plate collisions and rifting are not mutually exclusive processes (see Sengor, Burke and Dewey, 1976).

attributing particular crustal movement to either a Caledonian or early Ellesmerian event is most difficult, particularly with reference to basement uplifts close to the Pearya Geanticline. Although it remains to be proven, a reasonable working hypothesis is that the various basement uplifts are consanguineous, their deformation in Late Silurian time being triggered by a common mega-tectonic event.

The Boothia Uplift itself may be more complex than previously supposed. The basement fault blocks of indefinite north-south strike length may in fact have been traversed by at least one (if not several) major east-west faults (e.g. Muir and Rust, 1983), and in all likelihood the Uplift consists of a complex of semi-independently interacting basement blocks.

The final question that must be addressed is, whether or not the Boothia Horst and Cornwallis Fold Belt have retained their relative mutual positions following Siluro-Devonian tectonism. Some circumstantial evidence suggests a dextral strike-slip motion of between 60 and 100 km on an east-west trending fault of the Parry Submarine Rift Valley. A component of horizontal shear can be associated with graben structures (e.g. Bederke, 1966; Illies, 1981), and pending definite evidence to the contrary, transcurrent motion remains a distinct possibility.

2.12 CONCLUSIONS

1. Overall, the Douro Formation can be considered the result of nearly continuous and rapid deposition on a carbonate ramp. The upper 10 m of this formation in the vicinity of Goodsir Creek, Cornwallis Island, is a regressive sequence that commences with interpreted deep-water, unfossiliferous basin slope facies and terminates with strata deposited somewhat above storm wave base in the photic zone. The banded lime mudstones and thin phosphorite layer suggest low sedimentation rates and intermittently anoxic conditions within the basin at the close of Douro deposition.
2. The upper contact of the Douro Formation in the study area is an event horizon, signalling a major pulse of the Cornwallis Disturbance in late Ludlow time.
3. The lower member of the Barlow Inlet Formation consists of dark-colored mudrocks (clayshales to siltstones) with black shale affinities.
4. The lower member represents a wave-influenced, tide-dominated high-suspended load delta, deposited in a partially anoxic basin. Prodelta and beach foreshore facies are preserved. The siliciclastic detritus of the lower member was derived from the mass-wasting of proximally-located epeirogenic uplands that include crystalline basement terrane.

5. Several factors argue against the choice of the Boothia Horst as a source for the siliciclastic sediments of the lower member. However, no data are presently available that is suggestive of a more suitable alternative.
6. Dextral strike-slip motion along faults in Parry Channel is tentatively proposed to reconcile a Boothia Horst source for the lower member with present clastic facies disposition.
7. The Late Silurian crustal movements affecting the Boothia Uplift can probably be related to major plate interactions responsible for either the Caledonian or Ellesmerian orogenies.

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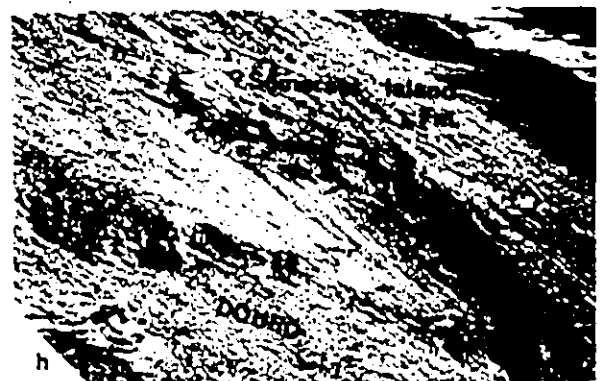
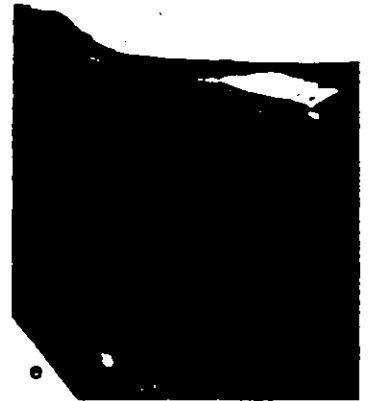
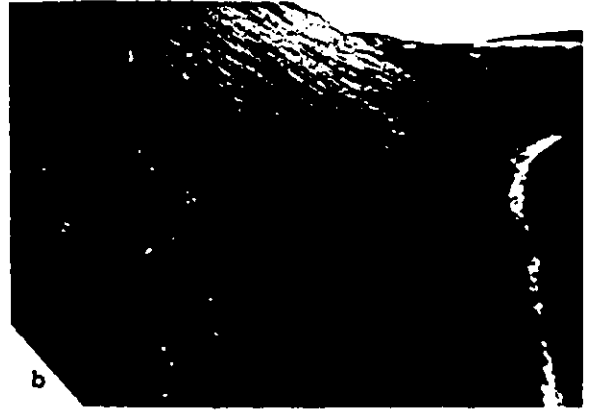
2.14 PLATES

Plate 2-1

Oblique views of principal sections.

Small arrows indicate positions of lower and upper contacts of the Lower Member of the Barlow Inlet Formation. In figures a through e, and g, the lower member is underlain by the Douro Formation and overlain by the upper member of the Barlow Inlet Formation.

- a View to the north along Cape Hotham Escarpment, with resistant carbonates of the Upper Member overlying poorly exposed mudrocks of the lower member. Arrows indicate exposures of lower member. Sequence dips at 30° to east. Escarpment top is approximately 90 m above valley floor.
- b Type section of the lower member (54 m) at Goodsir Creek (loc. 5). Note sampling trenches. Large arrow points to prominent hardground near top of the Douro Formation. View to the south.
- c Gullied exposure of the lower member (65 m) near loc. 7. View to the south.
- d Well-exposed and easily accessible section of the lower member (38 m) at loc. 3. View to the east.
- e Reference section of the lower member (52 m) at loc. 4. View to the south.
- f Douro/lower member contact at loc. 8. Arrow points to base of disturbed bedding at the top of the Douro Formation. Hammer head rests on the contact.
- g Southernmost exposure of the lower member (8.8 m) on Cornwallis Island, at Barlow Inlet (loc. 2).
- h Lower member (6.8 m) as exposed at Cheyne Point, Griffith Island (loc. 1). The lower member immediately overlies biohermal rocks (B) of the Douro Formation.

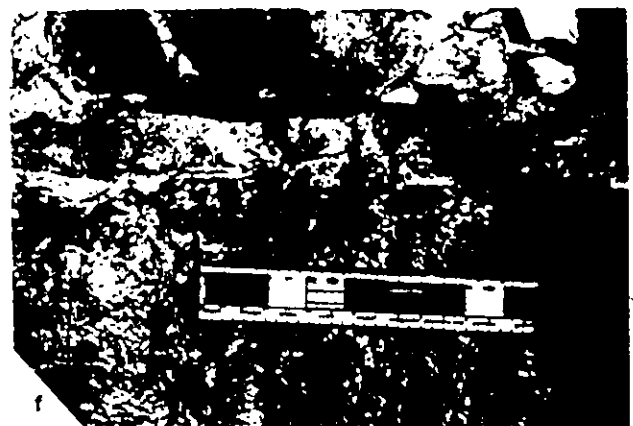


PL2-1

Plate 2-2

Uppermost Douro Formation beneath lower member
reference section.

- a Excavated lower contact of lower member (foreground). Arrow points to 10 m thick resistant ledge of uppermost Douro Formation.
- b Close-up of resistant ledge. Note predominance of thin, even, planar bedding.
- c Prominent hardground 8.2 m below Douro/lower member contact. Note thinning of beds to left (arrows) above hardground. Note also the massive appearance of interval immediately beneath hardground. Vertical field of view is 2 m.
- d Uppermost beds of Douro Formation (arrow points to lower member mudrocks). Beds above hammer head contain numerous gutter casts and brachiopod floatstones.
- e Discoid holdfast attached to Fe/Mn encrusted hardground surface depicted in figure c, above. Scale 1 cm.
- f Close-up of section through hardground showing (1) relief of surface, and (2) muddy (rather than rudaceous) nature of overlying beds. Scale 15 cm.



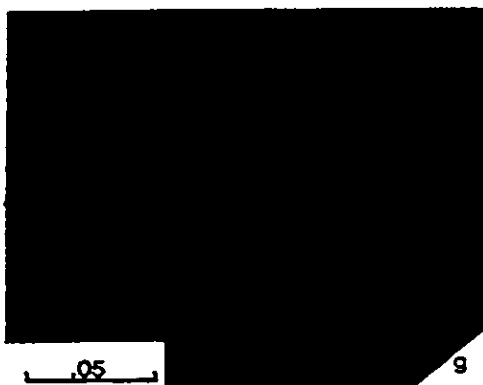
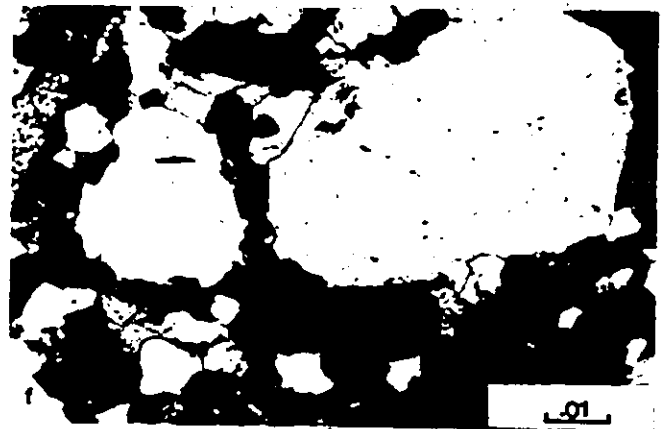
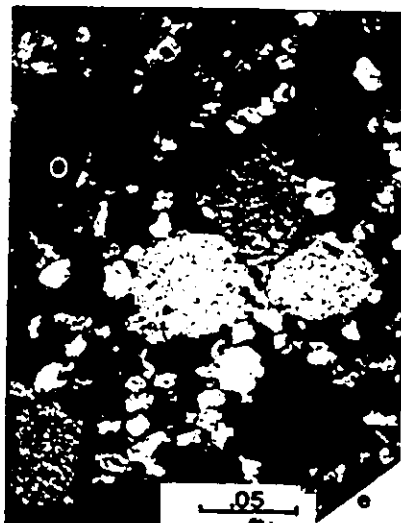
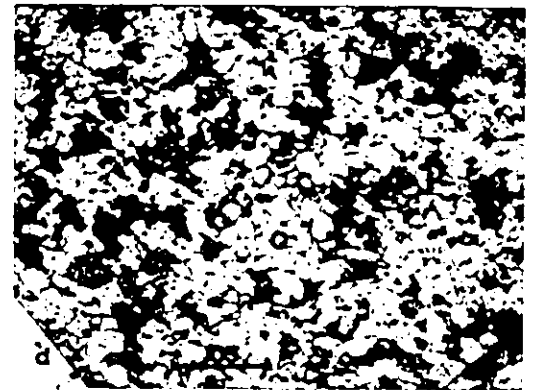
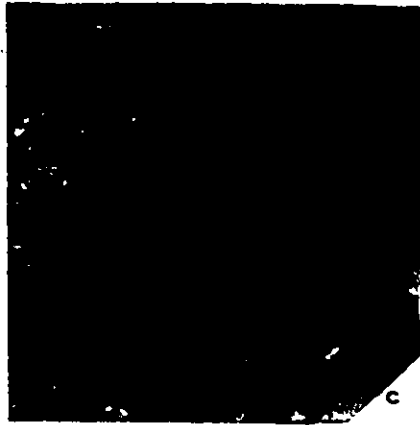
PL2-2

Plate 2-3

Petrography of Lower member rocks.

Bar scales are 1 cm (unlabelled) or hundredths of 1 cm (labelled).

- a Fine color lamination in conchoidally-fracturing clayshale.
- b Photomicrograph of clayshale exhibiting pyrite spherules, large illite/mica flakes (light-colored platelets), and quartz silt in an illite clay mineral matrix.
- c Bioturbated siltstone (?Chondrites).
- d Thin section of equigranular, fine subarkose etched by hydrofluoric acid fumes in order to distinguish quartz (clear-Q), and feldspar (cloudy-F).
- e Bimodal texture of heterostracan-rich subarkose. Arrow points to fish fragment.
- f Interlocking mosaic of predominantly quartz grains. Note euhedral rhomb of dolomite just above coarse quartz grain at left; this suggests dolomitization preceded competitive syntaxial quartz cementation. Note also the apparent increased roundness of larger grains and optical continuity of detrital cores and cement.
- g,h Thin section of subarkose in plane light (g) and under cathodoluminescence (h). Luminescent grains are feldspar; the more brilliant ones (e.g. Y,X) are plagioclases.



PL2-3

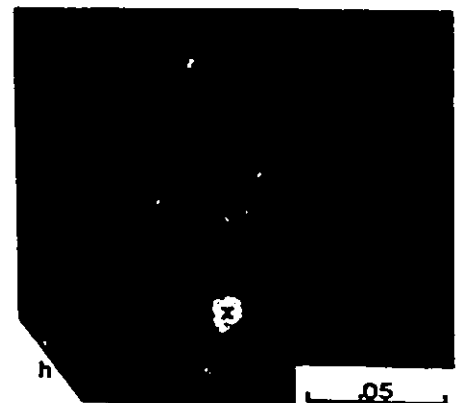
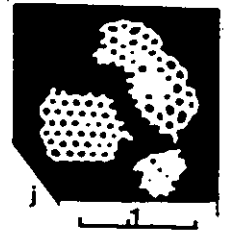
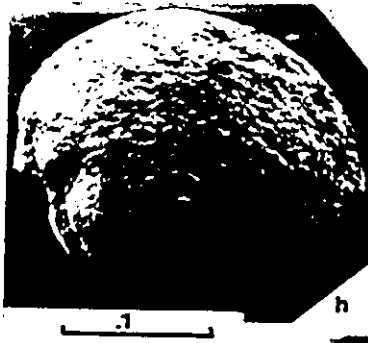
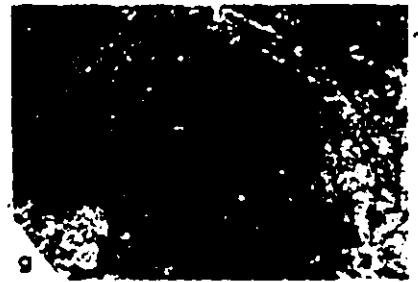
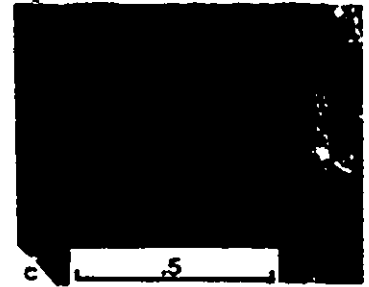
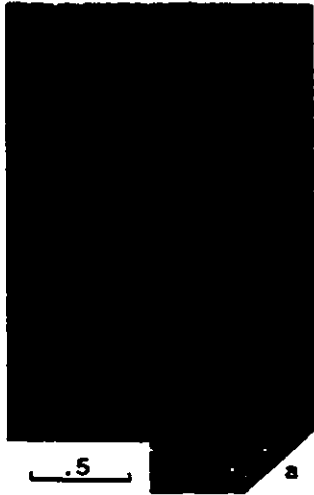


Plate 2-4

Arthropods, ostracoderms, echinoderms and graptolites
from mudrock facies.

Bar scales are 1 cm (unlabelled) or tenths of 1 cm
(labelled).

- a Paraaulacopleurid trilobite. Lack of free cheeks may indicate specimen is an exuvia. loc. 5, 10 m above base.
- b Paraaulacopleurid trilobite showing nature of free cheek and eyes. loc. 5, 10 m above base.
- c Cephalon of indeterminate trilobite from loc. 3, 3 m above base.
- d Dorsal shield of ?Ariaspis n. sp. loc. 5, 10 m above base.
- e Dorsal shield of indeterminate cyathaspidid ostracoderm (box enlarged in figure g), loc. 7, 1 m above base.
- f Dorso-lateral scale of indeterminate cyathaspidid. loc. 5; 10 m above base.
- g Detail of polygonal ornamentation of cyathaspidid in figure e, above.
- h Entomozoid ostracod: Elpezoe ?borealis Copeland (photograph courtesy of M.J. Copeland, GSC, Ottawa).
- i Assorted crinoid plates (columnals, brachials, pinnulars) and spines from sample of disaggregated mudrock. loc. 5, 10 m above base.
- j Possible holothurian sieve plates. loc. 5, 10 m above base.
- k,l Monograptus n. sp. (cf. M. ultimus Perner) loc. 6, 10 m above base.
- m,n Microcrinoids. Pelma(P), brachials(A), arms(1), calyx(2), stalk(3). loc. 7, 55 m above base.



PL2-4

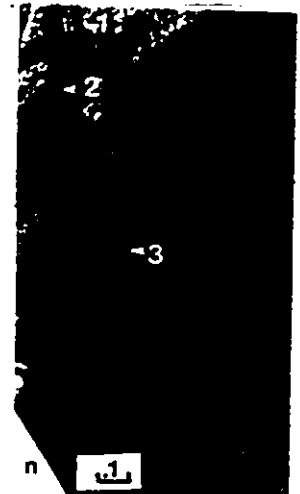
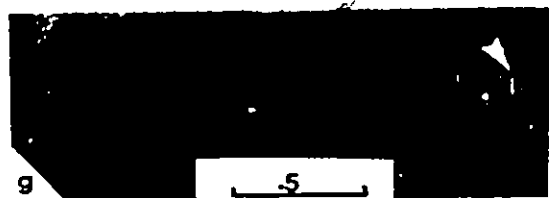
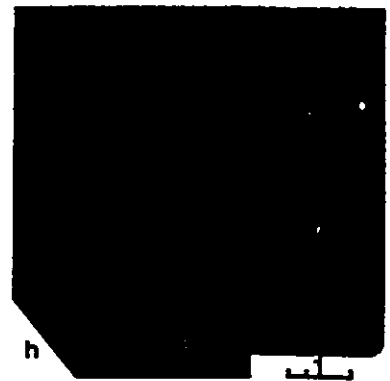
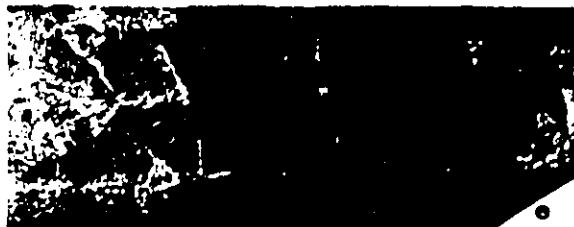
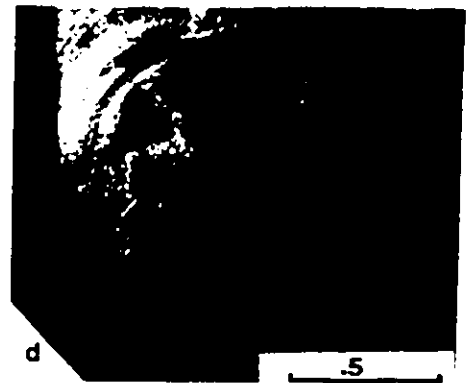
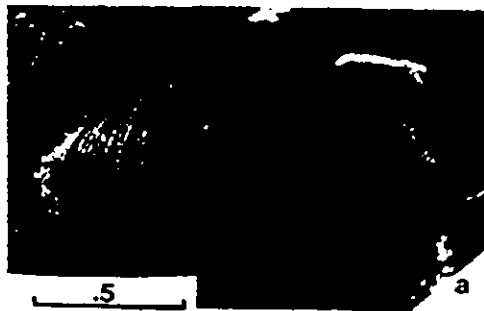


Plate 2-5

Mollusks from mudrock facies.

Bar scales are 1 cm (unlabelled) or tenths of 1 cm (labelled).

- a,b Bivalve sp. A. (multiple horizons, loc. 5).
- c Bivalve sp. D. (Pectinacea -- cf. Cornellites sp.)
loc. 7, 1 m above base
- d Articulated open valve pair of bivalve sp. B. loc. 5,
10 m above base.
- e Compressed, diminutive orthoconic nautiloid exhibiting
both (1) external ornamentation of concentric rings,
and (2) internal septa. loc. 7, 1 m above base.
- f Steinkern of orthoconic nautiloid in association with
lingulid brachiopod. loc. 5, 30 m above base.
- g Flattened orthoconic nautiloid with uncompressed apex
(arrow). loc. unknown.
- h Assorted small gastropods (plus two smooth ostracodes)
from disaggregated mudrock sample. loc. 5, 10 m above
base.

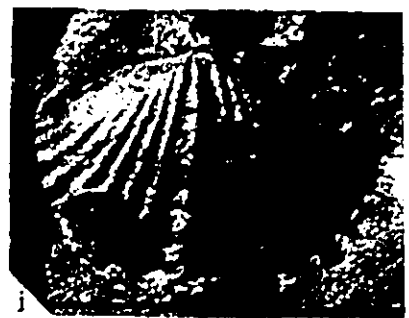
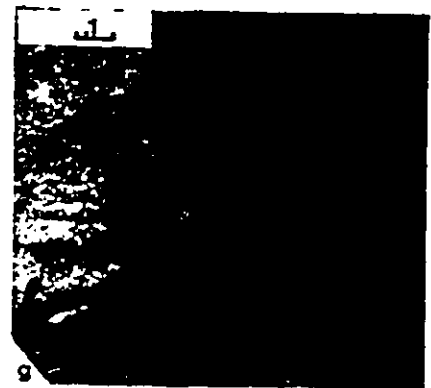
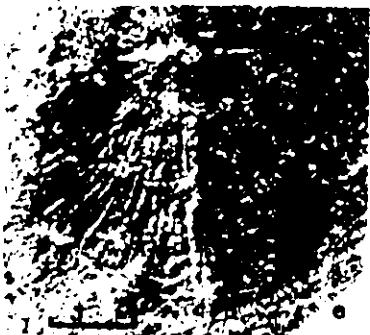


PL2-5

Plate 2-6

Brachiopods from mudrock facies.
Bar scales are tenths of 1 cm.

- a,e,h Single valves of brachiopod sp. A. loc. 5, 10 m above base.
- b Lingulid from loc. 5, 27 m above base.
- c ?articulated valve pair of brachiopod sp. A. loc. unknown.
- f,i Brachiopod sp. B. (possible rhynchonellids). Note remnant calcareous shell material in figure i, Loc. 7, 1 m above base.
- g Rhynchonellid brachiopods. loc. 5, 30 m above base.
- j Cyrtinid brachiopod (brachial valve of ?Tannuspirifer) loc. 4, 1 m above base.

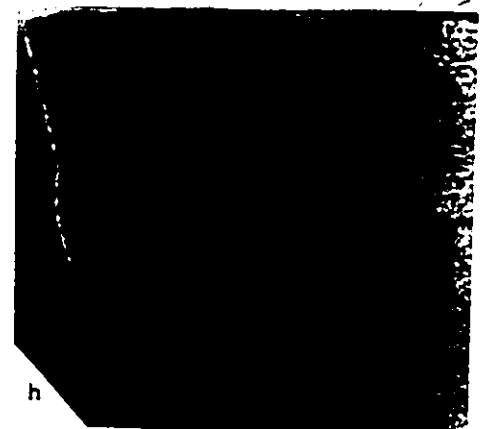
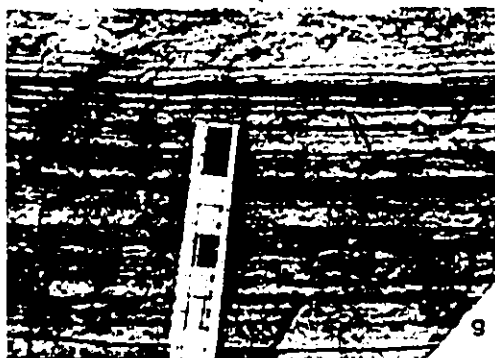
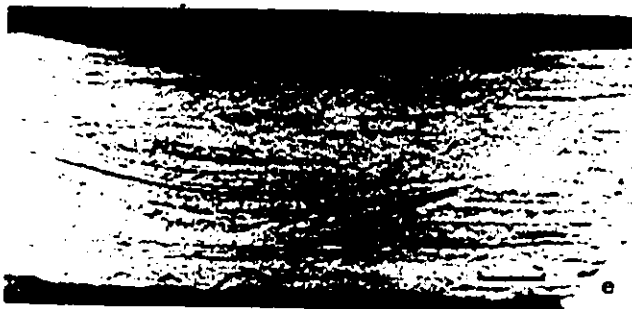
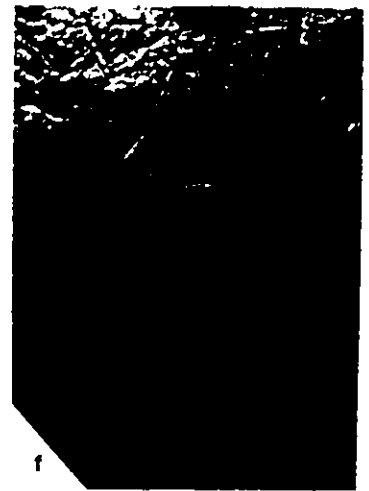
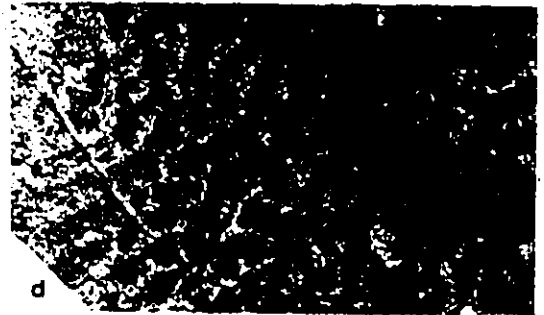
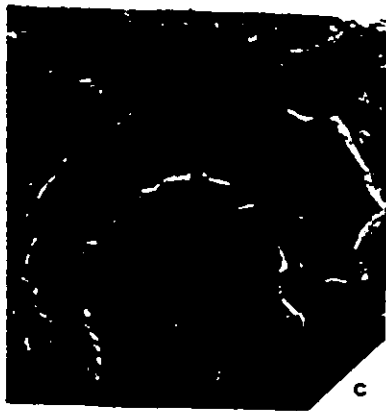
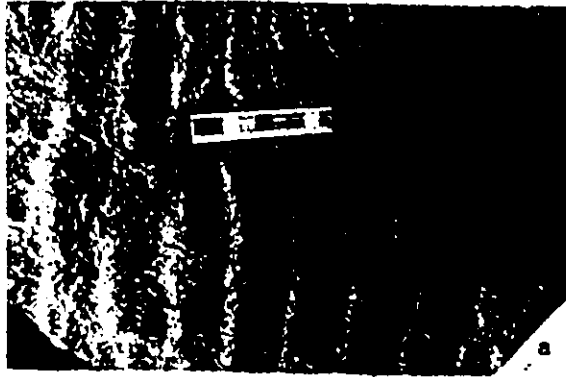


PL2-6

Plate 2-7

Structures of the subarkose facies.
Bar scales are 1 cm.

- a Weakly asymmetric, straight-crested, low-amplitude ripples. Scale 15 cm.
- b Polished slab exhibiting (1) unidirectional foreset laminae of weakly asymmetric ripples, and (2) gently-inclined parallel lamination.
- c Synaeresis cracks.
- d Runzelmarken.
- e Small-scale channel structure.
- f Desiccation polygons on bedding sole. Note abundant Cruziana traces.
- g Plane lamination. Scale 15 cm.
- h Low angle inclined lamination.

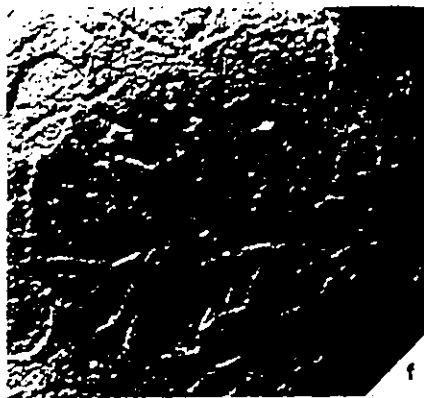
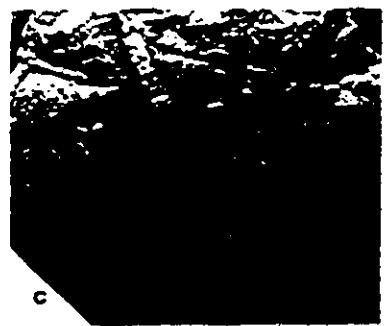
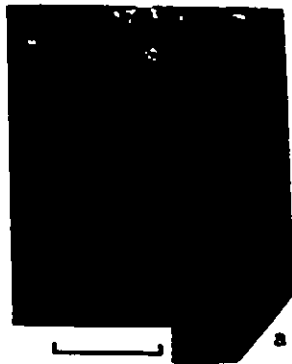


PL2-7

Plate 2-8

Lebenspurren and soft-sediment deformation structures.
Bar scales are 1 cm.

- a Bedding plane view of Skolithos (sand filled burrows in a coarse siltstone).
- b Bedding plane view of Planolites.
- c Bedding sole with Cruziana (c) and arthropod scratch marks (sm).
- d Alternating textural laminae of fine sandstone, siltstone, and clayshale. Sand-filled (?burrows, or ?synaeresis cracks, or ?injection dyklets) deformed by loading.
- e Bedding sole view of Cruziana.
- f Bedding plane view of Chondrites.
- g Load cast.
- h Loaded burrow on bedding sole.
- i Synsedimentary micro-fault. Bedding sole to top of picture.



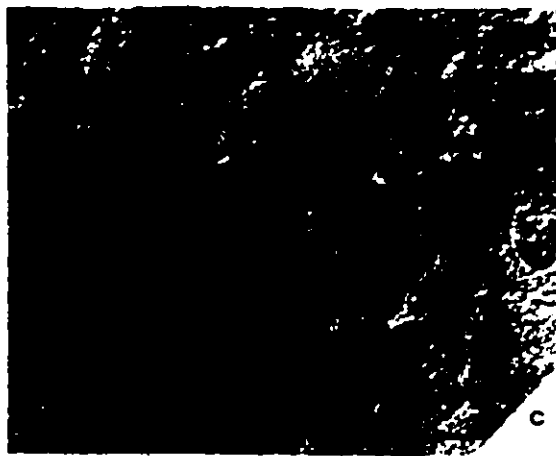
PL2-8

Plate 2-9

Fauna of the subarkose facies.

All specimens from scree at either loc. 4 or loc. 5.
Bar scales are 1 cm (unlabelled) or tenths of 1 cm
(labelled).

- a,c Poorly preserved rhynchonellid brachiopods.
- b Lingulid brachiopod with partial preservation of
chitino-phosphatic shell.
- d Bedding plane with abundant comminuted heterostracan
skeletons.
- e Bivalve sp. E.
- f Bedding plane rich in bivalve sp. D. (?modiomorphids).



PL2-9

III

PAPER III: SHALLOWING-UPWARD CARBONATE
HEMICYCLES IN AN UPPER SILURIAN MIOGEOCLINAL
SHELF SUCCESSION, ARCTIC CANADA

Abstract

The Upper Silurian Barlow Inlet Formation contains a significant proportion of peritidal lithofacies arranged in hemicyclic series. Two shallowing-upward sequences are recognized; a Restricted Infratidal (RI) Hemicycle and a Tidal Flat (TF) Hemicycle. Each complete cycle contains four members, representing higher energy subtidal lagoon or proximal back-barrier facies through to tidal flat facies.

The RI hemicycles consist of 1) a Biostromal Member, 2) a Sparsely Fossiliferous Mottled Dolomitic Carbonate Member, 3) a Massive Carbonate Mudstone Member, and 4) a Laminated Member. The TF hemicycles consist of 1) an Oncoidal Carbonate Member, 2) a Mottled Carbonate Member, 3) a Massive Carbonate Member, and 4) a Laminated Member. The hemicyclic sediments were deposited in an arid climate on a carbonate platform rimmed by crinoid sand shoals and or stromatoporoid-dominated biostromes and buildups, subject to minimal wave stress.

The muddy lithofacies and overall setting of the platform compares favourably with present-day Florida Bay, as well as ancient back-reef sequences described from the Devonian and Triassic systems.

The TF and RI hemicycles represent co-existing and intergradational environmental regimes. Depths likely did not exceed 14 m anywhere on the shelf. Hemicyclicity is best explained in terms of an episodic subsidence and tidal flat progradation model; eustatic sea-level rises need not be invoked.

3.1 INTRODUCTION

Uppermost Silurian peritidal carbonates are widely exposed in the central Canadian Arctic Archipelago. On Somerset, Griffith, and Cornwallis Islands, stratigraphic intervals assignable to lagoonal and tidal flat facies (collectively considered the platform interior magnafacies), represent a substantial proportion of the Somerset Island and Barlow Inlet Formations (FIG. 3-1).

This magnafacies, developed in Late Silurian time,⁷ affected by two regional tectonic controls: subsidence and epeirogenic uplift. The Franklinian Miogeocline⁸ is an area of folded shelf and basinal sediments (largely carbonates) 2.4 to 3.9 times thicker than nearby time-equivalent, flat-lying to gently inclined strata of the Arctic Platform (or Central Stable Region) (FIG. 3-2a,b). In contrast to, and possibly as a consequence of, increased subsidence in the miogeocline, episodic block-fault uplift of the Boothia Horst in Late Silurian through early Devonian time, resulted in emergence of a north-south trending landmass (the Boothia Axial Landmass -- FIG. 3-3). The record of continental syntectonic sedimentation on its flanks has been previously documented

⁷ Studied strata span the latest Ludlow and entire Pridoli (biostratigraphically this encompasses the latialata, eosteinhornensis, and crispa conodont Zones).

⁸ The Franklinian Miogeocline includes the Parry Island and Central Ellesmere Fold Belts, as well as part of the Cornwallis Fold Belt (FIG. 3-2a)

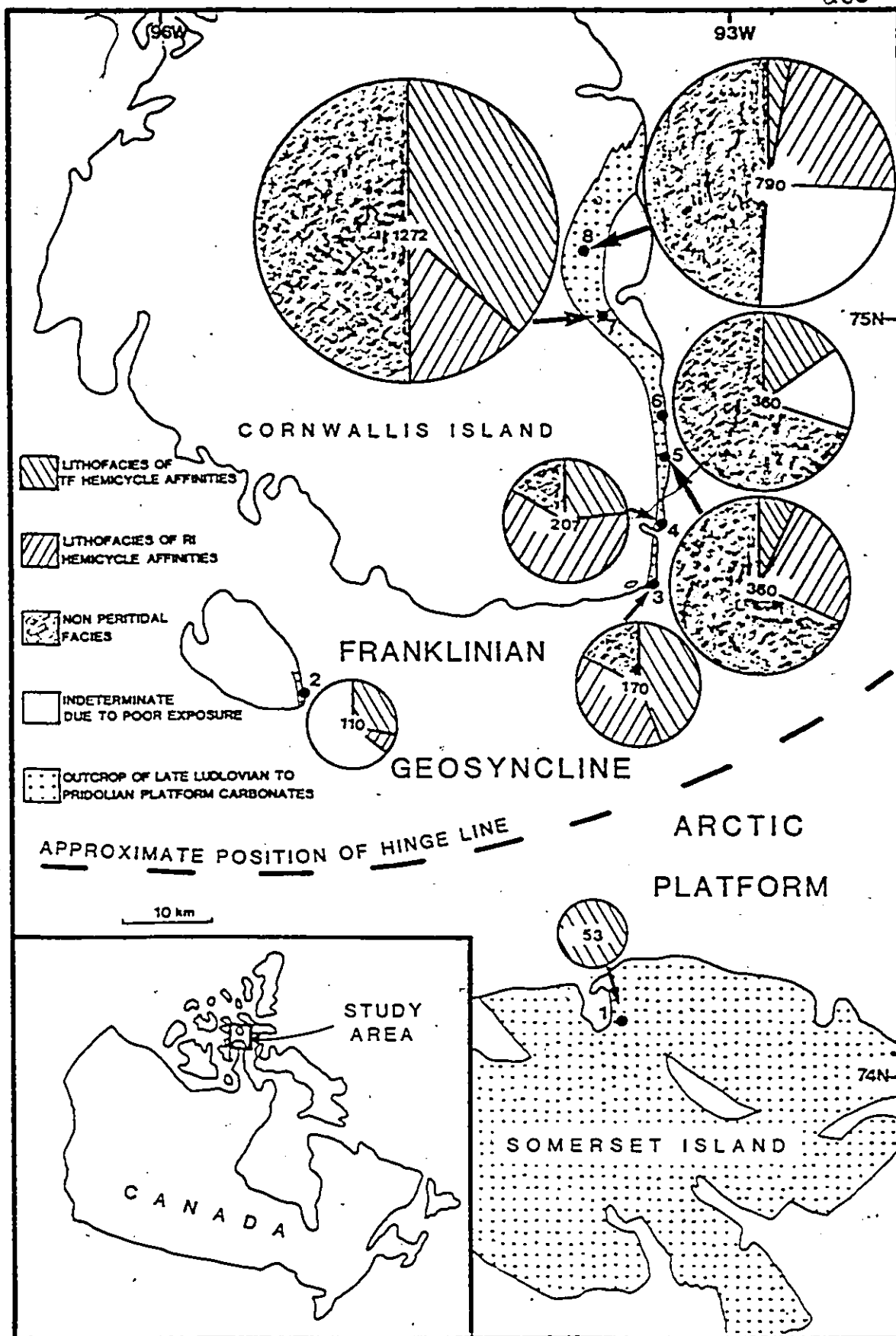


FIG. 3-1 Location map of study area showing proportion of peritidal facies. Numbers at the centre of the pie diagrams represent the total thickness in metres of the measured sections.

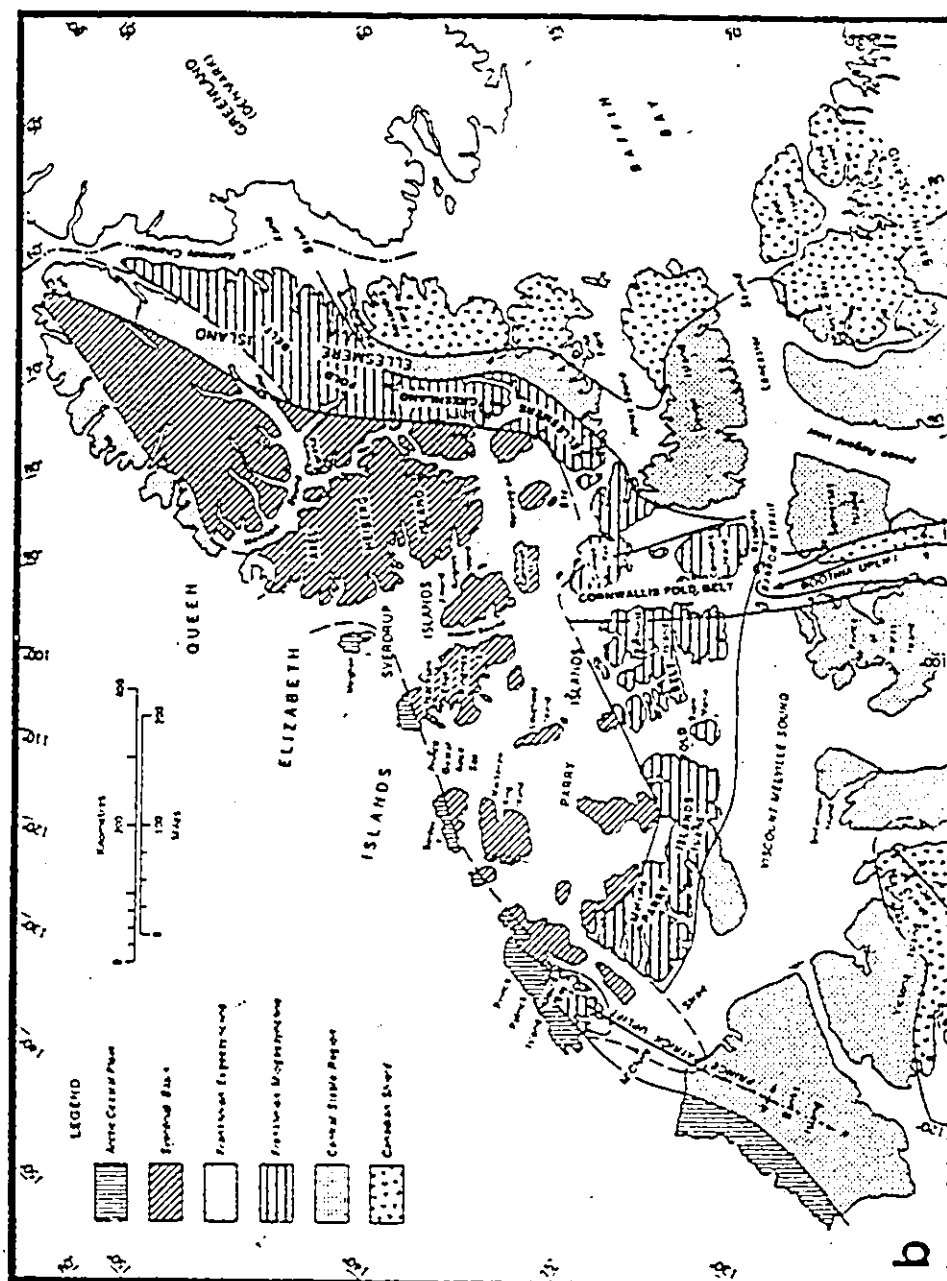


FIG. 3-2a Geological provinces of central arctic archipelago.

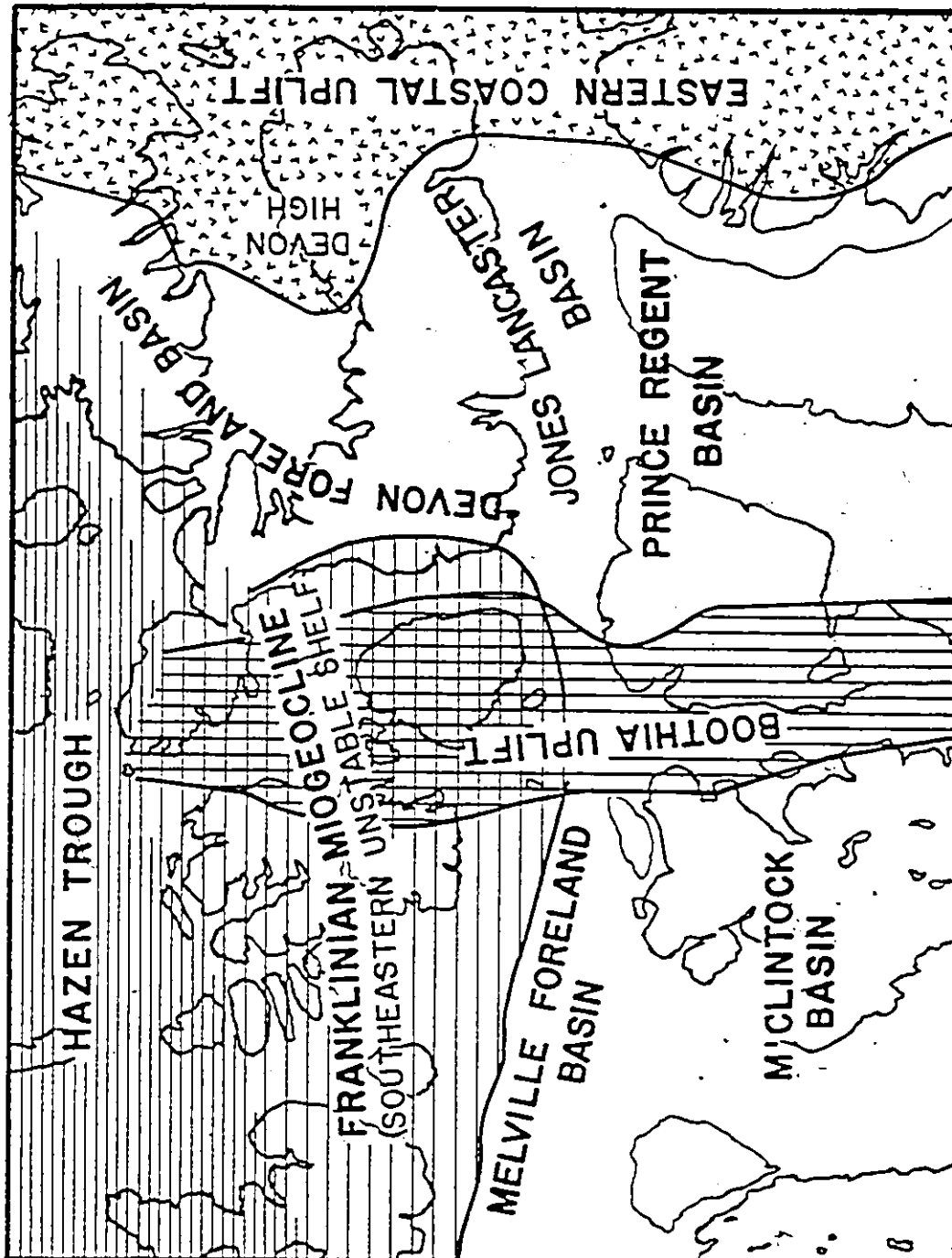


FIG. 3-2b Tectono-depositional elements of central arctic archipelago.

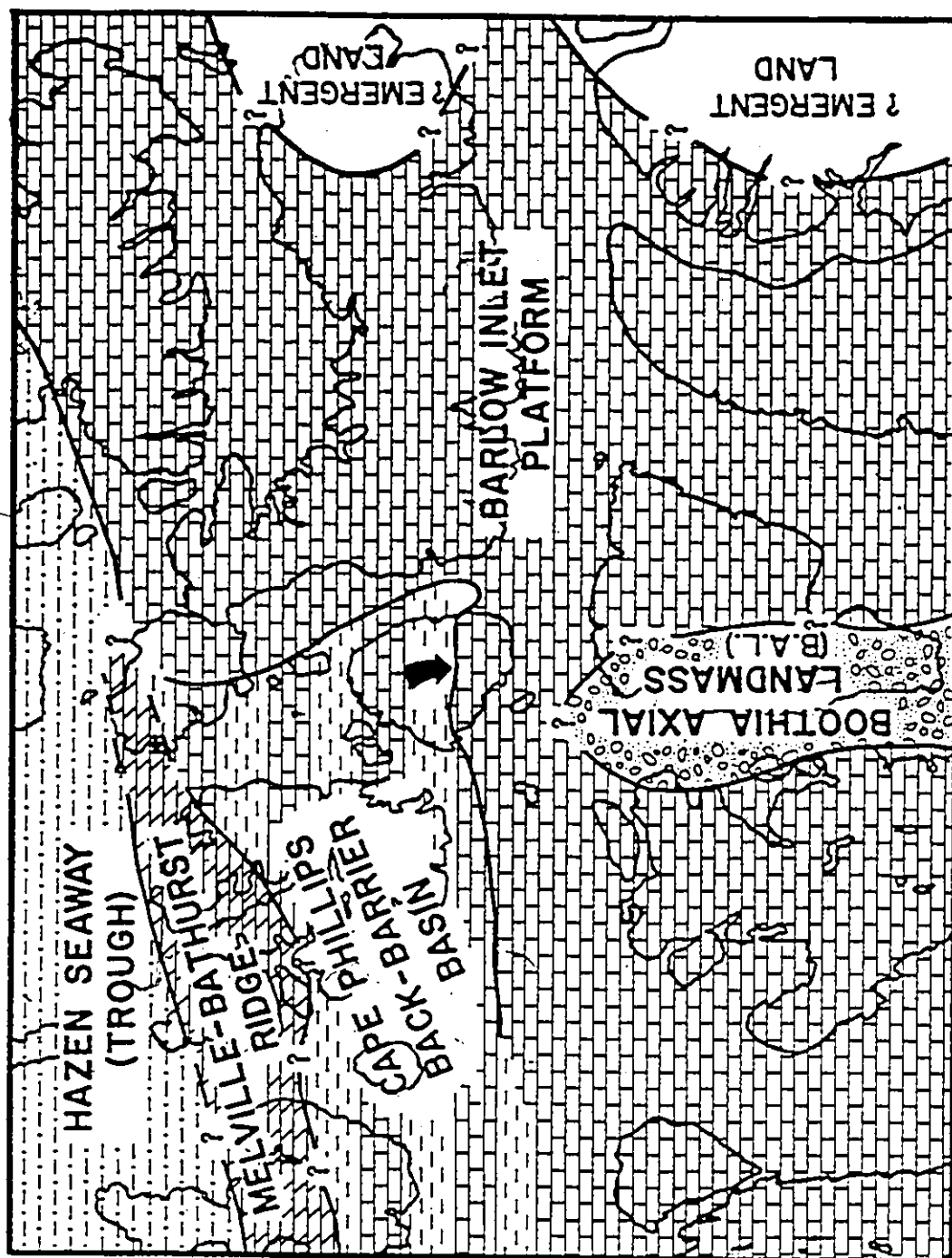


FIG. 3-3 Siluro-Devonian paleogeographic elements of central arctic archipelago.
 → CASBLFC (Cape Atry-Snowblind Bay line of facies change)

(Miall, 1970a,b; Gibling, 1978; Miall and Gibling, 1978). An extensive peripheral marine transitional zone (tidal flat) developed, prograded northward and encroached upon the area now represented by southern Cornwallis Island. The resulting stratigraphic record of tidal flat-shallow subtidal shelf carbonate sedimentation accumulated during rapid basin subsidence and ongoing epeirogenesis, and consists largely of vertically-stacked, shallowing-upward hemicycles.'

Historically, the formal recognition of sedimentary cycles dates back to work by Phillips in the first half of the 19th century (see Weller, 1964, p.12). With regard to the study of cyclicity in carbonate rocks, Dixon and Vaughan (1911) contributed one of the pioneering investigations, but sophisticated facies analyses of regressive shallow-water carbonate sequences were made only following the elucidation of Holocene peritidal environments in the late 1950's and early 1960's (Fischer, 1964; Aitken, 1966; Roehl, 1967; Laporte, 1967).

Cyclic shallow marine carbonates have now been described from a legion of stratigraphic levels and geographic localities (see Ginsburg, 1975). Whereas previously cyclic sequences were considered to be relatively rare, the

' The rhythmic successions of sediment discussed in this paper are referred to as the 'Barlow Inlet' hemicycles. This is done for brevity, as the hemicycles occur in both the Somerset Island and Barlow Inlet Formations where these formations intertongue on southern Cornwallis Island (see Paper 5, this thesis).

pendulum of geological doctrine has reversed, and now it is the absence of cyclicity more than its presence that demands explanation. The observation by Twenhofel (as quoted by Duff et al., 1967, p.1) that, "Essentially all deposition is cyclic or rhythmic", is far less a radical generalization now, than it was in 1939 when originally penned. As stated by Wilson (1975, p.43), "One of the most interesting stratigraphic results of carbonate study in the last twenty years has been the recognition that most thick limestone and dolostone sections consist of cyclically repeated strata". Nonetheless, although examples of cyclic carbonate sequences have been well-documented, the actual mechanism responsible for the vertical stacking of successive shallowing-upward cycles has remained controvertible or problematic. Furthermore the literature contains a paucity of research on Silurian shallow carbonate shelf, hemicyclic sequences.

The intent of this paper is to describe the various Barlow ~~inlet~~ hemicycle lithofacies, define a preferred vertical cycle succession, provide a paleoenvironmental synthesis, and discuss mechanisms responsible for the asymmetric cyclicity of the sequence.

3.1.1 Terminology

The terminology applied to cyclic sedimentation and sedimentary units has grown both in volume and ambiguity (see discussion by Duff et al., 1967, p.3-8). It is useful

to consider the term cycle in both concrete ("real") and abstract ("constructed") modes. The following terms and their meanings shall be observed in this paper (acceptable synonyms appear in parentheses).

series -- a continuous sequence of superposed cycles (term from Phillips, 1836).

real cycle -- a cycle that occupies real space and physically occurs in a rock succession.

constructed cycle -- a cycle formulated by induction or chance, but not physically existing in the rock record.

cycle -- a rhythmic succession of the ABCDCBA type.

asymmetric cycle (hemicycle) -- a rhythmic succession of the ABCDABCD type.

modal (typical) cycle -- a real cycle that occurs with great (or greatest) frequency.

idealized (model) cycle -- a constructed cycle based on statistical manipulation, theoretical consideration, or other devices, that strives to approximate a probable normative succession of lithofacies.

composite cycle -- any cycle, real or constructed, that contains additional or fewer cycle members than those defined in the modal or idealized cycle descriptions.

complete cycle -- a real cycle judged to contain a full complement of normal cycle members, arranged in logical sequence.

cycle member (member, element) -- a discrete stratigraphic interval of singular lithofacies aspect, that forms the basic unit of a cycle. Strictly speaking a member is characterized by a particular lithofacies (i.e. the latter is an attribute of the former), and hence the terms are not synonymous. However, 'member' and 'lithofacies' are used in this text interchangeably, for reasons of practicality.

normal member -- a member found in a modal or idealized cycle.

cryptic cycle -- a cycle whose individual identity cannot be distinguished from subjacent or suprajacent cycles.

cryptic cycle coalescence -- the joining of two or more cycles, that results in misidentification of this amalgamated product as a single thick cycle.

retrenched cycle -- a real cycle that is reduced in thickness or complement of members, or both, as the result of post-depositional erosion. (Generally in association with ravinement during the transgressive phase of the ensuing cycle).

aborted (incomplete) cycle -- a real cycle whose terminal member(s) was/were never deposited due to a reinitiated transgression.

infratidal -- shallow subtidal (in the sense of Roehl, 1967). Zone below mean low water mark; exposure index probably less than 17 (see Hardie, 1977).

subtidal -- Zone below spring low water mark, exposure index probably less than 10.

peritidal -- including shallow subtidal, intertidal and supratidal.

3.1.2 General Characterization of Barlow Inlet Hemicycles

The importance of cyclic shallow marine sediments within the Upper Silurian of the shelf area is graphically portrayed in FIG. 3-1. On Cornwallis Island peritidal facies occur within sections from the southeastern tip of the island (Cape Hotham, loc. 3), north to the line of facies change separating shelf from basinal depositional environments (Sophia Lake, loc. 8). The exposed Barlow Inlet Formation thickens northward, and only at locs. 7 and 8 do sections contain both lower and upper contacts of the formation. Successions of apparently well-developed hemicycles are most completely exposed and accessible in sections at Cape Hotham and Barlow Inlet (logs. 3 and 4 respectively) and the ensuing discussion is based largely on

data collected from these two localities. However the thickest series (560 m) of hemicyclic, peritidal sediments occurs at loc. 7 and 8, but here the continuity of outcrop is poor.

Individual complete hemicycles range in thickness from less than 1 m to 23 m, but all effectively represent discrete transgressive/regressive events. Cycle thickness variation is attributed largely to differential shelf subsidence or cryptic cycle coalescence. All principal hemicycle lithofacies are fine-grained (generally muddy). Calcilutites and calci/dolosiltites predominate, and with few exceptions, sand-size comminuted skeletal debris or other allochems are rare. The carbonates are relatively pure, with some cycle members exceptionally so. Coarse silt-sized terrigenous detritus is present, but never in excess of 5 %. Macrofossil diversity is generally low, although certain cycle members contain large numbers of individuals. Euryhaline and metahaline forms predominate.

Complete Barlow Inlet hemicycles contain a basal, higher energy, shallow subtidal carbonate member, and a terminal, low energy intertidal member, features held in common by many shallowing-upward (or shoaling upward) sequences described in the literature (e.g. James, 1978; Ginsburg, 1975). Shallowing-upward cycles are asymmetric in the sense that the transgressive depositional record is severely diminished, and most of the cycle represents sedimentation during relative sea-level fall.

The Barlow Inlet shallowing-upward cycles can be divided into two broad groups, each characterized by a particularly distinctive or predominant hemicycle member. These two groups are the Restricted Infratidal (RI) Hemicycles and the Tidal Flat (TF) Hemicycles. Their distribution is shown in FIG. 3-1. Most Barlow Inlet cyclic sequences can be assigned with confidence to one group or the other, but as the principal characteristics of each group are not mutually exclusive, some hemicycles appear to be transitional between the two categories.

3.2 RESTRICTED INFRATIDAL HEMICYCLES

The RI hemicycles contain up to 4 members or lithofacies, but are rarely complete and usually are typified by just two, a Biostromal Member and a Massive Carbonate Mudstone Member. The massive nature and relative resistance to weathering of these two predominant members, allow the RI series to form substantial cliffs (PL. 3-1g), that, where dolomitized, can be confused with core facies of large carbonate buildups. The other two members are a Sparsely Fossiliferous Mottled Dolomitic Carbonate Member and a Laminated Member. All four members and their variants are described below.

3.2.1 Biostromal Member

3.2.1.1 Field Appearance

The basal high(er) energy member of an RI hemicycle, is typically a stratiform, slightly argillaceous floatstone, with fossils of colonial metazoans making up 10 to 50% of the unit by volume (PL. 3-1a,b,c).

Where undolomitized, these biostromes are commonly rubbly weathering at their base and increasingly massive towards the top. Bedding is generally poorly-developed, ranging from thin wavy to nodular, to massive. Member thickness ranges from .3 to 7.7 m, and averages 2.8 and 2.2 m at localities 3 and 4 respectively. Individual biostromes are laterally extensive, and apparently constant in thickness

over distances of 50 to 200 m, although the gradational nature of the upper contact makes this difficult to confirm. Ninety percent of basal contacts observed are abrupt, but in the absence of scours, or of incorporation of clasts of the underlying lithology, lower contacts cannot be considered erosional.

3.2.1.2 Lithofacies Description

This member's identifying property is the abundance of its contained fossil fauna, discussed in detail below. The enclosing lithology is rubbly argillaceous limestone type II (Jones et al., 1979) mudstone to wackestone with relatively low clay content (see Narbonne, 1981). Even in principally limey hemicyclic sequences, both the matrix and carbonate layers or nodules are appreciably dolomitized (PL. 3-1d,e). This partly obscures original sediment texture, but apparently considerable sandy and rudaceous bioclastic material was set in a finer-grained matrix.

Many hemicycles at loc. 4 have been intensely dolomitized and Biostromal Members are represented by massive fine-grained dolomite containing large hemispherical vugs.

3.2.1.3 Macrobiota and Biogenic Structures

Rare Chondrites are the only discrete bioturbation structures in the Biostromal Member. Hemispherical stromatoporoids are predominant, although lamellar and

irregular growth forms are also common, as are algal oncoids. Hemispherical, cylindrical and subspherical favositids are less abundant, and solitary and fasciculate rugosans as well as dendroid stromatoporoids are rare (PL. 3-1d,f). Conspicuous by their rarity in, or absence from, most biostromal units are crinoid debris and shelly fauna. Of 104 colonial fossils observed in one biostrome (loc. 3), 59% were in growth position, 12% were apparently overturned, and the remaining 29% were indeterminate. Of these fossils, 89% were hemispherical to subspherical stromatoporoids, and 11% were favositid corals. Solitary and fasciculate rugosans were minor constituents, and a single cerioid rugosan was observed. The proportion of upright colonials appears to increase away from the base of the biostrome. Although variable, the predominance of stromatoporoids and of upright colonies can be considered typical of the biostromal units.

Preferred vertical zonations of faunas are not consistently developed, although oncoids, where present, rarely occur above 1 m from the base. The number and size of metazoan colonies tends to decrease upward. Irregular fenestrae are rarely developed in the Biostromal Member (PL. 3-1f).

3.2.2 Sparsely Fossiliferous Mottled Dolomitic Carbonate Member

3.2.2.1 Field Appearance

This member is generally unbedded, massive, and weathers a distinctly mottled medium grey and buff to grey-brown (PL. 3-2a,b). Units appear stratiform and parallel-sided, are laterally extensive, and average less than 1 m thick. Both basal and upper contacts are variable (PL. 3-2a,b,d).

3.2.2.2 Lithofacies Description

This lithofacies consists of discrete lozenge-shaped to highly irregular and interconnected lime mudstone nodules set in dolomitic matrix, and can be classified as Type I and labyrinthine mottled dolomitic limestones (Jones et al., 1979). The nodules are composed of peloidal lime mudstone to wackestone. Micritic subcircular peloids, 30 to 100 microns in apparent longest dimension, are separated by a mosaic of irregular, blocky spar with crystals 5 to 40 microns in size (PL. 3-2c). Peloid-spar contacts are either sharp or diffuse, which does not indicate whether the spar is an aggrading neomorphic product, or a recrystallized inter-allochem cement (see discussion on grumeleuse texture by Bathurst, 1975). The peloids are not easily discriminated and commonly, patches of micrite exhibit no peloidal texture. Bioclasts are generally recrystallized. Identifiable fossil allochems include fragments of ostracodes, gastropods, and brachiopods.

Solution seams (PL. 3-2f) are intensely dolomitized (PL. 3-2e). The dolomite is roughly equigranular and subhedral. Corroded unidentifiable allochems make up to 50% of the matrix. Sand size or larger terrigenous clasts are rare.

The distinction between the Mottled Dolomitic Carbonate Member and 'faintly-mottled' Massive Carbonate Mudstone Member is arbitrary where mottling of the former is less distinct. Where these lithofacies are juxtaposed with gradational contact, this distinction is particularly vague (see PL. 3-2a).

3.2.2.3 Macrobiota and Biogenic Structures

No discrete biogenic structures have been identified. The simple presence of mottling is not taken here to imply intense bioturbation (e.g. Braun and Friedman, 1969), although this remains a possibility. Body fossils are common to very rare, and include upright hemispherical to subspherical stromatoporoids; dendroid stromatoporoids, lamellar stromatoporoids, algal oncoids, Coenites, hemispherical favositids, solitary rugosans, high-spired gastropods, articulated ostracodes, and smooth spiriferid brachiopods. Stromatoporoids and algal oncoids predominate.

3.2.3 Massive Carbonate Mudstone Member

3.2.3.1 Field Appearance

The characteristic member of the RI shallowing-upward cycle is massive, largely structureless and unfossiliferous, carbonate mudstone with several variants or subfacies of basically similar field appearance.

Units of this member are massive and resistant and without primary bedding surfaces. However, some units show conspicuous, parallel, stylolitized fractures that mimic bedding, and that may be controlled therefrom (PL. 3-3a,b). The carbonate mudstone lithofacies commonly displays conchoidal fracture, attesting to a high degree of textural homogenization (PL. 3-3c). This lithofacies typically weathers uniform light to medium-dark grey. A medium grey-light buff faint mottling is rare, but detectable in all variants of this member (e.g. PL. 3-3f), and is consistently associated with one particular subfacies (PL. 3-8a,c).

Within the limits of field observation (<200 m along strike), unit thickness does not vary significantly. The significance of very slight pinch and swell of at least one unit of this lithofacies (PL. 3-3e), is discussed below in 'Environmental Synthesis'.

Of 25 basal contacts observed, most are gradational. The upper contact is variable (52% abrupt, 48% gradational). Units assigned to this lithofacies range from .2 to 20.1 m

in thickness, averaging 1.3 m for loc. 3, and 4.7 m for loc. 4.

Three principal subfacies constitute the Massive Carbonate Mudstone Member. These include:

1. Dense Mudstone Subfacies
2. Loferite Subfacies
3. Megalodont Bivalve Subfacies

Only the last subfacies contains macrofauna in any appreciable abundance.

3.2.3.2 Dense Mudstone Subfacies

This subfacies is composed of featureless micrite (PL. 3-4a) with subordinate diffuse patches of microspar (PL. 3-4c), and up to 5% scattered euhedral to subhedral dolomite rhombs (15-50 microns). Disseminated, fine-grained pyrite is an ubiquitous accessory mineral. Some polished sections show faint mottling (PL. 3-4b) of uncertain origin. The pattern suggests either intense sediment reworking by organisms, or complex zones of relative oxidation associated with degrading organic matter. However, neither acetate peels nor thin sections reveal textural differences or concentrations of disseminated sulphides at mottle boundaries.

Microstylolites are invariably associated with this subfacies, and in combination with non-sutured seam solution, produces distinctive fitted fabrics (PL. 3-4e) (Buxton and Sibley, 1981).

Fossils are rare to very rare and are almost exclusively gastropods (PL. 3-4f), amphiporids (PL. 3-4d) and articulated and single valves of ostracods (leperditids) (PL. 3-4g). No trace fossils have been discovered.

3.2.3.3 Loferite Subfacies

Field observations suggested that between 30 and 70% (depending upon the study locality chosen) of massive carbonate mudstone units from the RI hemicycles, exhibit some fenestral fabric (PL. 3-5a). Subsequent examination of polished sections, acetate peels and thin sections has indicated that, where fenestrae comprise 5% or less of cross-sectional area, they were consistently overlooked in the field. Hence the relative proportion of non-fenestral to fenestral carbonate mudstones is probably higher than indicated. Other than the presence of fenestrae, the Loferite Subfacies is identical with the Dense Mudstone Subfacies. Fenestrae vary considerably in orientation, shape, and type and proportion of infilling material (cements and internal sediment), and can best be systematically described in terms of these variables. The abundance (density), size, degree of interconnection, and proportion of types of fenestrae are all attributes of the fenestral fabric.

Most fenestrae have complex cross-sectional outlines that defy simple description or classification. Schemes that

have been developed, by necessity greatly generalize the shape of fenestrae, belying their complex, essentially intergradational nature. By convention more than relative applicability, the scheme adopted here follows Logan (1974), who established a simple tripartite classification for fenestrae, that has been subsequently utilized by others investigating birds-eye structures (e.g. Grover and Read, 1978). It is based on shape (effectively cross-sectional outline) and orientation of fenestrae to bedding (original horizontality). Logan (1974) recognized the following types:

1. laminoid -- fenestrae with irregular lensoid shapes, and oriented parallel or subparallel to bedding.
2. tubular -- tube-shaped fenestrae, simple or bifurcating, and oriented perpendicular to bedding.
3. irregular -- fenestrae subspherical to highly irregular in outline, and with no preferred orientation.

Polished sections exhibiting a predominance of each 'end-member' fenestral type, are shown in PL. 3-5b,c,d. More typically (e.g. PL. 3-5e) the Loferrite Subfacies requires a regrettably subjective application of Logan's tripartite classification. Of 52 randomly selected, oriented, polished slabs of massive carbonate mudstone, 30 showed some degree of fenestral development. Of these only eight (5 laminoid, 3 tubular) had a majority of fenestrae

with recognizable and consistent orientation. Twelve samples consisted exclusively of irregular fenestrae, and the remaining 10 contained principally irregular fenestrae mixed with the other types. In apparent width/length measurements, irregular fenestrae average .5/1.0 mm and attain maximum values of 15/20 mm. Tubular fenestrae have average widths of .5 mm, and do not exceed 3 mm across. Laminoid fenestrae have width/length measurements averaging 1.0/4.0 mm, but some exceed 20 mm horizontally. The vast majority of fenestrae are thus of fine void size, and rarely of medium void size (Logan, 1974, Table 2). In general fenestrae on any one slab are of similar size, regardless of the mixture of types, with dimensions of voids generally not varying more than one order of magnitude.

Internal geopetal sediment commonly floors fenestrae (PL. 3-5h). Its distribution is highly variable, but it appears to be preferentially associated with laminoid and tubular fenestrae. FIG. 3-4, shows marked decrease in void-filling sediment where fenestral area exceeds 25%. Usually where fenestrae are pervasive, they contain no particulate fill. Similar observations have been made by Grover and Read (1978, p.465). Four types of internal sediment have been identified, and include, 1) micrite-microspar, in either 1 or 2 generations (PL. 3-5g), 2) micrite-microspar with abundant euhedral dolomite rhombs 30-50 microns in size (at low magnification,

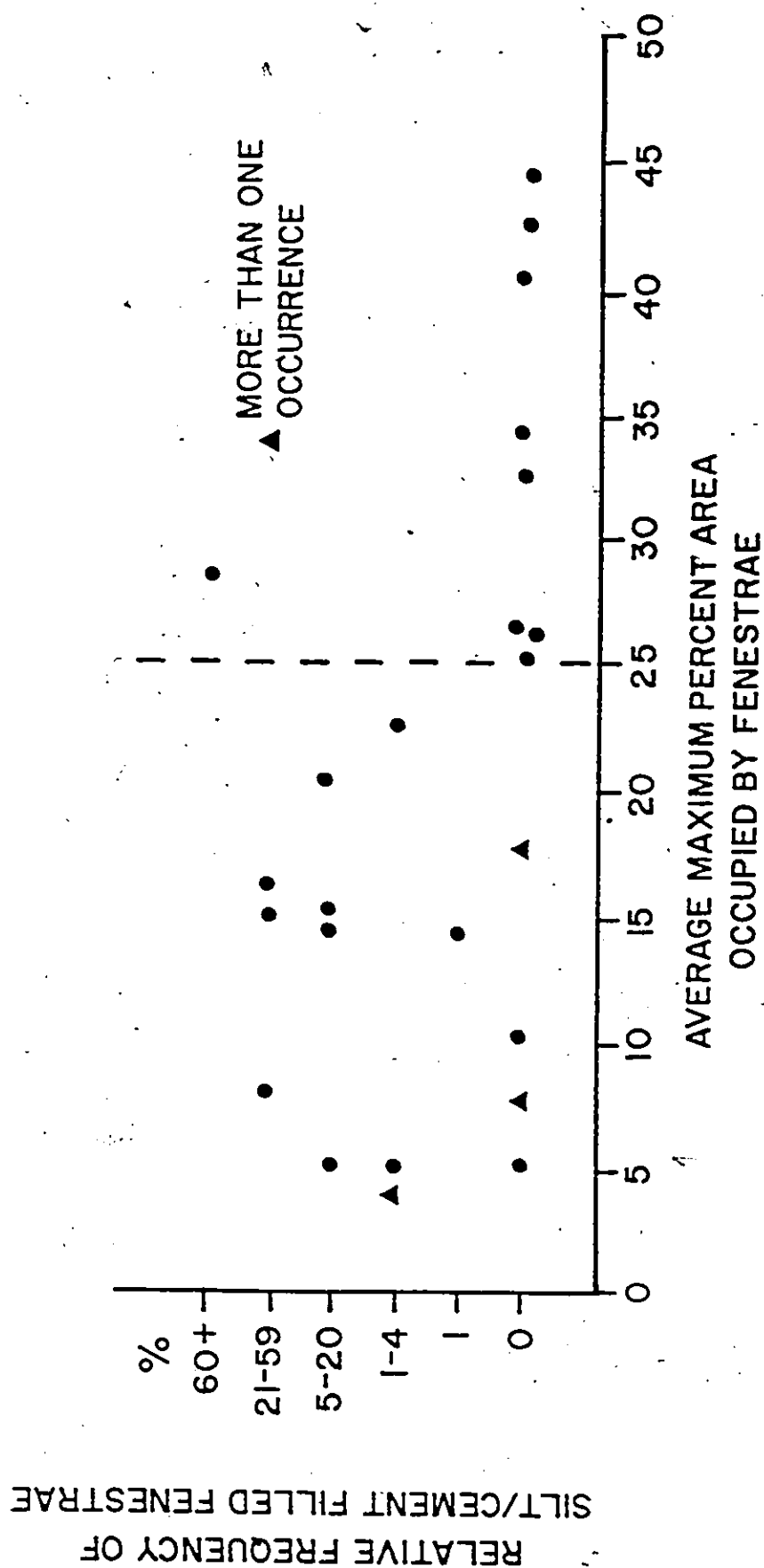


FIG. 3-4 Distribution of internal sediment in fenestrae.

od 826 such sediments appear to be uniform calcisiltites, but are in fact texturally bimodal), 3) intra(bio)micrite, generally in very large fenestrae or possible fractures, and 4) a mosaic of intergrown, equant, 30-50 micron dolomite crystals (PL. 3-5f). Inasmuch as the fourth void-filling material does not have a 'clastic' texture, its interpretation as a sediment remains dubious. It is retained in this category because it appears to have formed secondarily, after a precursor particulate sediment, as suggested by remnant wavy-lamination (PL. 3-5f). The dolomitic mosaic bears some similarity to the crystal or vadose silt described by Dunham (1969) (i.e. sediment of hypidiotopic equant, silt-size crystals that is different texturally from surrounding matrix, is overlain by blocky cement, lacks recognizable bioclasts, and commonly has a non-horizontal upper surface and traction-current structures such as wavy or cross-lamination). The dolomitic mosaic differs, however, in the following important aspects:

1. it is dolomite not calcite.
2. it does not overlie an isopachous encrustation of early fibrous calcite cement.
3. it has a crystal size (30-50 microns) that is coarser than the median values for Dunham's (1969) vadose silts (10-25 microns).

The most common geopetal sediments are micrite-microspar and dolomitic mosaic. Total occlusion of void space by sediment is rare.

The Lofelite Subfacies exhibits a variety of void-filling cements (Table 3-1). The majority of fenestrae are filled by simple clear, coarse, equant non-ferroan calcite (PL. 3-6a,b). This cement abuts and conforms to the irregular fenestral walls, and clearly postdates emplacement of internal sediment. Other cement fabrics are well-represented, and often co-occur in similar fenestral-types within a single sample. Centripetal coarsening of spar (drusy cements of many authors) occurs not only in intra-allochem voids (PL. 3-6h), but is also developed to varying degrees in fenestrae (PL. 3-6d). The resulting spectrum of cement fabrics does not lend itself to reliable deciphering of cement stratigraphy and discrimination of cement generations. Many fenestrae exhibit contiguous splays of coarse bladed spar normal to pore walls, with no significant centripetal coarsening (PL. 3-6e). These cements occur circum-fenestrally, but are not isopachous. In other instances bladed-spar hangs preferentially from pore-roofs (pendant cements), whereas fine equant spar develops on sediment-filled fenestral floors (PL. 3-6c). Are these crystal morphologies truly geopetally-influenced, or are they affected to a greater degree by original substrate mineralogy or crystallinity?

Table 3-1

Void-filling cements and internal sediments of Lofelite subfacies.

ORIGINAL SEDIMENT	LITHIFIED INTERNAL SEDIMENT	PROCESSES AND INTERMEDIATE PRODUCTS DOL.-dolomitization NEO.-neomorphism	CEMENTS OR ALTERED SEDIMENTS
			pyrite (9) coarse, equant, clear, calcite spar(8) coarse, equant, dolospar (7) medium, bladed clear spar (6) (in places pendant) centripetally-coarsening clear calcite spar (5) fine, equant, clear, calcite spar (4) coarse, equant, Fe- dolospar (3) mosaic of fine equant dolomite(2)
calcite silt --->	intramicrite (D) calcisiltite	-(DOL.)	
calcite mud --->	micrite(B)	-bimodal mixture of micrite and fine dolomite rhombs (C) (DOL.) -microspar (NEO.)	inclusion-rich, in part drusy, pseudospar (1)

	A	B	C	D	1	2	3	4	5	6	7	8	9
A		x	x	x	x	x	x	-	x	x	x	x	x
B	x		-	-	x	-	-	x	-	-	-	x	-
C	x	-		-	x	-	-	-	-	-	-	x	-
D	x	-	-		-	-	-	-	-	-	-	-	-
1	x	x	x	-	-	-	-	-	x	-	-	x	-
2	x	-	-	-	-		x	-	-	-	-	x	-
3	x	-	-	-	-	x		-	-	-	-	x	-
4	-	x	-	-	-	-	-	-	-	-	-	x	-
5	x	-	-	-	x	-	-	-	-	-	-	-	-
6	x	-	-	-	-	-	-	-	-	-	-	x	-
7	x	-	-	-	-	-	-	-	-	-	-	x	-
8	x	x	x	x	x	x	x	x	-	x	x		x
9	x	-	-	-	-	-	-	-	-	-	-	x	

A = micrite/microspar
matrix

x = direct contact
between void-
filling materials

The abrupt replacement of fine equant spar with coarse spar, and the presence of poikilitic textures, suggests discontinuous cementation leading to final occlusion of fenestral pore-space. Conversely, the centripetally-coarsening spars suggest relatively continuous cementation in adjacent pore spaces. Many fenestrae are surrounded by zones of inclusion-rich, fine to coarse, non-equant, calcite spar (PL. 3-6c,f). Were it not for its remarkably sharp contact with the surrounding micrite matrix, and its ubiquitous association with fenestrae, it could be considered a neomorphic fabric. It appears to predate the deposition of internal sediments (PL. 3-6c). Its origin remains enigmatic; perhaps it represents an original cryptocrystalline aragonitic cement that has been subsequently recrystallized. Rarely, coarse, equant, ferroan-dolospar precedes coarse equant calcite spar, although more commonly non-ferroan coarse, equant dolospar post-dates the precipitation of centripetally-coarsening spar.

Spar and sediment-filled solution molds of gastropods (PL. 3-6g) are common in the Lofelite Subfacies. Patches of idiomorphic equant dolomite are associated with stylolites, and have the same characteristics as the dolomitic mosaic sediment. Scattered individual euhedral dolomite rhombs form an integral, if minor part (<1%) of the matrix. Pyrite is a late-forming cross-cutting mineral phase that occurs in

trace quantities (PL. 3-7a). Stylolites are ubiquitous, and some result in fitted fabrics (PL. 3-7e).

The matrix of the Loferite Subfacies generally has a structureless micritic texture (PLS. 3-6b, 3-7a), or is rarely intraclastic. Peloids cannot be discriminated within the matrix, although in intimate association with fenestrae are spherical micrite bodies ranging in apparent maximum diameter from 50 to 2000 microns (PL. 3-6b,c,d). These are identical to the entities described by Fischer (1975) as "peds" (pedological term for the basic structural unit or particle of soils), and their formation may be related to circum-granular cracking. The pervasive development of fenestrae and peds partially transforms a calcilutite mudstone (or micrite) into a pseudo-peloidal grainstone (pelsparite). Rare fossil allochems include gastropods and ostracodes. Exceptionally, the Loferite Subfacies includes Amphipora wackestone (PL. 3-7f).

3.2.3.4 Megalodont Bivalve Subfacies

A less common, but nonetheless distinctive and significant variant of the Massive Carbonate Mudstone Member, is the Megalodont Bivalve Subfacies, typified by the occurrence of large megalodont bivalves. In some respects it differs strikingly from either the Dense Mudstone or Loferite Subfacies, for the Megalodont Bivalve Subfacies is fossiliferous; typically faintly to distinctly mottled, and

has a matrix rich in recrystallized allochems. It is included within the Massive Carbonate Mudstone Member for the following reasons:

1. it forms massive units, principally of fine-grained carbonate,
2. locally it has a fenestral fabric, and most importantly,
3. invariably it is stratigraphically bounded by units of either of the other two subfacies.

Approximately 10% of the RI hemicycles contain units of the Megalodont Bivalve Subfacies. Together these units account for only 3% of cumulative RI hemicycle thickness, although they may constitute a significantly larger proportion (up to 20%) of individual complete hemicycles. Typically the subfacies consists of concentrated articulated large, thick-shelled bivalves in a lime mudstone matrix (PL. 3-8a,b). Shell lengths average 8.5 cm and range between 5 and 15 cm. Population densities reach an observed maximum of 50 per square metre. The bivalves are invariably extensively recrystallized. Three dimensional specimens cannot be liberated from their matrix, and this prevents precise identification. They are tentatively assigned to Megalomoidea, the only Silurian megalodont bivalve genus.

Articulated valves of Megalomoidea are most commonly oriented with their beaks down, and the commissure plane roughly perpendicular to paleohorizontal (PL. 3-8a). Single

valves appear to be preferentially concave-up (PL. 3-8e). Articulated valves are randomly scattered throughout massive units, and valve pairs are not concentrated on, or parallel to discrete bedding planes. The orientation and distribution of articulated valves suggest an infaunal to semi-infaunal mode of life.

The mottling commonly shown by the Megalodont Bivalve Subfacies is the result of patchy distribution of dolomite (lighter-colored areas in PL. 3-8a,c). The matrix is composed of biopelmicrite, biopelsparite (PL. 3-8f), and micrite. Recognizable bioclasts comprise shell fragments, whole fossils of ostracodes, gastropods, bivalves, brachiopods, and indeterminate algae (PL. 3-8g), and sponge spicules. Bioclasts are generally recrystallized, and most are unidentifiable. Textures range from wackestone to grainstone. Peloids are distinguishable in packstones and grainstones, where they appear poorly sorted (40-800 microns), but cannot be discriminated in mudstone and wackestone matrices.

Macrofauna recorded from field observation of 19 units with Megalomoidea indicate that 56% of the units also contain hemispherical to globose stromatoporoids of small to moderate size (less than .3 m across), 31% contain small to large high-spined gastropods (murchisonids), and 13% have ostracodes (mostly leperditids). Less commonly associated faunal elements include lamellar stromatoporoids,

hemispherical favositids, solitary rugosans and crinoid debris.

Most basal contacts are abrupt but non-erosional, and most upper contacts are gradational. The Megalodont Bivalve Subfacies is very rarely seen to directly succeed stromatoporoid biostromal units, and is never overlain by laminites.

3.2.4 Laminated Member

3.2.4.1 Field Appearance

The Laminated Member generally weathers recessively due to its inherent fissility. This character is not developed universally, and massive variants or subfacies are present rarely. One of the principal slip-planes responsible for the Cape Hotham rockslide, is contained within this lithofacies (PL. 3-13c). Weathering colors are uniform and range from medium grey to buff, depending upon the degree of dolomitization.

Sixty seven percent of basal contacts of this member are gradational, whereas 70% of its upper contacts are abrupt. Due to its recessive nature, this lithofacies is difficult to trace laterally; no significant thickness variation shows on oblique air photos over distances of 150 m. It is assumed that this lithofacies forms a relatively extensive sheet-like body. Thicknesses average .5 m at both locs. 3 and 4 and the Laminated Member is present in 43% of real cycles of RI aspect.

3.2.4.2 Lithofacies Description

In field exposures and polished sections the laminae reflect a) compositional variations (PL. 3-9a), generally mm alternations of dolomite-rich and dolomite-poor laminae, or b) presence of parallel, closely-spaced stylolites (PL. 3-9c), or c) alignment of laminoid fenestrae (PL. 3-9d), or d) some combination of the above (e.g. PL. 3-9b). Frequently stylolites and/or laminoid fenestrae are superimposed on the compositional-textural laminae, producing second-order layering between 3 and 10 mm thick. Commonly fenestrae and stylolites clearly exploit inherent planes of weakness at boundaries between compositional (primary?) laminae. In other instances, no compositional layering can be detected, and fenestrae, or more commonly, stylolites, alone define the location (or former presence) of laminae.

Desiccation polygons are rare in this member within the RI hemicycles; more commonly bedding planes are smooth and structureless, or show minor papillae (PL. 3-9i) or crenulae that coalesce and form intricate irregular to zig-zag patterns (PL. 3-9f). The widths of the crenulae do not exceed 2-3 mm. Less commonly, subparallel, very low amplitude crenulae occur (PL. 3-9j). In cross-section laminae tend to be continuous, planar to somewhat crinkly, and generally parallel to unit boundaries. Minor discontinuities and irregularities in lamination appear to be more frequent where laminae are associated with a

fenestral fabric (PL. 3-9b), but occur as well in non-fenestrate rocks (PL. 3-9e,g). Rarely, laminae are gently domed over lateral distances of several 10's of cm, but discrete stromatolite structures are not apparently present (one possible exception is figured in PL. 3-9h).

In thin section, compositional laminae are composed of either a) subhedral, idiotopic dolomite (and minor terrigenous detritus) (PL. 3-10a), or b) amorphous organic and/or clay-rich material (PL. 3-10c), or c) a mixture of calcite microspar (or possibly silt-sized bioclasts) and dolomite rhombs. Calcite is usually subordinate to dolomite (PL. 3-10a). Couplets of a-b or a-c laminae produce the observed compositional layering, but the three laminae types are rarely associated in the same sample. Calcareous and/or organic rich laminae range between 50 and 300 microns in thickness, and average 150 microns. Dolomitic laminae are generally thicker, but rarely exceed 500 microns (PL. 3-10d,e). Dolomite grains making up the laminae average between 30 and 50 microns in apparent longest diameter (i.e. coarse silt), but some samples of this lithotype contain dolomitic laminae whose constituent grains are very fine sand-size (60-75 microns). Terrigenous detritus of fine to coarse silt-size rarely exceeds 10% of total sediment volume, and may represent wind-blown material (e.g. Schenk, 1967). In thin section, most laminae boundaries are very diffuse (PL. 3-10a), and recognizable only at low

magnification. However in some samples of laminite with a-b couplets, contacts are sharp (PL. 3-10c,e).

Overall the characteristics of this lithofacies are in accord with descriptions of ancient laminites in the voluminous literature on peritidal carbonates (e.g. Schenk, 1967; Laporte, 1967; Gebelein and Hoffman, 1974).

3.2.4.3 Macrobiota and Biogenic Structures

No body or trace fossils have been identified in the Laminated Member within RI hemicycles. Similar lithofacies from the TF hemicycles contain rare algal oncoids, ostracodes, indeterminate spiriferids and bivalves. Biogenic disruption of laminae is rare, but discrete trace fossils are present and include the genera Planolites, Sinusites and Chondrites. At loc. 7, comparable laminites principally contain rare ostracodes and comminuted ostracoderm debris, along with lebensspuren such as Skolithos and Polarichnus.

3.2.5 Analysis of Succession

In certain sequences of the Upper Silurian carbonates on Cornwallis Island, the limited number of lithofacies (4) tend to occur in a recognisable, ordered succession, bearing considerable resemblance to shallowing upward carbonate sequences described by James (1978). Limits to the stratigraphic extent of these individual partial or complete hemicycles were based on the following assumptions:

1. Lithofacies are not repeated within a single cycle.
2. The cycles are asymmetric.
3. The base of each cycle tends to start with the highest energy lithofacies (most open marine character), with succeeding lithofacies exhibiting progressively more restricted faunas, and more pervasive presence of phenomena indicative of subaerial emergence.
4. Erosional or abrupt contacts bound hemicycles, gradational contacts do not.
5. The lithofacies have been somewhat generalized to accommodate minor variations (e.g. the subfacies of the Massive Carbonate Member are not treated separately).

Tabulation of hemicycles defined on this basis are presented in Table 3-2. A modal cycle consisting of two lithofacies (A and C) accounts for 48% of all hemicycles identified. Eighty two percent of all hemicycles are

represented by just four ordered combinations of lithofacies (A-C, C-D, A-C-D, A-D-C), none of these containing all four lithofacies. The single complete cycle (A-B-C-D) bears little resemblance to the modal hemicycle, (A-C), and the relationship between all four lithofacies remains ambiguous. Markov Chain analysis was employed to elucidate the relationship. Miall (1973) and Jones (1975) presented useful summaries on the application of Markov chain analysis to sedimentary successions. The method used herein is the 1st order (or embedded) Markov chain analysis, which emphasizes transitions between lithofacies, and disregards the relative proportion of the succession that each lithofacies represents.

The results of the application of Markov analysis are shown in FIG. 3-5, and indicate that transitions A-B, B-C, C-D, and D-A occur with greater than random frequency. Modified Chi-squared testing (after Bonham-Carter) suggests rejection of the Null hypothesis (that the transitions are random) at the 95% confidence level. An appropriate constructed model cycle is therefore ABCD. Transitions tend to be gradational except at the bases of hemicycles, where the contacts are either abrupt or, rarely, erosional.

Hemicycle statistics

TF Hemicycles			RI Hemicycles		
Observed lithofacies sequences	# cycles	% total # cycles	Observed lithofacies sequences	# cycles	% total # cycles
abcd	4	14	abcd	1	3
abc	1	4	abc	1	3
abd	1	4	ab	1	3
ab	1	4	acd	4	14
acd	1	4	acb	1	3
ac	8	31*	ac	14	49*
ad	1	4	adc	2	7
a	1	4	bcd	1	3
bcd	3	12	cd	4	14
cdc	1	4		29	99
cd	3	12			
dc	1	4			
	26	101			

Mean TF Hemicycle thickness				Mean RI Hemicycle thickness			
locality	Different Measures			locality	Different Measures		
	I	II	III		I	II	III
3	2.7	3.9	3.0	3	4.3	6.4	3.0
4	3.6	5.2	5.7	4	6.9	7.5	8.4
7	12.0	16.9	12.7	7	12.2	19.1	-
8	6.4	6.4	6.4	8	8.1	12.0	-

* Modal hemicycle

I. total of all real cycles/# of real cycles

II. sum of average thickness for each lithofacies

III. total of all complete cycles/# of complete cycles

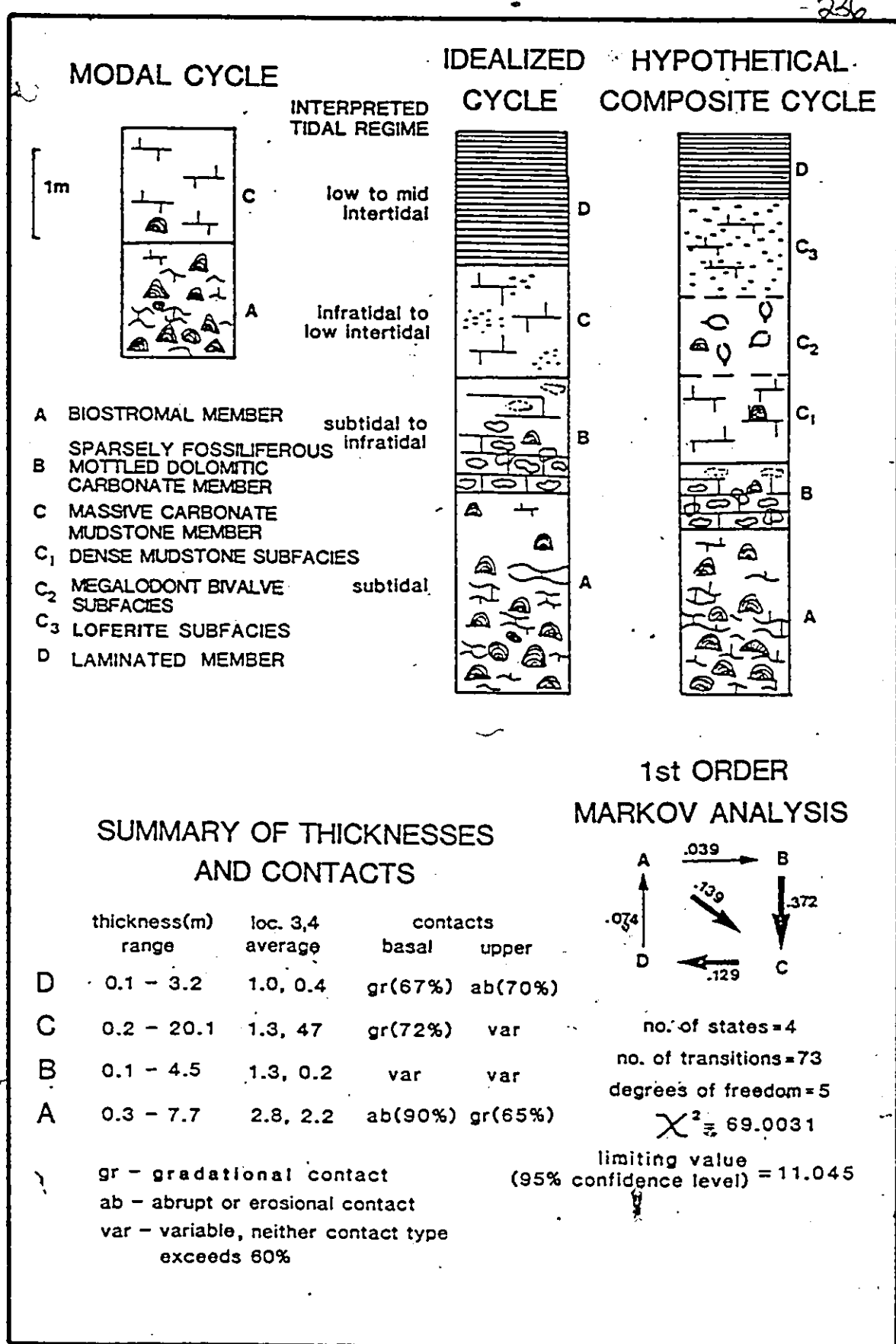


FIG. 3-5 Restricted Infratidal hemicycle model.

3.2.6 Calibration and Facies Interpretation

Empirically, based on complete hemicycles, and statistically, based on Markovian tendencies of lithofacies transitions, it is possible to reconstruct a model shallowing-upward cycle or norm (e.g. FIG. 3-5). The correct interpretation and environmental significance of members of the model depend on accurate calibration of some part of the cycle to a known datum, in this case sea-level. As stated by James (1978, p.127), "The thread that binds all such [shallowing-upward] sequences together is the presence of the distinctive intertidal unit, which, once recognized, allows one to interpret the surrounding lithologies in some kind of logical sequence."

In RI hemicycles there are few unambiguous criteria with which to differentiate supra, inter- and infratidal deposits. The result, unfortunately, is dependence on absence of features to delineate environmental zones.

3.2.6.1 Laminated Member

The Laminated Member presents such a problem. The laminae are almost certainly cryptalgal as indicated by lack of traction current structures, presence of surface papillae and crenulae, fenestrae and desiccation polygons, dolomite, small-scale disconformities, as well as other characteristics outlined by Aitken (1967) and Gebelein and Hoffman (1974). Absence of major surface structures (e.g.

blisters, domes), the fine laminoid fenestrae, and the stratiform nature of units of the Laminated Member, suggest that the cryptalgal fabric developed in association with growth of a smooth mat (Logan et al., 1974).

Many authors have placed similar cryptalgal facies in the supratidal zone (e.g. Laporte, 1967; Braun and Friedman, 1969), perhaps in part due to overzealous application of the Andros Island (humid climate) tidal flat analog where well-laminated and algal mat-colonized sediments occur in supratidal levee backslopes of tidal creeks (Shinn et al., 1969), and intertidal sediments are bioturbated. In the past 10 years, more thorough evaluation of different Holocene carbonate tidal flat environments has resulted in recognition of extensive deposits of algal-laminated sediments in various littoral settings (Hoffman, 1973). For example, such deposits are found in the upper intertidal (e.g. Trucial Coast, Snider, 1975). As well, smooth algal sheets extend "...continuously for distances of several km's along shore..." in the lower intertidal zone of Shark Bay, western Australia (Logan et al., 1974, p.157), and less extensive algal mats occur in the shallow subtidal zone (e.g. Bermuda, Gebelein, 1969).

The position of algal-laminated sediment in the littoral range reflects the presence/absence of invertebrate grazers (principally cerithid gastropods in the Holocene), which are in turn controlled by relative exposure and water salinity.

Hypersalinity, due to arid climate and hydrographic conditions (basin geometry), provides a barrier to potential grazers and allows algal mats to colonize intertidal and infratidal substrates.. The original adaptive radiation of burrowers and grazers, even into less stressful stenohaline to metahaline intertidal environments, may not have taken place until the mid-Paleozoic. As a consequence, "In Precambrian and lower Paleozoic deposits the characteristic tidal flat features such as desiccation polygons, well-laminated sediments and fenestrae will occur at the base of the intertidal zone" (James, 1978, p.130).

An arid to semi-arid climate is indicated by evidence from other nearly contemporaneous stratigraphic units within the region (see 'Environmental Synthesis' below). Despite the inferred aridity, such features as extensive polygon development, solution breccias, evaporite minerals or pseudomorphs thereafter, salt casts, teepee structures, and patterned carbonate are absent. Calcrete, intraclastic deposits, and extensive vadose diagenetic features are also not developed. In the absence of these diagnostic supratidal criteria, the laminated member is considered to have been deposited in the low intertidal to possibly mid-intertidal zone. The common presence of dolomitic lamination probably reflects selective dolomitization of Mg-enriched algal mat layers (Gebelein and Hofmann, 1974), and has no synsedimentary spatial significance (unlike

penecontemporaneous supratidal protodolomites of Shinn et al., 1965). No evidence was found to support a detrital origin for the dolomite, as has been suggested for dolomite in contemporaneous and environmentally-similar peritidal deposits on Somerset Island (Gibling, 1978). Tidal flat sediments are predominantly derived from the adjacent lagoon, and hemicycle facies of interpreted subtidal origin contain no dolomite that is not clearly due to secondary recrystallization (i.e. associated with solution seams, or as isolated euhedral rhombs) of the calcareous matrix.

3.2.6.2 Dense Mud Subfacies

The Dense Mud Subfacies contains no truly diagnostic environmental structures or allochems. The presence of articulated ostracodes and whole gastropods set in homogeneous medium grey lime mudstone compares with shallow subtidal sediments (adjacent marine) of the Bahamas described by Shinn et al. (1969). The presence of disseminated pyrite suggests incomplete oxidation of organic matter, which might be anticipated in subtidal sediments. Lack of bedding or lamination is likely the result of bioturbation, although the original depositional condition of these fine grained sediments may not have produced any primary layering (i.e. continuous rain of clay-sized carbonate).

3.2.6.3 Loferite Subfacies

The Loferite Subfacies is considered to have been deposited in infratidal to low intertidal environments.

This conclusion is based on the following observations:

1. it tends to occur below the laminated facies, which is thought to have been deposited in the low to mid-intertidal zone.
2. it is commonly medium to dark grey, suggestive of incomplete oxidation of original organic matter, more typically a subtidal than an intertidal attribute.
3. it contains (albeit rarely) a marine benthic fauna consisting of Amphipora, gastropods and ostracodes.
4. it is intimately interbedded with, and save for fenestrae, identical with, the Dense Mudstone Subfacies which is interpreted to be infratidal in origin.
5. the Megalodont Bivalve Subfacies also (rarely) shows a fenestral fabric, and is considered to be subtidal.

Whether the fenestrae themselves formed in a subaqueous environment is a moot point. Shinn (1968) advanced the hypothesis, that birds-eye structures are supratidal indicators, but subsequent research suggests that fenestrae do not exclusively form, nor are they preserved, in supratidal environments (e.g. Logan, 1974; Grover and Read, 1978). Additionally, Ginsburg (1975, p.198) has stated that "... there is reason to suspect that fenestral voids

(birds-eye loferites) may develop in subtidal sediments", a contention cautiously supported by other works on different fenestral types (e.g. Grover and Read, 1978; Flügel, 1982, p. 488, Zamarreno, 1975). Although different fenestrae-types reported in the literature are feasibly polygenetic, evidence from the Loferite Subfacies does not suggest independent origins for its fenestrae of different shapes and orientations.

The simplest explanation for the formation of all fenestral types of the Barlow Loferite Subfacies, is as a result of shrinkage of a cohesive sediment. Orientation of the fenestrae will reflect the fabric of the sediment (e.g. presence of laminae or not) and perhaps the confining pressure (e.g. depth of burial). The size and degree of fusion of fenestrae could be the result of the intensity, duration, or repetitive nature of the process creating the pores. Isotropic textures will produce irregular fenestrae or a mixed assemblage of fenestrae. Anisotropic textures will yield laminoid fenestrae. Tubular fenestrae will develop in accordance with predominantly horizontal tensile stress. Fenestral pores of different shapes and orientations, can be compared approximately to larger scale structures like prism cracks and sheet cracks, that have fundamentally different orientations but are due to the same process (shrinkage). Compare fig. 5 of Grover and Read (1978), fig. 7 of Read (1973), and PL. 3-5e (this paper)

with figs. 10, 11, 13, and 20 of Fischer (1964) and note the essential similarity of all fenestrae and the inevitable subjectivity involved in categorizing them.

No preferred stratigraphic sequence of fenestral-types was observed (e.g. Havard and Oldershaw, 1976), nor is there clear evidence of an order in pore genesis (e.g. Grover and Read, 1978). A linear relationship between fenestral density and proportion of area occupied by fenestrae, suggests that increased fenestral porosity reflects increased numbers of pores, rather than enlargement of pre-existing pores (FIG. 3-6). Maximum cross-sectional fenestral pore size does not show a preferred relationship with fenestral density. These facts imply either that a finite limit to fenestral size existed, or that formation of fenestrae was a relatively short-term process, and fenestral density was related to the intensity of this process, rather than to repeated exposure to it. Bleached rims and pyritized halos, similar to those described by Read (1973) and Havard and Oldershaw (1976) surrounding tubular fenestrae, rarely envelop fenestrae of the Loferite Subfacies. Havard and Oldershaw (1976) interpreted these as supratidal features, although the formation of pyrite in sediment due to reducing conditions created by organic decay is more common in the subtidal zone (e.g Logan, 1974, p.242). The bleached zone may reflect oxidation of reduced iron by slow flushing

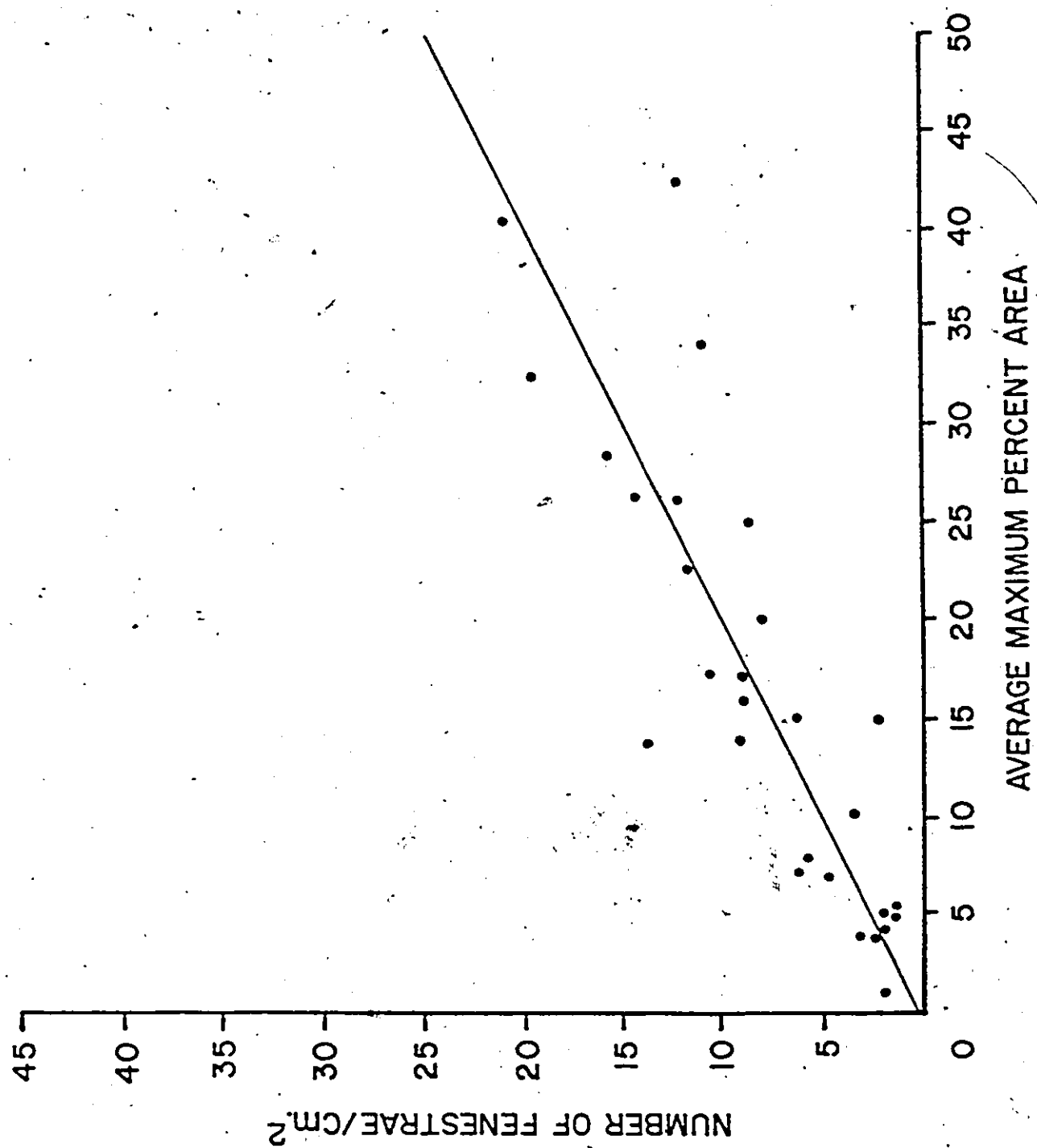


FIG. 3-6 Plot of abundance versus area occupied by fenestrae.

of fenestral pore systems with oxygenated marine water, driven by tidal currents. The preferred occurrence of internal sediment in samples of low fenestral density remains enigmatic, particularly if the sediment is a vadose feature (Dunham, 1969) where one would anticipate increased not decreased numbers of fenestrae. The absence of internal sediment in samples with high fenestral density does suggest that these pore systems developed rapidly, and did not remain subject to infiltration for long.

In conclusion, the fenestrae are thought to have developed by shrinkage. Apart from the fine laminoid fenestrae, they are not obviously associated with algal mats. Shrinkage most likely was accomplished by desiccation, although synaeresis cannot be ruled out. The shape/orientation of fenestrae is thought to have been controlled principally by relative anisotropy of the sediment, and predominantly extensional stress in the sediment. Sediments of the Loferite Subfacies were likely initially deposited subtidally, and later subjected to desiccation (by mechanisms discussed below). Alternatively the fenestrae may have developed in the subtidal environment by 1) synaeresis, 2) expansive subtidal algal mat growth, 3) organic decay and gas bubble generation, or 4) some combination of the above.

Some diagenetic features of the Loferite Subfacies do suggest vadose diagenesis (e.g. dissolution molds of

gastropods, pendant cements). The superimposition of vadose diagenetic fabrics on subtidal sediments is possible by invoking mechanisms that do not necessitate major eustatic sea-level drops, or tectonic uplift of the platform. These mechanisms include the following:

1. Progradation of the tidal flat wedge concomitant with shoreward downdrop of the tidal ground water table, which can place the subtidal sediments in the vadose zone (the meteoric phreatic zone may be absent or considerably diminished in an arid climate).
2. Vertical ground water table fluctuations due to tidal, seasonal or longer term changes (Grover and Read, 1978).
3. Periodic lowering of sea-level during solar-tidal cycles (e.g. 18.6 year cycle discussed by Wells and Coleman 1981; exposure of Florida Bay mudbanks semi-annually, Turmel and Swanson, 1976).

3.2.6.4 Megalodont Bivalve Subfacies

The Megalodont Bivalve Subfacies is considered to be subtidal on the basis of an associated in-situ fauna of globose to hemispherical stromatoporoids, tabulate and rugose corals amongst other forms. Portions of larger massive units which contain Megalomoidea also locally exhibit well-developed fenestral fabric. This intimate association between megalodont bivalves and loferite facies

has been documented by Fischer (1964, 1975) and Mountjoy (1975) in Triassic and Devonian shelf successions respectively, and suggests a shallow subtidal or infratidal habitat for the clams. Kobluck (1975) found a statistically significant association between encrusting stromatoporoids, Amphipora, and megalodont bivalves. Amphipora in particular, is regarded as diagnostic of Devonian distal back-reef (sublagoonal) facies. Sodero and Hobson (1979) reported megalodont bivalves from Silurian proximal back-reef sediments in Arctic Canada.

In contrast to this infratidal, back-reef interpretation, several authors (e.g. Klován, 1964; Trettin, 1979; Savelle, 1979) contended that megalodont bivalves are indicative of reef, proximal fore-reef, and/or high energy shoal environments. Perhaps this latter interpretation is due to their large size and gregarious nature which results in shell concentrations casually reminiscent of rudistid and oyster banks or biostromes. Indeed some authors contend that the rudistids evolved from the Megalodontidae (see Dechaseaux, 1969).

Megalomoidea itself has generally been regarded as having inhabited reef or shoal environments. Boucot (1975, p.256) stated that "the reef-associated Megalomus (= Megalomoidea) Community (is) a quiet water entity as shown by the commonly articulated valves". 'Quiet' and 'reef' are not mutually exclusive environmental descriptors by any means, but known

intracratonic Silurian bioherms (or biohermal stages) of this aspect do not contain Megalomoidea. Similarly, despite the occurrence of numerous and diverse carbonate buildups in the Barlow Inlet Formation, Megalomoidea is not part of the fauna of the buildups or associated flank beds.

The megalodont bivalves of the Barlow Inlet Formation are here considered to have led a semi-infaunal existence in a moderately calm-water, shallow lagoonal setting. These conclusions are based on the following reasons:

1. Megalomoidea is an equivalved, uncoiled, unattached bivalve that is associated with fine-grained sediment, forming floatstone (never framestone) texture. As such, these bivalves and their thanatocoenosis are strikingly dissimilar to rudistids or ostreids, and their deposits.
2. the synecology of Megalomoidea itself is not known, but comparisons to the closely related genera Megalodon and Conchodon suggest a semi-infaunal mode of life with beaks down and hinge line close to vertical (Zapfe in Fischer, 1964). The absence of rugae or costae on Megalomoidea supports a semi-infaunal existence for this genus, although its rather inflated shell suggests that it had a shallow burrowing capability. The relatively thick shell may have been more a means of protection against predatory gastropods (whose remains are commonly

associated with this clam), than an adaption to turbulent water conditions (see Kaufman, 1969).

3. the associated restricted fauna is not suggestive of open marine conditions (complete absence of stenohaline forms).
4. the fine-grained matrix and the absence of traction current structures are not suggestive of high energy conditions.

The rejection of a shelf-edge or proximal fore-slope environment for megalodont bivalves, follows Playford (1969) who concluded that such clams found in fore-reef settings are allochthonous, and were ultimately derived from a restricted back-reef biocoenose. Megalomoidea is associated with reef or platform complexes (term usage follows Burchette, 1981), but more specifically this bivalve occurs in the restricted shelf lagoon of the platform interior.

3.2.6.5 Sparsely Fossiliferous Mottled Dolomitic Carbonate Member

The Sparsely Fossiliferous Mottled Dolomitic Carbonate Member contains a limited fauna of restricted subtidal aspect. Algal oncoids and stromatoporoids (hemispherical and stick-forms) predominate. Algal oncoids are generally considered to represent "...morphological and ecological adaptation of encrusting organisms to soft-bottom and/or low energy environments" (Flügel, 1982, p.137). The member's principal characteristic is its mottled appearance, which

probably reflects early submarine lithification of peloidal muds to form nodules (Jones et al., 1979). This fabric has been subsequently modified by pressure seam solution and dolomitization (Wanless, 1978). In many respects the mottled dolomitic carbonate facies is intermediate between the biostromal and massive carbonate facies. The biostromal facies can also exhibit mottling, although this feature is often obscured by the abundance of massive stromatoporoids and the rubbly argillaceous nature of this member. The absence of distinct mottling in the Loferite and Dense Mudstone Subfacies remains problematic, but may reflect differences in initial porosities (micrites versus calcisiltites to very fine calcarenites), sedimentation rates, and/or degree of lithification prior to dolomitization. Although the mottling is a diagenetic phenomenon, it was apparently governed by some pre-existing environmental constraints. The Sparsely Fossiliferous Mottled Dolomitic Carbonate Member was probably deposited shoreward of the Biostromal Member in shallower, more restricted conditions, resulting in a decreased biota.

3.2.6.6 Biostromal Member

The Biostromal Member contains the greatest abundance and faunal diversity of any facies within the RI hemicyclic sequences. However, the fauna is impoverished in stenohaline and open marine organisms. The biostromes are

dominated by hemispherical to subspherical (globose) stromatoporoids. Both growth forms inhabited a variety of platform complex environments, ranging from fore-reef to back-reef (Jamieson, 1971; Kobluk, 1975; Burchette, 1981), and are thus not particularly facies-diagnostic. The relatively small massive stromatoporoids, the paucity of laminar forms and the local co-occurrence of dendroid stromatoporoids suggest a back-reef environment (Burchette, 1981; Tsien, 1971). Average concentrations of colonial metazoans is low, and the biostrome structure is therefore sparse to dense (terminology after Riding, 1981). The biostrome is essentially similar to other biostromes described in Devonian cyclic shelf interior sequences (Laporte, 1967; Read, 1973; Havard and Oldershaw, 1976; Kasig, 1981; Burchette, 1981). At the seaward edge of the lagoon the biostromal facies probably merged with stromatoporoid-frame biostromes, stromatoporoid-crinoid rubble buildups, and crinoid shoals that constituted the platform margin magnafacies (see Paper 4, this thesis).

3.3 TIDAL FLAT HEMICYCLES

Beneath the stratigraphic interval that embodies the RI hemicycles at locs. 3 and 4. there are carbonate sequences (61 and 40 m thick respectively), consisting of distinct shallowing-upward cycles that lack (or have diminished) Biostromal and Massive Carbonate Mudstone Members. In their stead, the calci/dolosiltite Laminated Member is more fully developed, and the hemicycles commence with an Oncoidal Carbonate Member. Further comparative characteristics of these rhythmic successions are as follows:

1. facies specific dolomitization is more pervasive than in the RI hemicycles, and dolomite mottling is fainter but more consistently developed.
2. the hemicycles are generally thinner than those of Restricted Infralittoral affinities.
3. a massive calci/dolosiltite member is present, whose origin is apparently unrelated to that of the massive carbonate mudstones of the RI hemicycles.
4. not only is the laminite facies proportionately thicker, but it is commonly composed of coarser carbonate and siliciclastic grains, and desiccation features are more prevalent.

In light of the importance of the Laminated Member and its paleoenvironmental interpretation. (discussed later in the text), these shallowing-upward cycles are called Tidal Flat (TF) hemicycles. The differences between the TF and RI

hemicycles outlined above, should not obscure the fact that they are broadly similar in many aspects of petrology and structure. Consequently, the following descriptions, largely emphasize different and distinctive aspects of each lithofacies.

3.3.1 Oncoidal Carbonate Member

3.3.1.1 Field Appearance

The oncoidal carbonate lithofacies is generally massive, non-bedded, structureless, mottled carbonate. Mottling is either faint or distinct, and is a product of the contrast between light grey weathering calcareous fine-grained nodules (in three dimensions) and enclosing buff-weathering matrix (PL. 3-11a). Mottling also results from the appearance of abundant, dark brown-grey weathering oncoids in the lighter colored dolomitic matrix, yielding a pattern described as 'organic mottling' (PL. 3-11a,b). Lithofacies units representing this member appear laterally continuous with limited thickness variation within the limits of field observations and outcrop extent (generally less than 20 m). Basal contacts are almost always abrupt or less commonly erosional, but rarely gradational. Upper contacts tend to be gradational. Average member thickness varies in different study areas. At loc. 3 average member thickness is .6 m, and at loc. 4 the same member is 2.0 m.

3.3.1.2 Lithofacies Description

The characteristic constituents of this member are algal oncoids. The oncoids are cyanophycean macro-oncoids (average greater than 1 cm), with moderately poorly defined concentric laminae (from .5 to 2 mm in thickness), porostromate texture (PL. 3-11d), and predominantly type C (PL. 3-11b) to rarely type R shape (PL. 3-11c) (Flügel, 1982). Oncoids range from .5 cm to 5 cm in apparent maximum diameter. Their shape is principally inherited from the oncoid nuclei, which are either peloidal lime mudstone lithoclasts or individual bioclasts (PL. 3-11f). Size and shape of nuclei vary considerably (see PL. 3-11b). Individual laminae are differentiated on the basis of relative proportions of micrite to microspar (algal filament molds).

Oncoids rarely constitute more than 5% of this lithofacies. The bulk of the member is composed of irregular to lenticular, faint to distinct calcareous nodules surrounded by dolomitic solution seams (Wanless, 1979). The proportion of dolomitic zones to nodules ranges from approximately 1:4 to 4:1, but dolomitic zones generally make up around 60% of the lithofacies. The nodules are composed of homogeneous micrite (pelmicrite) and biomicrites with mudstone and wackestone textures (PL. 3-11i). Local zones within the nodules are biopelsparite (PL. 3-11h). Peloids range from 100 to 3000 microns, and are undoubtedly

polygenetic (micritized bioclasts, rounded intraclasts, fecal pellets). Some peloids appear to have a diffuse microspar core, suggesting centripetal aggrading neomorphism; others show sharp micrite-spar contacts suggestive of micrite envelopes and arrested micritization of bioclasts. Other recognizable allochems include fragments of ostracode and mollusk shells. The dolomitic solution seams contain up to 50% solution-corroded bioclasts and peloids set in a matrix of equant subhedral to euhedral dolomite crystals (10-25 microns), and other insolubles forming a resistate fraction. Silt to sand-sized terrigenous clasts are negligible.

3.3.1.3 Macrobiota and Biogenic Structures

The predominant fossils apart from cyanophyte oncoids, are dendroid stromatoporoids (cf. Amphipora), small to large high-spined gastropods (PL. 3-11e), and ostracodes (PL. 3-11g). Rare fragments of cyathaspidid ostracoderms, small, irregular to hemispherical stromatoporoids and favositids, fasciculate rugosans, and smooth spiriferid brachiopods are also present. The degree of bioturbation in these sediments is difficult to assess. The mottling does not appear to reflect burrow patterns. Discrete trace fossils include Chondrites and Sinusites.

3.3.2 Mottled Carbonate Member

3.3.2.1 Field Appearance

The Mottled Carbonate Member is broadly similar to the Oncoidal Carbonate Member in appearance, except that oncoids are fewer, and hence 'organic mottling' is not developed. Bedding tends to be massive, although nodular bedding is present rarely where surfaces of more argillaceous units have been extensively weathered. Generally mottling is faint (PL. 3-12a,b), and the boundaries between calcareous nodules and dolomitic matrix are often diffuse. Mottled dolomitic limestone Type I (discrete mottles) predominates (56% of units); Type II mottled dolomitic limestone (semi-continuous micrite layers) is subordinate. The relative percentage of nodules to matrix, and nodule size vary considerably. Units of this member do not vary in thickness, or in lithofacies characteristics along strike within the limits of accessible outcrop. Member thicknesses at locs. 3 and 4 are .7 and .6 m, respectively. Basal contacts are 70% gradational, 30% abrupt, and upper contacts are over 80% gradational.

3.3.2.2 Lithofacies Description

Nodules are composed of pelmicrosparite (PL. 3-12e) with well-sorted subspherical micritic peloids averaging 125 microns. The texture is suggestive of an original pelmicrite packstone. Sharply defined, if somewhat

irregularly shaped patches of non-peloidal micrite (PL. 3-12b) represent either intraclasts or burrow fills. Horsetail stylolites (anastomosing solution seams) surround and penetrate the nodules, isolating mudstone clots at their margins (PL. 3-12a,d). Subhedral dolomite crystals in the solution seams are generally equigranular and 20 to 75 microns in apparent maximum diameter. Contrary to observations made by Wanless (1979) on nodular limestones, there is no appreciable difference in dolomite crystallinity or size between the nodules and solution seams. Allochems other than peloids and intraclasts include rare ostracode and gastropod shell fragments and indeterminate recrystallized bioclasts (PL. 3-12c). Preserved sedimentary structures are not abundant. Rare samples of faintly mottled carbonate exhibit both patterned carbonate structure (PL. 3-12g) and sheet cracks coated by non-isopachous, discontinuous rinds of fine, equant calcite (possible meniscus cement ;PL. 3-12f). More commonly faint lamination and desiccation polygons are associated with this lithofacies.

3.3.2.3 Macrobiota and Biogenic Structures

Oncoids, dendroid stromatoporoids and ostracods are the most abundant and consistently present organisms. Smooth spiriferids, small high-spired gastropods, ostracoderms, solenoporids, and small irregular to subspherical

stromatoporoids and favositids complete the faunal list. A diverse assemblage of trace fossils is present in this lithofacies. Chondrites is particularly abundant, along with forms such as Planolites, Uchirites, Skolithos, Bifungites, Helicodromites, and Sinusites.

3.3.3 Massive Carbonate Member

The Massive Carbonate Member is represented by one or both of two (unrelated?) lithofacies. The more common is a massive, unfossiliferous light to medium grey weathering calcilutite mudstone, similar to the dense mud subfacies of the RI hemicycles. The reader is referred to the more complete description of this subfacies given previously. The second lithofacies, discussed in detail below, is a massive light grey-buff to orange-buff-weathering dolomitic unfossiliferous calci/dolosiltite.

3.3.3.1 Field Appearance

Units of the member are unbedded, or exhibit apparent bedding 7 to 20 cm thick due to horizontal fractures (PL. 3-10f,g). Very weakly-developed lamination is evident locally in some units. Generally the weathering color of this lithofacies is uniform, although faint mottling is apparent in some units. Member thicknesses average 1.5 m for both locs. 3 and 4 (this figure is based on either or both lithofacies). Lateral pinch and swell or facies change

within individual units has not been observed. Sixty eight percent of lower contacts and 70% of upper contacts are gradational.

3.3.3.2 Lithofacies Description

The origin of the massive nature of this lithofacies (PL. 3-10f), is apparent in thin sections and acetate peels. Originally flat or undulatory cryptalgal laminae have been intensely disrupted (PL. 3-10b). Apart from this disruption, the lithofacies is similar to the Laminated Member. No bioclasts were identified. Desiccation polygons are associated with this lithofacies in 15% of units studied at loc. 3 and 4.

3.3.3.3 Macrobiota and Biogenic Structures

A very sparse fauna (and flora) associated with this member includes oncoids, gastropods, indeterminate bivalves, ostracoderms, smooth spiriferids, and pelmatozoans (indicated by a single pluricolumnal). Chondrites is the only trace fossil recorded from this unit.

3.3.4 Laminated Member

This lithofacies is developed in the TF hemicycles substantially more than in the RI hemicycles (see Table 3-2), but otherwise its characteristics are largely similar.

Desiccation polygons are more frequent in this lithofacies in the TF hemicycles than in the RI hemicycles, although this may be an artifact of the relative availability of exposed bedding planes. The faunal/floral components of this member have been discussed previously in conjunction with the RI hemicycles. Member thicknesses range from .2 m to 3.8 m, and average .9 m and 1.1 m at locs 3 and 4 respectively.

3.3.5 Analysis of Succession

The TF hemicycles were subjected to the same subjective and statistical appraisal applied to the RI sequences, to determine the preferred order of lithofacies succession (model cycle). The results are summarized in FIG. 3-7.

Whereas complete cycles make up only 3% of all real cycles in the RI series, complete cycles constitute 16% of the TF hemicyclic succession. Markov analysis indicates the following non-random transitions (at the 95% confidence level): A-B, B-C, C-D, D-A. An appropriate constructed model cycle is therefore ABCD. Intracycle transitions are gradational; intercycle contacts tend to be abrupt or erosional.

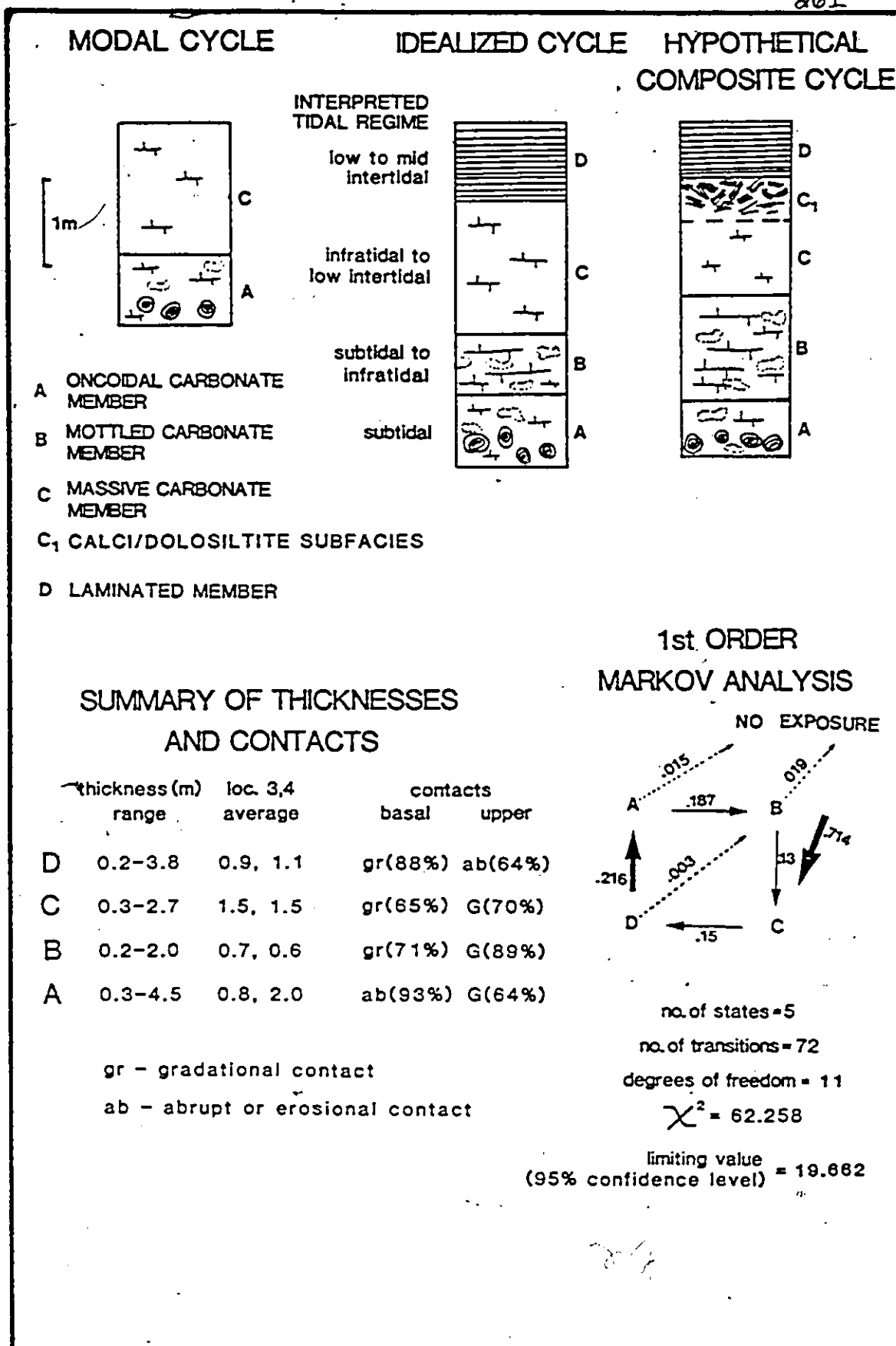


FIG. 3-7 Tidal flat hemicycle model.

3.3.6 Calibration and Facies Interpretation

Interpretation of the TF Laminated Member follows that of RI hemicycle laminites, which are considered to have been deposited in the low intertidal, to possibly mid-intertidal zone. The massive calci/dolosiltites stratigraphically beneath the laminites were clearly produced by disruption of a pre-existing laminated sediment (e.g. by displacive precipitation of gypsum, or by bioturbation). In the absence of pseudomorphs after evaporite minerals, or evidence of solution collapse, the disrupted texture is considered to have resulted more likely from biogenic action in the lowest intertidal to infratidal zones. The massive calcilutite, similar in all respects to the dense mud subfacies of the RI hemicycles, is considered to have been deposited in infratidal environments. The absence of loferite subfacies in the TF hemicycles remains, at present inexplicable.

Interpretation of the Mottled Carbonate Member is difficult because the mottling is a post-depositional effect of unknown relationship to the original sedimentary environment. The author follows Jones et al. (1979) in suggesting that the nodules formed due to synsedimentary submarine and substratal cementation. Unlithified portions of the matrix were subsequently subjected to intense seam solution (cf. Wanless, 1978). The presence of a restricted, lagoonal fauna supports a subtidal origin for these

sediments. That they were periodically if rarely exposed subaerially, is suggested by the presence of patterned carbonate, sheet cracks and possible meniscus cements.

The Oncolitic Carbonate Member represents the highest energy facies of the hemicycles, and was deposited in the relatively deepest environments, most distal from the paleoshoreline. The contained fauna is generally of restricted lagoonal aspect (dendroid stromatoporoids, gastropods, ostracode fragments, small globose stromatoporoids), but brachiopods suggestive of marginally more open circulation are locally abundant (smooth spiriferids such as Atrypa phoca, A. foxi). Oncoids, local grainstone textures in peloidal sediments, and the fragmented nature of most of the allochems attest to some degree of water agitation. Nonetheless, the predominantly muddy texture of these rocks, the lack of traction current structures, absence of intraclastic sediments, and the occurrence of oncoids (Flügel, 1982) all indicate that this highest energy facies was not subject to sustained water turbulence.

3.4 ENVIRONMENTAL SYNTHESIS

3.4.1 Regional Considerations

Boucot and Gray (1979) indicated in their global paleogeographic reconstructions for the Late Silurian that the southern arctic regions formed part of a broad mid-cratonic evaporitic domain. Extensive evaporite deposits are known from the Cape Crawford Formation, and to a lesser degree from the Leopold Formation (both Ludlovian) on Somerset Island (Reinson, 1978; Dixon and Jones, 1978). South of Barrow Strait, the Somerset Island Formation (latest Ludlow) has abundant gypsum nodules locally, along with rare halite casts. The formation of these salts has been compared to the genesis of evaporites on the Trucial Coast (Miall and Gibling, 1977, p.105). On Devon Island the Sophia Lake Formation (early to mid- Lochkovian), conformably overlying the Barlow Inlet Formation, consists in part of very thin-bedded gypsum and gypsiferous shale (Thorsteinsson, 1980). The Cape Storm and Somerset Island Formations, which respectively predate, or equate with, the Barlow Inlet Formation, show distinctive hypersaline geochemical signatures (Veizer et al., 1978). All these data suggest that the Barlow Inlet Formation was deposited in an arid climate.

The Barlow Inlet Formation represents deposition on a carbonate platform of possibly asymmetric physiography (Paper 5, this thesis), an interpretation based on the

thickness and character variation of shelf margin facies developed at different localities within the outcrop belt. A shelf 'rim' invariably separated open marine carbonate slope facies from facies of the hemicyclic series. This rim was composed predominantly of crinoid sand shoals, and dense stromatoporoid biostromes and buildups. Ecologic reefs probably formed a subordinate proportion of the shelf margin facies. These platform edge barriers were sufficient to restrict circulation in the platform interior, although the impact of both normal and storm-wave surge at the shelf rim, was apparently not appreciable. The absence or minimal development of oolites, true framework reefs with massive skeletons, reef-flat facies, and cross-bedding of bioclastic sands, all suggest relatively low-energy conditions. Extensive development of shallow foreslope sponge mudmounds (cf. Wilson, 1975), and inferred stratification of waters of the Cape Phillips Back-barrier Basin (Packard and Dixon, 1983), both attest to seas with restricted circulation. These conditions are probably related to the presence of a barrier (Melville-Bathurst Ridge) separating the Hazen Trough from the Cape Phillips Basin (FIG. 3-3). The northern borderland to the Hazen Trough (Pearya), may have also further dampened open ocean currents and wave swell.

3.4.2 Holocene Analogs

Several Holocene shallow carbonate environments provide useful parallels for interpreting the muddy lagoonal sediments of the Barlow Inlet hemicyclic series. These analogs include Florida Bay (and back-reef lagoon), eastern Shark Bay in western Australia, and the Mediterranean coast of France. Their applicability to the Barlow Inlet hemicycles lies in the presence, at each locality, of thick, laterally extensive deposits of structureless fine-grained carbonate. These deposits are bank-like, having formed and been stabilized by the dense growth of marine grasses (e.g. Thalassia, Posidonia, Cymodocea). Maximum thicknesses of individual banks range from less than 1 m (some Florida Bay banks), to 7.6 m (average maximum for the carbonate bank, eastern Shark Bay). The shallowest portions of the upper surface of banks rise to .3 m above mean sea level, and are rarely greater than 1 m below. Deepest portions of the banks may reach 5.5 m below sea level. The area occupied by individual banks can approach 1100 sq. km. The carbonate bank studied by Davies (1970) ranges from 3 to 13 km wide, and is 130 km long. The dimensions of several studied Florida Bay and back-reef banks are in the order of 1500 to 3000 m in largest dimension (Howard et al., 1970; Basan, 1973; Turmel and Swanson, 1976). The average width of most Florida Bay banks is about 500 m (estimate based on fig. 171 of Ginsburg, 1956). Except where cut by tidal channels,

most banks slope gently into inter-bank depressions (Shark Bay) or mud lakes (Florida Bay). These data indicate that, regardless of section orientation, fossilized equivalents of these Holocene deposits need not exhibit bank morphology, unless hypothetical lateral exposure exceeds 300 m. Rather, these banks would appear as stratiform, parallel-sided units of structureless fine-grained carbonate.

The Barlow Inlet hemicycles are not served perfectly by either the Shark Bay or the Florida Bay analogs. The Shark Bay bank is composed of less than 25% matrix (silt and clay-sized particles), and the overall facies mosaic is shore-parallel (Davies, 1970). In contrast, the Florida Bay shallow-lagoon facies form a patchwork mosaic (see FIG. 3-8) and the banks are predominantly of aragonitic calcilutite. The ordered succession and lateral continuity of facies in the Barlow Inlet hemicyclic series, strongly suggest their development through progradation of shore-parallel depositional environments, similar to Shark Bay. The texture of Florida Bay banks, however, more closely resembles that of the micritic Massive Carbonate Mudstone Member.

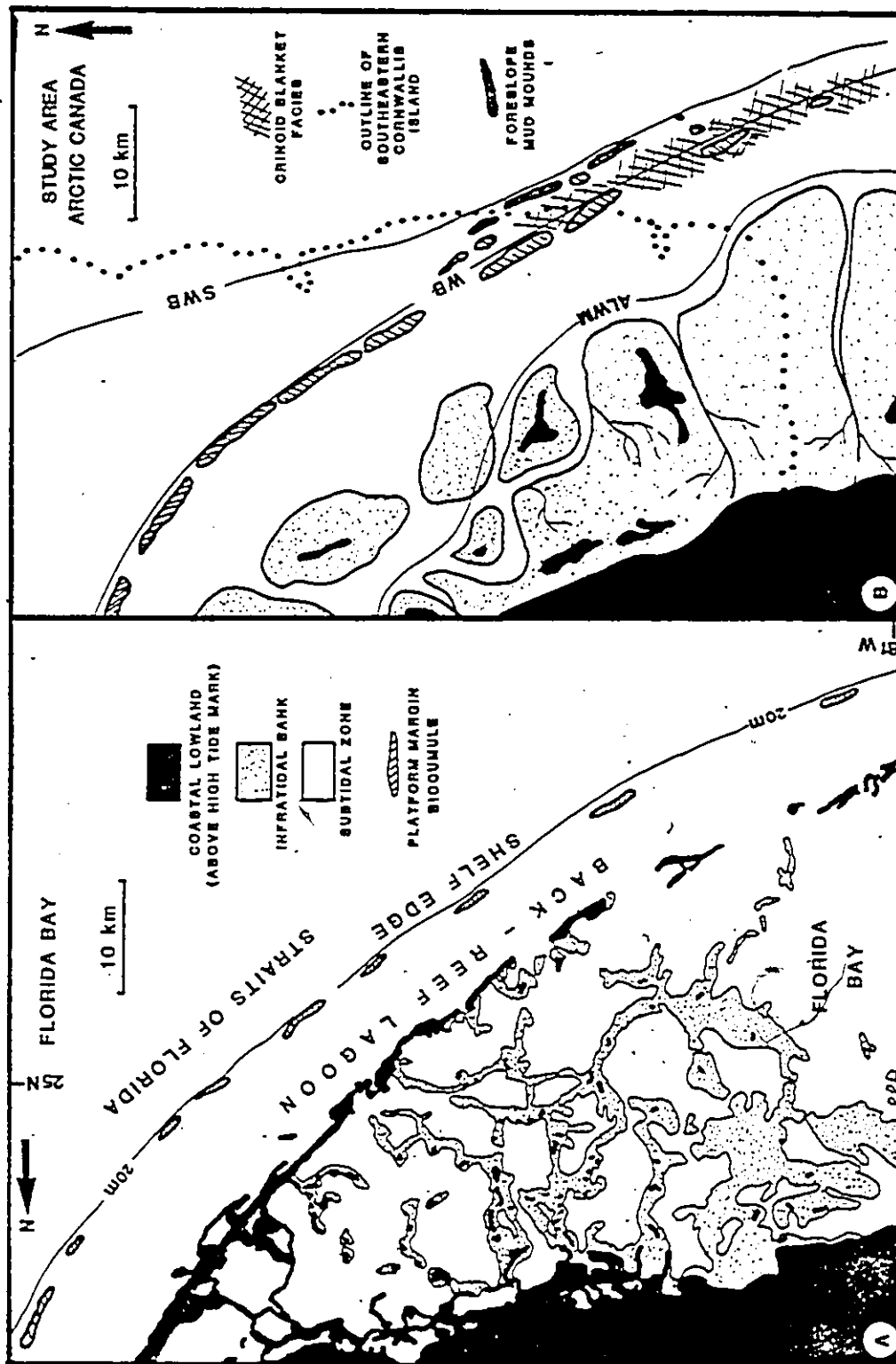


FIG. 3-8a Florida Bay region (after Ginsburg, 1956), note orientation, FIG. 3-8b Hypothetical distribution of facies on the Barlow Inlet Platform at some given geological instant. Differs from Florida Bay in having a shore-parallel rather than a patchwork facies distribution, and lacking a chain of keys or low islands. ALMM - annual low water mark, WB - wave base, SWB - storm wave base.

If the Massive Carbonate Mudstone Member does represent a subtidal bank, then one would anticipate observing at least rarely, some thickness variation, and the presence of cross-cutting channels. Some lateral pinch and swell of a massive carbonate mudstone facies has been previously noted (see PL. 3-13c). The probability of intersecting a tidal channel deposit is low. At Shark Bay, tidal channels are on average 1500 m apart (Davies, 1970, fig. 2). However, the dense mudstone subfacies at loc. 4 contains two large (PL. 3-13d) discrete, sharply-bounded, irregularly-shaped bodies that are inaccessible but evidently of different lithology (judging by their lighter color and their response to weathering). Their origin remains enigmatic, but they may represent channel fills within a massive calcilutite bank. Size and abundance of tidal channels will be proportional to the relative strength of flood and ebb currents, and also reflect the climatic humidity. The first two criteria cannot be gauged independently, but the arid climate inferred would suggest that few tidal creeks transected the Barlow Inlet tidal flats, and this may also reflect a paucity of tidal channels in shallow nearshore marine environments.

The Holocene banks are essentially self-sustaining. The volume of lime mud supplied by epibionts (melobesoid algae, Foraminifera, Bryozoa) attached to marine grasses, is probably sufficient to account for most of the fine-grained

carbonate retained within the banks (Land, 1970; Patriquin, 1972). Marine grasses also baffle and trap externally-derived mud (Ginsburg and Lowenstam, 1958). Their leaves are coated with an organic slime that further traps current-laden sediment. Jindrich (1969) has suggested that inorganic precipitation of aragonite may be induced by photosynthesizing or decomposing grass beds. Once sediments are deposited the extensive rhizome-network of the grasses effectively stabilizes the bank, to a degree such that hurricanes do not substantially erode grass-bound banks (Thomas et al., 1961).

The biofacies of the Florida carbonate banks has been discussed by Howard et al. (1970) and Turmel and Swanson (1976). A restricted sparse fauna of mollusks (mostly gastropods), ostracodes, sponges, codiacean algae, and miliolid Foraminifera occur within the mud bank biotope; and is broadly comparable to the restricted sparse fauna found in the Massive Carbonate Mudstone Member. Live grass root systems can extend to 40cm below the water/sediment interface, and decaying rhizomes and roots are ubiquitous through bank thicknesses. Immediate off-bank or bank margin biofacies include the red alga Goniolithon, Halimeda and other codiaceans, and the corals Porites porites and Siderastrea radians. This biofacies occupies muddy sand to sand substrates. The more skeletal-rich and calcarenitic nature of member B of the RI hemicycles may approximately

parallel the off-bank/bank margin facies of the mud banks (stick stromatoporoid/ P. porites analog?). At Shark Bay a predominantly infaunal bivalve community occupies more sandy portions of the low intertidal/infratidal zone (Megalodont Bivalve Subfacies equivalent?), shoreward of the marine grass dominated bank, and an algal-mat community predominates in the intertidal zone (Davies, 1970).

Abundant, massive colonial metazoans, that might correspond to the stromatoporoids of the Biostromal Member are absent in the immediate vicinity of the Florida mud banks, and are absent entirely from the Shark Bay bank environments. In the immediate back-reef environment of southern Florida, a community of large massive corals such as Diploria, Porites asteroides, Montastrea and Siderastrea, along with branching corals (Bathurst, 1975), may be more closely analogous to the Biostromal Member.

The fact that marine grasses did not evolve until the Late Cretaceous, raises important questions regarding the source and stabilization of the Barlow Inlet lime muds. The environment occupying the interior of the Barlow Inlet shelf may have been sufficiently tranquil that bank stabilization was unnecessary. However, the sharp channel profiles (PL. 3-13d), the purity of the calcilutite (i.e. no arenaceous phases) and the lateral continuity of the massive carbonate mudstone facies suggest that grain size sorting by hydrodynamic processes was not largely responsible for these deposits, and that an organic control is therefore likely.

Stabilization of the substrate might have been achieved by holdfast structures of chlorophytes, rhodophytes and/or phaeophytes, or by the growth of subtidal algal mats (predominantly cyanophytes). The ability of modern chlorophytes (e.g. Penicillus) to bind sediment is limited (Scoffin, 1970). Corallinacea (e.g. Goniolithon) would leave a tangible skeletal record of their presence, and phaeophytes are more common in Holocene rocky-shore habitats (at least one brown alga is, however, associated with carbonate banks of Shark Bay -- Davies, 1970). Subtidal algal mats may not leave a remnant laminated structure, but they should show a clotted texture and/or irregular patches of coarse and fine grains (Neumann et al., 1970). It is unlikely that algal mats could selectively trap only calcilutite sediments. Abbott (1976) has suggested that crinoids served as the principal baffles on lime mud banks in the Silurian. However, this view has been challenged by Riding (1981), and in any event crinoids are rare in the Barlow Inlet platform interior sediments.

The calcareous sheaths of modern chlorophytes disintegrate into clay and silt sized aragonitic particles when the organism dies. Such flora (e.g. Penicillus, Rhizocephalus) leave no recognizable trace of their existence (Stockman et al., 1967) but do contribute significantly to the mud fraction of lime sediments. Although it must remain speculative, Silurian codiaceans (or

dasyclads, or some other unknown biota), perhaps more efficient at sediment binding than their Recent relatives, may have served as both a sediment source and a stabilizing agent for the lime mud deposits of the Barlow Inlet platform.

3.4.3 Ancient Analogs

Portions of the lithofacies spectrum in the Barlow Inlet hemicycles are similar to many rock types described from shallow water limestone sequences throughout the Phanerozoic, but particularly in the Devonian and the Triassic. Fischer (1964, 1975) has described restricted lagoonal and intertidal facies from transgressive cycles in the Triassic Dachstein Limestone of the Austrian Alps. His subtidal megalodont limestones and intertidal loferites bear striking similarity to the Megalodont Bivalve and Loferite Subfacies of the Barlow Inlet Formation. The Triassic cycles exhibit caliche profiles indicative of sustained emergence (absent in the Barlow Inlet hemicycles), but are missing a subtidal biostromal facies. Fenestral limestone and megalodont bivalves are reported from back-reef cycles in Devonian buildups of the Canadian Cordillera (Mountjoy, 1975). These peritidal deposits lack well-developed cyclicity, do not contain biostromal beds, and do have a significant intraformational conglomerate facies. Restricted shelf facies of Devonian age from the Canning Basin contain

both fenestral mudstones and the bivalve Eumegalong (Playford, 1969). Read (1973) has described in detail the nature of the Canning Basin back-reef hemicycle sequences.

Other ancient examples that are comparable to the Barlow Inlet hemicycles, but do not apparently contain a megalodont facies, have been documented by Havard and Oldershaw (1976), Kasig (1980), and Burchette (1981). Biostromal facies characteristically develop in a reef-proximal position in the lagoons (e.g. Kasig, 1980).

3.5 SUMMARY

The facies of the Barlow Inlet hemicycles developed in restricted shelf lagoon and tidal flat environments. Predominantly muddy sediments were deposited in essentially shore-parallel facies belts. Progradation of these belts resulted in characteristic peritidal shallowing-upward cycles. Several Holocene carbonate bank environments are analogous, and particularly Florida Bay, where calcilutites with a very sparse fauna presently accumulate in tranquil shallow waters behind the Florida reef trend and keys (FIG. 3-8). Bathurst (1975, p.163) queried the possible appearance of "...a section through lithified [Florida] Bay sediments...", given the thick, structureless, biotubated lime mud units forming the mud banks. The massive carbonate mudstone lithofacies of the Barlow Inlet hemicycles is likely just such a lithified equivalent.

3.6 SIGNIFICANCE OF HEMICYCLIC SEQUENCES

3.6.1 Platform Evolution

Two important questions arise from this study concerning the evolution of the Barlow Inlet Platform.

1. Do the observed differences in hemicycle types necessarily reflect fundamental changes in platform morphology and shelf environments, or do they indicate synchronous records of deposition on different portions of the shelf?
2. What is the nature of the northward progradation of peritidal facies, and what implication does this have for platform evolution and regional tectonics?

To be convincingly answered, both questions require precise age control. Unfortunately, existing biostratigraphic data are extremely sparse, and until time lines can be drawn through the Upper Silurian section with confidence, interpretations of platform evolution must remain tenuous. The ideas presented below should be considered working hypotheses that will no doubt be modified as age data are secured.

The RI series succeeds a TF series at localities 3 and 4 at roughly the same stratigraphic position. The relative abruptness of the change from one series to the other is readily apparent in the weathering profile of sections containing both series (PL. 3-13e,f). At loc. 7 and 8 the same succession from TF to RI series is present, although the transition is less well-defined.

The TF-RI transition arguably marks an event horizon. What remains unclear is whether the two hemicycle types represent overlapping paleoenvironments, or substantially different facies reflecting significant (epeirogenic and/or organic) modification of the platform. It could be argued, for example, that the RI series represents cyclic sedimentation under the influence of a shelf-margin barrier. The restricted lagoonal interpretation that has been applied to the principal facies of this hemicycle type appears to support this. In the absence of a Biostromal Member and the diminished presence of the Massive Carbonate Mudstone Member, the TF hemicycles could be interpreted as more open shelf sediments (i.e. pre-barrier). The transition event could thus be growth and coalescence of carbonate buildups to form a barrier at the Barlow Inlet Platform shelf edge. The disposition of biocumule facies in the Barlow Inlet Formation does support some degree of Platform Margin evolution (see Paper 4, this thesis).

However, substantial platform modification need not be invoked to explain the differences between TF and RI hemicycles for the following reasons:

1. neither the faunal diversity, nor the distribution of biota suggests that the TF cycles were deposited under conditions of more open circulation than were the RI cycles (see Table 3-3).

Comparative biofacies statistics of hemicycles of
TF and RI affinities

	RI n = 14 loc. 3	RI n = 16 loc. 4	RI n = 13 loc. 8	TF n = 13 loc. 3	TF n = 13 loc. 4	COMPO- SITE RI	COMPO- SITE TF
MEAN MAXIMUM ABUNDANCE ¹ /CYCLE	4.5	2.9	3.8	3.2	3.7	3.7	3.5
MEAN MAXIMUM DIVERSITY ² OF BODY FOSSILS/CYCLE	4.3	1.1	2.8	1.7	1.5	2.7	1.6
MINIMUM - MAXIMUM DIVERSITY	0 - 9	0 - 4	0 - 7	0 - 3	0 - 3	0 - 9	0 - 3

1. abundance figure is based on a quasi-logarithmic scale as follows :
(in theory % allochems/unit volume rock; in practice % allochems/unit area of rock);

6 - 60% plus
5 - 20-59%
4 - 5-19%
3 - 1-4%
2 - <1%
1 - NONE
0 - NOT RECORDED

2. cycle number or lithofacies with the greatest diversity of biota is used. Trace fossils are not counted.

Table 3-3 (concluded)

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		PERCENT OF TOTAL RI CYCLES EXAMINED (43) WITH A PARTICULAR BIOTIC ELEMENT	PERCENT OF TOTAL TF CYCLES EXAMINED (26) WITH A PARTICULAR BIOTIC ELEMENT
STENOHALINE MACROBIOTA	BRACHIOPODS	14	19
	CRINOIDS	14	4
EURYHALINE MACROBIOTA	MASSIVE STROMATO- POROIDS	70	12
	DENDROID TABULATES	21	0
	MASSIVE TABULATES	12	8
	RUGOSE CORALS	32	8
MACROBIOTA TYPICALLY ASSOCIATED WITH METAHALINE- HYPERHALINE CONDITIONS	ONCOIDS	28	73
	DENDROID STROMATO- POROIDS	14	31
	GASTROPODS AND OSTRACODES	16	19

2. The first major buildups occur in the stratigraphic record only 26 m above the base of the upper member of the Barlow Inlet Formation at loc. 5. It seems likely that a partially rimmed shelf margin existed early in the development of the platform¹⁰ and that circulation in the platform interior was restricted from that time onward.
3. At loc. 5 equivalent strata (in terms of height above base) to the TF cycles of locs. 3 and 4, are pervasively dolomitized and severely recrystallized. Precise environmental interpretation of the representative lithofacies is not possible, although the units appear stratiform, and consist of either vuggy, coarse-grained dolostones with common to abundant molds of stromatoporoids, or massive, non-porous, fine to medium-grained dolostones which commonly have white, irregularly-shaped, sparry calcite blebs (.5-3 mm) scattered throughout the rock. These lithofacies are suggestive of the Biostromal and Massive Carbonate Mudstone Members of the RI hemicycles, and may represent the RI lateral equivalents to the TF hemicycles.

¹⁰ the lateral extent of the buildup is difficult to assess. Nonetheless the lowest buildups at loc. 5 are equivalent in magnitude and form to any younger reefoid structures in the Barlow Inlet Formation.

4. The estimated maximum bathymetric ranges for the two hemicycle types given below, suggest that TF hemicycles were usually initiated at somewhat shallower depths. This would explain in part the absence of the Biostromal Member. The difference in proportions of the remaining members in the two hemicycle types, and in particular the diminished representation of the Massive Carbonate Mudstone Member, can best be explained in terms of deviation from a simplified accumulation-subsidence model.

The model assumes that the average rate of subsidence increased linearly across the platform. The model also assumes that the stratigraphic thickness of any lithofacies is proportional to the geographic width of the zone it occupies on the shelf. Zones are parallel, and the depositional slope is constant. In this context the subsidence-corrected lithofacies thickness (TL) can be calculated as follows:

$$TL = L (1+S/A-S)$$

where L = total lithofacies thickness = $\sin a$ (cosa W)

and S = subsidence rate

A = accumulation rate

G = depositional gradient

W = width of lithofacies zone

a = angle of depositional slope

The stratigraphic thickness of a complete hemicycle (HT) can be expressed

$$HT = \sum T_n$$

where n represents each lithofacies of the complete hemicycle.

As illustrated in FIG. 3-9, if A, W, and G are held constant for each lithofacies, and if the subsidence rate at any one locality on the shelf is constant, then hemicycles should contain the same proportion of members, regardless of their geographic position. However real hemicycles show considerable variation in proportions of members, both between cycle types and between examples of the same cycle type at the same or different localities. Apart from the problem of subjective interpretation (e.g. cryptic coalescence), the principal reasons for lack of proportionality are as follows: a) the widths of facies belts are not constant. The assumption that W is constant for each lithofacies does not account for changes in facies belt width (and type) as the strandline approaches the effective seaward limit of tidal flat progradation (see FIG. 3-13 below). b) Subsidence rates are likely not constant but probably follow a decay curve, such as is illustrated

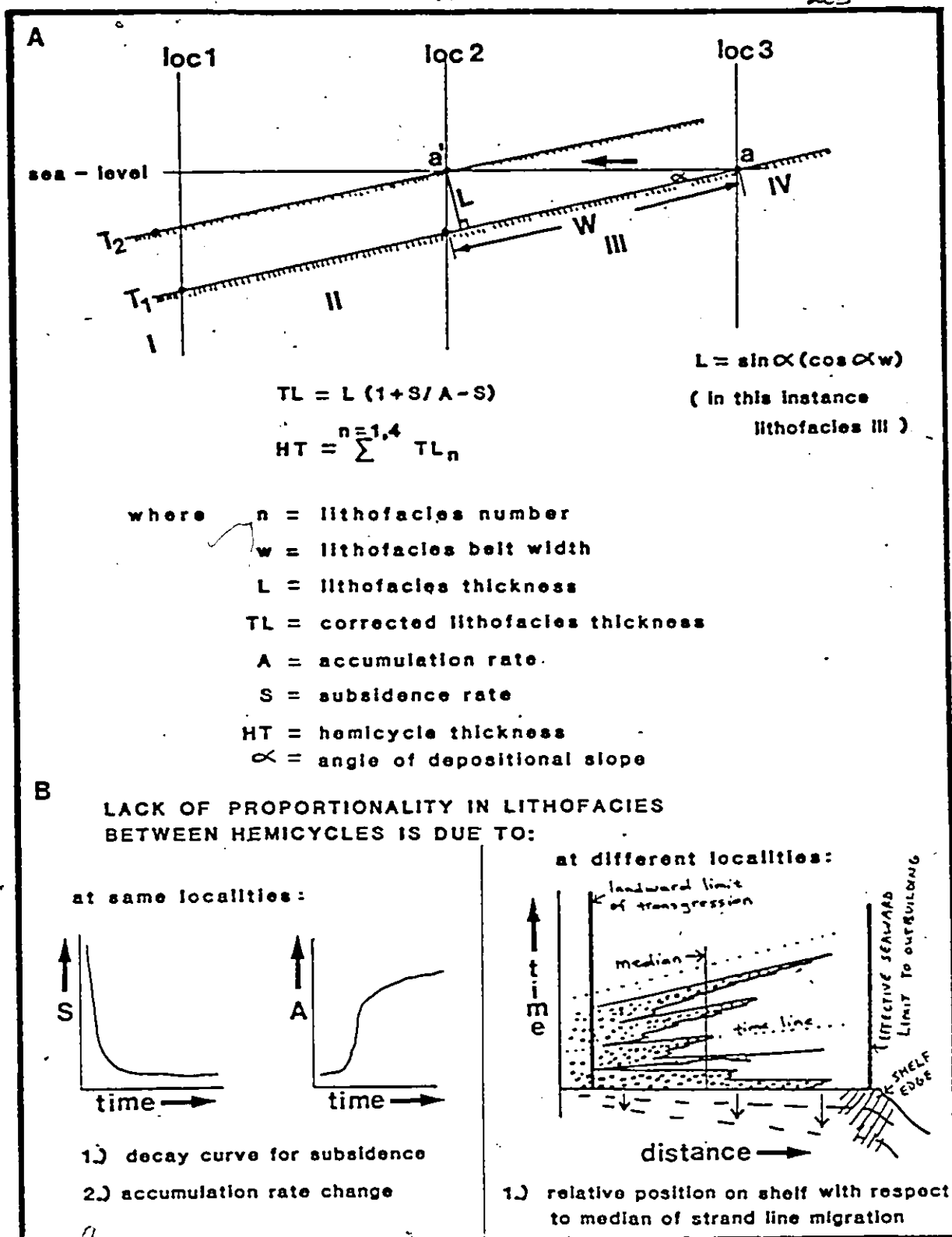


FIG. 3-9 Accumulation/subsidence model for hemicycles (A), and deviations from it (B).

in FIG. 3-9. Accumulation rates as well may vary with time (taking into account initial start-up phase that follows transgression). As facies prograde seaward, over one fixed location on the shelf a unique vertical record will accumulate that reflects these changing rates of subsidence and accumulation. Complete hemicycles therefore will likely exhibit different proportions of members, depending upon their relative position on the shelf (e.g. with respect to shelf margin or average mid-point of strandline migration, see FIG. 3-9). The TF hemicycles developed on a portion of the platform less frequently transgressed, whereas the RI cycles are thought to have developed in a mid-platform to shelf-margin proximal position.

5. If the RI and TF cycles are to be considered 'multi-facies' equivalents, then it is reasonable to suspect that both cycles intermediate between these two types, and transitions from RI to TF cycles, should exist. Examples of both can be found at locs. 7 and 5 respectively.

The abrupt transition from TF to RI series at locs. 3 and 4 represents an important event in the history of the Barlow Inlet Platform. At equivalent stratigraphic horizons at loc. 5 a significant marine transgression is recorded with deep-water carbonate slope-facies overlying shelf-margin

biostrome and crinoid shoal facies (see Paper 5, this thesis). It is suggested that the abrupt transition from TF to RI cycles in the platform interior is the result of a sudden southward retreat of the shelf margin, perhaps due to accelerated fault block subsidence of the Barlow Inlet Platform.

Therefore in response to the first question, there is no conclusive evidence that the RI cycles superseded TF cycles on a platform-wide basis due to some gross change in the paleoenvironments of the Barlow Inlet Platform. Instead, the two series probably accumulated at the same time but on different parts of the differentially-subsiding platform. The abrupt transition between series at locs. 3 and 4 is, however, the result of a major drowning of the seaward margins of the Barlow Inlet Platform. The cause of the drowning remains unknown.

This change in hemicycle character according to relative platform position, has also been documented by Burchette (1971, p.125) who noted "Cycles become thinner and restricted facies more evident in the direction of the paleoshoreline, and thicker, less restricted and less distinct distally".

The first occurrence of strandline facies (cryptalgal laminites) is at progressively higher positions in the Barlow Inlet Formation going from south to north. This is not surprising, given the inferred subsidence rate increase

to the north and the bathymetry of the pre-existing northerly inclined Douro Ramp (see Paper 1). Basin-filling is a prerequisite to tidal-flat progradation, and clearly the more northerly and deeper parts of the Douro Ramp-Cape Phillips Back-barrier Basin, required significantly more sediment to fill to sea level. What is anomalous, is the fact that the first occurrences do not follow a simple curve, but are paired into three distinct groupings (FIG. 3-10). Cryptalgal laminites appear at approximately the same height above base in some adjacent sections, even where separating distances are as great as 12.5 km (locs. 5 and 6). Between other sections 10km apart, the first occurrences of cryptalgal laminite are separated stratigraphically by 300 m (locs. 4 and 5). It would appear that long periods of equilibrium or stasis separated major regressive events affecting large areas of the platform more or less simultaneously.

Cyclic sedimentation generally occurred as a steady-state process operating spatially within the limits of existing platform dimensions. Short periods of platform aggradation or outbuilding, or periods of reduction in platform area (back-stepping or drowning), punctuated these equilibrium conditions, resulting in marked migration of the shelf-margin. These periods of disequilibrium and substantial platform genesis, are probably closely related to more intense periods of epeirogenic activity on the Boothia Uplift.

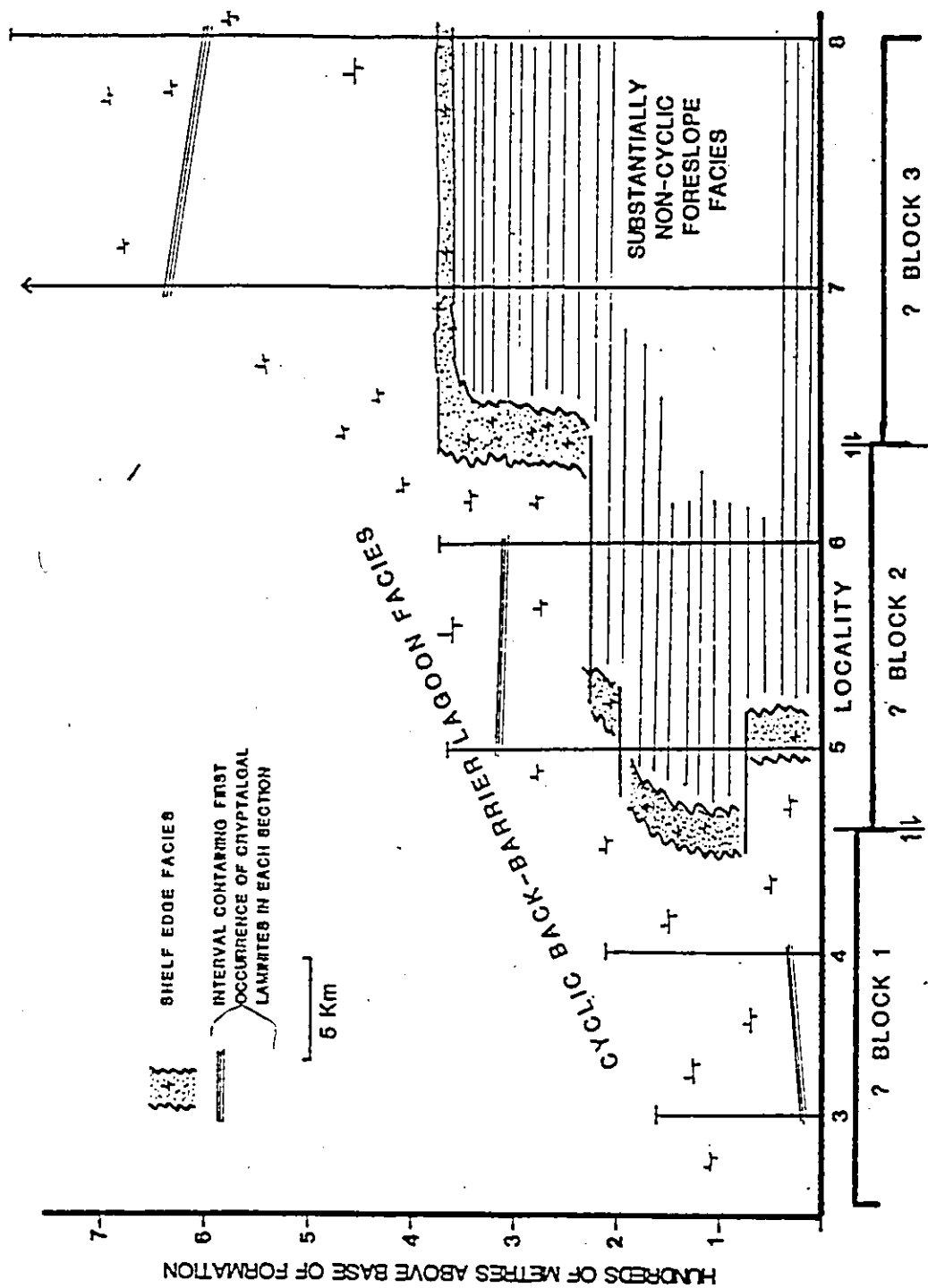


FIG. 3-10 First stratigraphic occurrence of cryptalgal laminites.

If, as is suggested here, the late Ludlow-Pridoli stratigraphic record of marine carbonates now exposed on Cornwallis Island, faithfully reflects pulses of tectonism affecting the basement fault blocks of the Boothia Uplift, then this record may provide some insight into the history of epeirogenesis of the uplift. Gibling and Narbonne (1977, p.1154) stated that "...the tectonism that produced the Boothia Uplift extended northward during the Pridolian and Gedinian." The manner in which epeirogenesis ultimately affected parts of the Miogeocline remains controversial. Thorsteinsson (1980) has suggested that three periods of diastrophism affected Boothia Uplift in late Silurian and Devonian time. He suggested that tectonic pulses of late Ludlow-early Pridoli and early Lochkovian ages affected areas south of Barrow Strait, and a crustal disturbance of late Lochkovian age caused uplift north of Barrow Strait. Miall (1983, p.495) has argued for "...a single, extended period of diastrophism, that gradually extended northward along the uplift to Cornwallis Island".

The evidence presented here would suggest that neither viewpoint is entirely appropriate. If the modified Bott model discussed below is accepted as a mechanism for producing the hemicyclicity of platform interior sediments, then mild ongoing uplift, and concomitant subsidence, is prerequisite. Long-term periods of 'relative' stasis existed, with no significant enlargement or extension of the

Platform. Short periods of more intense fault-block movement were responsible for significant areal enlargement or reduction of the platform. The number of these rather more intense diastrophic movements is not clear, but at least 9 can be identified from the late Ludlovian and Pridolian marine record (see Paper 5, this thesis).

3.7 ESTIMATING RELATIVE PLATFORM SUBSIDENCE

Given that equivalent lithofacies commence and terminate similar hemicycles throughout the platform, it is reasonable to assume that the bathymetric range represented by these cycle-types does not vary significantly from one study locality to another. Therefore, the differences between observed cycle thicknesses (composite and mean)¹¹ from area to area, must reflect first-order variation in subsidence rates on the platform. FIG. 3-11 plots mean and composite cycle thickness against distance between study localities. An increase in cycle thickness is observed from loc. 1 to loc. 8, although the magnitude of change varies between cycle types. At Cunningham Inlet south of the principal study localities, TF hemicycles from the time-equivalent lower member of the Somerset Island Formation are not significantly thinner. North of locs. 3 and 4, both TF and RI cycles in the type section of the Barlow Inlet Formation show appreciable increase in thickness. This trend would be anticipated, if subsidence principally governed cycle thickness, given the relative south-north disposition of these localities normal to the assumed Platform-Miogeocline flexure line.

¹¹ It is argued that with a sufficiently large sample of partial to complete cycles in each area, it is reasonable to compare mean cycle thicknesses from different localities as those processes resulting in 'coalesced' or 'aborted' cycles should have had similar net effect on all parts of the shelf lagoon.

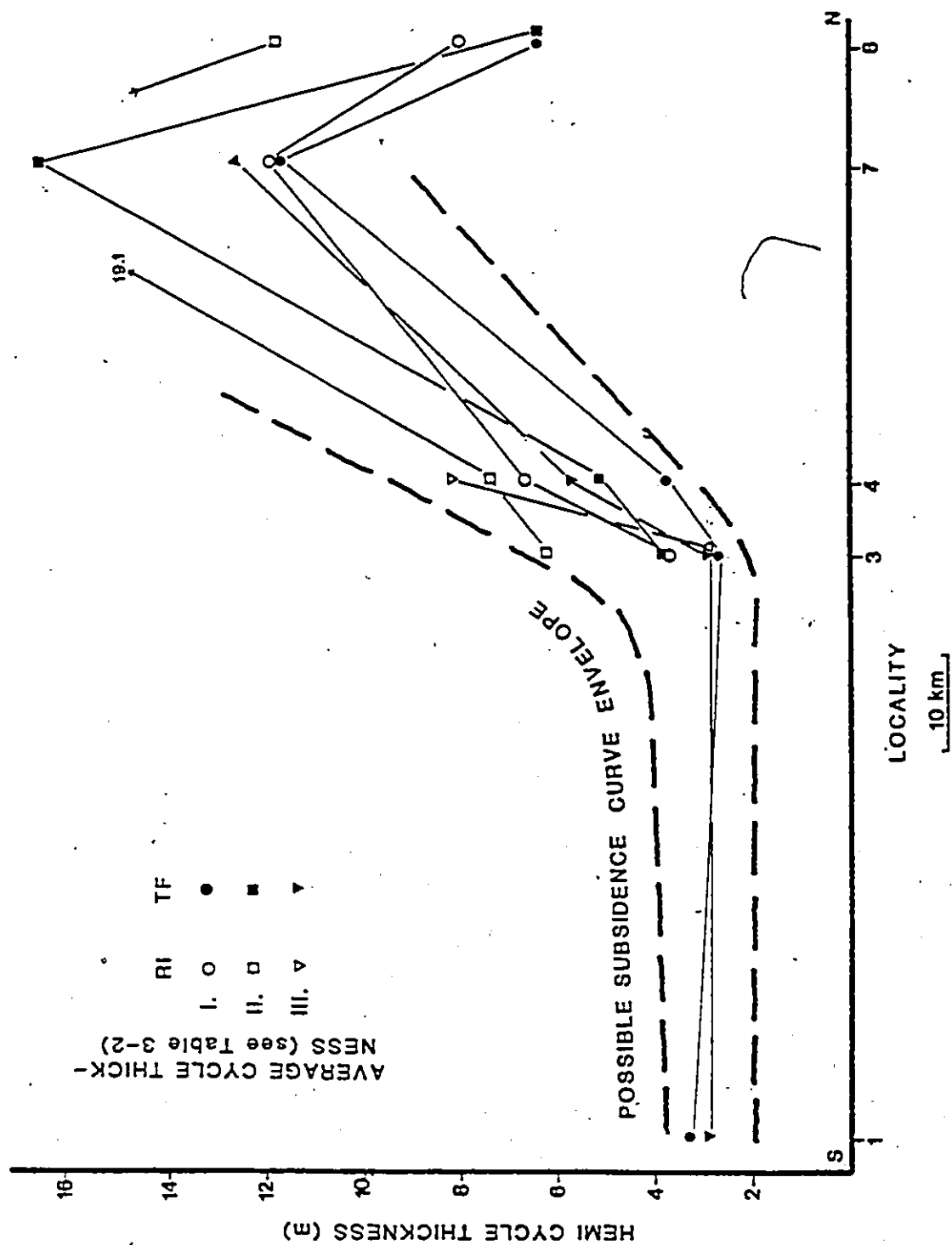


FIG. 3-11 Trends in average hemicycle thickness across the platform.

Curiously, north of the type section at loc. 7, limited data suggest a pronounced decrease in mean cycle thickness (FIG. 3-11). The decrease (33-48%, depending upon cycle type) is roughly proportional to the decrease in total section thickness (29-37%). It is not known why the portion of the platform represented at loc. 8 should have subsided less rapidly than areas immediately to the south. However, this phenomenon probably reflects the independence of movement of fault blocks within the Boothia Horst (Paper 2, this thesis; see also Muir and Rust, 1983).

The example given above underlines a potential application for cycle thickness statistics. If a mobile belt -- stable craton flexure remains relatively fixed such that local subsidence fluctuates minimally through time about some mean, then first approximations to total formation thickness can be derived from average cycle thicknesses. Of course a complete formation thickness must be available from a section that contains a reasonable number of well defined cycles, and comparisons can only be made to approximately time-equivalent successions of cycles of comparable type. Exposures of the Barlow Inlet Formation containing both upper and lower contacts are available only at the type section (loc. 7) and at Sophia Lake (loc. 8). The thickness of this formation on southeastern Cornwallis Island cannot be ascertained, as the upper contact lies beneath Wellington Channel or has been removed by erosion.

Using the subsidence curve shown in FIG. 3-11, crude¹² estimates of total formation thickness at locs. 3 and 4 are 325 m and 420 m, respectively. The subsidence rates illustrated in FIG. 3-11 are compatible with known average accumulation rate trends across the flexure since the Late Ordovician.

¹² In light of possible independent fault block movement, the author acknowledges that derivation of thickness or subsidence rate information solely on the basis of the subsidence curve presented in FIG. 3-11 will yield only first-order approximations, and considerable caution should be exercised in their interpretation. The example given is largely to emphasize the potential of the method.

3.8 APPROXIMATION OF THE BATHYMETRIC RANGE OF CYCLE LITHOFACIES

If, as has been argued, the base of the laminated lithofacies in both hemicycle types corresponds to the lower limit of the intertidal zone, then the position of this sealevel datum can be used to estimate bathymetric ranges for other cycle lithofacies (method after Read, 1973). Several factors can potentially modify the relationship between true bathymetry of the depositional environment of a particular lithofacies, and its vertical position in the hemicycle vis-a-vis the Laminated Member. These factors are compaction, cryptic cycle coalescence, initial bathymetry due to relative sea-level rise, and subsidence.

Fine-grained carbonate sediments have initial porosities of up to 70% (Bathurst, 1975), but as a consequence of early cementation, undergo little or no compaction (Pray, 1963; Zankl, 1969). The preservation of an "uncollapsed" primary fenestral fabric in lime muds of the Barlow Inlet Formation hemicycles attests to the negligible effect of loading on these sediments. Thus for the purposes of estimating depths of deposition, the compaction factor can be ignored.

Cryptic cycle coalescence refers to two or more cycles being mistakenly treated as a single cycle, because of the absence of a distinct, vertical lithofacies change suggestive of cycle initiation or termination. Relative sea-level rises that do not exceed the bathymetric range of a particular lithofacies, will often result in cryptic cycle

coalescence with a concomitant increase in apparent thickness of that cycle member. The very nature of cryptic coalescence makes it undetectable and therefore uncorrectable. It is reasonable to assume that this condition will be aggravated with elevated subsidence rates.

Following transgression, the depth of the seafloor below the sea-level at any locality will limit the maximum attainable thickness of the initial lithofacies. Unlike lithofacies B and C, lithofacies A, being the initial cycle member, is not bounded at its base by deeper water lithofacies. Therefore its thickness will depend not only on subsidence, but also on the level, within the complete bathymetric range for that lithofacies, at which accumulation commenced following sea-level rise (see FIG. 3-9). The lower bathymetric limit of lithofacies A cannot be estimated reliably.

It is clear that subsidence did not exceed sediment accumulation, otherwise transgressive rather than regressive hemicycles would have developed. However, for any subsidence rates greater than zero, cycle member thicknesses will be increased by keeping the depositional interface at a particular depth for a longer period. It can be concluded that minimum member thicknesses of unaborted cycles most closely approximate the true bathymetric ranges for each generalized lithofacies.

FIG. 3-12 provides several different measures of the relative positions of lithofacies (C,B,A) and inferred sea level (base of lithofacies D). The constructed cycle (case 1) represents the cumulative total of average thicknesses of cycle members. It is uncorrected for subsidence or coalescence, and therefore serves to maximize thicknesses, and hence bathymetric ranges. It is extremely improbable that the depth of the shelf interior ever exceeded 14 m. Case 2 represents the minimum thicknesses of the generalized lithofacies. Case 3 indicates the lithofacies positions within a complete real cycle with minimum aggregate thickness. Case 4 shows depth ranges of lithofacies A,B, and C beneath lithofacies D, in cycles that contain D. Case 5 is similar to case 4, except that only minimum values are plotted. No one measure produces a more consistent pattern, but there is reasonable agreement between measures. Table 3-4 summarizes probable maximum depths and bathymetric ranges for generalized lithofacies A through C.

BATHYMETRIC RANGE OF CYCLE MEMBER LITHOFACIES RESTRICTED INFRATIDAL HEMICYCLES TIDAL FLAT HEMICYCLES

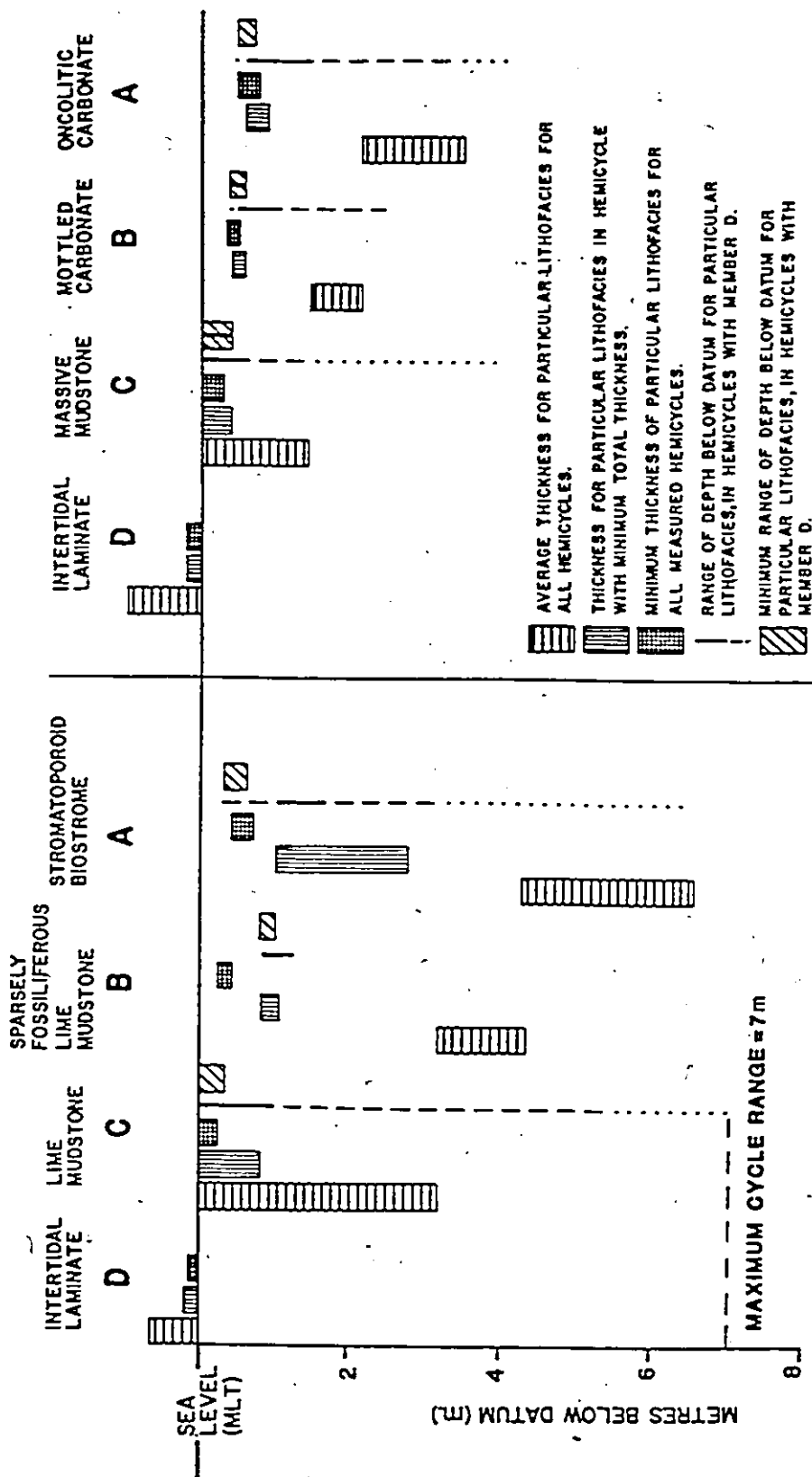


FIG. 3-12 Bathymetry determined by lithofacies position and thickness.

Table 3-4

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Probable bathymetric range for hemicycle lithofacies

MEMBER OR LITHOFACIES	Restricted Infratidal		Tidal Flat	
	DEEPEST LIMIT	POSSIBLE RANGE	DEEPEST LIMIT	POSSIBLE RANGE
C	< 1m	0 - 1m	< .6m	0 - .6m
B	< 1.2m	.2 - 1.2m	< 1.0m	.3 - 1.0m
A ²	° 2.8m	.4 - 2.8m	° 1.6m	.5 - 1.6m

1. below mean low tide (MLT) level. In order to minimize the effect of subsidence the figures are derived by placing greatest emphasis on complete hemicycles with minimum thickness.
2. for reasons cited in the text, the bathymetric range for lithofacies A is difficult to estimate.

If the above figures are correct, then it would be highly unlikely that shelf depths of the Platform Interior much exceeded 3 metres.

3.9 PRINCIPAL CONTROLS ON HEMICYCLIC SEDIMENTATION

Many models have been proposed to explain cyclicity in sediments. Various local, global, and even cosmic factors have been invoked to explain rhythmicity in the stratigraphic record. Review articles by Weller (1964) and Duff et al. (1967, chapter 12) present details of various models.

Strict application of any of these models to the Barlow Inlet Formation hemicyclic succession, denies the essentially speculative nature of the models. The cycle mechanisms advanced herein, although appropriately reflecting the bias of available data, are also essentially simple, and free from more regional, or even global influences. This, however, does not exclude more esoteric hypotheses. The critical assumptions employed in determining the 'most likely' control on the Barlow Inlet hemicyclic series, concern periodicity, eustasy, subsidence rates, basin dynamics, and accumulation rates.

3.9.1 Periodicity (absolute cycle intervals)

Biostratigraphic data are limited, but Uyeno (1977) reported conodonts suggestive of the transition between the latialata and eosteinhornensis Zones from a reefoid lithology 174 m above the base of the Barlow Inlet Formation at loc. 5. The lower limit of the latialata Zone occurs within or at the base of the lower member of the Barlow Inlet Formation (Thorsteinsson, 1980). The reefal development at loc. 5 is considered coeval with, and likely essential to, the development of the interior shelf hemicycles. The TF hemicycles at loc. 4 overlie the lower member of the Barlow Inlet Formation, and are therefore not older than the latialata zone. When corrected for a possible subsidence rate differential (see FIG. 3-11) the 174 m of latialata zone at loc. 5, equate approximately to 82 to 130 m at locs. 3 and 4, made up of at least 15 cycles (of either TF or RI type). Given a 6 to 8.5 million year duration for the Ludlow Epoch (McKerrow et al., 1980), the latialata Zone, as one of four European conodont zones in the Ludlow, does not likely exceed 2 or 3 million years in length. Therefore a maximum value for the average period of an asymmetric cycle is estimated to be about $(3 \times 10^6 / 15) = 200,000$ years. The range 50,000 to 200,000 is considered reasonable, and certainly correct within an order of magnitude.

3.9.2 Subsidence

Crustal accommodation to sediment load in the study area was considerable. At loc. 5, there is evidence of emergence at the base of the Cape Storm Formation and the top of the lower member of the Barlow Inlet Formation. The 800 m of strata between these two horizons accumulated during a period of relative epeirogenic quiescence. Reasonable limits to the duration of the siluricus Zone, to which all but the 50 m of the lower member can be assigned, range from 1.5 to 5.6 million years. Therefore disregarding compaction, which would tend to increase the following figures somewhat, the crust must have subsided on average between 16 and 61 cm/1000 years. In contrast, values for long term basin subsidence commonly range from 1 to 10 cm/1000 years (Schlager, 1981).

If possible eustatic sea-level changes are ignored (see next section), then interpreted deepening in the upper part of the Douro at loc. 5, (Narbonne, 1981; see Paper 2), must reflect periods of increased subsidence (100's cm/1000 years?) capable of exceeding carbonate accumulation.

In summary, that portion of the miogeocline proximal to the stable craton underwent rapid subsidence during the Late Silurian. Average subsidence rates may have been as high as 61 cm/1000 of years (at loc. 5), and rates as high as several 100's cm/year may have been achieved periodically. Even these substantial values may have been exceeded in the



latest Silurian, as a consequence of intense epeirogenesis to the south.

3.9.3 Eustasy

A wide variety of mechanisms potentially capable of adjusting eustatic sea-level have been advanced (see Table 3-5); of these only tectono-eustasy and glacio-eustasy warrant serious consideration.

Major 1st and 2nd order sea-level changes within the Phanerozoic, as determined by Vail et al. (1977) using seismic stratigraphy, are considered to be the result of a volume change of ocean basins due to the relative elevation of mid-ocean ridges. This in turn can be related to rates of seafloor spreading (Pitman, 1978). First and 2nd order cycles of tectono-eustasy (or oceanic epeirogeny) have periods of at least 10×10^6 years (Vail et al., 1977) and associated rates of sea-level change are on the order of 1 cm/1000 years (Berger and Winterer, 1974). Assuming that Barlow Inlet hemicyclic sedimentation had a period of 50,000 to 200,000 years, and given Holocene accumulation rates for lagoonal and tidal flat carbonates of 50 to 100 cm/1000 years (Schlager, 1981; Wilson, 1975), tectono-eustasy can confidently be disregarded as a significant cyclic mechanism.

Glacio-eustatic processes are responsible for 3rd order, and possibly 4th and 5th order cycles with periods

Table 3-5

Mechanisms capable of producing cyclicity in sedimentary strata.

a) Weller (1964) grouped theories on cyclic mechanisms into 3 broad categories :

- 1) DIASTROPHIC THEORIES
 - A. Continuous subsidence
 - B. Intermittent subsidence
 - C. Alternate subsidence and uplift
- 2) CLIMATIC THEORIES
 - A. Glaciation
 - B. Humid-arid alternations
- 3) SEDIMENTATION THEORIES
 - A. Peneplanation--reduced availability of sediment
 - B. Deltaic or fluvial avulsion
 - C. Current strength alternations
 - D. Compaction
 - E. Differential Settling
 - F. Breaking of barrier and flooding of restricted basins

b) Cause-effect relationships(--->): TECTONISM (UPLIFT, SUBSIDENCE)

COSMIC INFLUENCE --> CLIMATE --> SEDIMENTATION
 (via arrows from TECTONISM)

c) Short term cause and effect relationships (periods less than 500,000 years):

CYCLICAL AGENT PROCESS	ABSOLUTE EUSTATIC CHANGES	EFFECT ON SOURCE AREA	EFFECT ON DEPOSITIONAL BASIN	FEEDBACK EFFECTS OR
SUBSIDENCE	negligible	negligible	- profound, drowning of productive shelf	- induces uplift (Walcott, 1970) - may reduce autochthonous production of sediment UPLIFT
negligible	profound	- rate of	- induces infilling - basin emergence due to tectonism	subsidence (Bott, 1964), - promotes sedimentation, affects climate (eg. rainshadow)
CLIMATE	potentially considerable	rate of denudation	- potentially profound, modification of thermo-cline and productivity of skeleton-producing organisms	- promotes or reduces sedimentation, affects type of sediment
negligible TATION	- profound	- induces	SEDIMEN-	negligible subsidence

Mechanisms responsible for relative changes in sea-level
(adapted from Kendall and Schlager, 1981)

1. EUSTASY

A. TECTONO-EUSTASY

(oceanic epeirogeny)
sea-level rise due to a change in
volume of mid-oceanic ridges
concomitant with change in sea
floor spreading rates.

B. GLACIO-EUSTASY

volume change in hydrosphere due
to climate induced phase change.

C. ADDITIONAL EUSTATIC
MECHANISMS

- a) addition of juvenile waters.
- b) addition of water from hot springs.
- c) meteorite impacts.
- d) desiccation of an enclosed basin.
- e) flooding of a basin, of continental area.
- f) short-term draw-down in evaporite basins.
- g) continental collision.
- h) ocean sedimentation.
- i) ocean trench subsidence.

2. CRUSTAL SUBSIDENCE OR UPLIFT

3. SEDIMENTARY INFILL

as short as 10,000 to 100,000 years (Kendall and Schlager, 1981). Although cycle periodicity is comparable, a glacio-eustatic mechanism for the Barlow Inlet hemicycles is rejected for the following reasons:

1. There is no evidence worldwide of Late Silurian glaciation (e.g. tillites, dropstones, glacial striae).
2. Late Silurian paleoclimates are considered to have been essentially wet and warm (Frakes, 1979) relative to the present
3. If one assumes that the late Paleogene and Quaternary model of frequent climatic fluctuations is appropriate to the Paleozoic, then short term (20,000 years) periodicity of stadials with attendant sea level drops of up to 80 m, should be anticipated (see Adey, 1978, fig. 4). Karstic surfaces symptomatic of low-stand platform emergence have not been recognized in the Barlow Inlet succession.

3.9.4 Accumulation Rates

The siliciclastic content of the hemicycle lithologies is insignificant, and thus sedimentation can be viewed strictly in terms of accumulation of fine-grained carbonate sediments accompanying in-situ growth of a restricted shallow shelf macrobiota. Analogous Holocene sediments (tidal flat, lagoon) accumulate at rates of 50 to 100/cm/1000 years

(Wilson, 1975; Schlager, 1981). However, such figures probably represent minimal values inasmuch as the sedimentation potential is limited by rates of glacio-eustatic sea-level rise. The Holocene growth rate of organic rims (reef belts) is in the order of 50 to 2000 cm/1000 years. There is no evidence to suggest that Late Silurian accumulation rates were substantially different.

3.9.5 Basin Dynamics

The study area is situated on a portion of the southeastern margin of the Franklinian Geosyncline, a major NE-SW trending structural basin that developed from late Proterozoic to mid-Paleozoic time, and formed part of the Innuitian Orogen (Kerr, 1981). A basic tectono-sedimentary history of the Franklinian Geosyncline has now emerged through the efforts of the Geological Survey of Canada (e.g. Trettin, 1972; Thorsteinsson in Douglas, 1970). However, the origin and tectonic significance of the Franklinian Geosyncline remain poorly understood, and in particular the structural mechanics of subsidence at the geosynclinal margins have never been elucidated. The principal element of the geosyncline was a deep-water linear basin, the Hazen Trough, coincident with the axis of the geosyncline. The Hazen Trough was flanked to the southeast by an unstable shelf, which in turn was separated from the craton (Arctic Platform) by a "... hinge or broad flexure, which shifted

with time and generally is difficult to locate." (Trettin in Trettin and Balkwill, 1978). Transverse to the southeastern margin (hinge) of the geosyncline are fault-bounded basement uplifts that serve to isolate several smaller structural basins (indicated in FIG. 3-2b).

Of significance to the cyclic sedimentation of the Barlow Inlet Formation, is the interplay between the episodically-rising Boothia Uplift to the south (FIG. 3-2b), and the rapidly subsiding platform north of the craton-geosyncline flexure. Pertinent to the relationship between the Uplift, subsiding shelf, and stratigraphic cyclicity, is an understanding of the nature of the craton-geosyncline flexure or hingeline. Trettin (in Trettin and Balkwill, 1979) has suggested that the Hazen Trough developed by subsidence dependent on crustal extension (rifting). Both Trettin (op.cit., fig. 8) and Drummond (1973, fig. 5) indicated fault-bounded margins for the Hazen Trough. The southeastern edge of the geosyncline has not been interpreted in a similar manner, and largely by default (no pun intended), has been illustrated as a simple downflexing of the crust (e.g. Kerr, 1981; Thorsteinsson in Douglas, 1970; Thorsteinsson, 1980). Sedimentation and subsidence can be visualized in terms of two end-member models, namely 1) flexural loading of a relatively strong, rigid crust, and 2) depression of an Airy-type crust, wherein fault blocks react independently to local sediment

load (Watts and Ryan, 1976). The wholesale application of any model to the Franklinian Geosyncline's southern margin is well beyond the scope of this paper, and would require much new investigative work. Within the study area, however, there is evidence to suggest that faults exist parallel and proximal to the southeastern margin of the geosyncline. Muir and Rust (1983) postulated that E-W faults transect the Boothia Uplift and may have controlled the disposition of a major line of facies change. Paper 2 of this thesis postulates that the Boothia Horst is a complex of semi-independent basement blocks delineated by both E-W and N-S trending faults. A faulted Airy-type crust with transverse uplifts, would considerably resemble the Atlantic margin described by Sheridan (1974).

The relationship between proximally-located basin subsidence and basement uplift is not simply fortuitous. Bott (1964) and Walcott (1970) have both published on the complimentary isostatic response linking sediment load and crustal uplift, although their approaches were radically different. Bott (1964) viewed basin subsidence as the natural consequence of flow of ductile mantle material away from a subsidence site, and towards an area of crustal uplift caused by isostatic inequilibrium. In contrast, Walcott (1970) considered basement uplift to be the appropriate isostatic response to adjacent sediment loading and sediment-loaded depressions to form part of a

"compensated" topography (i.e. in isostatic equilibrium). These two interpretations leave us with the tectonic conundrum "which came first"? Isostasy is clearly involved, but whether in a 'cause' or 'effect' role remains controvertible. Bott (1976, p.2) has stated that "...the main role of sediment loading is probably that of amplifying subsidence caused by other primary mechanisms...", and in this light it is unlikely that high sedimentation rates were responsible for the interpreted degree and rate of subsidence of the Franklinian unstable shelf.

In either model normal-faulting can play a fundamental role in isostatic crustal reequilibration. In Walcott's model, faulting occurs when the flexured lithosphere reaches its elastic limit (e.g. 1 km sediment thickness difference in 100km lateral distance). Subsequent collapse of the basin margin produces a fault angle depression. Bott (1964) suggested that the position of the subsiding basin will be determined by pre-existing lines of weakness in the basement, and that as subsidence by faulting represents the greatest potential loss of gravitational energy, it will be the favoured mechanism. However faulting is dependent on external (tensional or extensional) crustal stresses.

3.10 SYNTHESIS

Two models can be advanced that help to explain the observed hemicyclicity of the Barlow Inlet carbonate platform sediments. Neither model fully explains all aspects of the rhythmic successions in a regional tectonic or depositional context. However, they are considered the simplest and most plausible hypotheses, given the available evidence. The two models are as follows:

1. Steady subsidence and tidal flat progradation hypothesis (or the 'Ginsburg' model).
2. Episodic subsidence and tidal flat progradation hypothesis (or the 'modified Bott' model).

3.10.1 Steady Subsidence and Tidal Flat Progradation Hypothesis

In 1971 Ginsburg proposed a model for the formation of repeated regressive (shallowing-upward) cycles in shallow water carbonate sequences. The principal tenets of Ginsburg's proposal (slightly modified herein) are as follows:

1. Voluminous quantities of fine-grained carbonate sediment are generated in lagoonal and open shelf areas (in the Holocene by breakdown of skeletal aragonite and calcite).
2. Much of this fine sediment is transported shoreward to be deposited on tidal flats, or trapped and stabilized within subtidal lagoons. The transporting mechanism may be wind or tide-driven currents, with

the greatest volume of sediment being moved when these currents are amplified by storms.

3. Net accumulation of sediment outpaces steady shelf subsidence, resulting in seaward progradation of tidal flat/lagoon facies belts.
4. Seaward progradation of the shoreline (see FIG. 3-13), has two consequences: 1) the original sediment source area is reduced concomitant with enlargement of the tidal flat/lagoon wedge, and 2) as the width of the supratidal flats increases less carbonate mud is supplied (by storms) to the flats, and more distal regions of the flats receive proportionately less sediment.
5. Ultimately, the rate of supply and accumulation of sediment falls below the subsidence rate of the shelf, and the tidal flat becomes submerged. Due to an inherent lag time in carbonate production following drowning, the shelf subsides to some depth before sedimentation again outpaces subsidence, and a new regressive cycle commences.

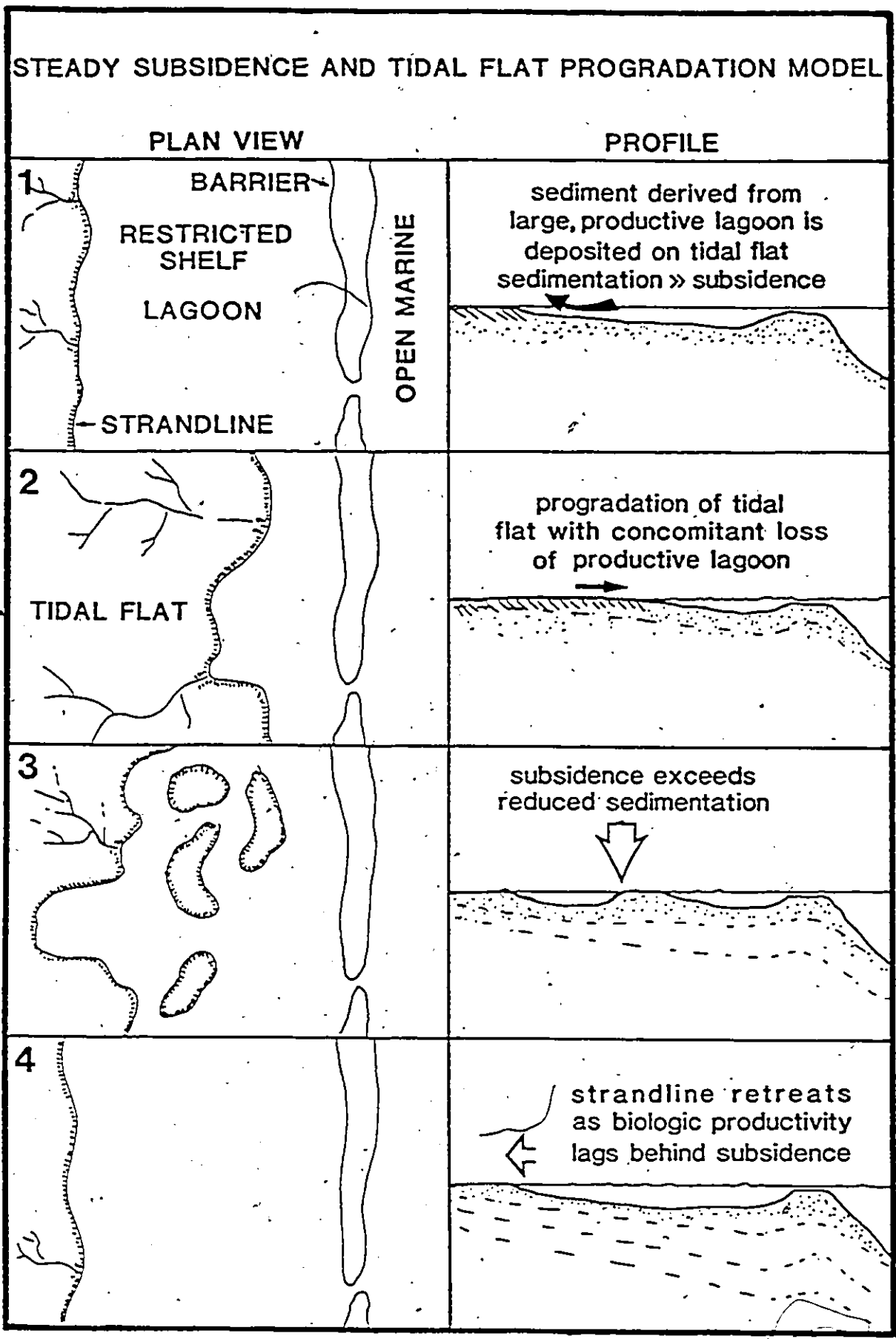


FIG. 3-13 Hemicyclicity due to steady subsidence and tidal flat progradation.

The Ginsburg model is by nature autocyclic (Beerbower, 1964), and as such, is not radically different from cyclic delta sedimentation (Moore, 1959) or cyclic "sabkha" evaporite-carbonate sedimentation (Mossop, 1975), both of which are considered to be self-limiting due to feedback mechanisms intrinsic to the depositional process. The appeal of the Ginsburg model is its simplicity and its independence from external influences such as spasmodic basin subsidence (a uniform flexural-type of subsidence is envisaged), and eustatic sea-level rises. Wong and Oldershaw (1980) have devised a model for cyclic carbonate sedimentation in Devonian back-reef lagoons that closely parallels the Ginsburg model.

In many respects the Barlow Inlet hemicycles are in accord with the 'Ginsberg Model'. There are, however, two significant discrepancies, as follows:

1. Uniform basin subsidence rates are not compatible with the history of spasmodic epeirogenic block-faulting within the region.
2. Kendall and Schlager (1981) suggested that the "start-up phase" of a carbonate platform following drowning, will show a considerable lag period before optimum carbonate production is achieved. The duration of this lag-period can be quantified accurately only in the Pleistocene-Holocene. Adey (1978) estimated the lag-period for reefs to be on

the order of 500 to 1000 years. How these figures relate to lagoonal environments is not clear. The initial high-energy member of the Barlow Inlet hemicycles was deposited in waters probably 2 to 4 m deep. In accordance with the production lag theory, even at the maximum average subsidence rate of 40 cm/1000 years, a lag period of 5000 to 10000 years would have to have expired before appreciable carbonate sedimentation occurred. In the author's opinion, this period of suppressed organic growth is unrealistic. For example, somewhat over 4000 years ago the Florida Bay area was inundated during the Flandrian transgression, and carbonate sediments were introduced for the first time. Exploitation of this newly created shallow marine area by carbonate-secreting or stabilizing organisms must have been exceptionally rapid, as the rate of sediment filling was greatest in the first 1000 years and decreased considerably thereafter (Scholl, 1966). Average maximum sedimentation rate for the mud banks in Florida Bay is about 45 cm/1000 years. The steady subsidence of the Ginsburg model should produce complete transgressive-regressive cycles, not asymmetric cycles indicative only of shallowing-upward trends.

3.10.2 Episodic Subsidence and Tidal Flat Progradation Hypothesis

In order that a relative sea-level rise does not leave an extensive sedimentary record (transgressive deposits), the rise must be rapid. The "modified Bott" model outlined below credits periods of greatly increased subsidence to accelerated movement on normal faults bounding the Boothia Uplift during times of isostatic re-equilibration. Peak subsidence rates so produced, would have exceeded carbonate production potential, and the sea-floor would have been depressed some distance below sea-level until such time as sediment accumulation outpaced a slackening subsidence rate. Several authors have embraced the concept of episodic subsidence (in conjunction with steady sedimentation) to explain asymmetric cycles in shallow water carbonates. Punctuated aggradational cycles are considered to have developed in this manner (Warshauer and Smosna, 1980). Anderson et al. (1978) and Kasig (1980) also attributed hemicyclicity to episodic subsidence.

The model follows closely the faulted basin model of Bott (1964), and the sequential steps are given below:

1. Mass deficiency in a portion of the crust, results in isostatic inequilibrium.
2. Response to this inequilibrium is uplift, which induces ductile flow in the mantle towards the area of uplift.

3. Following a lag time of thousands or tens of thousands of years (Crittenden, 1967), subsidence occurs in the area of withdrawn subcrustal (mainly mantle) material. Crustal deformation takes place by faulting along pre-existing lines of weakness in the basement. In order for normal faults to develop, the region must be influenced by tensional stress.
4. Isostatic readjustment is complicated by frictional resistance between fault blocks, and as a consequence basin subsidence is episodic. Accelerated subsidence along fault blocks marginal to the Boothia Uplift, could have produced transgression on the platform.
5. Subsidence rate decreases with time allowing a regressive sequence to develop (progradation as in the Ginsburg model). As the basin develops, simple sediment loading also contributes to crustal downwarping, which provides a background subsidence rate.
6. Erosion on the uplift leads to isostatic inequilibrium, and the process repeats itself.

There are several difficulties in applying the 'Modified Bott' model to the Barlow Inlet hemicycles, as follows:

1. A necessary element of the model is an uplifted region proximal to the basin. Such an active uplift should provide considerable volumes of terrigenous sediment to adjacent marine environments (e.g.

carbonate-sand facies of Miall, 1970), but the Barlow Inlet platform succession contains negligible siliciclastic detritus. This might be explained in terms of paleodispersal patterns, although the lower member of the Somerset Island Formation on Somerset Island (also interpreted as a peritidal carbonate succession, and much closer to the postulated uplands), also contains little siliciclastic material.

2. The fact that ~~the~~ hemicycles consistently commence at approximately the same depth suggests that the amounts of individual uplift, and hence subsidence, were very regular. This would seem to be unrealistic for a natural process dependent on so many variables. The Barlow Inlet Platform does, however, record several major back-stepping and fore-stepping events which might correspond to the anticipated irregular tectonic activity.

In light of the available evidence, the 'Episodic Subsidence and Tidal Flat Progradation' model appears to be more appropriate to explain the observed hemicyclicity of the Barlow Inlet peritidal sediments.

3.11 CONCLUSIONS

1. Carbonate platform interior facies form a considerable proportion of the Upper Silurian Barlow Inlet Formation, and are typically arranged in hemicyclic series. Two partially intergradational hemicycle types are recognized, a Restricted Infratidal Hemicycle and a Tidal Flat Hemicycle. Hemicycle thicknesses range from 1 to 23 m in series up to 560 m thick.
2. Complete RI hemicycles contain four members, arranged in shallowing upward sequences. In order, and commencing with the interpreted deepest water member, there is a Biostromal Member, a Sparsely Fossiliferous Mottled Dolomitic Carbonate Member, a Massive Carbonate Mudstone Member and a Laminated Member.
3. The Laminated Member is here interpreted to represent low to mid-intertidal zone deposits, and the Loferite Subfacies of the Massive Carbonate Mudstone Member represents sedimentation in the infratidal to low intertidal zone. The Megalodont Bivalve Subfacies occurs in the shallow subtidal, moderately tranquil back-barrier lagoon. The Biostromal Member is considered to have developed in a proximal back-barrier position within the platform interior lagoon.

4. Complete TF hemicycles contain four members arranged in a shallowing upward sequence. In order, and commencing with the interpreted deepest water member, there is an Oncoidal Carbonate Member, a Mottled Carbonate Member, a Massive Carbonate Member and a Laminated Member.
5. The hemicyclic series were deposited under an arid climate, in the interior of a carbonate platform that was subjected to minimal hydrodynamic stress. The platform was rimmed by crinoid sand shoals and stromatoporoid-crinoid buildups and biostromes.
6. The muddy lagoonal lithofacies of the Barlow Inlet hemicycles can be compared favourably with Florida Bay mud bank and lake sediments, but unlike this Holocene analog the Barlow Inlet facies probably developed in a shore-parallel, as opposed to patch-work, fashion. The Barlow Inlet hemicycles are similar to both Triassic and Mid to Upper Devonian peritidal successions described in the literature.
7. An explanation for the differences between the TF and RI hemicycles need not invoke substantial modification of platform environments. The TF and RI hemicycles contain lithofacies that coexisted, within the platform interior in positions that on average were more distal, and more proximal to the platform margin, respectively.

8. The hemicycles can, ideally, be used to determine relative rates of platform subsidence.
9. Using hemicycle thicknesses, it is estimated that the shelf depths of the Platform Interior were generally shallower than 14 m.
10. An episodic subsidence and tidal flat progradation model best explains the observed hemicyclicity of the Barlow Inlet Platform interior lithofacies. Eustatic sea-level rises are not required by this model.

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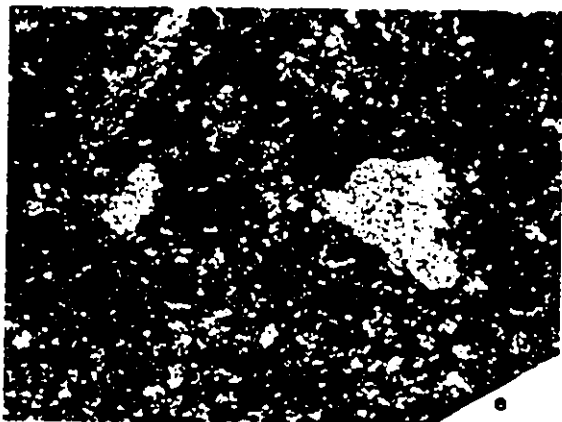
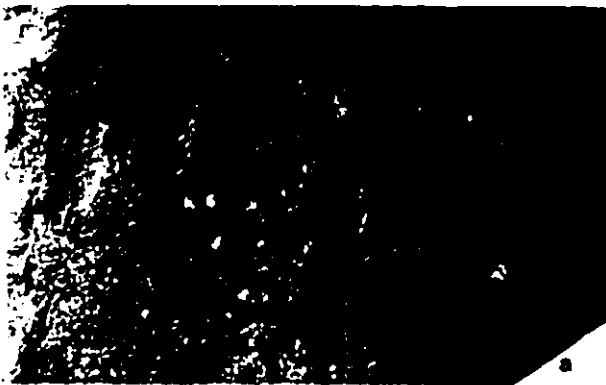
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3.13 PLATES

Plate 3-1

RI Hemicycles : Biostromal Member.

- a Two metre thickness of sparse biostrome of hemispherical to subspherical stromatoporoids at loc. 3. Note rubbly-weathering matrix.
- b Three successive biostromal intervals (indicated by black bars) at loc. 3. Jacob's staff with 0.1 m intervals.
- c Biostromal interval of sparse structure at loc. 3. Note poorly developed bedding. Randy Hyrniuk for scale.
- d Polished slab of biostromal lithofacies exhibiting hemispherical stromatoporoids (S), solitary rugosans (R), and appreciably dolomitized matrix (light colour).
- e Photomicrograph of peel of dolomitized matrix with scattered sand- to rudite-sized allochems of indeterminate origin. Scale 0.05 cm.
- f Polished slab of biostromal lithofacies exhibiting irregular fenestrae. Scale 1 cm.
- g Resistant, cliff-forming massive mudstones (4), overlying a basal biostromal member (3), a transitional packstone (2), and underlying strata of TF series aspect (1). The massive nature of the mudstone could be confused at a distance with reef core facies. Combined thickness of intervals 2 and 3 is 25 meters.

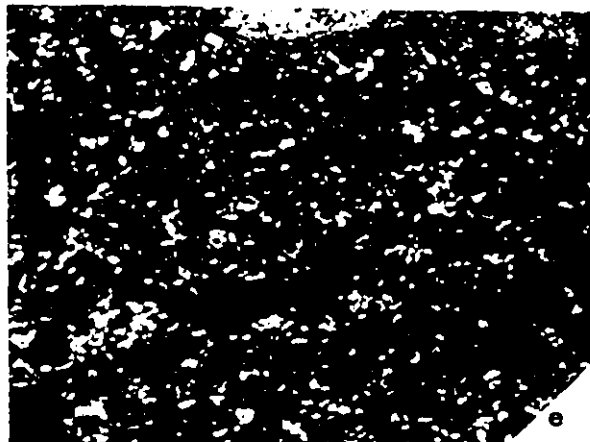


PL.3-1

Plate 3-2

RI Hemicycles : Sparsely Fossiliferous, Mottled Dolomitic Carbonate Member.

- a Gradational transition from sparsely fossiliferous mottled dolomitic carbonate lithofacies to partly mottled, massive carbonate mudstone lithofacies. Scale 15 cm.
- b Abrupt contact (arrows) between mottled and unmottled lithofacies. Scale 15 cm.
- c Acetate peel of the peloidal matrix of lime mudstone nodules. Note subhedral dolomite rhombs and recrystallized ostracode fragment. Scale 0.05 cm.
- d Polished slab exhibiting abrupt contact between mottled and fenestral mudstone lithofacies. Scale 1 cm.
- e Acetate peel showing dolomitization of matrix. Note subhedral rhombs are of equal size in both nodule (top of plate) and matrix. Scale 0.05 cm.
- f Polished slab exhibiting solution-seam in inter-nodule matrix. Scale 1 cm.

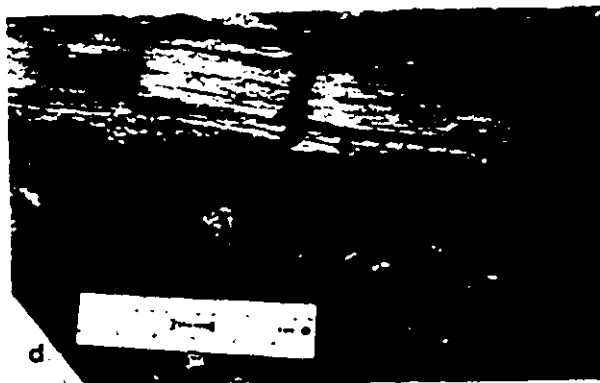


PL.3-2

Plate 3-3

RI Hemicycles : Massive Carbonate Mudstone Member.

- a Field appearance of massive carbonate mudstone member at loc. 4. Note parallel fractures. Guru Packard for scale.
- b Subparallel lenticular fractures in massive mudstone lithofacies at loc. 8. Height of section portrayed is approximately 10 m.
- c Fractured surface of homogeneous massive lime mudstone. Hammer 31 cm long.
- d Gradational contact (arrow) between underlying massive carbonate mudstone and overlying laminite. Laminations emphasized by subparallel stylolites). Scale 15 cm.
- e Cliff of RI series at loc. 3. Arrow points to unit of massive mudstone that exhibits some pinch and swell in its thickness over a lateral distance of 200 metres. Total stratigraphic thickness is approximately 80 metres.
- f Gradational transition (arrow) from faintly mottled to unmottled massive carbonate mudstone; latter overlain abruptly (at hammerhead) by sparsely fossiliferous mottled dolomitic carbonate lithofacies. Hammer 31 cm long.



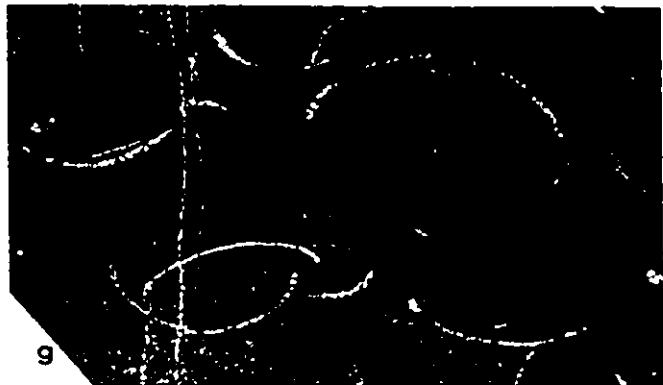
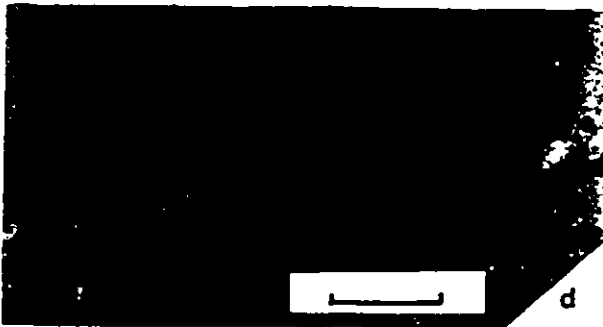
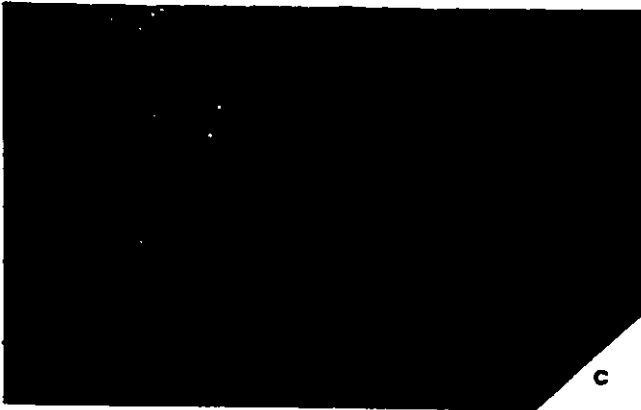
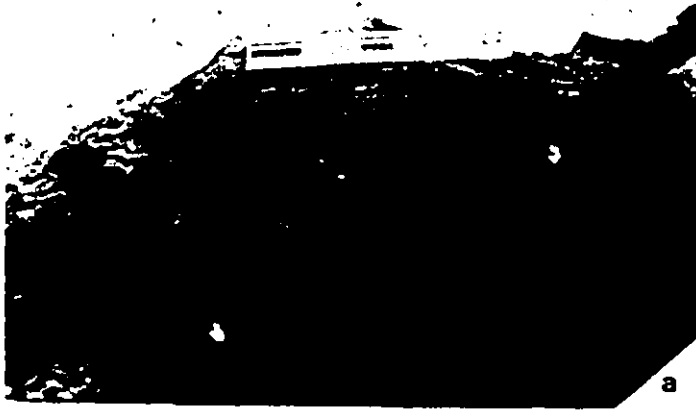
PL.3-3

Plate 3-4

RI: Hemicycles : Dense Mudstone Subfacies.

Bar scales are 1 cm (unlabelled) or tenths or hundredths of 1 cm (labelled).

- a Sample of typical dense mudstone subfacies from loc. 5. Scale 15 cm.
- b Faint mottling associated with dense mudstone subfacies (polished slab).
- c Acetate peel of micrite/microspar of dense mudstone subfacies.
- d Amphiporids in a floatstone with a dense mudstone matrix.
- e Fitted fabric developed in dense mudstone matrix (polished slab).
- f Solution mold of low-spined gastropod in dense mudstone subfacies (polished slab).
- g Abundant single valves of ostracodes in dense mudstone matrix (acetate peel).



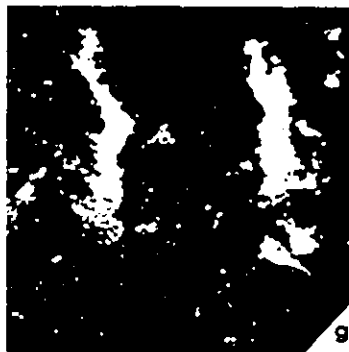
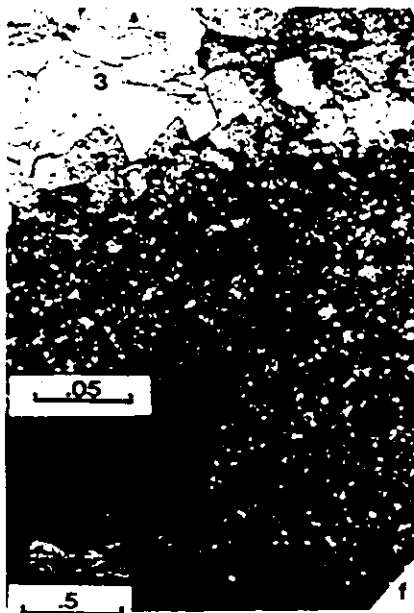
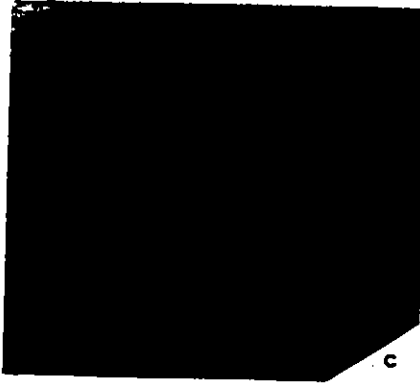
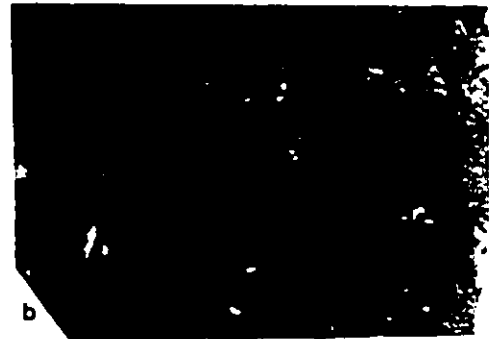
PL.3-4

Plate 3-5

RI Hemicycles : Loferite Subfacies - 1

Bar scales are 1 cm (unlabelled) or tenths or hundredths of 1 cm (labelled).

- a Field photograph of typical fenestral mudstone. Scale 15 cm.
- b Irregular fenestrae (polished slab).
- c Laminoid fenestrae (polished slab).
- d Tubular fenestrae (polished slab).
- e Mixture of fenestral types (polished slab).
- f Dolomite mosaic geopetal sediment (1), coarse equant dolospar cement (2), and clear coarse equant calcite spar (3). (Acetate peel) Inset shows wavy lamination in dolomite mosaic (arrow).
- g Micrite/microspar geopetal sediment. Arrow points to contact between 2 generations of sediment (thin section).
- h Polished slab showing laminoid, tubular and irregular fenestrae, all filled with dolomite mosaic sediment.

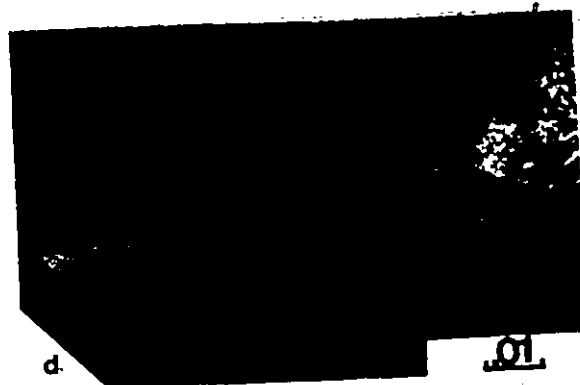
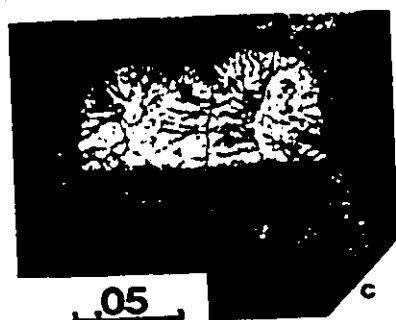
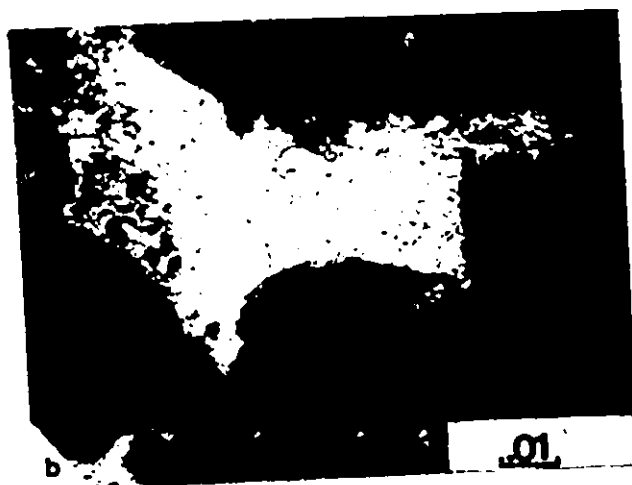


PL.3-5

Plate 3-6

RI Hemicycles : Loferite Subfacies - 2
Bar scales are tenths or hundredths of 1 cm.

- a Clear, coarse equant calcite spar filling fenestral pore, partially filled by microspar internal sediment (thin section).
- b Clear, coarse, equant calcite spar in direct contact with irregular fenestral walls (thin section).
- c Void with different sediment and cement generations. Micrite matrix (1), ?pseudospar (2), geopetal sediment (3,4), pendant calcite spar (5). Final occlusion of void is by clear, coarse, equant spar. Note fine, equant spar layer resting on sediment layer 4 (acetate peel).
- d Fenestrae(?) filled by drusy spar (acetate peel).
- e Two generations of void-filling spar. Splays of bladed spar (1), are enclosed by coarse equant spar forming local patches (2) of poikilitic texture (acetate peel).
- f Micrite (1) and ?pseudospar (2) surrounding pore filled by drusy spar (acetate peel).
- g Solution mold of gastropod (note absence of shell wall) (acetate peel).
- h Drusy spar in intra-allocherm (ostracode) cavity (acetate peel).



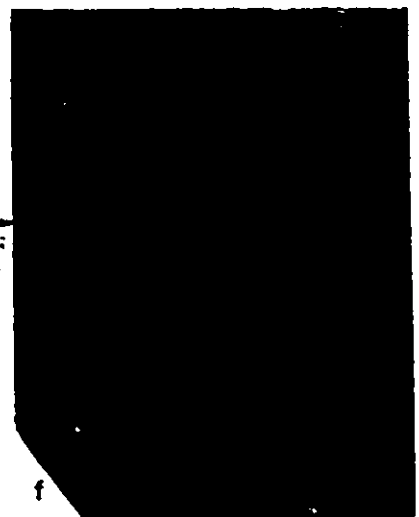
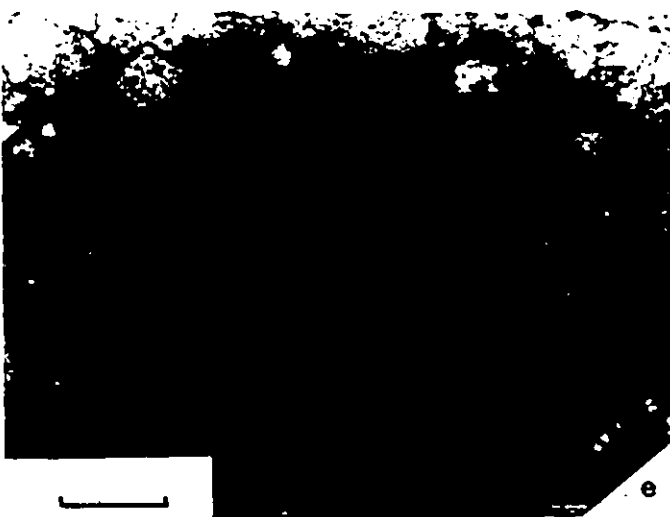
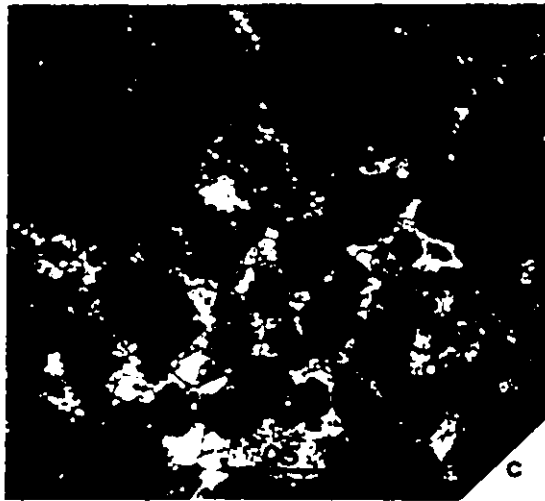
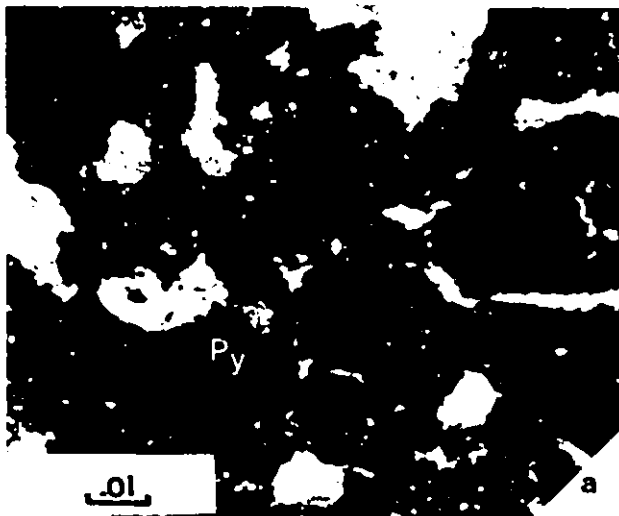
PL.3-6

Plate 3-7

RI Hemicycles : Loferite Subfacies -3

Bar scales are 1 cm (unlabelled) or tenths or hundredths of 1 cm (labelled).

- a Pyrite (Py) partially occluding pore space and disseminated in matrix (thin section).
- b Peds (or possibly pseudopeloids) intimately associated with fenestral development. Note bleached rims (arrow) of peds (polished slab).
- c Peds and fenestral development. Matrix does not appear to have a relic (compressed) peloidal texture (acetate peel).
- d Close-up of figure 3-7c showing subequant calcite spar associated with fenestrae (acetate peel).
- e Fitted fabric of fenestral mudstone (polished slab).
- f Fenestral wackestone rich in amphiporids (polished slab).



PL.3-7

Plate 3-8

RI Hemicycles : Megalodont Bivalve Subfacies.
Bar scales are 1 cm (unlabelled) or tenths or hundredths of
1 cm (labelled).

- a Bedding plane exposure of in situ articulated megalodont bivalves. Note subvertical orientation of commissure planes and the mottled appearance of matrix. Scale 15 cm.
- b Field occurrence of megalodont bivalves at loc. 5. Hammer 31 cm long.
- c Partial dolomitization (light colored mottles) of matrix between recrystallized valves of megalodont bivalve (polished slab).
- d Articulated valves of Megalamoidea (polished slab).
- e Vertical section through paired and single valves of megalodonts. Note single valves are preferentially oriented concave-up. Loc. 8.
- f Typical peloid-rich (pellets possibly from clams) packstone (pelmicrite to biopelsparite) with ostracode fragments (thin section).
- g Possible alga (A) (acetate peel).



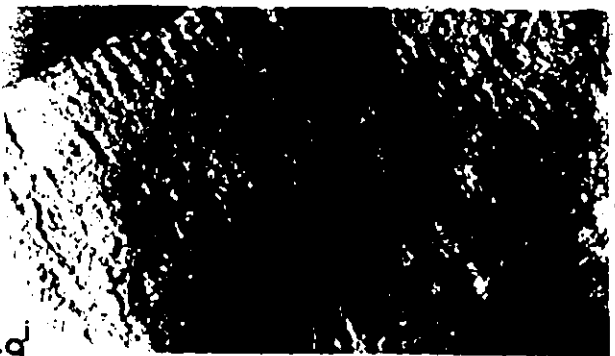
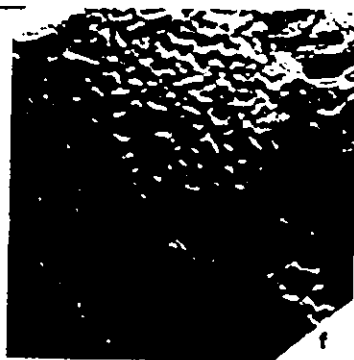
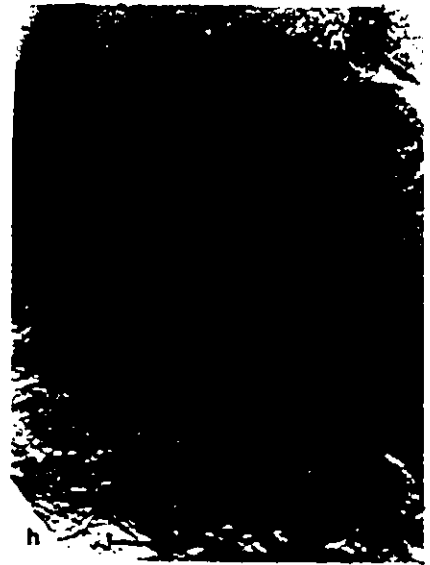
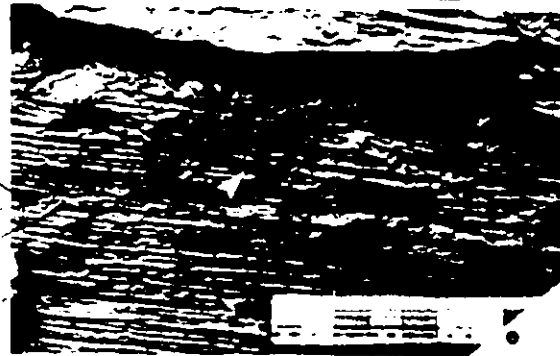
PL.3-8



Plate 3-9

RI and TF Hemicycles : Laminated Member - 1
Bar scales are 1 cm.

- a Compositional laminae (polished slab).
- b Laminoid fenestrae superimposed on compositional laminae (polished slab).
- c Subparallel stylolites superimposed on compositional laminae to produce 2nd order laminae (polished slab).
- d Laminoid fenestrae filled in part by calcite spar (dark) and dolospar (light) (polished slab).
- e Large-scale disconformity (arrow) within algal-laminated sediments at loc. 7. Scale numbered in mm below.
- f Surface crenulae with zig-zag pattern on cryptalgal mat (hand specimen).
- g Minor doming and truncation (arrow) of laminae (polished slab).
- h Upper surface of cryptalgal-laminated dolosiltite exhibiting subcircular (rounded polygons?), low-amplitude stacked inverted hemispheres. Hemispheres may have been isolated by desiccation of originally continuous smooth mat.
- i Surface papillae of cryptalgal mat.
- j Parallel crenulae on surface of cryptalgal mat.



PL.3-9

Plate 3-10

RI and TF Hemicycles : Laminated Member - 2
Bar scales are tenths or hundredths of 1 cm.

- a Alternated laminae of subhedral, idiotopic dolomite (1), and a mixture of calcite spar and dolomite rhombs (2). Note diffuse boundaries between laminae (thin section).
- b Disrupted lamination, possibly representing biogenic reworking of sediment (thin section).
- c Alternated laminae of amorphous organic and/or clay-rich material (1) and subhedral idiotopic dolomite (2). Note lack of grading and abrupt contact (arrow) (thin section).
- d Alternated dolomite (light) and calcite-rich (dark) laminae. Note small-scale doming of laminae at top (thin section).
- e Alternated organic/clay-rich and dolomite laminae. Note erosional discontinuity at top (top section).
- f Outcrop of massive calci/dolosiltite at loc. 7. Hammer 31 cm long.
- g Exposure of laminated dolosiltite (1), massive dolosiltite (2), and oncolitic carbonate (3). Note gradational contact between 1 and 2, and the abrupt contact between 2 and 3. Rod marked in 0.1 m intervals.

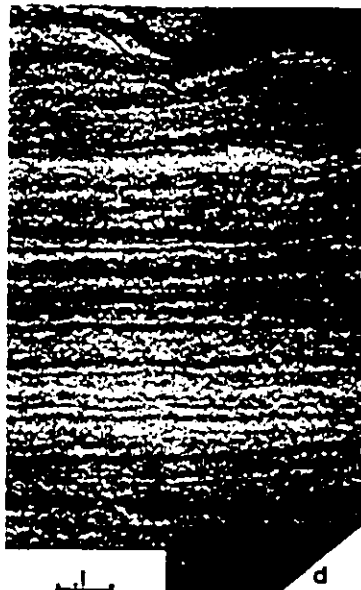
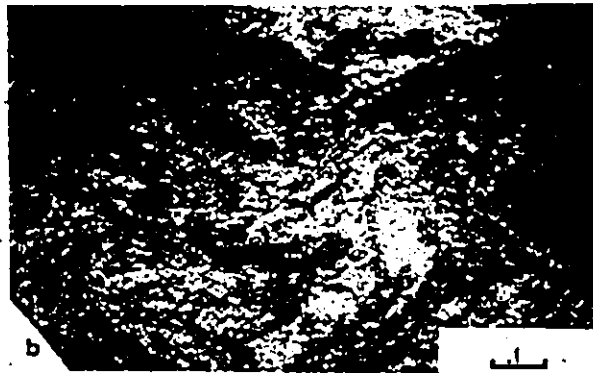
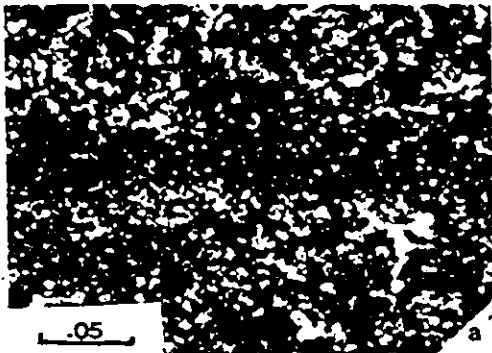
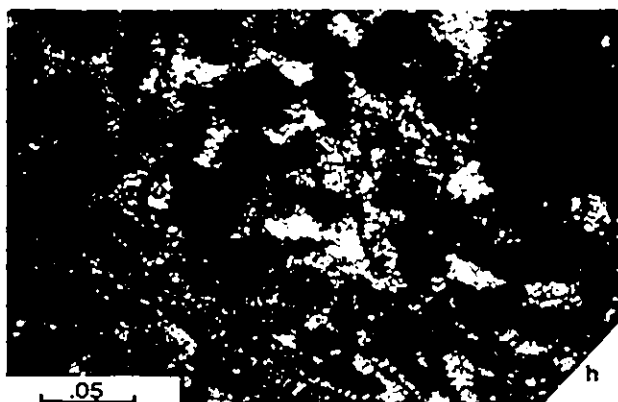
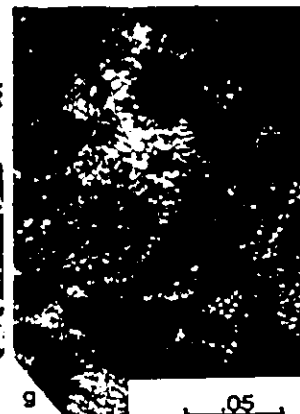
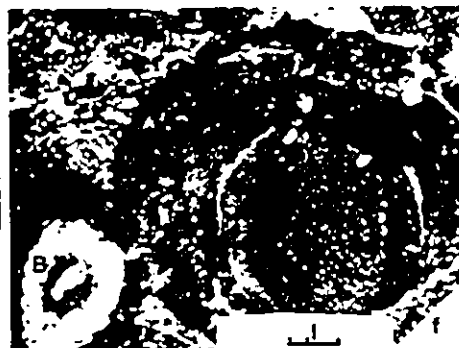
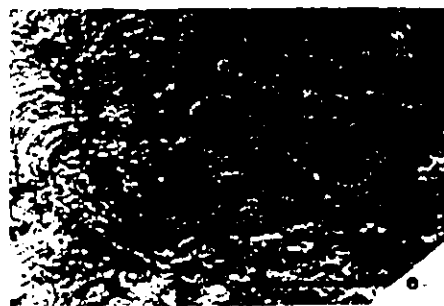
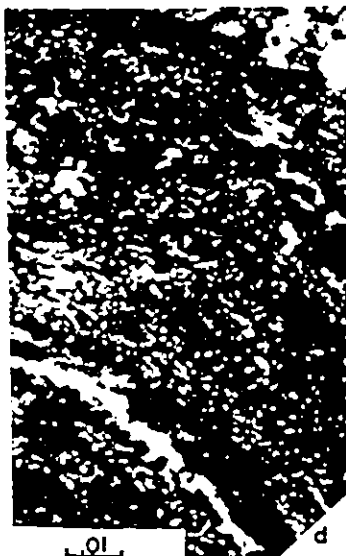
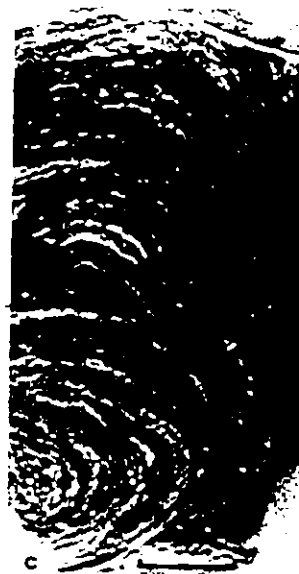


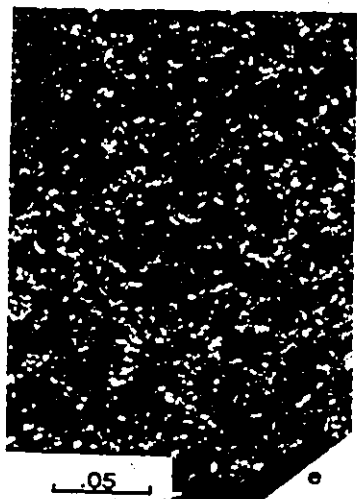
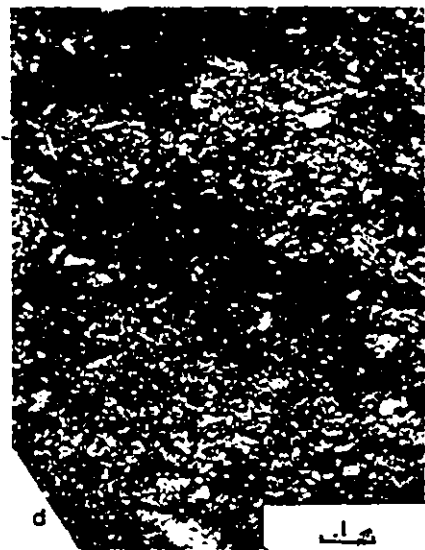
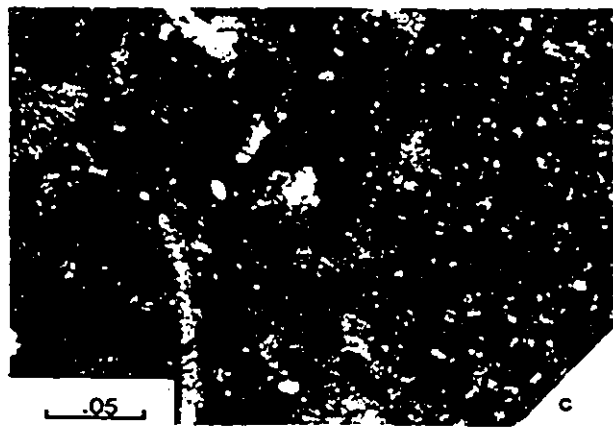
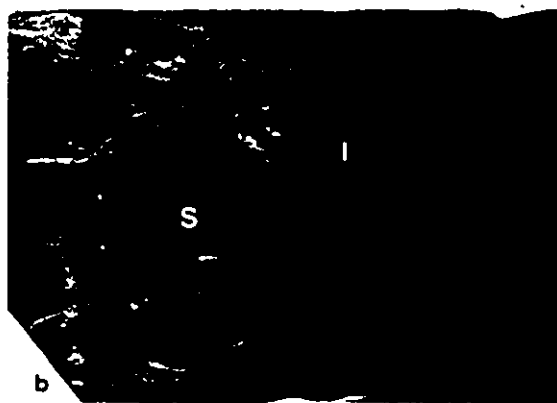
Plate 3-11

TF Hemicycles : Oncoidal Carbonate Member.

Bar scales are 1 cm (unlabelled) or tenths or hundredths of 1 cm (labelled).

- a Field photo of oncoidal carbonate facies exhibiting algal oncoids (1), lime mudstone to packstone nodules (2), and dolomitic matrix (3). The abundance of oncoids yields an 'organic mottled' appearance to the rock.
- b Type C oncoids with variably-shaped nuclei (hand specimen).
- c Type R oncoid (polished slab).
- d Molds of algal filaments within oncoid laminae (acetate peel).
- e Type C oncoid with a highly-spired gastropod shell serving as a nucleus (hand specimen).
- f Two oncoids with intraclastic (I), and skeletal (B) nuclei (acetate peel).
- g Ostracodes (O) and indeterminate peloids cemented by isopachous, fibrous calcite (arrow). From limestone nodule (acetate peel).
- h Ostrocode carapace (O) and peloid (P) in nodule composed of biosparite (acetate peel).
- i Dolomitic matrix with corroded bioclasts and peloids (acetate peel).





PL.3-12

Plate 3-12

TF Hemicycles : Mottled Carbonate Member.

Bar scales are 1 cm (unlabelled) or tenths or hundredths of 1 cm (labelled).

- a. Hand specimen of mottle carbonate lithofacies. Arrow points to pods of lime mudstone matrix isolated at the margin of a nodule, by seam solution. Vertical burrow tubes (S) are filled by coarser bioclastic packstone.
- b. Polished slab of mottled carbonate lithofacies exhibiting encrusting stromatoporoid (S), and either micrite intraclasts (I) or burrow fills (e.g. Chondrites) in a pelsparite nodule.
- c. Bioclastic wackestone nodule (clotted appearance of matrix is suggestive of pelsparite) in contact with dolomitic matrix at right (acetate peel).
- d. Photomicrograph of lime pods separated by solution seams as in figure a (acetate peel).
- e. Pelsparite matrix of limestone nodule. Note subhedral to subhedral dolomite rhombs (acetate peel).
- f. Possible meniscus cements (arrows) in a sheet crack (acetate peel).
- g. Solution seam and patterned carbonate (pc) in polished slab of mottled carbonate lithofacies.

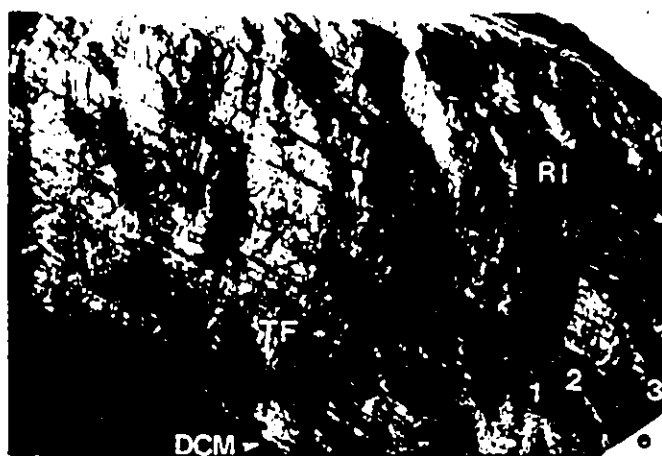
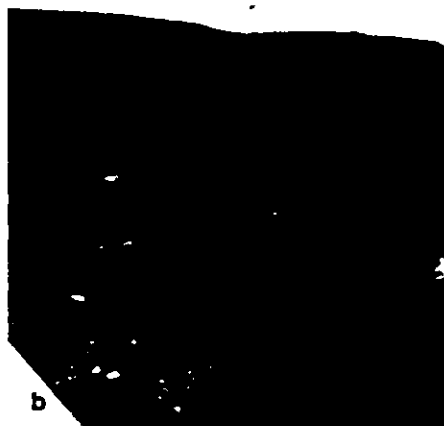
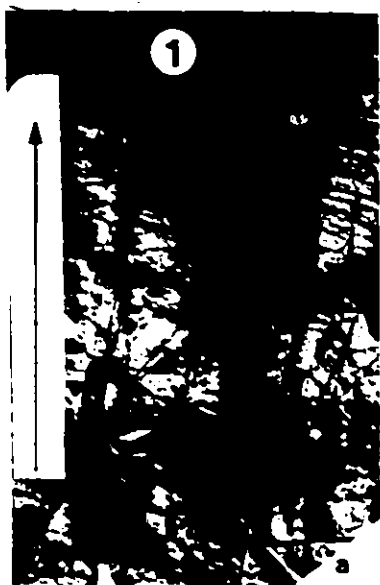


Plate 3-13

Hemicycles : Environmental Synthesis.

- a. Complete TF hemicycle with oncoidal carbonate (1), mottled carbonate (2), massive carbonate (3), and laminated members (4). Jacob's staff with 0.1 m intervals
- b. Four superposed hemicycles (TF affinities) at loc. 7. Height of section is 20 m.
- c. Pinch and swell in massive lime mudstone unit (arrows) at loc. 3. Major slip plane just above this unit, is coincident with a cryptalgal laminite lithofacies. Stratigraphic thickness of cliff is 80 m.
- d. Possible channel structures in massive lime mudstone lithofacies at loc. 4. Arrow marks lower contact of one of ?channel structures. Stratigraphic thickness shown is approximately 15 m.
- e. TF and RI series at loc. 4. Upper contact of TF series (1), lower contact of first biostromal lithofacies (2), and lower contact of first massive lime mudstone facies (3) are indicated. DCM (diverse community mud mound) indicated by white arrow. Note well-stratified nature of entire succession. Total stratigraphic thickness is 180 m.
- f. TF-RI transition at loc. 4. Numbers as in figure e. Height of cliff approximately 60 m.

IV

PAPER IV: PRELIMINARY SURVEY OF BIOGUMULE
(BUILDUP/BIOSTROMAL) FACIES FROM THE UPPER
SILURIAN OF ARCTIC CANADA

Abstract

The upper member of the Upper Silurian Barlow Inlet Formation of Arctic Canada represents a major carbonate platform complex, over 1.2 km thick. Biocumules, herein defined as variably-shaped deposits of concentrated in-situ organisms and contiguous sediment (collectively embracing biostromes and buildups), occur throughout this complex. They are considered to have developed in depositional environments ranging from medial foreslope to restricted lagoon. Ten biocumule facies are recognized and include: lithistid sponge dominated mud mounds, mucophyllid mud mounds, diverse community mud mounds, fasciculate rugose coral biostromes, stromatoporoid-crinoid rubble skeletal buildups, dolomitized massive skeletal (?crinoid) biocumules, lamellar stromatoporoid biostromes, massive stromatoporoid biostromes, dendroid stromatoporoid biostromes, and Coenites -capped fenestral mud mounds. In addition several biocumules exhibit internal vertical zonation of biofacies, and are here described under the term 'zoned biocumules'.

Buildups and biostromes of Late Silurian-Early Devonian age are poorly represented in the literature, and detailed

descriptions of complete platform complexes from continental margin settings are unknown. Biocumules of the Platform Margin are most closely comparable to the reef succession of Gotland, Sweden. Platform Interior facies bear close resemblance to Mid- and Upper Devonian backreef sequences. Basin Slope biocumules show similarities to Late Silurian carbonate ramp buildups of Arctic Canada and foreslope sponge mud mounds of the European Jurassic.

4.1 INTRODUCTION

The Upper Silurian Barlow Inlet Formation, occurring within the Innuitian Orogen of Arctic Canada, contains a diverse assemblage of hitherto unstudied stratigraphic reefs. Over 70 structures broadly classified as either buildups or biostromes (Table 4-1), were examined in the course of stratigraphic studies on eastern Cornwallis Island (FIG. 4-1). These structures, together with associated and volumetrically superior lagoonal and foreslope facies, form the 1300 m thick Barlow Inlet rimmed-shelf carbonate complex (Barlow Inlet Platform). Sediments of this shelf complex range in age from latest Ludlow to earliest Gedinne (latialata to woschmidtii conodont Zones - Thorsteinsson, 1980). Within the study area, reef development ceased in earliest Devonian time due to emergence of the Barlow Inlet shelf concomitant with a major pulse of the Cornwallis Disturbance (FIG. 4-2).

The contribution of this study lies in two domains. Firstly, few studies have focused on carbonate buildups within the thick sedimentary succession of the Innuitian Orogen¹³ and adjacent areas of the Arctic Archipelago and Greenland. Permian mounds have been examined by Davies and Nassichuk (1973), and Davies (1974, 1976); Upper Devonian

¹³ The Innuitian Orogen is composed in part of an older geosyncline (the Franklinian Geosyncline), Proterozoic to Devonian in age, and a younger successor basin (the Sverdrup Basin). A considerable proportion of the sedimentary succession is carbonate in nature.

reefs by Embry and Klovan (1971); and Silurian buildups by Mayr (1974), Narbonne and Dixon (1978, in press), Packard and Dixon (1979, 1982), and Hurst (1980, 1981, 1982).

Definitions

- Carbonate buildup - a carbonate mass which differs in nature to some degree from the surrounding deposits and overlying rocks, is typically thicker than equivalent carbonates, and probably stood topographically higher than the surrounding sediment during some time in its depositional history (Heckel, 1974, p.91).
- Reef - a buildup that displays 1) evidence of either potential wave resistance or growth in turbulent water which implies wave resistance, and 2) evidence of control over the surrounding environment (Heckel, 1974, p.96).
- Reef Complex - aggregate of reef, forereef and back-reef facies (Playford, 1969, p.8).
- Biostrome - purely bedded structures ... consisting of and built by sedentary organisms that do not swell into a mound-like or lens-like forms (James, 1983, p.379). A bedded structure ... which was built by sedentary organisms. They lack moundlike or lenslike form (Dictionary of Geological Terms, American Geological Institute, 1976, p.45).

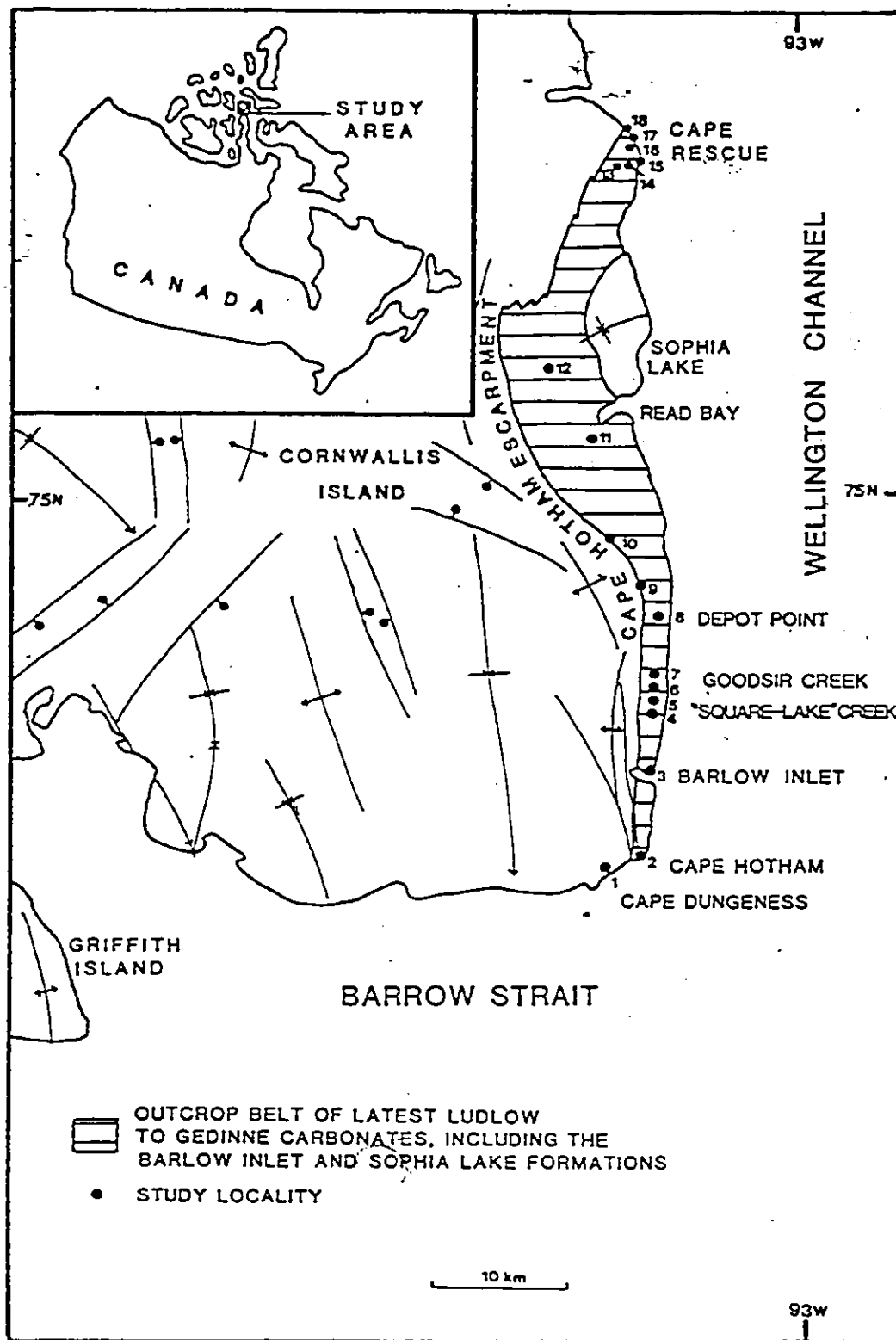


Fig. 4.1 Study area showing localities of examined biocumules.

More generalized information on carbonate platform margin facies in the arctic has been published by Trettin (1979), Sodero and Hobson (1979), Hurst (1983), Mayr (1974), McGill (1974), and Poey (1982). The interpretation of shelf margin facies is an integral part of understanding overall basin evolution (e.g. Trettin, 1979; Read, 1981; Hurst, 1983).

Secondly, our knowledge of Silurian stratigraphic reefs¹⁴ has been biased by attention paid to classic localities in the Great Lakes region (North America), Baltic Region (particularly Gotland, Sweden), and Wenlock Edge, England. Only recently have Silurian reef studies been undertaken outside these areas (e.g. Bourque, 1979; Hurst, 1980; Narbonne and Dixon, in press). Consequently, although considerable data are available on intracratonic (and generally pre-Pridolian) buildups and biostromes, little work has been done on latest Silurian stratigraphic reefs at cratonic margins. As stated by Wilson (1975, p.105), "Little is actually known about Silurian shelf margins". In particular, reefs considered¹⁵ to be latest Silurian-earliest Devonian in age are very poorly represented in the literature, and perhaps in the stratigraphic record (e.g. see Heckel, 1974; Burchette, 1981; Wilson, 1975, p.119). This gap hampers the analysis

¹⁴ as documented in the English language literature.

¹⁵ Some of the Great Lakes buildups traditionally viewed as Middle Silurian may be as young as Late Silurian (see Shaver, 1977).

of evolving organic strategies and adaptations within the reef environment, between mid-Silurian and mid-Devonian time.

The purpose of this paper is to describe the structure, composition, and paleoenvironments of the principal 'reefal' facies within the Barlow Inlet Formation, to provide an integrative model for the biostromes and buildups of the Barlow Inlet rimmed-shelf complex, and to compare and contrast this model with known Silurian and Devonian carbonate complexes.

4.1.1 Methods and Terminology

This study represents a preliminary survey of 'reefal' facies in the Barlow Inlet Formation. Time constraints and the scale of the project did not permit truly detailed study of the many buildups and biostrome structures intersected by stratigraphic sections. Nevertheless every attempt was made to trace these structures laterally to ascertain their (2-D) geometries, to record variation (vertically and laterally) in their biotic components, and to systematically sample all facies observed. In select instances, a) semi-quantitative compositional data were obtained by point counts on outcrops using rectangular grids, b) unbiased samples were taken at predetermined, closely-spaced stratigraphic intervals for the entire thickness of some structures (up to 20 m at 1 m intervals), and c) small mounds were excavated from overriding talus cover and mapped to determine degree of coalescence.

Retrieved samples were cut, polished and etched, and large acetate peels were made (5X5 cm., 5X3's, 3X3's). Select thin-sections and polished sections were stained with Potassium ferricyanide and Alizarin-red. In excess of 400 standard thin-section-area equivalents were examined. Point counts on large peels were used to confirm field estimates of proportion of constituent elements. The carbonate classifications of Dunham (1962) and Jones et al. (1978), or modifications thereof, are used to describe 'reefal' and 'non-reefal' lithologies (Table 4-2).

Descriptive biocumule terminology

Sedimentological roles of
biocumule-associated
organisms:

Bioclasts
Sediment bafflers
Inertial stabilizers
Encrusters
 a) Sediment binders
 b) Cement encrusters
 c) Organism encrusters
 (living, necros)
Symbionts

Biocumule Structures
(from Riding, 1977):

Solid
Frame
Dense
Sparse

Biocumule limestones
(modified after Embry &
Klovan, 1971):

Autochthonous: Bafflestone
 Bindstone
 Framestone

 Inertiastone

Allochthonous: Floatstone
 Rudstone

The 'reef' terminology used in this study is essentially in agreement with that presented, and argued for, by Heckel (1974), Wilson (1975, p.20-24)¹⁶ and Longman (1982, Table 1). Skeletal growth forms are described using a composite nomenclature based on Abbott (1973), Kobluk (1975), Tsien (1971), amongst others. Pertinent terms and their definitions are given in Table 4-3..

A clear distinction is not always made between form (growth) and function (sedimentological) of colonial skeletal metazoans, resulting in ambiguity in the usage of certain terms. For example, isolate massive colonial metazoans are often categorized as framebuilders even though they do not form an integral part of any frame structure. Conversely, many baffling or binding organisms that do construct a frame structure are not labelled as framebuilders, even though they fulfill this role. There is a tendency to confuse the stabilization strategy or effect of individual organisms, with the product of the intergrowth of a community of such organisms. Frame construction is the (fortuitous?) result of intimate intergrowth of organisms and is the collective responsibility of organisms which, individually, seek to stabilize their position within a given habitat by encrusting or binding neighbouring substrates, or baffling introduced sediment. In this sense frame construction is a secondary role of in-situ organic

¹⁶ The term 'platform' is here used in the sense of Wilson (1975), not Read (1981).

Descriptive terminology for skeletal growth forms.

MASSIVE	Hemispherical columnar conical Subspherical Bulbous Irregular Tabular
CYLINDRICAL (greater than 2cm in diameter)	
DENDROID*	Rod-like (few to no branches) Cactoid (few branches) Ramoses (many branches)
ADNATE	Lamellar Subconcentric

* in this paper, several qualifying terms are used to describe how massive the skeletal structure is, particularly with regards to dendroid forms.

e.g. protruberant

robust

delicate

growth (Table 4-4). Primary sedimentologic functions (often involving stabilization strategies or effects of organisms), are listed in Table 4-4. The term 'inertial stabilizer' is introduced to convey the ability of some organisms to stabilize subjacent sediment, simply by their mass. Table 4-5 lists the interpreted sedimentological functions of the principal skeletal constituents in one particular buildup type, using the symbols from Table 4-4.

The reef limestone classification of Embry and Klovan (1972) is modified (Table 4-2) to incorporate lithologies with conspicuous and volumetrically significant autochthonous skeletal elements that form sparse to dense, but not frame, structures. The terms floatstone and rudstone are retained for predominantly allochthonous sediments. A new term, 'biocumule', is introduced herein, to provide a collective label that embraces any concentration of macroscopic, in-situ organisms and contiguous sediment, regardless of structural wave resistance or original topographic relief. A formal definition of this new term follows;

biocumule - a discrete stratigraphic or lithofacies unit that is clearly distinguished from supra-, sub-, and adjacent units on the basis of either 1) its concentration of predominantly autochthonous skeletal material (generally of rudite size or larger), or 2) evidence of inferred binding and/or baffling

Table 4-4

Categories of sedimentological function for
biocumule-dwelling organisms.

Stabilization Strategy (Primary Sediment- ological Roles)	: Sediment Bafflers	Bf
	Inertial Stabilizers	Is
	Encrusters	E
	Sediment Substrate Binders	Es
	Cement Substrate Encrusters	Ec
	Organism Substrate Encrusters	Eo
	Infaunal Dwellers	Di
	Borers	Dbo
	Burrowers	Dbw
	Symbionts	Ds
Potential Secondary Function	: Bioclast	Sc
	Promotion of Substrate Cementation	Cs
	Framebuilder	Fm

activity of a concentration of non-skeletal organisms.

The term biocumule emphasizes the aspect of in-situ organic growth, but places no restrictions on the geometry of the resulting sedimentary body. Thus the term includes biostromes¹⁷ (whereas the term 'carbonate buildup' does not), but does not include loose, hydrodynamically-piled sediment mounds (whereas 'carbonate buildup' does). FIG. 4-3 diagrammatically illustrates the relationship between the two terms. Heckel (1974) has interpreted 'stratigraphic reef' to encompass both biostromes and buildups, and thus this term would also include biocumules. The original definition of stratigraphic reef includes the phrase "thick laterally restricted masses of ... carbonate rock" (Dunham, 1970), and it is questionable whether such a definition can be appropriately applied to essentially sheet-like banks and biostromes. Nevertheless pending further revision of the terminology, it is accepted here as the all-embracing generic term.

4.1.2 Distribution and Classification of Biocumules

Biostromes and buildups occur throughout the north-south outcrop belt of the Barlow Inlet Formation (FIG. 4-4). Excellent exposure allowed uniform spacing of studied stratigraphic sections.

¹⁷ as defined in the AGI glossary, and as used by Nelson et al. (1962), James (1983).

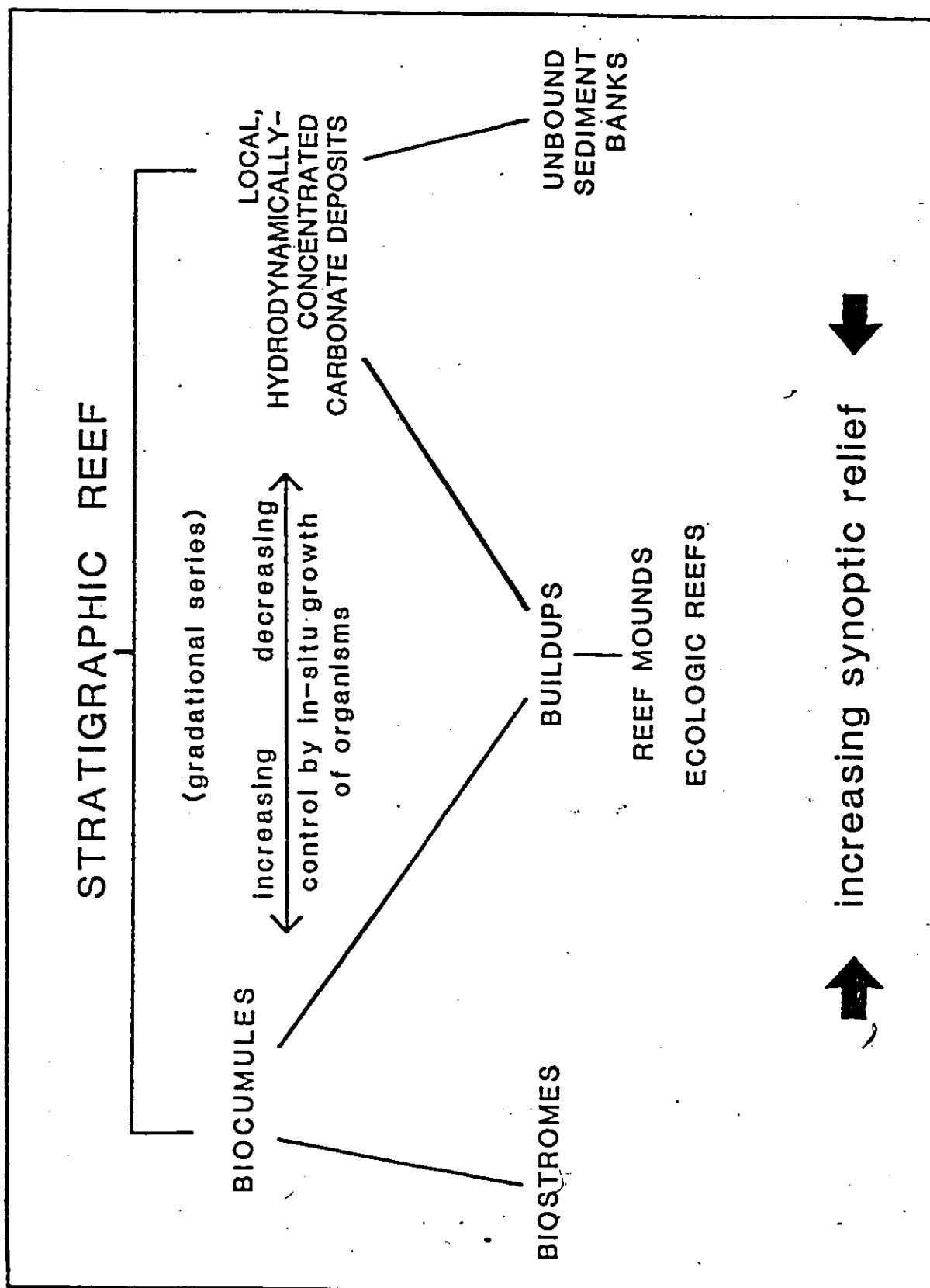
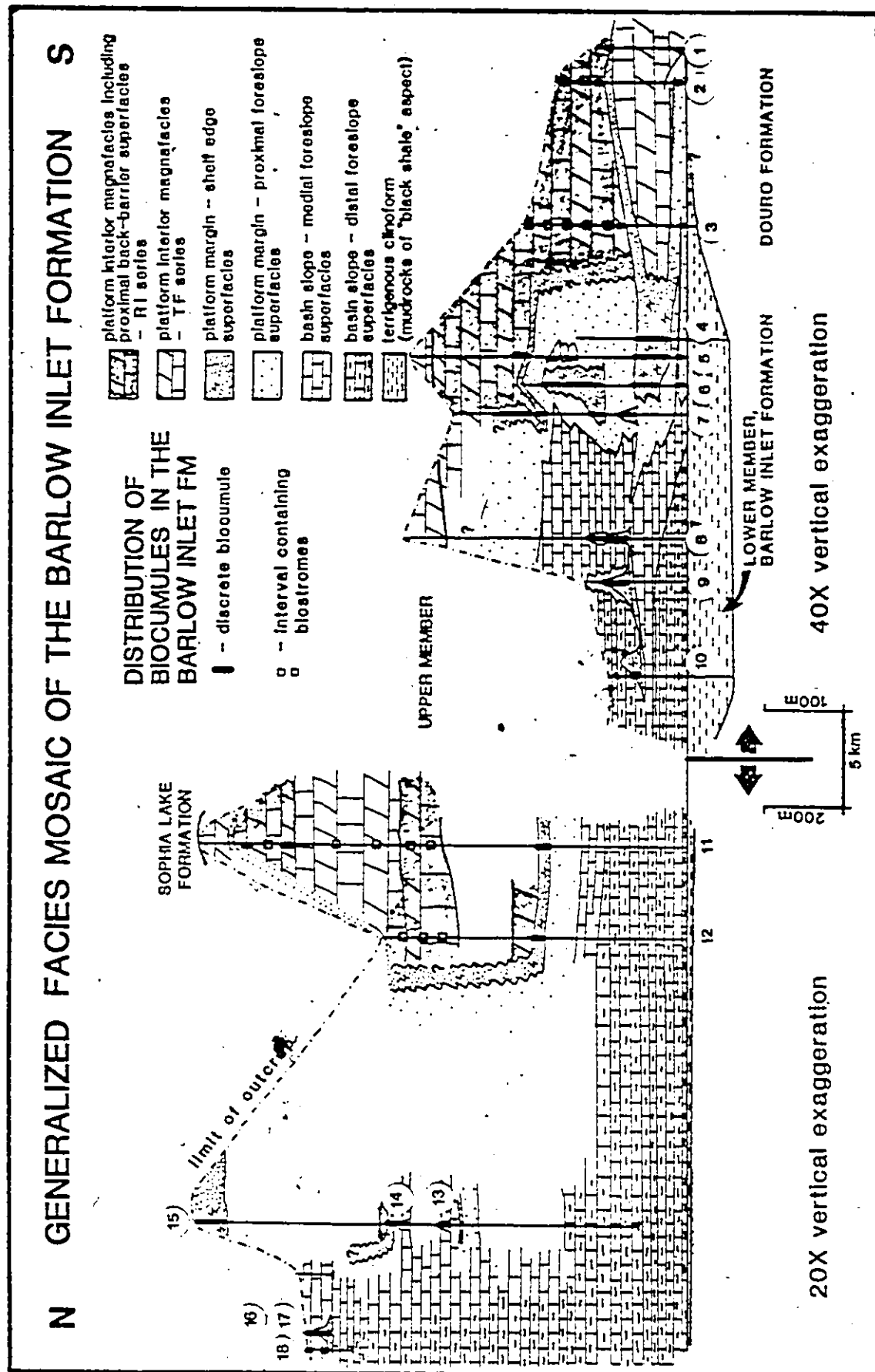


FIG. 4-3 Relationship of biocumule to other "reef" terminology.



Nonetheless the upper part of the Barlow Inlet Formation does not outcrop at or south of loc. 10, and therefore most studied biocumules occur in the lower half of the formation. With the exception of loc. 10, sections were not chosen specifically for their known or postulated abundance of 'reefal' structures, but virtually all sections intersect some biocumules (FIG. 4-4). The reader is referred to Paper 5 for a detailed description of the stratigraphic framework for the biocumules of the Barlow Inlet Formation.

The more than 70 biocumules examined vary considerably in predominant lithological constituents, geometry (observed and assumed), size, skeletal/matrix ratios, nature of biotic components, and degree of diagenetic alteration. Difficulties in attempting to categorize biocumules are analogous to those faced by paleontologists who must decide whether certain fossil forms are sufficiently different to warrant erection of a new species, or in other instances, whether it is valid to erect a new species based on a unique but solitary, and perhaps poorly-preserved specimen. Given the preliminary level of investigation; it has been decided to err on the side of oversimplification and the following 11 biocumule categories are recognized:

1. Lithistid sponge-dominated Mud Mounds
2. Mucophyllid Mud Mounds
3. Diverse Community Mud Mounds
4. Fasciculate Rugose Coral Biostromes

5. Stromatoporoid-Crinoid Rubble Skeletal Buildups
6. Dolomitized Massive Skeletal (?Crinoid) Biocumules
7. Lamellar Stromatoporoid Biostromes
8. Massive Stromatoporoid-Dominated Biostromes
9. Dendroid Stromatoporoid Biostromes
10. Coenites -capped, Fenestral Mud Mounds
11. Zoned Biocumules

The eleventh category is particularly artificial as it incorporates diverse, vertical biocumule successions, considered together solely for descriptive purposes. The above listed categories are not regarded as discrete 'reef types', but rather as intergradational biocumule facies in the rimmed shelf margin mosaic.

4.1.3 Regional Basin Evolution and Paleogeographic Setting

During Douro Formation (FIG. 4-2) time, marine carbonate deposition in the vicinity of present-day Somerset and Cornwallis Islands took place on a northerly-inclined ramp (the Douro Ramp) that bordered and imperceptibly merged downslope with a large, deep, back-barrier basin (Cape Phillips Back-Barrier Basin) (FIG. 4-5). Epeirogenic rise of a major north-south trending basement uplift (the Boothia Uplift) in latest Ludlow time, initiated profound changes in the structural configuration of the carbonate ramp.

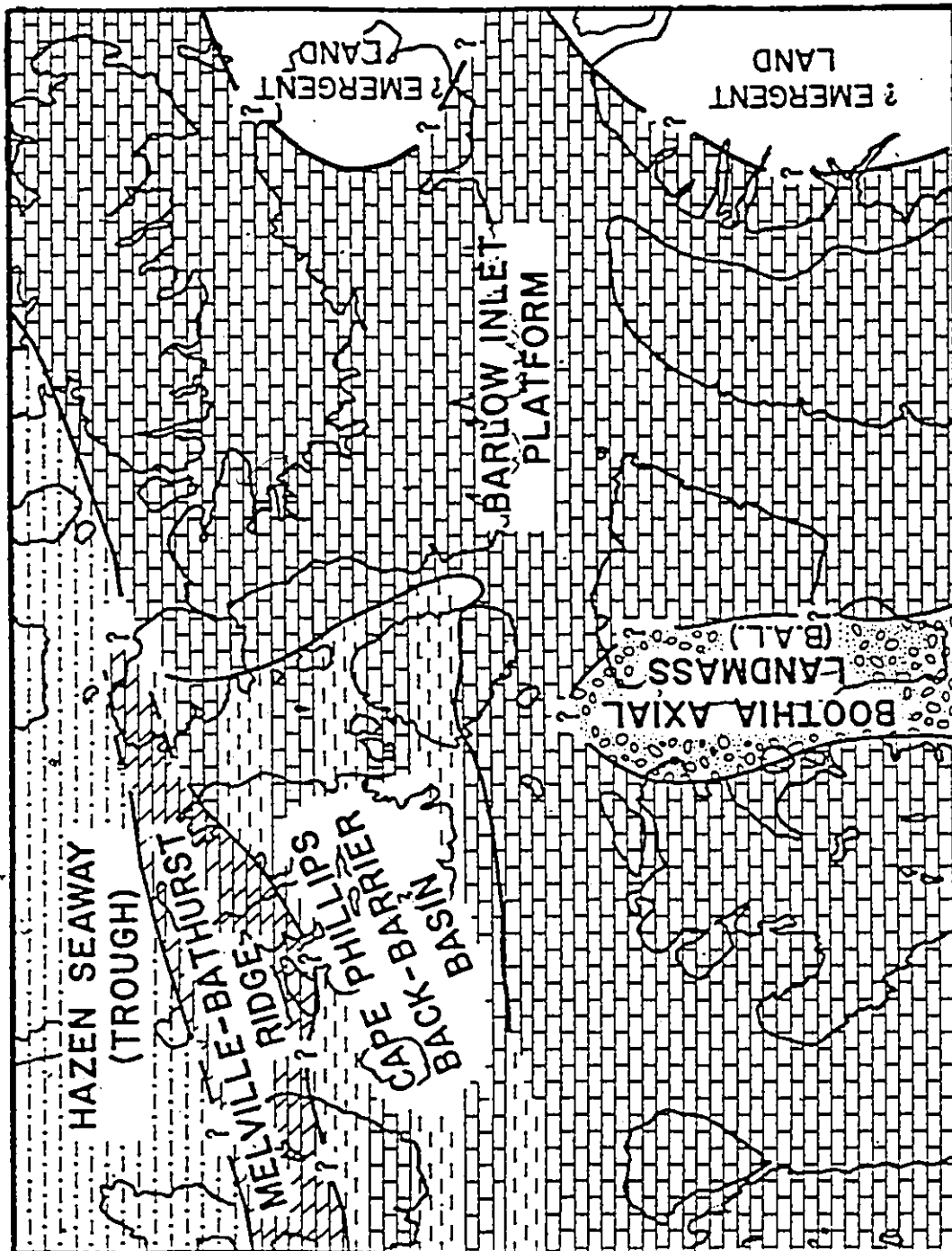


FIG. 4-5 Siluro-Devonian paleogeographic elements of the central arctic archipelago.

Mass-wasting of continental uplands produced by this epeirogenic uplift (a pulse of the Cornwallis Disturbance) resulted in deltaic sediments being deposited over wide areas of the evolving ramp. Following the cessation of terrigenous sediment influx, north of present Barrow Strait, there developed a platform with a rimmed shelf (the Barlow Inlet Platform), whose structural form and as a consequence, facies disposition, were strongly influenced by north-south (predominantly?) and east-west trending major normal faults bounding the basement uplift (or horst). Continued pulses of epeirogenic uplift to the south resulted in isostatic readjustment and subsidence of the platform. Episodic increase in subsidence rates coupled with normal carbonate accumulation, produced the characteristic PAC's (punctuated aggradational cycles) within the platform succession. In Gedinian time, the locus of uplift moved northward, and the platform was exposed and marine carbonate deposition ceased.

The Cape Phillips Back-Barrier Basin that bordered the northern side of both the Douro Ramp and the Barlow Inlet Platform was a stratified, in part anoxic basin, separated from a larger marine body of water (the Hazen Seaway) by a major (reefal? structural?) barrier, the Melville-Bathurst Ridge. Regional evidence suggests that the Barlow Inlet Platform was an interior shelf protected from excessive wave stress, with the biocumules of this shelf growing in the aerobic surface waters of an otherwise anoxic water

mass. These tentative conclusions are echoed by paleoenvironmental evidence suggested by the biocumules, and are discussed in the section entitled 'Synthesis', and in Paper 5 (this thesis).

4.2 DESCRIPTION OF BIOGUMULE FACIES

4.2.1 Lithistid Sponge-dominated Mud Mounds

Lithistid sponges are faunal constituents of several biogumule facies, but only occur as the principal 'reef-builders' in mud mounds outcropping at Depot Point (loc. 8), Cape Rescue (locs. 13, 14 and 18), and Goodsir Creek (loc. 6).

4.2.1.1 Size and Shape

The sponge mud mounds examined in the Cape Rescue area range from 2.2 to 18.8 m thick. The present topographic expression of these mounds suggests that their three dimensional shapes are highly variable, ranging from subspherical, to prolate, to oblate. Given the close proximity of four prolate structures at loc. 13 (PL. 4-1b), it is reasonable to assume that they coalesce, at least locally, in the third dimension. This would give rise to highly irregular biogumule morphologies reminiscent of those figured by Narbonne (1981, plate 3) for Ludlovian lithistid sponge mounds of the Douro Formation on Somerset Island. The much larger complex of interconnected mud mounds at loc. 8 (Depot Point), further substantiates the view that these structures developed intricate morphologies (PL. 4-1a) through lateral expansion and subsequent fusion of individual buildups. The stratigraphic relief on these mounds attains a maximum of 90 m. At Goodsir Creek one

small lithistid mud mound has been examined. It has an indeterminate shape and is 1.8 m thick.

4.2.1.2 Composition and Structure

The biocumules are composed overwhelmingly of lime mudstone. Discounting algal and lithistid sponge material, sandsize or larger skeletal allochems constitute, on average, less than 5% of the bulk rock. Only rare, local, and apparently randomly distributed pockets within the biocumules exhibit boundstone and floatstone textures comprising non-sponge/algal biota.

The lime mudstone lithologies of these biocumules may be divided into 5 subfacies.

1. lime mudstone subfacies (micrite to microspar with few to no sponge spicules or desma).
2. spicular mudstone subfacies (as above but randomly scattered spicules are common).
3. spicular framework subfacies (spicules are more regularly arranged and confined to clearly defined areas).
4. sponge mudstone subfacies (boundaries of individual sponges recognizable and growth form often discernible).
5. algal boundstone subfacies (lime mudstone with irregular, faint, crinkly laminae assigned to spongiostromate algae).

All five subfacies are intergradational and commonly occur within a single hand sample. The first three are generally mottled, with irregular lighter green-grey and darker grey patches of uncertain origin. Neither mottle color is preferentially associated with spicules or spicular frameworks (PL. 4-2a) and mottle boundaries follow no perceptible fabric. However, the darker mottles reflect a preponderance of micrite, whereas the lighter mottles define areas of microspar concentration. Highly irregular mottle outlines are not suggestive of burrowing, and bear little resemblance to obvious burrow structures in this biocumule which are typified by smooth, sulphide-encrusted burrow margins (PL. 4-2c). Mottle boundaries typically represent gradual change in the intensity of an enveloping halo, suggestive of a reaction interface for aggrading neomorphism (micrite to microspar). The mottling may be an early diagenetic (neomorphic) effect controlled in part by organic decomposition of sponge tissue.

The matrix is not peloid-rich, but in certain thin sections of relatively unaltered spicular framework boundstone, concentrations of tuberoids (Fritz, 1958) (or ?peloids of Gingac and Bourque, 1979) are associated with spicular framework, and probably represent a decay product of the sponges (Flügel and Steiger, 1981).

The sponge mudstone subfacies occurs less commonly than either the spicular mudstone or spicular framework subfacies, attesting to the susceptibility of the lithistid

sponge structure to biodestruction, organic decay and diagenesis (PL. 4-2f). No three dimensional sponge fossils were observed on weathered surfaces of the biocumules or within associated flank beds. In polished sections and peels, massive growth forms appear to predominate. These exhibit irregular outlines reflecting an encrusting habit. The size of individual sponges with this growth habit cannot be determined. Where preservation is good, sponges compose up to 80% of the rock. Frameworks of irregular forms are commonly developed (PL. 4-2d). Conical and cylindrical forms (up to 3 cm in diameter) are less common, and are rarely upright. No detailed taxonomic work has been done on these sponges, although two distinctive sizes of desma suggest the presence of at least two lithistid species. The spicules (originally siliceous) are invariably replaced by small (5-10 microns), equant, calcite spar crystals.

The algal boundstone subfacies consists of crinkly laminated micrite that probably represents cyanophycean spongiostromate algae. Commonly the algae and lithistid sponges are intimately intergrown (PL. 4-2g), and locally the algal boundstone predominates.

Other organisms that contribute to the biocumules form a low diversity assemblage comprising (in approximate order of abundance) crinozoan holdfasts (of the cirriferous type -- PL. 4-2e), pelmatozoan skeletal debris, syringoporid corals, cornulitids, low domical stromatoporoids, irregular massive

favositids, small atrypoid brachiopods, ostracodes, bryozoans, mucophyllid and stylopleurid rugose corals, cerioid rugosans, pentamerid and rhynchonellid brachiopods and possible dasycladacean algae.

No consistent vertical zonation of fossils has been observed. Algae and holdfasts locally occlude lithistid sponges, but such occurrences are apparently random spatially. Stylopleurid-rich boundstone forms a metre-thick capping to one portion of the mud mound complex at loc. 18. Limited observations suggest that non-sponge/algal biota (e.g. massive skeletal metazoans) increase in the upper southwestern-most portions (PL. 4-I) of the Depot Point buildup complex.^{1*} At loc. 18 syringoporid corals appear to predominate over sponges in the higher parts of several mounds, though this may simply reflect a greater degree of alteration at higher levels of the structure.

Both growth and solution cavities occur within the biocumule facies. Growth cavities are typically fenestrae of stromatactis type (Flügel, 1982) filled by clear, equant calcite-spar, and occur commonly in the mud mounds. They are elongated parallel to the preserved growth surface of the biocumule and have smooth bottoms and digitate or irregular roofs (PL. 4-2d). They rarely exceed 2 cm in length, but are generally larger than 5 mm. Stromatactis

^{1*} solitary rugosans (mucophyllids) and massive stromatoporoids comprise 25% of the biocumule, the matrix is substantially a crinoidal wackestone.

roofs are commonly formed of spicular framework, whereas cavity floors are constructed of either sediment fill or algal laminae.

Solution cavities are large, irregular, multigeneration cement-filled voids (PL. 4-2a). These cavities can exceed 10 cm in diameter, and show little fabric specificity. The cement generations include, in their apparent order of precipitation,

1. fine, equant, isopachous dolospar cement.
2. coarse, equant, zoned, Fe-dolospar (white in reflected light):
3. coarse, equant, clear, calcite spar.

Volumetrically the coarse white dolospar predominates. This cement is similar to that described by Beales (1971).

Very large scale (2-3 m wide) synsedimentary cavities may also have existed. At loc. 8, a massive unit of very finely laminated micrite, is enclosed on three sides by lithistid sponge boundstone. This massive unit bears no resemblance to inter-mound lime mudstone beds, and may represent the sediment fill (plus organic encrusters?) of a large interreef cavity.

4.2.1.3 Enclosing Sediments

Lithistid sponge mud mounds became established on at least two types of substrate. More commonly, they succeeded

polymictic carbonate breccias (PL. 4-2i). These breccias are partly mud-supported, but the clast density is high (60%). The breccias are generally poorly sorted, and have no internal structure. One such breccia yielded the following clast types; 'non-reefal' hemipelagic lime mudstone (75%), crinoid stems (10%), micritic boundstone (5%), crinoidal wackestone and packstone clasts (5%), fragments of massive colonial metazoans (mostly favositids) (<2%), and solitary rugosans (<2%). Some clasts exceed 1 m in diameter. Many clast types are angular and clearly behaved as coherent blocks during transport (including the micritic boundstone -- PL. 4-2j), whereas other clasts were so severely plastically deformed that they cannot be distinguished with confidence from matrix. The process responsible for depositing the breccia was capable of eroding underlying, bedded, hemipelagic mudstones, a fact readily supported by the abundance of this clast type in the breccias. In one sample, a partially brecciated bed is encrusted by non-calcareous algae, that in turn serve as a substrate for lithistid sponges and ?Sphaerocodium (PL. 4-2h).

Lithistid sponge biocumules also directly overlie thin-bedded wackestone and mudstone that contain a sparse fauna of low domical favositids, heliolitids, and crinoid debris (PL. 4-3g).

Both flanking strata and suprajacent units are composed of:

1. planar-bedded mudstone with intercalated marl seams containing a sparse to negligible fauna.
2. sparsely fossiliferous rubbly argillaceous carbonate (see Jones et al., 1979).
3. thin to medium bedded crinoidal wackestone with a normal marine fauna (brachiopods, bryozoans, trilobites, tabulates).
4. crinoidal packstone to grainstone (often mudstone-lithoclast rich, and with gutter casts on bedding soles (PL. 4-3d).
5. carbonate breccias similar to those underlying the mud mounds, but with oligomictic varieties as well.

Bioclast-rich flanking strata commonly wedge out or thicken towards the margins of the mud mounds (PL. 4-3c), which are in all cases sharply defined (see PL. 4-3a). For the most part a vertically continuous 'bioclastic halo' is not developed around the mud mounds. Most flank beds immediately adjacent to the biocumule are mudstone. Interspersed with the mudstone beds are crinoidal packstone lenses (PL. 4-3b), that are associated with in-situ holdfast-rich pockets within the buildup itself. The density of ossicles is related more to proximity of source, than to hydraulic sorting. Grainstones are rare; most lithologies so identified in the field, are in fact

diagenetically altered packstone. Packstone beds commonly pinch out rapidly away from the biocumule facies (PL. 4-3b); those crinoidal beds that are more widespread (10's of metres) and parallel-sided, are also generally lithoclastic (rudite-sized, subangular lime mudstone clasts), tighter-packed, and finer-grained (greater degree of pluricolumnal disarticulation). They also exhibit gutter casts on their soles, and are interpreted as locally storm-reworked-flank beds. There is no evidence to suggest a (distal) upslope derivation for this lithofacies, although this possibility exists.

Rare evidence of mass transport and slope failure (PL. 4-3e,f) include debris flow deposits (with rafted blocks of plastically-deformed strata, polymictic clasts and randomly oriented colonial metazoans in a muddy crinoidal matrix) and soft-sediment folds.

4.2.1.4 Environmental Interpretation

The sequence underlying the biocumules at loc. 8, consists of unfossiliferous to sparsely fossiliferous thin, even, planar-bedded couplets of lime mudstone and marl. Fossils present represent environmentally tolerant or open marine organisms. These ribbon limestones are in places affected by synsedimentary deformation (slump folding). Intercalated with this predominantly mudstone lithology are lithoclastic crinoidal packstone/grainstones, and carbonate

breccias. This succession is suggestive of moderately-deep foreslope environment, periodically subject to the effects of wave stress (e.g. storms). The mud mounds grew in this moderately-deep foreslope position.

The sponge buildups were rigid structures with well-defined margins. They were capable of standing at least 5 m above the surrounding seafloor, and had depositional slopes up to 35 degrees, perhaps even vertical (PL. 4-3a,g). The original topographic relief and rigidity of these organic structures, are indicated by the presence of angular 'reefal' blocks within the carbonate breccias immediately underlying and adjacent to the mud mounds. The rigidity of the boundstone can be attributed in part to the binding activity of the irregular encrusting lithistid sponges and both spongiostromate and porostromate algae. However, the presence of burrows (PL. 4-2c), and the loss of much of the originally coherent spicular frameworks, suggest that most solidification took place after the death of the organisms. Klappa and James (1980) and Narbonne (1981) have suggested that much of the lime mud in lithistid sponge buildups may have been produced by disintegration of calcified algal tubules (e.g. Girvanella). This process however, would not substantially lithify the mound core. Alternatively, much of the pervasive inter-spicular micrite may be cement, possibly related to the decomposition of sponge tissue as suggested by Gingac and Bourque (1979).

4.2.2 Mucophyllid Mud Mounds

Only three examples of this biocumule facies have been discovered within the Barlow Inlet Formation, all at loc. 18.

4.2.2.1 Size and Shape

The mucophyllid biocumules are roughly equidimensional (subspherical), and between 4 and 5 m in diameter.

4.2.2.2 Composition and Structure

The principal faunal element is the saucer-shaped solitary rugose coral, Mucophyllum sp. nov. (pers. comm., W. Parkins, 1983). The corals rarely exceed 8 cm in diameter. Densities as high as 100 individuals per square metre are present (PL. 4-3i), but it is unlikely that the corals formed true frameworks. Skeletal/matrix ratios approach 15%. The corals are evenly distributed, relatively whole, and uniformly oriented within the matrix. The corals are enclosed by a finely crystalline dolomitic matrix with rare scattered poorly-preserved crinoidal ossicles.

One of the biocumules exhibits some vertical zonation in faunal content. A lower mucophyllid biofacies is overlain by 1 to 2 m of crinoidal dolomudstone (locally wackestone) containing tabular stromatoporoids.

4.2.2.3 Enclosing Sediments

No flank or cap beds are exposed. The three mucophyllid mud mounds rest abruptly on a larger dolomitized mud mound (PL. 4-3h) composed principally of crinozoan holdfasts, spicular micrite and syringoporid corals. The mucophyllid mounds must have had original topographic relief (and rigidity), as the growth directions of most of the observed corals are parallel or subparallel to the paleohorizontal, a condition most simply explained by in-situ growth of the corals on the sides of a 'patch-reef' raised above the sediment/water interface.

4.2.2.4 Environmental Interpretation

Beneath and lateral to the sponge-holdfast-syringoporid and the mucophyllid mud mounds, are slumped periplatform ribbon lime mudstones, carbonate debris flows, and allochthonous blocks of biolithite, all suggestive of a foreslope environment. The sponge-holdfast-syringoporid mud mound rests conformably on crinoidal wackestone with a normal marine biota. Geopetals suggest that the larger mud mound is in place. The mucophyllid mud mounds are satellite patch reefs that are in part enclosed by the parent structure (see PL. 4-3h), suggesting that they did not colonize a 'dead reef'. The fact that both biocumule facies are dolomitized whereas exotic foreslope reef blocks are not, supports the interpretation of in-situ growth. The

absence of crinoid shoal facies, or extensive massive stromatoporoid-tabulate biocumules, suggests that the biocumules grew in calmer waters downslope from a shallow water shelf-break. Lithistid sponge mud mounds in the Barlow Inlet Formation, also interpreted as foreslope structures, similarly have a final biofacies involving cup-shaped rugose corals.

A puzzling and as yet unexplained attribute of the mucophyllid mud mounds, is that the coral calices uniformly face toward the mound centre (see PL. 4-3i).

4.2.3 Diverse Community Mud Mounds

The least conspicuous biocumules in terms of size are, paradoxically, the most impressive Barlow Inlet biogenic structures with respect to faunal/floral diversity and textural preservation. Altogether eleven diverse community mud mounds have been studied at locs. 2, 3, 5 and 6. Biocumules of this category are restricted to a stratigraphic interval 1.6 to 30.5 m above the base of the formation. In addition several biocumules that may represent severely dolomitized variants of the diverse community mud mounds are exposed up to 60 m above the base of the formation at locs. 5 and 6.

4.2.3.1 Size and Shape

The mud mounds are irregular ellipses in cross-section. Diminutive but discrete mounds as small as 0.4 m wide by 0.2 m thick have been recognized. The coalescence of several such small mounds or saccoliths (Smith, 1981) is in part responsible for compound structures that reach 5.4 m in thickness, and exceed 10 m in lateral extent. However, most of the mounds examined in this study are less than 8 m² in cross-sectional area (PL. 4-4a).

4.2.3.2 Composition and Structure

The diverse community mud mounds are characterized by heterogeneity in both composition and structure. Significant variation occurs within mounds, and between mounds found in similar environmental and stratigraphic settings. Systematic biotic zonation is apparently absent except in the largest of the structures, where a poorly developed succession is recognized.

The uneven distribution of organisms, and the generally poor surface expression of the biota in field exposures, poses serious obstacles to qualitative analysis of the mound constituents (see Scoffin, 1971). Semi-quantitative data based on peels (230 standard thin section equivalents) and polished sections suggest that the mud mounds had a predominantly loose to dense structure, but localized pockets of framestone are common (generally associated with halysitid and alveolitid corals). The proportion of lime

mudstone matrix (< 30 microns) varies from less than 5% to 90%, but is generally between 45 and 60%. The matrix consists of microspar with up to 5% euhedral dolomite (10-20 microns) and is locally enriched in lithistid sponge desmae. Within the boundstones, sand-size terrigenous detritus is very rare. Clay-silt insolubles tend to be associated with irregular patches of idiopathic dolospar (20-60 microns), that contain microstylolites and corroded bioclasts, suggesting that their origin can be related to late diagenetic intrastratal solution of an originally homogeneous micrite/microspar matrix. Irregular lenses of crinoidal rudstone are present but uncommon within the mud mounds; for the most part this subfacies is restricted to enclosing sediments.

Common to all the studied mounds, are trepostomate bryozoans (both dendroid and encrusting, PL. 4-4g), lithistid sponges (encrusting and cylindrical, PL. 4-4g), and pelmatozoan debris. Additionally, massive to stoutly cactoid solenoporacean algae (Solenopora -- PL. 4-4i) and rhizoidal holdfasts are present in mounds at locs. 6, 5 and 3, and although they have not been identified in limited samples from loc. 2, it is reasonable to assume they are present, given the abundance of algal balls and crinoid debris in flanking strata.

Corals, particularly tabulates, are important sediment binders and bafflers in mounds at loc. 6, where Catenipora,

Alveolites, syringoporids, proporids, favositids (Favosites, Coenites) are present and are locally the predominant mound fauna (see PL. 4-4b,c,d,e,f). Halysitids, favositids and auloporids are also present at loc. 3, whereas only the latter two coral groups are represented at loc. 2 (see Table 4-6).

Porostromate algae such as Girvanella and Sphaerocodium are ubiquitously present as accessory frame encrustors (PL. 4-5b), and spongiostromate algae occur as oncoids at loc. 3. Possible dasyclads occur as rare bioclasts at loc. 2 and 6.

Stereoxyloides (fasciculate rugose coral -- PL. 4-4f) forms local frameworks in the mounds but in general rugosans are not major faunal constituents.

Stromatoporoids (largely low domical, but including lamellar and dendroid growth forms) occur as important sediment encrusters, bafflers, and inertial stabilizers in the upper portion of the mound at loc. 3 (PL. 4-4f). No stromatoporoids are recorded at loc. 6, and only one encrusting labechid stromatoporoid has been identified from the mound at loc. 2.

Massive skeletal metazoans are invariably small, generally less than 0.15 m in maximum dimension. The biota present and their sedimentological roles are summarized in Table 4-5. Table 4-6 categorizes the similarities and differences between mounds at locs. 6, 3 and 2.

Table 4-5

Observed and interpreted sedimentological function of diverse
community mound-dwelling organisms
See Table 4-4 for translation table.
(broadly applicable as well to other biocumules in the
Barlow Inlet Formation)

GROUP	FUNCTION

ALGAE	
dasyclads	? -> Sc
solenoporids	E -> Sc ; E -> Fm
Girvanella	E -> Cs ; E -> Sc
Sphaerocodium	Es
spongiostromate algae	Es ; Es
SPONGES (all lithistids)	E/Bf -> Sc ; E/Bf -> Cs
TABULATE CORALS	
alveolitids	Es -> Fm
cateniporids	Bf -> Fm
Coenites (dendroid)	Bf -> Sc
favositids (massive)	E/Is
poropids	E/Is
syringoporids	Bf/?E -> Fm
RUGOSE CORALS	
solitary forms	E -> Sc
Mucophyllum sp.	E
fasciculate forms	Bf? -> Sc
STROMATAPOROIDS	
massive	Is/E -> Fm
adnate (lamellar)	E -> Fm
robust dendroid	Bf? -> Sc
BRYOZOANS	
adnate	E
delicate dendroid	Bf -> Sc
ECHINODERMS	
rhizoidal holdfasts	Es/Bf -> Cs?
pelmae and calyces	Sc
BRACHIOPODS	Sc
MOLLUSKS	Sc
ARTHROPODS	Sc

Comparison of diverse community mud mounds from different localities.
(1 mound from loc.2, 2 from loc.3, 2 from loc.5, 6 from loc.6)

T R A I T	L O C A L I T Y		
	2	3	5 and 6
Minimum biotic diversity	13	21	25
Minimum coral diversity	2	4	6
Lithistid sponges	YES	YES	YES
Dendroid tabulates and bryozoans	YES	YES	YES
Solenoporids	N/O, but in flank beds	YES	YES
Rhizoidal holdfasts	N/O, but cri- noid debris abundant in flank beds	YES	YES
Alveolitid corals	---	---	YES
Halysitid corals	---	YES	YES
Stromatoporoids	YES	YES	---
Suprajacent beds	Crinoidal packstone	Marlstone	Marlstone
Subjacent beds	N/O	Commonly crinoidal rudstone	Rarely lime mudstone/ wackestone
Structure	Dense- frame	Sparse- frame	Dense- frame
Matrix	Lime mudstone	Lime mudstone	Lime mudstone
Degree of Dolomitization	Limited	None	None

Note: N/O - Not Observed

Most primary mound cavities are represented by intra-allochem voids. Three lines of evidence suggest the former presence of growth cavities (now largely occluded by sediment). Firstly, accessory frame encrusters (e.g. solenoporids and Sphaerocodium) occur on the bottom surfaces of massive stromatoporoids (PL. 4-5d); their outward growth could not have been inhibited by sediment. Secondly, Mn-oxide crusts exist on all sides of sediment-filled patches that do not have burrow outlines (PL. 4-5h) (direct contact with sea-water may be a prerequisite to the formation of such crusts - see Schlager et al., 1976, p. 117). Thirdly, sediment-floored cavities exist, whose upper surfaces do not truncate allochems (PL. 4-5g) or show evidence of boring activity. Nonetheless, primary cavities do not form a substantial proportion of the mound lithofacies.

Diagenetic cavities (PL. 4-5e) exhibit three cement generations: an inclusion-rich, drusy calcite spar (microspar to pseudospar) that is in direct contact with the surrounding micrite or microspar of the cavity walls, a younger dolospar cement that shows progressive iron enrichment and a final, clear equant coarse calcite spar. The origin of such cavities is not known, but probably involves a complex succession of solution and precipitation. Tension cracks or fissures infilled by coarse blocky to slightly drusy spar represent the final stage of cement

precipitation within the mounds. Microstylolites crosscut these fissure-fillings, and are therefore the youngest diagenetic phenomena observed.

Zonation of diverse community mud mound biota is recognizable only at loc. 3 (FIGS. 4-6 and 4-7), where 5 crude zones (biofacies) are vertically arranged as follows, 1) a basal mudstone with rhizoidal holdfasts, solenoporacean and large crinoid pluricolumnals, 2) rhizoidal holdfasts and pluricolumnals, 3) hemispherical stromatoporoids, rhizoidal holdfasts and bryozoans, 4) hemispherical stromatoporoids, and 5) lamellar stromatoporoids.

4.2.3.3 Enclosing Sediments

The diverse community mud mounds generally rest on crinoid floatstone to rudstone lenses (0.3 to 1.0 m thick, in excess of 5 m wide), suggesting that the biocumules were preferentially established on gravelly rather than muddy substrates.¹ This observation is in accord with the findings of many Silurian reef workers (Scoffin, 1971; Abbott, 1976; Narbonne and Dixon, in press) and reflects a common trend in Paleozoic reef mounds (James, 1977, 1983). Apart from comminuted pelmatozoan skeletons, solenoporacean 'algal balls', pentamerid brachiopods (Nanukidium), and solitary rugosans (including Mucophyllum)

¹ Nevertheless, biocumules evidently flourished on top of muddy sediments in some instances (PL. 4-5a), perhaps by lateral accretion from original growth centres that were established on rudaceous bioclastic deposits.

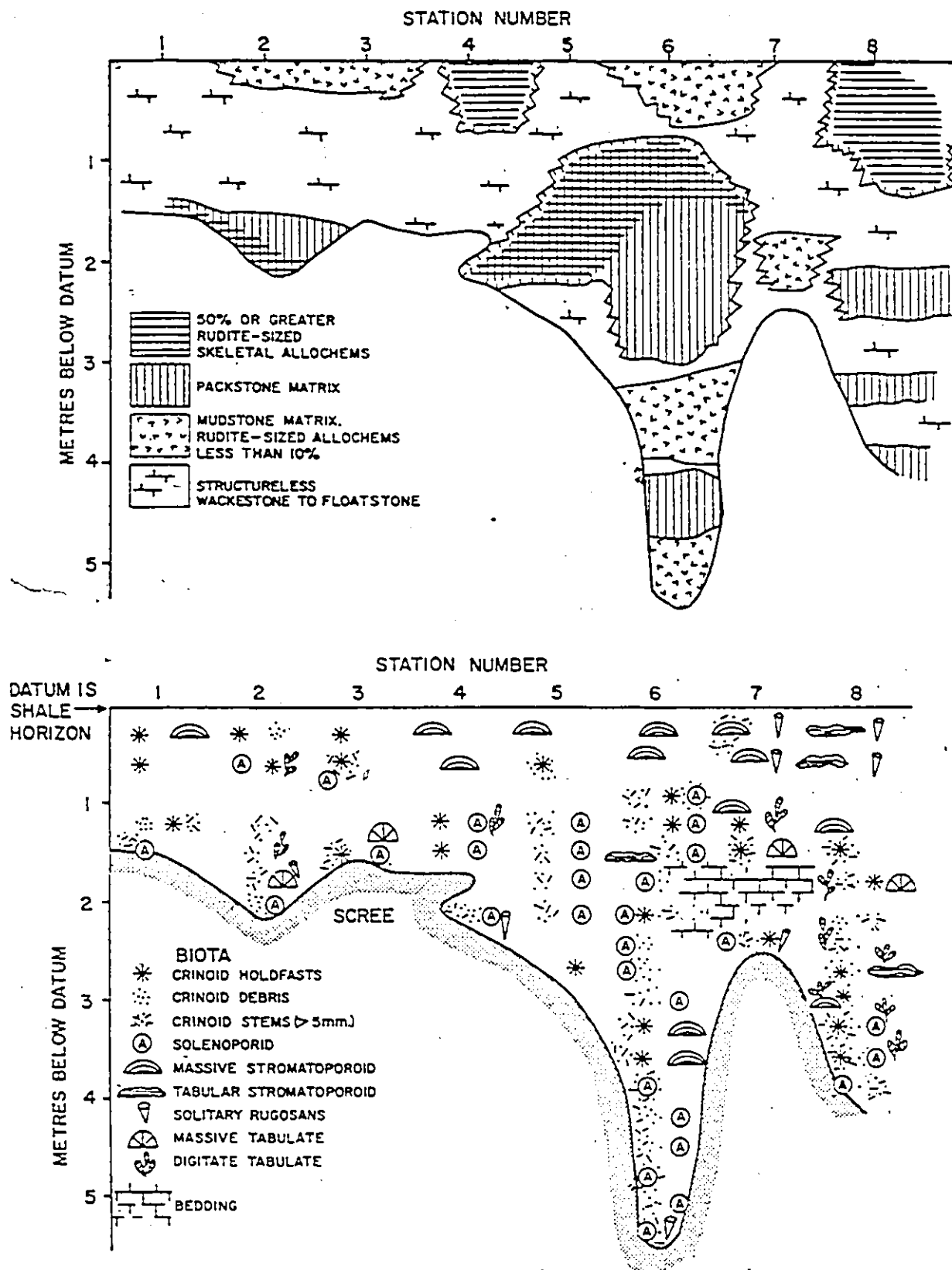


FIG. 4-6 Distribution of bio- and lithofacies in large diverse community mud mound at loc. 3.

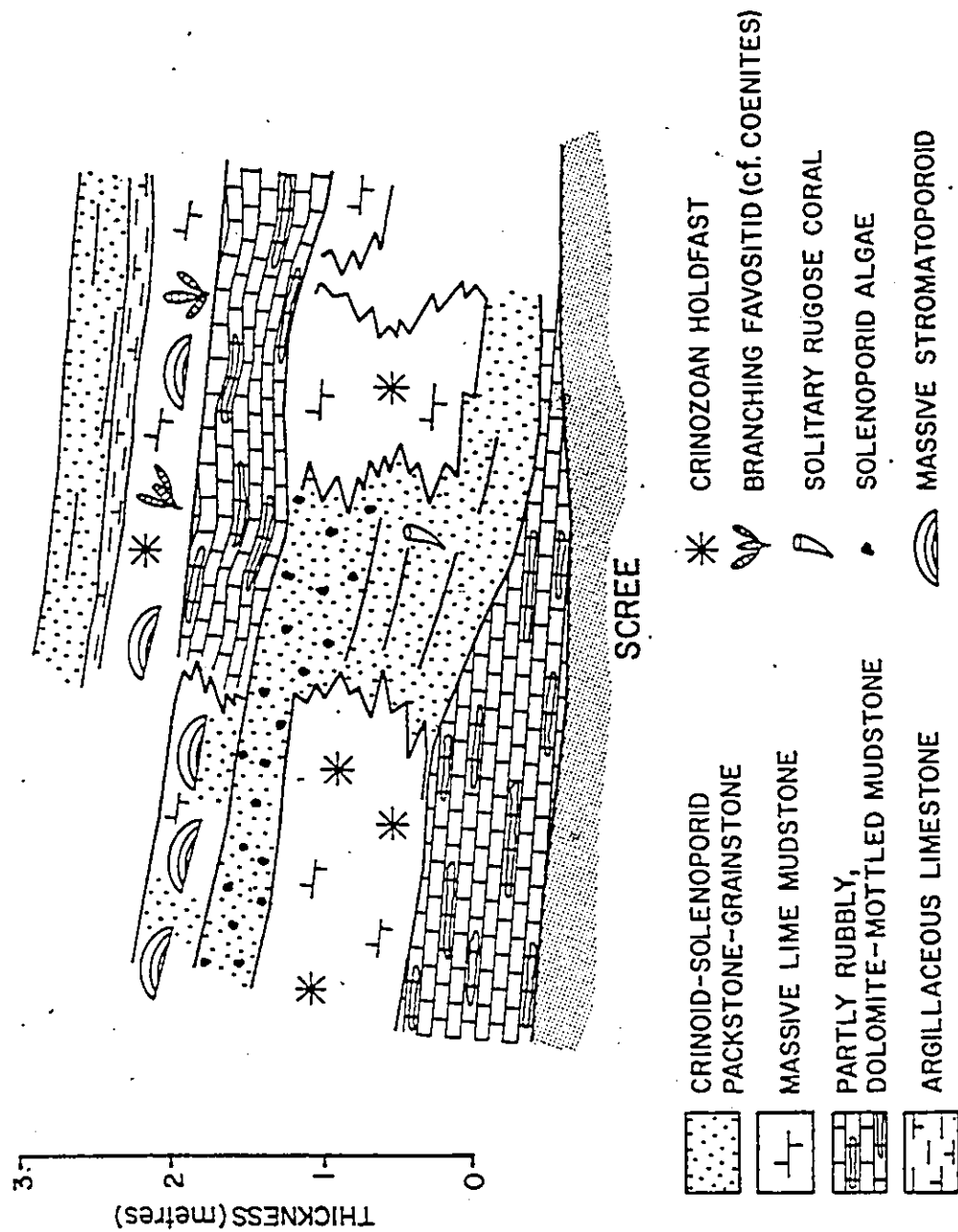


FIG. 4-7 Margin of large diverse community mud mound at loc. 3.

are common fossils in both basal and flank beds.. Dendroid bryozoans and favositids, Sphaerocodium, Girvanella (PL. 4-5i), along with fragments of massive stromatoporoids are other bioclasts in this lithofacies. Bioclasts commonly display micrite envelopes. One small mound lies upon an inversely-graded crinoidal limestone sequence (PL. 4-5c), with mound lithofacies directly overlying a coarse selvage of crinoid pluricolumnals.

For the most part the enclosing sediments are medium to thick-bedded crinoidal packstone, grainstone, and floatstone (texturally and compositionally similar to the subjacent beds), interbedded with argillaceous lime mudstone/wackestone. The crinoidal sediments are likely the result of storm-winnowing of pelmatozoan rubble largely derived from the mound flanks, whereas the mudstone/wackestone represents relatively continuous background sedimentation in the depositional environment. Locally, solenoporacean 'algal balls' form as significant a volumetric proportion of bioclasts as does pelmatozoan debris (PL. 4-4j and FIG. 4-7). No large blocks of mound material, or indeed lithoclasts of associated bedded facies are incorporated in the debris halos.

The uppermost mounds at locs. 5 and 6, and the top surface of the large mound at loc. 3, are covered by argillaceous, conchoidally-fracturing lime mudstone with a sparse fauna consisting of microcrinoids, eurypterids (cf.

Pterygotus pterygotus), strophomenid, orthid and rhynchonellid brachiopods, and bivalves (Pterinea). At loc. 2, the mounds are capped by ?cross-bedded crinoidal grainstone. At locs. 5 and 6, biocumule growth terminated at a stratigraphic level coincident with the occurrence of the argillaceous lime mudstone, although it is apparent that sediment influx was not the cause of the demise of the biocumules. Proof of this conjecture lies with the fact that a large proportion of the outer surface of these upper mounds is coated by a thin skin of Fe/Mn-oxide (PL. 4-5j) that truncates fossils within the biocumule, suggesting that prior to burial, the mounds suffered a period of subdued sedimentation concomitant with seafloor solution. Synoptic relief of one of the structures (PL. 4-5j) can be demonstrated to be 0.8 to 0.9 m, based on the following rationale:

1. ~~no~~ organisms colonized the Mn-oxide crusts, therefore development of this crust ended biogenic growth of the structure.
2. assuming sediment cover inhibited oxide crust formation, all encrusted portions of the mound must have been in direct contact with seawater (and likely above the sediment/water interface); oxide crusts occur at least 0.8 m below the top of the structure.
3. Rhizoidal holdfasts form an ubiquitous part of the mounds, and it is therefore reasonable to assume that much of the crinoidal material in adjacent beds was

derived locally. No crinoidal material is found in flank beds less than 0.8 m below the top of the structure.

Ultimately the stratigraphic interval containing the diverse mud mound biocumules is superceded by crinoidal packstone/grainstone facies.

4.2.3.4 Environmental Interpretation

The diverse community mud mounds occur within a coarsening-upward sequence, interpreted to represent shoaling on the proximal foreslope of a platform margin (see Paper 5).²⁰ The abundance of algal clasts and algal-bored (micritized) skeletal grains (PL. 4-5i) in sediments immediately underlying the biocumules suggests that the mounds grew within the euphotic zone (effectively 150 m), a view further substantiated by the abundant in-situ flora of the biocumules themselves.²¹ The ubiquitous presence of sponges and the rarity of massive stromatoporoids and corals (with the exception of the upper part of the zoned biocumule at loc. 3), suggests a more turbid environment (cf.

²⁰ The stratigraphic interval containing the diverse community mud mounds commences with hemipelagic foreslope lime mudstone rhythmites and terminates below either dolomitized stromatoporoid biostromal units or peritidal sediments.

²¹ Generalizations on the depth significance of higher taxonomic groups of algae must be subject to some skepticism (see Riding, 1973), and it is the abundance and diversity of algae that are considered here suggestive of the euphotic zone.

Narbonne, 1981). The argument is supported by the principal mound-dwelling organisms, all of which were potentially efficient at either shedding (dendroid and fasciculate bryozoans and solenoporids), and baffling (sponges, halysitids), or binding (Alveolites, rhizoidal holdfasts) or otherwise stabilizing loose fine-grained sediment. Atrypoides foxi, considered to have inhabited deeper water substrates (Jones and Narbonne, in press) occurs immediately above and below the mound interval at loc. 6. The presence of solitary and fasciculate rugosans suggests the subturbulent zone of Pedder (in Jackson, Pedder and Lenz, 1978) and Tsien (1974). Syringoporids predominate in faunas of pioneer communities (deeper, calm-water environments?) of other late Silurian reefs (Lowenstam, 1950; Bourque, 1979). The occurrence of spongiostromate oncoids is not taken here to indicate high energy conditions (see Peryt, 1977), but their presence in association with winnowed crinoid and solenopod bioclastic halos, does suggest at least periodic turbulent conditions on the seafloor. The lime mudstone and wackestone equivalent to the mound facies, situated beyond the outer limits of the debris halos, are representative of the predominantly calm-water conditions extant during mound growth. Collectively these data suggest that the diverse community mud mounds developed in a turbid, protected or deeper water environment subject to periodic turbulence.

4.2.4 Fasciculate Rugose Coral Biostromes

The fasciculate rugose coral biostrome category actually embraces two biocumule subfacies that are interpreted to occupy quite different paleoenvironments. One subfacies is characterized by dense phaceloid growth of the rugose coral Stereoxyloides, represented at locs. 2,3 and 11. The second subfacies consists of less dense concentrations of phaceloid to dendroid colonies of stylopleurid rugosans, with examples exposed at locs. 4,6 and 7.

4.2.4.1 Size and Shape

The stylopleurid biostromes are generally 1 to 5 m thick, although associated stratiform units where stylopleurids are both common and the predominant faunal element, attain thicknesses up to 14.6 m. The biostromes are at least 30 m wide and appear to be lenses of very low relief. The Stereoxyloides biostromes (PL. 4-6e) range in thickness from .3 to 1.6 m and extend along strike at least 30 m.²² They are thought to be sheet-like.

4.2.4.2 Composition and Structure

The stylopleurid biostromes have generally low skeletal/matrix ratios (<20%), and consist of a loose framework of sister corallites (buds) (PL. 4-6a,b)

²² accessibility of outcrop (primarily), availability of exposure, and time constraints prevented thorough lateral tracing of these structures.

reinforced by connecting processes and stolons. Maximum widths of mature corallites are generally about 3 cm. The matrix is carbonate mudstone (commonly dolomitized). In peels of specimens with relatively undolomitized interstitial lime mudstone, the presence of a significant quantity of spicular micrite suggests intimate intergrowth of stylopleurids and sponges (?lithistids). Other characteristically associated fauna include Nanukidium sp. (a pentamerid brachiopod), large (up to .5 m across) hemispherical favositids, cylindrical favositids, atrypid and strophomenid brachiopods, bryozoans (cf. Fenestrellina), hemispherical to tabular stromatoporoids, indet. solitary rugosans, and crinoid debris. Also associated is a diverse assemblage of gastropods including the ornate specialized genus Spinicharybdis (Rohr and Packard, 1981). In general the associated fauna (apart from the sponges) occurs only where the density of stylopleurids is relatively low. The upper and lower boundaries of the stylopleurid biostromes appear quite diffuse.

In contrast the Stereoxyloides biostromes usually have high skeletal/matrix ratios (up to 65%), are rarely extensively dolomitized, and have a non-spicular inter-corallite micrite matrix (PL. 4-6f). Neighbouring corallites are commonly fused, and individual mature corallites do not exceed 2 cm in width. Commonly the densely-packed rugosans preclude an associated fauna, but lamellar and dendroid stromatoporoids do co-occur.

Where concentrations of Stereoxyloides are lower, the above-mentioned fauna, as well as oncoids, infaunal smooth bivalves, hemispherical stromatoporoids and favositids, and ostracodes are associated. The upper and lower contacts of the Stereoxyloides biostromes are generally sharply defined.

4.2.4.3 Enclosing Sediment

Stylopleurid-rich units are typically preceded by lime mudstone and wackestone with a normal marine fauna. At loc. 8 densely concentrated stylopleurids occur in topmost facies of a lithistid sponge mud mound (PL. 4-6d). In general, their concentration increases gradually over several metres. Commonly conformably overlying beds to the biostromes are crinoidal packstone to grainstone.

The Stereoxyloides biostromes abruptly overlie a variety of lithologies, including poorly fossiliferous mottled dolomitic limestone, brachiopod-rich packstone, and crinoidal grainstone. Suprajacent units are also diverse, including cryptalgal laminite and crinoidal packstone.

4.2.4.4 Environmental Interpretation

The associated diverse, stenohaline fauna, and in particular the presence of sponges, strongly suggests an open shelf environment for the stylopleurid biostromes. The open framework produced by colonies of these rugosans, and the fine sediment size of the interstitial matrix, both

indicate calm water conditions. The relatively deep calices of the stylopleurids are suggestive of the subturbulent zone (Pedder in Jackson, Lenz and Pedder, 1978). An overturned block of biostromal material 2 X 1 m (PL. 4-6c) associated with an intraformational truncation surface at loc. 7. Suggests that the block was derived from upslope, and that the biostrome achieved some degree of cohesion (submarine cementation?), at an early stage. At locs. 7 and 6 in-situ stylopleurid biostromes are superceded stratigraphically (within 5 m) by stromatoporoid and crinoid-rich biocumules, considered to represent shallow shelf marginal facies. Collectively these data suggest that the stylopleurid biostromes developed in subturbulent proximal foreslope environments.

The Stereoxyloides biostromes at loc. 2 occur at the base of shallowing-upward hemicycles of the tidal flat, platform interior type (see Paper 3). Their intimate association with a generally restricted fauna, and cryptalgal laminite facies is suggestive of protected lagoonal-tidal flat conditions. Conversely, these biostromes, at locs. 3 and 6, occur a few metres above a complex of holdfast-tabulate-solenoporacean mud mounds and crinoidal grainstone beds. Nanshanophyllum (the solitary form equivalent of Stereoxyloides) is a common accessory in shelf-marginal buildups. On this evidence, it is suggested that dense biostromal concentrations of Stereoxyloides

accumulated in the immediate back-barrier portion of the lagoon. Smaller clusters of Stereoxyloides and Nanshanophyllum are distributed over a wide range of lagoonal and shelf-margin facies, implying a considerable degree of environmental tolerance.

4.2.5 Stromatoporoid-Crinoid Rubble Skeletal Buildup

Of all the exposed stratigraphic reefs within the Barlow Inlet Formation, none are more "reef-like" in external form than the stromatoporoid - crinoid rubble skeletal buildups (e.g. PL. 4-7a). Four examples at locs. 7, 9, 10 and 17 have been examined.

4.2.5.1 Size and Shape

These buildups range in maximum thickness from 20 m (loc. 17) to approximately 40 m (loc. 10) and are typically mound-like to irregular lenticular (PL. 4-8c). The buildup at loc. 10 has a maximum height/length ratio of 0.2 and cross-sectional area of approximately 3600 m². Apparent depositional dips for the buildup growth surface average 18 degrees, although gradients are locally much higher.

4.2.5.2 Composition and Structure

Primary textures are commonly obscured by intense dolomitization, particularly in the centralmost portions of the buildup (PLS. 4-7b, 4-8a,g). Nevertheless, locally

preservation is sufficient to elucidate in general terms the core structure. The buildup is constructed of massive colonial metazoans set in coarse skeletal matrix. The large massive colonials locally form frameworks, but are more commonly isolate (PL. 4-8f) (dense to frame structure of Riding, 1977). Stromatoporoids predominate, with growth forms tending to massive domical, but ranging to tabular and lamellar (PL. 4-8i). Favositid corals are less common, but are concentrated into local frameworks, with the corals tending to have an irregular growth habit (PL. 4-8j). Cerioid rugose corals are rare. Minor faunal constituents include Coenites, atrypoid and pentamerid brachiopods, orthoconic nautiloids, solitary rugose corals, and delicate dendroid stromatoporoids. The massive stromatoporoids encrust favositids and pentamerids, and one species of stromatoporeid hosts a commensal relationship with solitary cystiphyllid rugosans. Table 4-7 presents the results of in-field tabulations on the composition of part of a buildup core at loc. 17.

The matrix varies considerably in texture, but is monotonously consistent in its allochem content of comminuted, typically granule-sized crinoid debris (PL. 4-8h). Wackestone, packstone, floatstone, grainstone and rudstone textures are all present, but the last is most prevalent throughout the buildups. At loc. 9, a portion of the buildup free of massive skeletons exhibits well-defined graded beds

Table 4-7

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In-field tabulation of reef-core constituents
(Stromatoporoid-Crinoid rubble skeletal buildup, loc. 17)
418 point counts on typical reef-core facies. Thirty-two equally-spaced
traverses over an area, 1 X 2 m.

CONSTITUENT	PERCENT OF TOTAL COUNTS
Fine calcareous dolomite matrix	39
Coarse dolomite	1
Crinoid stems	6
Crinoid debris (<2 mm)	21
Rhizoid holdfast	<1
Irregular massive favositids	9
Lamellar to domical stromatoporoids	20
Other	4

of crinoidal packstone. Grading probably resulted from avalanching (by grain flow) down mound flanks (cf. Halley et al., 1983, fig. 4-10). Grading of bioclasts within buildups has also been attributed to storm activity (Tsien, 1977).

No growth cavities are evident in these buildups. Zonation within the buildup core has been observed only at loc. 9, where the core can be divided into a basal, mudstone-rich matrix with scattered colonials, crinozoan holdfasts, and dendroid tabulates (PL. 4-8e), and an upper, dense inertiastone with packstone matrix.

4.2.5.3 Enclosing beds

These buildups are generally underlain by stratiform units of weakly-bedded crinoidal packstone. These are in turn underlain by wackestone to sparse bafflestone rich in Coenites or fasciculate rugose corals (Stylopleura).

Flank beds are also crinoidal, grading (at loc. 9) from bedded packstone close to the buildup core to crinoidal wackestone some 10 m distant. Buildup contacts are generally gradational over distances less than 2 m, but sharply defined margins are common (PL. 4-7a).

Sparse stromatoporoid boundstone with mudstone matrix and slightly bedded appearance flank the buildup at loc. 17 (see PL. 5-20d of Paper 5).

At loc. 10, the buildup complex is overlain by sparsely fossiliferous ribbon limestone exhibiting numerous slump

folds (PL. 4-7c,d). At localities 7 and 9, fossiliferous crinoidal strata with domical stromatoporoids and favositids, gastropods, solenoporacean fragments, and solitary rugosans succeed the buildups (PL. 4-8b).

4.2.5.4 Environmental Interpretation

These buildups appear to occur at the apogee of large scale shallowing-upward sequences (Paper 5). Crinoidal blanket deposits (that in some instances succeed stylopleurid biostromes), served as substrates for planulae attachment of large metazoan colonials. Necessary prerequisites for successful and pervasive colonization of this substrate must have been reduction in sedimentation, and decreased mobility of surface sediments of the blanket. Siluro-Devonian buildups composed in part of free-lying massive stromatoporoids in a rudaceous matrix have been noted by Vopni and Lerbekmo (1972) and Mesolella et al. (1974), amongst others. Vopni and Lerbekmo (1972) suggested that massive stromatoporoids were capable of stabilizing mobile substrates and should be considered sediment binders (i.e. inertial stabilizers as used herein).

The buildups probably had mild topographic relief, as is indicated by draped suprajacent strata, locally sharp buildup margins, and development of bioclastic halos. Whether or not the buildups were 'wave-resistant' is a moot point. Limited observations suggest that dense rather than

frame structure characterizes the buildups. Blocks of crinoidal matrix, suggestive of early lithification of the buildup core, have not been identified in flank beds, and the potential for early cementation as a mechanism for symsedimentary buildup stabilization remains speculative. The efficiency of isolate tabular to hemispherical stromatoporoids in binding a granular, crinoidal substrate can be debated, but the mound form of the buildup is developed only in association with abundant stromatoporoid coenostea, suggesting, at least circumstantially, a relationship between reef upbuilding and stromatoporoid colonization. It is the writer's belief that organic stabilization (binding by stromatoporoids) coupled with local development of framework, and the mass of the deposit itself were collectively capable of resisting moderate and continuous wave stress. The moderate sorting of crinoid allochems, the paucity of lime mud, the presence of bioclastic flank beds, and the predominance of a massive stromatoporoid fauna, suggest growth of the buildups above wave base.

The source of the crinoidal debris in the matrix is conjectural: Despite a considerable volume of this sediment, holdfast structures have been identified only where mudstone matrices predominate, such as the lower portions of buildups at locs. 17 and 9. In other buildups there is no significant basal mudstone facies, and in any case, accumulation of

crinoidal debris clearly continued long after this initial buildup phase was buried. Much of the crinoid material may have been contributed to the buildup hydrodynamically, being derived from proximal portions of adjacent crinoidal blanket deposits. The question of the origin of crinoid debris, is amplified in the section entitled "Synthesis", below.

4.2.6 Dolomitized Massive Skeletal (?Crinoid) Biocumule

The most diagenetically altered and hence enigmatic biocumules within the Barlow Inlet Formation, are also the largest. The dolomitized massive skeletal (?crinoid) biocumules (DMS biocumules) form a significant proportion of the stratigraphic sections at locs. 4,5,6, and 7. Six examples are considered in this paper.

4.2.6.1 Size and Shape

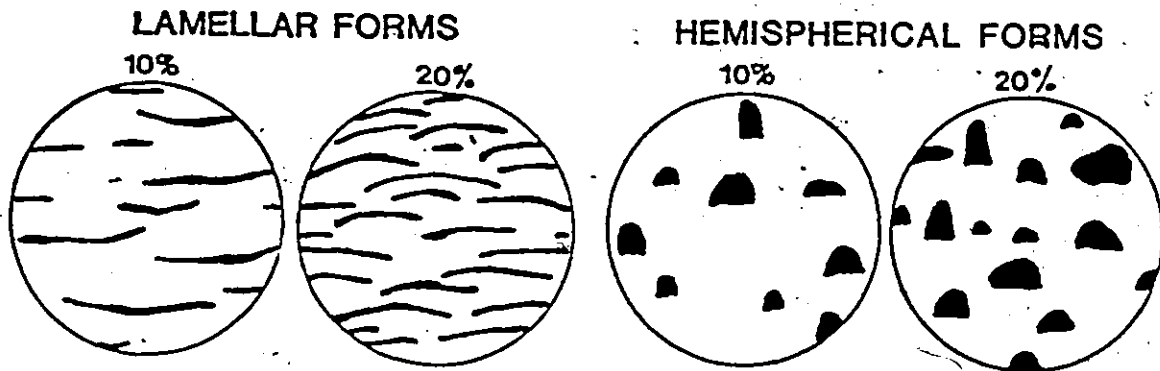
The 6 biocumules have thicknesses of 30 (loc. 6), 40 (loc. 15), 43 (loc. 5), 47 (loc. 4), 60 (loc. 7) and 70 m (loc. 6). Individual biocumules are actually collective packages of strata (units) (PL. 4-9b). These units have largely lost their individual identity through pervasive dolomitization, although their unique character remains recognizable from the size, abundance and arrangement of macropores and vugs, and the presence of casts and molds of massive skeletal metazoans. Unit contacts are generally defined by prominent bedding-parallel fractures that may

exploit original planes separating contrasting lithologies (PL. 4-9e). Unit thicknesses average 4m and have a median value between 2 and 3m (n=75). The shape of these structures cannot be ascertained on individual cliff exposures. On an outcrop scale (e.g. PL. 4-9b), the subparallel arrangement of units does not suggest stratigraphic relief, but suggests rather that these thick dolomitized intervals represent coalesced biostromes. Some of these structures clearly possess stratigraphic relief regionally (FIG. 4-8), with gradients on their margins locally exceeding 5 degrees. If biostromes are distinguished on the criteria of breadth/length ratios of 1/30 or less (see Nelson et al., 1962), then many of the DMS biocumules are not themselves biostromes, although they may be composed of same (e.g. Read, 1981).

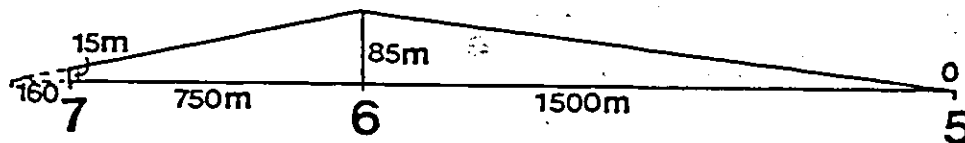
4.2.6.2 Composition and Structure

Due to the pervasive dolomitization, neither fossils nor matrix are readily identifiable. Molds and casts of massive stromatoporoids are most common, with domical (PL. 4-9d), irregular, and tabular growth forms predominating. Units commonly contain all three growth forms randomly juxtaposed. Poorly preserved hemispherical favositids and fasciculate rugosans, along with

REFERENCE CIRCLES FOR RELATIVE AREAS OCCUPIED
BY STROMATOPOROIDS (modified after Flugel, 1982)



GEOMETRY OF MAJOR DOLOMITIZED
BIOCUMULE AT LOCALITIES 6, 7



$$\text{minimum height / breadth ratio}^1 = 85 / 2410 = 0.03526$$

$$\text{limiting value for biostromes}^2 = 1 / 30 = 0.03333$$

— Based on this single N-S transect, this biocumule would not be considered a biostrome based on the given (1/30) height/breadth ratio criteria.

¹ minimum value because calculations assume that the biocumule extends to just north of locality 5

² as proposed by Nelson et al (1962)

FIG. 4-8 Aspects of dolomitized massive (? crinoidal) biocumules.

solitary rugosans and crinoid pluricolumnals are also common. Rarely identifiable remains of pentamerid and rhynchonellid brachiopods, mucophyllid corals, and dendroid favositids and stromatoporoids complete the faunal list. The proportion of rudite-sized massive skeletons to matrix reaches 35% or more, but difficulty in distinguishing the extent of skeletal molds makes this estimate uncertain. Large vugs or molds of massive colonial metazoans are absent from some units (PL. 4-9e) and presumably this reflects original paucity of these organisms, and is not an artifact of recrystallization. Small vugs are frequently infilled by coarse white dolospar such as that described by Beales (1971). The matrix is composed of medium and coarse crystalline dolostone. Calcareous impurities are rare. The dolostones are rarely mottled (PL. 4-9g), and are generally a relatively uniform light buff brown.

Unresolved is the relative contribution of crinoidal debris to the matrix of the massive dolostone. Where dolomitization is less extreme, molds of larger pluricolumnals can be detected (PL. 4-9c). Smaller ossicles, however, are indistinguishable in the coarse dolomite, pervasive throughout much of this biocumule facies (e.g. PL. 4-9f).

4.2.6.3 Enclosing Sediments

Partially dolomitized crinoidal packstone and grainstone in units 1 to 5m thick, underlie most dolomitized DMS biocumules. Similarly, crinoidal wackestone to grainstone invariably overlies the biocumules. In one instance (loc. 6), dolomitization and recrystallization have transgressed the basal contact of the DMS, obscuring the relationship of the biocumule to the subjacent unit (PL. 4-9h,i). Overlying the capping crinoidal calcarenites and also flanking the biocumules, are planar to nodular thin-bedded lime mudstone to wackestone, containing normal marine shelf fauna and characterized by the trace fossil Zoophycus. Portions of this succession show clear evidence of paleoslope processes, including slumped bedding, slide structures, and intraformational truncation surfaces.

4.2.6.4 Setting and Environmental Interpretation

These biocumules are intimately interbedded with, and surrounded by, foreslope deposits. The crinoidal sediments that immediately envelope these biocumules represent proximal foreslope debris aprons and/or shelf edge sand shoals. The intense dolomitization of these biocumules probably also reflects a platform marginal position (i.e. Dorag dolomitization model). The biocumules are not homogeneous and massive, but rather are composed of stacked parallel-sided units of slightly differing lithofacies. The fact that unit thicknesses and percentage of vugs and/or

colonial metazoan fossils of the biocumules and the platform interior cycles are comparable (see Table 4-8), suggests that:

1. a relationship exists between the DMS and platform interior superfacies and
2. the biocumule units may also be arranged rhythmically, although such cyclicity has yet to be discerned.²³

The dolomitized massive skeletal (?crinoid) biocumules are the principal shelf edge facies.

4.2.7 Lamellar Stromatoporoid Biostrome

Well-exposed biostromes in which lamellar stromatoporoids are extremely abundant, and constitute the only major component, are rare within the Barlow Inlet Formation, and restricted to loc. 5 and environs. This biocumule facies is also poorly-developed or exposed at loc. 2 and 7.

4.2.7.1 Size and Shape

At loc. 5 the lamellar stromatoporoid biostrome facies ranges from 2 to 5 m in thickness, can be traced for 100 m north-south and may²⁴ extend continuously in excess of 130 m east-west.

²³ Generally, shelf lagoonal cycles are not reflected directly in the shelf marginal facies (Kendall and Schlager, 1981).

²⁴ Uncertainty arises due to severe dolomitization, and/or poor continuity of exposure

Table 4-8

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Comparison between Restricted Infratidal (RI)
series of the Platform Interior and Dolomitized
skeletal (?crinoid) biocumules. (DMSB)

	LOCALITY					
	2(RI)	3(RI)	4(DMSB)	5(DMSB)	6a(DMSB)	6b(DMSB)
# of field units	41	68	14	16	5	9
total thickness of biocumule or RI series	77.5	138.8	48.5	42.9	29.5	82.5
average % area occupied by VCF ¹	15.5	4.6	5.5	6.6	9.1	12.6
total thickness of field units with VCF	58.3	10.5	31.0	24.4	-	45.2
average % area occupied by VCF *	20.7	9.0	9.0	11.6	-	13.2
# of field units with VCF / % total number of units	26/ 63	25/ 37	8/ 57	9/ 56	5/ 100	12/ 63
average field unit thickness	1.9 $\pm$ 1.8	2.0 $\pm$ 1.9	3.5 $\pm$ 4.2	2.7 $\pm$ 2.5	5.9 $\pm$ 5.5	4.3 $\pm$ 4.7
adjusted ² average field unit thickness	1.9 $\pm$ 1.8	1.9 $\pm$ 1.2	2.5 $\pm$ 2.2	2.2 $\pm$ 1.5	3.7 $\pm$ 2.6	3.0 $\pm$ 2.2

1. VCF stands for vugs and colonial fossils. Vugs are here considered macropores greater than 2cm in diameter.
2. adjustment involves disregarding those field units greater than 10m in thickness. This correction can be rationalized as very few field units measured are, in fact, greater than 5.0m, and those that are, generally could (and should) have been subdivided.

* for field units with VCF.

4.2.7.2 Composition and Structure

Lamellar, rather morphologically-variable stromatoporoids (?Actinostroma) constitute up to 80% of the biostromal units (PL. 4-10a,c). The individual stromatoporoids range in thickness from 4 mm to 10 mm and are arranged in subparallel fashion, mutually-fused to form framework structures. Other fossils, in approximate order of abundance, are fasciculate rugosans (Stereoxyloides), dendroid stromatoporoids (both Stachyodes and Amphipora cf. ramosa), crinoid pluricolumnals and debris, and Coenites. The associated fauna rarely exceeds 5% of total rock volume. Locally there appears to be an intimate relationship between lamellar and dendroid stromatoporoids (PL. 4-10b), and limited evidence suggests the 'rooting' of dendroid Stachyodes in lamellar stromatoporoids. Fischbuch (1968) has also suggested that Stachyodes may be simply a phenotypic variant of other growth forms. Further investigation is required to elucidate the relationship between lamellar and dendroid growth forms, their respective astogeny, preferred orientation and means of support.

The light grey-green weathering matrix is a medium crystalline dolostone (40-200 microns). Although the matrix is invariably dolomitized in studied examples, the stromatoporoids resisted all but minor external impingement dolomitization (Lucia, 1962).

4.2.7.3 Enclosing Beds

The biostromes overlies either relatively unfossiliferous bedded dolomudstone, or bioclastic rudstone, packstone and grainstone chiefly composed of crinoid fragments, solenoporaceans, dendroid stromatoporoids and favositids and broken fragments of lamellar stromatoporoids. No flank beds are recognized. The lamellar stromatoporoid biostromes are overlain by massive, coarsely crystalline ?crinoidal dolostone, representing the basal units of dolomitized massive skeletal (?crinoid) biocumules. At loc. 3, a poorly exposed biostrome is overlain by tidal flat hemicycle sediments (see Paper 3).

Laterally-equivalent beds include undifferentiated, massive medium crystalline dolostones (see PL. 5-13a of Paper 5, this thesis).

4.2.7.4 Environmental Interpretation

The paleoenvironmental significance of relatively thin, sheet-like stromatoporoids (tabular, lamellar, laminar of some authors) is the subject of much controversy (see Stearn, 1975, 1982). Some researchers have concluded that this growth form represents, low-energy deeper-water conditions (e.g. Fischbush, 1968; Klován, 1964; Embry and Klován, 1972), whereas others (e.g. Mayable, 1979; Stearn, 1975) consider the sheet-like morphotype to have developed in shallow turbulent conditions. Tsien (1971, 1974) recorded

thick lamellar stromatoporoids from the subturbulent fore-reef, and thin encrusting laminar forms from restricted back-reef environments of a mid-Devonian reef complex in Belgium. Kershaw (1981) has cautioned that genetic rather than environmental factors may control stromatoporoid growth form, a conclusion reiterated and critically discussed by Stearn (1982).

Several factors collectively suggest that the Barlow Inlet lamellar stromatoporoid biostromes developed in shallow, turbulent environments, at or proximal to the shelf edge (i.e. the shelf edge itself or the shallow proximal foreslope):

1. in places the biostromes overlies bioclastic crinoidal rudstone with considerable quantities of abraded and randomly-oriented fragments of lamellar stromatoporoids. The presence of such a talus apron is suggestive of at least periodic wave stress acting on the biostrome.
2. the suprajacent units are either dolomitized massive skeletal (?crinoid) biocumules, or tidal flat hemicycles. The latter are indisputably representative of back-reef and tidal flat environments, and although the depositional setting of the former cannot be precisely ascertained, it does not represent deep water conditions.

3. the biostrome is in part dolomitized. Dolomitization appears to be largely facies controlled, and generally restricted to shelf adge and proximal back-barrier environments.
4. no sequences of the Barlow Inlet Formation interpreted as deeper water deposits, contain lamellar stromatoporoids.

4.2.8 Massive Stromatoporoid-Dominated Biostromes

Numerically and volumetrically the most significant biocumule facies in the Barlow Inlet Formation, are the massive stromatoporoid-dominated sparse to dense biostromes exposed at locs. 1,2,3,5,11,12, and 14. As these biostromes have been previously discussed in detail in Paper 3, only a brief summary of their characteristics is reiterated here.

4.2.8.1 Size and Shape

Several of the structures considered in this category, are indeed stratiform and parallel-sided bodies that can be traced laterally for considerable distances (up to 250 m). Other structures, however, are not exposed sufficiently to ascertain lateral continuity, but as they are otherwise similar to the well-exposed examples, they are included in this category. Thickness variation between individual biostromes is considerable, ranging from .4 to 12 m. Thickness variation within biostromes is difficult to

ascertain, particularly where rudite or larger colonial metazoans are relatively low in concentration. Lower contacts are commonly abrupt, whereas upper contacts are generally gradational.

4.2.8.2 Composition and Structure

The biostromes have an inertialstone texture, with large colonial metazoans generally comprising less than 15% (and rarely, up to 35%) of the bulk rock, producing a sparse to dense structure (PL. 4-10d,e). The most conspicuous faunal elements are massive stromatoporoids, which principally exhibit hemispherical, subspherical and tabular growth forms. Individual colonies range from 7 to 25 cm in maximum apparent diameter, and are generally between 10 and 15 cm. Massive skeletons are predominantly in growth positions. Other common rudite-sized biotic elements include massive favositids (Favosites), cylindrical favositids, solitary and fasciculate rugose corals (e.g. Nanshanophyllum), cyanophyte oncoids, robust dendroid tabulates (Coenites), and delicate dendroid stromatoporoids. Other fossils are rare and include indet. bivalves, Megalomoidea, gastropods, atrypoid brachiopods, auloporids, Alveolites, solenoporaceans, crinoid ossicles, and ostracodes. The matrix is rubbly-weathering, slightly argillaceous mudstone to wackestone that commonly is substantially dolomitized (PL. 4-10f). Vertical biostromal zonation, if present, is

weakly developed. Oncoids tend to decrease in abundance upward, and the size and abundance of massive stromatoporoids also gradually diminish higher in the biostromal unit. Delicate dendroid stromatoporoids are concentrated in randomly spaced lenses within biostromal successions.

4.2.8.3 Enclosing Sediments

The biostromes commonly abruptly overlies cryptalgal laminite, or massive fenestral lime mudstone. Less commonly these biocumules commence with rubbly wackestone and mudstone (with few to no rudite allochems) that abruptly overlies cross-bedded crinoidal packstone. A sparse to dense structure of massive skeletons is achieved in the biocumules .5 to 2 m above the crinoidal packstones.

Suprajacent beds are generally in gradational contact with the biostrome, and consist of sparsely fossiliferous mottled dolomitic carbonates with a fauna/flora similar to that in the biostromes, but containing fewer individuals in differing proportions.

4.2.8.4 Environmental Interpretation

The biostromes form the basal member of restricted-infratidal hemicycles (see Paper 3), that are interpreted to represent rhythmic platform interior (lagoonal-tidal flat) sedimentation. The fine-grained

matrix, small percentage of overturned massive skeletons, and paucity of stenohaline organisms (brachiopods, crinoids), suggest restricted environments.

The relative abundance of the massive stromatoporoid fauna, the position of the biostrome within shallowing-upward cycles, and the association (if rarely) with cross-bedded crinoidal sands, all suggest an immediate back-barrier position within the lagoon (see also Paper 3).

4.2.9 Dendroid Stromatoporoid Biostromes

A common biocumule facies within the Barlow Inlet Formation is the dendroid stromatoporoid biostrome. Notable occurrences of this facies have been examined at locs. 2,3,5 and 12. Grainstone and packstone of finely comminuted and sorted amphiporid allochems are not here considered biocumulate, but nevertheless probably reflect proximity to source biocoenoses of these delicate branching stromatoporoids.

4.2.9.1 Size and Shape

Typical concentrations of amphiporids are discontinuous small lensoid bodies generally less than 0.2 m high or 1 m long (PL. 4-10g,h). These pockets of amphiporid floatstone to rudstone lie within stratiform units of argillaceous and/or mottled dolomitic lime mudstone from .2 to 3.4 m thick. The thickness range for units containing only amphiporids is .4 to 2.9 m.

4.2.9.2 Composition and Structure

The dendroid stromatoporoids form rudstone (predominantly) to floatstone textures. Individual elements are large (commonly up to 10 mm in length), often aligned and in some instances the taphonomy suggests single colonies shattered in place (PL. 4-10b). The principal sedimentological role of the amphiporids and more robust dendroid stromatoporoids was clearly as a source for rudite bioclasts. They may have acted as fine sediment bafflers, but Jamieson (1971) points out that occurrences of amphiporids in an upright position have not been found, which is anomalous if they indeed functioned as bafflers. Commonly the dendroid stromatoporoids are fused (PL. 4-10j) suggesting that they were mutually-supporting. The sediment-binding potential of amphiporids is not known; no holdfasts have yet been identified. In some instances stick-like stromatoporoids appear to arise from lamellar to irregular growth forms, and the common association of lamellar stromatoporoids with Stachyodes does suggest a close association between these growth forms. Possibly slight environmental changes promoted a radical growth form change within single species of stromatoporoids (lamellar to stick-like) analogous to the phenotypic variability seen in the Holocene scleractinian Monastrea annularis (domal to plate-like). No pronounced zonation of biotic elements is observed in these biocumules.

Amphiporids occur either as the only faunal element (PL. 4-10h,i), or in direct association with Coenites, crinoid pluricolumnals and debris, rugose corals, robust dendroid stromatoporoids (Stachyodes), and lamellar stromatoporoids (PL. 4-10g). These two biofacies are considered to reflect different environmental settings.

4.2.9.3 Enclosing Sediments

Lithologies immediately beneath and above amphiporid packstone lenses are typically unfossiliferous peloidal to micritic lime mudstone, crinoidal packstone, or lamellar stromatoporoid biostrome. Less commonly the amphiporid biocumules are interbedded with stromatoporoid-dominated sparse to dense biostromes. Where amphiporids are associated with robust dendroid stromatoporoids, crinoid debris and Coenites, enclosing sediments are generally mottled dolomitic wackestone or crinoidal packstone or lamellar stromatoporoid boundstone.

4.2.9.4 Environmental Interpretation

Within an overall shoaling succession, amphiporid skeletal biostromes invariably occur stratigraphically above (and in many instances just above) holdfast-tabulate-solenoporacean mud mounds, cross-bedded crinoidal sands and/or zoned biocumules considered to represent proximal foreslope and platform-marginal facies.

Concentrations of amphiporids only are restricted to TF and RI shallowing-upward hemicycle series of the platform interior lagoon (see Paper 3). The impoverished associated biota, the muddy matrix, the stratigraphic juxtaposition (albeit rarely) of cryptalgal laminite and amphiporid wackestone, and the rare development of fenestral fabric within this biocumule facies, all attest to a shallow, protected depositional environment (back-reef lagoon). This interpretation is consistent with numerous citations in a voluminous literature on the distribution of Devonian stromatoporoid growth forms (e.g. Kobluk, 1975; Jamieson, 1971; Burchette, 1981).

Concentrations of amphiporids in association with Coenites, crinoids, Stachyodes, lamellar stromatoporoids and rugose corals, are considered to represent deposits of shallow proximal foreslope environments. The amphiporids may be allochthonous but their relative abundance and large clast size does not suggest an extensive transport history. Furthermore deposits exhibit floatstone textures that are not suggestive of storm sedimentation. Storm-generated currents are the principal means of transporting back-reef sediment to the foreslope environment.

4.2.10 Coenites-capped, Fenestral Mud Mounds

The only recognizeably mound-shaped biocumule found in the type section (loc. 11) of the Barlow Inlet Formation, is a unique and somewhat enigmatic structure 1114 m above the base of the formation.

4.2.10.1 Size and Shape

The studied example has a distinctive low mound shape (PL. 4-11), and constitutes but one of several similar structures locally exposed in three dimensions over an area of 540 square metres. The mound is at least 18 m wide parallel to regional strike, and attains a maximum stratigraphic relief of 3.2 m. The mound flanks dip locally as steeply as 15 degrees (corrected for regional dip) (see PL. 4-12g).

4.2.10.2 Composition and Structure

The bulk of the biocumule is composed of fenestral lime mudstone (PL. 4-12a,b). Fenestrae are irregular in shape and small (cf. Logan, 1974). They show densities of 12.3/square cm and comprise 26% of the total cross-sectional area of examined hand specimens. Fenestrae are filled by coarse, idiotopic or drusy calcite spar. In many fenestrae calcite spar overlies an equigranular medium-grained mosaic of dolomitic ?sediment/cement (see Paper 3, this thesis). Within the micritic matrix, amphiporid stromatoporoids and

complete valves of ostracodes make up to 5 and 2%, respectively, of the bulk rock.

Of particular significance is the occurrence of whole, small, smooth shelled ostracodes within fenestral cavities. The ostracodes are not associated with internal sediment, and their size effectively precludes fortuitous post-mortem introduction into the fenestrae. The geopetal occurrence and lack of remnant micrite around the ostracodes, suggest that the fenestrae did not develop around pre-existing shells, and that rather the ostracodes led a coelobiontic existence. The fenestral mudstone is overlain by unbedded branching tabulate (cf. Coenites) bafflestone, with subordinate small encrusting favositids, amphiporids, and lamellar stromatoporoids (PL. 4-12c,d,e,f). Skeletal elements are densely packed and generally form a bioclast-supported framework whose interstices are filled with micritic sediment (40-60%). The nature of the original contact between fenestral lime mudstone and dendroid tabulate bafflestone is unfortunately obscured by pressure solution, but was apparently very abrupt.

4.2.10.3 Enclosing Sediments

The biocumule lies conformably over lime mudstone with cryptalgal lamination (PL. 4-12h). The cryptalgally-laminated lime mudstone is in turn, interbedded with apparently stratiform, massive fenestral lime mudstone and bedded mottled dolostone with an impoverished fauna.

Flanking strata thin towards, and drape over the mud mound. - The unbedded Coenites cap appears locally to be gradational with bedded, bioclast-rich flank beds. Bioclastic content on extensively-exposed bedding planes, decreases dramatically (by up to 90%) away from the mound core (PL. 4-12i,j,k), over distances of less than 5 m (change from packstone to mudstone). The presence of this bioclastic halo suggests that the mound core profoundly influenced adjacent sedimentation, an inference confirmed by the overwhelming proportion of branching tabulates among flank bed allochems (PL. 4-12i). Other bioclasts include solitary rugosans, solenoporaceans, crinoid ossicles, atrypoid brachiopods, and orthoconic nautiloids. Particular flank-bed strata have a matrix that weathers a distinctive blue, perhaps reflecting elevated Co content in clay minerals. Beds directly overlying the mud mound show decreasing bioclastic content stratigraphically upwards, in a similar fashion to the allochem variation in single beds lateral to the core. The uppermost beds are unfossiliferous thin-bedded mottled dolomitic limestone, similar in all respects to the mottled lithology underlying the mound (PL. 4-12g). The gradational decrease in skeletal fragments in the overlying beds suggests progressive deterioration in conditions suitable to growth of thickets of the dendroid tabulates.

4.2.10.4 Environmental Interpretation

The biocumule occurs within a sequence of carbonate strata that are impoverished in stenohaline organisms (brachiopods, crinoids), and display poorly-developed shallowing-upward hemicycles similar to those described in Paper 3. This suggests that the biocumule grew within the platform interior where circulation was restricted. That the fenestral lime mudstone facies was deposited in very shallow water is suggested by the underlying cryptalgal laminite (intertidal?) and the irregular fenestrae of the core facies (shrinkage -- possibly desiccation phenomena). It is difficult to envisage a process capable of producing a mound-shaped body of lime mudstone subaerially. Furthermore the interpreted in-situ ostracodes within the fenestrae would have required at least periodic submergence (leperditid ostracodes are considered capable of temporary adaptation to subaerial exposure -- Berdan, 1968). The absence of other fauna (e.g. Coenites) suggests a high stress environment (very shallow subtidal, possibly intertidal). The proliferation of dendroid tabulates, encrusting favositids, amphiporid and lamellar stromatoporoids centred about the fenestral mudstone core, suggests that these organisms readily exploited the areally limited substrate presented by the raised mound surface following its complete submergence (see FIG. 4-9).

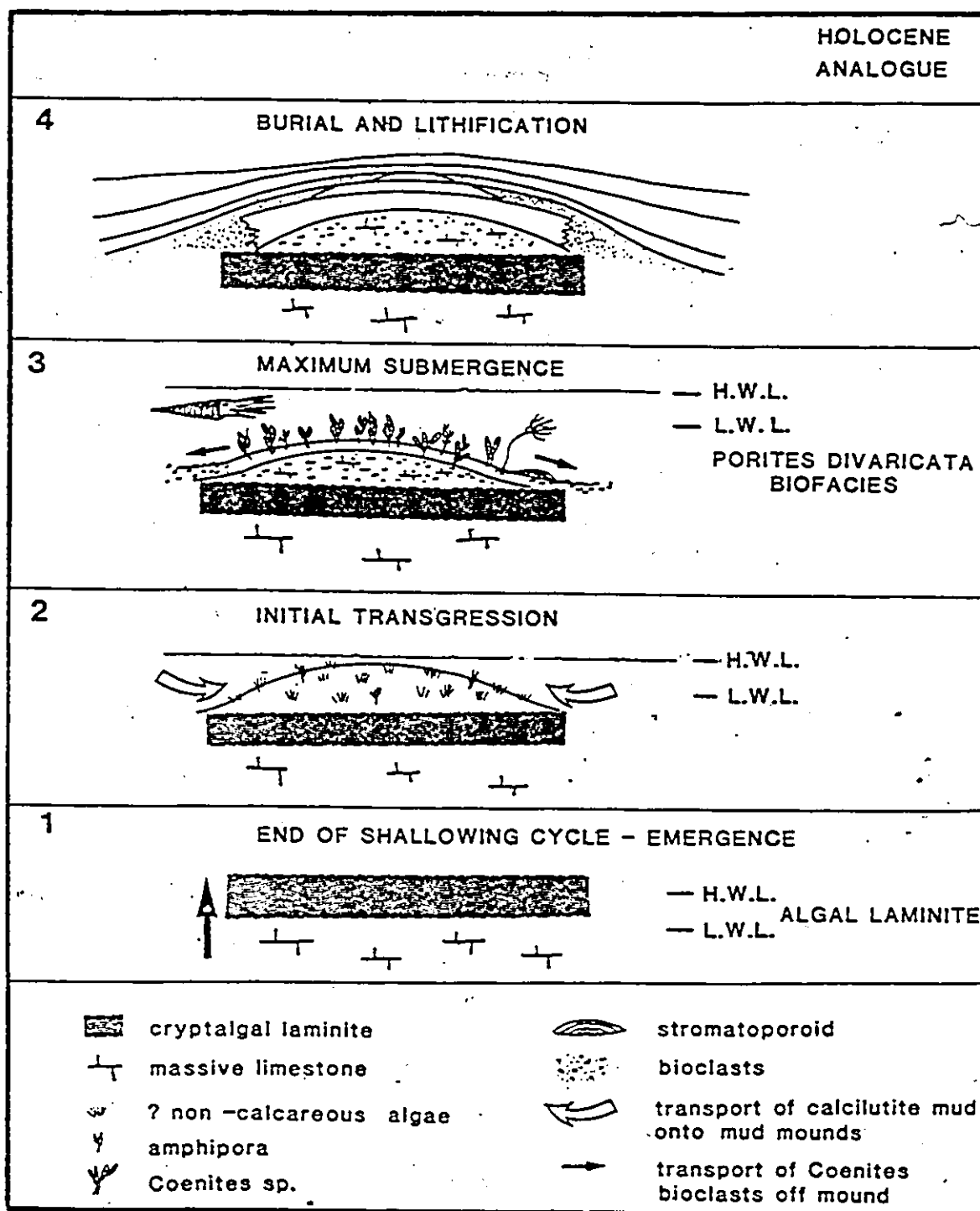


FIG. 4-9 Development sequence for the *Coenites*-capped fenestral mud mound. 1. Algal mats colonized tidal flats. 2. Surface is submerged and fine carbonate mud forms low mounds (bound by non-calcareous algae?). Periodic emergence results in fenestrae. 3. Dendroid tabulates colonize the raised surface of the submergent mound. 4. Synoptic relief of mound decreases and coral thicket is reduced in size and ultimately buried.

The dendroid tabulate thickets were incapable of lateral growth. Normal sediment accumulation gradually decreased the area of colonization of these organisms, until all growth was essentially terminated.

The postulated developmental history for the biocumule is illustrated in FIG. 4-9. The model clearly emphasises the importance of suitable substrates and inherited topographic relief to seafloor colonization. The origin of the initial lime mud mound remains speculative; possibly it resulted from the baffling and binding activity of non-calcareous algae (see Paper 3).

The Coenites -capped, fenestral mud mound exhibits remarkable similarities to modern Florida Bay and lagoon mud banks. Comparison to Rodriguez Bank (Turmel and Swanson, 1976) serves to underline these similarities:

1. both mud mounds commence on laminated (cryptalgal) lime mud(stone).
2. stratigraphic scale is similar: in the Barlow Inlet Formation the base of laminites is at most 7.1 m below the top of the Coenites cap; in the Florida Bay Holocene sediments the base of laminites is 7 m maximum below the bank top (see Turmel and Swanson, 1976, fig.17).
3. in both mounds laminites are superceded by massive lime mud(stone) forming the bulk (volumetrically) of the mound or bank.

4. on Rodriguez Bank, the lime muds are (locally) overlain by windward bank margin sediments consisting of dense accumulations of the finger coral Porites divaricata and the branching calcareous alga Goniolithon sp. The Barlow Inlet biocumule is capped by bafflestone consisting of dendroid tabulates (cf. Coenites) and amphiporids (delicate dendroid stromatoporoids).
5. in both cases the mound or bank is surrounded by a halo (or crescent) of bioclastic gravel or sand.
6. in both cases offbank sediments contain a restricted and impoverished autochthonous fauna/flora.

4.2.11 Zoned Biocumules - Introduction

Several studied biocumules exhibit, to a greater or lesser degree, vertically stacked successions of communities. This biocumule 'type' is non-genetic in the sense that the included examples are obscurely related, if at all. What they do share in common is an autochthonous biofacies succession. It is convenient to divide this diverse group of biocumules into categories that reflect 1) a degree of successional continuity, and 2) directional uniformity of community evolution (Table 4-9). A continuous successional change is one where the in-situ organic constituents change with respect to type and/or relative abundance in the absence of significant

Descriptors applied to zoned biocumules.

Continuity of Succession	: Continuous Discontinuous Hybrid
Directional Uniformity	: Progressive Reversible
Number of Recognizable Communities	: Bizoned Trizoned Quadrizoned Polyzoned
Delineation of zone	: Good Moderate Poor

discontinuities. Significant discontinuities would include such features as karstic surface, hardgrounds, omission surfaces, and units of clearly allochthonous origin (minor shale seams excepted). Covered intervals are also treated here as equivalent to discontinuities. Continuous successional change is generally marked by diffuse community boundaries.

Hybrid zoned biocumules contain both continuous and discontinuous biofacies successions. Progressive sequences proceed in a single direction, and communities do not repeat themselves within any one zoned biocumule. Additional characteristics used to define zoned biocumule categories include number of recognizable communities, and degree of delineation of zones.

Zoned biocumules are particularly important as they can serve genetically and spatially to relate biocumule facies occurring separately elsewhere. Only four zoned biocumules are considered in detail here (A,B,C,D -- FIG. 4-10). The community succession of each zoned biocumule is unique, and is therefore described separately below.

4.2.12 Zoned Biocumules-Example A

4.2.12.1 Size and Shape

Example A is located in a river gorge at loc. 12. The biocumule is 19.9 m thick (PL. 4-13a), the basal 8.8 m of which can be considered bedded strata (PL. 4-13b). The geometry of the biocumule is indeterminate, given

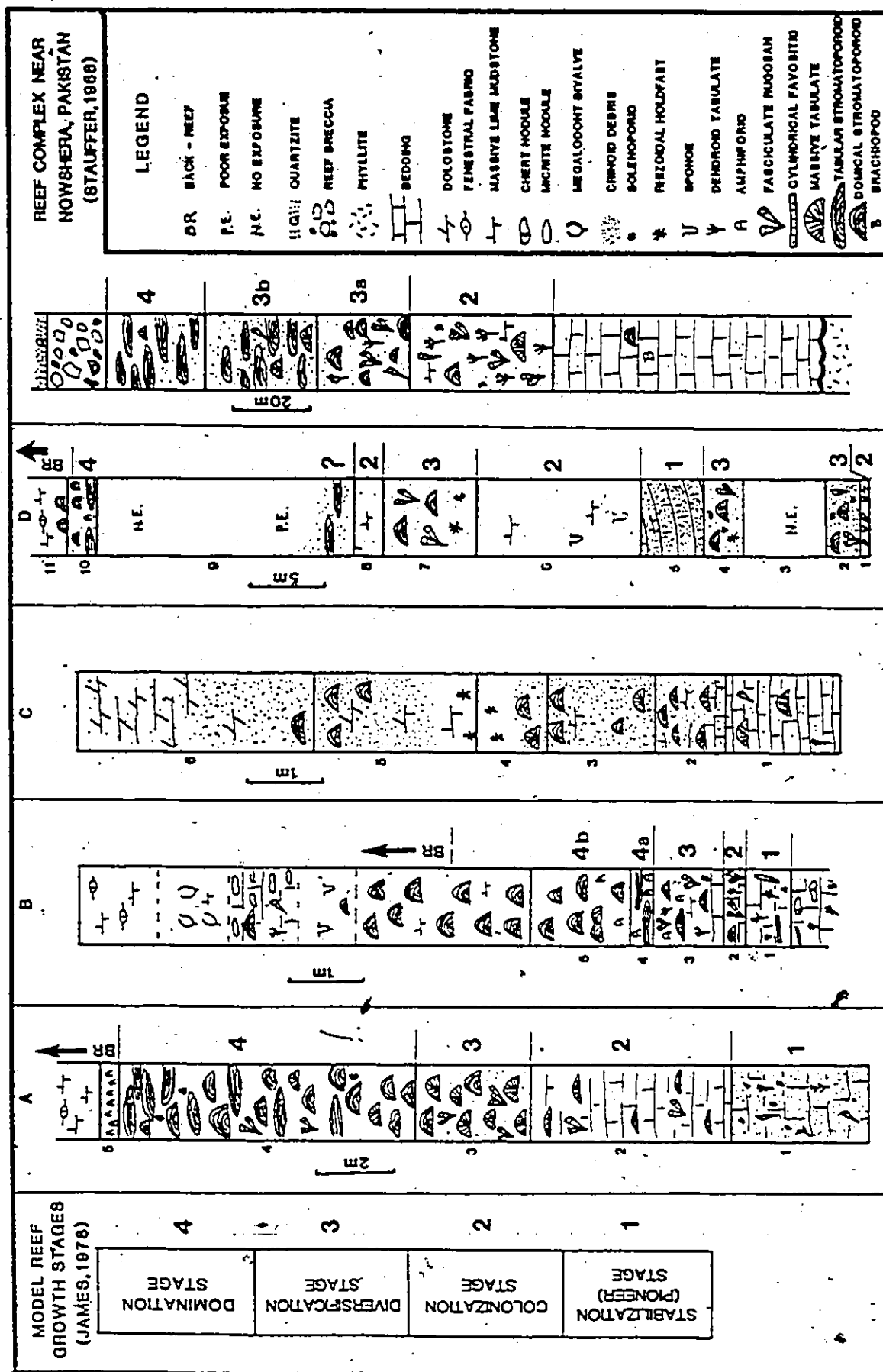


FIG. 4-10 Zoned biocumules: Successions of bio- and lithofacies. Where possible, biocumules are divided into growth stages. Siluro-Devonian reef complex in Pakistan is closely analogous.

the limited exposure, although it is associated with a north-south ridge at least 11 km long that may be an expression of underlying resistant reefal strata. The fact that the biocumule stratigraphically separates open marine proximal foreslope from restricted lagoon facies, suggests that the biocumule formed part of a laterally extensive organic barrier.

4.2.12.2 Community Succession

The basal zone (3.6 m) consists of interbedded bimodally-sorted floatstone with a grainstone matrix, and faintly mottled wackestone. The predominant rudite-sized allochems are crinoid debris, Coenites, low domical favositids and solenoporids. Solitary rugosans, strophomenid brachiopods, orthoconic nautiloids, ostracodes, gastropods and brachiopod shell hash, constitute the other organic remains. The matrix is composed of moderately-sorted peloids, and shell and crinoid skeletal fragments.

The second zone (5.2 m) consists of poorly to massively-bedded, slightly rubbly weathering crinoidal wackestone. Crinoid debris, solitary rugosans, low domical favositids, and solenoporids predominate.

The third zone (3.0 m), is composed of unbedded framestone to inertialstone (dense to frame structure), built principally by Favosites (2 species), Stereoxylodes, an indet. rugose coral, and adnate stromatoporoids (PL.

4-13d,f). Crinoid debris, whole ostracodes and brachiopod shell hash are found in microspar matrix, with mudstone to wackestone texture. Stromatoporoids are accessory frame encrusters in this zone.

The fourth zone (7.7 m) is a framestone composed predominantly of stromatoporoids and Solenoporaceae (PL. 4-13d,f), but including small massive tabulates (Favosites), Stereoxylodes, Coenites, ostracodes and brachiopod shell hash. The proportion of stromatoporoids increases to about 30% of total rock volume near the top, where the growth form is largely lamellar. Stromatoporeid growth forms near the base are low domical to tabular. The matrix is fine crystalline dolomite (80-360 microns) that includes numerous corroded bioclasts.

The fifth zone (0.4 m) is well-sorted, recrystallized packstone composed almost exclusively of amphiporids (cf. A. ramosa) (PL. 4-13h). Solenoporeid algal balls and Coenites are minor constituents. The matrix is micro-to pseudospar. Most allochems exhibit pressure-welded contacts and dolomite is abundant along microstylolites.

Above the fifth zone is a thick sequence of massive, fenestral and non-fenestral lime mudstone (PL. 4-13g) with local concentrations of amphiporids and bulbous stromatoporoids.

4.2.12.3 Environmental Interpretation

The 350 m sequence beneath the biocumule grades from thin, even-bedded, unfossiliferous lime mudstone with a Phycodes-Taenidium-Skolithos trace fossil assemblage at the base, upward into wackestone with a Zoophycus trace fossil assemblage and a diverse stenohaline fauna. This sequence is regarded as representing distal to proximal basin slope facies respectively (Narbonne and Packard, 1982). Massive lime mudstone above the biocumule is interpreted as a restricted back-reef lagoon deposit based on the low diversity biota, abundance of amphiporids, the darker colour of the rocks and the massive bedding (homogenized by burrowing). Krebs (1974) described both a "dark fossil-poor micrite facies" and a "dark Amphipora facies" from the Devonian of central Europe. He considered these to represent quiet, more restricted or deeper areas within the back-reef subtidal environment. The local concentration of bulbous stromatoporoids found in this study are probably analogous to the isolated shelf lagoon patch reefs or "dark bulbous stromatoporoid- Amphipora facies" of Krebs (1974).

Given the environmental constraints imposed by subjacent and suprajacent lithofacies, the zoned biocumule must represent a barrier reef. Zone 5 probably represents amphiporid gravels of the reef flat or proximal back-reef. The pattern of community succession exhibited by this zoned biocumule, bears strong resemblance to the idealized stages

of reef growth proposed by Alberstadt and Walker (1975) (see FIG. 4-10).

4.2.13 Zoned Biocumules-Example B

4.2.13.1 Size and Shape

This zoned biocumule at loc. 7 is only 3.4 m thick. Its geometry is unknown, but as with Example A, the fact that it isolates restricted shelf sediments from open marine foreslope deposits, suggests that it must have been a laterally extensive organic barrier.

4.2.13.2 Community Succession

Beneath the biocumule is faintly mottled, poorly fossiliferous, chert-bearing wackestone containing low domical favositids, cylindrical favositids, brachiopods, Coenites, solitary rugosans and crinoid debris. Units with the trace fossil Zoophycus occur 10 m below the base of the biocumule.

The lowermost zone of the biocumule (0.6 m) consists of faintly-mottled, poorly-defined beds, with a fauna consisting predominantly of Coenites, cylindrical favositids, and solenoporid algal balls. Crinoid debris is rare, and a solitary large domical stromatoporoid was recovered from this zone. Rudite-sized allochems account for 20-30% of the bulk rock. The matrix is micrite.

The second zone (0.3 m) is bafflestone with abundant Stylopleura, and common small low domical, cylindrical and dendroid favositids. Dendroid stromatoporoids (Stachyodes) are rare, as are Solenoporaceae and pentamerid brachiopods. The micritic matrix composes 70% of the bulk rock. Bedding is obscured by pervasive fractures.

The third zone (0.9 m) contains all the organisms listed above in floatstone (30% rudite allochems) with a lime mudstone matrix. Small hemispherical favositids are particularly common, although other faunal elements predominate locally. Areas of relatively unfossiliferous mudstone are also present. Domical stromatoporoids reappear. Towards the top all sense of bedding is lost, and the rock is distinctively mottled. Favositids and fasciculate rugosans predominate in the upper portion of this zone.

The fourth zone (1.6 m) consists of non-bedded, rubbly-weathering, mottled orange buff-light grey rock in which robust dendroid stromatoporoids predominate, and are associated with irregular to tabular stromatoporoids.

The fifth zone (2.5 m) is an inertialstone consisting of irregular to hemispherical stromatoporoids set in dark grey fine-grained matrix. Dendroid stromatoporoids are minor faunal constituents.

Overlying the biocumule gradationally is a succession of predominantly lime mudstone facies that includes, in stratigraphic order, a sponge-rich mauve grey mudstone with

rare irregular stromatoporoids; a distinctively mottled lime mudstone with rare rugose corals, Coenites, and irregular to lamellar stromatoporoids; a Megalamoidea floatstone; and finally a light grey fenestral mudstone.

4.2.13.3 Environmental Interpretation

In contrast to Example A, the pioneer community (stabilization stage) is represented by Coenites floatstone and not crinoidal packstone. The paucity of crinoids in the biocumule as a whole remains enigmatic. Coenites and Solenoporaceae are common to both Examples A and B. The second and third stages are comparable to the colonization and diversification stages of biocumule A. The fourth and fifth zones both represent community domination by single taxa or morphotypes, featuring dendroid and irregular to hemispherical stromatoporoids respectively. The very gradational upper contact of the biocumule, the presence of two domination substages, and the absence of a 'reef-crest' tabular, stromatoporoid framestone or reef pavement gravel, suggest that the environmental settings for biocumules A and B were significantly different. This difference is reflected as well in the suprajacent beds. Of the facies above the biocumule, three are common to the 'Restricted Infratidal' sequences of the back-reef lagoon (Paper 3). The sponge-rich facies, however, is unique and its paleoenvironmental significance is not yet known. The spicular framework is

quite dissimilar to those exhibited by the lithistid sponges of the foreslope mud mounds.

Overall the faunal diversity in the back reef of biocumule B, in conjunction with a very gradational reef-back reef contact, suggest that biocumule B was not a fully mature barrier reef, and probably did not grow in a particularly hydrodynamically-stressful environment.

4.2.14 Zoned Biocumules-Example C

Example C is a continuous reversible-polyzoned biocumule at loc. 16, stratigraphically equivalent to, and laterally 600 m to the south of, a stromatoporoid-crinoid rubble skeletal buildup.

4.2.14.1 Size and Shape

The shape of the biocumule is unknown due to limited outcrop. A total stratigraphic sequence of 9.8 m is exposed.

4.2.14.2 Community Succession

The basal 1.5 m is bedded crinoidal wackestone with common low domical favositid colonies and solitary rugose corals. From 1.5 to 2.4 m, bedding becomes less apparent, the relative abundance of favositids increases and the matrix is crinoidal rudstone to packstone. The ensuing interval (2.4 - 4 m) is unbedded, inertialstone-framestone with crinoidal rudstone matrix. Low domical favositids

predominate in the lower 2/3 of this interval, whereas massive stromatoporoids are more common in the upper third. A 1.0 m interval of rhizoidal holdfast bindstone (75% lime mudstone) follows and is then gradationally overlain by unbedded crinoidal rudstone to packstone rich in stromatoporoids (1.8 m). The uppermost exposed units (3.0 m) consist of thickly-bedded, dolomitic crinoidal rudstone to floatstone, with an apparent decrease in crinoid ossicles upsection.

4.2.14.3 Environmental Interpretation

Overall this zoned biocumule is similar to Example A (shelf margin barrier reef), with a bedded favositid- and rugosan-dominated colonization stage, passing upwards into an unbedded zone (diversification to domination stage) dominated by massive stromatoporoids set in crinoidal matrix. The latter zone is represented twice in Example C, separated by rhizoidal holdfast bindstone. It is not known at present, whether the holdfast bindstone represents a widespread return to a colonization stage for this particular biocumule, or whether the interval is, in three dimensions, an isolated pod similar in development to the 'crinoid nests' of the crinoid blanket deposits (see 'Synthesis'). Although in itself inconclusive, the presence of thoroughly dolomitized rudstone lends weight to the interpretation of the biocumule occurring at the shelf

margin.²³ Unlike Example A the biocumule culminates in crinoidal gravel and sand, not back-reef mudstone. Likewise the stromatoporoid crinoid rubble skeletal buildup at loc. 17 terminates in dolomitized crinoidal sand. Unfortunately, overlying sediments are not exposed. Regionally both biocumules (locs. 16 and 17) are laterally surrounded by Cape Phillips Formation strata (basinal and basin slope facies). This suggests that both biocumules were isolated 'pinnacle' reefs (or one single reef). If the uppermost dolomitized crinoidal rudstone was laterally extensive, it could represent the fusion of the contiguous shelf margin with the fore-reef 'pinnacle' reef facies. Alternatively, the rudstone could represent shoals of bioclastic sands localized on top of the biocumules at a time close to the peak of shallowing experienced by the biocumule.

4.2.15 Zoned Biocumules-Example D

4.2.15.1 Size and Shape

Example D is a discontinuous-reversible polyzoned biocumule at loc. 14. Boundstone facies occur discontinuously throughout approximately 50 m of stratigraphic section. The bulk of the zoned biocumule appears in weathered relief to be a vertically elongate, irregular mound-shaped structure (PL. 4-14a). The uppermost

²³ pervasive dolomitization in the Barlow Inlet Formation is generally restricted to shelf edge and proximal back-barrier facies.

boundstone facies is likely biostromal with a minimum lateral extent of 400 m for thickness of at least 6 m.

4.2.15.2 Community Succession

The lowest exposed unit (see FIG. 4-10) is a 0.6 m thick dolomitized fasciculate rugosan biocumule with a muddy matrix. It is overlain abruptly by rudstone (2.1 m) consisting of tabular to hemispherical stromatoporoids, hemispherical favositids, fasciculate rugosans, and solenoporids set in crinoidal packstone (predominantly) to mudstone matrix. Sixty-six percent of the colonial metazoans appear upright. Following a covered interval of 5.3 m, there is a unit (2.7 m) of inertialstone to framestone, with tabular to hemispherical stromatoporoids, hemispherical favositids, Solenoporaceae, fasciculate rugosans, rhizoidal holdfasts, and hemispherical bryozoans in a wackestone matrix (PL. 4-14b). Allochthonous components include crinoid debris, dendroid bryozoans, trilobite and brachiopod fragments. The wackestone matrix is a microspar (5-10 microns). Abruptly overlying this unit is crinoidal grainstone to packstone with the following characteristics:

1. rare? planar cross-bedded strata with cross-sets up to 15 cm thick. Positively identified cross-strata appear to be restricted to units with mean crinoidal grain size less than 1 mm (PL. 4-14d).

2. rare graded-bedding (PL. 4-14e) can probably be related to sorting during grain flow on the avalanche faces of small dunes responsible for the observed cross-bedding.
3. larger beds 10 - 50 cm thick (PL. 4-14f) are composed of massive, moderately-sorted crinoid-Solenoporaceae rudstone (grain size range 1 to 22 mm, average 2 - 4 mm). These large beds are not apparently cross-stratified, although they may be wedge-shaped. No graded bedding was observed. The top of each bed has a thin layer (less than 1 cm to 5 cm thick) of lime mudstone to wackestone. Vertically and laterally, these rudstone beds interfinger with massive to poorly bedded lime mudstone (PL. 4-14c).

Overlying the crinoid-Solenoporaceae rudstone is a 10.5 m interval of massive, lime mudstone, with very rare bryozoan, brachiopod, gastropod, and crinoid debris. Patches of spicular framework are present, but generally positive identification of sponge material is not possible. Apart from the aggradation of micrite to microspar and the growth of euhedral dolomite rhombs (generally less than 2% of rock), textures and allochems are well-preserved. If sponges are principally responsible for constructing this portion of the biocumule, then the loss of spicules must be ascribed to rather early diagenetic (silica dissolution) processes.

The gradationally overlying zone (6.0 m) is a massive unit similar in all respects to unit 3, with fasciculate rugosans, tabular and hemispherical stromatoporoids and rhizoidal holdfasts forming a framestone (PL. 4-14g).

Above unit 7, the relative abundance of skeletal fragments decreases considerably and the lithology of unit 8 is a massive crinoidal mudstone/wackestone similar to more skeletal-rich portions of unit 6. Following a thick, poorly-exposed interval (16.5 m) (which contains in its lower portion, tabular stromatoporoid and crinoid debris), the first well-exposed unit is a tabular stromatoporoid inertialstone with scattered crinoid debris and dendroid stromatoporoids. This unit is 2 m thick, and is separated by a hemispherical stromatoporoid inertialstone that passes gradationally upward into massive fenestral lime mudstone (PL. 4-14h) with megalodont bivalves.

4.2.15.3 Environmental Interpretation

Three discontinuities are present within the succession: two covered intervals and an interval of allochthonous crinoidal sediment. The significance of the covered intervals cannot be gauged, but they do not appear to separate drastically different facies. The abrupt contact between units 3 and 5 suggests that the underlying biocumule (unit 4) ceased growth because of encroachment of loose crinoidal sediment. Assuming most of the featureless lime

mudstone above the crinoid-Solenoporaceae rudstone was originally sponge boundstone, the faunal succession is similar to that in Upper Silurian lithistid sponge mud mounds described by Narbonne (1981). The latter are considered to have developed in calm water, periodically subject to wave turbulence (Rigby and Dixon, 1979) with elevated levels of turbidity (Narbonne, 1981), probably between storm and normal wave base.

The zoned biocumule at loc. 14 presents conflicting evidence on the relative influence of allogenic versus autogenic controls on faunal succession. As illustrated in FIG. 4-10, most of the sequence can be interpreted in terms of a simple autogenically-governed sequence, with a line of section commencing within a colonization stage of an underlying mound and including the stabilization, colonization and diversification stages of an overlying biocumule. There is no necessity for the change in the overall environment or bathymetry of deposition, as the faunal succession (up to unit 4) can be satisfactorily explained in terms of substrate specificity of organisms, and increasing availability and complexity of micro-habitats. However, one postulate of intrinsic community development and succession, is that it is progressive and cannot be reversed, as long as external conditions remain relatively constant. The facies change (units 3 - 4) is suggestive of retrogression to a

colonization stage, and must therefore be the result of an externally-modified environment (such as influx of fine terrigenous clastics).

The ultimate shallowing, as exhibited by intertidally-deposited fenestral mudstone (unit 11), clearly reflects allogenic control on facies development. Changing environmental conditions resulted from either the growth of the biocumule to sea-level, or regional tectonic uplift of the platform on which the biocumule was developing, or some combination of both.

4.3 COMPARISON TO OTHER SILURO-DEVONIAN BUILDUPS

As stated by Wilson (1975, p.105) and reiterated by others (Smosna and Warshauer, 1979; Isaacson and Curran, 1981), "Carbonate buildups of latest Silurian and earliest Devonian age are very rare, despite extensive Middle Paleozoic exposure in the northern hemisphere". Recent studies by Sodero and Hobson (1979), Bourque (1979), Narbonne and Dixon (1977, 1983), and Packard and Dixon (1979, 1982) on hitherto little-studied carbonate sequences in relatively less accessible areas, have begun to address the information gap on latest Ludlovian and Pridolian biocumules. In addition, many Great Lakes region 'reefs' are more youthful than previously supposed (Shaver, 1977), and this information can be integrated into a more complete synthesis on latest Silurian biocumule facies. Lower Devonian reefs remain poorly understood, although in the past few years effort has been directed toward their study, as is evident in the works of Smosna and Warshauer (1979), and Isaacson and Curran (1981).

Unfortunately it is not possible to compare the Barlow Inlet rimmed shelf complex in its entirety, with other individual Silurian carbonate platforms. Few studies on Silurian reefs have incorporated more than interpretation of isolated buildups, or concentrations of buildups, that occur on largely undifferentiated shelves (e.g. Great Lakes region, Gotland, Wenlock Edge). Sodero and Hobson (1979)

have studied the shelf and shelf margin carbonates of the Allen Bay Formation, (Lland.- Ludlow) of Arctic Canada, but pervasive dolomitization in these rocks makes comparison to Barlow Inlet Formation lithofacies difficult. Bourque (1979) and Hurst (1980, 1981) have published preliminary data on Silurian shelf and shelf margin facies from extensive and well-preserved exposures in the Gaspé peninsula (Canada) and N.W. Greenland, respectively. Still, currently ~~available~~ paleoenvironmental interpretations for these latter carbonate complexes is too limited to contrast them properly with the biocumule facies of this study. As a result, the most favourable comparisons based on the complete spectrum of platform facies present, can (and must) be made with the well-studied Middle and Late Devonian reef complexes of Europe, Australia, and Western Canada (see below).

Taken individually, or in smaller, environmentally-related groups, the biocumules can be usefully compared to a variety of Siluro-Devonian 'reef' and 'reef-related' facies described in the literature, as follows:

1. The lithistid sponge dominated mud mounds are closely comparable with lower sponge facies of small buildups that occur in late Ludlovian carbonates of the Douro Formation on Somerset Island, Arctic Canada (Narbonne, 1981; Narbonne and Dixon 1977, in press; Rigby and Dixon, 1979). The two biocumule facies

differ in that the Douro mounds have well-preserved sponges with a variety of growth forms, whereas the Barlow Inlet examples generally show poor preservation of principally irregular encrusting sponges which are commonly associated with spongiostromate lamination. The Barlow Inlet mounds are generally larger, lack the distinctive bioclastic halo of their Douro counterparts, and have a less diverse community of preserved organisms.

The Douro mounds occur on a carbonate ramp, interpreted as between storm and normal wave base, whereas the Barlow Inlet sponge mounds, although probably similarly situated bathymetrically, occur basinward of a rimmed shelf margin. Other Silurian lithistid sponge mounds have been described by Gingac and Bourque (1982), in Ludlovian strata of the Gaspé Peninsula. These resemble the Barlow Inlet mounds in biotic makeup and in size, but are more distinctively domical in outline (Bourque, 1979, fig. 7), and are more diagenetically altered. The featureless foreslope mud mounds described by Hurst (1980) could be constructed by lithistid sponges, and would therefore be clearly analogous to the Barlow Inlet biocumules.

Stromatactoid structures of the Barlow Inlet mounds (PL. 4-2d), and more pervasive stromatactis in

the Gaspé sponge mounds, may both be related to dissolution of 'sponge framework, or enlargement of growth cavities (see Gingac and Bourque, 1982). A similar origin may be advanced for stromatactoid mudstones within Niagaran reefs of the Great Lakes region. Significantly, co-occurrence of stromatactoid structures, sponges, and syringoporiid corals characterizes mud mounds or basal biohermal mudstone facies of Silurian age in the Great Lakes, Gaspé and arctic regions.

The lithistid sponge dominated mud mounds can also be compared to siliceous sponge mounds of Early and Mid-Ordovician age (Rigby, 1965, 1971; Stevens and James, 1976; Toomey and Nitecki, 1979; Klappa and James, 1980) and Mid- to Late Jurassic age (Aldinger, 1968; Palmer and Fursich, 1980, and Flügel and Steiger, 1981). The mound described by Flügel and Steiger (1981), formed by intergrowth of algal crusts and lithistid sponges in a basinal setting below normal wave base, is particularly reminiscent of both the principal biotic constituents and paleogeographic setting of the Barlow Inlet biocumules.

2. To the writer's knowledge, mud mounds composed predominantly of saucer-shaped solitary rugose corals (mucophyllids), are unique to the Upper Silurian of Arctic Canada. Packard and Dixon (1983) reported

mucophyllid-rich mud mounds in basin slope deposits of the upper part of the Douro Formation on Cornwallis Island.

3. The diverse community mud mounds are strikingly similar to Axelsro-type reefs, and to a lesser extent Hoburgen-type reefs described by Riding (1981) in Wenlock to Ludlow age strata of Europe. These reef types collectively embrace all reefoid structures of Wenlock Edge (see Scoffin, 1971; Abbott, 1976; Riding, 1981) and many of the biocumules of Norway, Estonia and Gotland (see Manten, 1971; Riding, 1981). In addition, the coral bioherms described by Philcox (1970) in the Silurian Hopkinton Formation of Iowa, can probably be considered of Axelsro type. Table 4-10,11 contrasts the diverse community mud mounds with Axelsro and Hoburgen reef types. There is in fact no sharp demarcation between these two, and the diverse community mud mounds appear to share characteristics of both. For example, stromatoporoids are subordinate to corals in the biocumule biota (Axelsro characteristic) but solenoporids are extremely abundant locally (Hoburgen characteristic). The size range of the Barlow Inlet biocumules exceeds that of Axelsro-type reefs, but maximum sizes of the diverse community mud mounds do not approach those achieved by Hoburgen-type reefs

Comparison of Axelsro and Hoburgen reef types, and
diverse community mud mounds

CHARACTERISTIC	AXELSRO REEFS		HOBURGEN REEFS, GOTLAND ²	DIVERSE COMMUNITY MUD MOUNDS
	WENLOCK EDGE ¹	GOTLAND		
Shape (cross-sectional outline)	irregular oval	irregular lensoid to conical	irregular elongate lensoid	irregular ellipse
Range of height/length ratio	(.2-.9)?	(.3-.9) ² (.2-1.0) ⁴	(.1-.8) ² (.02-1.0) ⁴	(.3 -.7)
Average size (exposed ht. x l.)	4.5x12m	-	-	2x4m
Max. size	20x100m	5x5m	-	5.4x10+m
Cross-sectional area (avg./max.) in m ²	54/<2000	<10/?<30	100/?<6500	<8/54 plus
Synoptic relief	.5-3m	<---- up to	10m ---->	.8m(for 1 mound)
Structure	sparse	sparse-dense	dense-frame	sparse-frame
Mound Matrix	micrite	marly	strongly marly	lime mudstone to wackestone
Growth cavities	yes	-?	?	yes
Diversity of preserved biota ³	?(22 min)	10 - 60	20 - 200	13 - 25
Shape of stromatoporoid colonies	variable; domical and laminar	domical ³ rather flat ⁴	variable, laminar to domical lenticular ⁴	predominantly lamellar, tubular to domical in largest structure

(continued)

Relative proportion of corals to corals plus stromatoporoids	<--- 25 - 80% --->		10-60%	50 to 100%
Consistency of biofacies within and between mounds	poor	poor	poor-moderate	poor
Systematic vertical zonation	no	no	yes	only in largest structure
Subjacent facies	bioclastic gravels	marlstone or bioclastic gravel	bioclastic sands and gravel	bioclastic gravels, rarely marlstones
Adjacent/suprajacent facies	nodular limestone, crinoidal mudstone to packstone	marlstone	bioclastic sands, Küppen-type reefs	marlstones to crinoidal wackestones and packstones
Biotic constituents	see Table 4-11			

1. data from Scoffin(1971), Abbott(1976), Riding(1981)
2. data from Manten(1971), Riding(1981)
3. data from Riding(1981)
4. data from Manten(1971)
5. Diversity at the species level. Diversity totals are biased due to greater degree of study of European reefs.

Biotic constituents of Axelsro reefs and diverse community mud mounds.
Presence(x), questionable record(?), or absence(-) shown.

BIOTA	Axelsro- type reef of Wenlock Edge	Axelsro- type reef, Gotland	Diverse Community Mud Mounds (DCM).	Remarks pertinent to DCM
ALGAE (and algal problematica)	x	x	x	
<u>Girvanella</u>	x	-	x	
<u>Rothpletzella</u>	x	x	x	<u>Sphaerocodium</u> sp.
<u>Wetheredella</u>	x	x	x	? <u>Wetheredella</u> sp.
spongiostromate algae	x	-	x	oncoids only
solenoporids	-	-	x	<u>Solenopora</u> sp.
SPONGES	-	x	x	
TABULATE CORALS	x	x	x	
<u>Halysites</u>	x	x	x	<u>Catenipora</u> sp.
<u>Alveolites</u>	x	-	x	
<u>Thecia</u>	x	x	-	
<u>Favosites</u>	x	x	x	
<u>Heliolites</u> ²	x	x	x	poroid corals
<u>Coenites</u> ³	?	?	x	
syringoporids	-	x	x	
RUGOSE CORALS	x	x	x	
solitary forms	x	x	x	largely <u>Nanshanophyllum</u> sp.
fasciculate forms	x	x	x	largely <u>Stereoxylloides</u> sp.
mucophyllids	-	-	x	
<u>Acervularia</u>	-	x	-	
<u>Entelophyllum</u>	x	-	-	
<u>Ketophyllum</u>	-	x	-	
STROMATOPORIDS	x	x	x	
massive	x	x	x	
adnate(lamellar)	x	-	x	
robust dendroid	-	-	-	
<u>Stromatopora</u>	x	x	-	
<u>Actinostroma</u>	x	x	-	
<u>Labechia</u>	-	x	x	

(continued)

Table 4-11 (concluded)

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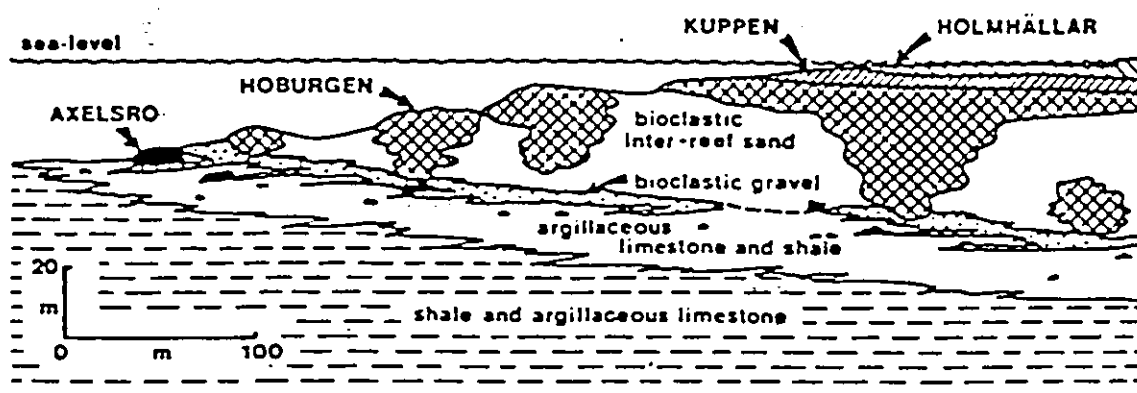
BRYOZOANS	x	x	x	both adnate and dendroid growth forms
<u>Hallopora</u>	x	-	-	
<u>Rhombopora</u>	x	-	-	
<u>Fistulipora</u>	x	x	-	
<u>Fenestella</u>	x	x	-	
<u>Coenites</u> ²	?	x	?	
ECHINODERMS	x	x	x	
rhizoid holdfasts	-	-	x	
pelma and calyx debris	x	x	x	
BRACHIOPODS	x	x	x	
rhynchonellids	-	x	x	
<u>Atrypa</u> sp.	-	x	-	
<u>Leptaena</u> sp.	-	x	-	
atrypoids	-	-	x	
MOLLUSKS	x	x	x	
bivalves	x	x	-	
gastropods	-	x	x	
ARTHROPODS	x	x	-	
ostracodes	x	x	-	
trilobites	x	x	-	

1. the above list is very incomplete even at the generic level. In particular the rugose coral, stromatoporoid, brachiopod and echinoderm faunas associated with Axelsro reefs have been the subject of detailed taxonomic investigations. The data given above is drawn principally from Scoffin (1971), Manten (1971) and Riding (1981). Inasmuch as systematic work on fossils from the DCMs has been completed, the only fruitful comparisons can be made at higher taxonomic levels.
2. included in the tabulates for convenience
3. Riding (1981) discussed a bryozoan called Coenites!

Hoburgen reefs contain greater species diversity, and different proportions of the larger taxonomic groups. They can be differentiated (on the basis of biofacies) from Axelsro reefs on Gotland, by the presence of abundant algae (Solenoporacea), and relatively more stromatoporoids with laminar growth form.

(Manten, 1971; Riding, 1981). The common development of flank beds around diverse community mud mounds is suggestive of Hoburgen-type reefs, but unlike these reefs, the Barlow Inlet mounds do not exhibit a climax community of dendroid rugosans and solenoporaceans in a tight framework. The structure of the biocumule, the predominant baffling and binding habits of the corals, and the variability of organic framework indicate greater affinities to Axelsro-type reefs.

Axelsro-Hoburgen reef types are sequentially developed in shallowing upward successions (Riding, 1981; see FIG. 4-11) with Axelsro reefs developed in waters below normal wave base, and Hoburgen reefs in a zone affected by wave turbulence (Manten, 1971).



INTERPRETED RELATIONSHIP BETWEEN GOTLAND
REEF TYPES (FROM RIDING, 1981)

CORRESPONDENCE BETWEEN SILURIAN BIOCUMULES
OF GOTLAND AND ARCTIC CANADA

GOTLAND	AXELSRO	HOBURGEN	KUPPEN	HOLMHALLAR
ARCTIC CANADA	3	3a	7	6

3 DIVERSE COMMUNITY MUD MOUNDS

3a LARGE, ZONED DCM AT LOC. 3

7 DOLOMITIZED MASSIVE SKELETAL (? CRINOID) BIOCUMULE

6 LAMELLAR STROMATOPOROID BIOSTROMES

FIG. 4-11 Gotland reef sequences and comparison with Barlow Inlet Formation biocumules

The diverse community mud mounds of Arctic Canada are thought to have developed at, or just below, normal wave base.

A prominent feature of the diverse community biocumules is their abundance of rhizoidal holdfasts. This is not a feature commonly noted in Paleozoic reef papers. Narbonne (1981) and Kissling (1969) have both stressed the important binding role played by these organic structures.

4. Biostromal deposits of fasciculate rugose corals are rarely discussed as such, but concentrations of fasciculate rugosans form important biofacies of uncertain geometries in several stratigraphic reefs. Bourque (1979) recorded fasciculate rugosans that form a pioneer community beneath sponge-stromatactoid mud mounds in the Upper Silurian of Gaspé. Williams (1980) recorded a fasciculate coral (Acinophyllum) forming laterally extensive 'pre-reef' facies, as well as contributing substantially to muddy reef core of a mid-Devonian buildup in New York State. Isaacson and Curran (1981) reported 'dendritic' rugosans set in micritic matrix as comprising the bulk of either first or second development stages in Lower Devonian buildups in New York.

In contrast to representation in pioneer communities, fasciculate rugosans in conjunction with

dendroid solenoporaceans constitute the climax community in Hoburgen reefs (Riding, 1981). It is clear from the present study that genotype more than morphotype dictates the environmental preference of various rugosans with fasciculate growth form. The stylopleurids are consistently recorded from strata regarded as calm to subturbulent, whereas Stereoxyloides appears more environmentally tolerant. Wilson (1975, p.142) cautioned that overgeneralization of the paleoenvironmental significance of fasciculate rugosans is easy. Evidently, much further work is required in elucidating the controls on semi-colonial coral growth.

5. The stromatoporoid-crinoid rubble skeletal buildups fall under the general category of organically-bound skeletal debris buildups (Heckel, 1977). Examples of massive stromatoporoid biocumules with crinoidal rudstone to grainstone to wackestone matrix, have been described in the Wenlockian barrier reef complex of the Michigan Basin (Mesolella et al., 1974), and in Llandoveryian buildups (crinoidal calcarenite association) of N.W. Greenland (Hurst, 1980). The Silurian Kuppen and Holmhällar biostromes of Gotland (Sweden), contain a coarse crinoidal matrix in association with a dense to frame structure of

stromatoporoids (Riding, 1981). These deposits, however, are biostromal, not mound-shaped, and in the Kuppen biostromes the principal stromatoporoid morphotype is tabular to laminar. The stromatoporoid floatstone biostromes described by Hurst (1980) also contain a crinoidal rudstone matrix, but have predominantly hemispherical stromatoporoid morphotypes. Devonian buildups do not appear to exhibit as strong an association between crinoid debris and massive stromatoporoids. However Burchette (1981) stated that coarse unsorted bioclastic debris (at least in part crinoidal) usually fills inter-skeletal areas in the reef core of European buildups.

Large buildups composed predominantly of bioclastic crinoidal debris, with few metazoan colonials have been described by Philcox (1970b) in the mid-Silurian strata of Iowa, and Bourque (1979) in the Upper Silurian Cap de l'Enfer facies of the Gaspé.

6. Distinctive biofacies that feature tabular-lamellar stromatoporoids are reported by Mesolella et al. (1974) (Silurian pinnacle reefs of the Michigan Basin), Riding (1981) (Holmhällar reef types of Gotland), Embry and Klovan (1972) (mid- Devonian buildups, Arctic Canada), Jamieson (1971) (Upper

Devonian reef complex, western Canada), and Playford and Lowry (1966) (Upper Devonian barrier reefs in Canning Basin, western Australia). Ecologic placement of this biofacies ranges from deep fore-reef, to reef crest, to back-reef. There remains no consensus on the environmental significance of this growth form (Stearn, 1965) partly because a satisfactory distinction is rarely made between tabular and adnate laminar morphotypes (see Abbott, 1973), and little account is taken of genotypic variation (Kershaw, 1981; Stearn, 1981).

7. Due to the pervasive coarse dolomitization of the dolomitized massive skeletal (?crinoid) biocumule, comparisons unfortunately reflect diagenetic textures rather more than primary depositional features. Those units within these biocumules which are stromatoporoid-rich commonly exhibit a diverse morphotype assemblage, reminiscent of the biostromes described by Kershaw(1981) in the Ludlovian shallow-water Kuppen reefs of Gotland. The non-diagnostic massive dolostones that form part of this biocumule facies can be compared to the featureless Siluro-Devonian shelf-edge dolostones of Sodero and Hobson (1979), and Mountjoy (1969), and possibly the structureless micrite described by Playford (1969). Many authors confronted with this

pervasively dolomitized facies have assumed, on the basis of enclosing and interpretable facies, that the dolostones are organically-bound shelf margin buildups. Wilson (1975, p.137), commenting on the surface expression of Devonian buildups in western Canada, stated that, "... margins of massive light-colored, coarse, sucrose dolomite must represent organic growths of algae and stromatoporoid".

8. Sheet-like concentrations of stromatoporoids with massive but generally small coenostea, have been described in many Devonian proximal back-reef facies; the reader is referred to Paper 3 for a selection of pertinent papers. The only difference noted between Mid-Late Devonian and Late Silurian stromatoporoid biostromes, is that, whereas bulbous growth forms are common in the former, hemispherical, subspherical and/or irregular morphotypes are more prevalent in the latter.
9. Biostromes and isolated pockets of dendroid (generally amphiporid) stromatoporoids have been routinely described and placed in facies context by researchers ever since Andrichuk (1957) originally assigned paleoenvironmental significance to Amphipora. The voluminous literature on Mid-Late Devonian reef complexes contains numerous references

that deal in part with this biocumule facies (e.g. Jamieson, 1971; Kobluk, 1975; Krebs, 1974; Havard and Oldershaw, 1976). To the writer's knowledge, only Bourque (1979) has noted dendroid stromatoporoids in Silurian back-reef facies. Given that their known stratigraphic range extends from the Silurian (Lecompte, 1956), this circumstance must be credited to a paucity of preserved Silurian lagoonal facies.

The monospecific concentrations of amphiporids in the Barlow Inlet Formation are strikingly similar to biostromes described by Krebs (1974) in Devonian back-reef facies of Europe. The association of lamellar stromatoporoids with Stachyodes has been documented previously by Jamieson (1971) and Fischbuch (1967). The co-occurrence of dendroid tabulates (e.g. Coenites, Thamnopora), Stachyodes, and amphiporids has been noted by St. Jean (1971). The autochthony of foreslope amphiporids, as advanced herein, has not been previously documented.

10. There are few ancient analogs to the Coenites-capped, fenestral mud mounds, although they bear considerable resemblance to Holocene, shallow-water mud banks (see above). Hurst (1980) has described dolomitized fenestral buildups in the Llandovery of N.W. Greenland that bear some resemblance to the Barlow Inlet biocumules (presence of fenestrae and

cryptalgal lamination), although they are an order of magnitude larger (up to 150 m in stratigraphic relief) and are as yet unstudied in detail. The calcilutite mud mounds described by Heckel(1974) from the Mid-Upper Devonian Tulley Limestone of New York, bear some resemblance to the Barlow Inlet biocumules in that 1) they are composed largely of calcilutite mudstone, 2) the sparse fauna consists principally of auloporids and slender tabulates such as Cladopora (= Coenites) and Striatopora, 3) stromatactoid structures are common, 4) the mounds are associated with both a muddy restricted shelf lithofacies and a more open marine bioclastic lithofacies, 5) the mounds are no thicker than 5m although they are laterally extensive (30 X 300m?).

Other small, stromatactoid lime mud mounds are described from the Middle Devonian of the Spanish Sahara by Dumestre and Illing(1967).

11.° For discussion purposes, it is useful to consider the lithistid sponge-rich zoned biocumule separately from biocumules that apparently do not contain abundant sponges.

If much of the undifferentiated massive lime mudstone in the zoned biocumule of example D was formerly lithistid sponge, as is inferred here, then the zonation seen in this biocumule would be

comparable to the sponge mud mounds described by Narbonne (1981) in the Ludlovian Douro Formation on Somerset Island. The principal difference would be that the coral-stromatoporoid facies of the Barlow Inlet biocumule is overlain by restricted shelf fenestral mudstones, whereas the Douro Formation mud mounds apparently did not evolve beyond the diversification stage represented by a similar coral-stromatoporoid facies. The Barlow Inlet Formation biocumule clearly developed within an overall regressive sequence, whereas the Douro mounds are interpreted to have grown at the peaks of transgressive cycles (Narbonne, 1981). Flügel and Steiger (1981) reported that the shallowest facies of Upper Jurassic sponge mounds, tend to be more coral-rich, a faunal succession that mirrors the change in biofacies seen in the regressive Barlow Inlet biocumules.

The other zoned biocumules compare favourably to generalized models of Siluro-Devonian 'reef' biofacies succession outlined by James (1983); a tabulate-rich basal biofacies is succeeded by a stromatoporoid-dominated biofacies. Michigan Basin pinnacle reefs (Silurian) described by Mesolella et al. (1974) have a basal crinoidal zone, a middle coral-algal zone and a terminal laminar

stromatoporoid zone which may equate to the principal stages of the zoned biocumules at loc. 7, 12 and 16. However, several facies are recognized in the (discontinuously) polyzoned Michigan Basin pinnacle reefs that are not recognized in the Barlow Inlet biocumules. Jamieson (1971) described a proto-reef to reef succession in the Upper Devonian of western Canada that features a basal dendroid favositid facies, a cerioid coral facies, and an upper stromatoporoid-algal facies, broadly comparable to the succession seen in zoned biocumules at locs. 7 and 12.

Large Hoburgen reefs (Silurian of Gotland) display an upward faunal succession commencing with a crinoid gravel, then a tabulate-rich facies, succeeded by a stromatoporoid-rich facies (Riding, 1981).

The reef biofacies zonation that is most similar to the Barlow Inlet sponge-poor zoned biocumules, is in the Nowsherra reef complex of latest Silurian-earliest Devonian age in Pakistan (Stauffer, 1969). The Nowsherra sequence of biofacies, illustrated in FIG. 4-10, is considered a shallowing upward succession.

Taken in its entirety, the Barlow Inlet Platform compares favourably with Devonian reef complexes of Emsian to Frasnian age.

The volume of reef literature concerned with this time period is formidable, but fortunately a number of detailed summaries on Australian, Canadian and European Devonian reef complexes is available in the English language. The reader is referred to Playford (1980) -- Canning Basin, Klován (1974) and Davies (1975) -- western Canadian Basin, and Krebs (1974) and Burchette (1981) -- Europe, for excellent regional syntheses or collections of papers on the structure and facies of Devonian buildups. In addition, useful comparisons between Australian and Canadian Devonian reefs on one hand, and Canadian and European reefs on the other, are available in Playford (1969) and Krebs and Mountjoy (1972), respectively.

The Barlow Inlet Platform is similar in thickness, and approximate duration to many Devonian reef complexes (Table 4-12). It is probably comparable in areal extent to Devonian barrier reef complexes (e.g. Canning Basin), but considerably larger than the classic central European and western Canadian atoll reef complexes. The predominance (volumetrically) of lagoon and tidal flat facies over rocks of the reef proper is similar in both. Common to the 'back-reef' facies of the Barlow Inlet Platform and Devonian reef complexes are fenestral lime mudstone, cryptalgal laminite, amphiporid rudstone, megalodont floatstone,

Comparison of the Barlow Inlet Platform and
Devonian reef complexes

EXAMPLE	REEF COMPLEX	PRESENT LENGTH	PRESENT WIDTH	APPROX. DURATION ¹ AREA (SQ.KM)	THICKNESS ²
Barlow Inlet Platform	barrier	>50km	>50km	>2500	2.5m.y. 700m
Redwater Reef (Alta.)	atoll	30km	25km	580	<1.0m.y. 320m
Canning Basin (Australia)	barrier	306km	60km	18000	2.5m.y. 1000m
Attendorn Reef (Germany)	atoll	10-16km	1.5-5.5km	70	1.5m.y. 950m

1. figures given are approximate and are based on stage or series durations available in Ziegler (1978) and McKerrow et al. (1980).
2. thickness of Redwater Reef includes the Cooking Lake Formation.

massive (domical to bulbous) stromatoporoid sparse to dense biostromes, a paucity of stenohaline organisms (crinoids and brachiopods), an absence of back-reef evaporites (with some exceptions), and the common cyclical arrangement of beds.

Within the organic reef facies, the predominance of massive domical to tabular stromatoporoids in sparse to frame structures, and the possible presence of an enigmatic massive and/or dolomitized facies, are commonly held to be features of both late Silurian and Devonian shelf margins.

Solenoporids, crinoid debris, brachiopods, fasciculate rugosans and sponges characterize some but not all Devonian foreslope environments. The same organic elements populate the Barlow Inlet foreslope.

There are, of course, dissimilarities as well (Table 4-13), but these are appreciably outweighed by the characteristics held in common.

A composite ancient analog (FIG. 4-12) best serves to portray the biocumule facies distribution of the Barlow Inlet Platform. The platform interior magnafacies closely parallels Mid- to Upper Devonian backreef sequences, as discussed above. The medial foreslope, characterized by lithistid-sponge dominated mud mounds, has no precise counterpart in other described Siluro-Devonian sequences. The settings for either or both Somerset Island sponge mud mounds (Narbonne, 1981) and N.W. Greenland mudstone mounds (Hurst, 1980) may be comparable. Broadly similar environmental conditions may have prevailed during

Similarities and dissimilarities between Silurian and Devonian reef and reef-associated facies.

ABSOLUTE DIFFERENCES :

- 1) The problematic cyanophytes Renalcis, Epiphyton, and the proto-coralline alga (or possibly stromatoporoid) Keega, important frame encrusters and builders in the Devonian, are apparently not present in the Silurian.
- 2) Halysitid corals, important Silurian sediment bafflers, and mucophyllid corals, are not found in Devonian biocumules.
- 3) Calcispheres, particularly abundant in the Devonian back reef facies, are not reported from Silurian rocks.

QUANTITATIVE DIFFERENCES :

- 1) Crinoid-rich flank beds are less well-developed in the Devonian.
- 2) Lithistid sponges and bryozoans are more important reef organisms in the Silurian.
- 3) Alveolitids and cerioid rugosans are more predominant in Devonian buildups.
- 4) Amphiporids may have had a wider environmental distribution in the Silurian.

PERPETUATED MISCONCEPTIONS :

- 1) No amphiporids existed in the Silurian and early Devonian (St. Jean, 1971)
- 2) No large fasciculate rugose corals existed in the Silurian (Tsien, 1977).

SIMILARITIES :

	<u>SILURIAN</u>		<u>DEVONIAN</u>		
	<u>M</u>	<u>U</u>	<u>L</u>	<u>M</u>	<u>U</u>
1) Stromatactis	x	x	-	x	x
2) Massive stromatoporoids predominate in reef crest zone	x	x	x	x	x
3) Back-reef megalodonts	-	x	-	-	x
4) Back-reef amphiporids	-	x	-	x	x
5) Corals predominate in forereef or pioneer reef communities	x	x	x	x	x
6) Solenoporids in reef and forereef	x	x	-	x	x
7) Sponges present in forereef or pioneer reef communities	x	x	-	-	x
8) Large fasciculate rugosans in forereef or protoreef	-	x	x	x	?
9) Massive stromatoporoid-dominated biostromes in proximal back-reef	-	x	-	x	x

x reported in the literature
 - not presently reported.

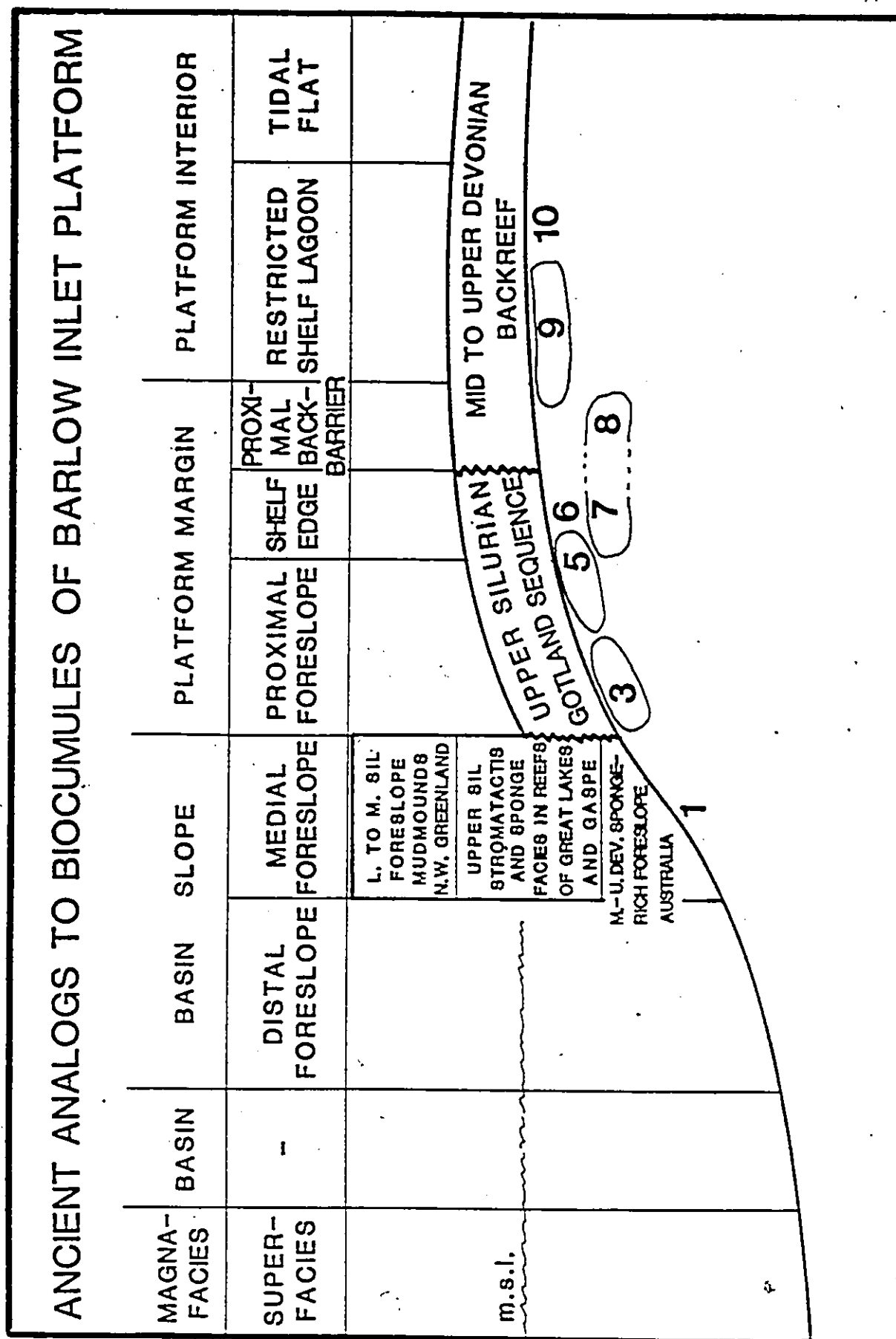


FIG. 4-12 Relationship of Barlow Inlet Platform to other studied "reefold" successions. Numbers refer to biocumule facies listed at the beginning of this paper.

deposition of the stromatactoid mudstone facies of the Great Lakes and the Gaspé, and on the sponge-rich portions of the Canning Basin foreslope. The platform margin magnafacies exhibits striking similarities to the shallowing-upward reef spectrum (Axelsro to Kuppen type) described by Riding (1981) on Gotland (FIG. 4-11).

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4.4 SYNTHESIS

The construction of a platform model that accurately portrays the disposition of the biocumule facies described herein, is made difficult by their discontinuous distribution, both areally and stratigraphically. For example, the diverse community mud mounds are restricted to the basal 31 m (possibly 60 m) of the formation, and are not found further north than loc. 6. Even within this distribution range they have not been identified at loc. 4. Similarly, the lithistid sponge-dominated biocumules were recognized in medial foreslope facies at loc. 8 and loc. 13, but not in intervening environmentally equivalent strata. The lamellar stromatoporoid biostrome facies appears to be singularly well-developed at loc. 5. Mucophyllid mud mounds are known only at loc. 18. It is thus apparent that the platform margin was not uniformly constructed along its breadth and length. Furthermore there is evidence to suggest that the nature of the platform margin evolved with time and is represented by somewhat differing biocumule facies stratigraphically. The origin of this variance cannot be proven given the presently available biostratigraphic data, although it is reasonable to speculate that the following factors contributed to a heterogeneous platform margin:²⁴

²⁴ such a condition is only to be expected in nature given the multiplicity of variables influencing marine biofacies distributions; Holocene reef tracts all show some degree of lateral variability.

1. the relative position of any portion of the shelf with respect to prevailing winds and principal storm tracks.
2. antecedent seafloor topography.
3. synsedimentary tectonism (e.g. position of growth faults).
4. nutrient supply (position of shelf relative to upwelling currents).
5. influx of toxic materials (terrigenous detritus, fresh or hypersaline waters).

All five variables can be spatially or temporally restricted. Unfortunately discriminating temporal from areal significance is not possible, particularly in light of the poor age control available for the Barlow Inlet Formation.

Therefore the models presented here may depict contemporaneity of juxtaposed facies, that did not in fact exist and conversely biocumules illustrated on separate models may have closer genetic relationships than suggested.

Two lines of evidence are used to reconstruct the platform margin. The most reliable is the biofacies successions revealed in the zoned biocumules. Examples A and B clearly reflect shallowing-upward sequences and can be viewed as transects through the platform margin. The other line of evidence is based on the environmental interpretation of each biocumule, and the tendency for biocumules to be spatially and stratigraphically associated.

Two critical problems must be addressed in attempting to construct a reasonable platform margin model.

1. Several biocumule facies have considerable concentrations of massive to lamellar stromatoporoids, and all are interpreted to occur at or near the shelf edge. What is the relationship between stromatoporoid-crinoid rubble skeletal buildups, dolomitized massive skeletal-(?crinoid) (DMS) biocumules, stromatoporoid dominated biostromes, and lamellar stromatoporoid biostromes?
2. Volumetrically significant, stratiform crinoidal deposits (grainstones, rudstones, packstones) occur throughout the Barlow Inlet Formation. What environments do these blanket deposits represent, and where did the pelmatozoan debris originate?

It has been shown that the DMS biocumules have characteristics in common with the Restricted Infratidal hemicycle successions of the platform interior (Table 4-4). They differ, however, in that at least some DMS biocumules demonstrably achieved stratigraphic relief (FIG. 4-8), they proportionately contain much more crinoidal debris and framework structure, and contain more and larger low domical to lamellar stromatoporoids (in contrast to the hemispherical and subspherical morphotypes of the back-reef biostromes). The stromatoporoid-crinoid rubble skeletal buildups are related but incorporate still more crinoid

material, are considerably thinner and more obviously mound-shaped, and are not visibly divisible into strata.

FIG. 4-13 graphically portrays the envisioned relationship between three of the biocumules. What are considered DMS biocumules are actually dolomitized platform margin sediments that contain both Kuppen-type barrier biostromes at the front and intensely dolomitized back-reef sediments (stromatoporoid-dominated biostromes and other facies of the Restricted Infratidal hemicyclic sequence). The stromatoporoid-crinoid rubble skeletal buildups occur as isolated mounds on the shallow proximal foreslope. The lamellar stromatoporoid biostromes appear to represent a special case, but the conditions requisite for their development are not known. Their discontinuous distribution suggests that they grew in particular isolated zones of higher current activity or substrate mobility (e.g. shelf edge channels). The Holmhällar reefs, which are similar, are considered to be high(er) energy linear reefs.

The crinoidal grainstone to rudstone lithofacies can give a misleading impression of the energy of the environment. Much apparent grain support seems to have been secondarily achieved by pressure solution. In addition, much of the intra-allochem cement is inclusion-rich pseudospar with abundant recrystallized peloids. Many of the crinoidal limestones of the Barlow Inlet Formation were probably

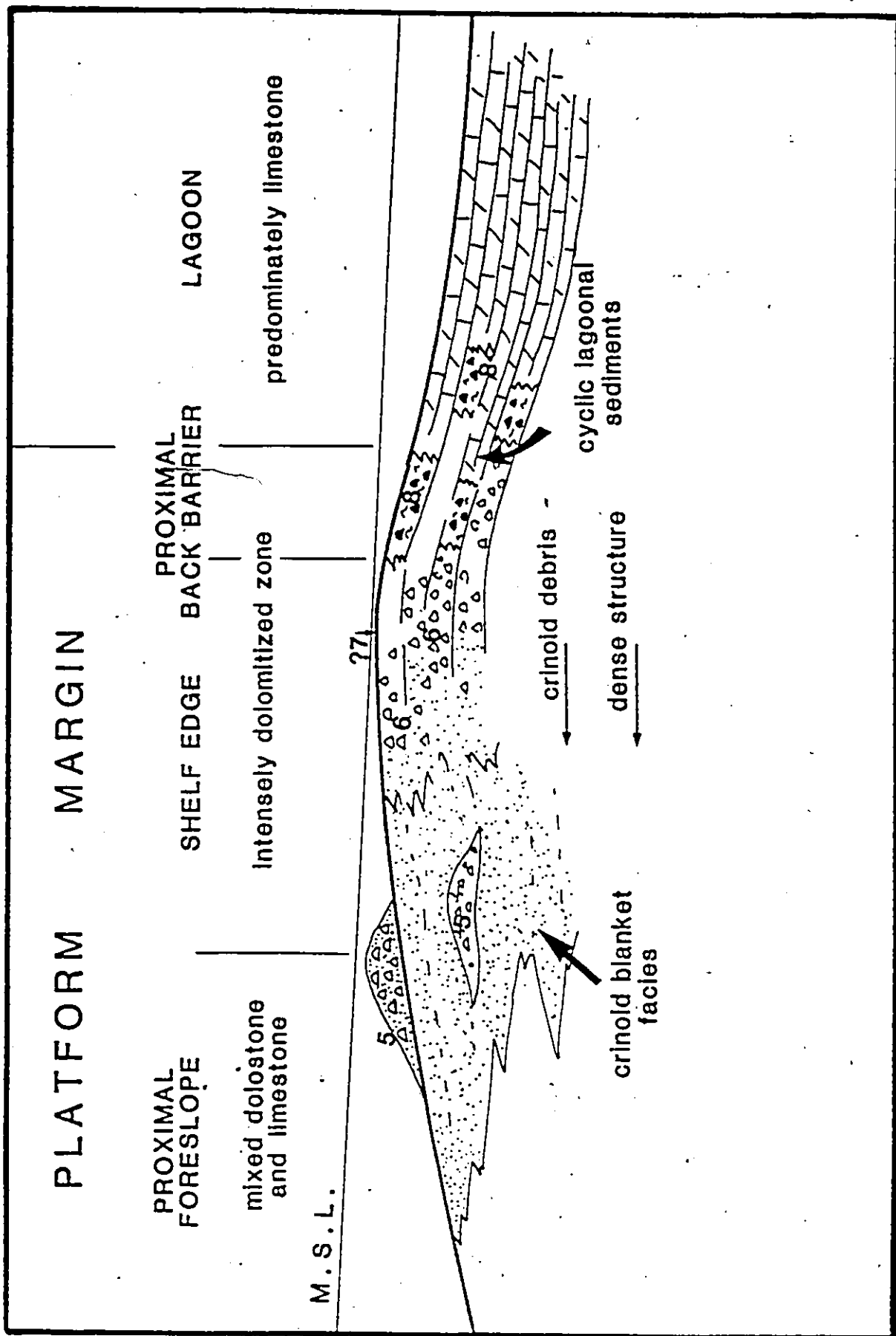


FIG. 4-13 Relationship of biocumules of the Platform Margin.

originally bioclast-rich crinoidal wackestone to packstone with a peloidal muddy to coarse silty matrix.

The grainstones are for the most part moderately to well-sorted, but this too can be deceiving. The pelma of crinoids tend to disarticulate spontaneously, and the resultant product is well-sorted for inherently biological reasons, not hydrodynamic ones. Recognizeable cross-beds are rare within the crinoidal blanket facies, and their occurrence seems to be restricted to particularly well-sorted and well-washed medium-grained crinoidal sands. Fortuitous weathering may highlight cross-sets (see PL. 5-5, Paper 5, this thesis), but normally such features remain obscure (see also Rhurmann, 1971). The presence of cross-stratification in otherwise massive crinoidal grainstone, may be indicated by conchoidal fracturing of these rocks. No coated grains have been recorded from either the crinoidal blanket deposits, or indeed any other facies of the Barlow Inlet Formation.

The lack of pervasive bed-load transport structures remains enigmatic. Certainly the abundance of abraded fragments and clasts with complete micrite envelopes, attests to grain interaction and mobility. The low density of the crinoid endoskeleton, coupled with the cylindrical to disk shapes of its constituent elements, may have promoted saltation rather than traction carpet transport.

In conclusion, the environmental significance of the blanket deposits is ambiguous and it is probable that these sedimentary bodies were developed under a range of hydraulic settings. At one extreme the deposits were only periodically disturbed by currents (at or just below wave base), and at the other extreme, the deposits may represent mobile, well-washed shoal sandstone, deposited in continuously agitated water.

The origin of the crinoidal bioclasts is not immediately clear. Significantly, holdfasts have not been identified within crinoidal packstone/rudstone lithotypes (intra or extra-reef), nor with lamellar stromatoporoid biostromes or DSM biocumules, all facies occupying the platform margin. Possibly this is a problem of perception (white weathering holdfasts are conspicuous against the grey-green background of enclosing lime mudstone) but evidently the upper Ludlovian Barlow Inlet Platform margin crinoids (which judging from pluricolumnal morphologies were restricted to few species), preferred to root themselves in muddy substrates. This is strongly suggested at loc. 6, where discrete pods of lime wackestone/floatstone occur randomly scattered throughout a predominantly crinoidal grainstone/packstone lithofacies (PL. 4-14i,j). Some margins of these pods are quite abrupt (although this sharpness may be exaggerated by solution), and this lends initial credence to interpreting the pods as exotic blocks.

However, careful examination shows that the vast majority of contacts are gradational. Sponges, rhizoidal holdfasts, halysitid corals, and solitary rugosans constitute the bulk of the in-situ organic constituents of these pods, which can best be described as 'incipient' diverse community mud mounds. A developmental model for these pods (and applicable to the diverse community mud mounds) is presented below.

Shallow swales on the predominantly sandy seafloor may have preferentially accumulated carbonate mud (perhaps due to the influence of a pioneer, sand-encrusting community of mud-baffling organisms, such as sponges). Once the patches of muddy substrate were colonized by other bafflers and binders (rhizoidal holdfasts and halysitids), the sites became loci for increased lime mud deposition. Some of the lime mud may not have been suspended sediment, but rather micrite cement precipitated in response to skeletal resorption by crinoids (see Paper 5). Due to the gregarious nature of the echinoderms, extremely large volumes of crinoidal debris were generated; volumes were sufficient for pelmatozoan bioclasts to predominate overwhelmingly in surrounding sediments, and perhaps sufficient to bury local crinoid nests.

The source for the crinoid debris in the blanket deposits is summarized in FIG. 4-14. Some crinoid material was probably derived from biocumules of the

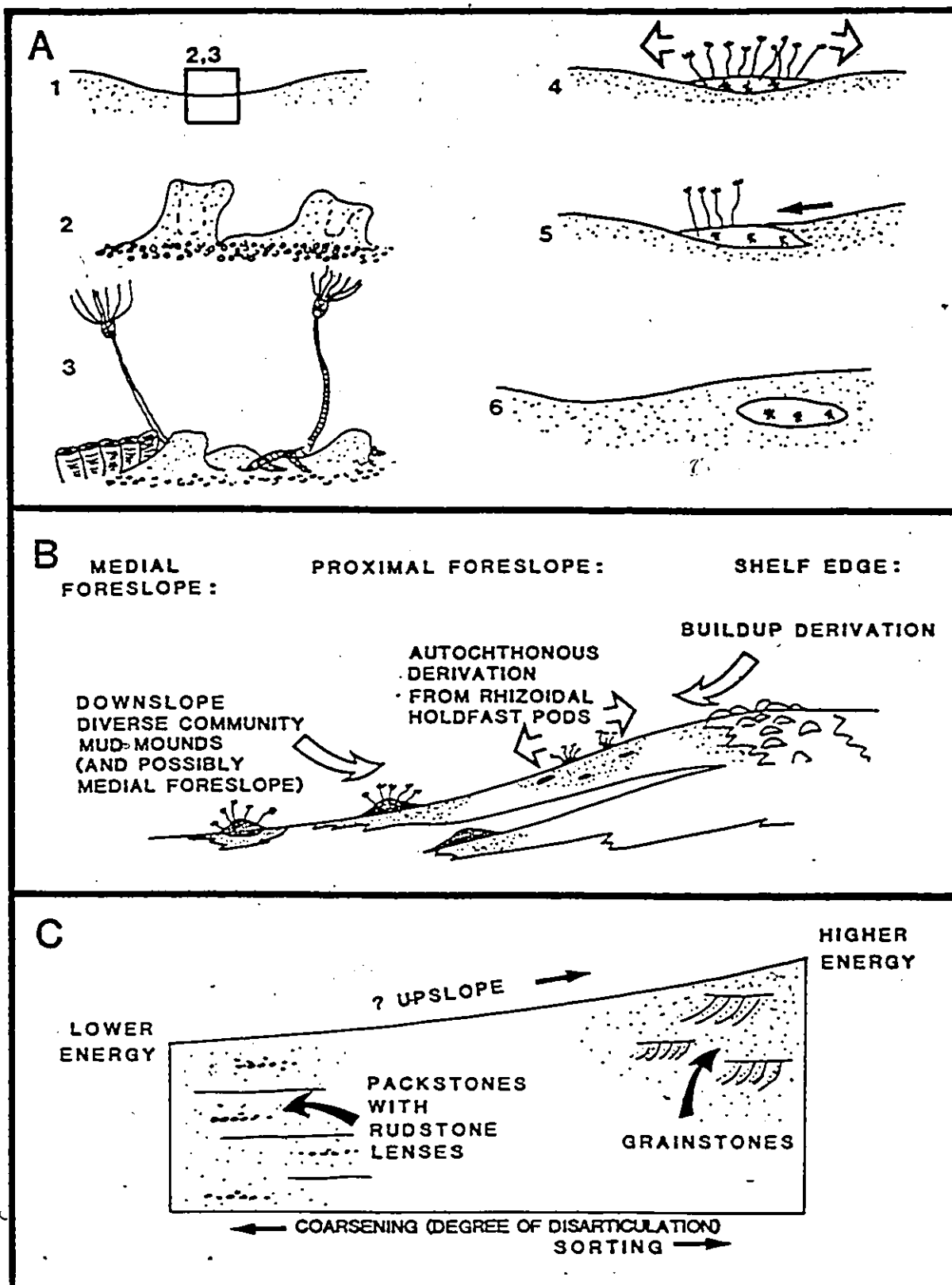


FIG. 4-14 Rhizoidal holdfast pods and the source of crinoid debris.
 A. Postulated development of rhizoidal pods: (1) a swale (hydrodynamically created?) is colonized by (2) a pioneer community (perhaps sponges). Lime mud is baffled or precipitated and a more diverse community predominated by crinoids and catenoporoids flourishes, contributing substantial volumes of crinoid debris to the surrounding area (4). Local pod is swamped by encroaching body of mobile sediment (5) and buried (6).
 B. Source of crinoid bioclasts for crinoid blanket facies.
 C. Spectrum of crinoidal limestones of crinoid blanket facies.

platform margin; the presence of crinoidal bioclastic halos around the reefs appears to require this. Other crinoidal debris may have been derived from the medial and distal foreslope, but this conjecture remains unproven. A deeper water origin for encrinite shoal bioclasts has been vigorously argued for by Rhurmann (1971), who studied Lower Devonian crinoidal deposits in Spain. Probably the crinoid bioclasts were derived from local concentrations of crinozoans, now represented by rhizoidal bindstone pods within the blanket deposits.

The platform margin profile is presented for convenience as three block diagrams (FIGS. 4-15,16,17) which portray the following:

1. the preferred stratigraphic and spatial associations of the biocumule facies,
2. the heterogeneity of the platform margin (continuous organic rim in some places; crinoid shoals in others), and
3. a possible evolutionary sequence (15 to 17) reflecting increased maturity of the platform margin.

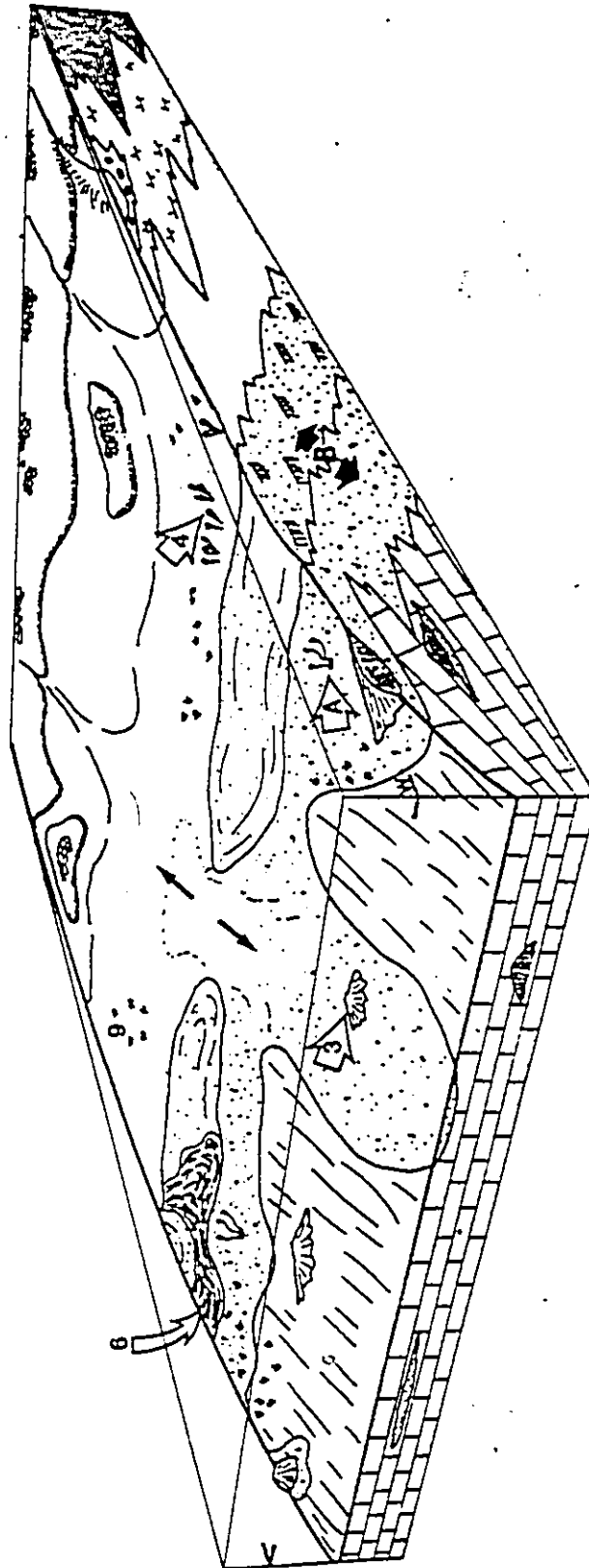


FIG. 4-15 Block diagram of biocumule facies associations. Numbers refer to biocumule facies listed at beginning of paper. Other symbols are as follows: W - wave base, A - rhizoidal holdfast pods situated in, B - crinoid blanket facies.

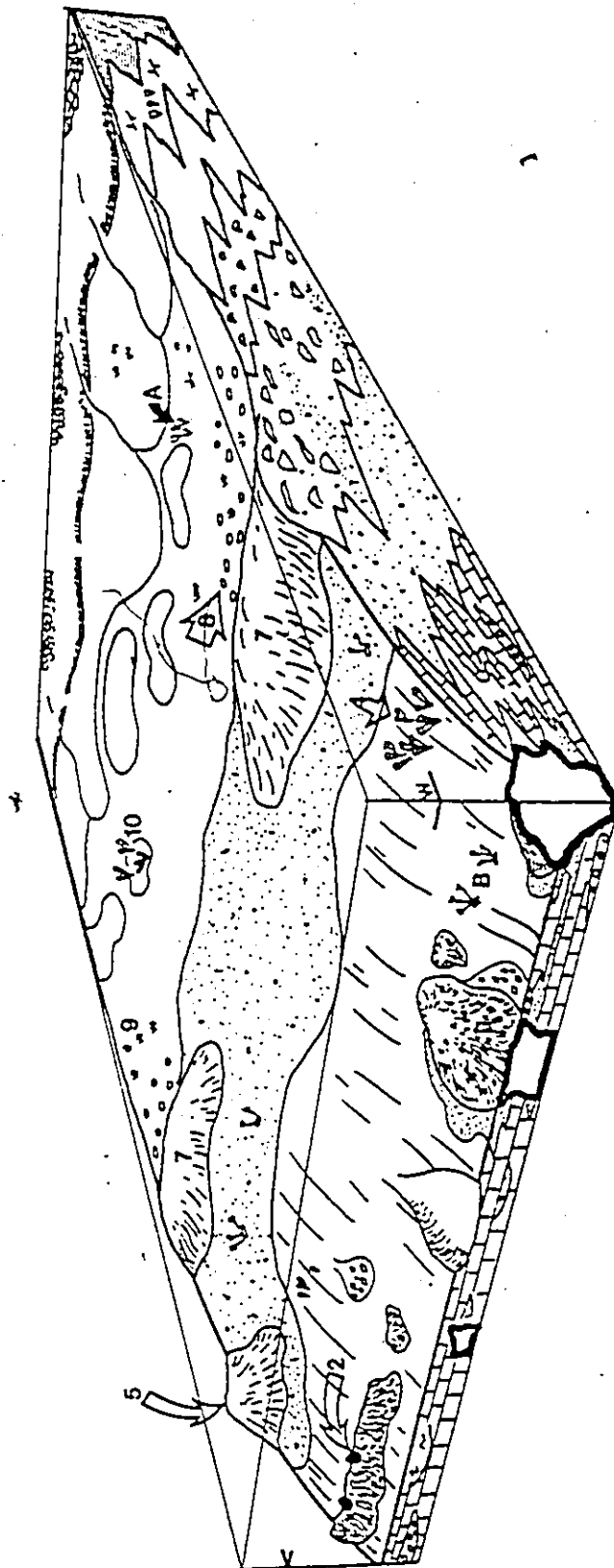


FIG. 4-16 Block diagram of biocumule associations. Numbers refer to biocumule facies listed at beginning of paper. Other symbols are as follows: w - wave base, A - concentrations of megalodont bivalves, B - sediment stabilization by crinoids inhabiting the medial foreslope.

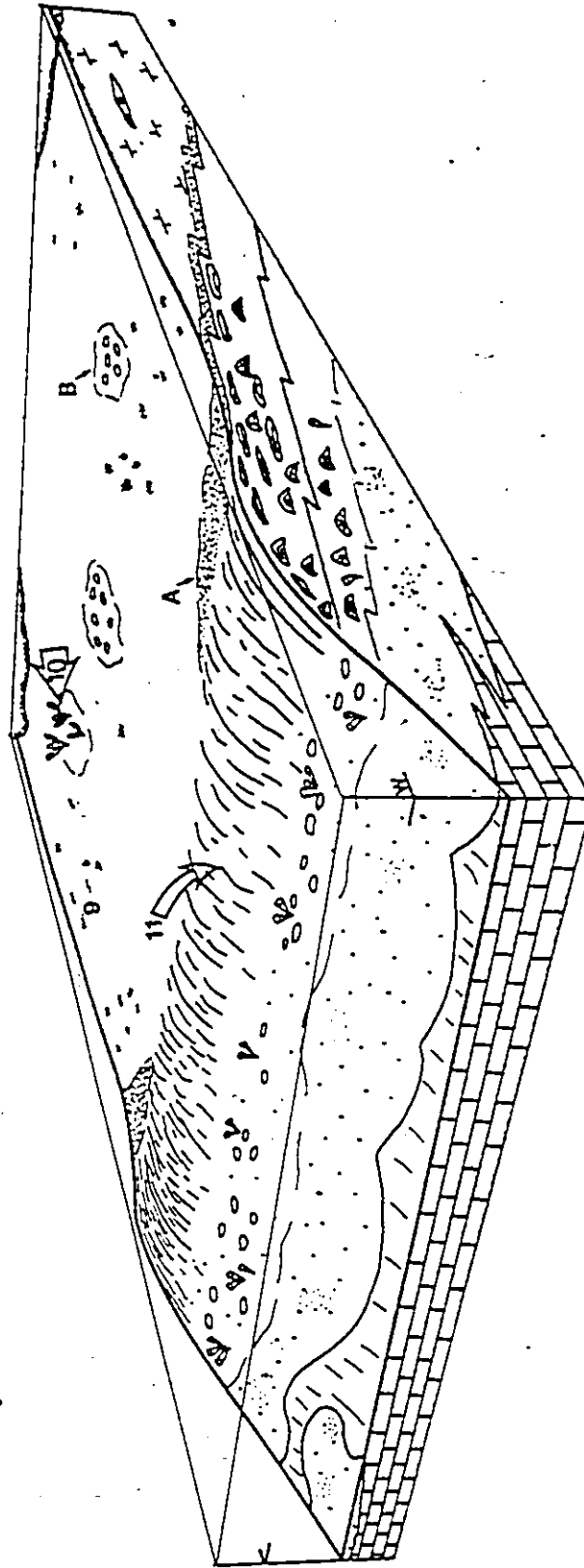


FIG. 4-17 Block diagram of biocumule associations. Numbers refer to biocumule facies listed at beginning of paper. Other symbols are as follows: W - wave base, A - amphiporoid packstone in immediate back-reef position, B - small massive stromatoporoid mud mounds in lagoonal setting. Zoned biocumule is a barrier reef similar to example A discussed in text.

4.5 CONCLUSIONS

1. the upper member of the Barlow Inlet Formation contains numerous different biostromes and buildups, here collectively considered under a newly-defined term, biocumule.
2. ten biocumule facies are recognized, including lithistid sponge-dominated mud mounds, mucophyllid mud mounds, diverse community mud mounds, fasciculate rugose coral biostromes, stromatoporoid-crinoid rubble skeletal buildups, dolomitized massive skeletal (?crinoid) biocumules, lamellar stromatoporoid biostromes, massive stromatoporoid biostromes, dendroid stromatoporoid biostromes, and Coenites -capped fenestral mud mounds. Several biocumules, not included in the above list, show a zonation of biofacies and lithotypes.
3. the biocumules occupied many different parts of the platform complex, ranging from medial foreslope to restricted shelf lagoon (FIG. 4-12).
4. the spectrum of biocumule facies of the Barlow Inlet Formation is closely comparable to known reef complexes of Mid- to Late Devonian age. It follows that the constructional paradigm for biocumules did not alter significantly from Late Silurian to Late Devonian time. The important biotic constructors, their abundances and roles, and the structure and

successional position of the reef communities they formed, remained relatively unchanged for the 75 m.y. prior to the close of the Devonian Period.

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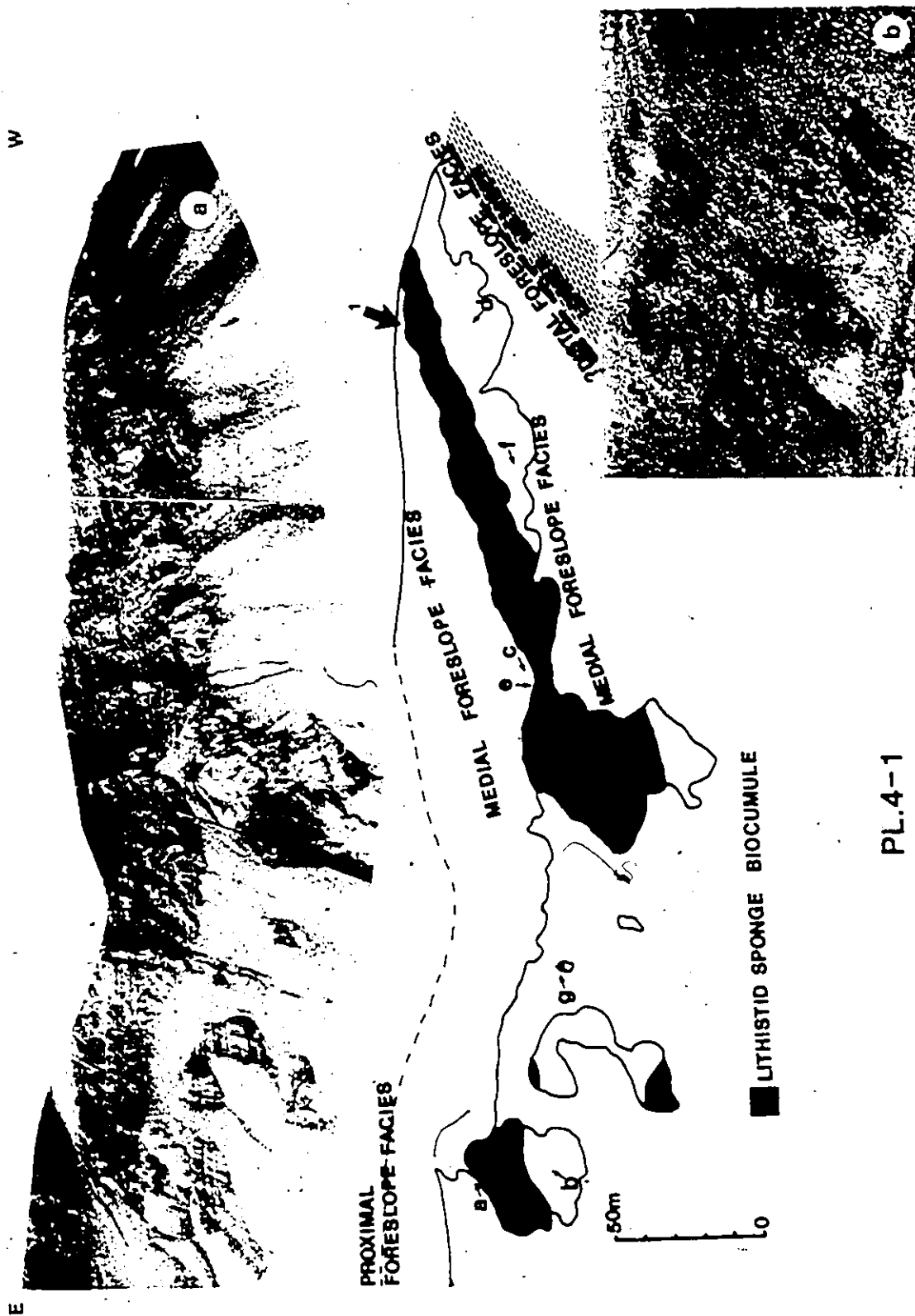
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4.7 PLATES

Plate 4-1

Lithistid sponge-dominated mud mounds - 1

- a View looking south at loc. 8 (Depot Point), with interpretive sketch beneath. Small letters refer to photographs figured on PL4-3. Uppermost part of biocumule in the west (1), has a greater abundance of massive skeletal organisms.
- b Three lithistid sponge dominated mud mounds at loc. 13 (Cape Rescue). Geologist at top, left of centre, for scale.

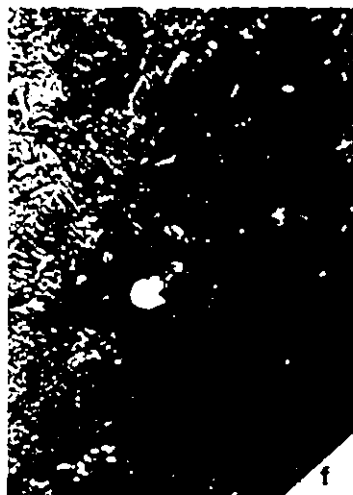
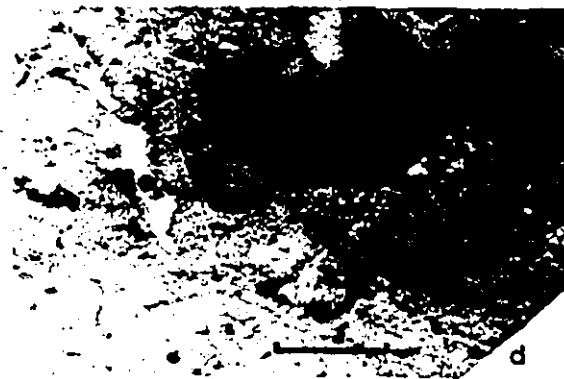


PL.4-1

Plate 4-2

Lithistid sponge-dominated mud mounds - 2
Bar scales are 1 cm.

- a Cement generations in a solution cavity. An isopachous rim of medium, equant, Fe-poor dolospar (1) is in direct contact with matrix. Coarse equant Fe-rich dolospar (2) exhibits gradually-increasing Fe content towards the centre of the cavity (a --> b). Final occlusion of cavity space is achieved by precipitation of coarse, blocky, clear, Fe-poor calcite (3). (polished slab)
- b Mottles (arrow) within spicular mudstone facies. (polished slab)
- c Burrows with pyritized margins in spicular mudstone.
- d Lithistid sponges with irregular growth form forming a framework. Note geopetal sediment in stromatactoid structures (arrow). (polished slab)
- e Rhizoidal holdfasts (RH) intergrown with favositid corals (F) forming local bindstone pockets in mud mound. (acetate peel).
- f Partially dissolved spicular network of irregular lithistid sponges. (acetate peel).
- g Intergrowth of spongiostromate algae (A) and lithistid sponge from loc. 18 (polished slab).
- h Partially-brecciated bed encrusted by spongiostromate algae (A), and in turn overgrown by lithistid sponges and calcareous algae (L).
- i Hemi-pelagic lime mudstones (1), overlain by polymictic carbonate breccia (2), overlain by sponge mud mounds (3).
- j Clasts in breccia include blocks of spicular mudstone (1), blocks of crinoidal packstone (2), and disoriented colonial metazoan fossils. (3).



PL.4-2

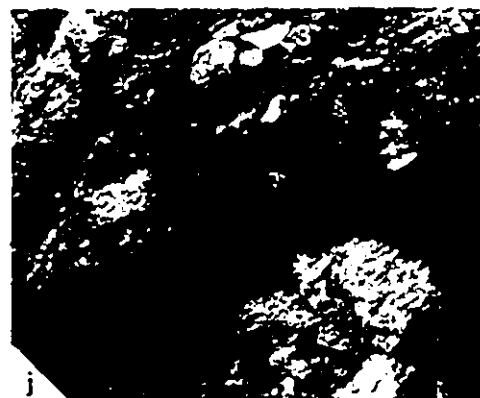


Plate 4-3

Lithistid sponge-dominated mud mounds and
mucophyllid mud mounds.

Figures a to g at loc. 8 (see Plate 4-1)

- a Steeply-dipping beds of lime mudstone adjacent to well-defined margin of mud mound. Hammer 31 cm long.
- b Rapid pinchout of crinoid debris beds (2) on either side of holdfast-rich portion of mudmound (1). Jean Dugal for scale.
- c Strata on top of (and adjacent to) sponge-dominated mud-mound. More resistant wedge-shaped beds are crinoidal. Arrow points to carbonate debris flow shown in figure e. Scale 200 cm.
- d Gutter casts on the sole of a lithoclast-rich crinoidal packstone bed. Hammer 31 cm long.
- e Carbonate debris flow with rafted (and plastically-deformed) lime mudstone beds (1), disoriented fossils of favostids (2), and blocks of biocumule mudstone (3). Hammer 31 cm long.
- f Slump folds, a common feature in flanking strata. Hammer 31 cm long.
- g Isolated stack featuring sub-biocumule beds of lime mudstone, wackestone and floatstone (1), overlain by mud mound mudstone (2), and unconformably overlain by crinoidal packstone (3). Note high angle between crinoidal flank deposits and subjacent bedded flank deposits and subjacent bedded limestone.
- h Occurrence of three mucophyllid mud mounds (arrows) astride a lithistid sponge dominated mud mound. Note that the small, satellite mound farthest to the left is partially enclosed by the parent structure. Locality 18.
- i Concentration of mucophyllids on mound surfaces. Note uniform orientation with apices pointed towards the viewer. Bar scale 1 cm.

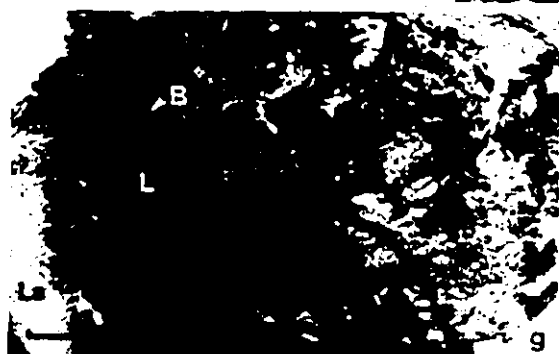
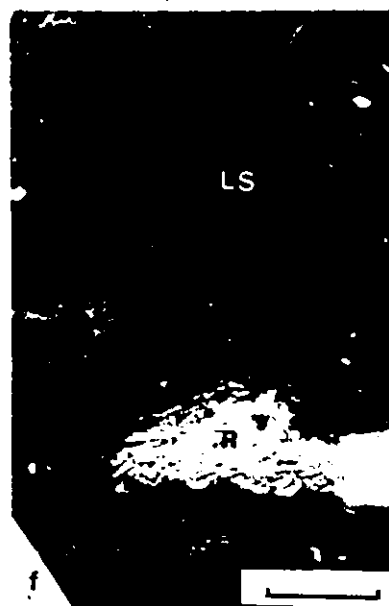
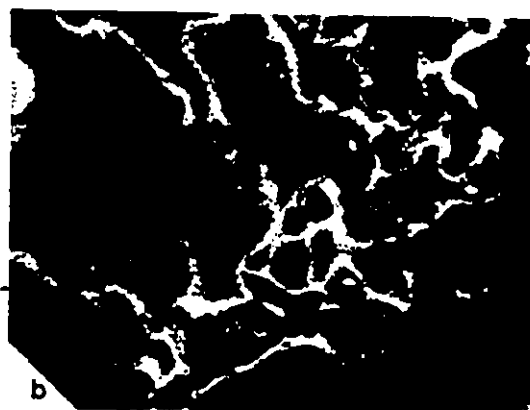


PL4-3

Plate 4-4

Diverse community mud mounds -1
Bar scales are 1 cm.

- a Diverse community mud mound composed of saccoliths, loc. 6. Boundary of biocumule outlined by string. Hammer 31 cm long.
- b Alveolitid coral bindstone (polished slab).
- c Proporid (P) and syringoporid (Sy) corals (polished slab).
- d Longitudinal section through syringoporid coral colony (acetate peel).
- e Cateniporid corals (C) (polished slab).
- f Rugose corals (R) and lamellar stromatoporoids (LS). (polished slab).
- g Lithistid sponge (L), encrusting bryozoans (B), and rugose corals (R). (polished slab).
- h Small, subspherical favositid corals (F) and dendroid tabulate corals (D). (polished slab).
- i Dendroid solenoporid alga (field photo).
- j Bioclasts of solenoporid algae (dark) and crinoid ossicles in flanking strata of diverse community mud mound. (polished slab).



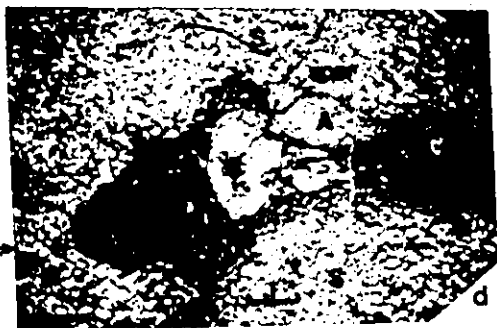
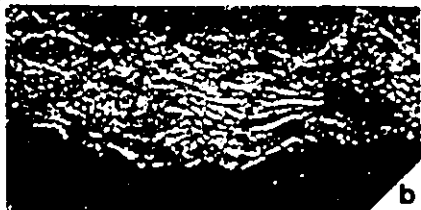
PL4-4



Plate 4-5

Diverse community mud mounds - 2
 Bar scales are 1 cm (unlabelled) or tenths or
 hundredths of 1 cm (labelled).

- a Diverse community mud mounds resting directly on thin-bedded lime mudstone and wackestone at loc. . Scale 15 cm.
- b Girvanella sp., a common component of diverse community mud mounds. Note that transverse lines visible in tubules are crystal boundaries, not divisions of the thallus.
- c Coarsening-upward sequence ranging from lime mudstone at the base (hammer head) to crinoidal rudstone and packstone at top. Diverse community mud mound rests directly on selvage of coarser pluricolumnals (arrow). Scale 15 cm.
- d Growth cavity (now largely occluded by sediment), with solenoporid (A) and encrusting stromatoporoid (S).
- e,f Solution? cavity filled by inclusion-rich, drusy spar (2), zoned, alternately Fe-rich/Fe-poor, coarse, blocky dolospar (3), coarse, clear, Fe-poor equant calcite (3). Cavity rimmed by micrite (1).
- g Geopetal sediment in ?growth cavity.
- h Sediment-filled cavities (1), surrounded by Fe/Mn encrusted cavity wall.
- i Rounded clast of Girvanella sp. Beds subjacent to diverse community mud mounds at loc. 6
- j Mn/Fe crust (1) coating upper part of mound at loc. 6. Highest level at which crinoidal sediments accumulated indicated by arrow (2). Biocumule entombed by argillaceous lime mudstone.

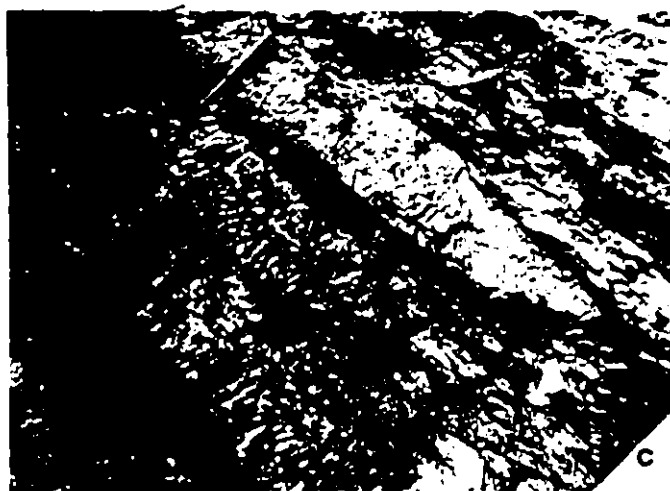


PL.4-5

Plate 4-6

Fasciculate rugose coral biostromes.
Bar scales are 1 cm.

- a Concentration of stylopleurids (s) and stolon structure (st) from local zone of rugose coral enrichment at top of lithistid sponge mud mound at loc. 8. (polished slab)
- b Stylopleurids(s) and stolon structure(st) from local zone of rugose coral enrichment at top of lithistid sponge mud mound at loc. 8 (polished slab).
- c Allochthonous block of biostromal material at loc. 7. Note growth orientation of stylopleurids (arrows) is directed down stratigraphically (i.e. block overturned). Hammer 31 cm long.
- d Stylopleurids (S) in mudstone matrix from same locality as example in figure. (polished slab).
- e Biostrome of densely-packed Stereoxyloides sp. Biostrome extends at least 20 m to the right. Scale 15 cm.
- f Transverse section of Stereoxyloides sp. (acetate peel).



PL.4-6

Plate 4-7

Stromatoporoid-Crinoid Rubble Skeletal Buildups - 1

- a Buildup at loc. 10. Arrows indicate locations of field photos on this plate. Note abrupt basal contact.
- b Intensely dolomitized core of buildup. Subjacent beds visible at lower left. Gary Graf for scale
- c Local relief of upper surface of buildup. Overlying strata are thin-bedded lime mudstone and calcareous shale, locally plastically-deformed (arrow).
- d Slumped beds of suprajacent strata. Hammer 31 cm long.

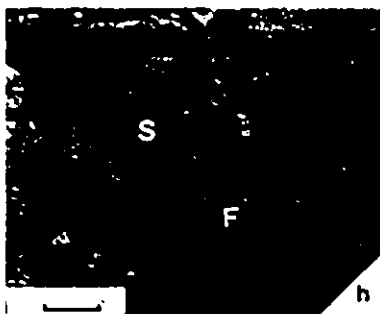
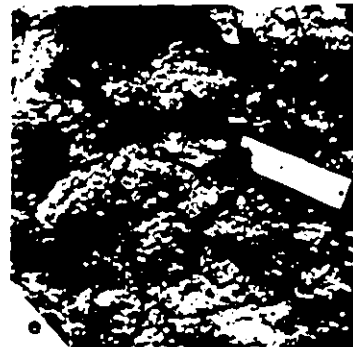
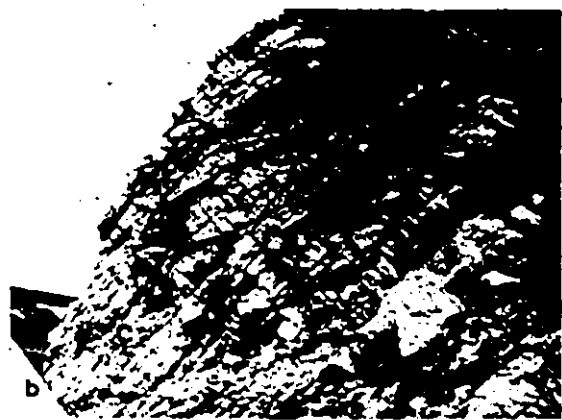


PL.4-7

Plate 4-8

Stromatoporoid-Crinoid Rubble Skeletal Buildup - 2

- a Dolomitized core (C) of buildup at loc. 9.
- b Dolomitized core (C) and suprajacent crinoidal packstone/grainstone beds of buildup at loc. 7. Jacob's staff is 1.5 m.
- c Two successive buildups at loc. 9. Arrow points to view in figure d.
- d Muddy basal facies (1) and crinoid-rich upper facies (2) of buildup at loc. 9
- e Rhizoidal holdfast-rich muddy basal facies in buildup at loc. 9. Scale 15 cm.
- f Stromatoporoid inertiastone with a crinoidal rudstone matrix.
- g Moldic porosity in dolomitized buildup core. Hammer 31 cm long.
- h Acetate peel of core facies featuring stromatoporoids (s), and favostid corals (F) in crinoidal packstone.
- i Assemblage of chaotically arranged and in part fragmented colonies of massive stromatoporoids.
- j Boundstone of favostid corals (F) with a crinoidal matrix.

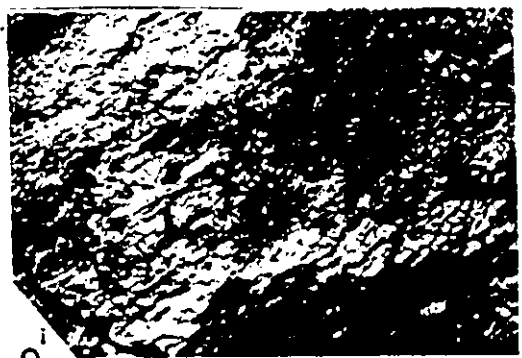
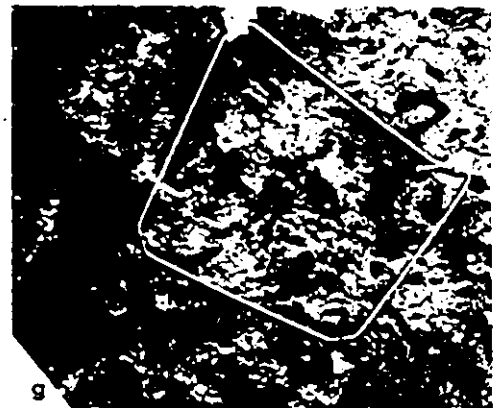
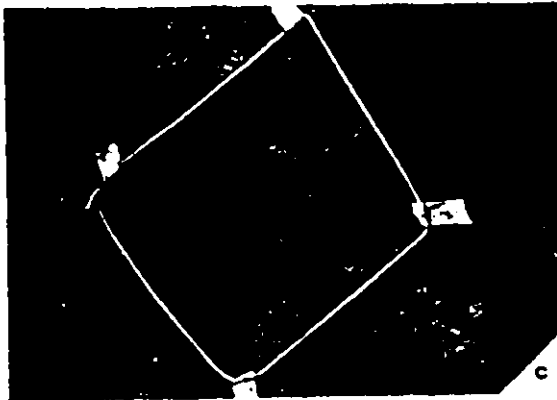
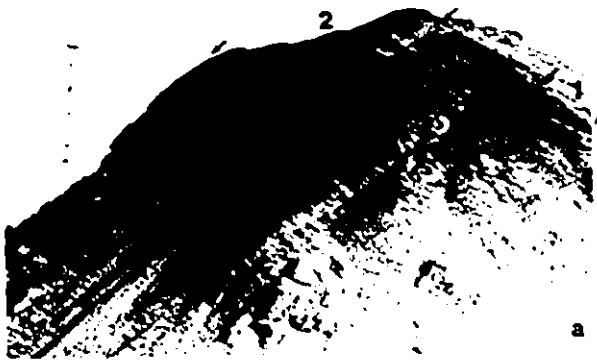


PL.4-8

Plate 4-9

Dolomitized massive skeletal (?crinoid) biocumules.

- a Two biocumules (1 & 2, between arrows) at loc. 6. Biocumule 2 is 70 m thick.
- b Dolomitized massive skeletal (?crinoid) biocumule at loc. 5. Note parallel nature of internal stratification. Interval is 43m thick.
- c Pervasively dolomitized crinoidal grainstone/packstone. Moldic porosity after crinoid pluricolumnals evident. Longest side of rectangle is approximately 20 cm.
- d Dolomitized hemispherical stromatoporoid with outlines of latilaminae preserved.
- e Contrast between ?fenestral dolomudstone(1) overlain abruptly by a stromatoporoid-rich coarsely crystalline dolostone (2). Hammer 31 cm long.
- f Molds of massive skeletal organisms in dolomitized biocumule at loc. 15. Hammer 31 cm long.
- g Stromatoporoid molds (irregular growth-form) in mottled medium crystalline dolostone. Longest side of rectangle is approximately 20 cm.
- h Transition upwards into dolomitized massive skeletal (?crinoid) biocumule (lower contact of biocumule 2, seen in figure a). Rubbly weathering, nodular beds (1) beneath biocumule (2) are partially dolomitized. Note contact zone runs at a high angle to bedding. Box enlarged in figure i.
- i Lateral transition from nodular-bedded dolomitic lime wackestone (3), to massive crystalline dolostone (1).

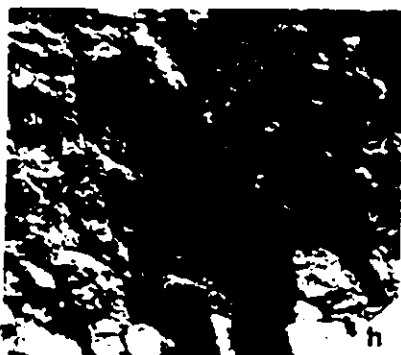
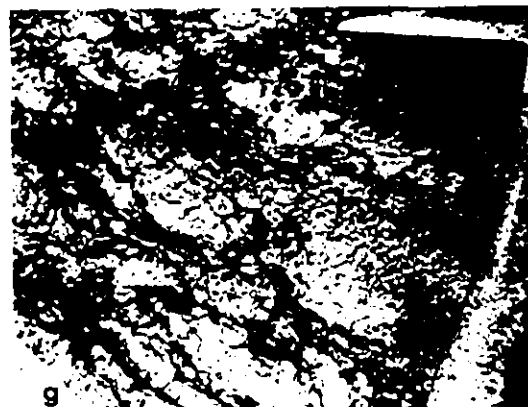
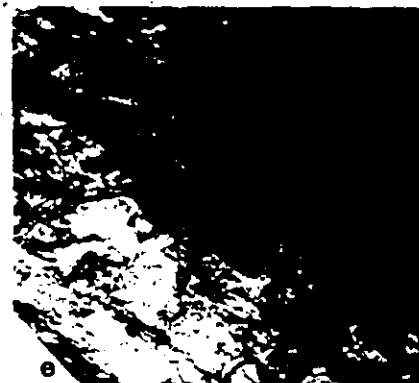
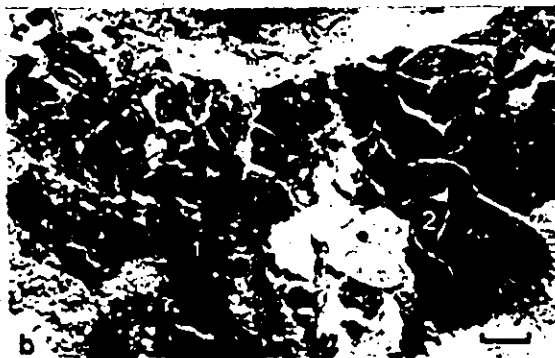


PL.4-9

Plate 4-10

Lamellar, massive, and dendroid stromatoporoid biostromes.
Bar scales are 1 cm.

- a Lamellar stromatoporoid.
- b Close association of robust dendroid (1) and lamellar (2) stromatoporoids.
- c Typical field appearance of lamellar stromatoporoid biostrome. Note that while the stromatoporoids have remained unaltered, the matrix is thoroughly dolomitized. Scale 15 cm.
- d Massive stromatoporoid-dominated biostrome of sparse structure. Note nodular bedded appearance of biostrome. Vertical field of view is 2 m. loc. 2.
- e Massive stromatoporoid-dominated biostrome of dense structure. Note some of the colonies are overturned. loc. 5.
- f Polished slab exhibiting hemispherical growth form of stromatoporoids and the partially dolomitized matrix.
- g Rubbly-weathering, mottled lithofacies with heterogeneous biofacies including delicate and robust stromatoporoids, crinoid pluricolumnals, and solitary rugose corals. Loc. 5.
- h Homogeneous biofacies of delicate dendroid stromatoporoids (amphiporids). Loc. 3.
- i Amphiporid floatstone (top direction to left), loc. 12. (polished slab).
- j Robust dendroid stromatoporoid (cf. Stachyodes), loc. 2 (polished slab).



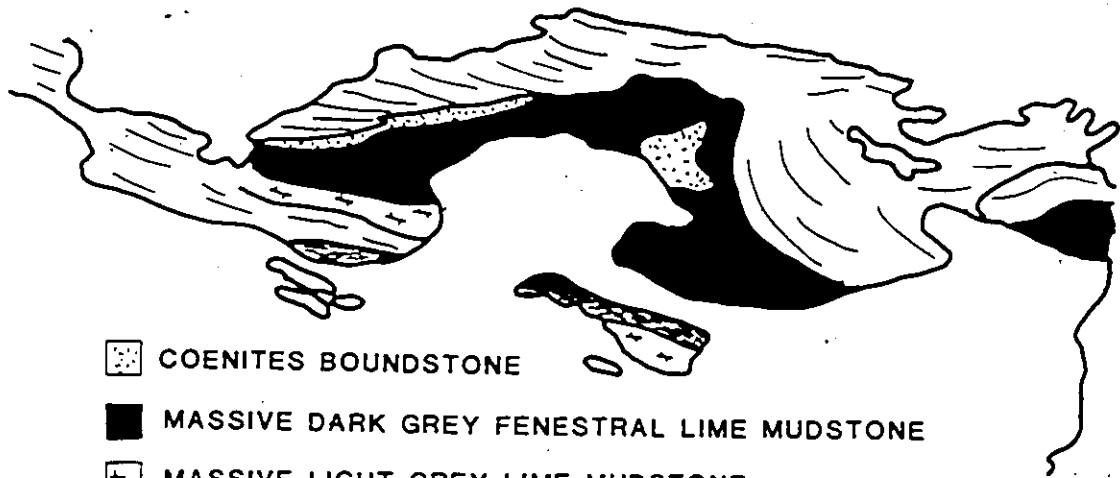
PL.4-10

Plate 4-11

Coenites-capped fenestral mud mound.

Aerial view of mud mound (a). Regional dip is shallow to the northeast (top right of picture). Lateral field of view is approximately 50 m. Cartoon below photograph depicts distribution of principal lithofacies. locality 11.





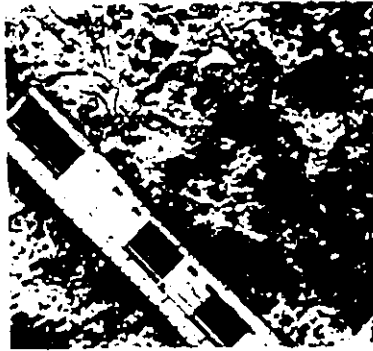
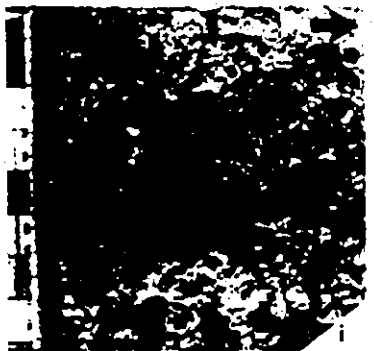
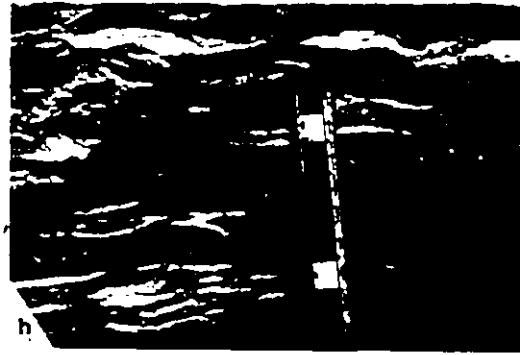
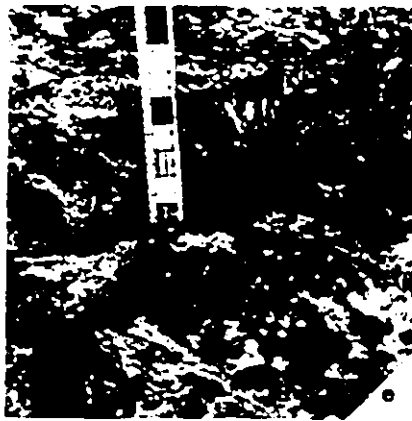
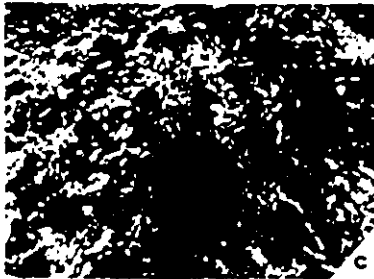
-  COENITES BOUNDSTONE
-  MASSIVE DARK GREY FENESTRAL LIME MUDSTONE
-  MASSIVE LIGHT GREY LIME MUDSTONE
-  BEDDED MOTTLED DOLOMITIC LIMESTONE
-  LAMINITE

PL.4-11

Plate 4-12

Coenites-capped fenestral mud mounds.

- a Dark grey fenestral lime mudstone of mound core. Flanking strata are in the foreground. Jean Dugal for scale.
- b Close-up of core pictured in figure a. Scale 15 cm.
- c Coenites-rich floatstone, capping core. Lens cap 6 cm across.
- d Acetate peel of capping facies featuring massive favositid corals (1), Coenites (2), and solitary rugose corals (3). Bar scale 1 cm.
- e Contact between fenestral lime mudstone and overlying Coenites bafflestone/floatstone. Scale 15 cm.
- f Acetate peel of the contact in figure e. Note irregular, stylolitized contact and fenestrae and disarticulated ostracode valves in the mudstone facies. Bar scale 1 cm.
- g Flank of mound featuring fenestral lime mudstone core (1), Coenites cap-rock (2), and mottled dolomitic flank beds (3). Hammer is 31 cm long.
- h Cryptalgal-laminated calci/dolosiltite. Scale 15 cm.
- i,j,k Single bedding surface of flank bed showing progressive decrease in bioclasts away (arrow) from the mound. Scale 15 cm.

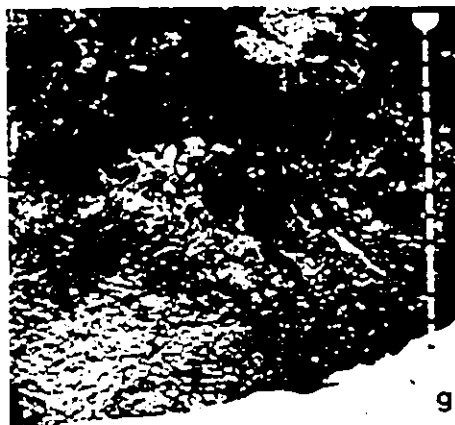
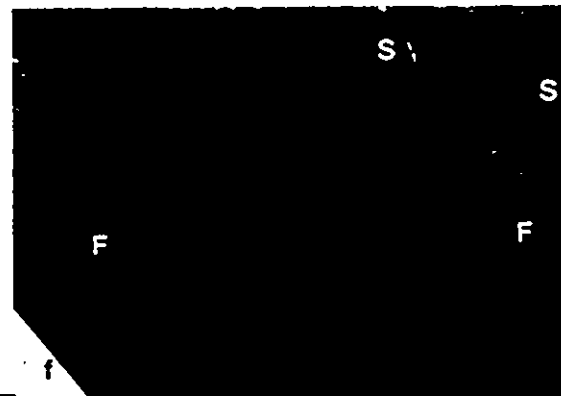
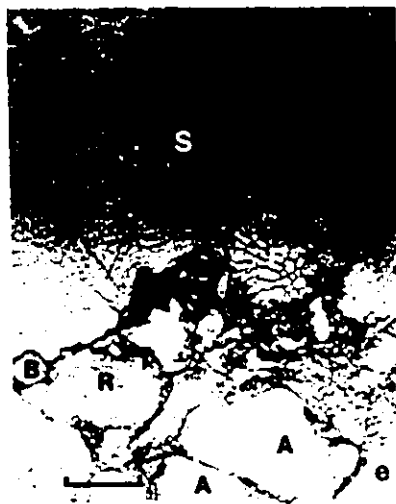


PL4-12

Plate 4-13

Zoned biocumules.
Bar scales are 1 cm.

- a Aerial view of barrier reef at loc. 12 (example A, zoned biocumules). Cumulative stratigraphic thickness of zones 3,4 and 5 is 11.1 m.
- b View of zones 2,3 and 4 at loc. 12. Note bedding becomes more pronounced towards base. Prone assistant for scale.
- c Tubular massive stromatoporoids (S) and low-domical favositid corals (F) in zone 4.
- d Favositids (F) and rugose corals (R) of zone 3.
- e Acetate peel of typical zone 4 facies, indicating stromatoporoid (S), bryozoan (B), solitary rugose coral (R) and solenoporiid algae (A).
- f Acetate peel of typical zone 3 facies featuring massive favositids (F), and subordinate encrusting stromatoporoids (S).
- g Massive lime mudstone of back-reef lagoonal facies (loc. 12). Jacob's staff with 0.1m intervals.
- h Amphiporid packstone of zone 5.



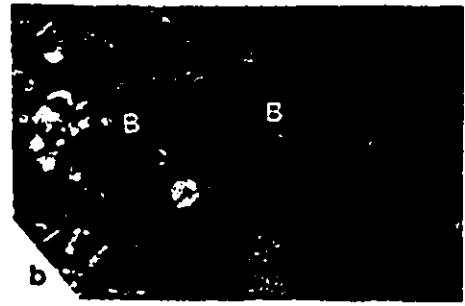
PL.4-12



Plate 4-14

Zoned biocumules and crinoid blanket deposits.
Bar scales are 1 cm.

- a Zoned biocumule (example D) at loc. 14. Lateral field of view is approximately 300 m. Featured are crinoidal flank beds (1), zoned mud mound (2) and back reef, restricted shelf facies (3).
- b Rich biota (in rudstone) of biocumule zone at loc. 14 immediately beneath allochthonous crinoidal beds. Favositid corals (F), calcareous algae (A), bryozoans (B), and stromatoporoids (S) are indicated.
- c Diffuse contact (arrow and dashed line) between allochthonous crinoidal facies (1) and sponge? mud mound facies (2) at loc. 14.
- d Cross-bedded crinoidal calcarenite grainstone of flank beds. Dashed line represents horizontal bedding. Scale 15 cm.
- e Graded bed in cross-stratified crinoidal sediments in figure d.
- f Ungraded crinoidal rudstone beds capped by lime mudstone. Photograph taken of area denoted by "1" in figure c. Scale 15 cm.
- g Rhizoidal holdfast and fasciculate rugose coral-rich zone within mud mound (loc. 14)
- h Massive dark grey lime mudstone of restricted shelf facies (loc. 14). Scale 4 in (10 cm).
- i Rhizoidal holdfast pod (P) in crinoidal packstone/grainstone at loc. 6. Note abrupt contact. Scale 15 cm.
- j Second pod (P) from same locality as figure i. Scale 15 cm.



PL4-14



V

PAPER V: STRATIGRAPHY AND FACIES DISPOSITION OF
THE UPPER SILURIAN CARBONATE MEMBER OF THE
BARLOW INLET FORMATION

Abstract

Eight major facies associations, herein called superfacies, are recognized in the Upper Silurian carbonate member of the Barlow Inlet Formation on Cornwallis Island. These superfacies represent environments ranging from distal foreslope (in contact with anoxic basal water mass of the Cape Phillips Basin), through rimmed shelf edge to tidal flat. The environments were associated with a large attached interior platform straddling a flexure between stable craton and miogeocline. Medial foreslope, proximal foreslope, shelf edge and proximal back-barrier superfacies contain diverse buildups and biostromes, that represent biota that apparently flourished in oxygenated near-surface waters of the basin.

The platform (Barlow Inlet Platform) evolved from a precursor ramp (Douro Ramp) by structural modification commensurate with subjacent block faulting of the Boothia Horst in latest Ludlovian time. Continued spasmodic epeirogeny affecting this basement structure, controlled periodic areal enlargement and drowning of the platform. These principal forestepping and backstepping events were superimposed on longer term accretionary outbuilding and

regional subsidence. In early-mid Lochkovian time the platform emerged during a third pulse of the Cornwallis Disturbance.

The distal basin slope and cyclic platform interior facies are largely indistinguishable from the C member of the Cape Phillips Formation of northern Cornwallis Island, and the lower member of the Somerset Island Formation of northern Somerset Island, respectively. However mapping contingencies dictate that formational boundaries remain unchanged.

5.1 INTRODUCTION

The Upper Silurian upper member of the Barlow Inlet Formation, and correlative beds of the Read Bay Group (FIG. 5-1) of shallow carbonate shelf aspect, arguably represent one of the most extensive carbonate platform deposits in the ancient record and certainly do with respect to the Upper Silurian. Carbonate platform lithofacies of the Read Bay Group are recognized in the subsurface, or in outcrop, from Banks Island to northwestern Ellesmere Island in the Canadian Arctic Archipelago, a distance of approximately 1400 km.²⁷

Despite the areal extent, appreciable stratigraphic thickness²⁸ and hydrocarbon reservoir potential, the rocks of the Read Bay Group have been little studied.²⁹ This paper describes and interprets the upper member of the Barlow Inlet Formation (hereafter referred to simply as the Barlow Inlet Formation) in one area of the Arctic Archipelago centred about Cornwallis Island (FIG. 5-2; PL.5-1).

²⁷ by way of comparison the rimmed shelf complex of the Great Barrier Reef is 1200 km long.

²⁸ the upper member of the Barlow Inlet Formation is greater than 1.2 km thick at its type section.

5.1.1 Regional Geological and Stratigraphic Setting

The study area is located in the Boothia Uplift structural province (FIG. 5-3), a region of basement block-faulting and supracrustal folding during Siluro-Devonian epeirogenesis.

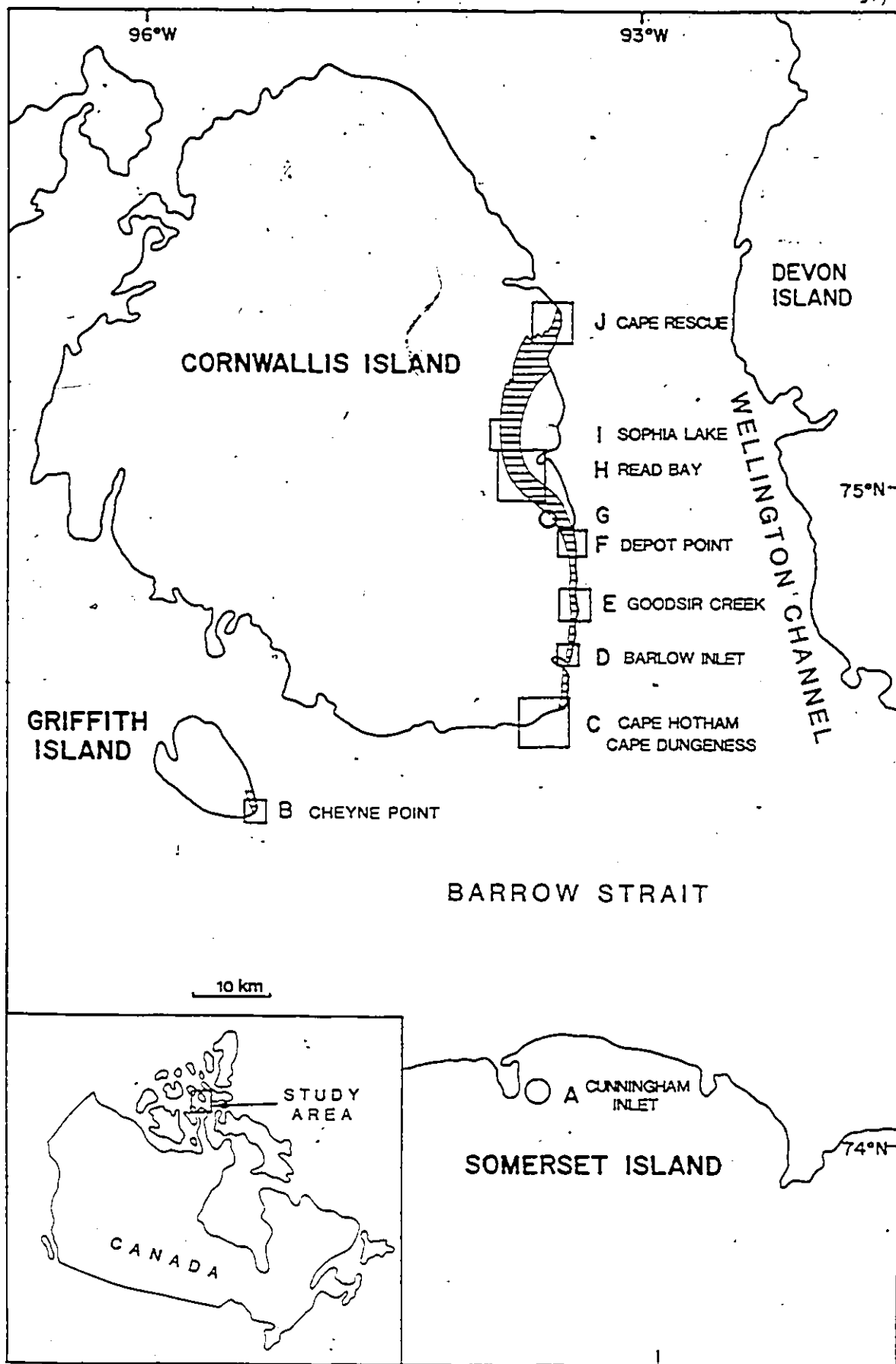


FIG. 5-2 Location map of study area.

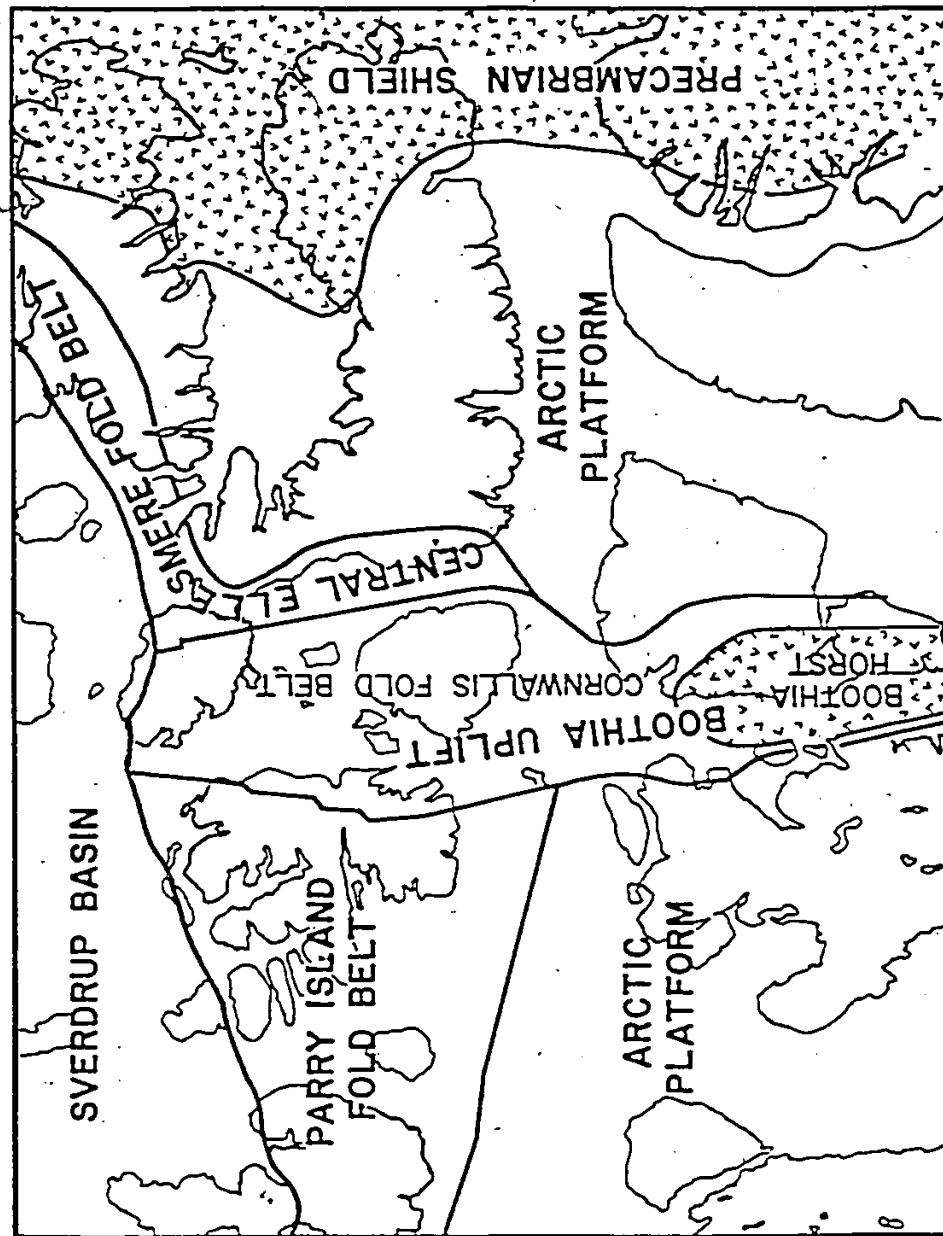


FIG. 5-3 Phanerozoic structural provinces of central arctic archipelago.

Cornwallis and Griffith Islands lie within the domain of the southeastern unstable shelf of the Franklinian Basin or Miogeocline (FIG. 5-4), a tectono-depositional element of increased subsidence and sediment accumulation from Proterozoic to Devonian time. To the south of the Franklinian Basin lay the stable cratonic interior (Arctic Platform) that is divided into a number of basins and basement highs as indicated on FIG. 5-4.

The southeastern shelf of the geosyncline contains an intrashelf depression herein called the Cape Phillips (Back-Barrier) Basin (FIG. 5-5). The Barlow Inlet Formation was laid down on a shallow carbonate platform (Barlow Inlet Platform) bordering the southern margin of the Cape Phillips Basin. The Cape Airy - Snowblind Bay line of facies change (CASBLFC) indicated on FIG. 5-5, records the approximate position of a transition between graptolitic facies of the Cape Phillips Formation to the north, and shelly facies of the Allen Bay - Cape Storm Formations and Read Bay Group to the south.

On Cornwallis Island the upper member of the Barlow Inlet Formation is underlain by a thin interval (approx. 50 m) of principally dark siliciclastic mudrocks that constitute the prodeltaic lower member of the Barlow Inlet Formation (Thorsteinsson, 1980; Paper 2 this thesis; FIG. 5-1). The Formation as a whole is underlain by carbonate ramp lithofacies of the Douro Formation, which is 300 to 450 m thick on Cornwallis and Griffith Islands.

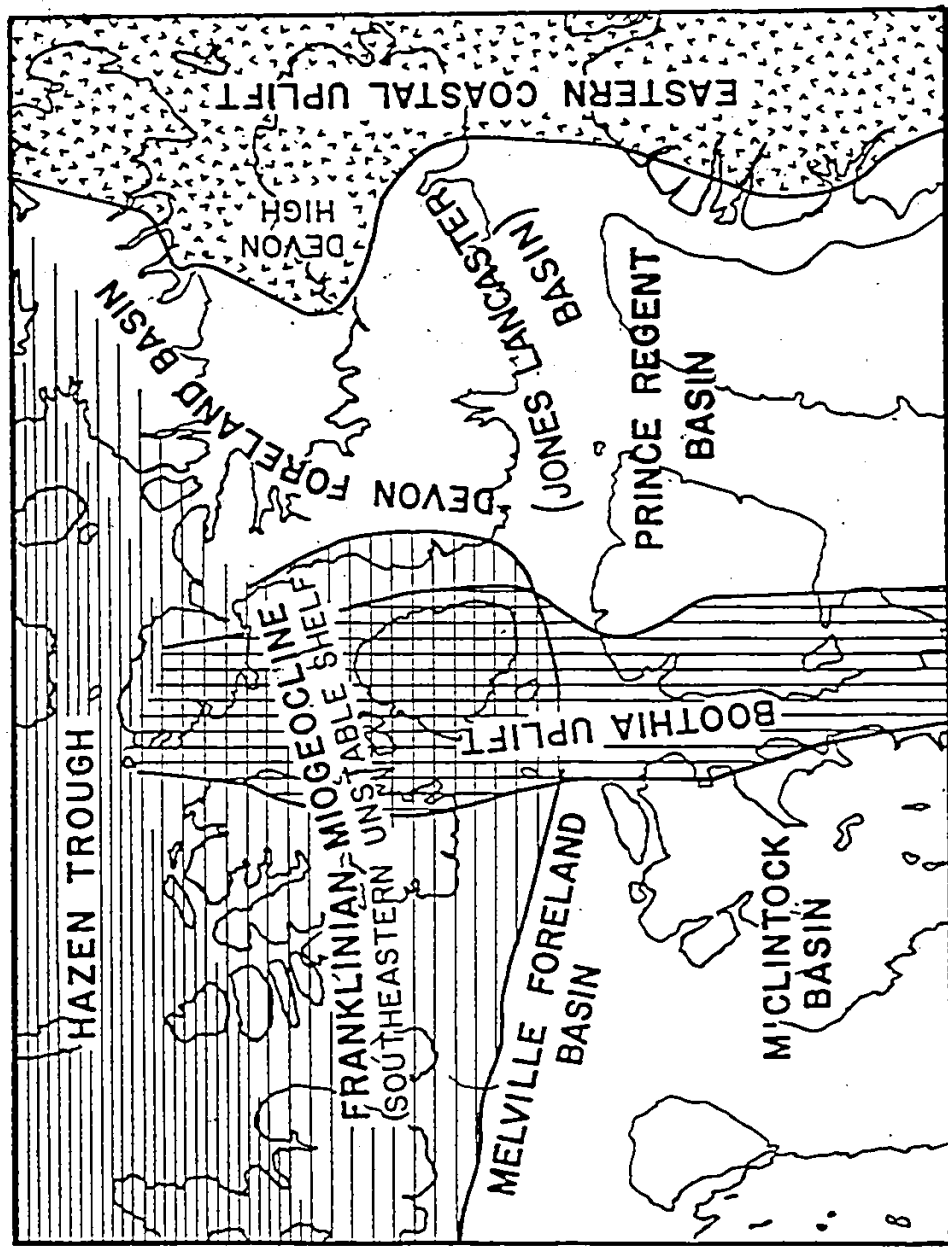


FIG. 5-4 Siluro-Devonian tectono-depositional elements of the central arctic archipelago.

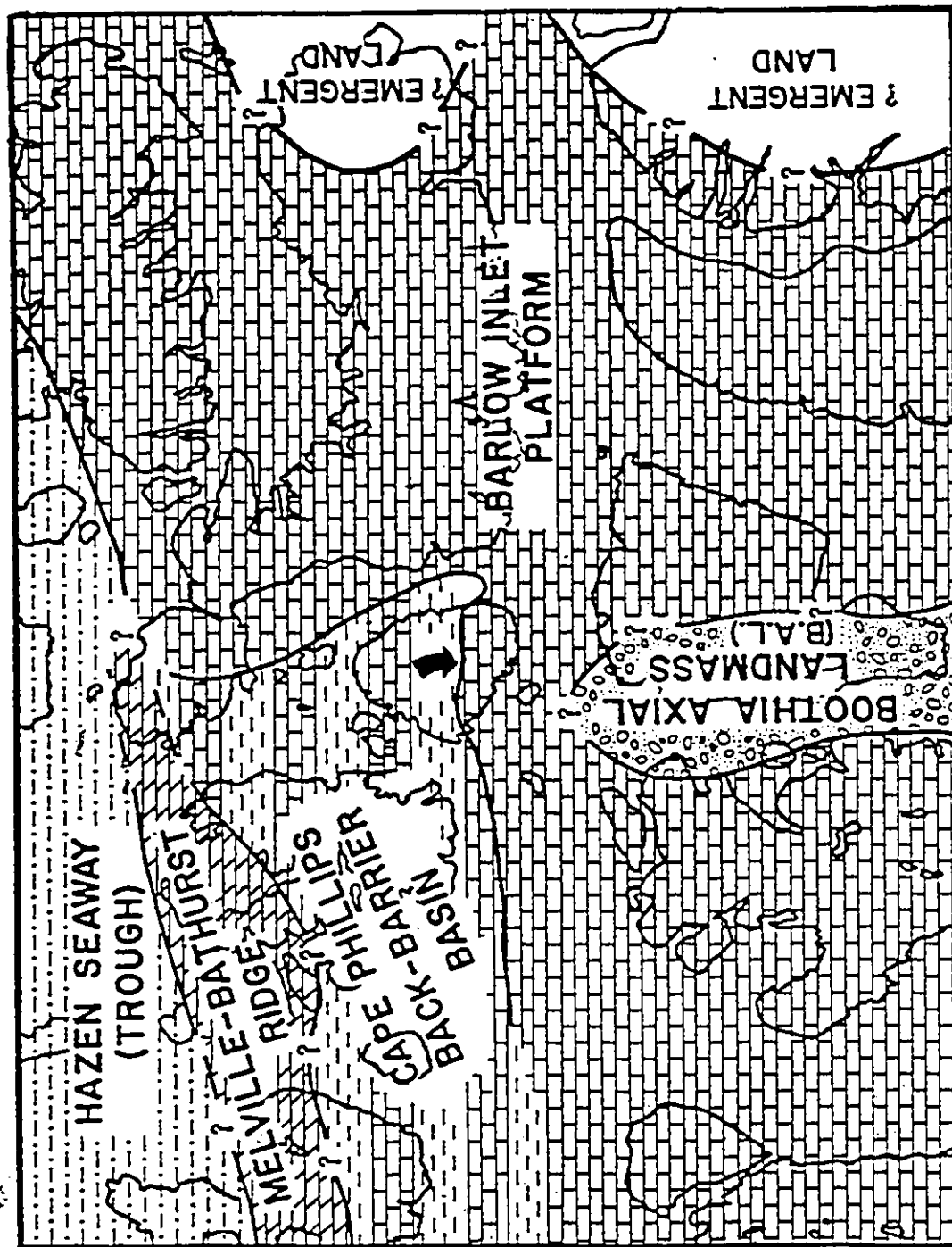


FIG. 5-5 Siluro-Devonian paleogeographical elements of central arctic archipelago.
 → CASBLFC (Cape Airy-Snowblind Bay line of facies change)

Conformably overlying the Barlow Inlet Formation, is the 560 m thick (at its type section) Sophia Lake Formation of marginal marine aspect. To the north of the facies change is the deeper water Cape Phillips Formation (composed of carbonate and siliciclastic mudrocks of basin and basin slope aspect). On Somerset Island, equivalent rocks belong to the peritidally-deposited lower member of the Somerset Island Formation.

The upper member of the Barlow Inlet Formation is 1272 m thick at its type section at Read Bay (loc. H, FIG. 5-2), and is composed of a wide variety of shallow to deeper marine carbonate facies. A more detailed synopsis of regional stratigraphic and facies relationships is given in section 5.4.3 'Regional Correlation', below.

5.1.2 Age

Based on conodont age determinations, the Barlow Inlet Formation ranges in age from latest Ludlovian to earliest Lochkovian (Uyeno, 1977; Thorsteinsson, 1980). There is at present no evidence of major disconformities within the succession, although minor hiatuses are doubtlessly associated with forestepping and backstepping events affecting the Barlow Inlet Platform (see part 5.4.2, 'Platform Design and Evolution', below).

The macrofauna supports a Late Silurian age (Table 5-1), but pending more detailed taxonomic investigations, the

utility of this fauna for biostratigraphic purposes remains limited.

5.1.3 Previous Work

The upper member of the Barlow Inlet Formation was established as the C member of the Read Bay Formation (Thorsteinsson, 1958). The change in nomenclature follows Thorsteinsson (1980); neither the boundaries of the member nor the location of its type section are altered.

Thorsteinsson (1958) first recognized and described the C member following reconnaissance field work on Cornwallis Island in 1950, 51 and 52. He chose a type section at Read Bay (loc. H), one of few localities where a complete, if somewhat poorly exposed, section of the member existed. Due to "...unforeseen circumstances..." (op.cit., p.65) the type section was never formally measured or described; this has resulted in the existence of a 1.2 km thick formal stratigraphic unit of rather enigmatic characteristics. Published sections are presently available for only the lower 220 m of the member (op.cit., p.65-67).

Published references to the Barlow Inlet Formation are limited to

Age ranges of selected macrofauna from the upper member,
Barlow Inlet Formation

MACROFOSSIL	STRATIGRAPHIC RANGE ABOVE BASE AND LOCALITY	AGE RANGE

CORAL		
<u>Catenipora</u> sp.	12-20m, loc.D, sect.4 1-27m, loc.E, sect.9	Ord. - Sil.
<u>Mucophyllum</u> sp.	15-30m, loc.E, sect.9 95-100m, loc.F, sect. ?980m, loc.J, sect.29	Mid to Late Sil.
BRACHIOPOD		
<u>Atrypoides foxi</u>	5-218m, loc.E, sect.9 highest occurrence is 280-285m, loc.F, sect.17	Late Ludlow - Early Pridoli
<u>A. erebus</u> ¹	12-15m, loc.E, sect.9 highest occurrence is 280-285m, loc.F, sect.17	
<u>Nanukidium</u> <u>cunninghamensis</u> ¹ (subfamily Pentamerinae)	15-30m, loc.E, sect.9 95-105m, loc.E, sect.7	Late Ord. - Early Dev.
ARTHROPODS		
<u>Hemiarces</u>	178-194m, loc.E, sect.8	Late Sil.
<u>agilonius</u> ¹	166-188m, loc.E, sect.11 225-230m, loc.F, sect.16	
<u>H. bigener</u> ¹	1192-1194m, loc.H, sect.20	Latest Sil - ?Early Dev.
STROMATOPOROIDS		
amphiporids	23-52m, loc.D, sect.4 385-408m, loc.I, sect.22	Sil. - Jur.
<u>Stachyodes</u> sp.	15-30m, loc.E, sect.8	Mid Dev. - Late Dev.
OSTRACODERMS		
<u>Ariaspis</u> sp.	8-12m, loc.D, sect.5 9-13m, loc.E, sect.7	Early Dev.

1. these are endemic forms, and the utility of their apparent age ranges for biostratigraphic purposes is limited.

reports and maps of a more regional nature: Thorsteinsson and Kerr (1968), Thorsteinsson in Douglas (1970), Thorsteinsson (1973), Gibling and Narbonne (1977), and Thorsteinsson (1980). Mayr (1978) has published on the subsurface equivalents of the Barlow Inlet Formation in the E-45 wildcat well on Devon Island.

Studies dealing primarily with the taxonomy of macro- and trace fossils collected at least in part from the Barlow Inlet Formation, include Bolton (1966), Jones and Packard (1979), Rohr and Packard (1981), Jones (1979, 1981a, 1981b), Narbonne and Packard (1983), and Parkins (in press). Papers concerned with the Read Bay Group and relevant to the present work on the Barlow Inlet Formation include Mayr (1974), McGill (1974), Poey (1982), Narbonne and Dixon (1982; in press), and Jones and Dixon (1978).

Specific aspects of the stratigraphy and sedimentology of the Barlow Inlet Formation as exposed on Cornwallis Island are outlined in Packard and Dixon (1979, 1980, 1982, 1983).

5.1.4 Methodology

Forty sections totalling 6.2 km and encompassing 1402 field units, were examined over three field seasons, in 1977, 1978 and 1980 (see Tables 1-1, 1-2 of Paper 1, this thesis). Standard stratigraphic techniques were employed (generally Jacob's staff; fixed-length chain where necessary). Data were recorded digitally on 80 character

unit cards modified by the author from an original Geological Survey of Canada design. Lithologic samples were taken from virtually every field unit. Individual fossils and fossiliferous slabs were collected systematically throughout the formation. Lithologic samples were slabbed, polished and stained², with stained thin sections being prepared from representative samples of both typical and clearly anomalous lithofacies. Field data were processed by Fortran IV programs designed by the author, using the CMS (Conversational Monitor System) available on the University of Ottawa's IBM 370.

Carbonate rock classifications utilized in this study, include a modified version (see Table 4-2 -- Paper 4 this thesis) of Embry and Klovan (1971), and the rubbly carbonate classification scheme of Jones et al. (1978).

5.1.5 Format of Presentation

The environmental analysis of carbonate rocks is now well-established and a number of excellent syntheses on carbonate facies and microfacies are presently available (Wilson, 1975; Flügel, 1982; Scholle et al., 1983). Inasmuch as there is a broad consensus on the criteria used in the recognition of various carbonate depositional environments, the author has decided to forego the more traditional 'descriptive stratigraphy followed by

² combined Alizarin red and potassium ferricyanide stain described by Friedman (1959).

environmental analysis' format. Instead the major depositional realms (herein called superfacies) of the Barlow Inlet Formation are described in general (part 2.0 below), with aspects of these superfacies emphasized and their sequences examined and locally correlated in eight major study localities (C to J, FIG. 5-2; PL.5-9,5-11,5-12,5-15, and 5-17 to 5-21) (Appendix B). The final part of the paper deals with physical correlation across the study area, highlights the evolution of the Barlow Inlet Platform, and considers the regional relationships of the Barlow Inlet Formation with coetaneous stratigraphic intervals on neighbouring islands (part 3.0 and Appendix C).

Detailed stratigraphic descriptions and schematic logs are available on file in the Department of Geology, University of Ottawa. Numbers assigned to the stratigraphic sections in the field are used in Paper 1 (this thesis) and in data on file. A simplified system of section numbers (Table 5-2) has been adopted in Paper 5.

Table 5-2

530

Index of stratigraphic sections

Paper 5 section number	Original section numbers as recorded in field notes and Paper 1
1	15b
2	15c
3	12
4	11
5	14b, 14c
6	14a
7	3
8	2a, 2b, 2c
9	1a
10	1b
11	4b
12	4a
13	7a
14	7b
15	7c
16	7d
17	8
18	6
19	9
20	5a
21	5b
22	10
23	22
24	21
25	20a
26	20b
27	20c, 20d, 20e
28	16a
29	16b
30	17
31	18
32	19

5.2 DESCRIPTION OF SUPERFACIES

The Barlow Inlet Formation is remarkable as a formal stratigraphic unit in that it contains such a heterogeneous assemblage of lithofacies, representing carbonate environments ranging from tidal flat to distal foreslope. Equally, it is rare that such a complete spectrum of lithofacies is so well exposed and preserved in one outcrop belt, a function of the structure and physiography of the study area, coupled with the cold dry arctic climate (minimal vegetation cover or contemporary karstification). The Barlow Inlet Formation is thus ideal for examining the makeup of an ancient carbonate platform. The value of such a study is enhanced in that carbonate platform facies of latest Silurian age have received scant attention in the available literature (Wilson, 1975, pp.105,119).

The nomenclature observed in this paper is illustrated in Table 5-3. A hierarchy of five facies or facies assemblages is recognized: Magnafacies, Superfacies, Facies, Subfacies and Varieties. Correlations are principally made at the level of magna- and superfacies. It will be noted that superfacies of the Platform Margin and Basin Slope do not contain a 'reef' label, to avoid the ambiguity associated with this term (see Heckel, 1974). Several other terms utilized in this text require clarification.

Facies hierarchy

MAGNAFACIES and SUPERFACIES	FACIES (including BIOCUMULES)	SUBFACIES (non-inclusive list, examples only)
<u>BASIN SLOPE (BS)</u>		
DISTAL FORESLOPE (BSdf)	BSdf1. Thin bedded, planar laminated lime mud- stone rhythmically intercalated with marl (ribbon limestone)	BSdf1a. As BSdf 1. but with abun- dant peloid rich, silty laminae
	BSdf2. Faintly mottled thin to medium wavy-bedded lime mudstone	
	BSdf3. Mollusk grainstone/ packstone	
	BSdf4. Hardgrounds	
MEDIAL FORESLOPE (BSmf)	BSmf1. Ribbon limestone (fine lamination less com- monly preserved)	BSmf6a. Lithistid sponge- spongiostromate algae-dominated biocumule
	BSmf2. Nodular to wavy, thin bedded lime mudstone/ wackestone (approaching mottled dolomitic limestone type 2)	
	BSmf3. Lithoclastic crinoidal packstone	
	BSmf4. Oligomictic mud-supported carbonate intraformational breccia	
	BSmf5. Polymictic carbonate mega- breccia	
	BSmf6. Lithistid sponge dominated biocumule	
	BSmf7. Mucophyllid mud mound	
<u>PLATFORM MARGIN (PM)</u>		
PROXIMAL FORESLOPE (PMpf)	PMpf1. Thin planar bedded lime mudstone (wackestone)	PMpf3a. Crinoidal- solenoporid rudstone
	PMpf2. Wavy to nodular lime mudstone (wackestone)	
	PMpf3. Massive to thick bedded crinoidal packstone/ rudstone	
	PMpf4. Crinoidal-brachiopod floatstone/rudstone	
	PMpf5. Fasciculate rugose coral biostrome	
	PMpf6. Dendroid stromatoporoid rudstone	
	PMpf7. Diverse community mud mound	
	PMpf8. Rhizoidal holdfast boundstone (continued)	

PLATFORM MARGIN (PM) - continued

SHELF EDGE PMse PMse1. Dolomitized massive skeletal biocumule
 PMse2. Zoned biocumules
 PMse3. Massive thick cross-bedded crinoid packstone/grainstone
 PMse4. Lamellar stromatoporoid biostrome
 PMse5. Stromatoporoid-crinoid rubble skeletal buildup
 PROXIMAL BACK BARRIER (PMbb) PMbb1. Massive stromatoporoid-dominated biostrome (sparse to dense structure)
 PMbb2. Oncoidal carbonate
 PMbb3. Fasciculate rugose coral biostrome
 PMbb4. Amphiporid grainstone to floatstone

PLATFORM INTERIOR (PI)

RESTRICTED SHELF LAGOON (Pisl) Pisl1. Massive stromatoporoid biostrome with sparse structures
 Pisl2. Oncoidal carbonate
 Pisl3. Sparsely fossiliferous mottled dolomitie carbonate
 Pisl4. Amphiporid floatstone
 Pisl5. Massive dense lime mudstone
 Pisl6. Coenites-capped mud mound
 Pisl5a. Megalodont bivalve floatstone
 Pisl5b. Fenestral lime mudstone
 TIDAL FLAT (Pitf) Pitf1. Cryptalgal laminite
 Pitf2. Gastropod-oncoid rudstone
 Pitf3. Massive weakly calci/dolosiltite
 Pitf4. Massive dolosiltite
 Pitf5. Dense lime mudstone (fenestral)
 Pitfla. Laminite with laminoid fenestrae

1. margin versus edge -- platform margin is used here to refer to the marginal area of the carbonate platform, and as such incorporates the shelf edge, proximal foreslope, and proximal back-barrier. Edge refers more specifically to the shelf-slope break commonly occupied by sand shoals and ecologic reefs.
2. carbonate platform complex -- similar in use to 'reef complex' (e.g. Burchette, 1981), or 'rimmed-shelf complex'. A platform complex would include basin slope, platform margin and platform interior magnafacies.
3. platform -- 1) informal or abbreviated usage of the term carbonate platform complex. 2) refers to a 3-dimensional, positive physiographic feature representing a relatively long geological history of carbonate deposition, erosion, and possible structural modification and/or igneous activity.
4. carbonate shelf (or simply shelf) -- essentially 2-dimensional upper surface of a platform or carbonate ramp. Ideally, although not practically, a shelf in geological terms represents a single synchronous surface.
5. biocumule -- a term introduced in Paper 4 (this thesis) and defined as 'a discrete stratigraphic or lithofacies unit that is clearly distinguished from supra-, sub-, and adjacent units on the basis of

either 1) its concentration of predominantly autochthonous skeletal material (generally of rudite size or larger), or 2) evidence of inferred binding and/or baffling activity of a concentration of non-skeletal organisms.' Thus the term biocumule emphasizes the aspect of in-situ organic growth without implying the geometry of the resulting sedimentary body.

6. infratidal zone -- very shallow subtidal zone.
7. peritidal zone -- incorporating infratidal, intertidal, and supratidal zones.

The Platform Interior Magnafacies embraces all littorally-influenced environments landward of rimmed-shelf edge; some or all of these environments were subject to limited exchange with the open ocean, rapid sediment production and accumulation, fluctuating or extreme salinities, and periodic emergence.

Strata of the Platform Interior Magnafacies are arranged cyclically in relatively thin (1-20 m) shallowing-upward sequences (PL. 5-8h). Lithologies representative of proximal back-barrier, restricted shelf lagoon and tidal flat environments occur within these hemicycles, although interpreting the limits for each lithofacies, within what is commonly a gradational succession, is problematical. In general the peritidal lithofacies of the Platform Interior are sparsely fossiliferous, and even where concentrations of

skeletons occur (e.g. PL. 5-8b,e), the biota exhibits low diversity and few stenohaline forms. Two intergradational and generalized hemicyclic sequences are recognized, and have been named after the characteristic or conspicuous environmental zone within each sequence type. These two types are, the Tidal Flat (or TF) hemicycle, and the Restricted Infratidal (or RI) hemicycle (FIGS. 5-6, 5-7). The nature and origin of hemicyclicity in the Platform Interior Magnafacies is thoroughly examined in Paper 3 (this thesis).

The Platform Interior Magnafacies is composed of two superfacies, the restricted shelf lagoon, and the tidal flat superfacies (Table 5-4), discussed below.

The Platform Margin Magnafacies incorporate all facies occurring at, or immediately adjacent to, the shallow water shelf-slope break (commonly occupied by reefs and sand shoals). The proximal foreslope, shelfedge, and proximal back-barrier superfacies are included. This broad grouping of facies is usefully-employed in situations where intense dolomitization has obliterated the distinctions between the three superfacies listed above, and yet available data does warrant assigning a particular stratigraphic interval to the platform marginal paleogeographic position.

The Basin Slope Magnafacies embraces carbonate slope environments transitional between platform margin and basin floor, but excluding peri-reefal (or shoal) environments

immediately basinward from, and significantly influenced by, the shelf edge proper. The Basin Slope Magnafacies includes distal and medial foreslope superfacies.

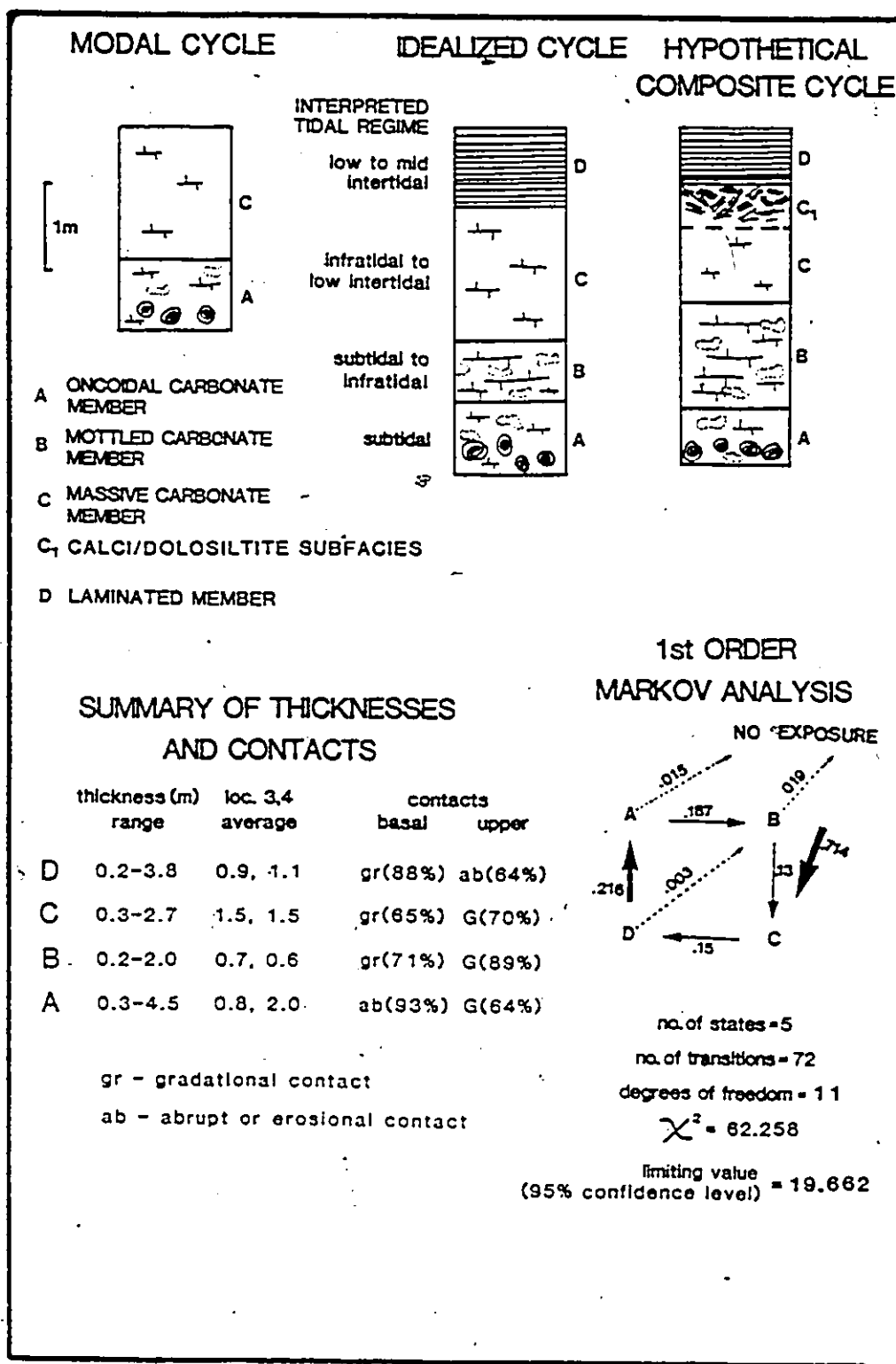


FIG. 5-6 Tidal Flat hemicycles.

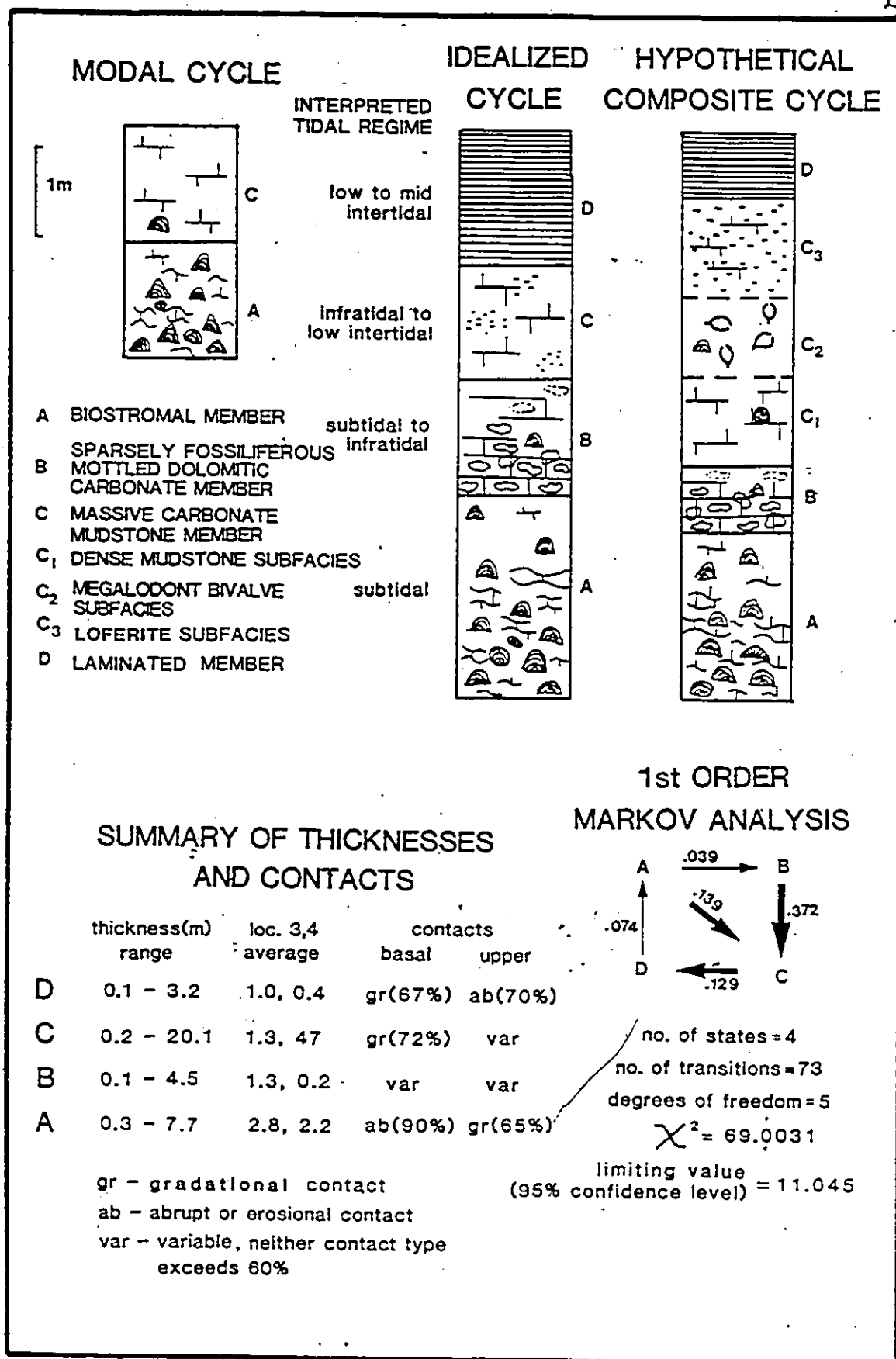


FIG. 5-7 Restricted Infratidal hemicycles.

Table 5-4

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Characteristics of Superfacies

SUPERFACIES	LITHOFACIES (refer to table 5-3) 1.MOST PREVALENT 2.OTHERS COMMON	BEDDING 1.THICKNESS 2.STYLE	MACROBIOTA 1.DISTRIBUTION 2.ABUNDANCE OF SKELETAL MACROBIOTA 3.AUTOCHTHONY OF BIOFACIES WITH HIGH CONCENTRATIONS OF FOSSILS ¹	BODY FOSSILS 1.DISARTICULATED FRAGMENTS 2.WHOLE OR NEARLY SO
DISTAL FORESLOPE	1.BSdf1 2.BSdf2	1.Thin 2.Planar	1.very localized 2.very rare 3.D	1.Shell debris 2.Orthoconic nautiloids, small bivalves, trilobites, gastropods
MEDIAL FORESLOPE	1.BSmf1 BSmf2 BSmf3 2.BSmf4 BSmf6	1.Thin- medium 2.Wavy to Nodular	1.Localized 2.Very rare to abundant 3.C	1.Crinoid debris 2.Lithistid sponges, rugose corals, low domical favostids, brachiopods
PROXIMAL FORESLOPE	1.PMpf2 PMpf3 PMpf7 2.PMpf1 PMpf5 PMpf6	1.Thin- thick 2.Wavy to massive	1.Variable, tending to widespread 2.Rare to very abundant 3.B.C.	1.Dendroid stromatoporoids, crinoid debris, solenoporids, brachiopod debris, bryozoans, algae 2.Tabulate corals, stromatoporoids, brachiopods, rugose corals

(continued)

1. A - in situ

B - autochthonous (from immediate environment)

C - allochthonous (proximal)

D - allochthonous (distal)

Table 5-4 (contd.)

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SUPERFACIES	LITHOFACIES (refer to table 5-3)	BEDDING 1.THICKNESS 2.STYLE	MACROBIOTA 1.DISTRIBUTION 2.ABUNDANCE OF SKELETAL MACROBIOTA 3.AUTOCHTHONY OF BIOFACIES WITH HIGH CONCENTRATIONS OF FOSSILS	BODY FOSSILS 1.DISARTICULATED FRAGMENTS 2.WHOLE OR NEARLY SO
SHELF EDGE	1.PMse1 PMse3 2.PMse5 PMse4	1.Thick-non- bedded 2.Lenticular, wedge-shaped, massive	1.Widespread 2.Common 3.B,C	1.Crinoid debris, solenoporids 2.Massive & lamellar stromatoporoids, massive favostids, rugose corals
PROXIMAL BACK BARRIER	1.PMbb1 2.PMbb4	1.Thin- medium 2.Wavy, nodular, massive	1.Widespread 2.Common to abundant 3.A,B	1.Brachiopod debris 2.Massive stromatoporoids, amphiporids, rugose corals
RESTRICTED SHELF LAGOON	1.Pisl2 Pisl3 Pisl5 2.Pisl4	1.Non-bedded to thick 2.Massive	1.Variable 2.Very rare to abundant 3.A	1.Ostracodes, amphiporids 2.Megalodont bivalves, other bivalves, stromatoporoids, amphiporids, gastropods, dendroid tabu- lates, oncoids
TIDAL FLAT	1.Pitf1 2.Pitf3 Pitf5	1.Non-bedded to thin, finely laminated 2.Planar to slightly wavy	1.Localized 2.Very rare 3.A (algae), C	1.Ostracoderms, ostracodes, amphiporids 2.Ostracoderms, ostracodes, gastropods, spongiostromate algae

(continued)

Table 5-4 (contd.)

SUPERFACIES	BIOTURBATION 1.DISTRIBUTION 2.AVERAGE DEGREE	DISCRETE TRACE FOSSILS 1.CONSPICUOUS 2.OTHER COMMON	PALEOSLOPE INDICATORS 1.ABUNDANCE 2.TYPE	EVIDENCE OF SYNSEDIMENTARY EMERGENCE (PERIODIC, IF NOT FREQUENT)
DISTAL FORESLOPE	1.Localized 2.Very limited	1. <u>Phycodes</u> 2. <u>Nov. Gen.</u> (aff. - <u>Phycodes</u>) <u>Cruziana</u> <u>Pelecypodichus</u> sp. A <u>Tanaedium</u> <u>Skolithos</u> <u>Palaeophycos</u>	1.Rare 2.Intraform- ational truncation surfaces	None
MEDIAL FORESLOPE	1.Variable, tending to Common 2.Partial	1. <u>Pilichnia</u> 2. <u>Zoophycos</u> <u>Chondrites</u>	1.Common 2.Intraform- ational Truncation Surfaces Slide deposits Mass flow deposits	None
PROXIMAL FORESLOPE	1.Widespread 2.Partial to thorough	1. <u>Zoophycos</u> 2. <u>Chondrites</u> <u>Palaeo-</u> <u>phycos</u> <u>Skolithos</u>	1.Rare 2.Intraform- ational truncation surfaces Slide deposits Talus deposits None	None
SHELF EDGE	?	1. - 2.no discrete traces recorded	None	If DORAG dolomitization model is adopted, then subaerial emergence of part of plat- form is required.
(continued)				

Table 5-4 (contd.)

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SUPERFACIES	BIOTURBATION 1.DISTRIBUTION 2.AVERAGE DEGREE	DISCRETE TRACE FOSSILS 1.CONSPICUOUS 2.OTHERS COMMON	PALEOSLOPE INDICATORS 1.ABUNDANCE 2.TYPE	EVIDENCE OF SYNSEDIMENTARY EMERGENCE (PERIODIC, IF NOT FREQUENT)
PROXIMAL BACK BARRIER	1.Widespread 2.Partial	1. - 2.no discrete traces recorded	None	? Fenestral fabric occurs in some back- barrier biostromes ?
RESTRICTED SHELF LAGOON	1.Widespread 2.Partial to thorough; possibly very limited	1. <u>Pelecypodichus</u> sp. B 2. <u>Skolithos</u> <u>Planolites</u> <u>Chondrites</u>	None	
TIDAL FLAT	1.Localized 2.Limited to partial	1. <u>Polarichnus</u> 2. <u>Chondrites</u> <u>Uchirites</u> <u>Skolithos</u> <u>Arthraria</u>	None	Fenestral fabric common- ly occurs in massive calci- lutite - Desiccation - Adhesion ripples

(continued)

Table 5-4 (concluded).

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SUPERFACIES	1.AMOUNT OF DOLOMITE 2.DEGREE OF DOLOMITIZATION	OTHER STRUCTURES OR CHARACTERISTICS	INTERPRETED RELATIVE HYDRAULIC ENERGY.
DISTAL FORESLOPE	1.Small 2.Insignificant	-Fe/Mn Nodules -Hard grounds are most common in the distal foreslope superfacies	Low
MEDIAL FORESLOPE	1.Small 2.Insignificant to limited	-Local stabilization of medial foreslope substrate by rhizoidal holdfast cementation -Gutter casts (tempestite indicator)	Low
PROXIMAL FORESLOPE	1.Moderate 2.Moderate to pervasive	Biocumules are particularly conspicuous in this superfacies	Low - moderate
SHELF EDGE	1.Large 2.Pervasive	-Crossbeds in calcarenites -Absence of ooids -Pods of micrite-bound rhizoidal holdfasts	Moderate to high
PROXIMAL BACK BARRIER	1.Small to moderate 2.Limited to pervasive	Biostromes generally have a sparse structure, and rubbly weathering	Low to moderate
RESTRICTED SHELF LAGOON	1.Small to moderate 2.Insignificant to pervasive	-Cyclic arrangement of strata -Omission surfaces -Possible subtidal channel structures	Low
TIDAL FLAT	1.Moderate to large 2.Moderate to pervasive	-Wavy to ripple-cross lamination -surface blisters -channel structures	Low

5.2.1 Distal Foreslope Superfacies

The distal foreslope superfacies (or basin margin of many authors) is typified by very sparsely fossiliferous, rarely bioturbated, fine grained carbonate sediment that exhibits thin, even, planar-bedding, and is disrupted (rarely) by slope failure phenomena. Except for work done by Trettin (1979), there have been few sedimentologic studies of the Cape Phillips Formation, and consequently the basin floor facies of the Cape Phillips back-barrier basin are not well understood. Apart from the structures (above) suggestive of a paleoslope, there may be little to distinguish deep undathem or. fondothen (basin floor) from distal foreslope facies.

5.2.1.1 Characteristic Lithofacies

The principal lithofacies is medium-dark grey, generally laminated, thin-bedded, calcilutite mudstone with rare body or trace fossils (PL. 5-2b). Beds are intercalated with thin argillaceous lime mudstone (PL. 5-2a).

The thickest laminae are composed of micrite to microspar with up to 20% dolomite rhombs (less than 50 microns) and 5% fine angular quartz silt. These fine-grained laminae are up to 5 mm thick and constitute the bulk of the laminated mudstone facies. Thinner laminae of coarser allo- and orthochems, rarely constitute more than 30% of the cross-sectional area of any given sample, but locally predominate

in some thin sections (PL. 5-2e). The most common grain type is an indeterminate calcareous (bio?)clast. Peloids and quartz silt are also common. The grains are moderately-sorted, ranging from 10-60 microns with a mode of 20-40 microns. These coarser grained laminae can be planar, wavy, or somewhat wispy and discontinuous. Thicker coarse grained laminae exhibit sharper upper and lower contacts and typically contain coarser constituents, with recognizable crinoid, brachiopod and gastropod elements. Clasts are commonly rimmed by opaque (?Fe-Mn) crusts. Basal contacts are locally erosive (PL. 5-2e). No grading of ortho- or allochems is present.

A second lithofacies is weakly-mottled, somewhat wavy to predominantly planar, thin to medium-bedded lime mudstone to wackestone (PL. 5-2j). Lamination is weakly developed or absent. Mottling, reflecting selectively greater dolomitization of diffuse zones within the rock is typically only faint on bedding surfaces. Shaley partings between beds are thin or non-existent, and the amount of argillaceous impurities within the mudstone is low. Allochems include fragments of crinoids, brachiopods, trilobites, and ostracodes.

Ungraded, well-sorted packstone and grainstone are present rarely within the distal foreslope superfacies. Packstone of diminutive high-spiral gastropods is associated with hardground development and Fe-Mn encrustation.

Brachiopod shell packstone and grainstone in lenticular deposits .1 to .5 mm thick, with undulating upper surfaces, (PL. 5-3c) are associated with syndimentary cementation phenomena. These deposits probably represent storm lags, in which inherently high initial permeabilities aided in submarine lithification. Hardgrounds are also associated with highly bioturbated limestone (PL. 5-2h), with patchy distribution of peloidal calcilutite packstone and calcilutite mudstone.

5.2.1.2 Structures

Apart from lamination, the only significant structure is bedding. In laminated calcilutite mudstone, beds are typically planar to very slightly wavy, thin and even-bedded. Bed thicknesses generally range from less than 1 cm to 5 cm.

The origin of bedding in "ribbon" limestones continues to be a subject of much debate (see Einsele and Seilacher, 1982 and papers therein). Enhancement of primary compositional layering by mechanical compaction and diagenesis reconciles most of the observed features of this class of limestones. Despite extensive field observations and sampling graded bedding (on any scale) has not been identified in the distal foreslope superfacies.

Intraformational truncation surfaces are rare, but conspicuous, features of the distal foreslope superfacies

(PL. 5-2g), and represent the only reliable indicators of slopes (however gentle). The beds overlying truncation surfaces are not themselves plastically or elastically deformed; their discordant attitudes merely reflect the inclination of beds filling subjacent hollows (slump scars?). The whereabouts of the slumped sediments remains enigmatic; this situation parallels that described by Davies (1977) for structures seen in carbonate slope deposits of Permo-Pennsylvanian age in the Sverdrup Basin.

5.2.1.3 Body Fossils

Body fossils are very rare, particularly so in the laminated planar-bedded lithofacies where no clearly autochthonous benthos has been identified. Fragmentary remains of orthoconic nautiloids, trilobites, gastropods, ostracodes and echinoderms are present. In the weakly mottled lithofacies strophomenid, rhynchonellid, and smooth spiriferid brachiopods, along with bryozoans are present in addition to the fauna mentioned above. Allochthonous packstones and grainstones contain shell debris (brachiopod), crinoid debris, pentamerid brachiopods, and solitary rugose corals.

Conspicuous by their absence from this superfacies, are graptolites, phyllocarids, and sponges.

5.2.1.4 Trace Fossils

Skolithos, Phycodes (PL. 5-2d), Pelecypodichnus, and Taenidium are the most commonly occurring ichnogenera. Discrete trace fossils are generally rare, but can occur in great abundance within restricted stratigraphic intervals (often single beds). Phycodes is the most characteristic ichnogenus of the superfacies.

5.2.1.5 Other Considerations

Ribbon limestones are deposited below wave base. Original compositional layering (Einsele, 1982) is generally attributed to rhythmic if minor climatic change. The origin of the textural laminae that constitute the beds is not clear. The fine grained laminae probably represent continuous suspension sedimentation, whereas the coarse-grained laminae are the product of sedimentation by distal sediment gravity flows generated either by slope failure or storm activity. The absence of grading and rare presence (poorly preserved) of foreset laminae, suggest bedload deposition (i.e. C division of Bouma cycle only).

The Na content of the soluble portion of ten representative samples of distal foreslope facies, ranges from 147 to 216 ppm placing these samples in the normal marine salinity field (Viezer et al., 1978). Insoluble residue contents range from 12.8 to 20.8% by weight.

Small, irregular Fe-Mn nodules (PL. 5-2i) occur as clasts within packstone-grainstone deposits. Bioclasts are commonly impregnated with opaque minerals (pyrite) and hardground surfaces, burrow margins and skeletal solution cavities are similarly encrusted. However, the color of the limestones is medium to medium-dark grey, and limited observations suggest that organic carbon does not exceed .5%. The generally well-preserved fine laminae, lack of benthic fossils, and sulphide-encrusted surfaces suggest that the distal foreslope facies was deposited under predominantly anaerobic conditions. This apparent contradiction may be explained in part by sediment dilution of the unoxidized organic matter (i.e. high sedimentation rate of non-organic, or previously oxidized sediment).

Overall the distal foreslope superfacies represents carbonate deposition under anaerobic (to disaerobic) conditions on a paleoslope situated below wave base.

5.2.2 Medial Foreslope Superfacies

The medial foreslope superfacies is typified by thin-bedded lime mudstone and wackestone, generally planar to wavy bedded but commonly mottled or nodular, with a limited benthos of open marine aspect (crinoidal debris predominates). Deposits or structures indicative of slope failure and mass transport are common. Indiscriminate bioturbation is generally pervasive but discrete traces are

also locally well-preserved and abundant. Lithistid sponge and mucophyllid mud mounds occur within this superfacies.

5.2.2.1 Characteristic Lithofacies .

The calcilutite mudstone and faintly-mottled wavy bedded lithofacies of the distal foreslope superfacies are present as well in the medial foreslope. Delicately preserved coarse-grained laminae are rare, attesting to increased sediment homogenization by burrowing infauna.

The most common lithofacies are mottled, thin planar to wavy bedded mudstone and wackestone with discrete lozenge-shaped nodules or more continuous hardbands of micrite (PL. 5-3f). The nodules and hardbands are considered to represent synsedimentary submarine lithification (see Jones et al., 1979), and are similar (although occurring in fine-grained rocks) to nodular sediments presently forming on the Bahamian platform foreslope (Mullins et al., 1980).

Crinoidal packstones are an important if subordinate lithofacies in the medial foreslope. Field units of this limestone type range from .1 to .5 m thick, and are both massive and ungraded. A common variety contains abundant subangular clasts of calcilutite mudstone in a crinoidal matrix. The presence of gutter casts (PL. 5-3j) on the soles of some crinoidal packstone units suggests a degree of sediment reworking by storms. Other packstones may have

originated as matrix-poor debris flows or grain flows.³⁰

Carbonate breccias, both polymictic and oligomictic, are associated with biocumules and slide structures. Clasts of boundstone as large as 1 m across are found in polymictic carbonate breccias proximal to lithistid sponge biocumules (PL. 5-3e). Other clast types include stromatoporoid and coral heads, calcilutite mudstone and crinoidal packstone. The bases of these breccias are commonly scoured.

The maximum observed thickness of polymictic carbonate breccia is 4 m. Oligomictic carbonate breccias are more commonly associated with slide structures (PL. 5-3a) and are composed of randomly arranged subspherical to tabular clasts set in dolomitic calcilutite (PL. 5-3f) or crinoidal packstone matrix (PL. 5-3c). As most of the deposits are matrix-rich they probably were transported as debris flows. Transformation of a slide into a debris flow is lucidly described by Cook and Mullins (1983).

Two biocumule types are considered to have developed on the medial foreslope. Lithistid sponge mud mounds, composed almost exclusively of micrite, spongiostromate algae and concentrations of lithistid sponge desmae, occur at several localities within the Barlow Inlet Formation outcrop belt (PL. 5-3i). These structures attain a stratigraphic relief

³⁰ Lowe (1976) has outlined theoretical constraints for grain flow deposits. In the absence of such diagnostic criteria as floating grains or reverse grading, the crinoidal packstones of the medial foreslope are unlikely to be grain flow deposits.

of up to 90 metres (Paper 4, this thesis), and formed discrete raised structures on the sea floor. Mucophyllid mud mounds occur only at locality J. They apparently grew as satellite reefs on the upper surfaces of lithistid sponge-dominated biocumules (see Paper 4, this thesis).

5.2.2.2 Structures

Structures in the medial foreslope superfacies include lamination (rare), gutter casts (PL. 5-3j), intraformational truncation surfaces (PL. 5-3a), interstratal recumbent folds (PL. 5-3d), and groove casts. Planar cross-beds, accretion foresets, and graded beds occur rarely within bioclastic halos of some lithistid sponge mud mounds (example B, zoned biocumules, Paper 4 this thesis), but are not recorded elsewhere on the medial foreslope. Bedding is invariably thin to medium (see Ingram, 1954). The groove casts (PL. 5-19g) are associated with carbonate breccia bedding soles, and may be the result of rigid plug sliding during debris flow transport (cf. Middleton and Hampton, 1976).

5.2.2.3 Body Fossils

The bulk of the fragmentary allochems are crinoidal. Whole crinoids and ophiuroids are conspicuous if rare elements of the fauna. Brachiopods are locally common, in particular strophomenids, pentamerids, rhynchonellids and atrypids. Atrypoids are relatively rare, and only

Atrypoides foxi and diminutive reef-dwelling atrypoids occur in this superfacies. Lithistid sponges, spongiostromate algae, low domical favositids, mucophyllid and stylopleurid rugose corals, orthoconic nautiloids, ostracoderms and bryozoans are common, if sporadically distributed fossils of the medial foreslope.

5.2.2.4 Trace Fossils

The ichnogenera Skolithos, Zoophycos, Pilichnia, Palaeophycus, and Chondrites occur in the medial foreslope superfacies and Pilichnia, in particular, is characteristic (PL. 5-3b). Certain beds of calcilutite mudstone are pervasively penetrated by these burrows (PL. 5-14e).

5.2.2.5 Other Considerations

The bioturbated nature of the lithofacies and the presence of a more diverse open marine shelly benthos, suggest that deposition in the medial foreslope took place in predominantly aerobic waters. Although hardgrounds have not been recorded in this superfacies, the abundance of nodular limestones, and angular, tabular clasts of foreslope lithofacies in debris flows, attests to the fact that near surface submarine cementation and the production of hardbands were ubiquitous. Micritic nodules and hardbands frequently exhibit narrow, tapering spar-filled fissures. These fissures have various orientations, but most taper

towards the soles of beds. The origin of these fissures is not known although they may be related to post-burial synaeresis of semi-consolidated micrite.

Slopes were colonized and to some degree stabilized by crinoids with rhizoidal holdfasts (PL. 5-3g). Associated with the network of rhizoids are micrite bodies, similar in size to the radicular cirri of the crinoids, that exhibit distinctive concentric color bands (PL. 5-3h). It is hypothesized that these micritic bodies are cementation zones that formed concomitant with endoskeletal resorption. The crinoids were thus bio-catalytic agents promoting lithification of slope sediment, adjacent to the holdfast structure.

5.2.3 Proximal Foreslope Superfacies

The proximal foreslope superfacies is typified by abundant bioclastic limestone exhibiting a very diverse open marine biota. Structures indicative of slope failure are present but less common than in the medial foreslope superfacies. Biocumules of diverse facies composition abound.

5.2.3.1 Characteristic Lithofacies

The predominant rock-type is medium to thick-bedded to massive crinoidal rudstone and packstone (PL. 5-4e). Bioclasts are principally crinoid pluricolumnals (and other

ossicles) and solenoporid algae. Some packstones contain abundant micritized bioclasts and peloids, but no coated grains. Interbedded with the crinoidal limestone are lime wackestone, floatstone, and rudstone with a diverse shelly fauna predominated by atrypid and atrypoid brachiopods. Lime mudstone with cyathaspidid fish, orthoconic nautiloids and pockets of Atrypoides erebus and A. foxi occurs in strata assigned to this superfacies, although this mudstone facies is possibly transitional to the medial foreslope. Mass flow deposits are rare in this superfacies.

Biocumule facies include diverse community mud mounds (PL. 5-4c), stromatoporoid-crinoid rubble skeletal buildups (PL. 5-4g), fasciculate rugose coral biostromes, dendroid stromatoporoid biostromes, and possibly lamellar stromatoporoid biostromes (see Paper 4, this thesis).

5.2.3.2 Structures

Intraformational truncation surfaces and interstratal slump folds are present but not common. Planar crossbeds in crinoidal grainstones have been recorded from a single locality. Ripple cross-lamination has been observed rarely in calcisiltites of this superfacies. Plane lamination is very rarely preserved in mudstone and wackestone, suggesting the presence of an active burrowing infauna.

5.2.3.3 Body Fossils

Of all the superfacies, the proximal foreslope contains the most diverse and abundant preserved biota. A minimum generic diversity of 63 can be tabulated, using detailed taxonomic investigation of only select collections of solitary rugose corals (W.G. Parkins, Univ. of Ottawa), articulate brachiopods (B. Jones, Univ. of Alberta) and gastropods (D. Rohr, Sul Ross State Univ.). Fragmentary remains are predominantly of crinoids and solenoporid algae (PL. 5-4d). Articulate brachiopods, and in particular the atrypoids and atrypids (PL. 5-4a) predominate in thanatocoenoses of relatively whole body fossils. A diverse coral assemblage, including favositids, syringoporids, alveolitids, cateniporids, proporids, and solitary and fasciculate rugose corals is found in the diverse community mud mounds and fasciculate rugose coral biostromes of the proximal foreslope. Lamellar, hemispherical and dendroid stromatoporoids occur insitu or as allochthonous bioclasts (PL. 5-4h) in various biocumules of the proximal foreslope superfacies.

5.2.3.4 Trace Fossils

Bioclastic wackestone and packstone contain a diverse assemblage of trace fossils including such eurybathic forms as Skolithos, Palaeophycus, and Chondrites. the characteristic ichnogenus in this superfacies is Zoophycos (PL. 5-4b).

5.2.3.5 Other Considerations

Although much of the proximal foreslope superfacies may have been deposited above wave base, wave-generated structures have been rarely preserved within the superfacies. The high degree of breakage and disarticulation of skeletal material in the bioclastic limestones is suggestive of reworking and transport, but much of the coarser crinoidal material may simply be the result of spontaneous disarticulation and limited transport from nearby biocoenoses. Pods of lime mudstone bound by rhizoidal holdfasts occur within massive crinoidal rudstone and packstone (PL. 5-4f); these are interpreted as in-situ nests of crinoids that contributed substantial volumes of skeletal material to the superfacies (see Paper 4, this thesis).

5.2.4 Shelf Edge Superfacies

The shelf edge superfacies collectively embraces all those facies formed in proximity to a "shallow-water" shelf-slope break separating open marine foreslope from restricted shelf lagoon. The shelf edge superfacies is invariably dolomitized, often intensely, thereby obscuring depositional textures.

Shelf edge facies apparently varied laterally along the platform margin, reflecting such variables as local subsidence rates and wave energy. As a consequence the shelf edge superfacies is not typified by any particular

lithofacies, but is represented by several, including dolomitized crinoidal grainstone, and stromatoporoid framestone, inertiastone, and bindstone.

5.2.4.1 Characteristic Lithofacies

The most commonly observed lithofacies of the shelf edge is a medium to coarsely crystalline vuggy dolostone with enlarged moldic porosity. Molds and casts of stromatoporoids are common but indifferently preserved. Molds of crinoid ossicles are also common, but the degree to which the original sediment was either crinoidal or stromatoporoid-rich is uncertain.

Interbedded with this predominant lithology are stratiform units of other severely dolomitized but commonly less porous carbonates; together, these units form the dolomitized marine skeletal (?crinoid) biocumules (PL. 5-5a,b,c,d). These biocumules are bounded laterally by foreslope facies and can have non-biostromal height/length ratios (see Paper 4, this thesis).

Dolomitized cross-bedded crinoidal grainstone (PL. 5-5g) occurs at the shelf edge, but also in extensive sheets that formed across the platform during major, high energy transgressive periods.

Stachyodes and lamellar stromatoporoid boundstone occur in shallow portions of the proximal foreslope and in shelf edge superfacies (PL. 5-5f).

Several zoned biocumules separate foreslope from restricted lagoonal facies, and are therefore considered to be barrier reefs. The best exposed of these occurs in section 22 (PL. 5-5e), where the principal reef growth stages consist of a basal massive tabulate coral-fasciculate rugosan inertiasstone to framestone (forereef), and an upper tabular stromatoporoid framestone (reef crest). The framestone is capped by amphiporid packstone (reef flat) in turn overlain by restricted lagoon lime mudstones.

5.2.4.2 Structures

The dolomitized massive skeletal (?crinoid) biocumules contain a series of parallel discontinuity surfaces (PL. 5-5b) that produce a layering that could be considered massive bedding. A crude "organic" bedding is imparted by lamellar stromatoporoids in the biostromes they constructed. Planar (predominantly) and trough cross beds are present. There is some indication of larger bed forms (megaripples, dunes?) in the lenticular bedded deposits of crinoidal dolarenite figured in PL. 5-5g.

5.2.4.3 Body Fossils

Crinoidal debris is ubiquitous. Hemispherical, tabular, irregular, and lamellar stromatoporoids, massive to irregular tabulates, fasciculate rugosans, and solenoporid algae are common. Stachyodes, cerioid rugosans, pentamerid brachiopods and mucophyllids are rare.

5.2.4.4 Trace Fossils

No trace fossils have been recorded from this superfacies. Their paucity can be attributed chiefly to their poor preservation potential in severely dolomitized rocks.

5.2.4.5 Other Considerations

The dolomitized massive skeletal (?crinoid) biocumules (DMSB) are difficult to interpret; the reader is referred to Paper 4 (this thesis) for a more thorough discussion of this facies. Hurst (1981) has interpreted what appears to be a similar rock type, as a (back-reef) skeletal sand facies. Sodero and Hobson (1979) considered massive vuggy dolomites to represent a shelf edge facies. The DMSB of the Barlow Inlet Formation represent, in fact, the intense dolomitization of an assemblage of superimposed facies that range from proximal lagoon to proximal foreslope.

5.2.5 Proximal Back-barrier Superfacies

The proximal back-barrier superfacies is transitional between shelf edge and restricted shelf lagoon superfacies, and hence its characteristics are largely intermediate between these two. The proximal back-barrier superfacies is typified by rubbly-weathering, partially dolomitized, muddy inertiastone, bafflestone, packstone, and floatstone.

5.2.5.1 Characteristic Lithofacies

The characteristic lithofacies is a rubbly-weathering, massive, domical to subspherical stromatoporoid-dominated biostromal inertialstone of sparse (predominantly) to dense structure (PL. 5-6a,b). The matrix is commonly dolomitized wackestone to mudstone (PL. 5-6c). These biostromes have been described in greater detail elsewhere in this thesis (Papers 3,4). Other lithofacies include fasciculate rugose coral bafflestone (PL. 5-6d,e) and amphiporid floatstone (PL. 5-6f). Amphiporid packstone occurs immediately above tabular stromatoporoid framestone in a zoned biocumule at section 20 (the biocumule is interpreted as a barrier reef -- Paper 4, this thesis). The amphiporid packstone is considered to represent a back-reef sand flat.

5.2.5.2 Structures

Bedding is ill-defined in the back-barrier superfacies, and the wavy to nodular appearance of this lithofacies (e.g. PL. 5-6b) is a function of seam solution more than primary compositional layering.

5.2.5.3 Body Fossils

Biotic diversity is generally low, although skeletal abundance is locally high (e.g. PL. 5-6f). Massive stromatoporoids, amphiporids, and solitary and fasciculate rugose corals are the most conspicuous faunal elements.

Other fossils common in the proximal back-barrier superfacies include oncoids, massive favositids, Coenites, ostracodes, and gastropods. Crinoid debris is rare.

5.2.5.4 Trace Fossils

None recorded.

5.2.5.5 Other Considerations

Crinoidal facies are evidently transitional to the proximal back-barrier in the absence of shelf edge biocumules (e.g. sections 4 and 5, where crossbedded crinoidal calcarenites form the shelf edge), but where barrier reefs are present (e.g. section 22), crinoids are almost entirely absent from the back-barrier and lagoon superfacies. Thus, unlike Halimeda, whose fragments cover much of the proximal foreslope and reef flats of Holocene carbonate complexes (e.g. see fig. 15 of James, 1983), and whose thalli grow abundantly in both lagoonal and fore-reef zones, the late Silurian crinoids of the Barlow Inlet Platform were important benthos and sediment types only at, and below the shelf edge.

Dolomitization within the proximal back-barrier is ubiquitous, but never as pervasive as within the shelf edge facies.

5.2.6 Restricted Shelf Lagoon Superfacies

The restricted shelf lagoon superfacies includes those lithofacies deposited subaqueously in the infratidal and subtidal zones in a position landward of the shelf edge or proximal back-barrier zone, and seaward of the strandline (i.e. mean low tide mark). This superfacies is typified by medium-bedded to structureless, generally sparsely-fossiliferous mottled carbonates and massive calcilutite mudstone.

5.2.6.1 Characteristic Lithofacies

In the TF hemicycles, the predominant subtidal lithofacies is a poorly fossiliferous mottled dolomitic limestone. Types I and II and labyrinthine patterns are all present (see Jones et al., 1979), but type I and labyrinthine are the most common. The degree of contrast between mottle and matrix, the size and shape of mottles, and the degree of seam solution affecting the mottle/matrix contact, are all highly variable characteristics (PLS. 5-7a, 5-16a). The principal subtidal lithofacies of the RI hemicycles is dense unfossiliferous medium to dark purple-grey calcilutite mudstone (PL. 5-7e). Other lithofacies include oncoidal mottled carbonate, mottled dolomitic wackestone (PL. 5-7c), massive and faintly mottled dolo/calcsiltites (PL. 5-7d), and megalodont bivalve floatstone and inertiasstone.

5.2.6.2 Structures

Bedding is generally uneven, medium to thick to massive. Lamination is not preserved although disruption of formerly laminated sediment is in part responsible for some of the unfossiliferous, massive faintly mottled dolo/calcsiltite. Fenestral fabric is developed in many of the lithofacies but its environmental significance remains controversial. Synaeresis cracks (PL. 5-10e) are locally abundant in some massive dolo/calcsiltites. Small, shallow swales or channels are overlain by oncoidal carbonates, suggesting increased associated water agitation (PL. 5-13d, unit 2). Omission surfaces are rare, but important, structures within the superfacies (PL. 5-7f), and are commonly capped by intraclastic skeletal packstone.

5.2.6.3 Body Fossils

The predominant body fossils in this superfacies include: ostracodes, dendroid stromatoporoids, gastropods, massive generally hemispherical to subspherical stromatoporoids, megalodont and other indeterminate bivalves, solitary (predominantly) and fasciculate rugose corals, brachiopods (largely smooth spiriferids), and auloporid corals. Dendroid tabulates, massive tabulates (Favosites), and crinoid fragments are relatively rare.

5.2.6.4 Trace Fossils

The subtidal and infratidal zones are characterized by two dwelling burrows, Pelecypodichnus sp. B. (PL. 5-7g), and Skolithos sp. (PL. 5-7d). Planolites and Chondrites are also present.

The undolomitized massive calcilutite mudstone contains no discrete ichnofossils. Its structureless nature could be attributed to very thorough mixing of the fine-grained sediment by a soft-bodied burrowing infauna. However, the absence of fecal material and remnant structure in the rock remains puzzling. Perhaps the homogeneity of this lithofacies merely reflects continuity of deposition.

5.2.6.5 Other Considerations

Synsedimentary submarine lithification clearly played an important role in the diagenetic evolution of these lithofacies as witnessed by the abundance of mottled dolomitic carbonates, the preservation of fenestral structure, and the occurrence of omission surfaces (hardgrounds). Mottled lithologies occur in nearly all superfacies, but nowhere as prominently as in the restricted shelf lagoon superfacies, and nowhere is the dolomitization as pervasive in mottled lithologies, as in this superfacies. Jones et al. (1979, p.246) suggested that the Mg for dolomitization "...may have been derived from the overlying seawater especially in areas of restricted circulation with

a high initial Mg/Ca ratio." The interpreted restricted lagoonal setting for these mottled dolomitic limestones, appears to lend credence to this hypothesis. Nevertheless the mottling remains a superimposed diagenetic fabric with as yet uncertain environmental affinities.

5.2.7 Tidal Flat Superfacies

The tidal flat superfacies incorporates all lithofacies deposited above low tide mark within the platform interior. This superfacies is typified by dolomitized cryptalgal laminites, and structures suggestive of periodic subaerial emergence.

5.2.7.1 Characteristic Lithofacies

The characteristic lithofacies is plane to gently wavy laminated unfossiliferous dolo/calcisiltite (PL. 5-8a). The presence of laminoid fenestrae (PL. 5-8c), alternation of compositional laminae, thickness constancy over irregular substrates, lack of traction current structures, surface ornamentation (PL. 5-8g), and desiccation polygons (PL. 5-8d), all suggest that these laminated sediments are algal in origin (hence cryptalgal laminites). Other lithofacies considered within this superfacies include massive calci/dolosiltites with desiccation polygons (PL. 5-8e), and skeletal rudstones with channelized bases (PL. 5-8b) interpreted as sheet-like storm lags. Massive fenestral

calcilutite conceivably was deposited in the intertidal or supratidal zones. Wavy bedded, ripple-laminated calcisiltite intercalated with shale (PL. 5-8f), is rare, but strongly reminiscent of wavy bedding seen in siliciclastic tidal mud flats (Reineck and Singh, 1980).

5.2.7.2 Structures

Lamination (thin laminated stratification -- Ingram, 1954) is the predominant structure. Bedding is generally medium to massive. Desiccation polygons are relatively rare, as are wavy bedding and wavy ripple lamination. Laminoid fenestrae are common; irregular to tubular forms are less so. Channel structures and discontinuity surfaces are rare.

5.2.7.3 Body Fossils

Leperditid ostracodes, small and large murchisonid gastropods, and ostracoderm debris are predominant in a generally extremely impoverished shelly fauna. Local concentrations can generally be attributed to storm winnowing.

5.2.7.4 Trace Fossils

Planolites, Sinusites, Skolithos, Uchirites, and Arthraria are relatively rare ichnogenera of the tidal flat superfacies. Chondrites is the most common trace fossil in

these rocks, but the most conspicuous (and diagnostic) is Polarichnus garnierensis (PL. 5-8d).

5.2.7.5 Other Considerations

Placing lithofacies within the tidal flat superfacies presupposes that rigid criteria exist enabling one to differentiate between infratidal, low intertidal and mid-intertidal depositional environments. Unfortunately most criteria are dependent upon factors other than relative emergence.¹¹ Furthermore over longer periods of time, infratidal deposits commonly do experience periods of subaerial exposure (e.g. 18.6 year cycle described by Wells and Coleman, 1981). Finally, the superimposition of a diagenetic fabric can obfuscate features diagnostic of the original environment of deposition.

The rationale employed in this study is admittedly assailable, but lends itself to more consistent interpretations than other approaches. The critical assumption is that algal mats probably colonized lower intertidal substrates in the Silurian (see James, 1977). The cryptalgal laminites do not contain evidence of supratidal deposition (coeval formations do have sabkha-like

¹¹ this has been addressed by the creation of a sliding scale of criteria reflecting not tidal zones, per se, but rather the likelihood of encountering particular structures or phenomena, given the proportion of time (averaged over several years) a depositional surface has been emergent (see Hardie, 1977). Unfortunately this system has not received wide acceptance by carbonate workers.

evaporites). Other lithofacies within the shallowing-upward cycles have been interpreted in compliance with this assumption (notwithstanding major discrepancies in environmentally diagnostic evidence).

5.3 CORRELATION AND DEVELOPMENTAL HISTORY OF PLATFORM

5.3.1 Physical Correlation Between Study Localities

FIG. 5-8 presents the interpreted physical correlation between study localities and individual sections within the principal outcrop belt of the Barlow Inlet Formation on eastern Cornwallis Island. Correlation of the lithostratigraphy has been attempted at the superfacies level only. Even with this degree of generalization, major problems and discrepancies remain. Complexities arise, in part, due to the obliquity of the line of section with depositional strike (i.e. platform margin), and variance in subsidence rates across the study area.

South of section 20, only the lower 350 m (at most) of the formation is exposed but this basal interval shows the greatest lateral facies variability. As a consequence FIG. 5-8 has been scaled differently for parts of the facies mosaic north of, and including, section 20 (20 versus 40 times vertical exaggeration for sections to the south).

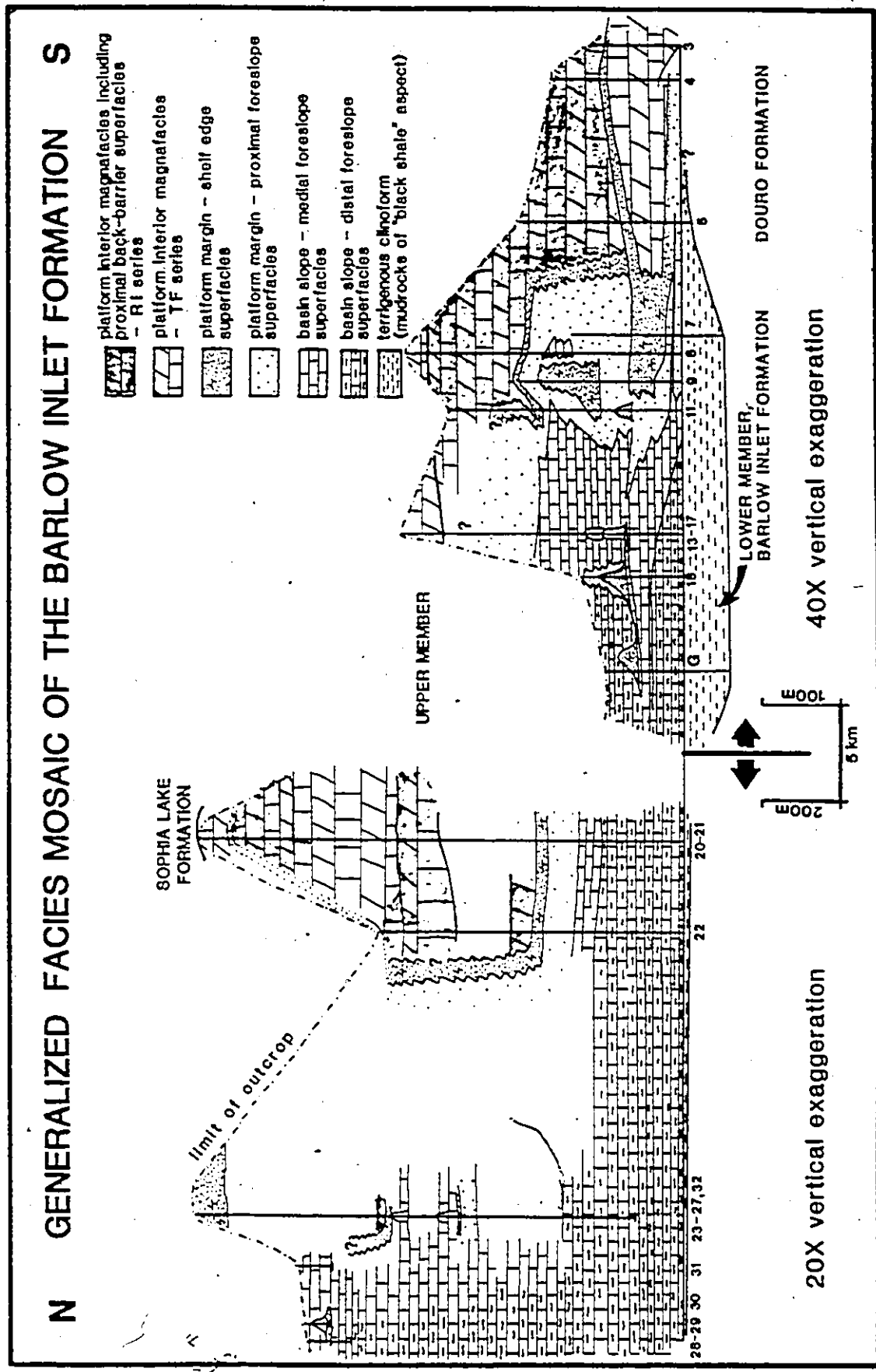


FIG. 5-8 Facies mosaic for the upper member, Barlow Inlet Formation.

5.3.1.1 Critical Assumptions

1. Structural simplicity. The outcrop belt is part of a simple homoclinal sequence, unaffected by major faults that might substantially affect correlation. Minor tear and normal faults are common at many localities but observed displacements are minimal. Exceptions to this may exist in the Cape Rescue and Cape Hotham areas where vertical airphotos show conspicuous lineaments. cursory examination of some of these features on the ground has led the author to believe that major offsets are not involved.
2. Presence of shelf rim on platform. The cyclic peritidal deposits of the Barlow Inlet Formation include thick massive calcilutite mudstones interpreted to have been deposited infratidally in a restricted lagoon. Implicit in this interpretation is the presence of a barrier restricting open marine circulation. Such shelf edge facies are therefore represented schematically between sections that contain proximal or medial foreslope superfacies and Platform Interior Magnafacies at approximately the same stratigraphic levels (FIG. 5-8).

5.3.1.2 Problems in Correlation, Rationalized and Unrationalized:

1. The lower member is not exposed in sections 4 and 5. In the absence of this valuable datum, basal parts of

section 4 are correlated to section 5, and sections further north, by the presence of a) similar brachiopod-crinoid biofacies below diverse community mud mounds, and b) lamellar-dendroid stromatoporoid and rugose coral biofacies lying above the DCM.

2. The Platform Interior Magnafacies of TF aspect is thinner at locality 5 than at locality 4. Variations in hemicycle thickness (see Paper 3, this thesis) would suggest that subsidence increased to the north, and therefore one would anticipate a thicker not thinner record, particularly if the TF-RI transition is considered a synchronous pan-platform event (see part 5.3.2.4, Platform Margin Migration and Platform Events, below).
3. In section 8 proximal foreslope facies are overlain by proximal back-barrier facies with no apparent intervening shelf edge. A rapid forestepping event is envisaged, and the condensed interval is correlated with shelf edge biocumules in section 11.
4. A problem similar to (3) is associated with sections 16 and 17; again rapid forestepping is called upon to explain the missing shelf edge facies.
5. Isolated stromatoporoid-crinoid rubble skeletal buildups in sections 18 and 19 (proximal foreslope) occur in the approximately equivalent stratigraphic interval to, but further north than, medial foreslope

sponge mud mounds in sections 14 to 16. In general for any given stratigraphic interval deeper water facies tend to occur to the north, but the occurrence of medial foreslope facies south of proximal foreslope facies probably can be attributed to the obliquity to depositional strike of the line of section, and the postulated presence of a platform margin re-entrant at sections 14 to 16. At even gentle slopes of $1/2$ degree, considerable bathymetric differences can be achieved over the lateral distances involved (e.g. 7.2 km at $1/2$ degree = 62 metres elevation difference). In the context of this problem it should also be noted that the lower-upper member contact in section 19 is quite diffuse so interval heights above base are subject to interpretation.

6. Correlation is made between the first biocumule facies seen in sections 20 and 22, although the character of these facies is somewhat different (see above). However, differences may be exaggerated by the predominance of bedding plane exposure in section 20, versus vertical section exposure in section 22, and the correlation is considered essentially sound.
7. In the absence of intervening shelf edge facies, strata of open marine aspect (crinoidal wackestones) overlie thick intervals of structureless, massive

calclutite in sections 20, 21, and 22. This relatively thin crinoid-rich interval (<10 m) is used here as a correlative unit. However, the crinoidal beds of section 22 occur 406 m beneath equivalent beds in section 21. The lower contact of the Sophia Lake Formation also occurs substantially lower in section 22 than in section 21. Graphic determinations also support a large differential in thickness. Possibly the platform area in the vicinity of sections 20/21 subsided appreciably faster than did the region encompassing section 22 (see Paper 3, this thesis). The accelerated subsidence may have occurred late in the history of the platform, as other stratigraphic intervals lower in the formation correspond well between these sections without substantial differences in stratigraphic position above the datum.

8. Extensive covered intervals with a corresponding paucity of structural attitude information, and the absence of the lower member within the Cape Rescue region, has made the stratigraphic placement of sections tenuous at best. Heights given are based on extrapolation of the lower-upper members contact, with the assumption that the relative positions of the studied stratigraphic intervals within the sections have not been structurally affected.

5.3.2 Platform Design and Evolution

5.3.2.1 Position of the Platform Margin

Conventionally, the Silurian shelf-edge on Cornwallis Island is inferred to coincide with the mapped position of the Cape Airy - Snowblind Bay line of facies change which separates "shelly" facies to the south from "graptolitic" facies to the north. Work by Sodero and Hobson (1979) would suggest that from Llandovery to Ludlow time, the CASBLFC does closely approximate to a relatively fixed position for the shelf-edge for the Allen Bay and Cape Storm Formations. Both Thorsteinsson (1980, p.21) and Muir and Rust (1982) have suggested that this facies change, at least in the vicinity of Snowblind Bay, is structurally controlled. Unfortunately the static, mapped position of the facies change cannot be applied except in a most general fashion, to the shelf-slope break of the Barlow Inlet Formation, or indeed the ramp-basin transition of the underlying Douro Formation. Basin slope facies of the Douro Formation are recognized as far south as Cape Hotham (southeast corner of Cornwallis Island) and Griffith Island, and shelf facies of the Barlow Inlet Formation occur as far north as Cape Rescue. This implies that from late Ludlow to late Pridoli time, basin and basin slope facies were progressively excluded from more southerly parts of what is now Cornwallis Island and the position of the shelf edge was in a state of flux. In this context the CASBLFC only reflects in post

mid-Ludlow time, the northernmost extent of observed shelf facies in the study area.

5.3.2.2 Orientation of Platform Margin

The very linear north-south trending belt of Barlow Inlet outcrop on Cornwallis Island effectively yields but a 2-dimensional transect across the platform margin, and therefore is not ideally suited for purposes of determining depositional strike of the shelf edge facies. In a broad sense, interpreted basinal facies proportionately increase to the north, and within the Barlow Inlet Formation are absent from exposures at study localities C and D in the south. However these generalities do not serve to elucidate depositional strike of the platform margin. Arguments are presented below promoting the hypothesis that south of loc. H, the shelf edge was oriented closer to north-south than east-west:

1. the presumed structure of the Cornwallis Fold Belt (Kerr, 1977, 1981), suggests that easterly dipping paleoslopes would develop on the flanks of homoclines generated by block faulting at depth, along bounding faults of the Boothia Horst.²²

²² the present eastern coast of Cornwallis Island may well be fault-controlled (Thorsteinsson, 1958). Although presumably an Eureka structure, it may represent movement on an older fault or line of structural weakness. Certainly north-south trending structures within the Cornwallis Fold Belt are common.

2. regional paleogeographic reconstructions by Miall and Gibling (1979) for Pridolian age sedimentary environments in the vicinity of the Boothia Axial Landmass, suggest that continental and near-shore facies were developed in parallel, north-south trending belts on either side of the B.A.L. Presumably marine facies would likewise be principally aligned with the N-S axis of the B.A.L.
3. limited slip line determinations of slide structures (see Woodcock, 1979) indicate easterly-dipping paleoslopes at loc. E.
4. the distribution of facies at loc. E, is more easily interpreted in terms of a N-S depositional strike than one east-west, as basin slope facies laterally equivalent to the principal dolomitized massive skeletal (?crinoid) biocumule, are situated both north and south of an interpreted shelf-edge reef (FIG. 5-8).

5.3.2.3 Origin of the Platform Margin

The Douro Formation of late Ludlovian age, was laid down on a largely undifferentiated carbonate ramp (Paper 2, this thesis). The lower member of the Barlow Inlet Formation represents the marine depositional product of continental mass-wasting. This synrogenic deltaic sedimentation was probably related to, if not directly associated with, block

faulting of the Boothia Horst (Paper 2, this thesis). In contrast to the Douro Formation the upper member of the Barlow Inlet Formation was clearly deposited on a highly differentiated carbonate shelf (see Papers 3 and 4, this thesis). Moreover differentiation of the shelf was apparently complete at a very early stage in the evolution of the platform (certainly by latest Ludlow time). An important problem to be addressed is, whether the ramp to platform transformation was primarily predicated on structural changes, or on biogenic productivity and construction (e.g. Wilson, 1975, p.21).

The available evidence suggests that, although organic accretion at the shelf edge was an important adjunct in shelf differentiation, the site of the platform margin was predetermined by extrinsic (tectonic) means. The following observations support a structural origin in the development of a differentiated shelf:

1. Table 5-5 presents for the upper member the heights above base at which the first biocumules (diverse community mud mounds) and first significant crinoidal blanket deposits (shelf edge shoals) occur. These figures range from 4.9 to 24.8 m. Above these levels (at locs. C and D) is the first record of restricted lagoonal sedimentation, implying the existence of a rimmed (and hence differentiated) shelf. These rather minimal thicknesses are

Table 5-5

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First Stratigraphic Occurrence of Crinoid blanket
deposits and/or diverse community mud mounds in
Barlow Inlet Formation.

	LOCALITY		
	C	D	E
Height above base (in metres) of first occurrence of diverse community mud mounds	4.9	14.1	16.0
Height above base (in metres) of first thick interval of crinoidal packstone/grainstone	8.4	19.8	24.8

considered to be the product of deposition during, and following, the basal transgressive event of the upper member. The lithofacies conceivably, although not necessarily, represent open-marine, pre-differentiated shelf sedimentation. Nevertheless, given the high rates of deposition in the Franklinian Geocline, these basal intervals probably do not represent a significant time period. Given that the Douro Ramp did not evolve into a rimmed shelf (despite the time involved in depositing up to 450 m of sediments) during a period of relative tectonic quiescence, it is unlikely that an organic accretionary model could primarily account for the 'rapid' shelf differentiation seen in the Barlow Inlet Formation, following the basal transgression,

2. Krebs and Mountjoy (1972) listed 4 criteria useful in determining whether or not ancient reef growth is related to synsedimentary ~~faulting~~.²² These criteria are as follows:

- a) linear reef trends
- b) reef trends parallel to known faults
- c) abrupt facies changes
- d) Pb-Zn mineralization in buildups.

²² there is no evidence of submarine fault scarps (i.e. by-pass margins) during deposition of the Barlow Inlet Formation. Synsedimentary faulting at depth is envisaged to effect subsidence at surface, and hence produce large scale modifications of pre-existing ramp morphology.

Ludlovian to Pridolian shelf edge biocumules occur from loc. D to loc. J in a N-S trend that parallels known and presumed faults in the Cornwallis Fold belt and Boothia Horst. The facies changes associated with Platform biocumules and shoals are in all cases abrupt, and minor Pb-Zn and marcasite mineralization (a common gangue mineral in Pb-Zn deposits) is associated with buildups in the Cape Rescue - Snowblind Bay area.

In summary, the inherited late Ludlovian seafloor topography in the vicinity of the Boothia Uplift, probably contained a pronounced shelf-slope break, produced through epeirogenesis (Pulse 2 of the Cornwallis Disturbance). During the initial deposition of the upper member, this shelf-slope break was further enhanced by continued post-epeirogenic subsidence of marginal fault blocks and organic accretion at preferred sites along the structurally-initiated break in slope.

5.3.2.4 Platform Margin Migration and Platform Events

The thickness of the interpreted shelf edge facies, at different localities, varies considerably. This variation is seen not only in maximum continuous shelf edge thicknesses, but also cumulative shelf edge facies thickness (FIG. 5-9). Given that depositional or accretionary rates of 'reefoid' margins likely

THICKNESS DISTRIBUTION OF SHELF EDGE SUPERFACIES

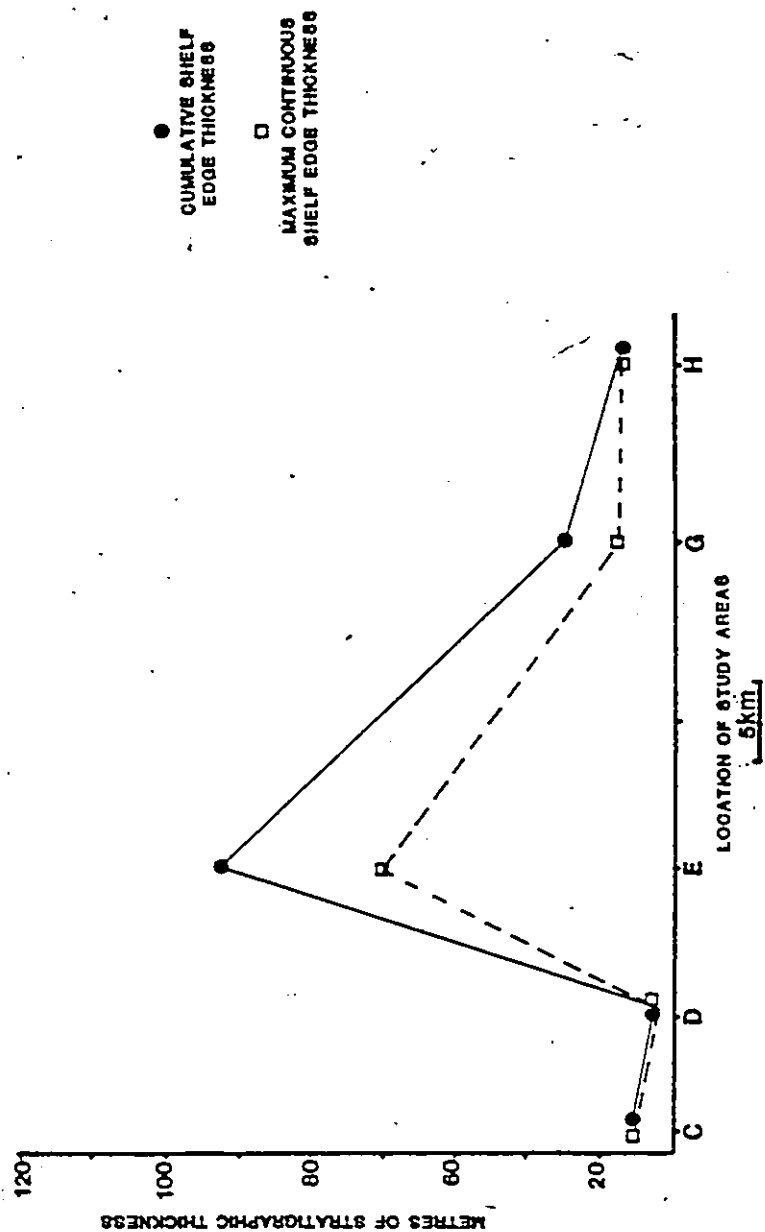


FIG. 5-7 Thickness distribution of shelf edge superfacies.

did not vary significantly from position to position on the platform, thickness differences must be accounted for by differential subsidence rates. It is reasonable to assume that where subsidence rates are high, shelf edge facies tend to build up vertically (hence achieve greater thicknesses), whereas at platform positions with low subsidence rates, lateral outbuilding will predominate, yielding a thinner record of the shelf edge (see Kendall and Schlager, 1981). FIG. 5-10 presents a model of the Barlow Inlet Platform that seeks to accommodate both basement fault block control on platform morphology, and observed thickness trends of shelf edge facies. Progradation was most rapid where subsidence was less, whereas progradation virtually ceased at fault-controlled shelf-slope breaks in areas of high relative subsidence.

Evidence presented elsewhere (Paper 3, this thesis) suggests that major platform forestepping events accompanied reduced subsidence (and possibly mild uplift) on basement fault blocks underlying the platform (see also FIG. 5-10). In this context overall enlargement of the platform area, and the progradation of platform margin and interior facies northward is seen as being both progressive and episodic. Outbuilding of the platform concomitant with basin filling was an ongoing and intrinsic process episodically accelerated by epeirogenic activity (platform events, see below) affecting the Boothia Uplift.

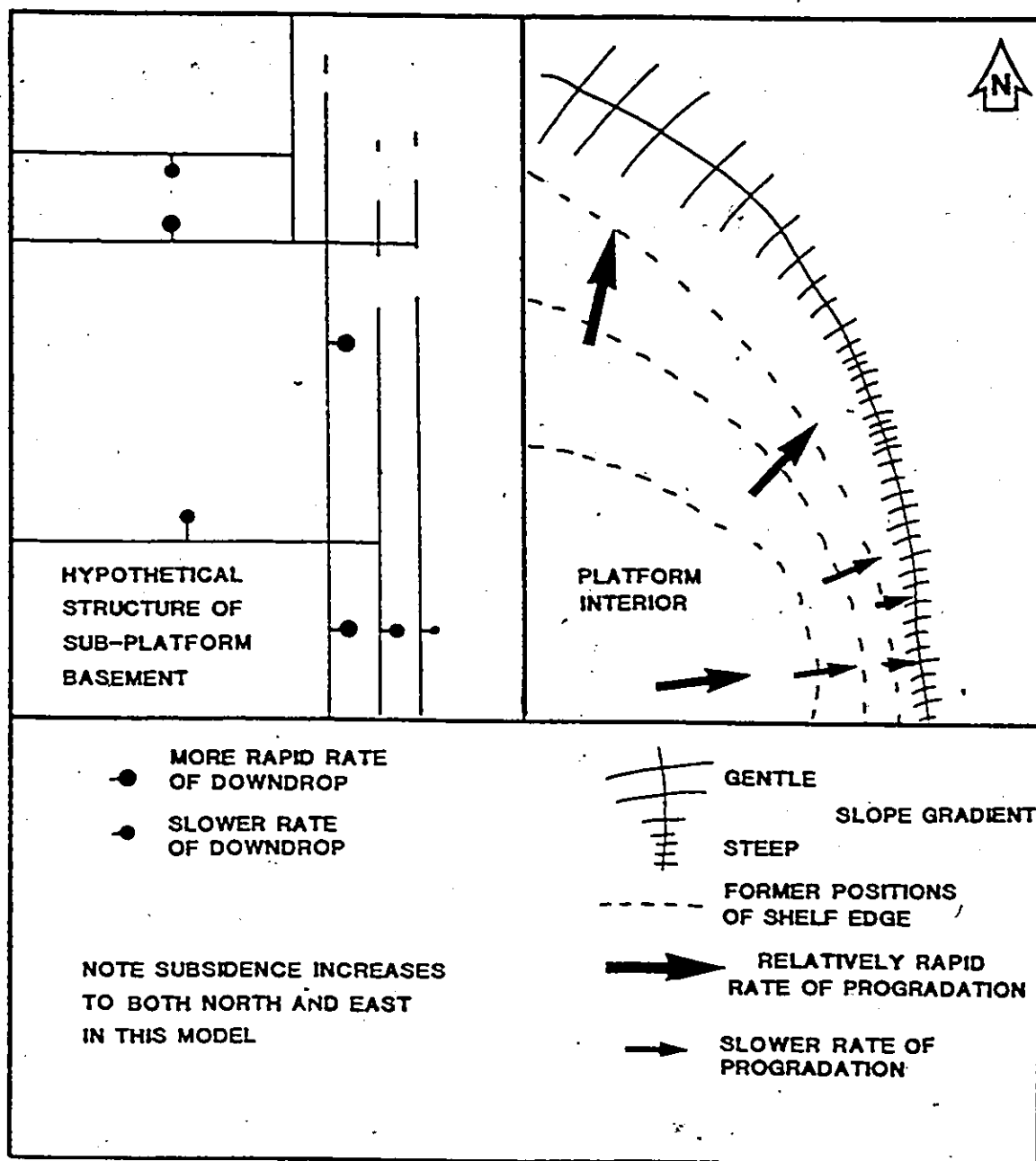


FIG. 5-10 Platform margin migration.

Within the Barlow Inlet Platform succession, there are a number of major, anomalous lithofacies transitions that are generally abrupt, and cannot be predicted based on transition tendencies in subjacent sequences. Given continuity of exposure, similar anomalous transitions are seen to occur at approximately the same stratigraphic positions in adjacent sections. Such lithofacies transitions are considered to represent significant 'events' in the overall evolution of the platform. These events involved major periods of forestepping (regression) and backstepping (transgression) and likely affected widespread portions of the platform simultaneously.²⁴ Table 5-6 lists and ranks the 9 major events presently identified, and FIG. 5-11 graphically depicts their position on the facies mosaic of the Barlow Inlet Formation. It is speculated that the origin of most, if not all, of these platform events lay in spasmodic episodes of more intense block faulting of the Boothia Horst. Corroborative evidence for episodic block fault activity beneath the platform is suggested as well by the stratigraphic first occurrence in different sections of cryptalgal laminites (see Paper 3, this thesis). The principal result of these events was the enlargement (or reduction) of the platform area.

²⁴ pending detailed biostratigraphic work, synchronicity of similar events between sections remains untested.

Events affecting the Barlow Inlet Platform

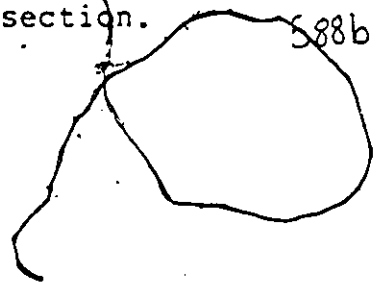
EVENT NUMBER	TRANSGRESSION OR REGRESSION	RANK ¹	EVENT DESCRIPTION
1.	T	B	Basal transgressive interval of upper member. Open marine carbonates directly overlie subarkosic beach shoreface facies of lower member.
2.	?T	B	Sparsely fossiliferous marlstones abruptly overlie diverse community mud mounds in sections 5, 8 and 9 and are associated with termination of mound growth.
3.	T	B	a) Transition from TF ² to RI series in sections 4 and 5. b) Biocumule drowning in sections 8 and 9. c) Biocumule drowning at loc. G.
4.	R	B	a) Transition from foreslope to lagoonal facies in the apparent absence of shelf edge facies in section 8. b) Transition from foreslope to lagoon/tidal flat facies in the apparent absence of shelf edge facies in section 17. c) Transition from foreslope to lagoonal facies with very thin intervening shelf edge facies in section 11.
5.	T	C	Transition from TF to RI hemicycle series in section 20.
6.	T	B	Occurrence of open marine (crinoidal) facies overlying lagoonal facies of RI hemicycle aspect in sections 20 and 21.
7.	R & T	C	Abrupt occurrence of desiccated, siliciclastic-rich, cryptalgal laminates overlying open-marine proximal foreslope facies in section 25, followed by abrupt return to open marine facies.
8.	R	C	Medial foreslope facies transitional to restricted lagoonal facies through a thickness of only 20m in section 27, with no significant shelf-edge facies preserved.
9.	R	B	Abrupt appearance of red beds of Sophia Lake Formation.

1.

Rankings are as follows :

- A - confirmed event horizon, present in 2 or more sections plus proof of synchronicity.
- B - probable event horizon, present in two or more sections.

C - possible event horizon, present in single section. 588b



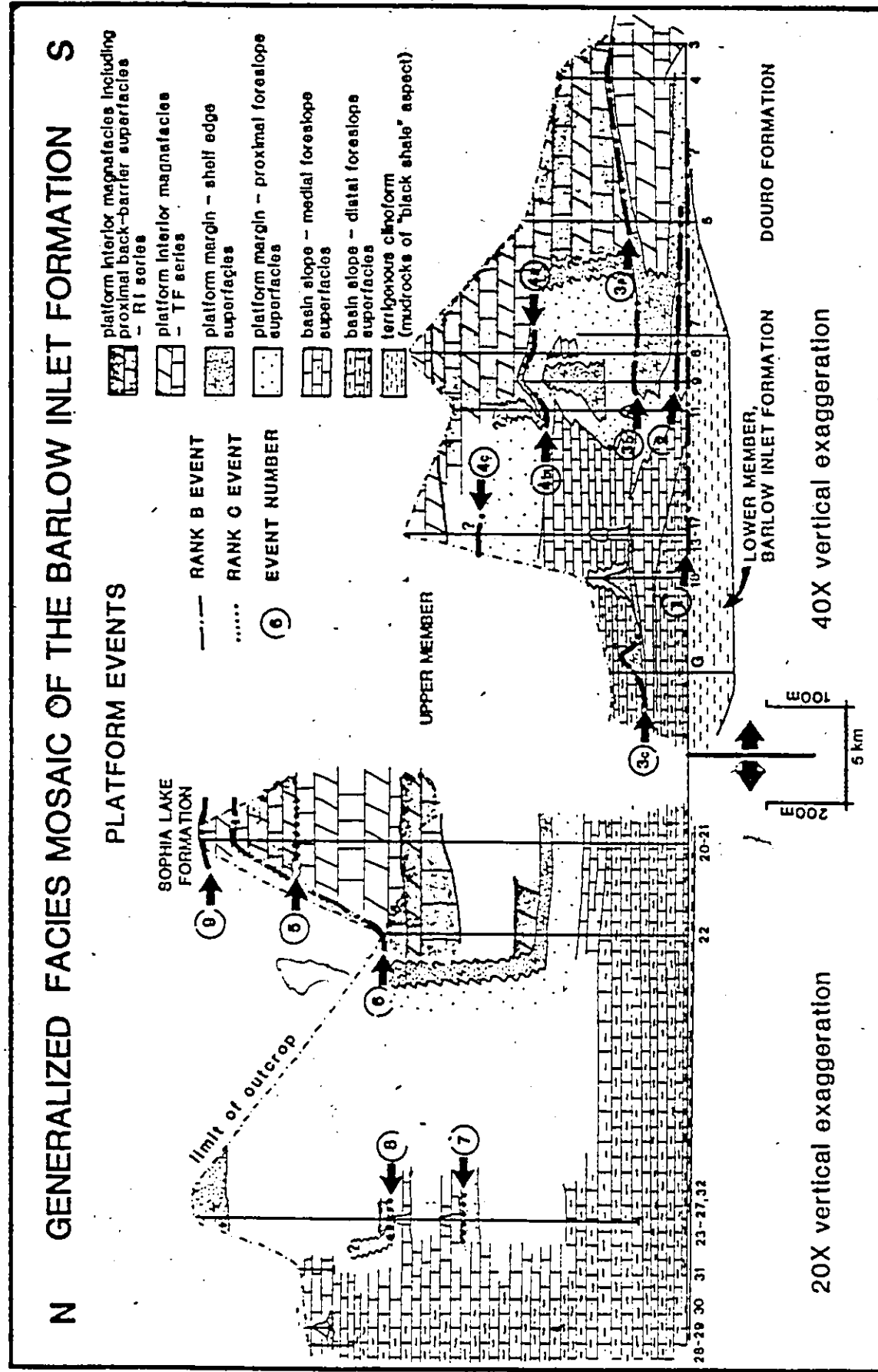


FIG. 5-11 Platform events.

5.3.2.5 Early Devonian Emergence of the Barlow Inlet Platform

Platform evolution terminated in early Lochkovian time due to elevation and emergence of much, if not all, of the subtidal shelf in the study area. That this emergence was in response to epeirogenic uplift rather than simple tidal flat accretion and outbuilding, is suggested by the following rationale:

1. In general, if sedimentation exceeds subsidence in paralic basins, then the result will be vertical and lateral accretion of a coastal wedge of sediments up to, or slightly above, base level (and hence emergence of formerly subtidal depositional sites, and areal enlargement of the contiguous landmass bordering the basin). Such outbuilding by simple sedimentary processes will, assuming eustatic constancy, both decrease average gradients within the depositional system, and further remove marine or near-marine depositional sites from the nearest upland sources of clastic detritus. As a consequence, the volume and mean size of terrigenous allochems is unlikely to increase under these circumstances. In fact quite the contrary, the siliciclastic/carbonate ratio would be expected to decrease in the tidal flat prism. In the uppermost 50 m of the Barlow Inlet Formation, a substantial quantity of quartz silt and fine sand is incorporated in apparently non-cyclic

peritidal lithofacies. Limited sampling of the overlying Sophia Lake Formation suggests that an even greater quantity of fine quartz sand occurs in this formation (PL. 5-16j). This suggests either development of a new proximally-located source, or increased relief of an original source area.

Furthermore, the abrupt but conformable contact²² between the Barlow Inlet and Sophia Lake Formations is marked by the first occurrence of dusky red colored siltstones and silty calcisiltites (PL. 5-22). Above the contact, red beds are a persistent lithofacies of the Sophia Lake Formation. Hematite staining is generally considered to represent the product of subaerial oxidation of ferromanganese minerals in an arid to semi-arid climate (e.g. Walker, 1976). The abrupt introduction of red beds suggests the tapping of a new source rich in ferromanganese minerals (e.g. crystalline basement terrain).

2. the uppermost beds of the Barlow Inlet Formation are early Lochkovian in age (Uyeno, 1977). The Snowblind Bay Formation (synorogenic molasse) that overlies the Sophia Lake Formation may be, in part, as old as

²² in the vicinity of loc. H this contact can be traced for over 10 Km, and appears to occur everywhere at approximately the same stratigraphic level (i.e. the color change marking the contact parallels regional strike).

mid-Lochkovian (Thorsteinsson, 1980). Given the close proximity in age between these two formations, it is reasonable to argue that emergence of the Barlow Inlet Platform was directly related to the diastrophism associated with Snowblind Bay sedimentation (pulse 3 of the Cornwallis Disturbance). As summarized in FIG. 5-12, the Barlow Inlet Platform increased in size by episodic fore-stepping in a northerly direction. As argued above, platform enlargement was accomplished by a combination of intrinsic basin infilling and platform margin progradation, along with extrinsic fault block uplift. Although the platform margin prograded northward (for most of the Ludlow and Pridoli), it cannot be assumed that emergence of the platform occurred as a result of progressive uplift from south to north. In fact, there is a body of evidence to suggest that final platform emergence accompanied a limited period of intense fault block uplift north of study locality G. Supporting this contention are the following arguments:

- a) The early Lochkovian Sophia Lake Formation, is reported from Baillie Hamilton Island (Thorsteinsson, 1980; Smith, 1980) some 80 km north of Read Bay. There the upper half of the formation, at least, represents marginal marine/coastal plain depositional environments

(see Gibling and Narbonne, 1977, for a discussion of the environmental significance of the Sophia Lake Formation). The underlying formations are of shallow subtidal to deep basinal origin, and together with the Sophia Lake Formation form a conformable shallowing upward succession. This shallowing trend suggests that uplift affected a broad area north of, and including, loc. H, prior to pulse 3 of the Cornwallis Disturbance

b) Muir and Rust (1983) have demonstrated that the principal transport direction in the coarse alluvial facies of the Snowblind Bay Formation was from north to south, which implies the greatest topographic relief induced through uplift occurred to the north. Older stages of this same period of diastrophism may have been responsible for the emergence of the Barlow Inlet Platform.

c) Northerly portions of the Boothia Uplift have been involved in either i) independent uplift (e.g. Pulse 1 of the Cornwallis Disturbance that affected Grinnell Peninsula -- Kerr, 1977), or ii) reduced crustal subsidence (the Cape Phillips Formation thins to the north -- Thorsteinsson in Douglas, 1970). The Barlow Inlet Formation itself may thin north of Loc. H, see Paper 3 this thesis). Both factors circumstantially point to differential fault block movement in the north, in part independent of epeirogenic activity in the south.

The emergence of the Barlow Inlet rimmed shelf complex forms part of the sequence of diastrophic events associated with pulse 3 (Pulse 2 of Kerr, 1977) of the Cornwallis Disturbance, wherein a large portion of the Franklinian Geocline was converted into part of the Arctic Platform (see Kerr, 1977, fig. 5).

5.3.2.6 The Barlow Inlet Platform defined: reasonable constraints and postulations

There remain several hitherto unanswered questions regarding the character of the Barlow Inlet Formation; some of these questions can be answered confidently, while responses to others are speculative. This section considers both regional and study-area-specific data in attempting to characterize the Barlow Inlet Platform. This is accomplished by providing plausible answers to the following questions:

1. Does the Barlow Inlet Formation represent a platform or a ramp, and if the former, was it predominantly open or rimmed?
2. Does the Barlow Inlet Formation represent an isolated platform (or series of platforms), or a single attached platform?
3. What was the nature of the platform margin?
4. How does the platform margin conform to Wilson's (1974) spectrum of shelf margin profiles?

5. Was the platform margin subject to severe, moderate or slight hydraulic stress?
6. What regional factors may have profoundly influenced platform design and development?

These questions are addressed individually below.

1) Notwithstanding that the author's conclusion to this question is emblazoned on section titles and throughout the foregoing text, differentiating between platform and ramp is not straightforward. Ambiguity arises from their definitions and use in the literature. Ahr (1973, p.222) originally defined the term ramp to mean a "... sloping topographic surface on which carbonate facies are deposited while subject to open ocean conditions from the surf zone to depths of hundreds of feet". Furthermore the ramp as defined, had the following key characteristics:

- a) grainy facies are deposited near the surf zone and adjacent mainland, and
- b) reef trends are not continuous.

Ahr (op.cit.) noted that "...in the ramp model, grainstones and packstone are landward facies and the sediments become muddy ... seaward. In the typical shelf model the landward facies are muddy. They pass seaward into shelf-margin grainstones and boundstones."

Read (1982) recently divided ramps into homoclinal and distally-steepened types. Both types have cyclic peritidal facies and offshore sand shoals. Additionally, homoclinal

ramps have lagoonal lime muds shoreward of sand shoal facies (op.cit., Fig. 1a), concepts which are, in the author's opinion, contrary to what was originally intended by the term "ramp". In describing Lower Silurian carbonates of northwest Greenland, Hurst and Surlyk (1983, p.493, 496) have used the term "ramp" and "ramp margin" (=shelf-slope break?); the latter is a label, in part, for the site of "...an enormous reef tract with reefs maintained at sea level throughout most of the Llandovery". No mention was made of the character of near-shore facies by Hurst and Surlyk (1983). Again the usage appears diametrically opposite to Ahr's (1973) ramp concept.

If current usage is applied, then the Barlow Inlet Formation could represent in part deposition on a carbonate ramp. However, in consideration of the original definition, the Barlow Inlet Formation with its interpreted shelf edge biocumule and crinoidal blanket facies, cannot be considered a ramp."

The platform had a rimmed shelf throughout much, if not all, of its history. At different times, and possibly coevally, the barrier forming the rim consisted of sand shoals together with isolated stromatoporoid inertiasone buildups, and barrier reefs.

" as discussed previously, the lowermost strata of the upper member may have been laid down on a carbonate ramp.

2) Without evidence to the contrary, the Barlow Inlet Platform is regarded as an attached platform, contiguous with the B.A.L. However, the interpreted independent movement of horst blocks, would be a favourable mechanism by which to produce isolate platforms (similar to the Bahamas). As stated by Read (1981, p.205), "...isolated platforms appear to develop above horsts on faulted, rapidly subsiding platforms of extensional continental margins", a situation that conceivably bears considerable resemblance to the late Silurian Barlow Inlet Platform of Arctic Canada.

3) McIlreath and James (1979) proposed a 4-Part classification of platform margins, consisting of 'Depositional' and 'By-Pass' margins with either shelf edge organic reefs or lime sand shoals. Read (1981) further modified this scheme by subdividing By-Pass margins into escarpment and gullied slope types and added a third category, the Erosional Margin. Extensive proximal foreslope crinoidal calcarenites, absence of distinctive periplatform talus, or of channelized deposits (gullied slope), lack of evidence for a sharp break in slope (e.g. steepened depositional gradient in association with shelf edge-proximal foreslope transition) and apparent absence of any submarine unconformity at the shelf edge, all suggest a depositional margin (or 'accretionary' margin of Read 1981). The shelf edge itself was occupied by both sand shoals and shallow-water biocumules (see (1) above).

4) Wilson (1974) described three empirical types of shelf margin profile: i) downslope lime mud accumulation, ii) knoll reef ramps or platforms, and iii) organic reef rims. None of these types is fully applicable to the Barlow Inlet Platform, and aspects of each are displayed at different paleogeographic and stratigraphic positions. Foreslope mud mounds are locally well developed in association with shelf edge sand blanket facies (type i aspect suggestive of quiet seas). Barrier reefs with no known foreslope mud mounds are developed elsewhere (type iii aspect suggestive of steep profiles and rough seas), whereas the development of inertiastone and bafflestone biocumules (as opposed to framestone composed of marine framebuilders) in association with crinoid blanket facies, is suggestive of type ii shelf margin profiles (quiet to moderate seas, gentle profiles). It is probable that the Barlow Inlet Platform margin was laterally variable, as a consequence of its orientation to prevailing winds, and its structural underpinning. Two basic profile margins are envisaged (FIG. 5-12): the first combines aspects of Wilson's types i and ii, and the second is a modified version of Wilson's type iii.

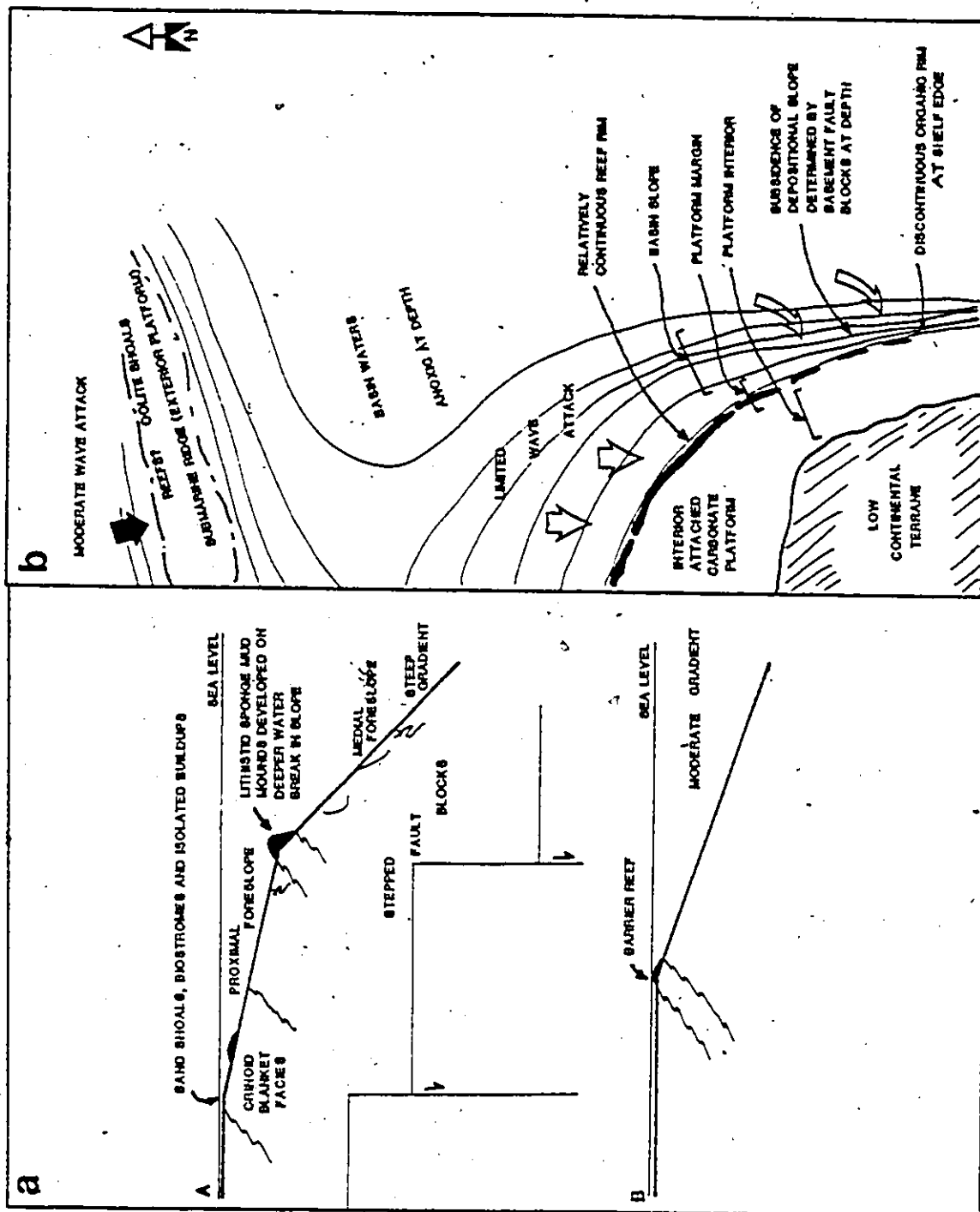


FIG. 5-12 Characteristics of the Barlow Inlet Platform.

5) The platform margin area was unlikely to have been subject to high wave stress, for the following reasons:

a) neither massive frame construction nor continuous reefs trends characterize most of the shelf edge superfacies studied.

b) Coated grains, major bar forms, and cross-bedding are absent to rare in the shelf edge crinoid blanket facies (although the paucity of cross-bedding may be a preservational phenomena).

c) Exotic (storm-dislodged?) blocks of shallow water biocumules are largely absent from the proximal and medial foreslope deposits.

d) Where barrier reefs do exist (section 22), the back-reef sand flat is limited in extent and contains no allochthonous reef-crest material (i.e. there is no dead reef pavement zone).

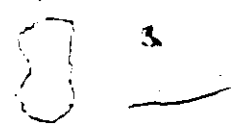
e) The presence of downslope mud mounds is a characteristic of tranquil seas (Wilson, 1974).

f) Although in part dependent on the efficiency of the shelf edge barrier, the fact that the restricted shelf lagoon and tidal flat lithofacies are virtually devoid of intraclastic deposits, or shelf edge derived sands, is suggestive of very low prevailing hydraulic energies in the lagoons and possibly limited storm surge effects.

g) Indications of water stratification in the Cape Phillips basin suggest that the prevailing winds, and currents they induced, were insufficient to mix or turn over basin water; only surface waters were thoroughly oxygenated.

6) As alluded to previously, the position and nature of the margins of the Barlow Inlet Platform were largely controlled by the bounding faults of the horst, and the degree of subsidence on individual horst blocks. However, no submarine fault scarp developed. This situation is similar to that described by Hurst (1982) for the Lower Silurian platform of north Greenland: a fault control for this margin is envisaged, but only some faults had surface expression (op.cit. p.116). The Barlow Inlet Platform margin may be a close parallel to Hurst's (op.cit.) stepped margin profile, which features downstepped fault blocks with variable dips, foreslope mud mounds (commonly as linear tracts) and isolate reef masses developed on the platform.

The occurrence of a linear trend of foreslope mud mounds at locality F (see PL. 4-1a,b, this thesis) may be related to a deeper, water shelf slope break (see FIG. 5-12). The paucity of slope failure phenomena in the proximal foreslope in contrast to their abundance in the medial foreslope, possibly reflects such a change in gradient, although alternatively, it may simply represent the different responses to shear exhibited by predominantly coarse



particulate (proximal foreslope) and muddy carbonate (medial foreslope) sediments. The presence of conspicuous carbonate breccias associated with the foreslope mud mounds (and their absence in shelf edge biocumules) also suggests heightened gradients. Debris flows were more likely generated by seismic shock than storm activity, and would have preferentially affected the least stable parts of the slope.

Major backstepping and forestepping 'platform events' are envisaged as responses to accelerated fault block movement superimposed on longer-term subsidence and progradation. Relative sea level changes are thus viewed primarily as a result of interplay between subsidence and sedimentation, with eustasy constant. Kendall and Schlager (1981) outlined three responses by carbonate platforms to relative sea-level change: a) drowning, b) catch-up, and c) keep-up. The first and last are represented in the Barlow Inlet Formation. Platform drowning is often marked by "...the presence of basal lime sands and gravels that result from migration of a high-energy transgressive environment over the platform..." (Read, 1982, p.206). The cross-bedded crinoidal sands that occur between RI and TF cycles in sections 4 and 5 (platform event 3a) represent such a deposit. Shelf margin progradation attests to the growth potential of the platform interior and rim relative to subsidence. Accelerated progradation (e.g. platform event 4) may signify mild uplift in addition.

As an external mechanism is invoked to explain the cyclicity of platform interior sediments, a mild version of the catch-up response may also be involved. Normally this response leads to development of an atoll or barrier reef with a deep lagoon. There is no evidence to suggest that the platform interior was ever submerged more than 7 m, but accelerated subsidence of the platform may have had a more adverse affect (drowning) on the platform interior than on the platform margin.

Regional factors (paleogeography, paleoclimate, supply of terrigenous materials) can have profound effects on platform design and development. The Barlow Inlet Platform bordered a deeper-water, stratified basin (that the Cape Phillips Formation represents deposition in euxinic conditions has been proposed by Kerr, 1974; Smith, 1980; Trettin, 1979; amongst others). The mechanism by which anoxia developed in the basal basin waters is not presently known (possibly density stratification due to hypersaline brines), but there is some suggestion that part if not much of the basin was silled by the Melville-Bathurst Ridge (see FIG. 5-5; see also Thorsteinsson in Douglas, 1970, Fig. X-4). The Barlow Inlet Platform would thus have been separated from the Hazen Seaway by both an intrashelf depression and an outer carbonate barrier. The Barlow Inlet Formation therefore represents an attached interior platform.

The climate in the latest Silurian was predominantly dry (Miall and Gibling, 1978), possibly arid. The evaporitic nature (Sabkha-like) of many nearby Upper Silurian formations (e.g. Cape Crauford, Leopold, Sophia Lake, and Somerset Island Formations) attests to greater evaporation rates. The study area falls within Boucot and Grey's (1979) Silurian mid-cratonic evaporitic domain. Lower Devonian paleolatitude determinations based on paleomagnetic analyses of red beds from the Somerset Island, Peel Sound and Snowblind Bay formations, place the study area close to the equator (Lapointe and Dankers, 1982).

Although tempestite deposits are rare within the Barlow Inlet Formation (e.g. PLS. 5-2c, 5-3j, 5-8b) other formations of similar age in the surrounding area contain appreciable storm-deposited material (e.g. Jones and Dixon, 1976). The perceptible impact of storms on the platform interior facies may have been lessened by the surprising resistance stabilized mud banks have to storm erosion (see Thomas et al., 1961). The prevailing paleo-wind direction for the study area is not known. Narbonne (1981) concluded that much of the fine terrigenous detritus in the underlying Douro Formation (late Ludlovian) was derived from the north, suggesting that prevailing currents ran approximately from north to south. It can be speculated that prevailing winds also moved from north to south (e.g. northeasterlies or northwesterlies).

Miall and Gibling (1978) and Miall (1983) argued that erosion and clastic sedimentation in the B.A.L. continued without apparent break from the latest Ludlow to possible mid-Devonian time. The paucity of terrigenous siliciclastic detritus in the upper member of the Barlow Inlet Formation, and the abundance and diversity of biocumules associated with the platform margin, all suggest that clastic pollution in the study area was minimal, and that either

- a) sediment dispersal remained directed westward throughout most of the Pridolian, or
- b) tidal forces effectively contained fluvial discharge and redistributed much of the terrigenous material in tidal flats and coastal plain facies marginal to the B.A.L. Equally, much of the total river load was carbonate detritus, both susceptible to solution and difficult to recognize as exotic material in a carbonate depositional system.

5.3.2.7 Synthesis

The Barlow Inlet Formation represents carbonate deposition on an attached interior platform that bordered stratified basin waters of a partially enclosed basin. Wave stress (prevailing and storm) was not intense at an accretionary platform margin variously occupied by crinoidal sand shoals, isolated skeletal buildups, and barrier reefs. The platform developed in a semi-arid to arid

climate. Its evolution and design were controlled largely by block faulting concomitant with ongoing epeirogenesis affecting the Boothia Uplift. Its origin lies in the structural modification of a pre-existing carbonate ramp. Platform drowning and platform outbuilding were probably related to accelerated fault block movement.

Two basic platform margin designs are envisaged (FIG. 5-12). Platform margins in the vicinity of localities E and F are composed of crinoid sand shoals, isolated stromatoporoid-crinoid rubble skeletal buildups, dolomitized massive skeletal (?crinoid) biocumules, and downslope lithistid mud mounds (possibly developed preferentially on a second shelf slope break). This margin type was relatively static and migrated appreciably only during platform events. The second margin type is represented at locality I, and comprises a tabulate-stromatoporoid barrier reef. Migration of the margin is viewed to have been relatively continuous with outbuilding occurring over a foreslope of moderate gradient.

Ongoing clastic sedimentation did not appear to have had any detrimental effect on carbonate productivity on the Barlow Inlet Platform, and little terrigenous material is incorporated in the carbonate sediments.

5.3.3 Regional Correlation.

Exposures of late Ludlovian to Pridolian age strata occur on most of the islands immediately surrounding Cornwallis Island, and at scattered localities west of the principal east coast outcrop belt on Cornwallis Island (FIG. 5-13). Unfortunately, outcrop is limited in some instances (e.g. Drake Bay Formation of Russell and Prince of Wales Islands). Comparisons, based on lithostratigraphic characteristics, are by necessity extremely generalized. Only in three areas (4, 8 and 10 of FIG. 5-13) are stratigraphic sequences known in adequate detail to warrant direct comparison with Barlow Inlet superfacies. Other areas contain lithofacies either not found in the Barlow Inlet Formation, or with descriptions insufficient for comparison. A detailed examination of time-equivalent strata from surrounding areas is given in Appendix C.

It is concluded that, the Barlow Inlet Formation is transitional, and in part lithologically equivalent to, other formations in the central Arctic Archipelago. In the south, the Somerset Island Formation (particularly the lower member) is strikingly similar to the TF series of the Platform Interior Magnafacies. To the north, the Cape Phillips C member is indistinguishable from the distal foreslope superfacies of the Basin Slope Magnafacies. Given the broad spectrum of facies within the Barlow Inlet Formation, these interpreted equivalencies should only be expected; a contrary situation would indeed be anomalous.

On the basis of the time-equivalent depositional environments summarized above, the disposition of generalized paleoenvironments for the Late Silurian can be depicted (FIG. 5-13).



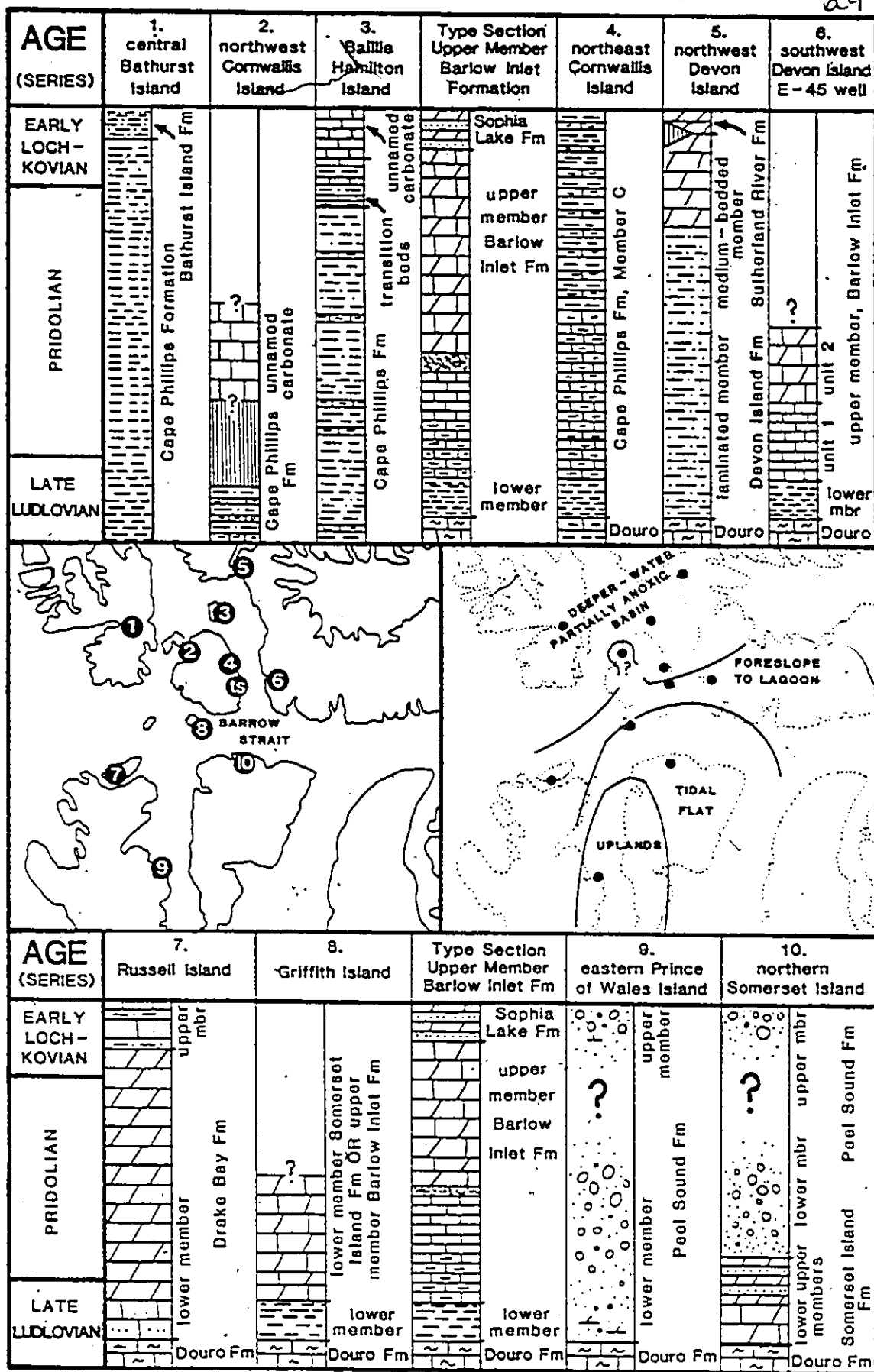


FIG. 5-13 Regional lithostratigraphic correlation.

5.4 CONCLUSIONS

1. The upper member of the Barlow Inlet Formation on eastern Cornwallis Island, Arctic Canada, a hitherto largely unstudied carbonate succession of late Ludlovian to early Lochkovian age, can be satisfactorily described in terms of three magnafacies incorporating 7 superfacies, as follows: Basin Slope Magnafacies (distal foreslope and medial foreslope superfacies), Platform Margin Magnafacies (proximal foreslope, shelf edge, proximal back barrier superfacies) and Platform Interior Magnafacies (restricted shelf lagoon and tidal flat superfacies).
2. The diagnostic characteristics of each superfacies are summarized in Table 5-4. Of particular note is the rich and diverse facies of biostromes and buildups recorded in the medial foreslope to restricted shelf lagoon superfacies.
3. The lower 360 m of the Barlow Inlet Formation is excellently exposed from Cape Hotham north to Depot Point. Fair exposure of the entire upper member is available at Read Bay and Sophia Lake. Much of the upper member is discontinuously exposed in the vicinity of Cape Rescue.
4. Physical correlation between study localities at the superfacies level suggests that, a) facies changes

were abrupt and in places complex, and b) major backstepping and forestepping events evidently affected the platform.

5. The platform margin shifted considerably in position through the geological history of the platform, and was oriented in places oblique and not normal to, the present north-south strike of the outcrop belt.
6. Platform margin migration and platform evolution was the combined result of accelerated block fault movement, progradation and regional subsidence. Eustatic sea-level changes are not invoked. At least nine major forestepping and backstepping platform events can be recognized. The events record periods of pan-platform drowning or uplift.
7. Pronounced uplift in early Lochkovian time, possibly with a locus to the north of the type section, effectively terminated platform growth and evolution in the study area.
8. The carbonate mass represented by the Barlow Inlet Formation was an attached interior platform with a rimmed, accretionary shelf margin. Locally the margin may have been stepped. Carbonate deposition occurred in equatorial latitudes under an arid to semi-arid climate. Wave stress was moderate to minimal. Pollution by terrigenous detritus was minimal. The production of clay- to silt-sized carbonate was evidently appreciable.

9. The Barlow Inlet Platform evolved primarily due to structural modifications of the pre-existing Douro Ramp. The Barlow Inlet Formation upper member was laid down over the deltaic deposits of the lower member
10. Despite the presence of coeval formations on virtually all surrounding islands, marked facies changes and the lack of detailed stratigraphy in the correlative formations makes comparison difficult. The lower member of the Somerset Island Formation is lithologically and environmentally equivalent to the TF series of the Platform Interior Magnafacies. Similarly the C member of the Cape Phillips Formation is the equivalent of the distal foreslope superfacies.

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5.6 PLATES

Plate 5-1

False color Landsat image of Cornwallis and Griffith Islands.



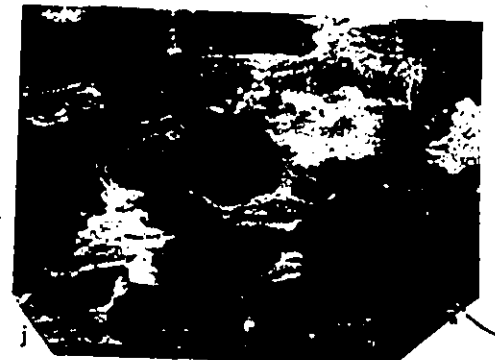
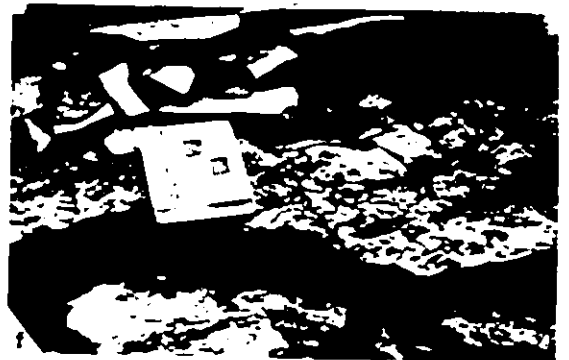
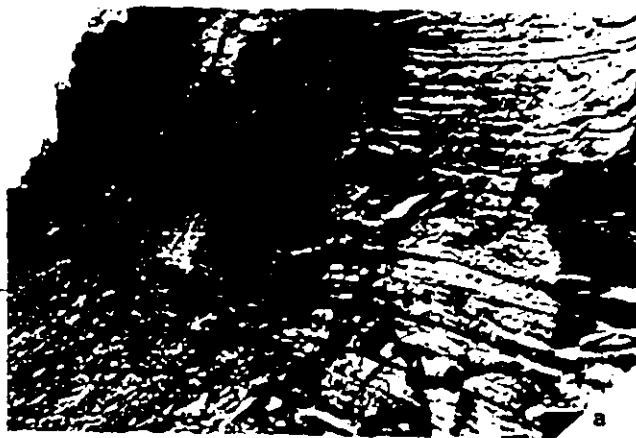
PL.5-1

Plate 5-2

Distal foreslope superfacies.

Bar scales are 1 cm (unlabelled) or tenths of 1 cm (labelled).

- a Rhythmic intercalation of lime mudstone beds and marl seams. Hammer 31 cm long.
- b Typical fine laminae of lime mudstone beds. Lighter-colored laminae are composed of calcilutite; darker laminae are composed of quartz silt, peloids, calcite silt, and rarely dolomite. Right half of sample is stained by Alizarin-red. Laminae are bent, or completely disrupted in the vicinity of the two feeding burrows (polished slabs).
- c Massive unit (.3 m thick) of shell debris grainstone and packstone, within otherwise monotonous sequence of thin, planar to slightly wavy-bedded lime mudstones. Arrow points to undulose solution contact separating grainstone (above) from packstone (below).
- d Phycodes sp., a common ichnogenus of the distal foreslope superfacies.
- e Photomicrograph of calcilutite lamina (1) separated by scoured base from overlying coarse-grained lamina (2) composed of calcite, quartz silt, and peloids (negative image).
- f Fe/Mn encrusted hardground surfaces (at note book level, and ledge beneath).
- g Intraformational truncation surface.
- h Polished slab exhibiting rip-up clasts and eroded hardband at base, highly bioturbated middle zone (light areas are calcilutite, dark zones are peloid-rich), and Fe/Mn emcrusted hardground surface (arrow).
- i Irregular Fe/Mn nodule (polished slab).
- j Nodular to wavy beds with dolomite mottles (lighter colored patches of coarser texture). Tapering spar-filled cracks may be due to syneresis (polished slab).



PL5-2

Plate 5-3

Medial foreslope superfacies.

- a Intraformational truncation surface associated with slump folded strata. Rick Zuran for scale
- b Pilichnia sp., a common ichnogenus of the medial foreslope superfacies. Scale 15 cm.
- c Oligomictic carbonate breccia with crinoidal rudstone matrix. Scale 15 cm.
- d Large scale soft-sediment overfolds in thin-bedded lime mudstone. Hammer 31 cm long.
- e Large angular blocks of sponge mud mound boundstone set in lithoclast-rich lime mud matrix. Interpreted megabreccia (or debris flow) associated with in situ sponge biocumules. Hammer 31 cm long.
- f Slide deposit(s) featuring in-situ brecciation of coherent beds (right); and random orientation of clasts (left). Deposit at left verges on a debris flow deposit and may truncate earlier slide deposit.
- g Rhizoidal holdfast network within medial foreslope superfacies. Lens cap 6 cm across.
- h Concentric color bands of micrite tentatively considered to be the product of cementation concomitant with endoskeletal resorption in rhizoidal holdfasts. White areas are non-resorbed holdfast skeleton (polished slab). Bar scale 1 cm.
- i Lithistid sponge-dominated mud-mounds.
- j Gutter casts on the soles of intraclast-rich, crinoidal packstone beds, representing either storm-reworked flank beds, or mass flow deposits (mud-poor debris flows). Hammer 31 cm long.

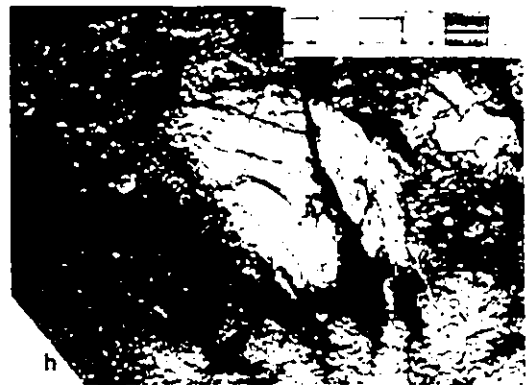
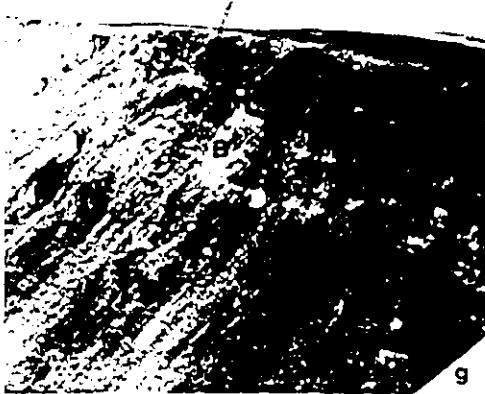
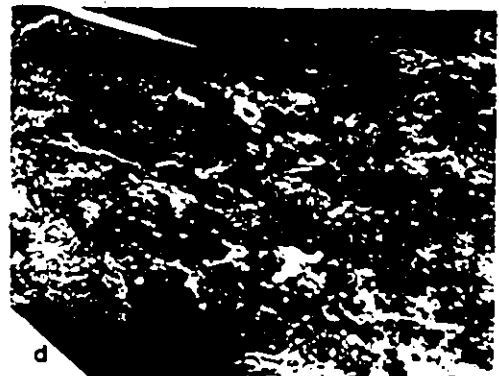


PL5-3

Plate 5-4

Proximal foreslope superfacies.

- a Rich thanatocoenose featuring favositid corals, atrypid (1) and atrypoid (2) brachiopods. Biofacies of the proximal foreslope show the greatest abundance and diversity of fossils. Bar scale 1 cm.
- b Zoophycos sp., a common ichnogenus of the proximal foreslope superfacies. Bar scale 1 cm.
- c Diverse community mud mound (DCM). Scriber 14 cm long.
- d Solenoporid, bioclast-rich flank beds to a DCM. Scriber 14 cm long.
- e Layers of crinoidal rudstone in a massive bed of crinoidal packstone (crinoid blanket deposit). Scale 15 cm.
- f Rhizoidal holdfast pod (P), within a crinoid blanket deposit. Arrows point to pod's basal contact. Lens cap 6 cm across.
- g Stromatoporoid-crinoid rubble skeletal buildup (B). Light color is the result of preferential dolomitization of these structures.
- h Hemispherical stromatoporoid in crinoidal packstone/rudstone matrix. Typical core facies of stromatoporoid-crinoid rubble skeletal buildup. Scale 15 cm.

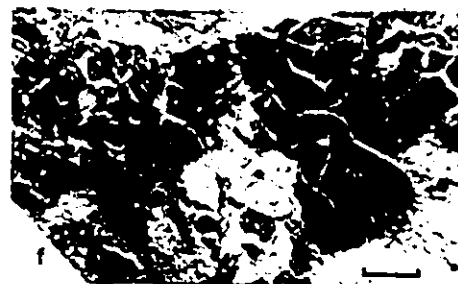
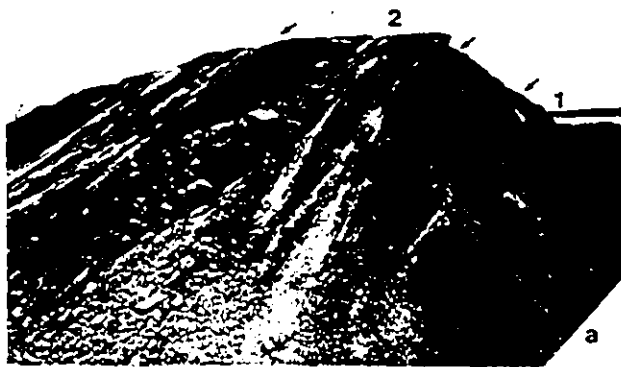


PL5-4

Plate 5-5

Shelf-edge superfacies.

- a Dolomitized massive skeletal (?crinoid) biocumules (1 and 2) of the shelf-edge superfacies. Biocumule 2 is 70 m thick.
- b Dolomitized massive skeletal (?crinoid) biocumule 43 m thick (between arrows). Note internal parallel stratification.
- c Massive stromatoporoids in partially dolomitized matrix. Jacob's staff with 10 cm divisions.
- d Severely dolomitized biocumule with pervasive moldic porosity (larger vugs probably reflect the former presence of massive colonial skeletons). Hammer 31 cm long.
- e Shelf-edge barrier reef (1) and back-reef lagoonal facies (2). Stratigraphic thickness of visible outcrop of reef is approximately 10 m.
- f Association of robust dendroid and lamellar stromatoporoids in shelf-edge and/or shallow proximal foreslope superfacies. Bar scale 1 cm.
- g Partially dolomitized crinoidal calcarenite grainstone. Conchoidal fractures probably follow original boundaries of crossbed sets. Height of outcrop is approximately 3 m.
- h Undolomitized, cross-stratified crinoidal calcarenite grainstone. Hammer 31 cm long.

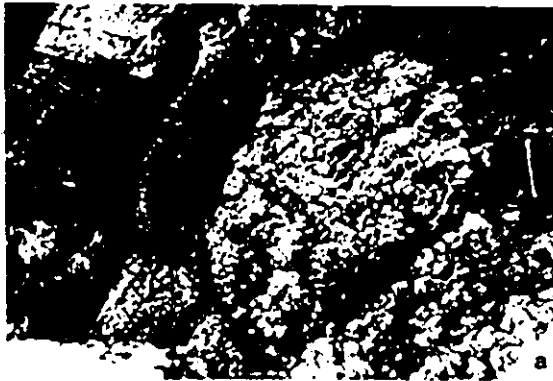


g PL5-5 h

Plate 5-6

Proximal back-barrier lithofacies.

- a Massive stromatoporoid biostrome of sparse structure. Stromatoporoid growth forms are predominantly hemispherical and subspherical. Note rubbly weathering of biostrome. Hammer 31 cm long.
- b Similar biostrome to that figured above. Randy Hyrniuk for scale.
- c Polished slab of biostromal facies exhibiting stromatoporoids (S), and solitary rugosans (R) set in a partially-dolomitized matrix (light-colored areas).
- d Biostrome of fasciculate rugosans (Stereoxyloides sp.). Biostrome extends laterally for several ten's of metres (beyond position indicated by arrow at right edge of photo). Scale 15 cm.
- e Polished slab of Stereoxyloides sp., principal rugose coral in the back-barrier and lagoon superfacies. Bar scale 1 cm.
- f Concentrations of delicate dendroid stromatoporoids (amphiporoids), common in both the back-barrier and lagoon superfacies.



PL.5-6

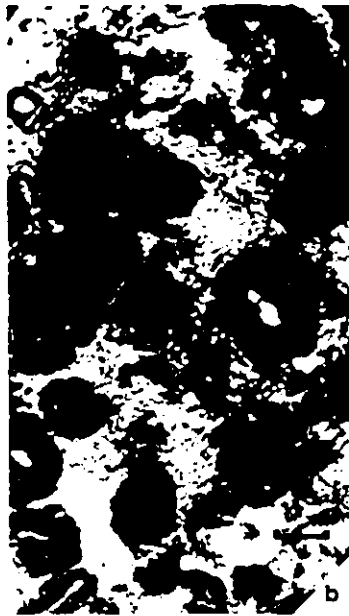
Plate 5-7

Restricted shelf lagoon superfacies.

Bar scales 1 cm (unlabelled), tenths of 1 cm (labelled).

Ruler scale 15 cm.

- a Medium-bedded, sparsely-fossiliferous mottled dolomitic carbonate lithofacies typical of lagoon superfacies. Note density and size of micrite mottles varies considerably.
- b Oncoid-rich carbonate, common in higher energy facies of the restricted shelf lagoon superfacies.
- c Skeletal-rich wackestone with abundant dendroid favositids. Although lagoonal facies can contain abundant bioclasts, faunal diversity is generally low, with rare stenohaline forms.
- d Faintly mottled massive dolosiltite with abundant vertical burrows (Skolithos sp.). Lithofacies may represent either low intertidal or infratidal zone.
- e Massive calcilutite mudstone deposited under restricted, low-energy infratidal conditions.
- f Omission surface within lagoon superfacies. Note truncated stromatoporid (arrow) (polished slab).
- g Pelecypodichnus sp. B, a common ichnospecies of the restricted shelf lagoon superfacies.
- h Photomicrograph of peloidal, ostracode fragment-rich packstone associated with higher energy facies of the lagoon superfacies.
- i Articulated valves of a megalodont bivalve, concentrations of which occur in the infratidal zone.

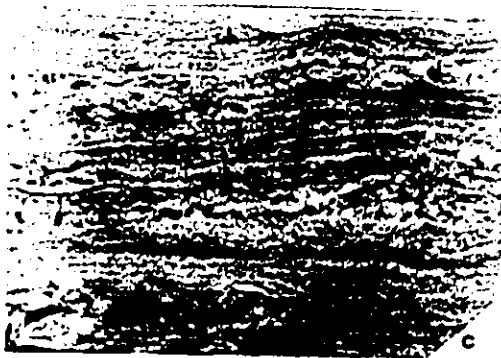


h PL5-7

Plate 5-8

Tidal flat superfacies.

- a Slightly wavy lamination within cryptalgal laminite. Bar Scale 2 cm.
- b Erosively-based rudstone of large, high-spined murchisonid gastropods, representing a tempestite lag deposit. Underlying unit is massive, dolo/calcsiltite of tidal flat origin. Scale 15 cm.
- c Laminoid fenestrae within cryptalgal laminite. Bar scale 1 cm.
- d Massive dolo/calcsiltite with desiccation polygons and the environmentally diagnostic trace fossil, Polarichnus garnierensis (arrow). Hammer 31 cm long.
- e Leperditid ostracodes on a desiccated bedding surface. Bar scale 2 cm.
- f Wavy to lenticular-bedded, wavy-laminated calcsiltite, interbedded with shale. Scale 15 cm.
- g Zig-zag crenulations on the surface of a cryptalgal mat. Field of view is 4 cm across.
- h Successive shallowing-upward cycles of lagoon-tidal flat paleoenvironmental aspect, at location H. Middle cycle figured is 4.3 m thick.



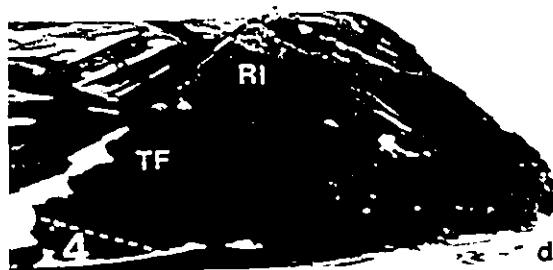
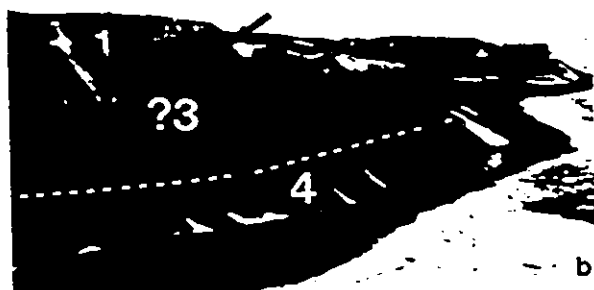
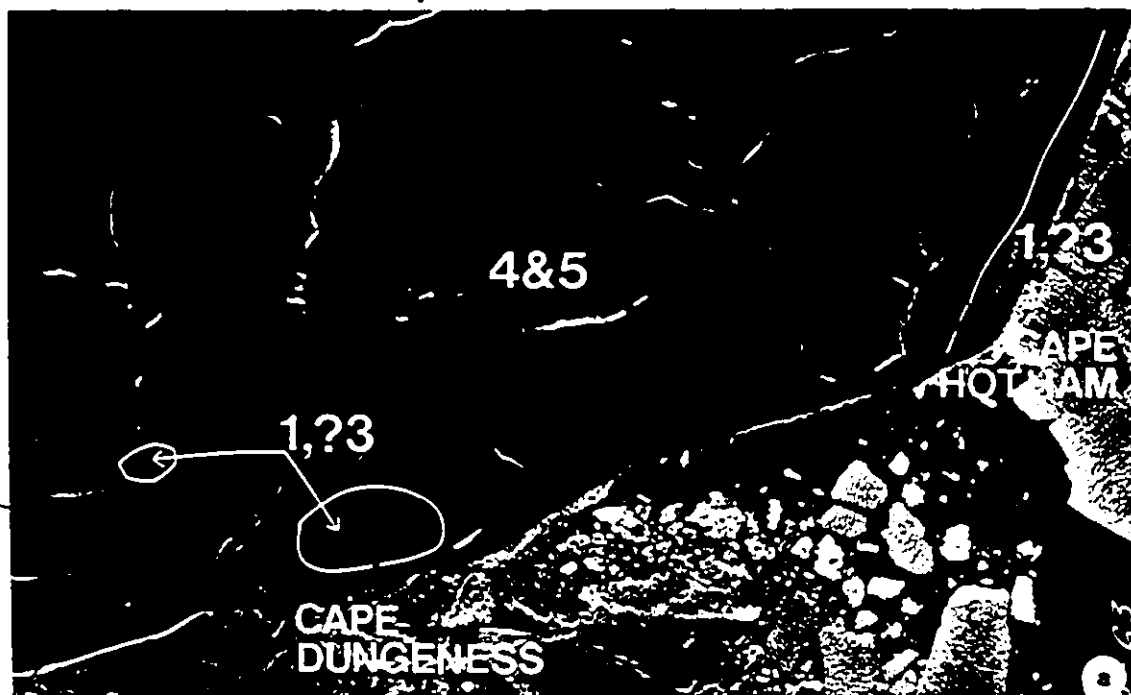
g PL5-8 h

Plate 5-9

Cape Dungeness and Cape Hotham study localities - 1

The major stratigraphic divisions are numbered as follows, on this, and ensuing plates:

- 0 - Sophia Lake Formation.
 - 1 - upper member, Barlow Inlet Formation.
 - 2 - lower member, Barlow Inlet Formation.
 - 3, ?3 - lithofacies equivalent to the lower member, Somerset Island Formation. Stratigraphic intervals consisting of TF series on southernmost Cornwallis Island, are here tentatively considered equivalent to 3, and are labelled ?3.
 - 4 - Douro Formation.
 - 5 - Cape Storm Formation.
 - ?6 - Cape Phillips Formation (C member). and lithofacies equivalents in the Barlow Inlet Formation.
-
- a Vertical airphotograph of study locality C. Dark lines with black numbers represent stratigraphic sections.
 - b Oblique view of Cape Dungeness (section 3). Arrow points to the stratum containing the vuggy dolostone in Figure c.
 - c Intensely dolomitized crinoidal rocks, near the top of section 3. Scale 15 cm.
 - d Oblique view of Cape Hotham (section 4). The TF series lies between a basal, diverse community mud mound and crinoid grainstone interval (1), and the first massive stromatoporoid biostrome (2), which marks the beginning of the RI series.
 - e Diverse community mud mound (outlined by ink) enclosed by dolomitized, in part cross-bedded, crinoidal packstones/grainstones. Section 4.
 - f Polished slab of sample from diverse community mud mound (section 4) featuring a lamellar stromatoporoid (top orientation as shown) encrusting massive calcilutite mudstone with dark patches of micritic bodies (similar to the alga, Renalcis). Bar scale 1 cm.



PL.5-9

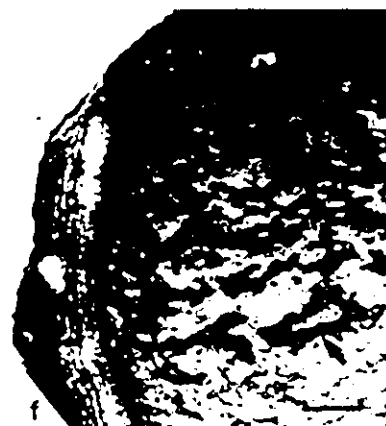


Plate 5-10

Cape Dungeness and Cape Hotham study localities - 2
Bar scales 1 cm. Jacob's staff 1.5 m.

- a Crinoidal rudstone flanking diverse community mud mound at base of section 4, featuring crinoid, solenoporid (S), and solitary rugose coral (R) bioclasts.
- b Articulated robust dendroid stromatoporoid (Stachyodes), in covered interval beneath TF series, section 4.
- c Lamellar stromatoporoid associated with fasciculate rugose coral (Stereoxylodes), in covered interval beneath TF series, section 4. Scale 15 cm.
- d Complete TF cycle. Hammer point on basal contact of oncoidal carbonate member; top of cycle is just below clinometer on Jacob's staff. Cycle is 1.2 m. thick (section 4, TF series).
- e Spindle-shaped cracks (synaeresis cracks?), common within TF series.
- f Upper part (up to arrow point) of cross-bedded crinoidal calcarenite interval, that separates the TF and RI series at section 4.
- g Torpedaspis cf. T. elongata Broad and Dineley, an ostracoderm in the TF series. Scale in mm.
- h Massive stromatoporoid biostromal facies overlying cryptalgal laminites in RI series, section 4.
- i Stratiform units of the RI series at section 4. Most are composed either of massive, calcilutite mudstone, or stromatoporoid biostromes. Thin, light-colored units are dolomitized cryptalgal laminites. Cliff height is 80 m.
- j Alternating biostromal (1,2) and massive lime mudstone lithofacies of the RI series (section 4).

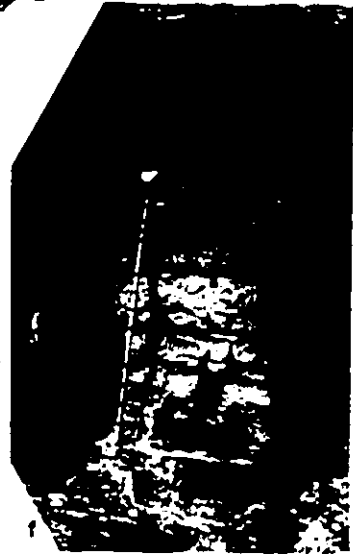
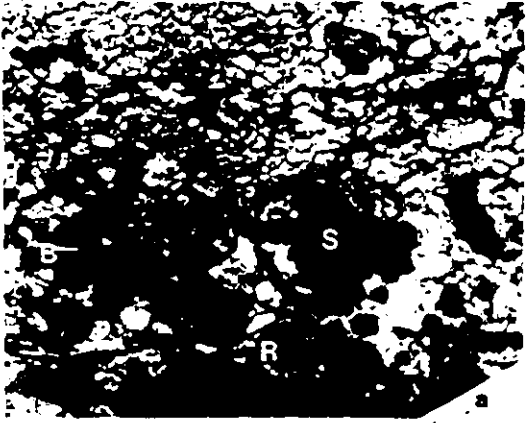
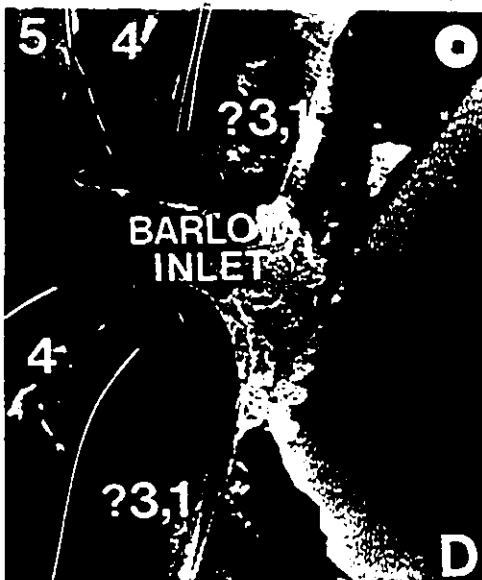


Plate 5-11

Barlow Inlet study locality.

- a Vertical airphotograph of study locality D. See Plate 5-9 for explanation of white numerals. Measured sections indicated by black lines and numerals.
- b Oblique view to the north of cliff face where sections 5 and 6 are located. Thickness of the lower member is 8.8 m.
- c Uprooted dististele portion of a crinoid column, associated with alveolitid and favositid corals, and brachiopods. Sample is from basalmost portion of the upper member of the Barlow Inlet Formation. Bar scale 1 cm.
- d Sparsely-fossiliferous, argillaceous lime mudstone (1) overlain abruptly by crinoid-solenoporid rudstone (2) developed on the flanks of a diverse community mud mound, 12 m above the base of the upper member. Scale 15 cm.
- e Massive-weathering diverse community mud mound (DCM) capped by thin, recessive unit of marlstone (hammer). Marlstone occurs 20 m above the base of the upper member, and just below the first cycles of the TF series. Hammer 31 cm long.
- f Transition between TF and RI series in section 5. Rubbly-weathering crinoidal packstones/wackestones predominate between arrowheads (20 m).
- g Massive lime mudstone of RI series in section 5. The writer for scale.
- h First occurrence of bedded units with a stenohaline (largely crinoidal) biota above the RI series in section 5. Height of outcrop is approximately 5 m.

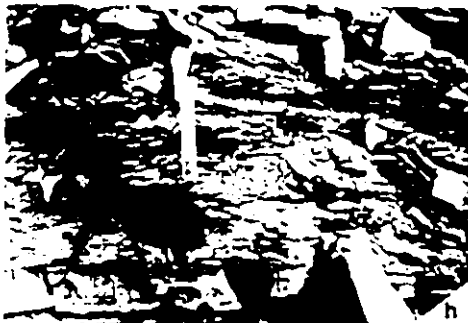
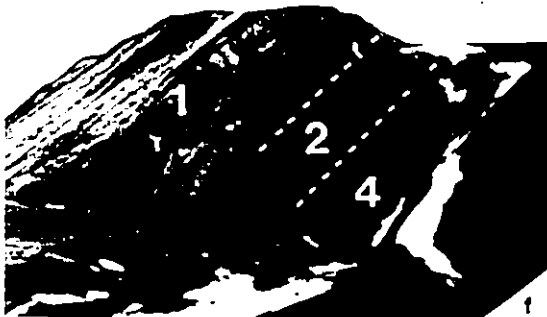
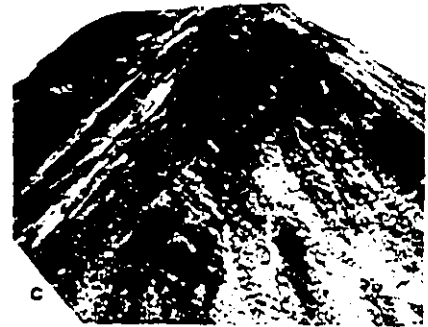


^gPL5-11^h

Plate 5-12

Goodsir Creek and 'Square-Lake' Creek study localities - 1
See Plate 5-9 for explanation of white numerals.

- a Vertical airphotograph of study locality E showing the location of sections 7 to 12 inclusively. The first occurrence of cryptalgal laminites is indicated by 'AL'. Measured sections indicated by black lines and numerals.
- b Oblique view to the north of section 11. Total stratigraphic thickness of upper member within field of view is 305 m.
- c Oblique view to the south of sections 10 and 9. Total stratigraphic thickness of upper member within field of view is 175 m.
- d Oblique view to the north of part of section 8. Total stratigraphic thickness of upper member within field of view is 215 m.
- e Oblique view to the northwest of most of section 8. The section that traverses the major cliff exposures in the background, is continued in the river gorge in the lower right foreground (arrow points to the 1st exposures in gorge).
- f Oblique view to the south of section 7. The lower member (2) is 52 m thick.
- g Fasciculate rugose coral (Stylopleura) in biostrome in section 7. Bar scale 1 cm.
- h Thin, platy-weathering marlstone interval above the diverse community mud mounds of section 8. (8 m above the base). Scale 15 cm.
- i Transition between lamellar stromatoporoid biostrome (1) and intensely dolomitized crinoidal rudstone (2) in section 8, 24 m above the base. Scale 15 cm.



PL.5-12

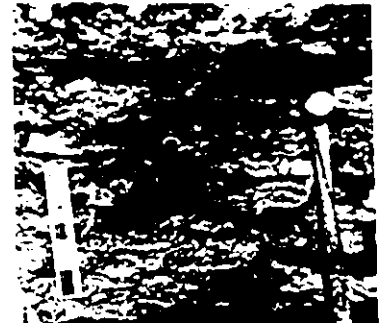


Plate 5-13

Goodsir Creek and 'Square-Lake' Creek study localities - 2

- a Severely dolomitized variant of the diverse community mud mound biocumule facies. Section 8, 15-20 m above the base.
- b Geologist stands at top of dolomitized massive skeletal (?crinoidal) biocumule within section 8. Arrow points to conspicuous crinoidal dolerudite unit of laterally-variable thickness. The base of this latter unit occurs 73.5 m above the base of section 8.
- c Major slide structure (the base of which is indicated by arrows), affecting medial foreslope mudstone and wackestone beds in section 8, approximately 165 m above the base.
- d Succession of infratidal and intertidal facies in section 8, commencing 332 m above the base with calcisiltite mudstone, in part laminated (1), overlain by oncoidal carbonate (2), in turn succeeded by cryptalgal laminite and massive dolosiltite (3), capped by oncoidal and weakly mottled carbonates (4). Note erosional contact between 1 and 2 (arrow), and less conspicuously, between 3 and 4. Total stratigraphic thickness in field of view is 10 m.
- e Diverse community mud mound (DCM) overlain by platey-weathering marlstone in section 9. Arrow indicates the highest stratigraphic level (adjacent to mud mound) at which significant crinoidal debris accumulated (21.3 m above the base of section 9). Vertical field of view is approximately 1 m.
- f Oblique view to the east of sections 10 and 9. Contrast between stratigraphically equivalent, intensely dolomitized (1) and partially dolomitized (2) biocumule facies, is discussed in text. Bar (1) is 37 m thick.
- g Intraformational truncation surface (arrows) in section 11 (94 m above the base). Included in beds above the truncation surface, is an overturned block of fasciculate rugosan (FG) biocumule.
- h Abundant chert nodules in lime mudstone beds 97 m above the base of section 11. Hammer 31 cm long.
- i The trace fossils Chondrites (1), and Palaeophycus (2), from beds 180 m above the base of section 11. Scale 15 cm.

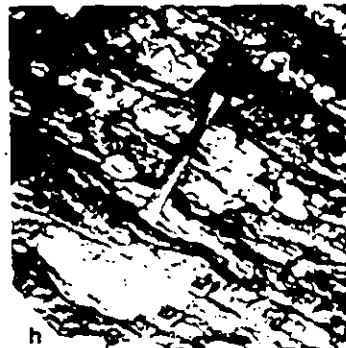
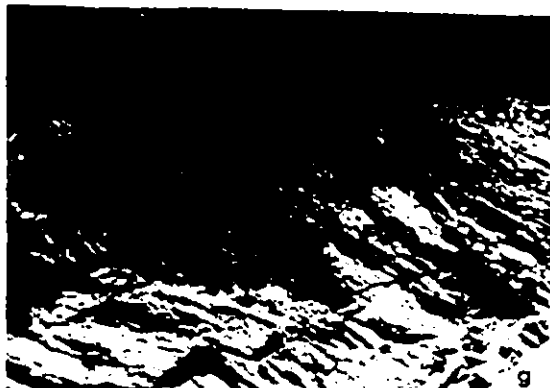
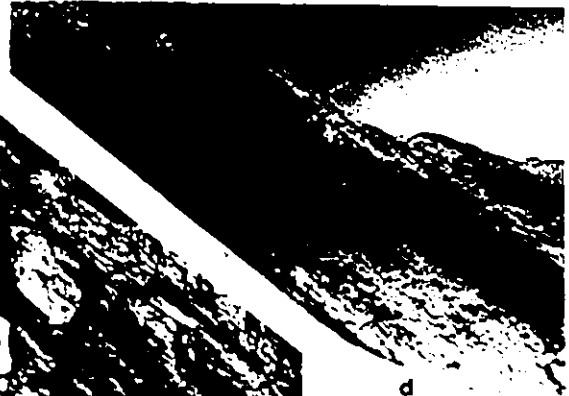
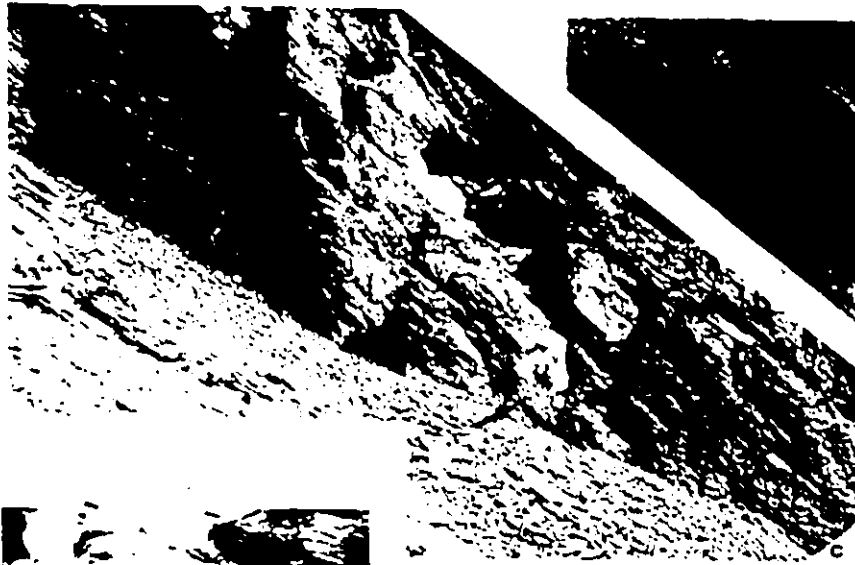


Plate 5-14

Depot Point study locality.

- a. Vertical airphotograph of study locality F, showing the positions of sections 13 to 18, inclusively. 'AL' indicates the position of the first occurrence of cryptalgal laminite. See Plate 5-9 for explanation of white numerals. Measured sections indicated by black lines and numerals.
- b. Bedding plane view of Atrypoides foxi (smooth spiriferid brachiopod) and articulated crinoid pelmae and calyces set in calcilutite matrix. Section 18, 50 m above base. Bar scale 1 cm.
- c. Oblique view to south of sections 13, 14, 15 and 16. Lithistid sponge mud mound interval indicated by arrows. Plateau is roughly 200 m above ice surface. 1 and 2 are lower and upper members, respectively.
- d. View looking up at cliff face in Figure c. Unfossiliferous, thin-bedded lime mudstone (1), is overlain by carbonate megabreccia (2), in turn succeeded by lithistid sponge dominated mud mound (3), capped by breccia, lime mudstone/wackestone, and crinoidal packstone beds of the medial foreslope superfacies (4).
- e. Abundant Pilichinia burrows in thin-bedded lime mudstone (interval 1 of Figure d). Scale 15 cm.
- f. Oblique view of strata in the vicinity of section 16. White-weathering crinoidal packstone beds flanking lithistid sponge-dominated mud mound (LSDM) are common, some of which are laterally persistent (arrow) and may represent deposits of storm-redistributed bioclasts. Thickness of indicated bed is not greater than .5 m.
- g. Massive crinoidal rudstone with calcilutite matrix, immediately adjacent to mud mound facies rich in rhizoidal holdfasts. Portion of scale shown is 12 cm.
- h. Upright dendroid favositids. Section 17, 340 m above the base of the upper member.
- i. Oncoidal carbonate. Section 17, 333 m above the base of the upper member. Bar scale 1 cm.
- j. Rippled bedding surface of laminated dolo/calcsitite facies. Section 17, 348 m above base of upper member. Scale 15 cm.

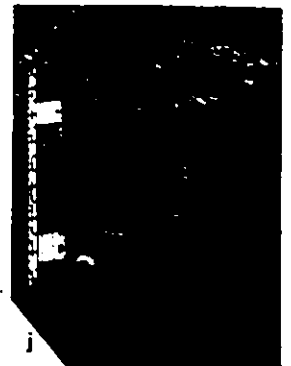
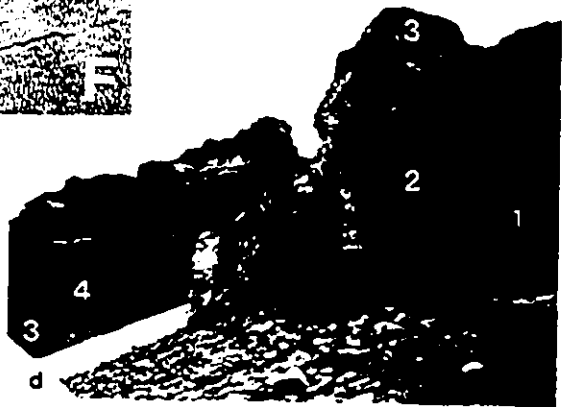
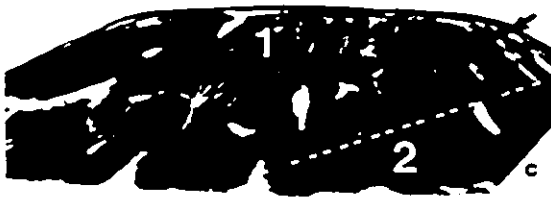
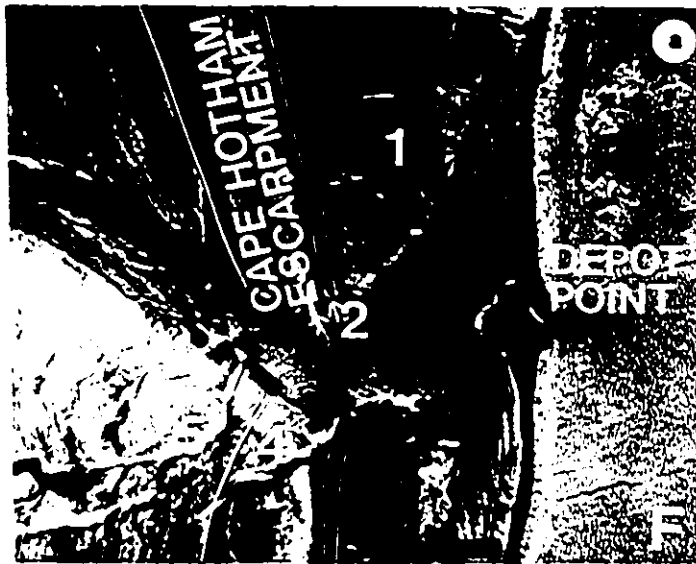


Plate 5-15

Read Bay (type section) study locality - 1
See Plate 5-9 for explanation of white numerals.

- a Vertical airphotograph of study locality H including section 20 and 21. There is slight overlap between sections 20 and 21, the former terminating and the latter commencing in the position of the dashed line indicated by an arrow. 'AL' marks the first occurrence of cryptalgal laminite, 'RF' points to a Coenites-capped, fenestral mud mound. A, B, C, D1, and D2 are informal stratigraphic intervals discussed in text.
- b Oblique view to the northeast. River gorge containing the type section indicated by open arrow.
- c Oblique view to the southwest of the upper part of the type section with Sophia Lake-Barlow Inlet formational contact in the foreground.
- d Stromatoporoid biostrome of frame structure in interval C of section 20, 365 m above the base. Hammer 31 cm long.
- e Close-up of massive stromatoporoid within biostrome in Figure d. Hammer 31 cm long.
- f Shallowing-upward cycles of TF aspect (interval D1), 785 m above the base of section 20. Second cycle from the bottom is 4.3 m thick.

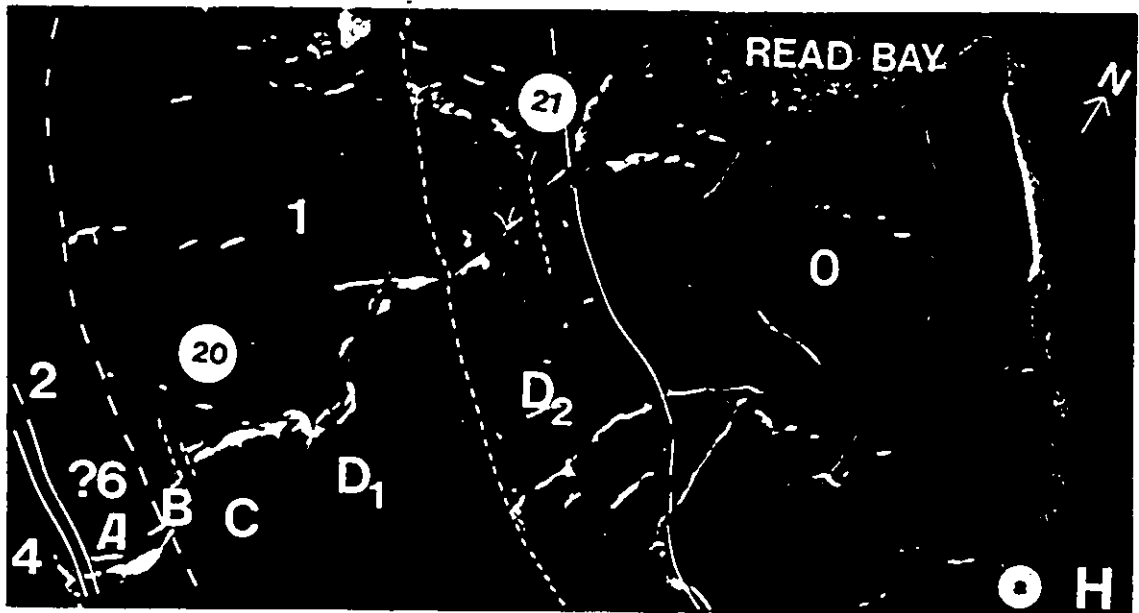


Plate 5-16

Read Bay (type section) study locality - 2

- a Mottled dolomitic lithofacies characteristic of peritidal deposition. Skolithos (vertical burrow), indicated by arrow. Note highly variable mottle patterns between beds. Interval D1, 945 m above the base of section 20. Hammer 31 cm long.
- b Concentration of Pelecypodichus sp. B on bedding plane, interval D1. Lens cap 6 cm across.
- c Concentration of murchisonid gastropods in conspicuous bed between 998.7 and 1000.5 m above the base of section 20 (interval D1). Scale 15 cm.
- d Biostrome of hemispherical, lamellar and adnate stromatoporoids (s), dendroid and massive favositids (c), and solitary rugosans (r), base of D2 interval, 1036.7 m above the base of section 20 (polished slab). Bar scale 1 cm.
- e Sequence at contact between D1 and D2 intervals, section 20. In order, this sequence consists of mottled dolomitic limestone type 2 with a restricted fauna (1), cryptalgal laminite with desiccation polygons (2), abruptly overlain by a mottled dolomitic limestone type 1 with megalodont bivalves, hemispherical stromatoporoids, solitary rugosans and large murchisonid gastropods (3), overlain by stromatoporoid biostrome (see description for Figure d) (4), overlain by fenestral calcilutite mudstone (5). The contact between D1 and D2 interval lies between 2 and 3. Unit 4 is 7.2 m.
- f Fenestral mud mound core (1), overlain by Coenites bafflestone (2), 1114.2 m above the base of section 20. Scale 15 cm.
- g Oblique view of Coenites -capped, fenestral mud mound (arrow). Mound is 7 m thick.
- h,i,j Photomicrograph of typical lithofacies near the Barlow Inlet-Sophia Lake formational contact. Peloidal coarse calcisiltite (h), argillaceous fine calcisiltite (i) are from the uppermost beds of Barlow Inlet Formation (section 21), whereas the calcareous, coarse quartz siltstone (j) is from the lowest beds of the Sophia Lake Formation. (thin sections, all at same scale 0.05 mm).

Plate 5-16
(Continued)

Read Bay (type section) study locality - 2

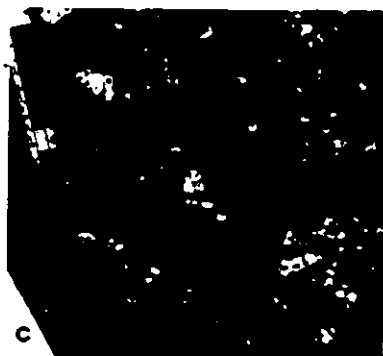
- k Uppermost sequence in section 20 featuring massive calcilutite fenestral mudstone (1), massive calcilutite floatstone with megalodont bivalves, murchisonid gastropods and hemispherical stromatoporoids (2), massive calcilutite mudstone with rare hemispherical stromatoporoids (3), and mottled dolomitic limestone (4) with a diverse stenohaline fauna. Laterally equivalent to unit 4 in section 21 are crinoidal wackestones and packstones. Unit 2 is 1.2 m thick.
- 1 Massive calcilutite mudstone (1) of RI aspect, overlain by a thin sequence of wackestone to packstone beds (<14 m) of more open marine aspect (common crinoids and brachiopods) and a thicker interval (70.7 m) of peritidal aspect (2). Basal contact of Sophia Lake Formation is just above topmost beds visible in plate. Height of cliff is approximately 100 m.



a



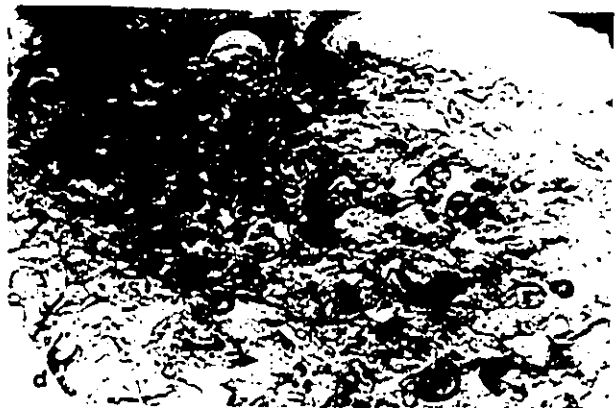
b



c



d



e



f



g



h

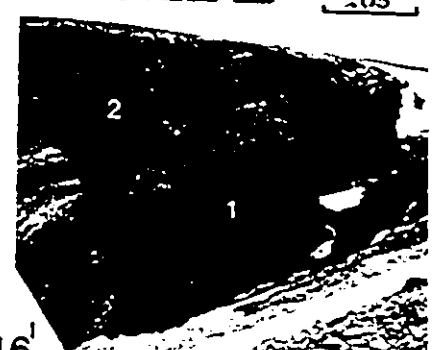
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204

1

PL.5-16^l



2

1

Plate 5-17

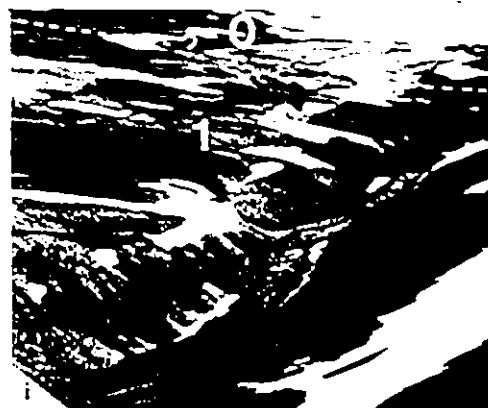
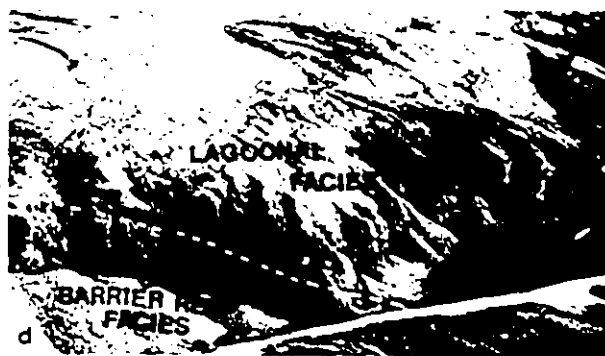
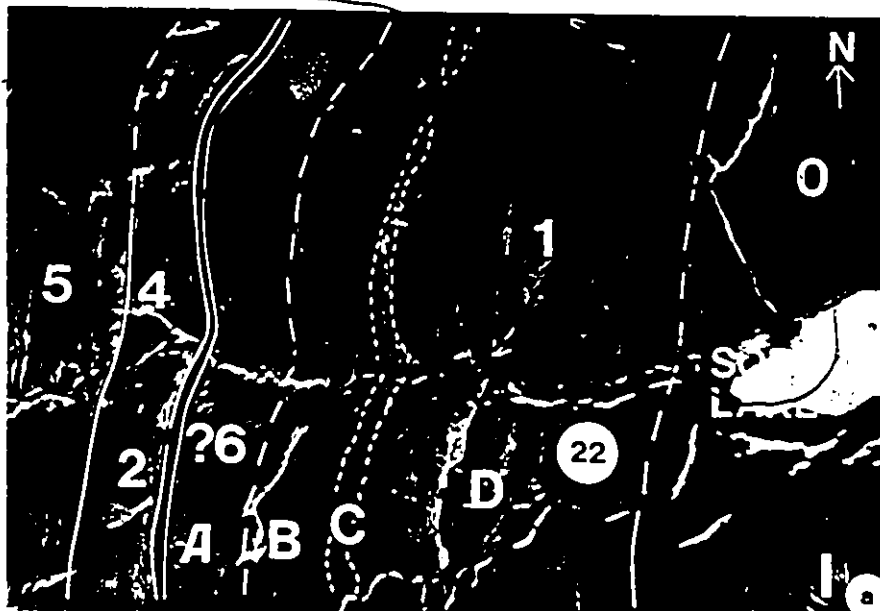
Sophia Lake study locality.
Bar scales 1 cm.

- a Vertical airphotograph of study locality I (section 22). '2 o/c' indicates the northernmost outcrop of the lower member examined, 'cr' is a thin dolomitized crinoidal grainstone, near the base of the upper member, and 'AL' is the first occurrence of cryptalgal laminites in the section. A, B, C, and D are gross, informal stratigraphic intervals within the upper member. See Plate 5-9 for explanation of white numerals.
- b Finely-laminated lime mudstone with burrows whose margins are coated with sulphides. Sample from interval A, section 22 (polished slab).
- c Finely-laminated, unfossiliferous lime mudstone exhibiting evidence of syndimentary lithification. Upper surface is suggestive of labyrinthine nodule development (nodule margins can be traced on the polished surface). Spar filling of Skolithos tubes (arrow), suggests partial consolidation took place relatively soon after deposition. Interval A.
- d Oblique view to the northeast of interval C, a barrier-reef complex. Small black arrow points to first occurrence of fossiliferous mottled dolomitic lithofacies of shallow subtidal lagoon aspect. Stratigraphic thickness of exposed reef facies is approximately 10-12 m.
- e Amphiporid floatstone of lagoonal facies (interval C).
- f Mottled dolomitic lithofacies (indicated in Figure d). Jacob's staff with 10 cm subdivisions.
- g Massive, calcilutite mudstone overlain by cryptalgal laminite, two lithofacies characteristic of interval D. Scale 15 cm.
- h Megalodont bivalve shell concentration in lime mudstone matrix, a conspicuous if rare lithofacies of interval D. Vertical field of view is .6 m.

Plate 5-17
(Continued)

Sophia Lake study locality.
Bar. scales 1 cm.

- i Oblique view to WNW. of river gorge containing the upper part of interval D of section 22. Approximate position of Barlow Inlet-Sophia Lake Formation contact is shown, as is the position of crinoidal lithofacies above interval C. The bulk of the outcrop in this view is massive to thick-bedded, unfossiliferous lime mudstone (RI series affinities). Distance from viewer to contact about 1 km. See Plate 5-9 for explanation of numerals.

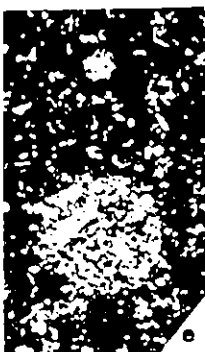
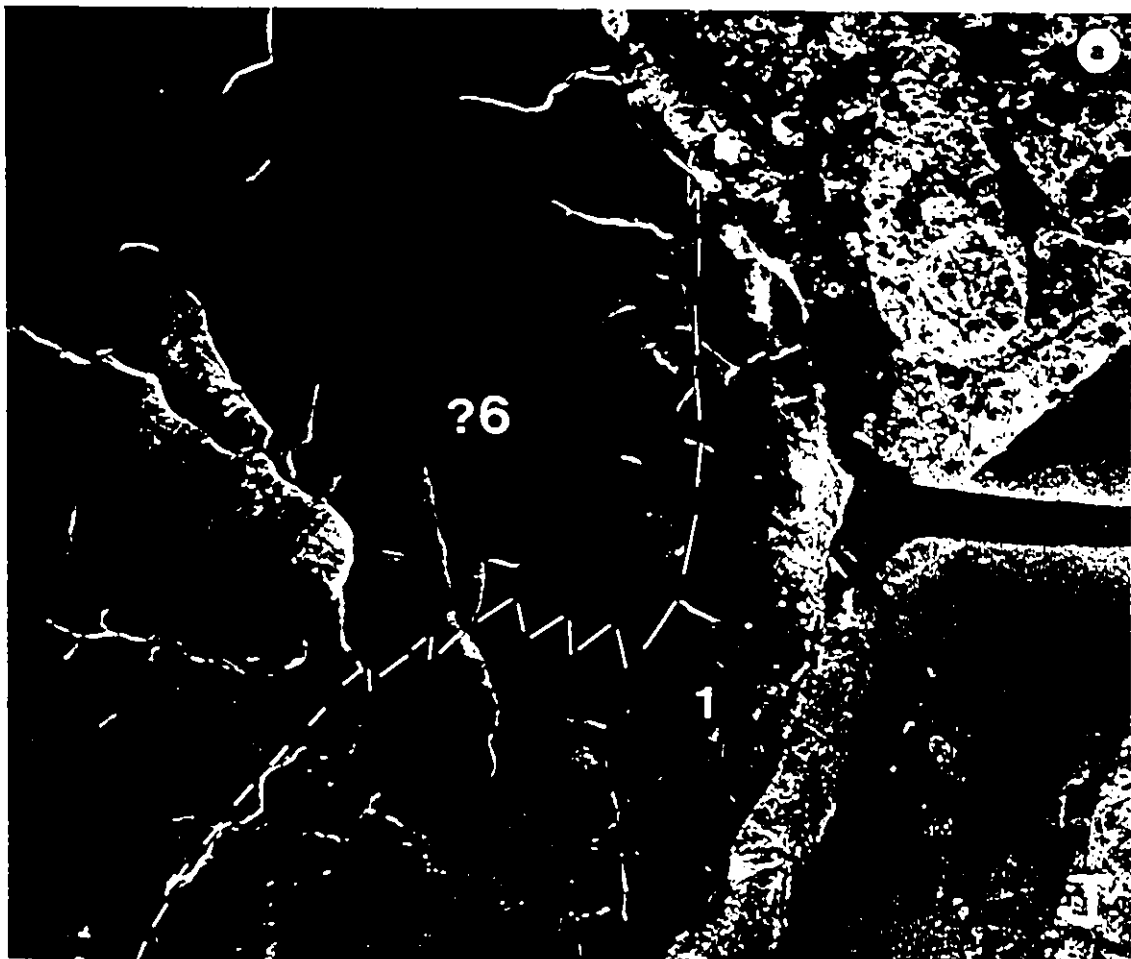


PL.5-17

Plate 5-18

Cape Rescue study locality - 1
Bar scales are 1 cm (unlabelled) or tenths of 1 cm
(labelled).

- a Vertical airphotograph of study locality J. Position of facies boundary (dashed white line) is approximate, and roughly follows Thorsteinsson (1973). 'RF' refers to a small zoned biocumule discussed in the text. Dips in the area vary, but are generally shallow to the east or northeast. See Plate 5-9 for explanation of white numerals. Measured sections indicated by black lines and numerals.
- b Planar, thin-bedded lime mudstone of the basal part of section 23.
- c Fe/Mn encrusted hardground developed on a packstone composed principally of small gastropods. Section 24, 36 m above the base.
- d Spar-infilled mold of small bivalve. Note shell material has been leached, and in its stead, there is an isopachous rim of pyrite (arrow). Section 23, 5 m above base.
- e Spherules composed of calcite silt (in some instances, possible pseudospar). Such spherules are quite common in the distal basin slope facies and may represent calcispheres or recrystallized radiolarians. An origin as burrow fills is unlikely, as tube-shaped cross-sections are not observed.
- f Encrustation by the porostromate alga, Sphaerocodium sp. in association with hardground in Figure c. Section 24, 36 m above the base.
- g Oblique view of section 24. View to the east.



PL5-18

Plate 5-19

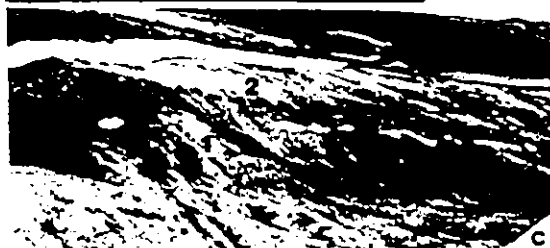
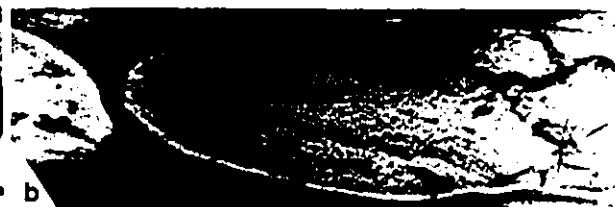
Cape Rescue study locality - 2
Bar scales are 1 cm. Hammer 31 cm long.

- a Fossiliferous mottled dolomitic limestone from basal part of section 25 (polished slab).
- b Faintly mottled, thin-bedded lime mudstone from interval 57 m above the base of section 25. Note extreme rounding of nodule (polished slab).
- c Oblique view to the north, of section 27 and environs. A zoned biocumule (1) is overlain by stratiform units of restricted shelf lagoon superfacies aspect (2).
- d Crinoidal-intraclastic orthoconglomerate overlying eroded lime mudstone substrate. Well-rounded nature of intraclasts may, in part, reflect derivation by exhuming diagenetic nodules, rather than rounding by transport. Origin of concentric bands in intraclasts is not known. Section 25, 65 m above the base (polished slab).
- e Crinoid blanket facies, section 28, 16 m above the base. Jacob's staff 1.5 m.
- f Oblique view to the west of sections 28 and 29. Crinoid blanket facies (1), facies of the medial foreslope (2), lithistid sponge dominated mud mound (3) and mucophyllid mud mound (4) are indicated.
- g Groove marks on bedding plane of lime mudstone, overlain by an interpreted carbonate debris flow. Section 28, 42 m above the base.
- h Carbonate megabreccia overlying and adjacent to, deformed, thin-bedded lime mudstones. Section 28, 50 m above the base.
- i Crinoidal dolarenite from basal part of section 28.
- j Intergrowth of spongiostromate algae (1) and lithistid sponge (2) in boundstone facies, 36 m above the base of section 28 (polished slab).
- k Wavy to nodular bedded crinoidal mudstone/wackestone from medial foreslope facies of section 28 (polished slab).
- l Exotic blocks of boundstone in medial foreslope facies featuring stromatoporoids (S), favositid corals (F) and rhizoidal holdfasts (RH) (polished slab).

Plate 5-19
(Continued)

Cape Rescue study locality - 2
Bar scales are 1 cm. Hammer 31 cm long.

- m Lithistid sponge dominated mud mound (#3 of Figure f),
section 29, 25.8 m above base. Jean Dugal for scale.

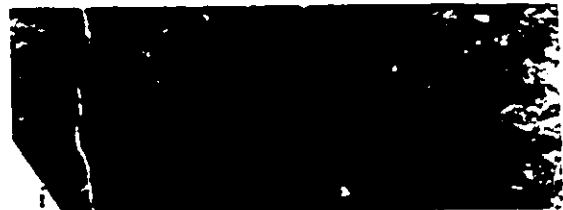
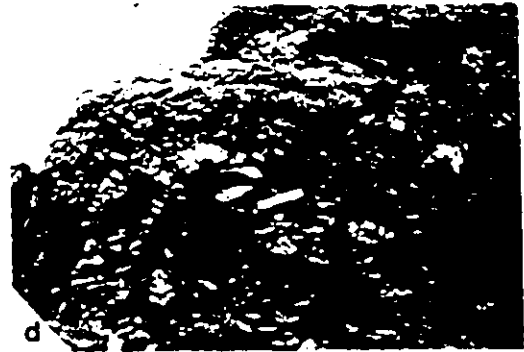
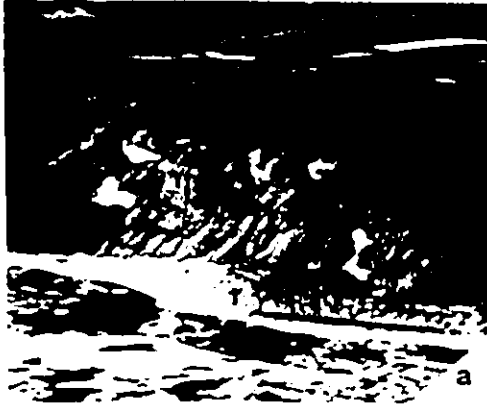


PL.5-19

Plate 5-20

Cape Rescue study locality - 3

- a Oblique view of the stromatoporoid-crinoid rubble skeletal buildup (2) at section 30. Crinoid blanket facies (1), and stromatoporoid inertiastone biostrome (3) are indicated.
- b Same structure as in Figure a, viewed from below. Note massive appearance, and sharp break (arrow) half way up buildup.
- c Core facies of buildup (section 30), featuring stromatoporoid framestone with crinoidal rudstone matrix. Scale 15 cm.
- d Massive stromatoporoid inertiastone biostrome of sparse structure, lateral to principal buildup in section 30. Scale 15 cm.
- e Grading of crinoid and solenoporid bioclasts from crinoidal rudstone facies of the buildup core at section 30. Bar scale 10 cm.
- f Dolomitized crinoidal rudstone capping buildup at section 30. Marcasite mineralization commonly infills fractures and solution molds. Scale 15 cm.
- g Intensely dolomitized stromatoporoid biostrome (dolomitized massive skeletal (?crinoid) biocumule) at section 32. Hammer 31 cm long.
- h Oblique view of section 32 (arrow indicates route of section). Laterally extensive dolomitized biostromes (as in Figure g) are indicated (1).
- i Dolomitized fenestral carbonate mudstone in section 32. Bar scale 1 cm.



PL5-20

Plate 5-21

Cheyne Point and Cunningham Inlet study localities.
See Plate 5-9 for explanation of white numerals.

- a Vertical airphotograph of study locality B, with position of sections 1 and 2 shown. Measured sections indicated by black lines and numerals.
- b Section 1 is located on the downthrown side of indicated fault and includes the uppermost Douro Formation, the lower member of the Barlow Inlet Formation (6.8 m thick), and strata equivalent to the lower member of the Somerset Island Formation.
- c Study locality A at Cunningham Inlet, northern Somerset Island. View is to the east, on strike with the northern bounding fault of the Cunningham Inlet Graben. Bar represents section (53.4 m) measured through a part of the lower member of the Somerset Island Formation (section 13 of Miall and Gibling, 1978). See Plate 5-9 for explanation of numerals.
- d Top and base of shallowing-upward hemicycles at study locality A. Lower unit consists of planar to wavy laminated dolo/calcsiltite. The overlying unit (the base of which is in line with the hammer head) consists of mottled dolomitic limestone. Hammer 31 cm long.
- e Tidal flat succession in section 2 (study locality B). Sequence consists of unfossiliferous, rubbly weathering lime mudstone (1), dolomitic cryptalgal laminite (2), oncoidal limestone (3), and mottled dolomitic carbonate (4). Unit 2 is 2 m thick.
- f Two orders of desiccation polygons, study locality A. Scriber 6 mm diameter.
- g Slight doming of cryptalgal laminites to form laterally-linked stromatolites. Section 1. Scale in inches.
- h Wavy-bedded, wavy to cross-laminated calcsiltite. Study locality A. Bar scale 3 cm.

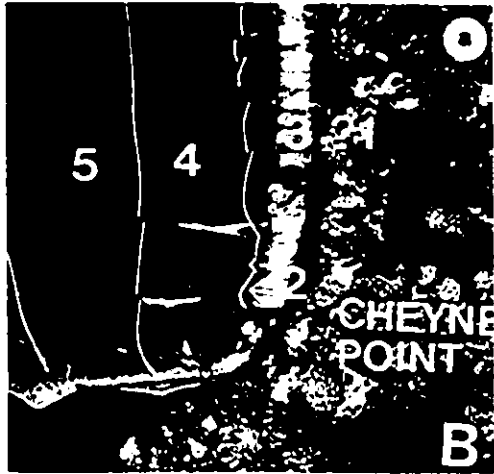
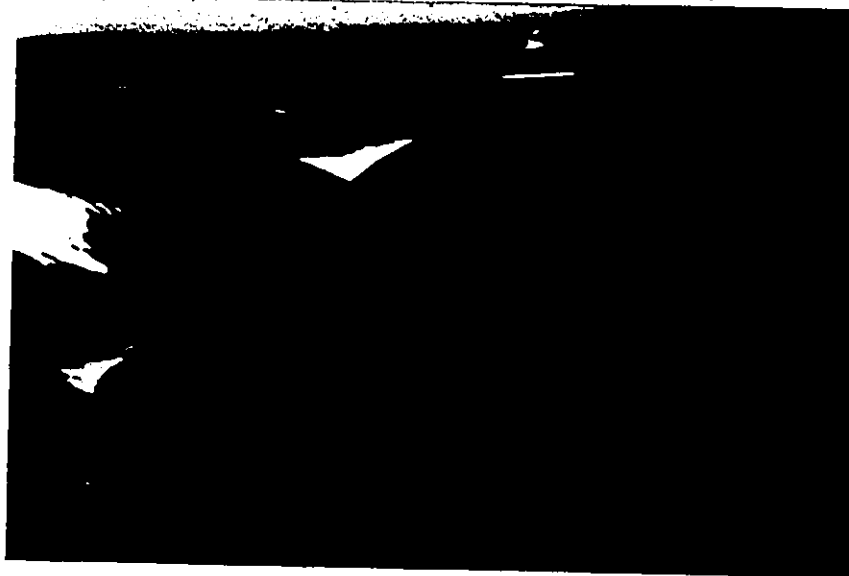


Plate 5-22

Contact between Barlow Inlet and Sophia Lake Formations.

Base of Sophia Lake Formation is placed at first major occurrence of dusky red-colored, quartzose siltstone. View is to the northeast, and shows cliff exposures along the gorge that contains the type section for both the Sophia Lake Formation, and the upper member of the Barlow Inlet Formation.



PL.5-22

Appendix A

PROBLEMS WITH THE PROVENANCE OF LOWER MEMBER DETRITUS

A.1 THE GRIFFITH ISLAND PROBLEM

The 12 m thick Cheyne Point exposure of the lower member (FIG. 2-2) is the most southerly and westerly occurrence of this member examined by the writer. It consists of dark grey mudrock (marginally so -- 54 % by weight of insoluble residue), with several limestone lenses containing large numbers of the brachiopod Atrypa erebus. The mudrock contains palynomorphs suggestive of marine depositional conditions (McGregor, pers. comm., 1981). Absent is the low-diversity restricted fauna of bivalves and inarticulate brachiopods typical of the lower member on eastern Cornwallis Island. Also conspicuously missing is the subarkose lithofacies. Together these features suggest that the Cheyne Point locality, although certainly affected by deltaic sedimentation, was more distal from the principal depocenter. If the uplifted arch provided the siliciclastic detritus, the thickness and inferred distal position of the lower member on Griffith Island is difficult to reconcile with its relative proximity to the Boothia Arch. Several arguments could explain such an anomaly but all, in turn,

raise other problems in this particular situation, as follows:

1. if fluviatile sediments entered the Cape Phillips Back-Barrier Basin considerably farther east than the drainage basin on the eastern flank of the uplift, the Griffith Island section would cease to be anomalous. This however would amplify problems discussed below under 'Missing Fluviatile Deposits'.
2. sediment dispersal patterns can be modified substantially by longshore drift, and this might account for the bulk of the detritus being deflected eastward. In this context Miall and Gibling (1978) described paleocurrent measurements from giant cross-beds of Lochkovian age from member 3 of the Peel Sound Formation on Somerset Island, that show a mean orientation to the northwest. Although the evidence was inconclusive, an eolian origin for these structures was advanced by the authors. Miall and Gibling (1978, p.114) were unable to reconcile these paleocurrent directions with postulated prevailing wind patterns based on paleogeographic reconstructions by Woodrow (1973). In recent paleomagnetic work, Lapointe and Dankers (1982, fig.4) revised the Early Devonian paleolatitude fix for the Peel Sound region to 10 degrees S. latitude, placing it under the influence of southeast trade

winds, and supporting an eolian origin for the giant cross-beds. If prevailing winds during the latest Silurian were to the northwest, concordant offshore currents may well have operated, but this would serve to increase, not decrease, detrital sedimentation in the Griffith Island area.

3. A complex paleo-shoreline comprising embayments, islands, and spits, can locally afford protection from influx of deltaic sediments. Such an intricate coastline would be more typical of a drowned shoreline, than of an emergent shoal surrounded by extensive tidal flats, the interpretation given the east-facing coast of the B.A.L., during late Ludlow - early Pridoli time.

A.2 MISSING FLUVIATILE DEPOSITS

The oldest exposures of rocks on Somerset Island that reflect emergence of a continental landmass in Late Silurian (post- siluricus Zone) time, are those strata making up the Somerset Island Formation (Miall et al., 1978). An accurate age assignment is difficult for the Somerset Island Formation. Miall et al. (1978, fig.7) considered it to be almost entirely Pridolian, whereas Thorsteinsson (1980) regarded it as restricted to latest Ludlow. Biostratigraphic data are far from conclusive (see Thorsteinsson 1980, p.26-27), and additionally the lower

contact is in part diachronous (Miall et al., 1978). For the purpose of this report, however, the basal beds of the lower member of the Somerset Island Formation are assumed to be late Ludlovian in age (latialata Zone), and therefore correlative with the lower member of the Barlow Inlet Formation. The various lithofacies of the lower member of the Somerset Island Formation, chiefly planar laminated, platy weathering fine-grained dolostones and limestones, are interpreted to have been deposited on intertidal to shallow subtidal mudflats (Miall et al., 1978). According to Miall and Gibling (1978, p.114), "...tidal flat conditions probably persisted through earliest Devonian time, and there is little or no evidence that fluvial sand bodies built out far enough to the east to enter the sea and form deltas...". This would appear to preclude the possibility that a major trunk river system capable of shedding appreciable volumes of sediment eastward into the Cape Phillips Back Barrier Basin existed on the eastern flank of the arch in Late Silurian time.

Given the reconnaissance nature of geological investigation in this region, a major fluvial deposit could have been missed. Furthermore, the upper member of the Somerset Island Formation does include strata attributed to fluvial sheet flood processes on supratidal flats (Miall and Gibling, 1978, p.105). The upper member rests directly on the Read Bay Formation (Douro) at localities closest to the

Boothia Horst (Miall et al., 1978, fig. 5). Therefore, the westernmost exposures of the upper member fluvial facies could conceivably be equivalent in age to the lower member of the Barlow Inlet Formation.

The Boothia Uplift has an asymmetry of structure and of disposition of time-correlative clastic wedge facies (Kerr, 1977; Miall and Gibling, 1978), on its eastern and western sides. Greatest paleotopographic relief occurred on the western margin, and significantly gentler slopes to the east. Apparently coeval with widespread tidal flat deposits on Somerset Island, the preserved sedimentary record of the western margin of the B.A.L. represents laterally variable coastal environments. Insular uplands with alluvial fans bordered directly on the sea. Coastal barriers and isolated lagoons developed in marine environments peripheral to the coastal fans (see Miall, 1969, fig.47). Miall (1969, p.181) stated that "Red beds ... are not important in the Lower Peel Sound, $\sum$ indicating that fluvial conditions were developed on only a very localized scale". The youthful topography, source proximity, and localized, ?intermittent fluvial conditions are not suggestive of the prolonged weathering and attrition (transport) necessary to produce large volumes of fine siliciclastic detritus. Furthermore, the clay-sized detritus likely would have been carried seaward into the marine environments to the west. With these interpretations, the western margin of the B.A.L. is

problematical as a source for the lower member. Compounding this problem is the uncertain time equivalence between the lower member and basal beds of the Peel Sound Formation exposed on Prince of Wales Island. This point is discussed more fully below.

A.3 TIMING OF EXHUMATION OF CRYSTALLINE BASEMENT

Multi-method modal analyses and normative calculations based on whole rock chemistry suggest that feldspars constitute between 10 and 30 % of the fine-grained sandstone facies of the lower member. In the absence of lithic fragments, this classifies this sandstone as a subarkose (predominantly) to arkose (Pettijohn, 1975). Although substantial feldspar content does not in itself signify first cycle derivation, a high content of extremely fresh feldspars (both microcline and oligoclase) in an interpreted beach deposit, lack of lithic fragments, particularly chert, and presence of unaltered detrital muscovite, are strong evidence for a crystalline source.

Within the pre-upper Ludlovian Paleozoic succession in the region, there are several arenaceous units (see Miall and Kerr, 1980; Thorsteinsson 1980; Soderö and Hobson, 1978), but all are reported to be quartzose. The Helikian Aston Formation of northeastern Somerset Island contains less than 10 % feldspar grains (Tuke et al., 1966). It is

"the basal beds of the Aston Formation that directly overlie the crystalline basement, are immature pebbly

difficult to conceive of a mechanism that could preferentially concentrate feldspar grains derived from quartz-rich sediments. It is also extremely improbable that muscovite could survive more than a single sedimentary cycle, or an extensive transport history. It is concluded therefore that the subarkose facies was derived from a 'relatively proximal' crystalline source or sources, of probable acidic composition. The question therefore arises as to whether or not such a source was exposed in the Boothia arch region during lower member deposition.

The Upper Silurian-Lower Devonian clastic wedge flanking the arch, shows a vertical distribution of clast lithotypes that corresponds in an inverse sense to the stratigraphy of the Uplift itself. This concept of inverse or reversed stratigraphy has been discussed at length by Miall (1969) and Miall and Gibling (1978) amongst others. The time when erosion first penetrated the Aphebian basement, should therefore be reflected in the first occurrence within the clastic wedge, of horst-derived detritus.

On the western margin of the arch, Miall (1969) studied 8 sections of the lower member of the Peel Sound Formation. This member shows extreme variability in thickness and stratigraphic sequence, and no faunal and limited physical correlation is possible between sections. The relative ages

arkoses (Dixon et al., 1971, p.737). If lower levels of the Aston Formation were being eroded, it is extremely probable that the basement itself was also available as a source.

of certain horizons within and between sections remain enigmatic. Nonetheless, Miall (1969) reported that in all the lower member sections, only three conglomerate units contain any basement-derived gneiss clasts, and all three units are close to the contact with the upper member of the Peel Sound Formation. Most sections contain oligomictic conglomerate units with sedimentary clasts only. No samples of sandstone from the lower member of the Peel Sound Formation contained more than 1 % feldspar. Gneiss and gabbro clasts occur in the lowest conglomerate beds in a section of the lower member of the Peel Sound Formation measured by Thorsteinsson (1980, fig.22). However, these beds occur approximately 200 m above the base of the formation and cannot be dated more precisely than late Ludlow - early Pridoli.

Biostratigraphic data are sparse and controvertible for the lower member of the Peel Sound Formation on Prince of Wales Island (see Miall, 1969, p.217-223). The most workable and abundant fossil faunas of both invertebrates and vertebrates, occur in a section at Wilson's Gorge (Broad, 1968). At this locality approximately 10 m above the base of the Peel Sound Formation, a unit rich in heterostracans contains the cyathaspidid ?Ariaspis sp., similar to specimens of this genus in the lower member of the Barlow Inlet Formation. The first clast clearly derived from crystalline basement occurs in a conglomerate 154 m

farther above the base. Vertebrate fossils suggest that this conglomerate bed was deposited in Gedinian time, but this age is brought into question by invertebrate fossils of Silurian aspect 3 m above the conglomerate. These ambiguous age determinations from Prince of Wales Island preclude clearly establishing when erosion first penetrated basement rocks of the Horst.

On Somerset Island, large clasts of igneous and/or metamorphic origin are reported only from the upper conglomerate facies (member 4) of the Peel Sound Formation (Miall and Gibling, 1978, p.97). However the sandstone of member 1 locally contains small metamorphic lithic fragments (Gibling, 1978). Gibling (1978) reported that the sandstones of the underlying Somerset Island Formation are composed exclusively of well-rounded and sorted quartz grains, suggesting derivation from older clastic units such as the Aston Formation.

On Somerset Island, the base of the Peel Sound Formation (member 1) is taken to be earliest Pridolian on the basis of the occurrence of the ostracoderm Hemicyclaspis murchisoni (Thorsteinsson, 1980).

In summary, the Precambrian crystalline basement was breached at least by early Pridoli time; the available evidence neither substantiates, nor negates the possibility that this event occurred earlier.

A.4 CONTRASTING MINERAL SUITES

Petrographic studies by Blackadar (1967) and Brown et al. (1969), indicate that the crystalline basement of the Boothia Horst is uniformly of upper amphibolite to granulite facies metamorphic grade. The medium grade - high grade boundary is defined in part by the breakdown of muscovite in the presence of quartz, and therefore muscovite is absent from granulite terranes (Winkler, 1974). No muscovite, apart from sericitic alteration products of feldspar, is reported from the Boothia Horst. In contrast, the principal non-opaque mineral in the subarkose facies is muscovite, as apparently detrital grains, in fresh flakes up to 200 microns in length. Muscovite is present as an accessory component (less than 1 %) in all thin sections studied.

According to Miall (1969, p.131), garnet is a rather common accessory detrital mineral in the Peel Sound Formation, a logical consequence of its presence as a mineral phase in both felsic and mafic gneisses of the Horst (Miall, 1969, table vii). Garnet's high optical relief and isotropic nature make it readily identifiable in thin section, but no grains of this mineral were discovered in careful examination of 10 thin sections of subarkose facies.

Many quartz grains exhibit 'shadowy' extinction in the Aphebian felsic gneisses of the Boothia Horst (Blackadar, 1967). Miall (1969) reported that quartz grains with undulose extinction are fairly common in the upper Peel

Sound sandstone facies. Within the subarkose facies, only a small fraction of the detrital quartz shows uneven extinction. This discrepancy may indicate that the horst did not serve as a source for the lower member sediments, or, it may simply reflect the small grain size of the subarkose, and hence an increased probability of individual clasts exhibiting an apparent uniform extinction (see Blatt, Middleton and Murray, 1973, p.268).

Collectively, the above evidence suggests that the source of the lower member subarkose facies was not the Boothia Horst. It must be stressed, however, that the horst is heterolithic and the exploitation of a single source rock type of perhaps limited areal exposure, could substantially bias a resulting clastic deposit. It is interesting to note that Gibling (1978, p.79) reported muscovite in Peel Sound sandstones from Somerset Island, and Miall (1969) recorded 'weathered' accessory muscovite in arenites of his sandstone facies. Possibly these reports represent misidentifications, or a source rock with a muscovite mineral phase was/is present within the Boothia arch region.

A.5 ALTERNATIVE SOURCES

The limited evidence available does not allow positive identification of an alternative source area or areas. However several possibilities can be considered, and these are herein discussed. To facilitate analysis of provenance, the following assumptions are made;

1. at locs. 1 and 8, the siliciclastic lower member is thin, lacks coarse detritus, has an open marine character, and is considered to have been distal to the principal region of sediment influx. Therefore source areas do not lie to the north or west, but rather to the south and/or east."
2. the source area consisted, at least in part, of crystalline basement rocks (for reasons discussed previously).
3. apart from the Boothia Horst, and some scattered, volumetrically insignificant Proterozoic gabbroic dykes and Phanerozoic kimberlite pipes, no other crystalline rocks are exposed on Somerset Island. Strata equivalent in age to the lower member on southeastern Somerset Island," the Somerset Island Formation, lie within a structurally undisturbed conformable succession. Therefore in the absence of

" to the northwest on Cornwallis Island in the vicinity of Marshall Peninsula (FIG. 2-1, loc. c), Jackson et al. (1978, p.37) report an unnamed limestone of mid-Pridolian age resting with 'angular unconformity' on the Cape Phillips Formation. The beds beneath the unconformity are Late Silurian in age (pers. comm., R. Thorsteinsson, 1983). The significance of this period of ?localized uplift and erosion in Late Silurian time, and its relationship to the epeirogenesis responsible for lower member deposition is not clear, but does suggest that a northwesterly source for the detritus of the lower member, cannot be completely discounted.

" Evidence presented by Jones and Dixon (1977) that suggests that the Leopold Formation of northeastern Somerset Island is Pridolian in age remains questionable; for the purposes of this paper, the Leopold Formation is considered to be entirely older than the lower member.

unconformities and/or protruding igneous/metamorphic bodies, the area occupied by Somerset Island is not considered to be a potential source, and the lower member source area must be farther east or southeast.

4. the areas presently represented by Devon, Baffin, Cornwallis and Somerset Islands, are assumed to have not moved laterally relative to one another, since the Lower Paleozoic.

The idea that during Late Silurian time, a source for terrigenous detritus lay immediately to the east of the study area, is not novel. To explain observed trends in siliciclastic content of the Leopold Formation, Jones (1975), Jones and Dixon (1975, 1977) and Dixon et al. (1981) postulated the existence of an igneous-metamorphic source terrane of low relief to the east or northeast of present-day Somerset Island. The exotic detrital components of the Leopold Formation comprise principally, angular to subangular quartz, relatively fresh plagioclase feldspar, along with lesser amounts of angular to subangular microcline, muscovite, polycrystalline quartz and chert (Dixon and Jones, 1978).

Potential source areas with crystalline basement rock presently exposed, occur on eastern Devon Island, and northwestern Baffin Island. The basement complex is composed of a variety of rock types (principally mafic and felsic gneisses) that have attained upper amphibolite to

granulite facies regional metamorphism, and are similar in most respects to rocks of the Boothia Horst. The basement complex, of interpreted Aphebian to Archean age, has undergone multiple deformation, culminating in the Hudsonian Orogeny. The petrology of these rocks has been described by Blackadar (1967) and Brown et al. (1969) (Boothia Horst); Roots (in Fortier et al., 1963, p.189-194), Christie (1969, 1970), and Kruprica (1973) (Devon Island); Lemon and Blackadar (1963), Blackadar (1970), and Jackson (1974a,b) (NW Baffin Island).

A survey of reported mineral assemblages of the various rock types in these high grade metamorphic terranes, suggests that that other shield areas besides the Boothia Horst, are equally poor candidates for the provenance of sand-sized detritus of the lower member. All the basement highs, including the Boothia Horst, were tectonically uplifted in Proterozoic time. However it is difficult to ascertain whether any or all of the present shield areas (see FIG. 2-10), were similarly exposed during the Late Silurian. Certainly on a regional scale, no Paleozoic sediments younger than late Middle Silurian are exposed on either Brodeur or Borden peninsulas, on Bylot Island, or on eastern Devon Island, suggesting they either were never deposited, or were deposited and subsequently removed by erosion. Cretaceous and younger sediments are locally preserved in fault block basins, and represent the first

record of post-Silurian deposition in the area. Pre-Upper Silurian successions commonly thin toward the basement highs, and there is no reason to suggest that in Late Silurian and younger time the fundamental paleotopographical situation was different. Kerr (1980) postulated that the Devon Island High continued to be mildly active during early Paleozoic time, but that sedimentation was continuous through Late Silurian and Early Devonian time. Evidence for the latter was not advanced.

A distinctive magnetic anomaly in western Devon Island (McGrath and Fraser, 1981) possibly represents either a discrete buried basement high (FIG. 2-14), or a western extension of the Devon Island High. Unfortunately at present there are no available data on the Siluro-Devonian succession on southeastern Devon, and thus thickness trends and conformability of the strata are unknown. Hypothetical Late Silurian epeirogenic uplift and erosion of the Devon Island High, and/or possibly the Krag Mountains High (Navy Board High), and/or ancestral Byam Martin High, could have resulted in the funneling of fine siliciclastic detritus westward. Given our present state of knowledge, these uplifts represent equally tenable alternatives for the provenance of the lower member.

Appendix B

DESCRIPTION OF PRINCIPAL STUDY LOCALITIES

B.1 INTRODUCTION

The Barlow Inlet Formation is characterized by heterogeneous and variable (both laterally and vertically) assemblages of lithofacies. In general the proportion of Basin Slope Magnafacies increases to the north, and decreases upsection, but within this scheme the number, proportion and position of superfacies between and within sections is variable. The resultant facies mosaic is locally complex, and would be difficult to predict based on an incomplete data base. This part of the paper strives to document facies variation within sections and to provide a physical correlation between sections at the superfacies level. Physical correlation at the field unit level⁴⁰ proved impractical with rare exception. The schematic logs provided (e.g. FIGS. B-1 to B-3) are extremely generalized but include, for convenience of cross-referencing, silhouettes of larger-scale sections on file in the Geology Department, University of Ottawa. The reader is referred to these for detailed lithostratigraphy of the formation.

⁴⁰ field units are defined in Paper 1, and average 3.7 m thick (1400 or so measured).

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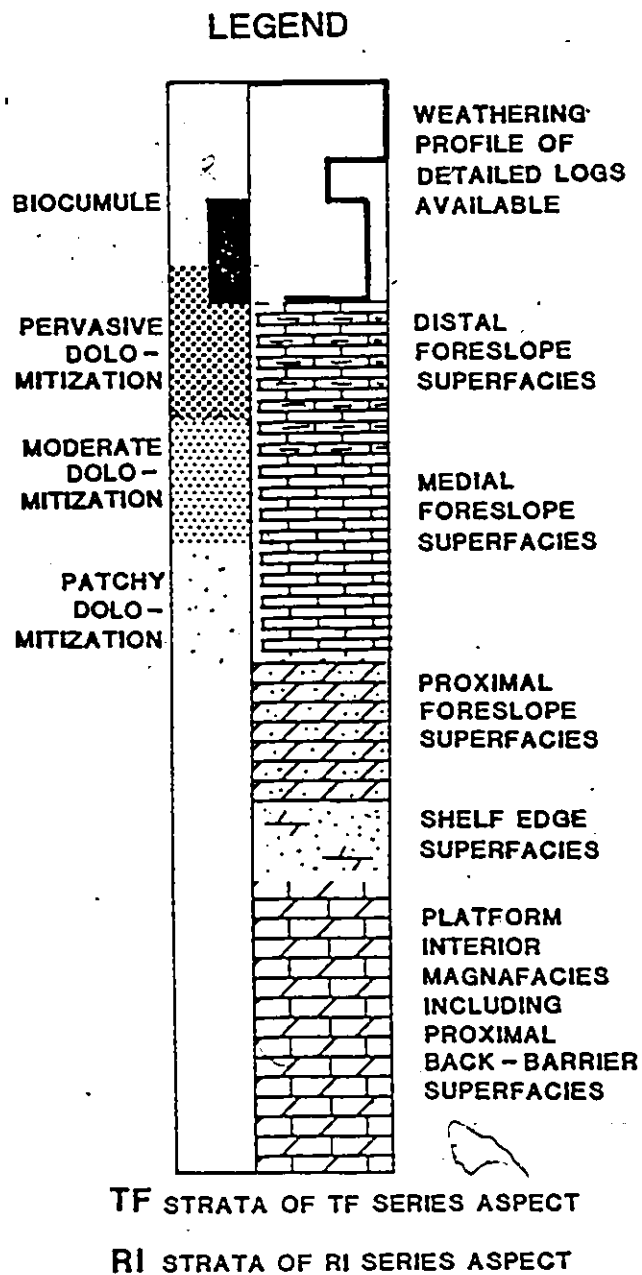


FIG. B-1 Legend for figures B-3, 4, 6, 7, 9 and 11.

TRACE FOSSILS

- Polarichnus 
- Zoophycos 
- Phycodes and related genera 
- Pilichnia 
- Skolithos 
- Pelecypodichnus 

BODY FOSSILS

- Lithistid sponges 
- Calcareous algae 
- Oncoids 
- Bryozoans 
- Articulate brachiopods 
- Gastropods 
- Orthoconic nautiloids 
- Megalodont bivalves 
- Solitary rugosans 
- Fasciculate rugosans 
- Favositid corals 
- Auloporid corals 
- Alveolitid corals 
- Halysitid corals 
- Dendroid tabulate corals 
- Domical stromatoporoids 
- Tabular stromatoporoids 
- Lamellar stromatoporoids 
- Dendroid stromatoporoids 
- Crinoid debris 
- Rhizoidal holdfasts 
- Whole crinoids 
- Trilobites 
- Ostracodes 

STRUCTURES

- Channel 
- Cross-bedding 
- Ripple cross-lamination 
- Graded bedding 
- Fenestrae 
- Desiccation polygons 
- Cryptalgal lamination 
- Hardground 
- Gutter casts 
- Intrastratal folds 
- Intraformational truncation surface 

FIG. B-2 Symbols for figures B-3,4,5,6, and 8.

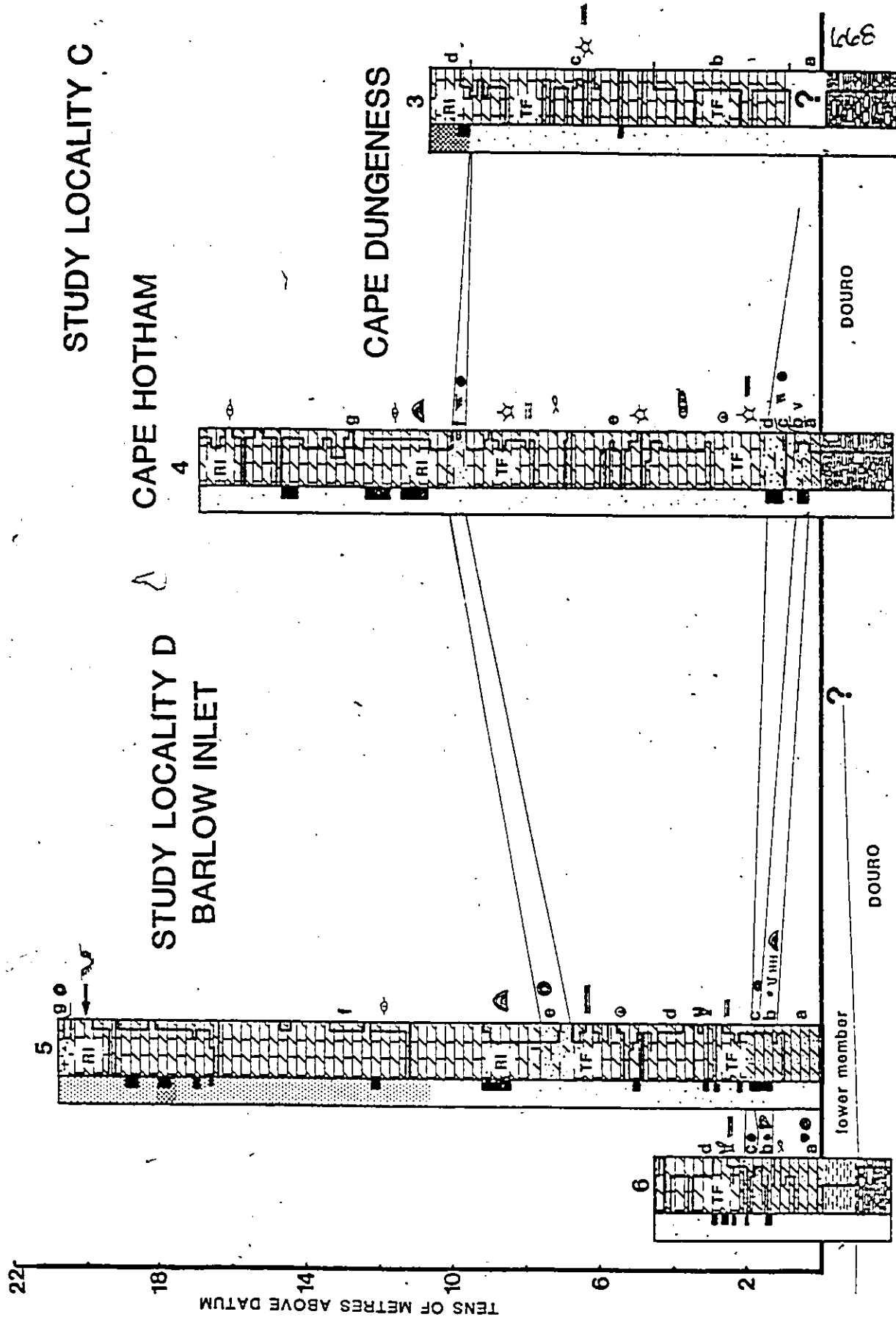


FIG. B-3 Correlation of superfacies in study localities C and D.

B.2 STUDY LOCALITY C -- CAPE HOTHAM -- CAPE DUNGENESS AREA

B.2.1 Exposure and Structure -- section 3 (Cape Dungeness):

Cape Dungeness (FIG. B-3; PL. 5-9a,b) is a small butte formed by resistant dolomites of the upper member overlying less resistant strata of the Douro Formation and a recessive interval with scree lithofacies similar to the lower member of the Somerset Island Formation. Strata dip approximately 10 degrees to the north.

B.2.2 Succession of facies -- section 3:

Unfortunately, interval a (10 m) is covered, and the nature and precise position of the upper contact of the Douro Formation remains unknown. Interval b' (37.4 m) consists of slightly rubbly-weathering faintly mottled dolomitic limestone (Type I) with a sparse fauna of Atrypoides foxi, encrinurid trilobite fragments, solitary rugose corals, Favosites, crinoid debris, orthoconic nautiloids and small high spired gastropods. Trace fossils include Chondrites, Palaeophycus, and Planolites.

The fauna is not environmentally diagnostic, but the presence of A. foxi does suggest this interval lies above the Douro Formation. In consideration of the fact that very similar lithologies in the succeeding interval are interbedded with cryptalgal laminites, it is suggested that interval b was deposited in a shallow, restricted, subtidal environment.

Interval c (51 m) consists predominately of rubbly faintly mottled dolomitic to argillaceous limestone, calcilutite mudstone and desiccated cryptalgal laminite. Oncoids, A. foxi, solitary rugosans, dendroid stromatoporoids, ostracodes and gastropods represent the biotic elements present. Chondrites and Skolithos are the only trace fossils. Crinoidal wackestone and stromatoporoid boundstone are rare lithofacies. Overall this interval resembles the Platform Interior Magnafacies of TF series affinities.

Interval d (11.9 m) consists of massive stromatoporoid boundstone, a pervasively dolomitized, vuggy, crinoid and stromatoporoid rich interval (PL. 5-9c), and massive dololutite. This interval is reminiscent of the proximal back barrier and restricted shelf lagoon superfacies of RI aspect.

B.2.3 Exposure and Structure -- section 4 (Cape Hotham):

Cape Hotham (FIG. B-3; PL. 5-9a) occurs at the southern end of the Cape Hotham Escarpment, a prominent cuesta that parallels the east coast of Cornwallis Island. The Barlow Inlet Formation is particularly well-exposed at Cape Hotham due to a major rock fall that produced an 80 m vertical cliff (PL. 5-9d). Several normal faults of very limited throw are present in section 4, but resultant offset on all faults does not exceed 5 m.

B.2.4 Succession of facies -- section 4:

The contact with the underlying Douro Formation is abrupt but otherwise unnoteworthy. Interval a (4.9 m) consists of crinoid wackestone with few rugose corals, Coenites, strophomenid and rhynchonellid brachiopods, alveolitid corals, and trilobite fragments. This diverse, open marine subtidal biota is suggestive of the proximal foreslope superfacies. Interval b (2.5 m) includes a diverse community mud mound enclosed by crinoidal grainstone and packstone (PL. 5-9 e). Much of the mud mound is recrystallized or composed of sponges, rugose corals, bryozoans and Labechia. However, one polished slab from the mud mound exhibits a stromatoporoid encrusting lime mudstone with a clotted fabric that has micritic structures superficially resembling Renalcis (PL. 5-9f). The surrounding bioclastic rocks have abundant crinoid clasts and solenoporids (PL.5-10a). Interval c (3.4 m) consists predominately of cross-bedded dolomitized crinoidal grainstone (see PL. 5-5g).

Interval d (20 m) is covered. However, in the scree of the basal portions of this interval, are abundant Stachyodes (PL.5-10b), amphiporids, Stereoxyloides (PL.5-10c), and lamellar stromatoporoids. - These concentrations and the absence of units upsection containing the same faunal assemblage, suggest that the fossils were derived from strata beneath interval d. Both intervals c and d are

assigned to the shelf edge superfacies. Interval e (61 m) consists of Platform Interior Magnafacies of TF aspect, contained in a series composed of 13 stacked hemicycles, one of which is illustrated (PL. 5-10d). Synaeresis cracks (PL. 5-10e), cryptalgal laminites and desiccation polygons are common. Oncoids, amphiporids, ostracodes and gastropods are common; ostracoderms such as Torpedaspis elongata (PL. 5-10g) are rare.

Cross-bedded crinoidal calcarenites (PL. 5-10f) constitute interval f (4.1 m). These sands probably represent a basal, transgressive, high energy deposit (see Read, 1982).

The overlying interval (g) is 68.2 m thick (PL. 5-10i) and contains strata assignable to the proximal back-barrier, restricted shelf lagoon, and tidal flat superfacies. This interval is characterized by RI hemicycles featuring massive stromatoporoid biostrome and structureless dense calcilutite mudstone lithofacies (PL. 5-10h,j).

B.3 STUDY LOCALITY D -- BARLOW INLET

B.3.1 Exposure and Structure -- sections 5 and 6:

Sections 5 and 6 (FIG. B-3; PL. 5-11a,b) were measured on a south-facing cliff on the north side of Barlow Inlet. The strata dip uniformly at 30 degrees to the east, and no major structures affect this homoclinal sequence.

B.3.2 Succession of facies -- Sections 5 and 6:

The basal 10.8 m of interval a (14.1 m in total) are covered. The exposed beds consist of rubbly argillaceous to mottled dolomitic limestone (Type II) with crinoid debris, solitary rugose corals, gastropods, solenoporid algae, massive and branching favositids, ostracodes, and indeterminate brachiopods and Catenipora. This boitic assemblage is open marine in character but carries no particular bathymetric significance. Interval b (5.7 m) is a relatively large, weakly-zoned, diverse community mud ~~mound~~ that contains sponges, rhizoidal holdfasts, tabulate corals and stromatoporoids. Stromatoporoids occur in higher concentrations in the upper portions of the mound. Flank beds are crinoid and solenoporid rudstone (PL. 5-11d). The mound is capped by .3 m of argillaceous lime mudstone (PL. 5-11e), in turn overlain by crinoidal grainstone and packstone (interval c) 1.7 m thick. Intervals a through c represent a transition from proximal foreslope to shelf edge superfacies.

Interval d (46.9 m) exhibits characteristics of the Platform Interior Magnafacies of TF aspect, although curiously desiccation polygons have not been recorded from this sequence. Cryptalgal lamination and algal oncoids are very common.

Interval e (4.5 m) consists of crinoidal grainstone and packstone, and is similar to interval f of section 4. The

overlying interval (f) is 133.3 m thick and comprises cyclic strata of RI aspect. The transition from TF to RI series is figured in PL. 5-11f. The RI series of section 5 is dolomitized in most places. The crystallinity of the dolomite varies greatly ranging from medium grey dololutite (otherwise indistinguishable from calcilutite, PL. 5-11g) to coarsely crystalline dolomite associated with stromatoporoid biostromal lithofacies. Several large channel-like bodies occur within massive mudstone at the top of the RI series (see Paper 3, this thesis). Lime mudstone and crinoidal grainstone compose the topmost strata of section 5. Interval g is 2.8 m thick, is distinctively thin-bedded (in contrast to subjacent lithofacies) and has a fossil assemblage of more open marine aspect, including crinoid debris and pluricolumnals, solenoporid algae and Coenites.

Section 6 includes 45.4 m of the upper member. The basal interval of section 6 includes a covered unit 1.5 m thick that immediately overlies subarkosic sandstone of the lower member. Interval a (11.6 m in total) also includes lime wackestone/floatstone (PL. 5-11c), argillaceous lime mudstone and crinoidal packstone. Flora and fauna include crinoid debris and strophomenid, atrypid, and rhynchonellid brachiopods, solitary and cerioid rugosans, Coenites, Alveolites, Ariaspis sp., gastropods and solenoporid algae.

Interval b (8.8 m) is laterally equivalent to the diverse community mud mound of section 5, and consists of rhizoidal

holdfast-fasciculate coral boundstone, crinoid-solenoporid rudstone (PL. 5-11d), and sparsely fossiliferous mottled dolomitic limestone and lime mudstone. Both intervals a and b represent proximal foreslope superfacies. Interval c (1.5 m) comprises crinoid-solenoporid grainstone and possibly represents a thin record of shelf edge superfacies. Interval d (16 m) exhibits characteristics typical of TF series lithofacies, in particular the common occurrence of cryptalgal lamination, algal oncoids and amphiporids. Dendroid stromatoporoids are particularly common in pods and lensoid concentrations near the base of the interval.

B.4 STUDY LOCALITY E -- GOODSIR CREEK - 'SQUARE LAKE CREEK' AREA

B.4.1 Exposure and Structures -- sections 7 to 12:

All sections contain relatively continuous outcrop, and except for the upper 150 m of section 8, exposure is restricted to north and south facing cliffs on the sides of Goodsir Creek and 'Square Lake' Creek (see FIGS. B-4, B-5; PL. 5-12a,b,c,d and f). The upper part of section 8 commences in a stream cut between 'Square Lake' and Wellington Channel (PL. 5-12e). Minor scissor faults present do not greatly affect stratigraphic relationships. Strata dip at 45 degrees to the east.

B.4.2 Succession of facies -- section 7:

Intervals a and b are composed of thin-bedded dark green lime mudstone with scattered crinoid debris, ostracoderm plates, nautiloids and atrypoid brachiopods. These intervals are assigned to the medial foreslope superfacies. Interval c (12 m) comprises crinoidal wackestone (predominately) and lime mudstone with scattered crinoidal debris, overlain by lamellar stromatoporoid bindstone. Atrypoid brachiopods are the only other recorded fossils.

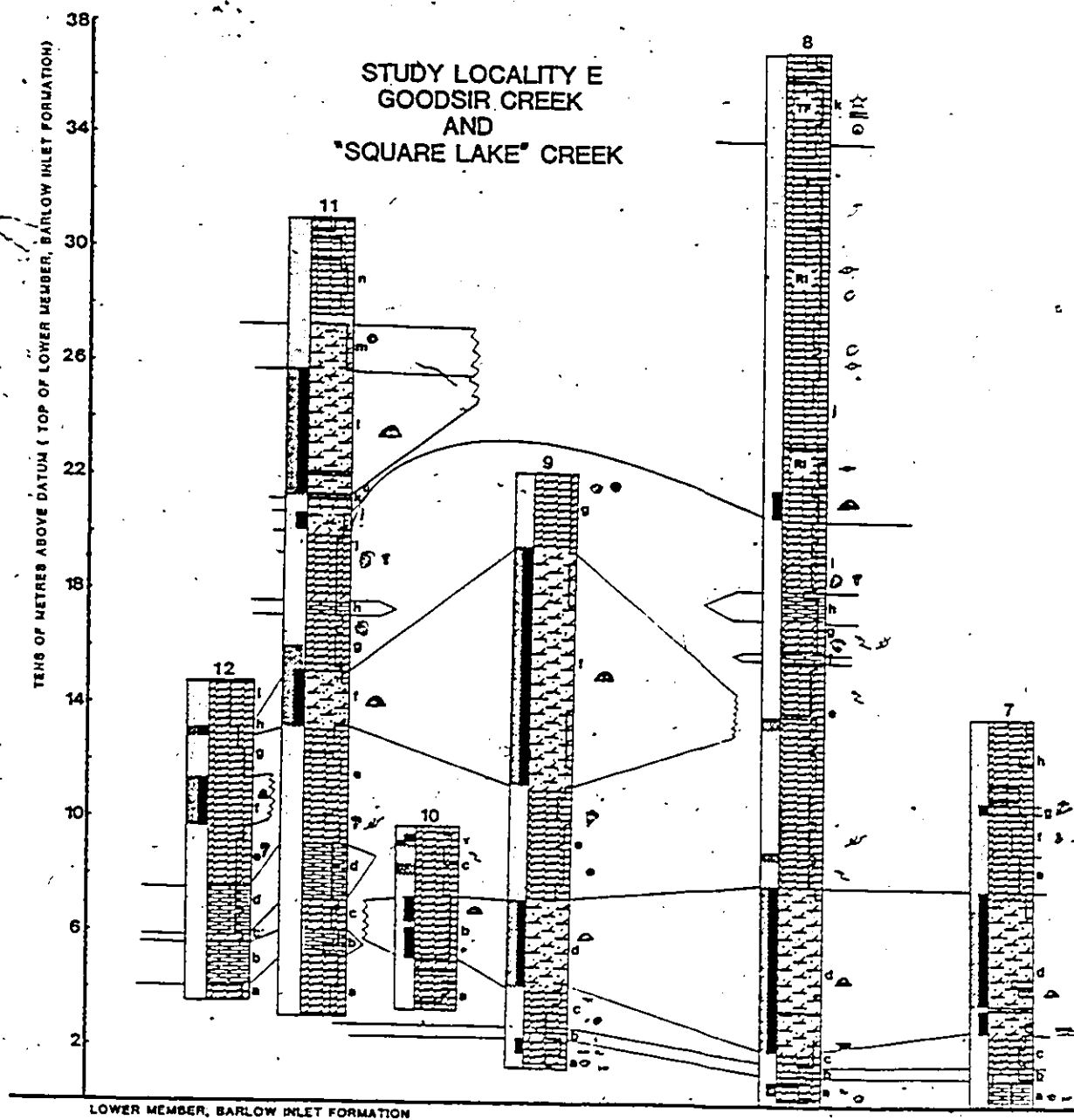
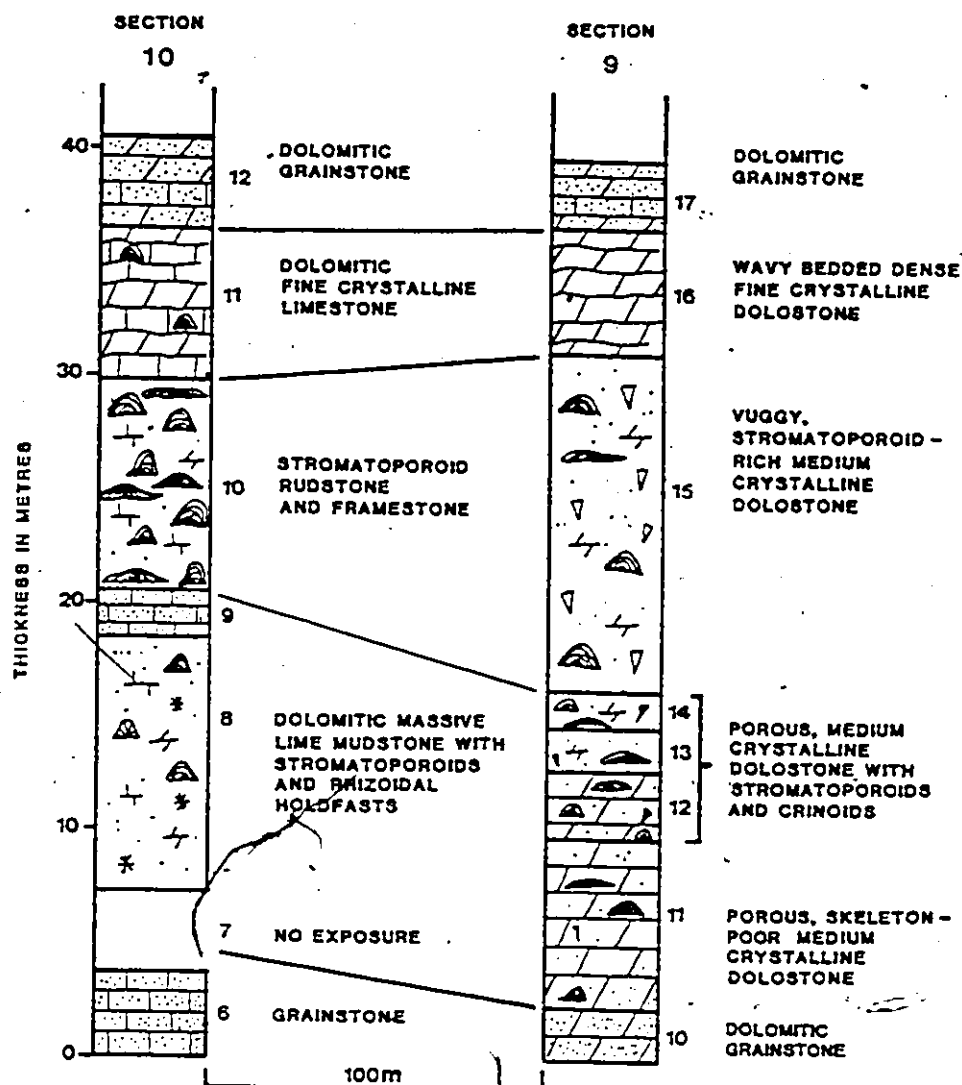


FIG. B-4 Correlation of superfacies in study locality E.

**CORRELATION OF FIELD UNITS
BETWEEN PORTIONS OF SECTIONS 9 AND 10**



▽ SOLUTION CAVITIES (VUGS) AFTER FOSSILS OF LARGE SKELETAL METAZOANS.

CRINOID BIOCLASTS

OTHER SYMBOLS STANDARD, OR USED IN FIGURES IN THIS PAPER.

FIG. B-5 Correlation of field units between portions of sections 9 and 10.

The stratigraphic position, and contained lamellar stromatoporoids of interval c identify it as proximal foreslope superfacies. Interval d (48.5) is a massive skeletal (?crinoid) biocumule with a predominance of lamellar stromatoporoids at the base. Crinoid debris, fasciculate rugosans, domical and dendroid stromatoporoids, and Coenites constitute poorly preserved (dolomitized) fauna. The reader is referred to Paper 4 (this thesis) for a discussion of the possible environmental significance of this biocumule type. Interval d is considered to represent, at least in part, shelf edge superfacies. Interval e (13 m) comprises rubbly to blocky weathering, thin-bedded, unfossiliferous dolosiltite. Apart from its lack of fauna, interval e is similar to the gradationally overlying interval (f) and is assessed a similar environmental significance. Interval f contains a rare to common biota enclosed in partly cross-laminated dolosiltite matrix. Fossils include a remarkable, diverse assemblage of gastropods (see Rohr and Packard, 1982, p.327), crinoid debris, sponge spicules, stylopleurid (fasciculate) rugose corals, massive domical favositids (up to .4 m across) and brachiopods (Nanukidium, Atrypa, Leptaena, strophomenids). This fauna suggests open marine conditions of shallow to moderate depth, and the 9.8 m of interval f are assigned to the proximal foreslope superfacies. Interval g (3.4 m) is biostromal and consists almost exclusively of fasciculate

rugose coral boundstone (PL. 5-12g) with associated pentamerid brachiopods and favositid corals. Thin to medium-bedded, rubbly to blocky weathering, dolomitized but fossiliferous wackestone and packstone with an open marine fauna constitute interval h (32 m). Intervals e through h are assigned to the proximal foreslope superfacies.

B.4.3 Succession of facies -- section 8:

Interval a comprises thin-bedded lime mudstone, crinoidal wackestone with a diverse brachiopod fauna, and diverse community mud mounds (see Paper 4, this thesis) built in part by a variety of tabulate corals including Catenipora and Alveolites. Interval a is assigned to the proximal foreslope superfacies. Directly overlying the mud mounds and their flanking strata are 3.3 m of extremely thin-bedded (PL. 5-12h) marlstone with a very sparse fauna of atrypoid and rhynchonellid brachiopods. The abrupt appearance of this facies may have a non-bathymetric significance (e.g. increased fine clastic sedimentation), but it is here interpreted as an incursion of deeper-water medial foreslope facies. This argillaceous lime mudstone interval can be traced as far south as section 5 and north to section 9 (see part.4.2, 'Platform Design and Evolution', below).

Interval c (6.5 m) represents a return to the proximal foreslope, as suggested by the presence of crinoid-solenoporiid packstone with an associated fauna of lamellar

and dendroid stromatoporoids, Coenites, pentamerid brachiopods and domical stromatoporoids and favositids. Interval d (58 m) is a massive skeletal (?crinoid) biocumule. At its base are dolomitized lamellar stromatopoid boundstone and crinoidal rudstone (PL. 5-12i), as well as zones of dolomitized boundstone and crinoidal rudstone reminiscent of diverse community mud mounds (PL. 5-13a). Interval d is composed of remarkable stratiform units (see PLS. 5-12d, 5-13b). The upper contact of interval d is abrupt (PL. 5-13b) and although interval e does contain a few massive dolomitized crinoidal units (PL. 5-13b), it is predominately interbedded, thin, planar to wavy-bedded calcilutite mudstone, wackestone and crinoidal lime packstone/mudstone. Evidence of slope failure (intrastratal folds, intra-formational truncation surfaces) are present, but carbonate breccias are not. Fossils tend to be concentrated on only certain bedding planes, but where present they constitute a diverse and abundant assemblage including atrypid, atrypoid, pentamerid, and strophomenid brachiopods, domical stromatoporoids, favositids, both cerioid and solitary rugose corals, gastropods, Coenites, and crinoid debris. This interval is assigned to the proximal foreslope superfacies. Intervals f through i are similar to interval e, but f and h are particularly unfossiliferous, thin-bedded lime mudstone intervals that may represent deeper-water conditions. Major

intraformational truncation surfaces associated with plastically deformed strata (PL.5-13c), and the trace fossil Zoophycos, occur in intervals g and i.

Interval j (135 m) consists predominately of fenestral calcilutite mudstone and megalodont bivalve floatstone and inertiasstone. Massive stromatoporoid-dominated inertiasstone biostromes occur at the base of this interval, which in total represents platform interior and back-barrier facies of RI aspect. Interval k (32 m) comprises oncoidal carbonate, massive and laminated calci-dolosiltite and faintly-mottled rubbly-weathering unfossiliferous mudstone, lithofacies characteristic of platform interior facies of TF aspect. PL. 5-13d illustrates one TF hemicycle observed in section 8.

B.4.4 Succession of facies -- section 9:

Intervals a,b,c, and d (21.3, 3.5, 12.5, 37 m, respectively) of sections 8 and 9 are essentially similar. As in section 8 a thin bedded mudstone envelopes the diverse community mud mounds of interval a (PL. 5-13e). Crinoidal rudstone and packstone are more extensively developed in section 9 below the first massive dolostone. Interval c contains rhizoidal holdfast mudstone pods and cross-bedded units not observed in the equivalent interval in section 8.

Interval e (28 m) contains crinoid-solenoporid packstone, grainstone, and slightly-dolomitized, thin-bedded wackestone

and floatstone containing fasciculate rugose corals, pentamerid brachiopods, crinoid debris, domical stromatoporoids and gastropods. Slope failure phenomena are present. Interval e is assigned to the proximal foreslope superfacies.

Interval f (69.3 m) is identical to interval d and is a shelf edge dolomitized massive skeletal (?crinoid) biocumule. Interval g (43.4 m) is a finely crystalline dolostone that contains locally, a remarkably well-preserved and diverse silicified open marine benthic fauna. Atrypid and atrypoid brachiopods are particularly abundant. Atricolated crinoids, rugose 'button' corals, stromatoporoids, gastropods and orthoconic mautiloids represent other faunal remains. Interval g is assigned to the proximal foreslope superfacies.

B.4.5 Succession of facies -- section 10:

Section 10 is located only 100 m to the north of section 9 (PL. 5-13f). Interval a (12.8 m) of section 10 is identical in all respects to interval c of sections 9 and 8. Interval b (22.0 m) is equivalent in stratigraphic position to a dolomitized massive skeletal (?crinoid) biocumule (DMSB) of section 9 (interval d), but is not as severely dolomitized. Comparison (summarized in FIG. B-5) between these two intervals is useful as it provides further insight into DMSB makeup. The differential dolomitization of these

two intervals may not necessarily be fortuitous: it could indicate differences in original makeup (e.g. muddier or sandier matrix in the respective biocumules), or even paleobathymetric position. Therefore the equivalencies indicated in FIG B-5 should be considered provisional. Interval c (13.1 m) is similar to interval e of section 9. The only notable difference in detail is the presence of a thin unit (less than 1 m) of lithistid sponge boundstone in interval c of section 10.

B.4.6 Succession of facies -- section 11:

Intervals a through d (22, 9, 11, 20 m) are the product of interfingering of proximal and medial foreslope facies represented by crinoid-solenoporid packstone and wackestone, and planar thin-bedded unfossiliferous mudstone, respectively. Interval e (40 m) is composed of domical stromatoporoid-rich crinoidal packstone and rudstone and thin to medium, planar to wavy-bedded lime wackestone and mudstone (the latter commonly with chert nodules -- PL. 5-13h), with a diverse open marine fauna. Intraformational truncation surfaces and exotic boundstone blocks are associated (PL. 5-13g). Interval f (20 m) is similar in all respects to unit d of sections 8 and 9.

Intervals g through i (20, 5, 23 m) can also be interpreted as the product of interdigitation of proximal and medial foreslope superfacies. Fossiliferous units are

rich in a diverse open marine fauna, and also contain the trace fossils Palaeophycus, Chondrites, (PL. 5-13i) and Zoophycos in some abundance.

Interval j is the thin stratigraphic record (5.0 m) of a stromatoporoid shelf edge barrier reef that exhibits a conspicuous, vertical fauna zonation. This reef is described in detail in Paper 4 (this thesis). Overlying the reef are fenestral lime mudstone and megalodont bivalve floatstone of a restricted back reef lagoon. This lagoon interval (j) is 4.6 m thick. Interval l (45 m) is a DMSB that approaches a stromatoporoid-crinoid rubble skeletal buildup (see Paper 4) in makeup and preservation of fauna. Interval m (15 m) consists of dolomitized, bedded crinoidal packstone/rudstone with abundant domical stromatoporoids, and cylindrical and domical favositids. Interval m represents either a shelf edge shoal or proximal foreslope blanket deposit.

Interval n (35 m) comprises discontinuous sparsely fossiliferous mottled dolomitic limestone, dolosiltite, and calcisiltite mudstone. Faunal diversity decreases upwards. In this regard interval n bears some resemblance to the lower part of section 18 (study locality F) which is tentatively considered to represent Platform Interior Magnafacies. The faunal paucity of interval n possibly suggests a restricted lagoonal environment.

B.4.7 Succession of facies -- section 12:

Section 12 is located some 200 m south of section 11. Intervals a through d correspond to the same intervals of section 11 with some thickness variation (5, 15, 3, 18 m, respectively). Interval e (19.5 m) is composed of argillaceous floatstone (basal 15 m) with fasciculate rugose corals, domical stromatoporoids and crinoid debris, and crinoidal packstone/grainstone (upper 4.5 m) with atrypid brachiopods and domical, lamellar and dendroid stromatoporoids. The upper part of interval e resembles interval c of sections 8 and 9. The fasciculate rugosans occur locally in boundstone concentrations, and dislodged blocks of this material are present in interval e of section 11 (see above). Interval e represents proximal foreslope environments. Interval f (17 m) is the core of a stromatoporoid-crinoid rubble skeletal buildup (proximal foreslope superfacies). Interval h (3.5 m) is a similar buildup, whereas the intervening, and suprajacent intervals g and i (14 and 16 m, respectively) are composed of bedded crinoidal packstone, crinoidal mottled dolomitic wackestone, and crinoidal dolarenite. Lamellar, dendroid and domical stromatoporoids, as well as atrypid and pentamerid brachiopods are associated with the predominant crinoid debris in this lithofacies. Intervals e through i all represent deposition in the proximal foreslope environment.

B.5 STUDY LOCALITY F -- DEPOT POINT

B.5.1 Exposure and Structure -- sections 13 to 18:

Sections 13 through 16 are located on a north-facing cliff on the south side of the lake near Depot Point (FIG. B-6, PL. 5-14a) Section 17 is in the stream gorge between the lake and Wellington Channel. Section 18 is on the Cape Hotham Escarpment just north of the lake. All six sections contain relatively continuous exposure, with strata dipping 30 degrees to the east. Stratigraphic offsets on minor scissor faults are negligible.

B.5.2 Succession of facies -- sections 13 and 14:

Intervals a, b and c of sections 13 and 14 are similar and can be considered collectively. The lowest exposed stratum of section 13 is estimated to be approximately 20 m above the top of the lowest member. Interval a (43.8, 31.9 m) is composed of thin bedded mudstone (predominantly), wackestone, carbonate breccia, lithoclastic crinoidal grainstone and rubbly

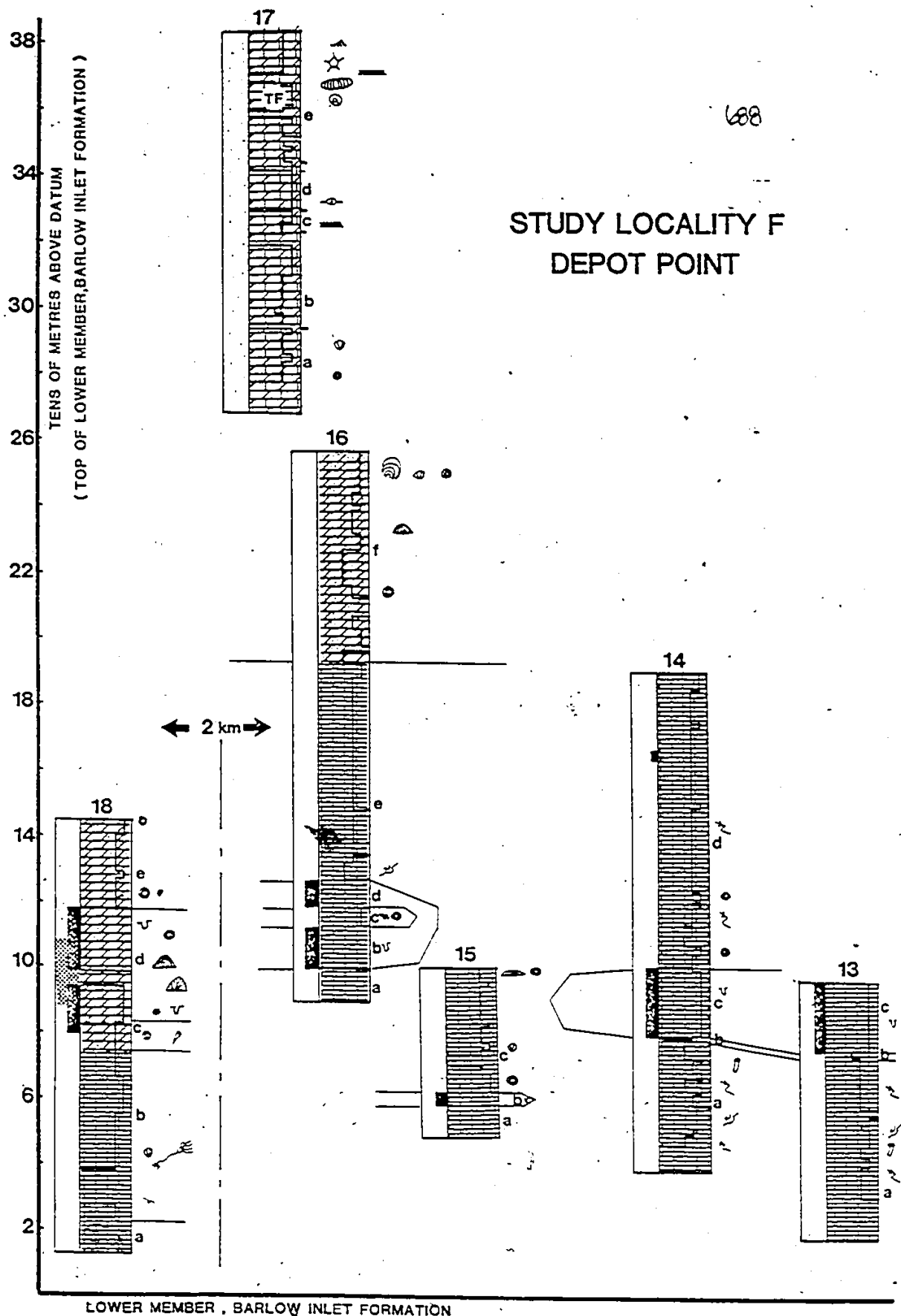


FIG. B-6 Correlation of superfacies in study locality F.

argillaceous limestone, type II. Principal biotic constituents include crinoid debris, domical favositid corals, solitary and atrypid brachiopods. Less common fossils include Coenites, pentamerid, rhynchonellid, strophomenid and atrypoid brachiopods, bryozoans, mucophyllid corals, cylindrical favositids and stromatoporoids. The principal trace fossils are Pilichnia (PL. 5-14e), Skolithos and Planolites. Soft sediment deformation structures are abundant. Gutter casts occur on the bases ~~of~~ lithoclastic crinoidal grainstone/rudstone units. Interval a is assigned to the medial foreslope superfacies.

Interval b (2.0, 1.0 m) consists solely of polymictic, ungraded and unsorted, matrix-supported(?), carbonate breccia. The predominant clast type is lithistid sponge boundstone, but skeletal metazoans, and tabular clasts of grainstone and hemipelagic mudstone are also present. The breccia has a scoured base and an irregular upper surface. It likely represents deposition from a debris flow rather than rock fall in light of the absence of shelter voids and laminated sediment fill between larger clasts, the presence of an erosive basal surface, and the incorporation of non-reefal, hemipelagic mudstone clasts from subjacent strata.

Interval c (20, 19 m) is a lithistid sponge-dominated mud mound. The reader is referred to Paper 4 (this thesis)

for a detailed description of this biocumule facies. Sponge mud mounds are here considered typical of the medial foreslope.

Interval d (88 m) of section 14 is similar to interval a and is also considered to be part of the medial foreslope superfacies. Plate 5.14d is a view looking up at section 14, encompassing all 4 intervals described above.

B.5.3. Succession of facies -- section 15:

Interval a (10 m) is largely covered. Interval b (3.6 m) consists of crinoid packstone, lateral to lithistid sponge boundstone. Where crinoid-rich sediments accumulated immediately adjacent to sponge mud mounds, they are commonly unsorted, and muddy (PL. 5-14g) with beds tapering abruptly away from the mound margin, which is itself generally rich in rhizoidal holdfasts. Laterally extensive, thin sheets of crinoidal packstone/ rudstone (PL. 5-14f, arrow) are interpreted to represent storm reworked flank bed material, or less likely mud-poor mass flow deposits.

Interval c (37 m) is similar to intervals a and d of section 14. Crinoidal rudstone and packstone are notable lithofacies and predominate at the top and base of the section, accounting for 33% of total section thickness.

B.5.4 Succession of facies -- section 16:

Interval a (10 m) is covered. Intervals b and d are lithistid sponge boundstone (12.6 and 8.5 m, respectively) separated by 5.6 m (interval c) of planar cross-bedded lithoclastic crinoidal grainstone. Interval d (approximately 70 m) is similar to interval d of section 14 with predominantly rubbly to blocky weathering mudstone and wackestone with an open marine fauna of limited diversity. Rare slope failure phenomena are recorded, but beds of lithoclastic crinoidal packstone are common. The boundary between intervals e and f is arbitrary. Sixty-six percent of the thickness of interval f (57.8 m) contain greater than 5% fossil material⁴¹ whereas 77% of interval e contains less than 5% skeletal allochems. Interval f contains a diverse benthos and the trace fossil Zoophycos in its uppermost 15 m. Interval f contains no slope failure structures, and is tentatively assigned to the proximal foreslope superfacies.

B.5.5 Succession of facies -- section 17:

A covered interval 10 to 20 m thick separates the top of section 16 and the first exposure of section 17. The latter section commences approximately 270 m above the base of the upper member. Interval a (15.6 m) comprises mottled dolomitic floatstone and wackestone (Types I, II and

⁴¹ based on field observations, and therefore reflecting the the distribution of macroscopic (i.e. medium sand size or {larger} allochems.

labyrinthine), with a predominantly brachiopod fauna (Atrypoides foxi, A. erebus, atrypids, strophomenids, along with gastropods, bryozoans and crinoidal fragments). The strata are thin to medium bedded. The litho- and biofacies are relatively non-diagnostic, although the predominance of stenohaline organisms does suggest environmental conditions of more open circulation.

Interval b is equally enigmatic, featuring thin-bedded, rubbly mottled limestone with sparse atrypoid brachiopods, gastropods, ostracodes and orthoconic nautiloids, but no crinoid debris. These lithofacies are vaguely reminiscent of shallow ramp lithofacies of the Douro Formation, but are anomalous facies with respect to the Barlow Inlet Formation. Interval c (6.8 m) consists of cryptalgal laminite interbedded with stromatoporoid-tabulate coral floatstone. Interval c represents the first occurrence of tidal flat superfacies in study locality f. Interval d (12.5 m) is well-exposed but largely inaccessible. It consists of resistant, (well-bedded?) calcilutite mudstone with fenestral fabric developed in the accessible uppermost portion of the unit. Interval e (40 m) is composed of rubbly argillaceous and mottled dolomitic mudstone and rare wackestone, oncoidal packstone (PL. 5-14i) and cryptalgal laminite. The trace fossils include Chondrites, and Polarichnus, and structures include lamination, desiccations polygons, and small symmetric ripples (adhesion ripples? --

PL. 5-14j). Of note is the preservation of unbroken Coenites in upright positions, suggesting exceptionally rapid burial (PL. 5-14h). The presence of oncoids, mottled lithologies and cryptalgal laminite are characteristic of TF hemicycles, and interval e is so assigned. Intervals d and c have questionable RI affinities (stromatoporoids, fenestral fabric) but these features are too ill-defined to warrant designating these intervals more specifically than to the Platform Interior Magnafacies. Intervals a and b are also provisionally included in the Platform Interior Magnafacies, although their lithofacies are atypical. The absence of shelf edge superfacies between, or in, sections 16 and 17 remains a critical problem. The presence of a major forestepping event at the time represented by the beds transitional from 16 to 17 is invoked to reconcile this discrepancy.

B.5.6 Succession of facies -- section 18:

Interval a (10 m) is poorly exposed and consists of unfossiliferous thin-bedded lime mudstone tentatively assigned to the distal foreslope superfacies. Interval b (51.3 m) comprises lime mudstone, wackestone, floatstone and faintly mottled dolomitic wackestone. The fauna is dominated by crinoids and brachiopods and includes whole crinoids and atrypoid brachiopods (PL. 5-14b), orthid, strophomenid, atrypid and rhynchonellid brachiopods, domical

and dendroid favositid corals, bryozoans, ostracoderms, and orthoconic nautiloids. Only the ichnogenus Palaeophycus is recorded from this interval. Most of the fauna is relatively whole and unbraded, and the predominance of a lime mud matrix, argues for accumulation below wave base. This interval is assigned to the medial foreslope superfacies. Interval c (7.5 m) occurs just beneath the first biocumule of section 18, and is composed of crinoid and Coenites -rich wackestone. Interval d (34 m) comprises two successive stromatoporoid crinoid rubble skeletal buildups, separated by bedded crinoidal wackestone. Interval e (26 m) consists of bedded units of crinoidal wackestone to packstone with solenoporid algae, domical stromatoporoids and favositids, solitary rugose corals, and atrypoid brachiopods. Intervals c through e are assigned to the proximal foreslope.

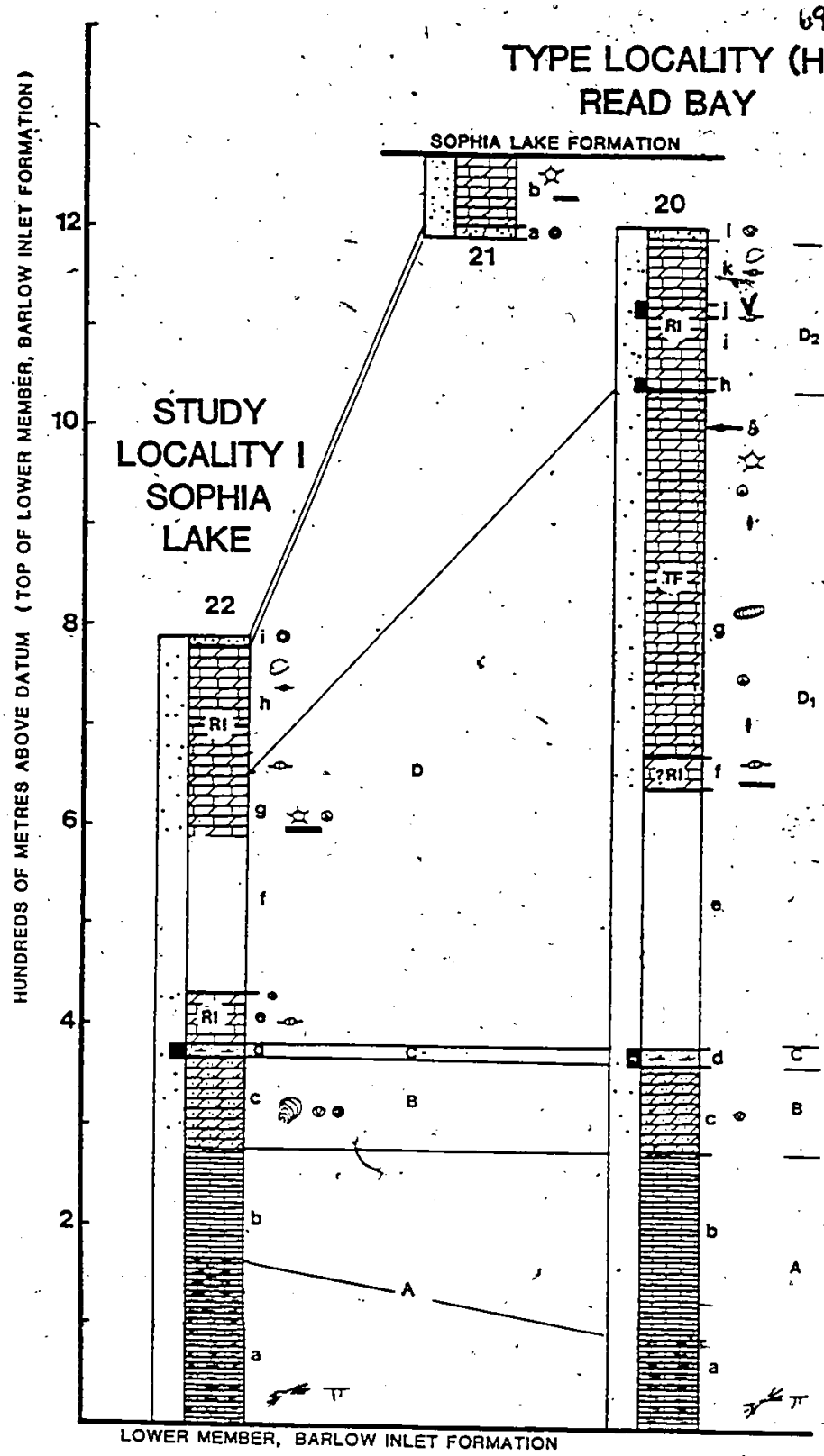
B.6 STUDY LOCALITY H -- READ BAY

B.6.1 Exposure and Structure--sections 20 and 21 (upper member type section)

The type section of the upper member is in the gorge of a stream that flows to the northeast from the Cape Hotham Escarpment, emptying into the southwestern part of Read Bay (FIGS. B-7,8; (PL. 5-15a,b). The base of the section is at the Cape Hotham Escarpment, which in this area forms a low bluff only 10 m or so above the surrounding plateau. The top of section 21 is marked by an abrupt color change at the first occurrence of dusky red-colored siltstones of the Sophia Lake Formation (PL. 5-22). The continuity of outcrop is variable. It is particularly poor for the basal 638 m, whereas for the remaining 634 m there is 73% outcrop (i.e. the gorge cuts deeper, see PLS. 5-15c, 5-16k,l). Fortunately facies variations are minimal in the lower part of the section and the felsenmeer provides adequate control for the gross lithostratigraphy of this interval. No major structures disturb this homoclinal sequence. Strata gently dip to the ENE. Inclinations of 19 degrees were measured at base and top of section, and dip measurements do not vary by more than 5 degrees (15 to 20 degrees).

B.6.2 Succession of facies -- sections 20 and 21:

Interval a (approximately 85 m) of section 20 is composed of unfossiliferous, rarely bioturbated, even planar-bedded calcilutite mudstone with very thin argillaceous partings.



Phycodes and Taenidium are the predominant trace fossils. This interval is assigned to the distal foreslope superfacies. Interval b (approximately 225 m) is similar to interval a but contains rare to common interbeds of wackestone with an open marine fauna including crinoid debris, strophomenid and atrypid brachiopods, solitary rugosans, low domical favositids, bryozoans, trilobites (Hemiargus sp.) and orthoconic nautiloids. Skolithos is the predominant trace fossil. Interval b is assigned to the medial foreslope. It must be emphasized that the transition between distal and medial foreslope superfacies is extremely gradational; for this reason interval thicknesses are approximate only. Interval c (70 m) consists of faintly-mottled lithofacies with yet more diverse and abundant benthos, including the first occurrence of calcareous algae (Solenopora), atrypoid brachiopods, and stromatoporoids. The shelf edge facies (interval d -- 5 or more metres) is composed of tabulate corals, lamellar to hemispherical stromatoporoids and cerioid rugosans in biostromal deposits (PL. 5-15d). The frame construction of adnate laminar stromatoporoids (PL. 5-15e) possibly suggests adaptation to turbulent, low sedimentation conditions (James, 1983, Fig.59).

Interval e (258 m) is largely covered. Isolated outcrops within this interval do contain either stromatoporoid boundstone, or mottled carbonate lithologies with a diverse

stenohaline fauna. This interval conceivably represents the interdigitation of various platform margin facies, but this remains unverified.

Interval f (25 m) commences with a unit of cryptalgal laminite, the first stratigraphic occurrence of this facies in section 20. Occurring above the laminite are units of fenestral calcilutite mudstone, and mottled dolomitic limestone with rare, small domical stromatoporoids. These lithofacies are reminiscent of platform interior magnafacies of RI aspect. The ensuing 371 m (interval g) consist of platform interior hemicycles of TF affinities (PL. 5-15f). Shallowing upward sequences of section 20 differ from TF hemicycles described in sections 4 and 5 in the following ways:

1. Hemicycles are thicker -- average of about 12 m.
2. Although their basal member is oncoidal, it is commonly rich in small domical stromatoporoids, rugose corals, amphiporids, gastropods and indeterminate bivalves.
3. Mottling is more distinct.
4. Massive calcilutite does not occur.
5. Ostracoderm debris is somewhat more prevalent in laminated and massive calci/dolosiltites.

Fossils of stenohaline organisms are very rare. The principal trace fossils are Skolithos (PL. 5-16b),

Polarichnus and Pelecypodichnus sp. B⁴² (PL 5-16b).

Recrystallized rudstone to floatstone of large muchisonid gastropods and oncoids form rare but conspicuous units (PL. 5-16c); one such unit serves as a useful reference bed within the type section, 1000 m above the base (its position is indicated by an arrow on FIG. B-7).

Interval h (8.0 m) is principally a biocumule with abundant domical and adnate lamellar stromatoporoids, and solitary rugosans (PL. 5-16d). Local framework is present, particularly towards the top of the biocumule. The geometry of this biocumule is not known, but the absence of crinoid debris, the co-occurrence of hemishperical stromatoporoids and Nanshanophyllum (solitary rugose coral), and the presence of fenestral lime rudstone and megalodont bivalves in the enclosing units, are all proximal back-barrier traits. Frame construction and the presence (but not predominance) of adnate lamellar stromatoporoids are more characteristic of shelf edge biocumules, suggesting that interval h was deposited in an environment transitional between these two superfacies. PL. 5-16e illustrates that portion of section 20 containing interval h.

Interval i (70 m) contains fenestral mudstone, massive skeletal biostromes, and mottled dolomitic limestone (Type I). Domical tabulate corals, solenoporids and crinoid

⁴² distinguishing between sections through steikerns of infaunal bivalves, and sediment-filled dwelling burrows (possibly with burrow margins), of the same clams, is difficult. In all likelihood, both are present.

debris are common faunal constituents in the upper part of this interval, suggesting that more open marine conditions prevailed. Brachiopods are conspicuously absent.

Interval j (8.5 m) consists of a Coenites -capped fenestral mud mound, described in detail in Paper 4 (this thesis), and illustrated in PL. 5-16g). This mound-shaped deposit of fenestral calcilutite is covered by floatstone of dendroid tabulate corals. Interval k (65.5 m) comprises fenestral calcilutite, massive stromatoporoid inertiastone, megalodont bivalve inertiastone, and mottled dolomitic limestone (Types I and II). The transition from interval k to interval l is illustrated in PL. 5-16k,l. Intervals h through k exhibit affinities toward RI lithofacies of the Platform Interior Magnafacies.

Interval l (13.6 m) is composed of thin medium-bedded mottled dolomitic limestone, with a biota suggestive of ameliorated circulation on the shelf (transgression?). Atrypoides polaris, Hemiarhes bigener, strophomenid brachiopods, orthoconic nautiloids, domical and cylindrical favositids, and Coenites are present. This interval is equivalent in stratigraphic position to interval a (7.7 m) of section 21 which is composed of crinoid-solenopodid packstone to rudstone with adnate lamellar and small domical favositids, and brachiopod shell debris. Small pods (.3 x .2 m) of massive lime mudstone occur within this facies. Adjacent beds are draped over these bodies, which are

reminiscent of rhizoidal holdfast pods of the proximal foreslope superfacies. Interval n is interpreted as a transgressive deposit marking a major backstepping event affecting the Barlow Inlet Platform. Interval b (71 m) of section 21 consists predominately of thin-bedded calcilutite mudstone and wackestone. The basal 35 m is unfossiliferous, generally laminated, and commonly exhibits desiccation polygons; the upper 35 m is also commonly laminated although evidence for subaerial emergence was not noted in these beds. Hardgrounds are present and 54% of field units are fossiliferous containing a largely brachiopod fauna (strophomenids, rhynchonellids, smooth spiriferids, and orbiculoids). Trilobite fragments, ostracodes, bryozoans, orthoconic (nautiloids, oncoids, large murchisonid gastropods, ostracoderm fragments, and cerioid rugosans complete the biota. Apart from the principal calcisiltite lithofacies (PL. 5-16i), both quartzose peloidal dolarenite grainstone (PL. 5-16h) and intraclastic rudstone are present. The quantity of silt and fine quartz sand incorporated in these lithofacies is highly variable (up to 20%), but is insignificant when compared to the first red-colored calcareous siltstone of the Sophia Lake Formation. This unit which conformably overlies the last upper member unit of section 21, contains 40 to 60% fine angular quartz sand (PL. 5-16j). The total thickness of the upper member of the Barlow Inlet Formation as its type section is 1272 m.

B.7 STUDY LOCALITY I -- SOPHIA LAKE

B.7.1 Exposure and Structure -- section 22:

Section 22 is in a stream gorge that commences at the Cape Hotham Escarpment and terminates just above the westerly inlet to Sophia Lake (FIG. B-7; PL. 5-17a). The succession contains no major displacements. Strata dip on average 17 degrees to the east (range 17 to 20 degrees).

B.7.2 Succession of facies -- section 22:

Intervals a (155 m) and b (84 m) are virtually identical to the similarly labeled intervals of section 20, and their description can be little embellished here. The distal foreslope superfacies is finely laminated and contains hardgrounds, and evidence of patchy submarine-lithified, but unexhumed, sediments. Uncollapsed, spar-filled Skolithos penetrate calcilutite mudstone that exhibits diffuse dolomitized zones (PL. 5-17c). The highly irregular weathered upper surface neatly resolves the distribution of pure micrite and dolosiltite. The upper surface is neither bored nor encrusted and was therefore unlikely to have been a hardground. The uncollapsed nature of Skolithos attests to an originally cohesive sediments (partially lithified?), and the patchy dolomitization suggests the original presence of permeable and impermeable (i.e. cemented) zones in the sediment. Fe/Mn encrusted burrow margins are also common (PL. 5-17b). One anomalous lithofacies that occurs within

the distal foreslope lithofacies is an Fe-rich crinoidal dolarenite grainstone. This unit does not exceed 2 to 3 m in thickness but can be traced on airphotos for the equivalent of 2 km by virtue of its lighter-colored and resistant nature. In all likelihood this stratum is also present in section 20, but was missed due to poor exposure. Possibly this unique stratum can be related to the regressive-transgressive platform event responsible for the development and then drowning of crinoid blanket and stromatoporoid-crinoid buildup facies at locality G, some 14 km to the south. Interval c (85 m) is somewhat better exposed than its counterpart in the type section. It consists of faintly mottled dolomitic limestones with a rare to very abundant biota consisting in part of whole and disarticulated crinoids, atrypid, strophomenid, rhynchonellid, and spiriferid (but not atrypoid) brachiopods, massive favositids typically with a low domical growth form, solitary rugose corals, and the trilobite Hemiargus. The trace fossil Zoophycos is common. Interval d (20 m) comprises a stromatoporoid-tabulate coral barrier reef (PL. 5-17d), described in detail in Paper 4 (this thesis). Immediately above the barrier reef are massive fenestral calcilutite mudstone, Amphipora floatstone (PL. 5-17e) and massive stromatoporoid boundstone of lagoonal aspect, that constitute most of the interval e (49 m). The upper 6 m of interval e is composed of mottled dolomitic

limestones (PL. 5-17f) containing a restricted biota of oncoids, Coenites, bivalves, solitary rugosans, and domical and tabular stromatoporoids. The topmost unit is resistant, massive fenestral dolostone with stromatoporoids and rugosans at its base. The predominance of fenestral carbonates in interval e suggests lithological affinities to Platform Interior Magnafacies of RI aspect, although cyclicity is not apparent.

Interval f (160 m) is covered. Interval g (56 m) also includes a substantial covered interval (upper 40 m). The basal 16 m comprises cryptalgal laminite and oncoidal mottled dolomitic limestone with gastropods, solitary rugosans, ostracodes, amphiporids, and the trace fossil Skolithos, and Pelecypodichnus sp. B. This lithofacies and fossil assemblage is similar to that described for the TF series in section 20.

Interval h (136 m) is composed of massive, fenestral and dense calcilutite mudstone, cryptalgal laminite (PL. 5-17g), floatstone of megalodont bivalves (PL. 5-17h), mottled dolomitic limestone (Type I) and rare grainstone and rudstone. Possible brachiopod shell debris occurs in only 1%, and crinoid debris in 23% of all units in this interval. This interval is assigned to platform magnafacies of RI aspect. Interval i (3 or more metres) occurs above 50 m. of calcilutite mudstone in the underlying interval h, and comprises crinoidal wackestone and packstone similar to

interval a of section 21, with which it is correlated. Exposure is extremely poor above this interval, and the Sophia Lake Formation lower contact is drawn (PL. 5-17a,i) at the first reddish-colored felsenmeer. This contact coincides approximately with the boundary indicated by Thorsteinsson (1980, Fig. 4).

Thickness estimates of the covered interval at the top of section 22 range from 50 to 175 m, and therefore the total upper member thickness at section 22 does not exceed 965 m.

B.8- STUDY LOCALITY J -- CAPE RESCUE AREA

B.8.1 Exposure and Structure -- Sections 23 to 25, 27 to 32:

The northernmost exposures of the Barlow Inlet Formation occur in the Cape Rescue area (FIG. B-9). Nine sections were measured in this area and include rocks assigned to all three magnafacies. The mapped position of the facies boundary that separates the Cape Phillips Formation from the Barlow Inlet Formation (PL.5-18a) is approximate. Thorsteinsson (1973) has included more of the western part of the study area in the Barlow Inlet Formation (e.g. sections 23, 24); in either case, the boundary is diffuse and its position is arbitrary. The interpretation offered here is based on the conclusion that the basin foreslope superfacies is the lithological equivalent of the C member of the Cape Phillips Formation.

From a crude graphic estimate, the base of section 23 is about 125 m above a datum represented by the Douro-Barlow Inlet contact extrapolated north of Snowblind Creek. The rocks are thus all younger than the siluricus zone (late Ludlow). Conodonts from beds 540 m above the base suggest a mid-Pridolian age (W.G. Parkins personal communication, 1983). The uppermost exposed beds in the study locality are estimated by graphic means to be approximately 1300 m above the datum. Their age is unknown.

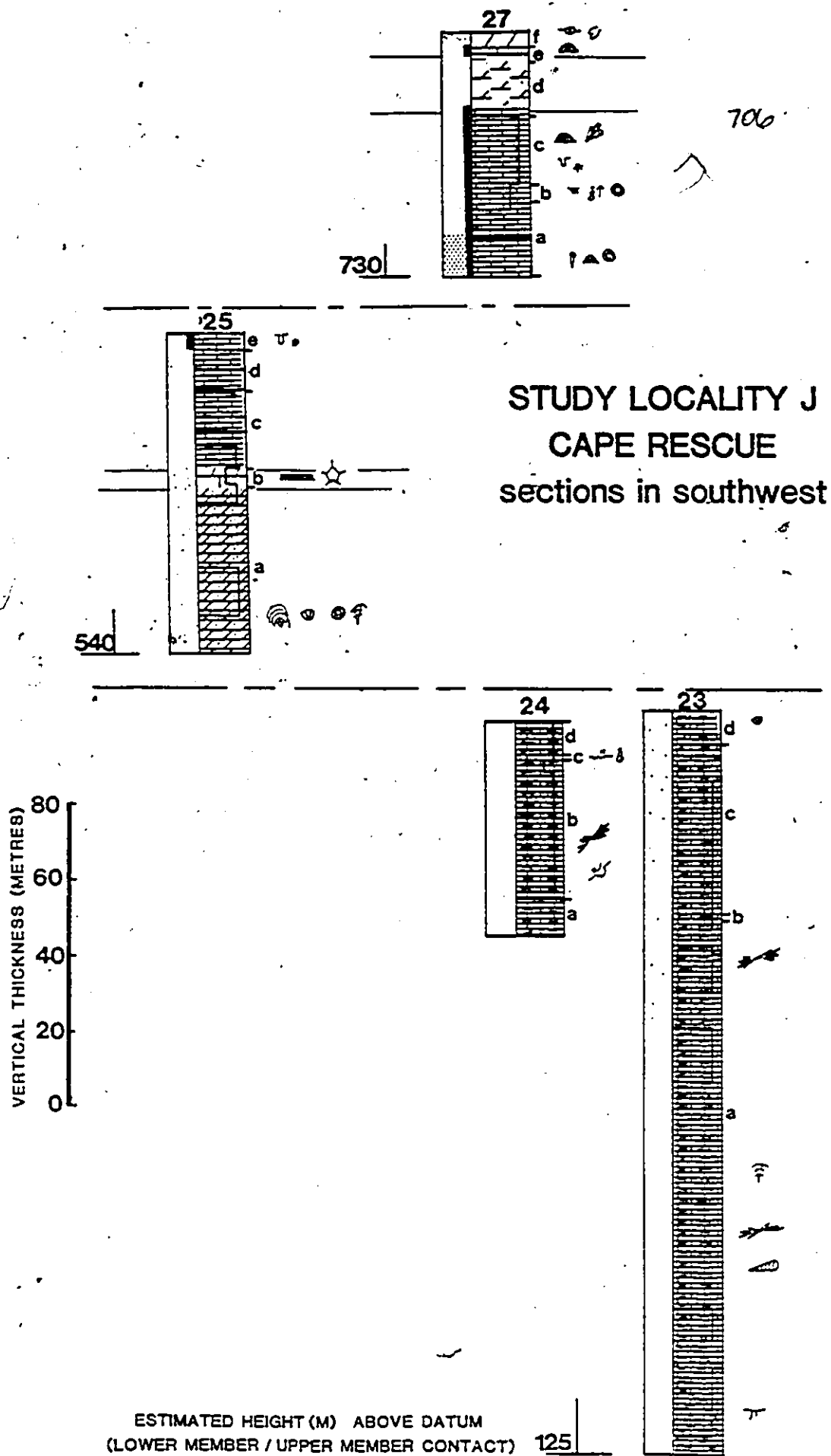


FIG. B-9 Correlation of superfacies in study locality J, southwestern sections.

Section 23 commences with the western-most exposure in the bed of a creek indicated on PL. 5-18a. Section 27 commences approximately 100 m above the top of section 25.

Sections 28 and 29 are coastal exposures, and represent the northern-most outcrop of the Barlow Inlet Formation on the east coast of Cornwallis Island. Outcrop is discontinuous on the stream-washed slope where much of section 28 is located (see 2 of PL. 5-19f), whereas only the uppermost portions of section 29 (see 3 of PL. 5-19f) are well exposed. No faulting is evident in the area, and beds dip gently (15 to 17 degrees) to the northeast.

Section 30 is a coastal exposure (PL. 5-20a). Strata dip moderately (20 to 26 degrees) to the ESE. The sequence is not apparently affected by faults.

Section 31 is in a small stream cut near, and at right angles to, Wellington Channel (PL. 5-18a). Strata dip gently to the east (less than 20 degrees). Exposure is discontinuous.

Section 32 is associated with coastal exposure 1.75 km south of section 31, but significantly higher stratigraphically. Section 32 contains the youngest rocks examined in the Cape Rescue area. Strata dip gently to the ESE. Several small faults cut the exposure (PL. 5-20h), section 32 being on the side of one of these faults.

B.8.2 Succession of facies -- section 23:

The first 135.6 m (interval a) is dominated by planar to wavy bedded lime mudstone of the distal foreslope superfacies (PL. 5-18b). Fossils are very rare and include orthoconic nautiloids, small indeterminate bivalves, indeterminate shell debris (brachiopod and ostracode), encrusting bryozoans, rhynchonellid brachiopods, small high-spired gastropods, trilobite fragments, and smooth spiriferid brachiopods. Ichnogenera include Skolithos, Phycodes, Pilichnia, Taenidium and Pelecypodichnus.⁴³

Above interval a is a lenticular unit, interval b, (3 m thick maximum) consisting of brachiopod shell packstone and grainstone (PL. 5-2c). The basal contact is sharp although not apparently erosional. This unit is terminated laterally by two small normal faults. The lower half of the unit is composed of a shell packstone with broken and predominately smooth brachiopod shells. Pentamerid brachiopod shell fragments, solitary rugosans, rounded clasts of massive bryozoans, orthoconic nautiloids, and crinoidal debris represent other bioclasts. Size range of particles is from 0.5 mm to 10 mm. The cement is inclusion-free, coarse, equigranular and slightly drusy. Shell fragments rarely

⁴³ Molds of small bivalves are commonly spar-filled, with the shell material replaced by Fe-Mn opaque minerals (PL 5-18d). This suggests complete solution and concurrent replacement of the predominantly aragonitic shell material. The calcilutite mudstone also contains problematical calcite-filled spherules (PL. 5-18e), possibly referable to calcispheres, or more likely recrystallized radiolarians.

show an isopachous rim fine-grained blocky spar.

The micrite/spar contacts are very abrupt, suggestive of primary sediment-void interfacies. The common impregnation of fossils by FeMn opaques, and the uncollapsed nature of cavity walls suggest that the sediment was lithified shortly after deposition. Cavities within the sediment possibly represent borings. Common allochems are irregular to subrounded FeMn nodules that may reach 5mm in maximum diameter (PL. 5-12i). Similar nodules have been described by Garrison and Fischer, 1969) in Jurassic deep water carbonate facies of Austria.

The upper surface of the packstone locally has a low amplitude wave-form (PL. 5-2c). The contact between packstone and overlying grainstone is stylolitic. The overlying grainstone is also composed of shell fragments smooth brachiopods, minor crinoid debris, and FeMn nodules. The postulated origin of this unit is given in point form below:

1. Storm-generated currents winnowed brachiopod-rich basin slope sediments. The partially-winnowed sediment may have been transported as a low amplitude dune, thus giving rise to the lenticular nature of the deposit. The undulating upper surface of the packstone may be the result of superimposed smaller scale bed forms, or alternatively, of subsolution or subsequent storm scouring.

2. The high porosity and permeability of the deposit and its raised topography on the sea floor, promoted submarine cementation, FeMn encrustation and possible solution of the upper surface.
3. A second storm deposit, perhaps in part derived from the underlying packstone, was laid down. This second deposit was almost completely winnowed to form grainstone.

Above the interpreted storm deposits, there are 37 m (interval c) of wavy, thin to medium-bedded lime mudstone and faintly-mottled mudstone and wackestone, in general with similar biota to interval a. Towards the top of the section, the benthos gradually increases in abundance and diversity, strata become more distinctly mottled, and argillaceous content decreases. Fossils include crinoid debris, Hemiarges, rhynchonellid, atrypoid and strophomenid brachiopods, and indeterminate bivalves. This interval (d) is 8 m thick. The increased benthos and degree of sediment bioturbation are viewed as reflecting ameliorating conditions (possibly shallowing) within the basin slope environments.

B.8.3 Succession of facies -- section 24:

Between and laterally equivalent to sections 23 and 24 are exposures with several intraformational truncation surfaces (PL. 5-2g). At section 24 (PL. 5-18g) four

intervals are recognized: a basal, weakly mottled lime mudstone (a), a laminated, planar bedded calcilutite mudstone (b), a gastropod packstone and hardground (c), and an upper argillaceous, wavy-bedded lime mudstone (d). Interval thicknesses are 6.2, 30.2, .2, and 10 m, respectively. All four are assigned to the distal foreslope superfacies. Except for interval (c), all are unfossiliferous but contain the trace fossils Phycodes, Palaeophycus and Skolithos.

The upper surface of the gastropod packstone is a hardground, heavily impregnated with Fe-Mn opaques (PL. 5-18c). Allochems are also impregnated: fossils include gastropods and Sphaerocodium (PL. 5-18f). The uppermost surface truncates allochems suggesting that a period of solution accompanied lithification (cf. Garrison and Fischer, 1969, p.27).

B.8.4 Succession of facies -- section 25:

The base of section 25 is estimated to be 540 m above the datum. Interval a (34.6 m) is a weakly-mottled mudstone to fossiliferous wackestone (PL. 5-19a) with an exceptionally diverse benthos. These strata contain the trace fossil Zoophycos and are crinoid-rich. The open marine character and diversity of the benthos, the commonly fragmentary nature of the fossils, local packstone textures, and the presence of possible dasycladacean (Ischadites), all suggest

a stenohaline subtidal environment of moderate to shallow depth. The interval is assigned to the proximal foreslope superfacies. The overlying interval b is 4.1 m of unfossiliferous, quartzose silty calci/dolosiltite with fine laminae and desiccation polygons. The possible significance of this abrupt shallowing and emergence is discussed below in part 4.2 'Platform Design and Evolution'. The interval is assigned to the tidal flat superfacies, although it conceivably represents emergence of an offshore bar, rather than a coastal tidal flat.

Interval c (21.1 m) contains weakly-mottled, slightly rubbly-weathering lime mudstone. Faunal abundance is low, and faunal diversity decreases gradually upsection. Dendroid bryozoans, stromatoporoids, crinoid debris, strophomenids, low-domical favositids, solitary rugosans, orthoconic nautiloids, and alveolitid corals are present. Skolithos is the only trace fossil recorded. The sparse but open marine fauna suggests a return to progressively deeper subtidal conditions. This interval is assigned to the medial foreslope superfacies.

Interval d is 10.3 m thick and consists of thin-bedded lime mudstone with lozenge-shaped nodules (PL. 5-19b) and a sparse fauna of crinoid debris, stems and holdfasts. Interbedded with this lithology are intraformations conglomerates, some with poorly sorted but very well-rounded, calcilutite intraclastics up to 30 mm across, that

are grain-supported and fringed by isopachous rims of prismatic to fibrous spar (PL. 5-19d). Associated with these intraclasts as either bioclasts or an encrusting epifauna are rhizoidal holdfasts, stromatoporoids and sponges. Crinoid debris is ubiquitous. The calcilutite intraclasts appear to have a concentric growth pattern (PL. 5-19d) and are considered to represent storm-exhumed nodules of submarine-cemented calcilutite (FIG. B-10). This explanation serves to reconcile the extreme rounding of the intraclasts (a diagenetic feature), the clastic texture of the deposits, and their interpretation as deeper water facies. This lithofacies appears to have many similarities with the Adneter Scheck facies described by Garrison and Fischer (1969).

Overlying interval d are 26.8 m (interval e) predominately of lithistid sponge-dominated mud mounds (PL. 5-3i) similar to those described at study locality 'F'. Minor facies include polymictic, matrix-supported carbonate breccia, crinoidal packstone and thin-bedded calcilutite mudstone.

DERIVATION OF INTRACLASTS IN SECTION 25

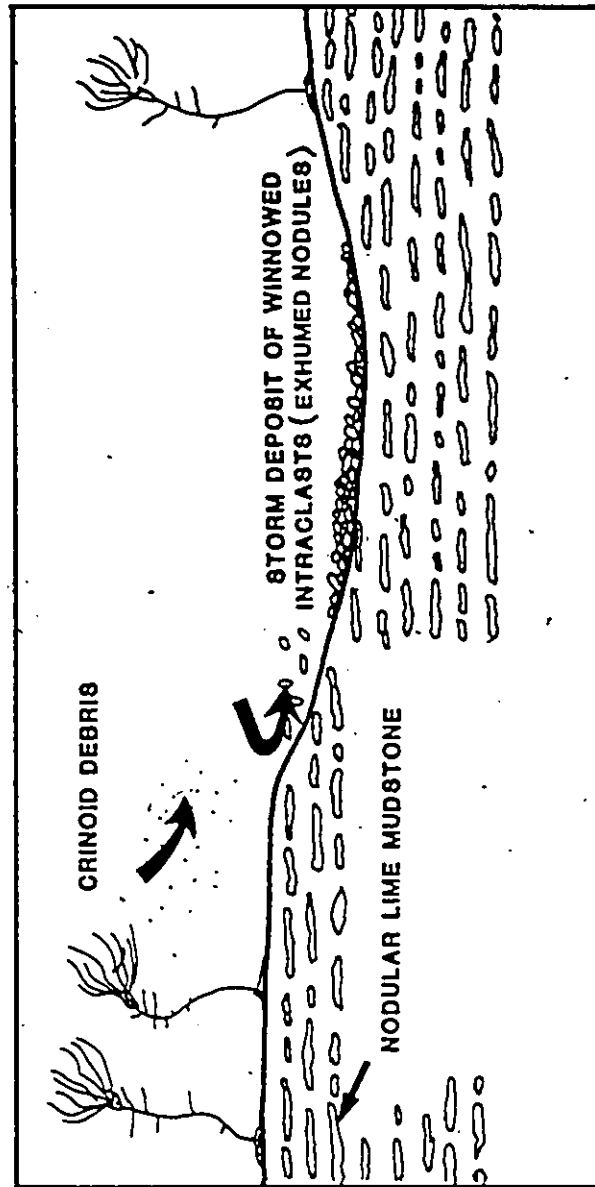


FIG. B-10 Derivation of intraclasts in section 25.

A more thorough investigation of these biocumules is found in Paper 4 (this thesis). Both intervals 'd' and 'e' are considered to belong to the medial foreslope superfacies.

B.8.5 Succession of facies -- section 27:

Section 27 represents mostly various facies of a discontinuous reversible zoned biocumule (Paper 4, this thesis). It cannot be ascertained whether individual facies (?sponge boundstone, crinoidal packstone/grainstone, fasciculate rugosan-stromatoporoid boundstone) possess bathymetric significance or whether the different facies are intrinsically governed by habitat and substrate complexity (indexed to biocumule maturity). Interval a (10.7 m) is composed of dolomitized fasciculate rugosan boundstone, and calcareous coral-stromatoporoid-solenoporid rudstone, inertialstone and framestone. Interval b (4.0 m) comprises graded, cross-bedded crinoidal grainstone, massive crinoid-solenoporid rudstone, and crinoidal wackestone. Interval b clearly represents bioclastic sediments of the reef flank. The crinoidal flank beds are overlain by biocumule facies of interval c, which include muddy sponge(?)-rhizoidal holdfast boundstone, and stromatoporoid fasciculate rugose coral-rhizoidal holdfast framestone. The muddy facies occurs both below and above the framestone facies, in this interval which is 18 m thick. Interval d (16.5 m) is poorly exposed, but at its base there are

tabular stromatoporoids and crinoid debris. A similar fauna occurs within inertiasstone of the basal part of interval e (2 m), along with dendroid stromatoporoids. Interval f (3.1 m) consists of a domical stromatoporoid inertiasstone that grades upward into massive fenestral calcilutite with megalodont bivalves.

Intervals a through c cannot be placed more accurately than within medial to proximal foreslope superfacies. Interval d and the basal part of interval e may belong to the shelf edge superfacies, whereas most of interval e and all of f can be placed confidently in the proximal back barrier and restricted shelf lagoon superfacies, respectively.

Plate 5-19c illustrates the irregular mound shape of the foreslope biocumule facies versus the ledge-forming strata of the shelf-edge and platform interior.

B.8.6 Succession of facies -- section 28:

Interval a (16 m) of section 28 (FIG. B-11) is composed of faintly mottled, sparsely fossiliferous lime mudstone to floatstone with low domical favositids, crinoid debris, whole sponges, bryozoans, and spiriferid brachiopods. The trace fossils Zoophycos, Chondrites and Pelecypodichnus are present. The interval is suggestive of the medial foreslope superfacies. The overlying interval b (15.1 m) consists of crinoidal floatstone, rudstone, grainstone and packstone, all of which are pervasively dolomitized. This particular crinoid blanket facies outcrops discontinuously for at least 2 km along the east coast. Similar dolomitized encrinites occur elsewhere along the coast in the Cape Resuce area, and have been figured and discussed by Thorsteinsson (1958). Bedding is absent from these rocks although cemented bands respond selectively to weathering, giving a semblance of stratification (PL. 5-19e).

Cross-stratification is not recorded, but encrinites are notoriously poor in preserving such structures, and its apparent absence is not diagnostic. Finer grained dolarenites (PL. 5-19i) are evidently grain supported but both floatstone and rudstones have a fine grained dolomitic matrix that presumably follow a lime muddy precursor. The crinoid blanket facies probably represents partially-

STUDY LOCALITY J CAPE RESCUE sections in northeast

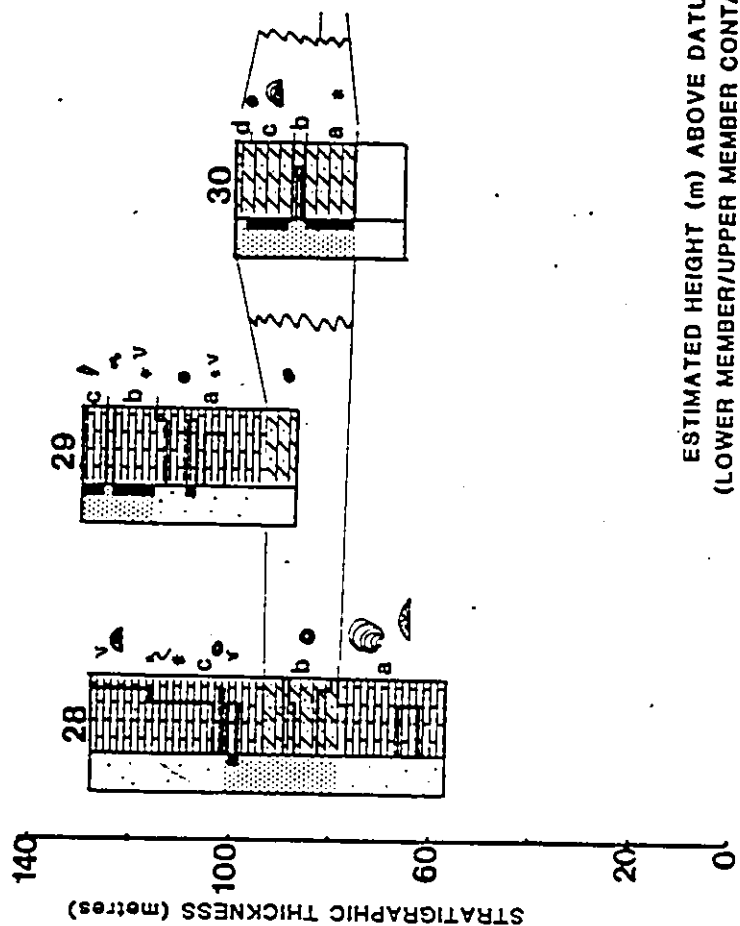


FIG. B-11 Correlation of superfacies in study locality J, northeastern sections.

winnowed bioclastic deposits of transported material, as well as more nearly autochthonous accumulations (see Paper 4, this thesis). Interval b is thought to have been deposited on the proximal foreslope. Interval c (33 m) contains a diverse assemblage of lithofacies belonging to the medial foreslope. These facies include: lithistid sponge-spongiostromate algae boundstone (PL. 5-19j); carbonate breccias (debris flows) featuring scours in underlying lithologies (PL. 5-19g), plastically-deformed rafts of strata, and commonly a crinoidal wackestone matrix; faintly-mottled, thin wavy-bedded wackestone exhibiting small irregularly distributed patches of hematite, synaeresis cracks (PL. 5-19k), a greater degree of elastic than plastic deformation of hard bands, and limited fauna of crinoids, favositid corals including Coenites, solitary rugosans, bryozoans, and some smooth spiriferids; rhizoidal holdfast boundstone (PL. 5-3g) that incorporates bio-catalyzed cementation zones; intraclastic crinoidal packstone; and exotic blocks of boundstone (up to 4 x 4 m) composed of massive favositids, stromatoporoids, sponges, syringoporids and rhizoidal holdfasts (PL. 5-19l). Soft sediment deformation of strata is common and deposits transitional between slides and mass flows are also present (PL. 5-19g). Overall, the carbonate breccia units increase in thickness upsection, as does the maximum clast size of these breccias.

The succession of facies represented in section 28 can be interpreted in two ways (FIG. B-12), as follows:

1. Interval b could represent a toe-of-slope deposit with sands and gravels derived from shallower-waters, by-passing the steep, largely non-accretional muddy foreslope (interval c) (e.g. by-pass margin with gullied slope). The succession would thus reflect continuity of deposition in a shallowing-upward, progradational platform margin model.
2. The favored interpretation involves 2 steps. Initially crinoid sands and gravels accumulated largely in place on the proximal foreslope. Subsequently, the platform drowned, with possible steepening of the foreslope gradient.

The latter interpretation is favored for several reasons, as follows:

1. there is no grading of crinoid bioclasts, nor interbedding of hemipelagic mudstones in interval b.
2. channelized deposits of crinoidal material are absent from the foreslope (although this could be a function of the available exposure).
3. the blanket sands are laterally extensive, stratiform rather than lobate bodies, and do not contain exotic intraclasts or other material derived from upslope.

DEVELOPMENTAL MODELS FOR SECTION 28

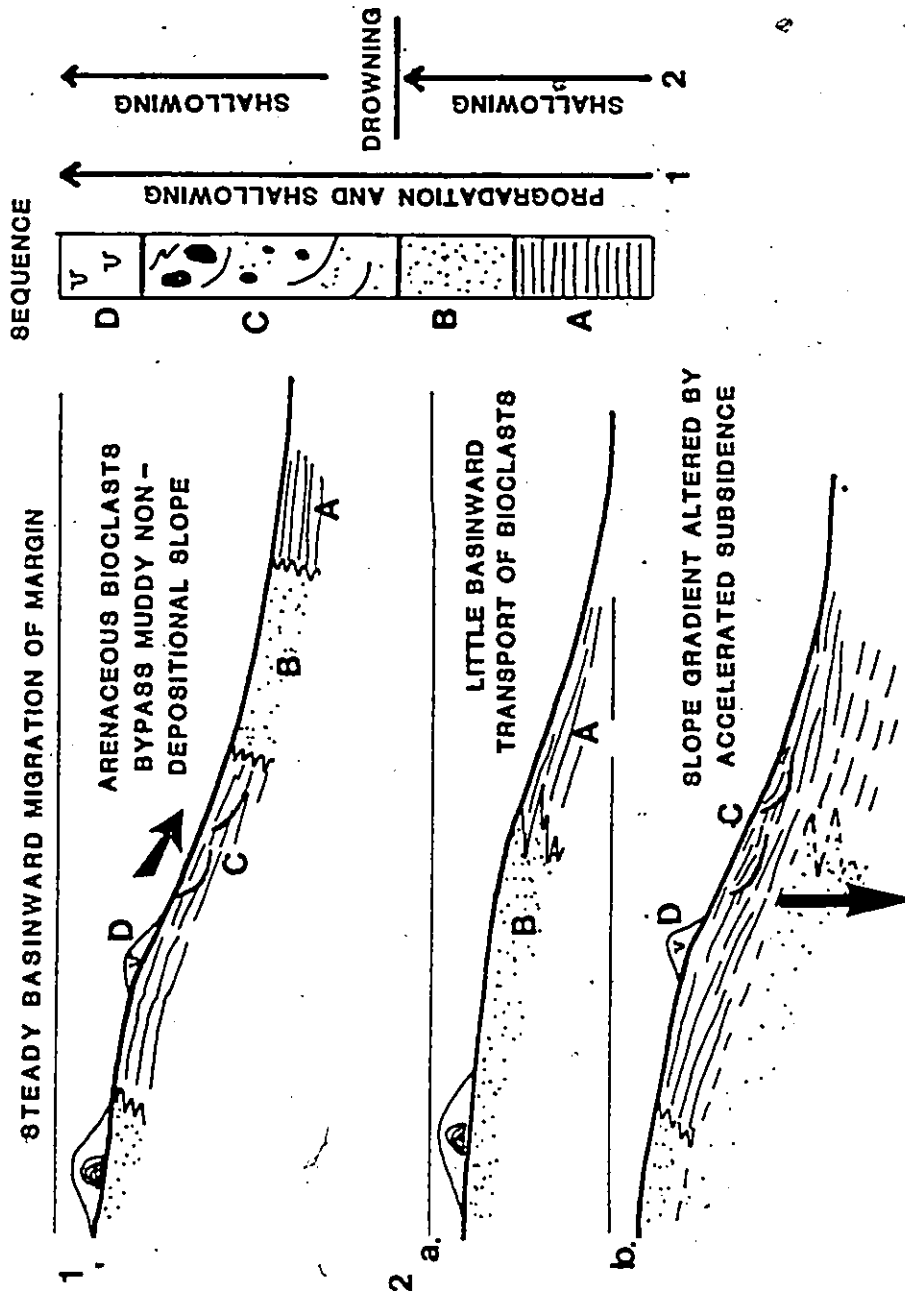


FIG. B-12 Developmental models for section 28.

4. in section 30 the blanket facies is laterally equivalent to a stromatoporoid-crinoid rubble skeletal buildup considered to represent a biocumule of the proximal foreslope.

B.8.7 Succession of facies -- section 29:

Interval a (23.6 m) of section 29 (FIG. B-11) commences above the same crinoid blanket facies as in section 28 (i.e. interval b) and consists of sparse exposure of faintly-mottled dolomitic wackestone, and thin units of boundstone featuring Favosites, solitary rugosans, rhizoidal holdfasts, lithistid sponges, and smooth spiriferid brachiopods. Interval b (11.7 m) comprises a large dolomitized lithistid sponge-syringoporid coral-rhizoidal holdfast mud mound and subjacent crinoidal wackestone (PL. 5-19f,m). Interval c (4.5 m) is represented by mucophyllid coral-rich satellite reefs that apparently 'colonized' the sponge mound (PL. 5-19f: see also Paper 4, this thesis). Intervals a, b and c are all considered to belong to the medial foreslope superfacies.

B.8.8 Succession of facies -- section 30:

The same crinoid blanket facies present in section 28 outcrops along the shoreline (1 of PL. 5-20a). The base of the biocumule that forms the bulk of section 30 (FIG. B-11), lies on strike and updip from the crinoid blanket facies.

Interval a (11.5 m) comprises dolomitized inertiasstone to framestone with crinoid debris, rhizoidal holdfast, hemispherical stromatoporoids and favositids set in a dolomitic mudstone to wackestone matrix. Interval b (9.0 m) is separated from interval a by a thin (.8 m) grey-green weathering lime mudstone to crinoidal wackestone (arrow in PL. 5-20e). The significance of this hiatus in reef growth is uncertain: it may reflect the platform drowning event that is suggested to have affected the facies succession in section 28 (1.3 km to the north). Interval b is composed of both stromatoporoid framestone (PL. 5-20e) and pockets of graded crinoidal rudstone (PL. 5-20e). Lateral to the main buildup are domical stromatoporoid inertiasstone (sparse-dense structure) biostromes (3 in PL. 5-20a). Overlying the buildup are at least 2 m of severely dolomitized, vuggy crinoidal rudstone with minor sulphide (predominantly marcasite) mineralization. The buildup in section 30 is a typical stromatoporoid-crinoid rubble skeletal buildup (described in detail in Paper 4, this thesis), that probably occupied a position on the proximal foreslope.

B.8.9 Succession of facies -- section 31:

Section 31 (FIG. B-11) is composed entirely of thin-bedded mottled dolomitic limestone (predominantly types I and II, rarely labyrinthine). Mottled lithofacies of this

section can be divided equally into those with rare to common fossils of diverse type (including orthid, strophomenid, and atrypoid brachiopods, solenoporid algae, crinoid debris, Ischadites, low domical favositids, fasciculate rugosans, Mucophyllum, bryozoans, and gastropods) and the trace fossil Zoophycos and those with very few fossils or reduced diversity (crinoid debris, low domical favositids, strophomenids, auloporids, orthoconic nautiloids, ostracodes, gastropods, and solitary rugosans) or no fossils at all. The characteristics of these rocks are intermediate in character between medial and proximal foreslope superfacies, suggesting that section 30 lithofacies were deposited in either transitional slope environments, or on a slope subject to vacillating environmental conditions.

B.8.10 Succession of facies -- section 32:

The entire sequence is pervasively dolomitized (FIG. B-11) and rocks of this genre in the immediate area have been previously considered 'reefoid' (Thorsteinsson, 1958). As Plate 5-20h illustrates, the dolostones are clearly stratiform, and true carbonate buildups have not been identified along the Cape Rescue coast south of section 31.

Section 31 does contain, nevertheless, laterally extensive 'biostromal' units (1 in PL. 5-20h), that typically characterize dolomitized massive skeletal (?crinoid)

biocumulés; biocumules, which are considered an inetgral part of the shelf edge superfacies. Interval a (31 m) comprises medium, fine and very finely crystalline dolostone with varying color, resistance to weathering, and degree of porosity. No organic remains can be identified in these rocks, but of considerable importance is remnant lamination and fenestral fabric (PL. 5-20i) seen in the very finely crystalline dolostone. Interval b (6.4 m) was evidently biostromal; as it contains abundant molds of massive stromatoporoids and favositid skeletons (PL. 5-20 g). Interval c (6.8 m) is similar to interval b but large vugs after massive skeletons are considerably fewer.

The lithofacies of section 32 are the result of intense dolomitization of both platform interior and shelf edge facies, and thus entirely analogous to massive skeletal (?crinoid) biocumules of sections 7,8,9 and 11 discussed previously, and in Paper 4 (this thesis).

Appendix C

REGIONAL CORRELATION

C.1 INTRODUCTION

For purposes of physical correlation (or contrast), it is useful to consider the regional sections in four groups, according to their gross paleo-environmental affinities: deeper-water basin and basin-slope, shallow water shelf, infratidal lagoon and tidal flat, and alluvial fan/alluvial plain.

C.2 DEEPER-WATER "BASIN AND BASIN-SLOPE" FACIES

The entire Late Silurian sequence on Bathurst Island (1 of FIG. 5-13) is composed of largely non-calcareous black mudstones and shales of the Cape Phillips Formation. At Twilight Creek,** the upper 65 m of the Cape Phillips Formation is shale interbedded with laminated, calcareous to argillaceous siltstone. The Cape Phillips Formation is overlain conformably by siltstones of the Bathurst Island Formation (the basal contact of which lies within the early Lochkovian). The Cape Phillips Formation is considered to represent largely pelagic sedimentation in an euxinic environment (Kerr, 1974). The Bathurst Island Formation

** section 25 of Kerr (1974).

represents a flysch-like deposit (clinoform subject to a large influx of terrigenous sediment). The non-calcareous shales, and calcareous siltstone of Bathurst Island have no direct counterparts in upper member superfacies. The Cape Phillips Formation on Bathurst Island may have been deposited in environments basinward of the deepest portions of the distal foreslope on Cornwallis Island.

On Baillie Hamilton Island (3 of FIG. 5-13) most of the late Ludlovian-Pridolian time span is represented by black, graptolitic, calcareous, laminated shale with minor siltstone and limestone of the Cape Phillips Formation. Smith (1980) regarded these lithofacies as representing deposition in a deep-marine, partially-restricted stagnant basin below wave base. The Cape Phillips Formation is gradationally overlain by transition beds (mudstone, siltstone, and limestone interbeds) which are in turn conformably succeeded by an "Unnamed carbonate unit". The lower beds of the unnamed unit consist of thin-bedded argillaceous lime mudstone with common argillaceous seams and abundant brachiopods and gastropods. Tabulate corals and echinoderm debris are locally abundant. These beds are early Lochkovian in age. As on Bathurst Island, the paucity of limestone in the Cape Phillips Formation of Baillie Hamilton Island precludes lithostratigraphic correlation with any Barlow Inlet superfacies. The transition beds described by Smith (1980) may be genetically related to the

distal foreslope superfacies, and the basal part of the unnamed unit has some characteristics in common with the medial foreslope superfacies, but these comparisons are extremely tenuous (Smith recorded few of the features noted in Table 5-4).

On Grinnell Peninsula (4 of FIG. 5-13), the Devon Island Formation is the biostratigraphic correlative of the Barlow Inlet Formation. Morrow and Kerr (1977) recognized 2 lithofacies: a) a laminated lithofacies (which makes up most of the Devon Island Formation) consisting of black, silty, thinly laminated, somewhat fissile limestone and dolostone, and b) a medium-bedded lithofacies of medium yellowish-brown and dark brown, thin- to medium-bedded dolomite with rare trough crossbeds. The former lithofacies has been interpreted by Morrow and Kerr (1977) to represent a clinotherm deposited in a reducing environment. This lithofacies bears considerable resemblance to fissile, silty, argillaceous limestone of the lower member of the Barlow Inlet Formation just north of section 22 in the study area. The two stratigraphic intervals are in part coeval, and likely were contiguous. The medium-bedded lithofacies of the Devon Island Formation was considered by Morrow and Kerr (1977) to represent an undathem; it has no known Barlow Inlet equivalent.

The Devon Island Formation is both conformably and unconformably overlain by the Lochkovian Sutherland River

Formation, consisting of fossiliferous and unfossiliferous dolomites of tidal flat aspect. The Sutherland River Formation does not apparently resemble any presently identified facies of the upper member.

On northeast Cornwallis Island (3 of FIG. 5-13) late Ludlovian to Pridolian strata north of the CASBLFC belong to the C member of the Cape Phillips Formation. The lithofacies of the C member have the following characteristics (Thorsteinsson, 1958, p.82):

- a) "alternating calcareous shales, argillaceous limestones, limestones and shales". The uppermost beds are increasingly argillaceous.
- b) strata are thin-bedded to laminated. "Shale partings are common and generally spaced at 2 to 5 inch intervals (5-13 cm) and in consequence the interbedded limestones weather to flagstone".
- c) textures of the limestone beds range from "sublithographic to fine crystalline".
- d) the rocks have a conchoidal fracture.

Referring to the Cape Phillips Formation as a whole (the C member makes up 75% of the formation), Thorsteinsson (1980) noted that:

- e) the rocks are ordinarily medium dark grey in color, and
- f) the rocks contain occasional beds of fossil debris (mass flow deposits) and soft-sediment deformation structures.

The author has examined one section (23) in a region mapped as Cape Phillips Formation (Thorsteinsson, 1973). The lithologies of this section conform well to the description given above. Furthermore, the identical lithologies compose and characterize the distal foreslope superfacies of sections 18,19,20 and 22, all occurring south of the CASBLFC. Evidently there is no reasonable means to differentiate between the C member of the Cape Phillips Formation and the distal foreslope superfacies of the Barlow Inlet Formation. The same environmental conditions are doubtlessly represented by both stratigraphic units. As it is difficult to distinguish on airphotographs between the distal and medial foreslope superfacies, mapping the Cape Phillips Formation south of its present position would be difficult to do with precision. Regardless of mapping contingencies, it should be emphasized that the distal foreslope superfacies is the lithofacies (and hence environmental) equivalent of the Cape Phillips C member. For this reason, on PL. 5-15a, 5-17a, the distal foreslope superfacies is indicated by map unit ?6, implying an identity with the Cape Phillips Formation.

C.3 SHALLOW-WATER SHELF FACIES

On Cornwallis Island, near Marshall Peninsula (2 of FIG. 5-13), a thin, middle Pridolian bioclastic carbonate unit rests unconformably on late Ludlovian Cape Phillips Formation (Jackson et al., 1978). The significance and areal extent of this angular unconformity remains uncertain (see Paper 2, this thesis). It is not known whether the shallow shelf facies represented by this carbonate unit formed part of, or was isolated from, the Barlow Inlet Platform.

The geology of southern and southwestern Devon Island is presently under investigation by officers of the Geological Survey of Canada and their results have not yet been released. Mayr (1978) has examined drill-cuttings, core, and logs of the IOE Panarctic et al. Devon E-45 wildcat (6 of FIG. 5-13) well which penetrates strata of the Read Bay Group, and commences at surface in Barlow Inlet upper member equivalents. Unfortunately no core is available through the Barlow Inlet interval, which totals 452.9 m according to Mayr (1978). The two units described from the Barlow Inlet interval in this well (Table C-1), do not apparently correspond to any divisions of this formation on Cornwallis Island. Mayr (1978) reported several characteristics in his units, that were not observed by this author in exposures on Cornwallis Island, such as: intervals of abundant very fine grained sand, very calcareous shale, and possible anhydrite

pseudomorphs. Absence of such diagnostic facies as the crinoid blanket, sponge biocumule, dolomitized massive skeletal biocumule, cryptalgal laminite, and fenestral mudstone facies, precludes physical correlation of the well intervals with the formation on Cornwallis Island.

Table C-1

Litho-facies characteristics
of strata in the Devon E-45 well equivalent
to the Barlow Inlet Formation, upper member (from Mayr, 1978)

LAND SURFACE

UNIT 2 : PURE AND SILTY LIME MUDSTONE, IN PART DOLOMITIC AND LOCALLY CONTAINING POSSIBLY ANHYDRITE PSEUDOMORPHS AND STROMATOLITIC-ALGAL LIMESTONE, CONSTITUTE THE PREDOMINANT ROCK TYPES. THE LIME MUDSTONE IS INTERBEDDED WITH PELOIDAL, INTRACLASTIC AND SKELETAL WACKESTONE AND PACKSTONE. VERY CALCAREOUS SHALE IS PRESENT IN LOWER PART. INTERBEDDED DOLOMITE IS SLIGHTLY SILTY, CALCAREOUS, AND VERY FINELY CRYSTALLINE	SURFACE TO 274.1m
UNIT 1 : SILTY TO SANDY, IN PART DOLOMITIC LIME MUDSTONE WITH INTERBEDS OF SKELETAL OR LUMP GRAINSTONE, AND LESS COMMONLY CALCAREOUS SHALE. BRACHIOPODS AND OSTRACODES ARE COMMON THROUGHOUT; CORAL AND GASTROPOD FRAGMENTS OCCUR IN TWO INTERVALS.	274.1-452.9m (178.8m)
LOWER MEMBER, BARLOW INLET FORMATION	BELOW 452.9m

On Russell Island (7 of FIG. 5-13) the correlative of the Barlow Inlet Formation is the lower member of the Drake Bay Formation (Mayr, 1978), the base of which is late Ludlovian (Thorsteinsson, 1980). The lowest part of the upper member is possibly early Devonian based on brachiopods (Mayr, 1978). The lower member is 460.2 m thick and consists of "...white to very light brown, uniform, medium crystalline dolomite ...plus rare aphanocrystalline dolomite" (Mayr, 1978, p.25). It is interbedded with light brown lime mudstone to skeletal wackestone, in part dolomitic in the lower part of the member. The lower part (first 153 m) also contains zones of abundant crinoid fragments and at the very base silty limestone. The original depositional environment of the lower member is not known, and the uniform medium crystalline dolomite does not correspond to known Barlow Inlet lithofacies. Possibly it represents the dolomitized equivalent of the massive carbonate lithofacies of the RI series.

C.4 INFRATIDAL LAGOON AND TIDAL FLAT FACIES

Two sections on Griffith Island (8 of FIG. 5-13; study locality B) include strata above the lower member of the Barlow Inlet Formation. Section 1 (PL. 5-21a,b) includes parts of the upper Douro Formation, all of the lower member (6.8 m), and 24 m of relatively continuous exposure above the lower member. These overlying beds are predominantly

thin-bedded calci/dolosiltite mudstone, that weathers light buff-gray, and contains sparse ostracodes, gastropods, Atrypoides erebus, and brachiopod debris. One mottled dolomitic unit yielded crinoid debris, dendroid bryozoans, and atrypoid brachiopods. Chondrites is the only ichnogenus recognized in the succession. Thirty percent of the units contain either cryptalgal lamination, or desiccation polygons, or both. Laminae are distinctively domed (laterally-linked low-domical stromatolites??) in one unit (PL. 5-2lg). The basal units are somewhat argillaceous and intraclastic, and are gradational with the underlying lower member. The lithofacies are not apparently cyclically arranged, but are clearly indicative of sustained peritidal conditions.

Planar stratified to laminated calci/dolosiltite is characteristic of the Somerset Island Formation (lower member), but is also an integral part of the TF series of the Barlow Inlet Formation. In the absence of a larger proportion of fossiliferous and oncoidal mottled dolomitic lithofacies, the sequence's affinities perhaps lie more closely with the Somerset Island Formation, but assignment to either formation is arbitrary.

The second section is further north (PL. 5-2la), and commences about 90 m above the top of the lower member. A predominantly covered interval lies between sections 1 and 2. Section 2 consists of cyclically-arranged strata, with

one complete hemicycle (PL. 5-21e) containing exactly the succession of lithofacies of the idealized TF hemicycle. Apart from oncoids, the mottled dolomitic lithofacies contains brachiopod shell debris, Atrypoides, solenoporids, crinoid debris, lamellar, hemispherical and irregular stromatoporoids, solitary and cerioid rugosans, Coenites, cylindrical and massive favositids, and small high-spired gastropods. Trace fossils include Chondrites, Beaconites, and Planolites. The strata of section 2 can confidently be placed in the Barlow Inlet Formation on the basis of stratigraphic position and close affinities to the TF hemicycles of the Platform Interior Magnafacies.

Correlative beds of the Barlow Inlet Formation on northern and northwestern Somerset Island include the Somerset Island and Peel Sound Formations. The latter formation and its attendant controversial age assignments are discussed below (5.4.3.4). Although the Peel Sound Formation on Somerset Island differs somewhat in its formal stratigraphic divisions and spectrum of facies from the same formation on eastern Prince of Wales Island (see Miall and Gibling, 1978), it remains principally the product of alluvial fan, braided river and possibly eolian depositional processes. Much of the Barlow Inlet Formation was deposited synchronously with the Peel Sound on Somerset Island, but these two formations bear no similarity to one another.

The Somerset Island Formation (Miall, Kerr, and Gibling, 1978) is latest Silurian (late Ludlow to ?Pridoli) in age, and is possibly entirely restricted to the latialata conodont zone and hence of latest Ludlow age (Thorsteinsson, 1980, p.26). The formation represents deposition on intertidal and supratidal flats, influenced by both minor marine transgression and alluvial sheet floods. The formation is divided into lower and upper members. The upper member is characterized by interbedded grey laminated dolomite and limestone, red quartzose siltstone and red dolosiltite, and rarely contains horizontally-laminated sandstone, and evaporite casts and nodules. The presence of sand, evaporites and red-colored to variegated lithofacies clearly distinguishes the upper member from any lithofacies of the Barlow Inlet Formation. The upper member evidently represents deposition on those portions of the Barlow Inlet Platform tidal flats least transgressed by the sea, and closest to the distal fringe of the B.A.L. alluvial plain.

The lower member of the Somerset Island Formation is characterized by mottled dolomitic, in part bioclastic, carbonates, and laminated dolosiltite and calcisiltite. The laminated lithofacies are cream to pale grey in color and contain desiccation cracks, ostracodes, and gastropods (Miall and Gibling, 1978; Miall, Kerr and Gibling, 1978). The lithofacies of the lower member are cyclically-arranged (see FIG. C-1) and thus, overall, bear considerable resemblance to the TF series of the Barlow Inlet Formation.

SOUTHERN SOMERSET ISLAND
MIAL AND GIBLING (1978)

CUNNINGHAM INLET
THIS PAPER

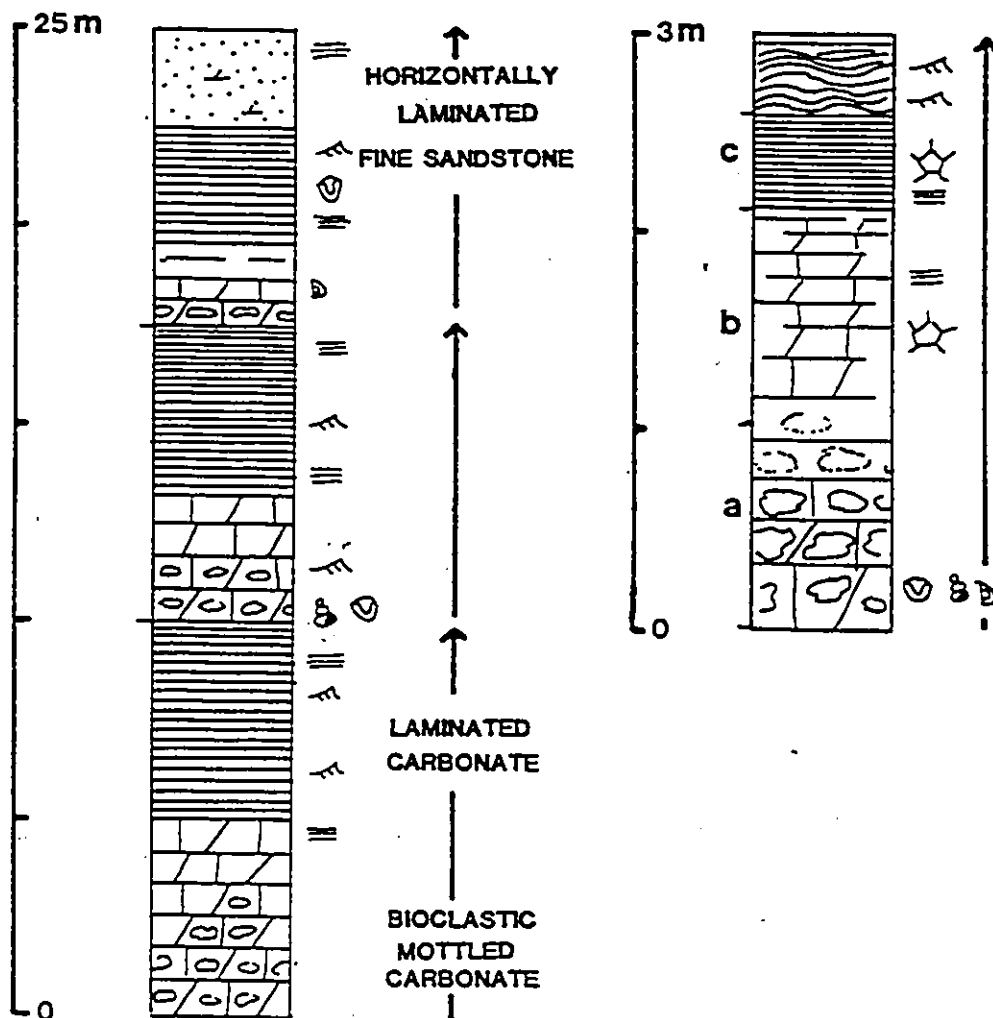


FIG. C-1 Hemicycles of Somerset Island Formation, lower member (lithofacies described in text).

The author investigated the lower member at Cunningham Inlet at a locality (study locality A) corresponding to section 13 of Miall, Kerr and Gibling (1978) and Miall and Gibling (1978). Miall and Gibling (1978) implied that section 13 is in conformable contact with the underlying Douro Formation. It is in fact a faulted contact⁴³ (PL. 5-21c), and the precise stratigraphic position of this section is therefore not known. Nonetheless on lithological and faunal grounds,⁴⁴ this section most probably does contain lower member strata. The lithofacies are cyclically-arranged and the modal hemicycle (FIG. C-1) is described as follows:

- a) an abrupt (rarely scoured) base is overlain by mottled dolomitic limestone type II, with large tightly-packed micrite nodules giving the lithofacies an overall massive medium-dark grey to purple-grey rather than a mottled appearance. Nodular bands are 3 to 5 cm thick, but fractures give this lithofacies a medium-bedded appearance (PL. 5-21d). Whole fossils were not noted. Brachiopod and ostracode debris predominate. Interbeds of crinoid-solenopodid grainstone are rare. This lithofacies is gradationally overlain by

⁴³ the section measured by the author commences at the lowest exposures available north of the northerly bounding fault of the Cunningham Inlet Graben.

⁴⁴ the section contains Atrypa foxi, which does not occur in the Douro Formation, and yet is no younger than Late Silurian.

b) thinly-bedded to weakly laminated silty calci/dolosiltite that weathers medium grey-brown. Ostracode fragments are common and the ichnogenera Polarichnus, Chondrites, Skolithos, and Palaeophycus are present. Desiccation polygons are common (PL. 5-2lf). This lithofacies is overlain gradationally by c) unfossiliferous, well-laminated light-medium grey-brown calci/dolosiltite.

In 2 hemicycles (of 7 observed) wavy to ripple cross-laminated calcisiltite gradationally overlies the laminite (PL. 5-2lh).

The lower member bears considerable resemblance to the TF series of the Barlow Inlet Platform Interior Magnafacies on southern Cornwallis Island. Furthermore, both stratigraphic intervals probably span the same age.⁴⁷ There are differences in lithofacies, notably that: a) oncoids were not observed in the Cunningham Inlet lithofacies nor were they reported by Miall and Gibling (1978), and b) wavy ripple lamination has not been observed above (?cryptalgal) laminite in TF hemicycles. Nonetheless, overall similarities outweigh differences, and it is reasonable to conclude that the TF series at locs. C and D represent a tongue of the

⁴⁷ both intervals commence above the 'event horizon' represented by top of the Douro Formation (see paper 2, this thesis) marking the termination of the siluricus zone, and both units contain the heterostracan Torpedaspis elongata, strengthening both environmental and time correlation between these two stratigraphic intervals.

Somerset Island Formation that extended to southern Cornwallis Island. Whether for mapping purposes this distinction is worth making remains unresolved, and for this reason PLS. 5-9a, 5-11a show the presence of map unit ?3 (i.e. possibly referable to the lower member of the Somerset Island Formation).

C.5 ALLUVIAL FAN/ALLUVIAL PLAIN FACIES

Strata equivalent in age to the Barlow Inlet Formation on eastern Prince of Wales Island belong to the predominantly continental deposits of the Peel Sound Formation. On Prince of Wales Island this formation is divided into two members (lower and upper), but their ages are the subject of some debate (see Thorsteinsson, 1980 vs. Miall, 1983). The lowermost beds of the lower member are probably of latest Ludlow age and are thus correlatives of the basal part of the Barlow Inlet Formation. Controversy surrounds the possible diachroneity of the contact between the members. The author tentatively follows Thorsteinsson (1980), and considers the upper member to be entirely Lochkovian or younger in age (FIG. 5-13). The facies represented by the lower member include: a) alluvial (or coastal) fan deposits, b) minor fluvial redbed deposits (suggesting a relatively narrow alluvial plain), c) marginal marine deposits consisting of siltstones, grey quartz sandstones and sandy dolomites, and d) open marine facies (nodular bioclastic limestones).

The influx of continentally-derived clastic sediment into the sea on the western margin of the B.A.L. would likely have inhibited the development of an organic (reef) shelf rim. Open marine conditions probably existed on a shallow shelf contiguous with the margin of the B.A.L., a situation in marked contrast to the rimmed shelf complex with restricted Platform Interior Magnafacies extant further north. Hence the open marine facies of the Peel Sound Formation do not have an exact environmental analog in the Barlow Inlet Formation; the faunal makeup of the proximal foreslope superfacies may be somewhat similar.

C.6 COMPARISON TO OTHER AREAS WITHIN THE ARCTIC ARCHIPELAGO

Areas beyond the islands immediately surrounding Cornwallis also contain carbonate strata formally or informally assigned to the Read Bay Group, and in part equivalent to the Barlow Inlet Formation. Very little work, and no comprehensive facies analysis, has yet been done on the Read Bay Group (or equivalents) of Banks, Victoria and Melville Islands (Mayr et al., 1980; Tozer and Thorsteinsson, 1964; Thorsteinsson and Tozer, 1962; Miall, 1976).

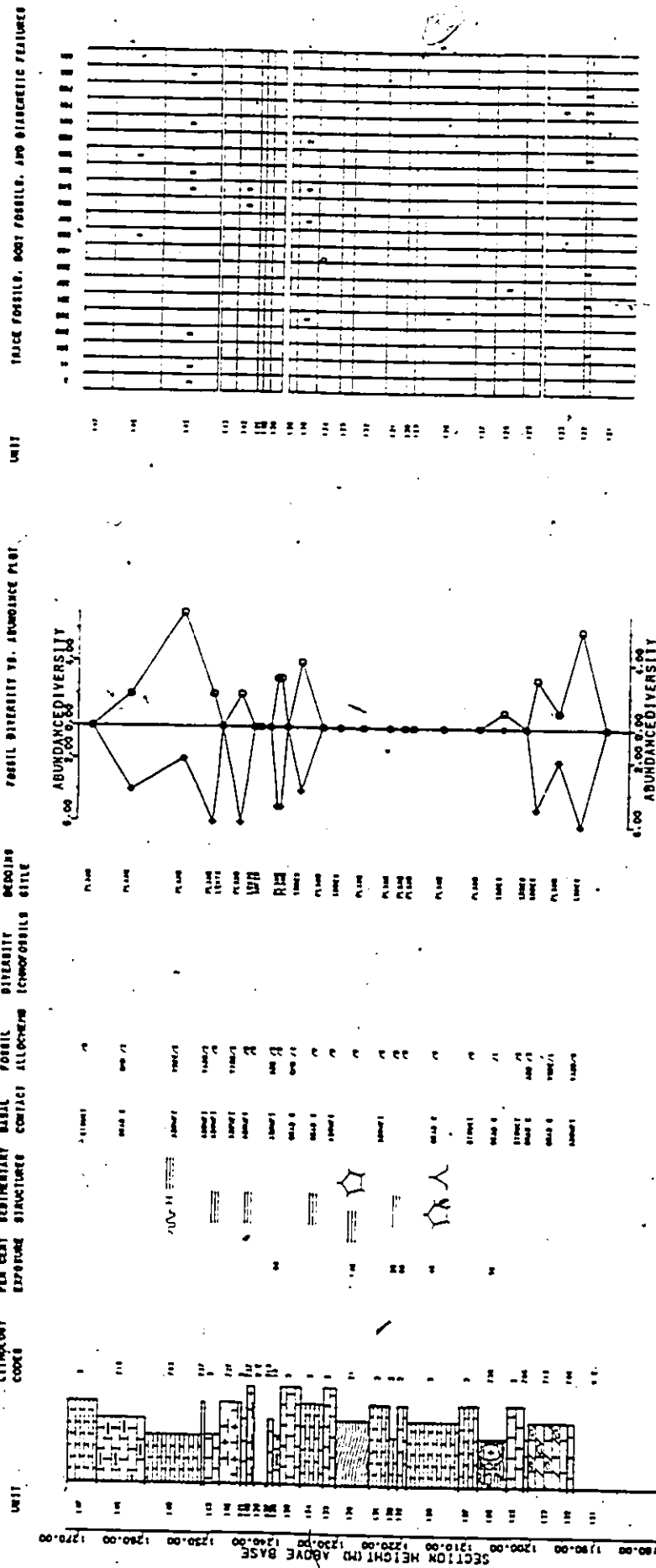
On Ellesmere Island, more substantial published reports exist for the Read Bay Group (Mayr, 1973; McGill, 1974; Kerr, 1976), and basinal equivalents (Trettin, 1979; Poey, 1982). The Silurian platform-basin transition on southern

Ellesmere Island, is currently being investigated in detail by J.-L. Poey of the University of Ottawa in collaboration with U. Mayr of the Geological Survey of Canada. Available published data are not amenable to even gross (subformational) lithostratigraphic correlation with Cornwallis Island. However, several superfacies described herein do have approximate counterparts in Siluro-Devonian facies of Ellesmere Island. These are listed in Table C-2.

Comparison of Silurian facies assemblages
of Cornwallis and Ellesmere Islands

Facies and/or facies assemblages (this study)	Facies and/or facies assemblages from Silurian strata of Ellesmere Island	Reference
distal foreslope superfacies	interlaminated lime mudstone and shale of Hazen Formation lithologies 2 and 4 of "Basinal Environment"	Trettin(1979) Mayr(1973)
medial foreslope superfacies	"Basin Slope Environment" marine-derived carbonates of Cape Phillips Formation	Mayr(1973) Trettin(1979)
platform interior magnafacies	"Unwinnowed Shelf Environment"	Mayr(1973)
1.massive calcilutite	1.?lime mudstone	
2.cryptalgal laminite	2.laminated peloidal grainstone	
3.fenestral mudstone	3.fenestral dolomite	
4.oncoidal carbonate	4.?oncolitic dolomitic packstone	

SECTION JP5B, UPPER PART OF TYPE SECTION FOR BARLOW INLET FM., READ BAY.



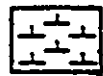
LEGEND

PRINCIPAL LITHOLOGIES CARBONATES



LIME MUDSTONE

-THIN-BEDDED AND/OR WITH
SHALE INTERBEDS
-MEDIUM-BEDDED



LIME MUDSTONE

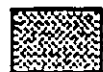
-MASSIVE, WITH OR
WITHOUT FENESTRAE



WACKESTONE



PACKSTONE



GRAINSTONE



BOUNDSTONE

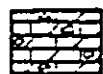
(UNDIFFERENTIATED, BINDSTONE, BAFFLESTONE)



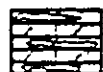
BOUNDSTONE

(FRAMESTONE)

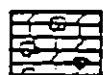
RUBBLY CARBONATES (8-13)



MOTTLED DOLOMITIC LIMESTONE TYPE I



MOTTLED DOLOMITIC LIMESTONE TYPE II



MOTTLED DOLOMITIC LIMESTONE TYPE III



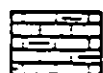
RUBBLY ARGILLACEOUS LIMESTONE TYPES I & II



RUBBLY, RECRYSTALLIZED AND/OR SHALE-
POOR CARBONATE



RUBBLY, RECRYSTALLIZED AND/OR SHALE-
POOR, FOSSILIFEROUS CARBONATE



MARLSTONE

(ARGILLACEOUS LIMESTONE)



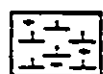
LAMINITE

(CRYPTALGAL)

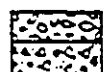
RECRYSTALLIZED CARBONATES (16-17)



VERY FINE TO FINE CRYSTALLINE



MEDIUM TO COARSE CRYSTALLINE



INTRAFORMATIONAL

-CONGLOMERATE
-BRECCIA

-IF THE ROCK IS PREDOMINANTLY
DOLOMITIC, THEN THE ABOVE
PATTERNS ARE MODIFIED BY
SLANTED VERTICAL LINES..

SEDIMENTARY STRUCTURES PRIMARY STRUCTURES



CROSS-BEDDING



PLANAR CROSS-BEDDING



TROUGH CROSS-BEDDING



RIPPLE CROSS-BEDDING



ASYMMETRICAL RIPPLES



SYMMETRICAL RIPPLES



RIPPLES



LAMINATION



GRADED BEDDING



GUTTER CASTS



CHANNEL

PENECONTEMPORANEOUS STRUCTURES



DESICCATION POLYGONS



SYNERESIS CRACKS



BLISTER MARKS



FENESTRAE
(BIRDS-EYE STRUCTURE)



BIOTURBATION



DISH STRUCTURES



PRIMARY DEFORMATION
(GENERAL CATEGORY OF STRUCTURES)



INTERSTRATAL SLUMP FOLDS



INTRAFORMATIONAL SUBMARINE
TRUNCATION SURFACE



RIP-UP CLASTS



HARDGROUND

DIAGENETICALLY-ALTERED STRUCTURES AND TEXTURES



POROSITY

1 of



BOUNDSTONE
(UNDIFFERENTIATED, BINDSTONE, BAFFLESTONE)



BOUNDSTONE
(FRAMESTONE)

RUBBLY CARBONATES (8-13)



MOTTLED DOLOMITIC LIMESTONE TYPE I



MOTTLED DOLOMITIC LIMESTONE TYPE II



MOTTLED DOLOMITIC LIMESTONE TYPE III



RUBBLY ARGILLACEOUS LIMESTONE TYPES I & II



RUBBLY, RECRYSTALLIZED AND/OR SHALE-
POOR CARBONATE



RUBBLY, RECRYSTALLIZED AND/OR SHALE-
POOR, FOSSILIFEROUS CARBONATE



MARLSTONE
(ARGILLACEOUS LIMESTONE)



LAMINITE
(CRYPTALGAL)

RECRYSTALLIZED CARBONATES (16-17)



VERY FINE TO FINE CRYSTALLINE



MEDIUM TO COARSE CRYSTALLINE



INTRAFORMATIONAL - CONGLOMERATE
- BRECCIA

- IF THE ROCK IS PREDOMINANTLY
DOLOMITIC, THEN THE ABOVE
PATTERNS ARE MODIFIED BY
SLANTED VERTICAL LINES..

NON-CARBONATES



SHALE



SANDSTONE

CODES FOR RELATIVE ABUNDANCE OF BODY FOSSILS
(VOLUME PERCENTAGE OF TOTAL FOSSIL ALLOCHEPS)

- 2 - LESS THAN 1 PERCENT
- 4 - LESS THAN 4 PERCENT
- 1 - 5-19 PERCENT
- 3 - 20-59 PERCENT
- 6 - GREATER THAN 60 PERCENT
- - OCCURRENCE IN OTHER THAN PRINCIPAL LITHOLOGY

TRACE FOSSILS AND DIAGENETIC FEATURES SHOW VALUES
AS PERCENTAGE TOTAL ROCK



SYMMETRICAL RIPPLES



RIPPLES



LAMINATION



GRADED BEDDING



GUTTER CASTS



CHANNEL

PENECONTEMPORANEOUS STRUCTURES



DESICCATION POLYGONS



SYNAERESIS CRACKS



BLISTER MARKS



FENESTRAE
(BIRDS-EYE STRUCTURE)



BIOTURBATION



DISH STRUCTURES



PRIMARY DEFORMATION
(GENERAL CATEGORY OF STRUCTURES)



INTERSTRATAL SLUMP FOLDS



INTRAFORMATIONAL SUBMARINE
TRUNCATION SURFACE



RIP-UP CLASTS



HARDGROUND

DIAGENETICALLY-ALTERED STRUCTURES AND TEXTURES



POROSITY



VUGOSITY (PORES > 2mm)



POROSITY AND VUGOSITY



VUGOSITY ALIGNMENT



BANDED POROSITY



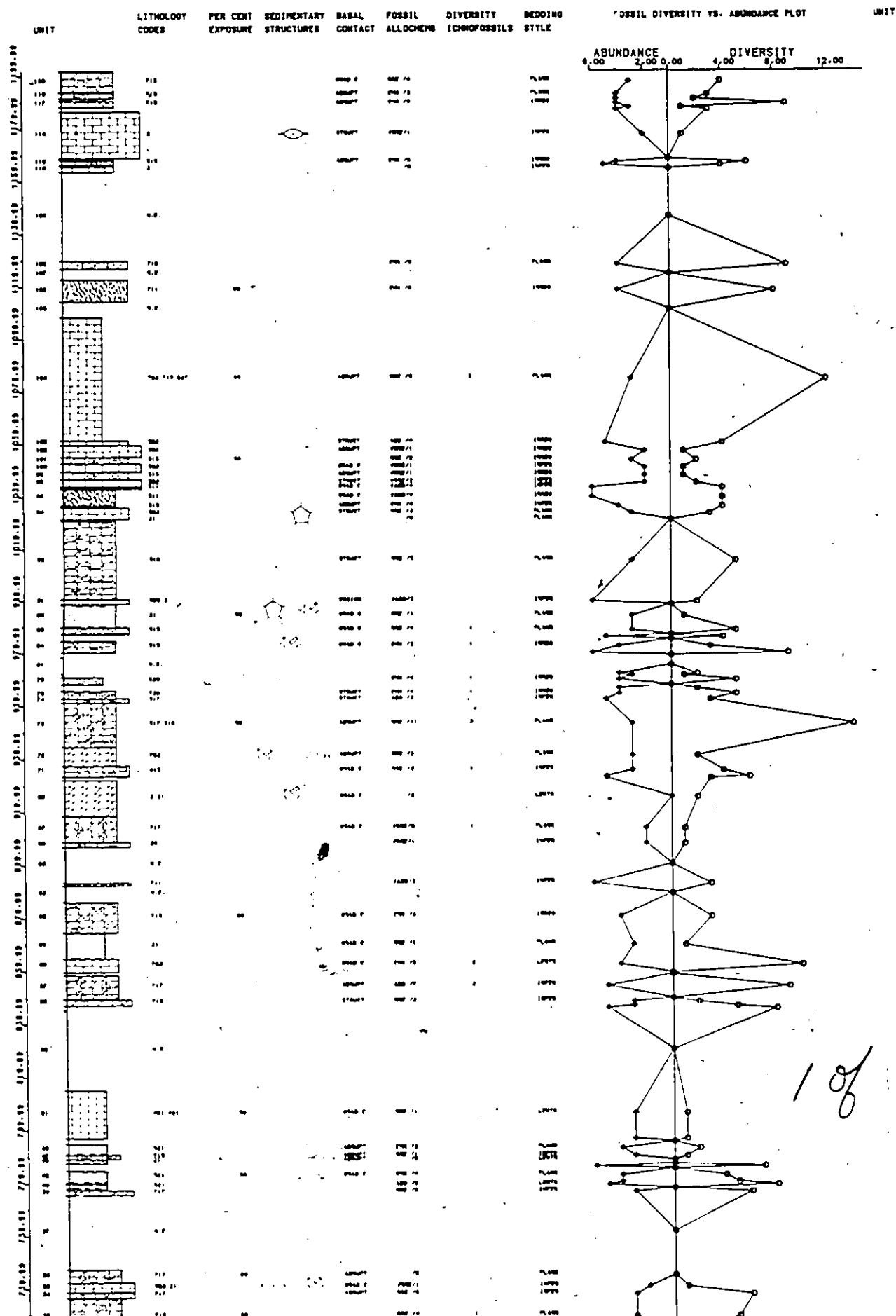
BANDED POROSITY AND VUGOSITY

2 of 2

FIG. B-8b Legend for figure B-8a

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SECTION JP5A. LOWER PORTION OF THE TYPE SECTION OF BARLOW INLET FM., READ BAY

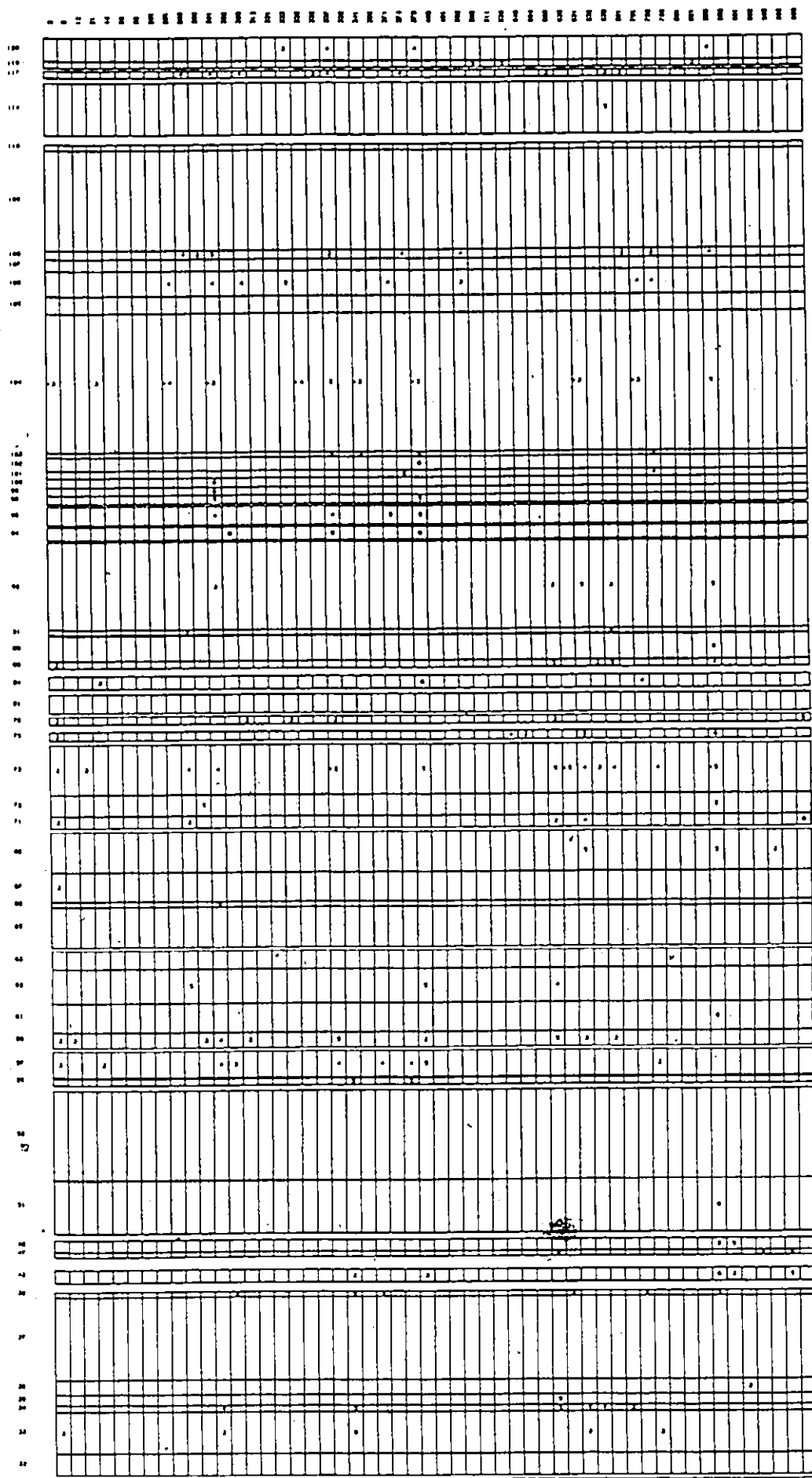
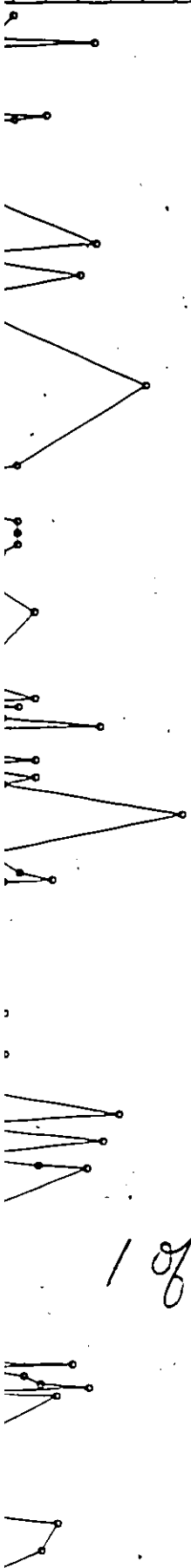


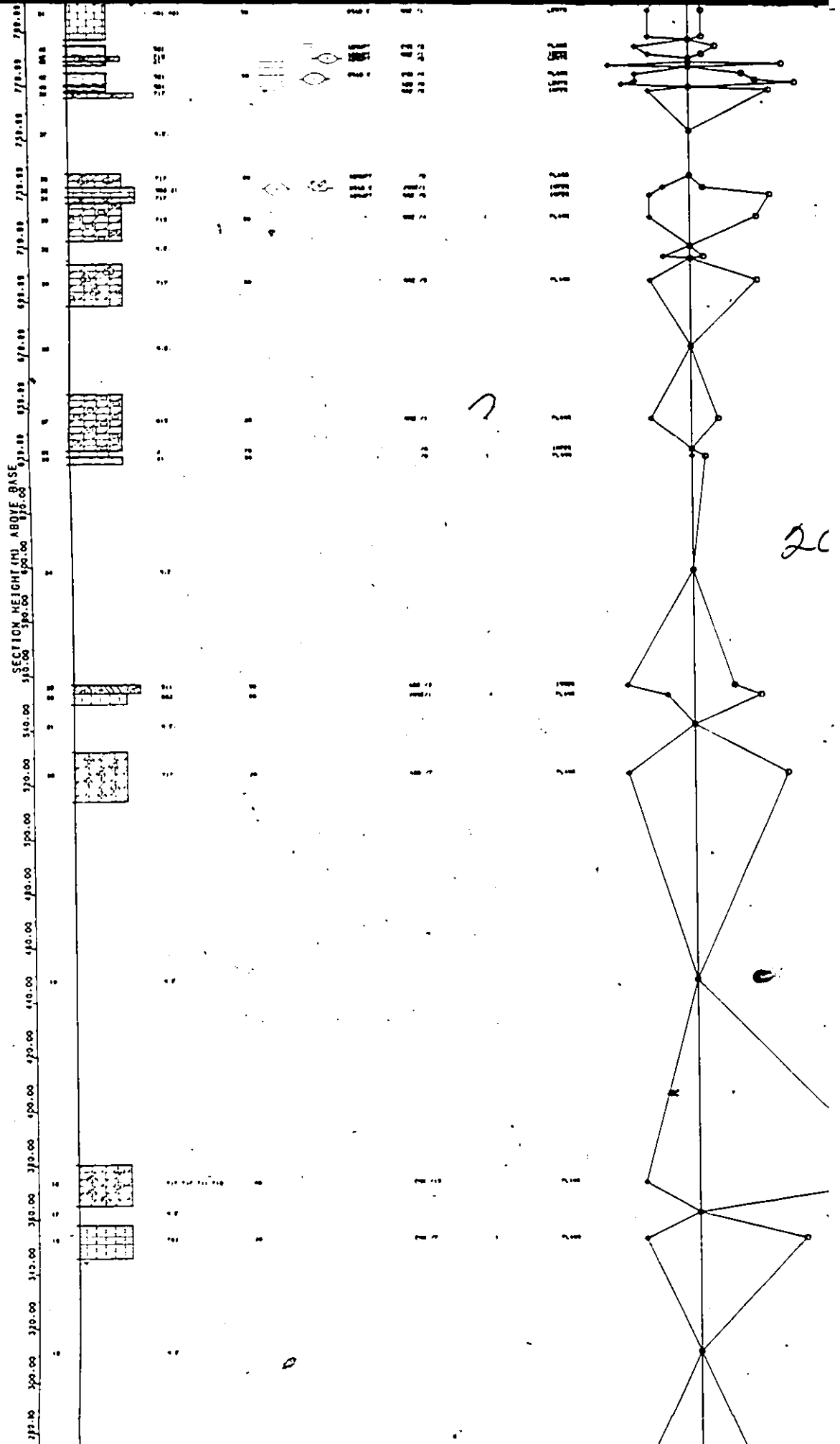
5. ABUNDANCE PLOT

UNIT

TRACE FOSSILS, BODY FOSSILS, AND DIAGENETIC FEATURES

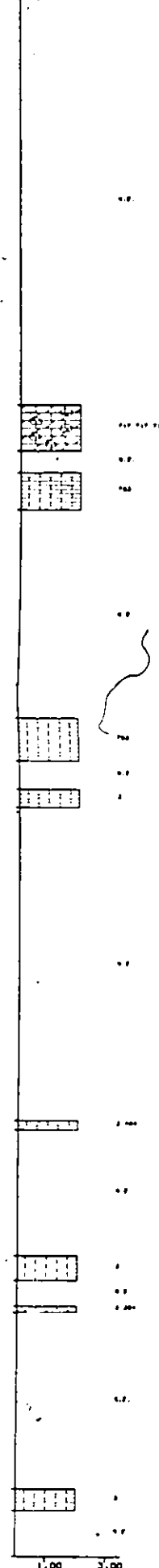
DIVERSITY
4.00 8.00 12.00



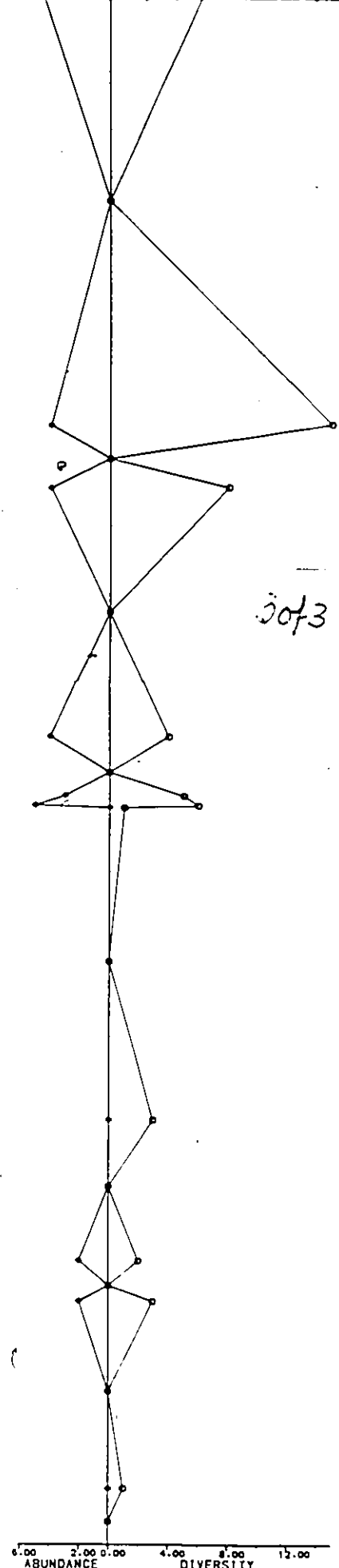


[illegible]

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WEATHERING PROFILE



ABUNDANCE DIVERSITY

3073

0.00 12.00
DIVERSITY