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Stratigraphy and depositional environments of an Upper Ordovician to Lower Devonian shelf-to-basin transition, Svendsen Peninsula, Ellesmere Island, N.W.T..

by

Jean-Luc Poey

A thesis submitted to the School of Graduate Studies in partial fulfillment of the requirements for the degree of Master in Science in geology.

University of Ottawa
Ottawa, Canada, 1988

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ISBN 0-315-46838-6

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Stratigraphy and depositional environments of an Upper Ordovician to Lower Devonian shelf-to-basin transition, Svendsen Peninsula, Ellesmere Island, N.W.T..

Abstract

The Upper Ordovician to Lower Devonian sequence on Svendsen Peninsula shows marked lateral and vertical facies changes. Vertical facies changes represent the establishment during the Early Silurian of a well differentiated shelf-to-basin transition, due to fault-created collapses of an extensive portion of the marginal shelf. For the remainder of the Silurian, interfingering facies record a series of shallowing-upward sequences interpreted as reflecting periodic outbuilding of the carbonate shelf alternating with minor collapses along the shelf-edge. A major latest Silurian collapse of the shelf-edge resulted in an extensive southeastward migration of basinal conditions.

Lateral facies changes reflect progressive deepening from a carbonate shelf northwestward into the Hazen Trough, the axial portion of the Franklinian Basin. Before the Early Silurian collapse, the Irene Bay Formation and overlying lower portion of the Allen Bay Formation accumulated on a ramp-like shelf which gradually deepened northwestward up to an abrupt gradient marking the margin of the laterally restricted Hazen Trough. The Irene Bay Formation and the lower portion of the Allen Bay Formation consist mostly of argillaceous and dolomitic mottled limestones, that change gradually upsection to massive, siliceous and bituminous dolostones, and reflect upward shallowing conditions. Slope conditions predominated within the study area for the remainder of the Silurian, as evident from the appearance of mass flow deposits. Interfingering proximal slope and shelf-edge deposits have been designated the Allen Bay Formation - Read Bay Group undivided. Dominated by calciturbidites, they contain abundant debris flows, as well as spectacular, though rare, olistostromes incorporating undolomitized reef blocks. Distal slope deposits of the Cape Phillips Formation consist mostly of graptolitic hemipelagic sediments alternating with fine calciturbidites which are distal equivalents of the coarser mass flows. The major collapse of the latest Silurian resulted in the spread of basinal "flysch" silty carbonates of the Devon Island Formation throughout the study area.

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Chapter 1
Introduction

1.1 Location and general setting

Within recent years, petroleum and mineral exploration companies have shown great interest in the Canadian Arctic Archipelago, since it contains a thick sequence of rocks representing two major ancient basins. During the Lower Paleozoic, the Franklinian Basin (Fig. 1.1) was situated between the Arctic Platform (part of the craton) to the southeast, and an inferred metamorphic and igneous terrane to the northwest, the Pearya Mountains (Trettin, 1973). The well known Upper Paleozoic-Mesozoic Sverdrup Basin is a successor basin to the Franklinian Basin (Embry & Klovan, 1976).

The Franklinian Basin comprised three different elements; a deeper axial portion termed the Hazen Trough trending generally northeast-southwest, a narrow mixed clastic-carbonate shelf to the northwest, and finally a much wider carbonate shelf to the southeast which partly encroached on the stable platform but was mostly a rapidly subsiding shelf (Trettin & Balkwill, 1979) (see Fig. 1.2; Fig. 1.3 shows the same three elements but for the study area alone). The Central Stable Region or Arctic Platform lay shoreward of the shelf and was an area of much slower subsidence and restricted sedimentation.

The study area is on the Svendsen Peninsula just north of Baumann Fiord, on west-central Ellesmere Island (see the location maps, Figs. 1.4 & 1.5). It exposes Upper Ordovician to Lower Devonian carbonate strata of the Franklinian Basin, that represent a range of depositional environments from shallow shelf to deep trough. It is an area with exceptional exposure and is accessible for at least the two summer months. The carbonate strata form commonly cliffs or very continuous river cuts; more argillaceous interbeds or facies equivalents are more easily weathered and less well exposed.

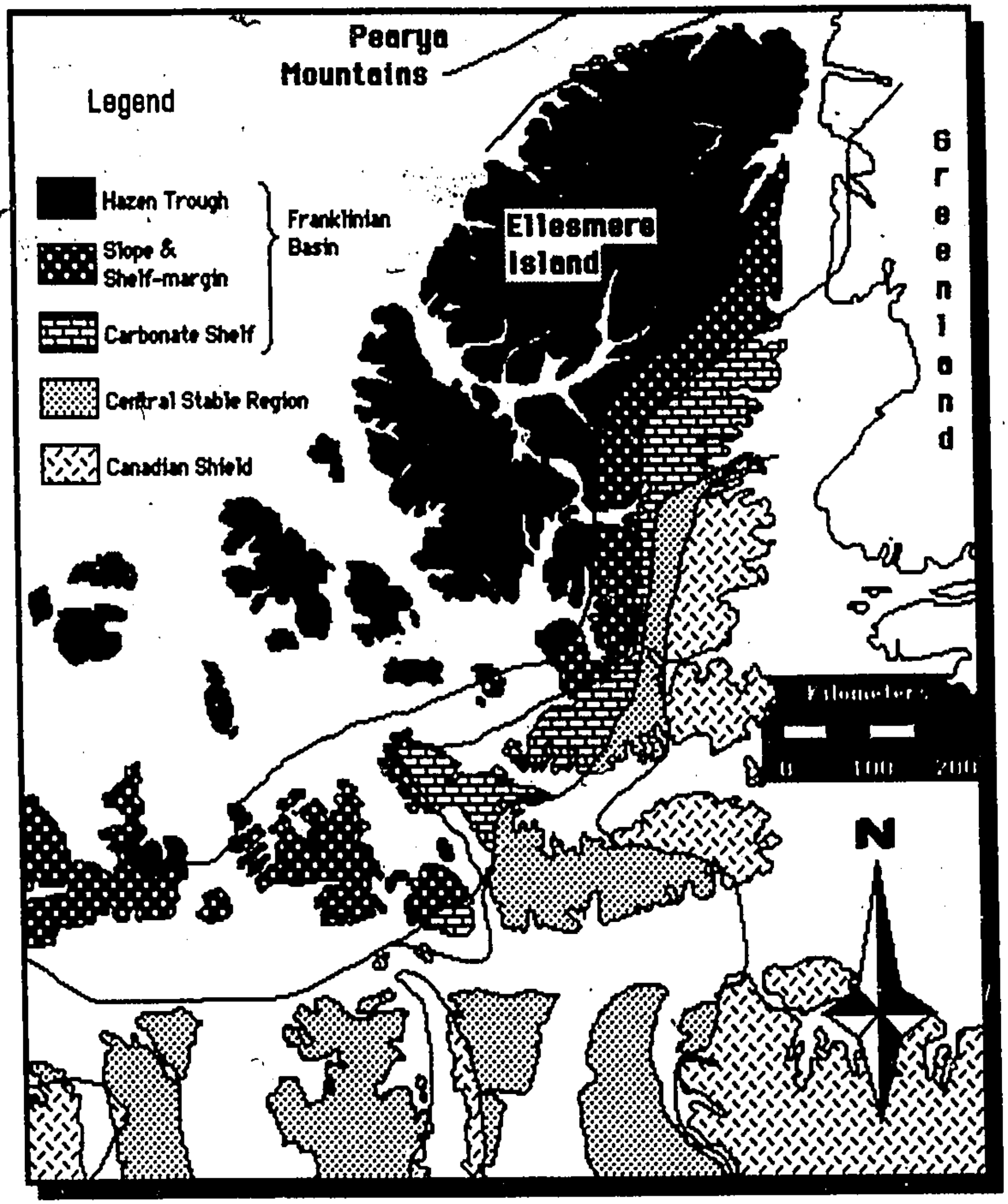


Figure 1.1 Tectono-stratigraphic provinces, Lower Paleozoic and older, of the Canadian Arctic Archipelago. (adapted from Kerr, 1975)

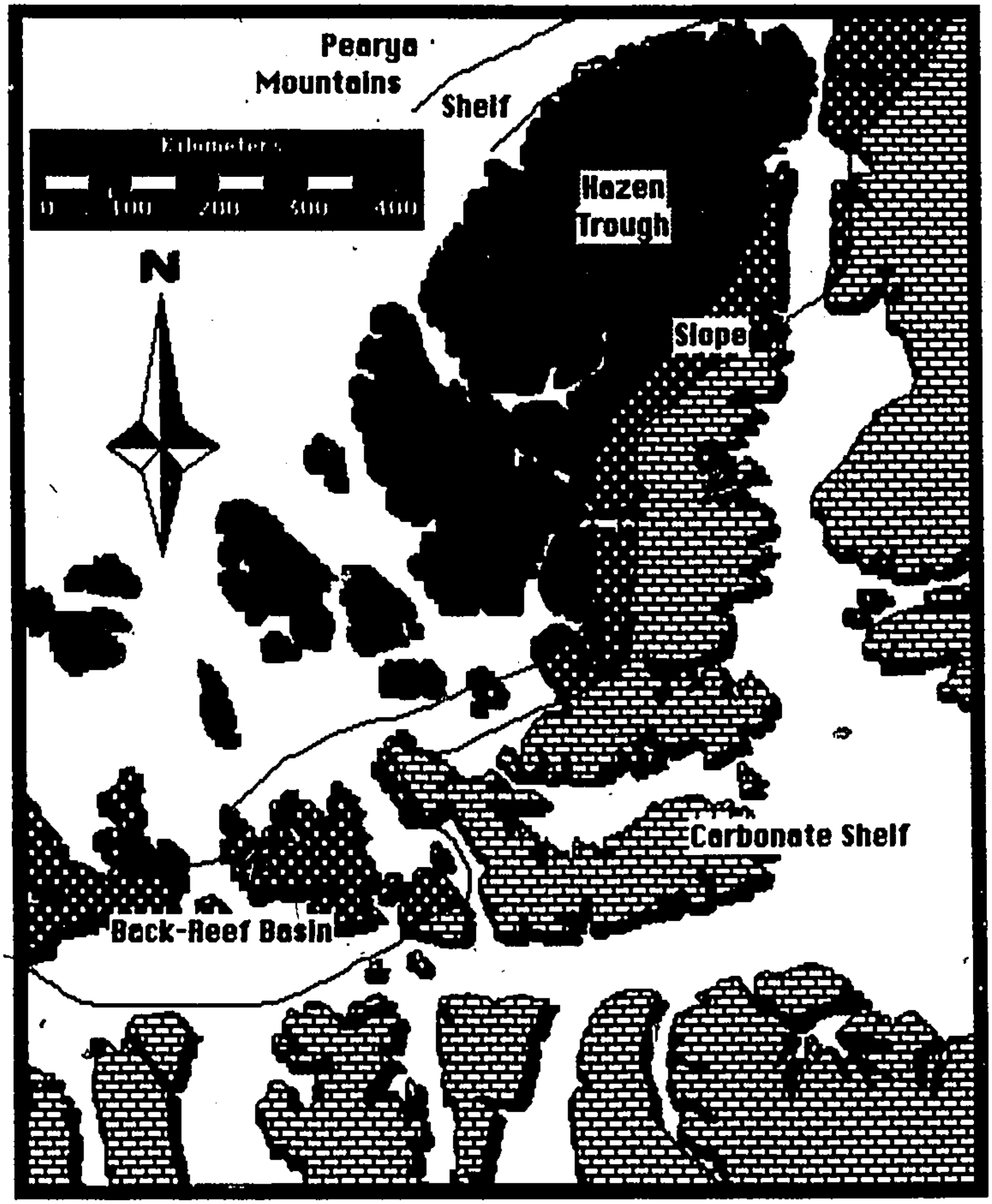


Figure 1.2 Paleogeographic map of the Canadian Arctic Archipelago for the Silurian. (adapted from Trettin, 1979 and Morrow & Kerr, 1973)

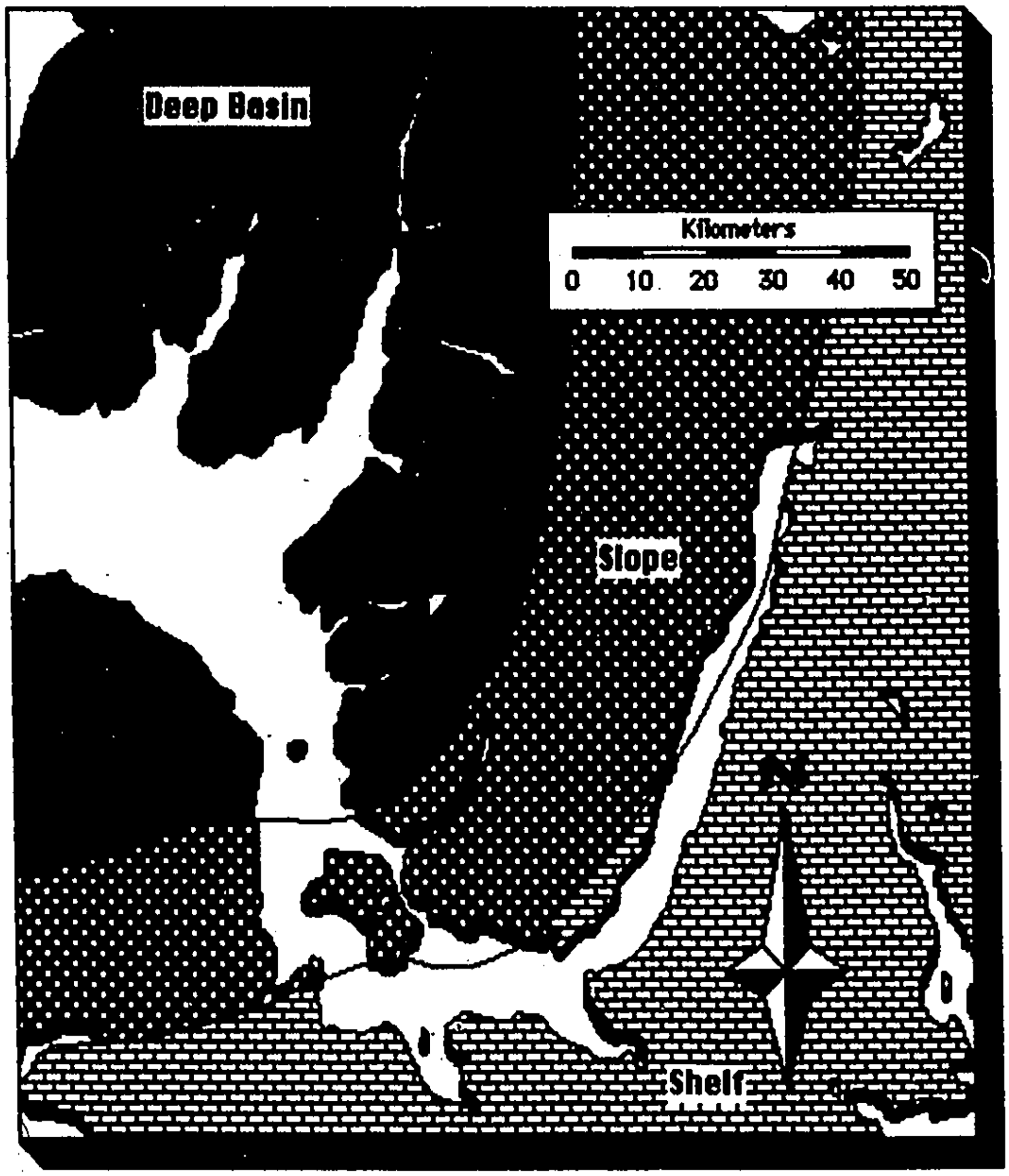


Figure 1.3 Distribution of Silurian depositional environments in the study area. (adapted from Mayr, 1974)

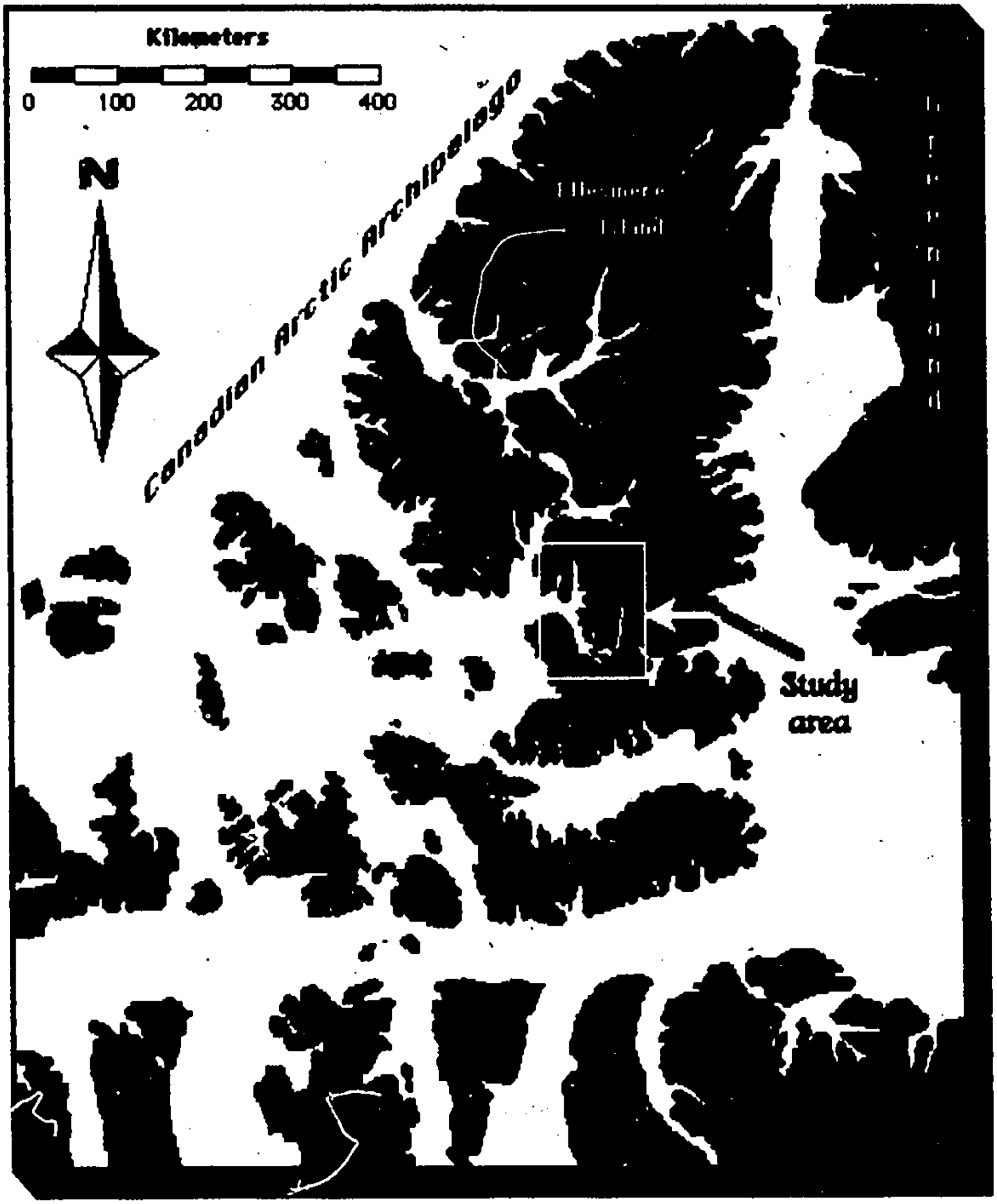


Figure 1.4 Location of the study area.

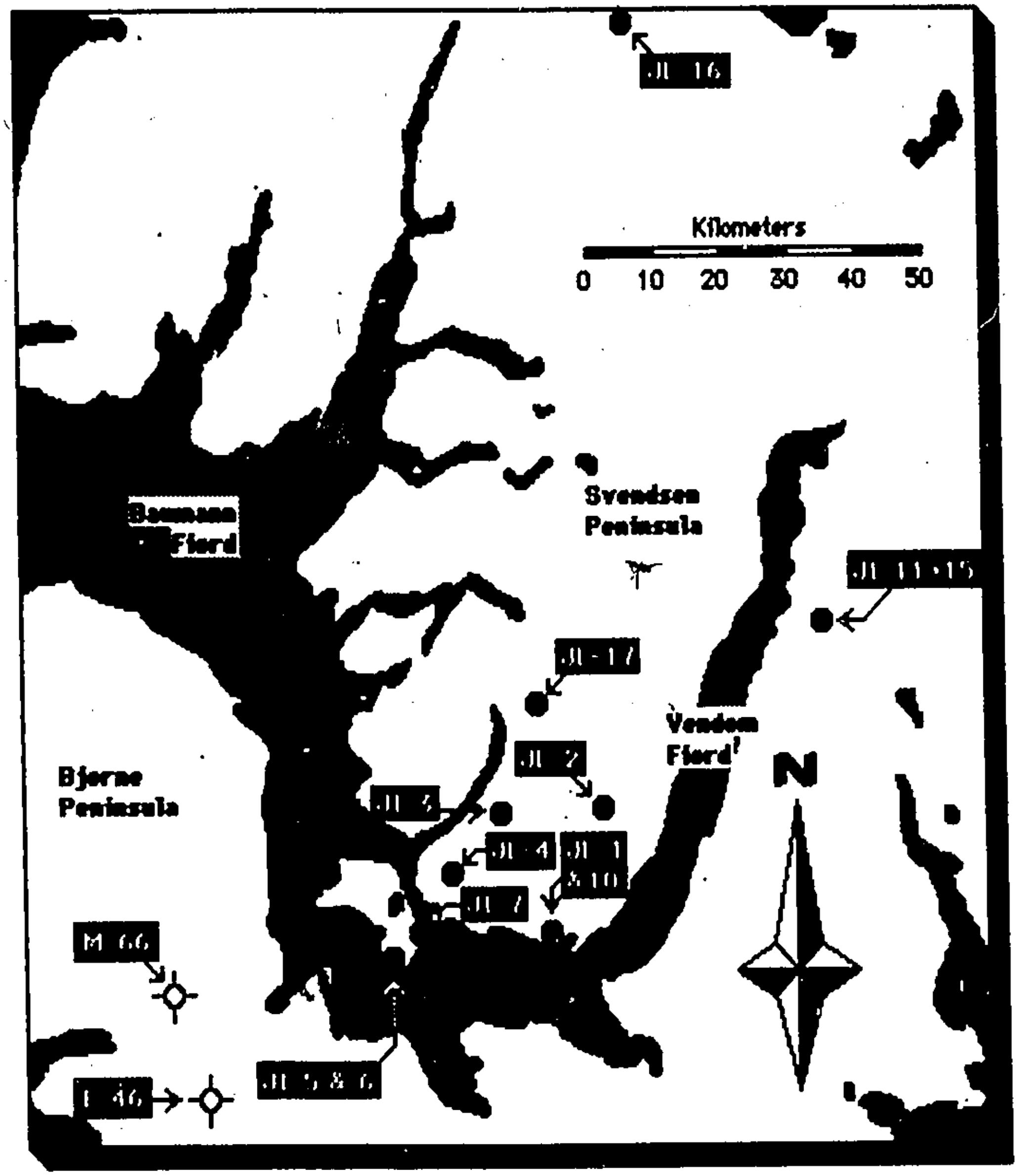


Figure 1.5 Location of measured sections (JL 01 to JL 17) and 2 dry wells (E-46 & M-66).

Few studies, and very few of them detailed, have been carried out in this large area (~ 2500 km^2 + few sections outside of that area, see Fig. 1.5), and in the Arctic Archipelago generally. Few institutions and private companies have conducted geological studies there because of the inaccessibility of the area, the logistics and preparations needed, and the enormous cost. This area preserves, however, an exceptionally complete transitional sequence between a carbonate shelf and a deeper trough. It is a sequence not commonly completely documented in the geological past, and represents a range of environments and processes not fully understood in modern depositional settings. Such transitional areas are presently of considerable interest in hydrocarbon exploration.

The formations included in this study display various lithologies, and show the complex and often abrupt lateral facies changes characterizing such transitional environments (see Fig. 1.6). Upper Ordovician to Upper Silurian carbonates are represented by the Irene Bay and Allen Bay formations and the Read Bay Group. Shoreward, the Cape Storm Formation, representing a more restricted and shallower environment, is intercalated between the Allen Bay Formation and the Read Bay Group (Kerr, 1975). A range of other formation names has been applied to the rocks transitional from the more open marine carbonates of the Allen Bay Formation and Read Bay Group, to highly restricted carbonates and clastics proximal to the coast line. These other formations lie outside of the study area; good summaries of the complex correlation of these shelf and cratonic rocks are given in Mayr (1978 & 1980), Miall (1976), Miall and Kerr (1980), and Trettin (1969). The shelf sequence passes basinward into relatively narrow belts of sedimentary facies running parallel to the paleo-shelf-margin. These represent shelf-margin and proximal slope environments (facies of the Allen Bay Formation-Read Bay Group undivided), and distal slope and rise environments represented by facies of the Cape Phillips and Devon Island formations. These transitional facies are often interdigitated making it difficult to define boundaries between them. The Cape Phillips and Devon Island facies ultimately pass basinward into flysch and starved basin facies of the Hazen and Imina formations, well beyond the study area.

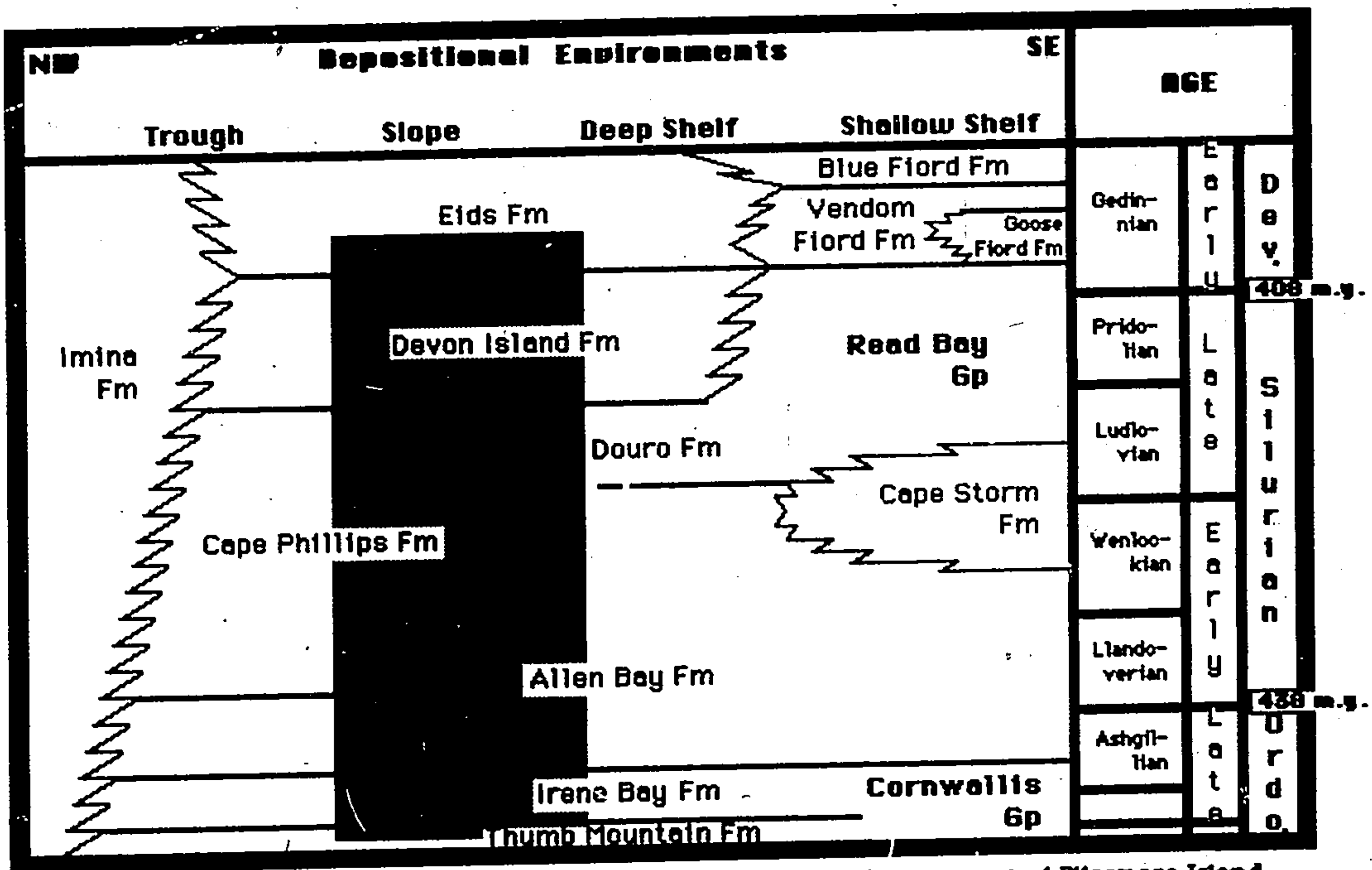


Figure 1.6 Stratigraphic relationships of Lower Paleozoic formations on central Ellesmere Island, based on Kerr (1975), Mayr (1978) and Trettin (1979). Shading represents formations in the study area.

1.2 Objectives

In comparison to other areas, geological work in the High Arctic has barely "scratched the surface". For this reason, most projects cover a variety of subjects necessary for a general interpretation of geologic history. This study on Svendsen Peninsula of west-central Ellesmere Island, is a general stratigraphic-sedimentological report with the following specific objectives:

a) The first detailed documentation in this area of the stratigraphy of Upper Ordovician to Lower Devonian shelf-margin, slope and basinal facies.

b) Detailed physical correlation of sections representing shelf-margin, proximal and distal slope, and basinal depositional environments.

c) Paleoenvironmental interpretation of the different facies.

d) Establishment of a time framework for the study using primarily conodonts (information provided by Dr. T.T. Uyeno, personal communication, 1982, 1984; from samples collected by the author) and graptolites.

e) Documention and interpretation of the development of shelf-margin reefs, and their influence on basin margin deposition, as well as their response to the evolution of the Franklinian Basin.

f) Assessment of early diagenetic processes, including early cementation, neomorphism, silicification, and dolomitization, and their very close association with, and influence on, shelf-margin and slope sedimentation.

g) Brief discussion of hydrocarbon potential of the studied formations in the area, probable hydrocarbon source beds, relative time of migration, and possible influence on diagenetic processes.

1.3 Previous Work

The limited published work available on the area consists largely of reports of reconnaissance mapping by the Geological Survey of Canada (G.S.C.). Mayr (1973) carried out reconnaissance studies in the study area as part of a G.S.C. mega-project. Trettin (1967, 1969a, 1969b, 1970, 1972 & 1973) has studied time-equivalent deposits of Early Paleozoic age, much farther north on Ellesmere Island. Most of those rocks, especially the axial Hazen Trough deposits, are very different from the sediments in this area. However, Trettin's study (1979) in the Cañon Fiord area included rocks very similar to those in the study area. Kerr (1967, 1968 & 1976) also briefly inspected the region, although he concentrated more on typical shelf sections in south-central and southeastern Ellesmere Island. Various studies by G.S.C. personnel on other areas of the Arctic (e.g. Kerr, 1974, 1975; Thorsteinsson, 1958, 1963, 1980; Thorsteinsson & Kerr, 1968; Thorsteinsson & Tozer, 1960; Mayr, 1978; Christie, 1967; McLaren, 1963; Tozer, 1963; Norris, 1963a, 1963b; and Morrow & Kerr, 1977) provide important stratigraphical information on correlative rocks.

The Arctic Geology Group at the University of Ottawa has carried out studies of time-equivalent rocks further south, on Devon, Cornwallis and Somerset islands. These dealt with deposits representing various shelf, shelf-margin and slope environments, commonly with extensive reefal development. Representative studies include those on reefs and reefal faunas (Dixon, 1979; Rigby & Dixon, 1979; Narbonne & Dixon, 1984), on Silurian stratigraphy and sedimentology (Narbonne, 1979; Narbonne & Dixon, 1983; Oldershaw & Narbonne, 1979; Gibling & Narbonne, 1977; Jones, 1974; Jones & Dixon, 1975; Dixon & Jones, 1977, 1978; Packard & Dixon, 1979, 1983) and on sedimentary geochemistry (Veizer et al., 1978; Brand & Veizer, 1979, 1980 and 1981).

The carbonate shelf and basinal deposits of the Franklinian Basin can be correlated across the Arctic Archipelago, and across Nares Strait into a sedimentary belt on northern Greenland. Only very recently has that belt been studied systematically, and only at a reconnaissance level (Hurst, 1980; Hurst & Surlyk, 1980, 1982 and 1983; Hurst et al., 1983; Surlyk, 1982; Christie & Peel, 1977; Mabillard, 1980; Mayr, 1976). A few studies tentatively correlated the rocks on Ellesmere Island and Greenland (Hurst, 1981; Hurst & Kerr, 1982).

A predicted world-wide hydrocarbon shortage in the 70's resulted in drilling of a few exploratory holes in Paleozoic strata of the High Arctic. However, these high risk ventures didn't bring the hoped-for dividends and so only a few reports have been published. These give only very approximate overviews of hydrocarbon potentials (Stuart-Smith & Wennekers, 1977; Sodero & Hobson, 1979).

In summary, very little detailed work has been done in south-central and southeast Ellesmere Island. An indication of this is the fact that major lithostratigraphic units such as the Allen Bay Formation, Read Bay Group and to a lesser extent the Cape Phillips and Devon Island formations remain undivided there.

1.4 Methods

With the objectives, above, in mind, a variety of standard field and laboratory techniques were used to carry out the stratigraphic/ sedimentological study. Fifteen stratigraphic sections (Fig. 1.5) were measured, described and sampled over two field seasons. Two months of field studies were carried out in 1981 in conjunction with a G.S.C. field mapping program in the region, and were supported out of a base camp on southern Svendsen Peninsula. Two months of field work from several 2-man fly-camps across the remainder of the study area, were completed in 1982. The documented sections provide the stratigraphic framework for the study. Detailed logs of these stratigraphic sections are on file in the Department of Geology, University of Ottawa.

Sections were measured using a 1.5 m Jacob's staff. Lithology, textures, bedding, unit thickness, sedimentary structures, fossil components and weathering characteristics were recorded systematically. Lithological sampling was carried out as follows: first, samples were collected representative of every lithology identifiable in the field, even minor ones; secondly, samples were taken of interesting or problematic rock types, even minor components (e.g. clasts and larger blocks), and included ones needed to verify characteristics associated with particular lithologies (e.g. basal portions of mass flows).

Representative macrofossils were collected throughout the sections, particularly groups such as graptolites in the basinal deposits, with the best potential for dating and correlation. Other representative macrofossils were collected, sufficiently well preserved for general identification. In addition, 1 kg rock samples, to be processed for conodonts, were collected at ~ 10 m intervals through a few selected sections and at closer intervals through more critical portions such as near major contacts. These samples were processed and the conodonts identified by Dr. T.T. Uyeno at the Geological Survey of Canada. Some sections were partly re-examined or re-sampled in greater detail during the second field season in order to resolve particular problems.

All lithological samples were examined in the laboratory with a binocular microscope, some as cut and polished hand specimens. Three hundred standard thin sections were prepared from selected hand specimens, stained with alizarin red (Friedman, 1969). These were the basis for microfacies analysis, a critical aspect of the study that commonly resulted in modification of, or correction of field observations of lithology and texture, especially for the fine-grained rocks that predominate in the sequence. The thin section study permitted more confident identification of finer fossil material and of colonial organisms such as calcareous algae, corals, bryozoans and stromatoporoids. The thin sections were also essential to study cement generations and for the discrimination of multigeneration clasts in the slope deposits.

Two sets of samples were analysed geochemically, one to study the relationship of dolomitization to primary zonation in a reef, and the other to document step-by-step dolomitization of deep shelf deposits of the Allen Bay Fm. Both involved XRF analysis following procedures from a University of Ottawa lab manual prepared by Dr. J. Veizer (1979).

1.5 Tectonic setting

The Lower Paleozoic sediments of this study on south-central Ellesmere Island are part of a much more extensive belt of rocks (Fig. 1.1), generally all similarly situated paleogeographically relative to the Franklinian Basin. This basin has been classified as a back-arc basin by Bally and Nelson (1980), and a modern counterpart might be the back-arc basin that has evolved since middle Tertiary time in the Java-Sumatra area (Kingston et al., 1983).

The Franklinian Basin developed from Early Cambrian to Late Devonian time. It is part of a larger tectonic province, the Innuitian Mobile Belt, that in the early Paleozoic included the Pearya Mountains, a postulated forearc basin beyond the Franklinian Basin to the northwest, and a subduction zone dipping towards the Central Stable Region of North America, beneath which an oceanic plate was engulfed (Trettin & Balkwill, 1979; Trettin, 1979). All evidence points toward a late Proterozoic origin for the mobile belt; a major uplift, the Eurekan Orogeny, in late Cretaceous to Miocene time terminated it.

In a tectonic model such as the one outlined by Dietz and Holden (1974), the Franklinian Basin would be a geosyncline, and the Pearya Mountains a geanticlinal ridge or a magmatic arc (Trettin et al., 1972). Drummond (1973) attempted to apply such a model to the Arctic Archipelago, dividing the Franklinian Basin into a miogeosynclinal trough (carbonate shelf) and a eugeosynclinal trough (Hazen Trough), with a miogeanticlinal ridge, as yet unindentified, in between. According to the geosynclinal model, the eugeosynclinal trough should be underlain by oceanic crust (Dietz & Holden, 1974), but there is strong evidence that the Hazen Trough is underlain by continental crust (Trettin & Balkwill, 1979). It is evident, therefore, that various terms have been applied in the literature to components of the Franklinian Basin, according to different tectonic interpretations of the mobile belt. In this study, the general terms shelf, shelf-margin, slope, and basin have been adopted and will be used throughout to refer to studied portions of the southeast flank of the Franklinian Basin.

The rocks of the Innuitian Mobile Belt record different periods of subsidence and uplift, often diachronous from one region to another and of varying significance between sedimentary belts (Trettin, 1972). The uplifts and deformations were related to three major periods of orogeny. Firstly, a Mid-Paleozoic Epeirogeny, from Late Silurian to Early Devonian time (Drummond, 1973), was related to the Caledonian Orogeny (Thorsteinsson & Tozer, 1970; Trettin, 1972) that marked closure of the Iapetus Ocean east of the North American craton (McKerrow, 1982; Hurst et al., 1983). Evidences for this Mid-Paleozoic Epeirogeny in the Franklinian Basin consist mostly of a series of uplifts (Kerr, 1977; Okulitch et al, 1986). Secondly, the Ellesmerian Orogeny, Late Devonian to Early Mississippian (Thorsteinsson, 1970) marked the end of the Franklinian Basin. The Pearya Mountains were uplifted diachronously from northeast to southwest, as well as portions of the Arctic

Platform, leading to exposure of the shelf area (Trettin, 1974). The Hazen Trough acted as a trap for sediments from those newly exposed areas, until it was filled and then exposed (Trettin, 1974; Smith & Stearn, 1982). Thirdly, the Eurekan Orogeny, Late Cretaceous to Miocene (Tozer, 1970), terminated the Sverdrup Basin, a successor basin to the Franklinian Basin, that began in Pennsylvanian time (Drummond, 1973).

Only the Mid-Paleozoic Epeirogeny influenced deposition of the studied sediments. This epeirogeny was mainly responsible for formation of the Cornwallis Fold Belt. Although the belt itself is too far south to have been a major influence in the study area, some deformation, uplift and block-faulting in the vicinity of the study area correspond in time to pulses of deformation of the Cornwallis Fold Belt. Kerr (1977) divided the formation of the fold belt into four main pulses, of which three can be related to the Mid-Paleozoic Epeirogeny or movements prior to but related to it. Kerr's two earliest pulses correspond to major periods of deformation recorded in the study area, Pulse 1 in the late Early Silurian, and Pulse 2 in the early Early Devonian. Much further to the northeast, studies in northern Greenland also showed probable block-faulting during the Early Silurian related to pre-Caledonian movements (Hurst, 1980; Hurst & Surlyk, 1983; Hurst et al., 1983; Surlyk, 1980 and 1982). Extensive block-faulting has been documented for the Cañon Fiord area in central Ellesmere Island (Trettin, 1973; Trettin & Balkwill, 1979) and for northwestern Devon Island (Morrow & Kerr, 1977). Sedimentation in the study area was influenced by this widespread epeirogeny and the evidence and regional relationships are discussed in more detail in subsequent parts of the thesis.

1.6 Generalized Geological History

Deposition of sediments in the Franklinian Basin can be divided into five phases, involving transgressive-regressive cycles or deepening-shallowing upward sequences: (1) late Proterozoic to Early Cambrian, (2) Middle to Late Cambrian, (3) Early to early Middle Ordovician, (4) late Middle Ordovician to early Middle Devonian, and (5) Middle to Late Devonian (Trettin, 1972 and 1973). Rocks of the late Proterozoic to Early Cambrian phase rest on the Precambrian basement and represent deposition prior to development of the Hazen Trough. They include siliciclastic and carbonate rocks deposited on a shelf under restricted conditions (Kerr, 1967; Trettin & Balkwill, 1979).

The Middle to Late Cambrian second phase of the Franklinian Basin recorded the differentiation of the Hazen Trough from shelf areas to the northwest and southeast (Fig. 1.1). The sediments of the Hazen Trough are those commonly associated with deep basinal euxinic conditions: mostly fine carbonates, some resedimented carbonates, and minor siliciclastics and cherts. They constitute the lower part of the Hazen Formation (Trettin, 1971 and 1979). Their shallow shelf equivalents can be divided into two groups or belts: an outer belt of limestones and fine to coarse siliciclastic rocks (Thorsteinsson, 1963) representing more open circulation, and an inner belt of somewhat similar, but generally coarser grained, deposits and some evaporites, representing restricted peritidal environments (Packard & Mayr, 1982; Christie & Embry, 1981).

The trough sediments of the Early to early Middle Ordovician third phase of the Franklinian Basin are also included in the Hazen Formation (Trettin, 1971 and 1979). The shelf equivalents begin with deposition of open marine mottled calcareous cherty dolostone of the Cape Clay Formation. Overlying limestones of the Baumann Fiord Formation (Early Ordovician) represent sabkha environments in part, as indicated by the presence of abundant gypsum and anhydrite (Kerr, 1967; Mossop, 1979; Packard & Mayr, 1982). The third phase concluded with deposition of open marine mottled dolomitic limestone, quite fossiliferous in places, of the Eleanor River Formation (Packard & Mayr, 1982; Kerr, 1968).

The late Middle Ordovician to early Middle Devonian fourth phase of the Franklinian Basin was the longest, with sedimentation influenced by the Mid-Paleozoic Epeirogeny (Drummond, 1973). Early Silurian collapse of the shelf-edge due to block-faulting is documented for the major part of the Franklinian Basin, from Cornwallis Island (Kerr, 1977) to northern Greenland (Hurst et al., 1983). The Silurian-Devonian boundary was marked by important movements on the Boothia Uplift (Thorsteinsson & Uyeno, 1980; Kerr, 1977), the Rens Fiord Uplift (Trettin, 1972), and the Bache Peninsula Uplift (Kerr, 1976) or Inglefield Uplift (Smith and Okulitch, 1987) and was also partly linked to shelf-edge block-faulting (Trettin and Balkwill, 1979). Later in the Early Devonian, Kerr's (1977) third pulse of the Cornwallis Disturbance resulted in major siliciclastic input on the carbonate shelf and completed the fourth phase.

The fourth phase of deposition is represented by a thick and diverse sedimentary sequence, part of which is the subject of this thesis. In shelf areas it started with deposition of restricted shelf argillaceous dolostone and anhydrite of the Middle Ordovician Bay Fiord Formation (Packard & Mayr, 1982; Kerr, 1967). In the study area, open marine conditions followed and persisted for the remaining period. The Bay Fiord Formation is overlain by fossiliferous, dolomitic, mottled limestone of the upper Middle to lower Upper Ordovician Thumb Mountain Formation (Kerr, 1968; Morrow & Kerr, 1977; Mayr, 1978). These are overlain by argillaceous, fossiliferous, dolomitic, mottled limestone of the Upper Ordovician Irene Bay Formation (Figs. 1.6 & 1.7) deposited on a subsiding shelf (Morrow & Kerr, 1977; Mayr, 1978). The latter formations contain the typical "Arctic Ordovician" fauna (Morrow & Kerr, 1978).

From Late Ordovician to Early Devonian time, the Allen Bay Formation, Cape Storm Formation and the Read Bay Group (Figs. 1.6 & 1.7) were deposited (Kerr, 1976; Mayr, 1973). These represent shelf and shelf-margin conditions with a wide variety of associated carbonate sediments. Some have been described in general terms in the literature (e.g. Mayr, 1973; McGill, 1974; Kerr, 1976; Trettin, 1979; Poey, 1982) and these rocks in the study area are discussed in detail in subsequent chapters.

In the Hazen Trough deposition of fine basinal sediments of the Hazen Formation continued into Early Silurian time. It was succeeded by flysch sedimentation, forming the Middle and Upper Silurian Imina Formation and post-flysch deposition of calcareous siliciclastic rocks of the Lower Devonian Eids Formation.

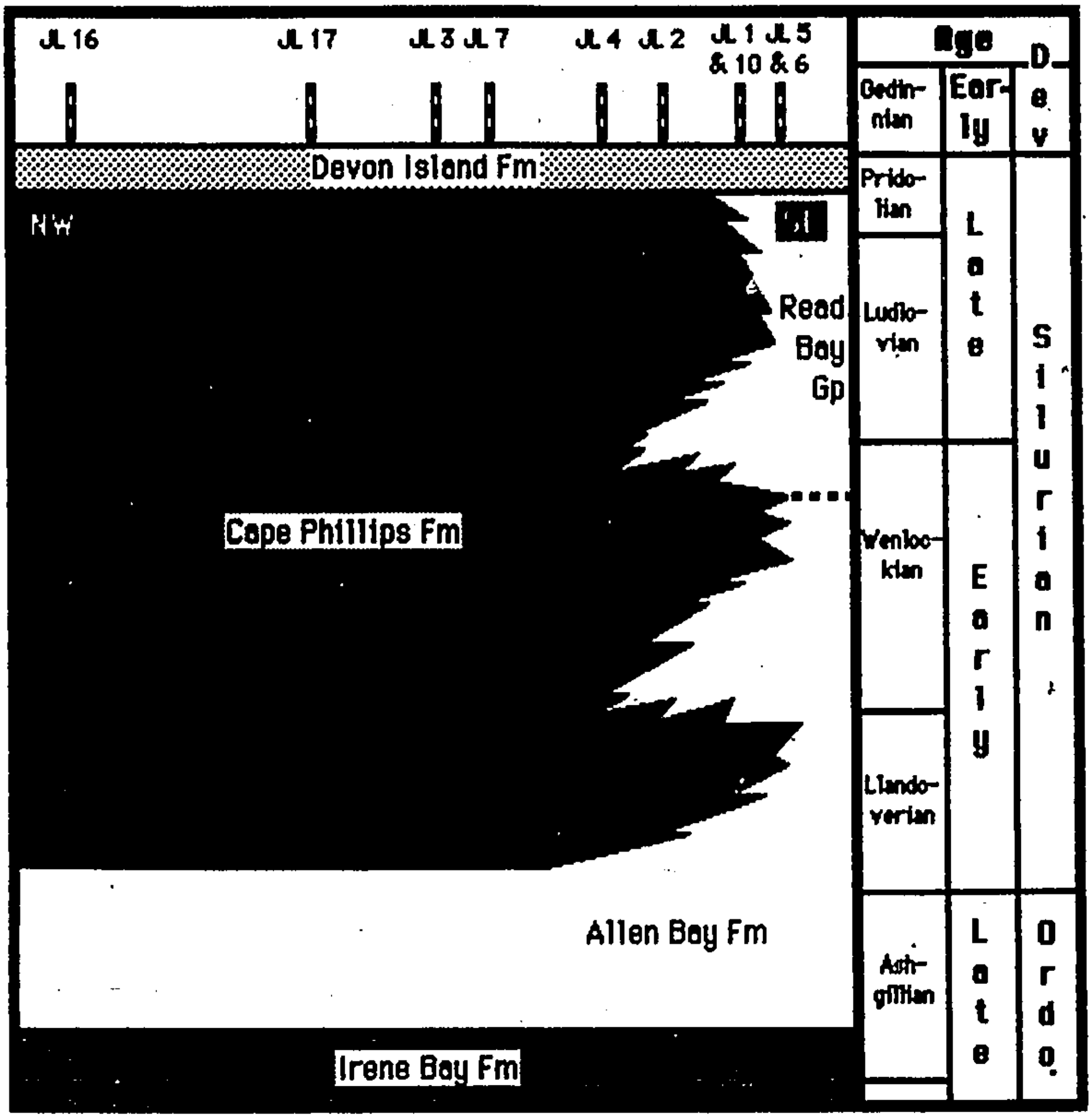


Figure 1.7 Stratigraphy of the study area showing the positions of sections across a generalized NW-SE profile. (adapted from Mayr, 1974)

Transitional between the deep basinal facies of the Hazen and Imina formations and the shelf-edge and shelf facies of the Allen Bay Formation-Read Bay Group undivided, are deposits of a slope environment and/or a back-reef basin. They are mostly marlstones interbedded with very fine grained limestones and some coarser resedimented carbonates, forming the Upper Ordovician to lowest Devonian Cape Phillips Formation (Kerr, 1976;

Morrow and Kerr, 1977; Trettin, 1979). In shelf areas, beds comparable to the Cape Phillips Formation, in the transition into Eids facies, have been distinguished as the Devon Island Formation. This Upper Silurian to Lower Devonian formation consists of dark, argillaceous, silty, fine grained limestone interbedded with calcareous shale (Morrow & Kerr, 1977) and, in part, succeeds limestones of the Read Bay Group in the study area (Figs. 1.6 & 1.7).

The fifth and last phase of the Franklinian Basin was marked by major uplift and erosion of the shelf area, and filling-in of the trough (Embry & Klovan, 1976; Smith & Stearn, 1982; Kerr, 1976). The uppermost unconformity represents termination of deposition in the Franklinian Basin (Drummond, 1972).

1.7 Acknowledgements

Many individuals and organisations contributed to the successful completion of this study. The study was part of a larger scale mapping program on southern Ellesmere Island by the Geological Survey of Canada (Institute of Sedimentary and Petroleum Geology) which contributed the idea for the project, as well as material and monetary support for the two summers spent in the field. I am grateful to Drs. U. Mayr, R. Thorsteinsson, and T. T. Uyeno, of the Institute, for their contributions to stratigraphy, graptolite identification and conodont processing and identification, respectively. Dr. C. Harrison, also with the Institute, made available part of a Petro-Canada report in the area, which was greatly appreciated. The study was financially supported by Natural Sciences and Engineering Research Council Grant A 5121 (to O.A. Dixon) and by funds from the Department of Indian and Northern Affairs, administered by the Northern Research Group, University of Ottawa. Continued logistic support in the field by the Polar Continental Shelf Project, under difficult field conditions, was greatly appreciated.

I would like to thank also a number of persons at the University of Ottawa. Mr. E. Hearn provided valuable advice on photography and drafting. Mr. R. Taylor skilfully prepared the numerous thin sections used. The assistance given by Ms. B. Bilodeau in some of the geochemical analysis was of great help; the discussions of those analyses with Dr. J. Veizer contributed to the evaluation of diagenesis, especially dolomitization, of those sediments. I would like to express my appreciation to Gary Graf, Iain Muir, Jeff Packard and Bill Parkins, my fellow students of the "Paleontology lab" for the numerous informal discussions on thesis and non-thesis topics. I would also like to thank Dana Naldrett for his useful advice on computer applications. The assistance given by 5 field assistants, Gord Perkins, Liz Ferguson, Dorian Sylvestre, Suzanne Sylvestre, and especially Luc Lamarche, contributed both to the completion of the project and to making those summers enjoyable.

A contributor without whom this thesis wouldn't have been possible is Dr. O.A. Dixon, my thesis advisor. In addition to advice on a large number of subjects, his corrections of the thesis, his obvious interest in the subject, and support for field trip and conference travel further enhanced my comprehension of the subject and are greatly appreciated.

Chapter 2

Irene Bay Formation

2.1 Stratigraphy and Age

The Irene Bay Formation is widely used as a marker unit in correlating Lower Paleozoic sequences in the Canadian Arctic. Its highly characteristic greenish-grey colour, bedding style, and sediment types contrast with formations above and below. These characters, together with its relatively constant thickness over a wide regional extent (discussed below), and its contained invertebrate fossils, the so-called "Arctic" Ordovician fauna (Kerr, 1968), all make the formation readily recognizable and useful both in surface mapping and subsurface correlation.

The Irene Bay Formation is the uppermost formation of the Cornwallis Group which also includes the Bay Fjord and Thumb Mountain formations (Thorsteinsson, 1958). It is generally accepted that the group ranges in age from Middle Ordovician (latest Llanvirnian) to Late Ordovician (earliest Ashgillian). The Irene Bay is the thinnest of these formations and has the shortest time range. Using macrofossils, Thorsteinsson (1958), and Thorsteinsson and Kerr (1968) situated the Irene Bay Formation in the latest Caradocian. Subsequent conodont studies (Barnes, 1974) indicated latest Caradocian and early Ashgillian, the presently accepted range. A late Ordovician age was also indicated by conodonts from a few samples (appendix No. 1) collected in this study (Dr. T.T. Uyeno, personal communication, 1982). The upper contact of the Irene Bay Formation with the Allen Bay Formation-Read Bay Group undivided, is considered to be essentially a time horizon regionally (Kerr, 1968).

2.2 Distribution, Thickness and Contacts

The Irene Bay Formation is 83 m thick at its type locality, Irene Bay, on southwestern Ellesmere Island (Kerr, 1967). The formation is widely consistent in thickness (comparable to the type section), but shows large variations at shoreward and basinward extremities. For example, Kerr (1968) recorded thickness variations between 37 and 320 m on Ellesmere Island; Trettin (1979) concurred and recorded a thickness of 43 m at Cañon Fiord, approximately 250 km north of the Svendsen Peninsula study area. In the study area (further basinward from areas reported by Kerr and Trettin) the Irene Bay Formation shows considerable thickness variations, from a minimum of 7 m (at section JL 16, Fig. 3.4e) to a maximum 40 m at the southeast margin (see Fig. 3.4c), closest to the shoreline (see Fig. 1.5). Field lithologies failed to reflect progressively deeper depositional environments basinward. However, the thinning of the Irene Bay Formation northwestward suggests deepening, and was very useful in interpreting the relative positions of the different sections. According to Trettin (1979) and Kerr (1978), the Irene Bay Formation also thins shoreward from the subsiding shelf to the stable Arctic Platform.

The lower and upper contacts of the Irene Bay Formation are conformable in the study area, as in areas farther to the southeast and south (Kerr, 1968; Morrow and Kerr, 1977). The lower contact with the Thumb Mountain Formation is gradational lithologically but abrupt in colour, from darker brown rocks below to light greenish-grey rocks above. This colour change is thought to be due to a higher bitumen content in less argillaceous and more dolomitic interbeds and "internodular" deposits between the more massive micritic limestones of the Thumb Mountain Formation. The upper contact of the Irene Bay Formation with the Allen Bay Formation is more gradational, but the transitional zone leading to the colour change is still quite thin. However, slightly above this contact, an approximately 5 m thick unit of limestone belonging to the basal Allen Bay Formation, is exactly the same light, greenish grey colour that characterizes the Irene Bay Formation. This Irene Bay Formation-like unit often leads to misidentification of this contact. The Irene Bay Formation is usually distinguished by having a much higher fossil content within its thin argillaceous interbeds than either the underlying or overlying formations.

2.3 Lithology

The Irene Bay Formation consists of thin-bedded lime mudstone or wackestone alternating with fossiliferous calcareous shale partings and beds. This alternation, apparently complicated by several generations of intense burrowing, gives to these beds a characteristic nodular or mottled appearance (Morrow and Kerr, 1977) (see Fig. 2.1). This recessive formation has a rubbly weathering character but is particularly characterized by its light greenish-grey weathering colour, contrasting with the darker brownish grey limestone of the underlying Thumb Mountain and overlying Allen Bay formations. This greenish-grey colour appears to be the result of weathering of the calcareous shale, as interpreted by Kerr (1968).

The proportions of nodular or bedded limestone and calcareous shale partings vary. In sections southeast of the study area (toward the paleo-shoreline), shale, as thin partings, is a minor constituent (Kerr, 1968). This compares with some sections in the study area (especially JL 01) in which the ratio of limestone to shale is 10 to 1. In sections to the north-northwest representing a deeper, though not basinal setting, the rocks are much more argillaceous, with limestone to shale ratios reduced to 1 : 1. A gradual change in the limestone-shale ratio was also noted by Kerr (1968).

The nodular and bedded limestone consists mostly of micritic lime mudstone, or rarely wackestone. It is fairly pure and homogeneous, lacks clay minerals, and rarely contains sand-sized skeletal fragments (mostly ostracods) and minor drusy mosaic spar cement filling some irregular primary openings and some fractures. The edges of the "nodules" or "nodular beds" vary in showing the results of neomorphic recrystallization of the micrite, namely microspar and pseudospar, that coarsens outward. This transitional aureole is usually not more than 100 μm in width, as shown by Plate 7-2 and Fig. 2.2.

The "matrix" surrounding the "nodules" or interbedded with the "nodular beds", is fossiliferous calcareous shale or marlstone that varies considerably in composition. Clay minerals predominate, but skeletal fragments are also very important and much more abundant than within the "nodular" limestone. The remaining constituents are disseminated,

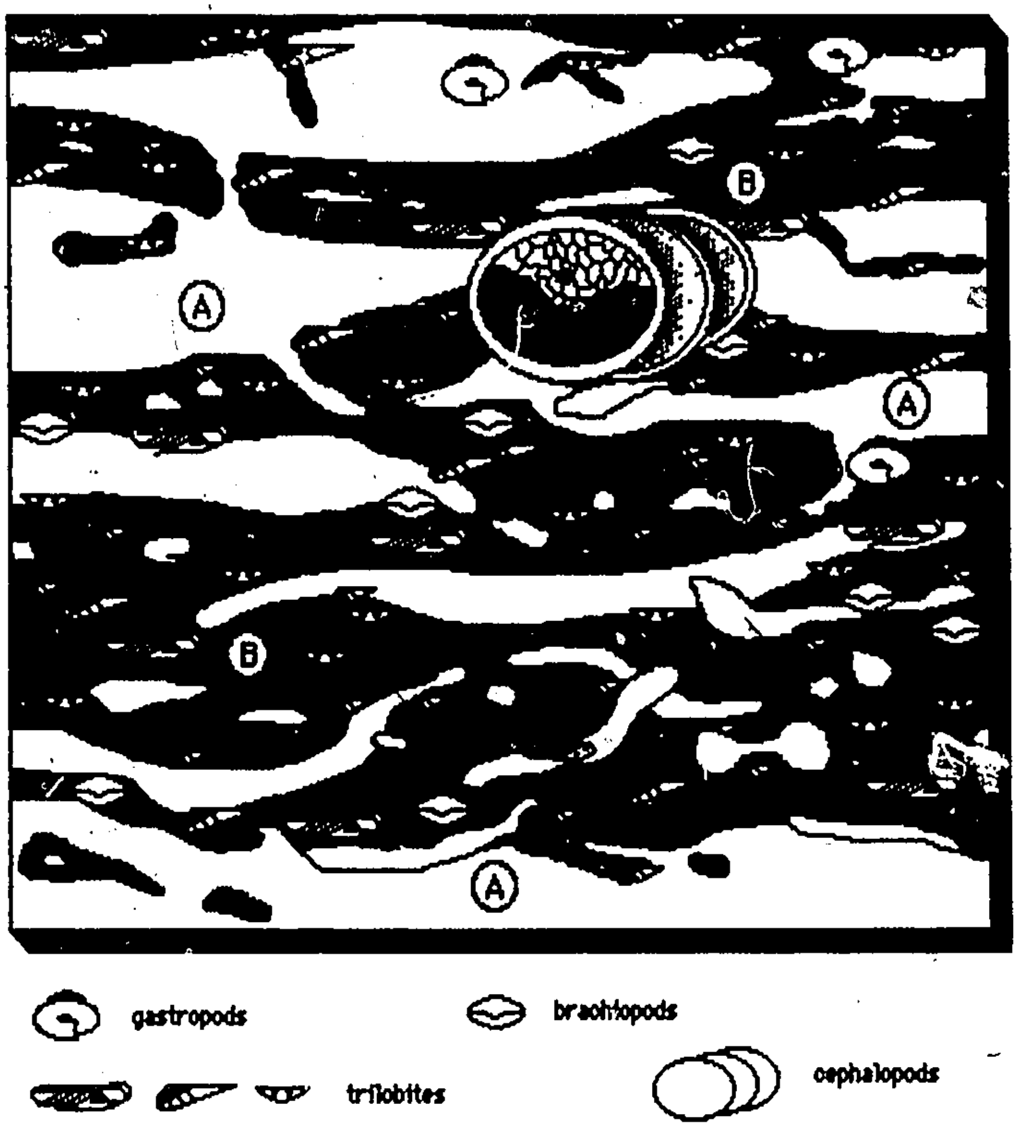


Figure 2.1 Diagrammatic representation of nodular limestone — Type I (discrete nodules) in the Irene Bay Fm or the bottom of the Allen Bay Fm. **A** represents nodular lime mudstone and **B** argillaceous lime mudstone matrix.

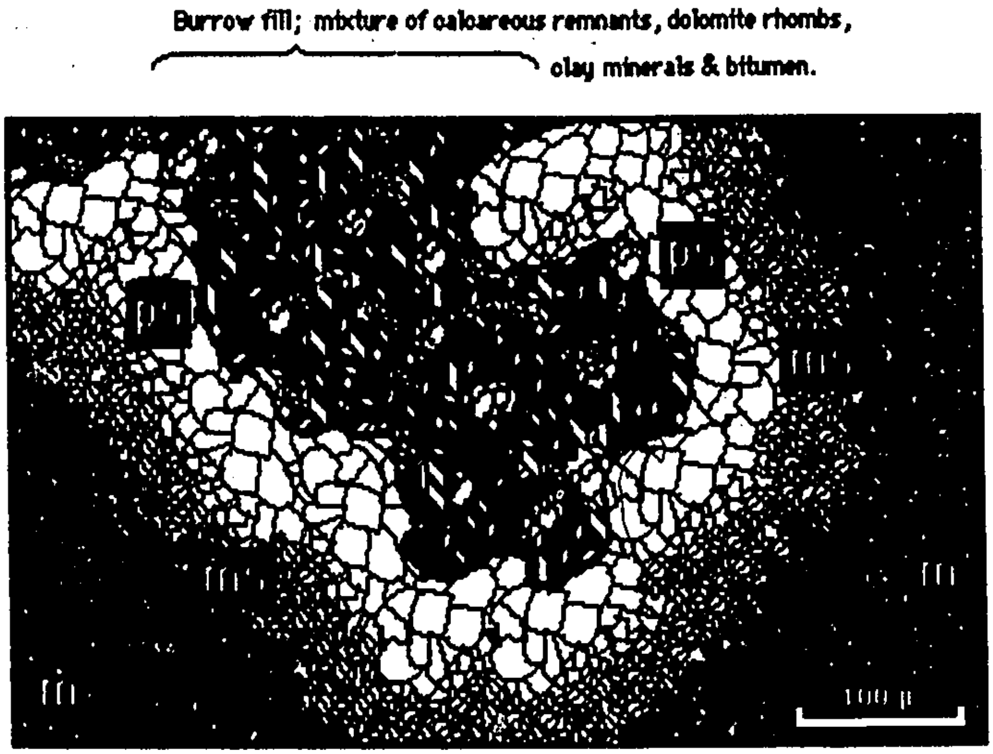


Figure 2.2 Diagrammatic representation of a neomorphic aureole around a burrow fill, Irene Bay Fm (m=micrite, ms=microspar, ps=pseudospar).

homogeneously distributed, recrystallized calcareous bodies, interpreted as calcareous relics of original burrow fill components, and very finely crystalline dolomite rhombs, replacing part of the calcareous sediment. Usually the dolomitization is far from complete, and the calcareous relics predominate over the dolomite rhombs. Also evident in thin sections, are augen-like bodies of mostly neomorphically transformed calcareous sediment, cut by anastomosing wisps of argillaceous matrix. The bodies vary from a few 10's of micrometers to a few millimeters in width.

The clay minerals are laths or plates about 5 μm long, that are parallel or subparallel to each other and tangential to "nodule" contacts, rather than to bedding. Since the "nodules" are generally elongated parallel to bedding, most of the clays are as well, except for those at the extremities of the "nodules" which are often oriented perpendicular to the bedding. Their orientation is consistent with compressional reorientation of clay minerals around more coherent "nodular" limestone.

The dolomite rhombs are very fine (< 1/16th mm), euhedral and equicrystalline. They are commonly disseminated homogeneously, indicating pervasive nucleation within the neomorphically transformed calcareous relics. They have a "dirty" appearance, reflecting numerous inclusions within the crystals.

Skeletal fragments in the "matrix" are much coarser than in the "nodular" limestone, the "matrix" often being a dolomitized floatstone rather than a wackestone. Most fragments are well preserved and show no evidence of dissolution. These skeletal components are mostly fragments of trilobites, ostracods, crinoid ossicles, brachiopods, gastropods and orthoconic cephalopods. A more detailed list follows.

2.4 Fossils

In addition to the features discussed earlier, the relative abundance and variety of fossils in the Irene Bay Formation compared to the formations below and above, help to make it a good marker, even though the formation is not as fossiliferous as some reported in the literature. The Irene Bay Formation, more than any other of the Cornwallis Group, is described as containing a good assemblage of the "Arctic Ordovician" fauna (Kerr, 1968).

Most of the fossils occur in the calcareous shale that encloses the predominant "nodules" and "nodular beds" of lime mudstone. Occasionally the fossil fragments are in grain-to-grain contact. The high fossil concentration might have been emphasized by post-depositional processes, as discussed later, but probably largely reflects original concentration. The fossils are mostly fragmented, and those, rarely, that are non-fragmented, are not in life-position. The fragments are commonly rudite size and larger, randomly oriented and quite poorly sorted.

The following fossils are present in the Irene Bay Formation:

- crinoids
- brachiopods
- ostracods
- trilobite fragments, probably belonging to *Ceraurus* sp.
- solitary rugosans, including *Streptelasma* and *Paleophyllum* sp.

- sponges (only as calcite spar-filled molds of individual spicules, probably siliceous originally)
- orthoconic cephalopods, some quite large, including *Billingsites* sp. and *Probillingsites* sp.
- bryozoan colonies, branching and bulbous
- *Catenipora* sp.
- *Receptaculites* sp.
- gastropods
- dasycladacean algae
- codiacean algae

2.5 Depositional environments and processes

The lithologies represented in the Irene Bay Formation are homogeneously distributed, with very little vertical variation and only minor lateral change. The recorded thinning, and increased argillaceous content of the formation toward the northwest, can be attributed to regional deepening toward the Hazen Trough.

The rocks resulted mostly from deposition of lime muds interrupted periodically by influx of subordinate clays. The rocks represent very low energy conditions typical of environments below fair-weather wave-base or in a protected lagoonal setting. No evidence of lagoonal setting, such as faunal restriction, shoaling-upward sequences, desiccation cracks or very fine lamination is apparent, and furthermore, a lagoonal setting would be inconsistent with the wide geographical distribution of such homogeneous sediments. Morrow and Kerr (1977) concluded that the Irene Bay Formation was deposited in a substantially deeper environment than was the underlying Thumb Mountain Formation and represented brief, relatively deep submergence and drowning of the Thumb Mountain shelf-lagoon complex.

The alternation between limestone and marlstone layers could have several causes. Firstly, in view of the adverse effect that rise of relative sea level would have on carbonate production, cyclic increases in turbidity would further strain the weak balance, resulting in decrease in carbonate sedimentation and higher concentration of clay minerals. Clays were probably the reason for the increase in turbidity levels, as they evidently represent the background sedimentation in the "compressed" marlstone layers. The "compression" of these layers would explain the higher concentration of allochems and burrows, though some associated organisms, such as trilobites, gastropods and cephalopods, did not need special light conditions for subsistence. However, fossil remains of solitary rugosans and crinoids are also slightly more abundant in those layers. Since at least the solitary rugosans are known to have needed firm attachment, they were probably attached to at least partially cemented limestone layers and not associated with marlstone deposition.

Secondly, this interbedding could have been solely the result of episodic influx of terrigenous muds in suspension, requiring no significant change in the rate of *in situ* carbonate production. Such increase in the sedimentation rates of terrigenous components would not explain, though, the associated increase in biogenic activity as described above, since higher sedimentation rates would tend to dilute rather than concentrate the evidence of biogenic activity, assuming of course no change in faunal concentration.

Thirdly, periodic increase in turbidity and/or terrigenous sedimentation could be the product of upwelling coming from the Hazen Trough. The amount of components introduced could have varied with time depending on supply of terrigenous muds to the trough and circulation patterns in the trough. Upwelling currents are also well known to be rich in nutrients, and if cyclic could have induced corresponding changes in biogenic activity. Often cooler in temperature than the ambient water of the shelf, upwelling currents can also promote penecontemporaneous lithification of lime muds, a probable diagenetic process of the Irene Bay Formation, as discussed below.

Evidence for alternative explanations for the limestone-marlstone alternations is lacking. Certainly there is no evidence of turbidites or mass flows or cyclicity generated by storm activity.

Similar nodular calcareous rocks occur in the Upper Cambrian of the southwest Virginia Appalachians, where Markello and Read (1982) described them as dolomottled ribbon rocks. In those deposits mottled irregular layers of lime mudstone, wackestone and skeletal packstone are disrupted by slightly dolomitized burrow fills. Read (1980) described similar Middle

Ordovician deep-ramp carbonate facies in the Virginian Appalachians. Although more bituminous (reflecting lower and fluctuating O_2 levels according to Read) and less argillaceous than the Irene Bay Formation, they show very similar characteristics: lack of shallow-water sedimentary structures, presence of diverse biota, well-preserved but variably fragmented fossils, abundant burrows, abundant unwinnowed fine carbonate and terrigenous material, and nodular or ribbon distribution of the main deposit. The difference in argillaceous content could be due to variations in amounts coming from their source areas, i.e. the Canadian Shield, in comparison to the Virginian Appalachians, maybe reflecting a more proximal location, or differences in basinal current circulation between the two basins.

Markello and Read (1982) interpreted such deposition to have occurred between fair-weather and storm wave-base. Little evidence exists for maximum depth limit except that in the study area a well-oxygenated environment is suggested by the light greenish-grey colour of the sediments and their low bitumen content (in some burrows, containing bitumen, the time of migration of the organic matter and its source are not clear).

The varied fauna also provides information about environments of deposition. First, the fossil fragments show no evidence of preferred orientation, which suggests that currents neither influenced deposition nor re-oriented shells after deposition. This implies deposition below tidal influence or even below wave-base.

Secondly, though most of the shells and exoskeletons were fragmented, the fragmentation was not extreme, with a large number of the fossils still whole or in large fragments. The allochems are very poorly sorted with grain sizes ranging widely from miniscule spicules to orthoconic cephalopod shells up to 18 cm in diameter. This also points to insignificant current activity. Furthermore, most of these poorly sorted fragments are present in calcareous shale layers or inter-nodular areas (interpreted as burrow fills). It is unclear if this intense biogenic activity during deposition of the argillaceous layers and the associated concentration of fossil fragments represent increased biota during those periods, or simply reduced carbonate deposition relative to the argillaceous sedimentation, and concentration through compaction. In either case, distribution and poor sorting suggest that the fragments were produced by biogenic activity rather than transport.

Burrows provide a third piece of evidence regarding depositional environments. In the Irene Bay Formation, the most important burrow systems by volume are those filled by strongly argillaceous sediment. Due to subsequent differential compaction, only those components extending vertically and interrupting the horizontal layering of the lime mudstone, remain evident, but probably the argillaceous interbeds were also intensely burrowed. The vertical burrows responsible for the nodular structure of the lime mudstone layers were not sufficiently abundant to destroy the layering in these beds. Because of deformation little else can be concluded about these burrows although they are common. Burrowing of the argillaceous sediments also took place during lime mud deposition. Although this burrowing was apparently less intense, the decreased burrow frequency might, alternatively, be interpreted as reflecting more rapid deposition. These lime mud-filled burrows are very well preserved, only slightly compressed and are almost exclusively horizontal. They also have much sharper, though not abrupt, contacts with the host sediment, the argillaceous interbeds. They include the ichnogenera *Phycodes* and *Planolites* with a range of diameters averaging 1.5 - 2 cm. The association of horizontal and vertical burrows is said to be characteristic of subtidal environments (Rhoads, 1967).

Additional important evidence for determining the depositional environment is the large diversity of the Irene Bay biota. Most were bottom dwellers, but some were nektonic and they exemplify a variety of ecological niches and inferred life habits: suspension feeders (brachiopods, crinoids, sponges, rugose corals and bryozoans), sediment feeders or grazers (ostracods, trilobites, gastropods), and carnivores (gastropods and cephalopods). Such a wide variety of organisms could have survived only in an open marine setting, unstressed by other factors such as wave action, abnormal salinity, high turbidity, rapidly changing conditions or unstable sedimentary substrates. A subtidal environment is clearly represented (Walker and Laporte, 1970). The lower limit of water depth can be determined by the presence of calcareous algae such as *Receptaculites* sp. and dasycladacean green algae. Because of their need for light for photosynthesis, the dasycladacean algae can only grow in water less than about 92 m (300 ft) deep (Nitecki, 1970). It was previously concluded that currents and transport were insignificant during Irene Bay Formation deposition, so the algae must have been preserved *in situ.* However, in comparison to the underlying Thumb Mountain Formation (Morrow and Kerr, 1977) the fragments of calcareous algae are much less common, suggesting an environment of deposition towards the deeper end of the range proposed by Nitecki.

2.6 Post-depositional processes

Many post-depositional processes affected the Irene Bay Formation, most of them quite early in the burial history. Represented are the effects of burrowing, early cementation, neomorphism, dolomitization, recrystallization, differential compaction and hydrocarbon migration.

Cementation - The lime mud layers were cemented early enough to have an impact on burrowing patterns. While vertical burrowing penetrated the layers, it was not intensive enough to completely destroy that layering. The contacts between argillaceous burrow fills and surrounding lime mud are mostly fairly abrupt, though not razor sharp, indicating some degree of cementation of the host sediments. However, the character of burrows formed during lime mud deposition (therefore lime mud-filled) indicates that, by then, the preceding lime mud layer (only a few cm beneath the surface) was completely cemented. The lime mud-filled burrows penetrate downward from the tops of argillaceous layers but, where they extend through those layers, only penetrate similarly argillaceous older burrow fills in the lime layer, not the lime mud itself. This relationship was also used by Kendall (1977) in determining the time of cementation. In addition, the lime mud-filled burrows show little or no compaction even though the surrounding argillaceous sediment was compacted and deformed. The lime mud-filled burrows show very abrupt contacts with the surrounding sediments as a result of this differential early cementation and compaction.

Selective early cementation of the lime mud components might have been due to better original permeability and circulation of $CaCO_3$-rich fluids than in the argillaceous sediments in which porosity and permeability could have been considerably reduced by abundant clay minerals. However, after cementation of the lime muds, diagenetic fluids could have been restricted to more limited pore spaces in the argillaceous layers and clay-rich burrow fills, as suggested by Morrow and Kerr (1977) and also partly by Kendall (1977).

The evidence for this early cementation lies in selective neomorphism of the lime mud. Lime mud in argillaceous layers and clay-rich burrow fills was completely aggraded to microsparite and pseudosparite. In comparison, the lime mud of the mudstone layers and of other type of burrows, mostly remains as micrite with only very thin (100 μm in width or less) neomorphic aureoles at the contacts between lime mud-rich and clay-rich layers or burrow fills. These neomorphic aureoles may indicate very minor permeability in the lime mud layers, but could also have resulted from diffusion. These differences in neomorphic aggradation suggest that diagenetic fluids followed the argillaceous layers and burrow fills preferentially and, therefore, that the lime mud layers were already cemented. In addition, the lime mud layers show no indication of compaction, and some original pore spaces are filled by mosaic calcitic spar cement.

Early cementation is being recognized increasingly as an important process in deeper depositional environments. Sea floor crusts have been reported in the Mediterranean and Red seas and near Barbados, where very slow sedimentation prevails (McHargue and Price, 1982). Analogous conditions of slow sedimentation could possibly be inferred for the argillaceous layers of the Irene Bay Formation, though slower carbonate deposition is only one of the possible mechanisms referred to in this study. Many legs of the Deep Sea Drilling Project reported very early cementation, where little water circulation and slow deposition persisted (e.g. Matter, 1974). Reasons for this early cementation in stagnant marine phreatic zones, such as in the deeper water parts of gently sloping carbonate ramps, are unclear, but bacteria and microboring algae removing organic films and/or increasing the pH may have contributed to intergranular cementation in very small pore spaces (Longman, 1980).

Dolomitization - Dolomitization was locally quite important, and produced randomly distributed rhombs averaging 20-30 μm in size. Their occurrence only in the argillaceous sediments is further evidence to support early cementation and early reduction of permeability in the lime muds. According to Morrow and Kerr (1977), this dolomitization might have accompanied neomorphism, and resulted from release of Mg^{2+} ions. Alternatively, clays might have been the source for the Mg^{2+} ions (Land, 1983; McHargue and Price, 1982). The mechanism would be the early expulsion of connate water and associated ions due to early compaction or a more direct interaction with the ions attached to clay surfaces. A later stage of dolomitization would follow the expulsion of structural water through compaction. These processes will be examined in more detail in Chapter 3 for the much more extensively dolomitized Allen Bay Formation. Morrow (1978) proposed another theory to explain specifically the partial dolomitization of burrow fills in the Irene Bay and Thumb Mountain formations. This theory combines the hypersaline and mixing models (Land, 1973 and 1983; Badiozamani, 1973) and would require a permeability contrast between relatively permeable burrow fills and impermeable surrounding material. Seasonally hypersaline fluids were invoked to produce early proto-dolomite which would have served as nucleation sites for late-stage dolomite growth. This late stage dolomite growth could have been fed by ions diffusing into the permeable burrows from mixed meteoric-marine pore fluids moving along bedding planes (Morrow, 1978).

Neomorphism - As mentioned earlier, neomorphism accompanied processes of early cementation and dolomitization. First, it enormously modified the grain size of the calcareous portion of the argillaceous sediments, transforming it almost exclusively to microsparite and pseudosparite. In addition, neomorphic aureoles averaging approximately 100 μm in thickness, enclose bodies of lime mud sediment. The very early cementation of the lime mud, blocking intergranular pores, presumably formed a barrier that gradually stopped the migration of diagenetic fluids. Diffusion must have been important, but was limited in its depth of penetration. At the contacts between argillaceous sediments and lime mud, grain size within the neomorphic aureoles diminishes from pseudosparite, next to the argillaceous sediment, to microsparite, and then to dark micrite which predominates in the lime mud (see Plate 7-2 and Fig. 2.2). Evidently the diagenetic fluids were channellized by argillaceous burrow fills and/or argillaceous interbeds, presumably because biogenic reworking left them loosely packed and permeable. Fresh or brackish water was probably involved because neomorphism requires removal of intercrystalline Mg^{2+} ions (Folk, 1974). Morrow and Kerr (1977) postulated that fresh groundwater, possibly derived from the Canadian Shield to the southeast, invaded the formation during deposition.

These early post-depositional processes, in addition to burrowing and differential compaction, created the very characteristic nodular pattern observable today. It began with the deposition of alternate layers of lime mud and argillaceous sediment (for reasons discussed earlier). Burrows penetrating the lime mud layers were filled by argillaceous sediments, and burrows following the argillaceous layers were filled by lime mud. Very early differential cementation of the lime muds, making them impermeable, restricted subsequent movement of diagenetic fluids. This resulted in partial dolomitization of the argillaceous sediments and important neomorphism of those sediments and the borders of adjacent lime mud bodies. Substantial compaction took place while the argillaceous material was still uncemented (possibly inhibited by clay minerals (Jenkins, 1974; Noble and Howells, 1974; Scholle et al, 1983; and Wanless, 1979)). This involved differential compaction of the argillaceous interbeds with loss of initial porosity, realignment of clay minerals, and deformation of the burrow fills. A nodular pattern resulted.

Recrystallization - Some fossils show evidence of recrystallization. Some groups, such as the gastropods and cephalopods, with shells probably composed originally of high Mg-calcite (Bathurst, 1975), are much more commonly recrystallized than others and are now composed of coarse mosaic sparry calcite. The timing of the recrystallization is not clear but in view of the evidence of diagenetic fluids migrating through those sediments quite early in their burial history (early cementation), this transformation was also probably early. It would have involved release of Mg^{2+} ions that could have contributed to the partial dolomitization.

Another important recrystallization left mosaic sparite-filled spicules, presumably of lithistid sponges. Its timing is not clear but it would have released a substantial amount of silica considering the abundance of such originally siliceous sponge spicules in these sediments. Silicification is not evident in the Irene Bay Formation but is fairly common in the underlying Thumb Mountain (Kerr, 1976; Morrow and Kerr, 1977) and overlying Allen Bay formations (Mayr, 1974; Morrow and Kerr, 1977). It will be examined further in Chapter 3.

Hydrocarbon migration - Minor hydrocarbon migration is indicated by bitumen in some intergranular and intercrystalline pore spaces in the argillaceous sediments. It did not originate in these sediments as they appear to represent oxygenated environments of deposition. The hydrocarbon migration followed most of the other post-depositional processes, except, perhaps, compaction. Its relationship with compaction is unclear. However, all the diagenetic fluids needed initial porosity for their migration, prior to the filling of those pores by the bitumen. The source of the bitumen could have been basinal equivalents in the Hazen Trough.

2.7 Conclusions

The sediments of the Irene Bay Formation were deposited in an open marine depositional environment, as evident from the varied fauna, termed the "Arctic Ordovician" fauna by Kerr (1968). The various niches and life habits that these organisms represent can only signify a well oxygenated environment with normal salinity. A shallow shelf swept by weak wind and tide-induced currents is thus inferred (Morrow and Kerr, 1977). However, all the studied data indicate a subtidal setting, certainly below fair-weather wave-base and possibly even below storm wave-base. No indications of current activity, such as reorientation of grains or sorting, were observed. The fossils are certainly fragmented but the rudite-size fragments, the very poor sorting, the random orientation of fragments, and the fact that they are mostly in intensely bioturbated argillaceous layers, suggest that biogenic activity caused the fragmentation. With no evidence of bottom transport, the presence of several genera of calcareous algae indicates photic conditions. However, the low concentration of these calcareous algae compared to the Thumb Mountain Formation probably indicates a depth range near the maximum limit of 92 m.

The Irene Bay Formation can easily be identified due to its very distinctive nodular structure and its greenish-grey weathering colour. The colour is evidence of deposition under oxygenated conditions. The nodular structure reflects a combination of depositional and post-depositional processes: alternation of lime mud deposition with mostly argillaceous deposition; early cementation of the lime mud layers; and subsequent mottling and mixing by biogenic activity, intense in the argillaceous layers but only partially interrupting the layering of the lime mudstone beds. The relationships of sediment types and textures to trace fossils representing more than one phase of burrowing provide the important evidence of early cementation of lime mudstones. These in turn had a major control on successive diagenetic processes such as neomorphism, dolomitization, recrystallization, hydrocarbon migration and differential compaction, some of which further enhanced the distinctive nodular character.

Although the Irene Bay facies characterize a broad shelf area, the facies do change gradually toward the northwest toward the Hazen Trough. The proportion of clay to lime mud increases substantially and the sequence thins in that direction. This can be explained either in terms of uniform background clay sedimentation, with the more argillaceous layers representing periodic slow-down in carbonate production, or in terms of cyclical variations in the supply of clays, presumably from the Hazen Trough, with differential compaction responsible for most of the thinning. These is, nevertheless, no evidence of abrupt facies change or of organic buildups or banks that would define a shelf break or margin. The Irene Bay shelf, therefore, is more representative of a ramp than of a rimmed shelf.

The burrows are mostly horizontal, transect particularly the argillaceous layers, and are themselves mostly filled by this clay-lime mud mixture. However, a few organisms managed to burrow through the lime mudstone layers, creating the nodular pattern. Their sharp burrow contacts with the host sediment can only be interpreted as evidence for early cementation of these lime mudstone layers, thus partly explaining their different bioturbation level from the argillaceous layers. Furthermore, the minor lime mud-filled burrows are also found within the argillaceous layers following older, clay-filled burrows even where these older burrows penetrate underlying lime mudstone layers, indicating the preference of these organisms for burrowed sediments over consolidated ones. In addition, not only are fragments of the burrowers' shells more common in the argillaceous layers and burrows, but remains of some suspension feeders are also slightly more abundant in these zones. Most of the suspension feeders, such as crinoids, bryozoans, brachiopods and rugose corals needed a firm substrate to anchor themselves and that is what an early-cemented lime mudstone layer would have provided. At their death, they rested above the lime mud, only to be buried by the clay-dominated sedimentation. This association of early cementation and burrowing formed the basic characters over which other diagenetic processes such as neomorphism, dolomitization, recrystallization and hydrocarbon migration were imprinted. The components of the argillaceous zones were most affected since early cementation restricted the migration of diagenetic fluids to those zones.

Chapter 3

Allen Bay Formation-Read Bay Group undivided and Cape Phillips Formation

3.1 Introduction

This chapter deals with the core of this study, a sequence of Upper Ordovician (Ashgillian) to Upper Silurian (Ludlovian and/or Pridolian) shelf to proximal slope carbonates previously designated Allen Bay Formation-Read Bay Group undivided (Mayr, 1974). Equivalent distal slope and basin margin deposits belong to the Cape Phillips Formation. These deposits are among those considered to have the highest potential for hydrocarbon accumulation in the Franklinian Basin. They are also particularly interesting for the wide variety of depositional environments represented. In contrast to the underlying Irene Bay deposits, representing shallow restricted shelf and ramp leading very gradually to deeper depositional environments, these mostly Silurian deposits represent more distinctly differentiated carbonate-forming environments: an open shallow-to-deep shelf, an abrupt shelf-break, a slope steep enough to generate mass flows, and a generally starved deep basin. Shoreward, a stable platform is transitional between this mobile belt and coastal areas (Kerr, 1976; see Fig. 1.2). These Silurian deposits are also exceptional for their wide distribution in generally parallel belts, southwest to northeast across the Arctic Archipelago, from Melville Island to northern Greenland. Their accumulation coincided generally with the apogee of Hazen Trough development. The Franklinian Basin appears to have been well differentiated into deep basin, slope and unstable shelf (eugeosyncline-miogeosyncline according to Trettin and Balkwill, 1979), but to date only deposits associated with the shelf, slope and shallower part of the basin (miogeosyncline) have been clearly identified and divided consistently stratigraphically. The origin and stratigraphy of the deep basinal (eugeosynclinal) deposits, further to the northwest, are still problematic.

The study area is located in this shelf-to-basin transition (deepening to the northwest) with the sections studied strategically situated through shelf,

shelf margin and slope settings (see Fig. 1.3). Numerous facies and facies changes are represented in different sections and at different stratigraphic levels within sections, reflecting constantly changing and commonly alternating depositional environments.

Due to the transitional geological situation of the study area, the stratigraphy of these mostly Silurian deposits is not simple. The formations individually are distinctive in character and depositional environments represented. The Cape Phillips Formation regionally, for example, is a characteristic sequence of argillaceous and silty lime mudstone (or marlstone) representing distal slope and/or back-reef basin deposition. None of these terms is applicable to typical deposits of the Allen Bay Formation-Read Bay Group as originally defined in shelf areas. However, the lateral transition between the Cape Phillips Formation and the Allen Bay Formation-Read Bay Group undivided is very gradual, and the boundary is difficult to define. Furthermore, the position of the shelf-slope break changed repeatedly during the Silurian in response to progradation or eustasy. In addition, the slope's gradient made mass flow a common depositional process, that often blanketed the whole slope as well as part of the basin. Relatively shallow water carbonates of the Allen Bay Formation-Read Bay Group undivided were carried and deposited in this fashion among the much finer deposits of the Cape Phillips Formation. With very little to indicate the distances travelled by these mass flows, it is very difficult to distinguish proximal from distal slope facies and the boundary between them is necessarily arbitrary.

An additional fundamental problem is the possible stratigraphic subdivision of the Allen Bay Formation-Read Bay Group undivided. Further south, the Allen Bay Formation can be clearly distinguished from the Read Bay Group, particularly where the dolostone and evaporites of the Cape Storm Formation intervene. The Read Bay Group itself can be divided into three formations, for example, on Cornwallis Island (Thorsteinsson, 1980). Several things make this impossible on Svendsen Peninsula. Firstly, during the Late Silurian, slope deposits of the Devon Island Formation encroached over the shelf, and it isn't yet certain how much the Read Bay Group was curtailed by this event. Certainly the group thickens to the south and the east. Secondly, the Cape Storm Formation is unrepresented in the study area.

The designation of a contact between the Allen Bay Formation and Read Bay Group thus remains an important and pressing problem in this area. Essentially the problem is that most of the facies studied here are unrepresented further south, and thus a direct correlation between areas requires biostratigraphic or other time-significant criteria. In this area, the equivalent of the Allen Bay Formation-Read Bay Group contact may be represented, not by a facies change, but by a change in the frequency of the facies changes, or by changes in the relative proportions of deposit types believed to represent particular facies.

The first part of this chapter describes the distribution, thickness, age and contacts of these formations. The following part describes the wide variety of rock types represented in this very thick sequence in the study area, and includes their fossil content. For ease of representation, the stratigraphic sequence is divided into eight parts (as shown in the section correlation diagram, Fig. 3.1), interpreted as representing major events in the depositional history, that were recorded in both basinal and shelf sections. The rocks in these divisions are categorized as part of the Cape Phillips Formation or of the Allen Bay Formation-Read Bay Group undivided.

An important marker unit that approximates the position of the Ordovician-Silurian boundary in these rocks, is a distinctive bitumen-rich dolostone and marlstone containing substantial nodular and bedded chert. The marker is of lowermost Llandoverian age. It is one of the important criteria used to correlate these sections regionally (Fig. 3-1) and is described in more detail later in the chapter (e.g. section 3.3.5.1). The descriptions of rock types and common associations, are followed by interpretations of depositional and post-depositional history. Excellent exposure and detailed microfacies analysis, made it possible to document and to relate in sequence a diverse array of depositional and diagenetic processes, especially those involving slope deposits.

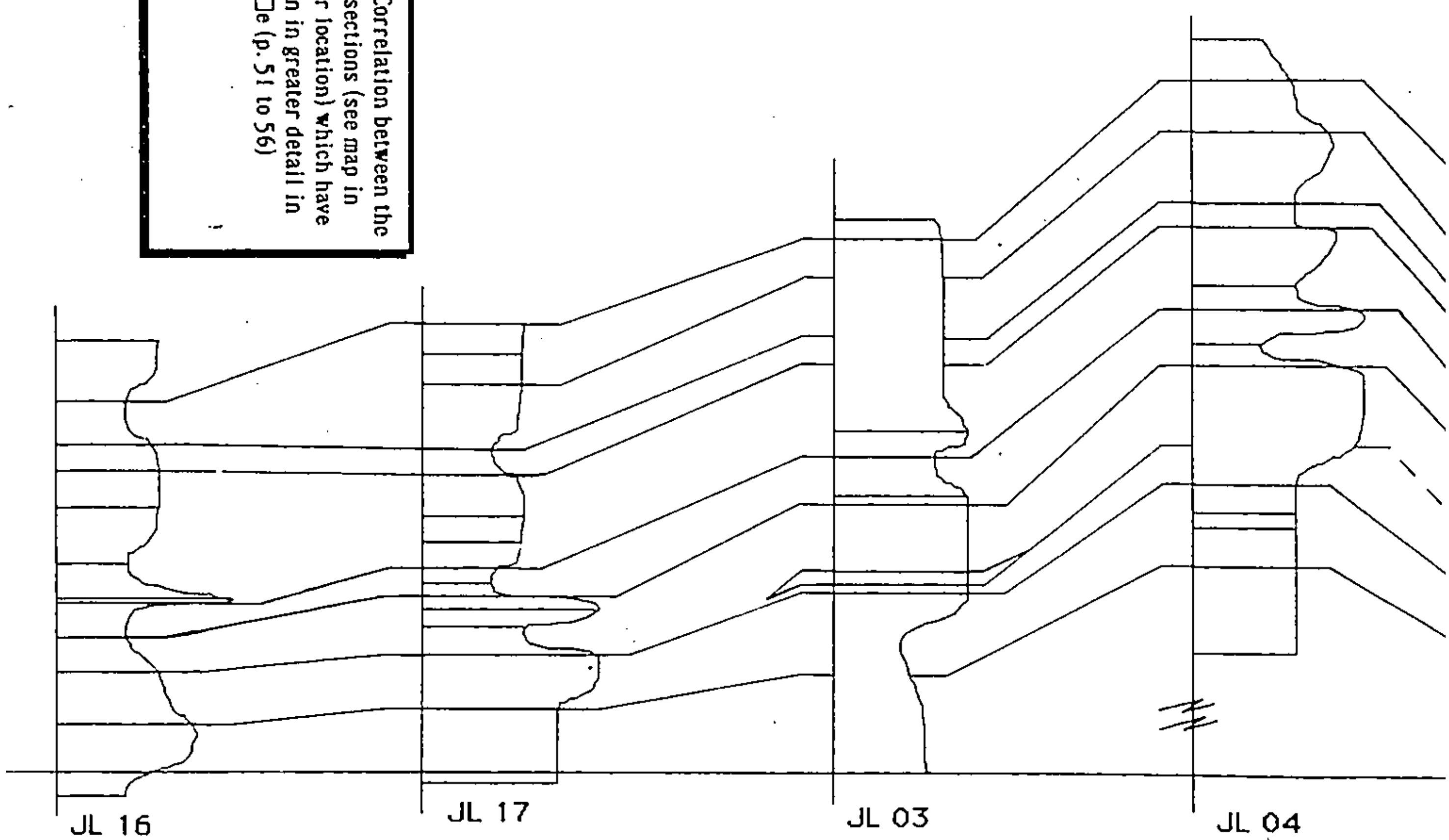
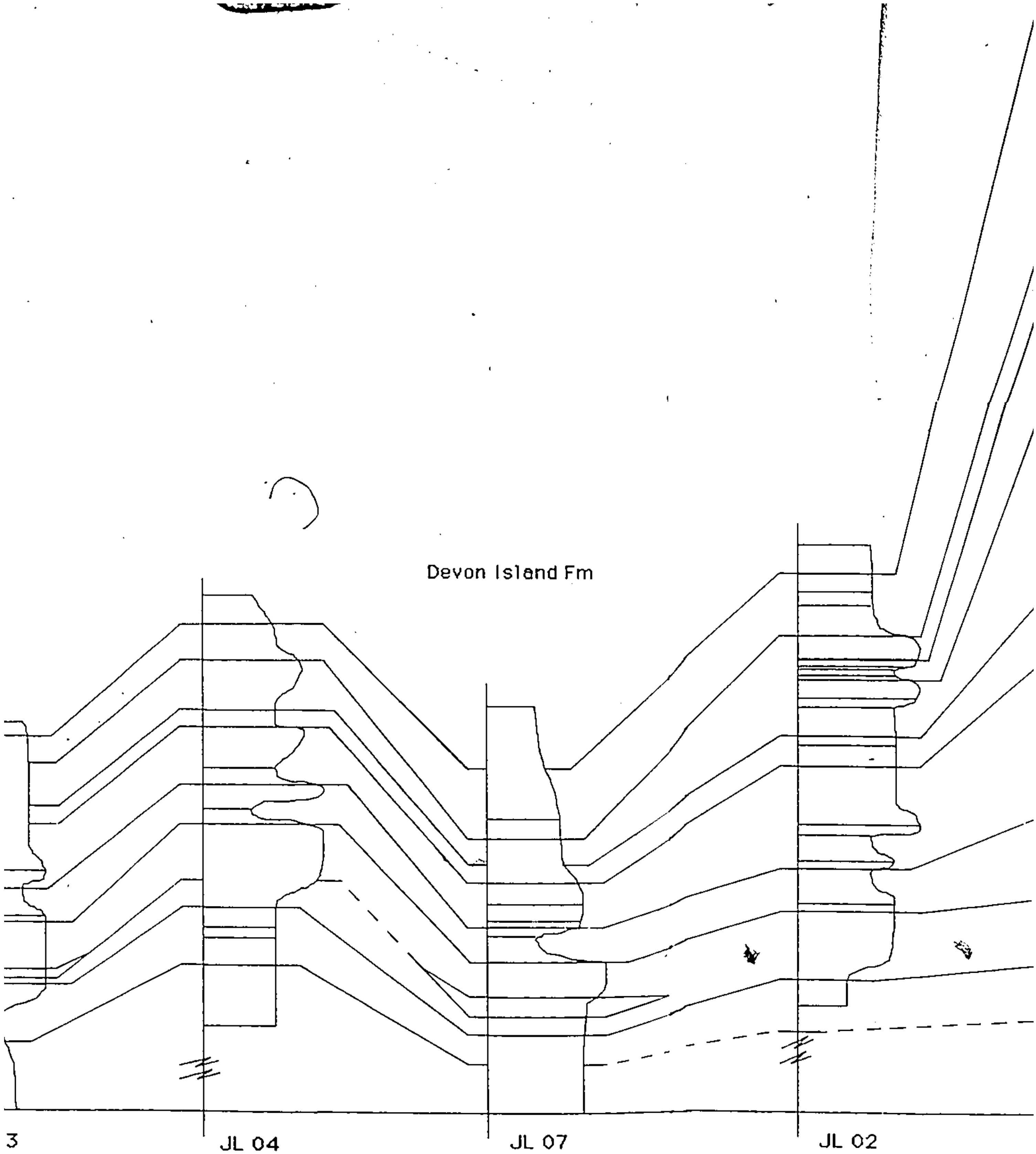
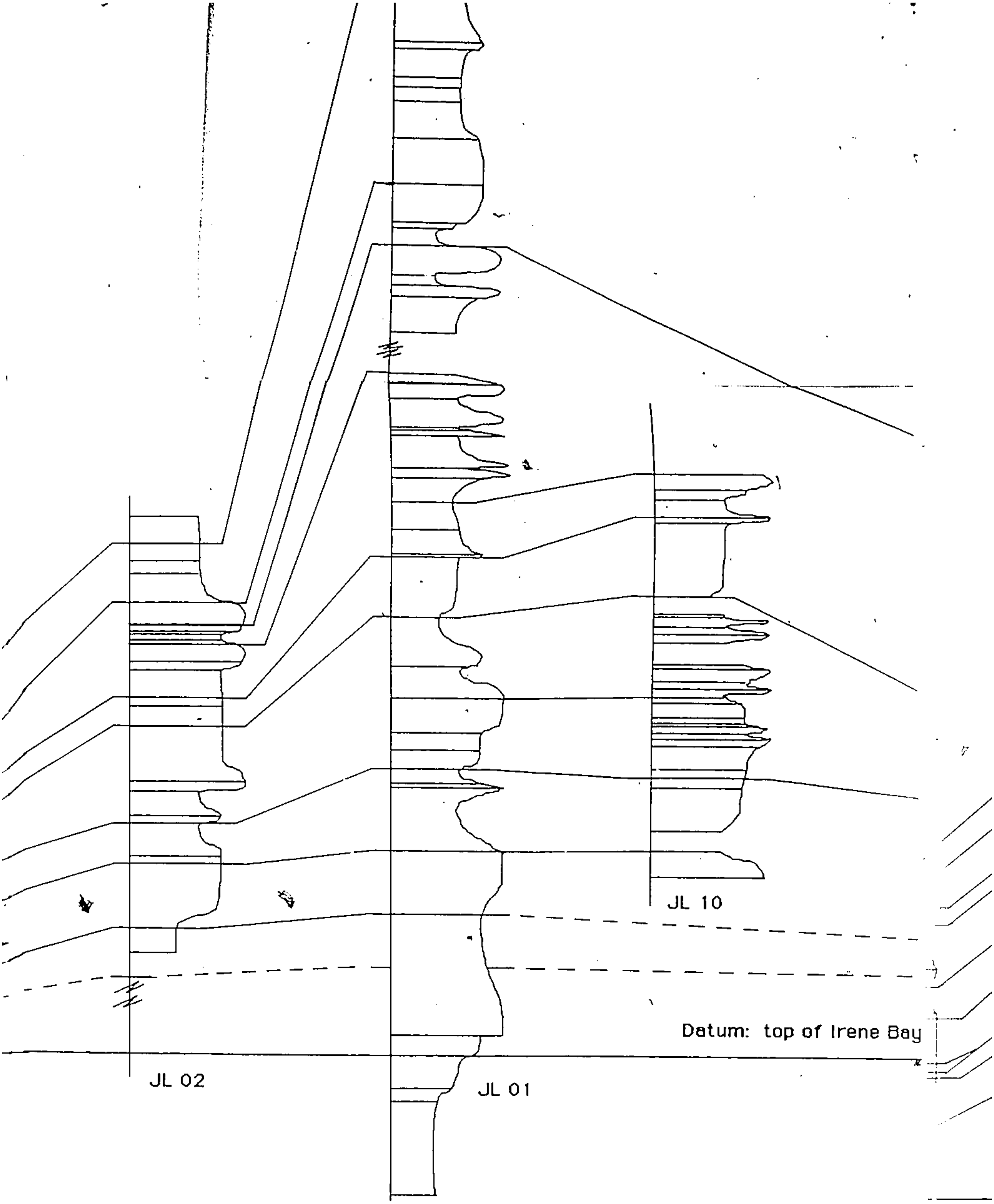
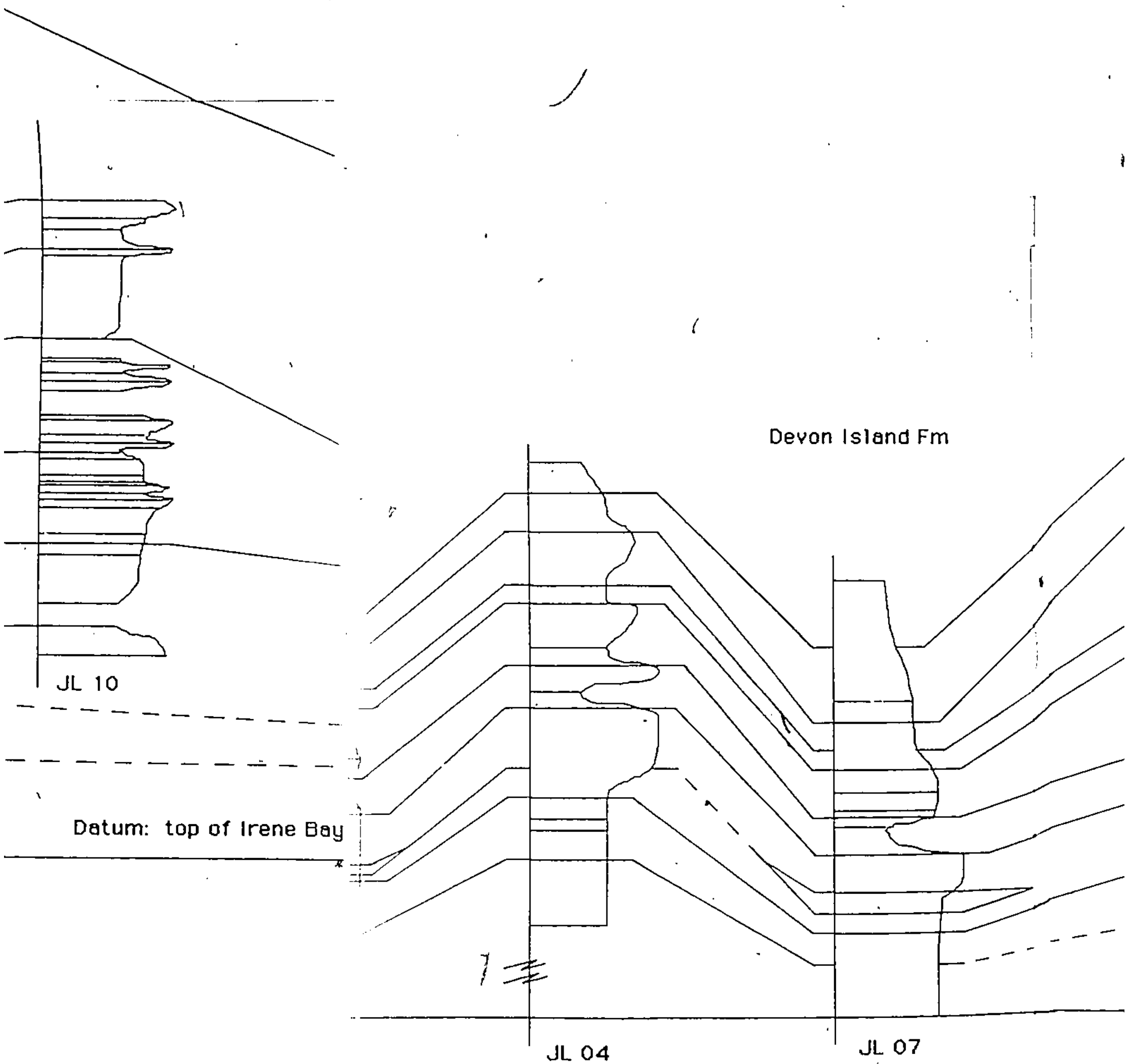
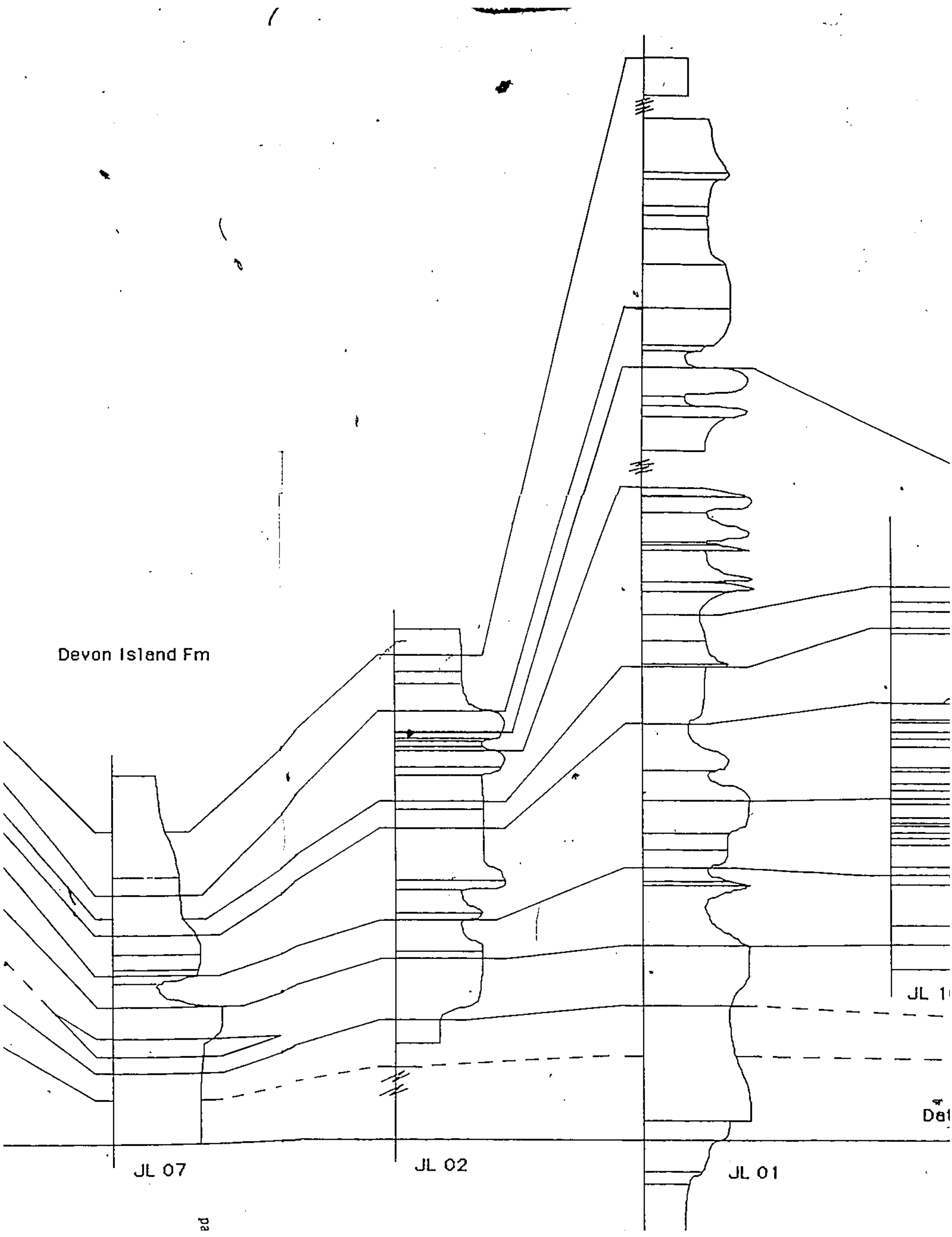


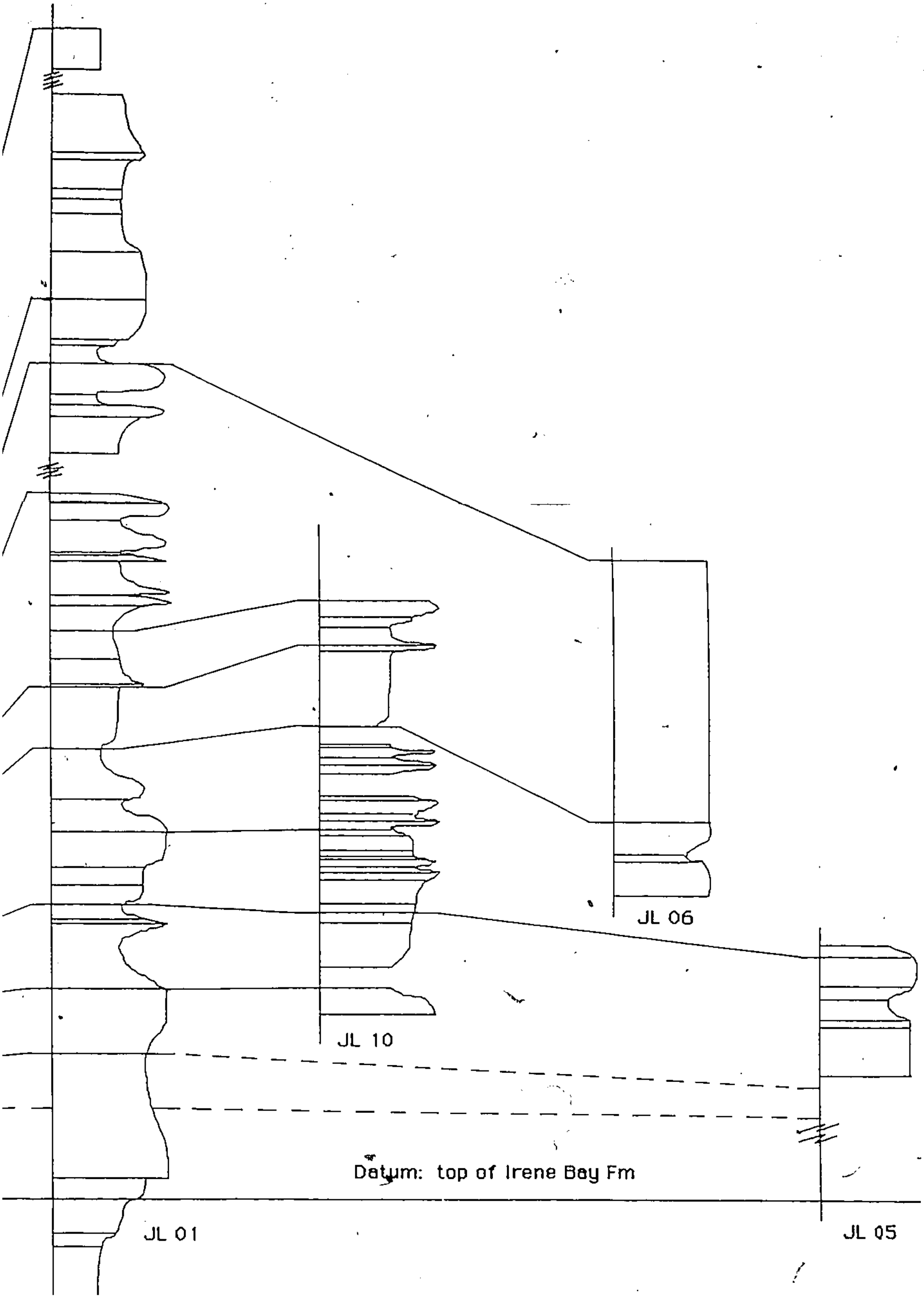
Fig. 3.1 Correlation between the measured sections (see map in Fig. 1.5 for location) which have been drawn in greater detail in Figs 3.4 a–e (p. 51 to 56)


Devon Island Fm
3
JL 04
JL 07
JL 02


JL 10
Datum: top of Irene Bay
JL 02
JL 01


Devon Island Fm
JL 10
Datum: top of Irene Bay
JL 04
JL 07


Devon Island Fm
JL 07
JL 02
JL 01
JL 1
Da


JL 06
JL 10
JL 05
Datum: top of Irene Bay Fm
JL 01

3.2 Distribution (age & contacts)

3.2.1 Age

The ages of these formations vary greatly from area to area. This is a reflection of the numerous facies changes along the shelf-slope-basin transition. These facies represent such dynamic depositional environments in a relatively narrow belt, especially at the shelf-margin, that facies changes are not synchronous between areas since they are in response to local anomalies and paleogeographic positioning. It is generally accepted though, that the complete sequence, whether it includes shelf deposits of the Allen Bay Formation and the Read Bay Group or distal slope and marginal basin deposits of the Cape Phillips Formation, covers a maximum time range from the Late Ordovician (Ashgillian) to the earliest Devonian (Gedinnian) (Thorsteinsson, 1980). It is also generally accepted that the lower contact between the Irene Bay Formation and the Allen Bay Formation or rarely, the Cape Phillips Formation (northwest of the Boothia Uplift; Kerr, 1974) is synchronous everywhere (Trettin, 1979). It is situated in the early Ashgillian, close to the Caradocian-Ashgillian boundary. However, if the whole sequence is divided into individual formations the contacts at higher levels are highly diachronous (Trettin, 1979). In the study area, the upper contact of the Cape Phillips Formation or of the Allen Bay Formation-Read Bay Group undivided is believed to be early Pridolian (Mayr, 1974).

Figure 3.2 represents the simplified stratigraphy at different locations in the Arctic Archipelago along the shelf-basin transition (see Fig. 1.1 for the location of this transitional belt). In areas where the deposits are old enough to incorporate the Irene Bay Formation, the top of this formation is represented as synchronous throughout. Although the Allen Bay Formation and Read Bay Group are generally equivalent to the Cape Phillips Formation, apart from minor regional variations, their upper contacts are highly diachronous (see Fig. 3.3 and section 4.2 of chapter 4). The overlying stratigraphy is much more highly diverse and complex, with almost unique nomenclature in each region. This clearly reflects the individual response of those areas to increasing activity of the Caledonian Orogeny during the Early Devonian (Thorsteinsson, 1970; Trettin, 1972; and Kerr, 1977).

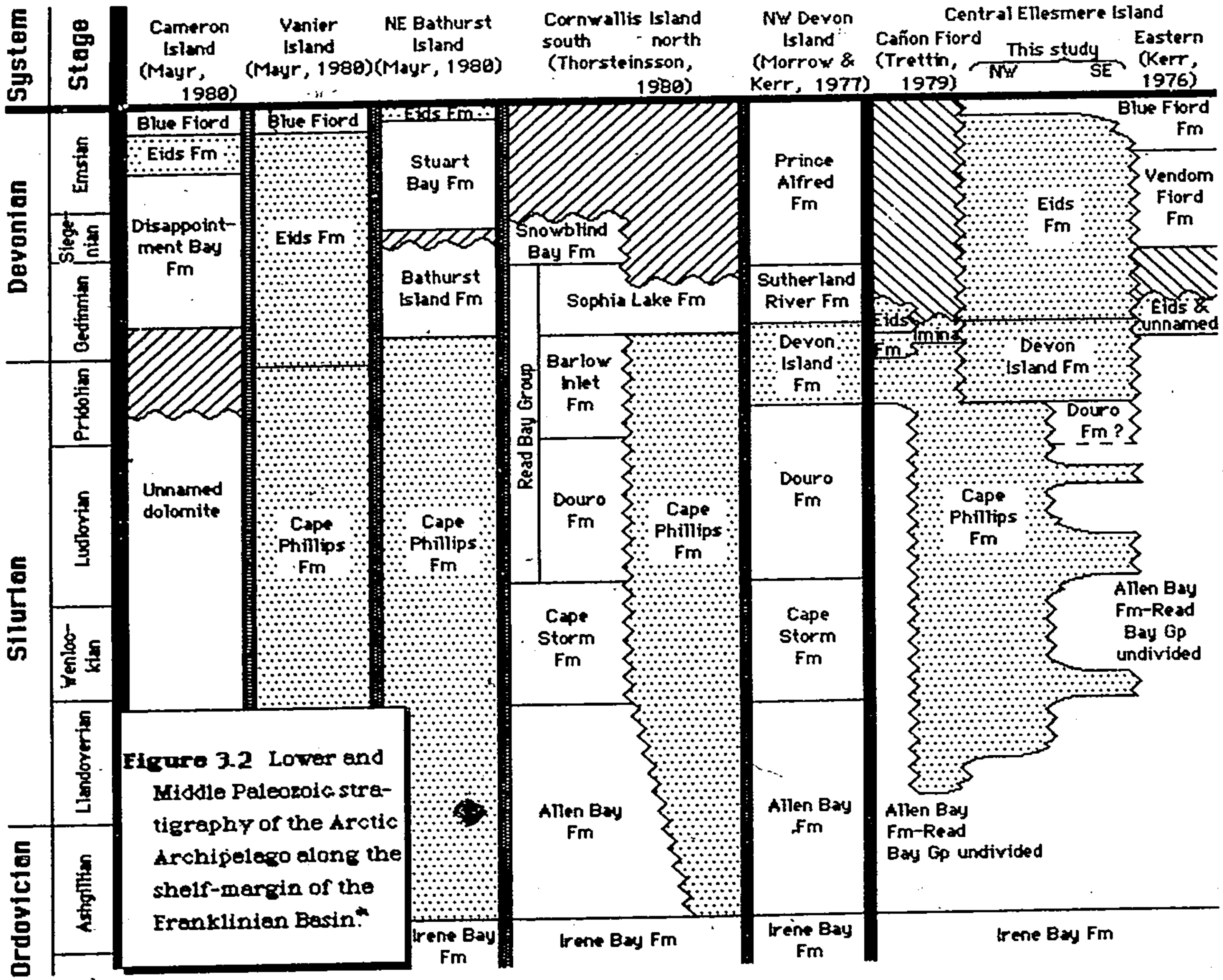


Figure 3.2 Lower and Middle Paleozoic stratigraphy of the Arctic Archipelago along the shelf-margin of the Franklinian Basin.*

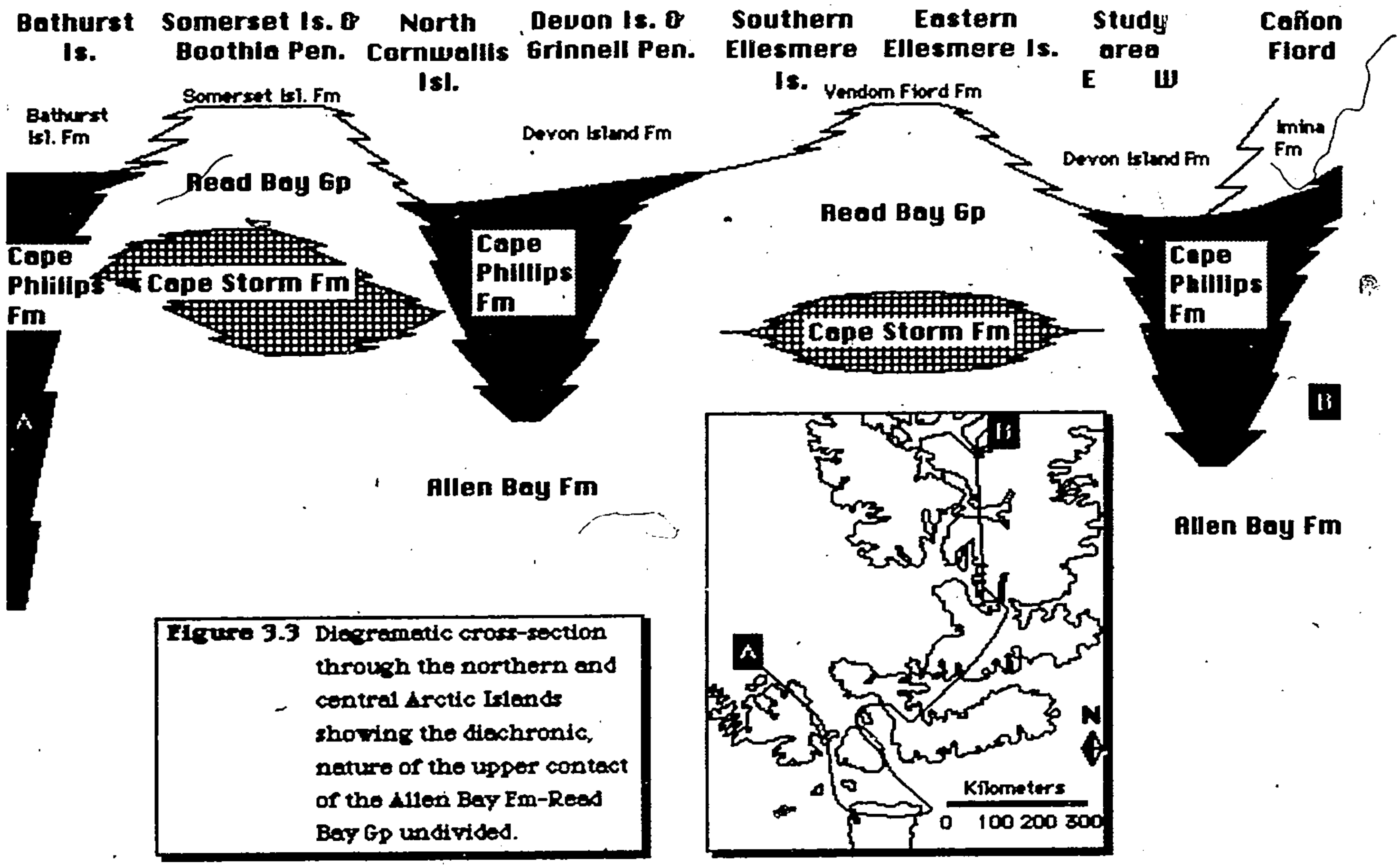


Figure 3.3 Diagramatic cross-section through the northern and central Arctic Islands showing the diachronic, nature of the upper contact of the Allen Bay Fm-Read Bay Gp undivided.

The diachronous nature of the tops of the Cape Phillips Formation and Read Bay Group is due to several factors. Gradual lateral shift of adjacent facies belts on a paleoslope (e.g. related to progradation or eustasy), is commonly extensive enough to entail a formational or stratigraphical change. Diachroneity can also be due to regional unconformities such as the ones associated with the Boothia and Bache Peninsula (or Inglefield, Smith and Okulitch, 1987) uplifts, or simply to the absence of diagnostic characteristics, and difficulty in differentiating closely related formations.

Even in the relatively small area of central Ellesmere Island, several stratigraphic models have been invoked. In part this is due to legitimate differences in depositional environments such as the progressively shallower environments represented on the stable platform of eastern and southern Ellesmere Island (Kerr, 1976; McGill, 1974; Packard, 1986). In part this is because in some areas (e.g. Cañon Fiord area (Kerr, 1976; and Trettin, 1979)) genetically and/or lithogically closely related formations, such as the Cape Phillips, Devon Island and Imina formations, lack the diagnostic characters that allow them to be differentiated elsewhere, such as in the study area. Nevertheless, the geological history interpreted for the study area is closely related to and will be compared to the known histories in these other areas of Ellesmere Island.

3.2.2 Contacts

In the study area, and along most of the unstable shelf-to-basin transitional zone defined earlier, the lower contact is usually the boundary between the Irene Bay and Allen Bay formations. As this lower contact is generally easy to identify, and is apparently synchronous (Trettin, 1979), it is an important marker in the Lower Paleozoic sequence of the Arctic. In areas surrounding the study area, the contact is reported to be sharp but conformable (Kerr, 1976 (Central Ellesmere Island); Trettin, 1979 (Cañon Fiord); Mayr, 1978 (Southern Ellesmere Island); Morrow & Kerr, 1977 (Grinnell Peninsula and northwest Devon Island); and Thorsteinsson, 1958, Mayr, 1974 (Cornwallis Island)). In some areas, as on the southern coast of Ellesmere Island, where the contact is not obvious, the appearance of the trilobite *Pseudogygites latimarginatus* was used (Thorsteinsson, 1958) to mark the beginning of the Allen Bay Formation. Table 3-1 summarizes the major changes across this contact.

In general, the lower contact shows a sharp colour change when observed from a distance. Closer inspection in the study area shows, however, that the change is gradational over a few centimeters to approximately one meter. The transitional zone is thicker if it accommodates all the lithological changes indicated in Table 3-1. Generally, within the study area, this zone thickens to the southeast or shoreward. Fragments of *Pseudogygites* sp. have been found at some localities, but only after careful sampling of beds already presumed to be basal Allen Bay Formation. In the study area, the distribution of this fossil is not consistent or widespread enough to be a reliable field indicator but is useful for confirmation where careful sampling is done.

Table 3.1 Irene Bay Formation/Allen Bay Formation Contact

- *Nature of contact*: gradational over 1 to 5 m
- *Colour changes, upward*:
  Weathering - from light greenish grey to darker brownish grey
  Fresh - from medium grey to dark grey
- *Weathering style changes, upward*: from a rubbly, less resistant lithology, to a more massive and blocky lithology
- *Lithological changes, upward*: from a nodular limestone (discrete nodules) with greater amount of argillaceous and dolomitic lime mud matrix, to a nodular limestone (bedded) with lesser amount of argillaceous and dolomitic lime mud matrix
- *Faunal changes, upward*: from a more diverse fauna with orthoconic cephalopods especially conspicuous, to a less diverse fauna containing the trilobite *Pseudogygites*

An additional complication in the thicker southwestern sections is that, between about 5 and 13 meters above the contact, there is a limestone sequence with the same weathering color as the Irene Bay deposits. It is slightly less argillaceous than the Irene Bay deposits, and is also slightly more resistant to erosion, more like Allen Bay strata. Nevertheless, where the Irene Bay Formation is covered, the Irene Bay-Allen Bay contact could erroneously be placed above this anomalous sequence. Since this sequence does not exist everywhere, and because it is inconsistently expressed mainly as a weathering phenomenon, it is rarely use as a secondary marker.

Still close to the shelf-basin transitional zone, in areas on the northwestern side of the Boothia Uplift, such as on Bathurst Island, deposits of the Cape Phillips Formation directly overlie the Irene Bay Formation. This contact on Bathurst Island, is sharp but conformable (Mayr, 1980).

The Allen Bay Formation is overlain by either the Read Bay Group or the Cape Phillips Formation. Both the Read Bay Group (Douro Formation) and the Cape Phillips Formation are overlain by the Devon Island Formation and this upper contact of the sequence examined in this chapter is almost certainly early Pridolian. Table 3-2 presents the characteristics of these four contacts in greater detail.

Table 3.2 Internal and Upper Contacts of Allen Bay Formation-Read Bay Group Undivided and Cape Phillips Formation

Allen Bay Fm/Cape Phillips Fm Contact

- *Nature of contact* : from gradational to abrupt but conformable
- *Colour changes, upward* :
  Weathering - from light grey to dark grey
  Fresh - from medium grey to dark grey
- *Weathering style changes, upward* : from massive to less resistant and flaggy deposits

- *Lithological and stratification changes* : from more massive deposits of the Allen Bay Fm (wide variety dependent on stratigraphical position of contact (see text), but in basinal NW sections, the last Allen Bay beds are very bituminous and cherty dolostone) to planar-bedded and argillaceous lime mudstone
- *Fossils* : the Cape Phillips Fm is graptolitic

Allen Bay Fm/Read Bay Gp Contact

- *Nature of contact* : very gradational over several metres
- *Colour changes, upward* :
  Weathering - from light grey to dark brownish grey
  Fresh - both are dark grey
- *Weathering style changes, upward* : from massive and planar to less resistant mottled strata
- *Lithological and stratification changes, upward* : from planar bedded lime mudstone or massive coarsely crystalline dolostone to mottled, though still bedded, lime mudstone
- *Fossils* : More varied fauna in the Read Bay Gp which is also marked by *Atrypoidea arcticus*

Cape Phillips Fm/Devon Island Fm contact

- *Nature of contact* : gradational over several metres
- *Colour changes, upward* :
  Weathering - from dark grey to light grey or yellow and rare orange dolostone beds
  Fresh - both are dark grey
- *Weathering style changes* : both flaggy, though the Devon Island Fm is slightly more massive
- *Lithological and stratification changes, upward* : from planar-bedded bituminous and argillaceous lime mudstone to less bituminous, more dolomitic and silty argillaceous lime mudstone, also planar bedded

- *Index fossil* : *Monograptus bohemicus* in the Devon Island Fm

Read Bay Gp/Devon Island Fm contact

- *Nature of contact* : abrupt but conformable
- *Colour changes, upward* :

  Weathering - from medium grey to light grey or yellow and rare orange dolostone

  Fresh - from medium brownish grey to dark grey
- *Weathering style changes, upward* : from more resistant and rubbly strata to flaggy beds
- *Lithological and stratification changes, upward* : from nodular, dolomitic limestone to bituminous, dolomitic and silty lime mudstone
- *Index fossil* : *Monograptus bohemicus* in the Devon Island Fm

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In general, the contact with the base of the Devon Island Formation is thought to be synchronous throughout the study area (Mayr, 1973), and in its vicinity (Morrow & Kerr, 1977). The age relationship at the Allen Bay Formation-Read Bay Group contact is less sure as the contact has been examined in only one section (JL 01). The contact was not previously defined either in the study area (Mayr, 1973), or elsewhere on Ellesmere Island (Kerr, 1976). For this reason the entire sequence in this area on Ellesmere Island is for convenience referred to as the Allen Bay Formation-Read Bay Group undivided.

The Allen Bay Formation-Cape Phillips Formation contact is highly diachronous, especially at inflection points (e.g. between shelf and upper slope) such as in the study area (Mayr, 1973) and on Cornwallis Island (Jones & Dixon, 1980; Mayr, 1980). In the study area, this contact varies in age from Llandoverian to Ludlovian. Though the Cape Phillips Formation is the lateral equivalent of the Allen Bay Formation and Read Bay Group, it generally prograded over the former, but not during deposition of the upper Ludlovian Read Bay Group (Poey, 1982). This progradation resulted in the

Cape Phillips Formation overlying progressively younger Allen Bay rocks toward the southeast. The contact in northwestern sections is abrupt (Table 3-2), but toward the southeast is clearly gradational and generally drawn where shales, marlstones or argillaceous limestone become predominant in upward succession (Kerr, 1976). The diachronous nature of this contact was also noted by Trettin (1979) further north in the Cañon Fiord area.

Although Table 3-2 enumerates criteria that distinguish the Cape Phillips-Devon Island formational contact, the changes listed are generally difficult to recognize. Though they are lithologically slightly different, field appearances do not reflect this and both appear dark, recessive and flaggy. Most conspicuous, although rare in proportion to other lithologies, are the orange-weathering dolostone beds in the Devon Island Formation. This contact is equivalent to the Cape Phillips-Imina formational contact in the Cañon Fiord area (Trettin, 1979). Trettin positioned it at the base of the lowermost sandstone bed or where sandstone becomes dominant. Usually the term Imina Formation is used for deposits of the axial portion of the Hazen Trough, equivalent in age to all the formations examined in this study, while deposits of the Devon Island Formation are generally interpreted as slope deposits (Morrow & Kerr, 1977). Both formations are similar lithologically.

The Read Bay Group-Devon Island Formation contact is very easy to distinguish and abrupt. All criteria necessary for its identification are listed in Table 3-2. Among the studied sections, only section JL 01 has this contact exposed, and there only on the other side of the ridge from where most of JL 01 was measured. The only other studied exposure of the contact was on Grinnell Peninsula (Morrow & Kerr, 1977), where the contact is very similar in age and nature and designated the Douro-Devon Island formational contact.

3.3 Lithological Descriptions

This section on lithological descriptions will be followed by sections discussing particular lithological associations and their significance, and the distribution of the various facies according to depositional environments.

The rock types presented below are grouped in five main categories: Limestone predominates and is the most diverse lithology in sections in the southeastern part of the study area. The first main lithological category includes all those limestones preserving bedding character and other primary depositional features. These have been divided mainly according to the classification of Dunham (1962) as modified by Embry and Klovan (1972). Within these divisions, the principal rock types (as distinguished on drawn sections—see Figs 3.4 a→e) are identified further by their principal components (e.g. bioclasts, peloids, intraclasts, and types of cement and/or matrix) using Folk's classification (1962) in conjunction with Dunham's. Some of these principal rock types are further distinguished on structure (e.g. lamination, bedding), or on minor components (e.g. silt, dolomite). Because of the importance and diversity of limestone conglomerates and breccias, these rocks are assessed separately in this first category.

Limestones that have been substantially altered post-depositionally form a second major category. Types are distinguished by a variety of forms of mottling and nodule development. Dolostones as the third main category are also post-depositionally altered. Designated types are distinguished on structure (massive, bedded or nodular character) and composition (fossil and bitumen content).

Argillaceous rocks, as the fourth category, include individual rock types distinguished by composition (calcareous, dolomitic, bituminous, silty, graptolitic contents). Finally, the fifth category includes siliceous deposits, occurring as nodular or bedded chert, or as various forms of replacement of other rock types.

The stratigraphic distributions of the principal rock types are represented on the drawn sections (Figs 3-4 a→e). More detailed stratigraphic sections are on file in the Department of Geology, University of Ottawa.

Legend for Figures 3-4 a→e

(the profile shown in the sections reflect their relative erosional profile)

Abbreviations:

ls ogle = limestone conglomerate	pokst = packstone	mdst = mudstone
grnst = grainstone	rdst = rudstone	mrlst = marlstone
	wokst = wackestone	

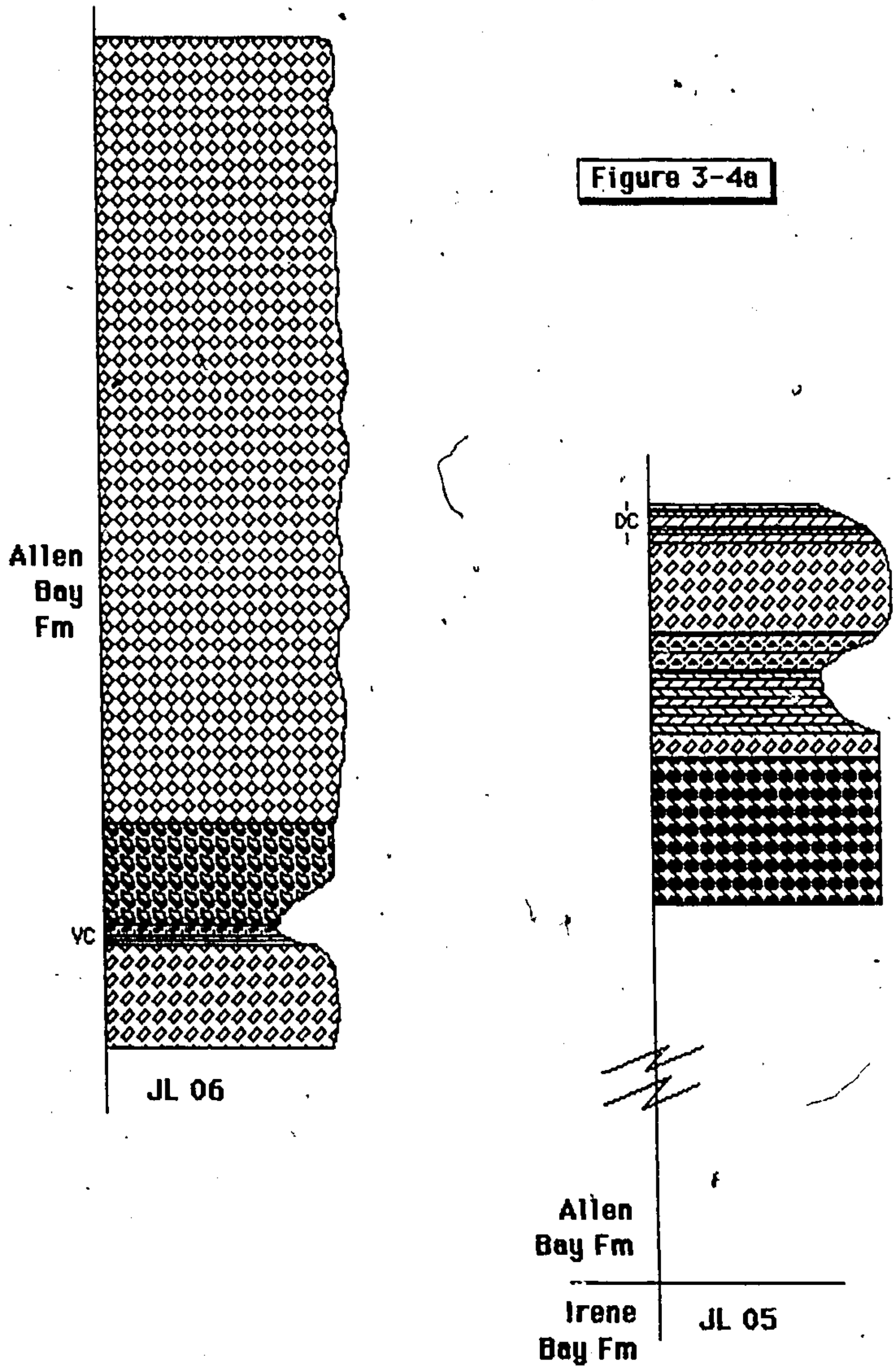
Indicates missing section of uncertain thickness

Vertical scale: 1 : 2500
1 cm = 25 m

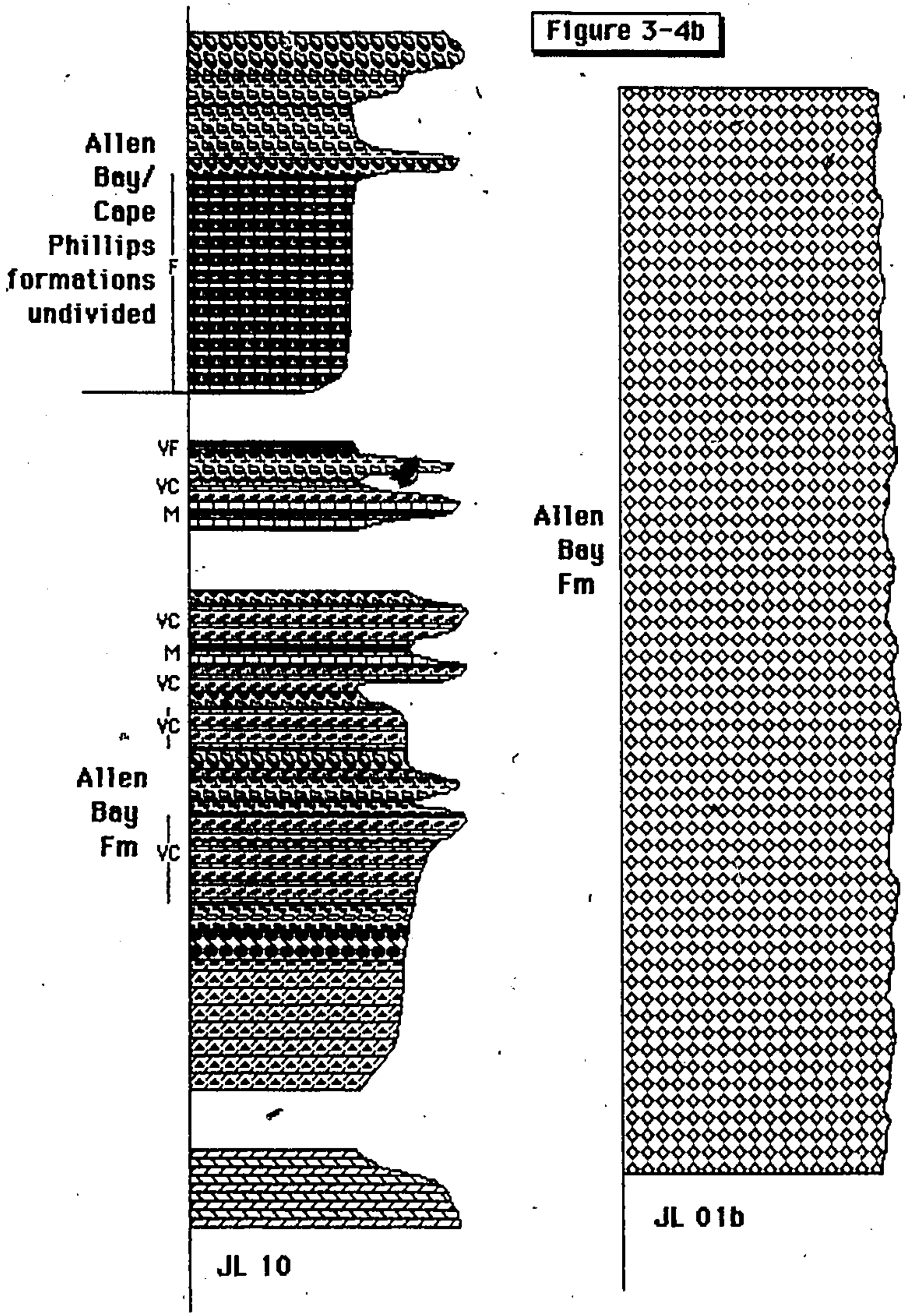
- Argillaceous nodular limestone
- Non-argillaceous nodular limestone
- Ribbon limestone
- Interbedded oligomictic flat-pebble conglomerate and ribbon limestone
- Coarse polymictic limestone conglomerate
- Polymictic limestone conglomerate inter-bedded with marlstone
- Coarse bioclastic pokst, grnst or rdst
- Bedded lime mudstone
- VC — Very coarse grade fining-upward sequences (ls ogle to mrlst)
- C — Coarse grade fining-upward sequences (grnst/rdst to mrlst)
- M — Medium grade fining-upward sequences (pokst to mrlst)
- F — Fine grade fining-upward sequences (wokst to mrlst)
- VF — Very fine grade fining-upward sequences (lime mdst-mrlst couplets)
- Calcareous marlstone
- Dolomitic marlstone
- Interbedded very finely crystalline dolostone and lime mudstone
- Nodular dolostone
- Massive dolostone
- Dolomitic breccias
- Bedded dolostone
- DC — Dolomitized coarse grade fining-upward sequences
- DF — Dolomitized fine grade fining-upward sequences
- Interbedded dolostone and chert
- Interbedded lime mudstone and chert

• Allen Bay Fm-Read Bay Gp undivided and Cape Phillips Fm •

Figure 3-4a

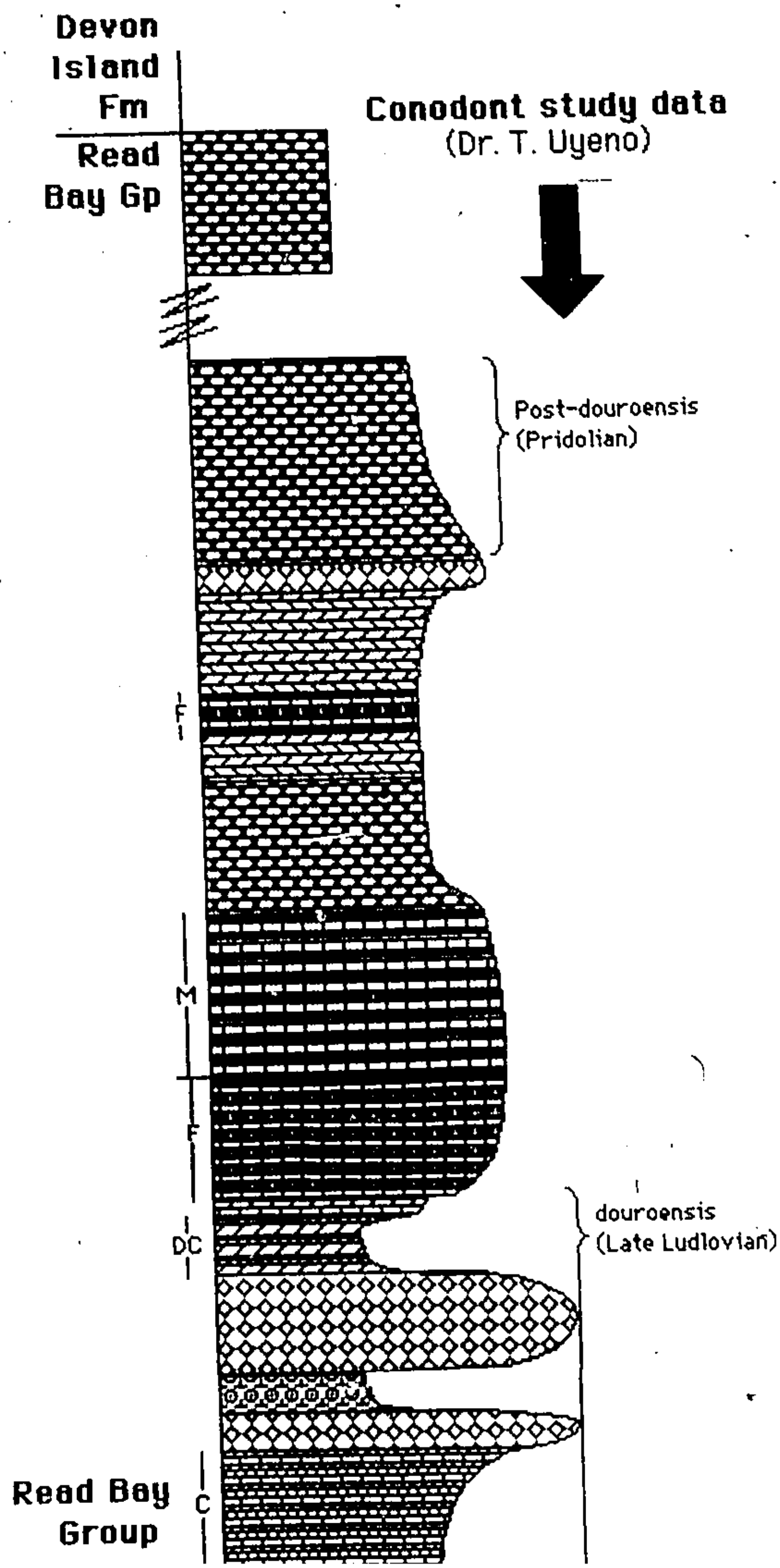


• Allen Bay Fm-Read Bay Gp undivided and Cape Phillips Fm •



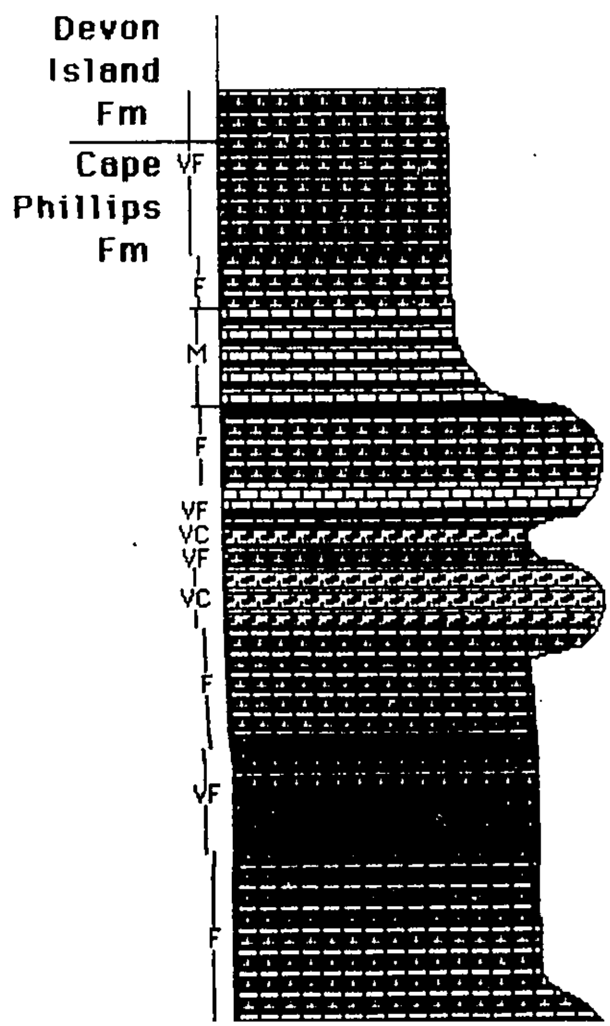
• Allen Bay Fm-Read Bay Gp undivided and Cape Phillips Fm •

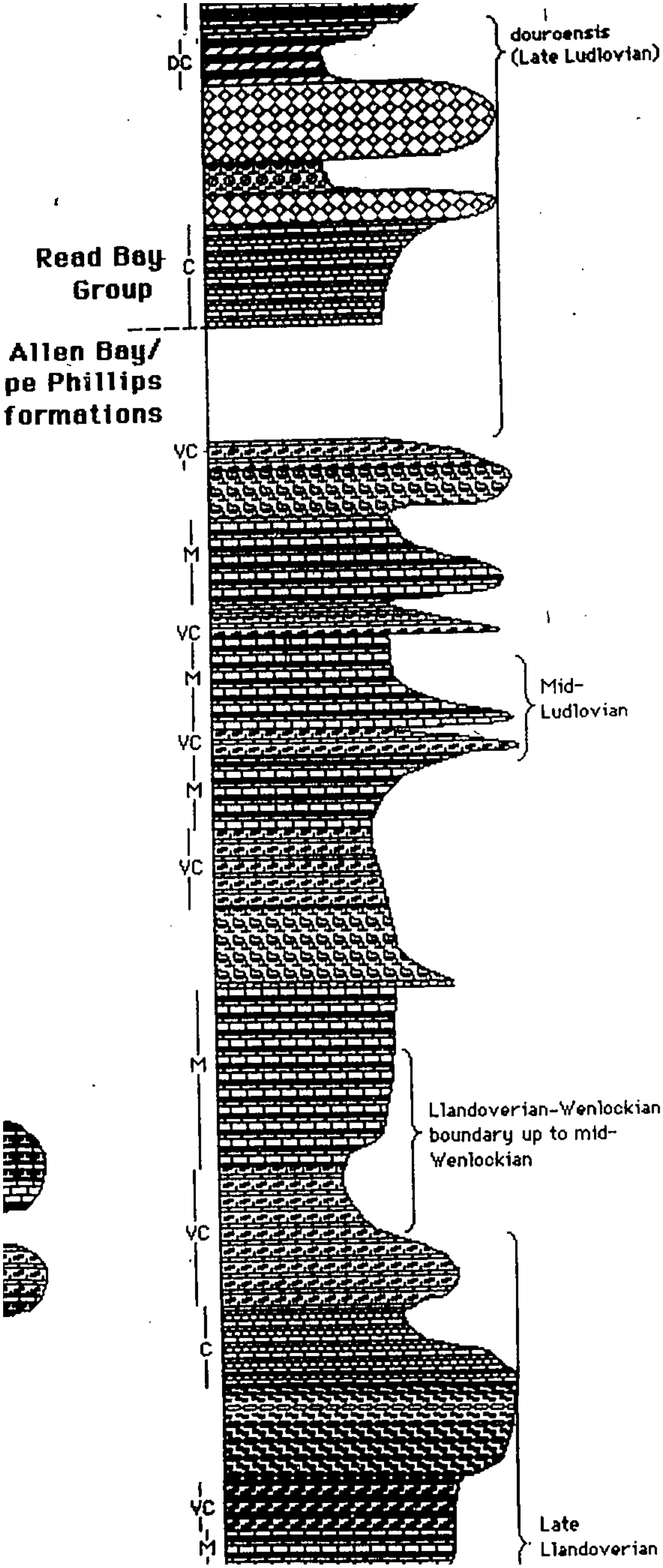
FIGURE 3.4c

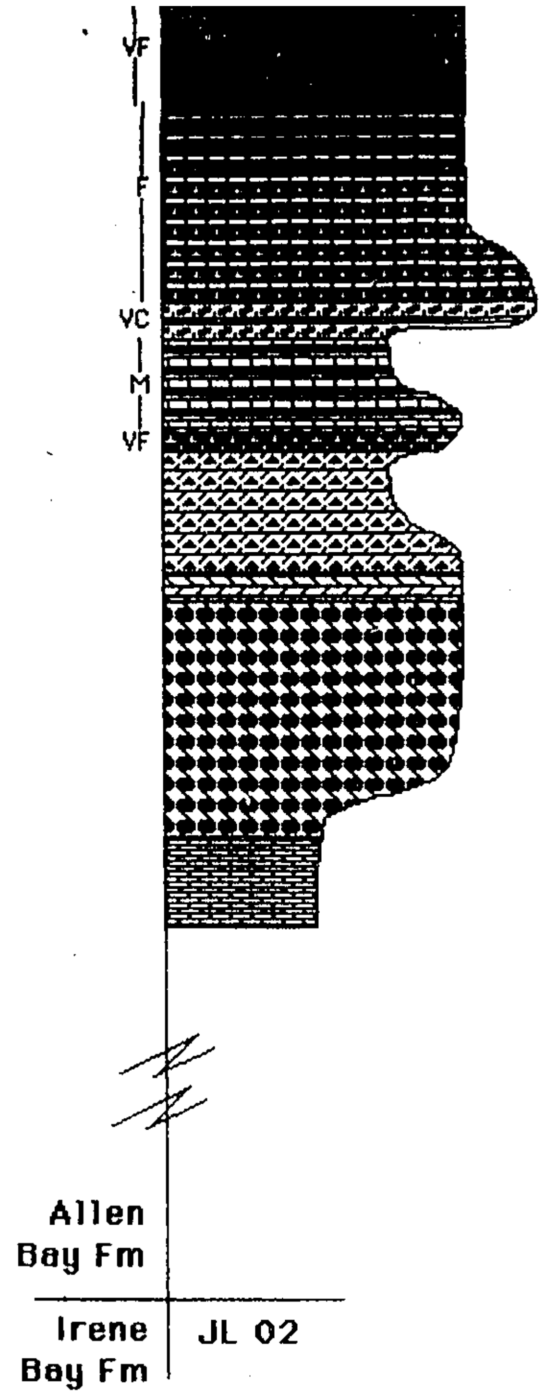

Devon
Island
Fm
Read
Bay Gp
Conodont study data
(Dr. T. Uyeno)
Post-douroensis
(Pridolian)
F
M
F
DC
douroensis
(Late Ludlovian)
Read Bay
Group
C

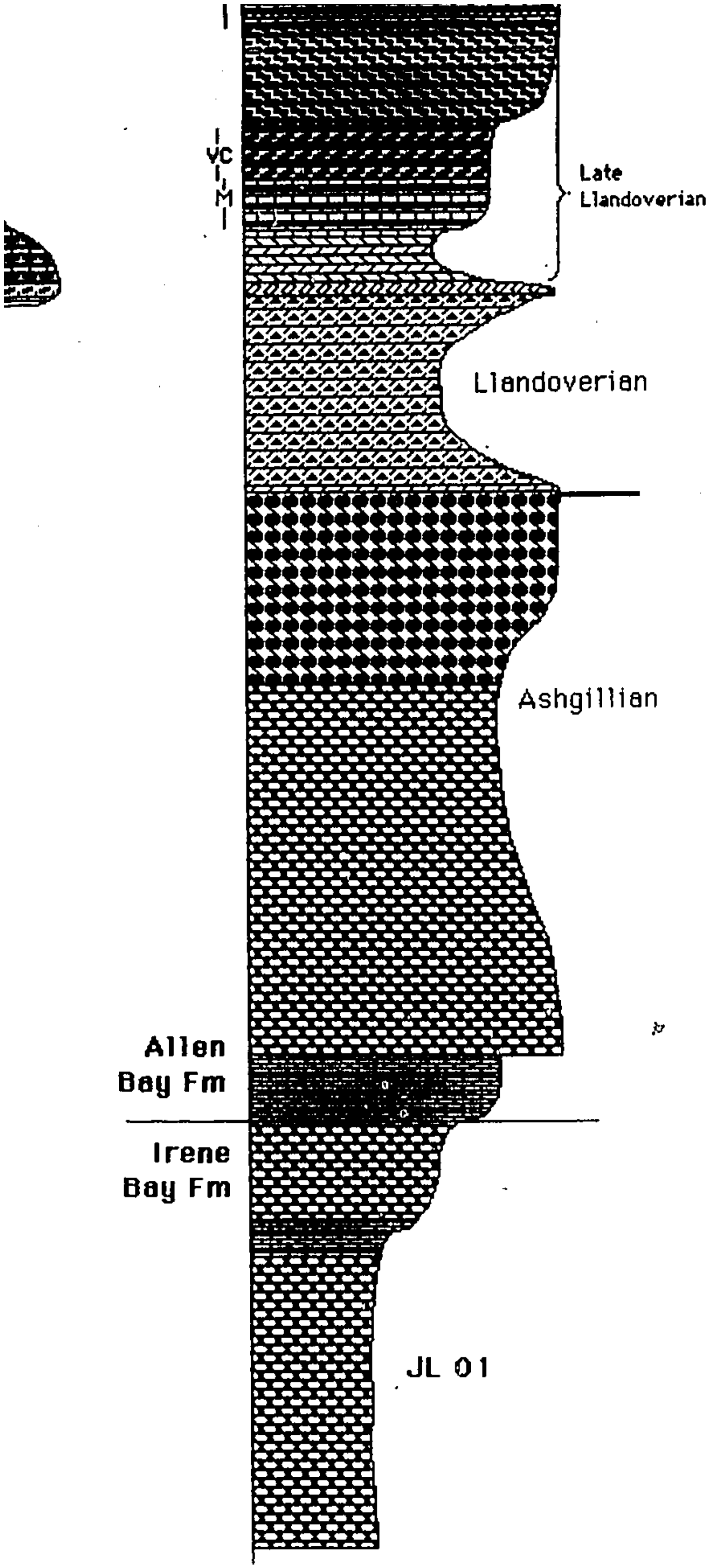
Read
Gr

Allen
Cape Phil
format




douroensis
(Late Ludlovian)
DC
C
Read Bay
Group
Allen Bay/
pe Phillips
formations
VC
M
VC
M
Mid-
Ludlovian
VC
M
VC
M
Llandoverian-Wenlockian
boundary up to mid-
Wenlockian
VC
C
VC
M
Late
Llandoverian


VF
F
VC
M
VF
Allen
Bay Fm
Irene
Bay Fm
JL 02



VC
M
Late Llandoverian
Llandoverian
Ashgillian
Allen Bay Fm
Irene Bay Fm
JL 01

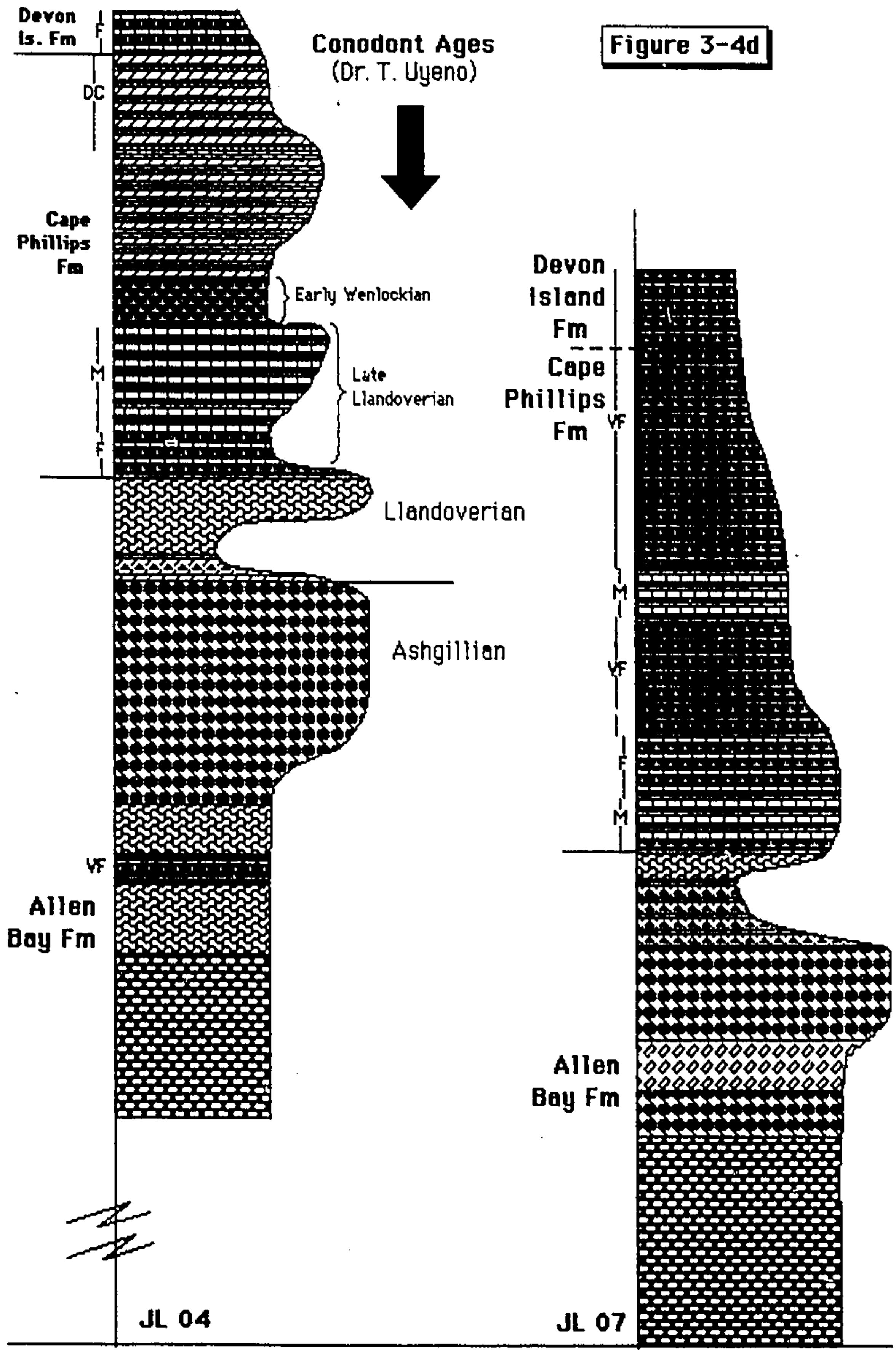
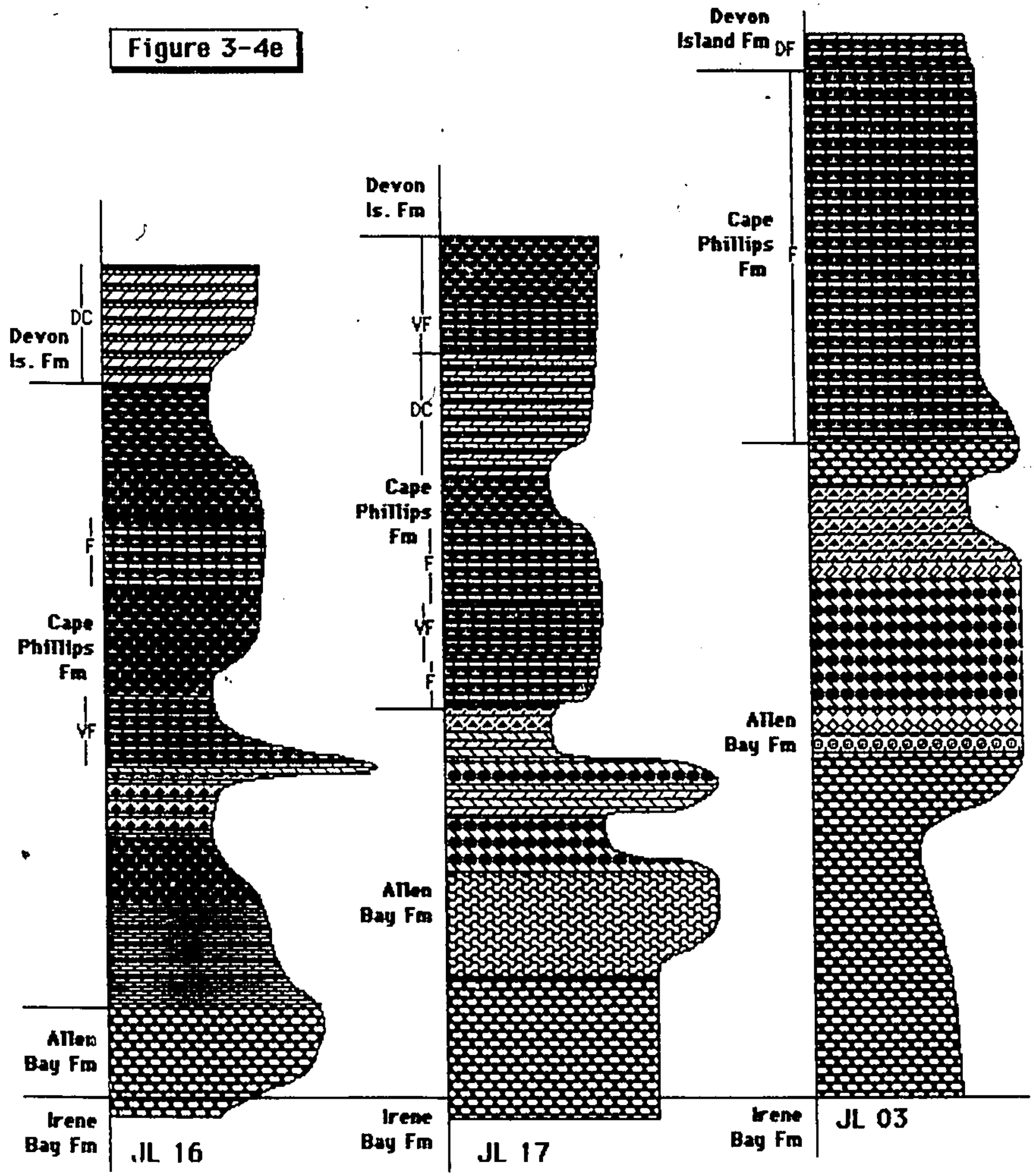


Figure 3-4d
Conodont Ages
(Dr. T. Uyeno)
Devon Is. Fm
F
DC
Cape Phillips Fm
Early Wenlockian
M
F
Late Llandoverian
Llandoverian
Ashgillian
VF
Allen Bay Fm
JL 04
Devon Island Fm
Cape Phillips Fm
VF
M
VF
F
M
Allen Bay Fm
JL 07



• Allen Bay Fm-Read Bay Gp undivided and Cape Phillips Fm •

3.3.1 Limestones with mainly primary characters

3.3.1.1 Lime mudstones

The major components of lime mudstone are not only micrite, but also microspar and pseudospar. Despite this, mudstone has been retained as a useful general term in this study. Granular components are not abundant enough to make prefixes (such as in "biomicritic") useful. The mudstones have been subdivided into *laminated*, or *massive*, *bedded*, or *massive*, *unbedded* types; the first commonly overlies the second, which in turn generally overlies the third, in fining-upward sequences. It is possible to distinguish varieties of mudstone that are silty, dolomitic, bituminous, siliceous, argillaceous (described as marlstone on subsequent pages), or pyritic, based on minor proportions of these components, or mottled (described below as mottled limestone), or combinations of these varieties. Silica occurs in these rocks as minor cryptocrystalline chert nodules, or selectively silicified grains, especially allochems. Pyritization of graptolites in these rocks is extremely rare. Dolomite as a replacement of micrite varies greatly in amount. The bitumen content also varies greatly, and is sometimes the cause of faint lamination in the mudstone beds where the bitumen occurs as thin "wispy" or "filamentous" lenses. However, the bitumen is more commonly intergranular, imparting a dark colour to the micritic sediment.

Some mudstone beds contain discoidal calcareous concretions. They vary from bodies a few centimeters in diameter (measured along bedding; see Fig. 3.5) and appearing as swellings of the associated thin-bedded lime mudstone, up to one meter wide "cannon balls" (see Plate no. 7-1). These concretions, especially the "cannon balls", are more common in marlstone beds, or where lime mudstone alternates with marlstone. They are micritic, have very gradational contacts with the lime mudstone deposits of similar compcsition, and the smaller concretions, especially, have an apparent ball-and-pillow structure. However, their concretionary origin is indicated by their lack of fossils such as occur in surrounding sediments, and by their concentric growth rings (see Fig. 3.5).

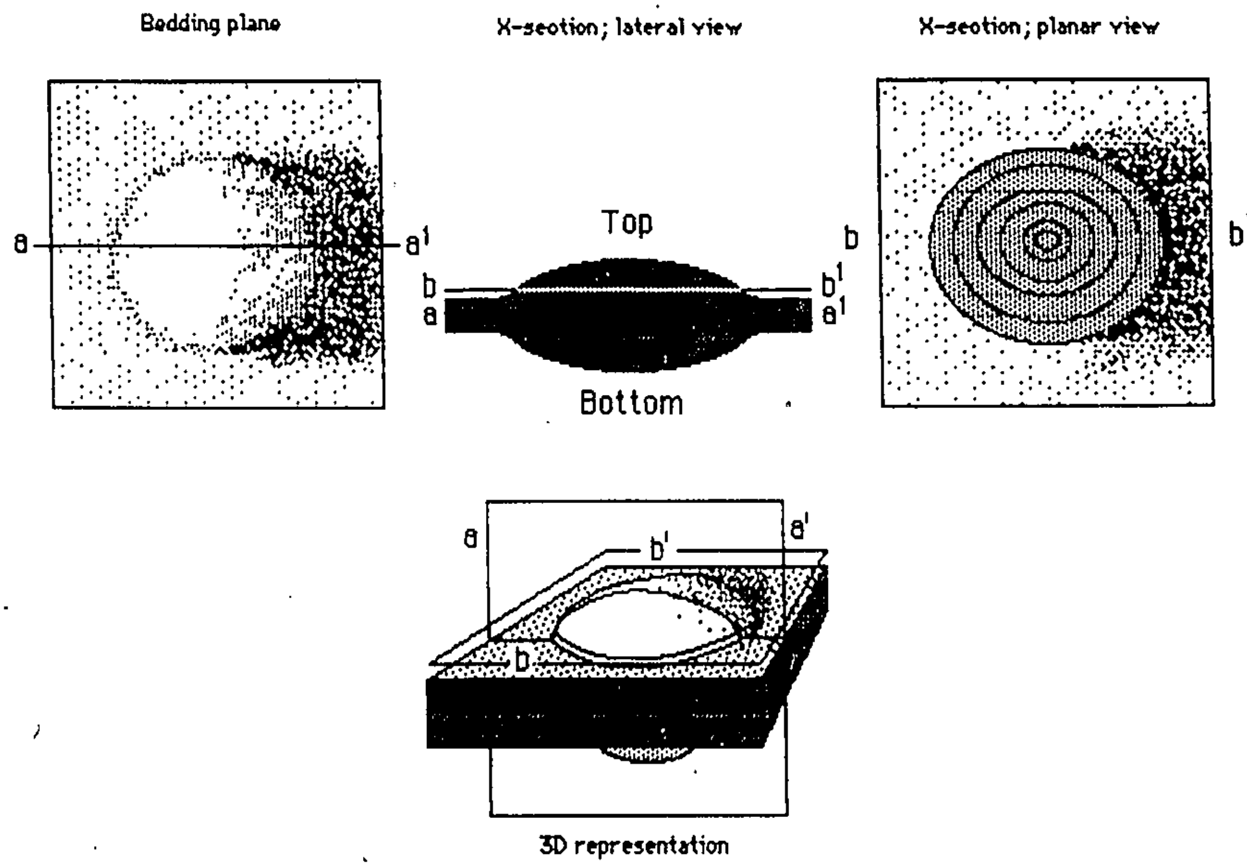


Figure 3.5 Diagrams of limestone concretions in the Llandoverian and Wenlockian of the Allen Bay Fm (section JL 01, unit 35).

These lime mudstones rarely contain disseminated, silt-sized quartz and feldspar. These siliciclastic grains are commonly concentrated in thin lenses that give a laminated appearance to these beds (see description below, under "Laminated lime mudstone"). They form generally less than two percent of the total components, rarely up to ten percent. The grains are generally silt sized; finer grains are also common but coarser grains are extremely rare. They are generally well sorted, and subangular to angular. These rarely occurring lime mudstones, with significant amounts of siliciclastic grains, have been termed "silty".

3.3.1.1.a Laminated lime mudstone

The laminated character does not imply an argillaceous content (clay-rich and clay-poor laminae) in these rocks, as is commonly the case. The lamination instead usually reflects the alternation of thicker lime mudstone layers with thin layers or very elongate minute lenses that are grain-rich (see also section 4.3.2). Most of these grains are siliciclastic, but some are bioclastic (usually crinoidal) or peloidal. The interstices in these grain-supported layers are filled by a very fine grained matrix, usually micrite, or, less commonly, by sparite cement.

The lamination, secondly, can be due to variation in bitumen content, where the bitumen occurs mostly in very thin, elongate lenses (generally about 10 × 100 μm). These bitumen-rich lenses form up to 30 percent of these strata, though generally less (approximately 10 percent). The interlaminated sediments, generally micritic, can lack bitumen.

The lamination can also reflect rare allochem-rich (or -richer) laminae in a mostly lime mudstone deposit. They apparently represent fining-upward sequences, but the grains are commonly too small to be distinguishable in the field. Some of these layers exhibit sharp lower contacts. Gradual up-sequence disappearance of these grain-supported lenses and/or laminae has been observed. These deposits will be examined further in subsequent description of fining-upward sequences.

A final type of laminated lime mudstone is composed mostly of homogeneous micrite with enigmatic lenses of pseudospar of various sizes (generally less than 100 μm thick) and length-to-width ratios. The lenses generally form less than 10 percent of the deposit. Disseminated monocrystalline pseudosparitic "calcispheres" varying from 50 to 100 μm in diameter are also associated in the micrite.

The bitumen content of all the laminated mudstones varies but is generally higher than in other deposits examined in this study, with the possible exception of some dolostones and some cherty sediments.

3.3.1.1.b Massive, bedded lime mudstone

Massive, bedded lime mudstone is quite distinct from the laminated lime mudstone, especially in its more massive field appearance and resistant weathering character. The fine calcareous grains that constitute this rock are homogeneously distributed. This homogeneity and absence of platy or argillaceous minerals results in very distinctive conchoidal fracturing. Very few fossils were observed in this rock type, but it does rarely contain spectacularly well preserved brachiopods, cephalopods, graptolites, some colonial corals (favositids, heliolitids), bryozoans, stromatoporoids, and fish plates. This lime mudstone is often associated with other rock types in fining-upward sequences. In these situations it grades upward, as the upper part of a massive bed, from a bioclastic sediment, or even a limestone conglomerate. The massive, bedded lime mudstone passes more abruptly upward into laminated lime mudstone, generally in a separate bed.

These deposits are generally bituminous and mostly composed of very fine micrite. Very little bioturbation disturbs its bedding or very faint lamination. Most of the evidence for bioturbation occurs in the overlying laminated lime mudstone (bioturbation is still not important enough to destroy the laminations) of these fining-upward sequences.

3.3.1.1.c Massive, unbedded lime mudstone (in limestone blocks)

The Ludlovian portions of sections JL 01 and JL 10 include large allochthonous blocks up to 5 m in diameter. Most are boundstone and will be described in section 3.3.1.5. However, approximately 10 % of the blocks in the younger strata are lime mudstone quite different from any mudstone described above. Firstly, they are very massive, homogeneous, unbedded, and commonly conchoidally fractured. Secondly, they are very light in colour and contain no bitumen, in marked contrast to the dark marlstone and lime mudstone. Thirdly, allochems are absent, except for extremely rare halysitid colonies.

They occur not only as large individual blocks embedded in laminated sediments, but also as small blocks and intraclasts within massive debris flows or olistostromes consisting mostly of boundstone (for definition see Embry and Klovan, 1972) blocks and intraclasts in a richly bioclastic matrix.

They are therefore easily distinguishable. Their association with bioclast-rich deposits presumably indicates a common origin. This mudstone is one of the most rarely occurring lithologies in the study area.

At first glance, these light coloured lime mudstone blocks, commonly with a sucrosic weathered texture, appear to be dolomitic. However, detailed petrographic examination shows that they are very finely cryptocrystalline to pseudosparitic limestones lacking dolomite. They contain no intergranular, pore-filling or grain-coating organic matter, in contrast to the previously described massive, homogeneous, conchoidally fractured and allochem-poor lime mudstone.

3.3.1.2 Wackestones and floatstones

Wackestones, according to Dunham (1962) are mud- supported limestones containing over ten percent grains. In floatstones the grains are rudite size. In this study wackestone occurs mostly in fining- upward sequences and rarely as an independent deposit. It is described further in subsequent descriptions of associations and especially fining-upward sequences. In all types the matrix is or was micritic and only the main grain types change from one type to another. *Bioclastic*, *peloidal* and *intraclastic* types of wackestone or floatstone (for definition see Embry and Klovan, 1972) are distinguished on the following pages. Their matrices can be micritic, microsparitic or pseudosparitic, and are usually a mixture.

In the field, wackestones are difficult to identify as such for several reasons, but especially because most are part of fining-upward sequences. In these sequences the upward reduction in grain content defines the transition from packstone to wackestone. However, in general this gradation is also accompanied by fining upward, and therefore it is common for the bioclasts to be too small to be identified in the field. Locally, even fine packstone could be mistaken for lime mudstone. In addition, if the main grain components are small micritic peloids floating in micritic matrix, they are very difficult to distinguish.

3.3.1.2.a Bioclastic wackestone

Crinoidal fragments usually predominate as grains in wackestone. Exceptionally, trilobite fragments (pygidia of one species) dominate the granular fraction, while in other examples, brachiopods, ostracods or colonial corals predominate. In some deposits, elongate fossil fragments (mostly brachiopods, ostracods and trilobites) oriented parallel to bedding give a laminated structure to these otherwise massive beds.

3.3.1.2.b Peloidal wackestone

Peloidal wackestone is common and peloids are common as well in other types of deposits, such as bioclastic wackestone. The peloids are consistently micritic, even where the original lime mud matrix has been completely neomorphically transformed to microspar and/or pseudospar. Even where this matrix has been completely dolomitized or silicified, the peloids commonly have kept their original composition. The peloids vary from 50 to 200 μm in diameter, though 70 μm is about the mean.

In fining-upward sequences, peloidal wackestone commonly overlies bioclastic wackestone with a very gradual transition, and peloids are commonly an integral part of the bioclastic layer as well.

3.3.1.2.c Intraclastic wackestone - calcarenite group

Deposits referred to here as intraclastic wackestone are extremely rare, principally because of the classification scheme used in this study. Most deposits dominated by intraclasts, including those matrix-supported, have been classified separately as types of limestone conglomerate or breccia to facilitate the depositional interpretation of this genetically and volumetrically very important group of rocks. Their unique mode of formation warranted a separate grouping in this chapter. Also, where the intraclasts are so few that the rock is not considered a conglomerate, they are generally secondary in proportion to bioclastic grains in the granular fraction of wackestone or packstone. Only rarely does a limestone conglomerate grade into a lime mudstone, and thus disseminated intraclasts are seen to float in a lime mudstone matrix, without an intervening bioclast-dominated transitional zone. What are referred to as intraclastic wackestone here are rocks in which the intraclasts are too small to be distinguished in hand specimens. They appear as granular deposits, with grains easily mistaken for peloids without petrographic examination.

3.3.1.3 Packstones and rudstones

Packstones and rudstones have grain-support and at least ten percent matrix. In rudstone most of the grains are of rudite size (Embry and Klovan, 1972). Packstones commonly occur as the bottom beds in fining-upward sequences.

These massive and resistant beds can be divided into three major groups distinguished by the principal grain components, i.e. bioclasts, peloids or intraclasts. In addition, depending on whether micrite (and its neomorphic equivalents) or cement predominates in intergranular voids, each of these groups can also be described in terms of being micritic or sparitic following Folk's terminology (1962), though these are not always easy characters to determine in the field, and are therefore only used with petrologic samples.

Packstones with large lithoclasts identifiable in the field, will be described subsequently under "limestone conglomerates and breccias". The intraclast-dominated packstone described here is a deposit with sand-sized intraclasts, averaging only 100 - 200 μm in diameter, and therefore too fine to qualify as a conglomerate or breccia. A fine granular texture and good sorting are evident in hand specimens, but the origin of this granular texture is not apparent without careful petrographic examination.

3.3.1.3.a Bioclastic packstone

Most of the packstones in the Allen Bay Formation-Read Bay Group undivided, including those in fining-upward sequences, are bioclastic. Two groups are recognized depending on the matrix-to-cement ratio: biomicritic and biosparitic packstone. Commonly, the bioclasts belong to one particular fossil group such as the crinoids, trilobites, branching bryozoans, or brachiopods (comparable to the coquinas of the grainstones section), or alternatively include several groups, e.g. colonial corals, calcareous algae and stromatoporoids, diagnostic of particular environments of deposition.

Most of the packstones are biomicritic, and those that can be considered biomicritic rudstone are rare. Some packstones are so fine grained that they can be identified as such only through petrographic examination. The matrix varies from micrite to pseudosparite, presumably all originally lime mud.

Normally, the bioclasts are barely in contact. However, some units show fitted or condensed textures representing dissolution at grain-to-grain contacts prior to cementation.

The grains are generally randomly oriented, only occasionally parallel to bedding. Alignment on bedding planes was rarely seen. One example was of branched bryozoans closely packed and more or less parallel on a bedding plane. Another exceptionally elegant example is shown in unit JL 01-61, where orthoconic cephalopods and pentamerid brachiopods are aligned, and taper in the same direction, indicating paleocurrent direction.

In biosparitic packstone, calcite spar cement predominates in intergranular areas, although micrite matrix constitutes at least ten percent. In most examples, sparite is abundant throughout and especially in sheltered pore spaces. Small pockets of sparite are also randomly distributed throughout the deposit. In other deposits, especially those with abundant elongate rudite-size allochems, such as brachiopod fragments, the lime mudstone matrix fills only sheltered areas, such as underneath concave-downward brachiopod shells. The latter can be termed biosparitic rudstone and is proportionally more abundant in the biosparitic packstone group than biomicritic rudstone is in the biomicritic packstone group. In brachiopod-rich deposits where matrix or sparite predominates in sheltered areas, the shells or fragments are commonly oriented in convex-up positions, parallel to bedding.

The cements are mostly calcitic, and in the form of single generation, isopachous mosaic crystals, although some deposits have a first generation of very fine dogtooth crystals as a very thin coating (10 to 20 μm) around bioclasts such as brachiopod shells. In richly crinoidal deposits, the cement is commonly in syntaxial overgrowths. As in the biomicritic packstones, some grains show fitted or condensed textures, reflecting partial dissolution prior to cementation.

Even though the intergranular areas are cement-dominated, the presence of "patches" of micrite matrix as well as micritic inclusions in cement crystals gives a dirty appearance to these rocks in outcrop, similar to the appearance of biomicritic packstones. It is thus quite difficult to distinguish between biosparitic and biomicritic packstones in the field, and only laboratory examination can help in making the distinction.

3.3.1.3.b Peloidal packstone

Peloids mostly occur in fining-upward sequences, usually initially mixed with bioclasts and, as peloid-dominated deposits only above the bioclast-rich bottom deposit. However, most commonly the peloid-dominated deposits lack grain-support, and are peloid- dominated wackestone rather than packstone. Exceptions will be examined below.

The peloids have very sharp contacts with the matrix or cement, and are subspherical. Like those in the mudstones and wackestones their average size is 70 to 80 µm, although some are up to 500 µm in diameter. The peloids are micritic and very little altered, diagenetically.

Pelmicritic packstone, a grain-supported deposit with micrite matrix predominating over spar cement, is present at the base of some fining-upward sequences, instead of bioclastic or intraclastic deposits. Pelsparitic packstone, a grain-supported deposit with at least 10 percent matrix but with cement predominating in interstices, occurs only rarely. It typically has fewer allochems than pelmicritic packstone, and the peloids are well sorted. Micritic matrix present tends to fill spaces between more closely crowded peloids.

3.3.1.3.c Intraclastic packstone - calcarenite group

Finer intraclastic rocks contain intraclasts generally too fine to identify in the field, and they are similar in texture and occur at the same stratigraphic levels as other types of packstones. Only petrographic examination can reveal their intraclast support.

The intraclasts are commonly mixed with bioclasts and locally with peloids. They are difficult to distinguish from peloids as both are commonly micritic and subspherical or ellipsoidal. Commonly the intraclasts have more irregular shapes and are slightly larger (average 100 - 200 µm). Minute masses showing a higher degree of neomorphic aggradation (i.e. to microspar and/or pseudospar) have been interpreted as intraclasts, since the peloids apparently had much greater resistance to diagenetic transformation, as described previously. Also, in fining-upward sequences, intraclasts, including those of miniscule sizes, commonly differ in distribution from bioclasts and peloids. In well differentiated sequences, the intraclasts occur below bioclast-dominated layers, and the intraclasts show a fining-upward trend. Where mixed, the bioclasts occur with the finer intraclasts where intraclast sizes are segregated, or within the entire intraclast-dominated lower zone. They do not occur among the coarser intraclasts without being present in the upper finer intraclast zone. In well differentiated fining-upward sequences, peloids occur above bioclast-dominated layers, and thus are separated from the intraclast-dominated layer. This sequential relationship can aid in distinguishing very fine intraclasts and peloids in the field.

The intraclastic packstone is quite different in texture from other packstones. The intraclasts commonly show a very condensed or fitted fabric with numerous microstylolitic contacts. Any remaining matrix is very bituminous micrite with scattered very fine euhedral dolomite rhombohedra.

Intramicritic packstone, with micrite matrix predominating over cement, is the more common variety of very fine intraclast-dominated packstones. Often, at first glance through a petrographic microscope, this type of deposit appears as a lime mudstone cut by an unusually large number of microstylolites. The identification of the intraclasts is made difficult by the fitted fabric and the unusual shapes imparted to the intraclasts. Although it is unusual to find such a well sorted deposit with intraclasts this small, its stratigraphic position within fining-upward sequences leaves little doubt as to its origin. In deposits where micrite matrix is important volumetrically, the micritic intraclasts can be differentiated because the matrix commonly shows a higher degree of neomorphic aggradation.

Intrasparitic packstone has been observed only rarely. Usually, an intraclast-dominated deposit with cement filling the interstitial space does not have enough micrite remaining to be classified as intrasparitic packstone and is instead a grainstone.

3.3.1.3.d Intrasparitic packstone - calcirudite group

A number of distinct grain-supported (and matrix-supported) intraclastic calcirudites can be recognized in the study area. They cannot readily be classified further using Embry and Klovan's scheme. For convenience, and because of their importance and diversity, this group is described below under the general heading "Limestone conglomerates and breccias".

3.3.1.4 Grainstones

Grainstone is the rarest of the limestone categories in the study area. It occurs either as part of fining-upward sequences or as separate beds bounded by abrupt erosional contacts, in association with these fining-upward sequences. In fining-upward sequences they are far from being the coarsest grained beds: many packstones and/or rudstones contain far larger grains, even though they contain more micrite matrix than the grainstones. The grainstones are distinguished by their excellent sorting and scarcity of micritic matrix.

The grainstones show distinctive silicification more consistently than any other limestone. Despite silicification of grains and cement, the original textures of the grainstones are commonly preserved by variations in crystal size, crystal shapes, growth orientation and types of chert, as described below in the section dealing with diagenesis, and therefore the components are still very easily identifiable. Different from other grain-supported deposits, the grainstones very seldom show compressed or fitted textures.

3.3.1.4.a Bioclastic grainstone

Bioclastic grainstone is rare compared to peloidal grainstone. Even where predominant, bioclasts are commonly mixed with peloids. The bioclastic components are relatively well sorted and elongate fragments are parallel to bedding. Diverse faunas are represented: crinoids, brachiopods, trilobites, ostracods and some calcareous algae are the principal allochems.

3.3.1.4.b Bioclastic grainstone - coquina

Among the bioclastic grainstones this is a rudstone which is a distinctive coquina. The term has in the past been applied mostly to shell-supported beach sediments. Obviously not a beachrock here, it has nevertheless the appearance of one. The supporting framework consists of well preserved smooth-shelled pentamerid brachiopods. The shells lack orientation and show little fragmentation although they are not hinged. The shells are large (averaging 4 X 6 X 3 cm in length, width and height) and heavy (the thickness of an individual shell averages 0,75 cm but can be up to 1,25 cm, at the hinge). Other fossils present include crinoids, other types of brachiopods, favositid colonies, solitary and colonial rugosans and calcareous algae.

Coarse mosaic calcite spar cement predominates in intergranular areas, but some micrite matrix (usually less than 10 percent) occurs in sheltered sites. Nevertheless, primary porosity and permeability are excellent with little evidence of compaction.

These beds are rare and show no consistent sequential or other relationships to other types of sediments. They stand alone as distinctive massive beds with resistant weathering profiles.

3.3.1.4.c Peloidal grainstone

Peloidal grainstone is the most common type of grainstone. The sorting is excellent and the subspherical peloids generally show no evidence of compaction. Other allochems, including ooids, of comparable sizes are commonly mixed with the peloids. The ooids commonly show radial and concentric structure, even where completely silicified. Those not silicified are generally very finely micritic and distinctly darker than associated miniscule micritic intraclasts. Figures 3.6 and 3.7 diagrammatically show such grainstone as part of thin fining-upward sequences (see also Plate. 5-5).

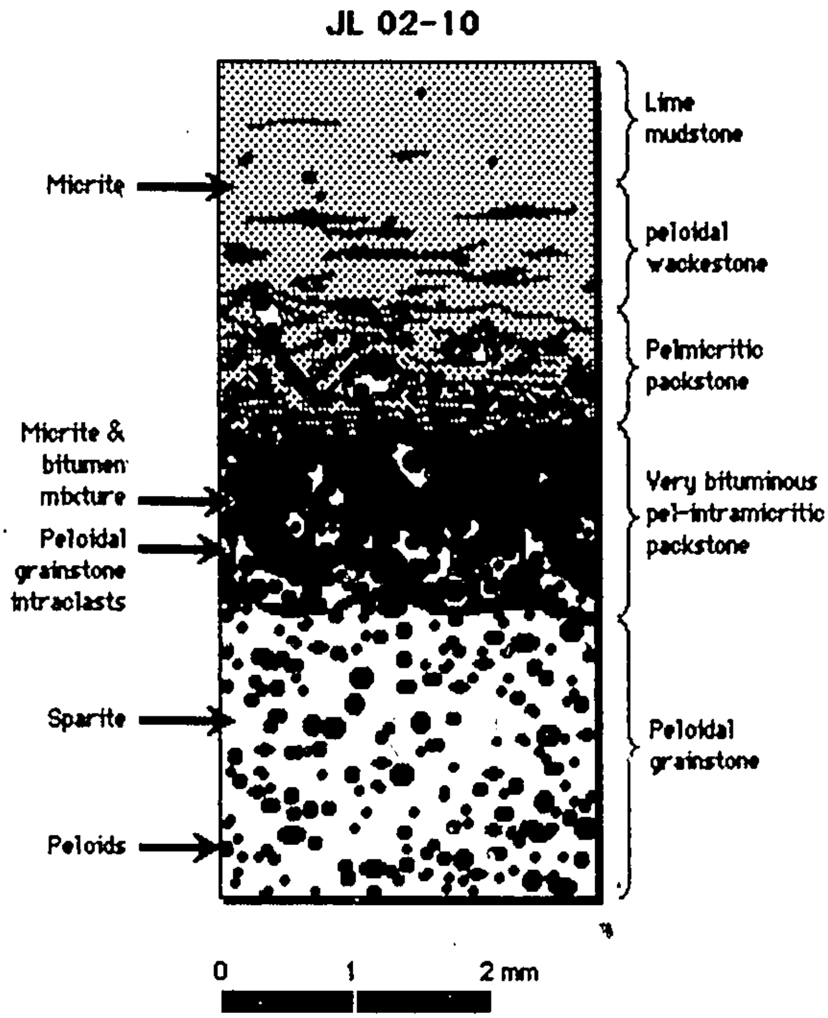


Figure 3.6 Diagrammatic representation of a very thin and fine grained limestone turbidite beginning with a peloidal grainstone, such as the ones found in section JL 02, unit 10.

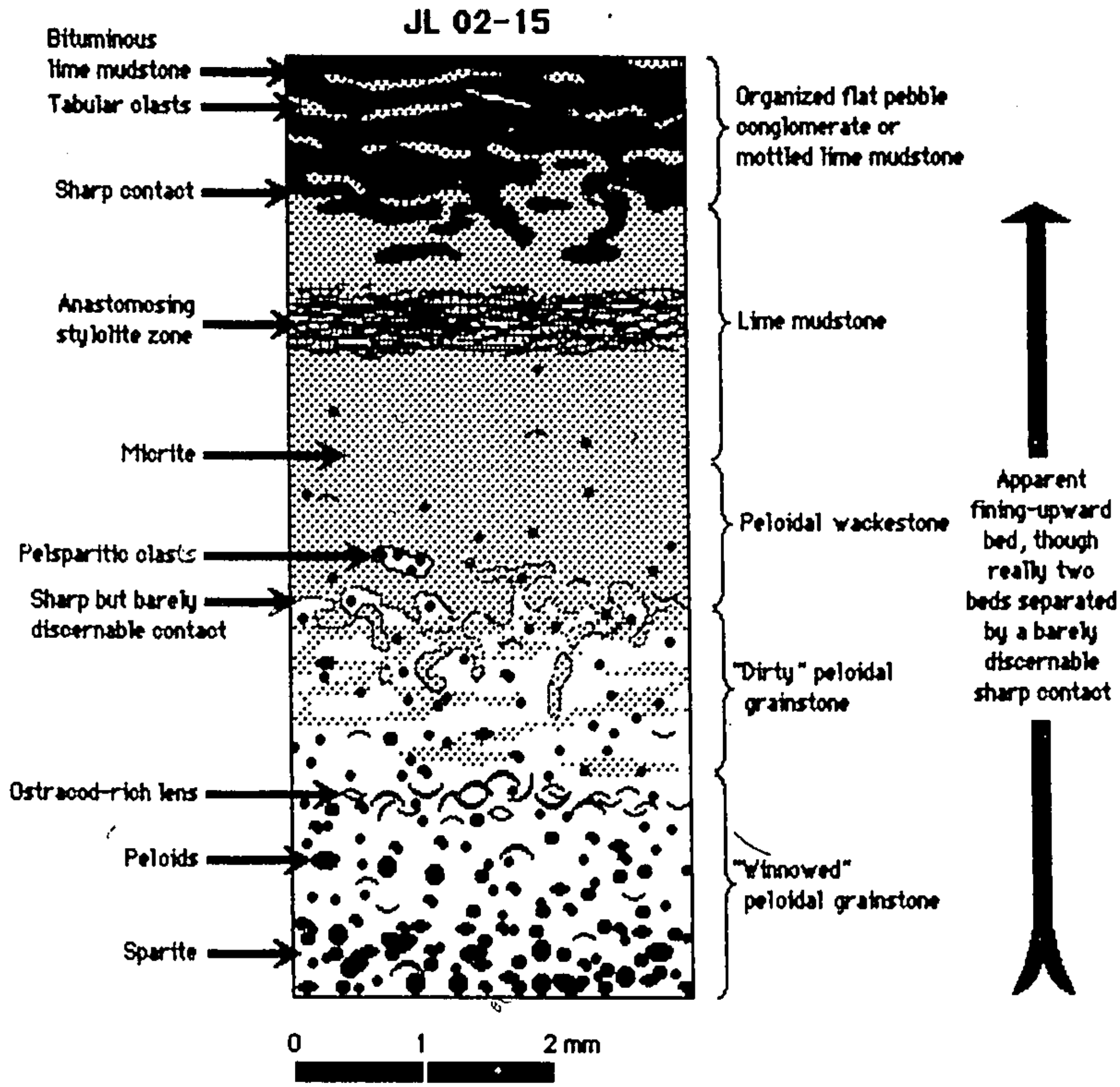


Figure 3.7 Diagrammatic representation of unit 15 of section JL 02, in which a peloidal grainstone apparently fines into lime mudstone but in reality these two beds are separated by a bored hardground.

3.3.1.4.d Intraclastic grainstone

Intraclastic grainstone is extremely rare. The very fine micritic intraclasts in these grain-supported rocks can be difficult to distinguish from peloids that occur with them. The typical excellent sorting commonly results in both being similar in size. However, the well rounded intraclasts generally have more irregular shapes than the subspherical peloids, so that at least one axis of the intraclasts is larger than the peloids. The most readily distinguished intraclasts are those in which original components such as peloids or other allochems are identifiable.

3.3.1.5 Boundstones (in limestone blocks)

In addition to the micritic limestone blocks described in section 3.3.1.1.c, there are boundstone limestone blocks. They are far more common than micritic limestone blocks (approximately 90% of all blocks), especially in older upper Wenlockian to middle Ludlovian strata. Individual blocks generally occur at stratigraphically lower positions than blocks in olistostromes, just the converse of the occurrence of lime mudstone blocks. In these very massive olistostromes, where most of the boundstone blocks occur, boundstone is the main lithology in intraclasts and blocks. These isolated blocks clearly appear out of place in the laminated or thinly bedded marlstone and lime mudstone that surrounds them and therefore could not be minor lower slope buildups. Discordant growth orientations of included *in situ* fossils confirm this.

These blocks are massive and show a wide size variation, ranging from few cm to 5 m in diameter (similar to the lime mudstone blocks). However, in olistostromes, their maximum diameters do not exceed 3 m. The fact that they are boundstones clearly indicates that they must have been detached from buildups. However, all the buildups examined in the study area, and beyond (Mayr, 1973; Kerr, 1976; and Thorsteinsson, pers. com. 1982), are composed strictly of coarsely crystalline and vuggy dolomite with very little of the original texture still identifiable. In contrast, the boundstone blocks show little dolomitization or other evidence of diagenetic transformation.

Although olistostromes will be examined in greater detail below, it is important to note here that in olistostromes these blocks are the most conspicuous components and appear to control the thickness of the flows (see Plate 3-1). They are subordinate, however, to a matrix of polymictic limestone conglomerate with much smaller clasts.

It is not uncommon to find that, during emplacement, these blocks disrupted the planar bedding of surrounding laminated marlstone and lime mudstone. The underlying layers were locally folded, with folds overturned rarely, and the overlying deposits are draped over the blocks.

These boundstone blocks consist mainly of framestone and bindstone according to the classification of Embry and Klovan (1972). In neither are the binding or framework organisms the major component. Lime mudstone matrix (especially in bindstone) and/or bioclastic matrix (especially in framestone) are more important volumetrically.

3.3.1.5.a Bindstone

In the bindstones, calcareous algae and stromatoporoids are the main binders. The calcareous algae include red coralline algae (Rhodophyta), especially some solenoporids, and green algae (Chlorophyta), especially dasycladaceans. Cyanophyte bacteria (blue-green algae; Cyanophyta) are represented especially by girvanellids and *Renalcis*. The stromatoporoids mostly have tabular and bulbous forms.

3.3.1.5.b Framestone

In the majority of framestones, colonial corals are the main faunal components. Branching and bulbous bryozoans are also major constituents, and rarely are the main frame builders. The colonial corals include favositids, halysitids, heliolitids, syringoporids and rugosans, generally in decreasing order of importance. Various solitary rugosans are also present. Stromatoporoids and crinoids are also quite important, and in some blocks crinoids are the main components. The algae in the bindstones are also present in the framestones but are not as important volumetrically.

3.3.1.6 Limestone conglomerates and breccias

The limestone conglomerates and breccias include some of the most spectacular rock types of the Allen Bay Formation-Read Bay Group undivided. In most stratigraphic sections these beds tower or hang over most other types of deposits, and portions of sections including more of these beds are clearly better exposed. Although they are a major lithology, they are generally subordinate in volume to interbedded source beds or to finer-grained limestones that overlie the intraclast-dominated deposits in fining-upward sequences.

Most of these beds have sharp, erosional lower contacts, and intraclasts concentrated in their lower portions. Their intraclasts decrease in abundance and/or size upward. Three main groups of intraclast-dominated deposits can be distinguished, based on their intraclast composition, intraclast shape and matrix composition. These are flat pebble conglomerates, conglomerates with a wider variety of intraclasts and matrix, and breccias. These can further be subdivided based on other criteria, into the types described below.

The intraclasts vary extremely in composition, a good reflection of the spectrum of rock types occurring in source areas. In addition, some conglomerates have other conglomerates as source beds. Up to three generations of intraclasts have been observed in a single thin section. Various evidence can be used to distinguish the different generations. Older generations tend to show progressively greater neomorphic aggradation. For example, a pseudosparite matrix might be the basis for distinguishing an older generation conglomerate from a younger one that might have a micritic matrix. Also, intraclasts in older generation conglomerates commonly have thin (less than 10 μm) micritic rims that are not developed around younger intraclasts. The forms of intraclast boundaries can also provide clues, with straight or curved margins reflecting the different competence of clasts of different generations. Intraclasts vary in size from less than 50 μm up to blocks 3 m in diameter (described above).

Most of these conglomerates are matrix supported, and thus are paraconglomerates. The matrix commonly consists of homogeneous micritic lime mudstone, with diagenetic transformations only occasionally interrupting its homogeneity. Bioclasts in the matrix vary from extremely rare to very abundant, as documented below. Of all the components, the one occurring most consistently and responsible for the distinctive black colour of the matrix in most conglomerates, is intergranular bitumen that coats micritic grains and dolomite crystals

In addition to variations in intraclast composition, there are large textural variations. Generally the matrix and/or intraclasts of these deposits are graded, very rarely inversely graded. As well, the intraclasts can be regularly oriented or randomly oriented in the matrix. Some deposits apparently incorporate two compositionally and texturally different gravity flows. Other textural diversity is discussed subsequently in this chapter.

3.3.1.6.a Oligomictic, organized flat-pebble conglomerate

Of all the types of intraclast-dominated limestones, flat-pebble conglomerate is among the most important. It is certainly the predominant conglomerate type in the Llandoverian part of the sequence, occasionally even predominating over all other lithologies. However, at higher stratigraphic levels, other types of conglomerates are more important and eventually predominate over flat-pebble conglomerate.

Organized flat pebble conglomerate has the intraclasts regularly oriented, commonly parallel to bedding. Usually they have gradational lower contacts with ribbon or nodular limestone. On average, about 50cm of the limestone beneath a conglomerate bed shows disruption that increases upward. The boudins, elongated lenses or "ribbons" of the ribbon limestone are increasingly fractured, folded, broken and rotated or displaced until eventually layering is no longer evident. No erosional break need be present, and partially detached portions of layers can be bent and incorporated in the overlying conglomerate.

The intraclasts, even multigeneration intraclasts, are exclusively homogeneous lime mudstone. These tabular intraclasts average 1,0 × 2,5 cm in size, but some are up to 3,0 × 15,0 cm. They are generally oriented parallel to bedding, but the topmost and bottommost portions of a deposit can show a slightly less organized fabric.

The matrix is also commonly lime mudstone. It differs slightly from intraclastic lime mudstone in being more neomorphically altered and partially dolomitized (~ 20-30 μm randomly distributed rhombohedra). Locally this matrix is bioclastic (generally crinoidal) wackestone or packstone, although usually only towards the tops of conglomerate units. This increase in abundance of bioclasts is not accompanied by any increase in size, nor is there any change in intraclast size or abundance.

In general, flat-pebble conglomerate changes upward to paraconglomerate due to gradual decrease in the number of intraclasts, and then to massive, homogeneous lime mudstone before development of ribbon limestone resumes. However, in the flat-pebble conglomerate units that show upward increase of bioclastic material in the matrix, this bioclastic enrichment continues above the conglomerate to form biomicritic wackestone or packstone. With subsequent decrease of bioclastic content, the rock then grades into lime mudstone and eventually to ribbon limestone. The bioclastic sediment overlying the conglomerate shows upward fining.

3.3.1.6.b Oligomictic, disorganized flat-pebble conglomerate

This type of conglomerate resembles organized flat-pebble conglomerate, but contains randomly oriented intraclasts (see Plate 3-5). The lime mudstone intraclasts are very similar in composition to, although slightly smaller and more fragmented than, those in organized flat-pebble conglomerate. They are less tabular, and commonly bent, folded or distorted into irregular shapes, with more rounded edges. Some intraclasts are clearly tabular, with more angular corners and abrupt contacts defining rectangular shapes.

Intraclasts rarely of different composition are mostly multigeneration clasts, as described previously. The first and second generation intraclasts are more commonly peloidal or biomicritic or biosparitic wackestone to packstone, with the micrite more intensely neomorphically aggraded than last generation intraclasts. They are also more spherical or ellipsoidal or irregular in shape.

The matrix, generally lime mudstone, differs from the matrix of organized flat-pebble conglomerate in being more diverse in composition. Bioclastic matrix is a lot more common and where present occurs throughout the flow, not just near the top. In such examples, the entire sequence grades upward from disorganized flat-pebble conglomerate, to paraconglomerate, to fining-upward biomicritic packstone or bioclastic wackestone or lime mudstone, to marlstone or ribbon limestone. The matrix is also generally very bitumen-rich.

Organized flat-pebble conglomerate, and to a lesser extent disorganized forms, are typically interbedded with ribbon limestone, and occasionally with some types of nodular limestone. Alternatively, conglomerate beds are commonly stacked in series, directly on top of each other, or alternate with massive lime mudstone. The latter is continuous with the conglomerate matrix and represents topmost gravity flow material lacking intraclasts. It also occurs commonly in the upward transition from conglomerate to ribbon limestone.

Both types of flat-pebble conglomerates show little diversity in intraclast and matrix composition. The thicknesses of individual intraclasts closely approximate the thicknesses of layers in interbedded ribbon limestone, indicating the close relationship between these two deposits. The matrix of the conglomerates is commonly lime mudstone as well, but is more altered by neomorphism, and locally more dolomitized and silicified, than in the intraclasts. Selective silicification more commonly affects the bioclasts. The locally increased dolomitization of the conglomerate matrix is similar in proportion to increased dolomization of the matrix in ribbon limestones. Locally, the transition between organized flat pebble conglomerate and underlying ribbon limestone or nodular limestone, can be seen to involve intermediate stages between a "ribbon" and derived intraclasts as well as continuity between internodule matrix and conglomerate matrix.

3.3.1.6.c Oligomictic, non-flat-pebble conglomerate

These oligomictic conglomerates, and other conglomerate types described below, appear generally at higher stratigraphic positions than the flat-pebble conglomerates. At these levels they are the principal intraclast-dominated limestone, although usually a minor lithology overall. Similar to those in the flat-pebble conglomerates, the intraclasts are mostly lime mudstone, but the oligomictic conglomerates differ in intraclast shape and matrix composition. In the oligomictic conglomerates, the intraclasts are mostly spheroidal or ellipsoidal. Other minor intraclasts are of peloidal wackestone, pelmicritic packstone, or pelsparitic packstone. Some conglomerates contain multi-generation intraclasts; they are far more common than in disorganized flat-pebble conglomerates, but still minor in proportion.

Usually the matrix represents original lime mudstone that commonly is more highly neomorphically aggraded than in the flat-pebble conglomerates. A compositionally varied, usually crinoid-rich, bioclastic matrix can also be present.

These conglomerates generally grade upward into lime mudstone or a matrix-like deposit, as the intraclasts gradually become fewer and/or smaller and eventually disappear. Their lower contacts are abrupt and probably erosional. The entire sequence is generally quite thick and massive with few breaks that can be interpreted as bedding. Its lowermost portion occasionally shows a condensed fabric with fused intraclasts.

3.3.1.6.d Oligomictic, bimodal conglomerate

Bimodal conglomerate, typically with intraclasts of two sizes, appears in the field to be most similar to flat-pebble conglomerates since the larger intraclasts are usually tabular lime mudstone pebbles. Closer examination reveals other intraclasts that are much smaller, usually by two orders of magnitude, or more. The larger intraclasts average a few centimeters in length while the smaller intraclasts average only 50 to 100 µm in mean diameter. In contrast to other conglomerates, there are no intraclasts of intermediate sizes.

These two types of intraclasts are consistently mixed together, but the smaller ones might better be considered as part of the matrix. In a fining-upward bimodal conglomerate, showing decreasing size and abundance of larger intraclasts, the smaller intraclasts do not decrease in size or abundance, as was the case with bioclasts in previously described deposits.

Although the larger intraclasts are composed mostly of lime mudstone, there are also intraclasts of other lithologies, similar to those of other oligomictic conglomerates previously described. The smaller intraclasts are composed exclusively of bituminous micrite. They are generally spherical although deformation due to compaction, and sutured, stylolitic contacts, the result of pressure solution due to compaction, are not uncommon. These smaller intraclasts are similar to peloids and can be mistaken for them.

However, where peloids and smaller intraclasts occur together, the intraclasts are generally larger, more bituminous and more intensely altered diagenetically.

The matrix is much poorer in allochems than in polymictic conglomerates described below. However, the variety of bioclasts present is comparable. Excluding the smaller intraclasts, the matrix is commonly mudstone or wackestone. If the smaller intraclasts are considered as part of the matrix then they are a major matrix component. In addition to bioclasts and the smaller intraclasts, the remaining matrix is composed mostly of highly neomorphically aggraded, substantially dolomitized mudstone.

These deposits commonly have condensed fabrics. In extreme cases, they appear to consist mostly of large lime mudstone intraclasts with minor intergranular micrite. However, the lime mud is cut by a network of microstylolites of various orientations that represent the fused contacts of smaller intraclasts. Furthermore, in areas where micrite is still present, it has been intensely neomorphically aggraded and dolomitized.

In general, silicification is not as important as in the polymictic conglomerates, but where present it is similar in texture.

3.3.1.6.e Polymictic conglomerates and olistostromes

Of all the conglomerates these are by far the most diverse, in both intraclast and matrix properties. The deposits are thick (above 1 m on average) and massive, with few breaks that can be interpreted as bedding. Most are simple, non-graded deposits with abrupt boundaries, within thick sequences of thin-bedded marlstone or lime mudstone (see Plate 3-1). Some fine upward from erosional basal contacts into matrix-like deposits, some of which in turn fine upward into marlstone or massive lime mudstone. These conglomerates occur mostly in sections JL 01 and 10.

Intraclast lithologies include a diverse range of mudstones, wackestones, packstones, grainstones and boundstones representing most of the limestone types described earlier in this chapter. Thus, the intraclast diversity is far larger than in the other types of conglomerate. Many of the types of intraclasts occur in a single deposit, in equal proportions or dominated by only a few types. The tabular lime mudstone intraclasts predominant in flat-pebble conglomerates also occur.

Intraclasts of massive, homogeneous, unfossiliferous lime mudstone (see section 3.3.1.1.c) are common. They are micritic to pseudosparitic and conspicuous for their very light colour, reflecting the absence of intergranular bitumen. They are extremely irregular in shape and vary from fractions of a mm to 5m in diameter.

Multi-generation intraclasts are present, but by far the most common types of intraclasts are bioclastic wackestones and packstones and less commonly grainstones. Framestones and bindstones are also present, more commonly as larger intraclasts. Most of the allochems are fossils, especially crinoids, though peloids are also fairly common. Ooids, however, occur extremely rarely. These intraclasts commonly have irregular shapes and their sizes vary enormously, from 50 μm to reefal blocks larger than 7 m in diameter. Some of these beds contain enough of these "reefal" blocks to be considered olistostromes, described separately below.

The polymictic conglomerates have the coarsest matrix, commonly wackestone or packstone, of all the conglomerates. The granular fraction is mostly poorly sorted fossil fragments of the same groups as occur in the intraclasts. Crinoids are best represented, but also present are brachiopods including particularly well preserved and numerous thick-shelled pentamerids, various calcareous algae, including green algae (dasycladaceans), and red algae (solenoporids), encrusting cyanophytes (*Girvanella* and *Renalcis*), various solitary and colonial corals including rugosans, favositids, halysitids, heliolitids and syringoporids, stromatoporoids, bryozoans, ostracods and trilobites.

Some of these deposits have a relatively open framework with up to 20-25% intergranular pore space. Few of the bioclasts and the intraclasts show sutured contacts. Intergranular spaces are partly filled by spar cement, mostly calcitic, less commonly dolomitic or siliceous, and to a minor extent bituminous. These are the only conglomerates to contain consistently dolomitized matrix material. In the field, these beds with high primary porosity and permeability are friable and recessive in the midst of otherwise massive resistant beds.

These conglomerates contain significantly more cement than others. This is mostly drusy mosaic calcite or locally isopachous calcite. In some areas, two generations of calcite cement comprise an initial very thin (averaging 10 μm) zone of dogtooth spar around bioclasts, and subsequent coarse mosaic spar filling remaining pore spaces. Coarse mosaic dolomite cement occurs locally. Siliceous cement occurs extremely rarely and mostly in significantly silicified deposits. It consists mostly of concentrically layered chalcedony, less commonly of equidimensional, anhedral quartz micro-crystals.

Dolomite occurs, not only as cement, but also as replacement of micrite. As micrite increases in proportion up-section, the proportion of dolomite generally increases up-section as well. Small (average 30-40 μm) dolomite rhombohedra usually have replaced < 10% of micritic materials, but locally this replacement is pervasive.

There is more evidence of neomorphic aggradation in the polymictic conglomerates than in any other conglomerate. There is also more evidence of selective silicification than in other conglomerates, but silicified components, nevertheless, are not common, and completely silicified deposits occur only rarely. Although the order of susceptibility of components to silicification varies somewhat from bed to bed, usually the fossils, especially crinoids, brachiopods, corals, bryozoans and dasycladacean algae, were most susceptible. Other silicified components are, in order of importance, lime mud portions of the matrix, spar cement crystals, especially their margins, and the margins of intraclasts. In beds where the silicification was complete, it was apparently a slow, ordered process that preserved original textures and other properties of components. Thus, for example, micrite was replaced by silica crystals smaller than one micron, therefore similar in size to the original micrite. The lower portions of the conglomerates are more silicified.

Olistostromes have been distinguished here as a particular variety of polymictic conglomerate with large amounts of "reefal" blocks. The olistostromes consist of these larger blocks and a so-called "matrix" generally of polymictic limestone conglomerate, as just described (see Plates 3-3 and 5-4). Intraclasts in this "matrix" are much smaller than the blocks and are much more diverse compositionally; the large blocks consist only of lime mudstone, bindstone, and framestone. Because of this bimodal character of larger blocks and smaller intraclasts, the latter are included as part of the matrix.

Porosity in the olistostromes is very good, despite partial occlusion by several generations of cement. Four types of cement are evident, corresponding to the four recognized in finer polymictic conglomerates.

The olistostromes are more dolomitized and silicified than any other deposits. The lime mudstone of their matrix is much more dolomitized than mudstone matrix in finer polymictic conglomerates. Selectively silicified bioclastic fragments (mostly solitary rugosans, crinoids, bryozoans, calcareous algae and some colonial corals, in order of importance), other allochems (such as the peripheries of smaller intraclasts), and even some micrite and calcitic cement crystals, are also more common than in other conglomerates. Progressively younger olistostromes are more dolomitized and silicified. These features are shown by the "matrix" but not by similar components in the larger blocks.

3.3.1.6.f Condensed polymictic breccias

These breccias are associated with bimodal and polymictic conglomerates. They resemble the latter in clast composition and size, but show markedly condensed fabrics (see Plate 5-3). They consist largely of well-fitted, angular intraclasts, most of them in stylolitic contact. The generally smaller lime mudstone intraclasts show more pronounced effects of pressure solution than other intraclasts. The little remaining matrix, apparently originally micritic, is now mostly dolomite and bitumen containing various proportions of allochems, mainly fossils. These breccias are rare in comparison to the limestone conglomerates.

3.3.1.6.g Tabular clast breccias

Tabular clast breccias are known principally in sections in the southeastern part of the study area, where they occur rarely. Most are in the predominantly dolomitized lower Llandoverian part of these sections and are described, therefore, in section 3.3.3 (Dolostones). Less completely dolomitized examples occur very rarely at higher stratigraphic levels. These resemble the previously described flat-pebble conglomerates: their intraclasts are similar in size (generally between 0,5 × 1,0 in width and 1,5 × 5,0 cm in length), but rather more angular, are randomly oriented, and are composed of more dolomitic micritic lime mudstone. Some intraclasts are internally laminated. The matrix is much more dolomitic than in the flat-pebble conglomerates and dolomite commonly predominates. Some of the 30 μm dolomite crystals show evidence of breakage. Fossils and other allochems are extremely rare in both matrix and intraclasts.

Silica is also more common than in the flat-pebble conglomerates. Fragments of conchoidally fractured, bituminous chert nodules, similar to those in the underlying bituminous dolostone units, occur locally. Some components, especially fossils, are rarely selectively silicificied, but not all the fragments of a particular group are silicified in the same rock. Some intraclasts are silicified peripherally.

3.3.2 Limestones substantially altered after deposition

A second major group of limestones shows substantial affects of a variety of post-depositional processes. One result is a complex mixture of lithological types within rock units. Another is that bedding is commonly obscure or even absent, and aspects of mottling or nodule development are more conspicuous. Three main categories have been distinguished for purposes of classification: argillaceous and non-argillaceous nodular limestones, and ribbon limestones. The classification of Embry and Klovan (1972) can still usefully be applied to components within these rocks, as for example, a nodular limestone can consist of burrow fills of dolomitic bioclastic wackestone in a lime mudstone matrix.

3.3.2.1 Argillaceous nodular limestones

Argillaceous nodular limestones are distinctive in consisting generally of two very contrasting lithologies, giving them a mottled structure. A rubbly-weathering character usually results where the nodule-forming lithology is more resistant than the enclosing matrix. These nodular areas are ellipsoidal, or augen-like masses.

The nodular portions are mostly homogeneous, dark, micritic lime mudstone, and locally are neomorphically transformed lime mudstone.

Allochems are rare. The nodular components are rarely wackestone, and locally are spicular (especially in the lower part of the Allen Bay Formation). The sponge spicules are all recrystallized to monocrystalline calcite spar. Also present rarely are discrete calcite spar cement fillings of primary pore spaces, varying in size and shape. Microfractures filled with calcite spar cement also occur rarely, restricted to the nodular portions. Local low amplitude microstylolites, apparently defined by bitumen concentrations, are similarly restricted to the nodular portions.

The internodular portions are far more diverse compositionally, and generally consist of a mixture of lime mudstone (mostly microspar and pseudospar), clay minerals, dolomite crystals, bitumen and allochems. The allochems are commonly more abundant than in the nodular portions and especially so where trilobite fragments predominate.

Internodular lime mudstone occurs randomly in small masses through which dolomite crystals are distributed. These crystals are generally very fine, varying between 15 and 50 μm in size, and averaging 25 μm. They appear to be euhedral rhombohedra, commonly with altered, probably dedolomitized edges. Dolomite crystals vary in proportion from negligible concentrations up to 85 percent of the internodular portions, averaging about 30 percent. Beds at lower stratigraphic levels contain less dolomite, and the proportion increases gradually upward stratigraphically, until dolomite completely replaces internodular lime mudstone.

Clay minerals vary in concentration but are never main components. Even in minor concentrations, and occasionally in association with wisps of bitumen, these platy minerals give a foliated appearance to internodular areas where they are oriented tangentially to nodule surfaces. Bitumen occurs rarely as wisps or anastomosing, very elongated lenses (10 X 100 μm). More commonly it is intercrystalline in the lime mudstone. It varies in concentration from being usually only a minor component, rarely up to 40 percent but is always more abundant in, and produces far darker colours in, internodular areas.

Neomorphic rims, 50 to 200 μm thick, averaging 100 μm, generally separate nodular and internodular portions (see Fig. 2.2 and Plate 7-2). These are especially conspicuous as lighter coloured rims around the generally much darker nodular lime mudstone. These represent a very gradual increase in grain size from micritic lime mudstone of nodule cores, into enclosing microsparite-dominated and pseudosparite-dominated zones. Dolomite crystals occur locally in the outermost pseudosparite zone. Despite the presence of neomorphic rims, the contacts between the nodular and internodular portions are generally abrupt.

Neomorphic zoning also occurs rarely in internodular areas and is seen as cores of mixed microsparite and pseudosparite locally enclosed by zones predominantly of pseudosparite.

Neomorphic rims are present usually where the nodular and internodular portions are similar mineralogically, such as in the lowermost beds of the Allen Bay Formation. In those examples, even where a nodular structure is clearly visible in the field, petrographic examination reveals only an extremely gradual, scarcely perceptible transition between the almost exclusively micritic lime mudstone of the nodules and the slightly coarser and more bituminous internodular lime mudstone. Such homogeneity generally reflects the absence of clay minerals from internodular portions.

In some argillaceous nodular limestones, the concentration of clay minerals and dolomite increases markedly, at the expense of the lime mudstone, from internodular areas lateral to nodules to those above and below closely-fitted nodules. Commonly accompanying this change is a gradual disappearance of neomorphic rims in the same direction. The outermost pseudosparite zone disappears first, and the tops and bottoms of the nodules eventually show extremely sharp contacts between the internodular and nodular lithology.

It is usually evident that the nodular lithology represents original host sediments, while the internodular portions, especially those in nodule-rich beds, are burrow fills. However, different generations of burrow fills can be distinguished, as there are those containing typical internodule lithology, as well as fills of lime mudstone with very little clay content. The latter can appear nodule-like in section.

Jones, Oldershaw and Narbonne (1979) recognized two major types of rubbly limestone, Types I and II, which equate with Types I and II nodular limestone, respectively, in this study. Their Type I has discrete lumps (in this study, distinct nodules). Type II rubbly limestone has continuous and/or semi-continuous layers, or is bedded. These two major types of rubbly limestone were further subdivided on the basis of major matrix components (in this study, internodular components), into argillaceous and dolomitic types. This designation according to end members in a continuous series implies an inverse relationship between argillaceous and dolomitic contents. This disagrees with Dunham and Olson (1980), McHargue and Price (1982) and Wanless (1979), who reported very slow dolomite crystal growth in association with increased argillaceous content, which does not appear to be applicable in this study. Dolomite content instead shows a much more consistent relationship to stratigraphic level. For this reason if the amount is sufficiently high, the terms argillaceous and/or dolomitic were instead added to nodular limestone. What is referred to here as Type III nodular limestone (labyrinthine mottling) was referred to as "mottled dolomitic limestone-labyrinthine" by Jones, Oldershaw and Narbonne (1979).

3.3.2.1.a Type I (discrete nodules)

Type I has the highest proportion of internodular lithology of all types seen in this study, and locally the volume of argillaceous internodular lithology even exceeds nodule volume. It most commonly includes mottled limestone with fossil-rich internodular portions. Nodules are generally ovoid and rarely linked. Layering is obscure as, commonly, vertical offset of the nodules produces a fitted fabric.

3.3.2.1.b Type II (nodular layers)

Internodular lithology is less important volumetrically in Type II nodular limestone, and mostly occurs as thin undulating layers that only locally interrupt the more massive lime mudstone beds. The lime mudstone beds show a diagnostic pinch and swell morphology that resembles boudinage.

3.3.2.1.c Type III (labyrinthine mottling)

Type III nodular limestone commonly has the least volume of argillaceous internodular lithology. However, these portions are connected vertically and horizontally, and appear as honeycomb or labyrinthine patterns in vertical sections. This is the only type in which burrowing is clearly responsible for the mottling, confirming the findings of Jones, Oldershaw and Narbonne (1979).

3.3.2.2 Non-argillaceous nodular limestone

Non-argillaceous nodular limestone appears superficially to be quite distinct from the argillaceous limestones described above. Both lithologies creating the nodular structure appear massive and have comparable resistance to weathering in contrast to the differential weathering character of argillaceous limestones. Only a conspicuous colour contrast on fresh surfaces reveals the nodular structure, and the colour contrast tends to be obscured by weathering.

The distinctive colouring is due to differential distribution of bitumen in pore spaces: brown to black bitumen-rich portions contrast with typically lighter coloured bitumen-poor portions. The bituminous lithology has irregular contours and in part consists of sinuous and multi-branching lenses, some preferentially elongated horizontally. The bituminous portion generally constitutes the nodular fraction, defined here as those portions apparently not connected with others of similar composition. Locally, where bituminous lithology predominates, the non-bituminous portion constitutes the "nodular" fraction and forms similar irregular-shaped discrete bodies. These two end-members, locally, gradually intergrade vertically so that it is possible to observe that it is not the bitumen which changes location but the bituminous portion which becomes more, or, less important volumetrically. These gradual up-sequence changes in the proportion between the two rock fractions that constitute these deposits are further discussed in section 3.4.2.2 (also see Plates 6-2 to 6-5). For this reason, it is much more appropriate to refer to these two fractions by lithological terms (bituminous dolomitic fraction and lime mud fraction), and use the terms nodular and matrix here only to signify their relative abundance.

Units of this lithology are usually massive and unbedded or very weakly thin-bedded. Fossils are very rare and of low diversity: only scattered crinoid fragments, and brachiopod and ostracod shells were recognized. This nodular limestone occurs mostly in the lower Allen Bay

Formation, though stratigraphically higher than the argillaceous nodular limestones.

Although similar in composition, the two fractions differ in more than their bitumen content. The lighter-coloured lime mud fraction is homogeneous micritic lime mudstone, except for enclosing neomorphic rims. It also contains minor sparite- and bitumen-filled primary pores. In the darker coloured bituminous dolomitic fractions, bitumen is important as an intercrystalline and minor, wispy component, constituting up to 40 percent of the lithology. Lime mudstone and dolomite are the other principal components. At lower stratigraphic positions in the Allen Bay Formation, the bituminous fraction is distinguished in having coarser grained lime mudstone than the non-bituminous fraction. The coarser grain size represents neomorphically aggraded microspar and pseudospar. Dolomite content increases at the expense of lime mudstone at progressively higher stratigraphic levels and eventually dolomite replaces the non-bituminous lime mudstone as well. The dolomite crystals are randomly scattered, and even where largely replacing lime mudstone, the rhombohedra do not much exceed 20 - 30 μm.

As in the argillaceous nodular limestones, the selective distribution of bitumen and dolomite are responsible for very sharp contacts between the two fractions. Coarsening-outwards neomorphic rims around the lime mudstone fractions, although generally thinner than in the argillaceous nodular limestones, suggest that the contacts were not originally as sharp.

3.3.2.3 Ribbon limestone

Ribbon limestone has regular bedding and/or lamination. The beds are thin to very thin and consist of alternating dark bitumen-rich and light bitumen-poor layers, although these layers are, in fact very elongated lenses of these lithologies. In this study, the use of the term ribbon limestone corresponds to Read's (1980) usage in the Ordovician ramp models of Virginia and West Virginia (also see Markello, 1981; Markello and Read, 1982). Bedding is present though not continuous. Weathering commonly leaves the bitumen-poor layers buff or light grey, while the bitumen-rich layers usually become much lighter. The latter are commonly yellowish, reflecting the oxidation of iron in the contained dolomite, a very important component of these bituminous layers.

Both types of lithology forming the ribbon limestone are massive and relatively resistant to weathering, due to a lack of clay minerals. Although intercrystalline bitumen is important, locally exceeding 20 percent of the components of the bituminous layers, it is subordinate to lime mudstone and/or dolomite. In the bituminous layers, the lime mudstone is mostly microspar and pseudospar. Very fine dolomite rhombohedra, averaging 30 μm in size, are randomly distributed.

The main component in bitumen-poor layers is dense, homogeneous micritic lime mudstone. Locally, these lime mudstone layers are interrupted by spar-filled fractures restricted to the layers. Minor bitumen occurs locally in pore spaces. Both the bituminous and the lime mudstone layers generally lack allochems, although ostracods and crinoid fragments were rarely identified.

Contacts between the bituminous and lime mudstone layers are very abrupt, enhanced by the selective bitumen and dolomite distribution. Neomorphic rims are not evident.

The layering in ribbon limestone is locally interrupted, sometimes to the extent that a series of nodular masses is formed at a distinct horizon. This concentration of generally elongate, nodular, lime mudstone masses at distinct horizons, without offsets, and bounded by fairly continuous bituminous layers, gives the impression of very good layering (see Plate 6-6).

Ribbon limestones in the lower Allen Bay Formation, are stratigraphically higher than the nodular limestones in the formation. They are commonly interbedded with organized and/or disorganized flat pebble conglomerates. They commonly grade upward from the massive lime mudstone forming the last unit of fining-upward sequences overlying the conglomerates, and precede the next conglomerate. They are thus above the chert-rich marker of the Allen Bay Formation, while the nodular limestones are mostly below, except for some beds in the upper part of the Read Bay Group.

3.3.3 Dolostones

In the central and southern islands of the Arctic Archipelago, the Allen Bay Formation consists mostly of massive dolostone within which very few original textures or structures are identifiable (Mayr, 1980). In the study area dolostone is important, but not the main lithology in the upper Ashgillian to lower Llandoverian segment of the Allen Bay Formation-Read Bay Group undivided, i.e. below, within and just above the chert marker unit. Bioherms and associated flankbeds are abundant in the Wenlockian and Ludlovian portions of the southernmost sections but are pervasively dolomitized. Rocks that partly dolomitized, are described in other parts of this chapter.

Properties of the dolostones appear to vary significantly with the amount of intercrystalline bitumen. Dolostones with little of no bitumen have a coarser anhedral mosaic of crystals; richly bituminous dolostones usually have finer subhedral to euhedral crystals, and are usually much more silicified.

Although, there is some evidence of selective and sequential dolomitization, a wide variety of lithologies have been dolomitized to some extent. Even in those completely dolomitized, the original presence of particular structures or components (e.g. nodularity, intraclasts, fossils) is commonly reflected by contrasting crystal sizes and shapes.

The dolostones are categorized into 6 major groups, as described below.

3.3.3.1 Bituminous, massive dolostone

Volumetrically, bituminous massive dolostone is the second most important type, second only to nodular or "marbled" dolostone (see section 3.3.3.5). It is commonly the main constituent of the cherty unit in the Ordovician-Silurian boundary beds in the area (see Plate 1-1). It is present in most sections, although less important than non-fossiliferous, bedded dolostone (see section 3.3.3.4) in most of the northern sections.

The rock consists of unfossiliferous, finely crystalline dolomite and intercrystalline bitumen. The dolomite crystals are subhedral and average 50 μm in size, although in the most bituminous deposits, the crystals are smaller (20-30 μm) and more euhedral. The bitumen content (average 5-10 percent, exceptionally up to 40 percent) is higher than in any other carbonate rocks in the area. It is primarily intercrystalline, but also occurs concentrated in wisps < 10 μm thick, parallel to bedding, and rarely in lenses 30 μm × 2 mm. Coarsely crystalline dolomite spar locally fills vugs.

The rock is typically homogeneous and massive, although locally discontinuous, faint nodular structure is evident, as well as occasional very faint laminations. Nodular and/or bedded chert is commonly incorporated. Some of the nodules are fractured, with the slightly displaced fragments separated by bituminous dolostone, like the enclosing rock.

3.3.3.2 Non-bituminous, massive dolostone (buildups)

Mound-shaped buildups of resistant, very light buff to white dolostone occur in several of the southern sections. In sections JL 01 and JL 06, Wenlockian to Ludlovian buildups are very large (lengths and heights about 500 × 50 m in former and 1500 × 200 m in latter). Smaller later Ludlovian buildups occur in section JL 01 (not fully exposed but averaging 40 × 10 m in size), and one of Ashgillian age in Section JL 03 (10 m thick).

The dolomite in these buildups is very pale due to the lack of bitumen, and is usually medium to coarsely crystalline (average 200 μm or locally coarser). The rock typically has 5-10 percent (rarely higher) isolated vuggy porosity, and no permeability. Some vugs are up to several cm across. Some larger ones are partly filled by coarse dolomite spar and smaller ones are more completely filled. The smaller buildups are generally less vuggy.

The buildups are largely massive and structureless, with no allochems evident. However, fossils, mostly as ghost structures and molds, can be recognized locally in some marginal areas of larger buildups, and include crinoids, stromatoporoids, bryozoans, colonial corals (rugosans, syringoporids, favositids, heliolitids, halysitids) and possible calcareous algae. Smaller mounds are commonly less coarsely dolomitized and preserve somewhat more evidence of original textures and structures. The 4 mounds in JL 01 were apparently originally boundstone in which stromatoporoids or halysitid corals predominated, accompanied by favositids, syringoporids and heliolitids. The mound in JL 03 was apparently a stromatoporoid-dominated framestone or boundstone with a variety of *in situ* crinoid, halysitid and favositid coral, and sponge fossils.

Beds enclosing the buildups also provide important information about their original characters and relationships. Dolomitized debris aprons surround smaller mounds in section JL 01 (described below in 3.3.3.3). One of these mounds is capped by only slightly dolomitized stromatoporoid boundstone and the mound in JL 03 rests on undolomitized (stromatoporoid) framestone with abundant and varied fossils, including colonial rugosoans, favositids, halysitids and brachiopods. These framestone or boundstone deposits are possibly more representative of the original characters of the massive dolostone of the associated mounds.

3.3.3.5 Fossiliferous, bedded dolostone

Fossiliferous, bedded dolostone is less important volumetrically and occurs mostly as dolomitized flankbeds surrounding the smaller Ludlovian buildups of section JL 01. Locally, fining-upward of abundant fine bioclasts can be recognized. The rock varies from medium crystalline dolostone to finely or very finely crystalline dolostone. In general, however, only a finely granular texture is evident, produced by a mosaic of closely fitted anhedral crystals. These deposits are thin to medium bedded with bedding defined commonly by microstylolitic seams of clay minerals, bitumen, and very fine dolomite rhombohedra.

Some vuggy porosity remains, but most original porosity is now filled with coarse dolomite sparite, slightly more transparent than surrounding dolomite. Some larger vugs have also been partly filled by bitumen. Greater bitumen content, either peripheral or central to the vugs, commonly has much more euhedral dolomite spar associated. The little remaining pore space was filled, in sequence, by calcite spar, silica cement and a second generation of dolomite spar. Exceptionally, empty vugs form up to 15% of rock volume.

Some cylindrical vugs of uniform diameter (usually less than 0,5 cm) represent molds of colonial fossils such as syringoporids, or rugosans. Fossils are also represented by characteristically shaped bodies of finer dolomite or by ghost-like outlines that are slightly darker than the dolomite. Crinoid ossicles, the most common allochem, can be identified this way. Silicified fossils are more obvious, but rare in occurrence and the silicification appears to be only a surface phenomenon. A wide variety of fossils, also dolomitized, can be observed in most fossiliferous bedded dolostone: crinoids, pentamerid, spiriferid and other types of brachiopods including *Atrypoidea* sp., ostracods, orthoconic cephalopods, gastropods, trilobites, branching and bulbous bryozoans, stromatoporoids, calcareous sponges, *Receptaculites* sp., calcareous algae including dasycladaceans, solenoporids, *Girvanella* sp., and phylloid (e.g. *Renalcis* sp.) algae, colonial rugosans, and alveolitid, heliolitid, favositid, syringoporid and halysitid corals. Even though considerable faunal diversity is typical, some beds such as in unit JL 01a-61, have concentrations almost exclusively of one or very few types of fossils, such as aligned bryozoan branches, or pentamerid brachiopods and orthoconic cephalopods.

Other allochems include peloids, some unidentifiable fossils, probably brachiopod shells, now consisting mostly of a mosaic of 30 μm dolomite crystals, darker than the surrounding dolomite, and extremely rarely, very finely crystalline intraclasts with very abrupt contacts. In those allochem-rich deposits, intergranular spaces originally filled with micrite are slightly darker and much finer than those filled with sparite.

Dolomitized fining-upward sequences, show a gradual change upward from massive, light-coloured dolostone having a sharp lower contact, to laminated, darker-coloured dolostone. The lower portion contains very minor and minute lenses of light-coloured chert and the top portion is extremely bituminous (up to 40 percent) with bitumen lamination. Almost pure bitumen is distributed partly as thin (~ 10 μm), elongated, wispy filaments, and as an intercrystalline deposit. The dolomite crystals fine-upward, generally from 50 to 10 μm. The sequences locally change upward from medium crystalline dolostone to finely crystalline dolostone, to nodular dolostone (40 percent non-bituminous medium crystalline dolomite, and 60 percent bituminous fine dolomite, as described below), to laminated and bituminous very finely crystalline dolostone. Fining-upward deposits with only the micritic matrix dolomitized, and the grain-dominated lower portion still mostly calcareous are fairly common.

Another type of fossiliferous, bedded dolostone is exclusively associated with the smaller mounds of JL 01. These medium-bedded deposits have better preserved original structures and textures. They commonly consist of medium crystalline dolostone, representing original intraclastic and peloidal bioclastic packstone or rudstone, fining-upward into silty and very bituminous, very finely crystalline dolostone with dolomite crystals less than 15 μm in size.

These fossiliferous dolostones are commonly interbedded with dolomitic marlstone (see section 3.3.4.4) and others with lime mudstone. The margins of lime mudstone bed are dolomitized with dolomite content decreasing to zero across a 30 cm transition zone, leaving central portions of thicker beds untouched by the dolomitization.

3.3.3.4 Non-fossiliferous, bedded dolostone

The non-fossiliferous, bedded dolostone occurs mainly in the northwestern sections (JL 16 and JL 04) where it is stratigraphically equivalent to the chert-rich, lower Llandoverian, bituminous, massive dolostone typical of most sections farther south. The dolomite is euhedral, very finely crystalline (sizes from 15 to 50 μm, averaging 30 μm), and richly bituminous (up to 30 percent). Most of the bitumen occurs as thin (10 μm) wisps, but the resulting vague lamination is only apparent in thin sections because of the overall dark appearance of the dolomite. Although largely unfossiliferous, the rocks extremely rarely contain three dimensional graptolites, some of them pyritized.

The rocks usually show very thin to medium planar bedding, locally wavy bedding where interbedded with marlstone or chert. However, their overall aspect is massive and homogeneous, even where interbedded, because of the common intergradation between chert and dolomite, their common conchoidal fracturing and their uniformly dark colour.

Non-fossiliferous, bedded dolostone also occurs in section JL 01, just above the chert marker unit. It differs in being only slightly bituminous and more obviously laminated, and in containing 5 to 10 percent vug-filling, coarse dolomite spar. This rock type also occurs rarely where lime mudstone has been dolomitized, for example where it occurs at the tops of limestone turbidites or between beds of fossiliferous dolostone.

3.3.3.5 Nodular or "marbled" dolostone

Nodular or "marbled" dolostone is the most important dolostone in the area, by volume. Most of it is Ashgillian or Llandoverian, and occurs mostly below, and to a lesser extent in and above the lowermost Llandoverian chert marker unit. It is more common and abundant in, but not restricted to, the southeastern sections. It directly overlies non-argillaceous nodular limestone (3.3.2.2), and the dolomitized portions of the latter are similar to the bituminous portions of the nodular dolostone. The nodular character is also similar in both rock types, although some nodular dolostones are more like argillaceous nodular limestones in structure. The structure is most conspicuous on fresh surfaces and commonly is obscured by weathering.

The nodular or "marbled" structure is defined by two components: dark bituminous finely to very finely crystalline dolostone, and light, mostly non-bituminous finely to medium crystalline dolostone. With increasing bitumen content in these rocks in northern sections, the lighter portions also become somewhat bituminous and there is less contrast between the two. Depending on their proportions in a rock, both components can form either matrix or nodular fractions. These sequential changes are further described in section 3.4.2.2.

The contacts between nodular and matrix fractions appear abrupt in outcrops but are gradational in thin sections, involving gradual changes in bitumen content, and in dolomite crystal size and shape. Bitumen content varies from less than 1% in the light fraction to between 2 and 30 percent (average 10 percent) in the dark fraction. Dolomite crystals in the light fraction are mainly anhedral (closely fitted) and 40-200 μm (average 110 μm) in size, and in the dark fraction mainly euhedral (loosely fitted) and 20-100 μm (average 50 μm) in size. The rocks contain few identifiable fossils; only poorly preserved, coarsely dolomitized crinoid and brachiopod material has been recognized.

Open vugs and vug-fillings occur mostly in the light non-bituminous dolostone; open vugs form less than 5% of total volume. Vug fillings of coarse anhedral clear dolomite spar contain intercrystalline bitumen rarely, especially peripherally.

These deposits form massive units without internal stratification, except for some concentration of "nodules" at particular levels that suggests a thin to medium bedding.

Another type of nodular dolostone that occurs rarely more resembles argillaceous nodular limestone. Much of it consists of more elongate nodular dolostone bodies surrounded by slightly less resistant, laminated, internodular dolostone. The nodular bodies are somewhat more bituminous, but not significantly different in crystal size (70 μm) from internodular portions. The rock is unfossiliferous and can contain vug-filling dolomite spar.

3.3.3.6 Dolomitized conglomerates and breccias

Dolomitized conglomerates and breccias are least important volumetrically in the area. They are mainly Llandoverian, occurring just above the lowermost Llandoverian chert unit. They are overlain by ribbon limestone and flat pebble conglomerate in southeastern sections, and by limestone turbidites of the Cape Phillips Formation in northwestern sections. They are especially common in section JL 05, but absent from most other sections.

The dolostones are massive and dolomitization has apparently destroyed most evidence of original bedding. However, they show features that suggest they are the dolomitized equivalents of disorganized flat-pebble limestone conglomerates (see section 3.3.1.6). Like the flat-pebble conglomerates overlying them, they are sparsely fossiliferous: only crinoids, brachiopods and sponges were recognized. The fragmented nature is usually clear on polished surfaces and in thin sections, but indistinct on weathered surfaces and easily mistaken for a nodular structure.

There are two principal matrix types. The most common consists of anhedral, medium to coarsely crystalline dolomite with little or no interstitial bitumen and therefore a closely fitted texture (see Fig. 3.8 and Plate 5-1). The dolomite is fairly clear and "cement-like", but contains patches of even coarser and clearer dolomite spar that make up 10-15 percent of the rock, and resemble vug-fillings. The centres of some of the larger patches contain euhedral dolomite crystals with interstitial bitumen. A second common matrix consists of bituminous, usually finely to medium crystalline, subhedral dolomite with fewer patches of coarser dolomite.

Clasts in the dolomitized conglomerates and breccias differ from the intraclasts of limestone conglomerate in being generally smaller, although this may be due merely to most being identified in sectioned specimens rather than in field exposures. The tabular clasts also differ in being more angular and having sharper contacts with the matrix (see Fig. 3.8 and Plate 5-1). They are composed of unfossiliferous, very finely to medium crystalline (usually finely crystalline) dolomite with as much as 20 percent interstitial bitumen (average 10 percent). Some have internal lamination defined by lenses of bitumen; some are cut by internal microstylolites that do not extend into the matrix; some have coarse dolomite spar-filled fractures, also restricted to the intraclasts.

Less commonly, the intraclasts are subrounded, or very irregular. Non-bituminous dolostone intraclasts and bituminous chert intraclasts in bituminous finely crystalline dolostone matrix are rare.

Another type of dolomitized conglomerate occurs rarely in unit JL 05-13 (see Fig. 3.9). The matrix is mostly a very bituminous mosaic of very finely euhedral dolomite crystals. Bitumen content varies from less than 5 to 20 percent in alternating laminae, averaging 10 percent overall. The bitumen in the richer laminae is segregated as thin wisps. The not too common lens-shaped intraclasts that they surround were probably molded into such shapes post-depositionally. Even though they are similarly constituted of euhedral to subhedral very finely crystalline dolomite, they are much less bituminous than the surrounding matrix. Deposits similar in composition to the intraclasts underlie these dolomitized conglomerates, separated by a sharp contact. These presumed source deposits are commonly the top member of a fining-upward sequence that started with dolomitic conglomerate.

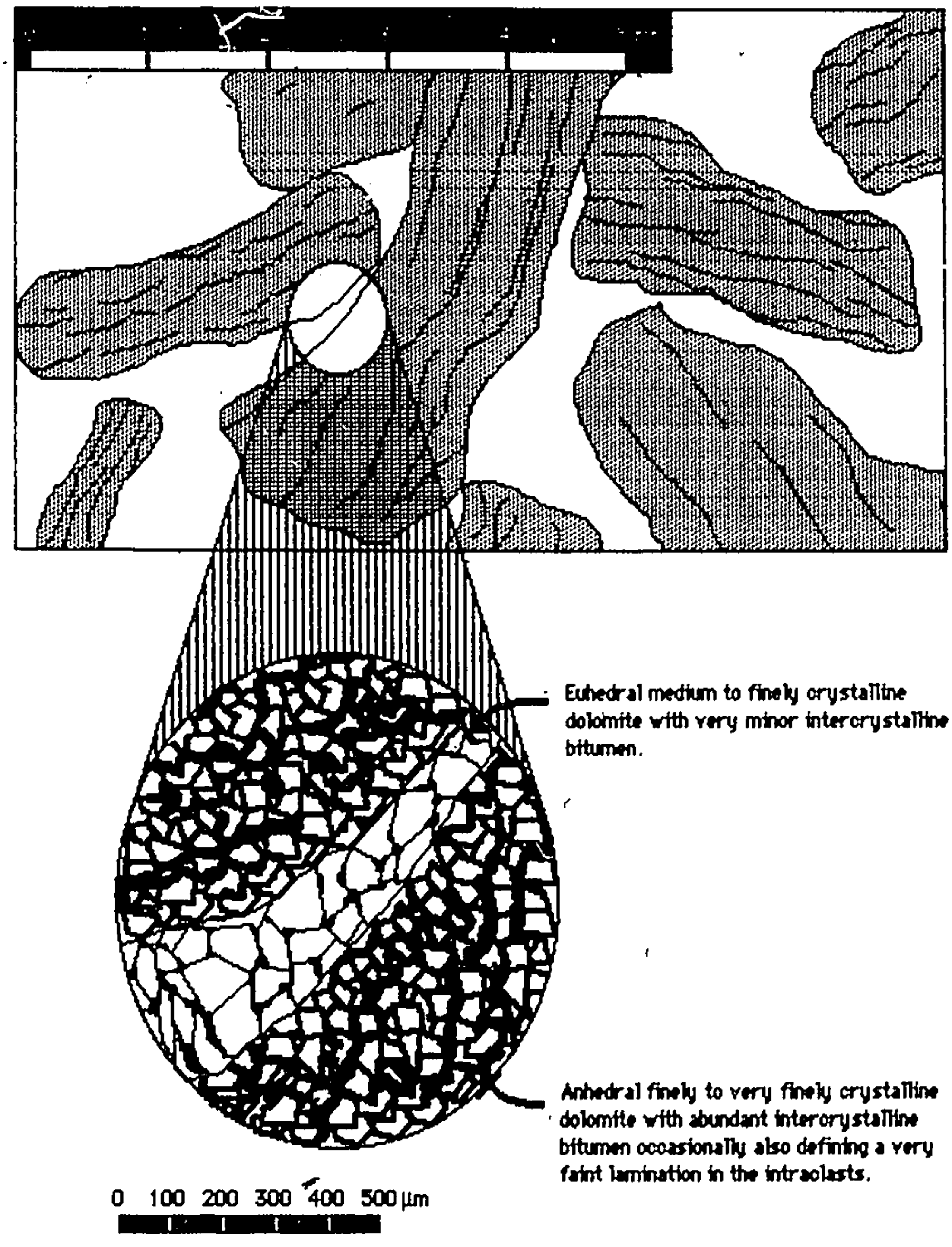


Figure 3.8 Diagrammatic representation of dolomitized breccia in the Llandoverian of the Allen Bay Fm (section JL 05, unit 15).

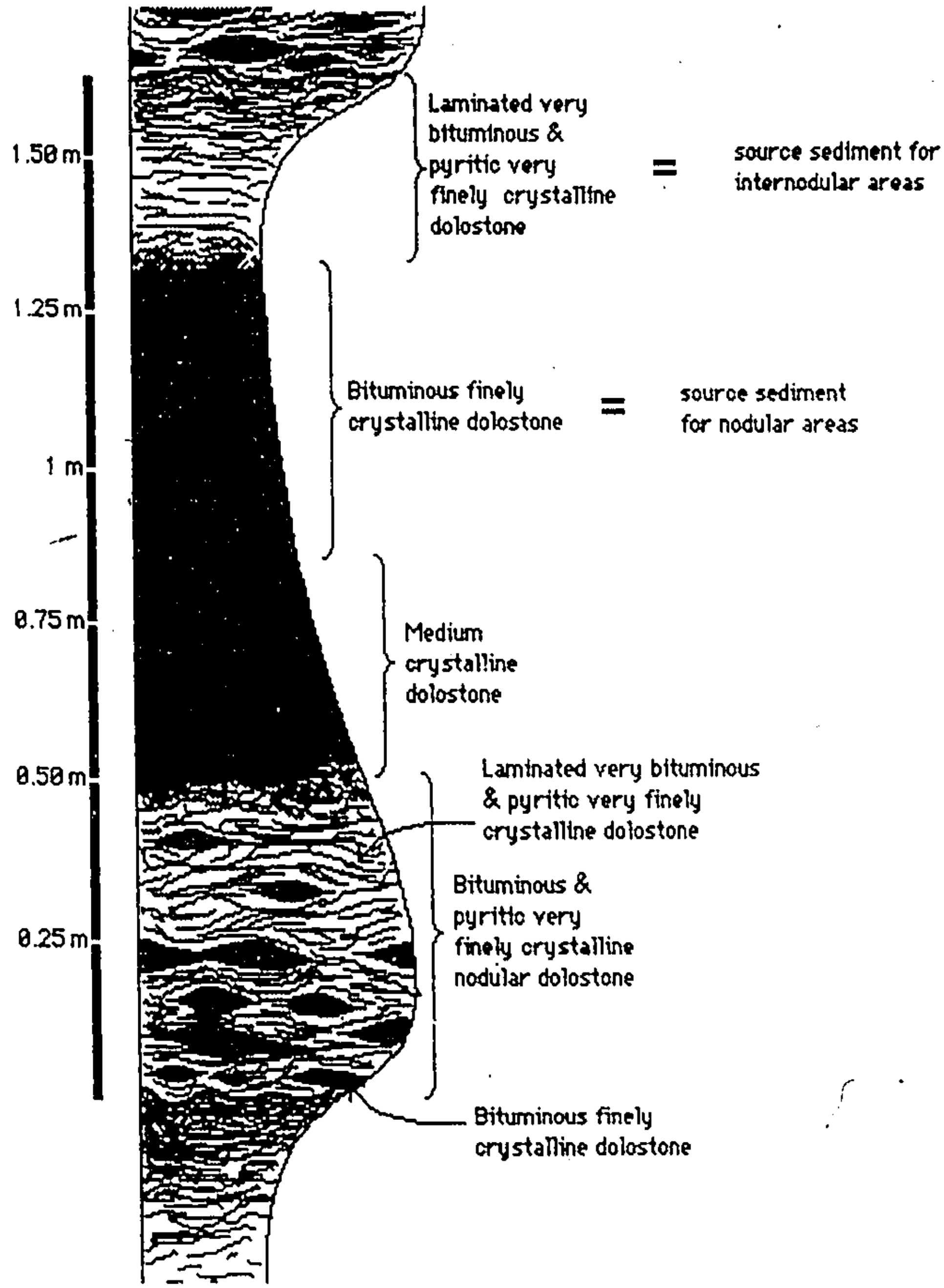


Figure 3.9 Nodular dolostone, possibly dolomitic breccia, peculiar to the lower Allen Bay Fm, and its possible origin.

3.3.4 Argillaceous rocks

Argillaceous rocks or marlstones are the main lithological group in the Cape Phillips Formation, especially in studied sections furthest to the northwest. In that direction it is also evident that the Allen Bay Formation-Cape Phillips Formation contact becomes gradually older. In some prior studies (Norris, 1963; Tozer, 1963; Kerr, 1976) these argillaceous rocks were mistakenly identified as shales, but careful petrographic examination indicates that they can be more correctly termed marlstones. Intervals of these lithologies that are abnormally rich in graptolites have been termed the "graptolitic shale facies" of the Cape Phillips Formation (Tozer, 1963). These are called graptolitic-calcareous marlstone in this thesis. Usually calcite and/or dolomite, and even bitumen contents exceed clay mineral content.

Although the marlstones are distinctly laminated and fissile, clay minerals vary in importance in determining fissility, and locally clay minerals are virtually absent. Several factors can be responsible for the lamination. Firstly, bitumen is fairly abundant in these rocks and constitutes up to 40 percent of rock volume, increasing generally up-section. Where bitumen content is more than 10 percent, in part or all of a bed, the bitumen is at least partly segregated into thin, wispy laminae (10 × 100 μm) in addition to its usual intergranular or intercrystalline distribution. Where even more abundant, the laminae coalesce horizontally although not as perfectly planar sheets along which the rocks tend to split. The different distributions of bitumen content are not usually evident in the field

Secondly, silt content can also contribute to a laminated appearance. Silt-size (rarely sand- or clay-size), angular quartz grains (feldspar grains are also common) are generally distributed randomly throughout these marlstones. However, local concentrations of these grains in minute, elongate lenses, or more rarely, in very thin laminae, result in a laminated structure. These tend to be concentrated in basal parts of what appear to be fining-upward sequences. Like bitumen, quartz grains increase in abundance towards the top of the Cape Phillips Formation. In other study areas (e.g. Cañon Fiord; Trettin, 1977), quartz is abundant enough to form siltstone or sandstone beds.

Thirdly, lamination can be the result of alternation, on a millimeter scale, of calcareous and dolomitic marlstone, the latter generally younger and increasingly abundant towards the Devon Island contact.

Finally, the laminated appearance of most marlstones is due to alternation of pure, massive lime mudstone or very finely crystalline dolostone laminae with more argillaceous laminae. Such deposits probably represent the finest and most distal portions of fining-upward sequences. Although the transition from purer lime mudstone or dolostone to more argillaceous marlstone appears abrupt in outcrops, the clay content increases gradually upward from the purer lime mudstone into the fissile argillaceous deposits. Where the argillaceous content alternates repeatedly, some of the purer carbonate interbeds have been deformed into boudin-like structures.

This small scale cyclicity of purer carbonates and argillaceous deposits occurs in many different settings. It forms regular, monotonous "banded" deposits several 10's of meters thick in most of the northwestern sections. These banded deposits can be grouped into a larger scale cycle in which argillaceous beds and/or bitumen content increase in proportion upward. In the southeastern sections, such cyclic carbonates and argillaceous deposits are much less important proportionally, but occur in the topmost portion of some fining-upward sequences. Some argillaceous deposits enclose an entire olistostrome or a single reefal block, and in these occurences constitute an important lithology.

Other rarely occurring components of these rocks include generally very well preserved graptolites, phyllocarids, trilobites, brachiopods, colonial corals (such as favositids and heliolitids), solitary rugosans, orthoconic cephalopods, bryozoans and sponges (detached spicules and complete specimens), mostly transported in but some of which can be considered to be *in situ*. In the extremely bituminous deposits of the northwestern sections, pyrite nodules (now hematite) are quite common and very thin layers of siderite or sphalerite are present locally. Calcareous concretions are also present, either as small swellings of purer carbonate layers (see Fig. 3.5) or as huge (up to 60 cm. in diameter) "cannon balls" found only in a few of the northwestern sections, especially JL 16 (see Plate 7-1).

Burrows are also present, but are more conspicuous in the purer carbonates than in interbedded marlstones. Nevertheless, different burrow assemblages are apparent. In lime mudstone layers, burrows are predominantly vertical (*Skolithos* sp., *Thalassinoides* sp., *Arenicolites* sp.), but in marlstone layers more are horizontal (*Phycodes* sp., *Planolites* sp., *Chondrites* sp., *Terebellina* sp., *Thalassinoides* sp.). The possible reasons for this difference will be discussed below.

3.3.4.1 Bituminous, calcareous marlstone

Bituminous, calcareous marlstone is the most common type of marlstone present, especially in older parts of the stratigraphic sequence, and increasingly towards the northwest. It is commonly interbedded with lime mudstone or other types of limestone in monotonously repeated cyclic couplets of generally less than 15 cm in thickness, some of which are spectacular as in JL 04 (see Plate 4-4). The proportion of marlstone to limestone varies greatly from one section to another. The interbedded limestone is generally quite massive, in contrast to the fissile marlstone. This marlstone is also typically the topmost lithology in different types of mass flows.

Fissility and lamination in the bituminous, calcareous marlstone are probably the combined effect of variations of clay mineral content, and high bitumen content. The large amount of bitumen, mostly as thin wispy laminae, tends to obscure the presence of clay minerals. The less bituminous and less argillaceous interbedded lime mudstone has a calcareous matrix which shows more neomorphic aggradation than the typical microspar matrix in the marlstone beds.

Very few fossils are evident in these marlstones. Graptolites are the commonest and some trilobites occur locally. An unusual trilobite concentration in a bed in JL 16-07 consists entirely of fragments of *Pseudogygites* covering every cleavable surface. Ninety percent of the fragments are pygidia, eight percent cephala, some with their genal spines, and only two percent thoracic fragments. Well preserved brachiopods, sponges, orthoconic cephalopods and colonial corals, especially favositids, are also present rarely. Placoderm (fish) plates and phyllocarids are rare and occur mostly in lime mudstone interbeds.

Pyrite (and hematite) is distributed mostly as small globular to lensoid nodules not more than 1 cm in length. Exceptionally well preserved, pyritic graptolites also occur very rarely. The most bituminous and pyritic beds, in the most northwestern sections, contain, extremely rarely, thin beds and laminae of a brownish, earthy-textured mineral, possibly derived from sphalerite.

3.3.4.2 Graptolitic, calcareous marlstone

Graptolitic, calcareous marlstone occurs mostly in the more northwestern sections, and is usually bituminous, pyritic, argillaceous and very fissile. This lithology is more abundant in Wenlockian and Ludlovian portions of these sections. It resembles shale in being highly fissile and more recessive than other marlstones, and has been included in "graptolitic shale facies" of the Cape Phillips Formation by some workers (Norris, 1963; Tozer, 1963; Kerr, 1976). Most examples of graptolitic marlstone are bituminous, calcareous marlstone, but, consistent with previous work, they are set apart because graptolites are conspicuous components and some units are not bituminous.

Graptolites are the predominant or exclusive fossils in these rocks; other types of fossils are rare. Although the graptolite fauna overall is diverse, particular parts of sections are commonly dominated by only one or two species. Most specimens are only thin carbon films on similarly dark bedding surfaces and are therefore very hard to distinguish. Exceptionally well preserved, 3-dimensional, calcite-filled specimens, occur locally as well as some extremely rare pyritized specimens in very bituminous strata.

3.3.4.3 Silty, calcareous marlstone

Silty, calcareous marlstone occurs mostly in the upper portions of the sections. Silt content increases generally up-section, is especially conspicuous in Ludlovian and Pridolian strata, and reaches a maximum at the Cape Phillips-Devon Island formational contact. However, the trend is not gradual and continuous through the entire Ludlovian and Pridolian strata. Sections with the greatest amount of silt lie between the thick sections of the southeastern corner of the study area and the most northwestern sections.

Although predominantly silt-size, the grains vary from clay to fine sand in size. These angular to sub-angular grains are mostly quartz but feldspar (both K-spar and plagioclase) is also common. Associated with these grains are miniscule laths (2 × 10 μm), possibly of gypsum or anhydrite, generally oriented parallel to bedding. Their small size and colour close to the background colour of the rock make it difficult to estimate their proportions compared to clay minerals.

The silt is generally well sorted, but only moderately sorted in some deposits. Where lime mudstone is interbedded, it generally contains more silt and the silt decreases gradually up-sequence into marlstone.

Silt grains are distributed randomly throughout most beds although some show decrease or increase upward. The silt is also concentrated locally in small lenses (~ 100 × 500 μm) mostly, but not entirely, in the basal portions of beds. These lenses have abrupt, although not erosional, contacts with surrounding sediments in having calcite spar cement rather than matrix in intergranular spaces, even though the quartz grains, forming 20 to 60 percent of the lenses, scarcely appear to be grain-supported. Silt also forms basal lag deposits of some marlstone beds although this is more typical in lime mudstone interbeds. Like the lenses, these very thin silt laminae exclude the marlstone matrix.

Silty calcareous marlstone contains a moderate amount of clay minerals, and minor silt and bitumen, but by far the main components is micrite and to a lesser extent minute dolomite crystals. The micrite shows progressively greater neomorphic aggradation in more silty rocks and where silt lenses and laminae contain some micritic material.

Allochems generally occur very rarely, but are more common, especially peloids and ostracods, in the silt lenses and laminae. The fauna is generally much more diverse and evenly distributed than in other types of marlstone and not simply dominated by graptolites. Ostracods are fairly important and also present are whole sponges, favositid corals, small orthoconic cephalopods (generally not more than 10 cm in length), and a wide variety of well preserved brachiopods. The graptolites present are generally much more poorly preserved than in other marlstones.

3.3.4.4 Bituminous dolomitic marlstone

Bituminous dolomitic marlstone superficially resembles bituminous calcareous marlstone, but differs in several respects. First, very fine dolomite crystals predominate over micrite. The dolomite crystals are generally less than 30 μm in size, averaging 15 to 20 μm, but as small as 5-10 μm, and are often altered at their edges. In some deposits the remaining calcitic micrite masses (~ 30 μm in diameter) are distributed such that an homogeneous fabric is kept. Secondly, these dolomitic marlstones are much less argillaceous than bituminous calcareous marlstone. The laminated and fissile nature of such deposits is solely dependent on the bitumen content and reflects alternation of darker layers, containing most of the bitumen laminae, and light layers with much less bitumen. The bitumen content is the highest among the marlstones.

This type of marlstone is interbedded with very bituminous, very finely crystalline dolostone and chert of the lower Llandoverian marker unit. It also occurs with gradually increasing frequency upward in the upper parts of these sections, as was the case with the silty calcareous marlstone. Above the Cape Phillips Formation, bituminous dolomitic marlstone is common in the Devon Island Formation and is apparently the predominant marlstone type.

In the upper parts of these sections, calcareous and dolomitic marlstone are commonly interbedded with dolomitic marlstone, increasing in proportion up-section. The boundaries between the interbeds are typically gradational. Each dolomitic marlstone grades upward and downward (Fig. 3.10) into calcareous mudstone and the couplets possibly represent repeated episodes of dolomitization.

Pyrite nodules occur in some of the more highly bituminous dolomitic marlstone. Of all the marlstones, this type has the lowest amount of allochems, especially fossils. Only a few graptolites were recognized.

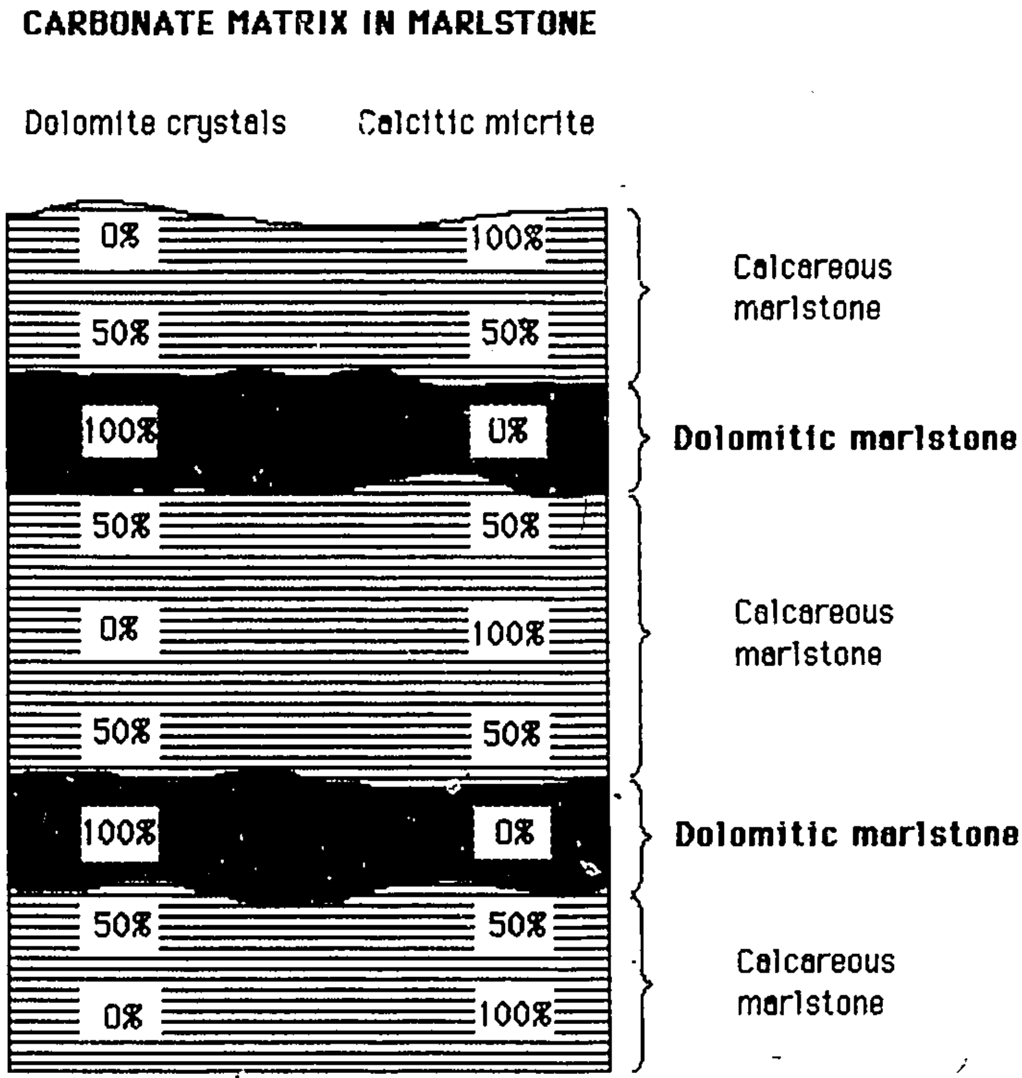


Figure 3.10 Diagrammatic representation of interbedded calcareous marlstone and dolomitic marlstone, such as the ones found in Ludlovian strata of the Cape Phillips Fm of most sections, with gradual transitions from a micritic to a dolomitic matrix. The dolomitic marlstone also contains more bitumen.

3.3.4.5 Silty, dolomitic marlstone

Some beds of silty, dolomitic marlstone occur in the top portion of the Cape Phillips Formation but this marlstone is more typical of the Devon Island Formation. It is the least important volumetrically of all the marlstone types.

Silt content and distribution are similar in these and the silty calcareous marlstone, and silt-sized, angular quartz and feldspar grains are either dispersed randomly, or concentrated in lenses, or in multiple lenses, forming very thin laminae of silt constituting 3% to 40% of the deposit. The lenses generally diminish in abundance up-sequence. Bitumen content can be as high as 30 percent but rarely is it distributed as thin, wispy laminae. As a consequence, this marlstone is rarely laminated or very fissile since clay mineral content is also very low.

The same fossils that appear in the bituminous, dolomitic marlstone, i.e. graptolites, are present here. However they are not as well preserved as in the calcareous marlstone. Trace fossils are present, and in some beds are very abundant. The commonest are *Phycodes* sp., *Planolites* sp., and *Chondrites* sp., of which several have various diameters and are of multiple generations.

3.3.5 Siliceous rocks

Siliceous rocks are a significant rock type only in the lowermost Llandoverian marker unit. However, other rock types, particularly the coarser-grained and purer limestones and dolostones, are partly silicified.

Siliceous rocks have been divided into two major groups: a) those pervasively silicified, and b) those selectively silicified. Examples of pervasively silicified rocks are the nodular and bedded cherts of the lowermost Llandoverian marker unit mentioned above. Within otherwise homogeneous dolostones and limestones, chert nodules occur at particular stratigraphic levels and some coalesce to form discontinuous beds. In the northwestern sections, entire planar beds and laminae of chert occur in the Cape Phillips Formation.

Selective silicification can range from very minor, partial replacement of one or a few fossil groups, to extensive replacement of most allochems, cements and matrix. This type of silicification, although generally insignificant in proportion and usually recognizable only petrographically, nevertheless importantly affected limestones in the southeastern sections.

The following four major types of silica have been recognized:

1) microcrystalline quartz — a rarely occurring isopachous mosaic of crystals filling minor pores
2) chalcedony — somewhat less rarely occurring, pore-filling cement consisting of elongate, subparallel crystals in concentric zones distinguished by slight colour differences
3) cristobalite (also known as porcelanitic chert or lussatite) — minute, disordered lepispheres that are volumetrically the most important, forming most of the pervasively silicified rocks and replacing the matrix and some allochems in selectively silicified rocks
4) "flame-type" chalcedony — with fibrous crystals much less consistently subparallel and equi-crystalline than in type (2). This type is intimately associated with, and as important as, cristobalite in replacing cements and fossils in selectively silicified rocks, but doesn't occur in the pervasively silicified rocks studied.

3.3.5.1 Nodular and/or bedded chert

Nodular and/or bedded chert is an important lithology of the lowermost Llandoverian marker unit, whether the unit consists mostly of very bituminous, very finely crystalline dolostone or bituminous, micritic lime mudstone, and is the most abundant type of chert in the study area. Very little of this type of chert is present above and below this marker unit. There is more nodular and/or bedded chert in the southeastern sections but this chert forms a higher proportion (up to 30% of some units) of the marker in the northwestern sections. This type of chert, except for the smaller nodules in the younger sediments, is usually very bituminous, similar to the enclosing, mostly very finely crystalline dolostone. Where nodular chert is abundant at particular stratigraphic levels, the nodules are commonly coalesced to form beds. This is much more common in the northwestern sections than in the southeastern ones. While the presence of chert beds indicates higher chert contents at particularly horizons, it does not necessarily reflect a higher silica content in the entire unit. For example, in some sections the silica content dramatically increases up-sequence (e.g. from 2% to 15%), while chert beds are found only in the bottom portion.

Also, much smaller lensoid nodules of chert (1 × 15 mm to 0,5 × 3 cm lenses) occur occasionally in younger strata, either in lime mudstone interbedded with minor marlstone, or in some ribbon limestone beds, or in upper portions of limestone turbidites, or fragmented in some mass flow deposits.

Except for the smaller nodules in the younger sediments, this type of chert is usually very bituminous, similar to the surrounding very finely crystalline dolostone. However, different weathering processes, can result in the chert nodules appearing either lighter or darker than the surrounding dolostone. No appreciable differences in degree of silicification and/or bitumen content appear to be the cause of the different weathering colours.

In addition to chert intraclasts that occur rarely in some breccia and conglomerate beds, some chert beds and nodules of the marker unit (JL 05-07, JL 01-21 and JL 10-20) are fractured, but the fragments are not extensively displaced and the host sediments show no evidence of fracturing.

This type of chert differs from the others in lacking fossils, except for the occasional graptolite. The chert and its enclosing sediments are the least fossiliferous deposits of the entire study area.

The silica composing the chert is mostly structureless, amorphous, cryptocrystalline cristobalite. Some of the chert contains an important amount of randomly distributed very small dolomite rhombohedra (less than 30 μm in size). This dolomite content consistently increases from the centres towards the edges of the chert bodies (see Plate 8-2).

Contacts between silicified areas and host sediments are of two types. Generally the contacts are gradational and defined by very gradual increase in dolomite crystals and decrease of silica crystals. The bitumen content does not generally vary significantly, but locally increases along with dolomite content. This transitional zone varies in thickness from approximately 1 mm to 1 cm.

Nodule contacts appear rarely abrupt due to enhancement by an enclosing layer (up to 0,5 cm in thickness) of very pure and coarse, white, mosaic dolomitic sparite that contrasts strongly with the black or dark brown background (see Fig. 3-11 and Plate 8-1). These enclosed chert nodules are separated from the sparite by transitional zones very similar to those surrounding nodules with gradational contacts. Bituminous, very finely crystalline dolostone with no chert surrounds the sparite layer. This host lithology is homogeneous except for a much higher bitumen content closer to the sparite. Bitumen content, however, gradually decreases to average amounts, i.e. approximately 10 percent of the total volume, within 1 or 2 cm of the contact.

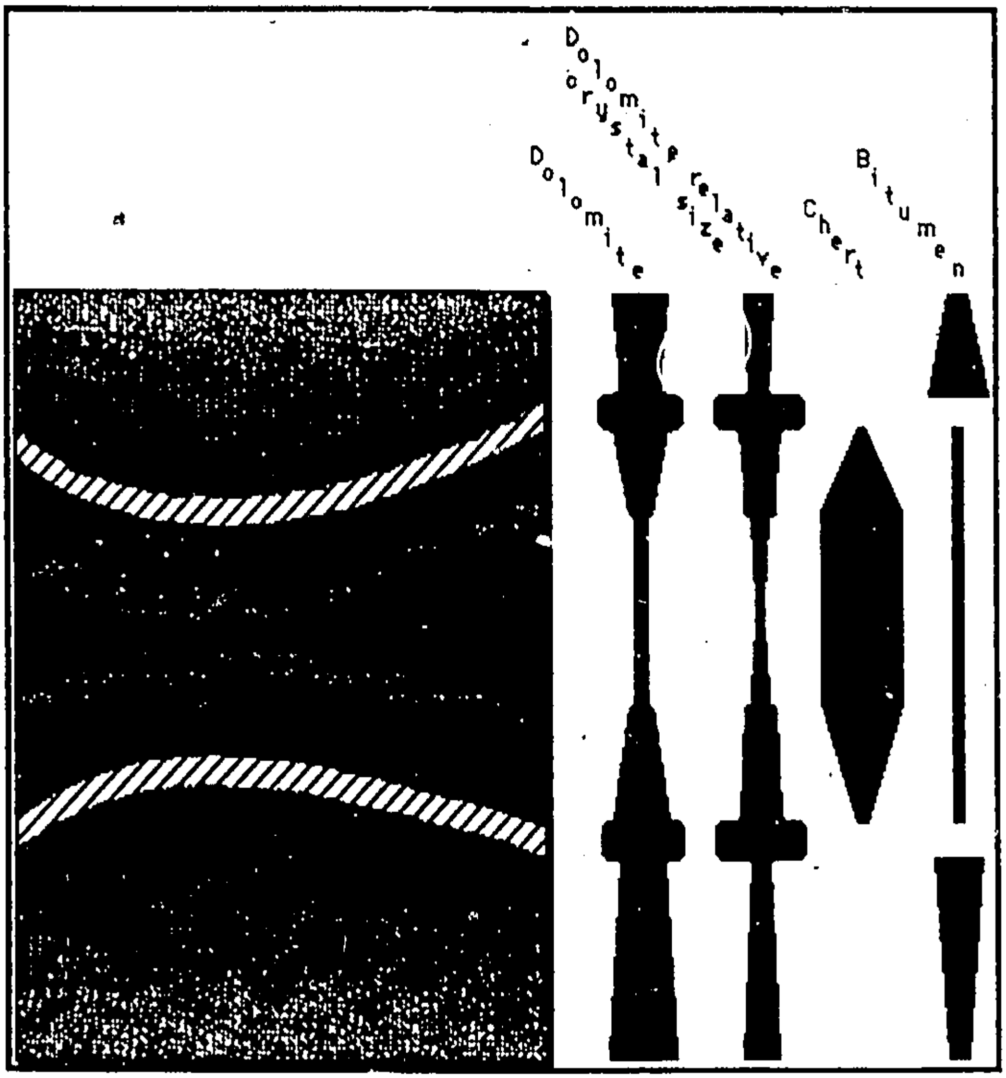


Figure 3.11 Diagrammatic representation of a bituminous chert nodule such as the ones found in the dolostone of unit 21 of section JL 01, showing the different zones. The distribution and relative abundance of components are represented to the right of the cross section.

The above compositional changes are usually accompanied by colour changes partly related to bitumen content. Three main zones exist: a) a black chert, b) a dark brown transition, and c) a lighter brown, very finely crystalline, dolomitic host sediment. Where the sparite shell is present, it encloses the two first zones. Occasionally, variation in degree of silicification produces colour zonation within the chert nodules.

Where the host sediment is bituminous, micritic lime mudstone, the chert nodules are coarser microcrystalline quartz rather than amorphous cristobalite. These quartz crystals are homogeneously distributed with micrite up to a transitional zone along the periphery of the nodule. At that point, the micrite content increases gradually towards the edge of the nodule, but although such a transitional zone is present, the contact is nevertheless abrupt.

Very rarely central portions of chert nodules, which are usually amorphous cristobalite, have been transformed to coarser, lighter-coloured "flame-type" chalcedony. Associated bituminous material is concentrated in lenses (on average 100 × 200 µm) and as a thin film surrounding the chalcedony core and separating it from the surrounding cristobalite.

A distinctive thin bed of pure blue chert occurs in unit JL 07-08, in the middle of what is apparently a small, downslope, coral-framework reef. The diverse *in situ* coral colonies are also selectively silicified.

3.3.5.2 Selectively silicified rocks

In contrast to the nodular and bedded chert, selectively silicified rocks are much more widely distributed stratigraphically. This silicification, as well as pervasive silicification (below), increases upwards from the lower or middle Ludlovian units in all sections, except in JL 01 where the increase ends approximately at the assumed contact between the Allen Bay Formation and the Read Bay Group.

Usually only certain minor, components are silicified, most evident from petrographic examination. However, examples can be recognized of all stages of progressive silicification up to completely silicified rocks. These pervasively silicified deposits, however, also contain portions parallel to bedding, but not necessarily separated by bedding planes, which were left untouched by silicification. The contacts are extremely abrupt, so that completely silicified rock is in direct contact with rock of the same deposit, still 100 percent calcite. Good examples are pelsparitic grainstones (see Plate 8-3). No break in the homogeneity of the deposit appears to control the distribution of the silicification front.

Original components that have been silicified remain clearly distinguishable because of size, shape and orientation of silica crystals in different components. For example, fine micritic calcite matrix has been usually transformed to very fine, almost amorphous, cristobalite, while cement is generally replaced by "flame-type" chalcedony. Among allochems, silicified peloids are typically cristobalite that is darker in colour than the matrix. Intraclasts can be a mixture of cristobalite and "flame-type" chalcedony but differ from peloids in their irregular shapes. Different groups of fossils can be distinguished, partly by shape and partly by differences in replacement silica. Hence, silicified crinoids are mostly cristobalite, while trilobites and brachiopods are mostly fibrous chalcedony. The outlines of most silicified allochems are further enhanced by thin bitumen coatings.

There is evidence of a general order of silicification related to different susceptibilities of different components. Fossils were apparently more susceptible and tend to have been silicified first and to greater extents; matrix, intraclasts, cement and peloids show progressively less evidence of silicification and presumably were less susceptible to diagenesis. Among the different fossil groups, sponges and bryozoans are almost always silicified; crinoids, brachiopods and calcareous algae are also commonly silicified; ostracods, solitary rugosans and colonial corals are not commonly silicified; trilobites are rarely, and then only partly silicified. However, surprisingly, even in extensively silicified deposits, individual spicules, probably from sponges, consist of a mosaic of coarse, sparry calcite.

Intraclasts are rarely entirely silicified, and generally only outer rims of various widths were affected. In unit JL 16-21, smaller intraclasts are almost completely silicified, while larger ones are at most only partly silicified peripherally.

Less commonly a different order of silicification can be recognized. In some fine limestone turbidites, a fine calcareous matrix and/or cement but not allochems have been silicified (see Plate 8-5). Therefore, since generally the fine fossil fragments are concentrated in the bottom portion of these limestone turbidites, the silicification increases upward as seen in JL 16-29. In other coarser-grained fining-upward sequences, only patches of matrix were silicified, some bordered by allochems but others more commonly in sharp contact with unaffected matrix. In some lime mudstone, silicification can be affected as much by position as by type of components. There are examples where progressive silicification of matrix was blocked by large allochems. In addition, in lime mudstones containing lenses of coarser packstone or grainstone, allochems in the lenses can be intensely silicified whereas similar allochems in the surrounding matrix are not. In larger lenses the degree of silicification decreases, in proximity to the surrounding matrix or host sediment (JL 02-24).

In other deposits, silicification evidently progressed along contacts, such as along bedding contacts where only immediately adjacent components were silicified, or along grain-matrix contacts such that only portions of fossils and matrix immediately at the contact were intensely silicified (especially the fossils), or along crystal boundaries in cements. From these contacts, silicification progressed inward and in some portions of these beds, silica almost entirely replaced the components so affected.

Most examples where only fossils are selectively silicified, represent reefal blocks or deposits closely related to a reef. In coarser fining-upward sequences, i.e. debris flows, or limestone conglomerates or breccias, selective silicification of the grains and cements has resulted in bottoms that are more highly silicified. Although some silicified fossils can be recognized in outcrops, petrographic examination commonly reveals that it is only a surface, i.e. recent diagenetic phenomenon.

As stated earlier, there is clear evidence that silicification started with amorphous cristobalite and progressed, by increase in size of silica crystals, eventually to "flame-type" chalcedony. However, fossils and micritic matrix show different results. In silicified fossils, the "flame-type" chalcedony is peripheral, and the cristobalite central. In contrast, in silicified patches of micritic matrix, the coarser chalcedony is central and surrounded by cristobalite.

3.3.5.3 Pervasively silicified rocks

Individual components in pervasively silicified rocks remain distinct, either due to differences in the character of silica crystals in different components, or to films of bitumen coating allochems. Pervasive silicification tends to be much more local than selective silicification. For example, the bottom portion of a bioclastic layer can be entirely silicified, but not the top portion of the same layer, the two now emphasized by a layer of bitumen (30 - 50 μm).

It can be difficult to determine whether a completely silicified layer was selectively or pervasively silicified. However, the latter typically results in "flame-type" chalcedony throughout. Silica crystal orientation and the relocation of almost all of the bitumen around the grains efficiently preserved structures and components.

Pervasively silicified rocks occur almost exclusively in sections located in the southeastern corner of the study area (JL 01, JL 02 & JL 05). They represent fining-upward sequences, and include coarse bioclastic conglomerates or breccias, and finer bioclastic and peloid-rich limestone turbidites. Typically, silicification followed particular bedding planes from which it spread downward and upward. For example, in Unit JL 02-37, the bottom but not the top part of the basal pelmicrite is silicified. In other examples, where there was truncation of an earlier fining-upward sequence, both the top of the truncated sequence and the base of the overlying sequence are silicified. Depending on the amount of truncation, the silicified top of the truncated sequence could represent either a finer micritic upper portion or the top of a coarser basal pelsparite lag, and have beneath it an unsilicified portion. Such relationships provide useful insight into the timing and mechanism of silicification.

3.3.5.4 Laminated silicified rocks

In general, laminated silicified rocks in the northwestern sections are stratigraphically equivalent to the bedded and nodular cherts of the southeastern sections. Unlike the bedded and nodular cherts, cherts in the laminated siliceous rocks show well preserved original structures such as laminations, are typically continuous, and usually occur in limestone rather than dolostone. Laminated silicified rocks are strongly bituminous, and pyrite nodules are commonly associated with them. They contain very few allochems, even graptolites.

The laminae reflect not only an original sedimentary lamination but also show the imprint of different degrees of silicification. The silica is distributed partly as cristobalite that pervasively replaces micrite crystals but varies substantially in amount from lamina to lamina. The transition from one to another is very gradual, considering the thickness of the laminae (1 mm to 1 cm). Where individual micritic or microsparitic crystals are not entirely replaced, much smaller cristobalite crystals commonly have replaced the edges. Overall, the cristobalite forms 10 to 20 percent of the unit, enough to give it chert characteristics such as hardness and conchoidal fracture.

The laminated appearance is also enhanced by chert distributed as highly siliceous lenses (on average 100 μm in thickness but up to 500 μm) within laminae (see Fig. 3.12). These lenses alternate with very bituminous micritic mudstone, in which the bitumen is concentrated in very thin wispy filamentous lenses (10 × 100 μm) and as intercrystalline material. Concentration of these chert lenses in some laminae with very little associated bitumen produces a lighter colour; other laminae with less chert and more bituminous lime mudstone are much darker.

Highly siliceous lenses (on average 100 μm in thickness but up to 500 μm) also contribute to the laminated appearance. These lenses are separated by very bituminous micritic mudstone, in which the bitumen is concentrated as very thin wispy lenses (10 × 100 μm) and as intercrystalline material. The lenses are concentrated in some laminae and since very little bitumen is associated, they are lighter in colour than the interlaminated, dark, bituminous lime mudstone.

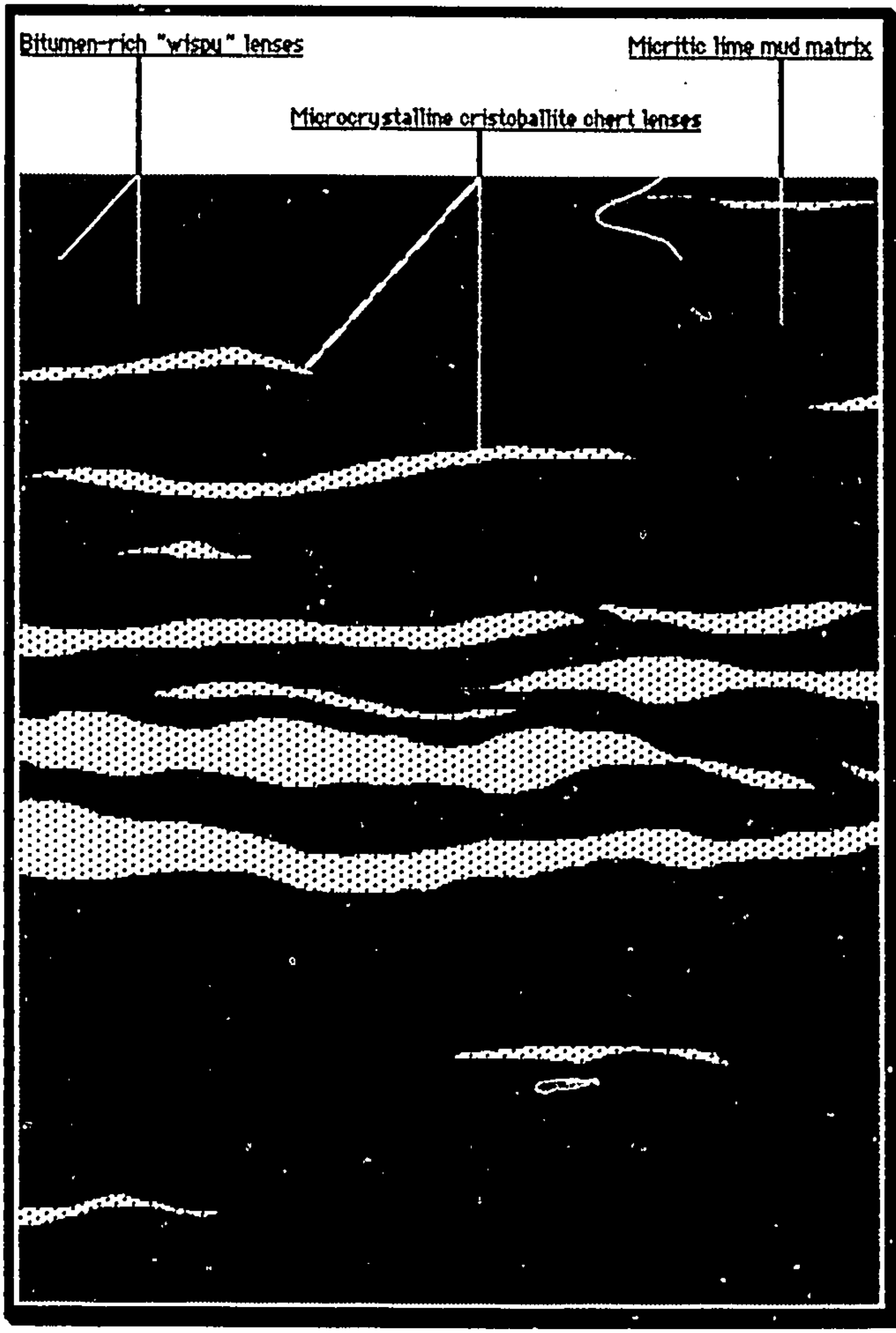


Figure 3.12 Diagrammatic representation of laminated, bituminous and cherty lime mudstone of the Ludlovian of the Allen Bay Fm.

This lensoid distribution of laminated chert probably does not reflect original lamination. In the northwestern sections, such deposits are found not only in the Llandoverian marker unit but also above it, in limestone turbidites. Some very fine limestone turbidites show a structure superficially like inverse grading, with a sharp erosional bottom contact, followed by laminated sediments that are in turn overlain by massive, homogeneous and non-laminated micritic lime mudstone. Closer inspection reveals that the bottom portions of these limestone turbidites are more silicified, with lenses of silica concentrated in some laminae as described above. The anomalous position of the laminated rock beneath a massive structureless portion, suggests that the lamination is of secondary origin. This is particularly evident in some units of section JL 16, the most northerly section.

Above the Wenlockian strata, laminated silicified deposits decrease rapidly in occurrence and volume, even in section JL 16. Only minor lensoid and/or nodular chert (usually not longer than 2 cm) occurs at the bottoms of very fine limestone turbidites.

3.4 Significance of particular lithological associations

3.4.1 Limestone conglomerates and fining-upward sequences

The most important lithological group of these sections, and ultimately the best evidence of a shelf-to-basin transition, are the limestone conglomerates and fining-upward sequences. A broad spectrum of deposits is represented, ranging from oligomictic to polymictic conglomerates, from very coarse to very fine and unsorted to well graded deposits. These are discussed and interpreted below as *debris flow deposits*, representing downslope mass flow of bed load material, and *limestone turbidites*, representing downslope transport of suspended load by turbidity currents. Sudden downslope mass movement of incompletely consolidated sediment commonly also generates turbidity currents, and as a consequence the two groups of deposits can be intimately associated both stratigraphically and geographically. Figs 3.13a→d illustrate the transition from debris flow to limestone turbidites downslope and the influence that different source sediments had on the final results. A diagrammatic representation of an ideal and complete sequence deposited on the middle to lower slope might look like Fig. 3.14.

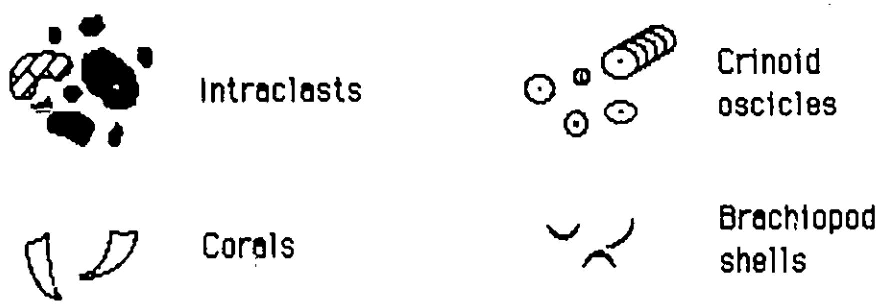
3.4.1.1 Limestone conglomerates as debris flow deposits

A debris flow is defined as "A moving mass of rock fragments, soil and mud, more than half of the particles being larger than sand size" (Bates and Jackson, 1980, p. 161). Subaqueous debris flows can be triggered by earthquakes, storms, tsunamis, etc. on any slope where instability is created by sediment accumulation.

Flat-pebble limestone conglomerates are a major group of rocks in the study area, particularly in the upper Llandoverian to mid-Ludlovian interval. Clear evidence of derivation from ribbon limestones includes the gradual transition of some ribbon limestone upward into flat-pebble conglomerate units of extremely similar lithological types and proportions. Such sequences clearly display undisturbed source material passing up into slightly moved material with limestone layers partly separated into pebbles, apparently cemented earlier and initially more cohesive than the matrix. This in turn passes gradually up into substantially moved material, although commonly with matrix-supported flat pebbles still largely horizontal (see Fig. 3.15). The conglomerate formed represents an early stage of mass movement in which essentially parallel flow was retained. It commonly grades up into a coarse, clast-rich fining-upward sequence, apparently representing closely associated turbidity current deposition. Fig. 3.16 illustrates the complete sequence.

Figures 3.13a→d (Following pages) Diagram representing 3 different types of mass flows and associated turbidity flows with their downslope gradational equivalents as they might have occured in the study area (Fig. 3.13a): a debris flow with ribbon limestone as its source (Fig. 3.13b), a coherent debris flow with ribbon limestone as its source (Fig. 3.13c), and an olistostrome with shelf-margin buildups as its source (Fig. 3.13d).

Figure 3.14 (see p. 123) Diagrammatic representation of an ideal fining-upward sequence starting with a tabular clast breccia. The corresponding bar chart roughly illustrates the relative importance of the different components, where present in the deposit. Black bars are for the components commonly associated with such deposit; grey bars, components occasionally found; and empty bars are for components rarely found. For a key to symbols found in these diagrammatic sections, see the following:



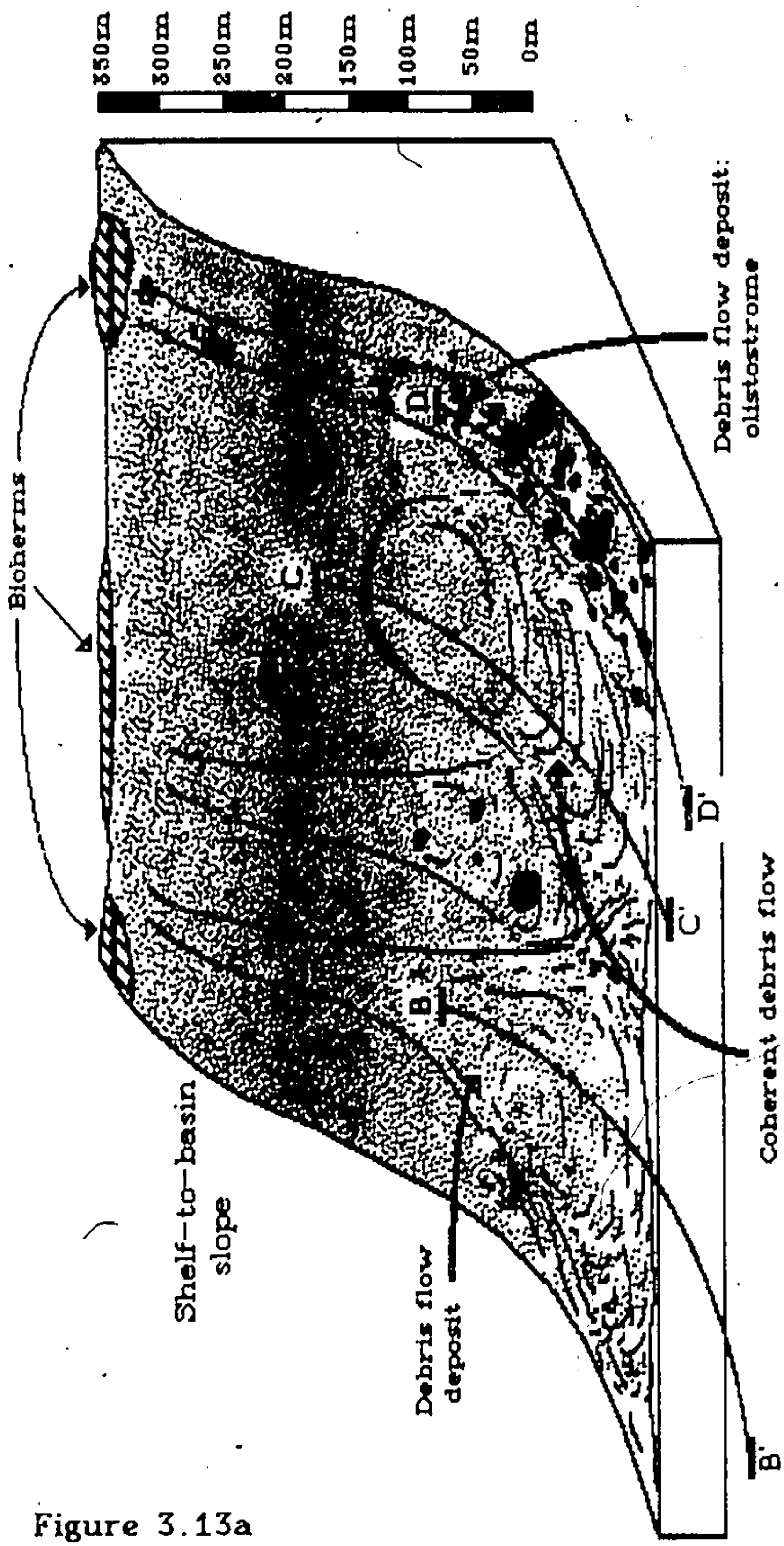


Figure 3.13a

• The Allen Bay Fm-Read Bay Group undivided and Cape Phillips Fm •

Figure 3.13b
B-B' — Gradational downslope changes in a debris flow deposit

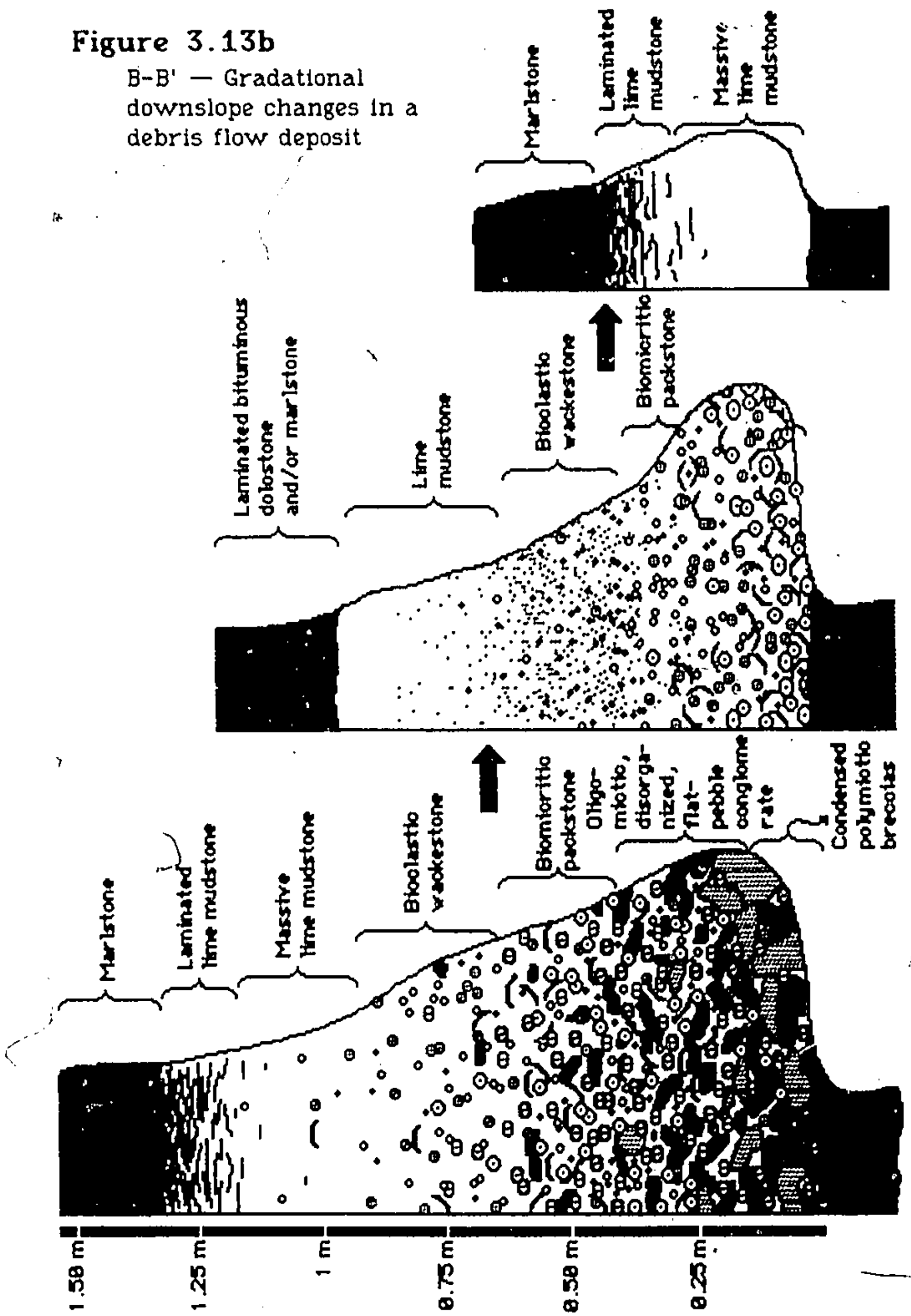


Figure 3.13c
C-C' — Gradational downslope changes in a slide deposit

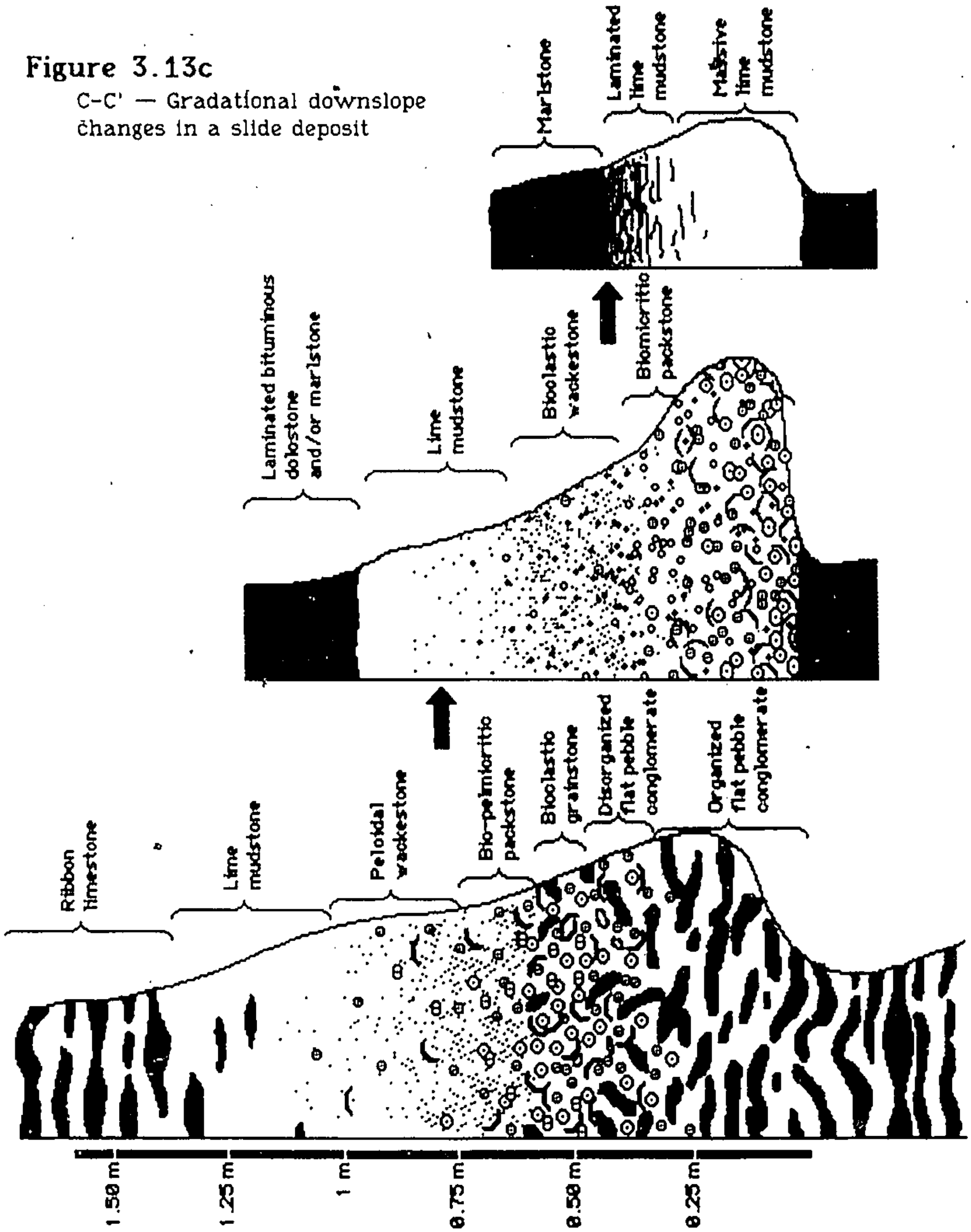


Figure 3.13d

D-D' — Gradational downslope changes in an olistostrome

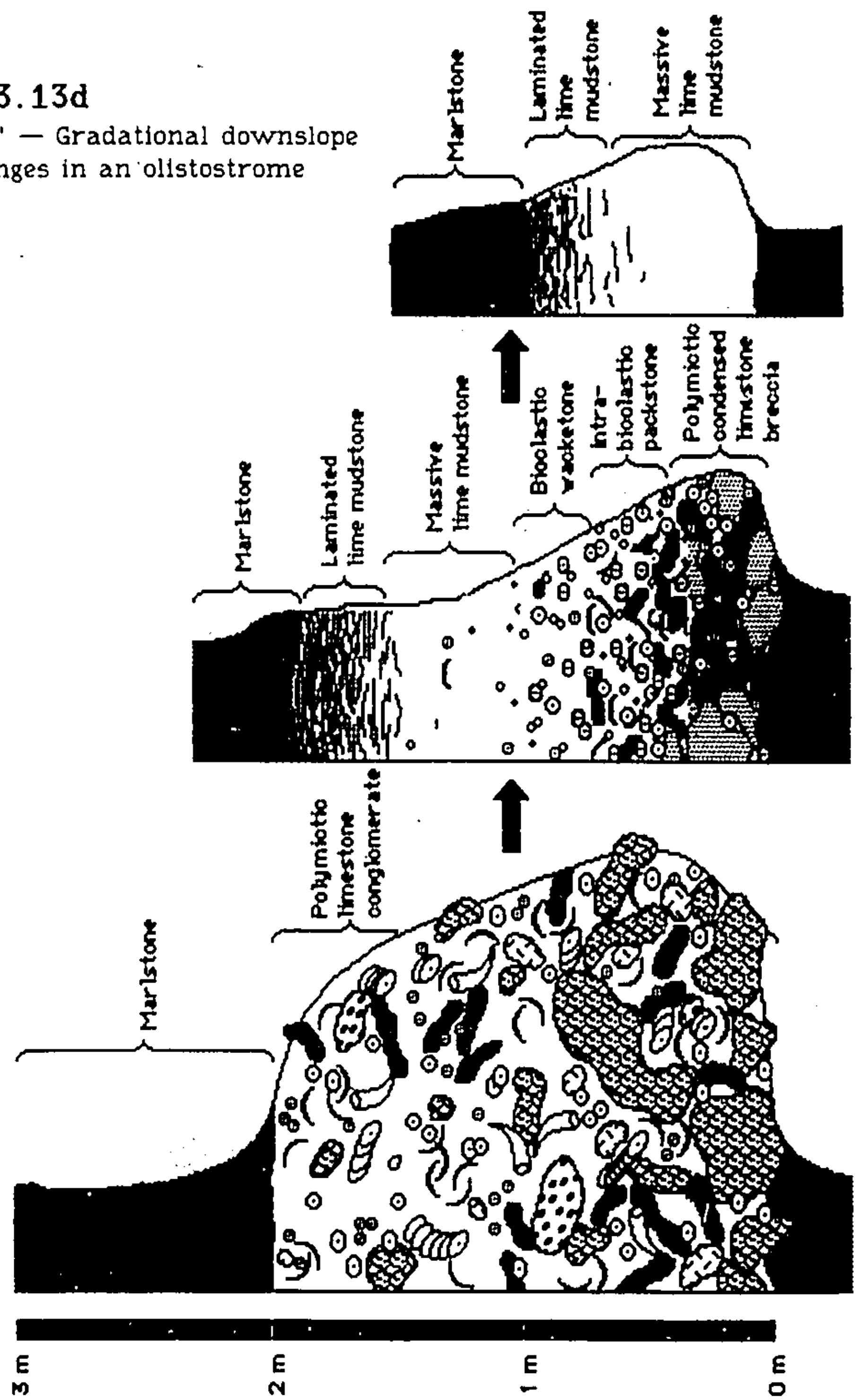
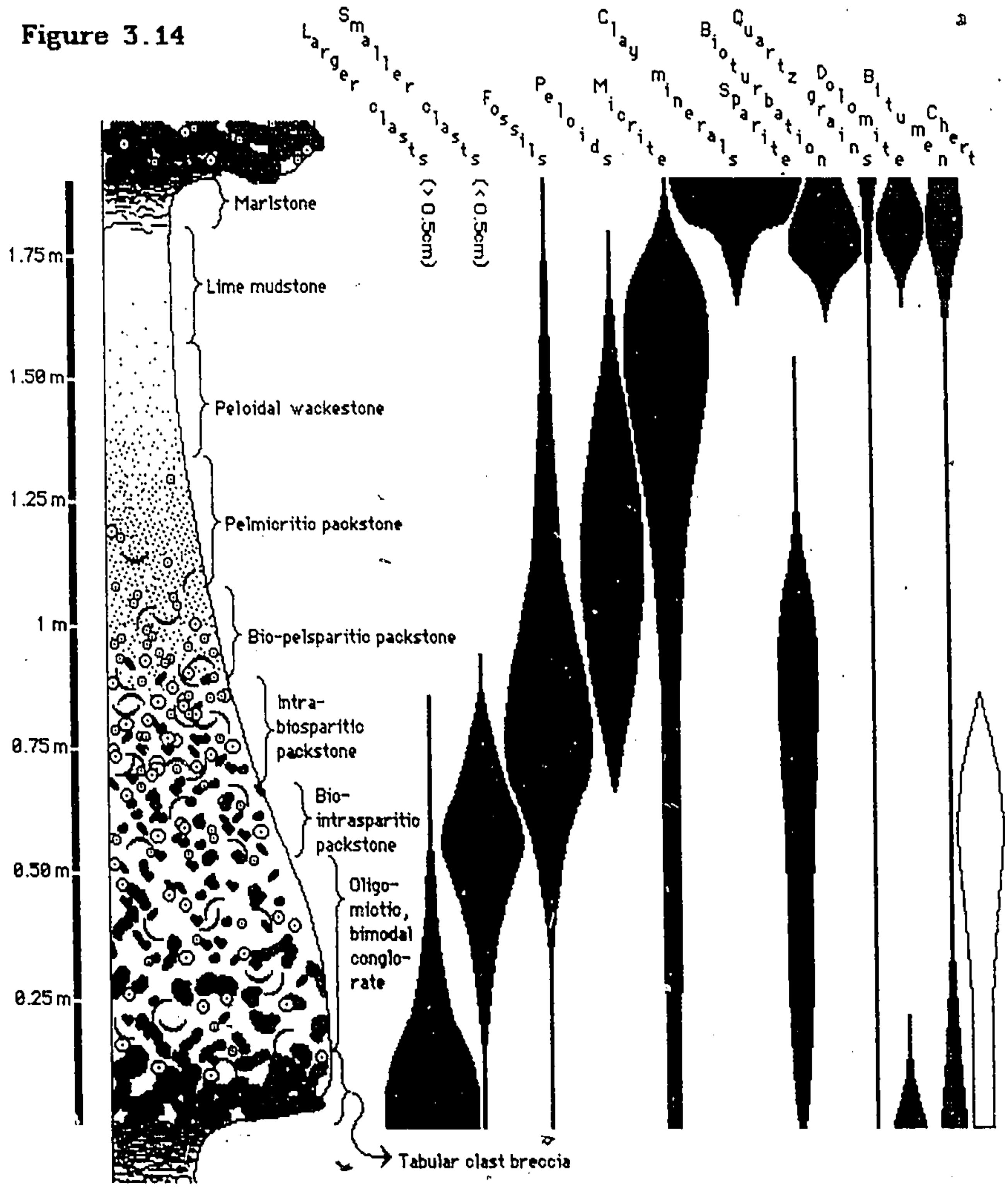


Figure 3.14



Larger clasts (>0.5cm)
Smaller clasts (<0.5cm)
Fossils
Peloids
Micrite
Clay minerals
Sparite
Bioturbation
Quartz grains
Dolomite
Bitumen
Chert
1.75 m
1.50 m
1.25 m
1 m
0.75 m
0.50 m
0.25 m
Marlstone
Lime mudstone
Peloidal wackestone
Pelmicritic packstone
Bio-pelsparitic packstone
Intra-biosparitic packstone
Bio-intrasparitic packstone
Oligomictic, bimodal conglo-rate
Tabular clast breccia

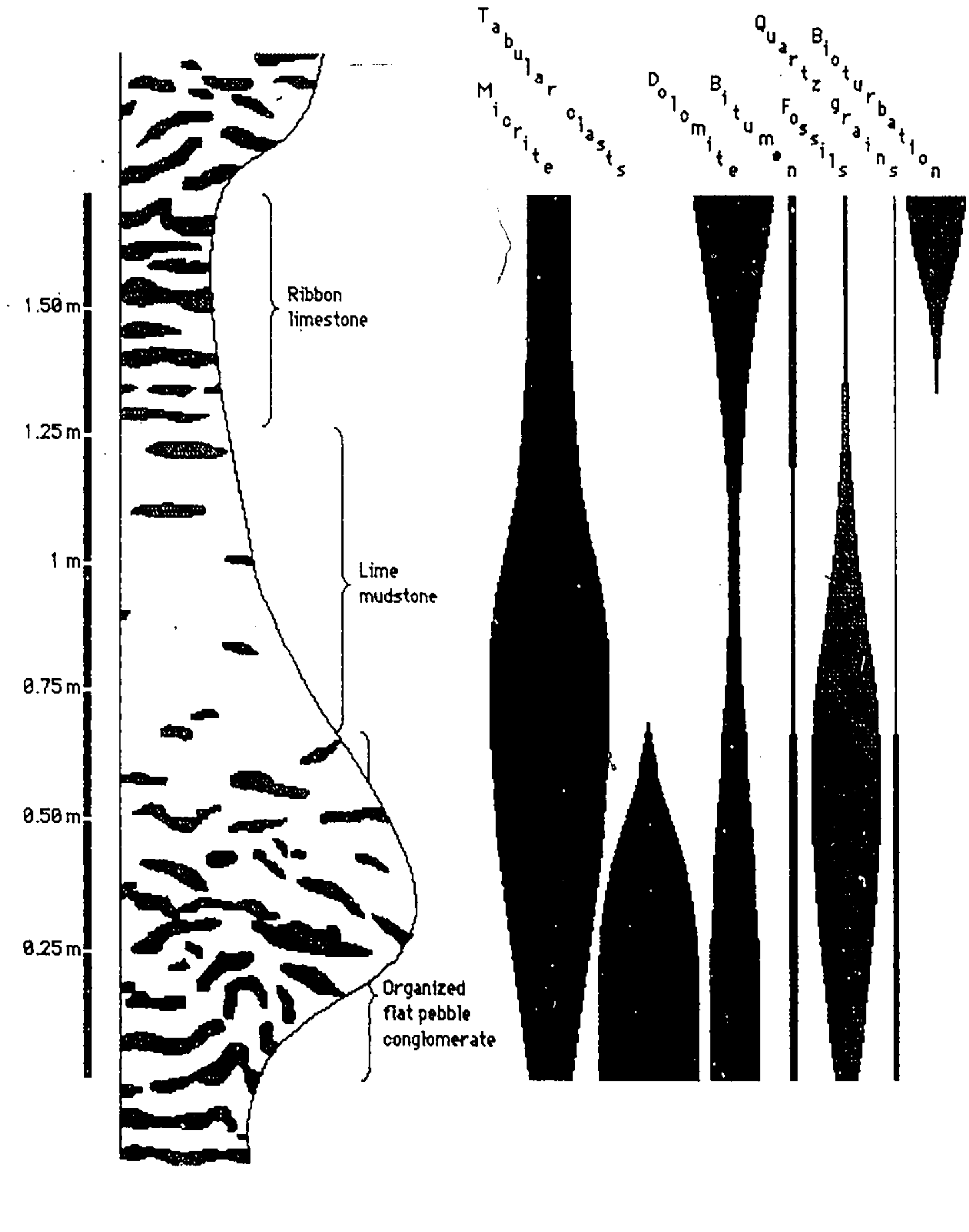


Figure 3.15 Diagrammatic sequence of a fining-upward organized flat-pebble conglomerate. Key to symbols p. 118, Fig. 3.14.

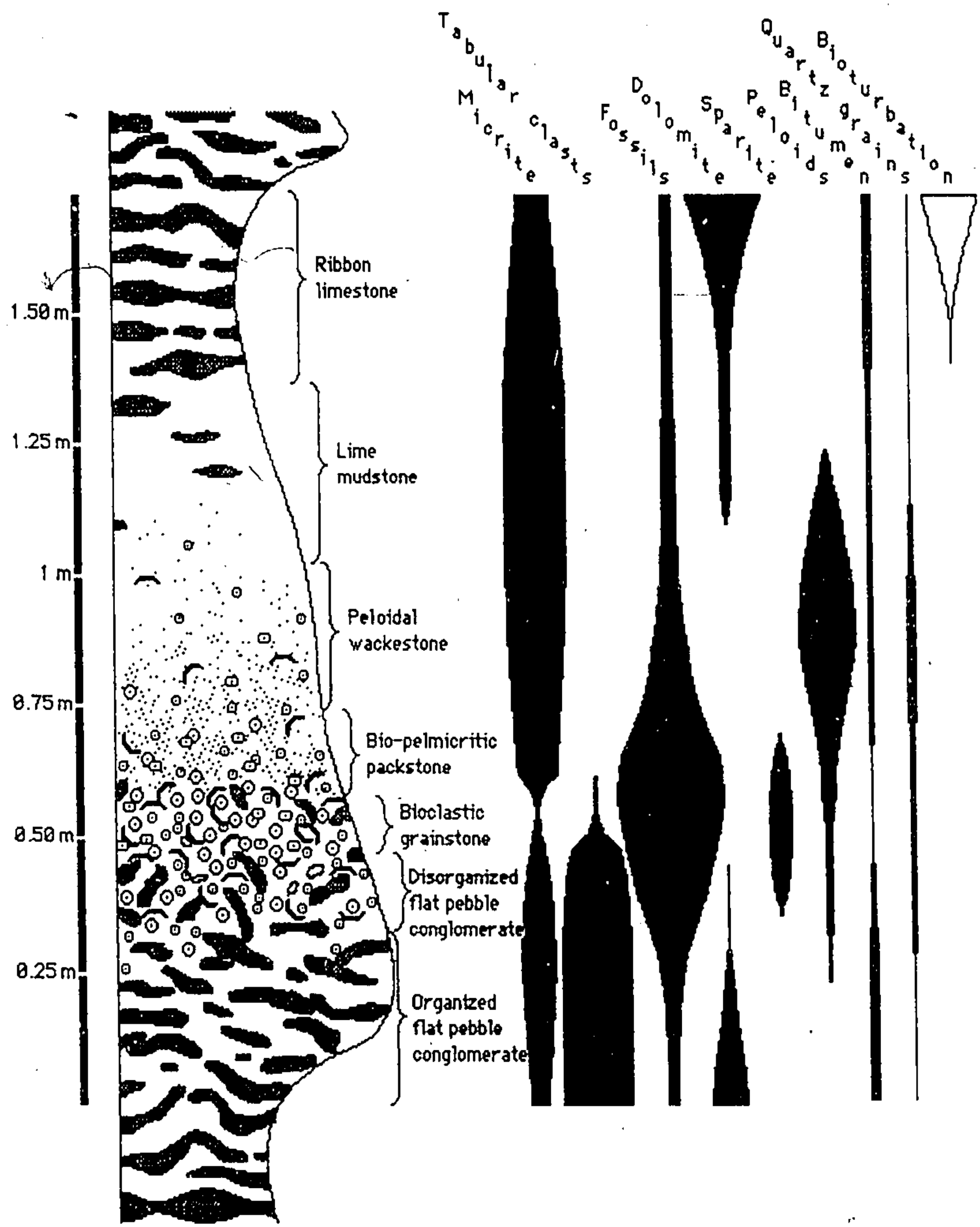


Figure 3.16 Diagrammatic sequence of organized flat-pebble conglomerate, limestone turbidite and ribbon limestone. For key to symbols see p. 118.

Lithologically similar conglomerates with sharp erosional bases and more chaotically arranged flat pebbles (see Figs. 3.17, 3.18 & 3.19) apparently represent more extensive turbulent mass movement. Unlike storm-deposited limestone conglomerates on shelf environments, these conglomerates typically show little or no preferred clast orientation, sorting or grading, and are predominantly matrix-supported. Also unlike storm deposits, they occur in a relatively narrow geographic belt: they are partly responsible for the greater thicknesses (1450 m in JL 01) of the southeastern sections in which they occur, compared to the northwestern sections (e.g. 400 m in JL 16). These massive limestone conglomerates also grade up into finer deposits that are darker, more laminated and less resistant, and these presumably represent turbidity current deposition that succeeded each mass flow event. Some of these limestones, especially in section JL 05, have been completely dolomitized, often obscuring evidences of their origin, therefore making them almost impossible to distinguish from nodular dolostones since they originated from them and contain the same components. However, careful examination of an entire sequence (from nodular dolostone to the top of the fining-upward dolomitic conglomerate, again a nodular dolostone, as illustrated in Fig. 3.20) helps in separating the nodular dolostone from the dolomitic conglomerate, since the proportion of darker areas to lighter areas dramatically increases across an abrupt contact.

Limestone conglomerates of more diverse clast composition are increasingly important volumetrically in Wenlockian and Ludlovian portions of the studied sequence. They illustrate increased diversity of available source materials, and multigeneration clasts are evidence of a sequence of cementation, breakage, transport and deposition repeated as many as 4 times over a prolonged history. They suggest slopes of significant inclination and length. These polymict limestone conglomerates form the matrix-supported, very poorly sorted bases for fining-upward sequences that commonly succeed each other for 10's of metres. Again, this appears to represent abundant debris flow deposition alternating with more organized deposition from closely associated, waning turbidity currents (see Figs 3.21, 3.22 & 3.23).

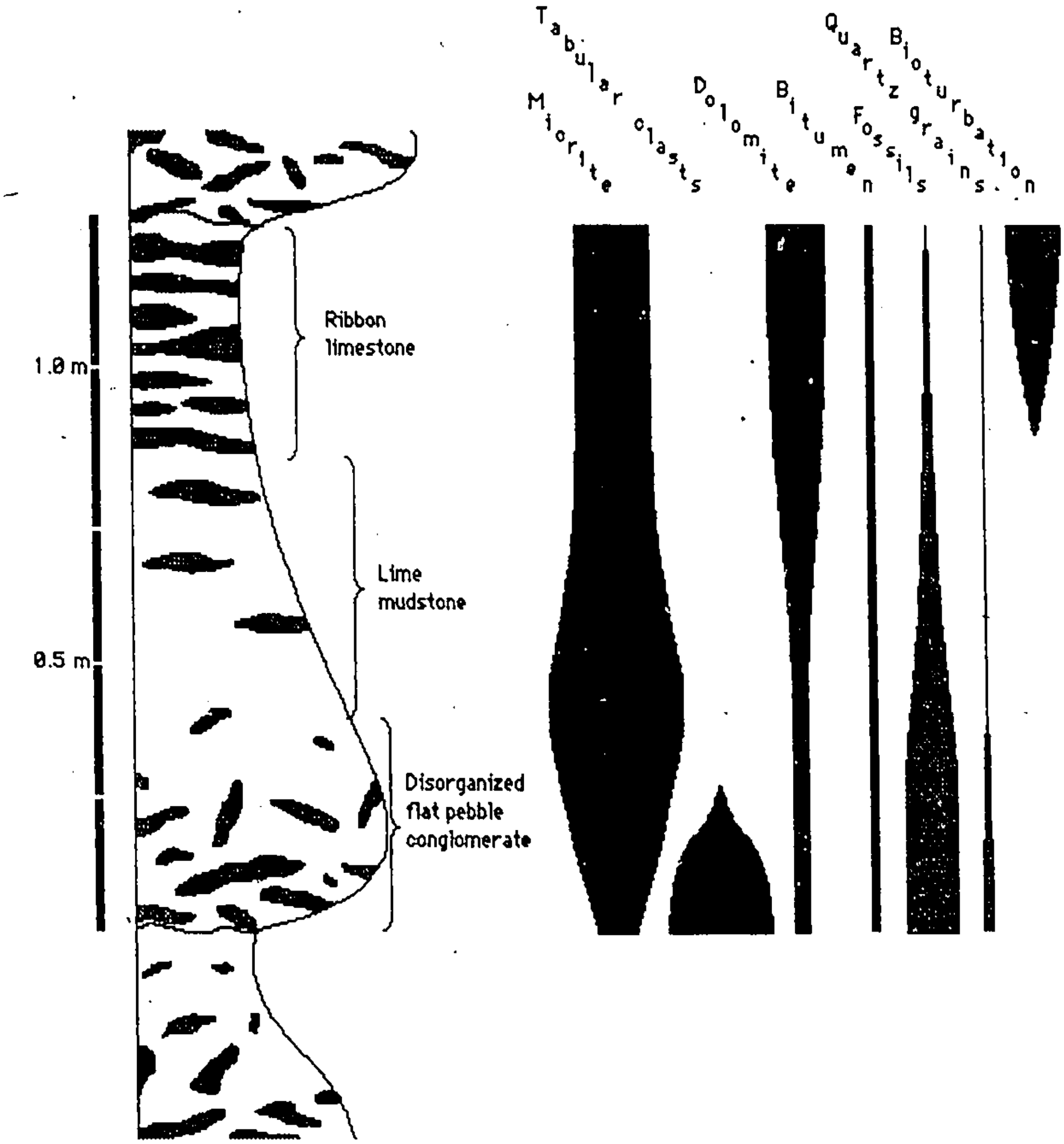


Figure 3.17 Diagrammatic sequence of flat-pebble conglomerate, lime mudstone and ribbon limestone. For key to symbols see p. 118, Fig. 3.14.

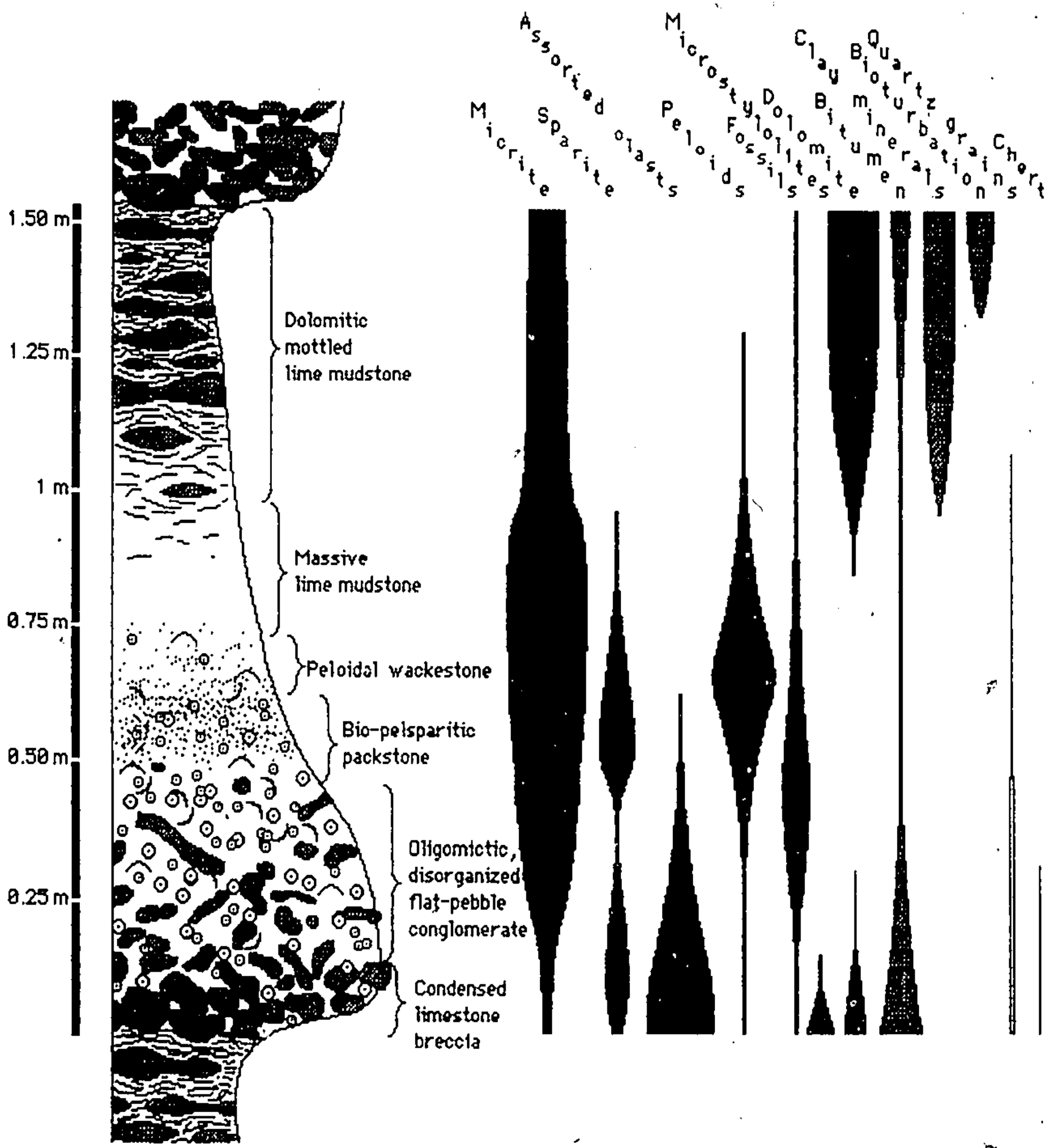


Figure 3.18 Diagrammatic bioclast- and peloid-rich fining-upward sequence, beginning with condensed limestone breccia and oligomictic disorganized flat-pebble conglomerate. Key to symbols on p. 118.

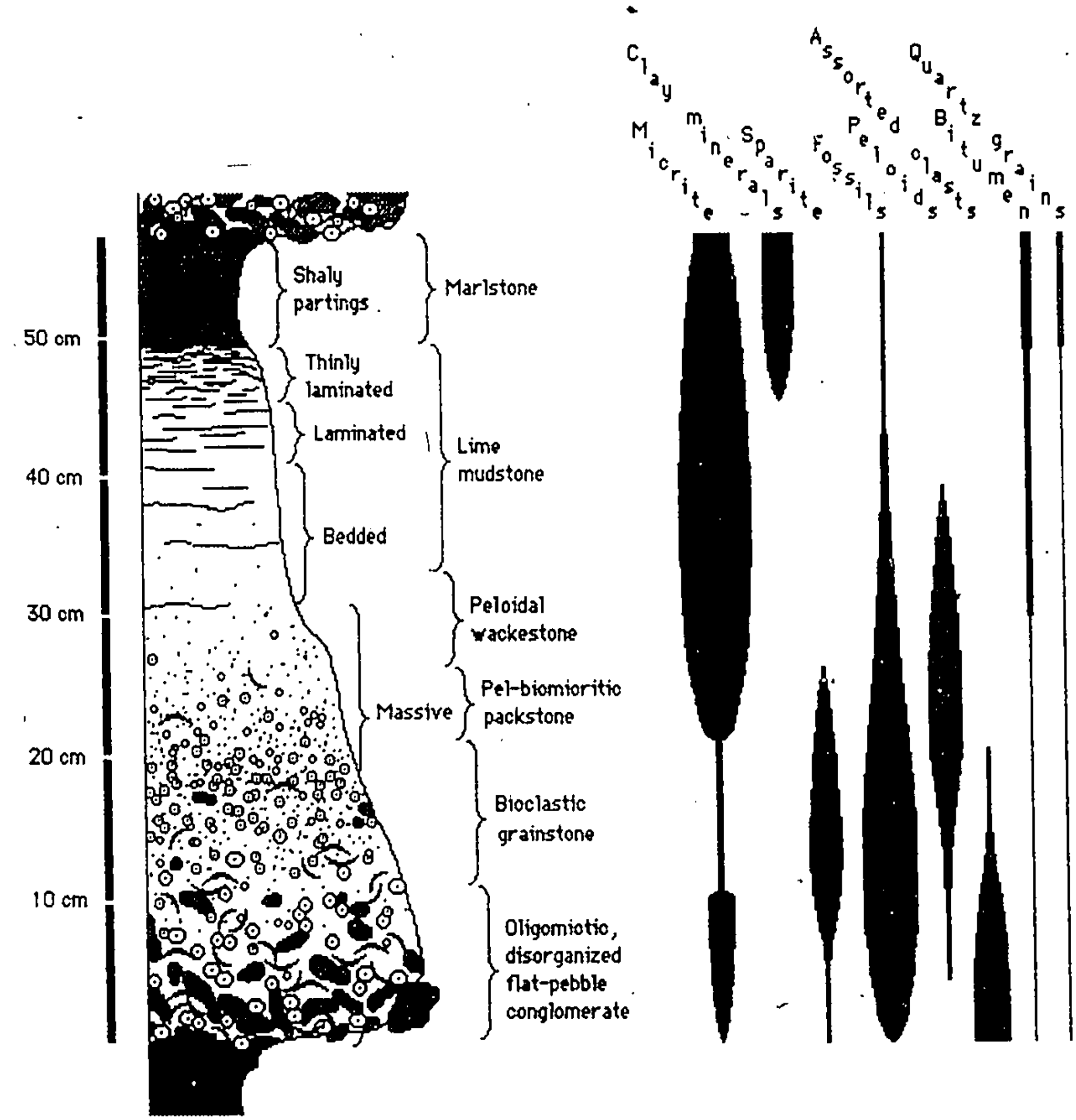


Figure 3.19 Diagrammatic bioclast- and peloid-rich fining-upward sequence, beginning with oligomictic disorganized flat-pebble conglomerate. Key to symbols on p. 118.

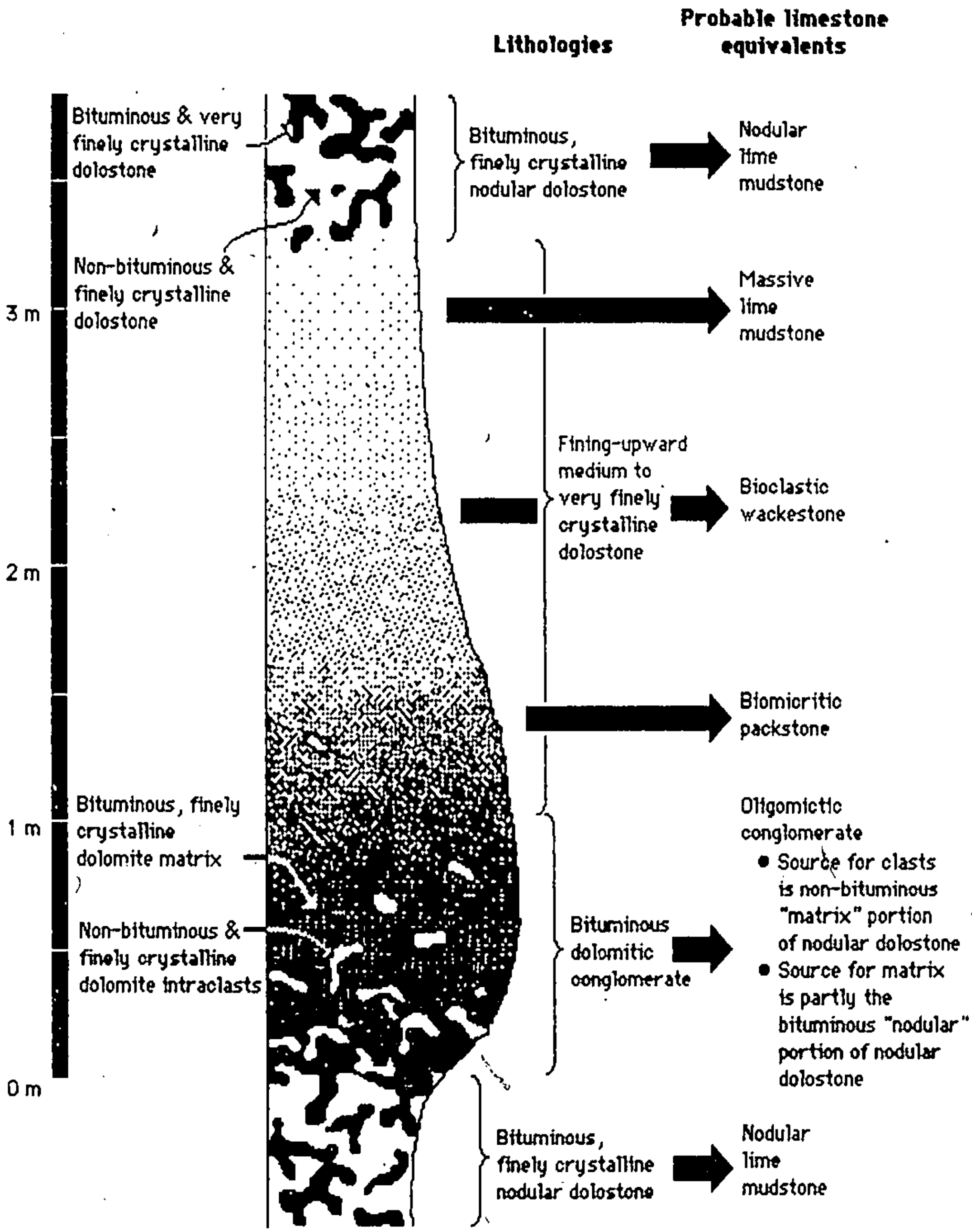


Figure 3.20 Diagrammatic relationships of nodular dolostone and dolomitic conglomerate, and probable equivalents.

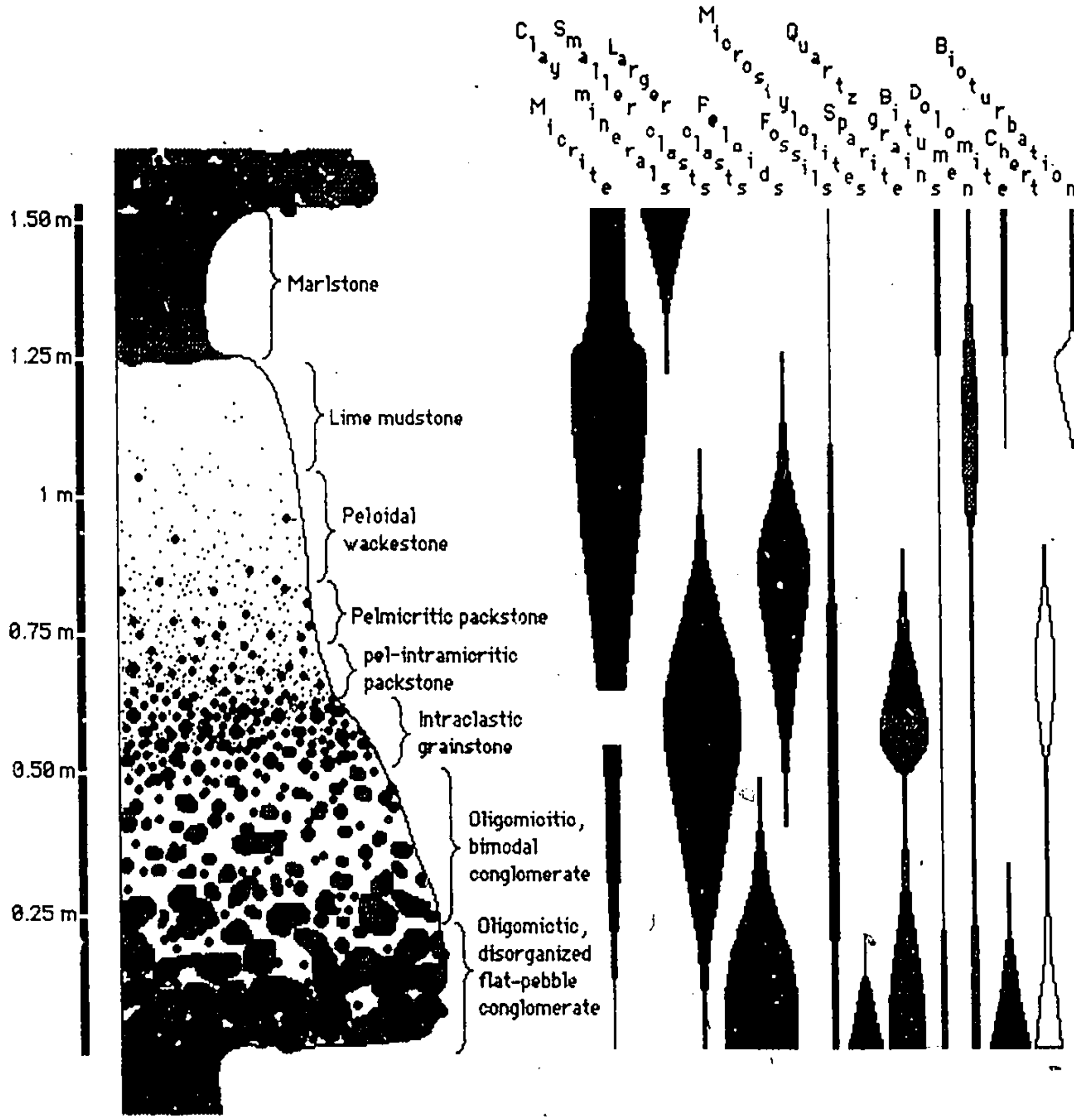


Figure 3.21 Diagrammatic intraclast- and peloid-rich fining-upward sequence, beginning with oligomictic disorganized flat-pebble conglomerate and oligomictic bimodal conglomerate. Key to symbols on p. 118.

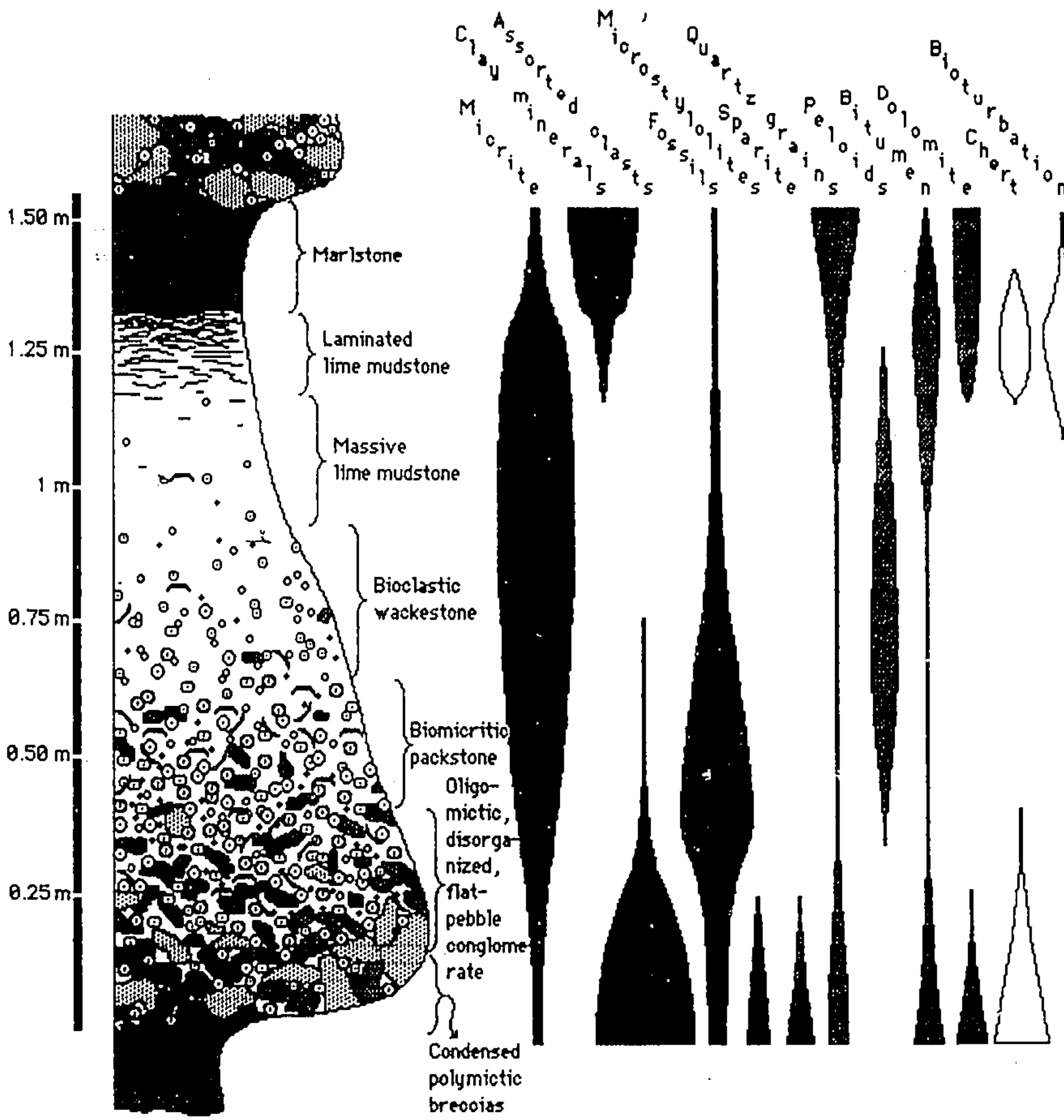


Figure 3.22 Diagrammatic bioclast-rich fining-upward sequence, beginning with polymictic conglomerate. Key to symbols on p. 118

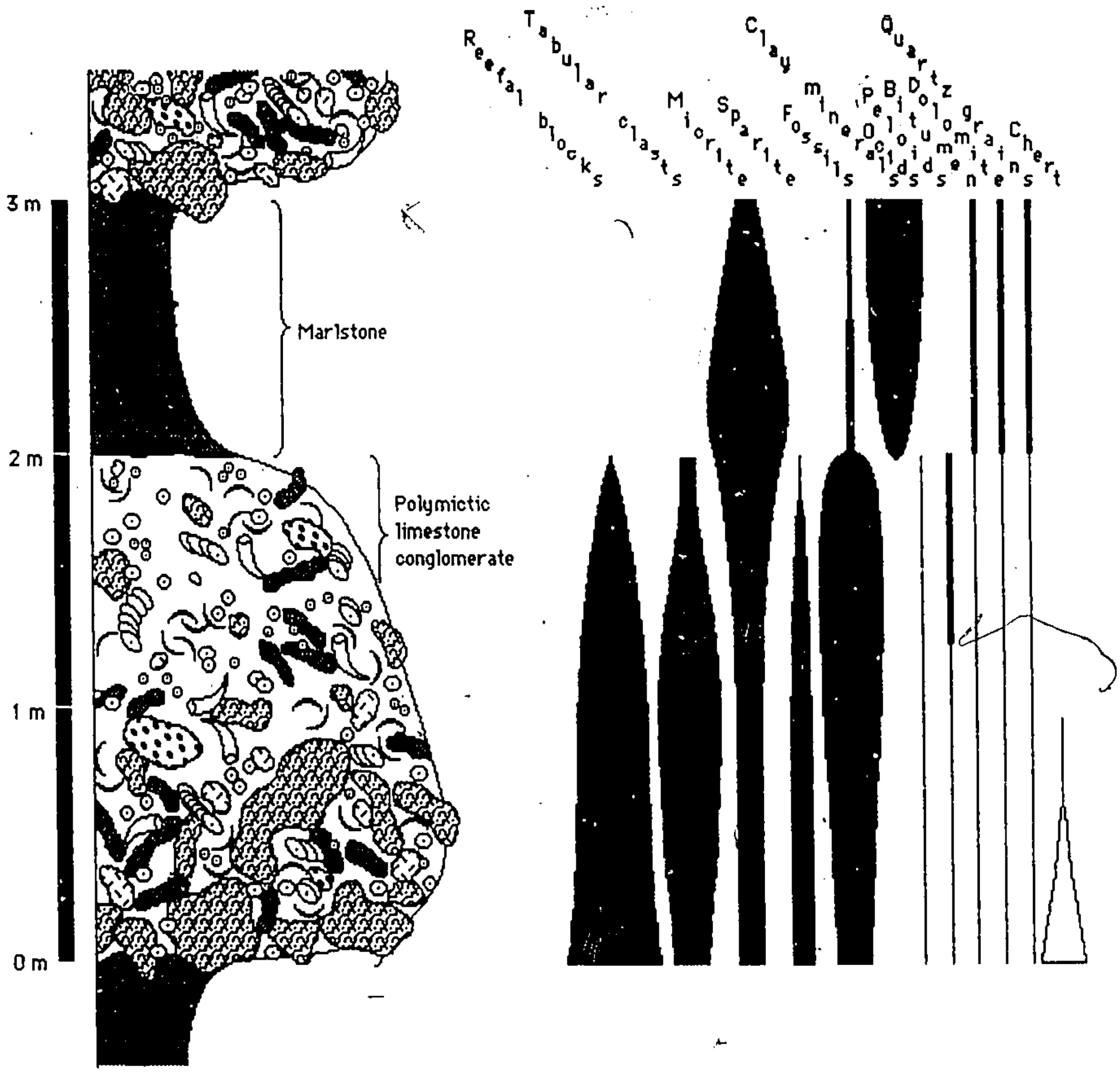


Figure 3.23 Diagrammatic representation of an olistostrome-(polymictic limestone conglomerate) marlstone couplet. Key to symbols on p. 118.

In summary the great variety of limestone conglomerate can be interpreted most reasonably as debris flow deposits, transported downslope as bed load and deposited largely chaotically. The clast components represent sediment sources ranging from reefoid buildups of the shelf margin to fine-grained, bituminous ribbon limestones of deeper fore-reef and slope environments. The evidence in the study area is of substantial slopes that persisted through at least the Silurian.

3.4.1.2 Fining-upward sequences or limestone turbidites

Interbedded with and gradationally overlying the limestone conglomerates, interpreted as debris flow deposits, are fining-upward sequences, interpreted as limestone turbidites. Four main zones can be recognized clearly in what appear to be complete limestone turbidites (also see Fig. 3.24):

i) basal grain-supported (mostly bioclastic, rudite size) zone
ii) transitional finer bioclastic zone (sand size) in which grain support is gradually lost
iii) massive lime mudstone zone, usually most important volumetrically
iv) laminated zone, usually grading imperceptibly up from lime mudstone into marlstone.

Most of the primary sedimentary structures that characterize Bouma's (1962) zones A to E of the classic siliciclastic turbidite are missing from the carbonate fining-upward sequences, possibly due largely to post-depositional destruction (Enos and Moore, 1983), and because of the different hydrodynamic behavior of the carbonate allochems. However, there is distinct grading, expressed as upwardly decreasing grain size and grain-to-matrix ratios, as well as some segregation of grain types. Bouma's zone A is presumed to be equivalent to zone i) because both have abrupt basal contacts and are the coarsest parts of each sequence. Bouma's horizontally laminated zone B compares in grain size with zone ii) in which horizontal lamination is seen rarely. Bouma's ripple cross-laminated zone C bears little resemblance to zone iii) but is tentatively correlated with it because of position in sequence. Zone iv) probably correlates with Bouma's zone D as both show parallel lamination. An almost imperceptible transition from laminated lime mudstone into marlstone in zone iv) possibly is equivalent to the transition into Bouma's pelitic zone (see also representation of turbidites in the following chapter — section 4.3.2).

Features other than grading, that collectively characterize these as turbidites, include the following:

- their greater importance generally somewhat northwest of the principal accumulation of debris flow deposits (i.e. in an inferred basinward direction)
- their common occurrence gradationally above debris flow deposits (southeastern sections)
- the penetration of burrows only from the topmost surface of a graded unit, implying "single-event" deposition (Scholle, 1971)
- the rare occurrence of a complete sequence of 4 turbidite zones as well as consistent order of zones in less complete sequences
- the presence of coarsest and thickest examples in the southeastern sections
- the prevalence of finer and thinner examples (i.e. those missing the bottom-most zones) in northwestern sections
- their monotonous repetition, without interruption in thick sequences, particularly toward the northwest
- their wide lateral extent although some gradation exist.

The lime mudstone-marlstone couplets (see Fig. 3.25) that are most important in the Wenlockian to Pridolian portions of the northwestern sections can be simply interpreted as the finest and most distal portions of the limestone turbidites. The more bituminous examples (graptolitic "shales" of previous authors) are predominantly marlstone, but all are considered turbidites because of the following:

- they have abrupt basal contacts
- fining-upward can be discerned petrographically (see Figs. 3.6 & 3.7)
- the predominant occurrence of the couplets is northwest of more complete limestone turbidites (i.e. in an inferred downslope direction).

The interbedding of lime mudstone and marlstone is thought to represent the repeated interruption of basin-derived "background" marl sedimentation by turbidity currents carrying fine carbonate derived from the shelf and slope. Somewhat similar couplets can be formed in pro-deltaic settings but are typically less bituminous, more fossiliferous and intensely burrowed.

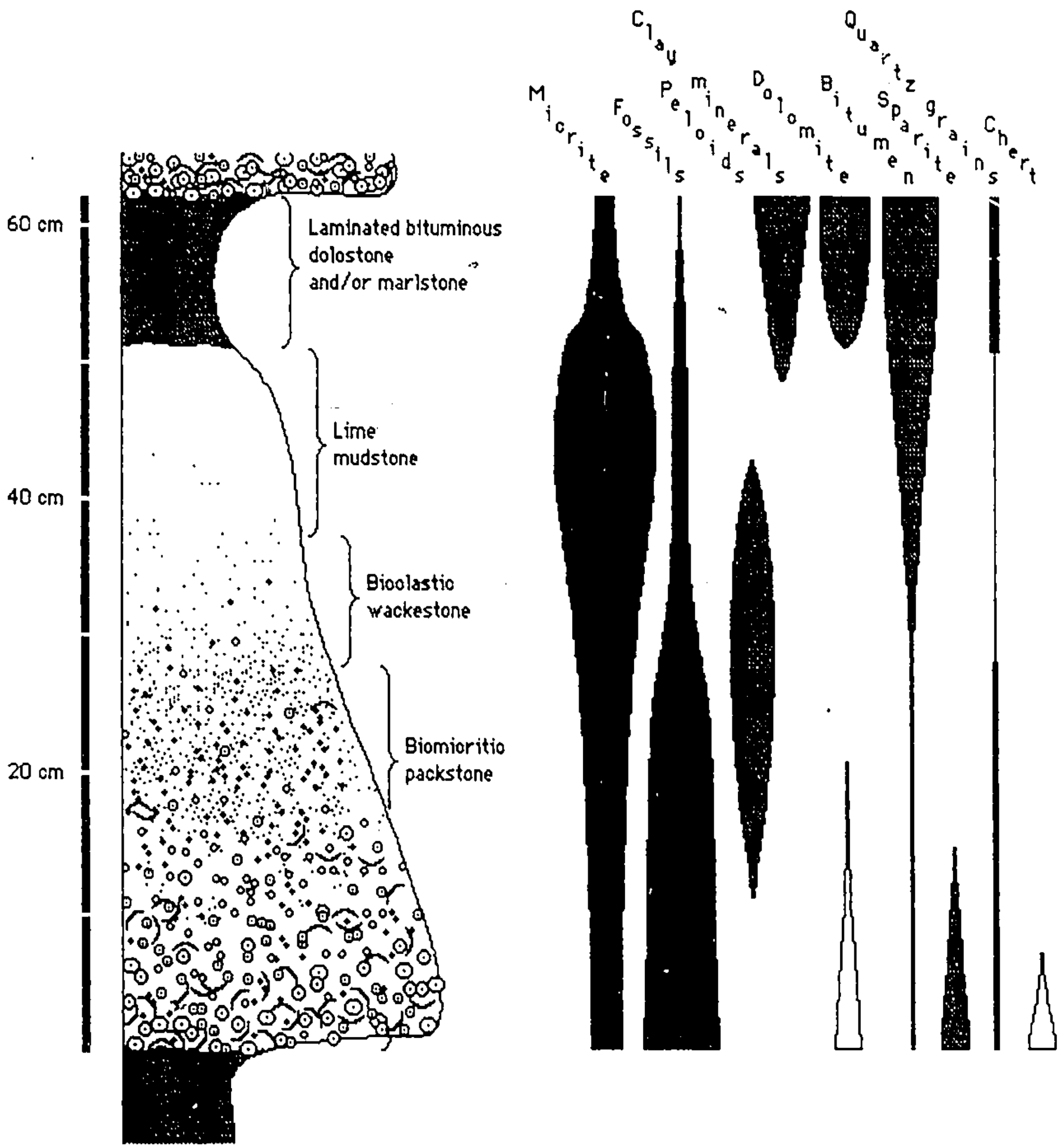


Figure 3.24 Diagrammatic representation of bioclastic limestone turbidite grading into lime mudstone and marlstone. Key to symbols on p. 118.

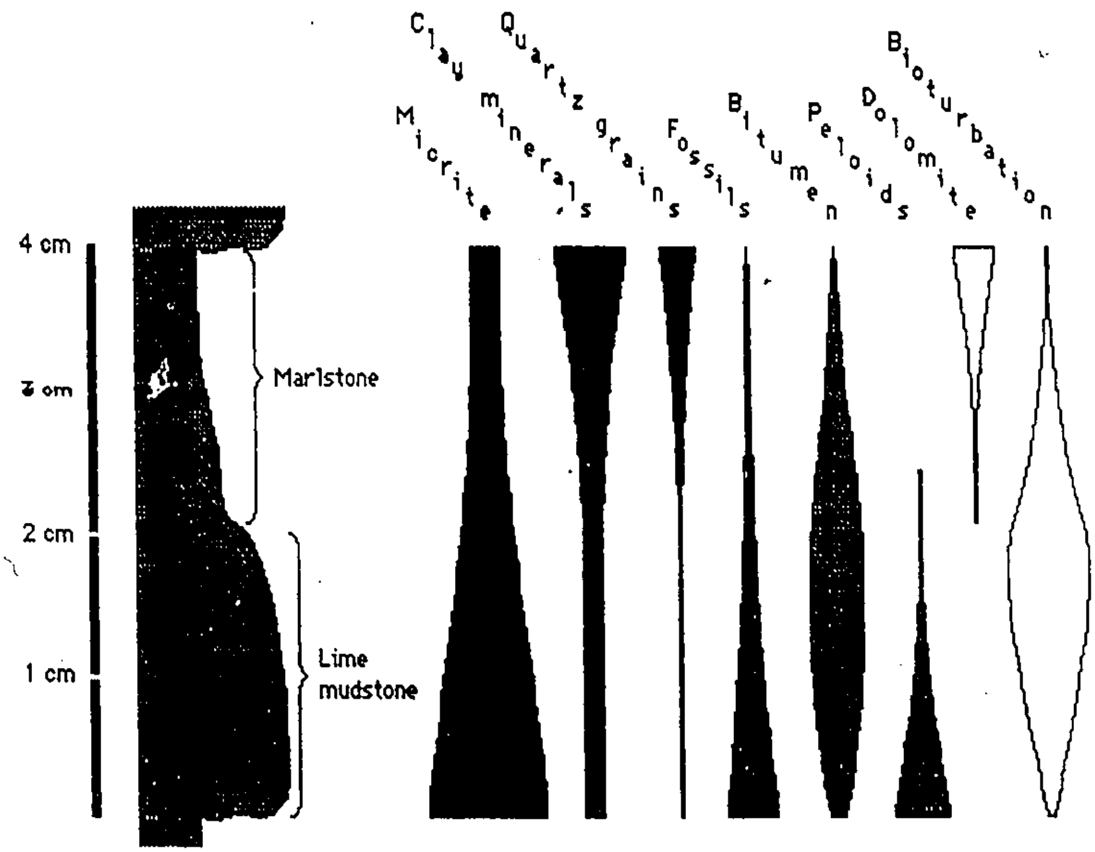


Figure 3.25 Diagrammatic representation of a fining-upward lime mudstone-marlstone couplet. Key to symbols on p. 118.

There is little doubt, therefore, that the fining-upward sequences also represent conditions of slope deposition. Their intimate association with presumably proximal debris flow conglomerates, their distal fining and thinning, and their textural properties are all consistent with them being carbonate turbidites, despite their lack of most of the sedimentary structures that characterize siliciclastic turbidites. These sequences provide clear evidence in the Silurian in this area of major downslope transport and deposition of finer sediments by turbidity currents.

3.4.2 Nodular carbonate and ribbon limestones: bioturbation or early cementation?

Nodular limestones are common over a range of shallow to deep shelf and slope environments (Dvorak, 1972; Garrison and Fischer, 1969; Hurst, 1981; Mullins et al., 1980) and are important for interpreting depositional and early post-depositional history. The origin of the nodular character has been variously attributed to bioturbation (e.g. Dvorak, 1972; Kendall, 1977), or partial early cementation (e.g. Hallam, 1967; Jones et al., 1979; Noble and Howells, 1974), or a combination of the two (Mullins et al., 1980; Wanless, 1980). Evidence that relates to this, from the diverse nodular limestones and dolostones and ribbon limestones of the study area, is discussed below.

3.4.2.1 Argillaceous nodular limestones

Argillaceous nodular limestones are the most common nodular limestones in the study area, and are the predominant lithology in Ashgillian portions of most sections. The 3 types recognized (see section 3.3.2.1 above) contain very similar components, differing only in the proportions and arrangements of these components.

Different types of nodular limestone have been attributed variously to different original argillaceous content (Wanless, 1979), to different degrees of early cementation (Garrison and Fischer, 1969), to compaction-induced diagenetic segregation (Hallam, 1967) or to complex interrelationships between early cementation, compaction and pressure solution (Jenkins, 1974; Wanless, 1979). Evidence of cementation of nodules, prior to compaction of internodule sediments, includes the following:

- neomorphic rims that usually surround the nodules, show evidence of dissolution toward tops and bottoms of some nodules
- fossils are better preserved in nodules than in internodule portions
- minor pores remained open in the nodules, to be filled subsequently by late diagenetic cement.

However, other critical evidence is sparse or lacking. For example, microstylolites are generally absent or not common enough to account for substantial dissolution of internodule material. Such local concentrations of insoluble materials might be expected at nodule/internodule boundaries (Wanless, 1979), but are extremely rare in these rocks. Also, no evidence was seen of exposure of nodules on the sea floor. Nodules encrusted by fossils, or with micritized upper surfaces were interpreted by Jones et al. (1979) as evidence of winnowing of unconsolidated internodular sediment, to expose early-cemented nodules formed slightly below the sediment-water interface.

Extensive bioturbation, including burrows of several generations, in argillaceous internodular portions, but scarcity of burrows or internal deformation in nodular portions can have more than one implication. This pattern in bioturbation would be consistent with early cementation of the nodules and consequently more prolonged access by burrowers to unconsolidated internodular portions. It is possible that burrowing was responsible for the initial disruption of continuous limestone layers and establishment of a nodular structure that was accentuated by early cementation (cf. Wanless, 1980).

In summary, there is evidence of both early cementation and bioturbation in these nodular argillaceous limestones, but no conclusive evidence yet, of their relative importance in nodule formation. A possible modern analogy for the inferred processes of nodule formation may be that reported by Mullins et al. (1980) from modern slopes in the Bahamas. They found that the predominant nodular carbonate sediments accumulating there are caused by penecontemporaneous burrowing, submarine cementation and bottom currents. Continuous layers of lime mud, accumulating while bottom currents inhibit settling of suspended clay minerals, are disrupted by burrowing.. These currents tend to increase pore water circulation and promote early cementation. With slightly decreased current intensity, clays can settle to form more marly layers.

3.4.2.2 Non-argillaceous nodular limestone

Non-argillaceous nodular limestone occurs mostly in the uppermost Ashgillian and rarely in the lower Llandoverian parts of the studied sections. It is closely associated in sequence with argillaceous nodular limestone, beneath, and nodular dolostone, above. A close similarity in nodular structure suggests that nodular dolostone resulted from dolomitization of non-argillaceous limestone, and intermediate stages are well documented (Figs. 3.26a→d).

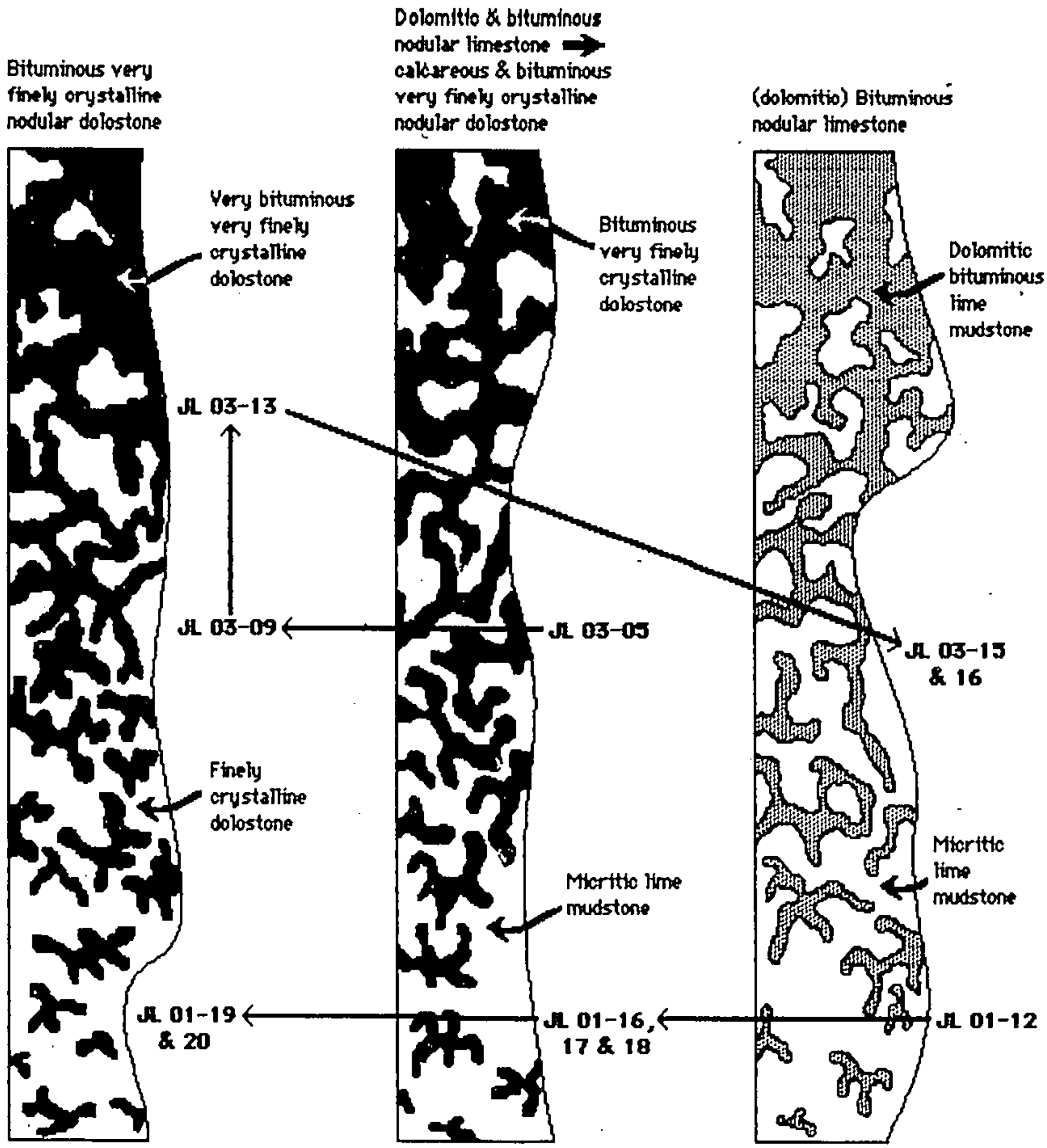


Figure 3.26a Diagrammatic representation of the nodular limestone gradational increase of their dolomite content up-sequence (follow the arrows) in sections JL 01 and JL 03. These diagrams generally show a gradual up-unit increase in the darker portions of these rocks. Finally the darker these dark portions the higher their bitumen content, a general up-sequence change as well.

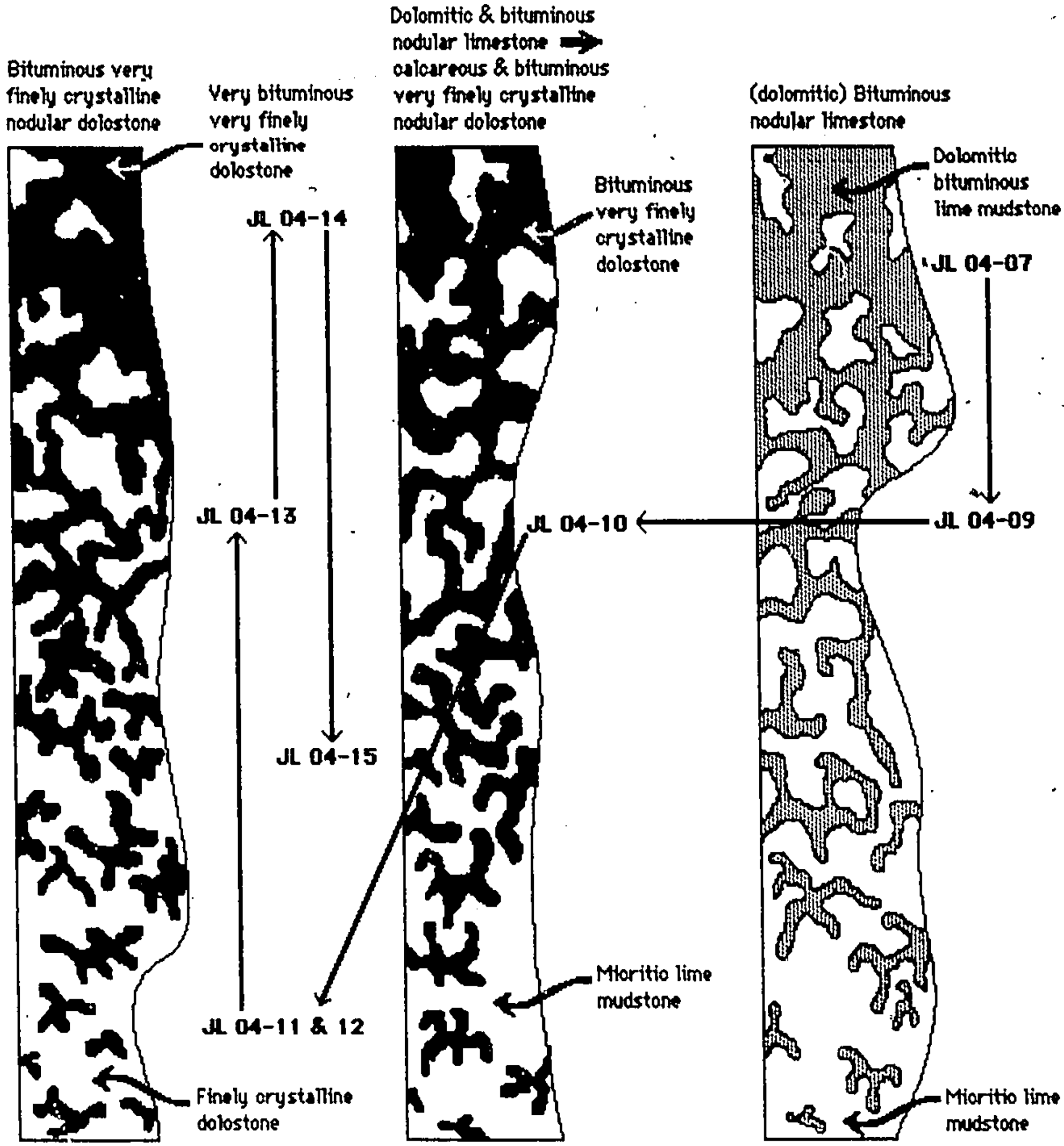


Figure 3.26b Diagrammatic representation of the general up-sequence progression from nodular limestone to nodular dolostone in section JL 04 (see caption for Figure 3.26a for more information). The thickness of strata throughout which these changes occur vary from section to section.

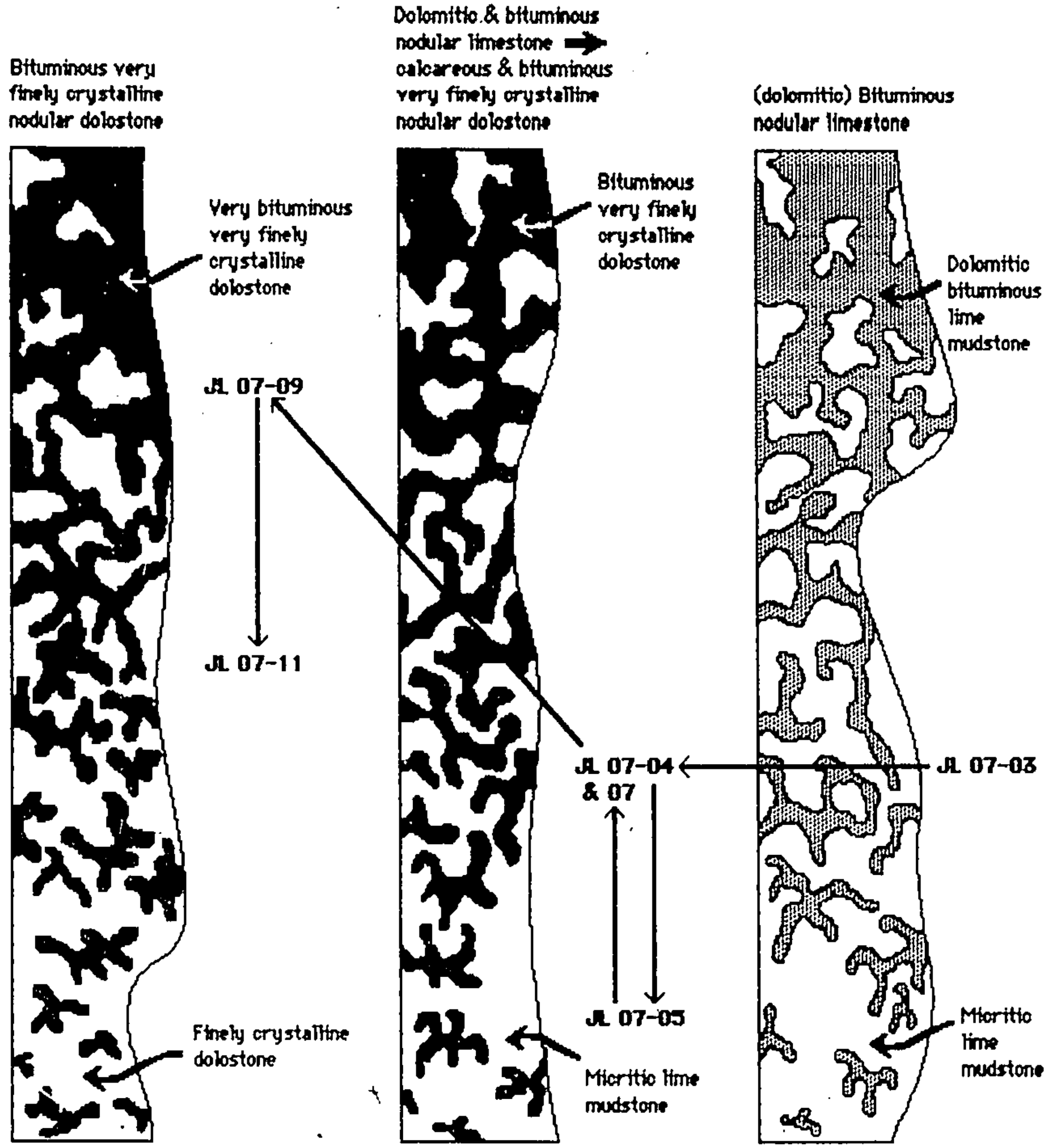


Figure 3.26c Diagrammatic representation of the general up-sequence progression from nodular limestone to nodular dolostone in section JL 07 (see caption for Figure 3.26a for more information).

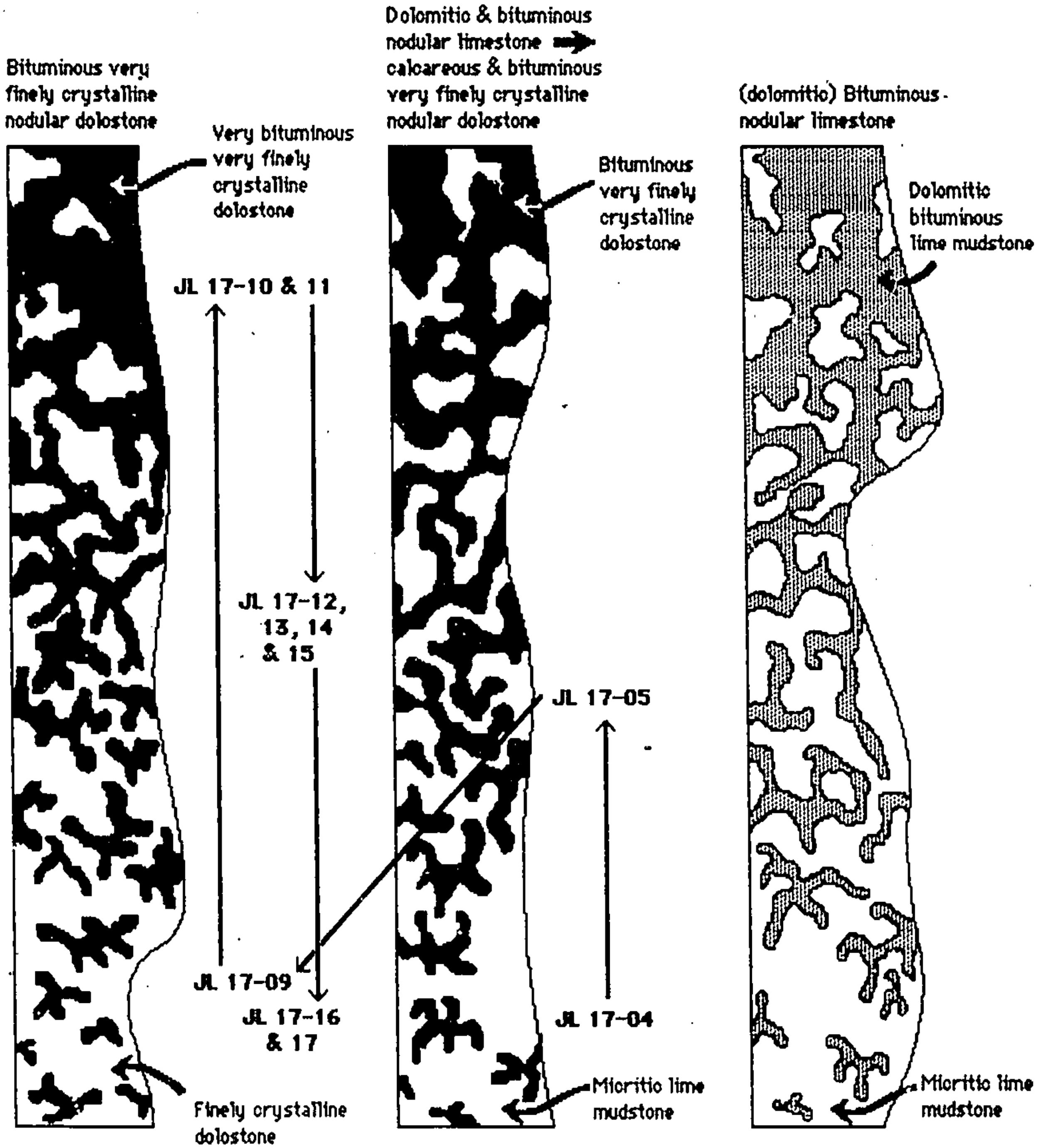


Figure 3.26d Diagrammatic representation of the general up-sequence progression from nodular limestone to nodular dolostone in section JL 17 (see caption for Figure 3.26a for more information).

The characteristic "marbled" appearance of this massive and largely unbedded, non-argillaceous, nodular limestone appears to be related more to early cementation than to bioturbation. There is no clear evidence of burrowing, and indeed the component rock types do not appear to be sufficiently continuous laterally or vertically to represent burrows. Lack of macrofossils in the limestone also suggests that benthic activity was reduced if not absent.

The lighter and darker portions in the limestone are thought to have been initially very similar. The lighter portions, now micritic lime mudstone, are interpreted as parts that were cemented early, closing intergranular porosity. Bitumen, then introduced at an early stage, filled remaining intergranular pore spaces to define the darker portions of the rock, but was blocked from entering early-cemented portions. Different degrees of early cementation could introduce considerable variety of appearance with greater early cementation resulting in greater proportions and continuity of the lighter-colored portions. The reason why the early cementation was patchy is not apparent.

The presence of these two rock components, one with cement-filled pore space and the other bitumen-filled, apparently had a marked effect on subsequent diagenesis. This can be seen particularly in rocks representing stages of progressive dolomitization of non-argillaceous nodular limestone toward nodular dolostone. Initially dolomitization affected the darker portions and resulted in very fine euhedral dolomite distributed through the bitumen, seen in rocks that can be termed dolomitic, non-argillaceous nodular limestone. The bituminous areas appear to be slightly darker than before, possibly reflecting enhanced porosity related to dolomitization, and additional bitumen filling.

After complete dolomitization of the darker areas, the lighter areas were progressively dolomitized, producing a mosaic of larger (finer crystalline), anhedral dolomite crystals, now seen in the nodular dolostone (discussed further in section 3.4.2.4 below & illustrated in Figs 3.26a→d)

The lack of evidence of benthic activity and fossils indicates some restrictive conditions in the depositional environment. As will be discussed below, deposition of the overlying nodular dolostone took place in progressively shallower environments during the Ashgillian, and indeed the resulting restrictive conditions might have started with deposition of these non-argillaceous nodular limestones as might be indicated by the partial or complete dolomitization of its darker portion. The absence of clays might also indicate winnowing on a current swept shallow-shelf.

3.4.2.3 Ribbon limestone

Ribbon limestone occurs mostly in Llandoverian parts of the studied sections. It is a lithology well documented elsewhere, for example on deeper portions of low gradient carbonate ramps (e.g. Read, 1980; Markello and Read, 1981).

Ribbon limestone is similar lithologically to non-argillaceous nodular limestone in combining lighter (bitumen-poor) and darker (bitumen-rich) rock types, but differs in that the two are regularly interbedded in very elongate lenses or beds of uniform thickness ("ribbons"). The continuous, thin, planar bedding of these largely unfossiliferous rocks is a clear indication of lack of bioturbation.

The formation of ribbon limestone here is considered to have been similar to the formation of non-argillaceous nodular limestone, i.e. partial early cementation of continuous layers, followed by bitumen migration locally into non-cemented layers, progressive dolomitization of lime muds in the darker layers, and progressive filling of presumably enhanced porosity by bitumen. Ribbon limestone has alternatively been attributed to compaction and pressure-solution (Wanless, 1979), but these processes do not appear to be applicable in this study because of the absence of pressure-solution features.

3.4.2.4 Nodular dolostone

Common examples of the upward transition of non-argillaceous nodular limestone into nodular dolostone, and their very similar nodular structure, are strong evidence that the latter formed through complete dolomitization of the former (see 3.4.2.2 above and Figs 3.26a→d). Massive nodular dolostone is characteristic of the Allen Bay Formation beneath the bituminous chert marking the Ordovician-Silurian boundary in the study area, and is also typical of the formation further south and southwest.

The Allen Bay Formation, from near its base up to the bituminous chert marker, appears to be a thick shallowing-upward sequence. It shows several trends, including upwardly diminishing bitumen content, increasing dolomite crystal size in bituminous portions, and increasing vuggy porosity in non-bituminous portions (the last especially in the top 20 m of this interval in the southern sections). Despite the extensive, destructive dolomitization, structures resembling hardgrounds or erosional surfaces are evident in nodular dolostone of the topmost part of this interval in section JL 01.

Similar nodular dolostone (e.g. Mayr, 1980, p.12, Fig. 10) characterizes much of the Allen Bay Formation further south. There it has been interpreted as shallow subtidal to supratidal (Morrow and Kerr, 1977; Mayr, 1979; Sodero and Hobson, 1979), based on occurrence of a very sparse, restricted fossil assemblage, vuggy porosity (now mostly spar-filled), complete dolomitization, and hardgrounds and/or karst features. Morrow and Kerr (1977) inferred that fresh water was involved in the dolomitization on Grinnell Peninsula, and the vuggy porosity and dissolution features support this. Analogous features, particularly toward the top of the nodular dolostone in the study area, suggest shallowing-upward trends, and at least part of the dolostone there appears to represent shallow shelf environments.

3.4.3 Dolostone buildups — probable reefs

Two large masses of pale dolostone and smaller associated ones occur in the Wenlock-Ludlow interval in sections JL 01 and JL 06 (see section 3.3.3.2, above), and a smaller one in section JL 04. Two smaller but lithologically similar buildups in sections JL 03 and JL 07 will be discussed further in section 3.5.2, below.

Very little evidence of original properties remains in these largely coarsely dolomitized, massive and structureless dolostone bodies. Their origin, therefore, can be inferred only partly from the buildups, themselves. Some of the best evidence of original components, properties and processes is provided by adjacent coarse bioclastic and intraclastic beds, although some of them, too, were extensively dolomitized near to the buildups.

The buildups appear to be lensoid and of very limited lateral extent, especially in section JL 06. Mound form is apparent, but it is difficult to distinguish original topography because the enclosing fine-grained rocks are substantially compacted differentially (Plate 1-1) and stratification was not seen to continue into the buildups from surrounding rocks.

Ghost structures and characteristic molds of crinoid fragments, gastropods, brachiopods, bryozoans, stromatoporoids, calcareous algae and compound rugose and tabulate corals are preserved very locally in the buildups, mostly peripherally. A similar diverse but fragmented fossil assemblage, presumably derived from the buildup, occurs in coarse bioclastic beds that thicken towards, and end abruptly against, the buildups. These bioclastic deposits, unlike most others in this study, are not part of fining-upward sequences and tend to have limited lateral extent (e.g. not extending from JL 01 to JL 02 in a presumed basinward direction). Their substantial porosity and/or intergranular cement, and the presence locally of current-aligned bryozoan fragments, orthoconic nautiloids and pentamerid brachiopods (e.g. units JL 01a-61 and 62), are evidence of vigorous sorting and winnowing.

Debris beds, interpreted as olistostromes (see Fig. 3.23 and Plate 3-1), adjacent to the buildup in JL 01, incorporate large and small blocks of massive, unbedded limestone, presumably from the buildup. The various clast types present, suggesting a variety of original rock types in the buildup, were generally not dolomitized, probably because of their isolation from the main body of the buildup. Included blocks of framestone and bindstone are reefoid in character but are disoriented and therefore cannot be interpreted as small *in situ* mounds. According to their composition and the variety of fossils shown, they most likely represent dislodged fragments of the more fossiliferous periphery of the buildup. Clasts of various sizes (up to 7 m) of distinctively pale, homogeneous, micritic lime mudstone lacking allochems, that appear in the highest of these debris beds, could represent the nature of original material from back of the periphery of the buildup. This would be consistent with the uniformly unfossiliferous character of the buildup core, now completely altered to crystalline dolostone. The buildup dolostone shows no geochemical differenoes (XRF—St/Mg) from periphery to core and this, too, suggests that the largely unfossiliferous nature of the core reflects the original character rather than extent of dolomitization.

The buildups occur only in what is interpreted as the shallowest, most shelfward part of the study area. Erosional features (possible karst) on top of the buildup in section JL 06 suggest that these structures were emergent at least intermittently.

The evidence collectively indicates that these large dolostone buildups in fact represent reefs with substantial original topography, as indicated by the amount of downslope debris flows derived from them and the size of some of the incorporated blocks. They apparently had pronounced fore-reef slopes, down which reef-derived debris of large and small size tumbled, as discrete talus blocks and as massive debris flows, interrupting the slope accumulation of finer lime mudstone and marlstone. There is some indication that the slightly fossiliferous margins and unfossiliferous cores, now entirely dolomitized, represent original zoning, analogous to zoning reported in reefs elsewhere (Packard, 1984). If this is so, then erosion and collapse of the reef margin became more substantial through time, eventually in the Ludlovian contributing blocks of reef core material to the flanking aprons of debris.

3.5 Major environments represented

In comparison to other formations included in this study, and indeed in comparison to the Allen Bay and Cape Phillips formations and Read Bay Group of other regions to the south and southwest, the lithological variety in the Allen Bay Formation-Read Bay Group in the Baumann Fiord area is extraordinarily large. Furthermore, these rocks or rock sequences alternate repeatedly, and sometimes cyclically, even those not thought to be genetically related. Some of them evidently represent narrow facies belts. The diverse components and processes represented in these beds, are evidence that various factors were responsible for the rapid alternation of these facies. Such amounts and variety in the cycles is thought to reflect unstable environments with narrow and diverse belts of deposition. The rapidity with which these depositional conditions succeed each other is characteristic of deposition across a shelf-to-basin transition. Such a model has been proposed by many workers (Christie and Peel, 1977; Hurst, 1981; Hurst and Surlyk, 1982; Hurst and Surlyk, 1983; Kerr, 1976; Morrow & Kerr, 1977;

Trettin, 1979; Mayr, 1978 & 1980; Packard, 1985) for such deposits throughout the Arctic Islands. The type of shelf margin, and thus the type of transitions more problematic. Not enough studies have been done to formulate a model that would explain consistently regional differences in the evolution of the Franklinian Basin, from northern Greenland to Banks Island.

In addition to the frequent alternation of rock types and sequences, several factors collectively substantiate this interpretation of a depositional model. The Silurian section as a whole thins spectacularly toward the northwest to a quarter of the thickness present in the southeastern sections. The fact that bitumen permeates almost every lithology is a characteristic of basinal and basin margin environments, but not of shallow shelf sequences of this duration. Extensive slopes, some of substantial gradient, are indicated by large volumes of flat pebble conglomerate, derived by downslope mass movement from ribbon limestone, of finer turbidites, and of reef-derived conglomerates and breccias.

Although a multitude of components and rock types are arranged in a great number of sequences, representing a dynamic ecological and structural evolution, it is nevertheless possible to note common characteristics and typical arrangements. Six major groups of associated lithologies or facies were recognized, namely *shallow* and *deep shelf*, *reefal shelf margin*, *upper* and *lower slope*, and *basin*. These are examined below.

3.5.1 Deep shelf

In this study, deposits of the deep shelf formed below storm wave-base and thus exhibit very little evidence of current influence, such as sorting, grading or alignment of allochems. In all sections the lower Ashgillian deposits, i.e. the argillaceous nodular limestones, are interpreted as deep shelf deposits. Although there are similar fine grained limestones in the uppermost Ludlovian and Pridolian of section JL 01, they are included with "reefal shelf margin deposits" (section 3.5.2) in view of their close association with buildups. Some ribbon limestones of the Llandoverian of JL 01 also possibly represent the deep shelf, as was the case for Virginian Cambro-Ordovician deposits studied by Markello and Read (1981), but here they are more characteristically upper slope deposits, as in Packard's (1984) study.

Although below wave-base, these deposits were nevertheless oxygenated as demonstrated by the intense burrowing and persistence of the varied "Arctic Ordovician" fauna (Morrow & Kerr, 1977). However, the fossils are sparse, indicating some degree of restriction compared to subsequent shallower water fossil concentrations. The rarity of light-dependent calcareous algae suggests deposition at depths below 40 m.

The predominantly very fine grain size of these deposits is an indication of calm conditions; coarse fossil grains present are considered to be *in situ* or to have been disturbed by bioturbation only. The extensive lateral distribution of these lower Ashgillian deposits, throughout the study area and widely on islands further south and southwest, is consistent with deposition in shelf environments.

3.5.2 Shallow shelf

Shallow shelf non-argillaceous nodular limestone and nodular dolostone are present in the Ashgillian of most sections. In these sections, non-argillaceous nodular limestone directly overlies the deeper shelf argillaceous nodular limestone, and is overlain by nodular dolostone, thought to represent still shallower environments in which the whole rock was dolomitized since at least part of the dolomite occured early in these rocks formation because some detrital dolomite grains were found in slightly younger mass flow deposits. Therefore the entire Ashgillian portion of the Allen Bay Formation up to the bituminous chert marker at the Ordovician-Silurian boundary is a single shallowing-upward sequence. The uppermost part of this sequence represents the shallowest environments in the study area, except for the large dolomite mounds. The following features together tend to substantiate this shallowing upward:

- decrease in evidence of bioturbation from argillaceous to non-argillaceous nodular limestone, and body fossils becoming less diverse and increasingly rare, are interpreted as indicating increasingly restricted conditions at higher stratigraphic levels

- upward increase in degree of dolomitization and size of dolomite crystals in southernmost sections (JL 01, 02, 05, 07, and 10) are interpreted as reflecting more intense and rapid dolomitization at higher levels
- vuggy porosity, such as in the top 20 m of this shallowing-upward sequence in the southernmost sections (e.g. units JL 01-19 and 20) is generally viewed as evidence of fresh water solution
- an erosion surface (possibly karstic) at the top of the sequence (unit JL 01-20) appears to indicate at least temporary subaerial exposure.

Morrow and Kerr's (1977) mixed water model for dolomitization of the Allen Bay Formation appears to be applicable in the study area. They attributed the dolomitization to lenses of fresh water invading the sequence from scattered more positive areas on the shelf. These could have been discontinuous stromatoporoidal or oncolitic mounds (Mayr, 1974; Morrow and Kerr, 1977) developed near the shelf margin in front of a shallow shelf lagoon. No appropriate mound structures have been recognized in the more southern part of the study area. However, in sections JL 03 and 07, somewhat to the northwest, and just above the nodular dolostone interval, are two mounds, 17 m and 21 m thick, incorporating stromatoporoid-compound coral floatstone, and now completely dolomitized. These and similar mounds could have been involved in the introduction of fresh water to the sequence according to the model of Morrow and Kerr (1977). Significantly, all the nodular dolostone sections south and east of these mounds (i.e. shelfward) show up-sequence changes as described above. In contrast, the 2 equivalent sections north of the mounds show minor or no up-sequence change — only partly dolomitized non-argillaceous nodular limestone in JL 17 and equivalent fine slope and basinal deposits of the Cape Phillips Formation in section JL 16. The mounds appear, therefore, to delineate the Ashgillian shelf margin (Fig. 3.27).

3.5.3 Reefal shelf-margin

Large, massive dolostone buildups, interpreted as reefs, occur in the Wenlockian to middle Pridolian parts of sections JL 01 and JL 06 (see 3.3.3.2 and 3.4.3), and smaller Ashgillian buildups in sections JL 03 and JL 07 (see

3.5.2). They are considered to define the contemporary shelf-margin. The reefs were discontinuous both through time and laterally along the shelf margin. Only Ludlovian reef development is represented in JL 06, but three stages of development are evident in JL 01: an extensive Wenlockian to middle Ludlovian stage and two subsequent short stages in the latest Ludlovian and the beginning of the middle Pridolian.

The buildups, particularly the large reefs, were responsible for accumulation of a belt of distinctive reef and associated facies at or near the shelf margin. The larger reefs evidently built up to sea level as shown by the abundant calcareous algae and stromatoporoids. Erosion of their margins contributed substantial and diverse debris to fore-reef slopes (see section 3.4.3). The reefs were lithified at the sea floor as indicated by the rigidity and coherence of eroded limestone blocks of all sizes transported downslope. The cemented, but undolomitized blocks provide major evidence about original components and properties of the reefs. The upper portions of these debris aprons also show abundant evidence of near-surface mechanical reworking by current and wave action — well sorted, porous grainstones, rudstones (coquinas) and packstones.

Increased slopes, accompanying development of the reefal shelf margin, can be inferred from the upward change to more polymict conglomerate, from increased evidence of multigeneration clasts, and from the increased proportions of fining-upward sequences that cap these debris flow deposits. The increasing magnitude, higher in the sequence, of collapses of the reef margin, as indicated by progressively larger collapsed blocks, and blocks that incorporate reef core as well as reef margin also suggest increased relief of the reef margin in section JL 01 from Wenlockian to middle Ludlovian.

Lithification of the reefs was followed by early dolomitization, possibly resulting from exposure of their upper portions and infiltration of fresh water. The dolomitization also subtantially affected closely adjacent porous beds in the debris aprons. Evidently porosity and permeability barriers prevented the dolomitizing solutions from extending into enclosing lime mudstones and marlstone that intertongue with the debris aprons.

Possible inter-reef or back-reef environments are represented in section JL 01 between the middle and latest Ludlovian portions of the reef. The predominant mudstones represent mainly low energy conditions, but are interbedded with minor, well-sorted, crinoidal and peloidal grainstone, packstone and wackestone (unit JL 01a-54). The sorting and spar cement in the grainstone are not consistent with mass flow deposition. The infrequent occurrence of these coarser deposits, some of them lenticular, similar to channel fills, possibly indicate storm deposition. Stratigraphically equivalent deposits downslope in section JL 02 are mostly typical limestone turbidites that incorporate a varied fauna of stromatoporoids and corals, apparently reef-derived. The source of this reefal material, presumably could have been either between JL 01 and JL 02, or lateral to JL 01 along the shelf margin. The mudstones and minor coarser deposits of unit JL 01a-54, therefore, could represent either back-reef or inter-reef environments. The mudstones are unusual in lacking evidence of bioturbation and in containing well preserved, complete ostracoderms, unlike the exclusively fragmentary remains in equivalent and earlier slope deposits.

The shelf margin is much less well defined where only small isolated buildups occur along it, and the distinctive flanking debris is minimal or absent. In the absence of reefs or other buildups, shelf environments can pass imperceptibly into slope environments, and facies typical of deep shelf or upper slope can be closely associated (e.g. interbedded) over a broad belt, without any clear demarcation of a shelf/slope break.

3.5.4 Upper slope

Deposits of the upper slope form the transition between deposits of the Allen Bay Formation-Read Bay Group undivided and those of the Cape Phillips Formation. They occur principally in sections JL 01, JL 05, JL 06 and JL 10. Similar deposits in sections further northwest are extremely rare among much finer downslope deposits and possibly represent more substantial mass flow and turbidity events than usual, rather than facies migrations. Upper slope deposits are best developed in upper Llandoverian to upper Ludlovian portions of the sequence.

Two distinct groups of upper slope deposits can be recognized: those associated with non-reefal shelf-margins, and those associated with reefal stages. Usually the former are fossil-poor oligomictic flat-pebble conglomerates interbedded with ribbon limestone that represent the repeated alternation of mass flow deposits with the *in situ* beds from which they were derived (late Llandoverian and Wenlockian of section JL 01). Only locally is crinoidal material, from shelf-margin or upper slope environments, incorporated as matrix in the conglomerate and as fining-upward debris in limestone turbidites.

During reefal stages, the upper slope deposits characteristically contain substantial reef-derived bioclastic and intraclastic debris. Massive polymictic limestone conglomerates and overlying fining-upward sequences predominate, representing the prevalence of mass flows and turbidity currents on the increasing gradients of the upper slope. Although the finer end-members of these fining-upward sequences are quite like limestone turbidites emplaced on the lower slope, the coarse limestone conglomerates underlying rarely occur in the northwestern sections. In addition, the massive lime mudstone beds that terminate many of the sequences on the upper slope, are commonly intensely bioturbated, compared to the extremely rare bioturbation in lower slope deposits.

3.5.5 Lower slope

The deposits of the lower slope environment are typical Cape Phillips Formation deposits, and incorporate most of the beds in the Llandoverian-to-Pridolian segment of sections JL 02, JL 03, JL 04, JL 07, JL 16 and JL 17. It is also the dominant depositional environment represented briefly by Wenlockian deposits of JL 01 and JL 10.

The lower slope deposits consist mostly of graded deposits, interpreted as limestone turbidites, that are finer than in upper slope deposits. They typically grade up from bioclast-rich limestone into lime mudstone or marlstone. They also differ from upper slope deposits in forming rhythmic sequences, some more than 100 m thick, almost without interruption by other deposits. They were deposited under more or less continuous anoxic conditions as demonstrated by their high bitumen content, generally restricted fossil assemblage of graptolites and ostracods, and minimal bioturbation. Very minor coarser-grained turbidites contain fossil material from upper slope and/or shelf-margin areas. The deposits of the lower slope have been commonly referred to in the past as "graptolitic shale" of the Cape Phillips Formation (Tozer, 1963).

3.5.6 Basin

Basinal and/or toe-of-slope deposits have been recognized only in JL 16, the most northern section, where they constitute most of the Ludlovian to Pridolian portion. These deposits resemble lower slope deposits but rarely contain allochems and even more rarely consist of fining-upward sequences with basal bioclastic (crinoid-rich) beds. The deposits are extremely bituminous, rhythmic, lime mudstone-marlstone couplets that represent the finest of fining-upward limestone turbidites, equivalent to coarser and thicker turbidites on the upper slope.

Anoxic conditions are indicated by the high bitumen content, the occurrence of pyrite nodules and the extremely restricted allochthonous assemblage of graptolites, some of them pyritized. Selective silicification and dolomitization are rare, but much more common than in lower slope environments. This is analogous to diagenesis of deep oceanic deposits reported in the Deep Ocean Project (Matter, 1974; Heath and Moberly, 1971) and in other studies (Berger, 1975) and partially explained by anoxic-acidic conditions, a good silica source and compaction of the clay minerals (Dietrich et al., 1963; Land, 1983; Mattes and Mountjoy, 1980; McHargue and Price, 1982; Wanless, 1979).

The source for the silica might have been fine sand and silt size quartz and feldspar in these deposits, some of them rich enough to be termed calcareous siltstones, and perhaps opaline silica from sponge spicules as well. Since quartz grains are not found in such abundance in any deposits of shallower environments in the study area, it is doubtful that their source was the shelf. More likely, they were transported north to south in the Hazen Trough from a land mass on the other side of the Trough (Trettin, 1979), and were not transported up-slope significantly.

Alternation of lime-mudstone laminae with well sorted but not fining-upward laminae and lenses that are quartz grain-supported, is consistent with basinal (e.g. contour) current rather than turbidity current processes. Also the large concentration of trilobite pygidia in unit JL 16-07, indicates a mechanism of sorting that is unlikely to have been achieved by turbidity currents, and contour or cross-basinal currents are suggested.

3.6 Geological history

3.6.1 Ashgillian (Fig. 3.27)

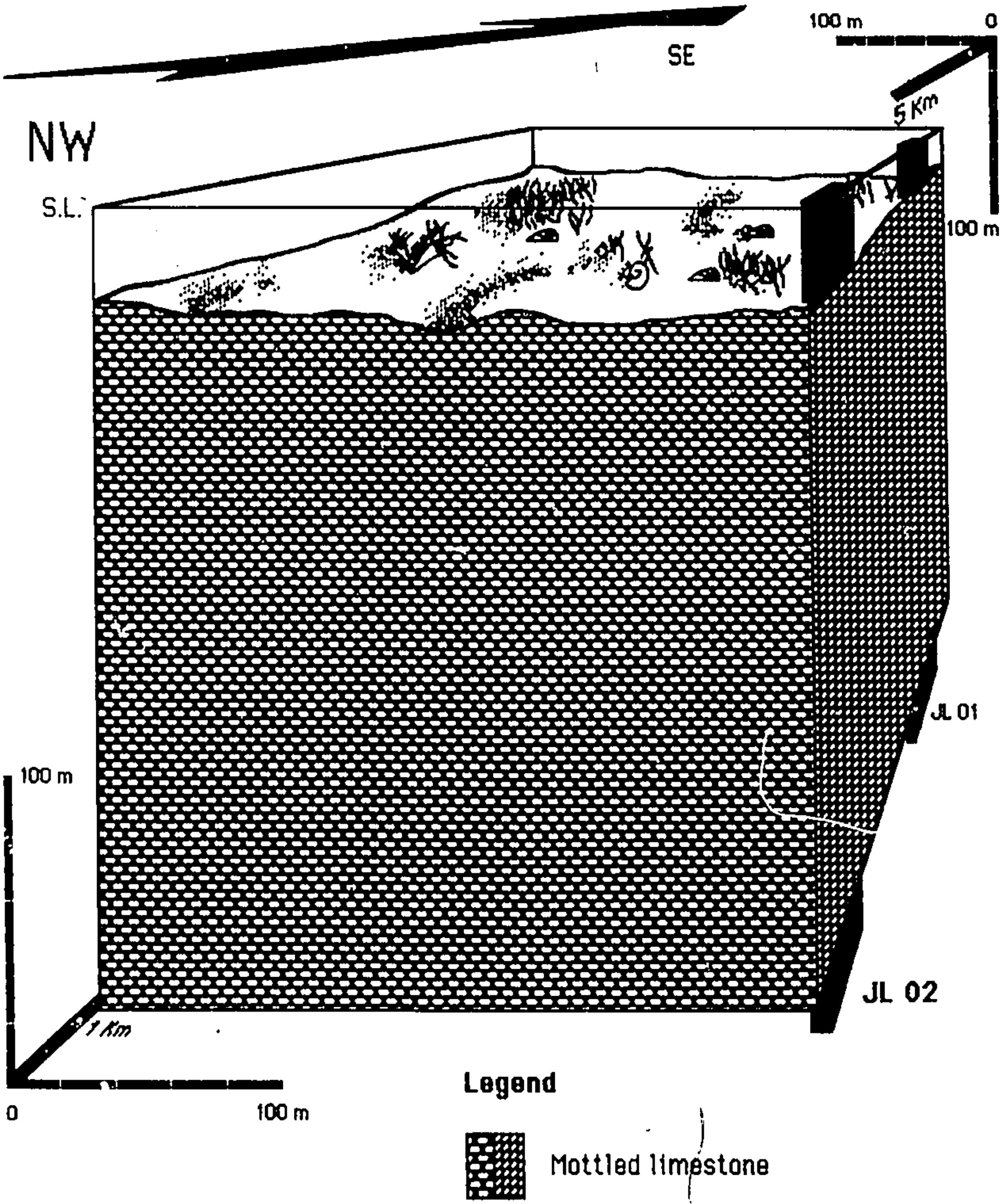
The Ashgillian portion of the Allen Bay Formation was characterized by depositional conditions similar to those of the underlying Irene Bay Formation as demonstrated by the abundance and variety of fossils, and the intense bioturbation . The little variations noted between sections attests to a wide, gently sloping carbonate shelf throughout the study area, probably a ramp which extended much further south. Fig. 3.27 is a representation of the area at the time.

This wide carbonate shelf had its margin much further to the northwest than interpreted from younger deposits, although probably poorly defined, typical of a ramp. Therefore it is no coincidence that sections JL 03 and JL 07, at the western edge of the study area, each contain small bioherms, highly dolomitized and silicified.

The gradual up-sequence faunal restriction, and the gradual changes from nodular limestones to bituminous nodular dolostones, culmulating in what appear to be hardgrounds and/or erosional surfaces defined by a vuggy porosity just underneath very irregular surfaces, reflect some fresh water influence and subaereal exposure, therefore a possible general shallowing-upward cycle.

Figure 3.27

Ashgillian



• The Allen Bay Fm-Read Bay Group undivided and Cape Phillips Fm •

3.6.2 Early Llandoverian (Fig. 3.28)

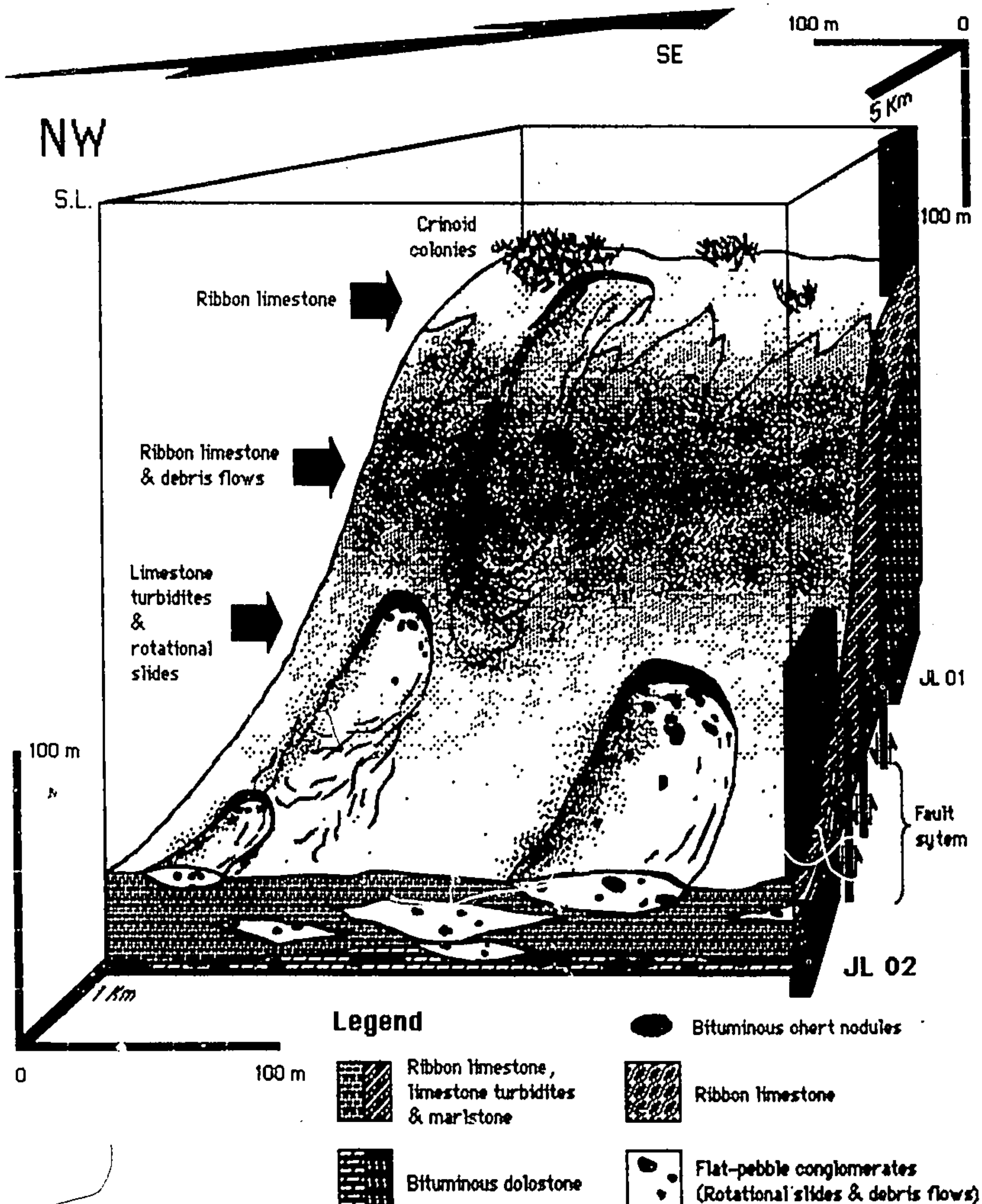
At approximately the Ordovician-Silurian boundary, an abrupt change is recorded through the area from Grinnell Peninsula on Devon Island to the northern tip of Greenland. This change is indicated in the study area by a unit of very bituminous and cherty dolostone, commonly interbedded with black chert. This unit thins dramatically basinward (northwestward) and is very deprived of fossils. Some potassium phosphate (apatite) is also present. Above this unit the sections become more heterogeneous as the shelf-margin backstepped abruptly to the southeastern corner of the area.

This sudden change can be interpreted as the collapse of a major portion of the outer shelf, and is registered not only in this area but throughout the Arctic Islands and northern Greenland (Hurst, 1980; Hurst and Kerr, 1982; Hurst and Surlyk, 1983). Just prior to this collapse, the shelf was at its shallowest, as indicated by dolomitization of these transitional sediments. Large amounts of organic matter that would normally have been oxidized at shallow water depths, were preserved as the sudden collapse quickly brought these shelf deposits into anoxic, much deeper environments. Earlier dolomitization produced porosity in which reducing microenvironments encouraged the precipitation and replacement by silica (chert nodules sometimes coalesced into beds).

3.6.3 Llandoverian to early Wenlockian (Fig. 3.29)

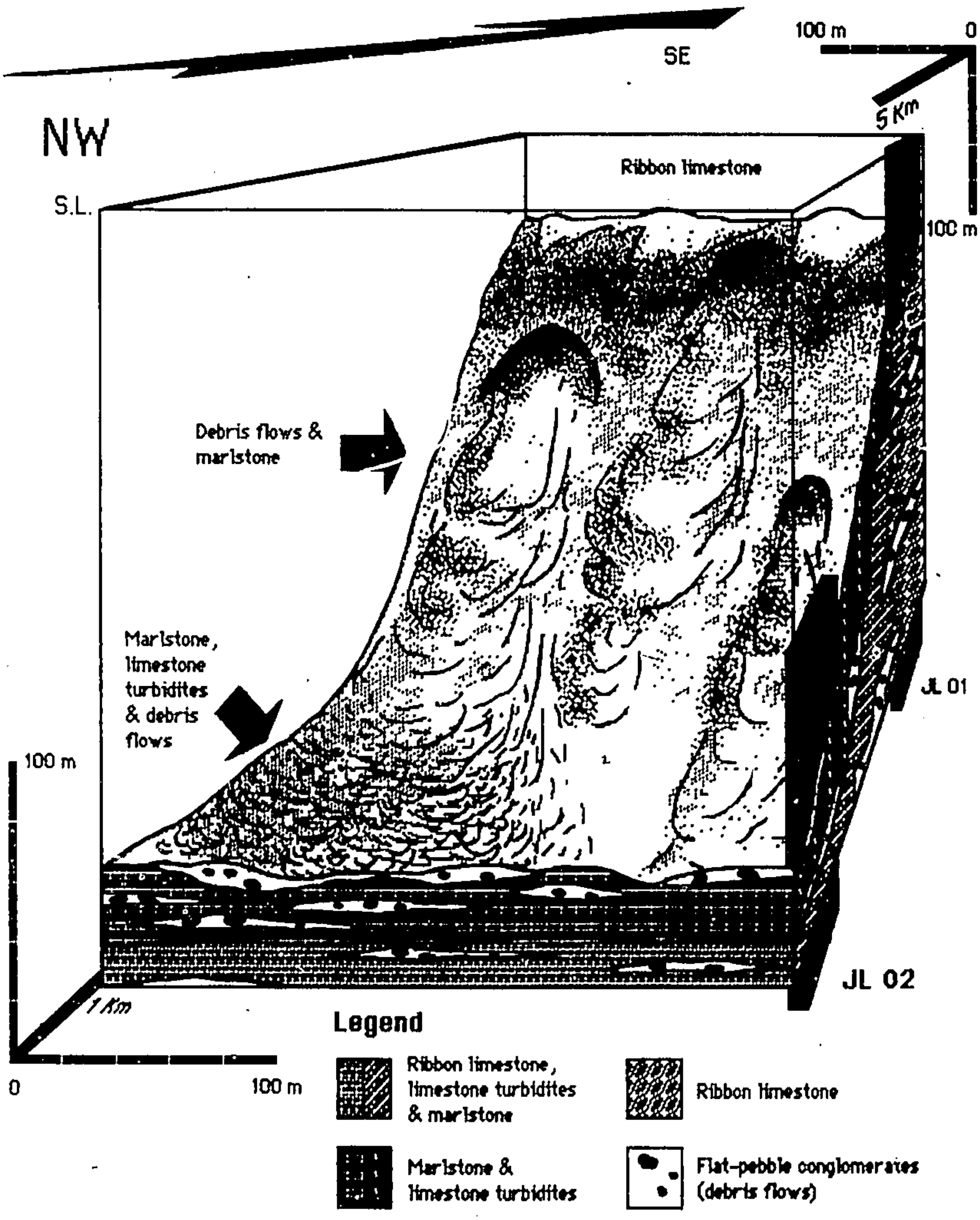
Only in JL 01 & JL 10 are shelf-margin or upper slope facies still represented in the Llandoverian and early Wenlockian in the study area. Ribbon limestones are characteristic and were the source for interbedded oligomictic flat-pebble conglomerate. The very restricted fauna, the partial dolomitization of these ribbon limestones and extreme dolomitization still further south (Mayr, 1978; Morrow and Kerr, 1977) are consistent with the persistence of shallower conditions toward the south after the collapse of the outer shelf. Sections further down slope are not dolomitized and dolomite occurs only as an allochthonous component. Because of low gradients, some flat pebble conglomerates have organized fabrics resulting from slow down slope creep of slides, without significant mixing of matrix and clasts.

Figure 3.28 Early Llandoverian



• The Allen Bay Fm-Read Bay Group undivided and Cape Phillips Fm •

Figure 3.29 Llandoverian to early Wenlockian



In lower slope sections (JL 02 & JL 04) most of the deposits of this age are finer distal turbidites equivalent to the coarse debris flows and slides (flat-pebble conglomerate) originating on the upper slope. Although few of the larger intraclasts travelled this far down the slope, much finer (less than 1 mm) intraclasts, peloids and organic debris that were entrained are now arranged in series of varied fining-upward sequences, or limestone turbidites. Fig. 3.29 illustrates the gradational character of these deposits. The turbidites are commonly interbedded with nodular limestone, resulting from bioturbation of their upper portion, and/or marlstone.

Sections still further into the basin (JL 03, JL 07, JL 16 & JL 17), contain the same types of deposits as above except that they are finer. These limestone turbidites mostly consist of bituminous lime mudstone and marlstone couplets with very few observed structures and components. Anoxic conditions prevailed as demonstrated by the large amount of bitumen preserved, the lack of fossils except for graptolites and the absence of bioturbated lime mudstone.

Towards the northwest, presumably in deeper environments, the following changes are evident:

- deposits are finer grained
- individual beds and the interval overall decreases in the thickness
- fauna is much more restricted and bioturbation is less
- more bitumen is preserved,
- autochthonous dolomite decreases
- there is no evidence of early cementation such as occurs in ribbon limestone of the upper slope.

Fig. 3.29 illustrates the depositional conditions prevailing for the Llandoverian to early Wenlockian.

3.6.4 **Early to middle Wenlockian** (Fig. 3.30)

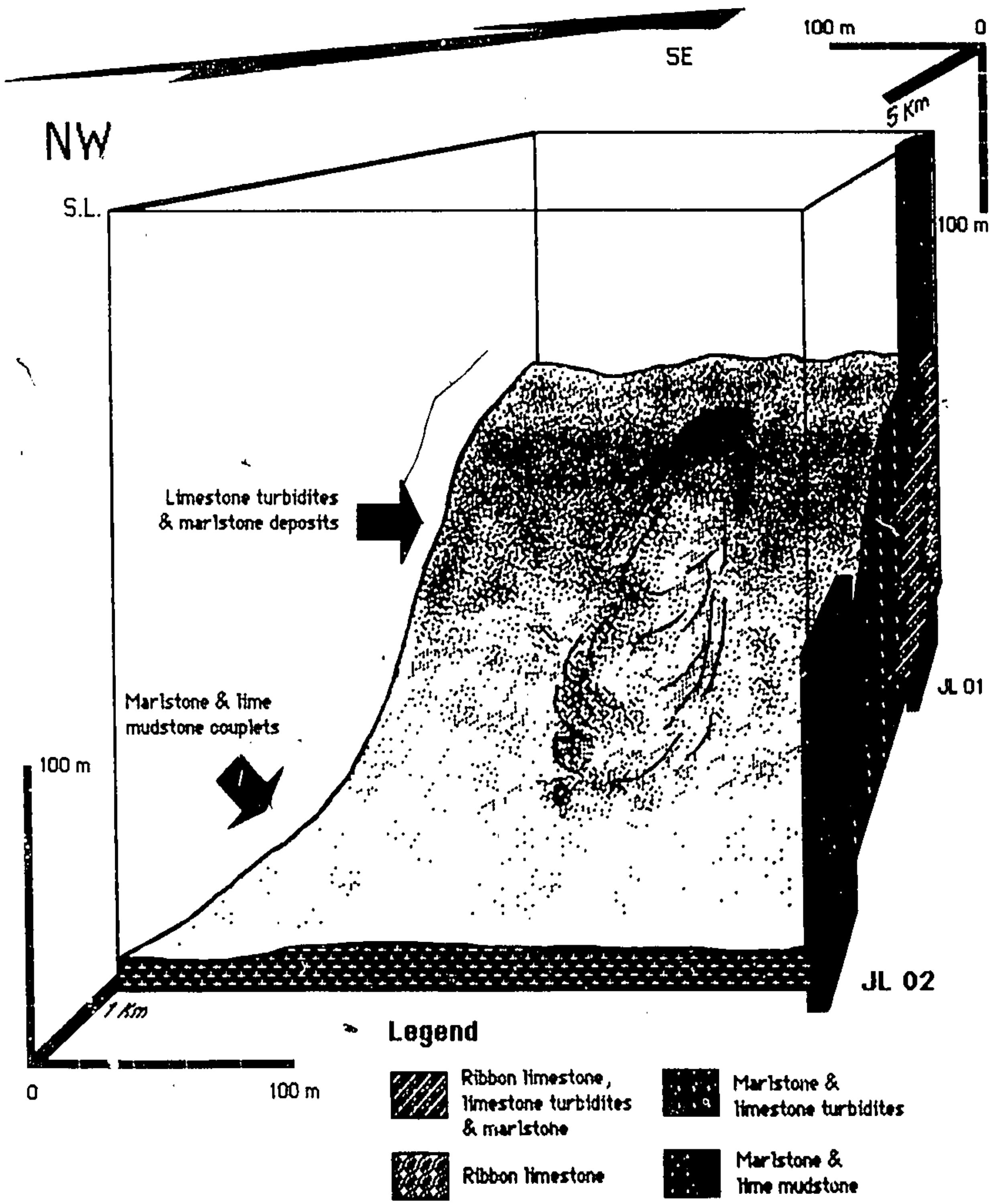
Most sections show evidence of increasing depth in the early to middle Wenlockian. However, sections further south (Mayr, 1978; Morrow and Kerr, 1977) are still coarsely dolomitized and show no evidence for deepening. Therefore it appears that only slope areas were affected, possibly through growth faults, following the original collapse. However, this contrast could also possibly be the result of slight rise of sea level, sufficient to have a dramatic effect on the fragile slope facies which are much slower to build up to their original levels since they were probably plunged under anoxic conditions, but not quite enough to affect the shelf depositional environments for a prolonged period of time, because the shelf quickly built up close to sea level again.

This apparent deepening is best demonstrated in sections at the southeastern corner of the study area (JL 01 & JL 10). There the underlying coarse flat pebble conglomerates gave way to monotonous lime mudstone-marlstone couplets in the early to middle Wenlockian, although on closer inspection the lime mudstones are in fact very fine and varied bioclastic limestone turbidites. Occasionally, in the middle Wenlockian interval, blocks of reefal origin (framestone and bindstone) in these very fine grained deposits reveal that bioherms were forming at the shelf-margin, although at that time apparently beyond the southeastern boundary of the study area.

Sections further northwest also consist of lime mudstone-marlstone couplets, but the slightly coarser-grained limestone turbidites quickly disappear, and marlstone becomes predominant. The sections furthest away are extremely bituminous and contain pyrite nodules, and together with an extremely restricted fauna, mostly graptolites, represent reducing conditions.

Overall, anoxic conditions prevailed over most of the slope and appear to have resulted from increased water depth. Fig. 3.30 demonstrates conditions prevailing at the time.

Figure 3.30 Early to middle Wenlockian



3.6.5 **Middle to Late Wenlockian** (Fig. 3.31)

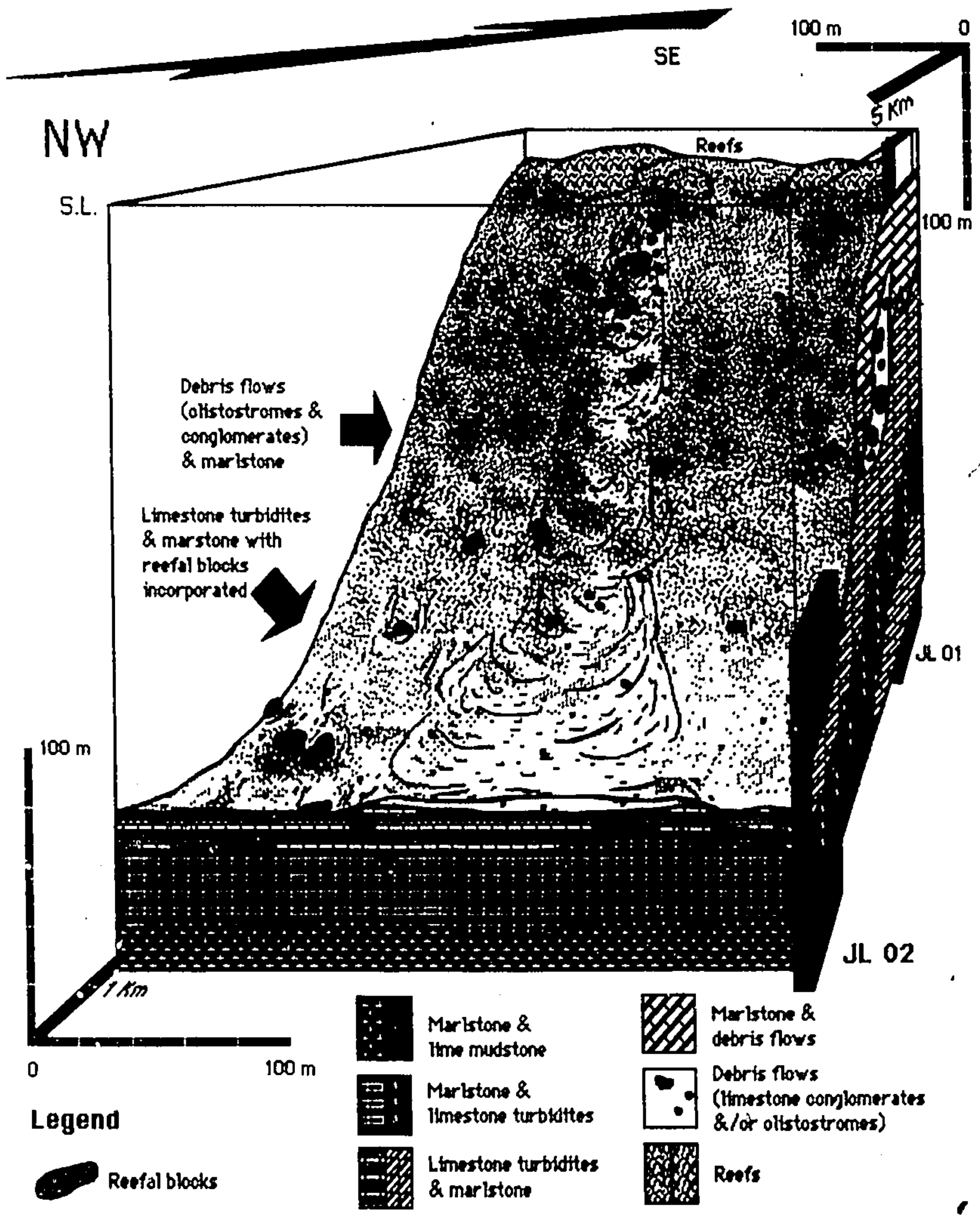
Following deepening, the principal reefal stage developed in the middle to late Wenlockian interval. It appears that sea level rise encouraged bioherm development, probably as the result of improved water circulation. During that time the shelf-margin returned to a position near the southeastern sections (JL 01 & JL 10) and a very large reef developed at section JL 06, the southernmost in the study area. At the bases of sections JL 01 and JL 10, a large or several smaller bioherms can also be observed. Several other large bioherms on Ellesmere Island are also of this age. In view of their size and the large amounts of debris, including olistostromes that formed downslope, these bioherms apparently formed at the shelf-margin itself. However, in section JL 04, a much smaller dolomitized bioherm, probably represents a downslope position.

Rapid outbuilding of the shelf-margin at this time created instability and resulted in derivation of bioclast-rich, thicker-bedded, and much coarser debris flow deposits, even olistostromes containing blocks up to 10 m in diameter, from the reef margin and occasionally the reef core.

Parts of the coarser fraction of limestone turbidites reached middle slope positions in sections JL 02, JL 04 and JL 07. However, lime mudstone-marlstone couplets remained important components in lower slope depositional environments.

In the remaining northwestern sections, lime mudstone-marlstone couplets are the most distal remains of the limestone turbidites, some of which started at least 40 km away. However, in contrast to the couplets deposited earlier, these are much purer lime mudstone beds and as a result are more resistant overall. They are also much less bituminous. The greater frequency of debris flows and turbidity currents, probably resulting from higher gradients and rapid outbuilding of the bioherm-dominated shelf-margin, would have promoted oxidizing conditions on this part of the slope and basin. Fig. 3.31 illustrates this model.

Figure 3.31 Middle to late Wenlockian



• The Allen Bay Fm-Read Bay Group undivided and Cape Phillips Fm •

3.6.6 Late Wenlockian to earliest Ludlovian (Fig. 3.32)

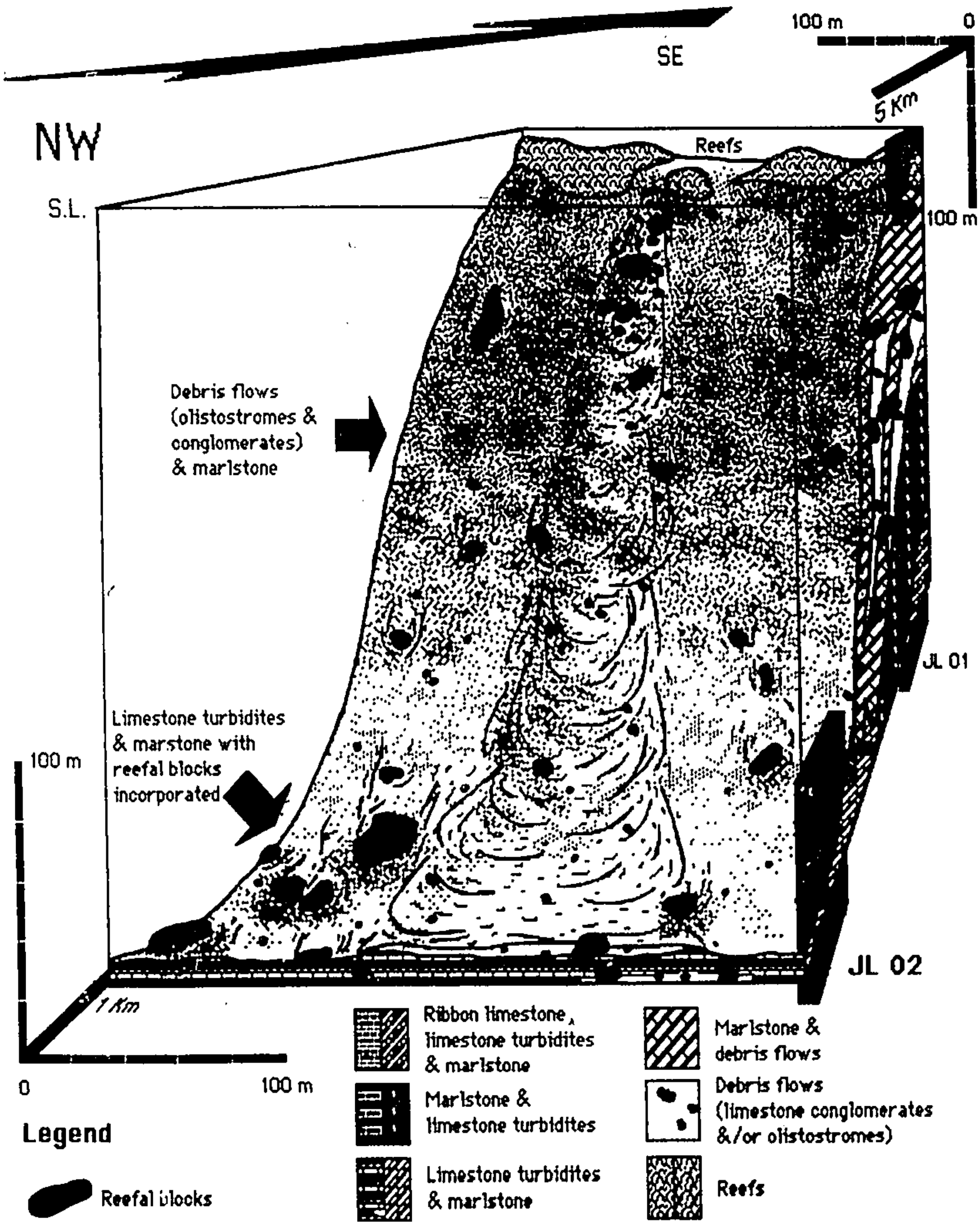
The late Wenlockian to earliest Ludlovian interval represents the climax of the reefal stage in the southeastern sections. Most of the bioherms in the study area appear to be of this age. The equivalent upper slope sediments are mostly massive olistostromes with large reef blocks and a matrix composed mostly of very coarse bioclastic packstone, representing principally the material from debris aprons surrounding the bioherms. Other finer types of debris flows occur in the slope deposits, although the intraclasts are similar in composition to the larger blocks, and are in a similar matrix. Commonly coarse limestone turbidites, beginning with a coarse bioclastic lag, gradationally overlie these debris flows.

The coarse debris flows extend only short distances down slope, suggesting that they were much more viscous and probably resulted from grain-flow rather than turbid flow. The sections representing intermediate slope positions are mostly very fine grained and commonly an order of magnitude thinner than the shelf-margin sections, as illustrated in Figures 3.4 and 3.32. They are not distinguishable from the sections composed of lime mudstone-marlstone couplets in the preceding interval.

3.6.7 Early Ludlovian (Fig. 3.33)

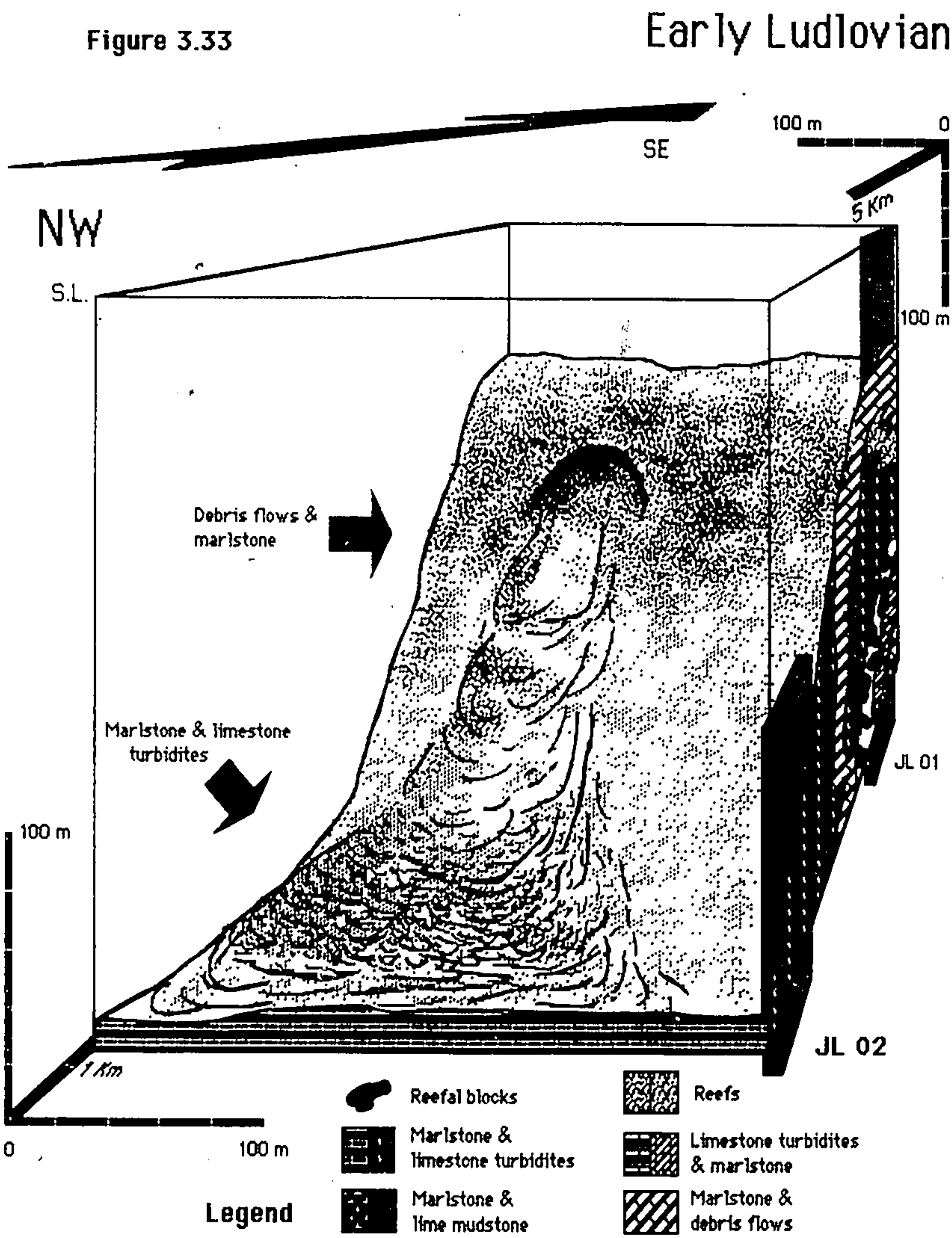
The shelf margin retreated briefly during the early Ludlovian, shown principally by rocks in sections JL 01 and JL 02. It appears that the relative rise of sea-level was sufficient only to influence sections at or near the shelf-margin. In section JL 02, for example, the coarse limestone turbidites interbeddded with minor debris flows changed to much finer limestone turbidites more similar to lime mudstone-marlstone couplets of more deeply located sections. In JL 01, the preceding extremely coarse and resistant debris flows are abruptly overlain by units of recessive fine-grained, limestone turbidites not unlike the ones lower down the slope. It appears that this elevation in relative sea-level, although insufficient to influence slope deposition, was nevertheless sufficient to interrupt or drown temporarily the bioherms growing at the shelf-margin, as represented by Fig. 3.33.

Figure 3.32 Late Wenlockian to earliest Ludlovian



• The Allen Bay Fm-Read Bay Group undivided and Cape Phillips Fm •

Figure 3.33


Early Ludlovian
NW
SE
100 m
0
5 Km
100 m
S.L.
Debris flows & marlstone
Marlstone & limestone turbidites
100 m
JL 01
JL 02
1 Km
0
100 m
Legend
Reefal blocks
Reefs
Marlstone & limestone turbidites
Limestone turbidites & marlstone
Marlstone & lime mudstone
Marlstone & debris flows

3.6.8 Late Ludlovian (Fig. 3.34)

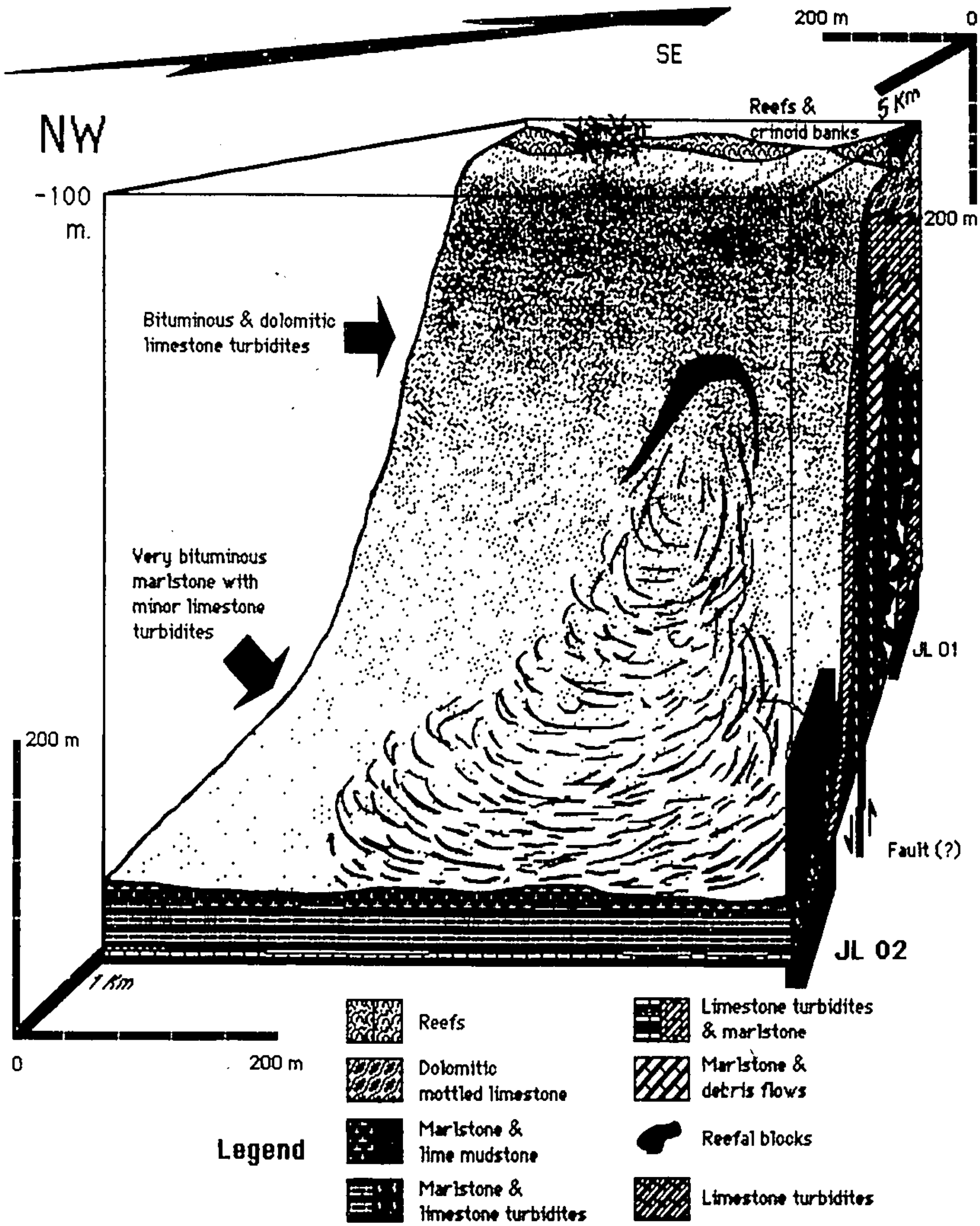
By late Ludlovian, bioherms were reestablished and some prograded into the study area. Apparently they were less well developed at the shelf-margin than previously, and the few that extend into the study area are less than 10 m in thickness.

Non-reefal deposits, instead, predominate in this interval. Bioherms in section JL 01 are enclosed by a wide variety of fine to coarse-grained limestones, some of them well sorted, many bioturbated. The rocks contain abundant and diverse body fossils, including some representing *in situ* benthos, and a trace fossil fauna that includes vertical burrows as well as the more widely common horizontal traces. These features collectively suggest shelf sedimentation with effective reworking by currents, good circulation, and at least periodically rapid accumulation of sediment. The absence of fining-upward sequences and of chaotic mass flow deposits is in striking contrast to underlying and laterally equivalent deposits of slope origin.

Equivalent rocks in sections JL 02, JL 07 and JL 04 are characteritically slope-deposited limestone turbidites of various coarseness, thickness and components. Some, such as in section JL 04, are turbidites of remarkably regular texture and uniform thickness. Often the limestone turbidites start with a thin lag of bioclastic fragments coming from a varied fauna although most of them are crinoidal. Sections further to the northwest are still mostly composed of lime mudstone-marlstone couplets, in the field apparently indistinguishable from underlying beds. However, these beds from both the middle slope and basinal sections contain more dolomite, silt-size quartz grains and silicification than underlying rocks, and are very bituminous, reflecting important changes in slope and shelf-margin conditions.

Figure 3.34

Late Ludlovian



The contact between the Allen Bay Formation and the Read Bay Group (Douro Formation) is considered to be in the late Ludlovian in this area, although the index fossils necessary to date it precisely are lacking. The contact is placed arbitrarily at the base of a thick unit of thin lime mudstone-marlstone couplets that appear nodular in their upper portions because of intense bioturbation. This unit also contains the brachiopod *Atrypoidea arcticus*, the first appearance of which is considered to be characteristic of the Read Bay Group.

3.6.9 Early Pridolian (Fig. 3.35)

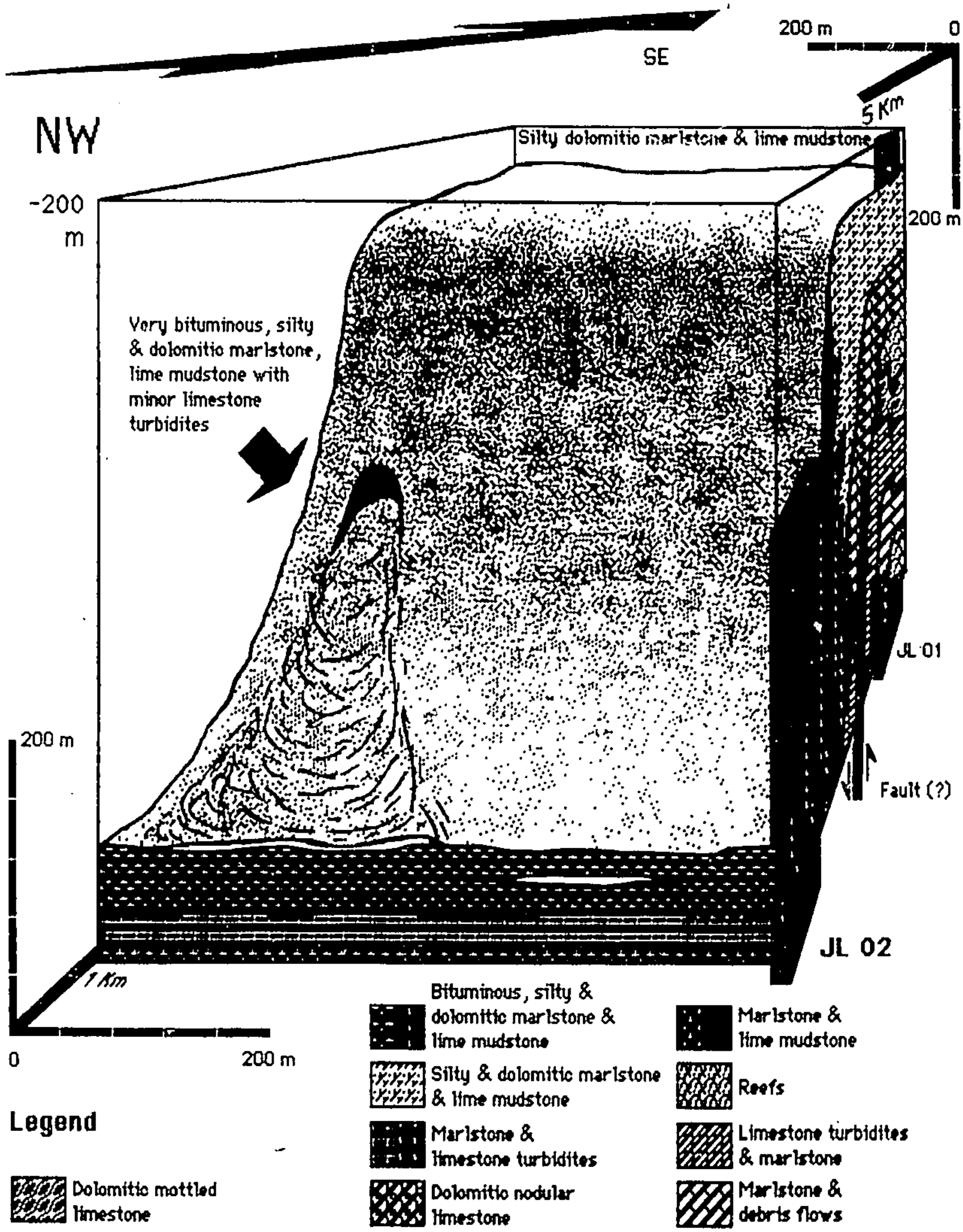
As Figure 3.35 illustrates, deposition during the early Pridolian, (and continuing with deposition of the Devon Island and Eids formations), appears to have been on a clinothem with relatively rapid sedimentation. As a result the slope eventually prograded, slowly filling in this basin.

The southeastern sections still contain shelf-margin facies, although probably representing somewhat deeper water environments. Minor coarsely dolomitized reefs continue to occur occasionally in section JL 01, accompanied by dolomitized debris aprons retaining identifiable structures and components. However the main lithology is a dolomitic nodular limestone with distinct nodules (mottled limestone type II), probably resulting from partial early cementation rather than from intense bioturbation, although the latter, too, is evident. This thick and homogeneous lithology is similar to the lithology of the Douro Formation in areas further south. The fossil fauna is varied. The lack of coarse and well sorted sediments reflects the absence of any sort of sorting mechanism, as would be expected on a deep shelf sea floor.

Farther north in the study area, fining-upward sequences of limestone turbidites continue to be the main lithologies. Although thin in comparison to previous limestone turbidites, a large number of them start with crinoid-rich bioclastic lag deposits.

Figure 3.35

Early Pridolian



• The Allen Bay Fm-Read Bay Group undivided and Cape Phillips Fm •

In sections situated mostly in basinal settings during this period, the main lithology is still marlstone, and for that reason the early Pridolian rocks are much thinner than in other sections. Although the shelf-margin does not appear to have migrated during this period, the increased water depth suggested by the shelf section also appears to have affected sedimentation in much deeper basinal areas. These deep basinal deposits are not as dolomitic or silty as the deposits underlying or overlying them or laterally equivalent to the southeast. It appears that silt content and dolomitization were shelf-related and occurred farther southeast as a result of deepening on the clinothem at this time. The toe of the slope and the basin were again very anoxic, as illustrated by the high amount of bitumen contained by these deposits and the very restricted fauna, entirely of graptolites.

Chapter 4

Devon Island Formation

4.1 Introduction

First described by Thorsteinsson (1963) at the type locality on Devon Island, the Devon Island Formation has been a source of controversy since, in particular concerning its lateral extent and age and its stratigraphic relationships with other formations. Its depositional environment is also not agreed upon and its designation as a distinct and mappable unit is questioned by some workers. However, some of these problems simply reflect difficulties of field identification due to weathering appearances in common with adjacent formations, and to very gradual lithological and paleoenvironmental transitions between the formations.

These other formations comprise (1) the Cape Phillips Formation, a distal slope facies mostly underlying the Devon Island Formation but which is also partly laterally equivalent basinward and in certain peculiar depositional environments, (2) the Imina Formation, the basinal "flysch" facies of Trettin (1979), a lateral equivalent, and (3) the Eids Formation, the slope and "post-flysch" facies of Trettin (1979), mostly overlying the Devon Island Formation. Even in the vicinity of the study area, the Devon Island Formation has rarely been identified as such, and equivalent beds have rather been incorporated as an upper member of the Imina Formation by workers such as Trettin (1979) and Kerr (1976) in the Cañon Fiord area, north of the present study.

Most studies in southern Ellesmere Island incorporating sequences of Late Silurian to Early Devonian age were large scale reconnaissance mapping projects, and so, understandably, the relatively thin Devon Island Formation, if measured sections in the study area and type locality can be taken as typical, was a casualty of the scale of mapping. More detailed field and laboratory studies, using fossils and thin sections, as undertaken here, leave very little doubt as to its mappability relative to surrounding formations.

Generally described as a very dark, silty, bituminous, thinly laminated limestone and dolostone or calcareous/dolomitic shale (Morrow, 1981), the Devon Island Formation also contains a few thicker, more massive beds of silty dolostone and bioclastic limestone, especially towards the top in regions east and south of the present study area (Kerr, 1976), although, as seen above, rarely identified as such. The high dolomite content gives it generally a very light weathering colour in comparison to the underlying Cape Phillips Formation, especially its upper member. These general characters also resemble, though, some parts of the formations mentioned earlier. The high dolomite and silt content cannot readily be distinguished in the field.

4.2 Distribution and Contacts

The Devon Island Formation is Late Silurian to Early Devonian in age (Morrow, 1981). In sections between the study area and the type locality, approximately 350 km to the south, it generally overlies the Douro Formation, part of the Read Bay Group (Morrow, 1981; Thorsteinsson, 1963). However, towards the northwest and the north-northwest, in deeper depositional environments, the Devon Island Formation conformably overlies slope and basinal deposits of the Cape Phillips Formation, and in those areas the lower contact as well as the formation itself become increasingly difficult to differentiate. Slightly lighter weathering colors, and higher silt and dolomite content (mostly recognizable in laboratory examinations) are the main criteria used in its identification. In the southeastern part of the study area (Figs. 4.1 and 3.4 a→e), it overlies the Allen Bay Formation-Read Bay Group undivided, and although the contact was not seen exposed there, Kerr (1974) has identified a conformable contact further north and northeast, although he referred the deposits of the Devon Island Formation to the Imina Formation. On Devon Island, the basal contact with the Douro Formation is sharp and unconformable (Morrow and Kerr, 1977), a local variation possibly reflecting proximity to the Boothia Uplift (Kerr, 1977).

The Devon Island Formation is overlain by different formations in different regions. In the study area it is overlain conformably by the Lower Devonian Eids Formation (Fig. 4.1), a sequence of orange-weathering, silty, dolomitic and bituminous limestone. In more basinal sections, towards the northwest, the Eids Formation takes on characteristics of the underlying Devon Island Formation, and thus becomes darker. In that area fossils are useful in mapping the contact (discussed below).

East and south on Ellesmere Island, the Devon Island Formation is overlain by the Vendom Fiord Formation (Fig. 4.2), part of a suite of siliciclastic sediments related to the Bache Peninsula (Kerr, 1976) or Inglefield Uplift (Smith and Okulitch, 1987).

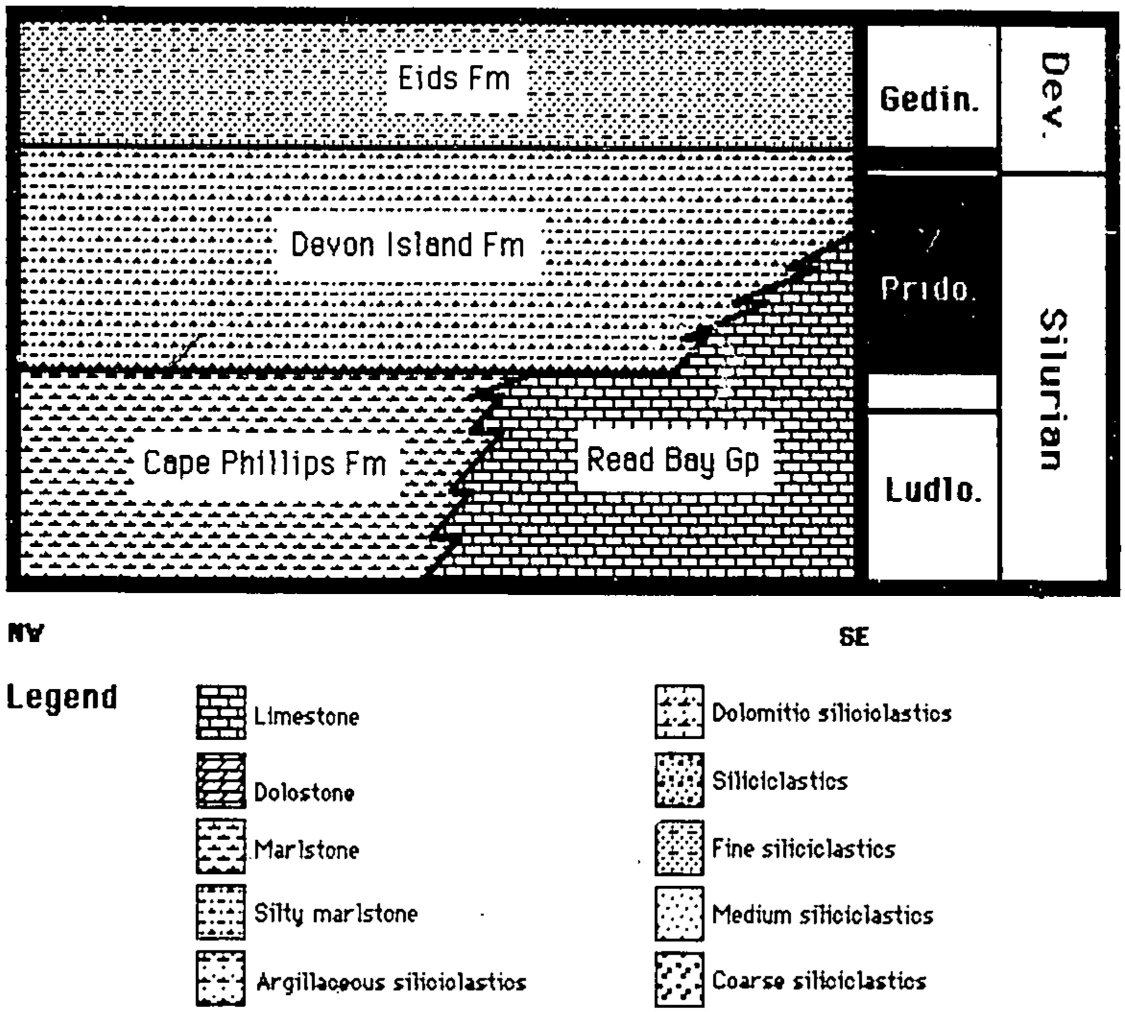


Figure 4.1 Stratigraphic relationships of the Devon Island Fm in the study area on Svendsen Peninsula. Black area represents age range of Devon Island Fm and equivalent deposits.

Towards the southwest, on northwestern Devon Island, the Devon Island Formation is conformably overlain by the Gedinnian Sutherland River Formation (Fig. 4.3) composed of very light grey, finely crystalline dolostone, variably fossiliferous or unfossiliferous (Morrow and Kerr, 1977). The two formations are separated by an unconformity on the part of northwestern Devon Island most affected by the Boothia Uplift.

The Devon Island Formation also has lateral equivalents throughout the Arctic Archipelago. This is partly due to lateral facies changes, and partly because in many areas the formation was not differentiated but incorporated with other formations. To the north, in the Cañon Fiord area, Trettin (1979) (Fig 4-4) included deposits of similar age in the early Gedinnian Imina Formation conformably overlying the Cape Phillips Formation.

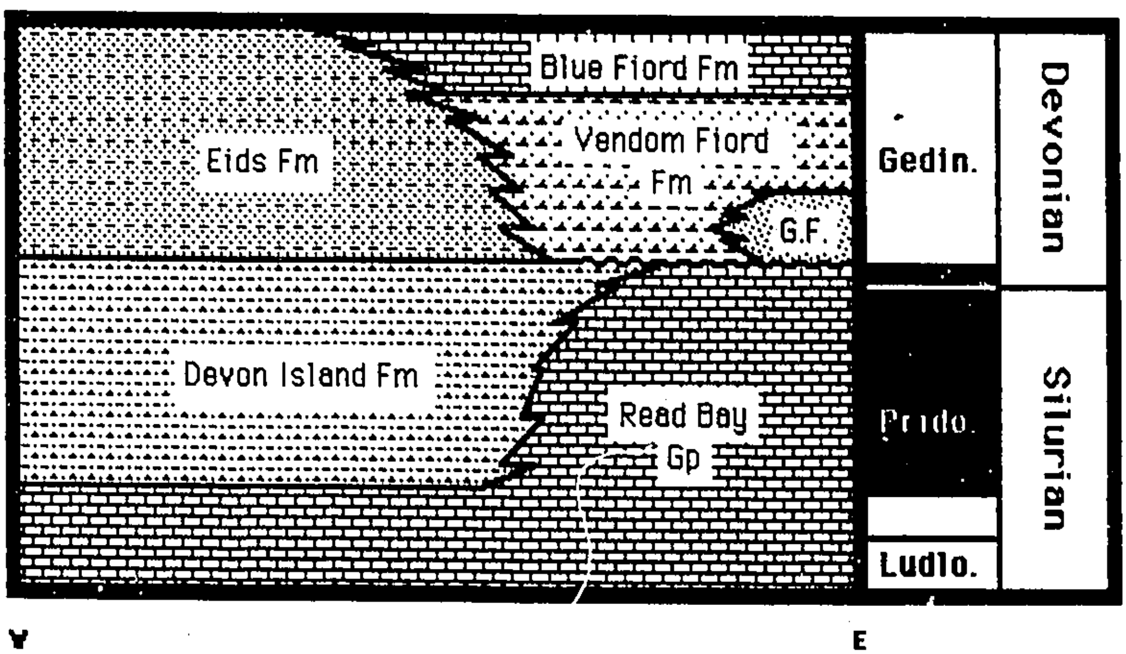


Figure 4.2 Stratigraphic relationships of the Devon Island Fm on Ellesmere Island, east of the study area (adapted from Kerr, 1976; G.F. = Goose Fiord Fm). Black area represents age range of Devon Island Fm and equivalent deposits. For key to symbols see Fig. 4.1 p. 176.

Similarly, the Devon Island Formation has not been distinguished toward the axis of the Hazen Trough, west of the study area. For example, at the northwest corner of Trold Fiord, Tozer (1963) recognized only two formations, the Cape Phillips and Eids formations, without differentiating an intervening formation. East of the study area (Kerr, 1976) the upper beds of the Read Bay Group (Fig 4-2) and possibly the very lowest beds of the Vendom Fiord and Goose Fiord formations are lateral equivalents of the Devon Island Formation.

South of this area studies are more numerous and detailed and the Devon Island Formation apparently has lateral equivalents in several formations on the southern flank of the Franklinian basin, as follows; the lower Eids Formation of southern Ellesmere Island (McLaren, 1963); the upper Cape Phillips Formation of northern Cornwallis Island (Mayr, 1978; Fig. 4.5); the Barlow Inlet and Sophia Lake formations of southern Cornwallis Island (Mayr, 1978; Packard, 1985); and the upper Cape Phillips Formation of Bathurst Island (Mayr, 1980; Fig. 4.6). Further south, in areas identified as the shallow shelf of the Franklinian Basin and/or the Stable Platform, other types of deposits represent the Pridolian and the Gedinnian, but are less relevent to the present study.

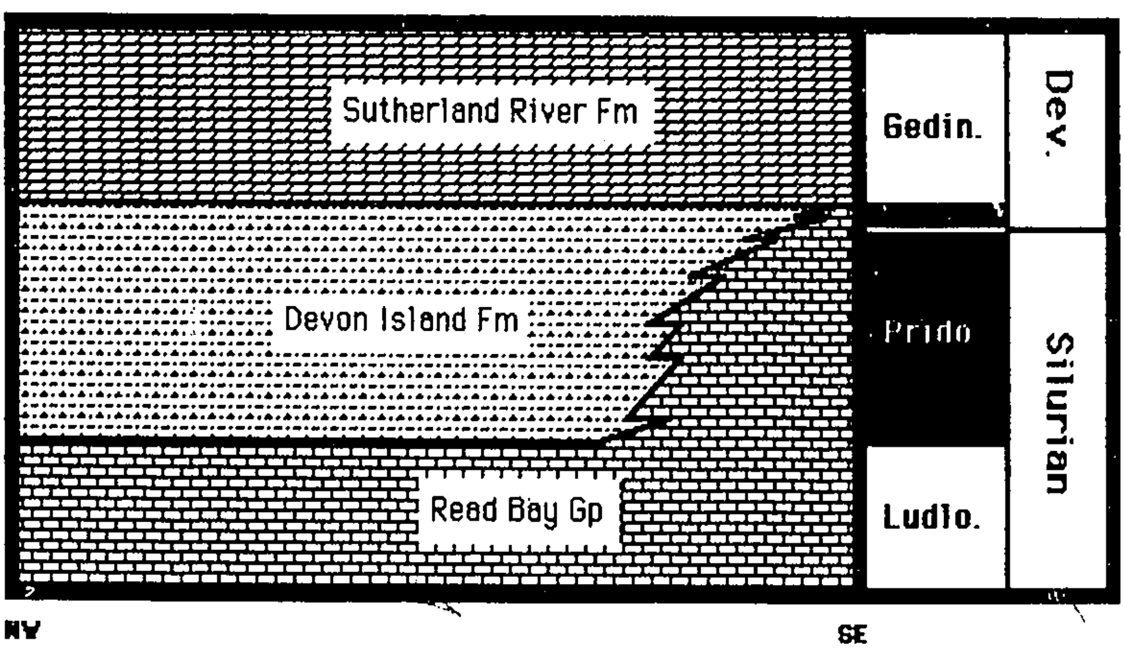


Figure 4.3 Stratigraphic relationships of the Devon Island Fm on Grinnell Penisula, Devon Island (adapted from Morrow and Kerr, 1977). Black area represents age range of Devon Island Fm and equivalent deposits. For key to symbols see Fig. 4.2 p. 176.

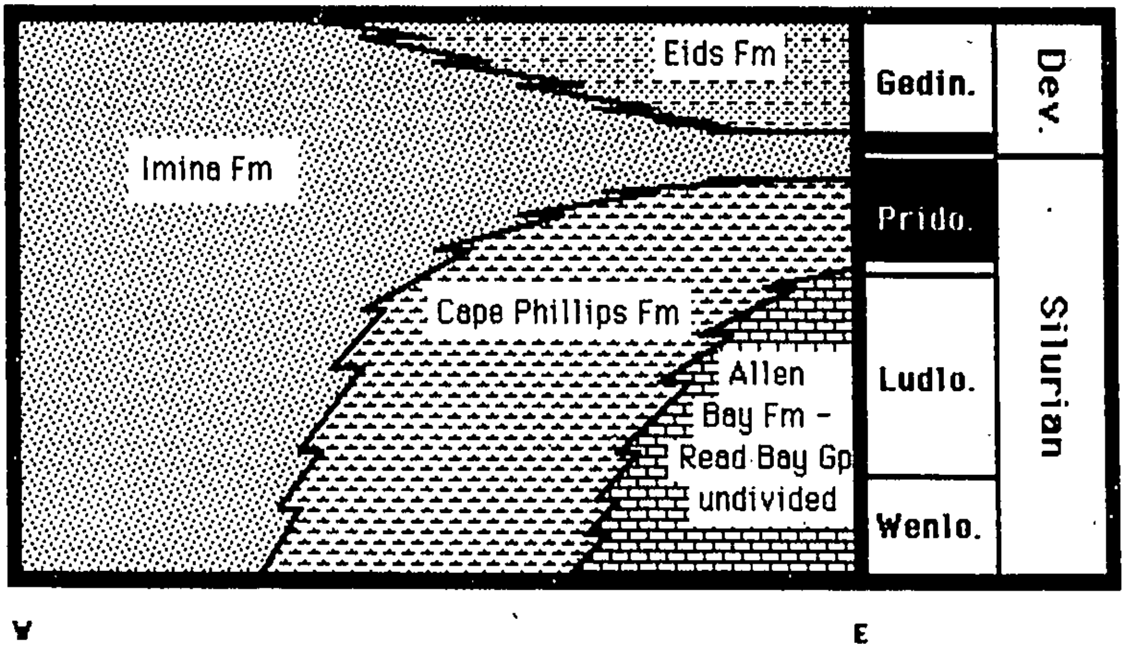


Figure 4.4 Stratigraphy in the Cañon Fiord area, Ellesmere Island (adapted from Trettin, 1979). Black area represents age range of deposits equivalent to the Devon Island Fm. For key to symbols see p. 176.

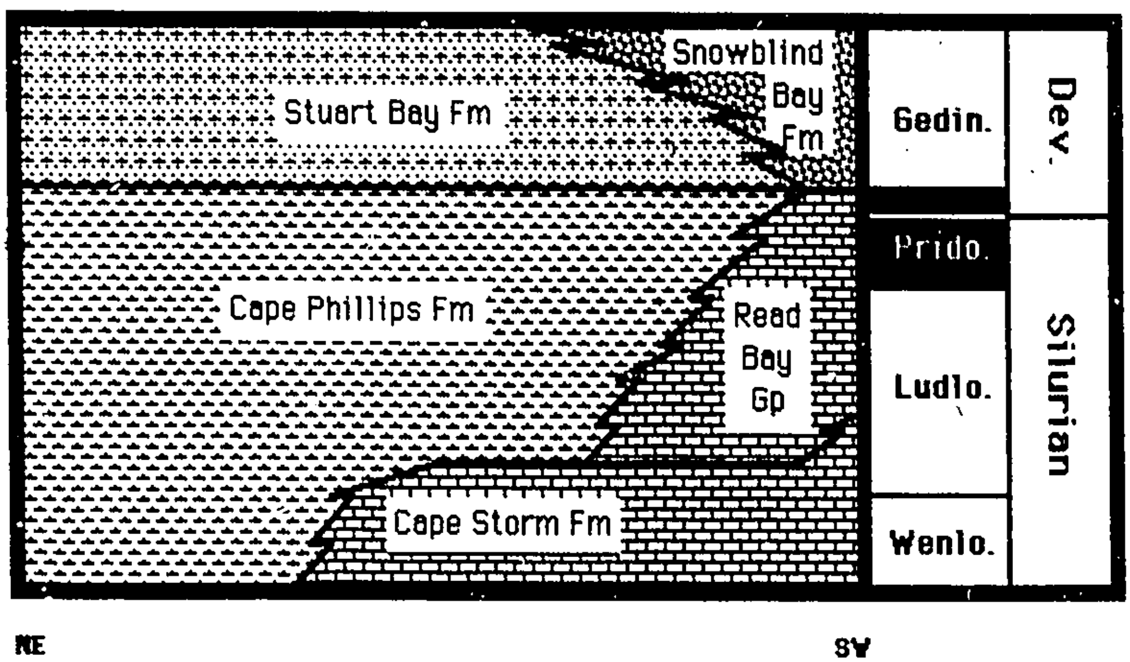


Figure 4.5 Stratigraphy on western Devon and Cornwallis islands (adapted from Mayr, 1978). Black area represents age range of deposits equivalent to the Devon Island Fm. For key to symbols see p. 176.

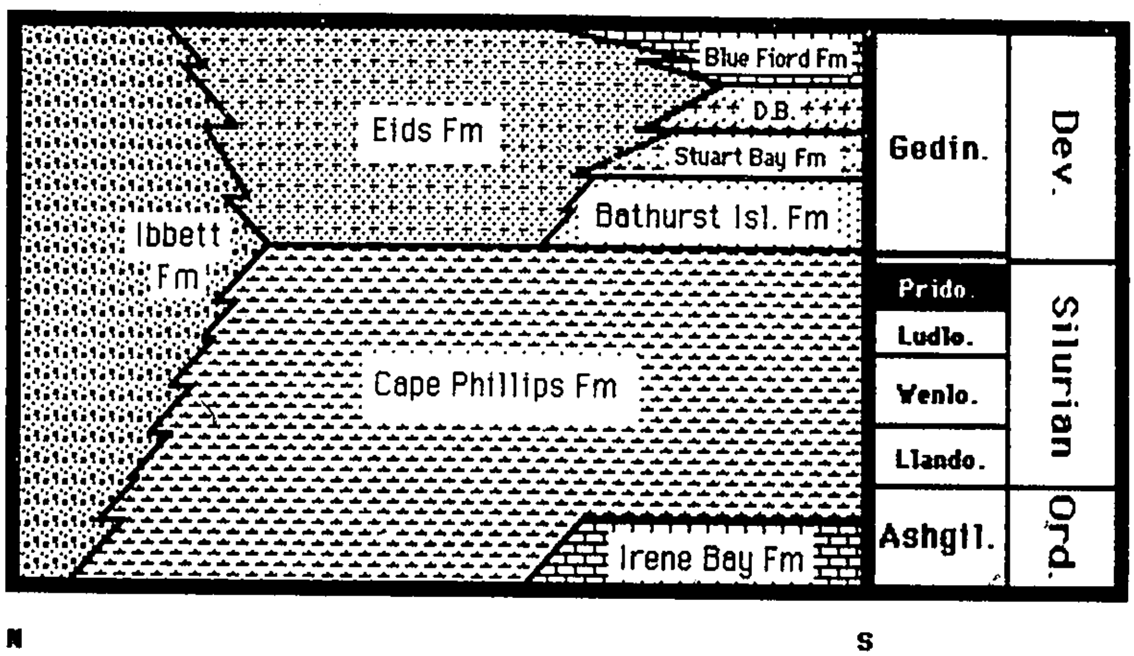


Figure 4.6 Stratigraphy of Bathurst Island (adapted from Mayr, 1980; D.B. = Disappointment Bay Fm). Black area represents age range of deposits equivalent to the Devon Island Fm. For key to symbols see p. 176.

4.3 Description

4.3.1 Field characteristics

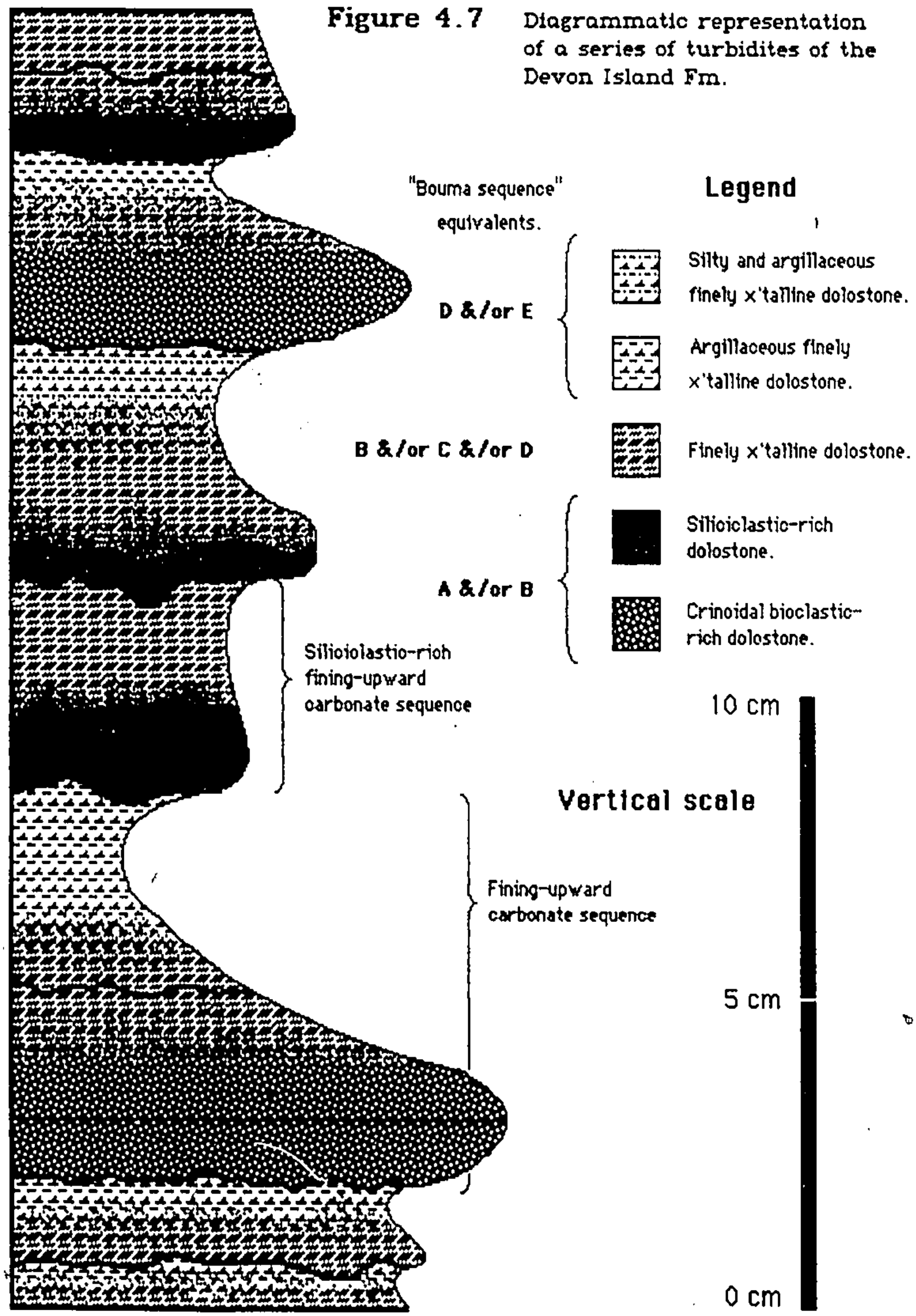
The sequence of beds designated the Devon Island Formation within the study area corresponds to the formation as described by Morrow (1981). Very thinly laminated to bedded carbonates with planar contacts predominate. The carbonates have thin and minor interbeds of rocks that, in the field, were designated calcareous and dolomitic shale or argillaceous limestone and dolostone. These argillaceous interbeds are more prevalent in, and form a transition zone in, the lower part of the formation where it rests on the distal slope-deposited Cape Phillips Formation. The argillaceous interbeds decrease markedly in abundance towards the middle part of the formation and again become a major constituent of the top half of the measured section. Only section JL 16 covered this upper part; most of the sections in the Devon Island Formation (JL 01, JL 03, JL 04, JL 07) exposed only its lower portion (see stratigraphic sections in Figs. 3.4 a→e). The argillaceous interbeds or marlstones are dark grey on fresh surfaces and slightly lighter where weathered, reflecting high interstitial bitumen content.

Carbonate layers are predominant in the lower and especially middle portions of this formation, and common in the top portion. They range from laminated to medium bedded, and all have sharp lower contacts. They have a very low argillaceous content, even where underlain and overlain by bituminous, dolomitic, and argillaceous siltstone, or marlstone. The absence of clay minerals is indicated by conchoidal fracturing, characteristic of beds without oriented grains or with equidimensional grains. The carbonate beds have a high bitumen content and freshly fractured surfaces emit a strongly petroliferous odour. The bitumen is intergranular and intercrystalline and gives a very dark brownish colour to fresh surfaces. Weathering surfaces vary from light cream to dark orange and the orange beds, even though subordinate in abundance, typify the Devon Island Formation, and contrast with the underlying Cape Phillips Formation. Some beds of the overlying Eids Formation have similar weathering colours. This characteristic weathering colour is due to dolomitization, as most of these more massive deposits are actually calcareous dolostone. The dolostone beds are commonly silty.

4.3.2 Petrography

Four main rock types can be distinguished in thin section: (1) fining-upward carbonates, (2) fining-upward carbonates containing a high proportion of siliciclastics, (3) finely crystalline dolostone, and (4) argillaceous carbonates. Figure 4-7 is an example of the field appearance and relationships between those four main rock types. Fining-upward carbonate sequences are volumetrically most important but each of the four types is sufficiently abundant to be important in interpretation of the depositional environments. The four types are intergradational but certain gradations seem to be more common then others. They are also often combined in units that are interpreted here as single mass flows.

Figure 4.7 Diagrammatic representation of a series of turbidites of the Devon Island Fm.



The *fining-upward carbonate sequences* generally have sharp basal contacts, but no sole marks were observed. A complete sequence starts as a thin basal grain-supported deposit consisting mostly of calcareous allochems. This basal portion is rarely thicker than 1 cm and mostly thinner than 0,5 cm. The allochems are mostly crinoid ossicles with rare ooids and more common dark micritic peloids. The allochem-rich layer is abruptly succeeded by a nearly completely dolomitized matrix-supported layer of fine (20 μm) equigranular and euhedral dolomite rhombohedra that makes up the remainder of each sequence. The matrix between the grains of the lower grain-supported portion was dolomitized, but the grains, including the peloids, remained unaffected even though they were probably originally similar in composition to the matrix. In some areas calcareous grains are isolated in the matrix of dolomite crystals. Bitumen filling a network of very fine micropores and some intercrystalline spaces gives a very dark colour to these deposits. Disseminated silt-size quartz grains constitute no more than 5% of the rock. Pyrite is present in minor amounts. The edges of crinoid fragments are occasionally silicified. The fining-upward carbonate sequences, even where complete, usually are less than 10 cm thick, and average only 3 cm. They are mostly thin-bedded and flaggy weathering. They weather light greenish grey to orange and are light to dark grey on fresh surfaces, depending on the bitumen content. Fig. 4.8 summarizes the up-sequence changes in relative abundance of its components.

Siliciclastic-rich fining-upward carbonate sequences are difficult to distinguish from the previously discussed fining-upward carbonate sequences without petrographic examination because the quartz and feldspar grains are mostly silt-size and are not the main component. Their weathering colour is typically more orange than the purely carbonate sequences, but not exclusively so. Upward gradation within these siliciclastic sediments is expressed more by gradual change in components than by change of grain size. These deposits typically start with a siliciclastic-rich bottom layer, with grains of sand- to silt-size quartz and minor feldspar forming up to 60% of the components. The grains either form a supporting framework organized as a series of very thin (less than 100 μm) and elongate lenses of siltstone and fine sandstone cemented by calcite or dolomite and also containing large concentrations of peloids as shown in Fig. 4.9, or are isolated among dolomite rhombohedra and are mostly well sorted and subrounded to subangular. The basal layer is transitional into an upper layer consisting of as much as 100 percent dolomite rhombs. This dolomitic upper part forms the main body of these siliciclastic-rich fining-upward carbonate sequences as shown in Fig. 4.10. Its siliciclastic content is, however, more diagnostic of this type of deposit. The dolomite rhombs are uniform in size throughout each sequence, mostly very finely crystalline, fairly euhedral, and homogeneously distributed. Coincidentally, they are approximately the same size as the siliciclastic grains.

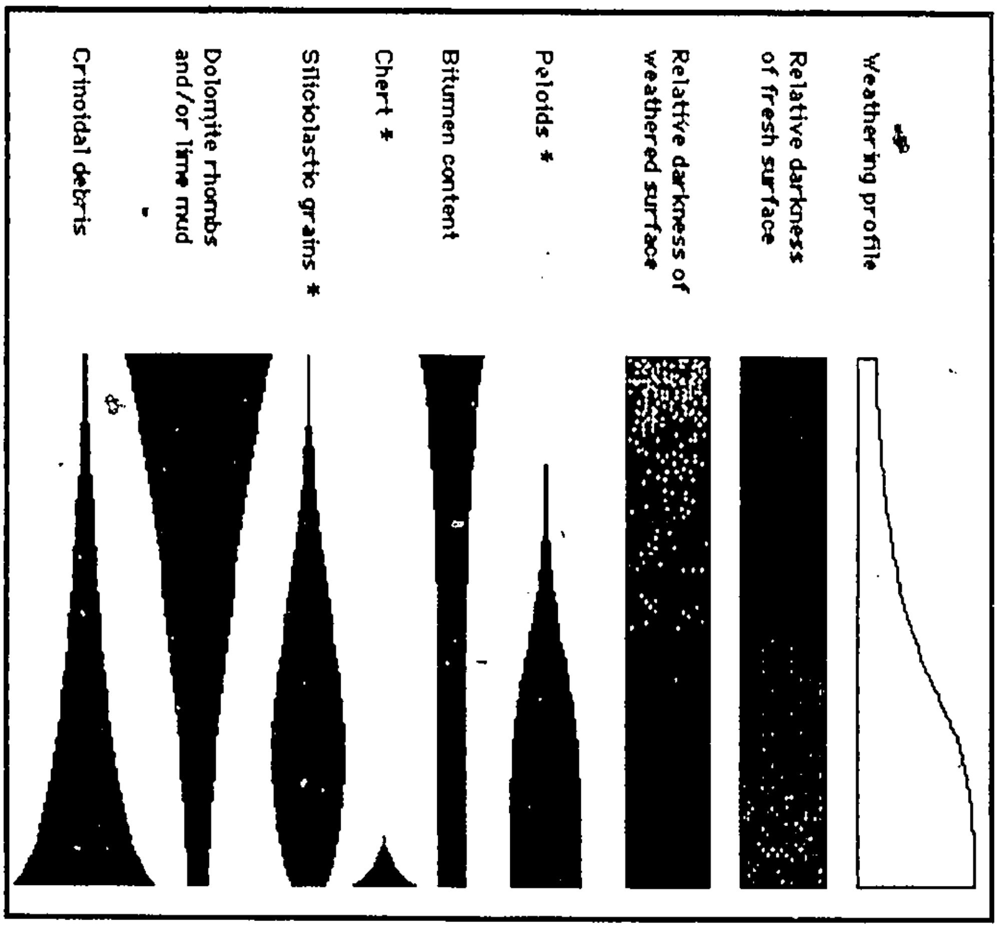


Figure 4.8 Relative abundance of components in a bioclast-rich fining-upward carbonate sequence of the Devon Island Fm (* - components sometimes absent). The width of the bars is exponentially exaggerated towards the bottom to show the trends clearly.

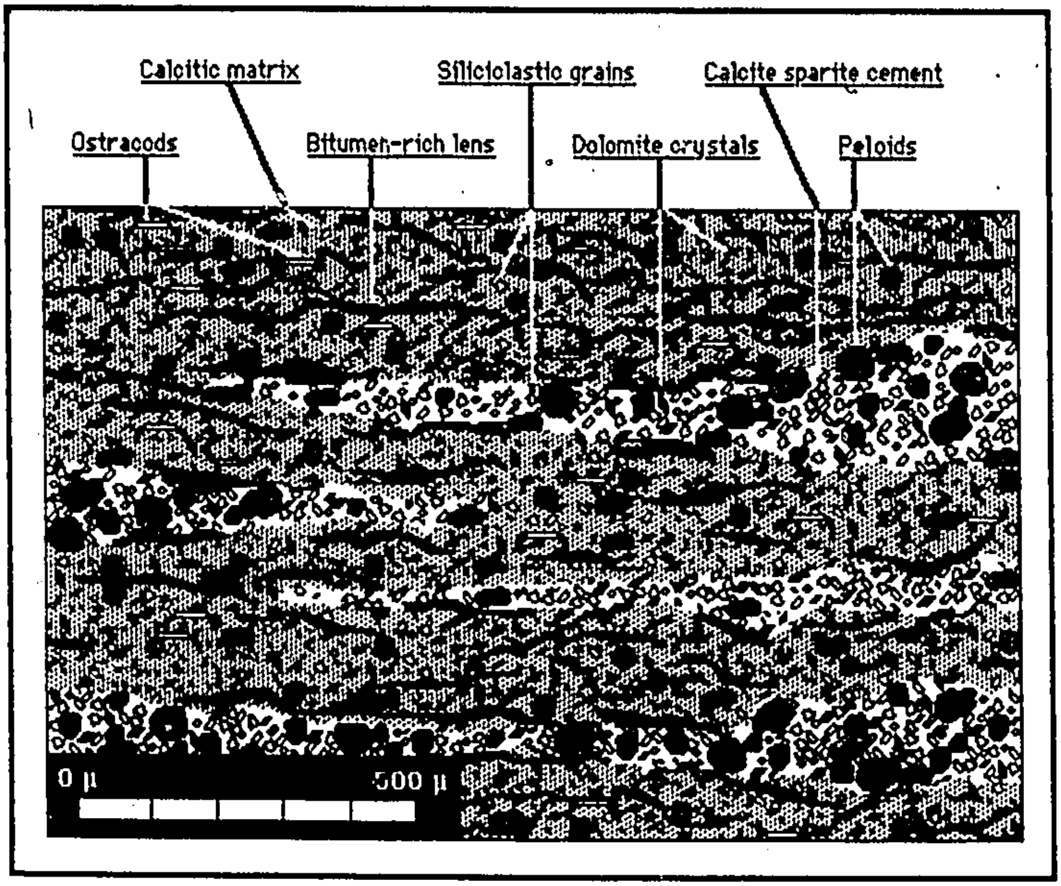


Figure 4.9 Diagrammatic representation of the bottom portion of siliciclastic-rich fining-upward carbonate sequence, where most of the grains are concentrated in thin, elongate lenses.

Other components in the siliciclastic-rich carbonate sequences include calcareous fossils, mostly crinoid fragments, concentrated in the lower portion. Horizontal or subhorizontal burrows occur in the dolomitic portions, generally are more abundant upward, and in some beds extensively disrupt layering. They can be distinguished by their generally darker weathering colour, reflecting greater intercrystalline bitumen content. Bitumen also fills minor pore spaces between dolomite crystals outside the burrow fills. Some of these sequences have a fairly high calcite content, either as patches of carbonate mud distributed through the dolomite or as cement. In the field, the reaction of this calcite to acid made it difficult to determine that these beds are in fact substantially dolostone.

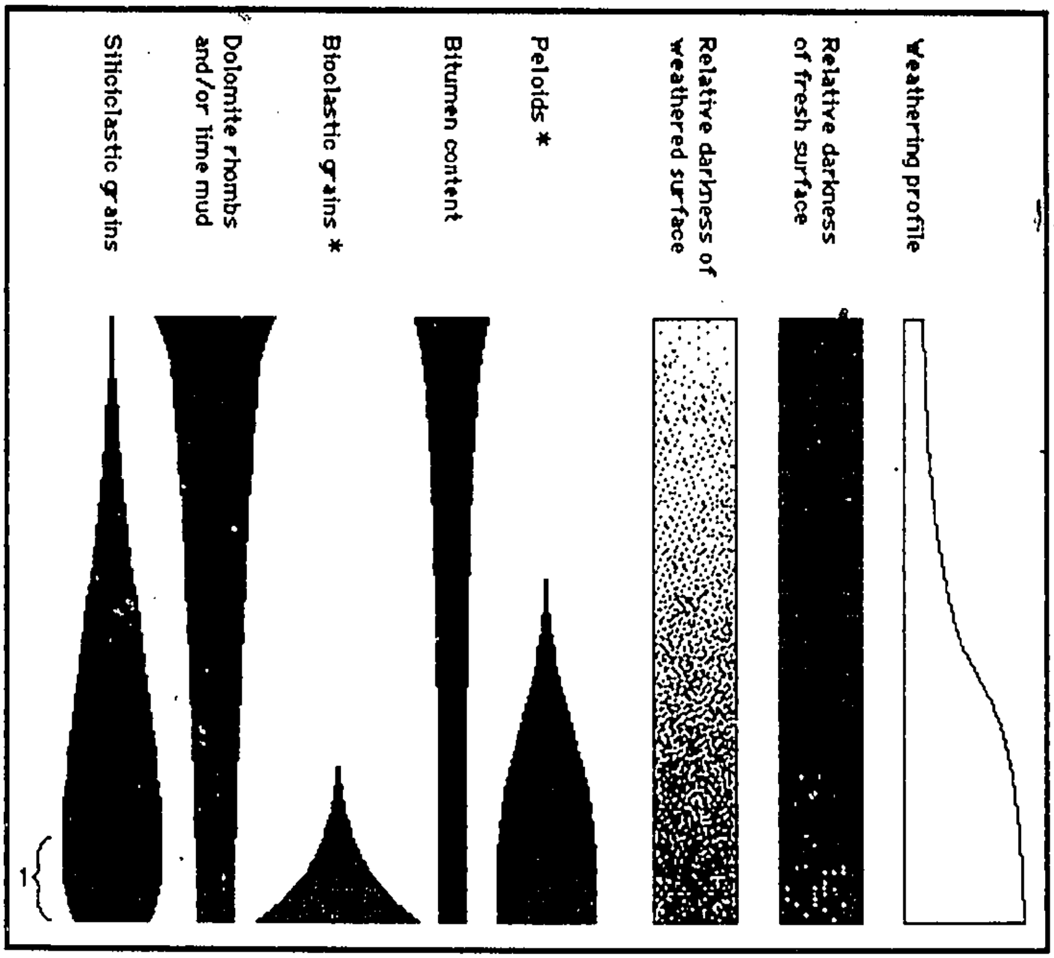


Figure 4.10 Relative abundance of components in the siliciclastic-rich fining-upward carbonate sequences of the Devon Island Fm (* - components sometimes absent; 1 - transitional zone with a bioclastic-rich bottom bed). The width of the bars is exponentially exaggerated towards the bottom to clearly show the trends.

The lower contact of each siliciclastic-rich carbonate sequence is generally non-erosional and distinguished in thin section by a moderately sharp transition from a dolostone to a siliciclastic-rich dolostone. Individual sequences vary greatly in thickness but average about 2 cm. This constitutes thin bedding, and internal lamination is commonly evident with a petrographic microscope. In the field, these rocks appear to be medium or thick bedded since sequences are commonly grouped and cannot readily be distinguished because grains are too small to distinguish and generally no sharp contacts separate the sequences. These flaggy to massive weathering beds are the most resistant portions of the Devon Island Formation.

The two types of fining-upward sequence described above share a common diagnostic character that serves to identify their gradual up-sequence changes even though their components are generally too fine to identify in the field: the upward increase of dolomite content is responsible for up-sequence colour changes from very dark grey to light cream weathering, despite the fact that bitumen content increases up-sequence.

A third, and volumetrically important rock type that can be distinguished in thin section is a thin bedded, bituminous, *finely crystalline dolostone*. It commonly forms discrete, conchoidally fracturing beds overlying either type of fining-upward sequence, or forms the uppermost portions of the few complete fining-upward sequences, or even forms individual beds not directly in contact with the fining-upward sequences. Although this dolostone resembles in composition and grain size the dolostone in the upper parts of the sequences first described, it distinctively contains more abundant and less fragmented fossil allochems, and fossils such as graptolites can be identified readily.

Finally, *argillaceous carbonates* are intercalated with the other rock types and commonly alternate with the pure carbonate and siliciclastic-rich carbonate sequences. They are the least resistant of all the rock types, are generally very thinly laminated and flaggy-weathering, and resemble calcareous shales more than argillaceous carbonates. Most are very dark in colour, although some are buff coloured and more resistant and fracture conchoidally. These buff coloured beds are mainly composed of calcite with less clay mineral content evident in thin section than was expected from field appearances. They are rich in bitumen, mostly distributed in thin, minute lenses. The calcite is mostly neomorphic pseudospar, of two distinct forms: most occurs as small spheres 10 - 150 μm (mean 30 μm) in diameter and the remainder as elongate lenses averaging 150 μm thick. The spheres are mostly homogeneously distributed, but concentrated in some beds more than others. They show no evidence of compaction, but clay minerals are wrapped around them. They appear to be diagenetic but their mode of origin is unknown. Dolomite rhombs are minor components in most of the beds but locally are abundant enough to form argillaceous dolostone.

These argillaceous carbonates incorporate most of the graptolitic fauna of the Devon Island Formation, but other body fossils are extremely rare. Numerous trace fossils are evidence, however, of a flourishing benthic fauna during periods of calm and diminished deposition. The filling of the burrows is mostly bituminous dolostone similar to deposits overlying these argillaceous carbonates.

4.4 Fossils

In most of the sections measured, the Devon Island Formation overlies the Cape Phillips Formation. Above the contact faunal diversity increases, accompanied by a minor increase in abundance of body fossils. This is the main factor in recognizing the contact since little lithological differences exist between the deposits of the Cape Phillips and Devon Island formations. Where the Devon Island Formation overlies the Read Bay Group (Douro Formation), the upward changes are mainly lithologic and involve either little change or even a decrease in fossil abundance and diversity.

Fossil are typically rare to very rare in occurrence in the Devon Island Formation. However, most fossils are whole or fragments large enough to be identified confidently, and this gives the impression that fossils are more abundant than in the underlying beds.

Graptolites, the predominant fossils in the Cape Phillips Formation, continue, but are reduced in abundance and diversity in the Devon Island Formation. This may be a reflection of their ubiquitous gradual disappearance during the Devonian, but may also be related to changing environments and rates of deposition. They are generally poorly preserved, often only as carbon films, and rarely as pyrite replacements. Among the more common and important graptolites are; *Monograptus priodon*, *Monograptus flemingii*, *Monograptus vomerina vomerina*, *Cyrtograptus lundgreni*, and *Stromatograptus* sp.. Among other species of *Monograptus* is *bohemicus tenuis* (also known as *Pristiograptus bohemicus* (Barrande, 1850) or *Bohemograptus bohemicus* (Pribyl, 1967)), an important index fossil thought to mark the base of the Devon Island Formation with its first occurrence (Thorsteinsson, pers. com., 1985), and approximately the Ludlovian-Přidolian boundary.

Brachiopods in the Devon Island Formation are generally smaller and more thin shelled than those in coarser beds of the Allen Bay Formation-Read Bay Group undivided or the Cape Phillips Formation. They are evident on bedding planes, generally oriented parallel to bedding and are usually very well preserved, commonly still hinged. Some are costate. Some very fragmented shells also occur in the basal portions of mass flow deposits. Similarly small, very thin shelled, smooth bivalves are also present in minor amounts. They too, are excellently preserved and oriented parallel to bedding, but are mostly unhinged.

Ostracods are more common than representatives of the other two bivalved groups. Most are not apparent in the field, except for local very thin shelled and very well preserved giant ostracods up to 0,8 cm in length. The basal portions of some mass flow deposits also contain fragmented ostracod shells. Well preserved spines and, less commonly, shells of phyllocarid arthropods occur rarely on bedding planes in the Devon Island Formation.

Orthoconic cephalopods, which were not observed in the Cape Phillips Formation and are rare in the Allen Bay Formation-Read Bay Group undivided, are common in the Devon Island Formation. Cephalopods in similar abundance, have been observed elsewhere in the study area, only in the Irene Bay Formation. In contrast the Devon Island cephalopods are much smaller, averaging only about 4 cm long, and are largely compressed forms on bedding planes. Some well preserved specimens are pyritized.

Lithistid sponges occur rarely but are excellently preserved. They include cylindrical forms averaging about 4 cm in diameter and of uncertain length, that are oriented either perpendicular or parallel to bedding. Their good preservation without compression suggests early recrystallization. They are now completely composed of chalcedony, with the silica probably derived from original spicules of opaline silica. Despite recrystallization, the central osculum, and an original radial pore pattern are distinct. Individual spicules recrystallized to chalcedony or replaced by coarse calcite spar are evident only rarely.

Fine crinoidal debris is present, but none is large enough to be identified in the field. Most occurs as the main constituent in grain-supported basal portions of turbidites (a minor part of the succession). Evidently crinoids did not live in these depositional environments and their remains are entirely allochthonous.

Trace fossils are common and provide some of the best evidence for biological diversity and activity during deposition of the Devon Island Formation. Unlike the burrows of underlying formations, where modification of more argillaceous fillings by differential compaction, dissolution, or various diagenetic processes commonly left little evidence of original biological activity, burrows in the Devon Island Formation have generally kept their original shapes and are similar in composition to the surrounding sediments. A darker weathering colour shown by the burrow fills perhaps reflects a higher bitumen content, but this was not confirmed by petrographic examination.

Species of *Phycodes* are most common and *Planolites* sp. and *Chondrites* sp. occur more rarely. Each of these genera includes forms with distinctly different diameters, and the assemblage as a whole represents a more diverse community of burrowers than initially apparent.

Except for the transported crinoidal debris, the fossils of the Devon Island Formation differ from those of immediately underlying formations, in their typically excellent preservation and minor fragmentation. The Devon Island assemblage is also more diverse. Some fossils (sponges) are upright and perhaps in life position. Most, however, lie flat, but with no preferred orientation. The fossils were affected by post-depositional processes including silicification (important for the sponges), pyritization (rarely in graptolites and cephalopods), and minor recrystallization.

4.5 Depositional environments and processes

4.5.1 Significance of organic material

The most pervasive character of the lower Devon Island Formation is its dark color, caused by a substantial content of organic material. Several factors can contribute to the preservation of organic matter in sediments, but not all are easy to assess in the geological record. Influential factors include:

- biological productivity and related aspects of climate and ocean circulation
- sedimentation rates
- tectonic activity
- oxygen levels in oceans

Firstly, *biological productivity* on land is favoured by humid climates. Humid climates also contribute to fluvial transport of resulting organic detritus to the marine realm, where it can settle as hemi-pelagic detritus at all water depths. Substantial biological productivity in the oceans requires well-oxygenated waters. Usually the principal result is a "rain" of pelagic organic detritus from agitated, near-surface waters. Nutrient replenishment is also required, either from land or through upwelling. The specifics of land derivation or of ocean circulation, however, can be difficult to evaluate in the geological record. The large quantities of preserved organic matter in the lower Devon Island Formation do imply, however, substantial contemporary biological productivity.

Secondly, *rates of sedimentation* can influence the preservation of organic detritus in several ways. Rapid sedimentation and burial tends to reduce the chances of biological degradation on or near the sea floor, and increase the chance that organic matter will be preserved. Conversely, rapid sedimentation tends to "dilute" the accumulating organic material by spreading it through a greater volume of sediment. Higher rates of sedimentation, usually of coarser sediment, also commonly accompany increased water circulation, improved oxygenation, and consequently increased biological degradation of organics. The predominance of fine argillaceous carbonates in the lower Devon Island Formation tends to indicate low energy conditions and relatively slow deposition, except for the minor episodes when bioclastic and siliciclastic fining-upward units were introduced.

Thirdly, *tectonic activity* influences accumulation of organic material through its influence on rates of erosion on land, introduction of siliciclastic sediment to ocean basins, and rates of sedimentation. North (1979) noted that, assuming potential hydrocarbon source sediments could have been formed uniformly through the Phanerozoic, they became hydrocarbon-rich, effective source deposits only during particular episodes of extreme tectonic activity or environmental conditions. He cited as examples, the Caledonian, Hercynian, Cimmerian and Alpine Orogenies, which were characterized by:

- high tectonic activity, with extended orogenesis
- high rates of sedimentation in deepening basins
- little non-marine sedimentation
- humid climates
- abundant carbonaceous detritus and planktonic faunas

Finally, *oxygen levels* in the oceans are a major influence on preservation of organic detritus. Strongly reduced oxygen levels or anoxic conditions are certainly implied for most of lower Devon Island Formation deposition, collectively by large organic content and commonly preserved lamination of the argillaceous carbonates and the predominance of planktonic fossils in a fauna of low diversity. In anoxic environments, organic detritus can accumulate because organisms responsible for aerobic degradation are excluded. Anaerobic degradation by bacteria and algae is also known to be effective, but this influence is diminished without burrowers to rework and expose the sediments, or with a continued substantial supply of pelagic detritus.

Many acknowledged hydrocarbon source rocks have been correlated with upwelling (North, 1979). Periodic upwelling of anoxic but nutrient-rich bathyal waters to upper slope or deep shelf settings would have the initial effect of killing benthic epifauna and infauna, but it would encourage organic productivity higher in the water column where mixing with oxygenated, near-surface waters took place. The resulting increased "rain" of pelagic organic detritus could be preserved with prolonged anoxic bottom conditions (Jones, 1983).

Anoxic or oxygen-poor conditions are commonly, but not exclusively, associated with greater depths in the oceans. Oxygen generated in surface waters by photosynthesis is consumed by animals at all depths, and if not replenished by circulation, is progressively depleted, particularly at greater depths. An oxygen-minimum layer, therefore, can be formed and its upper limit can lie anywhere from ocean basin depth up to shelf depth. The upper limit can also change in position, for example, with sea level changes or changes in patterns of ocean circulation. An oxygen-minimum layer can also be underlain by more oxygenated bottom waters introduced by currents related to turbid flow or thermohaline circulation. A modern example has been reported on the continental slope off the western shelf of India (Stackelberg, 1972).

According to a theory of Berry and Wilde (1978), the abundance of black shale or lime mudstone in the Silurian could be due to warmer global temperatures at the time. The resulting absence of polar ice caps and of related thermohaline circulation, could have resulted in persistent anoxic bottom conditions, an oxygen-minimum layer extending up into shallow depths, and more widespread preservation of organic deposits.

In summary, the rich organic content only provides evidence that the Devon Island Formation was deposited generally in oxygen-depleted or anoxic, low energy conditions, but does not indicate depth of water. Additional information on depositional environment can be derived from the associated sediments.

4.5.2 Significance of other deposits

Although the remaining deposits in the lower Devon Island Formation are minor in proportion to the argillaceous carbonate beds, they provide useful additional evidence of depositional environments. These lighter-weathering, thicker-bedded, fining-upward carbonate sequences and more homogeneous finely crystalline dolostone beds apparently represent sediments transported and deposited by turbidity currents. They generally do not, however, show the entire range of sedimentary structures characteristic of the Bouma sequence, that was based on siliciclastic turbidites (Bouma, 1962). Usually only grading and some lamination are present. In this sense, they are like turbidites reported by MacKenzie (1970) in Devonian slope deposits of the MacKenzie Mountains, by Wright and Wilson (1984) in the Jurassic of

Portugal, or by Davies (1977) in the Upper Paleozoic Hare Fiord Formation in the Sverdrup Basin.

Some workers (e.g. Wright and Wilson, 1984; Evans and Kendall, 1977; and several others) have recognized the various structures of the Bouma sequence in carbonate turbidites but this is apparently less typical (Wilson, 1969; Cook and Mullins, 1983; Enos and Moore, 1983). The fewer sedimentary structures in carbonate turbidites have been variously explained as follows:

- the structures were not formed, possibly if currents were of low velocity and low turbulence (Scholle, 1971a). This is unlikely here since the type of mass flow deposits indicate that the slope had too high a gradient.
- the structures were possibly destroyed by bioturbation (Bondel, 1974). However some evidence for these structures should still exist, which is not the case in this study.
- the structures (and textures) could have been modified or destroyed by various diagenetic processes (Scholle, 1971b) such as dissolution of carbonate components carried below the carbonate compensation depth, dolomitization of lime mud matrix, or compaction and pressure solution resulting in reorientation of components, cementation and solution-welding.

The turbidites in the Devon Island Formation are evidently sheet-like and extend considerable distances into basinal settings. Turbidites in sections JL 16 and 17 are 15-20 km basinward of the most shelfward section (JL 01) in this study. They could have been deposited as much as 60 km from the contemporary shelf edge during deposition of the lower member, judging from information in Kerr (1976) and sections examined by the author outside the study area, east of Vendom Fiord. The turbidites evidently were deposited on extensive slopes of low gradient, and considerable depth is implied. They are minor (5-10%) among the predominant bituminous pelagic sediments of the lower member. Davies (1977) and Scholle (1971a) interpreted such a proportion to indicate deposition in lower slope and basin floor settings.

The fossil fauna of the lower member is not abundant but is more diverse than in similar deposits of the Cape Phillips Formation. Although many of the fossils represent pelagic/planktonic organisms that would not have been influenced by bottom conditions, the presence of burrows such as *Phycodes*, *Planolites* and *Chondrites* mostly in the finely crystalline dolostone layers is direct evidence of at least temporary oxygenation of bottom waters. Scholle et al. (1983) and others have suggested that turbidity currents can be important in oxygenation and promote short-lived benthic activity. Although contourites could also have been the reason for higher oxygen levels, there are no indication in the area that they existed, such as cross-bedding.

The crinoidal debris concentrated in basal lags of the bioclastic turbidites suggests derivation from concentrations up-slope, possibly crinoidal banks or mounds. Such accumulations are known to occur over a range of shelf/slope environments but an upper slope source area is envisaged here, probably above storm wave base to account for both oxygenation and periodic generation of turbidity currents. Proximity to the upper boundary of an oxygen-minimum layer might explain the restricted crinoid-ostracod fauna and would help to explain the abundant organic matter evidently caught up in the turbid flow and deposited in the upper layers of the turbidites. Crinoid mounds in upper slope deposits, rather than in more usual shelf margin settings, have been noted, for example, by Packard (1984) in deposits of similar age on Cornwallis Island.

The siliciclastic, fining-upward carbonate sequences, also presumed to be turbidites, are less well known because they appear to be less common (although they are more difficult to recognize in the field) and tend to occur mainly in the upper member, that was examined in less detail in this study. Lenses of peloid-rich sandstone with calcite spar cement in the basal portions of some of these sequences, suggest periodically effective winnowing. Currents such as contour currents could have been responsible, as analogous sorting and winnowing by bottom currents have been recorded along open seaways north of Bahama Bank (Cook and Mullins, 1983).

The presence of two types of turbidites suggests discontinuous distribution of source material across the upper slope. Crinoid accumulations, by nature, could have had a patchy occurrence, with siliciclastic sediments intervening and possibly downslope from them. Siliciclastic sediments in lower slope settings in the Cañon Fiord area (Trettin, 1979) have been interpreted as sediments transported across the Hazen Trough by bottom currents. With occurrence of siliciclastic sediments downslope from a belt of crinoid accumulations in the study area, the siliciclastic turbidites could simply represent turbidity currents generated somewhat downslope from the crinoid mounds. The peloids, too, could have originated slightly below the crinoid mound belt and represent a soft-bodied organism tolerant of somewhat less oxygenated waters (Scholle, 1983b).

4.5.3 Regional considerations

Following shelf progradation represented by the Douro Formation at section JL 01, and to a lesser extent by coarser grained deposits within the Cape Phillips Formation in sections north and northwest of JL 01, the study area witnessed a return to deeper depositional environments for most of the Pridolian, as indicated by deposits of the Devon Island Formation. The abrupt lower contact indicates that the return to deeper environments was quite rapid. Furthermore, the Devon Island Formation seems to be consistent in composition and character, independent of variations shown by underlying deposits. Also of importance, this lower contact is fairly synchronous throughout the Arctic Archipelago: on Grinnell Peninsula, it is earliest Pridolian (Morrow & Kerr, 1977); further south on Devon Island, it is latest Ludlovian (Thorsteinsson & Uyeno, 1980); on northeastern Cornwallis Island, it is latest Ludlovian (Packard, 1984); in southern Ellesmere Island, it is thought to vary from earliest to late Pridolian (McLaren, 1963; Kerr, 1976), although the contact itself is not identified. On the southern coast of Ellesmere Island, this contact does not exist since the statigraphy is similar to that on Cornwallis Island and parts of Devon Island, i.e. representing a shallow shelf with no significant regressions (Packard, 1984). Most workers, though, interpreted these deposits as Cape Phillips Formation in this study area (Kerr, 1976; McGill, 1974; Mayr 1974). In the Cañon Fiord area, equivalent deposits, again referred to the Cape Phillips Formation by Trettin (1979), are clearly earliest Pridolian, and finally in the study area, conodonts and graptolites also suggest an earliest Pridolian age (Uyeno, pers. com., 1982 and 1984). Variations in thickness of the Devon Island Formation partly reflect a highly diachronous upper contact, and this was also seen on northwestern Devon Island by Thorsteinsson & Uyeno (1980).

In the study area, as shown in section JL 16, turbidity current deposition increased importantly upward in the formation, became the predominant process that formed the lower part of the upper member, and then diminished sharply thereafter. This represents, initially, a shallowing-upward, basin-fill sequence in which predominantly basinal facies — the richly bituminous marlstone of the lower member — are followed by prograding lower slope facies — the more grainy lower part of the upper member. Morrow and Kerr (1977) recognized clinothem and undathem depositional settings in the formation on Devon Island. What are termed basinal and slope deposits in this study, however, appear to be equivalent to their clinoform; fully basinal or undathem sediments (i.e. lacking the distal turbidites) do not appear to be exposed in the study area. This basin-fill sequence, largely derived from the carbonate shelf to the south and east, prograded northwestward, progressively filling the Hazen Trough (Morrow and Kerr, 1977).

An analogous Pridolian basin-fill sequence resting abruptly on the Read Bay Group in the Cañon Fiord area, is referred to the Cape Phillips Formation by Trettin (1979), but is presumably the lateral equivalent of the Devon Island Formation (lower member and lower part of upper member). Significantly, this sequence in the Cañon Fiord area is overlain by coarser siliciclastic flysch deposits, referred by Trettin (1979) to the Imina Formation. This flysch, derived from the north or northeast, rapidly filled the Hazen Trough, progressively from northeast to southwest. It evidently was restricted to axial portions of the trough, initially, and only in the earliest Gedinnian was able to spread fully across the collapsed shelf margin and cover the initial basin-fill sequence of the Devon Island Formation. The bituminous, fine, siliciclastic sediments referred to the uppermost Devon Island Formation in the study area are interpreted as distal flysch, and these sediments are apparently not recorded further south on Devon Island.

4.6 Post-depositional Processes

Many post-depositional processes affected these deposits. But they seem to have had very little influence on depositional or redepositional processes. There is more evidence of *burrowing* in the Devon Island Formation than in the underlying Cape Phillips Formation. Most burrows are horizontal and occur in the finely crystalline dolostone or dolomitic lime mudstone beds, probably partly because of slower deposition. Also, these burrowed deposits probably represent higher oxygen levels brought about temporarily by the turbidity currents. This periodic oxygenation would have encouraged benthic activity briefly before the background deposition of hemipelagic and pelagic muds resumed, and oxygen again became depleted. Burrowing was not a major process, as stratification and other primary features are commonly well preserved in the formation.

The timing of *cementation* is uncertain, although some clues are given by other diagenetic processes (see below). In the study area, no evidence was recognized of early cementation during the Pridolian. This contrasts with the Devon Island Formation of southern Ellesmere Island, in which some degree of early cementation is indicated by incorporation of intraclasts of shelf-margin and upper slope deposits in fining-upward mass flow deposits that cut into Devon Island sequences (Packard, 1986, pers. com.). Similar evidence of early cementation on the upper slope was described earlier (see chapter 3) in the transition between the Cape Phillips Formation and the Allen Bay Formation/Read Bay Group undivided. The carbonate turbidites so common in the Devon Island Formation of the study area contain no such intraclasts and are very sparse in southern Ellesmere Island. The differences possibly relate not just to derivation of turbidites potentially from different positions on the paleoslope, but also to earlier cementation in the southern area.

The coarser, mosaic, calcite spar cement in the grain- and peloid-supported lenses of siliciclastic-rich carbonate deposits can be interpreted as either an original fresh water phreatic cement (Mathews, 1971; Land, 1970) or as recrystallization of an earlier cement (Bathurst, 1975).

Neomorphism was important in the Devon Island Formation and transformed most original lime mud to microspar and pseudospar. The latter is especially conspicuous in some marlstone beds in which lenses and calcispheres probably resulted from dissolution of lime mud and subsequent reprecipitation of pseudospar.

Minute *pyrite nodules* occur in minor amounts, particularly in the top portions of the carbonate turbidites, i.e. the finely crystalline dolostone beds, and the marlstone. The association of this eogenic mineral with the more bituminous portions of the deposits reflect its origin, a result of bacterial sulfate reduction of organic matter (Scholle et al., 1983).

Selective silicification was an integral, but minor, part of the diagenesis, affecting mostly the crinoidal grains of lag deposits. In thicker lag deposits more components are silicified near the abrupt bottom contact. Such contacts were presumably the migration paths of silica-rich fluids, just as they were for similar deposits of the Cape Phillips Formation and Allen Bay Formation-Read Bay Group undivided. The occurrence of recrystallized sponge spicules (now calcite spar) and of well preserved sponges, now recrystallized to chalcedony in the finely crystalline dolostone and marlstone, probably indicates a biogenic source for the silica. The sponges are the only fossils silicified in these finer deposits.

Migration of silica-rich fluids along bedding planes was invoked by Von Rad and Rösch (1974) for silicification of pelagic sediments at the bottom of the modern Atlantic Ocean. They also concluded that the presence of organic-rich layers might have favoured the precipitation. This association of organic matter and silicification is related to organic acids that form during anaerobic degradation of the organic matter (Folk & Pittman, 1971; Berger, 1975). Resulting negative Eh and pH, and other factors such as foreign cations (pore water chemistry), *in situ* temperature, partial pressure, permeability of the sediments and especially time, facilitated the precipitation of silica (Von Rad & Rösch, 1974). These crinoidal lag deposits are interbedded with finer hemipelagic sediments probably deposited under anaerobic conditions and containing high amounts of interstitial organic matter. With conditions suitable for silicification, the crinoidal calcite was possibly less stable than the neomorphically transformed lime mud (Bathurst, 1975). Originally lime mud is more susceptible to diagenesis as shown by its ready dolomitization (Murray & Lucia, 1967). Morrow and Kerr (1977) also noted minor silicification to length-slow chalcedony of some bioclasts in biopelmicrite lag deposits of bioclast-rich fining-upward carbonate sequences in the Devon Island Formation on Grinnell Peninsula.

The most important diagenetic phenomenon was *pervasive dolomitization* of original lime mud. This was especially important in the bottom and middle portions of carbonate turbidites, and rarely affected the upper hemipelagic marlstone in which remobilization and recrystallization to calcite pseudospar was more important. Morrow and Kerr (1977) noted upward increase in dolomite content in the Devon Island Formation of Grinnell Peninsula. They proposed that this was produced by downslope (northwestward) movement of a fresh water lens through the upper part of the formation, as this clinoformal basin-fill sequence gradually built up into shallower water. The water would have moved through, and dolomitized more permeable layers, presumeably the coarser basal parts of turbidites that were sealed above and below by impermable hemipelagic muds. With sufficient hydraulic head, the fresh water could conceivably have displaced marine waters for considerable distances basinward (Morrow and Kerr, 1977). Choquette and Steinen (1980) proposed a similar model to explain dolomitization at shallow burial depths in the Mississippian Ste-Genevieve Limestone of the Illinois Basin.

The Devon Island Formation in the study area shows a shallowing-upward trend, as the proportion of coarser end members of the turbidites increases up-section. As these are the parts commonly dolomitized, the proportion of dolomite also increases overall up-section. Although sampling is less representative at higher levels (with much less field exposure), the proportion of dolomite in individual turbidites does not appear to be different at other levels. The intensity of dolomitization apparently did not increase up-section as suggested by the fresh water lens-mixing model of Morrow and Kerr (1977). Also it seems unlikely that a fresh water lens could have been driven laterally the exceptional distances that these thin, sheet-like turbidites extend across the basin.

Determination of the timing is necessary for a better understanding of the processes involved. Unfortunately, without geochemical data, there is no strong evidence of timing. The near-absence of dolomite crystals from the marlstone and from some coarser portions of carbonate turbidites could be the result of permeability control, if dolomitization occurred before cementation and complete destruction of primary porosity with deep burial. The lack of compressed bioclastic grains and peloids in the lower portions of carbonate turbidites, and the good preservation of shells in the middle portions indicate cementation prior to significant compaction. However, some burrow fills were somewhat compressed. The lenses of clean peloidal siltstone and sandstone were loosely packed when cemented by coarse mosaic calcite spar. The dolomite crystals are not associated with stylolites or other pressure-solution structures, which are in any case quite rare in the Devon Island Formation, and so could not be a response of the lime mud grains to stress, as Wanless (1979) suggested. All evidence points toward dolomitization non-related to deep burial.

These dolomitic carbonate turbidites are enclosed by argillaceous lime mudstone or marlstone. A mesogenic, or intermediate diagenetic model for dolomitization relates pervasive dolomitization of lime mud that is enclosed by argillaceous deposits or associated with clay minerals, to compaction. Various depths of burial can be involved but must be at least sufficient for lithostatic pressures to alter the chemical composition of the clay minerals. Magnesium ions so released from interlattice positions of chlorite, vermiculite and especially montmorillonite can then be involved in dolomitization of lime mud. Silica, iron, calcium and sodium ions, also released from the collapsed clays, could later be involved in silicification and cementation (McHargue & Price, 1982). This model of burial dolomitization has been used to explain dolomitization of the Upper Devonian Miette Buildup in Alberta (Mattes & Mountjoy, 1980). However, evidence of intermediate burial, such as stylolitization, pressure-solution seams, reorientation and/or breakage of grains (Mattes & Mountjoy, 1980; McHargue & Price, 1982) were not observed in the Devon Island Formation. Relatively early cementation could explain their absence.

Eogenic dolomitization of deep shelf and basinal deposits has been reported from the top portions of shallowing-upward sequences (Nichols & Silberling, 1980) and mostly related to reflux and mixing water models (Land, 1983). Dunham and Olson (1980) attributed dolomitization of shelf deposits to the mixing waters model but noted pervasive, minor dolomitization of basinal carbonates that could not be explained by lateral extension of a fresh water lens. They suggested that neomorphic transformation of original lime mud (probably mostly high-Mg calcite) to low-Mg micrite, microspar and pseudospar could possibly release sufficient Mg^{2+} ions for minor dolomitization of the lime mud, but this is controversial. This process would also occur at shallow burial depths, concurrently with cementation and neomorphism (Dunham & Olson, 1980; Folk & Land, 1975). Neomorphism was certainly important in the Devon Island Formation. This eogenic model could also explain the absence of evidence for compaction, and support the hypothesis of cementation prior to intermediate burial. Nevertheless, the observed distribution of dolomite crystals was evidently determined somewhat by original permeability. Early dolomitization at basinal water depths could also be related to sulfate reduction as suggested by Scholle et al. (1983). Sulfate reduction was evidently active in the Devon Island Formation, resulting in syngenetic pyrite.

Dolomite rhombs in deep water turbidites might also be interpreted as detrital grains transported from their shallower original environments, such as the shelf-margin (Scholle, 1971b). However, no breakage or abrasion that might be attributed to transport has been observed on dolomite crystals in the Devon Island Formation. Furthermore, dolomitization at shallower water depths by the mixing model (Badiozamani, 1973) should involve early cementation, which should be evident from the presence of intraclasts in mass flow deposits. None were observed. Interestingly enough, Scholle (1971b) reported authigenic overgrowths on transported crystals (up to 50% of the components, but more normally between 10% and 20%) due to release of Mg^{2+} and Fe^{3+} ions from clays during compaction. This process would have been appropriate to the Devon Island Formation, and might obscure surface damage on transported crystals.

In conclusion, lack of geochemical work on the dolomite and sampling that may not be representative because of poor exposure, make it difficult to decipher the processes responsible for partial dolomization of these deposits. However, early cementation at burial depths too shallow for the formation of structures such as stylolites and compressed granular textures, and the distribution of dolomite, at least partly according to primary porosity, could only correspond to dolomitization at shallow or intermediate depths of burial. Eogenic and mesogenic dolomitization models are therefore possible. None of the observations gathered point conclusively to a single model. Morrow and Kerr (1977) opted for the mesogenic model for the Devon Island Formation further south, on Grinnell Peninsula.

4.7 Conclusions

The base of the Devon Island Formation represents abrupt regional deepening. This has been attributed by Morrow and Kerr (1977) and Trettin (1980) to major faulting and collapse of the outer margin of the shelf, possibly associated with the Caledonian Orogeny.

The formation includes an initial, shallowing-upward, basin-fill sequence that prograded northwestward into the Hazen Trough (Morrow and Kerr, 1977). In the study area, basin floor facies are succeeded by lower slope facies, both considered here to be part of the "clinothem" of Morrow and Kerr (1977). The lower member of the formation, predominantly laminated, bituminous marlstone, represents mostly calm basin floor settings in which pelagic and hemipelagic sediments and organic materials accumulated. Fine, distal, carbonate turbidites were introduced intermittently into these calm, anoxic environments, temporarily improving oxygenation and encouraging benthic organisms.

The carbonate turbidites increase in frequency up sequence in the study area, and became the predominant facies, as slope deposits — the lower part of the upper member — prograded northwestward over the basinal deposits. More frequent episodes of oxygenation are represented, but background sedimentation was evidently still below an oxygen-minimum threshold. There is evidence of derivation of the turbidites from two different sources in more oxygenated water up slope: basal lag deposits rich in crinoid debris were derived possibly from discontinuous crinoid accumulations on the upper slope; lags rich in siliciclastic sand were derived from other upper slope deposits, possibly between, or slightly below, the crinoid accumulations.

The uppermost part of the formation appears to represent a return to mainly fine-grained hemipelagic sedimentation in anoxic conditions. It was not recognized on Grinnell Peninsula in the area of Morrow and Kerr's (1977) study, and is poorly understood in this study area because of inadequate exposure and sampling. It possibly is equivalent to the Imina Formation in the Cañon Fiord area, representing the onset of earliest Gedinnian flysch sedimentation (Trettin, 1979). If so, it comprises fine siliciclastics (distal flysch) that prograded from the north or northwest over the mostly carbonate shelf-derived, basin-fill sediments lower in the Devon Island Formation.

Post-depositional processes that affected the Devon Island Formation include: 1) minor burrowing, 2) early neomorphism of lime muds to microspar and pseudospar, 3) early pre-compaction cementation in some grainy deposits, but generally not early enough to provide lithoclasts to slope-derived turbidites, 4) selective silicification, particularly of crinoid material in turbidite lags, and, most importantly 5) dolomitization, largely of lower and middle portions of turbidites, that took place at shallow or intermediate burial depths.

Chapter 5

Conclusions

The study area on Svendsen Peninsula exposes Upper Ordovician to Lower Devonian rocks representing a transition from shelf to slope to basinal depositional environments. Five main phases of sedimentation can be distinguished. These represent changes in water depths, in submarine gradients, and in positions of the shelf margin through this period of time.

Initial sedimentation during the Ashgillian took place on a ramp (term according to Read, 1980; see also James and Mountjoy, 1983) that sloped very gently northwestward into the Hazen Trough. Nodular limestones containing a diverse marine fauna, the so-called "Arctic Ordovician Fauna" (Kerr, 1968), are typical of this first phase. They make up the Irene Bay Formation and lower part of the Allen Bay Formation. They are of similar character throughout the study area except that small bioherms occur in sections toward the west (JL 04, JL 07) inferred to represent somewhat deeper settings. The nodular limestone differ from the ribbon limestones of analogous ramp settings elsewhere (e.g. Read, 1980), presumably due to intense burrowing of the differentially cemented sediments. The first phase of sedimentation culminated near the Ordovician-Silurian boundary with a period of shallowing, and a change by the early Llandoverian to nodular dolostone with a restricted fossil fauna. No clear evidence of subaerial exposure had been identified, although it was possibly obscured by intense dolomitization.

During the second phase of sedimentation, the Hazen Trough widened markedly, presumably through collapse of its southeastern margin along a series of growth faults, such as those interpreted on Greenland (Hurst, 1980; Hurst and Kerr, 1982). Abrupt deepening in the early Llandoverian is marked by a unit of richly bituminous, cherty, finely crystalline dolostone that is widespread in the study area. Gradients increased, and the ramp was rapidly transformed into shelf, slope and basinal environments. These remained distinct from late Llandoverian to early Pridolian time and are represented by the upper Allen Bay Formation-Read Bay Group, undivided.

Slope sedimentation during the late Llandoverian second phase was characteristic of a depositional shelf margin and upper slope (cf. James and Mountjoy, 1983; McIlreath and James, 1984). Ribbon limestones predominate, interrupted intermittently by slope-derived mass flow limestone conglomerate and minor limestone turbidite units that thin and fine basinwards.

During the Wenlockian-Ludlovian third phase of sedimentation, a by-pass margin and upper slope (cf. James and Mountjoy, 1983; McIlreath and James, 1984) developed. The development of reefoid buildups and associated early cementation was accompanied by steeper gradients and the contribution of coarser, reef-derived mass flow deposits, including olistostromes, to the upper slope. Maximum reef development occurred during middle Wenlockian-earliest Ludlovian, and a lesser development in the late Ludlovian.

With disappearance of the large shelf margin reefs in the late Ludlovian, a depositional shelf margin and upper slope again developed for the fourth phase of sedimentation, and continued into early Pridolian time.

Repeated minor fluctuations in gradients, water depths, and the position of the shelf margin, during phases two to four, are reflected by variations in component types, proportions and grain sizes through the slope succession. These are more clearly recorded in sections toward the southeast, representing shallower environments where these changes would have had more pronounced effects. The slope deposits of phases two to four pass toward the north and northwest into uniform lower slope and basinal facies of the Cape Phillips Formation — lime mudstone/marlstone couplets representing continuous starved basin deposition which varied little through the Llandoverian to Pridolian interval.

Sedimentation during the fifth phase took place on a depositional shelf margin and slope, but the characteristic lime mudstone, marlstone and limestone turbidites of the Pridolian-Gedinnian Devon Island Formation indicate much more uniform conditions than previously. This corresponds to the McIlreath and James (1984) model of a depositional margin downslope from a shelf consisting of shallow water lime sand shoals and crinoidal banks, possibly accounting for the amount of dolomitization of these sediments.

The rocks of phases four and five represent deposition on a clinoform analogous to that interpreted by Morrow and Kerr (1977) on Grinnell Peninsula. Deposition on this clinoform progressively filled the Hazen Trough from the south. Shallow shelf deposits of the Read Bay Group (Douro Formation) prograded into the study area (section JL 01) by the Ludlovian. After a brief deepening in the Pridolian, progradation of basin fill deposits of the Devon Island Formation continued through Pridolian and Gedinnian times.

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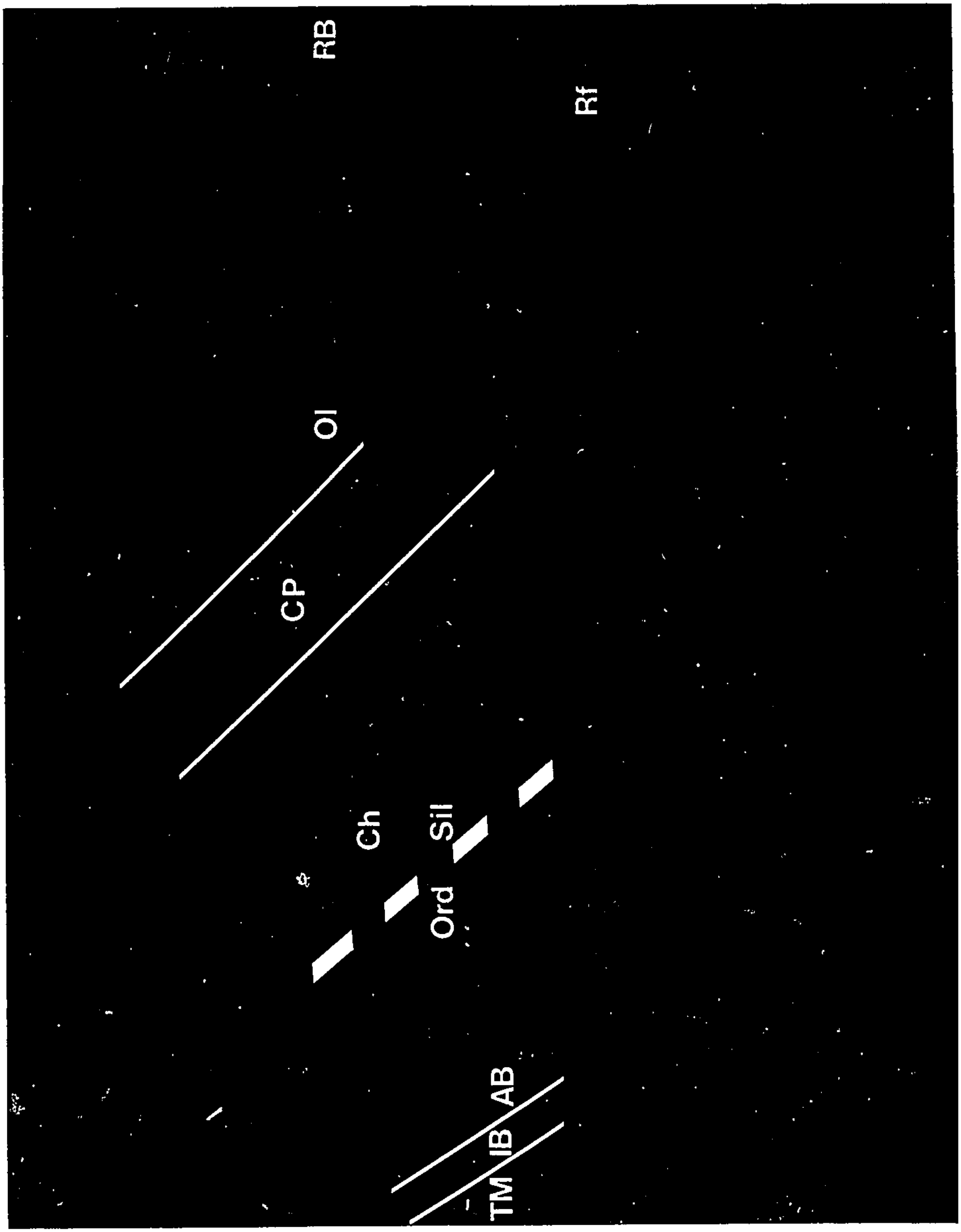
Plates

Plate 1 Air photograph of sections JL 01 and JL 10 (scale: the base of the section is approximately 1200 m wide). The Upper Ordovician Thumb Mountain Formation (TM) is at the left edge of the photograph, followed by the Irene Bay Formation (IB) and the Allen Bay Formation (AB) which constitutes most of the section. It is interrupted briefly by the Cape Phillips Formation (CP) and towards the top of the section (right edge) is overlain by deposits of the Read Bay Formation (RB). Also recognizable in this photograph is the bituminous cherty dolostone unit (Ch) marking the Ordovician(Ord)-Silurian(Sil) boundary. Olistostromes (Ol) are also visible as well as reefal deposits (Rf), the probable source material for the olistostromes.


RB
Rf
OI
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Ord
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IB
TM

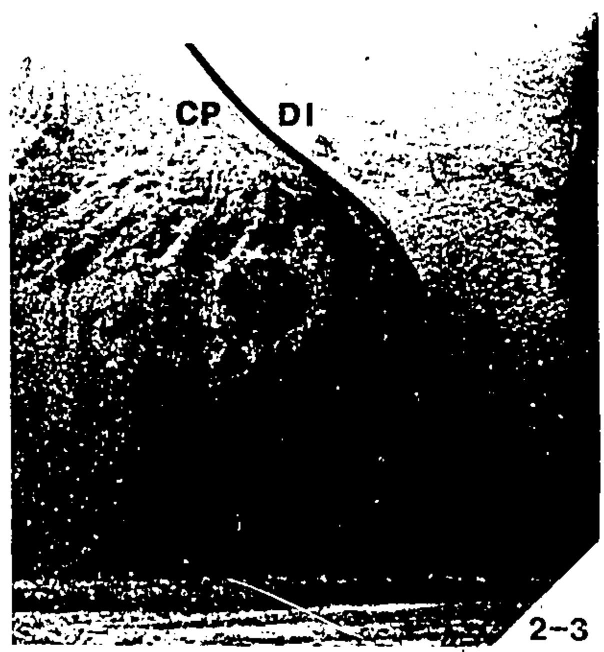
Plate 2-1 Portion of section JL 17, showing the gradual transition between the Allen Bay (AB) and Cape Phillips (CP) formations.

Plate 2-2 Lower portion of section JL 17 showing the Thumb Mountain Formation (TM), Irene Bay Formation (IB) and Allen Bay Formation (AB). Irene Bay Formation approximately 12 m thick.

Plate 2-3 Part of section JL 02 showing the gradational Cape Phillips Formation-Devon Island Formation contact. The base of the section is approximately 15 m wide.

2-1

2-2



2-3

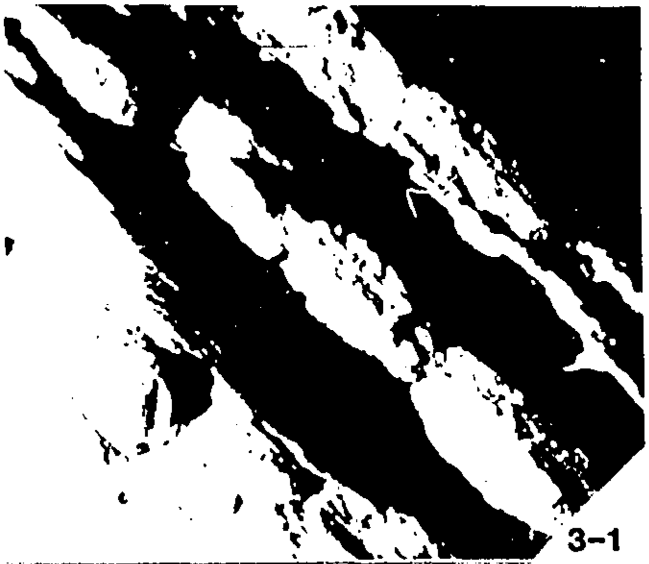
Plate 3–1 Thick-bedded olistostromes in the Allen Bay Formation of section JL 01. The middle flow is approximately 4 m thick.

Plate 3–2 Large reefal block within much finer-grained upper Wenlockian slope deposits (lime mudstone-marlstone couplets) of section JL 01. The block is approximately 7 m in diameter and is composed mostly of light coloured lime mudstone, possibly the main reef core component. The surrounding sediments are lime mudstone and marlstone characteristic of the Cape Phillips Formation.

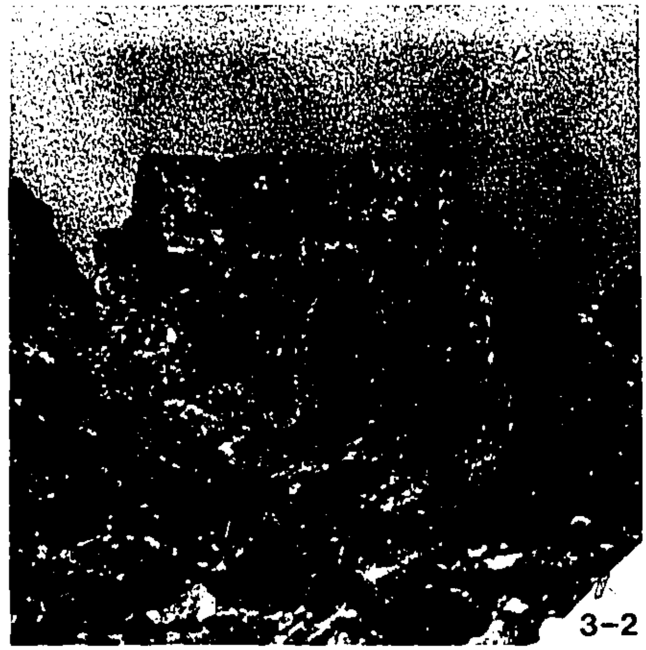
Plate 3–3 Limestone breccia of unit JL 10-78. The intraclasts are composed mostly of lime mudstone in a bioclastic wackestone matrix. Because of the large amount of fossils in the matrix, many of which are corals, the source material is thought to be from the shelf-margin, possibly from debris around bioherms. The base of this photograph is approximately 6 cm in width.

Plate 3–4 Flat-pebble conglomerate in JL 10-16, part of a debris flow. The tabular intraclasts are composed of lime mudstone and are sub-parallel to each other and to the bedding. The matrix, also lime mudstone, is slightly dolomitic.

Plate 3–5 Close-up of another flat-pebble conglomerate in unit JL 10-26. The intraclasts are much more disorganized than in the previous photograph and the matrix is much more bituminous, hence its dark colour. The divisions on the ruler are cms.



3-1



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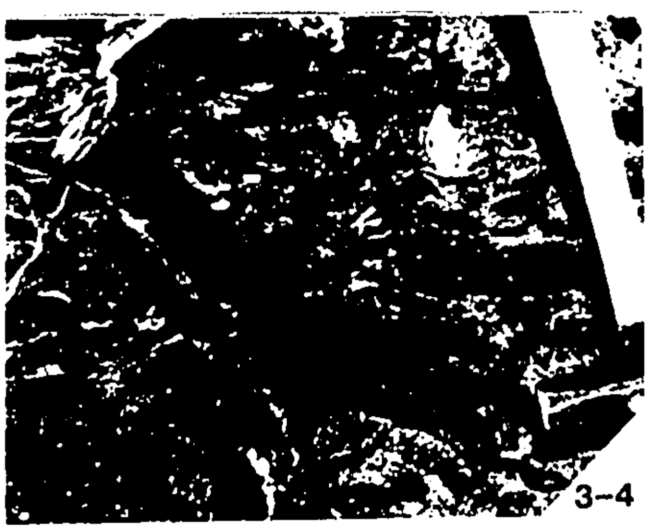


3-4

3-5

Plate 4–1 Limestone turbidite grading from crinoidal bioclastic packstone (PK) to marlstone (ML) in unit JL 10-61. Up-sequence, the rudite size bioclasts quickly diminish in size. The relatively narrow packstone-marlstone transition and the thickness of the packstone bed are probably representative of a proximal flow compared to the deposits shown below.

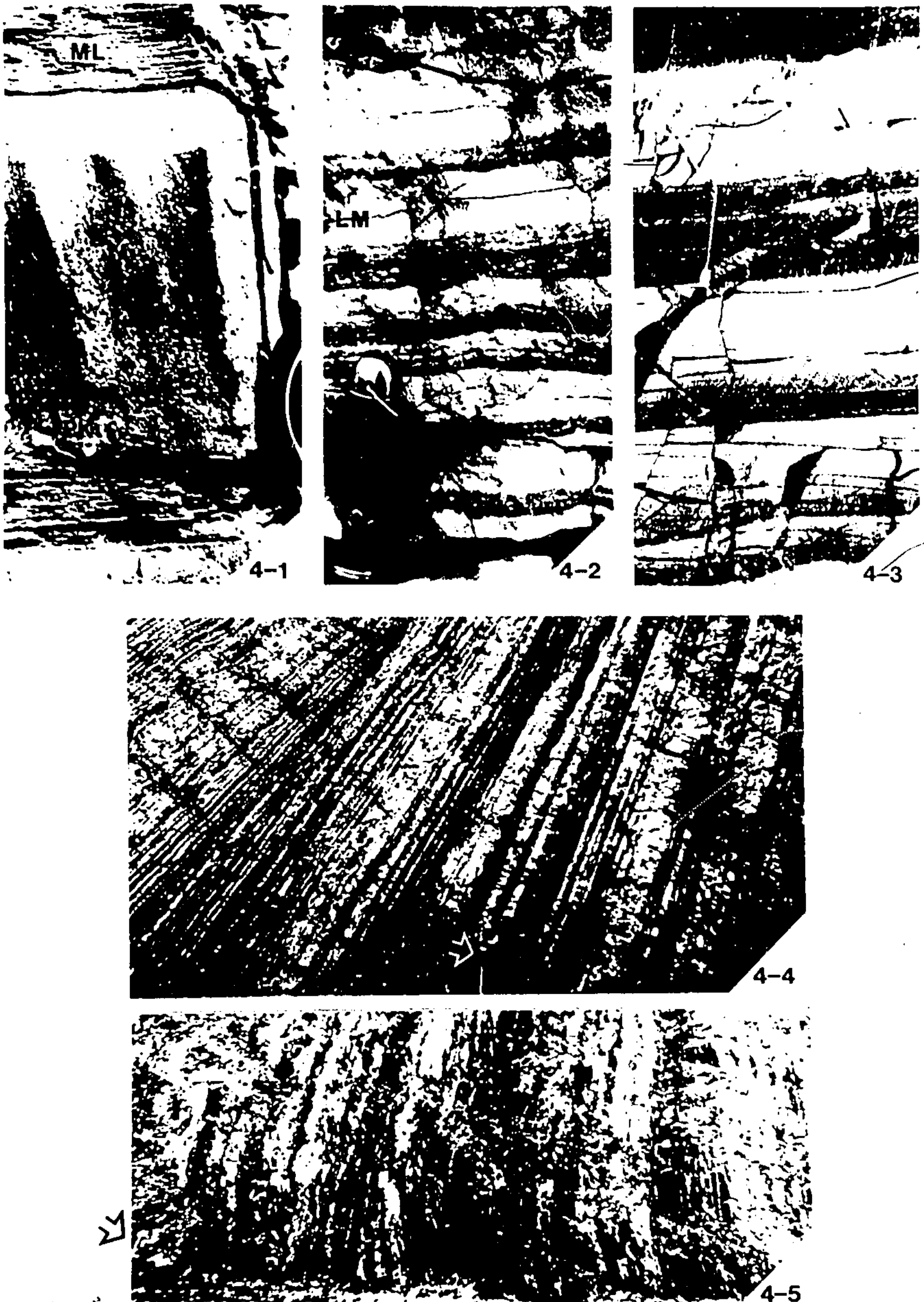
Plate 4–2 Limestone turbidites in section JL 04-31 beginning with crinoidal bioclastic packstone (PK) and terminating as massive lime mudstone (LM). This turbidites are much thinner than the one in the previous photograph, and they succeed each other more frequently, typical of a more distal deposit.

Plate 4–3 Limestone turbidites in unit JL 16-34, apparently entirely lime mudstone but showing a gradation in components and grain size. Basal fine to very fine-grained dolomitic crinoidal packstone grades up into very finely crystalline calcareous dolostone. The basal limestone is darker as it contains slightly more bitumen in larger pore spaces, and the dolostone is characteristically lighter due to weathering. This unit occurs in the sections representing the most distal depostional environment. The base of the photograph is approximately 40 cm in width.

Plate 4–4 A thick sequence in section JL 04 of thin, fine limestone turbidites, each comprising a couplet of apparently dark-coloured and light-coloured lime mudstone similar to those in Plate 4-3. Field assistant with 1,5 m pogo stick at bottom centre.

Plate 4–5 Top part of section JL 16 showing the most distal limestone turbidites, lime mudstone-marlstone couplets. 1,5 m pogo stick at lower left corner.

• Plates •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LM
4–1
4–2
4–3
4–4
4–5

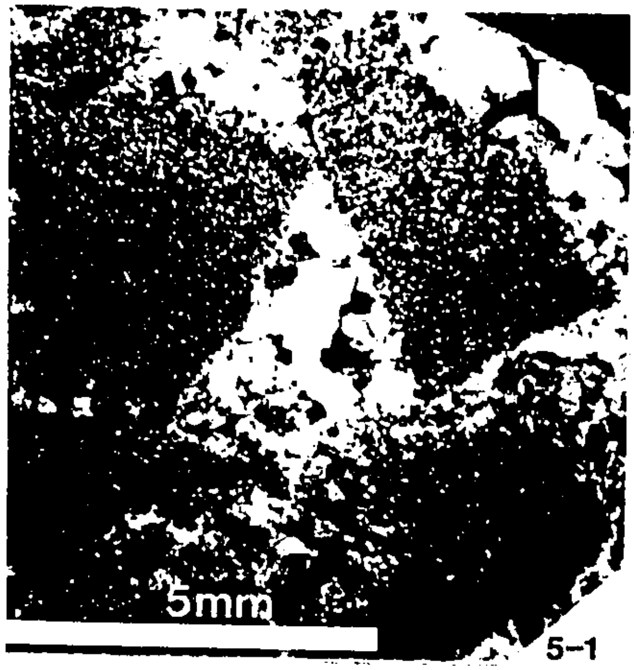
Plate 5–1 Dolomitized breccia of unit JL 03-10. Two distinct periods of dolomitization are suggested by the very fine anhedral dolomite crystals and abrupt contacts of the darker, more bituminous intraclasts, compared to the medium crystalline euhedral dolomite in the less bituminous enclosing cement. Early dolomitization of source deposits, followed by mass flow, and then precipitation of the dolomite cement, are inferred.

Plate 5–2. Partly dolomitized disorganized flat-pebble conglomerate in unit JL 02-11. The very fine dolomite crystals (D) occur only in the bituminous matrix. The crystals are much more anhedral than dolomite crystals normally found in partly dolomitized mass flow deposits, and many have rounded edges suggesting some transport. The matrix is thought to be derived from the partly dolomitized but uncemented portion of ribbon limestone; the intraclasts are fragmented ribbons of this source rock, made impermeable to dolomitization by early cementation.

Plate 5–3 Condensed polymictic breccia from unit JL 10-43. The presence of some oosparitic intraclasts (OO) identifies the shelf-margin as the source.

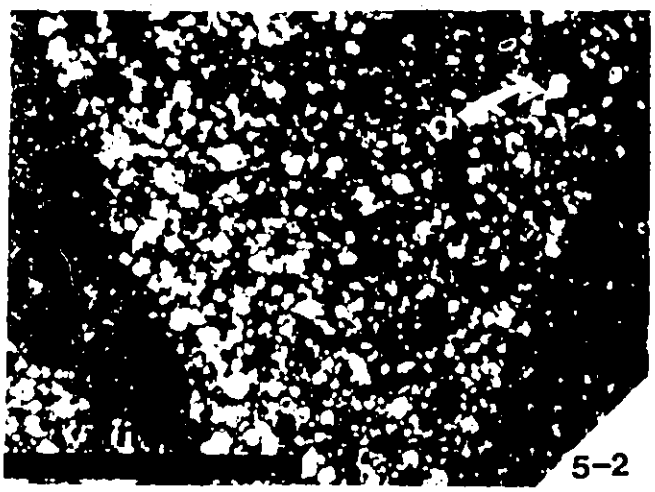
Plate 5–4 Example of polymictic limestone conglomerate of unit JL 02-12. Presence of pelsparite indicates that the mass flow originated near the shelf-margin. The fabric is somewhat condensed as demonstrated by stylolitic margins on intraclasts.

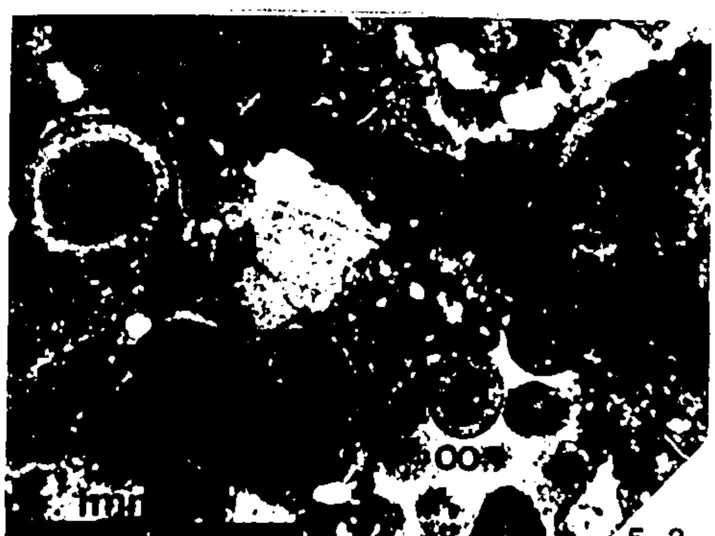
Plate 5–5 Fine limestone turbidite in unit JL 02-37. Pelsparite grainstone begins just above a sharp, scoured contact, grades upward into pelmicritic packstone and wackestone. Peloids disappear upwards, and the rock becomes a bituminous lime mudstone (shown in the lower portion of the photograph at the top of an underlying turbidite).



5-1

5-5


5-2


5-3

5-4

Plate 6-1 Argillaceous nodular limestone Type II (nodular layers) in the Ludlovian Read Bay Group of section JL 01. The dark coloured areas represent the early cemented fraction of the deposit; the light coloured areas are the uncemented fraction which was more susceptible to dolomitization, and now appears lighter in colour due to weathering.

Plate 6-2 Non-argillaceous nodular limestone of unit JL 14-07. Although both parts of this limestone were probably originally of the same composition, i.e. lime mudstone, lighter portions are attributed to dolomitization prior to cementation, as explained for Plate 6-1.

Plate 6-3 Non-argillaceous nodular limestone from unit JL 10-73 (polished surface parallel to bedding). Nodular structure is due to biogenic activity as demonstrated by the spreites within the burrow fills. The light coloured burrow fills have been partially dolomitized. Burrow fills average about 1,5 cm in width.

Plate 6-4 Nodular dolostone of unit JL 01-44. The rock consists of a very bituminous, very finely crystalline dolomitic portion which is dark colour here, and a light coloured, non-bituminous, finely crystalline dolomitic portion. This type of rock appears to be the result of complete dolomitization of non-argillaceous nodular limestone. The photograph represents a 4 cm wide polished section.

Plate 6-5 Nodular dolostone of unit JL 17-17. Although the components are the same as in the previous photograph, because of the irregular form of the bituminous dolomitic portion, it is much less evident that the non-argillaceous nodular limestone is the original rock.

Plate 6-6 Ribbon limestone in unit JL 10-16. The early cemented portion of this type has a darker weathering colour; the later-cemented portion appears lighter because of dolomitization (see Plate 6-1 caption). Commonly interbedded with flat pebble conglomerate, ribbon limestone is also the source of the intraclasts and dolomitic matrix of the conglomerate. (base of photograph = 22 cm)

6-1

6-6

6-2

6-5

6-3

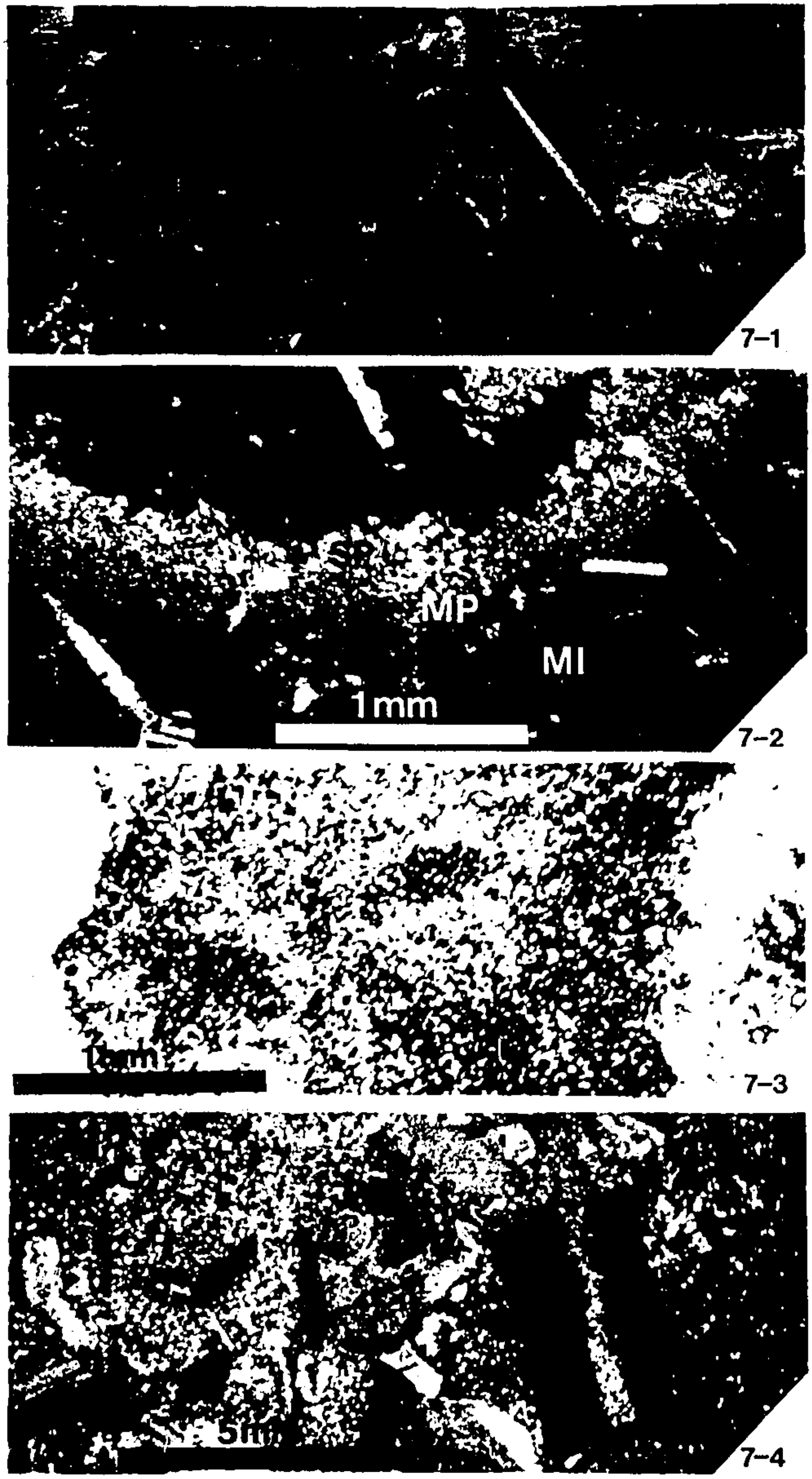
6-4

Plate 7–1 Large calcite concretions within massive lime mudstone of unit JL 17-30. Concretion growth deformed adjacent bedding.

Plate 7–2 Bioturbated "nodular" portion of a non-argillaceous nodular limestone of unit JL 17-09. The burrow fill (darker) is dolomitic, very bituminous lime mudstone, while the host sediment is a spicular micritic lime mudstone. The contact between the two is delineated by a neomorphic rim within the host sediment, grading from pseudospar (SP) closest to the contact, through microspar (MP) into regular dense micrite (MI). The rock is generally very similar to nodular dolostone, with irregularly shaped "nodular" and "matrix" fractions, such as in Plates 6–4 and 6–5, supporting the theory that the nodular dolostone simply represents formerly bioturbated limestone.

Plate 7–3 Burrow fill (darker), in nodular dolostone in unit JL 01a-75. The burrow fill is calcareous (white grains within—arrow) and bituminous finely to medium crystalline dolostone, and the matrix is dense finely crystalline dolostone. The tubular shape and vertical orientation of the bituminous dolostone portion indicate a dolomitized burrow fill, and this is substantiated by spreites in many of these structures. This deposit is part of reef debris interbedded with bioherms of medium crystalline dolomite, and was presumably partly dolomitized where in direct contact with biohermal tongues.

Plate 7–4 Biosparitic packstone portion of a coarse limestone turbidite in unit JL 01-36. Most of the matrix has been dolomitized, probably after deposition, since the dolomite crystals are euhedral and no lime mudstone is left or was mixed in.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7-4

Plate 8–1 Cherty, very bituminous finely crystalline dolostone in the Ordovician-Silurian boundary marker unit (JL 10-06). A chert nodule (CH) is enclosed by layers of pure calcite sparite (SP). Colour bands in the nodule represent different replacement levels of the dolostone. The central medium grey portion has been almost completely silicified; the lighter coloured zones represent transitional zones between the chert and much darker dolomitic host rock (DS). An incompletely formed chert nodule just above shows these different zones better.

Plate 8–2 Contact (arrow) between a chert nodule (CH) and the host finely crystalline dolostone (DS) in unit JL 10-73. The chert nodule probably grew by replacing the dolomite crystals, some of which remain in the nodule (white crystals).

Plate 8–3 Pervasively silicified intra-pelsparitic grainstone of unit JL 02-37. The extremely abrupt contact (arrow) between silicified (below) and non-silicified portions is sub-parallel to bedding. The original structures of the completely silicified portion are very well preserved, due to differences in the silica: sparite is represented by coarser and lighter-coloured chert; small intraclasts and peloids, by finer and darker-coloured chert, and these grains are further defined by a coat of bitumen. No silicification occurred above this contact.

Plate 8–4 Selective silicification in unit JL 01-39. A non-silicified *Girvanella* encrusts a completely silicified rugose coral. Intragranular cement within the coral was also untouched by silicification.

Plate 6–5 Selective silicification in a limestone turbidite (unit JL 16-34), affecting only the micritic lime mudstone matrix. In the bottom portion of this limestone turbidite, peloids and spar cement (SP) were not affected by silicification, while aggregates of lime mudstone matrix were entirely silicified (CH). This selective silicification is more important in the top portions of these limestone turbidites, in which lime mudstone has been replaced entirely by chert, as shown here by the top of a turbidite underlying the sharp contact (arrow).

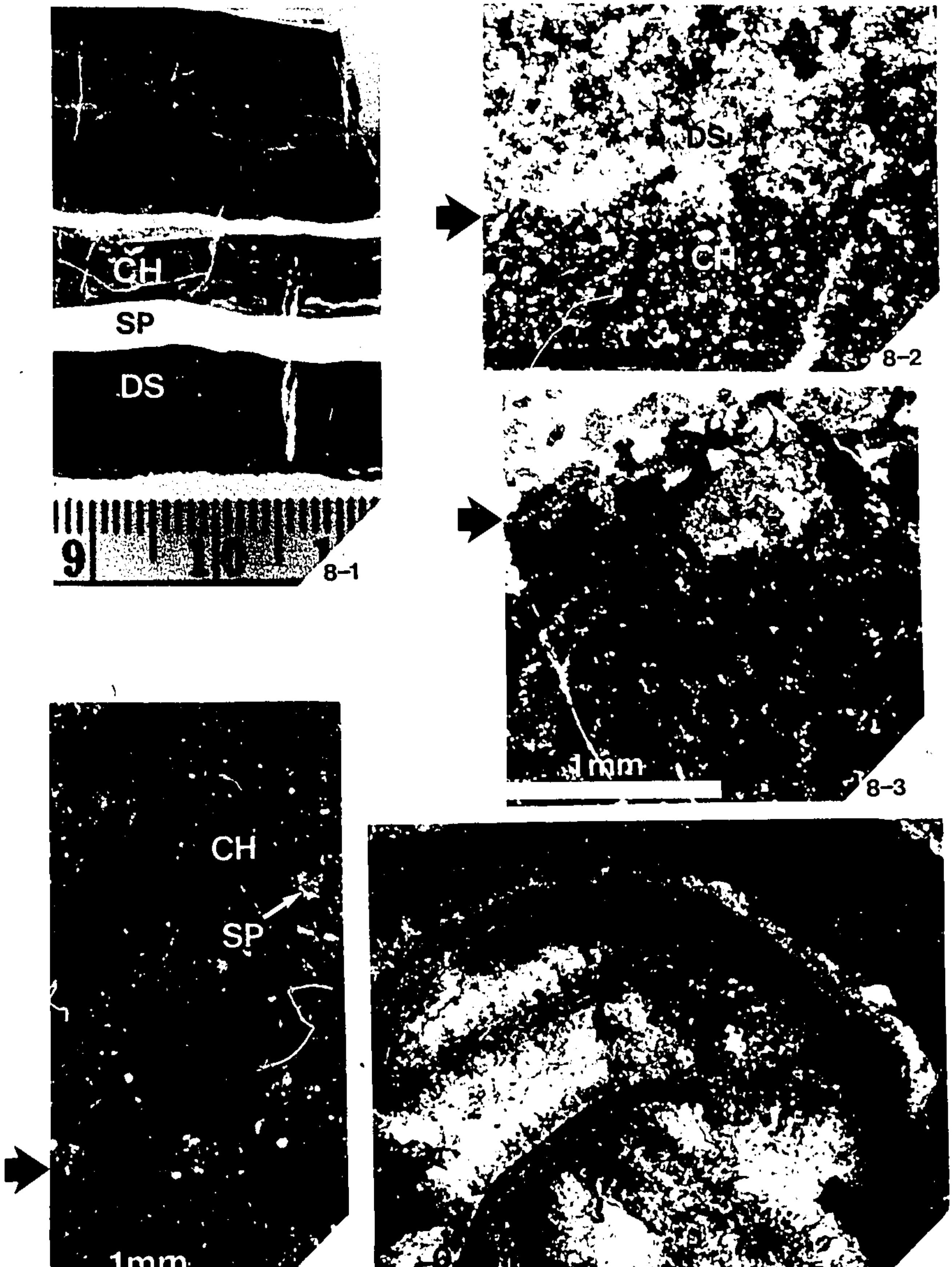

CH
SP
DS
8-1
DS
CH
8-2
1mm
8-3
CH
SP
1mm
8-5
8-4

Plate 9–1 Trace fossils (*Planolites* and *Rusophycus*) on a bedding plane of unit JL 16-01 (Irene Bay Fm). Such bioturbation was intense within the deposits of the Irene Bay Fm and the lower portion of the Allen Bay Fm.

Plate 9–2 Walking traces on bedding plane (JL 01a-63), probably created by a large trilobite (?) of which large fragments were found associated. The photograph represents an area 5 cm long.

Plate 9–3 High concentration of burrows (*Thalassinoides*) on a bedding plane of unit JL 10-51. Such evidence of bioturbation is common in the top bed of fining-upward sequences or in beds interlayered with fining-upward sequences.

Plate 9–4 Two generations of trace fossils in a thin section of unit JL 07-13. The first generation is a series of large burrows (BU^1); part of the width of one burrow covers most of this photograph. The burrow fill is calcareous very finely crystalline dolostone in gradational contact with darker micritic host rock (MI). The second generation is an incomplete *Paleophycus* burrow fill (BU^2) which follows commonly the central portion of previous burrow fills. The more abrupt contact delimiting these indicate that the first generation burrow fills were cemented and dolomitized early. The second generation burrows are partly filled with a dolomitic spar cement.

Plate 9–5 A trilobite fragment showing boring, in the bottom bituminous and bioclastic mass flow deposit of a fining-upward sequence in unit JL 01a-78. The borings (arrows) are filled with micrite.

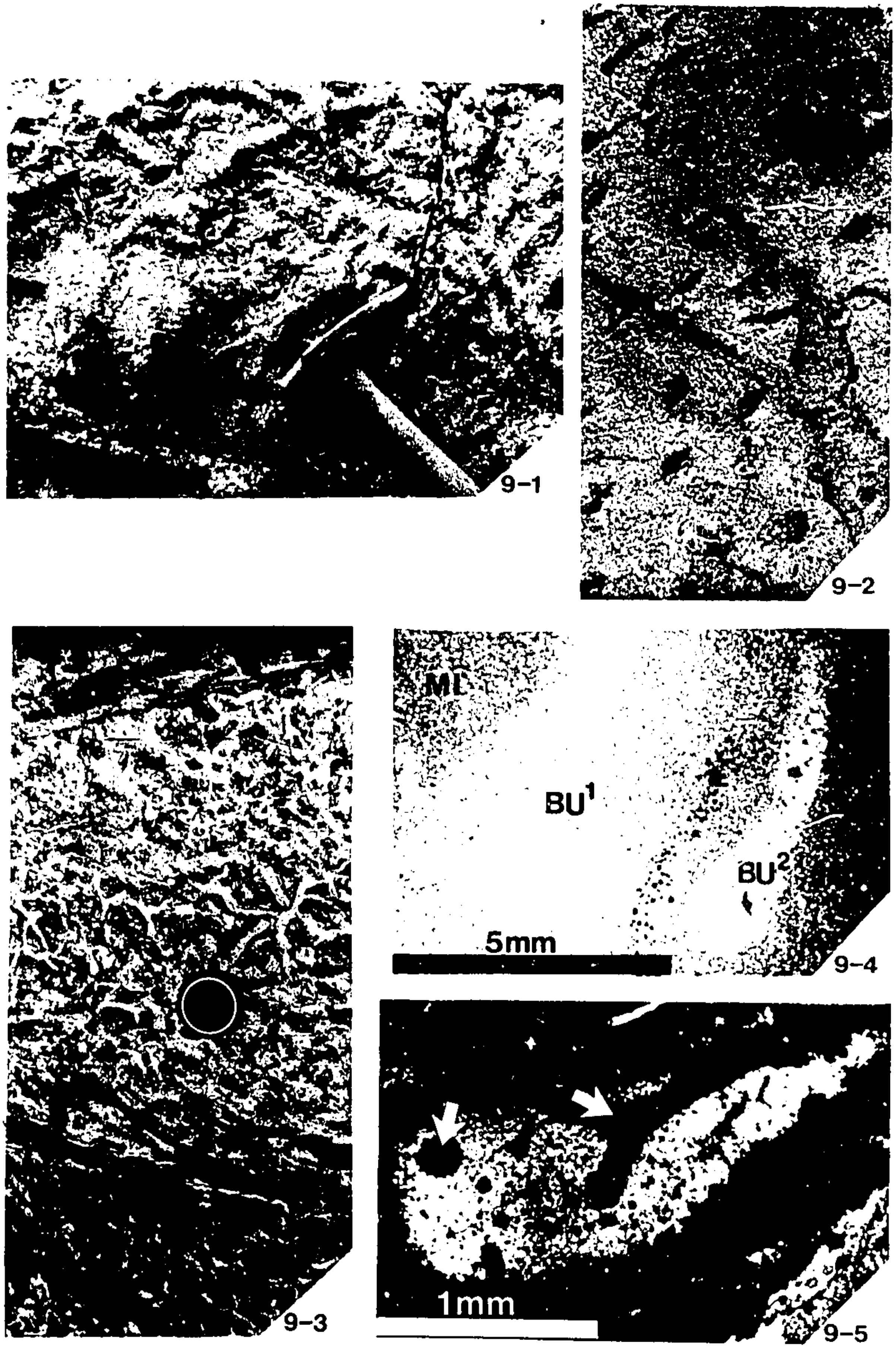
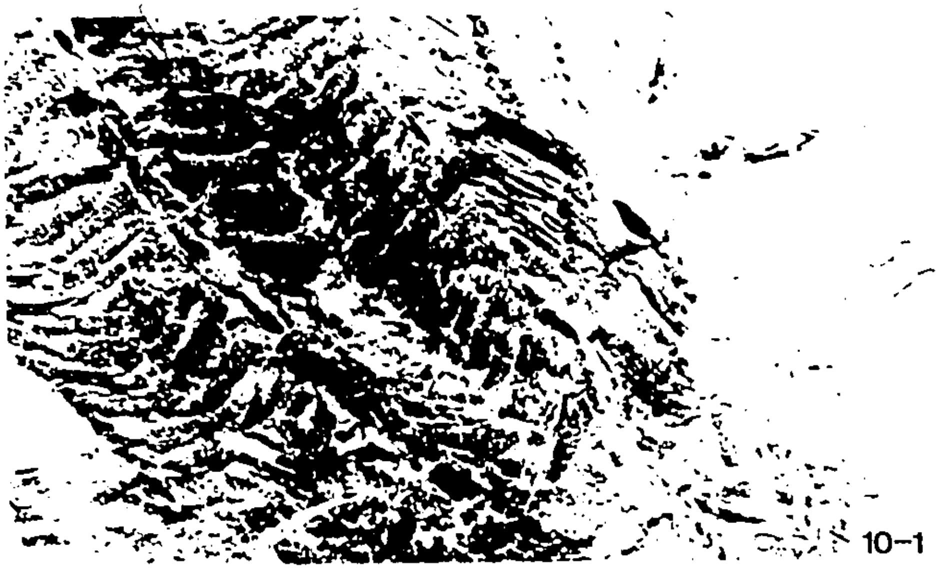

9-1
9-2
9-3
ML
BU1
BU2
5mm
9-4
1mm
9-5

Plate 10–1 A very rarely occuring ostracoderm (only 2 specimens found) possessing a multitude of small plates and lateral fins. This specimen was found in unit JL 01-44. The photograph represents an area 6 cm long.

Plate 10-2 Inferior plate of a more commonly found ostracoderm (from unit JL 02-24). The specimen is 3,2 cm in length.

Plate 10–3 *Monograptus spiralis,* a very common graptolite of the Cape Phillips Fm. Specimen from unit JL 17-30, enlarged 2×.



10–1

10–2

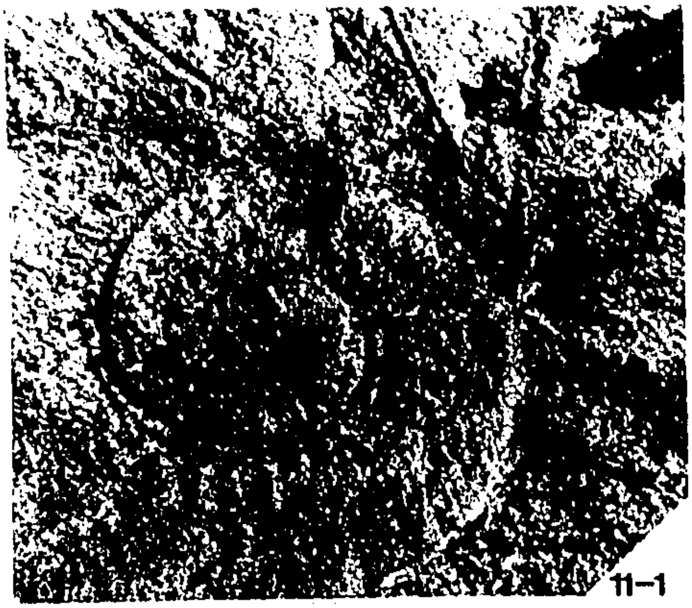
10–3

Plate 11–1 *Cyrtograptus sakmaricus*, an important graptolite, indicating Cape Phillips Fm beds of top Llandoverian age. Magnified 2×, this specimen was found in unit JL 04-28.

Plate 11–2 *Monograptus testis*, occurs mostly towards the top of the Cape Phillips Fm, close to the Devon Island Fm contact. These specimens from unit JL 04-30 are magnified 5×.

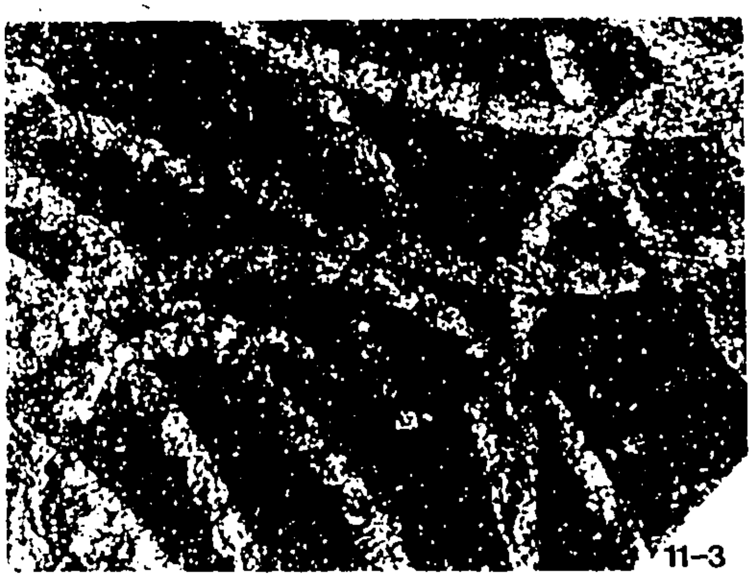
Plate 11–3 *Monograptus bohemicus*, one of the most important index fossils of Pridolian age deposits. It appears in the basal Devon Island Fm, helping to define the Cape Phillips Fm-Devon Island Fm contact. Here magnified 3×, these specimens are from unit JL 17-31.

Plate 11–4 *Monograptus priodon*, here magnified 2×, is common in beds of Late Silurian age such as in unit JL 04-27.



11-1

11-2



11-3

11-4