GEOLOGICAL SURVEY OF CANADA · 1842 ·

GEOLOGICAL
SURVEY
OF
CANADA

DEPARTMENT OF ENERGY,
MINES AND RESOURCES

BULLETIN 234

EVOLUTION OF A MIDDLE AND UPPER DEVONIAN SEQUENCE FROM A CLASTIC COASTAL PLAIN - DELTAIC COMPLEX INTO OVERLYING CARBONATE REEF COMPLEXES AND BANKS, STURGEON - MITSUE AREA, ALBERTA

L. F. Jansa and N. R. Fischbuch

Ottawa
Canada
1974

Price $5.00

EVOLUTION OF A MIDDLE AND UPPER DEVONIAN SEQUENCE FROM A CLASTIC COASTAL PLAIN - DELTAIC COMPLEX INTO OVERLYING CARBONATE REEF COMPLEXES AND BANKS, STURGEON - MITSUE AREA, ALBERTA

1,400–1973–6358–7

Critical Readers
A. W. NORRIS
R. W. MACQUEEN

Editor
MARGUERITE RAFUSE

Text printed on Georgian offset smooth (brilliant white)
Set in Times Roman with
20th Century captions

Artwork by CARTOGRAPHIC UNIT GSC

GEOLOGICAL SURVEY
OF CANADA

BULLETIN 234

EVOLUTION OF A MIDDLE AND UPPER DEVONIAN SEQUENCE FROM A CLASTIC COASTAL PLAIN - DELTAIC COMPLEX INTO OVERLYING CARBONATE REEF COMPLEXES AND BANKS, STURGEON - MITSUE AREA, ALBERTA

By

L. F. Jansa[1] and N. R. Fischbuch[2]

[1] Atlantic Geoscience Centre,
Geological Survey of Canada.

[2] Fischbuch Consultants Ltd.,
Calgary, Alberta.

DEPARTMENT OF
ENERGY, MINES AND RESOURCES
CANADA

Available by mail from Information Canada, Ottawa,
from Geological Survey of Canada, 601 Booth St., Ottawa,
and at the following Information Canada bookshops:

HALIFAX
1683 Barrington Street

MONTREAL
640 St. Catherine Street West

OTTAWA
171 Slater Street

TORONTO
221 Yonge Street

WINNIPEG
393 Portage Avenue

VANCOUVER
800 Granville Street

or through your bookseller

A deposit copy of this publication is also available
for reference in public libraries across Canada

Price: $5.00 Catalogue No. M42-234

Price subject to change without notice

Information Canada
Ottawa, 1974

Southam Murray
02KX-23387-6455

PREFACE

The depositional history of the rocks that form the present-day reservoirs of oil and gas is a matter of considerable interest and concern to petroleum geologists. The interpretation of the paleogeography with its involved relationships of reefs, "off reefs," and other features of the Devonian seas has provided information of great value in the development of the Devonian oil and gas fields of western Canada.

The present study describes the sequence of pre-reef growth terrigenous and evaporite sedimentation, and traces its development through later stages of carbonate reef-bank complexes. This case-history approach to a specific area was made possible by the extensive drilling program that began in the Swan Hills area in 1957 and continued in the Mitsue area, and which provided a considerable footage of cores for detailed study. By means of control studies such as this, especially in a commercially productive area, it is possible to determine relationships and test concepts that can be used by analogy elsewhere in correlating similar events and processes related to the provenance and exploitation of oil and gas.

S. C. ROBINSON
Acting Director, Geological Survey of Canada

OTTAWA, April 30, 1973

CONTENTS

Illustrations

EVOLUTION OF A MIDDLE AND UPPER DEVONIAN SEQUENCE FROM A CLASTIC COASTAL PLAIN—DELTAIC COMPLEX INTO OVERLYING CARBONATE REEF COMPLEXES AND BANKS, STURGEON–MITSUE AREA, ALBERTA

Abstract

Middle Devonian rocks of the Sturgeon–Mitsue area of west-central Alberta consist of a sequence of strata ranging from a lowermost evaporitic unit (Muskeg Formation) upward through a clastic delta complex (Watt Mountain Formation, Gilwood Member), upon which lie shales and evaporites (Fort Vermilion Formation). Over these shales and evaporites were deposited in succession: a basal carbonate blanket, a marginal reef-bank complex, and finally areally restricted solitary reef complexes (Swan Hills Formation). Sedimentation and paleotopographic configuration of the Gilwood delta complex appear to have had considerable effect on both the location and development of subsequent Swan Hills reef-banks.

During late Muskeg time evaporites and carbonates were deposited on broad tidal flats over much of Alberta, producing a surface of relatively low relief. Epierogenic uplift of the Peace River Arch, which lies immediately adjacent and to the northwest of the study area, resulted in the deposition of broad clastic halos around the Arch. The most extensive of these clastic halos (Gilwood Member) developed as a delta along the coastal plain. Beyond the delta open-marine shales (Watt Mountain Formation) were deposited. Following Watt Mountain time a period of regression and stagnation resulted in the deposition of evaporites (Fort Vermilion Formation) in the shallow lagoons and embayments behind the outer barriers of the Gilwood delta complex. Differential compaction of the Watt Mountain and Fort Vermilion sediments appears to mirror and somewhat subdue the original delta topography. Carbonate sedimentation during the Swan Hills transgression that followed appears to have been influenced by this topography as well as by such factors as wind and current direction, water salinity, water temperature, and tidal action. This interaction of organic growth with physical and hydrodynamic processes resulted in first, the development of a basal carbonate blanket composed predominantly of pellets, oölites, algal material, skeletal debris, and micrite; and secondly, the development of a marginal reef-bank complex consisting of a narrow discontinuous outer biohermal barrier built mainly of tabular and subspherical stromatoporoids, with an inner shelf lagoon developed landward. Vertical depositional cyclicity as well as lateral facies variation is evident in the marginal reef-bank complexes. Thirdly, areally restricted solitary reef complexes developed on the platform formed by the marginal reef-bank complexes and are skewed in a southwesterly direction.

Résumé

Les roches du Dévonien moyen de la région de Sturgeon–Mitsue (centre–ouest de l'Alberta) se composent d'une série de strates qui s'échelonnent d'une unité d'évaporites (formation de Muskeg) à la base jusqu'à un complexe deltaïque clastique (formation de Watt Mountain, niveau de Gilwood) que recouvrent des schistes argileux et des évaporites (formation de Fort Vermilion). Sur les schistes argileux et les évaporites se sont successivement déposés une couche de base de carbonates, un complexe de banc corallien bordier et enfin des complexes de bancs isolés et limités (formation de Swan Hills). La sédimentation et la configuration paléotopographique du complexe deltaïque de Gilwood semblent avoir eu un effet considérable sur l'emplacement et la formation des bancs coralliens subséquents de Swan Hills.

Vers la fin de la période Muskeg, les évaporites et les carbonates se sont déposés sur de larges veys sur presque toute l'Alberta, créant ainsi une surface au relief relativement bas. Le soulèvement épirogénique de l'arc de la Rivière de la Paix qui se trouve tout près et au nord-ouest de la région étudiée a été formé par le dépôt de larges halos clastiques autour de l'arc. Le plus étendu de ces halos clastiques (niveau de Gilwood) s'est transformé en delta le long de la plaine côtière. Au-delà du delta se sont déposées des argiles marines (formation de Watt Mountain). La période de stagnation et de régression qui a suivi la période de Watt Mountain a entraîné le dépôt d'évaporites (formation de Fort Vermilion) dans les lagunes et les baies peu profondes, au-delà des limites extérieures du complexe deltaïque de Gilwood. Les différentes compacités des sédiments de Watt Mountain et de Fort Vermilion semblent refléter et quelque peu déterminer la topographie du delta d'origine. La sédimentation de carbonates pendant la transgression de Swan Hills, qui a suivi, semble avoir été influencée par cette topographie et par des facteurs comme la direction du vent et du courant, la salinité de l'eau et sa température et la marée. L'interaction de la croissance organique et des phénomènes physiques et hydrodynamiques a premièrement amené la création d'une couverture de fond de carbonates principalement composée de graviers, d'oolithes, d'algues, de débris squelettiques et de micrites; elle a aussi amené la formation d'un complexe de banc corallien bordier qui consiste en une étroite barrière extérieure discontinue de récifs vrais principalement constituée de stromatoporoïdes tabulaires et subsphériques et qui entoure une lagune bordière intérieure qui évolue en direction des terres. On peut voir clairement dans le complexe de banc corallien le cycle des dépôts verticaux ainsi que la variation du faciès latéral. On note troisièmement la formation de complexes de bancs isolés et limités sur la plate-forme créée par le complexe de banc corallien bordier et qui sont orientés en direction sud-ouest.

INTRODUCTION

In general, the aims of this investigation were twofold: first, to systematically describe Middle Devonian pre-reef growth, terrigenous and evaporite sedimentation in a specific area of Alberta; and secondly, to examine the incipient stages of subsequent overlying carbonate reef-bank complexes (for explanation of the term reef-bank complex *see* p. 46). Readers who do not wish to examine all the details of the petrographic description can read the brief summary at the end of each section of the paper, and then proceed directly to the evolution of delta and surrounding environments, and follow it by reading the depositional history of the marginal reef-bank complexes.

The project was initiated during tenure of a National Research Council Fellowship (1969–70) awarded to the senior author and held at the Institute of Sedimentary and Petroleum Geology, Geological Survey of Canada, Calgary.

The Sturgeon–Mitsue area of west-central Alberta (Fig. 1) was selected for the study because data are readily available and sedimentary structures are relatively well preserved. The lowermost unit studied is the Middle Devonian upper Muskeg Formation, an evaporitic sequence lying behind a reef barrier far to the north (Fig. 2). Overlying the upper Muskeg evaporites is a clastic succession known as the Watt Mountain Formation (Fig. 4), which contains the Gilwood Sandstone Member, a delta complex fringing the Peace River Arch; above the Watt Mountain Formation lie the laminated anhydrites and dolomites of the Fort Vermilion Formation. Finally, the overlying reef complexes of the Swan Hills Formation are examined, with reference to the configuration and sedimentary processes involved in the underlying deposits. From this sequential study it is possible to determine that the initiation of organic buildups was related to the pre-existing topography, and to establish the role of individual processes controlling the evolution of the Lower Swan Hills marginal reef-bank complex, and the localization of the subsequent Upper Swan Hills isolated organic structures.

It was necessary initially to evaluate the data available, since many wells in the general Swan Hills area have cored sections only to the base, or near the base, of the Swan Hills Formation. In the Mitsue area many cores are available from the base of the Swan Hills Formation to well below the Gilwood Member. The Gilwood delta complex was examined in detail, since it contains a variety of sedimentary structures formed during several cycles of deposition, and appears to have had considerable influence on subsequent Swan Hills carbonate reef-bank sedimentation.

Previous Work

Intensive drilling activity began in the Swan Hills and Mitsue areas in 1957 and 1964 respectively, and has continued since that time. This exploration and development

Original MS. submitted: 26 September, 1972
Final version approved for publication: 4 May, 1973

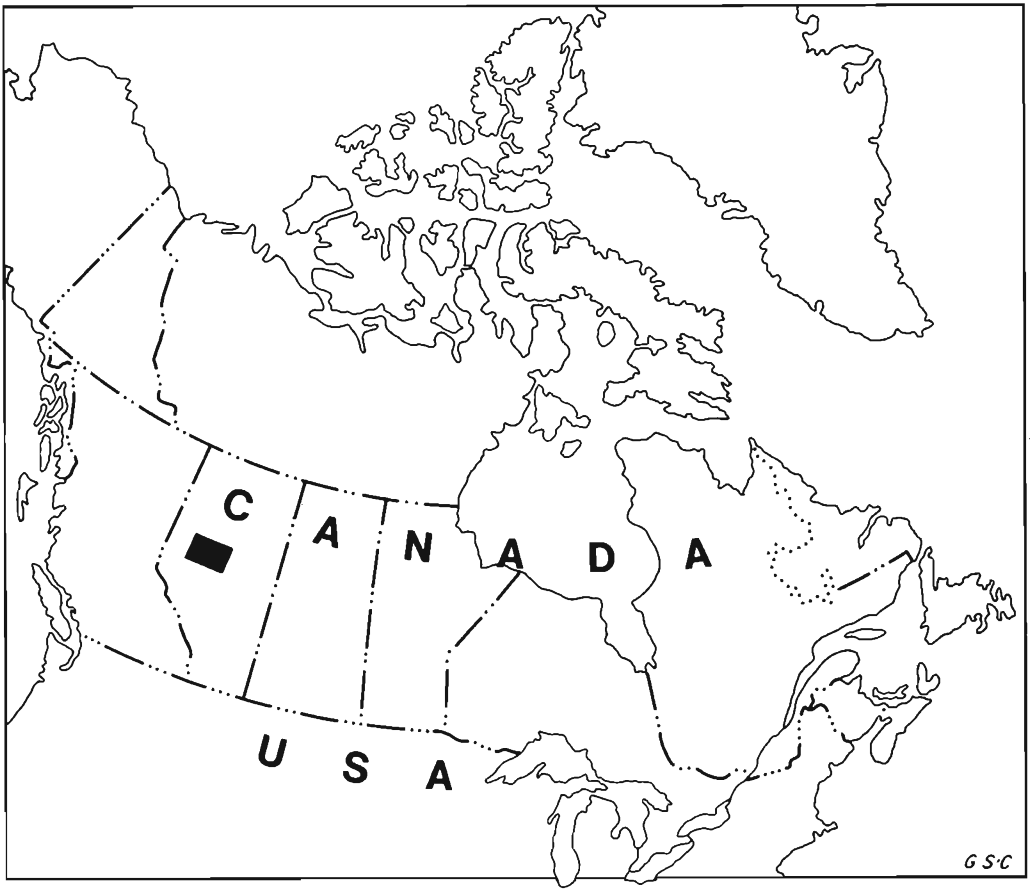


FIGURE 1. Location map of study area.

have produced many hundreds of feet of core that have received considerable attention from stratigraphers, paleontologists, and sedimentologists. Fong (1959) initially proposed the name for the oil-producing carbonate as the Swan Hills Member of the Beaverhill Lake Formation. These units were later raised in status to Swan Hills Formation and Beaverhill Lake Group by Leavitt and Fischbuch (1968). During the interim several comprehensive sedimentological analyses were presented. Edie (1961), Thomas and Rhodes (1961), Carozzi (1961), Fischbuch (1960, 1962, 1968), Murray (1965, 1966), Jenik and Lerbekmo (1968), Leavitt (1968), Hemphill *et al.* (1970) have described stratigraphic and sedimentological aspects of individual reef complexes and the interrelationship of the various reef complexes of the Swan Hills Formation. Paleontological studies have been published by Stearn (1963), Brown (1963), McGill (1966), and Fischbuch (1969, 1970a, 1970b).

The Gilwood clastic unit was first defined by Guthrie (1956), and since that time further sedimentological studies have been done by Kramers and Lerbekmo (1967), and Shawa (1969).

Study Area

Geographically the area lies in west-central Alberta (Fig. 1) about 150 miles northwest of Edmonton, between the fifth and sixth meridians (rges. 1 to 26 W 5th mer.), and extends from township 59 in the south to township 75 in the north (54°00′N to 55°20′N and 114°00′W to 118°00′W). Since the area extends from Sturgeon Lake in the western part of the study area to Mitsue on the eastern margin, it is referred to in the text as the Sturgeon–Mitsue area (*see* Fig. 3).

Geological History

The western Canadian sedimentary basin is underlain by the westward continuation of Precambrian rocks of the Canadian Shield. The oldest Cambrian basin on the Shield rocks was restricted to the Cordilleran trough, from which lower Paleozoic seas transgressed eastward over the cratonic shelf.

During Caledonian tectonic activity, the study area was uplifted and the pre-Devonian sediments removed; thus the basal Devonian strata unconformably overlie the eroded Cambrian section and in the northwest lap onto the Precambrian granitic terrane of the Peace River Arch. At the beginning of Early Devonian time a moderately subsiding basin was established south of the Tathlina uplift (Fig. 2). In the basin

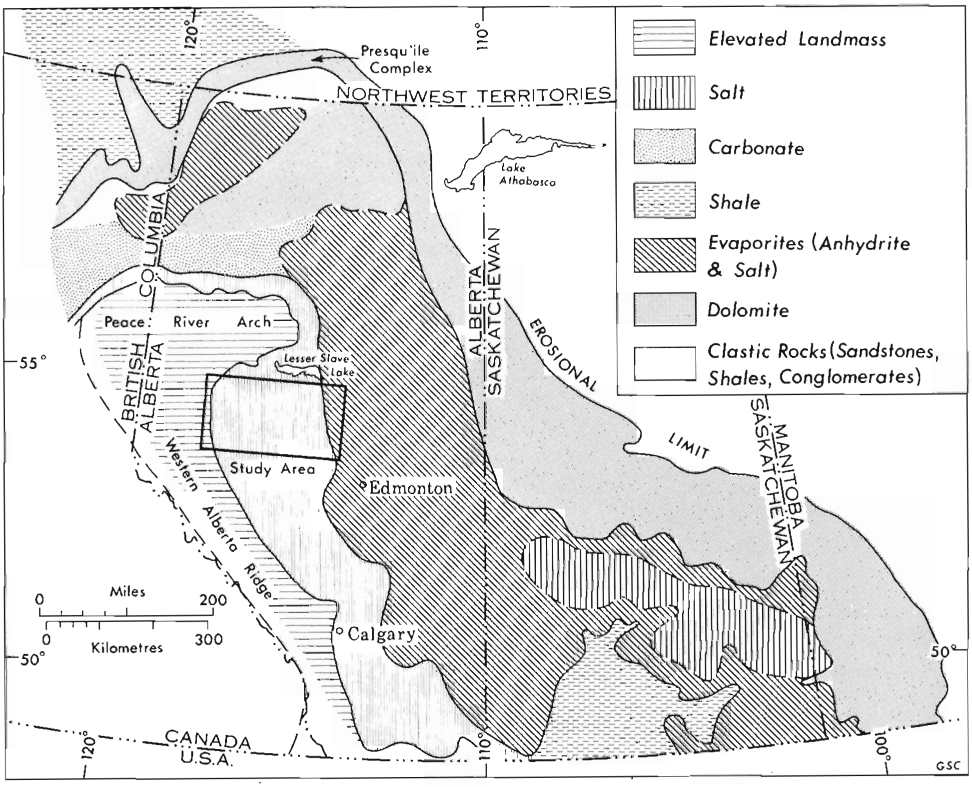


FIGURE 2. Lithofacies map of Upper Elk Point Basin (Late Givetian) (modified after Grayston, Sherwin, and Allan, 1964; and Douglas *et al.*, 1968).

600 to 1,300 feet (185 to 400 m) of clastic, evaporite and carbonate sediments were deposited. Restricted sea-water circulation in the Upper Elk Point basin led to a predominately evaporite (Muskeg Formation) and clastic (Watt Mountain Formation) accumulation. The Fort Vermilion Formation evaporites were deposited during standstill or slight regression of the sea before the main transgressive phase of late–middle Devonian began (Fig. 4).

During Late Devonian transgression in the study area the Swan Hills Formation was deposited (Fig. 3), which consists of a basal carbonate blanket. marginal reef-banks (for discussion of the term reef-bank *see* p. 46) and subsequently reef complexes.

Shaly deposits of the Waterways Formation accumulated in the deeper parts of the basin. As a result of a slowly transgressing sea, carbonate bank, reef, and shale deposition continued throughout Upper Frasnian into Famennian time; during the Famennian open-marine limestones and marine shales were deposited. Devonian rocks in the study area are approximately 3,600 feet (1,100 m) thick, comparable with 4,500 feet (1,350 m) of sediments deposited in the centre of the Elk Point Basin (east–central part of Alberta). The Devonian sedimentary sequence thins markedly (to approximately 300 feet, 90 m) over the Peace River Arch. For a more detailed Devonian history of the western Canadian sedimentary basin *see* Belyea *et al.* (1966), Bassett and Stout (1967), and Douglas *et al.* (1970). The basic stratigraphy of the area is discussed under general descriptions of the individual formations.

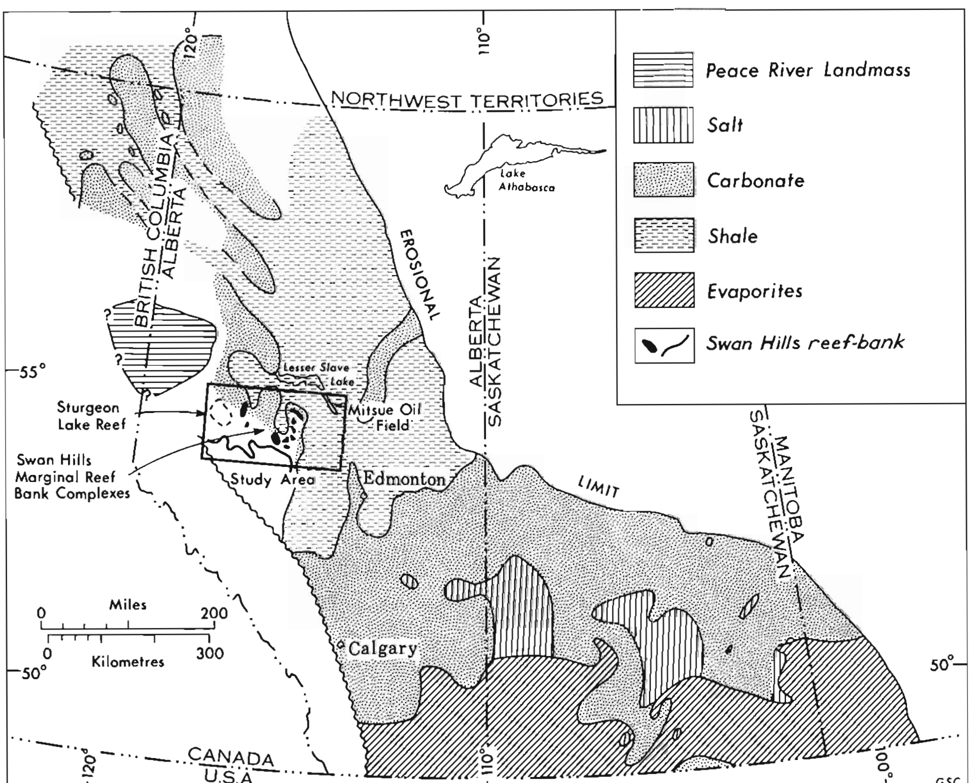


FIGURE 3. Lithofacies map of the Late Givetian – Early Frasnian Basin (southern part modified from Grayston Sherwin, and Allan, 1964).

Method of Study

More than 1,500 mechanical logs from wells in selected parts of the Sturgeon–Mitsue area were studied to obtain thicknesses of individual lithostratigraphic units. These data were plotted and isopach maps of the Gilwood Member, Fort Vermilion Formation, and Swan Hills marginal reef-bank complexes were constructed. Wells with continuous core were subsequently chosen for a more detailed study in a line perpendicular to the strike of sedimentation in an attempt to cover the complete environmental spectrum. Cores from 15 wells were studied in detail for lithological changes, structures, and contacts, in the Watt Mountain and Fort Vermilion Formations; as well as cores from 15 wells from the Swan Hills and Waterways Formations. Four hundred and ninety-five hand specimens were collected. Slabbed carbonate rocks were polished and etched; approximately one third of the polished area was stained by Dickson's method (1965, p. 587) to distinguish calcite, ferro-calcite, dolomite, and ferro-dolomite. Stained and unstained peels also were prepared.

Grain size, sorting, roundness, and composition were recorded. Peels were further examined under a petrographic microscope. Three hundred thin sections, also partly stained by Dickson's method, were prepared and studied under a petrographic microscope. In addition, some samples were analyzed by X-ray diffraction to determine percentage and composition of feldspars, carbonates and clay minerals in the fine-grained sediments.

Petrographic terminology is based on Pettijohn's (1957) classification of terrigenous rocks and Dunham's (1962) classification of carbonate rocks. Dunham's mudstone class was modified according to Folk (1962). Roundness values were estimated by using Krumbein's (1941) visual chart, and a sorting chart was used for routine sample description (Folk, 1968, p. 104). In selected samples sorting coefficient was calculated from thin sections, using procedures of Folk and Ward (1957).

Procedure

The petrographic parameters used for description of terrigenous rocks (sorting, roundness) are not consistent with those used for the description of carbonate rocks. Texture of carbonate rocks is influenced not only by the transported material but also by the skeletal material produced in situ which is modified by biogenic activity (e.g., Swinchatt, 1965). Skeletons and whole shells can be disaggregated by mechanical action as well as by the action of living biota, which, along with the destruction of binding organic components, can lead to disintegration with release of individual carbonate crystals into the sediment. Other processes of textural modification lead to the enlargement of particles; for example, the formation of aggregate particles, pellets, composite grains, mechanical coating (oölites and pisolites), or biological coating (oncolites). Therefore, parameters such as coefficient of sorting and skewness used for clastic grain size distribution are unreliable for carbonate grains. Similarly the coefficient of roundness has limited use in carbonate rocks because of the influence of biological breakdown on the shape of grains. A common approximate index of relative current strength and environmental indicator is the percentage of silt- and clay-sized fractions in carbonate rocks (Ginsburg, 1956; Folk, 1962; Purdy, 1963a, b). However, Davies (1970) has shown that sea grass can act as a baffle and trap a substantial amount of fine-grained carbonate material in a relatively high energy environment. It can be inferred that during

Devonian time populations of other biota such as algae, corals, sponges, or crinoids could have trapped fine-grained material much as sea grass does today. Primary crystallization of micrite in fossil voids in the reefs of Bermuda has been demonstrated by Ginsburg *et al.* (1971, p. 472) and from Red Sea reefs by Freidman and Amiel (1970, p. 555). Micrite in the interstices of the reef framework can accumulate to near 50 per cent of total sediment volume (Mesolella *et al.,* 1970) both by entrapment and the in situ production of micrite by the biological breakdown of skeletal material; nevertheless, associated energy conditions may be high. It can be concluded, therefore, that environmental and hydrodynamic interpretations based solely on the percentage of fine-grained carbonate material could lead to erroneous conclusions. For these reasons the authors based hydrodynamic and environmental interpretation of the biohermal carbonates mainly on paleoecological aspects of the biota itself and lateral facies interrelation rather than on the ratio of micrite to the carbonate grains.

Acknowledgments

The Alberta Conservation Board permitted examination of cores and provided samples for laboratory studies; Helen Belyea and A. D. Baillie guided the preliminary phases of this project. A. E. H. Pedder and A. W. Norris kindly undertook the task of paleontological identification of megafossils collected by the senior author. A. W. Norris and R. W. Macqueen critically read the original manuscript, and their profound knowledge of the subject led to its improvement. Discussions with members of the Geological Survey and geologists of the petroleum industry in Calgary were most helpful.

MUSKEG FORMATION

Introduction

The Muskeg Formation (Fig. 4) was first defined by Law (1955, p. 1954) for a series of anhydrites and fine-grained dolomites in the Steen River area of northern Alberta. He defined the formation as representing those rocks occurring between the top of the Keg River Formation and the base of the Watt Mountain clastics. This definition, however, later was modified by Belyea and Norris (1962, p. 10), Hriskevich (1966, p. 249), and McCamis and Griffith (1967, p. 441), who regarded the upper part of Law's Muskeg Formation as the Sulphur Point (Bistcho) Formation. The Sulphur Point Formation loses its identity in the upper part of the Muskeg Formation to the south and east, and in the Sturgeon–Mitsue area it is considered as part of the Muskeg Formation.

The Muskeg evaporites grade laterally northward into the Presqu'ile dolomites of northeastern British Columbia, northern Alberta, and the southern part of the Northwest Territories (Fig. 2). They extend southeastward through central Alberta and Saskatchewan to Manitoba and North Dakota, where they are known as the Prairie Evaporite Formation.

The total thickness of the formation ranges from a zero depositional edge along the eastern flank of the Peace River Arch and Western Alberta Ridge to a maximum of 900 feet (270 m) near the axis of the Elk Point basin. Only the uppermost part of the Muskeg Formation, or that part that lies just below the base of the Gilwood clastic unit, was examined in this study. The maximum core penetration of the Muskeg Formation in the Sturgeon–Mitsue area is approximately 30 feet (9 m), and it is from this interval that the following description and interpretation were made.

Composition

In the Sturgeon–Mitsue area the upper part of the Muskeg Formation is composed of alternating beds of bedded nodular (Fig. 5d) and mosaic (Fig. 5c) anhydrite, finely crystalline and sucrosic dolomite, and minor occurrences of halite. Highly bioturbated, dark grey calcilutite with interbeds of dark grey shale and minor amounts of red and green, mottled, calcitic or dolomitic shale also are present, with the shales being most common toward the base of the examined interval. The strata are unfossiliferous.

The argillaceous calcilutites are unfossiliferous, dark grey, and contain one per cent or less of floating silt-sized quartz grains. There is no obvious internal structure, but a vague bioturbation swirl pattern is apparent.

The microcrystalline dolomites commonly are grey, although in some places they have a green or reddish hue. Well-rounded sand grains are scattered throughout the dolomites.

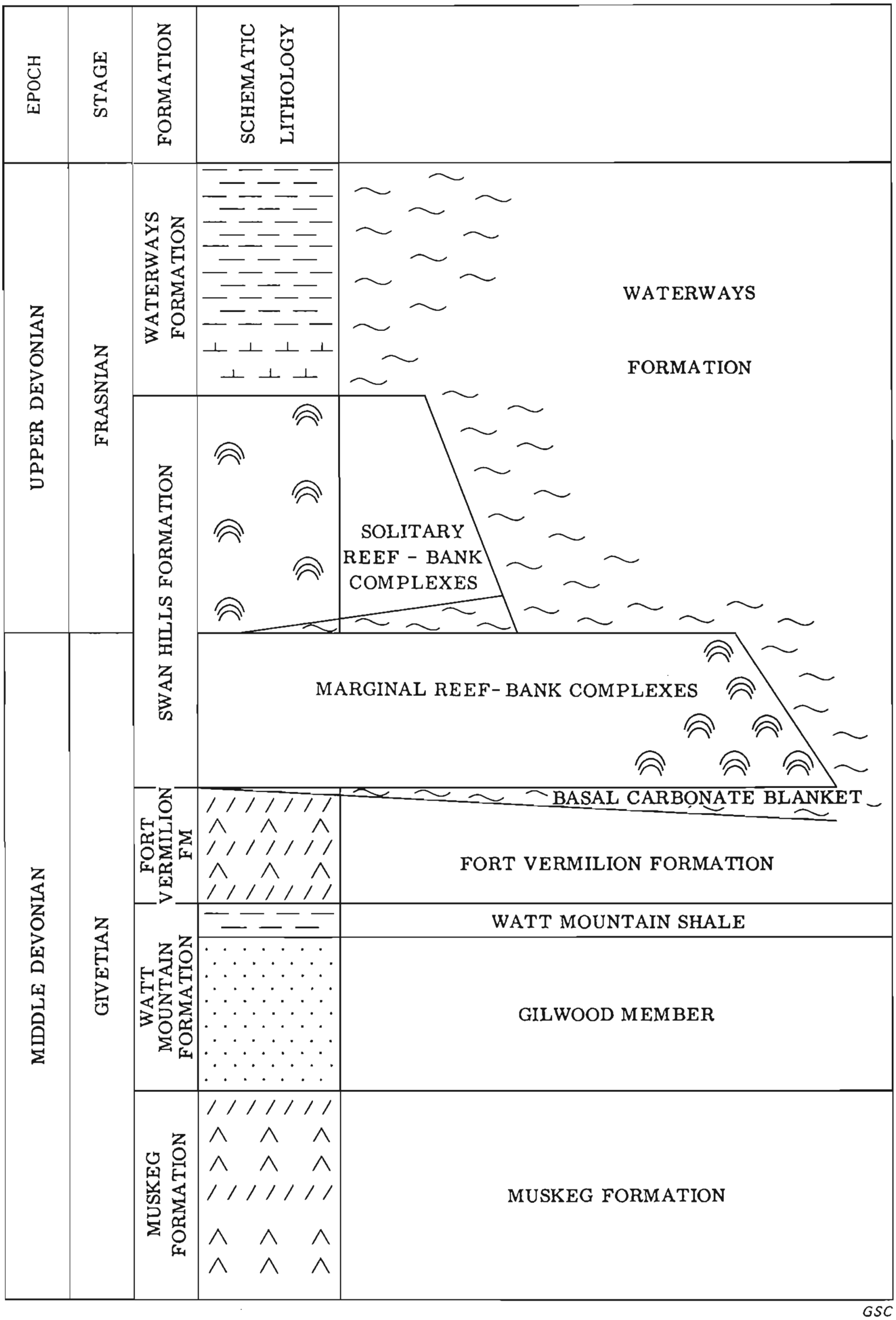
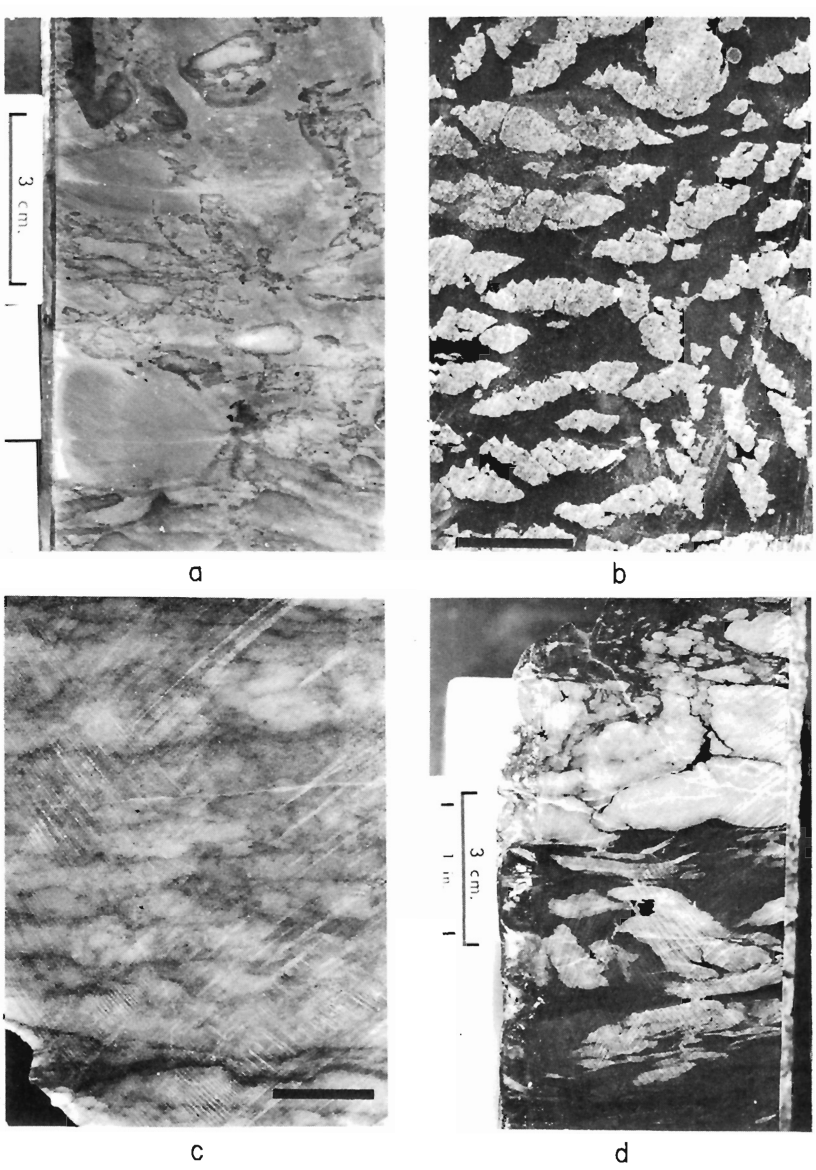


FIGURE 4. Devonian stratigraphic nomenclature, west–central Alberta, Canada.

FIGURE 5



a. Solution breccia at the top of the Muskeg Formation; brecciated shaly limestone enclosing purple coloured anhydrite nodules; Muskeg Formation. Core photograph, 4-28-71-4 W5, depth 6,060 ft (approx. 1,844 m).
b. Red coloured anhydrite pseudomorphs after gypsum crystals in a dolomitic shale; Muskeg Formation. Core photograph, 10-13-68-16 W5, depth 9,234 ft (approx. 2,815 m). Bar scale is equal to 3 cm.
c. Weakly bedded, mosaic anhydrite; Muskeg Formation. Core photograph, 2-6-70-10 W5, depth 7,666 ft (approx. 2,340 m). Bar scale is equal to 2 cm.
d. Red coloured nodular anhydrite in a grey shale; Muskeg Formation. Core photograph, 10-13-68-16 W5, depth 9,234 ft (approx. 2,815 m).

Anhydrite occurs in beds up to 5 cm. and is interbedded with microcrystalline dolomite. It also occurs in the form of nodules and as isolated lath-shaped crystals scattered throughout the dolomites. Anhydrite pseudomorphs after gypsum crystals are rare (Fig. 5b). The anhydrite crystals, which are as much as 3 cm long, occur in the dolomites and dolomitic shales with no apparent orientation. Crystal size of the anhydrite in the nodules, beds, and fine laminations is much smaller, and is commonly 0.01 to 1.0 mm. The texture of the anhydrite ranges from subfelted (Maiklem *et al.*, 1969, p. 223) in the nodules to a mosaic type in the beds and laminations. Locally the bedded anhydrite displays an aligned felted pattern, and may contain small patches and partings of organic material.

Finely disseminated pyrite (from 2 to 3 per cent) occurs as globular aggregates, or as halos forming oxidized rims around anhydrite nodules.

Sedimentary Structures

Significant structural patterns apparent in the Muskeg Formation are mudlumps, mottlings, burrow trails, desiccation cracks, horizontal laminations, and subrounded lithoclasts. The flat lithoclasts are as much as 4 cm long, and are composed of finely crystalline dolomite. Brecciated dolomite crusts in a green shale matrix and solution (collapse) breccias (Fig. 5a) occur near the top of the formation.

Discussion

Environment of Deposition

During the late Muskeg time the area from the Peace River Arch to Swan Hills appears to have been an extensive tidal flat which prograded eastward toward Mitsue to a slightly deeper and more open-marine environment. The sediments deposited in the hypersaline, oxygen-starved lagoons and ponds of this foreshore tidal flat are represented by varicoloured argillaceous dolomites. Embedded layers of brecciated dolomitic crust, anhydrite and the presence of desiccation cracks, all of which occur in close association with the red coloured shales, suggest intermittent exposure in a supratidal environment. In places where dark grey calcilutite occurs, slightly deeper waters are inferred which would indicate a subtidal environment. The absence of organisms, with the exception of burrowing animals, indicates a higher salinity typical of shallow semirestricted basins. The detrital quartz sand and silt appear to be of eolian origin and, together with the evaporites in tidal flat areas, suggest an arid or semiarid climate.

In general, sedimentary conditions in the upper part of the Muskeg Formation in the Sturgeon–Mitsue area can be compared with recent sedimentation along the Trucial Coast (Kendall and Skipwith, 1969) and to the Shark Bay in Western Australia, especially with the area between the Wooramwel Delta and Hamelin Pool (Logan *et al.*, 1970, p. 34–35). In this semiarid climate red beds have been deposited over broad tidal flats in semirestricted hypersaline or metahaline waters.

WATT MOUNTAIN FORMATION

Introduction

The Watt Mountain Formation was first defined by Law (1955, p. 1951) from the Steen River 2-22 well in northwestern Alberta. Law delineated 155 feet (46.5 m) of shale, siltstone, sandstone, arkose, limestone breccia, anhydrite, and dolomite as the Watt Mountain Formation, and equated it with Campbell's (1950, p. 90) Amco Shale of the Pine Point area in the Northwest Territories. This clastic unit is represented in the Elk Point area of central Alberta by 55 feet (approx. 17 m) of red and green shale and siltstone, which in turn is correlative with a green dolomitic, silty shale known as the First Red Bed unit in Saskatchewan and Manitoba (Grayston *et al.*, 1964, p. 54).

In the vicinity of the Peace River Arch a terrigenous clastic unit known as the Gilwood Sandstone (Guthrie, 1956, p. 227) interfingers with the Watt Mountain shales, and is known as the Gilwood Member of the Watt Mountain Formation.

The Gilwood Member is one of the highest and most widespread of a series of clastic wedges that emanate from the Peace River Arch (Fig. 3). Adjacent to the Arch these wedges coalesce to form a nearly continuous clastic unit which commonly is referred to as granite wash.

The Gilwood Member extends irregularly north and east from the Peace River Arch for a maximum of 150 miles (approx. 241 km). There it pinches out rapidly and in places forms traps for oil. Thickness of the member ranges from 60 feet (18 m) in the Mitsue area to 95 feet (approx. 30 m) at Nipisi (Shawa, 1969, p. 394), and to 155 feet (48 m) north of the Arch at the Manning 8-4-92-23 W5 well. The sandstones north of the Arch sometimes are informally referred to as the Manning sandstone (Mason, 1960, p. 215). Lateral thickness variations of the Gilwood Member and locations of wells used in this study are shown in Figure 6 (*in pocket*).

In the area of study a lower boundary is distinct in places where Gilwood sandstones overlie Muskeg Formation dolomites (Fig. 12, *in pocket*). The boundary is less well defined, however, where Watt Mountain shales unconformably overlie similar rocks of the Muskeg Formation, and thus gives the impression of a gradational contact.

Composition

The Gilwood Member is composed of interbedded conglomerates, sandstones, and shales. A characteristic feature of the Gilwood clastic unit is a simple to composite cyclic repetition of lithotypes from 5 to 10 feet thick (Fig. 12). Each cycle commonly is represented by a lowermost conglomeratic sandstone bed from 0.5 to 1 foot thick followed by several feet of medium-grained, poorly sorted sandstone. In the area from Sturgeon to Snipe Lake the basal part of the cycle consists mainly of pebble-sized grains (Fig. 7a). To the east, toward the Mitsue area, the pebbles are less common,

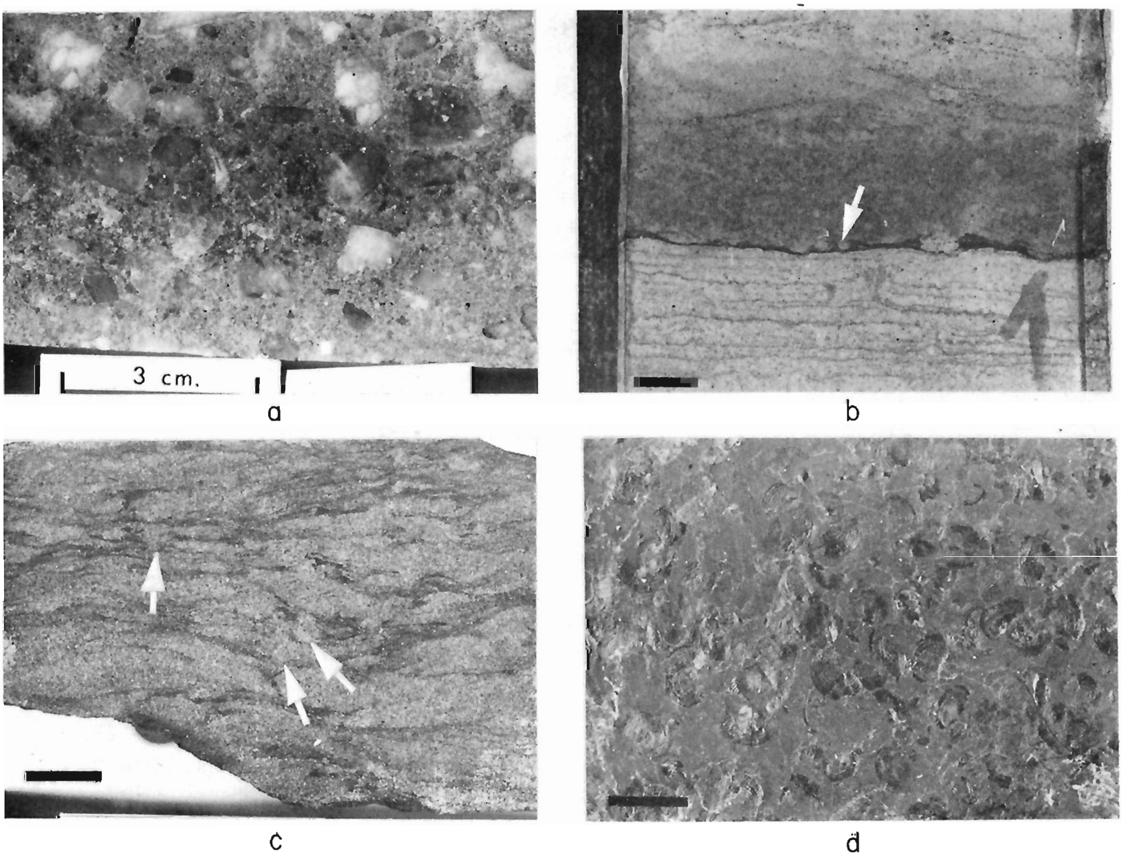


FIGURE 7. a. Arkosic, sandy conglomerate, with subangular quartz pebbles; Gilwood Member. Core photograph, 4-22-69-18 W5, depth 9,455 ft (approx. 2,880 m).

b. Horizontally bedded, well-sorted, fine-grained sandstone, overlain by small-scale multidirectional trough bedding; note bioturbation in thin, dark shale layer between units; Gilwood Member. Core photograph, 2-6-70-10 W5, depth 7,642 ft (approx. 2,340 m). Bar scale is equal to 1 cm.

c. Red coloured sandy siltstone with argillaceous laminae disrupted by burrowing; Gilwood Member. Core photograph, 4-22-69-18 W5, depth 9,467 ft (approx. 2,880 m). Bar scale is equal to 1 cm.

d. Dark grey shale with *Lingula* sp. and the conchostracan *Asmussia* concentrated along bedding planes; Gilwood Member. Core photograph, 10-35-71-5 W5, depth 5,999 ft (approx. 1,820 m). Bar scale is equal to 1 cm.

as much as 1.5 cm in diameter, and float in a medium- to coarse-grained sandstone matrix; however, local concentrations of granule-sized material occur throughout each cycle in beds up to 5 cm thick. The upper part of each cycle commonly consists of fine-grained sandstones (Fig. 7b) that grade upward into siltstones and silty shales (Fig. 7c). These shales in the eastern part of the area contain the conchostracan *Asmussia* (Norris, 1970) and linguloid brachiopods (Fig. 7d) which are typical of brackish-water deposits. The contact with the overlying, or younger cycles, in most places is erosional. In the composite cycles there may or may not be an erosional surface; whether erosion did or did not take place appears to have been dependent upon the flow regime. A gradual increase of the flow velocity would produce an erosional contact. Grain-size distribution in the cycles is mostly normal, although a reverse vertical grain-size distribution is apparent in some places. In the study area a lateral variation in grain size also can be noted, with grain size diminishing eastward. This is best demonstrated by lateral variations of sandstone occurrences: in the Sturgeon area 53 per cent of the Watt mountain clastic unit is composed of sandstone; at Snipe Lake, 32 per cent; and west of Swan Hills less than 1 per cent; at Mitsue there again is an increase to 60 per cent sandstone in the total section. East of Mitsue the Gilwood sandstones pinch out completely among the Watt Mountain Formation shales.

The Gilwood clastic unit has a broad spectrum of textural components, grain size, and rock types deposited in an environment that shifted vertically and laterally from marine to fluviatile, thus causing much mutual overlap. Environmental factors controlling sedimentation, therefore, are difficult to assess from limited descriptive petrographic data. Of the many parameters examined, only five were found to have the necessary lateral continuity to be traced throughout the study area. From these parameters gross differences in the original sedimentary environment are suggested. Minor local fluctuations undoubtedly also took place, but are not discernible owing to lack of core control. Factors examined in this study are particle size, per cent matrix, per cent feldspars, sorting, rounding, and sedimentary structures. Lateral changes of the above parameters and the related environmental significance are summarized in Figure 8. Figure 9 summarizes the biogenic parameters.

Sandstones

In grain size there is a complete gradation from very fine to very coarse. In the Sturgeon area median diameter of the grains is 1.32ϕ (phi units); to the east in the Snipe Lake to Swan Hills area the median diameter decreases to 3.64ϕ; and at Mitsue there is a substantial increase to 1.5ϕ.

Quartz—The Gilwood Member is composed of from 70 to 95 per cent quartz of the following varieties, described in order of their relative abundance. (1) Single crystals, which portray a straight or slightly undulose extinction; they are clear or slightly cloudy and rarely contain vacuoles. A logical source for this type of quartz would be acid plutonic rocks. (2) Crystals characterized by moderately undulose extinction, contain rare vacuoles, and display growth laminations. This variety is typical of vein type quartz. (3) Composite grains with sutured and crenulated margins that outline individual crystals, and have an undulose extinction. These grains are typical of pressure metaquartzites formed under dynamothermal metamorphism (Todd and Munroe, 1968, p. 1031). This category also would include indi-

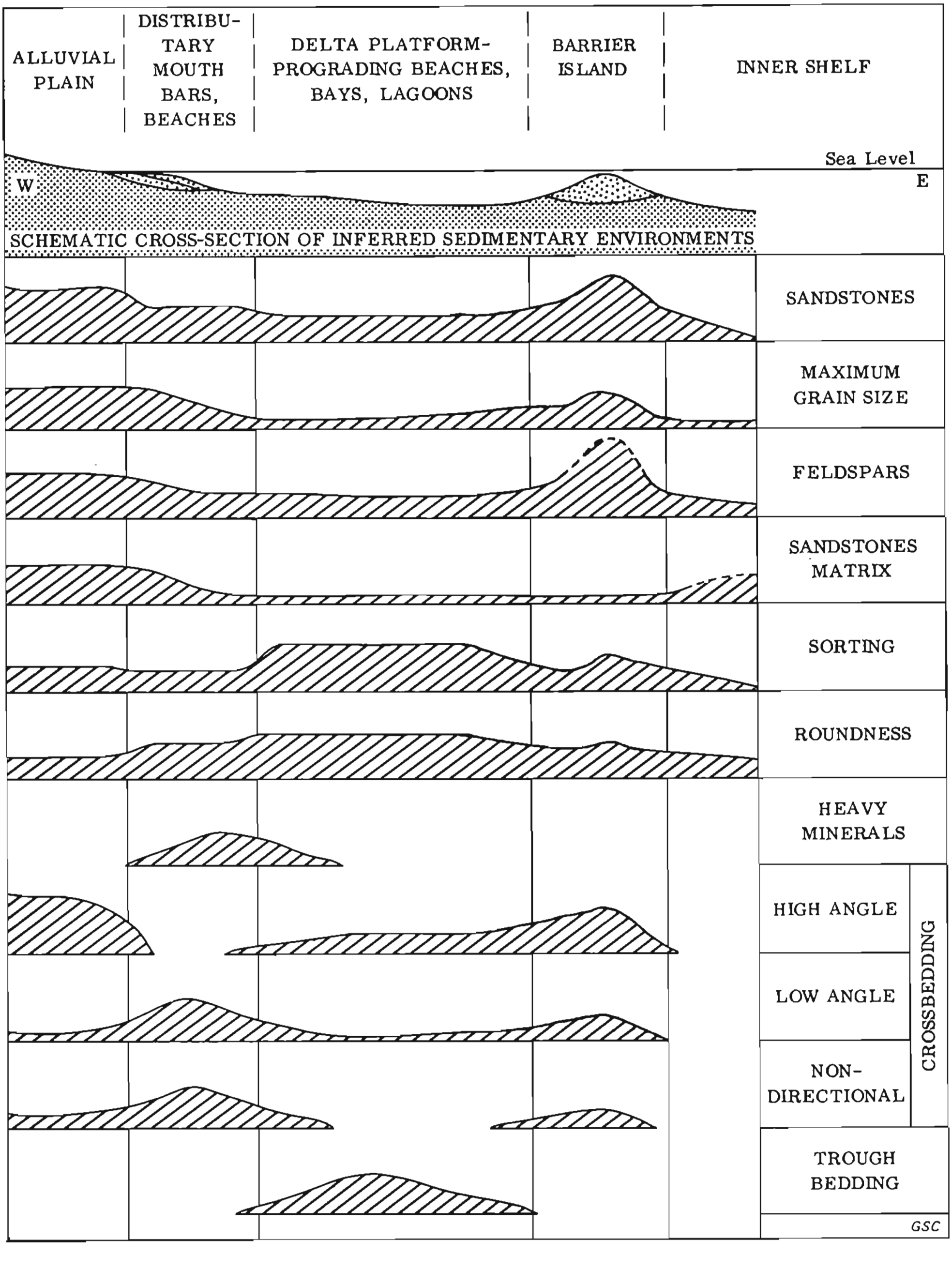


FIGURE 8. Generalized west–east variations in abundance and compositional parameters of sandstone and sedimentary structures in the Gilwood Member; inset above shows schematic cross-section of inferred sedimentary environments.

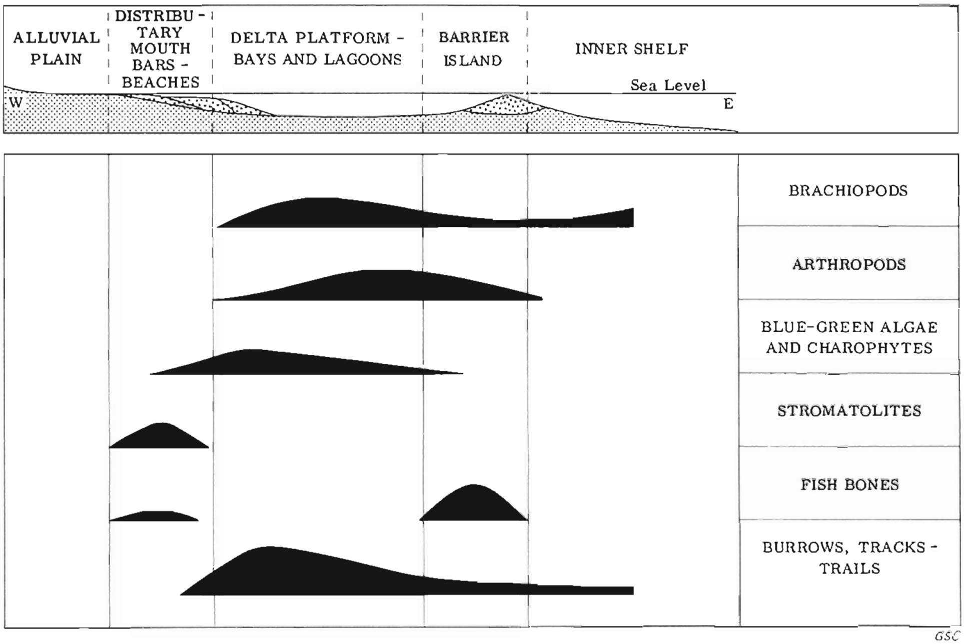


FIGURE 9. Generalized west-east variations in relative abundance of preserved fossils in the Gilwood Member inset above shows schematic cross-section of inferred sedimentary environments.

vidual grains that have a highly undulose extinction, as well as those grains with fibrous microlitic inclusions of sagenite, which are characteristic of low-grade metamorphic slate. A metamorphic source rock is inferred for this type of quartz.

Polycrystalline aggregates commonly form large grains and pebbles in the conglomeratic sandstone, and are composed mainly of quartz varieties (2) and (3). Boundaries of those grains with moderately undulose extinction are blocky with straight margins; whereas, the metaquartzite grains have highly sutured boundaries, and elongate grains are oriented parallel to the long axis.

In the study area the abundance of composite grains in the Gilwood Member decreases eastward. For example, at Sturgeon, composite grains make up 45 per cent of the total quartz, whereas at Mitsue they make up only 7 per cent. Decrease in percentage of composite grains could be explained by fracturing of the composite grains along crystal boundaries during transportation (Blatt, 1967, p. 422) resulting in the formation of more individual crystals and a general decrease in grain size.

Feldspar—Since the thin sections used in this study were stained only for ferrocalcite, calcite, and dolomite, and not for K-feldspars it was difficult to distinguish with certainty K-feldspars, in the finer grained sandstones. The Gilwood sandstones contain microcline and orthoclase, but no plagioclase. In the Nipisi area to the north, however, Shawa (1969, p. 395) reported up to 1 per cent plagioclase, and Kramers and Lerbekmo (1967, p. 358) noted minor amounts. In the study area feldspar content decreases eastward from an average of 14 per cent at Sturgeon to an average of 9 per cent in the Snipe Lake to Swan Hills region. This eastward decrease probably is a result of the relative mechanical instability of feldspars which, during transportation, would disintegrate more rapidly than the quartz. In the Sturgeon area feldspar grains generally are larger than quartz grains, and all grains are angular to subangular (Fig. 10a). Alternatively, at Mitsue feldspar grains are slightly smaller than quartz grains and all grains are subrounded to rounded (Fig. 11b). At Mitsue, however, feldspar content has increased to an average 14 per cent. This may be the result of transportation of sediment by longshore currents from the Nipisi area to the north where Kramers and Lerbekmo (1967, p. 352) found an average of 32 per cent feldspars in the sandstones.

Little alteration is evident in the feldspars. Microcline, the feldspar most resistant to weathering, appears to be fresh or only slightly altered. The orthoclase, however, displays some vacuolitic alteration (Folk, 1955, p. 357), and sericitization is apparent along cleavage surfaces. Occurrences of weathered K-feldspars are rare; they could, however, represent a separate population. Authigenic overgrowths are very common on most of the feldspar grains (Fig. 10c), and, consequently, many grains exhibit a hypidiomorphic or idiomorphic form.

Mica—Traces of muscovite, biotite, and chlorite micas rarely reach more than 0.5 per cent of total rock volume. Higher quantities are present where the Gilwood Member becomes shaly or silty, with the mica concentrated into very thin laminae. In the Mitsue area, biotite is the most common mica and commonly occurs as small flakes, as opposed to muscovite which consists of relatively large plates. These flakes could be transported farther by currents than the larger muscovite plates. In most places there is little alteration of the micas, although rarely the biotite is altered to chlorite.

Heavy mineral composition has been determined only from thin section examination. Rutile, zircon, tourmaline, and apatite are recognized, and represent a stable mineral association. They occur mainly as accessory components, and as individual grains or rarely as concentrations into thin laminae. Heavy minerals commonly compose less than 0.5 per cent of the total rock volume and rarely exceed 2 per cent.

Phosphates—Fragmented fish remains are most common in the Mitsue area, and have been tentatively identified by R. Thorsteinsson (pers. com.) as belonging to the Piscean Order Arthrodira. Phosphatic remnants of fish scales and bones range up to several centimetres in diameter. Pores in the bones commonly are filled with either secondary silica or anhydrite.

Glauconite—Silt to fine-grained, sand-sized glauconite grains are rare in the Gilwood Member. Most of the glauconite is present in the area from Snipe Lake to Mitsue where, rarely, it may compose up to 8 per cent of the total rock volume (Fig. 11a).

Hematite and pyrite—Finely disseminated hematite and iron oxide coatings on the clay minerals (Fig. 10b) are found in variable amounts throughout the Gilwood Member. They are most common, however, in the Sturgeon area where hematitic cement comprises up to 10 per cent of some samples. The hematite in the Gilwood sandstones appears to be of secondary origin and may have been produced by weathering processes similar to those found today in arid and semiarid climates where mafic minerals and biotites are broken down (Walker, 1967, p. 353). A minor amount of pyrite, up to 0.5 per cent of total rock volume, also is present as an accessory mineral.

Matrix—Matrix of the Gilwood sandstones is composed of fine-grained, silt-sized particles of major sandstone constituents mixed with argillaceous minerals (Fig. 10a).

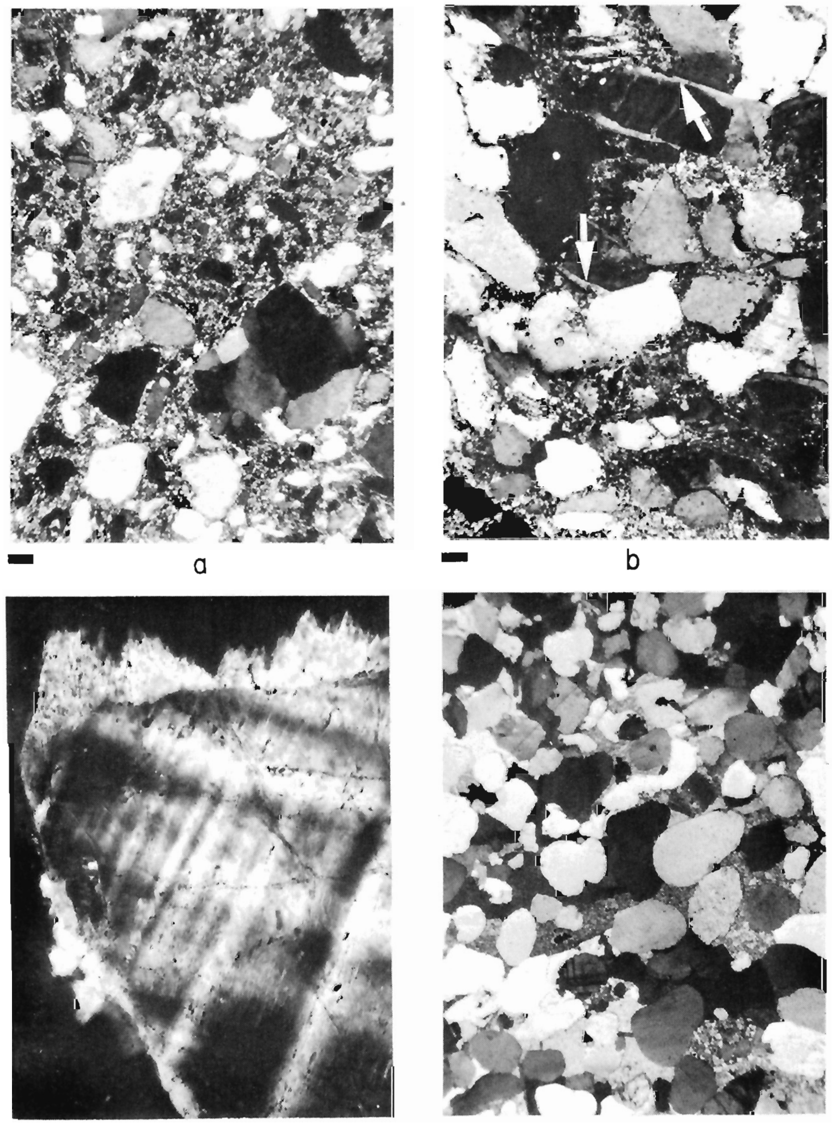


FIGURE 10

a. Greywacke; feldspathic, poorly sorted grains are angular to subangular; matrix is argillaceous; Gilwood Member. Thin section—crossed nicols. 10-14-72-23 W5, depth 9,815 ft (approx. 3,000 m). Bar scale is equal to 0.1 mm.

b. Feldspathic sandstone, medium grained; quartz grains angular to subangular, unsorted; argillaceous matrix is pigmented by hematite; note authigenic overgrowth of feldspar grains; Gilwood Member. Thin section—crossed nicols. 10-14-72-23 W5, depth 9,815 ft (approx. 3,000 m). Bar scale is equal to 0.1 m.

c. Authigenic feldspar overgrowth on microcline grain (marked by arrow); Gilwood Member. Thin section—crossed nicols. 4-22-69-18 W5, depth 9,472 ft (approx. 2,887 m). Bar scale is equal to 0.1 m.

d. Protoquartzite, coarse grained, well sorted, well rounded, cemented by anhydrite and silica; Gilwood Member. Thin section—crossed nicols. 2-6-70-10 W5, depth 7,634 ft (approx. 2,340 m). Bar scale is equal to 1 mm.

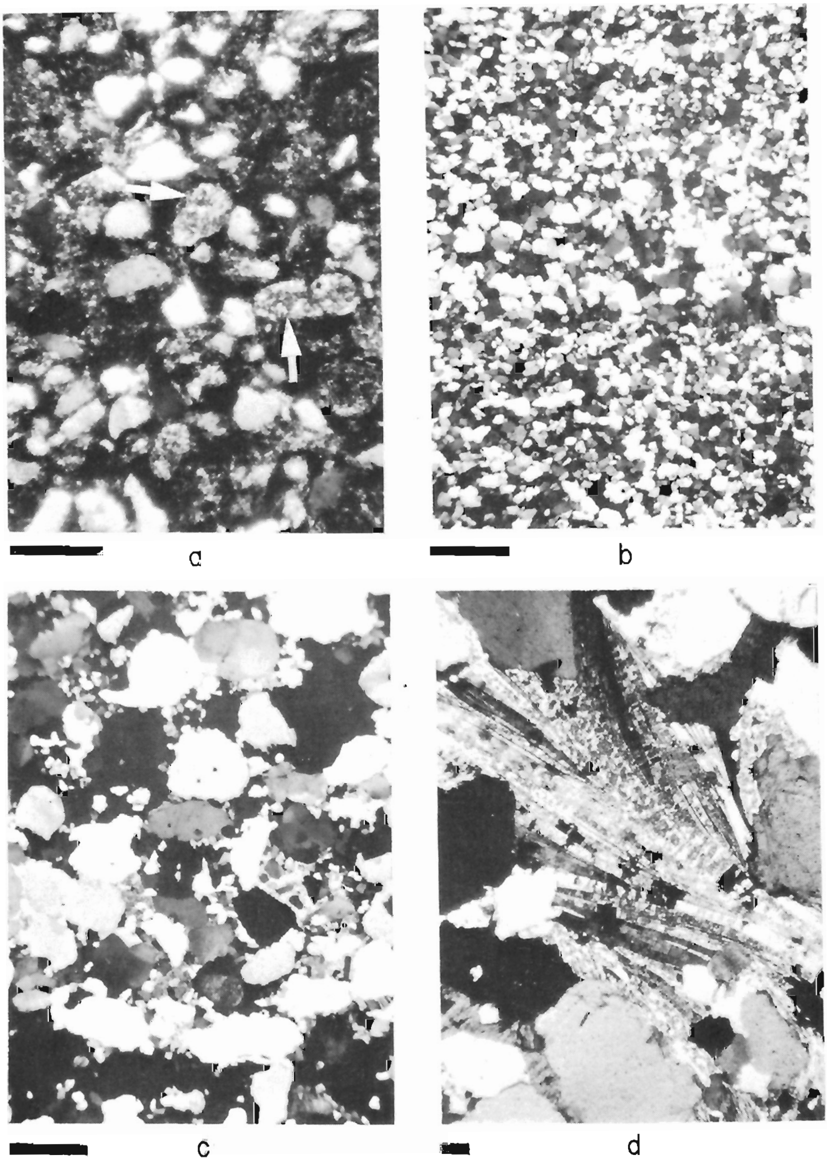


FIGURE 11

a. Poorly sorted glauconitic sandstone; subangular quartz grains and glauconite grains (marked by arrow) deposited in argillaceous matrix; Gilwood Member. Thin section—ordinary light. 4-28-71-4 W5, depth 6,039 ft (approx. 1,840 m). Bar scale is equal to 0.1 mm.
b. Glauconitic protoquartzite; grains are fine grained, well sorted, subrounded to rounded, and cemented by silica; coarser laminae in middle of picture are positively graded; Gilwood Member. Thin section—crossed nicols. 10-35-71-5 W5, depth 5,944 ft (approx. 1,800 m). Bar scale is equal to 1 mm.
c. Protoquartzite, coarse grained, grains well sorted, bimodal, cemented by sparry calcite and anhydrite; Gilwood Member. Thin section—crossed nicols. 2-6-70-10 W5, depth 7,632 ft (approx. 2,340 m). Bar scale is equal to 1 mm.
d. Fan-shaped aggregate of anhydrite crystals cementing grains in medium-grained feldspathic sandstone; Gilwood Member. Thin section—crossed nicols. 10-21-70-18 W5, depth 8,949 ft (approx. 2,730 m). Bar scale is equal to 0.1 mm.

The clay-sized fraction of some sandstones contains disaggregated micaceous fragments mixed with altered clay minerals. X-ray analysis reveals a mainly illitic composition of the clay minerals with a small amount of kaolinite and traces of chlorite. The matrix of the Gilwood sandstones in the Sturgeon area is hematite stained and fine pigmentation by hematite also occurs. Percentage of argillaceous matrix is highest in the Sturgeon area where the clay content of the matrix averages 15 per cent. Argillaceous content of the total matrix in the sandstones decreases eastward, and in the Mitsue area is less than 5 per cent of total rock volume.

Considering the argillaceous matrix content, these sandstones could be classified as immature and submature in the Sturgeon area with maturity increasing eastward. The most easterly deposited sandstones (Mitsue area) appear to be mature and submature. Sandstones of the Gilwood Member can be classified as feldspathic sandstones (Pettijohn, 1957), less commonly as arkoses and feldspathic greywacke (Fig. 10a), and rarely as orthoquartzites.

Cement—The Gilwood sandstones are cemented by anhydrite, calcite, dolomite, and silica, which range from 0 to 25 per cent of the total rock volume. The cement occurs both as a void filler in intergranular spaces, and as contact cement along grain boundaries. Anhydrite, calcite, and dolomite are present as void-filling cement, whereas silica occurs mainly as contact cement.

Anhydrite occurs (1) as large lath-shaped crystals to 3 mm long, commonly portraying a poikilotopic fabric; (2) as rosettes composed of anhedral crystals and filling intergranular voids (Fig. 11d); and (3) as nodules composed of small crystals displaying an aligned felted and subfelted fabric, with the alignment commonly parallel to the bedding plane. Anhydrite cements are most common near the base of the Gilwood Member where this formation grades into the underlying evaporites of the Muskeg Formation.

The calcite and dolomite contents are of both high and low iron content varieties, although the dolomite cement generally has a lower iron content. Texture of the calcite and dolomite is a fine to coarse crystalline mosaic, with crystals up to 0.6 mm in diameter.

Calcite occurs as a void-filling cement and also partly metasomatically replaces quartz grains and silica cement.

Cementation by silica is common in the cleaner (low matrix) sandstones, where diagenetic silica overgrowths are common in quartz grains. Silicification appears to be most prominent where the grains are tightly packed, and pressure solution appears to have occurred at some grain contacts.

Sorting and Roundness

Coefficient of sorting (Folk and Ward, 1957) has been estimated by measurement of grain sizes in thin sections. Gilwood sandstones in the Sturgeon Lake area (Fig. 10a) are poorly sorted (σ=1.5–1); to the east in the Swan Hills area they are moderately to well sorted (σ=0.63–0.35); Figure 11c shows the well-sorted bimodal sandstone from this area. In the Mitsue area the sandstones are moderately sorted (σ=0.6) with occasionally interbedded, well-sorted (Fig. 11b) sandstone beds.

For roundness (R) Krumbein's (1941) classification scale has been used. Roundness of clastic grains follows a pattern similar to sorting, and increases from the western to the eastern part of the study area (Fig. 8). Roundness of sandstone grains to the west in the Sturgeon area averages R=0.39; in the Swan Hills area it ranges R=0.63 to 0.8; to the east in the Mitsue area it again decreases slightly with R averaging 0.50.

Siltstones

Siltstones up to 4 feet thick commonly are interbedded with silty shales and sandstones. They are medium grey, and in some places have a greenish or reddish hue.

Quartz, which is the main constituent, has a wide range of grain sizes, and commonly is subangular to subrounded. Feldspar content ranges from 0 to 10 per cent of total rock volume, and mica is common. Gilwood siltstones of the Sturgeon–Mitsue area are characterized by glauconite. Fine-grained phosphatic debris derived from fish bones and fish scales is rare in siltstones. The matrix of the siltstones commonly is argillaceous, less commonly calcareous or dolomitic, and rarely anhydritic. Clay minerals determined by X-ray analysis are illite, kaolinite, and less commonly chlorite.

Shales

Shales of the Gilwood Member in some places contain silt and sand grains, occasionally they are calcareous, and range from medium to dark grey, to green, and rarely to red. The shales occur mainly in beds one foot to 2 feet thick, but rarely in beds up to 10 feet thick. Shales commonly are interbedded with siltstones, dolomites, and less commonly with fine-grained sandstones, and have a massive blocky appearance with a weak parting developed parallel to the bedding. In some places the grey and dark grey Gilwood shales are fossiliferous, with shells concentrated in fine-grained coquinas. Charophytes identified by Norris (1970) as *Eochara* cf. *E. wickendeni* Choquette, and fish scales are present in green shales that lie at the top of the Gilwood Member, in the Swan Hills and Mitsue areas. No fossils were found in the Sturgeon area.

Mineralogical composition of the shales was determined by X-ray diffraction analysis (Jansa). Clay minerals present are mainly illite and less commonly kaolinite. In the red and green shales that lie at the base and at the top of the Gilwood Member, the kaolinite content increases significantly, with the ratio of illite to kaolinite in the order of 3 to 1. Almost all shales reveal at least small amounts of K-feldspars, although no feldspars were found east of the Mitsue area. Small amounts of quartz also are apparent.

Sedimentary Structures

Only small-scale sedimentary structures are recognized in cores. In this study crossbedding has been separated into two categories: high-angle crossbeds with angles greater than 15 degrees; and low-angle crossbeds with angles less than 15 degrees. The most common low-angle crossbeds range from 5 to 10 degrees. Since all measurements are from cores, planar and tangential crossbedding have not been differentiated and only small-scale (approx. less than 2 feet thick) multidirectional crossbedding can be recognized. For similar reasons burrows, tracks, and trails are grouped together. Flaser bedding occurs rarely. Ripple-marks and horizontal bedding were not recorded since they have no clear environmental significance. The west to east distribution of five key sedimentary structures is summarized in Figure 8. The changing patterns in the lateral and vertical distribution of sedimentary structures throughout the study area are valuable indicators and suggest significant changes in sedimentary environment through time.

Discussion

Source Area

The virtual absence of sedimentary rock fragments and the abundance of feldspar indicate that sedimentary particles were derived primarily from an igneous-metamorphic terrane, an interpretation which has also been suggested by Kramers and Lerbekmo (1967, p. 366). The most important source rocks probably were acidic plutonic rocks, mainly of granitic composition (e.g., hornblende granite). This type of rock would contribute the quartz, potassium feldspar, zircon, rutile, and mica found in the Gilwood Member.

Associated with this crystalline source there appears to have been hydrothermal aplitic and pegmatitic veins which produced vein quartz and some microcline. The second major rock type appears to have been a metamorphic complex that produced metaquartz, tourmaline, chlorite, and mica. The high percentage of microcline in the Gilwood Member could indicate acidic gneisses in the source area. Little evidence of a polycyclic depositional history is seen in the Gilwood sandstones; the feldspars and quartz grains are uniformly abraded and variations in grain rounding are minor. Minor reworking of older sediments could, however, have occurred.

Generally, it is assumed that the source for the Gilwood sediments was granitic Precambrian rocks.

Climate and Topography

Climatic conditions and topographic relief of the source area at the time of deposition of Gilwood sandstones are apparent from the weathered state of unstable mineral species, particularly feldspars and less commonly micas; the occurrence of hematite; and general grain-size distribution. Feldspar grains do not display a prominent degree of alteration. Microcline, which is the most stable feldspar under weathering conditions (Folk, 1968, p. 83), mainly is fresh. Orthoclase, on the other hand, displays a broader range of alteration, with the grains fresh to cloudy and brownish. According to Folk (1955, p. 357) this is a vacuolitization type of weathering. Slight sericitization of orthoclase grains is rare. Fresh, angular and subangular feldspathic grains are found in areas of moderately low relief and mainly in arid climates. The presence of K-mica, however, indicates restricted removal of potassium, because of limited leaching conditions (Todd, 1968) which could be explained by low precipitation in the source area. Vacuolitization weathering is associated with weathering processes in the source rock and is most common in hydrous rocks such as granites and pegmatites (Folk, 1968, p. 84).

The concentration of iron and hematite pigment in mudstones, shales, and the argillaceous matrix of the Gilwood sandstones is readily explained by initial concentration of iron from iron-rich biotite, hornblende and ore minerals, and subsequent alteration during weathering processes. According to Walker and Honea (1969), this type of alteration could occur in sediments derived from desert sources where weathering is minimal and soils are poorly developed. The ephemeral streams and rivers and other sedimentological properties would indicate that the climate during Gilwood time probably was semiarid, with minor seasonal rainfall, a climate comparable to that of recent depositional areas in the northwest part of the Gulf of California (Thompson, 1968) and the southeast part of Shark Bay, Western Australia (Logan *et al.,* 1970).

Environment of Deposition

In Middle Devonian time a marine regression resulted from epeirogenic uplift of the Peace River Arch; rate of flow of drainage systems undoubtedly increased substantially. Streams prograded over piedmont and alluvial plains and deposited clastic sediments on the coastal plain over the surface of the evaporitic sequences of the Muskeg Formation. Several ephemeral streams and rivers appear to have been active in the vicinity of the Peace River Arch. The area dealt with in this study was situated at the southeastern margin of this broad fluviomarine system. The marginal position of the area in this system is suggested by the occurrence of a broad zone of tidal flats which prograded seaward for at least 40 miles (64 km), and by decreasing thicknesses of clastics southward.

The depositional site of the Gilwood Member in the study area is characterized by a fluviomarine offlap sequence that can be divided into four main physiographic provinces. These are (1) alluvial plain, (2) coastal and deltaic plain, (3) barrier island, (4) prodelta and open-marine shelf (the latter term is substituted by the shorter term 'inner shelf' in Figs. 8, 9, 13). Each of these provinces constitutes a distinct complex of depositional environments (Fig. 13; for location of cross-section *see* Fig. 6) in which sediments of different character have accumulated. Provinces are subdivided into individual environments on the basis of composition, texture, minor structures, and faunal content. Emphasis is placed on three dimensional relationships and facies substitution (Jansa and Tomsik, 1961).

Alluvial Plain

The alluvial plain appears to have formed a transition between the higher elevated land areas and piedmont plain of the Peace River Arch to the west and low-lying coastal plain to the east. Width and morphology of the alluvial plain could not be determined and depend on the proximity of bedrock protuberances to the coastal zone. Sediments covering the alluvial plain are characteristically diverse in type and texture (Fig. 12, *in pocket*). Data are limited, however, as cored intervals are not common from the lower part of the Gilwood section. Mechanical logs were used to aid in tracing the general distribution of the deposits of the alluvial plain.

In the Sturgeon–Snipe Lake area the lower part of the alluvial plain appears to have consisted of poorly sorted and subangular, medium- to coarse-grained arkosic sandstones with interbedded gravelly sandstones (Figs. 7a, 10a). The sandstones are interbedded with red silty shales in a cyclic manner. The lower part of each cycle contains graded beds that have an erosional surface at the base. High-angle crossbedding is common and no faunal remains were found. The coarse-grained sandstones and conglomerates are interpreted as being channel deposits. The overlying ripple-marked and horizontally laminated siltstones and shales apparently are the product of an upper flow regime (Harms and Fahnestock, 1965, p. 108) and are interpreted as overbank and flood-plain deposits. The occurrence of carbonate concretions in the red silty shales (Fig. 14a) suggests leaching conditions with secondary concentration of carbonates in nodules, which is a process similar to one occurring in soils observed on the Indus alluvial plain and other semiarid regions (Allen, 1970, p. 143). The low percentage of fines in the sandstones reflects an incomplete source-rock weathering in a semiarid climate, or a low suspension/bed load ratio in the streams. The high percentage of hematite and the presence of red beds (Fig. 14b) suggest periods of low-water level on the alluvial plain.

Coastal and Deltaic Plain

East of Snipe Lake the alluvial plain appears to have graded into an area of diverse depositional environments associated with coastal and deltaic sedimentary regimes (Fig. 13, *in pocket*). The diversity of facies is complicated by progradation of environments. The coastal and deltaic plain of the Gilwood Member appears to have been formed approximately at sea level, and the plain margin was partly intertidal.

River channels in present-day delta plains are subjected to tides and function essentially as tidal streams, especially during low-water river discharges (Gagliano and McIntire, 1968). Distributaries, distributary mouth bars, levees, interdistributary bays and lagoons, beaches, mudflats, swamps, and marshes compose the delta plain area of modern rivers. Most of these environments are represented in the Gilwood Member: however, the deposits of levee, marsh, and swamp environments were not recognized.

Distributary Mouth Bars

In modern deltas sand that is carried as bed load by a river is deposited as a result of a decrease in velocity and reduction in carrying power just behind the mouth of a distributary. Deposition takes place rapidly and the sediments are quickly reworked by currents and waves (Coleman and Gagliano, 1965; Moore, 1966).

The occurrence of distributary mouth bars can be recognized east of the edge of the alluvial plain (east of Snipe Lake) mainly on the basis of a substantial decrease in the amount of argillaceous matrix in the sandstones from 15 per cent in interpreted alluvial plain sediments to less than 5 per cent in the distributary mouth bar environment. Furthermore, there is a higher degree of quartz grain rounding, multidirectional crossbedding, low-angle crossbedding, convex cross-lamination, and interbedding with tidal flat deposits (Figs. 12, 13, 14c). Thickness of individual sand bodies is less than 10 feet. Phosphatic fish remains in the Gilwood distributary mouth bars are rare. The Gilwood distributary sandstones are submature, and poorly to moderately sorted. The poor sorting probably reflects the ephemeral character of streams and the high rate of accumulation during flash floods when river load was deposited beyond the river mouth. Active reworking was limited to the separation of fines and reworking of the upper parts of the bars. The distributary mouth bars appear to have been associated laterally with beaches and tidal flats. As the distributaries prograded seaward, the bar advanced over littoral deposits (Fig. 12) similar to recent deposition in the northwest part of the Gulf of California (Thompson, 1968).

Beaches

Beaches can be defined as the comparatively narrow zones along the coast which lie above the low-water mark. They are formed along shores of seas, bays, lagoons, and lakes. Owing to the sorting effect of the waves, most beach deposits are well sorted and are composed of fine-grained to gravelly and bouldery material. Most beach structures

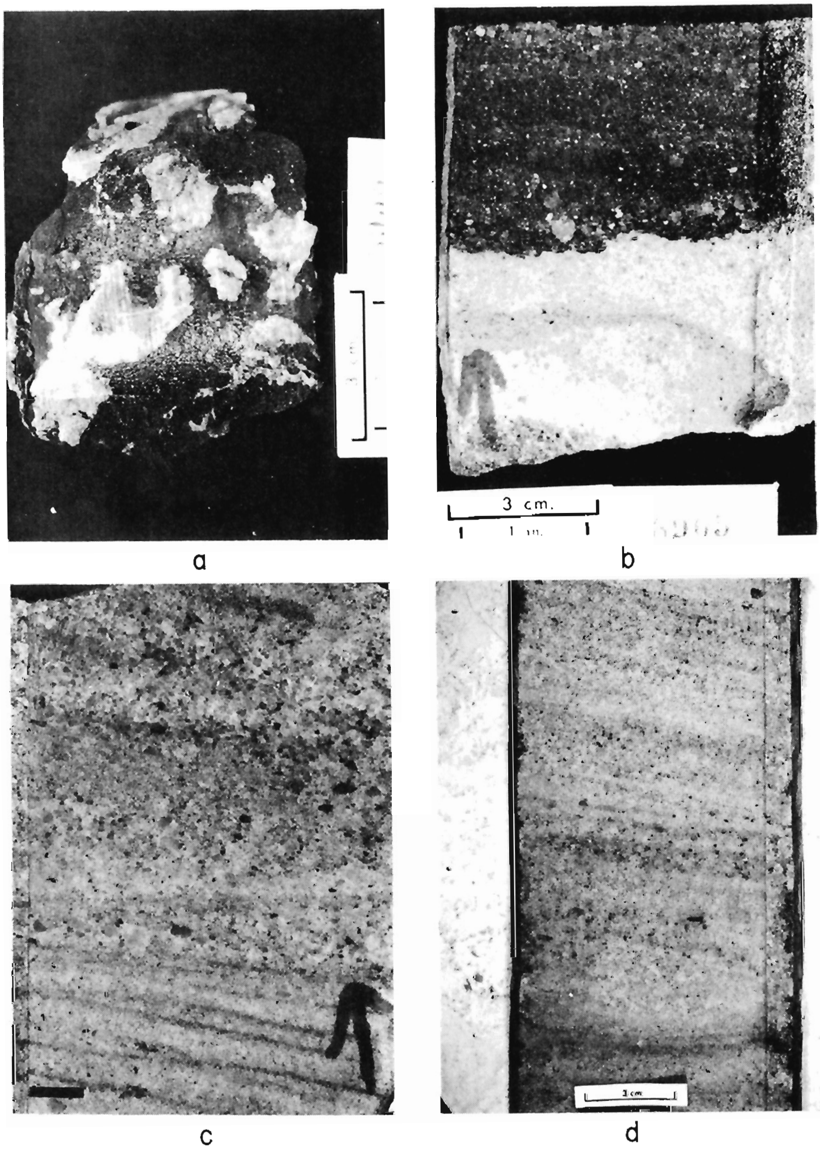


FIGURE 14

a. Irregularly shaped limestone nodules in red silty shale; Gilwood Member. Core photograph, 10-21-70-18 W5, depth 8,958 ft (approx. 2,730 m).

b. Quartzitic sandstone, coarse grained, poorly sorted and rounded, crossbedded; upper part of core red, lower part light grey; Gilwood Member. Core photograph, 10-21-70-18 W5, depth 8,965 ft (approx. 2,730 m).

c. Fine-grained feldspathic sandstone with low-angle convex crossbedding (ripple drift cross-lamination), overlain by crossbedded granular sandstone; Gilwood Member. Core photograph, 10-13-68-16 W5, depth 4,223 ft (approx. 1,287 m). Bar scale is equal to 1 cm.

d. Protoquartzitic sandstone, medium grained, quartz grains medium rounded and sorted, cemented by anhydrite, silica, and dolomite; note low-angle multidirectional crossbedding; Gilwood Member. Core photograph, 8-26-69-13 W5, depth 8,489 ft (approx. 2,588 m).

are primarily of a parallel or laminated type, with laminae dipping slightly toward the sea at the upper shoreface (Reineck, 1967; Thompson, 1937; van Straaten, 1959). The angle of dip is directly related to the slope of the foreshore surface, which in turn is largely dependent upon the mean size of the beach sediment (Andrews and Van der Lingen, 1969). Gilwood sediments interpreted as upper shoreface beach deposits contain all of the features mentioned above. Gilwood upper shoreface sandstones are well sorted ($\sigma=0.35$), mean size 0.15 to 0.17 mm, and grains are well rounded ($R=0.7$). Stratification is parallel, and beds are internally laminated. Parallel laminated sandstones with zones of low-angle cross-lamination and laminae of shales (Fig. 7b) probably represent the backshore zone of beaches. Low-angle cross-lamination, with angle of crossbedding averaging 7 degrees, and small-scale multidirectional trough bedding, with vertical burrows also are apparent (Fig. 15a, b, c) and represent a middle shoreface environment. Seaward, beach deposits grade into highly bioturbated sandstones of the lower shoreface (Fig. 15d).

The thickness of the Gilwood Member beach deposits commonly is less than 2 feet, but rarely may reach 5 feet. These deposits can be recognized in the area east of Snipe Lake and to the west side of Swan Hills (Fig. 13).

Gilwood Member beach sediments are embedded in mixed tidal flats sequences. The lateral extension of prograding beaches appears to extend to a maximum of 6 miles perpendicular to the direction of progradation. Some of these beaches could be lagoonal beaches similar to those on the recent East Australian coast (Hails and Hoyt, 1968, p. 38).

Marginal Tidal Flats

Present-day marginal clastic tidal flats consist of widespread accumulations of mud, silt, and in places sand. Finer grained sediments accumulate on the tidal flat surface close to high-water level; whereas, the coarser grained, sandy sediments lie at low-water level (Reineck, 1967, p. 193). This process is opposite to the sorting in a shallow sea. Areas of fine-grained mudflat deposition fringe the shorelines in regions marginal to the delta proper (Coleman and Gagliano, 1965). Muds also accumulate as extensive marginal interchannel tidal plains, and are reworked and redistributed by tidal and longshore currents.

Sedimentary features are dependent on the type of sediment, the physical processes and their intensity, and the rate of sedimentation. There is a broad variety of sedimentary features as summarized by van Straaten (1959), Reineck (1967), and recently by De Raaf and Boersma (1971). Horizontal, ripple, flaser and lenticular lamination, rhythmically graded lamination and bedding, mudcracks, and zones of active burrowing in the intertidal zone have been observed on recent tidal flats. In tidal channels shell or clay pebble deposits can accumulate as lag deposits.

The Gilwood Member tidal flat deposits are composed of silty shales with horizontally laminated ripple-bedded thin sandstone interbeds that have rare flaser bedding (Fig. 16a). Positive grading in the laminae and organic burrowing are common. Trough bedded sandstones which are embedded in bioturbated dark grey shales appear to be tidal channel fill (Fig. 16b).

The sandy bottom of the intertidal zone in places provided favourable conditions for algal growth. Dendroid stromatolites up to a foot high were found in the tidal zone of the Gilwood Member (Figs. 13, 14c). Recent stromatolites comparable to these Devonian types are present in hypersaline pools of Shark Bay, Western Australia,

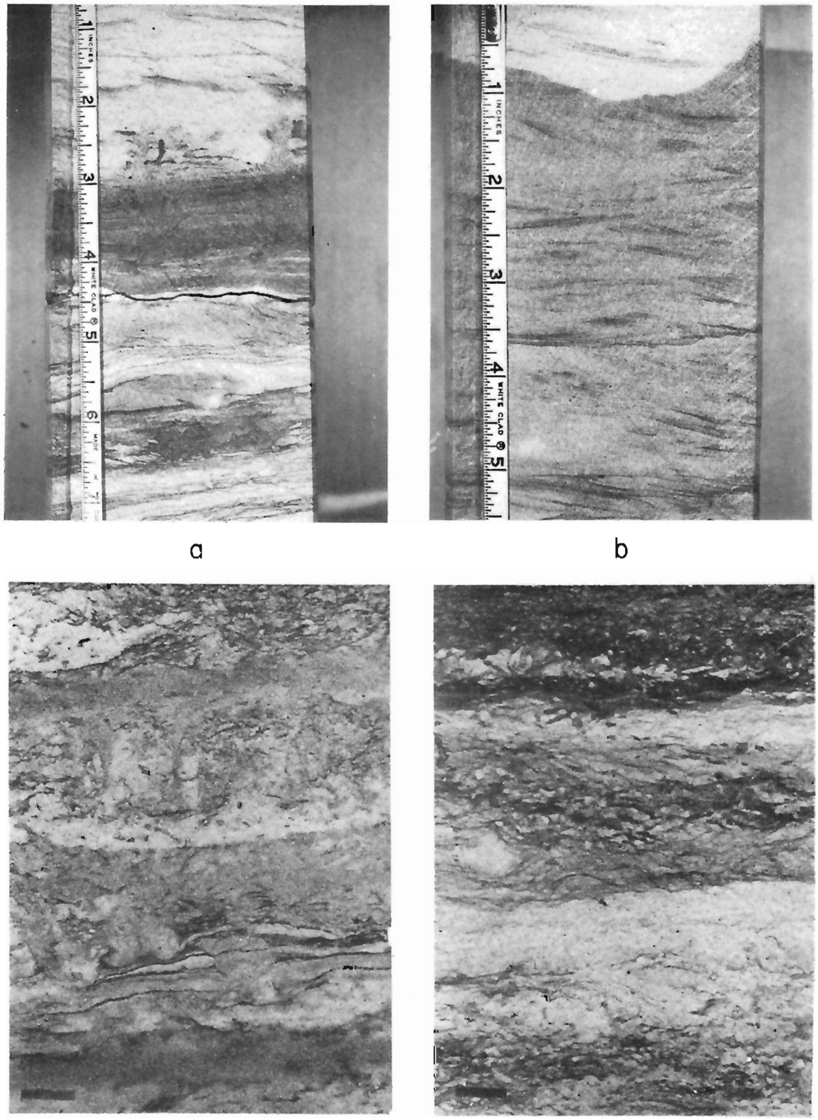


FIGURE 15

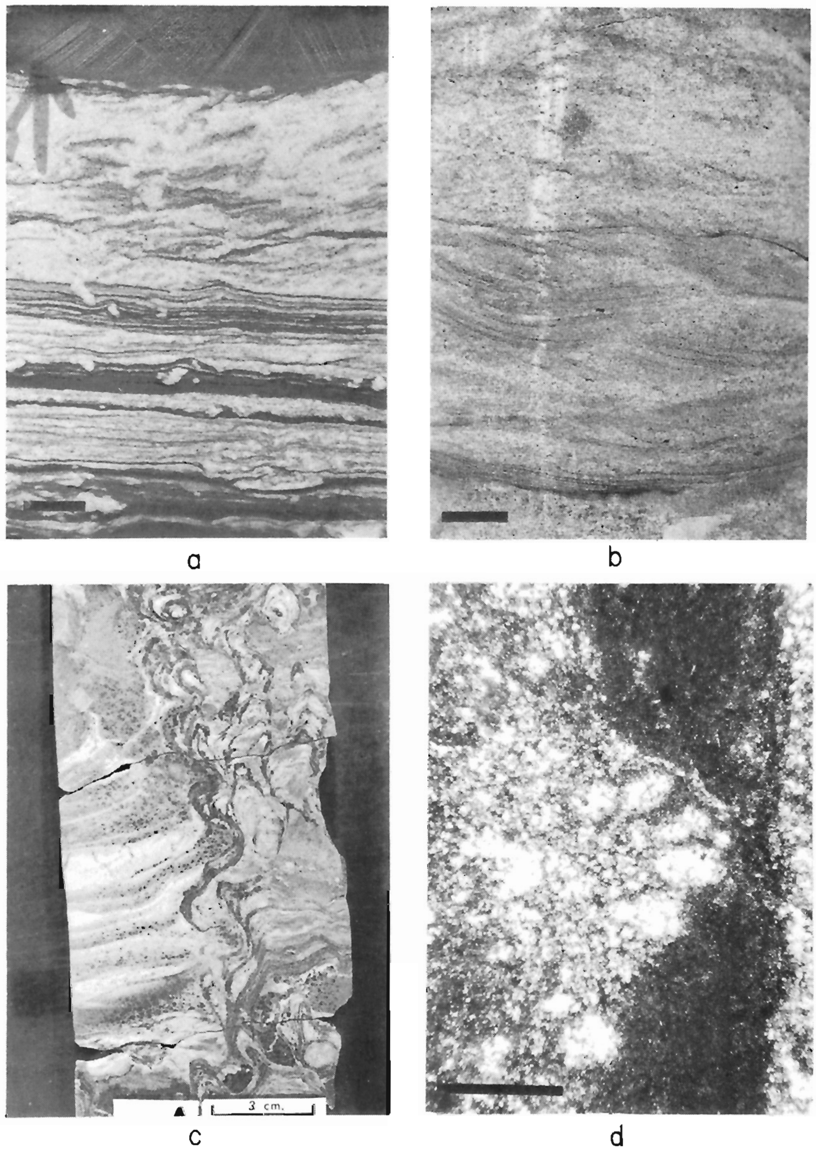
a. Glauconitic protoquartzite, with thin beds and laminae of shaly siltstone, and trails of vertically burrowing organisms; lamination in the mudstone is irregular, horizontal and at lower part of core small-scale crossbedding is apparent; quartz grains well sorted, subrounded; Gilwood Member. Polished core slab, 11-22-70-9 W5, depth 6,819 ft (approx. 2,080 m).

b. Glauconitic protoquartzite, fine grained; small-scale multidirectional trough bedding; Gilwood Member. Polished core slab, 11-22-70-9 W5, depth 6,820 ft (approx. 2,080 m).

c. Well-sorted, fine-grained sandstone with thin shaly laminae; lamination distorted by vertically burrowing organisms of *Scolithus* type (sandstone overlain by algal stromatolite); Gilwood Member. Polished core slab, 2-5-69-10 W5, depth 8,226 ft (approx. 2,500 m). Bar scale is equal to 1 cm.

d. Dark grey shale, grading upward into fine-grained sandstone, reworked by horizontally burrowing organisms; Gilwood Member. Polished core slab, 2-6-70-10 W5, depth 7,650 ft (approx. 2,340 m). Bar scale is equal to 1 cm.

FIGURE 16



a. Quartzitic siltstone with laminae and beds of dark grey shale; bedding horizontal; burrow trails in shaly beds filled with siltstone; Gilwood Member. Polished core slab, 2–6–70–10 W5, depth 7,648 ft (approx. 2,340 m). Bar scale is equal to 1 cm.

b. Fine-grained sandstone with small-scale trough bedding embedded in bioturbated dark grey shale containing linguloid brachiopods; Gilwood Member. Polished core slab, 2–5–69–10 W5, depth 8,222 ft (approx. 2,500 m). Bar scale is equal to 1 cm.

c. Dendroid stromatolite in fine-grained dolomitic sandstone; Gilwood Member. Polished core slab, 2–5–69–10 W5, depth 8,225 ft (approx. 2,500 m).

d. Detail of sporangial(?) chamber in outer wall of stromatolite; Gilwood Member. Thin section—ordinary light. 2–5–69–10 W5, depth 8,226 ft (approx. 2,500 m). Bar scale is equal to 0.5 mm.

where they grow in the intertidal zone. A similar hypersaline environment for the Gilwood tidal pools is suggested by the occurrence of microcrystalline dolomites associated with stromatolites.

The maximum thickness of Gilwood tidal flat deposits appears to be about 10 feet. The Gilwood tidal flats appear to have prograded over a distance of about 40 miles (64 km); they grade seaward into interdistributary bays and lagoons (Fig. 13). Tidal flat deposits were found as far west as east of Snipe Lake; they extend eastward to Mitsue.

Interdistributary Bays and Lagoons

Interdistributary bays and lagoons include areas of semirestricted shallow basins that are partly open to the sea, or connected with the sea by tidal channels or inlets (Coleman and Gagliano, 1965). Many such areas developed behind marginal barrier islands and are protected from the action of waves. In such protected areas, deposits consist mainly of mud with some sand. These basins are seldom more than a few feet deep and have the most diverse faunal assemblage in a delta system (Scott and Fisher, 1969).

Interdistributary bay deposits in the Gilwood Member are composed of medium to dark grey shales and rarely green shales that are silty in part. The shales are generally structureless, but in places have very fine, poorly preserved horizontal laminations. The Gilwood sediments are highly fossiliferous (Fig. 7d), with very common conchostracan, fish remains, and linguloid brachiopods (Norris, 1970).

The maximum thickness of Gilwood interdistributary bay and lagoon sediments is approximately 17 feet (5.1 m); these sediments appear to have accumulated near the lee side of a marginal sand barrier (Figs. 12, 13). In a landward direction, toward the Peace River Arch, the bay sediments thin to an average thickness of 2 feet and overlie tidal flat deposits. Lagoonal deposits also occur near the base of the Gilwood clastic unit and attain a maximum thickness of 4 feet. Depths of lagoons and interdistributary bays can be estimated on the basis of lateral facies relations and appear to have been from 5 to 8 feet at the beginning of deltaic sedimentation, increasing to 18 feet near the end of deltaic deposition.

Interdistributary bay deposits apparently have a wide areal distribution, and interfinger with marginal sand barriers in a seaward direction. Toward land (toward the Peace River Arch) the bay deposits terminate in intertidal mudflats.

Marsh or swamp environments that commonly consist of peats and peaty clays were not recognized in the Gilwood Member. Their absence could be explained by a high degree of oxidation that destroyed organic remains, or by a semiarid climate. A combination of both factors is possible in the Sturgeon–Mitsue area.

Barrier Island

A barrier island is a shallow, submerged, elongate sand body bordering many coastal lands, and in most places flanked on both sides by muddy sediments. Barriers are common around deltas, particularly along those parts that are not actively prograding. Wherever delta plains are slowly sinking, barriers will form at the margin and at times may grow upward, keeping pace with submergence (Shepard, 1960). Although modern barrier islands are easily recognized, ancient ones can be difficult to identify because of insufficient structural, textural, and compositional evidence.

Two main types of significantly different barrier islands can be recognized in the ancient record: (1) beach-barrier, lagoon (bay) system, and (2) barrier-tidal channel, lagoon (bay) system.

Beach-barrier type is well known from a detailed sedimentological investigation by Bernard *et al.* (1962) at Galveston Island. A beach-barrier system is built by lateral rather than vertical accretion and progradation of beach deposits, commonly in the form of beach spits. Berg and Davies (1968) recognized ancient examples of a beach-barrier type in the Lower Cretaceous Muddy Formation in Montana.

Shepard and Moore (1955) noted the presence and significance of inlet deposits in barrier island formations and the difficulty in recognizing them in older sediments. Hoyt and Henry (1967) used Sapelo Island as an example, and presented a lithological model of a tidal-influenced barrier island. Both Shepard and Moore and Hoyt and Henry recognized a positive grain-size gradation, with the coarse-grained crossbedded sediments at the base, and silty clays and clays at the top of the sand barrier. The resulting sedimentary features are similar to those of fluvial deposition but they differ in composition, fabric, and by distinct distribution of sedimentary structures.

Barrier island tidal channel–bay deposits, the second type noted above, can be recognized in the Gilwood Member and are part of the coastal plain delta system. Gilwood Member barrier island deposits form an elongate body 48 miles long and 12 miles wide with a maximum thickness of 54 feet. On the landward (western) side, the barrier sediments interfinger with bay sediments, and to seaward grade into prodelta and open-shelf marine sediments (Figs. 12, 13). The larger part (70 per cent) of the barrier consists of clean, moderately to well-sorted, gravelly to fine-grained sandstone. Sorting of the sandstones of the barrier is poorer than sorting of the beach sediments, and roundness is less pronounced (Fig. 8).

The Gilwood barrier island sediments are made up of several sandstone units, which are interbedded with silty shales, and probably represent a composite body, formed by several overlapping stages of sand barrier development. The characteristic feature of individual sandstone units is that each becomes finer from bottom to top (Fig. 12). At the base of the sequence, there is a sharp contact (scoured surface?) which is overlain by poorly sorted gravelly sandstone or coarse-grained sandstone with high-angle foreset beds and trough bedding (Fig. 17b). Coarse fragments of fish bones are common in this zone (Fig. 17a). This part of the sequence was either deposited within migrating tidal channels, or represents the tidal delta sequence which reworked and redeposited older fluviomarine sediments during time of sea level standstill or during slow transgression. The coarse-grained sandstone zone grades into fine to very fine grained sandstone, moderately to well sorted and rounded, with small-scale trough bedding, cross-lamination, multidirectional cross-lamination and rarely with zones of abundant vertical borings and heavy mineral laminations. Horizontal bedding and ripple-marks (Fig. 17c) are rare. These sediments indicate deposition in the middle to the upper shoreface environment, and in part could represent washovers. The sequence is terminated upward by silty shales deposited from suspension in well oxygenated brackish barrier-back ponds. Silty shales are dark grey, greenish or reddish and have both gradational and sharp contacts with the underlying sandstone.

The dark grey fossiliferous shales, which are found on the landward side of the barrier island, are lagoonal deposits embedded in the barrier sequence as result of washovers and landward migration of the barrier island complex. In an offshore direction

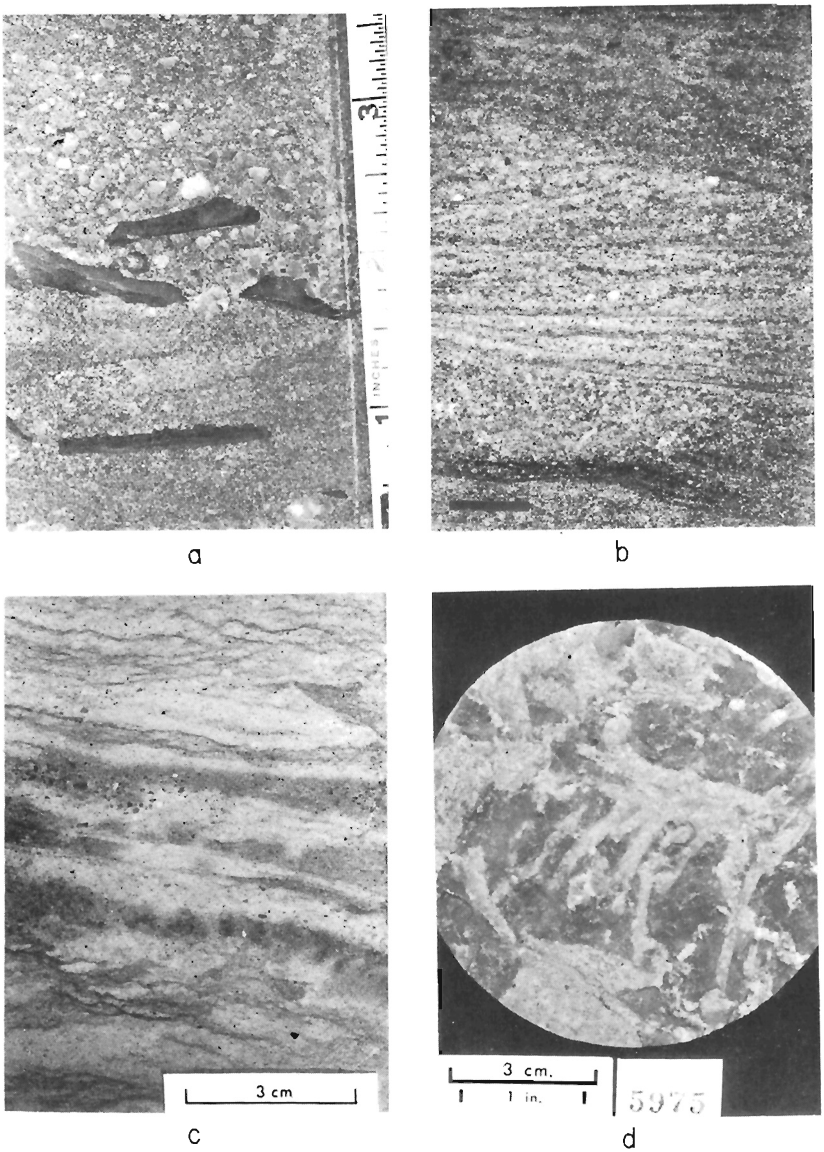


FIGURE 17

a. Poorly visible low-angle, multidirectional trough bedding in coarse-grained sandstone; graded bed of conglomerate encloses reworked, worn debris of phosphatic fish bones; Gilwood Member. Core photograph, 4–28–71- 4 W5, depth 6,017 ft (approx. 1,835 m).

b. Multidirectional trough bedding in poorly sorted, coarse-grained sandstone; Gilwood Member. Core photograph, 4–28–71–4 W5, depth 6,025 ft (approx. 1,835 m). Bar scale is equal to 1 cm.

c. Moderately sorted fine-grained sandstone with pockets of coarse-grained sandstone; irregular wavy lamination is outlined by green shale laminae; Gilwood Member. Core photograph, 10–3–71–4 W5, depth 5,983 ft (approx. 1,820 m).

d. Planar view of unidentified trace fossil in green shale interbedded with well-sorted, fine-grained sandstone; Gilwood Member. Core photograph, 10-35–71–5 W5, depth 5,975 ft (approx. 1,820 m).

the barrier deposits grade into bioturbated and mottled greenish and dark grey silty shales (Fig. 17d). The Gilwood barrier island can be in part compared with the recent sediments described by Kraft (1971, p. 2153) as a transgressive sequence of a baymouth barrier on the Delaware coast.

Where recent barrier flats are a result of hurricane washovers, for example, barrier islands of the central Texas Coast, the flats are composed exclusively of sand. A high percentage of fine-grained sediment in the Gilwood barrier flat sediments suggests that at the time of barrier island formation hurricanes did not have a strong influence on sedimentation.

Growth of the Gilwood barrier island appears to have been strongly influenced by longshore currents. There is a much greater lithological similarity in the direction parallel to the elongation of the barrier than perpendicular to the long axis of the barrier; also, the differences in the feldspar ratio between the sandstones of the barrier and landward delta facies are quite prominent. For a distance of about 7 miles (11 km) along the strike of the Gilwood barrier no significant lithological changes were noted. Perpendicular to the strike of the Gilwood barrier, however, over a distance of approximately 1.5 miles (2.4 km), significant changes were noticeable (Fig. 13). Overall sediment distribution suggests the direction of longshore currents from NNW to SSE during the deposition of the Gilwood barrier. Presumably, the barrier island developed from the submarine bar on which a beach-barrier was subsequently formed, although formation of the barrier island by spit growth and breaching cannot be ruled out.

Prodelta and Open Shelf

The sediments of the prodelta and open shelf sedimentary province were not studied in detail because of a lack of core material. Interpretation concerning this province is based on mechanical log studies which indicate that the Gilwood Member eastward from the Mitsue area is composed predominantly of shales. Shales are greenish grey, noncalcareous, as indicated by a conventional core from 6 miles northeastward of the Gilwood sand barrier (Fig. 6, loc. 15), where shale overlies a thin subaqueous sand-dune sequence. In the delta front areas where siltstone beds are present in a shaly sequence, according to the mechanical logs, the depositional environment is interpreted as a prodelta.

Evolution of the Delta and Surrounding Environments

Gilwood sedimentation in the Sturgeon–Mitsue area began by the development of extensive muddy tidal flats and offshore bars composed of glauconitic siltstone, as a response to the clastic sedimentation which started earlier to the north (Nipisi area). A shallow lagoon developed beyond the bar and simultaneously the first fluvial clastics were deposited west of Snipe Lake on the edge of the tidal flats. The river mouth bar facies prograded eastward over tidal flats which migrated eastward toward the barrier island (Fig. 13). During this progradation the main course developed two distributaries toward the south. Thus, the sedimentation pattern developed its final lobate form (Fig. 6). The main distributary channel probably was connected with a tidal inlet to the east. The prograding delta built a broad, shallow platform, and during a time of standstill of sea level the lagoon, on the landward side of the barrier, was nearly filled by prograding clastics. Following the time of standstill, sea level began to rise either by gradual sea level fluctuation or by emergence of the whole area; or by a drop in the supply of terrigenous clastics. As sea level rose destruction and winnowing of the previously deposited delta sediments by marine processes took place. Distributary mouth bars and beaches that had been drowned were eroded and the sand was redistributed by wave and current action. Part of the sandy sediment had been deposited along side and over the previously formed sandbar and the building of the offshore sand barrier continued. The sediment supply for the offshore barrier was in part transported by longshore currents from the Nipisi area to the north and west.

The reworking action of waves and longshore currents redistributed the delta platform deposits into 2- to 3-mile-wide sand belts oriented oblique to the shoreline These geomorphological low-relief, ridge-type of features can be compared in part with sand ridges of the present day on the wave dominated Rhone River delta (Kruit, 1955). As sea level slowly rose, formation of the barrier was able to keep pace with subsidence of the abandoned distributaries. As a result of the growth of the barrier, a protected bay was established over the deltaic platform.

After building of the barrier ceased, the bay subsequently was filled with clay and silt-sized sediments. Finally, the whole Sturgeon–Mitsue region became an area of extremely shallow brackish and freshwater ponds with extensive vegetation.

The sediments of this sequence consist of green waxy shales with green algae (charophytes), and are commonly known as the Watt Mountain shale *senu stricto* which has a very large areal distribution making it a useful correlation marker. The thickness of these shales in the studied area ranges from 2 to 4 feet. With the deposition of the green shales, the period of clastic sedimentation in the Sturgeon–Mitsue came to a close. However, pulses of clastic sedimentation continued in the western part of the area during deposition of the subsequent evaporitic unit known as the Fort Vermilion Formation.

The Role of Cementation

Areal distribution of different types of cements is related to environmental and geochemical diagenetic processes. Sandstones deposited on the alluvial plain (Fig. 18) are cemented by calcite with rare occurrences of anhydrite and silica. In the vicinity of distributary mouth bars silica appears to be the main cementing agent with minor amounts of carbonate and anhydrite. In the offshore sand-barrier environment cementation by anhydrite is common, although lesser amounts of carbonates and silica also are present.

Paragenetic order of cementation suggests that the silica was formed first, then carbonates, and finally anhydrite. These cements appear to be of diagenetic origin, with some of the silica being formed very early in the diagenetic process. The first stage probably occurred when lower pH river water came into contact with the saline alkaline waters of the delta front facies. The contact between fresh and marine water probably took place in the vicinity of distributary mouth bars, where presumably the pH gradient was sufficiently high to precipitate silica from solution. Electrolytes could cause silica to be absorbed on the surface of suspended inorganic particles, or facilitate co-precipitation of silica with such inorganic particles. This process causes a significant depletion in dissolved silica and was observed by Bien *et al.* (1958) in Mississippi River waters entering the Gulf of Mexico.

The second diagenetic stage is marked by carbonate precipitation. Carbonates can be deposited from groundwater and migrating interstitial fluids in a slightly alkaline environment.

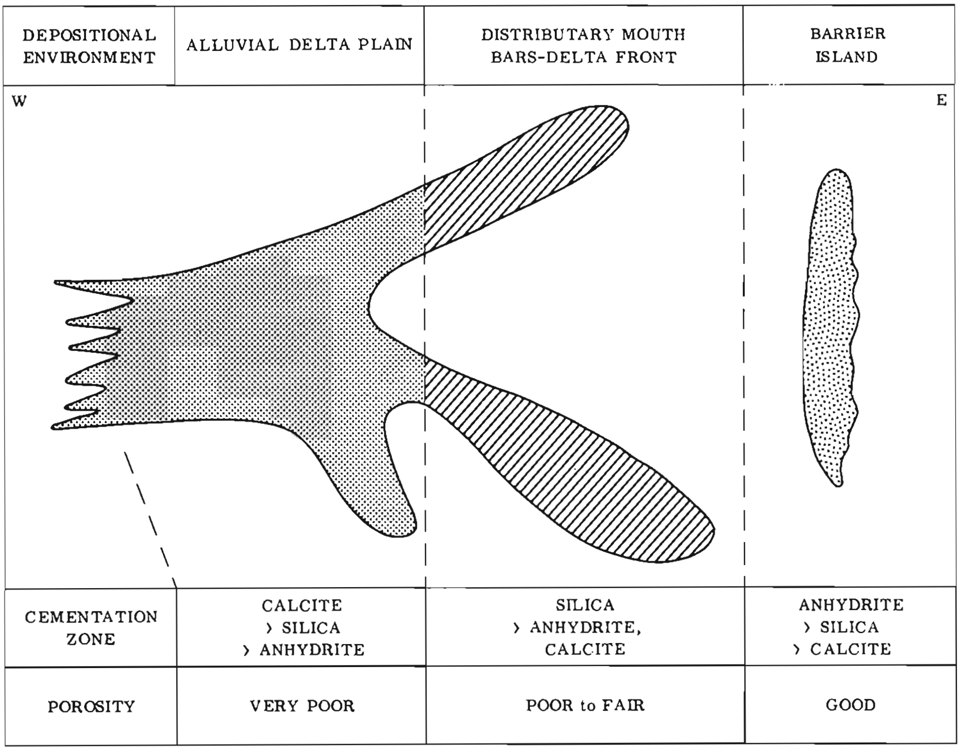


FIGURE 18. Schematic diagram of sandstone cementation in the Gilwood delta complex. Plan view.

Anhydrite appears to be the final cement precipitated and probably originated from the overlying or underlying evaporitic sediments of the Fort Vermilion or Muskeg Formation respectively. Migrating groundwater from the overlying Fort Vermilion Formation during subaerial exposure of the upper surface could have been a source of evaporitic brines. On the other hand the brines could be related to dewatering and compaction processes affecting the underlying Muskeg Formation. Lesser amounts of anhydrite, that which has poorly developed "gypsum rosette" and "sand crystal" form, could be of early diagenetic origin.

The intensity of cementation was found to be lowest in the area of the offshore sand barrier where poorly cemented sandstones are oil reservoirs. The poor cementation could have been related to incomplete primary cementation processes or to secondary solution of primary carbonate cement, or to early oil migration preserving effective porosity.

A zone of silica cementation in the vicinity of the distributary mouth bar deposits could have been a sufficient barrier for the updip migration of oil since no significant oil reservoir has been found to date updip from this zone of silica cementation. Similarly Fisher and Brown (1969, p. 58) found that in the lower Wilcox Formation of Texas (Eocene) there was an almost complete lack of oil in updip distributary channels, suggesting that oil migration was prevented by early cementation.

FORT VERMILION FORMATION

Introduction

Law (1955, p. 1945) introduced the term Fort Vermilion Member of the Slave Point Formation for 23 feet of anhydrite and dolomite that lies at the base of the Slave Point Formation in the California Standard Steen River 2-22-117-5 W6M well. Norris (1963, p. 59) raised the Fort Vermilion to formational rank and used the name Fort Vermilion Formation for a sequence of evaporites underlying the Slave Point Formation near Gypsum Cliffs. Murray (1965) extended the use of the term Fort Vermilion Formation to the Swan Hills area.

The Fort Vermilion Formation consists of a carbonate-evaporite rock sequence, which conformably and gradationally overlies the Watt Mountain Formation and is overlain abruptly by carbonates of the Swan Hills Formation (Fig. 4).

Thickness of this formation is somewhat erratic; it ranges from 38 feet (11.5 m) in the Swan Hills area to 5 feet (1.5 m) at Snipe Lake. To the west, in the Sturgeon Lake area, it decreases to nearly zero. There the unit grades into the terrigenous deposits that surround the Peace River Arch. Eastward from the Swan Hills field, in the Mitsue area, thickness of the formation decreases to approximately 15 feet (4.5 m).

Composition

The Fort Vermilion Formation is composed of thin alternating beds and laminations of pelletoidal limestone, microcrystalline dolomite, calcareous and dolomitic shales, and anhydrite. The anhydrite occurs as nodules, laminae, and beds up to 2 feet thick (Fig. 23b). It is pale grey to brownish grey; whereas, the dolomites and limestones are light grey to dark grey. The shales commonly are dark grey; near the boundary of the underlying Gilwood Member they are greenish grey. Also, near the lower boundary of the Gilwood Member the dolomites are interbedded with beds of sandy dolomite, and with siltstone and sandstone (Figs. 19b, 21c). Distribution of individual Fort Vermilion lithological types as well as biogenic parameters is summarized in Figure 20.

Laminated Pelletoidal Packstones

Laminated pelletoidal packstone is the most common lithotype in the Fort Vermilion Formation. Thickness of individual laminae ranges from 0.8 mm to 1 cm. The lower contact is commonly erosional, with small scours apparent (Fig. 19d). Pelletoidal packstone laminae grade upward into laminae of micrite or microcrystalline dolomite, which is in sharp contact with overlying pelletoidal laminae. Pellets are silt-sized, lithic, and subangular to rounded. Composite grains and superficial oölites are rare. The pelletoidal packstone is cemented by sparry calcite, coarse-grained dolomite,

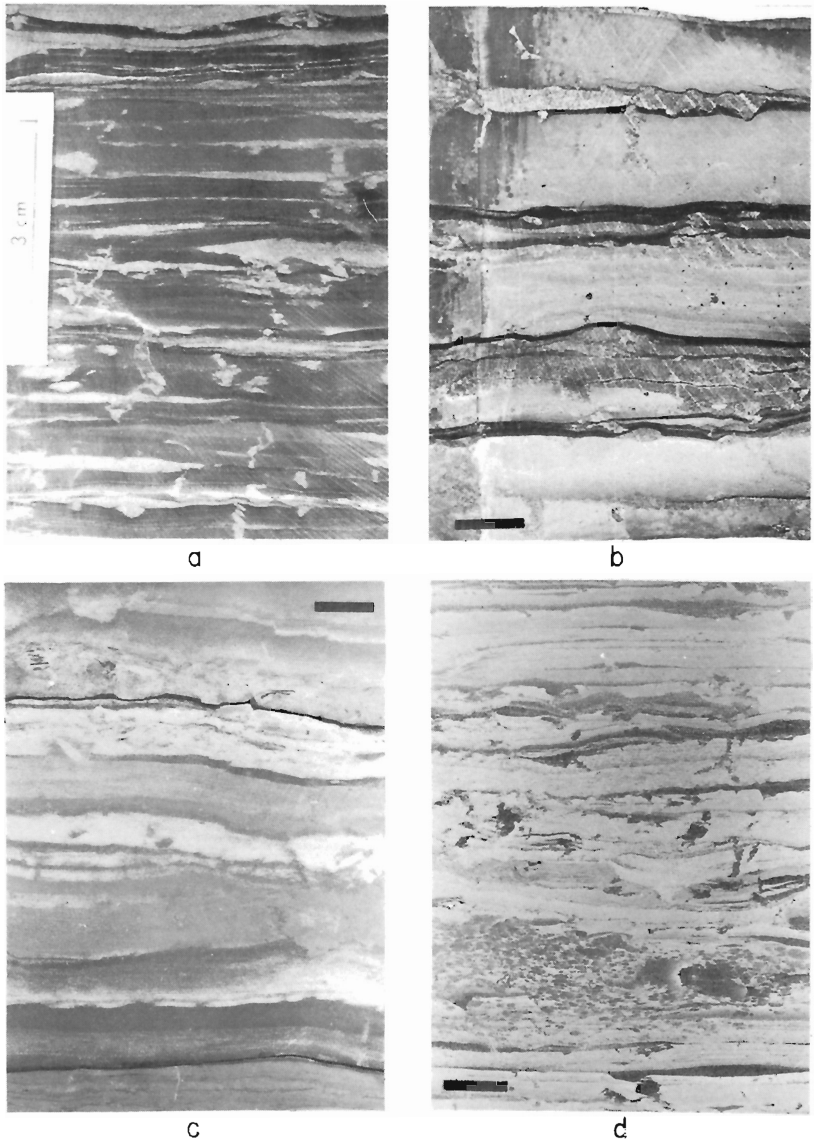


FIGURE 19

a. Flaser bedding; medium to dark grey shale interbedded with thin beds and laminae of fine-grained sandstone; vertical burrows common; Fort Vermilion Formation. Polished core slab, 4–28–71–4 W5, depth 5,992 ft (approx. 1,826 m).

b. Thin-bedded microcrystalline dolomite interbedded with laminae of dark grey shale and thin siltstone beds; vertical burrows common; Fort Vermilion Formation. Polished core slab, 10–21–70–18 W5, depth 8,934 ft (approx. 2,724 m). Bar scale is equal to 1 cm.

c. Graded bedding in microcrystalline dolomite caused by increase of argillaceous matter in the upper part of the individual laminae; note sharp-edged desiccation cracks; (slabbed rock highly etched by HCL and stained by Dickson's (1965) method; argillaceous laminae bluish in colour, with dolomite remaining white); Fort Vermilion Formation. Polished core slab, 2–6–70–10 W5, depth 7,615 ft (approx. 2,340 m). Bar scale is equal to 1 cm.

d. Flaser bedding in the laminated anhydritic dolomite; pellets at base of laminae grading into microcrystalline dolomite at top of laminae; small ripples and burrows at middle of picture; anhydrite of equidimensional nodules; note the similarity with Figure 17a; Fort Vermilion Formation. Polished core slab, 2–6–70–10 W5, depth 7,612 ft (approx. 2,340 m). Bar scale is equal to 1 cm.

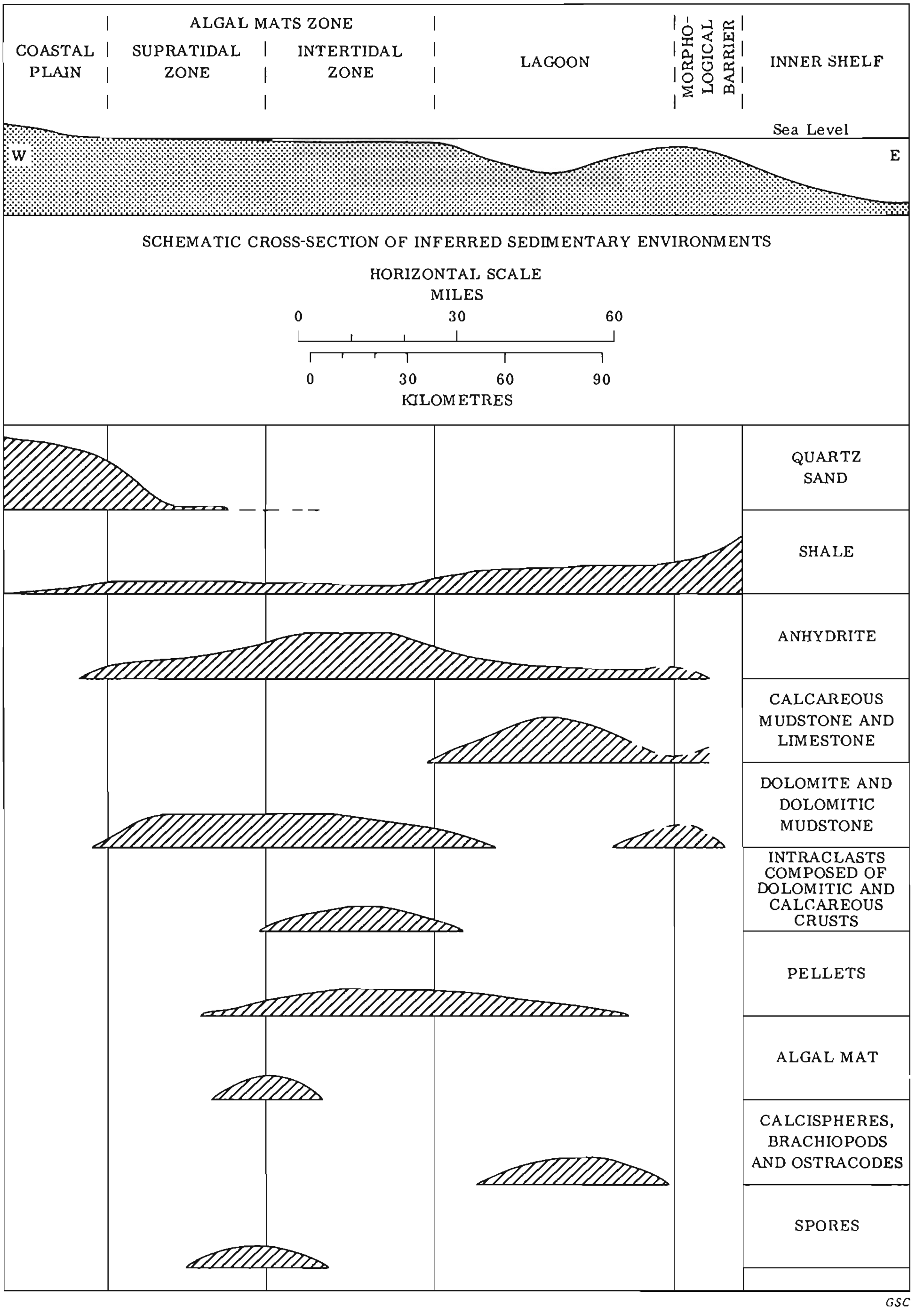


FIGURE 20. Generalized west–east variations in abundance of individual rock types and fossils of the Fort Vermilion Formation; inset above shows schematic cross-section of inferred sedimentary environments.

ferro-dolomite or anhydrite. Fossil fragments include ostracodes, calcispheres, and small gastropods, which in places occur in thick pelletoidal laminae. The laminated pelletoidal packstones as well as the overlying micritic laminae commonly are completely dolomitized, or less commonly, only the micritic laminae are dolomitized, and the pellets have remained undolomitized (Fig. 21b).

Dolomites

Most of the dolomites originated by the secondary alteration of pelletoidal packstones and calcilutites. This is evident in zones of incomplete dolomitization. Coarse euhedral dolomite crystals and small lithoclasts occur at the base of each lamina with an upward gradation into darker coloured microcrystalline dolomite. In the Snipe Lake area, the basal coarse-grained dolomite laminae are partly cemented by anhydrite; dispersed silt and sand-sized quartz grains are present (Fig. 21c). Microcrystalline dolomites also are interbedded with laminae and beds of dark grey shale. Dolomites which can be considered as primary (of early diagenetic origin) display aphanitic texture and are associated with anhydrite.

Laminated Calcilutites

Gradational change in colour outlines lamination in calcilutites of the Fort Vermilion Formation. Generally the lower part of each lamina is light grey and the upper part is dark grey. Changes in colour are influenced by a higher argillaceous and organic matter content in the upper part of the laminae. Thickness of the laminae commonly is less than 4 mm. Calcilutites are often interbedded with dark grey laminae and less commonly with greenish grey shales.

Shales

Shales are calcareous or dolomitic, and dark grey. Clay minerals consist mainly of illite with a small amount of chlorite. Quartz silt-sized grains are rare, and shales have interbedded thin pelletoidal laminae and laminae of microcrystalline dolomite. The occurrence of salt casts, fish scales and the charophyte *Gyrogonites* is rare. Charophytes occur mainly in the greenish grey shale near the boundary with the underlying Watt Mountain shale.

Limestone and Dolomitic Intrarudites

Laminae and thin beds composed of sharp-edged dolomitic intraclasts are very common. Dolomite clasts range in size from an average of 1 mm up to a maximum of 2.5 cm; they are composed of microcrystalline dolomite which, in some places, has vague horizontal laminations (Fig. 22a). Clasts have not been transported far from their original point of deposition, and some are tilted only a few centimetres (Fig. 22b). They are embedded in shale, anhydrite or dolomite. Unsorted accumulations of sharp-edged dolomite clasts are very rare (Fig. 22c); they probably represent a storm deposit.

Laminated limeclasts (Germann, 1969, p. 258), however, are rare in the Fort Vermilion Formation. These irregular limestone pebbles are subrounded to rounded and are composed of calcilutite. The limeclasts range from an average of 1 mm to a maximum of 1.5 cm in diameter. The limestone conglomerates form beds in dark grey shales

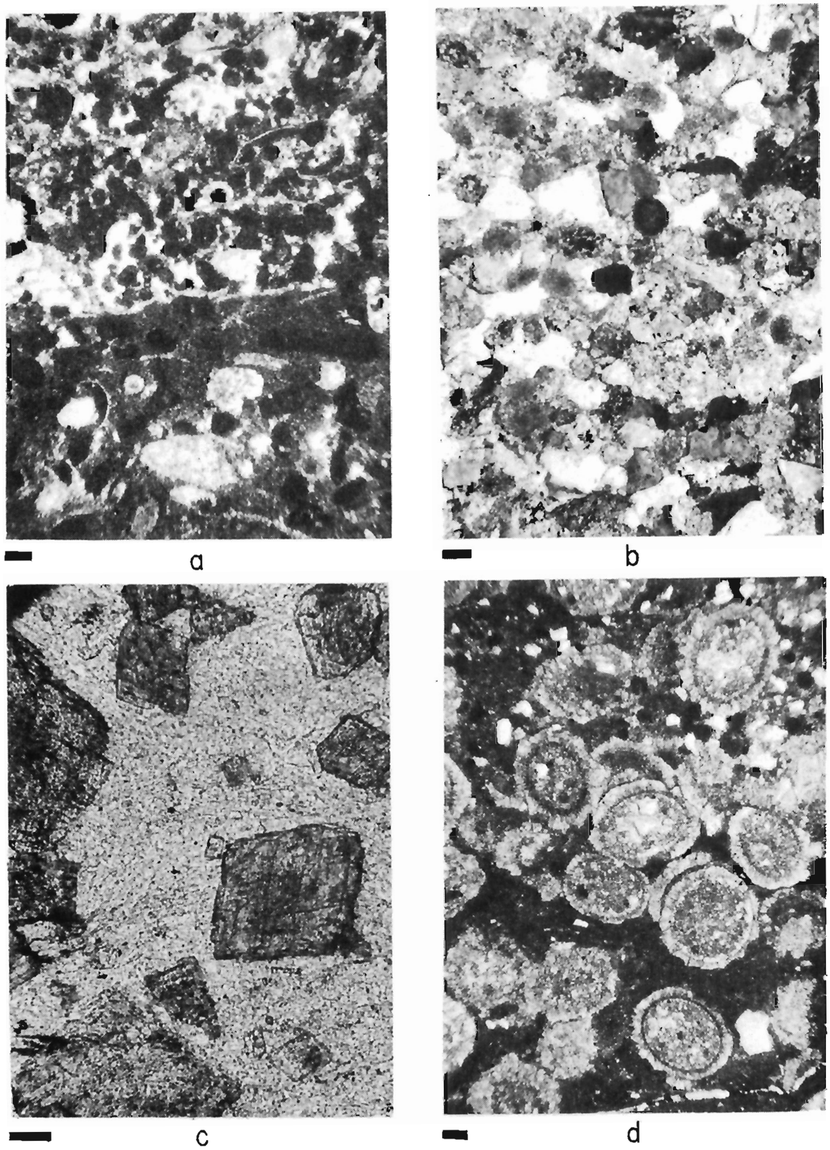


FIGURE 21

a. Laminated pelletoidal packstone; contact of grainstone laminae with underlying micritic laminae is sharp; micrite completely dolomitized, although pellets are composed of calcite; grainstone laminae composed of pellets, thin pelecypod shells, ostracodes and foraminifers cemented by anhydrite and euhedral dolomite; stained with Alizarin Red S; Fort Vermilion Formation. Thin section—ordinary light. 2–6–69–10 W5, depth 8,172 ft (approx. 2,490 m). Bar scale is equal to 0.1 mm.

b. Subangular quartz grains and pellets cemented by anhydrite and ferroan dolomite; Fort Vermilion Formation. Thin section—ordinary light. 4–28–71–4 W5, depth 5,985 ft (approx. 1,821 m). Bar scale is equal to 0.1 mm.

c. Felted anhydrite enclosing zonated euhedral crystals of ferroan dolomite; stained by Dickson's (1965) solution; Fort Vermilion Formation. Thin section—ordinary light. 10–23–71–19 W5, depth 8,785 ft (approx 2,678 m). Bar scale is equal to 0.1 mm.

d. Chara packstone; main components are spherical slightly abraded *Eochara* cf. *E. wickendeni* Choquette; matrix consists of calcareous pellets, small lithoclasts, feldspar, quartz grains, and dolomitic mudstone; note bladed sparry calcite in outer wall of charophytes, and coarse blocky sparry calcite filling internal voids; sparry calcite is in places replaced by fine crystalline subhedral dolomite; stained with Alizarin Red S; Watt Mountain Formation. Thin section—ordinary light. 2–6–70–10 W5, depth 7,622 ft (approx. 2,340 m). Bar scale is equal to 0.1 mm.

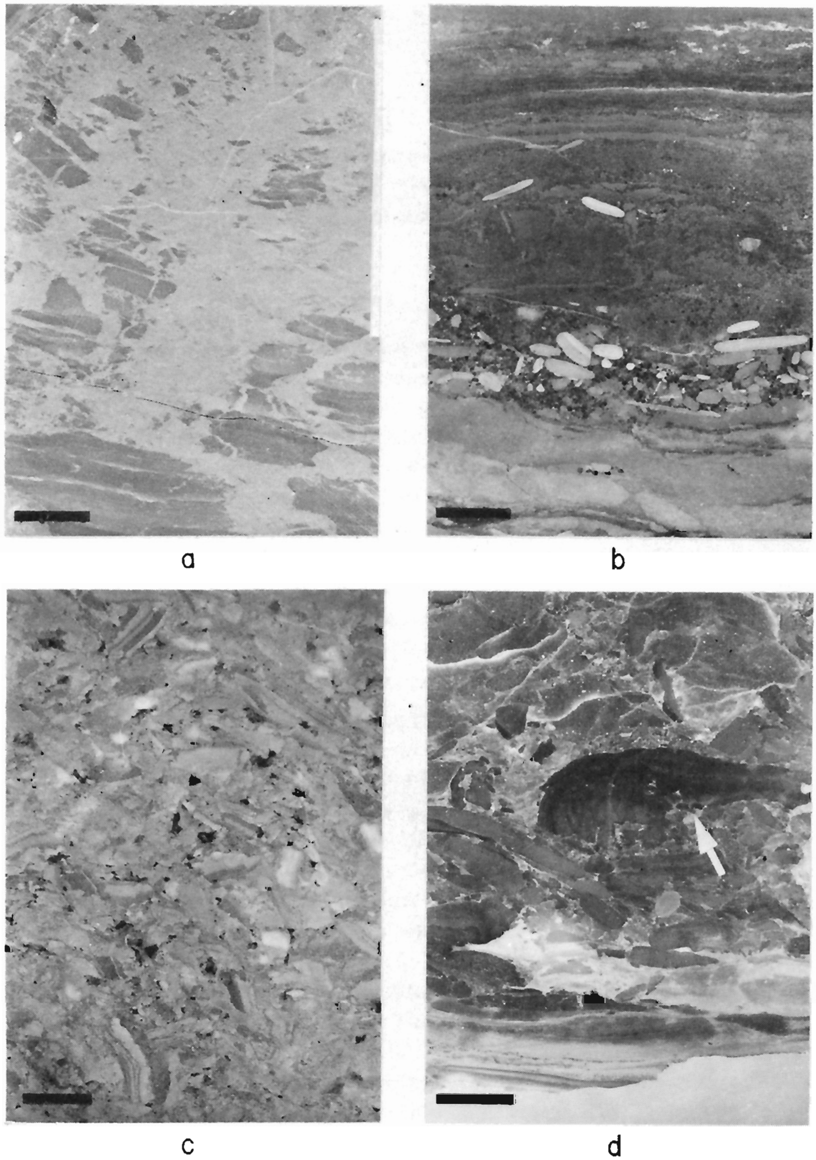


FIGURE 22

a. Sharp-edged lithoclasts of laminated dolomite in a green coloured shale; Fort Vermilion Formation. Polished core slab, 4–28–71–4 W5, depth 5,987 ft (approx. 1,825 m). Bar scale is equal to 1 cm.

b. Flat pebble-conglomerate bed in laminated dolomite; note scour surface; rock slab deeply etched by HC; Fort Vermilion Formation. Polished core slab, 4–22–69–18 W5, depth 9,422.5 ft (approx. 2,870 m). Bar scale is equal to 1 cm.

c. Breccia, composed of angular dolomite clasts; Fort Vermilion Formation. Polished core slab, 2–5–69–10 W5, depth 8,182 ft (approx. 2,500 m). Bar scale is equal to 1 cm.

d. Flat pebble-conglomerate embedded in a graded-bedded dolomite; pebbles up to 3 cm in diameter are elongate, less often equidimensional in shape and consist of poorly laminated unfossiliferous micrite, coated clasts and clasts of blue-green algae (arrow) matrix composed of pellets, quartz, feldspar grains, and micrite cemented by sparry calcite and anhydrite; conglomerate is overlain by laminated pelleted dolomite; Fort Vermilion Formation. Polished core slab, 10–13–68–16 W5, depth 9,183.5 ft (approx. 2,800 m). Bar scale is equal to 1 cm.

and in microcrystalline dolomite. Some of the limeclasts provided the substrate for algal growth. Lithoclasts produced by the erosion of laminar algal mounds rarely occur between pebbles (Fig. 22d). Silt and sand-sized quartz grains are present in the intrarudite matrix.

Anhydrite

Anhydrite occurs in nodular, distorted nodular, distorted bedded mosaic, bedded mosaic, massive, bedded massive, contorted mosaic, brecciated, and single crystal form (Fig. 23). Thickness of the anhydrite nodules ranges from several millimetres to several centimetres; maximum of the anhydrite beds is approximately 2 feet (0.6 m). Most nodules are closely packed, and exhibit mosaic and bedded mosaic structure. In other places the anhydrite layers have a well-developed contorted structure (Fig. 23a). In closely packed laminae the anhydrite nodules are separated by a thin film of brown organic and siliceous material. Anhydrite is interbedded with laminated dolomite (Fig. 23a), dolomitic shale, and pelletoidal limestone. Problematic blue-green algal mounds embedded in anhydrite sequences are rare (Fig. 23c).

Nodular and Bedded Mosaic Anhydrite

Nodular and bedded mosaic anhydrite are the most common forms in the Fort Vermilion Formation. The nodules are clearly visible where they are rimmed by a coat of impurities. Elsewhere the anhydrite rocks have a massive aspect in hand specimens, but microscopic examination reveals the nodular texture. The nodules are composed of felted (Fig. 21c), subfelted, and aligned felted masses of minute anhydrite laths (Maiklem *et al.,* 1969). Length of individual crystals averages 20 microns, but ranges from several microns to several hundred. Nodules frequently have a patchy internal fabric, with patches composed of differently oriented crystals.

Most nodules are flattened (Fig. 23b), with their long axis in the horizontal plane. Equidimensional nodules are less common, and are enclosed in highly organic host sediments (Fig. 23c). Flattened nodule shape was considered by Kinsman (1966) as a growth form, probably influenced by inner growth tension producing a preferred orientation of laths. "Flow" texture is apparent in hand specimens as well as in thin sections, and is not associated with fracturing; only slight bending of individual crystals has been observed. Almost all examined nodules are free of host sediment and display a notable displacement of surrounding sediment. The internal structure of nodules suggests that the process of nodule growth is one of internal expansion, by growth of a new lath inside the system between laths earlier developed, as suggested by Kinsman (1966), and Shearman and Fuller (1969). Nodular anhydrite commonly grades into bedded mosaic anhydrite, which has the same internal fabrics as the nodular anhydrite.

The second type of anhydrite occurring in the Fort Vermilion Formation is coarse lath-shaped crystals that range in size from 5 mm to 1 cm; they fill voids, mudcracks, fractures and interparticle space in intrarudites. Long needle-like crystals of anhydrite oriented parallel with probable algal surface layers are rare. Large anhydrite crystals commonly have a poikiloblastic texture, with the sediment incorporated into the crystals. Host sediment is assimilated to different degrees, and commonly is preserved at least as a ghost structure. This type of crystal growth does not produce displacement of surrounding sediment. Cross-sections of these crystals show pseudocubic outlines, suggesting that the crystals had formed primarily as anhydrite.

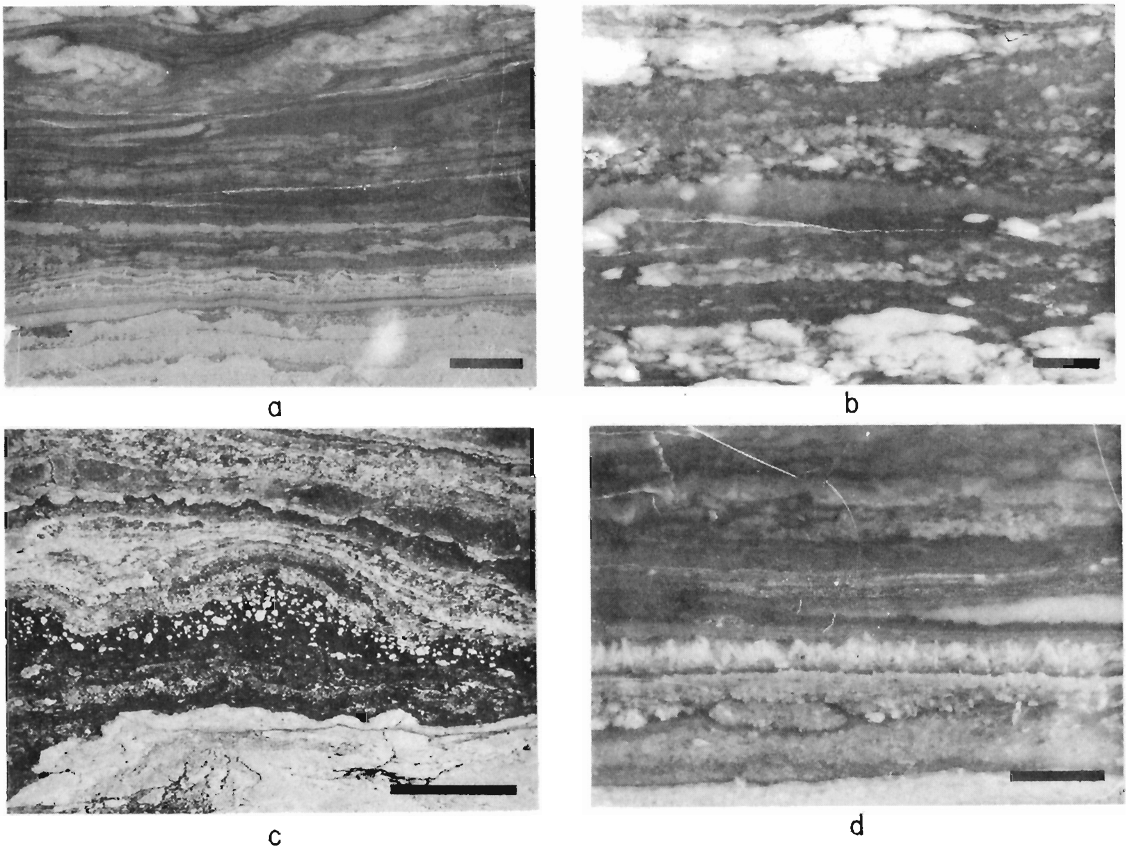


FIGURE 23. a. Laminated dolomite overlain by contorted, bedded anhydrite; each of the carbonate laminae consists of calcitic pellets capped by a layer of a microcrystalline dolomite; small scours at the base of laminae are filled with pellets; Fort Vermilion Formation. Polished core slab, 2–6–70–10 W5, depth 7,612 ft (approx. 2,320 m). Bar scale is equal to 1 cm.

b. Bedded, nodular mosaic anhydrite with patches and laminae of microcrystalline dolomite and micrite pellets; Fort Vermilion Formation. Polished core slab, 4–22–69–18 W5, depth 9,410.5 ft (approx. 2,869 m). Bar scale is equal to 1 cm.

c. Bed of subfelted anhydrite overlain by the laminae of blue-green algal mat; domal-shaped algal mat laminations are rich in organic material; filamentous algal casts (?) poorly preserved; small anhydrite nodules in upper part of algal mat are equidimensional in shape; Fort Vermilion Formation. Thin section—ordinary light. 10–13–68–16 W5, depth 9,185 ft (approx. 2,800 m). Bar scale is equal to 1 cm.

d. Thin bed of anhydrite pseudomorphs after gypsum crust in bedded nodular mosaic anhydrite; microcrystalline dolomite matrix visible as white dots and patches; note discoid and lenticular shape of crystal pseudomorphs; Fort Vermilion Formation. Polished core slab, 2–6–70–10 W5, depth 7,591 ft (approx. 2,314 m). Bar scale is equal to 1 cm.

The third type of anhydrite has a granular texture. It forms cement in pelletoidal packstones and microcrystalline dolomites, filling the interparticle voids.

Origin of the Anhydrite

Although the various habits of recent gypsum recorded from the Trucial Coast, Quatar, and Gulf of California (Kinsman, 1969; Shearman, 1966; Butler, 1970) are distinctive, careful examination of thin sections of the Fort Vermilion anhydrites did not reveal pseudomorphs after gypsum. An exception is the occurrence of laminae formed by vertically oriented crystals, some of which are lenticular and discoid and may represent anhydrite pseudomorphs after gypsum crusts (Fig. 23d).

Lithological evidence suggests the Fort Vermilion Formation evaporite sequence was deposited mainly on broad tidal flats, with maximum concentration of the anhydrite in the intertidal zone (Fig. 20). The nodular anhydrite could have originated under the sediment surface of the intertidal flats as described from present-day coastal sabkhas (Kinsman, 1969). The occurrence of equidimensional nodules of anhydrite supports the idea of early diagenetic origin of anhydrite and also early lithification of sediments. Whether the nodular anhydrite was a primary deposit or originated by gypsum replacement remains problematic. Growth of the nodules probably continued in later diagenetic stages.

Elongate anhydrite crystals with poikilitic fabrics as well as granular anhydrite cement may have formed from percolating groundwater and interstitial brines at any time, but in all likelihood are of late diagenetic origin. An epigenetic origin for some poikilitic anhydrite cannot be ruled out.

Sedimentary Structures

Graded lamination (Fig. 19c), laminae of angular and rounded dolomite and calcilutite intraclasts (Fig. 22b), small scours, erosion surfaces, desiccation cracks, horizontal lamination, low-angle crossbedding, and burrow trails (Figs. 19b, d) are common. There also are rare occurrences of salt casts and "gypsum" crust debris (Fig. 23d); storm breccia is rare (Fig. 22c).

Paleontology

Calcispheres are the most common fossils. There are rare occurrences of *Eochara wickendeni* (Peck and Morales, 1966, p. 309), spores, ostracodes, small gastropods, tufted algal mats and possibly skeletal algae and blue-green algal mats. One sample contains fish scales.

Discussion

Depositional History

The Sturgeon area, which is the most westerly part of the study area, was the site of a broad exposed alluvial plain covered by terrigenous sandy sediments during deposition of the Gilwood Member. During Fort Vermilion time, this was an area of nondeposition, and may have been a deflation region. This zone of terrigenous clastics grades eastward, toward Snipe Lake, into an area that was only partly under the influence of terrigenous sedimentation from the west. Consequently, beds of sandstones and shales are interbedded with dolomites and anhydrites. The anhydrites contain spores, mudcracks, erosion surfaces, dolomitic brecciated crusts, and rare blue-green algal mats. These rocks appear to have accumulated in a coastal zone with associated supratidal flats, and now form a zone about 30 miles wide (Fig. 20).

Eastward from the Snipe Lake area a lower supratidal–intertidal facies about 35 miles (56 km) wide is apparent. It is characterized mainly by the occurrence of laminated, normal graded, pelletoidal packstones, and layers of fine-grained dolomitic intraclasts. Desiccation cracks, and micritic crusts of the vadose zone suggest intermittent subaerial exposure of the surface during deposition of the Fort Vermilion Formation. In this zone anhydrite composes up to 40 per cent of the rock. To the east this intertidal zone extends approximately to the vicinity of the Freeman oil field and to the west side of the Swan Hills field.

The easternmost zone, which is 40 to 50 miles (64 to 80 km) wide, extends from the eastern edge of the intertidal zone (west of the Swan Hills field) to the Mitsue area. This zone is composed of mudstones, micritic argillaceous limestones, dolomites and minor amounts of anhydrite; extends from the intertidal flats (Snipe Lake to Swan Hills) to the site of the Gilwood sand barrier (Mitsue field); and is believed to have accumulated in a lagoon system. Salinity of the water was such that ostracodes and brachiopods could flourish. Anhydrite in the lagoonal zone occurs as void-fillings and probably is of late diagenetic origin. The physiographic setting of the Fort Vermilion Formation compares well with that of the present-day Trucial Coast where, according to Kinsman (1969) and Kendall and Skipwith (1969), the lagoons contain no evaporite-depositing or free-standing brine bodies. The lagoonal sediments grade landward into coastal flats up to 10 km wide, which grade landward into a 100-km-wide belt of continental sabkhas.

The formation of the evaporite sequences of the Fort Vermilion Formation probably was a result of a slight regression of the sea. A complete lack of terrigenous quartz grains in lagoonal deposits and only rare quartz grain occurrences in the intertidal flats suggest a predominant onshore wind direction. Also, a low-rainfall and semiarid climate can be assumed from the evaporites.

Contact of the overlying carbonates of the Swan Hills Formation with Fort Vermilion Formation is sharp, and in places erosion surfaces and fine-grained basal "conglomerate" layers are evident, suggesting that during the final stage of Fort Vermilion deposition there was subaerial exposure in some places. Transformation of most of the gypsum into anhydrite could have occurred at this time, and also much of the diagenetic anhydrite could have been formed.

Paleotopography of the Post Fort Vermilion Surface

Topographic elevations on the Fort Vermilion surface were obtained from the isopach map of the cumulative thickness of the Watt Mountain and Fort Vermilion Formations (Figs. 24, *in pocket*; 25). The top of the Muskeg Formation was used as a datum since it is considered to have been a broad tidal flat of low relief. The relief thus obtained from the cross-section and the isopach map is considered to be a close approximation of the configuration of post Fort Vermilion topography.

The process leading to the formation of relief on the post Fort Vermilion surface can be explained mainly by differential compaction. Sandstones of the distributary belts and sandstone ridges of the Gilwood Member were less compactable than shales filling interdistributary bays and lagoon basins (Figs. 6, 24). This topographic relief

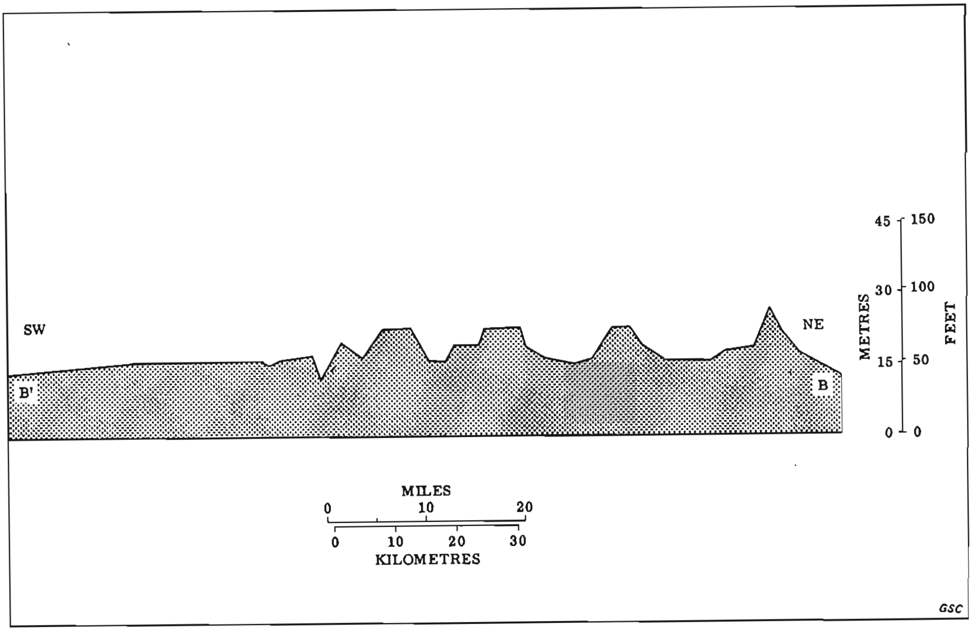


FIGURE 25. Paleo-relief prior to deposition of Swan Hills Formation; cross-section oriented SW–NE; time line top of Muskeg Formation; for orientation of cross-section see Figure 24.

probably was further modified by deflation during a short time of exposure of the Fort Vermilion surface. Consequently, on freshly weathered surfaces, mud particles were more easily removed than sand particles by wind erosion. The relief thus established had an elevation of no more than 30 feet (9 m); however, this was enough to significantly influence carbonate deposition.

SWAN HILLS FORMATION

Introduction

The oil-producing carbonate of the Swan Hills area was first defined as the Swan Hills Member of the Beaverhill Lake Formation by Fong (1959). Later, Leavitt and Fischbuch (1968, p. 294) raised the Swan Hills to formational rank. Fischbuch (1968, p. 452) separated the Swan Hills Formation into nine informal units, and designated these laterally persistent units from bottom to top as Divisions I to IX. The part considered in detail in this study is the lower part of the Swan Hills Formation, or Divisions I to IV, termed here as the marginal reef-bank complex.

The basal part of the Swan Hills Formation, which includes all of Fong's (1959) dark brown unit and part of the overlying light brown unit, can be correlated with the Slave Point Formation of northern Alberta, northeastern British Columbia, and the southern part of the Northwest Territories.

These correlations coupled with the paleontological evidence of McGill (1963, 1966), Fischbuch (1968), and W. K. Braun (pers. com.) indicate a Middle Devonian age for the lower part of the Swan Hills Formation. The upper part of the formation is considered to be Upper Devonian, and probably is correlative with part of the nodular off-reef shales of the Waterways Formation. Pedder (1971), however, from studies of the coral fauna, suggested a Late Devonian age for the upper part of the Swan Hills marginal reef-bank complexes.

Total thickness of the Swan Hills Formation ranges from 500 feet (150 m) for full carbonate buildups to less than 70 feet (21 m) in off-reef areas. The surrounding and enclosing Waterways shales, consequently, thin perceptibly over full Swan Hills buildups and reach a maximum thickness of 500 feet (150 m) in off-reef sections.

Terminology

Inorganic plateau—A geomorphological term used in this study for any comparatively flat surface plain which stood above the shelf floor prior to organic sedimentation. These elevations are composed of terrigenous clastic deposits, non-skeletal carbonates, and evaporites, and represent a drowned coastal plain inundated by the sea during late Givetian time (end of Fort Vermilion time).

Basal carbonate blanket—A thin veneer of carbonate sediments of skeletal and non-skeletal composition which covers the inorganic plateau in the form of a thin, sheet-like deposit. Near the edges of the Swan Hills inorganic plateau there were open-shelf type sediments deposited, however, bank-type sediments are predominant and consist mainly of pellets, grapestones, and oölites.

Bank—Banks as here defined are flat-topped structures rising above the surrounding terrain. Formation of such carbonate banks is envisaged as a mechanical accumulation of bioclastic debris, where the biota did not have the ecological potential to erect rigid, wave-resistant structures.

Formation of banks is a result of favourable biological and hydrodynamic-mechanical processes. Processes involved in bank formation are: first, differential biological productivity in the area, with the highest biological productivity over the bank shoals, and secondly, biological binding, trapping, or hydrodynamic stabilization of sediments.

Marginal reef-bank complex—The marginal reef-bank complex consists of (1) marginal reef-banks, and a (2) shelf lagoon (interior platform).

Marginal reef-banks consist of biohermal mounds in the lower part of the Swan Hills Formation and were formed along the outer, or seaward, edge of the carbonate platform to a distance of approximately 110 miles (177 km) from the Peace River Arch. Since they were marginal topographic features along the edge of the platform, they had a restricting effect on the interior parts of the bank or platform and acted as shallow submerged "barriers."

Shelf lagoons (interior platform) were formed over broad areas as a result of the development of bioherms along the outer perimeter of the marginal reef-bank complex. The Sturgeon–Mitsue shelf lagoon carbonates are interbedded with biohermal carbonates near the margins of the reef-bank complex.

Because of the overlapping and interfingering nature of the biohermal and shelf lagoon carbonates the total sequence is referred to as a marginal reef-bank complex.

Stratigraphy

For the purpose of this study the Swan Hills Formation is divided, on the basis of composition, colour, and geometry, into three informal units: a basal carbonate blanket, marginal reef-bank complexes, and solitary reef-bank complexes (Fig. 4).

The basal carbonate blanket, a widespread pre-Swan Hills unit, is a relatively continuous, thin, sheet-like cover over an inorganic plateau. The sediments composing the basal carbonate blanket are wackestones, packstones, and grainstones formed of skeletal grains, pellets, oölites and algal-coated grains.

The marginal reef-bank complexes form the lower part of the Swan Hills Formation (Div. I-IV, Fischbuch, 1968) and consists of marginal reef-banks (morphological highs) developed along the margins of the carbonate platform and morphological lows (shelf lagoons) in the interior of the platform. Thickness of the marginal reef-banks ranges from 80 to 130 feet (24 to 40 m). Areal distribution of the carbonate platform is similar to that of the underlying basal carbonate blanket (Fig. 26, *in pocket*). Rocks of the carbonate platform are mainly dark brown, but light brown sediments are common. Separation on the basis of colour (Fong, 1959) was not found to be a reliable criterion in distinguishing the carbonate platform from solitary reef-banks since dark brown beds are common in the upper part of the Swan Hills Formation.

The solitary reef-bank complexes, termed reef-fringed carbonate banks by Murray (1966, p. 1), constitute the upper part of the Swan Hills Formation (Fig. 4). These complexes developed during a relatively continuous transgression on elevated parts of the underlying marginal reef-bank complex. Thickness of the solitary reef banks (Div. V-IX incl.) ranges from 200 to 250 feet (60 to 75 m). Areal distribution of the solitary reef-bank complexes is much more restricted than that of the underlying marginal reef-bank complexes (Fischbuch, 1968). At the end of Division IX time the rate of submergence exceeded the rate of reef growth, bringing Swan Hills deposition to a close.

Basal Carbonate Blanket

Composition

The basal carbonate blanket is a thin, sheet-like veneer of carbonate sediments that covers large areas of the inorganic plateau (Fig. 27). On the basis of lithological similarities the basal carbonate blanket is correlated with part of the Slave Point Formation (Norris, 1963, p. 61). Thickness of the unit ranges from a zero edge west of Snipe Lake to 13 feet (4 m) at the eastern margin of the inorganic plateau in the area of the Swan Hills field. Because of the uniform, sheet-like distribution of the sediments, the descriptive term "blanket" is appropriately applied. The term "blanket" was similarly used by Logan *et al.* (1969, p. 35) for the description of shelf sediments of Yucatan.

Contact of the blanket deposits with the underlying Fort Vermilion Formation is sharp where biomicrites and skeletal and oölitic calcarenites overlie the anhydritic dolomites of the Fort Vermilion Formation. The occurrence of semilithified crusts, breccias, flat pebble-conglomerates, and desiccation cracks suggests a short period of exposure of the Fort Vermilion Formation prior to the deposition of the basal carbonate blanket. Although, in the "deeper" lagoonal basins sedimentation appears to have continued without interruption. Eastward from the edge of the inorganic plateau, basal carbonate blanket deposits grade into deeper water shaly sediments of the Waterways Formation.

The rocks of the basal carbonate blanket are divided into five lithofacies. The outer, or offshore, edge of the blanket is composed mainly of open-marine micrite and biomicrite; landward, there is a gradation into skeletal, pelletoidal, oölitic, dolomitic, or algal limestones (Fig. 27, *in pocket*).

Fossiliferous Micrite Lithofacies

This rock type consists of dark grey micrite which in places contains scattered faunal debris (Fig. 28a). The fossiliferous micrite lithofacies occurs near the eastern edge of the inorganic plateau (near the Swan Hills field) and composes approximately 23 per cent of the basal carbonate blanket sediments.

Composition

This lithofacies consists of dark grey micrite that commonly is recrystallized into microsparite, and contains less than 10 per cent skeletal particles. The most common bioclasts are small crinoidal fragments and small gastropods. Articulated and disarticulated ostracode carapaces are common. Pelecypods, brachiopods, calcispheres, foraminifers, and *Chara oogonia* are rare. Large angular fragments of brachiopod and pelecypod shells indicate near in situ accumulation. Nonskeletal particles are limited to rare occurrence of oölites, which probably were retransported from the oölitic shoals. The fossiliferous micrites are in places interlaminated with layers of pellets.

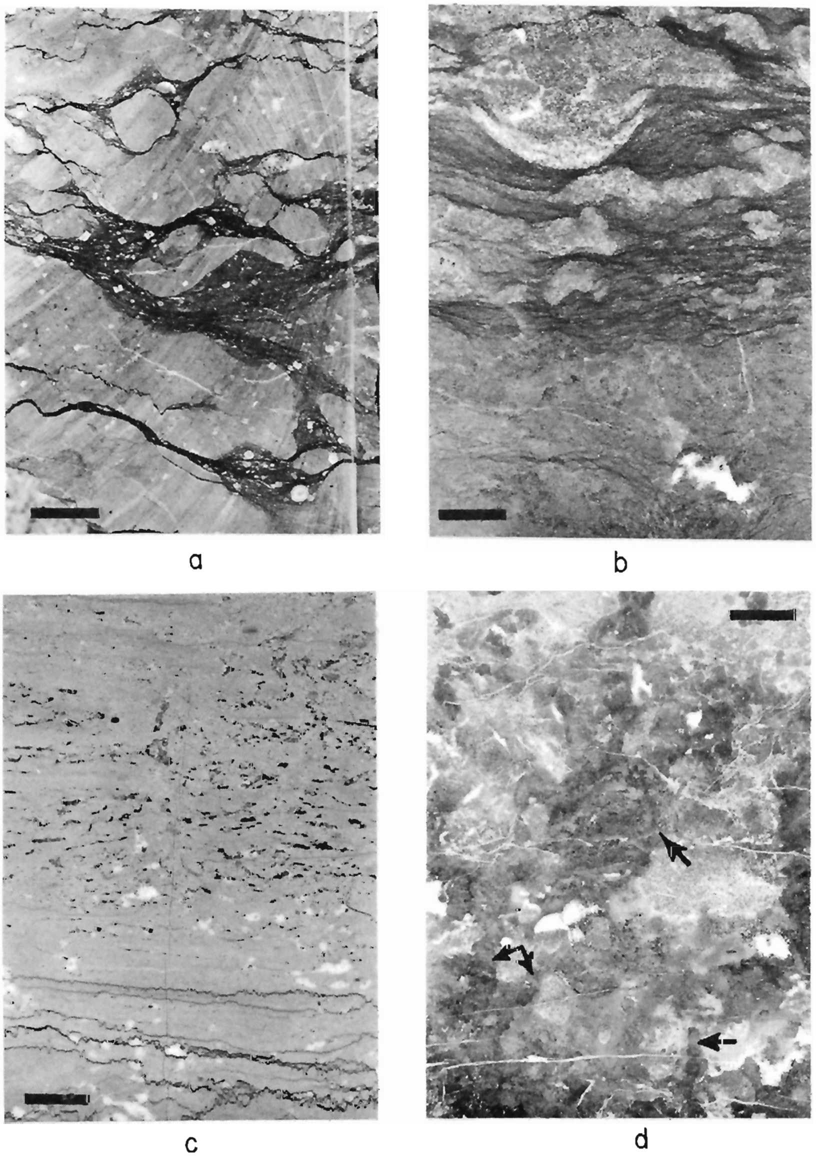


FIGURE 28

a. Slightly nodular fossiliferous micrite with crinoid bioclasts; irregular stylolitic laminae bounded by concentration of organic and argillaceous material; fossiliferous micrite lithofacies, basal carbonate blanket; Swan Hills Formation. Polished core slab, 4–7–70–10 W5, depth 7,717 ft (approx. 2,352 m). Bar scale is equal to 1 cm.

b. Scour in skeletal wackestone filled with well-sorted dasycladacean algae grainstone; grainstone cemented by sparry calcite; skeletal lithofacies, basal carbonate blanket; Swan Hills Formation. Polished core slab, 2–6–70–10 W5, depth 7,576 ft (approx. 2,920 m). Bar scale is equal to 1 cm.

c. Linear fenestral fabrics laterally grading into vertically oriented fenestra (teepee structure) in a poorly laminated micrite; fenestra in upper part of photograph unfilled, in lower part filled with anhydrite; note microstylolites in lower part of photograph; biomicrite lithofacies, basal carbonate blanket; Swan Hills Formation. Polished core slab, 10–23–71–19 W5, depth 9,779 ft (approx. 2,981 m). Bar scale is equal to 1 cm.

d. Digitate stromatolite in dolomitic matrix; note internal lamination of individual dome-shaped, dark grey protuberances and geopetal fabrics; basal carbonate blanket; Swan Hills Formation. Polished core slab, 2–5–69–10 W5, depth 8,158 ft (approx. 2,455 m). Bar scale is equal to 1 cm.

Sedimentary Structures

This lithofacies generally has a massive texture. The sediments commonly are bioturbated, and in places are horizontally laminated or have nodular texture. Post-depositional disturbance, probably due to the burrowing action of scavengers, in places rearranged the organic debris into a spiral pattern and produced local accumulations of skeletal particles. Laminations formed by higher concentrations of argillaceous material also are present. There are rare occurrences of linear fenestral fabric (Fig. 28c).

Environment

Fossiliferous micrites of the basal carbonate blanket were deposited below wave base in a shallow open-marine environment on the slopes of the inorganic plateau. Pelletoidal laminae, rare oölites, and dasycladacean algae suggest proximity to a carbonate shoal. Occurrence of fenestral fabric suggests that some micrites could have been deposited on a tidal flat.

Pelletoidal Lithofacies

The pelletoidal lithofacies is composed of well to poorly laminated, dark grey, fine-grained calcarenites, and composes about 30 per cent of the basal carbonate blanket. This lithofacies occurs generally to the west of the fossiliferous micrite facies. In the inner part of the carbonate platform, however, pelletoidal limestones overlie most of the inorganic plateau from about the Swan Hills field to Snipe Lake (Fig. 27). This lithofacies is most common at the base of the basal carbonate blanket.

Composition

Pellets compose 25 to 60 per cent of the rock and consist of micrite; they are commonly dark grey and ovoid (Fig. 29c). Diameter ranges from 0.06 to 0.15 mm, with an average of 0.12 mm. They generally are devoid of internal structure, but rarely display a vaguely clotted internal fabric and in places enclose spicules. This clotted pellet fabric and the presence of spicules suggest a faecal origin. The homogeneous pellets, on the other hand, could have been formed by mechanical aggregation, by micritization of bioclastic grains, or by disintegration of lithoclasts; also, a faecal origin can not be ruled out. Pellets produced by disintegration of lithoclasts commonly are lighter coloured than those formed by other processes, but because of the difficulty in recognizing individual types, all were tabulated together. Associated with pellets in the pelletoidal lithofacies are skeletal particles consisting mainly of fragmented brachiopods, pelecypods, gastropods, ostracodes, foraminifers, calcispheres, and algae (Fig. 29b). These skeletal grains compose from 3 to 15 per cent of total rock volume. Sparry calcite cement appears as very thin granular crusts less than 10 microns thick enveloping individual particles, with intergranular spaces filled with a coarse-grained drusy mosaic sparry calcite. Up to 25 per cent of the rock is micrite. Dolomite, in places, replaces microsparite. Void-filling anhydrite and silica are rare.

Structures

The pelletoidal lithofacies commonly is laminated, with laminae from 1 to 10 mm thick. The lamination is horizontal, and the basal coarser grained laminae have sharp,

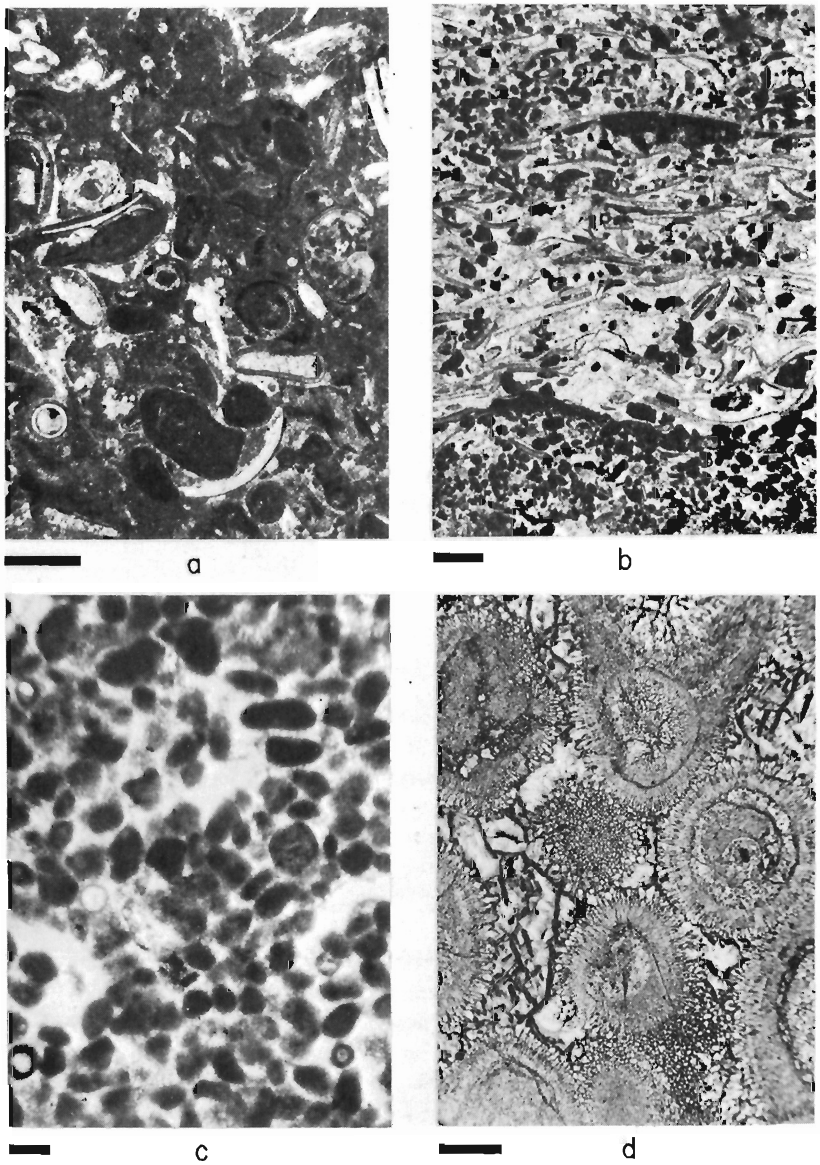


FIGURE 29

a. Black-pigmented grains in a skeletal packstone. The grains consist of gastropods, pelecypods and foraminifers, some of which have short fibrous oölitic coatings (superficial oölites), lithoclasts, and micrite matrix. Grains are partly coated and/or contain patches of black coloured pigment (iron monosulfide?); skeletal lithofacies; basal carbonate blanket; Swan Hills Formation. Thin section—ordinary light. 8–26–69–13 W5, depth 8,445 ft (approx. 2,574 m). Bar scale is equal to 0.5 mm.

b. Skeletal-pelletoidal grainstone laminae in pelletoidal packstone; bioclasts consist of thin-shelled planktonic pelecypods cemented by sparry calcite; small irregularly shaped microclasts and pellets are composed of micrite; pelletoidal lithofacies, basal carbonate blanket; Swan Hills Formation. Thin section—ordinary light. 8–26–69–13 W5, depth 8,435 ft (approx. 2,571 m). Bar scale is equal to 0.5 mm.

c. Pelletoidal grainstone composed of ovoid-shaped pellets, calcispheres, and pelecypod shells cemented by sparry calcite; pelletoidal lithofacies, basal carbonate blanket; Swan Hills Formation. Thin section—ordinary light. 2–5–69–10 W5, depth 8,151 ft (approx. 2,484 m). Bar scale is equal to 0.1 mm.

d. Well-sorted oölitic grainstone cemented by blocky sparry calcite and rarely anhydrite; oölites partly dissolved as a result of subaerial exposure(?); oölite layers composed of radial-fibrous sparry calcite; mixed oölitic lithofacies, basal carbonate blanket; Swan Hills Formation. Acetate peel—ordinary light. 8–26–69–13 W5, depth 8,444 ft (approx. 2,574 m). Bar scale is equal to 0.1 mm.

erosional contacts with underlying laminae. Interlaminations of pellets, micrite, bioclasts, and intraclasts are common and in places display a vertical grain size gradation. Mudcracks and small-scale ripples are present.

Environment

The pelletoidal lithofacies of the basal carbonate blanket probably was deposited on tidal flats and in shallow lagoons and bays. The laminated pelletoidal limestones, however, appear to have been deposited in an intertidal zone.

Skeletal Lithofacies

The skeletal lithofacies consists of dark yellowish brown and pale yellowish brown, fine- to medium-grained calcarenites (Fig. 28b). The areal distribution is similar to that of the underlying pelletoidal lithofacies, which in turn lies on the inorganic plateau (Fig. 27).

Composition

The skeletal lithofacies can be subdivided into three subfacies: skeletal wackestones, skeletal packstones (Fig. 29a), and skeletal grainstones, which occur in the ratio of 2:1:1. About 17 per cent of the sediment of the basal carbonate blanket is composed of skeletal wackestones. About 20 per cent of these wackestones is made up of fine- to medium-sand sized, unsorted and unoriented skeletal particles. The particles consist of brachiopod debris, small gastropods, ostracodes, and crinoids. *Tentaculites,* debris of green algae (charophytes), blue-green algae (*Girvanella*), and fragments of the problematic alga *Kamaena* are rare. Some of the skeletal grains are black pigmented. Nonskeletal particles consist of pellets (Fig. 29b), mud chips, and rounded mud pebbles (maximum diameter, 8 mm). The matrix is composed mainly of micrite. Composition of the skeletal packstones is similar to that of the wackestones, and differs only in the higher percentage of skeletal particles—40 to 60 per cent. The skeletal grainstone facies is typified by more subrounded to rounded particles, which range from poor to well sorted, and are cemented by sparry calcite. Gradation from one subfacies to another is complete, and can occur in a vertical distance of 1 cm. Compositional differences of this sort could have resulted from local variations in hydrodynamic conditions, by winnowing, or by redistribution by burrowing action of benthonic organisms.

Black-pigmented Carbonate Grains

Some of the skeletal grains are pigmented probably by very finely disseminated black iron monosulphide (Fig. 29a). Black rims developed around some of the lithoclasts, and are associated with very fine grained pyrite. Several workers have described blackening of the skeletal grains and their discoloration from modern sediments. Discoloration of calcareous grains with specks of pyrite and iron monosulphide was described by Macintyre (1970, p. 12) as being the result of reducing conditions in an organic-rich, brackish environment of stagnant ponds. Black-pigmented grains occur in seagrass-bank sediments of eastern Shark Bay, Western Australia (Davies, *in* Logan *et al.,* 1970, p. 139) and in shallow bays of the Florida Bay. Reducing conditions in the Florida Bay area were found in some places to be only several millimetres below the water sediment interface. Maiklem (1967), in his studies of the southern part of the Great Barrier Reef, explained the blackening of grains by the presence of land-derived iron and manganese which alter to sulphides under shelf-reducing conditions.

Most of the blackened grains observed in the basal carbonate blanket of the Sturgeon–Mitsue area occur in wackestones near the seaward edge of the inorganic plateau. Blackening of the grains suggests existence of reducing conditions near the edge of the organic plateau and low-water circulation during the time of carbonate deposition. The grains themselves could have been derived from the shelf bottom, where anaerobic, reducing conditions could exist near to the sediment-water interface, similar to that of Florida Bay.

Structures

Small-scale scour and cut-and-fill structures occur in the wackestone facies. With the exception of bioturbation, no other structures were observed.

Environment

The skeletal lithofacies appears to have been deposited in a wide spectrum of environments. Wackestones with *Tentaculites* and crinoids near the edges of the inorganic plateau are considered to have been deposited in a subtidal environment. Bioturbated skeletal wackestones commonly with green algae appear to have been deposited in very shallow subtidal to intertidal environments similar to those of the carbonate mud banks of Florida (Ginsburg, 1956), or Shark Bay (Davies, *in* Logan, 1970). The grainstones could represent either tidal channel fill or nearshore skeletal bank material. A nearshore origin can be inferred from the high content of green algae in some samples. Well-sorted grainstones could be in part a beach deposit.

Mixed Oölitic Lithofacies

The mixed oölitic lithofacies is a group of rock types consisting mainly of pelletoidal packstones, skeletal grainstones and wackestones where oölites compose more than 5 per cent of the sediment. Maximum oölite content in the mixed oölitic lithofacies is 60 per cent of total rock volume; the average is not more than 15 per cent. The mixed oölitic lithofacies occurs near the base of the carbonate blanket (Fig. 27), mainly in the central part of the Sturgeon–Mitsue area.

Composition

The oölites are developed as oölitic coatings around nucleii, with from one to four envelopes. Oölites with foraminiferal nucleii are elongate, whereas those with aphanitic pelletoidal grains as nucleii are circular. Individual concentric envelopes are rather poorly developed, and are composed of short radially fibrous calcite crystals. Oölites in the grainstone facies are coated with a thin crust of granular sparry calcite, and cemented by coarse, drusy, sparry calcite (Fig. 29d). Some have core linings of fibrous calcite, which suggests that the oölite core may have been dissolved probably due to temporary exposure of the oölites (Shearman and Skipwith, 1965). The oölites in the wackestones have partly abraded rims, suggesting a reworking of the oölites. Most of those in the mixed oölitic lithofacies can be classified as superficial (Illing, 1954, p. 36).

Environment

Oölites in the mixed oölitic lithofacies suggest the presence of high-energy shoals from which the oölite grains were retransported into surrounding sediments. The occurrence of oölite sediments can be related to the paleotopography of the underlying inorganic plateau with oölites being produced in the most elevated (shallowest) area of the platform. The main concentration in the study area lies between Snipe Lake and the Swan Hills field, and it appears that the shoal areas ranged from about 3 miles (4.8 km) to a maximum of 6 (9.6 km) in width, with water depth of several feet.

Algal Stromatolite Lithofacies

Modern classifications of the algal stromatolites by Logan *et al.* (1964) and Hofmann (1969) were designed primarily for algal structures with internal laminated fabrics, i.e., stromatolites as defined by Kalkowski (1908). The occurrence of recent nonlaminated algal mats was discussed by Neumann *et al.* (1970) who found that algal mats that developed in subtidal conditions lack internal fabrics, which they believe are characteristic of an intra- and supratidal environment. P. Hoffman (pers. com.) from studies of recent stromatolites in Hamelin Pool, Western Australia, emphasizes that textural composition of algal mats is controlled not only by environmental conditions but also by composition of the biological community.

Algal stromatolites are rare in the basal carbonate blanket (Fig. 27). They appear to be in a growth position and form beds up to 20 cm thick. Stromatolites are dark to very dark grey, have a digitated turbinate shape with straight, inclined, and partly linked branches (Fig. 28d). The internal structures differ from those of the stromatolites that occur in the underlying Gilwood Member in that they are unlaminated and have a sponge-like inner structure (spongiostromatolites?). Stromatolites occur near the base of the basal carbonate blanket, mainly in the area from the Swan Hills field to Snipe Lake. The algal stromatolite lithofacies overlies the mixed-oölitic lithofacies and the pelletoidal lithofacies. Stromatolites in the basal carbonate blanket of the Sturgeon—Mitsue area are similar in their internal structure and appearance to the genus *Aphostroma,* described by Gürich (1906) from the Viséan of Belgium.

Composition

This lithofacies is composed of 40 to 60 per cent digitated algal biomasses which have the spaces between the finger-like algal protuberances filled with skeletal wackestone. The finger-like branches measure approximately 16 mm long and 7 mm wide. The inner structure is spongelike, non-laminated, and in places displays poorly preserved vertical filamentous algal casts. Spongy fabrics occur mainly in the centres and near the top of the algal fingers. Around the algal fingers is a tightly packed rim of micrite. They also contain voids that are filled with coarsely granular sparry calcite. Rare occurrences of short thick tubes, 0.15 mm long and 0.03 mm wide, probably were internal structural channels. The algal stromatolite lithofacies probably contained several algal species. A large quantity of foreign material is present and consists mainly of foraminifers, *Kamaena*[1], and other unidentifiable shell debris. This high quantity of faunal material and its distribution suggest that the stromatolite was built mainly by sediment binding processes.

[1] Reading and Jansa (*in press*) have shown that the tubiform septate algae tentatively included to *Kamaena* Antropov is now attributed to the *Uraloporella variabilis.*

Voids between stromatolite fingers are filled with skeletal wackestone, consisting of a micrite matrix and the same biota that is enclosed in the stromatolites. Sand and silt-sized debris of stromatolites can be distinguished by the dark grey colour of the micrite matrix.

Environment

Stromatolites similar to those of the basal carbonate blanket have been described from the Devonian of Western Australia by Playford and Cockbain (1969, p. 1008), and are interpreted by them as being a product of deep-water nonskeletal algae, probably red-pigmented cyanophytes. Algal stromatolites in the Swan Hills area colonized submerged parts of slopes of the inorganic plateau which were protected from the direct influence of the waves. Stromatolites overlie shallow-water sediments and it is believed that they were formed in a much shallower environment than that proposed by Playford and Cockbain. Stromatolites of the basal carbonate blanket of the Swan Hills Formation appear to have been formed in a shallow subtidal environment similar to the recent subtidal mats at Abaco in the Bahamas (Neumann *et al.,* 1970).

Depositional History of the Basal Carbonate Blanket

Following deposition of the inorganic plateau sediments a marine transgression flooded the elevated area of the plateau, which probably was at least in part subaerially exposed prior to transgression. Shallow-marine sediments were then deposited as a thin veneer over the inorganic plateau (Fig. 27). Relief formed by preferential compaction of the Watt Mountain and Fort Vermilion lagoonal mud deposits and by deflation during periods of emergence, appears to have been as much as 30 feet (9 m) (Fig. 25) and, dependent on local bathymetric conditions, a variety of sediments were deposited. The highest areas of the plateau, where water depths probably did not exceed 6 feet (1.8 m), were sites of oölite formation. The deepest parts of the shallow sea flooding the plateau probably were no more than 40 feet (12 m), and the bottom was covered by fossiliferous carbonate mud. The coastline at the time of basal carbonate blanket deposition was at or just west of Snipe Lake. The Snipe Lake area was one of broad tidal flats, and the presence of dolomite suggests the existence of widespread supratidal conditions. The surface of the inorganic plateau appears to have been tilted eastward with a gradient of approximately 1 foot (0.3 m) to 6 miles (9.6 km). During transgression the environments migrated and interfingered slightly to the westward with a pronounced tendency for each environment to be oriented parallel to the strandline.

Deposition of the basal carbonate blanket undoubtedly was affected by a high production rate of carbonate-secreting organisms, and by a high "chemical precipitation" rate in the shallow seas overlying the inorganic plateau relative to the adjacent deeper shelf bottom. The occurrence of oölites suggests chemical precipitation of aragonite on the shallow platform, similar to that suggested for the Bahamas (Illing, 1954). Chemical deposition in the Bahamas is apparently related to the sweeping of cooler water from the deep ocean over the banks with each flood tide. As the water is warmed by the sun's rays, part of its dissolved carbon dioxide is lost and it becomes supersaturated with calcium carbonate. Furthermore, the process of precipitation can be intensified by biochemical extraction of carbon dioxide by marine plants, algae, and/or diatoms. The basal carbonate blanket probably was deposited under similar conditions.

Marginal Reef-bank Complexes of the Swan Hills Formation

The term marginal reef-bank complex is here used to represent a multiple and synchronous development of coalescent biohermal bodies along the outer margin of the carbonate platform and bank-type deposits behind (landward) of those biohermal bodies or barriers.

As used in this study the term marginal reef-bank complexes includes rocks referred to by Fong (1959) as the "lower dark brown member," by Leavitt (1968) as the "reef platform," and by Fischbuch (1968) as Divisions I to IV (Fig. 4). It is a distinctive unit consisting mainly of dark brown clastic limestone, and is widespread throughout most of the study area. Areal distribution is shown in Figure 26; thickness ranges from 70 feet (approx. 22 m) in the Snipe Lake area to approximately 160 feet (approx. 48 m) in the vicinity of Carson Creek (Hemphill *et al.*, 1970); there is a general thinning from east to west.

Despite an apparent sedimentary and organic complexity, the marginal reef-bank complexes can be divided into two distinct sedimentary cycles (Fig. 27). The lower cycle corresponds to Fischbuch's (1968) Divisions I and II, the second cycle to Divisions III and IV, with the thickness of each cycle approx. 45 to 50 feet (13 to 15 m). Deeper water biohermal limestones occur near the base of each cycle, and these are followed in turn by shallow-water lagoonal or bank-type deposits. The upper cycle appears to have been terminated by submarine or subaerial erosion. Each of the cycles represents one event of building and progradation of the reef-bank complex under conditions of a slow continuous transgression. Both vertical and lateral lithofacies and environmental zonation are apparent. As biohermal structures developed near the edge of the complex, lagoon-type sediments were deposited in the semirestricted areas of the inner part of the platform; they interfinger with biohermal sediments produced during the development of the reef-bank complex.

Organisms as Agents of Sedimentation

The Swan Hills reef-bank complex demonstrates the influence of organisms on sedimentation (Fig. 30). The marginal reef-bank complexes and the subsequently deposited solitary reef complexes are composed of both organisms that secrete preservable hard parts, and the associated sediments, which are of organic as well as inorganic origin. However, many of the most commonly represented organisms in modern reef and inter-reef communities lack preservable parts, and thus the record of organic life in fossil reefs probably is incomplete. Organic communities associated with the reef-bank complex can be divided into three major groups—primary frame-builders, secondary frame-builders, and bank communities; open-shelf communities played an additional and subordinate role.

Primary frame-builders are skeletal calcareous organisms that build a rigid framework, which maintains its form if the intersititial sediment is removed. The rigidity of the framework is related to the destructive force of waves and currents, and is lower in lower energy and sheltered areas that are protected from the full force of an open sea (Kornicker and Boyd, 1962). Accretion of the framework is achieved by a succession of overgrowths during which a new colony can spread over the in situ skeletons or debris of their predecessors. The primary frame-builders in Swan Hills marginal reef-banks were stromatoporoids and less commonly algae. Frame-builders

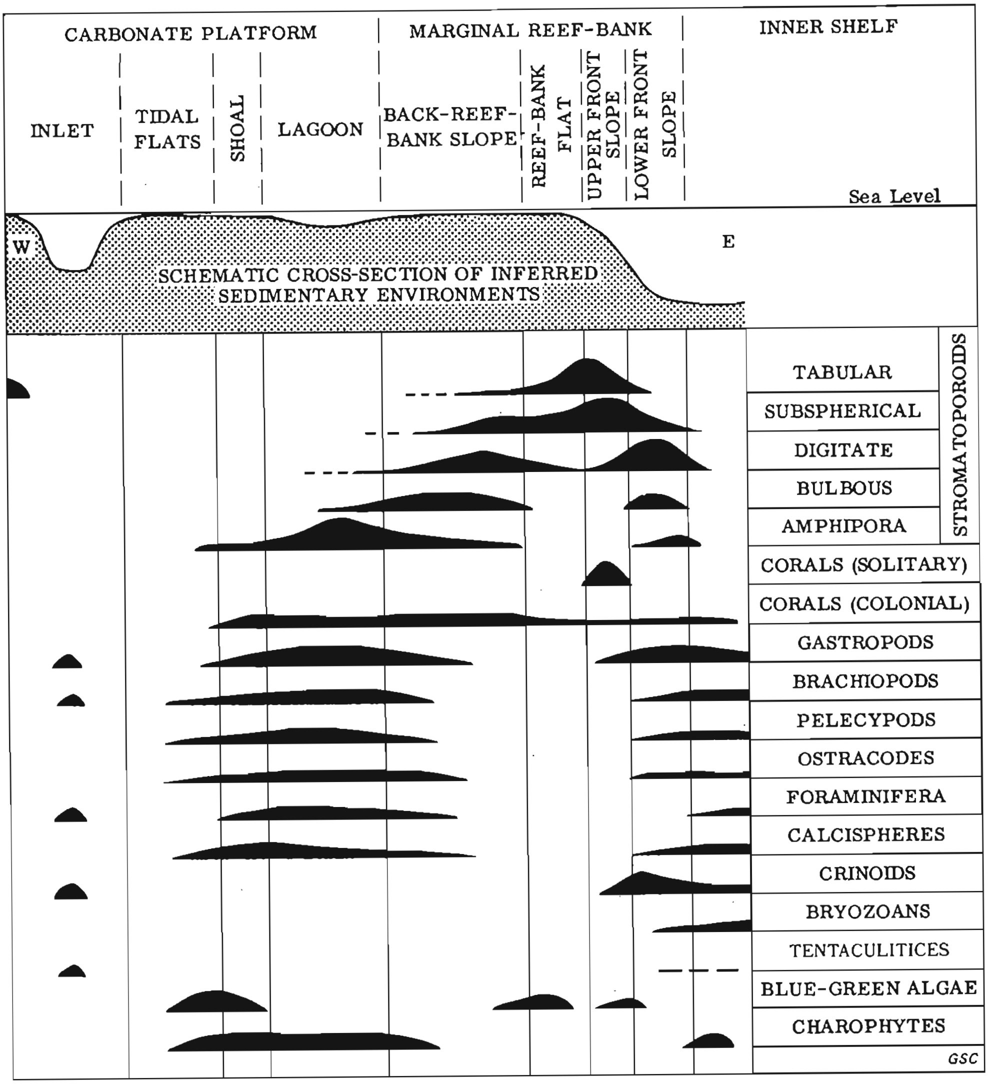


FIGURE 30. Schematic west–east variations in distribution and abundance of preserved fossils in the Swan Hills marginal reef-bank complex; inset above shows schematic cross-section of inferred environments.

were tabular, subspherical (Fig. 34d), and digitate stromatoporoids. Their collective rigidity was supplemented by encrusting algae, which bound together pieces of debris and helped cement them into a framework, although the algae themselves were not an important frame-builder in the Swan Hills marginal reef-bank complexes. Secondary frame-builders occur, in the form of debris, with primary frame-builders. Bank fossil communities acted as interstitial filling in the framework.

Secondary frame-builders include bulbous stromatoporoids, *Amphipora,* some digitate stromatoporoids (Fig. 35a), and colonial and solitary corals. The secondary frame-builders by themselves did not have potential wave resistance and did not build rigid frameworks. Consequently, they are most common on the lee side of rigid frameworks and barriers, and thus were protected from destruction by waves and currents. The faunal community acted mainly as a sediment trapper, and sedimentary particles were trapped between growing sessile organisms. Blue-green algal laminations among secondary frame-builders are rare and primary frame-builders commonly are associated with secondary frame-builders. Ramose forms are common in the lagoons with the exception of rare tabular forms such as *Euryamphipora* (Fischbuch, 1970a); secondary frame-builders generally do not have an encrusting habit. Mostly they are fragile forms seldom found in growth position. The steady accumulation of secondary frame-builders provides a continual supplement to the rigidity of the reef barrier.

Bank communities appear to have been a good source of skeletal particles and produced calcispheres, foraminifers, brachiopods, pelecypods, gastropods, ostracodes, and dasycladacean algae (Fig. 38c). Open-shelf communities contributed such forms as crinoids, bryozoans, tentaculitids, and foraminifers; these faunas were subordinate biological agents producing only skeletal particles and had no noticeable effect on the construction of the marginal reef-bank complexes. Quantitative abundance of different organic forms and their occurrence in the marginal reef-bank complexes are summarized in Figure 30. In Figure 31 additional compositional, textural, and structural parameters of inferred reef-bank environments composition are summarized.

Composition

In this study carbonates of the lower part of the Swan Hills Formation are separated into eight lithofacies on the basis of faunal content and constituents. These lithofacies are:

Marginal reef-bank facies:
- tabular stromatoporoid lithofacies
- coral lithofacies
- subspherical stromatoporoid lithofacies
- digitate stromatoporoid lithofacies

Shelf lagoon facies:
- bulbous stromatoporoid-*Amphipora* lithofacies
- *Amphipora* lithofacies
- oncolite-intraclast lithofacies
- skeletal-pelletoidal lithofacies.

Tabular Stromatoporoid Lithofacies

Although not volumetrically significant, this lithofacies played an important role in the initiation of the marginal reef-bank complex, and it occurs as a well-defined unit along outer margins of the reef-bank complex (Fig. 27). Its thickness ranges from 10 feet (3 m) at Virginia Hills to 23 feet (7 m) at House Mountain to 30 feet (9 m) in the Swan Hills field. It is limited in areal distribution and forms discontinuous zones less than 2 miles wide along the seaward edge of the inorganic plateau, and even narrower zones along the windward sides of channels and inlets of the inner platform.

Contact of this lithofacies with the underlying basal carbonate blanket commonly is distinct, whereas contact with the overlying beds is gradational. The tabular stromatoporoid lithofacies is composed mainly of tabular stromatoporoids and by definition is a true biolithite (Fig. 32a). The tabular stromatoporoids nearly always are in a growth position but in rare instances are broken, with overturned fragments which may have subsequent stromatoporoid overgrowths. At the edge of the marginal reef-banks

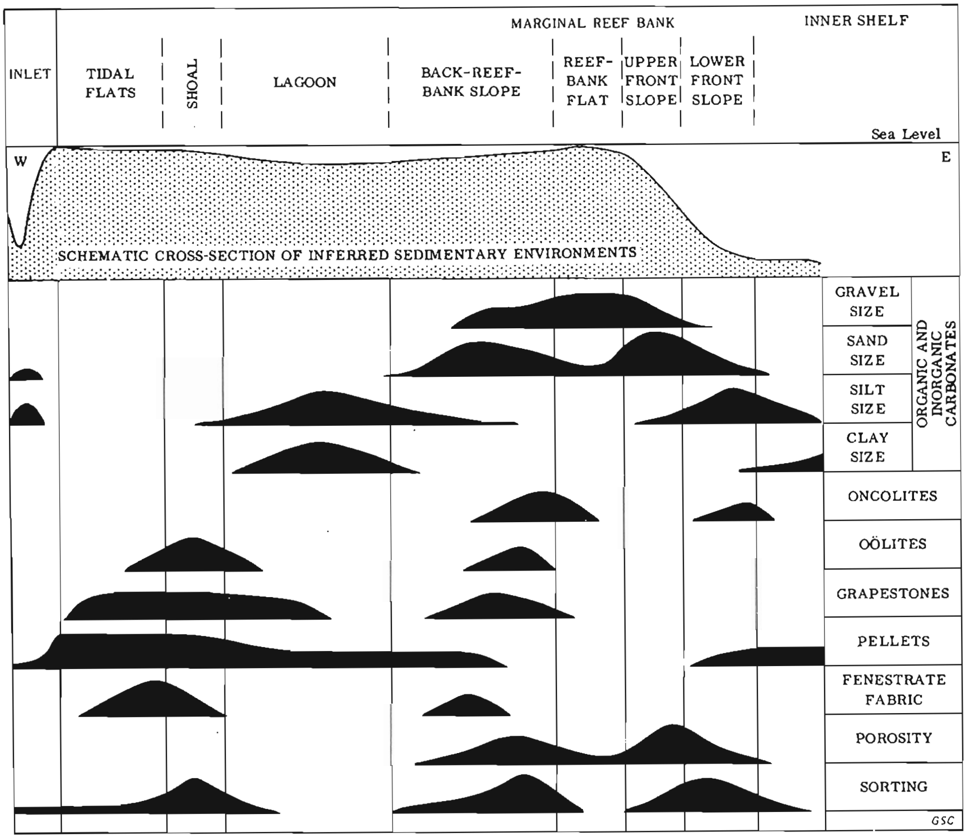


FIGURE 31. Schematic west–east variations in distribution and relative abundance of petrographic components and textural properties of carbonates in the Swan Hills marginal reef-bank complex; inset above shows schematic cross-section of inferred environments.

inclination of stromatoporoid latilaminae was found to be in the order of 20 degrees, which probably reflects the original inclination of the biohermal slope. Subspherical, digitate, and bulbous stromatoporoids are found associated with tabular stromatoporoids (Fig. 30). *Amphipora* skeletons occur on the leeward side of the tabular stromatoporoid mounds (Fig. 32c). At the front slope, crinoids (Fig. 32b) and thick-shelled corals (Fig. 32d) are prominent. These mounds are capped by a subspherical stromatoporoid zone containing poorly sorted skeletal calcarenite and stromatoporoid "rubble," and range from 5 to 20 feet (6 m) in thickness.

Composition

Tabular stromatoporoids compose 60 to 90 per cent of the rock. Thickness of individual coenostea ranges from 1 to 5 cm. Areal extent of the coenostea is not known due to lateral limitations of core diameter.

Coenostea commonly directly overlie one another, and are separated by pressure solution contacts which are outlined by black bitumen-filled stylolites. Matrix, which is rare, and fills the interstromatoporoid areas, is dark grey, and is composed of

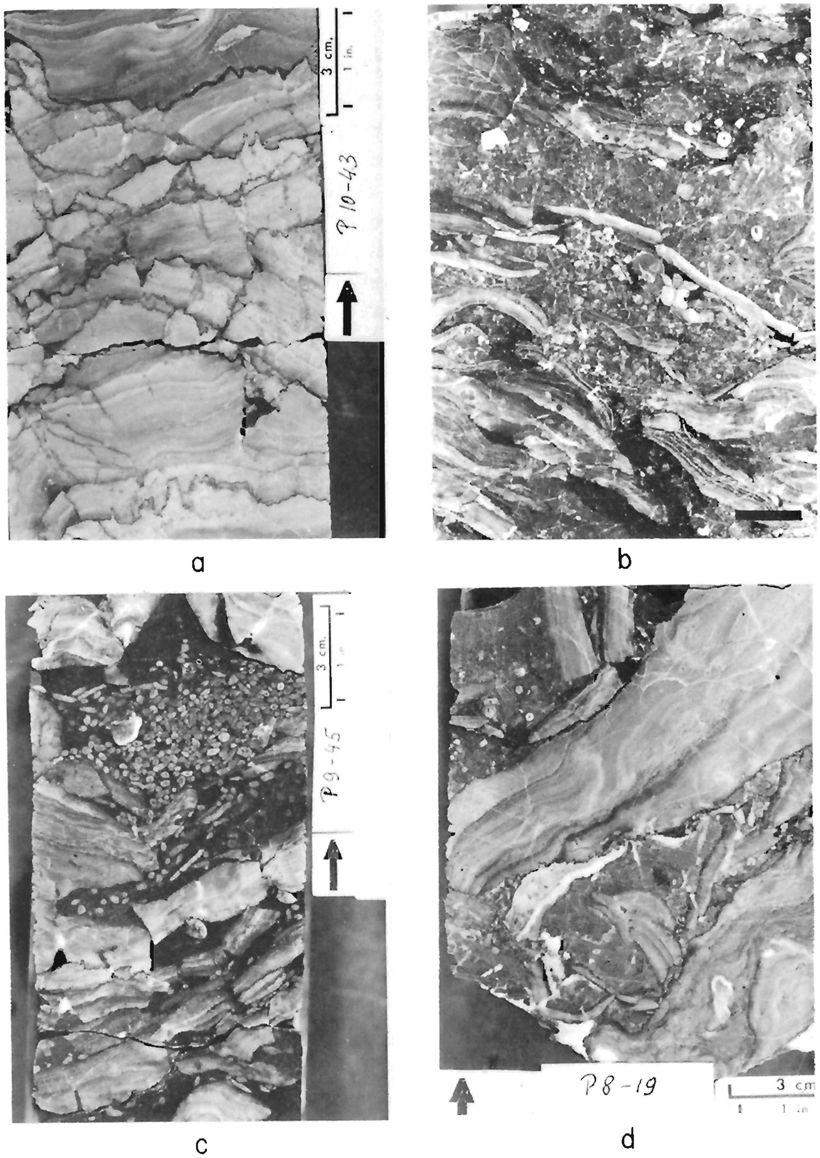


FIGURE 32

a. Tabular stromatoporoids with deep pressure-welded contacts between colonies bounded by black organic laminae; individual stromatoporoid colonies fragmented as a result of compaction(?); tabular stromatoporoid lithofacies, marginal reef-bank complex; Swan Hills Formation. Polished core slab, 2–6–70–10 W5, depth 7,562 ft (approx. 2,305 m).

b. Tabular and hemispherical stromatoporoids in a skeletal-pelletoidal grainstone matrix; clasts in matrix composed of crinoid and thick-shell pelecypod fragments; tabular stromatoporoid lithofacies; Swan Hills Formation. Polished core slab, 4–29–66–13 W5, depth 9,807 ft (approx. 2,990 m). Bar scale is equal to 1 cm.

c. Tabular and hemispherical stromatoporoids, fragmented, some overturned, *Amphipora* and solitary corals in micrite matrix; tabular stromatoporoid lithofacies, Swan Hills Formation. Polished core slab, 2–5–69–10 W5, depth 8,140 ft (approx. 2,480 m).

d. Tabular and hemispherical stromatoporoids with some colonies fragmented to form a skeletal packstone matrix; note abundant crinoid and *Grypohyllum* sp. nov. (Pedder); thick-walled solitary coral fragment near lower left corner of photograph was overgrown by the stromatoporoid colony; tabular stromatoporoid lithofacies, Swan Hills Formation. Polished core slab, 4–22–66–13 W5, depth 9,811 ft (approx. 2,991 m). Bar scale is equal to 2.5 cm.

skeletal debris consisting of fragmented stromatoporoids, gastropods, crinoids, brachiopods, and ostracodes. Large faecal pellets, 0.4 mm in diameter, rarely occur near gastropods and could have been excreted by them. Dark coloured micrite does not exceed 20 per cent of the total matrix content and is composed in part of coalesced silt-sized pellets which can be recognized in intrafossil cavities. Sparry calcite cement occurs as bladed crusts in pelecypod shell voids, as a granular cement among skeletal grains, and as sparry calcite mosaic filling fractures. Fracture-filling calcite is of the high iron variety and is of epigenetic origin. In places dolomite fills voids in fossils and fractures, and commonly is concentrated around fractures and stylolites. Void-filling and replacement by anhydrite laths are rare. In one sample, replacement of anhydrite crystals by dolomite was noted, which would suggest that dolomitization occurred after the anhydrite had been formed.

The upper surface of stromatoporoids commonly is coated with nonskeletal algal laminations, and several thin algal laminae may occur in one tabular stromatoporoid coenosteum. Symbiotic association of algae and stromatoporoids is evident in other Swan Hills limestones; stromatoporoid algal consortia were noted by Fischbuch (1968, p. 502) in the Swan Hills reef complexes, and by Klovan (1964, p. 38) in the Redwater reef complex.

Environment

The tabular stromatoporoid lithofacies in the Swan Hills Formation is on the sloping edges of an incipient carbonate platform, and is apparent on the windward sides of minor inlets and passes (Fig. 39). The sharp contact with the underlying basal carbonate blanket suggests that the substrate probably was at least partly cemented by the beginning of stromatoporoid colonization. Tabular stromatoporoid mounds appear to have been initiated in a subtidal environment, in water probably not exceeding 40 feet deep. Corals colonized the shallow lagoons formed behind the mounds; no evidence of exposure was found on the upper surface of the mounds and presumably these surfaces were never exposed even at low tide. The upper limit of tabular stromatoporoid growth probably occurred in a shallow subtidal zone. When the growth of the mounds reached the shallow subtidal zone the stromatoporoid growth ceased and current and wave action reduced some of the stromatoporoids to sand-sized material forming a calcarenite cap. However, some of the pellet-sized material may have been derived from the inner shelf lagoon. Low-angle slope and favourable hydrodynamic conditions would have kept the detrital carbonates near the crest of the mounds. The "barrier" effect of the tabular stromatoporoid mounds, therefore, had a strong influence in modifying the surrounding environments.

Pre-existing passes and channels related to the paleotopography of the inorganic plateau were narrowed and deepened by the growing mounds, which strengthened turbulence and velocity of the water flow. After the development of the tabular stromatoporoid mounds the floors of the passes and channels probably remained barren of sediment throughout most of the time of deposition of the marginal reef-bank complexes.

The hydrological effect of the tabular stromatoporoid lithofacies on the adjacent environments confirms the concept that tabular stromatoporoids flourished in a subtidal environment and may have penetrated as high as the shallow subtidal zone. A low- to medium-energy depositional environment, therefore, is suggested for the tabular stromatoporoid lithofacies. Stromatoporoid students do not agree on the depositional environment of tabular stromatoporoids. A high-energy environment for their growth in the Upper Devonian Miette reef has been suggested by Noble (1970, p. 522), and from the Jeffersonville limestone by Perkins (1963, p. 1341). Galloway (1957), Lecompte (1951), and Fischbuch (1968, p. 500), on the other hand, suggest that tabular forms thrived in relatively deep, quiet waters. Devonian tabular stromatoporoid mounds can in part be compared with modern platy and encrusting corals (*Montastrea annularis*), which generally occupy the lower reef front (Maxwell, 1968, p. 159; Mesolella, 1968, p. 59).

Coral Lithofacies

The coral lithofacies occurs as a well-defined biostromal unit at the base of the carbonate platform (Fig. 27). Its thickness ranges from 0 to 18 feet (5 m). The distribution generally is limited to the area of the basal carbonate blanket. The coral lithofacies does not extend beyond the western limit of the inorganic plateau, and its maximum development is restricted to the inner part of the platform behind tabular stromatoporoid "barriers" (Fig. 30).

Its contact with the underlying basal carbonate blanket is sharp, and in several places erosional (Fig. 34a); contact with the overlying lithofacies is gradational. The name of this lithofacies is derived from the abundance of the ramose tabulate corals, *Thamnopora* and *Coenites*. The coral lithofacies was described at Judy Creek by Murray (1964); from the Swan Hills field by Edie (1961), and Fong (1959); from the Swan Hills archipelago by Fischbuch (1968, Div. I); and from the Goose River area by Jenik and Lerbekmo (1968).

This lithofacies is composed mainly of dusky yellowish brown to pale yellowish brown coral–stromatoporoid calcirudite, with a poorly sorted and unsorted skeletal packstone or wackestone matrix. Thin bituminous laminae and stylolitic contacts are common. It also is typified by a random distribution of the coral genera *Thamnopora* and *Coenites*, and in the area close to the edge of the reef-bank complex by the occurrence of primary frame-builders such as tabular, subspherical, digitate, and bulbous stromatoporoids found interspersed among the corals (Fig. 34b). Corals commonly exceed 30 per cent of total rock volume and aid in tracing this lithofacies throughout the general Swan Hills area. They range in size from 1 to 3 cm in diameter. Corals represent an in situ accumulation; in-place disaggregation of the biota probably took place with little transportation, since broken coral fragments have sharp edges and are unsorted.

Composition

The major constituents of the coral lithofacies are secondary frame-builders, skeletal grains, and micrite. Minor constituents are primary frame-builders, sparry calcite, dolomite, anhydrite, and pyrite.

The frame-builders consist of 5 to 80 per cent colonial corals, 0 to 20 per cent tabular, digitate, and bulbous stromatoporoids, and in places up to 15 per cent *Amphipora*. Silt- and sand-sized skeletal grains are the dominant matrix constituent; they are poorly sorted to unsorted and probably were derived from frame-builders. Some of the grains may have been derived from other organisms such as brachiopods, crinoids, ostracodes, gastropods, Foraminifera, and calcispheres that also inhabited the carbonate banks; grains from these sources, however, commonly do not exceed 10 per cent of the sand-sized fraction. A full range of grain sizes, from sand to very fine silt, occurs in the

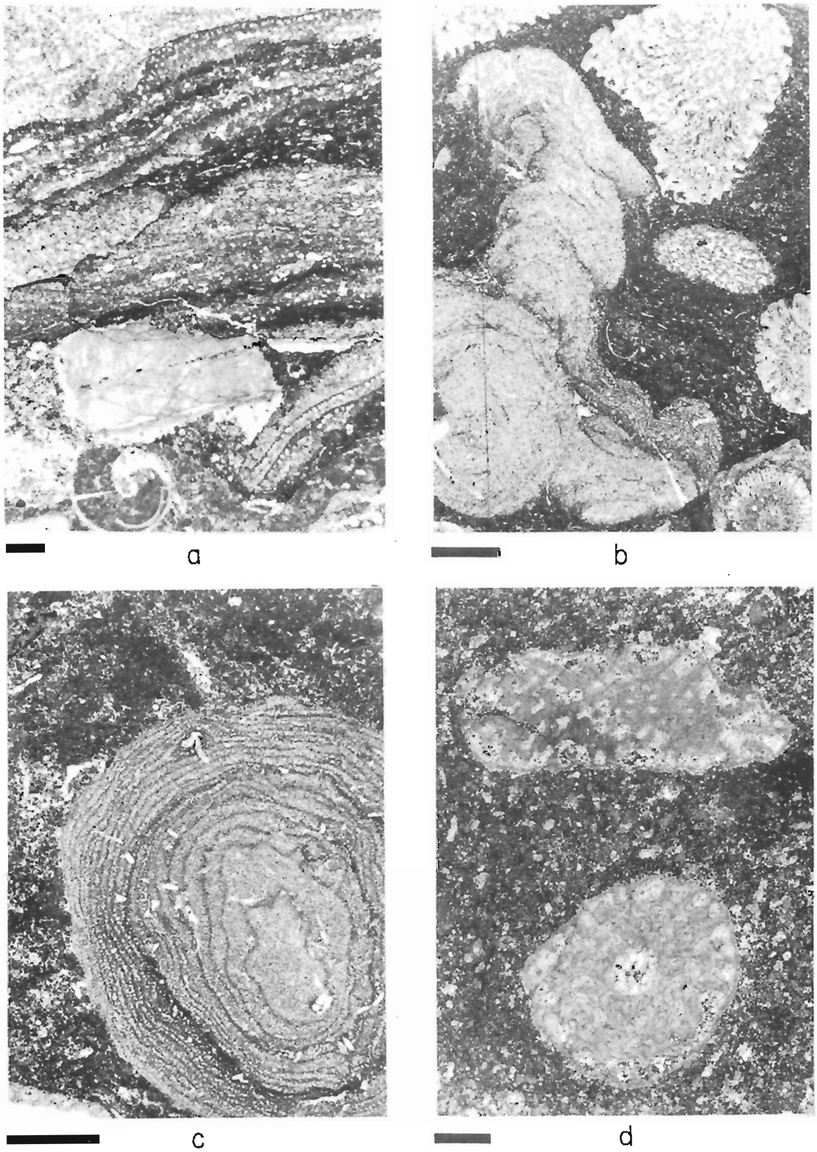


FIGURE 33

a. Subspherical stromatoporoids, crinoids, pelecypods, gastropods, and foraminifers in a pelleted mud matrix; sparry calcite cement in upper part of picture and erosion of surface of stromatoporoid colony may indicate current erosion; hemispherical stromatoporoid lithofacies; Swan Hills Formation. Thin section—ordinary light. 4–29–66–13 W5, depth 9,807 ft (approx. 2,990 m). Bar scale is equal to 1 mm.

b. *Thamnopora* and digitate stromatoporoids in a silt-sized skeletal packstone matrix; matrix composed of silt-sized debris of corals, stromatoporoids, algae, foraminifers, and micrite; voids in corals filled with sparry calcite, dolomite, chalcedony, and rarely lath-shaped crystals of anhydrite; note encrustation of coral by stromatoporoid in the lower left corner of photograph; coral lithofacies; Swan Hills Formation. Thin section—ordinary light. 4–22–69–18 W5, depth 9,467 ft (approx. 2,890 m). Bar scale is equal to 5 mm.

c. Bulbous stromatoporoid and *Amphipora* in a fine sand-sized, skeletal-pelletoid grainstone consisting of foraminifers, ostracodes, pellets, dasycladacean algae, and calcispheres cemented by blocky sparry calcite; bulbous stromatoporoid-*Amphipora* lithofacies; Swan Hills Formation. Thin section—ordinary light. 10–23–71–19 W5, depth 8,720 ft (approx. 2,660 m). Bar scale is equal to 5 mm.

d. Transverse and axial section of *Amphipora* in poorly sorted skeletal grainstone matrix; skeletal grains consist of disintegrated *Amphipora* cemented by blocky ferroan sparry calcite and dolomite; note well-preserved thin outer rim of the *Amphipora*. Skeletons stained with Alizarin Red S; *Amphipora* lithofacies; Swan Hills Formation. Acetate peel—ordinary light. 10–21–70–18 W5, depth 8,906 ft (approx. 2,715 m). Bar scale is equal to 1 mm.

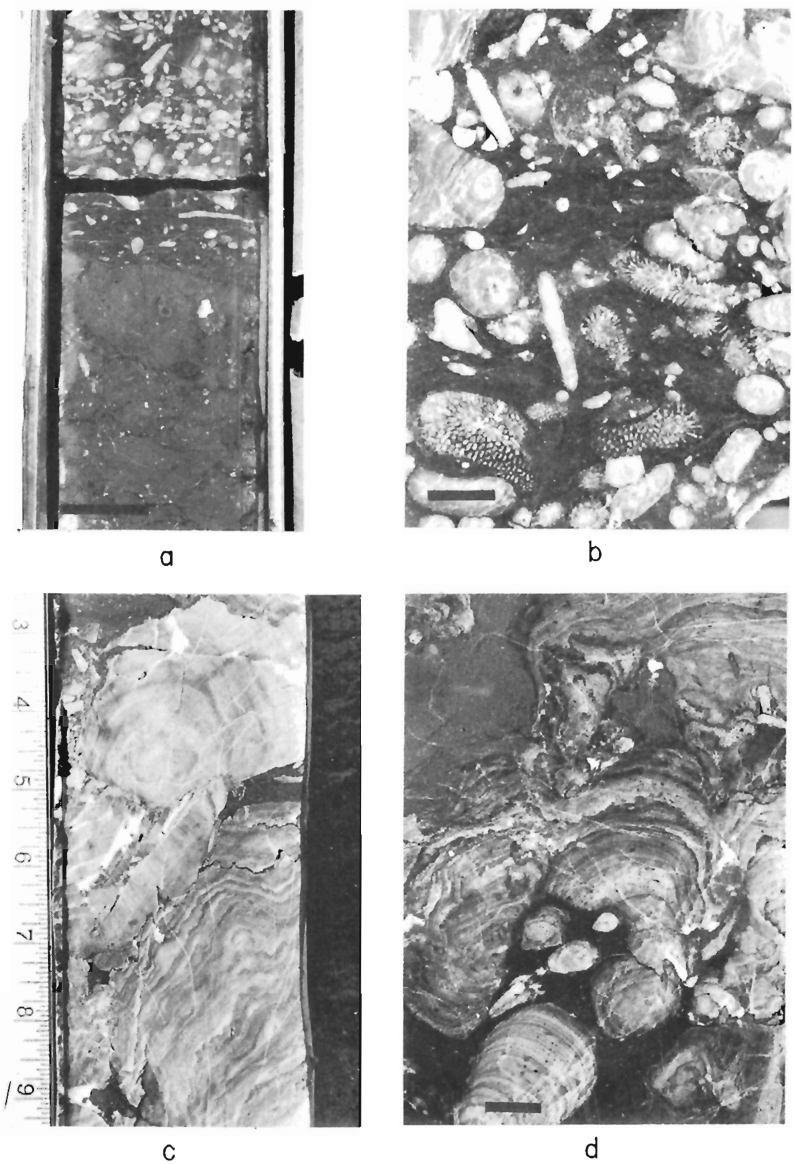


FIGURE 34

a. Distinct sharp contact between crinoidal biomicrite of the basal carbonate blanket and the coral lithofacies of the marginal reef-bank complex; Swan Hills Formation. Core photograph, 4–6–70–10 W5, depth 7,716 ft (approx. 2,352 m). Bar scale is equal to 3 cm.

b. Thamnoporid corals, digitate stromatoporoids, and crinoids in silt-sized wackestone matrix; coral lithofacies; Swan Hills Formation. Polished core slab, 4–22–69–18 W5, depth 9,387.5 ft (approx. 2,862 m). Bar scale is equal to 1 cm.

c. Subspherical stromatoporoids in a skeletal packstone. Note deep pressure-solution contacts; subspherical stromatoporoids lithofacies; Swan Hills Formation. Core photograph, 4–8–70–10 W5, depth 7,677 ft (approx. 2,340 m).

d. Subspherical stromatoporoid colonies in a poorly sorted silt to fine sand-sized, skeletal-pelletoidal packstone. Note the encrusting character of the stromatoporoids, producing a rigid framework; subspherical stromatoporoid lithofacies; Swan Hills Formation. Polished core slab, 10–21–70–18 W5, depth 8,874 ft (approx. 2,702 m). Bar scale is equal to 1 cm.

coral lithofacies with the silt-sized grains prevailing in most of the studied samples (Fig. 33b). Grains display little abrasion which suggests a complete in situ degradation of particles down to mud-sized material. Micrite composes up to 30 per cent of the rock and fills voids between individual grains. The micrite is stained to a dark brown by organic pigmentation.

Sparry calcite is a minor constituent of the coral lithofacies; it fills void spaces in bioclasts or their moulds, and is present in small fractures and veins. Total sparry calcite content is less than 2 per cent of total rock volume. A slightly ferrous variety of dolomite fills voids and fractures. Anhydrite rarely occurs as a void filler and may partly replace skeletons of corals and stromatoporoids. Anhydrite crystals are elongate with a well-developed lath shape up to 2.5 mm long. Silica filling voids in the fossils is rare, and filling short veins is very rare. Pyrite is scattered throughout the micrite matrix. The coral lithofacies has subsequently undergone intensive solution which is evident from abundant stylolites and pressure-solution contacts between bioclasts. Organic borings in stromatoporoids and corals, which are filled by silt-sized pellets, are similar to those noted by Murray (1966, p. 57) and Fischbuch (1968, p. 556).

Sedimentary Structures

With the exception of minor orientation of some organic debris parallel to the depositional plane, no structure was observed. Microstylolites or horizontally oriented thin bituminous laminae are present, and appear to be a later diagenetic feature. Disorientation and random occurrence of frame-builders is a characteristic structural feature of the coral lithofacies.

Environment

The coral lithofacies is a shoal-like deposit that appears to have had incipient tabular stromatoporoid mounds along its outer margins.

The surface of the incipient barriers probably was well below mean sea level, which permitted the low-energy waves to sweep over them and into the shallow inner lagoon that was populated by thickets of ramose corals, digitate stromatoporoids, and thin sheets of tabular stromatoporoids. The frame-building biota appears to have been more common near the margins of the coral unit. The unsorted character of the matrix of the coral lithofacies suggests low-energy conditions and mainly in situ deposition and disintegration of the corals. The fine-grained muds formed by the breakdown of skeletal particles were from time to time retransported toward the open sea or redeposited in shallow sheltered embayments where small meadows of *Amphipora* grew in the quieter waters.

Modern fauna and sediments resembling the coral lithofacies are found in the recent Bahama Banks and Florida Keys, within the *Porites* coral zone. The coral lithofacies is probably similar ecologically to the *Porites* zone. Windward lagoons of the Bahamas are, according to Newell and Rigby (1957, p. 35) covered by coral meadows. Composition of these Bahaman lagoons is approximately 30 per cent coralline algae, 25 per cent corals, and 10 per cent Foraminifera (Goldman *in* Newell and Rigby, 1957). In comparison, the coral lithofacies of the Swan Hills area is composed of approximately 30 per cent stromatoporoids, 20 per cent corals, and 10 per cent Foraminifera and other bank organisms. The high percentage of mud in the lower

Swan Hills lagoons can be explained partly by a lower energy environment and partly by micritization of clastic grains.

The disappearance of colonial corals in subsequent stages of marginal reef-bank development (Fig. 39) probably was the result of a greater constriction of inlets and passes and by the increased "barrier" development along the outer bank margins which restricted inflow of nutrient-bearing waters (Fig. 27). Carbonate mud then began to accumulate which, coupled with a restriction of free water exchange between lagoons and the open sea, produced unfavourable conditions for coral development and consequently resulted in their disappearance. Corals occurring in later stages of reef-bank evolution are thick-walled solitary corals such as *Grypophyllum* (Pedder, 1971), which appear on the upper fore-reef slopes (Fig. 32b).

Subspherical Stromatoporoid Lithofacies

This lithofacies is composed of olive-brown calcirudites and calcirudite-sized biolithite that consists of up to 90 per cent subspherical stromatoporoids (Fischbuch, 1968, p. 458) in growth position as well as fragmented debris with voids filled with poorly sorted skeletal calcarenite (Fig. 34c). Grains and fragments derived from stromatoporoids are abundant and are intermixed with fragmented crinoids, brachiopods, and solitary corals typical of the open-bank communities. The subspherical stromatoporoids facies overlies the tabular stromatoporoid lithofacies and interfingers with the digitate stromatoporoid lithofacies in the fore-reef area (Fig. 27). Toward the back reef-bank area the hemispherical stromatoporoid facies intertongues with the digitate and bulbous stromatoporoid-*Amphipora* facies. The matrix of the subspherical stromatoporoid facies in the back reef-bank area is more micritic and has a greater amount of poorly sorted wackestone (Fig. 32d) in comparison to the grainstone and packstone matrix of the fore-reef. Of the wells examined in this study none was situated in an exact reef-bank crest position, therefore, it was not possible to fully establish size and composition variations of this environment. Maximum thickness of the subspherical stromatoporoid lithofacies does not exceed 15 feet.

Composition

In the fore-reef zone (Fig. 30) subspherical stromatoporoids compose up to 65 per cent of total rock volume and have a poorly sorted calcarenite matrix cemented by sparry calcite (Fischbuch, 1968). Micrite is conspicuously absent from the matrix. The interstitial voids make this facies an important oil reservoir. Subordinate components of the subspherical stromatoporoid lithofacies are: 0 to 35 per cent digitate stromatoporoids, and 0 to 15 per cent *Amphipora*. Solitary corals, pelecypods, crinoids, ostracodes, brachiopods, and gastropods occur as minor constituents and rarely exceed 5 per cent of total rock volume. *Tentaculites* and the algae *Girvanella* and *Sphaerocodium* are extremely rare. Skeletal and pelletoidal grains are of a silt-to-coarse-sand size, with an average diameter of 0.12 mm. Grains are unsorted and commonly tightly packed. Micrite is apparent in most of the back reef-bank areas and increases from 20 per cent to 60 per cent of total rock volume toward the interior of the reef-bank complex. This also was noted by Murray (1966, p. 60) at Judy Creek. Sparry calcite and dolomite fill voids between grains. Pyrite is rare, and occurs only as individual crystals scattered throughout the matrix. Lath-shaped anhydrite crystals that fill cavities are rare.

Sedimentary Structures

The subspherical stromatoporoid lithofacies has no prominent structural features. It is a rather structureless accumulation of in situ frame-building biota, where large dome-shaped subspherical stromatoporoids engulfed each other as well as intervening sediment, thereby creating an interlocking, somewhat rigid structure, with the voids filled by skeletal sand- and silt-sized material (Fig. 34c).

Environment

Subspherical stromatoporoids appear to have occupied the surface of the upper reef-bank slope and the upper surface of the reef-bank (Fig. 30) which were immediately below low-water level. The heads of stromatoporoids are closely spaced and interlocked; they overgrew younger colonies and thus formed a massive wave-resistant skeletal framework. The combined encrusting habit of stromatoporoids and algae allowed these organisms to form a bulwark strong enough to considerably modify the environment. The domal shape and encrusting habit of subspherical stromatoporoids probably were more suitable for colonization of a sandy substrate than the laterally encrusting habit of tabular stromatoporoids. The absence of micrite in the fore-reef matrices suggests that these areas were subjected to wave surges leading to the winnowing out of fine-grained particles. The subspherical stromatoporoid facies can be compared to the recent *Diploria–Montastrea* reef communities of the recent Florida reef-tract. The surface of this lithofacies probably was not exposed at mean sea level although subspherical stromatoporoids probably had the capacity to grow into the surf zone.

Digitate Stromatoporoid Lithofacies

The digitate stromatoporoid lithofacies is a transitional unit which lies between the subspherical stromatoporoid lithofacies and the calcarenite lithofacies of the fore-reef slopes, and also occurs intermittently between the subspherical stromatoporoid lithofacies and the bulbous stromatoporoid–*Amphipora* lithofacies of the back-reef bank areas (Fig. 30). The name digitate stromatoporoid lithofacies is derived from the abundant finger-like, dendroid stromatoporoids in this rock type (Fig. 35a); it was recognized by Murray (1966) at Judy Creek and by Fischbuch (1962, 1968) in the Kaybob and other Swan Hills solitary reef complexes. In the marginal reef-bank complexes this facies is aelatively poorly developed (Fig. 27). The digitate stromatoporoid facies consists of light olive-grey limestones containing digitate, subspherical, or tabular stromatoporoids. *Amphipora* is common in places; corals are rare. In several cores it was observed that digitate stromatoporoids were attached to a common tabular base (Fig. 37a), which had spread in an encrusting form over underlying sediment. Forms of this type have been noted and formally described by Fischbuch (1970, p. 72). These forms grew from an irregular base in finger-like, wedge-shaped, conical protuberances, and generally are vertically oriented. The finger-like protuberances probably were fragile and easily broken. Redeposited stromatoporoid "fingers" formed what now represents digitate stromatoporoid calcirudites. Solenoporid algae were noted in this lithofacies by Murray (1966, p. 51) in the Judy Creek area.

Composition

The main constituents of this lithofacies are digitate stromatoporoids which commonly occur with subspherical stromatoporoids, and rarely with tabular stromatoporoids

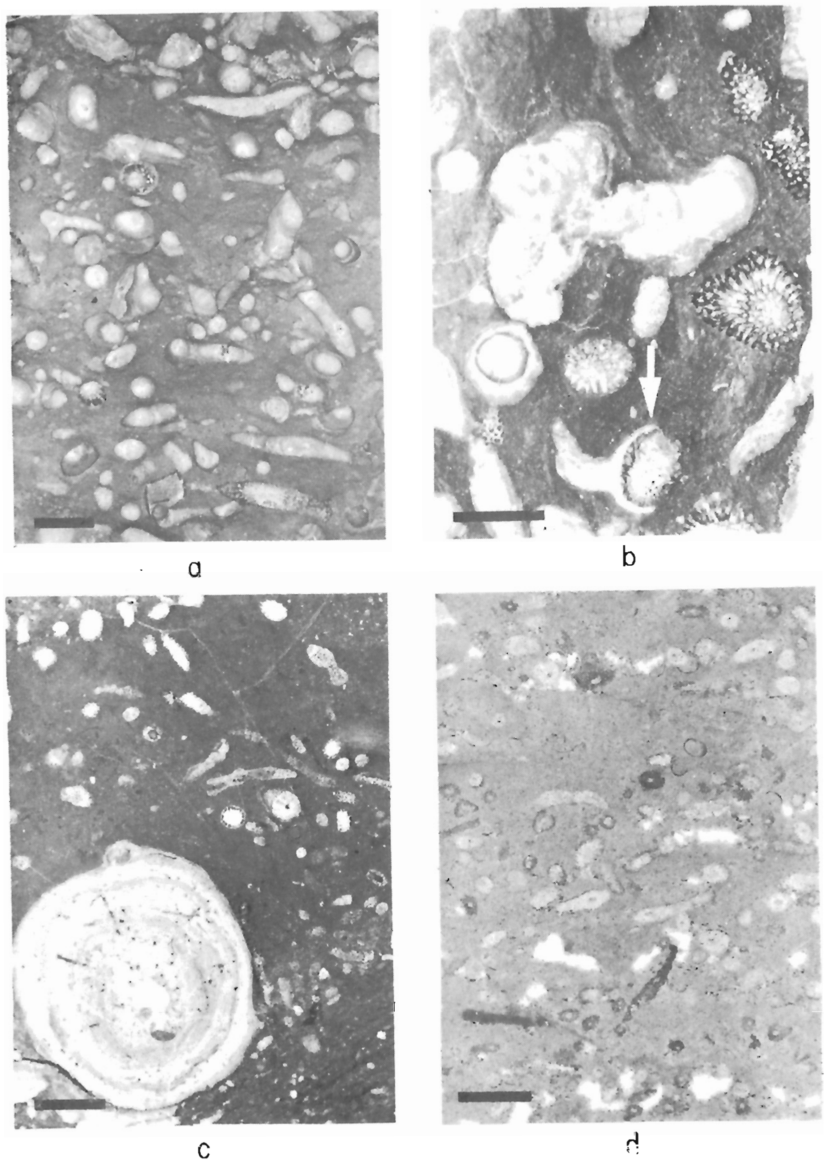


FIGURE 35

a. Slightly reworked fragments of digitate stromatoporoids, *Thamnopora* and pelecypod shells in a micrite matrix. Note lack of preferential orientation of skeletal debris and branching form of some of the stromatoporoids; digitate stromatoporoid lithofacies; Swan Hills Formation. Polished core slab, 10–13–68–16 W5, depth 9,161 ft (approx. 2,793 m). Bar scale is equal to 1 cm.

b. Overturned digitate stromatoporoid attached to the coral, in poorly sorted skeletal-pelletoidal packstone matrix; coral lithofacies; Swan Hills Formation. Polished core slab, 4–22–69–18 W5, depth 9,396 ft (approx. 2,865 m). Bar scale is equal to 1 cm. (Top of the picture to the right.)

c. Bulbous stromatoporoids and *Amphipora* in a dark grey coloured matrix composed of silt-sized skeletal grains and micrite. Note well-preserved *Amphipora* coenostea suggesting in-place deposition. Considerable solution is indicated by densely concentrated microstylolites (lower right corner); bulbous-stromatoporoid-*Amphipora* lithofacies; Swan Hills Formation. Polished core slab, 10–23–71–19 W5, depth 8,720 ft (approx. 2,660 m). Bar scale is equal to 1 cm.

d. Slightly reworked *Amphipora* coenostea and pelecypod fragments in a pelletoidal-skeletal wackestone matrix. Poorly developed birdseye vugs are filled with coarse sparry calcite; light grey *Amphipora* lithofacies; Swan Hills Formation. Polished core slab, 10–13–68–16 W5, depth 9,130 ft (approx. 2,783 m). Bar scale is equal to 1 cm.

or corals, and make up from 10 to 40 per cent of total rock volume. Individual stems range from 6 to 10 mm in diameter and are up to 4 cm long. In reworked digitate stromatoporoid calcirudites *Amphipora* is common in some places; bulbous stromatoporoids are rare. The matrix is composed of silt- to fine sand-sized skeletal grains which commonly are obliterated by micritization; in some places most of the matrix is composed of micrite. Grains are irregular in shape and poorly sorted. Crinoids, ostracodes, brachiopods, and calcispheres are minor constituents. Anhydrite laths are present in some digitate stromatoporoid skeletons.

Sedimentary Structures

The common occurrence of solution compaction features, such as microstylolites, suggests that the close spacing of grains could be the result of intense solution and compaction. Digitate stromatoporoids commonly are unoriented suggesting in-place disintegration (Fig. 35a).

Environment

The digitate stromatoporoid lithofacies is closely related to the subspherical stromatoporoid lithofacies. Digitate stromatoporoids appear to have been partly encrusting forms (thin basal attachments, Fig. 35b) and partly sediment trapping forms, with the sediment being trapped between the finger-like protuberances. Packstone fabrics, occurrence of micrite, association with subspherical, tabular stromatoporoids, and *Amphipora,* suggest that this lithofacies was deposited in a relatively deep water environment where normal wave action was limited mainly to oscillatory sea bottom movements. This environment probably was subjected to direct wave force only on rare occasions. It probably occupied the lower front slope and the rear slope of the reef-bank (Fig. 30). The digitate stromatoporoid lithofacies can be compared to the *Acropora cervicornis* zone of the Barbados, which is composed of broken and unoriented branches that are rarely in growth position and grow in the depth range from 15 to 40 feet (4.5 to 12 m) (Mesolella, 1968, p. 56). Toward the lagoon the digitate stromatoporoid lithofacies appears to have occupied a shallower environment.

Bulbous Stromatoporoid–Amphipora Lithofacies

Bulbous stromatoporoids and *Amphipora* are found mainly in the inner part of the marginal reef-bank complex (shelf-lagoon, Fig. 27). The name is derived from the abundance of small spherical stromatoporoids, which were termed bulbous by Murray (1966). The average diameter of these stromatoporoids is about 4 cm. They commonly are associated with *Amphipora,* digitate, and subspherical stromatoporoids, and very rarely tabular stromatoporoids and tabulate corals (Fig. 33c). Thickness of this lithofacies ranges from 20 to 30 feet (6 to 9 m) and composition remains relatively consistent throughout the area (Fig. 27). Near the margins of the reef-bank complex the percentage of secondary frame-builders commonly increases up to 80 per cent (Fig. 30). Primary frame-builders also are present, and the percentage of micrite matrix decreases. The bulbous stromatoporoid–*Amphipora* lithofacies grades into, and is interbedded with, the digitate stromatoporoid lithofacies, the oncolite-intraclast lithofacies, or the subspherical stromatoporoid lithofacies. Lagoonward it grades into the *Amphipora* lithofacies, or the mixed skeletal-nonskeletal lime sand facies. Bulbous stromatoporoid–*Amphipora* lithofacies were described at Judy Creek by Murray (1964, p. 59–70) as the *Amphipora*-stromatoporoid beds, and by Fischbuch (1968, p. 478–481) in Division II of the Swan Hills Formation.

Composition

The bulbous stromatoporoid–*Amphipora* lithofacies is composed primarily of secondary frame-builders, skeletal grains, pellets, micrite matrix, and sparry calcite cement (Fig. 33c). Frame-builders consist of from 10 to 30 per cent bulbous stromatoporoids, from 15 to 30 per cent *Amphipora,* from 10 to 15 per cent digitate stromatoporoids, and less than 10 per cent subspherical stromatoporoids. In the bank environment other fauna such as gastropods, corals, brachiopods, ostracodes, crinoids, foraminifers, and calcispheres occur in quantities commonly less than 5 per cent. Silt- to sand-sized skeletal grains are angular, poorly sorted and commonly micritized; in places they are very abundant and constitute up to 40 per cent of some samples. Rare angular pellets were probably produced by the disintegration of poorly cemented carbonate mud. The micrite matrix is an important constituent and in places averages up to 35 per cent. Sparry calcite fills interstitial voids. Dolomite and ferro-dolomite appear to be later void-filling cements, and partly replace the sparry calcite. Anhydrite occurs rarely replacing fossils. The matrix between bulbous stromatoporoids and *Amphipora* can be classified as skeletal wackestone and packstone. Algal borings and micrite envelopes around bioclasts are very common in this lithofacies. Fragments of *Amphipora* and digitate stromatoporoids in places have a thin, brownish grey, structureless coating.

Sedimentary Structures

Generally this facies is relatively structureless and fragments are unoriented. Rarely, however, there is a vague horizontally bedded aspect to the rock.

Environment

A high percentage of silt, fine sand-sized grains, and micrite as well as the occurrence of *Amphipora* suggests deposition either in a semiprotected, sheltered area of a reef-bank, or in a lagoon. In the surf zone on the Alacran reef, abrasion produces a high percentage of silt, which subsequently is deposited in moats and deeper parts of the lagoon (Hoskin, 1963). In the Swan Hills area the algal-coated skeletal debris in the bulbous stromatoporoid–*Amphipora* lithofacies, together with micritization of skeletal grains, algal borings, etc., suggests a shallow-water environment, and water depth comparable with the Alacran reef can be estimated as having been from approximately 10 to 60 feet (3 to 18 m).

Amphipora Lithofacies

The name of this rock type is derived from the abundance of *Amphipora,* which in places is the only macrofossil (Fig. 35d). Thickness of the lithofacies ranges from 2 to 45 feet (approx. 0.6 to 14 m), and it is interbedded with calcarenites and the bulbous stromatoporoid–*Amphipora* lithofacies (Fig. 27). *Amphiphora* beds occur in two subfacies which are separated on the basis of colour—the dark brown and the light brown beds. These commonly are interbedded with each other and have mutually transitional boundaries. In the marginal reef-bank complex dark brown *Amphipora* beds are most common, but light brown beds also occur.

Dark Brown Amphipora Subfacies

This subfacies is composed of *Amphipora* skeletal packstones and *Amphipora* gastropod pelletoidal wackestones. Most *Amphipora* skeletons are oriented horizontally, and the fragile outer rims are preserved suggesting little transportation (Fig. 33d). Where transportation has occurred this thin outer rim of the skeleton is disrupted or absent (Fischbuch, 1970, p. 67). Thus, the presence or absence of the outer rim provides a reliable criterion for determining degree of fossil transportation. *Amphipora,* because of its small size and fragile character, was easily transported by currents and is apparent in most reef facies. In the dark brown *Amphipora* subfacies *Amphipora* coenostea commonly have the outer rim intact and probably indicates in-place, or nearly in place, accumulations.

Amphipora Skeletal Packstone Microfacies

Composition

The *Amphipora* skeletal packstone microfacies (Fig. 33d) is dark brownish grey and contains up to 70 per cent *Amphipora. Amphipora* skeletons are up to 4 mm wide and to 2 cm long, and generally are horizontally oriented. Approximately 30 per cent of the coenostea in this microfacies are reworked and fragmented. Sand-sized detrital grains of *Amphipora,* ostracodes, calcispheres, foraminifers, rare pelecypods and kamaenid algae constitute the matrix. Micritization of *Amphipora* debris by boring algae is common and some of the skeletal debris is completely micritized. Micrite fills voids between skeletal particles, as well as *Amphipora* galleries and vesicles. In other places skeletal voids of *Amphipora* are filled with coarse-grained blocky sparry calcite which grades into granular crusts adjacent to skeletal walls, and rarely by subhedral dolomite or silica.

Sedimentary Structures

In intervals of alternating lithology horizontal stratification is relatively well developed.

Amphipora Gastropod Pellet Wackestone Microfacies

Composition

This microfacies is dusty yellowish brown and contains from 10 to 15 per cent *Amphipora.* The matrix is composed of approximately 30 per cent unsorted angular skeletal grains of coarse silt to fine sand size (0.05–0.25 mm). Skeletal fragments constitute 10 to 20 per cent of total rock volume and consist of fragmented *Amphipora,* gastropods, ostracodes, pelecypods, foraminifers (*Parathikinella*), calcispheres (*Calcisphaera spinosa*), rarely algae (*Girvanella*), and rare clasts of algal mats. Silt-sized pellets commonly merge to form a dark grey micrite matrix. Blocky sparry calcite of both high- and low-iron variety occurs in voids filling *Amphipora* galleries and in moulds after gastropods and pelecypods. In places where fenestral fabrics are present fenestra are filled with high-iron, coarse-grained, blocky, sparry calcite. Dolomite is rare and in places fills fossil voids.

Sedimentary Structures

Stratification is not apparent, although in places vague horizontal bedding is suggested by concentrations of *Amphipora* or bedded calcarenitic matrix. Horizontal pseudolaminations resulting from closely spaced microstylolites are common. Fenestral fabrics are rare.

The light brown *Amphipora* subfacies is characterized by *Amphipora* skeletal grainstones that are pale yellowish brown to dark yellowish brown. Linear fenestral fabrics are common, with fenestra filled by bladed sparry calcite, and in places with internal sediment at the bottom of cavities. Internal sediment consists of silt-sized pellets. Very thin laminae of micrite in places line *Amphipora* galleries.

Environment

The depositional environment for the dark brown *Amphipora* subfacies probably was a quiet, shallow lagoon (Fig. 30). Interlamination of *Amphipora* beds with pelletoidal sediments and associated fenestral fabrics suggest deposition on muddy tidal flats. Fenestral fabrics could also have originated during low water levels in the shallow lagoon.

The prevailing dark colour of sediments indicates poorly oxygenated bottom conditions. The light brown *Amphipora* subfacies, which is most common in solitary reef complexes, probably was deposited under somewhat higher energy and better oxygenated conditions. Close association of the light brown *Amphipora* microfacies with fine-grained skeletal calcarenites and sparry calcite cement suggests deposition under medium energy conditions. Lack of primary frame-builders in the *Amphipora* lithofacies suggests that deposition took place some distance from the barrier (Figs. 27, 30). The light brown subfacies appears to have been deposited in better oxygenated shoals from where the fine fraction was winnowed out. Occurrence of algal-coated *Amphipora* skeletons suggests an affinity with the oncolite–intraclast lithofacies.

Oncolite–intraclast Lithofacies

The oncolite–intraclast lithofacies (Fig. 27) is a dark yellowish brown limestone composed of algal nodules and coated intraclasts in a skeletal packstone or grainstone matrix. The oncolites are regularly or irregularly ovoid to rounded, and range from 1 to 3 cm in diameter (Fig. 36a). They are associated with intraclasts of similar size that may or may not be coated (Fig. 37a). The intraclasts are composed of biomicrite, skeletal packstone (Fig. 36b) or pelletoidal limestone and are rounded to well rounded. Some oncolites which have partly preserved algal structures are taken to be of algal origin. Other coated grains and clasts have a homogeneous micritic coating; these are similar to reworked vadose pisolites (Read, 1970, p. 11). Actually, both oncolites and vadose pisolites could have been formed during the development of the marginal reef-bank complex.

The oncolite–intraclast lithofacies is not uniformly distributed and is found mainly in the shelf lagoon (interior platform) of the marginal reef-bank complex (Fig. 31). It is common in the Snipe Lake area, where it composes beds up to 10 feet thick (Fig. 27). Similar facies were described from the Goose River field by Jenik and Lerbekmo (1968, p. 51) as "rubble reef zone," and by Leavitt (1968, p. 333) from the Carson Creek field as a dead reef flat facies.

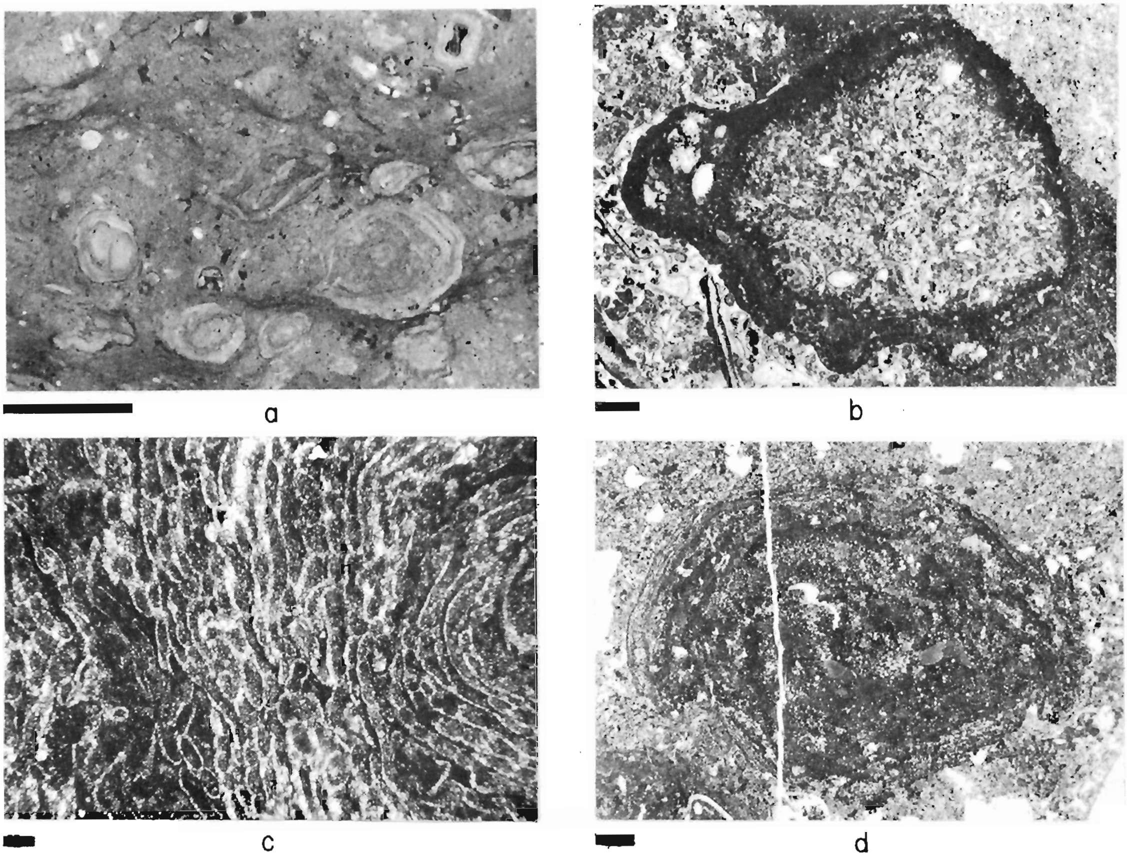


FIGURE 36. a. Oncolites formed by the algae *Girvanella* and *Sphaerocodium* encrusting micrite and crinoid cores in a pelletoidal-skeletal packstone matrix. Note the flattened shape of the oncolite in middle of the picture; oncolite-intraclast lithofacies; Swan Hills Formation. Polished core slab, 4–22–69–18 W5, depth 9,289 ft (approx. 2,830 m). Bar scale is equal to 1 cm.

b. Cross-section of oncolite with a skeletal grainstone core encrusted by red crustose algae. Note the poorly preserved algal texture in the upper part of the oncolite. Matrix consists of a skeletal grainstone with patches of skeletal wackestone at the lower left corner of the photograph; oncolite-intraclast lithofacies; Swan Hills Formation. Thin section—ordinary light. 4–22–69–18 W5, depth 9,372.5 ft (approx. 2,860 m). Bar scale is equal to 1 mm.

c. Oncolite formed by *Sphaerocodium magnum* showing tabular filaments cut in various orientations; oncolite-intraclast lithofacies; Swan Hills Formation. Acetate peel stained with Alizarin Red S. 10–23–71–19 W5, depth 8,750 ft (approx. 2,868 m). Bar scale is equal to 0.1 mm.

d. Oncolite with a sponge-like internal fabric; core formed by poorly preserved cluster of *Girvanella* and fragments of pelecypod shell; internal cavities in oncolite filled with micrite and finely crystalline euhedral dolomite; outer algal rim well developed; note inverted stacked hemispheroid structure of oncolite; matrix of crinoidal-pelletoid wackestone composed of crinoids, gastropods, and brachiopods. Micrite in part replaced by finely crystalline euhedral dolomite; oncolitic intraclast lithofacies; Swan Hills Formation. Thin section—ordinary light. 4–22–69–18 W5, 9,289 ft (approx. 2,830 m). Bar scale is equal to 1 mm.

Composition

Rocks of this lithofacies are composed of up to 40 per cent oncolites which have a maximum diameter of 3 cm (average, 1 cm). The shape of the oncolites is irregularly rounded to circular or elliptical, and rarely flattened. The oncolites have a concentrically laminated structure, with the laminae less than 1 mm thick; they are arranged in spheroid, or less commonly in hemispheroid shells (Fig. 36d). Oncolites with an uneven thickness of laminae have thicker laminae at the upward pointed surface and thin perceptibly at the bottom. The domal shape of some oncolites probably is associated with periods of non-movement when the upper laminae continued to grow.

In some of the laminae original algal structures can be observed. Nodules are composed of algal consortia consisting of *Girvanella, Sphaerocodium* (Fig. 36c), and other skeletal(?) and blue-green(?) algae. Oncolites can be subdivided into two types: those without nucleii, and those with nucleii. Oncolites without nucleii are composed of concentrically laminated, rounded, regular to irregular bodies with some of the nodules flattened. Unbroken laminae in flattened nodules suggest that the nodules may have been somewhat elastic. Nodules are 1 to 1.5 cm in diameter. The surface texture of some nodules is smooth, and the texture of others is coarsely nodular.

In cross-section some of these nodules are composed of two parts: (1) An inner core composed of massive micrite. However, in places "skeletal coralline algal?" structures are apparent, and several cores have spongy fabrics with cavities filled with silt-sized pellets, micrite, sparry calcite, or idiomorphic dolomite crystals; (2) an outer envelope 1 to 2 mm thick which encloses the core and is composed of very fine laminae of *Girvanella,* in places well preserved. Algal nodules originally were highly porous, and some of them have been bored, with borings filled by pellets.

Nucleii of coated intraclasts can be pelecypod fragments, brachiopod fragments, pellets, or intraclast debris (Fig. 36b). Skeletal intraclasts of the wackestone or packstone texture are composed mainly of bank-type biota such as: gastropods, ostracodes, calcispheres, foraminifers, and fine sponge spicules. The coated rim is several millimetres thick and composed of dense brown micrite. No internal structure is apparent. The composition of coated intraclasts is the same as that of uncoated intraclasts.

Environment

Algal oncolites of the Sturgeon–Mitsue area are similar to those described from the Yucatan shelf by Logan *et al.* (1969, p. 98), which are composed of *Lithothamnium–Lithoporella* community. At Yucatan they are present in areas outside the breaking-wave zone, where water movements are strong but more oscillatory than unidirectional.

Mesolella (1968, p. 49) found a horizon of carbonate nodules up to 4 cm in diameter constructed mainly of coralline algae on the reef flats behind the reef-crest at Barbados. Algal nodules also are found in back-reef areas of the Florida reef tract on the surface of some of the biohermal banks. Johnson (1961) presented a summary of data concerning algal growth in the vicinity of reefs and concluded that nodular forms occurred most commonly in the shallow waters of the reef flat and reef front where they are found mainly outside of the breaking-wave zone.

Algal nodules and coated intraclasts in the Swan Hills reef-bank complexes originated on reef-bank shoals presumably on the inner parts of the "reef-bank flats" (Fig. 31). This occurrence is similar to the algal coatings observed in the dead coral zone on recent reef flats by Maxwell (1968, p. 115). Oölites in some of the oncolitic rocks suggest prevailing oscillatory movement of the water on the reef-bank flats. Oncolites without nucleii, and thus porous and easily floatable, could have been transported through tidal channels or passes during storms and deposited in the front of the reef-bank complexes on the reef-bank slopes. Origin of the coatings of many coated intraclasts is uncertain, since the intraclasts consist mainly of dense micrite and lack internal structure. Intensive leaching and the occurrence of coarse-grained, blocky, high-iron calcite cement suggest exposure of the sediment to the vadose cementation zone. Some of the coated intraclasts could have originated as vadose pisolites which were eroded, incorporated into sediment, and concentrated with algal nodules either at the shelf lagoon margin behind the reef flat, or at the front of the reef-bank flats. Uncoated intraclasts probably were produced by erosion of the reef-banks (Fig. 37a).

Skeletal–pelletoidal Lithofacies

The skeletal–pelletoidal lithofacies (Fig. 27) is composed of a mixture of skeletal and nonskeletal particles which range from silt to sand size. Skeletal grains were derived mainly from bank and open-marine organisms: fragments of frame-building biota are absent or very rare. Nonskeletal particles include pellets and composite grains. The skeletal-pelletoidal lithofacies commonly is interbedded with the *Amphipora* lithofacies or the stromatoporoid–*Amphipora* lithofacies. Rocks of the skeletal–pelletoidal lithofacies are medium grey and pale yellowish brown, and are up to 20 feet thick. The skeletal–pelletoidal lithofacies in front of the reef-bank is composed mainly of skeletal debris (Fig. 38a).

In the inner shelf lagoon area and immediately behind the reef-banks, however, skeletal debris is rare (Fig. 38b).

On the basis of petrographic composition and texture the skeletal–pelletoidal lithofacies is subdivided into pelletoidal and skeletal microfacies.

Pelletoidal Microfacies

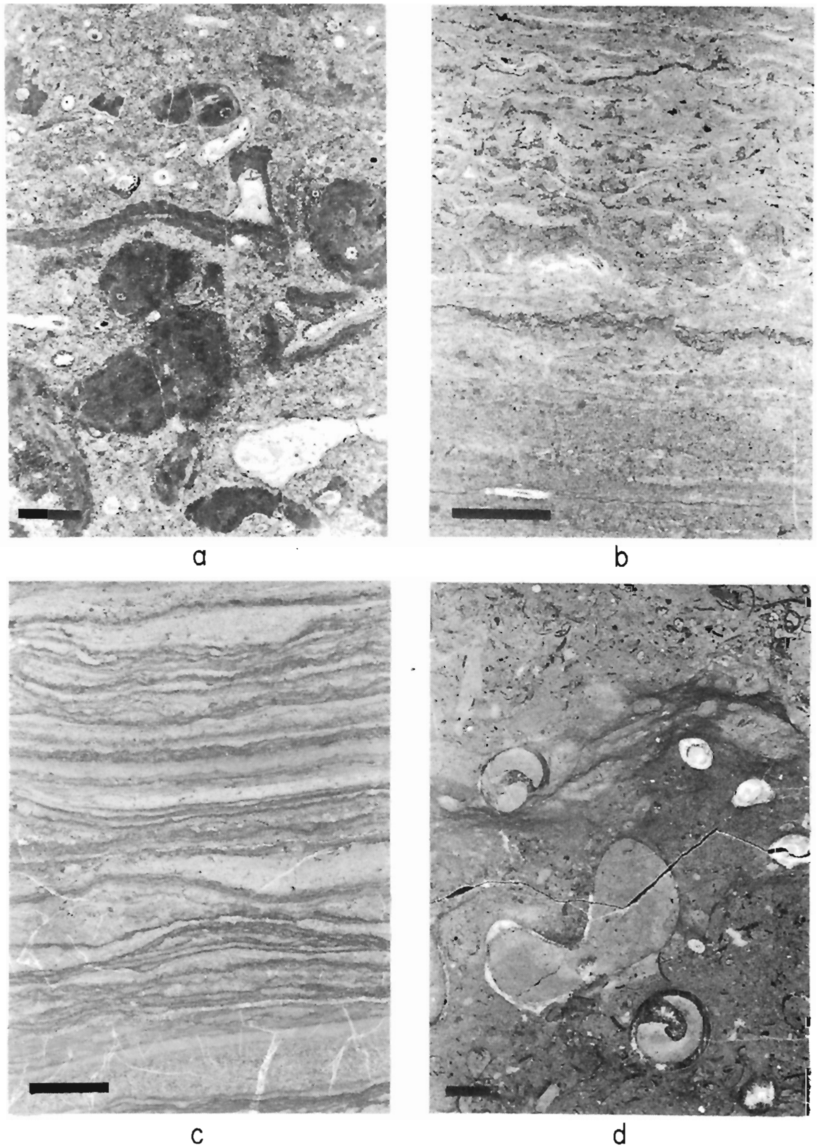
Composition

Pellets (Folk, 1962) are a distinctive feature of this microfacies: they compose up to 80 per cent of total rock volume, consist of micrite, and are elliptical or irregular in shape. Grain size ranges from silt to fine-grained sand size (0.04–0.16 mm). Some of the pellets in this subfacies may be of faecal origin: many, if not most, however, probably are small intraclasts, and accretionary grains.

Composite grains (grapestones, Fig. 38a) compose up to 20 per cent of the sediment and typically range in diameter from 0.4 to 8 mm (average, 0.8 mm). The outer rim of most composite grains has a white chalky appearance, and consists of micrite, which is restricted mainly to grain interstices. The deposition of chalky cement is dominantly a surface phenomenon. The internal voids in composite grains are cemented by acicular calcite, developed at the grain contacts, with the remaining space filled by granular sparry calcite with patches of micrite. The origin of the cement binding pellets into lumps can be explained by chemical precipitation of either aragonite (Illing, 1954, p. 30) and/or magnesian calcite (Taylor and Illing, 1969, p. 69) from sea-derived water.

A minor constituent is skeletal grains which, on the average, compose 12 per cent of the sediments. Skeletal grains consist of fragmented Foraminifera, calcispheres,

FIGURE 37



a. Rounded intraclasts of dark brown biomicrite enclosing *Amphipora*, digitate stromatoporoids, foraminifers and calcispheres in a pelleted-skeletal grainstone matrix; grains in places coated by granular crust of a sparry calcite and remaining space filled with coarse mosaic ferroan sparry calcite; peripheral zone of *Amphipora* skeletons strongly micritized; oncolite-intraclast lithofacies; Swan Hills Formation. Polished core slab, 10–23–71–19 W5, depth 8,731 ft (approx. 2,662 m). Bar scale is equal to 1 cm.

b. Poorly laminated pelletoidal stone, overlain by crinkled algal mats showing fenestral fabrics; fenestra are filled with coarse-grained sparry and ferroan calcite; note isolated skeletons of *Amphipora* and low-amplitude stylolites in lower part of the picture; skeletal-pelletoid lithofacies; Swan Hills Formation. Polished core slab, 10–23–71–19 W5, depth 8,736 ft (approx. 2,663 m). Bar scale is equal to 1 cm.

c. Laminated pelletoidal packstone with each lamina composed of graded pellets capped by aphanitic carbonate; lamination is locally obliterated by solution and development of microstylolites; skeletal-pelletoid lithofacies; Swan Hills Formation. Polished core slab, 10–13–68–16 W5, depth 9,129 ft (approx. 2,783 m). Bar scale is equal to 1 cm.

d. Small and large gastropods, pelecypods, and rare digitate stromatoporoids in a pelletoidal-skeletal wackestone matrix; skeletal-pelletoid lithofacies; Swan Hills Formation. Polished core slab, 10–23–71–19 W5, depth 8,753 ft (approx. 2,667 m). Bar scale is equal to 1 cm.

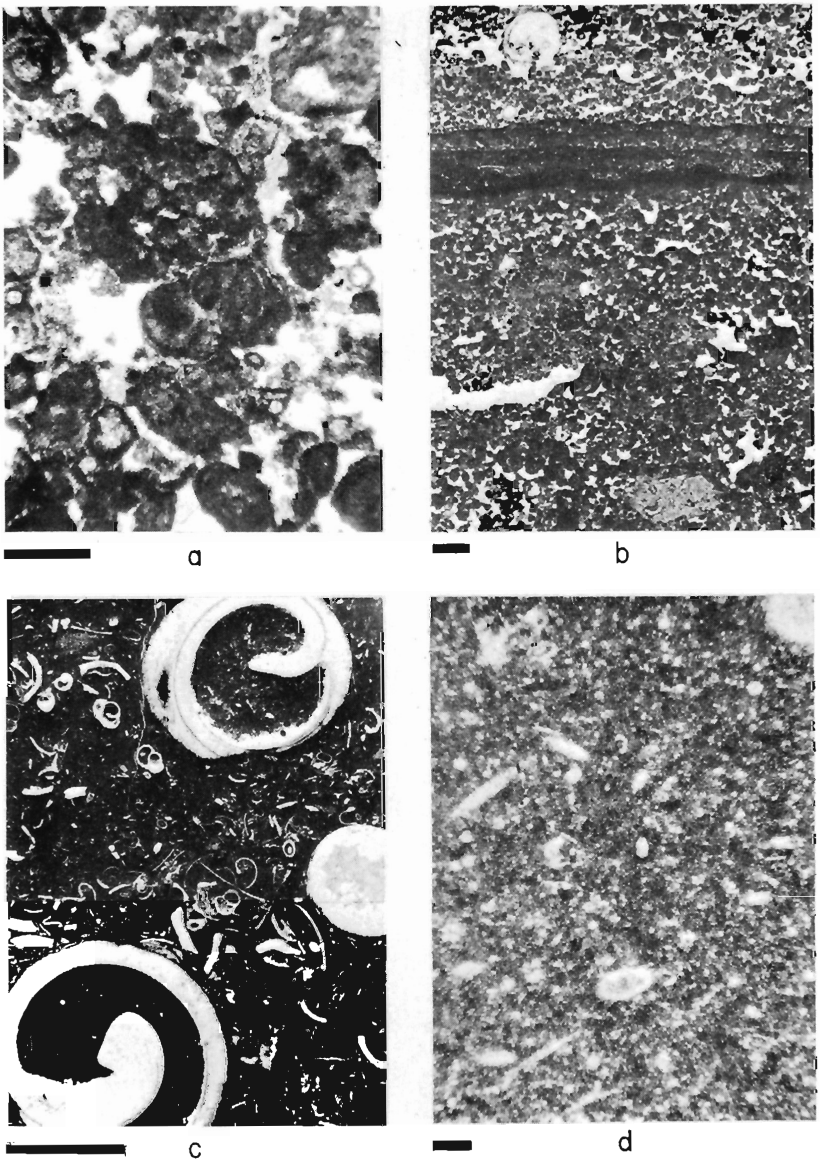


FIGURE 38

a. Composite grains (grapestones) pellets, and calcispheres cemented by coarse-grained, blocky sparry calcite; pelletoidal microfacies; Swan Hills Formation. Thin section—ordinary light. 10–23–8–19 W5, depth 8,739 ft (approx. 2,663 m). Bar scale is equal to 0.5 mm.

b. Micrite crusts in well-sorted pelletoidal grainstone cemented by blocky, ferroan sparry calcite and dolomite; rare abraded *Amphipora* skeletons concentrated in lower part of picture; linear vugs filled with coarse, blocky sparry calcite; crust in upper part of picture consists of multiple layers of micrite; grains between layers cemented with bladed sparry calcite; crust is interpreted as zone of early lithification in the vadose(?) zone; pelletoidal microfacies; Swan Hills Formation. Thin section—ordinary light. 2–5–69–10 W5, depth 8,103 ft (approx. 2,473 m). Bar scale is equal to 1 mm.

c. Poorly sorted gastropod-pelletoidal packstone; cross-section of gastropods shows replacement of wall by coarse sparry calcite; shells replaced by zoned ferroan calcite and blocky, low iron sparry calcite; highly fragmented gastropod shells and merged pellets compose matrix; pelleted nature of rock preserved in internal cavities of gastropods forming geopetal fabric; skeletal microfacies; Swan Hills Formation. Thin section stained by Dickson's (1965) methods, ordinary light. 10–23–71–19 W5, depth 8,753 ft (approx. 2,667 m). Bar scale is equal to 1 cm.

d. Spiculitic-pelletoidal wackestone composed of spicules, silt-sized pellets, and unidentifiable skeletal debris in a micrite matrix; skeletal microfacies; Swan Hills Formation. Thin section—ordinary light. 4–22–69–18 W5, depth 9,309 ft (approx. 2,838 m). Bar scale is equal to 0.1 mm.

ostracodes, pelecypods, and rarely reworked debris of *Amphipora* skeletons. A characteristic constituent of this microfacies are "blue-green" algal mats (Fig. 37b). They are poorly preserved and occur as thin, horizontal laminae, commonly distorted by desiccation and fenestral fabrics, and are similar to the recent buried smooth algal mats on the supratidal flats of western Andros Island (pers. observation—Jansa). In the pelletoidal microfacies dasycladacean and girvanellid algae also are present. Some grains have a thin rim of poorly developed superficial oölitic coating composed of low iron, granular sparry calcite, and in places, of short fibrous calcite. Lithified broken crusts up to 2 mm thick are composed of light grey micrite similar to that of the grapestone rims. Grains in the crusts are cemented by acicular calcite. Coarse-grained drusy mosaic high-iron calcite fills voids, cavities, and moulds which are in places partly filled by anhydrite crystals. In places dolomite replaces calcite and does not exceed 2 per cent of total rock volume.

Sedimentary Structures

The pelletoidal microfacies has a characteristically laminated structure. Small-scale multidirectional crossbedding is rare. Thickness of the individual laminae ranges from 1 mm to 1 cm and individual laminae have sharp basal contacts with positive gradational sorting apparent (Fig. 23c). The basal part of each lamina is composed of coarse silt to fine-grained, sand-sized carbonate material cemented by sparry calcite, and the upper part of each lamina is composed of micrite. Laminar fenestral fabrics are common, with the fenestra elongate, horizontally oriented, and filled with coarse-grained, high iron calcite and dolomite. There are rare occurrences of micritic crusts composed of multiple micrite laminae (Fig. 38b).

Environment

The pelletoidal composition, gradational sorting, lamination, fenestral fabrics, and occurrence of blue-green algal mats strongly suggest deposition in the upper intertidal zone of a semiprotected lagoon. The composition of the pelletoidal–grapestone microfacies is similar to that described by Purdy (1963b, p. 479) from the Andros platform. The higher percentage of micrite in the Swan Hills area, however, suggests a somewhat quieter environment. The abundance of grapestones is evidence of relatively low bottom agitation which allowed the grains sufficient time to become cemented. At Great Bahama bank grapestone facies are found in shallow parts of the bank platform, where in some localities the grapestones are exposed at low tide (Purdy, 1963b, p. 479). The pelletoidal–grapestone microfacies of the Swan Hills area probably was deposited in the tidal zone of shallow banks and shoals of the shelf lagoon. The crossbedded pelletoidal grainstone probably was formed in a turbulent, high-energy environment and would represent a tidal channel deposit. Association of the multiple micrite laminae with fenestral fabrics in Figure 38b suggests cementation in the vadose zone.

Skeletal Microfacies

This microfacies consists of dark yellowish brown gastropod–pelecypod wackestone as well as spiculitic–pelletoidal wackestones (Fig. 27) and is up to 10 feet thick. The shells and shell fragments commonly are unoriented and unsorted.

Composition

Bioclasts compose up to 40 per cent of the rock. Moulds after gastropods constitute up to 20 per cent of the sediment. Two types of gastropods are apparent: those of small diameter (1 mm), and those of large diameter (1 cm). Thick-shelled, fragmented pelecypods compose up to 5 per cent of the rock (Fig. 38c). Other constituents of this microfacies are ostracodes, calcispheres, crinoids, foraminifers, charophytes, spicules (Fig. 38d), and algae. Silt- and sand-sized faecal pellets in places compose up to 30 per cent of the sediment. The dark grey micrite matrix is in places partly dolomitized. Sparry calcite cement is rare and occurs mainly as fillings of moulds after gastropod shells.

Sedimentary Structures

In rare instances the gastropod–pelecypod skeletal packstone microfacies displays vertical organic burrows.

Environment

In the Snipe Lake area the skeletal microfacies is interbedded with the pelletoidal–grapestone microfacies and bulbous stromatoporoid–*Amphipora* beds. This facies represents beds deposited somewhat below wave-base in a lagoon or open-bank environment, or may represent tidal channel fill. Similar recent deposits occur in an off-bank environment of the Rodriguez Key, Florida (pers. observation—Jansa), in water depth not exceeding 10 feet (3 m).

Discussion

Depositional History of the Marginal Reef-bank Complexes

The present-day configuration of the Swan Hills reef-bank complex (Fig. 27) is a product of several phases of development, each of which accumulated and was modified under a wide range of environmental conditions. Paleotopography, character of substrate, current direction and intensity, wind direction, temperature, clarity of water, salinity, tidal range, and nutrient content acted as modifying factors which are not all obviously reflected in the stratal record. One object of this study is to gain an understanding of the conditions under which reef-banks are started.

Fischbuch (1968) divided the Swan Hills reef complexes into nine vertically separable rock-stratigraphic units (Divisions I to IX) which represent successive stages of reef development, each division being composed of several distinct facies. This divisional breakdown is followed in this paper with the exception of the definition of Division I. In this study Division I represents only the coral zone and its time equivalents. The basal carbonate blanket is defined as a separate lithological unit. The marginal reef-bank complexes as defined here consist of Divisions I to IV inclusive (Fig. 39). Each division represents a single stage of reef-bank growth with evidence of shallowing in Divisions II and IV. Subaerial and submarine erosion appears to have taken place and modified the original reef-bank profile.

Division 1

Division I ranges from 10 to 20 feet (3 to 6 m) in thickness, is the most widespread unit in the Swan Hills marginal reef-bank complexes, and covers much of the

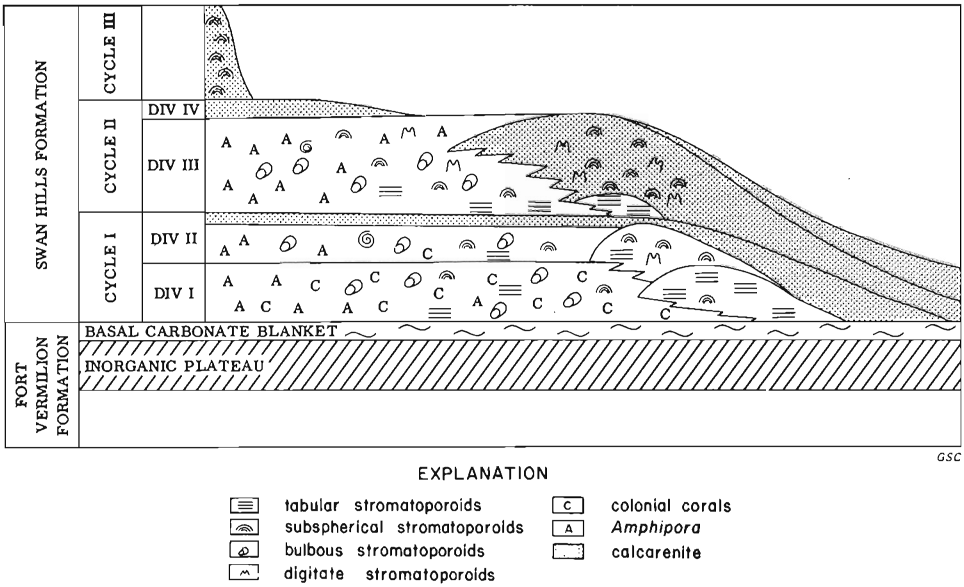


FIGURE 39. Schematic diagram portraying evolution of the edge of the marginal reef-bank complex of the Swan Hills Formation.

surface area of the incipient carbonate platform (basal carbonate blanket) (Fig. 26). The first bioherms, which were established in Division I time, modified the shallow open-bank environment and created an intricate series of microenvironments that eventually formed a broad marginal reef-bank complex. Without the modifying influence of the biohermal mounds the area would probably have continued as a biostromal and non-skeletal bank similar to the underlying basal carbonate blanket.

This division consists of three major lithofacies: a tubular stromatoporoid lithofacies, most common near the outer edge of the reef-bank complex; a coral lithofacies mainly in the inner shelf lagoon; and an *Amphipora* lithofacies containing *Amphipora* interbedded with pelletoidal packstones in the most distal western edge of the shelf lagoon (Figs. 27, 39).

Development began with tabular stromatoporoid colonization of the upper surface of the basal carbonate blanket. These frame-building organisms were localized on the easternmost bank slopes and edges of inlets and channels which occurred in the path of the prevailing winds and waves. Low-relief bioherms developed and were composed predominantly of overgrowing, interlocking, and encrusting tabular stromatoporoids. Dead colonies provided ample substratal foundations for new colonization and blue-green algae appear to have helped stabilize the loose carbonate sand between the stromatoporoid colonies. These tabular stromatoporoid mounds probably were wave resistant to a certain extent, since they appear to have originated in subtidal waters and may have grown to the upper subtidal zone (*see* p. 60). Surfaces of the tabular stromatoporoid colonies, however, do not appear to have been exposed even at low tide; these mounds probably had a maximum relief of not more than 20 feet (6 m). In gross form the structures probably formed discontinuous elongate ridges in a zone up to 2 miles (3.2 km) wide, which was completely submerged. Although these stromatoporoid mounds were of relatively low profile, they appear to have been capable of dissipating and absorbing a certain amount of wave energy, and thus made subtle modifications to the environment of adjacent areas. The degree of modification brought about by any particular barrier was dependent on many factors such as wave and current intensity, water depth, and the physiography of the barrier (Logan *et al.*, 1969, p. 187). The degree of modification of the depositional environment in most of the reef-bank complex suggests that the ridges consisted of a series of mounds that locally coalesced to form a discontinuous chain up to several miles long. These were situated at the windward edge of the reef-bank complex as submerged barriers forming zones up to 12 miles (19.3 km) or more long and 1 to 2 miles wide.

Semi-protected shelf-lagoon environments appear to have developed along the leeward side of the mounds. The depth of these lagoons corresponds to the height of the mound and the paleotopography of the reef-bank complex, and probably was from 20 to 30 feet (6 to 9 m). Shelf lagoon bottoms were colonized by colonial corals (*Thamnopora, Coenites*). Primary frame-builders such as tabular, digitate, and bulbous stromatoporoids formed the biohermal mounds. Since these tabular stromatoporoid bioherms at no time were exposed, there was free exchange of open-marine and lagoonal waters. Suitable depth conditions, protection against strong waves, and free exchange of sea water probably were the main factors for extensive coral growth in this time. A higher degree of restriction at the end of Division I time led to the decrease of water circulation and dissolved oxygen supply which, combined with other factors such as carbonate mud deposition over the shelf lagoon floor, resulted in the disappearance of the colonial corals. Colonial corals are not found in subsequent stages of reef development. Coral zone deposits of the Swan Hills Formation can be compared with those of recent windward lagoons of the Bahama Bank (Newell and Rigby, 1957) and a similar windward position is suggested for the coral lithofacies of Division I. At the end of Division I time the growth of tabular stromatoporoid mounds ceased. The surface of the mounds at this stage could be comparable to the recent dead surfaces of the Molasse Reef of the outer Florida reef tract where the dominant process is reef erosion.

Division II

Division II overlies Division I with gradational contact, but areally is less extensive (*see* Fig. 26). Its thickness, as defined in this study, does not exceed 40 feet (12 m). Although Fischbuch (1968, p. 478) found it to be 110 feet (approx. 33 m) thick at Virginia Hills, there could be some misinterpretation since only limited core data were available at Virginia Hills. Division II differs from Division I in that it lacks distinctive corals; only rare thick-walled solitary corals (*Grypophyllum* n.sp., Pedder, 1971) are apparent in Division II.

Facies are poorly developed and not easily defined. Subspherical stromatoporoids commonly overlie the tabular stromatoporoid ridges, and appear to have migrated slightly basinward where they are interbedded with off-reef skeletal calcarenites (Fig. 27). The beginning of Division II is characterized by accelerated accumulation of a skeletal silt and sand-sized material around the tabular stromatoporoid ridges. Skeletal debris was produced by disintegration of the biohermal biota of the reefs and supplemented by carbonate particles produced in the shelf lagoon with a minor influence of planktonic open-shelf biota. Skeletal particles were transported by wave- and tide-generated currents across the shelf lagoon and accumulated in the vicinity of the biohermal ridges (incipient reefs). The prevailing onshore water movement kept the calcareous particles close to the ridges, and did not allow them to be transported into the deeper offshore environment. The hydrodynamic conditions at this time appear to have been similar to those producing recent carbonate sandbars (Ball, 1967), where biohermal bodies in the cores of the "bars" stabilize the structures and prevent lateral migration. The developing marginal reef-bank slopes were subsequently colonized by subspherical stromatoporoids and by digitate stromatoporoids on the upper and lower reef-bank slopes respectively. Bulbous stromatoporoids appear to have flourished on the shallow slopes on the protected sides of the reef banks and *Amphipora* with rare digitate stromatoporoids in the deeper more restricted lagoons west of Snipe Lake. Development of the reef-bank slopes was, therefore, a process of an interaction between hydrodynamic-mechanical processes (detrital carbonate transportation and accumulation) and biological (bioherm forming) processes.

During Division II time the inlets and channels probably were narrowed and modified by further tabular and subspherical stromatoporoid growth along the reef-bank complex margins. Currents would have prevented deposition in the channels and inlets through most of the time of reef-bank complex development.

The close of Division II time is marked by a distinct lithological zone that is up to 10 feet (3 m) thick and consists mainly of pellets, grapestones, superficial oölites, and oncolites (Figs. 27, 39). At this time accumulation of sediments in the shelf lagoon exceeded upward growth of the marginal reef-bank front. Fenestral fabrics, leaching, erosion, micrite crusts, and coarse, blocky, high-iron calcite cement suggest intermittent exposure of the surface in some areas. A bank type of environment probably prevailed over the whole area for short periods of time, and marked the end of this first cycle of reef-bank development. At this time most of the Swan Hills area was a broad shallow platform with extensive tidal flats and some supratidal areas where chemical precipitation of carbonates was dominant over biological production. This type of environment is comparable to present-day sedimentation on the Bahama Banks.

Division III

The areal extent of Division III again is somewhat less than that of Division II (Fischbuch, 1968, and Fig. 26). The upper boundary is a sharply truncated erosional surface, which is marked in some places by a distinct 6- to 8-inch-thick (10 to 20 cm) bed of dark grey open-marine shale; this shale marks the beginning of the Waterways transgression. Thickness of this division is 40 to 60 feet (12 to 18 m) with a maximum of 70 feet (21 m) in the Kaybob area (Fischbuch, 1968, p. 482). Division III marks the first appearance of well-separated facies and major reef-bank environments (Figs. 27, 39). Reef-bank slope clastics are a volumetrically important constituent of Division III with some calcarenites up to 20 feet (6 m) thick. The calcarenite is interbedded with fragmented subspherical stromatoporoids on the reef-bank front slope in a manner similar to that of Division II, and extends partly into the upper lagoonward reef-bank slope. The lower lagoonward bank slope was populated by bulbous and digitate stromatoporoids and progressively grades into *Amphipora*-rich beds toward the lagoon. Algal-coated, gravel-sized bioclasts were deposited near the top of Division III on the inner reef-bank flats and could represent a rubble zone as was suggested by Jenik and Lerbekmo (1968, p. 51). The surface is truncated at the Swan Hills field as well as in the Snipe Lake area by an erosional unconformity (Fig. 40a).

The erosion is apparent only on elevated parts of the marginal reef-bank complex. In the channels and inlets cut into and through the marginal reef-bank complex, erosional surfaces are not apparent. This suggests that if the reef-bank complex was subaerially exposed, the exposure did not affect the deeper channels.

Erosional surfaces, however, can be produced by both subaerial and submarine erosion. The basal 6 inches (15 cm) of the overlying Waterways Formation contains crinoids, brachiopods, and glauconite, and appears to have been highly leached (Fig. 40b). Fossil moulds are filled with coarse-grained, high-iron sparry calcite and iron-rich dolomite. They are cemented by microcrystalline calcite with granular crusts around skeletal particles, with the remaining space filled by coarse-grained, blocky, mosaic calcite. In other places the upper few inches of Division III has voids filled by patches of anhydrite. The types of carbonate cements near the surface of Division III suggest that cementation occurred in the meteoric vadose zone (Land, 1970, p. 184), which, together with anhydrite, could mean that the surface was in some places subaerially exposed at the beginning of Waterways time.

Division IV

Severe erosion of the surface of the marginal reef-bank complexes, following the accumulation of Division III sediments, produced a considerable amount of calcarenite which appears to have been deposited in the low-lying areas of Division III topography (Fig. 39). In some places (in the Snipe Lake area) redeposited vadose pisolites commonly are found in the calcarenites. The thickness of the calcarenites ranges from 15 to 30 feet (4.5 to 9 m). Fischbuch (1968, p. 486) included these calcarenites in Division IV. Genetically this unit can be taken as the upper part of Division III, and is similar to the calcarenite beds that lie at the top of Division II. The greater thickness of calcarenite beds probably is the result of a longer period of erosion following Division III time. The end of deposition of Division IV sediments marks the end of the second and final cycle of marginal reef-bank complex development.

Subsequently the reef-bank complex was covered by the transgressing Waterways Sea and the channels and inlets of the marginal reef-bank complex were filled with open-marine shelf carbonates. At this time, however, further Swan Hills reef-bank growth continued in the form of solitary reef complexes which are found at Swan Hills, Judy Creek, Carson Creek, Virginia Hills, Kaybob, Goose River, and Snipe Lake; these complexes have been described in detail by Edie (1961), Murray (1965, 1966), Jenik and Lerbekmo (1968), Fischbuch (1962, 1968), Leavitt (1968) and others, and, therefore, were not considered in detail in this study.

Some authors have used colour as a criterion to distinguish marginal reef-banks from solitary reef complexes (dark brown and light brown respectively); this is not an objective manner of separation and was not used in this study. Colour of sediments is related to the degree of aeration and oxidation. Well-oxygenated sediments in the marginal reef-bank complex are the same light brown as the well-oxygenated sediments of the solitary reef complexes. The greater oxygenation of sediments in the solitary reef complexes appears to have been related to the smaller size of these structures, whereas, the underlying marginal reef-bank complex was well aerated only on the open-marine edge.

Factors Controlling Development of Marginal Reef-bank Complex

From the study of recent reefs it has been established that reef growth and reef shape depend on the initial development of reef organisms and their subsequent nourishment in an environment where bathymetric conditions permit their expansion. Factors controlling expansion include availability of nutrient-rich water, oxygenation, salinity, suitability of temperature, and amount of light penetration. These factors, consequently, depend on geographical latitude and water circulation.

Bathymetric Conditions

Bathymetric conditions are related to the paleotopography of the sea bottom and they are one of the main controlling factors in reef-initiation. Some workers believe that large-scale basement tectonic features controlled the initiation and configuration of reef growth in the Swan Hills area (Martin, 1967, p. 370; Keith, 1970, p. 854). Evidence obtained during this study, however, indicates that it was the configuration and sedimentary history of the underlying Gilwood delta complex and the Fort Vermilion evaporites which played a major role in the initiation of Swan Hills reef-banks and controlled their location. Nevertheless, such linear trends as South Kaybob could reflect subtle influences of underlying structural displacements.

Reefs founded on pre-existing topographic highs that project above the floor of the surrounding shelf are the rule for most oceanic as well as shelf reefs. In the West Indian province the Andros barrier reef is developed as a skeletal capping over a ridge-like mass of oölitic limestone (Newell and Rigby, 1957, p. 36). Reef-building in the British Honduras is, according to Purdy (1965, p. 157), also controlled by topographic highs, as are the Yucatan shelf reefs which are localized on cemented eolianite dunes (Logan *et al.*, 1969, p. 134). Fringing reefs of the Barbados follow the surface of a pre-existing Tertiary foundation (Mesolella *et al.*, 1970, p. 1901) and reefs of Mahe, Seychelles (Lewis, 1967, p. 140) are examples of reef capping on an irregularity of volcanic origin. The Swan Hills marginal reef-bank complex similarly appears to be localized on the drowned topographic highs of a delta platform complex (Fig. 41, *in pocket*).

In the Sturgeon–Mitsue area the Swan Hills Formation was deposited over a basal inorganic plateau which reflected the paleotopography of the previously deposited Gilwood delta complex (Fig. 41). The inorganic plateau had local relief which was probably in the order of 30 feet (9 m) (Fig. 25). Ridges and plateaux corresponded to the areas of thick sandstone accumulation along delta channel belts and over transverse sandstone ridges which developed during the destructional phases of deltaic sedimentation (Fig. 6). Differential compaction and pre-Swan Hills erosion enhanced the topographic reflection of the sandstone bodies. The Swan Hills transgression flooded the inorganic plateau, and water depth over the highest parts probably did not exceed 5 feet (1.5 m). This assumption can be made since the shoal area of the overlying basal carbonate blanket consists of oölites and pellets.

During the time of deposition the inorganic plateau probably was tilted slightly to the east; in the Sturgeon Lake area water depth over the inorganic plateau probably was near zero. On the eastern edge (Mitsue area) where biomicrites were deposited, water depth could have been as much as 100 feet (30 m).

Along the eastern margin of the shelf (Mitsue area) the surface of the inorganic plateau rose approximately 80 feet (24 m) above the surrounding sea bottom.

This relief can be estimated by the difference between the thickness of the terrigenous deposits of the delta complex and the estimated thickness of the synchronous shaly deposits in the basin. A thin veneer of nonskeletal carbonates and a mixture of skeletal and nonskeletal carbonates were deposited over the inorganic plateau (basal carbonate blanket) and mark the incipient stage of the carbonate platform sedimentation. After carbonate deposition was established in the area the topographic highs were colonized by frame-building biota, and subsequently tabular stromatoporoid bioherms were formed.

The Effect of Wind and Waves

Although the initiation of any given reef is subject to existing ecologic conditions and its location is determined by geomorphological factors, the most striking control of reef development is exerted by winds and currents (Fairbridge, 1950, p. 356).

Little information is available concerning the physical processes active during Devonian time. Some clues, however, can be obtained by considering the sedimentation patterns of the underlying formations, provided they are also representative of the wind and current regime of Middle Devonian time. Most conclusions, nevertheless, must be inferred from comparison of Middle Devonian sediments with modern reef settings. In present-day reefs oxygen and other nutrients are brought to the frame-building biota from the windward side. Hence growth is rapid and tends to be compact. On the leeward side of present-day reefs Kuenen (1950, p. 440) found conditions to be less favourable due to a dearth of food and oxygen and the greater amounts of sediment (Wells, 1957, p. 621). As a result the organisms assume branching, delicate growth forms and are widely spaced. In the Swan Hills area most of the massive tabular stromatoporoid mounds and ridges are on the northeastern sloping sides of the marginal reef-bank complex.

Wind direction during the deposition of the underlying Gilwood Member appears to have been from the north–northeast as derived from the orientation of the sand barrier (Mitsue area). In the succeeding evaporites of the Fort Vermilion Formation the lack of quartz grains from the alluvial plain to the west also suggests a landward direction for the prevailing winds. At the time of reef-bank initiation the coastal plain sediments consisted mainly of quartz sand deposits which were exposed about 50 miles (80 km) west of Snipe Lake; surprisingly no quartz grains are present in the marginal reef-bank carbonates. This further suggests a continuation of an onshore wind direction throughout Swan Hills time. The prevailing wind direction in the Sturgeon–Mitsue area appears to have been from a general northeasterly direction perpendicular to the elongation of the marginal reef-banks (Fig. 41).

Onshore winds are important in the successful development of reefs. Fairbridge (1950, p. 354) demonstrated that coral larvae are transported to the shoreline by onshore winds, where they are able to establish new colonies. Where offshore winds prevail, the larvae are transported to the open sea. Stromatoporoids may also have had free swimming juvenile medusae, and onshore winds would have had favourable effects on the establishment and expansion of stromatoporoid colonies.

Wind also is an important factor in the generation of waves and currents. Waves not only are a controlling factor in the configuration of the reef, but also play an important role in sediment distribution. The Swan Hills marginal reef-bank complex appears to consist of shelf-type "reefs" developed near the border of the Peace River landmass in a shallow epicontinental sea. This sea was devoid of reef knolls or island chains and the wind could generate waves easily. Strength of the waves and wind-generated currents can be estimated from the grain-size distribution of the sediments and character of the reef flat and reef slope deposits. Slope deposits of the marginal reef-bank complex are composed mainly of sand-sized skeletal debris, but gravel- and pebble-sized skeletal fragments also are present. The reef-bank structures were flat topped and the upper reef-bank slope did not have algal-coated surfaces typical of present-day high-energy oceanic reefs; this, together with grain-size distribution, suggests that the marginal reef-bank complex was not subjected to giant rollers of the type that impinge on the fronts of present-day Pacific and Indian Ocean reefs. Consequently, wave energy was comparatively low during deposition of the Swan Hills marginal reef-bank complex. Recent examples of reefs influenced by low-energy waves are those of the Yucatan shelf (Logan *et al.*, 1969, p. 134) which have grain-size distribution similar to that of the Swan Hills marginal reef-bank complex.

Often it is difficult to separate currents produced by wind from those of tidal origin (Bird, 1969, p. 22); consequently, it is impossible to separate them in the fossil record. Wind- and tide-produced currents, therefore, are referred to in this study only as surface currents. Velocity of the surface currents at the time of marginal reef-bank development can be estimated from the size of debris transported through channels and passes within the marginal reef-bank complex. As the passes and channels that intersected the complex were narrowed by the encroachment of fringing bioherms and biohermal ridges, current velocity would increase substantially. The floors of the channels and passes in the Swan Hills marginal reef-bank complex probably were places of nondeposition during marginal reef-bank sedimentation (Fig. 27). Minimum velocity of these currents can be estimated by considering that the largest frame-builders available for transport were subspherical stromatoporoid heads which have an average diameter of from 10 to 15 cm. To transport material of this size through the channels would require, at least at times, a current velocity of 150 to 200 cm/sec. (Nevin, 1946; Kuenen, 1950, p. 260).

Wells (1957, p. 611) too has shown that currents are important factors in the development of reefs since reefs do not flourish at sites where circulation does not maintain equitable temperatures and nutrients. Near-surface nutrients are most abundant in areas that have access to nutrient-replenished waters rising from oceanic depths. Depth of the Elk Point basin in the vicinity of the marginal reef-bank complex probably did not exceed several hundred feet, however, the location of stromatoporoid mounds along the outer edges of the complex (Fig. 41) would suggest onshore water movement at that time.

Recent corals are very sensitive to salinity changes and find optimum growing conditions in waters of normal salinity. Near-normal saline conditions probably were required by stromatoporoids, since stromatoporoid colonies rarely are found in growth position in restricted environments where salinity would be higher than normal. If one can assume that there were shoreward currents along the outer margins of the marginal reef-bank complex, then it can be further concluded that sea water in the shallow Elk Point Basin could not have been density stratified with hypersaline brines in the bottom layer at the time of the Swan Hills marginal reef-bank complex development.

The Effect of Tides

Currents produced by the ebb and flow of tides have a velocity related to the tide range. In the Swan Hills marginal reef-bank complex, although sediments of the upper reef-bank slope were deposited under relatively medium- to high-energy conditions, sedimentary features suggesting exposure above sea level were not found. Exceptions, however, may be the upper surfaces of individual cycles, where the reef-bank had been exposed for short periods of time. It is assumed that the tidal range during Swan Hills time was in the order of several feet. This assumption is based on lithological and petrographic similarities between the Swan Hills and Alacran reef on the Yucatan peninsula, where the tidal range is ±1 foot (Hoskin, 1963).

Temperature and Salinity

Climatic conditions during Middle Devonian time in Alberta can be assumed from evaporite deposition which preceded Swan Hills sedimentation, and the climate probably was hot and arid similar to the present-day Persian Gulf. In the Abu Dhabi area, offshore fringing reefs are separated by sheltered saline lagoons from supratidal sabkha deposits which are deposited at an average temperature of 30°C. A hot climate is characteristic for most of the recent hermatypic coral reefs, and the temperature ranges from 24° to 30°C (Logan *et al.*, 1969, p. 156; Maxwell, 1968, p. 90; Purdy, 1963a, p. 338; Evans, 1966, p. 689). The association of evaporites and red beds with some Devonian reefs of western Canada further corroborates the concept that temperature of the Devonian seas in the study area was similar to the climatic conditions of the Persian Gulf.

Surface salinity of the sea water in recent coral reefs is close to that of normal sea water. Salinity is slightly higher on Andros Island (Purdy, *loc. cit.*) and in the Persian Gulf (Evans, *loc cit.*) where it ranges from 36 to 38‰. As previously mentioned, stromatoporoids probably grew in sea water of near-normal salinity. Salinity of the surface waters in the Swan Hills area is assumed to have been similar to that of the Persian Gulf. Evaporites, found replacing some frame-builders and filling cavities, are related to time of exposure and could be, in part, of late diagenetic origin.

Initiation of the Marginal Reef-Banks (Summary)

The inorganic plateau appears to have had a relief of up to 30 feet (9 m) corresponding to the depositional pattern of the underlying clastic delta. It was flooded by a shallow Middle Devonian sea of near-normal salinity during a time of subtropical, semiarid climate. Northeasterly prevailing onshore winds generated waves and currents and caused movement of the surface water toward shore (Fig. 41). The waves struck the shoreline at an angle and produced longshore currents that travelled in a southerly direction. The eastward-inclined slopes of the elevated inorganic plateau faced the waves and currents of the open sea and, following incipient carbonate blanket deposition, were colonized by colonial corals and tabular stromatoporoids. Areas of accelerated organic colonization by biohermal type of biota corresponded to the offshore slopes of the distributary mouth bars and elongate finger-like sand bodies of the underlying Gilwood delta complex. The tabular stromatoporoids which colonized the upper part of the slopes built biohermal mounds up to 20 feet (6 m) high which later developed into zones of elongate ridges or low barriers 6 to 12 miles long and 1 to 2 miles wide.

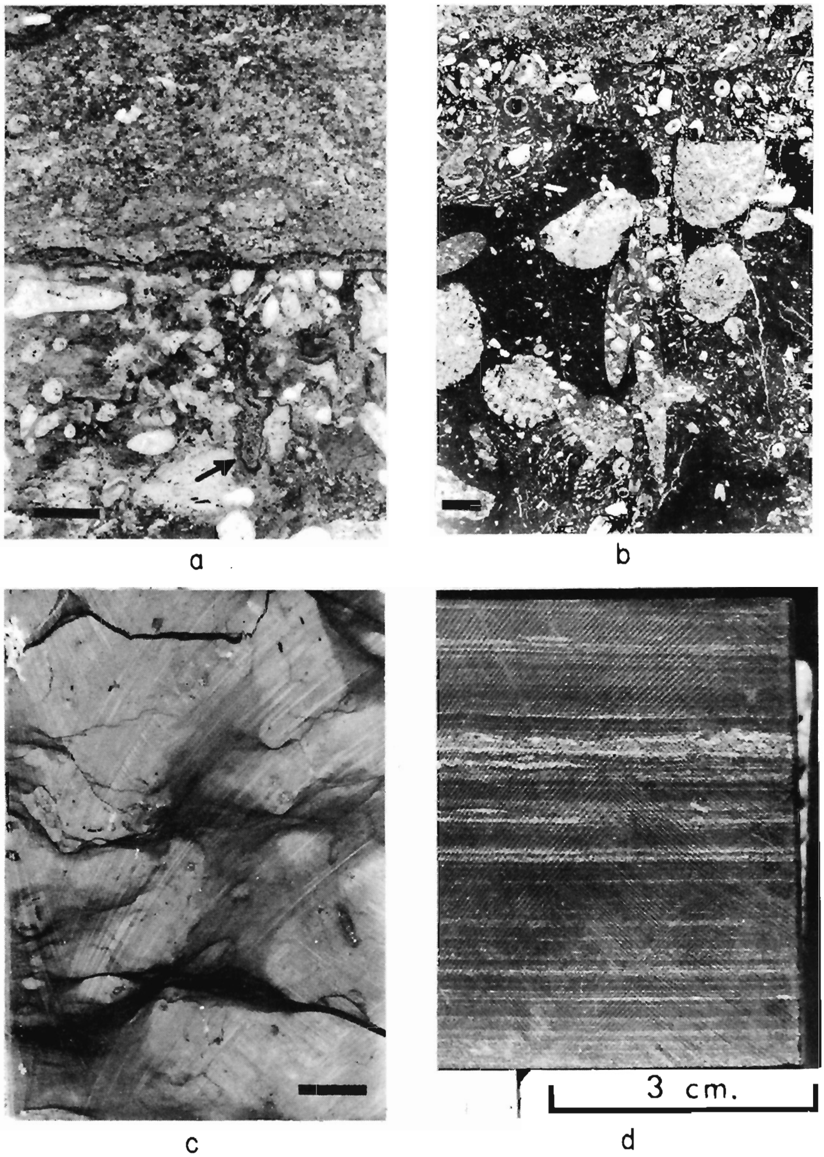


FIGURE 40

a. Erosional contact of marginal reef-bank complex with overlying Waterways Formation; note truncated digitate stromatoporoids at contact, thin crust overlying the erosion surface and borings or solution pipes into upper surface of the reef-bank. Waterways Formation is composed of crinoid-pelecypod packstone, with rare *Amphipora* debris; Swan Hills Formation. Polished core slab, 2–6–70–10 W5, depth 7,468 ft (approx. 2,277 m). Bar scale is equal to 1 cm.
b. Erosional contact of Swan Hills marginal reef-bank complex with overlying Waterways Formation; note truncation of digitate stromatoporoid at contact and intensive bioturbation; voids after borings and burrows filled with Waterways-type sediment; base of Waterways Formation is delineated by crinoid-pelecypod packstone, cemented with coarse blocky, ferroan sparry calcite; Swan Hills Formation. Thin section—ordinary light. 2–6–70–10 W5, depth 7,468 ft (approx. 2,277 m). Bar scale is equal to 1 mm.
c. Nodular fossiliferous micrite; Waterways Formation. Polished core slab, 4–7–70–10 W5, depth 7,624 ft (approx. 2,324 m). Bar scale is equal to 1 cm.
d. Horizontally laminated, grey coloured argillaceous, micritic limestone; Waterways Formation. Polished core slab, 10–13–68–16 W5, depth 9,032 ft (approx. 2,753 m).

These ridges were discontinuous and intersected by tidal channels and inlets and were located on the outer margin of the inorganic plateau. The interaction of bioherm- and bank-forming processes (Fig. 27) formed the marginal reef-bank complex (House Mountain, etc.), upon which isolated reef complexes (Swan Hills–Carson Creek) later developed. To the west, similar marginal reef-bank development took place on the windward side of channels, passes and re-entrants that were cut into the inorganic plateau, upon which other isolated reef complexes (Virginia Hills–Snipe Lake) later grew.

Cyclicity of the Marginal Reef-Bank Complex under Conditions of Continuous Transgression

The presence of well-defined units, which are repeated in a cyclic manner, is one of the most notable characteristics of the Swan Hills marginal reef-bank complex. A cyclic pattern of the Swan Hills deposition was noted by Carozzi (1961, p. 497) who divided the Swan Hills Formation into seven rhythms. Five cycles were recognized in this study, of which two form the marginal reef-bank complex (Fig. 39), and the remaining three form the solitary reef complexes. Thickness of the individual cycles is related to the position of the cycle in the reef-bank complex, and averages about 45 feet (approx. 13.5 m).

The cycles generally consist of deeper water sediments at the base with gradation upward into shallower water deposits, suggesting that these cycles originated by "regression" of the sea. The deeper water part of each cycle appears to have developed in a shallow subtidal environment (tabular stromatoporoid lithofacies, biomicrites of the open-marine shallow environment deposited on the outer shelf lagoon, or biomicrites deposited in a relatively deep lagoonal environment of the inner shelf lagoon). These beds in turn grade into beds representing a shallow reef-bank or tidal flat environment (subspherical stromatoporoid lithofacies, oncolitic–intraclast lithofacies, and skeletal–pelletoidal lithofacies). The lower part of each cycle is composed mainly of skeletal carbonates, and the upper part of predominantly nonskeletal material such as oölites, superficial oölites, grapestones, and pellets. The top of each cycle commonly is marked by the occurrence of fenestral fabrics, leaching, solution features, erosional contacts, vadose pisolites, and micrite crusts. Each cycle represents a vertical accretion of the marginal reef-bank complex by an interaction of biological and hydrodynamic processes.

Mechanisms controlling cyclic development can be related to the sea level fluctuation, caused by local tectonic processes as the Peace River Arch subsided (de Mille, 1958), or could have been the result of regional tectonic or climatic changes. Pulse-like periods of subsidence which were followed by times of standstill allowed the reef-building organisms sufficient time to grow up into the high-energy zone. A new period of submergence associated with a rise in sea level would initiate a new cycle. The slight areal restriction of each cycle can be explained by a greater amount of submergence than the organic development could tolerate.

A third theory suggested for cyclic development of the Swan Hills marginal reef-bank complex is based on the interrelation of organic and hydrodynamic processes that would have been in equilibrium with each other during a period of continuous transgression (or continuous subsidence). Requirements for this model are location on a shallow continental shelf or in an epeiric sea, with a slightly tilted sea bottom in an offshore direction, and the presence of barriers, either morphologic (bioherm, reef, bar, or bottom elevation) or hydrodynamic (onshore currents and waves). Applying this model to the Swan Hills carbonate sedimentation, each cyle would have started in a subtidal environment followed by the construction of a low barrier (Fig. 42). Shelf lagoons developed on the interior of these barriers. Onshore wind-driven waters would have deposited part of the fine-grained sediments generated in the shelf lagoon on the tidal flats. Part of the lagoon-produced sediments would be transported by tidal and wind generated currents to the barrier and deposited in the vicinity of the marginal bioherms.

Onshore water movement at the barrier front combined with the effect of a shallow sea at the reef-bank complex front retained the carbonate detritus, produced by fragmentation of biohermal biota and lagoonally derived carbonate particles, in the vicinity of the reef-banks. Part of the sediments accumulated at the reef-bank front would have been redeposited, from time to time, as washover on the back slopes of the reef-banks. Because of the combined "sealing" effect of hydrodynamic conditions and shallow sea bottom, the reef-bank complex and the shelf lagoon would be operating as a closed system which would not allow the carbonate particles, generated in the shelf lagoon and marginal reef-banks, to escape to the open-shelf area. The off reef-bank slopes were built upward by the accumulation of skeletal as well as nonskeletal carbonates. The over production of carbonate material relative to the rise of sea level expanded the reef-bank slopes laterally by sediment accretion, thus forming a carbonate platform. Width of these platforms in the Swan Hills area ranges from about 2 to 12 miles (approx. 3 to 19 km) across (Fig. 27). At this stage the Swan Hills carbonate platforms were comparable to the mature platform reefs of Wells (1957, p. 625). Due to very shallow water conditions over the carbonate platform

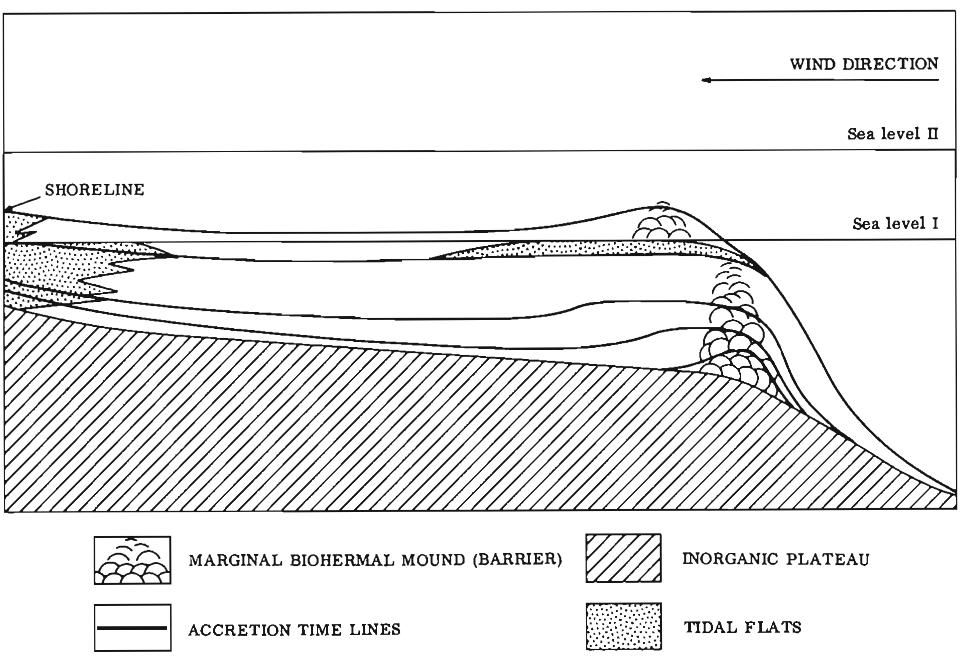


FIGURE 42. Schematic model of a cyclic development of the marginal reef-bank complex by sediment accretion in conditions of continuous, slow transgression.

surfaces, tidal conditions prevailed over most of their surface area. The tidal environment did not provide favourable conditions for growth of reef-builders, and only on the platform margins was there an appreciable growth of the frame-building biota. This period is marked by deposition of mainly chemically precipitated carbonates (bank stage) over the carbonate platform surface. Pellets, oölites, algal coated grains, oncolites, leaching, fenestral fabrics, micritic crusts (formed in the vadose zone of cementation) and anhydrite, represent deposits of this stage. The large quantity of shifting sand-sized particles deposited around the bioherms, the more saline waters from the platform flats, the higher turbidity of the water, or the combined effect of these factors diminished or stopped the growth of bioherms.

Decrease in skeletal carbonate production inhibited vertical reef-bank growth, and associated chemogenic carbonate precipitation could no longer keep pace with the rising waters. The area was slowly flooded by the sea; subsequently, favourable conditions for growth of skeleton secreting organisms were again established. This led to the initiation of new biohermal mounds at the hinge lines of the carbonate banks. Development of this new phase of stromatoporoid mounds marked the beginning of a new biohermal cycle within the reef-bank complex. This process of cyclic development would have been self-controlling and self-perpetuating and could have been accomplished under conditions of continuous transgression. Therefore, fluctuation of sea level, caused by intermittent tectonic subsidence, tectonic processes, or climatic changes was not necessarily responsible for cyclic development of the marginal reef-bank complexes of the Swan Hills area. Recently, Ginsburg (1971, p. 340) has proposed a similar generalized model for formation of regressive carbonate cycles.

WATERWAYS FORMATION

Warren (1933, p. 149) proposed the name Waterways Formation for those subsurface Devonian rocks in northeastern Alberta that unconformably underlie the Lower Cretaceous and are underlain by the Slave Point Formation. According to Norris (1963, p. 64), Kindle (1928, p. 16) used the term Peace River Point beds for a similar suite of rocks, consisting of fossiliferous blue shale and thin-bedded limestones overlying dolomitic limestones of the Slave Point Formation and underlying the Pleistocene and Recent deposits in the Gypsum Cliff area. Peace River Point beds thus have priority over Warren's Waterways Formation. Norris (*loc. cit.*) proposes to retain Kindle's name as a member of the much thicker Waterways Formation known from the outcrops and subsurface in the Athabasca–Clearwater area. In the Gypsum Cliffs area a period of erosion separated the Peace Point beds from the Slave Point Formation (Norris, 1963, p. 65).

The Waterways Formation in the Sturgeon–Mitsue area is 400 to 500 feet (120 to 150 m) thick and is composed of grey limy calcareous shales, interbedded with dark grey nodular limestones (Fig. 40c) and thinly laminated argillaceous limestones (Fig. 40d); it is more calcareous near the carbonate reef-bank complex than in more distal areas and envelops the marginal reef-bank complexes of the Swan Hills Formation (Fig. 27).

The Waterways Formation is a basin-filling, open-marine transgressive sequence. Some of the Waterways open-marine deposits may in part be time equivalents of the Swan Hills marginal reef-bank complex. Those Waterways sediments deposited immediately on the surface of the marginal reef-bank complex can be distinguished by a glauconite-crinoid-brachiopod rich calcarenite (Fig. 40b). This basal calcarenite appears to be similar to the "Coquina zone" of Jenik and Lerbekmo (1968, p. 53), and grades upward into nodular argillaceous limestones that commonly contain *Tentaculites* (Fig. 40c). The laminated limestones intergrade upward with grey, calcareous limy shales. The Waterways sequence represents a gradual marine transgression associated with a deepening of the basin.

The Waterways Formation appears to have been deposited in an open-marine epicontinental basin in which various reducing conditions prevailed, as can be assumed from the common occurrence of pyrite.

CONCLUSIONS

1. The uppermost Muskeg carbonate–evaporite sequence in the Sturgeon–Mitsue area was deposited on broad tidal flats, west of the Mitsue area; to the east was a shallow marine basin.

2. Epeirogenic uplift of the Peace River landmass spread terrigenous clastics beyond the toe of newly uplifted areas and deposited them over the evaporites of the Muskeg Formation. On the coastal plain the lobate Gilwood delta system developed with associated mudflats, shoreline beaches, and on the seaward side, lagoons, bays, and shallow open sea. The Sturgeon–Mitsue area had a marginal position in the fluviomarine complex that developed on the eastern side of the emergent Peace River landmass. Tidal flats prograded seaward for about 40 miles (64 km) and total delta progradation extended about 65 miles (104 km) eastward from the Peace River landmass. Delta progradation thus formed a broad inorganic platform.

3. Barrier island–bay-belt-flanked deltaic plain. A stacked offshore barrier island has been recognized in the Mitsue area. The barrier is of the barrier bar–tidal channel–bay type, and its formation related to the destructional stage of the delta. With the decrease in the sediment supply and continuous transgression, the deltaic sediments were partly reworked and the sand barrier, as well as transverse sandbars, was formed. After the building of the barrier ceased the bay was filled by clay and silt-sized sediments. Well-sorted barrier island sands retained high porosity.

4. Subsequently the carbonates and evaporites of the Fort Vermilion Formation were deposited as a result of semiarid climatic conditions and a cessation of terrigenous deposition over the drowned delta platform. A zone approximately 35 miles (56 km) wide provided optimum conditions for intertidal evaporite depositions. Calcareous shales were deposited in a subtidal environment east of the Mitsue area.

5. The differential compaction of interdistributary bay and lagoonal shales around the sandstone distributary belts and transverse sandbars resulted in a slight relief, which may have been magnified somewhat by deflation processes during short periods of exposure of the Fort Vermilion surface. These elevations were about 30 feet (9 m) high and were large enough to influence subsequent reef-bank initiation and facies distribution of the Swan Hills Formation.

6. The Swan Hills marine transgression flooded the inorganic plateau, and shallow-water carbonates were deposited over the plateau as a thin veneer (basal carbonate blanket), marking the first stage of the marginal reef-bank complex development. Depending on local bathymetric conditions and paleotopography, oölites, biomicrites, fossiliferous micrites, and pelletoidal carbonates were deposited on shoals, tidal flats, and broad shallow basins. Occurrence of black pigmented grains suggests that reducing and anaerobic conditions existed near the sea-bottom surface along the outer edge of marginal reef-bank complex toward the Waterways basin. Composite grains and oölites suggest that high chemical precipitation occurred in some areas of incipient reef-bank growth.

7. Algal stromatolites grew in a shallow subtidal environment which was protected from the direct influence of waves in water depth similar to that of recent subtidal mats at Abaco in the Bahamas (Neumann *et al.*, 1970).

8. Biohermal mounds of a tabular stromatoporoid developed at the outer perimeter of the basal carbonate blanket and inside the basal carbonate blanket on the slopes of the inlets, passes, and re-entrants. The developing biohermal mounds coalesced into elongate, discontinuous submerged ridges (barriers) which were several miles long, up to 2 miles (3 km) wide, and 30 feet (9 m) high. These belts can be compared in part to recent shelf-edge reef tract of the Florida Keys (pers. observation—Jansa).

9. The main element that influenced initiation of bioherms was pre-existing topography. Elevated areas, colonized by tabular stromatoporoids, were closely related to the position of the underlying distributary sandbars and distributary sand belts of the Gilwood Member.

10. The interfingering and overlapping of the biohermal and carbonate bank sediments built marginal reef-bank complexes 2 to 12 miles (approx. 3 to 19 km) wide, up to 20 miles (32 km) long with an average thickness of 100 feet (30 m). Location of these complexes was predetermined by the location of the tabular stromatoporoids mounds.

11. A prevailing northeast onshore wind and water movement established a hydrodynamic barrier at the front of the marginal reef-banks which prevented the sedimentary particles, generated in the reef-bank complexes and shelf lagoon, from being transported to the open shelf and basin areas. Thus, the reef-banks with associated shelf lagoon operated as a closed system during the upward growth of the reef-banks.

12. Other factors that influenced reef-bank development were: a low tidal fluctuation with a range of approximately 2 feet (0.6 m); medium rather than high overall energy conditions; and water salinities near that of normal sea water. A hot, semiarid climate similar to that of the recent Persian Gulf also is postulated during Swan Hills time, and is inferred from time equivalent synchronous evaporite deposits in the Williston Basin of Saskatchewan.

13. Two "regressive" cycles are recognized in the marginal reef-bank complex, with an average thickness of 45 feet. Deposition of these cycles can be explained by the combined effect of hydrodynamic processes such as wind direction, currents, waves, etc., on carbonate particles formed in the shallow shelf lagoon and marginal reef-banks. The cycles developed during a slow continuous transgression.

14. Two zones of hydrocarbon accumulation developed in a clastic coastal plain deltaic sequence and overlying carbonate reef-bank complexes of the Swan Hills area: (1) Weak and incomplete cementation preserved a high primary porosity in the well-sorted Gilwood sand barrier. This, together with a sealing effect of surrounding shales and overlying Fort Vermilion Formation evaporites, developed a highly potential stratigraphic trap. The early silica cementation of the sandstones in the distributary mouth bar zone formed an effective barrier to updip oil migration into deltaic and fluvial sands of the Gilwood Member. (2) The second zone developed in the marginal reef-banks of the Swan Hills Formation. Diagenetic processes, influenced by a short period of exposure into the vadose zone led to aragonite alteration, preferential solution, leaching, and incomplete filling of voids. Porosity is highest in the subspherical stromatoporoid facies and associated carbonates and facies on the upper front slope

of the reef-bank complexes and in the inner (lagoonward) slopes of the marginal reef-banks. Consequently, these facies are the most prolific oil reservoirs of the Swan Hills Formation.

REFERENCES

Allen, J. R. L.
1970: Physical processes of sedimentation, an introduction; London, Allen and Unwin, 248 p.

Andrews, P. B. and Van der Lingen, G. J.
1969: Environmentally significant sedimentologic characteristics of beach sand; N.Z.J. Geol. Geophys., v. 12, p. 119–137.

Ball, M. M.
1967: Carbonate sand bodies of Florida and Bahamas; J. Sediment. Petrology, v. 37, p. 556–591.

Bassett, H. G. and Stout, J. G.
1967: Devonian of western Canada; *in* International Symposium on the Devonian System, D. H. Oswald, *ed.,* v. 1, p. 717–752; J. Alberta Soc. Petrol. Geol., Calgary, Alberta, 1377 p.

Belyea, H. R., Moyer, G. L., Gray, F. F., and Grayston, L. D.
1966: Upper Devonian: Part 1—Beaverhill Lake Formation; *in* Geological history of western Canada, R. G. McCrossan, and R. P. Glaister, *eds.,* 2nd ed., 1966; Alberta Soc. Petrol. Geol., p. 60–66.

Belyea, H. R. and Norris, A. W.
1962: Middle Devonian and older Paleozoic formations of southern District of Mackenzie and adjacent areas; Geol. Surv. Can., Paper 62-15, 82 p.

Berg, R. R. and Davies, D. K.
1968: Origin of Lower Cretaceous Muddy Sandstone at Bell Creek field, Montana; Bull. Amer. Assoc. Petrol. Geol., v. 52, p. 1888–1898.

Bernard, H. A., LeBlanc, R. J., and Major, C. F.
1962: Recent and Pleistocene geology of southeast Texas, field excursion No. 3; *in* Geology of the Gulf Coast and Central Texas and guidebook of excursions; Geol. Soc. Amer. Ann. Mtg. Guidebook, p. 175–224.

Bien, G. S., Contois, D. E., and Thomas, W. H.
1958: Removal of soluble silica from fresh water entering the sea; (reprint from Geochimica et Cosmochimica Acta 14): a symposium; *in* Silica in Sediments, H. A. Ireland, *ed.*; Soc. Econ. Paleont. Mineral., Spec. Publ., no. 7, p. 20–35, 1959.

Bird, E. C. F.
1969: Coasts; Cambridge, Mass., M.I.T. Press (An introduction to systematic geomorphology), v. 4, 246 p.

Blatt, H.
1967: Original characteristics of clastic quartz grains; J. Sediment. Petrology, v. 37, no. 2, p. 401–424.

Brown, P. R.
1963: Some algae from the Swan Hills reef; Bull. Can. Petrol. Geol., v. 11, no. 2, p. 178–182.

Butler, G. P.
1970: Holocene gypsum and anhydrite of the Abu Dhabi Sabkha, Trucial Coast: an alternative explanation of origin; *in* Third Symposium on Salt, J. L. Rau and L. F. Dellwig, *eds.*; Northern Ohio Geol. Soc., Cleveland, p. 120–152.

Campbell, N. L.
1950: Middle Devonian of the Pine Point area, Northwest Territories; Proc. Geol. Assoc. Can., v. 33, p. 87–96.

Carozzi, A. V.
1961: Reef petrography in the Beaverhill Lake Formation, Upper Devonian, Swan Hills area, Alberta, Canada; J. Sediment. Petrology, v. 31, p. 497–513.

Coleman, J. M. and Gagliano, S. M.
1965: Sedimentary structures: Mississippi river deltaic plain; *in* Sedimentary structures and their hydrodynamic interpretation; Soc. Econ. Paleontol. Mineral., Spec. Publ., no. 12, Tulsa, p. 133–148.

Davies, G. R.
1970: Carbonate bank sedimentation, Eastern Shark Bay, Western Australia; *in* Carbonate Sedimentation and Environments, Shark Bay, Western Australia; Amer. Assoc. Petrol. Geol., Mem. 13, Tulsa, p. 85–168.

de Mille, G.
1958: Pre-Mississippian history of the Peace River Archipelago; J. Alberta Soc. Petrol. Geol., v. 6, no. 3, p. 61–68.

De Raaf, J. F. M. and Boersma, J. R.
1971: Tidal deposits and their sedimentary structures; Geol. en Mijnbouw, v. 50, no. 3, p. 479–504.

Dickson, A. A. D.
1965: Modified staining techniques for carbonates in thin section; Nature 205, p. 587.

Douglas, R. J. W., Gabrielse, H., Wheeler, J. O., Stott, D. F., and Belyea, H. R.
1970: Geology of western Canada, *in* Geology and economic minerals of Canada; R. J. W. Douglas, *ed.*, Geol. Surv. Can., Econ. Geol. Rept. 1.

Dunham, R. J.
1962: Classification of carbonate rocks according to depositional texture; *in* Classification of Carbonate Rocks, W. E. Ham, ed.; Amer. Assoc. Petrol. Geol., Mem. 1, p. 108–121.

Edie, R. W.
1961: Devonian limestone reef reservoir, Swan Hills oilfield, Alberta, Canada; Can. Mining Met. Bull., v. 54, p. 447-453.

Evans, G.
1966: Persian Gulf; *in* Encyclopedia of Oceanography, R. W. Fairbridge, *ed.*, Reinhold Pub. Corp., N.Y., p. 689–694.

Fairbridge, R. W.
1950: Recent and Pleistocene coral reefs of Australia; J. Geol., v. 58, no. 4, p. 330–401.
1966: Encyclopedia of Oceanography, R. W. Fairbridge, *ed.*; Reinhold, N.Y., 1966, 1021 p.

Fischbuch, N. R.
1960: Stromatoporoids of the Kaybob reef, Alberta; J. Alta. Soc. Petrol. Geol., v. 8, no. 4, p. 113–131.
1962: Stromatoporoid zones of the Kaybob reef, Alberta; Alberta Soc. Petrol. Geol., v. 10, no. 2, p. 62–72.
1968: Stratigraphy, Devonian Swan Hills reef complexes of Central Alberta; Bull. Can. Petrol. Geol., v. 16, no. 4, p. 446–587.
1969: Devonian stromatoporoids from central Alberta, Canada; Can. J. Earth Sci., v. 6, no. 1, p. 167–185.
1970a: *Amphipora* and *Euryamphipora* (Stromatoporoidea) from the Devonian of western Canada; Paleontology, v. 13, Part I, p. 64–75.
1970b: Devonian reef-building stromatoporoids from western Canada; J. Paleontol., v. 44, no. 6, p. 1071–1084.

Fisher, W. L. and Brown, L. F., Jr.
1969: Delta systems and oil and gas occurrences; *in* Delta Systems in the Exploration of Oil and Gas: a research colloquim; Bur. Econ. Geol., Univ. of Texas, Austin, p. 54–66.

Folk, R. L.
1955: Note on the significance of "turbid" feldspars; Amer. Mineralogist, v. 40, p. 356–357.
1962: Spectral subdivision of limestone types; *in* Classification of Carbonate Rocks, W. E. Ham, *ed.*; Amer. Assoc. Petrol. Geol., Mem. 1, Tulsa, p. 62–84.
1968: Petrology of Sedimentary Rocks; Hemphill's, Austin, Texas, 170 p.

Folk, R. L. and Ward, W. C.
1957: Brazos river bar: a study in the significance of grain size parameters; J. Sediment. Petrology, v. 27, no. 1, 3–26.

Fong, G.
1959: Type section Swan Hills Member of the Beaverhill Lake Formation; J. Alberta Soc. Petrol. Geol., v. 7, no. 5, p. 95–108.

Friedman, G. M. and Amiel, A. J.
1970: Submarine cements in modern Red Sea reef rock; Geol. Soc. Amer. Abstract with programs, v. 2, p. 555.

Gagliano, S. M. and McIntire, W. G.
1968: Reports on the Mekong River Delta; Baton Rouge, Louisiana State Univ. Coastal. Stud. Inst., tech. rept. 57, 143 p.

Galloway, J. J.
1957: Structure and classification of the Stromatoporoidea; Bull. Amer. Paleont., v. 37, no. 164, p. 341–480.

Germann, K.
1969: Reworked dolomite crusts in the Wettersteinkalk (Ladinian, Alpine Triassic) as indicators of early supratidal dolomitization and lithification; Sedimentology, v. 12, no. 3-4, p. 257–277.

Ginsburg, R. N.
1956: Environmental relationships of grain size and constituent particles in some south Florida carbonate sediments; Amer. Assoc. Petrol. Geol., Bull., v. 40, p. 2384–2427.
1960: Ancient analogues of recent Stromatolites; Int. Geol. Congr., Copenhagen, XXI Session, Part XXII, p. 26–35.
1971: Landward movement of carbonate mud: New model for regressive cycles in carbonates; AAPG Abstracts, v. 55, p. 340.

Ginsburg, R. N., Marszalek, D. S., and Schneidermann, N.
1971: Ultrastructure of carbonate cements in a Holocene algal reef of Bermuda; J. Sediment. Petrology, v. 41, no. 2, p. 472–482.

Grayston, L. D., Sherwin, D. F., and Allan, J. F.
1964: Middle Devonian: *in* Geological history of western Canada; R. G. McCrossan and R. P. Glaister, *eds.;* Alberta Soc. Petrol. Geol. Calgary, p. 49–59.

Gürich, G.
1906: Les spongiostromides du Viséan de la province de Namur; Mem., Mus. R. Hist. Nat. de Belg., v. III, no. 4, p. 1–56.

Guthrie, D. C.
1956: Gilwood sandstone in the Giroux Lake area, Alberta; Alberta Soc. Petrol. Geol., v. 4, no. 10, p. 227–231.

Hails, J. R. and Hoyt, J. H.
1968: Barrier development on submerged coasts: problems of sea-level changes from a study of the Atlantic Coastal Plain of Georgia, U.S.A. and parts of the east Australian coast; Z. Geomorpho., v. 7, p. 24–55.

Harms, J. C. and Fahnestock, R. K.
1965: Stratification, bed forms, and flow phenomena (with an example from the Rio Grande); Soc. Econ. Paleontol. Mineral. Spec. Publ. no. 12, Tulsa, p. 84–115.

Hemphill, C. R., Smith, R. I., and Szabo, F.
1970: Geology of Beaverhill Lake reefs, Swan Hills area, Alberta; *in* Geology of Giant Petroleum Fields—Symposium, 53rd. Ann. Mtg. Oklahoma City, 1968; Amer. Assoc. Petrol. Geol., Mem. 14, p. 50–90.

Hofmann, H. J.
1969: Attributes of Stromatolites; Geol. Surv. Can., Paper 69-39, 58 p.

Hoskin, C. M.
1963: Recent carbonate sedimentation on Alacran Reef, Mexico; Nat. Acad. Sci./Nat. Res. Counc., Publ. 1089, Foreign Fields Res. Progr. Rept. 19, 160 p.

Houbolt, J. J. H. C.
1957: Surface sediments of the Persian Gulf near the Qatar Peninsula; Mouton & Co., Netherlands, 108 p.

Hoyt, J. H. and Henry, V. J. Jr.
1967: Influence of island migration on barrier island sedimentation; Bull. Geol. Soc. Amer., v. 78, no. 1, p. 77–86.

Hriskevich, M. E.
1966: Stratigraphy of Middle Devonian and older rocks of Banff–Aquitaine Rainbow West 7-32 discovery well, Alberta; Bull. Can. Petrol. Geol., v. 14, p. 241–265.

Illing, L. V.
1954: Bahaman calcareous sands; Bull. Amer. Assoc. Petrol. Geol., v. 38, no. 1, p. 1–95.

Imperial Oil Ltd., Geological Staff
1950: Devonian nomenclature in Edmonton area, Alberta, Canada; ibid., Bull. Amer. Assoc. Petrol. Geol., v. 34, no. 9, p. 1807–1825.

Jansa, L. F. and Tomšik, J.
1961: Facial-cyclic analysis and its use in the Ostrava Coal basin; Geol. Methods. Prague, Nu. 1, 68 p.

Jenik, A. J. and Lerbekmo, J. F.
1968: Facies and geometry of Swan Hills Reef Member of Beaverhill Lake Formation (Upper Devonian), Goose River field, Alberta, Canada; Bull. Amer. Assoc. Petrol. Geol., v. 52, no. 1, p. 21–56.

Johnson, J. H.
1961: Limestone-building algae and algal limestones; Colo. School Mines Publ., Boulder, p. 297.

Kalkowsky, E.
1908: Oolith und Stromatolith in Norddeutschen Bundsandstein; Deutche Geol. Gesell. Zeit., v. 60, p. 68–125.

Keith, J. W.
1970: Tectonic control of Devonian reef sedimentation, Alberta; Amer. Assoc. Petrol. Geol., abstracts of ann. conv., Calgary, p. 854.

Kendall, Ch. G. St. C. and Skipwith, P.A. d'E.
1969: Geomorphology of a Recent shallow-water carbonate province; Khor Al Bazam, Trucial Coast, southwest Persian Gulf; Bull. Geol. Soc. Amer., v. 80, p. 865–892.

Kindle, E. M.
1928: The occurrence and correlation of a Devonian fauna from Peace River, Alberta; Contrib. to Canadian Paleontol.; Geol. Surv. Can., Mus. Bull. 49, p. 14–18.

Kinsman, D. J. J.
1966: Gypsum and anhydrite of recent age, Trucial Coast, Persian Gulf; *in* Second Symposium on Salt; Northern Ohio Geol. Soc., Cleveland, v. 1, p. 302–306.
1969: Modes of formation, sedimentary associations and diagnostic features of shallow-water and supratidal evaporites; Bull. Amer. Assoc. Petrol. Geol., v. 53, no. 4, p. 830–840.

Klovan, J. E.
1964: Facies analysis of the Redwater reef complex, Alberta, Canada; Bull. Can. Petrol. Geol., v. 12, no. 1, p. 1–100.

Kornicker, L. S. and Boyd, D. W.
1962: Shallow-water geology and environments of Alacran reef complex, Campeche Bank, Mexico; Bull. Amer. Assoc. Petrol. Geol., v. 46, no. 5, p. 640–673.

Kraft, J. C.
1971: Sedimentary facies patterns and geologic history of a Holocene Marine transgression; Bull. Geol. Soc. Amer., v. 82, no. 8, p. 2131–2158.

Kramers, J. W. and Lerbekmo, J. F.
1967: Petrology and mineralogy of Watt Mountain Formation, Mitsue-Nipisi area, Alberta; Bull. Can. Petrol. Geol., v. 15, no. 3, p. 346–378.

Kruit, C.
1955: Sediments of the Rhone delta: 1. Grain size and microfauna; Kon. Ned. Geol. Mijnbouw, Gen. Verhand., v. 15, p. 357–514.

Krumbein, W. C.
1941: Measurement and geologic significance of shape and roundness of sedimentary particles; J. Sediment. Petrol., v. 11, p. 4–72.

Kuenen, P. H.
1950: Marine Geology; N.Y., Wiley, 568 p.

Land, L. S.
1970: Phreatic versus vadose meteoric diagenesis of limestones: evidence from a fossil water table; Sedimentology, v. 14, p. 175–185.

Law, J.
1955: Geology of northwestern Alberta and adjacent areas; Bull. Amer. Assoc. Petrol. Geol., v. 39, p. 1927–1975.

Leavitt, E. M.
1968: Petrology, paleontology, Carson Creek North reef complex, Alberta; Bull. Can. Petrol. Geol., v. 16, no. 3, p. 298–413.

Leavitt, E. M. and Fischbuch, N. R.
1968: Devonian nomenclatural changes, Swan Hills area, Alberta, Canada; Bull. Can. Petrol. Geol., v. 16, no. 3, p. 288–297.

Lecompte, M.
1951: Les stromatoporoïdes du Dévonien moyen et supérieur du Bassin de Dinant; Inst. Roy. Sci. Natur. Belg., Mem. 116 (1951), p. 1–215.

Lewis, M. S.
1967: The morphology of the fringing coral reefs along the east coast of Mahe, Seychelles; J. Geol., v. 76. p. 140–153.

Logan, B. W., Davies, G. R., Read, J. F., and Cebulski, D. E.
1970: Carbonate sedimentation and environments, Shark Bay, Western Australia; Amer. Assoc. Petrol. Geol., Mem. 13, Tulsa, p. 1–223.

Logan, B. W., Harding, J. L., Ahr, W. M., Williams, J. D., and Snead, R. G.
1969: Carbonate sediments and reefs, Yucantan Shelf, Mexico; Amer. Assoc. Petrol. Geol., Mem. 11, Menasha, p. 1–198.

Logan, B. W., Rezak, R., and Ginsburg, R. N.
1964: Classification and environmental significance of algal Stromatolites; J. Geol., v. 72, no. 1, p. 68–83.

Macintyre, I. G.
1970: Sediments off the west coast of Barbados: diversity of origins; Marine Geology, 1970, v. 9, no. 1, p. 5–23.

Maiklem, W. R.
1967: Black and brown speckled foraminiferal sand from the southern part of the Great Barrier Reef; J. Sediment. Petrology, v. 37, no. 4, p. 1023–1030.

Maiklem, W. R., Bebout, D. G., and Glaister, R. P.
1969: Classification of Anhydrite: a practical approach; Bull. Can. Petrol. Geol., v. 17, no. 2, p. 194–233.

Martin, R.
1967: Morphology of some Devonian reefs in Alberta: a paleogeomorphological study; *in* International Symposium on the Devonian System, Calgary, p. 365–385.

Mason, A. D. M.
1960: Manning Sand: Watt Mountain Formation, Elk Point Group; *in* Lexicon of Geologic Names in the western Canada Sedimentary basin and Arctic Archipelago, J. Alberta Soc. Petrol. Geol., Calgary, p. 215–216.

Maxwell, W. G. H.
1968: Atlas of the Great Barrier Reef, Elsevier, Amsterdam, 1968, 258 p.

McCamis, J. G. and Griffith, L. S.
1967: Middle Devonian facies relationships, Zama area, Alberta; Bull. Can. Petrol. Geol., v. 15, p. 434–467.

McGill, P.
1963: Upper and Middle Devonian ostracods from the Beaverhill Lake Formation, Alberta, Canada; Bull. Can. Petrol. Geol., v. 11, no. 3, p. 299–303.
1966: Ostracods of probable Late Givetian age from the Slave Point Formation, Alberta; Bull. Can. Petrol. Geol., v. 14, no. 1, p. 104–133.

Mesolella, K. J.
1968: Uplift reefs of Barbados: physical stratigraphy, facies relationships and absolute chronology; Ph.D. thesis, Brown Univ., Providence, R.I., 736 p.

Mesolella, K. J., Sealy, H. A., and Matthews, R. K.
1970: Facies geometries within Pleistocene reefs of Barbados, West Indies; Bull. Amer. Assoc. Petrol. Geol., v. 54, no. 10, p. 1899–1917.

Milliman, J. D.
1969: Carbonate sedimentation on four southwestern Caribbean atolls and its relation to the "oolite problem"; *in* Geology of the American Mediterranean; Gulf Coast Assoc. Geol. Soc. Trans., v. 19, p. 195–206.

Moore, D.
1966: Deltaic sedimentation; Earth Sci. Rev., v. 1, no. 2-3, p. 87–104.

Murray, J. W.
1964: Some stratigraphic and paleoenvironmental aspects of the Swan Hills and Waterways Formations, Judy Creek, Alberta, Canada; Ph. D. thesis, Princeton Univ., 121 p.
1965: Stratigraphy and carbonate petrology of the Waterways Formation, Judy Creek, Alberta, Canada; Bull. Can. Petrol. Geol., v. 13, no. 2, p. 303–326.
1966: Oil producing reef-fringed carbonate bank in the Upper Devonian Swan Hills Member, Judy Creek, Alberta; Bull. Can. Petrol. Geol., v. 14, no. 1, p. 1–103.

Neumann, A. C., Gebelein, C. D., and Scoffin, T. P.
1970: Composition, structure and erodability of subtidal mats, Abaco, Bahamas; J. Sediment. Petrology, v. 40, no. 1, p. 274–297.

Nevin, C.
1946: Competency of moving water to transport debris; Bull. Geol. Soc. Amer., v. 57, p. 651–674.

Newell, N. D. and Rigby, J. K.
1957: Geological studies on the Great Bahama Bank; *in* Regional aspects of carbonate deposition, R. J. LeBlanc and J. G. Breeding, *eds.*, Soc. Econ. Paleontol. Mineral., Spec. Publ. no. 5, Tulsa, p. 15–72.

Noble, J. P. A.
1970: Biofacies analysis, Cairn Formation of Miette reef complex (Upper Devonian), Jasper National Park, Alberta; Bull. Can. Petrol. Geol., v. 18, no. 4, p. 493–543.

Norris, A. W.
1963: Devonian stratigraphy of northeastern Alberta and northwestern Saskatchewan; Geol. Surv. Can., Mem. 313, 168 p.
1970: Report on Devonian fossils from the subsurface of the Swan Hills area, northern Alberta; Report No. D2, 1970, AWN (unpubl.), GSC, Calgary, p. 1–3.

Peck, R. E. and Morales, G. A.
1966: The Devonian and Lower Mississippian charophytes of North America; Micropaleontology, v. 12, no. 3, p. 303–324.

Pedder, A. E. H.
1971: Twelve lots of subsurface Middle Devonian fossils from the Swan Hills area of Alberta; Report of GSC, Calgary, No. LFJ, 24 AEHP, 1971, p. 2 (unpubl.).

Perkins, R. D.
1963: Petrology of the Jeffersonville limestone (Middle Devonian) of southeastern Indiana; Bull. Geol. Soc. Amer., v. 74, no. 1, p. 1335–1354.

Pettijohn, F. J.
1957: Sedimentary Rocks; Harper and Brothers, N.Y., 718 p.

Playford, P. E. and Cockbain, A. E.
1969: Algal stromatolites: deepwater forms in the Devonian of western Australia; Science, v. 165, no. 3897, p. 1008–1010.

Purdy, E. G.
1963a: Recent calcium carbonate facies of the Great Bahama Bank; 1. Petrography and reaction groups; J. Geol., v. 71, no. 3, p. 334–355.
1963b: Recent calcium carbonate facies of the Great Bahama Bank; 2. Sedimentary facies; J. Geol., v. 71, no. 4, p. 472–497.
1965: Structural control of recent calcium carbonate deposition in British Honduras; Geol. Soc. Amer., Abstracts for 1964, p. 157.

Read, J. F.
1970: Carbonate bank development and associated sedimentation, Edel Province, Shark Bay, Western Australia; part 2: Calcrete deposits and Quaternary sediments; Unpubl. Ph.D. thesis, Univ. Western Australia, Perth.

Reineck, H. E.
1967: Layered sediments of tidal flats, beaches, and shelf bottoms of the North Sea; *in* Estuaries, G. H. Lauff, *ed.*, Amer. Assoc. Petrol. Geol., Wash., p. 191–206, Publ. no. 83.

Scott, A. J. and Fisher, W. L.
1969: Delta systems and deltaic depositions; *in* Delta Systems in the Exploration for Oil and Gas: a research colloquim, W. L. Fisher, L. F. Brown, and A. J. Scott; Tex. Bur. Econ. Geol., Univ. Tex., Austin, p. 10–29.

Shawa, M. S.
1969: Sedimentary history of the Gilwood sandstone (Devonian) Utikuma Lake area, Alberta, Canada; Bull. Can. Petrol. Geol., v. 17, no. 4, p. 392–409.

Shearman, D. J.
1966: Origin of marine evaporites by diagenesis; Bull. Inst. Mining Met., 1717, p. 207–215.

Shearman, D. J. and Fuller, J. G.
1969: Anhydrite diagenesis, calcitization, and organic laminates, Winnipegosis Formation, Middle Devonian, Saskatchewan; Bull. Can. Petrol. Geol., v. 17, no. 4, p. 496–525.

Shearman, D. J. and Skipwith, P. A. d'E.
1965: Organic matter in Recent and ancient limestones and its role in their diagenesis; Nature, 208, p. 1310–1311.

Shepard, F. P.
1960: Gulf coast barriers; *in* Recent Sediments, Northwest Gulf of Mexico, F. P. Shepard, F. B. Phleger, and T. H. Van Andel, *eds.*, Bull. Amer. Assoc. Petrol. Geol., Tulsa, p. 197–220.

Shepard, F. P. and Moore, D. G.
1955: Central Texas coast sedimentation: characteristics of sedimentary environment, recent history and diagenesis; Bull. Amer. Assoc. Petrol. Geol., v. 39, no. 8, p. 1463–1593.

Stanley, D. J. and Swift, D. J. P.
1968: Bermuda's reef-front platform: bathymetry and significance; Marine Geol., v. 6, p. 479–500.

Stearn, C. W.
1963: Some stromatoporoids from the Beaverhill Lake Formation (Devonian) of the Swan Hills area, Alberta; J. Paleont. v. 37, no. 3, p. 651–668.
1966: Microstructure of stromatoporoids; Paleontology, v. 9, Pt. 1, p. 74–124.

Swinchatt, J. P.
1965: Significance of constituent composition, texture and skeletal breakdown in some recent carbonate sediments; J. Sediment. Petrology, v. 35, no. 1, p. 71–90.

Taylor, J. C. M. and Illing, L. V.
1969: Holocene intertidal calcium carbonate cementation, Qatar, Persian Gulf; Sedimentology, v. 12, p. 69–107.

Thomas, G. E. and Rhodes, H. S.
1961: Devonian limestone bank-atoll reservoirs of the Swan Hills area, Alberta; J. Alberta Soc. Petrol. Geol., v. 9, no. 2, p. 29–38.

Thompson, W. O.
1937: Original structures of beaches, bars, and dunes; Bull. Geol. Soc. Amer., v. 48, p. 723–752.

Thompson, R. W.
1968: Tidal flat sedimentation on the Colorado River Delta; Amer. Geol. Soc., Mem. 107, Boulder, p. 1–133.

Todd, T. W.
1968: Paleoclimatology and the relative stability of feldspar minerals under atmospheric conditions; J. Sediment. Petrology, v. 38, no. 3, p. 832–844.

Todd, T. W. and Monroe, W. A.
1968: Petrology of Domengine Formation (Eocene) at Potrero Hills and Rio Vista, California; J. Sediment. Petrology, v. 38, no. 4, p. 1024–1039.

van Straaten, L. M. J. U.
1959: Minor structures of some recent littoral and neritic sediments; Geol. Mijnbouw, v. 21, p. 197–217.

Walker, T. R.
1967: Formation of red beds in modern and ancient deserts; Bull. Geol. Soc. Amer., v. 78, no. 3, p. 353–368.

Walker, T. R. and Honea, R. M.
1969: Iron content of modern deposits in the Sonoran Desert: a contribution to the origin of red beds; Bull. Geol. Soc. Amer., v. 80, p. 535–544.

Warren, P. S.
1933: Age of the Devonian limestone at McMurray, Alberta; Can. Field Natur., v. 47, no. 8, p. 148–149.

Wells, J. W.
1957: Coral reefs; Geol. Soc. Amer., Mem. 67, v. 1, p. 609–631.

Index

BULLETINS

Geological Survey of Canada

Bulletins present the results of detailed scientific studies on geological or related subjects.

Some recent titles are listed below (Information Canada No. in brackets):

200 Part I: Biostratigraphy of some Early Middle Silurian Ostracoda, eastern Canada: Part II: Additional Silurian Arthropoda from arctic and eastern Canada, *by* M. J. Copeland, $1.50 (M42-200)

201 Archaeocyatha from the Mackenzie and Cassiar Mountains, Northwest Territories, *by* R. C. Handfield, $2.00 (M42-201)

202 Faunas of the Ordovician Red River Formation, Manitoba, *by* D. C. McGregor, *et al.*, $2.00 (M42-202)

203 Geology of lower Paleozoic formations, Hazen Plateau and southern Grant Land Mountains, Ellesmere Island, *by* H. P. Trettin, $3.00 (M42-203)

204 Brachiopods of the Detroit River Group (Devonian) from southwestern Ontario and adjacent areas of Michigan and Ohio, *by* J. A. Fagerstrom, $2.00 (M42-204)

205 Comparative study of the Castle River and other folds in the Eastern Cordillera of Canada, *by* D. K. Norris, $2.00 (M42-205)

206 Geomorphology and multiple glaciation in the Banff area, *by* N. W. Rutter, $2.00 (M42-206)

207 Geology of the resurgent cryptoexplosion crater at Mistastin Lake, Labrador, *by* K. L. Currie, $2.00 (M42-207)

208 The geology and origin of the Faro, Vangorda and Swim concordant zinc-lead deposits, Central Yukon Territory, *by* D. J. Tempelman-Kluit, $3.00 (M42-208)

209 Redescription of *Marrella splendens* (Trilobitoidea) from the Burgess Shale, Middle Cambrian, British Columbia, *by* H. B. Whittington, $3.00 (M42-209)

210 Ordovician trilobites from the central volcanic mobile belt at New World Island, northeastern Newfoundland, *by* W. T. Dean, $2.00 (M42-210)

211 A Middle Ordovician fauna from Braeside, Ottawa Valley, Ontario, *by* H. Miriam Steele and G. Winston Sinclair, $2.00 (M42-211)

212 Lower Cambrian trilobites from the Sekwi Formation type section, Mackenzie Mountains, Northwestern Canada, *by* W. H. Fritz, $4.00 (M42-212)

213 Sequence of glacial lakes in north-central Alberta, *by* D. A. St-Onge, $2.00 (M42-213)

214 Classification and description of copper deposits, Coppermine River area, District of Mackenzie, *by* E. D. Kindle, $4.00 (M42-214)

215 Brachiopods of the Arisaig Group (Silurian-Lower Devonian) of Nova Scotia, *by* Charles W. Harper, Jr., $5.00 (M42-215)

217 The geology and petrology of the alkaline carbonatite complex at Callander Bay, Ontario, *by* John Ferguson and K. L. Currie, $2.00 (M42-217)

218 Keweenawan volcanic rocks of Michipicoten Island, Lake Superior, Ontario (41N): An eruptive centre of Proterozoic age, *by* R. N. Annells, $5.00 (M42-218)

219 Lower Cretaceous Bullhead Group, between Bullmoose Mountain and Tetsa River, Rocky Mountain Foothills, Northeastern British Columbia, *by* D. F. Stott, $6.00 (M42-219)

221 Chitinozoa and Acritarcha of the Hamilton Formation (Middle Devonian), southwestern Ontario, *by* J. A. Legault, $4.00 (M42-221)

222 Contributions to Canadian Paleontology, *by* D. E. Jackson, *et al.*, $6.00 (M42-222)

224 Carboniferous and Permian stratigraphy of Axel Heiberg Island and western Ellesmere Island, Canadian Arctic Archipelago, *by* R. Thorsteinsson, $6.00 (M42-224)

231 *Yohoia* Walcott and *Plenocaris* n. gen. arthropods from the Burgess Shale, Middle Cambrian, British Columbia, *by* H. B. Whittington, $3.00 (M42-231)

232 Conodonts of the Waterways Formation (Upper Devonian) of northeastern and central Alberta, *by* T. T. Uyeno, $4.00 (M42-232)

234 Evolution of a Middle and Upper Devonian sequence from a clastic coastal plain – deltaic complex into overlying carbonate reef complexes and banks, Sturgeon – Mitsue area, Alberta, *by* L. F. Jansa and N. R. Fischbuch, $5.00 (M42-234)