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SEDIMENTOLOGY AND DIAGENESIS OF THE WINNIPEGOSIS FORMATION, SOUTHEASTERN SASKATCHEWAN

by

ROBIN LAUREL STANFORD ©

A THESIS

SUBMITTED TO THE FACULTY OF GRADUATE STUDIES AND RESEARCH
IN PARTIAL FULFILMENT OF THE REQUIREMENTS FOR THE DEGREE
OF MASTER OF SCIENCE

DEPARTMENT OF GEOLOGY

EDMONTON, ALBERTA
FALL, 1989

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FACULTY OF GRADUATE STUDIES AND RESEARCH

The undersigned certify that they have read, and recommend to the Faculty of Graduate Studies and Research, for acceptance, a thesis entitled SEDIMENTOLOGY AND DIAGENESIS OF THE WINNIPEGOSIS FORMATION, SOUTHEASTERN SASKATCHEWAN submitted by ROBIN LAUREL STANFORD in partial fulfilment of the requirements for the degree of MASTER OF SCIENCE in GEOLOGY.

Supervisor

Date 89-09-11

Abstract

The Winnipegosis Formation is composed of carbonate rocks deposited in the southern Elk Point Basin during Middle Devonian time. The formation includes the Lower, Upper, and Ratner members.

The Lower Member is a basin-wide deposit which is overlain by isolated buildups of the Upper Member. The Ratner Member, which also overlies the Lower Member, accumulated inbetween buildups. The sediments of the Lower Member accumulated below wave base on a carbonate ramp. Buildups of the Upper Member initiated as codiacean algae mounds. With growth into wave base, a thin reefal framework of branching corals and stromatolites developed on the crests and flanks of the buildups. The Ratner Member accumulated under anoxic, hypersaline bottom conditions caused by stratification of the basin waters. The laminated basinal sediment represents incremental deposition of platform-derived lime mud, uncalcified pelagic organisms, and turbidites sourced from adjacent buildups. In the final stages of basin sedimentation, which may postdate buildup growth, increasing restriction led to perioidic precipitation of calcium sulfate from basin waters.

The rocks of the Winnipegosis Formation underwent a complex diagenetic history. Early diagenesis involved precipitation of marine, isopachous cements in primary pores in buildups. The development of hypersaline stratification in the basin waters led to formation of small, anhydrite nodules in the basinal muds. Regional type 1 dolomitization of the rocks occurred through subtidal seepage reflux, whereby dense basin waters flowed downward through underlying strata.

Middle stage diagenesis, which is characterized by dissolution and diagenesis involoving highly variable water chemistries, probably occurred in a surface or near surface diagenetic environment. The middle stage is represented by blocky dolomite cementation, anhydrite cementation, anhydrite replacement and at least two phases of dissolution.

Progressive burial of the rocks by deposition of overlying strata initiated late stage burial diagenesis. Pressure-dissolution and recrystallization of early dolomites occurred in response to burial. Ions released by pressure-dissolution reacted with limestones to form late type 2 dolomite. Limpid dolomite cement precipitated slowly under nearly static flow conditions. Anhydrite cementation and replacement, which occurred throughout the diagenetic history of the rocks, continued in the deep burial environment.

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TABLE OF CONTENTS

LIST OF FIGURES

LIST OF TABLES

I. INTRODUCTION

A. OBJECTIVES

The Winnipegosis Formation is now recognized as a prolific hydrocarbon reservoir in the subsurface of Saskatchewan. This study will document the depositional and diagenetic history of rocks in the Winnipegosis Formation in southeastern Saskatchewan in order to increase understanding of this complex reservoir.

The specific objectives are as follows:

1) to reconstruct an integrated depositional model, emphasizing correlation between members in the Formation;

2) to identify the major diagenetic phases in the rocks and determine their stratigraphic distribution, timing, and environment of formation;

3) to reconstruct the diagenetic history of the rocks.

B. LOCATION OF STUDY AREA

The study area is located in the subsurface of southeastern Saskatchewan near the town of Estevan (Fig. 1). The area extends from Township 1 to Township 4 and across Ranges 4 through 10 west of the second meridian (102° longitude).

C. METHODS

Six core (194 m) were logged in detail and sampled for thin sections at intervals of 1.5 to 5 m. The core locations were specifically chosen to include relatively complete sequences with representative sections through all three members of the Winnipegosis formation. Preference was given to locations that exhibited transitions from limestone to dolostone within the formation.

Polished core and core samples were examined under a binocular microscope. One hundred and fifty thin sections were surface impregnated with Rhodomin-B epoxy and then

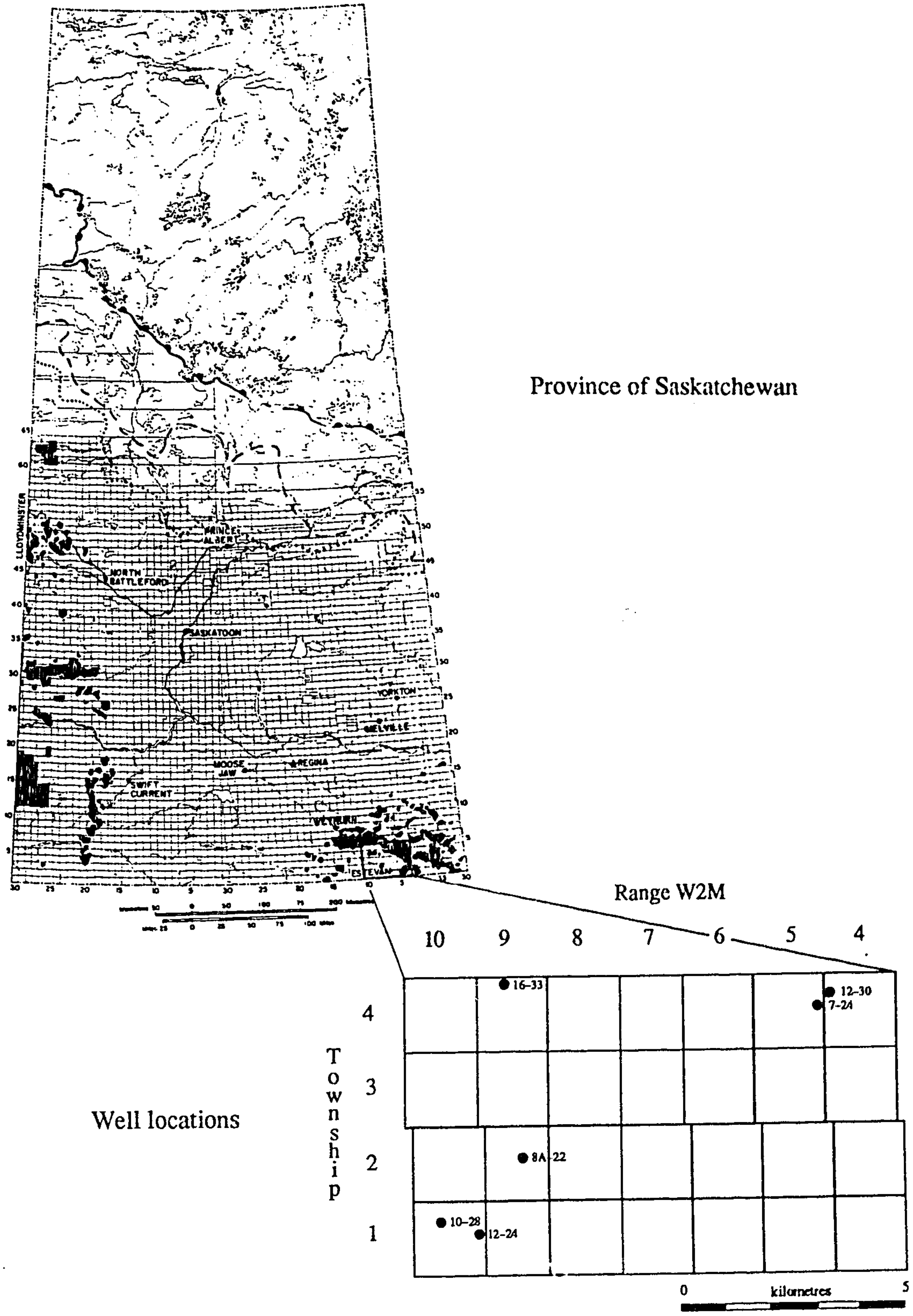


Figure 1. Map showing location of study area, modified from Simpson (1984).

finely polished. All thin sections were stained with Alizarin Red-S to distinguish between dolomite and calcite (Evamy, 1963) and potassium-ferricyanide to distinguish between ferroan and nonferroan carbonates (Dickson, 1965).

All thin sections were examined on a conventional petrographic microscope, a Jenamed™ fluorescent microscope, and a Technosyn™ cold cathodoluminescent microscope (model 8200 Mark II). The fluorescent and cathodoluminescent microscopes were used to assist in differenciation of diagenetic mineral phases. The fluorescent microscope employed blue light excitation. Operating procedures for the cathodoluminescence microscope were kept as constant as possible throughout the study to allow qualitative comparison of color and intensity of emission between samples. The operating conditions for cathodoluminescence work were 15-20 Kv voltage, 500-600 µA gun current, and a 0.05-0.10 torr operating vacuum.

Fifteen samples were chosen for X-ray diffraction analysis to determine dolomite stoichiometry. After washing off the drilling mud, the samples were manually crushed with a mortar and pestle. Each sample was spiked with an internal quartz standard NBS-28 and mounted on a glass slide with alcohol. The analyses were run on a Phillips™ X-ray diffractometer operated at 50 kV and 20 mA, using CoK_{∞} radiation (wavelength of 1.790210 angstroms). Samples were scanned from 27°-42° at 1°2 theda per minute.

Dolomite nonstoichiometry was calculated by measuring the $d_{(104)}$ spacing of the dolomite relative to a quartz internal standard. Goldsmith and Graf (1958b) determined that this d-spacing varies linearly with mole % $CaCO_3$. The relationship between the d-spacing and composition was derived from the linear equation in Lumsden (1979).

D. STRATIGRAPHY

Introduction

In southeastern Saskatchewan, the Devonian Elk Point Group consists of the Ashern, Winnipegoisis, and Prairie formations, in ascending order. The Elk Point Group contains prolific hydrocarbon reservoirs and has been studied extensively, with numerous revisions of the stratigraphic nomenclature and boundaries (Fig. 2).

The Elk Point Formation was originally named by McGehee (1949) for the evaporites, dolostones, and shales lying between Cambrian strata and the Upper Devonian Waterways Formation, in exposures in east-central Alberta and western Saskatchewan. The Elk Point Formation was subsequently correlated into Manitoba and elevated to group status by Belyea (1952). In Saskatchewan, the Elk Point Group was divided into the Ashern, Winnipegosis, and Prairie Evaporite formations by Baillie (1953b) which correlate to the Contact Rapids, Keg River, and Muskeg formations, respectively of the Elk Point Group in northern Alberta (Burrowes and Krause, 1987).

Ashern Formation

The Ashern Formation consists of barren, red to orange, argillaceous dolostones (Baillie, 1951). The absence of body fossils in the Ashern Formation makes direct determination of its age impossible. Stratigraphic positioning, however, suggests a Middle Devonian (Eifelian) age for the Ashern Formation in Saskatchewan since correlative strata in Northern Alberta are both overlain and underlain by Eifelian age strata (Lobdell, 1984).

An angular unconformity separates the Ashern Formation from the underlying Silurian Interlake Group throughout Saskatchewan. There are conflicting views, however, on the nature of the boundary between the Ashern Formation and overlying Winnipegosis Formation. This boundary has been described as gradational (Jones, 1965; Fuzesy, 1975; Meijer Drees, 1986), unconformable (Baillie, 1951; Norris and Uyeno, 1971; Norris *et al.*, 1982; Perrin, 1982; Rosenthal, 1987), and paraconformable (Lobdell, 1984).

In the study area a significant change in lithology occurs at the boundary between the Ashern and Winnipegosis formations. The Ashern Formation consists of barren mudstones (dolostone) that contains 5-10% quartz silt. The mudstones are massive to laminated with rare *Zoophycus* and interbeds of mudstone intraclasts (Plate 1A, B). The overlying Winnipegosis Formation differs sharply in composition from the Ashern

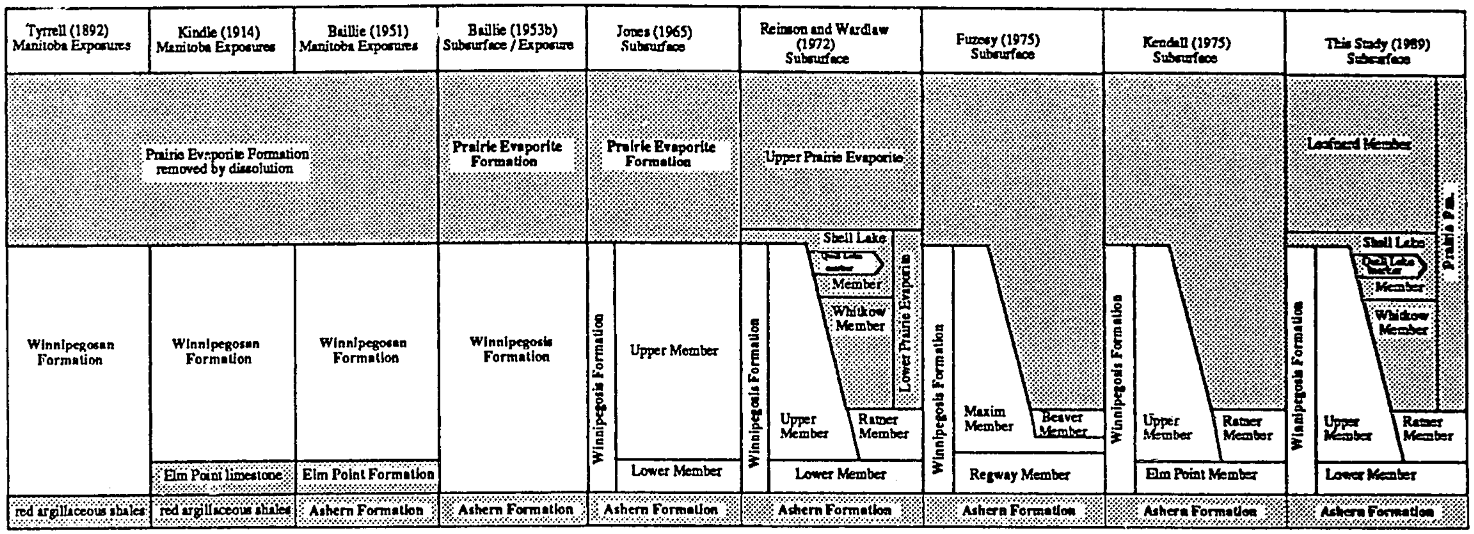


Figure 2. History of stratigraphic nomenclature in the Middle Devonian Elk Point Group in Saskatchewan and Manitoba.

Formation and consists of packstones to mudstones (limestone or dolostone) with a crinoid-brachiopod fauna. Quartz silt is absent in the carbonates of the Winnipegosis Formation. Based on the lithologic change over the boundary (in some cases from dolostone to limestone), the boundary between the Ashern and Winnipegosis Formations in the study area is interpreted as unconformable although the depositional hiatus represented by this boundary was probably short-lived (Perrin, 1982).

Winnipegosis Formation

Tyrrell (1892) first introduced the name "Winnipegosan Formation" for dolostones exposed in Manitoba along the shores and on the islands of Lakes Winnipegosis and Manitoba (Fig. 2). Limestones underlying the Winnipegosan Formation were named the Elm Point Limestone by Kindle (1914) and later renamed the Elm Point Formation by Baillie (1951). Since the Elm Point Formation proved to be undifferenciable in the subsurface, Baillie (1953b) combined the Elm Point and Winnipegosan formations and renamed them the Winnipegosis Formation (Fig. 2). The stratigraphic division known as the "Elm Point Formation" is still used in studies of the Manitoba exposures, although the Elm Point Formation is considered a lateral equivalent of the Winnipegosis Formation (Norris *et al.*, 1982; McCabe, 1987).

Baillie (1951) first recognized that large variations in the thickness of the Winnipegosis Formation in the Manitoba exposures were due to local development of thick, carbonate buildups. The buildups formed cliffs of highly fossiliferous, dense to thick bedded dolostone while the interbuildup strata consisted of low lying, poorly fossiliferous dolostone with local interbeds of shale. Jones (1965) formally divided the Winnipegosis Formation into Lower and Upper members (Fig. 2). The Lower Member consisted of variably bituminous and argillacous carbonates that were sparsely fossiliferous. The top of the member was defined as the top of a regionally developed, 1 to 2 m thick bed of argillaceous and carbonaceous mudstone, that was rich in tentaculitids. The overlying Upper Member varied in composition and Jones (1965) identified three main lithologies:

1) biofragmental carbonates with or without intervals of organic framework, 2) variably fossiliferous carbonates, and 3) finely laminated carbonates. The biofragmental carbonates and variably fossiliferous carbonates formed thick carbonate buildups. In positions between buidups, the Upper Member thinned and was composed predominantly of the finely laminated carbonates lithology.

Reinson and Wardlaw (1972) modified the two-fold division of Jones (1965) to include a third member named the Ratner Member (Fig. 2). Reinson and Wardlaw (1972) redefined the Upper Member of Jones (1965) to exclude the laminated carbonates and then grouped the laminated carbonates with overlying interlaminated carbonates and anhydrites to form the Ratner Member. They did not alter the definition of the Lower Member. Thus, under the three-fold division of Reinson and Wardlaw (1972), the Winnipegosis Formation consisted of the Lower Member which was overlain by either the carbonate buildups of the Upper Member or, the laminated carbonates and evaporites of the Ratner Member, in positions between buildups. The Ratner Member onlaps the flanks of buildups of the Upper Member but was not deposited over the crests of the buildups (Fuller and Porter, 1969a, 1969b; Wardlaw and Reinson, 1971; Reinson and Wardlaw, 1972).

Two alternative schemes for dividing the Winnipegosis Formation into members were advocated by Fuzesy (1975) and Kendall (1975) but they failed to gain widespread acceptance and will not be discussed further.

Several modifications have been made to the three-fold member division of Reinson and Wardlaw (1972). In particular, the use of an argillaceous marker to define the top of the Lower Member has been questioned. The first problem is that there are several argillaceous intervals in the Lower Member which may be mistaken for this marker (Streeton, 1971). The second problem is that the argillaceous marker at the top of the Lower Member is usually absent where overlain by the Upper Member (Kendall, 1975; Rosenthal, 1987). The marker grades laterally into flanking beds of buildups of the Upper Member (Kendall, 1975). Since the argillaceous interval at the top of the Lower Member is laterally equivalent to the Upper Member, it cannot be used to define the boundary between the two members. In the study area, the boundary between the Lower and Upper members is placed at the change in lithology from crinoid-brachiopod packestones-mudstones to overlying, algae-peloid packestones, floatstones, and grainstones. The argillaceous marker at the top of the Lower Member is included with the overlying Ratner Member following Norris *et al.* (1982). Apart from these minor modifications, this study follows the division of the Winnipegosis Formation outlined in Reinson and Wardlaw (1972).

Brachiopod and conodont zonations indicate that the Winnipegoisis Formation is late Eifelian to early Givetian in age (Norris and Uyeno, 1971; Norris *et al.*, 1982). The boundary between the Winnipegosis Formation and the overlying Prairie Formation is placed at the change in lithology from carbonates, or interlaminated carbonates and evaporites, to overlying massive evaporites (Reinson and Wardlaw, 1972). The contact is commonly sharp where the Prairie Formation overlies the Upper Member but is gradational in positions overlying the Ratner Member (Holter, 1969).

Prairie Formation

The term "Praire Evaporite" was introduced by Baillie (1953a) for halite, anhydrite and potash salts overlying the Winnipegosis Formation. The Prairie Evaporite Formation was formally defined by Baillie (1953b) but was renamed the Prairie Formation to conform to binomial nomenclature by Harris and Mallin (1957). Outside of the study area, the Praire Formation has been locally removed through post-depositional dissolutioning and the Winnipegosis Formation is unconformably overlain by the carbonates and shales of the Dawson Bay Formation.

Holter (1969) identified lower and upper salt units in the Prairie Formation in Saskatchewan. The lower salt consisted of halite and anhydrite whereas the upper salt contained interbedded halite and potash (sylvite and carnallite commonly intermixed with halite). The contact between the two salt units was placed at the base of the lowest potash bed. The lower salt unit filled the basinal areas between buildups in the Winnipegosis Formation and with continued deposition, onlapped the buildups, enclosing them completely in evaporites. In basinal positions the lower salt unit is up to 90 m in thickness but thins to less than 10 m through onlap onto the buildups (Reinson and Wardlaw, 1972; Oglesby, 1987). The upper salt is up to 210 m thick (Meijer Drees, 1986).

The lower salt unit was formally divided into the Whitkow and Shell Lake members by Reinson and Wardlaw (1972). Meijer Drees (1986) designated the upper salt unit as the Leofnard Member. In basinal areas away from buildups in the Winnipegosis Formation, the Whitkow Member consists of halite while the overlying Shell Lake Member is predominantly anhydrite and the boundary between the two members is taken at the change in lithology from halite to overlying anhydrite. Adjacent to buildups, however, the Whitkow Member grades into anhydrite making the Whitkow and Shell Lake members undifferenciable (Reinson and Wardlaw, 1972; Kendall, 1987; McCabe, 1987).

A distinctive carbonate unit named the Quill Lake marker beds (Jordan, 1967, 1968; Reinson and Wardlaw, 1972) occurs in the Whitkow-Shell Lake members adjacent to buildups. The Quill Lake marker beds occur at the stratigraphic level coinciding with the tops of buildups in the Winnipegosis Formation (Fig. 2). They thicken towards the buildups, reaching up to 15 m in thickness, and grade basinward into anhydrite (Wardlaw and Reinson, 1971). The Quill Lake marker beds consist of massive to laminated, peloidal packestones and grainstones with local *Amphipora* wackestones, stromatolites, and pisoids (Jordan, 1967, 1968; Wardlaw and Reinson, 1971; Meijer Drees, 1987).

Stratigraphic positioning between two formations of Givetian age establishes the age of the Prairie Formation as Givetian. The contact with the overlying Dawson Bay Formation is sharp and unconformable (Norris *et al.*, 1982).

E. PALEOGEOGRAPHY

The Winnipegosis Formation was deposited in the Elk Point Basin, an intracratonic basin in the Western Canadian Interior that extended over 3000 km from the North West Territories to North Dakota (Fig. 3). The Elk Point Basin was bounded by the Tathlina Arch to the north, the Transcontinental and Central Montana Arches to the south, the Severn Arch to the east, and the West Alberta Arch to the west (Basset and Stout, 1967; Porter *et al.*, 1982; Podruski *et al.*, 1987). The Peace River Arch and Meadow Lake Escarpment transected the basin (Van Hees, 1958), dividing it into the Northern Alberta, Central Alberta , and Saskatchewan Sub-basins (Grayston *et al.* 1964). In Middle Devonian time, the Severn and Tathlina arches subsided allowing normal marine waters to enter the Elk Point Basin from the northeast and northwest (Williams, 1984; Burrowes and Krause, 1987). The Meadow Lake Escarpment also subsided making the Central Alberta and Saskatchewan Sub-basins a continuous feature.

A eustatic rise in sea level occurred during the Middle Devonian through a series of small transgressive and regressive pulses (Vail *et al.*, 1977; Hallam, 1984; Johnson *et al.*, 1985). Basin deposition was also influenced by noneustatic fall in sea level which led to restriction of the basin and subsequent infill by evaporites of the Prairie Formation. The restriction probably resulted from the growth of a carbonate barrier complex which extended across the northwestern entrance to the basin, during deposition of the Keg River Formation (Basset and Stout, 1967). As was illustrated by Williams (1982), however, other interpretations of basin circulation and restriction are possible, depending on the assumptions made about the eroded eastern margin of the basin.

From paleoclimatic reconstructions, southeastern Saskatchewan was within 5° latitude north of the equator during the Devonian (Habicht, 1979). This paleolatitude should have had prevailing wind directions from the north east, based on modern global wind patterns (Campbell, 1987).

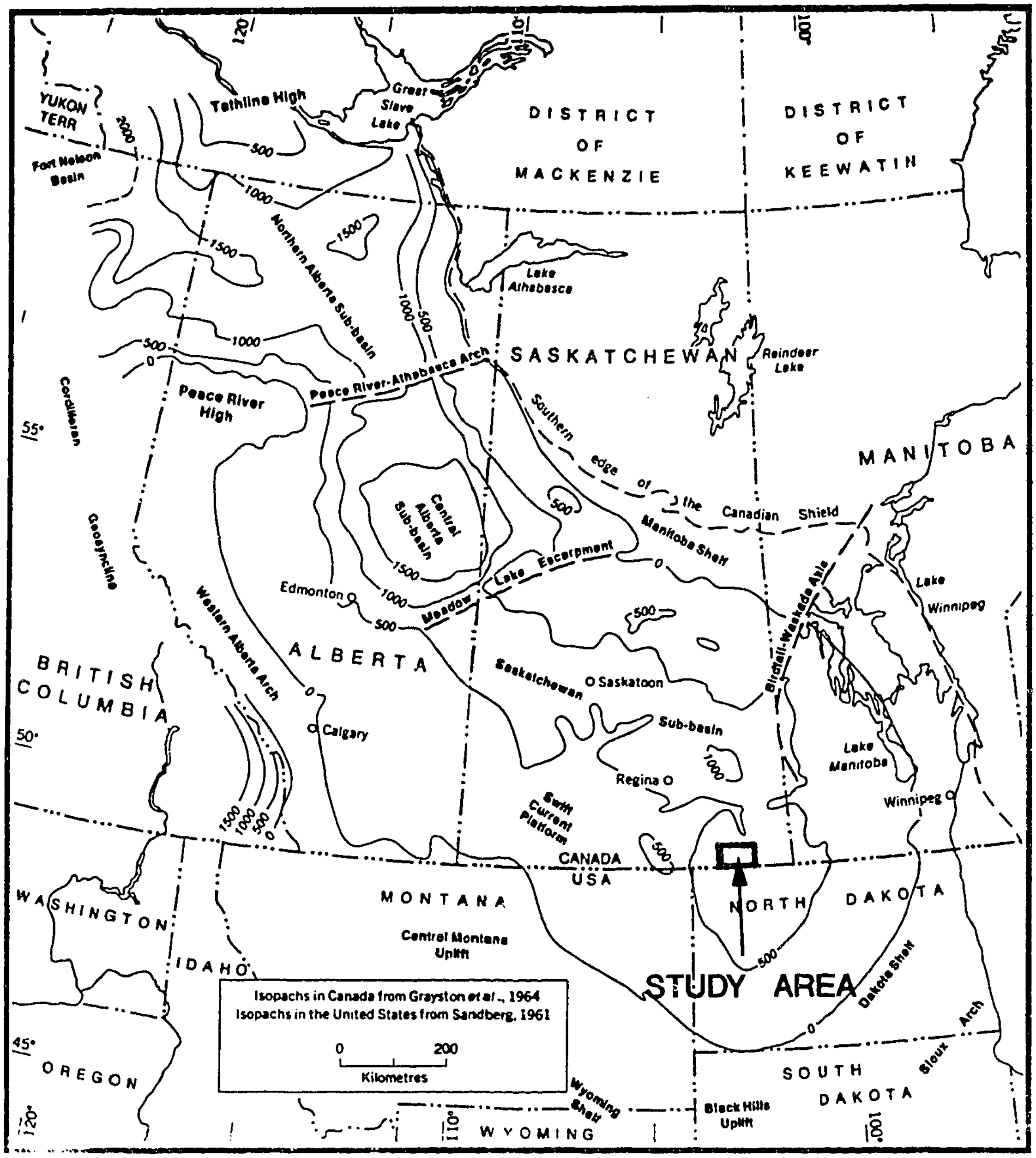


Figure 3. Paleogeographic map of the Devonian Elk Point Basin showing related tectonic features, isopach of Devonian basin fill, and position of study area (Modified from Norris *et al.*, 1982).

F. BURIAL HISTORY

Deposition of the Winnipegosis Formation ceased in the Middle Devonian and the carbonates were encased by the evaporites of the Whitkow and Shell Lake members of the Prairie Formation. Deposition of these evaporites probably coincided with subaerial exposure of the buildups in the Winnipegoisis Formation. The halites in the Whitkow and Shell Lake members have been interpreted as shallow-water deposits (Shearman, 1970; Brodlyo and Spencer, 1987; Kendall, 1984; 1987) although considerable controversy remains concerning water depth during deposition of the Prairie Formation (Oglesby, 1987). This is the only period when the Winnipegoisis Formation could have been directly exposed to meteroic waters since the strata were subsequently buried by the marine carbonates and shales of the Middle Devonian Dawson Bay Formation.

In the study area, the Winnipegosis Formation is currently under 2300-2700 m of overburden. These figures probably represent minimum burial depths since unconformities occur in overlying strata. Post-depositional tectonic disturbance of the Devonian in southeastern Saskatchewan has been minimal. Regional dip is approximately 2.8 m per km to the southwest (McCabe, 1987).

G. CLASSIFICATION SYSTEMS AND TERMINOLOGY

To avoid ambiguity over the intended meaning of terms, the classification schemes and definitions used in this study are specified below.

Depositional textures in the rocks are classified according to Embry and Klovan (1971). In petrographic descriptions, the crystal size divisions follow Folk (1959) with crystal shapes classified after Friedman (1965). Anhydrite fabrics and structures defined by Maiklem *et al.* (1969) are used and cement fabrics follow the classification by Flugel (1982). The porosity classification follows Choquette and Pray (1970).

The definitions of terms are used in this study are as follows.

Buildup - a non-genetic term used to describe a laterally restricted body of carbonate rock which differs in composition and internal fabric from surrounding and overlying deposits and displays topographic relief above equivalent, typically thinner, strata (Heckel, 1974). Buildups include reefs, mounds, banks, and composite features.

Reef - a massive or layered buildup which formed through organic accumulation and was potentially wave resistant, being stabilized syndepositionally by organic framework and/or submarine cementation (James, 1987).

Mound- a buildup lacking large skeletal metazoans, which formed through organic accumulation rather than purely mechanical accumulation (Wilson, 1975).

Bank - a buildup accumulated through mechanical piling of detrital carbonate sediment by waves and currents and partly by trapping or baffling (Wilson, 1975).

Recrystallization- a diagenetic process affecting a single mineral species in which the mineralogy remains unchanged after the reaction, apart from minor adjustments in trace element or isotope abundances (Folk, 1965). An increase in crystal size or a change in crystal shape are common examples of recrystallization.

Replacement- a diagenetic process affecting a single mineral species in which a new mineral exists after the reaction (Folk, 1965). Dolomitization of limestone is a common example of replacement.

Shallow burial- a near-surface environment in which diagenetic processes are dominated by surface-derived waters. Shallow burial is equivalent to the eogenetic zone defined by Choqette and Pray (1970).

Deep burial- an environment below the zone of near-surface diagenesis (shallow burial) and above the realm of low-grade metamorphism (Choquette and James, 1987). Diagenetic processes in deep burial environments are dominated by connate and compactional waters. Deep burial is equivalent to the mesogenetic zone defined by Choquette and Pray (1970).

II. SEDIMENTOLOGY

A. INTRODUCTION

During the Middle Devonian, the Elk Point Basin of the Western Canada Interior was the site of extensive carbonate sedimentation. Shallow carbonate platforms developed around the basin perimeter while carbonate buildups and lime mud accumulated in the basin interior. The study area occurs in the interior of the Saskatchewan Sub-basin. Using the position of the platform margins in the Winnipegosis Formation as defined by Ehrets and Kissling (1987), the study area was approximately 20 km basinward of the western platform margin.

B. LOWER MEMBER

Introduction

The Lower Member is 7-12 m thick in the study area and laterally continuous, underlying both the buildups of the Upper Member and the basinal strata of the Ratner Member. The Lower Member is commonly 3-4 m thicker in positions underlying the Upper Member than in positions underlying the Ratner Member (Fig. 4).

Several metres of topographic relief may have existed on the sea floor after deposition of the Lower Member, with enhanced water circulation on the topographic highs promoting organic colonization. Thickening of the Lower Member in positions underlying buildups of the Upper Member was also noted by Rosenthal (1987) in the Manitoba exposures, and by Langston and Chin (1968) in the stratigraphically equivalent Keg River Formation in northern Alberta. Better well control is needed in the study area, however, to verify that increases in the thickness of the Lower Member are not equally common in positions overlain by the Ratner Member.

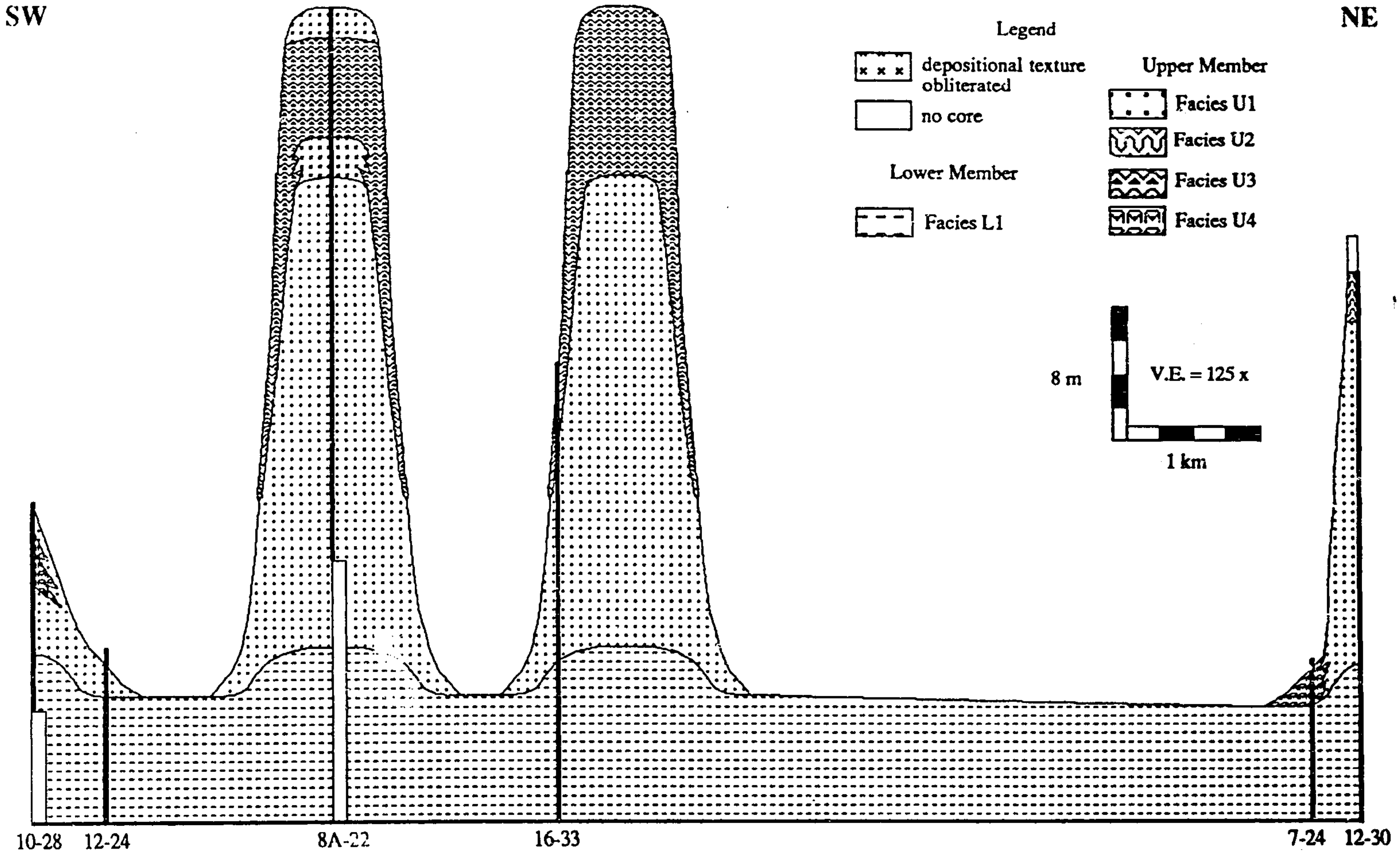


Figure 4. Stratigraphic cross-section showing facies distributions in the Upper and Lower members of the Winnipegosis Formation. Datum is base of the Lower Member. Interpolation between core locations is based on the following assumptions: 1) buildups locally grew to a similar height (McCabe, 1987); 2) the same facies patterns and geometries are repeated throughout the study area. See Figure 1 for map of well locations.

Facies descriptions

In the study area, the Lower Member consists of crinoid-brachiopod mudstone to packstone, designated as facies L1. A detailed description of the facies is given in Table 1.

Facies L1 commonly exhibits a distinctive mottled texture which is interpreted as a product of burrowing which has undergone extensive diagenetic alteration (Plate 1C, D). The burrows are predominantly horizontal, 1-2 cm in diameter, and up to 10 cm long. The burrows have undergone preferential pressure-dissolution and recrystallization which suggests that they differed in composition and/or texture from the surrounding carbonate groundmass. The severe diagenetic alteration does not permit identification of specific ichnogenera in the study area. Elsewhere in the basin, however, *Chondrites*, *Paleophycus*, and *Thalassinoides* have been identified in the Lower Member (Ehrets and Kissling, 1987; Rosenthal, 1987).

Environmental interpretation

The Lower Member is relatively homogeneous in the study area, suggesting that depth and energy conditions remained constant during deposition. Crinoids and brachiopods are the dominant skeletal elements and this assemblage is characteristic of open water conditions with normal marine salinities (Jamieson, 1970; Wilson, 1975). The abundant carbonate mud and poor sorting attest to minimal reworking by waves or currents and deposition of the Lower Member probably occurred in a low energy environment below fair weather wave base. The predominance of biogenic structures over wave or current structures in the Lower Member supports deposition under relatively deep, subtidal conditions (Rosenthal, 1987).

The Lower Member was deposited throughout the Saskatchewan Sub-basin (Edie, 1952). A similar stratigraphically equivalent deposit also occurs in the remainder of the Elk Point Basin underlying buildups in the Keg River Formation (Langston and Chin, 1968). Isopach maps of the Lower Member show an absence of pronounced regional variations in

Facies	Rock type	Accessory mineralogy	Allochems	Fauna		Sedimentary structures	
				major	minor	syngenetic	epigenetic
L1. Crinoid-brachiopod wackestones and packestones							
Subfacies a	dolostone	siliclastic clays	skeletal material (C-S)	crinoids brachiopods	pelecypods fenestrate bryozoans colonial corals *Tentaculites* trilobites sponge spicules	—	nodular bedding burrows?
Subfacies b	dolomitic limestone	siliclastic clays 10-20% dolomite	as above	as above	as above		as above
U1. Codiacean algae-peloid packestones and rare grainstones and floatstones.							
Subfacies a	dolostone	—	skeletal material (A) peloids (C)	codiacean algae	pelcyepods dasycladean algae bryozoans gastropods crinoids colonial corals branching stromatoporoids	—	—
Subfacies b	dolomitic limestone	10-20% dolomite	as above	as above	as above		
U2. *Renalcis?* boundstones and floatstones in a peloid grainstone groundmass	dolostone	—	skeletal material (A) peloids (A)	*Renalcis?*	gastropods ostracodes calcispheres	planar beds with apparent dips of 35-40°	—
U3. Coral - stromatolite boundstones, floatstones, and framestones.	dolostone	—	skeletal material (A) peloids (S)	*Thamnopora* cyanobacteria	pelecypods gastropods encrusting bryozoans Syringopora fenestrate bryozoans	—	—
U4. Oncoid floatstones in a peloid wackstone to packstone groundmass							
Subfacies a	dolostone	—	oncoids (C) peloids (C) skeletal material (S) intraclasts (R)	Oncoid nuclei of intraclasts, and fragments of brachiopods, green algae and colonial corals. Oncoid cortices contain *Sphaerocodium* and *Girvanella* Groundmass contains *Tentaculites*, *Styliolina*, gastropods, and crinoids		—	—
Subfacies b	dolomitic limestone	10-20% dolomite	as above	as above			

Table 1. Facies descriptions for the Lower and Upper members. Abbreviations: A- abundant, C- common, S- sparse, R - rare.

thickness, suggesting that deposition occurred on a level bottom or gently sloping ramp (Perrin, 1982; Ehrets and Kissling, 1987).

The fauna, depositional textures, and regional distribution of the Lower Member suggest deposition occurred on an open marine carbonate ramp, which was below fair weather wave base in the study area.

C. UPPER MEMBER

Introduction

The Upper Member forms buildups overlying the Lower Member. Published seismic surveys show that the buildups in Saskatchewan are 0.5 to 6 km in diameter with steep margins and flat tops (Gendzwill, 1978). McCabe (1987, 1988) argued that buildups in the Winnipegosis Formation locally grew to a uniform height, with regional variations in thickness as a function of positioning in the basin. In southeastern Saskatchewan, the location of this study, the buildups in the Winnipegosis Formation are estimated to be 70 m thick (Martindale and Orr, 1988b). Northwest of the study area, near Saskatoon, the buildups are over 90 m thick (Wilson, 1985).

The Upper Member exhibits abrupt variations in thickness in the study area. In 12-30-4-4w2M, the Upper Member is 45 m thick while less than 1 km away in 7-24-4-5w2M, the Upper Member is only 3.5 m thick (Fig. 4). Beds flanking a buildup core become progressively thinner towards the buildup to basin transition and the abrupt thinnning observed in the Upper Member may indicate penetration of flanking beds.

Facies descriptions

In the study area, the Upper Member can be divided into 4 facies, herein designated as U1 to U4. Detailed descriptions of the faces are given in Table 1.

Codiacean green algae, including *Litanaia* and *Lanicula*, are a major component of facies U1 (Plate 2A) and also occur in facies U4 (Plate 3). Ancient codiacean algae were probably similar to modern codiacean algae which are a morphologically diverse group of erect, benthic marine plants (Bathurst, 1975; Babcock, 1986).

The codiaceans have a tubular thallus (body) consisting of 4-10 loosely interwoven tubular filaments surrounded by a peripheral meshwork of smaller branching filaments. Both the thalli and tubular filaments are round in cross-section. According to Johnson (1964), *Litanaia* and *Lanicula* are morphologically similar except for the shape of the thallus. The thallus in *Litanaia* is straight-sided to slightly irregular in longitudinal section, whereas that of *Lanicula* consists of stacked, bowl shaped elements in which the top of the bowl is approximately twice the diameter of the base.

In facies U2, an encrusting organism composed of irregular, micrite-walled chambers is tentatively identified as the cyanobacteria *Renalcis* (Plate 1F, 2C). The identification remains tentative due to poor preservation of the organism. Extensive pressure-dissolution and recrystallization hampers analysis of chamber morphology but several well-preserved chambers reach 125-500 μm in diameter (Plate 2D). These features resemble *Renalcis* which has a simple internal structure composed of micrite-walled chambers ranging from 50-500 μm in diameter, and commonly has an encrusting habit (Pratt, 1982).

The branching coral in facies U3 are encrusted and intergrown with laminar to hemispherical stromatolites and rare encrusting stromatoporoids and bryozoans (Plate 2B). The organic framework is commonly brecciated but appears to be locally *in situ* in 8A–22-2–9w2M between 2750-2780 m depth. Depositional fabric in the groundmass of facies U3 is obscured by extensive recrystallization and leaching.

Environmental interpretation

Facies U1 with local interbeds of facies U2 is a major constituent of the buildups in the study area (Fig. 4). Facies U1 occurs at the base of the buildups and may reach over 30 m in thickness.

The abundance of lime mud and poor sorting (Plate 1E) suggest that facies U1 accumulated *in situ* with thorough homogenization of the sediment through bioturbation.

Deposition probably occurred below fair weather wavebase since facies U1 typically lacks evidence of current or wave reworking.

There is, however, rare development of grainstone fabrics in facies U1. The grainstones grade into packestones and are not present as distinct beds. The local presence of grainstone fabrics is interpreted to result from winnowing by normal (thermohaline) bottom currents. The gradation from grainstone to packestone suggests the grainstones are unrelated to storm or resedimentation events.

The allochem assemblage in facies U1 is dominated by codiacean algae and peloids. Modern codiacean algae inhabit water depths of less than 1 m to over 100 m and ancient codiaceans probably existed over a similar range in depth (Jamieson, 1970). Hence, the presence of codiacean algae is not indicative of growth in a specific water depth. The paucity of robust, frame-building organisms in facies U1 is significant. James (1983) argued that without the presence of frame-building organisms, a buildup cannot exist in the zone of constant turbulence, usually wave induced, because the smaller and more delicate organisms are broken and swept away. The fauna thus supports growth below wavebase.

The peloids in facies U1 are nonuniform in size and shape. This nonuniformity is characteristic of peloids with polygenetic origins (Beales, 1965). Disaggregation of codiacean algae produces peloids (Wolf, 1965) and the abundance of associated codiacean algae supports this origin for some of the peloids in facies U1. Other potential origins of peloids include disaggregation of calcified cyanobacteria (Coniglio and James, 1985), disaggregation of cyanobacterial-bound mud (Flugel, 1982), direct precipitation (Macintyre, 1985; Chafetz, 1986; Reid, 1987), accumulation of fecal pellets, micritization of skeletal fragments (Bathurst, 1966), and obliteration of internal structures in skeletal fragments through recrystallization (Blatt *et al.*, 1972) or dolomitization.

The accumulations of facies U1 are best described as mounds, which implies autochthonous, organic accumulation of sediment (Wilson, 1975). Sedimentary fabrics and mound fauna suggest growth occurred below fair weather wave base with local winnowing of mound sediment by normal bottom currents. The closest modern analogues to the facies U1 mounds are *Halimeda* buildups growing on the outer shelf of the nothern Great Barrier Reef and on the Shonda Shelf slope in the eastern Java Sea. The modern buildups are 20-50 m thick and composed predominantly of fragments of the codiacean algae *Halimeda* with minor amounts of foraminifera, gastropods, bivalves, echinoids, corals, and bryozoans (Davies and Marshall, 1985; Orme, 1985). The buildups are *in situ* organic accumulations growing in low energy, shelf to slope environments at water depths of 20 to 100 m (Davies and Marshall, 1985; Drew and Abel, 1985; Phipps *et al.*, 1985; Roberts *et al.*, 1985).

An environmental interpretation of the facies U1 mounds must also account for the local presence of interbeds of facies U2. Post-depositional tectonic disturbance of the Devonian in Saskatchewan has been minimal and thus, the apparent dips of 30-40° exibited by the bedding in facies U2 are indicative of deposition on an inclined surface. The encrusting *Renalcis?* probably colonized the flanks of the mounds of facies U1. A similar example of cyanobacterial colonization of the flanks of Devonian-age buildups was documented by Playford and Cockbain (1969) in the Canning Basin, Western Australia.

Facies U3 occurs on the flanks and crests of the mounds of facies U1 and U2 (Fig. 4). The branching corals, stromatolites, bryozoans, and stromatoporoids in facies U3 are intergrown into boundstone and framestone fabrics which could provide structural resistance to turbulent water. The organic framework in facies U3 is commonly brecciated, forming floatstone fabrics which is a strong indication of growth in the surf zone (James and Macintyre, 1985). Martindale and Orr (1988b) noted that branching corals generally occur towards the interior of the buildup, with stromatolites and red algae forming the higher energy, crest. This differenciation into higher energy crest and protected interior also demonstrates that the buildups had grown into wavebase during deposition of facies U3.

In locations where facies U4 occurs, the Upper Member is less than 5 m thick (Fig. 4). This dramatic thinning, alone, suggests that facies U4 occurs within the transition from buildup to basinal deposition. Facies relationships support this interpretation. Facies U4 is gradational between buildup and basin deposits. It interfingers with facies U1 towards the buildups (Fig. 4) while in a basinward directions, it interfingers with and is gradationally overlain by facies R1 of the Ratner Member (Fig. 5).

Oncoids are the dominant allochem in facies U4 and the oncoid cortices are concentric, indicating they formed through uninterrupted growth with continuous rolling in currents (Flugel, 1982). Some of the oncoids are abraded (Plate 3), suggesting that they have been reworked and/or transported. The oncoids, however, are associated with tentaculitids and accumulation of these delicate, planktonic organisms would require settling from quiet waters. The tentaculitid shells would probably be destroyed during transport and they are interpreted as autochthonous deposits.

Facies U4 is characterized by intermixing of allochems which formed in turbulent, well oxygenated waters and delicate allochems which settled from water column. The oncoids and other coarse skeletal debris in facies U4 probably formed in shallow, higher energy waters on the buildups and were transported down the buildup slopes into deeper, low energy waters. The tentaculitids settled out from the slowly circulating basinal waters in periods between influxes of buildup-derived debris. Mixtures of authochthonous basinal material and allochthonous debris derived from the buildup are characteristic of deposits marking the buildup to basin transition (Clough and Blodgett, 1984). Sedimentologic evidence thus supports the interpretation of facies U4 as a buildup to basin transition initially proposed on the basis of abrupt thinning of the Upper Member and facies relationships.

D. RATNER MEMBER

Introduction

The Ratner Member is laterally discontinuous and overlies the Lower Member in positions between buildups of the Upper Member. The Ratner Member onlaps buildup flanks but was not deposited over the crests of buildups (Jones, 1965; Wardlaw and Reinson, 1972; Kendall, 1975). In the study area, the Ratner Member increases in thickness from 7 m in the east to 15 m in the west (Fig. 5).

Facies descriptions

In the study area, the Ratner Member is divided into facies R1 to R5. Detailed descriptions of the facies are given in Table 2.

Gamma-ray logs suggest that siliclastic clays are present in the Ratner Member. The presence of clays was verifed by XRD analysis and dissolution of rock samples in hot HCl acid. Laminations and disseminated blebs of micritic, organic-rich material are also abundant (Plate 4A). Edie (1952) first recognized the high organic content in the Ratner Member and subsequent analyses by Wardlaw and Reinson (1971) showed that the Ratner Member contains up to 5 weight % organic carbon. Biomarker correlation of crude oils suggests that organic-rich strata in the Ratner Member are the source of hydrocarbons trapped in buildups of the Upper Member (Brooks *et al.*, 1988; Martindale and Orr, 1988b).

Facies R1 occurs at the base of the Ratner Member, directly overlying the Lower Member (Fig. 5). On buildup flanks, facies R1 interfingers with and gradationally overlies facies U1 and U4.

Environmental interpretation

Settling from suspension was the dominant depositional process in the Ratner Member and produced the characteristic laminated fabric. Tentaculitids, uncalcified organic material, siliclastic clays, and lime mud in facies R1 settled out from the water column in a low energy environment, undisturbed by waves or currents. The laminations are

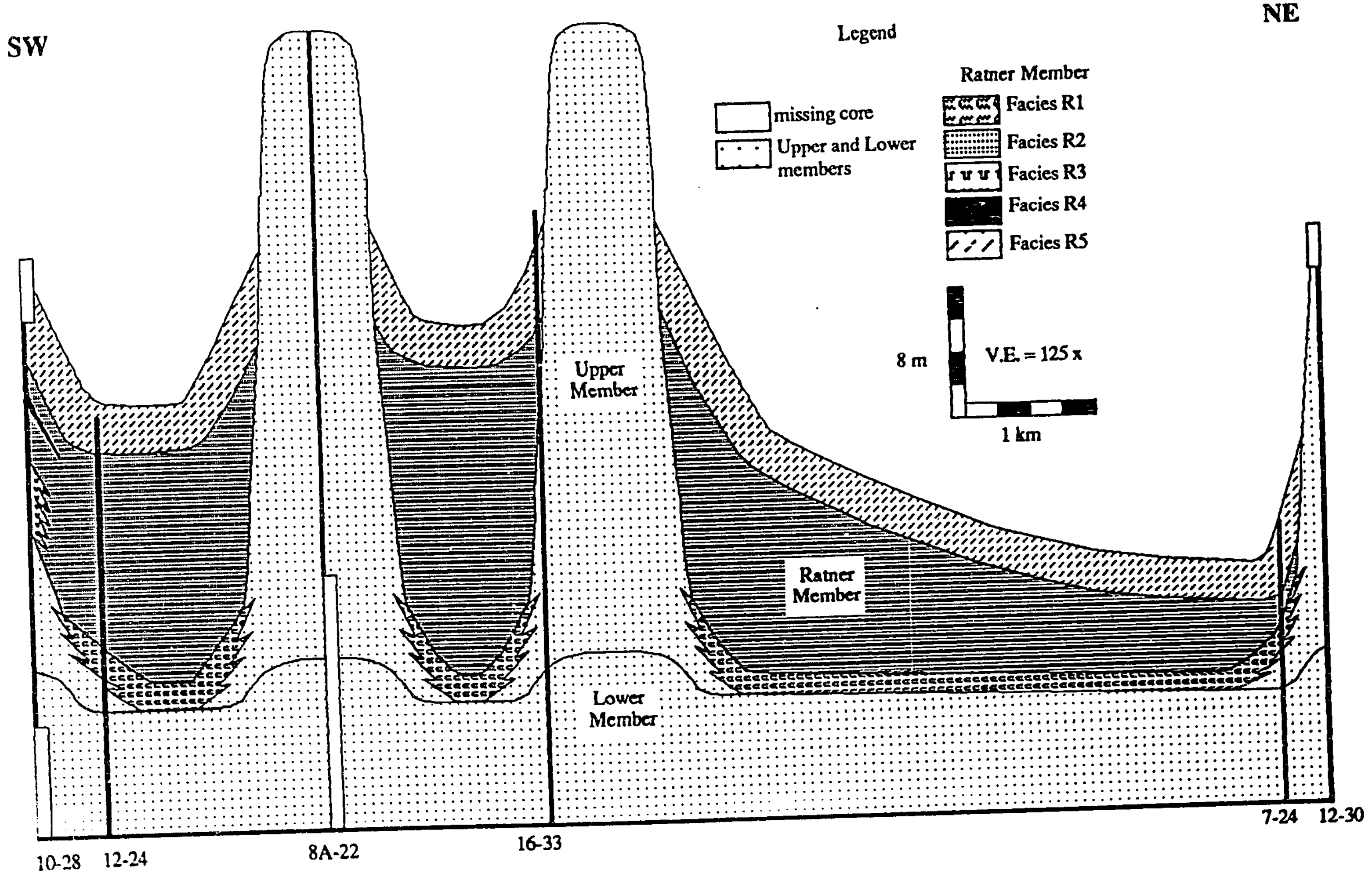


Figure 5. Stratigraphic cross-section showing facies distributions in the Ratner Member of the Winnipegosis Formation. Datum is base of Lower Member. Interpolation between core locations is based on the assumption that facies patterns and geometries are repeated throughout the study area. See Figure 1 for map of well locations.

Facies	Rock type	Accessory mineralogy	Allochems	Fauna		Sedimentary structures	
				major	minor	syngenetic	epigenetic
R1. Tentaculitid mudstones							
Subfacies a	dolostone	organic-rich material	skeletal material (S)	*Tentaculites* *Styliolina*	—	planar to slightly wavy laminations	—
Subfacies b	dolomitic limestone	organic-rich material 10-20% dolomite	as above	as above			
R2. Laminated mudstones							
Subfacies a	dolostone	organic-rich material 1- 5% anhydrite	—	—	—	planar to slightly wavy interlaminations of organic-rich material	anhydrite nodules
Subfacies b	dolomitic limestone	organic-rich material 10-20% dolomite					—
R3. Mottled mudstones	dolostone	organic-rich material 1-5% anhydrite	skeletal material (R)	articulated crinoid stems	—	—	mottling
R4. Peloidal packestones	dolostone	—	peloids (C)	—	—	3-5 cm thick, sharply-based beds	—
R5. Interlaminated mudstones and anhydrite	dolostone	5-30% anhydrite organic-rich material	—	—	—	planar to convolute laminations	anhydrite nodules

Table 2. Facies descriptions for the Ratner Member. Abbreviations: A- abundant, C- common, S- sparse, R- rare.

undisturbed by bioturbation and benthic organisms are notably absent. These features, together with the preservation of organic material provide evidence for the existence of anoxic conditions at and below the sediment surface (Demaison and Moore, 1980).

The presence of planktonic organisms in facies R1, however, and interfingering of facies R1 with buildup debris demonstrate that anoxic water conditions did not exist throughout the entire water column. Shallow, oxygenated waters supported buildup growth and planktonic organisms while bottom waters in the basin were anoxic. This implies an absence of mixing within the water column and the presence of water stratification.

It is difficult to establish the water depth during deposition of facies R1. Yochelson and Lindeman (1986) pointed out that mudstones and wackestones containing tentaculitids are commonly interpreted as deep water deposits, yet there is no evidence that tentaculitids were restricted to pelagic habitats. Stratigraphic evidence is also inconclusive. Kendall (1975) argued that the height of buildups provides an estimate of depositional water depths for laterally equivalent strata and on this basis facies R1 would have accumulated in water depths of 60-70 m. This approximation is probably invalid, however, since it assumes that the buildups had grown into wave base during deposition of equivalent strata and it also ignores the effects of mechanical compaction.

Facies R2 gradationally overlies facies R1 (Fig. 5) and differs from facies R1 in that tentaculitids are absent. The preservation of organic material in facies R2 and the absence of body fossils or bioturbation indicates that anoxic conditions persisted at and below the sediment surface during deposition of facies R2.

Interlaminations of organic-rich material and carbonate mud in the Ratner Member, analagous to facies R2, were interpreted as shallow water deposits by Fuller and Porter (1969a, 1969b) and Shearman and Fuller (1969) based on a resemblence to modern stromatolites from shallow subtidal to intertidal environments along the Trucial Coast. Wardlaw and Reinson (1971) and Davies and Ludlam (1973), however, demonstrated that individual laminations in the Ratner Member could be correlated over distances of up to 25 km. These authors argued that laterally continous laminations were inconsistent with a shallow water origin and advocated a low energy, subtidal environment of deposition. By analogy, a similar depositional environment is proposed for the laminated mudstones of facies R2.

Facies R2 locally grades into facies R3 (Fig. 5) which is similar to facies R2 except that laminations in the mudstones are locally disrupted by very poorly preserved burrows (Plate 4E). The presence of burrowers attests to oxygenated conditions on the basin floor but the paucity of burrows and lack of epifauna are typical of deposition under oxygen-reduced conditions in the water column and within the sediment (Stoakes and Creany, 1985). Facies R3 is locally developed in locations where the Ratner Member onlaps buildup flanks (Fig. 5). This suggests that slightly shallower waters on the flanks of the buildups may have been better oxygenated than waters deeper in the basin. Facies R3 grades upwards into facies R2, however, showing that oxygenated conditions did not persist and these locations were engulfed by anoxic waters during later deposition.

Interpretation of facies R2 must also account for the presence of thin (3-5 cm), discontinuous interbeds of facies R4. The stratigraphic relationship between the two facies suggests that deposition of facies R4 packestones periodically interrupted the background sedimentation of facies R2 mudstones. Although upper contacts on interbeds of facies R4 are gradational, the basal contacts are sharp (Plate 4C, D) which indicates rapid deposition (Kriesa, 1981). Deposition during waning flow conditions may be inferred for beds with sharp basal contacts and gradational upper contacts, even though normal grading is not well developed (Collinson and Thompson, 1982). Sedimentary structures thus indicate that episodes of facies R4 deposition occurred rapidly, under waning flow conditions.

The peloids and skeletal fragments comprising facies R4 clearly formed in a well oxygenated environment, in contrast to the surrounding barren mudstones of facies R2 which accumulated under anoxic conditions. Interbeds of facies R4 clearly represent allochthonous sediment which was periodically transported into the anoxic basin environment. Although preservation of facies R4 allochems is poor, the allochem assemblage of peloids and skeletal fragments resembles that of facies U1 of the buildups. These factors, together with the fact that facies R4 was only observed in positions where the Ratner Member onlaps buildup flanks (Fig. 5), suggest that facies R4 sediment was derived from buildups.

Transport of sediment from the buildups into the adjacent basin would probably be gravity-induced but facies R4 has none of the characteristics of slumps, debris flows, and grains flows which are typically poorly sorted and/or exhibit reverse grading (Selley, 1976). Facies R4 does bear evidence of deposition under waning flow conditions and this is characterisitic of sedimentation from turbid flows (Walker, 1984). Only the massive A division of the Bouma sequence can be recognized in facies R4 but this is common in carbonate turbidites and is primarily due to poor preservation (McIlreath and James, 1984). Carbonate turbidites were also identified in the Ratner Member in northern Alberta by Davies and Ludlam (1973).

The laminations of mudstone and organic-rich material in facies R5 strongly resemble those of facies R2 and probably formed under similar conditions i.e. a low energy, anoxic, subtidal environment. The presence of anhydrite laminations, however, is unique to facies R5. Like the mudstone and organic-rich laminations, anhydrite laminations in the Ratner Member can be correlated over distances of 25 km (Davies and Ludlam, 1973) to 160 km (Wardlaw and Reinson, 1971). Thus, the laterally continuous anhydrite laminations are also interpreted as sedimentary accumulations, deposited through uniform input and settling of particles over a wide area. Given the vast areal extent of deposition, a detrital source for the anhydrite in facies R5 is unlikely. Instead, the widespread anhydrite deposition in the Ratner Member probably resulted from direct precipitation of calcium sulfate from concentrated seawater (Klingspor, 1969; Wardlaw and Reinson, 1971; Davies and Ludlam, 1973).

The origin of the lime mud in the Ratner Member is constrained by the Devonian age of the strata. Planktonic foraminifers and coccoliths are the primary source of carbonate mud formed in modern basinal environments but these organisms probably did not evolve until the Jurassic (McIlreath and James, 1984). Lime mud is the dominant constituent of the Ratner Member and the westward increase in thickness towards the platform margin (Fig. 5) suggests the lime mud was sourced from the platform. Prolific quantities of carbonate mud are commonly generated on carbonate platforms, principally through bioerosion and physical disintegration of green algae (Bathurst, 1975). The lime mud could have been transported from the platform to the basin by either storm currents or tides. In the study area, no sedimentary structures are preserved in the sequence to assist in discriminating between these transport processes.

The organic-rich laminations in the Ratner Member are probably comprised of uncalcified, pelagic organisms which inhabited the oxygenated surface waters Ratner Member deposition. Accumulation of this organic-rich material represents autochthonous basinal sedimentation which was periodically interrupted by influxes of allochthonous lime mud. In later stages of deposition, basin waters became sufficiently concentrated to precipitate calcium sulfate.

Stratification of basin waters created anoxic bottom conditions during deposition of the Ratner Member. Potential causes of water stratification include changes in suspended sediment content, temperature, or salinity (Sonnenfeld, 1984). The presence of sedimentary anhydrite in the Ratner Member strongly suggests that water stratification throughout deposition was mainly related to hypersalinity.

E. A DEPOSITIONAL MODEL FOR THE WINNIPEGOSIS FORMATION

Deposition of the Winnipegosis Formation occurred in Middle Devonian time, during which an overall rise in eustatic sea level occurred (Vail *et al.*, 1977; Hallam, 1984; Johnson *et al.*, 1985). Deposition in the Elk Point Basin, however, was greatly influenced by noneustatic restriction which caused a relative drop in sea level throughout the basin. The change from open marine to restricted marine conditions is represented by the Winnipegosis Formation. Deposition of the Winnipegosis Formation was initiated in an open marine shelf environment and was terminated in an anoxic, hypersaline basin environment.

Deposition of the Winnipegosis Formation began with the Lower Member which accumulated under open marine conditions on a carbonate ramp. The ramp was below fair weather wave base in the study area and a rapid rise of water depth in the basin probably triggered the growth of buildups of the Upper Member on the ramp. Localization of the buildups may have been influenced by the presence of subtle paleotopographic highs in the underlying Lower Member.

Buildup growth initiated with accumulation of codiacean algae mounds (facies U1 and U2) in a relatively deep subtidal environment influenced by normal bottom currents. With growth into wave base, a reefal framework dominated by corals and stromatolites (facies U3) colonized the crests of the mounds. Turbulent conditions led to extensive brecciation of the framework. Mixtures of allocthonous buildup sediment and autochthonous basinal sediment (facies U4) accumulated along buildup to basin transitions.

The Ratner Member accumulated on the basin floor under anoxic, hypersaline conditions caused by stratification of the basin waters. The distinctive laminated fabric in the Ratner Member reflects incremental deposition of suspended sediment. Organic rich layers probably represent accumulation of uncalcified, planktonic organisms which inhabited shallow, oxygenated layers of the water column. This background basinal deposition was periodically interrupted by influxes of lime mud transported from adjacent platforms into the basin. During early stages of deposition poorly oxygenated conditions were locally present at the sediment surface, possibly associated with paleotopographic highs, and in these locations the laminated fabric is locally disrupted by burrowing organisms.

Thin, sharply based interbeds of skeletal material and peloids occur in locations where the Ratner Member onlaps the flanks of buildups of the Upper Member. These interbeds are interpreted as turbidites sourced from the buildups. Stratigraphic positioning of turbidites and interfingering of buildup and basinal sediment along buildup flanks in 12-24-1-10 w2M and 7-24-4-5 w2M suggest that most of Ratner Member deposition occurred contemporaneously with buildup growth. The presence of a stratified water column would permit growth of buildups in shallower, oxygenated waters concurrently with depositon under anoxic, hypersaline conditions in the deeper basin.

During final stages of Ratner deposition, which may postdate buildup growth, bottom waters became sufficiently hypersaline for intermittent precipitation of calcium sulfate to occur. Basin-wide precipitation of calcium sulfate was still periodically interrupted by influxes of lime mud. Deposition of the Winnipegosis Formation ended with increasing restriction of the basin and initiation of massive evaporite deposition of the Prairie Formation.

III. DOLOMITIZATION

A. INTRODUCTION

Dolomitization is a complex process and the origin of pervasively dolomitized rocks continues to be debated (e.g. Morrow, 1982a, 1982b; Land, 1985; Machel and Mountjoy, 1986; Hardie, 1987). Part of the complexity arises from the fact that direct observation of the physical and chemical conditions required for dolomitization is hindered by an inability to experimentally precipitate ordered, stoichiometric dolomite from solutions at less than 100° C (Bathurst, 1975). Identifying conditions that favor dolomitization is also complicated by the range in composition and structural ordering in the mineral dolomite which results in a corresponding range in stability (Graf and Goldsmith, 1956).

Dolomite can form in many different environments (Land, 1986). Waters known to be capable of dolomitization include marine-derived brines (Behrens and Land, 1972; Patterson and Kinsman, 1982), continental-derived brines (von der Borsch *et al.*, 1975; Andrews *et al.*, 1987), mixtures of marine- and continental-derived brines (Pierre *et al.*, 1984), normal seawater (Saller, 1984), and mixtures of seawater and meteoric water (Land, 1973). Machel and Mountjoy (1986) concluded that dolomitization may occur in waters of virtually any salinity as long as they are supersaturated with respect to dolomite. In addition, they suggested that hydrologic parameters such as flow rate and permeability ultimately determine whether or not massive dolostones form in any given situation.

The Winnipegosis Formation illustrates the diagenetic complexity of ancient dolostones. In the study area, two phases of replacive dolomite were identified: early, pervasive dolomitization (type 1); and late, partial dolomitization associated with pressure-dissolution (type 2). The two phases formed at different times during the diagenetic history of the rock under very different chemical and hydrological conditions. The record of dolomitization in the Winnipegosis Formation was further complicated by subsequent diagenetic alterations of the replacive dolomite.

B. TYPE 1: EARLY PERVASIVE DOLOMITIZATION

Petrography

In transmitted light, type 1 dolomite is light brown and contains numerous micrometer-sized inclusions, probably from the precursor carbonate (Plate 5B). The dolomite crystals are anhedral to subhedral, with relatively sharp extinctions. Crystal size ranges from very fine to medium but is commonly fine to very fine. The cathodoluminescence signature of type 1 dolomite is a dull, reddish orange (Plate 8A,B); while under ultraviolet fluorescence, type 1 dolomite is a dull brown (Plate 7E).

Allochem preservation is variable in type 1 dolomite but for the most part the allochems are nonmimically replaced with retention of outlines but poor microfabric preservation (Plate 5A). The poor preservation of allochems in the dolomite is partly due to diagenetic alterations which followed dolomitization, including dissolution, dolomite recrystallization, and anhydrite replacement. The variation in allochem preservation between rock types illustrates this influence. Allochem microfabrics are generally much better preserved in dense lithologies such as mudstones, wackestones, and packestones compared to porous grainstones. Since most diagenetic reactions in carbonate rocks require water to transport the reactants and products of the reaction, low permeability rocks are less susceptible to diagenesis. Hence the greater destruction of microfabrics in the porous lithologies in the Winnipegosis Formation.

Distribution

In the study area, most of the rocks in the Winnipegosis Formation are replaced by type 1 dolomite. Based on estimates from logged core in the study area, over 85 % (by volume) of the rocks in the Formation have been replaced by type 1 dolomite. The only intervals of the Winnipegosis Formation that did not undergo type 1 dolomitization were dolomitic limestones along the base of the Formation (Fig. 6). These limestones, however, were partially replaced by type 2 dolomite later in the diagenetic history.

In the study area, limestones occur along the base of the formation. In 12–24–1–10w2M and 7–24–4–5w2M, the uppermost strata (facies R5 of the Ratner Member) underwent type 1 dolomitization while the underlying strata (facies R3 of the Ratner Member and the entire Lower Member) were not dolomitized (Fig. 6). The limestones are interpreted as localized features since in both locations, the limestones were completely dolomitized in core from less than 1 km away. With the sparse core control , however, the lateral extent of the limestones cannot be determined with any degree of certainty. Data from recent wells drilled in the study area suggests that basal limestones may be common in the Winnipegosis Formation (Bill Martindale, personal communication, 1988).

A similar pattern of massive dolomitization, with local preservation of limestones along the base of the Winnipegosis Formation, was reported for other areas in the subsurface of Saskatchewan by Jones (1965), Wardlaw and Reinson (1971), and Fuzesy (1975). The Winnipegosis Formation was extensively dolomitized in the interior of the Saskatchewan Sub-basin but in other locations, dolomitization was less extensive. A regionally preserved limestone is present in Manitoba (Norris and Uyeno, 1971; Norris *et al.*, 1982; McCabe, 1987). The limestone is stratigraphically equivalent to the Lower Member. The Winnipegosis Formation is only partially dolomitized in North Dakota (Ehrets and Kissling, 1987) and northern Alberta (Schmidt *et al.*, 1985).

Timing

Type 1 dolomitization occurred relatively early in the diagenetic history of the rocks. The only diagenetic phases to form before type 1 dolomitization were isopachous cementation in the Upper Member and the formation of nodular anhydrite in the Ratner Member. The isopachous cements were completely replaced by type 1 dolomite while the nodular anhydrite was partially replaced by dolomite.

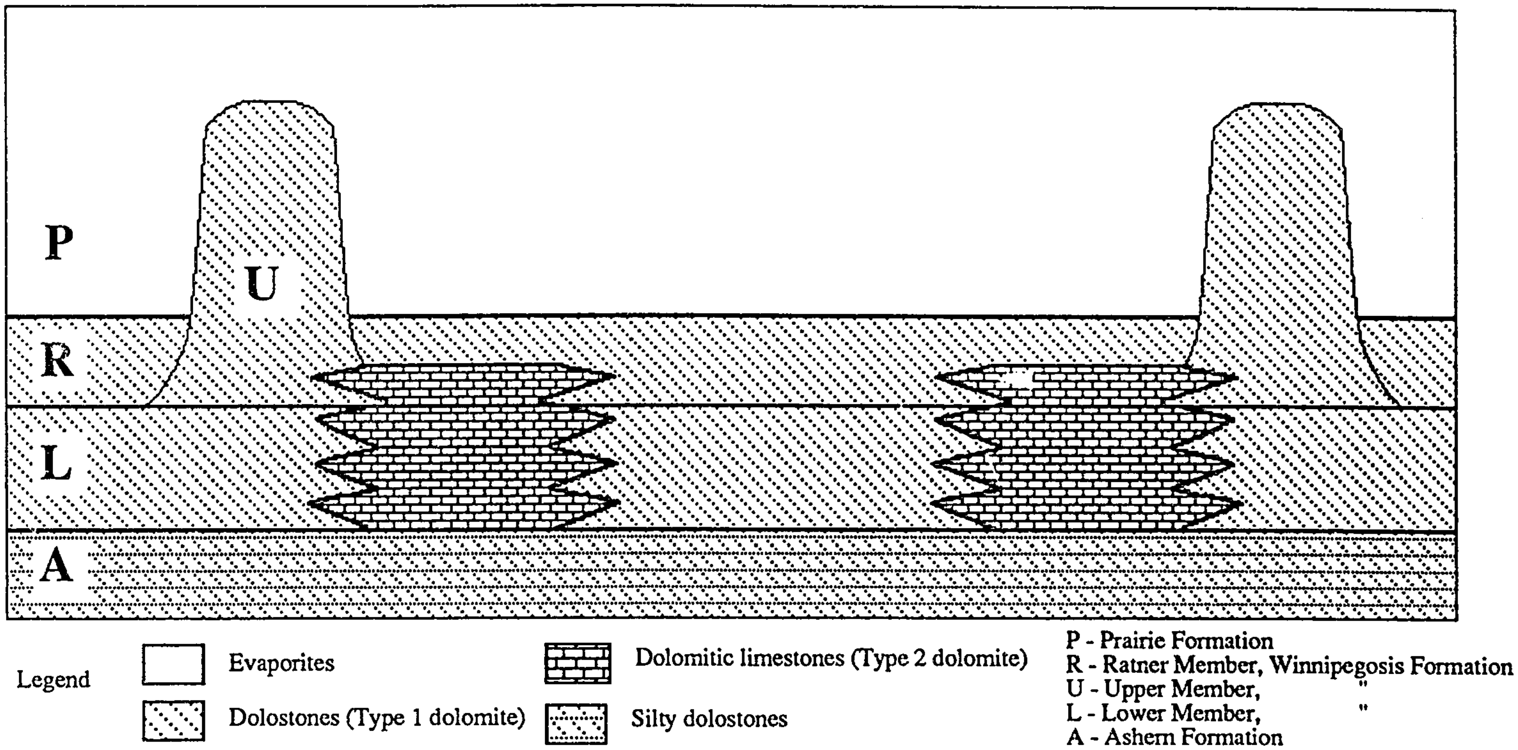


Figure 6. Sketch of lithologic distribution in Upper Elk Point Group in the study area.

Wardlaw and Reinson (1971) also concluded that massive dolomitization occurred early in the diagenetic history of the Winnipegosis Formation. They cited the the lack of deformation of the early evaporite nodules in the Ratner Member as evidence that the carbonates surrounding the nodules were lithified through dolomitization prior to compaction.

Dolomite nonstoichiometry

Nonstoichiometry, expressed as mole % $CaCO_3$, measures the departure of a dolomite composition from the ideal (stoichiometric) composition of 50 mole % $CaCO_3$. The nonstoichiometry of a dolomite reflects the Mg/Ca ratio of the dolomitizing waters (Goldsmith and Graf, 1958) and is thus useful in unravelling water chemistry during dolomitization.

The composition of the dolomite was calculated using powder x-ray diffraction to measure the $d_{(104)}$ spacing of the dolomite relative to a quartz internal standard. This d-spacing varies linearly with the mole % $CaCO_3$ in the dolomite (Goldsmith and Graf, 1958) and the linear equation derived by Lumsden (1979) was used to quantify the relationship. The error of this linear approximation is less than ± 2.0 mole % $CaCO_3$ (Rosen *et al.*, 1988). The error results from the fact that cation disorder and structural effects may also influence the cell size of dolomite, although composition is still the major control (Reeder and Sheppard, 1984). Iron is the only element other than calcium that may be abundant enough to influence the unit cell size in dolomites (Runnels, 1974) and Goldsmith and Graf (1958) determined that greater than 3 mole % FeO is required to produce a measurable change in the size of the unit cell of dolomite. Iron substituition has not altered the unit cell size in type 1 dolomites since they do not stain with potassium ferricyanide and thus contain less than 1 mole % FeO (Lindholm and Finkleman, 1972).

Type 1 dolomite has a slightly magnesium-rich to stoichiometric composition, ranging from 47 to 50.5 mole % $CaCO_3$ (Fig. 7). The composition of the dolomite is similar from

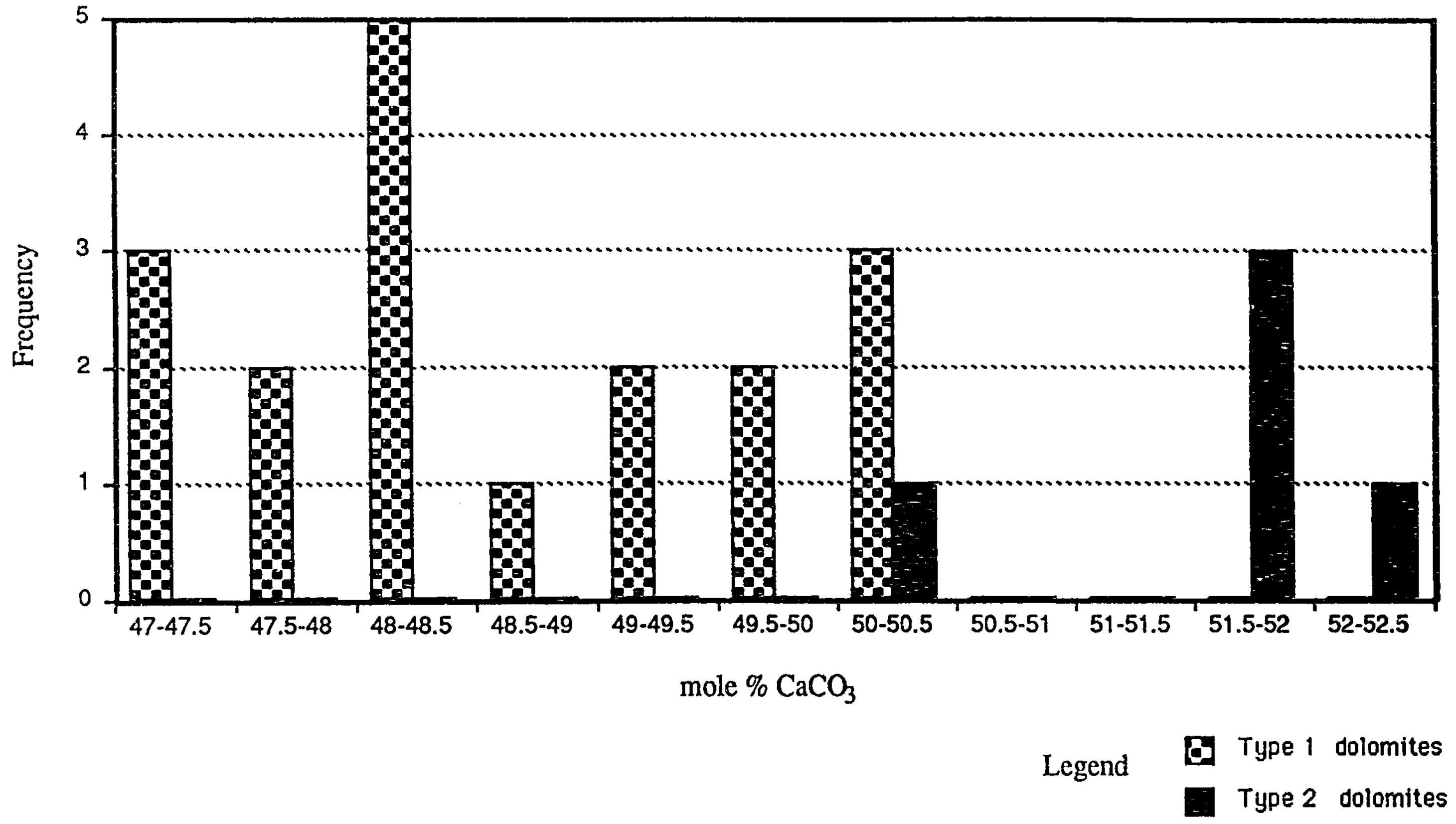


Figure 7. Histogram showing calculated values of dolomite nonstoichiometry

all three members of the Winnipegosis Formation. There are two probable causes for the observed range in stoichiometry. First of all, powder x-ray diffraction is a bulk analysis. Although type 1 dolomite comprised over 90% of the samples, as estimated from corresponding thin sections, minor amounts of other dolomite phases do occur. In addition, any interpretation of the chemistry of ancient dolomite must consider the possibility of diagenetic alteration of the original chemistry. There is abundant textural evidence of recrystallization in type 1 dolomite and the second cause for the range in stoichiometry may be partial alteration of the original dolomite chemistry.

A mineral recrystallizes to achieve a more stable form in response to a change in environment. Recrystallization of dolomite is common, particularily in rocks which have undergone burial (Bein and Land, 1983; Sperber *et al.*, 1984; Banner *et al.*, 1988) and the dolomites in the Winnipegosis Formation are currently buried to depths of 2300-2700 m in the study area. A stoichiometric composition is the ideal and hence, most stable form of dolomite. Thus, the compositional range in type 1 dolomite could reflect partial alteration of an unstable Mg-rich dolomite towards a more stable stoichiometric composition in response to burial.

A magnesium-rich to stoichiometric dolomite composition is characteristic of formation from dolomitizing waters with a high Mg/Ca ratio (Goldsmith and Graf, 1958; Lumsden and Chimahusky, 1980; Sperber *et al.*, 1984; Gunatilaka *et al.*, 1987; Chafetz *et al.*, 1988). Although the chemistry of type 1 dolomite may have been altered, it has retained a slightly magnesium rich composition which suggests formation from waters with a high Mg/Ca ratio.

Water chemistry

The Ratner Member demonstrates an association between the formation of nodular anhydrite and type 1 dolomitization. In intervals where nodular anhydrite is present, the Ratner Member is dolomitized. Conversely, nodular anhydrite is very rare where the Ratner Member remains a limestone. Both type 1 dolomitization and the formation of nodular anhydrite occurred relatively early in the diagenetic history of the rocks, with dolomitization following the formation of nodular anhydrite. This association with evaporite diagenesis, suggests that type 1 dolomitization also involved hypersaline waters.

The nonstoichiometry of type 1 dolomite supports a hypersaline water chemistry. The Mg-rich to stoichiometric composition of type 1 dolomite is characteristic of precipitation from dolomitizing waters with a high Mg/Ca ratio. Marine waters, whether modified or unmodified, are probably the only abundant natural solutions which are rich in magnesium (Land, 1985; Given and Wilkinson, 1987). An elevated Mg/Ca ratio is produced in marine-derived brines by the precipitation of calcium sulfate. Thus, the nonstoichiometry of type 1 dolomite supports formation from a hypersaline brine that had undergone precipitation of calcium sulfate.

Hypersaline waters have a strong chemical drive for dolomitization. Dolomitization is thermodynamically and kinetically favored by higher Mg^{2+}/ Ca^{2+} ratios and higher $CO_3{}^{2-}$ / Ca^{2+} ratios (Lippman, 1973; Carpenter, 1976; Morrow, 1982a; Machel and Mountjoy, 1986) and these ratios are increased in hypersaline brines by calcium sulfate precipitation. Magnesium ions are partially dehydrated in hypersaline waters, which reduces a potential kinetic barrier to dolomitization (Carpenter, 1976). Although elevated salinities will thermodynamically promote dolomitization (Carpenter, 1976), Folk and Land (1975) argued that dolomitization was kinetically inhibited by hypersaline conditions. In a review of this controversy Morrow (1978) concluded that the increased Mg/Ca ratios in hypersaline fluids would probably compensate for any potential kinetic inhibition resulting from high salinities.

Further determination of the water chemistry during type 1 dolomitization may prove challenging. Isotope and trace element geochemistry are powerful tools for identifying the nature of dolomitizing fluids. Type 1 dolomite, however, is extensively recrystallized which makes the presence of the original trace element and oxygen isotope signatures questionable since these components are commonly altered during recrystallization (Dix and

Mullins, 1980; Land, 1980, 1983, 1985; Bein and Land, 1983; Hardie, 1987; Chafetz *et al.*, 1988; Coniglio *et al.*, 1988). Analysis of rare earth element patterns may prove useful as it would provide information on the water chemistry during dolomitization and rare earths are less susceptible to alteration during recrystallization (Banner *et al.*, 1988; Dorobek and Filby, 1988). Cross-contamination from later dolomite phases is also a concern in geochemical analyses of the dolomite in the Winnipegosis Formation. In addition to type 1 dolomite, two phases of dolomite cement and a second phase of replacive dolomite are present in the rocks.

Hydrogeological Reconstructions

Hydrogeology is an important consideration in the analysis of ancient dolostones since vast amounts of water must flow through carbonates in order to supply the magnesium required for dolomitization. The direction of fluid movement during dolomitization can be reconstructed from the resulting distribution of limestones and dolostones (Barrett, 1986; Purser and Schroeder, 1986; Saller and Moore, 1986). Geographic trends in the trace element composition of dolostone bodies have also been used for this purpose (e.g. Machel, 1986). Since dolomite chemistry is commonly altered by later diagenesis, Hardie (1987) argued that lithologic distributions are probably the most reliable evidence for reconstructing the hydrogeological setting in which ancient dolostones formed. Before lithologic distributions can be interpreted, however, it must be determined if the dolostones are the product of a single hydrological regime or a composite of several different regimes.

A distinctive pattern exists in the distribution of limestones and dolostones in the Winnipegosis Formation in the study area. Where the basal beds of the Ratner Member are limestone, the underlying Lower Member is also limestone (Fig 6). In adjacent locations, however, complete dolomitization of the Ratner Member is associated with complete dolomitization of underlying strata. The pattern is not coincidental since it is repeated throughout the study area. The lithologic distribution demonstrates that the

Ratner Member and underlying strata were dolomitized under the same flow system. Where the dolomitizing waters were unable to flow through the base of the Ratner Member, they were also prevented from dolomitizing underlying strata.

The fact that limestones are preferentially preserved along the base of the Winnipegosis Formation strongly suggests that the dolomitizing waters entered through the top of the Formation. Organic-rich mudstones along the base of the Ratner Member (facies R1 and R2) locally blocked the flow of the dolomitizing waters. Low permeabilities are common in mudstone lithologies (Freeze and Cherry, 1979; North, 1985) due to compaction. The effects of compaction generally increase with depth and thus the mudstones at the base of the Ratner Member were less permeable than those at the top. The flow of the dolomitizing waters was predominantly in a vertical direction, downward through the strata. Lateral flow would have resulted in the replacement of the basal limestones by dolomitizing waters that penetrated the base of the Ratner Member in adjacent completely dolomitized locations. The preservation of the limestones argues against a strong lateral component to flow of the dolomitizing waters.

There is no observable difference in depositional texture between limestones and dolostones in the Lower Member and the presence of the limestones does not appear to reflect the presence of permeability barriers in the Lower Member. Instead, the distribution of the limestones in the study area results from the interaction of a vertically-oriented flow regime and permeability barriers along the base of the Ratner Member. Variations of this general model of dolomitization could account for the presence of other limestones preserved along the base of the Formation. Any low permeability strata, such as well-cemented rocks from the buildup margins, could also inhibit dolomitization of underlying strata through a sheltering-type effect.

Dolomitization of the Winnipegosis and Ashern formations occurred independently. The Ashern Formation is uniformly dolostone, even where overlain by limestone in the Winnipegosis Formation. Thus, dolomitization of the Ashern Formation was not influenced by the vertically-oriented flow regime that was active in overlying strata. These observations are consistent with the model of syndepositional dolomitization in the Ashern Formation advocated by Lobdell (1984).

The presence of a thick succession of evaporites overlying the Winnipegosis Formation constrains the timing of the dolomitization. Thick evaporite deposits, such as the Prairie Formation, are effective barriers to flow (Freeze and Cherry, 1979; Dutton, 1987) and their presence would certainly inhibit regionally developed flow into the underlying carbonates. The flow of waters downward into the Winnipegosis Formation must have occurred before the Prairie Formation effectively formed a barrier to flow.

Determining when the Prairie Formation first became a barrier to vertical flow in the study area is difficult. In theory, the accumulation of any thick evaporite deposit should require the formation of a bottom seal which restricts the outflow of the basin waters. Otherwise evaporative concentration of the waters is inhibited by low residency times (Sonnenfeld, 1984; Jennyon, 1986). This theory is supported by the fieldwork of Logan (1987) who studied the hydrogeology of the Quaternary MacLeod Basin of Australia. Logan (1987) confirmed that significant thicknesses of evaporites did not begin to accumulate in the MacLeod Basin until after the formation of a basin-lining aquiclude. By analogy, a basin-lining seal probably formed in the Saskatchewan Sub-basin during early stages of evaporite deposition of the Prairie Formation. The timing of the vertical flow of dolomitizing waters is thus constrained to occur prior to the formation of an impermeable seal in the overlying evaporites of the Prairie Formation.

The hydrogeological model of dolomitizing waters moving downward through the rocks is based on evidence from the Ratner Member and underlying strata. Where the Ratner Member onlaps the flanks of buildups of the Upper Member, the pattern of dolomitization is consistent with this model. The buildup cores, however, are not onlapped by the Ratner Member and in these locations the Winnipegosis Formation is completely dolomitized and the directions of fluid movement during dolomitzation cannot be reconstructed. The complete dolomitization of the buildup cores and underlying strata probably results from the lack of an impermeable layer to impede flow of the dolomitizing waters moving downward through the Formation. Type 1 dolomitization of the buildup cores has the same relative timing as dolomitization of surrounding strata and there is no observable difference in the nonstoichiometry of dolomites from the buildup core and those from the Ratner Member. Since there is no evidence to support an alternative model of dolomitization in the buildup cores and the vertically orientated flow system which dolomitized surrounding strata was operating on a regional scale, it is likely that this system was also responsible for dolomitization of the buildup cores. The evidence for a direct correlation between massive dolomitization in the buildups cores and massive dolomitization of surrounding strata, however, remains inconclusive and it is possible that the buildup cores were dolomitized under a different hydrological system than surrounding strata.

Dolomitization model

A model for type 1 dolomitization must account for :

1) Massive dolomitization on a regional scale,

2) Vertical flow of dolomitizing waters downward through the rocks,

3) Dolomitizing waters with a high Mg/Ca ratio,

4) An association between dolomitization and evaporite diagenesis,

5) Dolomitization early in the diagenetic history of the rocks, constrained by hydrogeology to occur prior to the formation of an impermeable seal in the overlying evaporites of the Prairie Formation.

The subtidal seepage reflux model of dolomitization meets all of these requirements. Subtidal seepage reflux dolomitization was first popularized by Adams and Rhodes (1960) and there continue to be many examples of massive dolomitization which are interpreted to result from seepage reflux (e.g. Sears and Lucia, 1980; Cercone and Lohmann, 1986; Coniglio *et al.*, 1988; Theriault, 1988). In subtidal seepage reflux, dolomitization occurs as basin waters flow downward into underlying strata. The flow is induced by gravity and requires the basin waters to be of greater density than the underlying displaced porewaters. Theoretically, this type of flow could occur in either hypersaline or normal marine basin waters (Simms, 1984). Seepage reflux, however, is commonly associated with hypersaline conditions and precipitation of evaporites (Morrow, 1982b; Land, 1985; Jenyon, 1986). The progressive increase in salinity that leads to evaporite precipitation enhances the potential for a density difference between basin waters and underlying porewaters. Ionic diffusion between the basin waters and underlying porewaters works to correct this imbalance but it is a slow process relative to evaporative concentration. Once the difference in density between the two waters reaches some critical level, flow will occur. Flow is disrupted by the presence of impermeable beds or the removal of the density differential by mixing with lower salinity waters and/or mineralization. Influxes of normal marine seawater into the basin may also disturb the concentration gradient and impede the process. The influxes of fresh marine seawater renew the supply of magnesium and with continued evaporation, a density difference is re-established and flow resumes.

Hsu and Seigenthaler (1969) questioned the ability of seepage reflux to induce massive dolomitization. They argued that the small increases in the density of basinal waters from evaporative concentration would be too small to induce regional flow. Using scaled laboratory models and numerical analysis, however, Simms (1984) showed that *any* upward increase in fluid density in a porous medium is an unstable state and will cause flow to occur. Simms further concluded that in the absence of permeability barriers, potentially giant flow systems extending over 300 m in depth could be formed through seepage reflux.

From this model, it is inferred that dolomitization of the Winnipegosis Formation probably occurred as a series of flow events rather than a single continuous event. The hypersaline conditions required for dolomitization were first established during deposition of the Ratner Member. The potential for dolomitization would persist until the buildups were completely encased in the evaporites of the Prairie Formation which formed a barrier to vertical flow.

The idea that the Winnipegosis Formation was dolomitized through seepage reflux is not new. In modelling the deposition of Prairie Formation, both Klingspor (1969) and Schmalz (1969) suggested that the underlying Winnipegosis Formation could have been dolomitized by refluxing brines. Seepage reflux was also considered as one of the several possible models for dolomitization of the Winnipegosis Formation by Fuzesy (1975), Precht (1986), Ehrets and Kissling (1987), and Rosenthal (1987). McCabe (1987) considered seepage reflux to be the only possible explanation for pervasive dolomitization of the Winnipegosis Formation. All of these workers recognized the strong potential for dolomitization resulting from the hypersaline system directly overlying the Winnipegosis Formation, but they did not present further evidence in support of the theory.

C. TYPE 2: LATE PARTIAL DOLOMITIZATION

Petrography

In transmitted light, type 2 dolomite is light brown and contains minor amounts of micrometre-sized inclusions, probably from the precursor carbonate. Type 2 dolomite partially replaces limestones where it occurs as isolated rhombs to centimetre-sized patches of dolomite mosaic (Plate 9F). The crystal size distribution is nearly unimodal, consisting of fine to medium sized crystals and the crystals are euhedral to subhedral with relatively uniform extinctions. Type 2 dolomite is a dull red under cathodoluminescence, in sharp contrast to the limestones which are light orange. The cathodoluminescence signature of type 2 dolomite is similar to that of type 1 dolomite.

In general, type 2 dolomite preferentially replaces the mudstone groundmass and allochems remain as calcite (Plate 9F). Where partially dolomitized allochems do occur, neither allochem shape nor microstructure is preserved by the dolomite.

Distribution

Type 2 dolomite has a highly restricted distribution, occuring near or within products of pressure-dissolution (Plate 9E). Type 2 dolomite replaces 20–30% of the limestones in the Winnipegosis Formation (Fig. 7). Based on these occurrences, type 2 dolomite is a minor phase and constitutes less than 5 % of the total rock volume.

In describing the distribution of type 2 dolomite, pressure-dissolution and its products must also be discussed. Pressure-dissolution is induced by overburden or tectonic stress and thus occurs in burial environments. The stress is concentrated at contact points or surfaces between grains, crystals, or larger entities. The increased solubility of the stressed portion of the rock results in dissolution and the accumulation of insoluble residue. The products of pressure-dissolution are remarkably diverse (James and Choquette, 1987) and there are many different classifications of these features (e.g. Stockdale, 1922; Park and Schot, 1968; Logan and Semeniuk, 1976; Wanless, 1979; Buxton and Sibley, 1981; Bathurst, 1987). The classification by Wanless (1979), which distinguishes between sutured seam and non-sutured seam products of pressure-dissolution, is used in this study. A sutured-seam is a jagged interface between two rock faces and is generally coated by insoluble residue. Sutured seams have a sutured appearance in sections normal to the seam and are characterized by the presence of interlocking pillars and sockets (Choquette and James, 1987). Sutured seams are commonly referred to as stylolites. A non-sutured seam consists of a smooth, undulose seam of insoluble residue. Care must be taken to distinguish between non-sutured seams and primary clay laminations (Shinn and Robbin, 1983) and positive identification of a non-sutured seam requires evidence of dissolution. Non-sutured seams are also known as microstylolites (Mossop, 1972; Marshak and Engelder, 1985), horsetails (Roehl, 1967), solution seams (Garrison and Kennedy, 1977; Buxton and Sibley, 1981), and dissolution seams (Bathurst, 1987).

Mosaics of type 2 dolomite occur in non-sutured seams in limestones and the distribution of the dolomite parallels the distribution of the seams. Theoretically, the dolomite could have accumulated as insoluble residue since dolomite is less soluble than calcite under burial conditions (Mattes and Mountjoy, 1980). Type 2 dolomite, however, is rare in positions away from the seams and the abundance of dolomite in the seams is considerably greater than would be expected from passively accumulating residue. In addition, type 2 dolomitization is clearly associated with partial dissolution of the limestone substrate. This relationship is displayed by the corrosion and partial dolomitization of calcite allochems within the non-sutured seams.

Timing

Type 2 dolomitization occurred relatively late in the diagenetic history of the rocks in association with pressure-dissolution. The absolute timing of type 2 dolomitization depends on the depth of burial that caused pressure-dissolution in the rocks.

A multitude of factors influence the depth at which pressure-dissolution initiates in a rock including lithology, non-carbonate content , previous diagenetic history, saturation of water for carbonate, temperature, porosity, permeability, and flow rates of pore water (Wanless, 1979; Bathurst, 1987; Choquette and James, 1987). Dunnington (1967) suggested that pressure-dissolution became effective at burial depths of 600–900 m but this range was disputed by Land and Epstein (1970), who documented sutured grain contacts at burial depths of less than 3 m. In general, dissolution along particle-to-particle contacts may begin within the first few 10's to 100's of metres of burial but advanced stages of dissolution involving non-sutured and sutured seams are believed to form at greater burial depths below the influence of shallow meteoric circulation (Neugebauer, 1973, 1974; Coogan, 1974; Bathurst, 1975; Elliot, 1982; James and Choquette, 1987).

The above analysis places general constraints on the timing of type 2 dolomitization. Although some types of pressure-dissolution in the Winnipegosis Formation may have begun at a relatively shallow depth of burial, the extensive dissolution associated with the formation of non-sutured seams and type 2 dolomite probably occurred at a later stage of relatively deep burial.

Dolomite nonstoichiometry

The nonstoichiometry of type 2 dolomite was calculated by the same method used for type 1 dolomite and the same restrictions on interpretation of the data apply. The average unit cell size of type 2 dolomite corresponds to a stoichiometric to slightly calcium-rich composition, ranging from 50 to 52.5 mole % $CaCO_3$ (Fig. 7). Like type 1 dolomite, type 2 dolomite does not stain with potassium ferricyanide and thus the observed expansion in the dolomite unit cell is not due to iron substituition.

Dolomites with stoichiometric to slightly calcium-rich composition typically form from dolomitizing waters with a relatively low Mg/Ca ratio (Goldsmith and Graf, 1958; Lumsden and Chimahusky, 1980; Mattes and Mountjoy, 1980; Sperber *et al.*, 1984). There is no textural evidence to suggest that the original nonstoichiometry of type 2 dolomite has been altered by recrystallization. If type 2 dolomite formed in a burial environment, then it should be less susceptible to the recrystallization that commonly affects surface-formed dolomites on burial.

There is little overlap in the calculated values of nonstoichiometry for the two dolomite phases (Fig. 7) and the two types of replacive dolomite in the Winnipegosis Formation appear to differ in composition. Type 1 dolomite tends to be slightly magnesium-rich while type 2 dolomite is slightly calcium-rich. Elementary statistical analysis employing an F-test and a t-test shows that the two dolomite populations are statistically different to significance level, alpha, of 0.01. The statistical results, however, must be interpreted with caution since the accuracy of the data is within ± 2 mole % $CaCO_3$. Petrographic evidence, which shows that the two types of dolomite formed under different conditions at different times in the diagenetic history of the rocks, supports the interpretation that the two phases of dolomite are chemically different.

Water Chemistry

Incomplete replacement of a limestone substrate during dolomitization demonstrates that the supply of magnesium was insufficient for the process to go to completion (Sperber *et al.*, 1984). This could result from 1) waters with low magnesium concentrations, 2) restricted flow through either inadequate hydraulic head or the presence of a low permeability substrate, or 3) inadequate time for the process to go to completion. Since type 2 dolomitization occurred in a burial environment, where conditions may remain relatively stable for long periods of time, insufficient time is unlikely to have played a major role in limiting type 2 dolomitization.

The nonstoichiometry of type 2 dolomite is characteristic of formation from waters with a low Mg/Ca ratio. This relative depletion of magnesium is typical of subsurface waters (Hitcheon et al., 1971; Chaudhuri et al., 1987) and suggests that the extent of type 2 dolomitization may have been limited by relatively low magnesium concentrations in the dolomitizing waters.

The substrate is typically the dominant source of magnesium for dolomitization associated with pressure-dissolution (Logan and Semeniuk, 1976; Wanless, 1979; Mattes and Mountjoy, 1980; Land, 1985). The limited amounts of magnesium available for type 2 dolomitization probably reflect low levels of magnesium in the substrate.

"The problem of the supply of magnesium for dolomitization is an important constraint in burial environments " (Morrow, 1982b, p. 97). Kinetic factors, however, compensate for low concentrations of magnesium in subsurface waters, allowing dolomitization to proceed. In particular, increased temperatures at depth reduce the Mg/Ca ratio that is required for dolomitization (Udowski, 1968). Higher temperatures also lessen potential kinetic barriers to dolomitization by increasing the reaction rate and inducing partial dehydration of magnesium ions. In addition, Zenger (1983) pointed out that the great amounts of time that rocks commonly spend in the burial environment combined with the stable conditions in this environment may indirectly promote dolomitization.

Hydrogeological Reconstructions

The distribution of type 2 strongly suggests that dolomitization resulted from fluid movement and ionic transport associated with pressure-dissolution. Pressure-dissolution occurs over a water film (Weyl, 1959) which may or may not be in hydrological connection with the bulk porewater system of the rock (James and Choquette, 1987). During pressure-dissolution, ionic transport occurs through diffusion along the water film and/or the bulk porewater system (Baker *et al.*, 1980; Shinn and Robbin, 1983). The net direction of ionic transport may be either parallel or perpendicular to the imposed stress direction (Wanless, 1979) i.e. either parallel or perpendicular to the pressure-dissolution seams. The direction of ionic transport is strongly influenced the permeability of the substrate (Bathurst, 1975) and may be limited to centimeter-scale distances due to low permeability substrates (Shinn and Robbin, 1983; Clifford *et al.*, 1987).

Dolomitization Model

A model for type 2 dolomitization must account for:

1) Incomplete dolomitization,
2) Restricted flow of dolomitizing waters,
3) Dolomitizing waters with a low Mg/Ca ratio,
4) An association between dolomitization and pressure-dissolution,
5) Dolomitization relatively late in the diagenetic history of the rocks.

Type 2 dolomite formed by a reaction between the substrate and ions released during pressure-dissolution. The dominant direction of ionic transport during dolomitization was parallel to seams. The transport of ions into the rock, away from the seams, was probably inhibited by a low permeability substrate. Dolomitization occurred relatively late in the diagenetic history and the permeability of the substrate had been reduced by mechanical compaction. The nonstoichiometry of the dolomite suggests the dolomitizing waters had a relatively low Mg/Ca ratio as is typical of subsurface waters. Type 2 dolomitization probably occurred as a single, progressive event accompanying pressure-dissolution.

Incomplete dolomitization of the substrate resulted from an insufficient source of magnesium and restricted ionic transport.

The ability of pressure-dissolution processes to dolomitize on a regional scale is controversial. Logan and Semeniuk (1976) and Wanless (1979, 1982) proposed that pressure-dissolution could pervasively dolomitize on a regional scale. Land (1985), however, questioned the ability of pressure-dissolution processes to supply enough magnesium for massive dolomitization. In the study area dolomitization associated with pressure-dissolution has only replaced 20–30% of the limestone substrate and was clearly incapable of massive replacement.

D. DIAGENETIC ALTERATIONS OF THE DOLOMITES

"Dolomite is commonly extensively modified during diagenesis" (Land, 1985, p. 112). Type 1 dolomite has been extensively recrystallized and has also undergone local dissolution. In contrast, there is no evidence of recrystallization or dissolution of type 2 dolomite.

This difference in the susceptibility of the dolomite phases to alteration has two probable causes. Firstly, type 1 dolomite formed early in the diagenetic history of the rocks in a near surface environment while type 2 dolomite formed much later in a burial environment. Type 1 dolomite may simply be less stable and more prone to alteration under burial conditions than type 2 dolomite. Secondly, rocks composed of type 1 dolomite are typically more porous than rocks containing type 2 dolomite. Since most diagenetic reactions in carbonate rocks involve water, porous rocks are more susceptible to further diagenetic alterations.

Recrystallization

In the past, evidence for recrystallization has been well-documented in the study of limestones (e.g. Bathurst, 1958; Beales, 1965; Folk, 1965). Although recrystallization of dolostones was identified by Graf and Goldsmith (1956), it has only been recognized as a common diagenetic process in the last decade (e.g. Land, 1980, 1983, 1985; Fairchild, 1980; Gregg and Sibley, 1984; Sperber *et al.*, 1984; Hardie, 1987; Dorobek and Filby, 1988).

Textural evidence for recrystallization of type 1 dolomite is abundant, particularily in association with pressure-dissolution features. Isolated patches of dolomitized micrite and/or poorly-preserved allochems are commonly surrounded and partially replaced by coarser dolomite crystals (Plate 5C, D). The micrite and allochems appear to float in the surrounding coarser crystals and the coarser dolomite crystals truncates depositional fabrics preserved in the finer ones. The embayment of the coarser dolomite into the finer dolomite indicates that the coarser crystals postdate the finer ones. In addition, this relationship shows that the recrystallization occurred in the dolomites. Since the recrystallization fabric involves two dolomite phases it was not formed in the limestone precursor and then texturally preserved by dolomitization.

Another common fabric in type 1 dolomite which is indicative of recrystallization consists of dolomite mosaics with irregular and patchy variations in crystal size, and partial to complete obliteration of depositional fabrics. Although this type of fabric is clearly related to recrystallization (Bathurst, 1958), it cannot be determined if the recrystallization affected the dolostones, their limestone precursors, or both.

Cathodoluminescence signatures also provide evidence of recrystallization in type 1 dolomite. Variations in the cathodoluminescence signature of a rock or mineral provide a qualitative measure of variations in trace-element chemistry (e.g. Sippel and Glover, 1965; Meyers, 1974; Frank *et al.*, 1982; Fairchild, 1983; Machel, 1985). Recrystallization has a homogenizing effect on trace element distributions resulting in a homogeneous cathodoluminescence signature (Sommer, 1972; Dunham and Olson, 1980; Coniglio, 1986; Lee and Friedman, 1987). The cathodoluminescence signature of type 1 dolomite is homogeneous and for the most part, allochems and groundmass in type 1 dolomite are undifferenciable under cathodoluminescence (Plate 8A). Thus, cathodoluminesce signatures support the textural evidence for extensive recrystallization of type 1 dolomite.

A mineral recrystallizes in order to achieve greater stability. "The basic processes causing recrystallization are analgous to dolomitization in the sense that fluids move through the rocks, importing and exporting material, and causing chemical alteration" (Land, 1980, p. 97). Type 1 dolomite is slightly magnesium-rich to stoichiometric and formed relatively early in the diagenetic history in a near surface environment. Dolomites formed in near surface environments are commonly unstable in burial environments (Land, 1980, 1983, 1985; Hardie, 1987; Dorobek and Filby, 1988). The chemical potential for recrystallization is enhanced by higher temperatures in burial environments, which accelerate the rate of reaction . Increased lithostatic pressure also promotes recrystallization to coarser crystal sizes since this lessens the surface free energy and relieves strain at crystal contacts (Baker *et al.*, 1980; Gratier, 1987).

Recrystallization and pressure-dissolution are commonly associated in type 1 dolomite (Plate 5C). This association occurs since both are induced by lithostatic pressure (Wanless, 1979; Land, 1985).

Dissolution

Dolomite dissolution is analgous to calcite dissolution (Drever, 1982) and thus, the wealth of information on calcite dissolution may be applied to dolomite dissolution. The main difference between the two processes is that under equivalent conditions, the rate of dolomite dissolution is much slower. Bathurst (1975) summarized the dissolution and precipitation of calcium carbonate by the reaction:

$$CO_2 + H_2O + CaCO_3 — Ca^{2+} + 2HCO_3^- .$$

From this reaction it is evident that carbonate dissolution is dependent on P_{CO_2}.

Because meteroic waters are in equilibrium with atmospheric CO_2, they are common agents in the dissolution of carbonate rocks. A large variety of waters, however, are known to dissolve carbonate rocks including mixtures of meteoric and marine waters

(Bogli, 1964; Runnels, 1969; Plummer, 1975; Back *et al.*, 1984), or mixtures of reducing and oxygenated marine waters (Paull and Neumann, 1987; Luther, 1988), or formation waters with anomalously high concentrations of CO_2 supplied by thermal reactions involving hydrocarbons (Moore and Druckman, 1981; Elliot, 1982).

Fossil molds occur locally in type 1 dolomite. They are particularily abundant in the Lower Member and facies U3 of the Upper Member. Fossil molds are rare, however, in equivalent undolomitized strata in the Lower Member and this implies that formation of the molds was related to dolomitization. Preferential dissolution of fossils requires a difference in solubility between fossils and groundmass. Molds in dolomite usually form through selective dissolution of either primary aragonite or calcite constiuents encased in a dolomite groundmass (Flugel, 1982). Dissolution of the fossils may have occurred during or after dolomitization. The molds in the Lower Member were subsequently reduced to occluded by dolomite and anhydrite cements.

Type 1 dolomite is locally dissolved forming vugular and solution-enlarged primary porosity. The formation of vugs and solution-enlarged porosity in type 1 dolomite is most common in the Upper Member. The distribution of vugs is not facies specific. Rather, it reflects the distribution of primary porosity. Undersaturated waters flowed preferentially through porous strata, enhancing the porosity through dissolution. The vugs are smooth-sided and dolomite crystals surrounding the vug are truncated. This truncation shows that dissolution followed dolomitization. If the vugs had formed prior to dolomitization, the dolomite crystal faces would project into them (Sears and Lucia, 1980; Ruzyla and Friedman, 1985). The vugs were subsequently reduced to occluded by dolomite and anhydrite cements. Rare evidence exists for the formation of vugs later in the diagenetic history. This later dissolution affected dolomite and anhydrite cements as well as dolomite groundmass (Plate 5E).

Intercrystalline to microvuglar porosity was locally created throughout all three members of the Winnipegosis Formation in type 1 dolomite (Plate 5F). Although intercrystalline porosity is common in many dolomites, the mechanism through which it forms remains widely debated. One school of thought argues that the development of intercrystalline to microvugular porosity in a dolomite results from partial dolomitization of the substrate followed by dissolution of the remaining limestone (e.g. Murray, 1960; Wardlaw, 1979; Randazzo and Zachos, 1984; Sibley and Gregg, 1987). An alternative theory is that the development of porous dolomite fabrics represents a modification of porosity that was present in the precursor limestone. Davies (1979) attributed the formation of intercrystalline porosity in dolomites to textural homogenization of porosity in the precursor limestones. Barrett and Hardie (1986) proposed that the pore size and distribution of porosity in the precursor limestone control whether or not the replacive dolostone is porous. They argued that a porous dolomite will form from a precursor limestone with an irregular distribution of large pores while a nonporous dolomite will form from a limestone with a regular distribution of small pores. There is no evidence from the Winnipegosis Formation dolomites to support one theory over another and the mechanism through which intercrystalline to microvugular porosity formed in type 1 dolomite remains unknown.

E. SYNOPSIS

1. Two types of dolomitization occurred in the Winnipegosis Formation at different times in the diagenetic history of the rock and under different chemical and hydrological conditions.

2. Type 1 dolomitization occurred on a regional scale and most of the rocks were replaced early in the diagenetic history by type 1 dolomite. The distribution of limestones and dolostones shows that the dolomitizing waters moved downward through the Formation. The dolomite nonstoichiometry, the association with evaporite diagenesis, and stratigraphic constraints on the timing of downward flow suggest that type 1 dolomitization involved hypersaline waters and occurred in a near-surface environment early in the

diagenetic history of the rocks. Type 1 dolomitization is interpreted as subtidal, seepage reflux dolomitization.

3. The petrography, composition, and relative timing of type 1 dolomitization in the cores of the buildups of the Upper Member is identical to that in the surrounding dolostones and there is no evidence to suggest that the buildup cores were dolomitized under different conditions than surrounding strata. The directions of flow during dolomitization of the cores, however, cannot be reconstructed and it is thus possible that a different flow system was operative in these locations.

4. Type 2 dolomitization partially replaced limestones late in the diagenetic history of the rocks. The distribution of the dolomite shows that dolomitization was related to pressure-dissolution. Restricted flow and an insufficient source of magnesium resulted in incomplete replacement of the substrate.

5. The two types of dolomite appear to be chemically different. Type 1 dolomite has retained a slightly magnesium-rich to stoichiometric composition whereas type 2 dolomite is slightly calcium-rich to stoichiometric.

6. Type 1 dolomite was extensively altered through recrystallization and dissolution but there is no evidence of alteration of type 2 dolomite. This difference in the stability between the two types of dolomite has two probable causes. Type 1 dolomite formed in a near surface environment and is thus more unstable than type 2 dolomite under burial conditions. In addition, rocks composed of type 1 dolomite are more porous and hence more susceptible to diagenetic alteration than rocks containing type 2 dolomite.

IV. CARBONATE CEMENTATION

A. INTRODUCTION

Carbonate cementation is a complex and variable process involving precipitation of aragonite, magnesian calcite, calcite, or dolomite. Diagenetic alteration of the original mineralogy is common and serves to complicate the process even more. Carbonate cements precipitate from a wide range of water compositions and diagenetic environments. Harris *et al.* (1985, p. 81) emphasized this, noting that carbonate cement forms in "... the marine environment during deposition of the sediment, in a near-surface position where a freshwater lens penetrates the depositional environment, in a subaerial setting produced by sea level fall, in a mixing zone between waters of varying chemistries, or in the brines of the deeper subsurface."

In the study area, there are three generations of carbonate cement in the Winnipegosis Formation 1) a first generation, dolomitized isopachous cement; 2) a second generation, blocky dolomite cement; and 3) a third generation, limpid dolomite cement. All three generations of cement are dolomite but they formed at different times through different diagenetic processes. The first generation of carbonate cement is not primary dolomite. The cement formed early in the diagenetic history and was then completely replaced by dolomite during pervasive, type 1 dolomitization of the rocks (Fig. 8). The second and third generations of cement, however, are primary dolomite precipitates that formed after pervasive, type 1 dolomitization.

B. FIRST GENERATION DOLOMITIZED, ISOPACHOUS CEMENT

Petrography

The isopachous cement is composed of turbid brown dolomite crystals, 50–200 μm long, which have highly irregular to serrated intercrystalline boundaries. The boundary between the cement and substrate is sharp (Plate 6A). Locally, the cement exhibits a

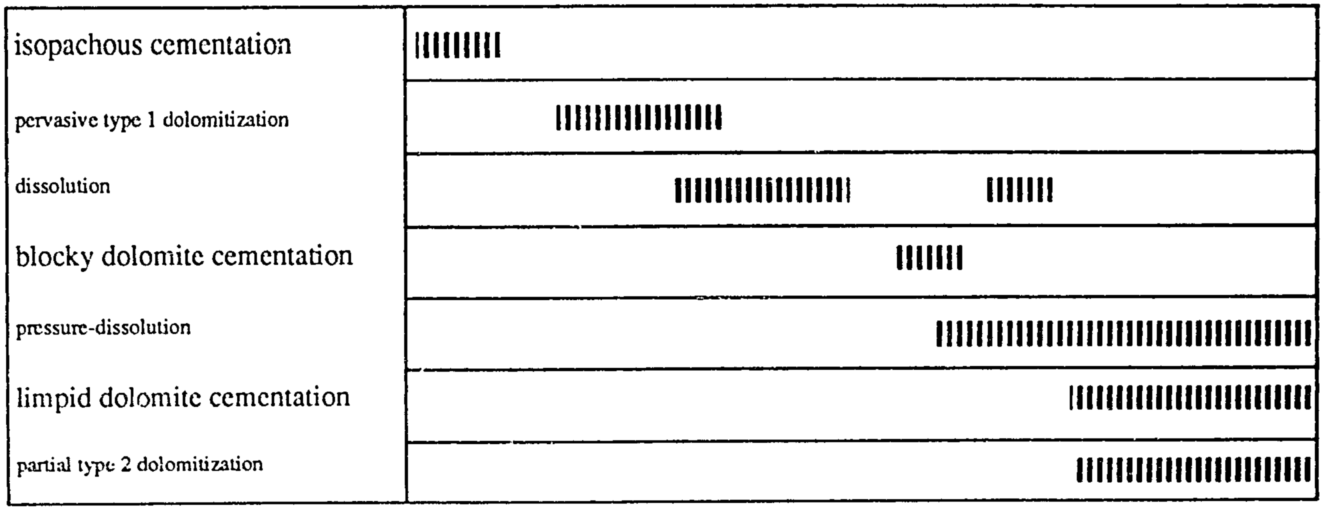


Figure 8. Chart illustrating the relative timing of carbonate cementation with respect to dolomitization and dissolution. Relative time increases to the right.

sweeping extinction along the length of the pore (Plate 6E) indicating that the optic axes of the cement crystals are aligned perpendicular to the pore wall. The isopachous cement lines pores but does not fill them completely and may be followed by second and/or third generation cements.

Pore-lining fabric, the presence of multiple generations of crystalline pore fillings, the sharp boundary with the substrate, and the preferred orientation of the optic axes normal to the substrate demonstrate genesis as a cement (Bathurst, 1975; Chilingar *et al.*, 1979; Kaldi and Gidman, 1982; Jones *et al.*, 1984). Bathurst (1975), however, documented that irregular intercrystalline boundaries, like those in the isopachous cement, are products of neomorphic replacement. The partial obliteration of the cement fabric also suggests that the dolomite comprising the isopachous crusts is neomorphic and has replaced the original cement mineralogy.

The dolomitized, isopachous cement occurs in rocks that were completely replaced, early in the diagenetic history, by type 1 dolomite. The cathodoluminescence and ultraviolet fluorescence signatures of the dolomitized, isopachous cement are identical to those of the dolomitized groundmass (Plate 8A,B). Similarity in cathodoluminescence signatures supports formation from waters of similar chemistry (e.g. Lee and Friedman, 1987) and it is likely that the isopachous cement was dolomitized at the same time as the surrounding rocks.

In the study area, evidence of the original mineralogy of the isopachous cement was destroyed by dolomitization. The cement fabric has been partially obliterated and the chemical signatures of the cement are secondary. Wilson (1985) reported a similar isopachous cement from the Upper Member of the Winnipegosis Formation in central Saskatchewan. She interpreted the original cement mineralogy as magnesian-calcite since in her area, some of the cement was still calcite and there was no evidence to suggest an aragonitic precursor. The isopachous cement in the study area may also have originally precipitated as magnesian-calcite.

Distribution

Isopachous cement has a restricted distribution. It only occurs in primary porosity in facies U2 and U3 in the buildups of the Upper Member. Isopachous cement lines intraskeletal and growth-framework porosity in these facies and is estimated to reduce the porosity by up to 30%.

Timing

Isopachous cement formed early in the diagenetic history. It was the first generation of cement and formed only in primary porosity. Isopachous cementation predates pervasive, type 1 dolomitization of the rocks which occurred early in the diagenetic history in a near surface environment. There are rare hair-line fractures in the isopachous cement which are healed by third generation limpid dolomite cement or anhydrite cement. Fracturing of the isopachous cement would occur in response to mechanical compaction during burial. These relationships suggest isopachous cement formed prior to burial in a surface or near surface environment.

Cementation model

Any model for isopachous cementation must account for:

1) the isopachous, pore-lining fabric,
2) cementation restricted to primary porosity in buildups,
3) the facies-controlled distribution of the cement in the buildups.
4) precipitation early in the diagenetic history, in a surface or near surface environment,

The development of an isopachous fabric indicates precipitation from water-filled pores in a phreatic zone. Halley and Harris (1979) showed that either marine or meteoric phreatic zones may precipitate isopachous cements.

Although the cement fabric is not indicative of a specific diagenetic environment, the distribution and timing of isopachous cementation provide evidence for precipitation from marine waters. The isopachous cement is early diagenetic and formed exclusively in primary porosity in carbonate buildups. Longman (1980), Moldovanyi and Lohmann

(1984), Prezbindowski (1985), and Freeman-Lynde *et al.* (1986) argued that this style of cementation is characteristic of precipitation shortly after deposition from marine waters.

The fact that isopachous cements in the Winnipegosis Formation only occur in the buildups is significant. James and Choquette (1983) suggested that early diagenetic cements restricted to buildups are indicative of marine cementation. Buildups are commonly preferred sites for marine cementation since they are associated with water turbulence and organically stabilized sediment which appear to promote the process (James *et al.*, 1976; Longman, 1980; James and Choquette, 1983; Bathurst, 1986).

The final line of evidence in support of precipitation from marine waters is the facies-controlled distribution of the isopachous cement. A facies-controlled distribution suggests that cementation occurred early enough to be influenced by the environment of deposition. James *et al.* (1976) showed that modern marine cementation is strongly facies specific and thus, the facies- controlled distribution of the isopachous cement supports precipitation from marine waters.

C. SECOND GENERATION BLOCKY DOLOMITE CEMENT

Petrography

Blocky dolomite cement consists of dolomite crystals, 70–150 μm long. The crystals are light brown, equant, and euhedral to subhedral. The cement fabric varies between 1) isolated crystals that discontinuously line pore walls, 2) aggregates of crystals which are patchily distributed around pore walls, and 3) crystal mosaics that loosely fill pores. The boundary between the cement rhombs and the substrate is sharp but the rhombs are randomly oriented and do not exhibit a preferred orientation with respect to the substrate (Plate 6F, 7E). There is no evidence of gravitational alignment in the cement fabric.

Blocky dolomite crystals are rooted on dolomitized groundmass or first generation isopachous cement. The cement is distinctly coarser and more euhedral than the groundmass dolomite (Plate 9A) although both are a similar color under plane light, ultraviolet fluorescence, and cathodoluminescence. Locally, however, the blocky dolomite crystals exhibit a distinctive blotchy luminescence (Plate 9B). In the central portion of the crystal, patches luminesce dark red while the remainder of the crystal luminesces an orange color identical to the cathodoluminescence signature of type 1 dolomite. Blocky dolomite crystals are interpreted as cement based on the presence of pore-lining fabric, multiple generations of crystalline pore fillings, a sharp boundary with the substrate, and planar intercrystalline boundaries.

Distribution

Blocky dolomite cement has a patchy distribution in the rocks that is not related to facies. It is locally developed in secondary and cement-reduced, primary porosity throughout the Upper and Lower members. The abundance of this cement varies from pore to pore. On average, it is estimated that blocky dolomite cement reduces porosity by less than 10%. In rare cases, however, blocky dolomite cement occludes pores.

Timing

Blocky dolomite cement is earliest generation of cement in secondary porosity but follows dolomitized, isopachous cement in primary porosity. Blocky dolomite cement occurs in rocks that were completely replaced by type 1 dolomite, early in the diagenetic history. The cement occurs in vugs in type 1 dolomite, which shows that blocky dolomite cementation postdates both type 1 dolomitization and subsequent dissolution of the dolomite (Fig. 8).

Locally, the outer edges of blocky dolomite crystals are slightly corroded and central portions of the crystals have been dissolved and subsequently filled by limpid dolomite cement (Plate 8C, D). This indicates that blocky dolomite cementation was followed by dolomite dissolution and limpid dolomite cementation. Available evidence poorly constrains the timing of blocky dolomite cementation and the cement may have precipitated at virtually any stage in the diagenetic history, ranging from a near-surface environment to a deep burial environment.

Cementation model

There is insufficient evidence to determine the nature of the diagenetic environment during blocky dolomite cementation. The only interpretation that can be made with certainty is that the cement precipitated from water-filled pores in a phreatic zone. Phreatic rather than vadose conditions are indicated by the fact that the cement lines the entire circumference of the pore walls, although discontinously. In addition blocky dolomite cement has euhedral crystal terminations which are rare in vadose cements. In vadose environments cement crystals stop growing at the air-water interface, producing irregular terminations that reflect the shape of the water film rather than crystal structure (Dunham, 1971; Flugel, 1982). The distribution of the cement around pore walls and the euhedral terminations indicate precipitation in the phreatic zone.

Blocky dolomite cement has been diagenetically altered. It has a blotchy cathodoluminescence signature, which is indicative of partial recrystallization (Dorobek and Filby, 1988). The relict patches which luminesce dark red probably represent the original luminescence signature. The original dark red luminescence signature has partially to completely recrystallized to a lighter orange color. The dark red relicts only occur in the central portion of the blocky dolomite crystals. This fact and the local dissolution of the cores of blocky dolomite crystals strongly suggests that the crystals were originally zoned. In zoned crystals, the crystal core has a different composition and stability than the rim which permits selective dissolution of the crystal cores to occur. The dissolution of the cores must have occurred prior to recrystallization of the blocky dolomite cement since recrystallization has a homogenizing effect and would lessen or eliminate compositional differences within the crystal.

Despite uncertainty as to its origin, blocky dolomite cement provides valuable evidence about general trends in the diagenetic history of the rocks. The presence of blocky dolomite cement indicates that after pervasive type 1 dolomitization and subsequent dolomite dissolution, the pore waters again became saturated with respect to dolomite. In addition, the partial dissolution of the blocky dolomite crystals is direct evidence that a second period of dolomite dissolution followed blocky dolomite cementation. Thus, this portion of the diagenetic history was characterized by fluctuation of the pore waters between supersaturation and undersaturation with respect to dolomite.

D. THIRD GENERATION LIMPID DOLOMITE CEMENT

Petrography

Limpid dolomite crystals are characteristically euhedral, water-clear, and lacking inclusions (Folk and Land, 1975). In the Winnipegosis Formation, third generation limpid dolomite cement is commonly a syntaxial overgrowth rooted on first and second generation cements or dolomitized groundmass (Plate 6C, 7C). The thickness of the overgrowth varies with the size of the pore and overgrowths range from less than 10 μm thick in small pores to over 1 mm thick in larger pores. The boundary between the cement and substrate is sharply defined. Although the substrate may be composed of anhedral dolomite crystals, the limpid overgrowths are euhedral with rhombic terminations (Plate 6D).

More rarely, limpid dolomite cement has a drusy mosaic fabric in which crystal size coarsens from 30 μm to 120 μm towards the centre of the pore (Plate 7A). Enfacial junctions are common in the drusy fabric and intercrystalline boundaries are planar.

Limpid overgrowths and crystals are interpreted as cement based on the presence of multiple generations of crystalline pore fillings, the sharp boundariy with the substrate, the increase in crystal size away from the substrate in the drusy mosaics, the planar, intercrystalline boundaries, and the absence of relict features indicative of neomorphism.

The ultraviolet fluorescence and cathodolouminescence signatures of limpid dolomite cement are distinctive. Limpid dolomite cement is nonfluorescent and appears black when viewed under ultraviolet fluorescence (Plate 7E). Under cathodoluminescence, the limpid dolomite cement is zoned. The zones are defined by variations in both the color and intensity of emission under cathodoluminescence, and range from red to yellow in color and from bright to very dull in intensity. The pattern of zoning varies between core locations and sometimes varies between adjacent pores in a single thin section (Plate 9D). Individual zones are a constant thickness but this thickness varies between zones, ranging from less than 5 µm to 300 µm. The number of zones that occur varies with the size of the pore. In small pores only one to two zones are present (Plate 9C) while in larger pores, up to eight zones may occur. There is no evidence of dissolution between succesive zones in the cement and boundaries between zones are planar indicating growth by regular, parallel increments.

Distribution

Limpid dolomite cement is abundant and has a widespread distribution in the study area. Limpid dolomite cementation is not facies-specific and it occurs in cement-reduced, primary and secondary porosity throughout the Upper and Lower member dolostones.

Limpid dolomite cement is common in rocks with average porosities of less than 5%, as determined by core analyses and visual estimation. The low porosities in these rocks are due to the combined effect of relatively low original porosities and occlusion by abundant limpid dolomite cement. By comparison, limpid dolomite overgrowths are thin and poorly developed in the porous dolostones in 8A–22–2–9w2M which have retained average porosities of 10-20%.

Timing

Limpid dolomite cement formed relatively late in the diagenetic history. It is the last generation of carbonate cement in the rocks and overgrows both dolomitized, isopachous cement and blocky dolomite cement. Limpid dolomite cement precipitated in vugs in type 1 dolomite and thus cementation postdates both pervasive, type 1 dolomitization and subsequent dissolution of the dolostones (Fig. 8). Limpid dolomite cement also infills the leached cores of blocky dolomite crystals, showing that limpid dolomite cement postdates a later phase of dolomite dissolution that partially dissolved second generation blocky dolomite cement. Limpid dolomite cementation both predates and postdates pressure-dissolution. Sutured seams cut molds filled by limpid dolomite cement (Plate 7A) but limpid dolomite cement also occurs in vugs which truncate sutured seams (Plate 7B).

The overlap in timing between limpid dolomite cementation and the formation of sutured seams suggests that the two processes occurred simultaneously. Some types of pressure-dissolution features begin forming in shallow burial environments under the influence of surface-derived waters. Sutured seams, however, represent an advanced stage of pressure-dissolution that usually occurs in deeper burial environments, well below the influence of surface-derived waters (Neugebauer, 1973, 1974; Coogan, 1974; Bathurst, 1975; Elliot, 1982; James and Choquette, 1987). The association between limpid dolomite cementation and pressure-dissolution suggests that cementation occurred in a burial environment.

Cementation model

Any model for limpid dolomite cementation must account for:

1) the precipitation of limpid crystals, commonly as syntaxial overgrowths,
2) the cathodoluminescence zonation in the cement,
3) the abundance and widespread distribution of the cement,
4) the anomalously poor development of limpid dolomite cement in the porous dolostones in 8A–22–2–9w2M,
5) cementation late in the diagenetic history in a burial environment.

The formation of limpid crystals requires slow crystallization (Folk and Land, 1975). Because slow rates of crystallization allow the formation of well-ordered and stoichiometric crystals, limpid dolomite characteristically has a much lower solubility than other types of dolomite (Folk and Siedlecka, 1974).

The limpid dolomite cement in the study area commonly forms syntaxial overgrowths. The prevalence of this growth form probably reflects the fact that the kinetics of crystallization are less demanding for crystal growth to continue on a previously formed nuclei than to initiate a new crystal (Lindholm, 1969). As there were abundant previously formed dolomite crystals to act as nuclei, limpid dolomite cement grew as a syntaxial overgrowth. In addition, the reduction of kinetic barriers to nucleation allows syntaxial overgrowths to precipitate from waters at lower degrees of supersaturation than those required for the growth of crystals which nucleate through random collisions (Sibley and Gregg, 1987).

Limpid dolomite cement is distinctively zoned under cathodoluminescence. Cathodoluminescence zoning in cement reflects spatial variations in concentrations of trace elements in the crystal which are caused by changes in pore water chemistry during precipitation (Meyers, 1978; Grover and Read, 1978, 1983; Pierson, 1981; Frank *et al.*, 1982). Cathodoluminescence zones, however, represent minor fluctuations in water chemistry rather than distinct changes in the diagenetic environment since the mineralogy and fabric of the cement are not altered (Purser and Schroeder, 1986).

Cathodoluminescence zoning can result from variations in the concentrations of Mn^{2+} and Fe^{2+} in the cement (e.g. Meyers, 1978; Pierson, 1981; Frank *et al.*, 1982). In some cases, however, this may be an oversimplification of the chemistry since large numbers of trace elements are known to participate in carbonate luminescence (Sommer, 1972; Marfunin, 1979; Machel, 1985). The particular element or group of elements responsible for the zoning in the limpid dolomite cement in the study area remains unknown and an accurate determination would require either a spectroscopic analyser for the cathodoluminescence microscope or access to an electron microprobe.

The pattern of cathodoluminescence zoning in the limpid dolomite cement varies widely in the study area, without well-defined regional trends. There are two possible interpretations. The variable pattern of zoning in the limpid dolomite cement could indicate partial isolation of the pore waters during cementation. Jones *et al.* (1984), Enos (1986), and Schroeder (1986) showed that variable patterns of cementation result from partial isolation of the porewaters. This would require the rate of precipitation of the cement to exceed the rate of change in the bulk water composition (Wong and Oldershaw, 1981).

Alternatively, the variability in the zoning patterns of the limpid dolomite cement could reflect different paths of flow during cementation. Different flow paths and corresponding differences in the rate of flow, produce localized patterns of chemical evolution during cementation (Searl, 1988). Variations in flow paths arise from differences in the permeability of the rock and/or the presence of permeability conduits such as faults. Majorwicz *et al.* (1988) published evidence of a major fault structure in the Precambrian basement that trends north through the study area and faulting associated with this structure may focus and deflect regional flow in the area. A deflection of flow during cementation would also account for the anomalously poor development of limpid dolomite cement in the highly porous dolostones in 8A–22–2–9w2M.

Limpid dolomite cementation was widespread and required large amounts of magnesium. The timing of limpid dolomite cementation in the study area suggests that precipitation occurred in a burial environment. Although subsurface waters are commonly supersaturated with respect to both calcite and dolomite (Machel and Mountjoy, 1986; Choquette and James, 1987; Hardie, 1987), they contain low concentrations of magnesium ions (Dickey, 1966, 1969; Hitcheon *et al.*, 1971; Chaudhuri *et al.*, 1987; Morton and Land, 1987) and cementation from waters with low ionic concentrations requires tremendous volumes of fluid to pass through the rock (Scholle and Halley, 1985). The original porosity was relatively low in the rocks in which limpid dolomite cement is abundant and it is unlikely that tremendous volumes of water would pass through this type of substrate.

The alternative hypothesis is that limpid dolomite cementation was associated with pressure-dissolution. Vast amounts of magnesium would be released during pressure-dissolution of earlier formed dolomite and a local source of magnesium is more compatible with cementation in a low porosity substrate. In addition, there appears to be a spatial relationship between pressure-dissolution features and limpid dolomite cement in the study area. Pressure-dissolution features are common in the dolostones which contain abundant limpid dolomite cement but conspicuously rare in the dolostones in which the cement is poorly developed. The poor development of limpid dolomite cement is the porous dolostones in 8A–22–2–9w2M thus may be related to the paucity of pressure-dissolutioning in this location.

Dunnington (1954) first proposed that material dissolved by pressure-dissolution might reprecipitate in nearby pores in the rock. Precipitation of ions derived from pressure-dissolution requires nearly static porewaters since if flow rates are too high, the dissolved ions will be flushed out of the pore system (Buxton and Sibley, 1981). Nearly static flow conditions would result in partial isolation of the pore system and the formation of the localized patterns of zoning observed in the limpid dolomite cement. A model of limpid dolomite cementation related to pressure-dissolutioning is favored since it is consistent with all available evidence.

E. SYNOPSIS

1. First generation isopachous cement formed early in the diagenetic history. The isopachous fabric indicates precipitation from water-filled pores in the phreatic zone. The cement probably precipitated from marine waters since cementation was facies specific and occurred exclusively in primary porosity in buildups. First generation isopachous cement was replaced by type 1 dolomite later in the diagenetic history.

2. Second generation blocky dolomite cement was a primary dolomite precipitate. The origin of this cement is uncertain, but its precipitation indicates that following type 1 dolomitization and subsequent dolomite dissolution, the pore waters again became supersaturated with respect to dolomite. The partial dissolution of the blocky dolomite crystals is direct evidence that a second period of dolomite dissolution followed blocky dolomite cementation. Thus, this portion of the diagenetic history was dominated by fluctuation of the pore waters between supersaturation and undersaturation with respect to dolomite. Blocky dolomite cement has been extensively recrystallized. The selective

dissolution of the cores of the crystals and the relict pattern of cathodoluminescence suggests that prior to recrystallization, the blocky dolomite crystals were zoned.

3. Third generation limpid dolomite cement was also a primary dolomite precipitate and formed late in the diagenetic history. The limpid crystals precipitated slowly and minor changes in pore water chemistry during precipitation produced zones. An interpretation of cementation related to pressure-dissolutioning is favored. Cannibalization of earlier formed dolomites would provide abundant magnesium ions and a local ion source accounts for the abundance of cement in rocks which had relatively low original porosities. Cementation related to pressure-dissolutioning occurs under nearly static flow conditions and thus the variable pattern of zoning in limpid dolomite cement may be attributed to partial isolation of the pore waters.

V. EVAPORITE DIAGENESIS

A. INTRODUCTION

As waters become increasingly concentrated, evaporite minerals begin to precipitate. The concentration of the water is commonly attained by evaporation at the air-water interface but can also be achieved by brine freezing or ion filtration (Kendall, 1984). The sequence of mineral precipitation varies with the ionic composition of the water. Evaporite minerals may precipitate from brines derived from marine waters, meteoric waters, subsurface waters, or mixtures of the above (Butler, 1966; Raup, 1970; Williams, 1970).

Evaporite minerals are particularily susceptible to diagenetic alteration due to high solubilites and the fact that they commonly precipitate as metastable hydrates. The identification of early diagenetic features in ancient evaporites is made difficult by successive mineral alterations and recrystallizations.

In the study area, two phases of evaporite diagenesis can be identified in the rocks of the Winnipegosis Formation: 1) anhydrite nodules, and 2) anhydrite cementation and replacement.

B. ANHYDRITE NODULES

Petrography

The anhydrite nodules are small, having diameters of less than 1 cm. Individual nodules are rounded or slightly elongate parallel to bedding and displace laminations in the carbonate groundmass (Plate 4B, 10A). Locally nodules coalesce to form layers of anhydrite which are discontinuous and irregular in thickness (Plate 10B). The nodules are composed of aggregates of lath-shaped anhydrite crystals. The laths are 80-200 µm long and randomly oriented. Maiklem *et al.* (1969) classified this type of anhydrite fabric as felted.

Distribution

The anhydrite nodules have a restricted distribution and only occur in the dolostones of the Ratner Member. Anhydrite nodules occur 10-12 m below the base of sedimentary anhydrite laminations in the Ratner Member but become more abundant towards the top of the Ratner Member, in facies R5.

Anhydrite nodules similar to those in the study area occur at the same stratigraphic level throughout the subsurface of Saskatchewan as documented by Edie (1959), Fuller and Porter (1969a, 1969b), Shearman and Fuller (1969), Wardlaw and Reinson (1971), Reinson and Wardlaw (1972), Bebout and Maiklem (1973), and Davies and Ludlam (1973).

Timing

The anhydrite nodules formed early in the diagenetic history of the rocks. The formation of the anhydrite nodules predates type 1 dolomitization since nodules are partially replaced by dolomite (Plate 10B). The displacement of laminations in the carbonate groundmass around the anhydrite nodules results from soft sediment deformation of the carbonate groundmass around more competent nodules during mechanical compaction (Murray, 1964; Shearman and Fuller, 1969; Wardlaw and Reinson, 1971; Davies and Ludlam, 1973). Thus, the nodules formed in soft sediment prior to burial.

Model of formation

Any model for the formation of nodular anhydrite must account for:

1) the felted anhydrite fabric,
2) formation of the nodules in soft sediment early in the diagenetic history,
3) formation of the nodules in a restricted stratigraphic interval on a regional scale.

In modern environments, nodules may precipitate as either primary gypsum (Kinsman, 1966; West *et al.*, 1979) or primary anhydrite (Kinsman, 1966; Kinsman, 1974; Butler *et al.*, 1982). Hence, the original mineralogy of the nodules in the Ratner

Member may have been either gypsum or anhydrite. Phase relationships indicate that anhydrite should be the stable phase of calcium sulfate in normal marine sea water above 40°C or at lower temperatures in more saline waters, while gypsum should be the stable phase in brackish or fresh waters and at lower temperatures in sea water (Deer *et al.*, 1966; Billo *et al.*, 1987). It is evident that primary precipitation of anhydrite requires extreme surface conditions of temperature and brine concentration. These conditions rarely occur in modern surface environments, where anhydrite forms dominantly through dehydration of gypsum (Murray, 1964; Shinn, 1983; Sonnenfeld, 1984). Given the extreme surface conditions required for primary precipitation of anhydrite, the nodules in the Ratner Member probably precipitated as gypsum which later dehydrated to anhydrite. Available evidence, however, is inconclusive. The rocks in the study area are presently buried to depths of greater than 2 km and nodules of either primary gypsum or anhydrite would probably appear much the same after deep burial (Warren and Kendall, 1985).

Gypsum nodules forming in modern hypersaline environments have a felted fabric similar to the nodules in the Ratner Member and thus they may provide an analogue for fabric development in the ancient nodules (Shearman and Fuller, 1969). The felted fabric in modern gypsum nodules results from displacive growth (Kinsman, 1966; Shearman, 1966) in which older crystals break along cleavage as they are pushed aside during new crystal growth. Breakage of the crystals forms lath-shaped cleavage fragments since gypsum has three directions of well-developed orthogonal cleavage. Disaggregation of modern nodules reveals that the outlines of the laths exhibit rectangular re-entrants controlled by the three directions of cleavage (Shearman and Fuller, 1969). Displacive nodule growth requires soft sediment and minimal overburden and thus, modern nodules form within 1 to 2 m of the sediment surface (Kinsman, 1969; West *et al.*, 1979). By analogy, a similar growth mechanism is interpreted for the nodules in the Ratner Member.

The formation of anhydrite nodules is not indicative of a specific diagenetic environment. Hypersaline pore waters and a soft substrate suitable for displacive growth occur in both supratidal sabkha environments and subaequeous environments below a brine body (Murray, 1964; Kinsman, 1966; Dean and Davies, 1975; Sonnenfeld, 1984; Warren and Kendall, 1985).

The origin of the small anhydrite nodules in the Ratner Member has been widely debated. Fuller and Porter (1969a, 1969b), Shearman and Fuller (1969), and Fuzesy (1975) interpreted the nodules as products of supratidal sabkha diagenesis. Their argument was based on 1) a similarity between nodules in modern sabkhas and those in the Ratner Member and 2) interpretation of the organic-rich laminations in the Ratner Member as intertidal algal mats. As was noted earlier, however, the formation of nodular anhydrite is not restricted to supratidal sabkha environments. The similarity in fabrics between modern and ancient nodules only suggests that both grew displacively in soft sediment. Of equal importance, however, is the fact that the continuity of laminations in the Ratner Member argues against intertidal deposition and instead supports deposition in a subtidal environment through widespread settling of sedimentary particles (Wardlaw and Reinson, 1971; Davies and Ludlam, 1973).

The anhydrite nodules are interpreted as subtidal in origin based on the sedimentological evidence for subtidal deposition of the enclosing carbonate sediments. Anhydrite nodules formed in the Ratner Member on a basin-wide scale and this early phase of evaporitic diagenesis documents the passage of hypersaline waters through the basin sediments.

C. ANHYDRITE CEMENTATION AND REPLACMENT

Petrography

Isolated crystals to irregularily-shaped aggregates of anhydrite crystals occur as cement and/or replacement of the carbonate groundmass (Plate 10E). The anhydrite is coarsely crystalline with crystals ranging from 50 μm to greater than 4 mm in size. The crystals are equant to columnar shaped and both the cement and replacive anhydrite commonly have planar intercrystalline boundaries. The replacive crystals locally contain abundant intracrystalline and intercrystalline inclusions of the carbonate groundmass (Plate 10C).

Anhydrite cementation and replacement are grouped together since 1) in most cases it is impossible to petrographically distinguish between individual phases and 2) rare anhydrite crystals exhibit evidence of both cementation and replacement. These anhydrite crystals typically nucleated in pore space on earlier generations of carbonate cement but in later stages of growth the anhydrite partially replaced the carbonate groundmass surrounding the pore. The boundary between the anhydrite crystals and the carbonate groundmass is diffuse and irregular which is indicative of replacement (Machel, 1986). At the same time, however, the anhydrite is associated with the presence of multiple generations of crystalline pore fillings which demonstrates genesis as a cement (Bathurst, 1975; Chilingar *et al.*, 1979; Kaldi and Gidman, 1982; Jones *et al.*, 1984).

Distribution

Anhydrite cementation and replacement commonly occurs along the base of the Lower Member and has a patchy distribution throughout the remainder of the Winnipegosis Formation. The anhydrite occurs in both limestones and dolostones and is not facies specific. Although anhydrite cement is widespread, it is not abundant and generally reduces porosity by less than 10%.

Timing

Anhydrite cemention and replacement occurred at several times during the diagenetic history of the rocks. Earlier replacive anhydrite is commonly poikilotopic, containing numerous inclusions of type 1 dolomite. The earlier poikilotopic phase has been partially replaced by a later, inclusion-free phase. Anhydrite cement to replacement may occur as the first cement phase in vugs in type 1 dolomite (Plate 6B) or it may postdate blocky and limpid dolomite cements (Plate 10F). Anhydrite crystals are cut by sutured seams (Plate 10C) but in other cases clearly postdate advanced stages of pressure-dissolution since 1) sutured seams can be traced into patches of replacive anhydrite where the insoluble residue occurs as inclusions in and between the anhydrite crystals (Plate 10D) and 2) anhydrite occurs in tension gashes formed along non-sutured seams. It can be inferred that anhydrite cementation and replacement occurred over a range of diagenetic environments, up to and including deep burial, based on the diverse anhydrite occurrences described above.

Model of Formation

Any model for anhydrite cementation-replacement must account for:

1) formation in a range of diagenetic environments,
2) a source of calcium sulfate.

The earlier anhydrite phases may have formed from hypersaline waters associated with deposition of the Ratner Member or Prairie Formation. Other evidence of hypersaline basin waters flowing through the rocks is present which makes this a plausible hypothesis. In this case, seawater-derived brines would have been the ultimate source of the calcium sulfate.

The later anhydrite phases formed in a burial environment. Late diagenetic anhydrite is common in carbonate rocks, particularily in those with associated early diagenetic evaporites (Choquette and James, 1987). The late diagenetic anhydrite represents remobilization of the earlier formed calcium sulfate (e.g. Elliot, 1982; Nahynybida *et al.*, 1982; Machel, 1986). There are numerous potential sources of calcium sulfate which could have been dissolved to re-precipitate as late diagenetic anhydrite including sedimentary anhydrite in the Ratner Member, the Prairie Formation, or underlying strata such as the Interlake Group. Earlier diagenetic anhydrite phases could also provide local sources of calcium sulfate.

D. SYNOPSIS

1). Small anhydrite nodules grew by displacement in soft sediment in the Ratner Member. Sedimentological evidence suggests the nodules grew subaqueously, in sediment flooring a brine body.

2). Anhydrite cementation and replacement occurred throughout the Winnipegosis Formation at several times in the diagenetic history.

3). Earlier phases of anhydrite cementation and replacement may have been related to flow of hypersaline basin waters which were present during deposition of the Ratner Member and Prairie Formation.

4). Later phases of anhydrite cementation and replacement occurred in a burial environment and probably involved remobilization of earlier formed sedimentary and/or diagenetic anhydrite phases.

VI. A DIAGENETIC MODEL FOR THE WINNIPEGOSIS FORMATION

A. INTRODUCTION

Paragenesis refers to the assemblage of minerals that compose a rock (Hurlburt and Klein, 1977). The diagenetic history of a rock is commonly expressed as a paragenetic sequence, which document the evolution of the mineral assemblage. The paragenetic sequence for the Winnipegosis Formation in the study area can be divided into three stages: early, middle, and late (Fig. 9). Diagenesis in the early stage was strongly influenced by depositional environment. Middle stage diagenesis was characterized by dissolution of mineral phases and the large variations in pore water chemistry in this stage suggest a surface or near-surface diagenetic environment. Processes related to burial dominate late stage diagenesis. It is important to note that transitions between stages were gradual and some diagenetic phases, such as pressure-dissolution, overlap stage boundaries.

B. EARLY STAGE DIAGENESIS

Early stage diagenesis includes the formation of isopachous cement, anhydrite nodules, type 1 dolomite, and skeletal molds (Fig. 9). Isopachous cement precipitated from marine waters into primary porosity in facies U2 and U3 in buildups of the Upper Member. Bottom waters in the basin were hypersaline and anhydrite nodules formed in soft sediment in the Ratner Member in this subaqueous, hypersaline environment. The periodic flow of dense basin waters downward through underlying strata was the driving mechanism for type 1 dolomitization through subtidal, seepage reflux. Type 1 dolomitization occurred on a regional scale, prior to the formation of an effective barrier to flow in the overlying Prairie Formation.

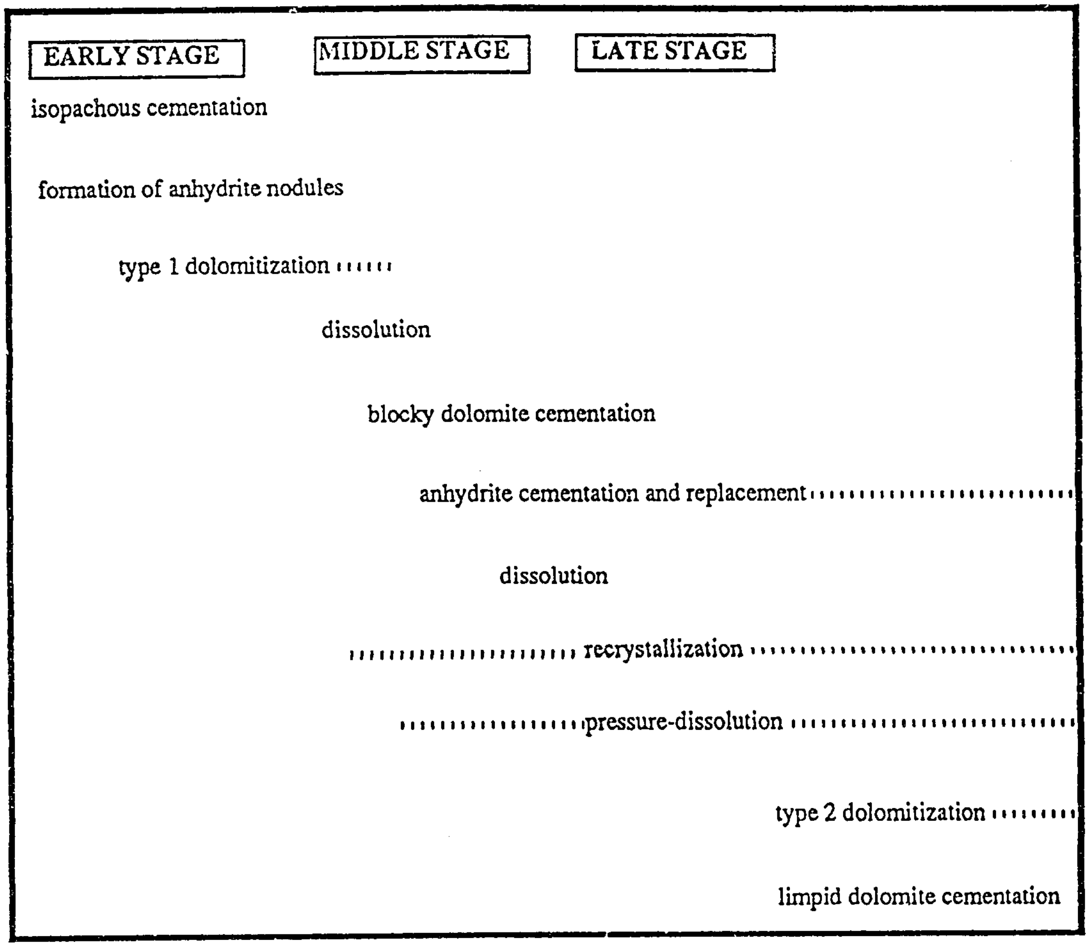


Figure 9. Paragenetic sequence for the Winnipegosis Formation illustrating the relative timing of major diagenetic phases. Relative time increases to the right. Hatched lines show range of formation.

Skeletal molds formed in type 1 dolomite through dissolution of calcareous skeletons during and/or after dolomitization. This dissolution marks the transition into middle stage diagenesis.

C. MIDDLE STAGE DIAGENESIS

Middle stage diagenesis consists of dissolution of type 1 dolomite followed by precipitation of blocky dolomite cement, anhydrite cementation and replacement, and a second phase of local dissolution involving type 1 dolomite, blocky dolomite cement, and anhydrite cement (Fig. 9). The anhydrite cement clearly precipitated from hypersaline waters. Water chemistry also fluctuated between undersaturation and supersaturation with respect to dolomite during this stage and this variability is typical of surface or near-surface diagenetic environments (Zenger, 1983). The dissolution may have been related to influxes of meteroic waters or mixing between marine, hypersaline, and/or meteoric waters.

Evidence for subaerial exposure of the buildups of the Winnipegosis Formation in southeastern Saskatchewan has been documented by MacDonald (1989). This period of exposure, which occurred through progressive restriction of basin waters, most likely correlates to middle stage diagenesis.

D. LATE STAGE DIAGENESIS

The rocks of the Winnipegosis Formation were progressively buried by deposition of overlying strata and late stage diagenesis occurred in this burial environment. Earlier-formed diagenetic phases altered in response to burial and new diagenetic phases formed. The late stage includes pressure-dissolution, recrystallization of type 1 dolomite and blocky dolomite cement, type 2 dolomitization, limpid dolomite cementation, and anhydrite cementation and replacement (Fig. 9).

Pressure-dissolution and recrystallization of type 1 dolomite and blocky dolomite cement may have begun during middle stage diagenesis but reached maximum development under increased temperatures and pressures at depth. Ions released by pressure-dissolution reacted with limestones to form type 2 dolomite. Type 2 dolomitization only partially replaced limestones due to an insufficient source of magnesium and restricted ion transport through a low permeability substrate.

Precipitation of limpid dolomite cement was a slow process, during which minor fluctuations in the porewater chemistry produced zones. The variable pattern of zoning in the cement is interpreted to result from nearly static flow conditions creating partial isolation of the pore system.

Anhydrite cementation and replacement occurred throughout the diagenetic history of the Winnipegosis Formation up to and including deep burial. Anhydritiziation postdates advanced stages of pressure-dissolution and represents one of the youngest diagenetic phases to occur in the rocks. The deep burial anhydrite probably formed through dissolution and re-precipitation of earlier formed calcium sulfate.

VII. SUMMARY AND CONCLUSIONS

1. The Middle Devonian Winnipegosis Formation accumulated in the Elk Point Basin during Givetian time. Initial transgression into the Saskatchewan sub-basin is represented by deposition of the underlying Ashern Formation. Following a minor regressive pulse, the sediments of the Winnipegosis Formation accumulated during renewed transgression and subsequent restriction of the basin waters. Basin restriction culminated with evaporite deposition in the overlying Prairie Formation.

2. The sediments of the Lower Member was deposited under open marine conditions below wave base on a carbonate ramp.

3. Buildups of the Upper Member initiated in response to accelerated subsidence rates in the interior of the Sub-basin. The lower part of the buildups is comprised of codiacean algae mounds which accumulated in relatively deep subtidal environment influenced by normal bottom currents. With growth into wave base, a reefal framework dominated by corals and stromatolites developed on the crests of the buildups. Growth in turbulent waters led to extensive brecciation of the framework. Mixed deposits of allocthonous buildup-derived debris and autochthonous basinal sediment accumulated in buildup to basin transitions.

4. The laminated mudstones of the Ratner Member were deposited in a low energy, subtidal environment. Hypersaline stratification of the basin waters resulted in anoxic bottom conditions. Most of the basinal strata of the Ratner Member were deposited during active buildup growth. Organic-rich laminations in the Ratner Member probably formed through accumulation of pelagic organisms which inhabited the upper layer of oxygenated basin water. Interlaminations of mudstone represent lime mud which was

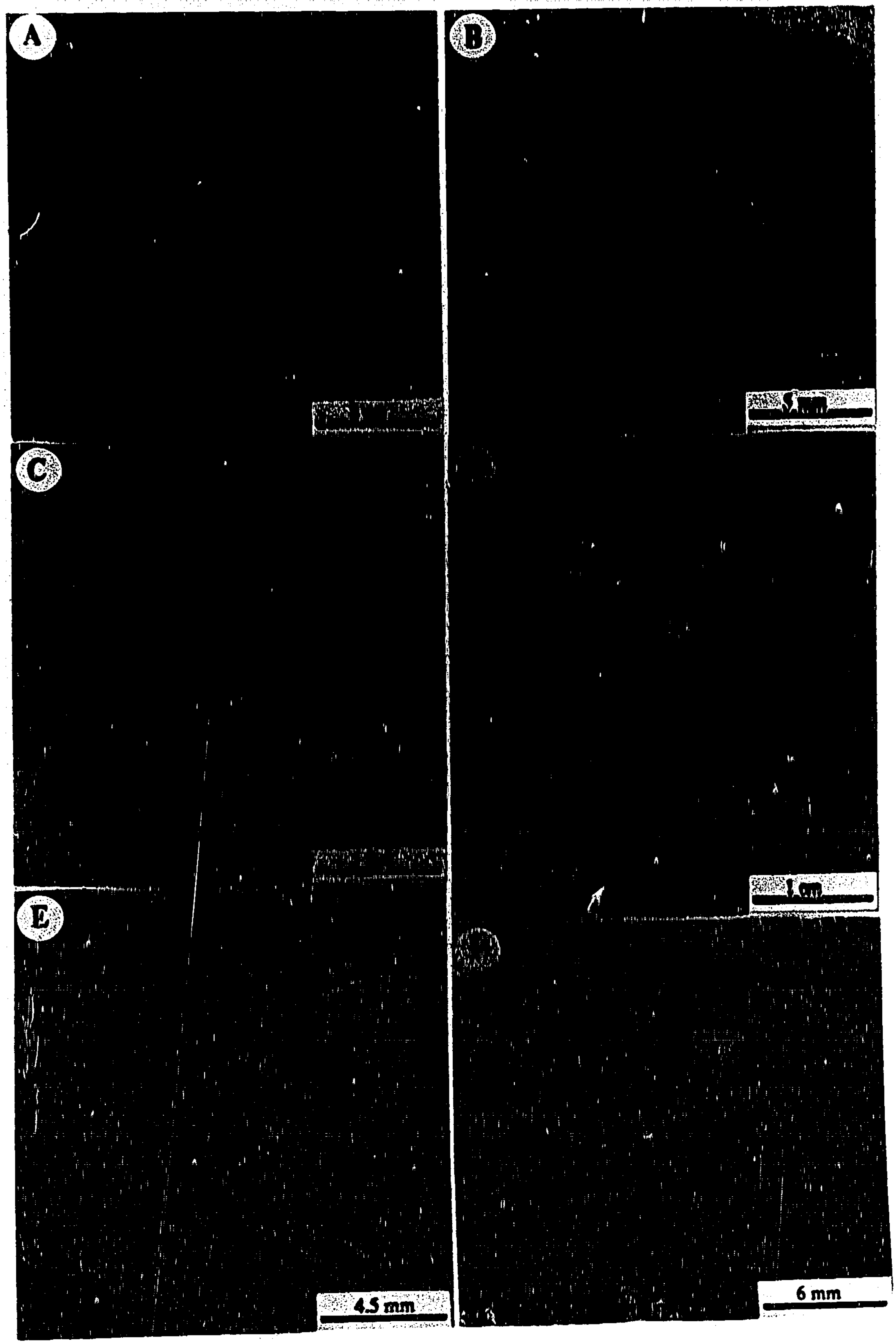
periodically swept into the basin from adjacent platforms. Sharp-based interbeds of peloids and skeletal fragments are interpreted as turbidites derived from adjacent buildups. Increasing restriction during later deposition led to direct precipitation of calcium sulfate from basin waters, forming laterally continuous anhydrite laminations.

5. Early stage diagenesis of the Winnipegosis Formation was influenced by depositional environment. Isopachous cements precipitated in the buildups of the Upper Member from marine waters. Stratified hypersaline basin waters led to the formation of anhydrite nodules in the muds of the Ratner Member and regional type 1 dolomitization through subaqueous, seepage reflux.

6. Middle stage diagenesis bridges the transition from early diagenesis influenced by depositional environment to late diagenesis related to burial. In this stage dissolution of type 1 dolomite was followed succesively by blocky dolomite cementation, anhydrite cementation to replacement, and dissolution of all dolomite and anhydrite phases. Middle stage diagenesis is dominated by dissolution and highly variable water chemistries and probably occurred in a surface or near surface diagenetic environment. Episodic anhydrite cementation and replacement continued into late stage diagenesis.

7. Late stage diagenesis, which occurred during progressive burial of the Winnipegosis Formation, was characterized by modification of earlier phases in response to burial and formation of new diagenetic phases. The transition between middle and late stage diagenesis was gradual. Pressure-dissolution and recrystallization of type 1 dolomite and blocky dolomite cement may have initiated during the middle stage but these processes reached maximum development under the elevated temperatures and pressures at depth. Ions released during pressure-dissolution reacted with limestones to form minor amounts of type 2 replacive

dolomite. Limpid dolomite cement precipitated slowly under nearly stagnant flow conditions in this environment.


A
B
5 mm
C
1 cm
E
4.5 mm
6 mm

VIII. PHOTOGRAPHIC PLATES

Plate 1

(A) Polished core sample of a mildly compacted *Zoophycus* in the silty mudstones of the Ashern Formation (12-24-1-10 w2M, 9105' KB).

(B) Polished core sample of mudstone intraclast beds common near the top of the Ashern Formation (12-30-4-4 w2M, 7606.5' KB).

(C) Polished core sample of crinoid-brachiopod wackestones and mudstones (facies L1) of the Lower Member. Note the predominantly horizontal alignment of the heavily altered burrows (16-33-4-9 w2M, 7992' KB).

(D) Polished core sample illustrating rare fragments of codiacean algae (arrow) present in the crinoid-brachiopod wackestones and packestones (facies L1) of the Lower Member (12-24-1-10 w2M, 9084' KB).

(E) Polished core sample of codiacean algae-peloid packestones (facies U1) of the Upper Member. Note the random orientation of the skeletal fragments and the abundance of lime mud (16-33-4-9 w2M, 7965' KB).

(F) Polished core sample of coral-stromatolite boundstones and framestones (facies U3) of the Upper Member. *Thamnopora* (C) are encrusted by stromatolites (S) (16-33-4-9 w2M, 7937' KB).

Plate 2

(A) Positive print of thin section illustrating codiacean algae-peloid packestones and grainstones (facies U1) of the Upper Member. Both codiacean algae (open arrows) and rare dasycladean algae (solid arrows) occur (12-24-1-10 w2M, 9066.5' KB).

(B) Positive print of thin section of coral-stromatolite boundstones and framestones (facies U3) of the Upper Member. Peloidal groundmass (I) fills growth-framework cavities. Note the isopachous cement (arrow) lining intraskeletal porosity in the corals (16-33-4-9 w2M, 7936.5' KB).

(C) Positive print of thin section illustrating *Renalcis?* boundstones (facies U2) of the Upper Member. Light colored areas represent irregular, growth framework cavities (12-30-4-4 w2M, 7498' KB).

(D) Enlargement of detail in Plate 2C. Photomicrograph of *Renalcis?* boundstones illustrating irregular, micrite-walled chambers of *Renalcis?* (12-30-4-4 w2M, 7498' KB).

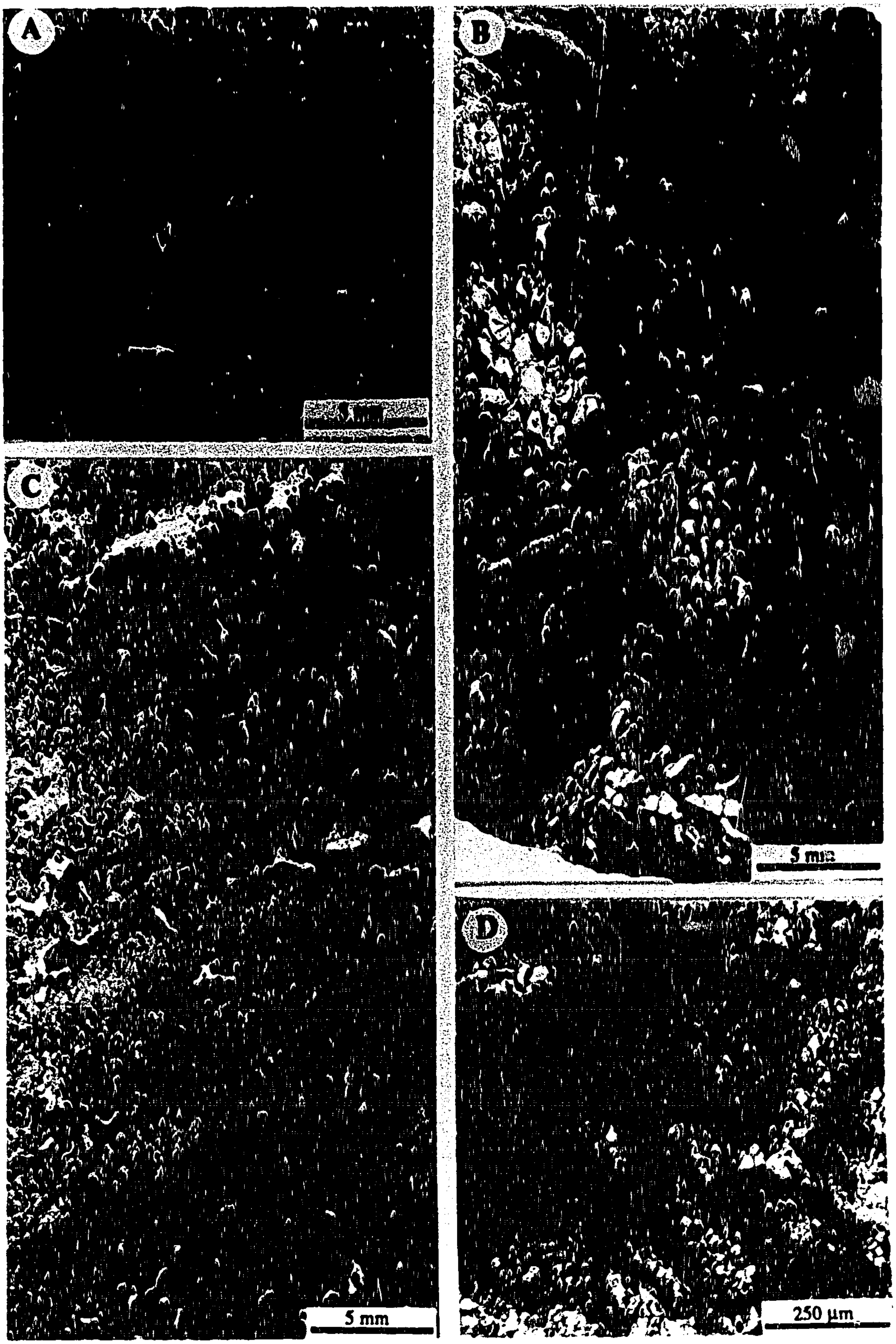

A
B
C
D
5 mm
5 mm
5 mm
250 µm

Plate 3

Positive print of thin section of oncoid floatstones in a peloid wackestone to packestone groundmass (facies U4) of the Upper Member. Tentaculitids (small arrow) are associated with abraded oncoids (large arrow)
(7-24-4-5 w2M, 7666' KB).

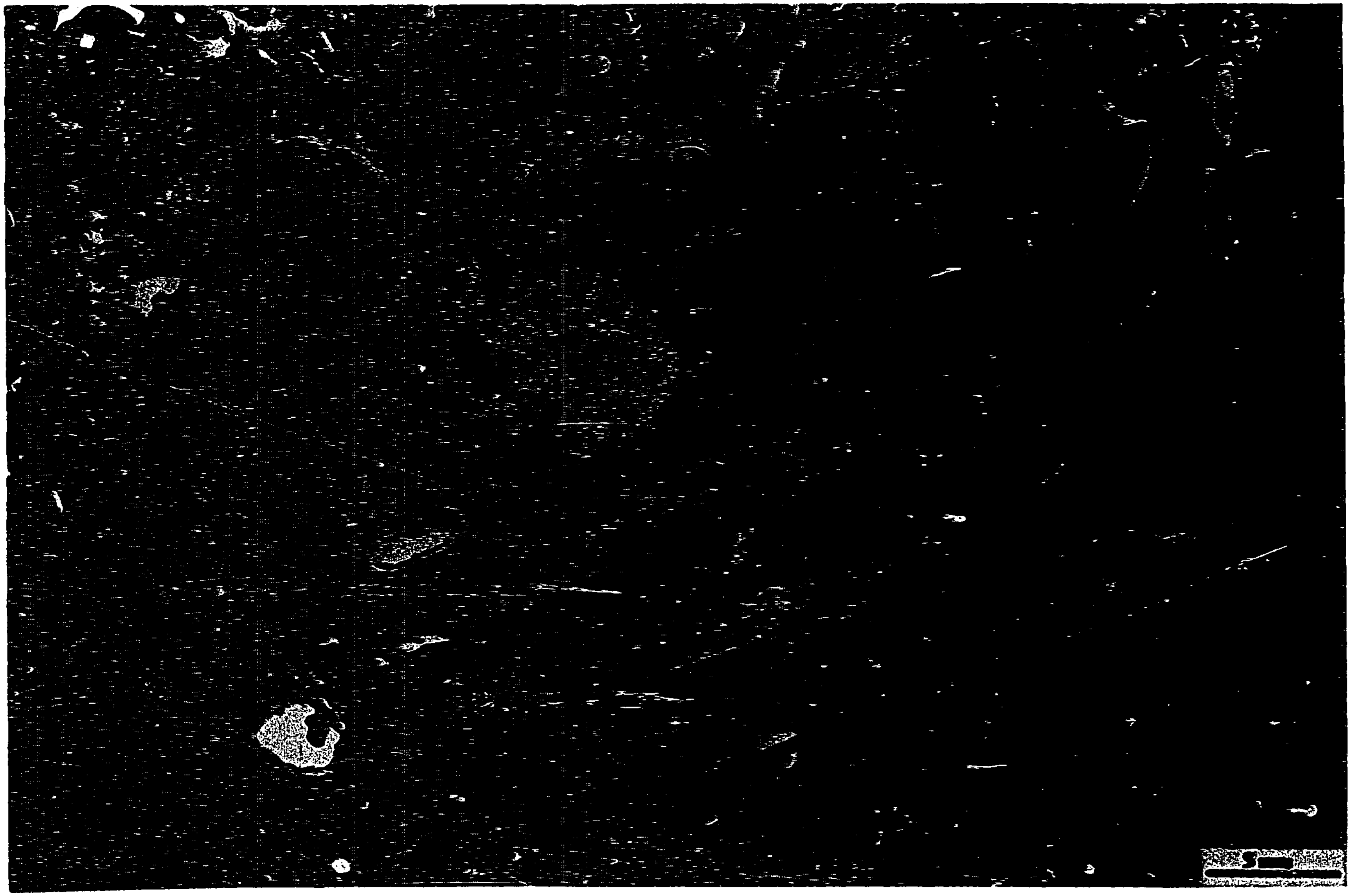

5

Plate 4

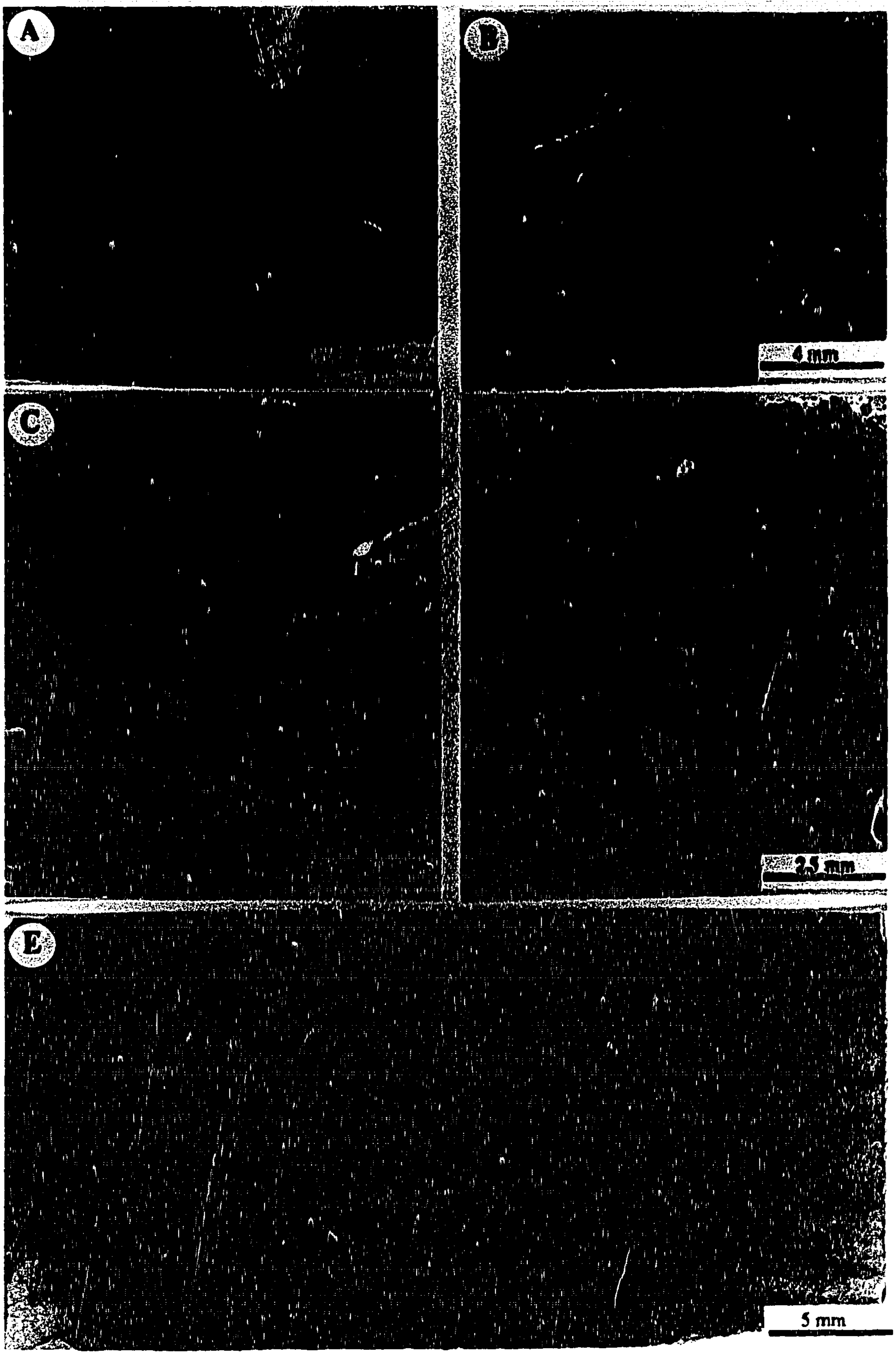
(A) Polished core sample of the undolomitized, laminated mudstones (facies R2) of the Ratner Member. Apparent truncation of laminations results from sutured-seams parallel to bedding (12-24-1-10 w2M, 9053' KB).

(B) Polished core sample of anhydrite nodules (light grey) in dolomitized, laminated mudstones (facies R2) of the Ratner Member. The laminated groundmass is displaced (arrows) around anhydrite nodules (7-24-4-5 w2M, 7640.5' KB).

(C) Polished core sample of peloidal packestones (facies R4) in the Ratner Member. The arrow denotes the sharp contact between peloidal packestones and the underlying laminated mudstones (10-28-1-10 w2M, 8968.5' KB).

(D) Enlargement of detail in Plate 4C. Negative print of thin section illustrating the allochem assemblage of peloids (P) and skeletal fragments (A) in the peloidal packestones (10-28-1-10 w2M, 8968.5' KB).

(E) Negative print of thin section of burrowed mudstones (facies R3) of the Ratner Member. Orientation of the burrows is horizontal (10-28-1-10 w2M, 8985' KB).


A
B
4 mm
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E
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Plate 5

(A) Negative print of thin section illustrating poor preservation in type 1 dolomite. Allochem ghosts (G) are floating in extensively recrystallized groundmass (10-28-1-10 w2M, 9020.5' KB).

(B) Plane polarized light (PPL) photomicrograph illustrating the extensive recrystallization common in type 1 dolomite. A relict skeletal allochem (outlined by arrows) floats in extensively recrystallized groundmass (10-28-1-10 w2M, 9002.5' KB).

(C) PPL photomicrograph of recrystallization in type 1 dolomite associated with pressure-dissolution. Dolomite crystals in the groundmass are embayed into the crinoid fragment (arrows). Note the abundance of sutured-seams (12-30-4-4 w2M, 7560.5' KB).

(D) PPL photomicrograph of recrystallization textures in type 1 dolomite. Floating coarse dolomite crystals (r) have partially replaced finer dolomite (12-30-4-4 w2M, 7564.5' KB).

(E) PPL photomicrograph of a vug truncating both type 1 dolomite groundmass and anhydrite cement (8A-22-2-9 w2M, 2587.5 m KB).

(F) PPL photomicrograph illustrating intercrystalline to microvuggy porosity in type 1 dolomite. Depositional fabrics are obliterated (8A-22-2-9 w2M, 2582.1m KB).

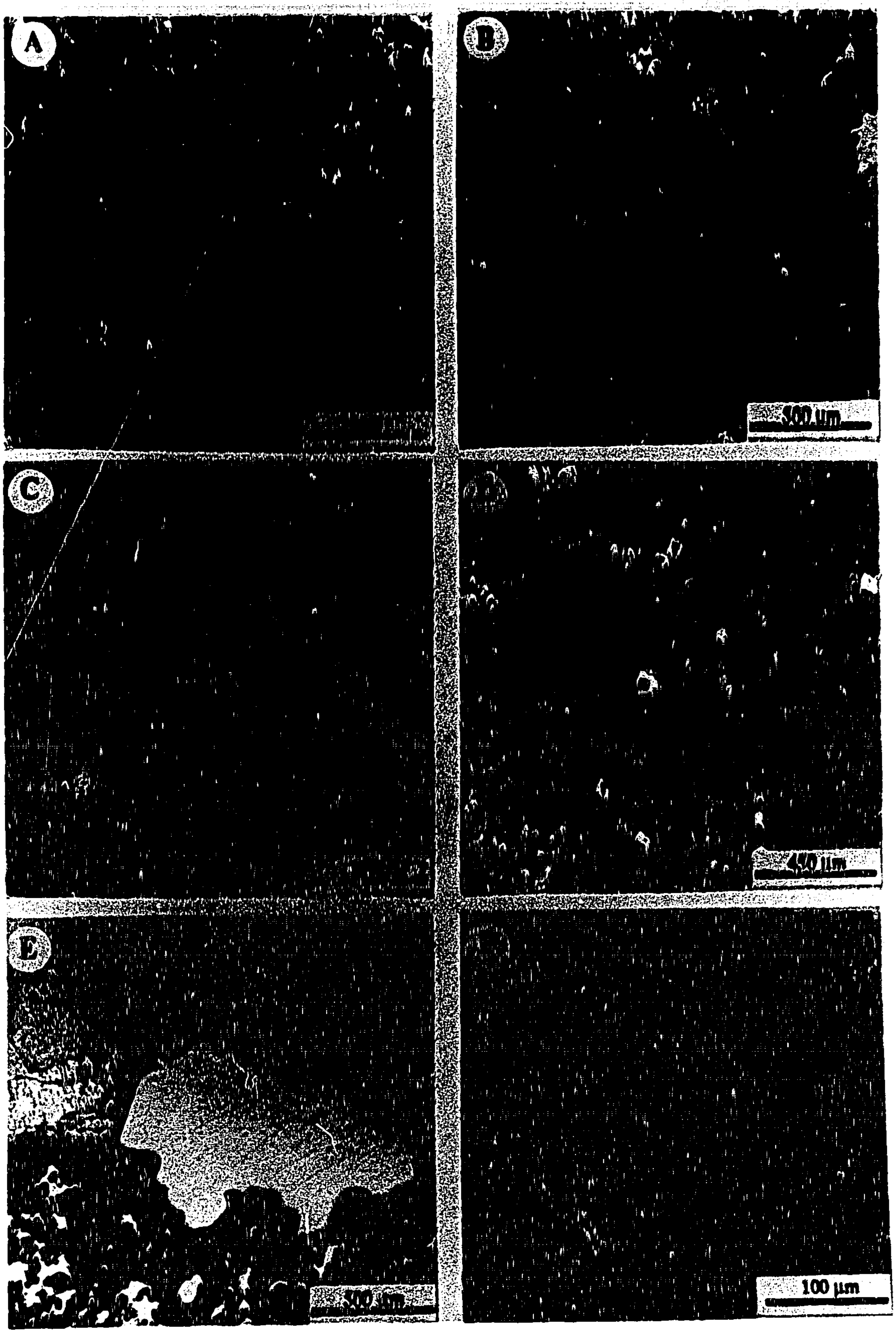

A
B
500 μm
C
450 μm
E
500 μm
100 μm

Plate 6

(A) Plane polarized light (PPL) photomicrograph of isopachous cement lining intraskeletal porosity in a well preserved coral. Note the sharp boundary between the cement and the substrate (arrow) (16-33-4-9 w2M, 7937' KB).

(B) PPL photomicrograph of a poorly preserved coral, also with isopachous cement lining intraskeletal porosity. Partial dissolution of the coral (arrow) was followed by anhydrite (A) cementation. Same sample as Plate 6A (16-33-4-9 w2M, 7937' KB).

(C) PPL photomicrograph of isopachous cement (white arrow) lining growth-framework porosity in *Renalcis?* boundstones. Overgrowths of limpid dolomite cement (black arrows) occur on the isopachous cement (12-30-4-4 w2M, 7498' KB).

(D) Enlargement of detail in Plate 6C. PPL photomicrograph illustrating euhedral terminations (arrow) on limpid dolomite cement overgrowths (12-30-4-4 w2M, 7498' KB).

(E) Cross polarized light photomicrograph of isopachous cement lining a growth-framework cavity? in coral-stromatolite boundstones and framestones. Note the partial preservation of sweeping extinction in the cement and the patchy variations n crystal size in the extensively recrystallized substrate (8A-22-2-9 w2M, 2578.0 m KB).

(F) PPL photomicrograph of blocky dolomite cement lining to partially occluding a vug in type 1 dolomite. Note the random orientations of the cement crystals (8A-22-2-9 w2M, 2576.2 m KB).

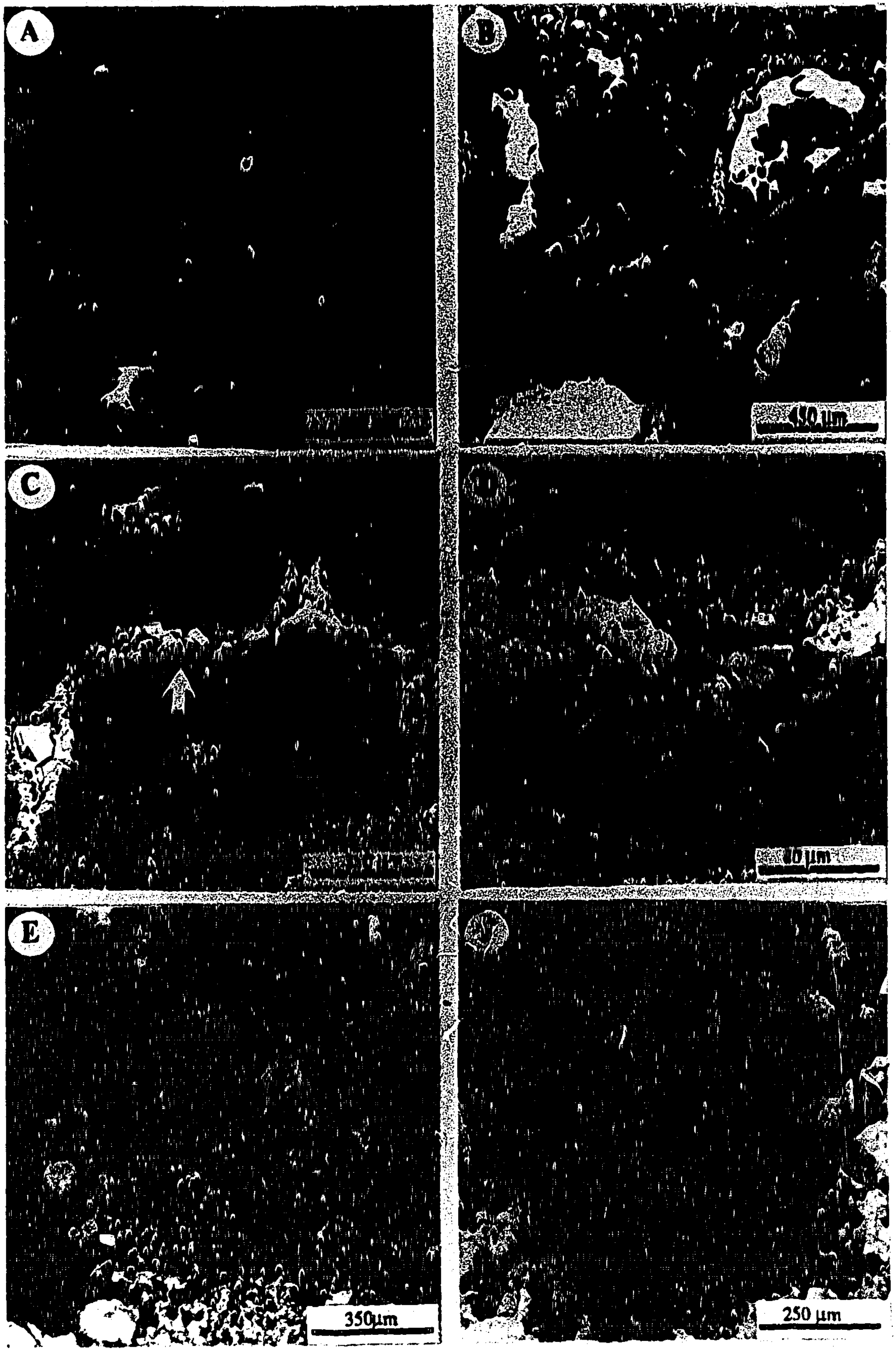

A
B
150 μm
C
E
350μm
250 μm

Plate 7

(A) Plane light photomicrograph (PPL) of limpid dolomite cement occluding fossil molds in type 1 dolomite. Note truncation of cement by sutured-seams (arrow) 12-30-4-4 w2M, 7590.5' KB).

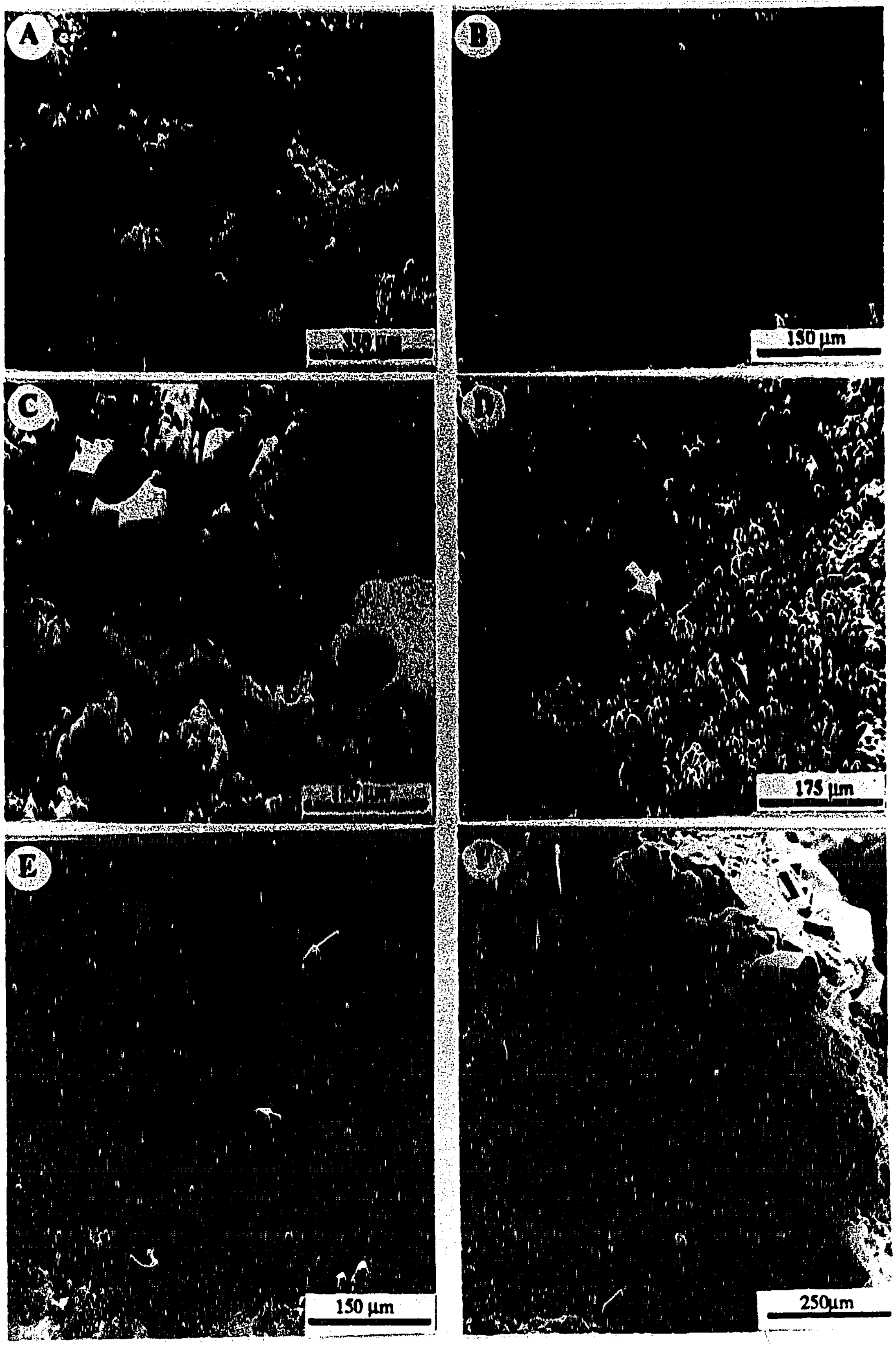
(B) PPL photomicrograph of limpid dolomite cement occluding vug in type 1 dolomite. Note that sutured-seam is truncated (arrow) by cement (12-30-4-4 w2M, 7590.5' KB).

(C) PPL photomicrograph illustrating limpid dolomite cement as an overgrowth (L) on type 1 dolomite. Anhydrite cement (A) also occurs in the vug (12-30-4-4 w2M, 7499.5' KB).

(D) PPL photomicrograph of limpid dolomite cement (arrow) as overgrowths on type 1 dolomite into intercrystalline porosity (10-28-1-10 w2M, 9006.5' KB).

(E) Fluorescent light photomicrograph illustrating nonfluorescent character of limpid dolomite cement (arrows). Cement overgrows dully fluorescent type 1 dolomite (12-30-4-4 w2M, 7560' KB).

(F) Electronmicrograph of blocky dolomite cement lining a vug in type 1 dolomite. Note the random orientation of the cement crystals (8A-22-2-9w2M, 2574.2m KB).


A
B
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C
D
175 µm
E
150 µm
250µm

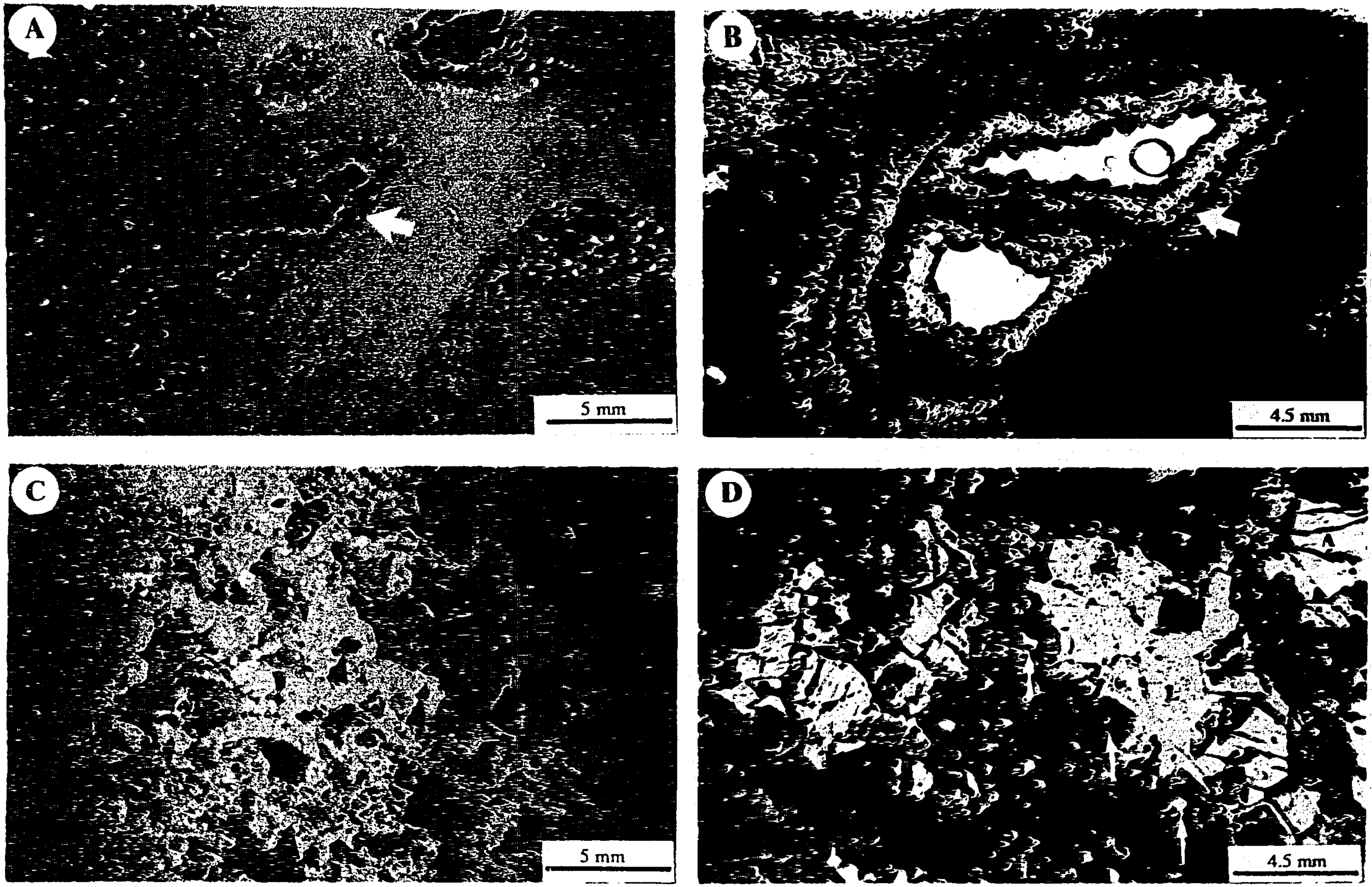
Plate 8

(A) Cathodoluminescence (CL) photomicrograph illustrating similarity in cathodoluminescence signatures between coral replaced by type 1 dolomite and dolomitized, isopachous cement. Arrow denotes the boundary between cement and coral (16-33-4-9 w2M, 7937' KB).

(B) PPL photomicrograph same view as Plate 8A.

(C) CL photomicrograph of cement occluded vug in type 1 dolomite. Cores of blocky dolomite cement crystals have been dissolved and infilled by limpid dolomite cement (arrows). Cement sequence is i) blocky dolomite cement , ii) limpid dolomite cement (L), and iii) anhydrite cement. Note the zones in the limpid dolomite cement (L). Blue dots are artifacts induced during slide preparation (16-33-4-9 w2M, 7965' KB).

(D) PPL photomicrograph same view as Plate 8C.


A
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B
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C
5 mm
D
4.5 mm

Plate 9

(A) Plane polarized light (PPL) photomicrograph of blocky dolomite cement (b) lining a vug in type 1 dolomite. Porespace is light blue (8A-22-2-9 w2M, 2578.8m KB).

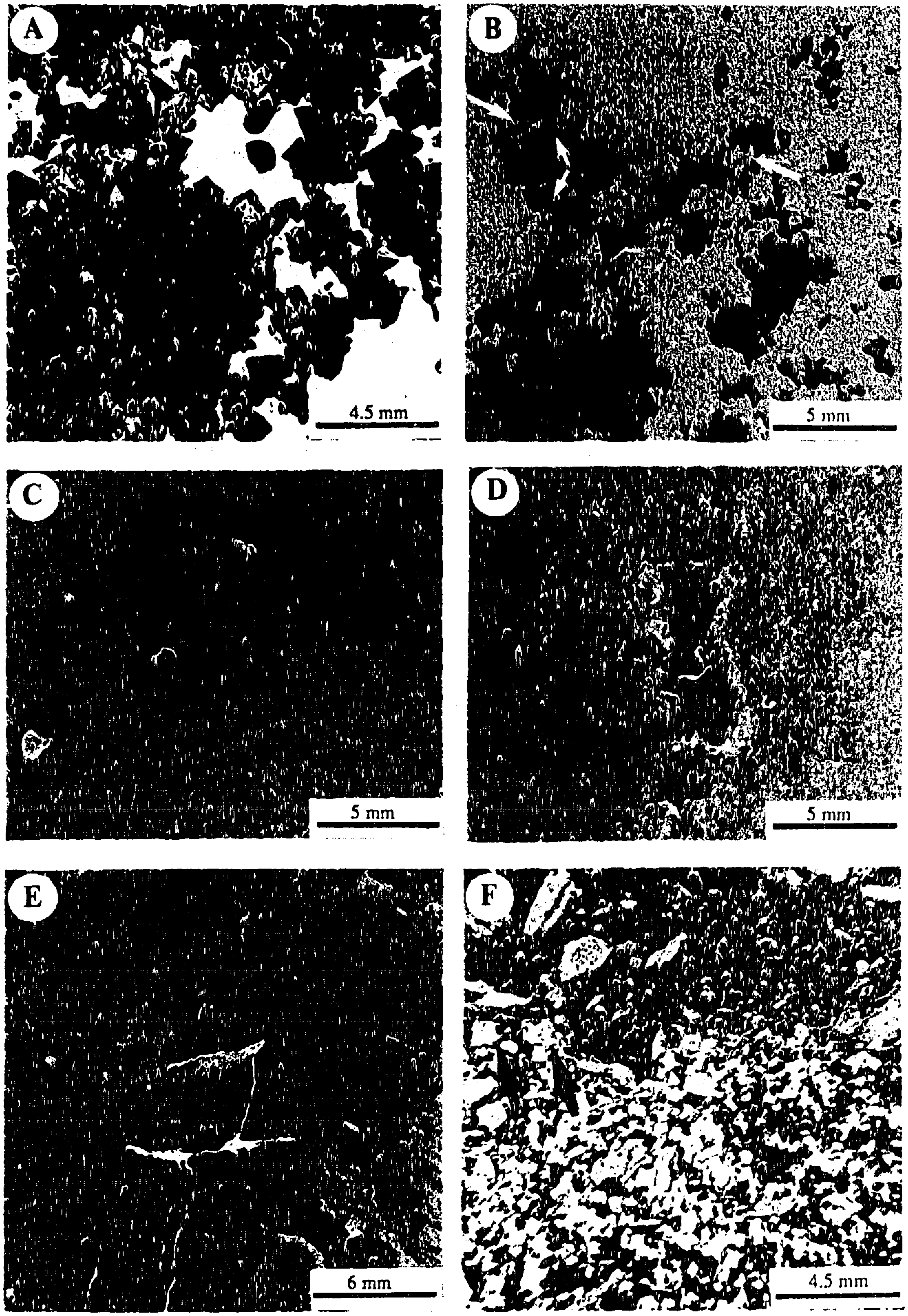
(B) Cathodoluminescence (CL) photomicrograph same view as Plate 9A. Blocky dolomite crystals display distinctive blotchy luminescence in which patches of the crystal core luminesce darker red (long arrows). Overgrowths of limpid dolomite cement (short arrows) occur on the blocky dolomite crystals.

(C) CL photomicrograph of limpid dolomite cement occluding intercrystalline to microvuggy pores in type 1 dolomite. Blue dots are artifacts induced during slide preparation (12-30-4-4 w2M, 7560' KB).

(D) CL photomicrograph of limpid dolomite cement occluding a vug in type 1 dolomite. Note that pattern of zoning differs from that in Plate 8C (10-28-1-10 w2M, 9016.5' KB).

(E) Photograph of thin section stained with Alizarin Red-S illustrating partial replacement of limestones by type 2 dolomite. Distribution of the dolomite parallels the unsutured-seams. Limestone is red. Dolomite is light brown. (12-24-1-10 w2M, 9073.5' KB).

(F) PPL photomicrograph of thin section stained with Alizarin Red-S illustrating type 2 dolomite. Note the association of the dolomite with black, insoluble residue. Note also the truncation of limestone allochems by the dolomite (arrows) and the relict limestone allochems floating in the dolomitized groundmass (12-24-1-10 w2M, 9106.5' KB).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D
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E
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F
4.5 mm

Plate 10

(A) Plane polarized light (PPL) photomicrograph of an anhydrite nodule in type 1 dolomite. Note the displacement of the organic-rich lamination around the nodule (arrow) and the felted anhydrite fabric (7-24-4-5 w2M, 7640.5 ' KB).

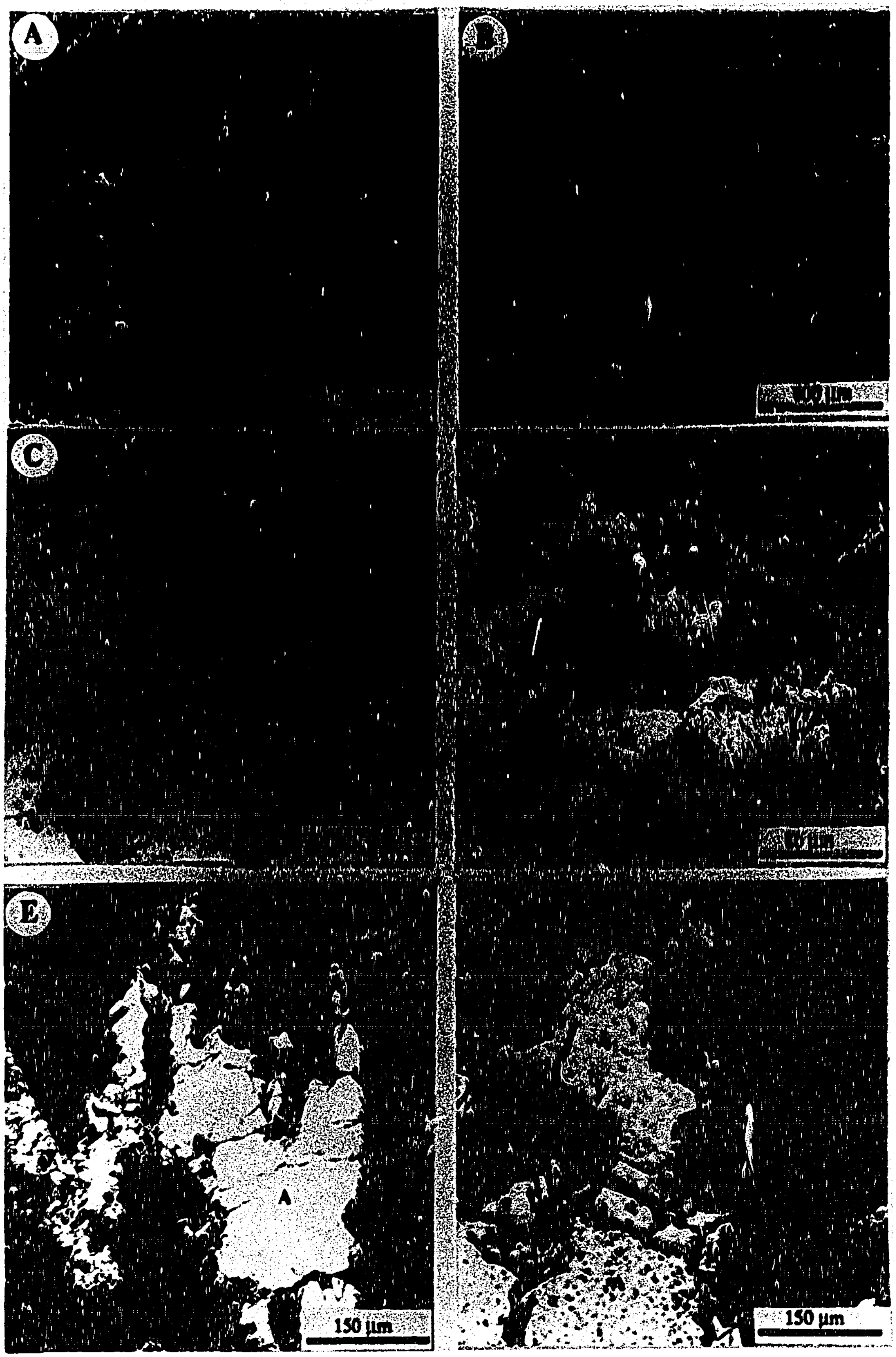
(B) PPL photomicrograph of a discontinuous and irregular anhydrite lamination formed by lateral coalescence of anhydrite nodules. Note the partial replacement of the anhydrite by type 1 dolomite groundmass (7-24-4-5 w2M, 7640.5' KB).

(C) PPL photomicrograph of anhydrite (A) partially replacing type 1 dolomite (D). Note that anhydrite crystal is truncated by sutured seam (12-30-4-4 w2M, 7602' KB).

(D) PPL photomicrograph of inclusions of pressure-dissolution residue in replacive anhydrite (12-30-4-4 w2M, 7566.5' KB).

(E) PPL photomicrograph of anhydrite partially replacing dolomitized skeletal allochem and groundmass (12-30-4-4 w2M, 7554' KB).

(F) PPL photomicrograph of vug occluded by i) limpid dolomite cement (L) and ii) anhydrite cement (A). The anhydrite locally exhibits a replacive character (arrow) (12-30-4-4 w2M, 7501' KB).


A
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A
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