UNIVERSITY OF ALBERTA

STRATIGRAPHY AND SEDIMENTOLOGY OF THE JURASSIC FERNIE FORMATION IN THE SYLVAN LAKE, MEDICINE RIVER AND GILBY FIELDS, SOUTH-CENTRAL ALBERTA

By

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The undersigned certify that they have read, and recommend to the Faculty of Graduate Studies and Research, for acceptance, a thesis entitled STRATIGRAPHY AND SEDIMENTOLOGY OF THE JURASSIC FERNIE FORMATION IN THE MEDICINE RIVER AREA, SOUTH CENTRAL ALBERTA submitted by IAN COLLAR in partial fulfillment of the requirements for the degree of MASTER OF SCIENCE.

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R B Rains ....

Date April 2 1990 ....

This thesis is dedicated to the memory of my mother, Mary Collar, who passed away, after a brief but brave battle with cancer, on January 15, 1990. Upon completion of my B.Sc. she had encouraged me to complete a Master's of Science degree. Although she passed away just as I finished the writing of this manuscript, I know she was proud of my work. I only wish she could have seen the final copy of this thesis and been able to attend my graduation.

ABSTRACT

The study area lies along the eastern subcrop edge of the Jurassic Fernie Formation. In this area it consists of the Nordegg Interval deposits, the Poker Chip Shale, the Rock Creek Member and, in places, the 'Upper Fernie Shale'. Each of these units is bounded above and below by major interregional unconformities, which coincide with global eustatic lowstands. The Nordegg interval contains two sequences, separated by a surface of subaerial exposure and erosion. The lower unit contains the typical Nordegg Member lithologies to the west, but grades into sandstones of the Medicine River Member to the east. Overlying this is a thin deposit of oyster banks, which signify the onset of renewed transgression. Above the Nordegg Interval are the organic-rich shales of the Poker Chip Shale. Thess indicate another eustatic pulse, where the shoreline lay far to the east and anoxic bottom conditions prevailed. The Rock Creek Member consists of a series of unconformity-bounded depositional units, each of which grades from limestones and sandy limestones, in the west, to quartz-rich sandstones in the east. At least three depositional sequences are identified in the area. At the top of the Rock Creek Member a black shale and siltstone unit, of Bathonian to early Callovian age, is often identified but its occurrence is sporadic, as it is often eroded below Ellerslie sediments.

The Fernie Formation was deposited on a stable passive margin with low subsidence rates. Because of these conditions the deposition of the Fernie Formation units was largely controlled by the rate of sediment input and the eustatic level of the Jurassic sea. Deposition of the Nordegg deposits took place at a time of low sediment input and these deposits are therefore characteristically carbonate-rich and clay-free. The Rock Creek

Member, on the other hand, was affected by a major input of terrigenous sediment in the Medicine River field area. Deltaic sedimentation resulted in clay-rich sediment accumulation in the central part of the area. Re-dispersal of muddy deltaic sediment by longshore currents deposited clean shoreface sands in the surrounding area. On the deeper parts of the shelf, away from the delta, the Rock Creek Member consists largely of skeletal sandstones. Seaward movement of clastic sediment, during dseposition of this sediment, was at a minimum.

The stratigraphic architecture of the Fernie Formation, near its subcrop edge, was dominated by subsidence rates which affected this area. Because of the very low regional rates of subsidence, the potential for preservation of shoreline deposits, and ultimately potential reservoir rocks, was relatively low. Local perturbations in this regional trend provided areas in which much thicker shoreline sediments were preserved. Such was the case for the Medicine River-Gilby Jurassic pools. Here, local subsidence, post-dating deposition of the Medicine River Member, allowed for the preservation of the reservoir rocks. To the south, such subsidence did not occur and all evidence of equivalent strata was removed before deposition of the Poker Chip Shale. In the Sylvan Lake Field, local subsidence became a significant factor during and after deposition of the Rock Creek Member. Syndepositional subsidence allowed deposition and preservation of thick deposits of clean shoreface sandstones to develop in this area. Post-lithification subsidence appears to have fractured these tightly cemented sandstones, allowing secondary porosity to develop reservoir capacity. In areas where such processes were absent, sandstones are thin and tightly cemented.

Reservoir development in the Jurassic strata depended upon a complex association of processes. Interpretation of the depositional environments, stratigraphic architecture and subsidence histories is essential in understanding these reservoirs and attempting to predict future exploration and development trends.

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STRATIGRAPHY AND SEDIMENTOLOGY OF THE JURASSIC FERNIE FORMATION IN THE MEDICINE RIVER AREA, SOUTH CENTRAL ALBERTA

INTRODUCTION

The study area is a twelve township block (T. 37-40; R. 3-5w5), which encompasses the Gilby, Medicine River and Sylvan Lake Fields. This area contains a number of prolific hydrocarbon reservoirs at many different stratigraphic levels. Oil and gas are produced from Devonian carbonates along the Leduc-Rimbey reef trend, a play that has seen renewed exploration efforts in recent years. Unconformity-style traps also host significant quantities of hydrocarbons in both upper Mississippian (Elkton and Pekisko reservoirs) and Jurassic (Nordegg, Rock Creek, J1-J3?, detrital and undifferentiated units) sediments. The seal to these unconformity traps is provided by impermeable shales of the lower Cretaceous deposits. The overlying Cretaceous section also hosts a number of stratigraphic traps associated with clastic reservoirs. Hydrocarbons are produced from reservoirs in Lower Mannville Ellerslie and Ostracod deposits, Upper Mannville Glauconite sandstones, as well as upper Cretaceous sediments in the Viking and Cardium units. Such an abundance of prospective zones has resulted in a high density of exploration and development wells, and therefore a large data base of geophysical logs, cores and drill stem tests. In all, 887 wells have been drilled to at least the Glauconite zone.

This work focuses on the Jurassic section, as this hosts some of the main productive reservoirs in this area. The study area lies along the north-south trending subcrop edge of the Jurassic Fernie Formation. This formation contains several formal and informal members. The lower Jurassic consists of the Sinemurian Nordegg Member (and its equivalents) and the Toarcian Poker Chip Shale. The middle Jurassic includes the Bajocian Rock Creek Member and a sporadic shale unit (the 'Upper Jurassic shale' of Marion, 1982), believed to be Bathonian to early Callovian in age. In this area the entire upper Jurassic section has been removed or was never deposited. As this area was also near the paleo-shoreline during much of Jurassic time (Rall, 1980; Marion, 1982 and 1984), several clastic reservoirs developed in this section.

Jurassic sediments produce oil from several large pools in sandstones equivalent to the Nordegg Member (Ter Berg, 1969; Rall, 1980) in the Gilby and Medicine River Fields. Production (predominantly gas) is also achieved from the Rock Creek Member. In the Sylvan Lake gas unit there are a number of clastic reservoirs which have been termed either Jurassic or Basal Mannville, but the distinction is neither clear nor, in many cases, correct.

Because of the numerous unconformities within the Jurassic Fernie Formation, and the complexities of the underlying and overlying unconformities, the stratigraphy of the Jurassic in this area is often very difficult to interpret, and therefore the nature and extent of the hydrocarbon pools in this area are difficult to determine. This is particularly true in the Sylvan Lake Field, where the subcrop of Jurassic sandstones (of undetermined age by industry) is dissected by deep sandstone-filled valleys of probable lower Mannville age. Both types of sandstones are productive and much confusion has arisen as to which reservoirs are Jurassic and which are Cretaceous in age.

The main objective of this work is to unravel the complex stratigraphic relationships of the Jurassic and lower Mannville units and provide a means by which the Jurassic clastic reservoirs can be separated from those of structurally adjacent, lowermost Cretaceous, incised valleys. Detailed geophysical log correlation, supported by lithologic core work and thin section petrographic analysis, is the main tool used to achieve this goal. The second objective is to examine the distribution of hydrocarbon accumulation and relate it to depositional and post-depositional processes.

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STRATIGRAPHY OF THE NORDEGG MEMBER: A NEW PERSPECTIVE ON THE ORIGIN OF THE LOWER JURASSIC MEDICINE RIVER AND GILBY OIL FIELDS, CENTRAL ALBERTA

INTRODUCTION

The Jurassic System of western Alberta consists of the Fernie Formation at the base and the lower part of the overlying Kootenay Formation (and its correlative Nikanassin Formation) at the top. Only the uppermost Jurassic stages are represented by the Kootenay/Nikanassin Formations and these units do not extend very far eastwards into central Alberta. The Fernie Formation, however, extends beneath the plains of central Alberta and locally contains several zones of quality reservoir sediments.

The Lower Jurassic consists of the Sinemurian age, Nordegg Member and its equivalents, overlain by the Toarcian age, Poker Chip Shale. The Nordegg Member has teased hydrocarbon explorationists since the mid 1950's, as it yielded very lucrative fields during early exploration efforts. Two areas of significant hydrocarbon production from Nordegg-equivalent strata are the Paddle River-McLeod area and the Medicine River-Gilby Fields (Springer et al., 1964), both of which are located along the eastern subcrop edge of the Jurassic System as unconformity traps. Attempts to follow up these discoveries, however, were disappointing and it appears that the reservoir facies is absent in most areas along the erosional subcrop of the Lower Jurassic strata.

This paper attempts to: 1) define the geologic circumstances which led to the development of the Medicine River-Gilby producing area, and 2) develop a new model for exploration of similar types of fields. Two main studies have dealt with this area (Deere and Bayliss, 1969; Rall, 1980) and have attributed production to the presence of the reservoir sandstone unit that forms the subcrop beneath the pre-Cretaceous unconformity. In neither case, however, was any explanation forwarded as to why only this area was productive. Using a detailed stratigraphic approach, which is intermediate between the very regional lithologic and petrographic study of Deere and Bayliss (1969) and the very detailed petrographic reservoir study of Rall (1980), this paper will define the geologic events which led to the deposition and preservation of the reservoirs in these fields and provide an explanation for the lack of reservoir units in surrounding areas.

The area of study (Figure 2.1) is located in central Alberta, approximately 50 km west of Red Deer. This area covers 12 townships and consists of townships 37, 38, 39 and 40 and ranges 3, 4 and 5 west of the fifth meridian. As shown in Figure 2.2, this incorporates parts of the Willesden Green, Sylvan Lake, Medicine River and Gilby Fields. This area contains prolific hydrocarbon reservoirs at many different stratigraphic levels and therefore drilling activity has been intense since the middle 1950's. Well densities range from as many as 14 wells per section, in range 3w5, to as few as 17 wells per township in the western part of the study area. Well density generally decreases from east to west, parallelling the hydrocarbon potential of the area. In all, 887 wells penetrate at least to the depth of the Glauconite member. Examination of more than one well in an L.S.D. was essential because of the stratigraphic complexity of this area. Because of the numerous unconformities, significant variation was observed over distances of several hundred meters.

Jurassic reservoirs were an early target for exploration, an endeavor which was rewarded with the discovery of the Medicine River Jurassic 'A' pool in 1956 (TerBerg, 1966a). Subsequent discoveries (1958) of the Medicine River Jurassic 'B' and Gilby Jurassic 'B' pools prompted the extensive development of the Jurassic System in this area. There are

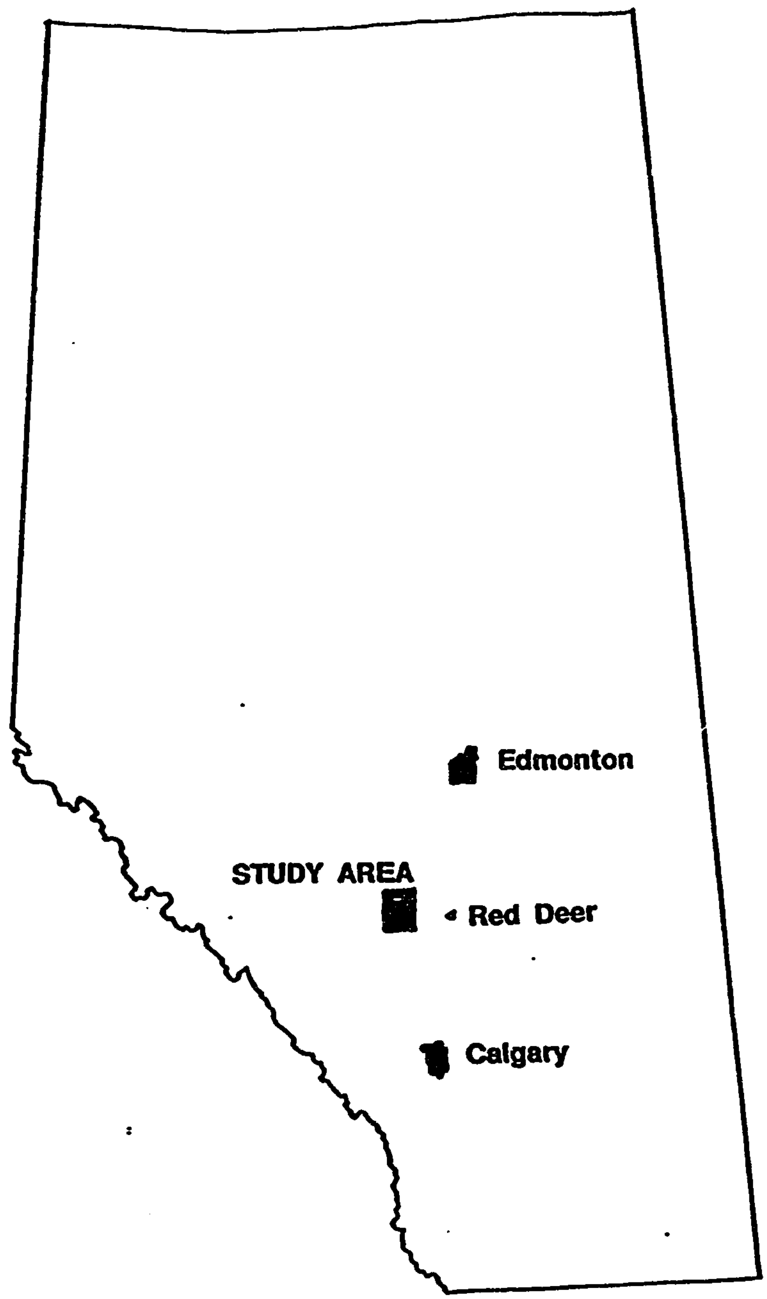

Figure 2.1. Location of the study area in Alberta.

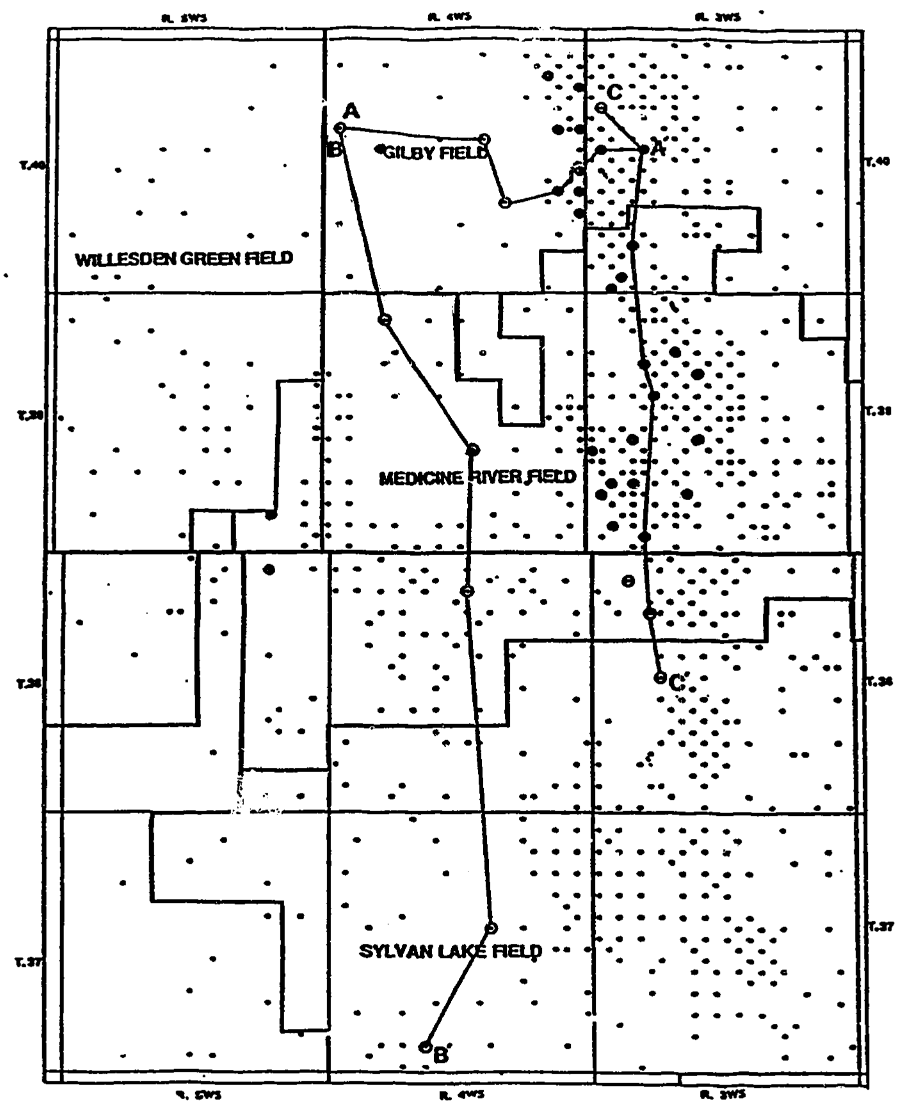

Figure 2.2. Detailed field map of the study area and location of the cross-sections illustrated in this text. Small circles indicate the wells examined; large filled circles indicate location of core examined.

currently six Lower Jurassic oil pools recognized as producing in the Medicine River and Gilby fields, most of which are under enhanced recovery, water-flooding schemes. Data presented here will show that these pools may be divided further into separate pools.
Nordegg-equivalent strata also contain gas pools, updip and in separate outliers of Lower Jurassic strata from the oil pools. One of these gas pools is producing, but has been recognized as the 'Basal Quartz D' pool. Production of both oil and gas from these pools is achieved from reservoirs formed in dolomitic sandstones, which are correlative to the Nordegg Member (Ter Berg, 1966; Watkins, 1966a; Watkins, 1966b; Deere and Bayliss, 1969; Rall, 1980).

REGIONAL STRATIGRAPHY

Before dealing with the detailed stratigraphic relationships in the subsurface of the study area, it is important to understand the regional stratigraphic relationships in the Lower Jurassic. Stratigraphic nomenclature of the Jurassic System is extended from the western outcrop belt, where it was initially developed and extensively described, into the subsurface of west-central Alberta.

LITHOSTRATIGRTAPHY

The term 'Fernie Shale' was first used to distinguish a shaly sequence from overlying Cretaceous sediments in the Blairmore-Frank coal fields of southwestern Alberta (Warren, 1934). These sediments were assigned a Jurassic age with the discovery of the ammonite *Cardioceras*. Formation status was assigned to the Fernie shales by McLearn in 1916 (Warren, 1934). In subsequent years, however, different authors have described the

Fernie as having both formation and group status. Since this unit contains a number of formally and informally defined 'members' and 'beds' (and no formally defined formations), and since the Fernie has never been formally defined as either a group or a formation, this study will follow the recommendation of Hall (1984) and refer to the Fernie as an informal lithostratigraphic unit of formation status.

Lithostratigraphy and biostratigraphy within the Fernie Formation have been dealt with extensively by a number of authors, but few formalized definitions of units exist. This has led to a nomenclature that often mixes lithostratigraphic and biostratigraphic principles, and applies both nomenclature systems to a variety of formal and informal subdivisions of the Fernie formation. These problems have recently been reviewed by Hall (1984) and Hall and Stronach (1984) and their recommendation of using strictly lithostratigraphic units, separated from biostratigraphic designations, will be utilized in this study.

In the northern part of the outcrop belt, the basal unit of the Fernie formation consists of dark grey to black, cherty and phosphatic limestones. This unit has been named the Nordegg (Black Chert) Member by Spivak (1949). The highly fossiliferous *Oxytoma* bed usually caps this unit and is included as part of the Nordegg Member (Hall, 1984). Erdman (1945) described the Oxytoma bed as a fossiliferous limestone unit containing abundant pelecypod shells; of which *Oxytoma* and *Ostrea* were particularly abundant. In the south, the base of the Fernie formation is marked by what Hall (1984) refers to as the Unnamed Basal Shale and Coquina. This unit is highly variable in composition and contains phosphatic conglomerates, conglomeratic sandstones, or coarse, phosphatic, bioclastic rudstones (Stronach, 1984). Since both this and the

Nordegg Member sediments yield Sinemurian fossils, these units are thought to be correlative.

The Red Deer Member of Hall (1984) is defined as resistant, yellowish-weathering, dark grey platy shales and dark limestones. This unit has an extremely limited areal extent and is said to be of Late Pleinsbachian age. There appears to be a major hiatus between the Sinermurian age Nordegg Member, and its equivalents, and the Late Pleinsbachian Red Deer Member.

The Poker Chip Shale (also called the Paper Shale by Frebold, 1976) is probably the most extensive and uniform unit in the Fernie formation. It consists of black, fissile to papery shales, and ranges in thickness from 10 to 38 m (Hall, 1984). This unit may be in contact with any of the Red Deer Member, the Nordegg Member (or its equivalent) and Triassic or Paleozoic sediments. A major hiatus is therefore indicated at this lower contact.

BIOSTRATIGRAPHY

Biostratigraphic zonation of the Jurassic around the world is based on the work of Arkell (1956), who developed the western European standard chart based on ammonite fauna. These standard ammonite zones are thought to be essentially chronostratigraphic units and are grouped into a number of stages (Figure 2.3). The difference between this chart (modified from Hallam, 1975 and Hall, 1984) and that of Arkell (1956) is the insertion of the Aalenian stage, which replaces Arkell's Lower Bajocian.

The dating and correlation of marine Jurassic formations in North America, is based mainly on ammonites in the lower and middle part of the Jurassic, and on ammonites and the bivalve genus *Buchia* in the Upper

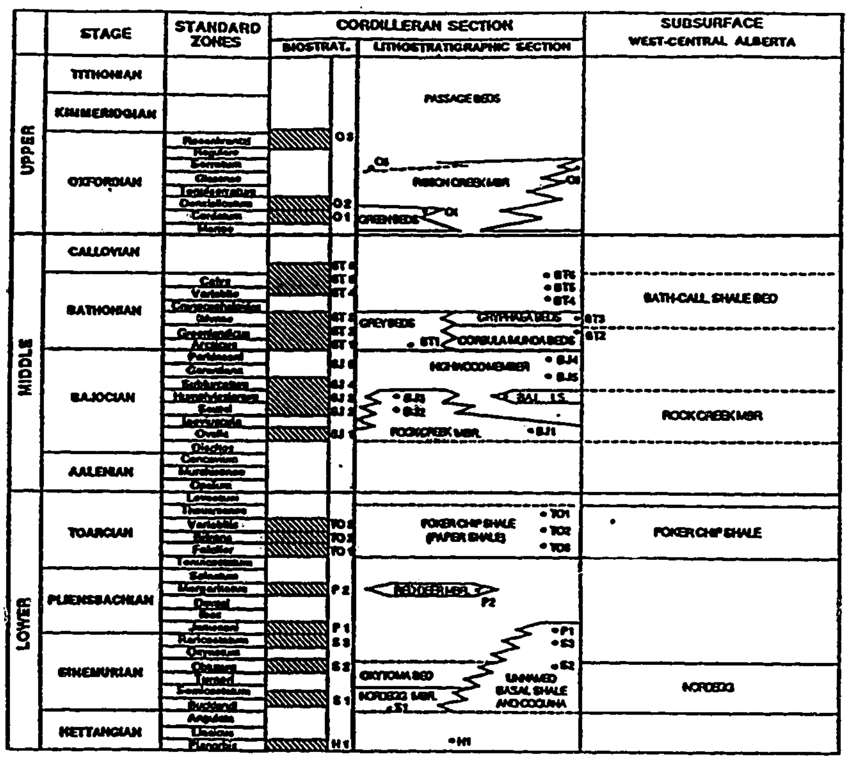

Figure 2.3. Stratigraphic chart of the Jurassic System in Alberta. Cordilleran section is modified from Hall (1984) and the subsurface section is modified from Marion (1984).

Jurassic (Poulton, 1984).

Hettangian age (H1) fossils are known only from the Peace River area; the base of the Fernie Formation is usually formed by Sinemurian or Toarcian age sediments. The oldest (S1) fauna is found in the Nordegg Member and correlates to the upper part of the Bucklandi Zone and the lower part of the Semicostatum Zone. Obtusium Zone (S2) faunas have been found in the unnamed basal coquina. Raricostatum Zone (S3) fossils have been collected in southeastern British Columbia, an area that has also produced fauna of Pliensbachian age, Jamesoni Zone (P1). Other Pleinsbachian age faunas have been equated with the Red Deer Member, and correlate to the Margaritatus Zone.

Ammonites representing the Falcifer (TO1), Bifrons (TO2) and Variablis (TO3) zones have been collected from the Poker Chip Shale throughout the outcrop area. These zones represent the majority of the Toarcian stage and other Toarcian zones are thought to be present but have yet to be proved.

Biostratigraphic and lithologic breaks indicate a number of unconformities in the Fernie formation. A major hiatus is indicated at the base of the Fernie formation, where, in all but a few areas in British Columbia, the Hettangian stage is missing. A second major hiatus occurs at the base of the Toarcian Poker Chip Shale. This erosional gap generally spans the lowermost zone of the Toarcian Stage and the majority of the Middle and Upper Sinemurian Stage. Locally, erosion during this time even removed the Sinemurian sediments, and the Poker Chip Shale forms the lowest unit of the Fernie formation.

JURASSIC IN THE SUBSURFACE OF ALBERTA

Published reports of the Jurassic System in the subsurface of west-central Alberta are relatively sparse and few recent publications exist. The extensive outcrop work on the Jurassic sections of the Cordillera were first extended into the subsurface by Spivak (1949). Based on lithologic characteristics and stratigraphic position, he identified a Fernie and a Kootenay section and subdivided the Lower Fernie into a basal Nordegg Member and the Poker Chip Shale. The Nordegg Member sediments were described as dark grey-brown cherty limestone or silty limestone and the Poker Chip shale as a black, calcareous shale.

Lackie (1958) described the Jurassic subsurface of the Peace River area in northwestern Alberta and northeastern British Columbia. Correlation to Frebold's foothills sections was based on similar lithologic characteristics, stratigraphic position and facies changes. Lackie (1958) described a "chocolate-brown to black shaly unit, which shows strong radioactivity, and which rests unconformably on Triassic or Paleozoic strata" as the Nordegg Member equivalent. Thin interbeds of dense to finely crystalline argillaceous limestones occur within this unit. These sediments change in lithology towards the southeast to typical Nordegg Member chert, dark shales and siliceous or cherty limestones. Overlying the Nordegg equivalent unit are grey-brown to black, fissile to papery shales with abundant pyrite, which were identified as the Poker Chip Shale.

Springer et al., (1964) described the Jurassic strata as extending east from the disturbed belt, under the western plains area, to an eastern erosion limit. This erosional edge strikes north-south, paralleling the disturbed belt, and terminating the Jurassic strata just west of Red Deer. The present study area lies along this erosional edge. Springer et al.,

(1964) described the Nordegg Member in the subsurface, as 'consisting of calcareous, speckled, dark brown bituminous shale and in places of sandy limestones'. In two areas, the Paddle River-McLeod River area and the Gilby Field (this study area), Springer et al., (1964) described the occurrence of fine, partly calcareous sandstones, which are equivalent to the Nordegg Member. Overlying the Nordegg Member, they described dark brown, platy, non-calcareous shales which they correlated to the Poker Chip Shale Member. The Poker Chip Shale was reported to contain a Toarcian age fauna in the Gilby field area, and these shales overly the Nordegg Member equivalent sandstone horizon of the Gilby Field (Springer et al., 1964).

Bovell (1979) described the Nordegg member, north of township 50, as consisting of sandy, phosphatic, fossiliferous packstones and grainstones, laminated biomicrite, cherts, fine grained calcareous sandstones, and dolomites. He concluded that these sediments were deposited on a stable, low relief shelf which was at least 200 to 500 km wide. The Nordegg Member was recognized on the basis of wireline log correlation and similarity of lithology and stratigraphic position (Bovell, 1979). Gas is produced in the Paddle River, Greencourt, Whitecourt and Blue ridge gas fields from secondary porosity in both the sandstone and chert lithologies, along the erosional edge of the Nordegg Member,

Previous Work

As noted earlier, oil was discovered in the Jurassic System (Nordegg equivalent sandstones) in the Medicine River area in October, 1956. The economic significance and good well and core control have prompted several detailed studies of the Jurassic sediments of this area.

Nordegg-equivalent sandstones in the Medicine River Jurassic 'A' and 'C' pools were discussed by Watkins (1966a, 1966b) and in the Gilby Jurassic 'B' pool by Fraser (1966). Watkins described these reservoirs as 'very fine carbonate "clastic", predominantly dolomite grains, with varying percentages of silica sand grains and carbonate matrix, and illustrated a truncation type of trapping mechanism. Fraser (1966) described a similar lithology, but he envisioned the sandstones as draping into an eastern valley, and a related structural closure as the trapping mechanism. Subsequent studies showed that truncation is the trap type in both areas. Original oil in place was estimated for the Jurassic A, B and C pool at $36x10^3$ BBLs, $58x10^3$ BBLs and $75x10^3$ BBLs respectively. Average porosity and permeability across the two fields is on the order of 15% and 30md, respectively (Fraser, 1966; Watkins, 1966a, 1966b).

In a more regionally oriented study, Deere and Bayliss (1969) described the lithology and mineralogy of the Lower Jurassic in an area that extended from the eastern subcrop (present study area) to the western disturbed belt. They distinguished three lithologic units in the Nordegg-equivalent strata; these are: 1) dark brown to black bituminous chert, 2) medium grey bioclastic limestone with abundant pelecypod shell debris and scattered detrital quartz, and 3) medium grey dolomitic sandstone consisting of equal amounts of detrital dolomite and quartz. These units are arranged in broad north to south-trending bands and grade from lithology one through three in a west to east direction. The reservoir-forming sandstone unit is only found in outliers along the eastern subcrop edge in the Medicine River and Gilby Fields. Deere and Bayliss (1969) interpreted the Nordegg-equivalent strata to represent a stable shelf environment in which the sandstone lithology represents the nearshore area with the source area to the east.

Rall (1980) completed a more detailed lithologic and diagenetic study of the sandstone reservoirs in the Medicine River and Gilby area. Here Nordegg-equivalent strata are described as consisting of fine grained "remarkably uniform" quartz, chert and dolomite grain sandstones which display no apparent coarsening or fining sequences and only rarely display recognizable sedimentary structures. These sandstones are capped by a "limestone coquina" interpreted by Rall (1980) to be oyster banks and part of the same depositional system as the underlying clastic unit. This upper limestone was said to indicate a decreasing clastic input into the area and the onset of regression of the Nordegg sea.

As in the outcrop area, the Nordegg Member and its equivalents are overlain by the Poker Chip Shale in the subsurface. Deere and Bayliss (1969) described the overlying Poker Chip Shale in central Alberta as a medium grey, non-calcareous and fissile shale with a mineralogical composition dominated by quartz and illite, with lesser amounts of kaolinite and scattered pyrite. Springer et al., (1964) indicated a Toarcian age for this unit, based on an apparent Toarcian faunal assemblage collected from the shale in the Gilby area. Rall (1980) confirmed this by assigning a probable Toarcian age, based on the occurrence of the dinoflagellate cyst *Nannoceratopsis gracilis senex* .

A deep valley system forms the eastern limit of several of the Nordegg-equivalent oil pools. Ter Berg (1966) describes a series of what he terms 'Upper Jurassic' sandstone reservoirs (UJ1, UJ2 and UJ3), which are unconformity bounded units filling the valley system, in the Medicine River Field. Rall (1980) discovered Poker Chip Shale equivalent (Toarcian age) shales between the lower J1 and the upper J2 sediments, thereby dating the J1 deposits and the valley system as Lower Jurassic in age. In addition, Nordegg lithologies were found in blocks in the J1 deposits indicating a post-Nordegg-lithification age for valley development. Ter Berg (1966) indicated an Upper Jurassic age for the J2 deposit, but Hopkins (1981) suggested a Middle Jurassic to Lowermost Cretaceous age for these deposits. The valley itself, however, can be realistically dated as a lower Jurassic feature, formed after Nordegg lithification and before Poker Chip Shale deposition. This valley will be referred to as the J-valley, named after the deposits that fill it.

SUMMARY

The lower part of the Jurassic in this area consists of, from base to top, the Nordegg member and its equivalents and the Poker Chip Shale. In previous work, the sandstone deposits in the Medicine River and Gilby Fields were referred to as the Jurassic Sandstones (Fraser, 1966), the Nordegg-equivalent sandstones (Deere and Bayliss, 1969; Springer et al., 1964) and the Nordegg sandstones (Rall, 1980; Watkins, 1966a, 1966b; Bovell, 1979). Since the Nordegg Member is defined as a lithostratigraphic unit that does not include a sandstone lithology (Spivak, 1949), the term 'Nordegg sandstone' is inappropriate. In this study, a return to the original lithostratigraphic nomenclature, as suggested by Hall (1984), will be followed. The Nordegg Member-equivalent sandstones, in the Medicine River and Gilby Fields, are therefore informally referred to as the Medicine River Member. It is implied that this unit is stratigraphically equivalent to the Nordegg Member. This name is derived from the field in which oil was first discovered in these deposits (which is in turn named after the river that runs through the field). The Medicine River Member is defined as fine grained, detrital dolomitic and quartzitic sandstones, with varying amounts of calcareous fossil debris. Typical sandstones of this interval are cored in the 06-19-40-3w5 well between 7090ft anf 7140ft; and in the 10-20-39-3w5 well between 7072ft and 7115ft. This definition does not include the 'coquinoid sandstone', which caps the Medicine River Member (described by Rall, 1980), for reasons that will become apparent in future discussion. The term 'Nordegg interval' will be used when the entire section of the Nordegg Member and its equivalent strata are referred to (including the Medicine River Member). The correlation of these strata to the section described in the foothills is illustrated in Figure 2.3.

METHODS

This study is based on detailed stratigraphic correlation of wireline logs, which is supported by core data. In all, 887 wells in the study area were examined (Figure 2.2), including the wells beyond the erosional edge of the Lower Jurassic strata. Twenty eight cores were examined. Most of these are located in the producing field areas, and only three were found west of these fields. Detailed petrographic work has been completed on this unit by Deere and Bayliss (1969) and Rall (1980). Two wells were sampled, in this study, and eight thin sections were examined (four from each well). Because of the previous detailed petrographic work, these sections were only examined on a qualitative basis, to get a 'hands-on' impression of this previous work. Numerous cross-sections were constructed; three of which are illustrated in this paper. The datum generally used for correlation and mapping of the Lower Jurassic strata was the base of the Upper Mannville Glauconite coal, as this is the most consistent and extensive marker horizon in the area. Where correlations extended westward into the Glauconite-Hoadley complex, a basal Glauconite, high gamma and high neutron density, shale marker was used to compensate for the thickening of the Glauconite zone in this area. Datums in the Mannville Group had to be used because of the lack of consistent markers in the Jurassic System, which is the result of a number of unconformities within the Jurassic strata. These datums worked fairly well, however, when correlating west to east. Westward dip discordances of approximately 2 m/km and 3 m/km had to be compensated for, in Jurassic and Mississippian correlation, respectively.

Since the Lower Jurassic rests on the Mississippian unconformity, paleotopography on this surface is important in understanding the geology of the Lower Jurassic sediments. Two isopach maps were generated to detail this paleotopography and will be discussed in more detail in the next section. In addition, isopach maps of the Nordegg Interval, of the Medicine River sandstone member, and of the Poker Chip Shale were generated. All production and core data were supplied from the data files of Dome Petroleum Ltd., and are updated to May, 1987.

RESULTS

MISSISSIPPIAN UNCONFORMITY SURFACE

Mississippian age carbonate rocks form the subcrop beneath the Jurassic strata in west-central Alberta. Four interregional erosional hiatuses: pre-Permian, pre-Triassic, pre-Jurassic and pre-Cretaceous, combined to form the present erosional surface. The effect was a peneplanation of the Mississippian strata below the Jurassic system and a series of parallel, deeply incised valley systems below the Cretaceous strata (Macauly et al., 1964). Westward-increasing subsidence rates during these erosional events resulted in progressively older strata exposed in an eastward direction.

In this study area, the rocks which subcrop at the post-Mississippian unconformity consist of, from base to top: shaly crinoidal limestones and grey shale of the Banff Formation; clean light-coloured bioclastic limestones, which are variably dolomitized, of the Pekisko Formation; fine grained silty dolomites and siltstones of the Shunda Formation and clean fossiliferous dolomites of the Elkton Member of the Turner Valley Formation. Paleotopographic relief on surface is as much as 140m over short distances (1 km) and is particularly irregular in the Sylvan Lake Field area and along the eastern subcrop edge of the Shunda Formation. Structural closures on porous dolomitized limestones of the Elkton and Pekisko Formations, because of the topographic relief beneath the unconformity, form prolific oil and gas reservoirs in the Sylvan Lake, Medicine River and Gilby Fields.

Paleotopographic relief on the post-Mississippian unconformity is important in its effect on controlling both depositional and preservational trends within the overlying Jurassic system. An isopach map of the Banff Formation top to the post-Mississippian unconformity surface (Figure 2.4.) does not show the true paleotopography of the unconformity, that existed during Jurassic time, since local subsidence (discussed later) within the Mississippian strata has produced local relief on the Banff Formation datum. In addition, average westward discordance of dip between Jurassic and Mississippian strata is approximately 1.1m/km (as estimated from structure maps not illustrated in this paper), further distorting the true paleotopography. It is, however, effective in illustrating the subcropping

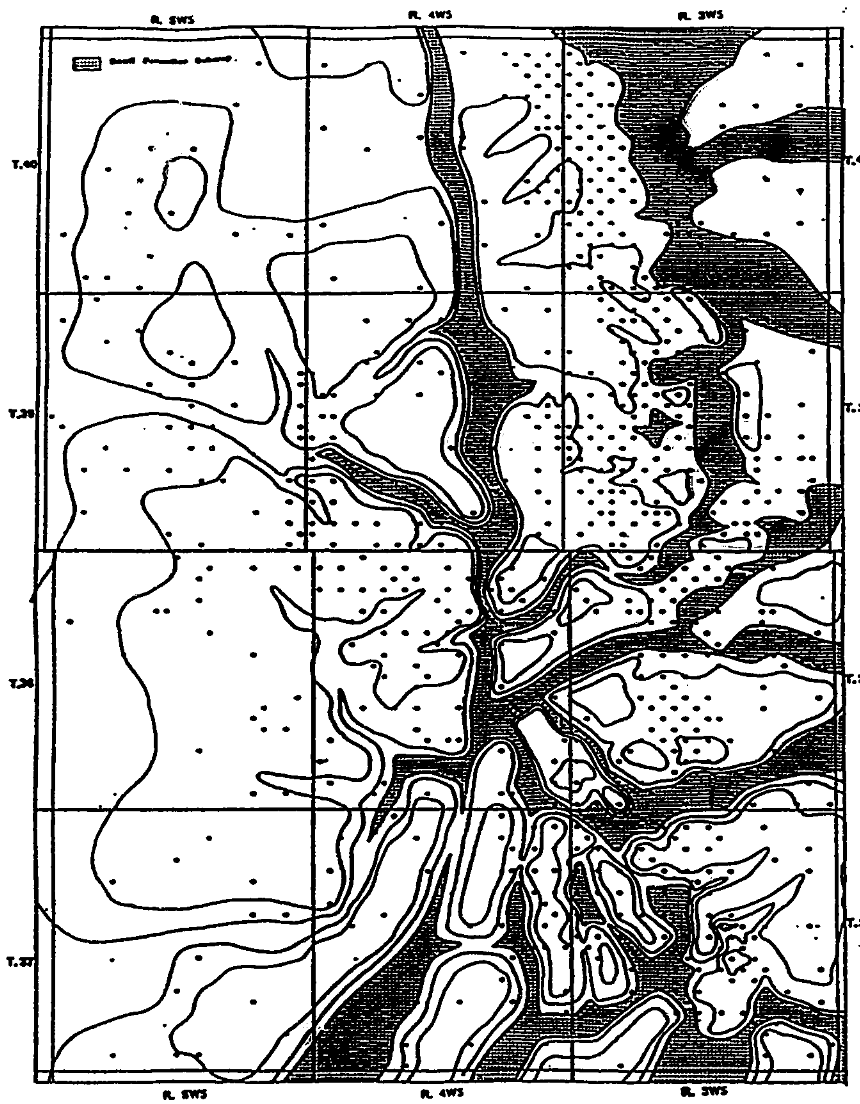

Figure 2.4. Isopach map from the top of the Banff Formation to the post-Mississippian unconformity. Contour interval=20m; stippled areas show where the top of the Banff Formation is eroded and therefore the edge of the 0m isopach.

formations. The Elkton Member is only found at the top of the Mississippian in the west, where westward subsidence of the Mississippian has allowed preservation of the greatest thicknesses of the Mississippian strata. Exceptions to this occur in the Sylvan Lake Field, where Elkton Member outliers form prolific oil and gas pools, east of the normal trend of the Elkton subcrop. The erosional limit of the Shunda Formation is also observed striking in a north-south direction through the central part of this area. The eastern part of the map is dominated by subcropping Pekisko formation, with the Banff Formation only exposed in the deepest portions of various valley systems.

Progressively older Mississippian formations subcrop towards the east, with Elkton and Shunda strata dominating the western subcrop and Pekisko and Banff formations dominating the eastern subcrop. This trend conforms to that observed by Bokman (1963) who described progressively older Paleozoic strata being exposed to the east. This trend indicates pre-Jurassic westward subsidence in response to the initial phases of the Colombian orogeny.

The second map of the Mississippian unconformity (Figure 2.5), an isopach from the Glauconite coal marker bed to the unconformity, approximates more closely Jurassic age topographic relief on this surface. This map compensates for pre-Jurassic local and regional subsidence of the Mississippian strata. However, since it includes Cretaceous strata, it is affected by post Jurassic subsidence. The effect of this subsidence was estimated from structure maps (not shown in this paper) on the Jurassic strata (Nordegg and Rock Creek Member tops) and on the Glauconite coal marker. Eastward convergence of dip between the Jurassic strata and the coal marker is estimated at approximately 2m/km. This discordance must be kept in mind whenever Mannville datums are used in correlating the Jurassic strata (as is the practice in this study).

This map (Figure 2.5) illustrates some active topographic features prevalent during Jurassic time. West of range 3w5, the pre-Jurassic surface was relatively flat and stable, with a slight westward dip. Dissection of this surface was relatively rare and probably occurred during early Cretaceous time. An axial high, approximately one mile wide, trends from north to south along the eastern edge of range 4w5. East of this ridge, in the Medicine River and Gilby areas, the post-Mississippian surface drops gently into a 5 km wide depression whose eastern limit is punctuated by the deep, lower Jurassic age, J-valley described by Rall (1980) and Hopkins (1981). Some parts of this valley are as deep as 100m below the top of the normal Mississippain unconformity. Still further east, the unconformity again rises and conforms to the regional dip of the Mississippian unconformity surface.

The Medicine River and Gilby Fields are, therefore, dominated by an open-to-the-north basin (the northern extent is outside of this area so the actual limits of the depression are not known), which is bounded by Mississipian highs to the east, west and south. Cross sections through this area show that the Banff and Pekisko strata dip into the depression (Figure 2.4) A structure map of the Banff Formation indicates lower than the regional average dips in this basin. Therefore, this basin developed as a result of local subsidence of Mississippian, and possibly older strata, and is not an erosional feature on the post-Mississippian unconformity.

The Nordegg interval throughout the area overlies the Mississippian strata; the Elkton and Shunda Formations to the west and the Pekisko Formation to the east. Only a few wells west of the axial high contain

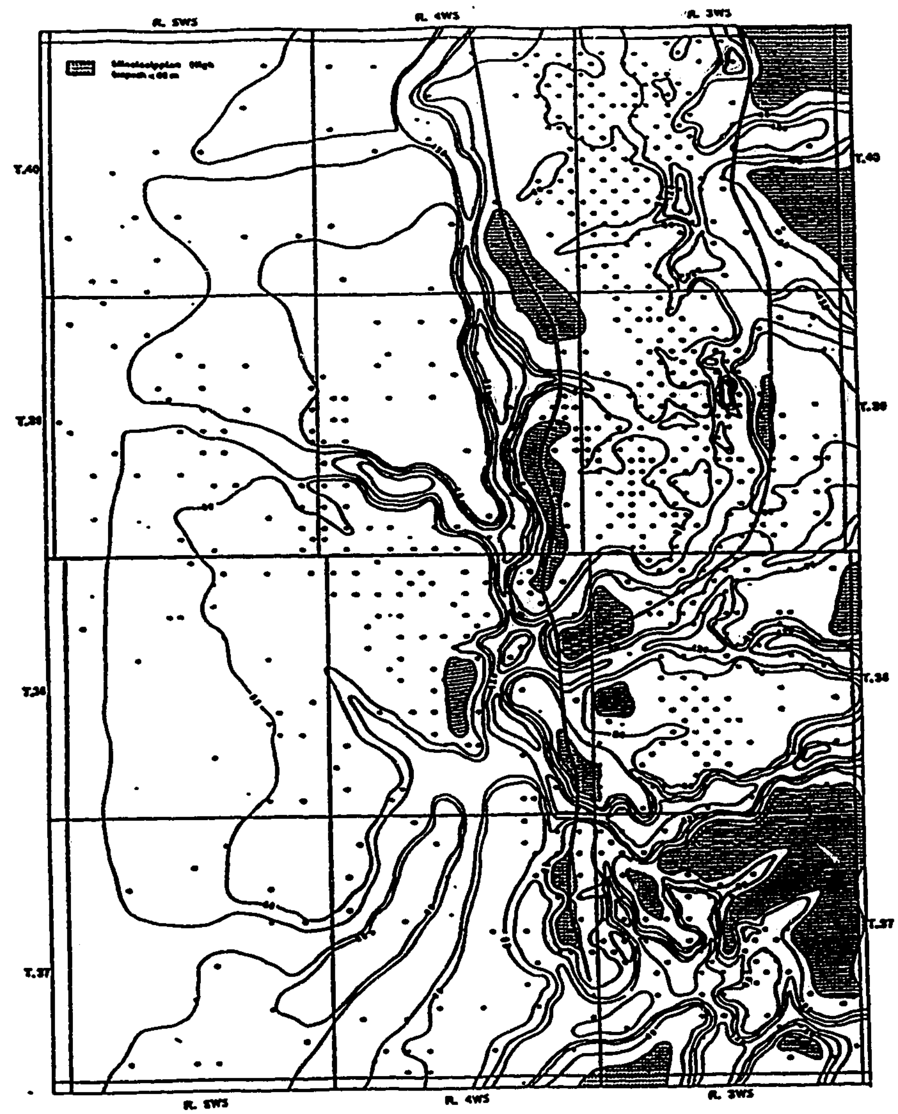

Figure 2.5. Isopach map from the base of the Glauconite coal to the post-Mississippian unconformity. Solid line on the west side indicates the location of the axial high while the line to the right indicates the eastern limit of the Medicine River-Gilby basin. Contour interval=20m.

core through the Nordegg/Mississippian contact. In all cases Nordegg Member limestones rest on sharp and irregular surfaces which show no evidence of a residual zone of dolomite breccias. In addition, logs showed little evidence (ie. gamma 'spikes') of any residual zone development across this contact. On the other hand, in almost every well east of the axial high region, this contact is marked by a pronounced spike in the gamma log and very low resistivity readings. Core through the contact in this area is abundant and reveals a zone of dolomite breccias and fractured Mississippian lithologies. The common lithology of this zone consists of large (from pebble to cobble size), angular blocks of dolomite in a matrix of either coarse dolomite, pebbly sand or green pyritic shales. These are often interbedded with massive, finely crystalline dolomite, which may be fractured; or with massive green shales containing rare angular chert or dolomite clasts. Occasionally green shales with dolomite clasts are enclosed in massive dolomite and apparently fill originally open cavities. Laminations within the green shale can often be observed conforming to the microtopography on dolomite in these cavity fills. These types of deposits are typical of karst terrains and are found as cave filling sediments, solution collapse breccias, and capping soil horizons. Some of these dolomite breccias appear to be paleosol horizons developed on the extensive exposure surface of the Mississippian unconformity while others appear to fill former cavities. Soil deposits were probably piped down into the cavities through joints, a process common to karst terrains (Jennings, 1975).

In one core of this contact (10-06-39-3w5; Plate 1.4), a near vertically dipping surface was observed between the Mississippian dolomite and the Nordegg Member-equivalent sandstones. Steeply inclined bedding within the overlying sandstone conforms to the dip of the contact, and it appears that this sediment was smeared along this contact. This surface is interpreted as a fault plane, which was active after the deposition of the lowest part of the Medicine River Member, but while the sediment was still relatively unlithified. Such structural features can be used to explain the development of the residual zone, found predominantly in the Medicine River-Gilby depression. Faulting in the basin area would result in the fracturing of the Mississippian strata, which in turn would allow the permeation of acidic surface waters into the Mississippian carbonates. Such waters would readily dissolve the carbonate rocks, developing a karst-type surface. In areas lacking faulting, as in the area west of the axial high, karstification would be minimal and the products of karstification more likely eroded during the transgression of the Nordegg sea.

Faulting in Mississippian strata appears to be the the primary process responsible for the development of the Medicine River-Gilby sub-basin. Solution collapse within the underlying Mississippian carbonates, related to the karst surface developed on these faulted strata, may have played an important role in the localized increase in subsidence rates.

NORDEGG INTERVAL

The Nordegg Member is identified on the basis of its stratigraphic position, between the Mississippian unconformity and the Poker Chip Shale, and on its characteristic lithology of bioclastic limestones or bituminous and phosphatic calcareous shales (Lackie, 1958; Spivak, 1949). The Medicine River Member occurs in the same stratigraphic position but consists of fine grained, variably fossiliferous and calcareous, dolomitic sandstones, which towards the west are interbedded with and grade into the bioclastic limestones of the Nordegg Member.

These characteristic lithologies are generally easily recognizable on geophysical logs, making regional correlation relatively straight-forward. Nordegg Member limestones are characterized by very high resistivity, which is in sharp contrast to low resistivity found in underlying shaly dolomites of the Shunda Formation and in the overlying Poker Chip Shale. Where the Nordegg Member limestone overlies dolomites of the Elkton Formation, they are identified on the basis of density differences on sonic and density porosity logs. This becomes particularly useful in the southwestern part of the area where thin (1-2m) Nordegg limestones overlie Elkton dolomites. Pe curves are extremely valuable tools in mapping Nordegg sediments, since this log makes the differentiation between silicate-, dolomite-, and calcite- based rocks very easy. Another characteristic feature of the Nordegg Member, and its equivalent sandstone, is a basal residual zone of dolomite breccias and green shale which commonly appears as a sharp 'kick' on the gamma log, particularly toward the east. The Medicine River Member shows an irregular and characteristically high gamma profile with a strong, positive SP deflection (Rall, 1980).

Based on core and geophysical log data, an isopach map (Figure 2.6) of the entire Nordegg interval was generated. Only the northwest part of the area is covered by the Nordegg interval and a general thinning trend toward the south and east is discernable. In areas west of the Mississippian axial high, the Nordegg interval thickness corresponds approximately to the depth of the unconformity below the Glauconite Coal marker bed (as seen from the map in Figure 2.6). The Nordegg interval is

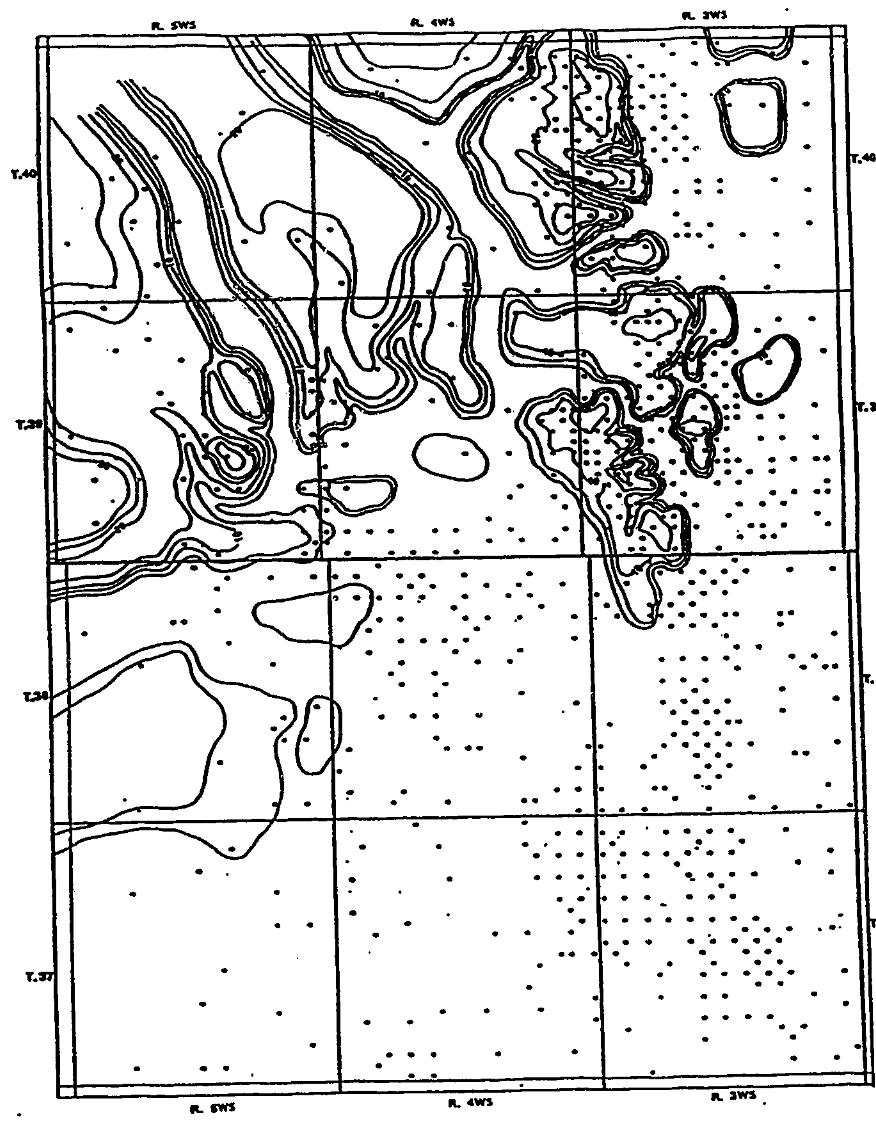

Figure 2.6. Isopach Map of the Nordegg interval; thickness variation is primarily due to erosion. Contour interval=5m.

planed off and is absent over areas where this depth is less than 80m. In areas where the unconformity is deeper than 80m below the coal, the thickness of the Nordegg interval approximately mirrors the structural trends of the Mississippian unconformity. Thickness trends in the Nordegg interval are independent of depositional strike, which is approximately north-south (Deere and Bayliss, 1969), and are purely a function of pre-Poker Chip Shale erosion. Thinning, therefore, appears to be in response to paleostructure (not necessarily paleotopography) on the Mississippian unconformity. In areas of greater pre-Poker Chip Shale subsidence, greater thicknesses of Nordegg interval sediments are preserved. Greater subsidence appears to have affected the northernmost and westernmost parts of the area.

The average thickness of the undissected Nordegg interval is in the order of 20m, with a marked increase in thickness along the westernmost area to over 30m. This coincides with the deepening of the Mississippian unconformity to greater than 100m below the Glauconite coal, which indicates a deepening of the Nordegg basin to the west. To the east, the Nordegg interval thins over the axial high, and is locally eroded beneath the Poker Chip Shale. Still farther east, the Nordegg interval thickens into the local depression on the Mississippian unconformity, where it hosts the prolific oil reservoirs of the Medicine River and Gilby Field Jurassic reservoirs. Maximum thickening here is up to 25m in the Gilby Field. The Nordegg interval thins from north to south within this depression.

The eastern edge is marked by numerous isolated outliers which are separated from the western subcrop by a deep north-south trending valley incision (R. 4w5), or locally by erosion of the entire Nordegg interval beneath the Poker Chip Shale over the axial high. The main north-south trending outiier is dissected into segments by subordinate Cretaceous valley systems, trending roughly perpendicular to the deep valleys.

NORDEGG INTERVAL STRATIGRAPHY

Stratigraphy within the Nordegg interval is more complex than indicated by previous workers. The internal stratigraphy of the Nordegg is best illustrated in the northern part of the area, where the thickest and most complete section of the Nordegg interval is preserved by greater post-Nordegg-deposition subsidence. Cross section A-A' (Figure 2.7) stretches from the western thick area, across the axial high and into the Gilby Field basin. This cross section shows that the Nordegg interval is clearly divisible into two distinct units with a sharp contact between. The lower unit contains typical Nordegg lithologies of bituminous calcareous shales and bioclastic limestones in the west, which grade into the fine sandstones of the Medicine River Member to the east. The upper unit consists of coquinoid limestones and rests with a sharp contact on the lower unit. This unit is laterally continuous and is usually dominated by the shells of the pelecypod genus *Ostrea*; therefore the upper unit is here informally referred to as the '*Ostrea* bed'.

Lower Unit

The lower unit contains three predominant facies which pass gradationally from one to one another. In a west to east direction, these are: a) very calcareous shales, b) bioclastic limestones and c) dolomitic sandstones.

The calcareous shales were only observed west of the axial ridge, and only in the northern part of the area. This facies was encountered in the

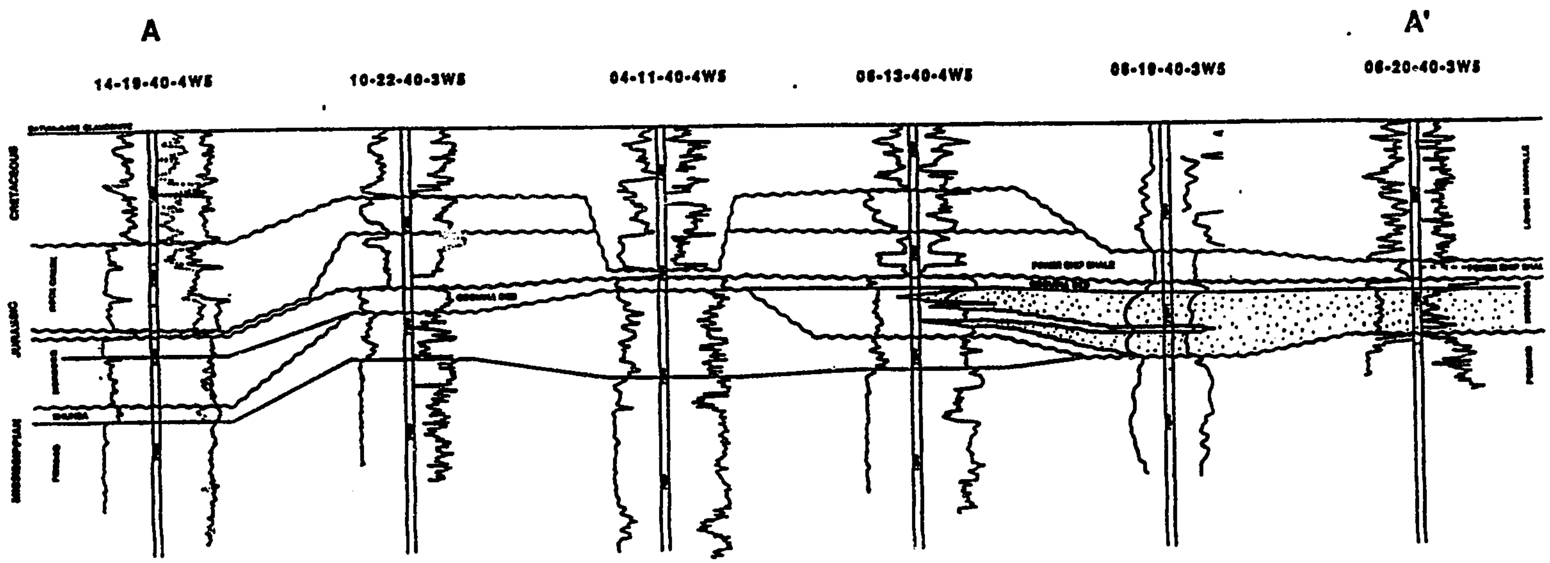

Figure 2.7. Cross-section A-A' shows the stratigraphy of the Nordegg interval across the Gilby field.

14-19-40-4w5 well (Figure 2.7) and observed in core in the 06-20-40-4w5 well. The latter well is an extremely old well, with very poor log quality, so the core had to be correlated to the good quality log in the 14-19 well. Since lithologies suggested by the latter log match up well to stratigraphic equivalent depths on the old log, and since these wells are only 2km apart, using this interchange of data between the two wells is reasonable. In the 6-20-40-4w5 this lower unit consists of massive dark grey to black, organic-rich, very calcareous shales with rare, extensively bioturbated interbeds of siltstone.

The dark massive nature of this unit and the abundance of preserved organic matter suggest a reducing and poorly oxygenated substrate which was inhospitable to most organisms. There is a general lack of megafossils, which suggests that poor circulation of bottom waters allowed for stagnation and therefore poor oxygenation at the sediment/water interface and in the substrate. These conditions suggest relatively low energy conditions. The shales are occasionally interrupted by thin, extensively bioturbated siltstone beds which suggest the sudden influx of terrigenous material and oxygenation of the bottom waters by higher energy events. Events such as these allow brief colonization of the substrate by benthic organisms and hence lead to the biogenic reworking of the coarser grained material. These higher energy events are probably related to storms which transported coarse sediment out onto the shelf. Evidence of storms was also identified in the shoreward deposits of the Medicine River sandstones.

To the east, the calcareous shales grade into light-grey skeletal grainstones which contain abundant fine-grained detrital quartz, dolomite and phosphate sand. These sediments are generally structureless and appear thoroughly bioturbated. Occasional 25cm sets of high-angle, planar cross-bedding, and trough cross-bedding with scour surfaces are observed. Abundant thick-shelled, and often articulated, pelecypods, with a prominent ribbed ornamentation (Plate 2.3b), are common in this unit, particularly in the bioturbated portions. This facies grades eastwards into the Medicine River sandstones, being interbedded with sandstones proximal to the axial high.

In thin section, abundant echinoid plates and pelecypod shells make up the majority of the framework grains, implying a normal marine environment. Occasional high angle cross bedding and a lack of carbonate mud suggest a relatively high energy regime. These sediments are interpreted to have been deposited on an open shelf which was affected by high energy waves or currents. Higher energy conditions provided a very mobile substrate, which inhibited infaunal organisms, and therefore allowed for the preservation of the cross bedding. In areas of slightly less energy and a more stable substrate, abundant infaunal organisms, including the prominently ribbed pelecyopods, reworked the sediment, destroying most evidence of physical sedimentary structures.

The Medicine River Member forms the eastern part of the subcropping Nordegg interval. These sandstones are generally buff to light brown in colour and consist of varying proportions of fine detrital dolomite, quartz and phosphate sand grains, with minor amounts of chert and traces of metamorphic polycrystalline quartz grains (Plates 2.7 and 2.8). These observations are consistent with those of Rall (1980) except for the presence of detrital phosphate. The abundance of moldic and vuggy porosity indicates that these were once fossiliferous sandstones. Occasionally, chert replacement of the fossil debris has preserved the broken pelecypod shells. This deposit is generally very uniform and massive in appearance, with little or no evidence of sedimentary or biogenic structures. The abundance of solution porosity, which can be as high as 20% of the rock volume, suggests that diagenetic alteration of this deposit was extensive. Solution of both silica (quartz and chert) and carbonate grains occurred, and in many cases molds were later filled with blocky calcite spar or chert (Rall, 1980). It is suggested here that the massive appearance of this unit is related to the intense, destructive diagenetic alteration, which obliterated both sedimentary and biogenic structures.

Occasional sharp-based, 10cm beds of coquinoid sandstone, with abundant broken, and randomly oriented, pelecypod shells, are present within these sandstones. The matrix is generally very sandy and is often the host of pervasive poikilitic calcite cementation. Chertification of this cement and of the fossils is common. Correlation of these beds between wells is not evident, indicating the localized nature of these deposits. Their sharp erosive bases indicate high energy conditions capable of winnowing out fine sand, concentrating large fossil fragments and transporting them basinward. Brenner and Davies (1973) described similar thin-bedded (3-30 cm thick) and laterally restricted (10m lateral continuity) deposits, with erosive bases and quartzose sandy matrices, as relatively nearshore storm lag deposits. Thus, the Medicine River Member coquinoid sandstones probably represent high energy storm events which eroded the shoreface, winnowed out coarse shell debris, and transported it shelfward.

As indicated above, the majority of the Medicine River Member consists of massive uniform sandstones with few observable sedimentary or biogenic structures. Sedimentary structures were rarely preserved by favorable diagenetic alteration. Large scale (>20 cm) sets of high-angle planar cross-lamination are observed in several wells, usually as the result of preferential oil staining along enhanced solution porosity in the laminations. These laminations were perhaps once composed of calcareous fossil debris, which has since been dissolved. Patchy, and probably early, cementation by chert has also preserved structures in some wells, particularly toward the south, where chert cementation is more common. These structuresinclude trough cross-lamination with bounding scour surfaces and ripple lamination.

In one well (12-17-39-3w5) chert cementation of fine grained sandstone over a 20 cm interval has preserved a sequence of sedimentary structures which consists of, from base to top: massive sandstone; a single 10cm long, 3 cm high, asymmetrical current ripple; small-scale trough cross-lamination; gently inclined parallel lamination, with low-angle discordant truncation surfaces; and a second bed of small-scale trough cross lamination. Bioturbation in this sequence is restricted to the massive sandstone at the base, where thorough churning of the sediment was probably responsible for the massive nature of this bed; and a single 2cm high, 1cm diameter escape trace. The low-angle lamination in this sequence is interpreted to be the result of a foreshore environment, where high energy flat bed conditions are the result of swash and backwash currents (Reading, 1978). The substrate in this environment is very mobile ,and therefore this is a very harsh environment for most marine organisms. Such an environment would accommodate only the most hardy and mobile of organisms, which were able to adjust to constantly shifting substrates and alternating periods of erosion and deposition. This accounts for the general lack of biogenic structures, and the single escape trace in Plate 2.2d represents an organism's response to increased sedimentation rates. The small-scale trough cross-bedding and the ripple structures were probably deposited in a shallow subtidal setting, where deeper water and slightly lower energy conditions allowed ripples to form.

In most wells the sandstones are very calcareous at the base, but contain little or no calcite towards the top. Eight Medicine River sandstone thin sections, from two wells (06-20-40-3w5 and 06-29-39-3w5), were examined. In one, calcite occurs as a patchy, pore-filling, spar cement which decreases upward and is present as a rare remnant pore-filler in the upper reservoir section. In this case, it appears that the trend of decreasing-upward calcite content was a response to diagenetic processes. In the second well, the lower portion is a very calcareous sandstone or very sandy limestone, which is composed of microcrystalline calcite mud, pelecypod shell fragments and echinoid plate fragments. In the calcite-free section, no evidence of fossil molds was observed. In this case the decrease in calcite appears to have been a response to depositional processes, probably related to an increase in terrigenous sediment supply. In the first case, the calcite cement probably also represents original depositional calcite (fossil debris and micrite mud), which was later dissolved and replaced by spar cement. It is suggested that the decreasing-upward calcite trend throughout the Medicine River Member, is a depositional trend, which resulted in a general increasing detrital input throughout the deposition of the lower clastic unit. This suggestion is supported by the westernmost wells of the Gilby field, where cores and logs indicate that only the upper 5m of the lower unit contains non-calcareous sandstones. Here, the base is dominated by very calcareous sandstones which are interbedded with, and grade into, the skeletal limestones of the Nordegg Member.

This decreasing-upward calcite trend is in contrast to work by Rall (1980), who used an increasing-upward carbonate trend to suggest the onset of regression and the decrease in terrigenous sediment supply. The discrepancy between this and the above observation probably owes its origin to the inclusion of the upper *Ostrea* bed, by Rall (1980), in the same depositional system as the lower unit. Since these belong to different depositional systems (see discussion below), a decrease in clastic input for the lower unit cannot be inferred from the calcareous nature of the upper limestone. This trend suggests that the clastic system was prograding onto a carbonate shelf environment.

Towards the west, the sandstones become more calcareous in nature and contain abundant skeletal debris. In some instances whole or broken pelecypods with similar ornamentation to those described from the bioclastic limestones, are observed. Many of these fossils are articulated, suggesting that they are infaunal species. The intense diagenetic alteration seen in the eastern and southern parts of the area is not as dominant here, and it appears that pervasive calcite cementation may have preserved much of the original structure and fossil material. Occasional low-angle parallel lamination, consisting of coarse skeletal debris, and rare 25cm sets of high-angle cross lamination are observed. These calcareous sandstones are, however, more commonly massive in appearance due to intense biogenic reworking. This deposit is transitional between the bioclastic limestones in the west and the Medicine River sandstones to the east.

The sandstone lithology, which is the reservoir facies of the Medicine River and Gilby Fields, is restricted to the east and south by erosion, and to the west by a facies change into the bioclastic limestone lithology (Figure 2.8). The transitional deposits trend north-south, suggesting that the depositional trends within the lower Nordegg unit were in this direction and that the shoreline lay to the east. Where the sandstones grade into the bioclastic limestones, well logs and core show the interbedding of the limestones and sandstones.

Although it is difficult to determine the precise depositional environment of a massive uniform sandstone deposit, the rare sedimentary stuctures, abundant fossil debris and occasional sharp-based fossil beds do give some indication of the depositional regime. Abundant echinoderm and pelecypod debris indicates a shallow, fully marine setting. The presence of the foreshore sedimentary structures gives evidence of at least some of the sand being deposited in a beach environment. The nature of the beach, whether along a strandplain or barrier island, is not clear, since the eastern erosional edge prevents the identification of associated landward facies of either environment or the mapping of sand body geometry. The high energy nature of the shoreface is indicated by the large scale cross lamination and the general paucity of biogenic structures. One striking feature of these sandstones is the general lack of very fine grained sediments (clay and silt) and the abundance of calcareous fossils. The proximity of these sandstones to a limestone dominated shelf suggests that there was little fine grained material available in suspension. High concentrations of suspended fine clastic material would essentially suffocate such a diverse and densely populated community of calcite-secreting organisms.

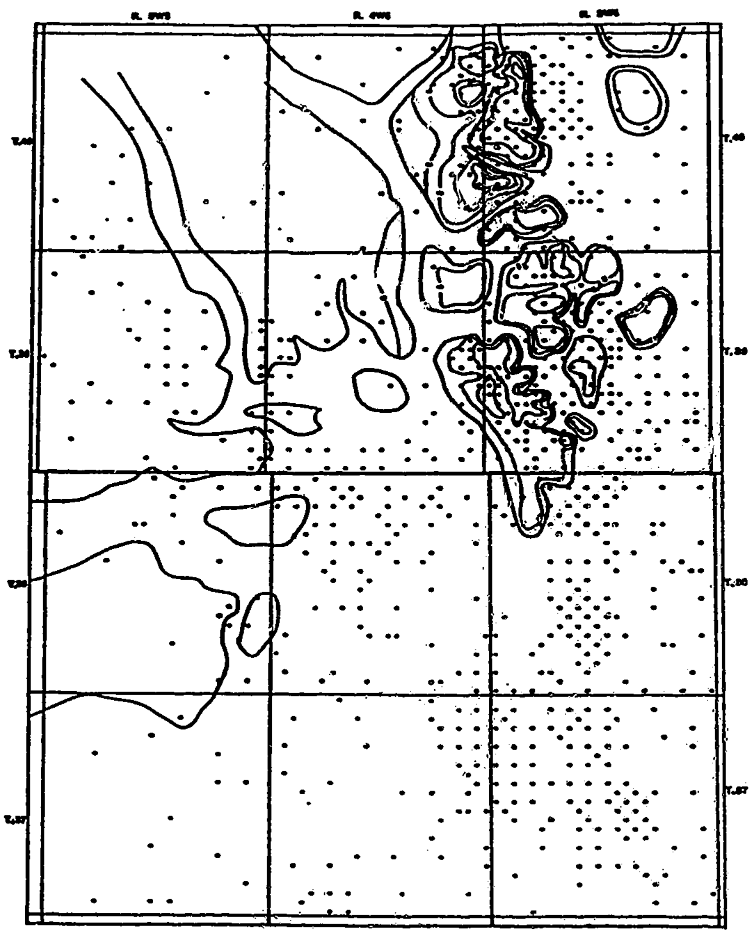

Figure 2.8. Subcrop of the Nordegg interval and the porous sand isopach of the Medicine River sandstone member; contour interval is 5m.

There must have been a fluvial source nearby to supply the clastic material in such quantities that prevented carbonate accumulation in the nearshore zone. The system was probably dominated by longshore drift which removed mud from the system and allowed the carbonate sediments to accumulate. Such a system is similar to parts of the Brazilian coast, where sand-dominated strandlines are prograding out over a carbonate shelf, along a wave and longshore current-dominated delta environment (Domineguez et al., 1987).

Summary

The lower unit of the Nordegg interval appears to have been deposited on a relatively high energy, stable continental shelf, where the deep basin lay to the west and the shoreline to the east. The shelf area was the site of prolific carbonate sediment production, suggesting a shallow, well lit and warm sea which was highly oxygenated. Close to the shore there was an abundant supply of terrigenous sediment, which dominated the system and prevented significant carbonate production. As this supply diminished westwards onto the shelf, calcareous sandstones and bioclastic limestones were able to accumulate. The coast was frequently affected by erosive storm events, which deposited coquinoid sandstone beds within the sandstone member.

Upper Unit: Ostrea Bed

The upper unit consists of limestone, with little terrigenous material, which is dominated by large whole and broken shells of the pelecypod genus *Ostrea*. Many of these fossils are larger than the core diameter (4in) and are up to 1.5 cm thick. These shells are generally found in a matrix of coarsely crystalline bioclastic limestone composed of broken shell debris and calcite spar cement. The matrix becomes finer grained and more organic-rich (darker colour) toward the west, suggesting lower energy and less oxygenated conditions in that direction. The oyster shells are generally oriented in the horizontal plane, although some vertical and random orientations are observed. In a few cases sponge borings, filled with the matrix sediments, were observed in these shells. In rare cases the upper unit consists of mostly light grey skeletal sand with only a few large *Ostrea* shells.

Rall (1980) favored a series of shallow water, high energy, discontinuous oyster beds for the depositional environment of the upper *Ostrea* bed ('limestone coquina' of Rall, 1980) based on the abundance of broken shells, the discontinuous nature and lack of 'basinward shifting facies boundaries'. In the more regional context of this study, the *Ostrea* bed does not appear to be discontinuous, but is a regionally extensive (within this area) deposit which thickens and becomes shalier and more organic-rich basinward (to the west). The patchy distribution of this unit appears to be entirely a function of pre-Poker Chip Shale erosion. In addition, the generally broken morphology of the shell debris described by Rall (1980) was subordinate to the occurrence of whole, and in some cases articulated, oysters seen in this study.

Bovell (1979) described similar sediments in the Nordegg Member to the north, and interpreted them to be large shell banks or oyster reefs, which are characteristic of shallow water (<3m) environments. Therefore, the *Ostrea* bed in the present study area probably suggests deposition in shallow water and the depositional limit of this unit was probably near the east. This interpretation is supported by the depositional thinning that is seen toward the east. In contrast to the underlying unit, the clastic component of this sediment is very minor, suggesting a sediment-starved type of deposit. The base of the *Ostrea* bed is often observed to contain pebbles of the underlying strata.

Deposition of the *Ostrea* bed therefore occurred during a period of low clastic input in a shallow marine, nearshore environment, which allowed for the accumulation of very concentrated deposits of whole and broken calcareous shell material. The pebbles at the base of this unit suggest that this sediment was deposited during a transgression, and sediment starvation resulted from this transgression. This unit is therefore interpreted as a nearshore condensed section.

Stratigraphic Relationship

The Ostrea bed appears to be a regional continuous deposit which thickens to the east (Figure 2.7). Thickening is probably both depositional, as the shaly marker in the middle of the 10-22-40-4w5 well can be correlated to the thinner eastern wells, and erosional due to pre-Poker Chip Shale erosion. In the western part of the area, this unit overlies calcareous shales of the lower unit as seen in the 14-19-40-4w5 well (and cored in the 06-20-40-4w5 well). The axial high area is shown in the 10-22-40-4w5 and 4-11-40-4w5 wells (Figure 2.7). Over this ridge, the *Ostrea* bed rests directly upon the Mississippian unconformity and even this unit thins over the highest points of the axial ridge. Locally the *Ostrea* bed is completely eroded over this high and the Poker Chip Shale rests directly on Mississippian strata.

East of the high, the aerial distibution of the Ostrea bed is controlled by the extent of erosion predating the Poker Chip Shale deposition, and

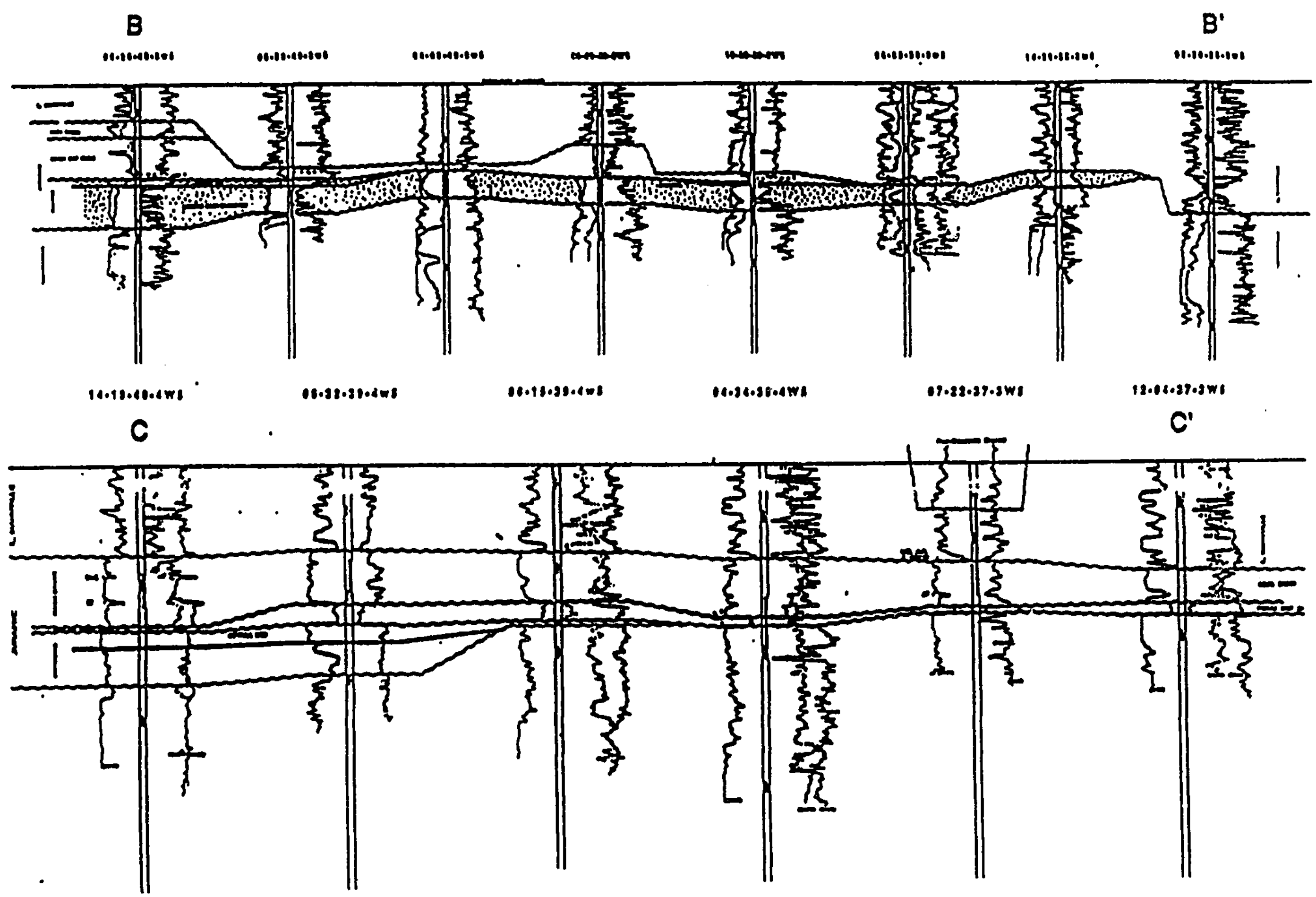

Figure 2.9. Cross-sections B-B' shows the thinning of the Medicine River sandstone, and the loss of the Ostrea bed due to erosion, from north to south, through the Gilby and Medicine River fields. Thinning is accompanied by a rising Mississippian unconformity surface. Cross-section C-C' shows the wedging-out of the Nordegg interval, through the study area, from north to south, due to a rising post-Mississippian unconformity surface.

this distribution is extremely patchy in nature. Preservation of the *Ostrea* bed is generally limited to the northwestern part of the Gilby Field (Figure 2.8), although exceptions do occur (eg. 06-05-39-3w5).

Although the lower unit is absent over the high, depositional trends within the lower unit do not appear to be affected by the axial high. Therefore this high does not appear to have been an emergent feature during the deposition of the lower unit. Absence of the lower unit over the high appears to be due to erosion, which predated the deposition of the *Ostrea* bed. Preservation of the Medicine River Member, east of the high, is therefore due to a locallized subsidence event which postdated the deposition and lithification of the lower unit and predated the *Ostrea* bed.

The unconformable contact between these units was observed, and confirmed in several cores. This contact was everywhere observed as a sharp surface showing several cm of irregular relief. In one well (14-24-40-4w5), where the *Ostrea* bed is in direct contact with the Medicine River Member, the top 10cm of the underlying sandstone contains 6cm diameter bore holes filled with skeletal limestone, similar in appearence to the matrix of the *Ostrea* bed. This suggests that the sandstone was at least partially lithified at the time of *Ostrea* bed deposition, and in places, was exposed to the sea floor for some period of time. In another well (06-12-40-4w5), the top of the sandstone is brecciated, and is overlain by a green shale, with clasts of the Medicine River sandstone contained in it. In 04-36-40-4w5 the top of the sandstone is very irregular and depressions in this surface are overlain by a one-grain-thick layer of rounded dolomite pebbles (1cm in diameter) and coarse sand grains. The dolomite clasts indicate erosion of a local exposure of the Mississippian strata, possibly from the axial high, or perhaps from beyond the eastern limit of the Medicine River sandstone. This is overlain by a green shale and a 1.75 m thick bed of calcareous, laminated green shale with minor interlaminations of fine sandstone. Laminations are generally horizontal with some apparent current ripple laminations; microfaulting is present in the sandier portions. No bioturbation features were observed in this deposit, even though sedimentary structures were readily observed. This type of sediment is consistent with non-marine deposition and probably indicates an alluvial flood plain type of environment. In either case it indicates emergent conditions and a significant hiatus. This, in turn, is overlain by the typical sediments of the *Ostrea* bed.

The *Ostrea* bed has a patchy distribution throughout the Gilby and Medicine River Fields, due to erosion of the top of the Nordegg interval. It is generally more common and laterally extensive in the northern area, where the Nordegg interval is thickest, but is almost entirely missing from the southern area. One exception is found in the 06-05-39-3w5 well, near the southern erosional limit of the sandstone member. Here the top of a 10m interval of Medicine River sandstone is found to be intensely weathered and brecciated, with green, very pyritic shale infilling cavities in the eroded sandstone top. This weathered horizon is overlain by a buff coloured siltstone, of probable nonmarine origin, and by a conglomerate of chert pebbles and partially chertified Medicine River sandstone pebbles, up to 7cm in diameter. Typical *Ostrea* bed sediments are found between this conglomerate and a thin grey-green shale of the Poker Chip Shale. The base of this unit also contains pebbles of the underlying strata, representing a transgressive lag deposit.

Although the *Ostrea* bed is only observed in one well in the Medicine

River Field (06-05-39-3w5), its occurrence at the top of a thin Nordegg interval is significant. It was noted earlier that the Medicine River Member thins from north to south, and that this thinning is erosional, since it is parallel to the depositional strike. The occurrence of the *Ostrea* bed in the southern well, over partially eroded Medicine River sandstones, indicates that erosional thinning was primarily the result of the pre-*Ostrea* bed erosional event and is only accentuated by the pre-Poker Chip Shale erosional event.

Because of the aerially limited nature of Rall's (1980) work, the regional distribution of this unit and its erosional contact with the lower clastic unit, particularly over the Mississippian axial high, was not identified. In addition, the indicators of a substantial unconformity were also not identified. Because of this, Rall (1980) suggested that the *Ostrea* bed indicates an onset of a regressional event of the same sea which deposited the lower clastic unit. These units were thereby included as part of the same depositional system. Data presented here suggest that the sea which deposited the lower clastic unit had long since regressed from the area and that substantial exposure and erosion had removed parts of the lower unit. The *Ostrea* bed is here interpreted as the initial deposits of a later transgression. This transgression eroded the surface of the Medicine River Member, removing most of the terrestrial deposits associated with the unconformity (except in the case of the 04-36-40-3w5 well). The pebbly nature of the base of the*Ostrea* bed supports the erosive nature of this transgressive bed. In some cases (14-24-40-3w5 well) not even a transgressive lag was developed on the lithified Medicine River Member, and this surface lay exposed on the sea floor, allowing boring organisms to inhabit the lithified substrate.

In areas where the *Ostrea* bed is eroded beneath the Poker Chip Shale, the Medicine River Member underwent a second period of Lower Jurassic exposure and erosion. Erosion appears to have been more extensive in the south and this is reflected in the nature of the Medicine River Member. Chertification of the sandstone becomes increasingly dominant from north to south, with chert only forming scattered nodules and preferentially replacing fossil beds in wells such as 10-20-39-3w5; but occurring as extensive, non-selective, replacement of the sandstone in wells such as 06-09-39-3w5. This chertification may also be in response to closer proximity to the pre-Cretaceous unconformity. In addition, the amount of calcite sand grains decreases from north to south with a concomitant tendency of the sandstone to become more massive in nature in that direction. These features combine to indicate an increase in the diagenetic alteration toward the south, which is probably related to the time that the Medicine River sandstones were exposed to the surface and subjected to acidic meteoric and vadose waters.

The Poker Chip Shale-Medicine River sandstone contact was observed in several core. The top of the sandstone was generally brecciated and clasts of the sandstone were often partially or completely replaced by chert. In some places cavities within the brecciated Medicine River sandstone are infilled with green shale (Plate 2.6c) in much the same way as the karst surface of the Mississippian unconformity. Since the Medicine River sandstone is composed of a large proportion of carbonate grains, its surface appears to also have undergone karstification, although to a lesser extent than the Mississippian strata. In some instances, a conglomerate of rounded chert granules and Medicine River sandstone pebbles occurs within or overlying the brecciated zone. The chert clasts are probably the transported erosional products of the Mississippian surface to the east. In one instance (10-20-39-3w5; Plate 2.6b) the top of the Medicine River Member is marked by a light brown, dense calcareous shale with abundant carbonaceous-lined root molds. This deposit is interpreted as a rarely preserved paleosol horizon, formed during the exposure period between the Poker Chip Shale and the *Ostrea* bed. This unit appears very similar on logs to the *Ostrea* bed and, therefore, care must be taken when correlating the *Ostrea* bed.

The Poker Chip Shale itself consists of black, fissile to papery shales which are organic rich, pyritic and generally massive in nature. On logs they exhibit very high gamma levels, very low resistivity, and very high neutron porosity and sonic readings. The density porosity generally shows an opposite trend to the neutron porosity which indicates the very dense but highly organic nature of the shale deposit. At its lower boundary, whether it is in contact with Nordegg sediments or with the Mississippian unconformity, the basal one meter or so is altered from a very black colour to a dull grey-green or bright olive green colour. This feature in shales has been attributed to a decrease in the Fe^{3+}/Fe^{2+} and later removal of the more soluble Fe2+, when rocks are above the water table (Potter et al., 1980). Such oxidation also affected the organic content of the Poker Chip Shale. Oxidation probably occurred during the exposure period associated with the pre-Cretaceous unconformity, where oxidizing meteoric fluids filled porous Medicine River sandstones and attacked the overlying shales. The impermeable nature of the Poker Chip Shale did not allow these oxidizing fluids to penetrate deeply into the unit, thus only the boundaries are affected. Similar alteration occurs in the shales at the top of the Poker Chip Shale, in the detrital shales, and in the shales which fill karst cavities in the Nordegg and the Mississippian strata.

The Poker Chip Shale deposit oversteps the erosional edge of the Nordegg interval to the south and east, except where pre-Cretaceous erosion has removed it from above (Figure 2.10). The base of the Poker Chip Shale is relatively flat and the post Nordegg interval erosion appears to have formed a peneplain surface with little or no incision. It appears that the Nordegg interval was only preserved in areas where there was significant subsidence predating planation, as is seen in the northwest portion of the study area.

The planation surface beneath the Poker Chip Shale is interrupted by the north-south trending valley system that marks the erosional limit to several of the oil pools in the Medicine River sandstone member. A lower Jurassic age of shales in this valley was assigned by Rall (1980). Clasts of Nordegg lithologies, and in particular, apparent clasts of chertified *Ostrea* bed deposits, were found in sediments underlying these shales. Lithified Nordegg blocks, and the steepness of the valley walls, indicate the Nordegg interval sediments were fully lithified at the time that this valley formed. This indicates that the valley system was formed in lower Jurassic time between the lithification of the Nordegg interval and the deposition of the Poker Chip Shale. This valley has been attributed to an incision event postdating the Nordegg sea regression. However, there are some problems with this interpretation. The valley trends parallel to the depositional strike, and therefore is perpendicular to the dip of the basin. It is unlikely that a valley system would be incised to 100m depth parallel to the depositional strike, and it is therefore concluded that this feature is a function of structural movements, which localized incision.

Over the eastern, and parts of the southern portions, of the Medicine

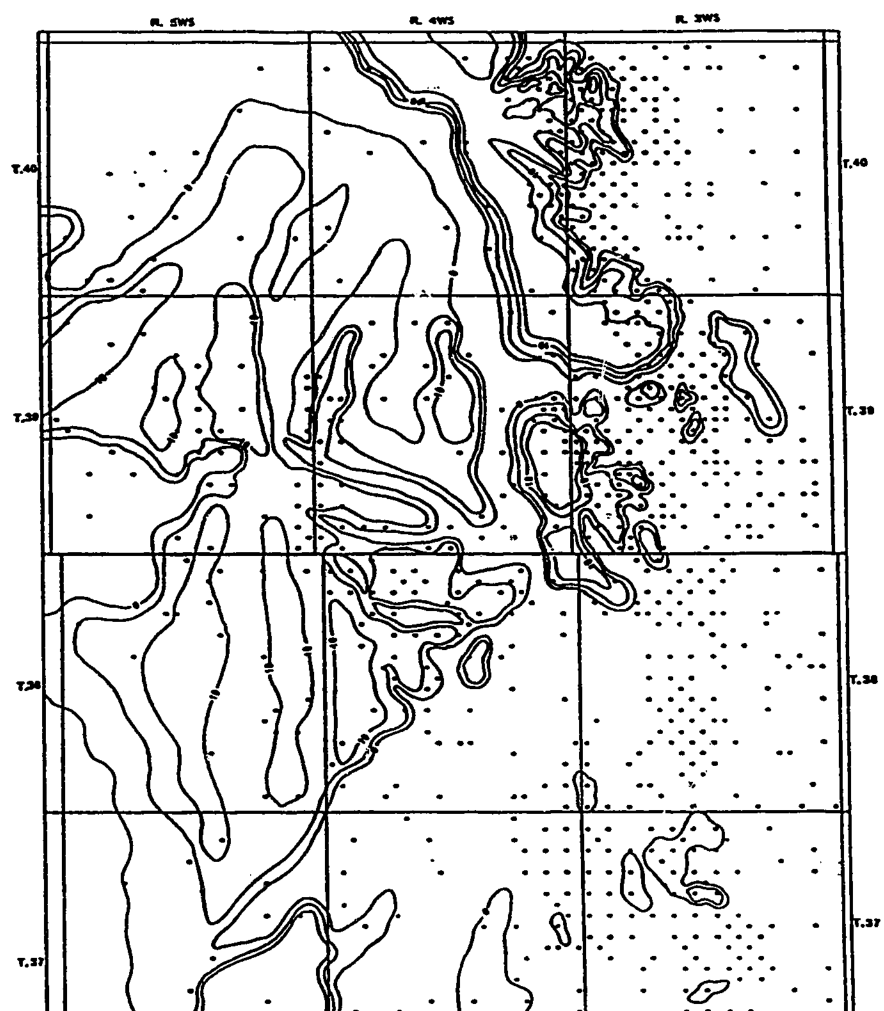

Figure 2.10. Isopach map of the Poker Chip Shale. Contour interval=5m.

River Member the Poker Chip Shale is eroded and Lower Cretaceous Ellerslie sediments lie directly on the Medicine River Member. In this case, the upper boundary is marked by brown carbonaceous shales or heterolithic deposits which appear to be non-marine paleosols or alluvial floodplain deposits. Abundant rooted horizons are often observed in these deposits. In one well (12-17-39-3w5), thick Ellerslie sandstones are in direct contact with the Nordegg interval. The contact is marked by a large stylolite, with several cm scale relief. Ellerslie sandstones are quartzose, with abundant chert and carbonaceous debris, are coarser grained than the Jurassic sandstones and exhibit a fining and shaling-upward trend. The different textures and mineralogy of the two sandstones makes them relatively easy to distinguish in cores and in relatively recent, good quality logs. The carbonate content of the Medicine River sandstones is obvious on Pe and CNL-CDL logs. In addition, the Medicine River sandstones contain abundant detrital phosphate, in which radioactive elements appear to cause high gamma readings in clean porous sandstones. High gamma readings in Ellerslie sediments indicate high clay content.

DISCUSSION

SEQUENCE STRATIGRAPHY PRINCIPLES

Lower Jurassic deposits in western Alberta accumulated on a low relief, stable passive margin, which abutted against a low-lying craton to the east, and deepened progressively to the west. The shelf at this time was extensive, stretching westwards at least 200 to 500 kilometers (Bovell, 1979). Since these deposits are bounded by, and contain several major unconformities, they are best described in the context of sequence stratigraphic principles. Sequence stratigraphy is a branch of stratigraphy which subdivides the rock record, using a succession of depositional seqences, which are constrained within a chronostratigraphic framework, defined by surfaces of non-deposition, erosion and correlative conformities (Haq et al., 1989; Van Wagoner, 1989). A depositional sequence is defined as a relatively conformable succession of genetically related strata, bounded by unconformities and their correlative conformities (Mitchum, 1977). These sequences are interpreted to be deposited during one cycle of sea-level fluctuation, between the inflection points on the falling limbs of the sea-level curve (Haq et al., 1989; Posamentier and Vail, 1989; Sarg, 1989).

An unconformity, as defined by Van Wagoner et al., (1989), is used here as a surface separating younger from older strata, along which there is evidence of subaerial exposure or subaerial erosional truncation, with a significant hiatus indicated. Two types of unconformities are recognized in the sequence stratigraphy literature. Type 1 unconformities occur where the rate of eustatic fall exceeds the rate of subsidence at the depositional shoreline break, and is therefore characterized by stream rejuvenation and fluvial incision, sediment bypass of the shelf and an abrupt basinward shift of facies and coastal onlap (Posamentier and Vail, 1989). Type 2 unconformities are more subtle and occur where the rate of eustatic fall is less than the rate of subsidence at the depositional shoreline break; this unconformity type is therefore characterized by a basinward shift in facies and coastal onlap, no stream rejuvenation, or fluvial incision, and by slow widespread subaerial erosion and a gradual denundation or degradation of the land surface (Posmantier and Vail, 1989). Type 1 unconformities are favored by low rates of subsidence and rapid rates of eustatic sea-level fall, whereas type 2 unconformities are favored by high rates of subsidence and slow rates of eustatic sea-level fall.

Sequences are built on a hierarchy of sediment packages which are stacked on a shelf during a eustatic sea-level cycle. The basic building block is the depositional system, which is defined as a three-dimensional assemblage of lithofacies (Van Wagoner et al., 1989). These are combined to form system tracts, which are the linkage of contemporaneous depositional systems (Brown and Fisher, 1977). Based on the type of the lower bounding unconformity, sequences are classified as type 1 or type 2 sequences; the former rests on a type 1 unconformity while the latter rests on a type 2 unconformity. Type 1 sequences are composed of lowstand-, transgressive- and highstand- system tracts, while type 2 sequences are composed of shelf margin-, transgressive- and highstand- system tracts. The difference is that in type 1 sequences, streams carry sediment to the shelf edge and develop lowstand fans and wedges on the shelf slope; whereas in type 2 sequences the streams debouch onto the middle shelf and deposit parallic or deltaic sediments on the shelf, with no associated lowstand fan on the shelf slope (Posamentier and Vail, 1989).

The arrangement and preservation of sytem tracts within depositional sequences, and of the sequences themselves, are dependent upon the interbalance between sediment supply, eustatic variation and subsidence (Jervey, 1989). The combination of subsidence and eustatic variation will affect the relative position of the sea level, which in turn controls the ability of a shelf to accommodate sediment (Jervey, 1989). Sediment supply controls the ability of shoreline deposition to keep pace with relative changes in sea level. Sedimentation rates lower than the rate of relative sea-level rise will result in transgression and onlap of the coastline; while sedimentation rates exceeding relative sea level rise will result in progradation and downlap of the coastline. A balance of these factors will cause the coast to aggrade vertically until an imbalance is reached. During times of transgression, sediment supplied to the shelf becomes trapped in flooded river valleys and therefore sediment supply to the coast becomes reduced. This results in a state of sediment starvation and the development of widespread marine condensed sediments over the shelf. Condensed sections are the result of extremely slow rates of sedimentation and are therefore characterized by a high concentration of pelagic sediment, glauconite or phosphatic minerals, burrowed or bored sufaces, or carbonate deposits. Condensed sections represent times of maximum flooding.

THE LOWER JURASSIC SHELF

The passive continental margin of western Alberta during Sinemurian time consisted of a low relief surface, which dipped gently westward for several hundreds of kilometers. The climate during this period was considerably more equable than that of today (Hallam, 1975) and the location of the Jurassic sediments in western Canada are considered to be significantly farther north than at the time of deposition (Bovell, 1979). Tropical climatic conditions probably prevailed during the deposition of the Nordegg interval and therefore favored carbonate accumulation on the shelf; particularly where there was no significant input of terrigenous clastic material.

Where clastic sediment supply was sufficient, thick accumulations of relatively mud-free sands accumulated along the shoreline. These sandstones are dominated by detrital dolomite and phosphate, suggesting that they were derived from cherty Paleozoic carbonate rocks and previously deposited Jurassic or Triassic phosphatic rocks, which outcropped along the low-lying margin of the stable craton. Sediment supply must have been low to allow for the removal of the mud from the system and to allow for the coeval accumulation of clastic and carbonate sediments. This setting is therefore similar to the continental margin along the east coast of Brazil (with the exception of the area of major sediment input from the Amazon drainage system) where small river systems supply sand to an otherwise carbonate-dominated shelf (Dominguez et al., 1987). Here, the dominance of longshore drift removes the fine grained sediment along the coast and therefore allows for the accumulation of carbonate sediments on the stable shelf. The clastic wedge along the Brazilian coast has prograded basinward over these carbonate sediments since Pleistocene time (Dominguez et al., 1987).

Subsidence rates on passive margin shelves are a function of lithospheric cooling and sediment loading (Jervey, 1989). Since the Lower Jurassic shelf was an area of carbonate accumulation and low sediment supply, sediment loading was not a significant factor. This shelf would therefore have been characterized by low, relatively stable rates of subsidence which gradually increased basinward. Sequence preservation is a function of the shelf's ability to accommodate sediment. Areas of low subsidence rates are more likely to develop type 1 unconformities and therefore undergo significantly greater levels of erosion (Posamentier and Vail, 1989). The relatively thin development of the Jurassic strata, and the abundance of unconformities, attest to the low rates of subsidence (and therefore accommodation) on this shelf. Localized perturbations in subsidence rates occurred but will be dealt with later in this text.

The Nordegg interval shelf was therefore characterized by warm shallow seas, low sediment influx and low and stable rates of subsidence. In such a stable setting, the pattern of facies distribution and preservation is primarily controlled by fluctuation in the eustatic sea level and secondarily by the local variation in subsidence rates.

The current theories of sequence stratigraphy originated in the Gulf coast area, where abundant sediment supply and rapid subsidence make for a very different system from that of the Jurassic system of western Canada (Hallam, 1989). Application of these principles to the Nordegg interval must therefore take into account the low sedimentation and subsidence rates which affected this shelf.

SEQUENCE STRATIGRAPHY OF THE NORDEGG INTERVAL

The Lower Jurassic Nordegg interval is bounded below by the post-Mississippian unconformity; a Type 1 interregional unconformity, which can be correlated around the world. This unconformity correlates to a global megacycle eustatic sea level fall, which exposed the Mississippian strata from Permian to lower Jurassic time. The upper bounding unconformity to the Nordegg interval occurs below the Toarcian age Poker Chip Shale. This, too, is probably a type 1 unconformity, as significant valley systems formed during the time of subaerial exposure of the Nordegg interval. Although these are not interpreted as incised valleys, Lower Jurassic fluvial deposits are found up to 30 m below the depositional surface of the Nordegg interval, suggesting that the base level was probably low enough to form a type 1 unconformity.

The Nordegg interval is divided into two separate depositional sequences by an unconformity surface. This surface shows evidence of significant subaerial exposure and truncation, but, valley incision is not observed. In a study of this local nature, it is difficult to establish the type of unconformity associated with this surface. The Oxytoma bed in the outcrop area has similar lithologic characteristics to the *Ostrea* bed here, suggesting that these units may be correlative. If so, the underlying unconformity may be extended into the outcrop area. If this correlation holds true, it would suggest that the unconformity extended out over the shelf and may be a type 1 unconformity. Angular truncation of the underlying sequence, due to local subsidence (Figure 2.7), also suggests that the unconformity is a type 1 surface.

These unconformities are sequence boundaries as defined by Van Wagoner et al., (1989), representing periods of prolonged subaerial exposure and significant erosion of the underlying sequence. They therefore supply good chronostratigraphic datums, between which Nordegg facies distribution patterns can be constrained.

The Nordegg interval, in the study area, can therefore be divided into two depositional sequences. Both of these show considerable truncation by the overlying unconformities and therefore, the full depositional sequences are not preserved in either case. This is a function of low subsidence rates, and therefore low preservation potential on this shelf. The present study area lies relatively close to the eroded depositional limit of the Nordegg interval, as suggested by the detrital constituents of the sandstones, and by the thin and nearshore nature of the deposits of both sequences. This area probably lay proximal to the landward edge of the Sinemurian age shelf; or as described by Posamentier and Vail (1989) near the hinge-line of this basin.

The lower sequence consists of nearshore clastic sediments, which includes sandstones deposited in foreshore and shoreface environments. This clastic depositional system prograded westward onto a carbonate-dominated shelf system. This environment is dominated by the influx of dolomitic sandstones, which were probably derived from the Mississippian subcrop not too far to the east. The lower sequence, in the Medicine River area, represents the progradational phase of the highstand system tract, as it consists of prograding nearshore sandstones near the basin hinge-line.

The upper depositional sequence is often eroded completely beneath the pre-Toarcian sequence boundary, but the basal deposits, where preserved, consist of the *Ostrea* bed. These deposits represent deposition in the nearshore zone where very little terrigenous material was supplied to the system. The base of this bed generally contains abundant Nordegg interval and Mississippian derived pebbles. The *Ostrea* bed is therefore interpreted as a transgressive deposit with a basal transgressive lag. Transgression, which postdated the lithification of the lower sequence, was probably relatively rapid and drowned low-lying areas and river valleys, essentially trapping terrigenous sediment inland. This allowed the nearshore zone to be dominated by a series of oyster banks. Extremely slow sedimentation rates are indicated by the boring of the underlying lithified sandstones and the high concentration of the oysters in this deposit. This bed is therefore recognized as a condensed marine section that reflects the time of maximum marine flooding during the second sequence. The *Ostrea* bed is therefore the transgressive system tract of the upper Nordegg interval sequence. The progradational phase of the highstand system tract of this sequence was probably removed by erosion during the pre-Toarcian erosional event.

The third Lower Jurassic depositional sequence was initiated by the Toarcian relative sea level rise. It overlies a type I sequence boundary and consists of organic-rich radioactive marine shales. The Poker Chip Shale fits the description of Galloway's (1989) condensed section, as it is a widespread radioactive marine mudstone, reflecting slow sedimentation and concentration of organic matter. This deposit therefore reflects a period of maximum flooding which is associated with the Toarcian depositional sequence.

The Nordegg interval in the Medicine River and Gilby areas is therefore dominated by two depostional sequences, separated by an erosional and subaerial unconformity. The first consists of the highstand system tract, containing shoreline terrigenous sedimentation, prograding over the carbonate-dominated shelf. The second is represented by a transgressive system tract deposit, dominated by a condensed accumulation of oysters. The top of the latter sequence was eroded in this area, and therefore the highstand deposits of this sequence are not preserved.

Comparison To Global Lower Jurassic Sequences

Subsidence on the Lower Jurassic shelf was relatively stable and rates relatively low. Sediment input was also relatively low. It is unlikely that either of these processes was strong enough to affect the relative sea level fluctuations and therefore, the sequence architecture seen in this study is the reflection of variation in eustatic sea-level.

Global curves of coastal onlap and eustatic fluctuations (Figure 2.11), in the Lower Jurassic series, were recently published By Haq etal., (1989). The Sinemurian period is shown to contain two third-order cycles of

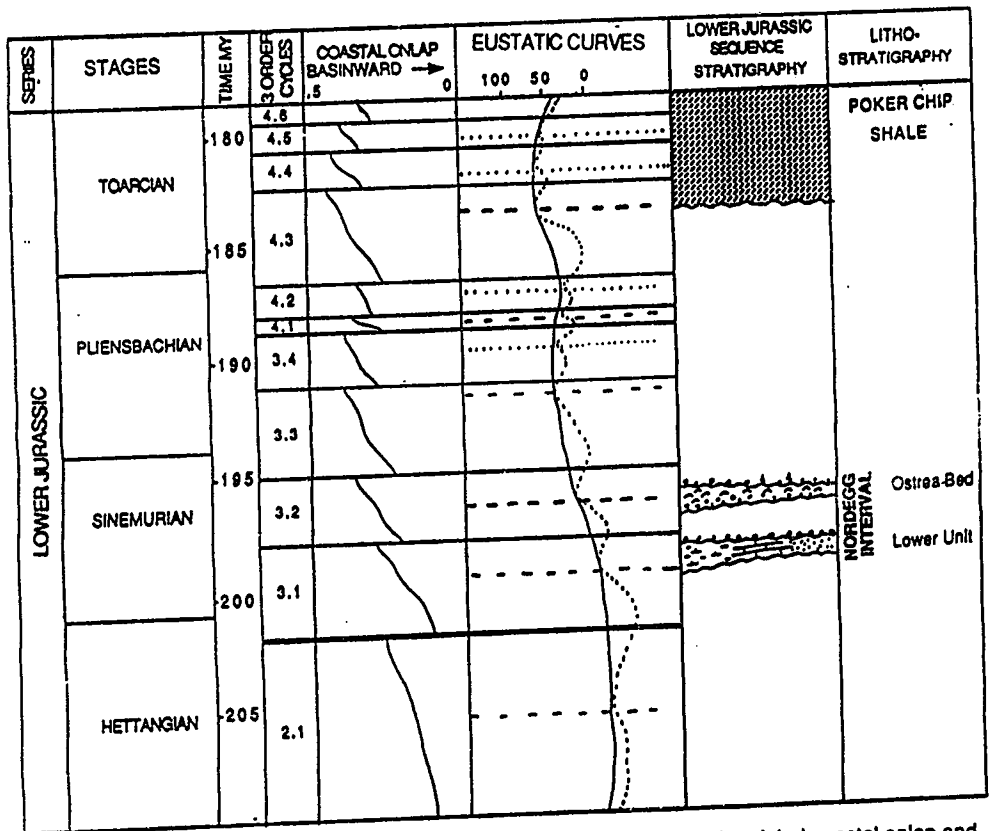

Figure 2.11. Comparison of the sequences described in this area to the global coastal onlap and sea-level curves of Vail and others (1989).

eustatic sea level fluctuation, which punctuate the overall transgressive arm of a fourth-order cycle. Fluctuation of the sea level associated with the shorter term cycles is on the order of 20m on this chart. The coastal onlap curve associated with the eustatic fluctuations shows two periods of onlap, the second extending farther landward than the first, separated by a 'minor' type 1 sequence boundary.

The right side of this diagram illustrates the sequences observed in the study area and the interpreted correlation to these global cycles. The placement of these two sequences is speculative, since there is no chronostratigraphic support. However, the sequence architecture suggests that such a representation is highly plausible. The lower unit is interpreted as the highstand deposits associated with the 199.5 m.y. transgression (Haq et al., 1989). Coastal onlap of the subsequent 196.5 m.y. transgression invaded farther landward, and therefore may have deposited the condensed section of the *Ostrea* bed over shoreline deposits of the lower sequence highstand system tract. Later erosion at the 196 m.y. sequence boundary may have removed the highstand progradational deposits of this upper sequence. Finally, in the Toarcian period, a major eustatic rise, approximately twice the magnitude of earlier Sinemurian eustatic fluctuations, caused the transgression of the Lower Jurassic sea. Coastal onlap was, globally as well as in this area, much more extensive than in the earlier systems. The Poker Chip Shale is therefore more extensive than the Nordegg interval and consists of condensed deposits that overstep the shoreline deposits of the Sinemurian period.

Although there is no biostratigraphic evidence to support correlation of this sequence to Haq et al's., (1989) sea level curve, it is logical to do so, since eustatic variation would appear to be the dominant factor affecting the sequence architecture of the Nordegg interval deposits. The two depositional sequences described in this area fit well with these curves, but a more regional correlation of the sequences and biostratigraphic zonation is required to confirm this correlation.

SUBSIDENCE

Subsidence refers to the sinking of the basin floor by any process, the most important of which are lithospheric cooling and sediment loading on passive continental margins (Jervey, 1989). In order for sediments to accumulate on such a stable shelf, there must be space available below base level. Base level is approximated by sea level in passive margin settings. Accommodation is therefore a function of both sea level fluctuation and subsidence, both of which affect the position of relative sea-level and therefore the accommodation potential (Jervey, 1989).

The Lower Jurassic margin appears to have undergone increasing subsidence to the west, since progressively older strata are exposed on the Mississippian unconformity in that direction. The Nordegg interval is generally eroded beneath the Toarcian, Poker Chip Shale throughout the majority of this area. A rise in base level, associated with the Toarcian eustatic sea level rise, appears to have been too late to preserve much of the Nordegg interval sediments in this area. In some areas, however, the Nordegg interval is preserved, and may attain thicknesses as great as 30m. Accommodation, and therefore preservation, of the Nordegg interval appears to be primarily a function of subsidence. Substantial subsidence in the northwest corner of the study area brought a thick section of Nordegg sediment below the base level and this sediment was preserved beneath the Poker Chip Shale. Subsidence progressively decreased to the east, therefore the preserved section of Nordegg interval sediments also decreases to the east. The same trend is observed to the south, where the Nordegg interval is truncated below the Poker Chip Shale.

An interesting and economically important anomaly to the eastward truncation of the Nordegg interval occurs in the Medicine River and Gilby fields. Here subsidence of the Mississippian unconformity created a small subsidence basin, which dropped a thick sequence of the Medicine River Member below the base level. This subsidence event, as demonstrated earlier, predated the pre-*Ostrea* bed erosional event and therefore preserved the reservoir sandstones in these fields. Outside of this local basin, where subsidence was low, the Medicine River Member is very thin or more usually absent, because it was truncated by pre-Poker Chip Shale and pre-*Ostrea* bed erosion. Preservation of these thick sandstone sections in the Medicine River and Gilby fields is a local anomaly which owes its origin to subsidence tectonics.

It is therefore important, from an exploration and development point of view, to determine the origin of Medicine River-Gilby subsidence basin. Downward structural movements in this basin area are indicated by structure trends within the Mississippian strata (discussed earlier), and this subsidence therefore owes its origin to movements in the basin floor. The presence of faults, observed directly in core, underlying this basin has already been discussed and subsidence appears to be at least in part due to downfaulting of the Mississippian strata. In addition, the relationship of the residual zone not only supports the idea of a faulted carbonate terrain underlying the basin, but also indicates that subsidence may also have been accentuated by collapse tectonics on the karst surface of the Mississippian unconformity.

This raises the questions of why faulting and subsidence are restricted to such a local area and why this area in particular? The answers to these questions may lie in the relationship of this area to the Leduc Reef trend. The reef trends approximately north-northwest beneath the area (Figure 2.12). During deposition of the Jurassic system, the Mississippian strata would have been relatively close to the surface and therefore acting as a rigid tabular block forming the bedrock beneath the shelf. Differential subsidence of Mississippian strata, around the Leduc Reef trend, is indicated in regional cross-sections presented in the Oil and Gas Maps of Western Canada (CSPG, 1981). If there was ongoing differential subsidence occurring around the reef during Lower Jurassic time, it would be expressed in the Mississippian strata as extensional faulting and there would be downdropped blocks parallel to the edge of the reef. The Medicine River-Gilby basin however, does not parallel the reef but instead parallels the strike of the Shelf. Although differential subsidence may be important to subsidence in some areas, it does not seem to be an adequate process to explain the origin of this basin. A variation of this mechanism, however, may explain the development of the Medicine River-Gilby basin.

At the time of Lower Jurassic deposition, the shelf was undergoing progressive deepening to the west because of increasing subsidence along the basin margin (possibly in response to the initial phase of subduction related to the western Cordillera). The reef would act as a rigid bench, trending obliquely to the stress field. The westward-increasing subsidence would cause a flexural stress regime on the rigid Mississippian crust, and therefore an extensional fracture system would develop in the Mississippian system. The fracture trend would be aligned

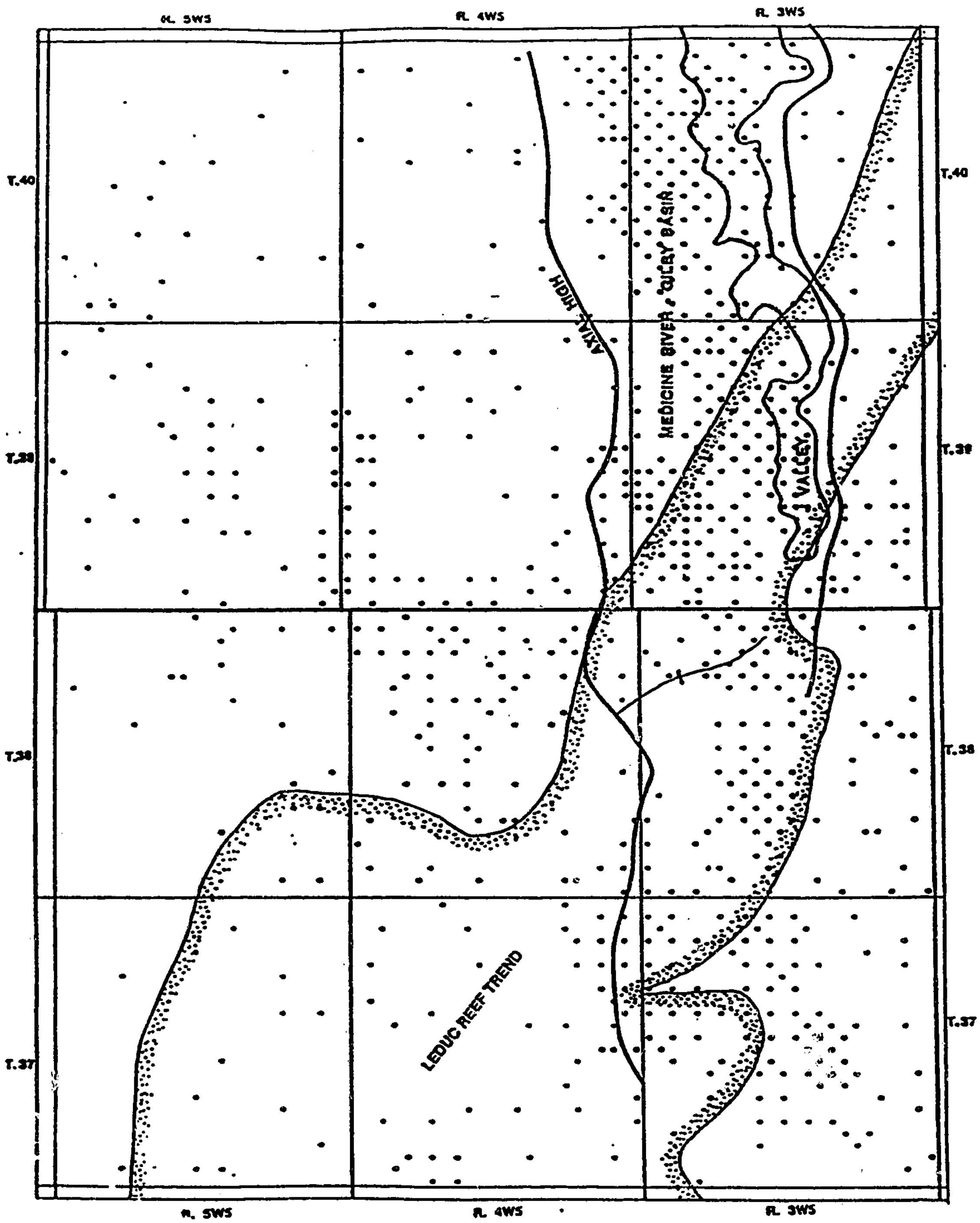

Figure 2.12. This map shows the relationship of the Medicine River-Gilby subsidence basin to the location of the Leduc reef trend.

parallel to the stress field, hence in the direction parallel to the Medicine River-Gilby basin. Faulting may have been active during the deposition of the Medicine River sandstone member and therefore directly related to the subsidence which allowed for the reservoir preservation. Alternatively, the fracture systems may have been old structures that were rejuvenated by sediment loading during deposition of the Nordegg interval.

IMPLICATIONS FOR HYDROCARBON ACCUMULATIONS

The Medicine River Member forms prolific oil producing reservoirs in the Medicine River and Gilby Fields. The sandstones range up to 30 meters in thickness, and generally thin east and south by truncation beneath the *Ostrea* bed and the Poker Chip Shale. The sandstones are fine grained and consist predominately of detrital dolomite, phosphate and quartz, with varying amounts of fossil debris. Porosity ranges up to 25%, while permeabilities may be as high as one darcy, but average 30md across the two fields. The porosity appears to be primarily secondary in nature and is formed by the dissolution of calcite cement, fossil debris, and to a lesser extent, by dissolution of the terrigenous framework grains (Rall, 1980).

Figure 2.13 shows the relationship of the separate pools to the Nordegg interval subcrop and to the distribution of the sandstone lithology. Also shown on this map are the cumulative production of oil to May, 1987, for each pool, the location of significant gas tests in the sandstone member and the structure of the Nordegg interval. The structure within the Nordegg interval rises to the east, toward the porous sandstones, and therefore gas filled sandstones are restricted to the easternmost subcrop area.

Oil was trapped in the sandstones of the Medicine River Jurassic A and

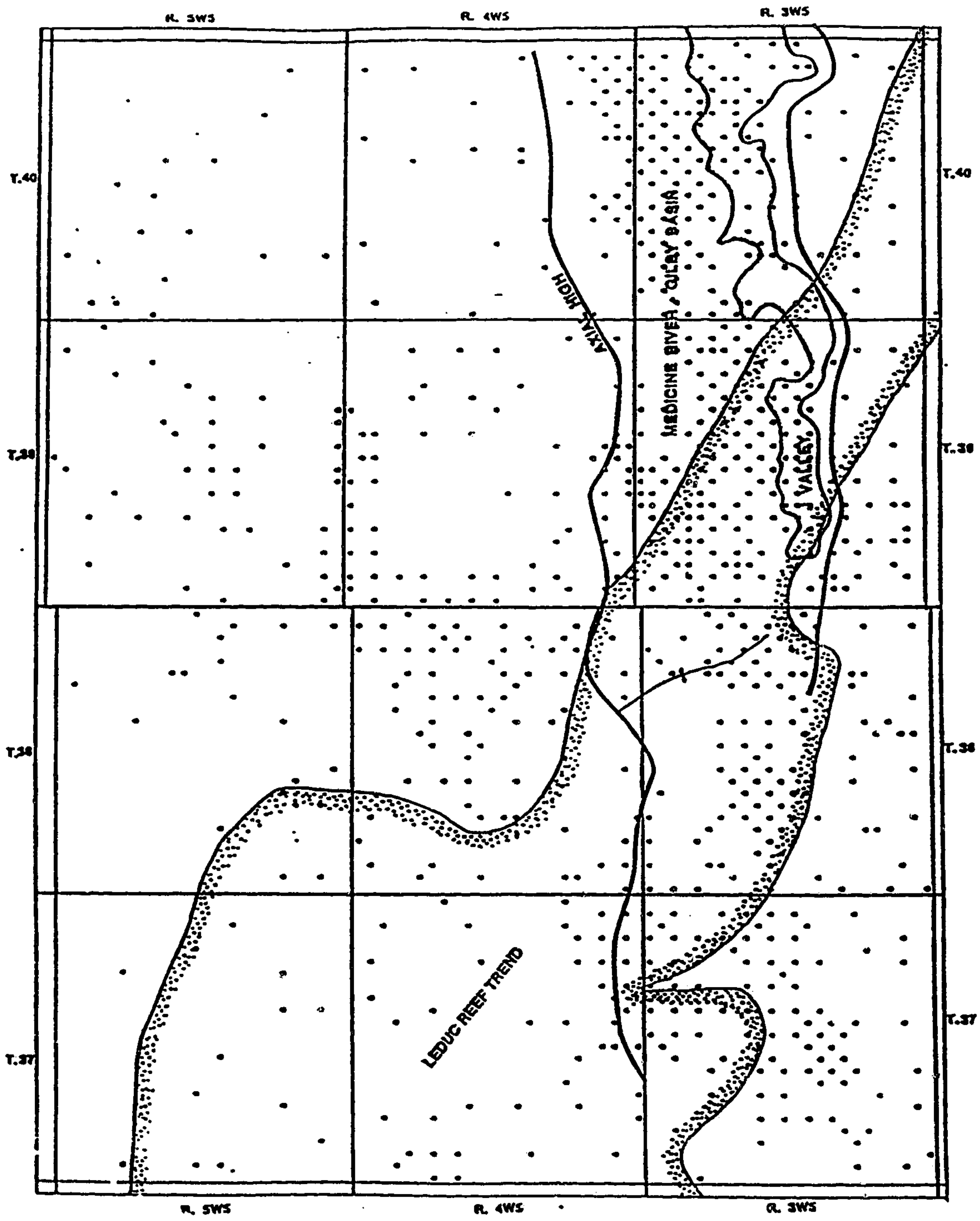

Figure 2.12. This map shows the relationship of the Medicine River-Gilby subsidence basin to the location of the Leduc reef trend.

A' pools and the Gilby Jurassic F and B pools, updip against the side of the J-valley (Figure 2.12), which formed after the lithification of the Nordegg interval sediments and before the deposition of the Poker Chip Shale. The upper part of the fill of this valley, which abuts against the Medicine River sandstone, is impermeable in most cases. Weathering of the Nordegg interval subcrop may also have formed the impermeable seal on the eastern edge of these pools. Where the fill of the upper part of the valley was permeable, leakage from the Medicine River sandstones may have filled these reservoirs. The updip edge of the Medicine River Jurassic C, K and B pools and the gas-filled, non-producing reservoirs in the Gilby area, were formed by a combination of truncation beneath the *Ostrea* bed and Poker Chip Shale, and by later incision during Lower Cretaceous fluvial channelling. The Medicine River Basal Quartz D gas pool is also interpreted to be a Medicine River sandstone reservoir.

Lateral seals of the pools were generally formed by erosion of the subcrop, during lower Cretaceous time, where narrow valleys cut through the Nordegg interval. In some cases these valleys may be only a one hundred meters wide and their presence may only be indicated by a discontinuity in the reservoir fluids (eg. variation in the oil water contacts). Lateral pool separation also occurs where the Nordegg interval is planed off over the Mississippian highs, beneath the Poker Chip Shale or *Ostrea* bed.

Figure 2.13 differs slightly from other published data (Ter Berg, 1966; Rall, 1980) in that the Medicine River A pool is separated into two pools by a structurally high Mississippian surface, over which the Nordegg is planed off beneath the Poker Chip Shale. This break is inferred from the 02-06-40-3w5 well, where core indicates that the Poker Chip Shale is in direct contact with Mississippian strata; and from the 08-06-40-3w5 well, where just a thin Nordegg interval, non-porous limestone, occurs between Mississippian and Poker Chip Shale sediments. In addition, an erosional break appears to divide the Medicine River Jurassic C pool. This break was inferred from a difference in elevation of the oil/water contact across this boundary, and from a 're-entrant' into the Nordegg interval subcrop (Figure 2.13). The break appears to be formed by a very narrow Cretaceous valley which is not observed in any of the pool wells. A similar type of valley separates the Gilby Jurassic F and Jurassic B pools.

The downdip limits of the pools in these fields are generally formed by a water leg in the reservoir. In the case of the pools east of the J-valley, however, this limit is formed by the J-valley itself. Other pools are completely bounded by erosion and are therefore outliers, filled with hyrocarbons.

Total combined oil production from the Medicine River sandstone in these two fields, to May 1987, was 37 million barrels of oil, and most of the reservoirs are currently under water-flood enhancement. It is therefore important that previously unrecognized lateral breaks in the pools be mapped and recognized so that maximum efficiency of the water-flood can be achieved by appropriate location of the injection wells.

IMPLICATIONS FOR FUTURE POTENTIAL

The Nordegg interval has teased explorationists since the discovery of the prolific reservoirs in this area. Only a few areas in Alberta, such as this one and the Paddle River Fields, actually produce from the Lower Jurassic. One of the objects of this study was to identify the reason that this area is anomalously productive and postulate why other areas are not. Both of these areas contain the nearshore sand deposits, but since the depositional trends essentially parallel the subcrop edge, it is not sufficient to suggest that reservoir development is a simple function of pre-Cretaceous erosion. It would seem unlikely that this sandstone is only preserved in two areas in Alberta, especially since it appears that these sandstones were cliff forming units (in the J-valley walls) and therefore relatively resistant to erosion. One would expect more areas of preservation if the sandstone was continuous along the Lower Jurassic shelf edge.

Using the sequence stratigraphic approach, this study has identified the unique conditions that made the preservation of the reservoir unit possible in the Medicine River-Gilby area. It seems that the general trend was for erosion of the lower sequence and preservation of at least the base of the upper sequence. Since this consists of the transgressive system tract it is a sand-poor deposit. Only in areas of greater than average shelf subsidence were the progradational deposits, and hence the porous sandstones of the lower sequence, preserved. Faulting associated with this local subsidence may have enhanced the development of secondary porosity in these sandstones, where open fractures in the sandstone would enhance fluid flow through the tightly cemented rock, and therefore encourage secondary porosity development. Faulting and related joint systems in the underlying Mississippian strata would also allow the free movement of fluids from these rocks into the sandstones, which would not only affect porosity development but also the migration of hydrocarbons.

The origin of these productive fields is related to localized subsidence which allowed the preservation of sandstone deposits. Exploration aimed at finding similar deposits, along the Jurassic subcrop edge, should be geared toward finding areas of similar subsidence relationships. For a similar type of preservational system, subsidence must have occurred after the deposition of the sand-prone sequence, but before erosion of this sequence could be achieved. In this area, subsidence was related to flexure over the reef and similar anomalies should be expected over other reef trends. One thing to keep in mind is that faulting over the reef in this area did not only occur in lower Jurassic time, but also (as is revealed in separate papers in this volume) occurred sporadically throughout the Jurassic, and influenced the preservation of later sediments.

In light of this sequence stratigraphic approach, other more subtle trap types become possible in the Nordegg Member. The erosional edge of the Nordegg interval is interpreted to be very close to the depositional limit of the Nordegg interval shelf, and that would therefore make these sandstone reservoirs part of the highstand system tract. Reservoir sediments should not just be expected to occur in the highstand deposits, but may also occur in shelf margin system tracts or in the transgressive system tract. Similar reservoir sandstones may therefore be associated with the Nordegg interval sequence boundary that separates the two sequences (upper Nordegg interval sequence boundary), shelfward of the erosional limit.

Finally, substantial potential for reservoir development was observed in the carbonate sediments of the *Ostrea* bed, where thin but highly porous intervals occur in conjunction with the overlying unconformity surface. Porosity development appears to be linked to secondary solution of the fossil grains at the unconformity surface. This unit is probably not in hydrodynamic communication with underlying sandstone reservoirs in the Gilby field, as impermeable shales or unaltered limestones overlie the contact between these units. Porosity developed in this area is heavily oil stained, but, the porous interval is very thin (usually <1m). Exploration efforts for this unit should be geared to finding areas in which thick intervals of porosity are developed.

In summary, the Nordegg interval contains very lucrative, although regionally sporadic, productive horizons. This paper outlines the further potential of what is considered a mature target horizon. With the potential for additional Nordegg interval reservoirs this interval becomes an extremely attractive exploration target. Trapping mechanisms for such reservoirs are very likely to exist, since very good seals are formed by the carbonate sediments of the Nordegg interval, and also by the impermeable shales of the overlying Poker Chip Shale. Source rocks are also not a problem, since the high organic content of largely unoxidized deep marine sediments of the Lower Jurassic are thought to be one of the most prominent source intervals for the entire western Canadian sedimentary basin.

CONCLUSIONS

The Nordegg interval in the Medicine River and Gilby field areas was deposited on a stable, slowly subsiding passive margin which was charaterized by slow rates of terrigenous clastic input. The shelf was dominated by carbonate accumulations, except where fluvially introduced clastics were reworked into clean beach and shoreface sands. The Nordegg interval is composed of two sequences, separated by a surface of subaerial exposure and erosional truncation. In the first sequence, dolomitic shoreline sands prograded onto a carbonate shelf during deposition of the highstand systems tract. The second sequence contains a condensed deposit of oyster shells, which are the nearshore expression of the maximum period of flooding during deposition of the second sequence. These two sequences correlate well with the eustatic sea level curves of Haq et al., (1989).

Slow subsidence and sedimentaion rates allowed for erosional processes to dominate the Nordegg interval, and rarely are appreciable sections of either sequence preserved. The development of the north-south trending Medicine River-Gilby subsidence basin allowed the preservation of thick sections of nearshore sandstones of the lower sequence. This subsidence predated the *Ostrea* bed deposition, and therefore did not allow the lower sequence to be as deeply eroded, at this sequence boundary, as in other areas. Subsidence in this local area appears to be the result of flexure of the rigid Mississippian carbonate rocks over the underlying Leduc reef, during Lower Jurassic time. This flexure developed a fault system which, when loaded by sediments, became an area of active subsidence. Preservation of the sandstone, and therefore development of the prolific Nordegg equivalent reservoirs, is directly in response to subsidence during Lower Jurassic time.

Areas of similar subsidence relationships, along the erosional edge of the Lower Jurassic, would make good prospective targets for future exploration. In addition, the recognition of a sequence boundary within the Nordegg interval opens the possibilities for the discovery of future reservoir sediments in shelf margin- and transgressive- systems tract deposits, basinward of the erosional edge.

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PLATE 2.1 Mississippian Unconformity Surface

2.1a.

Medicine River Sandstone member overlying steeply dipping surface on Mississippian dolomite. Note the steeply inclined bedding in the sandstones conforming to this, which is interpreted as a fault plane (10-06-39-3w5; depth).

2.1b.

Typical lithology of the fractured Mississippian dolomites (D), overlain by the brecciated dolomites in a green shale matrix of the residual zone (R); (06-05-39-3w5).

2.1c.

Contact between the Medicine River sandstone member and Mississippian strata (05-18-39-3w5).

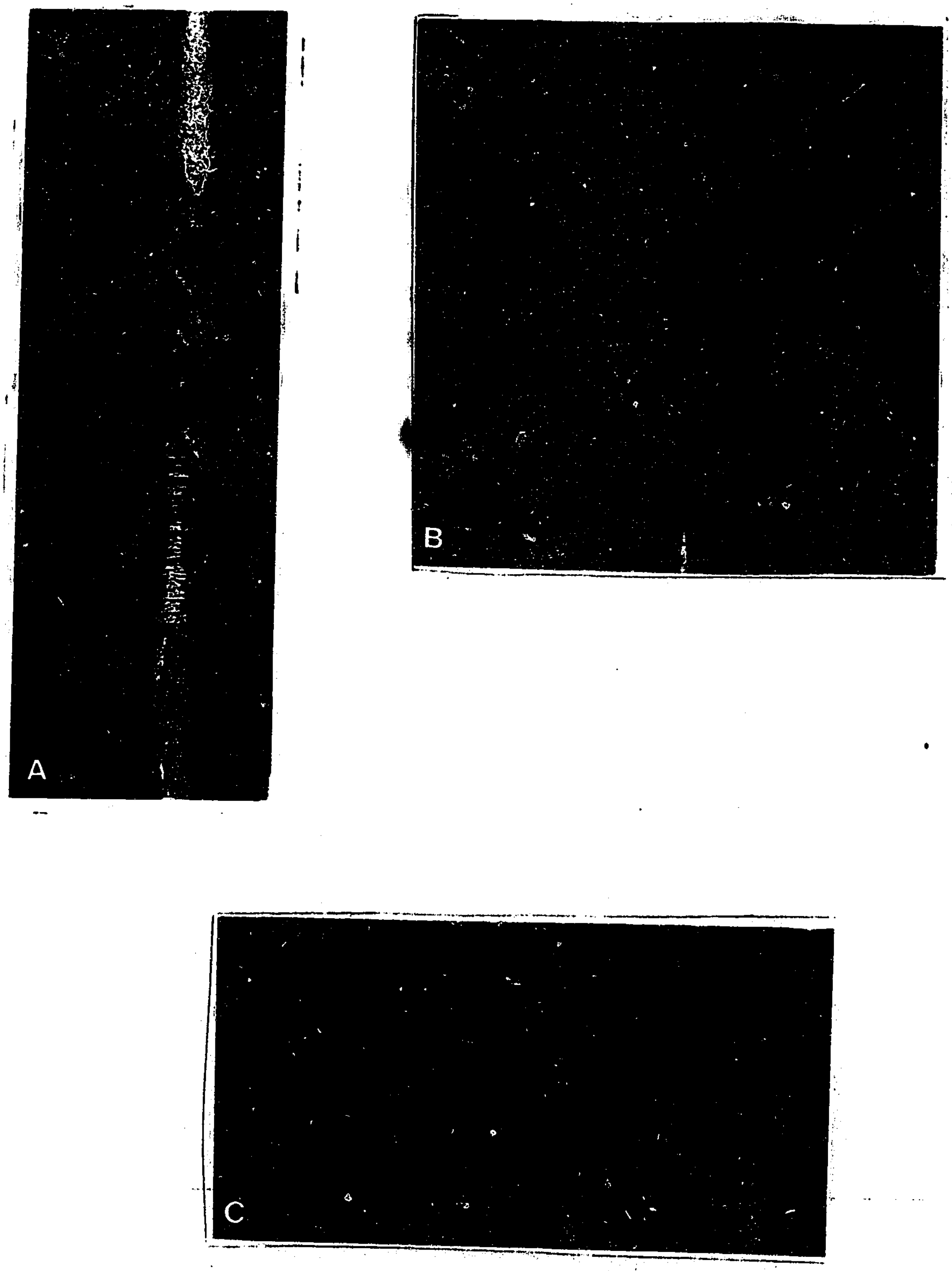

A
B
C

PLATE 2.2 Medicine River Sandstone Member Lithology

2.2a.

Typical structureless texture of the Medicine River sandstone member, note vuggy appearence (02-28-39-3w5).

2.2b.

Slabbed core of the massive sandstone lithology (12-17-39-3w5).

2.2c.

Planar cross-bedded sandstone, bedding enhancement results from preferential oil stain along porous bedding planes (16-24-40-3w5).

2.2d.

Sedimentary structures preserved by chertification of the sandstone. Massive, probably bioturbated structureless sandstone; overlain by a current ripple and trough cross-lamination; low angle planar lamination with low angle discordance surface; and more trough cross lamination (12-17-39-3w5). The escape trace at the top was formed in response to elevated sedimentation rates.

2.2e.

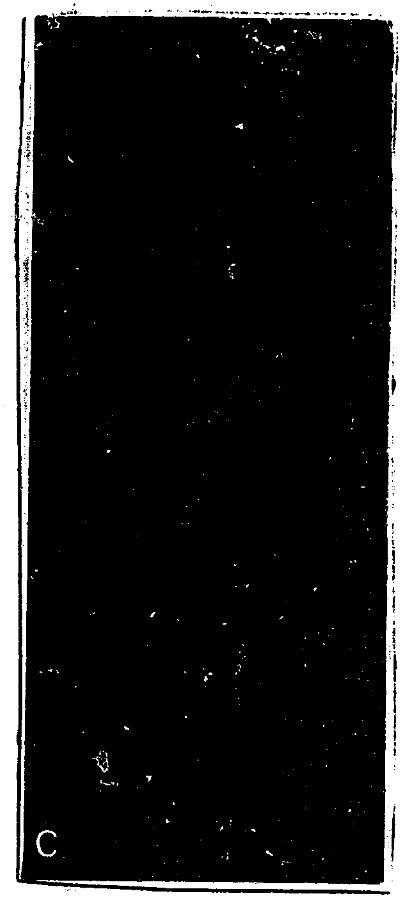
Rare bioturbated shaly sandstone with ripple mud drapes (12-17-39-3w5).

2.2f.

Sharp-based coquinoid sandstone which commonly occurs within the Medicine River sandstone member (16-25-40-3w5).


A


B


C


D
E
F

PLATE 2.3 Calcareous Sandstones

2.3a.

Cross bedding with erosional scour surface in calcareous sandstone (16-24-40-4w5).

2.3b.

Ribbed pelecypods commonly found in the calcareous sandstone lithology; note the massive texture of the surrounding sandstone (16-13-40-4w5).

2.3c.

Cross lamination in the calcareous sandstones; white grains are calcite sand grains (16-13-40-4w5).

2.3d.

Cross bedded skeletal limestones (12-11-39-5w5).

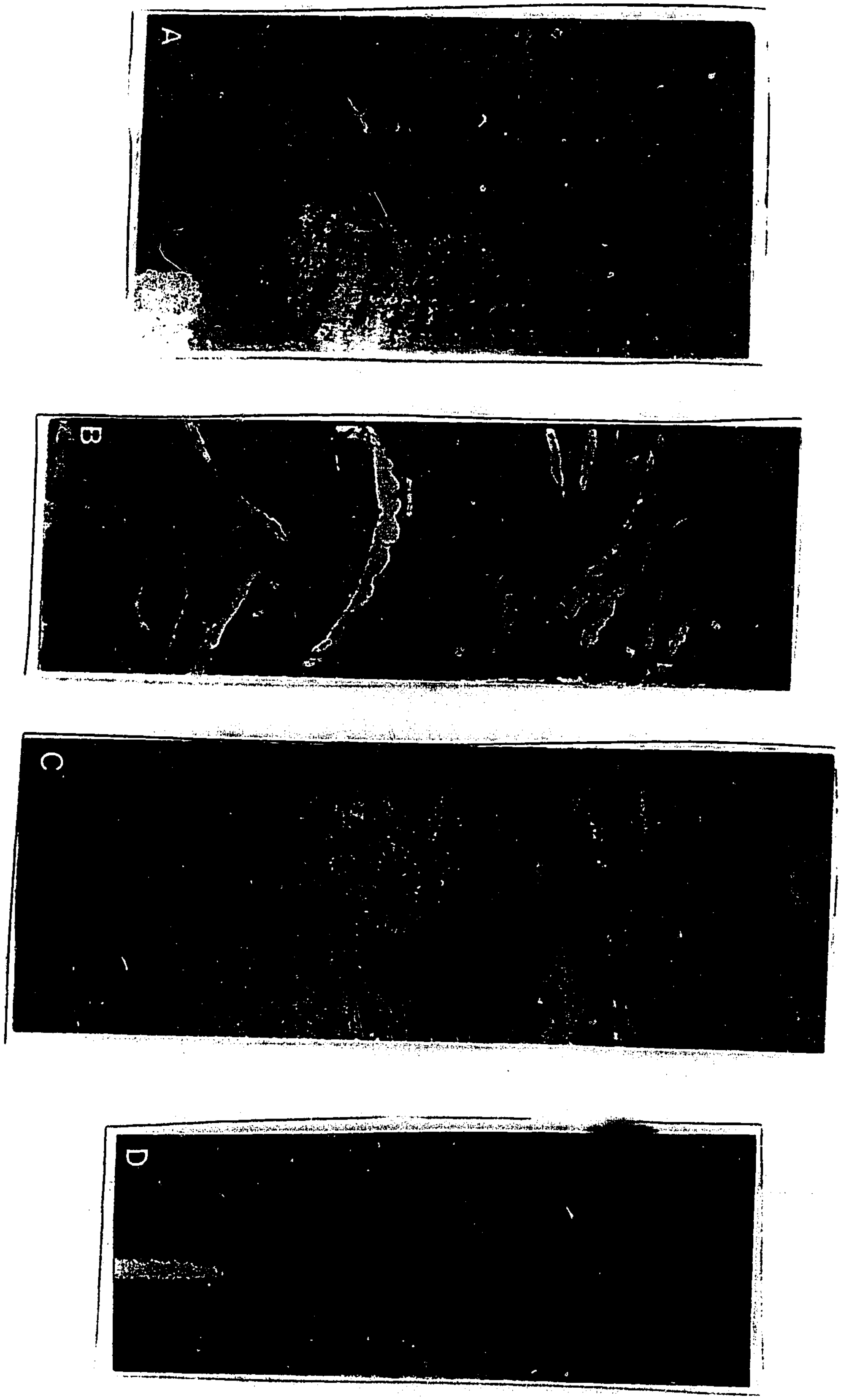

A
B
C
D

PLATE 2.4 Nordegg Interval Sequence Boundary

2.4a.

Top of the Medicine River sandstone member (SS) with dolomite pebbles (D) and laminated shale (Sh) above (04-36-40-4w5).

2.4b.

Laminated sand and shale, alluvial plain sediments (04-36-40-4w5).

2.4c.

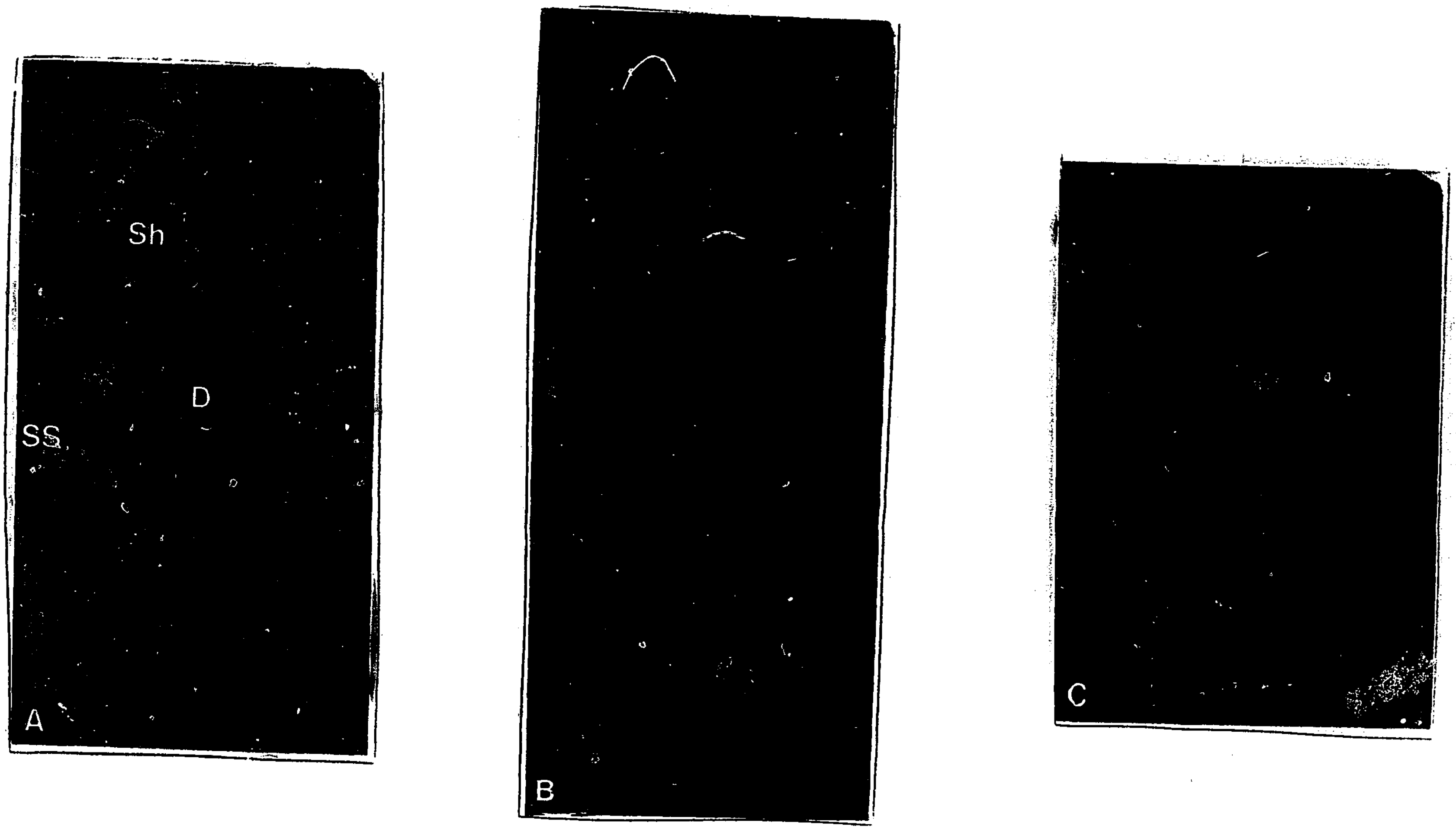
Transgressive surface at the base of the Ostrea bed; note the pebbles at the base (04-36-40-4w5).

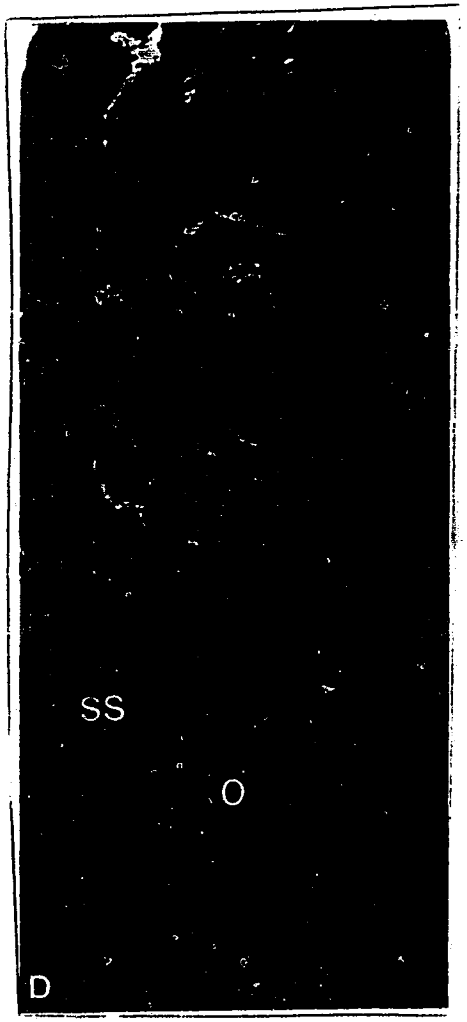
2.4d.

Bored Medicine River sandstone top (SS), with Ostrea bed sediment (O) infilling the boreholes; Ostrea bed above (14-24-40-4w5).

2.4e.

Transgressive base of the Ostrea bed with chert pebbles and large angular, partially chertified pebbles (SS) of the Medicine River sandstone member (06-05-39-3w5).


Sh
D
SS
A
B
C


SS
O
D

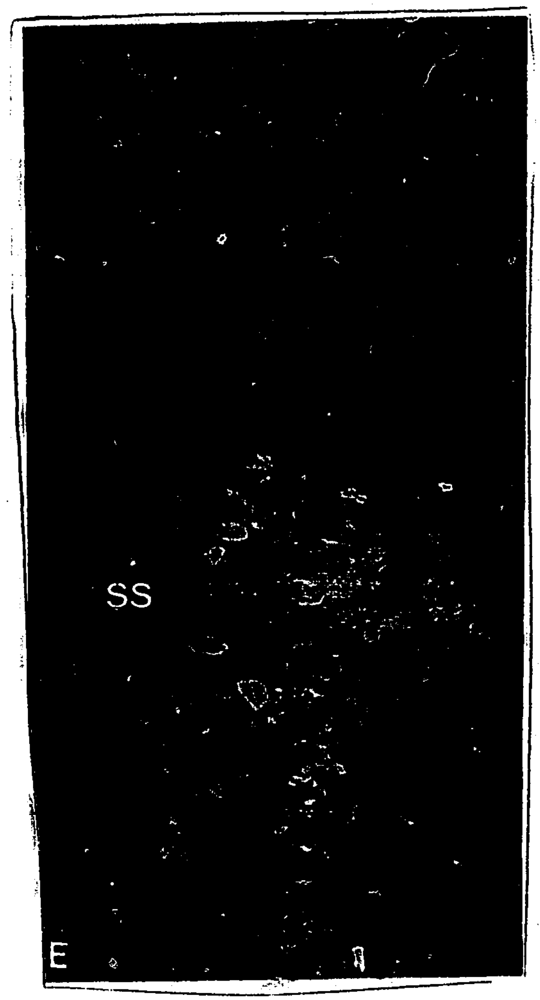

SS
E

PLATE 2.5 Ostrea bed Lithology

2.5(a-c). Large Oysters, some articulated, in a matrix of dark organic-rich limestone (A) or skeletal grainstone (B and C); (12-11-39-5w5).

2.5d. Altered Ostrea bed below the pre-Poker Chip Shale unconformity; large Oysters are still visible (O), but the matrix has undergone dissolution diagenesis and the limestone contains abundant moldic porosity (P); (14-24-40-4w5).

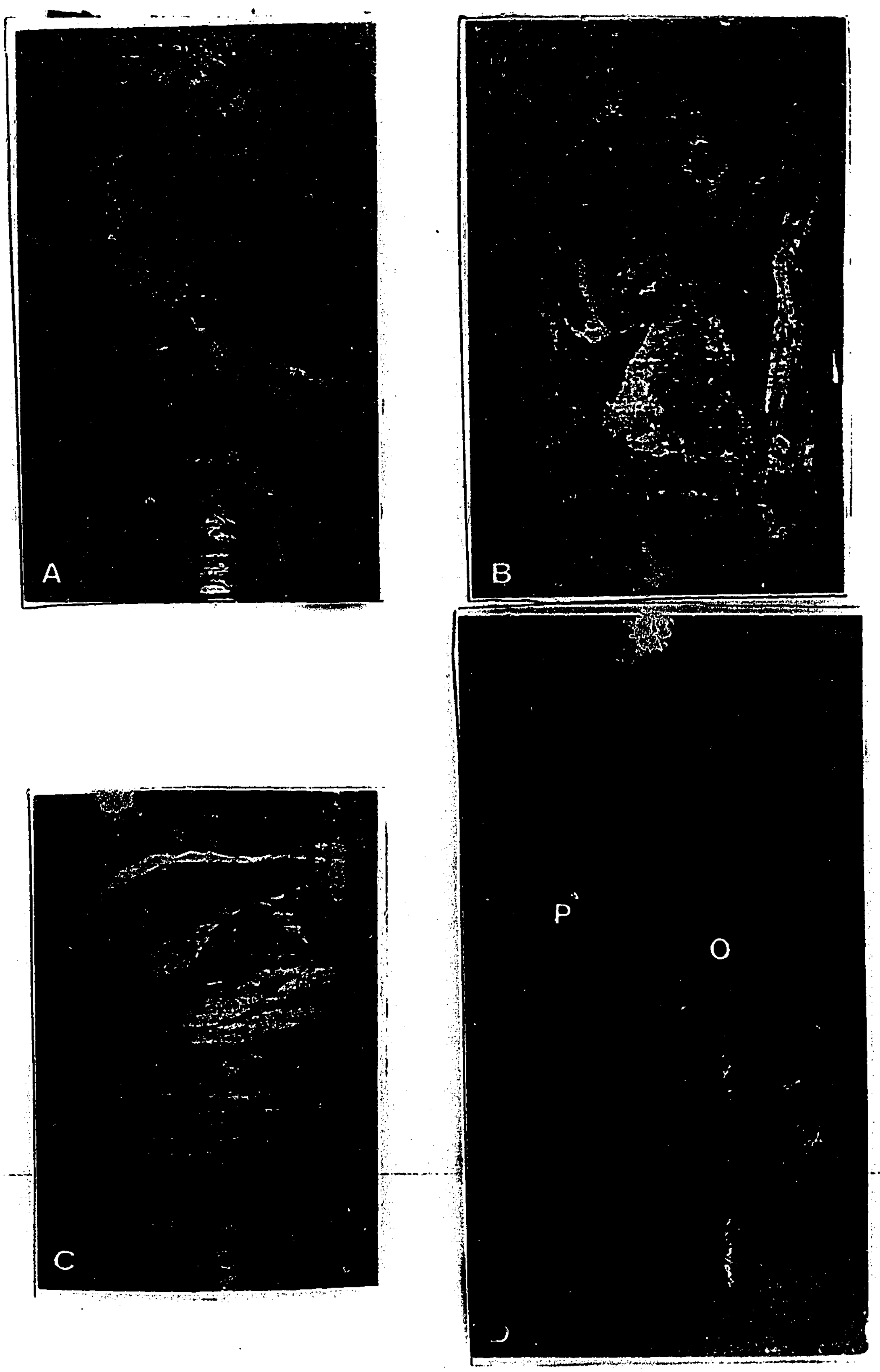

A
B
C
P
O

PLATE 2.6 Pre-Poker Chip Shale Unconformity

2.6a. Poker Chip Shale overlying the Ostrea bed with a pebbly horizon between. Pebbly zone consists of chert and Medicine River sandstone pebbles and green shale cavity fill sediment (06-05-39-3w5).

2.6b. Calcareous shale, with carbonaceous root? casts (R), which lies between the Poker Chip Shale and the Medicine River sandstone member; this deposit is interpreted as a paleosol horizon above the Medicine River sandstone member (10-07-39-3w5).

2.6c. Cavity in the top of the Medicine River sandstone member, filled with green shale; possible karst surface on the dolomitic sandstone (10-20-39-3w5).

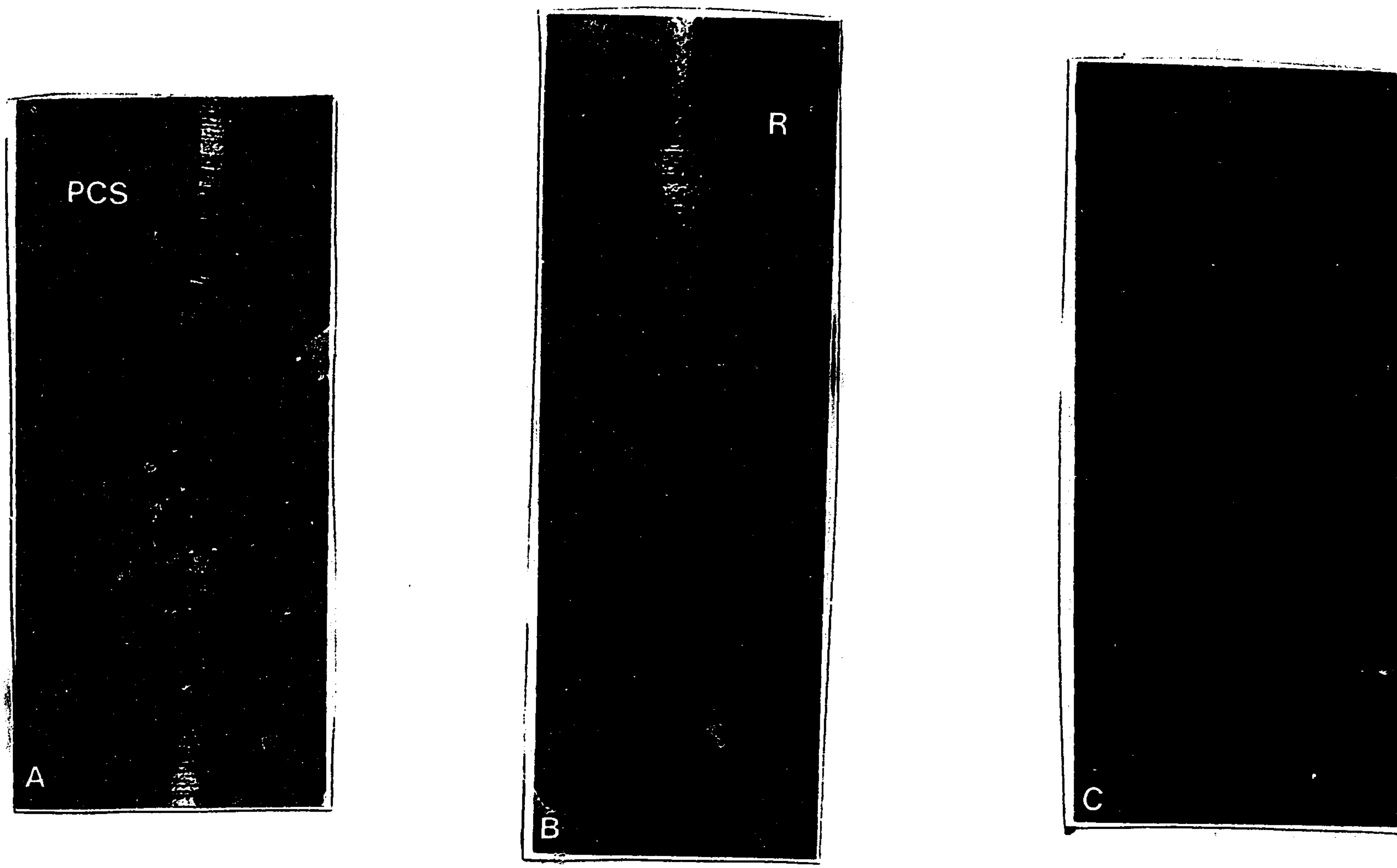
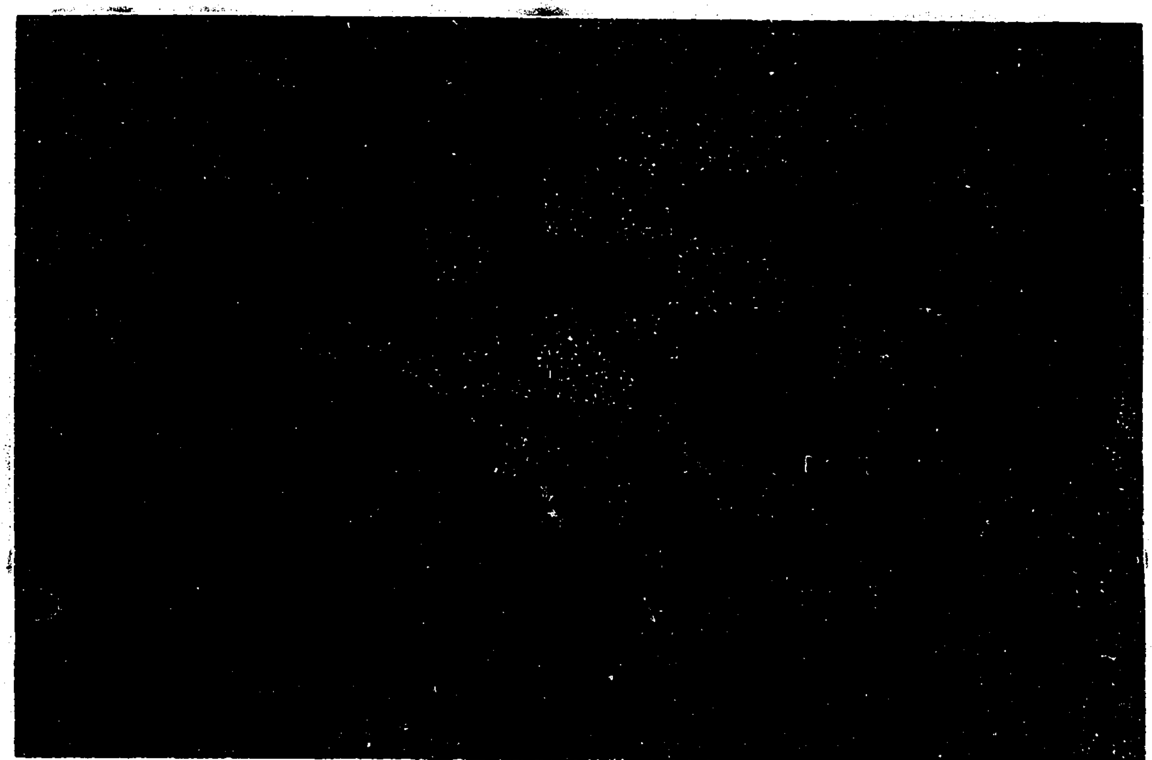

PCS
A
R
B
C

PLATE 2.7 Medicine River Sandstone Member Photomicrograph

2.7a. Porous reservoir sandstone of the Medicine River Member. Note the abundance of detrital phosphate (P), dolomite (D), chert (CT) and quartz (Q). Porosity appears to be secondary in nature and is related to dissolution of pore filing calcite cement (C) and detrital grains. (Sample from 6-20-40-3w5; 7021 ft).

2.7b. Same as above slide with cross-polars.



A

B

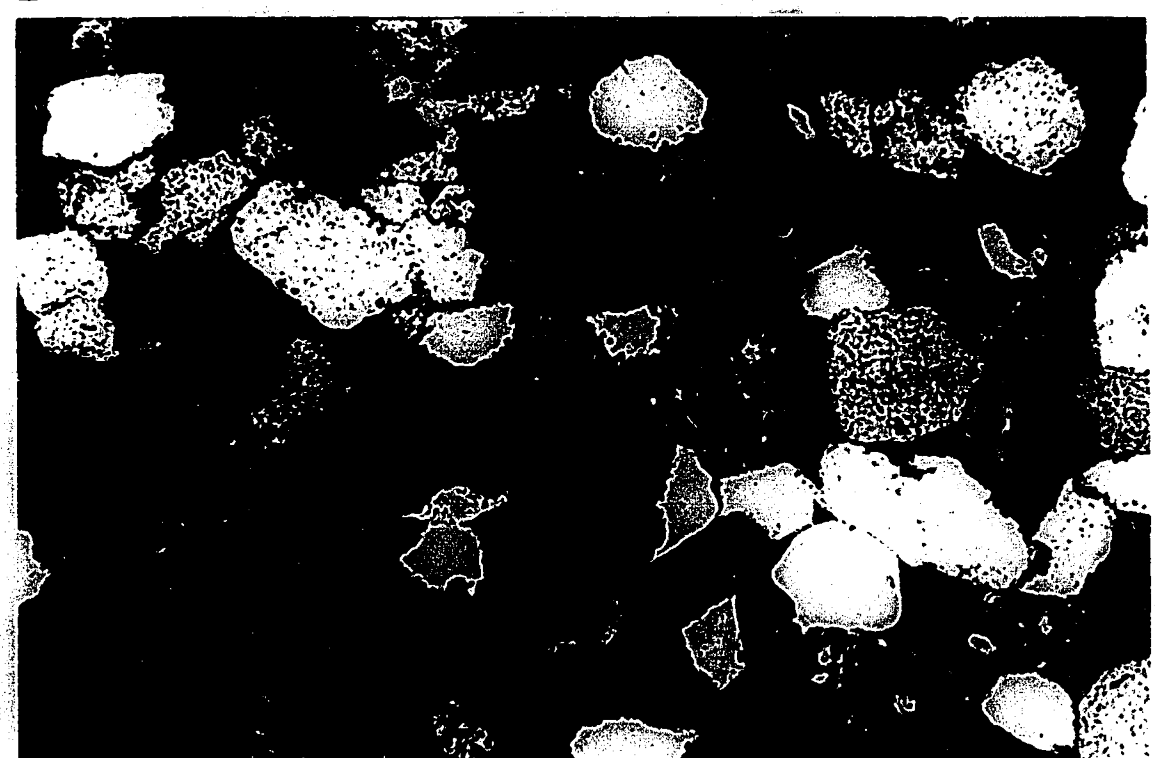


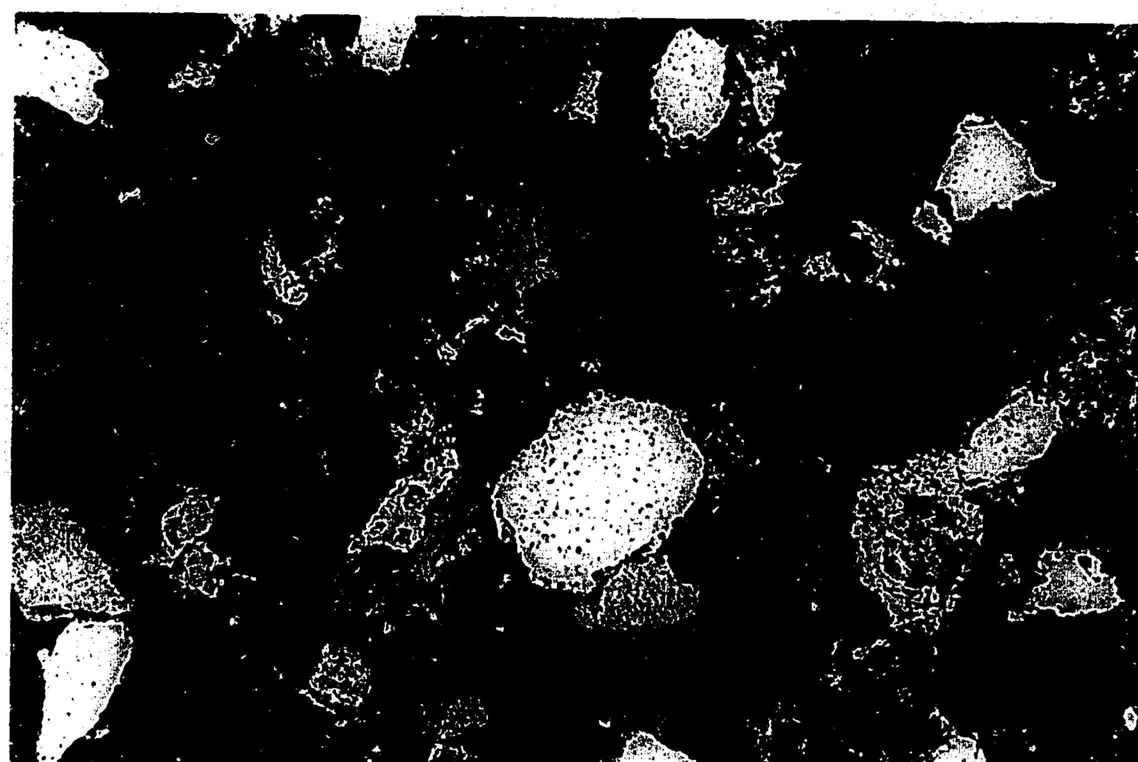
PLATE 2.8 Medicine River Sandstone Member Photomicrograph

2.7a. Porous dolomitic sandstones in which detrital dolomite (D) is the most common grain type. Also present in this sample are abundant detrital phosphate (P) and chert (CT) grains and microcrystalline dolomite cement (Sample from 6-29-39-3w5; 7124 ft).

2.7b. Same slide as above with cross-polars.

A

B



STRATIGRAPHY AND SEDIMENTOLOGY OF THE ROCK CREEK MEMBER IN THE MEDICINE RIVER, GILBY AND SYLVAN LAKE FIELDS; SOUTH-CENTRAL ALBERTA

INTRODUCTION

The study area (Figure 3.1) is located in central Alberta, approximately 50 km west of Red Deer. This area consists of townships 37, 38, 39 and 40 and ranges 3, 4 and 5 west of the fifth meridian and incorporates parts of the Willesden Green, Sylvan Lake, Medicine River and Gilby Fields (Figure 3.2). This area contains prolific hydrocarbon reservoirs at many stratigraphic levels and therefore drilling activity in this area has been intense since the middle 1950's. Well densities range from as many as 14 wells per section, in range 3w5, to as little as 17 wells per township in the western part of the study area. Well density generally decreases from east to west, parallelling the hydrocarbon potential of the area. In all, 887 wells penetrate at least to the depth of the Glauconite member, all of which were examined. Because of the stratigraphic complexity associated with the numerous unconformities in this section, significant stratigraphic variation is observed over short distances, up to several hundred meters. It was therefore essential to examine all wells in each Legal Subdivision.

Jurassic reservoirs were an early target for exploration, an endeavor which was rewarded with the discovery of the Medicine River Jurassic 'A' pool, in 1956 (TerBerg, 1966a). Subsequent discoveries in 1958 of the Medicine River Jurassic 'B' and Gilby Jurassic 'B' pools prompted the extensive development of the Jurassic in this area.

The Sylvan lake gas unit produces gas from the Rock Creek Member in several wells and several others have tested at significant gas flow rates (up to 9MMCF/D) but have yet to be completed in the Rock Creek Member. In addition, there are several oil producing wells in the Rock Creek member

Figure 3.1. Location of the study area in central Alberta.

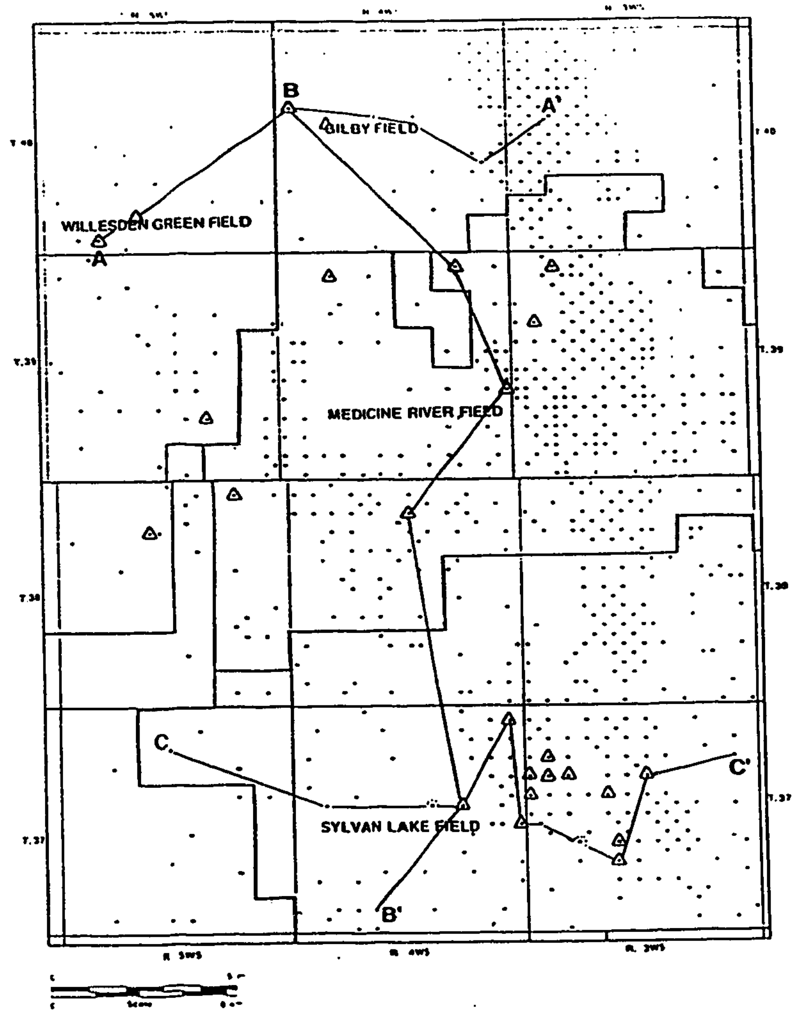


Figure 3.2. Map of the study area showing the wells examined, cored intervals and the regional cross-sections illustrated in this text. Also shown are the various fields within the study area.

of the Sylvan Lake Field. Several wells in the area have averaged a production rate of 8-12 m^3/d. Because of the stratigraphic complexity of this area, these Middle Jurassic reservoirs have been correlated inconsistently as Nordegg sandstones, Ellerslie sediments and even Ostracode deposits. Several significant pools are designated as basal Quartz pools while others are simply labelled Jurassic pools. Rock Creek reservoirs also produce hydrocarbons in several scattered wells in the Medicine River and Gilby Fields, and have significant hydrocarbon shows in other non-productive wells.

REGIONAL STRATIGRAPHY

Before dealing with the detailed stratigraphic relationships in the subsurface of the study area, it is important to understand the regional stratigraphic relationships in the Lower and Middle Jurassic strata. Stratigraphic nomenclature of the Jurassic System is extended from the western outcrop belt, where it was initially developed and extensively described, into the subsurface of west-central Alberta.

LITHOSTRATIGRTAPHY

The term 'Fernie Shale' was first used to distinguish a shaly sequence from overlying Cretaceous sediments in the Blairmore-Frank coal fields of southwestern Alberta (Warren, 1934). These sediments were assigned a Jurassic age with the discovery of the ammonite *Cardioceras*. McLearn (1916) assigned formation status to the Fernie shales, but, in subsequent years different authors have described the Fernie as having both formation and group status. Since this unit contains a number of formally and informally defined 'members' and 'beds' (and no formally defined formations), and since the Fernie has never been formally defined as either a group or a formation, this study will follow the recommendation of Hall (1984) and refer to the Fernie as an informal lithostratigraphic unit of formation status.

Lithostratigraphy and biostratigraphy within the Fernie Formation have been dealt with extensively by a number of authors, but, few formalized definitions of units exist. This has led to a confusion of nomenclature that often mixes lithostratigraphic and biostratigraphic principals, and applies both nomenclature systems to a variety of formal and informal subdivisions of the Fernie Formation. The problems caused by this have recently been reviewed by Hall (1984) and Hall and Stronach (1984) and their recommendation of using strictly lithostratigraphic units, separate from biostratigraphic designations, will be utilized in this study.

The lowermost beds of the Fernie Formation in the outcrop belt consist of the Nordegg Member and its equivalents. A more detailed stratigraphic description of this unit is dealt with in the previous paper and is not discussed further.

The Poker Chip Shale (also called the Paper Shale by Frebold, 1976) is probably the most extensive and uniform unit in the Fernie Formation. It consists of black, fissile to papery shales, and ranges in thickness from 10 to 38 m (Hall, 1984). This unit overlies the Red Deer Member, the Nordegg Member (or its equivalent) or even Triassic or Paleozoic sediments. A major hiatus is therefore indicated at this lower contact.

As defined by Warren (1934) the Rock Creek Member is a marker horizon of calcareous sandstone occurring 15 to 45 m from the base of the Fernie. Although the term Rock Creek has been used to describe all

Bajocian sediments by various authors, recent publications have called for the return of the term Rock Creek to mean precisely what Warren had intended it to mean. At the type section in Rock Creek, the member is a cliff-forming unit 22 m thick, consisting of parallel bedded sandstones, coarsening-upwards and capped by two beds of coarser sandstone with pebbles, small phosphatic pebbles, oxidized pyrite masses and poorly preserved bivalves. Throughout the central region of the outcrop belt, no sandstones or siltstones representative of the Rock Creek Member occur at the appropriate stratigraphic level. In continuous exposures through this interval, the black platy Poker Chip Shale passes upwards into either black oolitic limestones or grey, rusty weathering shales with beds of calcareous concretions and abundant belemnites representing the Highwood Member. Farther north, sandstones again occur in this part of the section, interbedded with grey rusty weathering shales. These sandstones are regarded as being the distal margins of the Rock Creek Member, which here interfinger with the finer offshore shales of the Highwood Member.

Towards the east the sandstones become much thicker (10-20m) and coarser. Thicker nearshore sands contain intense bioturbation, rippled surfaces, low angle planar cross stratification and layers of phosphatic pebbles, shell debris, mud clasts and abundant broken belemnite guards. Thickening and coarsening trends of the sandstone, and the decreasing age of the top of the sandstones towards the east, indicate an easterly source.

The contact between the Toarcian age Poker Chip Shale and the the overlying sandstone and siltstone of the Rock Creek Member is apparently conformable in the outcrop area, with progressive coarsening-upwards or an alternation of dark shales and sandstones. Although no abrupt lithological boundary is apparent in the foothills and mountain areas, biostratigraphy indicates a hiatus during the Aalenian stage between the Toarcian age Poker Chip Shale and Rock Creek Member.

The Highwood member consists of a sequence of dark grey, rusty weathering shales, in some places black, papery and organic-rich. This unit was previously included with the Rock Creek member but has since been separated by Stronach (1984). Also included are the belemnite 'battle fields' of Frebold (1957a) and the "Bajocian Limestone", the latter consisting of black, fossiliferous, oolitic limestone beds. The Highwood Member ranges in thickness from 17m to 57 m and is generally in sharp contact with the underlying Rock Creek sands or the Poker Chip Shale.

The sediments of the Highwood member are succeeded by any of three easily distinguishable and generally spatially separate units (Stronach, 1984). In the Fernie and Crowsnest Pass areas the Grey Beds overly the Highwood member and consist of grey shales which are more or less silty and often highly indurated. Near Blairmore, this unit is replaced by light grey-green or grey-brown shales of the *Corbula Munda* beds which contain bivalve-rich, sandy units. The entire unit is named after the characteristic species of this rich fauna, even though it only occurs in these sandy interbeds. In some areas the *Corbula Munda* Beds are capped by a thicker fossiliferous unit which has been named the *Gryphaea* Bed. Finally, in the Pigeon Creek area this interval is represented by parallel bedded alternations of sand and shale which are characteristic of the Pigeon Creek Member.

Abruptly overlying the *Gryphaea* Beds and the Pigeon Creek Member are dark grey shales with orange weathering sideritic concretions of the Ribbon Creek Member. This unit interfingers with, or is gradationally overlain by, the Green Beds or Passage Beds. The Green beds consist of dark to bright green, often friable, berthierine sandstone and siltstone with irregular interbeds of purplish grey siltstone and yellow brown weathering calcareous concretions (Stronach, 1984).

The uppermost lithological unit of the Fernie Formation consists of the Passage Beds. These sediments generally grade up from dark grey shales, with yellow-brown weathering concretions, to thickening and coarsening-upward silty and sandy bands with a ribboned appearance. As they become sandier upward, they exhibit parallel, rippled and hummocky cross-lamination, sole marks, and bioturbation.

Biostratigraphy

Biostratigraphic zonation of the Jurassic Period around the world is based on the work of Arkell (1956), who developed a western European standard chart based on ammonite fauna. These standard ammonite zones are thought to be essentially chronostratigraphic units and are grouped into a number of stages. The accepted standard stratigraphic chart for the Jurassic period is shown on the left side of Figure 3.3. The difference between this chart (modified from Hallam, 1975 and Hall, 1984) and that of Arkell (1956) is the insertion of the Aalenian stage which replaces Arkell's Lower Bajocian.

The dating and correlation of marine Jurassic formations, in North America, is based mainly on ammonites in the lower and middle part of the Jurassic System and on ammonites and the bivave genus *Buchia* in the Upper Jurassic (Poulton, 1984). The Fernie record is quite incomplete when compared to the western European standard section. Only 12 of the 59 ammonite zones present in Europe were identified by Frebold (1976). Although more ammonite zones are now recognized in the Fernie, there still remains numerous faunal gaps in the formation (Hall, 1984).

The relationship of the faunas collected from the Fernie section to the lithostratigraphic subdivisions of the formation and to the standard zones of Europe are shown in Figure 3.3. Hettangian age (H1) fossils are known only from the Peace River area and the base of the Fernie Formation is usually formed by Sinemurian or Toarcian age sediments. The oldest Sinemurian (S1) fauna are found in the Nordegg Member and correlate to upper part of the Bucklandi Zone and the lower part of the Semicostatum Zone. Obtusium Zone (S2) faunas have been found in the unnamed basal coquina. Raricostatum Zone (S3) fossils have been collected in southeastern British Columbia, an area that has also produced fauna of Pliensbachian age, Jamesoni Zone (P1). Other Pleinsbachian age faunas have been equated with the Red Deer Member, and correlate to the Margaritatus Zone.

Ammonites representing the Falcifer (TO1), Bifrons (TO2) and Variablis (TO3) zones have been collected form the Poker Chip Shale throughout the outcrop area. These zones represent the majority of the Toarcian period and other Toarcian Zones are thought to be present but have yet to be proved.

No Aalenian fauna have been found in Fernie outcrops, although a continuous gradational section appears to occur between the Toarcian and Lower Bajocian sediments. It is suggested that the Aalenian stage may lie within an unfossiliferous unit of shales and sands below the Rock Creek Member (Frebold, 1976).

Bajocian age fauna are represented in the Rock Creek Member, the Bajocian Limestone and the Highwood Member. The oldest known Bajocian Fauna (BJ1), as identified with the occurrence of the ammonite *Sonninia*

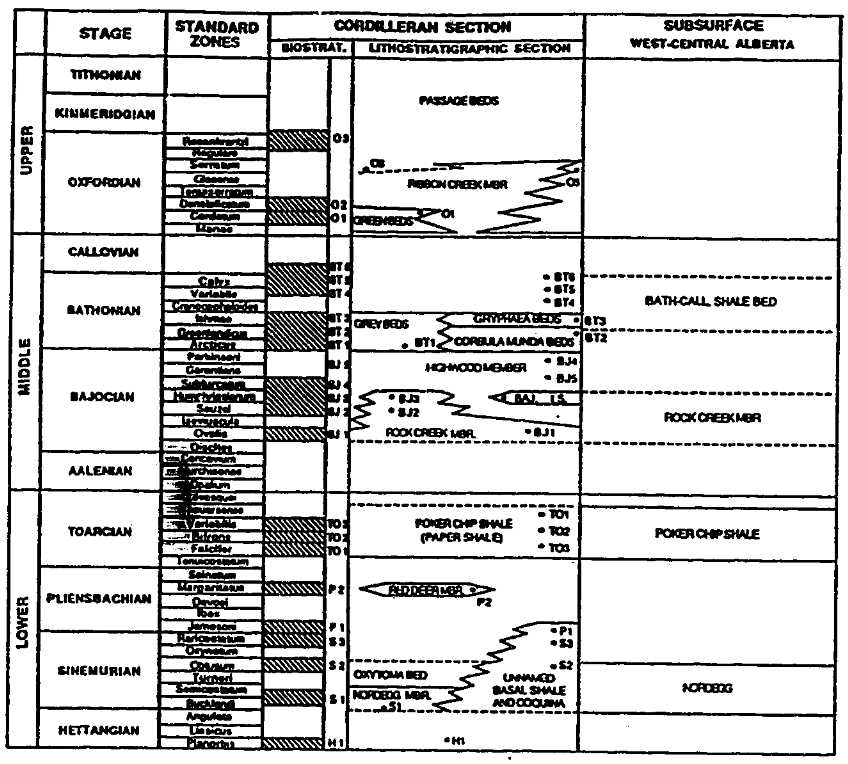


Figure 3.3. Stratigraphic chart of the Jurassic System in Alberta. Cordilleran section is modified from Hall (1984) and the subsurface section is modified from Marion (1984).

gracilis, is found in the Rock Creek Member and correlates to the Ovalis Zone. The Rock Creek Member also yields younger Sauzei Zone (BJ2) and Humphreisianum Zone (BJ3) fauna, the latter equating to the Bajocian Limestone. Upper Bajocian, Subfurcatum Zone (BJ4) and possibly even younger fauna have been retrieved from the Highwood Member.

Sparse and poorly preserved Bathonian fauna, which are difficult to correlate to the European Standard Zones, indicate the presence of most of the Bathonian zones. The Grey Beds, *Corbula Munda* Beds, and *Gryphaea* Beds all contain Bathonian age fauna (BT1-3), and shales overlying the *Gryphaea* bed contain younger Bathonian fauna.The Callovian stage appears to be missing in the Rock Mountain section. Oxfordian age fauna have been recovered from the sediments of the upper part of the Green Beds, in shales overlying the Green Beds and from the lower part of the Passage Beds.

Biostratigraphic and lithologic breaks indicate a number of unconformities in the Fernie Formation. A major hiatus is indicated at the base of the Fernie Formation, where in all but a few areas in British Columbia, all of the Hettangian is missing. A second major hiatus occurs at the base of the Toarcian Poker Chip Shale. This erosional gap generally spans the lowermost Zone of the Toarcian and the majority of the Middle and Upper Sinemurian. Locally, erosion during this time even removed Sinemurian sediments and the Poker Chip Shale forms the lowest unit of the Fernie Formation. A faunal gap appears between the Poker Chip Shale and the Bajocian sediments of the Rock Creek or the Highwood Members. However, it is not known whether this gap represents a true hiatus or just an unfossiliferous, undated zone. There appears to a continuous, lithologically conformable section, between these units in the Foothills region.

Rock Creek Member in the Subsurface of Alberta

Published reports of the Jurassic System in the subsurface of west-central Alberta are relatively sparse and few recent publications exist. The extensive outcrop work on the Jurassic sections of the Cordillera were first extended into the subsurface by Spivak (1949). Based on lithologic characteristics and stratigraphic position, he identified a Fernie and a Kootenay section. Spivak (1949) further subdivided the Fernie into a basal Nordegg-equivalent unit, the Poker Chip Shale, a unit of calcareous shales with thin sandstones and a 'Belemnite conglomerate' (probably equivalent to the Highwood Member), and an uppermost unit of green glauconitic, fine grained quarztose sandstones (probably correlative to the Green Beds of the outcrop area).

Lackie (1958) described the Jurassic subsurface of the Peace River area in northwestern Alberta and northeastern British Columbia. Correlation to Frebold's (1976) foothills section is based on similar lithologic characteristics, stratigraphic position and facies changes. Overlying the Nordegg-equivalent unit are grey-brown to black, fissile to papery shales with abundant pyrite, which were identified as the Poker Chip Shale. A series of blocky, micaceous shales and quartzose, well sorted sandstones, with abundant nodular and cementing iron oxides, occur above the Poker Chip Shale. This unit becomes sandier towards the east. Lackie (1958) labelled this unit as a Rock Creek equivalent, but, by the terminology of Hall (1984) this probably correlates to the Highwood Member in the west and becomes the Rock Creek Member in the sandier eastern section. Lackie (1958) indicates that microfossils of Bajocian age were collected from this unit, supporting these correlations. However, the age of these fossils was given with some uncertainty.

Deere and Bayliss (1969) described the Poker Chip Shale in central Alberta as a medium grey, non-calcareous and fissile shale with a mineralogical composition dominated by quartz and illite, with lesser amounts of kaolinite and scattered pyrite. Springer et al., (1964) indicated a Toarcian age for this unit, based on an apparent Toarcian faunal assemblage collected from the shale in the Gilby area. Rall (1980) confirmed this age by assigning a Lower Jurassic, and probable Toarcian age, based on the occurrence of the dinoflagellate cyst *Nannoceratopsis gracilis senex.*

The Rock Creek Member was shown to overlie the Poker Chip Shale in the westernmost part of Deere and Bayliss's (1969) area, but their schematic cross-section would indicate pre-Cretaceous erosion of the Rock Creek across the present study area. In a summary of the Sylvan Lake Elkton C pool, however, Sherwin and Hawryszko (1966) indicated the presence of Rock Creek Member sandstone reservoirs capping the Elkton oil pool.

More recently, Marion (1982, 1984) published work on the Rock Creek Member in the subsurface of central Alberta, approximately 60 miles to the northeast of the present study area. Marion described the Rock Creek Member as predominately fine grained quartz arenite sandstones, with lesser amounts of burrowed silty sandstones, coquinoid sandstones and a laminated green siltstone unit. A Bajocian age was assigned to these sediments based on the discovery of a single juvenile ammonite *Sonninia* sp., therefore correlating this unit to similar sediments in the outcrop belt (Marion, 1984). These sediments are very similar in both age and lithology to the Rock Creek Member in the type section, near Rock Creek, Alberta (Warren, 1934; Hall, 1984). The Rock Creek Member generally lies in contact with the pre-Cretaceous unconformity and therefore difficulty arises in the differentiation of Jurassic and Ellerslie sandstone units. Marion (1984) reported that subtle mineralogical differentiation of Rock Creek sandstones and overlying Ellerslie sandstones is possible based on detrital chert content; Rock Creek sandstones contain no more than 1.5% chert, whereas Ellerslie sandstones contain greater than 3.5% chert. In a more regional study, Rosenthal (personal communication) confirmed this mineralogical differentiation, indicating that the Rock Creek Member sandstones generally contained greater than 97% fine grained quartz. Although the Ellerslie sandstones generally contain more chert, this distinction becomes very difficult in the finer grained sandstones of these units(Rosenthal, personal communication).

Marion (1984) also discussed the presence of an upper Jurassic Shale unit overlying the Rock Creek Member, in places where it has not been removed by Lower Cretaceous erosion.

OBJECTIVES

The objectives of this research are threefold. The first is to unravel the stratigraphic complexities of the Jurassic, in particular the Rock Creek Member, in an area that is both stratigraphically complex and economically productive. It has been previously established that Jurassic sandstones and Lowermost Cretaceous sandstones are petrographically distinguishable (Rall, 1980; Marion, 1982; Rosenthal, personal communication). Mineralogy and micropaleontology are used in conjunction with geophysical log correlation to achieve this objective.

The second aim of this work is to develop a depositional model for the Rock Creek Member in this area. Combined with the stratigraphic framework provided herein, this model will allow explorationists to better understand the depositional trends of the Rock Creek Member and their control on the hydrocarbon productivity of the unit.

The final purpose of this study is to examine the hydrocarbon distribution within the Rock Creek Member and establish the controlling factors of reservoir development.

METHODS

As this is a prolific hydrocarbon producing area the well density penetrating the Jurassic strata is very high. In all, 887 geophysical logs were examined and correlated. Stratigraphic picks at the tops of the Mississippian strata, the Jurassic System (and each of its members), and the 'glauconite coal marker horizon' were recorded. The top of this latter marker bed was generally used as a datum, since it is the most laterally extensive marker in the area. Numerous stratigraphic cross-sections were constructed in a grid that covered the study area. The locations of the three examples in this paper are illustrated in Figure 3.1.

Twenty-four cored intervals were also examined to provide lithological support for the correlations and data to develop a depositional model for the Rock Creek Member. Seventy-six sandstone samples were collected and examined in thin section, lending further support for the lithostratigraphic separation of Ellerslie and Jurassic sandstones. In addition twenty-nine shale samples from Poker Chip Shale, Rock Creek and Ellerslie deposits were collected for micropaleontologic investigation.

The compilation of this data provided a good stratigraphic framework within which the nature of the Rock Creek Member and associated deposits could be described. The results of this data will first be outlined and then a discussion of the stratigraphic and depositional character of the Rock Creek Member will ensue.

RESULTS

PETROGRAPHY

Ellerslie Member

Twenty-four thin sections of Ellerslie sandstones, from twelve wells, were examined. Terrigenous framework grains were grouped into 1. monocrystalline quartz, 2. rock fragments, and 3. chert. The relative percent of each of these groups was normalized to 100 percent and plotted on the ternary diagram in Figure 3.4, along with the Jurassic samples for comparison. Ellerslie sandstone framework grain content varied a great deal, averaging 69.5 relative percent quartz (31-98% range), 10.2 relative percent rock fragments (0-21% range), and 20.4 relative percent chert (0-61% range).

Sixteen of the twenty-four samples contained a matrix material consisting of detrital clay and quartz silt (Plate 3.15), ranging in abundance from 0 to 58% of the rock volume. Three of these samples contained over 40% of the fine grained matrix material, and each of these samples contained a higher than average quartz content (>85%). Therefore, it appears that the quartz content of the Ellerslie sandstone is in part dependent on the overall grainsize and texture of the sediment.

Rock fragments averaged 6 percent of the total rock volume of the Ellerslie samples examined. Of these, the average composition was 28 % (0-75% range) metamorphic grains, 32% (0-71% range) sedimentary

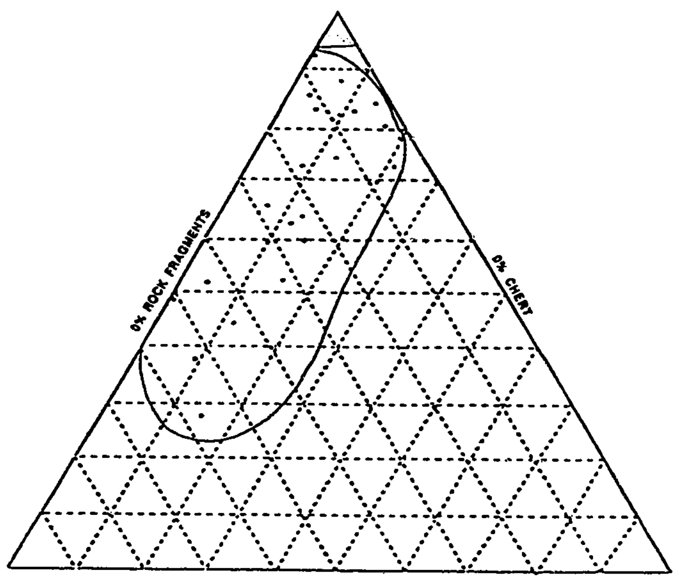


Figure 3.4. Ternary diagram of the framework mineralogy for the Rock Creek and Ellerslie member sandstones. The stippled triangle at the top indicates the field in which all Rock Creek member sandstones fall while the circled area indicates the Ellerslie Member field.

fragments and 42 % (0-71% range) polycrystalline quartz. The polycrystalline quartz group consisted of finely crystalline quartz, but it was unclear whether these were igneous in origin or whether they were fine grained highly cemented quartzites derived from the underlying Rock Creek sediments. The sedimentary rock fragments consisted of shale, siltstone and sandstone fragments. Very rare grains of volcanic origin were observed, but, these were generally highly degraded and difficult to recognize.

Other detrital grains were observed in trace amounts, the most common of which were mica flakes. Plagioclase feldspar and zircon grains were rare and pyrite was a common cementing and replacing mineral, forming up to 6% of the rock volume. In general, however, it was found at levels < 1%. Carbonaceous material and pyrobitumen were also common.

Rock Creek Member

Fifty sandstone samples of the Rock Creek Member, from eleven wells, were point counted. Several sandy limestone and limestone samples were also examined from this interval but will be discussed separately. The distinction between limestone and sandstone was arbitrarily set at approximately 50% carbonate material.

The terrigenous framework grains were grouped into: 1) monocrystalline quartz, 2) rock fragments, and 3) chert. The relative percentages of these groups (normalized to 100%) are plotted on the ternary graph in Figure 3.4 along with the Ellerslie samples for comparison. The Rock Creek Member sandstone framework grains show little variation in there mineralogy. Quartz ranges from 97.1 to 100 relative percent (average 99.2%), chert from 0-1.2 relative percent (average 0.15%) and rock fragments from 0-2.2 relative percent (average 0.68%). The rock fragment content is made up of predominately metamorphic grains (44%) and polycrystalline quartz fragments (39%). The remainder of this group (17%) consisted of sedimentary fragments; predominately shale grains.

Zircons occurred in trace amounts and were observed in three samples. Phosphate grains were common, particularly in samples collected from wells in the western part of the study area (T. 39-40; R. 4w5-5w5). Phosphate grains were generally associated with sandstones proximal to and transitional with carbonate sediments that form the westernmost part of the Rock Creek Member. In some cases, phosphate grains consist of aggregates of quartz silt and phosphate (Plate 3.11b) and are interpreted to be fecal material. Phosphate also occurs as thin coatings on sandsize quartz grains (Plate 3.12a) and these appear to be similar in origin to superficial ooids. In this case the phosphate probably replaced oolitic calcite coatings. Other phosphatic grains appear as massive rounded phosphate and are believed to be detrital phosphate grains, perhaps eroded from the underlying phosphate rich Nordegg member (Deere and Bayliss, 1969; Bovell, 1979). Feldspar grains were present, but were only rarely observed.

Thick successions of clean quartzose sandstones are often interrupted by thin intervals of highly mottled sandstone with green, randomly oriented shale streaks or streaks of white, apparently leached sandstone (Plate 3.3d.). These features bear no resemblance to known biogenic or physical sedimentary structures, and are difficult to interpret with the unaided eye. Such zones are particularly common in the Sylvan Lake area, where they separate series of stacked quartzose sandstones.

In thin section much of the mottling appears to be infill of porosity and previously open fractures by internal sedimentation or illuviation from above (Plate 3.13; 3.14b). Some pore spaces are filled with crescentically laminated clay and fine particles of quartz silt (Plate 3.14a). Similar features were described by Percival (1986) in Upper Carboniferous quartzites in England. These were interpreted to be the products of pedogenic processes, in particular illuviation of leached material. In this example, the laminated pore linings were formed by downward percolation of water, in sediments which lay above the water table. This water transported materials from the weathered zone above, and deposited these materials in the sediments below. Such a structure indicates the proximity of this sediment to a surface of active soil formation and therefore the surface of subaerial exposure.

Tangentially oriented clay linings of detrital quartz grains were also found in these mottled Rock Creek sandstone intervals. These clay linings are composed of a straw yellow birefringent clay, suggesting an illite or smectite composition. Percival (1986) described similar features, formed by mixed layer illite-smectite clays, in the sandstones of the Upper Carboniferous deposits discussed above. Again, these features are interpreted as the result of pedogenic processes. Plate 3.9c shows a fracture that is filled with a laminated clay material and oriented quartz silt. This feature indicates internal sedimentation which took place in open fracture systems, a result of downward percolating water.

The green shale and white, apparently leached, mottled horizons found within the quartz sandstone are therefore interpreted as the result of pedogenic processes. This suggests that these sediments were proximal to a surface of subaerial exposure and soil formation, indicating a possible unconformity. Open fractures, which were subsequently filled with illuviated material, suggest that at the time of soil formation the underlying sediment was relatively well lithified. The mottled horizons generally show up well on geophysical logs, appearing as a relatively high gamma, low porosity and permeability and high density streaks. In addition, they usually contain abundant pyrite and therefore appear as very low resistivity beds.

In the Rock Creek Member there is a complete range from non-calcareous, quartzose sandstones to calcareous sandstones (10-50% carbonate) and limestones (>50% carbonate). This transition generally occurs vertically downward in section and laterally westwards acoss this area.

Slightly calcareous sandstones generally contain small amounts of calcite spar cement that partially fills the primary pore spaces between the framework grains (Plate 3.12a). Occasionally, rare pelecypod or echinoderm fragments (Plate 3.11a) were observed in these sediments, but the majority of the calcite is diagenetic in origin, probably derived from the dissolution of more calcareous units nearby. More calcareous sandstones occur interbedded with the clean sandstones and are gradational with them. The carbonate content consists of pelecypod fragments, echinoderm debris, microcrystalline calcite mud, coarse calcite spar and, more rarely, ooids and pellets (Plate 3.12b). Sand size quartz grains vary in abundance from 40 to 63% of the rock volume.

Limestones in the Rock Creek Member are generally dominated by broken pelecypod and echinoid fragments and by calcite spar. In one sample, however, approximately 18.5% of the rock is made up of microcrystalline calcite mud. Quartz sand content in the samples examined varies from 17 to 36.5%. Phosphate grains as described above are very common in the calcareous sandstones and sandy limestones of the Rock Creek Member.

Petrography Summary

In summary, it is possible to separate Ellerslie Member sandstones and Rock Creek Member sandstones on the basis of subtle differences in framework grain mineralogy (Figure 3.4). Rock Creek sandstones plot in the greater than 97% quartz area (hatched), while the Ellerslie Member sandstones plot in a wide area (outlined) but never in the hatched zone. In addition, the presence of phosphate grains, echinoid fragments and pelecypod debris associated with the Rock Creek sediments indicates the fully marine conditions responsible for deposition of this unit. Phosphatization of ooids and fecal pellets indicates upwelling currents associated with open marine shelf conditions (Bovell, 1979). Ellerslie sediments contain little or no indication of marine deposition and are generally considered to have been deposited in fluvial, or possibly, slightly brackish settings.

MICROPALEONTOLOGY

Because of the complex nature of the stratigraphic relationships between the Jurassic and Lower Mannville strata, shale samples were collected from most of the observed core for micropaleontological analysis to support correlations made on the basis of lithologic, geophysical and stratigraphic characteristics. Of these, thirty-eight samples were selected and sent to the Bujak Davies Group for analysis. The samples were processed by Eric Young at the laboratories of the Bujak

Davies Group and palynological and kerogen analysis were performed by Dr. E.H. Davies.

The primary objective of this work was to determine the age of the various units and establish correlation within the study area and to the outcrop belt in western Alberta and eastern British Columbia. Particular emphasis was placed on correlation of the Rock Creek Member and the Poker Chip Shale unit, of Middle Bajocian and Toarcian age respectively in the outcrop area. Ages are assigned according to concurrent range of the marker species of dinoflagellates. As there is no standard section to which palynological assemblages can be related, there is a certain degree of uncertainty in correlation of these assemblages to the ammonite based divisions of the outcrop belt. Dr. Davies, however, expressed a good degree of confidence in his age assignments and they will be assumed correct in this study.

Some samples, usually highly oxidized and leached shales from the detrital member, yielded an indefinite or barren assemblage. In shales that were better preserved five main assemblages were recognized. Age assignments of the samples are shown in Table 3.1 and the marker species identified and the ages they were assigned are shown in Table 3.2. Assemblage 'A' is dominated by *Nannoceratopsis senex* var. A and is considered Lower Jurassic, Toarcian in age. Assemblage 'B' is characterized by *Bradleyella n.sp.*, *Gonglyodinium hoceratum* and *Dapcodinium semiabulatum*. This assemblage is considered lower Middle Jurassic- Aalenian to Bajocian in age. The presence of *Comparodinium* spp. and *Dichadogonyaulax norrisii* is used to distinguish subassemblages 'B1' and 'B2' respectively. B1 is thought to represent an Aalenian age, while B2 a Bajocian age. Assemblage 'C' is defined by a diverse assemblage of

LOCATION	SAMPLE	DEPTH	AGE
06-05-40-5W5	S100	2365m	BAJ.
14-04-40-5W5	S033	2236m	BATH.-E. CALLOV.
14-19-40-4W5	S102	2346m	AALEN.-BATH.
	S037	2355m	BAJ.
10-35-39-4W5	S038	7327ft	BAJ.
08-13-39-4w5	S040	7040ft	AALEN.-BAJ.
	S039	7054ft	AALEN
16-31-39-3w5	S044	7060ft	MID-LATE JUR.
04-34-38-4w5	S036	2285m	INDETERMINATE
	S035	2286m	TOARCIAN
	S034	2289m	INDETERMINATE
12-11-39-5W5	S031	7840ft	PROB BAJ.
11-28-38-5W5	S101		BAJ.-BATH.
	S029	2428ft	PROB. BAJ.
07-22-37-4W5	S026	7622ft	PROB. MID JUR.
	S106	7625ft	?E. CRET.
	S104	7665ft	TEN. TOARCIAN
10-36-37-4W5	S023	7360ft	AALEN.
	S022	7380ft	TOARCIAN
16-13-37-4W5	S024	2302m	BAJ.
	S025	2314.5m	PROB. BAJ.
10-30-37-3W5	S007	7358ft	INDETERMINATE
04-29-37-3W5	S041	7325ft	TOARCIAN
	S042	7320ft	JURASSIC
04-27-37-3W5	S003	2180m	BATH.-E. CALL.
	S001	2195m	TOARCIAN
14-09-37-3W5	S005	2301m	JURASSIC
04-27-36-3W5	S010	7162ft	NEOCOMIAN
10-18-40-2W5	S105	7064ft	APTIAN-ALBIAN

Table 3.1. Table of the shale samples and the ages assigned to them on the basis of micropaleontology.

LATE BATH. TO EARLY CALL.	Clenidodinium pachydermum Clenidodinium sellwoodii Eodinia pachytheca Gonyaulacysta jurassica Gonyaulacysta pectinigera Occisucysta thulia Occisucysta thulia bf. Sentusidinium asymmetrica Sirmiodinium gressii Sentusidinium erythrocumum
BAJOCIAN	Facetodinium #EA Gonglyodinium spp. lithdinium serrulata cf. Mendicodinium #EHyalina Nannoceratopsis gracilis Bradleyella #EA Dichadogonyaulax norrisii Gonglyodinium hocneratum Sentusidinium asymmetrica Valensiella vermiculata Nannoceratopsis senex var. A
AALENIAN	Bradleyella #EA Comparodinium aquilonium Dapcodinium semitabulatum Eyachia prisca Gonglyodinium hocneratum Lithodinium serrulata Sentusidinium asymmetrica Menocodinium #EHyalina
TOARCIAN	Nannoceratopsis senex. var C Dapcodinium semitabulatum Sentusidinium asymmetrica Comparodinium perpunctatum Mendicodinium #EHyalina Nannocertosis #Estega

Table 3.2. Table of the various marker species, and their corresponding age range, used to assign ages to the shale samples (from the Bujak Davies Group of Calgary, Alberta).

dinoflagellates which contains *Ctenidodinium sellwoodi, Gonyaulacysta jurassica* and *Sirmiodinium grossii.* This assemblage is considered to be upper Middle Jurassic, Bathonian to Early Callovian in age. Assemblage 'D' is somewhat uncertain, but close pollen affinities with Assemblage C and occurrence of such dinoflagellates as *Lithodinia jurassica*, indicate a probable Middle Jurassic age. Finally, assemblage 'E' is defined on the basis of spores such as *Cicatricosisporites* spp. and *Aequitriradites* spp. and abundant gymnospermous pollen. This assemblage is early Cretaceous in age and can be subdivided into a Neocomian age 'E1' assemblage, characterized by the presence of the miospores *Cicatricosisporites exiloides* sensu, *Corollina vignollensis*, *Tuberosittriletes montuosus*, and *Trilobosporites* spp., and an Aptian to early Albian age 'E2' assemblage, defined by the presence of the dinoflagellates *Ctenidodinium n. sp.* and *Vesperopsis mayi*, and the miospores *Januasporites spiniferous* and *Plicatella pseudomacrorhyza*. Figure 3.5 illustrates the age relationship of these various assemblages.

Table 3.1 shows the wells and depths of the samples which Dr. Davies analyzed. The fourth column in this table shows the ages assigned to the samples. There are essentially three groups of ages associated with Jurassic rocks. The oldest are Lower Jurassic-Toarcian in age and were collected from the Poker Chip shale. The second set were collected from shaly or calcareous zones in the Rock Creek Member and are Middle Jurassic Aalenian to Bajocian in age. The final group of samples was collected from the shale which unconformably overlies the Rock Creek sediments. These samples are Bathonian to Early Callovian in age. This latter group confirms the presence of an Upper Fernie Shale, described by Marion (1982), but a Middle Jurassic age is indicated for this unit.

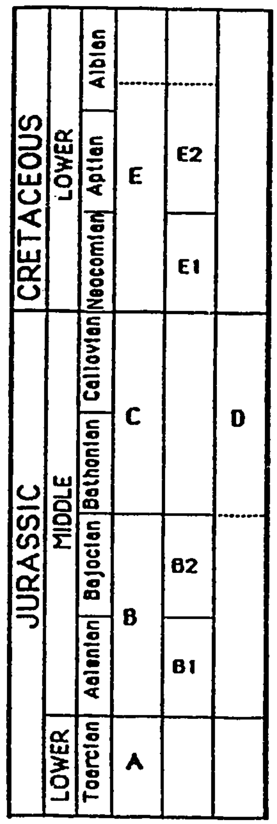


Figure 3.5. Stratigraphic chart of the Lower Cretaceous and the Upper part of the Lower and Middle Jurassic; and the corresponding Microfaunal assemblages identified in this study.

The paleoenvironments for each sample were estimated on the abundance and diversity of miospores and dinoflagellates observed in the samples. Most of the Jurassic shales examined contained a restricted assemblage of opportunistic dinoflagellates suggesting a stressed environment. This is particularly true for the Toarcian Poker Chip Shale which was probably deposited in an open sea, characterized by highly anoxic bottom waters. Dark organic-rich shales in the Rock Creek Member also contained a restricted assemblage of dinoflagellates suggesting similar conditions. Shale samples associated with the limestones and sandstones of the Rock Creek Member tended to contain a diverse assemblage of open marine dinoflagellates, suggesting normal marine salinities with good water circulation. A sample collected from the Rock Creek Member in 10-35-38-5w5 contains abundant opportunistic dinoflagellates, with fair diversity, suggesting a marginal marine environment. Such an interpretation is in keeping with the facies of this well, which indicate a deltaic setting (see sedimentologic discussion).

Shale samples collected from the Lower Mannville Ellerslie member generally yielded abundant miospores with no evidence of *in situ* dinoflagellates. These were, therefore, interpreted to be non-marine in origin. Some samples collected from thin shale breaks in thick sandstone sections yielded a low diversity assemblage of marine dinoflagellates and abundant miospores. Such an assemblage suggests brackish water conditions. Thus there was some marine influence involved in the deposition of parts of the Ellerslie Member.

STRATIGRAPHY OF THE STUDY AREA

POST-MISSISSIPPIAN UNCONFORMITY SURFACE

The current study area (shown in Figure 3.2) lies proximal to the subcrop edge of the Jurassic Fernie Formation. This formation unconformably overlies Mississippian carbonate sediments which are exposed along a gently westward sloping, post-Mississippian unconformity. This subcrop surface exposes progressively older Mississippian strata (Elkton, Shunda, Pekisko and Banff formations) toward the east. At the subcropping edge, exposure and karstification of the more porous Mississippian strata, such as the Elkton and Pekisko Formations, created highly irregular paleotopographic relief along this unconformity. Faulting in the Mississippian strata (discussed more fully in the previous paper), at the time of Fernie Formation deposition, also affected the paleotopography of this surface, which greatly affected the depositional and preservational aspects of the various units in the Fernie Formation.

An isopach map from the base of the Glauconite coal to the top of the post-Mississippian unconformity shows the paleotopography of the post-Mississippian unconformity during Jurassic time (Figure 3.6), only where the unconformity is covered by Jurassic sediments. In some areas, Early Cretaceous erosion has subjected the post-Mississippian unconformity to further denudation. Where the unconformity is covered by Jurassic strata, the surface is relatively flat and slopes gently westward toward the disturbed belt. There are two broad zones that can be distinguished on the map, separated by the north-south trending Mississippian axial high. To the west of this high, the subcrop surface is a relatively flat and stable surface. To the east, in the Medicine River basin feature and the Sylvan Lake Field, the post-Mississippian unconformity surface is one of high relief. Topographic relief on the post-Missississipian

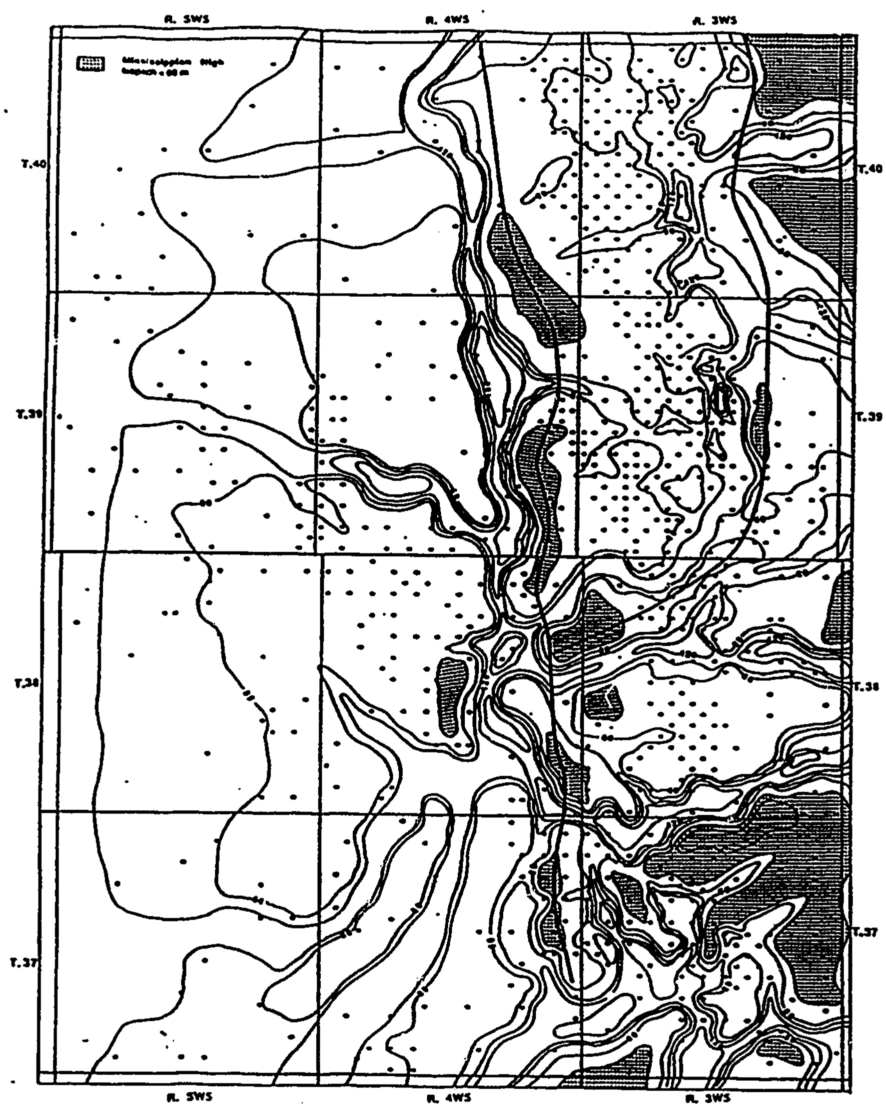


Figure 3.6. Isopach map from the base of the Glauconite coal to the post-Mississippian unconformity. Bold line running north-south through the middle of the diagram indicates the location of the axial high on this unconformity while the bold line which splays to the north east indicates the eastern limits of the Medicine River-Gilby basin. Contour interval=20m.

unconformity is particularly complex in the Syvan Lake area where up to 80 meters of relief are observed over as little as 1km. Jurassic sediments not only cover the high areas on the post-Mississippian unconformity, but also fill the low-lying valleys, therefore suggesting that this paleotopography was present during the deposition of the Jurassic section.

West of the axial high, paleotopography was relatively stable and the only major relief is caused by Early Cretaceous erosion. This surface now dips markedly to the west, but, in Middle Jurassic time this surface probably displayed a very low angle dip.

LOWER JURASSIC

When fully preserved the Fernie Formation consists from base to top of the Nordegg member and its equivalents, the poker Chip Shale, the Rock Creek Member and an informally named Upper Fernie shale (Marion, 1982). Each of these units is bounded by an unconformity and therefore thickness trends within the various units are a function of post-depositional erosion and do not generally reflect depositional trends. The Nordegg member was only found in the northwest part of the study area and is truncated to the south by pre-Poker Chip Shale erosion. This trend (Figure 3.7) reflects increased post-Nordegg/pre-Poker Chip Shale subsidence in that area, which allowed for the preservation of the Nordegg sediments (see previous paper).

The Poker Chip Shale consists of black, fissile to papery shales which are organic-rich, pyritic and generally massive in nature. On logs this unit exhibits very high gamma levels, very low resistivity, and very high neutron porosity and sonic readings. The density porosity generally shows an opposite trend to the neutron porosity, which indicates the very

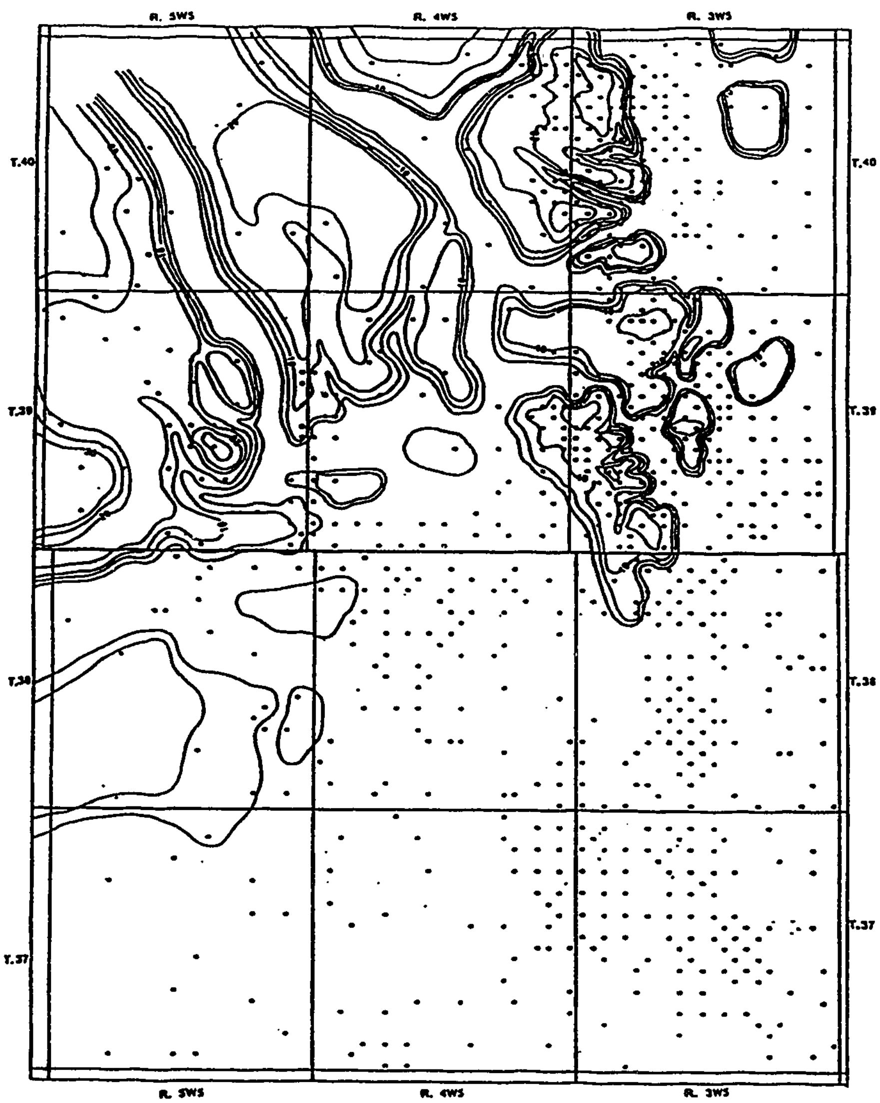


Figure 3.7. Isopach map of the Nordegg interval; thickness variations are primarily due to post-Sinemurian erosion. Contour interval=5m.

dense but highly organic nature of the shale deposit. These characteristics are indicative of a condensed marine shale and make the Poker Chip Shale an easily recognizable stratigraphic marker bed, separating Lower Jurassic from middle Jurassic strata. Lower Mannville Ellerslie shales were generally deposited in a terrestrial environment and are therefore highly oxidized, carbonaceous and relatively sandy. Gamma levels in these younger shales do not reach the high levels of the Poker Chip Shale, even when the latter is oxidized. Similarly, resistivity is not as low and the neutron porosity not as high. Shales of these different systems are thus easily distinguishable on geophysical logs and in core.

At its lower boundary, whether it is in contact with Nordegg sediments or with the Mississippian unconformity, the basal one meter or so of the Poker Chip Shale is altered from a very dark black colour to a dull grey-green or bright olive green colour. This feature in shales has been attributed to a decrease in the Fe^{3+}/Fe^{2+} and later removal of the more soluble Fe2+, when rocks are above the water table (Potter et al., 1980). Such oxidation also affected the organic content of the Poker Chip Shale. Oxidation probably occurred during the exposure period associated with the pre-Cretaceous unconformity, where oxidizing meteoric fluids filled porous Medicine River sandstones and altered the overlying shales. The impermeable nature of the Poker Chip Shale did not allow these oxidizing fluids to penetrate deeply into the unit, therefore only the boundaries are affected. Similar alteration occurs in the shales at the top of the Poker Chip Shale, in the detrital shales, and in the shales which fill karst cavities in the Nordegg and the Mississippian strata.

The Poker Chip Shale oversteps the erosional edge of the Nordegg interval to the south and east, except where pre-Cretaceous erosion has

removed it (Figure 3.8). Karstification of lithified Nordegg sediments below, and the presence of Nordegg pebble conglomerates, indicate the presence of a significant hiatus between these units. The base of the Poker Chip Shale is relatively flat and the post-Nordegg interval erosion appears to have formed a planation surface with little or no incision. The Poker Chip Shale thickens from the eastern pre-Cretaceous erosional edge, to a maximum of 20m along the northern part of the axial high. The unit again thins to the west and eventually pinches out between the Rock Creek and Nordegg Members. Therefore, a significant erosional hiatus is evident between the Poker Chip Shale and the Rock Creek Member.

DETRITAL

The Detrital Member is a deposit of green oxidized shales and weathered fragments of dolomite and chert, that accumulated as a zone of weathering on Mississippian carbonate deposits during periods of subaerial exposure and erosion. The Deville or Detrital Member is considered by many to be Late Jurassic to Lower Cretaceous in age (Williams, 1963), but Rall (1980) found the Deville to occur below Lower Jurassic Toarcian age shales in the Gilby Field. Detrital lithologies, in this study, were found beneath the Rock Creek Member, the Poker Chip Shale and the Nordegg Member. This suggests that the Detrital Member is a diachronous deposit, which formed during each of the many different exposure periods of Mississippian strata. Throughout most of this area the detrital underlies Jurassic deposits and is therefore probably Lower to Middle Jurassic in age. Shale samples collected from the Detrital were generally too oxidized to yield any microfossils, although a few Jurassic age fossils were observed. This is in agreement with the Lower Jurassic

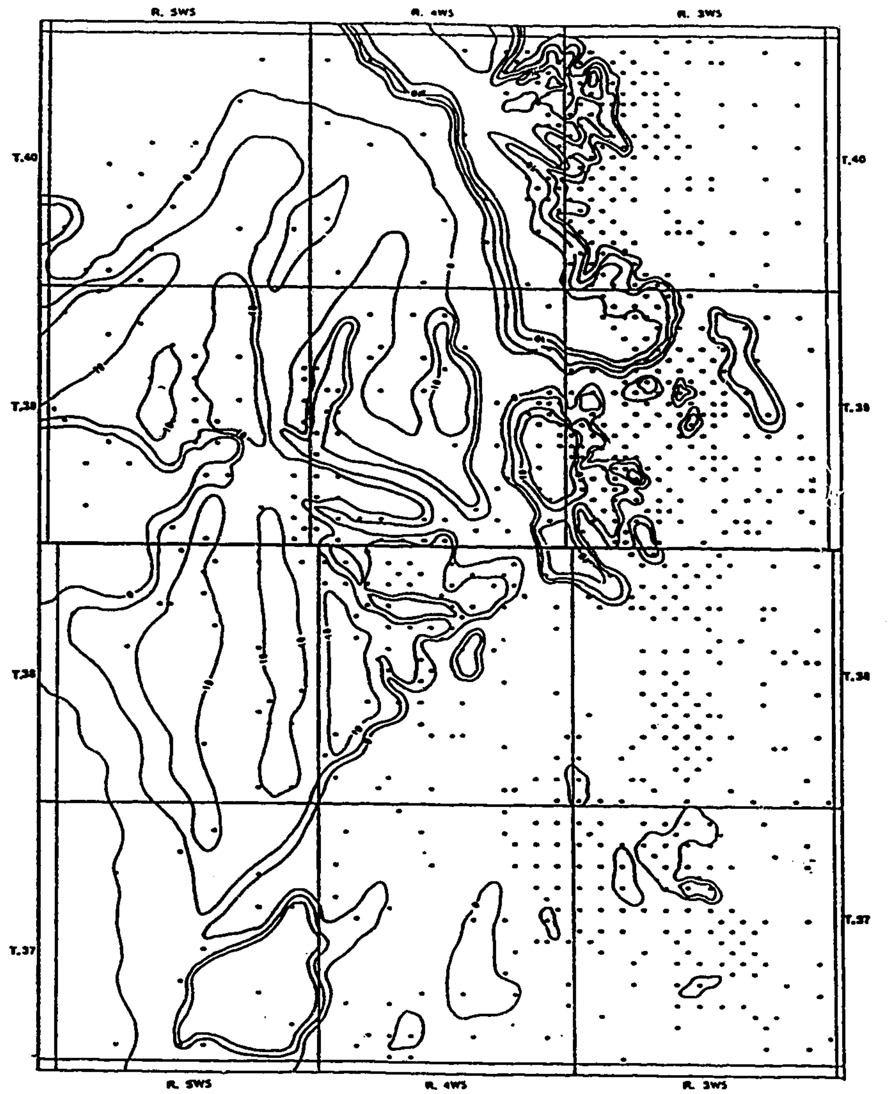


Figure 3.8. Isopach map of the Toarcian age Poker Chip Shale. Contour interval=5m.

age given by Rall (1980).

In the complex Sylvan Lake area, Poker Chip Shale lithologies are only found capping the Mississippian highs, resting directly on Mississippian carbonates. Toarcian microfauna collected from these shales confirm the correlation to the Poker Chip Shale of the outcrop belt. The Nordegg was either not deposited in this area, or more likely was removed by erosion because pre-Poker Chip Shale subsidence was not sufficient to preserve these sediments. The floors of the valley systems, in the Sylvan Lake area, are generally covered with a veneer of dolomite and chert breccias and by green, highly oxidized shales with floating chert and dolomite fragments (Plate 3.1). These are typical lithologies found in the Detrital Member (Rall, 1980).

The dolomite breccias in the Detrital sediments (Plate 3.1c.) contain very large angular blocks and are often grain-supported. Such features are typical of sediments that are formed by solution collapse in karst terrain (Jennings, 1975). Detrital deposits overly unbrecciated Mississippian dolomites. Sediment similar in composition to the Detrital deposits is often found filling cavities within these dolomites (plate 3.1b). The Detrital sediments therefore represent a weathering zone on the post-Mississippian unconformity, overlying a karsted terrain. Karstification can also explain the rough and highly irregular topography seen in the Sylvan Lake area. This area is probably an old karst terrain in which abundant solution valleys and sinkholes formed as the porous Elkton and Pekisko formations dissolved at the exposure surface.

MIDDLE JURASSIC ROCK CREEK MEMBER

Along the subcrop edge of the Jurassic in this study area, the stratigraphic relationships become very complex, particularly in the Sylvan Lake Field. In many instances Ellerslie sandstones sit directly upon sandstones of the Rock Creek Member. Ellerslie sediments may fill deep incisions into Jurassic strata and therefore, often lie structurally lower than the older sediments. This makes mapping of these units extremely difficult and the sandstone reservoirs of the two units are often confused or grouped together.

Several criteria were used in this study to identify and map the Rock Creek member. The first was stratigraphic position between the Poker Chip Shale and the pre-Cretaceous unconformity. The pre-Cretaceous unconformity was often difficult to define, particularly in the absence of core. However, this horizon usually displayed several distinctive features. The pre-Cretaceous unconformity was produced by a prolonged period of subaerial exposure, which subjected the underlying lithified Jurassic sediments to extensive weathering, oxidation and soil formation processes. This unconformity is therefore usually associated with a 0.3 m thick zone of mineralization at the top of the Jurassic. In the sandstone lithology, this zone is marked by dissolution of the framework grains, infilling of the porosity with illuviated clay material (see discussion of illuviated clay material in the petrography section) and replacement of the sandstone by pyrite and other oxide minerals. This is usually a zone of very high density and low porosity which is obvious on the porosity and sonic logs. The high content of the pyrite also causes a distinct low peak in the resistivity and the illuviated clay may cause high gamma levels.

In addition the shale-poor sediments of the Rock Creek Member are generally overlain by shaly sediments or paleosols of the Ellerslie Member producing a very sharp contact on geophysical logs. Such a contact may also occur where the Rock Creek Member is overlain by the Upper Fernie shale.

The Rock Creek Member is lithologically distinct from the overlying Ellerslie sandstones. Rock Creek sandstones are fine to very fine grained and very well sorted. Ellerslie sandstones range from fine to coarse grained and are generally poorly sorted. In some cases well sorted, fine grained sandstones do occur in the Ellerslie Member and are difficult to distinguish from Rock Creek sandstones. Ellerslie sandstones often display abundant lithoclasts and carbonaceous debris and often appear 'dirty'. In contrast, the Rock Creek sandstones are almost pure quartz, they display a sugary texture, may be calcareous and are often found interbedded with fossiliferous marine skeletal limestones. In addition, these sandstones often display features of marine bioturbation. Ellerslie sandstones, on the other hand, were generally deposited in a fluvial setting (or possibly a brackish environment) and therefore, the only bioturbation observed in these sediments included rare and small *Planolites* in shale partings. Finally, the two stratigraphically distinct units contain petrographically separable sandstones. Subtle differences in the chert content allow the distinction of Ellerslie and Rock Creek sandstones (Marion, 1982, 1984). Correlation and mapping of the Rock Creek Member was therefore done on a lithologic basis using all of these criteria.

'UPPER FERNIE SHALE'

The Upper Fernie shale, as defined informally by Marion (1982) is a dark, organic-rich shale, which caps the Rock Creek Member. In this area it is patchy in distribution, as it has generally been removed by early Cretaceous erosion. This deposit can be recognized as a high gamma bed, which displays a sharp contact with the underlying Rock Creek Member. The Upper Fernie shale often contains interbeds of fine rippled sandstone beds signifying low energy, but episodic deposition. Samples collected from these shales contain abundant gymnospermous pollen with common marine dinoflagellates. The Upper Fernie Shale is therefore considered to have been deposited in an open marine environment. The high gamma nature ('hot' shale) of this unit indicates a high organic content typical of transgressive shales. This is therefore a condensed interval which indicates renewed transgression at the end of Middle Jurassic time. This unit has yielded a distinct Bathonian to Early Callovian faunal assemblage which probably makes it correlative with the Grey Beds of the outcrop area.

LOWER MANNVILLE, ELLERSLIE MEMBER

The lower part of the Early Cretaceous in central Alberta is represented by the Lower Mannville Formation, which consists of the Deville Member, the Ellerslie Member (locally known as the basal Quartz) and the overlying Calcareous or Ostracod Member (Glaister, 1959). The diachronous Deville Member has been discussed and is not considered to be part of the Lower Cretaceous in this area. The term Ellerslie Member, as defined by Hunt (1950) contains all sediments deposited between the Deville and Ostracod zone. As the Deville is known to be Jurassic in age in this area, the Ellerslie will be considered here as all sediment deposited between Jurassic strata and the Ostracod Member. The top of the Ellerslie is traditionally defined as the top of the first quartzose sandstone below the Ostracod Member, since this is the easiest method to identify it on the logs. In this study, however, the base of the Ostracod, and therefore the top of the Ellerslie member are marked at the position in the core where the predominately terrestrial deposits of the Ellerslie Member grade up into the calcareous, bioturbated and fossiliferous brackish water deposits of the Ostracod Member.

The Ellerslie Member in this area can be divided into two discrete lithological assemblages. The first consists of a stacked series of fining-upward, fine to medium grained, massive and cross-bedded sandstones. The sandstones are erosionally based and contain abundant clasts of shale, sandstone and chert, particularly at the base. These sandstones generally grade up into interbedded fine sandstone and shale which display lenticular and wavy bedding with ripple lamination in the sandstone beds. Individual sandstone sequences are in the order of 3-4m thick but may attain thicknesses of 10m. These deposits are generally found filling deep valley systems, cut into the subcrop of the Jurassic strata. This assemblage is interpreted as fluvially deposited channel sandstones which fill the valleys of an old drainage system.

The second assemblage is dominated by finer grained material but may also contain fining-upward sandstone sequences. This assemblage consists of carbonaceous, shaly siltstones and sandstones, interbedded with brown mottled shales. Root mottling and preserved root molds are the most commonly observed structure in the sandstones, although rare parallel and ripple laminated sandstones are observed. These sediments are commonly associated with a series of interbedded fine sandstones and black shales that display highly distorted, and often nearly vertical bedding, with abundant evidence of soft sediment deformation and microfaulting. This assemblage is generally found overlying the Jurassic subcrop adjacent to the valleys. Thin sharp-based, cross-bedded sandstones are also associated with these deposits but are not as frequent or thick as those in the valleys. These deposits are interpreted as being predominately interfluvial flood plain sediments with occasional small deposits of fluvial channel sandstones. Coal beds may be associated with these assemblages.

Two shale samples were collected from the Ellerslie Member and analyzed for the micropaleontologic assemblages. One was from the top of a thick fluvial channel sandstone in 10-18-40-2w5 and yielded an Aptian to early Albian assemblage of microflora. The second was from a valley fill in 04-27-38-3w5 and yielded a somewhat older Neocomian aged microfloral assemblage. This indiates a much earlier age of the valley-filling sediments than those of the surrounding areas. The sample collected from the 10-18-40-2w5 well was taken from a shale break in the sandstone. This shale yielded opportunistic dinoflagellates representative of a marginally marine environment, suggesting that there is some marine influence in the deposition of the regional Ellerslie Member. This is not surprising as brackish water ichno-assemblages are observed in similar shale breaks in other Ellerslie cores. The second shale sample was collected from the valley-fill deposits. This shale sample yielded a diverse assemblage of miospores and no *in situ* dinoflagellates, and is therefore considered to be non marine in origin. Although these data are extremely limited, they does suggest that the Ellerslie Member was more greatly influenced by marine conditions up through the section.

STRATIGRAPHY AND DISTRIBUTION OF THE ROCK CREEK MEMBER

Figure 3.9 illustrates the type-well chosen to represent and correlate the Rock Creek Member throughout this area. This well was chosen because it contains good quality and up to date geophysical log data, core data to calibrate the lithology to the log data, and biostratigraphic data, correlating it to the outcrop belt. The Rock Creek Member in this well is divided into three depositional sequences separated by surfaces of subaerial exposure. The detailed sedimentology of these sequences will be discussed in more detail later. However, the nature of the bounding surfaces are relevent to the stratigraphy of this unit. The upper sequence boundary (Plate 3.2a) consists of a rubbly and mottled horizon, with pebbles of the underlying material at the top, overlain by a brownish green shale bed. The mottled zone is interpreted as an illuviated clay horizon (see discussion in the petrography section) and the pebbles indicate the surface of subaerial exposure. The shale bed contains rare opportunistic dinoflagellates suggesting a restricted marine depositional environment. This shale is therefore the condensed section of the overlying sequence and indicates renewed transgression, over the exposed surface.

The lower sequence boundary (Plate 3.2b) only contains the illuviated clay horizon, which is in sharp contact with the clean sandstones of the overlying middle sequence. This contact signifies the surface of subaerial exposure. At this sequence boundary the condensed section was either removed by erosion, as the high energy conditions of the shoreface migrated westward over the area, or it was never deposited. The occurrence of clean shoreface sandstones (see sedimentology discussion) over this surface suggests that the latter was probably the case. Such a series of events would require extremely low rates of subsidence, and

14-19-40-4W5

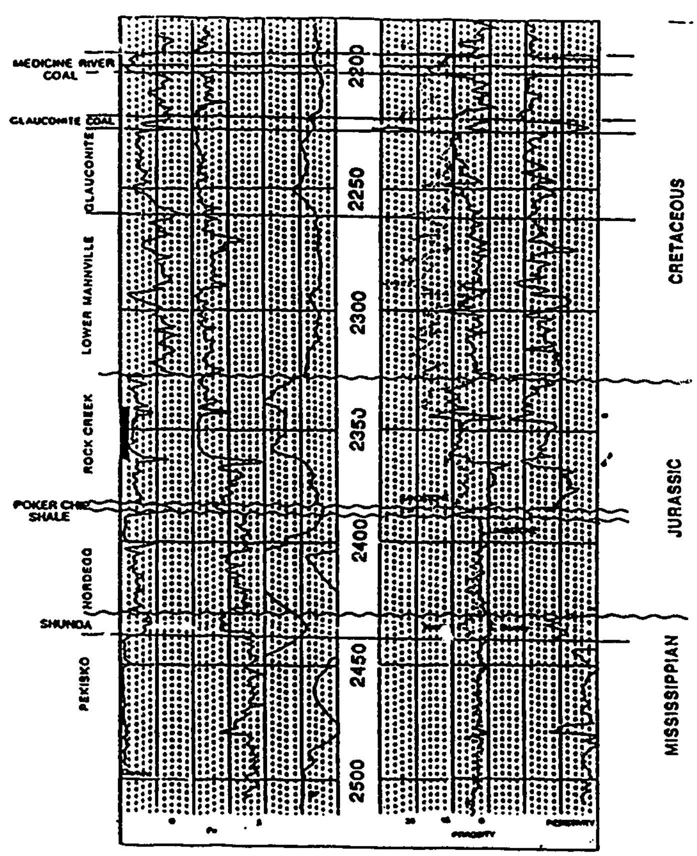


Figure 3.9. Type-log used in the correlation of the Rock Creek Member. Core interval examined is indicated by the black column on the left side of the gamma log. Shale samples used for micropaleontological analysis are denoted by the points to the right of the resistivity log.

therefore low accommodation potential.

As illustrated in the type-log (Figure 3.9), the zones of subaerial exposure are easily identified by their high density, Pe curve and gamma levels and by the low resistivity readings related to their pyrite content. In many parts of the study area the surfaces can be traced using these log characteristics. In some cases, however, the sequence boundaries can only be observed in core, as the zone of illuviated clay mottling may have been eroded. In this case, clean sandstones of one sequence rest unconformably on those of another. Plate 3.2c demonstrates that the underlying sequence was lithified and eroded prior to the deposition of the subsequent sequence. As no pedogenic or condensed section is preserved between these sandstones, this sequence boundary is virtually indistinguishable on geophysical logs.

In the central portion of the study area, where the Rock Creek becomes shalier, the sequence boundaries also become very difficult to trace without core data. For example, one such boundary is characterized by a 1cm bed of pebbly sandstone between silty shales (Plate 3.2c). Again, this type of boundary is very difficult to recognize when just geophysical logs are available.

Despite the difficulties expressed in the above discussion, attempts were made to correlate these sequences across the study area. Cross-section A-A' (Figure 3.10a) shows the correlation of the Rock Creek Member, through the type well, across the northern part of the study area. This section shows the thinning of the Rock Creek Member, from west to east, over the Mississippian axial high. The three sequences identified in the 14-19-40-4W5 well can be correlated through the western part of the area, although they may become conformable in the 06-05-40-5w5

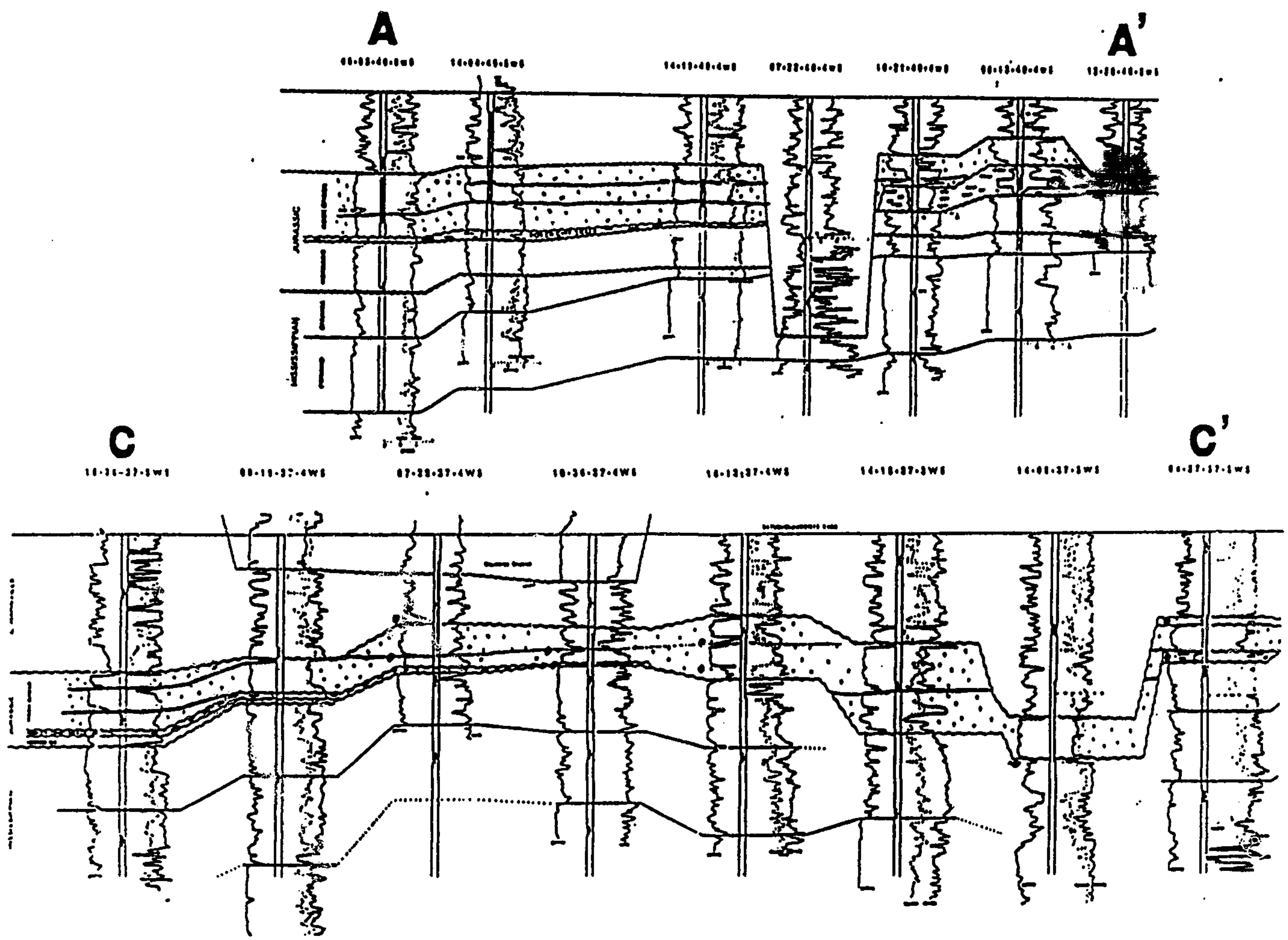


Figure 3.10. Cross-sections A-A' and B-B' are west to east sections through the Gilby and Sylvan Lake Fields respectively. Stippled portion denotes the Rock Creek Member while interval dashed shows the Poker Chip Shale. Black circles indicate shale samples analysed for microfauna.

well. These sequences progressively onlap the Mississippian high, with lower sequences being eroded at the expense of the overlying sequences. This suggests very low rates of subsidence, and therefore low accommodation, over the Mississippian high. East of the axial high, the Rock Creek Member generally has been eroded during Lower Cretaceous exposure.

The stratigraphic architecture of these Rock Creek sequences along the line of the A-A' section (Figure 3.11a) suggests that the Poker Chip Shale thickens in a step-like fashion over the axial high. The Rock Creek sequences appear to terminate against these step features, which may represent shoreface incision during periods of relative sea-level stillstand, in much the same way as envisioned by Plint et al., (1988) for the Cardium deposits.

Cross section B-B' shows the correlation of the Rock Creek Member from the type well south into the Sylvan lake field and cross-section C-C' shows the correlation of the Rock Creek Member through the Sylvan Lake Field (Figure 3.12). In the westernmost well of C-C' the expression of the three sequences is again developed. To the east, however, erosion at each of the unconformities, including the pre-Cretaceous unconformity, makes it difficult to distinguish the different sequences. Once again, erosion over the axial high has removed the lower sequences, developing an onlapping stratigraphy as in the northern section. In this area the entire Nordegg section has been removed prior to the deposition of the Poker Chip Shale, signifying very low rates of subsidence during the pre-Toarcian exposure period. In addition, the Poker Chip Shale is very thin and often eroded in this area, suggesting continued low rates of subsidence pre-dating Rock Creek deposition.

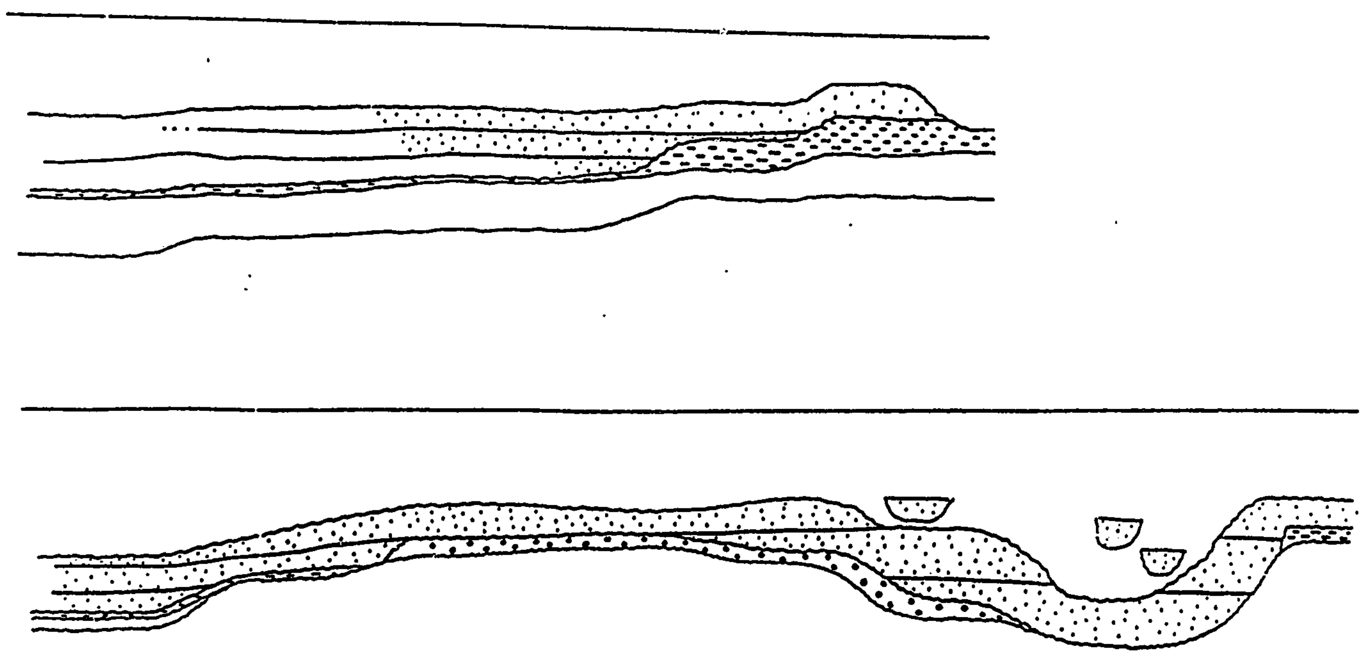


Figure 3.11. These are schematic sections following the lines of the cross-sections in Figure 3.8; the internal stratigraphy of the Rock Creek Member is shown on both sections. Fine stippling shows the Rock Creek Member sandstones, dashed pattern shows the Poker Chip Shale and the coarse stippling shows Detrital lithology.

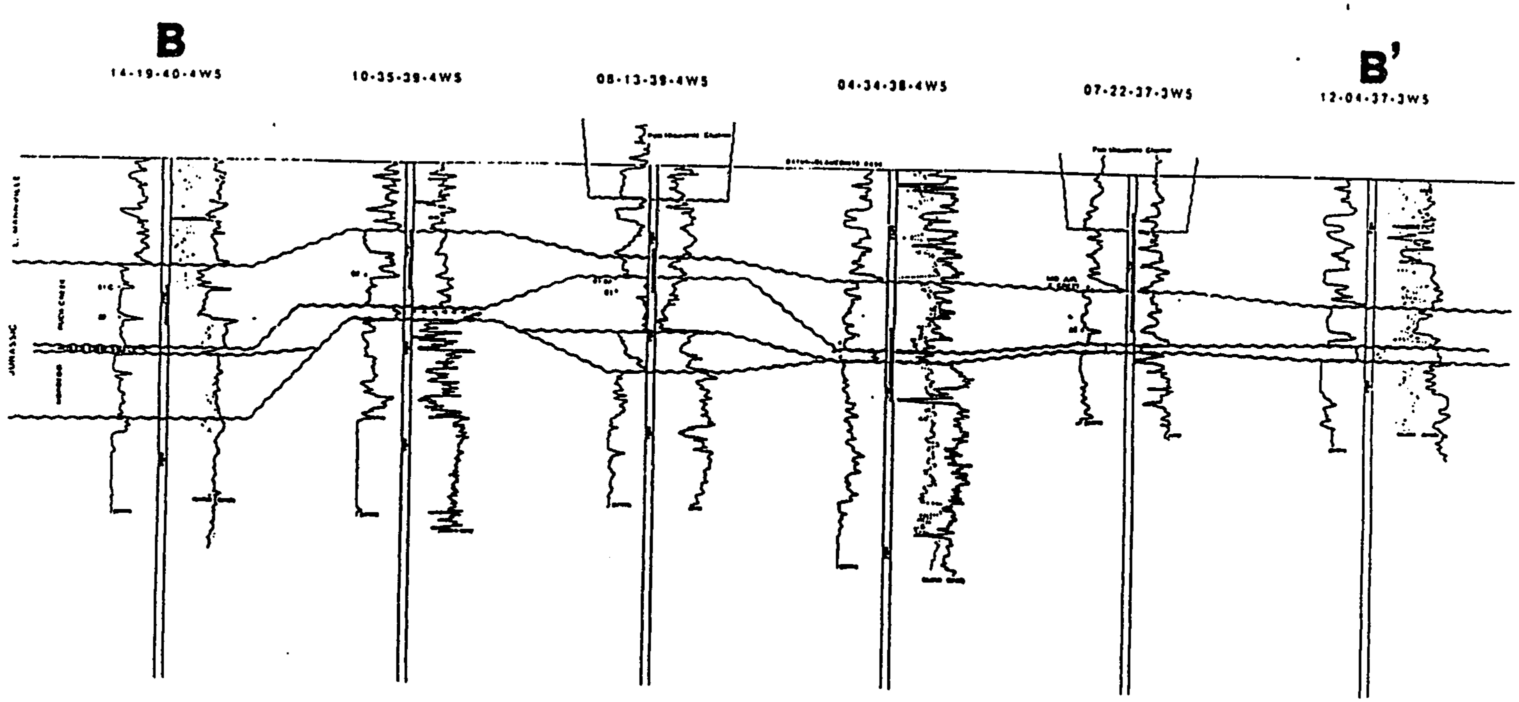


Figure 3.12. Cross-section B-B' shows the correlation of the Rock Creek Member from north to south, through the study area. Stipples pattern Indicates Rock Creek Member. Black circles indicate shale samples analyzed for micropaleontology.

East of the axial high the Rock Creek Member is not eroded, but progressively thickens into the low-lying areas. Once again the sequence architecture is developed in the Sylvan Lake valleys, with lower sequences terminating against valley walls, beneath the erosion surfaces of the overlying sequences. Lower Cretaceous drainage patterns appear to have followed these low areas. In the lowest parts of the valleys, incision into the Rock Creek Member has developed a series of complex and isolated outliers. The incision of the Rock Creek subcrop, and emplacement of Lower Cretaceous sandstones structurally adjacent to Rock Creek sandstones, have created a very difficult area to correlate and map (ie. Figure 3.13). Within section 30 (T. 37; R. 3w5), the Rock creek Member is present along the east and west boundaries, but a valley is eroded through the Rock Creek in the center of the section. This valley is filled with Lower Mannville sediments and the base of the valley is floored by Rock Creek sandstones. These basal sandstones, however, are part of a lower sequence and are not equivalent to the sandstones that form the valley walls. This example illustrates the difficulty that arises in correlating through just one section in the Sylvan lake field.

West to east cross-sections through the Sylvan Lake Field effectively illustrate the thickening from the axial high on the Mississippian strata, into the low-lying areas (Figure 3.14). At the eastern limit of Sylvan Lake Field, the pre-Rock Creek depositional surface returns to its regional elevation and only the uppermost Rock creek sequence is preserved (Section A). Just east of this section, the subcrop limit of the Rock Creek Member is reached and the entire Member is eroded. The lowest part of the area appears to be floored by a thick veneer of Detrital Member sediment, suggesting that the lows are formed

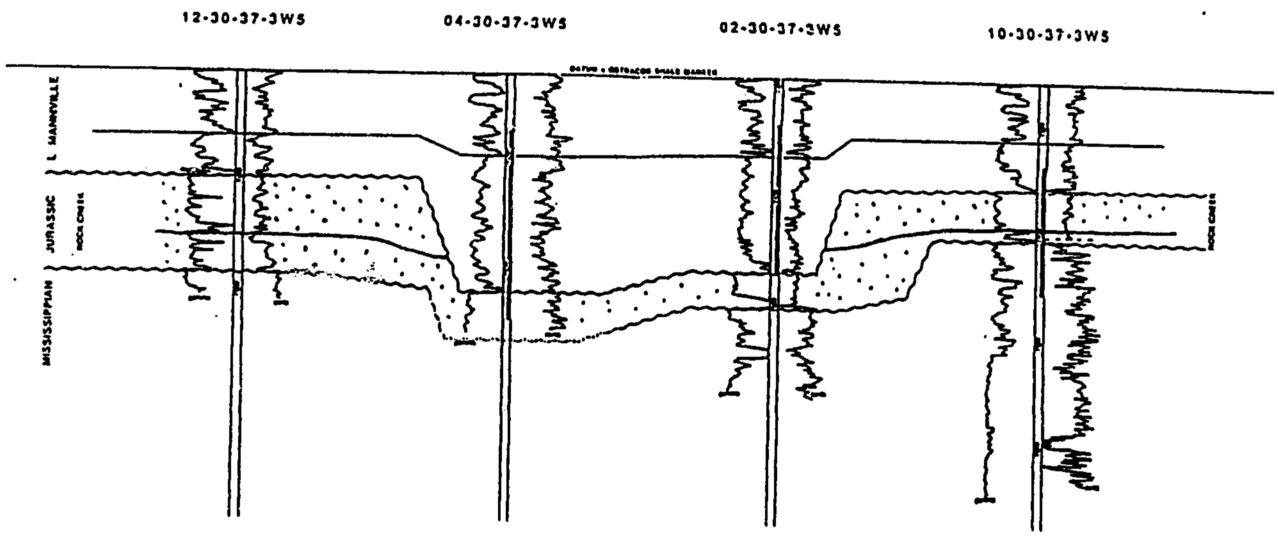


Figure 3.13 This cross-section shows the stratigraphic complexity of the Rock CreeK Member in the Sylvan Lake area. All wells shown are found in section 30 (37-3W5).

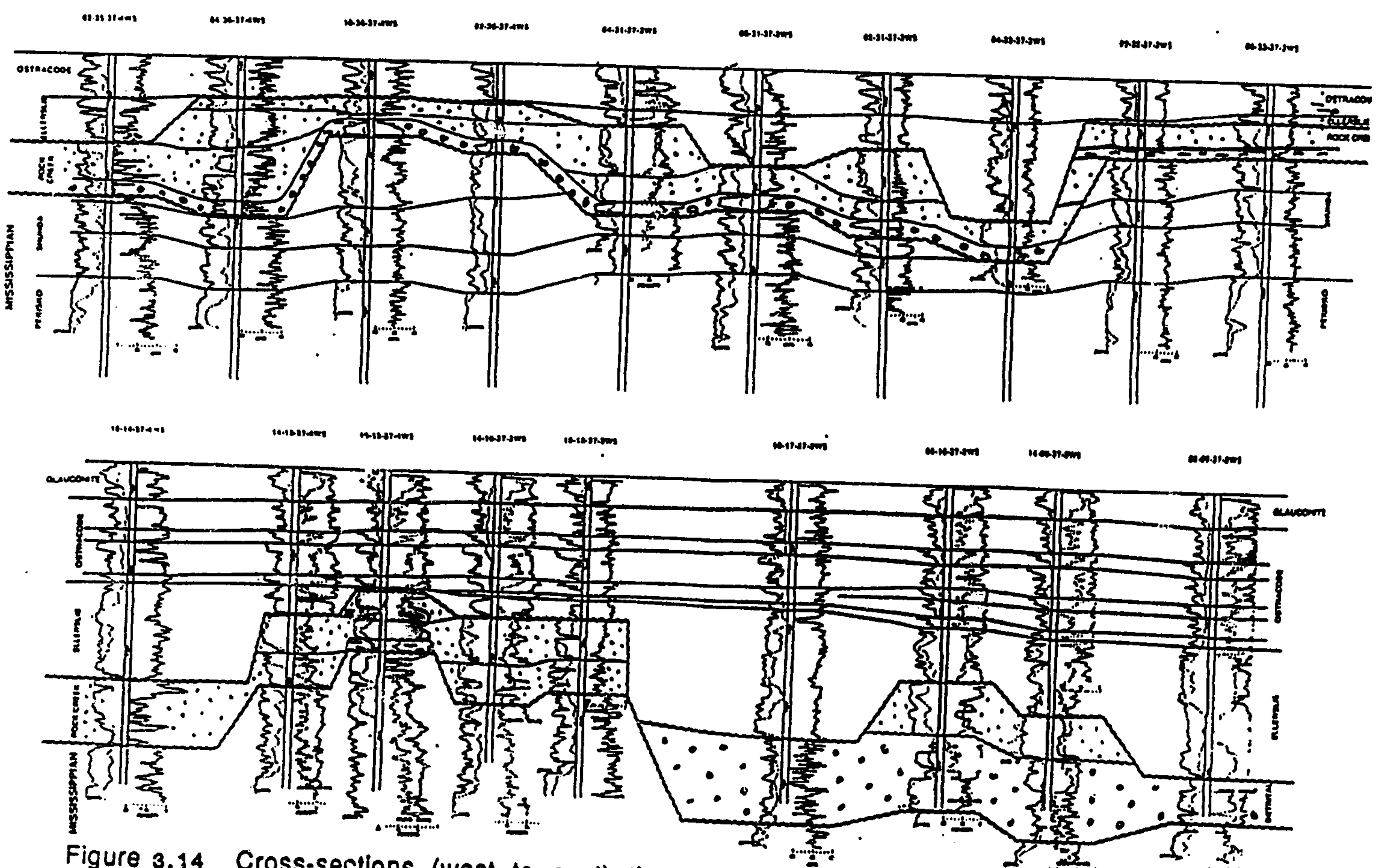


Figure 3.14 Cross-sections (west to east) through the Sylvan Lake Field, showing the stratigraphic complexity of the area. Fine stipples indicate the Rock Creek Member, while coarse stipples are detrital deposits.

over karst collapse debris in solution valleys (ie. Section B). Pre-Cretaceous erosion in the low area of section B has formed an oil-productive outlier within the lowest Rock Creek depositional sequence.

Figure 3.11b illustrates the schematic sequence architecture of the B-B' section. Termination of the lower sequences against the pre-Rock Creek Member unconformity is strictly a function of inter Rock Creek erosion, between the depositional sequences. This does not appear to be a function of non-deposition around ancestral highs (or islands) since depositional trends within the sequences are not affected. Therefore, it is suggested that the depositional surface of each sequence in the Sylvan Lake area was relatively flat and that the sequences were actually deposited on the highs. Subsidence rates over the high were lower than in the areas to the east and west, and each sequence was removed over the high before the deposition of the next. Relatively high rates of subsidence must have been active in the low areas of the Sylvan Lake Field, during the deposition of the entire Rock Creek Member to allow for the anomalously thick section of stacked Rock Creek sandstones.

Continuous vertical movement throughout the deposition and lithification of the Sylvan Lake Rock Creek sediments is indicated by the extensive syn-depositional and post-lithification disruption and fracturing of these sandstones (Plates 3.3a-c). This shows that fracturing and vertical movement probably occurred over a long period in this area, beginning very early in the history of the Rock Creek deposits.

The isopach of the Rock Creek (Figure 3.15) indicates that the interval is at its thickest in the westernmost part of the area where it reaches a maximum of 45m. Thickness varies little toward the east,

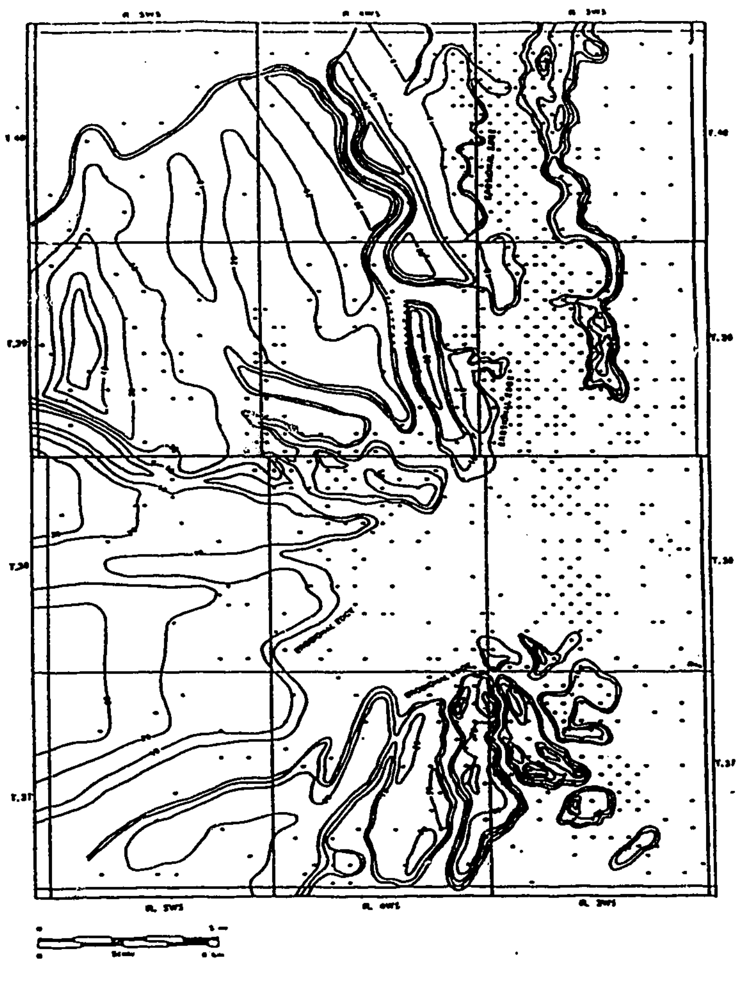


Figure 3.15. Isopach map of the Rock Creek Member. Most of the thickness variation is due to pre-Cretaceous erosion, although depositional thickening appears to occur toward the west. Contour interval=10m.

except where substantial early Cretaceous erosion has removed significant portions.

The Rock Creek interval in the northern part of the area is separated from the Sylvan Lake area to the south by a northeast-southwest trending valley system that widens to the northeast. In the central part of the area, the subcrop is dissected by several deep and narrow, branching valleys. A similar valley system separates the Gilby area Rock Creek from the Willesden Green area Rock Creek. These valleys are generally less than one mile wide but may be cut into the Jurassic subcrop as much as 125m. The sediment that fills the valleys generally appears shaly and easily differentiated from Jurassic sediments. These are typical of sediments found in the Ellerslie Member. Therefore, the valleys are interpreted to have been formed in Lowermost Cretaceous time. An example of one of these deep narrow valley systems is illustrated in the Cross-section A-A' (Figure 3.10; well 10-21-40-4w5).

An isopach of the total sand distribution within the subcrop limits of the Rock Creek interval (Figure 3.16) indicates that the sandstone lithology is restricted to a north-south trending belt which is limited to the east by erosion. The sandstones generally grade into clean skeletal limestones to the west, but in the central part of the area they grade into sandy shales and shaly limestones. In this central area, however, the sandstone lithology extends farther eastwards than in the other areas. This anomaly will be explained when the sedimentology of the unit is discussed. In the northern area the sandstone isopach thickens from west to east, and reaches a maximum of 15m just west of the axial high feature. The thickness of the sandstone then decreases dramatically over the axial high, reflecting the erosion of lower sandy sequences over the

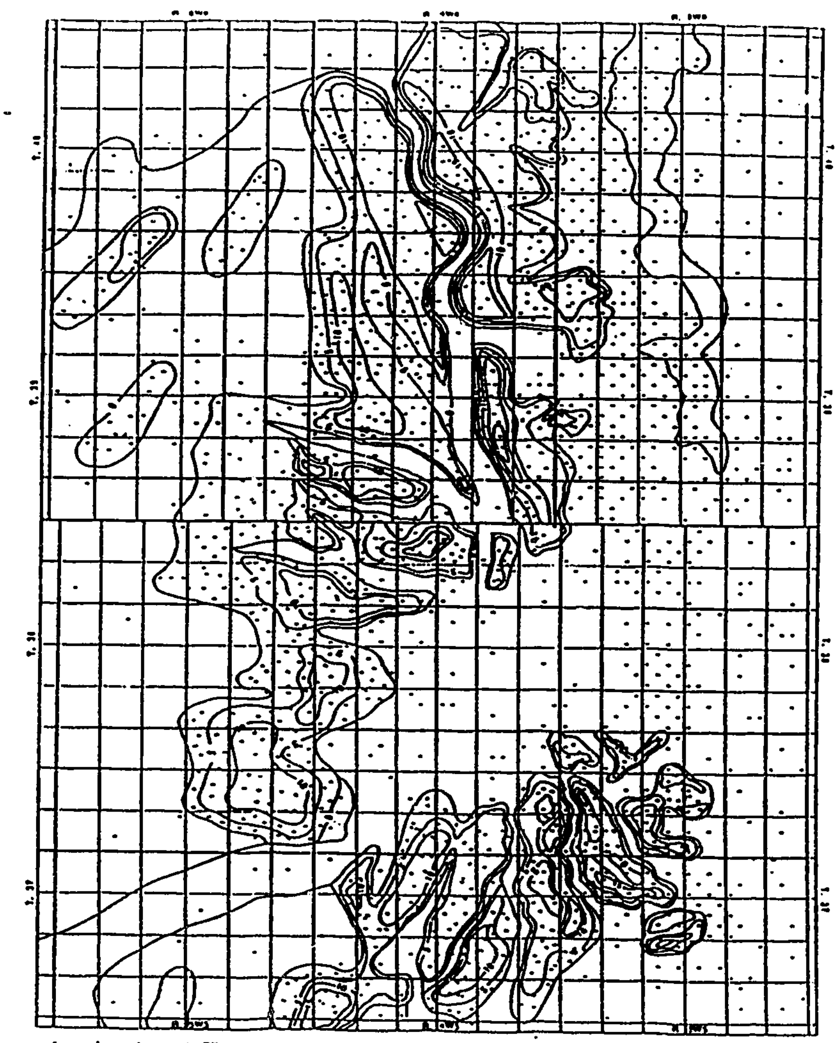


Figure 3.16. Gross sand isopach of the Rock Creek member indicating a transition from limestone to sandstone from west to east. Sandstones thicken up to the axial high area and then thin over the high. Complexity in the Sylvan Lake area is due to complex subsidence and erosional trends. Contour interval=5m.

high. East of the axial high the sandstone again thickens as a result of increased subsidence.

The distribution and thickness trends in the Sylvan Lake area are once again more complex than the surrounding areas. The sandstones thicken into the lows on the post-Mississippian unconformity except where they were eroded by Lower Cretaceous valley incision. Over the highs the sandstones thin markedly and in places were completely eroded.

SEDIMENTOLOGY

Most of the cores examined were interpreted as penetrating the middle depositional sequence and therefore the sedimentology of this sequence is best-documented. The other sequences were cored in only a few wells, but there appears to be very little difference in the processes responsible for their deposition. Therefore the model developed for the B sequence is considered applicable to the entire Rock Creek interval.

The Rock Creek Member in this area is divided into three lithologically and texturally distinct associations of facies which are laterally gradational with one another, and are often interbedded. Their time-equivalence therefore makes them representative of different depositional systems of the same depositional sequence (Posamentier, et al., 1988). The facies associations recognized here will be described and interpreted separately and then assimilated into a depositional model for the Rock Creek strata.

Facies Association 1. Fine grained quartzose sandstones.

This is the most commonly observed association of facies in the Rock Creek Member. It consists of extremely quartz-rich (generally greater than 97% of detrital grains), fine grained sandstones with varying degrees of bioturbation and shale content. Differentiation of the facies in this association is defined of the basis of shale content and on the degree and type of bioturbation.

Facies 1.1. (Plate 3.4)

This facies consists of predominately massive, fine grained, well sorted quartzose sandstones. Occasional 20cm sets of planar cross-bedding, with erosional bounding surfaces and dips in the order of 20° are observed. Bioturbation is generally absent or in rare cases consists of discrete dwelling structures of *Skolithos* and *Ophiomorpha*. More commonly, the facies is massive or highly altered by fractures and faults, or mineralization products associated with the unconformity surfaces. This is particularly true where the facies is found close to the Rock Creek Subcrop edge and in the Sylvan Lake area.

The cross-bedding indicates high energy conditions in which sand migrated across the sediment water interface in the form of megaripples. High energy conditions and a mobile substrate provided a relatively stressful environment for most organisms. Where present, bioturbation is restricted to simple dwelling structures of suspension-feeding marine organisms.

This facies is interpreted to have been deposited in a high-energy, upper shoreface environment. Marine conditions are indicated by the presence of the rare *Ophiomorpha* and *Skolithos* . High energy conditions in the shoreface are indicated by the general lack of bioturbation, which indicates frequent and vigorous wave and current reworking to overcome the effects of biogenic activity (Howard and Reineck, 1981). High energy conditions are also indicated by the presence of medium scale sets of erosionally based cross-stratification (Clifton et. al., 1971) and by the lack of fine grained sediment throughout the facies.

Facies 1.2. (Plate 3.5a-b)

Facies 1.2 consists of fine grained, quartz-rich sandstone which contains abundant discontinuous and stylotized shale laminations. Sedimentary structures are readily observed and consist of bidirectional ripple lamination and in some cases symmetrical ripple forms, preserved with a capping thin mud layer. Bioturbation is rare but, more abundant than in the previous facies, and consists predominately of dwelling structures such as *Ophiomorpha* and *Skolithos.*

Energy conditions were lower here than in the previous facies, but were still high enough to prevent biogenic reworking of the sedimentary structures. The ichnofossils are restricted to dwelling structures of the *Skolithos* ichnofacies and represent high energy tolerant forms which rely on a suspended food source. This sediment was deposited in the lower part of the upper shoreface, where energy conditions still dominated the fabric of the deposits. However, lower flow conditions resulted in the development of ripple lamination and allowed for more thorough biogenic reworking of the physical sedimentary structures.

Facies 1.3. (Plate 3.5c-d)

This facies varies from slightly shaly sandstones to shaly sandstones. All evidence of physical sedimentary structures has been obliterated by thorough biogenic reworking, which, coupled with the higher argillaceous content, suggests lower energy conditions. This facies contains a diverse and abundant assemblage of ichnofossils, including *Skolithos*, *Rosselia*, *Teichichnus*, *Palaeophycus*, *Planolites*, *Thalassinoides*, *Asterosoma*, *Chondrites*, *Diplocraterion* and *Bergauaria*.

The diversity of the ichnofossils, and the presence of forms such as *Diplocraterion*, *Asterosoma* and *Rosselia*, are indicative of fully marine conditions. This ichnofossil assemblage records a diverse group of both deposit-feeding and dwelling structures and contains a mixture of forms present in both the *Cruziana* and *Skolithos* ichnofacies. This suggests a relatively low energy environment, allowing the accumulation of organic detritus in the sediment, which deposit-feeding organisms mined. Energy levels were still high enough, however, to provide suspended food and support a number of suspension-feeding organisms. In addition the sediment is still dominated by fine grained quartz-rich sand which is consistent with moderate energy levels. The low content of fine grained sediment suggests that relatively little was available in the water column to accumulate on the sea floor.

Facies 1.4. (Plate 3.7c-d)

This facies is lithologically similar to the previous facies but is differentiated on the basis of its ichnological assemblage. It is typically intensely bioturbated with no preserved physical sedimentary structures. The ichnofossil assemblage is dominated by deposit-feeding structures such as *Teichichnus, Planolites, Thalassinoides,* and *Chondrites*. This facies differs from the previous one in that representatives of the *Skolithos* ichnofacies are generally absent, and grazing traces such as *Helminthopsis* become common. This assemblage is restricted to the *Cruziana* ichnofacies and representatives of the *Skolithos* ichnofacies are absent. The predominance of the *Cruziana* ichnofacies suggests that this facies was deposited in an offshore setting. The lack of abundant fine grained material suggests that very little shaly sediment was available for accumulation in this low energy environment.

Facies 1.4 generally forms the base of the vertical sequence and is succeeded progressively upward by facies 1.3 through 1.1. There is therefore a general coarsening-upward grain size trend, and a decrease upward in the diversity and degree of bioturbation. This sequence indicates an increasing energy level upward in the section as the upper shoreface sequence prograded over the lower shoreface and offshore sediments. The full sequence, however, is rarely preserved due to erosion of the underlying sediments by the high energy shoreface environment. There is also a general tendency for the upper shoreface sediments to dominate the eastern sections of the Rock Creek, while lower shoreface and offshore deposits dominate the western portions of this facies association. This suggests that the shoreline lay to the east and prograded westward over the shelf. The shale content of the lower shoreface and offshore facies also varies. This variation is related to the deposition of facies association 2 and will be discussed in a later section.

Facies Association 2. Heterolithic Shale and Sandstone Assemblage.

This facies association is locally restricted, only occurring in the central part of the area. Sediments of these deposits contain a variety of lithologies which range from clean fine grained quartzose sandstone to silty shale. The difference between this association and the previous one is the abundance of fine grained detrital sediment and the very different ichnological fabric.

Facies 2.1. (Plate 3.8a-b)

This facies consists of fine grained, low angle laminated sandstone with abundant styolites and millimeter thick shale laminations. In some cases sharp, scour-based, planar cross-bedded sandstones occur. Cross-bedding is in 10cm sets and individual beds dip at approximately 20°. Bioturbation is virtually non-existent in this facies although rare small *Teichichnus* and *Planolites* burrows may be observed.

Facies 2.2. (Plate 3.8d)

Facies 2.2 is composed of interbedded fine grained sandstone and silty shale. Original bedding is usually preserved and consists of 1-5cm beds of bioturbated fine sand with 0.5-3cm beds of silty shale between. Sandstone beds appear to be extensively bioturbated and no internal structure can be distinguished. Some biogenic mixing of the sandstone and shale lithologies has occurred, but the original bedding is still readily recognizable. Biogenic reworking of this sediment is therefore far from complete. The biogenic fabric is dominated by a low abundance and low diversity ichnofossil assemblage, which consists of simple feeding and dwelling structures such as *Planolites* and *Skolithos*, and a few escape traces. This low diversity and abundance of trace fossils, coupled with the relatively small size of the individual traces, suggests a stressed environment in which normal marine organisms found it difficult to thrive. This suggests an environment which was probably less saline than the normal marine waters. Stress may also have developed in response to periodic fluxes in sediment input, as suggested by the interbedding of the sand and shale lithology.

Facies 2.3. (Plate 3.8c)

This facies consists of finely interlaminated, fine grained sandstone and silty shale, with millimeter scale, very low angle to horizontal lamination. Bioturbation in this facies is virtually non-existent, despite the fine grained nature of the deposit. This environment was dominated by frequent short-lived influxes of coarse sediment, followed by periods of fine sediment deposition from suspension.

Although the fine grained nature of this deposit suggests a favorable site for biogenic reworking there is no evidence of bioturbation in these sediments. This suggests that the environment in which this facies was deposited in was highly stressful, probably due to high rates of sedimentation (see subsequent discussion) and possible fresh water influx.

Facies 2.4. (Plate 3.9)

Facies 2.4 consists of fine grained sandstone and silty shale as in the previous two facies, but all evidence of physical sedimentary structures was completely obliterated by biogenic reworking. The textural fabric of this facies is completely dominated by the ichnofossil *Teichichnus*. Individual *Teichichnus* burrows are large, up to 2cm in diameter and longer than the width of the core (4 in.), and may extend vertically through the core for as much as 25cm. The precursor sediment appears to have been similar to the previous facies, but the texture has been completely reworked.

Teichichnus is formed by a horizontally-burrowing deposit-feeding organism, probably a polychaete worm, which excavates the substrate from a more or less fixed vertical position with respect to the sediment/water interface (Pemberton and Frey, 1984a). Vertical movement of the burrow structure, and therefore development of the characteristic retrusive spreiten, occurs in response to the vertical movement of the sediment/water interface, resulting from sedimentation and burial of this surface. In facies 2.4 the burrows extend vertically over 25 cm. If the life span of the trace-making worm was 1.5 years, a reasonable estimate of a polychaete worm's life span, then this indicates minimum sedimentation rates of 17cm per year. This is probably an absolute minimum estimate as the sediment observed is highly compacted and the full extent of the burrows is not observable in these cores.

The dominance of the sediment fabric by one ichnogenera suggests that this sediment was deposited in a highly stressful environment, which inhibited colonization by normal marine organisms. The individual *Teichichnus* burrows are large, suggesting that they inhabited normal marine saline and oxygenation conditions. Therefore, the stress was probably the result of the very high rates of sedimentation indicated earlier. The organisms that inhabited this substrate were opportunistic species which colonized an environment few other organisms could. Facies 2.4 was deposited in a marine environment in which high sedimentation rates and high concentrations of suspended fine grained sediment inhibited the development of a diverse community of marine organisms.

This facies is interbedded with fossiliferous shaly and sandy limestones, which reflect periods of low sediment input. These may indicate periods of transgression, sediment entrapment onshore and colonization of the substrate by thick-shelled bivalves and other carbonate secreting organisms.

Interpretation

Facies 2.4, the *Teichichnus* beds, were only found in one core (11-28-38-5w5), but were persistent through most of this cored interval. This sediment reflects very high sedimentation rates, which the majority of marine organisms could not tolerate. This sediment is interpreted as being deposited in the very distal part of a delta complex where abundant fine grained sediment accumulated at rates which exceeded the system's ability to remove it. This area was a sufficient distance from the delta front that the affect of fresh water influx was not felt. Normal marine sediment-stressed conditions allowed opportunistic marine polychaete worms to colonize the high sedimentation environment and keep pace with the constantly rising sediment surface.

The other three facies occur just to the east of this well, in 10-35-38-5w5. Facies 2.1 and 2.2 indicate deposition in the distal delta front environment, where periodic floods deposited coarser grained sediment over fine grained sediment, and hence developed the interlamination and interbedding seen in these facies. Bioturbation in these facies indicates a brackish environment that would be expected along the distal delta front. In some cases the sedimentation stress was too great, and prevented any biogenic reworking of this sediment. Facies 2.1 is interpreted as delta front sandstones. This facies only occurs in one 5 ft. thick bed in this well and is overlain by sediments of the distal delta front facies. This suggests that the 10-35 well is the westernmost extension of the delta system and that the complete delta section lies to the east. There are a number of wells that penetrate the Rock Creek strata, east of the 10-35 well, but no cores of this interval were cut. The log character of the wells (as illustrated in 04-34-38-3w5) shows a thick series of interbedded clean sandstone and shale, which is unlike the thick sections of clean sandstone in most Rock Creek section logs. This suggests that a delta associated with the facies association 2, occurs just east of the two cored wells examined.

Facies Association 3. Limestones

Facies 3.1. (Plate 3.7b)

This facies consists of skeletal grainstones which consist predominately of broken, thin-shelled bivalves and echinoid plates. There is generally a high siliciclastic content which consists predominately of fine quartz sand and silt. In some cores, this facies becomes more micritic or more siliciclastic-dominated and may be gradational into the bioturbated sandstones of facies 1.3. These limestones are generally structureless and probably thoroughly bioturbated. In some instances, however, high angle planar cross-bedding was observed, indicating high energy conditions and the clastic nature of these deposits.

Facies 1.3 was probably deposited on an open shelf where high energy conditions and low terrigenous clastic influx allowed the accumulation of skeletal grainstones.

Facies 3.2. (Plate 3.7a)

This facies is less common than the previous one and consists of thick-shelled, randomly oriented bivalve shells, floating in a matrix of dark grey, organic-rich, shaly limestone or shaly skeletal limestone. The organic and shaly nature of this facies suggests a much lower energy depositional environment than that which deposited the skeletal grainstones. The thick shelled bivalves were probably introduced to the environment by storm resedimentation or were infaunal organisms that were not reworked and broken up in this low energy setting.

Relationship and Distribution of the Facies Associations

The distribution of the Rock Creek facies associations is shown in Figure 3.17. The Rock Creek Member sediments in this area consist predominately of the fine grained quartzose sandstone lithofacies assemblage. These sediments show a general gradation from east to west from the upper shoreface sandstones of facies 1.1 to offshore shelfal sandstones of facies 1.4. Further westwards this association grades into the limestone facies association and a full array of gradational deposits occurs between. In the more westward areas of sandstone deposition, there is also a gradation upwards from the limestone facies to the sandstone facies. The Rock Creek Member was therefore deposited predominately in a shoreface environment, probably associated with a strandplain system. This depositional system prograded out over a stable shelf, which was dominated by carbonate sedimentation. Both the paucity of fine grained siliciclastic detrital sediment, and the association of the limestone and sandstone facies, suggest that there was very little fine grained sediment available in the system. If there were, carbonate secreting organisms could not survive and lower energy deposits would be expected to contain high quantities of argillaceous sediment. Removal of fine grained sediment from the system was likely provided by strong longshore currents which transported suspended clay and silt along the coast.

The exception to the general lack of lack of fine grained sediment is

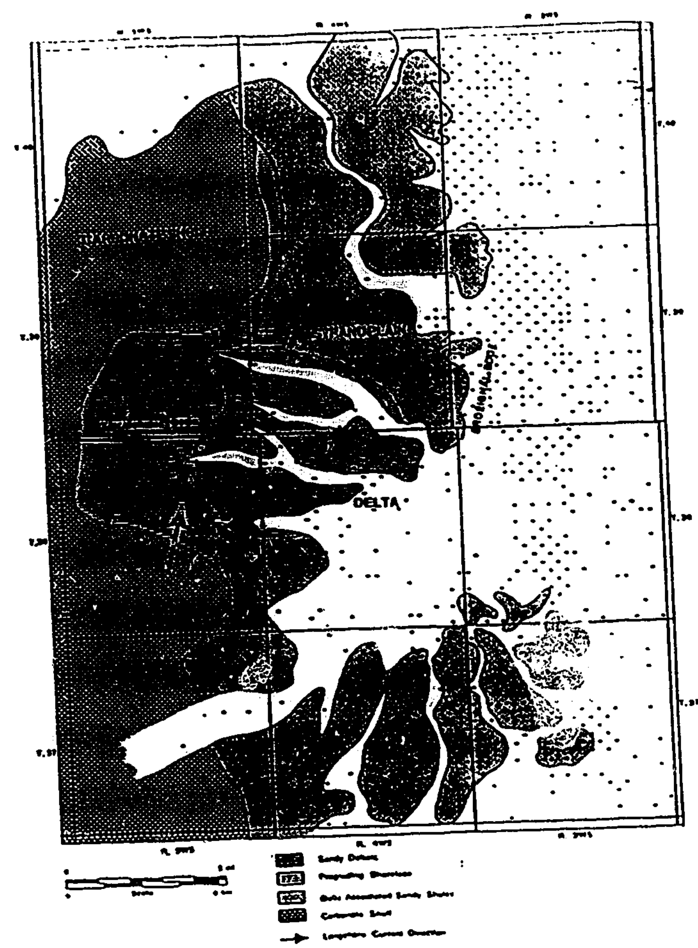


Figure 3.17 Interpreted depositional trends within the Middle sequence of the Rock Creek Member.

found in the Medicine River area, and is associated with the delta complex. Here abundant fine grained detritus is associated with delta front deposits. Although longshore currents were still active in this area, their effect was minimized by the extreme rates of fine detrital input from the delta system. In an area of such high discharge of fine suspended sediment, deposition of mud is more reflective of the concentration of the suspension than of the energy level (Clifton, 1986). In areas just to the north of the delta complex (eg. 06-32-39-4w5) the affect of this sediment input was still felt as the lower shoreface deposits in this area are muddier than in equivalent sediments further away from the delta. The shaly sandstone deposits, as interpreted from logs, appear to extend northward from the delta. To the south, in the Sylvan Lake area, there is little evidence of shaly sandstones and the effect of the delta appears to be minimal. Here shoreface sandstones interfinger directly with shelfal, fossiliferous limestones. These relationships suggest that the longshore currents responsible for the removal of the fine grained detritus flowed from north to south.

DISCUSSION

DEPOSITIONAL SYSTEMS

The Rock Creek Member was deposited as a mixture of carbonate and siliciclastic facies on a stable shelf setting. Although the literature is not extensive on such mixed sediment deposits, several examples of modern and ancient analogues can be found with similar characteristics to the Rock Creek Member sediments.

The modern day and relatively recent (Quaternary) eastern coast of Brazil in South America provides a good modern analogue for Rock Creek depositional systems. Here a stable continental shelf, up to 100 km wide, has been in existence since early Cretaceous rifting separated South America from Africa (Dominguez et al., 1987) and opened the Atlantic ocean. The Rock Creek Member was deposited on a similar passive margin which was even older (with respect to Middle Jurassic time) than this and is suggested to have been at least 20 to 500 km wide (Bovell, 1979). The Brazilian Shelf is a low angle feature of nearly constant depth (60-80m) with the outer shelf dominated by carbonate (reef) deposition and the inner shelf (<30m) by terrigenous sand deposition. The coastline is punctuated by several emerging drainage systems which build wave dominated deltas and their associated strandlines out onto the carbonate shelf. The system is dominated by longshore currents which transport most of the river-borne sediment along the coast, removing most of the mud from the system. Removal of this mud allows for development of a carbonate 'factory' to operate on the shelf and deposit largely carbonate sediment below 30m depth. The depositional system envisioned for the Rock Creek Member fits neatly to that of the Caravelas delta system (Figure 3.18).

In the wave-dominated deltas of the Brazilian coast there is an asymmetry of facies associated with the longshore drift. The downdrift sands are generally more poorly sorted and angular than the updrift sediment. Angularity of the Rock Creek Member framework grains does not appear to change across the Medicine River delta, probably due to the recycled nature of the sand grains. The Brazilian sands are derived from igneous and metamorphic sources and are therefore less mature. There is, however, a decrease in the sorting of the Rock Creek sandstones to the north of the delta system, since the clay content of the sediment shows a

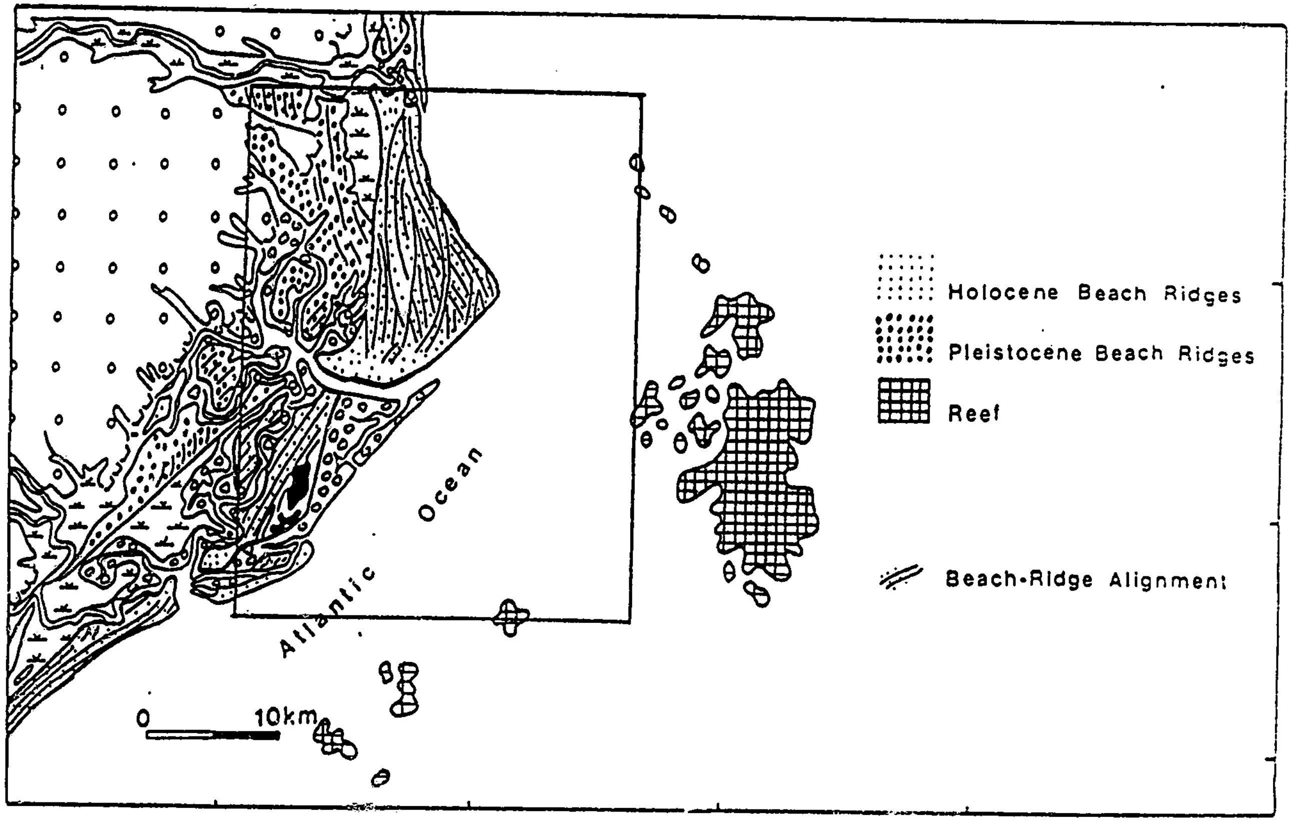


Figure 3.18. Geologic map of the Caravelas beach-ridge plain along the eastern coast of Brazil. This diagram shows the surface expression of two separate depositional sequences (Pleistocene and Holocene), as they are associated with a wave and longshore current-dominated delta system. Superimposed on this diagram is the relative size of the study area, indicating the similar scales of the this system to that described for the Rock Creek Member. (Modified from Domineguez et. al., 1987).

marked increase. This suggests a south to north current direction. Concomitant with this is the tendency for carbonate-dominated sediment to appear further eastwards, south of the delta system, in response to the lack of suspended fine grained clastic material.

An interesting feature of the Rock Creek delta system is the development of the silty and shaly sediments of the *Teichichnus* dominated facies in front of the Medicine River delta. The delta front sediments of the Brazilian coast were not discussed by Dominguez et al., (1987), but Clifton (1986) discusses the accumulation of mud on high energy shelves. In areas where great concentration of silt and clay are discharged from a river, as in the case of the large Amazon drainage system, accumulation of fine grained sediment becomes a function of concentration of sediment rather than the level of energy (Clifton, 1986). Along the shelf adjacent to this large drainage system, mud accumulates rapidly on the shelf even though the energy systems are much the same as in the case of the Calveras delta. The fluvial system which discharged into the Medicine River area during the deposition of the Rock Creek Member probably discharged a high concentration of fine grained sediment. At such high rates of influx the prevailing energy system was not able to redistribute all of this fine grained material and the delta front area was one of very high rates of silt and mud deposition. Such high rates of sedimentation and high concentrations of suspended fine sediment made this a very harsh environment for most organisms to inhabit. Suspension-feeding organisms could not cope with the suspended particles, which would have clogged their feeding apparatus, and deposit-feeding animals were generally not able to keep pace with sedimentation rates. Only a few specialized organisms were able to keep pace with sedimentation, and since these specialized opportunistic species thrived they dominated the fabric of the sediment in the delta front area.

An ancient analogue to Rock Creek deposition is provided by the Mississippian mixed carbonate/siliclastic deposits associated with the Damme Field, in southwestern Kansas. Shelf sedimentation was dominated by skeletal wackestones and ooid grainstones, reflecting the varying energy conditions across the shelf (Handford, 1988). With the influx of siliciclastic sediment, as a river debouched into the sea, progradation of the delta system and its associated strandplain occurred (Figure 3.19). This system, like the modern example, must have been dominated by longshore currents which removed any carbonate inhibiting, fine grained clastics, from the system.

A similar association of facies was described from the Rock Creek Member 100 km to the northwest by Marion (1984). These were described as being deposited on a stable, shallow marine shelf on which the depositional patterns were dominated by storm processes and tidal currents. No evidence was found for either of these processes in this study, although the description of the facies in both areas appear to be similar. Marion (1984) described burrowed silty sandstones, and illustrated large *Teichichnus* ('megaburrows') dominated sediments. This is identical to the *Teichichnus* beds of this study and it is suggested that the distal delta front facies, described here, also occurs to the north. This indicates that the sediments described by Marion (1984) were also deposited in a wave dominated shoreline setting, in delta and shoreface systems, rather than out on the shelf.

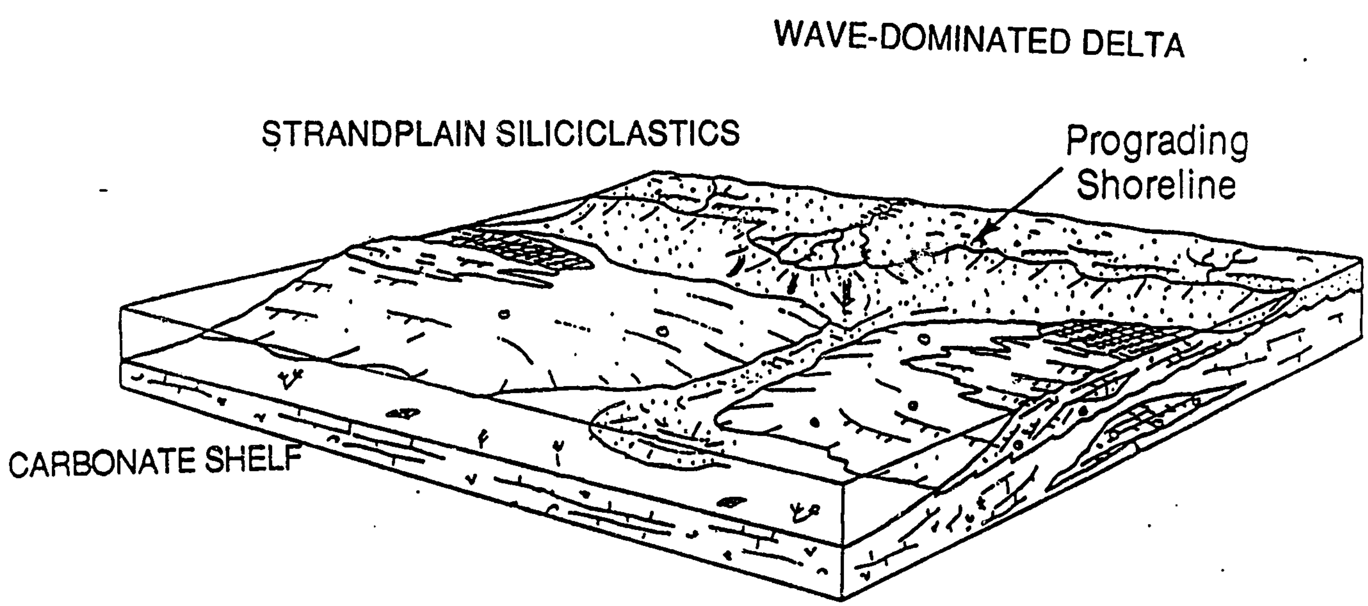


Figure 3.19. Block diagram illustrating the facies relationships interpreted from the mixed carbonate/siliciclastic system in the Damme Field, southwestern Kansas (after Handford, 1988).

SEQUENCE STRATIGRAPHY

The stratigraphic architecture of the Rock Creek Member consists of three unconformity-bounded deposition sequences (Figure 3.10), as used in the sense of Brown and Fisher, (1977), Vail e. al., (1977), Van Wagoner et al., (1987). Each sequence is bounded above and below by a zone of weathering and pedogenic alteration, and in rare instances by a single pebble layer derived from the underlying sequence. Generally, sequence boundaries separate stacked parallic sandstones and are often difficult to identify. In rare cases (14-19-40-4w5) these unconformities are overlain by a structureless fissile marine shale, which has all the attributes of a marine condensed horizon. Such a deposit marks the zone of maximum transgression. Rare opportunistic dinoflagellates in this shale indicate a stressed environment, probably due to decreased oxygen levels associated with the deposition of condensed sections (Galloway, 1989).

Each successive sequence erosionally truncates the previous sequence and oversteps it landward, creating an onlapping succession of stacked shoreface sandstones. Although only extensive examination of cores from the middle sequence was possible, it appears that the depositional mechanisms differed little from sequence to sequence. In the area of the Medicine River delta complex the entire Rock Creek interval appears to consist of interbedded sands and shales of probable deltaic origin. Since no cores were available through this interval it is difficult to define any sequence boundaries. However, the persistence of the log characteristics through the interval suggest that this area may have been the focus of fluvial input throughout the Rock Creek interval, and there may be several sequences present in this complex.

This is similar to the Brazilian coast where several wave-dominated deltas, associated with different episodes of eustatic movement are stacked in the same geographic area. For instance, two successive episodes of wave-dominated delta progradation occurred during the Quaternary in the area of the Calaveras delta (Figure 3.18).

Although it is difficult to correlate this area to the global sea level curves (Figure 3.20) of Haq et al., (1989) because of the imprecise nature of the biostratigraphic data, some general comparisons can be made. There appears to have been a major sea level lowstand in the early Aalenian stage which was probably responsible for the major sequence-bounding unconformity produced by erosion of the Toarcian age Poker Chip Shale.

The Middle Jurassic global curves suggest several third order cycles of coastal o[illegible] two are shown to affect the Aalenian to Bajociar[illegible]gests either a tectonic control on the Rock Creek seq[illegible]hat these are 4th order cycles that are not illu[illegible]988), curves. The latter is likely the case as the [illegible]not known for its active tectonic regime and as thi[illegible]imal to the depositional edge of the Rock Creek Member, higher-order eustatic cycles are likely to be evident.

SUBSIDENCE

The Middle Jurassic shelf lay on a stable and very old passive margin, which had persisted through most of the Paleozoic and early Mesozoic time (Bovell, 1979). In such a setting, subsidence levels are a function of lithospheric cooling and sediment loading (Jervey, 1988), both of which would have been relatively low on such an old continental margin (Cloetingh, 1988). Typically, basinward-increasing subsidence combined with eustatic rises and falls in sea level, are responsible for the general

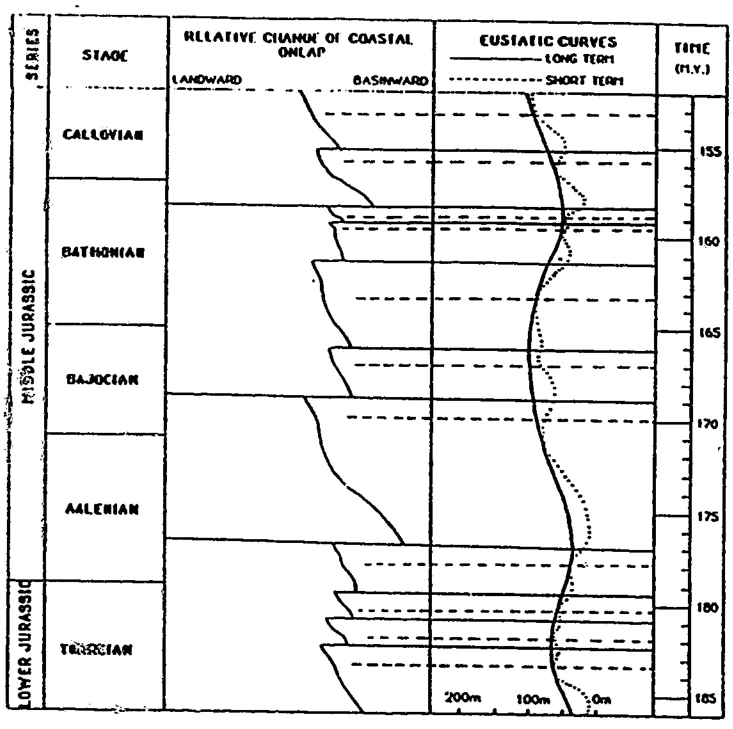


Figure 3.20. Diagram showing the global coastal onlap curves and the interpreted eustatic curves associated with the upper part of the Lower and the Middle Jurassic Series. (after Haq, et al., 1989).

onlapping nature of the Rock Creek Member. This onlapping sequence architecture is illustrated in a west-east cross section, across township 40 (Figure 3.10).

Accommodation is a function of sea-level fluctuation and subsidence (Jervey, 1989). Preservation potential of sediments on such a stable, slowly subsiding continental margin is relatively low as it is dependent upon the system's ability to accommodate that sediment. Low levels of subsidence, and therefore accommodation, explain the thin nature of the Rock Creek sequences and the erosive nature of the onlapping architecture.

In several localized areas, the simple stratigraphic architecture of onlapping Rock Creek sequences was interrupted. This was particularly common east of the Mississippian axial high. The Sylvan Lake Field lies just east of this high and the architecture of the sequences is illustrated in cross-section B-B' (Figure 3.10). The area west of the axial high is characterized by the typical onlapping nature of the sequences, and all but the uppermost sequences were eroded over the highest point of the post-Mississippian unconformity surface. East of this high, the lower sequences were again developed in localized low areas on the post-Mississippian unconformity and a dramatic thickening of the Rock Creek Member occurs in this area. Preservation of these sequences appears to have been a function of increased subsidence during the Aalenian and Bajocian stages. Depositional trends are independent of these lows and therefore the low areas do not appear to have been present before the deposition of the Rock Creek Member. In areas of higher than regional subsidence levels, the lower sequences of the Rock Creek Member were brought below the base level and were therefore spared from the extensive erosion that removed correlative sediments from the highs. As subsidence in the Sylvan Lake area continued through the period in which the Rock Creek Member was deposited, sedimentation kept pace with the subsidence and therefore little affect was present on the depositional environments.

This anomalously high subsidence appears to have begun after the deposition of the Poker Chip Shale as this unit is not found in low areas of the Sylvan Lake Field. The base of the lows is formed by highly fractured dolomites of the Mississippian strata with sandstone and green shale filling the cavities. These deposits are generally overlain by a highly altered green shale with chert and dolomite breccia horizons, typical of the detrital zone. If subsidence pre-dated the Poker Chip Shale thicker accumulation of the Poker Chip Shale would be expected in the lows. Continued subsidence appears to have directed lowermost Cretaceous drainage patterns, resulting in the incision of the Rock Creek Member and lateral juxtaposition of Ellerslie sediments against Rock Creek sandstones. This is responsible for much of the stratigraphic confusion that has affected mapping attempts of the clastic reservoirs in the Sylvan Lake gas unit.

Continued subsidence throughout the the deposition and lithification of the Rock Creek sandstones explains the highly altered and fractured nature of these deposits in the Sylvan Lake Field (Plates 3.3a-c) The structure in Plate 3.3a appears to be a fracture system that developed while the sediment was still relatively unlithified and therefore affected a relatively wide zone (1 in.) within the sediment. Conversely the fracture systems in Plates 3.3b and c occur as very narrow (mm scale) fractures that appear to have occurred while the sediment was relatively rigid and therefore well lithified. The ongoing structural deformationof the sediments was responsible for the destruction of most of the evidence of biogenic and physical sedimentary structures in these sandstones and therefore their characteristic massive appearance.

Greater than regional subsidence rates, also affected the Medicine River-Gilby area. The development of the Medicine River-Gilby subsidence basin was discussed in the previous paper. This area began subsiding after the deposition of the first Nordegg Member sequence and therefore pre-dates the subsidence that affected the Sylvan Lake area. Subsidence in this sub-basin continued into Middle Jurassic time, as indicated by the anomalous thickening of the Rock Creek Member in this area. Subsidence levels were not as substantial as in the Sylvan Lake Field and therefore only thin intervals of the Rock Creek Member are preserved. Other evidence of anomalous subsidence in this area is the preservation of deeper water sediment in the Rock Creek Member in 02-30-39-3w5. Subsidence appears to have brought offshore sediment below the base level in an area where such deposits are generally eroded. Wells to the east of this well contain only upper shoreface sediment preserved in the Rock Creek Member.

In summary, the architecture of the Rock Creek was not only influenced by the depositional systems and eustatic sea level variations that operated through Aalenian and Bajocian time, but were strongly affected by the rates of subsidence on the shelf and therefore the accommodation ability of this shelf. Regional subsidence rates were very low and therefore the entire Rock Creek interval is one of highly erosive processes. Preservation of any one sequence may be patchy depending upon the subsidence regime.

CONTROLS ON HYDROCARBON PRODUCTION

In the Sylvan Lake Field the Rock Creek Member has produced a total of $251x10^3$ m^3 oil and $744x10^6$ m^3 gas (up to 05/1987). The best gas well in the field (16-29-37-3w5) has produced 80% of the gas in the field, at average rates of $108x10^3$ m^3/D. There are several wells in the field which have not been completed to produce Rock Creek gas, but which have tested at rates greater than $176x10^3$ m^3/D, and as much as $317x10^3 m^3/D$. Oil production in the area is sporadic with many low-rate oil wells present. There are, however, several wells which produce from Rock Creek sandstones at their allowable of $11 m^3/D$. In addition, numerous low rate wells are producing oil from the Rock Creek Member along the subcrop edge in the Medicine River Field (section 13-39-4w5). In the Gilby Field there has been no production from the Rock Creek Member, but drill stem tests reveal that the Rock Creek subcrop contains several potentially prolific gas bearing reservoirs. Rock Creek sandstones in this area have tested at gas rates in the order of $388x10^3$ m^3/D.

Reservoirs within the Rock Creek Member are restricted to the sandstones of facies 1.1 and 1.2 . Because the Rock Creek sandstones were deposited as sheet-like sands in prograding shoreline environments, there is little opportunity for stratigraphic-trapping mechanisms within the Rock Creek itself. The entrapment of the hydrocarbons in this area therefore relies on erosional truncation of porous sandstone units and the development of structural closures, downdip from the truncation points.The play fairway in the Rock Creek member has therefore been primarily focused on the subcrop traps developed beneath the pre-Cretaceous unconformity. This study has demonstrated an onlapping sequence architecture, where sandy units are truncated beneath successive sequences, over the eastern Mississippian axial high. This opens the possibility for a new, and more subtle play type; one which chases erosional truncations within the Rock Creek Member itself. In addition, the presence of several separate sequences suggests that there may be more than one shoreline-related sand trend within the Rock Creek Member. Future subtle traps may be developed where such sand trends are truncated by the narrow, deep, Lower Cretaceous valleys, which cut through the Rock Creek Member. Bends in the path of the incised valley features would provide structural closures in the Rock Creek sandstone reservoirs, while the shaly and impermeable valley fill sediments would provide the lateral seal.

In addition to providing an adequate cap rock and lateral seals, the pre-Cretaceous unconformity also appears to have enhanced reservoir conditions in the Rock Creek Member. Sandstones in the Rock Creek are generally tightly cemented with quartz overgrowths (Plate 3.9), and therefore porosity and permeability are generally too low for economic extraction of hydrocarbons. Along the pre-Cretaceous subcrop edge of the Jurassic System, secondary porosity is well developed by the leaching of calcite cements, unstable grains (chert) and even quartz grains and their overgrowths (Plate 3.9). This effect is particularly enhanced in areas where fracture systems within the sandstone reservoirs acted as conduits for the leaching fluids, as in the case of the Sylvan Lake Field.

CONCLUSIONS

The Rock Creek Member in the Gilby, Medicine River and Sylvan Lake Fields was deposited in at least three sequences, each bounded above and below by a surface of subaerial exposure and erosion. This unit was deposited in three laterally transitional depositional systems. In the central portion of the area, a delta system deposited a succession of heterolithic deposits which contained abundant fine clastic detritus. This delta system was dominated by wave and longshore current energy, which redistributed much of the fine clastic material along the coast. Sand from this system was transported shorter distances and accumulated along the shoreface and beach system of the associated strandplain. Seaward, away from the influence of suspended clastic detrital material, the shelf area was dominated by carbonate sedimentation. Here pelecypod and echinoid grainstones accumulated in high energy areas and pelcypod wackestones and rudstones in lower energy settings.

The Rock Creek Member was deposited on an old and very stable passive margin where subsidence rates greatly affected the preservation of the deposits. In some areas, such as the Sylvan Lake Field, high subsidence rates allowed for the preservation of anomalously thick successions of Rock Creek sediment. Continued subsidence throughout the deposition of the Middle Jurassic sediments and into Lower Cretaceous time directed Lower Mannville drainage, and therefore erosion, through this field. The result was an area of complex stratigraphy in which Ellerslie and Jurassic sediments are juxtaposed, and therefore an area which is extremely difficult to map.

The main reservoir rocks in this unit are formed by the clean sandstones of the shoreface environment. In many areas these sandstones are tightly cemented so that the best reservoir conditions occur at the subcrop edge, where porosity and permeability are enhanced by secondary solution of cements and framework grains. In the Sylvan Lake Field continued subsidence caused pervasive fracturing in the Rock Creek sandstones which, combined with the proximity of this area to the pre-Cretaceous unconformity, created even better reservoir enhancement.

Future efforts in exploration of hydrocarbons in the Rock Creek Member should be aimed at mapping the separate sequences and predicting the shoreline sandstone trends within these individual sequences. Then mapping of Lower Cretaceous Valley systems and inter Rock Creek erosional surfaces will point to optimum locations for stratigraphic hydrocarbon traps within the Rock creek Member.

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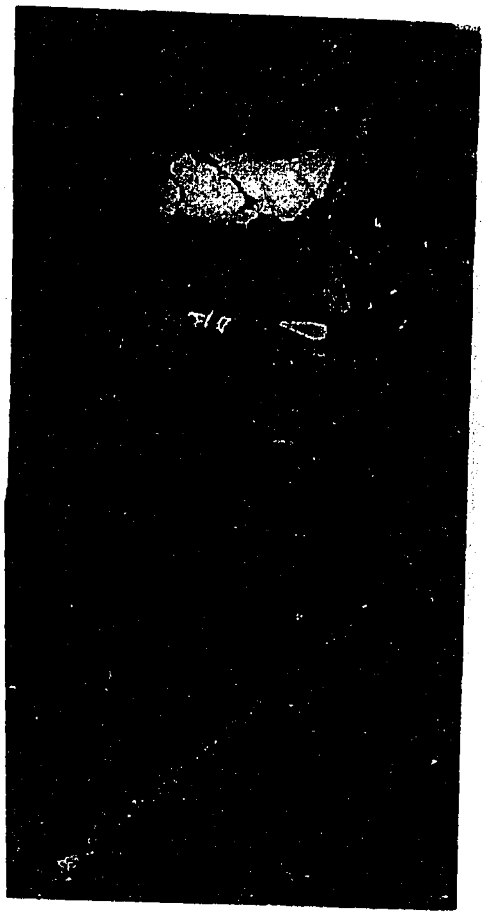
Plate 3.1. Detrital lithologies.

Plate 3.1a. Chert conglomerate with rounded and weathered chert pebbles. Both matrix-and grain-supported textures as well as lenses of laminated fine grained sandstone. This deposit appears to be water transported and not an *in situ* weathering product. (16-13-37-4w5; 5-32 m)

Plate 3.1b. Green shale and angular dolomite clasts infilling a cavity in Elkton dolomite. This indicates *in situ* weathering at the post-Mississippian unconformity and infilling of solution cavities with detrital-type lithologies. (16-13-37-4w5; 5-30).

Plate 3.1c. Dolomite breccia with a clean sandstone matrix. This deposit is predominately grain-supported and the angularity of the dolomite clasts suggests little transportation. This deposit is interpreted as a solution breccia which was later infilled with fine grained sandstone filtering down from above. (16-13-37-4w5; 5-31).



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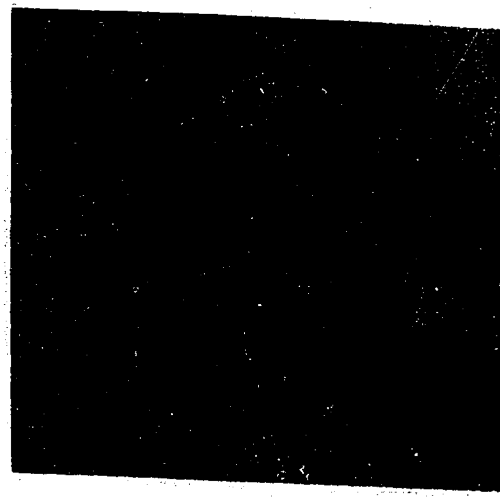


B

C

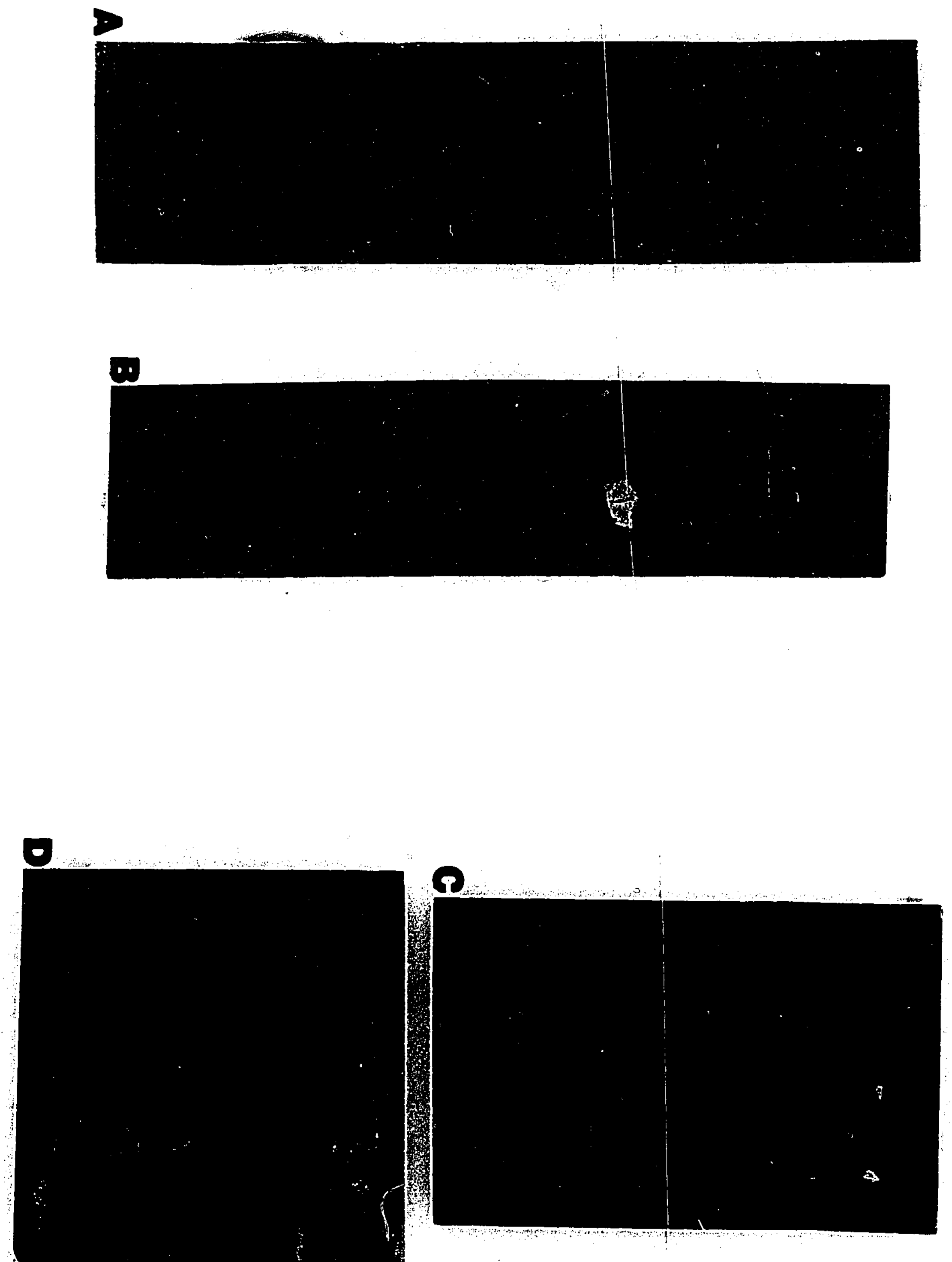
Plate 3.2. Sequence-bounding unconformity surfaces in the Rock Creek Member.

Plate 3.2a. Transgressive shale overlying possible sandstone pebble conglomerate and surface of pedogenically altered sandstone. (14-19-40-4w5; 2346m).

Plate 3.2b. Clean shoreface sandstone lying in sharp contact of the pedogenically altered zone of the underlying sequence. (14-19-40-4w5; 2355m).

Plate 3.2c. Unconformity surface between two Rock Creek Member sandstones. Surface is marked by a sharp contact with a pebble (center) of lithified sandstone, derived from the underlying sequence. Note the depositional drape of overlying sediment around the sandstone pebble. (10-35-39-4w5; 10-21m).

Plate 3.2d. The unconformity surface is composed of a veneer of pebbly sandstone overlying deeper water shelf shales and siltstones. (11-28-38-5w5; 17-28m).


A
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D

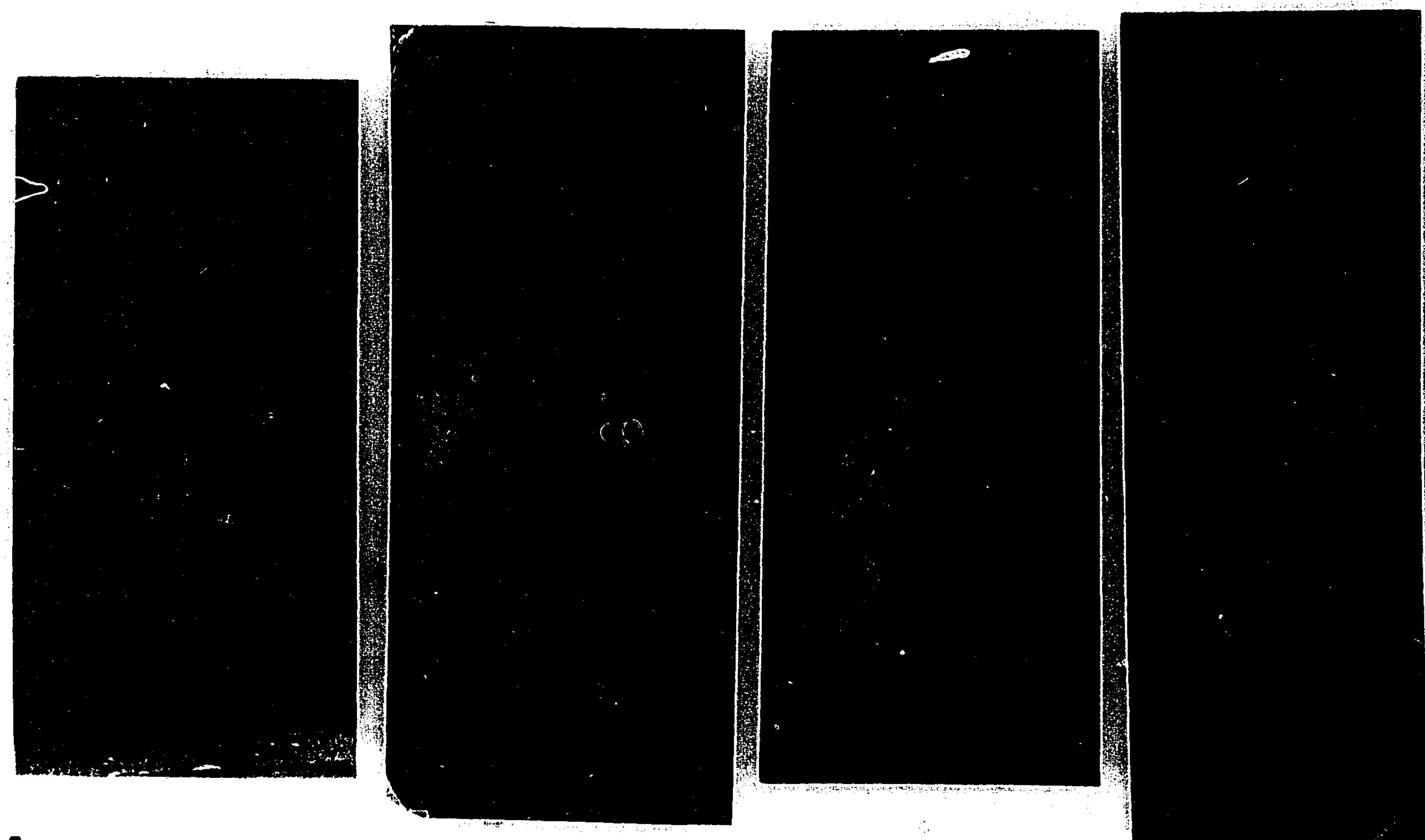
Plate 3.3. Typical fractured and weathered deposits associated with the Sylvan Lake Field Rock Creek sediments.

Plate 3.3a. Variously oriented fracture systems cross-cutting one another in a clean quartzose sandstone. Right-dipping fractures appear to offset, and therefore postdate, left-dipping fractures. (14-09-37-03w5; 2-17).

Plate 3.3b. Horizontal and obliquely dipping minor fractures cut by a 1cm wide, steeply dipping zone of system bifurcating and amalgamating fractures. (04-29-37-03w5; 5-27).

Plate 3.3c. Steeply dipping, wide band of apparent deformation due to vertical movements in a semi-consolidated quartzose sandstone. The sediments appears to have been somewhat soft as the band of deformation is wide and contains indications of grain flowage. (10-36-37-4w5; 5-27).

Plate 3.3d. Typical appearances of a weathered, highly quartzose sandstone, which occurs just below the pre-Cretaceous unconformity and separating sequences of stacked Rock Creek sandstones. (14-09-37-3w5; 10-15).



A B C D

Plate 3.4. Clean quartz arenite sandstones (Facies 1.1) associated with the Rock Creek, upper shoreface environment.

Plate 3.4a. Planar cross-bedding (14-19-40-4w5; 1-10)

Plate 3.4b. Massive quartzose sandstones with *Ophiomorpha* ichnofossils made visible by differential oil staining; Sylvan Lake Field. (Slabbed view, 12-21-37-3w5; 18-22).

Plate 3.4c. Cross-section through core showing massive sandstone with rare *Ophiomorpha* and *Diplocraterion* ichnofossils. (14-09-37-3w5; 2-20).

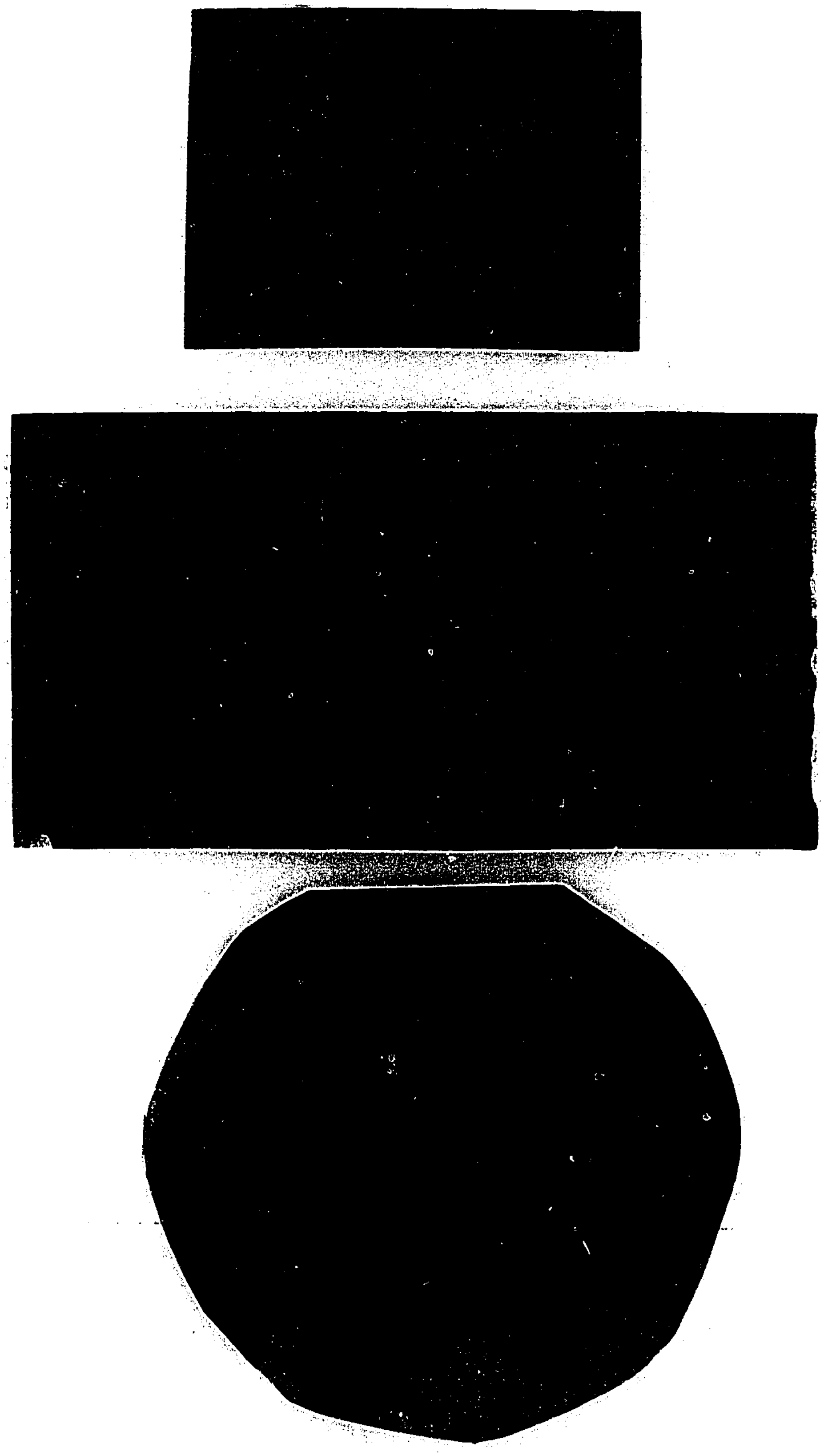


Plate 3.5. Lower shoreface and lower part of the upper shoreface facies (Facies 1.3 and 1.2 respectively).

Plate 3.5a. Lower shoreface shaly and heavily bioturbated sandstone. Note the abundance of horizontal burrows of deposit-feeding organisms (*Planolites* and rare *Teichichnus*). (11-28-38-5w5; 17-26).

Plate 3.5b. Upper part of the Lower shoreface with *Rosselia* (r), *Teichichnus* (t) and *Palaeophycus* ichnofossils. (06-20-40-4w5; 14-16).

Plate 3.5c. Lower part of the upper shoreface where abundant shale drapes on symmetrical ripples and bidirectional ripple lamination are present. Note the faint traces of *Ophiomorpha* ichnofossils. (06-05-40-5w5; 18-11).

Plate 3.5d. Ripple lamination and symmetrical mud drapes on ripple forms grade up into planar cross-bedded sandstones of the upper part of the upper shoreface facies. Note the general lack of bioturbation. (14-19-40-4w5; 10-15).

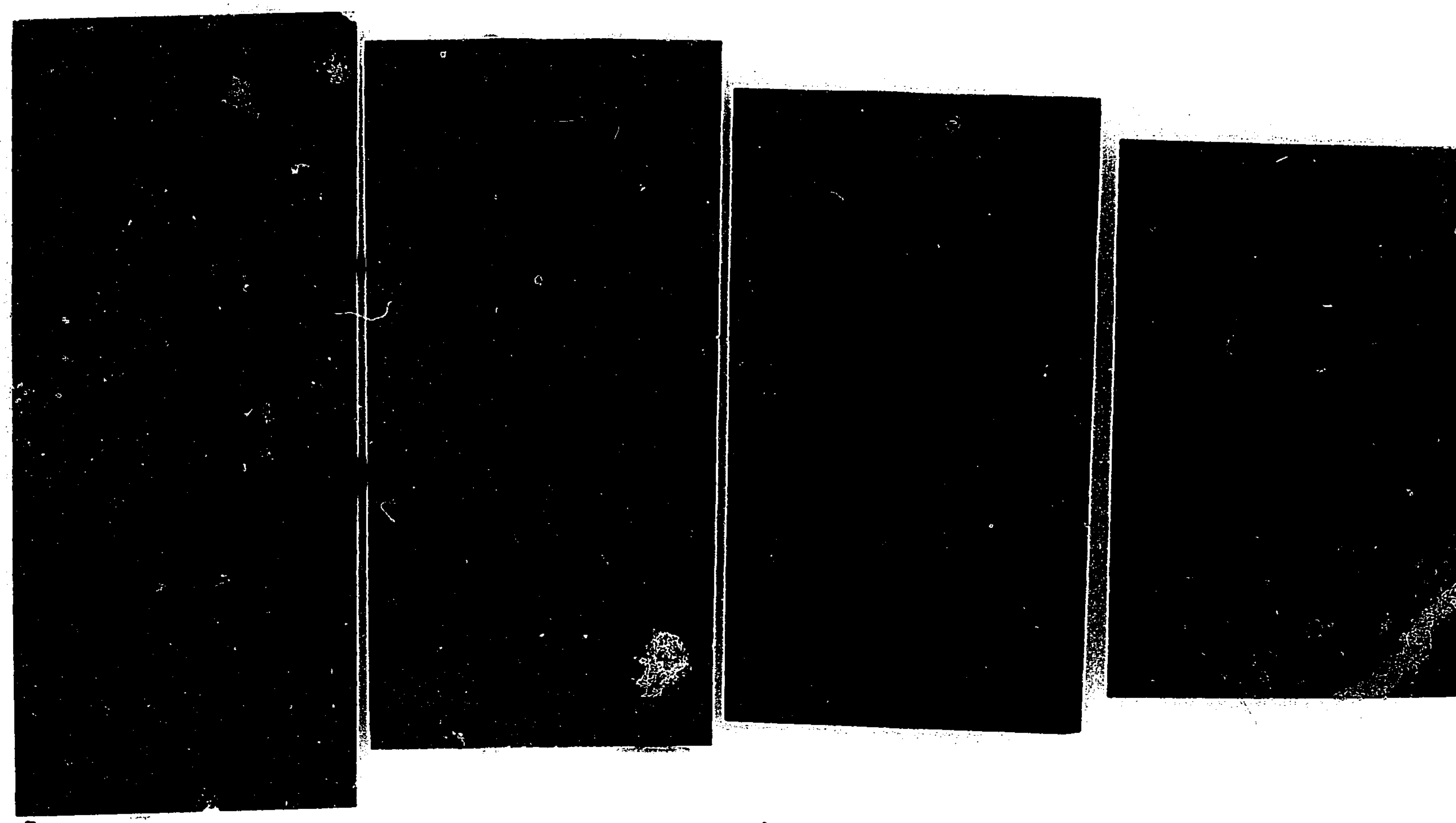

A B C D

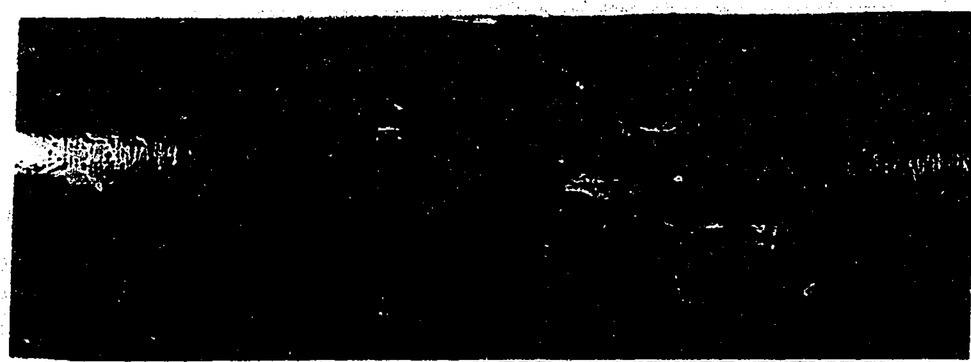
Plate 3.6. Rock Creek member carbonates and offshore siliciclastics.

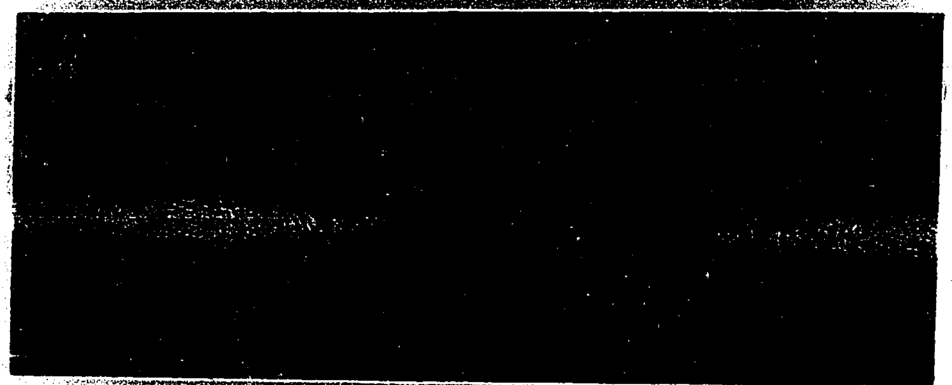
Plate 3.6a. Oyster shells in a sandy fine grained limestone matrix (14-04-40-04w5; 18-3).

Plate 3.6b. Coquinoid limestone. (06-32-39-04w5;14-4).

Plate 3.6c. Offshore muddy sandstone facies with abundant *Teichichnus* (t) and *Rosselia* (r) ichnofossils. (02-30-39-3w5; 10-34).

Plate 3.6d. Offshore muddy sandstone with *Planolites*, *Teichichnus* (t) and *Helminthopsis* (h). (14-19-40-4w5; 10-7).





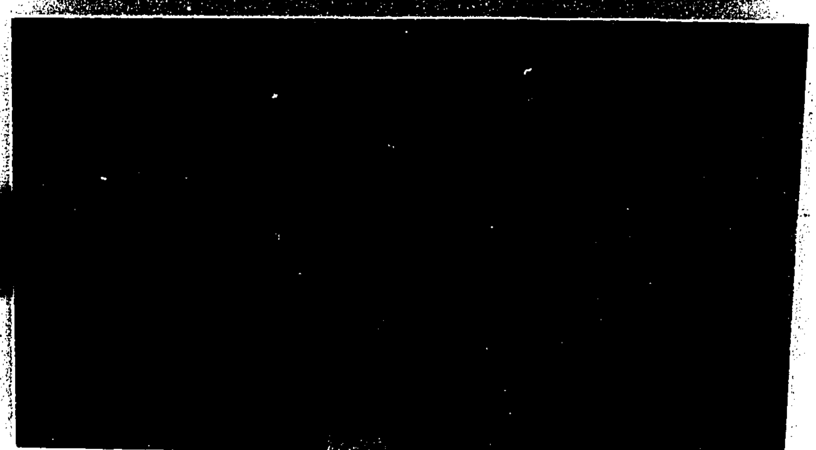


Plate 3.7. Heterolithic Facies Association.

Plate 3.7a. Sharp scour-based, planar cross-bedded sandstone with no evidence of bioturbation (Facies 2.1). (10-35-38-5w5;).

Plate 3.7b. Fine grained sandstone with mm scale shale laminae and stylolites (Facies 2.1). (10-35-38-5w5; 16-5).

Plate 3.7c. Finely interlaminated fine grained sandstone (white) and silty shale (dark) (Facies 2.3); note lack of bioturbation. (10-35-38-05w5; 16-10).

Plate 3.7d. Interbedded fine sand and silty shale (Facies 2.2) with a restricted assemblage of ichnofossils consisting of *Planolites* (p) and *Skolithos* (s). Rare escape structures (e) signify rapid rates of deposition. (10-35-38-05w5; 16-7).

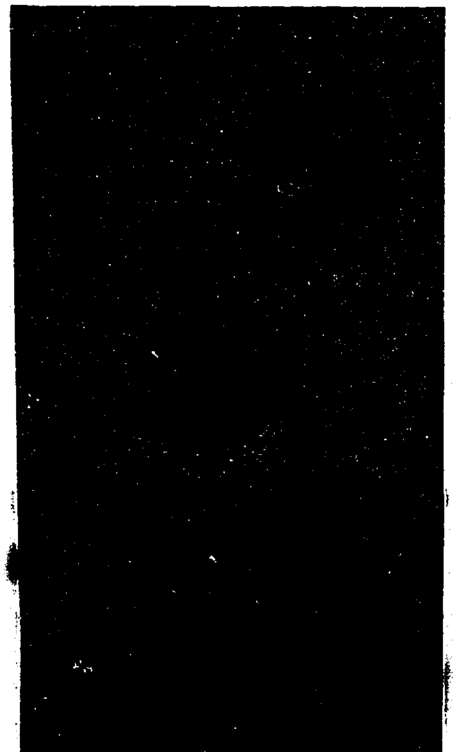
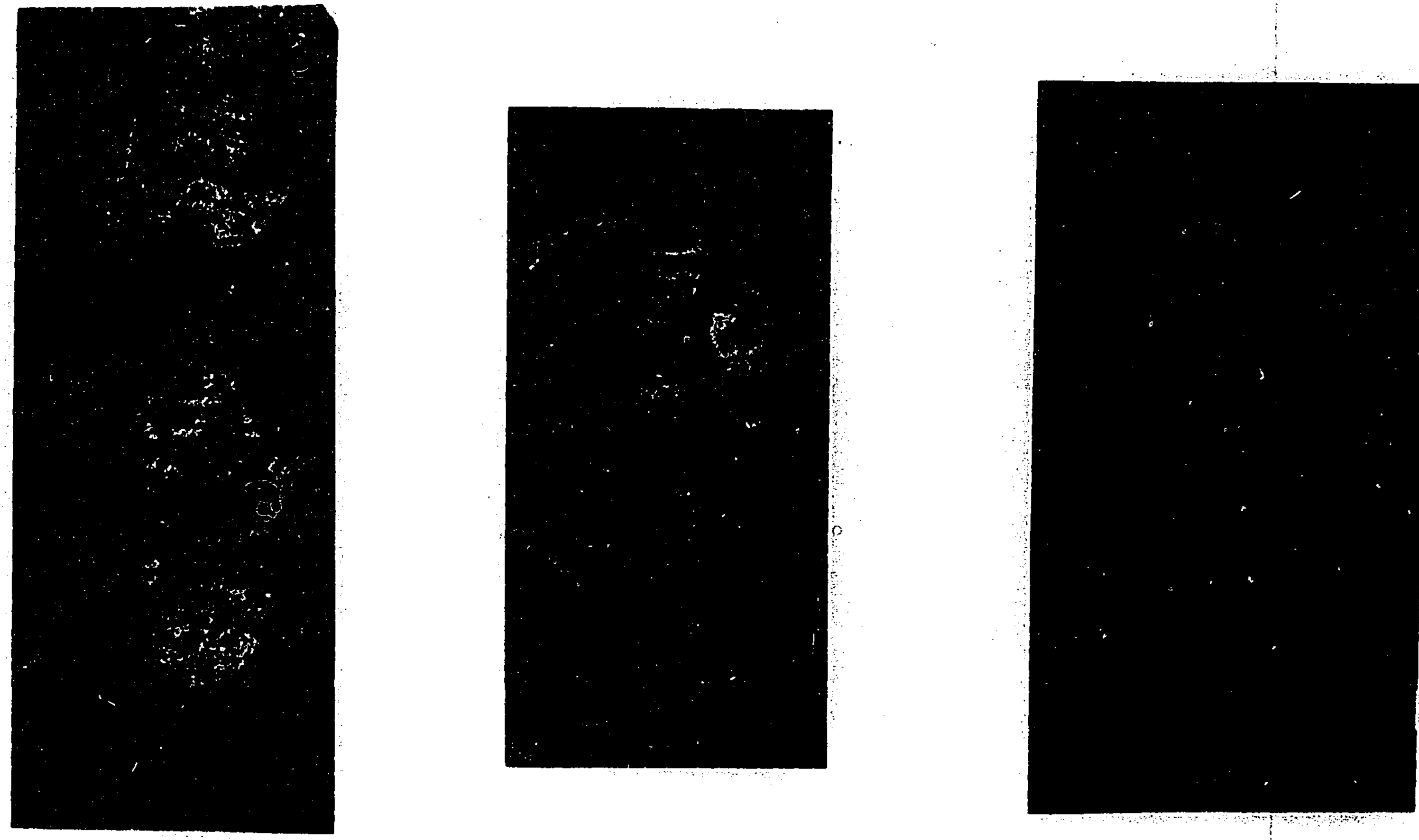



Plate 3.8. *Teichichnus* beds; Facies 2.43

Plate 3.8a and b. Sandy, silty shale with the texture completely dominated by large *Teichichnus* (t) trace fossils (Facies 2.4). The large size of the individuals suggests fully marine salinities while the long vertical height suggests very high rates of sedimentation. (10-35-38-05w5; 17-32 &17-31).

Plate 3.8c. Silty and sandy shales of facies 2.4 with large *Conichnus* trace fossil (c) again signifying very high rates of sedimentation. (10-35-38-05W5; 17-35).



A B C

Plate 3.9 Typical quartz arenite of the Rock Creek Member (14-19-40-4w5; 2351m). Note the predominance of this sandstone of detrital quartz grains. Silica cement occludes most of the pore space and leaves little effective porosity. A) PPL; B) cross-polars.

A

B

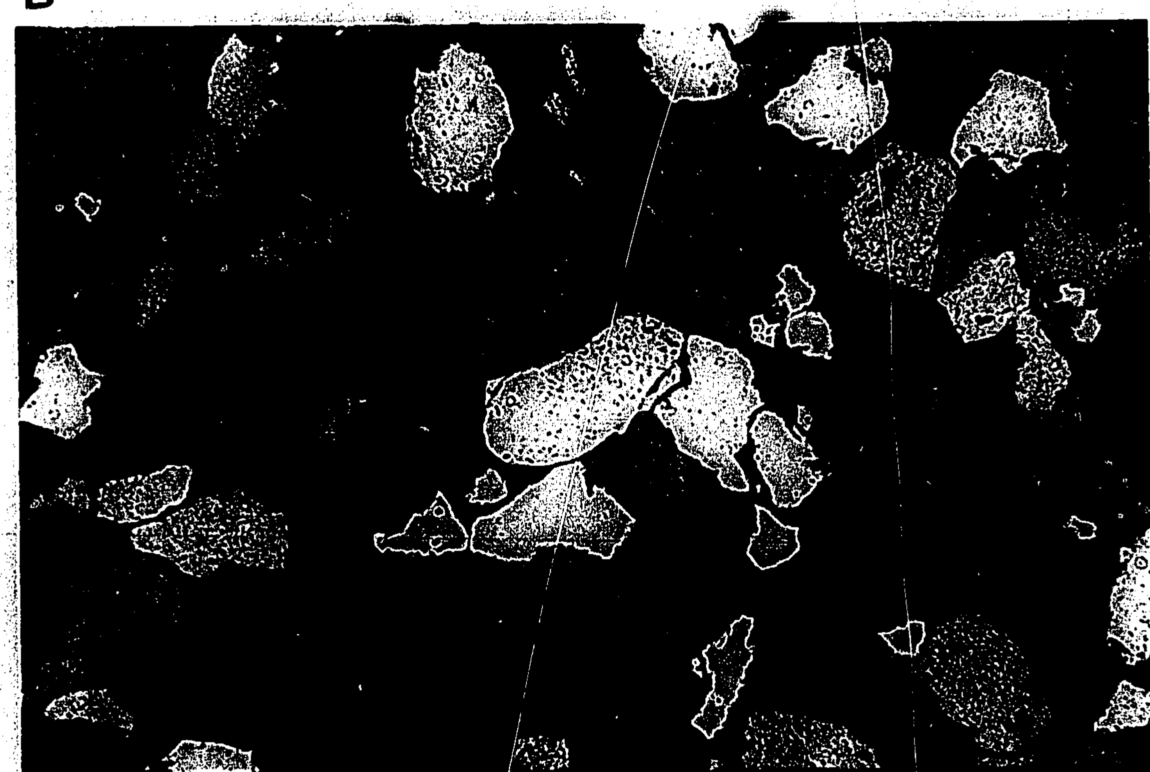


Plate 3.10 Rock Creek quartz arenite of similar composition to the previous slide (12-21-37-3w5; 2247m). In this case, extensive porosity networks are developed throughout the slide, providing a porous and permeable reservoir rock. Secondary solution porosity is indicated by the oversized pores (O), ragged grain boundaries (R), and remnant quartz (RM) shards floating in the pore spaces. A) PPL; B) cross-polars.

A

B

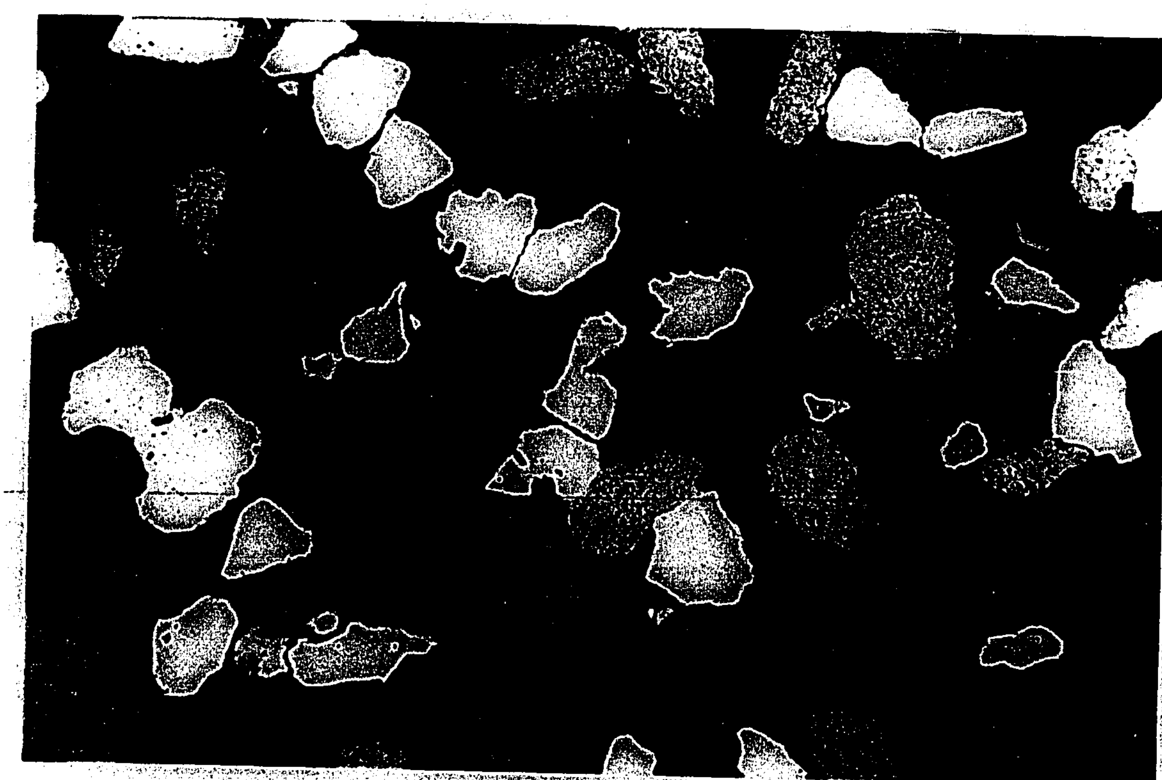
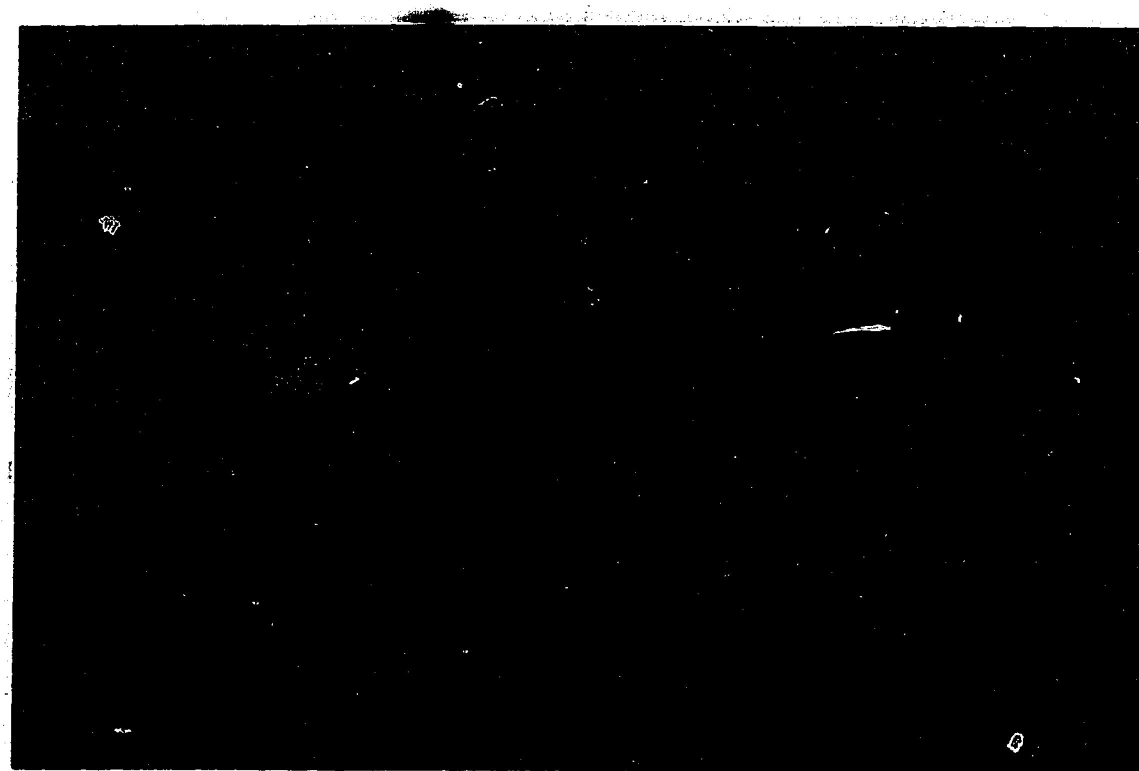


Plate 3.11a. Calcareous sandstone of the Rock Creek Member (10-36-37-4w5; 7378 ft.). The large calcite fragment in the center of the photo is an echinoderm plate. Porosity in this sample is completely occluded by calcite cement.

Plate 3.11b Calcareous sandstone of the Rock Creek Member (10-36-37-4w5; 7378 ft.). The phosphate grain (P) in the center of the photo is interpreted to be a fecal pellet. C = calcite cement; Q = quartz; E = echinoid plate.



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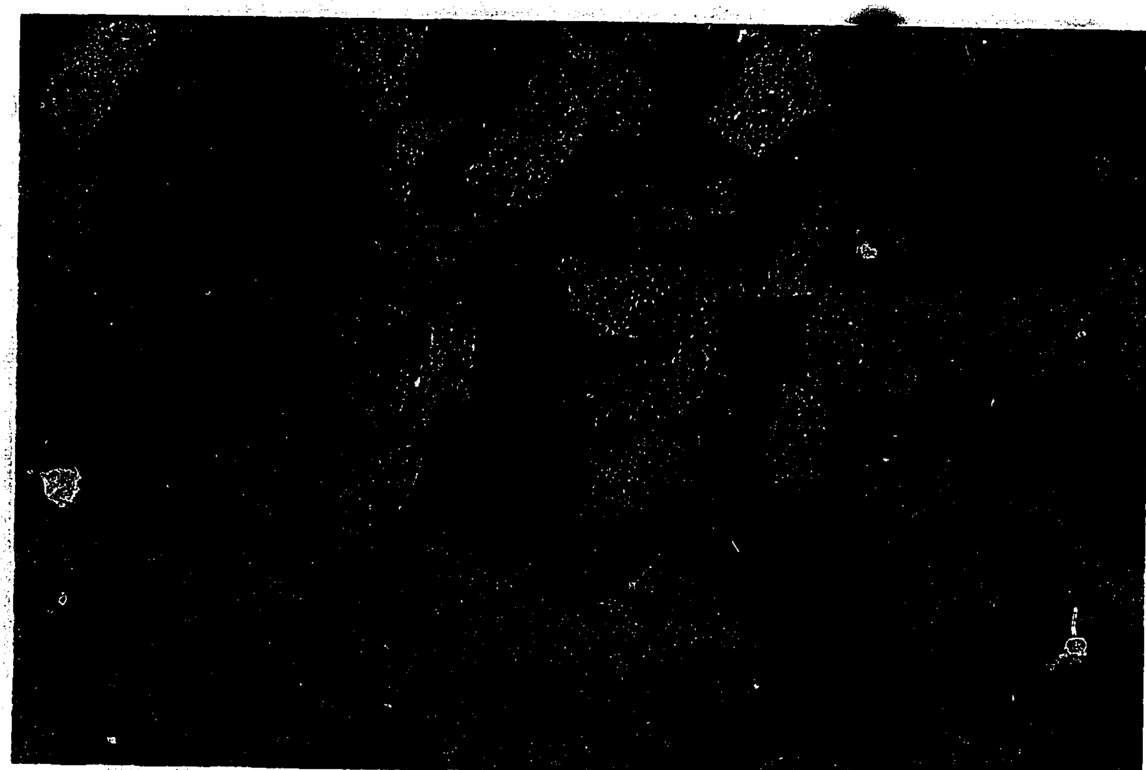
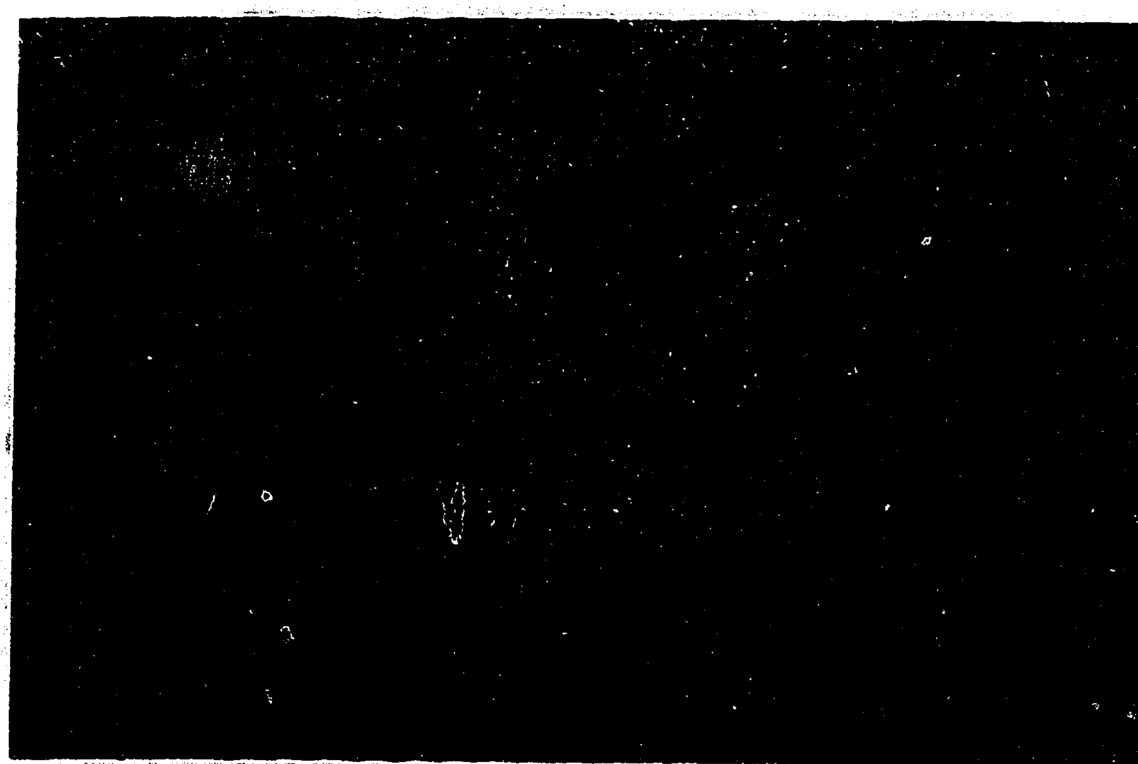


Plate 3.12a. Calcite cemented sandstone of the Rock Creek Member (11-28-38-5w5; 2430m). Note the abundance of phosphate ooids (PO) containing quartz silt nuclei. Porosity appears to have developed by the dissolution of calcite cement.

Plate 3.12b. Sandy bioclastic limestone of the Rock Creek Member (10-36-37-4w5; 7358 ft.). The predominant grains in the slide are large oyster fragments (OY), Crinoid fragments (CR), and fine grained quartz sand and silt (Q).



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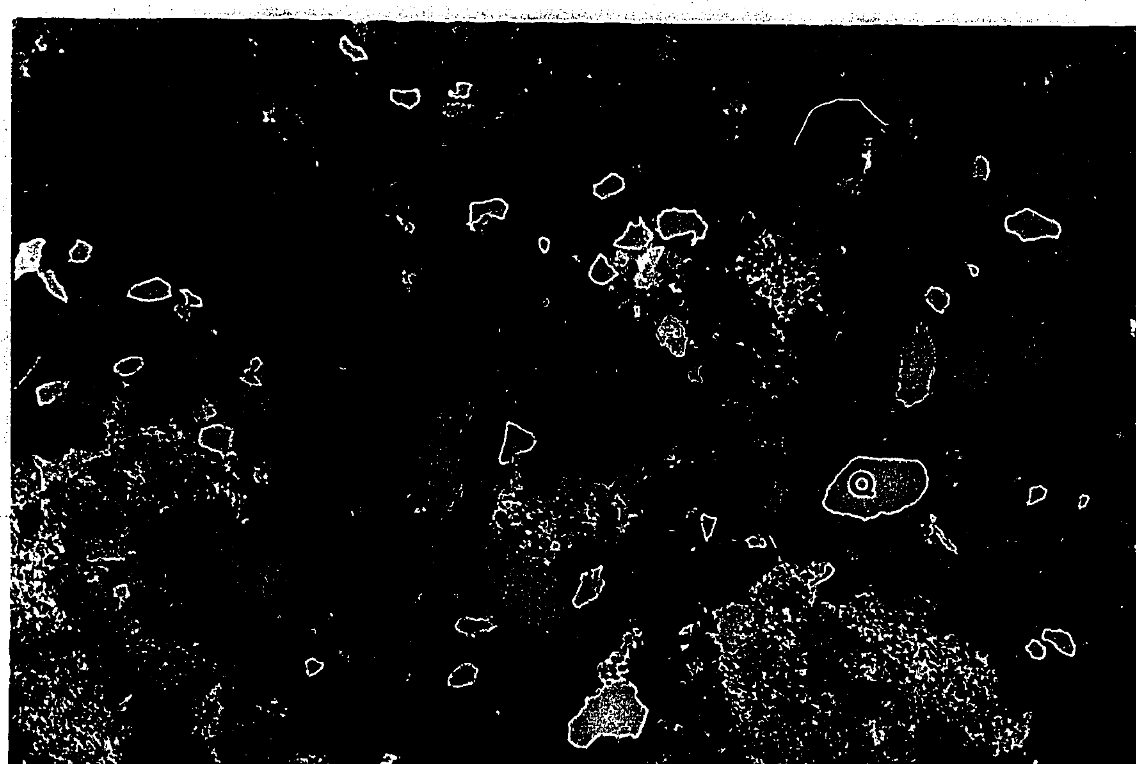
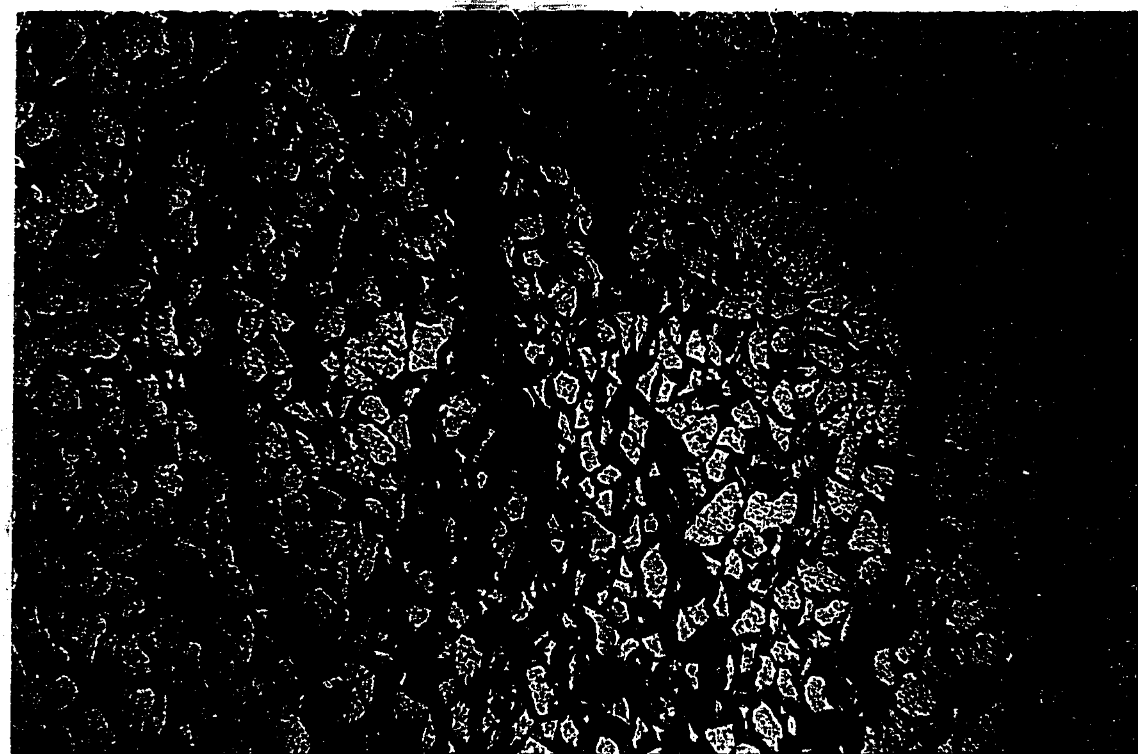


Plate 3.13a. Quartz arenite sandstone of the Rock Creek Member (14-09-37-3w5; 2286m). The quartz grains in this sample are highly corroded and the pore space filled with clay sized material. This slide shows the alteration of typical quartz arenites of the Rock Creek Member, which have been weathered and leached at an unconformity. The porosity was infilled with pedogenically emplaced clays. Laminations in the slide are interpreted as clay filled fractures in the sandstone.

Plate 3.13b. This slide shows the type of sediment found in the mottled horizons associated with the Rock Creek Member (12-21-37-3w5; 2245m). Quartz grains are highly corroded and leached and porosity is infilled with pedogenically emplaced clays. DQ is an odd shaped quartz grain illustrating the dissolution process which shaped this grain.



A

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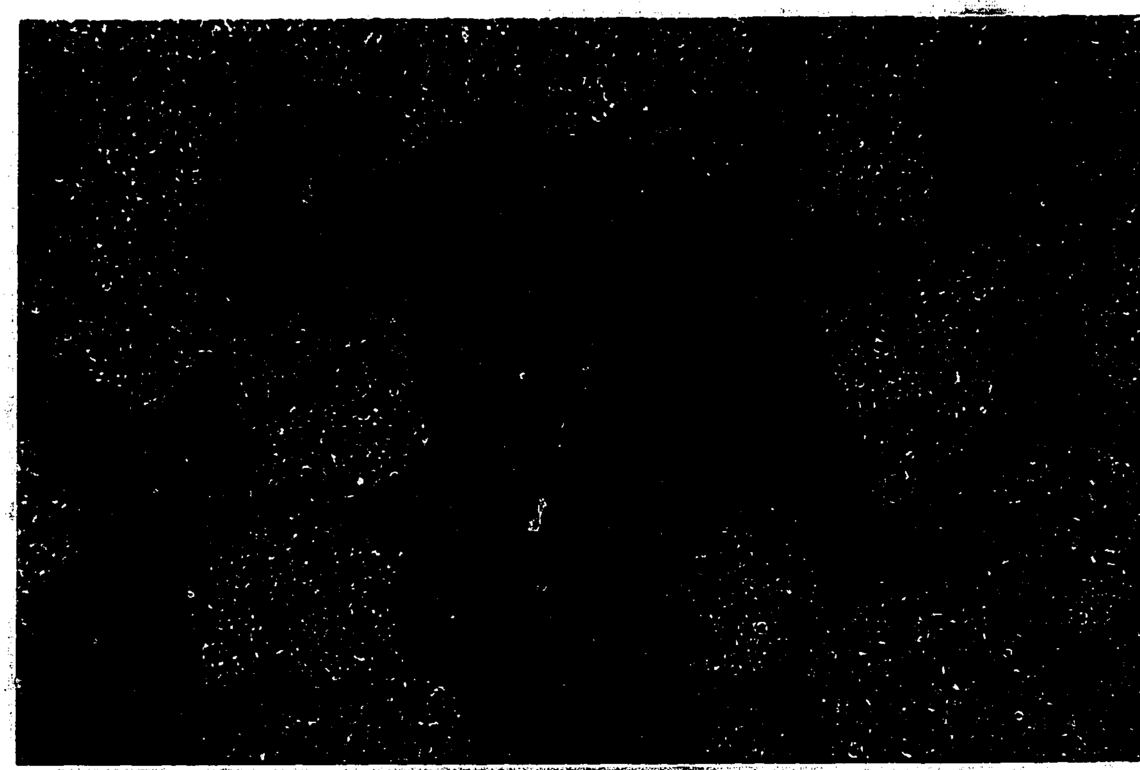
Plate 3.14a. The center of this slide shows a concentrically laminated clay filled pore space (14-09-37-3w5; 2286m). This feature indicates the deposition of clays in the pores by percolation of ground water downward, from a surface of subaerial exposure.

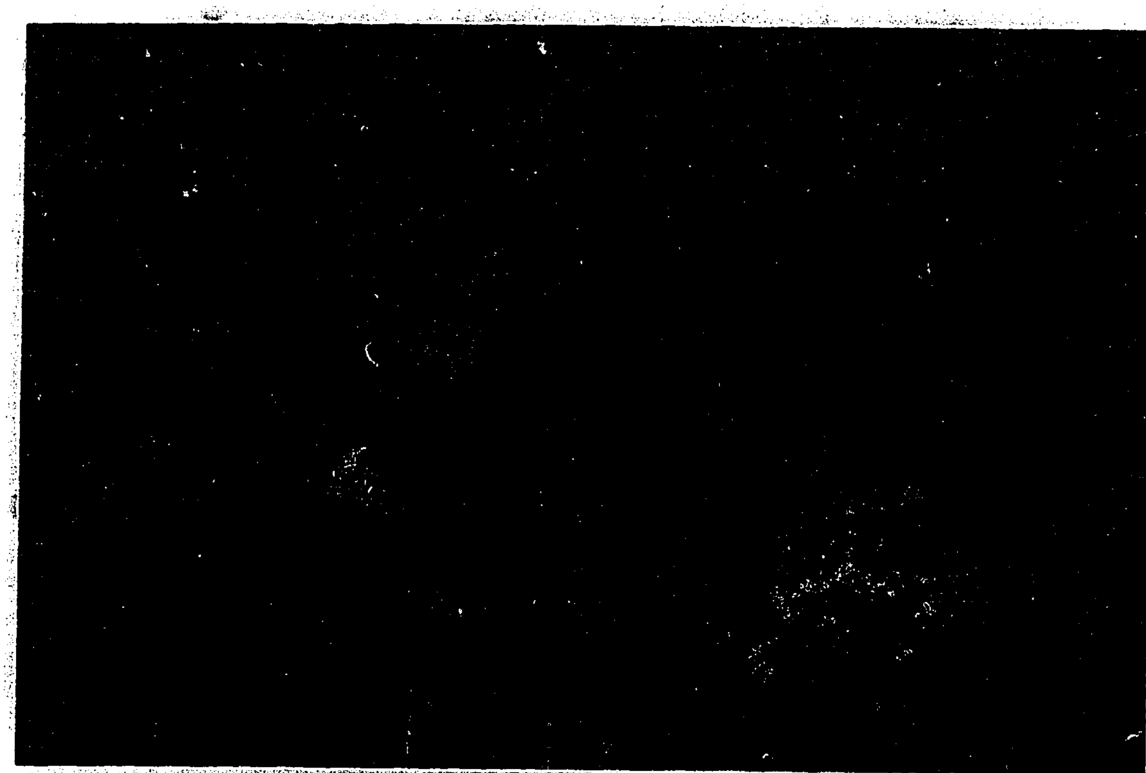
Plate 3.14b. This slide shows a fracture system in a tightly cemented quartz quartz arenite of the Rock Creek Member (14-09-37-3w5; 2281m). Parts of the open fracture system were filled with clay and silt sized material by percolating ground waters.



A

B

Plate 3.15. This slide shows chert (CT) rich poorly sorted sandstones of the Ellerslie Member (02-30-37-3w5; 7415ft.). This sample was collected from a valley which dissected the Rock Creek Member subcrop in the Sylvan Lake Field. Note the clay and silt matrix throughout much of the photo. A) PPL. B) Cross-polars.



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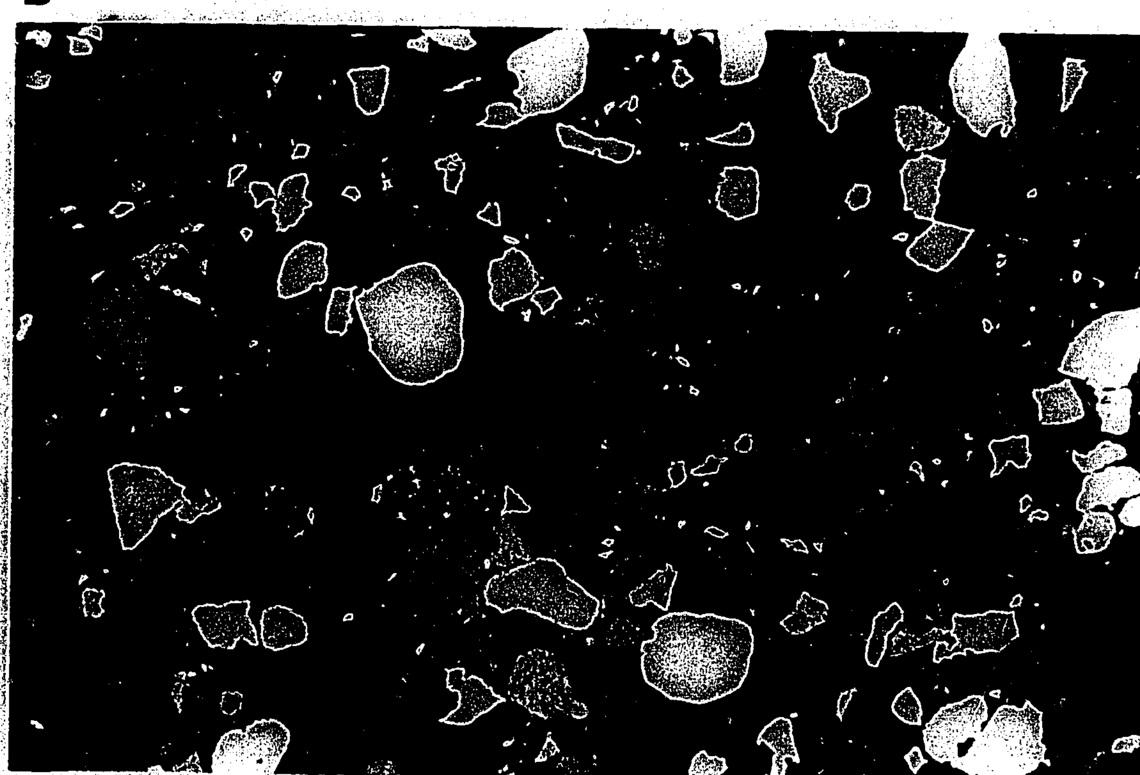
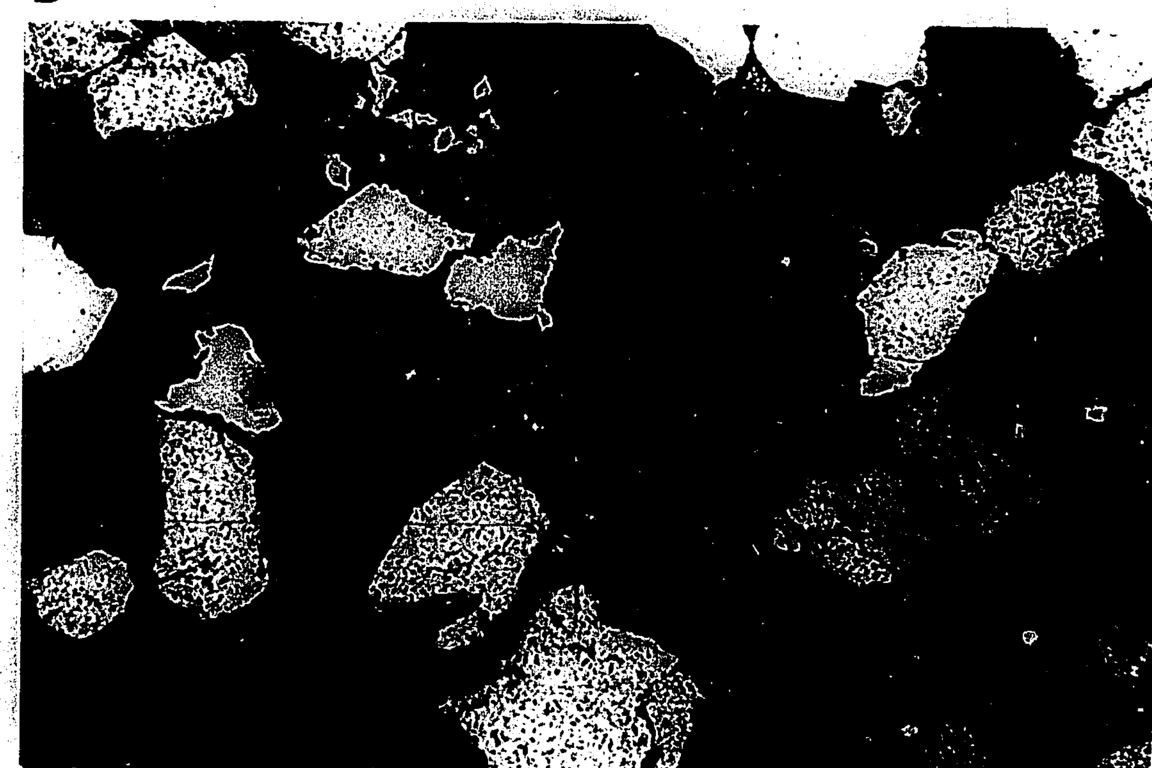


Plate 3.16. Chert rich sandstones of the Ellerslie Member (08-09-37-3w5). Porosity in this sample appears to be the result of dissolution of the chert grains. A) PPL. B) Cross-polars.

A

B



CONCLUSION

The Jurassic Fernie Formation in south-central Alberta was deposited on a west-facing, extremely old and stable passive margin. Subsidence rates on such margins are thermally controlled and, on one as old as this, are extremely slow. It is therefore not surprising that this formation, which is known to span most of Jurassic time, is very thin and dominated by erosional unconformities. In the study area, which lies relatively close to the depositional edge of several of the Fernie Formation's members, this unit consists of only a few of the members present to the west in the foothills of the Cordillera. The Fernie Formation in the study area is divisible into several discrete lithologic members which are: 1) sandy limestones and limestones of the Sinemurian age Nordegg Member, 2) stratigraphically equivalent sandstones of the Medicine River member, 3) oyster shell conglomerates of the '*Ostrea* Bed', which unconformably overlie the above units, 4) organic shales of the Toarcian Poker Chip Shale, 5) limestones, sandy limestones and sandstones of the Rock Creek Member and 6) shales of the Bathonian to early Callovian 'upper Jurassic Shale'. Major unconformity surfaces separate the Sinemurian, Toarcian and Bajocian deposits. Associated with this section is the detrital lithologic unit, which consists of weathering products of the various unconformities. As the detrital lithologies are observed beneath the Nordegg and Medicine River members, the *Ostrea* Bed, the Poker Chip Shale, the Rock Creek Member and the Cretaceous section, this unit is highly diachronous.

The Nordegg interval in the Medicine River and Gilby field areas was deposited on a stable, slowly subsiding passive margin which was charaterized by slow rates of terrigenous clastic input. The shelf was dominated by carbonate accumulations, except where fluvially introduced clastics were reworked into clean beach and shoreface sandstones. The Nordegg interval is composed of two sequences, separated by a surface of subaerial exposure and erosional truncation. In the first sequence, dolomitic shoreline sandstones prograded onto a carbonate shelf during deposition of the highstand systems tract. The second sequence contains a condensed deposit of oyster shells, which are the nearshore expression of the maximum period of flooding during deposition of the second sequence. These two sequences correlate well to the eustatic sea-level curves of Haq et al., (1989).

Slow subsidence and sedimentation rates allowed erosional processes to dominate the Nordegg interval, and rarely are appreciable sections of either sequence preserved. The development of the north-south trending Medicine River-Gilby subsidence basin allowed for the preservation of thick sections of the nearshore sandstones of the lower sequence. This subsidence predated the *Ostrea* bed deposition, and therefore did not allow the lower sequence to be as deeply eroded, at this sequence boundary, as in other areas. Subsidence in this local area appears to have been the result of flexure of the rigid Mississippian carbonate rocks over the underlying Leduc reef, during Lower Jurassic time. This flexure developed a fault system, which when loaded by sediments, became an area of active subsidence. Preservation of the sandstone, and therefore development of the prolific Nordegg-equivalent reservoirs, in this area, was directly in response to subsidence during Lower Jurassic time.

Areas of similar subsidence relationships, along the erosional edge of the Lower Jurassic, would make good prospective targets for future exploration. In addition, the recognition of a sequence boundary within the Nordegg interval opens the possibilities for the discovery of future reservoir sediments in shelf margin and transgressive systems tract deposits, basinward of the erosional edge.

The Rock Creek Member in the Gilby, Medicine River and Sylvan Lake Fields was deposited in at least three sequences, each bounded above and below by a surface of subaerial exposure and erosion. This unit was deposited in three laterally transitional depositional systems. In the central portion of the area, a delta system deposited a succession of heterolithic deposits which contained abundant fine clastic detritus. This delta system was dominated by wave and longshore current energy, which redistributed much of the fine clastic material along the coast. Sand from this system was transported short distances and accumulated along the shoreface and beach system of the associated strandplain. Seaward, away from the influence of suspended clastic detrital material, the shelf area was dominated by carbonate sedimentation. Here pelecypod and echinoderm grainstones accumulated in high energy areas and pelecypod wackestones and rudstones in lower energy settings.

The Rock Creek Member was deposited on an old and very stable passive margin where subsidence rates greatly affected the preservation of the deposits. In some areas, such as the Sylvan Lake Field, high subsidence rates allowed for the preservation of anomalously thick successions of Rock Creek sediment. Continued subsidence throughout the deposition of the Middle Jurassic sediments and into Lower Cretaceous time directed Lower Mannville drainage and therefore erosion through this field. The result was an area of complex stratigraphy in which Ellerslie and Jurassic sediments are juxtaposed, and therefore an area which is extremely difficult to map.

The main reservoir rocks in the Rock Creek Member are formed by the clean sandstones of the shoreface environment. In many areas these sandstones are tightly cemented so that the best reservoir conditions occur at the subcrop edge, where porosity and permeability are enhanced by secondary solution of cements and framework grains. In the Sylvan Lake Field continued subsidence caused pervasive fracturing in the Rock Creek sandstones, which, combined with the proximity of this area to the pre-Cretaceous unconformity, created even better reservoir enhancement.

Future efforts in exploration of hydrocarbons in the Rock Creek Member should be aimed at mapping the separate sequences and predicting the shoreline sandstone trends within these individual sequences. Then mapping of Lower Cretaceous Valley systems and inter Rock Creek erosional surfaces will point to optimum locations for stratigraphic hydrocarbon traps within the Rock creek Member.

The predominant controls on the deposition of the Jurassic sediments appear to have been a combination of sea-level fluctuation and sediment supply. Where sediment supply was low carbonates were the dominant sediment type. However, where sedimentation was high, deltaic and strandplain sedimentation prevailed. Preservation of the sediments was strongly controlled by the rates of subsidence on the shelf and by localized perturbations in the subsidence rates. The Medicine River and Gilby area is one of the only places in western Canada where the Sinemurian shoreline is preserved. This unique situation was created by local high rates of subsidence, prior to pre-Toarcian erosion, that brought these sediments below base level. Similarily thick sequences of Rock Creek sediments are preserved in the Sylvan Lake field because of high rates of local subsidence pre-dating the bulk of pre-Cretaceous erosion, but post-dating Nordegg Member deposition.

REFERENCES

Haq, B.U., Hardenbol, J., and Vail, P. R., 1989, Mesozoic and Cenozoic chronostratigraphy and cycles of sea-level change. In Wilgus, C.K., Hastings, B.S., C. G. St. G., Kendall, Posamentier, H. W., Ross, C. A., and Van Wagoner, J. C., eds., Sea-Level Changes: An Integrated Approach: Society of Economic Paleontologists and Mineralogists Special Publication No. 42, p. 71-108.

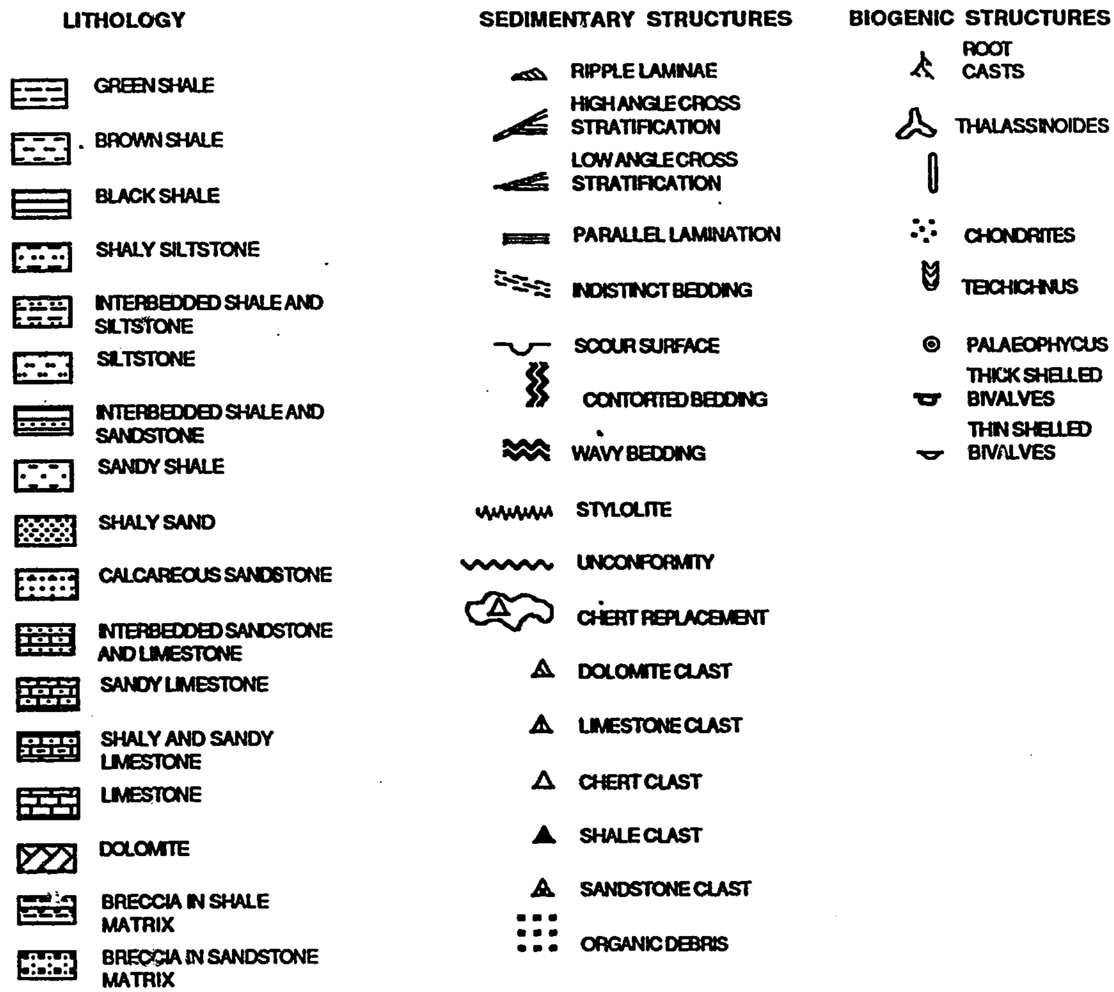
APPENDICES

APPENDIX

1. Nordegg Stratigraphy Core-logs
    i) List of Cores
    ii) Legend
    iii) Logs

2. Rock Creek Stratigraphy Core logs
    i) List of Cores
    ii) Legend
    iii) Logs

3. Mineralogy Tables
    i) Percentage of minerals counted
    ii) Relative percent of framework grains


LITHOLOGY
GREEN SHALE
BROWN SHALE
BLACK SHALE
SHALY SILTSTONE
INTERBEDDED SHALE AND SILTSTONE
SILTSTONE
INTERBEDDED SHALE AND SANDSTONE
SANDY SHALE
SHALY SAND
CALCAREOUS SANDSTONE
INTERBEDDED SANDSTONE AND LIMESTONE
SANDY LIMESTONE
SHALY AND SANDY LIMESTONE
LIMESTONE
DOLOMITE
BRECCIA IN SHALE MATRIX
BRECCIA IN SANDSTONE MATRIX
SEDIMENTARY STRUCTURES
RIPPLE LAMINAE
HIGH ANGLE CROSS STRATIFICATION
LOW ANGLE CROSS STRATIFICATION
PARALLEL LAMINATION
INDISTINCT BEDDING
SCOUR SURFACE
CONTORTED BEDDING
WAVY BEDDING
STYLOLITE
UNCONFORMITY
CHERT REPLACEMENT
DOLOMITE CLAST
LIMESTONE CLAST
CHERT CLAST
SHALE CLAST
SANDSTONE CLAST
ORGANIC DEBRIS
BIOGENIC STRUCTURES
ROOT CASTS
THALASSINOIDES
CHONDRITES
TEICHICHNUS
PALAEOPHYCUS
THICK SHELLED BIVALVES
THIN SHELLED BIVALVES

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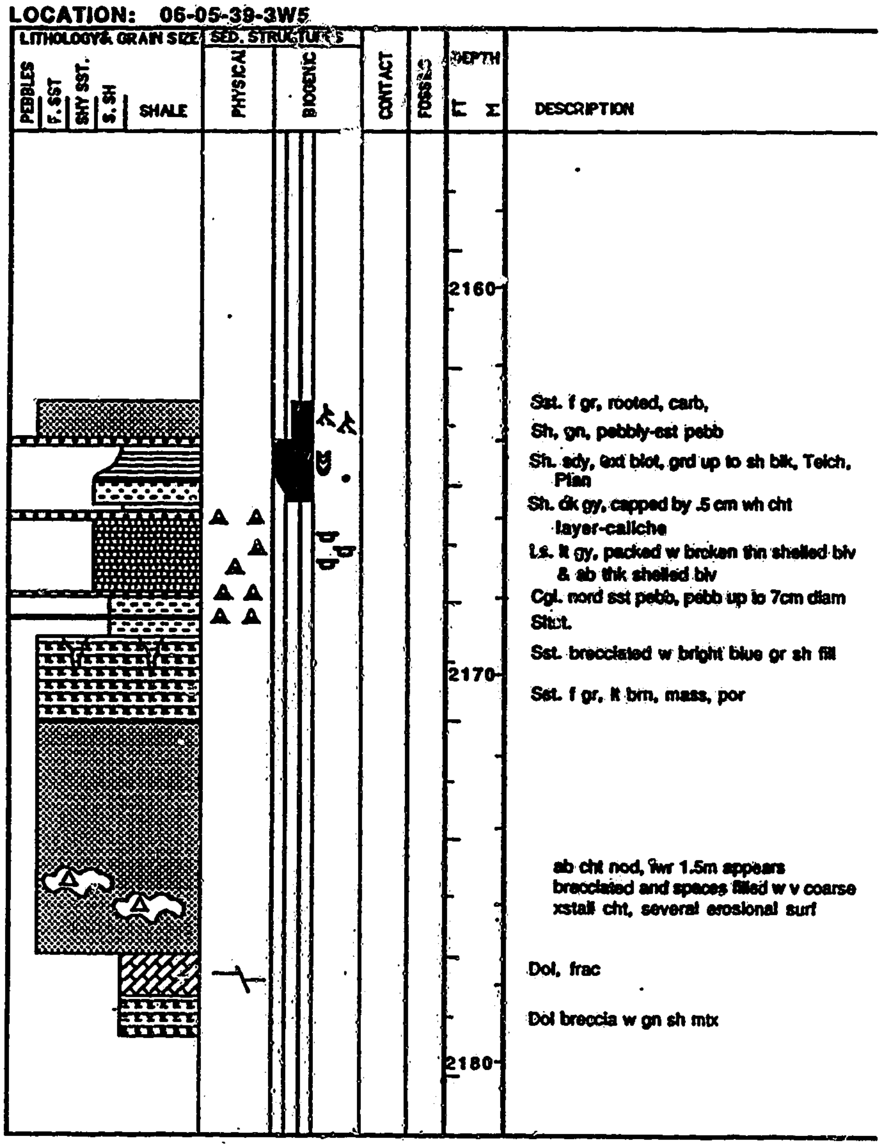

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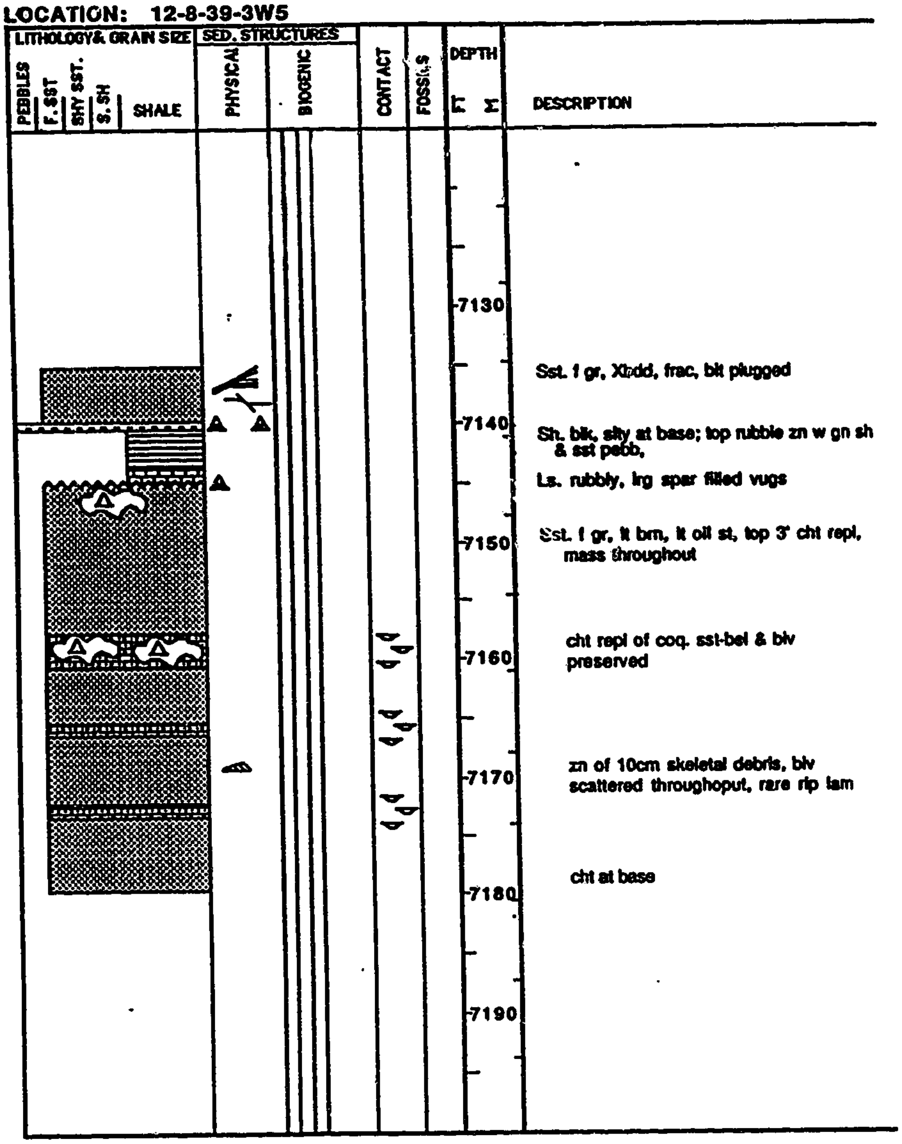

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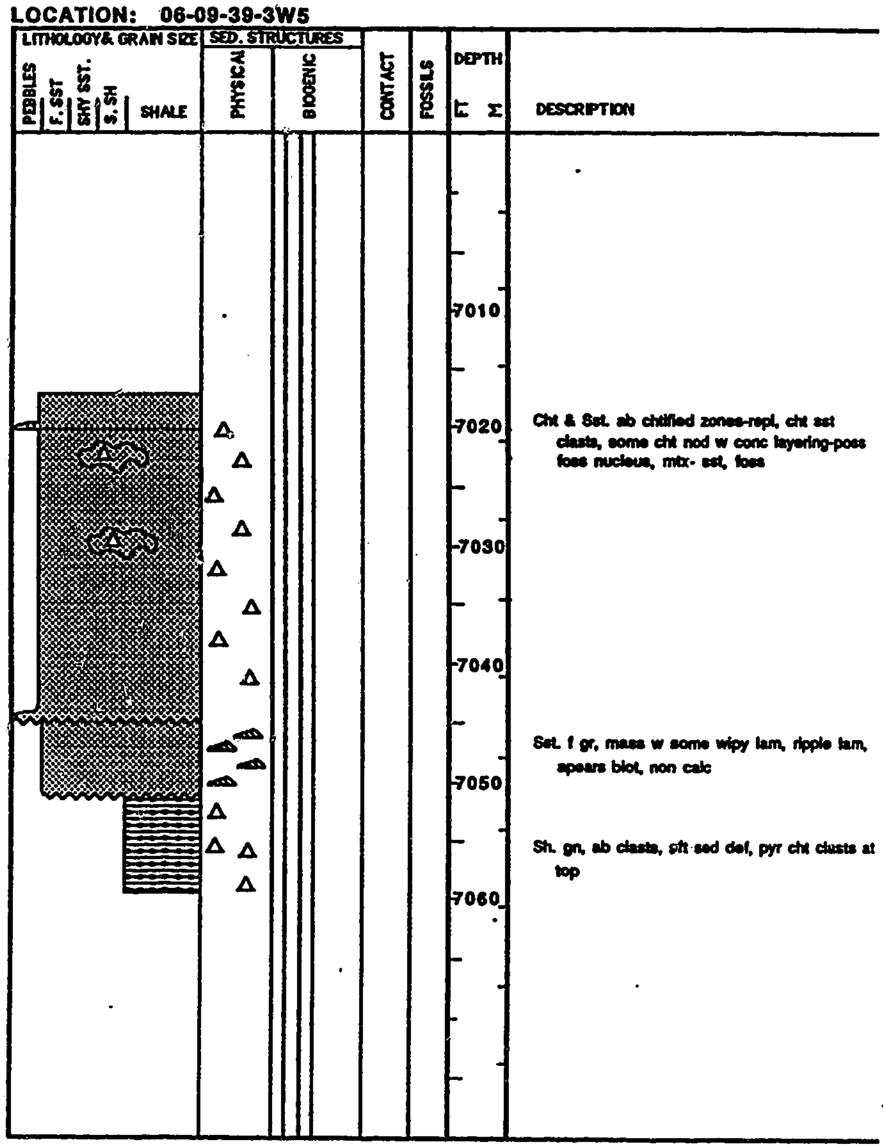

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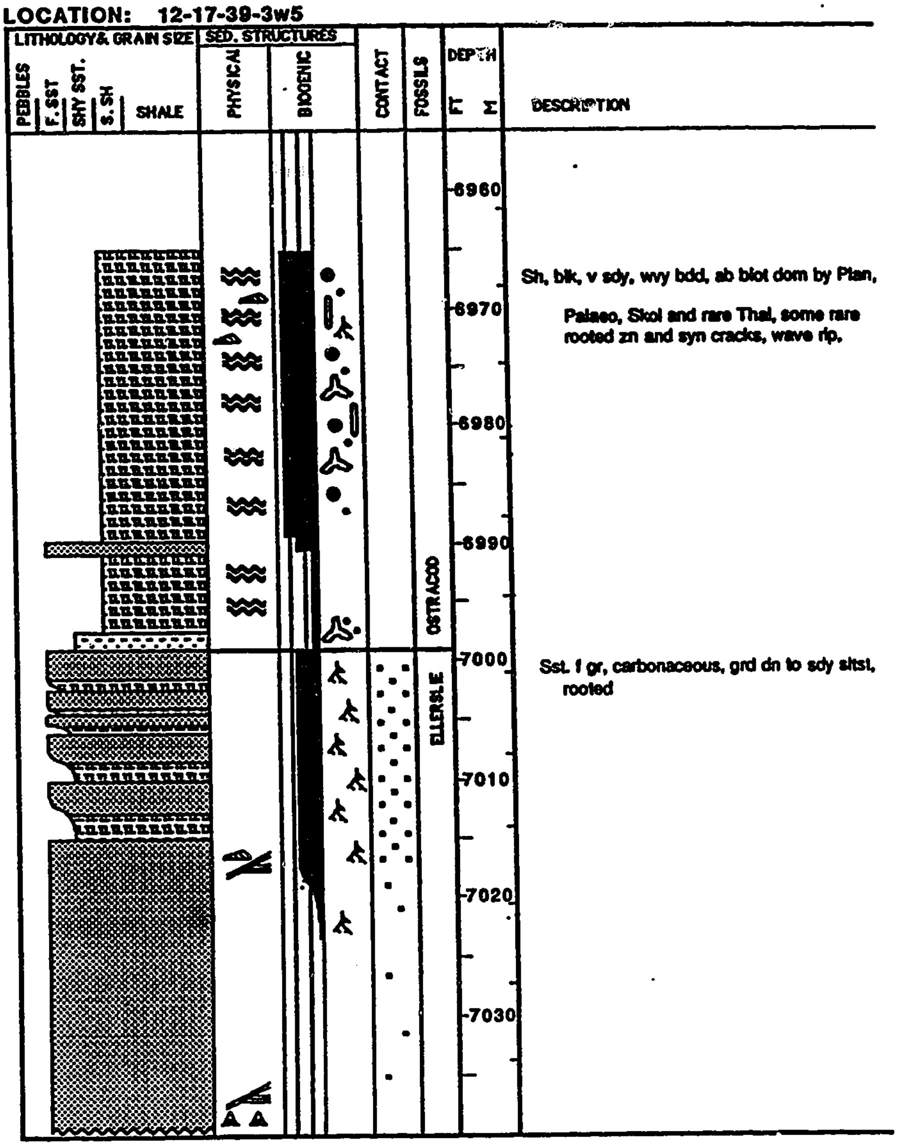

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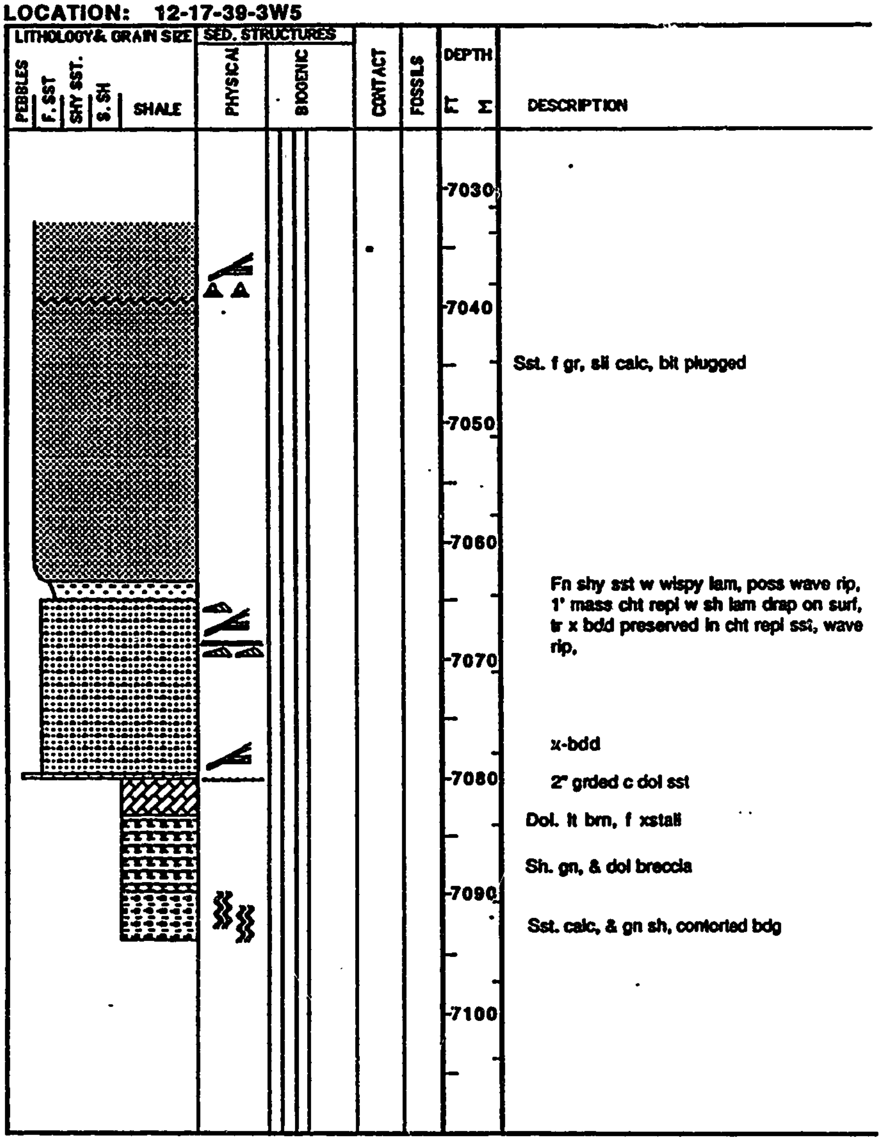

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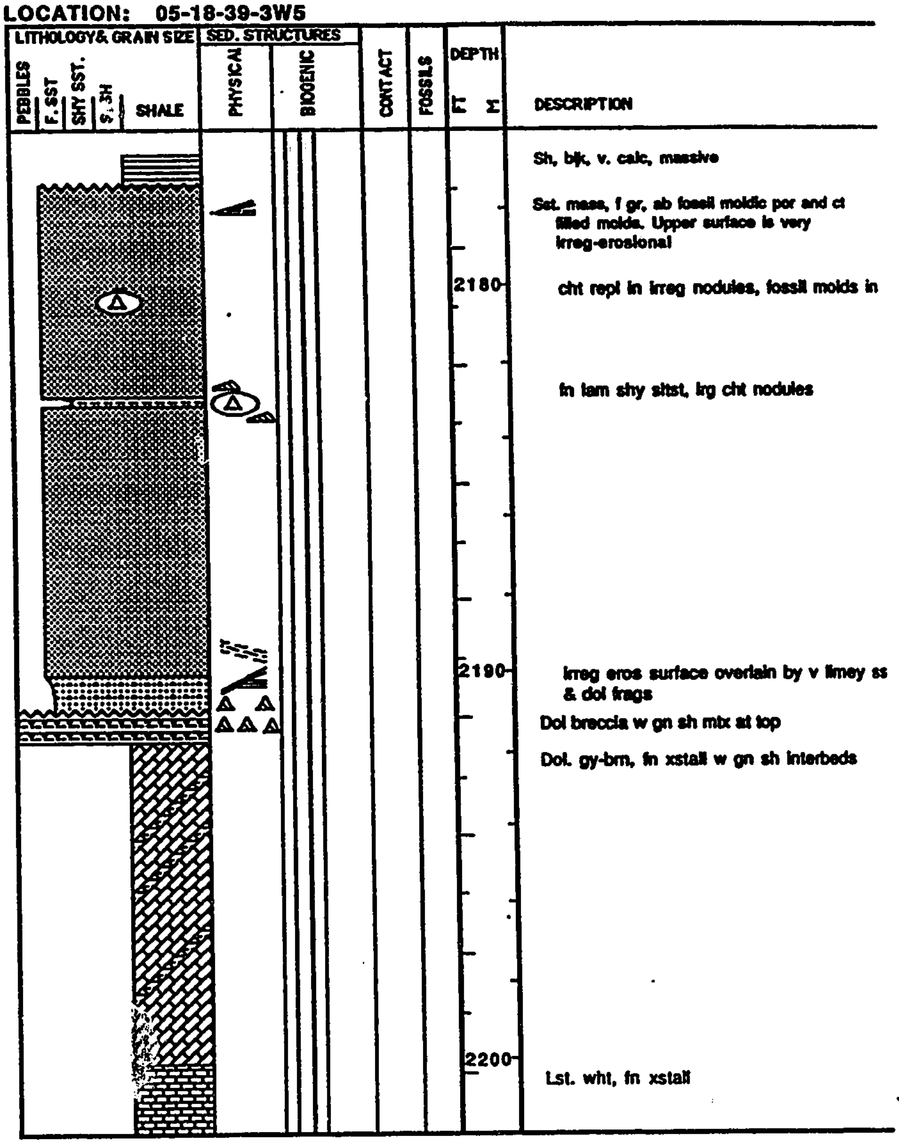

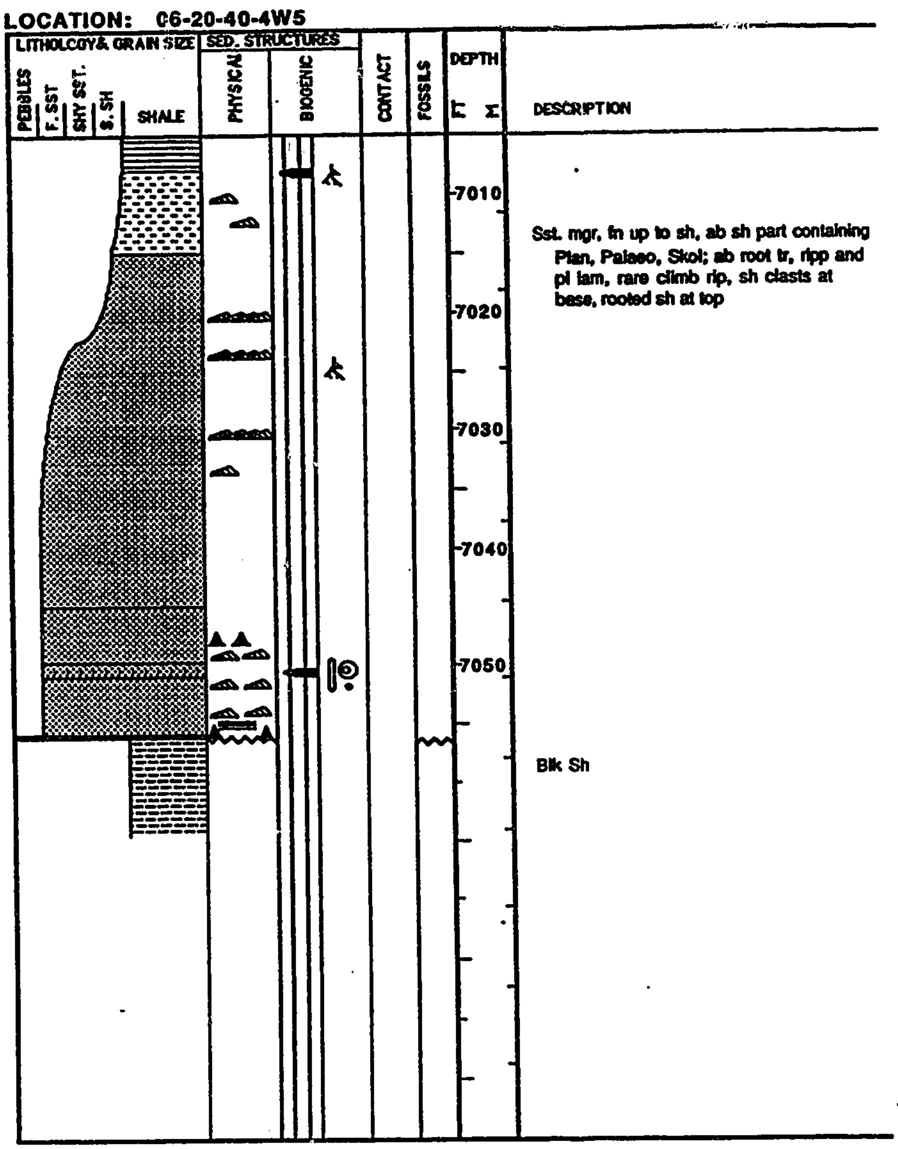
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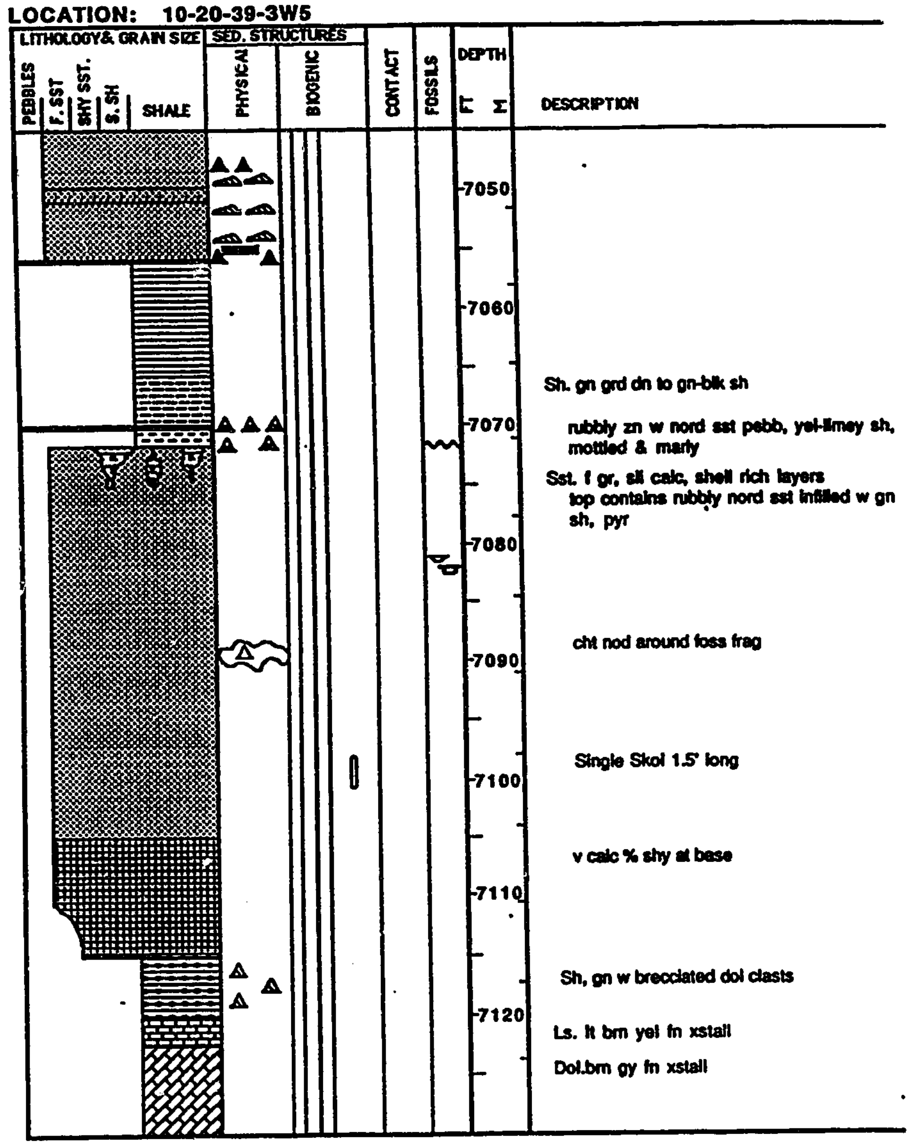
PEBBLES
F. SST
SHY SST.
S. SH
SHALE
SED. STRUCTURES
PHYSICAL
BIOGENIC
CONTACT
FOSSILS
DEPTH
FT M
DESCRIPTION
7010
7020
7030
7040
7050
Sst. mgr, fn up to sh, ab sh part containing Plan, Palaeo, Skol; ab root tr, ripp and pl lam, rare climb rip, sh clasts at base, rooted sh at top
Blk Sh

LOCATION: 10-20-39-3W5



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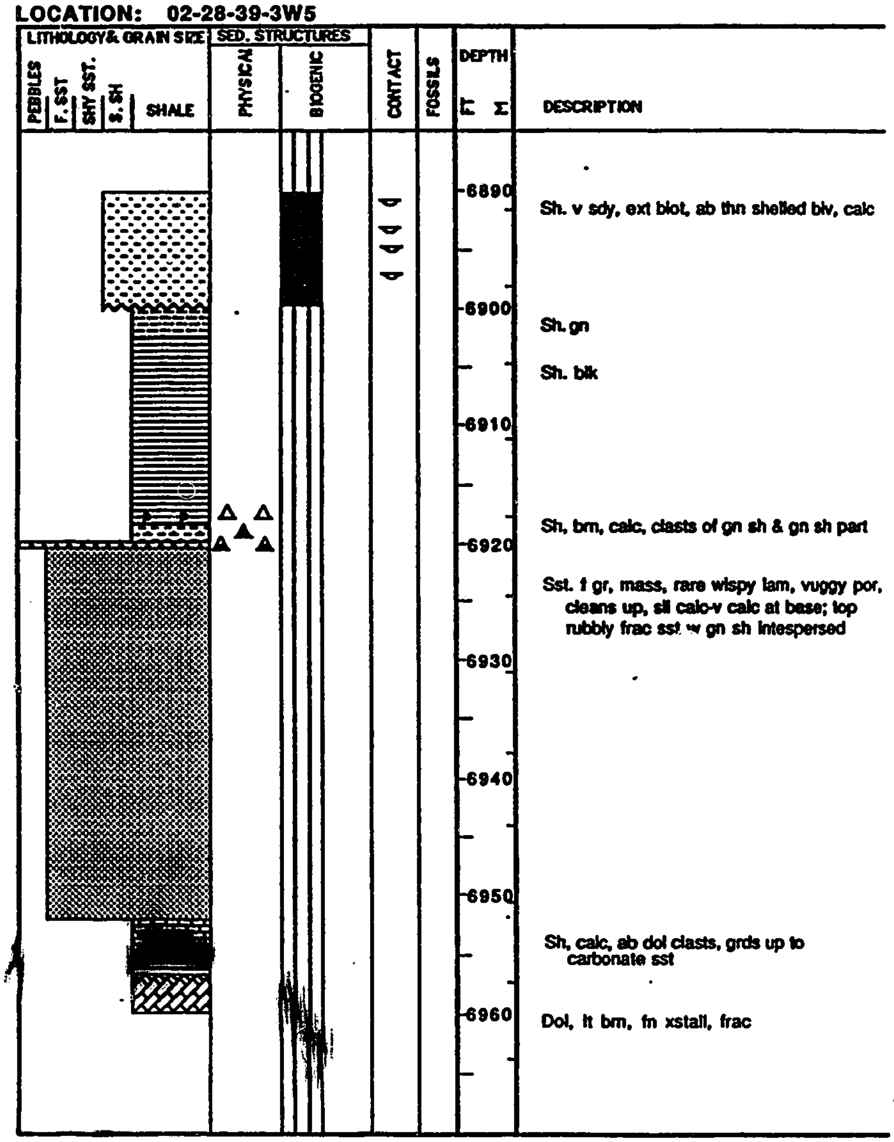

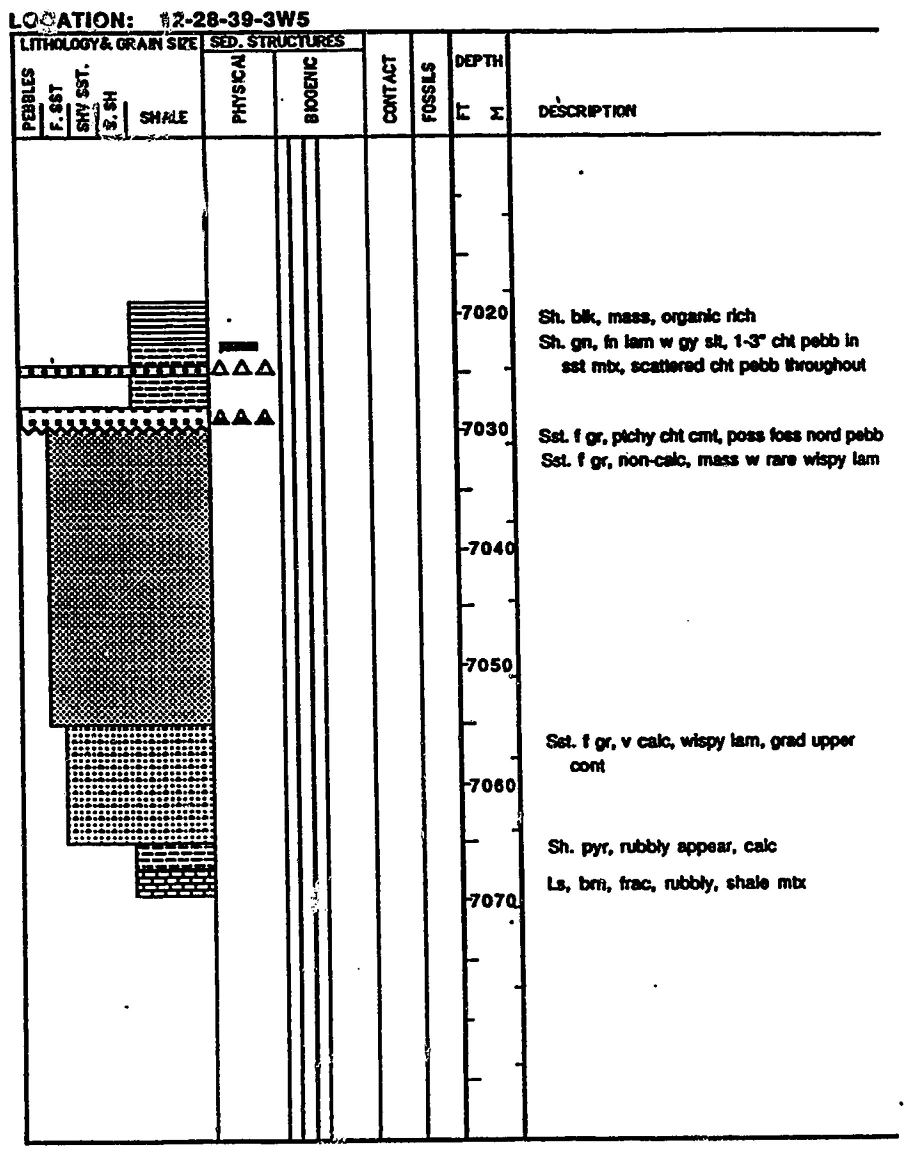
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LITHOLOGY & GRAIN SIZE
PEBBLES
F. SST
SHY SST.
S. SH
SHALE
SED. STRUCTURES
PHYSICAL
BIOGENIC
CONTACT
FOSSILS
DEPTH
FT M
DESCRIPTION
7020
7030
7040
7050
7060
7070
Sh. blk, mass, organic rich
Sh. gn, fn lam w gy slt, 1-3" cht pebb in sst mtx, scattered cht pebb throughout
Sst. f gr, pchy cht cmt, poss foss nord pebb
Sst. f gr, non-calc, mass w rare wispy lam
Sst. f gr, v calc, wispy lam, grad upper cont
Sh. pyr, rubbly appear, calc
Ls, brn, frac, rubbly, shale mtx

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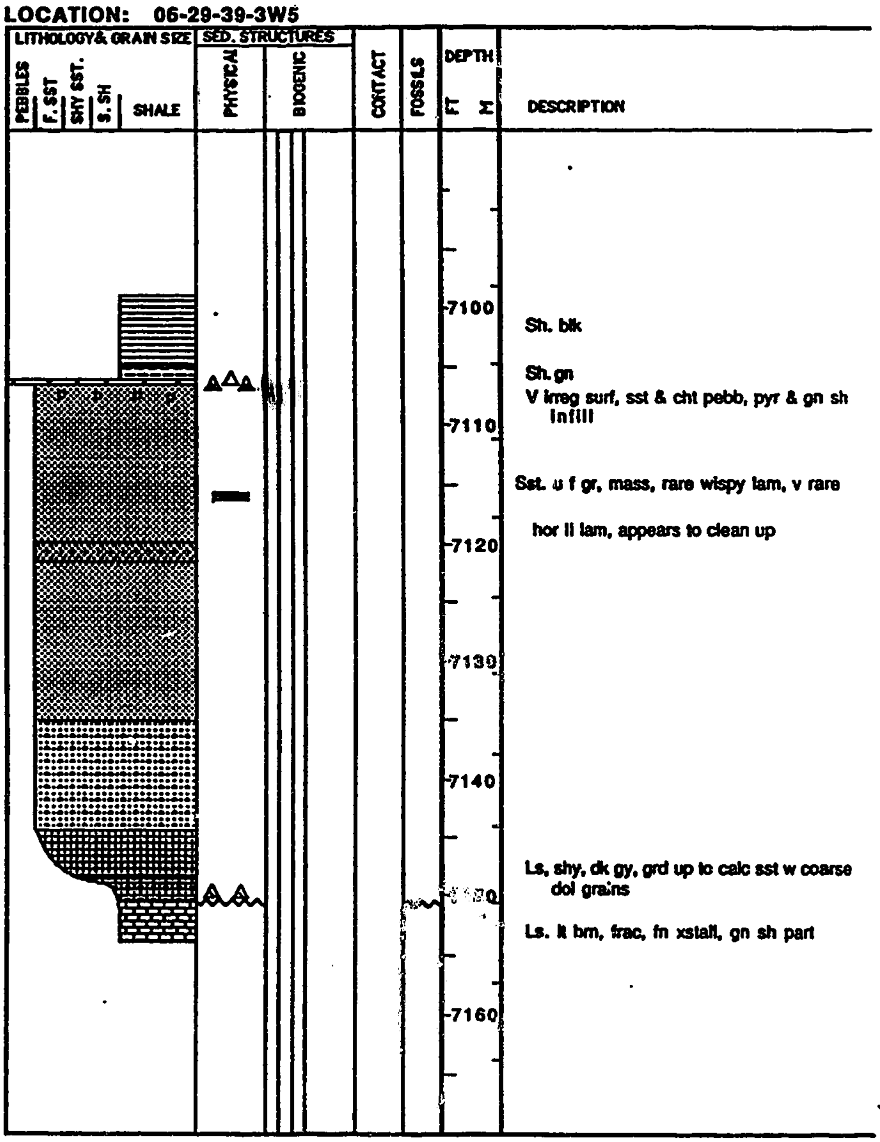

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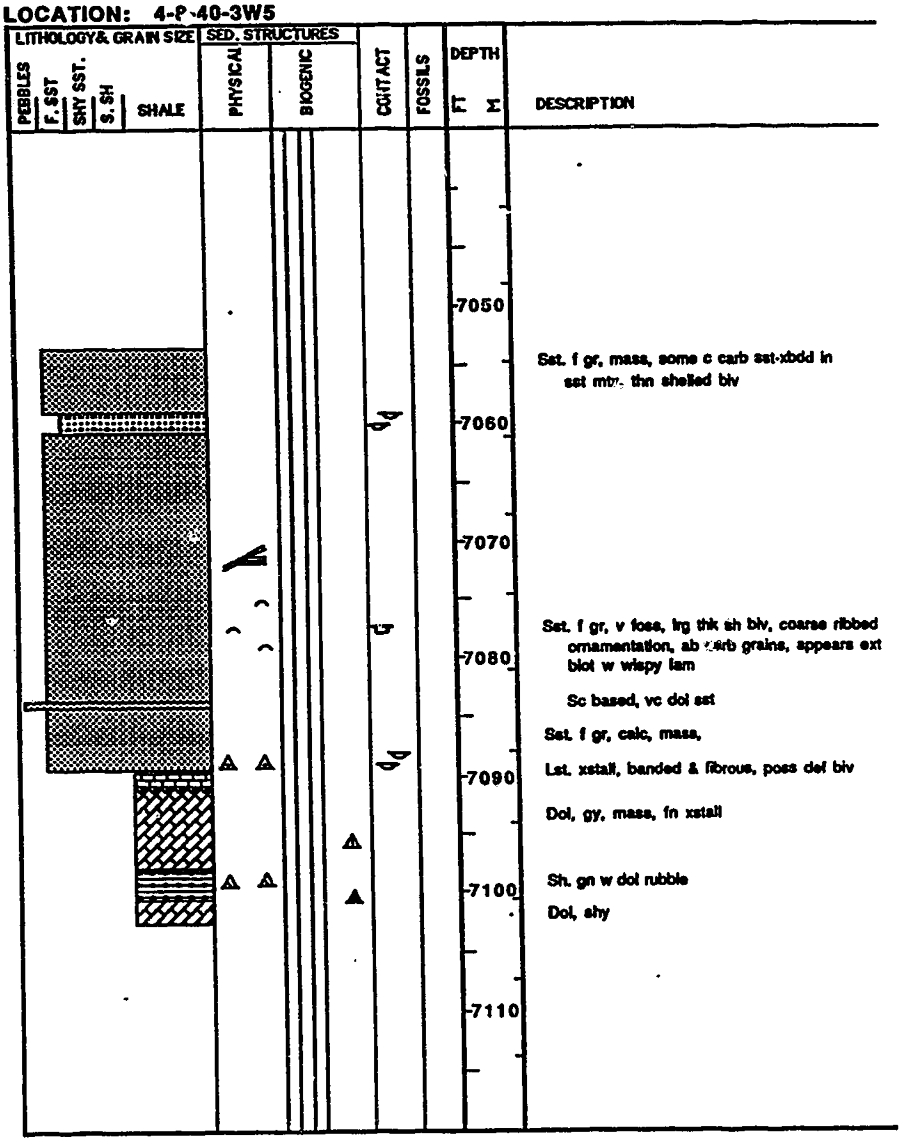

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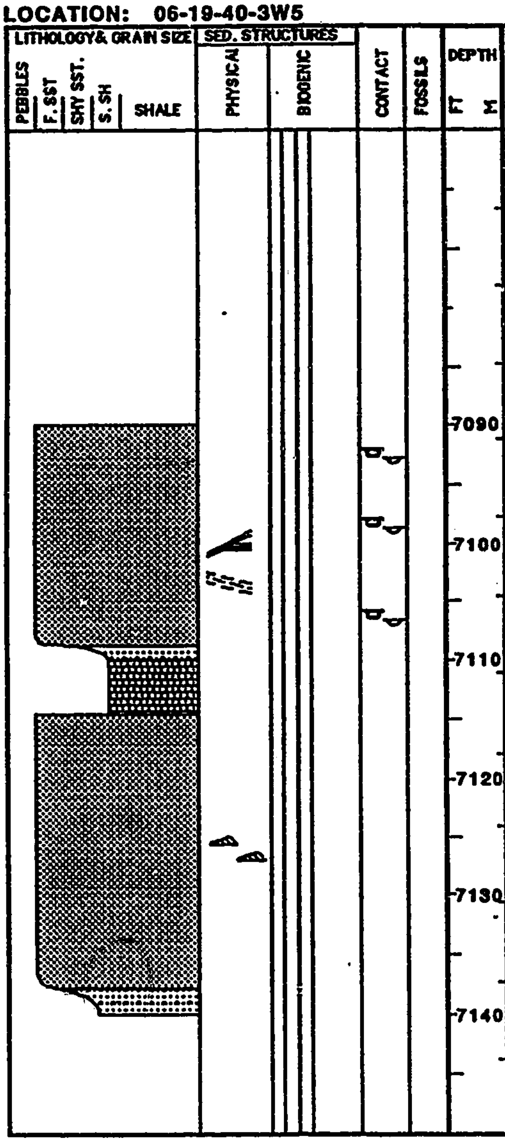

DESCRIPTION

Sst. u f gr, ab foss frag, thk shell biv, ribbe
biv, rare ind lam, Xbdg, skeletal sst, mostly massive

Ls. sdy, skeletal, xbdd, at base, l c gr foss debris in f gr sst mtx, sst content incr up.

Sst. u f gr, calc, mass, rippl lam, more calc at base, ext biot text

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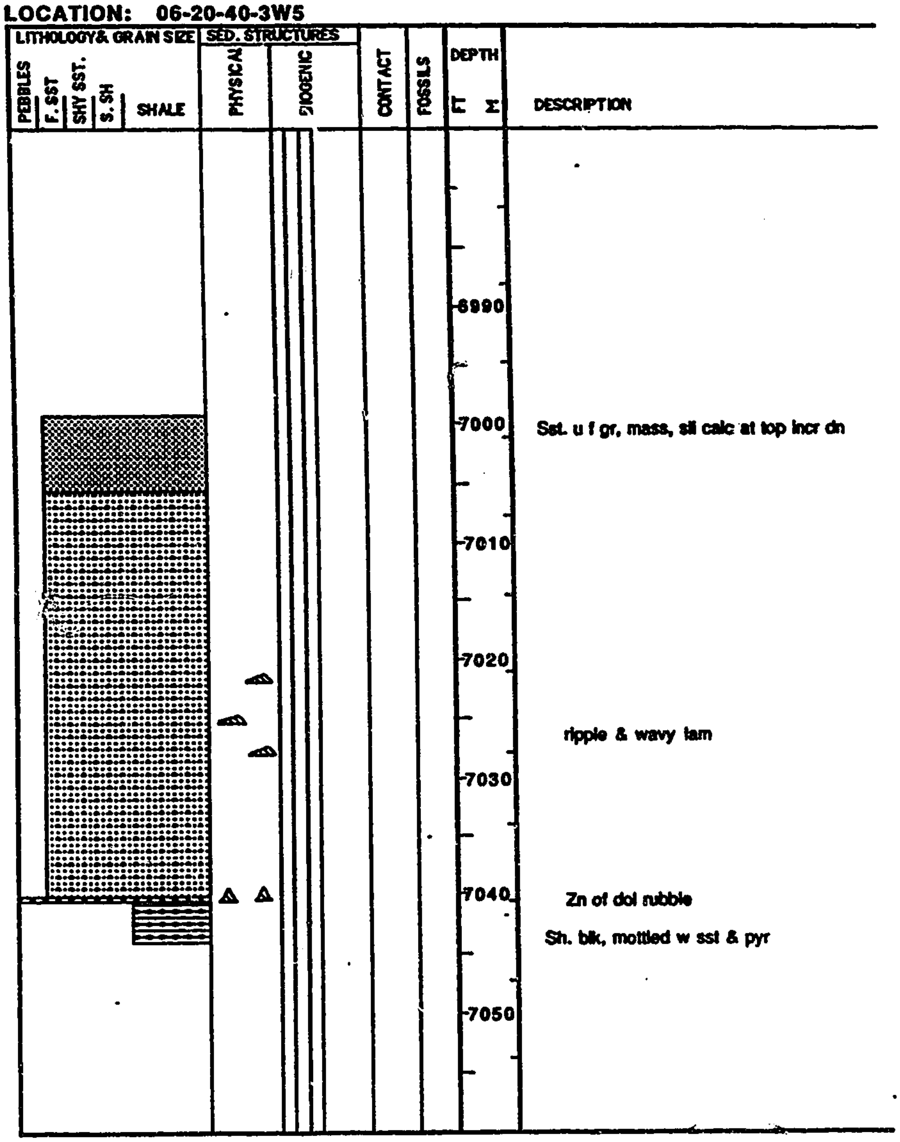

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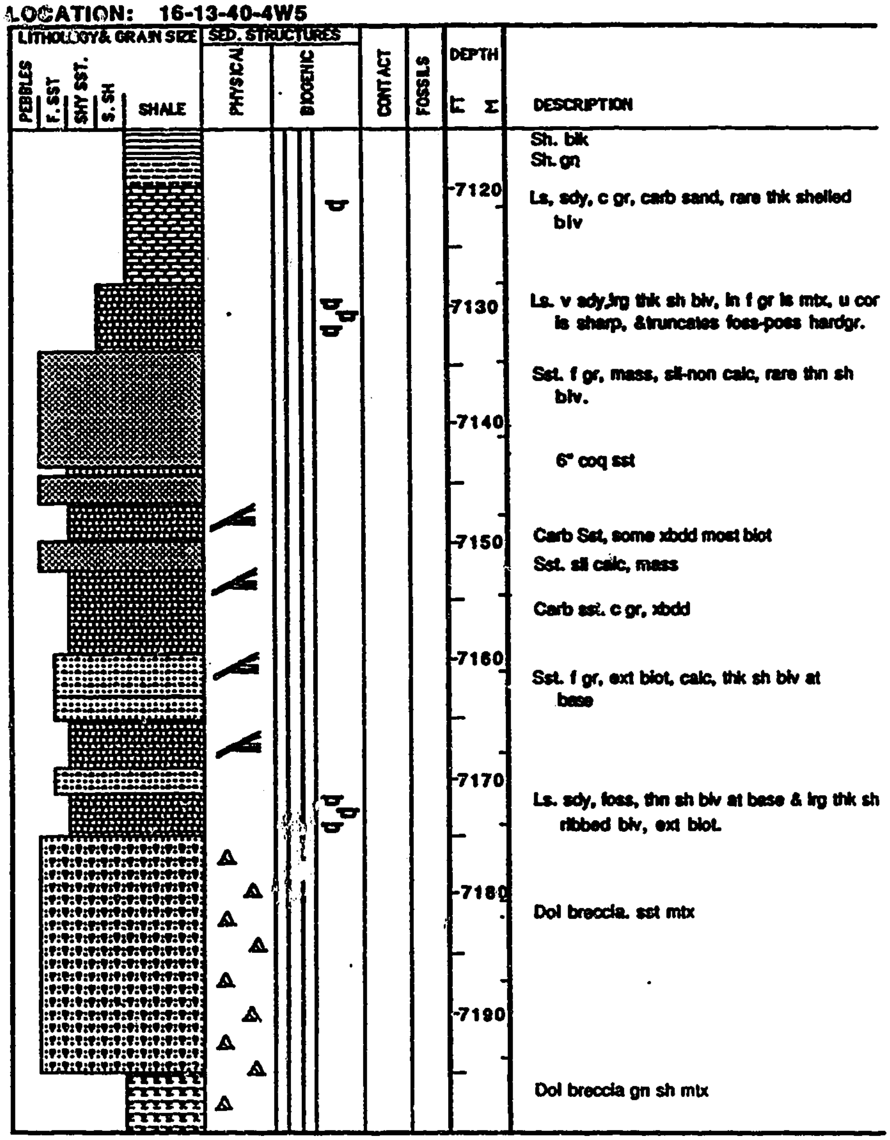

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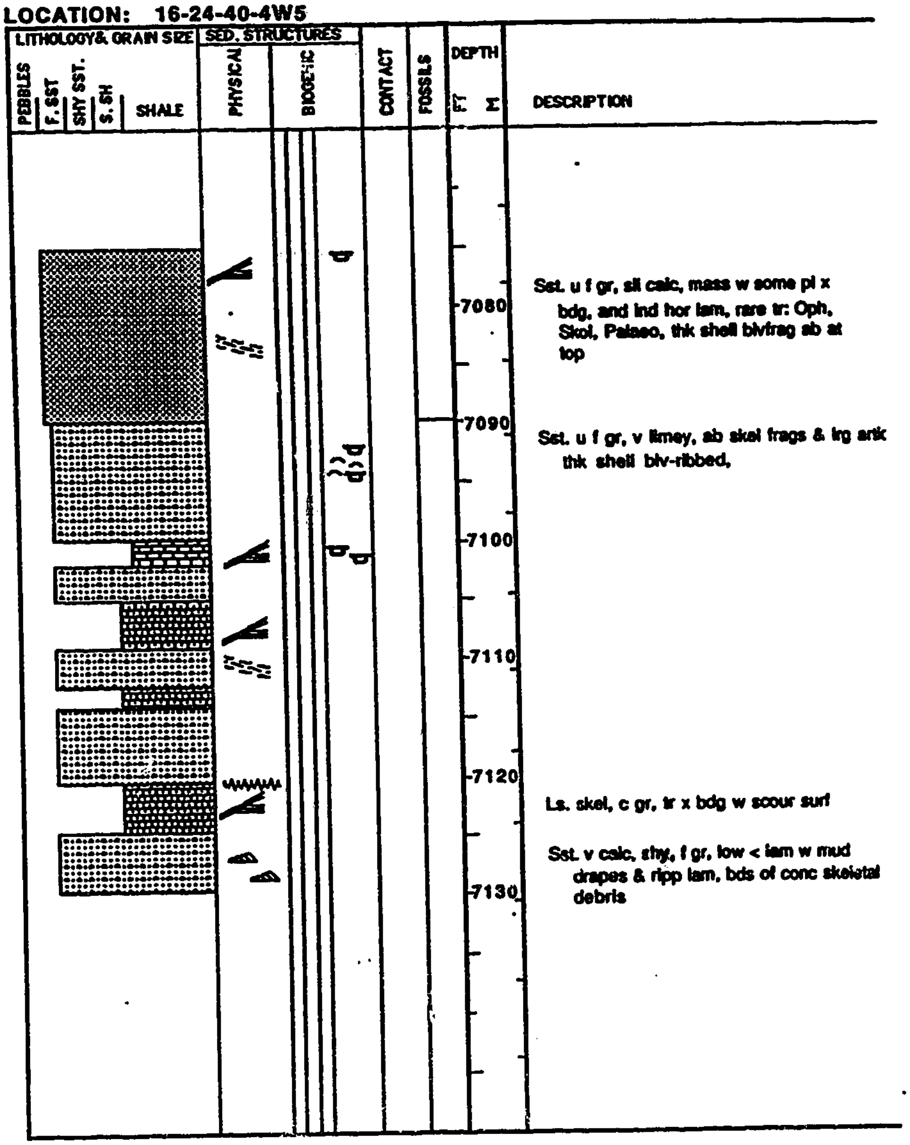

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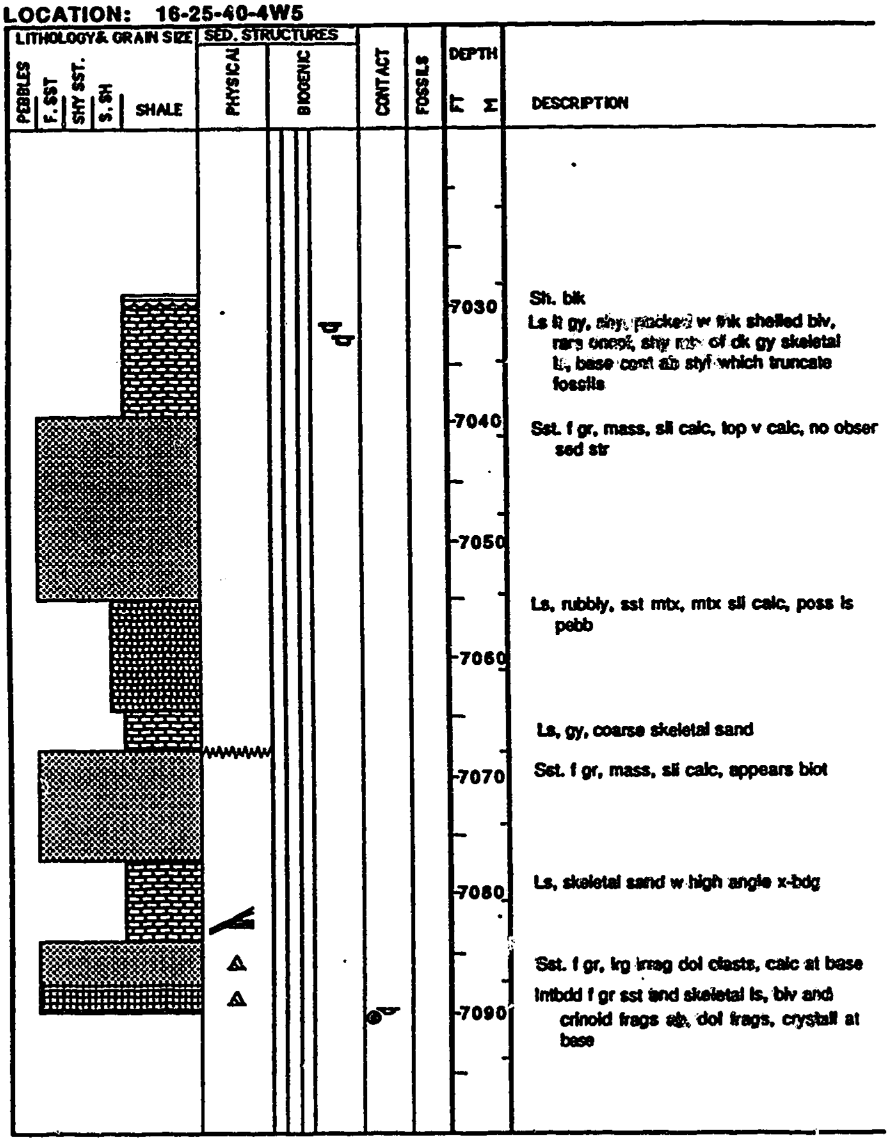

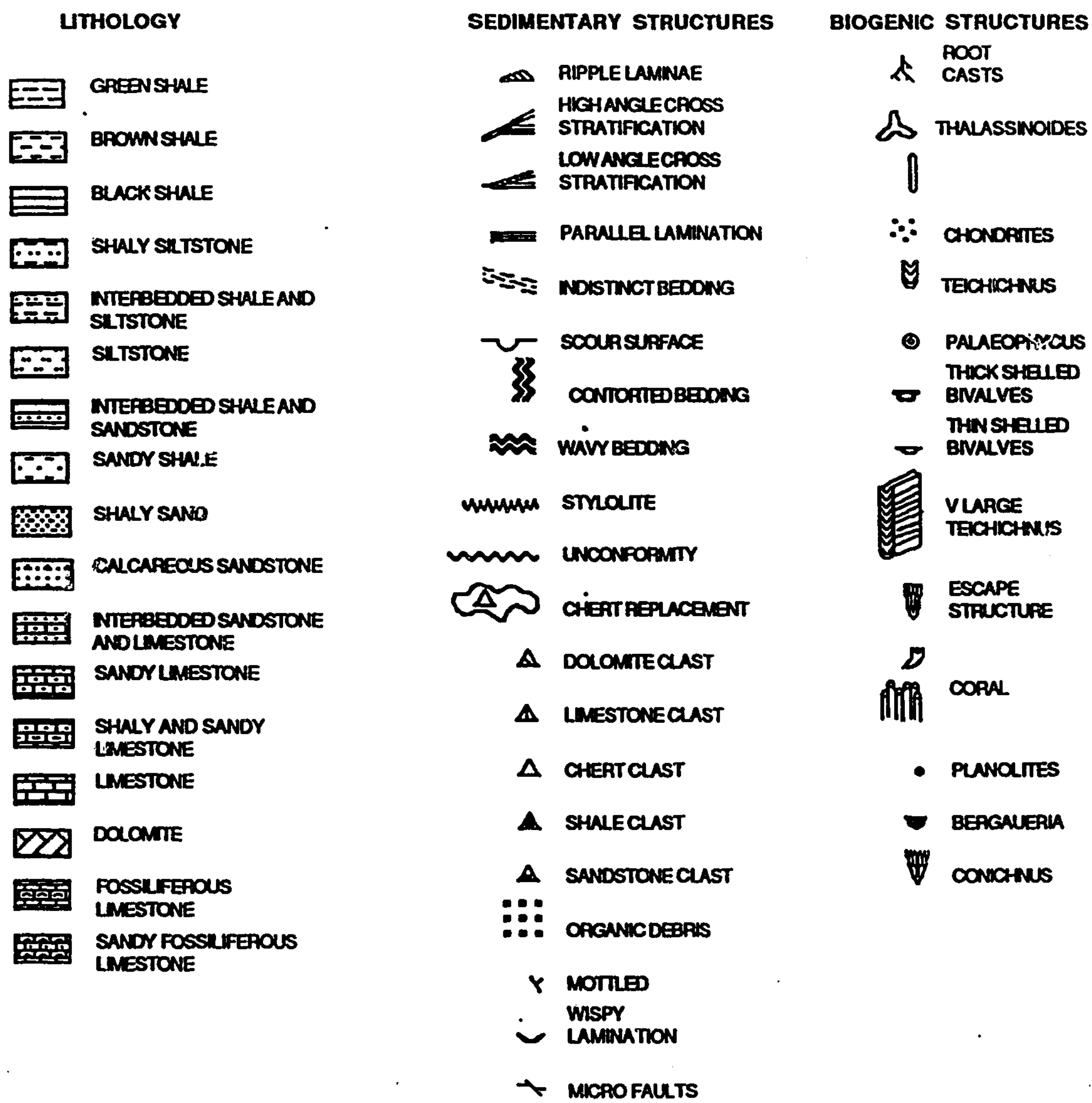
LITHOLOGY
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BROWN SHALE
BLACK SHALE
SHALY SILTSTONE
INTERBEDDED SHALE AND SILTSTONE
SILTSTONE
INTERBEDDED SHALE AND SANDSTONE
SANDY SHALE
SHALY SAND
CALCAREOUS SANDSTONE
INTERBEDDED SANDSTONE AND LIMESTONE
SANDY LIMESTONE
SHALY AND SANDY LIMESTONE
LIMESTONE
DOLOMITE
FOSSILIFEROUS LIMESTONE
SANDY FOSSILIFEROUS LIMESTONE
SEDIMENTARY STRUCTURES
RIPPLE LAMINAE
HIGH ANGLE CROSS STRATIFICATION
LOW ANGLE CROSS STRATIFICATION
PARALLEL LAMINATION
INDISTINCT BEDDING
SCOUR SURFACE
CONTORTED BEDDING
WAVY BEDDING
STYLOLITE
UNCONFORMITY
CHERT REPLACEMENT
DOLOMITE CLAST
LIMESTONE CLAST
CHERT CLAST
SHALE CLAST
SANDSTONE CLAST
ORGANIC DEBRIS
MOTTLED
WISPY LAMINATION
MICRO FAULTS
BIOGENIC STRUCTURES
ROOT CASTS
THALASSINOIDES
CHONDRITES
TEICHICHNUS
PALAEOPHYCUS
THICK SHELLED BIVALVES
THIN SHELLED BIVALVES
V LARGE TEICHICHNUS
ESCAPE STRUCTURE
CORAL
PLANOLITES
BERGAUERIA
CONICHNUS

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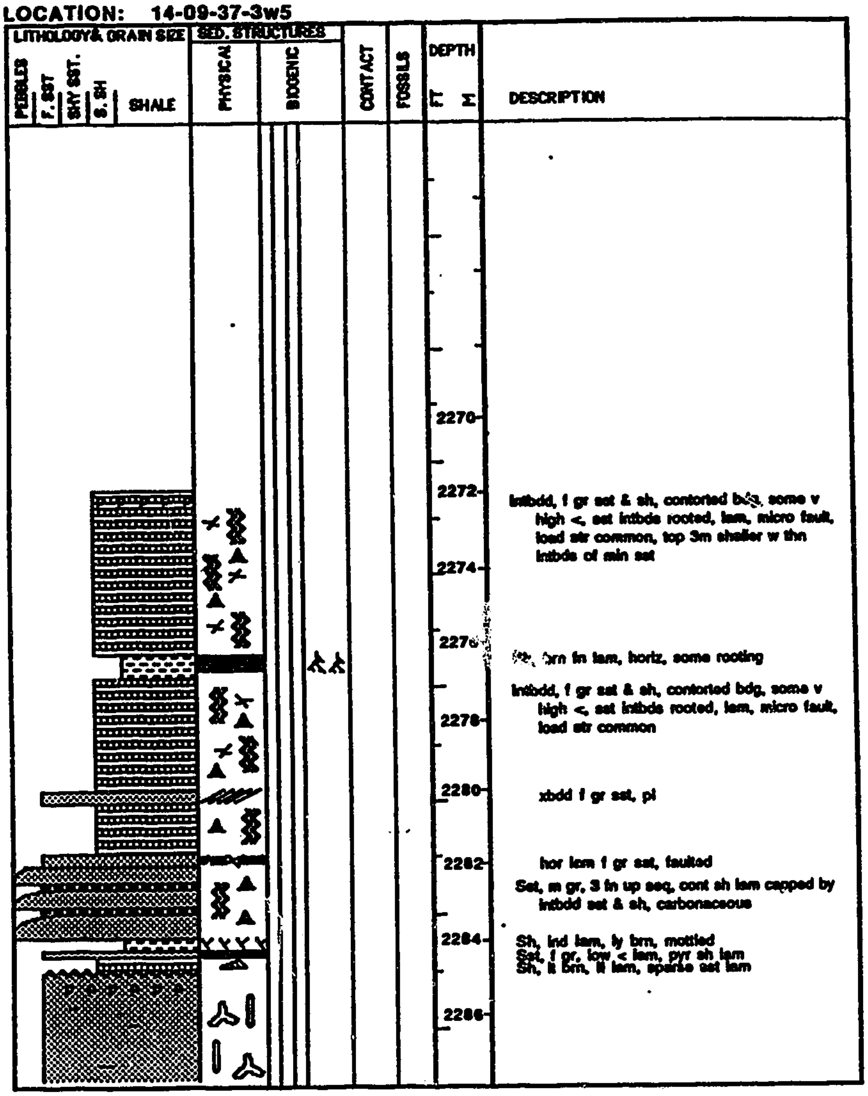

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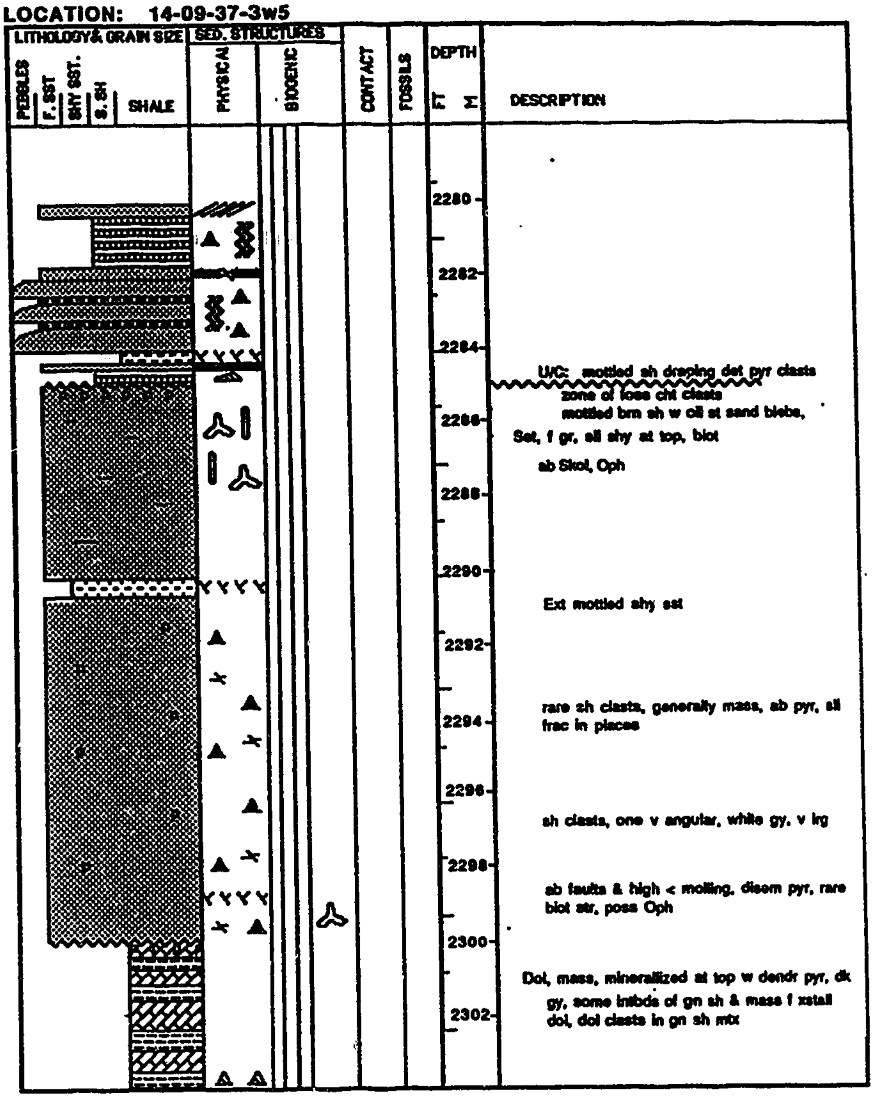

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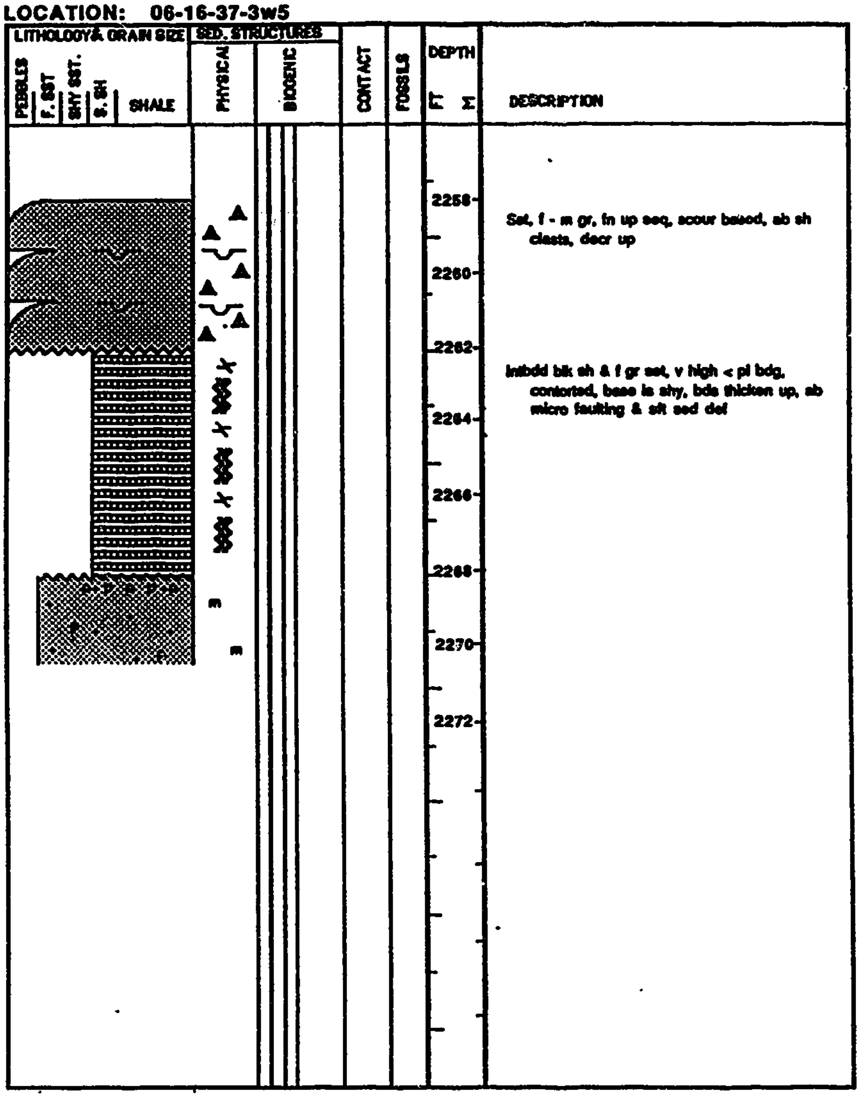

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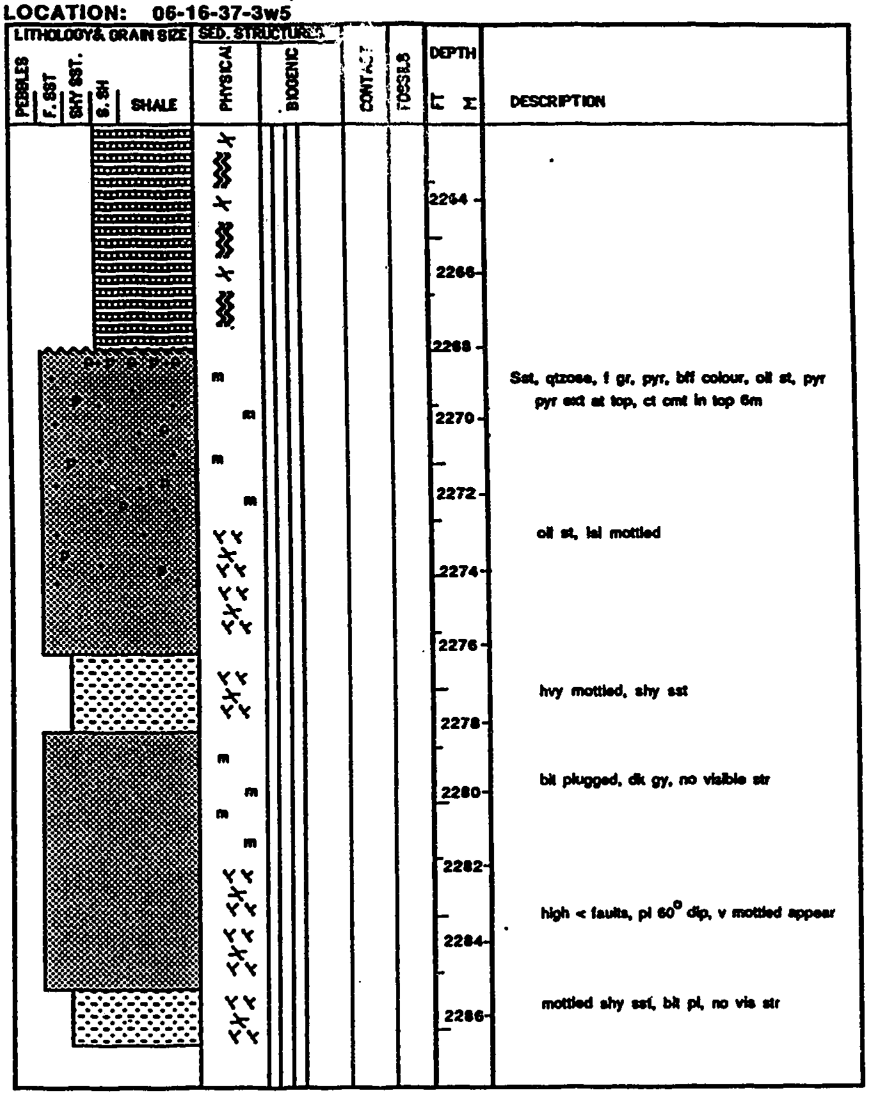

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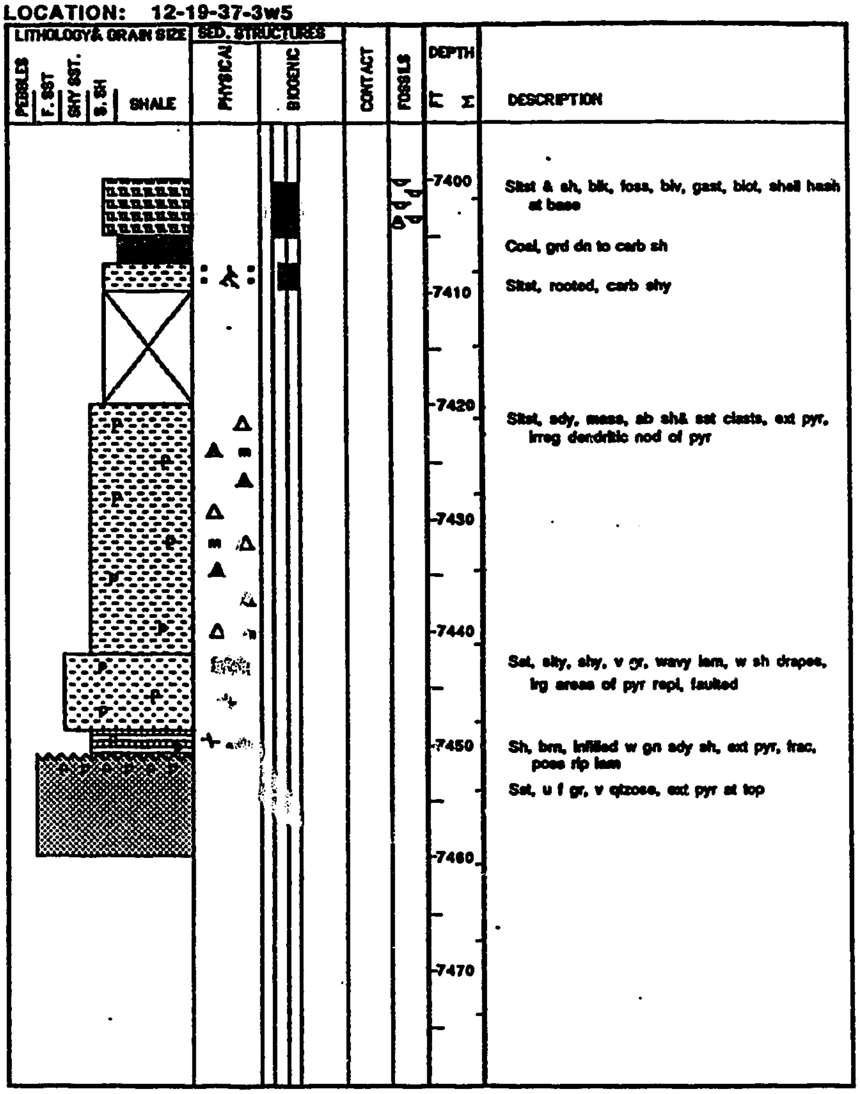

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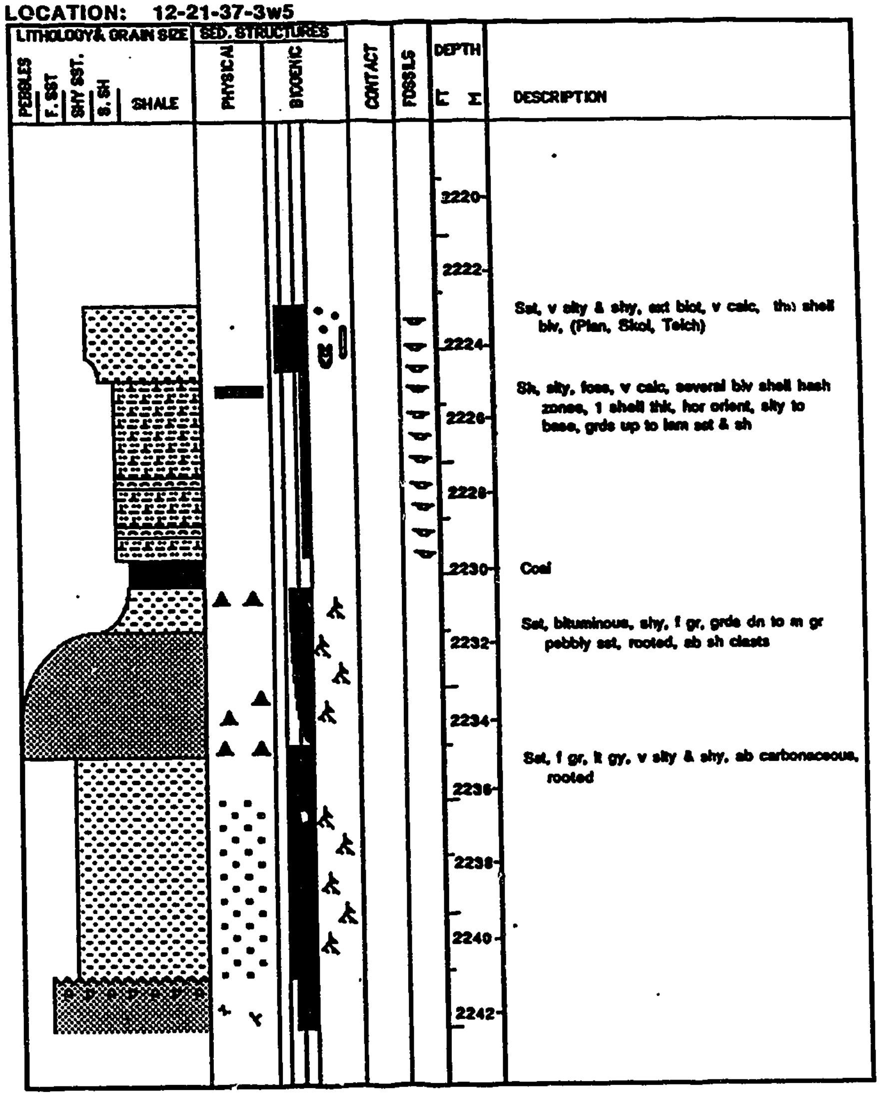

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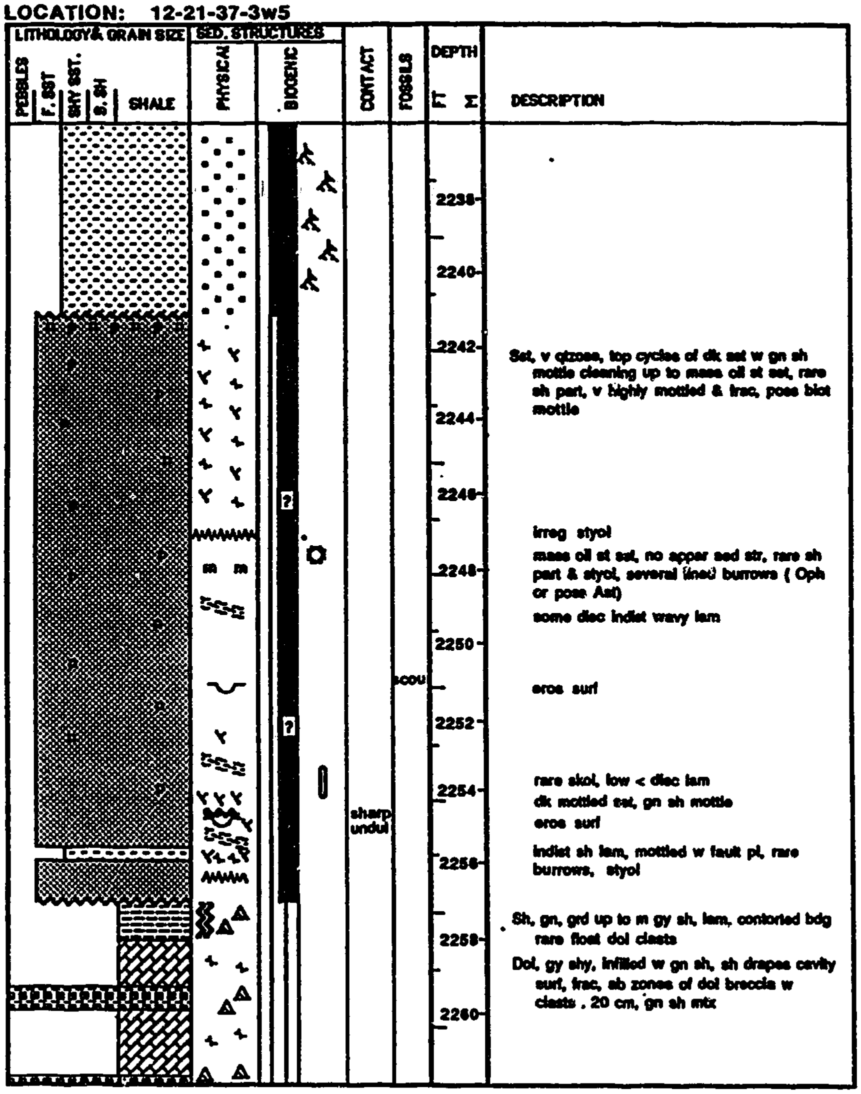

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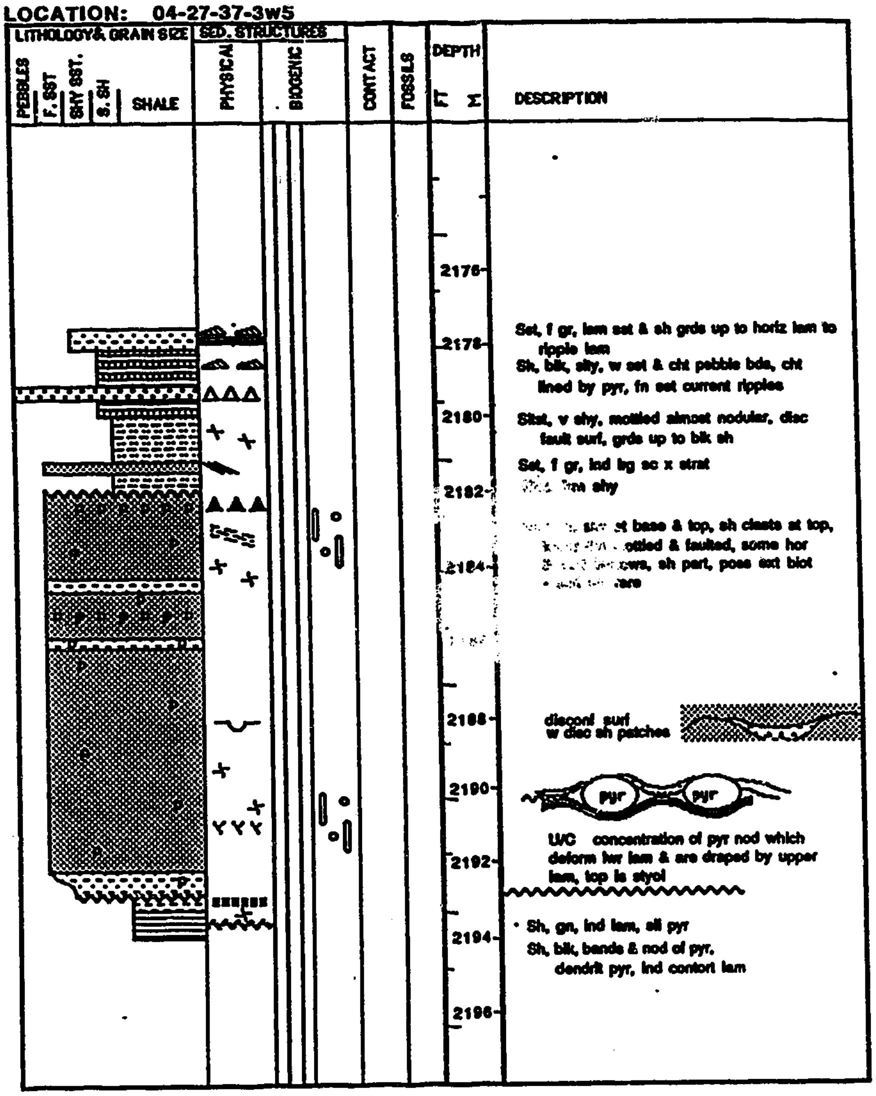

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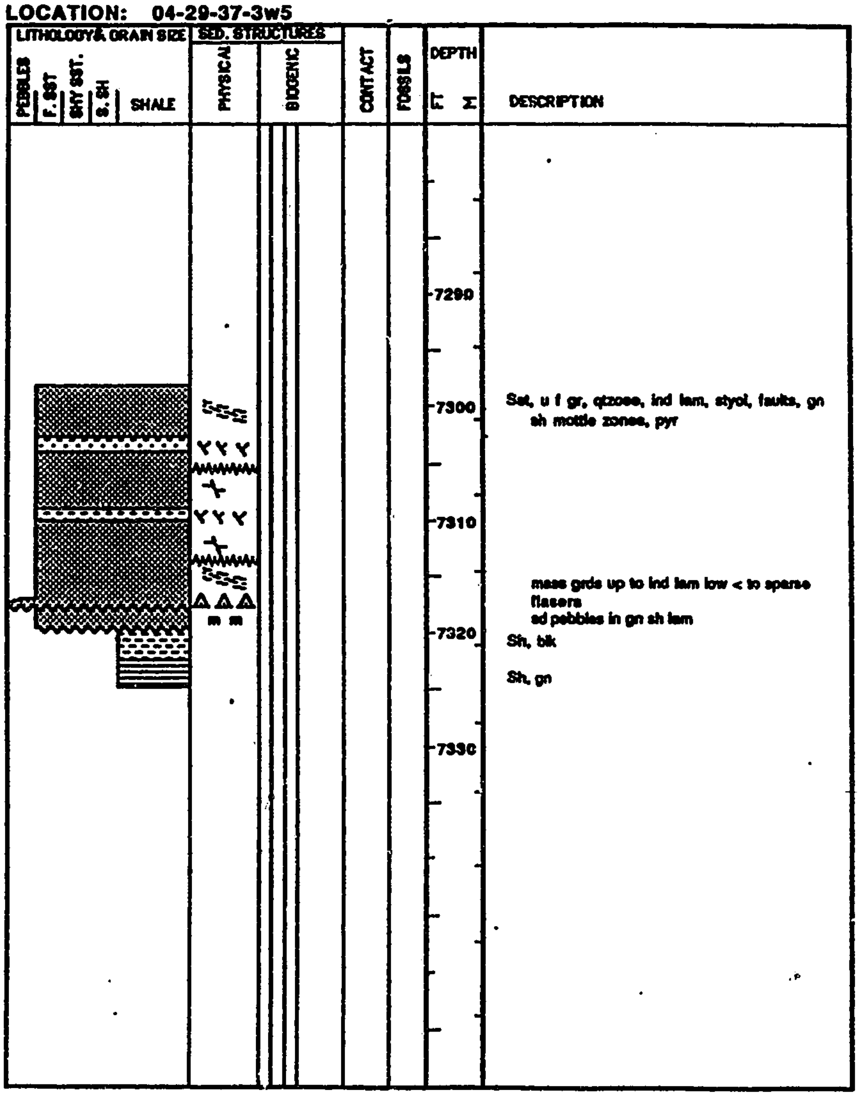

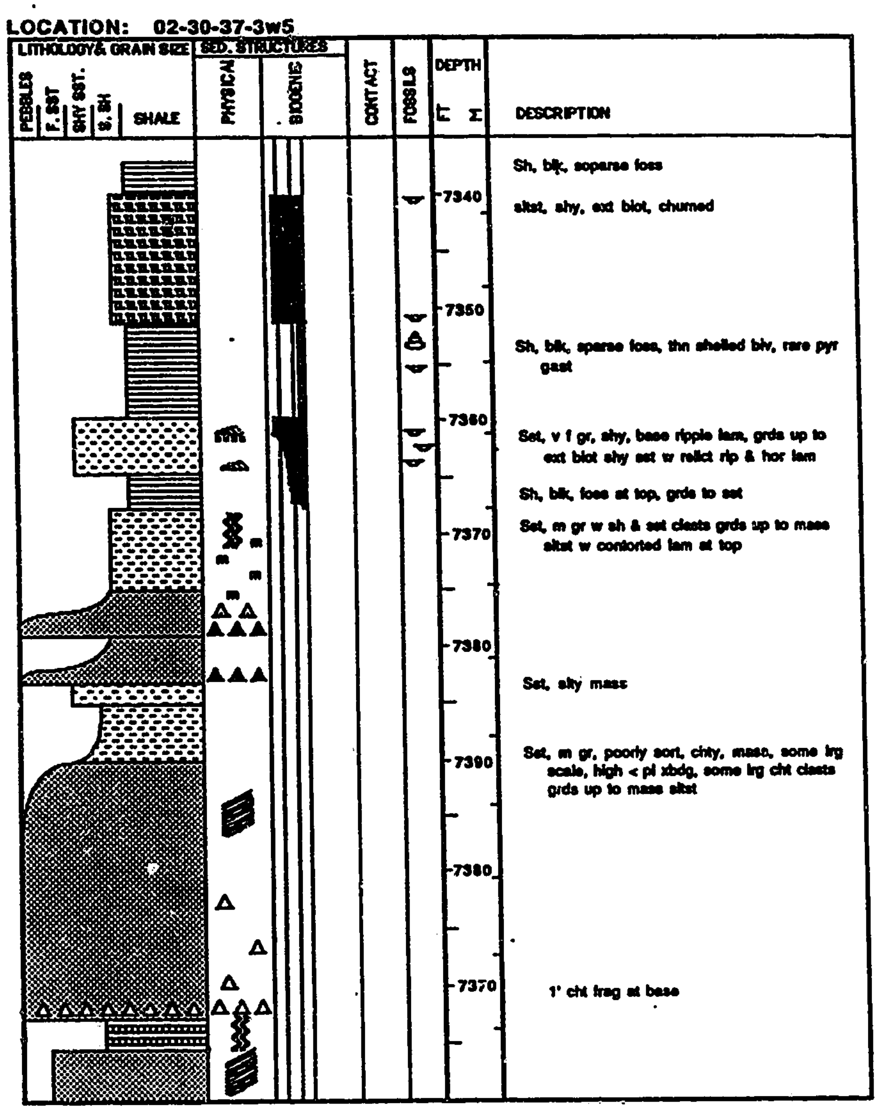
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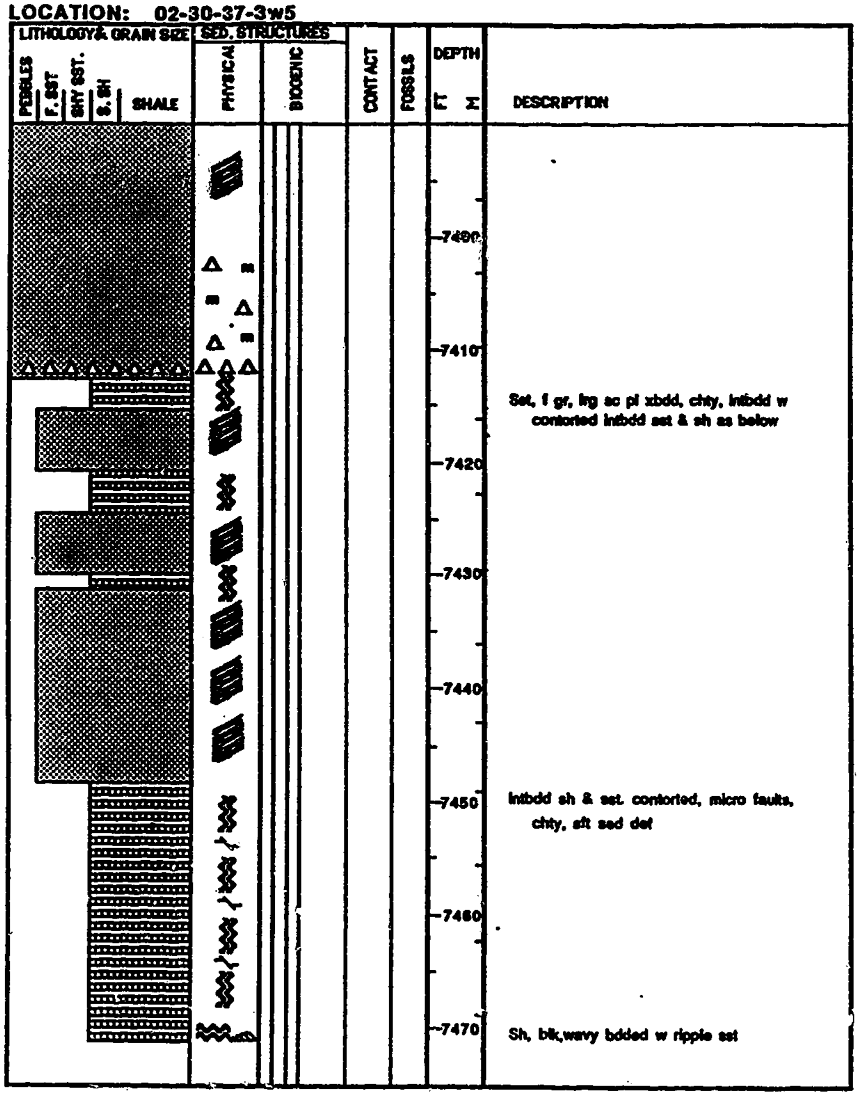
SED. STRUCTURES
PEBBLES
F. SST
SHY SST.
S. SH
SHALE
PHYSICAL
BIOGENIC
CONTACT
FOSSILS
DEPTH
FT M
DESCRIPTION
7340
7350
7360
7370
7380
7390
7380
7370
Sh, blk, soparse foss
sltst, shy, ext biot, churned
Sh, blk, sparse foss, thn shelled biv, rare pyr gast
Sst, v f gr, shy, base ripple lam, grds up to ext biot shy sst w relict rip & hor lam
Sh, blk, foss at top, grds to sst
Sst, m gr w sh & sst clasts grds up to mass sltst w contorted lam at top
Sst, slty mass
Sst, m gr, poorly sort, chty, mass, some lrg scale, high < pl xbdg, some lrg cht clasts grds up to mass sltst
1' cht frag at base

LOCATION: 02-30-37-3w5



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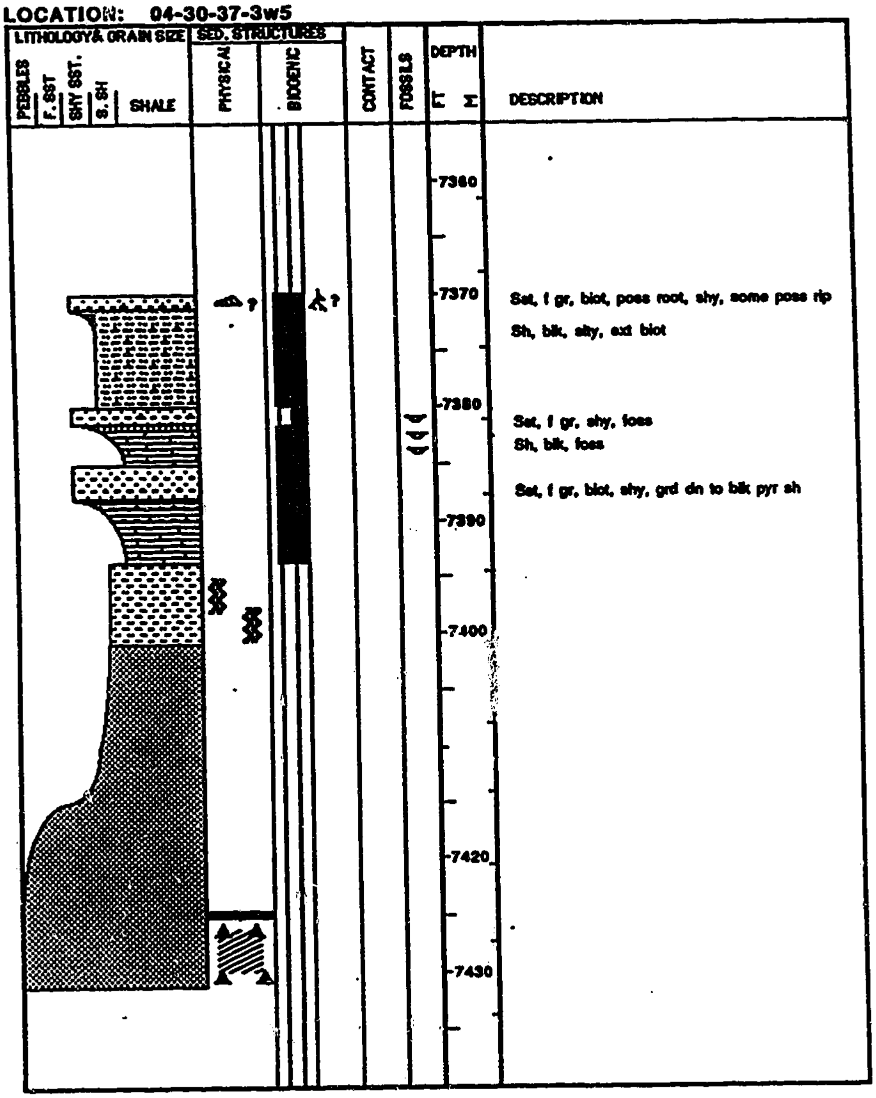

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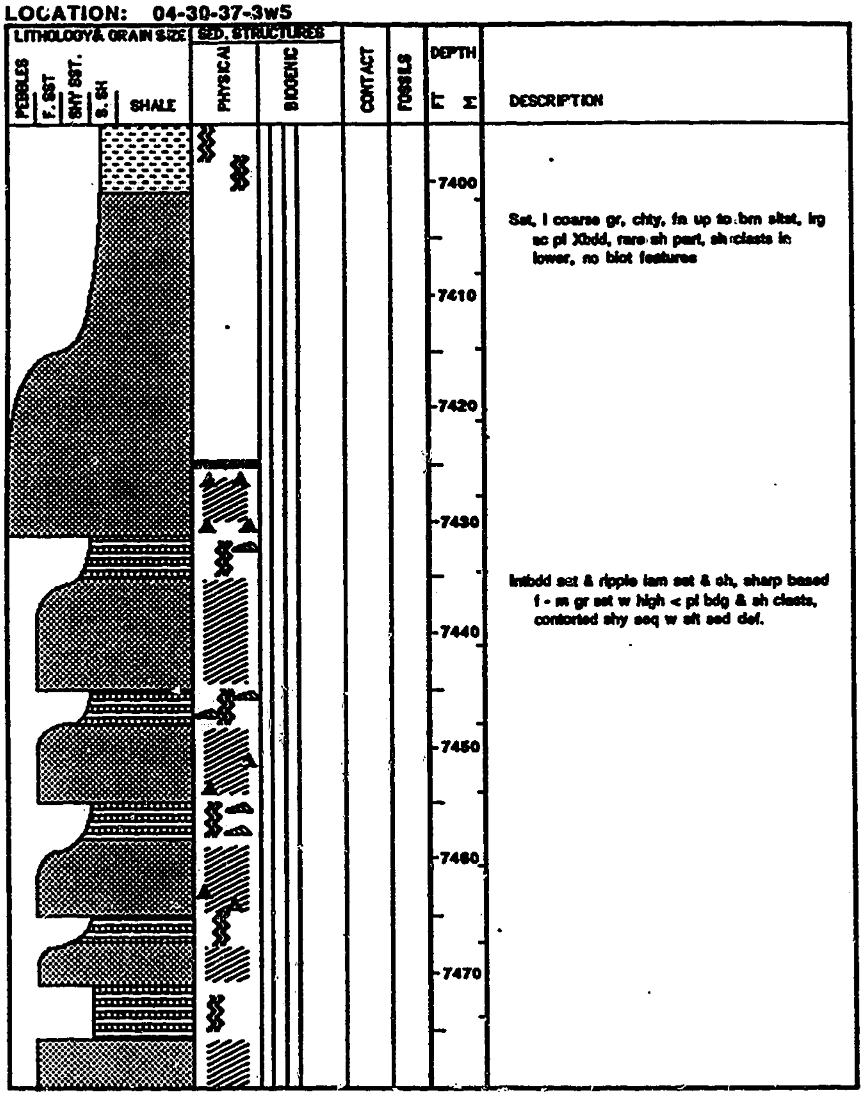

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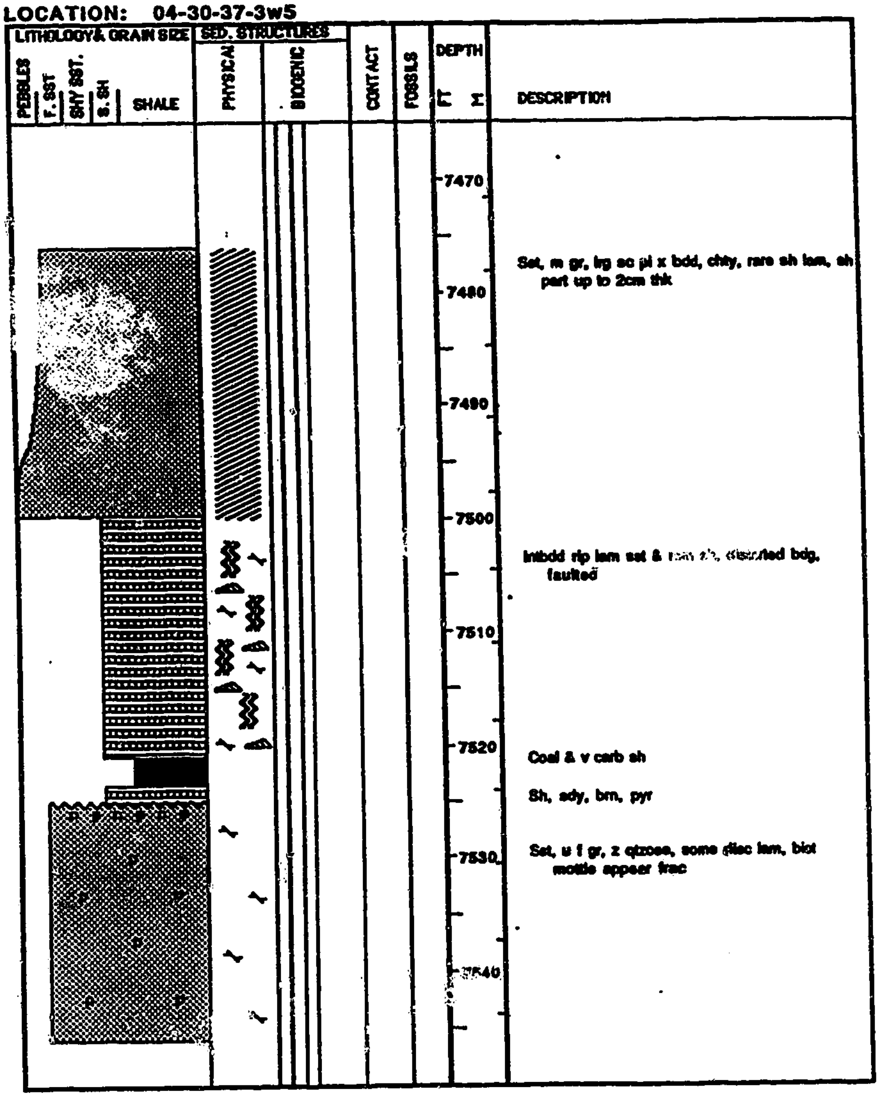

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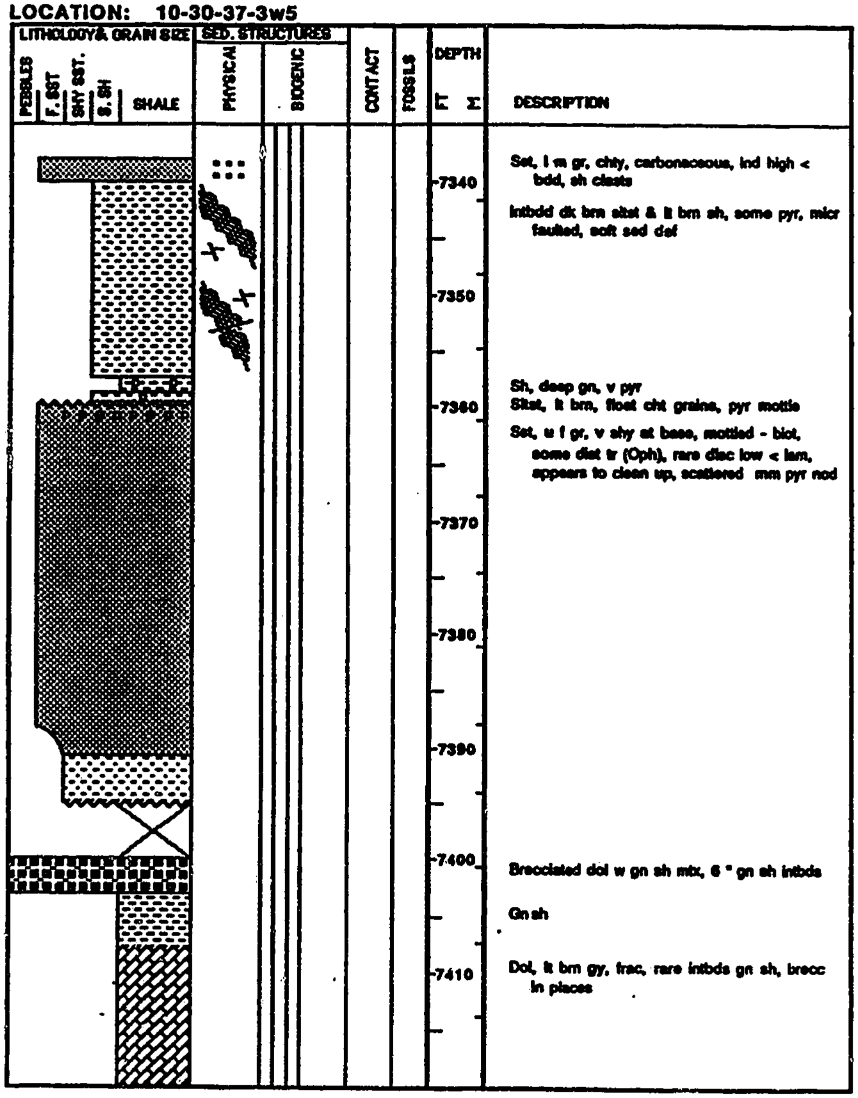

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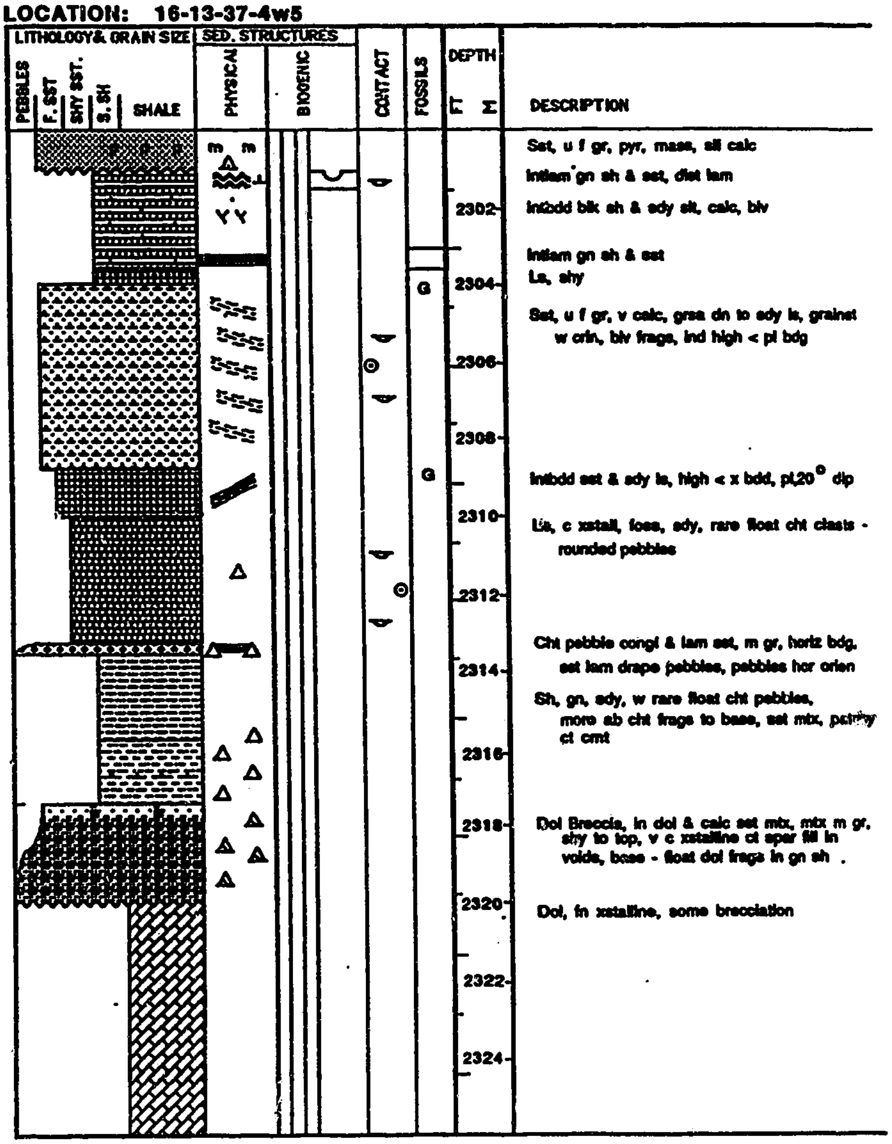

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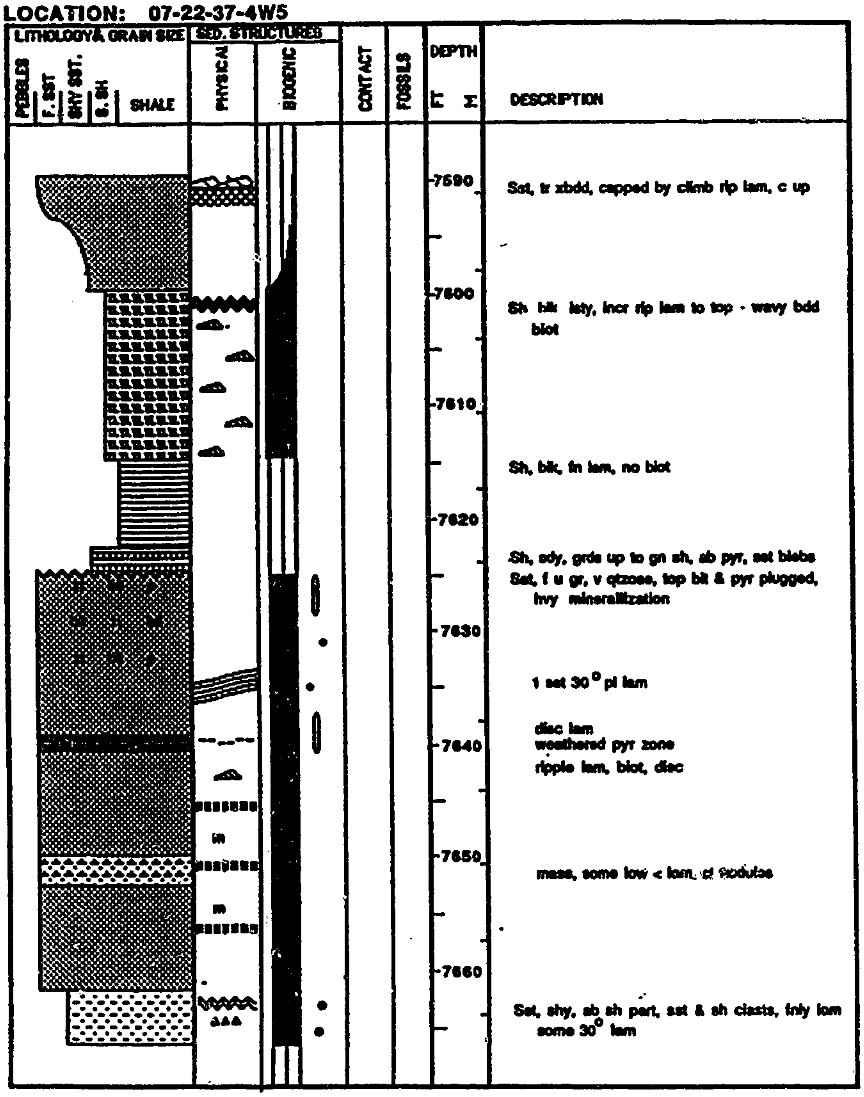

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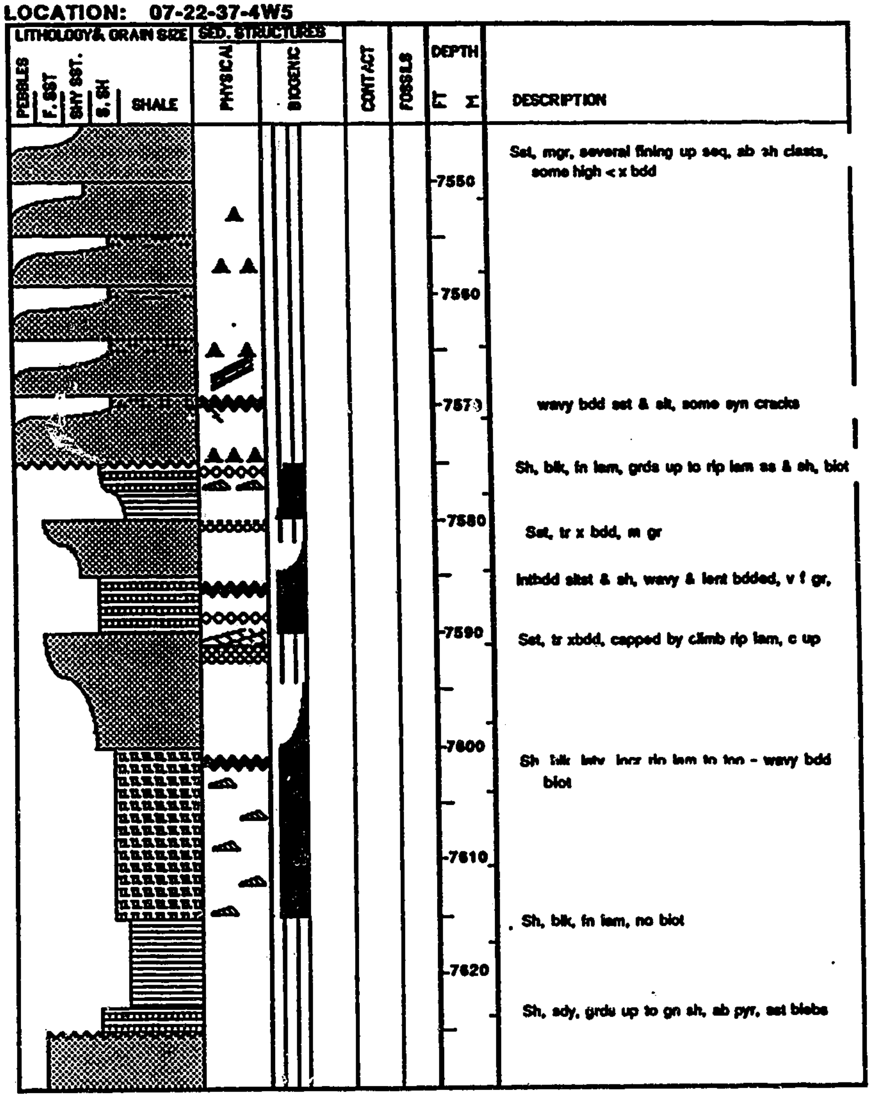

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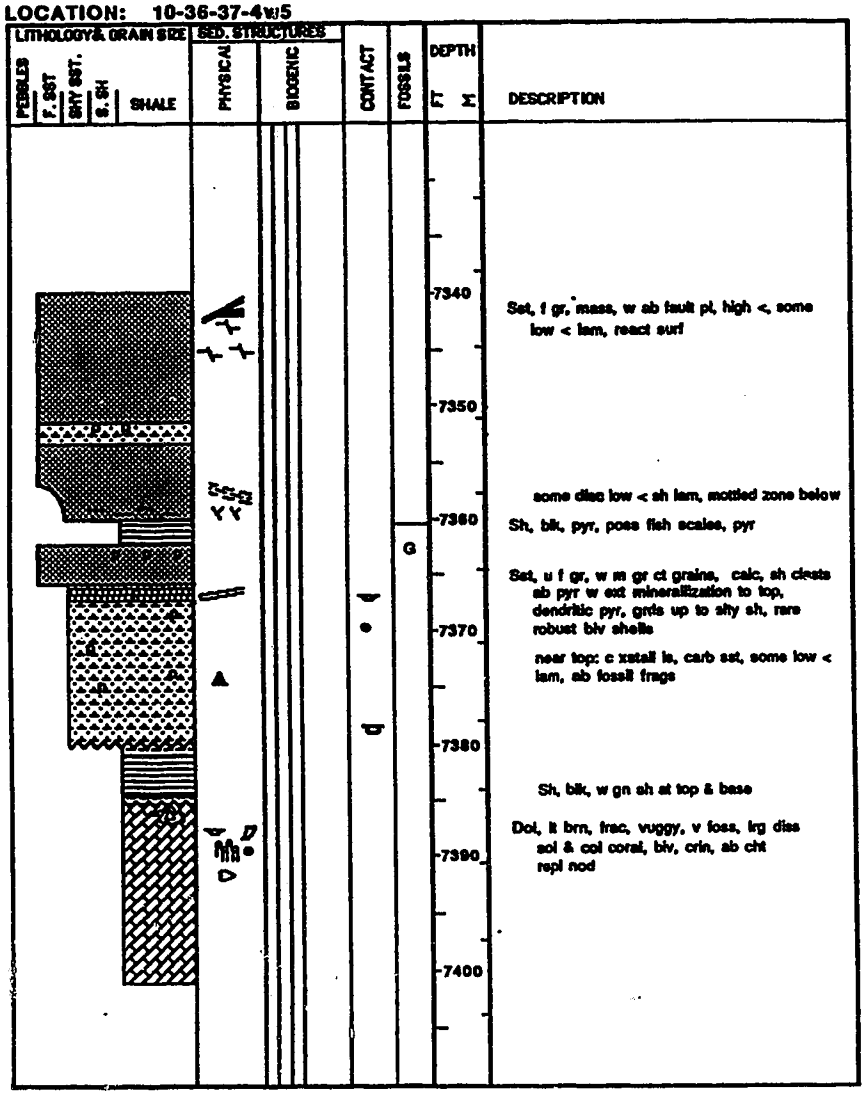

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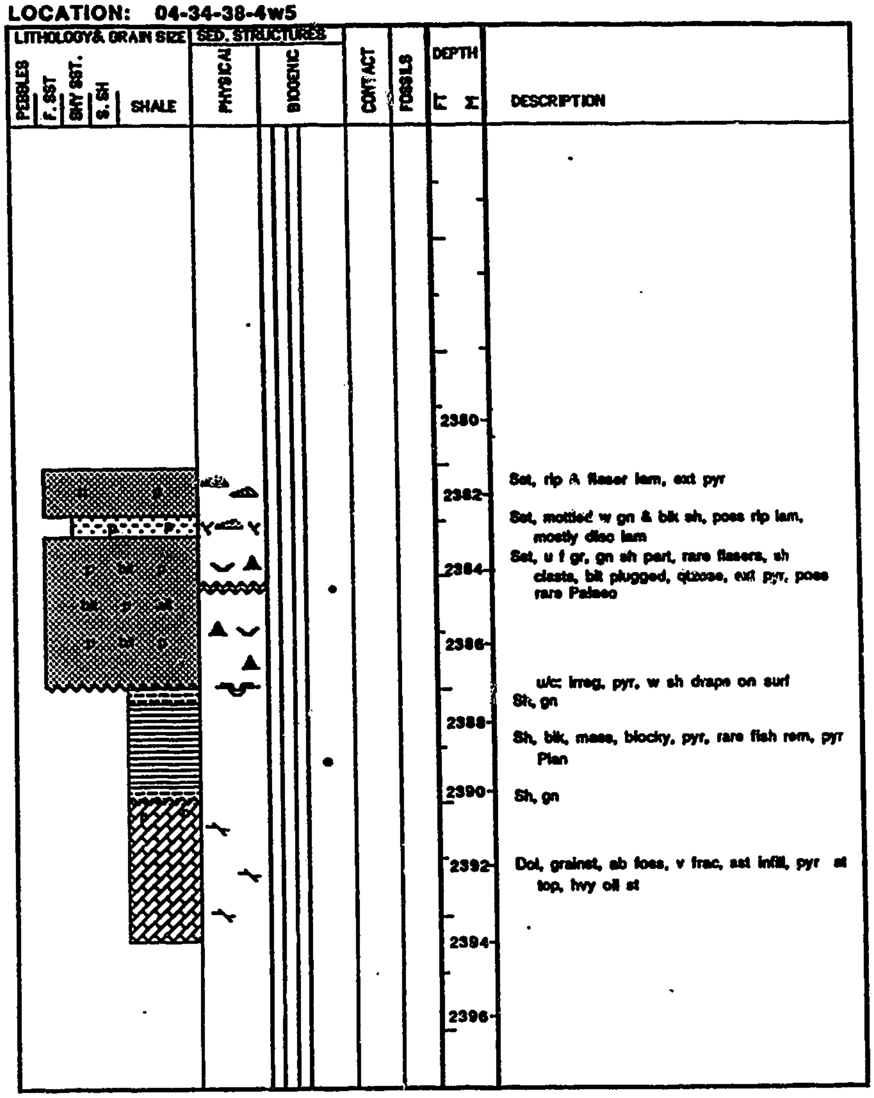

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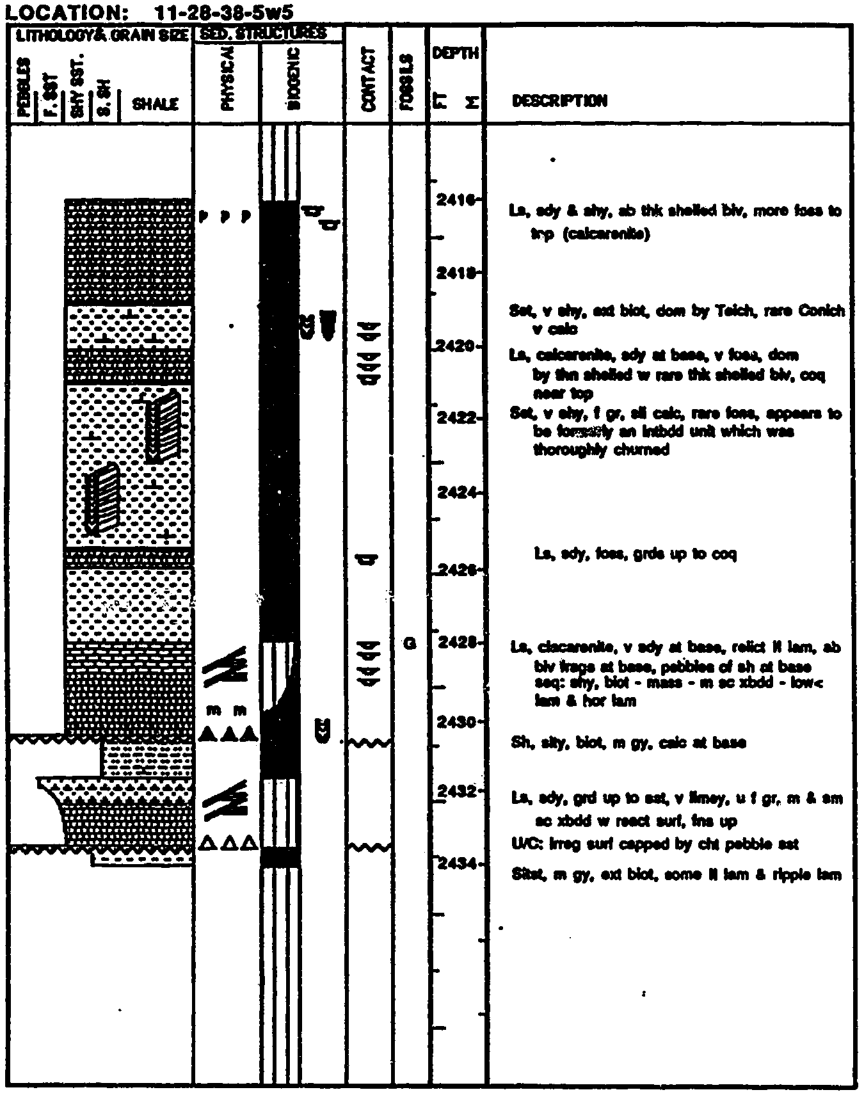

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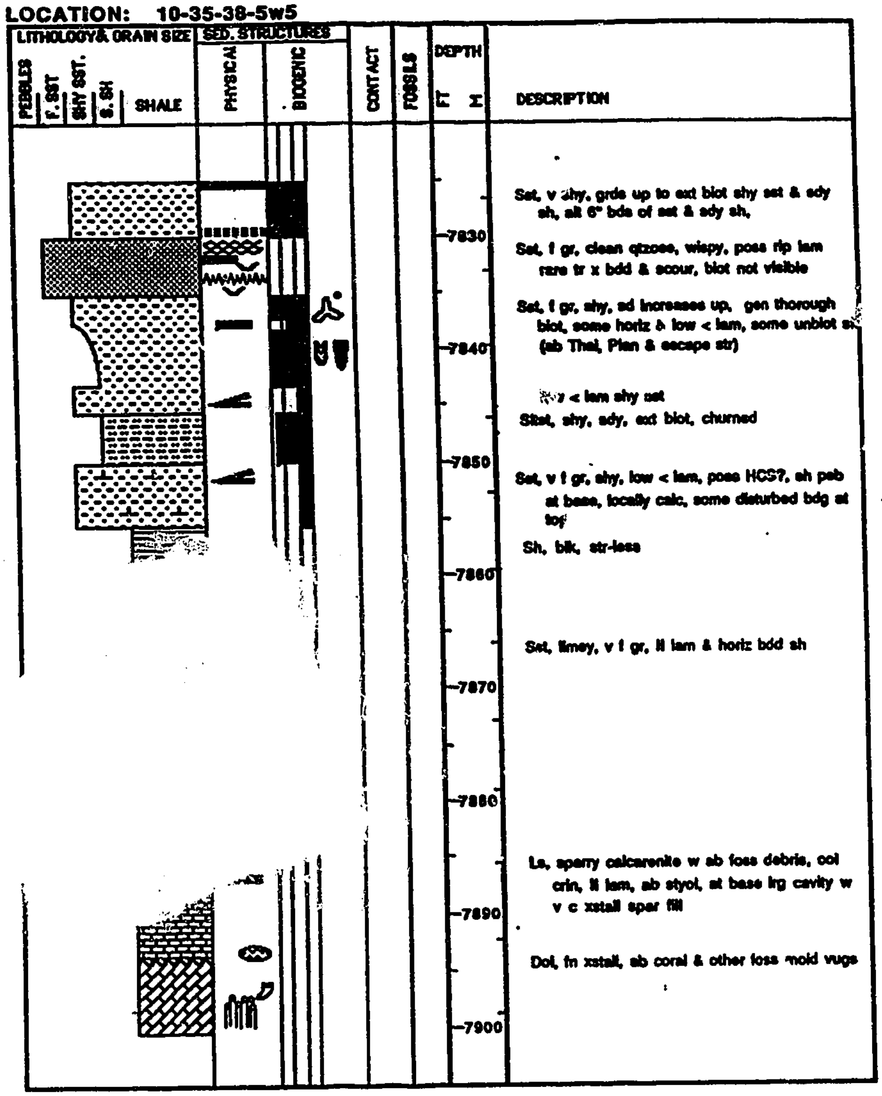
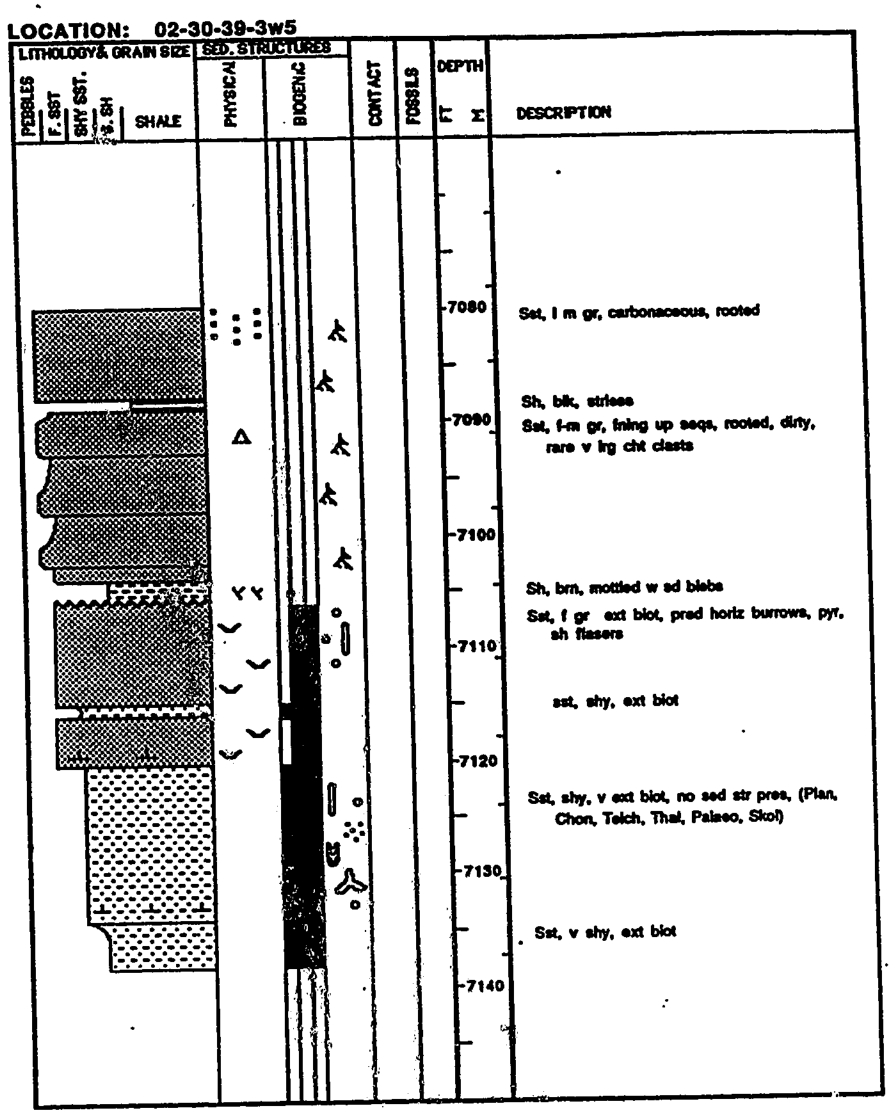
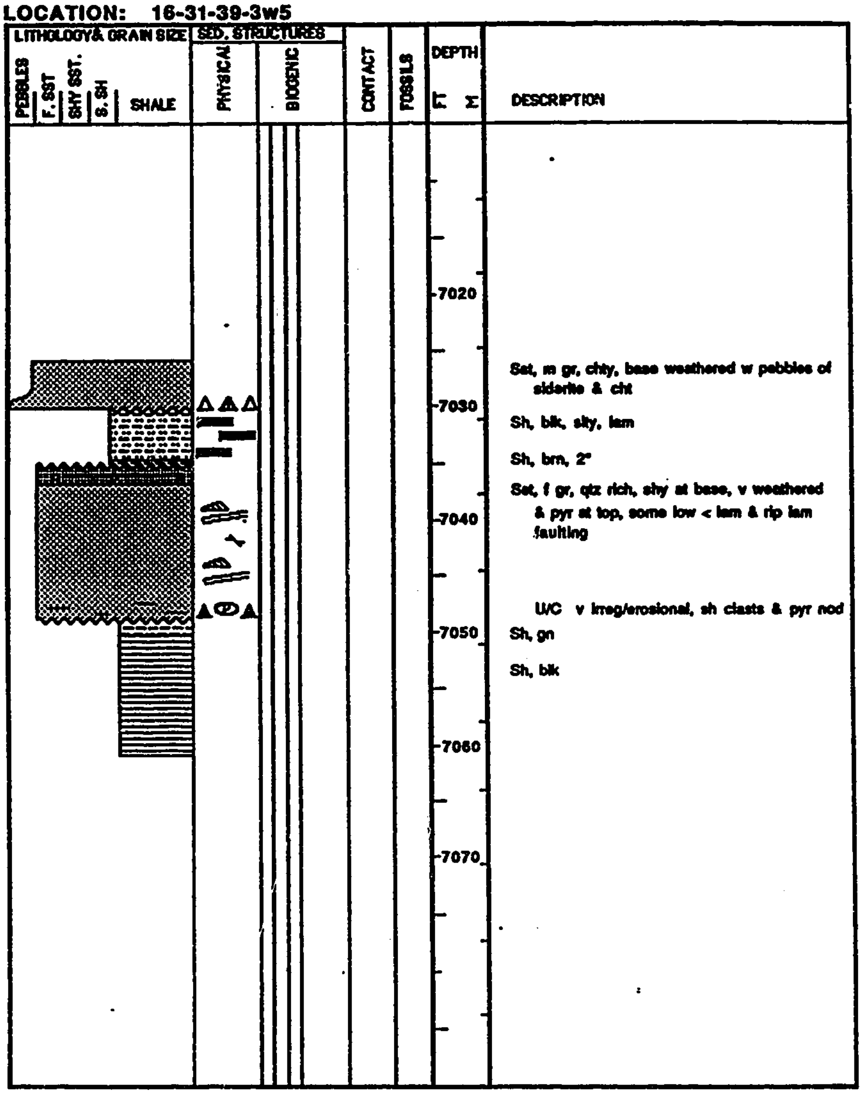



LOCATION: 02-30-39-3w5
LITHOLOGY& GRAIN SIZE
PEBBLES
F. SST
SHY SST.
SL. SH
SHALE
SED. STRUCTURES
PHYSICAL
BIOGENIC
CONTACT
FOSSILS
DEPTH
FT
M
DESCRIPTION
7080
7090
7100
7110
7120
7130
7140
Sst, l m gr, carbonaceous, rooted
Sh, blk, striae
Sst, f-m gr, fning up seqs, rooted, dirty, rare v lrg cht clasts
Sh, brn, mottled w sd blebs
Sst, f gr ext biot, pred horiz burrows, pyr, sh flasers
sst, shy, ext biot
Sst, shy, v ext biot, no sed str pres, (Plan, Chon, Teich, Thal, Palaeo, Skol)
Sst, v shy, ext biot

LOCATION: 16-31-39-3w5



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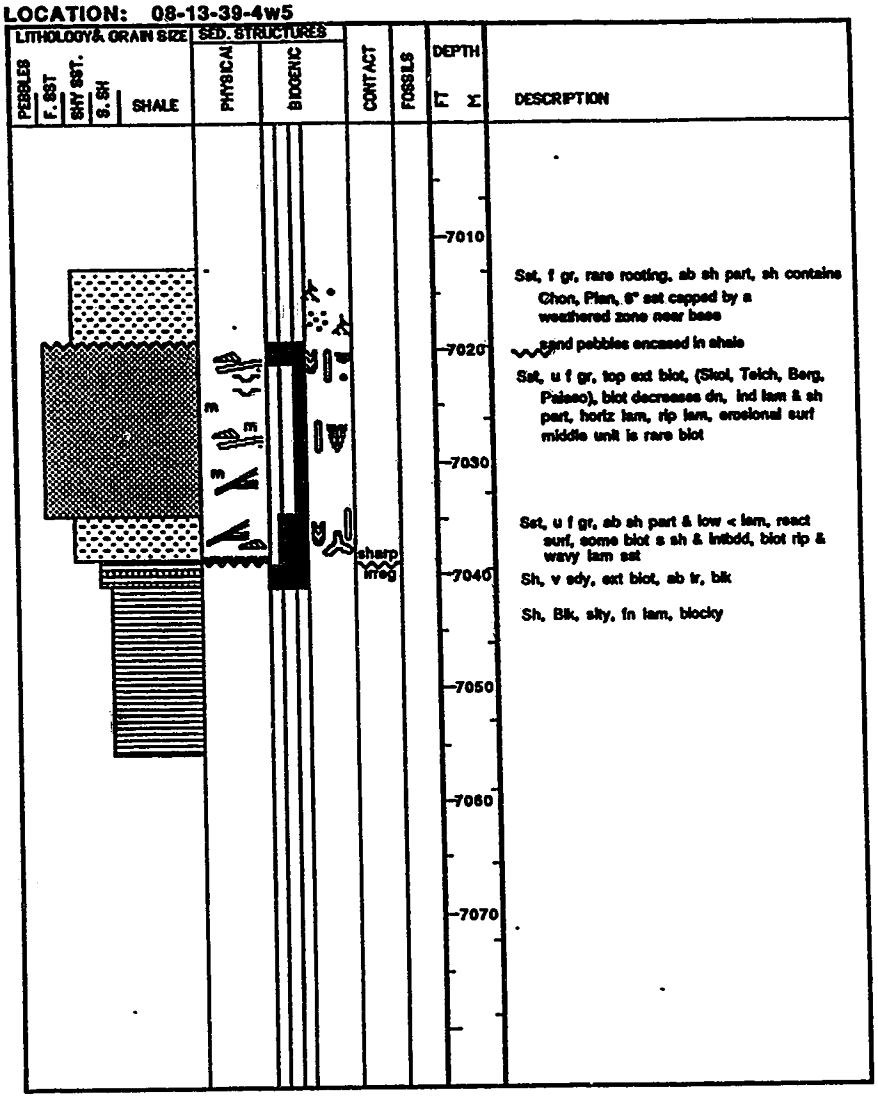

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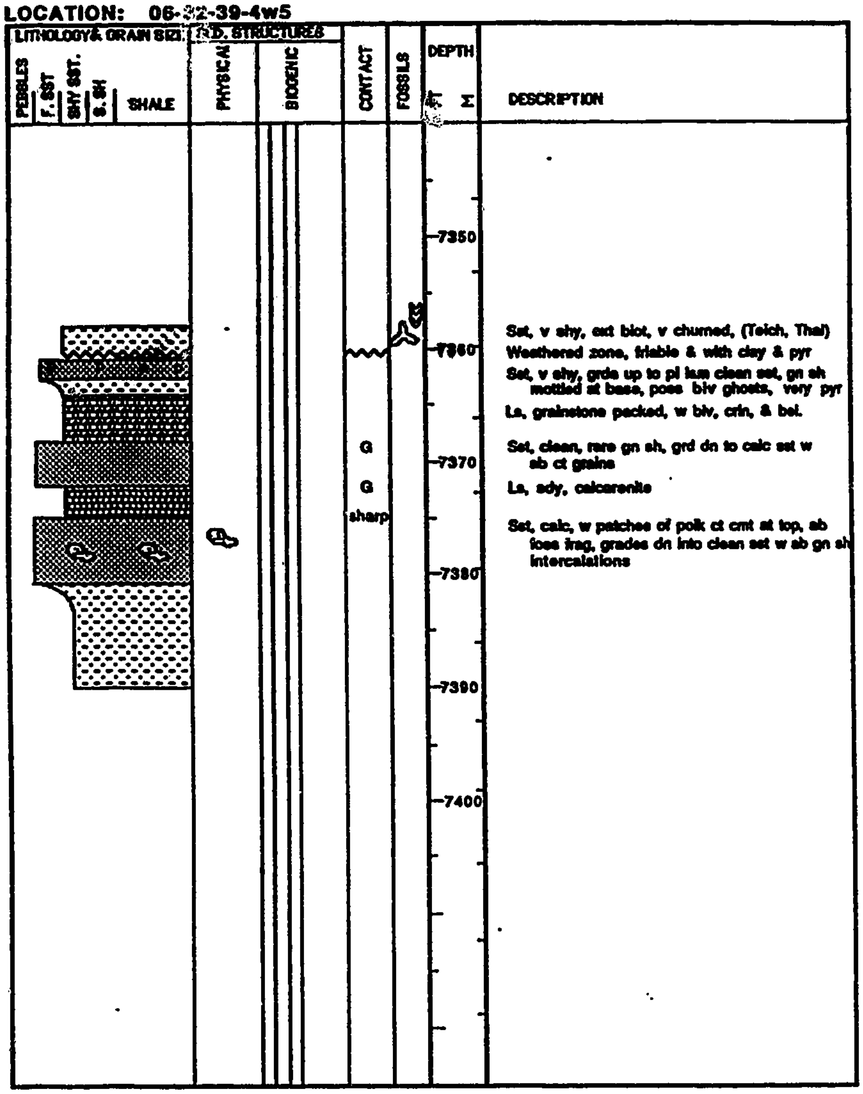

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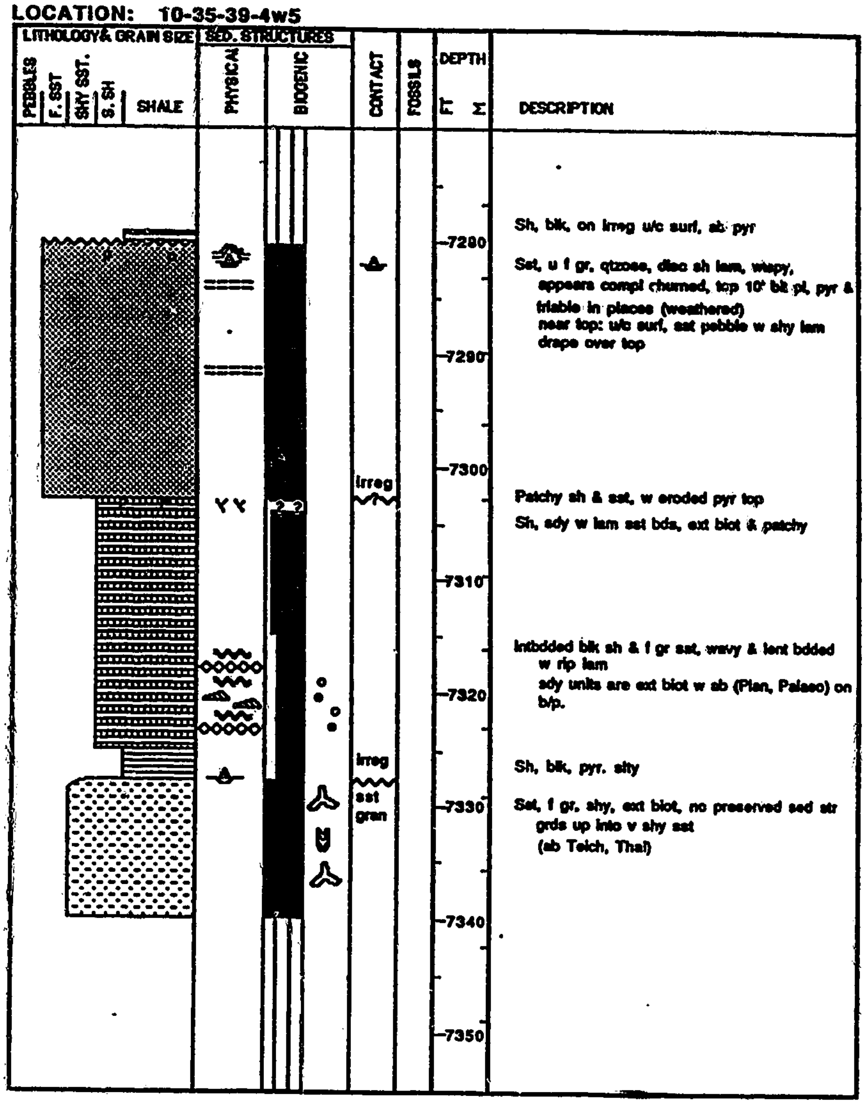

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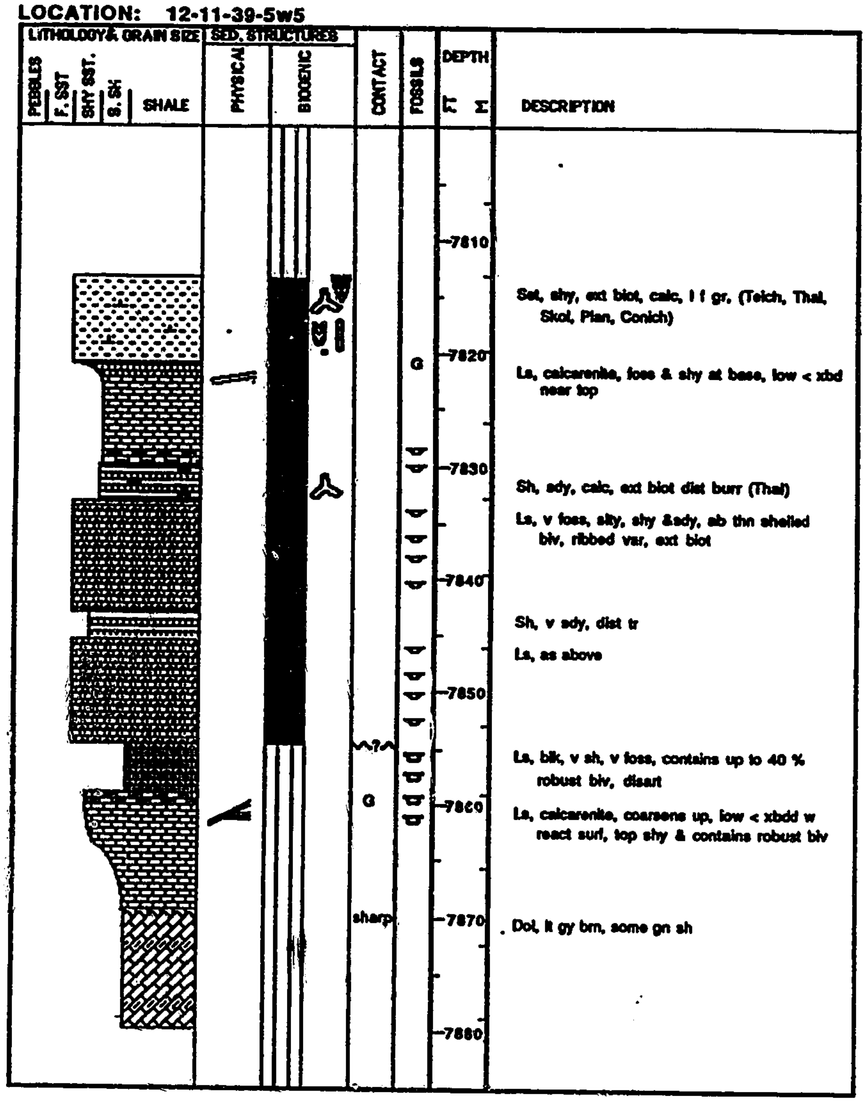

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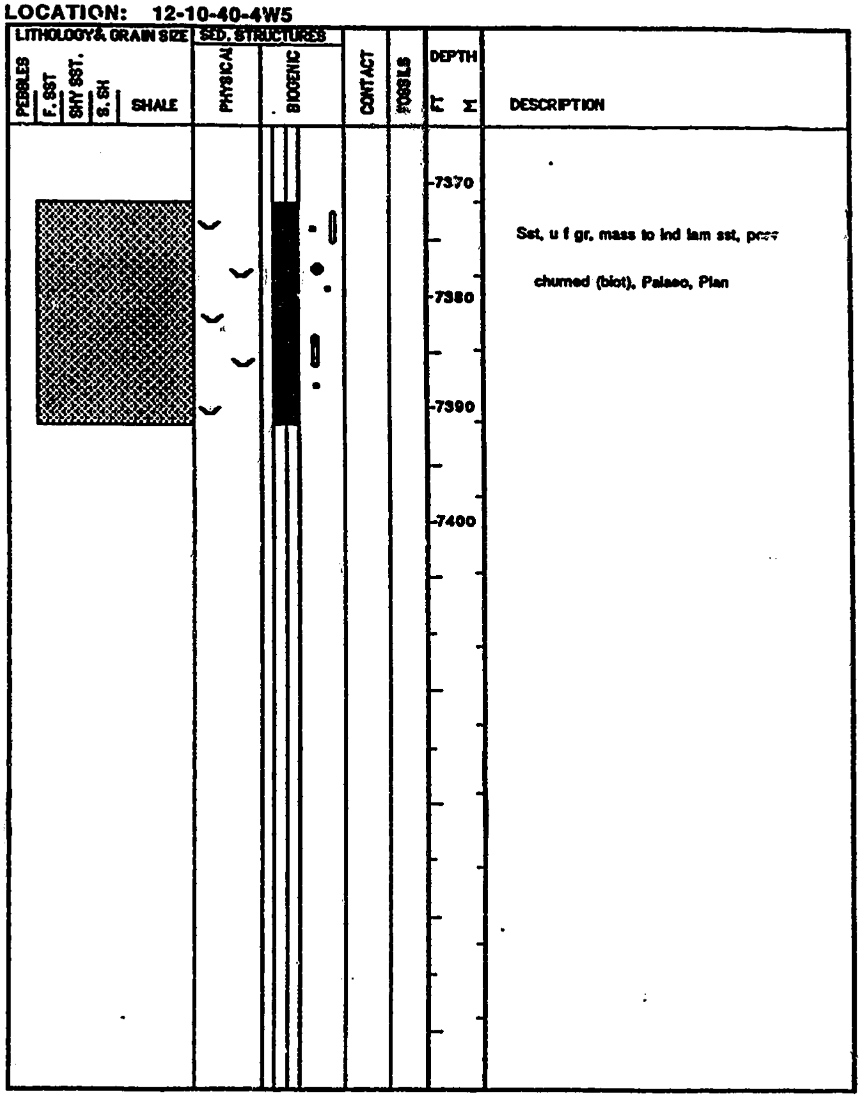

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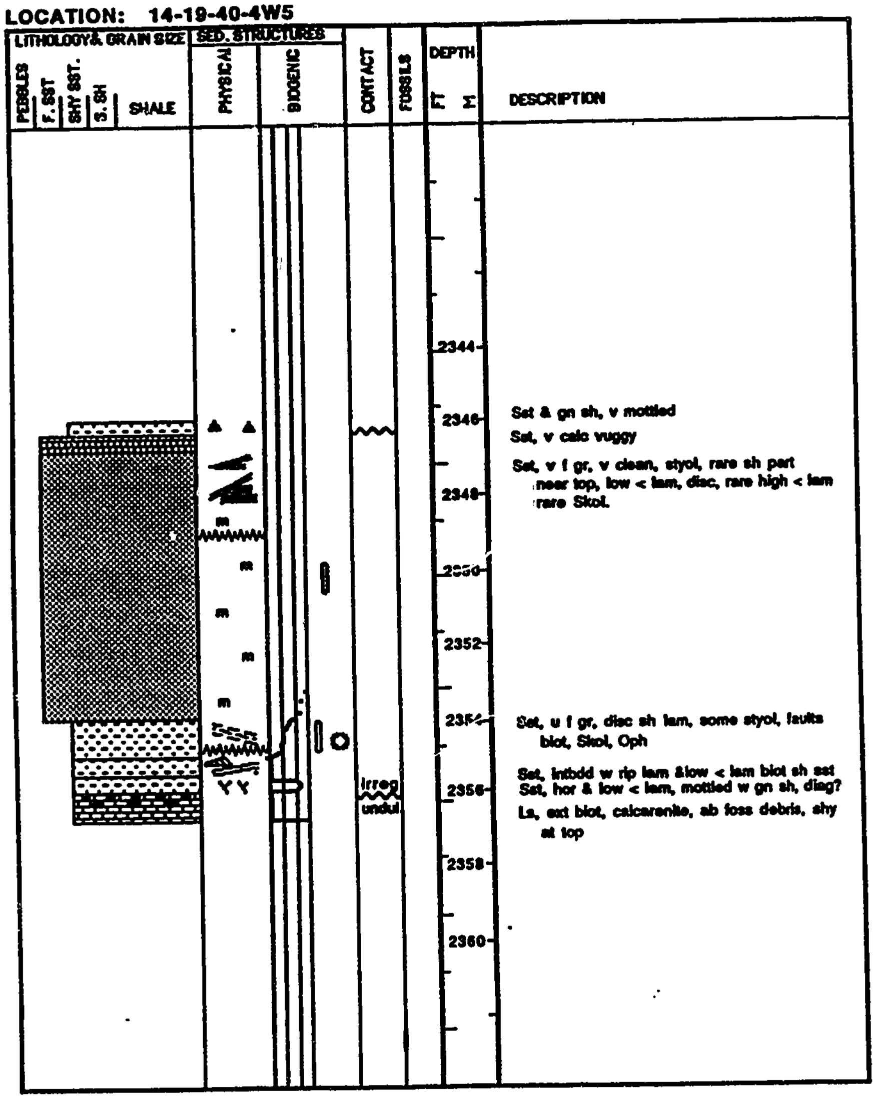

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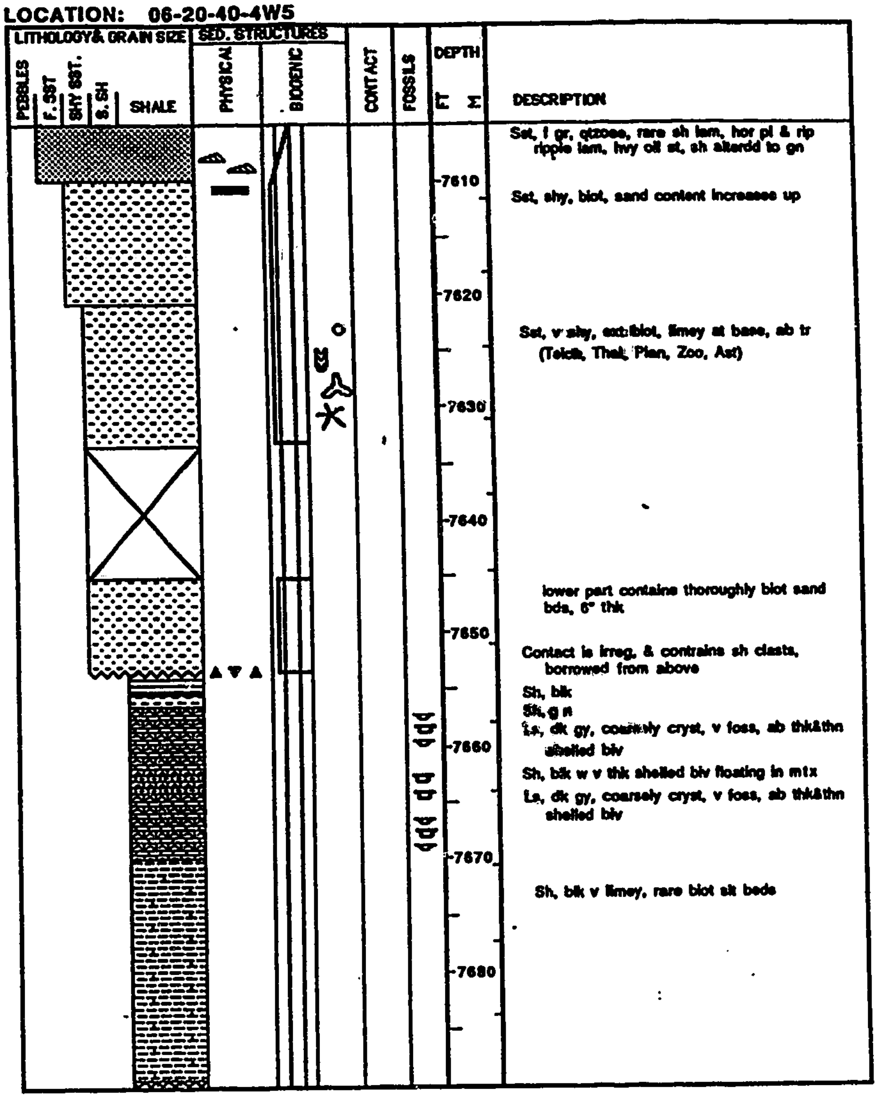

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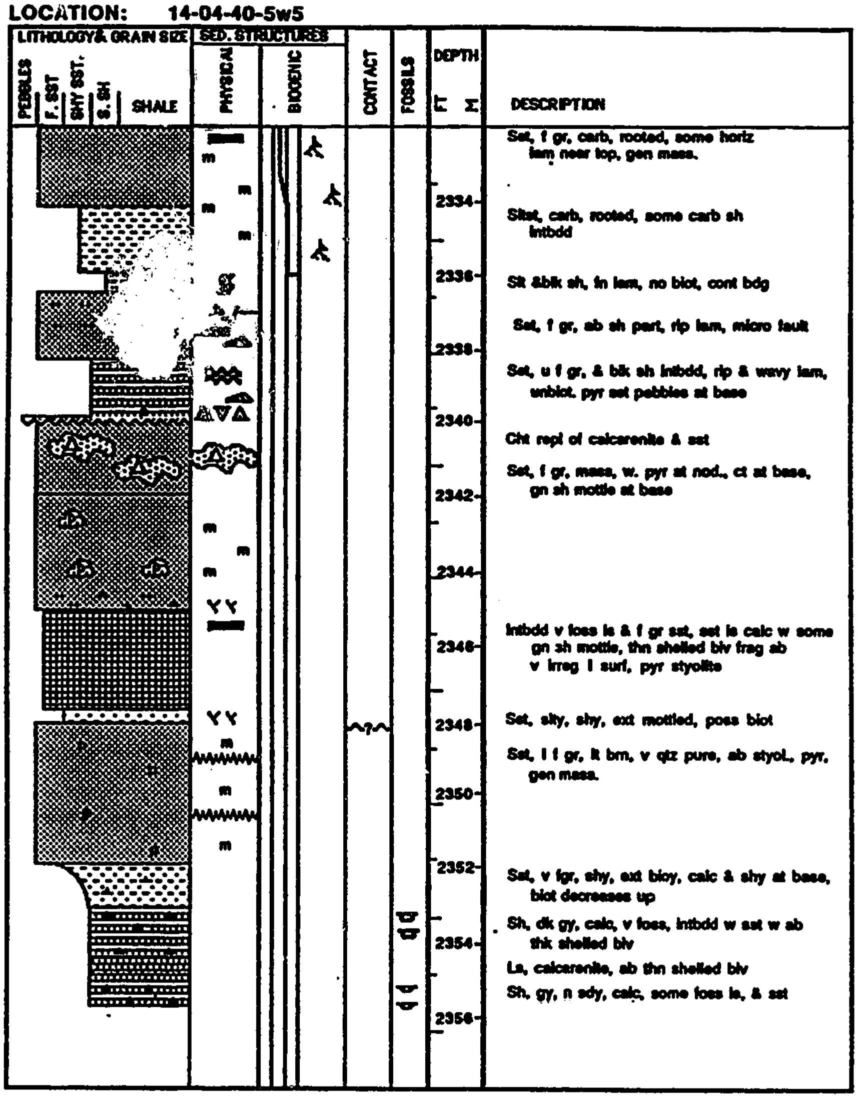

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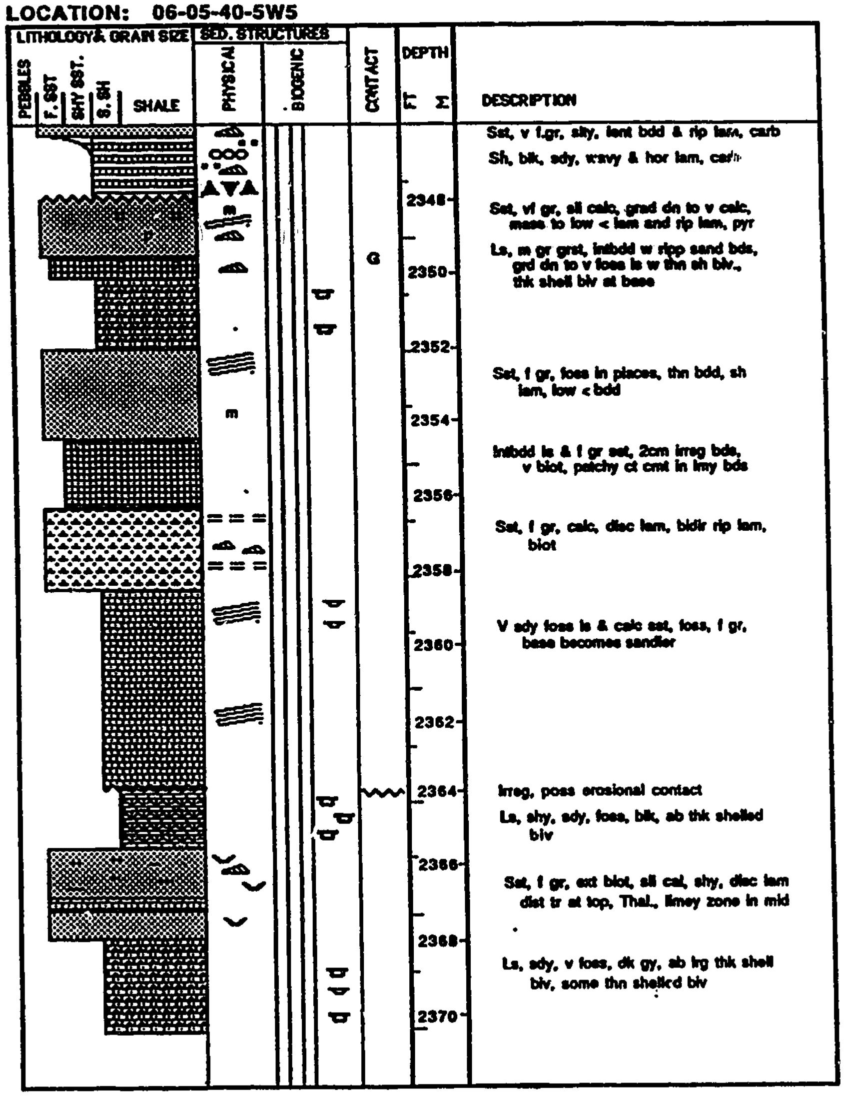

SAMPLE	LOCATION, DEPTH AND DESCRIPTION	QUARTZ	RF	CHERT	CARB	POROSITY	CLAY	ACCES
1001J	04-27-37-3w5 2189m	71	0.5	0	0	24.5	2.5	1.5
1002J	04-27-37-3w5 2186m pyr sst	81.5	1	0	0	11.5	0.5	5.5
1003J	04-27-37-3w5 2183m as above	77.5	1	0	0	7.5	13.5	0.5
1004E	04-27-37-3w5 2177m Ell. Micaceous sst.	67.5	5	4	0	5	18	0.5
1005J	12-21-37-3w5 2253m f gr shy sst	76.5	0.5	0.5	0	13	7	2.5
1007J	12-21-37-3w5 2249m clean oil st sst	76	0	0	0	23.1	0.5	0.4
1009J	12-21-37-3w5 2247m clean oil st sst	74	0	0	0	7.5	16.5	2
1010E	12-21-37-3w5 2244m tite pyr sst	46.5	2	7.5	0	1	41	2
1012J	12-21-37-3w5 2234m Ell carb sst	74	1	0	0	11	5	9
1015J	08-16-37-3w5 2281m mottled sst	53.5	0	0	0	1.5	0	45
1016E	08-16-37-3w5 2267m pyr sst top Jr	52.5	6.5	13.5	0	10	16.5	1
1017E	08-16-37-3w5 2259m Ell. oil st sst kaol	49.5	9	20	0	2	18	1.5
1018J	08-16-37-3w5 2258m Ell. kaol sst carb	81.5	0	0.5	0	10.5	7.5	0
1019J	14-09-37-3W5 2300m l contact	83	0	0	0	12.5	4.5	0
1020J	14-09-37-3W5 2289 clean sst	85	0.5	0.5	0	13	1	0
1021J	14-09-37-3W5 2281 clean sst	57.5	0	0	0	1	25.5	16
1023J	14-09-37-3W5 2288m pyr sst at top u/c	79	0	0	0	14	1	6
1024J	14-09-37-3W5 2283.5m Ell sst	76.5	0	0	0	14.5	0.5	8.5
1025J	10-30-37-3W5 7395' shy sst	79.5	0	1	0	11.5	3	5
1026J	10-30-37-3W5 [illegible]' cl sst	84.5	0	0	0	9.5	2	4
1027E	10-30-37-3W5 7371' cl sst	53	5	22.5	0	7.5	5	7
1028E	10-30-37-3W5 7362' upper cl sst	65.5	11	11	0	1	10.5	1
1029E	10-30-37-3W5 7340' Ell sst	73.5	12	1.5	0	11.5	1.5	0
1030E	10-20-38-3w5 7316' l sst	39	12	23	0	3	10	13
1030bE	10-20-38-3w5 7305' top xbdd sst	80.5	10	6.5	0	2	1	0
1031D	04-27-37-3w5 7147' shy carb sst	75.5	3.5	0.5	0	16	2	2.5
1039E	04-35-39-3w5 7040' cht breccia	76.5	6.5	0	0	15	2	0
1069J	12-15-39-3W5 7021' base of sst wh flasers	60.5	0	0	30	1	1	7.5
1070J	12-15-39-3W5 [illegible] cl oil st sst	50	1.5	0	30.5	1	3.5	13.5
1071J	10-36-37-4W5 7378' [illegible] calc sst	17	0.5	0	78	1	0	3.5
1072J	10-36-37-4W5 7376' as above	72	0	0	0	11.5	8.5	8
1073J	10-36-37-4W5 7358' ls c crystall	84	0	0	0	7.5	6.5	2
1074J	10-36-37-4W5 7358' vuggy sst	69	1	0	6	11.5	12	0.5
1075J	10-36-37-4W5 7343' lam sst	54	0.5	0	41	2	0	2.5
1076J	16-13-37-4W5 2310m c xstall w pebb	62.5	0	0.5	33.5	1.5	0	2
1077J	16-13-37-4W5 2308m cl oil st sst	57	0	0	23.5	15	0	4.5
1078J	16-13-37-4W5 2307m cl oil st sst	30	0	0	52	0.5	0.5	17
1079J	16-13-37-4W5 2306m oil st carb sst	68.5	0	0	0	20.5	7.5	3.5
1080J	16-13-37-4W5 2300m u sst	85.5	0	0	0	9.5	1	4
1083E	07-22-37-4w5 7576'	64	7.5	6.5	0	8	12.5	1.5
1084E	07-22-37-4w5 7556'	38.5	4.5	31	0	3.5	18.5	4
1085E	04-30-37-3w5 7455' Ell thn sst	37.5	9.5	35	0	5.5	11	1.5
1086J	04-30-37-3w5 7425' chan sst	92	0	0	0	7	1	0
1087E	04-30-37-3w5 7500' l chan sst	37	0	38	0	7	12.5	5.5
1088E	04-30-37-3w5 7540' lur sst	24	6.5	47.5	0	9	11.5	1.5
1089E	02-30-37-3W5 7415' Chan sst	27	7	37.5	0	1	20.5	7
1090RC	02-30-37-3W5 7405' chan sst	74.5	0.5	0	12	10.5	0	2.5
1091RC	02-30-37-3W5 7367' Ell sst	36.5	0.5	0	59.5	0	0	3.5
1092RC	11-28-38-5w5 2430m RC sst	68	0.5	0	15.5	5.5	0	10.5
1093RC	11-28-38-5w5 2419m calcarenite	84.5	0.5	0.5	6.5	6	1	1
1094RC	11-28-38-5w5 2431m lwst sst	70.5	0	0	23.5	4	0	2
1095RC	06-05-40-5w5 2358.6m 3rd RC sst	78	1	0.5	0	13	1.5	6
1096RC	06-05-40-5w5 2452.5m mid RC sst	80	1.5	0	4	11	0	3.5
1097RC	06-05-40-5w5 2348m Upper RC sst	65.5	1.5	0	13	18	1.5	2.5
1099RC	12-11-39-5w5 7825'	86.5	0	0	1.5	7.5	0.5	4
1100RC	14-04-40-5w5 2317m ost sst	81	1	0.5	0	9.5	1.5	6.5
1101RC	14-04-40-5w5 2352m mass sst	86	0	0.5	2.5	7	1.5	2.5
1102RC	14-04-40-5w5 2351m mass sst	64	0.5	0	0	1.5	31	3
1103RC	14-04-40-5w5 2349m mass sst	85.5	1	0.5	0.5	10	1	1.5
1105E	14-04-40-5w5 2340m u mass sst	49	10	18	0	16.5	2.5	4
1106E	14-04-40-5w5 2337m cht ls/sst	34	1.5	4.5	0	1.5	58	0.5
1107E	08-09-37-3w5 Ell oil stsst	49.5	0	1	0	0	47.5	2
1108RC	14-16-37-3w5 2240m Ell sst cl	77.5	0.5	0	0	13	0.5	8.5
1109RC	14-16-37-3w5 2245m Ell kaol pl sst	72.5	1	0	0	14	11.5	1
1110RC	04-34-38-4w5 2285m m sst	86	1	0	0	12	1	0
1111RC	04-34-38-4w5 2275m rippled sst	87.5	0.5	0	0	8.5	3.5	0
1112RC	14-19-40-4w5 2354m	90	0.5	0	0	7.5	2	0
1114RC	14-19-40-4w5 2349m	80	1	0.5	0	15	2.5	1
1115RC	14-19-40-4w5 2347m cht ls/ssr	72	1.5	0	0	12.5	14	0
1116RC	10-35-39-4w5 7300'	83	0.5	0	0	15	0.5	1
1117E	08-13-39-4w5 7025'	63.5	18	5	0	10	3	0.5
1118RC	08-13-39-4w5 7020' oil st sst	71	1	0	0	26.5	1	0.5
1121E	16-31-39-3w5 7021' RC cl sst top	79.5	0.5	0	0	18.5	0.5	1
1122RC	16-31-39-3w5 7061' Ell sst	48	12	20	0	1.5	16	2.5
1154E	10-20-39-3w5 7030' Ell sst	67.5	18	1	0	8.5	4	1
1155E	10-20-39-3w5 7048' Ell sst	65.5	14	2	0	10.5	7	1

Percentage composition of the samples as estimated from 200 point counts.

SAMPLE	REL % Q	REL % RF	REL % CHT
1001J	99.3	0.7	0
1002J	98.8	1.2	0
1003J	98.7	1.2	0.1
1004E	88.2	6.5	5.3
1005J	98.7	0.7	0.6
1007J	100	0	0
1009J	100	0	0
1010E	83	3.6	13.4
1012J	98.7	1.3	0
1015J	100	0	0
1016E	72.4	9	18.6
1017E	63.1	11.5	25.4
1018J	99.4	0	0.6
1019J	100	0	0
1020J	98.84	0.6	0.56
1021J	100	0	0
1023J	100	0	0
1024J	100	0	0
1025J	98.8	0	1.2
1026J	100	0	0
1027E	65.8	6.2	28
1028E	74.9	12.6	12.5
1029E	84.5	13.8	1.7
1030E	52.8	16.2	31
1030bE	83	10.3	6.7
1031D	95	4.4	0.6
1039E	92.2	7.8	0
1069J	100	0	0
1070J	97.1	2.9	0
1071J	97.1	2.9	0
1072J	100	0	0
1073J	100	0	0
1074J	98.6	1.4	0
1075J	99	0.92	0.08
1076J	99.2	0	0.8
1077J	100	0	0
1078J	100	0	0
1079J	100	0	0
1080J	100	0	0
1083E	82.5	9.62	7.88
1084E	52	6.08	41.92
1085E	46.8	11.6	41.6
1086J	100	0	0
1087E	49.4	0	50.6
1088E	30.8	8.3	60.9
1089E	37.8	9.8	52.4
1090RC	99.4	0.6	0
1091RC	98.7	1.3	0
1092RC	99.3	0.7	0
1093RC	98.8	0.6	0.6
1094RC	100	0	0
1095RC	98.1	1.3	0.6
1096RC	98.2	1.8	0
1097RC	97.8	2.2	0
1099RC	100	0	0
1100RC	98.2	1.2	0.6
1101RC	99.4	0	0.6
1102RC	99.2	0.8	0
1103RC	98.3	1.1	0.6
1105E	63.6	13	23.4
1106E	85	3.8	11.2
1107E	98	0	2
1108RC	99.4	0.6	0
1109RC	98.6	1.4	0
1110RC	98.9	1.1	0
1111RC	99.4	0.6	0
1112RC	99.5	0.6	0
1114RC	98.2	1.2	0.6
1115RC	98	2	0
1116RC	99.4	0.6	0
1117E	73.4	20.8	5.8
1118RC	98.6	1.4	0
1121RC	99.4	0.6	0
1122E	60	15	25
1154E	78	20.8	1.2
1155E	80.4	17.2	2.4

Relative percentage of framework grains.