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SEDIMENTOLOGY AND STRATIGRAPHY OF THE DISMAL LAKES GROUP,
PROTEROZOIC, NORTHWEST TERRITORIES

by

© Charles Kerans, B.Sc.

A thesis submitted to the Faculty of
Graduate Studies and Research in partial fulfilment
of the requirements for the degree of
Doctor of Philosophy

Department of Geology

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Ottawa, Ontario

August 27, 1982

The undersigned hereby recommend to the Faculty of
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ABSTRACT

The Dismal Lakes Group is a 2-km-thick middle Proterozoic sequence of predominantly shallow-marine strata composing the second major siliciclastic-to-carbonate depositional cycle within the Coppermine Homocline. These strata succeed the first-cycle Hornby Bay Group, and are overlain by 1.2 B. Y. Coppermine River Group basalts. Dismal Lakes strata record gradual stabilization of unevenly subsiding post-orogenic terrane of the Hornby Bay Group, thus establishing a stable north- and west-sloping shelf which rimmed the northwestern margin of the Canadian Shield.

Deltaic and marine sandstone and shale of the Fort Confidence Formation, and sabkha mudstones of the Dease Lake Formation, compose the fining-upward portion of the siliciclastic phase of the second cycle. Gradual eastward transgression and source-area stabilization led to development of a level depositional surface before onset of the carbonate phase. Kendall River Formation peritidal dolostones interfinger with Dease Lake sabkha deposits, recording continued gradual eastward transgression, and eventual development of an extensive carbonate platform.

Rapid transgression caused drowning of the Kendall River platform, and initiated deposition of deeper-water shales and dolostones of the basal Sulky Formation.

During this transgression, the platform differentiated into a slowly subsiding eastern area, the September Lake High, and the surrounding steadily subsiding platform. A major stromatolite buildup formed on the September Lake High during Sulky member B deposition, as did a distinctive buildup-crest shoal facies (member C).

Greenhorn Lakes Formation sedimentation was dominated by transgressive-regressive dolostone-biostrome sequences on the platform; coeval peritidal dolostones on the September Lake High indicate that this feature remained a depositional high through Greenhorn Lakes deposition. The axis of subsidence shifted westward, and the paleoslope changed from west to southwest during deposition of the Greenhorn Lakes Formation. Uplift in the area around the Muskox Intrusion exposed the platform to subaerial erosion near the end of Greenhorn Lakes deposition, reflecting an early pulse of Muskox Intrusion emplacement. Transgression and renewed peritidal dolostone deposition continued until interrupted by magmatism of the 1.2 B. Y. Mackenzie Igneous Event, comprising extrusion of conformably overlying Coppermine River Group basalts, and intrusion of Mackenzie diabase dykes and Muskox Intrusion.

Post-Dismal Lakes Group faulting led to detachment of Kendall River, Sulky and Greenhorn Lakes Formations from underlying units along a décollement zone within the Dease Lake Formation, followed by high-angle faulting which

involved dip-slip and possibly some reverse offset
with up to 1 km of displacement.

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- E. Photomicrograph of silicified compacted ooids, Kendall River Formation.
- F. Photomicrograph of dolomite rhombs replacive after chert.

INTRODUCTION

The Dismal Lakes Group is a well-exposed middle Proterozoic sequence of predominantly shallow-water marine siliciclastic and carbonate strata. The Group, a component of the Coppermine Homocline, crops out almost continuously in a 300-km belt from Dease Arm to Coronation Gulf (Figures 1 and 2). All outcrops are within the Arctic Circle and, except for the southernmost exposures near Great Bear Lake, are in tundra of the barrenlands.

The primary objective of this study has been to conduct a basin analysis in detail sufficient to allow synthesis of the tectono-depositional history of the Dismal Lakes Group. Starting with the stratigraphic framework established by Baragar and Donaldson (1973), this study was carried out in conjunction with the study of G. M. Ross (Carleton University) of the underlying Hornby Bay and basal Dismal Lakes Group siliciclastics. The purpose of these complementary studies was to produce a detailed sedimentologic-stratigraphic framework for the lower half of the Coppermine Homocline. During this basin analysis, mapping of the Dismal Lakes Group in the Bebensee Lake and Takatcho River map areas was completed at 1:100,000 scale (Ross and Kerans, in prep.). Revisions of the Dismal Lakes and Coppermine River 1:250,000 sheets, and of the existing stratigraphic

Figure 1. Regional geology of the western Canadian Shield. Major Middle Proterozoic basins (Athabasca, Thelon, and Elu Basins and Coppermine Homocline) are stippled. (from Ross, in press). GBL- Great Bear Lake, GSL- Great Slave Lake, CG- Coronation Gulf, AA- Athapuscow Aulacogen, KB- Kilohigok Basin, BF- Bathurst Fault, MF- MacDonald Fault.

Figure 2. Geologic setting (A) and geographic features (B) of the Coppermine Homocline region. '11-DL - Unit 11 through Dease Lake Formation, KR-GH - Kendall River through Greenhorn Lakes Formation. Modified from Kerans et al. (1981).

A.

B.

framework, were minor (see below).

The objectives of this study were achieved largely through detailed measurement of sections, paleocurrent analysis, and analyses of polished slabs and thin sections. In addition to sedimentologic and tectonic histories of the Dismal Lakes Group, detailed study of the diagenetic history of the Group was possible because of excellent preservation of primary structures and textures in the carbonate strata.

PREVIOUS WORK

The Dismal Lakes area was visited by several geologists in the early 1900's (Sandberg, in Douglas, 1913; O'Neill, 1924), but the primary interest of these geologists was native copper in the Coppermine River Group. The first map of the Dismal Lakes Group, as now defined, was produced by Baragar and Donaldson (1973) as part of a helicopter mapping project that covered the Dismal Lakes and Coppermine River 1:250,000 map sheets. Helicopter coverage was supplemented by foot-traversing in critical areas (Donaldson, 1969, 1971; Baragar and Donaldson, 1970) to produce what has proved to be a remarkably accurate map. In earlier maps, the Dismal Lakes Group (Figure 3) was considered to be part of the Hornby Bay Group, which was divided into the lower Hornby Bay Group sandstone (present Hornby Bay Group) and the upper Hornby Bay Group (present Dismal Lakes Group) by

FRASER ET AL. (1960)	BARAGAR AND DONALDSON (1973)	KERANS ET AL. (1981)	THIS STUDY
HORNBY BAY GROUP (DOLOSTONE)	DISMAL LAKES GROUP	16	GREENHORN RIVER FORMATION
	15	15	SULKY FORMATION
	14	14	KENDALL RIVER FORMATION
	13	13	DEASE LAKE FORMATION
	12	12	FORT CONFIDENCE FORMATION
	11	11	11
HORNBY BAY GROUP (SANDSTONE)	HORNBY BAY GROUP	10	10 ²
	9	9	9
	8	8	8

Figure 3. History of stratigraphic nomenclature of Dismal Lakes and Hornby Bay groups.

Fraser et al. (1960, 1970) and by Cook and Aitken (1971). On the basis of map-scale structural relationships, Baragar and Donaldson (1973) suggested that an unconformity existed between their map Units 11 and 12, and that the dolostone-dominated upper Hornby Bay Group of previous authors be designated the Dismal Lakes Group, comprising their map units 12-16 (Figure 3). Detailed mapping in the Kirwin Lake area by several mining companies (results summarized in Yeß, 1979) and later by Ross (1980) led to recognition of the exposed unconformity, which suggested that the unconformity be placed below Unit 11 (Figure 3).

Most recently, Kerans et al. (1981) published a summary of the tectono-depositional history of the Hornby Bay and Dismal Lakes Groups. In addition to the redefinition of the Hornby Bay-Dismal Lakes contact, the recognition of a disconformity of regional significance within Unit 16 (Greenhorn Lakes Formation of this study; see Figures 3 and 4) enabled refinement of regional correlation between the Coppermine Homocline and Bathurst Inlet areas (Kerans et al., 1981).

Several changes to the stratigraphic framework presented in Kerans et al. (1981) are proposed here, including the informal introduction of formational names for Units 12-16, redefinition of member C of Unit 14 (Kendall River Formation), and redefinition of the Unit 15-16 contact (Sulky Formation-Greenhorn Lakes Formation

contact; Figure 3).

Previous work of a more detailed nature has also been carried out on the distinctive stromatolites of the Sulky Formation (Donaldson and Jones, 1971; Donaldson and Taylor, 1972; Donaldson, 1973, 1976; Kerans and Donaldson, 1979), and the well-preserved microfossil assemblages (Donaldson and Delaney, 1975; Horodyski and Donaldson, 1980; Horodyski et al., 1980).

REGIONAL GEOLOGIC SETTING

The Dismal Lakes Group is a component of the undeformed middle to upper Proterozoic Coppermine Homocline, in the northwestern corner of the Canadian Shield (Figure 1). The sedimentary and volcanic strata of the Homocline were deposited over the 1.93-1.85 B.Y. Wopmay Orogen (Van Schmus and Bowring, 1980; Hoffman, 1981) following metamorphism, uplift and erosion of the Orogen to depths of several kilometres (St Onge et al., 1982).

The Coppermine Homocline dips gently to the west and northwest, exposing a sequence of shallow-marine and terrestrial sedimentary and volcanic rocks which includes, from base to top: the Hornby Bay Group (2 km), the Dismal Lakes Group (2 km maximum), the Coppermine River Group (4 km) and the Rae Group (2 km). Dismal Lakes Group strata conformably overlie the Hornby Bay Group west of Teshierpi Fault (Figure 2). Adjacent to the Teshierpi

Fault, basal Dismal Lakes Group arenites (Unit 11) rest unconformably on fault slices which expose all levels of Hornby Bay strata as well as granitoids and metavolcanics of Wopmay Orogen. East of Teshierpi Fault, Unit 11 sandstones were deposited unconformably on Unit 8 of the Hornby Bay Group, and on Wopmay Orogen rocks (Ross, 1980; Kerans et al., 1981). The Coppermine River Group, which conformably overlies upper Dismal Lakes dolostones, and Mackenzie diabase dykes, which cut all Dismal Lakes strata, yield Rb/Sr isochrons of approximately 1.2 B.Y. (Baragar, 1972; Patchett et al., 1978). This 1.2-B.Y. age provides an upper limit on the age of the Dismal Lakes Group. The conformable nature of the Hornby Bay-Dismal Lakes Group succession west of Teshierpi Fault suggests that this 3-4 km thick sequence lies closer to the 1.2 B.Y. upper age limit than to the lower age limit of 1.85 B.Y. provided by zircon geochronology of Wopmay Orogen (Van Schmus and Bowring, 1980). The tentative chronostratigraphic chart presented in Figure 4 illustrates suspected time relationships within the Hornby Bay and Dismal Lakes groups. This proposed correlation suggests a significant time gap between late-stage activity of Wopmay Orogen and initiation of sedimentation in the Coppermine Homocline.

The Hornby Bay, Dismal Lakes and Coppermine River groups are cut by NE- and NW-trending faults (Figure 2). These two trends dominate the structural grain of the

Homocline, and of the underlying Wopmay Orogen rocks as well. These faults are inferred to have originated as conjugate transcurrent faults during the terminal collision phase of Wopmay Orogen (Hoffman, 1980a; Hoffman and St Onge, 1981). Several faults of this set underwent syndepositional dip-slip movement during Hornby Bay Group deposition (Ross, 1980). Syndepositional fault activity had significantly decreased before Dismal Lakes Group deposition, but the Teshierpi Fault (Figure 2) and subsidiary faults in the Kirwin Lake area were reactivated as dip-slip faults during deposition of basal (Unit 11) Dismal Lakes sandstones (Kerans et al., 1981).

North-trending structures include the Herb Dixon and southern Canoe Lake Fault, as well as the Mackenzie diabase dykes and the Muskox Intrusion. All north-trending structures except the Leith Line (see below) post-date Dismal Lakes deposition, and may be related to development of an E-W tensional stress regime associated with the Mackenzie igneous event. The Leith Line is a major north-trending lineament manifested by a gravity high traceable from Leith Peninsula, along the eastern shoreline of Great Bear Lake, into the Lac LeRoux area (Figure 2). Facies changes within the Hornby Bay and Dismal Lakes Groups demonstrate that the Leith Line was an active syndepositional hinge zone, with subsidence and accumulation rates being greater west of this hinge (Kerans et al., 1981).

The East River Fault, which displays at least 1 km of north-side-down dip-slip displacement, is the only major east-trending structure in the area (Figure 2). This fault post-dates Dismal Lakes Group deposition, and locally displays moderate to intense fault-proximal folding of both Hornby Bay and Dismal Lakes Group strata. For reasons discussed in Chapter 7, this fault is most likely a reverse fault.

DEPOSITIONAL HISTORY OF THE DISMAL LAKES GROUP

Dismal Lakes Group strata comprise the second siliciclastic-to-carbonate depositional cycle within the Coppermine Homocline. To set the stage for discussion of individual formations, a synopsis of the preceding Hornby Bay Group siliciclastic-to-carbonate cycle is presented, as adopted from the synthesis of Ross (in Kerans et al., 1981), followed by a brief summary of Dismal Lakes Group deposition. This synopsis will serve to contrast the two cycles, and more importantly, will provide a lead-in to the detailed analysis of Dismal Lakes strata. This introduction is necessary because the first unit described in detail in this study, the Fort Confidence Formation, occurs half way through the siliciclastic phase of the second cycle. Units 10 and 11, the basal units of the siliciclastic phase of the second cycle (Figure 4), have been studied by Ross (1980; in Kerans et al., 1981).

Figure 4. Tentative chronostratigraphic chart for the Dismal Lakes and Hornby Bay groups. 8-11 are unit numbers of Baragar and Donaldson (1973: see Figure 3). FC- Fort Confidence Formation, DL- Dease Lake Formation, KR- Kendall River Formation, S- Sulky Formation, GH- Greenhorn Lakes Formation, a-e are members of the Greenhorn Lakes Formation. Cycles 1 and 2 are siliciclastic-to-carbonate cycles referred to in text.

The Hornby Bay Group records initial deposition of the Coppermine Homocline succession. In contrast to Dismal Lakes strata, Hornby Bay deposition took place in a tectonically active environment. Early stages of sedimentation took place in isolated continental basins, where alluvial and eolian processes dominated (lower Unit 8). Syndepositional faulting and crustal doming were controlling factors during this initial phase, causing facies changes to be rapid and localized. The transition to a more stable, homogeneously subsiding basement is recorded by broad braidplain and shallow marine siliciclastic deposits (middle and upper Unit 8). High-energy shallow-marine dolostones of Unit 9 mark the period of maximum stability and maximum transgression during Hornby Bay Group deposition. In comparison to platform carbonates of the Dismal Lakes Group, however, greater differential subsidence on this southwest-dipping platform produced rapid lateral facies changes, and more abundant higher-energy subtidal, ooid bar belts. Siliciclastic input from the craton was greatly reduced during this carbonate phase, indicating that the source area had been leveled before initiation of the second siliciclastic-to-carbonate depositional cycle.

Renewed uplift in the Teshierpi Mountain area signaled the beginning of the second clastic-carbonate cycle which caused deltaic mudstones and arenites of Hornby Bay Group Unit 10 to prograde over the platform

dolostones of Unit 9 (Figure 5A). Regional southwest tilting and dip-slip movement along the Teshierpi Fault produced a coarsening-up transition into fluvial conglomerates and arenites of lower Unit 11 (Figure 5B). As tectonism waned, some fluvial arenites were reworked by marine processes to produce the mixed fluvial and marine record of upper Unit 11, which is transitional into finer-grained facies of the Fort Confidence Formation (Figure 5C; Ross, in Kerans et al., 1981).

The Fort Confidence and Dease Lake Formations comprise the upward-fining upper half of the siliciclastic portion of this second siliciclastic-to-carbonate depositional cycle which was initiated by Unit 10 deposition (Figure 4). Syndepositional fault activity was less pronounced in the second depositional cycle, having essentially terminated before Fort Confidence deposition. The deltaic-marginal marine deposits of this formation contain significantly more shale than conformably underlying Unit 11 (Figure 6), and record decreased source-area relief and gradual eastward transgression. Dease Lake Formation mixed terrestrial-marine mudstones and dolostones (Figure 6) are transitional into the overlying carbonate-dominated portion of the cycle, recording continued gradual transgression and near-complete reduction in source-area relief.

Shaly-to-clean dolostones of the Kendall River

Figure 5. Schematic representation of tectonic events and depositional environments during upper Unit 9 (9C) of the Hornby Bay Group through the Fort Confidence Formation of the Dismal Lakes Group. View is from East River looking northeast along the Teshierpi Fault (modified from Kerans et al., 1981).

- A. Normal fault movement along the Teshierpi Fault exposed Unit 9 dolostone to subaerial silicification with erosion and transportation to the southwest.
- B. Regional uplift caused erosion of Units 9, 8 and basement. Drainage was in part controlled by oblique dip-slip movement towards the southwest along the Teshierpi Fault.
- C. The culmination of uplift resulted in further erosion of Hornby Bay Group and basement rocks and deposition on a north-northwest facing fluvial (Unit 11) - deltaic (Fort Confidence Formation) complex.

DISMAL LAKES STRATIGRAPHIC COLUMN

Figure 6. Generalized stratigraphic column for the Dismal Lakes Group. FC- Fort Confidence Formation, DL- Dease Lake Formation, KR- Kendall River Formation, S- Sulky Formation, GH- Greenhorn Lakes Formation. SP- supratidal, I- intertidal, SB- subtidal.

Formation, the lowermost deposits of the carbonate portion of the cycle, were deposited during continued gradual eastward transgression of the low-relief surface which existed at the end of siliciclastic sedimentation. Resultant facies differentiation is slight. Maximum transgression and differentiation of the carbonate platform into eastern buildup and western subtidal shelf divisions occurred during deposition of massive stromatolitic dolostones of the Sulky Formation (Figure 6). This differentiation of the platform was maintained through several transgressive-regressive pulses during Greenhorn Lakes Formation sedimentation, which left a record of cyclic dolostone-shale sequences west of Dismal Lakes, whereas peritidal dolostones were deposited continuously in the east.

Carbonate sedimentation was interrupted locally by doming and exposure of the platform just before the end of Greenhorn Lakes deposition. The ensuing transgression and deposition of uppermost Greenhorn Lakes dolostones preceded extrusion of the Coppermine River Group basalts, which terminated Dismal Lakes deposition.

CHAPTER 1. FORT CONFIDENCE FORMATION

Sandstones and "mixed facies" (lenticular-bedded sandstone and green-black flaser-bedded shale) of the Fort Confidence Formation overlie the shale-poor arenites of Unit 11 (terminology of Baragar and Donaldson, 1973) gradationally. Mixed facies make up roughly half of the formation, and crossbedded quartz arenites compose the remainder. The name Fort Confidence Formation is informally proposed for these sandstones and shales, which outcrop at the site of the abandoned Fort on the northeast shore of Dease Arm. The best exposures of the unit are along the north shore of Dismal Lakes, where small stream cuts create exposures of both resistant sandstones and recessively weathering shales. The Fort Confidence Formation interfingers with the underlying Unit 11 arenites, and the contact is arbitrarily defined as the base of the lowermost metre-thick shale in the Fort Confidence section. Due to the characteristic recessive weathering of the mixed facies, this contact can be picked accurately only in drill core.

The upper contact of the Fort Confidence Formation with the overlying Dease Lake Formation is remarkably abrupt, both in the area north of Dismal Lakes and in the western outcrop belt around Dease and Bebensee Lakes. North of Dismal Lakes, mixed shale-sandstone facies passes upward into anhydrite-pseudomorph-bearing red

dolostones of the Dease Lake Formation. In the Dease and Bebenssee Lake areas, the mixed facies passes upward into tan stromatolitic dolostones within less than one metre.

No complete sections of the Fort Confidence Formation are exposed. Typical exposed sequences, such as those on the north-central shore of Dismal Lakes, consist of discontinuous strike-parallel ridges of quartz arenite separated by valleys of recessive-weathering mixed facies. This recessive-weathering character of the mixed facies everywhere casts doubt on the position of the lower contact, making thickness estimates of the formation difficult. A minimum thickness estimate for the formation in the Dismal Lakes area is 200 m.

LITHOFACIES DESCRIPTIONS

Sporadic exposure and absence of distinctive marker beds prevent subdivision of the Fort Confidence Formation into formal or even informal members. Lithofacies subdivisions are used here to break the formation into distinctive lithologic units without implying specific stratigraphic position. The Fort Confidence Formation can be divided into three principal lithofacies based on sedimentary structures and sandstone-shale ratios. Lithofacies A and B are sand-dominated and are intercalated with mixed sandstone-shale of lithofacies C.

Lithofacies A:

Lithofacies A is best exposed near the top of the Fort Confidence Formation in a creek which enters the northwest corner of Dismal Lakes (section 2, Figure 7). Lithofacies A sandstones gradationally overlie lithofacies C, producing an overall coarsening-upward sequence.

The lithofacies consists of 60-100 cm bedded, fine- to medium-grained well-sorted quartz arenite, with interbeds of 20-60 cm bedded mixed sand-argillite. The argillite in lithofacies A is typically green. The sandstone beds are commonly amalgamated in the upper part of the lithofacies, forming packages up to 6 m thick.

Low-angle tabular planar (Sp of Miall, 1978) and wedge-planar cross-stratification are characteristic. Horizontal lamination (Sh of Miall, 1978) and low-angle cross lamination also are common (section 2, Figure 7). Tops of sandstone beds typically display small-scale wave and current ripple marks.

Argillaceous interbeds are sand-streaked (cf. de Raaf et al., 1977) and in places display lenticular bedding. Injection-modified desiccation cracks are ubiquitous. Some show clear surface-desiccation origin; others may be a form of syneresis crack (cf. Plummer and Gostin, 1981). Some mud-chip intraclasts are incorporated into overlying sandstone beds.

No systematic paleocurrent analysis was carried out on lithofacies A. Qualitative observations suggest

Figure 7. Typical sections of Fort Confidence Formation showing lithofacies C-A (section 2) and B-C (section 4) transitions. Characteristic sedimentary structures and paleocurrent patterns are labelled along right sides of columns. Sedimentary structure key as in Figure 6. Sections located on Figure 8.

diffuse to opposed paleocurrent patterns.

Lithofacies B:

Lithofacies B is composed of moderately- to well-sorted quartz arenite with subordinate granulestone. Lithofacies B sandstone-dominated sequences, in contrast to those of lithofacies A, contain only rare shale interbeds. These consist of 2-4 cm green argillite beds and intraclasts at the base of some crossbed sets. Basal contacts of lithofacies B on lithofacies C are erosional, commonly displaying a basal intraclast conglomerate.

These sandstone bodies typically fine upward, showing a vertical succession of Se or Sh to Sp to St (Miall, 1978), with a corresponding bed thickness decrease upward from 100 to 10 cm (section 4, Figure 7; Plate 1A). Parting lineation is common in the Sh facies, indicating upper flow regime currents (Harms et al., 1975). Smaller scale crossbedded caps (St) locally display herringbone cross stratification, mud drapes and interference ripples.

Paleocurrents from lower Sp and St beds display northeast- to northwest-directed unimodal transport (Figure 9). The upper small-scale St sets show polymodal or diffuse paleocurrent readings, commonly with current reversal patterns.

Uppermost St beds pass gradationally upward into mixed facies sediments characteristic of lithofacies C.

Rhythmic lithofacies B to C repetition has been noted in most sections (Figures 7 and 8). This repetition defines an overall scour-based fining upward sequence from the base of lithofacies B to the top of overlying lithofacies C, which in turn is truncated by the succeeding B unit.

Lithofacies C:

Lithofacies C (mixed sandstone-shale) is the most distinctive lithology in the Fort Confidence Formation. The lithofacies consists of thinly intercalated fine- to very fine-grained quartz arenite and green-black shale. Lenticular and subordinate flaser bedding dominate (Plate 1C and D). Symmetric, asymmetric and interference ripples, injection-modified mud cracks (Plate 1D), cut and fill structure, rills and small channels are common sedimentary structures. Convex crosslaminated, flat bottomed sandstone lenses with multiple reactivation surfaces are common (Plate 1C). One of these sand "bars" is composed of ooid grainstone. Rare 10-15 cm dololomite units can be correlated locally and used as markers where core control exists.

Paleocurrent patterns in this lithofacies show a minor northward mode but are generally diffuse (Figures 7 and 9), and less commonly bipolar.

Lithofacies C forms the upper portion of B-C fining-upwards sequences and the basal portion of the

coarsening-upward C-A succession (Figure 7).

INTERNAL STRATIGRAPHY OF FORT CONFIDENCE FORMATION

Although the lower contact of the Fort Confidence Formation is poorly defined, the contact with the overlying Dease Lake Formation is sharp and well exposed in several areas. As such, the upper contact has been used as a datum for correlation of sections. Relative to this datum, lateral discontinuity can be demonstrated over distances of several hundred to 1000 m for sandstone bodies of lithofacies A and B (Figure 8). Amalgamated sandstones of lithofacies A pinch out eastward from section 2 to section 3 (Figure 8), a distance of 4 km. Lithofacies A beds could not be traced laterally to their pinch out, but they appear to display laterally an interfingering relationship with lithofacies C, consistent with their gradational lower contacts. Numerous lithofacies B sandstone intervals along the north-central shore of Dismal Lakes also pinch out laterally over 700-1000 m. These sandstone lenses are scour-based and fine upwards. This vertical sequence, coupled with the characteristic lateral discontinuity, suggests a channel form rather than a bar form. Northeast of Dease Lake, the scoop-shaped base and flat top of a small-scale channel can be observed in a single outcrop (Plate 1B).

Figure 8. Measured sections of the Fort Confidence Formation which show the laterally restricted nature of sandstone bodies (lithofacies A and B are not differentiated). Recessive units are lithofacies C shales.

Figure 9. Paleocurrent rose diagrams for Fort Confidence Formation. Bracketed roses are taken from single measured sections and are in proper stratigraphic position. Number of readings per station given within each rose, except for total rose.

REGIONAL PALEOCURRENT PATTERNS

As previously mentioned, lithofacies A sandstones and lithofacies C mixed facies display diffuse paleocurrent patterns, occasionally being poly- or bimodal. Such paleocurrent patterns do not offer much insight into regional paleogeography.

Sequence B channel sandstones show consistent unimodal paleocurrents (Figure 9). The overall northeast-northwest paleoflow shows a change from the southwesterly-directed flow of lower fluvial Unit 11 (Kerans et al., 1981), but is consistent with that of the overlying Dismal Lakes Group formations.

Two outcrop areas north and east of Dease Lake show southeastward paleoflow (Figure 9), in opposition to the dominant regional trend. Such opposed paleoflow may represent local variation in paleoslope or in the currents responsible for dispersal of the sandstones.

INTERPRETATION

The interpretation of an offshore versus a coastal depositional environment for the Fort Confidence Formation is constrained by the interpreted depositional environments of the overlying and underlying formations. Lower Unit 11 is a fluvial deposit, but the upper portion is a mixture of fluvial and shallow high-energy marine facies. Unit 11 and the Fort Confidence Formation intimately interfinger, and must represent deposits of adjacent environments (cf. Walther, 1893). The basal

Dease Lake Formation from Dismal Lakes eastward is an evaporite-bearing mudstone of probable coastal sabkha origin, whereas in the Dease and Bebensee Lake areas, subtidal to intertidal dolostones rest directly on lithofacies C of the Fort Confidence Formation. The stratigraphic position of the Fort Confidence Formation, situated between two terrestrial to shallow-marine units, implies a relatively shallow-water setting for this unit as well.

Lithofacies C, which forms the matrix for the two sand-rich lithofacies, resembles both intertidal mud flat deposits (Reineck and Singh, 1975) and subtidal marine shelf clastics (cf. de Raaf et al., 1977) in terms of its predominant lenticular bedding. A subtidal origin can be ruled out, however, due to the presence of exposure features such as mud cracks, channels, some late-stage run-off features (rill marks) and abundant intraclasts. In addition, the ripple structures of the M1 and M2 facies of de Raaf et al. (1977) were predominantly wave-generated, whereas, based on characteristics presented by de Raaf et al., most of the ripple cross-stratification in lithofacies C was formed by traction currents consistent with tidal processes acting upon a periodically exposed flat. The convex-laminated, flat-based sand bars within lithofacies C are similar to wash-over fans described by Schwartz (1975) and Hayes (1967). The ooidic composition of one such bar indicates

an allochthonous onshore-directed storm origin for at least one of these deposits. Finally, the close lateral and vertical juxtaposition of lithofacies C and intertidal dolostone of the Dease Lake Formation supports the tidal flat hypothesis.

The coarsening-upward sandstone sequence of lithofacies A gradationally overlies and interfingers with tidal flat deposits of lithofacies C, and therefore must be of relatively shallow water origin. Tabular-planar crossbed sets may be analogous to shallow subtidal marine sand bars described by Meckel (1975) from the Holocene Colorado Delta. Alternatively, they may be analogous to ridge and runnel sets (cf. Hunter et al., 1979). The multitude of small-scale cross-stratification which modifies upper portions of the larger bedsets are consistent with such an origin.

Wedge-shaped sets of cross strata may be beach shoreface deposits, and low-angle to horizontally bedded sandstones are similar to deposits produced by spit accretion (cf. Hobday and Tankard, 1978).

Green argillaceous mixed facies intercalated with thick sandstone units in lithofacies A are inferred to represent low-energy deposits formed in troughs behind or between current barriers produced by the migrating sand bars (cf. laminated muds between marine sand bars on the shoreface of Colorado River Delta: Meckel, 1975).

Lithofacies B sandstone bodies are interpreted to be channel-fill sequences based on lenticular form, erosional base, fining-up and thinning-up trends, and unidirectional paleoflow. The presence of one outcrop-scale lithofacies B lens which displays channel form demonstrates the channel morphology of at least one of these sandstone bodies. The basal tabular-planar bed sets probably were formed by migration of sandwaves in a unidirectional current regime. Upper flow regime conditions forming parting lineation represent probable sheet flood phases of deposition. Upper portions of the channel-fill sequences are more thinly bedded and commonly display current reversals and mud drapes between bed sets. This unidirectional-to-bidirectional fining-upward channel-fill sequence closely resembles deltaic distributary channel-fills such as those described by Moore (1979) from the Cambrian of Australia and those of the Late Quaternary Niger Delta (Oomkens, 1974; Allen, 1965). An alternative origin for these channel-fill sequences is by migration of tidal creek point bars in an ebb-dominant regime (cf. Barwis, 1978). Tidal creek point bar deposition and fluvial distributary channel deposition represent two end-member processes. Lithofacies B channel-fill sequences probably represent deposits formed by a combination of these processes, as are those of the Niger Delta.

In summary, the Fort Confidence Formation is a

mosaic of tidal flats, marine shoreline sands and channel-fill sands (distributary channels or tidal creeks). This environmental mosaic overlies and interfingers with fluvial-marine sands of Unit 11, and is superseded by predominantly marine Dease Lake Formation mudstones and dolostones.

A delta emptying into a shallow epicontinental sea (similar to the setting described by Moore, 1979) is considered most analogous. Sediment transported into a shallow receiving basin would inhibit the development of a significant delta-front facies, explaining the absence of a large-scale coarsening-upward succession in the Fort Confidence Formation. Features and structures from lithofacies C mud flats and upper portions of lithofacies B channel-fill sands clearly indicate tidal reworking, whereas lithofacies A sands appear to be largely wave-generated sequences.

Regardless of the ebb-tidal or fluvial origin of the northeast- to northwest-directed paleocurrent trends (Figure 9), the north- to northwest-inclined paleoslope maintained during deposition of the succeeding Dismal Lakes Group units appears to have prevailed during Fort Confidence deposition.

PLATE 1. FORT CONFIDENCE FORMATION

A. Medium-scale tabular-planar crossbed in basal 2 m of a lithofacies B sandstone body south of Bebensee Lake. Note low angle of cross-lamination and poorly exposed Sh bed which underlies the tabular-planar set. Notebook is 17 cm long.

B. Concave-up base of lithofacies B quartz arenite channel-fill (dashed) cut into lithofacies C shale. Pinchout of sandstone near left hand side of photograph marks channel margin. Steeply inclined prominent fractures are joints. Person is 1.5 m tall.

C. Typical outcrop of lithofacies C mixed sandstone-shale. Some sandstone beds are relatively thick (4-6 cm) and continuous. Thinner beds are lenticular and flaser bedded. Thickest sandstone bed (near base of photograph) displays several reactivation surfaces (arrows). Notebook near centre of exposure is 17 cm long.

D. Thinly interbedded sandstone and shale of lithofacies C. Shale layers have compacted more readily, causing sandstone-filled mudcracks to deform into vermiform structures. Marking pen is 14 cm long.

PLATE 1

A

B

C

D

CHAPTER 2. DEASE LAKE FORMATION

Red to pink mudstones and tan to grey dolostones which conformably overlie the Fort Confidence Formation are informally named the Dease Lake Formation, which is equivalent to Unit 13 of Baragar and Donaldson (1973). The unit is best exposed along the north shore of a northeast-elongated lake, informally named Dease Lake by the Gulf Minerals camp of John Thompson (1974-1979). The Dease Lake Formation is abruptly conformable over the Fort Confidence Formation, and is in gradational contact through up to 30 metres with the overlying Kendall River Formation. West of Leith Line (see Figure 2) the Formation is divided into three distinct members which are, from base to top: a tan quartzose stromatolitic dolostone (member A), red mudstone and tan dolostone (member B), and, pink mudstone and grey dolostone (member C). In the area around Dease Lake, these three members total 150 m in thickness. East of Leith Line the Dease Lake Formation thins to 100 m and consists of a partial section of members B and C. These strata grade upward into the dolostone-dominated Kendall River Formation (Figure 10).

West of central Dease Lake, a 60-600 cm tectonic breccia separates member B from member C. This breccia thickens rapidly westward at the expense of member C strata, and west of the southwest end of Dease Lake,

Figure 10. Generalized cross-section of the Dease Lake Formation showing the distribution of the three members and extent of the member C deformed zone. Member A is restricted mainly to the area of increased subsidence west of the Leith Line. Member C thickens near section 10, as the southeasterly source area is approached. Section locations are shown in Figure 11.

member C is entirely deformed (Figure 10; see Chapter 7).

MEMBER A

Member A is the only marginal-marine carbonate deposit of the Dismal Lakes Group which is dominated by physical sedimentary structures rather than by trapping and binding organic mats. The thickness of the member in the Dease Lake area is about 75 m. As previously mentioned, the member pinches out east of Leith Line and occurs only in small fault slices west and north of Dease Lake. Where observed, contact between the Fort Confidence Formation and member B of the Dease Lake Formation is conformable, and is marked by a variable biostromal unit. The upper contact of member A with member B is transitional over approximately 5 metres, with red mudstones gradually becoming subequal to tan dolostone. Six detailed sections of this member are illustrated in Figure 11.

Member A is divided into three major lithofacies: a thin-bedded shaly dolostone; crossbedded dolarenite, and; stromatolitic biostromes.

Thin-Bedded Shaly Dolostone Lithofacies:

The thin-bedded shaly dolostone lithofacies consists of 2-5 cm beds of dolosiltite and sandy dololutite with rare 1-2 cm green shale partings. The thin-bedded lithofacies typically forms 40-80 cm thick units which overlie either biostromal or crossbedded dolarenite

facies and which almost everywhere are overlain by the crossbedded dolarenite lithofacies. Current-generated sedimentary structures include: parallel to wavy current lamination, ripple cross-stratification, oscillation and interference ripples (Plate 2A and 2B), thin (0.5 cm) stringers of medium-grained quartz sand and thin (3-7 cm) intraclast conglomerate beds. Crenulate organic mats, desiccation cracks (Plate 2B) and convolute lamination (Plate 2C) are common to this lithofacies. Ripple cross-stratified layers commonly show current reversal (Plate 2D). In-drift and in-phase climbing ripple lamination occurs repeatedly in 10-20 cm sets in sections 2 and 3 (Figure 11). At section 3, climbing ripples occur in 30-cm rhythmically-repeated sequences which consist of low-angle planar-to-horizontal cross-lamination, overlain by in-drift climbing ripples (type two of Jopling and Walker, 1968), overlain by in-phase (type one) climbing ripples (Plate 2E). Local 20-40 cm intervals of lenticular or linsen bedding occurs in association with shale interbeds near the base of the section. Shale interbeds are more common near the base of the section.

Throughout member A, rare to common 5-10 cm scour-based intraclast conglomerate beds with up to 40 percent mixed quartz sand and ooids are intercalated with the current and organic-laminated sediments.

Crossbedded Dolarenite Lithofacies:

Medium-bedded tabular-planar sets of cross-stratified quartzose-oolid dolarenite form units 70-140 cm thick. Upper and lower contacts of this lithofacies with either the thin-bedded or biostromal lithofacies are sharp and define tabular units. The crossbedded lithofacies has not been observed as a channel fill. Current reversal is common in the crossbedded lithofacies, and one quartzose unit shows a polymodal pattern, in agreement with the member as a whole (Figure 11). Reactivation surfaces (Plate 2F) are also common, as is herringbone cross-stratification. Clay drapes and desiccation features are absent.

Dolarenites are well sorted regardless of oolitic or quartzose composition. This lithofacies is generally oolitic, but quartz content increases eastward, approaching a level of abundance subequal to that of the ooids.

Biostromal Lithofacies:

Stromatolite biostromes form an important component of member A. They typically are 1 m thick, range up to 10 m, and are composed of both LLH-C domal types with diameters of 60-100 cm, and distinctive SH-V types with passive to radial branching (Plate 3A). Where exposed, the Fort Confidence-Dease Lake member A contact is marked by a biostrome which rests on lithofacies C shale of Fort Confidence Formation. The biostromes generally cap the

crossbedded lithofacies, but in places they overlie thin-bedded shaly dololutite.

Interspaces between columns are filled with either quartzose-oidic dolarenite or, less commonly, dololutite. Lamination profiles indicate synoptic relief of generally less than 15 cm for individual columns, and 20-40 cm for small bioherms. Some laminae within columns contain trapped quartz grains and/or ooids. Lamination microstructure is predominantly smooth, but crenulate textures also occur.

A unique biostrome consisting of continuously linked large-scale tufted organic mats occurs near the top of sections 4 and 5 (Figure 11; Plate 3B and 3C). This stromatolite morphology has not been reported elsewhere from the Proterozoic of the Bear Province. Laminations are smooth and continuous, but dololutite is the only intercalated sediment. This biostrome rests on ooid dolarenite. Overlying lithologies are not exposed at either locality.

Stromatolites within member A biostromes are symmetrical in plan view except for one example in section 8 (Figure 11) which displays marked south-southeast elongation. Ridges of the tufted mat biostromes typically form a polygonal pattern, giving an overall random orientation.

Interpretation: Member A

Thin-bedded shaly dolostones are ~~interpreted~~ as tidal flat deposits (probably both lower and middle tidal flat). Mud cracks and abundant intraclastic conglomerates are indicative of periodic exposure. Lenticular bedding and clay drapes on current ripples probably were formed by alternation of traction bedload deposition and suspension deposition during slack-water phases between ebb and flood tide (cf. Reineck and Singh, 1975). Flat organic-laminated sediments probably represent upper intertidal deposits characterized by lower sedimentation rates and increased substrate stability.

Convolute bedding is formed by sediment liquefaction in response to sediment instability as a result of rapid loading of water-charged sediment. Convolute bedding is relatively rare in intertidal deposits, but has been reported by Wunderlich (1970), Button and Vos (1977) and Tankard and Hobday (1977). The penecontemporaneous origin of convolute bedding in the member A example is demonstrated by erosional truncation of convolute laminae by the succeeding crossbed set (Plate 2C).

Climbing ripples can be formed by either wave or current ripple migration (McKee, 1965; Jopling and Walker, 1968). Although differentiation between these two processes is difficult, the basic requirement, deposition from sediment-charged water, holds true in both cases. Climbing ripple structure is common in fluvial overbank deposits and turbidites, but is not

particularly characteristic of tidal flat deposits (Reineck and Singh, 1975). A possible depositional mechanism for the repetitive horizontal-to-type 2-to-type 1 climbing ripple sequences is by rapid sedimentation out of sediment-charged storm surges which dump their load rapidly on the tidal flats. A unidirectional flood current, modified by oscillatory wave action, could have produced these climbing ripples, the type 2-to-type 1 transition representing an increase in the suspended load-bed load ratio, and a gradual decrease in current energy as storm surge waned. Such a waning current would be consistent with the model proposed by Jopling and Walker (1968) for the transition from type 2-to-type 1 ripple drift. The thin (2-10 cm) massive sheet-like intraclastic and/or ooidic or quartzose deposits are also interpreted as storm deposits on the tidal flats.

Metre-thick intervals of crossbedded quartzose-ooid dolarenite lithofacies must have been deposited under moderate energy conditions. Medium-bedded, medium- to coarse-grained tabular-planar crossbed sets were deposited by migration of low-sinuosity dunes and small sand waves. Current reversal is indicative of deposition by alternating ebb and flood tidal currents, and reactivation surfaces indicate alternation of low- and high-energy current regimes. Evidence of exposure and desiccation is lacking. A subaqueous bar environment just offshore of the tidal flats is proposed for these

sands. In this active tidal environment, ooids likely would form autochthonously, as opposed to having been transported landward to their site of deposition by storm currents.

Stromatolite biostromes, the third lithofacies, are interpreted as having formed in a subtidal environment as well. Typically the biostromes rest on crossbedded sand facies and display no signs of desiccation. The overall biostromal morphology is suggestive of a uniform, subtidal depositional environment. Intercolumnar fill consists of coarse, well-sorted quartzose-ooid detritus, and stromatolite branching is abundant, indicating a moderate- to high-energy current regime (cf. Horodyski, 1977). The columns closely resemble the Recent stromatolites which occupy a high-energy headland position along the coast of Hamblin Pool, Shark Bay (Logan et al., 1974).

Facies Sequences and Lateral Distribution:

Orderly vertical repetitions of lithofacies such as those described by James (1979) and Hoffman (1975) are not well developed in member A. The only example of vertical cyclicity occurs in section 8, where a sequence comprising shaly thin-bedded dolostone - clean thin-bedded dolostone - crossbedded dolarenite - biostrome, is repeated several times (Figure 11). A similar succession is described by Tucker (1977) from a

Late Precambrian mixed clastic-carbonate tidally-influenced succession in northern Norway. In the unit described by Tucker (1977), a cyclic vertical succession of thin-bedded dolostone - crossbedded dolarenite - biostrome - crossbedded dolarenite - thin-bedded dolostone was interpreted as the depositional record of transgression and regression, with the biostromes representing the deepest part of the succession. The lower transgressive half of Tucker's cycle is essentially equivalent to the thin-bedded dolostone - crossbedded dolarenite - biostrome cycles of member A. The significance of these possible transgressive sequences within member A is difficult to assess because of their apparent lateral restriction.

Lateral facies variations of member A cannot be documented properly, due to the small number of measured sections, their geographic proximity, and structural complexities which make accurate correlations difficult. In addition, no widespread marker beds have been recognized. Two apparent trends are the eastward pinchout of the entire member, and the marked westward thickening of the basal biostrome in section 2. These trends support a general north-trending depositional strike for this unit.

Paleocurrent diagrams for this shallow-marine unit are difficult to interpret in terms of paleogeography, due to the presence of shore-parallel, as well as

shore-perpendicular paleoflow.

MEMBER B

Member B mudstones and dolostones conformably overlie member A dolostones and pass gradationally upward into member C mudstones and dolostones, except west of central Dease Lake where the B-C contact is tectonic, being marked by a breccia unit 60-600 cm thick (Figure 10). The distinction between members B and C is clear at the type area around Dease Lake, but the distinction is extremely difficult in the Dismal Lakes area and eastward (Figure 12).

Member B consists of two lithofacies, a massive red mudstone and a tan silty dolostone-to-dolomitic siltstone. Subsidiary lithofacies restricted to the Dismal Lakes area include scour-based granulestones and nodular carbonate beds.

Mudstone Lithofacies:

Massive red mudstone beds, 20-150 cm thick, compose roughly 60 percent of member B, and are intercalated with tan dolostones which typically are 60 cm thick.

Current lamination, thin fine-grained sand laminae and less common wave ripple and climbing ripple cross-stratification are the dominant current features. Mud cracks and halite casts (Plate 3D and 3E) are common bedding surface features. Ripped-up mud-cracked layers have produced local intraclastic conglomerates.

Figure 12. Selected stratigraphic sections of members B and C. Lithologic symbols as in Figure 6. Mudstone coloration is as follows: r- red, g- green, gr- grey, b- black. General trend is from red oxidized mudstone to black reduced shale. Upward coarsening of member C in section 10 appears related to renewed or increased input from a nearby terrestrial source area.

Tan Silty Dolostone Lithofacies:

Twenty to 60 cm thick tan dolostone beds of this lithofacies, locally containing up to 50 percent well-sorted quartz silt and fine-grained sand, typically display parallel current lamination and small-scale ripple cross-stratification. Desiccation features are generally not present in these dolostone beds. Halite impressions are rare, but lenticular gypsum crystal pseudomorphs are common. Thicknesses of the dolostone beds are remarkably constant: the beds nowhere pinch out on the scale of individual outcrops.

Scour-based Lithic Granulestone-Sandstone Lithofacies:

The basal 40 m of member B in the Dismal Lakes area contains scour-based fining-upward sequences of lithic granulestone and sandstone which fill broad open channel(?) scours. Rare rock fragments are largely granitic; quartzite and quartz arenite are subordinate clasts. Farther west at Bebensee Lake, 6-9 m wide channels are filled with fine-grained sand and silt.

Nodular Dolostone Lithofacies:

Units dominated by nodular dolomicrite occur locally within the red mudstone lithofacies in the Dismal Lakes area. The carbonate layers grade downward from nearly pure carbonate to less than 20 percent carbonate over some 30 to 40 cm. Expansive precipitation of carbonates and multiple generations of brecciation of the carbonate

nodules and host mudstone has essentially obliterated all primary fabrics. The breccias are of jigsaw type, and obviously are autochthonous. Several of these nodular carbonate units are continuous for three to four kilometres. A hematitic-pisolite layer up to 60 cm thick is associated with the massive red mudstone facies in the Dismal Lakes area.

Lithofacies Interpretation: Member B

The member A-B transition is marked by an increase in fine-grained terrestrial siliciclastics, decrease in dolostone, and a complete absence of well-sorted medium- to coarse-grained quartz sand. Sedimentary structures in the mudstones (such as mudcracks and halite hoppers) are indicative of repeated flooding and drying. Current and wave ripple cross-stratification record low-energy currents during periods when the mudflats were flooded. Intraclastic conglomerates generated by storm reworking of desiccated, mud-flaked tidal flat sediments are common. Tan silty dolostone beds increase in abundance westward, probably reflecting a more marine-marginal setting relative to eastern extensions of the outcrop belt (Dismal Lakes area). The dolostones are dominated by parallel current lamination and current ripple cross-stratification. Desiccation features are rare, and evaporite casts are restricted to sparse lenticular gypsum pseudomorphs. Both sedimentary structures and

composition indicate a more distal, marginal-marine setting for these dolostones.

In addition to the eastward decrease in dolostone, the presence of nodular dolostone and scour-based granulestone lithofacies in the eastern part of the area implies an easterly clastic source. The concentration of scour-based lithic granulestones restricted to the eastern area suggest additional fluvial input by shallow braided streams and sheet wash deposits which are characteristic of wadi fans (cf. Glennie, 1970).

The nodular carbonate textures are interpreted as caliche horizons formed on fan surfaces during periods of non-deposition. They are analogous to those described by Allen (1960), Walls et al. (1975) and Harrison and Steinen (1978).

Analogous Holocene expansive arid mud flats with minor alluvial input are the Ranns of Kutch (Glennie and Evans, 1976), the western Gulf of California (Shearman, 1970) and parts of the Persian Gulf coast (Evans et al., 1969). The rarity of channel structures in member B, relative to the amount of clastics deposited, suggests that fine-grained clastics were first brought into a marine environment via major fluvial systems along strike of the present outcrop area, and later thrown back onto the flats during storm events. This presently is the major process depositing fine-grained sediment on the Ranns of Kutch (Glennie and Evans, 1976). The result is

an extensive low-relief plain underlain by evenly bedded sheet-like deposits displaying only minor wave reworking.

The carbonates, which are rhythmically interbedded with the siliciclastic mudstones, may have accumulated autochthonously in this coastal setting, being drowned periodically by onshore-directed floods carrying siliciclastic muds. Arid conditions indicated by the presence of gypsum and halite casts and caliche horizons make member B a likely candidate for a coastal sabkha environment.

MEMBER C

Member C represents the transition from mixed clastic-carbonate sediments to relatively clean dolostones of the overlying Kendall River Formation. Where not tectonically disturbed, upper and lower contacts are gradational (Figure 10). Although complete undisturbed sections of member C are nowhere exposed, a nearly undisturbed section occurs north of the eastern end of Dease Lake, in a broad west-plunging syncline. Here the member is approximately 50 m thick. This value is a minimum thickness, however, because minor faulting appears to cut out some of the section, and the upper contact is not exposed. Distinctive marker units which could be used to correlate sections are lacking in member

Member C consists of light grey dolostones and pink

mudstones in the Dease Lake area, making it distinct from the red mudstone and tan dolostone of member B. Eastward, the distinction between member B and C is less striking, being marked by a steady upward increase in dolostone, and the presence of intraclastic dololutite beds. The member contains four lithofacies: light grey dolostone and pink mudstone dominate from Great Bear to Dease Lake; tan dolostone and red-green-grey mudstone are the common lithofacies east of Dease Lake.

Light Grey Dolostone Lithofacies:

Light grey argillaceous dololutite beds (20-60 cm) are typical of member C west of section 9 (Figure 12). They may display lamination, but are typically massive. Stylolites are abundant in these beds. In the basal 5 m of the member, quartz silt laminae are present, but these die out upward. Hackly silicification is common, and some chert nodules appear to be silicified evaporites (on the basis of external "cauliflower" morphology of the nodules and anhydrite inclusions).

Pink Mudstone Lithofacies:

Pink mudstones are massive and display poorly developed parting. Individual beds are 10-30 cm thick, but contacts of mudstones with intervening dolostones are diffuse and nearly everywhere are modified by stylolites. No sediment coarser than clay size has been observed, and no sedimentary structures are visible.

Tan Dolostone Lithofacies:

Medium-bedded (10-30 cm) argillaceous dolostones in the upper half of the Dease Lake Formation in the Dismal Lakes area (sections 9 and 10, Figure 12) display faint parallel current and ripple cross-lamination. Higher in the section, halite casts, thin intraclastic conglomerates and dolostone increase in abundance.

Red to Green to Grey Mudstone/Shale Lithofacies:

Shaly mudstones are intercalated with the tan dolostones in the east (sections 9 and 10, Figure 12). Near the upper contact of the Dease Lake Formation, the shale colour changes from red to red-and-green and eventually to grey or black (Figure 12). Current structures are absent, but some halite pseudomorphs and rare mudcracks are visible on bedding surfaces in the upper portions of the section.

Lithofacies Interpretation: Member C

The grey dolostone-pink mudstone couplets characteristic of member C in the type area are essentially devoid of any environmentally diagnostic sedimentary structures. Intensive stylolitization in these lithofacies may partially explain the paucity of primary structures and textures preserved. The oxidized condition of the iron species (presumably complexed with clays) and the stratigraphic position of member C, separating coastal sabkha mudstones of member B from

peritidal dolostones of the Kendall River Formation, suggest that they are shallow-water deposits. The absence of desiccation features also suggests a shallow subtidal depositional environment. The persistent fine-grained nature of the lithofacies and the complete lack of current-generated sedimentary structures indicates that if deposited in a shallow subtidal environment, then it must have been a protected lagoon or embayment. The presence of silicified evaporites also is suggestive of a restricted-circulation environment.

In the east, member C represents the transition from lower siliciclastic mudstone-dominated deposits to dolostone-dominated deposits of the Kendall River Formation. A decrease in the abundance of halite hoppers and mudcracks (relative to member B) indicates less frequent periods of desiccation. The upward shale colour change from red-to-green-to-grey and black suggests a gradual transition to more reducing (subtidal ?) conditions. A marked increase in the percentage of dololite beds, a decrease in the quartz silt content and a marked increase in dololite intraclastic conglomerates are all indicative of a transition to a carbonate-dominated depositional environment. Decrease in siliciclastic input probably reflects gradual peneplanation of the source terrane, and the increase in intraclastic conglomerates reflects greater proximity to the effects of storm flooding from the adjacent

carbonate-dominated marine environment (Kendall River Formation). Fine- to medium-grained arenites increase in abundance upward south of Dismal Lakes (proximal to the Teshierpi Highlands, Figure 13A) but this coarsening upward trend is not reflected in sections north or west of Dismal Lakes.

PALEOGEOGRAPHIC RECONSTRUCTION OF DEASE LAKE FORMATION

Member A lithofacies accumulated in relatively high-energy tidal flat, subtidal shoal and biostrome settings which rimmed the siliciclastic delta front of the Fort Confidence Formation. The source of the well-sorted quartz arenites was most likely reworked Fort Confidence deposits. In essence, this laterally restricted carbonate member probably represents destructive-phase deposition on the seaward margin of the Fort Confidence delta (Figure 13a).

Paleocurrent measurements clearly indicate tidal reworking in a complex shallow-marine setting. However, due to the interference of shore-perpendicular and shore-parallel currents, such measurements cannot aid greatly in paleogeographic analysis. Based on the eastward pinchout of the member and crude northwest and southeast paleocurrent modes, the depositional strike during accumulation of member A was approximately northeast.

Lithofacies variations in member B also suggest a

Figure 13. Paleogeographic sketches showing inferred sequence of depositional environments for the Fort Confidence and Dease Lake Formations. Perspective looking southward. Members A-C are Dease Lake Formation members. See text for discussion.

north-trending depositional strike. Member B deposits are lacking in well-sorted quartz arenites, suggesting that the multicyclic marine terrigenous source had been covered before member B deposition. Medium-to-coarse siliciclastic detritus is rare throughout member B, occurring only around southern Dismal Lakes. This siliciclastic material is significantly less mature, both texturally and compositionally, than that in member A and the underlying Fort Confidence Formation, suggesting a probable first-cycle source to the south. Reactivation of the Teshierpi-Big Tree highland area (Kerans et al., 1981) is a likely source area. The coarse siliciclastic detritus decreases westward, and the percentage of quartz silt and dolostone increases. This compositional change is reflected in the change in interpreted depositional environments from wadi fan and continental sabkha in the southern Dismal Lakes area, to coastal sabkha in the Dease Lake area (Figure 13b).

Member C deposits reflect a near-complete stabilization of the siliciclastic source area, with only clay-sized and minor silt-sized siliciclastics being deposited. The percentage of dolostone increased gradually during deposition of the member, reflecting a gradational transition to the overlying carbonate-dominated depositional regime (Kendall River Formation). Although little can be said with certainty concerning the depositional setting of member C, a broad

Homogeneous, protected shallow subtidal(?) environment is indicated. The lack of significant facies variation in the member is interpreted to reflect a phase of minimal topographic relief during deposition of the Dismal Lakes Group (Figure 13c). The interfingering of intraclastic and rare ooidic storm deposits near the top of the succession signals the onset of transgression and deposition of the laterally time-equivalent offshore peritidal dolostones of the Kendall River Formation (Figure 13c).

PLATE 2. DEASE LAKE FORMATION

A. Bedding surface displaying interference ripples in the thin-bedded lithofacies (member A). Cap of marking pen is 5 cm long.

B. Irregular desiccation cracks formed in mud which accumulated in troughs of oscillation ripples. Variable thickness of mud drapes has controlled the irregular morphology of the desiccation cracks. Thin-bedded dolostone lithofacies of member A. Cap of marking pen is 5 cm long.

C. Convolute lamination in parallel laminated-to-ripple cross-laminated thin-bedded lithofacies (member A). Convolute lamination is sharply truncated by succeeding sediment layer, demonstrating the penecontemporaneous origin for the convolute lamination. Field of view is 30 cm wide.

D. Medium-bedded quartzose dolarenite of crossbedded dolarenite lithofacies (member A), in which 2-8 cm tabular-planar crossbed sets show current reversals. Laminations weathered in relief contain higher percentages of quartz sand. Hammer handle for scale.

E. Example of repetitive cross-lamination sequence consisting of low-angle tabular-planar cross-lamination (T) overlain by type 2 ripple drift lamination (2) which grades transitionally into type 1 ripple drift lamination (1). See text for discussion. From crossbedded dolarenite lithofacies of member A. Notebook is 17 cm long.

F. Multiple reactivation surfaces (arrows) on tabular-planar crossbed set from crossbedded dolarenite lithofacies (member A). Cap of marking pen is 5 cm long.

PLATE 2

A

B

C

D

E

F

PLATE 3. DÉASE LAKE FORMATION

A. Isolated bioherm of SH-C columnar stromatolites of biostromal lithofacies (member A), displaying passive radial branching. Bioherm is flanked by thin-bedded lithofacies which at least in part is organically laminated at this locality. Hammer for scale.

B. Cross-section through unique tufted biostrome of biostromal lithofacies (member A). Tufted peaks of continuous organic mats form ridges in the third dimension (see C of this plate). Synoptic relief of peaked ridges is 4-20 cm. Notebook is 17 cm long.

C. Plan view of tufted stromatolitic biostrome illustrated in B above. Note the irregular intersecting pattern of ridges, which, in conjunction with the continuous nature of internal lamination, demonstrates that the resemblance of these forms to desiccation polygons is only superficial. At this locality, the stromatolitic ridges are exhumed, displaying an original 4-20 cm synoptic relief. Notebook is 17 cm long.

D. Halite hopper casts in red mudstone lithofacies of member B.

PLATE 3

A

B

C

D

CHAPTER 3. KENDALL RIVER FORMATION

Conformably overlying the Dease Lake Formation is a sequence of thin- to medium-bedded grey dolostones. These dolostones are particularly well exposed along the west bank of the Coppermine River north of its junction with the Kendall River. The name Kendall River Formation is proposed here for these dolostones, which are equivalent to Unit 14 of Baragar and Donaldson (1973) as modified by Kerans et al (1981). Incomplete sections of the Kendall River Formation occur below the massive bluffs of the Sulky Formation, along the north shore of Dismal Lakes and throughout the outcrop area. The most complete sections are along the Coppermine River (Figure 15, section 7), in a gorge at the northwest end of Dismal Lakes (Figure 15, section 5), and in an open syncline due north of Dease Lake (Figure 15, section 2). West of Dease Lake, strata of this formation occur only in small exposures within fault slices.

The Dease Lake-Kendall River Formation contact, as described in the previous chapter, is gradational, and is arbitrarily designated as the base of the first stromatolitic dolostone that appears in the section (Figure 14, Figure 15). A biostrome of LLH and SH stromatolites (1-6 m thick) marks the top of the Kendall River Formation. This biostrome occurs throughout the Dismal Lakes outcrop belt as a massive-weathering tan to

Figure 14. Typical stratigraphic column of the Kendall River Formation showing three-member subdivision and relationships with underlying and overlying formations. Symbols as in Figure 6.

blue-grey bed which is sharply overlain by shale of the lower Sulky Formation.

The well-bedded dolostones of the Kendall River Formation show a decrease in siliciclastic shale as they grade upward from intercalated shale and dolostone of the Dease Lake Formation. Concomitant with the decrease in shale is an increase in clastic carbonates (ooid-intraclastic units) and cryptalgalaminites (cf. Aitken, 1967), marking the transition to shallow-water carbonate deposition. Shallow-water deposition was abruptly terminated by rapid transgression and deposition of lower Sulky Formation shales which drape the Kendall River upper biostrome (see Chapter 4).

In order to analyse the internal stratigraphy and depositional history of the Kendall River Formation, essential components, or lithofacies, must be described. East of Dease Lake, where exposure is adequate, the formation can be divided into three members on the basis of lithofacies distribution, a lower shaly dololutite-rich member (A), a clean clastic carbonate-dominated member (B) and an upper biostrome member (C). Vertical and lateral lithofacies variability is then assessed in order to interpret changes in depositional pattern and paleogeography during Kendall River deposition.

LITHOFACIES DESCRIPTIONS

Mud Dominated Lithofacies:

Siliciclastic Shale

Thin black to dark-green siliciclastic shale interbeds commonly separate massive dololutite beds in the lower portion of the Formation (Plate 4A). The only apparent sedimentary structures are rare halite hopper impressions. Shale beds commonly drape underlying lithologies and are nowhere associated with scour surfaces.

Interpretation

Thin siliciclastic shale beds are interpreted as subtidal, and their black to dark green colour is attributed to a reducing environment of deposition. The associated massive dololutites are also interpreted as subtidal, as discussed below.

Thinly Intercalated Siliciclastic Shale and Dololutite:

Black siliciclastic shale and light-weathering argillaceous dololutite occur as thinly intercalated platy weathering units 20-60 cm thick. Parallel current lamination and rare cut-and-fill are the only sedimentary structures observed in this lithofacies.

Interpretation

Thinly intercalated shale and dololutite with subtle current bedding, but with only rare erosional structures and no coarse clastic debris or desiccation features, are

interpreted to have been deposited in a low-energy subtidal environment. The delicate interlayering of carbonate and siliciclastic sediment is indicative of a transitional siliciclastic-carbonate environment.

Dololutite

Light-weathering argillaceous-to-clean dololutite occurs in 10-80 cm thick units, some of which can be traced for up to 75 kilometres (Plate 4A and 4B). Parallel current lamination, desiccation features, thin granule layers and small-scale (10 by 40 cm) intraclast-filled scour channels are common sedimentary structures associated with the dololutite. The channeled dololutite bed in Plate 4B can be traced for 75 km, showing a range of thickness limited to 40 to 70 cm. Laminar-fenestral porosity (cf. loferites of Fisher, 1964) occur in the dololutites, as do algal mat structures, but both are rare.

In the lower portion of the section, dololutites which are massive and rarely display clastic textures are merely thicker equivalents of dololutite in the shale-dololutite lithofacies described above (Plate 4A).

Interpretation

Light-weathering dololutite beds display evidence of being deposited in relatively shallow water.

Well-developed mud cracks, fenestral fabric, scour channels, parallel current lamination and dolarenite

laminae are indicative of alternating periods of low and high current energy and periodic subaerial exposure. Similar associations of sedimentary structures are typical of Holocene tidal flats, such as the Three Creeks belt on northwest Andros Island (Shinn et al, 1969; Hardie, 1977) and parts of the Abu Dhabi sabkha (Evans et al, 1969), or pond-marginal facies of the Coorong Lagoon (Muir et al, 1981). The major difference between dololutite beds of the Kendall River Formation and, for example, the Three Creeks region, is the lateral continuity of the former. This lateral uniformity of individual beds, such as the channeled dololutite bed described above, is characteristic of the dololutite lithofacies, and to a certain extent, the formation as a whole (e.g. upper biostrome). Laterally extensive deposits do not occur in most modern tidal flats, which is probably a result of the greater, although subtle, depositional topography in modern environments. Ginsburg and Hardie (1975) have demonstrated that even slight variations in depositional topography can produce significant differences in depositional texture in carbonate tidal flats. A remarkably uniform depositional surface is implied for accumulation of the dololutite lithofacies.

Massive dololutite beds without current lamination, which occur in the lower part of the formation, are interpreted as subtidal, based on the absence of exposure

features and the association with tidalite facies.

Clastic-Dominated Lithofacies:

Intraclastic Packstone-Grainstone

Two types of intraclastic packstone-grainstone are recognized. Two- to ten-centimetre thick beds containing greater than 90 percent angular, tabular mud flakes and algal-mat fragments, with interstitial micrite, form tabular beds which conformably drape over stromatolites (Plate 4C) and other lithofacies.

The second type of intraclastic lithofacies is packstone and grainstone consisting of well-sorted and rounded grains, and subordinate amounts of angular mud chips (Plate 4E). These dolarenites occur as sheet deposits and as channel-fill lithologies cut into dololutite (Plate 4D) and SH stromatolites (Plate 5C). Rounded dolarenite beds commonly are crossbedded and are between 10 and 110 cm thick (Plate 4D).

Interpretation

Intraclast drapes are interpreted as storm deposits. Their allochthonous origin is supported by the absence of associated scour channels or other high-energy features which would be expected if the intraclast beds represented in-situ high-energy accumulations. Instead, the drapes overlie fine-grained lithologies without erosion, indicating that deposition was from a

high-density suspension. Finally, the monomict composition of these intraclast breccias suggests little mixing before deposition. This lack of mixing is best explained by local derivation and minimum transport of clasts ripped up from desiccated mud flats. Thin sheet-like storm deposits consisting of flat mud chips are common on modern tidal flats (Hardie and Ginsburg, 1977) and appear to be similarly abundant in the Kendall River Formation.

Upper portions of intraclast beds may be reworked into edgewise conglomerates (cf. Hoffman, 1975) or stone rosettes (cf. Ricketts and Donaldson, 1979). Ricketts and Donaldson (1979) have shown that such edgewise and rosette patterns in flat-chip breccias are indicative of reworking and stacking of chips in the wave swash zone.

The absence of micritic matrix and the greater degree of sorting and rounding of the second variety of intraclast grainstones (Plate 4E) are indicative of relatively long-term reworking of sediments in a moderate- to high-energy environment. The SH morphology of stromatolites associated with these intraclast grainstones also supports a moderate- to high-energy current regime; lateral linkage of stromatolites by algal mats is prohibited by sediment movement in intervening channels. A wave swash zone or beach seems a likely environment for the formation and deposition of intraclast grainstones.

Mixed Oncolitic-Oolitic-Intraclastic Packstones

Another variation on the intraclast grainstones are mixed oncolitic-oolitic-intraclastic grainstones and packstones. These deposits are typically scour-based and may occur as intercolumnar fill in some stromatolitic units. Beds are 10-60 cm thick and rarely display cross-stratification.

Interpretation

Mixed grainstones and packstones occur as drapes, channel-fills and tabular beds. The bimodal grain size distribution (the ooids are well-sorted; the intraclasts are poorly sorted and much coarser) suggest that these deposits represent a mixture of two distinct lithologies. Mixing may have occurred in single events such as storms, or may have occurred through multicyclic reworking of single-event lithologies (intraclast or ooid storm drapes) by normal background tidal processes. The thin sheets are interpreted as storm layers, and channel-fills as high-energy intertidal deposits.

Ooid Grainstone-Packstone

Beds containing greater than 90 percent ooids are rare in the Kendall River Formation. Individual beds generally are thinner than 40 cm, averaging 10-20 cm (Figure 15). These thin beds are massive and non-graded. In the Bebensee Lake area, where thicker (40-100 cm) ooid packstone-grainstones occur, low-angle and herringbone

crossbedding are common.

Consistently thick ooid corticies are composed of fine concentric layers (Plate 4F). Superficial ooids (cf. Bathurst, 1975; Loreau and Purser, 1973) are rare. In thin section, well-preserved non-silicified ooids display a pseudo-isogyre extinction pattern, but no radial fracturing or other relationships indicating primary radial vs. tangential structure (cf. Halley, 1977; Sandberg, 1975; Wilkinson and Landing, 1978) were observed.

Interpretation

The critical question concerning these thin ooid sheets is their allochthonous or autochthonous origin. A storm-transported origin is proposed for all of the thin ooid sheets, based on their thin discontinuous nature and intercalation with intertidal and supratidal sediments. Loreau and Purser (1973) have described superficial ooids (ooids with corticies less than 1 mm thick) as typical of allochthonous storm-deposited ooids which have been removed from their site of formation, and hence ceased to accrete. In contrast, ooids of the Kendall River Formation are ubiquitously well developed. Modern environments of ooid formation require a water depth of up to several metres, and a high-energy shelf-edge setting (Loreau and Purser, 1973; Hine, 1979; Milliman, 1969). Ooid distribution maps of the Persian Gulf (Loreau and

Purser, 1973) and the Great Bahama Bank (Purdy 1963) have shown that analogous thin allocthonous ooid beds are common landward of ooid shoals, presumably having been deposited by onshore-directed storms. Thicker ooid beds (40-100 cm) are rare (Figure 17b; Plate 5C, and lithofacies distribution section), but where they occur repeatedly they may indicate greater proximity to areas of autochthonous ooid formation, or possibly small active ooid shoals.

Algal-Laminated Lithofacies:

Flat Laminated Algal Mats

A complete spectrum of algal mat fabrics occurs in the Kendall River Formation. Smooth, pustulose and palisade mat types (Logan et al., 1974; Hardie and Ginsburg, 1977) are most common. Smooth mats are intercalated with lenses of clastic carbonates and are commonly scoured. Pustulose and palisade mats (Plate 5A) are closely associated with tepee structures, syneresis features and mud cracks.

Some silicified algal mats in lower member B contain distinctive white nodular chert masses, creating a texture which superficially resembles chicken wire anhydrite (Plate 5B). In addition, white chert nodules do not display a pore-fill fabric, but rather one that would be expected if the chert were replacing a precursor fabric. No anhydrite inclusions have been found in the

nodules, but compression of mats by expansive growth of the nodules, prior to silicification, suggests an expansive early diagenetic origin for the nodules (Plate 5B).

Interpretation

Systematic studies on algal mat fabrics of Recent (Logan et al., 1974; Hardie and Ginsburg, 1977) and ancient (Ricketts, 1979) sequences have demonstrated that there is a correlation between algal mat texture, exposure index (Ginsburg and Hardie, 1975) and sedimentation rate. The more extensive the subaerial exposure and desiccation, the greater the tendency to develop pustular or blister mat morphologies. The greater the degree of flooding, and hence sedimentation, the greater the likelihood that introduced sediment will fill irregularities in the mat surface, while exposure time decreases. Such conditions would favour the generation of smooth mat textures.

A similar correlation applies for algal mat lithofacies of the Kendall River Formation. Although smooth and irregular mat fabrics may be closely interlayered, they are not juxtaposed along the same depositional horizons. Pustulose and palisade mats are more commonly associated with thin continuous flat algal mat beds which display fenestral porosity, desiccation and tepee structures. These deposits are characteristic

of higher intertidal and supratidal environments (James, 1979; Shinn et al., 1969; Shinn, 1968; Roehl, 1967; Laporte, 1967). Smooth mats link LLH domes, and are intimately interbedded with the lenticular and channel-fill intraclastic lithofacies. This mat fabric is characteristic of subtidal and lower intertidal environments. As noted by many authors (e.g. Garrett, 1970; Hoffman, 1974), competitive grazing by metazoans was not a factor in restricting algal mat distribution in the Precambrian, thus allowing subtidal algal mats to flourish.

Chicken-wire texture in algal mats of the Kendall River Formation has developed by expansive growth of nodules of unknown composition. Although modern chicken wire fabrics are of early diagenetic origin, they appear to be environmentally indicative in that they develop only in the supratidal zone (Bush, 1973; Kinsman, 1966). A similar supratidal sabkha origin is proposed for Kendall River chicken-wire texture, although no relic evaporite minerals remain. A silicified evaporite interpretation for the chert nodules is strongly supported by the early displacive nature of the nodules.

Stromatolite Mounds

Stromatolite morphologies are varied, but a useful first-order distinction can be made between LLH and SH morphologies. LLH domes are typically symmetric and

broad, with low (10-20 cm) synoptic relief (Plate 4C). Internal lamination is generally smooth, but palisade texture is locally important (Plate 5A). In the type section of the Formation, alternating palisade and smooth mat textures can be observed in a single dome. LLH domes may be spaced at distances of 50-100 cm, or may be closely packed (Plate 4C). Both spaced and close-packed domes have drapes of intraclastic dolarenite or dololutite as intercolumnar fill. Scour features are rare.

SH or club-shaped stromatolites are less common than LLH morphotypes, increasing in abundance in upper portions of the Formation. SH stromatolites may be of SH-V (turbinate of Hoffman, 1969) or SH-C (club-shaped of Hoffman, 1969) morphology, are bounded by scour channels of clastic dolarenite and typically are constructed of smooth mats (Plate 5C). SH stromatolites are generally small, being 10-30 cm high and 10-20 cm, wide and are both elongate and symmetrical in plan.

Beds consisting largely of oncolites occur in a few places (Plate 5D). Some oncolites became stabilized during growth, and developed into SH-C or LLH stromatolites.

Interpretation

Smooth-laminated LLH-S and LLH-C domes are interpreted as low-energy subtidal or protected

intertidal forms. Desiccation features are absent as are high-energy lithofacies such as intraclast grainstone channel-fills. When clastic material fills inter-areas it is as a drape, and does not truncate underlying algal laminations. LLH forms are more common in lower portions of the Formation, where shale and massive dololutite lithofacies are also common. This association is consistent with a low-energy shallow subtidal interpretation. A case can be made for some of the LLH stromatolites having been periodically exposed, because palisade mat texture occurs in the cores of some mounds (Plate 5A).

SH forms, based on their morphological similarity to high-energy headland forms described from Shark Bay (Logan et al., 1974; Hoffman, 1976) and on their association with higher energy clastic lithofacies, are interpreted as intertidal and upper subtidal forms which inhabited shorelines receiving moderate wave energy.

Oncolite wackestone-packstones formed in a wave-agitated environment. Such an environment would be required to roll these structures periodically to allow accretion of concentric laminae. A shallow subtidal environment is suggested. This interpretation is consistent with interpretations of similar facies in Phanerozoic deposits (cf. Halley, 1975; Wilson, 1975).

CYCLIC SEDIMENTATION

Many shallow-water carbonate sequences are known to be cyclic. The most common repetitive theme in peritidal dolostones is the shallowing-upward cycle (James, 1979). Briefly, this cycle consists of a basal lag which represents an abbreviated record of transgression, a subtidal bioturbated wackestone unit, an intertidal unit of variable current or algal-laminated sediment, and an upper supratidal unit of intercalated storm deposits and algal marsh and pond sediments. The process which generates this cyclic shoaling is one of transgression (subsidence and/or eustatically controlled) followed by progradation of rapidly accumulated carbonate.

Most well-developed cyclic shallowing-upward sequences described in the literature occupy continental margin settings (e.g. Hoffman, 1975; Reinhardt and Hardie, 1976; Mossop, 1979; examples in James, 1979) and tectonically active continental margin basins (e.g. Delaware Basin, Meissner, 1972; Italian Dolomites, Fischer, 1964), where periodic subsidence caused frequent punctuation of cycles. In stable intracratonic settings such as the Michigan and Williston Basins, subsidence rates were slow and steady, causing environmental belts to migrate laterally, obscuring any vertically repetitive prograding sequences. In these sequences, sedimentation kept pace with subsidence, producing a monotonous peritidal succession.

The near-absence of cyclicity (30 metres of total

700 m measured section; Figure 15) during Kendall River Formation deposition is interpreted to reflect the tectonic quiescence of this period of deposition, as will be discussed below.

Although cyclic sedimentation is rare in the Kendall River Formation, four small-scale cycles (perhaps more aptly termed repetitive sequences) have been recognized. These cycles are described below and give an indication of the styles of deposition in this peritidal carbonate unit.

"Sabkha Cycles" of the Coppermine Section:

Lower member B (see next section) of the Kendall River Formation along the Coppermine River (section 7, Figure 15) contains five 0.5-1.75 m shallowing-upward cycles. These cycles consist of a lower dololutite-dominated component, a central stromatolitic and dolarenitic component, and an upper silicified chicken-wire algal mat lithofacies. In these cycles, the shallowing trend is not uniform. Some dololutites are variably desiccated, and thus in part must be intertidal. Some crossbedded dolarenite beds display mudcracked upper surfaces, suggesting an upper intertidal position of final deposition. In general, however, the dololutite is subtidal, the clastic-stromatolitic section is intertidal, and the chicken-wire lithofacies is supratidal. These environmental interpretations are

based on lithofacies discussions presented above. Lithofacies and cycle interpretations are summarized in Figure 16. Similar sabkha cycles are reviewed by Kendall (1979) and Mossop (1973). Sabkha cycles discussed in these reviews typically have a thicker evaporitic cap, and also show evidence of gypsum in the cycle as well as anhydrite. A less intensive potential evaporation (cf. McKenzie et al., 1981) might explain the relatively thin evaporite development; the absence of gypsum is actually rather speculative, because all textures now are silicified. If the above-described cycles are sabkha cycles, then their presence in the central Kendall River Formation implies local extreme restriction coupled with moderately high potential evaporation, as would occur in an arid lagoon setting.

Tidal Flat Cycles:

The sequence dololutite, flat mat, erosion, SH-C and intraclastic dolarenite channel-fill, dololutite drapes (Figure 17a) is repeated frequently in member B, especially in the gorge section at the west end of Dismal Lakes (Figure 15, section 5). This repetitive sequence is interpreted as another variety of shallowing-upward cycle that was generated by shoreline progradation. The basal dololutite is massive and of probable subtidal origin; the overlying algal mat is smooth and is interpreted to have formed in a lower intertidal setting.

Figure 16. Measured section of "sabkha" cycles of lower member B from the Coppermine River section. Arrows denote individual cycles and bar graph on right illustrates interpreted depositional environments. Symbols as in Figure 15.

FIGURE 16

Figure 17. Clastic-stromatolitic cycles of member B. A) Schematic section of small-scale shallowing-upward cycles consisting of lutite-scour-stromatolite-intraclastic conglomerate. B) Measured section of large-scale deepening-upward (transgressive) cycles of upper member B from Rebensee Lake area. Cycles consist of mudcracked dololite-crossbedded ooid grainstone-biostrome. See text for further discussion. Arrows denote individual cycles. Bar graph to right of each column shows interpreted depositional environments. Symbols as in Figure 15.

FIGURE 17

Scouring of the algal mat-dololomite succession, subsequent growth of SH-C stromatolites, and deposition of clastic intercolumnar sediment indicates a higher energy depositional setting, perhaps on the upper shoreface. Final deposition of dolomite capping the dolarenite indicates a return to a lower energy subtidal environment as a result of transgression.

A closely analogous sequence of flat mats grading upward into LLH-S and finally into SH-C stromatolites with intercolumnar dolarenite channel-fill, is similarly interpreted to represent a gradual increase in current energy related to shoaling. These cycles are on the order of 40-100 cm thick, and occur in several sections in member B.

Biostrome Cycles:

More regularly developed cyclic sequences of larger scale occur near the top of the Kendall River Formation in the Bebenssee Lake area (Figure 17b; Figure 15, section 3). Cycles are 1-4 m thick and consist of subequal components of shaly dolomite, ooid grainstone, and a biostromal cap (Plate 5C).

Dolomites are thin-bedded and are commonly desiccated. Overlying ooid grainstones are low-angle cross-laminated, well-sorted beds up to 1 m thick. The ooid grainstones occur in both conformable and erosional contact with underlying dolomites. The upper

stromatolitic unit of LLH-S and SK-C stromatolites rests both conformably and erosionally upon underlying ooid grainstones. Where erosion of the grainstone has occurred prior to stromatolite growth, bioherms developed on topographically high erosional remnants. Where erosion was not significant, homogeneous biostromes developed on ooid beds.

Desiccated dololutites are of intertidal origin. Crossbedded ooid grainstones are interpreted as deposits of autochthonous ooid shoals in this sequence, being too thick and well-sorted to have been generated during a single storm event. If these beds are autochthonous, or at least proximal to their area of formation, a high-energy subtidal setting is implied. Stromatolite bioherms and biostromes indicate large-scale stabilization of ooid shoals, occasionally preceded by a period of erosion. Stabilization by algal mats required an initial decrease in current energy, and hence substrate mobility, allowing the mats to spread. After stabilization, stromatolites could have survived in either a higher- or lower-energy environment.

This lutite-ooid-stromatolite sequence might be explained either by steady transgression or by transgression followed by regression. Initial transgression is required by the tidal-flat-dololutite to subtidal transition. The transition from ooid shoal to stromatolite biostrome-bioherm may have occurred after

sea level rose to a level at which currents no longer could actively transport sediment, allowing algal mats to stabilize the substrate. This situation would be analogous to Holocene transgression on the northern Bahama Platform, where bank-marginal portions of some ooid shoals have been rendered inactive (Hine, 1979). The deeper seaward portions of these inactive banks are already being stabilized by grasses and subtidal algal mats (Hine, 1979). Alternatively, if ooid shoals had built up above wave base, protected back-barrier settings would have formed, and these could have been stabilized by stromatolites. Deposition during continuous transgression is favoured, because no good evidence of shoaling has been observed

VERTICAL LITHOFACIES VARIATIONS AND MEMBER SUBDIVISIONS

In the two best-exposed sections of the Kendall River Formation (sections 5 and 7, Figure 15), a clear distinction can be made between a lower mud-rich member and an upper member which consists predominantly of clastic dolarenite and cryptalgalaminite (cf. Aitken, 1967). The contact between these two members is gradational, and some lithofacies overlap occurs even well into the upper member. The distinction between the upper and lower member is sufficiently distinctive to be mappable at any scale, but subdivision of the upper member as suggested in Kerans et al. (1981) can be

accurately defined only in continuous sections, and is not considered a practical map unit. Although the middle member will not be subdivided, significant variations occur within it, and these will be discussed in sections concerned with vertical and lateral lithofacies variations.

The uppermost member of the Kendall River Formation is a laterally persistent biostrome which occurs in all sections from west of Limestone Point to east of All Night Lake (Figure 15). Designation of this biostrome as an informal member is justified by its importance as a stratigraphic marker throughout the basin.

Member A:

Member A is completely gradational with the upper Dease Lake Formation. This transition has been observed only in sections 5 and 7 (Figure 15). The transition is exposed north of Dease Lake, but structural complexities there confuse stratigraphic relationships. The Dease Lake-Kendall River transition is manifested by a change from subequal abundance of dololomite and siliciclastic mudstone, to grey dololomite with thin black shale partings (Plate 4A). The first appearance of stromatolites, interpreted to signal the transition from coastal sabkha to marine carbonate deposition, is used to define the Dease Lake-Kendall River contact.

Member A is a minimum of 11 m thick at section 7;

its true thickness probably approaches 20 m, as in section 5 (Figure 15). Massive- to current-laminated dololutite lithofacies dominates the member, occurring in 10-30 cm beds separated by thin black shale beds (Plate 4A). Subordinate lithofacies are symmetrical low-relief LLH-S domes and thin 2-10 cm ooid and intraclastic packstone lithofacies. Desiccation structures are rare, but mud cracks and halite hoppers occur in the upper parts of member A.

Member B:

The transition from dololutite-dominated member A to clastic-dominated member B is gradational. Stromatolite beds increase in number, and SH forms associated with clastic channel-fill material become subequal to LLH forms. Mixed oncolitic-ooidic-intraclastic packstones are an important component of member B, as are mud-cracked and fenestral dololutite lithofacies. Outlines of broad, shallow tidal channels (0.6 by 3 m) carved into tidal flat dololutite are preserved locally, as are storm berms of intraclastic conglomerate up to 20 cm thick. These large-scale structures are considered definitive evidence of tidal flat and shoreface deposition respectively.

In upper member B, the percentage of ooid grainstones shows a marked increase, both in the Coppermine section (section 7, Figure 15), and farther west (sections 3 and

5). Interbedded with these ooid beds are intertidal lithofacies such as SH-C stromatolites and scour-channel fills, mud-cracked and fenestral dololutite and flat crenulate algal mat. At Bebensee Lake (section 3) and Greenhorn Lakes (section 1), 0.5 - 1.5 m thick bioherms and biostromes are interbedded with the ooid packstones and grainstones.

Member C:

The upper biostrome of the Kendall River Formation is considered worthy of member status in that it is the most consistent marker bed in the lower half of the Dismal Lakes Group. Though lateral changes occur in this unit (Figure 15), it maintains a consistent stratigraphic position throughout the outcrop belt. In addition, its typically massive-weathering character provides generally good outcrop.

The biostrome everywhere is approximately 6 m thick. From September Lake eastward, a 4 m-thick current-laminated dololutite and flat algal unit separates a lower 1 m LLH biostrome from an upper biostrome composed of SH-V bioherms (Figure 15, sections 8-11); north of Dismal Lakes, the biostrome consists of 6 m of 10-20 cm diameter SH-C columns with less common broad LLH domes. The stromatolites are consistently symmetrical except at the extreme northwest end of Dismal Lakes, where a weak west-southwest (260°) preferred

elongation is developed. West of Dismal Lakes the biostrome is composed of SH-C stromatolites which locally deviate from a vertical growth axis (Figure 15, section 11).

INTERPRETATION

The overall coarsening-upward sequence from the upper Dease Lake Formation, through member A and lower and upper member B (Figure 14) is indicative of increased current energy during gradual transgression. The transition from coastal sabkha environment (member C of the Dease Lake Formation) to a low-energy subtidal environment during member A deposition marks the initial marine phase of transgression of the Dismal Lakes Platform. The absence of any intervening coarse clastic shoreline lithofacies between the marginal marine upper Dease Lake Formation and the subtidal shale-dolomite dominated lithofacies of member A indicates a protected low-energy environment. An offshore barrier, or simply an extremely low depositional slope, must have been present to damp wave activity, allowing accumulation of a muddy shoreline facies.

Member A, with its subtidal stromatolites and occasional storm-derived clastic drapes, is interpreted to have accumulated in a subtidal lagoon.

Member B lithofacies formed in both tidal flat and moderate- to high-energy subtidal environments.

Lithofacies changes are rapid (each lithofacies interval is rarely thicker than 60 cm), and show little evidence of consistent vertical change or cyclicity. A complex mosaic of peritidal depositional environments is indicated. Clearly intertidal flat (mud-cracked and fenestral dololutite, flat crenulate algal mat, storm-deposited packstones and grainstones) and shallow subtidal or lower intertidal (LLH and SH stromatolites, thick ooid and intraclast grainstones with occasional herringbone cross-stratification, oncolitic wackestones) environments predominated. This complex lithofacies mosaic is comparable to depositional patterns in modern carbonate tidal flats of Andros Island (Shinn et al., 1969).

A trend toward higher energy subtidal deposition in upper portions of member B is indicated by the higher percentage of ooid grainstone. However, water depths probably never exceeded 2-4 m, because ooid beds are intercalated with intertidal dololutites (e.g. Figure 17b). In addition, ooid grainstones nowhere are thicker than 1.5 m, suggesting that extensive development of subtidal ooid shoals did not occur. Clearly, these subtidal environments were short-lived. Ooid grainstones and associated bioherms and biostromes (dololutite-ooid-biostrome cycles of Bebensee Lake) probably accumulated slightly seaward of the tidal flat and high-energy shoreline lithofacies of lower member B.

The gradational contacts between member C of the

Dease Lake Formation and members A and B of the Kendall River Formation, and the complex interfingering of lithofacies therein, implies gradual transgression of time-equivalent facies belts (Figure 18). Coastal sabkha mudstones of the uppermost Dease Lake Formation graded seaward into protected subtidal lagoonal deposits of member A of the Kendall River Formation. Member B tidal flat deposits acted as a protective barrier for the member A lagoon, allowing only occasional storm-deposited drapes of intraclastic or ooidic material to breach this barrier. Offshore of the tidal flat complex, thin sheets of ooid sand formed in shallow high-energy subtidal settings which occasionally shoaled into the intertidal realm.

The lateral continuity of member C, which thins only slightly from west of Limes Point in Dease Arm to the eastern limit of exposure, a distance of 200 km, is remarkable. Some lateral variability is observed (Figure 15), such as the change from smooth algal mats to thrombolitic mats from west to east, and the change in capping lithologies from black shale west of September Lake to edgewise conglomerate in a red shale matrix east of this position.

A point of interest is the extent of time-transgression involved for deposition of this biostrome. Because the exposed cross-section of the Kendall River Formation is oblique to depositional

Figure 18. Schematic block diagram of depositional setting of the Kendall River Formation showing transgressive facies relationships and complex microenvironmental depositional pattern. S- coastal sabkha facies of Dease Lake Formation, member C. A- member A of Kendall River Formation. B- member B of the Kendall River Formation. Dotted outline of ooids and intraclasts represents a storm washover fan. See text for further discussion.

strike, the strike-perpendicular section must be only a fraction of the 200 km outcrop length. In addition, east of Bebensee and Dease Lakes (Figure 15, sections 4-12) the biostrome overlies peritidal sediments, whereas from these areas west it rests on subtidal lithofacies. This eastward overstepping of facies indicates that at least transgression and deposition of the biostrome were more rapid than that of underlying facies.

On the basis of its internal consistency and facies overstepping relationships, member C is inferred to be an essentially synchronous unit. Minor algal mat microstructure changes within the biostrome, and the changes observed in immediately overlying facies, are consistent with an eastward shoaling reflected in the formation as a whole. Singh (1979), in discussing his work on Proterozoic biostromes in India, suggested that biostromes showing extreme lateral persistence might be equivalent to time lines.

Such a homogeneous platform-wide depositional environment as that represented by member C could only have formed by rapid flooding of a very low-relief depositional surface. This is consistent with the general lack of lateral variability within the Kendall River Formation. Finally, major transgression of the platform which directly followed member C deposition (Chapter 4) suggests that the flooding that led to member C deposition was a minor precursor pulse of this later event.

The absence of cyclicity in this vertical sequence has been discussed above. A slow non-episodic subsidence rate and/or gradual eustatic rise allowed lateral facies migration to dominate the depositional record.

Scattered, minor progradational cycles (Figures 16 and 17) are exceptions, and probably indicate brief periods of either more active subsidence or eustatic rise, followed by shoreline progradation. In general, the extreme stability during accumulation of the Kendall River succession is interpreted as characteristic of a stable intracratonic setting. Thicker transgressive cycles occur only at Bebansee Lake (section 3), situated west of Leith Line where subsidence generally was more active throughout Dismal Lakes sedimentation.

LATERAL FACIES VARIABILITY AND PALEOGEOGRAPHY

A stable peritidal platform deposit such as the Kendall River Formation is not particularly well suited for paleogeographic analysis. As a first-order approximation, it can be said that during Kendall River deposition, the entire Coppermine Homocline region must have been a broad, featureless surface of deposition. Lateral continuity of the member C biostrome, which exhibits only minor facies changes across the entire area, exemplifies this environmental homogeneity (see below).

Although a slight westward thickening of the formation occurs (Figure 15), this may be a function of

slightly greater subsidence rate, and hence more rapid accumulation west of Leith Line. A minor environmental gradient from east to west is suggested by the greater percentage of ooid grainstones in the Greenhorn and Bebensee Lake areas (sections 12 and 10). In terms of relative water depth at any given time, however, an increase of only a few metres would have been sufficient to cause this change.

An overall eastward thinning of the formation, an increase in flat algal lithofacies and tepee structures, and a decrease in ooid-intraclast grainstone sheets are the most important eastward changes (cf. sections 1 and 3). In general, it can be said that the Kendall River Formation shoals eastward, but that the eastern limit of exposure is still considerably west of the depositional zero-edge.

The greater thickness and increase in higher energy lithofacies to the west and northwest of Dismal Lakes, and slight thinning and increase in percentage of supratidal lithofacies in easternmost sections, suggests that a northerly or northeasterly depositional strike existed during Kendall River deposition. Unfortunately stromatolite elongation trends and measurable crossbedding are rare, and thus cannot be used to aid in the paleogeographic reconstruction. Siliciclastic detritus, restricted to clay-sized material, is distributed homogeneously throughout the

formation, and thus cannot provide information regarding paleoslope or terrestrial source-area proximity. As previously mentioned, a remarkable aspect of this formation is the dearth of lateral facies variation.

PLATE 4. KENDALL RIVER FORMATION

A. Typical outcrop of member A, consisting of current-laminated to massive light grey dololutite lithofacies with thin interbeds of siliciclastic shale lithofacies. Rucksack for scale.

B. Current-laminated dololutite cut by scour channels filled with intraclastic debris.

C. LLH-C stromatolites with low synoptic profile, draped by an intraclastic bed which fills inter-dome lows.

D. Large rip-up clasts of dololutite incorporated in well-sorted and rounded intraclast grainstone. Dololutite lenses in upper portion of bed define flaser bedding.

E. Photomicrograph of rounded intraclast grainstone. Laminated grains probably are pieces of eroded algal mat. Width of field is 5 mm. Plane-polarized light.

F. Photomicrograph of silicified ooid grainstone. Cortices of well-preserved ooids show dense multiple lamination. Clear rim cement is silicified dolomite rhomb cement (see Plate 20C for close-up) which is succeeded by spherulitic chalcedony. Width of field is 4 mm. Plane-polarized light.

PLATE 4

PLATE 4.

A

B

C

D

E

F

PLATE 5. KENDALL RIVER FORMATION

- A. Palisade algal mat texture in core of LLH-C dome showing typical vertical growth fabric. Width of photo is 30 cm.
- B. Polished slab of silicified chicken-wire algal mat texture showing deflection of algal laminae around nodules. Lower member B.
- C. Biostrome cycle (1.5 m thick) in upper member B near Bebensee Lake. Lowermost flat-bedded dololite (D) passes upward into ooid packstone (O), which is scoured (contact dashed), and capped by LLH-C biostrome. Notebook is 17 cm long.
- D. Oncolite packstone of upper member B. Width of field is 25 cm.

PLATE 5

PLATE 5

CHAPTER 4. SULKY FORMATION

The Sulky Formation is the main cliff-forming unit of the Dismal Lakes Group (Figure 19). The name Sulky Formation is informally proposed for the platy to massive dolostone exposures along the north shore of Sulky Lake. The Sulky Formation, also well exposed in the Greenhorn Lakes area, is 50 to 80 m thick in the east on the September Lake High, and 100-300 m on Dismal Lakes Platform. The unit thickens gradually westward, as do all formations of the Dismal Lakes Group. The top of a laterally persistent biostratotype (Kendall River Formation, member C) everywhere marks the base of the Sulky Formation (Figure 20). The upper contact of the Sulky Formation is less well defined. In the east, a stratiform chert breccia bed (sections 15-18, Figure 19) and associated silcrete (section 11, Figure 19) are overlain by dark grey coarsely recrystallized dololutes of the Greenhorn Lakes Formation. In the west (sections 1-13, Figure 19), a distinctive conical stromatolite bed is overlain by dark grey dololutes and clastic carbonates of the Greenhorn Lakes Formation.

The Sulky Formation is divided into three lithologically distinct members (Figure 20). Significant lateral lithofacies changes occur within each of the members from the September Lake High (sections 15-20, Figure 20) onto the surrounding Dismal Lakes Platform

Figure 19. Sulky Formation distribution showing location of measured sections discussed in text and apex of September Lake High (SL, Sulky Lake; GL, Greenhorn Lakes; DeL, Dease Lake; BL, Bebessee Lake; DL, Dismal Lakes; Spl, September Lake).

Figure 20. Three-fold subdivision of Sulky Formation showing generalized facies subdivision within members A and B. Member C is schematic due to the small scale of facies interfingering. September Lake High, section 1 - 5; Transition zone, sections 6,7; Platform, sections 8 - 20. Symbols as in Figure 6.

(sections 1-14, Figure 20; see also Figures 21, 24 and 29). Each member is described and interpreted by comparing sections from the High and from the surrounding Platform.

MEMBER A

Member A consists of platy bedded dololutes with subordinate siliciclastic shale interbeds. The lower contact of member A is equivalent to the basal Sulky Formation contact described above. The upper contact is consistently sharp, commonly definable as a bedding surface, where light grey-to-buff massive stromatolitic dolostones of member B rest on recessive-weathering dark grey to grey-green platy dolostones of member A. Member A thickens westward from 30 to 140 m (Figure 21). Five distinctive lithofacies are distinguished (A1 - A5), only one of which occurs on both the High and the Platform (Figure 21). In contrast to other lithofacies described in this thesis, these lithofacies occupy a definite stratigraphic position.

A1 Lithofacies:

This lithofacies is 10-20 m thick on the September Lake High, pinching out west of section 16 (Figure 21). It rests directly on the Kendall River Formation and is gradationally overlain, and interfingers with, the A3 lithofacies. It consists of 3-5 cm platy bedded dololute with red shale partings (Plate 6A). Common

sedimentary structures near the base of A1 are ripple marks and current lamination, SH-V columnar stromatolites (Plate 6B), associated stone rosettes (cf. Ricketts and Donaldson, 1979) and edgewise conglomerates. Wavy and parallel current laminations become more abundant upward, and small LLH-C stromatolites occur locally (sections 14 and 15).

A2 Lithofacies:

The A2 lithofacies, consisting of black shale with subordinate grey dololomite interbeds (Plate 6C), is up to 22 m thick west of Limestone Point (section 1) and pinches out east of central Dismal Lakes (section 10, Figure 21). The lower contact of this lithofacies with the Kendall River Formation is sharp, with shales resting directly upon the member C (Kendall River Formation) biostrome. This lithofacies grades upward into the A3 lithofacies over 5 m.

A3 Lithofacies:

The A3 lithofacies occurs on both the Platform (0-40 m) and the September Lake High (12 - 26 m, Figure 21). On the High the transition from A1 to A3 is marked by a change from red to green coloration and an absence of sedimentary structures except for parallel current lamination. Westward, on the Platform, black pyritiferous shales of A2 grade upward into A3 green lutites which contain black shale partings throughout.

A4 Lithofacies:

Lithofacies A4, comprising rhythmically interbedded light grey and dark grey dololomite (Plate 6D) is the most abundant component of member A on the Platform. In gradational contact for a few metres with A2, this lithofacies thickens westward from 0 m (section 14) to at least 115 m along the north shore of Dease Arm (section 1, Figure 21). Light grey beds (3-8 cm) increase upward in abundance from 50-70 percent, relative to dark grey beds, and are generally more abundant in eastern sections (eg. section 12). Current lamination, normal grading, climbing ripples and small (20-60 cm) debris flow units are present locally in the light grey beds. These beds are internally laminated on a millimetre scale, with alternating light grey dololomite and black (organic rich?) recrystallized dololomite. Where not disrupted by stylolitization, individual mm-thick laminae can be traced for tens of centimetres. However, pervasive recrystallization to a pseudospar mosaic has obscured these features in most exposures.

A variety of early post-depositional to penecontemporaneous deformation structures characterize the A4 lithofacies. Pinch and swell bedding (Plate 6D) related to differential cementation, compaction and microscale stylolitization (non-sutured seam solution of Wanless, 1979), centimetre-scale calcite-filled en echelon fracture systems (largely restricted to light

grey beds), rare small-scale slump folds with a consistent southwestward sense of overturning (Figure 22) and small-scale (1-2 m) intraformational truncation surfaces (cf. Wilson, 1969; Plate 6E) are all present. Differential competence of light grey (ductile) and dark grey (brittle) beds is suggested by differing response to penecontemporaneous slumping (Plate 6F), suggesting that early differential cementation has preferentially lithified the dark grey layers.

A5 Lithofacies:

Small (1-2 m by 4-25 m) lenses of massive light grey algal-laminated carbonate (A5 lithofacies) occur near the contact between members A and B on the north shore of Dismal Lakes (sections 12 and 13). These lenses represent minor interfingering of member B algal carbonates with underlying member A sediments. Algal laminae within the lenses, consisting of continuous mats of linked concave-up cusps, probably represent incipient curly bedding (see discussion below).

Lithofacies Interpretation: Member A

All Member A lithofacies are mud-dominated, thin-bedded units displaying either parallel current-lamination or a total absence of sedimentary structures. These characteristics and features such as black pyritiferous shale (A2), graded bedding, mm-scale lamination, synsedimentary slump folds, intraformational

Figure 22. Plot of trends of fold axes of subhorizontal small-scale (10-30 cm amplitude) synsedimentary slump folds from facies A4 (section 18).

truncation surfaces and small-scale debris flow beds (A4) compare closely with features of other sequences interpreted as deeper water, sub-wave base accumulations (Wilson, 1969; Smith, 1977; Pfeil and Read, 1980; Cook and Taylor, 1977). Light grey beds are interpreted as episodic current-deposited muds, in part emplaced by turbidity currents. The dark grey organic-rich beds are interpreted as "background" sediments. Deposition of member A was initiated upon the extensive peritidal platform of the Kendall River Formation. The few shallow-water features at the base of the A1 lithofacies of the September Lake High record initial platform flooding. Flooding of the rest of the platform (sections 1-13) occurred rapidly, allowing shales, rather than a basal transgressive lag, to be deposited.

The lithofacies contrast between the September Lake High and the surrounding Platform (Figure 21) represents the first significant period of differential subsidence of the previously homogeneous Dismal Lakes platform, with the September Lake High subsiding at a lesser rate than the surrounding Platform.

The September Lake High-Dismal Lakes Platform transition, though marked by significant lithofacies changes, does not represent a shelf-to-basin transition such as the Tansill-Capitan-Lamar (Tyrell, 1969) or El Abra-Tamabara-Cuesta del Cura (Enos, 1974) lithofacies transitions. The absence of large-scale slump folds and

breccias (cf. Cook and Taylor, 1977) and major intraformational truncation surfaces (cf. Davies, 1977) suggests that member A more closely resembles open shelf accumulations (cf. Wood Canyon Member of Lodgepole Formation, Smith, 1977; Patterson Member of Shady Dolomite, Pfeil and Read, 1979).

The generalized paleogeography interpreted along the line of measured sections is illustrated in Figure 23.

MEMBER B

The unique stromatolite assemblage of the middle member of the Sulky Formation is the most variable and striking feature of the Dismal Lakes Platform. This well-exposed buff to light grey massive unit consisting almost entirely of stromatolitic dolostone, is an outstanding example of a Proterozoic carbonate platform.

Member B is 25-30 m thick on the September Lake High and 10-40 metres thick on the Platform. The lower contact with Member A is sharp, as previously described. The upper contact is more variable, being defined by the first appearance of medium-bedded (0.5-1.5 m) dolostones above the last massive algal bed. Lithofacies will be contrasted between the September Lake High and the Platform.

September Lake High:

On the September Lake High, a distinctive assemblage of stromatolite types occurs in a regular upward

Figure 23. Simplified paleogeographic cross section showing development of differential topography between September Lake High and western platform during member A deposition. Symbols for facies subdivisions as in Figure 21.

succession (Figure 25). The buildup attains a thickness of up to 30 m, with lamination inheritance continuous throughout, regardless of transitions between different stromatolite geometries. Less than 5 percent of the buildup consists of non-stromatolitic material (dololutite and fibrous cement).

The initial stromatolitic growth form which colonized member A dololutites was the branching conical-columnar stromatolite (jacutophyton: uncapitalized descriptive names are applied here, in accordance with the arguments presented in Donaldson, 1976). Growth initiates from a point, with branches diverging low in the vertical development, giving way to vertical orientations within a few metres of initiation (Plate 7A), which persist upward for an average of 7 metres. A plan view (Plate 7B) approximately 1.5 metres from the base of a jacutophyton bush shows greater asymmetry of the outer columns, reflecting the highly inclined growth component of these branches. Column diameters range from 5-15 cm, and the spacing between columns is 5-15 cm; distances between jacutophyton bushes is 3 to 5 metres. Inter-column and inter-bush fill is thinly bedded dololutite (Plate 7A). Synoptic relief of the jacutophyton clusters may have been as much as the full 7 metres during growth. A minimum relief of 10 cm is indicated by growth-increment outlines of individual columns.

PT

Figure 25. Idealized vertical sequence of stromatolite types of the September Lake buildup: 1, branching conical-columnar (jacutophyton); 2, spaced conical-columnar (conophyton); 3, laterally linked conical columnar; 4, laterally linked domal; 4a, laterally linked conical-columnar stromatolites on 4; 5, branching digitate columns.

Vertical growth of spaced conical columns, the second growth stage (Figure 25) is a continuation of the jacutopyhton stage and typically extends for approximately 4 metres. Column diameters increase to an average of 20 cm, with 20-30 cm intercolumnar spacing. Approximately 7m above the base, the conical-columnar branches spread to form an evenly spaced biostrome, the clustering effect of underlying jacutophyton bushes being lost. Intercolumnar fill consists of massive dololomite and fringing growths of fibrous submarine cement. Synoptic relief could not be accurately established, but was at least 40 cm. At 11 m (Figure 25), conical columns form a zone of tightly packed laterally linked conical stromatolites (Plate 7C) which extends for another 2-15 metres. Column diameters increase from 0.4 to 6 m through coalescence of individual columns. Synoptic relief ranges from 0.5-12 m (Plate 7D). Broad domal stromatolites (Plate 7E) overlie the conical sequence (Figure 25). The conical to domal transition is completely gradational, and again the trend is toward increasing diameter of individual structures by continued coalescence. Large domes develop maximum diameters of 40 m and have a synoptic relief of up to 15 m. Rare interdome spaces are filled with dololomite and fibrous cement. Smaller columns modify surfaces of the large domal structures.

The columnar stromatolites growing upon the large

domes have a distinctive polygonal outline, and the width of intercolumn spaces is a remarkably consistent 1 cm (Plate 7F). Intercolumn fill is dololutite with rare (less than one percent) fragments of algal mats. Column diameters average 10-15 cm, and decrease steadily upward from 60 cm to 5 cm. At the top of the sequence, the large domal surfaces upon which the small columns grow have almost no synoptic relief.

These small polygonal-packed columns are the final growth stage of the member B buildup on the September Lake High and are conformably succeeded by 10 metres of fenestral and pisolitic dolostone (member C).

The idealized vertical sequence (Figure 25) persists throughout the September Lake High, with only minor variations in stage thickness. Nowhere is growth interrupted by clastic sedimentation.

Transition Zone:

Exposed in the Coppermine River Valley (section 15, Figure 24) and the south flank of Teshierpi Mountain (section 14, Figure 24) are thin (10-15 m) sections of member B which display growth forms transitional between conical-columnar stromatolites of the September Lake High and the more chaotic "curly" bedding typical of the platform (Figure 26). These transitional structures are continuous cusped algal sheets with conical apices, creating a "circus tent" appearance. Structures range

Figure 26. Stromatolite morphologies transitional between conical columnar stromatolites of the September Lake High(D) and curly bedding of the platform(A). "B" (from section 14) shows more densely laminated apical zones reminiscent of conical-columnar stromatolites, with apical zones separated by continuous algal drapes. "C" (section 15) shows early development of curly bedding before grading into conical structures upsection. Note irregular apophyses in "D"(arrows) which appear analogous to curly bedding structure (sketches not to scale).

from a few centimetres to 4 m in wavelength (Figure 26C and D; Plate 8A). Some larger structures near Teshierpi Mountain show synoptic relief up to 8 m. Gaps between algal laminae in lows between apical peaks are lined by fibrous cement (Plate 8B) similar to that which fills intercolumnar spaces in the September Lake buildup. Locally, laminae are not continuous across apices, producing algal structures equivalent to the discrete concave-up algal plates or curly bedding discussed below (Figure 26A; Plate 8A). Other growth forms display a concentration of laminae in the apical zone, and sparse laminae connecting these zones, producing a structure equivalent to linked conical-columnar stromatolites (Figure 26B).

Platform:

The platform west of Teshierpi Mountain is dominated by a unique and highly variable stromatolitic association. Thickness of member B on the platform is also variable, ranging from 10-40 m, with individual units rarely traceable laterally for more than 2 km. The lower contact is typically sharp, but the upper contact is gradational. As on the high, the first exposure of well-laminated dolostone above the uppermost massive stromatolitic zone (here displaying curly bedding) marks the transition to member C. Above this contact, stromatolites are well-formed conical-columnar

types. Member B on the platform consists of a massive stromatolitic lithofacies (B1) and thin-bedded dololutite (B2).

B1 Lithofacies

The B1 lithofacies is massive to unbedded dolostone composed of concave-up discrete stromatolitic plates of 0.1-2.0 m diameter (curly bedding of Kerans and Donaldson, 1979; Plate 8C) and derived breccias (Plate 8D). Stromatolitic plates and breccias are either set in a dololutite matrix, or maintain an open framework which is filled by fibrous and equant spar cements (Plate 8B). An algal origin for the curly bedding structure can be demonstrated through similarities in lamination microstructure and intergrowth of curly bedding with conical-columnar stromatolites (Kerans and Donaldson, 1979). Such a growth continuum is best shown in the transition zone previously described (Figure 26).

Breccias derived from fragmentation of the curly bedded units are an essential component of this stromatolite association. The breccias typically consist of angular to subangular, subequant algal-laminated clasts ranging from 0.5 to 5.0 cm set in a dololutite matrix (Plate 8D). Breccias are largely restricted to the base of stromatolitic beds. They are poorly sorted and non-graded and display both intact and disrupted

frameworks. The breccia zones have sharp basal contacts and grade upward into undisrupted curly bedding, are extremely discontinuous laterally, and involve only curly bedding which was previously uncemented. Fragments of cemented stromatolitic material were not observed.

B2 Lithofacies

The B2 lithofacies consists of massive to vaguely parallel-laminated dololomite in 10-60 cm beds between sequences of massive curly bedding. These micrite beds are flat-lying, continuous, and of even thickness (Plate 8F).

Alternation of B1 and B2 lithofacies is irregular, and occurs on both regional and outcrop scales (Figure 24).

Lithofacies Interpretation: Member B

Stromatolitic growth which dominated member B deposition began on a quiet-water platform below wave base. Differential subsidence during deposition of Member A produced a local topographic high (September Lake High) surrounded by a west-sloping slightly deeper-water platform. The September Lake High dramatically influenced member B stromatolitic growth, as indicated by the contrasting stromatolite types which inhabited the high and the platform (Figures 24 and 28).

The total dominance of mud as intercolumnar fill, the absence of spur and groove channels or stromatolite

asymmetry and consistent lamination-inheritance (cf. Eriksson, 1977; Ricketts, 1979) all point to a sub-wave base origin for member B. In this sense the September Lake buildup is similar to the marginal-mound setting proposed by Dunham (1972) for the Capitan Reef. It is different from most Proterozoic carbonate shelf sequences which typically exhibit marked mound elongation (Hoffman, 1967, 1969; Young and Long, 1975; Eriksson, 1977; Cecile and Campbell, 1978; Ricketts, 1979; Tucker, 1977; see Table II). Similar low-energy profiles are the Hadrynian Noonday Dolomite (Wright et al., 1978) and the Aphebian Utsingi Formation (Hoffman, 1974).

On the platform, local accumulations of curly bedding (CB) stromatolites surrounded by flat-bedded dololutite created a patchy biohermal topography. The synoptic relief of these bioherms of curly bedding was negligible, as demonstrated by the flat-lying dololutite beds which overlie them.

The sharp basal contacts of breccia zones associated with CB accumulations argues against origin by solution collapse. In addition, solution breccias commonly have allochthonous matrix material (transported by dissolving solutions), and clasts within them are rounded as a result of solution (Middleton, 1961). The mud-supported fabric, lack of sorting and upward transition into undisrupted curly bedding (Plate 8E) negates an allochthonous current-transported origin for the breccias.

These synsedimentary breccias may have been produced by collapse due to loading. Where synsedimentary or penecontemporaneous cements filled spaces between algal plates (cf. Plate 8B), sufficient support would be provided to prevent collapse of the loose curly bedding framework (Figure 27a). On the other hand, loading by upward growth in the absence of synsedimentary cementation eventually would lead to an unstable configuration of incompetent algal plates, and collapse of the most-loaded lower portion, resulting in brecciation (Figure 27b).

It follows from this hypothesis that the breccias are indicators of the patchy distribution and early submarine origin of the fibrous cements.

The minimum topographic expression of the September Lake buildup relative to the platform is interpreted as about 15 m, based on the difference in synoptic relief between the contrasting stromatolite forms (0 for curly bedding; 15 m for domal phase of September Lake buildup). In addition to this difference, the unknown amount of relief inherited at the end of Member A deposition (Figure 23) must be considered. Thus a cumulative relief difference of 20-30 m does not seem unreasonable. The high-to-platform transition consists of a gentle fore-buildup slope (Transition Zone) rather than a steep escarpment. The interpreted paleogeography and distribution of stromatolites is shown in Figure 28.

Figure 27. Schematic illustration of synsedimentary collapse origin for curly bedding. See text for discussion.

Figure 28. Paleoenvironmental interpretation of member B showing maximum development of September Lake buildup in east on a topographic high (see Figure 24). The slightly deeper platform to the west passes gradually into the buildup zone.

MEMBER C

Member C consists of a sequence of medium- to thick-bedded cream- to dark-grey stromatolitic and non-stromatolitic dolostones. Member C thickness ranges from 10 m on the buildup crest to 100 m farther west on the platform (Figure 29). Again, lithofacies on the high differ significantly from those on the platform, and will be discussed separately.

September Lake High:

Bedded pisolite units, pisolite- and tepee-bearing fenestral dolostones, enterolithically folded dololutites, stratiform chert breccia and silcrete are the five lithofacies recognized within member C on the September Lake High.

Pisolite Lithofacies

Individual pisolites are laminated on a mm-scale (Plate 9A), and are much better ordered than typical caliche pisolites. Laminae are now composed of micrite, and bladed neomorphic dolomite pseudospar with c-axes oriented perpendicular to the pisolite laminae (Plate 9A). Nuclei are peloids, broken or compound pisolites and pisolite intraclasts. Many pisolites display circumgranular fractures (Plate 9A). These pisolites are similar to those described from the Yates Formation of the Permian Capitan complex (Dunham, 1965, 1968; Esteban and Pray, 1977; Scholle and Kinsman, 1974).

Two distinct pisolite associations are recognized: a bedded association consisting of 20-60 cm layers intercalated with mm-scale parallel-laminated dololomite, and a pisolite-tepee-fenestral dololomite association consisting of 0-60 cm lenses in dish-shaped depressions formed by tepees.

Bedded pisolite units (Plate 9B) are both normally and reversely graded, unstratified to crudely stratified, and moderately to poorly sorted. Polygonal packing and coalescence of individual pisolites occur only in the upper few centimetres of some units. Some pisolites display consistent downward elongation. The fabric is grain supported and matrix-free. Meniscus cements are not preserved; perching of sand-sized pisolites upon larger grains is common (Plate 9A).

The bedded association has characteristics of both the transported and in-place pisolite associations recognized by Esteban and Pray (1977). Reverse grading, polygonal packing and grain perching all indicate an *in situ* development phase. Moderate size sorting, absence of fines and a grain-supported framework point to an initial mobile or clastic stage of formation.

The second association, the pisolite-tepee-fenestral dolostone association displays well-developed pisolites (Plate 9C) within depressions up to 1 metre deep between tepee ridges. However, individual pisolites are significantly larger (to 12 cm). They commonly display

polygonal packing, downward elongation and apparent matrix support. The pisoliths are set in and/or interfinger with fenestral-peloidal dolostones. Grain perching is common, and geopetal cavity filling also is present.

The extremely poor size sorting, polygonal packing, downward elongation and the presence of fenestral carbonte textures indicate a strong vadose overprint for these pisolites (c.f. Dunham, 1969). The absence of evidence for host-rock weathering as an agent for production of these pisolites, and their marine-marginal association, suggest that they are primary accretionary grains formed in a marine-marginal environment. This interpretation is consistent with the prevailing interpretation for the Capitan pisolites (Scholle and Kinsman, 1974; Esteban and Pray, 1977).

• Fenestral-Tepee Dolostone Lithofacies

Fenestral to massive dolostone commonly disrupted by tepee ridges (cf. Smith, 1974; Assereto and Kendall, 1977) composes the bulk of Member C on the September Lake High. Tepee ridges are largely in the embryonic stage (Assereto and Kendall, 1977), with a few in the mature stage, having ridges affecting up to 2 m of section, and forming individual polygons up to 4-6 m in diameter. Non-tectonic fractures (Dunham, 1969), oriented both parallel and oblique to bedding, also are

associated with this lithofacies. Cellular crustal cements (Plate 17D; Assereto and Kendall, 1977) and botryoidal pisolitic texture (Plate 17C; Esteban and Pray, 1977) encrust host rocks of the tepee structures. Well-developed pisolite sequences occur preferentially within depressions between tepee ridges, much like those described from the shelf-crest lithofacies of the Permian Capitan Reef (Esteban and Pray, 1977). Massive aragonite fans associated with tepee structures, such as those attributed to submarine cementation (Loucks and Folk, 1976; Cys and Mazzulo, 1977; Assereto and Folk, 1980) were not observed in the Sulky Formation.

Enterolithically Folded Dololutite Lithofacies

Enterolithically folded and fractured dololutites are a third distinctive lithofacies on the September Lake High (Plate 9D). They are restricted to a 2 m-thick unit approximately 80 cm above the Member B-C contact at section 19. These chaotically contorted cream-coloured lithographic (5-10 u) dololutites have undergone a complex diagenetic history which involved expansive growth and dissolution of evaporite minerals. Expansive evaporite mineral growth is probably responsible for spreading apart of dololutite layers (Plate 9E), with folding occurring during post-solution collapse. Similar expansive growth of evaporites, resulting in the disruption of bioclasts, has been reported by Chowns and

Elkins (1974).

Although no erosional truncation of the folded dololutes is apparent (cf. Ricketts, 1979), a penecontemporaneous or early diagenetic origin for folding is supported by the stratiform nature of the occurrence (gradational above and below into fenestral dolostone) and the post-deformational agate cements which fill the abundant pore spaces. These agate cements are apparently related to a silcrete approximately 1 m above the enterolithically folded unit, indicating that the folds and solution-produced porosity must have developed before silcrete formation.

Silcrete

A stratiform chert unit 15 cm thick occurs near the crest of the September Lake buildup. It is composed of essentially pure chert in the upper 10 cm, grading downward into a mixed chert-carbonate rock with swirly texture, and finally into unweathered dolostone. Chert pebbles derived from the silicified evaporite breccia (see below) form a lag on the silcrete surface; to the east, the silicified evaporite breccia marks the top of member C (Figure 29). These reworked chert clasts indicate an early near-surface silicification of the evaporite unit discussed below.

Stratiform Chert Breccia

A 0.6-1.5 m thick stratiform chert breccia occurs

throughout the September Lake region (sections 16-18, Figure 29; Plate 9E) for a distance of 20 km. The lower contact is irregular and disrupts up to 1.5 m of the underlying dolostone. Angular clasts (2-30 cm) within the breccia are composed of white and orange-red chert and dolostone. Locally (section 16), the top of the breccia is reworked and a lag deposit of chert clasts rests on its upper surface.

That this chert-carbonate breccia formed by dissolution and partial replacement of evaporites is supported by:

- relic felted-lath texture preserved in chert (Plate 9F)
- small (20-50 μ) inclusions of anhydrite trapped within secondary quartz grains
- abundant length-slow chalcedony.

The felted-lath texture is similar in appearance to that illustrated by Chowns and Elkins (1974), and interpreted by them to represent silicified anhydrite.

Platform:

Member C is poorly exposed in sections 14 and 15 of the Transition Zone, but excellent partial sections are exposed on the Platform, particularly in the Bebensee, Greenhorn and Sulky Lake areas (sections 4-9, Figure 29). The member C Platform lithofacies are significantly different from those on the September Lake High. The

most prominent lithofacies consists of laterally persistent biostromes. Two distinctive biostrome types are domal-digitate stromatolites, and tightly packed conical-columnar stromatolites (0.6-4 m thick), which typically occur in the progression: flat mat to conical-columnar to branching conical-columnar (Plate 10A, 10B). Other lithofacies include thin ooid beds (10-60 cm; Plate 10C) and sooty grey dololomite with small-scale tepees, palisade-type algal lamination (c.f. Hardie and Ginsburg, 1977; Plate 10D) and small domal stromatolites.

Both conical and domal stromatolite biostromes are composed of individual stromatolites which display consistent asymmetry (Figure 30). The most common interpretation of stromatolite asymmetry is that prevailing currents influenced growth. This has been well documented in Recent sediments (Logan et al., 1974) as well as from numerous ancient examples (Hoffman, 1967, 1969; Young, 1974; Cecile and Campbell, 1978; Ricketts, 1979; Eriksson, 1977; Tucker, 1977). With one reported exception (Young and Long, 1975), stromatolite elongation is perpendicular to strandline, leading to an interpretation of shaping by tidal currents.

Elongations of member C conical and domal biostromes are all perpendicular (Figure 30). Elongation trends retain a near-perpendicular relationship even when the two biostrome types are interbedded on a 1-2 m scale.

Figure 30. Rose diagrams of linear trends from elongate stromatolites in biostromes of member C of the western platform (sections west of 11).

(Figure 31). These differences in stromatolite elongation trends may result either from opposed currents or abnormal variations in stromatolite growth morphology (cf. Cecile and Campbell, 1978).

From the geometry of the basin through time, and the more prominent elongation of the conical stromatolite biostromes, it is most likely that the conical trend (148°) represents the true paleoslope; the domal trend having been formed either by longshore currents (cf. Young and Long, 1975) or some independent growth phenomenon (cf. Cecile and Campbell, 1978).

Interpretation: Member C

Member C sediments accumulated during shoaling of the Dismal Lakes Platform from a sub-wave base submerged position (late member A-member B). The association of pisolite sequences (both in situ and reworked varieties), fenestral carbonates, tepees and associated non-tectonic fractures, enterolithically folded dololutes, exposure surfaces and silicified evaporites all point to subaerial and restricted hypersaline peritidal depositional environments on the September Lake High during deposition of member C. Pisolitic lithofacies north of September Lake (section 17) occur only 1.5 m above the final growth phase of the buildup, implying that shoaling occurred relatively rapidly. A buildup crest-shoal environment with local islands and hypersaline ponds is envisioned

Figure 31. Typical section of member C on western platform showing opposed elongation directions of interbedded conical and domal stromatolite biostromes. Numbers of measurements shown in circles. Directions are averaged means for each station.

(Figure 32).

The previously deeper-water platform (Figure 28) shoaled more gradually and accumulated a greater thickness (100 m as opposed to 10 m on the high). Ooid units and stromatolite biostromes are interpreted to be subtidal, whereas fenestral-cryptalgal sediments probably formed in intertidal and supratidal environments. Common alternation of biostromal, ooidic and fenestral-algal mat lithofacies indicates more active pulses of subsidence on the Platform, probably associated with progradational shallowing-upward depositional cycles (cf. James, 1979; Hoffman, 1976). Greater rates of subsidence typify the area west of Leith Line throughout the time of Dismal Lakes Group sedimentation. Apparent complexities in observed lithofacies cyclicity (Figure 31) may be related to nonpreservation or to rapid lateral lithofacies migration during progradation.

DISCUSSION

The preceding analysis of the Sulky Formation has rested almost entirely on analogy with well-documented Phanerozoic and Recent carbonate examples. Several well-exposed Proterozoic carbonate sequences have now been analyzed by the modern comparative lithofacies approach discussed by Ginsburg (1974) and it seems useful to compare the Sulky Platform to some of these other Proterozoic examples.

Figure 32. Paleoenvironmental interpretation of member C showing contrasting facies response of High and platform during shoaling of the region. Symbols as in Figure 29.

First, a classification scheme is needed to provide a basis for comparison. Wilson's (1974) classification (Table I) is used in this discussion rather than the schemes recently proposed by James and Mountjoy (1981) and Read (1981) because of the emphasis placed on the geomorphology of the shelf-basin transition. This transition is of particular interest for two reasons. It is generally the best reflection of the nature of the organisms in the associated buildup. It also is consistently better exposed than fore- and back-buildup lithofacies, and thus provides the best basis for comparison of the largest number of examples. One important addition to Wilson's classification scheme is the recognition of drowned platforms (Schlager, 1981). Any of Wilson's platform-margin types can be subjected to rapid submergence, and therefore the numerals are merely modified to indicate this condition by addition of a subscript (eg. Id).

Several associations are apparent from the tabulations in Table II. The Sulky, Utsingi, Noonday and possibly the Malmani Dolomite are seen to be examples of Type I platform margins. All are mud-dominated low-energy associations mostly lacking distinctive stromatolite elongation and channelization (Utsingi and Malmani mounds display subtle elongation perpendicular to the strandline). Platform-to-basin transitions for this group are both sharp (Noonday Dolomite: Wright et al.,

TABLE I. CARBONATE PLATFORM MARGIN CLASSIFICATION (after Wilson, 1974)

TYPE	POSITION ON PROFILE	RELATION TO WAVE BASE	FRAMEWORK, COMPOSITION, CHARACTER	ASSOCIATED FACIES
I	Downslope	Below	Bioclastic Lime mud $\pm$ organic boundstone	Profile-crest sand shoals
II	Outer shelf edge of ramp	Within, but moderate	Interreef sands greater than boundstone, mud accumulation possible	Profile-crest sand shoals or mud flats, foreslope sands
III	Topographic crest of profile	Within wave-break zone	Largely boundstone	Protected backreef lagoon, fore- slope breccias and slumps

TABLE II

EXAMPLE	AGE (BY)	STROMATOLITE GEOMETRY				INTER-STROMATOLITE FILL		FORE-BUILDUP FACIES		BACK-BUILDUP FACIES		TECTONIC SETTING	
		Size	Relief	First-Order Symmetry	Second-Order Symmetry	Mud	Sand	Coarset	Turbidite	Slump/ Talus	Starved Shale	Open Margin	Intra-Clastic
		Maximal Diameter	Synoptic										
Helmet	2.6	20m	5m	2:1-4:1	N.A.	x	rare					x	x
Delimita S. Africa	2.6	20m diameter											
Utsingf Formation	1.87	2x20m	<50m	up to 10:1	N.A.	x					x	x	10
Moonday Delimita	1.0	300x75m	<200	1	1	x					x	x	10
Salty Formation	1.2	20x40m	15m	1	1	x						x	10
Navor FN	1.05	100x60m	50m	20:1	5:1	x	rare					x	11
Kuonik FN	1.05	10x50m	50m	5:1	20:1	x						x	11
Party Bay FN	1.2		<2m	3:1	10:1	x					N.A.	x	11
Portuguese Delimita	0.7	1x0.6	1m	1	1	x	x?				N.A.	x	11
Unit 9 (R. Bay)	0.5	5m	3m	1	1	x					N.A.	x	11
Clareng Formation	0.7		10m	1:6	1:10	x					x	x	11
Talhellet Formation	1.87	27m x 60m	2m		6:1	x					x	x	11

B = Drained

N.A. = Not Applicable

1978) or completely gradational (Utsingi Formation: Hoffman, 1974). Although a buildup-to-open shelf transition is exposed on the Sulky Platform, no clear platform-to-basin transition has been recognized. Possibly a gradual or "ramp" style transition (Ahr, 1973) occurs westward, but has been unrecognized due to the subtle nature of the transition. According to Wilson (1969), gradual platform-to-basin transitions are characteristic of intracratonic basin settings such as that proposed for the Dismal Lakes Group.

According to Schlager (1981), rates of carbonate accumulation are at least one order of magnitude greater than long-term eustatic fluctuations. Following this argument, drowned platforms point to geologically anomalous events such as fault-related subsidence, glacio-eustatic fluctuations, or sudden deterioration of water-mass conditions leading to inhibition of carbonate production (Schlager, 1981).

Whereas both the Utsingi and Noonday platform margins were formed in tectonically active environments (margins of Athapuscow and Amargosa aulacogens respectively), the intracratonic setting of the Sulky Formation argues against a tectonic origin alone as an explanation for platform drowning. No evidence exists for water-mass deterioration, and no record of glaciation is known for this time period. Though relatively stable, a change in the pattern of regional subsidence

(differentiation of the September Lake High) does correspond closely to initiation of the transgression which led to drowning of the previously peritidal platform. A combination of increased differential subsidence and significant eustatic rise appears to be the most likely explanation for drowning of the Platform. A synchronous transgressive event in Elu Basin, 300 km to the east (Campbell, 1979; Kerans et al., 1981) demonstrates that the event may have been at least in part eustatic.

Critical to Schlager's (1981) argument that extraordinary short-term geologic processes are required to cause platform drowning, is the assumption that the rate of carbonate accumulation in the geologic past be equivalent to present rates (within an order of magnitude), keeping it outside the range of eustatic or subsidence-related processes. The fact that only one reported case of a Type III platform margin is noted in Table II does not lend support to this contention. Type III margins are characteristically steep rimmed, and result from reef systems outpacing sedimentation on forereef and backreef lithofacies belts as the reef keeps pace with rapid eustatic rise. Such rapid upward growth is characteristic of Mesozoic and Cenozoic hexacoral reefs, but only rarely has been recorded for Paleozoic reefs (Wilson, 1974). Two Paleozoic examples are the Devonian reefs of the Canning Basin, Australia (Playford,

1980), and the Cathedral Formation (Middle Cambrian) of Alberta (MacElreath, 1979).

The Taltheilei Formation, and the time-equivalent Kimerot Platform (Great Slave Lake and Bathurst Inlet areas respectively: Hoffman, 1973; Campbell and Cecile, 1981) appear to be the only Proterozoic examples of Type III margins to date. The rarity of Type III margins in stromatolitic Proterozoic carbonate shelf margins, and the absence of all but the blue-green algal components of the "carbonate machine" during the Early and middle Proterozoic, suggests that carbonate accumulation on platforms may have been significantly slower during these times. Slower accumulation rates would allow drowning by normal eustatic fluctuations, rendering drowned platforms less "anomalous" features.

A final comparison from Table II concerns the paleoecology of conical stromatolites. Branching-conical (jacutophyton) and conical (conophyton) stromatolites occur in the deeper-water lithofacies of both the Sulky Formation and the Utsingi Formation (as well as the basinal equivalent Blanchet Formation: Hoffman, 1974). This subtidal setting appears to be characteristic of these stromatolites (Hoffman, 1976; Donaldson, 1976) and they may be a common feature of drowned Proterozoic platforms.

PLATE 6. SULKY FORMATION

A. Thin (3-5 cm) platy bedded, current-laminated dololutite with red siliciclastic shale partings and small LLH stromatolites. Lithofacies A1 (section 15).

B. SH-V club-shaped stromatolites at base of member A (section 18). Lens cap is 60 cm in diameter.

C. Black shale with tan dololutite interbeds, lithofacies A2 (section 9).

D. Rhythmically bedded light-grey current-laminated dololutite beds and dark-grey organic-rich beds of lithofacies A4. Pinch-and-swell bedding is caused by differential cementation and compaction (section 2).

E. Small-scale intraformational truncation structure in lithofacies A4 (section 1). Pencil at centre-left is 17 cm long.

F. Early disruption of lithofacies A4 showing light-grey dololutite which, while unlithified, intruded brittle-deformed dark-grey beds. Several vermiform vertical injections contain a later generation of spar (section 1).

PLATE 6

PLATE 6

A

B

C

D

E

F

PLATE 7. SULKY FORMATION

A. Cross-section of branching conical-columnar (jacutophyton type) stromatolites initiated on, and surrounded by, platy dololutites of uppermost member A (A3 lithofacies). Branches weather darker due to differential silicification (growth stage 1 of Figure 25; section 15).

B. Plan view of branching conical-columnar (jacutophyton type) type stromatolite bush composed of individual 10-15 cm columns (grey) surrounded by dololutite (light grey). Cross-section is approximately 1.5 m above the member A-B contact. Note the greater asymmetry of marginal columns (growth stage 1 of Figure 25; section 15). Pencil is 17 cm long.

C. Tightly packed, laterally linked conical-columnar stromatolites with 60-80 cm of synoptic relief (growth stage 3 of Figure 25; section 15).

D. Large conical-columnar stromatolite (growth stage 3) illustrating 12 m of synoptic relief. Note mamillary surface texture modifying lower portions of the column (section 14). Man near centre provides scale.

E. Broad (25 m diameter) low-relief domal stromatolite (growth stage 4 of Figure 25). The inclinations of successive laminae gradually decrease until the synoptic relief is less than 2 m (where man is standing), at which point growth stage 5 (Figure 25) has been initiated (see F of the Plate). The cores of such domal structures are conical stromatolites. Photograph taken near section 17.

F. Culmination growth stage (stage 5 of Figure 25) of digitate columns on top of domal stromatolite shown in E of this Plate. Individual columns are 40-60 cm in diameter. Intercolumn spaces of uniform thickness are filled by dololutite. Photograph taken near section 17.

PLATE 7

PLATE 7

A

B

C

D

E

F

PLATE 8. SULKY FORMATION

A. Transitional morphology of curly bedding "dishes" in discrete stacks separated by vertical zones where laminae become subvertical and locally form conical inflections (see Figure 26C for comparison; base of member B, section 13. Hammer, bottom center, indicates scale.

B. Photomicrograph of laminae which form the transitional curly bedding structures shown in Figure A. Note isopachous fibrous cement lining a relatively open algal framework in troughs between inflection zones (section 13). Plane-polarized light. Width of field is 2.5 cm.

C. Typical curly bedding structure showing concave-up algal plates with low lamination inheritance in dololutite matrix (section 2). Compare with well-ordered stacks of plates in A of this Plate.

D. Angular algal-laminated clasts of curly-bed breccia in matrix of dololutite and late-diagenetic chert. Breccia is poorly sorted and non-graded, and has a sharp basal contact with thinly bedded dololutites of member A, approximately 60 cm below photograph (equivalent to lower portion of illustration at right of Figure 27B; section 11). Coin at bottom-center is 2 cm in diameter.

E. Chaotically arranged curly bedding, probably disrupted during formation of the breccia (D of this Plate) which underlies this outcrop by 5 m. This chaotic zone is equivalent to the upper portion of illustration at the right of Figure 27B (section 11).

F. Dololutite interbed of lithofacies B2 (20 cm thick). Lower contact is essentially horizontal, indicating little or no synoptic relief for the underlying curly bedded unit (massive bed at base of outcrop; section 7).

PLATE 8

PLATE 8

A

B

C

D

E

F

PLATE 9. SULKY FORMATION

A. Photomicrograph of pisolites of the bedded association showing irregular mm-scale lamination, radial recrystallization (R), and radial fractures (F) and perched grains (member C, section 14). Plane-polarized light. Width of field is 1 cm.

B. Outcrop of bedded pisolite association showing alternation of pisolitic and laminar zones. Cap of marking pen is 0.5 cm wide (section 14).

C. Large polygonally packed pisolites of tepee-fenestral-pisolite association. These pisolites occur within a tepee depression formed by the ridge illustrated in D of this Plate (member C, section 19). Width of field is 20 cm.

D. Enterolithically folded dololutite. Cream-white sediment is host rock. Grey dololutite and agate (clear) are later precipitates (member C, section 19).

E. Outcrop of stratiform chert breccia. Angular chert fragments float in recrystallized carbonate matrix. Chert clasts retain a felted lath texture after anhydrite (section 14). Notebook is 17 cm long.

F. Photomicrograph of felted-lath texture preserved in a clast of the stratiform chert breccia (E of this Plate). Plane-polarized light. Width of field is 1 cm.

PLATE 9

PLATE 9

A

B

C

D

E

F

PLATE 10. SULKY FORMATION

A. Oblique cross-section of flat-to-conical growth sequence from conical biostrome of member C of the Platform (section 2). Notebook is 17 cm long.

B. Branching-conical-columnar stromatolite (jacutophyton) as seen in a section parallel to bedding. This stromatolite form represents the culmination of a typical flat-mat to conical to branching-conical stromatolite growth-cycle (section 3). Lens cap is 60 cm in diameter.

C. Silicified ooid grainstone from member C of the Platform (section 11).

D. Dense algal lamination of member C of the Platform (section 7). Lens cap is 60 cm in diameter.

PLATE 10

PLATE 10

A

B

C

D

CHAPTER 5. GREENHORN LAKES FORMATION

Uppermost dolostones and shales of the Dismal Lakes Group, here termed the Greenhorn Lakes Formation, are well exposed in cliff sections below Coppermine River Group basalts and in gently dipping cuestas along Greenhorn Lakes and Dease Arm. The contact of the Greenhorn Lakes Formation with the underlying Sulky Formation is conformable; west of Dismal Lakes this contact is marked by a continuous conical stromatolite biostrome, and east of Dismal Lakes by a silicified evaporite breccia. The unit thins eastward from more than 600 m near Greenhorn Lakes to less than 100 m at the eastern limit of exposure. Thinning is most pronounced between Greenhorn Lakes and Bebenssee Lake.

A distinctive depositional break occurs within the formation east of central Dismal Lakes where large-scale karsting of the carbonate platform marks an internal disconformity. To the west, this erosional break becomes less pronounced and the members appear to be conformable west of Dismal Lakes.

Five informal members are recognized within the Greenhorn Lakes Formation (Figure 33, Figure 34) recording at least three transgressive pulses. From the base to the top these members are:

Figure 33. Generalized cross-section of the five members of the Greenhorn Lakes Formation (A-E). Southwestward thickening of all members is consistent with the westward-shale-out of the disconformity surface. Lithologic symbols are: 1) clean dolostone, 2) fissile shaly dolomite and dolomitic shale, 3) dolomite with minor shale partings, 4) oolitic intracrystalline dolomite, 5) quartz arenite, 6) LLH-C stromatolites, 7) conical-columnar stromatolites, 8) isolated LLH bioherms, 9) tungusoid biostrome.

MEMBER A

Member A consists of a 150 m sequence of monotonous grey to dark grey coarsely recrystallized dolostone which conformably overlies the uppermost shoal-water facies of the Sulky Formation (Figure 33). These grey dolostones are conformably overlain west of section 13 (Figures 33 and 34) by the basal biostromes of member B. From section 13 east, biostromal beds of member B are missing and the member A-B contact is undefinable (Figures 33 and 34). The absence of internal marker beds within member A render thickness estimates difficult in fault-repeated sections (e.g. section 14, Figure 34). Member A grades laterally from roughly 170 m near Greenhorn Lakes to 20 m near the eastern margin of the outcrop belt.

This member is best exposed in the type area of the Group, north of western Dismal Lakes (sections 8-10, Figure 34). Here the basal part of the member consists of up to 10 m of dolostone containing LLH stromatolites characterized by distinctive finely crenulate laminations, and less regular accumulations of edgewise conglomerate. Ooid beds (10-40 cm) occur in this portion of member A at sections 4, 6, 7 and 9 (Figure 34).

Above this lowermost sequence, most of member A consists of monotonous, uniform 20-40 cm bedded grey to dark grey dolostone containing pustulose and smooth algal mats, low-relief 10-20 cm diameter LLH-C stromatolites and flat-chip conglomerate. Oncolitic overgrowths on intraclasts are common. Tepee structures (Plate 11A) and

palisade-type algal mats (cf. Hardie and Ginsburg, 1977) are less common. Small-scale scours are present, but channeling was not noted. Nodular grey-black chert is common throughout the member.

The most remarkable feature of this unit is its lithologic homogeneity throughout the 150 m section. Part of a detailed measured section (Figure 35) illustrates the consistency of this member.

Interpretation: Member A

The basal subdivision of member A, characterized by distinctive crenulate-laminated bioherms, coarse intraclastic units and some ooidic beds, is interpreted to have formed in a moderate- to high-energy subtidal environment. The ooidic beds, 10-40 cm thick, are best interpreted as allochthonous sediments which have been transported from a more active shoal environment, presumably offshore to the west. Allochthonous ooids are common features of such modern carbonate settings like the Trucial Coast, Persian Gulf (Purser and Loreau, 1973). Tepee structures with up to 50 cm relief (Plate 11A) are interpreted as having formed during periods of desiccation (Assereto and Kendall, 1977). The local occurrence of edgewise conglomerate stacked against tepee ridges indicates a wave-swash zone setting. Taken together, these features can be interpreted as representative of a brief period of moderately

Figure 35. Representative detailed measured section of member A from north shore of Dismal Lakes (section 11) showing the predominance of dense algal mat fabric and associated peritidal sedimentary structures.

3

high-energy shallow-subtidal to intertidal deposition.

In contrast, the overlying bulk of member A is dominated by low-energy intertidal to supratidal sedimentation. Flat algal mats and less common LLH-C type stromatolites are the dominant algal structures. These, in association with pustulose (c.f. Logan et al., 1974) and palisade (cf. Ginsburg and Hardie, 1977) algal mat textures are analogous to structures typical of deposition in low-energy intertidal environments such as protected portions of the Trucial Coast (Purser and Evans, 1973), Shark Bay (Logan et al, 1974) and the leeward side of Andros Island (Hardie, 1977; Shinn et al., 1969). Intraclastic beds (5-20 cm) are the second most dominant lithology, probably representing storm-reworked desiccated mudflat sediments (c.f. Shinn et al., 1965). Two features which appear to suggest local hypersaline conditions are rare silicified halite hoppers and moderately common tepee structures. Together, these features clearly indicate an intertidal to supratidal environment of deposition for this unit.

The most remarkable feature of this member is the marked lateral and temporal constancy through which its depositional environment was maintained. The member records stable peritidal carbonate deposition across the Dismal Lakes Platform for the initial 150 m of deposition of the Greenhorn Lakes Formation. This interpreted continuity of environment suggests marked tectonic

stability during deposition.

MEMBER B

This member is best exposed and developed in the vicinity of Greenhorn Lakes in a series of west-dipping cuestas (Plate 11B). The basal contact of the member is defined by the first appearance of thick-bedded light-grey stromatolitic biostromes succeeding monotonous grey dolostones of member A. As illustrated in Figure 33, member B thins eastward from 190 m of cyclic biostrome sequences at Greenhorn Lakes to a single 2.5 m biostrome at Bebensee Lake (section 5), which is continuous to Coppermine River (section 12).

Member B east of the Coppermine River consists of monotonous peritidal dolostones which accumulated without apparent depositional break. On the September Lake High, peritidal deposition was continuous throughout members A through D, rendering the individual members indistinguishable (Figure 34).

The uppermost biostrome of member B is overlain with sharp conformity by a siltstone-shale unit (member C), which extends as far east as Teshierpi Mountain (just east of section 9, Figure 33). At Coppermine River (section 12, Figure 34), no capping lithology is exposed above the member B biostrome.

At Greenhorn Lakes the member consists of at least five well-developed biostromes which range from 5 to 22

m, and are separated by 5-40 m intervals of 20-40 cm bedded dark grey dolostone indential to that of member A (Plate 11B). Commonly, one or more laterally-restricted bioherms occur in individual sections. The main biostromes are distinctive cliff-formers of thickly bedded to massively bedded, light grey stromatolitic dolostone (Plate 11C). Bedding planes within biostromes appear to represent brief hiatuses in biostrome development. These surfaces are generally accentuated by minor changes in stromatolite column diameter and degree of elongation. No consistent vertical trends in column diameter or asymmetry could be determined through individual biostromes, or within the biostromes as a whole. Individual columns, however, maintain the same general SH-C form throughout all parts of all biostromes (Plate 11C; Figure 36). Length:width ratios of the stromatolite columns typically are 4:1 to 5:1.

Individual columns commonly are separated by a remarkably consistent centimetre-wide intercolumnar micritic infill.

Figure 36 shows a detailed section through a typical biostrome-to-bedded dolostone sequence. Particularly striking is the consistent shallow-water nature of the dolostone units which separate biostromes. Typical morphological variability of stromatolites within the biostromes also is shown in Figure 36.

Stromatolite elongations from these biostromes show a strong northeasterly trend southwest of Bebensee Lake,

Figure 36. Representative detailed measured section of member B biostrome and interbiostrome sediments from northeast of Sulky Lake (section 2).

and an easterly to southeasterly trend east of Bebenssee Lake (Figure 37). Figure 5 also shows the gradual E-W to N-S variation in elongation of successive biostromes in the best-exposed section east of Greenhorn Lakes (section 3, Figure 34). Elongation trends are consistently oriented perpendicular to the direction of thickening of the member (compare Figure 33 and Figure 37).

Thickening directions of individual biostromes cannot be determined due to the relatively homogeneous nature of the combined stromatolites. For this reason, the internal stratigraphy of member B is not well controlled. As shown in Figure 33, member B can be interpreted as either an onlap or an offlap succession. The stratigraphic position of member B, underlain by peritidal dolostones of member A, and overlain by subtidal shales and shaly carbonates of member C (see next section), indicates that onlap provides the more likely, but not compelling, interpretation.

Interpretation: Member B

Member B deposition is considered separately in terms of the distinctive biostromal units, and the intervening grey-dark grey dolostones.

Lithologies and sedimentary structures of the grey-dark grey dolostones are essentially identical to those described for member A. A minor difference is the greater abundance of tepee structures and thin (2-5 cm)

oolitic beds in member B. A similar intertidal to supratidal depositional environment is envisioned for these sediments. The critical observation, illustrated in Figure 36, is the uniformity of grey-dark grey dolostone sequences. Bed thickness, lithology and sedimentary structures are essentially constant, showing no recognizable trends as biostromes are approached. The lack of diagnostic sedimentary structures within member B biostromes makes an accurate depositional interpretation difficult. The key features of these stromatolites are their regionally consistent elongation, ubiquitous SH-C morphology and micritic intercolumnar sediment infill. Although the lack of coarse clastic intercolumnar fill suggests that current energy was not great, the consistent elongation of stromatolites in these biostromes over large areas (Figure 37) indicates the presence of regionally persistent currents. A supratidal setting is not considered to be compatible with these characteristics. Although it is difficult to exclude an intertidal setting for these stromatolites, their constant elongation over a large area would seem to support a shallow subtidal environment of deposition. Beds of peritidal origin above and below the biostromes indicate that the biostromes probably did not form in a significantly deeper subtidal environment.

Member B biostromes record a discrete change in depositional setting which affected large expanses of the

platform. The sharp contacts between subtidal biostromes and surrounding peritidal dolostones suggest pulse-like transgressive-regressive events, in contrast to gradual progradational regressive sedimentation such as that displayed by cycles in the Rocknest Formation (Hoffman, 1975).

Elongate stromatolites shown in Figure 37 suggest a southeast-trending strandline west of Bebensee Lake, if long axes are presumed to have been perpendicular to strand. (c.f. Campbell, 1979; Hoffman, 1969; Logan et al., 1974). This is consistent with direction of thickening of the Formation as a whole as shown in Figure 33. The marked northeastward thinning of member B from 190 m in the Greenhorn Lakes area to a single 2.5 m biostrome east of Bebensee Lake, clearly indicates a westward axis of subsidence for the Dismal Lakes Platform during deposition of this unit. The apparent hingeline during member B deposition was west of Bebensee Lake, well west of the long-lived Leith Line (Kerans et al., 1981).

The record of member B sedimentation on the September Lake High, east of Coppermine River (sections 13-18), is essentially continuous with that of member A and the overlying member C. A stable supratidal depositional environment must have persisted on the September Lake High during member B sedimentation. Transgressive pulses which resulted in widespread development of biostromes did not affect the September

Lake area.

MEMBER C

Where complete, member C consists of a lower shale unit and an upper massive stromatolitic unit (section 3, Figures 33 and 34; Plate 11D). In the Bebensee and Dismal Lakes areas (sections 5 and 7-10) the shale passes upward imperceptibly into member D. As with member B, equivalent sedimentation on the September Lake High is indistinguishable from underlying and overlying units.

Member C thins eastward from 120 m at section 2 to 5 m at section 10. Shale colour changes from predominantly black near the base of the section, to green, and finally to red. Flaggy bedded-to-fissile terrigenous shales are intercalated with parallel-laminated or rarely rippled 1-2 cm dolosiltite layers. Individual centimetre-thick laminae are parallel and continuous over tens of metres. Desiccation features are absent, and only one flat-chip conglomerate bed (3 cm) was noted near the eastern limit of the shale (section 8). North of Dismal Lakes (section 7), within the green central portion of the shale, black organic films on bedding planes of the green siltstone contain thick-sheathed filamentous cyanobacterial microfossils (Horodyski et al., 1980).

From Limestone Point (section 1) to section 4, the shale unit is capped by a massive cream-coloured dolostone containing conical stromatolites. These are

the only conical-columnar forms known within the Greenhorn Lakes Formation. North of Sulky Lake these stromatolites show slight east-northeast elongation, but as is typical of conical stromatolites in the Sulky Formation, intercolumnar fill is micritic.

This unit is 12 m thick north of Sulky Lake, thinning westward to 5 m at Limestone Point, and eastward to 4 m at section 4. The well-exposed shale-biostrome contact north of Sulky Lake is undulate, and shows about 2 m of relief due to loading and differential compaction (Plate 11D).

In the Bebensee and west Dismal Lakes areas (sections 5 and 6) the member C-D contact lies within a continuous shaly carbonate succession, making recognition of the contact arbitrary. Where the biostrome segment of member C is present, defining precisely the upper contact, the member is overlain by shaly carbonates which contain small stromatolite bioherms.

Interpretation: Member C

The member C shale-biostrome couplet is inferred to have been deposited during a major transgressive pulse, somewhat analogous to the Sulky Formation member A-B sequence. The lower shales and siltstones which rest abruptly on the upper member B biostrome are interpreted as subtidal, based on their mm-scale lamination, and the absence of desiccation features. Current-rippled

dolosiltite beds (2-5 cm) within the shale are likely washed-in, with the terrigenous shales representing continuous "background" sediments. The black-to-green-to-red upward change in shale colour is indicative of increasingly oxidizing conditions, and is interpreted as a consequence of shallowing of the basin.

The conical-columnar stromatolite biostrome capping member C west of Bebensee Lake records a continuation of quiet subtidal conditions in this area. The intercolumnar micrite and locally developed weak stromatolite elongation indicate a low-energy depositional environment. Finally, the generalization that most conical-columnar stromatolites accumulated in subtidal environments (Donaldson, 1976; Hoffman, 1976; Chapter 4) further supports a subtidal interpretation.

The member C shale-to-carbonate transition is perhaps equivalent to the core-shale to upper-limestone portion of the classic midcontinent Pennsylvanian cyclothems discussed by Heckel (1980). According to Heckel, core shales were formed by offshore transport of terrigenous fines from a shoreline source undergoing rapid transgression, with the overlying limestones accumulating when shoreline migration had stabilized. By analogy, member C shales would have formed during transgression, with the biostrome developing when transgression had stabilized and clastic input had decreased. Although the biostrome is restricted to the

western portion of the outcrop area, the lower member C shale extends as far east as Teshierpi Mountain (10 km east of section 10), preserving the record of a major transgressive event across the Platform.

As with preceding members, no apparent response to this transgressive event is recorded on the September Lake High, where deposition of peritidal carbonates was continuous.

MEMBER D

The member C-D contact is easily recognized where the member C biostrome is present. Likewise, the upper contact of member D east of section 8 (Figures 33 and 34) is clearly marked by a disconformity. However, the lower contact east of section 4 and the upper contact west of section 8 are both arbitrary, lying within a continuous, but poorly exposed, shaly dolostone (Figures 33 and 34).

The most characteristic lithologies of the member on the Platform are red shaly dolostones and isolated bioherms of LLH and SH type stromatolites. Elongation directions of these bioherms are consistent with those of underlying member B (Figure 37). Thick ooid beds (1-2 m) near the top(?) of member D at sections 4 and 5 are a third important lithology. These ooid beds are well-sorted medium-grained grainstones. No crossbedding is apparent in these grainstone beds.

These three lithologies dominate member D in

sections 1, 3, 4 and 5. In section 2, however, 50 m of 20-40 cm bedded dolostones containing flat-chip conglomerates, tepee structures and flat algal mats appear to structurally overlie the upper member C biostrome. Unfortunately, this 50 m section is structurally isolated, placing its stratigraphic position in doubt. These sediments, though, are remarkably similar to those of member A, and to interbiostromal portions of member B.

In sections 13-18 (Figures 33 and 34), member D consists of 20-40 cm bedded algal mat-intraclastic dolostones which are continuous from member A through D. As previously mentioned, contacts between members on the September Lake High cannot be identified because of lithologic similarities of the units.

North of Dismal Lakes (sections 7-10), member C shales grade upward into cream dololutes with abundant halite pseudomorphs and parallel current lamination, and typical intraclastic-stromatolitic dolostones (Plate 11E).

Interpretation: Member D

Shaly to flaggy bedded dolostones and isolated bioherms forming the bulk of member D from Bebensee Lake westward are interpreted as having been deposited in a quiet-water setting. Stromatolites show northeasterly elongation trends, parallel to those of the member B

biostromes, implying a similar paleogeographic setting and/or paleocurrent regime. Low synoptic relief and LLH-C stromatolite forms within these biostromes suggest relatively low current energy. In addition, intercolumnar fill (where present) and interbiohermal fill is micritic, also suggesting a low-energy environment of deposition. These data, in addition to the absence of desiccation features or current scour, imply a low-energy subtidal setting for most of member D in the west. Exceptions to this setting are the thick grainstones (sections 4 and 5) and small bioherms (section 1) cut by channels filled with ooidic-intraclastic sediments. Clearly, high-energy depositional conditions existed locally.

North of Dismal Lakes (sections 9-12) member D intraclastic-ooidic carbonates and SH-C and SH-V type stromatolites probably accumulated in a moderate- to high-energy environment. Mud cracks and salt casts on bedding surfaces indicate intermittent exposure. This sequence is best interpreted as being dominated by high-energy low intertidal deposition, with periods of desiccation indicating prolonged exposure. Ooids, mixed with intraclasts, are probably allochthonous, being brought from offshore shoals similar to those described above from the western Platform.

East of section 13 on the September Lake High, only grey dolostones interpreted as supratidal to intertidal

deposits are present.

DISCONFORMITY SURFACE

The disconformity surface within the Greenhorn Lakes Formation is exposed as far west as section 8 and as far east as section 18. The variability of the disconformity surface and its relation to paleodepositional relief, are evident in Figure 38. Westward from section 16 on the September Lake High, the erosion surface becomes progressively more subtle (Figure 38). At section 14, detailed mapping has revealed as much as 40 m of erosional relief related to karsting in underlying Greenhorn Lakes lithologies (Plate 12A; Figure 38). Roof pendants, large-scale collapse breccias (Plate 12A) and fluvial cave-floor chert granule sediments (Plate 12B) characterize this area. About three km farther west (section 13), the disconformity surface is marked by 2 m of reworked karstic dolostone breccia and by weathered joints which are filled from above with quartz arenite (grikes). These joints extend some 2 m into the underlying member D dolostone (Plate 12C). Joints in exposures 12 km farther west extend 1 m into the underlying dolostones, but here the infill material is a speleothemic flowstone (Plate 12D; Figure 38). At section 10, a regolith profile is developed below the disconformity surface (Figure 38). Here the underlying ooidic dolostone is leached and recrystallized to a

Figure 38. Details of the intra-Greenhorn Lakes disconformity surface.

Sketches 8, 9, 10, and 12 are 1 m wide and are keyed to section locations in Figure 33. Section 8) Sharp contact on dolostone-lithic conglomerate resting directly on unweathered member D dolostone. Section 9) Sharp contact of dolostone-quartz-lithic wackestone upon unweathered member D dolostone. Section 10) Regolithic weathering of ooid packstone of member D up to 60 cm below the disconformity surface. Geodic chert nodules (black) are associated with the regolith. Dashed area represents an area of weathered dolostone and clay. Section 12) Minor joint system developed in member D dolostones up to 1 m below the disconformity. Joints are filled with both speleothemic flowstone and quartz arenite. Section 14) Well-developed karst collapse breccia showing a portion of a preserved roof peddant. Blocks in the breccia are highly angular and range from 1-10 m in length. The matrix of the breccia consists mostly of quartz sand which sifted down from the disconformity surface, mixed with highly weathered dolostone. Section 16) Exhumed paleohorst of member D overlapped by basal sandstones and sandy dolostones of member E. A progressive decrease in quartz content is observed in the overlapping member E sediments near the top of the paleohorst structure.

FIGURE 38

micritic carbonate for some 40 cm below the erosional contact. Small 0.5 cm dyklets are down-filled with weathered carbonate debris, and hollow globular chert concretions (1-2 cm) are concentrated within the weathering zone. Felted needle texture pseudomorphic after evaporites (cf. Chowns and Elkins, 1974) is well preserved in the chert.

The disconformity surface at section 8, 5 km farther west, is marked by a sharp contact with essentially unweathered carbonates of underlying member D (Figure 38). Farther west, the disconformity surface is not exposed.

North and east of September Lake, the disconformity surface was observed at two localities. At the disconformity, section 18 consists of a typical reworked karstic breccia with breccia clasts incorporated in a quartz-sand matrix. At section 17, the disconformity is subtle, marked only by a centimetre-thick layer of quartz-chert wackestone and by shallow (5 cm) down-filled fractures (Figure 34).

Interpretation: Disconformity Surface

Figure 38 summarizes the morphology and detailed character of the disconformity surface. There is no reflection of the later shallowing, and emergence is reflected in underlying member D sediments, indicating that regression and platform exposure was rapid.

Another key observation is the localization of intense weathering and erosion in the area between September Lake and All Night Lake. From September Lake west, weathering profiles become increasingly subtle, until only a sharp discontinuity surface marks the contact at section 8. This variability in weathering profile, together with the reconstructed paleotopography of the disconformity surface (Figure 38; see discussion in Chapter 6) indicates that the September Lake area was a centre of uplift and exposure.

The rapid and localized uplift (doming?) which caused exposure and erosion of the Platform appears related to local tectonism. This interpretation is supported by the presence of syndepositional faults in the All Night Lake area (see Chapter 6). It is tentatively suggested that this local uplift is related to early-stage emplacement of the Muskox Intrusion (see Chapter 6).

West of section 8 the disconformity is no longer apparent in the poorly exposed shaly dolostones of the member D-E contact zone. At least some portion of the western shallow subtidal shaly dolostone-biostrome-oid sequence probably represents the record of continuous deposition offshore of the exposed September Lake High.

MEMBER E

The uppermost member of the Greenhorn Lakes

Formation is best exposed in sections 7-10 (Figures 33 and 34) below the Coppermine River Group basalts. From Dismal Lakes east, the lower contact of member E is the disconformity surface discussed above. West of this area, the recessively-weathering shaly dolostones of lower member E are generally poorly exposed. Here, lacking evidence to the contrary, the member D-E contact is presumed to be conformable. In this same area the ooid beds in member D occur near the member D-E contact, and could be considered part of either member.

Shaly stromatolitic carbonates at the base of member E grade upward into typical 20-60 cm bedded intraclastic-ooidic-stromatolitic dolostones, the thickest exposed section of which is 80 m, preserved below the Coppermine River Group outlier north of Sulky Lake (section 2, Figure 34).

East of section 8, member E consists of a lower siliciclastic fining-upward sequence overlain by a 20-60 cm bedded clastic-carbonate section which is equivalent to that of uppermost member E in the western outcrop area.

The complete fining-upward sequence for the lower siliciclastic portion of member E is as follows: 1) basal quartzose dolostone-chert lithic conglomerate (10-40 cm); 2) chert-dolostone lithic subquartz arenite (0-100 cm); 3) quartz arenite (0-100 cm); 4) flaser-lenticular bedded quartz packstones and intercalated mudstones (0-10 m,

Plate 12E); 5) red and green variegated siliciclastic shale (0-10 m). In sections 7 through 15, a distinctive tungussoid stromatolite marker is roughly coincident with the uppermost occurrence of siliciclastic material in the section (Figure 33; Plate 12F).

The basal sand-rich lithologies are either massive, or consist of 2-10 cm planar-tabular crossbed sets with dispersed paleocurrent directions (qualitative observation). Abundant carbonate cement is present in these units, and beachrock (c.f. Donaldson and Ricketts, 1979) occurs locally. The overlying flaser-lenticular beds are locally mud cracked and contain 2-4 cm intraclastic beds as well as rare 4-6 cm ooid packstone beds with sharp lower and upper contacts. The uppermost siliciclastic shales show abundant mud cracks, but do not contain rip-ups or intraclasts. Figure 39 shows the distribution of these facies with respect to the disconformity surface. All facies except a 2-cm basal quartz wackestone pinch out onto the paleotopographic high above the roof of the Muskox Intrusion (see Chapter 6).

As described previously, west of the exposed disconformity, shales and shaly dolostones grade upward into clean 20-60 cm bedded intraclastic-ooidic-stromatolitic dolostones. This clean dolostone unit persists to the contact with the overlying Coppermine River Group basalts. The clean dolostones

Figure 39. Detailed cross-section of lithofacies directly overlying the intra-Greenhorn Lakes disconformity surface. Sections are not keyed to section numbers in Figure 33 as they are more closely spaced. Facies changes display eastward onlap (see text). Lithologies are: 1) quartzose dolostone-chert lithic conglomerate, 2) quartz arenite, 3) dolomitic quartz arenite, 4) wavy bedded intercalated dololomite-quartzose dolomitic packstone, 5) dolomitic shale, 6) shale, 7) clean dolostone, 8) tungusoid biostrome. Black units are ooid and intraclast packstones and grainstones interpreted as storm deposits.

FIGURE 39

thin from 80 m at section 2 to 24 m at section 16, and continue to thin gradually northward to section 18, where they are only 16 m thick (Figures 33 and 34). The upper dolostone shows remarkably little facies variability, both vertically and laterally. Where exposed, dolostones immediately below the Coppermine basalts contain mixed ooidic and intraclastic detritus, as well as small 5-20 cm LLH-C stromatolites and flat algal mats.

The basalt-dolostone contact is everywhere conformable, and evidence of pre-extrusion erosion is lacking. In some areas, to a maximum depth of 70 cm, the dolostone below the contact is altered to a chalky carbonate. The basal basalt flow locally contains pipe vesicles, and much of it is extensively altered. Agate and calcite amygdules are common in the basal flow (Baragar and Donaldson, 1973).

Interpretation: Member E

The final phase of Dismal Lakes Group sedimentation involved transgression and inundation of the exposed eastern Platform and September Lake High, with peritidal deposition re-established well before extrusion of flood basalts of the Coppermine River Group, which resulted in termination of carbonate deposition.

Because of stratigraphic uncertainty of placement of the member D-E contact west of section 8, the precise nature of this transition remains uncertain. As

previously discussed, the contact is presumed to be conformable, based on the westward decrease in degree of erosion, as seen along the exposed portion (Figure 38).

Shaly dolostones containing isolated bioherms of LLH stromatolites, essentially identical to those of upper member D, occur in lower member E. These likewise are interpreted as moderate- to low-energy subtidal deposits. Some portion of this sequence must contain sediments which are time-equivalent to the transgressive fining-upward siliciclastic sequence that rests directly on the disconformity surface east of section 7. None of the exposed member E sediments west of section 7 contain siliciclastic material coarser than clay-size; apparent restriction of sand-sized terrigenous detritus to the nearshore areas of basal member E may be either a function of non-preservation or non-exposure, or a reflection of hydrodynamic concentration within the high-energy nearshore environment.

The basal fining-upward siliciclastic sequence in section 8-18 is interpreted as a transgressive marine shoreline deposit. That this sequence is of onlap type is clearly shown by the cross-section in Figure 39. Basal lithic-rich conglomeratic sandstones represent a thin transgressive lag, analogous to the classic basal Cambrian conglomeratic sandstones of the North American craton (Doty and Batten, 1977). That these basal sandstones are marine and not fluvial is supported by the

massive-to-small-scale planar-tabular crossbedding and dispersed paleocurrent directions of this unit. A moderate- to low-energy environment is the most likely depositional setting for this unit.

Flaser-bedded dololomite-quartz packstone overlying the basal transgressive lag is interpreted as a mixed tidal flat deposit. Multidirectional current bedding and flaser bedding are indicative of transport and deposition by tidal currents (Reineck and Singh, 1975). Mud-cracked bedding surfaces and associated beds of rip-up clasts record periods of desiccation and reworking by storm tides. Additional evidence of storm deposition are the isolated ooid packstone beds (Figure 39) which record essentially instantaneous depositional events. The preferred interpretation for these ooid beds is that they were transported onshore from subtidal ooid shoals during onshore-directed storms.

The final component of the siliciclastic fining-upward succession consists of 1-2 cm bedded mudstones and shales. The shales are pervasively mud-cracked, and commonly contain thin (2 mm) quartz sand stringers which delineate bedding. A tidal flat environment is postulated for these shales as well. A more restricted upper intertidal to supratidal environment may be suggested by the greater degree of desiccation, lack of sand-sized detritus or of storm-generated intraclastic and ooidic beds.

As illustrated in Figure 39, the complete fining-upward succession is present only in sections 8, 9, 10 and 11. East of this area, the basal quartzose transgressive lag is immediately overlain by a thin unit of quartz arenite and quartzose dololutite, and then typical peritidal dolostones. In section 16 (Figure 34), the only evidence of an erosional break is a 3 cm bed of quartz wackestone (see Figures 38 and 39). Apparently the siliciclastic source area was gradually cut off during transgression, resulting in resumption of dolostone deposition. If the siliciclastics were transported into the area by longshore currents rather than by fluvial processes, as is indicated by the sedimentary structures (see above), then a depositional model can be envisioned in which gradual transgression across an irregular shoreline eventually caused isolation of small islands. As a consequence, these islands would be cut off from the "river of sand" provided by longshore currents in a carbonate-producing environment, and quartz-free sediments would begin to accumulate around them. A similar process of isolation could have occurred around the Septemner Lake High.

The provenance of the quartz arenites of member E poses a problem. The compositional and textural maturity of these sands suggests a multicyclic origin (cf. Suttner et al., 1981). The abundant quartz-rich sediments in the lower Dismal Lakes and Hornby Bay Groups would need to

have 500 m eroded to be exposed, and are thus an unlikely source. A second possible source could have been the lower quartz-rich units of the lower Coronation Supergroup or its equivalents, assuming that they were still exposed. Campbell (pers. comm., 1981) recorded local southerly-directed paleocurrents in the lowermost Kanyuyak Formation siliciclastics of the Bathurst Inlet area. Perhaps these and member E quartzose clastics were derived from a northerly source such as Victoria Island, where outliers of Goulburn Group siliciclastics occur in windows through Paleozoic cover.

The uniformity of the tungusoid stromatolite biostrome just above the lower siliciclastic portion of member E (sections 7-15, Figures 33 and 34) suggests that depositional relief effectively had been eliminated by the time that dolostone deposition resumed. However, the paleohorst structure at section 16 appears still to have been a topographic high at this time (Figures 33 and 39).

Following deposition of transgressive siliciclastics, peritidal carbonate deposition again spread throughout the area of the Dismal Lakes outcrop belt. Sections 1-12 are dominated by locally channelled SH-V stromatolites and by thin ooid or mixed ooid-intraclastic beds characteristic of moderate- to high-energy depositional environments. These lithofacies everywhere dominate the upper 40-80 m of member E. The presence of sediments interpreted as peritidal everywhere

below the Coppermine River Group basalts, supports the persistence of regionally widespread shallow-water conditions throughout the Homocline at the time of initial Coppermine volcanism.

SUMMARY: GREENHORN LAKES FORMATION

Initial deposits of the Greenhorn Lakes Formation are monotonous medium-bedded grey dolostones which record extremely stable peritidal sedimentation throughout the area. Gradual eastward thinning of lowermost (member A) deposits reflects the thinning of the Group as a whole. The persistent shallow-water record of member A indicates a close balance between subsidence and accumulation.

Minor transgressive pulses, possibly reflecting overall increased subsidence of the Platform, resulted in deposition of a cyclic series of laterally extensive subtidal biostromes (member B). Consistent elongation of individual stromatolites within biostromes (Figure 37), coupled with thickening and facies changes (Figure 34), indicate a generally SSE-oriented strandline during member B deposition. A single major transgressive pulse succeeded member B, depositing the shale-biostrome couplet of member C. Terrigenous detritus was generated in a nearshore setting and transported offshore to accumulate on the now-subtidal portion of the Platform. As shoreline progradation stabilized and supply of terrigenous detritus to the Platform decreased, a conical

stromatolite biostrome developed, signaling the return to carbonate deposition. Member D deposition was predominantly in a shallow-water subtidal environment from Bebensee Lake west. Here, on the deeper portion of the Platform, stromatolite elongations remained consistent with those of member B biostromes, suggesting a relatively constant current régime and/or basin paleogeography. Moderate- to high-energy intertidal deposits accumulated in the central Dismal Lakes area, and the September Lake High persisted as a positive feature, maintaining a consistent peritidal position as recorded in a markedly thinned section.

Exposure of both the eastern platform and the September Lake High resulted in erosion and karsting east of central Dismal Lakes. This abrupt uplift may be related to tectonism associated with initial emplacement of the Muskox Intrusion. West of central Dismal Lakes, shaly dolostones of basal member E, identical to those of member D, were deposited continuously.

Transgression of the exposed portion of the September Lake High followed, with deposition of a transgressive lag of siliciclastic shoreline sandstones and tidal-flat shales. Siliciclastic input gradually decreased, and the entire area was again covered by peritidal dolostones. Dismal Lakes Group deposition ended with extrusion of the Coppermine River Group basalts.

PLATE 11. GREENHORN LAKES FORMATION

A. Tepee structure in grey algal-laminated dololotite, filled by light grey dololotite. This tepee ridge had 30 cm of topographic relief on the depositional surface. Photograph of member A(?) at section 14. Width of field is 1 m wide.

B. Northwest-dipping cuestas of member B biostrome-dolostone couplets exposed northeast of Greenhorn Lakes (section 3). Resistant biostromes cap cuestas (arrows), while grey-dark grey dolostones form slopes. Each cuesta has approximately 60 m of section exposed. Foreground is dip-slope of biostromal cuesta cap (section 3).

C. Characteristic light grey massive-bedded dolostone of member B biostrome (section 2). Broad SH-C columns are 20-40 cm wide and are separated by centimetre-thick micrite-filled interspaces (not visible in photograph).

D. Cliff face of member C with slope of red-tan shale and cap of massive light grey stromatolitic dolostone (section 2). Dolostone cap is 5 m thick.

E. Typical clastic-stromatolitic dolostone of member E with LLH-S and LLH-C low domal stromatolites locally scoured, and infilled by coarse dolarenite (arrow). Photograph from section 10.

PLATE 11

PLATE 11

A

B

C

D

E

PLATE 12. GREENHORN LAKES FORMATION

A. Large-scale karst collapse breccia composed of blocks of member D and possibly older units. Dotted line separates in-situ dolostone below from collapse breccia above. Figure 38 gives relative stratigraphic position of this photograph (section 14). Person (arrow) is 1.8 m tall.

B. Hammer rests on trough-crossbedded chert granulestone which is surrounded by karst collapse breccia. Chert granulestone represents insoluble residue from karst dissolution which has been reworked by cave-floor streams (section 14).

C. Sandstone-filled grike system developed in member D dolostone below disconformity surface. At this locality the grikes extend 10 m below the contact. Photograph taken at section 13. Notebook is 17 cm long.

D. Joint system which has been solution-widened, occurring approximately 60 cm below the disconformity surface. Here the joint system is filled with fibrous-to-prismatic dolomitic speleothem flowstone (section 12). Figure 38 gives the stratigraphic position relative to the disconformity surface.

E. Thin intercalated layers of quartz arenite (dark) and dololutite (light), probably representing a variety of lenticular bedding. Figure 39 gives stratigraphic position of this lithofacies. Pocket knife is 6 cm long.

F. Portion of tungussioid biostrome of member E. Column margins are irregular, and passive branching is common. Intercolumnar fill is ooid wackestone-packstone. Figures 33, 34, and 39 give stratigraphic position of this biostrome (section 9). Width of field is 25 cm.

PLATE 12

PLATE 12

A

B

C

D

E

F

ASPECTS OF THE TECTONIC HISTORY OF THE DISMAL LAKES GROUP

CHAPTER 6. TIMING OF EMPLACEMENT OF THE MUSKOKX INTRUSION

The Muskox Intrusion is a layered ultramafic body of middle Proterozoic age in the northwest corner of the Canadian Shield (Smith and Kapp, 1963; Irvine, 1970a; Figure 40). In cross section, the Intrusion is interpreted to be a N-S elongate oblate funnel. Both the Intrusion and the concordantly overlying middle Proterozoic Coppermine Homocline (Kerans et al., 1981) dip gently northward at 5 degrees. Aeromagnetic and gravity data suggest that the Muskox Intrusion continues northward beneath Coppermine Homocline for a minimum of 30 km, and possibly for as much as 120 km (Fraser et al., 1972; Hornal, 1968).

The Muskox complex was emplaced into the early Proterozoic Akaitcho Group metasediments and into parts of the Hepburn metamorphic-plutonic complex of Wopmay Orogen (Hoffman, 1981; Hoffman and St. Onge, 1981; Easton, 1981) which forms the floor and portions of the roof zone (Figures 40 and 41). The remainder of the roof rocks are basal sandstone units of the middle Proterozoic Hornby Bay and Dismal Lakes groups (see discussion below; Figures 45 and 46)

The Muskox Intrusion is the largest and best exposed layered ultramafic body in the Canadian Shield. It is also one of the better-studied intrusions of its kind, and is commonly cited as an example of fractional

Figure 40. General geology of the Coppermine Homocline and underlying Wopmay Orogen. Muskox Intrusion is shown in black. Diagonal striping- Coppermine River Group; dolomite pattern- Dismal Lakes Group; stippled pattern- Hornby Bay Group; GBB- Great Bear Batholith; HMPC- Hepburn metamorphic-plutonic complex; AFB- Asiatic Fold Belt; TB- Tree and Takijug Basins; S- Slave Craton; T- Teshierpi Fault; Bar-and-dot symbol is on downthrown side of faults. Location of Figure 45 is outlined.

Figure 41. Fault pattern of main Canoe Lake Fault and its east (E CLF) and west (W CLF) branches. Bar-and-dot symbols are on downthrown sides of post-depositional faults. Faults with double bar symbol on downthrown side are syndepositional. Numbered stippled areas along faults mark dolostone breccias. See text for further discussion. Light stipple is for Hornby Bay-Dismal Lakes groups. Dense stipple is for covered areas.

FIGURE 41

crystallization processes (e.g. Hyndman, 1972; Irvine and Smith, 1967, 1969; Irvine, 1980). In addition, the Muskox Intrusion is a key component of the Mackenzie Igneous Event (see Fahrig et al., 1981), other major components of which are the Mackenzie diabase dykes and the Coppermine River Group basalts (Fahrig et al., 1981; Fahrig and Jones, 1969; Robertson, 1969; Figure 42). Present radiometric dates provide only poor resolution of age relationships of these magmatic events, thus placing a great deal of importance on understanding the basic geologic relationships as a basis for establishing relative ages.

The primary purpose of this chapter is to attempt to resolve the question of the age relationships of the Intrusion relative to the Dismal Lakes and Coppermine River groups, a question which has been debated since the initial discovery of the intrusion (Smith and Kapp, 1963; Irvine, 1970a; Hoffman, 1980). The second part of this chapter will attempt to document in greater detail the early, pre-extrusion record of the emplacement of the Intrusion as preserved in uppermost sediments of the Dismal Lakes Group.

PREVIOUS WORK

Since initial discovery of the Muskox Intrusion in 1956, most work on the Intrusion has been done by geologists of the Geological Survey of Canada (Smith,

Figure 42. Distribution of some Mackenzie-age magmatic rocks within the Canadian Shield. 1) Coppermine River Group; 2) Muskox Intrusion; 3) Mackenzie diabase dykes; 4) Ekalulia Formation basalts (equivalent to Coppermine River Group in Bathurst Inlet area). Figure modified from Baragar (1977).

1962; Smith and Kapp, 1963; Smith, Irvine and Findlay, 1967; Irvine 1970a, 1970b, 1971, 1980; Irvine and Smith, 1967, 1969; Irvine and Baragar, 1972).

A major problem with the history of the Intrusion has been its age relationships with respect to overlying Dismal Lakes Group sediments and Coppermine River Group basalts. All previously published cross-cutting relationships and radiometric age data (all dates group around 1200 M. Y.) are equivocal.

Before 1970, the Muskox Intrusion was assumed to be coeval and probably comagmatic with the Coppermine River Group, based on geographic proximity, similar though poorly constrained K/Ar ages (Fahrig and Jones, 1969) and Rb/Sr ages (Smith et al., 1967; Baragar, 1972) and paleomagnetic data (Robertson, 1969; Patchett et al., 1978). In addition, Baragar (1969) was able to correlate minor-element trends of Cu and Ni in the Coppermine River Group basalts with trends observed in the Muskox Intrusion (Irvine and Smith, 1967, 1969), suggesting that it represents the feeder chamber for much of the basalt. The relatively simple history implied by this assumed relationship is illustrated in Figure 43a.

Irvine, however, during his detailed mapping of the exposed portion of the Intrusion and its cover, discovered that Dismal Lakes Group dolostones (which occur between the Muskox Intrusion and Coppermine River basalts) show no evidence of contact metamorphism in

Figure 43. Previous interpretations of the emplacement history of the Muskox Intrusion. A) pre-1970; B) 1970; C) 1980 (see text for discussion). Dashed pattern- Wopmay Orogen; stippled pattern- Hornby Bay Group; dolomite pattern- Dismal Lakes Group; vertical lines- Coppermine River Group (or other flood basalt sequences as in B); black- Muskox Intrusion. The fault illustrated is the Canoe Lake Fault.

FIGURE 43

A.

B.

C.

outcrops only 300 m from the roof of the Intrusion (Figure 46). In addition, he observed that the Canoe Lake Fault (CLF), a north-striking normal fault which offsets the intrusion 500 m in a west-side-down sense, did not appear to affect the overlying Dismal Lakes or Coppermine River groups (Irvine, 1970a). Here Irvine assumed that the west branch of the CLF (see Figure 41) was the true continuation of the fault. From these relationships, Irvine reconstructed the following series of events for the area: 1) intrusion of the Muskox complex cutting both Wopmay basement and Hornby Bay Group sandstone, and extrusion of an associated flood-basalt sheet (required by the mechanism of formation of the ultramafic layered portion of the intrusion); 2) offset of a part of the Muskox Intrusion and surrounding strata by the Canoe Lake Fault; 3) erosion; 4) deposition of the Dismal Lakes Group followed by extrusion of Coppermine River Group flows (Irvine, 1970a; see also Figure 43b). This series of events implies a significant time gap between emplacement of the Muskox Intrusion and extrusion of the Coppermine River Group, thus requiring two distinct periods of magmatism.

Hoffman (1980) suggested that Irvine's arguments could be circumvented if it could be demonstrated that the "east branch" of the Canoe Lake Fault was actually the true extension of the Canoe Lake Fault, as opposed to the west branch as was assumed by Irvine (1970a; see also

Figure 41). If this alternative hypothesis for the east branch of the Canoe Lake Fault is considered, the possibility that equivalent offset of Dismal Lakes and Coppermine River Group units has occurred, cannot be readily discounted (Hoffman, 1980). An additional consequence of this postulated easterly extension of the CLF is that the apparent proximity of dolostones unaffected by contact metamorphism can be explained by fault juxtaposition (Hoffman, 1980; see also Figure 46). Thus, by proposing a northeasterly extension of the Canoe Lake Fault, Hoffman was able to counter Irvine's (1970a) arguments, and suggest once again that the Muskox Intrusion and Coppermine River Group were coeval and possibly comagmatic (Figure 43c).

NATURE OF THE EAST BRANCH OF THE CANOE LAKE FAULT

A west-side-down displacement of at least 500 m along the proposed east branch of the CLF is required to attain a magnitude of displacement equivalent to that observed on the main portion of the CLF, and to place the unmetamorphosed dolostones (discussed above) a minimum of 500 m above the roof of the intrusion, thus removing them from the proposed minimum contact-metamorphic zone of the intrusion.

The primary goal of this aspect of the thesis was to obtain an accurate estimate of displacement along the hypothesized east branch of the CLF. Reconnaissance

mapping in the area (Figure 24 of Kerans et al., 1981) had indicated that faults subsidiary to the CLF may have been active during deposition of the upper Dismal Lakes Group, and this possibility also warranted further investigation.

The composite stratigraphic column and detailed map of the roof of the intrusion in the All Night Lake area are shown in Figures 44 and 45. The map demonstrates the nature of displacement along the east and west branches of the CLF. As indicated by Irvine (1970a), the west branch of the fault has only minor displacement, being in the opposite sense to the main CLF, and thus is not a likely continuation of the main fault. The east branch, however, shows west-side-down displacement consistent with that of the main branch of the CLF. The cross section in Figure 46 indicates that the magnitude of displacement is approximately 500 m, equivalent to that for the main CLF. The negligible displacement of the Coppermine River Group where the fault transects this sequence (Figure 45, point X) may be accounted for by the easterly splay 11 km north of the main splay which takes up most of the displacement. Thus, the fault which intersects the Coppermine River Group basalts at point X in Figure 45 is not the main continuation of the CLF.

Two conclusions are drawn from the map and cross section (Figures 45 and 46): 1) that the east branch is the continuation of the CLF, and; 2) that restoration of

Figure 44. Composite stratigraphic column for Hornby Bay and Dismal Lakes groups in the All Night Lake area. Column at right also serves as legend for Figures 45 and 46. M.I.- Muskox Intrusion. Black strata in left column are granophyre-rich sills, possibly related to the Muskox Intrusion (see text).

Figure 45. Geologic sketch map of the All Night Lake area compiled from mapping done for this study with additional data taken from Baragar and Donaldson (1973) and Hoffman and Pelletier (1982). Key to map units as in Figure 44. BL- Badlands Granite; MNDH- Muskox North Drill Hole (Smith et al., 1967). Faults with bar-and-dot are post-depositional, whereas those with fine symbols are interpreted to have been active during Dismal Lakes deposition (see text). A-A' is the line of cross-section illustrated in Figure 46. "X" marks the intersection between a splay of the east branch of the Canoe Lake Fault and the lower contact of the Coppermine River Group (see text). Location of map is shown in Figure 43.

FIGURE 45

Figure 46. Cross-section illustrating relationships of the east and west branches of the Canoe Lake Fault to the Muskox Intrusion and the Hornby Bay-Dismal Lakes cover sequences. Key to lithologies as in Figure 44. Fault movement arrows with the letter S are syndepositional faults. Stippled unit within Greenhorn Lakes Formation is the unconformity complex discussed in text. C.R. Gp.- Coppermine River Group. X- location of unmetamorphosed dolostones closest to the roof of the intrusion.

FIGURE 46

displacement on the east branch of the CLF is sufficient to have "removed" the unmetamorphosed dolostones from the effects of contact metamorphism. Thus, as suggested by Hoffman (1980), and refined here, geologic relationships appear to confirm the interpretation that the Muskox Intrusion and Coppermine River Group lavas are coeval.

THE HORNBY BAY GROUP-DISMAL LAKES GROUP UNCONFORMITY

Indirect evidence pertaining to Muskox-Coppermine age relationships is shown by the Hornby Bay-Dismal Lakes unconformity. Irvine's (1970a) work on cover relationships of the Muskox Intrusion was conducted simultaneously with mapping of the Hornby Bay and Dismal Lakes groups (Baragar and Donaldson, 1973). At this time it was clear that an unconformity existed between the two groups, but detailed relationships were unclear. Irvine (1970a) postulated a pre-Dismal Lakes Group emplacement of the Muskox Intrusion (Figure 43b), and he suggested that erosion associated with the Hornby Bay-Dismal Lakes unconformity removed the extrusive equivalents of the Intrusion. Such extrusive equivalents are required to explain olivine enrichment of the layered series of the Intrusion (Irvine, 1970a).

Detailed work on the unconformity (Ross, 1980; Kerans et al., 1981) has shown that only locally is it significant, and that west of Teshierpi Fault (Figure 40) the Hornby Bay and Dismal Lakes groups are in gradational

and conformable contact. In addition, Ross (in Kerans et al., 1981) has shown that paleoflow of the basal Dismal Lakes Group fluvial sandstone was to the southwest, draining the area to be occupied by the postulated basalt sequence. Thus, if a thick flood basalt sequence was deposited and then eroded, evidence of its source should be present in western basal Dismal Lakes sandstones. In fact, no evidence of a local basaltic source is present in the sandstones; this evidence supports a post-Dismal Lakes origin for the Intrusion and its extrusive equivalents.

POSITIVE EVIDENCE FOR THE AGE OF EMPLACEMENT

Cross-cutting Relationships:

Aside from the clear offsets of the east branch of the CLF, the most striking revision to earlier maps (Figure 45) is the distribution of the basal clastic units of the Hornby Bay and Dismal Lakes groups.

Locally, both Unit 8 of the Hornby Bay Group and Unit 11 of the Dismal Lakes Group (terminology of Baragar and Donaldson, 1973; Figure 44) are quartz arenites, making distinction difficult and accounting for previous critical misinterpretation of those units in the All Night Lake area (compare Figure 45 with Figure 2 of Irvine, 1970a). The most reliable criteria for distinguishing Unit 8 from Unit 11 are those based on stratigraphic relationships. Internal characteristics

such as lithology and sedimentary structures of both units vary significantly laterally, and thus have a low reliability for differentiation of the units (Ross, 1980).

Following detailed examination of stratigraphically known outcrops of Units 8 and 11, characteristic lithologic and paleocurrent signatures were developed for the All Night Lake area, and outcrops with no stratigraphic control could be mapped confidently. In the All Night Lake area, Unit 11 sandstones are white to pink, very well-sorted quartz arenites with scattered to bimodal paleocurrent patterns. Hornby Bay Unit 8 sandstones (probably Unit 8c of Kerans et al., 1981) are red, poorly sorted quartz arenites rich in vein quartz, with unimodal northwest-directed paleocurrents.

Sandstones exposed at the roof contact of the Intrusion in the vicinity of the Musko North Drill Hole (MNDH: Figure 45), previously mapped as part of the Hornby Bay Group, are reinterpreted here to be Dismal Lakes Group Unit 11 on the basis of the following criteria: 1) the sandstones exposed at the MNDH are well-sorted quartz arenites, and; 2) more importantly, they grade continuously upward over 30 m into interbedded quartz arenites and black shales typical of the Fort Confidence Formation. In outcrop, mafic dykelets cut both units, which show moderate recrystallization. In the MNDH core, the sandstones grade downward into

quartzites, which are brecciated at the contact with the Intrusion (Findlay and Smith, 1965). Tridymite pseudomorphs have been reported in some of the quartzites (Pouliot, 1968). On the basis of these data, the Muskox Intrusion is interpreted to be younger than the Fort Confidence Formation, and, by reason of depositional continuity, younger than the Group as a whole with the possible exception of the upper Greenhorn Lakes Formation (see below). This cross-cutting relationship is the most direct indication of relative ages, and it supports the contention that the Intrusion is coeval with the Coppermine River Group basalts.

Upper Dismal Lakes Group Doming and Faulting:

Character of Intra-Dismal Lakes Group Disconformity

An important feature of the Dismal Lakes Group is a distinctive local karstic disconformity developed within the Greenhorn Lakes Formation (Figure 38; Chapter 5). In the All Night Lake area, as in the vicinity of September Lakes, megabreccias occur below this disconformity. In the September Lakes area these breccias are related to karst collapse (see Chapter 5). However, in the All Night Lake area the breccias appear to have accumulated along minor syndepositional normal faults which show displacement of between 5 and 20 m (Figures 41 and 46). These breccias are interpreted to be fault-related talus breccias which subsequently were reworked during

transgression. Only two of the four breccia exposures can be demonstrated to lie along fault traces (exposures 1 and 2 in Figure 41). The larger of the two occurrences (exposure 1) in part downfills its fault zone and in part overlaps the downthrown block. The smaller of the breccias (exposure 2) is a fault-zone filling which occurs 20 m below the Kendall River Formation-Sulky Formation contact, which in turn is 90 m below the intra-Greenhorn Lakes disconformity surface (Figure 46). Kendall River Formation carbonates along the fault trace display small-scale open folds. The infill sediment is a mixture of dolostone fragments from the walls of the fault, and chert and quartz grains which have sifted down along the fault from the disconformity above. Localized folding and brecciation are interpreted to have occurred during fault movement, whereas sediment-downfilling and fault zone widening occurred during later solution-modification associated with the Greenhorn Lakes disconformity.

Disconformity Paleorelief

Assuming that the Dismal Lakes (Group-Coppermine River Group) contact represents a time plane, and knowing that the disconformity surface east of central Dismal Lakes was subaerially exposed (see above), the maximum paleorelief on the disconformity surface (relative to paleo-sea level) can be inferred by choosing the basal

Coppermine River Group contact as a horizontal datum (Figure 47).

Total relief above sea level is approximately 80 m over 108 km, a 0.00074 degree slope (Figure 47). But, the average slope on the disconformity between sections 15 and 16 (Figure 47) nearest the roof of the intrusion is 0.3 degrees. This slope is significantly anomalous when compared with both the average regional slope on the disconformity and typical depositional slopes on Recent carbonate platforms, which are typically less than a metre per kilometre. That this slope is significant is shown by the karst breccias formed on it, because regression not coupled to basement uplift presumably would have produced a shallow-dipping and widespread erosional surface. Thus, uplift or doming is implied by this reconstruction of paleorelief.

An abrupt local increase in uplift in the area of section 16 (Figure 47) with associated talus breccias and syndepositional faults delineates the area directly above the roof of the Muskox Intrusion north of All Night Lake (Figure 47). Preliminary evidence that the disconformity surface on the east side of the presumed paleokarst summit slopes eastward, also supports the contention that this region was the apex of doming (Figure 47). The spatial coincidence of maximum paleorelief and consequent karsting directly above the Intrusion strongly supports the interpretation that the relief developed during

Figure 47. Reconstruction of paleorelief along the Intra-Greenhorn Lakes Formation disconformity surface showing westward tilting of the surface, becoming particularly steep between sections 15 and 16. The dotted line is a continuous stromatolite marker that provides a secondary datum. The primary datum is the Coppermine River Group lower contact (see text for discussion). The western edge of the Muskox Intrusion would occur directly below section 16. Note that at section 7 the disconformity fades out and sediments above and below the surface become continuous. Thus, section 7 probably marks the western limit of regression and subaerial exposure. Outline map of the Dismal Lakes Group outcrop belt shows locations of the sections.

FIGURE 47

CORONATION
GULF

G.B.L.

doming was related to an early pulse of Muskox emplacement (see Figure 48).

Time Interval Between Doming and Muskox Emplacement

If the disconformity surface is assumed to be related to doming during an early pulse of Muskox emplacement, then the time interval separating this pulse from final shallow emplacement and associated extrusion (Coppermine River Group basalts) can be estimated. The maximum time interval separating the two events is represented by the time for accumulation of the thickest preserved interval of sediments separating the disconformity and the first Coppermine River Group flow. By extrapolating accumulation rates for Phanerozoic carbonates averaged from Schlager (1980: 100m/million years), an estimated 700,000 years is the period between these two events. Note that by using accumulation rates for ancient sequences, the diastem factor of Sando (1975) is accounted for as best as possible.

Without comparative data on rates of emplacement of similar ultramafic intrusions, it is difficult to assess the accuracy of this calculated time interval, and it should be considered only an order-of-magnitude figure.

Discussion

The relationship of local doming of Dismal Lakes strata to Muskox emplacement remains speculative. Data are not available on the east side of the Intrusion due

to the paucity of outcrops, but the disconformity discussed above extends to the northeast as far as Bathurst Inlet (Campbell, 1979; Kerans et al., 1981). Apparently the disconformity does not define a symmetrical dome about the intrusion. However, the relatively pronounced westward slope (see Figure 47), together with the local steepening directly over the intrusion, and minor local faulting, are best interpreted as a reflection of local doming related to emplacement of the Intrusion. As an extension of this hypothesis, it is conceivable that the eastern extension of the disconformity in the Bathurst Inlet area is related to regional-scale doming associated with the Mackenzie Igneous Event (Fahrig et al., 1981).

Irvine (1971) suggested that the funnel-like shape of the Muskox Intrusion indicates passive emplacement into an incipient graben which presumably would have involved both Wopmay basement and existing cover rocks (see Irvine, 1971, Figure 1). That the Intrusion was emplaced within a tensional stress regime is apparent from its basaltic composition (Irvine and Smith, 1967) and its apparent association with the geographically widespread Mackenzie dyke swarm. However, the graben hypothesis deserves reconsideration in view of the demonstrably post-Dismal Lakes Group age of the Intrusion. Any process of graben formation necessarily would be recorded in the cover sequence which now is

known to consist of both Hornby Bay and Dismal Lakes group strata. Although not exposed on the eastern margin of the Intrusion, these same rocks on the west show no evidence of the faulting, downwarping or extension typically associated with grabens.

Thus, known cover relationships do not support the graben hypothesis, but they do not exclude the possibility of minor doming of Coppermine Homocline cover during Intrusion emplacement.

The origin of the compositionally unique set of sills which intrude the Fort Confidence and Dease Lake Formations on the northwest margin of the Intrusion must be considered in light of the data presented here (Figure 45; Baragar, 1969; Baragar and Donaldson, 1973, p. 14). These sills are compositionally distinct from the Coppermine River Group rocks in that they are rich in micropegmatite (locally up to 20%) and contain ilmenite rather than magnetite as is typical of the Coppermine River Group lavas (Baragar, 1969). Allowing that the Muskox Intrusion is younger than the Fort Confidence and Dease Lake Formations, it now is possible to speculate that these sills were derived from the granophyre zone of the intrusion (Baragar, pers. comm., 1980), which would account for their distinctive composition. If these sills are related to the Muskox Intrusion, they provide additional evidence for forceful emplacement.

The combined evidence of syndepositional uplift,

faulting and erosion, and cross-cutting relationships suggest a five-part evolution of the cover sequence over the Muskox Intrusion, as follows: 1) deposition of Hornby Bay and Dismal Lakes groups Unit 8 - lower Greenhorn Lakes Formation, without consideration of the Hornby Bay-Dismal Lakes unconformity (see Figure 48a); 2) injection of a small sill-like precursor of the Muskox Intrusion (most likely along the Wopmay-Hornby Bay unconformity), causing slight local uplift, faulting and erosion of subaerially exposed Greenhorn Lakes sediments (Figure 48b); 3) deposition of upper Greenhorn Lakes sediments in a stable shallow-marine environment during transgression of the uplifted terrane (Kerans et al., 1981; see also Figure 48c); 4) emplacement of the main body of the Muskox Intrusion and associated sills and extrusion of Coppermine River Group flows (Figure 48d); and finally, 5) offset along the Canoe Lake Fault followed by erosion to the present surface (Figure 48e).

THE MACKENZIE IGNEOUS EVENT

The implications of results presented here to the nature of the Mackenzie magmatic event deserves consideration. The thickest part of the Coppermine River Group flood basalt sequence, and the most concentrated zone of the Mackenzie dyke swarm, are spatially associated with the Muskox Intrusion (Baragar, 1969; Baragar and Donaldson, 1973). The previous

Figure 48. Five-part summary of the emplacement history of the Muskox Intrusion. See text for discussion.
 C.R.G.- Coppermine River Group. GLF- Greenhorn Lakes Formation, KRF- Kendall River Formation, HB8- Hornby Bay Group, Unit 8, DLF- Dease Lake Formation.

interpretation that the Muskox Intrusion is significantly older than either the Coppermine River Group flows or the Mackenzie dykes obscured the importance of these correlations. Now the three igneous complexes may be viewed as components of a single major event with the Muskox Intrusion as a locus, as suggested by Robertson (1969) and by Fahrig and Jones (1969; see also Figure 42). The tectonic significance of this major regional magmatic event remains to be resolved (see Baragar, 1977; Patchett et al., 1978).

CONCLUSIONS

Reinterpretation of sandstone units in the vicinity of the roof zone of the Muskox Intrusion has led to the following conclusions: 1) The Canoe Lake Fault extends northeastward into the Dismal Lakes and Coppermine River Groups, displaying some 500 m of west-side-down displacement. This relationship effectively negates Irvine's (1970a) arguments for pre-Dismal Lakes Group origin of the Muskox Intrusion (see Hoffman, 1980); 2) The Muskox Intrusion cuts Unit 11 and the Fort Confidence Formation of the Dismal Lakes Group in the vicinity of the Muskox North Drill Hole, providing positive evidence for the post-Dismal Lakes age of the Intrusion.

Detailed mapping of the Dismal Lakes-Hornby Bay cover sequence has led to additional information relevant to tectonic events in the area. Local doming, most

pronounced above the roof of the Intrusion, is recorded in uppermost strata of the Dismal Lakes Group (Greenhorn Lakes Formation), reflecting an uplift event which preceded initial extrusion of the Coppermine River Group basalts, probably by several hundred thousand years. It is proposed here that this doming may reflect an early pulse in the emplacement of the Muskox Intrusion. An episode of gentle doming, implying sill-like emplacement of at least the initial magma body, is consistent with observed cover relationships.

The conclusion that the Muskox Intrusion is coeval, and most likely comagmatic with, the Coppermine River Group basalts and Mackenzie dyke swarm, provides clarification of a major magmatic event which occurred 1200 M.Y. ago, with a centre around the Muskox Intrusion.

CHAPTER 7. POST-DISMAL LAKES GROUP TECTONISM

The Dismal Lakes Group in particular, and the Coppermine Homocline as a whole, are remarkably devoid of structural complexities, making their area of occurrence ideal for lithostratigraphic and sedimentologic analyses of middle Proterozoic strata. Locally, however, three distinct but possibly related styles of deformation which affect Dismal Lakes Group units are recognized: 1) stratabound folding and brecciation of member C of the Dease Lake Formation; 2) linear zones of non-stratabound deformation involving member C of the Dease Lake Formation and all overlying units; and 3) fault-proximal folding.

STRATABOUND DEFORMATION OF DEASE FORMATION MEMBER C

Complexly deformed dolostones and mudstones of member C of the Dease Lake Formation stand out in marked contrast to otherwise gently dipping homoclinal strata of the Dismal Lakes Group. It is tempting to compare the member C deformed zone to the superficially similar Stark Megabreccia of Athapuscow Aulacogen (Hoffman, 1973) and the Omingmaktook Breccia of Kilohigok Basin (Cecile and Campbell, 1977). All three are lithologically selective, involving mudstones that contain evidence of relic evaporites. On the basis of preliminary examination of the member C breccias however, a different stress

mechanism for these breccias is inferred.

Stratigraphic Relationships:

Member C of the Dease Lake Formation displays a high mudstone-dolostone ratio along and west of Dease Lake, and also shows evidence of relic evaporites in the form of silicified bedded and nodular anhydrite (Plate 19F). Member C is gradationally underlain by member B of the Dease Lake Formation, which also is composed of mudstones and dolostones. Overlying member C are medium-bedded dolostones of the Kendall River Formation, and massive stromatolitic dolostones of the Sulky Formation.

Member C is the first mudstone-rich lithology below the massive and medium-bedded dolostones of the middle Dismal Lakes Group. In this sense the intense deformation of member C is analogous to selective decollement in shale-rich portions of the Rocknest Formation below a massive stromatolitic member (Tirrul, 1982; Grotzinger, 1982), and to shales which underlie structurally competent Mississippian carbonates in the northern Canadian Rockies (Thompson, 1979).

Both the lower and upper contacts of the deformed zone are sharp. Locally, strata of member B which directly underlie the breccia are folded into conjugate and box folds, but more commonly the basal breccia of member C rests sharply on unaffected member B strata (Plate 13A). The upper contact of the deformed zone with

the Kendall River Formation is also sharp. Locally, breccia-filled dykelets originating in the upper breccia intrude the otherwise undeformed overlying Kendall River dolostones. Where observed, the contact is undulose, defining a gentle dome and basin topography with hills of massive breccia surrounded by gently dipping dolostones (Plate 13F).

Areal Distribution:

Deformation in member C is restricted to areas west of the east end of Dease Lake (Figure 50). Along the north shore of Dease Lake, deformation consists of a stratiform breccia zone which separates otherwise undeformed members B and C. The breccia averages 2 m, thickening slightly westward. This basal breccia, (marked as a heavy black line in Figure 50), is nearly continuous from Dease Lake to the western outcrop limit. Approximately 3 km northwest of Dease Lake, the deformation zone is fully developed and well exposed. Here, member C is pervasively folded, and upper member B is deformed into distinctive box and conjugate folds (Plate 13B). Westward from this area, member C is everywhere pervasively deformed. The upper breccia and the contact with the Kendall River Formation (Figure 49) are well exposed in a gentle dome 2.5 km south of Sulky Lake, and in the hinge of an open anticline 4 km northwest of Sulky Lake (Figure 50). Additional

Figure 49. Schematic cross-section of structural subdivisions of the member C deformed zone. Vertical section represented is roughly 150m. B- member B of Dease Lake Formation, C1- basal breccia of member C of the Dease Lake Formation, C2- recumbently folded division of member C, C3- upright folded division of member C, C4- upper breccia of member C, K- Kendall River Formation. Keyed boxes refer to close-up illustrations in this chapter. See text for discussion.

exposures of the upper contact not recognized during mapping for this study may be present along the south shore of Greenhorn Lakes.

The stratabound deformation zone is restricted to a 20 km-wide belt which parallels the East River Fault (ERF; Figure 50). Member C displays only minor folding north of Falcon Lake Fault; however, the position of outcrops of member C north of this fault make it unclear whether this decrease in deformation implies a position on the eastern or northern margin of the deformed zone (Figure 50).

Structural Divisions of the Deformed Zone:

In areas where the deformed zone is well developed, five structural divisions can be recognized (Figure 49). These divisions are: 1) uppermost conjugate folded member B; 2) basal member C breccia (C1); 3) recumbently folded member C (C2); 4) upright tight to isoclinally folded member C (C3); and 5) upper member C breccia (C4) (Plate 13, A-E).

Uppermost Conjugate-Folded Member B

Folding of uppermost member B occurs only west of Dease Lake in areas where the deformed zone is fully developed. The fold style is distinct from that of overlying deformed member C (compare Plates 13B and C). Folds are of conjugate and box geometries, and occur on both meso- and macroscales (Plate 13B). Folds are

generally open, and shortening of greater than 50 % is rare. The few axial planes measured have variable orientations, but fold axes are consistently subhorizontal.

Basal Member C Breccia (C1)

The 2 m basal breccia is the most laterally extensive division of the deformed zone, occurring from eastern Dease lake to Dease Arm of Great Bear Lake. The C1 breccia rests on relatively undeformed member B strata. The contact is typically sharp (Plate 13A), and a parallel fabric is developed locally in the basal 40 cm (Plate 14A). The breccia is monomict, consisting of angular clasts of light grey dololomite in a matrix of pink-grey mudstone (typical member C lithologies). Clasts range from 1 m to granule size, and occur either dispersed in the mudstone matrix or in grain-to-grain contact. Many of the clasts retain visible bedding, and some display fold closures (Plate 14B). These folds, and most dololomite fragments, have a preferred orientation parallel to the basal contact with member B.

Although all detached folds in the breccia, and folds in the overlying recumbent C2 division, are of parallel style, brittle deformation in the form of jigsaw-type internal brecciation is common within the folded beds (Plates 14C and D). The dolostone bed in Plate 14C is essentially a parallel-folded, brecciated

dolostone.

Within the recessive-weathering mudstone matrix are zones of hackly-weathering silica, which morphologically resemble sulphate evaporites. Only two of these siliceous zones were sampled from the upper C4 breccia, but thin sections of both contain inclusions of anhydrite. Such silica pseudomorphs are common in all portions of member C west of Dease Lake.

Recumbently Folded Member C (C2)

A zone of recumbent folds defined by resistant-weathering light grey dololomite beds enveloped in pink-grey mudstone gradationally overlies the C1 breccia. This C2 division is developed from northwest of Dease Lake to Dease Arm of Great Bear Lake. The transition from recumbent folds to upright folds of the succeeding C3 division is not exposed, but in the principal outcrop area northwest of Dease Lake (Figure 50), a minimum thickness of 10 m is indicated.

Folds are typically recumbent isoclines essentially lacking cleavage (Plate 13C). Thickening of fold hinges is minor and irregular, and fold style is best described as parallel. Dolostone beds which define the fold geometries are 20-40 cm thick. Due to extreme flowage, the thickness of enveloping mudstone interlayers is impossible to estimate in areas of folding, but by analogy with undeformed equivalents at Dease Lake, these.

mudstone layers probably were 40-60 cm thick. Internal brecciation of dolostone beds is rare, in contrast to the common internal brecciation of detached dolostone folds in the underlying C1 breccia.

Fold amplitudes were not measurable at outcrop scale due to sparse exposure, but no large-scale nappe structures are suspected, due to the limited (10-20 m) thickness of this zone.

Upright Tight to Isoclinally Folded Member C (C3)

In the area between Dease and Greenhorn Lakes and in the vicinity of Sulky Lake, large areas of upright tight-to-isoclinally folded dololutes occur (Plate 13D). Lithologically this division contains much less mudstone, suggesting that it involves stratigraphically higher portions of member C (transitional upwards into clean dolostones of the Kendall River Formation). Dololite beds are 40-60 cm thick. In several areas this division rests structurally above the C2 division. The change from horizontal to vertical folds between the C2 and C3 divisions is not exposed sufficiently to determine whether recumbent folds have been refolded into an upright orientation, or whether there is a sharp break between the divisions. The upright folds pass gradationally upward through a zone of internally brecciated folds (Plate 13E), into the upper (C4) breccia.

Most structures observed are of mesoscopic scale, but the lack of reliable top indicators in the dololomite beds prohibits recognition of larger-scale structures which may be present. Fold axes are subhorizontal and the dominant axial trend is east-southeast (Figure 50).

Upper Member C Breccia (C4)

The deformed zone is capped by an upper breccia which is well exposed south and northwest of Sulky Lake in the areas where its upper contact with the Kendall River Formation occurs. This breccia zone is at least 30 m thick south of Sulky Lake. The breccia is massive and is composed of monomict dolomite clasts in a clay-rich micrite matrix. This matrix also contains silicified anhydrite remains, as mentioned above.

The breccia is gradational below, passing through a zone of internally brecciated folds (Plate 13E) into upright folds of the C3 division. South of Sulky Lake the contact of the C4 breccia with the Kendall River Formation is sharp (Plate 13F), but in the area northwest of Sulky Lake, blocks of undeformed dolomite, with bedding parallel to the upper contact, are intercalated with the breccia. "Block" margins cannot be defined on an outcrop scale, and they may represent structurally isolated but essentially in situ remnants of undeformed uppermost Dease Lake Formation, transitional into overlying undeformed Kendall River Formation.

(schematically represented in Figure 49). Locally at the upper contact with the Kendall River Formation, dykelets of C4 breccia cut up into Kendall River dolostones, but otherwise the contact is sharp, much like the member B/C1 contact.

Fold Orientation Data:

Measurements of fold axes were taken systematically only throughout the deformed zone. These measurements, separated into three domains, are summarized in three stereonet plots (Figure 50). Domain 1 is considered unique because of its proximity to a northwest-trending portion of the ERF, and fold axes are oriented in a south-southeast direction. Domains 2 and 3 represent the two main outcrop areas of the deformed zone, in the Sulky-Great Bear Lake and Greenhorn-Dease Lake areas respectively. Both plots include measurements for both C2 recumbent and C3 upright fold axes. Although measurements do not cluster tightly, most fold axes have shallow plunges to the south-southeast. This trend is parallel to the ERF and to large-scale non-stratabound folds which occur near the ERF along the north shore of Dease Arm (Figure 50).

Origin and Significance of Stratabound Deformation:

Origin of C1 and C4 Breccias

Two possibilities are considered for the origin of the breccia divisions: brittle deformation caused by

lateral shear stresses, and evaporite-related solution collapse. A shear stress brecciation mechanism with undeformed upper Dismal Lakes Group strata being pushed, or sliding over underlying units, making the deformed zone essentially a broad decollement zone, is favoured for the following reasons;

a) First, detached folds within the breccia maintain a consistent orientation, with axial planes parallel to bounding surfaces. Such a preferred fabric suggests orientation within a directed stress field. A process involving initial folding and internal brecciation (Plate 14C), followed by continued shear and gradual limb disaggregation (Plate 14B), would supply isolated breccia clasts, leaving isolated fold remnants. In addition, the strong preferred orientation of all clasts, with the plane of flattening parallel to the bounding surfaces of the brecciated units, is more likely to be a product of orientation within a stress field. Solution collapse should give a random clast orientation.

b) Secondly, collapse breccias typically display clast-support textures, but the C1 and C4 breccias contain matrix-supported zones as well as clast-supported zones.

c) Finally, association of these breccia zones with folded dolomite divisions which exhibit significant

shortening, is consistent with a lateral shear stress origin for the observed brecciation.

The combined evidence listed above implies a tectonic, shear stress-generated origin for the breccias.

Generation of Fold Divisions C2, C3 and Uppermost Member

B

Layer-parallel shortening is the best mechanism to explain the various fold styles of the deformed zone. In contrast to the random orientation of fold-axes orientations expected in response to solution collapse of semiconsolidated sediments, fold trends in the deformed zone are generally constant throughout (Figure 50). In addition, the stratigraphy within the deformed zone is essentially identical to that seen in undeformed areas north of the eastern portion of Dease Lake. This still-coherent stratigraphy rules out any large-scale removal of section required for collapse-related deformation.

Some additional constraints can be placed on the nature of folding and related deformation. The parallel geometry and the absence of cleavage in folds of the deformed zone creates an appearance similar to soft-sediment slump folds. However, the brittle internal brecciation of some dololutite layers (Plates 13E and 14A) indicates that sediments must have been partially lithified before deformation. The combination of a high

pore-fluid pressure, and a significant ductility contrast between unconsolidated mudstone and semilithified dolostone would permit this unusual combination of ductile and brittle deformation.

The amount of lateral translation to produce the observed strain must have been significant. Little can be said about the amount of translation related either to the breccia or to the recumbently folded division, but the C3 upright, isoclinal division is coherent, and therefore the degree of shortening for it can be estimated. The C3 division outcrops for 20 km normal to the direction of shortening (northward), and represents a shortening of 12 km if 60 percent shortening is assumed. This estimate of lateral translation is conservative, because it considers only the exposed portion of the deformed zone. When applied to the entire deformed zone, this estimate appears even more conservative because it does not account for shear stress accommodated by either the breccia or the recumbent divisions.

Timing of Deformation

The absence of onlap relationships, and the presence of crosscutting breccia dykelets at the deformed zone-Kendall River Formation contact, support a post-lower Kendall River timing of deformation. If deformation had occurred after deposition of lower Kendall River Formation carbonates, but before

termination of Dismal Lakes Group deposition, some interruption in deposition should probably have been recorded within this sensitive carbonate platform succession. Because no such interruption is recorded, deformation is inferred to postdate Dismal Lakes Group deposition.

Assuming a post-Dismal Lakes Group- pre-Coppermine River Group timing of deformation, a minimum cover of 1.3 km must have been present over the deformed zone at the time of deformation. If deformation postdated extrusion of Coppermine River Group flows, the thickness of the cover sequence would have been substantially greater.

The Deformed Zone as a Decollement, and the Origin of Shear Stress

In essence, the stratabound deformation zone of member C of the Dease Lake Formation represents a diffuse zone of decollement across which a minimum of 12 km lateral slip has been transmitted. Deformation was selective to this zone because of high pore-fluid pressures (related to high clay content of the unit) and to the presence of possibly significant amounts of evaporites. Both evaporite-rich and clay-rich units are common zones of detachment in other areas (cf. Tirrul, 1982; Thompson, 1979; Lemoine, 1973).

The two main questions with regard to origin of the stress regime are: 1) Was the lateral translation caused

by a push from the side or was it related to gravity sliding?, and; 2) Can the source of the push or slide be related to the spatially proximal East River Fault?

Similar styles of deformation have been recorded from the classic gravity slide areas of the Alps (Lemoine, 1973; Debelmas and Kenakhove, 1973) and from typical fold-thrust belts. The attraction of the gravity model is that it explains the deformation by sliding which requires only minor tilting to initiate. Such tilting could be caused by normal fault activity which was the typical style of post-Dismal Lakes Group tectonism.

Alternatively, the ERF may be interpreted as a thrust or reverse fault, by reason of extensive folding and compressional deformation proximal to the fault (Figure 50). Such extensive deformation is more typical of thrust or reverse faults than of normal faults. If the ERF is a reverse fault, then this fault is a likely candidate to have provided the initial push which caused upper Dismal Lakes strata to detach and migrate along the member C decollement.

The possibility exists that stratabound member C deformation is related to the ERF. In the gravity slide model, the south-side-up displacement of the ERF would have caused tilting prior to faulting, resulting in northward sliding of upper Dismal Lakes units. In the thrust model, northward compression related to thrusting

would have applied lateral stress to the Dismal Lakes succession north of the fault, causing detachment in the weakest part of the section (the clay- and evaporite-bearing member C strata). In either case, because the stratabound deformed zone is folded (along with the remainder of the Dismal Lakes and Hornby Bay groups) by drag folds along the north side of the ERF (Figure 50), stratabound deformation can be inferred to have predated major movement and folding related to offset of the ERF. Thus, if the ERF caused the stratabound deformation, then this deformation must have occurred early. Equally likely, the source of the "push" may be related to some feature no longer exposed.

Summary:

Stratabound deformation of member C, and minor folding in uppermost member B of the Dease Lake Formation, appear related to lateral translation of Kendall River and overlying strata, relative to the underlying Dismal Lakes and Hornby Bay units. Lateral motion was north or south, parallel to the axis of maximum shortening, and was on the order of 12 km. Further investigation of fold geometry is required to determine the transport vector. Selective detachment in member C is related to high clay content, which caused high pore-fluid pressure, and to the presence of evaporites. Both served to lubricate this diffuse zone of decollement by reducing friction due

to grain-to-grain interaction. High pore-fluid pressures also are required to explain the absence of cleavage in this zone.

Data concerning the cause of shear stress (whether due to gravity sliding or to tectonic compression), and the possible relationship to the ERF, thus are inconclusive.

LINEAR NON-STRATIFORM DEFORMATION ZONES

Two north-trending linear zones of intense deformation occur north of the east and west ends of Dease Lake. In contrast to the stratabound deformation described above, these zones of deformation involve member C and all overlying units exposed in this zone, specifically the Kendall River Formation and member A of the Sulky Formation. Units below the C1 breccia are not involved in this deformation. An additional contrast between linear and stratabound zones is that penetrative cleavage is developed in the former but not in the latter (Plate 14E).

Neither of the linear deformed zones can be related to fault lineaments. Both belts stand as areas of positive relief above the surrounding and underlying homoclinal succession. Member C strata within these linear ridges occur topographically and structurally higher than any surrounding equivalent undeformed exposures of this unit.

Interpretation:

The origin of the linear deformed zones cannot be related to simple fault-parallel folding, because deformation is selective to units above the C1 breccia. No fault offset or other deformation occurs in underlying units. This relationship suggests a genetic association of these linear deformed zones with the stratabound deformation zone. Furthermore, although faulting has not occurred, member C strata have been structurally uplifted in these zones relative to the surrounding undeformed equivalents. Such a relationship, without offset of underlying units, suggests injection or intrusion of these sediments upward, possibly along fractures in the cover sequence which has since been eroded.

The development of penetrative cleavage in the linear deformation zones, in contrast to the stratabound zones where cleavage is notably absent, implies a difference in pore-fluid pressure between these zones. Perhaps these linear deformation zones developed by fluid escape from overpressured stratabound portions of member C. Fluid escape along these zones might explain the difference in pore-fluid pressure, and the upward injection of member C strata. The linear nature of these zones might be related to fractures in the cover sequence which would not necessarily be present in strata below the C1 breccia. Similar, though much more dramatic features, are mud volcanoes such as those forming today

(cf. Snead, 1964; Higgins and Saunders, 1974) and those described by Cecile and Campbell (1977) from the Proterozoic Goulburn Group.

FAULT-RELATED FOLDING

Characteristics of fault-related folding, which are typified by the area paralleling the ERF on the north shore of Dease Arm (Figure 50) are:

- axes are subparallel to fault planes and are subhorizontal
- folds are not lithologically or stratigraphically selective
- a penetrative cleavage is generally developed in association with folding
- folds are only developed in areas proximal to major faults

Other areas aside from Dease Arm which display significant folding related to faults are shown in Figure 51.

All faults which display fault-marginal folding crosscut and postdate all units which they intersect, including the Coppermine River Group (cf. Herb Dixon and Teshierpi Faults; Figure 51).

Non-selective, penetrative fault-marginal folds with axes parallel to fault planes, and with subhorizontal

Figure 51. Map showing areas of significant fault-related folding within the Coppermine Homocline (arrows).
 ERF- East River Fault, FLF- Falcon Lake Fault,
 HDF- Herb Dixon Fault, SG- St. Germain Graben,
 SF- Sinister Fault.

axes, are considered here to be drag folds. The geometry of these folds implies dip-slip, and not strike-slip motion for the faults, but insufficient analysis has been conducted to constrain the vector of motion along the faults. Because fault-plane dips are not known for any of the faults in Figure 51, the normal, reverse, or thrust character of these faults cannot be established. Until now, all late-stage faulting in the Coppermine Homocline has been assumed to be of normal displacement, but more detailed examination of fold geometries and minor structures associated with faults (such as the East River Fault) may result in recognition of at least some reverse movement (as suggested in Figure 50).

PLATE 13. STRATIBOUND DEFORMATION ZONE OF DEASE LAKE
FORMATION

A. C1 breccia resting sharply on muddy dolostones of member B of the Dease Lake Formation. Stratigraphic position within the deformed zone is shown in Figure 49. Hammer for scale.

B. Conjugate and box fold geometries within uppermost member B, just beneath the C1 breccia. The upper 1 m of the cliff face and the rocks capping the hill are the C1 breccia. Cliff face is approximately 12 m high. Stratigraphic position shown in Figure 49.

C. Recumbent isoclinal folds of C2 division marked by resistant dololutite beds within fissile light-coloured mudstone. Dololutite beds lack cleavage. Fold closures can be seen within the mudstone in the top-center portion of the photo. Stratigraphic position shown in Figure 49.

D. Small (50 cm high) outcrop of upright folds in C3 division of thinly bedded dololutites and mudstones. Closure indicated by arrow. Stratigraphic position shown in Figure 49.

E. Internal brecciation in fold in upper portion of C3 zone which is transitional into C4 breccia. Although individual dololutite layers show intense brecciation, fold geometry is intact. Stratigraphic position shown in Figure 49.

F. Upper contact of deformed zone. Gently dipping undeformed Kendall River Formation dolostones are in tectonic contact above rounded hill of massive C4 division upper breccia. Kendall River dolostone section is approximately 2.5 m thick. Stratigraphic position shown in Figure 49.

4

PLATE 13

PLATE 13

A

B

C

D

E

F

**PLATE 14. STRATIBOUND DEFORMATION ZONE OF DEASE LAKE
FORMATION**

A. Close-up of the basal contact of the C1 breccia and underlying member B of the Dease Lake Formation (arrows) showing development of a parallel fabric within the breccia. Pencil near centre of photo is 17 cm long.

B. Truncated fold limb of detached fold in the upper portion of C1 breccia zone. Dololomite layers are surrounded by mudstone-rich breccia matrix. Notebook is 17 cm long.

C. Detached fold within C1 breccia showing parallel, recumbent geometry and absence of cleavage. Internally, the dololomite layer is extensively brecciated (see D of this plate: location is marked by the boxed area). Notebook is 17 cm long.

D. Close-up of dololomite layer shown in Figure C, showing detail of brecciation of this layer. Breccia fabric is angular and of jigsaw type. Long dimension of photo is 15 cm.

E. Penetrative cleavage developed in limb of an upright isocline of member C of the Dease Lake Formation. Photo taken within the western linear deformed zone (see Figure 50). Hammer for scale.

PLATE 14

PLATE 14

A

B

C

D

E

SUMMARY OF TECTONO-DEPOSITIONAL HISTORY OF DISMAL LAKES
GROUP

Strata of the Dismal Lakes Group formed on a stable cratonic basement of uplifted and eroded rocks of the Wopmay Orogen, and both conformably and unconformably overlie Hornby Bay Group sediments. Dismal Lakes Group deposition was the result of a second major siliciclastic-carboante depositional cycle, initiated by renewed uplift and faulting in the Dismal Lakes area. Unit 10 of the Hornby Bay Group and Unit 11 of the Dismal Lakes Group form the basal coarsening-upward portion of the second siliciclastic phase.

Fort Confidence Formation sandstone and shale lithofacies formed by a combination of fluvial and marine processes acting on a marginal-marine fan delta. The Fort Confidence Formation fines upward from Unit 11 mixed fluvial and marine facies, and records stabilization and gradual eastward transgression. Dease Lake Formation member A quartzose dolostones, which are restricted to the area west of Leith Line, formed by continued eastward transgression and reworking of the Fort Confidence delta. Continental and coastal sabkha lithofacies of Dease Lake members B and C display less lateral variation, and are dominated by silt and clay-sized siliciclastics. They also contain an upward-increasing dolostone component. These trends signal termination of the siliciclastic.

phase of deposition, and set the stage for deposition of stable carbonate platform sediments.

Kendall River Formation peritidal dolostones show westward thickening across Leith Line, and minor westward and northwestward deepening to a predominantly subtidal regime, but essentially represent deposits of an extremely level and stable peritidal platform.

Lithofacies variation is restricted to the scale of individual beds, reflecting control by microenvironmental topography (e.g. tidal channels and levees). The shallowing-up cyclicity so typical of continental margin peritidal carbonates is largely absent from this formation.

Sulky and Greenhorn Lakes Formation carbonates form the upper two units of the carbonate phase of the siliciclastic-to-carbonate cycle of the Dismal Lakes Group. In contrast to Kendall River carbonates, however, differential subsidence, both gradual and rapid, local and regional, played an important role in controlling vertical and lateral lithofacies distribution.

The platform-drowning event which resulted in deposition of black shales and deeper-water carbonates on the Kendall River member C biostrome is the first major eustatic/tectonic event of the carbonate phase.

Coincident with rapid transgression during early Sulky Formation deposition was differentiation of the heretofore homogeneous platform into the slowly subsiding

September Lake High and the more rapidly subsiding surrounding platform. This differential subsidence is reflected by localization of the September Lake buildup (member B) and the overlying buildup-crest evaporite-pisolite-tepee association of member C.

Greenhorn Lakes deposition is dominated by repeated transgressive-regressive pulses recorded by the grey dolostone-biostrome cycles of member B, and by the shale-biostrome couplet of member C. As for Sulky deposition, the September Lake High was a positive feature, containing no record of these early transgressive pulses within its remarkably uniform peritidal section. Subtle changes in the tectono-depositional regime are the westward shift of the area of maximum subsidence, from Leith Line to west of Bebensee Lake, and change from a west- to a southwest-inclined paleoslope west of Bebensee Lake. These changes are first apparent in member B, and persist throughout remaining Dismal Lakes Group units.

Localized uplift in the September Lake area at the end of member D deposition caused karsting and erosion of the platform in this area, while west of Dismal Lakes, deposition was continuous. This uplift/karst event, and minor associated syndepositional faulting, is most pronounced over the roof of the Muskox Intrusion, and it seems likely that this interruption in deposition reflects an early pulse of emplacement of the Intrusion.

Marine transgression followed, depositing a basal siliciclastic fining-up sequence which is succeeded by renewed peritidal dolostone deposition (member E). During this final period of dolostone deposition, peritidal environments persisted throughout the area, demonstrating that the platform had once again returned to a homogeneously subsiding shelf area.

Extrusion of Coppermine River Group basalts abruptly terminated dolostone deposition. These basalts, and associated Mackenzie diabase dykes and MuskoX Intrusion, comprise intrusions (broadly characterized as the "Mackenzie event") emplaced in an east-west tensional stress regime at 1.2 B. Y..

Post-Dismal Lakes Group stratabound deformation of member C of the Dease Lake Formation was initiated either by uplift and gravity sliding, or by a lateral push related to movement along the East River Fault. Due to retained high pore-fluid pressure within member C, this unit acted as a zone of decollement, allowing overlying Kendall River, Sulky and Greenhorn Lakes Formations to glide over underlying Dease Lake and lower units without significant deformation. However, along north-trending linear zones north of Dease Lake, strata overlying the member C decollement zone were deformed when fluids from this overpressured zone escaped upward.

After stratabound and linear deformation, movement

along several fundamental faults, most notably the East River, Sinister and Falcon Lake Faults, caused fault-localized folding of Hornby Bay and Dismal Lakes Group units. This folding also involved the stratabound deformation zone, demonstrating relative timing of these events.

REGIONAL CORRELATION

Dismal Lakes Group strata are closely correlative with the Ellice and Parry Bay Formations of the Bathurst Inlet area (Kerans et al., 1981), suggesting that these areas represent erosional remnants of a once extensive Helikian shelf which rimmed the northwestern edge of the shield, having a gross northerly paleoslope (Donaldson et al., 1982). This correlation has been based on closely comparable lithostratigraphy and unconformities (Kerans et al., 1981).

Attempts have been made to correlate Dismal Lakes and Hornby Bay Group strata with other isolated Helikian sequences in the Athabasca and Thelon Basins (Figure 1; Fraser et al., 1970), the Mackenzie Mountains (Young et al., 1979), and other parts of the North American Cordillera (Donaldson, 1973). Although rock types in the former two areas bear close resemblance to the Hornby Bay-Dismal Lakes succession, it also is possible that each area behaved as an independent depocenter. Detailed paleomagnetic and/or geochronologic studies in these

areas should increase reliability of the proposed correlations, but still will not solve the problem of whether the basins were coextensive or separate depocentres. A potentially more useful approach toward correlation between these isolated outcrop areas is the comparison of transgressive-regressive cycle charts. Donaldson et al. (1982) showed that the maximum transgression in the Dismal Lakes Group was rapid, causing platform drowning in both the Coppermine and Bathurst Inlet areas. This transgression was recorded in both areas by deeper-water shale and dolostone deposition followed by development of a distinctive conical stromatolite biostrome (Sulky Formation members B and C; Parry Bay Formation, member P3b). Such maximum transgressive events, particularly if rapid, may represent eustatic or tectonic events of widespread importance, and should be considered primary elements of correlation.

In fact a closer examination of the carbonate unit of the Athabasca Basin, the Carswell Formation, has revealed the presence of conical stromatolites there which are nearly identical to those of the Sulky-Parry Bay biostrome (K. Wheatley, pers. comm., 1982). Although no positive correlations can be made on this basis, it can be said that if the Carswell dolostone is correlative with any unit in the Coppermine Homocline, the most likely candidate would be that unit formed during maximum

transgression, the Sulky Formation. Thus a Carswell-Sulky Formation correlation would be consistent with eustatic and "biostratigraphic" information.

Correlation of the Dismal Lakes Group with the grossly time-equivalent Wernecke Supergroup (Young et al., 1979; Delany, 1981) is not yet feasible, due to the lack of detailed tectono-depositional information for this unit.

ASPECTS OF THE DIA GENETIC HISTORY OF THE DISMAL LAKES
GROUP

CHAPTER 8. DIAGENESIS OF THE SULKY FORMATION

Recent studies of penecontemporaneous or early diagenetic processes and products (see reviews in Bathurst, 1975; Milliman, 1974) have shown these fabrics to be useful for deciphering the syndepositional and early post-depositional history of carbonate sediments. The Sulky Formation contains a variety of well-preserved diagenetic fabrics which can be used to delineate the approximate syn- to post-depositional history of these sediments during the different depositional stages outlined in Chapter 4.

This study is based on textural data from polished slabs and thin sections studied under transmitted light and cathodoluminescence. Inferences regarding original chemical compositions, whether calcite, high-Mg calcite, aragonite or dolomite are based on morphological comparisons with modern cements. Present mineralogical distinctions were made using the staining technique devised by Dickson (1965) for calcite, ferroan calcite, dolomite and ferroan dolomite. More definitive interpretations of primary mineralogy must be based on detailed trace element and isotope analyses (cf. Davies, 1977; Walls et al., 1978; Meyers and James, 1978).

MEMBER A

Only minor diagenetic alteration of ~~water~~-water Member A (lithofacies A) micrites has been observed.

This can be seen in zones of localized cementation which display greater resistance to compaction, demonstrating their pre-compaction origin. Nodular cementation is typical of deep-water mudstones, commonly occurs quite early. None of the nodules have been reworked into succeeding deposits, such as occurs on the modern Bahamian slope (Mullins et al., 1980) or in slope facies of the Belcher Group (McEwen et al., 1978).

A second form of cementation is the differential lithification of dark and light grey interbeds of lithofacies A4. Recognition of this differential cementation is enhanced by syn-sedimentary slumping which caused dark grey beds to deform in a brittle manner, whereas light grey beds were deformed plastically (Plate 6F).

MEMBER B

Cementation:

Fibrous cement is the second most abundant component next to stromatolites in Member B of the Sulky Formation. The fibrous cement occurs as isopachous fringes on stromatolites, intergrown with stromatolitic laminae, and intercalated with micritic sediment of presumed marine origin. Such relationships are indicative of a submarine environment of precipitation for these fibrous cements. The various fibrous cement textures and succeeding pore-filling spar cements are described systematically

below.

Length-Fast Zebra Dolomite

Length-fast zebraic dolomite (LFZD), where it is present, consistently is the first generation cement on stromatolitic laminae. LFZD forms fibrous isopachous and botryoidal fringes, and is succeeded by laminar fibrous dolomite (LFD), blocky zoned ferroan dolomite spar, algal lamination or micritic sediment (Figure 53). LFZD has an irregular 0.1-0.2 mm scale banding which is visible with the unaided eye (Plate 15A), reflecting alternating fibrous and micritic bands which approximately parallel the substrate boundary (Plate 15A and 15B). The fibrous bands consist of 80x20u bladed crystals with slightly consertal boundaries (Figure 52). Bladed crystals are neomorphic after a fibrous precursor (= neospar of Mazzulo, 1977; 1980), with individual crystals displaying a sweeping extinction pattern, presumably inherited from the fibrous precursor. Inclusions within ray crystals impart a psuedo-pleochroism (Plate 15D; cf. Kendall and Broughton, 1978).

Micritic bands are composed of 4-10u crystals with irregular interlocking boundaries. Relic bladed crystal patches contained within micritic bands display an extinction pattern consistent with surrounding bladed bands (Figure 52). Micritic bands display a random

Figure 52. Schematic sketch of (A) alternating relic fibrous (now ray crystal), and (B) micritic bands of LFZD fabric under crossed nicols. Long axes of ray crystals are oriented north-south. The extinction pattern is oblique rather than parallel to crystal long axes. Relic fibrous zones in micritic bands (C) display similar extinction pattern. Ray crystal boundaries are consertal. Bar scale is 0.02mm.

Figure 53. Sequences of diagenetic pore-filling events for member B: 1) algal lamination. 2) precipitation of LFZD cement on algal lamination. 3a) alternating growth of algal lamination and LFZD. 3b) alternating micritic sediment and LFZD. 4) precipitation of LFD on algal lamination or LFZD. 5) precipitation of zoned ferroan dolomite on LFD, LFZD, micritic sediment or algal lamination. 6) precipitation of clear spar, euhedral quartz (Q), and hydrocarbon infill (H).

FIGURE 53

extinction pattern and are non-pleochroic.

The optic orientation of relic LFZD fibres may be analogous to the zebra habit of twisted quartz fibres (McBride and Folk, 1977; Frondel, 1978) in that a single fibrous band, oriented N-S or E-W, will not be totally extinct or illuminated, but rather will display a jagged extinction pattern (Figure 52; Plate 15B). This extinction pattern results from the twisting of original fibres or their constituent crystallites (cf. Frondel, *op. cit.*). Local variations in crystallite orientation can produce a length-slow extinction pattern, with the extinction angle being perpendicular to the fibre long axes. These length-slow zones also display higher relief than neighboring length-fast textures (Plate 15C).

The banded texture of LFZD is regularly spaced, but it is not incremental, or continuous within one cavity (i.e. it is not possible to trace individual fibrous or micritic bands around the perimeter of a pore). In fact, boundaries between micritic and fibrous bands are extremely jagged (Figure 52) and bands commonly form anastomosing patterns (Plate 15B). Zoning by fluctuating chemical composition of pore fluids in a phreatic environment (required by isopachous rim geometry) would produce bands or growth increments of an even thickness which would be continuous throughout the pore system.

The relic patches of ray crystals within micritic bands suggests that the micritic bands originated by

preferential neomorphism, possibly related to greater strain contained within original fibres due to crystal dislocations. Folk and Assereto's (1976) coconut meat calcite texture (their Figure 1, zone 1) displays similar jagged banding in plane-polarized light. That this feature is visible in a Holocene speleothem suggests that it may control later neomorphism, eventually resulting in a texture similar to the fibrous-micritic banding of LFZD. The origin of this jagged light and dark banding is not interpreted by Folk and Assereto; however, it is clear from their illustration (Folk and Assereto, 1976; their Figure 1, zone 1) that it is not related to growth increment banding which is smooth, parallel and evenly spaced.

Laminar-Fibrous Dolomite

Laminar fibrous dolomite (LFD) occurs as light tan translucent isopachous fringes over either stromatolitic laminae or LFZD (Figure 53; Plate 15D). LFD is evenly banded to unbanded (banding visible in hand specimen), in marked contrast to the jagged light (cloudy micritic) and dark (translucent fibrous) banding of LFZD. Ray crystals of LFD are elongate (8:1 to 10:1), perpendicular to the substrate, and average 500x50 μ . Crystals are length-fast with extinction slightly divergent away from the substrate (Plate 15D). Boundaries between ray crystals are consertal and locally micritized (Plate

15D). This fabric is similar to relic fibrous cement of Mazzullo (1977, his Figure 54) and to Mg-calcite palisade cement described by Schroeder (1972). Both are interpreted as submarine in origin.

Zoned Ferroan Dolomite

Coarse, euhedral, slightly cloudy dolomite spar succeeds either LFZD or LFD as a third cement type (Plate 16A). Rhombs are 1-2 mm across, increasing slightly in grain size away from pore walls. Locally crystals are epitaxial over LFD. This relationship has not been noted for LFZD. Dolomite rhombs contain either one or two zones of iron oxide-calcite aggregates which mimic crystal growth zones. These zones likely formed by dedolomitization of particularly Fe-rich dolomite growth bands (cf. Frank, 1981). Rhombs also display multiple luminescent bands which are related to fluctuations in concentration of the trace elements Mn and Fe. According to Fairchild (1980), the ratio between Fe and Mn is the factor that controls luminescence.

Localized dedolomitization of early growth zones, succeeded by unaltered growth zones, indicates that crystal growth was interrupted by solution (cf. Frank, 1981; Fairchild, 1980; Katz, 1971). Sulky Formation Fe-zoned spar also fills fractures which cut all previous fabrics, indicating a relatively late, post-lithification origin.

Limpid Dolomite

Clear, equant 3-5 mm dolomite rhombs form a final carbonate pore-fill stage where porosity is not occluded. Rhomb faces are flat as are twin planes. Only minor saddle dolomite (cf. Radke and Mathis, 1981) was observed. This clear spar is similar to the limpid spar described by Folk and Land (1975).

Quartz-Hydrocarbon

A rare late-stage pore-fill is euhedral quartz in association with hydrocarbons. Quartz crystals are both singly and doubly terminated and are clear. Hydrocarbons occur as rounded blebs or spheroids up to 2 mm across (Plate 16B).

• Paragenetic Sequence

The precipitation sequence of cement fabrics, and variations of that trend, are illustrated in Figure 53. LFZD is typically the initial cement stage (Figure 53, stage 2), followed by LFD (Figure 53, stage 4) and Fe-zoned spar (Figure 53, stage 5) in an uninterrupted precipitation sequence (Plate 16A). Where LFZD is absent, LFD may be the initial cement stage (Plate 15D), and where both of these cements are absent, Fe-zoned or clear dolomite lines the pores (e.g. Figure 54, section 3; Plate 16B).

Micrite sedimentation and/or algal growth locally interrupts this sequence during the first two

precipitation stages (Figure 53, stages 3a, b). LFZD is commonly intergrown with algal layers or sediment (Figure 53) but in only one case was LFD intercalated with micrite. Intergrown LFD and algal laminae were not noted.

Zoned ferroan dolomite succeeds either LFD, LFZD or algal substrate (Figure 53, stage 5). Solution fabrics such as development of secondary porosity, dedolomitization of Fe-rich dolomite growth zones and truncation of zoned dolomite spar indicate that at least one solution event preceded precipitation of limpid dolomite and the quartz-hydrocarbon association.

Cement Distribution

Figure 54 shows the regional distribution of cement types within Member B. LFZD is most abundant in the September Lake buildup area and in curly bedding sequences of the eastern platform and transition zone (see Chapter 4). Within the buildup LFZD is restricted largely to shelter porosity associated with branching conical-columnar and spaced conical-columnar stromatolites. On the eastern platform, LFZD lines primary pores of the loose curly bedding algal framework. LFZD is rare on the western platform, and occurs only in section 7.

LFD occurs only on the eastern platform (sections 11-13), where it is subequal to LFZD. As with LFZD, it lines primary shelter porosity within the open algal

framework.

Fe-zoned and clear spar occur ubiquitously wherever pore space is not occluded. Both are also present in adjacent formations. Euhedral quartz occurs sparsely in sections on the platform, particularly those associated with curly bedding, as a late but primary void fill. Hydrocarbons occur only in section 5.

Origin of Member B Cements:

Both LFZD and LFD are interpreted as former submarine cements based on the following criteria:

- first-generation cement on marine bioclasts (stromatolites)(LFZD and LFD)(cf. Davies, 1977)
- isopachous rim geometry (LFZD and LFD)(cf. Davies, 1977; Mazzullo, 1977; Newell, 1955)
- morphologically similar to Recent submarine cements (LFZD and LFD)(cf. Mazzullo and Cys, 1979; Davies, 1977)
- intercalation with marine sediment (LFZD and LFD)(cf. Davies, 1977; Krebs, 1969; Mazzullo and Cys, 1979; Otte and Parks, 1963)
- encrustation or intergrowth with marine organisms, as represented by the algae/bacteria of organic mats in stromatolites (LFZD)(cf. Davies, 1977; Krebs, 1969; Mazzullo and Cys, 1979)

Submarine cementation of carbonate sediments, such as those in the September Lake buildup, is a widely documented phenomenon (Bathurst, 1980; Bosselini and Rossi, 1974; Davies, 1977; Mazzullo and Cys, 1977, 1979; Mazzullo, 1977; Krebs, 1969; Walls et al., 1978, among others). The Sulky Formation cements illustrated here show many similarities both in fabric and facies distribution to other documented occurrences of submarine cementation in Phanerozoic rocks, although no equivalent Precambrian occurrences are known. The generally compact structure of most stromatolite types may have precluded the formation of comparably abundant submarine cements in other Precambrian buildups.

Isopachous Fe/Mn zoned spar cement is most commonly associated with phreatic precipitative environments. Manganese-rich/iron-poor zones (indicated by luminescence) require relatively reduced pore fluids for precipitation, a situation compatible with meteoric or burial, but not with marine, settings (Meyers, 1978; Carpenter and Oglesby, 1976). In addition, the ubiquitous distribution of this cement (Figure 54) and its occurrence as a fracture filling, suggest that it was precipitated in a groundwater system of regional homogeneity following burial, lithification and compaction. Solution events which interrupted growth of this cement stage also indicate periodic flushing by waters undersaturated with respect to dolomite. The origin of limpid spar appears

related to meteoric diagenesis with groundwater only slightly supersaturated with respect to dolomite (Folk and Land, 1975). The quartz-hydrocarbon association is likely related to greater burial temperatures, which would explain its restriction to deeper portions of the depositional basin.

A schematic summary of the diagenetic evolution of this sequence is presented in Figure 58 in coordination with diagenetic events for Member C.

Neomorphism:

Several neomorphic (cf. Folk, 1965) fabrics are common in Member B in association with stromatolites and fibrous cements. These fabrics will not be described in detail, but their mention is important because they can significantly alter the appearance of original depositional fabrics (e.g. Figure 56), and also may enhance otherwise unrecognized early diagenetic fabrics.

Recrystallization of Algal Laminae

Coarse (2000x500 u) palisade-type length-fast dolomite crystals occur within branching conical-columnar stromatolites of the September Lake buildup. This crystalline fabric is visible in both polished slabs and thin section. In both, a consistent orientation of crystal long axes perpendicular to lamination is apparent. (Plate 15B and 16C). A possible explanation for the control of this crystallographic orientation (suggested

by Noel James, pers comm., 1981) is that very fine bladed crusts were precipitated on algal laminae while stromatolites were growing (Figure 55a). During later diagenesis, this pre-existing fabric would be mimicked by coarser neomorphic spar to give the lamination-perpendicular orientation. The neomorphic origin of this coarse spar fabric is demonstrated by poikilotopic inclusion of multiple micritic algal laminae within single spar grains, consertal intercrystalline boundaries, and the radiaxial extinction patterns of single crystals (Plate 16C). If it is accepted that control of neomorphic recrystallization requires a precursor texture, then this fabric may indicate probable sites of early cementation.

Recrystallization of Micritic Substrate

The striking abundance of fibrous textures in some specimens from the September Lake buildup (e.g. Figure 56), locally composing up to 70 percent of some outcrops, is a feature typical of many originally highly porous boundstones (cf. Krebs, 1969; Bosselini and Rossi, 1974). These fibrous fabrics initially were interpreted as recrystallization textures after micrite (Tyrell, 1969). The present trend is to interpret all fabrics as neomorphosed cements (Schmidt and Klement, 1971; Mazzullo, 1980). Only a few authors have proposed both replacive and precipitative

Figure 55. Neomorphism of (A) intercalated algal laminae (a.l.) and bladed spar crust (b.c.) to (B) length-fast palisade dolomite (n.s.) texture. Bladed spar crust is hypothetical and crystal size is exaggerated for clarity.

Figure 56. Algal boundstone(?) fabric (stippled) encrusted by LFD (unpatterned). Apparent high degree of fibrous cement texture is caused by pseudo-fibrous neomorphism of primary algal texture. The local marked discordance between growth increment banding of LFD (dash-dotted lines) and substrate boundary (marked by R) is a result of this neomorphism (see text for further discussion).

origins for certain fibrous carbonate fabrics (Krebs, 1969; Fischbuch, 1968; Davies, 1977).

The existence of fibrous carbonate that may be neomorphic after micritic algal frameworks, is suspected for some Dismal Lakes stromatolites. The relationships which indicate a neomorphic origin are the irregularity of the cement-"framework" contact on a macroscopic scale (Figure 56) and the presence of relic patches of micritic framework surrounded by fibrous carbonate (Plate 16D). Typically, LFD cements form even isopachous rims which mimic original pore-wall geometry. In Figure 56, a marked divergence is apparent between the cement growth surface and the cement-"framework" boundary. This divergence is caused by framework recrystallization which progresses into the framework from the original pore-wall boundary. This recrystallization results in an apparent overabundance of cement relative to framework (e.g. Figure 56). Similar relationships have been noted by Krebs (1969), and this phenomenon has been observed in several Devonian reefs of the Alberta Basin as well. This recrystallization phenomenon probably is more common than presently recognized, leading to erroneous estimates of the percentage of primary porosity.

Recrystallization of LFZD

In addition to possible micritization-banding and ray-crystal growth in LFZD, two other varieties of late

aggradational neomorphism affect this fabric.

Elongate 1000-2000 x 100 μ length-fast ray crystals, which closely resemble palisade calcite of Folk and Assereto (1976), locally replace LFZD. These crystals are similar to those replacing algal laminae. Locally a single palisade crystal has replaced both LFZD and algal textures (Plate 15B). Where length-fast ray crystals replace fans of LFZD, they form fan aggregates, demonstrating control of neomorphic recrystallization by the primary radial fibrous fabrics (Plate 16E).

The second (and most abundant) neomorphic fabric replacing fibrous cements is an interlocking jigsaw-type psuedospar of 50-100 μ equant grains (Plate 16F). Grain boundaries are lined with micrite. Individual psuedospar grains are characterized by undulose extinction. Well-preserved LFZD or LFD texture can be traced laterally into this interlocking neomorphic texture within a space of 2 cm. The patchy distribution, interlocking grain boundaries and grain-size reduction at grain boundaries, all indicate a neomorphic origin for this texture.

MEMBER C

Regression during Sulky Formation deposition caused exposure of the September Lake buildup. This change resulted in transition from submarine to meteoric-vadose and phreatic diagenetic environments. Exposed reef

crests are typified by complex diagenetic histories (cf. Assereto and Folk, 1980; Esteban and Pray, 1977). An analogous diagenetic history probably is responsible for the formation of the multiple diagenetic carbonate fabrics described below from Member C lithologies on the September Lake buildup.

The complex fabric evolution of Member C is divided into precipitative (or aggradational), and deformational (or dissolution-related) events (Table III). This division is necessary in this complex environment where fabrics typically interpreted as diagenetic or "secondary" (ex. cement fans) become sufficiently abundant to form a major constructional portion of the rock. Fabrics will be discussed as listed in Table III diagenetic histories of the September Lake High and Dismal Lakes Platform are discussed separately.

September Lake High:

Host Rock Origin

Host lithologies (those considered to be deposited by physical sedimentation processes) consist predominantly of white sub-mm laminated micritic limestone, and a now-fenestral, crudely laminated micritic limestone. Subsidiary lithologies are stromatolitic units (10-40 cm thick) and rare intraclastic beds (5-20 cm thick). Primary bedded anhydrite can be inferred to have been present in

TABLE III. DIAGNETIC EVENTS FOR MEMBER C OF SULKY FORMATION, SEPTEMBER LAKE HIGH

PRECIPITATIVE STAGESPENECONTEMPORANEOUS DEFORMATION STAGES

IIe. Silcrete formation, precipitation of agate cements

IIb. Brecciation and clastic reworking of silicified evaporite

IIId. Evaporite silicification

IIc. Deposition of geopetal micrite

IIb. Precipitation of luminescent spar

IIa. Precipitation of non-luminescent bladed spar

IIa. Continued fabric disruption, brittle fracture of original host rock, plastic deformation of precipitated micritic coatings

I. Precipitation of micritic coatings and some fibrous cement in association with crustal-cellular cement and pisolites

I. Host rock brecciation, tepee formation, development of fenestral porosity

sections 16-18 from relic textures preserved in chert clasts within a stratiform breccia (see Chapter 4). Thin gypsum or anhydrite layers also may have been intercalated with white micritic limestones (see below).

Accumulation of "host rock" sediment occurred several times during Member C deposition, as a result of local incursions which interrupted diagenetic processes, adding fresh unaltered host rock to the system.

Deformation Stage I

Synsedimentary breccias (Plate 9F) and tepee structures (Plate 9D) were formed by subaerial weathering, desiccation and recrystallization of host lithologies. Angular breccias are similar to subaerial breccia horizons from the Pleistocene of Barbados (Harrison, 1977) and the Carboniferous of Kentucky (Harrison and Steinen, 1978). All three typically display a jigsaw pattern of host-rock clasts indicating an in situ origin. In the Dismal Lakes and the Kentucky examples, host-rock clasts appear to float in a matrix of finely laminated micrite. Origin of this micrite is discussed in the following section. In the Dismal Lakes example, some of the brecciation and collapse of host rock layers may be related to dissolution of intercalated evaporite layers. Stratiform evaporites are known from laterally associated facies, making this association tenable, but no relic evaporite textures were observed.

The origin of tepee structures was discussed in a previous section (Chapter 4). As outlined by Assereto and Kendall (1977), tepees are presumed to result from desiccation of, and expansive crystallization within, host rock during subaerial exposure. Enhancement of primary porosity and production of secondary porosity by formation of fenestral fabrics within host carbonates is a third type of alteration preceding precipitation stage I. This penecontemporaneous deformation related to breccia and tepee formation defines deformation stage I.

Precipitation Stage I

Precipitation, rather than deposition by current processes, is proposed here for the origin of much of the micrite which post-dates deformation stage I. Fibrous carbonate is the second precipitative component of precipitation stage I, alternating with micritic layers in pisolites and cellular crustal cements.

Precipitative micrite occurs in textures ranging from clearly precipitative rims around breccia fragments (Plate 17A) to laminated botryoids closely resembling stromatolites (Plate 17C). Non-laminated precipitative micrite is essentially indistinguishable from micrite deposited by currents where pore-lining relationships are not apparent (Plate 17B). Where pore-lining relationships are visible, micritic rims are uneven in thickness, locally forming geopetal stalactitic and

stalagmitic structures (Plate 17A).

Some examples of subaerial breccias with precipitative micrite pore-fill have been deformed a second time (see deformational stage II), disrupting both original host rock and micritic rims (Plate 17B). Host-rock breccia clasts were deformed in a brittle manner during this event, whereas micrite was deformed plastically, indicating that these coatings were cohesive but not totally lithified. Micritic structures resembling stromatolites (Plate 17C) have thin sub-mm scale lamination which is transitional to cellular-crustal cement (Plate 17D). Pseudo-stromatolite structures composed of cellular-crustal cement are illustrated by Assereto and Kendall (1977, their Figure 526), and are interpreted by them to be stalagmites. This inorganic origin is favoured for the Dismal Lakes structures as well, although in this instance, pseudo-stromatolites do not form in cavities. The reason for rejecting a biologic origin for these structures is that cellular-crustal cement is known only from inorganic stalagmites and stalactites, and similar lamination textures do not occur in any unequivocal Dismal Lakes stromatolites.

Cellular-crustal cements are composed of scalloped micritic layers, 50-100 μ thick, which enclose lenses of clear equant microspar-psuedospar (Plate 17D). A few spar lenses display relic radial extinction patterns inherited

from a fibrous precursor, but no square terminations indicative of an aragonitic precursor (cf. Loucks and Folk, 1976; Folk and Assereto, 1976; Assereto and Folk, 1980) are preserved.

Pisolites consist of concentric scalloped bands of micrite and clear equant spar. Relic fibrous texture within spar laminae testify to a fibrous precursor with radial c-axis orientation. A final early precipitative texture consists of pelleted micritic cement in the pisolitic grainstones. Similar pelleted micritic cements occur in Holocene carbonates, where they are composed mostly of high-Mg calcite (Mazzullo, 1977).

Deformation Stage II

Deformation stage II is essentially synchronous with precipitative stage I. This stage includes all phenomena which disrupt precipitative stage I fabrics (Plate 17B). This event is primarily identified by contortion of precipitated micrite layers, but also includes continued development of tepees.

Precipitative Stage II

Precipitative stage II is defined by two distinctive carbonate cement types: a cloudy bladed non-luminescent dolomite spar, and; a clear luminescently banded equant dolomite spar. Significant penecontemporaneous silica diagenesis also occurred during precipitation stage II.

Cloudy bladed dolomite spar occurs in most samples

of member C on the September Lake High, both as isopachous rims and as stalactitic growths. Bladed spar cements rim previously deformed micrite cements, indicating a post-deformational stage II origin for this cement type.

Clear equant spar cement is easily distinguished from bladed spar by morphology, luminescence and lack of inclusions. This cement type forms isopachous rims, except where growth is inhibited by synchronous micrite sedimentation. This luminescent cement is present in approximately 10 percent of samples studied.

Silicification of evaporites, formation of a silcrete weathering profile, and related precipitation of agate cements, all postdate precipitation of at least the bladed dolomite spar. Relative timing of agate cement and luminescently zoned spar cement is uncertain.

Clasts of silicified evaporite in a stratiform chert breccia (Plate 9E) display a well-preserved felted lath texture as described by Chown and Elkins (1974). Small (10x50 u) inclusions parallel to long axes of lath-shaped pseudomorphs are interpreted as anhydrite, the probable precursor mineral. Laths are typically replaced by monocrystalline quartz, but polycrystalline infills also occur. Laths are surrounded by banded chalcedony rims. Chalcedony is generally interpreted as a pore-fill rather than a replacive fabric (cf. Milliken, 1979) but the

presence of highly birefringent inclusions marking cloudy growth bands in chalcedony might indicate evaporite dissolution concomitant with chalcedony precipitation. Similar banded chalcedony (agate) occurs as cements throughout Member C on the September Lake High. Agate cements typically coat non-luminescent bladed spar, but have not been seen in contact with luminescent spar in any pore-fill sequence.

Qualitative field observations made before recognition of the silcrete profile show an increase in the silica-carbonate ratio upward through the section towards the silcrete horizon. Extensive leaching of carbonates before agate precipitation is apparent immediately below the silcrete, as would be expected during subaerial exposure and leaching. The silcrete horizon is 4-6 cm thick, consisting of massive chert with pockets of relic carbonate and secondary clay minerals (rare). All primary carbonate textures have been destroyed, and are now replaced by a swirly silica texture cut by abundant quartz and carbonate veins. Euhedral quartz crystals displaying flamboyant spectral extinction (Chowns and Elkins, 1974) are scattered throughout the silcrete. Such crystals have been reported from silicified evaporite deposits (Chowns and Elkins, 1974; Milliken, 1979) and from ancient silcrete (Rubin and Friedman, 1981). Clay minerals fill pore spaces lined with chalcedony, indicating a late

diagenetic origin relative to silcrete formation. This is the only occurrence of clay minerals in the Dismal Lakes carbonate section.

The final-stage pore infill post-dating the agate cements is a translucent Fe-calcite similar to the late-stage limpid spar of Member B (see previous section).

Platform:

Diagenetic events recorded in rocks of the Platform lithofacies of Member C are relatively simple in comparison with the complex history on the September Lake High. Penecontemporaneous diagenesis is restricted to pisolite and tepee formation, and to development of fenestral porosity. Equant spar is the only carbonate cement fabric preserved.

The sequence of post-depositional (burial) diagenetic events for member C Platform lithofacies can best be illustrated by observations of a widely occurring allochem of likely metastable primary mineralogy (aragonite or high-Mg calcite). Ooids are the only widely distributed allochem which fits these requirements in member C. Stages of ooid diagenesis are summarized in Figure 57. Initial diagenesis caused differential dissolution of alternating laminae within ooids, producing ring-in-ring structures which were gravitationally stabilized (Figure 57B). Differential

Figure 57. Diagenetic events in member C of Sulky Formation as recorded in ooid packstone fabric (A). Stage B illustrates dissolution of unstable laminae of ooid cortex. Stage C is micritization of micritic matrix and remaining layers of ooid cortex (upper illustration). Where previous dissolution of ooids has not occurred, the entire ooid is micritized (lower illustration). Stage D is the silicification of all fabrics randomly; open-space fill is by chalcedony (lined fabric) and equant megaquartz (stippled).

dissolution resulted from either compositional or microporosity-related (Donath et al., 1981) differences within the ooid cortex. Following selective dissolution, micritization of both undissolved and dissolved ooids and surrounding matrix-micrite occurs (Figure 57C).

Micritization has caused variable degrees of fabric loss. Some ooids have a pseudo-isogyre extinction pattern inherited from an originally radial-fibrous texture, whereas other micritized ooids show random extinction.

Silicification is patchy, occurring as 5-8 x 20-30 cm chert nodules, some of which display concentric, concretionary structure. Chert mimics the totally micritized carbonate texture; chalcedony forms isopachous to botryoidal rims around intragranular and secondary (ooidic) porosity.

The paucity of early diagenetic features, and the homogeneity of later dissolution and micritization, suggest that these sediments were not exposed to vadose or near-surface marine diagenetic processes for any appreciable time. Rapid subsidence and accumulation rates on this steadily subsiding platform probably were responsible for limiting diagenesis to the inferred phreatic-meteoric realm.

Origin of Member C Diagenetic Fabrics:

All precipitative and deformational fabrics of member C of the September Lake High, excluding late stage

ferroan calcite, have Recent and Phanerozoic counterparts in peritidal and subaerial marginal-marine environments. Fluctuating salinities of diagenetic waters as a result of flushing of the pore system by alternating fresh (rain and groundwater) and marine (onshore-directed marine flooding) waters produced cement stratigraphies with variable compositions (eg. calcite, dolomite, aragonite). In addition, water-table fluctuations resulted in the superposition of vadose and phreatic textures.

The slow subsidence rate of the September Lake High would have enhanced the effects of this diagenetic environment by increasing the length of time over which these sediments were exposed to the vadose diagenetic environment. Steady subsidence of the platform, during exposure of the September Lake High, restricted diagenesis in this portion of Member C to relatively late-stage dissolution, neomorphism and silicification.

SUMMARY

Progressive diagenetic stages recorded in the Sulky Formation are illustrated in Figure 58. Local early submarine cementation of quiet-water micrites occurred during deposition of member A (Figure 58A). Abundant precipitation of fibrous cements during formation of member B algal carbonates occurred in a moderately deep subtidal environment (Figure 58B). The abundance of submarine cement fabrics in this sub-wave base buildup

Figure 58. Summary of diagenetic history of the Sulky Formation. A) member A- localized cementation and formation of concretionary nodules in deeper-water micrites. B) member B- fibrous cementation of September Lake buildup and curly bedding zone of eastern platform. C) member C- vadose and phreatic-meteoric diagenesis of buildup-crest facies of September Lake High and restricted vadose diagenesis of the western platform. D) Post-burial precipitation of Fe-zoned spar and other late-stage fabrics throughout the Sulky Formation and surrounding units.

FIGURE 58

appears to be inconsistent with Matthew's (1974) contention that considerable wave pumping is essential to produce the pore-water circulation required for the supply of Ca^{++} and HCO_3^- ions to precipitation sites.

Shoaling and subaerial exposure of the buildup-crest localized vadose diagenesis and weathering, and produced a variety of diagenetic fabrics in member C of the September Lake High (Figure 58C). Steady accumulation of subtidal-to-supratidal sediments of member C on the Platform restricted subaerial diagenesis to localized zones (Figure 58C). Late-stage dissolution, micritization and silicification are the dominant diagenetic textures in this portion of member C.

Burial-related cementation of Fe-zoned spar is pervasive throughout the Sulky Formation, and in surrounding formations. These cements were precipitated in equilibrium with oxidizing groundwaters in a phreatic diagenetic environment. A relatively late, burial-stage origin for these cements is postulated (Figure 58D).

PLATE 15. DIAGENESIS OF THE SULKY FORMATION

A. Polished slab of curly bedding and distinctively banded LFZD cement. Banding of LFZD conforms to pore walls defined by stromatolitic laminae (S). Scale is graduated in millimetres.

B. Photomicrograph of LFZD cement coating stromatolitic laminae. Stromatolitic lamination is best seen in plane-polarized light (C of this Plate). Banding of LFZD is sub-parallel to substrate boundary, but the alternating fibrous (F) and micritic (M) bands anastomose, indicating that they do not represent growth-increment banding. Large blades of neomorphic palisade dolomite overprint both stromatolite and LFZD fabrics. Isolated botryoids within LFZD are interpreted as primary growth structures. Cross-polarized light. Width of field is 2 cm.

C. Plane-polarized light photomicrograph of B of this Plate. High-relief areas in stromatolite are length-slow dolomite. Low-relief areas making up most of the stromatolitic substrate are of length-fast palisade dolomite. Stromatolitic layering is marked by dense micritic clots and laminae (S). Width of field is 2 cm.

D. Photomicrograph of LFD cement showing isopachous morphology and coarse elongate ray crystals. Crystals show length-fast orientation with c-axes normal to substrate. LFD is succeeded by zoned ferroan dolomite as a final pore-fill. Cross-polarized light. Width of field is 1.5 cm.

PLATE 15

PLATE J5

A

B

C

D

PLATE 16. DIAGENESIS OF THE SULKY FORMATION

A. Zoned ferroan dolomite cement succeeding LFZD and LFD as a late-stage pore-fill. Plane-polarized light. Width of field is 1.5 cm.

B. Solution vug filled with clear and zoned ferroan dolomite spar, followed by hydrocarbon infill. Note that the faint stromatolitic lamination (S) is truncated by the vug. Hydrocarbon blebs cross-cut Fe-zoned dolomite (arrow). Plane-polarized light. Width of field is 1.5 cm.

C. Neomorphosed stromatolitic laminae with c-axes of pseudospar crystals oriented perpendicular to laminae (dotted). Cross-polarized light. Width of field is 1.5 cm.

D. Neomorphosed stromatolite laminae adjacent to LFD cement. Portions of the primary stromatolite-cement contact (dashed) are nearly destroyed due to aggrading neomorphism, which has left micrite islands (I) floating in neomorphic spar. Plane-polarized light. Width of field is 1.5 cm.

E. Fan aggregate of length-fast palisade recrystallization fabric replacing LFZD cement. Fan-like array of palisade crystals is controlled by orientation of primary LFZD fabric. Cross-polarized light. Width of field is 1.5 cm.

F. Jigsaw pseudospar, a common neomorphic fabric after fibrous cement. Crystal boundaries are mainly consertal and lined with micrite. Plane-polarized light. Width of field is 1.5 cm.

PLATE 16

PLATE 16

A

B

C

D

E

F

PLATE 17. DIAGENESIS OF THE SULKY FORMATION

A. Irregular pore in enterolithic breccia, filled with pore-lining and geopetal micrite of precipitative stage I and agate cement. Host rock micrite is darker than precipitative micrite and displays a sub-mm-scale lamination. Plane-polarized light. Width of field is 2 cm.

B. Enterolithic dolomite breccia showing alternating host micritic sediment and precipitative stage I micrite cement, here preferentially deformed by deformation stage II. Precipitative stage II equant spar cement fills voids. Contrast between sub-mm-scale laminated host sediment and massive precipitative micrite is apparent. Plane-polarized light. Width of field is 2.5 cm.

C. Pseudo-stromatolite structures composed of cellular-crustal cement. A sediment drape that fills low area in cement coating demonstrates the syndimentary origin of this cement fabric. Plane-polarized light. Width of field is 2 cm.

D. Cellular-crustal cement showing well-developed micritic-cellular structure. Coarser spar areas within micritic "cells" locally preserve radial-fibrous crystal texture. Plane-polarized light. Width of field is 2.5 cm.

PLATE 17

PLATE 17

CHAPTER 9. SILICA DIAGENESIS OF UPPER DISMAL LAKES GROUP DOLOSTONES

Chert nodules are abundant in the Dease Lake, Kendall River, Sulky and Greenhorn Lakes Formations. The presence of chert is particularly important in the latter three formations, because it commonly preserves many of the primary carbonate textures which otherwise might be destroyed by carbonate diagenesis.

The origin of secondary chert in shallow water carbonates has occupied the efforts of numerous geologists for many years (Ireland, 1959; Knauth, 1979). Recent DSDP research dealing with the formation of deep sea cherts has provided much useful information on the early stages of chert precipitation (Wise and Weaver, 1974). In particular, crystallization textures formed during the transition from amorphous silica to microgranular quartz (chert) have been documented. The transitional silica forms (opal-A, opal-CT and chalcedony) represent gradual stabilization from amorphous silica with a solubility of 40-60 ppm. The recognition of a sequence of quartz fabrics which can be related to initial solubilities of the precipitating waters is potentially applicable to petrographic studies which attempt to address the problem of the origin of chert (cf. Meyers, 1977) in much the same way that recognition of early diagenetic carbonate cements has

aided the study of limestone formation.

Specific problems that are approached in this chapter are:

- how do the chert textures observed in Dismal Lakes cherts compare with those described from DSDP studies?
- what do these chert textures indicate about the nature of silica diagenesis?
- what is the relationship between silica diagenesis and carbonate diagenesis?

STRATIGRAPHIC RELATIONSHIPS

Chert types from different stratigraphic intervals of the Dismal Lakes Group are distinct. Specifically, the abundant waxy-black chert nodules of the Kendall River and Greenhorn Lakes formations contrast with the less abundant white-grey chert of the Sulky Formation. Although the distribution of both chert types with respect to depositional fabric appears similar for the three formations, the microscopic chert textures and the timing of silicification relative to carbonate diagenesis differ significantly.

Chert from these formations will be compared with emphasis on diagenetic history. Kendall River and Greenhorn Lakes cherts are essentially identical, and therefore only Kendall River and Sulky Formation cherts

will be discussed in detail.

Kendall River Formation:

Chert in the Kendall River Formation occurs as nodules elongated parallel to bedding and ranges from black through grey to white depending on whether it replaces organic matter or whether it forms as pore-fill material respectively (Plate 18A). Commonly, silicification has been selective, replacing stromatolites or algal mat layers. Chert nodules also replace sediments with textures indicative of a high primary permeability such as ooidic grainstones or pustulose algal mat layers (Plate 18A). Preferential compaction of unsilicified carbonates around chert nodules (locally up to one-sixth original depositional fabric) demonstrates that at least some silicification preceded compaction (Plate 18B).

In contrast to the black cherts which predominate in most ~~sections~~ of the Kendall River Formation, the chert in sections south and west of Greenhorn Lakes is typically sugary-white or grey. A minor chert type in the Kendall River Formation is a coarse-grained green chert associated with silty carbonate layers near the base of the formation.

Sulky Formation:

Chert is generally absent from members A and B of the Sulky Formation, but is common in member C. Here it

typically is white, coarsely crystalline, and occurs as nodules parallel to bedding. Hollow concretionary chert nodules appear restricted to member C (Plate 18C).

On the September Lake High, two distinctive stratiform occurrences of chert, a stratiform chert breccia of evaporite-replacement origin, and a silcrete profile, are present in addition to the typical nodular chert. Small amounts of chert occur in member A in the A2 facies as silicified algal layers. Chert in member B occurs only on September Lake High where it replaces both algal laminations and carbonate cement fabrics, particularly in growth zone 1 (see Figure 25).

Compaction relationships in the Sulky Formation are as described for the Kendall River Formation, with carbonate fabrics surrounding chert nodules being strongly compressed relative to the corresponding fabrics within chert nodules, indicating precompaction silicification. Rarely, thin carbonate veins crosscut the chert nodules. These veins probably formed by injection of partially lithified or unlithified sediment into fractures in the early lithified chert nodules during post-lithification compaction (cf. Meyers, 1977).

SILICA DIAGENETIC FABRICS

Description:

The classification scheme proposed by Wilson (1966) for silica diagenetic fabrics is easily adaptable to the

present study (Figure 59). Modifications proposed for this study are the addition of zebraic chalcedony as a cement fabric, the recognition of relic lephispheres now recrystallized to microcrystalline granular quartz, and the absence of length-slow chalcedony (luetecite).

The author has also reordered Wilson's diagram such that the precipitative fabrics are arranged in the observed order of precipitation. This sequence can be related to decreasing silica solubility data for interpreted fabric precursors (Kastner, 1979). Fabrics are described systematically below, followed by documentation of fabric relationships and interpretation of the diagenetic sequence.

Chalcedony Overlays (CO)

Clear, isopachous chalcedony rims (10-20 μ) occur locally as first-generation cements coating allochems (Plate 18D). This chert type (CO) is most common in grainstone fabrics, but also occurs in silicified algal mats. The extinction pattern of this chalcedony is irregular, and although fibres are generally oriented perpendicular to the substrate, many fibres appear oblique, and few extend across the entire rim. The isopachous, non-radiating character of this chalcedony makes it distinct from spherulitic chalcedony. Where present, CO precedes the growth of all other precipitative fabrics, including lephispheric MGQ (see

Figure 59. Chert fabrics of the Dismal Lakes Group, classified according to the terminology of Wilson (1966). CO- chalcedonic overlay, MGQ- microcrystalline granular quartz, SC- spherulitic chalcedony, MQ- mosaic quartz, EQ- equant quartz. Cement fabrics are arranged in ideal precipitation order, presumed to correlate with decreasing silica solubilities (MQ and EQ- 6 ppm; CO- 40-80 ppm). Dashed areas represent clastic particles (e.g. ooids and intraclasts). Spheres in MGQ cement represent lephospheres. Upper sketch in SC cement with jagged banding represents zebrachalcedony. Fine rectangular laths in MQ replacement fabric represent evaporite pseudomorphs, as do prismatic inclusions in EQ replacement sketch.

3

FIGURE 59

CO	MGQ	SC	MQ	EQ

DECREASING SOLUBILITY				
CEMENT				
REPLACEMENT

below).

Microcrystalline Granular Quartz (MGQ)

This fabric category combines at least two fabric subtypes of similar grain size and texture, but of different colour, inclusion content and general occurrence (Plate 18E and 18F). MGQ may be either white or shades of brown. Grains are anhedral, equigranular and range in size from 20 μ to submicroscopic. The extremely fine grain size and presence of included organic material in brown MGQ produces a dense pseudoisotropic extinction character (Plate 18F). Brown MGQ typically contains dolomite and finely disseminated organic matter as inclusions (Plate 18E and 18F) and generally is finer grained than white MGQ. White MGQ contains no microscopically distinguishable inclusions (the white coloration is probably imparted by submicroscopic fluid inclusions (c.f. Folk and Pittman, 1971). Much of the MGQ in the Sulky Formation is transitional, in terms of coloration and inclusions between the white and brown varieties of chert described above. As discussed below, both the brown and light brown (transitional) MGQ textures are interpreted to be of replacement origin. The difference in coloration and inclusions between the brown chert and this transitional chert type probably is due to the amount of carbonate diagenesis that took place before replacement by silica

(see below).

White spheres 10-30 μ in diameter can be recognized in the white MGQ fabric where it forms rims on primary pore space boundaries (Plate 18E). Individual spheres are generally discernible towards the center of pore spaces, merging or grading into massive white MGQ towards pore walls, giving the impression that massive white MGQ formed by growth and coalescence of discrete spheres (Plate 18E). This relationship, though only locally preserved, is thought to be of general importance.

By analogy with precipitative silica fabrics observed in Recent cherts (e.g. Wise and Weaver, 1972), these structures are interpreted as relic lephispheres. Growth and coalescence of lephispheres is inferred to be the dominant process by which open-space-filling white MGQ formed. Coalescence of lephispheres has been noted by Carver (1980) and by Kastner (1979) in studies of Miocene and Recent siliceous sediments respectively.

Lephispheres characteristically occur as open-space-filling structures and not as replacive or displacive features. The size of the spheres described here, and their occurrence as early diagenetic pore-filling structures, supports the lephisphere analogy.

Spherulitic Chalcedony (SC)

Spherulitic chalcedony is generally clear in early growth stages, becoming distinctly banded in larger spherulites which coalesce to form agate-like pore-fills. Typical banding is an alternation of clear and brown layers, but rare blue coloration also has been observed. Individual spherulites and spherulitic crusts locally attain thicknesses of up to 1 cm, but more typically are 200-1000 μ thick (Plate 19A). All chalcedony is length-fast, with the zebraic type as a distinctive variety (Plate 19B).

Almost all chalcedony spherulites are nucleated on a substrate and occur as precipitative pore-fillings (Plate 19A and 19B). An exception to this general rule are spherulites (1-3 mm) which appear to be suspended in dolomite (Plate 19C). These spherulites coalesce to form irregularly shaped silicified patches within the host carbonate. They are restricted in occurrence to member B of the Sulky Formation on the September Lake High. These replacive spherulites overprint both algal and cement fabrics, and typically appear cloudy grey in plane-polarized light due to the presence of abundant carbonate inclusions. The presence of these inclusions causes the spherulitic radial extinction pattern to be irregular. Preservation of primary carbonate textures within zones silicified by these spherulites is poor.

Mosaic Quartz (MQ)

The silica of this fabric is colourless. It may contain inclusions of carbonate and sulphate, but more commonly is inclusion-free. The fabric is typically equigranular, except where it occurs as a pore-space lining, where it may take on a drusy elongation perpendicular to the pore boundary (Plate 19D). Grain size ranges from 50-2000 μ , and averages 1000 μ . Grain size gradation from margin to pore interior rarely occurs. Mosaic quartz exhibits randomly oriented straight extinction, but also displays either undulose or flamboyant spectral extinction (c.f. Chowns and Elkins, 1974). One occurrence of megaquartz spherules (c.f. Milliken, 1979, p. 250) was noted in a sample from the silcrete profile of member C of the Sulky Formation.

Euhedral Quartz

Euhedral quartz occurs both as radiating clusters and as doubly terminated crystals within a carbonate matrix, and as singly terminated crystals growing into open pore space.

The radiating-cluster variety of EQ is restricted to a single bed which provides a marker for the top of the Sulky Formation on the platform. These radiating clusters appear as porphyroblasts with a poikilotopic texture (Plate 19E). Abundant inclusions of anhydrite and dolomite occur in these crystals. Clusters attain

diameters of 2 cm, and isolated crystals have a maximum size of 1 by 0.2 cm. Extinction is straight in isolated crystals, and slightly to strongly undulose where crystals merge into clusters. Isolated crystals tend to be doubly terminated, whereas crystal clusters display a euhedral habit only on cluster perimeters (Plate 19E).

Paragenesis of Silica Diagenetic Textures:

Following the work of Wilson (1966) it has been possible to distinguish precipitative and replacive silica fabrics. In addition, a consistent order of precipitation (paragenetic sequence), has been determined for precipitative fabrics. To a limited extent it also has been possible to correlate silica diagenesis with carbonate diagenesis. Many workers have demonstrated both theoretically (Krauskopf, 1959; Knauth, 1979) and through observation that a delicate balance exists between stability of silica and calcium carbonate, and that this balance is related to fluctuations in pH and pCO_2 . The ability to correlate silica and carbonate diagenesis is useful, not only in outlining the overall diagenetic history, but also in constraining timing of silicification and source of silica.

Precipitative Fabrics (cements, open-space fillings)

Thin chalcedony overlays (CO) occur as the first-generation fabric coating allochems (ooids and intraclasts: Plate 18D). This silica texture is

restricted to grainstone fabrics. Chalcedonic overlays commonly are associated with the white variety of MGQ, with the overlays invariably preceding the MGQ. In these associations the white MGQ displays a well-developed relic lephishpere texture (Plate 18E). Where chalcedonic overlays are succeeded by SC, the contact between the two textures is sharp (Plate 20A), suggesting that these fabrics represent two distinct precipitation phases. CO has not been observed as a replacement fabric.

The occurrence of chalcedonic overlays as a pore-lining fabric restricted to grainstones implies a precipitative-cementation origin for this silica type. That CO occurs consistently in the sequences CO/white lephishpheric MGQ, or CO/SC, indicates a sequential timing of formation of the respective fabrics, and thus lends support to a precipitative rather than a replacive origin. The highly irregular extinction pattern of this chalcedony type may be related to recrystallization of a precursor polymorph such as opal or opal-CT. The isopachous morphology also is reminiscent of opaline cements.

The two main types of MGQ occur in distinctively different associations (Plate 20B). Brown inclusion-rich chert consistently forms a replacement fabric, and is discussed below. White MGQ only occurs as an interstitial fabric, and not in allochems. White MGQ is distinctly inclusion-free and slightly coarser grained

than brown MGQ. It commonly forms as a pore-lining fabric which is succeeded toward the pore interior by either spherulitic chalcedony or megaquartz. Where white MGQ occurs, it either composes the first-generation pore-fill texture, or is preceded by CO. Where white MGQ grades toward the pore interior into clear mosaic quartz, a transition zone displaying relic lephisphere texture may be preserved (Plate 18E). This chert type is interpreted as a pore-fill texture based on its consistent occurrence as an interstitial fill between primary grains, and on its systematic position in the sequence of chert textures as illustrated in Figure 59. In addition, the possible presence of relic lephisphere texture is consistent with the precipitative pore-fill origin of this fabric.

Spherulitic chalcedony also has a multiple origin. By far the most common habit of chalcedony is the botryoidal spherulitic crust texture which is restricted to grainstone fabrics. Spherulite nucleation sites are spaced from 0.2-0.5 mm apart along pore walls (Plate 19A). Spherulite fan boundaries are straight, and triple junctions form 120 angles typical of precipitative fabrics (Plate 19B). Spherulitic crusts may be first-generation precipitates or may be preceded by CO (Plate 20A), white MGQ, or locally, relic carbonate cements which are now silicified (Plate 20C). Spherulitic crusts may be succeeded toward the pore

interior by white MGQ (rare) or by mosaic quartz.

As for previous fabrics, the pore-lining geometry, position in a consistent pore-fill stratigraphy and the lack of inclusions or observed overprinting relationships all indicate a precipitative origin for the SC fabric.

Although Folk and Pittman (1971), McBride and Folk (1977) and Milliken (1979) have noted zebraic chalcedony in association with silicified evaporites, none noted replacement of evaporites by chalcedony. No association was observed between evaporite replacement textures and zebraic chalcedony in Dismal Lakes cherts.

Mosaic quartz showing equant or drusy fabrics is common, but is volumetrically less important than either MGQ or SC. This fabric is more abundant in cherts of the Sulky Formation than in those of the Kendall River or Greenhorn Lakes Formations. MQ occurs in pore-fill sequences typically preceded by either MGQ or SC, although rarely it is present directly lining pore walls (Plate 19D). Mosaic quartz in its drusy form generally occurs in clastic carbonates of high primary porosity, as is the case with the other precipitative chert fabrics. In addition, MQ may fill secondary porosity such as oomoldic porosity (Plate 19D). In general, MQ is the final precipitative phase. Locally, however, it may be followed by coarse dolomite spar.

As argued for the previous fabrics, the drusy pore-lining texture, absence of inclusions and relative

position in the previously established cementation sequence all indicate a precipitative origin for this fabric.

Euhedral quartz in vugs of both primary and secondary (leached) origin clearly displays a precipitative fabric of incomplete growth into open pore space. Vugs commonly are lined with coarse dolomite spar precipitated before quartz. The only pore-fill material which post-dates euhedral quartz is an unanalyzed hydrocarbon.

The idealized precipitation sequence for quartz types is shown in Figure 59. Although the complete sequence has not been observed in a single pore-fill, the relative precipitative order is well-established through comparison of numerous thin sections. Omission of one or more precipitative stages is common, but the order of precipitation is consistent.

Replacive Fabrics

Of all chert fabrics described above, four can be assigned to a replacive origin (Figure 59). They are the brown variety of MGQ, chalcedony spherulites, mosaic quartz and euhedral quartz. Common characteristics of these replacive fabrics are the presence of abundant carbonate inclusions (as well as inclusions of organic matter, evaporite minerals and finely disseminated liquid) and the tendency to overprint primary carbonate textures.

Brown microcrystalline granular quartz is typically cryptocrystalline and occurs within ooids or intraclasts of grainstone fabrics (Plate 18F). Less commonly, the organic portion of biogenic mats and the micritic matrix of packstone fabrics have been converted to brown MGQ. A replacive origin is indicated for both brown and tan MGQ by the preferential occurrence of this chert fabric within primary carbonate allochems (ooids, micritic sediment, intraclasts) or organic mats. In addition, the preferential occurrence of abundant carbonate inclusions in this MGQ fabric, in contrast to the white MGQ, is evidence for partial replacement of a primary carbonate host.

Spherulites of chalcedony are of two types: spherulitic crusts as described above; and isolated spherulites, 1-3 mm diameter, which occur only in member B of the Sulky Formation. Some isolated spherulites overprint primary carbonate fabrics, including stromatolite laminae and fibrous carbonate cements (Plate 19C). This overprinting relationship, and the abundant carbonate inclusions within these spherulites (in contrast to chalcedony of spherulitic crusts which is inclusion-free), is evidence of a replacive origin for this silica fabric.

Mosaic quartz forms a replacement texture as well as a precipitative texture. In contrast to the fabrics

previously discussed, mosaic quartz of replacement origin is difficult to distinguish from precipitative MQ, except where diagnostic inclusions are observed. In particular, MQ is recognized as replacive after anhydrite (gypsum?) as needle pseudomorphs. Here, prismatic anhydrite inclusions are contained poikilotopically within crystals of the quartz mosaic. Long axes of the anhydrite inclusions commonly are parallel to pseudomorph long axes (Plate 19F).

Euhedral quartz crystals that grew poikiloblastically in a dolomicrite matrix, and which contain abundant dolomite and anhydrite inclusions (Plate 19E), are clearly of replacement origin. Similar features have been noted by Nissenbaum (1967) and Friedman (1980) in Phanerozoic carbonate/evaporite sequences.

In summary, MGQ is the most common replacive silica fabric, the brown MGQ being common in the Kendall River Formation (imparting a dark brown to black appearance to megascopic chert nodules) and the clear-tan MGQ in the Sulky Formation. MQ, euhedral quartz and replacive SC are laterally restricted in occurrence and are not volumetrically significant.

SILICA/CARBONATE DIAGENETIC RELATIONSHIPS

Cross-cutting relationships, such as silicified pressure-solution fabrics (Plate 20E) and silicified, partially recrystallized textures (Plate 20C and 20D),

can provide a general impression of the relative timing of carbonate and silica diagenesis. General trends are described below, and are summarized for both the Sulky and Kendall River Formations in Figure 60. An outstanding problem with this analysis is the uncertainty of the timing of dolomitization. Many Precambrian dolomites have been shown to be similar to Phanerozoic dolomites in terms of their secondary, early to late diagenetic origin (e.g. Ricketts, 1979). However, Tucker (1982) has presented evidence to suggest that some Precambrian dolomites are of primary origin. Until geochemical or other evidence is available to prove otherwise, it will be assumed that dolomite in the Dismal Lakes Group formed by replacement of calcium carbonate.

In the Kendall River Formation, coarse rhombic pseudomorphs within a micritic matrix (rhombs and micrite now silicified; Plate 20D) suggest that dolomitization of an original lime mud has occurred, but this texture could also represent aggradational neomorphism of an original dolomicrite to dolomite spar. Tentatively, the relationship in Plate 20D is inferred to represent minor early dolomitization before silicification. Pervasive dolomitization occurred in the sediment surrounding this silicified area. Similarly, dolomitization apparently was not interrupted by silicification in areas surrounding chert nodules, so that extensive recrystallization occurred adjacent to chert nodules.

Compaction and associated formation of stylolites primarily are post-silicification processes, but local differential compaction of silicified fabrics indicates that silicification in part was selective (e.g. some ooids and not others), and that some compaction occurred before silicification was complete (Plate 20E).

Silicified stylolites also are an indicator of some later stage, post-burial silicification. Chalcedonic overlay (CO) and brown and white MGQ all appear to precede major compaction. Where present, they have preferentially resisted compaction (Plate 20E).

An aggradational neomorphic event, which caused much unsilicified carbonate in still-permeable sediments to recrystallize to pseudospar, overprints brown and white MGQ and SC fabrics (Plate 20F). This neomorphic event in turn was followed by precipitation of MQ. Megaquartz, as previously mentioned, commonly fills secondary porosity which probably was produced by volume loss and differential dissolution during dolomitization/neomorphism. Figure 60 summarizes the general carbonate/silica diagenetic relationships for the Kendall Formation.

The carbonate/silica diagenetic history of the Sulky Formation is somewhat different than that of the Kendall River Formation. Very little fine detail of the primary carbonate fabrics is preserved either in the carbonate or in the replacive chert. Most chert is coarser,

Figure 60. Relative timing of silica and carbonate diagenesis for Kendall River and Greenhorn Lakes Formations (A) and Sulky Formation (B). COMP- compaction, DOLO/NEOMORPH- dolomitization/neomorphism, SIL/EVAP- silicification of evaporites.

clear-to-tan replacive MGQ which likely formed after initial neomorphism of the host carbonates. This relatively later initial silicification is believed to be the cause of the lesser organic content (more white-tan vs. black chert) of the Sulky Formation cherts. The generally coarser grained texture of replaced micritic matrix and the lack of preservation of detail such as finely laminated ooids (compare Plate 19D with Plate 4F) also is accounted for by relatively late silicification. Because nearly all lithologies in the Sulky Formation are now dolomite, this major recrystallization episode probably was produced by regional dolomitization as opposed to a dolomite-to-dolomite aggradational neomorphic event.

Following this major carbonate diagenetic event, dolomite was replaced by tan MGQ and MQ, and drusy MQ was precipitated to fill the secondary porosity (Plate 19D). Minor EQ and hydrocarbon infill occurred as final pore-fill events.

A somewhat different silicification history is recorded in member C of the Sulky Formation on the September Lake High, where silcrete formed syndepositionally during subaerial exposure. Closely associated in space and time was silicification of a stratiform evaporite breccia. Timing of silicification of this breccia relative to carbonate diagenesis of surrounding units has not been determined, but

silicification must have occurred early in the depositional history, because clasts of chert from this breccia have been reworked to form a lag deposit capping the breccia.

All petrographic evidence indicates that silicification through the MQ precipitative phase was complete before the onset of major compaction and stylolitization.

To summarize, significant differences between Kendall River Formation black cherts and Sulky Formation white-tan cherts are apparent, in terms of texture, timing of formation and relation to carbonate diagenesis. Specifically, the early precipitative silica fabrics (CO and MGQ lephospheres) were formed only in the Kendall River Formation cherts, and the degree of preservation is much better in these cherts. Early silicification by fluids of relatively high dissolved silica concentration (40-60 ppm) apparently was important in silicification of the Kendall River Formation (and by inference, of the Greenhorn Lakes Formation), but not in silicification of the Sulky Formation. The reason for these differences probably is related to initial differences in the source of silica for these two units, as discussed below.

SOURCE OF SILICA

Without oxygen isotope data, definite conclusions about the nature of silica source for these

two formations cannot be presented, but some general statements nevertheless can be made. Obviously, the source of silica for Precambrian cherts was not biogenic. opal such as sponge spicules, the most commonly suspected source of silica for Phanerozoic shallow-water replacement cherts. Thus, the silica either must come directly from oceanic waters, implying early synsedimentary diagenesis, or it must have been provided by groundwater which would derive its silica as a result of dissolution of either clay minerals or detrital quartz within the sedimentary sequence. Although a synsedimentary origin of some of the chert in the Kendall River Formation is probable, as is suggested by the presence of interpreted opal-CT lephospheres, the relatively late, post-regional-diagenetic silicification in the Sulky Formation indicates that silica must have been introduced by formation waters. In addition, the near absence of quartz, either as clays or detrital silica in the upper two members of the Sulky Formation, suggests that silica must have been brought in from a relatively distant source. The transport distance required for the supply of silica to the site of precipitation implies the development of a major phreatic groundwater system before silicification. This observation is consistent with post-dolomitization silicification in the Sulky Formation.

PLATE 18. SILICA DIAGENESIS

A. Chert nodules elongate, parallel to bedding in Kendall River Formation.

B. Closeup of A of this Plate showing differential compaction of carbonate around early silicified chicken-wire-textured algal mat fabric. Chert nodule is 10 cm wide.

C. Hollow concretionary chert nodules from member C of the Sulky Formation. Note elongation parallel to bedding.

D. Ooid grainstone of Kendall River Formation showing early CO rim coating grains, followed by pore-fill of isolated lephospheric MGQ (light dots) in organic rich matrix. Plane-polarized light. Width of field is 3 mm.

E. Shelter pore in ooid-intraclast packstone of Kendall River Formation filled initially by pore-fill MGQ (PM) displaying lephospheric morphology towards the pore interior. MGQ is followed by SC (S) and finally by MQ (M). Note carbonate inclusions and overall darker nature of MGQ which replaces intraclasts (RM). Plane-polarized light. Width of field is 1 mm.

F. Replacive MGQ showing dense chert in ooid core and inclusion-rich rims. Pore-fill MGQ is white, and relatively inclusion-free. Plane-polarized light. Width of field is 3 mm.

PLATE 18

PLATE 18

A

B

C

D

E

F

PLATE 19. SILICA DIAGENESIS

A. Spherulitic chalcedony crust showing straight junctions between individual spherulites and 0.1 mm spacing between spherulite nucleation sites. Ooid grainstone of Kendall River Formation. Cross-polarized light. Width of field is 1 mm.

B. Distinctive fabric of zebraic chalcedony filling a pore in an ooid grainstone of the Kendall River Formation. Straight and 120 degree boundaries between spherulites are typical of pore-fill fabrics. Late dolomite rhombs overprint chalcedony. Cross-polarized light. Width of field is 2 mm.

C. Chalcedony spherulites overprinting fibrous carbonate cement in member B of the Sulky Formation. This photo shows only initial stages of growth and coalescence (two spherulites touching). Advanced-stage replacement covers several square centimetres of host carbonate. Cross-polarized light. Width of field is 5 mm.

D. Drusy MQ filling oomoldic porosity in ooid packstone of Member C of the Sulky Formation. Cross-polarized light. Width of field is 2 mm.

E. EQ rosette from upper marker bed of member C of the Sulky Formation. Core of rosette is actually MQ which grades outward into larger quartz prisms which contain abundant carbonate inclusions. Euhedral morphology is only well developed on the margin of the rosette. Cross-polarized light. Width of field is 5 mm.

F. MQ replacing relic evaporite lath texture. Relic lath-boundaries are outlined by fine dark inclusions. Larger light inclusions are anhydrite. Member C of Dease Lake Formation. Cross-polarized light. Width of field is 3 mm.

PLATE 19

PLATE 19

A

B

C

D

E

F

PLATE 20. SILICA DIAGENESIS

A. Pore in ooid packstone of Kendall River Formation, lined initially by CO isopachous rim, which is succeeded by SC. A distinct break can be seen between the two growth stages. Ooid grainstone of Kendall River Formation. Cross-polarized light. Width of field is 0.5 mm.

B. Contrasting nature of brown replacive MGQ (replacing intraclast) and white pore-fill MGQ which lacks carbonate inclusions. Intraclast grainstone of the Kendall River Formation. Plane-polarized light. Width of field is 5 mm.

C. Dolomite rhomb cement, now silicified and overlain by SC cement. The inferred paragenetic sequence is: dolomite cementation; SC cementation; replacement of dolomite cement and ooid by silica. Ooid grainstone of the Kendall River Formation. Plane-polarized light. Width of field is 0.7 mm.

D. Silicified dolomite rhombs in silicified micritic matrix. Fabric indicates that initial stages of dolomitization preceded silicification. See text for discussion. Dolomite of Kendall River Formation. Cross-polarized light. Width of field is 0.5 mm.

E. Pressure solution of differentially silicified ooid grainstone of the Kendall River Formation. The inferred paragenetic sequence is: differential silicification of some ooids; pressure solution; final silicification. Cross-polarized light. Width of field is 5 mm.

F. Euhedral dolomite rhombs overprint MQ pore-fill. See also Plate 19B. Ooid grainstone of the Kendall River Formation. Cross-polarized light. Width of field is 0.7 mm.

PLATE 20

PLATE 20

A

B

C

D

E

F

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Figure 11. Meas

CF-

2

LEGEND

thin-bedded dolostone

transitional lithology

crossbedded dol. arenite

SH biostrome

tufted biostrome

thrombolitic biostrome

beach rock

marker unit

CF

10F

Continuous with Fort Confidence Formation.

7
MEMBER B

3053

Figure 15. Measured section

15. Measured sections of the Kendall River Formation.

9

10

11

12

20 m cover

10m cover

50F

60F

14 m cover

70F

8 of

90F

7

14 m cover

10 of

8 m cover

20 m cover

25 m cover

LEGEND

LITHOLOGY

- shale
- thinly intercalated shale
- dolomite
- packstone-grainstone
- algal-laminated dolomite

STROMATOLITIC STRUCTURES

- LLH
- SH-C SH-V
- oncolite
- smooth mat
- pustulose, crenulate mat
- thinly interlayered algal mat
- chickenwire mat

SEDIMENTARY STRUCTURES

- current lamination
- crossbedding
- herringbone crossbedding
- ripple cross stratification
- channel
- intraclast conglomerate
- stone rosette
- intraclast conglomerate
- fenestrae
- halite hoppers
- mud cracks

11 OF

intercalated shale/dololomite

fine-grained limestone

laminated dolostone

STRUCTURES

SH-V

mat

120F

fine, crenulate mat

interlayered algal mat /dololomite

fine mat

STRUCTURES

lamination

bedding

fine crossbedding

cross stratification

conglomerate

settle

conglomerate lens

pebbles

INDEX MAP

130F

- intraclast conglomerate
- fenestrae
- halite hoppers
- mud cracks
- syneresis features (molar)
- tepee structure
- ooids
- rounded intraclastic dolomite

140F

intraclast conglomerate

stone rosette

intraclast conglomerate lens

fenestrae

halite hoppers

mud cracks

syneresis features (molar tooth)

tepee structure

oids

rounded intraclastic dol arenite

150F.15

300

10F

•

20F

190

Figure 21. Measured sections of member A of the Sulf

30F

0 km
|

STRUCTURES

- Ⓐ COLUMNAR STROMATOLITE
- / STONE ROSETTE
- ~ SMALL-SCALE FOLDING
- 5 - LITHOFACIES ABBREVIATIONS

of the Sulky Formation.

40F4

10F

Figure 24. Measured sections of membrane

100
1

11
V

12
V

13
V

14
V

15
V

of member B of the Sulky Formation.

PLATFORM LITHOFACIES

- CURLY BEDDING
 - CURLY-BED BRECCIA
 - THIN-BEDDED DOLOSTONE (B2)
- (B1)

30F

0 km

40F4

16 V

17 V

18 V

19 V

20 V

SEPTEMBER LAKE BUILD-UP

STROMATOLITE TYPES

BRANCHING CONICAL-COLUMNAR

COLUMNAR

CONICAL-COLUMNAR

FIBROUS CEMENT

DOMAL

300

10F

200

20F.

100

0 km
|

sections of member C of the Sulky Formation.

CIES

IGLOMERATE

STALLIZED DOLOSTONE

AL-COLUMNAR

R, INCLINED

SEPTEMBER LAKE HIGH LITHOFACIES

STRATIFORM CHERT BRECCIA

PISOLITE

TEPEE-FENESTRAL

ENTEROLITHICALLY-FOLDED DOLOLUTITE

40F4

10F

20F

3

4

E

D

C

B

A

30F

5

6

E

A

40F

150

Figure 34. Measured sections of the Greenhorn Lakes Formation.

50F

150

Figure 34. Measured sections of the Greenhorn Lakes Formation. A

60F

on. A-E are member subdivisions.

INSTONE

FRACLAST PACKSTONE

ORMITY

70F

300

15

16

17

8 of

300

17

340 km

18

300

C

400

500

B

600

700 m

A

90F

100

A

10 of

11 of

GREY DOLOSTONE

MEMBER B BIOSTROMES

SHALE

DOLOLUTITE

SHALY DOLOLUTITE

UPPER MEMBER C CONICAL-COLUMNAR BIOSTROME

ISOLATED BIOHERMS

OOID GRAINSTONE

OOID-INTRAGLAST PACKSTONE

DISCONFORMITY

KARST BRECCIA

QUARTZ ARENITE

QUARTZOSE DOL ARENITE

TUNGUSSOID BIOSTROME

120F

ONE

130F

140F/4

Figure 37. Stromatolite elongations from Greenhorn Lakes

member B (solid rose diagrams) and member D (open rose diagrams)

10 F

LEGEND

x - number of measurements

1,2 - biostrome number

□ - outcrop pattern, Greenhorn Formation

● - limit of cover

KM

20F

Lakes Formation
rose diagrams).

mation

30F

INDEX MAP

KM

0 5 10 15

Formation

BEBENSEE LAKE

G.B.L.

5 OF

INDEX MAP

6056

10F

20F

M

n = 19

100

n = 53

GREENHORN
LAKES

30F

n=53

Figure 50. Map showing geologic setting and distribution of stratabound and linear

- 1) Fort Confidence Formation, 2) member A, Dease Lake Formation,
- 3) member B, Dease Lake Formation, 4) member C, Dease Lake Formation
- 5) symbols for three domains of member C deformed zone, from left to
- 6) Kendall River Formation, 7) member A, Sulky Formation, 8) member B
- 9) linear deformed zones. HB-8 and HB-9 are Hornby Bay Group units

linear deformation zones.

n,

formation, with C1 breccia in black,

left to right, 3,2,1,

umber B, Sulky Formation,

nits

1

2

3

4

5

6

60F

6

7

8

9

km

70F

67°00'
118°15'

8

8 of 8

10
km

U

U

Figure 55. Distribution

10F

tribution of cement types, member B of Sulky Formation. Earliest cement is closest to column. Lithologic symbols

20F

geologic symbols as in Figure 24.

3053

END

06109184

FIN