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Geology and Depositional History of the Hornby Bay Group, Proterozoic, Northwest Territories, Canada

University — Université

Carleton University

Degree for which thesis was presented — Grade pour lequel cette thèse fut présentée

Ph.D.

Year this degree conferred — Année d'obtention de ce grade

1984

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GEOLOGY AND DEPOSITIONAL HISTORY OF THE HORNBY BAY GROUP, PROTEROZOIC,
NORTHWEST TERRITORIES, CANADA

by

© Gerald M. Ross, B.Sc.

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

Department of Geology

Carleton University
Ottawa, Ontario
October 28, 1983

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ABSTRACT

The Hornby Bay Group is a 3km thick succession of predominantly arenaceous rocks that compose the base of the Coppermine Homocline, a 10km thick sequence of shallow marine to subaerial sedimentary rocks in northern Bear Province. Rhyolite lavas in the upper part of the Hornby Bay Group and dates on basement rocks constrain the interval of deposition to between 1.84 to 1.66 Ga. The history of sedimentation recorded in the Hornby Bay Group is one of adjustment of fluvial drainage and sea-level to changes in the magnitude and orientation of post-orogenic crustal tilt that preceded formation of the overlying Dismal Lakes Group platform.

The lower part of the Hornby Bay Group comprises conglomerate and arenite deposited in fluvial and less common aeolian depositional systems. Deposits of the oldest system (Bigbear Formation) record infill of a north-trending, east-dipping basin by fluvial sediments and formation of an extensive ($>4000\text{km}^2$) land-locked aeolian sand sea (erg). Syndepositional dip-slip reactivation of basement faults, a consequence of crustal uplift, is manifest in the occurrence of coarse intraformational conglomerates and a change in paleodrainage from southeast to southwest in the upper part of the Bigbear Formation. Fault River Formation records deposition of fluvial facies that infilled a southwest-trending topographic trough that was due partly to syndepositional fault movement west of the Bigbear system. A change from uplift to subsidence of a southwest-dipping paleoslope led to burial of Bigbear and

Fault River Formations and the deeply incised Teshierpi-Bigtree peneplain by deposits of the Lady Nye fluvial system, a river system of significant dimensions (hundreds of kilometres long) that deposited an upward-fining sequence of conglomerate and arenite. The Lady Nye braidplain terminated in a north-northwest-facing delta complex developed in a high-energy mesotidal marine setting. The East River Formation, a relatively thin unit of stromatolitic dolostone, records drowning of the Lady Nye delta complex during marine transgression and initiation of carbonate deposition on a high-energy northwest-facing carbonate ramp. Regional uplift led to ramp shoaling and deposition of peritidal carbonates that were later buried by a clastic wedge composed of cannibalized East River Formation rocks. The Kaertok Formation, a sequence of regressive deltaic-marine siliciclastics, was deposited by a fault-controlled fluvial system that prograded over the East River ramp. The top of the Kaertok Formation marks the termination of Hornby Bay Group deposition and onset of a second siliciclastic-to-carbonate cycle, the Dismal Lakes Group.

The Narakay Volcanic Complex, a bimodal basalt-rhyolite volcanic succession in Dease Arm of Great Bear Lake, consists of mafic pyroclastic rocks and younger felsic pyroclastic rocks and lavas interbedded with the upper three formations of the Hornby Bay Group. A swarm of east-trending feldspar-phyric dykes may record the effects of north-south extension that led to formation of the Dismal Lakes platform.

ACKNOWLEDGEMENTS

It is a pleasure to acknowledge my advisor, Dr. J.A. Donaldson, who maintained a high level of confidence and enthusiasm in my work, and believed that it was indeed possible to transform a metamorphic petrologist into a sedimentologist. An NSERC grant to Dr. Donaldson and an Epstein Scholarship from Carleton University supported me during my tenure in Ottawa. Dr. F.H.A. Campbell, Dr. M.B. Lambert, and Dr. P.F. Hoffman, were among many friends at the Geological Survey of Canada who offered criticism and discussion of various aspects of this thesis. In particular, Dr. Campbell improved the quality of the thesis immensely through his incisive editing and commentary.

Field support was provided by Department of Indian Affairs and Northern Development, courtesy of W.A. Padgham, Chief Geologist. In addition, numerous mining companies and individuals provided helicopter support: Hudson Bay Oil and Gas (J. Whittstock), Gulf Canada (J. Thompson), BP Minerals (G. Campbell), Imperial-Esso (F. Hassard and G. Woollet), INCO (E. Debycki), and AGIP (M. Male). Charlie Kerans, my co-worker who mapped the Dismal Lakes Group, is thanked for his assistance and enthusiasm during this study and also set a high standard of excellence. Samuel Bowring (University of Kansas) performed the U-Pb zircon geochronology and also helped discuss various aspects of the geology of the Narakay Volcanic Complex and Wopmay Orogen. Jeff Chiarenzell and L.B. Aspler kept the office atmosphere lively and stimulating. Val

Chiarenzelli is thanked for her patient and excellent production of the photos used for the plates. Terri Smith helped with drafting and pitched in during the critical final days of the thesis.

Finally I would like to dedicate this thesis to two close friends, Dr. Daniel Ross, Jr., (my father) and Mrs. Elsy G. LeVino (my grandmother) who supported me throughout the duration of this thesis but whose untimely deaths did not allow them the satisfaction of seeing it completed.

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- E. Bedding in bedded facies of Narakay Lithic Tuff.
- F. Photomicrograph of crystal-rich rhyolite ash tuff.

Plate N6

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- A. Banding in welded lapilli tuff of Narakay Lithic Tuff.
- B. Devitrification(?) structures in Narakay Lithic Tuff.
- C. Photomicrograph of feldspar-phyric dyke.
- D. In situ explosion breccia.
- E. Breccia dyke in feldspathic arenite of Lady Nye Formation.
- F. Hand sample of intrusive breccia.

CHAPTER 1

INTRODUCTION

The Coppermine Homocline (also referred to as Amundsen Basin; Christie, et al., 1972) is a thick (>10km) sequence of sedimentary and volcanic rocks in the northwest Canadian Shield (Fig. 1-1). The Hornby Bay Group, a succession of unmetamorphosed and relatively undeformed Middle Proterozoic sedimentary and volcanic rocks, constitutes the lowermost 3-4km of the Homocline. The Hornby Bay Group occurs throughout an area of more than 5000km², from Coronation Gulf to Great Bear Lake, and traverses a variety of countryside, from densely wooded bush (Leith Peninsula; southern Great Bear Lake; Fig. 1-3) to barrenlands north of the Arctic Circle. Outcrop quality and continuity is quite variable; glacial meltwater channels in bedrock and roches moutonnées provide excellent exposures that are separated by large areas of gently rolling tundra covered by glacial till ("golf course") or a thin veneer of blocks of frost-heaved bedrock ("boulder deserts").

This study is part of a larger project in basin analysis that entailed mapping of the lower two groups (Hornby Bay and Dismal Lakes Groups) of the Coppermine Homocline. The project was initiated by J.A. Donaldson and field work by Donaldson, C. Kerans (Dismal Lakes Group), and G. Ross (Hornby Bay Group) was carried out during the summers of 1978 to 1981. Collaborative mapping of both the Hornby Bay and Dismal Lakes Groups

Figure 1-1 Generalized geology of the northwestern Canadian Shield. HB: Hudson Bay; LA: Lake Athabasca; GSL: Great Slave Lake; AA: Athapuscow Aulacogen; MF: McDonald Fault; BF: Bathurst Fault; KB: Kilohigok Basin; CG: Coronation Gulf; GBL: Great Bear Lake; Modified from Lewry and Sibbald (1980) and Hoffman (1980).

led to formulation of an integrated model of deposition and stratigraphy in this area (Kerans, et al., 1981). This thesis presents the results of in-depth sedimentologic and stratigraphic analysis of the Hornby Bay Group and builds a framework for future studies of more detailed nature.

The scale of mapping varied depending on the quality of outcrop. Well-exposed rocks in the Coppermine River and Fault River regions were mapped at a scale of 1:50,000, whereas areas of less contiguous exposure were mapped at 1:100,000 scale. The Narakay Islands, in Dease Arm of Great Bear Lake, contain exceptional exposures that were mapped at a scale of 1:5000. Access to the region is limited; camps were put in by bushplanes chartered from Yellowknife, about 600km to the south. Most traverses were conducted on foot or by boat although helicopter support was used to reach especially remote areas.

PREVIOUS WORK

The results of this thesis and the work of Kerans (1983) on the overlying Dismal Lakes Group, represent the first comprehensive regional analysis of these groups. The region was visited by several geologists after the turn of the century but most work was of reconnaissance nature and/or directed towards exploration for amygdules in the Coppermine River Group basalts, that are filled with native copper (Bell, 1901; O'Neill, 1924; Sandberg, in Douglas, 1913). In the late 1950s, the region was visited by Fraser, et al., (1960) as part of a

FRASER ET AL. (1960)	BARAGAR AND DONALDSON (1973)	KERANS ET AL. (1981)	THIS STUDY
HORNBY BAY GROUP (DOLOSTONE)	DISMAL LAKES GROUP	DISMAL LAKES GROUP	GREENHORN LAKES FORMATION
			BULKY FORMATION
			KENDALL RIVER FORMATION
			DEASE LAKE FORMATION
			FORT CONFIDENCE FORMATION
			MOUNTAIN LAKE FORMATION
HORNBY BAY GROUP (SANDSTONE)	HORNBY BAY GROUP	HORNBY BAY GROUP	KAERTOK FORMATION
			EAST RIVER FORMATION
			LADY NYE FM
			FAULT RIVER FM BIGBEAR FM.

Figure 1-2. Evolution of stratigraphic nomenclature for the Hornby Bay and Dismal Lakes groups. Numbers refer to informal unit nomenclature.

Geological Survey of Canada mapping project in the northwest barrenlands. At that time, the Hornby Bay Group was divided into a lower sandstone unit and upper dolostone unit (Fig. 1-2). Subsequent mapping by Donaldson (1969, 1971), Baragar and Donaldson (1970, 1973), and Cook and Aitken (1971) led to publication of two 1:250,000-scale map sheets that covered about half of the outcrop area of Coppermine Homocline rocks. Baragar and Donaldson (1973) subdivided the Hornby Bay Group (as defined by Fraser, et al., 1960) and gave the "dolostone" unit group status (Dismal Lakes Group). Based on map patterns these authors postulated that an unconformity was present between the Hornby Bay (Unit 11) and Dismal Lakes (Unit 12) Groups.

Initiation of the recently completed program of re-investigation and detailed mapping of the Hornby Bay and Dismal Lakes Groups occurred in 1978, about midway through the boom in uranium exploration in the region. As a consequence of local detailed drilling and mapping in the Teshierpi Mountain area (work summarized in Yeo, 1980), and regional mapping (Ross, 1980), the position and nature of the unconformity between the Hornby Bay and Dismal Lakes Groups was revised (Fig. 1-2). In general, the contact between the two groups is unconformable east of the Teshierpi Fault and conformable west of the fault, except for the region south of Teshierpi Mountain (which is west of the fault) where there is a pronounced unconformity between the two groups.

This thesis uses informal stratigraphic nomenclature that supersedes earlier informal designations of "systems" (e.g. Fault River system), "facies" (e.g. Bunn Creek facies), and "units" (e.g. Unit 10) (Kerans, et al., 1981). Formal nomenclature will be published to accompany publication of maps. Wherever possible, the informal proposal of names for units of formation and member status has followed the conventions of the recently published Code for North American Stratigraphic Nomenclature (North American Commission on Stratigraphic Nomenclature, 1983).

REGIONAL GEOLOGY

The Hornby Bay and overlying Dismal Lakes Group are the lowermost groups in the undeformed north-dipping Coppermine Homocline which in this area also contains the Coppermine River and Rae Groups (Fig. 1-3B). The Hornby Bay Group rests unconformably on Aphebian basement that ranges in age from 1.93-1.85 Ga (Van Schmus and Bowring, 1980) and is overlain with local unconformity by the Dismal Lakes Group. The Dismal Lakes Group is conformably overlain by basalt flows of the Coppermine River Group, dated at 1.2 Ga (Baragar, 1972). Both groups are cut by the Muskox Intrusion (Kerans, 1983) and by Mackenzie-age diabase dykes.

The Hornby Bay and Dismal Lakes Groups represent two cycles in excess of 4km thick, of terrestrial-siliciclastic to marine-carbonate deposition that took place over older stable cratonic crust (Wopmay Orogen: Hoffman, 1980). Deposition

occurred on a generally quiescent post-organic basement, although periodic movement along reactivated basement faults caused locally important facies changes.

The basement of the southern part of the Amundsen Basin includes the Aphebian Hepburn metamorphic-plutonic complex, Great Bear Batholith (and associated volcanic rocks) and sedimentary rocks of Coronation Supergroup (Hoffman, 1981a). The Hepburn complex is composed of foliated granitoid rocks of the Hepburn and Wentzel batholiths and metasedimentary rocks of the Akaitcho and Epworth Groups. Granitic and volcanic rocks of Great Bear Batholith and McTavish Supergroup (see Hildebrand, 1981) probably underlie most of the Hornby Bay and Dismal Lakes Groups and contributed significant detritus to the basal Hornby Bay Group.

Northwest- and northeast-trending faults significantly influenced the depositional history of both the Hornby Bay and Dismal Lakes Groups. According to Hoffman (1981b) and Hoffman and St. Onge (1981), these are conjugate wrench faults that formed as a result of Aphebian plate collision west of the present exposure of Wopmay Orogen and accommodated regional shortening through strike-slip and reverse motion. These faults were reactivated, with a dip-slip and oblique dip-slip sense of motion, during and following deposition of the Hornby Bay and Dismal Lakes Groups. The magnitude of dip-slip offset varies considerably along strike of individual fault, implying a hinged or "piano key" sense of movement. This style of fault movement is inferred to reflect the variation in orientation,

with respect to a vertical plane, of strike-slip fault planes, a feature that is commonly observed during experimental formation of strike-slip faults (Wilcox, et al., 1973).

North-trending faults, such as the Herb Dixon Fault, Canoe Lake Fault, and unnamed faults along Leith Line and farther west (Fig. 1-3), are also an important structural component of the Coppermine Homocline. These faults are characterized by dip-slip movement that was for the most part restricted to late- to post-Dismal Lakes Group deposition (except for Leith Line). These faults parallel the trends of Mackenzie-age diabase dykes, Coppermine River Group basalt feeders, and the Muskox Intrusion, which suggests that these features formed under the same stress system. A single major east-trending fault, the East River Fault, is a low-angle fault that displays up 1km of offset. It represents a décollement surface along which post-depositional sliding of the upper formations of the Dismal Lakes Group took place.

Leith Line (Kerans, et al., 1981) is a major north-trending zone which appears to have acted as a syndepositional hinge during accumulation of upper Hornby Bay Group rocks and the entire thickness of the Dismal Lakes Group. Leith Line extends from the south shore of Great Bear Lake (Leith Peninsula) north through the eastern (and deepest) part of Great Bear Lake, and continues northward through the Hornby Bay and Dismal Lakes Groups (Fig. 1-3). Indirect evidence of this feature is suggested by the concentration of north-trending (presumably Mackenzie-age) dykes in areas west of Leith

Line, lack of basement exposures west of the Line, pronounced magnetic quiescence west of the Line, a change in the sense of displacement across the Teshierpi Fault where it crosses the Line (from northwest-side down east of the Line to southeast-side down west of the Line), and a sharp change in the strike of map units across the Line (eastward in the east; northward in the west). Consistent westward thickening of map units and inferred deepening of depositional environments of western facies of both the Hornby Bay and Dismal Lakes Groups indicates greater subsidence west of Leith Line. Steeply-dipping to vertical and locally overturned beds of the Hornby Bay Group along this line (north of Teshierpi Fault) suggest locally significant post-depositional movement. The Leith Line is interpreted to reflect the presence of a fundamental tectonic boundary, probably between Great Bear Batholith (on the east) and basement of uncertain composition on the west (possibly part of zone 5 exotic terrane of Wopmay Orogen) such that the magnitude of crustal subsidence during deposition of the Hornby Bay and Dismal Lakes Groups was greater west of the Line.

The Hornby Bay Group is an approximately 3-4km thick succession of shallow-water sediments dominated by fluviatile and marine siliciclastics, with subordinate aeolian siliciclastics and marine dolostones (Fig. 1-4). At no one place can the entire thickness of 3-4km of sediment be observed. The basal unit of sandstone and conglomerate, formerly referred to as Unit 8 (Baragar and Donaldson, 1973; Kerans, et al., 1981),

Figure 1-4. Simplified stratigraphy and interformational correlations within the Hornby Bay Group. Bunn Creek conglomerate, at the base of the Lady Nye Formation, is a distinct conglomerate that also occurs at the top of the Bigbear and Fault River Formations. Black bar indicates duration of activity in Narakay Volcanic Complex. Columns are arranged in east to west sequence.

was deposited in three separate but interrelated and overlapping depositional systems that have been given provisional formational status (Bigbear, Fault River, and Lady Nye Formations; Fig. 1-4). Facies patterns and paleocurrents in these rocks do not reflect control by fault activity related to the final stages of compression in Wopmay Orogen; therefore, indicating a cessation of orogenic activity prior to deposition of Hornby Bay Group sediments. The Bigbear and Fault River systems were in part coeval, and both are conformably overlain by Lady Nye system deposits. Lady Nye Formation fluvial facies grade laterally and vertically into deltaic-marine siliciclastics and a distinctive unit of stromatolitic dolostone, the latter composing part of the East River Formation. Carbonate rocks were deposited along a northeast-trending shoreline that migrated to the east during marine transgression. Basement uplift resulted in progradation of deltaic-marine siliciclastics of the Kaertok Formation over carbonate algal marshes of upper East River Formation. Reactivation of a major northeast-trending basement fault (Teshierpi Fault) during regional uplift prior to Dismal Lakes Group deposition resulted in an unconformable relationship between the Hornby Bay Group and Dismal Lakes Group on the upthrown fault block. The Narakay Volcanic Complex, a sequence of bimodal rhyolite-basalt volcanic rocks, was deposited during deposition of the upper three formations of the Hornby Bay Group.

CHAPTER 2

BIGBEAR FORMATION

The Bigbear Formation, comprising sediments deposited in the oldest siliciclastic depositional system in the Hornby Bay Group, occurs as relatively continuous outcrop north and east of the Big Bend in the Coppermine River, and as remnants preserved in fault-bounded outliers along Wopmay Fault (St. Germain graben, Zephyr graben, Belleau graben) and Sinister Fault (First Mountain outlier) (Fig. 2-1). The formation is named for exposures in Bigbear Creek, and although not all members of the formation are exposed in the creek section, the name is retained from earlier informal usage (Ross, 1980; Kerans, et al., 1980, 1981). Lithostratigraphic correlation between the outliers and the main region of exposure indicates that the Bigbear Formation once covered an area of at least 4000km². Erosion of the Bigbear Formation occurred at least three times following deposition; 1) erosion of western parts of the formation during deposition of the Fault River and Lady Nye Formations; 2) erosion of the formation east of Teshierpi Fault (Fig. 1-2) during deposition of basal Dismal Lakes Group clastics (Kerans, et al., 1981), and 3) erosion during northward tilting of the Coppermine Homocline.

The Bigbear Formation rests unconformably on Wopmay Orogen rocks that are composed of granitoid and metamorphic supracrustal rocks east of Wopmay Fault (Hepburn metamorphic-plutonic complex; Fig. 2-1) and volcanic and plutonic rocks

FIGURE 2-1: Generalized geologic map showing setting and principal subdivisions of the Bigbear Formation. Note the post-Hornby Bay Group graben structures and the location of the Mist Mountain outlier. C=Lady Nye Formation.

west of the fault (Dumas Group and Great Bear Batholith, respectively). Although most of the Bigbear Formation is unconformably overlain by basal clastics (Unit 11) of the Dismal Lakes Group, exposures south of Mouse River (Fig. 2-3, section E) show that the Bigbear Formation is also conformably overlain by the Lady Nye Formation of the Hornby Bay Group (Fig. 1-4). This relationship provides the basis for regional interbasinal correlation, and establishes the relative timing of the different depositional systems.

The Bigbear Formation is composed almost exclusively of arenite, with subordinate monomictic and polymictic conglomerates and red mudstones. Seven map units (members) have been recognized (Fig. 2-2) and these are inferred to record three successive assemblages of basin fill that are referred to as "stages", although this does not imply that these units are distinct time-stratigraphic units. Stage 1 comprises basal alluvial fan-fluvial facies that are overlain by the largely eolian facies of Stage 2. Stage 3 deposits are a mixed fluvial-eolian(?) facies formed during conditions of tectonic activity, expressed as synsedimentary normal faults, that led to partial cannibalization of earlier deposits and drainage basin reorganization.

STAGE 1

Deposits of Stage 1 comprise a basal paleosol-silcrete unit and overlying units of arenite and conglomerate (members A1-A3) that define an upward-fining and thinning succession.

FIGURE 2-2: Composite stratigraphic column for the Bigbear Formation. Thickness (left side of column) is in metres.

The maximum measured total thickness for all three members is 225m (Fig. 2-3; section D). Thickness variations within the basin are difficult to accurately determine because of pinch-out (lap out) of members and generally poor exposure of the unconformity. Two observations are pertinent: at Mist Mountain outlier, Stage 1 deposits are drastically thinned and comprise only 10-20m of immature conglomerate interlayered with Stage 2 eolian arenites, and along the western edge of the main outcrop area (south of Bigbear Creek, Fig. 2-3), member A2 is absent and Stage 1 units are less than 60m thick (Fig. 2-3, section B). This suggests that both the Mist Mountain region and the western edge of the outcrop belt were paleohighs during Stage 1 deposition. Deposition of Stages 2 and 3 was influenced by a paleohigh south of Bigbear Creek, but not at Mist Mountain outlier. This implies the existence of a crudely north-trending depression during Stage 1 deposition and that, from evidence presented later, the northwestern limit of outcrop is close to a syndepositional basin margin.

Present exposures of basement, especially north of the Big Bend in Coppermine River (Fig. 2-3), where lower Stage 1 lithologies are best exposed, form topographically rugged knobs (Plate B1A). Members A1 and A2 onlap these knobs and, in addition, patches of paleosol and basal conglomerate (member A1) occur as remnant outliers in recesses on the flanks of the present-day highs. This suggests that the present topography is an exhumed mature paleotopography, somewhat modified by Pleistocene glaciation, that had syndepositional relief of as

much as 150m; exhumed basement paleohills are morphologically similar to bornhardtts (cf. Thornbury, 1963).

Member A1

The basal unit of the Bigbear Formation (member A1) is a thin (<20m thick) but regionally extensive unit that has been recognized in both the Mist Mountain outlier and St. Germain graben, and is widespread north of the Big Bend in Coppermine River (Fig. 2-3). Three lithofacies have been recognized within this member, but have not been mapped in sufficient detail to allow differentiation on the map in Figure 2-3. The lower facies is a massive paraconglomerate composed of locally-derived basement fragments dispersed in a sandy siliceous matrix (Plate B1B). The upper facies is also composed of paraconglomerate, but contains sandy interbeds and has tighter framework packing. The third facies comprises micaceous siltstones and granular arenites that occur in scattered outcrops and is of uncertain stratigraphic position.

Lower Facies

The lower facies of member A1 occurs as relict patches preserved within re-entrants in basement paleohighs and at the bases of inferred paleovalleys. Two subfacies have been recognized: massive siliceous paraconglomerate, which is most common, and a single occurrence of stratified siliceous mudstone, arenite, and conglomerate.

All of the rocks that compose this facies contain abundant evidence of syndepositional to early diagenetic silica

cementation, and are therefore referred to as silcretes. Descriptions and interpretations of these rocks are only preliminary; much more work is required to map out subfacies, fully describe and decipher petrographic features and mineralogy, and to compare these with Cenozoic and younger silcretes (cf. Australia: Langford-Smith, 1978). Silcretes are both indurated soils and sedimentary rocks, and the nomenclature of their textural features has not been adequately established. In the following descriptions, sedimentologic terminology will be used and, where appropriate, the relevant soil term (e.g. Brewer, 1964) will be included in parentheses. There is little agreement among workers as to a suitable definition of silcrete (see Langford-Smith, 1978, for range in definitions). Silcrete will be used to identify a silica-cemented, indurated sedimentary rock in which the cementation was essentially syn-sedimentary. In the examples described here, the synsedimentary origin of the cementation is obvious, because silcrete occurs as clastic grains (intraclasts) within basal member A1, and as pebbles in upper member A1 and in overlying members A2 and A3.

Basement rocks that underlie member A1 are surprisingly fresh up to less than 1m from the unconformity. Biotite, which is particularly susceptible to pedogenic alteration (Loughnan, 1969) occurs as nearly pristine grains in both biotite-bearing dacitic tuffs of the Dumas Group and high-grade metapelites of the Epworth Group that are in contact with member A1. Granitic rocks were, however, more susceptible to weathering because

Fe-silicates and potassic feldspars are altered to hematite, but only to depths of less than 5 metres. This difference in weathering is also reflected in the composition of clasts in the overlying conglomerates, i.e. relatively impermeable felsic volcanics are overwhelmingly more abundant than granitic clasts. Within 1m of the unconformity, basement rocks (except for granitic rocks) display a weak mottled texture defined by purple to red hematite patches (1-2cm wide) in a relatively bleached pale orange groundmass. Basement cobbles within A1 conglomerates display this same range in alteration, which suggests that the thinness of the zone of basement weathering reflects climatic and pedogenetic characteristics and is not the result of truncation by erosion.

Siliceous Paraconglomerate Subfacies. This subfacies, the predominant constituent of the lower facies of member A1, is a massive paraconglomerate composed of angular clasts of locally derived basement rocks dispersed in a matrix of silica-cemented arenite (Plate B1B). In many areas, coarse blocky dolomite comprises a significant portion of the matrix (calcareous paraconglomerate of Kerans, et al., 1981). The subfacies rests sharply on underlying basement. In one outcrop of the contact, large (1mx2m) irregular block-like bodies of rhyolite are bounded by thin (<5cm wide) fractures filled with silica-cemented sand identical to the matrix of the conglomerates. Similar sand vein-fill also occurs in vertical veins that may extend as much as 15m down from the unconformity. With the exception of the Mist Mountain outlier, clasts are composed

almost exclusively of Dumas Group felsic volcanic rocks. The clasts are extremely angular, show no signs of mechanical abrasion, and float in the matrix. In some samples, in situ fragmentation can be seen where delicate angular fragments have spalled off the parent clast.

The matrix to the clasts, which composes up to 70% of the rock, displays a variety of textures that attest to the complex history of cementation. Three textural-mineralogic components have been recognized (Plate BlC): a framework of silica-cemented sand grains that are disrupted by b) irregular vugs filled with microcrystalline silica and c) later veins filled with lamellar-botryoidal megaquartz, chalcedony, and quartzine.

The matrix framework is composed of angular to rounded, well-sorted sand grains, <1mm, that are cemented by silica. The sand grains are composed of monocrystalline (volcanic) quartz, feldspar (microcline, plagioclase, and untwinned feldspar), micaceous lithic fragments (reworked peds?), micas, and rare green tourmaline and zircon (Plate BlE). The feldspars range from pristine to those that contain an abundance of fine opaque inclusions that impart a dusty appearance. These constituents are cemented by syntaxial quartz overgrowths, in cases where the framework is relatively quartz-rich, or as is more common, isopachous chalcedonic rim cements (chalcedans; Plate BlE). These latter cements are composed of fibrous length-slow, and less commonly, length-fast chalcedony which has been partly recrystallized to flamboyant prismatic

microquartz (terminology of Milliken, 1979). Micaceous grains, which appear to be mostly muscovite and/or illite, occur as thin films around other framework grains (argillans), as framework grains with a weak regular preferred orientation (latisepic fabric), or as scattered "expanded" grains (concertina structure of Watts, 1978). In the latter case, fine (<5 μ) fibrous length-slow chalcedony (quartzine) has precipitated in the expanded a-b planes of the micas as fibrous crusts with c-axes perpendicular to the a-b plane (Plate BlE).

Irregular vugs (orthovugs?), that range in size from 10 μ up to 50mm in the widest dimension, are common in all samples (Plate BlF). Locally the smaller vugs are in such abundance as to impart a clotted to fenestral fabric to the rock. In general the vugs are smooth-walled, elongate parallel to the paleohorizontal (i.e. laminoid fenestrae), and are infilled with a very fine-grained intergrowth of low birefringent microcrystalline silica and scattered clay grains. This void-fill lacks internal stratification, although some vugs are floored geopetal cavities or display druses of prismatic mega-quartz that line vug walls. These void-fills clearly developed after chalcedony cementation of clastic framework, because cemented sand is crosscut and included in the microcrystalline vug-fill silica.

The third textural component of the matrix are irregular, almost pygmatic, veinlets of distinctive red, white and clear silica that crosscut earlier cements and void fills

(Plate B1C). The veinlets are composed of two textural varieties of silica that have a consistent geopetal relationship to one another, although the veinlets may be composed of only a single silica type. The margins of some vein walls are covered with a crust of coalesced spherules of length-slow chalcedony (Plate B2A). The botryoidal character of the crust is defined by concentric bands of brown inclusions (bubbles?, cf. Folk and Weaver, 1952) that impart a characteristic red colour to hand samples of this silica type. In some cases the spherules are composed of alternating length-fast and length-slow chalcedony. Inward from the chalcedony crust, crusts and spherules of prismatic megaquartz, with wavy undulose extinction, compose the bulk of the vein-fill material. (Crystals) display euhedral terminations that abut against the cemented sandstone host, suggesting a component of expansive crystallization (cf. Watts, 1978) in addition to passive infill of open fractures (Plate B2A). Brown inclusion-rich bands also mark prism terminations and record incremental growth of the megaquartz. Thin laminae (<10u) of length-slow chalcedony occur as bands within some megaquartz spherules. This type of vein-fill occurs as geopetal infill of the centers of earlier vugs that were only partly filled with microcrystalline silica, and as discrete bodies unrelated to the vugs. In one example, this type of silica defines a botryoidal-domal texture (Plate B2B) erroneously interpreted as stromatolitic (Kerans, et al., 1981). It is unknown whether these botryoids formed in a

cavity or at the surface, similar to "pillow" silcretes (cf. Wopfner, 1978) and bulbous silcretes (Ambrose and Flint, 1981).

In most outcrops of the massive paraconglomerate, coarse (2-4mm) blocky dolomite grains are so abundant in the matrix that outcrops have a deeply pitted surface due to differential weathering of the carbonate grains. Thin section observation indicates that the dolomite is relatively late in the cementation history of the conglomerates. It replaces matrix, framework, and cement, and generally obliterates earlier textures. Some of the spar is zoned as shown by inclusion-rich bands.

Stratified Subfacies. This subfacies comprises a single large outcrop that occurs in a paleovalley along the contact between Akaitcho Group metasedimentary rocks and younger Hepburn granodiorite (Fig. 2-3). This subfacies is distinctive and unique in that it is finer grained than the conglomerate subfacies, displays well-developed stratification (Plate B1D), and has strong hematite pigmentation, in contrast to the bleached character of the conglomerate subfacies. Beds are 1-10cm thick, show no internal stratification and are composed of: 1) finely laminated, siliceous argillite, 2) arenite, and 3) paraconglomerate (Plate B1D). The argillite is composed of silt-size and finer clasts in an indurated siliceous matrix. The arenite beds are similar to the arenite that forms the matrix in the conglomerate subfacies, except that the former commonly contain ferruginous oolites (sesquioxidic nodules). The conglomeratic layers are composed of angular breccia.

fragments of silcrete (intraclasts) and minor subrounded basement fragments dispersed in a silica-cemented sand matrix. Mud-filled vugs (metavugs), common in the conglomerate subfacies, are absent in this subfacies, but veins and cavities filled with clear to red and white silica (megaquartz and chalcedony) are common. The veinlets do not crosscut the rock fabric but tend to occur as fenestrae parallel to layering (Plate B1D).

Interpretation

The massive texture of the paraconglomerates and the strictly local composition of the clasts, in addition to the extreme angularity of the clasts, suggests that these conglomerates were developed in situ, presumably by pedogenic processes. The relatively thin zone of weathering within the basement implies an arid rather than humid paleoclimate (Loughnan, 1969). The arenaceous matrix in the conglomerates probably was transported by wind, as is suggested by the grain size (<1mm), high sphericity of many grains, and the lack of sedimentary structures characteristic of aqueous traction currents. Depositional breaks were not recognized within the conglomerate subfacies, but are common in the stratified subfacies. The conglomerate subfacies is interpreted as a scree slope deposit (cf. Twidale, 1981) and the stratified subfacies as "pan" or lacustrine sediments that accumulated in a paleolow. The framework components in both subfacies record an inferred sequence of pedogenic and diagenetic events:

- 1) Deposition by groundwater(?) of clay skins (argillans) around sand-size clasts. Illuviation/expansion of micas, probably through the replacement of interlayer K^+ by H_2O and oxidation of iron (cf. Isherwood and Street, 1976).
- 2) Cementation of sand-sized and finer material by isopachous, fibrous silica and precipitation of length-slow chalcedony in the a-b planes of expanded micas. In part, expansion of mica grains may have been contemporaneous with chalcedony precipitation, similar to precipitation of displacive calcite in calcretes (cf. Watts, 1978).
- 3) Disruption of lithified sediment, possibly by contraction during a phase change in authigenic silica phase(s) (cf. Kastner, 1979), to form irregular vugs. These were partially to completely infilled by massive microcrystalline silica (Plate B1E). The vugs are similar in appearance and mineralogy of the infill to those of inferred paleosol origin described from the Caballos Novaculite (McBride and Folk, 1977; Fig. 11).
- 4) Precipitation of vein- and vug-fill spherulitic chalcedony and megaquartz as infills of newly formed fractures and open spaces in partly filled earlier vugs. In the stratified subfacies, this material, rather than microcrystalline silica, infilled open spaces. The change in silica morphology in both the vein fill and framework cements (Plate B1D), from

fibrous chalcedony to megaquartz, may reflect a change in chemical composition of the pore fluids

(groundwater) (i.e. uSiO_2 , pH, etc; Wilson, 1966; Folk and Pitman, 1971; Keene, 1983).

- 5) Local replacement of silcrete by coarse dolomite spar and also recrystallization of fibrous silica (chalcedony) into prismatic to blocky megaquartz. The relative timing of these two processes is not certain, but textural relationships suggest that they were in part contemporaneous.

As silcretes are best developed in a region underlain or adjacent to felsic volcanic rocks of the Dumas Group, much of the authigenic silica in these rocks may have been derived by groundwater dissolution of silica-rich basement rocks. The silcretes in member A1 contain very fresh feldspars (Plate B1D) and do not appear to contain abundant TiO_2 . This implies that these silcretes formed by accretion of allogenic components (cf. Ollier, 1978) (both detrital and dissolved material) rather than by silicification of residual weathered basement. This is supported by the lack of a deep weathering profile in basement rocks and lack of concentration of refractory weathering products (e.g. Zr, TiO_2 , Al-rich clay minerals). In addition, the thinness of the zone of basement weathering and the low (absent?) abundance of TiO_2 suggest an arid-semiarid paleoclimate (cf. Summerfield, 1983). Length-slow chalcedony, which is a common cement in these silcretes, has been recently attributed to evaporite replacement by silica

(Folk and Pitman, 1971; Milliken, 1979). Although length-slow chalcedony is present in these rocks, no other evidence of evaporite minerals was observed. Folk and Pitman (1971) and more recently Keene (1983), suggest that the length-slow crystallographic orientation of the silica fibers reflects the pH and concentration of SO_4 of the diagenetic fluids and are not indicative of any particular depositional environment.

Upper Facies

The upper facies of member A1 ranges in thickness from 0-30m, and is composed predominantly of massive paraconglomerate with thin (<50cm thick) interbeds of arkosic arenite. The paraconglomerate occurs as massive irregular- to tabular-shaped bodies that mantle subjacent basement paleohighs (Plate B1A). It is composed of subequant, subangular cobble to boulder-sized fragments of locally derived basement rock (most commonly Dumas Group volcanic rocks) moderately packed in a sandy matrix. Arenite interbeds, which constitute less than 15% of this facies, consist of surprisingly mature arkosic to feldspathic, well-sorted, medium-grained sand cemented by hematite films and syntaxial quartz overgrowths. The arenites are mostly horizontally stratified, but in a few places, cross-laminated sets were observed.

Interpretation

The massive, matrix-supported texture typical of the conglomerates is suggestive of a mass flow (sediment gravity flow) mechanism of deposition. The occurrence of the conglomerates as mantles around basement paleohighs and local

derivation of cobbles implies that this facies accumulated as colluvial cones or rock fans on the flanks of basement paleohighs (cf. Twidale, 1981, p.11, for a modern example). The stratified arenite interbeds were probably deposited by aqueous runoff (sheetwash, sensu stricto, Hogg, 1982) with only rare development of bedforms.

Siltstone Facies

This facies occurs in isolated outcrops in the large valley immediately west of Wopmay Fault (Fig. 2-3) and has been intersected in drill holes to the east. The upper and lower contacts of this facies are not exposed; member A2 arenites are topographically above this facies in the valley west of Wopmay Fault, and therefore this facies is considered to be part of member A1.

The maximum thickness of this facies is 40m, as determined from drillcore (BP Minerals, unpublished data). It is composed of planar beds (5-20cm thick) of horizontally-laminated pebbly to granular micaceous siltstone and arenite. These rocks have a red to maroon pigmentation with green ("reduced") blotches, are arkosic in composition, and contain angular granules and pebbles of granitic and volcanic rock, as well as mudchips scattered along laminae. Bedding surfaces contain an abundance of white mica flakes, and commonly display parting (current) lineation.

Interpretation

This facies records deposition by periodic shallow flood events that transported poorly sorted immature clastic

detritus of local derivation. The parallel stratification typical of this facies suggests upper flow regime conditions; the lack of cross-stratification attests to the short duration of flood events such that bedforms did not develop. These deposits may represent distal equivalents of upper member A1 fanconglomerates and interbedded arenites.

Member A2

This member is composed of randomly interbedded, cross-stratified, coarse-grained arenite and, in the lower part of the member, massive polymictic paraconglomerate. The conglomerate-sandstone ratio, average grain size, and average bed thickness, all decrease upwards in the member as it grades into the finer grained arenites of member A3. The member sharply overlies member A1 lithologies, and also onlaps basement rocks. Member A2 occurs exclusively in exposures north and east of Coppermine River; it pinches out south of Bigbear Creek and north of Big Bend (Fig. 2-3). Its basinward extent is unknown because of lack of exposure in the Mouse graben; the member is absent in Mist Mountain outlier.

The conglomerates within this member invariably are sheets (<2m thick) of massive paraconglomerate that overlie shallow scour surfaces in the underlying arenite. They are composed predominantly of rounded cobbles of volcanic rock, subordinate granitoid rocks, and vein quartz, in a matrix of coarse-grained arenite.

Coarse- to medium-grained, moderately sorted and indurated lithic to arkosic arenites are typical of the lower part of the member; the abundance of feldspar and lithic fragments decreases upwards in the member. Mottled purple to mauve, pigmentation, typical of this member, grades into the mottled brick red and white coloration of member A3. Cross-stratification consists of weakly cyclic, stacked sets of large-scale (<1m thick) tabular to more commonly trough-shaped crossbeds (Plate B2D). In a few outcrops, climbing ripple stratification (type A and B of Jopling and Walker, 1969) was observed in finer grained beds, as was horizontal stratification. In general, mudstone occurs only as intraclasts at the base of larger crossbed sets and in conglomerates, but in one example, an in situ bed of dark red parallel-laminated mudstone was observed. This bed has a deeply scoured top surface that is littered with angular rip-ups; the bed pinches out over a lateral distance of less than 5m.

Inferred paleocurrents (Fig. 2-3, 2-5) generally are broadly unimodal towards the south-southeast. Local deviations from this trend, in which apparent paleoflow was to the southwest, occur in some outcrops adjacent to basement paleohighs.

The lower part of this member commonly contains concretions with maximum diameter from 3-15cm. The concretions are relatively resistant to weathering (Plate B2E), are tan with distinct white rims, and are composed of equant poikilotopic

subhedra of dolomite. An early post-depositional origin of these structures is indicated by the deformation of sand laminae around the concretions.

Interpretation

Member A2 is interpreted as valley-fill fluvial deposits similar to the Donjek and South Saskatchewan models of Miall (1977, 1978) and the braided-meandering fluvial model of Jackson (1978). Deposition was restricted to fluvial channels that occupied valleys within the topographically rugged landscape. Although paleoslope was to the southeast, as suggested by the dominant trends of paleocurrents, paleohighs caused local deflection of paleoflow. The lack of well-developed depositional cyclicity is attributed to the syndepositional removal of low-flow stage deposits and amalgamation of higher-energy deposits. In this process, mud that may have been a significant component of the fluvial system, was largely removed by erosion. An ideal depositional cycle consisted of streamflood deposition (cf. Bluck, 1966) of paraconglomerate (derived from volcano-plutonic basement rocks west of Coppermine River) or pebbly arenite by large-scale transverse bedforms followed by dune-sandwave deposition, climbing ripples and finally suspended-load deposition of mudstone. Upward-fining of this member into member A3 arenites indicates progressive source-area backwasting (retrogradation) and suggests that members A2 and A3 are probably proximal-distal facies equivalents.

Member A3

For the most part, member A3 is composed of indurated medium-grained subarkosic arenite that has a distinct mottled red and white pigmentation. Along the western edge of the main region of exposure, south of Bigbear Creek, conglomerate composes up to 50% of the member (Fig. 2-3, section B) (Plate B2D). In St. Germain graben (Fig. 2-1), the arenites are interlayered with calcareous paraconglomerate lithologically identical to member A1 conglomerates. The member is absent from the Mist Mountain outlier. The absolute thickness of this member is unknown; the thickest measured section (Fig. 2-3, section B) is 170m, but the lower contact is not exposed. In the Big Bend area, this member is in gradational contact with both underlying (member A2) and overlying (member A4) arenites. The contact with member A2 is arbitrarily defined on the basis of grain size (when medium- to fine-grained arenite comprises >50%), and pigmentation (purple to red member A2 changes to mottled red and white); the contact with member A4 is defined as the first appearance of thinly-bedded friable hematitic quartz arenite.

The arenaceous facies of this member is composed of moderate- to well-indurated subarkosic to feldspathic arenite that displays moderate to good sorting and contains abundant rounded framework grains. The feldspars are extremely fresh, compared to those in the Fault River and Lady Nye Formations. The arenites occur in 10-40cm thick stacked sets of tabular planar and trough cross-strata and minor horizontal

stratification (Plate B2F). The stacking of the sets does not appear to be organized with respect to variations in grain size, bed thickness or crossbed orientation. The "arms" of stacked trough crossbeds may create a "pseudoherringbone" appearance to the crossbeds. Crossbed set thickness and the abundance of feldspar decreases towards the top of the member. Bedding surface structures, such as linear asymmetric ripple marks and superposed ("ladderback") ripple marks, although not common, were observed in several exposures.

The conglomeratic facies, which is restricted to outcrops south of Bigbear Creek, occurs as solitary to compound beds of polymictic, basement-derived paraconglomerate interbedded with typical A3 arenites. Solitary, flat-based tabular sheets of paraconglomerate (<2m thick), composed of a framework of rounded cobble- to granule-size fragments of basement volcanic rocks and granule- to sand-size grains of polycrystalline quartz, altered feldspar, mudstone, and silcrete, outcrop as resistant cuestas. In thin section, these conglomerates have a well-developed microgranular cherty cement which accounts for their durability. In contrast, the thicker, compound paraconglomerates (Plate B3D), in sheets up to 40m thick weather recessively. These paraconglomerates are coarser grained than the solitary sheets (rounded cobbles up to 25cm), and commonly contain scoured lenses of stratified feldspathic arenite. Matrix-framework relationships are largely obscured by in situ disaggregation during compaction of weathered volcanic porphyries, although some beds have a sandy matrix that

laterally continuous (Fig. 2-3; sections A-D) and define a wedge-shaped sequence that thins to the north and south.

Inferred paleocurrents in the arenites are unimodal to the southeast throughout the basin (Fig. 2-3, 2-5), and they show no deflection around exhumed paleohighs.

Interpretation

The arenite facies is interpreted as a low- to moderate-energy fluvial system, although the facies does not bear resemblance to published facies models for sandy fluvial systems (e.g. Miall, 1978; Cant, 1978; Smith, 1970). Deposition probably took place during migration of sinuous to straight-crested large-scale ripples (cf. Harms, et al., 1982) within broad, shallow channels. The large scale of the channels may account for the difficulty in recognizing such structures (i.e. channels, if present, are probably recognizable only in the larger outcrops). Solitary conglomeratic beds represent discrete stream/sheetflood depositional events on an otherwise sand-dominated floodplain. The compound conglomerate units represent deposition on small(?) alluvial fans. Interfingering of fan and fluvial facies along the western margin of the basin implies that this area is close to the depositional fall-line (lapout).

The uniform southeast-directed paleocurrents indicate that most (or all) pre-existing paleotopography had been infilled, so that the paleocurrent pattern reflects only the regional paleoslope.

FIGURE 2-4: Schematic block diagrams that illustrate the three stages of deposition inferred for the Bigbear Formation.

STAGE 2

Hematitic arenites, which constitute more than 95% of Stage 2 lithologies, are inferred to be records of deposition within a land-locked erg or desert sand sea. Detailed sedimentologic analysis of the two members of Stage 2 (members A4 and A5) is presented elsewhere (Ross, 1983b); only a brief synopsis of the essential characteristics of these members as they pertain to the depositional history and evolution of the Bigbear depositional system will be presented here.

The maximum measured thickness of Stage 2 arenites is 460m (Fig. 2-3, section F); these members thin to less than 250m at Mist Mountain outlier, and also thin towards the west and north, although a lack of exposure precludes an accurate thickness determination. Stage 2 comprises two members: A4 consists of crossbedded, very friable, dark red quartz arenite; A5 is a much more indurated, less hematitic arenite that contains thin (<10cm thick) red mudstone and basement-derived volcanic paraconglomerate beds (<1m thick) along the western edge of the basin. Although not well exposed, the contact between fluvial A3 arenites and eolian A4 arenites appears to be gradational. The contact between Stage 2 and Stage 3 deposits is marked by the first appearance of monomictic intraformational boulder-cobble conglomerate, except in the Mountain outlier where fluvial pebbly arenites mark the onset of the third stage of deposition.

Major conclusions are:

- 1) Stage 2 deposition in the Bigbear system took place in

an erg (cf. Wilson, 1971; Cook and Warren, 1975) that covered an area of at least 4000 km². The western margin of the erg was close to Coppermine River, northward from Big Bend, where fluvial facies interfinger with eolianites.

- 2) Onlap of member A1 paraconglomerates and basement by member A4 eolianites in Mist Mountain outlier (Fig. 2-6) indicate that this area was a paleohigh during Stage 1 deposition. Stage 2 deposits are also thinner in Mist Mountain outlier than those exposed along the western side of Mouse graben.
- 3) Erg formation occurred as a consequence of the interaction of sand-laden prevailing winds and a topographic low, bounded, in part, on the west by the fall-line of the Teshierpi-Bigtrees peneplain and on the east by a paleohigh east of Mist Mountain outlier. These features trapped sand and initiated erg formation by topographic lee effects (on the west side) and by blocking sandflow out of the depression (on the east side).
- 4) Based on arguments presented elsewhere (Ross, 1983a) these eolianites formed in an arid climate. The available paleomagnetic data (Irving and McGlynn, 1979) indicate a southern latitude for the erg (25-40°S) similar in location to present-day southern latitude deserts such as the Kalahari and Simpson deserts.

FIGURE 2-5: Paleocurrent map for the main region of exposure of the Bigbear Formation. Arrows represent mean azimuths calculated from a minimum of 20 paleocurrent measurements per station.

STAGE 3

Stage 3 deposits comprise two members (A6 and A7) that are composed predominantly of mature quartz arenite, but which also contain significant amounts of intraformational conglomerate. This stage records the transition from predominant eolian deposition to fluvial deposition, and burial of the Bigbear Formation by the Lady Nye Formation.

Member A6 rests with apparent conformity on Stage 2 arenites, thus implying ongoing subsidence of the basin contemporaneous with cannibalism, and therefore uplift, in other parts of the basin. Member A7 grades, with gradual changes in both lithology and paleocurrent orientation, into pebbly arenites of basal Lady Nye Formation. A lack of marker beds and poor exposure of Stage 3 members makes accurate thickness determinations impossible.

Member A6

This member is composed of medium-grained hematitic quartz arenite (70%) and interbeds of intraformational conglomerate (30%). The base of the member is marked by the first appearance of conglomerate which rests sharply on member A5 arenites. The upper contact with member A7 is entirely gradational over 30m. The upper contact is arbitrarily chosen as the first occurrence of beds of coarse-grained, white weathering quartz arenite with vein quartz pebbles. Conservative estimates of thickness, based on map pattern and measured sections, suggest a thickness for this member of up to 735m (Fig. 2-3, section E) to almost 1000m (section at Mist Mountain outlier, Fig. 2-6). In Bigbear Creek, the correlative

lithologies comprise only 40m of section, although the top of the section is not exposed.

The arenites occur in stacked tabular sets (20-100cm thick) dominated by wavy horizontal stratification with uncommon tabular planar cross-stratification (Plate B3C). Bedding surfaces within horizontally stratified sets invariably display linear asymmetric ripple marks. Grain size and lithologic character of the arenites is remarkably consistent throughout the member, although desiccated red mudstones occur in the section in Bigbear Creek. Arenites in the upper part of the member contain a greater abundance of small-scale (10-30cm thick) cross-stratification than in the lower parts of the member.

The intraformational conglomerates occur in sheets up to 30m thick. The conglomerate beds are thickest low in the member and decrease up section in thickness, average and maximum grain size, and relative abundance. The conglomerates are composed almost exclusively of subequant, subround to angular cobble- to boulder-size clasts of mature red quartz arenite, lithologically identical to member A5 arenites. Rare pebbles of volcanic basement, vein quartz, and member A2 arenites have been observed. The cobbles form a tightly packed, commonly non-stratified framework supported by a sand- to mud-rich matrix (Plate B3A). In Bigbear Creek, the conglomerates rest directly on thin-bedded mudstone-sandstone couplets that display impressive soft-sediment deformation, presumably the result of loading during conglomerate deposition.

The top surfaces of such conglomerate sheets contain coarse boulders that protrude through a thin veneer of ripple-marked red muddy siltstone. Also in the Bigbear Creek section, single-cobble layers of intraformational conglomerate are interlayered with rippled and desiccated red mudstones and siltstones (Plate B3B). More commonly, the conglomerates in member A6 occur as discrete sheets within the arenites, and have sharp lower and upper contacts. The thickest conglomerate bed observed (30m thick) contains slabs of red arenite up to 2m thick and 5m long. In one place, the conglomerate rests on a scoured surface, with up to 5m relief, and is composed of a framework of well-rounded boulders that display imbrication at the top of the bed and cross-stratification in the channel thalweg (Plate B3C).

Most of the arenite clasts that comprise the framework of the conglomerate beds are lithologically identical, with respect to composition of sand grains and cement, to the matrix arenite in the conglomerates. The most common type of cement is composed of polygonally interlocking syntaxial quartz overgrowths. In several thin sections, however, intraformational arenite clasts displayed a laminar botryoidal cement composed of nearly isotropic microcrystalline silica (opal?) in addition to clasts that display a fibrous chalcedonic cement.

Although structures suitable for paleocurrent analysis are uncommon, a clear pattern is discernable. Paleoflow inferred from crossbeds within arenites of this member display a strong

southeast-directed unimodal pattern. Those conglomerates that display directional fabric indicate transport to the southwest (Fig. 2-3, 2-5).

A notable exception to this pattern of paleotransport and lithologic character occurs in arenites that overlie eolian facies in Mist Mountain outlier (Fig. 2-6). In this area, no conglomeratic facies are present, and no A7-type vein quartz pebble-bearing arenites occur. Instead, Stage 2 eolianites are overlain by up to 80m of low-angle cross-stratified to plane-bedded arenite that is overlain by 290m of coarse-grained, locally pebbly quartz arenite. The latter sequence of arenites displays spectacular large-scale trough crossbeds up to 10m thick (Plate B3D). Paleocurrents inferred from these crossbeds are unimodal to the southeast, parallel to the strike of the Sinister Fault (Fig. 2-6). The trough crossbedded arenite is gradationally overlain by almost 700m of thin-bedded (5-20cm) quartz arenite and subordinate mud-cracked red mudstone, the latter displaying trough to tabular cross-stratification, horizontal stratification, and in-phase (type B) climbing ripples.

Interpretation

The conglomerate beds in member A6 were deposited by debris flows (e.g. Plates B3A, B), streamflood, and channelized streamflow (Plate B3C). The interbedded arenites were deposited by fluvial processes on a shallow floodplain; small-scale crossbeds represent deposition by migrating transverse bed-forms, whereas the more common horizontal stratification

probably formed during lower flow regime deposition (cf. Smith, 1971; McBride, et al., 1975). The larger, low-angle crossbed sets may represent deposits of solitary eolian dunes but this is uncertain. The thick sequence at Mist Mountain outlier represents deposition exclusively by fluvial processes, although the genesis of the 80m thick section of plane-bedded arenite (transitional from Stage 2 to Stage 3) is unknown. The large-scale trough crossbed sets were deposited by large-scale sinuous-crested transverse (three-dimensional) bedforms. Thinbedded arenite, which composes the bulk of the Mist Mountain section, was deposited on a shallow, low-energy floodplain.

The facies patterns, paleocurrents, and origin of the intraformational conglomerate can be explained within the context of fault-influenced sedimentation. The cobbles, boulders, and slabs of red quartz arenite most probably were derived from earlier deposits of Bigbear Formation because no other lithology of similar character is known from Wopmay Orogen ((red quartz arenites of the Burnside River Formation, Campbell and Cecilie (1981) are lithologically similar, but crop out more than 400km away (east) of the Bigbear region)). The southwest-directed paleocurrents associated with appearance of these conglomerates suggests derivation from the northeast, a significant deviation from the regional southeast-dipping paleoslope inferred for older members. Normal fault movement is the simplest way to account for the paleoslope change, the random occurrence of conglomerate beds, and

composition of the conglomerates. Facies distribution and paleocurrents suggest southwest-side down movement along the northwest segment of Sinister Fault (Fig. 2-7). The intraformational conglomerate was derived from erosion of the uplifted block of the fault as coarse clastic detritus was shed towards the southwest, away from the fault, by massflow and streamflow processes. In contrast, the fault-parallel paleocurrents and lack of intraformational conglomerate in Stage 3 arenites at Mist Mountain outlier indicate that movement along the Sinister Fault in this region may not have created sharp syndepositional relief. The inferred along-strike difference in geomorphic expression of the Sinister Fault is interpreted as reflecting a transition from "stick-slip" movement, near the head of Mouse graben, to gradual creep near Mist Mountain outlier. The along-strike change in the style of movement of the Sinister Fault implies that the fault was "hinged" between the head of Mouse graben and Mist Mountain outlier. The occurrence of large slabs of quartz arenite in conglomerates along the west side of Mouse graben suggest that there may have been a syndepositionally active fault in this region in addition to the Sinister Fault. The intensity of syndepositional fault movement decreased during deposition of member A6, as suggested by decreased conglomerate clast size and, in Mist Mountain outlier, decreased bed thickness up section.

Member A7

This member is sporadically exposed between Bigbear Creek and the north end of Mouse graben; it also extends to the

Figure 2-6. Geologic map and measured stratigraphic section of Bigbear Formation preserved in Mist Mountain outlier. Paleocurrent rose diagrams are for all measurements taken for unit to the right of the rose. See Figure 2-1 for location of outlier.

FIGURE 2-7: Schematic block diagram to illustrate inferred paleogeography along Sinister Fault during deposition of member A6. Large arrows at the end of the block show relative sense of crustal tilt (rotation) along the fault. Formation of syndepositional graben structure east of Mist Mountain outlier is inferred because of the need to constrain channel depth necessary for formation of large-scale bedforms. Arrows on depositional surface are measured (solid) and inferred (dashed) paleocurrents.

north-northeast and is the part of the Hornby Bay Group that is intruded by the Muskox Intrusion (Kerans, 1983). At the north end of Mouse graben, this member records the complete transition from Bigbear to Lady Nye Formation sedimentation. The descriptions below are based largely on this section (Fig. 2-3, section E).

Member A7 is 176m thick at the northwest end of Mouse graben; thicknesses could not be measured in other areas because of the low relief of the outcrops, and truncation by the Dismal Lakes Group. In section E (Fig. 2-3), the member is composed of variable proportions of four lithologies: thin-bedded, medium- to fine-grained, indurated pink quartz arenite, red mudstone, coarse-grained pebbly quartz arenite, and minor intraformational conglomerate. The pebbly quartz arenite constitutes most of member A7. Fine-grained arenites predominate in the lower part of the section, occurring in stacked undulose tabular beds (<15cm thick) that commonly contain indistinct parallel lamination. Red mudstones are present only in the middle part of the section where they occur in 2-10cm thick beds with sand stringers and desiccation cracks. The coarse-grained arenite is characterized in this section by a white to tan colouration (due to kaolinite, rather than hematite-illite, cement, Plate B3E) and by the ubiquitous occurrence of scattered pebbles of white vein quartz. In areas to the west and northeast of Mouse graben, where a strong hematitic pigmentation is developed in this arenite, the vein quartz pebbles serve to distinguish this

arenite from others in the region (cf. Kerans, 1983). This arenite becomes prominent in the middle of the section, and the upper part of the member is composed exclusively of pebbly arenite. The contact with the Lady Nye Formation is marked by the first appearance of sandy conglomerate which contains cobbles of Tree River Formation (siliceous oolitic ironstone, Hoffman, 1981) or Mara Formation (Goulbourn Group, Campbell and Cecille, 1981), a distinctive clast lithology that is characteristic of the basal part of the Lady Nye Formation (Bunn Creek conglomerate, member C2a).

Paleocurrents in all outcrops are unimodal. Low in the section they are directed towards the south, shift to the southwest in the middle part of the member, and are directed southwest to northwest in the upper part of the member (Fig. 2-3, 2-5).

Interpretation

This member was deposited by fluvial processes, although it is difficult to find a modern analogue or suitable facies model. The tabular beds of pink arenite were deposited by low-energy sheet-like flow that was succeeded by channelized flow that gave rise to distinct channel and "overbank" facies (lenticular arenites and red mudstones, respectively). The increase in bed thickness, grain size, and abundance of large-scale crossbeds in the upper part of the member suggest that comparatively higher energy (perennial?) fluvial conditions prevailed.

This member records deposition during paleoslope change

from predominantly southeast- to west-dipping near the top of the member. The change in paleoslope was accompanied by an influx of extrabasinal detritus, in contrast to the intrabasinal origin for the cobbles in member A6. The change in depositional style within this member may reflect a climatic change from relatively arid to more humid. Although this conclusion is somewhat speculative, it is supported by the occurrence in this member of siliceous, ultradurable pebbles, a characteristic of humid environment gravels (West, 1978).

SUMMARY

Of critical importance to understanding tectonic controls on Bigbear Formation deposition and regional geologic history, is an evaluation of the relationship, if any, between compressive tectonics within Wopmay Orogen and patterns of sedimentation in the Bigbear depositional system.

The final phase of activity in Wopmay Orogen was the "remote terminal collision" during which plate collision west of Great Bear Lake led to conjugate transcurrent faulting and "pop-up" thrust tectonics (Hoffman, 1980; Hoffman and St. Onge, 1982). Conjugate northeast-northwest faults (Figs. 2-1, 2-3) were initially characterized by strike-slip movement that changed to reverse movement during progressive deformation and fault plane rotation. Some major faults within Wopmay Orogen outline large crustal "shingles" (Hoffman and St. Onge, 1982) that contain metamorphic mineral assemblages of contrasting pressure regimes. The amount of movement along these faults

can be constrained through the combined use of bathograds in metamorphic basement rocks (Carmichael, 1978; St. Onge and Carmichael, 1979) and the thickness of overlying Hornby Bay Group strata. In the case of Mist Mountain outlier, pre-Hornby Bay Group uplift along the Sinister Fault, after correction for minimum post-Hornby Bay Group dip-slip (1.3km, Fig. 2-6), is in the range of 2-3km (Hoffman and St. Onge, 1982). Within this tectonic scenario, three conditions could have influenced Bigbear Formation sedimentation: 1) formation of topographic highs and lows along compressive and releasing double bends in transcurrent faults (Crowell, 1974; Mitchell, 1981), 2) formation of a foredeep in front of active thrust blocks (Eisbacher, et al., 1974), and 3) relict topography created by erosion of either 1) or 2).

Several points argue against any relationship between compressional tectonics in Wopmay Orogen and sedimentation within the Bigbear depositional system:

- 1) Paleocurrents indicate a basinwide southeast-dipping paleoslope that transgresses shingle boundaries.
 - 2) Stage 3 intraformational conglomerates are best explained as being the result of normal rather than reverse movement along the Sinister Fault. Reverse movement along this fault, which forms the eastern boundary of a shingle, is implied by offset of bathograds across the fault; reverse movement must predate the onset of Hornby Bay Group deposition.
- Activation of the Sinister Fault during deposition of

member A6 signals the onset of a period of inferred regional extension.

- 3) The lack of cyclicity in Bigbear Formation fluvial deposits and the high compositional maturity of the arenites is more consistent with tectonic quiescence. Depending on source rock composition, basins formed in tectonically active environments are typically filled with thick distinctive cyclic sequences of immature clastics (cf. Steel and Gloppen, 1981; Reading, 1980; Heward, 1978).

The Bigbear Formation, and Hornby Bay Group in general, is therefore inferred to have been deposited after the final period of compression that affected Wopmay Orogen, as initially suggested by Hoffman (1980).

Stage 1 deposits record infill of rugged paleotopography by successively finer grained, and therefore lower energy, fluvial systems. Formation of a deep weathering profile in the basement prior to the onset of fluvial deposition was impeded by an arid paleoclimate and development of impermeable silcretes and duricrusts; these are among the oldest silcretes yet recognized (Chiarenzelli, 1983). The silcretes were formed by early chalcedony cementation of rock fans (colluvial cones) that developed adjacent to paleohighs. Donjek-type fluvial facies (member A2) drained from the northwest and flowed through paleovalleys around remnants of degraded basement spurs in the piedmont front (cf. Twidale, 1981). Back-wasting of the source area led to a decrease in grain size during

formation of a peneplaned source area (Teshierpi-Bigtree peneplain, discussed in chapter 4). Infill of paleotopography before the onset of member A3 deposition resulted in formation of an extensive braidplain ($>4000\text{km}^2$). A basin margin is inferred for the present northwestern limit of outcrop along Coppermine River, where member A2 pinches out and coarse fanglomerates occur in member A3. The consistent occurrence of basement-derived conglomerates within otherwise arenaceous rocks along the western edge of outcrop implies that this margin of the basin was long-lived and may be related to faulting, possibly along a fault parallel to the Coppermine River (the northward extension of faults that bound St. Germain graben is a possibility). The lack of Stage 1 deposits in Mist Mountain outlier (except for thin A1 paraconglomerates) indicates that this area was a paleohigh during Stage 1 deposition and suggests that a roughly north-trending basin existed during this stage of sedimentation.

Stage 2 eolianites were deposited in an arid-climate erg (sand sea) that developed within the Bigbear depositional system. Throughgoing regional sandflow from the west was blocked by the topographically negative Bigbear basin. After initiation of dune bedforms and sand build-up, the rough surface of the dune field acted as an effective sand trap that maintained erg accretion. Ephemeral fluvial systems prevailed along the western (upwind) margin of the erg. The change in depositional style from Stage 2 eolianites to fluvial facies of Stage 3 probably reflects a change in climate from arid to

humid. Normal fault movement along the reactivated Sinister Fault caused cannibalization of lithified Stage 2 arenites, and deposition of stream and mass flow intraformational conglomerates of Stage 3. A change in the style of movement along the fault, from stick-slip to creep, from the head of Mouse graben southeast to Mist Mountain outlier, respectively, is invoked to explain paleocurrents that are orthogonal to parallel to the strike of the fault, and to explain the inferred difference in geomorphic expression of the fault during sedimentation. The normal faulting is inferred to be an expression of regional westward tilting, which culminated in the burial of Bigbear fluvial and eolian deposits by the west-flowing Lady Nye fluvial system.

PLATE B1

B1A: Exhumed paleotopography north of Big Bend in Coppermine River. View is towards the southwest. Cores of knolls are composed of Dumas Group dacite tuff (foreground) and aphyric flow-banded rhyolite (background) and are largely mantled by colluvial and alluvial fanglomerate. Contact between dacite and mantling sediments is marked by dashed white line. Height of hill in foreground is approximately 75m.

B1B: Glaciated outcrop of massive siliceous paraconglomerate composed of angular clasts of aphyric rhyolite lava (Dumas Group) in a matrix of chalcedony-cemented arkose (silcrete). Field of view is about 40cm wide.

B1C: Hand sample of arenaceous matrix in massive siliceous paraconglomerate. This sample is composed largely of silica-cemented sand-size framework (1) with sparse intraclasts of siliceous mudstone (M), irregular laminoid fenestral voids filled with massive microcrystalline silica (2), and irregular veins and void-fills (3) of red to clear chalcedony and mega-quartz. Stratigraphic top is towards the right side of photo. 18mm-wide coin for scale.

B1D: Hand sample of stratified subfacies composed of inter-layered silica-cemented arenite (A) and peloidal-intraclastic granulestone (G). Note clotted fenestrae in arenite layers, filled with massive microcrystalline silica, and laminoid fenestral cavity (LF) at base of peloidal layer that is filled with red and clear chalcedony. Stratigraphic tops to the right. 18mm coin for scale.

B1E: Thin section photomicrograph of chalcedony-cemented arkosic arenite. Disrupted framework is composed of quartz (Q), feldspar (F), and micas (M). Note "concertina structure" in micas. Cement is composed of fibrous megaquartz and chalcedony. Scale bar is .50mm long. Crossed nicols.

B1F: Thin section photomicrograph of fenestral-clotted texture defined by irregular voids filled with microcrystalline silica (dark gray) within a framework composed of chalcedony-cemented arenite. Scale bar is 2.5mm long. Crossed nicols.

PLATE B1

A

B

C

D

E

F

PLATE B2

B2A: Thin section photomicrograph of vein fill within arenite matrix of conglomeratic silcrete that represents the final phase of silcrete formation. The vein-fill is composed of coalesced spherules of length fast chalcedony (C) overlain by prismatic megaquartz (M) that appears to have actively pushed aside (displacive crystallization) arenite matrix (A). Shape of original megaquartz grains, in part masked by later recrystallization, is preserved by concentric bands of inclusions. Scale bar is 0.5mm long. Crossed nicols.

B2B: Hand sample of calcified botryoidal silcrete. Chalcedony-cemented arenite (A) occurs at base of slab. Sandy layer at center of slab (S) appears to be a synsedimentary drape but serial section reveals that it infills a subhorizontal fracture. The sample is composed almost completely of sparry dolomite that has replaced earlier botryoidal chalcedony. Stratigraphic tops towards top of photograph. 18mm-wide coin for scale.

B2C: Stacked tabular-planar crossbeds composed of coarse-grained sublitharenite of member A2.

B2D: Stream cut through sequence of member A3 paraconglomerates along the western edge of the outcrop area (section B, Figure 2-3). Lenticular layers are scoured remnants of interconglomerate arenites. Height of cliff face is 10m.

B2E: Dolomite-cemented arenite concretions in member A2 sublitharenite. Note deformation of laminae around concretions. 18cm pencil for scale.

B2F: Stacked tabular-planar to gentle trough crossbeds in medium-grained quartz arenite of member A3. Two-dimensional section through divergent crossbed sets produces apparent herringbone pattern. Brunton compass in upper center for scale.

PLATE B2

A

B

C

D

E

F

PLATE B3

B3A: Debris flow paraconglomerate from member A6, Bigbear Creek. Subangular blocks of member A5 arenite float in mud-rich matrix that contains angular sand and pebble-size arenite clasts. Note the unblemished ripple surface of the arenite boulder that attests to the lack of interparticle collision and high matrix strength. 9cm knife in left center for scale.

B3B: Single-cobble-thick debris flow beds within parallel-laminated siltstones and mudcracked mudstone, Bigbear Creek. 9cm penknife (left center) rests on mudstone drape on debris flow bed.

B3C: Conglomeratic infill of channel cut into member A6 (A5?) arenites. Conglomerate is framework-supported and composed of rounded boulders of quartz arenite. Stratification (left side of photo) and boulder imbrication indicate that the conglomerate was deposited by southwest-flowing currents. Underlying arenites are composed of parallel-laminated beds and a single tabular-planar crossbed (X); note that the lamination in the crossbed dips in an opposite direction to that in the overlying conglomerate. Packsack at left (circled) for scale.

B3D: Megatrough crossbeds in pebbly quartz arenite of inferred member A6 correlative in Mist Mountain outlier. The field of view is dominated by a single large crossbed that is about 10m thick (stratigraphic tops to right) with top of the underlying crossbed exposed in the lower left corner.

B3E: Thin section photomicrograph of white-weathering vein-quartz pebble arenite of member A7. Framework of subrounded quartz grains displays well-developed quartz overgrowths (O) that form a subpolygonal interlocking mosaic. Later kaolinite cement (K) formed during dissolution of quartz overgrowths and framework. Scale bar is .5mm long. Crossed nicols.

PLATE B3

A

B

C

D

E

CHAPTER 3

FAULT RIVER FORMATION

The Fault River Formation is exposed in a series of northwest-dipping, strike-parallel cuestas that extend from Hornby Bay northeast along the Fault River valley for a distance of about 40km (Figure 3-1A and Plate 1A). At the northeast end of the river valley, the Fault River Formation occurs in narrow fault-bounded wedges. Scattered exposures also occur on islands along the northeast shore of Great Bear Lake and in a fault-bounded outlier at Elizabeth Lake (Fig. 3-1B). The formation is named for the exposures in the valley of the Fault River; a type section is here designated at 66°43'N and 117°23'W (Fig. 3-1C).

The juxtaposition of the Fault River Formation and Great Bear Batholith along the Fault River Fault indicates that at least 300m of dip-slip movement (northwest-side down) has occurred along this fault. Evidence is presented that suggests that some of the movement along the fault was syndepositional. North of the fault, the Fault River Formation rests unconformably on weathered volcano-plutonic rocks of Great Bear Batholith. Although post-depositional fault movement and Quaternary cover largely mask the unconformity, the contact can be observed in a thin slice of basement on the northwest side of the Fault River Fault. The unconformity also is exposed at the northeastern end of the outcrop belt, where members of the Fault River Formation, in ascending stratigraphic

FIGURE 3-2: Generalized stratigraphic column showing inferred depositional environments for the members of the Fault River Formation. LNF = Lady Nye Formation, the conformably overlying fluvial sequence.

order, onlap basement rocks. In addition, each member shows increased grain size and decreased thickness in this part of the outcrop belt. The Fault River Formation is conformably overlain by the basal conglomerate of the Lady Nye Formation (Bunn Creek conglomerate). Towards the northeast end of the Fault River Valley the Lady Nye Formation oversteps Fault River Formation and rests directly on basement. This relationship, in combination with regional paleocurrent trends and variations in grain size, suggests that the present Fault River Formation is only the northern remnant of the (at one time) more extensive Fault River fluvial system. The Fault River Formation is 370m thick in the type section and is subdivided into four informal members: the lower two members (B1 and B2) comprise a nearly continuous upward-coarsening succession of predominantly conglomeratic rocks that are sharply overlain by the basal arenites of a second upward-coarsening sequence, represented by members B3 and B4.

Member B1

This member (the Prospect Creek facies of Kerans, et al., 1981), which varies erratically in thickness from 0 to 90m, rests on weathered volcanic and granitic rocks of Great Bear Batholith. The volcanic rocks are mostly felsic ignimbrites and porphyries of the Sloan Group (Hoffman, 1978) that were weathered to an indurated, hematite- and silica-rich rock, with a hackly fracture and local bleached spots. The granitic rocks were weathered to a crumbly grus rich in hematite. The

FIGURE 3-3: Measured section through the thickest continuous exposure of member B1. UFR = upper flow regime; LFR = lower flow regime. The terms "distal", "medial", and "proximal" reflect the relative energies of the depositional environments rather than established proximity to source areas. Arrows to the right of the column are paleocurrent azimuths; the other symbols represent desiccation cracks.

thickness of the weathered zone could not be determined but the amount of weathering of Great Bear Batholith rocks appears to be most intense in the Fault River region, decreasing in intensity towards the east, south, and north (per. obs., and Hoffman, 1978).

In the thicker sections, member B1 coarsens upwards from mudstone-siltstone-arenite at the base into boulder paraconglomerate near the top where the member grades into the overlying conglomerate-arenite couplets of member B2. The lower part of this member is commonly absent from exposures of the unconformity, and the paraconglomerate of the upper part of the member rests directly on basement. In places where member B1 is entirely absent, member B2 paraconglomerates rest on basement. The apparent restriction of finer grained member B1 lithologies to the thicker sections (valley fill) and onlap of conglomeratic rocks on topographically higher areas suggests local paleotopographic relief of up to 50m.

The lower part of member B1 (Fig. 3-3) is composed of poorly organized interbeds of red mudstone and siltstone (Fl), arenite (Sp, Sh, St) and paraconglomerate (Gms) that occurs in tabular sheets and scour-fill lenses. Massive immature paraconglomerate composed of angular, locally-derived clasts in a mud-rich matrix, occur as flat- to scour-based sheets (20-50cm). Some of the sheets have a distinct basal zone of muddy siltstone that coarsens up abruptly (inverse grading). Tops of paraconglomerate beds commonly show a somewhat sorted (winnowed?) gravel layer that is draped by a thin layer (<5cm).

of parallel-laminated red mudstone and siltstone. Lithic to arkosic, hematite-rich arenite occurs as parallel-laminated (Sh) to cross-laminated (Sp, St) beds and lenses (<30cm) interlayered with desiccated red mudstones and siltstones (Fl).

Successively higher units within the member display an increased grain size, in addition to increased proportion of conglomerate, bed thickness, and vertical order to the beds. The middle part of the member is composed of stacked beds of mud-rich massive paraconglomerate (Gms) with rare interbeds of mud-cracked argillaceous siltstone (Fl) and cross-stratified litharenite (Sp). The upper part of the member is composed of coarse boulder paraconglomerates (Gms), with a sandy matrix, rare orthoconglomerate (Gp) and lenses of flat- to cross-laminated hematitic litharenite. The paraconglomerates in both the middle and upper parts of the member are composed largely of rounded clasts of weathered felsic volcanics and subordinate granitoid clasts; vein quartz and red quartz arenite cobbles are minor but conspicuous components of the uppermost conglomerates. The roundness, size of the cobbles, and percentage of sand in the matrix increases upward. The most altered cobbles have been deformed during compaction, forming irregular masses of hematite-rich "pseudomatrix" which obscures recognition of depositional matrix. Interbedded litharenites occur as lenses that may have either a scoured-base/planar-top or a planar-base/scoured-top. Nodules of poikilotopic dolomite cement occur locally within the arenites and

conglomerates: these nodules are randomly distributed, and because they lack geopetal fabrics and orientation, they probably are not pedogenic structures.

Paleocurrent indicators are sparse within this member. Combined measurements of cross-lamination and rare cobble imbrication indicate west-northwest transport, perpendicular to the strike of the Fault River Fault and related parallel faults (Fig. 3-4A).

Interpretation

Paraconglomerates within this member are inferred to have been deposited during floods by sediment gravity flow processes: debris flows (mud-rich matrix with inversely graded base) and stream/sheetfloods (sand-rich matrix). The waning stages of flow would have winnowed the tops of conglomerate sheets, and deposited lower energy parallel- to cross-laminated arenites and suspension-load mudstones and siltstones. The entire member is interpreted as an alluvial fan deposit because, even though paleocurrent and grain size data are not available to document a radial dispersal pattern (cf. Williams, 1969), the general facies succession resembles that of alluvial fans (cf. Bull, 1977; Nilsen, 1969, 1982). The lower fine-grained part of the member probably formed on the fan fringes. Such areas are typically the sites of ephemeral channel incision below the intersection point (Wasson, 1974) and infill of such channels results in lenticular bed geometry and poorly developed vertical cyclicity. The occurrence of debris flow

deposits on fan-fringe areas might imply a relatively small radius fan (cf. Gloppen and Steel, 1981). The upsection increase in grain size and bed thickness represents fan progradation. As the fan(s) prograded, fan-fringe facies would infill paleovalleys; further infill and basement onlap occurred during deposition of medial fan debris flow and streamflood paraconglomerates. The local provenance of clasts in the lower conglomerates, and paleocurrents normal to the Fault River Fault, strengthen the suggestion that these conglomerates formed on small alluvial fans along a fault scarp (Fig. 3-8). The upper conglomerates grade into compositionally similar and regionally more extensive conglomerates of overlying member B2; thus the upward coarsening cycle within this member reflects the onset of conglomerate deposition within the basin. The appearance, in particular, of red quartz arenite boulders, probably derived from Bigbear Formation, indicates the influx of detritus from an expanded drainage basin area, consistent with a model of fan progradation (cf. Denny, 1966). The high sphericity and rounding typical of the upper conglomerates, although not characteristic of alluvial fan conglomerates, probably formed by several processes: reworking of fan-head conglomerate during progradation, spheroidal weathering (cf. Williams, 1969), and possibly increased abrasion because of vigorous bedload transportation, as suggested by McGowen and Groat (1970) for first-cycle rounded cobbles in the Cambrian(?) Van Horn Sandstone.

FIGURE 3-4: Paleocurrent maps for members of the Fault River Formation. Arrows are vector mean azimuths at a station; the number adjacent to the arrow indicates the number of measurements. Rose diagrams are plotted as percent measurements in each 30° sector; the number of measurements is shown at the center of each rose. A: Members B1 (short arrows) and B2 (long arrows). B: Rose diagrams for stations in member B3. C: Member B4 (long arrows) and member C2 (of Lady Nye Formation, short arrows).

Member B2

This member (the Volcanic Paraconglomerate facies of Kerans, et al., 1981), which attains a maximum thickness of 30m (Fig. 3-5) has a gradational contact with the underlying conglomerates of member B1. The member is in sharp (paraconformable?) contact with the overlying arenites of member B3, which drape the top surface of the uppermost conglomerate bed. The member is composed of tabular sheets of paraconglomerate, and subordinate orthoconglomerate, with wedge- to tabular-shaped beds of cross-stratified arenite (conglomerate-arenite couplets).

Paraconglomerates (Gms, Plate FlB), as massive sheets (1-4m) with a tightened framework in the upper part of many sheets, are most common along with rare, massive to cross-stratified orthoconglomerates (Gm, Gp, respectively), and thin (<20cm) orthoconglomerate lags within the arenite beds (Fig. 3-5). The conglomerates are composed of well-rounded cobbles and boulders, up to 25cm in diameter, dispersed in a matrix of coarse-grained, quartz-rich litharenite. At the base of the member, most of the cobbles are relatively fresh felsic volcanic rocks and porphyries with subordinate vein quartz and red quartz arenite. Towards the top of the member, the conglomerates are composed largely (>2:1) of cobbles and boulders of red quartz arenite, lithologically identical to arenites of the Bigbear Formation (members A5-A6). The composition of the cobbles and the non-hematitic, sand-rich matrix serve to distinguish these conglomerates from those of member B1.

Streamflood Fan Facies

Gms - streamflood paraconglomerate

Gm - gravel bar orthoconglomerate

Gp - gravel bar foreset stratification

Sp - sandwave; bar edge sandwedge

St - dunes

FIGURE 3-5: Partial measured section of member B2, illustrating typical character of the member. Arrows are paleocurrent azimuths.

The arenites within this member occur as flat-based lenses with scoured tops and as undulose sheets. Both bedding styles have non-gradational to rarely gradational contacts with the conglomerate interbeds. The arenites are coarse grained, locally pebbly, distinctly non-hematitic, quartz-rich litharenite to quartz arenite, identical to the sand matrix in the conglomerates. The feldspar content is invariably less than 5% and the dominant lithic components are grains of cherty (devitrified) felsic volcanic rock. Low-angle ($<20^{\circ}$) planar cross-stratification is typical, although rare trough crossbeds also occur (Fig. 3-5). Pebbly lags occur at the base of some sets of arenite. Paleocurrent measurements (76, Fig. 3-4) were taken in the arenites because of the general lack of fabric in the conglomerates; they indicate a strong unimodal west-northwest transport direction.

Interpretation

This member bears resemblance to upper fan deposits of the Scott outwash fan (Boothroyd and Ashley, 1975) and the "proximal alluvium" facies of Steel and Aasheim (1978) and is therefore interpreted to have been deposited by a high-energy braided river or streamflood fan(?). This member represents the culmination of the coarsening-up trend observed in the underlying member, and continued fan progradation. The abrupt appearance of member B3 arenites at the top of member B2 marks the beginning of a second upward-coarsening cycle and implies relatively sudden fan abandonment in this part of the Fault River basin.

The conglomerate-sandstone couplets are interpreted to represent flood-cycle deposits. The high proportion of arenite in the matrix of the paraconglomerates suggests a streamflood/sheetflood mode of origin rather than deposition by true debris flows, although some authors contend that debris flows may have a sand-rich matrix (Gloppen and Steel, 1981). Most of the conglomerate was deposited suddenly during flood events with minor amounts deposited by topset and foreset accretion on gravel bars (Gm and Gp, respectively) during relatively steady flow conditions. Waning flow conditions caused local winnowing of the tops of the flood paraconglomerates, and migration of sandwaves (Sp) and dunes (St). Sand also was deposited in depressions on bar surfaces, and as bar-edge sand wedges (cf. Williams and Rust, 1969; Rust, 1979).

The composition of the cobbles (Sloan Group felsites and Bigbear Formation arenites) and paleocurrents indicate a source area to the east-southeast (Fig. 3-8). The high sphericity and rounding of the cobbles may be the result of abrasion by mutual collision of ultradurable lithologies (rhyolite and arenite) rather than long distance of transport (cf. Abbot and Peterson, 1978), which is consistent with the lack of less durable cobble lithologies. The high concentration of quartz and durable lithic fragments in the arenites may imply either a second-cycle origin (e.g. Suttner, et al., 1981) or possible removal of labile fragments as a result of the milling action of the ultradurable cobbles.

2

This member is inferred to represent flood-cycle deposits of proximal braided rivers, similar in fluvial style to modern humid fans, that drained an enlarged source area to the east-southeast. The appearance of boulders and cobbles of Bigbear Formation arenites indicates uplift and incision of earlier Bigbear deposits. Paleocurrents directed away from Fault River Fault suggest that this fault (or parallel structure) acted as a basinal fall line. The overall upward-coarsening represented by members B1 and B2 could represent either a period of autocyclic progradation (e.g. fan-head incision and trenching) or a period of allocyclic progradation in response to tectonic lowering of the basin floor. The appearance of abundant Bigbear arenite cobbles in this member, which may coincide roughly with syndepositional fault activity during Bigbear Formation deposition, tends to favour a tectonic origin for the upward-coarsening cycle.

Member B3

This member (the Elizabeth Lake facies of Kerans, et al., 1981), the base of the second upward-coarsening cycle in the Fault River Formation, has a maximum measured thickness of 30m, and pinches out towards the north-northeast (Fig. 3-1). Where exposed, the member rests sharply on a conglomerate lag at the top of member B2; it grades into the overlying pebbly arenites of member B4 (Fig. 3-6). The lack of interfingering and onlap of B3 arenite on B2 conglomerate implies that these members are not facies equivalents. Member B3 is composed

FIGURE 3-6: Section measured through the transition from member B3 to B4, Elizabeth Lake outlier. fs = fine sandstone; cs = coarse sandstone; cgl = conglomerate. Brackets to the right of the column indicate stratigraphic interval over which paleocurrent measurements were made.

predominantly of thin-bedded, medium- to fine-grained, hematitic, locally feldspathic quartz arenite (Plate FlF). The transition into member B4 occurs over an interval of about 10m, and is marked by the appearance of sheets of massive volcanic conglomerate, a loss of hematitic pigmentation, coarsening of grain size, increased bed thickness, and increased regional paleocurrent consistency (Fig. 3-6, 3-4).

The lower part of member B3 is characterized by thin (20cm) apparently non-cyclic beds of hematite-rich, friable quartz arenite, with a flaggy to platy weathering aspect, which display a variety of stratification styles and abundant bedding plane structures. This part of the member thins to the north-northeast and is replaced by beds typical of the upper part of the member (ie. facies equivalents). Horizontal stratification, in sets 10-20cm thick, is most common (Sh facies in Figs. 3-6, 3-7; Plate FlC). It is interlayered with sets of cross-strata and composed of regular to discontinuous laminae of medium- to fine-grained arenite. In some places, 2-3 grain thick laminae of lithic (volcanic) granules and/or red mud-chips occur. Tabular to rarely wedge-shaped sets of low-angle ($<10^\circ$) planar cross-strata are common, and low-angle sets commonly grade downstream into horizontal stratification (Fig. 3-7). Two scales of high-angle cross-strata are interbedded with the above facies. Large-scale (20-30cm) tabular-planar crossbeds are common, and thicker sets have reactivation surfaces; large-scale trough crossbeds also occur. The vertical stacking of the "arms" of trough crossbed sets

FIGURE 3-7: Sketch from an outcrop photo of member B3. Sp/St refers to tabular crossbeds with laminae that are locally tangential to the lower bounding surface. Key for other abbreviations is given in Figure 3-6.

produces an apparent bimodal herringbone appearance (Fig. 3-7; Plate FlC). High-angle cross-stratification also occurs in thin (2-10cm) tabular sets with pronounced planar foreset cross-stratification. Ripple stratification (type B climbing ripples of Jopling and Walker, 1969) was observed in several outcrops. Bedding plane structures, in order of relative abundance, include linear asymmetric ripple marks (cf. Picard and High, 1973), sand-filled desiccation cracks, parting lineation, and superposed (runoff or ladderback) ripple marks.

The upper part of member B3 contains mostly trough cross-bedded arenite and interbeds of volcanic granulestone and conglomerate with relatively minor amounts of small-scale cross-stratification and horizontal stratification. Bedding plane structures are much less abundant in this part of the member. The arenites are sublithic, coarser-grained and better indurated than below, and contain up to 15% rounded fragments of felsic volcanic detritus. Trough to undulose tabular beds are generally thicker (20-40cm) than in beds below, and have low-angle ($<20^\circ$) cross-stratification. Interbeds of massive, moderately sorted granulestone-conglomerate rest on planar to slightly scoured bases and are composed of a framework of subround granules and cobbles of felsic volcanic detritus packed in a matrix of hematite-rich sand. Conglomeratic beds increase in abundance as member B3 grades into member B4, concomitant with an increase in average bed thickness. Some of the conglomeratic sheets are composite and contain thin (<10 cm) layers and lenses of parallel laminated fine-grained arenite.

Paleocurrent patterns for member B3 are complex (Fig. 3-4) and have a wide range in single-station dispersal patterns and regional paleoflow. Some stations show a relatively strong unimodal pattern, whereas most stations show a broad unimodal pattern and, in one case, a distinct bimodal pattern. The westernmost exposures of member B3, at Elizabeth Lake, contain dominantly southward-directed paleocurrent indicators, whereas paleocurrents inferred for the central part of the outcrop belt record flow to both the south and northwest. As member B3 grades into B4, paleocurrents shift to a uniform west-directed transport throughout the outcrop belt.

Interpretation

P The lower part of the member is inferred to have been deposited by a low-energy shallow fluvial system as suggested by the thinness of the beds and fine grain size. The stratification sequence is unlike any described from fluvial facies models, although the relative abundance, scale and types of sedimentary structures most closely resemble those described from modern ephemeral streams (cf. Picard and High, 1973, Williams, 1968). Scour-fill structures and obvious cyclicity are notably absent.

Within the context of modern ephemeral stream bedforms and depositional processes, observed sedimentary structures are interpreted as follows: low-angle inclined stratification was formed by foreset migration of transverse bars and "sandwaves" (two-dimensional large ripples of Harms, et al., 1982) in

which the inferred shallow water depth retarded development of flow separation, and so steep foresets could not form (cf. Picard and High, 1973). Horizontal stratification may represent bar top (thin, <10cm sets), channel thalweg, sheetwash, or longitudinal bar accretion during upper flow regime and possibly lower flow regime conditions. The lack of microcross-lamination implies a relatively rapid loss in stream velocity ("flashy" behavior) such that lower flow regime ripple bedforms did not develop. The large-scale sets of inclined planar stratification also represent transverse bar or sandwave deposits. The reactivation surfaces that occur in these crossbeds probably formed during a rapid drop in flow stage, as suggested by the simple, rather than compound, character of the crossbeds (Harms, et al., 1982). The small-scale tabular planar sets represent deposition from transverse two-dimensional ripples. Linear asymmetric ripple marks, runoff ripples, and desiccation cracks all attest to the presence of channel margin/bar top (eg. topographically positive) areas, and in the case of the latter structure, frequent subaerial exposure. The terms "channel" and "bar" are somewhat misleading in this interpretation because no evidence of channels was observed (eg. scour surfaces with an intra-clastic and/or pebble lag). Possibly the depositional process involved brief inundation and flow during which the depositional surface became essentially a sheet with bedforms of varying scales, but no well-defined channels. Scour features, normally developed as a result of flow constriction, would

only develop after significant bedform topography had formed, a characteristic that is not indicated by the thin character of bedding in member B3.

The upper part of the member records deposition in a relatively high-energy fluvial environment as suggested by the increased grain size and bed thickness. The massive granule-stone/conglomerate sheets record brief periods of flooding when extrabasinal basement fragments, mostly Sloan Group felsic volcanic rocks, were swept into the river system. The lack of stratification in these beds suggests both shallow water and rapid deposition. The composite character of some sheets implies amalgamation of superposed flood deposits. The trough crossbeds formed by migration of large-scale three-dimensional ripples (dunes). The dominance of stratification formed by migration of these bedforms suggests relatively longer periods of inundation compared to the lower part of the member. The lack of recognized sand and/or gravel bar facies implies a poorly developed bar-channel morphology and a low-relief, sheet-like depositional surface, similar to the lower part of the member.

Single station variability of paleocurrent rose diagrams and high dispersion are expected features of ephemeral streams (Picard and High, 1973; Smith, 1971) and are attributed mainly to local modification of flow by bedform topography. The variability may also reflect bar and/or sandwave morphology i.e. bars may have had a concave-downcurrent "horned" appearance, similar to the cross-channel bars of the sandy South

Saskatchewan River (Cant and Walker, 1978), which would produce a dispersed unimodal to distinct bimodal transport pattern. The significance of the regional paleoflow pattern is difficult to determine, although one point seems clear: the southward-directed paleoflow for stations within the Fault River valley imply no geomorphic expression along most of the fault system during this period of deposition. The regional pattern is tentatively interpreted to record the effects of regional paleoslope (towards the south) and local deflection caused either by channel sinuosity (i.e. an internally braided meandering channel, cf. Rust, 1978; Jackson, 1978) or local paleoslope modification (i.e. the morphology of the abandoned lobe/sheet of member B1-B2 conglomerates or a structurally controlled (?) topographic high south of Fault River) (Fig. 3-9A). The change in regional paleoflow, from dispersed to unimodal west-directed at all stations in overlying member B4, indicates a different style of fluvial deposition and paleoslope orientation.

Member B4

Member B4 (the Volcanic Litharenite facies of Kerans, et al., 1981) is the thickest and most prominent member within the Fault River Formation, attaining a thickness of more than 200m. The member gradationally overlies member B3, and is sharply overlain, with apparent concordance, by coarse conglomerates of member C2 of the Lady Nye Formation (Bunn Creek conglomerate). As is the case with the underlying units,

this member pinches out along the north-northeast edge of the preserved basin. The only systematic vertical change within this member is a change in pebble composition: overall lithologic character, sedimentary structures, and inferred paleoflow are uniform.

This member is composed almost exclusively of moderately-sorted, coarse-grained, pebbly, quartz-rich litharenite to sublitharenite dominated by coarse grains of polycrystalline quartz derived from a deformed plutonic source area (Plate FlE). The most prominent lithic clasts are subrounded fragments of red felsic volcanic rock that generally comprise less than 15% of the rock, although some of the coarser beds may contain up to 50% volcanic rock fragments. Pebbles of volcanic composition predominate in the lower half of the member, but decrease upwards to less than 5%, and are replaced by well-rounded pebbles of white vein quartz. Beds of in situ mudstone, and arenite finer than medium-grained sand, are totally absent, although the former does occur as rare intraclasts. The member is generally non-hematitic, in contrast to the dark red colour of member B3, and has a tan-white pigmentation derived from authigenic clay minerals (Plate FlE). Sheets of loosely-packed volcanic paraconglomerate are restricted to the lower part of the member.

The only sedimentary structure displayed by the arenites is large-scale cross stratification. Individual sets are generally thicker than 50cm; one 15m crossbed was observed, although the average set thickness is typically 1-2m

(Plate F1D). Trough and tabular-planar crossbeds predominate. Trough crossbeds occur as sets of interwoven (stacked) simple, festoon sets (Plate F1D) and appear to grade into tabular-planar sets. The intermediary crossbed is a tabular-shaped set, up to 5m thick, that rests on a slightly scoured surface. Laminae are tangential to the lower bounding surface and in plan view have a gently concave downstream shape. The lamination in such sets is typically shallow (about 20°). Tabular-planar sets also are common and show a wide range in set thickness (0.5-15m). Foreset stratification is defined by alternating inclined layers of massive arenite, pebbly arenite, and rarely sandy conglomerate. Small-scale stratification is absent, although in a few examples, tabular-planar crossbeds less than 30cm were observed. Several features are notable in their absence such as a lack of vertical organization of crossbeds, no evidence of reactivation surfaces and/or bedform superposition within the large-scale crossbed sets, and the low percentage of conglomerate (<5%).

Conglomerate, as relatively continuous sheets composed largely of material coarser than 4mm, is surprisingly sparse within this member. In the lower part of the member, sheets of matrix-supported, sand-rich volcanic paraconglomerate and granulestone occur as interbeds in the arenites and comprise about 15% of the member. The upper part of the member contains very little conglomerate: most material coarser than sand occurs scattered along laminae and does not occur in significant concentrations. Exceptions include rare small sets

(<40cm) of tabular planar cross-stratified gravels. Both massive paraconglomerate and/or orthoconglomerate are absent from this part of the member.

Paleocurrent measurements are regionally uniform. All stations have a tight unimodal pattern that indicates flow to the west (Fig. 3-4). This is in contrast to the highly variable paleocurrents, both at individual stations and regionally, typical of member B3. Paleoflow for member B4 is essentially parallel to paleoflow in the overlying basal Dady Nye Formation (Fig. 3-4).

Interpretation

This member is interpreted as a high-energy fluvial deposit based on unimodal paleocurrents, coarse grain size, and bed thickness. Genesis of the cross-stratification is inferred on the basis of a comparison with bedforms in the Brahmaputra River (Coleman, 1969), which attain a large size and seem to be a reasonable analogue for the large-scale crossbeds in member B4. The change in morphology from trough-shaped to tabular-shaped crossbeds, including the commonly observed intermediary tabular-shaped sets with asymptotic foreset laminae and trough-like plan shape, represent variations in bedform crestline sinuosity and secondary flow turbulence. Tabular planar sets represent straight-crested bedforms such as two-dimensional ripples (cf. Harms, et al., 1982) and sandwaves (Coleman, 1969). Many sedimentologists have suggested that tabular planar sets in

sandy fluvial sequences represent linguoid or transverse bar migration (Smith, 1970; Miall, 1977; Cant and Walker, 1978). However, in the examples described by these authors, the azimuths of the foresets are at a high angle to the channel azimuths, suggesting a significant component of lateral bedform migration. This was not observed in member B4, and thus tabular planar sets are interpreted as the deposits of two-dimensional bedforms (cf. Harms, et al., 1982). With increasing turbulence and flow velocity, two-dimensional bedforms changed into three-dimensional bedforms with more sinuous crests, that deposited the trough-shaped crossbeds. Flow in the member B4 river system probably was relatively perennial although, alternatively, flood stage flow may have lasted long enough for large-scale bedforms to develop. The lack of cyclicity suggests stable/perennial flow and relatively continuous vertical aggradation, although low-flow stage deposits may not have been preserved (cf. Cant, 1976). The lack of cross-stratification less than 20cm thick is attributed to the coarse grain size of the member which probably suppressed ripple formation (Harms, et al., 1982). The lack of conglomerate sheets, usually interpreted as a bar or streamflood facies in fluvial deposits, is attributed either to nonavailability of this size material in the source area and/or a high sediment supply to the basin coupled with rapid subsidence (presumably this would result in rapid aggradation with insufficient time for winnowing of coarse sand).

The striking regional consistency of the paleocurrents indicates both a high-energy and ~~high~~-sinuosity fluvial system and formation of a regionally continuous paleoslope to the west (Fig. 3-9B). The west-dipping paleoslope continued throughout deposition of the rest of the Hornby Bay Group. The source area for the detritus contained a mixture of felsic volcanic lithologies and deformed granitoid rocks and was probably several kilometers west of Wopmay Fault (Bowring, pers. comm., 1982). The orientation of paleocurrents relative to the strike of the Fault River Fault system seems to indicate a period of quiescence along this fault system (Fig. 3-4C).

Summary

The Fault River Formation records two periods of progradational fluvial sedimentation that infilled a shallow southwest-trending basin superposed on a west-dipping paleoslope. Exposures of the Fault River Formation are largely fault-bounded remnants that, based on paleocurrents and lithologic correlation between outliers, are part of a once extensive (greater than 1800km²) fluvial system (Fig. 3-9A,B). Evidence for syndepositional movement along the Fault River Fault system is:

- A. Paleocurrents (members B1 and B2) are directed orthogonally away from the fault (transverse drainage).
- B. All member B1 conglomerates are of local derivation.

FIGURE 3-8: Schematic tectonodepositional reconstruction for members B1 and B2. The large open arrows represent the westward-dipping paleoslope.

C. Weathering of Great Bear Batholith is thickest in the Fault River region; in areas to the north, that are overlain by Lady Nye Formation, the weathered zone is very thin as a consequence of erosion before the deposition of the Lady Nye Formation. This suggests that the Fault River region must have subsided relative to surrounding areas.

D. Paleocurrents in member C2 (of Lady Nye Formation) are parallel to the trend of the fault system (longitudinal drainage).

E. The coarsest conglomerate in member C2, and the entire Lady Nye Formation, occurs in exposures adjacent to the Fault River Fault, which (in addition to member C2 paleocurrents that are parallel to the fault system) are compatible with the existence of a fault-parallel depression that controlled drainage.

The lower upward-coarsening cycle records progradation of conglomeratic alluvial fan facies (Fig. 3-8). These fans are inferred to have initially built out from a fault-controlled scarp, and were later buried by streamflood fan deposits (member B2). The more exotic composition of B2 conglomerates indicates incision and back-wasting of local topography, with resultant derivation of clasts from an expanded drainage area that included Sloan Group volcanic rocks and Bigbear Formation arenites. The abrupt appearance of B3 arenites records the onset of a period of deposition by low-energy ephemeral streams that occupied the "trough" during conditions of

FIGURE 3-9: A: Schematic diagram of measured (solid arrows) and inferred (dashed arrows) paleoflow for member B3. The "topographic high" offers a possible explanation for the paleocurrent pattern (see also Figure 3-4B). B: Schematic paleogeographic reconstruction for deposition of member B4. Sloan Group and deformed granitoids represent the source area lithologies. F.R.F. = Fault River Fault.

tectonic quiescence. The influx of volcanic detritus associated with the coarsening of B3 arenites represents rejuvenation of a basement source area to the east-northeast, and erosion of Sloan Group volcanic rocks and deformed granitoid rocks. This cycle of upward-coarsening culminated in deposition of basal Lady Nye Formation conglomerates, and burial of Fault River Formation strata.

The upward-coarsening cycles in the Fault River Formation could be autocyclic (fan head entrenchment and progradation of depositional lobes) or allocyclic (tectonic lowering of the basin floor and/or climatic change) (cf. Heward, 1978; Steel, et al., 1977). Inferred syndepositional movement along Fault River Fault system suggests upward-coarsening in response to the lowering of the basin floor, particularly for the lower members (members B1 and B2). Although the upper cycle appears to have been partly affected by fault activity, inferred to be the cause of basement rejuvenation, climatic factors also were influential, as implied by the change from ephemeral-style fluvial deposits of member B3 to perennial deposits of member B4 deposits. The predominance of streamflood paraconglomerates in the lower two members, and the inferred perennial character of member B4 rivers, implies a humid climate (Rust, 1979; Schumm, 1977) although member B3 deposits may have occurred during a period of relatively drier climate.

The stratigraphic position of the Fault River Formation, relative to member C2 of the Lady Nye Formation and the Bigbear Formation, invites correlation of Fault River

Formation deposition with deposition of members A6 and A7 of the Bigbear Formation. In both areas, deposition would have occurred under the influence of sporadically active normal faults. Such normal faulting is interpreted as a response to tensile stresses generated during regional westward tilting; the faults imposed occasional local control on fluvial drainage patterns within the context of overall westward tilting.

Plate FlA: Aerial view of the Fault River Fault (FRF) that separates Sloan Group volcanic rocks (S) from northwest-dipping (to the right side of the photo) deposits of the Fault River Formation. View is looking southwest towards Great Bear Lake.

FlB: Massive paraconglomerate typical of member B2. Moderately well-rounded cobbles are composed of rhyolite (R), Bigbear Formation arenites (A), and vein quartz (V). Matrix is coarse-grained quartz-rich sublitharenite.

FlC: Stratification within member B3 quartz arenite. Individual beds display either horizontal stratification (H) or small-scale cross-stratification (C).

FlD: Crossbedding typical of member B4 volcanic litharenite. Crossbed sets display trough shape (St) or tabular-planar shape (Sp). Note asymptotic shape to base-of-foreset laminae in Sp sets and apparent opposed paleocurrent sense where Sp set overlies right edge of St set (hip level of geologist).

FlE: Photomicrograph of coarse-grained sublitharenite of member B4. Note high proportion of polycrystalline quartz and well-developed quartz overgrowths. Dissolution of quartz cement and framework was contemporaneous with illite authigenesis (I). Scale bar is 0.5mm long. Crossed nicols.

FlF: Photomicrograph of intergranular cement in member B3 quartz arenite. Framework composed of fine-grained quartz (Q) with a cement composed of silica (S), hematite (opaque), and cubes of crandallite group mineral goyazite, a Sr-Al phosphate. Scale bar is 15u long. Plane-polarized light.

PLATE F1

A

B

C

D

E

F

CHAPTER 4

LADY NYE FORMATION

Deposits of the Lady Nye Formation, the youngest fluvial system of the Hornby Bay Group, comprise arenites and minor conglomerates that outcrop discontinuously over a large area (7000km²) between Dismal Lakes and Great Bear Lake. Smaller regions of outcrop also occur on Leith Peninsula and south of Mouse River. Although this formation has a greater areal extent it is not as well exposed as the Bigbear or Fault River formations. Most regions of outcrop are gently rolling hills, covered with "boulder deserts" of frost-heaved sandstone (felsenmeer), that expose little continuous stratigraphic section (cf. Plate L1A). This, in combination with lithologic homogeneity and a lack of marker beds, has frustrated attempts at precise measurement of stratigraphic thicknesses, and documentation of vertical and lateral facies relationships. Based on scattered stream cuts and cliff faces, a provisional lithostratigraphic succession has been recognized. In general, this subdivision can be used to successfully differentiate arenite members on a regional scale.

The Lady Nye Formation is composed largely of quartz arenite and feldspathic arenite, and at the base of the formation, conglomerate. It unconformably overlies and onlaps volcanic rocks of the Sloan Group and granitoid plutons of Great Bear Batholith. The formation rests conformably on both the Bigbear and Fault River Formations. The formation is

characterized by a gradual upward fining and thinning as it grades into marine dolostones of the East River Formation. A minimum thickness of about 2.5km has been calculated from map pattern and structure west of Leith Line; a substantially thinner succession (<500m) outcrops east of this structure.

The Lady Nye Formation has been subdivided into four informal members (Fig. 4-3): Member C1, an assemblage of feldspathic arenites and conglomerates, is restricted to regions south of Dismal Lakes, and underlies member C2. Member C2 (Bunn Creek facies of Ross, 1980; Kerans *et al.*, 1981) comprises two informal lithofacies, a lower conglomeratic facies and an upper sandy facies. The upper sandy facies grades into overlying member C3 arenites (Cape MacDonnell facies of Ross, 1980; Kerans *et al.*, 1981) which in turn grade into the "mixed facies" (sandstone and shale) of member C4.

The Lady Nye Formation is named after Lady Nye Lake, a small lake near the middle of the outcrop belt (Fig. 4-1). At no single locality can all four members be observed in full stratigraphic succession; a type area for each member will be designated as the unit is described.

BASEMENT WEATHERING AND PALEOTOPOGRAPHY

Lady Nye Formation unconformably overlies and onlaps a rugged, deeply dissected paleotopography developed in granitic and volcanic rocks of Great Bear Batholith and Sloan Group, respectively (Great Bear Magmatic Zone, Hoffman, 1982). In general, the effects of pre-Lady Nye pedogenic alteration of

FIGURE 4-2: Schematic stratigraphic sequence for Lady Nye Formation; because no absolute thickness values are available, thicknesses of members are drawn in relative proportions.

basement are minor, and apparently are restricted to the upper few meters of basement rocks below the unconformity. Nowhere was a deep weathering profile observed. Alteration consists of hematization of iron silicate minerals and, in volcanic rocks, ferruginous aphanitic groundmass. In several examples, granitic rocks less than 1m below the unconformity still contain pristine hornblende and biotite. The relatively thin character of the weathering zone is inferred to be due to erosion and truncation of the weathering profile by the Lady Nye fluvial system, and does not reflect the nature of the paleoclimate.

Basement rocks occur as topographically rugged inliers that range in size from tens of square meters to square kilometers (Plate L1A). Margins of the larger inliers are typified by long re-entrants commonly filled with Lady Nye Formation arenite. Present valleys between basement inliers are underlain by Lady Nye Formation arenites, suggesting that the present topography is a strong reflection of paleotopography during deposition of Lady Nye Formation. The maximum relief between valley floors and adjacent basement paleohighs is at least 260m. Although subsequent tectonic movement has in places obscured syndepositional topography, some inferences can be made on the basis of the present-day topography and its relationship to Lady Nye Formation paleocurrents.

The area between Teshierpi Fault and Bigtree-Fault River Fault (Fig. 1-2 and 4-1) is of particular interest because it forms a relatively coherent block unaffected by faults. From a

distance, the northwest side of this block appears to be a flat-topped plateau. From a vantage point immediately north of Lady Nye Lake, the planar plateau top is easily visible and can be extrapolated across paleovalleys to the tops of basement inliers to the east. On the northwest (down-thrown) side of Teshierpi Fault (northeast of Lac Rouviere), a large elliptical knob of granite is surrounded and overlapped by Lady Nye Formation arenites (Plate L1A). Paleocurrents from the arenite are subparallel to the margins of the inlier, and implies infill of a paleovalley that was parallel to the long axis of the inlier (Fig. 4-4).

Available evidence suggests that the paleotopography at the base of the Lady Nye Formation is the remnant of an incised planation surface. This surface has been referred to as the Teshierpi-Bigtree peneplain (Kerans et al, 1981); the term "peneplain" here is used in the literal sense of the word (ie. a region that is almost a plain) rather than the genetic sense (cf. Meyerhoff, 1976; Thornbury, 1964). The re-entrant morphology of inlier margins reflects bedrock spurs, whereas discrete knobs probably are detached and degraded spurs (cf. King, 1967). Paleocurrents (Fig. 4-4) indicate flow parallel to and around (divergent pattern) basement highs of what must have been a more or less through-flowing fluvial system. Incision of the Teshierpi-Bigtree peneplain is thought to record a period of uplift and west-southwest-directed tilting that led to lowered base level and erosion by the Lady Nye river system.

MEMBER C1

This member crops out only south and west of the west end of Dismal Lakes and south of Lac Rouviere (Fig. 1-2 and 4-1). Outcrops in both areas are included within this member because even though the outcrops are lithologically distinct, they occupy a similar stratigraphic position (below member C2a).

The outcrops south of Lac Rouviere are in a 60m thick sequence of arenite capped by 10 meters of paraconglomerate. The unit is overlain conformably by conglomerates of lower member C2a, and rests unconformably on slightly weathered granitic rocks of Great Bear Batholith. The lower part of the sequence is composed of moderate to poorly sorted arkosic and feldspathic dark red hematitic arenite. These rocks are ubiquitously crossbedded, with trough crossbeds the predominant type. Paleocurrents (Fig. 4-4) are unimodal to the south-southwest, directed away from the immediately adjacent granitic highlands, and represent a deviation of more than one quadrant from the westerly-directed paleocurrents of overlying member C2a. The sheet of paraconglomerate at the top of the member is massive, with a moderately packed texture, and is composed of subround cobbles (<10cm) of vein quartz and granitoid rocks in a hematitic sandy matrix.

Member C1 south and west of west Dismal Lakes comprises hematitic, red arenites and subordinate paraconglomerates of unknown thickness that rest unconformably on granitic and volcanic rocks. The member is assumed to grade into overlying member C2b pebbly arenites, but the contact is poorly exposed

and/or faulted. Dark red hematitic feldspathic arenites and pebbly arenites compose most of member C1 and display trough cross-strata (<50cm) and 30-50cm beds of horizontal lamination. Interbeds of paraconglomerate form less than 5% and occur as massive sheets less than 1m thick, composed of subangular boulders of massive vein quartz in a hematite-rich muddy matrix. Subangular cobbles also occur as discrete clasts within arenite beds. Paleocurrents for a single station (Fig. 4-4) are towards the north, parallel to those in overlying arenites of members C2b and C3.

Interpretation

Both occurrences of member C1 are interpreted as fluvial deposits of locally-derived clastic material. The local provenance of units is indicated by the composition of cobbles within conglomeratic beds: granitoid cobbles are common in outcrops adjacent to granitic basement rock, and vein quartz cobbles predominate in outcrops close to occurrences of giant quartz veins (not differentiated in Fig. 4-1). The inferred stratigraphic position of these rocks, below members C2a and C2b, and their restricted lateral extent of outcrop, suggests that this member represents remnants of earlier fluvial systems that existed before the formation of a regionally extensive fluvial system. If this hypothesis is correct, then this member in part may be time-equivalent with the upper member(s) of the Fault River Formation.

MEMBER C2

Member C2 of the Lady Nye Formation (Bunn Creek facies of Kerans, et al., 1981) has been subdivided into lower and upper lithofacies assemblages that are distinguished on the basis of grain size differences. The estimated thickness of member C2 east of Leith Line is about 350m. The lower lithofacies contains massive conglomerate irregularly interstratified with pebbly arenite, whereas the upper lithofacies is composed entirely of pebbly arenite. The contact between these lithofacies is gradational over 5-10m. Although the member outcrops west of Leith Line (Fig. 4-1; south of Dease River), the upper and lower contacts are not exposed.

Lower Lithofacies

This lithofacies ranges in thickness from 0-20m, and is composed of variable proportions of interbedded conglomerate, pebbly arenite, and rare muddy siltstone. It is well-exposed in several spectacular stream cuts (Bunn Creek and the headwaters of East River and Teshierpi River; Fig. 4-1) and underlies an extensive area of low-relief outcrop in the Fault River region. Exposures in the Fault River region are designated as a type area. The unit rests unconformably on Wopmay Orogen basement rocks, and displays a conformable relationship with the underlying Fault River Formation, Bigbear Formation and member C1.

The distinguishing characteristic of this lithofacies is the composition of the cobbles. In decreasing order of

FIGURE 4-3: Measured stratigraphic section through basal member C2a orthoconglomerates in a cliff section at the northeast corner of Hornby Bay. The arrows (right of the column) represent paleocurrent measurements.

abundance, prominent constituents are red to pale-coloured quartz arenite, vein quartz, quartzite, peloidal siliceous ironstone (jaspillite), and black stromatolitic chert. Cobbles of local basement provenance (ie. volcanic and granitic rock) are conspicuously subordinate to absent. The distinctive clasts of jaspillite are extremely useful for stratigraphic correlation and provenance determination. They probably were derived from siliceous ferruginous paleosols of the Tree River Formation (Hoffman, pers. comm., 1983) or equivalent Mara Formation, which outcrop in the Tree River Fold Belt of Wopmay Orogen and Kilohigok Basin, more than 100km and 300km to the east, respectively. Occurrence of these clasts has been used to establish interformational/interbasinal correlation shown in Figure 1-4. The clasts of quartz arenite probably were derived from the Bigbear Formation (members A5-A6), whereas the stromatolitic chert and quartzite pebbles were derived from the Rocknest and Odjick(?) Formations of Wopmay Orogen.

Although no systematic attempt was made to delineate regional variations in conglomerate clast size (mean and maximum), certain qualitative generalizations can be made. The coarsest grain size (40cm; with 23cm maximum for jaspillite clasts) and thickest accumulation of conglomerate (20m) occurs in the Fault River region. Clasts of equivalent grain-size crop out 40km to the northeast along strike, whereas to the northwest, the clast size seems to decrease. Thus the thickest and coarsest conglomerate occurs along a northeast-trending belt that defines the axis of an inferred depositional trough.

Conglomerates occur as massive sheets of weakly indurated ortho- to paraconglomerate, and rare sets of cross-stratified paraconglomerate (Fig. 4-3). Orthoconglomerates are most common, and comprise scour-based tabular sheets of tight-packed well-rounded cobbles and boulders with a matrix of coarse-grained arenite; in several outcrops, distinct imbrication was observed (Plate LlB). Paraconglomerates are finer grained and are composed of 20-50% cobbles that float in a matrix of purple hematitic quartz litharenite. The paraconglomerates commonly overlie undulose scour surfaces, and in some places have an intact framework at the top of the bed (Plate LlC). Where the unconformity is exposed in the headwaters of Teshierpi River, paraconglomerates display large-scale (2m) tabular planar crossbed sets.

Coarse-grained to pebbly arenites occur as lenticular to sheet-like interbeds within conglomerates. The arenites are dark red to mottled purple and white, and range from quartz litharenites to feldspathic arenites with remarkably fresh feldspars (Plate LlD; microcline, plagioclase, orthoclase). In outcrops dominated by orthoconglomerate, arenite occurs as non-pebbly lenses (10-50cm) that drape the underlying pebble surface, and have scoured top surfaces (Plate LlB); the arenite may have a massive (Sm), cross-stratified (Sp, St), or horizontally laminated (Sh) texture (Fig. 4-3). In outcrops dominated by paraconglomerate, the arenites occur as 0.5-2m thick tabular- to trough-crossbedded sets with scattered pebbles (Plate LlC). The pebbles are nowhere concentrated

along laminae, a feature considered to be hydrodynamically significant. At the Teshierpi River exposure of the unconformity, arenites occur in spectacular cosets of megatroughs in which individual sets are more than 2m thick, and up to 20m wide in plan view.

In Bunn Creek, poorly sorted medium- to fine-grained micaceous arenites are common (Plate LlC), and occur as lenses with planar-undulose bases that truncate underlying structures and a deeply scoured top surface. Internally they contain horizontal to low-angle ($<10^\circ$) planar lamination. Locally, these arenites contain a well-developed phosphate cement composed of goyazite and fluorapatite.

Mudstone and siltstone beds are rare, and were observed in but a single outcrop where they comprise several lenses of parallel-laminated tan-coloured muddy siltstone with well-developed convolute lamination, cusate current ripple-marks, and desiccation cracks.

Depositional Cyclicity

Depositional cyclicity is difficult to recognize in dominantly conglomeratic sequences. This is the result of amalgamation of successive conglomerate units as suggested by the lenticular character of the arenite interbeds. In the paraconglomerate sequences (cf. Bunn Creek, Plate LlC), cyclicity is more obvious and consists of an ideal sequence of massive paraconglomerate (Gms) overlain by cross-stratified pebbly arenite (St) and capped by plane-bedded micaceous

arenite (Sh). The latter facies is irregularly distributed through the sequence, and may succeed either, the Gms or St facies.

Paleocurrents

All paleocurrents are strongly unimodal towards the west (Fig. 4-4). In several outcrops, paleocurrents were measured from both cobble imbrication and cross-stratification, to assess the difference in orientation of the two fabrics. The difference is negligible ($<30^\circ$), and therefore subsequent paleocurrent measurements were obtained exclusively from crossbeds. The west-directed paleocurrents, in conjunction with the inferred source area for the peloidal ironstone clasts, indicate that these clasts have been transported at least 150km. In association with the coarse size of these clasts (up to 23cm), transport and deposition by a river system of significant power and dimensions would seem implicit.

Interpretation

This facies is interpreted as being a high-energy, dominantly gravel and coarse-sand-bedload fluvial deposit not unlike the modern proglacial Donjek River (cf. Boothroyd and Ashley, 1975; Miall, 1977) or Klifandi fan in Iceland (Tankard, et al., 1981, p.133). Orthoconglomerates are inferred to have formed by vertical accretion on longitudinal gravel bars; interbedded lenticular arenites represent remnants of

slack water bar-top sandy bedforms (cf. Rust, 1978). The lenticular shape of these arenites may be the result of erosion by subsequent floods, or it may reflect the hummocky upper surface of the conglomerate sheet (cf. Williams and Rust, 1969). Sand-rich multistory sequences dominated by paraconglomerate (cf. Plate L1C) formed by different processes. A typical depositional cycle records streamflood deposition of sand-rich paraconglomerate within broad channels, followed by dune migration (St) and/or winnowing of the upper surface of the conglomerate during falling water (cf. Plate L1C). In events of somewhat lower flood energy, large-scale sandy bedforms deposited crossbeds composed of pebbly arenite. As suggested by random occurrence of the pebbles, this detritus was deposited quickly, and deposition did not involve the migration of superposed bedforms of different grain size to the bedform brinkpoint (cf. Smith, 1970). The lateral extent of the planar erosion surfaces at the top of Gms and St facies (tens of meters) is in contrast to the relatively local extent of scour surfaces at the bases of trough sets, implying that the planar surfaces may have formed during a channel-wide drop in water level (shoaling) that caused bevelling and reworking of the channel topography. The horizontally laminated arenites (Sh) represent lower flow regime(?) slack-water deposits. Trends for both maximum clast size and conglomerate thickness suggest that the Fault River region was a depositional sag that channelled the early Lady Nye fluvial system.

Clasts of peloidal ironstone in member C2 conglomerates were derived from a source area that is presently at least 150km east of present outcrops of the conglomerate. These clasts were derived from units that were unaffected by thrusting in the adjacent Asiatic Fold Belt of Wopmay Orogen (i.e. the Tree River Formation is part of the autochthonous cover within the Orogen) and probably did not extend farther to the west than their present location (Hoffman, 1983, pers. comm.). Why are these clasts present in member C2 and how were they transported such tremendous distances? The ironstone clasts are extremely durable and could resist both mechanical and chemical abrasion which would thereby allow them to remain in the fluvial system for a long period of time. If the clasts had been transported by normal perennial stream flow, a significant gradient in slope should have existed between the Tree River Fold Belt and present outcrops of member C2 conglomerate. An alternate hypothesis invokes episodic transport of conglomerates during flood periods i.e. streamflood paraconglomerates. Most occurrences of ironstone-bearing conglomerates are, however, framework-supported, a fabric that is not expected from streamflood deposition. It is inferred that the orthoconglomerate fabric resulted from in situ winnowing following flood peaks. The paraconglomerate bed shown in Plate L1C is an example of this process. Similar processes of transport and deposition have been inferred for distinctive rhyolite clasts of the southern California borderland, some of which have been transported up to 320km from their source area (Kies and Abbot, 1983).

Upper Lithofacies

The upper lithofacies assemblage in member C2 consists of an essentially homogenous sequence of cross-laminated hematitic pebbly arenites of uncertain thickness (200m is a conservative estimate). It is the predominant lithology in most basement paleovalleys, and is widespread east of Leith Line. The unit onlaps basement paleohighs (cf. Plate L1A) and probably overlies member C1 arenites south of Dismal Lakes (Fig. 4-1). It grades laterally and vertically into member C3 arenites from which it may be distinguished on the basis of grain size, sedimentary structures and sequence, and to a lesser extent, colour. Because of its gradational relationship to member C3 some identifications are rendered ambiguous. Southwest of Dismal Lakes, this lithofacies grades over 10m into the mixed clastics and carbonates of the East River Formation, although the transition is not well-exposed. East of Teshierpi Fault, basal siliciclastics of the Dismal Lakes Group rest unconformably on this lithofacies (Fig. 4-1).

This lithofacies is composed of varicoloured hematitic, locally feldspathic, quartz sublitharenite. Colours range from pale orange-buff to dark red, and pigment homogeneity commonly is disrupted by irregular "bleach zones". The arenite is moderately to poorly sorted, and is cemented by silica, hematite, illitic clays, and rare authigenic feldspar. Pebbles of vein quartz and mudchips (up to 5%) are concentrated in discontinuous pebble-thick lags at the base of crossbeds or scattered irregularly throughout arenite beds (Plate L1E). In

general, the maximum and median grain size and the bed thickness decrease with increased stratigraphic height.

Most outcrops consist of flaggy weathering sets of trough crossbeds that are stacked without recognizable cyclicity. In some cases, variations in set thickness may be used to recognize upward thinning cycles (1-2m thick); however, outcrop-wide erosion surfaces (cf. Plate L2A) between sets and cosets are not apparent, and thus recognition of depositional cyclicity is difficult. The intensity of scour at the base of trough sets decreases up-section such that the coset geometry typical of this lithofacies grades into the distinctly stacked sheets of member C3:

Paleocurrents (Fig. 4-4) in all outcrops are unimodal towards the southwest to northwest, more or less parallel to paleocurrents in the under- and overlying arenites.

Interpretation

The upper lithofacies of member C2 represents a sand-dominated fluvial environment of lower energy than the underlying lithofacies, thereby implying a decrease in source-area relief and slope gradient. Decrease of depositional energy with time also is in accord with the up-section decrease in grain size, bed thickness, and depth of scour between sets. Trough crossbeds formed by migration of sinuous megaripples (large-scale three-dimensional ripples of Harms, et al., 1982); the tabular-planar sets presumably are the deposits of straight-crested bedforms. The lack of "bar facies"

(i.e. well-developed tabular-planar sets, cf. Smith, 1970; Collinson, 1970; Cant and Walker, 1978) is curious, and suggests that the fluvial bed surface was covered almost exclusively by transverse dune-like bedforms of different sizes. The lack of laterally extensive planation surfaces in the upper lithofacies of C2 is the result of destruction of this surface (if formed) by migration of successive bedforms. The up-section decrease in depth of scour at the base of trough crossbeds may be the result of decreased flood energy through time, or it may reflect decreased variability in bedform morphology with respect to crestline height (cf. Cant and Walker, 1978). Both hypotheses imply decreased current energy with time.

MEMBER C3

This member (Cape MacDonel facies of Kerans, et al., 1981) comprises spectacular bluff sections preserved under sills of resistant Western Channel Diabase along the north shore of Hornby Bay (Plate L1F) and in a stream cut south of Dease River (Plate L2A). The cliff sections along the north shore of Hornby Bay are designated as a type area for this member. Elsewhere the member crops out in rubbly hills that do not lend themselves to detailed sedimentologic analysis. The thickness of this member is unknown, as there are no areas in which both the upper and lower contacts are exposed. However, south of Dease River, more than 1.7km of apparently continuous stratigraphic section is exposed. The member is inferred to be

Figure 4-5. Detailed measured section of exposure of member C3 along the north shore of Hornby Bay (type area). Stream power decreases to the left, arrows to right of column represent paleocurrent azimuths. Tabular palmar sets with angular and tangential foresets are not differentiated.

the lateral facies equivalent of upper member C2; although the transition is not well-exposed, the contact with the overlying arenites of member C4 also appears to be gradational.

Lithologically, member C3 is rather homogenous and composed mostly of medium-grained, locally pebbly, quartz arenite (>90%), feldspathic arenite (<5%), and green to red micaceous siltstone and mudstone (<5%). The quartz arenites are extremely pure and display a moderately sorted to well-sorted framework cemented by quartz overgrowths (Plate L2D) and less abundant illitic clays. A pale green colour is common in many of the quartz arenites and is imparted by abundant (up to 3%) authigenic green tourmaline that occurs as intergranular films and spherulitic clusters (Plates L2F). Feldspathic arenite is texturally identical to quartz arenite but contains 10-15% fresh feldspar, commonly with overgrowths (Plate L2C). Feldspathic arenite has mostly been seen in outcrops of the upper part of the member from Dismal Lakes south to East River, although the alteration products of feldspars have been identified in other areas. Pebble- to granule-sized grains of vein quartz (<5%) occur as discontinuous lenses at the base of crossbed sets, and less commonly, as dispersed grains in association with mudchips. Micaceous green siltstone and green to red mudstone occur rarely between cosets of arenite, increasing in abundance and thickness upwards in the member. (Plate L2E). In uppermost parts of member C3, these mudstones commonly display mudcracks.

The arenites occur in laterally extensive (ten of meters) stacked sheets up to 1m thick (Plate L2A). Individual sheets are separated by low-relief undulose scour surfaces cut into underlying mudstone/siltstone or arenite; in outcrop, the mudstone layers between sheets appear only as thin (<5cm) recesses (Plate L2A). Each sheet is a coset of dominantly tabular-planar sets, with gently curved asymptotic foreset laminae (Plate L2B), and less common tabular sets with angular toesets and trough crossbeds. Marked grain-size variation between adjacent foreset laminae (cf. Smith, 1970; Collinson, 1970) occurs exclusively in tabular-planar sets containing cross-laminae that make an angular rather than tangential intersection with the lower set-bounding surface. Cosets that contain only one type of crossbed are composed of 2-5 sets that do not appear to vary systematically with respect to set thickness and orientation, either within cosets or between adjacent cosets. In cosets with more than one style of cross-stratification, horizontal lamination commonly occurs towards the top of the coset. With increasing stratigraphic height, horizontal lamination composes a greater proportion of each coset coincident with an upward thinning of cosets. Undulose truncation surfaces that bound cosets are overlain by parallel-laminated red to green micaceous siltstone and/or mudstone.

Paleocurrents in all measured outcrops are unimodal (Fig. 4-4) and towards the west, in accord with patterns recorded in underlying units. However, a distinct shift is apparent from

westerly to more northerly-directed flow with increased stratigraphic height, especially in outcrops between the south shore of Dease Arm and the north shore of Hornby Bay (Fig. 4-1).

Interpretation

Member C3 was deposited by a sand-rich-bedload fluvial system (braidplain) that drained towards the west-northwest. The stacked, multistory character of the cosets and their pronounced sheet-like geometry suggests discrete depositional pulses in shallow, probably very broad, channels such that flow was essentially unconfined. Greater preservation of coset geometry, smaller size of crossbeds, and greater proportion of mudstone/siltstone within this member, in comparison to member C2b, imply a less energetic depositional environment, in accord with the inferred proximal-distal relationship between these members. Scoured bases to cosets record the onset of flood conditions, whereas planar top surfaces of cosets most likely formed during channel-wide emergence. The latter conclusion is based on the ubiquitous occurrence of suspended load (low flow) micaceous mudstones above this surface. Tabular planar sets, with asymptotic foreset laminae that have no marked grain-size variation, formed by migration of sinuous-crested bedforms with effective lee side eddies (cf. three-dimensional large-scale ripples of Harms, et al., 1982; Jopling, 1965). Tabular-planar sets with angular foresets resemble crossbeds formed by the migration of transverse

• sandbars (cf. Smith, 1970; Cant and Walker, 1978; Miall, 1977; Collinson, 1970). The lack of grain-size variation between laminae in these sets may be due either to a homogenous / grainsize population or to a lack of superposed (second-order) bedforms (i.e. flood-stage bedforms with little modification by low-flow bedforms). If the latter reason is the cause of the observed textural homogeneity of the crossbeds, then an abrupt change in fluvial regime and flow energy as a consequence of rapid shoaling after floods is implied. Ideally, each coset formed by a) scour of muds during rising flood stage; b) initiation and migration of dune and rare bar bedforms; c) rapid decrease in flow depth (such that second-order degradational bedforms did not develop) with consequent planation of the depositional surface followed by d) deposition of suspended load mudstone/siltstone. Sets of horizontal lamination, which first appear at the top of cosets lower in the member, may represent either upper flow regime flow conditions during emergence (cf. Eriksson, 1978) or lower flow regime conditions during the waning stages of flood flow. Horizontally laminated arenite common in the upper part of the member probably formed during lower flow regime; its increased abundance up section, coincident with decreased average bed thickness, implies a general decline in fluvial energy with time.

MEMBER C4

This member ("unnamed" facies of Kerans et al., 1981)

comprises four distinct lithofacies assemblages that record marine transgression of the Lady Nye system. A lack of observed interfingering between adjacent lithofacies, absence of basinwide marker units, and poor exposure of the upper and lower contacts of lithofacies, precludes accurate paleogeographic reconstructions. However, lateral facies relationships can be constrained on the basis of apparently isochronous, or nearly so, storm deposits in lower East River Formation. Red arenite and mudstone (lithofacies A) is the predominant lithofacies of member C4 in the eastern regions of Lady Nye Formation and grades vertically (and laterally) into drab (white to gray) arenite and mudstone (lithofacies B). Lithofacies C is composed almost exclusively of white quartz arenite with minor green mudstone. Lithofacies D, a distinct assemblage of hummocky cross-stratified arenites and siltstones with black mudstone interbeds, occurs exclusively in the Narakay Islands (Fig. 4-1).

Lithofacies A, B, and D have been observed to grade into the overlying dolostones and siliciclastic rocks of the East River Formation; the boundary between the two formations is arbitrarily chosen as the first occurrence of dolostone (Kerans, et al., 1981). In sections east of Leith Line, East River dolostone first appears as beds of intraclastic conglomerate, whereas to the west, basal East River Formation is marked by oolitic dolarenite and/or stromatolitic dolostone. The appearance of East River Formation dolostone in a dominantly siliciclastic sequence marks the onset of decreased

sediment supply and inception of "clear water" carbonate sedimentation.

Lithofacies A: Red Arenite and Mudstone

This lithofacies constitutes all of member C4 that outcrops between Dismal Lakes and East River (Fig. 4-1), and also occurs in fault-bounded outcrops in Blood Creek (along with lithofacies C of this member). South of Ritch Island (Dease Arm, Great Bear Lake) a 30m-thick sliver of thin-bedded pink quartz arenites, not readily assignable to other lithofacies within this member, is tentatively placed within lithofacies A. Although the contact is poorly exposed, lithofacies A apparently pinches out east of Leith Line, where East River Formation dolostones rest almost directly on member C3 arenites (see Fig. 5-3). The outcrops south of Ritch Island are the westernmost exposures of this lithofacies, but this most likely reflects fault truncation rather than true paleogeographic extent. Towards the top of sections measured at Falcon Lake and East River (Fig. 5-3), the red pigmentation of these rocks gives way to a drab gray to brown colour similar to an observed lateral (westward) change in colour (drab colours predominate in member C4 outcrops in Dease Arm). The lower contact of lithofacies A with member C3 arenites is nowhere exposed, but is inferred to be gradational and essentially a continuation of the upward-thinning trend for member C3. The maximum measured thickness for lithofacies A is 40m; the contact with East River Formation is marked by beds of carbonate-bearing intraclastic paraconglomerate.

The predominant lithologies in lithofacies A are fine- to medium-grained, moderately sorted feldspathic quartz arenites containing disseminated fine hematite grains in a tightly packed framework of subangular to subround grains. Angular chips of red mudstone are a common constituent that occur at the bases of arenite sheets. The arenites occur as tabular to lenticular sheets (<20cm) irregularly interbedded with flat-laminated red mudstone (Plate L3A). Tabular sheets rest on slight scour surfaces, whereas lenticular sheets rest on deeper scours (*i.e.* channel-fill). Tops of thicker arenite beds are capped by undulose planar erosion surfaces (Plate L3A).

The only internal structure in these otherwise massive arenites is rare, poorly developed normal grading or horizontal lamination; cross-stratification is absent. Interlayers of red mudstone, which range from thin (<1cm) films between arenite sheets up to 20cm-thick beds, commonly display well-developed desiccation cracks (Plate L3E). Vertical variation within this facies consists of a gradual increase in the sandstone:shale ratio and decrease in lenticularity of sandstone beds.

The large exposure of pink quartz arenites south of Ritch Island, tentatively correlated with lithofacies A to the east, comprises a succession, at least 35m thick, of thin-bedded quartz arenite and green to brown mudstone. The arenites show an upward thinning of beds; they are mature fine-grained quartz arenites in tabular sheets 5-10cm thick, and display either a massive texture or a faint parallel lamination with local ripple cross-stratification. The mudstones occur as thin

films between arenite sheets, and decrease in abundance towards the top of the section. Bedding plane structures include sand-filled mudcracks and subaqueous shrinkage cracks (cf. Donovan and Foster, 1972) and linear symmetric wave ripple marks. Synsedimentary dewatering structures (pipes) and related anticlinal warps occur towards the top of the section.

Interpretation

The arenite-mudstone couplets (Plate L3A) probably record flood events in a low-energy distal fan/fan delta environment. The massive to graded arenites are inferred to have been deposited rapidly from fluvial sheetfloods (sensu stricto, Hogg, 1982). The moderately-sorted texture of these deposits, which is atypical of rapid deposition, is inherited from the inferred "upstream", facies-correlative, well-sorted arenites of member C3. Emergence of the depositional surface under conditions of upper(?) flow regime led to erosion of the depositional surface and formation of undulose-planar top surfaces of arenite sheets. This was followed by suspension deposition, and subsequent desiccation, of red mudstone.

Lithofacies B: Drab Arenite-Mudstone

This lithofacies is widespread in outcrops in and around Dease Arm, and also occurs within the upper few metres of Lady Nye Formation at Falcon Lake and East River. The maximum thickness of the lithofacies is unknown; a minimum of 40m has

been measured in a section in eastern Dease Arm, where the upper contact of this lithofacies with basal East River Formation was observed. Although similar strata are present on the Narakay Islands, sedimentary structures and arenite composition are sufficiently different to warrant inclusion of Narakay Island outcrops within lithofacies D rather than lithofacies B.

Lithofacies B is composed of interbedded white to greenish or pink, medium-grained subfeldspathic quartz arenite and green to brown-black siliceous mudstone. The scale of arenite/mudstone intercalation ranges from centimetre-scale (wavy to lenticular bedding; Plate L3C) to decimetre-scale (stacked tabular arenite sheets, up to 40cm, separated by thin mudstone films or a bed of wavy-lenticular strata (Plate L3B). Thicker arenite beds are massive or display low-angle ($<10^\circ$) to horizontal lamination, and trough- to tabular-shaped crossbeds. Thinner beds are massive or contain ripple cross-stratification and/or horizontal lamination. Tops of arenite beds, especially the thinner sets ($<10\text{cm}$), display low-relief interference ripple marks. The mudstones commonly contain subaqueous shrinkage cracks infilled with sand; on some bedding surfaces the infills form an interconnected subpolygonal network that probably is of subaerial origin. Podiform bodies of nodular quartz, inferred to be silicified "cauliflower" evaporites, (cf. Chown and Elkins, 1974), occur locally within this lithofacies (Plate L3B).

Interpretation

The thin bedded arenite-mudstone component of this lithofacies likely was deposited on a mixed sand-mud tidal flat (cf. Reineck and Singh, 1975). Thicker-bedded dominantly arenaceous sequences may represent deposits of sand-rich portions of the flat (e.g. shoal/flat margin, cf. Owenshine, et al., 1976); massive arenite beds may have been deposited by storm/washover processes (Schwartz, 1982). Unfortunately, paleocurrents are not available to confirm the influence of tidal currents on deposition. Because this lithofacies is transitional upwards to East River dolostones, it probably overlies arenite-rich lithofacies C.

Lithofacies C: Quartz Arenite

This lithofacies is best exposed in continuous outcrop along a strike-parallel west-trending ridge south of Dease Arm. Elsewhere it outcrops sporadically, and has not been observed east of longitude 119°0. To the west it disappears under the Quaternary cover of Cape MacDonnell peninsula, but it does outcrop on a single island north of the Narakay Islands (Fig. 4-1). The stratigraphic position of these arenites is uncertain; apparently they underlie hummocky cross-stratified arenites (lithofacies D) of the Narakay Islands, and underlie drab arenites and mudstones (lithofacies B) to the east. Unfortunately, critical exposures of lateral facies relationships are lacking. The thickness of this lithofacies is unknown; even though it is reasonably well-exposed south of

Dease Arm, fault repetition and lack of marker beds effectively precludes thickness determinations: 50m is a conservative estimate.

Lithofacies C is composed almost entirely of well-sorted medium-grained quartz arenite that is white to pale mauve or green, but also contains subordinate intraformational conglomerate and mudchip-bearing arenite. Homogeneous grain size, in addition to the affinity that lichen has for this particular lithology, makes recognition of sedimentary structures almost impossible. The arenites occur in continuous sequences to distinctly stacked cosets of cross strata; in the latter instance the cosets are separated by thin (<5cm) green mudstone layers. Crossbeds comprise noncyclical sets, 10-30cm thick, of trough, and less commonly tabular-planar, cross-stratification. Exhumed bedding surfaces, up to 400m², display abundant cusate current ripple marks and oscillation/interference ripple marks, both draped by thin green mud films. In one example, a long trough-like depression is covered with sinuous-crested asymmetric current ripple marks, with axes perpendicular to the axis of the depression, and the axes of a superposed set of small-scale wave ripple marks are perpendicular to the first set (Plate L3D). Additional structures include horizontal lamination and small-scale crossbeds (in sets 10-40cm thick), and large but sparse trough crossbeds (1-3m). Two occurrences of intraformational conglomerate were noted; one example (Blood Creek) is a 1m-thick orthoconglomerate bed of 5-10cm white quartz arenite cobbles in a sandy white

matrix. This conglomerate bed overlies trough-crossbedded arenite, and is overlain by a bed of horizontally-laminated arenite. In the other occurrence, south of Narakay Islands, subangular cobbles of quartz arenite "float" in a sandy super-mature matrix; the stratigraphic position of this outcrop is uncertain, and it may even be within member C3 (so indicated, for the sake of simplicity, in Figure 4-1).

Paleocurrents (Fig. 4-4) are commonly bimodal and in several examples are bimodal-bipolar. In the latter case, modes occur along a roughly north-south axis. A similar pattern is exhibited by the acute bimodal roses, most of which have modes in the northeast and northwest quadrants. The strong northern component is parallel to the apparent paleo-slope trend deduced for upper member C3.

Interpretation

A marine origin for this lithofacies is proposed primarily on the basis of bimodal to bimodal-bipolar paleocurrent patterns. Although the extreme maturity of this lithofacies is suggestive of marine reworking, this characteristic is considered inconclusive because of the high maturity of C3 fluvial arenites, which presumably represent an "upstream" sample of detritus that was transported into the marine environment. Because of the lack of sufficient detail no particular depositional environment can be proposed, but a tide-influenced sand-dominated setting is regarded as most appropriate.

The northern component of many of the paleocurrents is

parallel to paleocurrents in underlying fluvial arenites (member C3), and most probably reflects paleoslope-controlled shoreline-perpendicular bedform movement. This is interpreted as a reflection of ebb-and flood-tidal currents with deviations due to the interaction of bedform topography and the distribution of sand bodies (cf. Boothroyd, 1979; Hayes, 1980).

Lithofacies D: Hummocky Cross-Stratified Arenite

This lithofacies is exposed only on the Narakay Islands where it consists of at least 130m of feldspathic arenite/siltstone and black mudstone and grades upwards into clastic and stromatolitic dolostone of East River Formation. The directly subjacent lithofacies is unknown, because no lower contact with the Narakay Island succession is exposed. The nearest outcrop of member C lithologies is separated from the Narakay Islands by a deep (700m displacement on bounding faults) graben (Fig. 4-1).

Lithofacies D is composed of fine-grained feldspathic arenite, siltstone and black mudstone, commonly with thin streaks of arenite or siltstone. Irregular nodules and bands, up to 10cm thick, of very dense black chert occur locally at the top of both arenite and mudstone beds. Samples of arenite and siltstone display moderately well-sorted, tightly-packed frameworks of quartz, chert, feldspar, mica, zircon and tourmaline. The composition of the arenites appears to remain more or less constant upwards within the section, except for

FIGURE 4-6: Schematic section of facies D hummocky cross-stratified fine-grained clastics on Narakay Islands. Sections to the right of main column show detail of specified interval.

influxes of locally-derived volcanic detritus. The gradational contact with East River Formation is locally marked by the occurrence of "calcareous arenite", a recrystallized oolitic, quartz arenite with carbonate cement, and intraclastic to stromatolitic dolostone. Sequences of internal structures and geometry of arenite beds form a variety of subfacies that vary systematically in a vertical succession that can be interpreted in terms of changing bathymetric conditions (Fig. 4-6).

Arenite Beds

Lenticular Arenites: Lenticular-bedded arenite and siltstone predominate in the lower part of the lithofacies. These rocks occur as scour-based lenses interlayered with sand-streaked to ribbon-bedded sandstone/mudstone (Plate L4A). Lenticular arenite bodies are either massive or comprise cosets of hummocky to low-angle cross-stratification and horizontal lamination. Bed lenticularity decreases and bed thickness increases as this subfacies grades into the overlying tabular arenite.

Tabular Arenite: Tabular to wavy-undulose tabular beds of arenite are the most typical bedding type within lithofacies D. They are bounded above and below by scours, some of which have marked variations in relief. Arenite beds occur in couplets with sand-streaked mudstone or, as is more common, in amalgamated cosets (Plate L4B). Arenite beds rest on scour surfaces and comprise cosets of two or more sets bounded by

low-angle truncation surfaces. The tops of sets commonly display horizontal lamination parallel to the lower bounding surfaces and/or hummocky cross-stratification (Plate L4C). Flat lamination composes a larger fraction of each arenite coset with increased stratigraphic height. Amalgamated cosets form compound cosets up to 8m thick (Plate L4B) in which depositional breaks are marked by thin recessional intervals that correspond to mudstone layers (<5cm). Individual laminae (5-10mm) display subtle normal grading and a concentration of mica flakes at the top of each lamination (Plate L3F).

Ribbon-Bedded Arenite-Mudstone: This lithology occurs throughout lithofacies D and is referred to as "graded rhythmite" in Figure 4-6. It is interbedded with sand-streaked mudstone, and commonly occurs in beds that separate arenite cosets (Plate L4B). Each bed (5-20cm) consists of a scoured base and an overlying massive to faintly-graded arenite, commonly containing black mudchips at the base, that grades into rhythmic arenite-mudstone couplets 5-15mm thick (cf. graded rhythmites of Reineck and Singh, 1972; Nilsen, 1982). In some occurrences of this subfacies, the massive to graded division of beds is absent, and amalgamated sets of rhythmic beds occur instead (Plate L4E). Soft-sediment convolutions of laminae are locally present, and are truncated by the overlying set of rhythmites (Plate L4F).

Mudstone Beds

This brown to dense black lithology occurs as thin (<5cm)

scoured remannts between arenite cosets or as thicker beds (up to 20cm) higher in lithofacies D. The characteristics (texture, thickness, percent sand) of this lithology are best observed in wave-washed shoreline exposures; lichen cover and recessive weathering hamper work on inland exposures. The mudstones most commonly display parallel lamination draped over underlying topography (Plate L4D). Some of the thicker (5-20cm) mudstone beds, which occur exclusively in the upper 20-30m of lithofacies D, contain sand in the form of thin (1-3cm), wavy to lenticular sheets containing graded bedding, ripple cross-lamination, and "injection structures" (Plate L4D). On bedding planes the injection structures form an irregular "bird's foot" pattern instead of a regular polygonal pattern, indicating that they are more closely allied to subaqueous shrinkage cracks rather than subaerial shrinkage cracks (Plummer and Gostin, 1981; Donovan and Foster, 1972).

Interpretation

Lithofacies D formed by deposition during alternating periods of storm (hummocky cross-stratified and massive arenite subfacies) and fairweather (mudstone and sand-streaked mudstone subfacies) sedimentation. Bed shape, internal structure, and vertical sequence are nearly identical to storm-dominated deposits from the Phanerozoic described by Dott and Bourgeois (1982), Mount (1982), Bourgeois (1980), Clifton and Hunter (1982), Harms et al., (1975), and Goldring and Bridges (1973). Recognition of this depositional facies

provides important paleobathymetric constraints, because all the above authors agree that hummocky cross-stratified arenites are deposited in subtidal environments, although estimates of maximum water depths range from 40 to 80 metres (Dott and Bourgeois, 1982). In addition, the occurrence of hummocky cross-stratification within a significant portion of member C4 constrains the shoreline paleoenvironment to one that was strongly influenced by wave activity.

As a consequence of the recent popularity of storm-process depositional models for interpretations of the rock record, a well-defined facies model is available to test interpretations of the subfacies of lithofacies D (Fig. 4-6). The two cycles that are present within this facies (a lower upwards thickening and coarsening, overlain by an upwards thinning and fining cycle, Fig. 4-6) represent a progradational pulse followed by shoaling of a relatively mud-rich depositional environment, respectively. Overall shoaling is indicated by both decreased bed lenticularity with increased stratigraphic height, and the occurrence of mudcracks at the top of the succession. The lower upward-thickening and coarsening cycle is attributed to a local influx of coarser clastic material, analogous to depositional lobes in turbidite fans (Mutti and Ricci Lucchi, 1978; Ingersoll, 1978).

The lenticular arenite beds in lower lithofacies D represent the most distal offshore deposits within the succession. Their mode of deposition most probably involved a turbidite-like flow of low concentration (cf. Lowe, 1982) in which

storm-suspended (by wave oscillation and shoaling) fine-grained clastic sediment was transported offshore (cf. Brenchley and Newall, 1982; Cant, 1980). Overlying tabular arenite bodies are more typical of deposition above storm wave base (Dott and Bourgeois, 1982). Each coset represents deposition by a single storm event that involved sea-bottom erosion followed by oscillatory or rhythmic suspension deposition of hummocky cross-stratified sets. Shape-graded laminae within these sets is consistent with deposition from suspension, as has been suggested by other workers (Dott and Bourgeois, 1982; Clifton and Hunter, 1982). The common occurrence of scour-separated sets of hummocky cross-stratification within cosets may reflect fluctuations in storm and wave activity. Some of the cosets have deeply scoured top surfaces draped by mudstone that must have formed during waning storm activity. Amalgamated cosets probably were formed as a result of storm removal of fairweather mudstones. The graded rhythmite subfacies records small depositional pulses also related to storms and vigorous wave action (Reineck and Singh, 1972). Although this subfacies generally has been thought of as a "deeper" subtidal (below storm wave base) deposit (Dott and Bourgeois, 1982; Reineck and Singh, 1972), in the example described here it also occurs in the shallow-water part of the succession. The mudstone facies records tidal and wave-influenced "fairweather" sedimentation on a mud-rich sand-starved substrate (cf. de Raaf, et al., 1977). Flat-laminated arenite at the top of some hummocky

cross-stratified cosets mark a decline in wave energy and oscillatory flow, possibly related to storm passage (cf. Clifton and Hunter, 1982). Lithofacies D is therefore inferred to record subtidal shelf (below storm wave base) to lower "shoreface" storm-dominated deposition. The term "shoreface" is in quotation marks because, although most authors place the upper limit of hummocky stratification within the lower shoreface (and hence imply an upper shoreface/beach), no such facies were recognized in member C4.

Discussion of Member C4

Member C4 lithofacies record a complex mosaic of subtidal marine to terrestrial depositional environments. Broadly speaking, this assemblage of facies may be referred to as a delta complex in that deltas "are defined more broadly as coastal accumulations, both subaqueous or subaerial, of riverderived sediments adjacent to, or in close proximity to, the source stream, including the deposits that have been secondarily molded by various marine agents,..." (Wright, 1979, p.5). Furthermore, the Lady Nye system deltaic component could be referred to appropriately as a fan delta i.e. an alluvial fan that prograded directly into a standing body of water (Holmes, 1965; McGowen, 1970; Westcott and Ethridge, 1980). Because of the apparent lack (or nonrecognition?: Jackson, 1978) of meandering fluvial deposits in pre-Devonian rocks (Cotter, 1978; Rust, 1978) and the dominance of bed-load river systems, it is probable that most, if not all,

pre-Devonian fluvial systems terminated in fan deltas. Therefore the occurrence of fan delta deposits in pre-Devonian rocks could not be used to indicate tectonic setting except for sediments composed of coarse clastics of obvious local derivation (cf. Gloppen and Steel, 1981; Westcott and Ethridge, 1980; Dutton, 1982). Many models for deltaic sedimentation are based on the characteristics of modern deltas (cf. Coleman and Wright, 1976; Coleman and Pryor, 1981) that typically have a relatively narrow active channel system (compared to the size of the delta) that delivers sediment to a standing water body. The ubiquitous occurrence of stacked sheet-like depositional units within the fluvial facies of the Lady Nye Formation does not indicate the existence of a central or major (deep?) channel system (Plates L2A, L3A). Therefore, such models may be inappropriate for fan deltas in a nonvegetated setting, and hence no attempt is made to define the Lady Nye system delta in terms of analogous modern or post-Devonian models. However, it is possible to constrain models of Lady Nye delta sedimentation in terms of major processes that affect modern deltas (i.e. wave, tide, and river energy; Coleman and Wright, 1975; Galloway, 1976).

Wave Energy: The most important clue to the scale and frequency of wave activity that affected the Lady Nye system is provided by the occurrence of a thick (140m+) unit of storm-deposited hummocky cross-stratified fine-grained clastics (lithofacies D). Occurrence of this type of stratification over a stratigraphic interval of more than 140 metres implies

that the Lady Nye system coastline was dominated by wave action and storms.

Tide Energy: Tidal regime is inferred to be within the mesotidal (or lower) range (4m). This conclusion is based mostly on negative evidence i.e. the lack of arenaceous facies deposited by offshore/foreshore tidal sand bodies (cf. Anderton, 1976; Homewood and Allen, 1981) that are typical of macrotidal environments (Hayes and Kana, 1976; Wright, 1978), especially those undergoing marine transgression (Nio, 1977).

Fluvial Energy: Fluvial influence of the dynamics of the Lady Nye system coastline is inferred to have been minor. This is based on the thin-bedded character and fine grain size of terrestrial deposits (cf. lithofacies A) which implies a shallow low-energy fan delta/braidplain.

It is thus inferred that member C4 records deposition on a wave- and storm-dominated mesotidal shelf and coastline that fringed a low-energy north-northwest-facing fan delta. A proposed paleogeographic/paleoenvironmental model is shown in Figure 4-7C. In this model, each of the four lithofacies in member C4 are interpreted as isochronous. Fan delta/floodplain facies (lithofacies A) are inferred to have graded shorewards into tidally-influenced delta plain clastics of lithofacies B. Arenites of lithofacies C are interpreted as a high-energy sand-flat facies that was in part coeval and in part older than the tidal flat facies of lithofacies B. Hummocky cross-stratified arenites and siltstones of lithofacies D formed offshore. No shoal/beach facies was recognized in the Dease

Arm area, although clearly one must have existed in order to reduce incoming wave energy and supply sand to the offshore region. It is therefore proposed that the coarse-grained arenites of lithofacies C may have acted as submergent shoals that reduced wave energy. The apparent restriction of hummocky cross-stratified facies to the Narakay Islands region implies the existence of a shoal region east of the islands.

SUMMARY

The following sequence of events/processes is inferred for the Lady Nye Formation:

- 1) West-southwest tilting and regional uplift, perhaps in part contemporaneous with upper Bigbear and Fault River Formation sedimentation, caused incision of the Teshierpi-Bigtree peneplain and formation of a rugged paleotopography with as much as 250m+ relief (Fig. 4-7A). During this period, much of the softer pedogenically-altered basement rock was stripped away, leaving little preserved regolith.
- 2) As peneplain incision progressed and valley headwalls were extended "upstream", the river system merged with a drainage system that may have originated from as far as 150 km to the east-northeast (indicated by previously discussed Tree River jaspillite cobbles), although the predominant constituent of member C2a conglomerates, the Bigbear Formation arenites, argues for substantial contribution of local detritus. During this period, deposition was restricted to

Figure 4-7: Schematic representation of the evolution of paleogeography of the Lady Nye system. A. Uplift and incision of Teshierpi-Bigtree peneplain during deposition of Fault River Formation and the upper two members of the Bigbear Formation, coincident with the onset of regional westward tilting. B. Beginning of deposition of the Lady Nye Formation. Both the Bigbear and Fault River Formation have been buried. C. Marine transgression of north-northwest facing high-energy siliciclastic shore zone. Dashed diagonal line in A is trace of the not yet active Teshierpi Fault.

A

B.

C.

fault-controlled depocenters such as the Fault River region, and possibly in areas where member C1 is preserved.

3) Crustal movement changed from uplift to subsidence, still with a west-dipping paleoslope, and led to accumulation of coarse fluvial conglomerates composed only of extremely resistant clasts derived from regions to the east (member C2a). A structurally controlled topographic depression in the Fault River region controlled early drainage patterns of the Lady Nye system (Fig. 4-7B), as indicated by variations in conglomerate bed thickness, cobble size and paleocurrents. In addition, paleocurrent and basement-cover outcrop patterns imply the existence of southwest-trending paleovalleys between Teshierpi and Bigtree Faults, and northeast of Lac Rouviere.

4) Continued subsidence and accumulation of Lady Nye system arenites led to infill of paleotopography; possibly this was the time of initial activity along the Leith Line, because the sequence of arenites west of the Line appear to be dramatically thickened relative to the eastern counterparts. Relatively high-energy sandy braided streams in basement highlands (member C2b) graded downcurrent into the lower energy braidplain facies of member C3.

5) Decrease in source area relief and slope gradient produced the overall upward-fining and thinning trend of the Lady Nye Formation. Late in the history of C3 braidplain deposition, a distinct northward component to paleoslope developed, and seems to have persisted throughout deposition.

of marine arenites of member C4 (lithofacies C) and perhaps also East River Formation. Decreased fluvial energy in the Lady Nye system was coincident with the onset of marine transgression and formation of a wave-dominated mesotidal coastline.

Plate Ll.

A. Ground view of exhumed elliptical "island" (monadnock) composed of Great Bear Batholith monzogranite surrounded by in situ and frost-heaved member C2b pebbly arenite (A). View looking northwest from member C2b arenite northeast of Lac Rouviere. Field of view is about 5km wide.

B. Basal member C2a orthoconglomerate, about 5m above member B4 of Fault River Formation, north shore of Hornby Bay. Trough crossbedded arenite channel facies (St) is overlain by longitudinal gravel bar orthoconglomerate and arenite facies (Gm and Sh respectively). Note well-developed imbrication in conglomerate (paleoflow right to left, towards the southwest). Sh arenites are inferred to be bar top and bar edge deposits.

C. Sequence of massive paraconglomerate sheets (Gms), trough crossbedded pebbly arenite (St); and micaceous arenite (Sm) in paraconglomerate-rich part of member C2a. St beds are inferred to be deposits of "channel" dunes; Gms beds are streamflood paraconglomerates; top of lowermost Gms bed (above head of geologist in lower right) has a "tightened" framework formed during winnowing of streamflood deposit. Sm arenites are low-flow facies formed during streambed shoaling; note that they overlie planar truncation surfaces and have scoured tops.

D. Photomicrograph of member C2b arenite from south of Dease River. Moderately sorted framework of monocrystalline and polycrystalline quartz and alkali feldspar (grain in partial extinction above scale bar). Cement comprises compaction-related pressure solution welds and minor quartz overgrowths succeeded by patches of illitic clay minerals that formed contemporaneously with post-overgrowth quartz dissolution. Crossed nicols. Scale bar is 1mm long.

E. Coset of several small-scale tabular crossbeds in pebbly arenite of upper member C2b. Pen (upper right) for scale is 14cm long.

F. Panoramic view of cliffs of member C3 arenite capped by sills of Western Channel Diabase, north shore of Hornby Bay (type area for member C3). Pleistocene boulder beach ridges in the foreground. Cliffs in background are about 100m high.

PLATE L1

A

B

C

D

E

F

Plate L2

A. Stacked multistory arenite sheets of member C3, south of Dease River. Each sheet comprises scour-based S_t and S_p flood deposits overlain by low-flow mudstones (recessive weathering). Scale bar is 1.5m high.

B. Tabular planar crossbed sets with tangential foreset laminae within arenite coset of member C3. Scale bar is 30cm high.

C. Feldspathic quartz arenite from upper member C2 east of Lac Le Roux. Alkali feldspars display inclusion-rich subhedral rims which are interpreted as authigenic overgrowths; plane-polarized light. Scale bar (left side) is 1mm long.

D. Photomicrograph of supermature quartz arenite shown in Plates 3A and 3B. Note quartz overgrowths and sutured grain boundaries; crossed nicols. Scale bar (upper right) is 1mm long.

E. Stacked sets of horizontal to low-angle cross-laminated arenite interbedded with red and green mudstones (dark recesses in cliff face). Geologist in middle of photo for scale. Upper part of member C2, north shore of Hornby Bay.

F. Authigenic tourmaline in member C3 arenite. Tourmaline (High relief mineral) occurs in intergranular areas as spherical bundles of acicular crystals associated with illite (light gray) and hematite (opaque). Tourmaline crystals crosscut older quartz overgrowth cements and framework grains implying that tourmaline precipitation was associated with quartz dissolution. Plane polarized light, scale bar is 250 μ m long.

PLATE L2

A

B

C

D

E

F

Plate L3

A. Lenticular to tabular-shaped arenite sheets and interbedded red mudstone of inferred distal fan origin (member C4, lithofacies A, East River area). Pen for scale (upper right) is 14cm long.

B. Interbedded parallel-laminated and massive arenite beds of inferred delta plain origin (member C4, lithofacies B, Prospect Island). Hammer head rests on lens of silicified "cauliflower" evaporites.

C. Wavy- to lenticular-bedded white quartz arenite and green mudstone, typical of lithofacies B. Note sand-filled shrinkage cracks above pen. Pen for scale is 14cm long.

D. Bedding surface exposure that displays a gentle trough-shaped scour covered with asymmetric current ripple marks (paleoflow left to right) that have a set of superposed wave ripple marks in ripple troughs. Member C4, lithofacies C. Hammer in center of photo for scale.

E. Well-developed mudcracks within red mudstone of member C4, lithofacies A. Pen in background is 14cm long.

F. Polished slab of hummocky cross-stratified siltstone from lower part of lithofacies D, member C4. Light and dark laminae comprise size- and shape-graded couplets composed of silt-rich layer overlain by a micaceous layer. Coin is about 1cm wide.

PLATE L3

A

B

C

D

E

F

Plate L4

A. Lenticular-bedded feldspathic arenite/siltstone and ribbon-bedded silty mudstone. Base of lithofacies D, Narakay Islands. Hammer for scale.

B. Amalgamated tabular arenite sheets from middle part of lithofacies D; arrows point to 20cm bed of graded siltstone-mudstone rhythmites. Hammer (left center) for scale.

C. Sets of hummocky cross-stratified fine-grained feldspathic arenite that compose part of a coset. Pencil for scale is 14cm long.

D. Detail of mud-rich interbed near top of lithofacies D. Structures include starved ripples of siltstone, sand streaks, sand-filled shrinkage cracks, and irregular load(?) structures at top of lowermost sand bed. Pen for scale is 14cm long.

E. Two sets of graded rhythmites of siltstone-mudstone laminae that overlie a massive to flat laminated arenite. Note truncated soft-sediment folds at top of lower rhythmite bed. Compare this photograph, taken of upper (shallow-water) lithofacies D, with deep-water graded rhythmites shown by Dott and Bourgeois (1982). Field of view is about 40cm wide.

F. Example of an irregularly convoluted contact between massive fine-grained arenite and graded rhythmites in upper part of lithofacies D. Penknife for scale.

PLATE L4

A

B

C

D

E

F

CHAPTER 5

EAST RIVER FORMATION

The East River Formation comprises four distinct members that record i) marine transgression and submergence of a siliciclastic shoreline and shelf (members E1 and E2), ii) progradation of carbonate tidal flats during regression and coastal offlap (member E3) and, iii) cannibalization of older East River Formation lithologies during syndepositional fault movement (member E4). This latter phase of sedimentation reflects the onset of syndepositional movement along the Teshierpi Fault, which apparently continued throughout deposition of overlying Kaertok Formation and lower Dismal Lakes Group clastics (Kerans, et al. 1981). The overall history of sedimentation recorded in the East River Formation is similar to that observed in many platform carbonate sequences (cf. Wilson, 1975; Read, 1982).

The East River Formation gradationally overlies siliciclastics of the Lady Nye Formation; the contact is chosen as the first appearance of dolostone. The contact between the East River Formation and overlying Kaertok Formation is designated as the point at which quartz arenite beds constitute at least 50% of the section; in most areas the contact is gradational over a 10-15m interval. In the Falcon Lake area (Fig. 5-2), Kaertok Formation siliciclastics disconformably overlie member E3 peritidal dolostone. No evidence of a break in sedimentation (e.g. calcretes or karstic dissolution surfaces) was

FIGURE 5-1: Schematic stratigraphic column showing member subdivisions, thicknesses, and inferred depositional settings for the East River Formation. See Table 1 for key to lithologic symbols.

LITHOLOGIES

QUARTZ ARENITE
CALCAREOUS ARENITE
SILICEOUS MUDSTONE
DOLOSTONE
G	GLAUCONITE

OOIDS
DOLOSTONE INTRACRYSTS
SILICEOUS INTRACRYSTS
EDGEWISE CONGLOMERATE
SOFT-SEDIMENT DEFORMATION

STROMATOLITE STRUCTURES

FLAT MAT
LATERALLY LINKED DOMES
SH-V COLUMNAR
SH-C COLUMNAR
DIGITATE/MICROCOLUMNAR
ONCOLITES

MUDCRACKS
RIPPLE MARKS
SILICIFIED EVAPORITES
ROUNDED INTRACRYSTS
SILICIFICATION

SEDIMENTARY STRUCTURES

HERRINGBONE CROSSBEDS
RIPPLE CROSS-LAMINATION
TROUGH CROSSBEDS
TABULAR-PLANAR CROSSBEDS
HUMMOCKY CROSS-STRATIFICATION
PARALLEL LAMINATION

TABLE 1

observed along the contact in this section, and it is not believed to represent a significant break in sedimentation.

The East River Formation crops out from south of Dismal Lakes west-southwest to Dease Arm (Fig. 5-2). The formation does not occur east of Teshierpi Fault (the section of the fault north of $66^{\circ} 50'N$), where it has been eroded from the upthrown block and redeposited as member E4 clastics. The eastern limit of deposition (depositional zero-edge) is unknown. The formation also occurs in southern Great Bear Lake on both Leith Peninsula and the Superstition Islands (Fig. 1-1). Regional correlations, based on the stratigraphic position of distinctive stromatolite bioherms, indicates that the minimum area of submergence during East River Formation sea-level highstand was close to $27,000\text{km}^2$. The thickness of this formation and individual members is not well known because of incomplete exposure and large gaps between measured sections, but gross trends have been recognized (Fig. 5-2). In easternmost exposures (south of Dismal Lakes, cf. Fig. 5-2), the entire formation is less than 100m thick; it is at least 150m thick at East River; farther to the west (in the Narakay Islands) and south (on Leith Peninsula) the formation is close to 200m thick.

The East River Formation derives its name from particularly well-exposed sections in cliffs and stream cuts along East River, in the northwest corner of map sheet 86K (Sloan River map-area) (members E1, E2) and where East River is ponded into two shallow lakes along the southern boundary to map sheet 86O

(Bebensee Lake), (members E1, E2, E3, and basal E4) (Fig. 4-2). This formation was referred to previously as Unit 9 (Baragar and Donaldson, 1973; Kerans, et al., 1981).

Member E1

The basal member of the East River Formation has been recognized throughout the entire map-area. Member E1 comprises those lithologies that occur between the first bed of dolostone (the top of the Lady Nye Formation) and the first appearance of microdigitate domal stromatolite bioherms (member E2). In earlier publications, both members E1 and E2 were placed within the "Lower Clastic Facies" (Unit 9A) (Ross, 1980; Kerans, et al., 1981). However, the lithologic character of each unit is sufficiently distinct to warrant designation as two map units.

Member E1 is composed of five lithofacies that record the transition from terrestrial and marine siliciclastic deposition to carbonate ramp deposition (cf. Ahr, 1973; Read, 1980; see later discussion). In the easternmost sections, this member is only 10m thick, but thickens westward to 70m (Fig. 5-2). The five lithofacies, in order of decreasing proportion of siliciclastic detritus, are: i) thin-bedded arenite-mudstone, ii) quartzose oolitic dolostone, iii) intraformational conglomerate, iv) intraclastic dolostone, and v) stromatolitic dolostone (Fig. 5-5, 5-6). The intraclastic dolostone lithofacies is composed entirely of dolostone and is distinguished from the intraformational conglomerate lithofacies which

Figure 5-2. Simplified fence diagram showing regional changes in thickness and distribution of members in East River Formation. Jagged terminations of top and/or base of column indicates absence of control on that contact; jagged terminations within columns indicates fault truncation. The base of member E2 has been used as a horizontal datum. Numbers in circles at the base of the fence diagram indicate distances between sections in km; approximate geographic positions are shown in the sketch map. Solid line at the base and/or top of sections indicates observed contacts (with Lady Nye Formation and Kaertok Formation, respectively). Note wedge-shaped geometry of member E4 and pronounced thickening near southwest end of Teshierpi Fault. Arrows at base of sections mark positions of inferred syndepositional hinge lines (also shown as stippled regions on sketch map); Leith Line on the right and the break-in-slope east of the Narakay Islands on the left. T.F.= Teshierpi Fault, F.L.F.= Falcon Lake Fault, F.R.F.= Fault River Fault.

contains fragments of both carbonate and siliciclastic composition. These five lithofacies occur interbedded in vertical sequences that display rare cyclicity and an overall increase in the abundance of carbonate-rich lithofacies. The lateral distribution and continuity of lithofacies is irregular which suggests a complex facies mosaic (cf. Laporte, 1967).

Thin-bedded Arenite-Mudstone Lithofacies

In many respects this lithofacies is identical to underlying thin-bedded siliciclastics of the Lady Nye Formation (member C4, facies A,B) but is distinguished by the occurrence of dolostone interbeds. The reader is referred to detailed descriptions of member C4 that were presented in the previous chapter.

In the eastern part of the outcrop belt (Fig. 5-3), the thin-bedded arenite-mudstone lithofacies occurs in 1-2m thick beds interlayered with intraclastic dolostone lithofacies, stromatolitic dolostone lithofacies, and beds of massive dolomite. Each bed is a coset composed of 10-15cm thick sets of parallel-laminated to massive, medium-grained feldspathic arenite separated by thin green mudstone laminae. In addition to quartz and alkali feldspar, framework grains include glauconite peloids and mudchips and silicified ooids (glauconite is here used to designate the morphologic glauconite; preliminary XRD analysis indicates that the peloids are in fact composed of the mineral glauconite; cf. McRae, 1972; Buckley, et al., 1978). The ooids commonly have a nucleus composed of a

quartz or feldspar grain enveloped in a cortex of a few thin concentric laminae (e.g. superficial ooids of Loreau and Purser, 1973). An unusual feature of these ooids is that the cortex commonly is replaced by fine sprays of acicular green tourmaline (dravite; XRD analysis) and less commonly by barite.

In the Falcon Lake and East River sections, this lithofacies essentially is a continuation of lithofacies A of member C4 (Lady Nye Formation) except that; a) there is a decrease in thickness and abundance of red mudstone beds, b) the mudstones change colour upwards in sections from red to greenish-black coincident with decreased pigmentation in the arenites, c) intraformational conglomerate lithofacies is common, and d) uncommon occurrence of nodular silicified evaporites within green mudstones and arenites. The inferred evaporitic origin of these nodules is discussed at the end of this chapter.

To the west (Prospect Island, Dease Arm; section 8, Fig. 5-2) the thin-bedded arenite-mudstone lithofacies comprises wavy-lenticular bedded arenites and mudstones similar to lithofacies B (member C4, Lady Nye Formation) except that the mudstones are more calcareous, the arenite beds are thinner, and nodular silicified evaporites occur.

In the Narakay Islands, this lithofacies is composed of stacked cosets of fine-grained arenite to siltstone inter-layered with 2-30cm thick beds of wavy bedded sand-streaked mudstone (Plate E1A). Arenite cosets are composed of 10-30cm

thick sets that display flat bases and scoured tops and are separated by 2-5cm thick layers of black to green mudstone. Arenite sets commonly are massive but also display parallel-lamination and low-angle cross-stratification. Bedding plane exposures of the tops of arenite beds display symmetric wave ripple marks commonly with a less prominent superposed orthogonal set (Plate ElC). The sets of sand-streaked mudstone between arenite cosets are composed of 2-10cm flat to ripple cross-laminated, undulose to lenticular arenite layers and 2-5cm layers of mudcracked black mudstone (Plate ElB).

Interpretation

Although the rocks that have been placed within this lithofacies are lithologically similar, they probably represent a variety of microenvironmental settings, for which detail is unavailable for complete documentation.

The cosets of flat-laminated arenite in the eastern sections represent a mixture of framework grains of both terrestrial and marine derivation, although the exact process of deposition is uncertain. In the Falcon Lake and East River sections, the thin-bedded arenite-mudstone lithofacies was deposited on a marine-influenced coastal floodplain. The appearance of drab-coloured arenite and black to green mudstone implies the presence of reduced Eh waters (brackish) that may have resulted from mixing of fresh and marine waters. In the Dease Arm area, the thin-bedded arenite-mudstone lithofacies records deposition in an inferred sand-rich tidal

flat. The examples of this lithofacies from the Narakay Islands represent alternating sand- and mud-rich tidal flat/sand shoal deposition, perhaps similar to the clastic facies in the mesotidal shoals of Oosterscheide Estuary in the Netherlands (Nio, et al., 1978). The arenite cosets probably record periods of traction current deposition by megaripples and lower flow regime plane beds and storm deposition of massive to flat-laminated arenite (cf. Terwindt, 1981). Finely interbedded arenite-mudstone is nearly identical to Recent deposits on mud-rich portions of tidal flats (Van Straaten, 1959; Reineck, 1967, 1972). Mudcracks and interference ripple marks suggest periodic emergence and desiccation. Interlayering of arenite cosets and sand-streaked mudstone reflects the close spatial relation between these two subfacies, a situation that is characteristic of intertidal estuarine sand shoals and tidal flats (cf. Owenshine, et al., 1976; Terwindt, 1981)

Quartzose Oolitic Dolostone Lithofacies

This lithofacies ranges in composition from an oolitic quartz arenite to oolitic dolostone (ooid grainstone). Medium-scale (10-30cm) cross-stratification is a ubiquitous sedimentary structure (Plates E1D, E2A) and occurs in cosets up to 10m thick (cf. Faloon Lake and Lac Le Roux sections; Fig. 5-3). A thick quartz arenite in the East River 86K section is included in this lithofacies (Fig. 5-3) because a genetic relationship is inferred even though it is lithologically different from

Figure 5-3. Detailed sections within member E1 along Leith Line, locations of sections shown on inset map. Base of East River Formation is marked by the appearance of dolostone intraclasts and or in situ dololite. Base of member E2 is marked by the occurrence of digitate domal stromatolites. Sections are correlated based on both the position of beds of intraformational conglomerate lithofacies and the prominent interval of quartzose colitic dolostone lithofacies. Note that the onset of carbonate deposition was diachronous with respect to siliciclastic deposition. See Table 1 for key to symbols.

typical exposures of this lithofacies. Cosets of quartzose oolitic dolostone lithofacies in Dease Arm area are generally thinner ($<1\text{m}$) than those to the east. Crossbed types such as trough, tabular-planar, and herringbone tabular sets are common (Plate E1D, E2A). Sets with ripple cross-stratification also occur, as do sets with low-angle ($<10^\circ$) to horizontal-planar stratification. In general, no vertical sequence to size or shape of crossbeds was observed.

The quartzose oolitic dolostone lithofacies displays regional variations in composition with respect to the ratio of siliciclastic grains to ooids, and the abundance of diagenetic chert nodules. In sections from East River to Falcon Lake (Fig. 5-3), quartz and feldspar grains are common; they may compose as much as 80% of a bed (oolitic quartz arenite). Commonly, however, they occur as cores to superficial ooids in quartzose oolitic dolostone (ooid grainstones) (Plate E2B). In occurrences of this lithofacies in the Dease Arm area, quartz and feldspar grains generally comprise $<20\%$ of a bed. Chert nodules (Plate E2A) and silicified ooid grainstones (Plate E2C) are prevalent in the easternmost sections, and absent from most sections in Dease Arm. In addition, neomorphism of carbonate framework grains and cements has all but obliterated primary textures in Dease Arm. Recrystallization also affected oolitic dolostone in the east but to a lesser degree, as evidenced by more complete preservation of ooid cortex laminae and a less sparry appearance to the rocks.

Nearly all of the paleocurrent data for the East River Formation (Fig. 4-4) are within this subfacies. The rose diagrams range from broadly unimodal to distinct bimodal to polymodal. No consistent regional trend can be recognized. In one example (Lac Le Roux), paleocurrent measurements from successive arenite cosets yielded almost diametrically opposed modes (Fig. 4-4).

Interpretation

The ubiquitous occurrence of large-scale crossbeds in this lithofacies suggests deposition from lower flow regime bedforms, analogous to those described from modern ooid sand shoals (Ball, 1969; Hine et al., 1980). A strong tidal influence is implied by the common bimodal to polymodal character of the rose diagrams, and the occurrence of herringbone crossbeds; unimodal paleocurrent patterns may reflect ebb- and/or flood-dominated tidal sedimentation (cf. Hayes, 1980; Boothroyd, 1978). Some of the sequences of stratification bear strong resemblance to modern siliciclastic examples from tidal inlets and deltas (Hayes, 1980) and mesotidal-macrotidal ridge and runnel systems (van den Berg, 1977; Fox and Davis, 1972). The apparent lack of reactivation surfaces, commonly considered typical of tidal-intertidal sand bodies (cf. Klein, 1970; Ginsburg, 1975) is unusual; it may imply that bedforms were more or less continually migrating (i.e. they maintained equilibrium with the flow conditions).

Between Falcon Lake and East River, the quartzose oolitic dolostone lithofacies comprises a stratigraphically continuous sheet, 5-10m thick, over an area of 500km² which suggests that in this area, this lithofacies represents a belt of nearly continuous tide-influenced nearshore ooid sand shoals. The thickest and most siliciclastic-rich occurrences of this lithofacies also occur in this area, further suggesting that this area was proximal to a source area of siliciclastic detritus. Based on composition, stratigraphic position, and paleocurrents, one likely source is represented by the quartz arenite body in the East River 86K section (Fig. 5-3). Chert-rich beds of this lithofacies also are particularly common in this area, and thus the inferred source of silica for the chert nodules may have been meteoric waters derived from a siliciclastic aquifer (cf. Knauth, 1979; Muir, et al., 1980).

Intraformational Conglomerate Lithofacies

The intraformational conglomerate lithofacies occurs as massive sheets of paraconglomerate interbedded with the thin-bedded arenite-mudstone lithofacies. It occurs only in the lower part of member E1, before the appearance of abundant dolostone. This lithofacies is most common in the Falcon Lake and East River sections but also occurs in Blood Creek (Fig. 4-2). It does not occur in the sections from south-central Dismal Lakes region nor has it been found in outcrops west of Ritch Island. The distribution and unique composition of this lithofacies makes it useful for paleogeographic and time-line reconstruction.

FIGURE 5-4: Sketch based on tracing of slabbed surface showing SH-V stromatolite developed on a substrate of intraformational conglomerate lithofacies (sample is from first stromatolite bed in section 2, Figure 5-3). 1) Angular clast of cross-laminated oolitic dolostone, 2) detrital infill containing dololutite rip-ups (2a) and a layer of dololutite (2b), 3) channel-fill of dololutite rip-ups and millet-seed sand, 4) SH-V stromatolite, and 5) silicified cauliflower evaporites.

This lithofacies occurs as massive sheets interbedded with the thin-bedded arenite-mudstone facies. It occurs only in the lower parts of member E1, before the appearance of abundant dolostone. Conglomerate sheets overlie either siliciclastic beds or (rarely) in situ dolostone; the conglomerates commonly are draped by siliciclastic sediment, although in one locality the intraclasts acted as a substrate for stromatolite growth (Fig. 5-4). Each paraconglomerate sheet is composed of a dispersed framework of angular to subround intraformational lithoclasts in a matrix of subfeldspathic quartz arenite. The lithoclasts range in size from 2-20cm; carbonate lithoclasts are composed of cross-stratified quartzose oolitic dolarenite, dololutite, silty dololutite, and dololutite with poikiloblastic quartz euhedra. Lithoclasts of siliceous composition include red mudstone, glauconitic mudstone, distinctive feldspar-phyric porphyry and green laminated "mudstone", the latter composed of euhedral green tourmaline, quartz silt, and hematite. The matrix to the lithoclasts displays a marked bimodal grain size distribution and is composed of 10-60% coarse sand grains ("millet seed" grains) and fine-grained moderately sorted feldspathic arenite. The "millet seed" sand grains range in size from coarse to very coarse sand, are extremely well-rounded and spherical (Plate E1E), and composed mostly of coarsely polycrystalline quartz and chert. In several thin sections, "millet seed" grains are composed of chalcedonic quartz typical of replaced evaporites (Plate E1F) (xenotopic quartz of Chown and Elkins, 1974).

Interpretation

The massive, matrix-supported framework of the conglomerates, angular shapes of intraformational lithoclasts, and abrupt appearance of this lithofacies as interbeds in sub-aerial sediments, all argue for relatively sudden ("catastrophic") deposition. The occurrence of conglomerate with a framework composed of abundant clasts of marine derivation, within a succession of terrestrial and coastal plain clastics (member C4; Lady Nye Formation; thin-bedded arenite-mudstone lithofacies of member E1), implies that the conglomerates are deposits of onshore-transported debris. This lithofacies is therefore inferred to record storm surge and shoal washover deposition of contemporaneously lithified marine sediments onto the near-shore surface of the Lady Nye Formation flood-plain.

As concluded by Hayes (1967) in his study of the effects of hurricanes on the Texas coast, "hurricanes can mix environment-sensitive fauna from a variety of environments into a single sedimentary deposit" (*ibid.*, p.1). In the example described here, environment-sensitive lithologies have been mixed, transported, and deposited by storms. The dolostone lithoclasts are peritidal carbonates that underwent early cementation (*cf.* Dravis, 1979) before being ripped up; this facies is similar to the quartzose oolitic dolostone lithofacies which is widespread in Dease Arm (Plate E1D, E2A). The origin of the glauconite mudchips is uncertain; although probably of marine origin, in situ source beds of the

glauconite have not been observed in the map area. The millet-seed quartz grains are inferred to have formed as a result of continuous high-energy wave reworking and aeolian abrasion in a beach/sand shoal environment (cf. Folk, 1968; Taira and Scholle, 1979). Millet-seed arenite beds were seen in only a single outcrop on Ritch Island, but this lithology probably is more widespread. The feldspar-phyric porphyry fragments resemble the porphyries that occur in pyroclastic rocks of the Narakay Volcanic Complex; if indeed they are from this source, transport of at least 100km is implied (the fragments are extremely vesicular and probably could have floated). The occurrence of dolostone lithoclasts containing poikiloblastic quartz euhedra, which probably replaced evaporites (cf. Friedman, 1980; Friedman and Shukla, 1978), and millet seed grains of evaporitic chalcedony, suggest that silicification of evaporites was an early diagenetic event.

The intraformational conglomerate lithofacies is inferred to record depositional events in which onshore-directed storm surge currents transported and mixed blocks of marine dolostone, beach facies sands, and feldspathic arenite, and deposited them on the surface of the Lady Nye system floodplain. In one locality (Fig. 5-4), marine inundation persisted after the passage of a storm, as indicated by growth of a stromatolite biostrome that can be traced for at least 10km (Fig. 5-3). In view of the lack of biostratigraphic control in most Precambrian successions, this lithofacies provides a unique inventory of contemporaneous facies that can be used to

constrain paleogeographic reconstructions (Fig. 5-7).

Intraclastic Dolostone Lithofacies

This lithofacies composes about 20% of member E1. It occurs as massive 5-30cm sheets and as lenses of interdomal and dome surface clastics associated with SH-V and LLH stromatolites, respectively. Massive sheets of angular to subangular granule to cobble-sized fragments composed of algal- and current-laminated dololomite blanket underlying beds. These sheets rest on non-erosive to gently scalloped basal surfaces and serve as substrates for stromatolite growth. The lenses of intraclasts associated with domal stromatolites occur as edgewise conglomerates restricted to recesses in the domal surfaces.

Interpretation

This facies records disruption, transport and deposition of locally-derived contemporaneous carbonate lithofacies. The generally angular shape of the intraclasts indicates only minor transport and abrasion before deposition. The lutite clasts were likely derived from desiccated ponds (cf. Hardie, 1977), whereas fragments of clastic dolostone were derived from cemented parts of the quartzose oolitic dolostone lithofacies. Sheet-like intraclastic beds were deposited by storms and flood/washover rather than surf-zone processes (cf. Hoffman, 1975). Lenses and pockets of edgewise conglomerate are intraclasts that were trapped in recesses on the domal stromatolites; the intraclasts may have been "washed on" or

the chips may be the remnants of lutite layers on the stromatolites that were desiccated and lumped into the low areas. The upright packing of these chips is indicative of wave action (cf. Ricketts and Donaldson, 1979).

Stromatolitic Dolostone Lithofacies

Stromatolitic dolostone, containing a variety of stromatolite morphotypes, is common within member El. Stromatolite beds are commonly biostromal at the scale of the outcrop, but in only a few cases can a particular unit be traced between sections (10km max.). In all cases, the stromatolite heads appear symmetric in sections parallel to bedding, and do not display any elongation. In most cases, lamination within the stromatolites consists of irregular alternations of smooth to wavy algal* laminae (mm-cm scale) and wavy to lenticular 5-10mm dololutite laminae. Chert occurs in irregular patches (5-20cm) and paper-thin layers that mimic algal lamination within the stromatolite head. Three different stromatolite morphologies occur: columnar, domal, and uncommon planar cryptalgal laminites.

Columnar stromatolites occur in beds (<50cm) composed of

*Although the term "algal" is used throughout this paper to describe organic lamination within stromatolites, it is understood that the stromatolites were likely communities of microbial organisms that were not exclusively algal (cf. Krumbein, 1983).

3

spaced to tightly-packed 10-20cm diameter heads of squat (low height-to-width ratio) SH-V and less common SH-C varieties (terminology of Logan, et al., 1964), (Plates E2E-F). Biostromes commonly overlie beds of coarse clastic material (intraformational conglomerate and intraclastic dolostone lithofacies), (Fig. 5-4; 5-6C,D). Intraclastic dolostone occurs between heads as granule-size packstone and angular edgewise conglomerate (flakestone). Columnar stromatolites also occur as 10-30cm heads that nucleate within creases and furrows in larger domal structures. Columnar stromatolites commonly coalesce upwards into LLH domal structures (cf. Fig. 5-6C) and rarely planar cryptalgal laminates.

Domal stromatolites occur in beds (<50cm) of LLH-C domes or less commonly as large domal structures up to 5m thick, 3m wide, and with synoptic relief of as much as 2m. Planar cryptalgal laminites are uncommon; they occur as massive beds with faint planar-undulose lamination, irregular fenestrae and bird's eye structures filled with pinkish dolomite spar, and small silicified evaporite rosettes. "Cauliflower" silicified evaporites are common within the stromatolites of eastern Dease Arm and Leith Line sections (Plate E6C), and occur as nodules along lamination within stromatolites.

Compound stromatolite structures occur as bioherms within crossbedded oolitic clastics and as surface coatings on the tops of coarse intraclastic blocks (Fig. 5-6B,C). Bioherms within oolitic sands (Fig. 5-3) are flat-based lens-shaped bodies (1m thick X 10m+ wide) with an overall domal shape

composed of numerous tightly packed SH-C to LLH individuals¹⁶⁶ (5-20 cm high). Bioherms within Dease Arm are also composed of tightly-packed SH-C to LLH individuals, but the bioherms are smaller (1m X 1m) structures that grew off of the upper surface of coarse intraclastic blocks (Plate E2E). These bioherms also interfinger with contemporaneous interdomal channel clastics and have bound large (up to 10cm) dolostone slabs onto the bioherm surface.

Interpretation

With exception of the planar cryptalgal laminites, the stromatolites of this lithofacies are interpreted to have grown in an intertidal to subtidal environment. This conclusion is based on the character of algal lamination (smooth to undulose non-fenestral mats), shape of the heads in comparison with modern stromatolites (cf. Hoffman, 1974, 1975; Logan, et al., 1974); and the lack of features in the stromatolites indicative of long periods of emergence and desiccation (cf. Playford and Cockbain, 1976). As will be discussed later, the nodular evaporites are the result of precipitation from early diagenetic brines rather than surficial desiccation, and thus reflect the character of the paleoaquifer and/or hydrologic cycle rather than amount and duration of exposure. Maximum synoptic relief on domal stromatolites is 2m which is considered an average water depth. The common occurrence of columnar stromatolites suggests growth around a headland promontory subjected to moderate wave and tidal action.

(cf. Hoffman, 1976); the vertical change to LLH-type stromatolites is considered to indicate shoaling. Although much of the lamination in the stromatolites is biogenic in origin, the lutite layers attest to periods of more rapid deposition of suspended material, perhaps following storms (cf. Ginsburg and Hardie, 1975; Park, 1976). Recurrent periods of high-energy conditions are recorded by onlap and burial of stromatolite bioherms by coarse interdomal clastics and emplacement of slabs of dolostone on the bioherm surfaces.

Facies Associations, Cyclicity, and Paleogeography

In this section, vertical sequences and lateral associations between lithofacies are examined and used to infer a model for regional paleogeography and depositional environments of the member E1 depositional system.

Vertical Cyclicity

Examination of Figure 5-3, a roughly north-south cross-section of member E1 from East River to Falcon Lake, shows that member E1 is characterized by poorly developed vertical cyclicity. Of the several types of cycles recognized, none could be traced regionally (compare with carbonate units within the Rocknest Formation of Coronation Supergroup in which individual cycles can be traced for more than 200 km; Hoffman, 1973; Grotzinger and Hoffman, 1983). Reasons for the apparent lack of cyclicity within member E1 are: a) complex facies mosaic such that lateral migration of each facies does not produce distinct basinwide cycles, b) storm truncation of

FIGURE 5-6: Detailed section of an ideal siliciclastic-to-carbonate cycle in member E1, Narakay Islands. Sub- and inter- at top of column on far left refer to inferred position in tidal environment.

facies (cf. Radke, 1980), c) slow rate of transgression such that sedimentation more or less kept pace with subsidence and sea-level rise, d) non-depositional facies that may make cyclicity difficult to recognize, and e) presence of barriers that may cause juxtaposition of different energy facies (cf. Irwin, 1965). It is inferred that the first three processes were most influential in determining the character of member E1.

Although member E1 is typified by acyclic arrangement of lithofacies, several motifs of cyclic sedimentation are locally present: 1) siliciclastic to carbonate cycles, 2) storm-shoal cycles, and 3) stromatolite cycles.

Siliciclastic to Carbonate Cycles

These cycles are thickest and best developed in the Narakay Islands where they are 4-8m thick and display the vertical succession of: a) thin-bedded arenite-mudstone, b) quartzose oolitic dolostone, c) intraclastic dolostone, and d) stromatolitic dolostone (Fig. 5-5; Plate E1A-D, E2E). The complete succession of these lithofacies is only rarely developed and superposition/amalgamation of the same lithofacies is common. These cycles are interpreted to record the migration of siliciclastic tidal flat/shoal regions and intershoal depressions/channels. The poorly developed cyclicity and lack of lateral persistence of cycles suggests that they reflect

autocyclic migration of laterally equivalent facies rather than externally controlled (allocyclic) cyclicity (i.e. tectonic control).

The broad tidal channels described from the Persian Gulf are up to 10m deep and "several" kilometers wide (Loreau and Purser, 1973), dimensions that would make it difficult to recognize these structures except by detailed lithostratigraphic analysis. Although Persian Gulf channels and associated ooid facies and algal-coral bioherms are perhaps an attractive modern analogue for the quartzose oolitic dolostone and stromatolitic dolostone lithofacies, respectively, the shape and size of Persian Gulf channels is controlled by the position of reefal headlands which in turn reflects the site of Pleistocene topographic highs, an unlikely situation in member El. The quartzose oolitic dolostone lithofacies contains stratification that is similar to that found in modern intertidal siliciclastic shoals of the Wash and Oosterschelde estuary, in Britain and the Netherlands, respectively (McCave and Geiser, 1979; van den Berg, 1977). By analogy with these areas and modern Bahamian ooid shoals (Ball, 1969) stratification in the quartzose oolitic dolostone lithofacies would have been generated by migration of lower flow regime bedforms that migrated in response to tidal/wave-generated (?) currents. The presence of low-angle to horizontal stratification and sequences of stratification similar to that produced by ridge accretion (cf. Van den Berg, 1977; Fox and Davis, 1972) imply local areas of wave shoaling over a sandflat/bar. Early

cementation of oolitic sands is indicated by the presence of large rip-up blocks analogous to cemented sands of modern Bahamian ooid sands (cf. Dravis, 1979; Harris, 1979). Disruption of cemented oolitic sands, blocks of which became the site of stromatolite growth during quiet water periods, may have occurred during periods of abnormal current surge through channels and intrabar lows, perhaps related to storm activity.

Storm-Shoal Cycles

Although not abundant (<15%), storm deposits are an important component of member E1 that are marked by the occurrence of the intraformational conglomerate lithofacies and some beds of the intraclastic dolostone lithofacies. The former lithofacies is designated as "allochthonous" storm deposits because they represent marine deposits that have been transported considerable distances and deposited on a contemporaneous coastal-fluvial floodplain. Minimum distance of transport is inferred from correlation of transported and in situ lithoclast lithologies between sections 1 and 2 (10km) (Fig. 5-3). Maximum distances are inferred on the basis of the presence of clasts of Narakay Island volcanic porphyry (70km) and millet-seed sand grains; the only in situ outcrop of this latter lithology is on Ritch Island (50km west of East River). These figures probably are in excess of the actual distance of transport by a single storm; some of the Narakay porphyry is very vesicular, and fragments may have floated a considerable distance; the apparent restricted occurrence of millet-seed

sand grains is a reflection of the almost complete lack of outcrop of the East River Formation between East River and Ritch Island.

Allochthonous storm deposits are inferred to be the result of hurricane-generated onshore storm surge transportation. The size of the clasts (up to 8cm), massive texture of the conglomerates, and apparent long distance of transport of lithoclasts and millet seed sand grains attest to the intensity of storms and the probable low slope to the coastal zone (cf. Singh, 1980). Most of the beds of intraformational conglomerate lithofacies are overlain by floodplain arenite and/or mudstone, but in one example (Fig. 5-4), stromatolites grew on the clasts. This is interpreted to be the result of local flooding due to the storm destruction of a barrier, rather than a gradual inundation as a result of transgression (cf. Hayes, 1967).

Massive intraclastic dolostone beds are "autochthonous" storm deposits of marine-derived and deposited lithoclasts. In all occurrences of massive sheets of this lithofacies the clasts are angular and unsorted, which implies rapid deposition and a short residence time in active currents. These intraclastic beds almost invariably are overlain by stromatolites (Fig. 5-6C,D), suggesting rapid colonization and stabilization of intraclastic pavement by algal mats. Similar storm washover and flood deposits of intraclasts, overlain by algal mats, are described from the Bahamas (Hardie, 1977). The upright orientation of intraclasts in some beds (stone rosettes,

edgewise conglomerate) implies at least some wave action and clast packing (cf. Ricketts and Donaldson, 1979).

Stromatolite Cycles

Stromatolite cycles consist of various combinations of LLH and SH growth modes and uncommon flat mats (Fig. 5-6B-D; Plate E2F) that record either an upward-shoaling or upward-deepening of the depositional surface. Shoaling cycles comprise SH-C and SH-V stromatolites, commonly developed on an intraclastic substrate, that grade into overlying LLH-C stromatolites. In modern stromatolitic tidal flats this change in morphology is the result of shoaling during progradation of an intertidal to supratidal environment over an agitated subtidal to intertidal environment (cf. Logan, et al., 1974). Deepening ("drowning") cycles are recorded by alternations of LLH followed by SH mode stromatolites although these cycles could also reflect a change in the degree of agitation and rate of sedimentation.

Macroenvironmental Model and Paleogeography

The overall lithologic change in member E1 is an upwards decrease in the ratio of siliciclastic to carbonate beds. This is interpreted to be a reflection of long-term decrease in the rate of delivery of siliciclastic detritus to the basin, contemporaneous with marine transgression.

A major problem with paleogeographic reconstruction of Precambrian sedimentary rocks is the lack of biostratigraphic control, making it difficult to recognize isochronous facies.

FIGURE 5-7A: Schematic paleogeographic block diagram for deposition of upper Lady Nye Formation.

FIGURE 5-7B: Paleogeographic diagram for deposition of middle part of member El. Letters in parentheses refer to lithofacies described in the text. OOD: quartzose oolitic dolostone lithofacies; TBAM: thin-bedded arenite-mudstone lithofacies; SD: stromatolitic dolostone lithofacies.

This problem has been partly circumvented in member E1 by assuming that lithoclasts in intraformational conglomerate storm deposits are samples of isochronous facies.

Two models of isochronous paleogeography are inferred for the lower and upper parts of member E1 (Fig. 5-7A and B). A 60km gap in exposure between East River and Ritch Island limits the precision of the reconstructions. The orientation of the strandline is difficult to ascertain because the strike of the outcrop belt imparts an east-west bias to facies changes. In addition, paleocurrents from marine facies of upper Lady Nye Formation and member E1 of East River Formation are only of limited use because of the typically variable orientation of currents on epicontinental marine shelves (cf. Nelson, et al., 1982; Hayes, 1980). In these settings the combined effects of wave refraction, wind direction, tidal currents, interaction and deflection of currents with bedforms and shoreline configuration cause currents to bear little relationship to strandline orientation. A north-northeast striking, northwest-facing shoreline, inherited from the depositional slope recorded in upper Lady Nye system, is proposed and is consistent with most of the data.

Figure 5-7A is a reconstruction of inferred paleogeography during transgressive marine sedimentation in Lady Nye Formation, but is assumed to be applicable to the transition from Lady Nye Formation into East River Formation deposition. A low-energy fluvial floodplain (lithofacies A, member C4) extended from Falcon Lake south to East River, and possibly

down to Blood Creek. Floodplain clastics graded to the west (seaward) into the marine tidal flat/delta plain clastics of lithofacies B (member C4, Lady Nye Formation). Parts of the sand shoal/tidal flat complexes (channels and passes between sand flats) became the sites of ooid generation and growth of stromatolite bioherms (siliciclastic to carbonate cycles of member E1). The inferred restriction of carbonate lithologies to broad channels is in contrast to the common observation that siliciclastic shoals and sand bodies act as suitable areas for the generation of ooids and stromatolites (cf. Dutton, 1982; Ginsburg et al., 1982). In the region from east Dease Arm to Leith Line, there is a 60km gap in data; oolitic and evaporitic dolostone deposition is inferred from the occurrence of these lithologies in intraformational conglomerate deposits. Storm surge and waves transported these deposits unimpeded for as much as 50km.

Continued transgression and a general decrease in the supply of siliciclastic detritus resulted in the paleogeographic model depicted in Figure 5-7B. At this point in time a fluvial wedge of medium-grained quartz arenite had prograded onto the high-energy lagoonal carbonate tidal flats of upper member E1. The fluvial material was redistributed by tidal and refracted wave currents (wind-forced currents are also a possibility, cf. Morton, 1982). Some of the material was deposited as relatively pure quartz arenite, whereas many of the sand grains became nuclei for superficial ooids and were admixed with carbonate ooids. The ooids formed on tidal bars

and in channels that fed ooid deltas (cf. Loreau and Purser, 1973). Stromatolite bioherms developed on the inactive parts of ooid sand bars and between ooid ~~delta~~ lobes. To the west, ooid shoals graded into a complex facies mosaic of broad shallow channels, interchannel banks, islands, and emergent headlands. Some of the interchannel banks were siliciclastic sand-mud tidal flats; others were regions of stromatolite colonization and washover deposition of intraclastic dolostone.

The acyclic stacking of member E1 lithofacies is attributed mostly to the complex mosaic of non-linear facies belts (that would leave an irregular record of facies migration), storm truncation, and a slow rate of subsidence (equal to the rate of sedimentation). Alternatively, de Raaf and Boersma (1971) have pointed out that shallow tidal deposits are typically acyclic.

Member E2

Member E2 comprises interbeds of distinctive biohermal to biostromal stromatolites and oolitic to quartzose dolostone and subordinate calcareous quartz arenite. The unit is widespread, extending from Leith Peninsula north to Dismal Lakes throughout an area of at least 30,000 km². The base of the member is marked by the first occurrence of large (>1m diameter) microdigitate compound domal stromatolites. The top of the member is marked by the last occurrence of large domal stromatolites that have either a microdigitate or smooth

texture to the laminations. The member thickens towards the west from 20 to more than 80m (Fig. 5-2). The latter thickness is considered a minimum, because this measurement was obtained in the Narakay Islands where at least 20m of synsedimentary erosion of this member has occurred. The member has gradational contacts with both underlying and overlying members. In the upper part of the member, evaporites are present in the form of silicified nodules (<1cm) and as irregular layers that probably are silicified crystal mush.

Member E2 displays several vertical changes in lithologic character as it grades into member E3:

- a) A decrease in siliciclastic detritus concomitant with an increase in ooid content.
- b) A distinct upward-thinning trend in the Narakay Islands; this trend is less distinct in sections to the east.
- c) The stromatolites, which form large complex domes in the lower part of the member, decrease in internal complexity and abundance of microdigitate structure.

Stromatolite Lithofacies

This lithofacies is characterized by silicification of the stromatolites that locally obliterates all primary textures; in these cases the stromatolites occur as masses of vari-coloured (green, white, tan, gray, white, pink) chert that has a hackly fracture and faint stromatolitic lamination. Disseminated copper and manganese mineralization is associated with some stromatolite beds and was observed in nearly all

Figure 5-8. Examples of stromatolite and clastic successions in member E2. A. 1) Massive to hummocky cross-stratified calcareous arenite, 2) dololite layers in channel, 3) massive (due to recrystallization) dolostone, 4) bioherm of microdigitate stromatolites. Note undercutting of channel walls beneath stromatolite, discontinuities in stromatolite growth, and beveled top of bioherm (Narakay Islands). B. 1) Bioherms and biostrones (at scale 0- outcrop) of microdigitate stromatolites, 2) parallel-laminated to massive () oolitic dolostone, 3) coarse intraclastic-oolitic dolostone, 4) storm deposit of intraclast-bearing massive calcareous arenite overlain by hummocky cross-stratified calcareous arenite. Note compound character of biostrones (East River, 86K). C. Storm sequences in Narakay Islands. 1) Massive intraclast-bearing calcareous arenite, 2) hummocky cross-stratified calcareous arenite with detached pillows of contorted laminae (4); 3) parallel-laminated dololite, 5) microdigitate stromatolite crusts. Vertical arrows designate individual storm deposits (Narakay Islands).

exposures of this lithofacies east of the Narakay Islands.

Stromatolites occur as compound domal structures (domes composed of more than one type of stromatolite morphology) up to 5m wide and having as much as 1-2m of synoptic relief (Plate E3A). The largest stromatolites (5m in diameter) occur as LLH biostromes; they have an irregular distribution within the map area. Giant domal stromatolites occur in western Dease Arm (Prospect Island and unnamed island north of Narakay Islands), south of Teshierpi Mountain, along East River (86K) (Plate E3B) and on the Superstition Islands in Great Bear Lake. In plan view the stromatolites range from symmetric circular structures to irregular lobate "kidney bean" shapes; marked elongation was noted in only a single example in Dease Arm (northwest-southeast elongation). At the scale of individual outcrops, stromatolites occur as both bioherms and biostromes. In some outcrops, bioherms were observed to coalesce upwards to form biostromes (Plate E3A; Fig. 5-8). Outcrop scale biostromes cannot be correlated between measured sections, indicating that on a regional scale, the stromatolites are discontinuous.

Bioherms and biostromes invariably rest on scoured surfaces (Plate E3A, Fig. 5-8); these surfaces are commonly steep with up to 60cm relief. In some places the steep character of the scour surface has been enhanced by early substrate stabilization by stromatolites that allowed for increased undercutting of unprotected sediment (Fig. 5-8). The stromatolites occur as relatively thick (>1m) units or as thin (<20cm) crusts on

scour surfaces. Many of the stromatolites are composed of two or more growth phases separated by a discontinuity (Fig. 5-8). Stromatolite beds are draped by overlying beds of clastic detritus.

The stromatolites are composed of layers of morphologically different algal mat textures that occur in irregular vertical sequences. Microcolumnar to digitate (.5-2cm diameter) SH-C crusts, up to 10cm thick, are the most common form. Inter-column areas commonly contain micrograded beds of fine-grained clastic material (dolomite, quartz and feldspar silt, ooids; Plate E3D). Clastic grains can also constitute single-grain thick laminae within algal structures that attest to agglutination and entrapment of clastic debris by the stromatolites. Lateral linkage of columns by a continuous mat may occur several times within a layer (Plate E3D). Algal mats are composed of smooth to colloform laminae. Microdigitate layers may grade upwards into larger (5cm diameter) tight-packed SH-C stromatolites. The different stromatolite layers do not occur in any distinctly regular, vertical sequence, but rather as irregular alternations of smooth and colloform mats, microcolumnar to digitate layers, and larger SH-C stromatolite layers. Although silicification of the stromatolites is ubiquitous, only the algal layering is preserved; filamentous or other cellular textures have not been observed.

Clastic Lithofacies

Sheets of clastic debris occur as solitary beds or stacked cosets that separate stromatolite layers (Fig. 5-8B-C). This lithofacies ranges in composition from fine-grained calcareous feldspathic arenite, with admixed oolites and intraclasts, to ooid grainstones (Plate E3E). Coarse intraclastic dolorudites also occur, but are rare (Fig. 5-8B). Siliciclastic material is most common in the lower part of the member; it is absent from the upper part of the member and is also entirely absent from member E2 east of Leith Line. Dolostone beds commonly occur as dark brown structureless units in which an oolitic texture is inferred on the basis of the preservation of oolites in chert nodules, the presence of cross-stratification, and, in thin section, recognition of progressive neomorphism of ooids to coarse spar (Plate E3E).

The clastic lithofacies occurs in either solitary or stacked tabular to undulose sheets, generally 10-30cm thick (Plate E3F, Fig. 5-8), that drape the underlying stromatolites. Stacked multistory sheets may reach an aggregate thickness of up to 9m, but this is rare. Internally, the clastic sheets display a variety of cross-stratification, although hummocky cross-stratification is most common. In the Narakay Islands, the clastic beds are compound cyclic cosets composed of an ideal sequence of: a) 2-3 sets of planar to hummocky cross-stratification overlain by b) planar, horizontal to low-angle ($<10^0$) stratification, ripple cross-stratification, rare regressive ripples (cf. de Raaf and Boersma, 1971) and c) type B

climbing ripple stratification (Plate E3F). Stratiform synsedimentary deformation of entire beds into convolute folds (Fig. 5-8C) is common in the upper parts of cycles or, in rare examples, the entire cycle may be deformed. Of note is the occurrence of a particularly distinctive bed of this lithofacies within the Narakay Islands. This unit is 1.2m thick, can be traced throughout the islands, and contains a greater proportion of siliciclastic grains than either underlying or overlying beds. The sequence of structures is shown in Figure 5-8.

Interpretation

The regularly interbedded character of the clastic and stromatolitic lithofacies is interpreted to reflect alternating periods of storm and fairweather deposition, respectively. The lack of sedimentary structures indicative of emergence and desiccation imply a subtidal to lower intertidal setting.

The clastic units are interpreted as storm deposits on the basis of: a) the common occurrence of hummocky cross-stratification, b) vertical sequences of sedimentary structures indicative of decreased flow competency, c) the occurrence of soft sediment convolute folds, presumably formed by dewatering of rapidly deposited, fluid-rich sediment, and d) the manner in which the clastic units drape the stromatolites. In general the stromatolite lithofacies formed during fairweather conditions, although several features are suggestive of periods of high-energy turbulence such as: a) domes with two or more phases of growth separated by an erosional discontinuity

(Fig. 5-8A), b) the common occurrence of normally-graded clastic layers in interdomal channels and clastic laminae within stromatolites (indicative of turbid water) (Plate E2D), and c) deep undercutting of stromatolite-stabilized pedestals. A subtidal agitated environment has been proposed for morphologically similar stromatolites from the Precambrian Porsanger Dolomite (Tucker, 1977) and sublittoral bights in Shark Bay (Hoffman, 1976, p.269). In addition, the morphology of the algal mats is consistent with a subtidal setting (cf. Logan, et al., 1974).

Member E2 is inferred to represent more or less isochronous submergence of the East River platform at the time of maximum marine transgression. The upward changes in the character of member E2 and the appearance of member E3 peritidal facies record gradual shoaling and offlap. The source of the oolitic material in the clastic beds is a problem; are they autochthonous grains that were suspended and redeposited by storms or are they allochthonous and derived from ooid shoals elsewhere on the shelf? The sedimentary structures within the arenaceous beds (hummocky cross-stratification, etc.) are indicative of offshore transport during storm surge ebb (cf. Dott and Bougeois, 1982). This implies that extensive ooid shoals must have existed to the east. An ideal depositional cycle would be: a) wave suspension of ooids and siliciclastic material during storms, b) offshore-directed transport and oscillatory suspension deposition by low-concentration turbidite-like storm surge ebb, c) dewatering of fluid-rich sediment

d) scour of the depositional surface by waning storm currents and/or high-energy background currents, and e) stromatolite colonization and sediment stabilization interrupted by periodic agitation (not as intense as storms). Decreased energy with time is indicated by decreased bed thickness of the clastic lithofacies, and change from predominantly microdigitate stromatolites to smooth domal structures.

Member E3

This member is composed of interbedded stromatolitic and intraclastic to oolitic dolostone, dololutite, and siliceous mudstone (Fig. 5-9). It is at least 20m thick in the east, thickens to more than 80m at East River and then thins to less than 40m in the Narakay Islands (Fig. 5-2). It gradationally overlies member E2; the contact with member E4 is marked by the first occurrence of granules and pebbles of chert. In the Falcon Lake section, member E4 is absent and member E3 is directly overlain by red mudstones of Kaertok Formation. In the Narakay Islands, member E3 is lithologically distinct from sediments to the east; inclusion of these lithologies within member E3 is based on the position of these units above member E2 and below Kaertok Formation correlatives.

Stromatolitic Lithofacies

Stromatolitic dolostone is an upward decreasing component of member E3. The stromatolites are dominantly of LLH and SH mode, and form biostromes up to 5m thick. Stromatolites

FIGURE 5-9: Detailed measured section through part of member E3 east of Lac LeRoux. See Table 1 for key to lithologic symbols. Segments of columns are continuous from left to right.

also occur as 2-5cm drapes of smooth mats over irregular scour surfaces in underlying clastics, and as 2-10cm beds with flat lamination. Biostromes commonly overlie oolitic/intraclastic dolostone and are abruptly overlain either by this lithofacies or siliceous red to green mudstone (Fig. 5-9). Biostromes may display either single mat growth mode or cyclic alternations between LLH and SH modes (Fig. 5-9, 10). In addition to its occurrence as a stromatolite substrate, oolitic and intraclastic debris also occurs as channel-fill lenses, pockets of edgewise conglomerate (Plate E4C), and thin sheets washed over from adjacent channels (Fig. 5-10). SH-V to SH-C mode stromatolites vary from 10-100cm wide (even within a single biostrome) and have a circular, symmetric shape in bedding plane sections; they commonly coalesce upwards into LLH stromatolites. Many of the larger SH-V stromatolites have a squat "mushroom cap" shape (a low width to height ratio; Plate E4A). LLH stromatolites form sheets of low-relief (<20cm synoptic relief) heads, with 5-50cm between heads (Plate E4B), that commonly grade upwards into nearly flat, undulose mats. Some LLH stromatolites display an upward increase in tendency towards SH-V shapes. In the Narake Islands, biostromes comprise low-relief (<5cm) bioherms and crusts, with microdigitate to colloform lamination, that rest on scour surfaces cut into underlying clastic beds (Plate E5A).

Lamination within the stromatolites is composed of flat to undulose, pustulose to smooth algal mats, tufted palisade-like mats, dololutite, and thin (<1cm) layers of peloids and fine-

Figure 5-10. Outcrop sketch of relationships between stromatolitic and clastic lithofacies in member E3, (Falcon Lake). 1) Dololuite, 2) LLH, SH-V, SH-C stromatolites, 3) oolitic-intra-clastic dolostone. Note coalescence of SH-V and SH-C stromatolites into LLH forms. Lowermost oolitic layer is a washover deposit that infilled interdunal channel and buried stromatolites. Base of successive channel/washover sequence is marked by dashed line.

grained oolites. Smooth algal lamination has a distinct tan colour in outcrop and is common in SH and LLH stromatolites. Crude palisade structures (cf. Hardie, 1977) have a distinct "sooty-gray" color and are composed of alternations of crinkly algal mats, tufted algal "bushes", dololutite, and irregular laminoid fenestrae (Plate E4E). This style of lamination is restricted to low-relief LLH domes and also is associated with teepee structures (cf. Kendall and Assereto, 1977). Chert occurs as thin ribs and irregular patches within stromatolites (Plate E4A); in stromatolites with smooth mat textures the chert is tan to white, whereas in palisade-textured mats it is gray to black. Although most of the chert is replacive, several occurrences were noted, particularly in the palisade mats, of pore-fill isopachous botryoids of length-slow chalcedony.

Intraclastic-Oolitic Lithofacies

This lithofacies occurs in 10-50cm beds composed of various proportions of ooids, rounded to angular intraclasts, and oncolites. The beds are massive to rarely cross-stratified (tabular-planar sets) sheets that drape stromatolites. Intraclastic dolostone composed exclusively of angular lithoclasts (dololutite and rare oolitic packstone) occurs as interdomal clastics and as edgewise conglomerate between domes (Plate E4C). Oolitic packstone (Plate E4D) and poorly-sorted mixed oolitic-intraclastic packstone occur as massive sheets that locally contain oncolites, up to 6cm in diameter (Fig. 5-9), are locally present in these beds. In the Narakay Islands, the

clastic lithofacies associated with the stromatolites occur in 10-30cm thick cosets with flat to micro-hummocky cross-stratification. Massive sheets of oolites, up to 50 cm thick, also are present.

Nodular silicification is common in this lithofacies in exposures east of the Narakay Islands. Chert occurs as irregular-shaped nodules and patches of multicoloured pink to tan to black material of both replacement and pore-fill origin. In some cases individual silicified ooid grains are contained within an unsilicified dolosparite (neomorphic texture) matrix. In other nodules the oolites have been only partly silicified, but the carbonate matrix has been entirely dissolved and the pore space infilled by: a) isopachous length-slow chalcedony, b) botryoidal length-slow chalcedony, and c) equant megaquartz (Plate E4D).

Mudstone Lithofacies

Two types of mudstone are present within member E3. Finely-laminated to massive dololutite occurs as 2-5cm thick layers at the top of clastic sheets (Fig. 5-9). Green to red siliceous mudstone occurs as thin (<5cm) films and in beds up to 1m thick. The thin films drape ooid-intraclastic layers and stromatolites. Mudcracks are present but sparse within the siliceous mudstone.

Interpretation

Member E3 east of Narakay Islands bears a strong

resemblance to tidal flat facies as described for Shark Bay, Australia, and particularly the Three Creeks region of the Bahamas (Logan, et al., 1974; Hardie, 1977, respectively). SH-V and SH-C stromatolites mark the sites of tidal passes and channels along an intertidal headland (cf. Davies, 1970; Logan, et al., 1974); upward coalescence into LLH stromatolites reflects decreased current activity and shoaling (Fig. 5-10). The smooth lamination typical of many of these stromatolites is indicative of infrequency or absence of exposure. In contrast, the palisade-textured mats probably accumulated in supratidal algal marshes (cf. Hardie, 1977) by alternation of flood deposition during abnormal high tides and/or storms of dololomite and peloid/ooid layers and algal growth. Massive ooid packstones most probably are storm or levee(?) washover deposits of ooids and intraclasts derived from contemporaneous shoals. The angular intraclastic dolostones are locally derived deposits of mudchips stacked on algal marshes during flooding and wave agitation. Siliceous mudstones are interpreted as terrestrially derived material deposited during flood runoff that inundated the carbonate tidal flats. An analogous situation was observed by the author after heavy rains left a thick layer (5cm) of mixed siliclastic and organic material in a back-barrier hypersaline lagoon in Baja California (Laguna Figueroa, cf. Margulis, 1982; Horodyski, 1977). In summary, member E3 east of the Narakay Islands consists of cyclic peritidal carbonates deposited in an intertidal to supratidal setting.

The Narakay Islands contrasts with this proposed paleo-environment in that it is characterized by current-laminated clastics and colloform stromatolites typical of lower intertidal to subtidal environments. In particular, the micro-hummocky cross-stratification implies an agitated, wave-influenced environment. The nature of the transition between the two sub-environments of member E3 is unknown (the closest mainland outcrop of member E3 is 25km east of the Narakay Islands).

Member E4

Member E4 has only been observed in outcrops east of Dease Arm, specifically from East River northeast to Dismal Lakes; it is absent from Falcon Lake and the Narakay Islands. The member is inferred to be wedge-shaped, thickening from northwest to southeast and pinching out to the west (Fig. 5-2). In the Falcon Lake section, the contact between member E3 and Kaertok Formation is sharp (disconformable), although evidence of erosion or a significant time break (such as caliche and/or karstic dissolution of member E3 below the contact) is lacking. This indicates that member E3 in this region is time-equivalent to member E4 to the southeast. The largest chert clasts are in sections adjacent to Teshierpi Fault, and decreases to the west. This is a reflection of the influence of the Teshierpi Fault on the depositional geometry of this member.

The member is composed mostly of intraclastic and oolitic

packstones and rudstones, stromatolitic dolostone, and siliceous mudstone. Carbonate lithologies decrease in abundance upwards in the member as the abundance of chert and quartz grains increases. The chert grains are characteristic of this member, and distinguish member E4 intraclastic beds from those of member E3; the contact between these two members is vertically gradational. The contact with overlying Kaertok Formation is also gradational, and is defined as the first occurrence of arenite beds composed of greater than 50% quartz grains.

Stromatolitic dolostone composes approximately 20% of the member and contains stromatolites that are similar to member E3 stromatolites. LLH and SH-V are the most common varieties; they typically contain smooth algal laminae intercalated with thin dololutite and/or peloidal/oolitic clastic layers. Stromatolites also occur as thin (2-5cm) drapes on erosional surfaces between intraclastic beds. The stromatolites overlie massive to rarely cross-laminated intraclastic packstone and grainstone. Many of the stromatolites have grown on oncolites that occur at the top of the underlying clastic bed.

Clastic units, 20-50cm thick, are composed of ooids (mostly superficial ooids developed on nuclei of rounded intraclasts and less commonly quartz grains), rounded intraclasts (dololutite and oolitic/intraclastic dolostone), blocky to slabby clasts of cream-coloured dololutite (Plate E5D), rounded clasts of chert (silicified dolostone), and rare oncolites. Cross-stratified grainstones (<5% of the member) consist of well-sorted granule- to coarse sand-sized detritus that

displays small-scale (<10cm) tabular-planar cross-stratification. Crossbeds occur as solitary sets on top of massive intraclastic beds or as 3-4 stacked sets; herringbone stratification was observed in several outcrops. Red to green mudstones (10-300cm) are poorly exposed (Plate E5C). The thinner beds occur as drapes on dolostone beds; thicker beds contain 2-10cm thick intercalations of massive chert-dolostone granulestone.

Coarse calcareous chert-pebble conglomerates and granulestones near Teshierpi Fault occur as 10-20cm beds that commonly display a massive texture, but also display normal or inverse grading, cross-lamination, and wave ripples on the tops of beds. Red mudstones in these outcrops commonly are mudcracked, and in one example contain casts of halite crystals.

The composition and inferred provenance of the chert pebbles and granules is of particular interest. These grains first appear as sparse (<2%) granules in intraclastic packstone. Upwards within the member the grain size and relative percentage of these clasts increases. The chert grains are white, tan, red, and rarely green; commonly these colours (especially the red and white) occur in concentric rings parallel to the grain boundaries (Plate E5E). The grains are composed of silicified member E4 and E3 lithologies such as dololutite, oolites, and evaporites. Rare glauconite peloids occur in the granulestones and conglomerates near Teshierpi Fault.

Paleocurrent measurements from this member are sparse and taken from small-scale cross-stratification. Measurements (20) at two separate stations south of Dismal Lakes both give polymodal patterns with a strong mode to the north.

Interpretation

Member E4 comprises peritidal to terrestrial(?) deposits that contain increasing quantities of clastic debris derived by the erosion of members E3 and E4 (Fig. 5-11). The wedge-like shape of the member (Fig. 5-2) reflects a fan-like depositional geometry and a component of crustal tilt towards the southeast and southwest(?). These features suggest that the appearance, composition, and depositional pattern of this member was strongly influenced by syndepositional movement along the Teshierpi Fault and suggests that member E4 marks the beginning of normal(?) fault movement along the Teshierpi Fault.

The stromatolite beds in member E4 are indicative of an infrequently emergent intertidal to subtidal(?) environment. The typically massive texture of the intraclastic beds implies sudden deposition (by storms and washover flooding) and minor reworking by marine currents. The rounded character of most intraclasts and the presence of oolites suggests derivation of this material from a continuously-agitated marine environment. Angular blocks and slabs of dololomite were derived by erosion of nearby desiccated carbonate muds. The

Figure 5-11. Sketch of inferred paleogeography, looking northeast along the trace of the Teshierpi Fault, during deposition of member E4. Syndepositional movement along Teshierpi Fault was hinged (scissor fault) with northeast-end up, as indicated by arrows. Arrows on depositional surface signify directions of both marine and fluvial transport. Question marks indicate that the original extent of member E4 southeast of Teshierpi Fault is uncertain.

chert pebbles were derived from member E3; less resistant carbonate matrix to the diagenetic chert nodules was eroded during weathering and transport, and the more durable chert fragments were thus concentrated. Some of the chert grains may have been derived from parts of member E3 that were silicified during subaerial exposure (ie. silcretes). This is suggested mostly by the concentric colour banding and chert "skins" on dololomite clasts which resemble those developed on chert and limestone fragments, respectively, in the American southwest (northwestern New Mexico, personal observation). The polymodal paleocurrents and presence of glauconite within the chert pebble conglomerates indicates that they were deposited in a marine environment. Erosion of the chert pebbles may have occurred as a consequence of wave attack along a fault scarp and/or fluvial erosion and later delivery to a marine environment. The red mudstones are interpreted to be terrestrially derived flood deposits.

Evaporites

Chert nodules, geodes, and crystal pseudomorphs, all interpreted as silicified evaporites, are widespread throughout the East River Formation. They are particularly common in member E1, but also occur in the other members. Three types have been recognized: 1) "Cauliflower" geodes and chert nodules, 2) Quartz euhedra, and 3) Massive podiform to nodular cherts. The evaporitic origin and descriptive terms used are based on the work of Chown and Elkins (1974), Folk and Pitman (1971), and

Milliken (1979). Similar evaporites have been reported from Precambrian dolostones of the MacArthur Basin, Australia (Walker, et al., 1977; Muir, et al., 1981) Fennoscandia (Siedlecka, 1976; Tucker, 1976), and the Belt Supergroup of Montana, (Horodyski, 1976).

"Cauliflower" Geodes and Chert Nodules

This type of evaporite replacement is most common in member E1 and less common in member E2. It occurs as solitary to grouped nodules that range in size from 1 to 10cm (Plate L3B, Plate E2D, E6A-C). Nodules are commonly concentrated along stromatolite and/or cross-lamination and enclosing lamination wraps around the nodules (Plate E6C). The nodules are either geodes, with a crust of centripetal drusy quartz crystals in association with pyrite and dolomite crystals, or completely solid where the quartz crystals have grown together. The outer rind of the nodules is composed of white to banded pink chalcedony (Plate E6B). The exterior surface of the nodules is most commonly "lumpy" and convolute (and resembles the surface of a cauliflower), but in several well-preserved examples platy crystals display a rosette texture (Plate E6A).

Petrographically these nodules display a zonation in silica types and textures that is similar to that described by Milliken (1979) (Plate E6B, D,E). The outermost rind is composed of ragged interlocking crystal pseudomorphs composed of lutecite and length-slow chalcedony (Plate E6D). This material has a distinct "graphic" texture, referred to as xenotopic.

quartz by Chown and Elkins (1974), and has been recognized in millet-seed quartz grains of member E1 (Plate E1F). This grades into an inner zone of zoned spherulitic quartzine and spectral flamboyant megaquartz (Plate E6E). This is the layer that imparts the pink colour to the nodule rims and commonly displays concentric inclusion bands around spherical to ovoid to euhedral-hexagonal cores (Plate E6E). Inwards of this zone is a central layer of megaquartz, locally with a core of inclusions and numerous scattered "bright" inclusions (possibly anhydrite). The megaquartz displays straight to strip-like, banded extinction.

Euhedral Quartz

Euhedral poikilitic quartz crystals occur in oolitic dolostones of members E2 and E3 (Plate E6F). These crystals are essentially identical to euhedral quartz described by Friedman and Shukla (1980). They occur either as scattered single crystals or as a loosely-connected meshwork of randomly oriented crystals concentrated parallel to bedding.

Massive Podiform to Nodular Cherts

This type of chert occurs commonly in the intertidal facies of member E3, and less commonly in the lower parts of member E4. It occurs either as 10-15cm thick massive beds of dense white chert or as nodules and pods within the core and marginal regions, respectively, of teepee structures. These cherts are composed of spherulitic spectral flamboyant

and euhedral megaquartz. In addition, bladed ghost crystals are common in these cherts, as can be seen when viewed in plane-polarized light.

Interpretation

Type 1 cherts (cauliflower geodes and chert nodules) are silicified anhydrite nodules formed by the early diagenetic growth of gypsum (cf. Butler, 1969, for a modern example from the Trucial Coast, Persian Gulf). Poikilitic euhedral quartz (type 2) commonly has been interpreted to be the result of evaporite replacement (Friedman and Shukla, 1980). The massive to nodular cherts (type 3) are interpreted as evaporitic on the basis of the relic textures of bladed crystals and the microscopic character of the chert.

Presumably all of these evaporite occurrences imply a high rate of evaporation and concentration of SO_4 in the groundwater. Because this is essentially a diagenetic process, the presence of evaporites may not necessarily imply an arid climate.

In type 1 evaporites, the sulfate mineral was dissolved contemporaneously with silica replacement. The rate of dissolution of the sulfate was greater than the rate of silica precipitation, and hence a geode formed. The source and timing of silicification is interesting. The occurrence of silicified evaporites as intraclasts in member E1 intraformational conglomerates indicates that the silicification was extremely early. In Phanerozoic examples of silicified evaporites, a

biogenic source for the silica is commonly proposed (Chown and Elkins, 1974). Recent isotopic studies (Milliken, 1979) indicate that although some of the silica in such cherts is of marine origin, much of it is of meteoric origin. According to Knauth's 1979 model for the origin of chert nodules, the meteoric silica is derived from dissolution of biogenic silica in previously deposited carbonates. Clearly this is not the case in Precambrian silicified evaporites. It is proposed that the silica necessary for silicification was derived from: meteoric groundwater that flowed into the carbonate units through a siliciclastic aquifer, dissolution of quartz grains in interbedded siliciclastics, or possibly by the transition of smectite to illite (Hower et al., 1976; Eslinger and Sellars, 1981). The first two conditions are favoured; the last possibility is discounted because the smectite-illite transition cannot produce the amount of silica needed and it does not take place sufficiently early during diagenesis. Isotopic analysis of the different silica morphologies within the East River Formation should shed light on the source of silica.

Summary

The East River Formation constitutes a transgressive-regressive cycle, up to 250m thick, deposited during eastward onlap and subsidence of a west-northwest-facing shelf. The initial phase of transgression and marine redistribution of Lady Nye system clastics is recorded in the mixed siliciclastic and carbonate lithologies of member El. The

bathymetric maximum of the transgression covered the entire preserved limits of the East River shelf (30,000km²) and was characterized by the alternation of storm and fairweather deposition (member E2). Peritidal dolostone of members E3 and E4 formed during regional shoaling and coastal offlap. The latter phase of offlap (member E4) was associated with syn-sedimentary movement along the Teshierpi Fault, a structure that continued to influence facies geometry and depositional patterns throughout upper Hornby Bay Group and lower Dismal Lakes Group sedimentation.

The East River Formation is interpreted to have been deposited on a carbonate ramp (cf. Ahr, 1973; Read, 1980; 1982). This interpretation is based largely on negative evidence such as the lack of bank-margin build-ups and peri-platform allodapic carbonates (cf. McIlreath and James, 1979). Giant stromatolite biostromes in member E2 are the most likely candidates for a shelf margin facies; however, they occur randomly throughout the basin and do not resemble bank margin stromatolites (cf. Hoffman, 1974). It may be that the bank margin facies is exposed farther to the west, but the character of the lower two members of the East River Formation, acyclic deposits of a high-energy, wave-dominated unprotected coast/shelf, is totally unlike that expected in shelf lagoons (Wilson, 1975).

An "intraramp" break-in-slope was present between the Narakay Islands and regions to the east. This is inferred on the basis of: 1) the abrupt eastward thinning and pinch out of

hummocky cross-stratified arenites in upper Lady Nye Formation, 2) siliciclastic to carbonate cycles in member E1 are most complete in the Narakay Islands, 3) there is a pronounced change in member E3 from inter/supratidal to entirely subtidal in this region, 4) facies are uniformly "deeper" in the west, and 5) regionally extensive units are thickest in the west.

The west-northwest-facing East River carbonate ramp was an "open shelf" subjected to severe storm attack; storms were particularly intense during deposition of lower member E1 and throughout deposition of member E2. This coincides with a paleolatitude of 15-30° from the equator (south?) (Irving and McGlynn, 1979; 1981) which falls within the modern latitude of greatest frequency of hurricane and cyclone activity (Marsaglia and de Vries Klein, 1983).

With respect to the two most commonly cited modern analogues of carbonate sedimentation (Bahama Banks and Persian Gulf) the East River Formation is a "hybrid" in that it contains elements of each of these areas. It is similar to the arid Persian Gulf carbonate environment in that: 1) it had a ramp-like character to the shelf, 2) oolitic facies are nearshore, rather than bank-margin accumulations, and 3) evaporites and tepee structures are common. It is similar to the humid Bahama Bank carbonates in that it was: 1) an open shelf subjected to frequent unmitigated wave attack (windward margin of the Bahamas, cf. Hine, et al., 1981) and 2) member E3 algal mat facies are similar to those in Bahamian algal

marshes. It is concluded that the East River Formation was deposited on an open high-energy shelf in a strongly seasonal semi-arid to humid subtropical climate.

In comparison with numerous published descriptions of carbonate shelf and ramp sequences, the East River Formation is unique. Intracratonic and passive margin shelves invariably have some sort of distinct bank margin facies (MacIlreath and James, 1979; Wilson, 1975). Carbonate ramps, which appear to be typical of foreland basin carbonates (Read, 1980; Loreau and Purser, 1973), also are dissimilar to the East River Formation. In all of the reviewed examples, none is characterized by a thick sequence of high-energy storm deposits analogous to those described from open clastic shelves (cf. Dott and Bourgeois, 1982; Mount, 1982). The significance of these accumulations is not known at this time.

Plate ElA: Stacked arenite beds in lower part of siliciclastic-to-carbonate cycle. Note flat base and scoured top of massive (inferred storm deposit) arenite, and starved megaripples. Recesses between sand beds are weathered black mudstone layers. 7cm penknife (middle of photo) for scale.

ElB: Arenite-mudstone beds from siliciclastic-to-carbonate cycle. Arenites display flat to micro cross-lamination; mudstones have desiccation cracks and pull-apart structures. 15cm pencil for scale.

ElC: Bedding plane exposure of arenites in middle part of siliciclastic-to-carbonate cycle; calcareous arenites in upper part of cycle occur at the left side of the photo. The bedding planes are covered with interference (ladderback) ripple marks. Note that the axes of the most prominent ripple set have the same orientation throughout the sequence. Hammer (upper left) for scale.

ElD: Cross-stratification in calcareous arenite that consists of tabular planar (Sp), low-angle inclined (Si), trough (St), and ripple cross-lamination (Sr). Note current reversals. Matchstick (left side of photo, in crack) is 5cm long.

ElE: Thin section of millet-seed arenite from Falcon Lake. Moderately well-sorted feldspathic arenite with coarse sand-size grains of silicified oolite (O) and polycrystalline quartz (Q). Cement consists of pressure solution welds and replacive dolomite along grain boundaries. Scale bar is 1mm long. Crossed nicols.

ElF: Millet-seed sand grain displaying xenotopic extinction pattern, typical of quartz formed during evaporite replacement. Scale bar is 0.5mm long. Crossed nicols.

PLATE E1

A

B

C

D

E

F

Plate E2A: Nodular silicification in cross-bedded oolitic dolostone (East River, 86M). 14cm pen for scale.

E2B: Ooids and superficial ooids within cross-laminated quartzose oolitic dolostone lithofacies. "S" is silicified zone and displays poorly preserved ooid cortices composed of microgranular chert. Note isopachous chalcedony infill of inter-ooid pore space. "D" is dolomitized zone. Superficial ooid nuclei are composed of quartz and rare feldspar grains. Crossed nicols. Scale bar is 1mm long.

E2C: Thin section of silicified oolites from quartzose oolitic dolostone lithofacies. Ooids are composed of microcrystalline silica and dolomite that preserve relic radial and less pronounced concentric structures in ooid cortices. Pores have been infilled by fibrous isopachous length-fast chalcedony (F) and equigranular chert (C). Dolomite rhombs at ooid-cement contacts are post-silicification, as suggested by local cross-cutting of silicified ooid cortex by dolomite (A, upper left). Scale bar is 1mm long. Crossed nicols.

E2D: Member E1 (Lac Le Roux) SH-C stromatolites overlie intraclastic packstone and coalesce into LLH domal stromatolites. The stromatolites are overlain by quartzose oolitic dolostone with herringbone crossbeds. Note mass of silicified nodular evaporites (E) at base of crossbedded sequence. Brunton compass for scale.

E2E: Channel facies of SH-C gradational into LLH stromatolite bioherms on in situ and ripped-up blocks of oolitic dolostone. Note quartzose intraclastic debris in interbiohermal channel, and rip-up slab on stromatolite surface that has acted as a nucleus for subsequent stromatolite growth. Length of matchstick (right center) is 5cm.

E2F: Upward coalescence of SH-C into LLH stromatolites. Intraclastic dolostone occurs at the base of the succession. SH-V shape is accentuated by the angle of the outcrop surface, which slopes gently towards the observer. Field of view is 1m wide.

PLATE E2

A

B

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Plate E3A: Biostrome of microdigitate stromatolites (Ritch Island). SH-C domes rest on sharp scour surface and coalesce into LLH domes. The composition of the substrate is unknown because of poor preservation but appears to be calcareous. arenite. Hammer for scale.

E3B: Biostrome of laterally-linked domal stromatolites overlying stacked sets of massive (?) oolitic dolostone (East River, 86K). These stromatolites are composed of smooth, rather than colloform, laminations and are typical of the biostromes in the upper part of member E2. Height of cliff face is 10m.

E3C: Hand sample of silicified digitate/microcolumnar stromatolites typical of member E2 biostromes. Stromatolite interareas are filled with dololutite and arenaceous debris. Note common branching and local lateral linkage of columns.

E3D: Detail of microdigitate stromatolites as seen in thin section. Note numerous fining up siltstone-mudstone laminae in interareas and agglutinated silt grains on algal heads. Scale bar is 5mm long.

E3E: Thin section of recrystallized ooids showing neomorphism of dolostone that has resulted in extreme loss of original textures. Darker areas within ooids are composed of chert. Scale bar is 1mm long. Crossed ooids.

E3F: Clastic lithofacies between stromatolites (Narakay Islands). Undulose to hummocky cross-stratified quartzose oolitic dolostone is overlain by type B climbing ripples. Matchstick is 5cm long.

PLATE E3

A

B

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Plate E4A: Smooth mat LLH and SH-V (mushroom-shaped) stromatolites of member E3 (section 2, south of Dismal Lakes). Note ribbed and nodular pattern of silicification. 15cm pen for scale.

E4B: Bedding plane exposure of tight-packed LLH stromatolites (Dease Arm) in member E3.

E4C: Pocket of edgewise conglomerate on surface of sooty gray domal stromatolites (East River, 86M). Clasts are draped by algal mats that form second bed of domal stromatolites. 5cm matchstick (upper left) for scale.

E4D: Thin section of silicified oolite from East River area. Ooids are composed of chert that in several places preserves delicate lamination within the ooid cortex. Intergranular cement is composed of blocky dolomite spar (limpid dolomite) (D, upper right) and botryoidal pore rim chalcedony with chert in the core (C). Scale bar is 1mm long. Plane polarized light.

E4E: Detail of palisade algal mat textures in sooty gray stromatolitic dolostone of member E3 (East River, 86M). Lamination consists of darker crinkled algal mats, light-coloured lutite/peloid layers, and layers with tufted to clotted algal mats and irregular sparry fenestrae. Head of 4cm matchstick rests on patch of replacive black chert.

PLATE E4

A

B

C

D

E

Plate E5A: Smooth to colloform mat LLH stromatolite crusts on scoured current-laminated dolostone (member E3, Narakay Islands). Note intraclastic chips in some interdomal areas. Stromatolite crusts in center of photo are about 5cm thick.

E5B: Microhummocky cross-stratification in oolitic (?) dolostone (Narakay Islands). This lithofacies occurs as beds between stromatolites and contrasts with the massive texture of clastic beds typical of member E3 to the east.

E5C: Outcrop of member E4 south of Dismal Lakes (near Teshierpi Fault) showing interbeds of stromatolitic and intraclastic dolostone and recessively-weathering red mudstone. 1.7m person for scale.

E5D: Intraclastic dolostone of member E4 with crude parallel stratification (south of Lac Le Roux). Layer of laminated dololutite occurs at top of intraclastic bed. Lens cap for scale.

E5E: Hand sample of chert pebble conglomerate in uppermost member E4 near Teshierpi Fault. Note inverse grading and rinds on many of the clasts. Coin for scale is 15mm wide.

E5F: Photomicrograph of silicified evaporite rosettes in interbedded quartzose dolosiltite (light layers) and dololutite (dark layers). Rosettes are composed of polycrystalline prismatic megaquartz and spherular botryoids of chalcedony. Scale bar is 3mm wide. Plane polarized light.

PLATE E5

A

B

C

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F

Plate E6A: Detail of surface of cauliflower geode with well-preserved rosettes of blade-like pseudomorphs composed of xenotopic quartz. Coin is 15mm wide.

E6B: Polished slab of cauliflower geodes within partly silicified dolomite matrix from lower member E3, near Teshierpi Fault. Rims of geodes are composed of pink (gray in photo) spherulitic silica that grades into central region composed of poikilitic megaquartz. Coin is 15 mm wide.

E6C: Cauliflower geodes within stromatolite head from member E2 (Leith Peninsula). Stromatolite is overlain by base of hummocky cross-stratified quartzose oolitic dolostone. Coin is 15mm wide.

E6D: Photomicrograph of margin of cauliflower geode. Matrix is composed of fine-grained dolomite. Tabular pseudomorphs composed of nearly isotropic silica that grades inward to spherulitic chalcedony and megaquartz. Photograph is 5mm wide. Crossed nicols.

E6E: Photomicrograph of transition of chalcedonic spherules into megaquartz. Chalcedonic spherules are partly recrystallized to megaquartz but still retain radial extinction of concentric spherules. Chalcedonic spherules grade into spherules composed of elongate to equant poikilitic megaquartz with poor preservation of spherular texture. Scale bar is 1mm long. Crossed nicols.

E6F: Photomicrograph of oolitic dolostone with quartz euhedra in member E1 (Falcon Lake area). Well preserved ooids are silicified (left center). Poikilitic quartz euhedra are replacements of evaporites. Inter-ooid cement is composed of sparry dolomite. Note local replacement of chert ooids by dolomite rhombs (D). Scale bar is 1mm long. Crossed nicols.

PLATE E6

A

B

C

D

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CHAPTER 6

KAERTOK FORMATION

The Kaertok Formation is composed of interbedded mudstone and fine-grained arenite to siltstone. It occurs sporadically from the region south of central Dismal Lakes (Teshierpi Mountain area) as far west as Falcon Lake, south to East River and Blood Creek, and on into Dease Arm (Narakay Islands; Figure 6-1). It also occurs on the Superstition Islands (southern Great Bear Lake) and on Leith Peninsula. Although the formation is widespread, the high mudstone content typical of the unit leads to poor outcrop quality and continuity. Four nearly complete stratigraphic sections were observed; three of these (Kaertok Creek, East River, and Narakay Islands) are described in detail, as they cover the range of lithologic variations encountered within this formation.

The word "kaertok" is an Inuit word meaning "solid rock" (Thibert, 1969) and is proposed as the name for an unnamed creek that drains into southern Dismal Lakes. Red mudstone and arenite (Unit 10 of Baragar and Donaldson, 1973), is well-exposed in a steep-walled gorge that forms the lower reaches of this creek. It is here proposed that this creek section be designated as the type section of the Kaertok Formation.

At East River (86M), Falcon Lake, and south of west Dismal Lakes, the Kaertok Formation grades into Unit 11 (referred to informally as the Mountain Lake formation), the basal unit of the Dismal Lakes Group. South of Teshierpi Mountain, the

FIGURE 6-1: Location map showing distribution of the Kaertok Formation (horizontal rule) and location of measured sections in Figure 6-2. 1) Kaertok Creek section, 2) East River section, 3) Narakay Island section. F.L.: Falcon Lake; T.M.: Teshierpi Mountain.

Mountain Lake formation rests unconformably on thinned and densely silicified remnants of the Kaertok Formation; the Kaertok Formation does not occur east of the Teshierpi Fault (that section of the fault north of 67°N). In the Narakay Islands, the Kaertok Formation is overlain by the Narakay Lithic Tuff of the Narakay Volcanic Complex (see next chapter) and dololutes of uncertain stratigraphic position. No upper or lower contacts were observed in exposures from Blood Creek, Superstition Islands, or Leith Peninsula.

Kaertok Creek section

The lowest 35 metres of the Kaertok Formation exposed in the type section is composed of interbedded granulestone, arenite, and mudstone. The base of the formation is not exposed at this locality, but the lithologic sequence is similar to the basal 10 metres of Kaertok Formation that overlie the East River Formation south of Teshierpi Mountain. Beds are arranged in a vertical sequence that displays an overall upwards thinning of average sand bed thickness and an increased proportion of mudstone, concomitant with increase in grain size of the coarser clastic beds (Fig. 6-2). Low in the section arenite-mudstone couplets predominate; higher in the section, fining-up cycles (10-50cm thick) become dominant and are composed of, in ascending order, intraclastic granulestone, mudstone with siltstone streaks, and calcareous mudstone and dololite. At the base of the section, arenites occur as 5-30cm thick tabular sheets that rest on planar to gently undulose scour

surfaces. Upper surfaces of the arenites commonly display symmetric to asymmetric sinuous to catenary current/wave ripple marks. The arenites are mostly medium- to fine-grained, moderately sorted quartz arenites with <5% feldspar. Angular to subrounded clasts of red to brown mudstone are ubiquitous, being concentrated at the base of arenite sheets and along foreset laminae. Arenite sheets are cosets that consist of multiple smaller sheets with breaks between sets marked by shallow scours, mudfilms, or mudchips. Parallel lamination is the most common stratification within the arenites, but ripple cross-lamination is also common, and type A climbing ripple stratification was observed in several places. Paleocurrent measurements from both ripple axes and crossbeds give a polymodal rose diagram (Fig. 6-2).

Granulestones and mudchip-rich arenites (sandy flakestones) occur as massive 2-10cm scour-based sheets at the base of fining-upward cycles (Plate K1B). The framework is composed of moderately sorted, rounded granules and coarse sand grains of quartz, red chert, feldspar, and cherty green mudstone packed in a muddy matrix. Mudchips are angular to subrounded and composed of red and green siliceous mudstone and pale tan-coloured clasts of calcareous mudstone and dololutite.

The fine-grained portion of fining-upward cycles are composed of red mudstone and, at the top of the cycle, dololutite (Plate K1A, B). The mudstones are dark red, commonly with green pigment in irregular bands parallel to layering. Lenticles of siltstone and less commonly granulestone, occur in

the lower portion of the mudstone intervals. Mudstone layers are commonly disrupted by desiccation cracks that are infilled with sand-size clastic material. Calcareous mudstone and dololomite occurs at the top of complete fining-upwards cycles (Plate KlA, B). In the upper part of the section the red (siliceous) and white (calcareous) mudstones are interbedded on the scale of 5-10cm, with no interbeds of coarser material (Plate KlA). Desiccation has broken the mudstone layers into a series of disconnected slab-like fragments. The dolostones occur as white, chalky layers (<5cm thick) with a massive texture or parallel lamination. Desiccation caused disruption of lamination in the dololutes and in some places formed in situ breccias that have fractures that are infilled with overlying granulestone (Plate KlB).

The middle and upper parts of the Kaertok Creek section are composed of a monotonous succession of finely interbedded mudstone, siltstone, and fine-grained arenite that is about 120 metres thick. These lithologies occur in irregular fining- and coarsening-upwards cycles that range from 1.5 to 15 metres thick. Arenites and siltstones are rich in quartz and comprise massive to flat-laminated sheets and lenses 5-15 centimetres thick. Mudstones are entirely disrupted by numerous desiccation cracks that caused mixing of interbedded unlithified silts and muds, and formation of in situ breccias. The average grain size and bed thickness of the coarser clastic layers remains essentially unchanged for 100 metres; thereafter it coarsens and thickens upward, concomitant with decreased

abundance of mudstone, as the Kaertok Formation grades into basal clastics of the Dismal Lakes Group (Mountain Lake formation).

East River Section

Exposures of the Kaertok Formation in the East River (86M) region form north-trending ridges on the stoss side of a prominent roches moutonnees. The eastern part of the section has a moderate dip to the west (35°) that decreases towards the west to 10° , near the contact with the Mountain Lake Formation. The Kaertok Formation is much thicker here than in the section at Kaertok Creek or Falcon Lake; 400m was measured, but the base of the section is not exposed. In general, the succession comprises two upward-coarsening cycles from mudstone up to pebbly arenite (Fig. 6-2). Each coarsening-up cycle is composed of three lithofacies: 1) thin-bedded arenite-mudstone grading into 2) crossbedded arenite that fines up into 3) thin-bedded arenite. A fourth lithofacies, red mudstone and arenite, occurs only at the base of the section and is similar to exposures of the Kaertok Formation to the north and northeast.

Red Mudstone-Arenite Lithofacies

This lithofacies constitutes the lower 20m of the East River section. It is composed of red mudstone beds (20-50cm thick) and tabular to lenticular beds (5-30cm thick) of feldspathic hematitic quartz arenite (Plate K1C). Arenite beds rest on scour surfaces, have mudchips at the base of the bed,

and display parallel lamination and rare ripple cross-lamination. The tops of arenite beds are commonly rippled, and overlying mudstone drapes are mudcracked. Mudstone beds display parallel lamination and mudcracks. The gradation from this lithofacies into overlying drab-coloured arenites and green mudstones of lithofacies 1 is marked by a decrease in average sand/siltstone bed thickness and change from red to green pigmentation in both arenites and mudstones.

Thin-bedded Arenite-Mudstone Lithofacies

This lithofacies comprises the fine grained base of coarsening-up cycles, and occurs as 1-3m thick intervals within the thin-bedded arenite lithofacies. It is composed of interbedded fine-grained arenite to siltstone and mudstone. Arenite/siltstone beds are commonly 1-5cm thick, with rare beds up to 20cm thick, and are composed of moderately well-sorted quartz arenite, with minor feldspar (<5%), and a quartz-chlorite cement. They range in colour from green to white and occur as tabular, wavy, and lenticular beds (Plate K1D). Internally these beds may be massive, or they may display ripple cross-stratification, parallel-lamination, or rarely climbing ripple (type B) stratification. Interference and runoff ripple marks are common on bedding surfaces as are spindle to pustular-shaped casts that are inferred to be the result of syneresis (cf. Plummer and Gostin, 1981). Some stratigraphic intervals of this lithofacies have been affected by soft-sediment deformation that caused minor but widespread convolutions of beds

(Plate K1D). Mudstones occur as 1-5cm thick beds intercalated with the arenites; they range in colour from dark green to reddish. Upward coarsening within this lithofacies is marked by a gradual increase in sandstone:shale ratio, increased sand bed thickness, and in several examples the sudden appearance of abnormally thick scour-based sheets of arenite.

Crossbedded Arenite Lithofacies

Thin-bedded arenite, siltstone, and mudstone grades into dominantly arenaceous material of the crossbedded arenite lithofacies. The gradation occurs by an increase in percentage of sandstone and decrease in percentage of mudstone over a poorly-exposed 10-20m interval. The arenites range from well-sorted medium-grained quartz arenite to moderately-sorted granular to pebbly, subfeldspathic arenite, both with a blotchy drab to mauve pigmentation. Pink-coloured grains are conspicuous but minor (<2%) components of the arenites; thin sections reveal that these grains are invariably composed of reworked silicified oolites and chert (Plate K1F), derived from the East River Formation. Felsic volcanic fragments, probably derived from basement rocks, were observed in several samples.

Trough crossbeds (30-80cm thick) are the most common sedimentary structure within this lithofacies, followed by tabular-planar crossbeds and horizontal to low-angle lamination. No systematic vertical arrangement of these crossbed types was observed. Crossbed orientations within this

lithofacies are strongly unimodal (all measurements are within a single 90° quadrant) to the southwest and west. (Fig. 6-2).

Thin-bedded Arenite Lithofacies

This lithofacies gradationally overlies the crossbedded arenite lithofacies and grades into thin-bedded arenite-mudstone lithofacies. It is the uppermost lithofacies of coarsening-up cycles and represents a relatively thin fine-grained cap developed at the top of these cycles. It is composed of medium-grained quartz arenites that have a blotchy, mauve-coloured hematitic pigmentation; angular mudchips are common at the base of crossbed sets and along foreset laminae.

The arenites occur in 5-30cm sets of cross-strata that display tabular-planar and low-angle ($<10^\circ$) cross-stratification and parallel stratification (Plate K1E). Sets are stacked into cosets composed of 2-3 sets; cosets are interbedded with thin (<5 cm thick) mud films and, less commonly, lenticular to wavy-bedded arenite and red mudstone (<20 cm thick). Bedding surface exposures of arenite-mudstone contacts reveals the presence of a diverse array of syneresis structures that range from rectilinear spindles to wart-like protuberances.

Paleocurrent measurements from this lithofacies are either bimodal-bipolar or strong unimodal. In both cases (Fig. 6-2) a pronounced northeast mode is common, which is of opposite orientation to the mode from the underlying crossbedded arenite lithofacies. The bimodal-bipolar dispersion pattern is reflected in outcrops that display herringbone crossbeds

(Plate K1E). No systematic vertical arrangement to the cross-bed types was observed, although there is an up-section decrease in bed thickness as this lithofacies grades into the thin-bedded arenite-mudstone lithofacies.

Narakay Islands Section

At least 165m of Kaertok Formation lithologies are present in the Narakay Islands; this is a minimum thickness because the section is disrupted and truncated by faults and pyroclastic deposits. The section presented in Figure 6-2 is a composite section based on sections measured on three different islands within the Narakays. The Kaertok Formation in this region comprises four lithofacies. Thin-bedded siliciclastic and carbonate rocks compose the bulk of the section followed by stromatolitic dolostone, massive arenite, and oolitic/intraclastic arenite. Volcaniclastic deposits are also a common component of the Kaertok Formation in the Narakay Islands, but these will be described in the next chapter. This lithologic assemblage is quite different from the Kaertok Formation exposed to the east, in that dolostone is a prominent constituent, siliceous mudstone is absent, and the average grain size of the siliciclastic material is smaller. The Narakay Island section is correlated with other exposures of the Kaertok Formation because this formation represents regional scale reintroduction of terrigenous clastic detritus into a dominantly carbonate environment.

The base of the Kaertok Formation in the Narakay Islands

is marked by the first occurrence of thin-bedded siliciclastics and/or stromatolitic dolostone with SH-V mode stromatolites. The top of the formation is marked by the occurrence of a massive lithic tuff to tuff breccia (Narakay Lithic Tuff) which is widespread throughout the Narakay Islands. Up-section changes within the formation are a change in stromatolite morphology from SH-V to LLH, increased sorting and grain size of siliciclastics, and evidence of periodic emergence and desiccation of the depositional surface.

Thin-bedded Siliciclastic and Carbonate Lithofacies

This is the most common lithofacies of the Kaertok Formation. It is composed of centimetre-scale interbeds of variably calcareous, fine-grained arenite to siltstone and dolostone. As a result of differential weathering of these two rock types, outcrops display a distinct ribbed pattern (Plate K2C). Siliciclastic rocks are composed of a tightly-packed framework of well-sorted quartz grains, with subordinate feldspar, zircon, and tourmaline. The siliciclastic beds range in shape from tabular to wavy to lenticular (Plate K2A); internally they appear massive, but this is in part a reflection of homogeneous grain size. Dolostone layers occur as lenticular to tabular beds; ripple cross-lamination is locally common (Plate K2A), but most dolostone appears massive. The original grain size characteristics of the dolostone have been masked by recrystallization.

Upwards within the section, this lithofacies changes gradually into thinly-interbedded fine-grained quartz arenite and green siliceous mudstone (Plate K2F) with no dolostone interbeds. The arenites have a distinct pink colour, display parallel lamination, and contain intraclasts of green mudstone. The arenite beds may grade into the overlying mudstone layer or they may be abruptly overlain by mudstone. The mudstones occur as 1-2 cm thick beds that are commonly disrupted by mudcracks. Bedding surfaces of the arenites display well-developed mudcracks and wave-generated asymmetric ripple marks.

Stromatolitic Dolostone Lithofacies

Stromatolites occur in 10-60cm thick biostromes; individual biostromes cannot be correlated between outcrops and some biostromes were observed to pinch out over tens of metres. Stromatolites almost invariably overlie a scour surface that is commonly littered with intraclasts. They are most commonly draped by thin-bedded siliciclastic and carbonate lithofacies, but they are also overlain by massive beds of oolitic/intraclastic dolostone. In the lower part of the formation, the stromatolites occur as spaced to locally coalescent SH-V columns up to 1m wide, with 50cm of synoptic relief (Plate K2B). Most of the heads are symmetric in cross-section but some display a weak but consistent southwest-northeast elongation. Lamination within the stromatolites consists of scalloped, concave-down algal mats and lenses of dololomite. High-energy interdomal clastics, in the form of angular intraclasts.

and overturned blocks of stromatolites up to 15 cm wide, are locally present.

Upwards in the formation the average size and thickness of the biostromes decreases, and SH-V forms are replaced by LLH bioherms (Plates K2D,E). Bioherms, up to 2m in diameter, are composed of 5-10cm wide SH-C domal stromatolites that coalesce into polygonally packed LLH domal individuals. Clastic interdomal debris is absent. Some of the stromatolites display a pronounced southwest-northeast elongation (Fig. 6-2). Towards the top of the formation stromatolites occur as biostromes of undulose LLH domes (Plate K2D) and less commonly as flat to peaked mats and rare oncolites.

Massive Arenite Lithofacies

Fine-grained massive arenites are most common in the lower part of the section (Plate K2A), and are absent from the upper 50m. They occur as tabular sheets (10-50cm thick) with no internal structure; no evidence of scour was observed at the top or bottom of these beds. They are composed of a framework of fine-grained sand-size quartz, feldspar (<5%, microcline and plagioclase), zircon and tourmaline. Some of the massive arenites contain fragments of volcanic porphyry typical of the Narakay Volcanic Complex. Dolomite is ubiquitous and composes up to 25% of the rock; it occurs as scattered subhedral clots with no discernable primary texture (oolites, for example).

Oolitic/Intraclastic Lithofacies

Dolostone, composed of oolitic and rounded, sand-sized intraclastic debris, occurs as 30-50cm thick beds within the Kaertok Formation (Plates K2A,E). In examples where recrystallization has not obliterated primary textures, each oolitic bed is a coset of 3-7 tabular-planar crossbeds. Paleocurrents inferred from the crossbeds define a weak bimodal-bipolar to strong, unimodal pattern that indicates dominant transport to the west-southwest (Fig. 6-2). Beds composed exclusively of intraclastic material generally do not display cross-lamination, but occur as massive sheets. In the lower part of the formation these beds are thin and the intraclasts are angular; higher in the section they are thicker and composed largely of well-rounded dolostone fragments.

Interpretation

The character, vertical sequence, distribution, and thickness of lithofacies within the Kaertok Formation can best be explained within the context of a deltaic depositional system. The coarsening-up cycles in the East River section are similar to those described by Vos (1977) and Tankard and Barwis (1981) and are considered to be typical of deltaic systems (Miall, 1979). The three lithofacies that comprise each coarsening-up cycle in this section record progradation and abandonment of delta lobes. The thin-bedded arenite-mudstone lithofacies (1 in Fig. 6-2) is interpreted as tidally-influenced deposits of the delta plain/prodelta. They coarsen upwards into

inferred distributary channel facies (crossbedded arenite lithofacies) deposited during progradation of a southwest-flowing fluvial channel. Channel-abandonment was followed by reworking of the fluvial deposits by onshore-directed marine currents and deposition of the thin-bedded arenite facies. Paleocurrents for this facies are to the northeast, opposite to those in the underlying fluvial channel, and may reflect onshore surge of tidal- and wave(?) -generated currents up the abandoned channel.

The section described from Kaertok Creek and the lithologically similar section from Falcon Lake are inferred to be overbank-type deposits. The basal part of the Kaertok Creek section may be marine, based on the dispersed (polymodal) character of the paleocurrents, but this is based on only a few measurements. The bulk of the section is inferred to have been deposited by thin sheetfloods (cf. Hogg, 1982; Hubert and Hyde, 1982) that left upward-fining deposits of intraclastic arenite and mudstone. The muds were desiccated during emergence after flooding. Calcareous mudstones and dololomite beds at the top of cycles probably were formed during evaporation of playa-like ephemeral lakes (cf. Smoot, 1978).

Exposures of the Kaertok Formation in the Narakay Islands are inferred to have been deposited in a peritidal marine environment. Sedimentary structures within the thin-bedded siliciclastic and carbonate lithofacies resemble those formed by wave action in a subtidal to intertidal setting (cf. de Raaf, et al., 1977; Reineck and Singh, 1975). The

FIGURE 6-3: Schematic paleogeographic reconstruction for the Kaertok Formation. Section locations refer to sections in Figure 6-2. Note inferred control of drainage system and sedimentation pattern by the Teshierpi Fault.

siliciclastic component of this lithofacies is composed of reworked (allochthonous) terrestrial detritus, whereas the dolostone represents autochthonous material. The stromatolites within this section also formed in a subtidal to intertidal setting; biostromes appear to have been restricted to shallow channels developed on the depositional surface. The oolitic/intraclastic lithofacies was deposited on low-relief shoals that were affected by predominantly offshore-directed currents. The massive arenites were deposited suddenly, and may be storm surge deposits. Up-section changes within the formation are indicative of decreased energy and water depth (i.e. upward-shoaling).

It was mentioned in the chapter on the Lady Nye depositional system that most models for Phanerozoic delta systems would be inappropriate for non-channeled fluvial systems developed before the advent of vegetation. However, a "Phanerozoic type" model for deltaic sedimentation is believed appropriate for the Kaertok Formation because of the inferred structural control of drainage orientation by the Teshierpi Fault. Thus the Kaertok depositional system would have had a central trunk river that drained into a standing water body rather than a shallow and very wide channel system (as inferred for other fluvial units in the Hornby Bay Group). Synsedimentary movement along the Teshierpi Fault effectively restricted the location of the channel "thalweg" to regions close to the trace of the Teshierpi Fault. The drainage system debauched into marine waters in the East River region and in regions to the south.

The inferred control of Kaertok Formation drainage by the Teshierpi Fault is supported by the strong parallelism between distributary channel paleocurrents (Fig. 6-2, East River section) and the trace of the Teshierpi Fault. In addition, synsedimentary movement along this fault has already been documented during deposition of member E4 of the underlying East River Formation, and it would seem reasonable that this movement might continue during Kaertok Formation deposition. Structural control of Kaertok Formation deposition is also supported by the pattern of facies distribution and thickness variations. Overbank facies, as exemplified by the type section of the Kaertok Formation, are restricted to regions to the northwest of Teshierpi Fault. This facies thickens from northwest to southeast as the fault, and the coarse-grained facies of the contemporary delta facies, are approached. It is therefore concluded that synsedimentary movement along the Teshierpi Fault controlled the style and locus of deposition of the Kaertok Formation. This period of movement, which was initiated during deposition of member E4 of the East River Formation, is a reflection of regional regression and offlap of terrestrial siliciclastics.

Plate K1:

A. Overview of mudstone-rich portion of Kaertok Formation in the type section. Light-coloured layers are calcareous mudstone and dololutite. Height of cliff face is 3 metres.

B. Fining-upward cycle in lower Kaertok Formation, type section. Intraclastic granulestone forms base of one cycle and overlies red mudstone, calcareous mudstone, and dololutite at top of underlying cycle. Note that the granulestone infills desiccation fractures within the dololutite bed. The granulestone fines-up to intercalated siltstone-mudstone and desiccated red mudstone. Knife is 9cm long.

C. Scour-based, fine-grained lenticular arenites and red mudstones in lower East River section. Field of view is 1.5m wide.

D. Wavy- and lenticular-bedded siltstone and green mudstone of thin-bedded arenite-mudstone lithofacies (East River section). Note deformation of arenite/siltstone beds and sand-filled syaeresis cracks. Pen is 14cm long.

E. Stacked cosets of fine-grained hematitic arenite of thin-bedded arenite facies (East River section). Cosets are separated by films of red mudstone. Hammer for scale.

F. Photomicrograph of quartz arenite from upper part of upward-coarsening cycle in East River section. Arenite is composed largely of quartz grains (clear) with subordinate but conspicuous grains of alkali feldspar (light gray) and chert (dark). Cherty grains are composed of ferruginous chalcedony formed during silicification of evaporites in East River Formation. Note truncated ooid in upper right. Width of photo is about 5mm. Plane polarized light.

PLATE K1

A

B

C

D

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F

Plate K2:

A. Detail of thin-bedded siliciclastic and carbonate lithofacies. Resistant ribs of massive and parallel-laminated calcareous siltstone; note tabular to lenticular ("starved ripples") shape of beds (arrow). Recessive beds of ripple cross-laminated dolostone (oolitic?). Field of view is about 30cm wide.

B. Plan view of solitary to coalescent symmetric SH-V domes within stromatolitic dolostone lithofacies (Narakay Islands section). Interdomal material is mixture of volcanic sand and oolites. Hammer in center for scale.

C. Outcrop with three of the lithofacies within the Kaertok Formation (Narakay Islands section). Outcrop is composed of thin-bedded siliciclastic and carbonate lithofacies (1), oolitic/intraclastic dolostone (2), and massive arenite lithofacies (3). Note crossbeds within oolitic bed and characteristic ribbed pattern of siliciclastic and carbonate lithofacies. Hammer at right for scale.

D. Thin-bedded siliciclastic and carbonate lithofacies and interbed of massive intraclastic granulestone overlain by LLH bioherms. Arrows point to two bioherms composed of laterally linked smooth domal stromatolites. Each biostrome is about 10cm high.

E. Biostrome of laterally linked domal stromatolites overlain by massive intraclastic granulestone (I) and flat-laminated arenite (A). Uppermost Kaertok Formation (Narakay Islands section). Synoptic relief between heads on surface of biostrome is about 25cm.

F. Pink arenite and desiccated green mudstone of upper Kaertok Formation (Narakay Island section). Note mudstone intraclasts. Matchstick is 4cm long.

PLATE K2

A

B

C

D

E

F

CHAPTER 7

NARAKAY VOLCANIC COMPLEX

The Narakay Volcanic Complex is exposed on the Narakay Islands, a group of more than 20 islands along the south shore of Dease Arm, Great Bear Lake (Fig. 1-2; 4-2). They contain the westernmost exposures of the Hornby Bay Group, and the only known volcanic rocks within the group. In general, the complex comprises a north-dipping homoclinal succession of shallow marine rocks (>500m thick) of the Lady Nye Formation, East River Formation, and Kaertok Formation (Fig. 7-2). Mafic pyroclastic and derivative epiclastic rocks occur throughout the lower 350 metres of the succession and rhyolite lavas and lithic-rich felsic tuffs occur in the upper 150 metres. The entire complex is cut by a swarm of east-trending feldspar-phyric dykes and related diatremes and explosion breccias. The Narakay Islands are separated from mainland exposures of correlative volcanic rocks to the south by a graben bounded by faults with more than 500m normal displacement (Fig. 4-2).

The only work done on the complex prior to 1980 was the description of two samples of "porphyrite tuff" by A.E. Barlow, who accompanied J.M. Bell in 1899 (Bell, 1901), and a single traverse by J.A. Donaldson in 1969 (Donaldson, unpublished notes, 1983). A preliminary report on the complex was published in 1982, after 2 seasons of detailed mapping of the complex (Ross, 1982). As a consequence of their marked relief, the Narakay Islands served as a landmark to early travellers

FIGURE 7-1: Simplified geologic map of the Narakay Volcanic Complex. Individual dykes and structural data are not plotted for the sake of clarity. The islands that are inset in the lower left are the westernmost Narakay Islands and lie 4km west of the dogleg-shaped island. The mafic porphyry is illustrated in Figure 7-3.

Figure 7-2. Schematic stratigraphic section for the Narakay Volcanic Complex. Black rectangles to the right of the column indicate periods of deposition of volcaniclastic rocks; shaded box is pyroclastic event inferred on the basis of coarse grain size of epiclastic rocks. Arrow with open head at base of Narakay Lithic Tuff indicates stratigraphic position of radiometrically dated rhyolite porphyry. Sea level units: S-subtidal; I-intertidal; Su-supratidal. Overlap of lines represents uncertainty in inferred water depths. Siliciclastic grain size/ (arrow at base of column) decreases to the left.

in the region; the original name of the islands, as appears on their maps, is "Narakazae Islands", a Great Bear Lake Indian word of uncertain meaning.

Mafic Pyroclastic Rocks

Structural Setting and Outcrop Pattern

The rugged topography and relief of the Narakay Islands owes its origin to the resistant character of the pyroclastic rocks. Typically these rocks outcrop as rugged knobs with up to 100m of relief (Plate N1A); the upper contact between the pyroclastic rocks and overlying rocks is nowhere preserved. Exposed contacts between the tuffs and adjacent to subjacent sedimentary rocks are steep (70-90°) and discordant. (Fig. 7-1). Exposed contacts between tuff and wallrock are either smooth curvilinear boundaries or they are marked by an irregular re-entrant boundary. Although bedding in the sedimentary rocks abuts bedding in the pyroclastic rocks, no direct evidence of fault truncation (e.g. slickensides, tension gashes, etc.) was observed along contacts. Along irregular contacts between tuff and wallrock, large blocks of wallrock are included within the tuffs; such blocks of country rock are absent from tuffs in contact with wallrock along curvilinear boundaries. Country rock across both types of contact commonly displays brecciation, boudinage, and recrystallization.

Bedding Style and Lamination

Pyroclastic rocks occur as massive to bedded sequences 40-100m thick composed of stacked, tabular-shaped beds of ash to lapilli tuff and agglomerate/ tuff breccia (terminology of Fisher, 1966) that display poorly-developed internal stratification (Plate N1F). No regular vertical sequence of bedding styles and/or textures was observed (cf. Wohletz and Sheridan, 1983). Individual beds range in thickness from 10-300cm thick. Boundaries between sets are mostly planar and beds rest on unscoured basal surfaces although some contacts display convolutions as a consequence of loading. Individual beds are defined on the basis of changes in grain size. No systematic variation in grain size or ratio of lithic to pumiceous clasts was observed although detailed granulometric analyses (cf. Tasse, et al., 1978) were not performed. Massive bedding is characteristic of the tuffs but where stratification occurs, it is defined by thin (2-5cm) ash tuff beds (Plate N1D), single-grain-thick concentrations of lapilli, and less common normal and reverse graded bedding (Plate N1C). Stratification produced by traction currents (cf. Harms, et al., 1982) was not observed. Tuff beds invariably are poorly sorted; lapilli tuff and tuff breccia are most common and comparatively well-sorted ash tuff is less common. Most of the tuff beds display a loosely-packed to disrupted framework (Plate N2B); framework-supported textures are uncommon. In some of the ash-rich beds, coarse blocks and bombs have contorted the underlying lamination and produced bomb- or impact-sag features

(cf. Williams and McBirney, 1979). In some outcrops, beds of laminated ash tuff drape a block-size clast that forms the head to a train of clasts that decrease in size downstream (current shadow, Plate N1E).

Bedding commonly dips at angles of 20-30° although dips as steep as 71° and as shallow as 10° were recorded (Fig. 7-6). The bedding in outcrops ranges from constant ("homoclinal") to variable; the origin of bedding orientations can be understood by comparing them with tuff beds on the largest island. Bedded tuffs define a bowl-like depression that is filled with columnar-jointed porphyry (Plate N1B; Fig. 7-3, 7-6). The tuff beds dip radially inward at 10-25° and the axes of the columnar joints are perpendicular to the plane of bedding in the tuffs. The inward-facing tuff beds pass through a zone of variable dips before they become uniformly outward-dipping (quaquaversal). The entire structure is nested topographically within older to penecontemporaneous Kaertok Formation sedimentary rocks and tuffs, respectively.

This structure is interpreted as the rim and inner portion of a small tuff cone; the columnar-jointed porphyry is likely solidified lava that ponded within the tuff cone crater. The tuffs were deposited on both inward-dipping and outward-dipping slopes of the crater. The systematic change in the orientation of bedding is similar to that observed in late Cenozoic to Holocene maars and tuff rings (cf. Heiken, 1971; Womer, et al., 1981; Schminke, 1977). By analogy with the relationships

FIGURE 7-3: Schematic cross-section of relationship between columnar jointed mafic porphyry and bedded tuffs (see Plate N1B for photograph).

described above, outcrops of pyroclastic rock that display a constant (homoclinal) orientation to bedding represent remnants of tuff cone flanks. Outcrops that display variable orientations of bedding represent sections through the cone margins (transition zone from inward-dipping to outward-dipping beds). In outcrops where the zone of variable bedding orientation does not define a systematic pattern, syndepositional slumping of beds could account for changes in original bedding orientation. Direct evidence of this process is present in the form of beds, albeit uncommon, that display stratiform deformation of lamination and the occurrence of large (up to 10m thick and 30m X 40m in area) slabs ("intra-clasts") of brecciated bedded tuff (Plate N1D).

Grain Shapes and Sizes

Studies of the morphometric characteristics of pyroclasts from modern volcanoes of known eruptive style provide a basis for interpretation of the eruptive mechanism in ancient pyroclastic deposits (cf. Honnorez and Kirst, 1975; Walker and Croasdale, 1972; Heiken, 1974). In the Narakay Volcanic Complex, juvenile volcanic rock fragments comprise more than 90% of the framework of pyroclastic rocks; accessory lithic fragments, composed of sedimentary country rock, comprise the remaining 10%. Cognate lithic fragments are a minor component, generally recognizable only in thin section.

Lapilli- to ash-size clasts of volcanic debris display two morphologies (Plate N2B,E). Grains generally display an equant

shape and do not display any flattening, thereby implying that they were relatively cool upon emplacement. Grain boundaries are either curvilinear or display delicate apophyses and re-entrants (Plates N1F, N2B, E). Curvilinear grain boundaries are fracture surfaces that imply fragmentation by contraction during thermal shock and disruption of magma, and possibly solid lava, by steam explosions. The morphology of these grains is nearly identical to that described from tuffs generated by phreatomagmatic explosions (Heiken, 1971, 1974; Honnorez and Kirst, 1975; Self, et al., 1980). Grains bounded by delicate re-entrants and less commonly, bubble walls (Plates N2B, E) are morphologically similar to cinder produced by Strombolian eruptions of present-day volcanoes (Self, et al., 1980, Walker and Croasdale, 1972).

Block-size pyroclasts are present as bombs and breccia fragments. Bombs were recognized as such on the basis of their morphologic similarity to bombs from recent eruptions rather than inferring their temperature upon eruption (cf. MacDonald, 1972). A variety of bomb types occur, commonly as subordinate fragments over a thick stratigraphic section, and only rarely do they compose more than 20% of a bed (i.e. agglomerates, sensu stricto) (Plate N2A). Cored bombs are common and range in size from block-size to medium-grained lapilli-size, and display a complex history of formation (Plate N2A, C). Bombs are composed of an angular core fragment surrounded by a cortex of vesicular porphyry that displays concentric flattening of vesicles, and also contains pyroclasts agglutinated

Figure 7-4. Examples of cored bombs, sketched from 35mm slides.

- A. 1-Xenolith of East River Formation dolostone forms head of bomb. 2-First tail composed of layers of ash and lava. 3-Feldspar-phyrlic vesicular porphyry forms second tail of bomb. 4-Pyroclastic lapillus agglutinated in tail of bomb. Elongation of tail suggests that the dolostone fragment was the aerodynamic head of the bomb. During flight, syrupy lava collected on the tail and changed the centre of gravity. The bomb then fell into its present orientation, as suggested by the elongate tendrils of lava that parallel stratigraphic tops (arrow).
- B. 1-Bomb core composed of broken older bomb composed of core of dense porphyry (A) and cortex of lava (B), dashed line shows colour banding that is parallel to the bomb margin. 2-Agglutinated ash and lapilli pyroclasts. 3-Irregular bomb cortex of microlitic vesicular lava. 4-Accessory lithic fragment of chert. Note the asymmetry of the bomb cortex and current shadow marked by that side of the bomb that is thinner. Although the sample comes from float, the bomb is inferred to have traveled from left to right, based on orientation of vesicles, bomb asymmetry, and direction of elongation of lava tendrils.

during bomb travel (Plate N2A; Fig. 7-4A,B). Several bombs display an asymmetric cortex formed by aerodynamic moulding of the bomb surface (achneliths) (Fig. 7-4). Bomb cores are inferred to have been derived from a variety of sources: those composed of sedimentary rock were derived from country rock; dense, slightly vesicular porphyry are fragments of solidified lava from within the volcanic conduit; lapilli tuff was derived from crater walls as were fragments of older cored bombs (Fig. 7-4A). Slabs of broken ribbon bombs and fusiform spindle bombs (Plate N2D) also occur, but are not as common as cored bombs.

Block-size breccia fragments are locally abundant and commonly occur in the same bed with cored bombs (i.e. agglomerates) (Plate N2A). The blocks are composed of highly vesicular (scoriaceous) to relatively dense lava with few vesicles. Many of the blocks display rims that contain sparse vesicles that increase in abundance and size inwards from the block margin. This pattern of vesicles is commonly truncated on one or more sides of a block by a fracture surface (Plate N2C). The vesicles in all cases are symmetric and circular in shape; no elongate vesicles were observed, which implies that vesicles developed when the lava was still relatively liquid (i.e. low viscosity).

Pyroclast Composition and Alteration

Pyroclastic rocks comprise lithic lapilli tuffs to ash tuffs with subordinate lithic tuff breccia and agglomerate;

crystal-lithic tuffs also occur but are comparatively rare. The most abundant component of the tuffs are juvenile volcanic fragments that display slightly vesicular to pumiceous, and less commonly, scoriaceous textures. The fragments are composed of porphyry that contains sparse phenocrysts of potassic feldspar, plagioclase, and a chloritized mafic mineral (olivine? pyroxene?) in an aphanitic groundmass of 5-20% plagioclase microlites (An₂₀₋₃₀) (Plate N2G). Differences in groundmass mineralogy allows distinction of dark-coloured fragments, characterized by a groundmass rich in opaque oxides and subordinate chlorite, and pale green-coloured clasts that have a groundmass composed exclusively of chlorite (Plate N2B, E). In addition to differences in groundmass mineralogy, the dark-coloured clasts are characterized by a greater proportion of microlites, phenocrysts, and vesicles than the pale-coloured clasts. In both clast types, the plagioclase microlites display a hyalopilitic to pilotaxitic and trachytic texture (terminology of Williams, et al., 1982); glomeroblastic clots of microlites and swallowtail terminations (cf. Lofgren, 1980) to microlites are also common (Plate N2G). Potassium feldspar phenocrysts display a micropoikilitic core composed of an irregular network of isotropic material and chlorite and an inclusion-free rim (Plate N2G). The composition of the phenocrysts could not be optically determined because of the abundant fine inclusions of limonite, that cause grains to appear to be nearly isotropic, and the common partial to complete replacement of phenocrysts by carbonate. Vesicles in the

groundmass are filled with chlorite, carbonate (calcite, dolomite, and/or ankerite), and rarely quartz. Accessory lithic fragments are composed of angular fragments of sedimentary country rock. Rare cognate lithic fragments display a subophitic to diabasic texture and are composed of plagioclase, oxides, and inferred olivine pseudomorphs. The columnar jointed mafic porphyry, which appeared in the field to be the freshest rock encountered, displays extensive alteration. Clear, nonpleochroic amphibole, plagioclase, and clinopyroxene are sparse relict minerals; porphyroblastic epidote is common in addition to chlorite, sphene, pumpellyite, white mica, and carbonate.

All mafic rocks have been affected by post-depositional alteration that is particularly noticeable in the originally more porous pyroclastic rocks. Alteration can be recognized in the field by colour changes that reflect variations in mineralogy. Relatively fresh tuffs are dark green to black but grade into progressively lighter-coloured pinkish to tan and white tuffs (compare Plates N2B and N2C; also N2E). In several occurrences of pyroclastic rock, the colour change is concentric with respect to the contact with country rock (Fig. 7-6). The dark green colour of the least-altered tuffs is imparted by chlorite and celadonite(?) that occur as alteration products and vesicle-fill phases. Light-coloured tuffs are composed dominantly of carbonate and limonite; the latter appears to be a replacement of chlorite. Carbonate (dolomite, ankerite, and less common calcite) occurs as discrete euhedral

rhombs (<10%) in relatively fresh tuffs and as anhedral poikilotopic clots that constitute up to 90% of severely altered tuffs. The relationship of the alteration to the tuff-country rock contact (Fig. 7-6) implies that it took place after deposition of the tuffs, possibly as a consequence of the interaction of tephra with sea water.

The variation in pyroclast colour and groundmass mineralogy is inferred to reflect different crystallization histories for the two clast types. The dark clasts, which contain a greater abundance of microlites, vesicles, and opaque grains in the groundmass, represent a greater degree of crystallization prior to fragmentation. This conclusion is based on the inferences that: 1) the opaques in the groundmass are near-solidus exsolution products from a partly crystalline lava, 2) the greater abundance of vesicles is a reflection of more complete gas exsolution from the melt, and 3) the greater abundance of microlites is an indication of more complete crystallization. In contrast, lighter-coloured clasts did not undergo as complete crystallization before fragmentation. The wholly chloritic groundmass characteristic of these clasts formed from recrystallization of relatively glass-rich fragments (palagonite as a precursor?). Thus, dark-coloured clasts formed during the fragmentation of partly solidified magma within the vent whereas pale clasts formed by disruption of relatively fluid lava. Some authors refer to the former type

form during the same eruption that caused fragmentation, whereas the lighter-coloured clasts would be referred to as "juvenile" because they apparently formed synchronous with eruption. A strikingly similar variation in clast colour, mineralogy, and textures has been observed by the author in hand samples and thin sections of basaltic tuff from the Pleistocene Menan Buttes tuff cones of Idaho (cf. Hamilton and Myers, 1963).

Discussion

Contact relationships between pyroclastic rocks and adjacent sedimentary country rocks suggest pre- and syneruptive disruption of the country rock. Pre-eruptive doming, commonly observed prior to eruptions of modern volcanoes (MacDonald, 1972; Williams and McBirney, 1979; Black, 1974) formed a tensile stress field that caused boudinage of partly lithified sediments, brittle fracture of lithified layers, and formation of arcuate, high-angle fractures. Syneruptive collapse along these fractures (cf. Lorenz, 1973) created the steep contacts between tuffs and country rock (Plate N1A) and probably was also partly responsible for the formation of roughly circular(?) depressions that controlled cone shape and preservation. Explosive stoping of wall rock by phreatomagmatic blasts during eruptions, is evidenced by the incorporation of wall rock into the tuffs, and locally intrusive relationships between tuffs and wall rock and probably obscured

recognition of fault planes (cf. Self, et al., 1980; Lorenz, 1974). Both fluidization during eruptions and explosive stopping and excavation of wallrock could have contributed to post-eruptive subsidence.

Paleobathymetry inferred for sedimentary rocks associated with the tuffs (subtidal to intertidal; Fig. 7-2) implies that water was likely an important component of eruptions. Water- and/or steam-rich eruptive clouds are relatively viscous, compared to dry eruption columns. This results in airfall deposits that are poorly sorted as a consequence of "ash scavenging" by water (Walker, 1981). Eruption clouds of this type would be closer to mud in consistency compared to less dense eruption columns inflated by magmatic gases and superheated steam. In addition to formation of poorly sorted and stratified airfall deposits, the high viscosity and density of water-rich eruption clouds favours formation of pyroclastic flows with high cohesive strength that are essentially debris flows (cf. Middleton and Hampton, 1973; Lowe, 1982). Thus it is considered probable that there may be considerable overlap, in terms of structures and textures, between pyroclastic airfall and flows formed during water-rich (hydrovolcanic) eruptions.

Tuff beds in the Narakay Volcanic Complex are interpreted to have been deposited both by pyroclastic airfall and pyroclastic flow. The lack of vesiculated tuffs (cf. Lorenz, 1973), in addition to lack of welding and vapour phase crystallization, imply emplacement of relatively cool pyroclastic

ejecta, possibly at temperatures of $<100^{\circ}\text{C}$. Unequivocal airfall deposits display bomb-sag features; airfall tuff deposited from lower viscosity eruption clouds (due to less water and/or a higher content of magmatic gas and superheated steam) display crude parallel stratification, better sorting, and tighter packing of the framework (cf. Plate N1D). Inferred pyroclastic flow deposits display "deflation bedding" (Plate N1E) formed during gas escape after flow deposition (cf. Sheridan, 1979, p.131). In addition, some of the beds with "out-sized" clasts that float in a finer grained matrix and beds with reverse graded bedding (Plate N1C) may have been deposited by pyroclastic flows. Thick ($>2\text{m}$ thick) massive tuff beds are enigmatic; they may have been deposited by pyroclastic flows, airfall from a water-rich cloud, or by blasts of mud-rich material cleared from vents analogous, perhaps, to cypressoid jets of tephra-rich steam and/or water commonly observed in modern hydrovolcanic eruptions (Thorarinsson, 1966; Wohletz and Sheridan, 1983).

The occurrence, morphology, and distribution of bombs within pyroclastic rocks are important with respect to characterization of eruptive style. Cored bombs, as described above, form when a pyroclast does not clear the vent and falls back into ponded or fountaining lava within the vent and receives a coating of lava. Their occurrence within pyroclastic successions indicates that eruptions varied from hydromagmatic to Strombolian. The two eruptive styles could have been contemporaneous; modern volcanoes can display both phreatomagmatic

and Strombolian behavior synchronously (Self, et al., 1980). Fusiform bombs and ribbon bombs also are products of Strombolian eruptions. In conjunction with the morphology of lapilli ash, a phreatomagmatic (hydrovolcanic) to Strombolian eruptive style, that was largely subaerial, can be inferred. Because of the mixture of products typical of Strombolian eruptions with those of hydrovolcanic eruptions, the eruptive style is inferred to be transitional between Surtseyan and Strombolian depending on the water:magma ratio (cf. Wohletz, and Sheridan, 1983). Certain features considered to be typical of each of these eruption types (cf. Williams and McBirney, 1979; MacDonald, 1972) are, however, absent such as lava flows (typical of Strombolian eruptions) and an abundance of finely comminuted ash and generally fine-grain size of tuff deposits (typical of Surtseyan eruptions).

Based on the previous discussions an ideal eruptive cycle would consist of (Fig. 7-5):

- 1) Pre-eruptive tumescence during magma ascent and possibly gas pressure build-up (due to gas exsolution from the magma during decompression and/or steam generation) (cf. Woollsley, et al., 1975) created a tensile stress field. This resulted in boudinage of partly lithified sedimentary rocks, brecciation of lithified units, and formation of a system of arcuate fractures and faults.

- 2) Eruptions were initiated when water encountered hot rock, either hot country rock and/or magma. Water gained access to the hot rock through fractures formed during pre-

FIGURE 7-5: Cartoon of eruptive cycle:

- A. Pre-eruptive doming during ascent of mafic magma causes fracturing.**
- B. Steam explosions occur as magma comes into contact with water. Both wall rock and magma are disrupted by explosions. Gas escape and eruption causes subsidence along fractures formed during A.**
- C. Continued eruption infills depression and builds subaerial tuff cone during hydrovolcanic and Strombolian eruptions.**
- D. After eruption ceases, topographically positive parts of tuff cone are eroded and epiclastic debris is swept into shallow marine environment.**

Question mark in throat of cone denotes uncertainty in composition of plugged conduit. Massive tuff and solidified magma are possibilities.

A.

B.

C.

D.

4

eruptive tumescence and through permeable sedimentary aquifers (e.g. oolite beds). The rapid volume increase when water flashes to steam causes tremendous phreatomagmatic explosions that disrupt both country rock and magma. Brecciation of previously overlying and presently adjacent wall rock formed lapilli- to block-size sedimentary rock fragments that were extruded along with volcanic rock fragments; larger blocks of brecciated wall rock remained in situ and were engulfed in the tuffs. Curiously no phreatic breccia (cf. Wohletz and Sheridan, 1983) was deposited.

3) Steam- and/or water-rich clouds of pyroclastic debris deposited poorly sorted and massive to crudely stratified tephra that built up rapidly into a subaerial tuff cone (sensu stricto; Wohletz and Sheridan, 1983). The morphology of the pyroclastic grains suggests that although much of the eruptive activity was a consequence of hydrovolcanic explosions a significant amount of the activity was Strombolian. This is suggested by the occurrence of lapilli of cinder, fusiform and ribbon bombs, and breccia fragments of scoria. The mixture of light- and dark-coloured clasts implies brecciation of solidified lava (dark clasts) within the conduit by steam explosions generated during the ascent of new batches of magma (light clasts). In at least one area, subsequent explosions were not of sufficient intensity to remove ponded lava (columnar-jointed porphyry in Plate N1B). Variations in grain size, sorting, and framework packing of tuff beds are attributed to variations in eruptive intensity,

pyroclast:fluid ratio, and the ratio of water to magmatic gases and superheated steam in the eruption cloud. Deposition of relatively cool tephra occurred on the flanks of subaerial tuff cones by pyroclastic airfall and less common pyroclastic flow.

4) Following cessation of an eruptive cycle, topographically positive portions of tuff cones were removed by erosion and redistributed in the marine environment. This involved surface runoff (rainwater and wave spray) and wave and storm-wave destruction of the cone and redistribution as an apron of epiclastic debris (to be discussed later). Preservation of tuff cones is a function of syn- to post-eruption subsidence history and formation of synvolcanic depressions.

Two features, commonly associated with tuff cones and maar volcanoes formed by hydrovolcanic eruptions, but absent from the Narakay Volcanic Complex, are accretionary lapilli and base surge deposits. Accretionary lapilli are considered typical of subaerial eruptions in which some water, either rainwater and/or vaporized groundwater, was present during the eruption (Moore and Peck, 1965; Fisher, 1981). Their absence from tuffs of the Narakay Volcanic Complex is due to the high content of water in the eruption column; accretionary lapilli would be so dense that they would be destroyed on impact. Walker (1982) offers a similar explanation for the lack of accretionary lapilli in silicic phreatoplinian tuffs from the Taupo Volcanic Zone of New Zealand.

The lack of base surge and pyroclastic deposits is more

puzzling. As inferred in the following section, some surge deposits are in fact present in the Narakay Volcanic Complex but their abundance is much lower than is typical for tuff cones and maars (Wohletz and Sheridan, 1983; Lorenz, 1974a). Base surge associated with hydrovolcanic explosions comprises a laterally spreading "toroid" of steam and ash. In shallow marine environments (e.g. Taal, Phillipines) the surge cloud spreads out laterally across the top of the water surface; deposition of cross-laminated tuffs, typical of base surge, only occurs if a subaerial substrate is present, otherwise "base surge deposits" in such an environment would consist of fall-out through the water column after cessation of surge movement. An additional factor that must be considered, as first pointed out by Crowe and Fisher (1973) and later by Schminke, et al. (1973), is that the formation of traction current structures in surge clouds depends on the viscosity of the cloud. The high viscosity and cohesive character of water- and/or steam-rich ash clouds may preclude the formation of traction current structures typical of base surge and pyroclastic surge deposits.

Reconstruction of Cones

Reconstruction of original cone sizes, aspect ratios, and recognition of vent centres is hampered by incomplete preservation of cones and deformation of primary dips by compaction, slumping, and post-depositional faults. By considering the changes in the dip of bedding, pattern of alteration of tuffs,

FIGURE 7-6: Inferred vent centres on main island of the Narakays. Vents are labeled from oldest to youngest, VC1, VC2, and VC3.

the presence of intrusive breccia, and cone plugs, three eruptive centres have been inferred (Fig. 7-6); all occur on the largest island within the Narakays. The oldest cone in this area appears to have been contemporaneous with lower member E2 of the East River Formation (Fig. 7-2); cone remnants on Bluff Island (next section) are the oldest within the Narakay Volcanic Complex, as they occur within Lady Nye Formation arenites. The location of the vent centre (VC1; Fig. 7-6) is inferred on the basis of bedding dips (away from the centre), patterns of alteration, and the presence of a distinct intrusion breccia (described in later section). The intrusion breccia is a lopolith-shaped body that may have formed during sub-volcanic explosive activity. If all of the tuff beds on the eastern limb of the island (Fig. 7-6) are considered part of the same cone (they occur at a similar stratigraphic position), it would imply a crater with a diameter of almost 3 kilometres. On the western limb of the island, bedding relationships suggest smaller cones with a diameter of a kilometer or less. As discussed previously, tuffs adjacent to the columnar-jointed mafic porphyry define the inner margin of a small tuff cone (VC3); it appears from regional relationships that this cone is nested within an older sequence of tuffs and may represent a parasitic cone developed on the flanks of the larger edifice (VC2).

Bluff Island Section

An island west of the largest central island of the

FIGURE 7-7: Simplified section through pyroclastic and epiclastic deposits exposed on Bluff Island. Facies, composition, and sedimentary structures are shown to the right of the column and inferred processes on the left.

Narakays (informally referred to as Bluff Island) contains superb exposures of volcanoclastic rocks. These rocks record formation of a syneruptive fan-shaped wedge of mixed epiclastic and pyroclastic detritus developed during progradation of a cone flank. The entire section is more than 50m thick and comprises five different facies (Fig. 7-9); the section is underlain and overlain by feldspathic arenites of Lady Nye Formation (member C4; lithofacies D). Most of the volcanoclastic rocks are composed exclusively of volcanogenic fragments typical of mafic pyroclastic tuffs elsewhere in the Narakay Volcanic Complex, and also display a similar range in colour and mineralogy. Although the sequence of volcanoclastic rocks gradationally overlies Lady Nye Formation arenites, when the contact is viewed from a distance (500m) the bedding in the two units is discordant and intersects at a low angle ($<10^\circ$). Significantly, the bedding in the volcanoclastic rocks dips in the same direction as the direction (southwest) of paleoflow (and progradation) inferred from crossbed measurements and vertical lithofacies succession within the section.

1) Parallel-Laminated, Tuffaceous Arenite Facies: This facies marks the base of the section and records the gradual introduction of increasing amounts of volcanic detritus into a subtidal marine environment; all facies that overlie this facies are composed exclusively of volcanic fragments. Rocks of this facies display a distinct tan colouration, as a result of alteration of volcanic detritus to limonite, that distinguishes them from the typically gray arenites of the

underlying Lady Nye Formation. This facies is composed of 10-30cm thick sets with parallel lamination; sets are separated by undulose scour surfaces.

2) Graded-Laminated Facies: This facies is about 18m thick and is composed of two types of beds: massive to graded coarse ash to lapilli tuff beds and laminated fine ash tuff beds (Plate N3A). Massive and graded beds (<10cm) are interlayered with 2-10cm thick beds of laminated fine ash that display parallel lamination and ripple cross-lamination; the tops of some beds have been moulded into wave ripples. Thicker beds of laminated ash tuff display climbing ripple (type A and B, Jopling and Walker, 1969) stratification.

3) Crossbedded Facies: This facies is characterized by well-developed cross-stratification, although parallel-laminated and massive beds also occur. The average bed thickness and grain size is greater than that in the underlying facies; the contact between the crossbedded facies and the overlying facies is covered, but is presumed to be gradational. The lower part of this facies is composed of interbedded cross-laminated lapillistone and massive fine to coarse lapillistone. Massive to parallel-laminated beds are tabular sheets, up to 2m thick, composed of lapilli-size clasts and local concentrations of coarse lapilli and blocks/bombs (Plate N3E). Fragments display a pristine, unabraded surface morphology with delicate apophyses along the boundaries. Cored bombs are scattered throughout the coarser-grained beds (Plate N3E).

A variety of cross-stratification styles are present within

this facies; parallel-stratification is present but is subordinate. Tabular-planar to shallow trough crossbeds with low-angle ($<15^\circ$) to asymptotic foreset stratification are most common (Plate N3D). Foreset laminae may be composed of a single grain size (fine lapilli to coarse ash) or may display normal grading and rare reverse grading. Concave-down laminae also occur, and in one exposure climbing tabular-planar mega-ripples, in 50cm thick sets, was observed (Plate N3B). Although most beds are composed of lapilli-size clasts, fine lapilli- to ash-size layers were also observed. These display parallel lamination, type A climbing ripple stratification, and chute and pool structures with backset bedding (Plate N3C). Paleocurrents inferred from more than 40 measurements are strongly unimodal to the southwest, parallel to the dip of bedding within the entire sequence (Fig. 7-9).

4) Lenticular and Wavy Bedded Facies: This facies occurs near the top of the section; it is truncated and scoured by the overlying channel-fill facies. The facies is composed of interbedded coarse- to fine-grained arenite and silty mudstone (Plate N3F). Granulestones and arenites occur as 5-15cm thick beds that display normal grading, and the top surface of the bed is commonly scoured. Silty mudstones display parallel to ripple cross-lamination and contain lenses and "starved ripples" of cross-laminated fine-grained arenite.

5) Channel-Fill Facies: This facies is composed of granulestones and conglomerates that fill a deep ($>10m$ deep) scour in the underlying lenticular-bedded facies. The channel-fill is

composed of stacked sheets of (<1m thick) of massive and normal graded paraconglomerate. Although composed entirely of volcanic fragments, this facies contrasts with underlying facies in that the clasts are all well-rounded and the framework is tightly-packed.

Interpretation

The Bluff Island section is interpreted to record episodic deposition of volcanic detritus during a single eruptive cycle, composed of numerous explosive events. The vertical sequence of facies records a change in the depositional mechanism as a function of decreased water depth (i.e shoaling) and progradation of cone flanks during eruption. Parallel-laminated tuffaceous arenite at the base of the sequence records mixing of "background" arenites and tuffaceous detritus deposited as pyroclastic debris settled through the water column and was partly reworked by marine currents. The graded-laminated facies records deposition of lapillistone-ash tuff couplets by pulses of sediment gravity flows and settling of tephra through the water column. These deposits represent Bouma-like Ta-Td couplets; ripple cross-laminated beds are analogous to Bouma Tc beds. Wave rippled top surfaces of some of these beds probably formed during scour by marine currents; thus some of the beds inferred to be Bouma Tc and Td beds may have formed by traction currents associated with marine currents. Unabraded surface textures, typical of framework grains, implies limited reworking of erupted material prior to

FIGURE 7-8: Schematic facies model to illustrate inferred lateral relationships between lower three facies of the Bluff Island section.

deposition. Each graded-laminated couplet records deposition of a subaqueous pyroclastic flow ("pyroturbidites", Williams and McBirney, 1979), inferred on the basis of the pristine surfaces of the framework grains, the lack of evidence of reworking of detritus after deposition (as might be expected if units were deposited in an agitated marine setting), and the similarity of the depositional sequence to other inferred subaqueous pyroclastic flows (cf. Tasse, et al., 1978). The gradual coarsening- and thickening-upwards of this facies into the overlying crossbedded facies reflects progradation of proximal facies over relatively distal facies, and thus is inferred to record cone growth.

Crossbedded lapillistones and lapilli tuffs in the cross-bedded facies formed during unidirectional transport of pyroclastic debris by high-energy, upper flow regime bedforms developed on a depositional surface that had shoaled from subtidal into intertidal to subaerial(?). Low-angle foreset stratification that makes an asymptotic/tangential contact with the base of the crossbed set forms under conditions of high-energy turbulent flow (Jopling, 1965). Climbing ripple and particularly megaripple structures record conditions of rapid deposition from a sediment-charged fluid. Chute and pool structures and backset bedding imply deposition from a fluid in the upper flow regime, although the exact flow conditions responsible for the formation of chute and pool structures is uncertain (Schminke, et al., 1973). The complete absence of particle abrasion and the similarity of contained sedimentary

structures to those in some base surge deposits (cf. Wohletz and Sheridan, 1979; Schminke, et al., 1973; Waters and Fisher, 1971; Crowe and Fisher, 1973), particularly the occurrence of chute and pool structures and backset bedding, implies that much of the crossbedded facies was deposited by pyroclastic surges. Individual surge events may have been base surges (sensu stricto, Fisher, 1979) that accompany phreatomagmatic eruptions, or they may have been ground surges that accompany the passage of pyroclastic flows.

Assuming that the vertical succession from graded-laminated facies into crossbedded facies records lateral facies distribution (Middleton, 1973), then the crossbedded facies records subaerial ground surge associated with the passage of pyroclastic flows that entered the water and were deposited as subaqueous pyroturbidites (Fig. 7-8). Another possibility for deposition of some of the cross-laminated units is by upper flow regime currents associated with return (ebb) flow of waves and storm waves, a hypothesis that is consistent with the occurrence of hummocky cross-stratified arenites at the base of the section. Lapilli tuffs with a massive to faint parallel stratification could have been deposited by pyroclastic flows and/or airfall from water-rich eruption clouds.

The bedding style in the wavy bedded facies resembles that formed by tidal and wave action (cf. Reineck and Singh, 1973; de Raaf, et al., 1977). This facies is therefore interpreted to record traction current reworking of cone flanks during the waning phases of eruptive activity. Massive and graded beds of granulestone and arenite were deposited by sediment gravity

flow processes; sediment gravity flows are inferred to have been generated by the return flow of waves (cf. Nelson, 1982). The top of each bed was scoured before burial by traction current deposits of parallel- to ripple cross-laminated lenticular-bedded silty mudstone. The channel-fill facies was also deposited by sediment gravity flows but the framework grains clearly underwent reworking prior to transport. The channel is inferred to have formed by rip-current scour of the flanks of the cone (Fig. 7-10) during cone destruction after the cessation of eruptive activity.

Epiclastic Facies

Epiclastic volcanoclastic rocks formed during syn- and post-eruptive destruction of mafic tuff cones occur sporadically throughout the stratigraphic sequence of the Narakay Volcanic Complex (Fig. 7-2). The term "epiclastic" is applied to any surficial deposit of volcanoclastic material in which the depositional mechanism was not a direct consequence of a volcanic eruption (Fisher, 1981). The depositional mechanism for epiclastic and pyroclastic rocks may, in some cases, be the same (suspension deposition and sediment gravity flow deposition) and the textures produced during deposition will be similar. Distinction of an epiclastic from pyroclastic origin for a particular unit can be difficult, if not impossible, especially in proximal volcanic environments. In the Narakay Volcanic Complex the two types of deposits were distinguished on the basis of clast composition (freshness,

roundness, amount of admixed non-volcanogenic grains), bed geometry, stratification styles and stratigraphic setting. Some of the epiclastic rocks were deposited during and after known pyroclastic events but in other cases epiclastic rocks cannot be linked to known pyroclastic events (Fig. 7-2) which implies the existence of vents that are not exposed.

Epiclastic rocks occur as tabular to lenticular (channel- and scour-fill) bodies up to 14m thick. Proximal, distal, and "mixed" facies have been recognized; there appears to be a complete gradation between proximal and distal facies.

Proximal Facies

Proximal epiclastic rocks comprise lenticular channel-fill units that are inferred to grade down-current (and laterally?) into tabular-shaped distal facies. The channels are broad (up to 400m wide) structures cut into the underlying sedimentary rocks (Plate N4A). They display a distinct sequence of lithologies and sedimentary structures (Fig. 7-9).

The basal subdivision of the channel-fill, onlaps the walls of the channel (Plate N4B), and consists of parallel-laminated, limonitic, granule-bearing arenite, composed of moderately sorted volcanic arenite and mudstone with single-grain thick laminae of granule-size clasts. In some sections, small-scale (<15cm) cut-and-fill structures and ripple cross-lamination occurs within this subdivision. The parallel-laminated subdivision is abruptly overlain by massive to crudely graded volcanic paraconglomerate (Plate N4C). Soft-sediment

FIGURE 7-9: Schematic illustration of complete channel-fill sequence in proximal epiclastic facies.

deformation of the arenites below the conglomerate division is common, and apparently was produced by loading during rapid deposition of the conglomerate. Basal volcanic paraconglomerate occurs in tabular beds, 50-200cm thick, composed of unsorted and angular clasts dispersed in a mud-rich matrix. Beds are massive or display normal and less common reverse grading; large (5-10cm wide) clasts commonly float in a finer grained matrix at the top of a bed (Plate N4C). Clasts within the paraconglomerates are virtually indistinguishable, with respect to composition, surface texture, and proportion of juvenile to accessory fragments, from those found in pyroclastic tuff breccia. The basal paraconglomerates are overlain by granulestones and conglomerates that comprise 50-150cm thick trough-shaped beds that display normal grading of 5-15cm thick foreset strata. Compared to the underlying conglomerates, clasts within the trough-structures are finer-grained, display tighter packing, are rounded (subangular to subrounded), and display more advanced alteration of mafic porphyry clasts to limonite and carbonate. The cement in these conglomerates commonly is a sparry Fe-rich dolomite and/or ankerite. The trough bedded division is overlain by massive and tabular-planar crossbedded granulestones composed of moderately sorted and rounded volcanic clasts. The top surface of the channel-fill is scoured (up to 2m relief, although 50cm or less is most common) and overlain by stromatolites. In general, stromatolites occur at the top of almost every epiclastic bed in the Naray Volcanic Complex regardless of whether it is of

proximal or distal facies.

Paleocurrents inferred from bedding surface exposures of the trough structures are unimodal. Those inferred for the crossbedded granulestones are commonly asymmetric bimodal-bipolar along an axis that is parallel to the paleoflow azimuth for the underlying conglomerates. The mode with the greatest magnitude is in the same direction as that inferred for the underlying conglomerates.

Distal Facies

Epiclastic rocks inferred to have been deposited farther from the vent are characterized by finer grain size, thinner beds, and a tabular, rather than lenticular, shape to the deposit. Individual sheets range from 1-3m thick and are composed of beds that range from 5-50cm thick. Many of the sheets display a sequence of structures that is similar to that described from the proximal facies: a) parallel- and ripple cross-laminated granular arenite commonly with soft-sediment deformation; b) massive to graded granulestone and conglomerate, c) crossbedded granulestone, and d) stromatolites. Coarse paraconglomerate and trough-bedded structures are absent. Sheets also are composed exclusively of stacked turbidite beds, 5-15cm thick, that display Bouma Ta-Td, and less common Ta-Tc-Td, cycles (Plate N4E). Some beds display complete Bouma cycles less than 20cm thick. The tops of sheets are scoured, with up to 1.5m of synoptic relief, and overlain by stromatolites (Plate N4F). Sheets composed of stacked Bouma

beds resemble subaqueous pyroclastic flows; they are inferred to be epiclastic because they contain admixed quartz and rare microcline feldspar (derived from in situ marine clastic sediments), and volcanic framework grains display moderate rounding.

"Mixed" Facies

A third end-member facies in the spectrum of epiclastic rocks are "hybrid" rocks composed of reworked pyroclastic fragments and "background" marine detritus. These rocks commonly display structures formed by "background" marine currents although sediment gravity flow deposits also occur.

The mixed facies ranges in grain size from conglomerate to mudstone and is composed of variable proportions of volcanic and non-volcanic detritus. Volcanic clasts are commonly rounded and have a greenish to dark brown colour. Green-coloured clasts are composed of carbonate and acicular needles of green mica (probably celadonite); brown clasts are composed of limonite and carbonate. Admixed non-volcanogenic grains include mudchips, quartz and feldspar sand grains (Lady Nye Formation), quartz sand and ooid grains (East River Formation), and quartz and feldspar sand grains in the Kaertok Formation.

Beds of mixed facies are rare in Lady Nye Formation, and are most common in upper East River Formation and lower Kaertok Formation. They occur sporadically throughout the respective formations, and individual deposits do not display an orderly vertical sequence of structures and associations

(cf. proximal and distal facies). The most common occurrence of mixed facies rocks are wavy to lenticular-bedded units (upper East River Formation) and massive arenaceous beds (Kaertok Formation). Wavy to lenticular beds are composed of 2-5cm thick beds of massive to cross-laminated oolitic granulestone to arenite intercalated with beds (1-3cm thick) of wavy- to parallel-laminated tuffaceous siltstone to mudstone. In the Kaertok Formation, the mixed facies occurs in massive sheets (20-100cm thick) of quartzo-feldspathic arenite with 20-50% volcanogenic grains in the framework; convolute bedding is common within these beds.

Depositional Model

Any model to explain proximal-distal epiclastic facies must be consistent with the sedimentary environment proposed for the associated non-volcanogenic rocks. Sedimentary rocks of the Narakay Islands have been affected, to varying degrees, by wave, storm-wave, and tidal currents. The influence of topographically positive parts of tuff cones on the hydrodynamic character of background marine currents provides a framework in which to interpret the sequence of structures and inferred depositional processes in epiclastic rocks.

Inferred proximal deposits infill channels formed by rip currents and undertow generated during the interaction of incoming waves and the tuff cone. The lack of features indicative of subaerial exposure (calcretes, karstic dissolution surfaces) of the channel walls is negative evidence that

Figure 7-10. Model for cone destruction and dispersal of epiclastic detritus (not drawn to scale). "Channelized sediment flow" indicates location of proximal channel-fill facies developed in rip channel. "Unchannelized sediment flow" is represented by distal facies. Marine currents are indicated by two-headed arrows.

supports the idea that channels were entirely subaqueous. Tuff cones and maar volcanoes commonly have topographically irregular flanks characterized by undulose or swaley topography and saddles (cf. Thorarinnson, 1964). Such irregularities could act to control the location and intensity of wave and storm-surge run-off in two ways. Swales and gullies on the cone surface would control the location and intensity of wave (and rainwater) run-off. Breaches and sea level saddles in the cone flanks would also contribute significantly to the generation of powerful ebb currents. Storm-surge and rip currents in modern coastal areas are generated when water is "piled-up" along the beach by waves and/or wind (Davis, 1978; Morton, 1982). When the hydraulic head associated with these build-ups exceeds the ability of onshore wind and/or incoming waves to contain it, strong rip currents and storm-surge-ebb currents are generated. During periods of wave attack and wind activity, water could enter the cone through breaches and saddles and be "piled-up" inside the cone. Currents generated during water escape and surge through breaches and saddles could be of sufficient intensity to cause erosion of channels observed in proximal epiclastic facies in the Narakay Volcanic Complex.

The lower subdivision within channel-fill sequences onlaps the walls of the channel and imply that the channel was "oversized" with respect to subsequent flow volume. Parallel-laminated granule-bearing arenites at the base of channel-fill sequences are inferred to have been deposited by lower flow regime(?) currents in the channel thalweg. The occurrence of

intraclastic small-scale scour-fill features within this subdivision is further evidence that flow in the channels was "underfit" with respect to overall channel size. Coarse-grained paraconglomerates at the base of the conglomeratic subdivision are interpreted as cohesive, high-density sediment gravity flow deposits of unreworked pyroclastic debris that was swept off cone flanks. The overlying trough-shaped beds record infill of shallow scoop-like scours by turbiditic pulses that deposited the graded foreset stratification found in the troughs. Pyroclastic debris had, by this time, undergone some reworking as suggested by the roundness of fragments within these beds. Increased amounts of reworked volcanoclastic debris was delivered to the channel system as uppermost deposits display a well-rounded texture. Unimodal paleocurrents inferred from the orientation of trough structures imply unidirectional currents, inferred to be the result of wave- and stormwave-generated surge and/or rip currents. The crossbedded facies at the top of many sequences displays bimodal-bipolar paleocurrents that suggest deposition by alternating onshore-offshore currents that represent tidal and/or wave surge back and forth within the channels. The shape of the crossbed sets in this subdivision implies deposition by a sandwave-type of bedform (cf. Harms, et al., 1982). The scour surface at the top of the sequence was generated by high-energy background marine currents prior to growth of stromatolites with distinct subtidal to intertidal morphology (cf. Logan, et al., 1974). The affinity of stromatolites for the

top of epiclastic sequences is due to the increased substrate rugosity of these deposits and their topographic elevation above the sea floor (cf. Ginsburg, 1983).

Inferred distal facies record non-channelized sediment gravity flow deposition. Those deposits that display a similar sequence of structures as the proximal facies, but differ with respect to the shape of the deposit and grain size, are inferred to have been deposited by similar processes, i.e. sediment gravity flows of different density and abundance of unworked versus reworked pyroclasts. Those units composed exclusively of stacked Bouma cycles record deposition from low-concentration turbidity currents (cf. Lowe, 1982). The significance of two distinct styles of distal epiclastic deposition is unknown.

The mixed facies records deposition of epiclastic volcanic detritus subjected to thorough reworking and mixing with autochthonous non-volcanogenic marine detritus. Sedimentary structures reflect depositional processes of the contemporaneous marine environment. Epiclastic rocks in upper East River Formation contain structures formed by tide and wave action; epiclastic rocks in the Kaertok Formation were deposited by water-rich turbulent suspensions possibly related to storm activity. Individual beds of epiclastic detritus cannot be correlated between islands (sections), which implies an irregular distribution of volcanic fragments on the sea floor. Volcaniclastic material was introduced into the marine environment by pyroclastic processes (airfall and flows) and by

ebb and rip currents that swept debris off of the cone flanks. Many fragments remained active for a considerable amount of time as suggested by the well-rounded and sorted texture of many deposits and the occurrence of volcanic fragments with an oolitic cortex (superficial ooids).

Rhyolites and Felsic Pyroclastic Rocks

The upper part of the Narakay Volcanic Complex contains the products of felsic igneous activity in the form of flow-banded rhyolite lavas and endogenous dome(s), intrusive rhyolite porphyry, and a widespread pyroclastic unit containing abundant lithic fragments (Fig. 7-2; 4-2). Felsic volcanism did not occur in the complex until after the cessation of mafic igneous activity. The lithic tuff is herein accorded member status because of its distinctive lithologic character; it is composed of abundant accessory lithic fragments derived principally from the Kaertok Formation, with clasts of flow-banded rhyolite as a minor but stratigraphically very important type. In the type section (designated later) the lithic tuff overlies flow-banded rhyolite and is also interbedded with, and overlain by, units of identical composition as the upper Kaertok Formation. This is important because, although the contact between the rhyolites and the Kaertok Formation is not exposed, stratigraphic relationships suggest that the flow-banded rhyolites are contemporaneous with upper Kaertok Formation sedimentation. Establishing the relationship between the Kaertok Formation and felsic igneous rocks is important

because a rhyolite porphyry that intrudes the flow-banded rhyolites has been isotopically dated (U-Pb zircon geochronology; S.A. Bowring, University of Kansas) and, along with analysis of zircons from east-trending dykes, closely constrains the age of the Hornby Bay Group.

Flow-Banded Rhyolite

Flow-banded rhyolites are common in the northwestern and western islands of the Narakay Volcanic Complex (Fig. 7-1). They are invariably overlain by the lower facies of the lithic tuff member; underlying and adjacent rocks are inferred to be those of the Kaertok Formation, although contacts are not exposed (cf. Fig. 7-12). Outcrops are composed of durable, pink to red-coloured, largely aphyric, well-banded rhyolite (Plate N5A) that is referred to as "rhyolite" on the basis of high silica content (estimated in thin section) and the occurrence of phenocrysts of quartz and potassium feldspar. Flow-banding is defined by changes in colour between red, pink, and white that reflect variations in the abundance of feldspar, and hematite inclusions within the feldspar. Flow-banding is very coarse in inferred subterranean rhyolite, and individual bands are 2-10cm wide; these rhyolites commonly contain 5-10cm wide rotated blocks ("intraclasts") of flow-banded rhyolite (Plate N5A). Rhyolites of inferred extrusive origin display delicate flow-banding that is obvious only in thin section (Plate N5B), although rare outcrops display cm-scale flow-banding that forms chaotic isoclinal folds. The flow-banding

has a variable strike and generally steep dip ($70-90^{\circ}$); no systematic pattern to the orientation of the banding was observed.

Most of the rhyolites contain scattered phenocrysts (1-5%) of alkali feldspar and quartz. The thick-banded rhyolites are composed of a few phenocrysts of alkali feldspar and bipyramidal quartz in a granular hypidiomorphic groundmass of subhedral alkali feldspar and quartz. The pink colour typical of this rhyolite is due to the abundant inclusions of limonite and hematite that are present in the feldspars. Ragged subhedral clots of poikilotopic dolomite locally replace feldspar. Thin (1-2mm wide) stringers of polycrystalline megaquartz and chert occur in these rhyolites; the stringers are generally parallel to the megascopic flow-banding. Spherulitic and/or fibrous crystal textures indicative of devitrification are absent in thick-banded rhyolite, which implies that the "microgranitic" texture is primary. In contrast, rhyolites that display delicate flow-banding commonly contain axiolitic overgrowths on phenocrysts, spherulitic textures, and a groundmass composed of prismatic laths of alkali feldspar and quartz that mimic the flow-banding (Plate N5B). These textures imply the former presence of glass that has undergone devitrification.

Rhyolite Porphyry

On several of the westerly Narakay Islands, flow-banded rhyolite has been intruded by rhyolite porphyry. The porphyry occurs on three islands that covers an area of about 20km^2 ,

although it is uncertain if all occurrences of the porphyry represent a single intrusion. The porphyry cross-cuts flow-banded rhyolite lavas and is overlain by lithic tuff (Narakay Lithic Tuff, see below). Therefore the porphyry was intruded late during Hornby Bay Group sedimentation (Fig. 7-2), but before widespread lithic tuff deposition.

The porphyry has a purple to olive colour and is composed of anhedral to subhedral quartz phenocrysts and subhedral phenocrysts and glomeritic knots of alkali feldspar in an aphanitic to microphyric groundmass. Aphyric groundmass is composed of anhedral chlorite, dolomitic carbonate, chert, and anhedral clots of oxide; microphyric groundmass contains abundant plagioclase microlites. Along the western contact between the porphyry and flow-banded rhyolite, the porphyry contains spherical amygdules filled with saddle dolomite (cf. Radke and Mathis, 1980) and quartz.

Zircons occur mostly as microphenocrysts in the groundmass, although a few phenocrysts up to 1mm long were observed. U-Pb analysis of splits of the zircon population from several samples gives points that cluster close to the Concordia curve and indicate an age of 1690 ± 10 my. This is interpreted as the age of crystallization of the porphyry and implies that most of the Hornby Bay Group is older than 1.69Ga (S. Bowring, pers. comm.).

Narakay Lithic Tuff

Massive and bedded lithic tuff, composed largely of

sedimentary rock fragments, is a distinctive and widespread map unit (Fig. 7-1) and is therefore ascribed member status. A designated type section occurs on the dogleg-shaped island in west-central Narakay Islands. The tuff occurs on almost half of the Narakay Islands and also occurs in outcrops on the mainland south of the islands (Fig. 4-1), an area of 375km². The member is recognized by its massive matrix-supported texture, extreme durability, tawny yellowish-brown colour, and abundance of sedimentary rock fragments.

Stratigraphic Setting: The Narakay Lithic Tuff overlies flow-banded rhyolite, rhyolite porphyry, and occurs within the Kaertok Formation. Except for the type section, most outcrops of the member expose only the lower 50 metres or less of the member. In the type section (Fig. 7-11), lithic tuff overlies brecciated flow-banded rhyolite and contains conformable interbeds of dolostone and arenite that are essentially identical to the Kaertok Formation. The tuff is overlain by stromatolitic and intraclastic dolostone that contains a few thin (<50cm thick) interbeds of epiclastic conglomerate derived from the tuff. The section grades into more than 100 metres of thin-bedded largely structureless dololutite that is the stratigraphically highest unit exposed in the Narakay Volcanic Complex.

The contact between the tuff and subjacent lithologies is invariably sharp and underlying rocks are commonly brecciated and/or folded (Plate N5C). Flow-banded rhyolite displays largely in situ brecciation in which lithic tuff forms the

Figure 7-11. Measured section through Narakay Lithic Tuff in the type section. M $\frac{1}{2}$ massive facies, B bedded facies, T thin-bedded facies, E epiclastic facies. Letters separated by a slash indicate interbedding of facies. Arrows indicate axes of stromatolite elongations. Table 1 (Chapter 5) gives key to symbols used.

matrix to the clasts (Plate N5C). Some of the rhyolite breccias are cemented by black to green tourmaline; tourmaline also occurs as a fracture- and vug-fill phase, in association with quartz, in brecciated rhyolite below the contact with the tuff. The Kaertok Formation below the tuff also is brecciated, and large slabs of arenite and dolostone are common along the contact (Fig. 7-12,13). Some of the slabs display gentle domical folds and stratiform brecciation of dolostone layers. In addition, some outcrops of the Kaertok Formation below the contact display upturned leaves of "delaminated" sedimentary rock and convolutions of bedding apparently related to loading during tuff deposition. Vertical dykes, up to 1 metre wide, composed of tuff, extend downwards as much as 15m into the Kaertok Formation. Local areas within the tuff, along the contact with underlying rocks, are intensely tourmalinized and the normally yellowish-brown tuff has a dark green colour.

Contact relationships between the tuff member and the Kaertok Formation are well-exposed on an island in the central part of the Narakays; they are complex but also elucidate the eruptive/depositional mechanism associated with tuff emplacement (Fig. 7-13). Exposures at the western end of the island provide a north-south cross section of contact relationships.

Sedimentary rocks of the Kaertok Formation compose the southern third of the island and dip to the north-northwest at about 15°. These rocks are undeformed and are separated from brecciated Kaertok Formation rocks to the north by a north-east-trending normal fault (northwest-side down). The central

FIGURE 7-12: Map of island in east-central Narakay Islands illustrating the relationship between flow banded rhyolite (inferred dome rhyolites) and Narakay Lithic Tuff member. a-massive facies; b-bedded facies, c=thin-bedded facies. Note large slab of Kaertok Formation in south-central part of the island. Normal fault with more than 200m northwest-side down displacement separates lower Kaertok Formation from Narakay Lithic Tuff member.

third of the cliff face exposes the contact between the Kaertok Formation and massive tuff breccia of Narakay Lithic Tuff. North-dipping lower Kaertok Formation composes the lower part of the cliff face and is overlain by a 20m thick zone that contains large slabs of upper Kaertok Formation. The slabs display a variable orientation that ranges from near-horizontal gently folded slabs to slabs with vertical orientation. The slabs are tightly packed with a largely nondescript matrix of crushed dolstone and, in the upper third of the slabs, lithic tuff. The slabs are overlain by massive tuff breccia that infills and buries the tops of the slabs. The central third of the cliff face is separated from the northernmost third by a fault that is inferred on the basis of apparent lithologic discontinuity across a sharp topographic valley. Rocks north of the inferred fault are composed of interbedded massive and bedded lithic tuff that contains a large slab of the Kaertok Formation. It could not be determined if the massive breccia that caps the central third of the cliff face was the same as that at the top of the northern third. Also, it is uncertain whether the large slab of Kaertok Formation in the northern third of the cliff face was derived from underlying Kaertok Formation or whether it could be a sequence of interbeds (cf. Fig. 7-11) within the tuff that were dislocated during later eruptions. A schematic interpretation of the inferred sequence of events that produced these relationships is shown in Figure 7-13 and will be discussed later.

Lithologic Subdivision and Composition: Three facies have been distinguished within the Narakay Lithic Tuff: 1) massive lapilli tuff and tuff breccia, 2) bedded and uncommonly stratified and graded lapilli tuff, and 3) thin-bedded tuff. The thin-bedded tuff facies is not well-developed in the type section. On one island (Fig. 7-12), these three facies comprise an orderly vertical succession; in general, however, these three facies are randomly interbedded, although the massive facies is most common at the base of the member.

In general, the tuff member is composed of accessory lithic fragments, primary quartz and feldspar phenocrysts, and green-coloured siliceous fragments that are inferred to be felsic glass. The relative proportion of crystal to lithic components is variable; most beds are lithic tuffs (Plate N5C, N6A) but a few crystal and vitric-crystal tuffs occur (Plate N5F). Accessory lithic fragments (5-95%) are composed largely of angular to rounded clasts of arenite, dolostone, chert, and mudstone from the Kaertok Formation; angular to rounded fragments of rhyolite and mafic pyroclastic rocks also occur (Plate N5D). Rhyolite fragments are abundant only in tuffs that directly overlie in situ rhyolite. Clasts of Kaertok Formation range in size from sand-size fragments up to large slabs (5m thick over an area of 10x20m). Fragments of flow-banded rhyolite are generally of finer grain size (clasts rarely exceed 5cm in diameter). Quartz phenocrysts range from equant subhedra to irregularly shaped broken grains (Plate N5F). Feldspars are potassic feldspars but in most thin

sections examined they have been altered to clots of subhedral dolomite and hematite (Plate N5F); in hand sample this creates a porphyritic texture defined by brownish "phenocrysts". Wisp-shaped, green-coloured, rarely quartz-phyric clasts constitute less than 5% of the tuffs, and are composed of fine-grained white mica and microcrystalline silica. Some of these clasts are probably accessory lithics derived from siliceous green mudstones that occur in upper Kaertok Formation, but most are inferred to be glassy juvenile volcanic fragments. The groundmass of the tuffs is massive and aphanitic, but locally displays wispy banding (Plate N6B) possibly formed during flow and/or compaction of the tuff followed by devitrification. Although most tuffs are massive and display no fabric, some tuff beds display a weak parallel alignment of elongate lithic fragments and green glass shards and weak fluidal banding (Plate N6A). Vapour phase overgrowths on phenocrysts and spherulitic devitrification products are absent.

Massive Facies. Unsorted massive lapilli tuff to tuff breccia (Plate N5D) constitutes the lower part of the tuff member; at the base of the type section this facies is 50m thick, but also occurs in 5-10m thick beds higher in the section. Although there are subtle vertical variations in clast:matrix ratio (i.e. packing) and clast composition, no bedding and/or breaks in the section were recognized.

Framework clasts, composed largely of lithic fragments,

constitute less than 30% of the facies and float in a siliceous tan-coloured matrix (Plate N5D). The only fabric observed in this unit is weak to locally well-developed banding and alignment of clasts. Large slabs and blocks of the Kaertok Formation are common in this facies, especially near the base of the member.

Bedded Facies. This facies is most common in lower and middle parts of the member where it occurs in tabular-shaped beds 20-200cm thick (Fig. 7-11). Bedding is defined by grain size changes, thin (2-10cm thick) fine ash tuff beds, and weak parallel lamination. The overall composition is the same as the massive facies but the average grain size is smaller and large slabs of Kaertok Formation are absent. Some beds contain only sparse lithic clasts and are appropriately referred to as crystal and vitric-crystal tuffs (Plate N5F).

Normal and reverse graded bedding, lapillistone-ash tuff couplets, and massive bedding with a dispersed to tight-packed framework occur in this facies. In some outcrops with lapillistone-ash tuff couplets, the laminated ash tuff at the top of one couplet is deformed by large clasts at the base of the overlying lapillistone bed (Plate N5E). Such features may form by rapid loading and/or ballistic impact.

Thin-Bedded Facies. This facies is gradational into the bedded facies and is characterized by finer grain size and thinner beds. Individual beds range from fine lapilli tuff to ash tuff and display weak to well-developed normal grading in which crystals, vitric, and lithic fragments are concentrated

at the base of a bed. Most beds are composed of 10-15% framework grains in a siliceous aphyric groundmass; subhedral phenocrysts of quartz and carbonate-replaced feldspar occur, in addition to lithic clasts and angular blocky shards of chert and chalcedony (devitrified glass). This facies is commonly interbedded with the bedded facies; locally the two facies are deformed into irregular convolute, soft-sediment folds (Fig. 7-12).

Interpretation

The Narakay Lithic Tuff overlies and contains fragments of both the Kaertok Formation and flow-banded rhyolite. Assuming that deposition of the Narakay Lithic Tuff was more or less isochronous, this implies that formation of the underlying rhyolites was contemporaneous with upper Kaertok Formation sedimentation. Inferred paleobathymetry of the Kaertok Formation (intertidal to supratidal; <2m deep) provides constraints on eruptive mechanism of synchronous and subsequent felsic volcanism.

Flow-banded rhyolite was the first-formed product of felsic volcanism. Rhyolites with delicate mm-scale flow-banding and textures indicative of devitrification are probably extrusive flows. Flow-banded rhyolite with coarse banding and a hypidiomorphic granular texture is inferred to be a subsurface endogenous dome. The granular texture implies a greater degree of crystallization at slightly sub-solidus temperatures, a condition that might be expected to occur in an endogenous

FIGURE 7-13: Cross-section of relationship between Kaertok Formation and Narakay Lithic Tuff exposed in cliff face on the easternmost island with outcrops of Narakay Lithic Tuff. A-D are cartoons of inferred events that led to configuration shown at bottom of figure. Dotted pattern at left side of panels C and D indicates that the presence of these rocks are inferred because, as shown in the bottom panel, these rocks are not exposed.

A. Pre-eruptive Doming

C. Airfall-Fallback; Massive Facies

B. Massive Phreatoplinian Explosions

D. Pyroclastic Flow; Fall; Bedded Facies

dome rather than an extrusive rhyolite. Emplacement of shallow subsurface domes of highly viscous magma commonly is associated with deformation of overlying country rock, especially if the country rock is bedded sedimentary rock (cf. Williams and McBirney, 1979; Williams and Curtis, 1976). In the Sutter Buttes volcanic complex (Pliocene; Sacramento Valley, California) emplacement of rhyolite domes caused country rock (Cretaceous sedimentary rocks of Great Valley Group) to be uplifted as much as 500 metres over an area of 30km² (Williams and Curtis, 1976). Emplacement of rhyolite dome(s) in the eastern islands of the Narakay Volcanic Complex (flow-banded rhyolites in the western part of the complex are interpreted to be lavas) may have been in part responsible for the deformation and dislocation observed in the Kaertok Formation (Fig. 7-13).

Some of the deformation and brecciation of flow-banded rhyolite, rhyolite porphyry, and Kaertok Formation occurred just prior to and during eruption and deposition of the lithic tuff (Fig. 7-13). Breccia interstices were infilled with tuff (Plate N5C) and, in some areas, tourmaline. Open fractures were also infilled with tourmaline and quartz, both of inferred hydrothermal origin. The boron is inferred to have been derived from a magmatic source in addition to seawater and connate(?) brines that underwent "distillation" during contact with hot rock. Thus, disruption of the Kaertok Formation may have resulted from a variety of processes such as emplacement of rhyolite domes, explosions associated with eruption and deposition of Narakay Lithic Tuff, and synvolcanic faulting.

The Narakay Lithic Tuff is inferred to have been deposited from eruption clouds formed during intense phreatomagmatic explosions. Beds of crystal and vitric-crystal tuff probably represent the products of eruptions in which the driving force was a result of magmatic rather than phreatomagmatic activity. The widespread occurrence of the tuff (375km²) and its typically poorly sorted texture imply that the eruptive style may have been phreatoplinian (cf. Walker, 1981; Self, 1983). A phreatomagmatic-hydrovolcanic origin for parts of the tuff member (particularly the massive facies) is based on: 1) the high proportion of accessory lithic fragments which is considered characteristic of phreatomagmatic to phreatic eruptions (Williams and McBirney, 1979), 2) the low concentration of pumice, which indicates that violent vesiculation contributed only a minor amount of energy to eruptions, and 3) uncommon occurrence of textures formed by welding, lack of spherulites, and vapour phase crystallization textures which indicate that the tuff was relatively cool upon deposition. The rounding of lithoclasts (which led to suggestions of a sedimentary origin for the tuff member: Ross, 1982) probably formed by milling and mutual collision of particles during transport and by gas streaming and abrasion by particle-charged gases, typical of fluidized eruptions (Ollier and Joyce, 1974).

The three facies that compose the Narakay Lithic Tuff are inferred to reflect different depositional and eruptive styles. The massive facies, common in the lower part of the member, is interpreted to be the product of rapid airfall

airfall deposition from steam-rich eruption clouds and cypressoid blasts formed during initial phreatomagmatic eruptions as water encountered hot country rock and/or silicic magma. Beds of the massive facies in stratigraphically higher parts of the member may have been deposited by pyroclastic flow or airfall. The bedded facies was deposited by airfall and pyroclastic flow; the lack of stratification in individual beds of the thin-bedded facies implies that these beds were deposited by pyroclastic flows. Initial phreatic blasts caused disruption and incorporation of Kaertok Formation rocks. The decrease in abundance of accessory lithic fragments relative to juvenile components (phenocrysts and glass shards) with increased stratigraphic height suggests that eruptions changed from hydrovolcanic to more magmatic in character. This is also suggested by the increased tendency for the occurrence of bedding and internal fabric which implies increased inflation of the eruption cloud by magmatic gases and superheated steam, and a decrease in the abundance of "cold" steam and water in the eruption column.

As the intensity and frequency of pyroclastic eruptions decreased, intertidal to supratidal carbonate deposition resumed. Wave(?) attack of some beds of lithic tuff led to formation of beds of framework-supported epiclastic gravels composed of clasts of Kaertok Formation winnowed from tuff beds (Fig. 7-11). The thick succession of quiet-water dololutes that overlie the tuff records regional subsidence of the depositional surface below wave base (lacustrine or marine?)

following the cessation of eruptive activity.

The nature of the volcanic depression, and associated edifice(s), that formed during accumulation of felsic volcanics in the Narakay Volcanic Complex is difficult to infer. The Narakay Lithic Tuff does not dip more steeply than the regional dip of the Hornby Bay Group in Dease Arm. This suggests that the tuff member was deposited in a nearly horizontal attitude away from vent edifices. The occurrence of rhyolite domes and lavas and voluminous deposits of silicic tuff (150km^3) are consistent with a cauldron-like depression. Unfortunately, post-depositional movement along northeast-trending faults precludes documentation of syneruptive subsidence along the fractures. The eruptive pattern recorded by felsic igneous rocks of the Narakay Volcanic Complex is similar to that described from non-resurgent cauldrons i.e. emplacement of rhyolite domes, lavas, and a hyabyssal porphyry, followed by violent pyroclastic eruptions and eventual subsidence (MacDonald, 1972).

Dykes, Diatremes, and Explosion Breccias

The entire Narakay Volcanic Complex is cut by a swarm of east-trending (Fig. 7-1) plagioclase-phyric dykes. In several outcrops these dykes can be traced into explosion breccias (terminology of Parsons, 1969). As used in the following descriptions, an explosion breccia is a subterranean breccia formed by explosions in which the breccia remained immobile after it formed. Mobile breccias are classified as intrusive

breccias or diatremes; the latter type of breccia deposit is distinguished on the basis of shape and composition of the breccia.

Dykes

East-trending dykes in the Narakay Volcanic Complex are narrow (<50cm wide) tan-coloured bodies that are almost completely altered to fine-grained carbonate. Most of the dykes are vesicular and in rare cases dykes contain such a high proportion of spherical vesicles (infilled with brown sparry ankerite and/or dolomite) that the rock resembles a scoria. More commonly dykes display a zonal pattern of vesicles that are very small to absent along the chilled margins of dykes and increase in abundance and size in towards the core of the dyke. Less than 10% of the original mineralogy is preserved. Plagioclase, which is the only original mineral preserved, occurs as coarse (1-2mm wide) phenocrysts (1-2% of the rock) in a groundmass of plagioclase microlites that display a pilotaxitic to hyalopilitic texture (Plate N6C). Although the textures are well-preserved, both groundmass and phenocrysts have been replaced by fine-grained dolomite. In hand sample many phenocrysts have a greenish colour that is imparted by fine acicular needles of a micaceous mineral that is probably celadonite.

Explosion Breccias

The Narakay Islands contain a distinct suite of explosion

phyric dykes. The breccias are composed of angular blocks of in situ wall rock (Plate N6D). The breccias comprise irregularly-shaped pods developed sporadically along the length of a dyke although they also occur as isolated bodies unrelated to dykes. Those associated with dykes are composed of fragments of tan-coloured vesicular porphyry and locally derived sedimentary wall rock in a matrix of sparry ankerite and dolomite. Dykes do not cross-cut these breccias but can be traced into the breccia pod and out the other side. Breccia pods not associated with dykes and lacking in fragments of porphyry are composed exclusively of brecciated and altered wall rock. Alteration consists of recrystallization of dolostone to sparry, dark brown, manganiferous dolomite and/or ankerite, and vug-fills of barite, calcite, and chalcopyrite (Plate N6D). Breccias with magmatic fragments formed when dyke magma encountered pockets of connate fluids in the wall rock; breccias that are devoid of fragments of dyke porphyry probably formed when ascending superheated water underwent adiabatic decompression and flashed to steam.

Diatremes

Dyke- to pipe-like bodies of breccia occur on the Narāy Islands and on the mainland to the south. Diatreme bodies on the mainland are different from those on the islands. They are roughly circular structures (<1km wide) composed of well-bedded to massive mafic lapilli tuff; bedding dips radially in towards the center of the body. Tuff clasts are composed of

mafic plagioclase porphyries, with a dark green colour in the center of the diatreme that becomes brown to tan-coloured towards the diatreme rim. Comminuted arenite fragments, sand grains derived from disaggregation of arenite, and rare granitoid rock fragments also occur in the tuffs.

A single pipe-like diatreme occurs on the Narakay Islands; it is 200m wide, massive, and composed of tuff breccia of angular to rounded fragments of feldspar-phyric porphyry and wall rock in a lapilli tuff matrix. The porphyry fragments are tan to white-coloured; wall rock fragments are mostly of local derivation (Kaertok Formation) although clasts of white quartz arenite (Lady Nye Formation, member C4) and granitic rocks also occur. A vertically-oriented, tabular-shaped diatreme breccia occurs on Bluff Island (Plate N6E). The breccia is composed of sand- to boulder-sized clasts of white quartz arenite, siltstone, mudstone, and oolitic dolostone in a limonitic calcareous matrix; juvenile fragments of porphyry are only a minor component. The white quartz arenite boulders are interesting for two reasons: 1) they are extremely well-rounded, which attests to particle abrasion during diatreme emplacement (cf. Williams and McBirney, 1979; Ollier and Joyce, 1974) and 2) they were derived from lower member C4 of the Lady Nye Formation which, in combination with the provenance of other clasts, indicates derivation of breccia fragments from at least 600m of stratigraphic section. In this respect the granitoid rock fragments in other diatremes must have been derived from at least a kilometre below their

present stratigraphic position; they may have been derived from Wopmay Orogen basement or from conglomeratic layers in member C2 arenites of Lady Nye Formation.

Discussion

Plagioclase-phyric dykes were emplaced into a fracture system formed during north-south extension (cf. Baragar, 1977) and represent the final phase of igneous activity in the Narakay Volcanic Complex. The Dismal Lakes Group was deposited on a north to northwest-facing platform (Kerans, et al., 1981; Kerans, 1982) which suggests that this dyke swarm may record the orientation of tensile stresses that preceded formation of the Dismal Lakes Platform. The vesicular textures in most dykes are the result of exsolution of magmatic volatiles and suggest a shallow level of emplacement. Explosion breccias formed when ascending and laterally spreading dykes encountered connate fluids. Continued interaction of dyke rock with sea water after emplacement caused almost complete alteration of original mineralogy to carbonate and celadonite. Explosion breccias with no magmatically-derived fragments probably formed during adiabatic decompression of superheated fluids that flashed to steam; secondary alteration and mineralization of these breccias may be the result of interaction with either the superheated fluids and/or convected sea water.

Diatreme breccias were also generated by the interaction of fluids with hot magma but the depth of contact, as inferred from the stratigraphic position of contained clasts, was much

greater. Clast composition indicates depths at least 1km below the present stratigraphic level of the diatremes. Ascending magma, which appears to have been the same composition as the dyke magmas, encountered diagenetic fluids and/or a deep aquifer in Lady Nye Formation arenites. As the water flashed to steam the volume increase provided energy for fracture propagation and transport of breccia (cf. Woolsley, et al., 1975). Gas streaming and particle collision during transport rounded most of the clasts. Bedded, tuff-filled diatremes on the mainland may represent near-surface tuffs that subsided into the conduit after eruption and gas escape (cf. Lorenz, 1974, 1976). Mixing of fragments from different stratigraphic levels (i.e. breccias in upper Lady Nye Formation that contain fragments of lower Lady Nye Formation and East River Formation) reflects variations in gas pressure within the conduit, and hence periods of upward injection and downward subsidence of breccia.

Intrusive Breccia

A single body of intrusive breccia occurs on the east limb of the largest island of the Narakays (Fig. 7-6). Although it cross-cuts bedding, it appears to be more or less stratiform and is restricted to the middle part of member C4 of the Lady Nye Formation.

The breccia is a lapilli tuff composed of white to tan-coloured clasts of feldspar porphyry and angular fragments of Lady Nye Formation arenites, siltstones, and mudstones in a

brown-coloured matrix of disaggregated sand grains (Plate N6F); the porphyry fragments are lithologically identical to those in altered extrusive pyroclastic rocks. The tuff is cemented by granular chert and contains scattered anhedral dolomite. The intrusion is completely massive and homogenous with little variation in grain size or composition.

The intrusion outcrops over an elongate area of about 500m²; it occurs as lobate to irregular sill-like bodies that when considered as part of single intrusion define a lopolith-shaped body. The apparent source of the tuff is to the west of the area marked "VC1" on Figure 7-6. This area is underlain by a fault-bounded knob composed of tuff and Lady Nye Formation. When viewed from lake level (west of the knob) the tuff crosscuts bedding in the arenites, and then appears to have spread laterally, subparallel to bedding. In detail, the contact between the tuff and arenite is irregular and lobate; arenites in the contact zone display a variable dip to bedding and a blocky fracture as a consequence of low-grade contact metamorphism.

The tuff is inferred to have formed by explosive disruption and mixing of Lady Nye Formation with feldspar-phyric porphyry. The cause of the restricted stratigraphic range of the tuff is unknown. Spatial and stratigraphic relationships are consistent with its origin as a subvolcanic, near-vent intrusion formed during the eruption and deposition of lower East River Formation (member E1) pyroclastic tuffs.

SUMMARY

The Narakay Volcanic Complex is a well-exposed example of bimodal mafic-felsic volcanism in a high-energy marine environment. Largely subaerial tuff cones, composed of mafic pyroclastic debris, were formed during hydrovolcanic and Strombolian eruptions. Tuff cones were preserved in pre- to syneruptive depressions formed as a consequence of explosive removal of country rock and subsidence along fracture systems. Topographically positive parts of tuff cones were subjected to erosion by marine currents during return flow of waves and storm waves. Cone morphology (slope of flanks; location and amplitude of saddles and swales) affected local hydrodynamic conditions and controlled the location of wave-surge ebb and rip currents.

Mafic volcanism was succeeded by felsic igneous activity that involved emplacement of endogenous rhyolite dome(s) and lavas and eruption of mixed lithic-felsic pyroclastic ejecta. A rhyolite porphyry that intrudes the lavas, and is overlain by the Narakay Lithic Tuff, constrains the age of upper Hornby Bay Group sedimentation to 1.69 ± 0.10 Ga. The sequence of felsic rocks is inferred to represent formation and infill of a non-resurgent cauldron. Voluminous pyroclastic tuffs were deposited during violent phreatoplinian eruptions that dispersed tuffaceous detritus over an area of at least 400 km^2 .

The association of bimodal mafic-felsic igneous activity is typical of magmatism in regions of crustal extension (cf. Suneson and Lucchita, 1983). The temporal association of

mafic followed by felsic magmatism is commonly thought to reflect the process of ascent of mafic magma followed by crustal anatexis and formation of felsic melts (Emslie, 1980). The Narakay Volcanic Complex occurs at the west end of a prominent east-trending gravity high, the trend of which is parallel to the dyke swarm in the Narakay Islands. It is suggested that this gravity high reflects the presence of thinned lithosphere that formed during incipient rifting and north-south extension that eventually led to the formation of the north-facing Dismal Lakes Group Platform.

Plate N1A: Overview of contact relationships between bedded pyroclastic tuffs (V) and sedimentary rocks of East River and Kaertok Formations. Height of hill is about 100m.

N1B: Lake-level view of exhumed tuff cone and columnar-jointed mafic porphyry. Tuffs dip radially inward; contact with the porphyry is marked by dashed white line as is change from inward-dipping to outward-dipping bedding in the tuff. Trees on top of island are about 7m high.

N1C: Crude reverse grading and parallel stratification in lapilli tuff that is overlain by massive ash tuff bed. Height of outcrop is about 1m.

N1D: Brecciated intraclastic slab of interbedded lapilli and ash tuff. Fractures in tuffs are infilled with lapilli tuff derived from adjacent disrupted beds. Entire slab is enclosed in massive lapilli tuff to tuff breccia (not shown in photo). Scale bar is 50cm long.

N1E: Bedding drape and lee-side current shadow around lithic block in lapilli tuff. Bed of massive lapilli tuff (1) is overlain by weakly stratified lapilli tuff (2) and ash tuff (3). The ash tuff drapes the largest block in a discontinuous layer of block-size lithic fragments. The light-coloured clasts are composed of altered vesicular porphyry; the darker clasts to the left are composed of stromatolitic dolostone of the Kaertok Formation. Clast size decreases to the left of the largest clast, implying paleoflow from right to left (i.e. a current shadow). Penknife is 6cm long.

N1F: Poorly-sorted massive lapilli tuff bed with clasts of mafic porphyry (m) and accessory lithic material composed of arenite (a). Note the delicate fracture-bounded margin of the largest clast of mafic porphyry. Penknife is 6cm long.

PLATE N1

A

B

C

D

E

F

Plate N2A: Massive agglomerate composed of blocks of scoriaceous to dense, weakly vesicular lava, and cored bombs in a matrix of lapilli tuff. The light colour of the fragments is the result of alteration as is the dark colour of the tuff, which is due to the abundance of limonite and ankerite. Matchstick (left of top-centre) is 4cm long.

N2B: Hand sample of massive lapilli tuff typical of relatively fresh pyroclastic rocks in Narakay Volcanic Complex. Dark clasts are composed of mafic porphyry with a groundmass rich in opaque minerals. Light-coloured areas within these clasts are alkali feldspar phenocrysts and amygdules. Coarse, light-coloured fragments are accessory lithic clasts derived from country rock and composed of oolitic chert and quartz arenite. Note small light-coloured lava fragments which have a groundmass rich in chlorite. Fragment morphology suggests that the fragments formed during hydrovolcanic explosions. Coin is 15mm in diameter.

N2C: Detail of cored bomb in agglomerate. Apparently subspherical bomb is composed of a core of vesicular porphyry mantled by relatively dense lava. Note that the pattern of vesicles in the core is truncated, implying explosive derivation from solidified lava. Vesicles in the rim lava are flattened concentric to the margin (not visible in this photo). Scale bar is 4cm long.

N2D: Fusiform bomb at top of bed of lapilli tuff. The pattern of vesicles ("jelly roll structure") is nearly identical to the pattern found in cross-section of modern Strombolian bombs.

N2E: Photomicrograph of lapilli tuff. Tuff is composed of dark clasts (D) rich in microlites and groundmass opaques, light clasts with a chloritic groundmass and few microlites (L), and crystal fragments (X). "P" is clast of scoria (lower left). Inferred ash matrix is now composed of fine-grained carbonate. Scale bar is 5mm long. Plane polarized light.

N2F: Incipient alteration in thin-bedded ash tuff. The colour of the rock is various shades of green (black to grey on photo) except for irregular patches that have a brown colour imparted by the presence of post-depositional carbonate poikiloblasts (cf. bed to the left of the coin). Coin is 15mm in diameter.

N2G: Photomicrograph of dark-coloured mafic porphyry. Round areas are amygdules (V) filled with carbonate and chlorite. Large phenocryst (A) is potassic feldspar with micro sieve-textured core, composed of chlorite and opaques and inferred to be relic glass inclusions. (O) may be olivine (?) phenocryst that has been replaced by carbonate. Note abundant plagioclase microlites. Scale bar is 1mm long. Plane polarized light.

PLATE N2

A

B

C

D

E

F

G

Plate N3A: Interbedded massive to parallel-laminated lapilli tuff and ripple cross-laminated to parallel-laminated ash tuff of graded-laminated facies. Beds are interpreted as Ta-c, Ta-b, and Ta-b-c turbidites of inferred pyroclastic origin. Note rippled top of tuff below matchstick in centre of photo and climbing ripple cross-stratification in lowermost ash tuff.

N3B: Climbing megaripples in crossbedded facies. Lower and upper part of cliff face is composed of parallel-laminated lapilli-tuff (P). Crossbedded section (C) is composed of upward-thickening coset of tabular planar crossbeds that were deposited by bedforms with a moderate angle of climb such that foreset and topset strata were preserved.

N3C: Backset bedding in crossbedded lapilli tuff. Current moved from right to left. Backset bedding infills of chute-and-pool structure. Height of rock face is 20cm.

N3D: Low-angle cross-stratification of inferred base surge origin. Matchstick is 4cm long.

N3E: Parallel-laminated lapilli tuff. Lamination formed as fragments were deposited after settling through the water column (inferred from the absence of sag features below cored bomb, marked by upper arrow). Arrow in lower half of photo marks discontinuous layer of coarse cinder.

N3F: Lenticular-bedded oolitic volcanic arenite and siltstone of mixed facies in lower Kaertok Formation. Arenite beds are composed of massive to cross-laminated oolites and rounded volcanic clasts. Matchstick (lower left) is 4cm long.

PLATE N3

A

B

C

D

E

F

Plate N4A: Lens of epiclastic channel-fill (outlined in black) on the inferred flank of a tuff cone (contact marked by subvertical black line). The channel deepens from left to right from 1m to 12m over a distance of 300m. The channel cuts into upper East River Formation sedimentary rocks and is overlain by more epiclastic deposits and Kaertok Formation sedimentary rocks. Height of cliff face is 100m.

N4B: Parallel-laminated limonitic granulestone at base of proximal channel thalweg. White clasts are fragments of altered porphyry. Knife is 9cm long.

N4C: Crude graded bedding in proximal epiclastic conglomerate. Deposit is composed of angular, unabraded clasts of vesicular porphyry and large blocks of sedimentary rock. Note outsized clasts of Kaertok Formation at top of the bed. Scale bar is 20cm high.

N4D: Stacked sheets of massive epiclastic granulestone from "medial" facies. Note deformation of mudstone bed below epiclastics.

N4E: Epiclastic beds of massive and normal graded granulestone and mudstone overlain by a biostrome of SH-C stromatolites. The volcanic fragments are now composed entirely of limonite and carbonate. The stromatolites grew on a scour surface with 1.5m of synoptic relief. Hammer in right center.

N4F: Ta-e Bouma turbidite beds in tabular sheet of distal facies epiclastic rocks. Each bed is composed of granules of feldspar and porphyry that grade into sand-size clasts capped by silty mudstone. Note rippled tops of upper beds. Matchstick is 4cm long.

PLATE N4

A

B

C

D

E

F

Plate N5A: Outcrop of flow-banded rhyolite typical of inferred dome-forming rhyolite. Banding is defined by colour variations imparted by differences in the amount of feldspar and exsolved Fe-oxides in feldspar. Note blocks ("intraclasts") to the right of the knife that were formed by brecciation and rotation during lava flow. Penknife is 9cm long.

N5B: Photomicrograph of delicate flow-banding in rhyolite lava. Phenocrysts are composed of quartz (Q) and altered alkali feldspar (dark grains). Field of view is 3mm wide. Plane polarized light.

N5C: Brecciated rhyolite lava. Dark-coloured clasts are in situ fragments of brecciated rhyolite lava (shown in N5B). White matrix is infill of Narakay Lithic Tuff. Sledgehammer is 1.2m long.

N5D: Glacially-rounded outcrop of massive facies of Narakay Lithic Tuff. Note unsorted texture and high percentage of matrix. Hammer head is next to large block of mafic pyroclastic tuff derived from older volcanic deposits.

N5E: Bedding in bedded facies of Narakay Lithic Tuff. Lower bed of massive lapilli tuff is overlain by well-stratified ash tuff. Load structure to the left of the 9cm long penknife probably is due to ballistic impact (i.e. bomb sag feature). Fragments in this tuff are composed of dolostone (recesses), rhyolite, and chert (white).

N5F: Photomicrograph of crystal-rich rhyolite ash tuff. Phenocrysts are composed of quartz (grey to white) and alkali feldspar (F). Feldspars are altered to dolomite and hematite. Scale bar is 5mm long. Crossed nicols.

PLATE N5

A

B

C

D

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Plate N6A: Flow-banding in lapilli tuff of bedded facies of Narakay Lithic Tuff. Fragments are composed of dololomite (D), quartz phenocrysts (Q), and glass(?) shards (G) in a cherty aphanitic groundmass. Some of the quartz grains are not phenocrysts but were derived from disaggregation of Kaertok Formation arenites. Gray specks throughout thin section are tourmaline euhedra developed after tuff flowage and welding. Scale bar is 5mm long. Crossed nicols.

N6B: Devitrification(?) structures developed in ash tuff composed of quartz (Q) and altered alkali feldspar (A) phenocrysts in an aphanitic cherty groundmass. Light-coloured wispy patches are composed of chert and illite and are oriented parallel to bedding. Note how large wisps grade into very fine shreds. Siltstone lapillus in upper left. Scale bar is 5mm. Crossed nicols.

N6C: Photomicrograph of feldspar-phyric dyke. Coarse feldspar phenocrysts (F) are surrounded by pilotaxitic groundmass with feldspar microlites and opaque grains. Note dolomite-filled amygdulites (A). As a consequence of alteration, the entire rock is composed almost completely of fine-grained dolomite. Scale bar is 5mm long. Plane polarized light.

N6D: In situ explosion breccia composed of angular blocks of cross-stratified oolitic quartz arenite. The dark colour of the clasts is due to alteration of dolomite to sparry Mn-rich ankerite and/or siderite. White patches are vugs filled with barite and chalcopyrite.

N6E: Breccia dyke within feldspathic arenites of member C4 (Lady Nye Formation). The dyke is composed of rounded to angular fragments of quartz arenite, mudstone, dolostone and altered porphyry in a matrix of sparry ankeritic dolostone. Note rounded boulder above hammer head; this boulder is composed of very well indurated quartz arenite derived from about 600m below its present stratigraphic level. The well-developed rounding attests to intense fluidization and gas-streaming during emplacement.

N6F: Hand sample of intrusive breccia. The coarser fragments (white) are composed of altered mafic porphyry, feldspathic arenite (light grey), and mudstone (black), the latter two lithologies being derived from lithofacies D, member C4, Lady Nye Formation. The groundmass is composed of comminuted porphyry and arenite that is partly replaced by limonite and arenite. Coin is 15mm in diameter.

PLATE N6

A

B

C

D

E

F

CHAPTER 8

SUMMARY

The preceding seven chapters of this thesis have presented the results of descriptive research and interpretation of a relatively unmapped and little known sequence of Proterozoic rocks in the Northwest Territories of Canada. The purpose of this chapter is to attempt to take these interpretations beyond the present basin margins and speculate on the the Hornby Bay Group with respect to tectonic setting, extrabasin-al correlations, and geochronology. As a prelude to considering these questions, a brief synopsis of the depositional history of the Hornby Bay Group will be presented.

The Hornby Bay Group comprises a sequence of terrestrial and marine siliciclastics and marine carbonates deposited on continental crust. The Hornby Bay Group unconformably overlies Wopmay Orogen, and available evidence suggests that deposition post-dated tectonic activity in Wopmay Orogen. However, the pattern of deformation imparted to Wopmay Orogen by the final plate collision (e.g. the conjugate wrench fault system) significantly affected depositional patterns within the Hornby Bay Group.

Hornby Bay Group deposition was initiated as a north-trending topographic low was infilled by fluvial and aeolian sediments of the Bigbear depositional system. On the basis of grain size variations and facies patterns, west and east margins of the basin have been inferred in the region along

the Coppermine River (north of Bigbend) and east of Mist Mountain Outlier, respectively. Presently, Bigbear system deposits cover at least 4000km^2 but the original extent of this system is unknown. The first stage of deposition by the Bigbear system is recorded in an upward-fining sequence of conglomeratic and arenaceous rocks that infill rugged paleotopography. Colluvial cones and rock fans developed on the flanks of paleohighs as southwest-flowing braided rivers flowed through intervening valleys. Source area back-wasting led to decreased grainsize in the fluvial system and formation of an extensive sandy braidplain, possibly during formation of the Teshierpi-Bigtree peneplain in the source area. Bigbear erg, an aeolian sand sea that deposited more than a half kilometre of aeolian arenites, developed when sand-laden prevailing winds (from the west) encountered the braidplain. Erg formation was controlled by decreased wind competency on the lee side of the western margin of the erg and by impedance of through-going winds by the eastern margin of the basin. Intraformational conglomerates near Sinister Fault signal the end to a period of tectonic quiescence and onset of major reorganization of Bigbear system paleoslope and drainage. Upper parts of the Bigbear Formation record the change from southeast- to southwest-dipping paleoslope and burial of the Bigbear system by the Lady Nye fluvial system.

The Fault River Formation occurs as fault-bounded remnants of a once extensive ($>2000\text{km}^2$) fluvial system. The Fault River Formation is composed of two upward-coarsening sequences of

arenaceous and conglomeratic rocks about 350m thick. Drainage and depositional patterns were influenced, but not controlled, by syndepositional movement along a northeast-trending system of faults (Fault River Fault system). The lower upward-coarsening cycle records westward progradation of high-energy stream flood fan system that buried earlier fault scarp fan facies. Clasts in the conglomerates were derived from erosion of the Sloan Group (Great Bear Batholith) and outlying (westernmost) deposits of the Bigbear system. The second upward-coarsening cycle records uplift and basement rejuvenation east-northeast of present outcrops of the Fault River Formation that was followed by burial of the Fault River system by the Lady Nye system.

Inferred sporadic fault movement during Fault River Formation deposition, the occurrence of boulders of upper Bigbear Formation arenites, and the conformable occurrence of Lady Nye Formation conglomerates at the top of the Fault River Formation invites correlation of Fault River system deposition with upper Bigbear system deposition. Both systems were affected by faulting that commenced during a change from east-dipping to west-dipping paleoslope.

The Lady Nye Formation contains conglomerates and arenites that record infill of deeply incised basement paleotopography by a large-scale integrated fluvial system. The paleotopography formed during incision of Teshierpi-Bigtree peneplain, in response to uplift and westward tilting of basement, that was probably in part contemporaneous with Fault River

Formation deposition. Deposition of Lady Nye Formation commenced when uplift of the rejuvenated peneplain ceased and the region began to subside. Source area erosion during deposition of the second upward-coarsening cycle in the Fault River Formation breached a basement ridge that may have separated the Lady Nye fluvial system from the region west of Coppermine River, and thus allowed drainage of Lady Nye system into the Fault River region. The Lady Nye Formation was deposited by a southwest-flowing, high-energy river system that drained through basement paleovalleys and carried clasts derived from as far as 150km to the east. High-energy rivers in the region of basement highlands graded downstream into an extensive sandy braidplain. Dramatic thickening of members within the Lady Nye Formation west of a north-trending lineament (that also marks the western limit of basement highlands) is the first evidence of greater differential subsidence across Leith Line, a long-lived basinal hinge line inferred to represent a fundamental contact and change in the composition of basement rocks (i.e. tectonic boundary). Marine transgression of Lady Nye braidplain led to formation of a complex mosaic of braidplain, delta plain, and subtidal marine facies affected by severe storms. Paleocurrents from upper Lady Nye Formation indicate a northward paleoslope, perhaps due to crustal doming that may have preceded Narakay Volcanic Complex volcanism.

The East River Formation is a sequence of carbonate and subordinate siliciclastic rocks that record a transgressive-

regressive cycle of sedimentation that occurred on a west-northwest facing carbonate ramp. Mixed siliciclastic-carbonate sedimentation took place in a high-energy intertidal setting during clastic stabilization and decrease in the supply of siliciclastic detritus. Distinctive stromatolite bioherms and hummocky cross-stratified oolitic dolostone, occur over an area of at least 30,000km², formed on a storm-dominated ramp. Optimum conditions for carbonate deposition allowed sea-level build-up (shoaling) and progradation (regressive offlap) of tidal flat facies. This may have been the result of local tectonic control on sea-level rather than reflecting global-eustatic variations because shoaling of the East River Formation culminated in cannibalization of previously deposited East River Formation during the inception of dip-slip movement along the Teshierpi Fault.

Dip-slip movement on Teshierpi Fault was associated with regional uplift and initiation of siliciclastic sedimentation. The depositional pattern and facies geometry of fluvial-deltaic siliciclastics of the Kaertok Formation was controlled by syndepositional movement along the Teshierpi Fault. The Kaertok Formation was deposited by a fluvial system that flowed southwest along, or close to, the Teshierpi Fault, and terminated in a wave-dominated delta system. Kaertok Formation siliciclastics grade into basal siliciclastics of the Dismal Lakes Group (Mountain Lake Formation), which record a period of southwest-flowing fluvial sedimentation followed by marine transgression of a north-facing paleoslope.

The Narakay Volcanic Complex contains mafic and felsic pyroclastic rocks that are interbedded with sedimentary rocks of the Lady Nye, East River, and Kaertok Formations. Mafic hydromagmatic volcanism formed tuff cones and derivative epiclastic rocks deposited in a high-energy marine environment. Later felsic magmatism was associated with formation of a non-resurgent cauldron-like structure. Emplacement of endogenous rhyolite dome(s) and lavas during Kaertok Formation deposition was succeeded by powerful phreatoplinian eruptions that deposited the widespread Narakay Lithic Tuff. A thick sequence of quiet-water dololomite is restricted in occurrence to the Narakay Islands and was deposited during subsidence following eruption of Narakay Lithic Tuff. The final igneous event recorded in the Narakay Volcanic Complex was the emplacement of east-trending, feldspar-phyric dykes, which may be important strain markers related to formation of the north-facing Dismal Lakes Group platform.

Tectonic Setting

Early hypotheses for the tectonic setting of the Hornby Bay Group and Coppermine Homocline advocated a rift or aulacogen origin (Burke, 1976a; Fraser *et. al.*, 1972). Rifts are defined as "elongate depressions overlying places where the entire thickness of the lithosphere has ruptured in extension" (Gregory, 1921, *in* Burke, 1976a). Lithosphere-scale rupture in rifts involves formation of normal and listric-normal fault systems with kilometres of displacement (Wernicke and

Burchfiel, 1982). Fault activity during deposition of the Hornby Bay Group was restricted to relatively minor (tens to a few hundreds of metres) offset during reactivation of an existing fracture system (the conjugate wrench system in Wopmay Orogen). In view of the apparent absence of lithosphere-scale rupture in the Coppermine Homocline, and the dissimilarity of sedimentary rocks to those formed in rift valleys and tectonically active environments, it is perhaps desirable to abandon the idea that the Hornby Bay Group and Coppermine Homocline is a rift or aulacogen.

Before proposing an alternate tectonic model for the accumulation of the Hornby Bay Group the following main points should be considered.

- 1) Cratonic tilting to the east (early Bigbear Formation) implies uplift in the west. The low relief of the Teshierpi-Bigtree peneplain indicates that the magnitude of uplift was gentle and may have been focused west of the present western limit of exposed Hornby Bay Group rocks.

- 2) Paleoslope reversal, and crustal tilt towards the west (upper Bigbear Formation, Fault River Formation, and Lady Nye Formation). This change in paleoslope was associated with dip-slip reactivation of basement wrench faults and uplift in the Coronation Gulf area.

- 3) The Hornby Bay Group thickens to the west, indicating that the region of initial uplift (1) later became a region of subsidence, although it is difficult to decide whether the westward thickening is due exclusively to uplift-related

subsidence or to different mechanical behavior of Wopmay Orogen basement west of Leith Line.

4) The Narakay Volcanic Complex represents the westernmost exposures of Hornby Bay Group rocks, and their location may mark the region of early uplift that caused the east-dipping paleoslope recorded in the oldest Hornby Bay Group deposits.

The history of sedimentation recorded in the Hornby Bay Group is one of adjustment of fluvial drainage and sea-level to changes in the orientation and magnitude of crustal tilt that is inferred to have occurred during a period of mild crustal extension, as suggested by dip-slip reactivation of basement faults, and the occurrence of bimodal basalt-rhyolite volcanism. The cause of the extension is unknown. As Burke (1976b) points out in his discussion of sedimentation in the Chad Basin, an active intracratonic basin, the history of sedimentation in intracratonic basins may be controlled by structures and tectonic processes far removed from the locus of deposition. Hence, sedimentary rocks within such basins may reveal little of the tectonic processes responsible for their deposition. A similar situation is envisaged for the Hornby Bay Group, and therefore it is difficult to determine whether the inferred extension recorded in this group resulted from isostatic readjustment of Wopmay Orogen following final collision, rifting and plate separation elsewhere in the northwest Canadian Shield (to the west?), or the result of plate collision in regions far removed from the outcrop area of the Hornby Bay Group.

The extent of the Narakay Volcanic Complex does not, unfortunately, help in resolving this problem. The Narakay Volcanic Complex is marked by a prominent gravity high at the west end of a east-trending regional gravity high (+20 to +25mgal) (Fig. 8-1). Available data suggest that this trend does not extend across Great Bear Lake, and must close in Dease Arm. The origin of the regional gravity anomaly that extends from the Narakay Islands to east Great Bear Lake is unknown. If it is related to the Narakay Volcanic Complex, it may signify a linear region of thinned lithosphere and/or denser crust (i.e. intruded by basalt) that could have resulted from crustal extension prior to formation of the Dismal Lakes Group platform.

Geochronology and Extrabasinal Correlation

U-Pb zircon geochronology on zircons from the Narakay Volcanic Complex provide the first direct dates of Hornby Bay Group rocks. Hypabyssal rhyolite porphyry, inferred to predate deposition of the Narakay Lithic Tuff, yields an age of 1696 ± 10 my. Zircons from east-trending feldspar-phyrlic dykes, the youngest rocks in the Narakay Volcanic Complex, yield an age of 1660my. Therefore the age of the uppermost Hornby Bay Group is bracketed between 1660my and approximately 1700my. The youngest date (also U-Pb zircon analysis) in igneous basement rocks is from granitic rocks of Great Bear Batholith and yields an age of about 1840my (Van Schmus and Bowring, 1980; Bowring, 1982). Thus the Hornby Bay Group accumulated

Figure 8-1. Gravity map of the Great Bear Lake region (from Bouguer Gravity Map No. 64112, 1981).

during the period from about 1820my to about 1660my. The implications of this data are:

1) The Hornby Bay Group has commonly been correlated with other intracratonic sandstone basins in the northwest Canadian Shield (Fig. 1-1). The Athabasca Basin is filled with lithologically similar rocks; Rb-Sr age determinations of tuffaceous rocks in this basin indicate an age of deposition of around 1500my (Raemaekers, 1981). The Thelon Basin is also filled with a similar sequence of clastic and subordinate carbonate rocks. Recent U-Pb analysis of authigenic apatites formed during evaporitic lacustrine sedimentation indicate an age of 1690my (Chiarenzelli, et al., 1983). Therefore it would appear that the Hornby Bay Group is more closely correlated with Thelon Formation clastic rocks than those of the Athabasca Basin.

2) On the basis of westward-flowing paleocurrents recorded from the Athabasca, Thelon, and younger Hornby Bay Group rocks, earlier workers suggested that these basins were part of an integrated fluvial system that covered the northwest Canadian Shield (Fraser, et al., 1970). Geochronologic data suggest that only the lower Thelon Formation and Hornby Bay Group were contemporaneous. Clast composition of the west-flowing Lady Nye system indicates that this fluvial system probably was not connected with the Thelon fluvial system.

3) Rocks indicative of an arid paleoclimate occur at the base of both the Thelon and Bigbear Formations. In the Thelon Formation, these rocks include evaporitic lacustrine mudstone,

silcretes and more than 500m of aeolian arenites (Bigbear erg). If these two basins are contemporaneous, it would imply the existence of a significant area of arid climate sedimentation in the northwestern Canadian Shield around 1750-1700my ago.

4) Hornby Bay Group rocks are generally considered to be correlative with Belt-age rocks in the Canadian Cordillera (Young et al., 1978). It now appears that the Hornby Bay Group is significantly older than Belt Supergroup rocks, which are generally considered to have ages of around 1450my (Stewart, 1972).

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CAPTIONS FOR FIGURES IN POCKET

FIGURE 2-3: Detailed geologic map of the main region of exposure of the Bigbear Formation. Dased mostly on original map by the author; geology of parts of Wopmay Orogen is from Hoffman (1978).

FIGURE 3-1: Geology in the Hornby Bay region of Great Bear Lake.

A. Generalized geologic map of the Hornby Bay region, based on the mapping for this thesis and earlier work of Irving, et al. (1972), Hoffman (1978), and Hildebrand (1981). B and C: 1:50,000 scale maps of the outlier at Elizabeth Lake and the Fault River region, respectively. Sections referred to in text are numbered on maps.

FIGURE 4-1: Geologic map of the Lady Nye Formation and overlying East River and Kaertok Formations in the region between Hornby Bay and Dismal Lakes. Modified from Baragar and Donaldson (1973), Hoffman (1982), and from unpublished work of the author.

FIGURE 4-4: Paleocurrent map for the Lady Nye Formation and lowermost member of the East River Formation (member E1). Paleocurrents of member C2a, C2b, and C3 are not differentiated. Rose diagrams plotted for stations in members C4 and E1 (East River Formation) show both marine influence and the effect of paleoslope. The dolostone pattern marks the base of the East River Formation. Arrows in brackets are sequential up-section stations; average given at base of each bracket. Lithologic and fault symbols as in Figure 4-1. NVC= Narakay Volcanic Complex; PI= Prospect Island; TR= Teshierpi River; LR= Lac Rouviere.

FIGURE 6-2: Measured sections within the Kaertok Formation. Symbols for lithologies and sedimentary structures are given in Table 5-1. Numbers along left side of columns indicate thickness in metres; note that each section has been plotted at a different scale. In section "1" (type section, Kaertok Creek): TU/CU= thinning-up/coarsening-up cycles; TU/FU= thinning-up/fining-up cycle; ThU/Cu= thickening-up/coarsening-up cycle. The rose diagram at the base of the column shows measurements of crossbeds (inner rose) and measurements of ripple mark axes (outer rose). In section 2 (East River section) circled numbers along the left side of the column refer to lithofacies described in the text. "Red" and "green" along the right side of the column refer to the dominant colour of the mudstones and arenites. In section 3 (Narakay Islands section) most of the sandstone and siltstone units are calcareous. Rose diagrams are a compilation of crossbeds and stromatolite elongations from the entire section.

5 5

OF / DE

INTRUSIVE ROCKS

North-trending diabase dikes (Mackenzie age?)

Western Channel Diabase.

HORNBY BAY GROUP

LADY NYE FORMATION H_3

C2a- C2a Member Polymictic orthoconglomerate and pebbly sublitharenite. Cobbles of Bigbear Formation, vein quartz, and siliceous ironstone

FAULT RIVER FORMATION H_2

B4- B4 Member Pebbly to granular sublithic coarse-grained quartz arenite.

B3- B3 Member Hematitic medium-grained quartz arenite, locally feldspathic, with 15-20% volcanic paraconglomerate.

B2- B2 Member Interbedded paraconglomerate sheets and quartz arenite lenses. Cobbles of McTavish Supergroup volcanic rocks Bigbear Formation, and vein quartz.

B1- B1 Member Interbedded micaceous siltstone and mudstone at base, grades into lithic to arkosic arenite and locally derived volcanic-granitic paraconglomerate

Bu- Fault River Formation-undifferentiated

WOPMAY OROGEN

- Geologic contact: known, inferred
Normal fault,
teeth on downthrown-side
Strike-slip fault
(pre-Hornby Bay Group deposition)

FRF - Fault River Fault

B

37

HORNBY BAY
(GREAT BEAR LAKE)

4 of

36

ELIZABETH LAKE

NORTH

WOPMAY OROGEN

Granite plutons of Great Bear Batholith

McTavish Supergroup mostly felsic tuffs, lavas, and intrusive porphyries

A

1 of

2 of 1

MOUNTAIN LAKE FM.

MOUNTAIN LAKE FM.

PALEOCURRENT
MEASUREMENTS
FROM OOLITIC
GRAINSTONE

37

NT
S

NARAKAY IS.
LITHIC TUFF.

4 of 1

SAME
LITHOLOGIC
SEQUENCE

100

75

50

25

15

10

5

159

1 of 1

1
PH
D
1984
8
R-1969

878

Falcon
Lake

29

37

4 of 1

Thes.
Ph.D.
1984
K166b

59

LEGEND

DISMAL LAKES GROUP

HORNBY BAY GROUP

KAERTOK FORMATION

GE
(kr)

Dease
River

Ritch Is

COVER POOR OUTCROP

GEOLOGICAL CONTACT
(known inferred)

NORMAL FAULT

17 of 1

| 8 of |

LEGEND

DISMAL LAKES GROUP

HORNBY BAY GROUP

KAERTOK FORMATION
(mudstone & arenite)

EAST RIVER FORMATION
(Dolostone)

LADY NYE FORMATION

MEMBER C4 Arenite and Mudstone

C3 Arenite

C2b Pebbly Arenite

C2a Conglomerate

C1 Arenite

WOPM OROGEN

GREAT BEAR BATHOLITH

McTAVISH SUPERGROUP
(Volcanic Rocks)

C4 C3

C4/C3

C4/C3

107

COVER POOR OUTCROP

GEOLOGICAL CONTACT
(known inferred)

NORMAL FAULT

Blood
Creek

C3

C2

C3

C3

C3

C3

C3

12 of 12

1 of 1

Dease
River

East
River

12 16 18
20

13

9
15
20
23
82
15

6 of 1

BLOOM
CREEK

HORNBY BAY
(GBL)

35
33

10
12
29

20
15
16
15
24
90

22
42

| 4 of 1 |

B.

250

A₅

170

A₃

29

42

1 of

C

37

DISMAL LAKES GROUP (UNIT II)

A7

A7

A7

A7

A7

A7

A7

HORNBY BAY GROUP

LADY NYE FORMATION

Lc₁-polymictic paraconglomerate
(Hunn Creek conglomerate)

BIG EAR FORMATION

A₇- Crossbedded quartz arenite with
white vein quartz pebbles

A₆- Interbedded intraformational paraconglomerate and crossbedded red quartz arenite

A₅- Interbedded sets of cross-to planar bedded quartz arenite; thin mudstone and polymictic paraconglomerate

A₄- Fractured, non-tuffaceous quartz arenite with tabular planar cross-bedding

A₃- Crossbedded feldspathic to arenaceous quartz arenite, local

A₂- Interbedded massive polymictic conglomerate and crossbedded sublitharenite

A₁- Massive, locally derived, poorly sorted and uncommon arkosic interbeds

Dialase dyke (Mackenzie age?)
known, inferred

Geological Contact
known, inferred

Outcrop, composition unknown

Normal Fault, teeth on
downthrow side

Quaternary cover

4 of 1

*Sinister
Fault*

Lc₁ 030

RAY GROUP

WOPMAY OROGEN

FORMATION

G - GREAT BEAR INTRUSIVE SUITE:
hornblende biotite monzogranite

D - DUMAS GROUP: felsic tuffs and
lavas, minor basalt

H - HEPBURN INTRUSIVE SUITE:
biotite granodiorite to granite

A - AKAITCHIC GROUP: amphibolite
facies metasediments

paraconglomerate
ark conglomerate

FORMATION

bedded quartz arenite with
quartz pebbles

bedded intraformational para-
conglomerate and crossbedded
arenite

bedded sets of cross-to parallel-
bedded quartz arenite; minor red
bedded polymictic paraconglomerate

hematitic quartz arenite in
tabular planar crossbeds

bedded feldspathic to mafic
quartz arenite, locally conglomeratic

bedded massive polymictic para-
conglomerate and crossbedded hematitic
arenite

locally derived paraconglomerate
common arkosic interbeds

59

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E.

176

13

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6 f.

SINISTE Fault

Thickness
of units,
in meters

0-20

201

61

60

125

D.

A₅

A₄

A₃

A₂

A₁

8

14

6

57

Paleocurrent
mean azimuth,
number of
measurements

mst ss cgl

A₄

D

D

D

BIG BEND

129

6 15 29 29

END

2	9	H	1	0	H	8	5
---	---	---	---	---	---	---	---

FIN