SEDIMENTOLOGY AND STRATIGRAPHY
OF THE
NORTHERN AND CENTRAL METASEDIMENTARY BELTS
IN THE
BEARDMORE-GERALDTON AREA
OF
NORTHERN ONTARIO

by

Jonathan Ross Devaney ©

A thesis submitted in partial fulfillment
of the requirements for the degree of
Master of Science

LAKEHEAD UNIVERSITY

Thunder Bay, Ontario, Canada

January, 1987

ISBN 0-315-39582-6

ABSTRACT

The northern and central metasedimentary belts (NMB, CMB) are east-trending, regional (formation- to group-scale) lithostratigraphic units within the Beardmore-Geraldton Archean terrane of northern Ontario's Superior Province.

Gravelly braided rivers deposited the lithofacies assemblage of clast-supported polymict conglomerate and interbedded sandstone that comprises most of the NMB. Felsic volcanic pebbles and cobbles are the most abundant clast lithology.

The CMB contains several east-trending lithofacies assemblages that together form a generally northward-coarsening sequence. The southern CMB is composed of mudstone, iron formation and sandstone with common graded beds. This fine-grained assemblage is paralleled to the north by a horizon of rhythmically bedded and cross-bedded sandstone units, and a heterogeneous, conglomerate-rich assemblage. The northern CMB strata are mostly a conglomeratic assemblage similar in form (lithofacies) and composition (provenance) to the NMB. Some fine-grained units (mudstone, iron formation, graded sandstone) of relatively minor regional extent are present near the north margin of the CMB.

The CMB's southern (lower) fine-grained subaqueous facies are apparently capped at different sites by: 1) rhythmic delta front couplets; 2) sandy braided river deposits with very rare associated intertidal strata; and 3) a conglomeratic submarine fan or fan-delta front resedimented assemblage. The northern (upper) CMB is a gravelly braided river deposit, with minor aquabasinal facies.

The CMB is probably a 1 - 2 km thick structurally modified homoclinal sequence. Relative positions of its depositional paleoenvironments, as deduced from lithofacies assemblages, suggest that the epiclastic portion of the CMB is the record of a dominantly coarsening-upward, subaqueous to subaerial trend that was produced by a prograding clastic system(s), likely a number of fan-deltas.

The extreme eastern part of the CMB is composed of generally oligomict, coarse (conglomeratic) felsic pyroclastic and/or reworked volcaniclastic facies which are probably subaerial deposits.

Similar clast compositions and several sedimentological criteria, including average maximum deformed clast size, strongly suggest that highly proximal fluvial facies in the NMB and more distal fluvial and aquabasinal facies in the CMB were originally part of a continuous coarse clastic wedge or sheet.

TABLE OF CONTENTS

Page

LIST OF FIGURES

List of Figures (Cont'd)

List of Figures (Cont'd)

LIST OF TABLES

CHAPTER 1: INTRODUCTION

1.1 Purpose and Background

The purpose of this study is to examine two belts of metasediments and, through the use of sedimentary facies models, interpret their depositional paleoenvironments. The relative positions of the paleoenvironments may provide regional stratigraphic clues to the history of the Beardmore-Geraldton area (Fig. 1.1). A secondary and related aim is the investigation of conglomerate provenance by way of point counts of clast lithologies done on outcrops.

Studies of Archean (Early Precambrian) metavolcanic-metasedimentary terranes, or "greenstone belts" (Condie, 1981; Ayres *et al*, 1985) have provided many challenges to geologists. Some workers have focused their interests on the metasediments involved. Readers not familiar with Archean metasediments are urged to read Ojakangas' (1985) excellent review.

Within the metavolcanic-metasedimentary terranes, the most common metasediments are the Resedimented Facies Association, turbidites and slates of submarine fan origin, and an Alluvial Fan-Fluvial Facies Association, conglomerates and sandstones of braided river origin (Ojakangas, 1985). Felsic volcanic sources are thought to have supplied the detritus for proximal braided rivers and distal subaqueous basins (Ojakangas, 1985; Ayres and Thurston, 1985). Strata

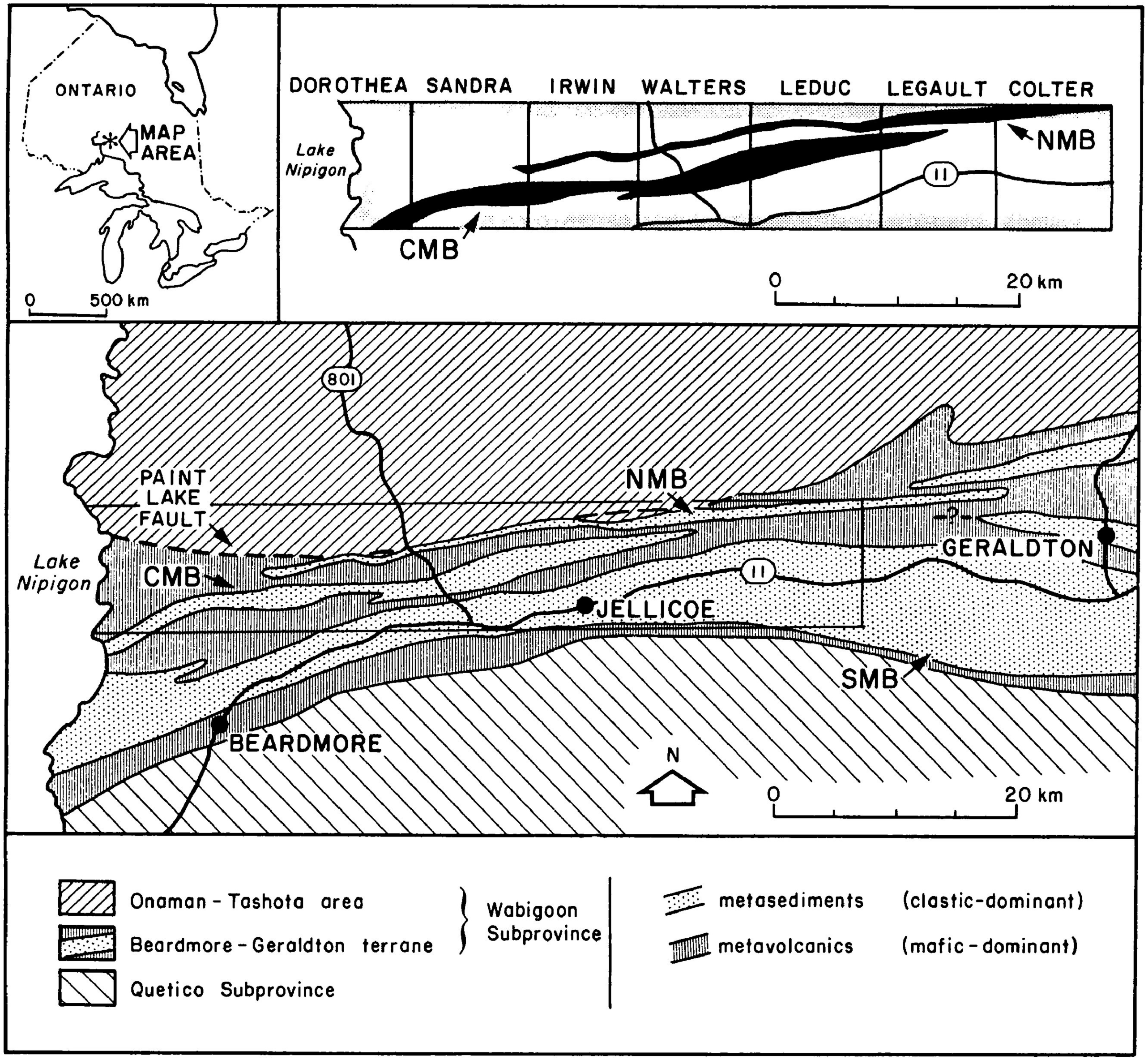

ONTARIO
MAP AREA
0 500 km
DOROTHEA
SANDRA
IRWIN
WALTERS
LEDUC
LEGAULT
COLTER
Lake Nipigon
NMB
CMB
11
0 20 km
801
PAINT LAKE FAULT
Lake Nipigon
CMB
NMB
?
GERALDTON
11
JELLICOE
SMB
BEARDMORE
N
0 20 km
Onaman - Tashota area
Beardmore - Geraldton terrane
Wabigoon Subprovince
Quetico Subprovince
metasediments (clastic-dominant)
metavolcanics (mafic-dominant)

Fig. 1.1: Simplified geological map of the Beardmore-Geraldton area, modified from Stott (1984a, b) and Mason *et al* (1985). Upper right inset shows the study area, Dorothea to Colter Townships.
NMB: northern metasedimentary belt;
CMB: central metasedimentary belt;
SMB: southern metasedimentary belt.

interpreted as shallow marine, deltaic, and coastal are extremely rare. However, recent research on fan-deltas (Wescott and Ethridge, 1980; papers in Koster and Steel, 1984) emphasizes that braided fluvial and turbiditic subaqueous fan deposits can be part of a single depositional system, a fan-delta; this point was neglected by Ojakangas (1985).

The presence or absence in the study area of any resedimented (submarine fan) conglomerates is of interest in light of some controversial discussion in the literature (Goodwin, 1977; Walker, 1978a) regarding Archean basin margin criteria. In some cases it can be quite difficult to differentiate resedimented from braided river conglomerates (Winn and Dott, 1977; Hein, 1984), especially in deformed metaconglomerate in which primary clast fabrics (Walker, 1984a, Fig. 11) have been destroyed.

Some previous studies of Archean coarse clastic metasediments were both exemplary and inspirational, those of Eriksson (1978, 1980, 1981), Teal (1979), Hyde (1980) and Wood (1980), the latter three concerning areas within Superior Province of the Canadian Shield.

Since

> "There is a deplorable lack of detailed stratigraphic and paleogeographic analyses on the basis of physical volcanology and sedimentology in most Archean greenstone belts ..." (Kroner, in Miall, 1984, p.451)

perhaps this thesis can help remedy the situation. Ayres (1969, p.312), in his classic dissertation on Archean geology, wrote that the "Beardmore-Geraldton belt...is probably the best place to attempt a sedimentological study" of the transitions at the edges of subprovinces in northern Ontario.

1.2 Nomenclature

The term "belt" has become vague due to its application to geological units of widely varying scales, from kilometres to hundreds of kilometres. The following terminology will be applied:

1) individual regional to subregional lithostratigraphic units are "belts";
2) "greenstone belts" are metavolcanic-metasedimentary (supracrustal) terranes or sequences, comprising portions or all of certain subprovinces (Anhauesser *et al*, 1969; Condie, 1981; Ayres *et al*, 1985); and
3) subprovince, a term which is preferred over "superbelt".

Belts are analogous to stratigraphic formations or groups, and greenstone belts to groups or supergroups. Stratigraphic nomenclature has not been applied to many greenstone belts because of their structural complexity and a general lack of sedimentological study.

The author believes that avoiding the common term "greenstone belt" can help avoid restrictive connotations that could hinder analysis of the Beardmore-Geraldton area. Use of the alternative term "terrane" emphasizes the atypical nature (Devaney and Frálick, 1985; Williams, 1985) of the Beardmore-Geraldton region. The Onaman-Tashota area is not named as a terrane because of a lack of detailed study.

The study area's northern metasedimentary belt (NMB) and central metasedimentary belt (CMB) (Fig. 1.1) comprise the informally defined "Namewaminikan Group" (Devaney and Fralick, 1985). The abbreviations NMB, CMB and SMB (southern metasedimentary belt) will be used throughout the rest of this thesis. The use of these terms conforms with the existing terminology for the study area (Mackasey, 1975, 1976).

Hereafter the prefix "meta-" will be omitted (except where necessary) as these greenschist facies rocks will be treated as sedimentary rocks

"Mudstone" refers to slate, argillite and siltstone. Coarse siltstone will be referred to as such.

The term "aquabasin" (Devaney and Fralick (1985, p.127) refers to a lake, sea or ocean, and implies a distinction from a purely fluvial basin. This term is useful for Precambrian sediments for which the distinction between lacustrine and marine deposits can be very difficult or impossible to make. "Aquabasinal" is an adjective for such deposits.

1.3 Geology of the Beardmore-Geraldton Area

The Beardmore-Geraldton area (Fig. 1.1) is an Archean metavolcanic-metasedimentary terrane within the Wabigoon

Subprovince of Superior Province. Recent reports and summaries of the regional geology are given by Mackasey (1975, 1976), Mason and McConnell (1983), and Mason *et al* (1985). The area is known for its gold deposits.

The Paint Lake Fault (Mackasey, 1975, 1976), a major structural and lithologic boundary in eastern Wabigoon Subprovince, separates the felsic volcanic-rich Onaman-Tashota area from the Beardmore-Geraldton terrane's long, linear east-striking belts of alternating sediments and mafic volcanics (Fig. 1.1; Stott, 1984a, b).

Mafic volcanic belts in the study area contain massive, pillowed and amygdaloidal flows. Felsic volcanics are uncommon. The three sedimentary belts are:

1) the NMB, a conglomerate-rich assemblage;
2) the CMB, a conglomerate-sandstone assemblage with minor mudstone and iron formation; and
3) the SMB, a sandstone-mudstone assemblage with minor iron formation and conglomerate.

The relative positions of these belts are thought to be the result of a regional fault or fold system (Mackasey, 1976; Williams, 1985). Mackasey (1975), Lavigne (1983), Macdonald (1984), Buck and Williams (1984), Williams (1985) and Anglin and Franklin (1985) have discussed the importance of dextral motions in the Beardmore-Geraldton terrane (Fig. 1.1). Major dextral faults are common throughout Superior Province (e.g. Schwerdtner *et al*, 1979).

The immediate Geraldton area is outside the study area of this thesis and displays a more complex arrangement of lithologic belts (Fig. 1.1; Beakhouse, 1984; Stott, 1984a, b) than near Beardmore. Metasedimentary belts similar to those in the Beardmore-Geraldton area are exposed west of Lake Nipigon, along regional strike, and may be lithostratigraphically correlative units (Tanton, 1935; Pye, 1968; Mackasey *et al*, 1974; Mackasey, 1975, 1976).

The NMB and CMB are the subject of this thesis. The SMB was not examined but because of its lithofacies, particularly its many graded beds, the SMB is likely of largely turbiditic origin (Mackasey, 1975, 1976; Carter, 1983; Kehlenbeck, 1983; Barrett and Fralick, 1985).

Rarity of granitic intrusives, the absence of large-scale diapiric structures and felsic volcanic centres, and the presence of regional-scale, strongly linear supracrustal units in the Beardmore-Geraldton terrane (Fig. 1.1) make it an abnormal greenstone belt, although it may be erroneous to analyze it separately from the Onaman-Tashota area.

The remaining discussion pertains to Fig. 1.1.

Minor dioritic (Archean?) intrusions, Proterozoic intrusions, and most of the faults shown on published geologic maps have been omitted from Fig. 1.1 for clarity and simplicity.

The position of the Wabigoon-Quetico Subprovince boundary is controversial (Card *et al*, 1980; Kehlenbeck, 1983; Williams, 1985) but the traditional and arbitrary position given in Fig. 1.1 is satisfactory for the purpose of this thesis.

The NMB is drawn as separate from the metasedimentary belt that passes through Geraldton (Fig. 1.1). This correlation is based on the uniform lithofacies and composition along the proposed strike length of the NMB, and a lack of evidence for joining the eastern extension of the CMB to the NMB, as Stott (1984a, b) and others have done. Extensive glacial overburden limits argument over these points.

1.4 Previous Work on the NMB and CMB

The early work on the study area's metasediments concerned the "iron ranges" and "Huronian", "Windigokan" and "Timiskaming" conglomerates (Bell, 1869; Parks, 1901; Coleman and Moore, 1907, 1908; Wilson, 1910; Burrows, 1917; Tanton, 1919). Langford (1929) noted the conglomerate units' regional thinning to the south, and postulated a source to the north. As part of an Ontario Department of Mines (ODM) mapping project, Laird (1937) described the southward fining and thinning of the conglomerate strata, and Bruce (1937) offered interpretations of the NMB and CMB: "streams were bringing down gravel and sand to be spread out as coalescing river fans" (Bruce, 1937, p.29).

Further mapping by the ODM resulted in field guides (Mackasey, 1970, 1972), geological reports (1975, 1976) and maps (Mackasey *et al*, 1976a, b). One of Mackasey's field assistants did a minor provenance study (Callander, 1970). Most of the above authors have commented on the highly polymict composition of the conglomerate clasts, and have listed clast lithologies.

Mackasey (1976, p.15) has described the metasediments in the study area (Fig. 1.1) as "three broad east-west striking belts separated by metavolcanic rocks", believing them

> "to have been related to the same depositional basin and to have formed a laterally continuous succession before folding"

and

> "The interfingering relationship exhibited by the conglomerate, feldspathic sandstone and greywacke in the central belt is interpreted to be the result of depositional facies changes. That is, the boundaries shown on the map represent primary sedimentary units rather than that forming by subsequent faulting and folding."
> (Mackasey, 1976, p.20)

Ayres (1969) and Mackasey *et al* (1974) included the Beardmore-Geraldton area's metasediments in their discussions of the nature and evolution of Superior Province's subprovinces and their boundaries.

Metasediments in the Beardmore-Geraldton area outside this thesis' study area have recently been of interest. Conglomerates and sandstones at the eastern limit of the NMB, near Geraldton, have been studied by Devaney (1982) and Beakhouse (1984). Barrett and Fralick (1985) have described the sedimentology of the fine-grained clastic and chemical facies present. Macdonald (1984), Kehlenbeck (1983), Anglin and Franklin (1985) and Williams (1985) have studied the structural history of the metasediments.

1.5 Scope of the Study

> "In Archean studies, the dice are loaded against the sedimentologist because of structural deformation, lack of availability of paleocurrent data, and general absence of long, unbroken, and continuously-exposed stratigraphic sections."
> (Turner and Walker, 1973, p.835)

The most sophisticated contemporary stratigraphic and sedimentological research is based on well-exposed, continuous sections, often in mountainous terrain. Compared to these detailed studies, this thesis is a reconnaissance scale study, with minor detailed sedimentology. This is because of the poor exposure offered by the Canadian Shield: small, erratically spaced outcrops of deformed metasediments. However, the work outlined herein is detailed compared to the usual reconnaissance to semi-detailed mapping that is done in Archean greenstone belts.

It was not the purpose of this study to perform a sedimentary basin analysis. A basin analysis requires:

1) knowledge of the three-dimensional geometry of formation-scale units;
2) the use of context, the relative position of units in more or less conformable sequences; and
3) a reliable chronology.

Such geometry and context are usually unknown for Archean supracrustal sequences. Deformation and the obvious lack of fossils preclude the usual chronologic assumptions and techniques. Only recently have U-Pb zircon dates become popular.

1.6 Difficulties in Mapping and Analyzing Archean Terranes

> "Stratigraphic reconstruction of greenstone belts is a difficult and time-consuming job. Reconstruction is hampered by isoclinal folding, numerous faults, scarcity of marker units, paucity of time controls, discontinuous nature of volcanic units, rapid facies changes related to volcanic centres, and uncertainties about Archean processes."
>
> (Ayres *et al*, 1985, p.3)

In deformed terranes such as Archean greenstone belts one cannot take the normally most basic assumptions (e.g. way up of a bed; stratigraphic continuity within a sequence) for granted; such criteria must be suggested or shown in detail.

Contacts between lithostratigraphic units, particularly formation-scale ones, are rarely exposed. Because of the generally low density of exposure in the Canadian Shield, one does not measure proper stratigraphic sections, precise unit thicknesses are unavailable, and many interesting paleo-subenvironments are no doubt covered by overburden between individual rock knobs. The limitations discussed in Section 1.5 preclude the execution of a highly sophisticated basin analysis in Superior Province.

It is important to carefully examine as many outcrops as possible, ideally all of them. The importance of a given facies need not be proportional to its aerial or volumetric extent; one can step over a potentially critical paleo-subenvironment.

In the Archean, isoclinal folding and transposition of bedding are common. Discontinuities in layered sequences may be marked by nothing more than a cleavage plane. For these reasons, each set of cross-strata, each graded bed, and other more subtle way-up criteria count as one top indicator each. Careful mapping over a large area (Figs. 1.1, 3.1) resulted in the observation/interpretation that the CMB contains a very large number of north-facing stratigraphic top indicators, a consequence of the assumption that stratigraphic continuity cannot be assumed in outcrop sections.

Even when enough outcrops are examined, deformation and small outcrop size lead to a lack of data regarding:

1) primary clast or grain imbrication;
2) primary shape sorting;
3) maximum particle size to bed thickness ratios;
4) maximum clast size to mean clast size (C/M) ratios;
5) lateral facies relationships (including the presence or absence of sheet-like units, at almost any scale); and
6) lack of complete confidence in any seemingly logical and orderly vertical trends, at almost any scale.

The Precambrian's general lack of body and trace fossils is also a hindrance.

Outcrops of Archean strata typically weather along foliation or joint planes, with the result that few bedding planes are exposed, making good paleocurrent data extremely rare (Ojakangas, 1985).

1.7 Field Methods

The author attempted to visit all of the outcrops in the NMB and CMB. Excepting Irwin and east Sandra Townships (Fig. 1.1), this was nearly accomplished. The resulting coverage of the study area is sufficient that any regional trends that are present should have been encountered.

Field work consisted of pace-and-compass traverses and canoeing lakes and rivers. Roads provided good access. Cut lines were rare. Published geological maps, not airphotos, were used for determining location.

Except for clean shoreline exposures, outcrops usually required extensive moss-peeling, cleaning and sweeping (broom and wisk). This was necessary for observing subtle features and obtaining uncluttered photographs. Weathered surfaces were vastly superior to fresh ones. Rarely, water and bleach were used for scrubbing the surfaces.

The scale of mapping varied, depending on the quality of the exposures and the variety of lithofaces present. The four scales used were:

1. bed-by-bed (mm to metre);
2. characteristics of lithologies/lithofacies in one small outcrop (mm, cm to tens of metres);
3. characteristics of lithologies/lithofacies in a set of outcrops (tens to hundreds of metres); and
4. reconnaissance (brief examination of lithology, clast size, bedding and foliation).

Systematic description of the outcrops was performed by using a checklist (Table 1.1) specially designed for Archean coarse clastics. Beds and clasts were measured with a tape measure. The exposed, apparent length in mm of the long (along strike of schistosity) and short (perpendicular to schistosity) axes of the largest clasts were recorded; the flat to rounded, two-dimensional exposures of the elongate clasts provided only rare three-dimensional views of oblate clasts. Apparent (horizontal) bed thicknesses and short axis clast dimensions were later converted to true thicknesses by multiplying the measurements by the sine of the dip of bedding and foliation respectively. Besides measuring the apparent long and short dimensions of an outcrop area's ten largest clasts (giving a D10 average) and noting their lithologies, the dimensions of the largest three quartz, chert and jasper clasts were recorded. Clasts of unusual composition were also noted. Average deformed clast size was a visual estimate, checked by measurements. Sandstone grain size was always carefully measured with a mm scale, and sorting (standard deviation of grain size) estimates were always checked against a standard reference diagram for consistency. Sedimentary structures were described in terms of both their actual (cm) and relative (thicker than ...) dimensions, and were usually photographed. Mesoscopic structural features, such as kinks, tension gashes, folds, faults, and the orientation of the schistosity were noted.

TABLE 1.1

1) LITHOLOGY
  1 clast-supported polymict conglomerate
  2 matrix-supported polymict conglomerate
  3 pebbly sandstone
  4 pebble-free sandstone
  5 pebble bands in sandstone
  6 massive sandstone (Sm)
  7 planar, horizontal and parallel laminated sandstone (Sh)
  8 cross-stratified sandstone (St, Sp, Sr)
  9 siltstone
  10 argillite
  11 slate
  12 banded iron formation (BIF), magnetite
  13 BIF, hematite
  14 breccia
  15 oligomict conglomerate

2) BEDDING
  1 massive)
  2 crude) conglomerate only
  3 well-developed)
  4 cobble bands)
  5 finer bands
  6 alternating coarse and fine layers
  7 bed sole(s)
  8 parallel, horizontal, planar)
  9 tapering) outcrop scale
  10 channeling, scouring)
  11 sandy horizons
  12 sandy horizons as a lateral extension of sandstone
  13 sandstone more pebbly as bed tapers
  14 sandstone bed ends abruptly

STRIKE AND DIP OF BEDDING AND SCHISTOSITY

3) STRAINED BED THICKNESS
  1 apparent thickness
  2 true maximum bed thickness, in cm
  3 tapers to other thicknesses, in cm
  4 bed length (and length : thickness ratio), in m

4) NORTH AND 5) SOUTH BED MARGINS
  1 sharp vs 2 irregular (note scale, lateral variability)
  3 gradational via % of clasts
  4 gradational via grain size
  5 planar
  6 concave north
  7 concave south

Table 1.1 (Cont'd)

6) STRAINED D10
   - in mm
   - list long and short diameters and lithology

7) APPROXIMATE AVERAGE DEFORMED CLAST SIZE
   1 cobbles (64-256 mm)
   2 very large pebbles (32-64 mm)
   3 large pebbles (16-32 mm)
   4 medium pebbles (8-16 mm)
   5 small pebbles (4-8 mm)
   6 granules (2-4 mm)
   7 very coarse sand (1-2 mm)
   8 coarse sand (0.5-1 mm)
   9 medium sand (0.25-0.5 mm)
   10 fine sand (0.125-0.25 mm)
   11 very fine sand (0.06-0.125 mm)

8) SORTING
   1 conglomerate framework
   2 conglomerate matrix
   3 sandstone
   4 non-sorted to extremely poorly sorted
   5 very poorly sorted
   6 poorly sorted
   7 moderately sorted
   8 well sorted
   9 very well sorted

9) GRADING
   1 ungraded
   2 coarse-tail
   3 distribution
   4 fines north
   5 fines south
   6 developed at base of bed
   7 developed at top of bed
   8 developed throughout bed

10) SEDIMENTARY STRUCTURES
   - quantitative (in cm's)
     and qualitative (relative to *what*)
   - paleocurrents

11) LAMINATION
   1 laminated vs 2 thinly bedded
     (1 cm thick boundary)
   3 maximum laminae thickness, in mm
   4 other laminae thicknesses, in mm
   5 planar vs 6 undulatory/wavy
   7 horizontal vs 8 inclined

Table 1.1 (Cont'd)

11) LAMINATION (Cont'd)
  9 parallel vs 10 convergent, non-parallel
  11 lateral continuity > 1 m
  12 poor lateral continuity
  13 concave north
  14 concave south
  15 coarser laminae thicker than finer ones
  16 finer laminae thicker than coarser ones
  17 coarser and finer laminae same thickness
  18 fines north
  19 fines south
  20 rippled

12) PEBBLE BANDS (relative position in sandstone bed)
  1 parallel to bed margins
  2 oblique to bed margins (give angle)
  3 laterally continuous
  4 laterally discontinuous
  5 tapering (where to?)
  6 typical number of pebbles thick
  7 bed thickness, in mm

13) STRATIGRAPHIC TOPS
  1 north
  2 south
  3 sharp conglomerate sole
  4 conglomerate fines upward
  5 sandstone fines upward
  6 cross-bedding
  7 channel
  8 scour

14) MISCELLANEOUS
  1 gneissic clast(s)
  2 multi-lithology clast(s)
  3 pillowed mafic clast(s)
  4 epiclastic sediment clast(s)
  5 largest clasts are granitoids
  6 vein quartz clasts present

15) STRUCTURAL GEOLOGY

Table 1.1 Checklist for field mapping of Archean coarse clastic metasediments.

Diagrams and stratigraphic sections of the best outcrops were drawn or recorded.

D10 measurements. There are some limitations involved in measuring the average size of the ten largest conglomerate clasts (D10) in an outcrop area. Although the apparent thickness (measured horizontally) of a strained clast is easily converted to its true thickness (measured perpendicular to schistosity) when the dip of the schistosity is known, one never knows how closely the true thickness measured in outcrop approximates the maximum dimension of the strained clast's axis perpendicular to the clast schistosity. The maximum dimension may be hidden in the rock or it may have been eroded away, a consequence of the "cut effect" (e.g. Pettijohn, 1975, Fig. 3-38).

Accurate estimation of clast shape is also a problem. Clasts in the typically horizontal to low-angle outcrop surfaces are elongate along the schistosity (excepting tectonically rotated clasts). Similar views in vertical planes passing through the line of dip, planes generally perpendicular to the outcrop surfaces, suggest that the clasts in the NMB and CMB are characteristically oblate. The few clasts that could be removed from the outcrops were oblate (plus very rare spherical granitoid clasts). Because of the cut effect mentioned above entire clasts are usually not seen, but assuming that the clasts are typically oblate seems reasonable in light of the three-dimensional outcrop views seen throughout the study area. Nevertheless, the presence of some prolate clasts could confuse the data.

There are probably over 10,000 clasts exposed by erosion surfaces in almost any given 40 square metre area of conglomerate exposed in the study area. At D10-measured sites it was thus easy to quickly scan many tens of thousands of clasts and measure the apparent dimensions of the largest fifteen or twenty clasts. Because so many clasts were scanned and measured, it is thought that: 1) the D10 measurements of bouldery and cobble-rich coarse, strained conglomerate should be larger than the D10 measurements of fine pebble, similarly strained conglomerate; and 2) the D10 measurements should quantitatively reflect what is obvious from outcrop observations (i.e. bouldery vs. very small pebble conglomerates).

Different degrees of deformational flattening of clasts according to the different clast lithologies (and thus competencies) is not a problem in the study area because the largest clasts (8, 9 or 10 of the D10 set) are usually granitoids. Thus, comparing the D10 measurements of outcrop areas involves mostly comparing the apparent dimensions of the coarsest, oblate (ideally) granitoid clasts.

Different degrees of clast deformation owing to different amounts of strain could confuse any apparent regional sedimentological trends. If the apparent deformed clast size (sub-)regional trends or patterns correspond directly with other objective sedimentological trends, it will become reasonable to assume that clast deformation has not seriously obfuscated any apparent clast size trend(s).

Measurements of D10 are often performed on undeformed conglomerates and have produced meaningful results (e.g. Pettijohn, 1975, p.515). It seems worthwhile trying the same for deformed conglomerates in order to see whether or not the results will be consistent with other sedimentary criteria.

1.8 Provenance Point Counting Field Methods

At 31 selected sites in the NMB and CMB (Fig. 10.1), point counting of conglomerates was performed. This required clean surfaces and often containers and bags of water had to be carried overland hundreds of metres from the nearest lake, stream or road. A few square metres were then rinsed and scrubbed free of dirt.

A movable grid of two perpendicular tape measures was set up at each site. One tape acted as a fixed base line, the other as a mobile traverse line. The grids were approximately 2 metres by 1 metre. A 5 cm spacing of grid points was most often used, as this allowed most of the clasts to be counted only once; the largest clasts spanned several grid points. Using code numbers for the clast's lithologies, 500 non-matrix points at each point count location were categorized. An assistant recorded the data on a grid sheet of 500 squares. The number of grid points lying on inter-clast matrix was also recorded, giving clast : matrix ratios. Usually 600 - 750 points needed to be counted to reach 500 non-matrix points. Early in the field season the point counts took two to three hours each, but in the latter half of the season 45 - 60 minutes was normal; these times do not include traversing and site preparation. Photos were taken of most of the sites.

CHAPTER 2: THE NMB

2.1 Extent and Character of the NMB

The northern metasedimentary belt (NMB; Figs. 1.1, 2.1) extends from Corrigan Creek in Sandra Township to the Volcan Lake area near Geraldton, a distance of approximately 60 km. Because the strata are usually very steeply dipping, the 150 - 900 m north to south extent shown on published geological maps approximates the stratigraphic thickness of the tectonized NMB.

Outcrops referred to in the text are shown in Fig. 2.1.

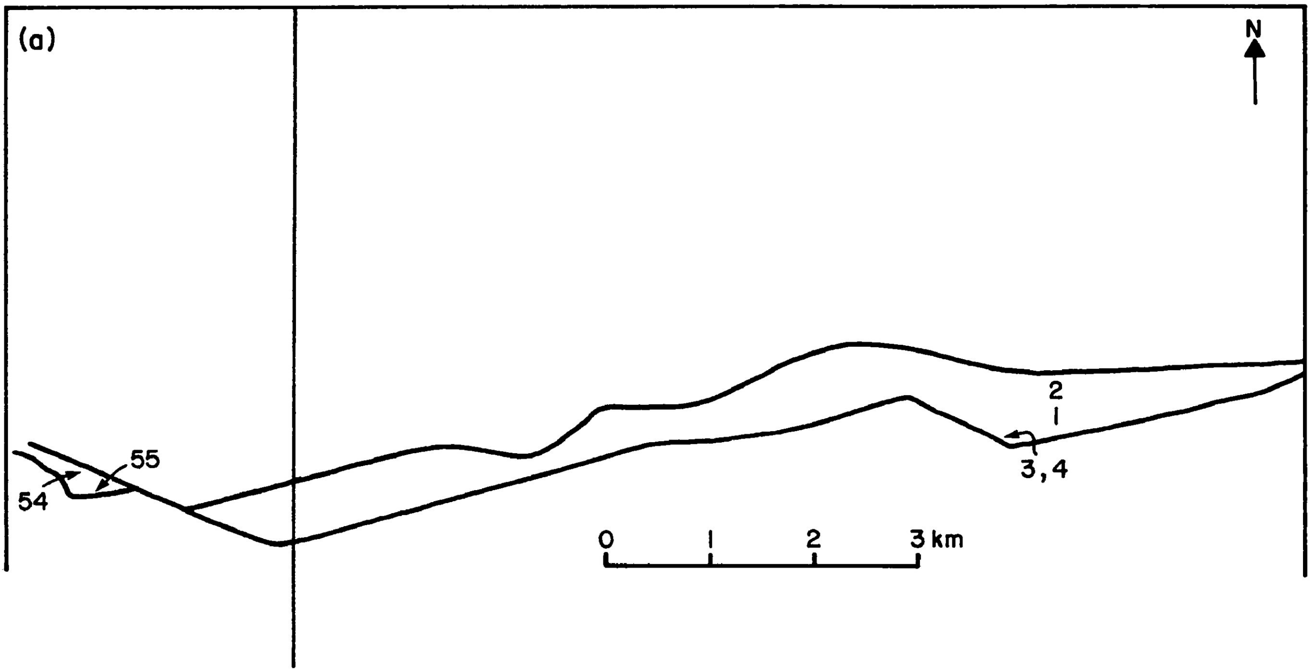
As Tanton (1935), Pye (1968), Mackasey *et al* (1974), and Mackasey (1975, 1976) have noted, a sequence of lithologic belts similar to that in the Beardmore-Geraldton area is exposed west of Lake Nipigon, along regional strike. Strata in these western belts include conglomerate very similar (identical?) to the NMB's, enhancing the possible long-range lithostratigraphic correlation.

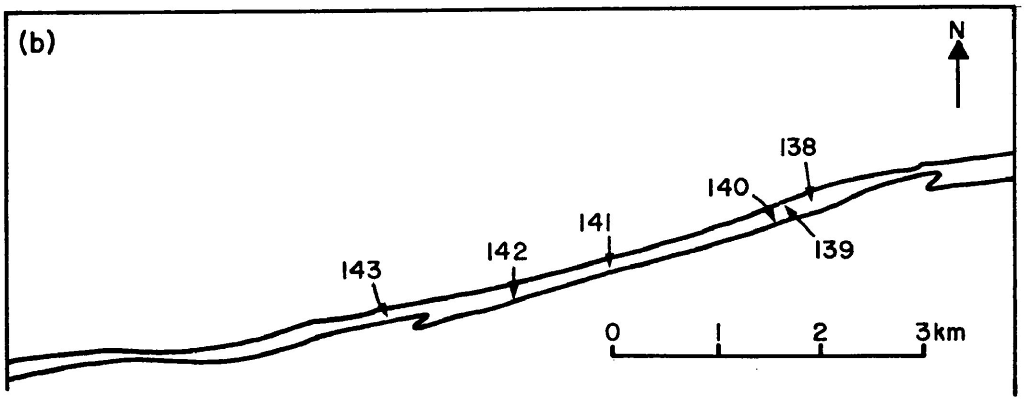
The lithofacies assemblage comprising the NMB is dominated by massive to crudely bedded, clast-supported polymict conglomerate, interbedded with a relatively minor amount of massive, plane laminated and cross-bedded pebbly sandstone (Figs. 2.2 to 2.5). In each exposure, large or small, sandstone beds usually constitute 0 - 10% of the section, rarely up to 33%.

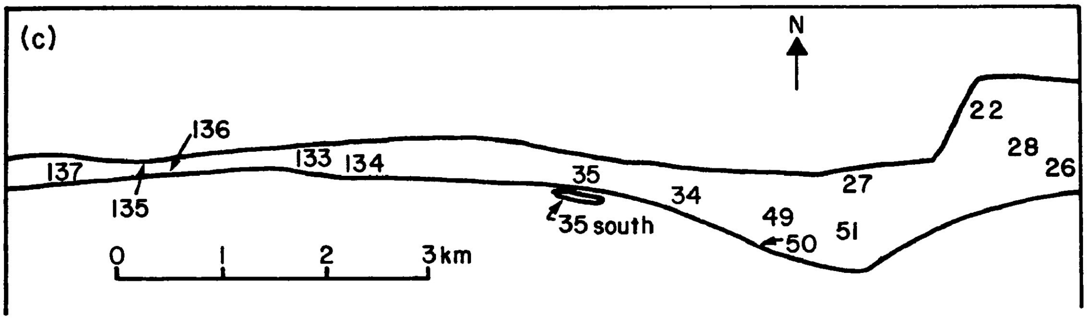
The polymict conglomerate is rich in felsic and volcanic

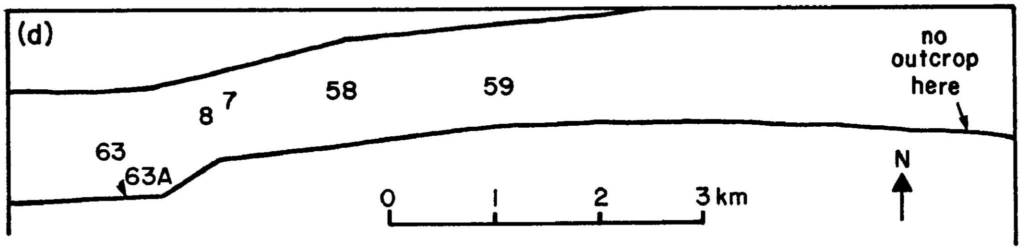
Fig. 2.1: Locations of numbered outcrops of the NMB mapped and referred to in the text (Chapters 2, 10). Modified from Mackasey (1975, 1976) and Mackasey *et al* (1976a, b). Refer to Fig. 1.1 for locations of the Townships.

a) eastern Sandra and Irwin Townships;
b) Walters Township; the Paint Lake Fault (Fig. 1.1; Mackasey, 1976) forms most of the NMB's north boundary here;
c) Leduc Township; the Jellicoe Fault (Mackasey, 1976) offsets the NMB on the northeast corner of the Township;
d) Legault Township; note that the NMB is here not correlated with the volcaniclastic sediments (Chapter 9) to the south of and parallel to the NMB, as Mackasey *et al* (1976a) did; and
e) Colter Township.


(a)
N
55
54
2
1
3, 4
0
1
2
3 km


(b)
N
138
140
139
141
142
143
0
1
2
3km


(c)
N
136
137
135
133
134
35
35 south
34
22
28
26
27
49
50
51
0
1
2
3km


(d)
8
7
58
59
63
63A
no outcrop here
N
0
1
2
3km

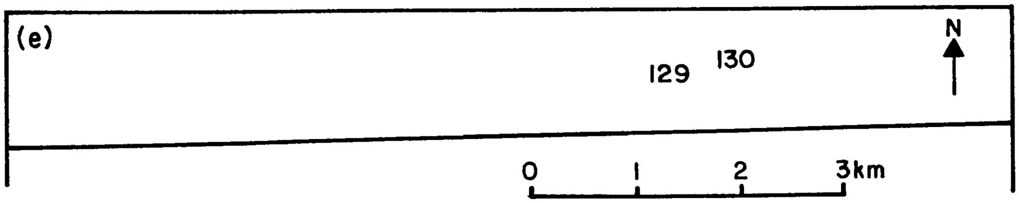

(e)
129
130
N
0
1
2
3km

Fig. 2.2: Clast-supported polymict conglomerate interbedded with sandstone. Middle sandstone "bed" is actually two beds, with the finer and darker one to the right and separated by an abrupt, undulose bed contact. Planar cross-bedded sandstone is poorly visible at lower right centre. Note the large cobble at the lower right. Daypack for scale. Outcrop #7.

Fig. 2.3: Typical schistose, clast-supported polymict conglomerate. Outcrop #35, site of point count #14.

Fig. 2.4: Crudely bedded, clast-supported polymict conglomerate. The lens of finer pebbles is parallel to nearby conglomerate-sandstone bed contacts. Outcrop #1.

Fig. 2.5: Bouldery, clast-supported polymict conglomerate. The largest clasts are mostly granitoids. Lichen obscures some of the smaller clasts. Outcrop #55.

CENTIMETRE
INCH

clasts, as discussed in Chapter 10.

Strata are generally east-striking (Mackasey, 1975, 1976; Mackasey *et al*, 1976a, b). Easily recognizable, well developed stratification is defined by sharp bed-to-bed transitions from conglomerate to sandstone (Fig. 2.2), and rarely by sandstone to sandstone transitions. More subtle stratification is described below under "crude bedding". The well developed and crude "horizontal" stratification outline the paleohorizontal.

Structural Features. The strike of schistosity and bedding are usually very similar. Schistosity dips are often steeper than bedding dips.

Small outcrop-scale folds, excepting kink folds, are very rare and are present mostly in the NMB's eastern exposures (outcrops #7, 8 and 59). Fig. 2.6 shows folded conglomerate and sandstone beds. In the study area, transposition of bedding along the schistosity rarely produces pebbly "bands", but such folding and transposition can usually be recognized in outcrop, thus avoiding confusion with primary pebble bands (see Section 2.5). In obviously folded areas, fold hinge regions and schistosity-transposed slices have been avoided in favour of the limbs.

The fabric of the conglomerate is described in Sections 2.2 and 2.4. The deformed clasts have consistently high L/S ratios (see Table 2.1) in Walters and west Leduc Townships, immediately along the Paint Lake Fault (Fig. 1.1).

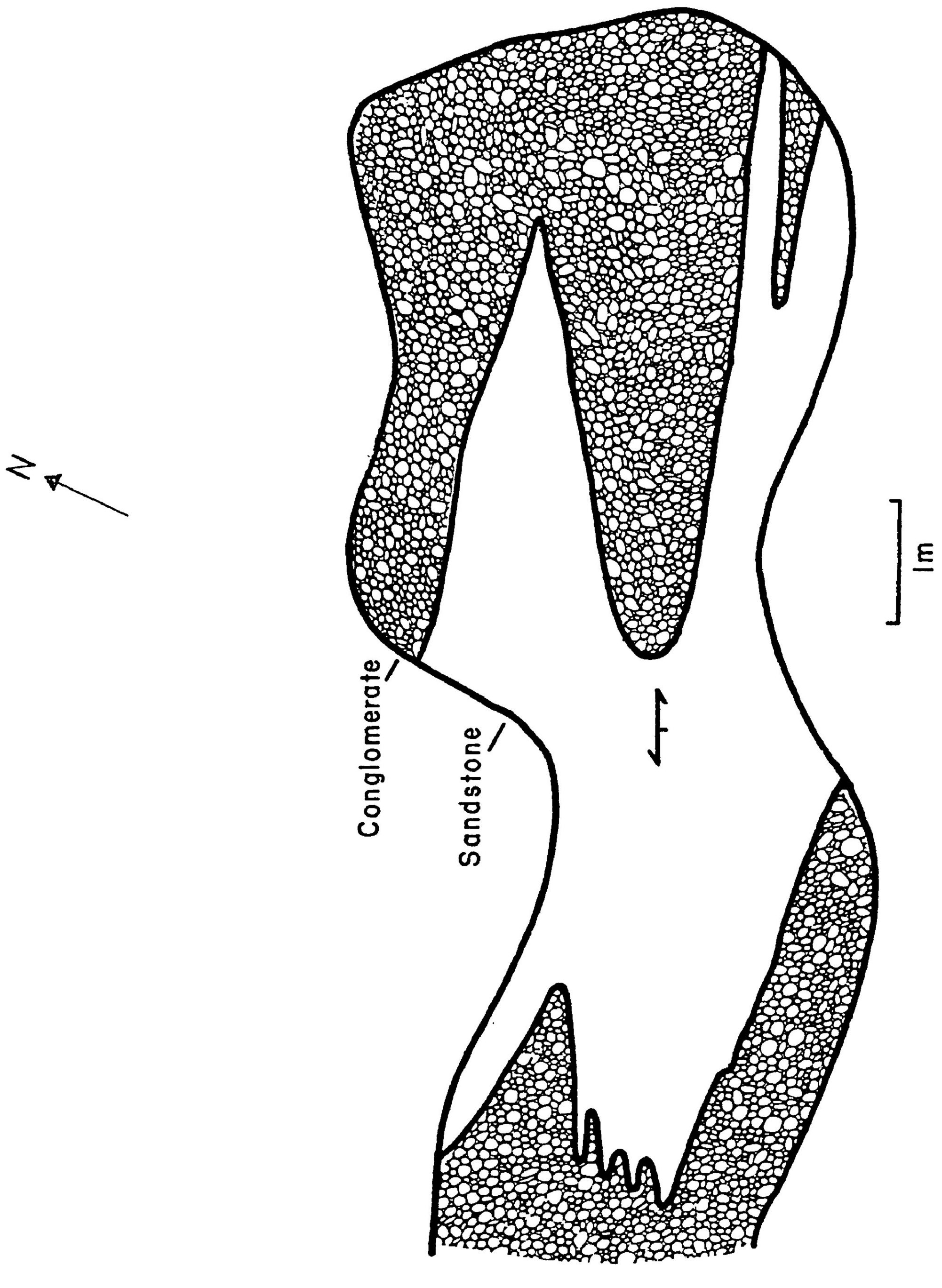

N
Conglomerate
Sandstone
1m

Fig. 2.6: Plan view of part of Outcrop #8, showing folded beds. The schistosity strikes at 64°.

2.2 Conglomerate Lithofacies

Textures. The conglomerate is always a tightly packed clast-supported framework of gravel clasts, modified by deformation (Fig. 2.3, 2.4). Primary fabrics such as imbrication have been tectonically obliterated.

Pebbles composed of softer, less mechanically competent lithologies are wrapped or bent around harder, more competent clasts. The softer clasts have greater long : short axis ratios. A few granitoid clasts appear to be perfect spheres.

The conglomerate is considered a bimodal assemblage of gravel clasts and sand grains; sand grains are not considered when estimating clast-supported conglomerate sorting.

Clast frameworks are poorly to moderately sorted. Less often, and usually where cobble-bearing, the conglomerate is very poorly sorted to unsorted. Changes in sorting can define subtle crude bedding.

The beds are nearly always ungraded. A previous study (Devaney, 1982) found some subtly fining-north beds within a 125 m section of the NMB east of the study area.

The matrix to the framework pebbles is poorly to moderately sorted, coarse or medium to fine-grained sandstone. It is always identical in texture and composition to nearby sandstone beds.

Stratification. Bed margins are sharp. There are no gradual transitions from conglomerate to sandstone by way of either clast/grain size or percentage of gravel clasts.

Occasionally the north margins of these east-west striking beds are sharper than the south margins, as shown by sandstone to conglomerate transitions that are slightly more abrupt in the percentage of gravel clasts per cm of section (Fig. 2.7).

Very few conglomerate bed thicknesses were recorded. This is a function of the small size of the outcrops, the beds often being as large or larger than the scale of aerially continuous outcrop (usually metres). Also, it is more difficult to identify subtle crude bedding in foliated, poorly sorted, coarse polymict conglomerate than in undeformed strata.

Crude bedding. This is defined by clasts of a specific size range that are different from the immediately surrounding clasts or beds.

Crude bedding can be vaguely to well-defined, depending on the magnitude and abruptness of the change in average and/or maximum clast size. Large and abrupt changes are more easily recognized, as opposed to minor changes representing very subtle layering. These changes occur both vertically, in section, and laterally, along strike, but the typically small outcrops display vertical changes more often.

Two types of crude bedding are present:

1) thin layers distinct from much larger surrounding areas. As this type becomes more subtle, it is transitional to massive, poorly sorted, clast-supported conglomerate; and

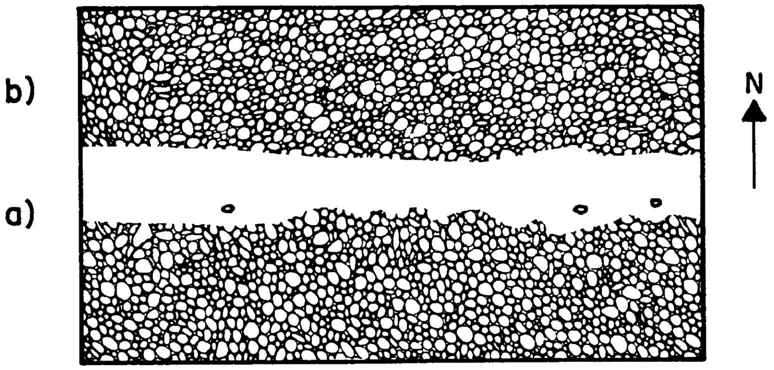

b)
a)
N

Fig. 2.7: Sand/sandstone bed with a sharp inferred top and a less sharp inferred base:

a) a non-erosively based bed of pebbly sand buried underlying gravel, followed by

b) a sharp, erosively based bed of gravel. The study area's (NMB, CMB in Fig. 1.1) east-striking fluvial sandstone beds interstratified with conglomerate usually have sharper north margins where a sharp vs. less sharp asymmetry is apparent; such beds suggest stratigraphic tops to the north.

2) thick, laterally consistent beds of different average and/or maximum clast size. This type is transitional to well-developed bedding.

Layers of type 1 (above) are more abundant. They consist of cobble bands within a finer background of clast-supported conglomerate, fine pebble (2-16 mm) bands within coarser clast-supported conglomerate (Fig. 2.4), and sand "pods".

Sand pods are the most minor of sand lenses, only cms thick and wide. In section they are abnormal concentrations of sandy conglomeratic matrix that act as sandstone micro-beds. They are usually parallel to nearby well-defined bedding. Sometimes two or more pods occur along strike, defining a sandy horizon separating conglomerate beds.

2.3 Anomalous Conglomerate Lithofacies

Fine felsic rudite. Outcrop #50, two small sites 100 m apart, contains anomalous, clast-supported "conglomerate" composed of felsic volcanic and quartz pebbles. No jasper pebbles were seen, which are a minor but highly recognizable and characteristic constituent of the NMB's usual polymict conglomerate. The abnormally fine-grained (average 2-8 mm, short axis), poorly sorted framework clasts have a D10 of only 20/51 mm (Table 2.1).

This outcrop may be part of a felsic volcanic, not sedimentary, unit.

Very coarse, mafic clast-rich conglomerate. The following is taken partly from Devaney and Fralick (1985, p.127).

TABLE 2.1

Township	Outcrop #	D10 (in mm)	L/S
		S / L	
Sandra	54	146 / 257	1.76
	55	209 / 358	1.71
Irwin	3	161 / 263	1.63
	1	109 / -	-
Walters	143	105 / 274	2.61
	142	201 / 360	1.79
	141	249 / 495	1.99
	139	201 / 428	2.13
	138	186 / 444	2.39
Leduc	137	133 / 301	2.26
	135	80 / 192	2.40
	136	183 / 308	1.68
	133	100 / 168	1.68
	134	187 / 338	1.81
	35	114 / 219	1.92
	35	112 / 208	1.86
	49	93 / 231	2.48
	50	20 / 51	2.55
	27	115 / 174	1.51
	28	162 / 271	1.67
	22	58 / 164	2.83
Legault	63	91 / 157	1.73
	63A	362 / 842	2.33
	8	180 / 300	1.67
	8	114 / 203	1.78
	7	101 / 154	1.52
	58	87 / 147	1.69
	59	109 / 258	2.37
Colter	129	82 / 124	1.51
	130	112 / 179	1.60

Table 2.1: Measurements of average maximum deformed clast size (D10) in conglomeratic outcrops of the NMB. Locations of Townships and outcrops are shown in Fig. 1.1 and 2.1. D10 measurements, the average size of an outcrop's ten largest clasts, are given in short (S) and long (L) axes, each in mm, with the short axis usually perpendicular to the schistosity and the long axis usually parallel to the schistosity (see Section 1.7). "L/S" is a measure of the ellipticity of the exposed cross-section of the coarsest clasts.

In Legault Township (Fig. 1.1), the contact between the NMB and the mafic volcanic belt to the south is exposed in one small outcrop (Mackasey *et al*, 1976a), outcrop #63A. The contact is a razor-sharp line on a flat surface. North of of the contact mafic boulders, cobbles and pebbles are abundant (up to greater than 50% of the clasts by area) within the clast-supported, polymict conglomerate. The smaller clasts are mostly felsic (felsic volcanics, and a few quartz pebbles). In this outcrop the ten largest clasts are all of mafic composition. It is the *only* outcrop in the entire NMB in which the ten largest clasts are not granitoids or felsic volcanics, and it is by far the coarsest conglomerate exposed in the NMB (Table 2.1). Because of the extreme coarseness, sorting is very poor to almost nonexistent. In places, conglomerate of more normal felsic-rich composition forms a matrix to the mafic mega-clasts.

Immediately south of the contact the mafic volcanics appear highly strained and veined, as might be expected at such a lithologic contrast in a deformed terrane. The contact is defined by contrasts in colour, composition, texture, fabric, and weathering. The conglomerate is slightly more resistant to weathering than the mafic volcanics just as the mafic clasts in the more typical conglomerate outcrops are recessively weathered. About 25 m south of the contact mafic pillow lavas are well exposed.

2.4 Clast Size

As discussed in Section 1.7, the conglomerate's gravel-sized clasts are considered to be oblate (ideally) spheroids flattened in the plane of the schistosity. Average size of the deformed clasts, measured perpendicular to the schistosity (true thickness of apparent short clast dimensions; see Section 1.7) is in the 16 – 64 mm range. The average size of the ten largest clasts at a site (D10) is quite variable: cobble-poor, cobble-rich, and bouldery conglomerate are present in the NMB.

A D10-measured "site" is an outcrop area, a set of small individual outcrops which have their data grouped together. Some outcrop areas are much smaller than others; the author sacrificed having a standard small sampling area in order to obtain regional coverage. There may be a slight bias towards larger D10 values from larger outcrops areas.

The largest eight, nine or ten clasts at a site are nearly always granitoids. Felsic volcanic clasts usually comprise the rest of the D10 set. Because of the nearly uniform D10 clast lithologies the results are consistent and outcrops can be compared.

Table 2.1 shows the regional variation in D10 measurements. The NMB's conglomerate clasts are very coarse in Walters Township (Fig. 2.8), and coarsest at outcrop #63A in Legault Township. Fig. 2.5 shows a boulder-bearing outcrop.

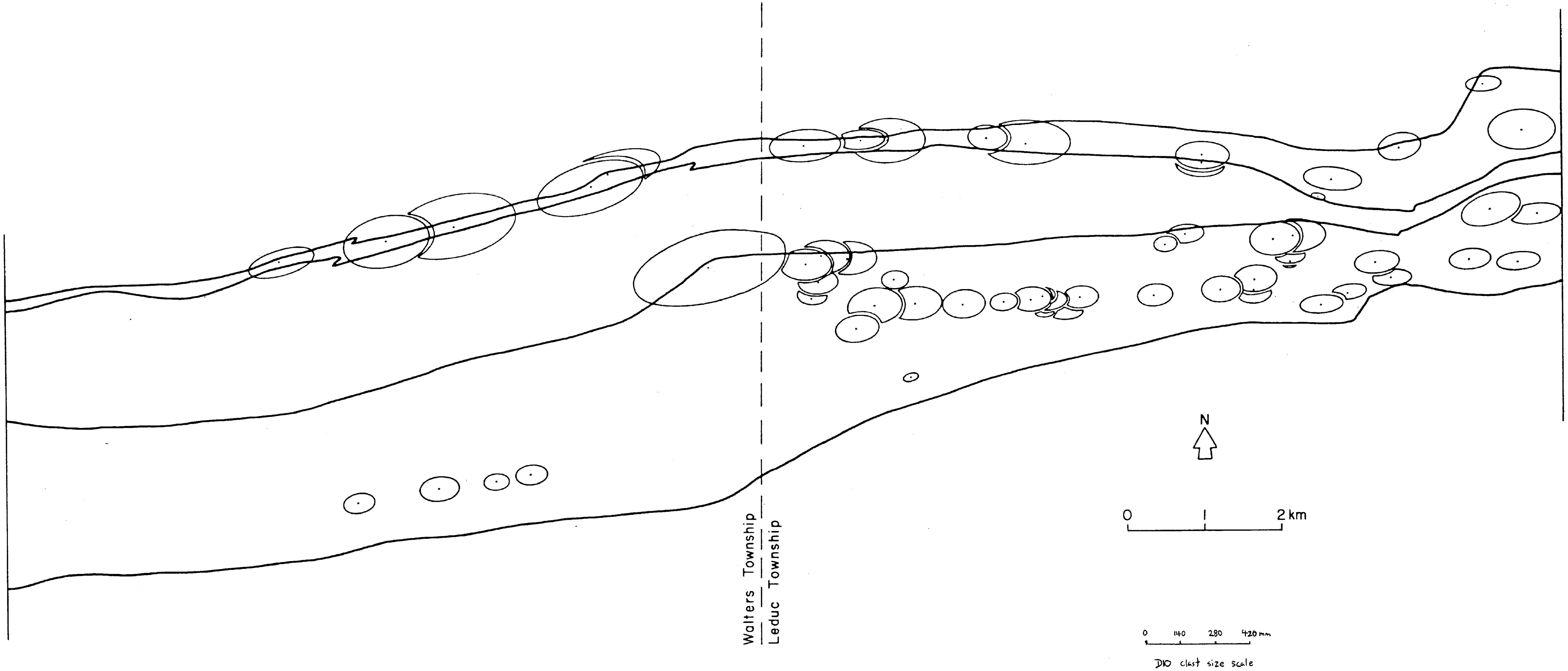

Walters Township
Leduc Township
N
0
1
2 km
0
140
280
420 mm
D10 clast size scale

Fig. 2.8: Measurements of D10 (average maximum deformed clast size of an outcrop area; see Section 1.7) in the conglomeratic outcrops of the NMB and CMB in Walters and Leduc Townships (Fig. 1.1; for outcrop locations see Figs. 2.1 and 3.1). Tables 2.1, 6.2 and 6.3 give the numerical data this diagram is derived from.

2.5 Sandstone Lithofacies

Lithology. Coarse to medium-grained, poorly to moderately sorted pebbly sandstone beds form a relatively minor part of the conglomerate-rich outcrop sections, usually less than 10% sandstone.

Very coarse to fine-grained, more poorly sorted beds are less common. Moderately sorted beds are surprisingly abundant. Some beds are pebble free.

The sand grains are angular to subround. Equant quartz grains have probably retained their primary shapes. Feldspar grains are equant to elongate, the latter along the foliation. Rock fragments are elongate along the foliation. These beds appear to be "arkosic", feldspathic arenites (classification of Okada, 1971). The arenite-wacke distinction (15% matrix) is not critical here, and in Archean sandstones metamorphism makes the distinction difficult, impossible or unreliable. Mackasey (1975, 1976) has given representative thin section descriptions and photomicrographs of this sandstone.

Stratification. Small exposures show beds with parallel margins (north and south contacts), while larger exposures often show a gradual lateral tapering. The actual lensing out of sandstone beds is frequently seen.

Strained bed thicknesses range from 10-60 cm at the thickest parts. Outcrops were usually too small to measure bed lengths/widths. Bed contacts tend to be sharp (see

Section 2.2).

Smaller sand lenses qualify as well-developed bedding because they are more laterally consistent (tens of cms or more along strike) and well defined (conglomerate to sandstone lithofacies change) than sand pods (see Section 2.2).

Pebble bands within sandstone beds are usually parallel to bed margins, less than 5 cm thick, and 1-5 pebbles thick. Laterally continuous pebble bands appear "well-developed", and discontinuous ones "poorly developed". The continuity has two scales:

1) small scale: bands with pebbles in contact with each other are well-developed; and
2) large scale: bands extending for metres along strike are well-developed, versus poorly-developed, laterally discontinuous pebbly horizons.

These pebble bands are nearly always composed of relatively small clasts, and their appearance is identical to such bands in similar undeformed strata, arguing against any possible structural origin (see Section 2.1).

<u>Sedimentary structures</u>. The beds are, in order of decreasing abundance, massive (structureless), plane laminated, and cross-bedded.

Plane laminated beds display planar, horizontal and parallel layers less than or equal to 1 cm thick, varying from faintly to well defined. Coarser (coarse-, medium-grained) and finer (medium-, fine-grained) laminae are revealed by careful examination, recessive weathering, and rarely by rusty weathering of pyrite.

Cross-bedding is very rare in the NMB. No trough cross-beds were found. (Differentiating trough and planar cross-beds can be difficult in these largely two-dimensional exposures, a problem discussed in Section 5.4.) Outcrop #7 contains a 20 cm thick coset of two sets of planar cross-strata. Excepting the inclination, foreset laminae have the same characteristics as the plane laminae described above. Concavity of the foreset-toeset laminae give tops to the north.

Some pebble bands may be low-angle (less than 10°) foreset deposits.

2.6 Anomalous Lithofacies Associations

Irwin Township's anomalous mudstone and graded sandstone Finer grained sandstone and mudstone are present in outcrop #2 and north of the interbedded conglomerate and sandstone in the southern part of outcrop #1. Mackasey's (1975) mapping of these sites is partly erroneous.

This finer, pebble-free sandstone is as coarse as medium to fine-grained. Lamination in the sandstone and mudstone is planar, horizontal and parallel. Some beds approximately 1 cm thick show grading via coarse tail concentrations along sharp bed bases, and show tonal darkening towards finer parts. Some of the grading is accompanied by parallel laminae, suggesting partial Bouma sequences (ADE). Many of the thinly layered strata appear highly deformed, with small (cms) isoclinal fold hinges.

It should be noted that this thinly-bedded to laminated mudstone, sandstone and graded sandstone assemblage has *not* been found/seen interbedded with the conglomerate to the east, west and south in the NMB. Unless purely structural displacement is solely responsible, these unusual strata are a result of primary facies changes, changes that are considered regionally minor because of their limited extent, 3-5 km along strike in east Irwin Township, compared to the rest of the 60 km long NMB.

Very few exposures of the NMB in Irwin Township were examined because of access, logistical and financial problems. Mackasey's (1975) descriptions of the conglomerate-sandstone lithofacies assemblage in this area show it to be the same as the rest of the NMB.

Intercalated sediments and volcanics. Mackasey (1976) shows a conglomerate lens over 400 m long within the mafic volcanics immediately south of outcrop #35 in Leduc Township. A discontinuously exposed section was measured across these units, from north to south:

1) (north) 0-18.5 m: typical NMB conglomerate and sandstone;
2) 18.5-57.5 m: mafic volcanics, tuffaceous in places;
3) 57.5-81.0 m: a 23.5 m thick lens (?) of typical NMB conglomerate and sandstone; and
4) (south): mafic volcanics.

The mafic tuff has a bluish weathered surface. It is coarse to medium-grained, with faint laminae of coarser grains, and laterally discontinuous pebble bands one pebble thick. These clasts are 5-10 mm thick (short axis) and are felsic to mafic (volcanic?).

In order to search for evidence of local derivation of clasts, such as abnormal enrichment in the percentage of a mafic component, point counts of conglomerate at the 0 m and 65.5 m levels (PC-14 and PC-26, respectively, in Table 10.2) were performed, showing the latter to have approximately twice as many mafic clasts by area, but not the highest percentage in the NMB. This result and a lack of structural control make the case for local derivation inconclusive.

Mackasey (1976) shows another conglomerate lens about halfway between outcrops #35 and 31, north of Leduc Lake. The eastern part of this lens could not be found, and a possible western part was briefly examined. It does not, as Mackasey's (1976) map legend implies, resemble the local conglomeratic units at all, and is likely a volcanic breccia.

CHAPTER 3: INTRODUCTION TO THE CMB

3.1 Extent and Character of the CMB

The lithostratigraphy of the central metasedimentary belt (CMB; Figs. 1.1, 3.1, 3.2) was mapped by Mackasey (1975, 1976) and his assistants. Mackasey (1975, 1976) recorded few sedimentological details and offered no environmental interpretations.

The CMB extends 43 km along strike, from Lake Nipigon to Legault Township. It may extend further east but there is little field evidence for this. Using Mackasey's maps (1975, 1976), it is estimated that the CMB is 1-2 km thick. Since a proper stratigraphic section cannot be measured across these discontinuously exposed, deformed rocks, a thickness estimate must suffice.

The generally east-striking strata dip moderately (40°-50°) to moderately steeply (60°-70°) northward, with some sites at or near vertical.

Mackasey (1975, 1976) mapped and commented on the CMB's coarsening-north trend (Figs. 3.1, 3.2). Overall it is well-defined for about 34 km along strike, containing:

1) a southern portion composed of mudstone (argillite/slate), sandstone, and minor iron formation;
2) a central sandstone-rich part, with conglomerate beds, pebble bands, and vary rare mudstone; and
3) a northern conglomerate-rich part, very similar in

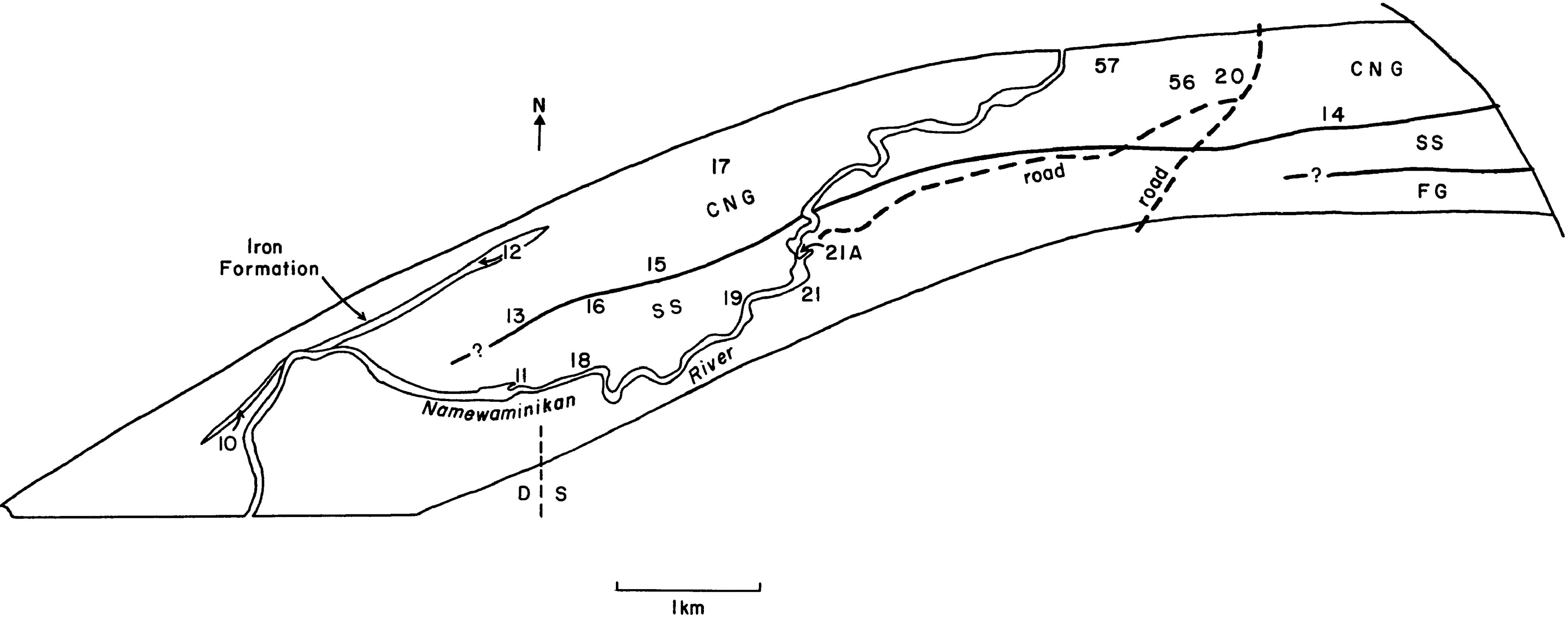

N
57
56
20
CNG
14
SS
?
FG
road
road
17
CNG
Iron Formation
12
15
21A
13
16
SS
19
21
?
11
18
Namewaminikan
River
10
D
S
1km

Fig. 3.1(a): Locations of outcrops studied and lithofacies assemblages of the CMB in Dorothea and Sandra Townships. Modified from Mackasey (1975).

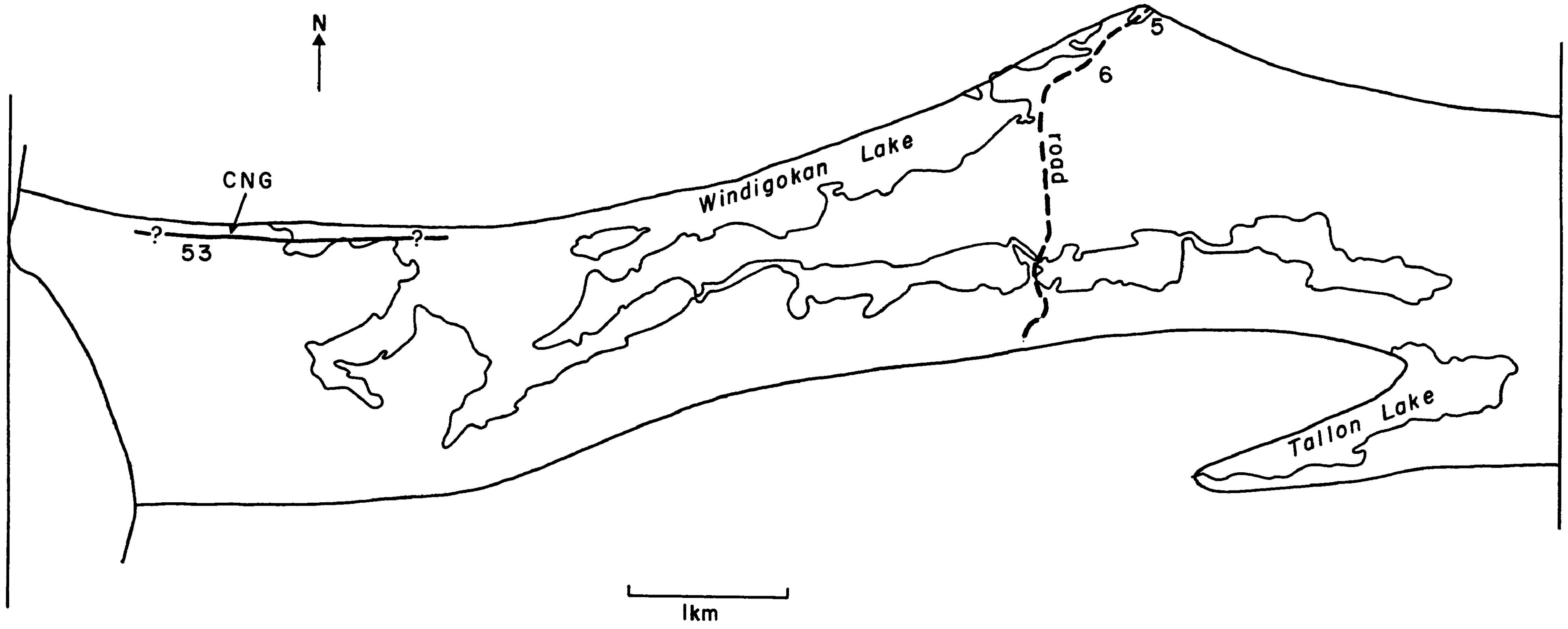

N
CNG
?
53
?
Windigokan Lake
road
5
6
Tallon Lake
1km

Fig. 3.1(b): Locations of outcrops studied and litho facies assemblages of the CMB in Irwin Township. Modified from Mackasey (1975).

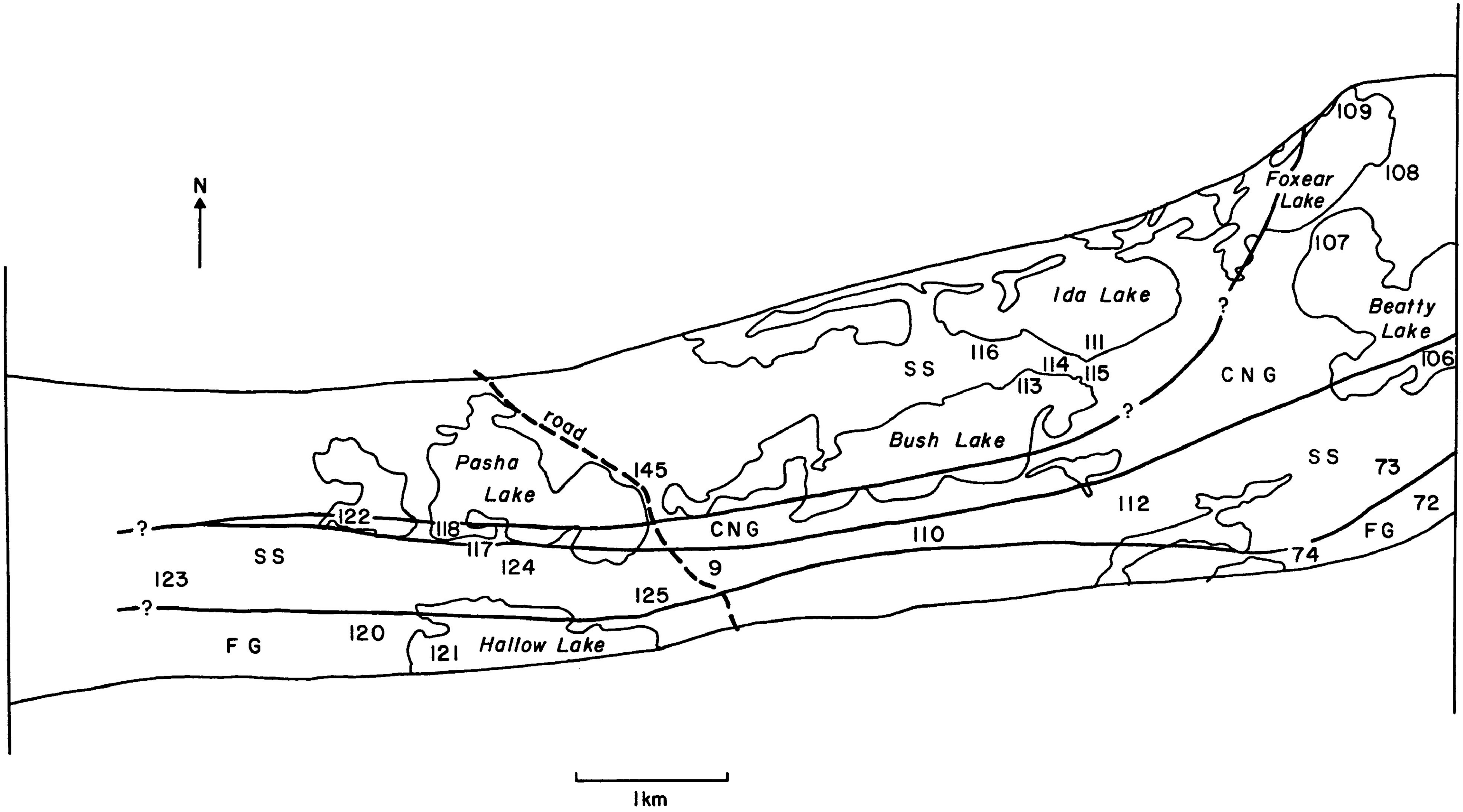

N
109
108
Foxear Lake
107
Beatty Lake
106
Ida Lake
111
116
114
115
113
SS
?
CNG
?
Bush Lake
SS
73
72
FG
112
74
road
Pasha Lake
145
122
118
CNG
110
-?
117
SS
124
9
123
125
-?
120
FG
121
Hallow Lake
1km

Fig. 3.1(c): Locations of outcrops studied and lithofacies assemblages of the CMB in Walters Township. Modified from Mackasey (1976).

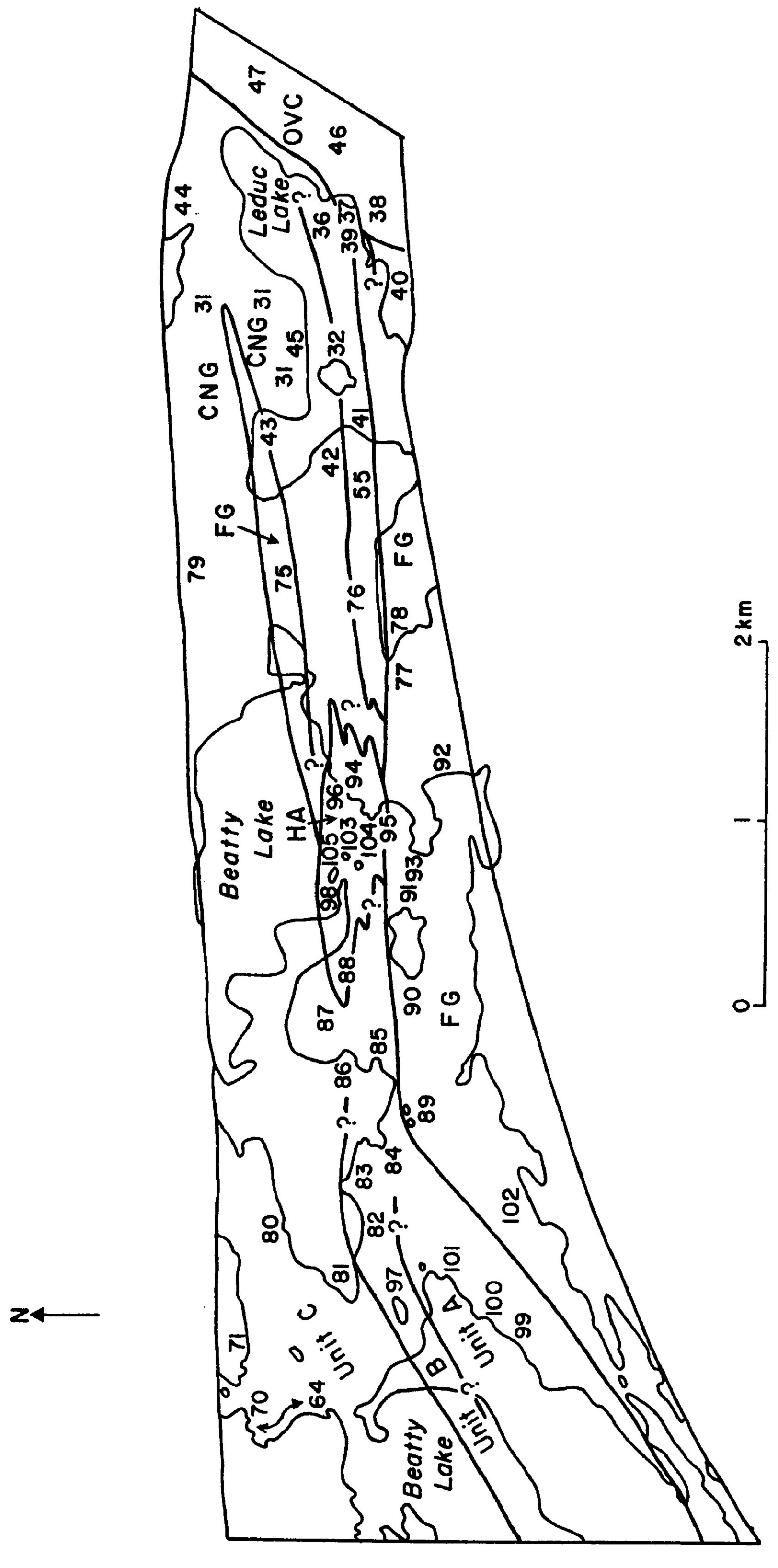

N
OVC
47
46
Leduc Lake
38
37
36
39
40
44
31
CNG
CNG
31
45
32
31
43
41
42
55
FG
79
75
76
FG
78
77
92
Beatty Lake
HA
96
94
105
103
104
95
98
93
91
88
90
87
FG
85
86
89
84
83
82
81
80
102
101
97
100
99
Unit A
Unit B
Unit C
71
70
64
Beatty Lake
0
1
2 km

Fig. 3.1(d): Locations of outcrops studied and litho facies assemblages of the CMB in Leduc Township. Modified from Mackasey (1976). The volcaniclastic part of the CMB east of Leduc Lake (outcrops #37, 38, 46 and 47 in this Figure) is the subject of Chapter 9.

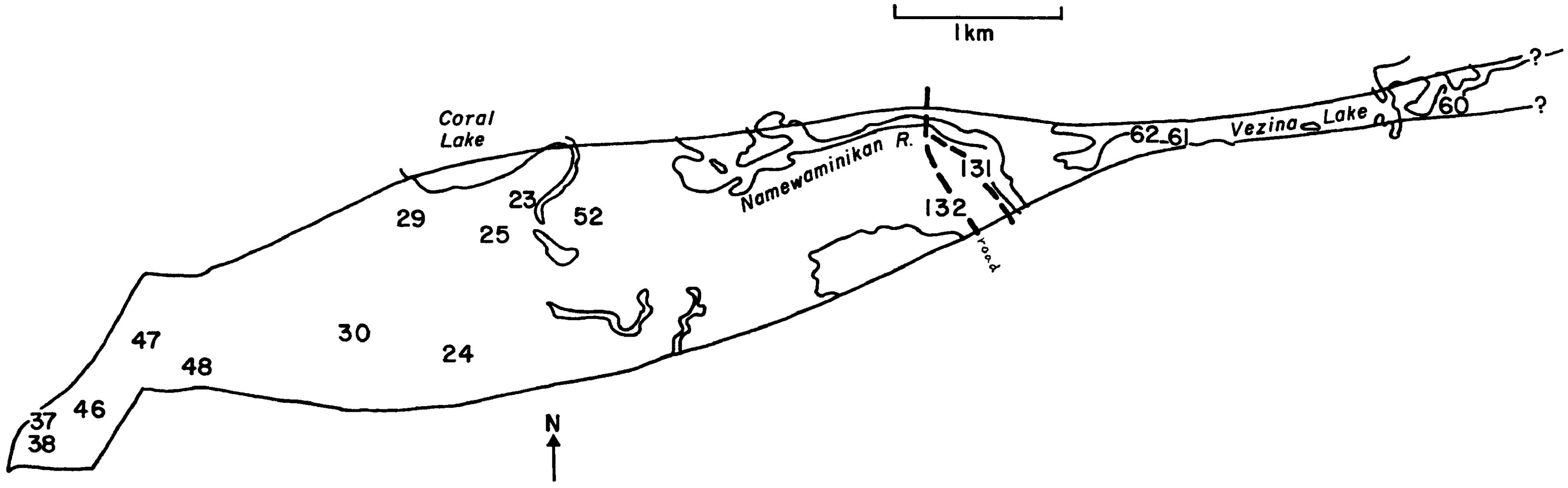

1km
Coral
Lake
Namewaminikan R.
Vezina Lake
road
?
?
N
29
23
25
52
131
132
62
61
60
47
48
30
24
46
37
38

Fig. 3.1(e): Locations of outcrops studied in Leduc and Legault Townships, the "Coral Lake volcaniclastic sediments" of the CMB. Modified from Mackasey (1976) and Mackasey *et al* (1976a).

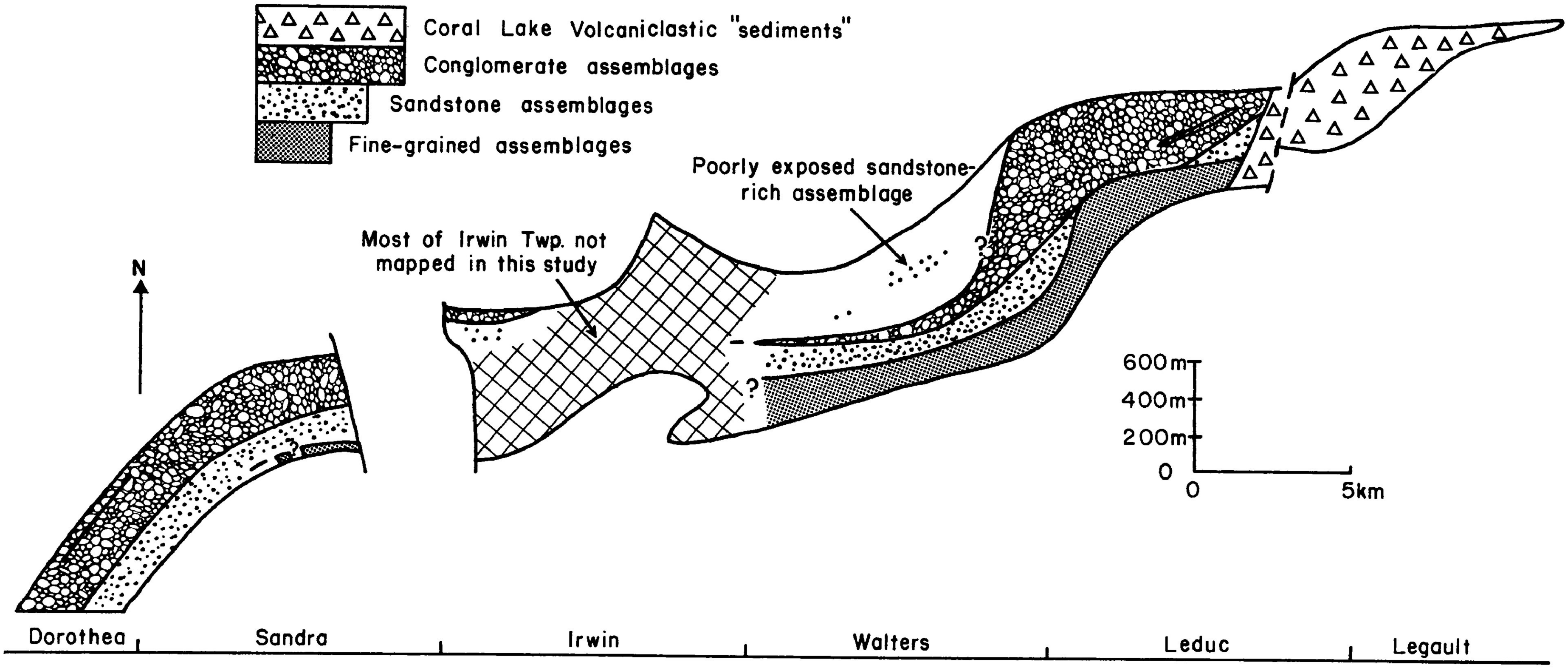

Coral Lake Volcaniclastic "sediments"
Conglomerate assemblages
Sandstone assemblages
Fine-grained assemblages
Poorly exposed sandstone-rich assemblage
Most of Irwin Twp. not mapped in this study
N
?
600m
400m
200m
0
0
5km
Dorothea
Sandra
Irwin
Walters
Leduc
Legault

Fig. 3.2: Lithostratigraphy of the CMB (Fig. 1.1). Modified from Mackasey (1975, 1976) and Mackasey *et al* (1976a). Composite stratigraphic sections are given in Fig. 12.1.

its lithofacies to the NMB.

This trend is outlined below in the form of segments of the CMB from west to east. No stratigraphic continuity has been assumed because of possible structural complications.

3.2 The CMB's Coarsening-North Trend

Dorothea and Sandra Townships. Despite a low density of exposure, the coarsening-north trend is well defined. The southern half is sandstone-rich, with mudstone preferentially located along the southern margin of the CMB. A critical exposure contains a smaller scale (100 m), subtly-defined coarsening-north trend within the sandstone horizon as well. The northern half of the sequence is conglomerate with an iron formation horizon in Dorothea Township. The CMB is covered by Proterozoic diabase in the eastern 2.5-3.5 km of Sandra Township, but obviously continues along strike into Irwin Township (Mackasey, 1975).

Irwin Township. The coarsening-north sequence is very poorly displayed, but is suggested by the presence of conglomerate immediately along the CMB's north margin.

Walters Township. A well-defined coarsening-north trend spans the southern half of the CMB. Mudstone, fine-grained sandstone and minor iron formation are present along the southern margin, separated from the usual conglomerate by a horizon of coarse to medium-grained sandstone. Glacial overburden covers nearly all of the CMB's northern half, except for a few lakeshore exposures of anomalous graded sandstones and associated lithofacies.

Western Leduc Township. Beatty Lake covers most of the CMB, but provides excellent shoreline exposures. The southern third is rich in mudstone and sandstone, the central third in conglomerate and sandstone, and the northern third in conglomerate. Sedimentological details such as maximum average deformed clast size (D10), mean grain size, and percentage of conglomerate in outcrop outline the CMB's best-developed coarsening-north trend.

East-central Leduc Township. In the Leduc Lake vicinity, mudstone is exposed near the CMB's southern margin, sandstone and conglomerate occupy the middle, and the northern CMB is conglomerate-rich, with a mudstone horizon that extends westward to Beatty Lake.

Leduc and Legault Township. East of Leduc Lake for about 7 km along strike, the sequence is "conglomeratic" and the provenance of the gravel-size clasts changes to nearly 100% felsic (to intermediate) volcanic. These strata are either pyroclastics (tuff-breccia, lapillistone, tuff) or reworked, coarse volcaniclastics. Because of their morphological similarity to the CMB's conglomerate-rich parts, Mackasey (1969a, 1976) at first failed to differentiate between these volcaniclastics and the CMB's more typical polymict conglomerates, but did later map the strata in Legault Township as volcaniclastic (Mackasey *et al*, 1976a).

No coarsening-north trend is present in this unit.

CHAPTER 4: THE CMB's FINE-GRAINED ASSEMBLAGE

4.1 Introduction

The geographic extent of the CMB's fine-grained lithofacies assemblage is shown in Figs. 3.1 and 3.2. Fig. 3.1 gives the locations of specific outcrops referred to in the text. For local details such as bedding and cleavage measurements the reader should examine Mackasey's (1975, 1976) maps.

The mudstone unit in east-central Sandra Township and most of the fine-grained assemblage in Irwin Township (Figs. 3.1, 3.2) were not examined owing to access, logistical and financial problems. Both Laird (1937) and Mackasey (1975) have recorded the presence of Sandra Township's mudstone unit. Mackasey's (1975) map of Irwin Township shows for this part of the fine-grained assemblage a suite of "sandstone, wacke, siltstone, argillite, slate and iron formation", and three outcrops with north-topping beds based on grading and cross-stratification.

4.2 Lithofacies Descriptions

The fine-grained assemblage is composed of mudstone, dominantly fine to very fine-grained non-pebbly sandstone, an anomalous conglomerate facies, and iron formation. Some interesting and unusual features of this assemblage are described below.

Outcrop #53. A traverse perpendicular to strike across outcrop #53 encountered mostly mudstone and medium to fine-grained sandstone, and revealed an interesting interval 50 to 70 m south of the local conglomeratic assemblage (Fig. 3.1). This interval has laminated siltstone, with two thick sandstone interbeds, in the south part, and graded sandstone and mudstone to the north.

The siltstone is thinly bedded to laminated, with laminae 1-10 mm thick. Red, non-magnetic layers are probably hematitic, and therefore possibly gradational to an iron formation. Note that Mackasey (1975) recorded iron formation 1.2 km to the east along strike, on the shore of Windigokan Lake. One tiny ripple is present in the laminated siltstone, its paleocurrent to the east. Thickening of the coarser laminae may represent starved ripples.

Two anomalously thick and coarse sandstone beds are interstratified with the siltstone. One is 25 cm thick and ungraded, and the other 15 cm thick and fining north from very coarse to medium-grained sand. The two beds are very coarse to medium-grained and poorly sorted. At the base of one of these beds, based on tops north from graded beds metres away (described below), there is a fragment of red mudstone, most likely a locally-derived rip-up clast.

The northern part of this site contains interbedded medium to fine-grained sandstone and silty mudstone. Laminae within these 2-20 mm thick beds are 0.1-2.0 mm thick. Some thin beds are graded, fining north from fine to very fine-grained sand to silty mud. Cleavage improves upward in these graded beds because of the increasing mud content. The beds have sharp soles (south contacts).

The next 15 m to the north contain an interbedded mixture of mudstone, very fine-grained and medium to fine-grained sandstone. One medium-fine sandstone bed has a sharp south contact with mudstone having faint sand laminae, the contact likely being a sharp (erosive?) base (tops north).

Outcrop #72. The sandstone in the southern part of outcrop #72 is fine-grained and moderately to well sorted. A minor amount of very fine-grained beds are present. Most of the strata are laminated.

Laminae are planar, horizontal to slightly inclined, parallel and defined by coarser and finer sandy layers. Yellow and recessive weathering mudstone laminae (and very thin beds) are interlaminated with the fine-grained sandstone.

Some horizons display alternating fine-grained sandstone laminae about 10 mm thick and mudstone laminae up to 5 mm thick. There is also a suggestion of faintly developed fining and thinning-northward (upward) sequences tens of cms thick, in which sandy laminae thin and muddy laminae increase in amount to the north (upward). Stratigraphic tops are via concave troughs, only metres away, as described below.

One sharp based, 18 cm thick bed fines northward (upward) from fine to very fine sand in its upper 4 cm.

A few metres of section contain broad, shallow trough cross-beds. The troughs are 10 cm thick, over 1.0 m wide and are concave north. They are erosively based, truncating at low angles the adjacent laminae. Laminae within the troughs are inclined up to 13° from the paleohorizontal. The troughs are a variation from the locally dominant horizontally to low-angle laminated sandstone, and resemble swaley cross-strata (Walker, 1984b). One small (1 m sq.) spot contains "arches", concave south low-angle laminae similar to hummocky cross-strata (Walker, 1984b).

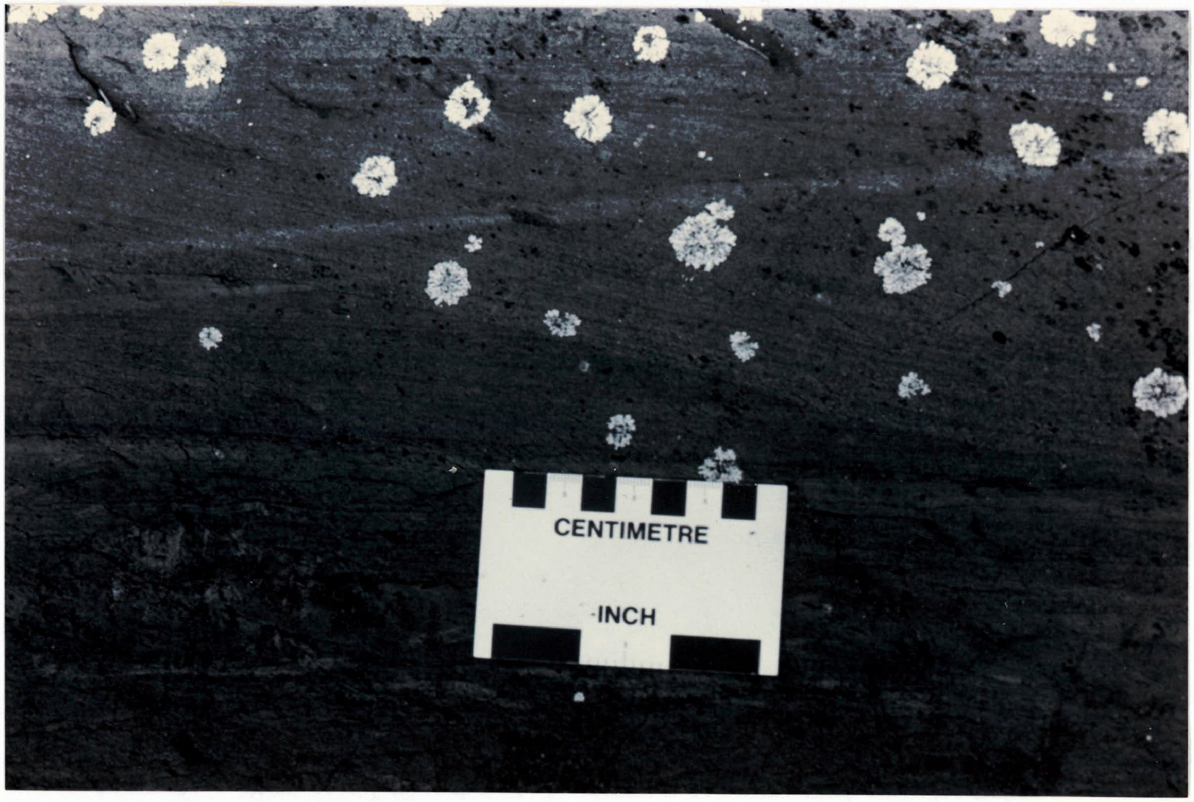
The northern part of outcrop #72 contains silty mudstone, very fine-grained sandstone, and interlaminated siltstone and iron formation with thicker (10 mm) silty and thinner (1-2 mm) magnetite laminae. Mackasey (1976) did not record any iron formation in outcrop #72, but did in outcrop #74 1.0 km to the west along strike.

<u>Leduc Township outcrops.</u> Mudstone (argillite/slate) and coarse siltstone are the assemblage's dominant lithologies in this area (Fig. 3.1). The siltstone has planar, horizontal and parallel laminae up to 4 mm thick. The coarsest laminae contain fine to very fine sand. Outcrop #39 shows lamination defined by recessive and resistant weathered layers and also colour variation, with some folded laminae at angles to the cleavage. The minor amount of fine to very fine-grained, pebble-free sandstone present is in beds, mostly graded ones, up to 5 cm thick (Fig. 4.1). These beds grade from fine (only rarely medium) to very fine sand or mud, and indicate stratigraphic tops to both the north and south, a result of

Fig. 4.1: Thinly bedded sandstone and siltstone. Some beds and laminae are graded. the lighter toned parts being coarser. Isoclinal folding and structural truncation of beds are evident. Outcrop #91.

Fig. 4.2: Middle: low-angle scour-fill laminae of coarser sandstone truncate underlying horizontal finer, silty sandstone. Scour is broadly concave up and resembles swaley cross-stratification in form. Top: horizontal laminae cross-cut underlying low-angle laminae. Outcrop #89.


CENTIMETRE
INCH


CENTIMETRE
INCH

isoclinal folding. Fig. 4.2 shows a relatively large sandstone-filled scour structure, similar in form to swaley cross-stratification.

The bedding present, particularly in the mudstone, is often laterally discontinuous at a scale of tens of cms, revealing many of the tapering laminae and thin beds to be tectonized shreds of strata.

Outcrop #102 contains a lithofacies not seen elsewhere in this assemblage. Oligomict fine-grained conglomerate is massive and ungraded, with a clast-supported, moderately sorted framework. The mean deformed clast size is about 5-10 mm, with a D10 of 30/61, by far the finest in the CMB (Table 6.2). All of the clasts are felsic volcanic pebbles. The interbedded sandstone varies from pebbly, coarse to medium-grained and very poorly sorted (ranges from granules to fine sand) to fine to very fine-grained and moderately to well sorted. Some of the very pebbly sandstone is texturally gradational to matrix-supported conglomerate.

CHAPTER 5: THE CMB's SANDSTONE ASSEMBLAGE

5.1 Introduction

The geographic extent of the CMB's sandstone lithofacies assemblage is shown in Fig. 3.1 and 3.2. Fig. 3.1 gives the locations of specific outcrops referred to in the text. For local details such as bedding and schistosity measurements the reader should examine Mackasey's (1975, 1976) maps.

As the assemblage varies considerably along strike, four separate descriptions (Sections 5.2, 5.3, 5.4, 5.5) will be given.

5.2 Sandra Township Area

The sandstone-dominant part of the CMB exposed in southwest Sandra Township (Fig. 3.1) is rich in rippled, fine-grained sandstone. Most distinctive are rhythmic couplets (Fig. 5.1), which dominate the outcrops. Fig. 5.2 shows the best exposure of these units, a continuously exposed stream section along a waterfall (Photo 1 in Mackasey, 1975).

This stream section contains about 30 m of superbly exposed strata. Ripples abound, there is no significant tectonic deformation, and no macroscopically visible foliation; Mackasey (1975, p.4) also noted these "non-foliated metasediments". Because of the 100% exposure of the section, its internally undisturbed aspect, abundant well-preserved sedimentary structures and bedding and many top indicators all facing north, including thousands of ripples,

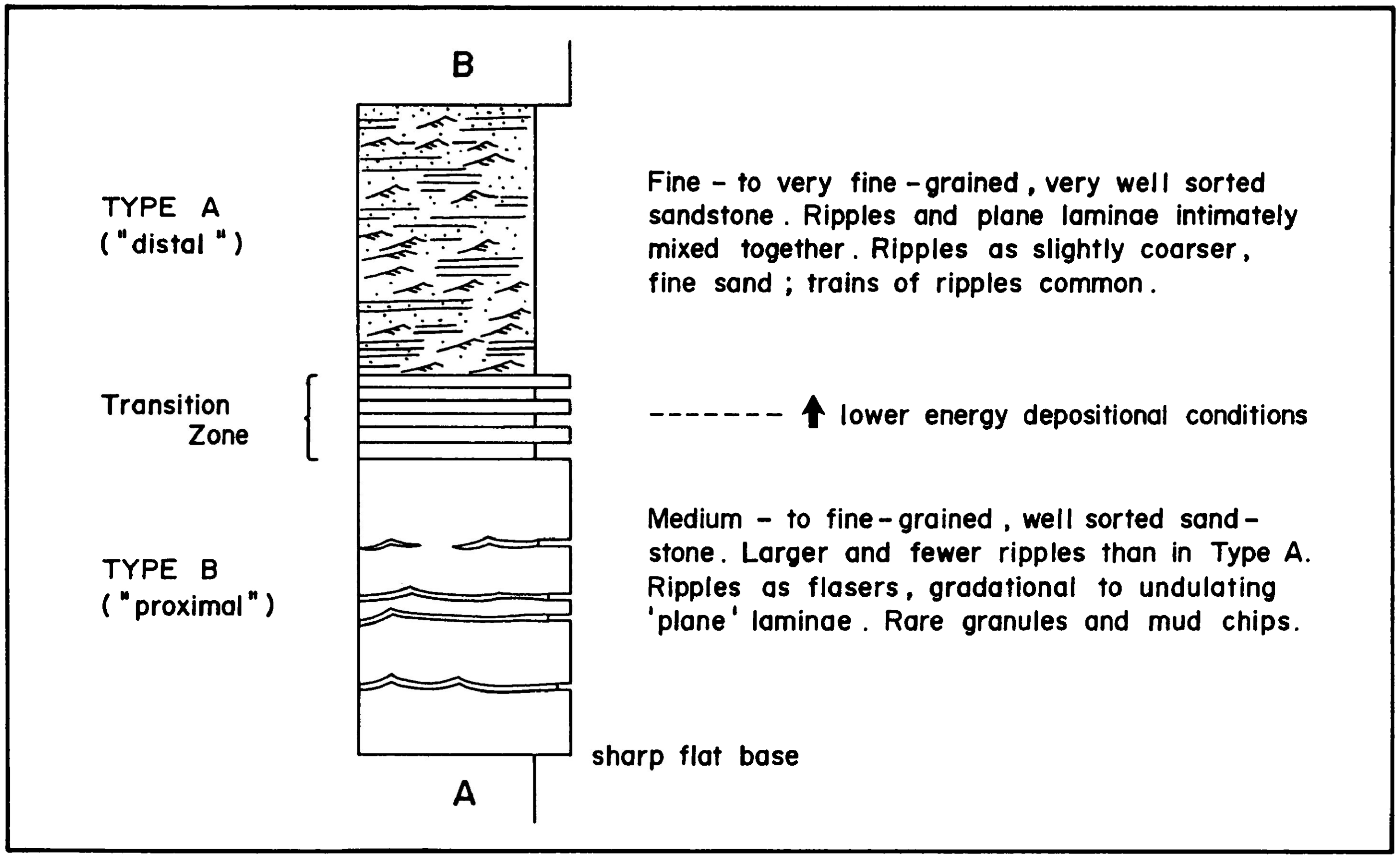

B
TYPE A
("distal")
Fine - to very fine -grained, very well sorted sandstone. Ripples and plane laminae intimately mixed together. Ripples as slightly coarser, fine sand; trains of ripples common.
Transition Zone
------- ↑ lower energy depositional conditions
TYPE B
("proximal")
Medium - to fine-grained, well sorted sandstone. Larger and fewer ripples than in Type A. Ripples as flasers, gradational to undulating 'plane' laminae. Rare granules and mud chips.
sharp flat base
A

Fig. 5.1: Idealized couplet (or rhythmite). Vertical scale varies from about 0.5 m to several metres.

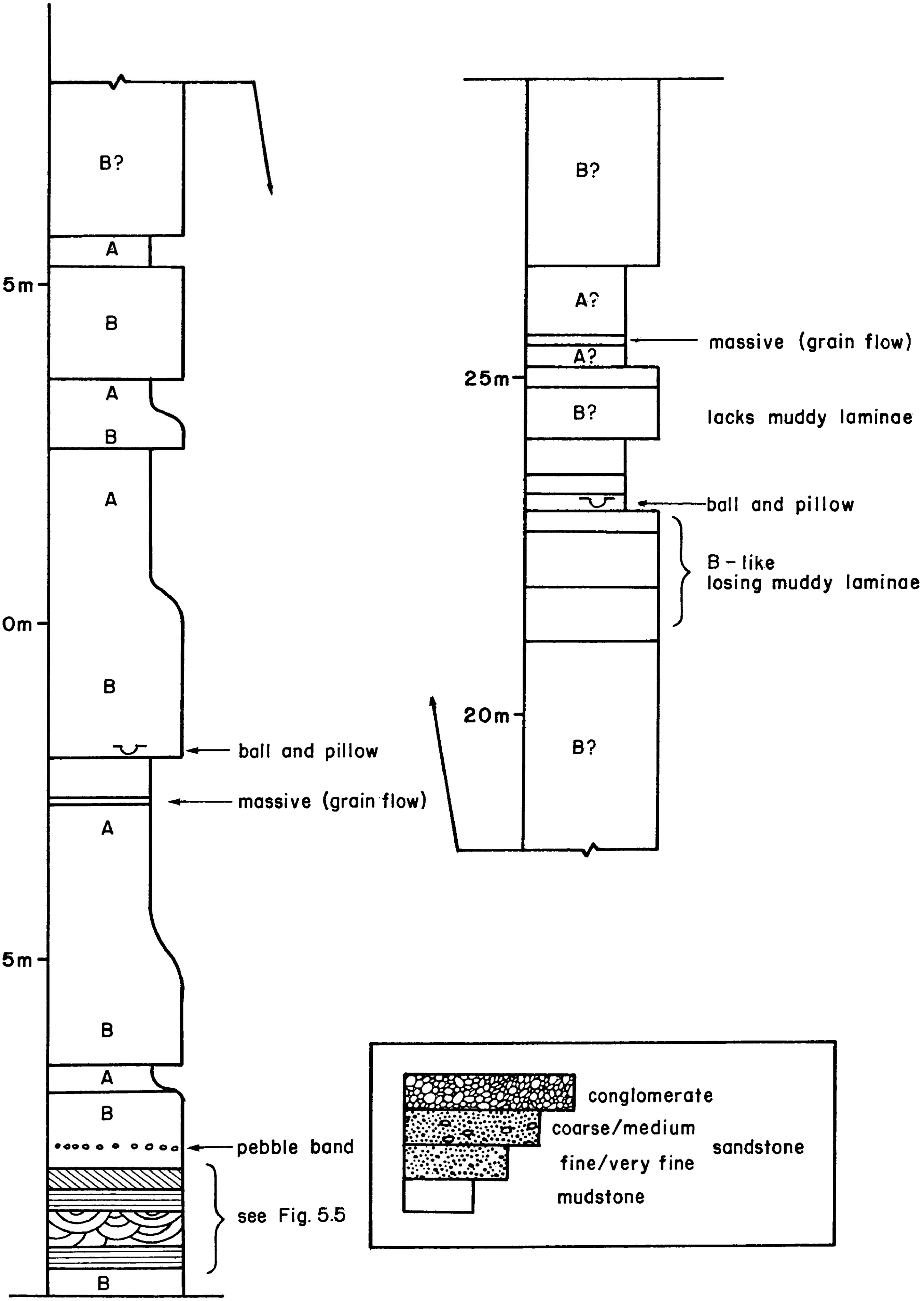

B?
A
15m
B
A
B
A
10m
B
ball and pillow
massive (grain flow)
A
5m
B
A
B
pebble band
see Fig. 5.5
B
B?
A?
massive (grain flow)
A?
25m
B?
lacks muddy laminae
ball and pillow
B - like
losing muddy laminae
20m
B?
conglomerate
coarse/medium
fine/very fine
sandstone
mudstone

Fig. 5.2: Continuously exposed stratigraphic section, waterfall and east bank of Namewaminikan River, south part of outcrop #21A. Type A and B units are described in Fig. 5.1 and in the text. Top to north.

the section is considered a continuous one, with conformable strata reflecting their hydrodynamic evolution. Further north (up), the section continues for about another 70 m, continuously exposed in places, and containing a cross-cutting mafic dike.

Outcrops #21 and 19 (Figs. 5.3, 5.4) also contain couplets. Sandstone-rich outcrops in east central Sandra Township (east of the road in Fig. 3.1) were not examined in detail because of their poor exposure and limited time.

Rhythmically-bedded couplets and associated lithofacies. Type B - Type A couplets or rhythmites (Fig. 5.1) range from 42 cm to several metres thick (Figs. 5.2, 5.3). Figs. 5.1 to 5.4 show the rhythmic nature of the couplet-dominated sequences. Cross-bedded and associated plane laminated sandstone forms a minor part of the sections (Fig. 5.2, 5.3, 5.5).

Type B facies/units. These are described in Fig. 5.1. Listed below are additional details.

Where ripples are not outlined by mud drapes (flasers, mud laminae) the sandstone is massive-looking to faintly rippled. Some rippled bedding planes are visible (Fig. 5.6). The exposures tend to weather along muddy, rippled bedding planes, producing a staircase-like section in outcrop #21A's north-dipping strata.

Fig. 5.7 shows the typical stratification within a Type B unit. The wavy flasers (Reineck and Singh, 1980, Fig. 183) are 0.1-2.0 mm thick, 5-10 cm long and up to 5 mm deep. They

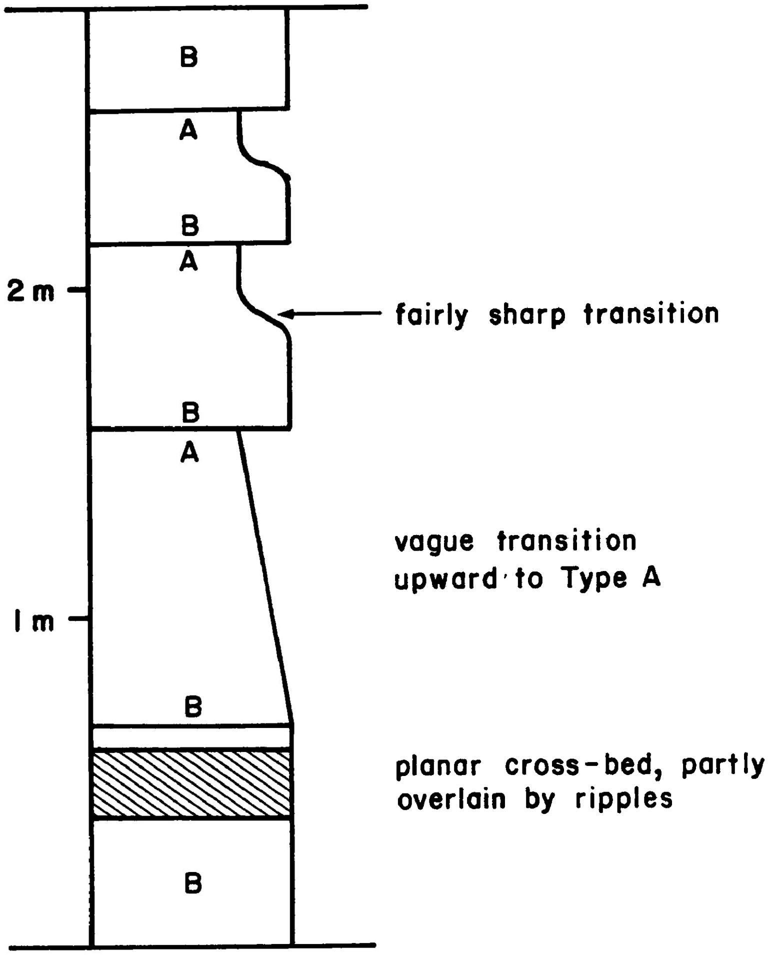

B
A
B
A
2 m
fairly sharp transition
B
A
vague transition upward to Type A
1 m
B
planar cross-bed, partly overlain by ripples
B

Fig. 5.3: Section of outcrop #19 and Fig. 5.4 Type A and B units are described in Fig. 5.1 and in the text. Top to north. Legend in Fig. 5.2.

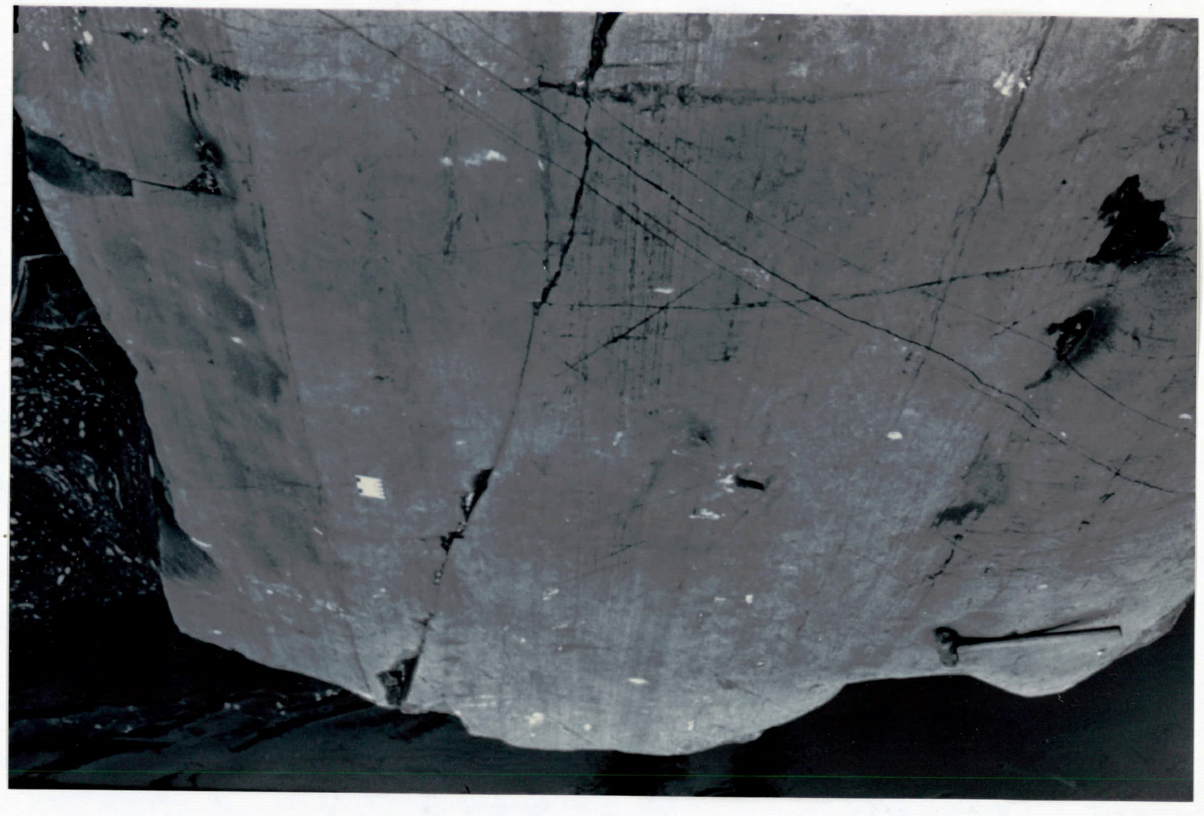


Fig. 5.4: Rhythmically bedded Type B – A couplets; top to left. Right margin of outcrop is near 1.5 m level of Fig. 5.3. Note sharp base of Type B unit, to right of white scale card. Middle and far right: darker, laminated Type A units. Outcrop #19.

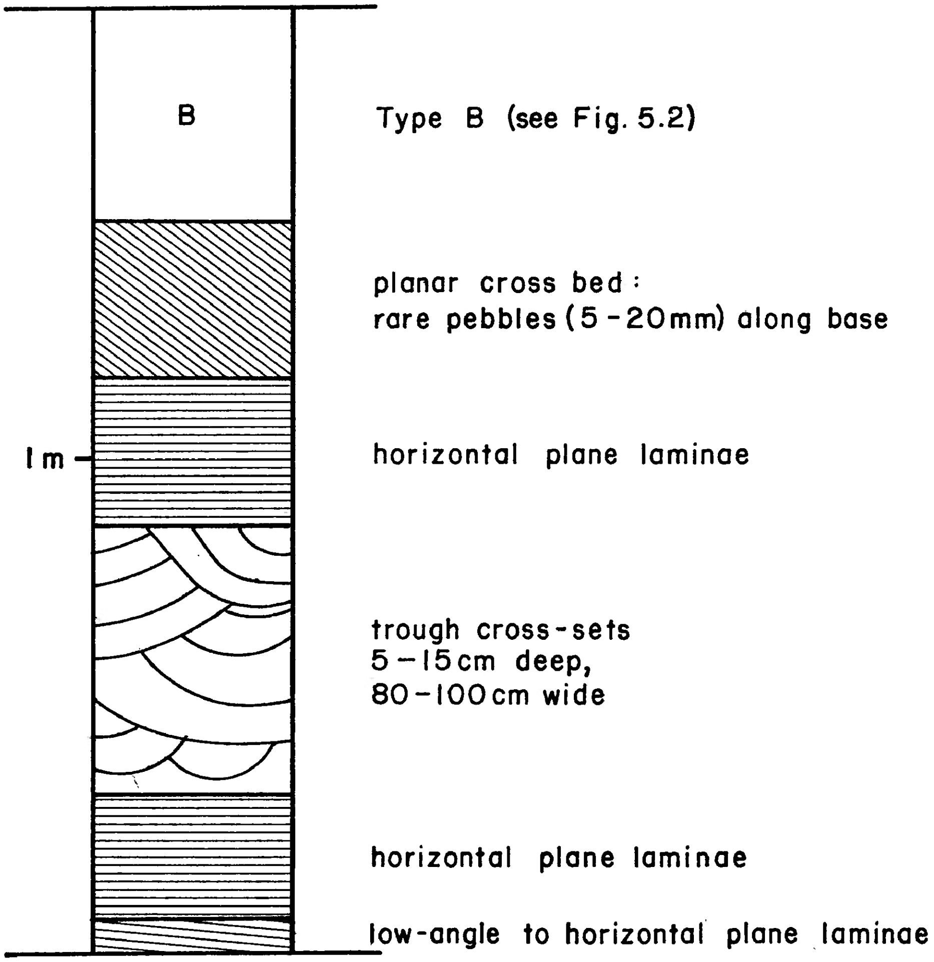
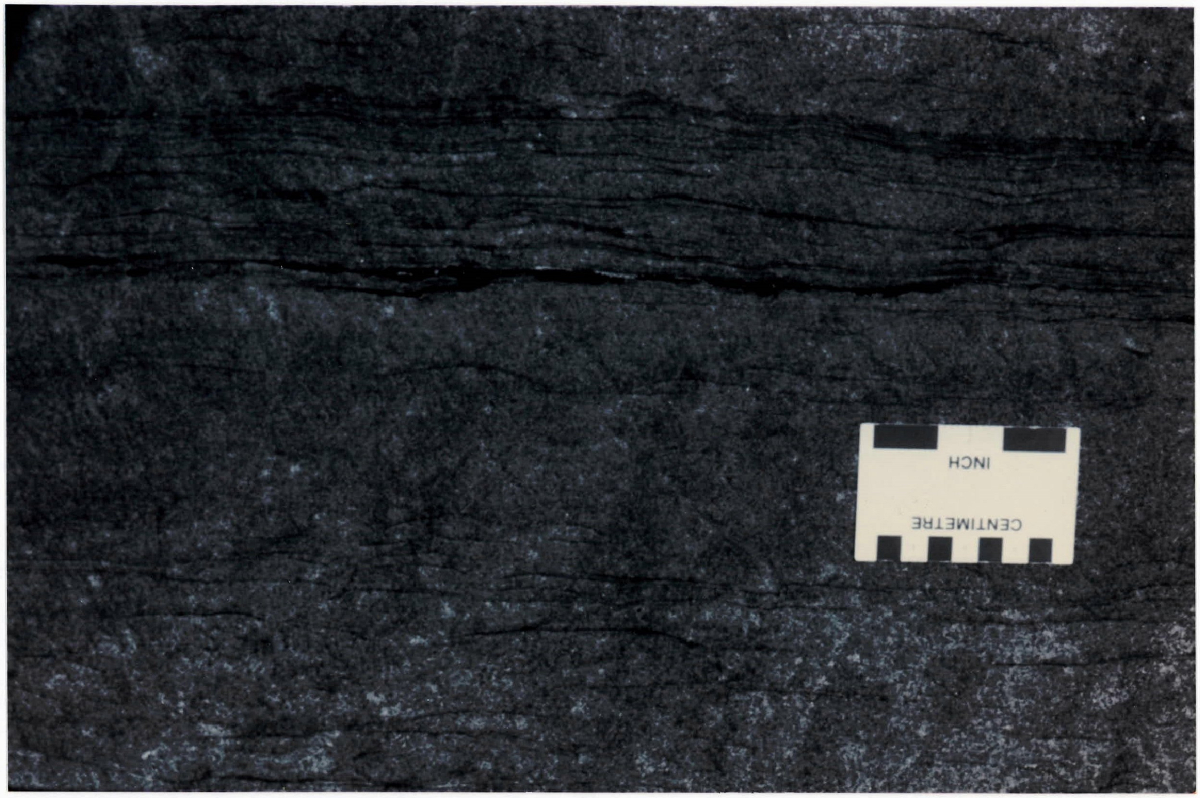
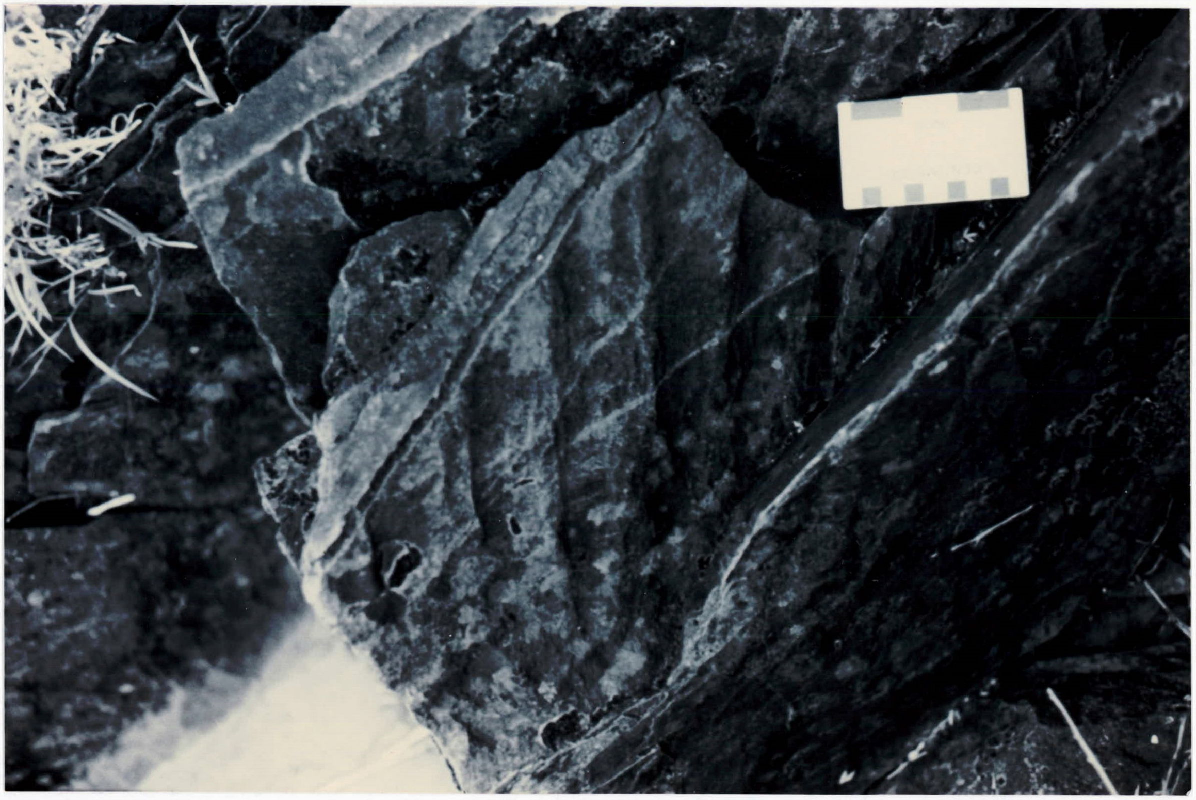

B
Type B (see Fig. 5.2)
planar cross bed :
rare pebbles (5 - 20mm) along base
1m
horizontal plane laminae
trough cross-sets
5 - 15cm deep,
80 - 100cm wide
horizontal plane laminae
low-angle to horizontal plane laminae

Fig. 5.5: Detailed view of lowermost part of section in Fig. 5.2. Top to north. Legend in Fig. 5.2.

Fig. 5.6: Rippled bedding plane within a Type B unit. Outcrop #21A.

Fig. 5.7: Type B stratification. Darker, fine (muddy) laminae are present as flasers (ripple drapes) and wavy flasers gradational to plane laminae. Bottom: note the dark, finely laminated band (wavy bedding). Outcrop #21A.


CENTIMETRE
INCH



are dark because of their mud content. Lateral continuity of muddy laminae of 0.5 m is common; poor lateral continuity is of 10 cm (dm) scale. The laminae are laterally undulating on a 10 cm scale where ripples are not well defined, a type of wavy flaser bedding (Reineck and Singh, 1980, Fig. 183). The laminae occur singly or in clusters of up to 10 laminae in 2 cm of section, the clusters being a type of wavy bedding (Reineck and Singh, 1980, Fig. 183).

One Type B unit has a sharp, erosive base (Fig. 5.8). High-energy inflow brought in very coarse sand and granules, scoured down into underlying muddy laminae and locally deposited muddy rip-up clasts.

Rare cm-scale sand dikes, 1-2 mm mud chips, ball-and-pillow structures up to 20 cm thick, and laterally discontinuous pebble bands are also present.

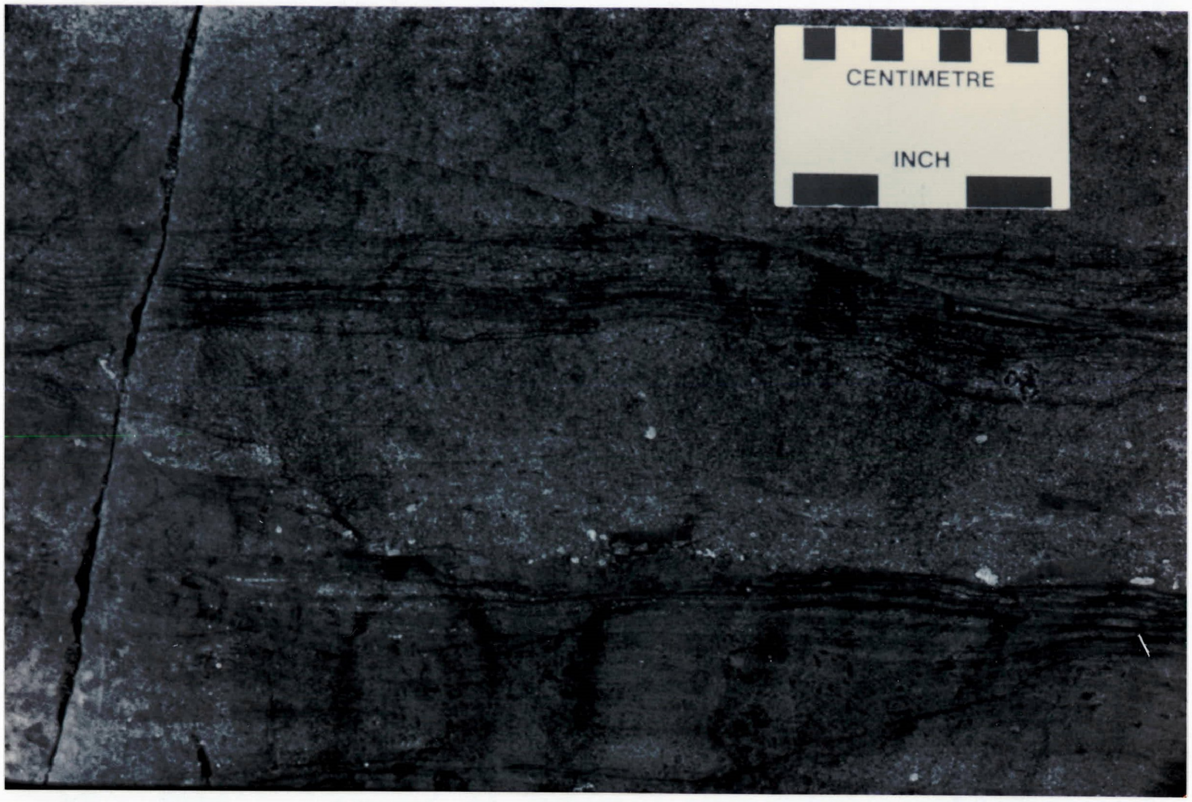
Outcrop #21 contains an abnormally coarse Type B unit of coarse to fine-grained sandstone with granules. The granules are rock fragments and mud chips, subround and elongate.

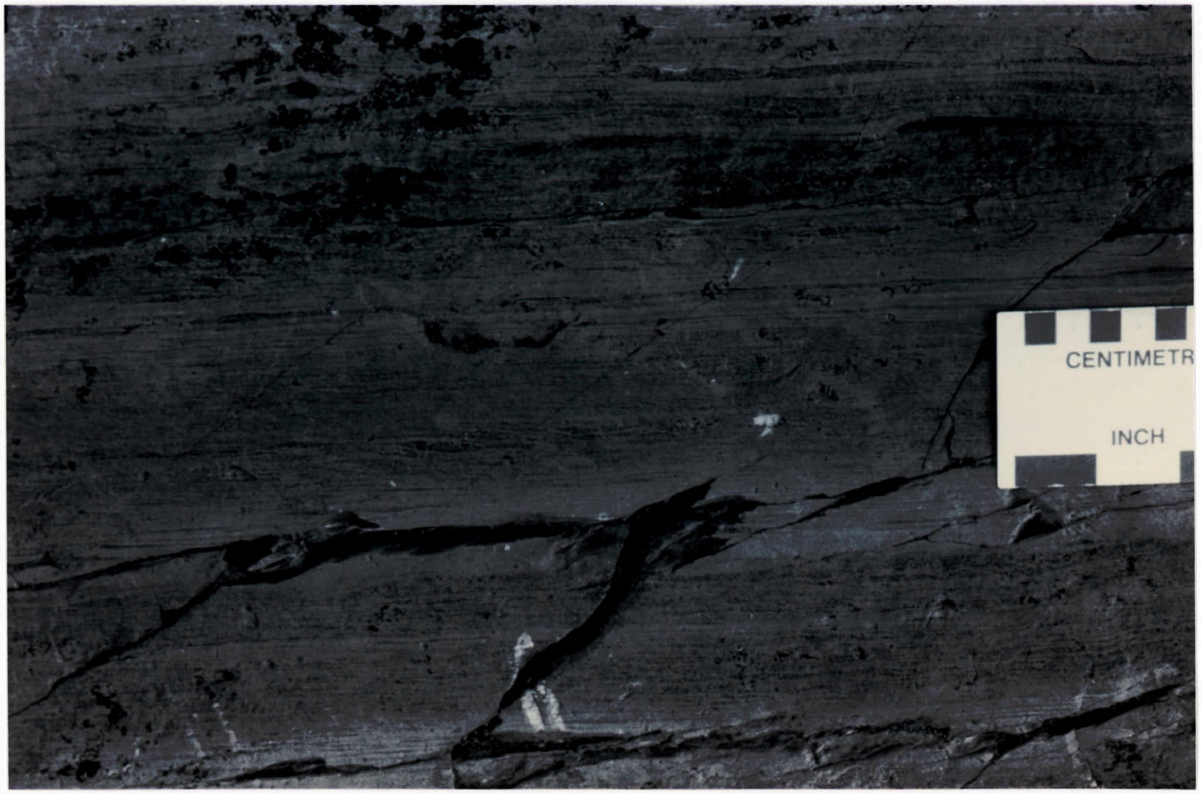
Granule laminae are 5-10 mm (2-5 granules) thick. They form a very minor part of this Type B unit. One granule lens is 3 cm thick and 6 cm long. Laterally discontinuous laminae have a 4-8 cm continuity and are regularly spaced along strike.

Some of the granules appear to be imbricated, with both contact and isolate imbrication. There is no visible foliation present; Mackasey (1975, p.4) described these Type B-A units as "resistant non-foliated metasediments". The

Fig. 5.8: Bottom: Type A laminated sandstone. Middle: sharp, erosive base of a Type B: note the very coarse sand, granules and dark, muddy rip-up clasts. A good example of wavy bedding is present below the scale card. Outcrop #21A.

Fig. 5.9: Type A stratification. Cross-laminated, fine to very fine-grained sandstone. Ripples are the slightly coarser, darker lenticles. Middle (above scale card): ripple train of eight or nine individual ripples. Outcrop #21A.


CENTIMETRE
INCH


CENTIMETR
INCH

apparent imbrication gives a paleocurrent approximately to the east, agreeing with ripples 5 cm above, and most other cross-stratification in the area (outcrops #19, 21A).

Type A facies/units. These are described in Fig. 5.1. Listed below are additional details.

Laminae are planar, horizontal and parallel, less than 0.1 mm to 1.0 mm thick (rarely cms thick), and undulatory over about 8 cm laterally. Laminae 1-2 mm thick swell to ripples up to 5 mm thick. Microlaminae are also rippled, with ripples less than 1 mm thick.

Laminae and ripples constantly grade into one another on cm-scales (Fig. 5.9). The ripples stand out as slightly coarser lenticles up to 1 cm thick. Groups of ripples have good lateral continuity, with individual ripples spaced 3-6 cm apart in trains. One Type A unit shows, in these ripple trains, ripples:

1) 4 cm thick are spaced 20 cm apart;

2) 10-15 mm thick are 10-15 cm apart; and

3) 5 mm thick are 6 cm apart.

Obviously the smaller (lower energy) ripples are more closely spaced in sequence. These ripples stand out well in outcrops because of preferential lichen growth on them.

Paleocurrents measured from ripples are all approximately to the east to southeast, a broadly unimodal pattern.

Two Type A units contain unusual interbeds within them. Two beds 8 cm and 16 cm thick (in Fig. 5.2) are sharp-based, massive, very fine to fine-grained sandstone.

The lateral continuity of Type A units across outcrop #21A is at least 10 m in places. The tops of these units can be excellent local marker horizons.

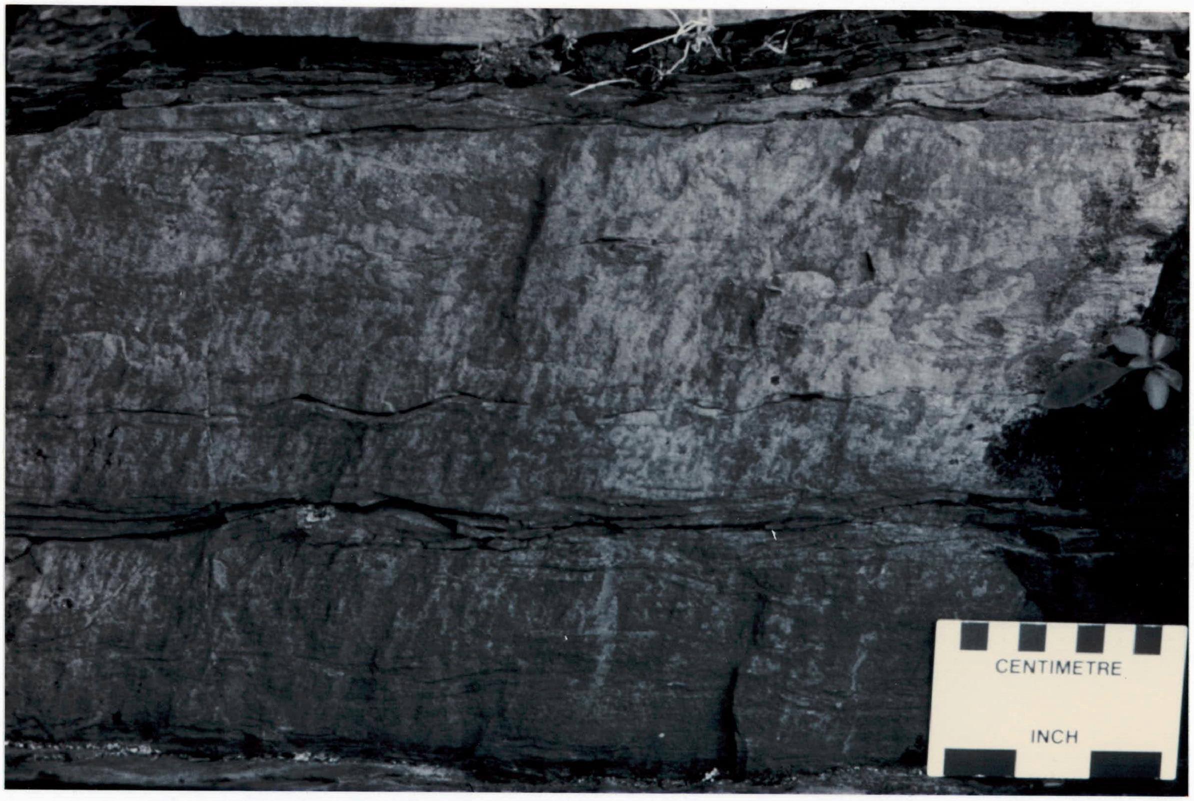
Transition zones between the Type B and A units are thin relative to the B-A couplets (Fig. 5.1). The transition is marked by the appearance of Type A style layers cms thick over 10-100 cm mid-couplet intervals.

Other units. In the upper 9 m of the section in Fig. 5.2, well-defined Type B and A units are not common. Units of similar thickness, grain size and sedimentary structures are present, which may represent very poorly developed Type B and A units, the results of similar but slightly different processes and preservation. Here B-style intervals are medium to fine-grained, with some coarse sand and very little to no mud content. Faint lamination is occasionally present. A-style intervals are fine to very fine-grained, with often vague plane laminae and ripples. One fine to very fine-grained bed contains symmetrical (oscillation?) ripples 1 cm deep and 3-7 cm wide (Fig. 5.10).

Some Type B and A units are present above these "poorly or undeveloped couplets", in the upper, northern part of outcrop #21A above the section in Fig. 5.2 (east of an old dam, and below strata described below).

Upper outcrop #21A sequence. North and east of the section in Fig. 5.2 there is sandstone similar to Type B units, coarse to fine-grained and moderately sorted. Farther north (at an upper waterfall), medium-grained pebbly

Fig. 5.10: Rippled fine to very fine-grained sandstone. Middle: two-dimensional view of symmetrical ripples. Outcrop #21A.


CENTIMETRE
INCH

sandstone (low percentage of pebbles; clasts smaller than 15 mm) and coarse to fine-grained sandstone are present. A well-exposed, straight-crested asymmetrically rippled bedding plane gives a paleocurrent to the southeast, as is usual.

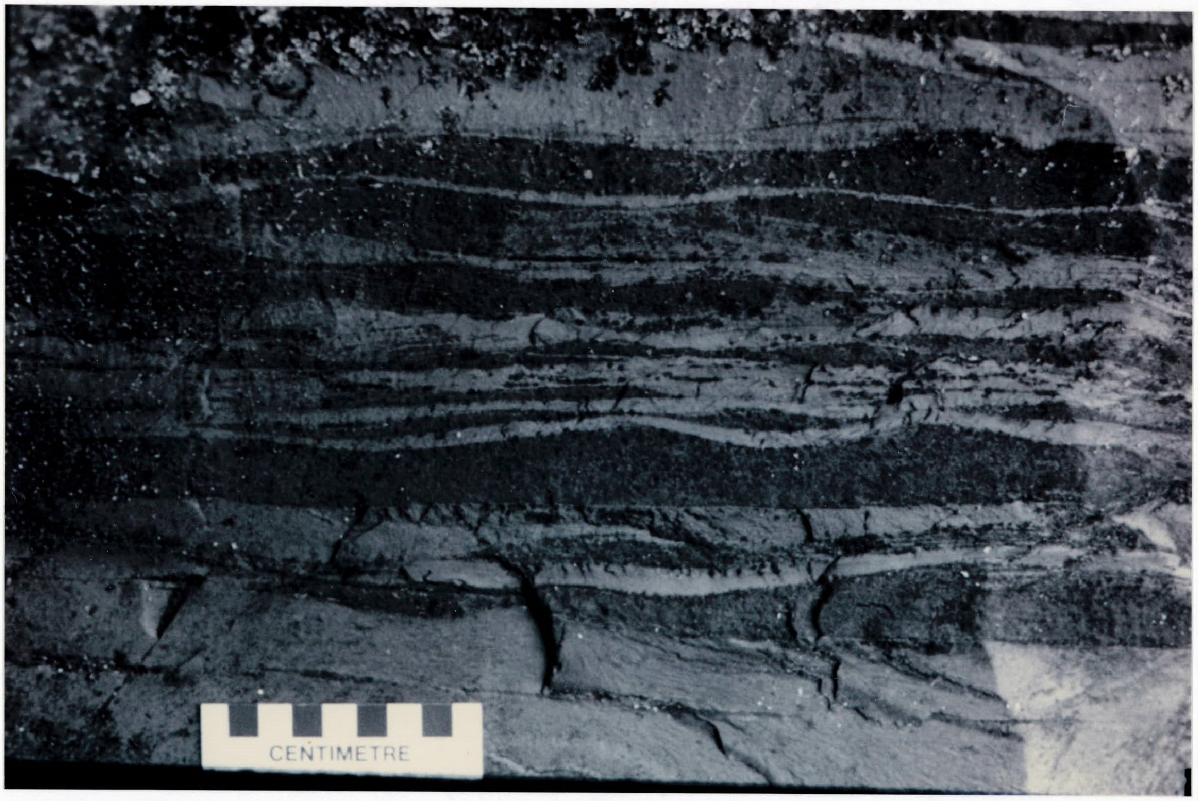
The northern tip of outcrop #21A is medium to fine-grained sandstone, with some coarse to fine-grained pebbly (rare 0.5 mm pebbles) parts. Faint plane laminae and ripples are present. No mud laminae are present.

In the three sites listed above, the presence of more coarse sand than in the couplets to the south (in Fig. 5.2) and the absence of mud and very fine-grained layers defines a subtle coarsening northward trend. As all the top indicators --cross-beds, ripples, sharp and erosive bed soles, and loading features--are to the north, this large outcrop contains a nearly continuously exposed, subtly coarsening-upward sequence about 100 m thick.

<u>Outcrop #18</u>. Outcrop #18 was not recorded by Mackasey (1975). Very thinly bedded, light to dark, laminated to rippled sandstone (Fig. 5.11) is directly downstrike from other rippled units, and has the same moderate dip (40-50°N) as other local outcrops, so it seems an obvious outcrop and not float.

The lighter, thicker (usually 2-6 cm) horizons up to 30 cm thick are very fine-grained silty sandstone, with some faint rippling. The darker and coarser layers are thin beds and laminae of rippled and plane laminated fine-grained sandstone. Ripples are mostly 0.5-1 cm thick, and up to 3 cm

Fig. 5.11: Rippled fine (dark) to very fine-grained (light) sandstone. Foresets, best seen in the upper ripples, indicate paleoflow to the right. Outcrop #18.


CENTIMETRE

thick. The thick rippled layers have a flat base and undulatory top, being 3 cm thick at crests and 2 cm at troughs, at wavelengths of up to 20 cm. One ripple train of laterally linked ripples is 5-10 mm hick with 7-9 cm wavelengths. One starved ripple train is thinner, individual ripples being 1-2 mm thick with 5-8 cm wavelengths and spaced 7-15 cm apart, the gaps along strike between ripples being about one ripple length. The thicker the ripples, the more laterally continuous they are likely to be.

Paleocurrents are all to the east; a few trough cross-laminae may be to the southeast.

Plane laminae are planar, horizontal and parallel. Some laminae pinch and swell laterally on a dm* scale. Parts of this very small section have a wavy bedding aspect, thin alternations of muddy and clean sand.

5.3 Walters Township Area

A lithostratigraphic horizon of massive, coarse to medium-grained and medium to fine-grained pebbly sandstone, with no conglomerate at all, extends across Walters Township between the CMB's fine-grained and conglomeratic horizons (Fig. 3.1). Some vaguely defined, laterally discontinuous pebble bands are present. Outcrop #73 contains mud chip clasts (2-3 mm thick, 4 cm long).

Strike-parallel zones of intense schistosity and 10 m - scale recessive weathering are probably faults.

* 10 cm

Outcrop #9 is anomalous, containing well-defined trough cross-beds. The troughs are concave north, 10-15 cm deep and 1-2 m wide. Adjacent sandstone beds are massive and plane laminated. The foreset laminae are 2-5 mm thick with good (greater than 1 m) lateral continuity. These laminae are defined by coarser and finer sand grains, and recessive weathering. They taper laterally towards the trough margins. The schistosity cross-cuts the trough laminae.

5.4 Beatty Lake Area, Leduc Township ("Unit A")

Fig. 3.1 shows the boundaries of this lithostratigraphic unit, termed "Unit A" by Devaney and Fralick (1985). Medium to fine-grained, well sorted sandstone is most common. It is often finer (fine to very fine-grained or silty) where rippled and plane laminated, and coarser (up to very coarse-grained and moderately to poorly sorted) in the larger cross-beds, foreset laminae, and scour bases (toesets and bed soles) (Fig. 5.12). The above applies to both outcrops and individual beds.

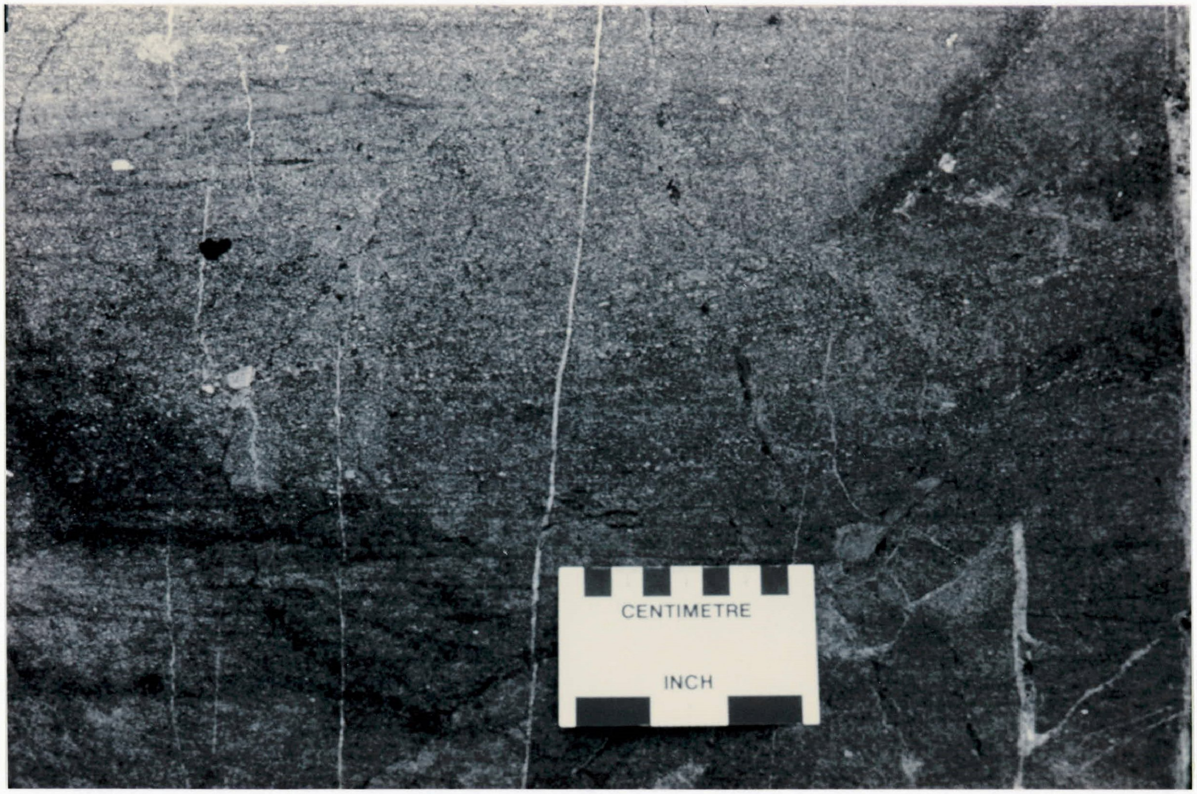
Beds and single cross-sets are up to 30 cm thick. Darker fine-grained laminae often outline the stratification.

All of the cross-beds and ripples with concave (foreset) laminae are concave north.

A few jasper grains were observed.

Exposure problems. Except at one small (metres square) possible site, no bedding plane views were seen in outcrops #99, 100 and 101. Without good three-dimensional exposure of

Fig. 5.12: Middle: very coarse sand grains outline the foresets and sweeping (tangential) toesets of a planar cross-bed. The largest clasts are intraformational silt chips. Note the low-angle lamina of finer sand at the upper left. Outcrop #100.


CENTIMETRE
INCH

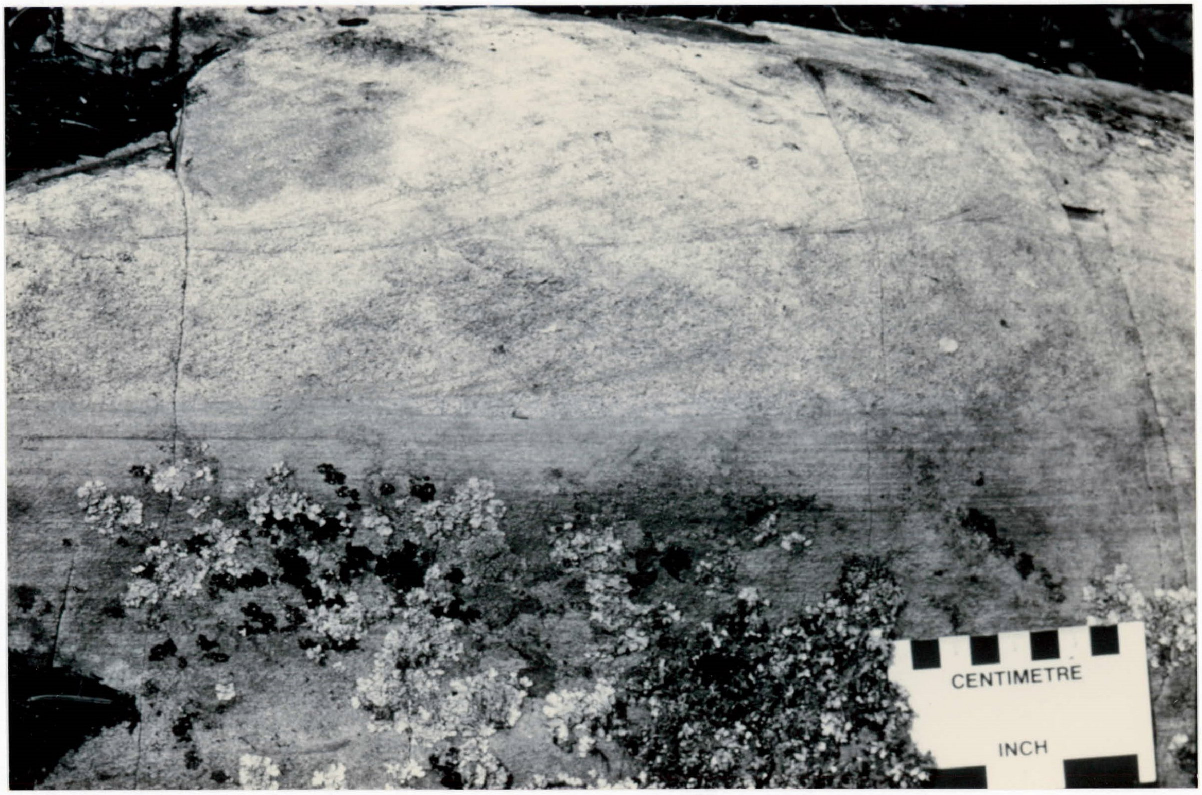
individual beds, differentiating between planar and trough cross-sets can be problematic. When planar foresets are laterally persistent over several metres, especially in tabular planar cross-sets stacked in cosets, such cross-sets are most likely not troughs.

As shown in Reineck and Singh (1980, Figs. 41, 44, 46, 52), in flow-perpendicular sections (taken along depositional strike) troughs are easily identifiable, but in flow-parallel sections (taken along depositional dip) trough cross-sets resemble some views of wedge-shaped planar cross-sets. The scale of the preserved units is important: a wedge-shaped planar cross-set may be a mere erosional remnant of a once much larger dune (i.e. small planar-looking part of a large trough).

Also, as paleocurrents vary vertically in section, a vertical plane of section will intersect successive cross-sets at possibly different/varying angles to each cross-set´s paleocurrent, the result being that cross-sets identical in three dimensions can be exposed in sectional views having different appearances.

Planar cross-bedding. Tabular and wedge-shaped cross-sets are up to 30 cm thick. Thinner (up to 10 cm) tabular sets in particular are often stacked in cosets several sets thick (Fig. 5.13). Foreset laminae vary from straight to concave north, with angular and tangential toesets, respectively. Pebbles are usually preferentially concentrated in the lower foresets and toesets.

Fig. 5.13: Finer planar, horizontal and parallel laminated sandstone capped by stacked sets of coarser planar tabular cross-strata with tangential toesets. Outcrop #101.


CENTIMETRE
INCH

Trough cross-bedding. Trough cross-sets are up to tens of cms thick, mostly 5-10 cm deep and 0.5-1.0 wide. The broad, shallow troughs (Fig. 5.14) typically have 1:10 depth : width ratios.

Ripples. Ripples have both trough (Fig. 5.15) and planar forms. Flasers outlining ripples reveal troughs 3-20 mm deep and up to 6 cm wide. Flasers are gradational to planar, horizontal to slightly undulose mud laminae 1-3 mm thick (Fig. 5.16). In rare cases fine-grained sand laminae drape ripples and cross-bed foresets.

The larger trough ripples, up to 5 cm deep and 20-30 cm wide, are transitional in size to trough cross-beds.

Plane lamination. Laminae are planar, horizontal and parallel, in beds up to tens of cms thick. By definition, the laminae are up to 10 mm thick, the thinnest being 1-2 mm thick. Both horizontal and foreset (inclined) laminae are defined by consistent grain size changes (coarser, finer laminae), and rare rusty layers. The best developed and most laterally continuous plane lamination tends to be in the finer (fine to very fine-grained) sandstone. The finer laminae are often darker.

Some plane lamination varies laterally to slightly undulose laminae and planar cross-lamination (ripples).

Very low-angle laminae may be:

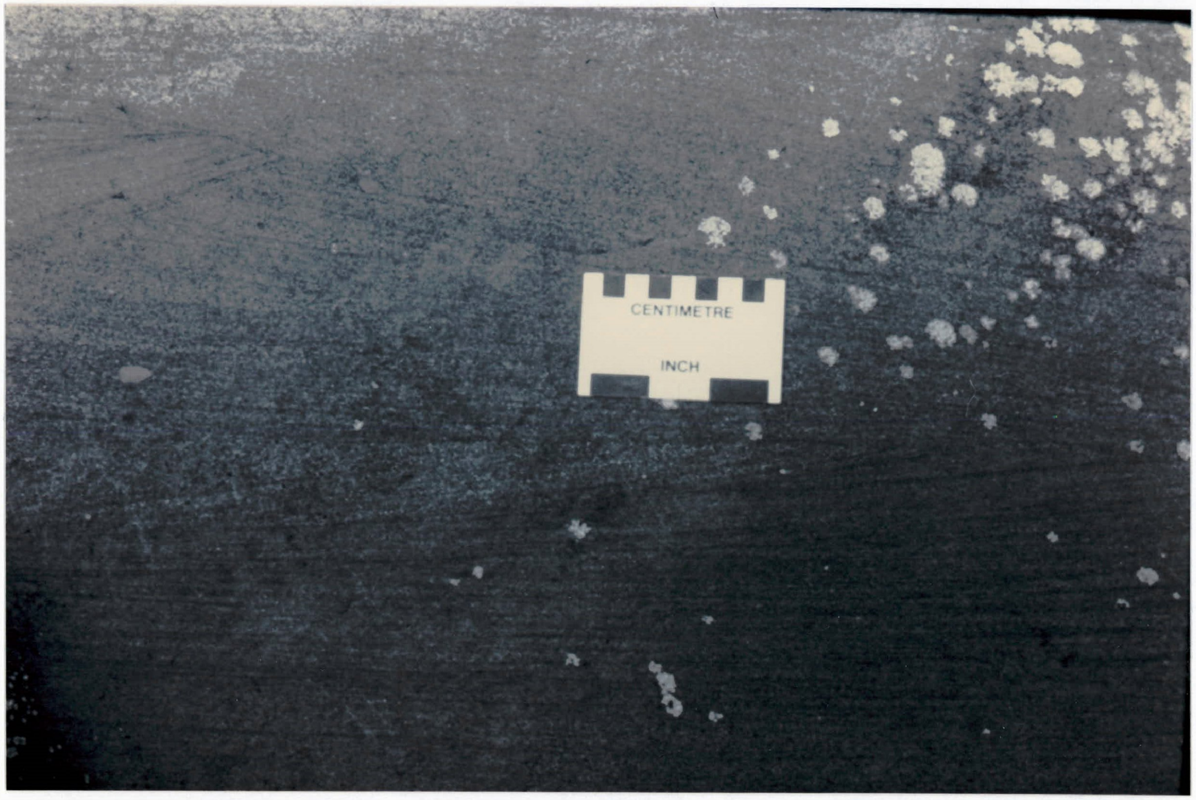
1) primary features (e.g. Miall's (1978a) Sl facies);

2) part of larger, broad trough cross-bed floors;

3) near-tangential toesets of planar cross-beds, in

Fig. 5.14: Top left: trough cross-sets. Outcrop #101.

Fig. 5.15: Middle: cross-laminated (trough ripples) sandstone, overlain by plane laminae at top. Outcrop #100.


CENTIMETRE
INCH

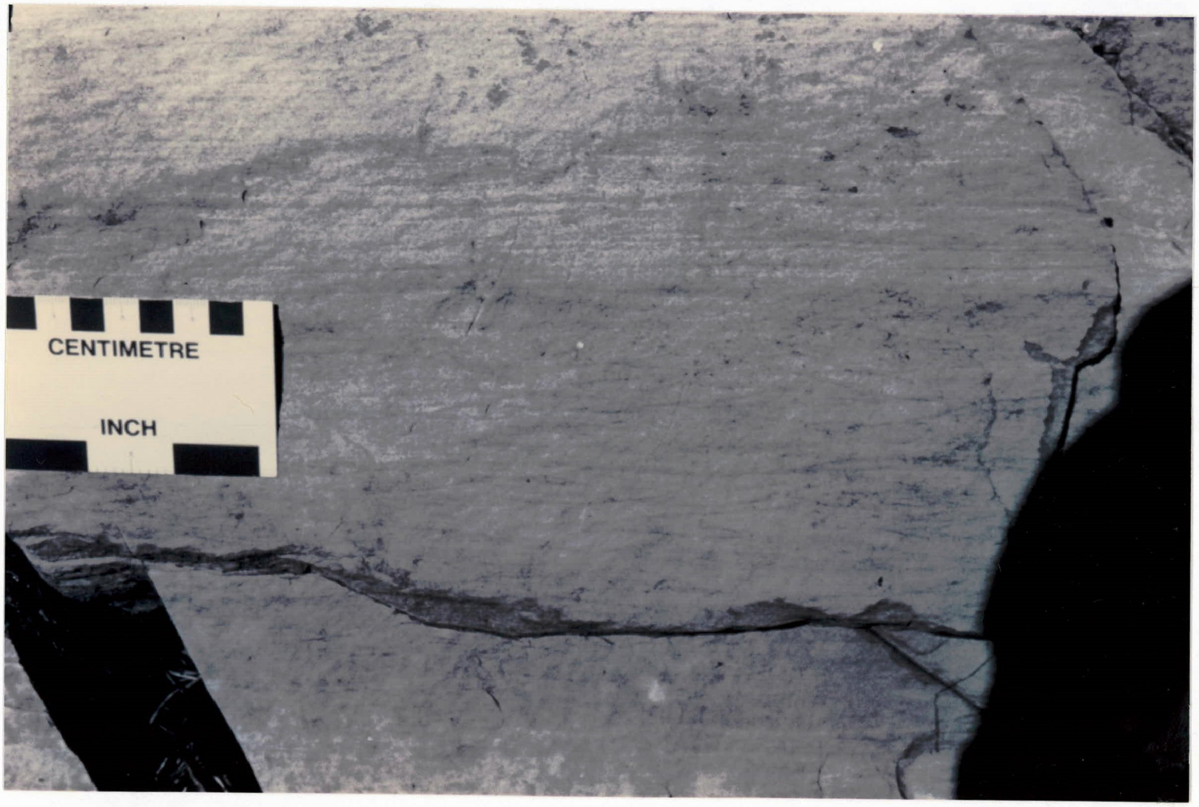
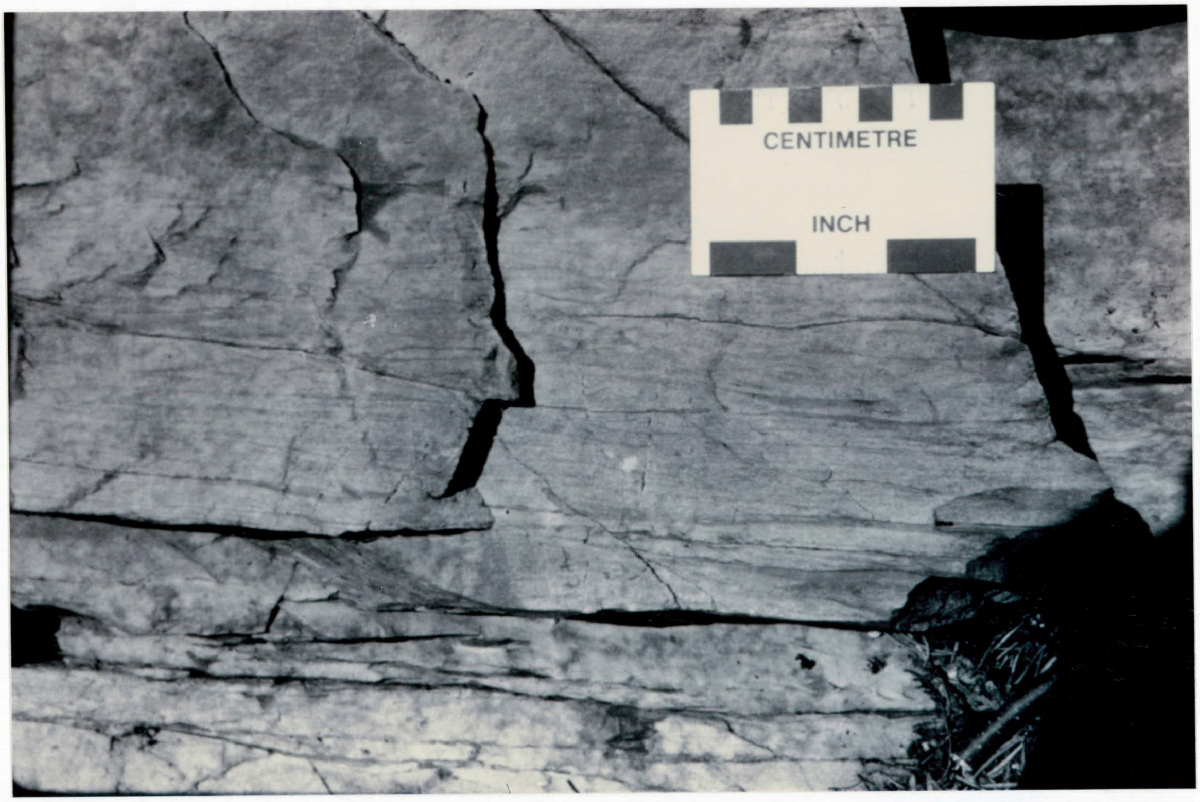

CENTIMETRE
INCH

Fig. 5.16: Fine-grained sandstone with horizontal, planar to slightly undulose mudstone (dark) laminae gradational to flasers. Outcrop #100.


CENTIMETRE
INCH

flow-parallel section; or

4) flow-perpendicular sectional views of planar cross-beds (Reineck and Singh, 1980, Fig. 41), in which laminae can also appear horizontal.

Convoluted stratification. Two convoluted horizons, one 40 cm thick, and very rare contorted laminae were noted. This soft-sediment deformation is usually in finer (fine to very fine-grained) sandstone.

Massive beds. These form a very minor part of Unit A.

Mudstone clasts. These are about 2-12 mm thick and are associated with coarser (very coarse or coarse to fine-grained) sandstone.

Herringbone cross-stratification. Figs. 5.17 and 5.18 show this very small-scale lithofacies association of cross-bedded, rippled and low-angle to horizontally plane laminated sandstone.

The cross-bed sets have an apparent herringbone (bimodal-bipolar) pattern in section, with four paleocurrent reversals in 55 cm (Figs. 5.17-5.19). The form of the exposure does not allow a three-dimensional view, only a cross-section, hence the apparent paleocurrents could not be examined fully.

Some beds in Fig. 5.18 have very vaguely defined horizontal and low-angle laminae, which could be cross-bed toesets or shallow trough margins. Most of the ripples (up to 3 cm thick) are wedge-shaped sets, not the tabular units shown diagrammatically in Fig. 5.18.

Several metres to the west approximately along strike, a

Fig. 5.17: Small exposure of sandstone showing apparent herringbone cross-stratification. There are four paleocurrent reversals within 55 cm of section. Outcrop #100. Details are shown in Figs. 5.18 to 5.20.


CENTIMETRE
INCH

Fig. 5.18: Section from outcrop #100 and Fig. 5.17. Herringbone cross-sets appear to be well-defined in the two-dimensional outcrop view. Three beds are at the cross-bed/ripple size boundary (5 cm thick). Top to north. Sandstone is medium to fine-grained and moderately to well sorted.

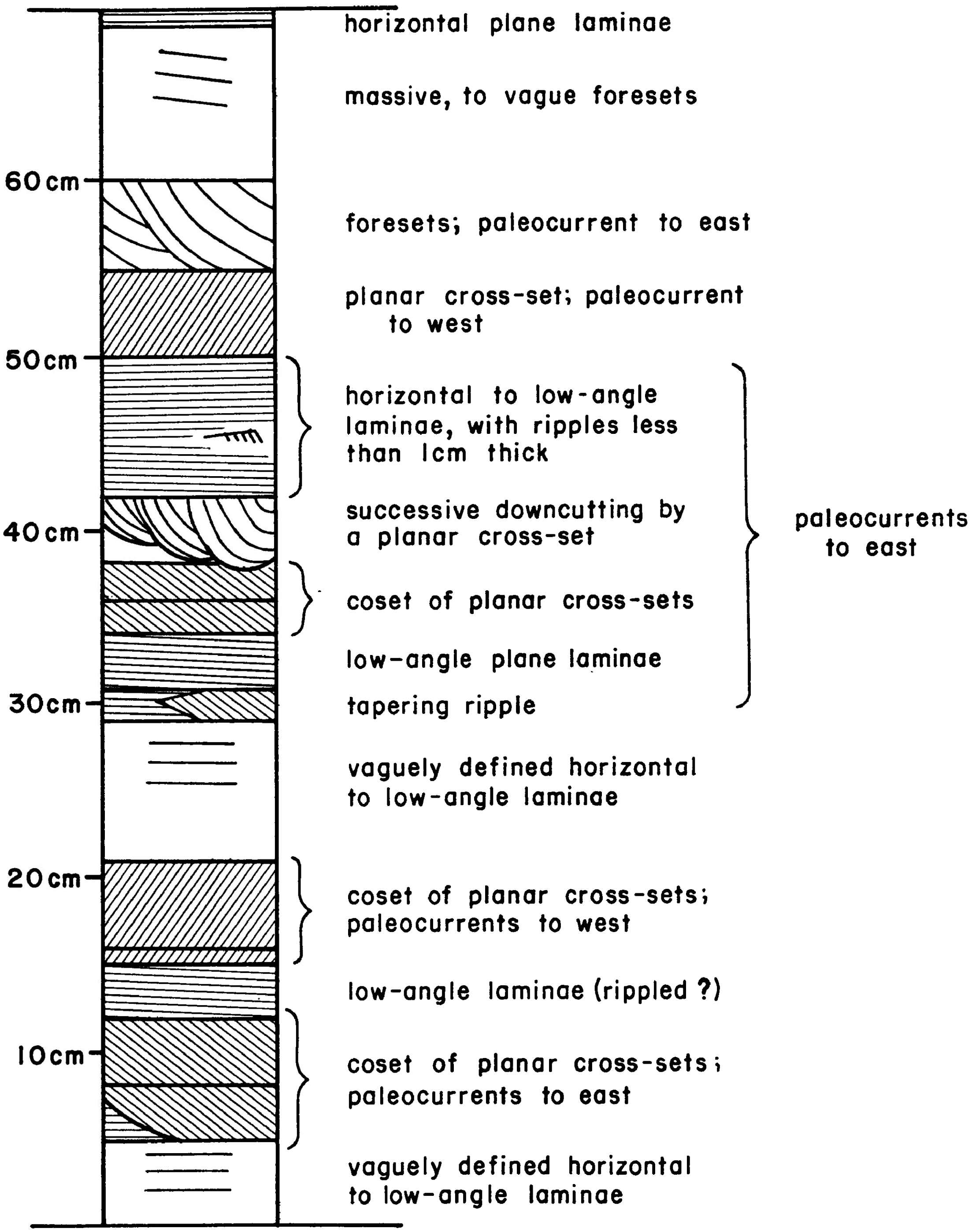
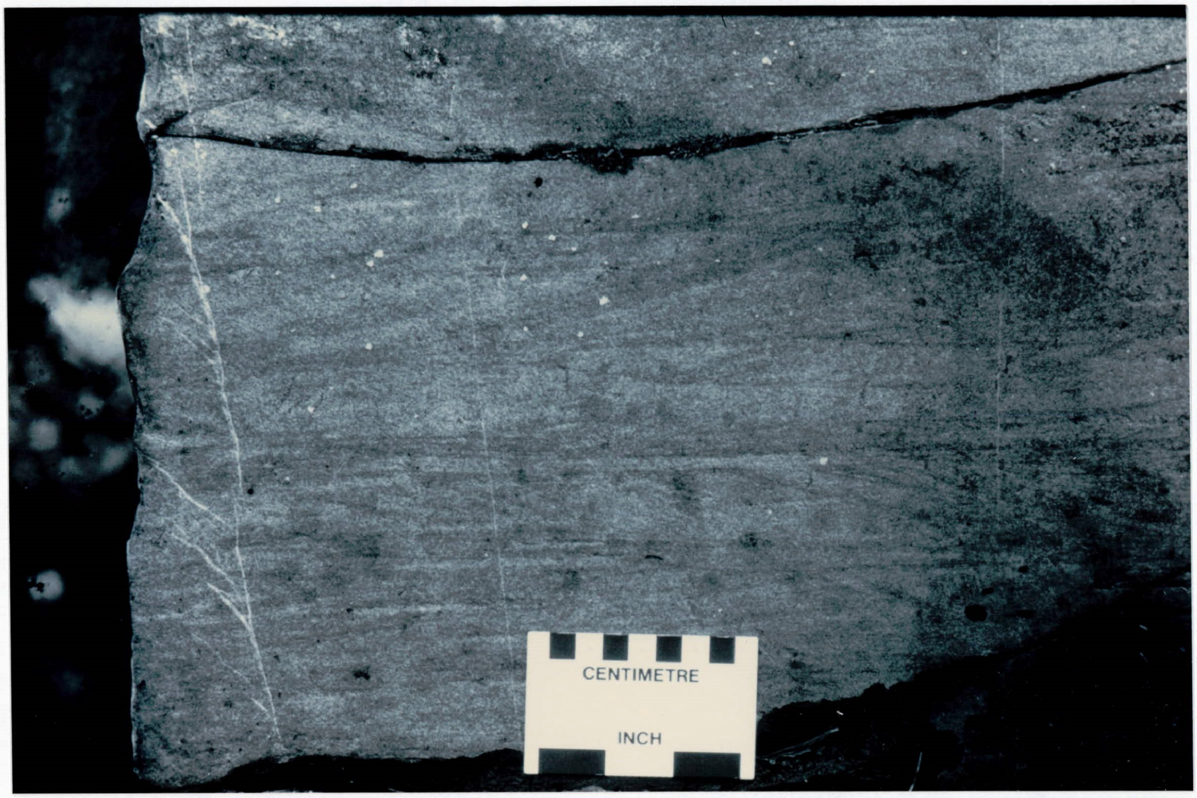
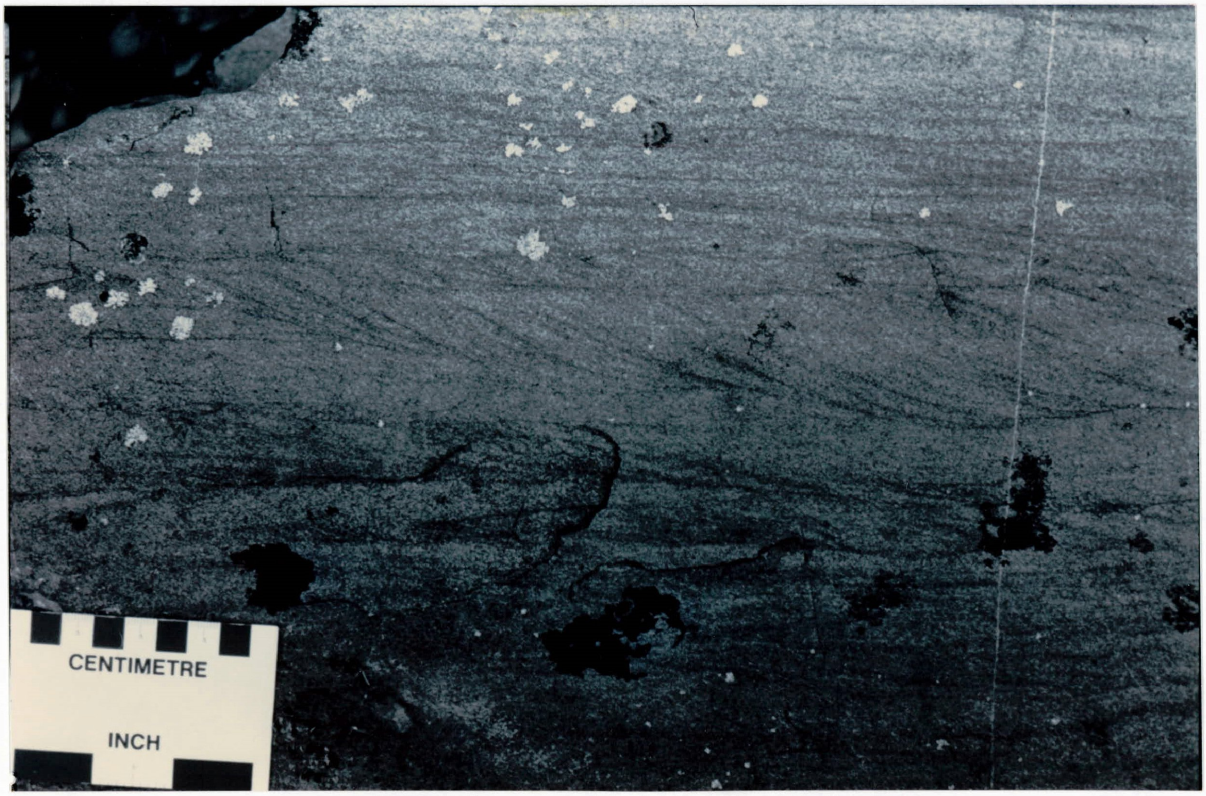

horizontal plane laminae
massive, to vague foresets
60cm
foresets; paleocurrent to east
planar cross-set; paleocurrent to west
50cm
horizontal to low-angle laminae, with ripples less than 1cm thick
successive downcutting by a planar cross-set
40cm
coset of planar cross-sets
low-angle plane laminae
30cm
tapering ripple
paleocurrents to east
vaguely defined horizontal to low-angle laminae
20cm
coset of planar cross-sets; paleocurrents to west
low-angle laminae (rippled ?)
10cm
coset of planar cross-sets; paleocurrents to east
vaguely defined horizontal to low-angle laminae

Fig. 5.19: Detail of lower parts of Figs. 5.17 and 5.18. Herringbone cross-strata (ripples) show one paleocurrent reversal.

Fig. 5.20: Detail of middle parts of Figs. 5.17 and 5.18. Rippled and plane laminated sandstone. Middle: note the successive downcutting (to the right) of the toesets. the result of increasing depth of leeside backflow scour. A paleocurrent reversal may be present above the downcutting foresets.


CENTIMETRE
INCH


CENTIMETRE
INCH

1.2 m thick interval of fine-grained sandstone is dominated (about 85%) by planar, horizontal and parallel lamination. Contacts between beds showing variations in the development of plane lamination and cross-lamination (ripples) are transitional. Finer background sediment varies from muddy flasers and laminae (Fig. 5.16) to dark, silty very fine-grained laminae.

<u>Trends within Unit A</u>. Unit A is an assemblage dominated by cross-bedding, with a subordinate amount of plane laminated and rippled sandstone. Where cross-beds and plane laminae are interbedded in horizons metres thick, cross-beds are thicker and more abundant. It is estimated that the section exposed along the shoreline of outcrops #99, 100 and 101 is approximately:

1) up to 70% planar cross-bedded;
2) 10-20% trough cross-bedded; and
3) 10-20% plane laminated, rippled and of other facies.

Because of the exposure problems previously discussed, the relative amounts of planar and trough cross-sets are very approximate.

A quick count of well exposed cross-sets along the shorelines of outcrops #99, 100 and 101, mostly apparent planar sets, gave 73% (77 of 106) to the west or southwest (after a simplistic single rotation to horizontal), a broadly unimodal pattern.

The southwest part of Unit A (outcrops #99, 100) is rich in sandstone that is medium to fine-grained, well to moderately sorted, pebble-free, plane laminated and cross-laminated (rippled). The herringbone cross-bedding and most of the mud laminae, mud chip clasts and flasers are in outcrop #100. This contrasts with the northern and northwest parts of Unit A in outcrop #101 where cross-bedded, coarse to fine-grained, moderately to poorly sorted, pebbly sandstone is most characteristic.

Progressing from the southwest (outcrop #99) to the northeast and north (outcrop #101), the sandstone:

1) is pebble-free, with very rare granules and small (mms thick) mud chip clasts;
2) contains the first pebbles seen, up to 45 mm (all clast measurements are of short axes) and preferentially located in foresets and toesets;
3) has 20-60 mm pebbles;
4) has 20-80 mm clasts; and
5) has 70-120 mm clasts (not 7-12 m clasts, a typographic error in Devaney and Fralick, 1985, p.128).

Pebbly cross-beds are more abundant to the north and northeast in Unit A. Pebble bands are rare and only vaguely defined, being laterally discontinuous and one clast thick. These become increasingly common northward. No thin conglomerate beds are present. The pebbles and cobbles, Unit A's coarsest fraction, define a subtle coarsening-north trend. As all of the stratigraphic top indicators are to the north, it seems reasonable to call Unit A a subtly coarsening-upward sequence.

5.5 Leduc Lake Area, Leduc Township

The section in outcrop #41 (Fig. 5.21) is dominated by coarse to medium-grained, moderately to poorly sorted pebbly sandstone. The more poorly sorted beds contain some fine sand. The sandstone is massive, cross-bedded and plane laminated. Thin conglomerate beds comprise about 14% of the section.

Planar cross-beds. The cross-stratified units in Fig. 5.21 are mostly planar cross-bed foresets outlined by pebbles (Fig. 5.22). Deformed pebbles 2-10 mm thick, mostly less than 6 mm thick, and very coarse sand grains outline most of the foresets. Coarser foresets contain 5-25 mm clasts. These pebbly foresets are usually 3-4 cm apart (Fig. 5.22). Some sandy foreset laminae 1-10 mm thick are defined by layers alternatingly rich and poor in coarse sand. Foreset angles are up to 20% from the paleohorizontal. Lateral-diagonal continuity of foresets is up to 30 cm. Some toesets are tangential and rich in coarser sand. All of the foresets give apparent paleocurrents to the west.

Plane lamination. Plane laminated sandstone displays the usual planar, horizontal and parallel laminae mms thick and defined by coarse to medium-grained layers versus medium-fine ones. Very low-angle laminae may be large cross-bed toesets.

Fine sandstone. The section also contains a small

Fig. 5.21: Section of outcrop #41. Top to north. Legend in Fig. 5.2.

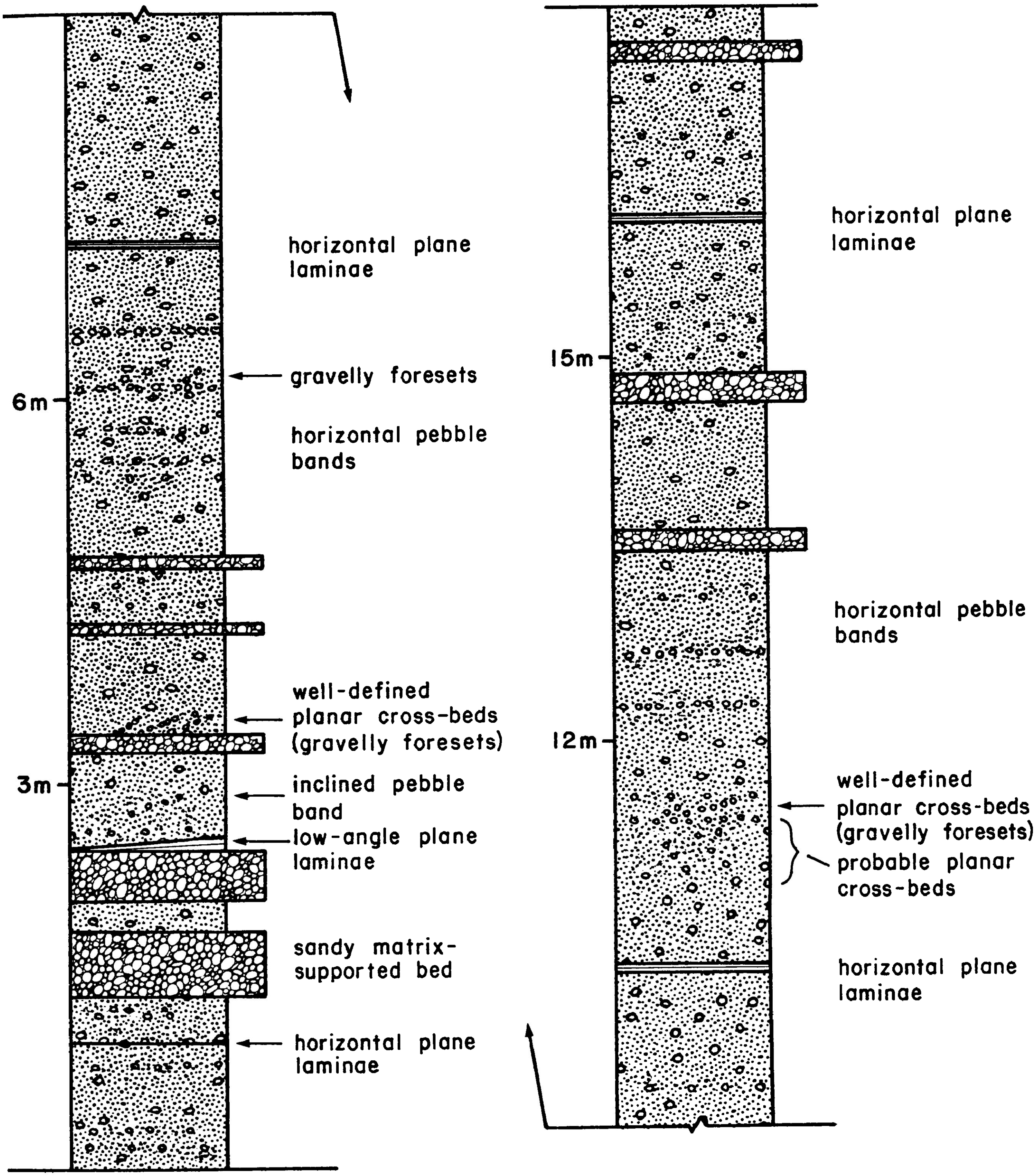

horizontal plane laminae
gravelly foresets
6m
horizontal pebble bands
well-defined planar cross-beds (gravelly foresets)
3m
inclined pebble band
low-angle plane laminae
sandy matrix-supported bed
horizontal plane laminae
horizontal plane laminae
15m
horizontal pebble bands
12m
well-defined planar cross-beds (gravelly foresets)
probable planar cross-beds
horizontal plane laminae

Fig. 5.22: Top: conglomerate-sandstone bed contact is parallel to the outcrop's bedding (in Fig.5.21), and outlines the paleohorizontal. Bottom: horizontal pebble band with the same strike as the schistosity (elongate clasts). Middle: pebbles and very coarse sand grains outline foresets at about 15° from the paleohorizontal and the schistosity. Outcrop #41.


CENTIMETRE
INCH

number of finer, better sorted sandstone beds. Two examples are beds 3 and 4 cm thick, fine-grained, well sorted and non-pebbly, in contrast to the usual beds described above.

Pebble bands. The planar and horizontal pebble bands are both "well developed", with good lateral continuity and gradational to thin conglomerate beds, and "poorly developed", with poor lateral continuity. Pebbly foresets are a distinct type of pebble band discussed above. Like the pebbly foresets there is usually a slight enrichment in coarser sand along the pebble bands, much like a coarser matrix without an intact pebble framework to fill.

Conglomerate beds. The beds in Fig. 5.21 are thinner and finer-grained versions of the CMB's usual coarse, polymict conglomerate (see Chapter 6).

The massive, clast-supported and ungraded beds are mostly up to 20 cm thick, with two thicker 40 and 50 cm beds. They are often too thin to show crude internal stratification, except for one bed with a 4 cm thick, 20 cm wide granule-rich lens.

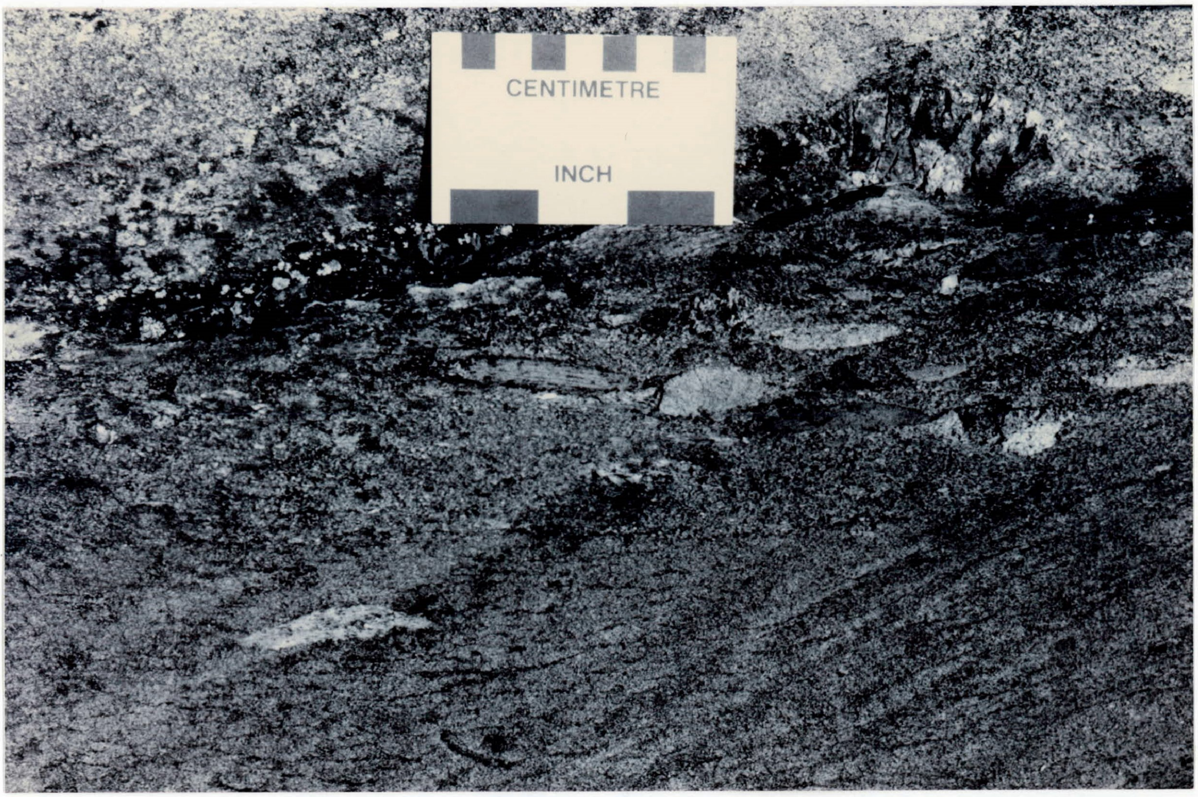
The northern parts of some beds are not tightly packed, being partly matrix-supported. Within a bed this can vary laterally, over metres, to clast-support.

Clasts in the pebble bands range from 2-42 mm (short axis). The thin conglomerate beds have D10 measurements of 17-42 mm (short axes only). Large cobbles are notably absent.

One sharp contact between finer and coarser "sandstone" beds is due to the coarser one being the sandy matrix component of a conglomerate bed with a sharp (erosive?) base (Fig. 5.23), giving tops north.

Similar facies. Outcrop #76, 1.1 km to the west along strike, contains a similar suite of coarse to medium-grained, moderately sorted pebbly sandstone, with pebble bands (clasts up to 85 mm) and thin (15 mm) conglomerate beds (clasts up to 50 mm).

Fig. 5.23: Middle: sharp-based coarser sandstone is the matrix component of the overlying conglomertate bed. Outcrop #41.


CENTIMETRE
INCH

CHAPTER 6: THE CMB's CONGLOMERATIC ASSEMBLAGE

6.1 Introduction

The geographic extent of the CMB's conglomerate-rich lithofacies assemblage is shown in Fig. 3.1 and 3.2. Fig. 3.1 gives the locations of specific outcrops referred to in the text. For local details such as bedding and schistosity measurements the reader should consult Mackasey's (1975, 1976) maps.

The CMB's conglomeratic assemblage is very similar to the NMB's, except that there is a greater variety of lithofacies in the former than in the latter. Therefore, details regarding characteristics such as the textures and stratification of this assemblage's conglomerate and sandstone facies will not be repeated here, and the reader is referred to Sections 2.1, 2.2, 2.4 and 2.5 for this information. The descriptions that follow [in this chapter] will concentrate on the interesting sedimentological details present in the CMB's conglomeratic assemblage.

Figs. 6.1 to 6.5 show representative stratigraphic sections from selected outcrops. Some measured sections (Fig. 6.3 to 6.5) are sandstone-rich, but the sections given in these figures are biased towards the most interesting sandstone-rich horizons. Exposures with 90-100% conglomerate are common but there is little point in illustrating such monotonous sections. Thinner (less than 20 cm) conglomerate

Fig. 6.1: (a) Section of southern part of outcrop #13.

(b) Section near hilltop summit of outcrop #13; top to north.

Legend in Fig. 5.2.

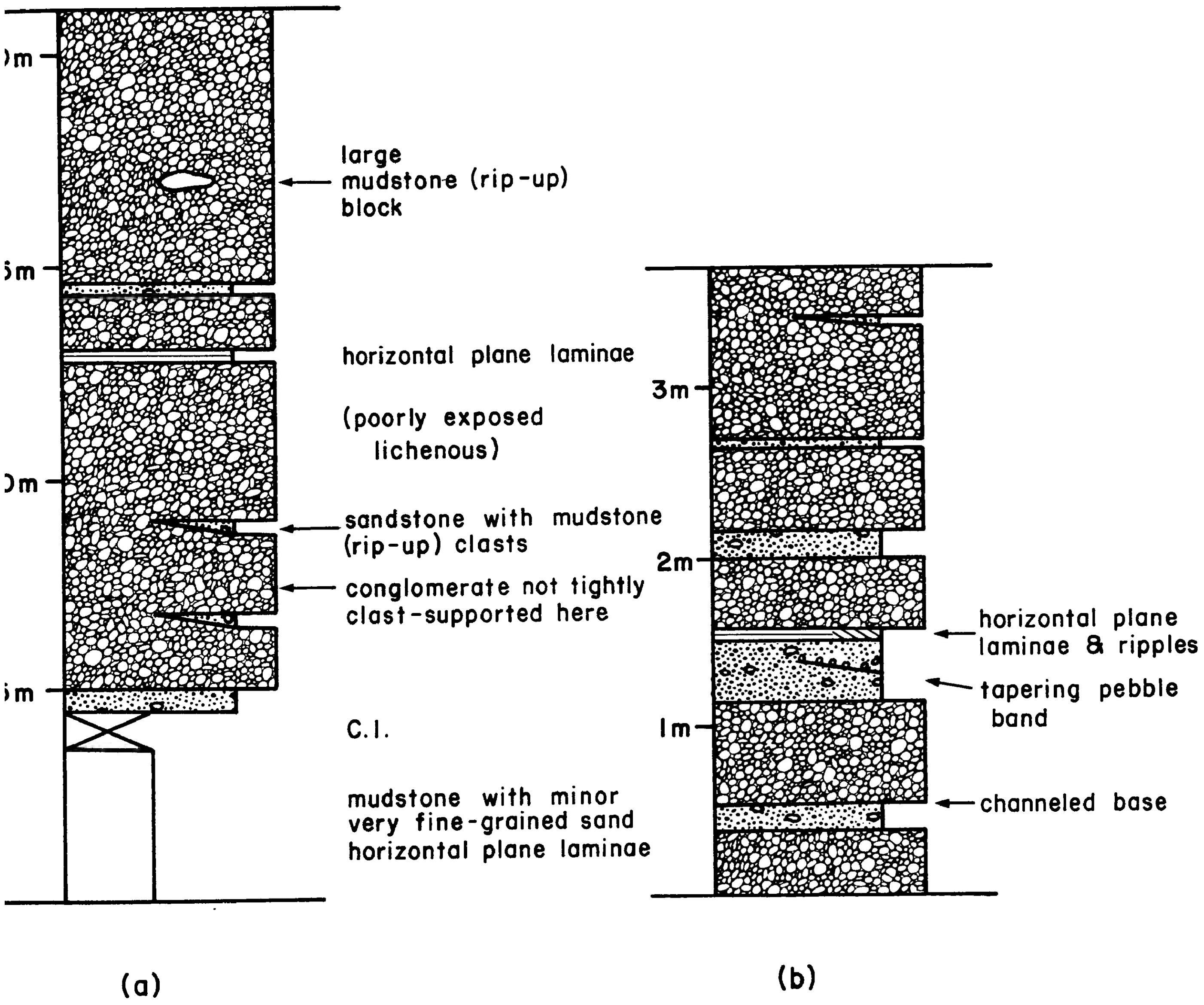

m
m
m
m
large mudstone (rip-up) block
horizontal plane laminae
(poorly exposed lichenous)
sandstone with mudstone (rip-up) clasts
conglomerate not tightly clast-supported here
C.I.
mudstone with minor very fine-grained sand horizontal plane laminae
(a)
3m
2m
1m
horizontal plane laminae & ripples
tapering pebble band
channeled base
(b)

Fig. 6.2: Section from outcrop #65. Top to north. Legend in Fig. 5.2.

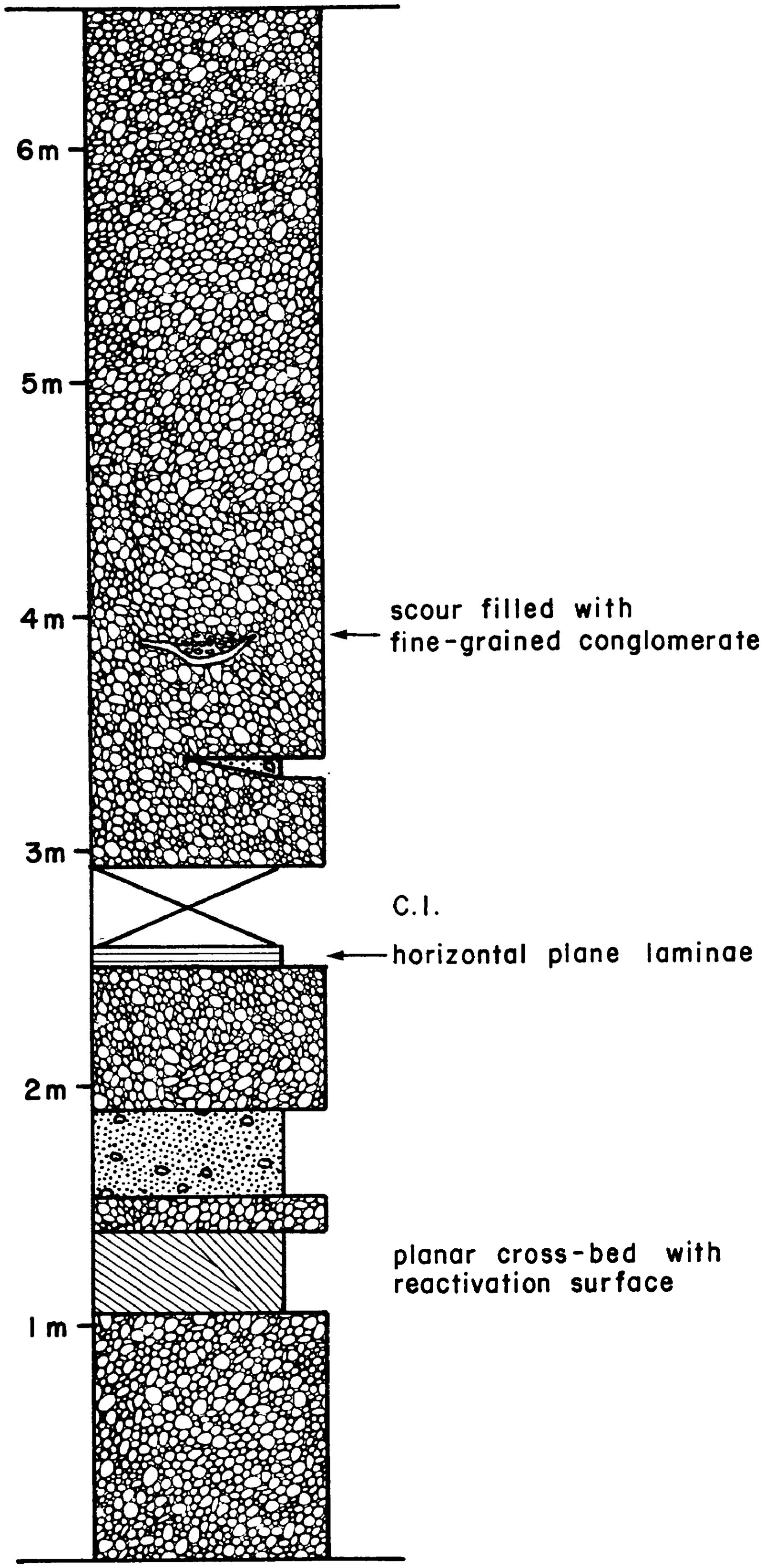

6m
5m
4m
scour filled with fine-grained conglomerate
3m
C.I.
horizontal plane laminae
2m
planar cross-bed with reactivation surface
1 m

Fig. 6.3: Section from outcrop #66. Top to north. Legend in Fig. 5.2.

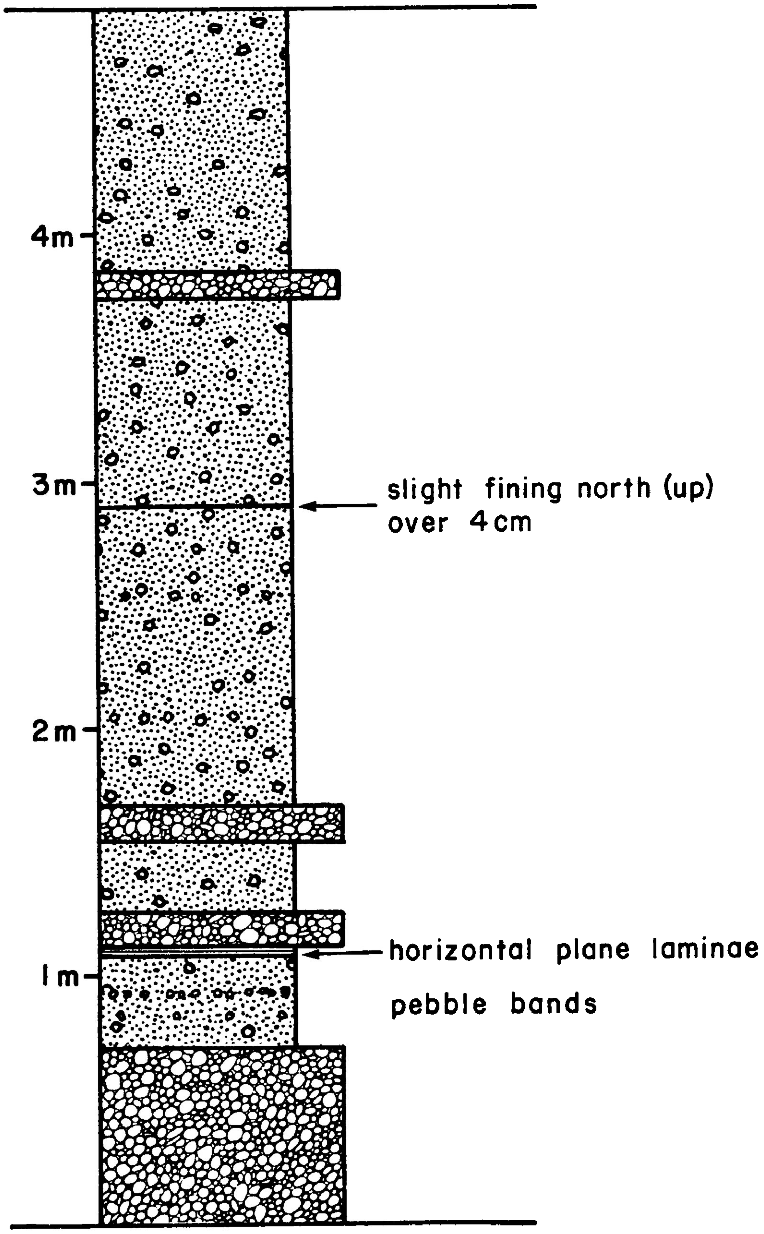

4m
3m
2m
1m
slight fining north (up) over 4cm
horizontal plane laminae
pebble bands

Fig. 6.4: Sections from outcrop #67. Top to north. Legend in Fig. 5.2.

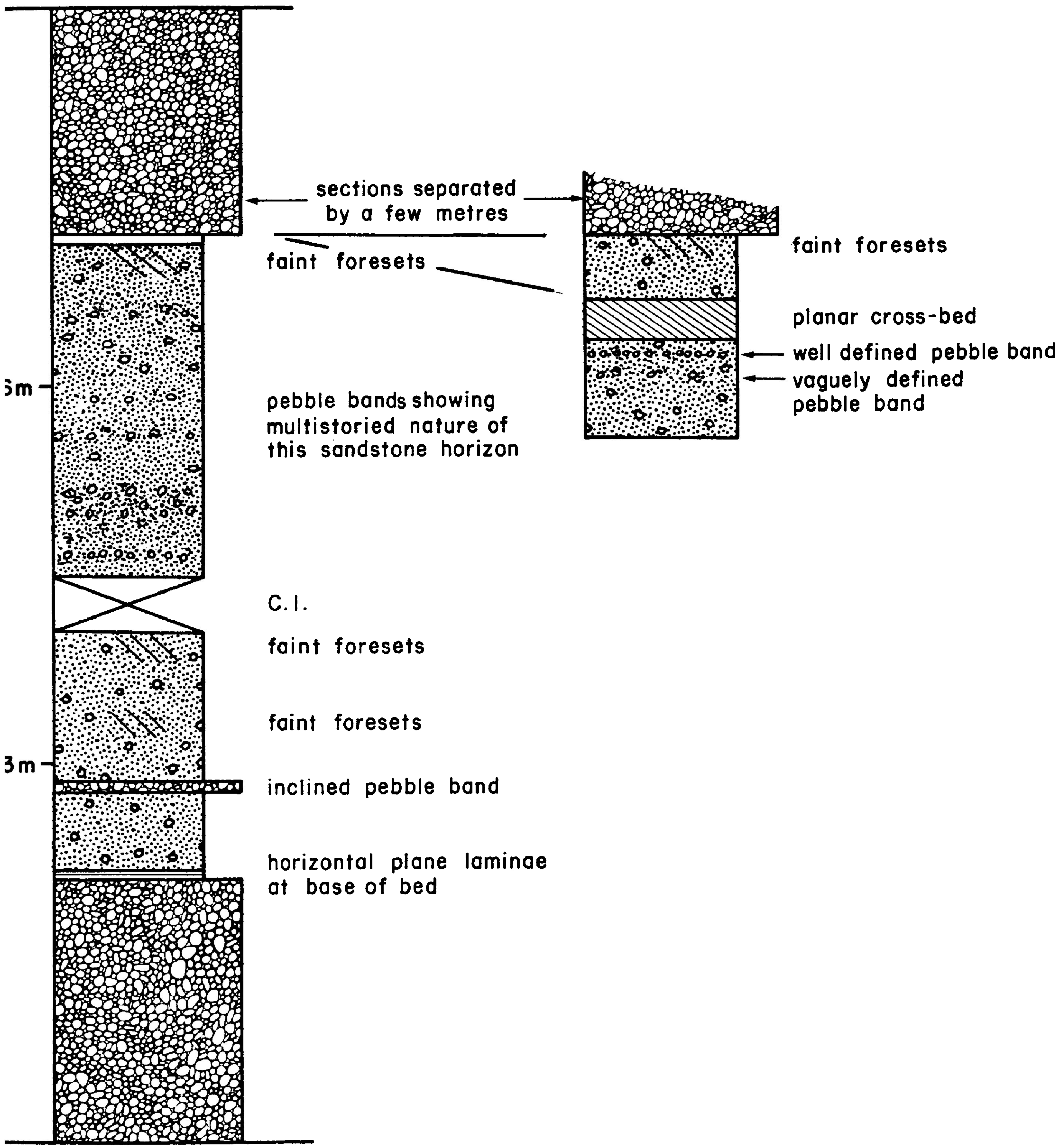
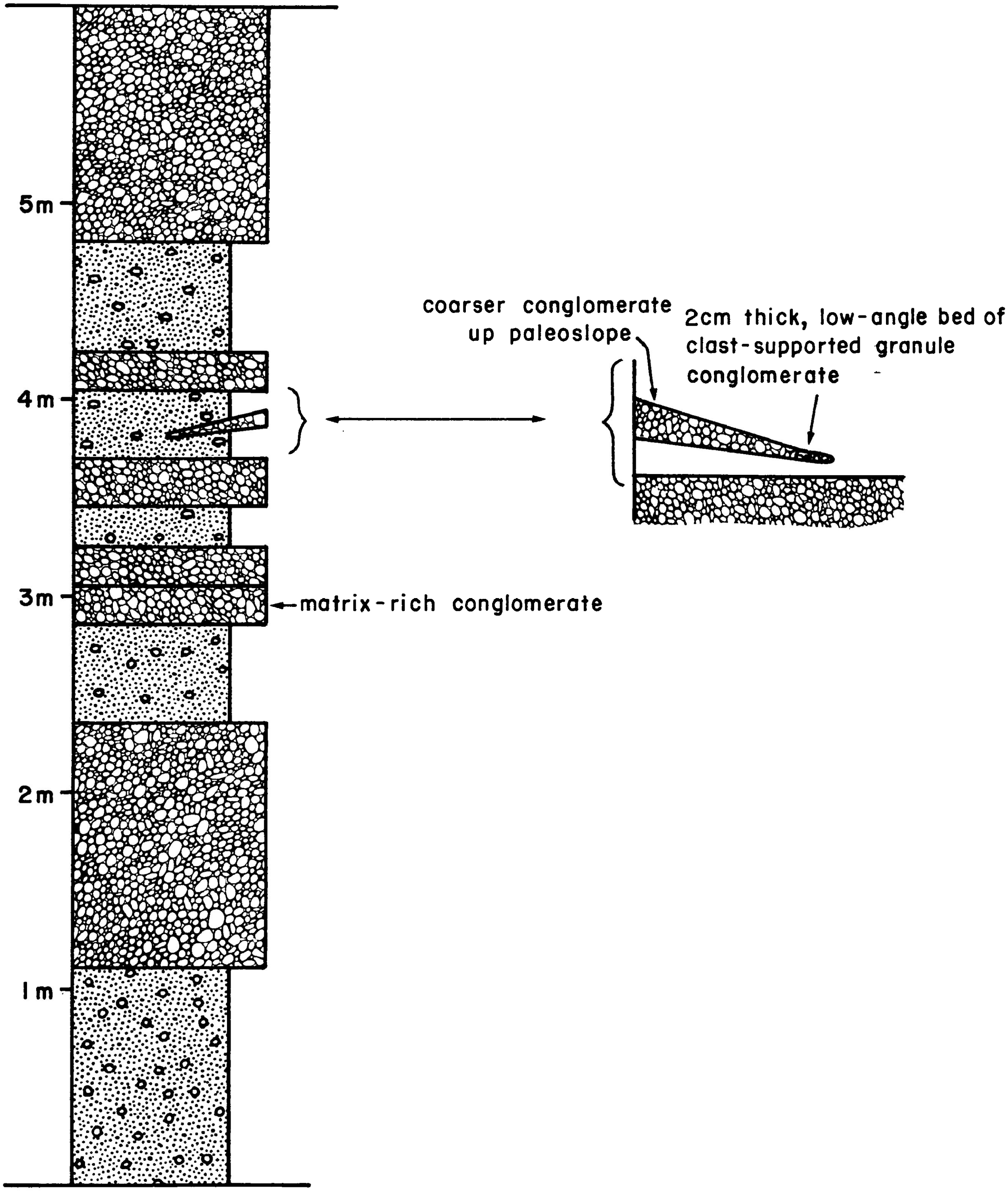

sections separated by a few metres
faint foresets
pebble bands showing multistoried nature of this sandstone horizon
5m
C.I.
faint foresets
faint foresets
3m
inclined pebble band
horizontal plane laminae at base of bed
faint foresets
planar cross-bed
well defined pebble band
vaguely defined pebble band

Fig. 6.5: Section from outcrop #80. Apparent horizontal thicknesses are given. Top to north. Legend in Fig. 5.2.



apparent thickness only
(close to true thickness)

beds are notable in sandstone-rich (50% of section) intervals several metres thick in outcrops #13 and 20.

Table 6.1 lists the assemblage's outcrops, for use with Table 6.2.

TABLE 6.1

Township	Outcrop #
Dorothea	12, 13
Sandra	16, 15, 17, 57, 56, 20, 14
Irwin	53
Walters	123, 122, 117, 9, 110, 109, 108
Leduc	64 to 71, 97, 80 to 86, 76, 79, 42, 32, 45, 31

Table 6.1: Outcrop areas in the CMB's conglomeratic assemblage. Outcrops are listed west to east.

TABLE 6.2

Township	Outcrop #	D10 (in mm) S / L	L/S
Dorothea	12	57 / 82	1.44
	13	81 / 116	1.43
Sandra	15	92 / 125	1.36
	17	94 / 123	1.31
	57	62 / 100	1.61
	56	93 / 153	1.65
	20	118 / 168	1.42
	14	86 / 151	1.76
Irwin	53	104 / 155	1.49
Walters	9	85 / 123	1.45
	110 west	94 / 167	1.78
	110 mid	70 / 95	1.36
	110 east	86 / 131	1.52
	109	281 / 638	2.27
Leduc	71	136 / 226	1.66
	70(B)	141 / 204	1.45
	70(A)	130 / 195	1.50
	67	94 / 161	1.71
	64	85 / 116	1.36
	80	74 / 110	1.49
	81	144 / 217	1.51

TABLE 6.2 (Cont'd)

Township	Outcrop #	D10 (in mm)		L/S
		S	L	
Leduc (cont'd)	97	111	176	1.59
	83	141	222	1.57
	86	120	172	1.43
	88	76	105	1.38
	98	106	145	1.37
	105	68	97	1.43
	103	90	142	1.58
	104	65	97	1.49
	96	85	157	1.85
	94	90	145	1.61
	102	30	61	2.03
	79 north	75	152	2.03
	79 south	65	96	1.48
	76	96	146	1.54
	31	131	206	1.57
	31	133	245	1.84
	31	67	126	1.88
	31	44	68	1.55
	45	108	183	1.68
	42	102	151	1.48
	32	103	152	1.48

(outcrops listed below are part of the Coral Lake volcaniclastic "sediments")

Leduc	37, 38	74	172	2.32
	46	66	151	2.29
	47	97	176	1.81
	48	64	158	2.47
	29	126	256	2.03
	30	83	170	2.05
	25	74	204	2.76
	24	72	182	2.53
Legault	52	82	244	2.98
	131	129	264	2.05
	132	137	383	2.80
	62	140	309	2.21
	61	79	189	2.39

Table 6.2: Measurements of average maximum deformed clast size (D10) in conglomeratic outcrops of the CMB. Township locations are given in Fig. 1.1. Outcrop locations are shown in Figs. 2.8, 3.1. D10 measurements, the average size of an outcrop's ten largest clasts, are given in short (S) and long (L) axes, each in mm, with short axes usually perpendicular to the schistosity and the

long axes usually parallel to the schistosity (see Section 1.7). "L/S" is a measure of the ellipticity of the sectioned area of the coarsest clasts.

6.2 Conglomerate Lithofacies

As in the NMB (Chapter 2), the CMB's conglomeratic assemblage exposures are composed of massive to crudely bedded polymict conglomerate with schistose, clast-supported, poorly to moderately sorted and ungraded frameworks (Figs. 6.6, 6.7). The matrix to the clasts is coarse to fine-grained, poorly to moderately sorted sandstone, and is usually identical in character to the sandstone interbeds.

Table 6.2 lists the D10 measurements (see Section 1.7) from the CMB; most of the data is from the conglomeratic assemblage (refer to Table 6.1 and Figs. 3.1 and 2.8).

The composition of the polymict conglomerate clasts is discussed in Chapter 10.

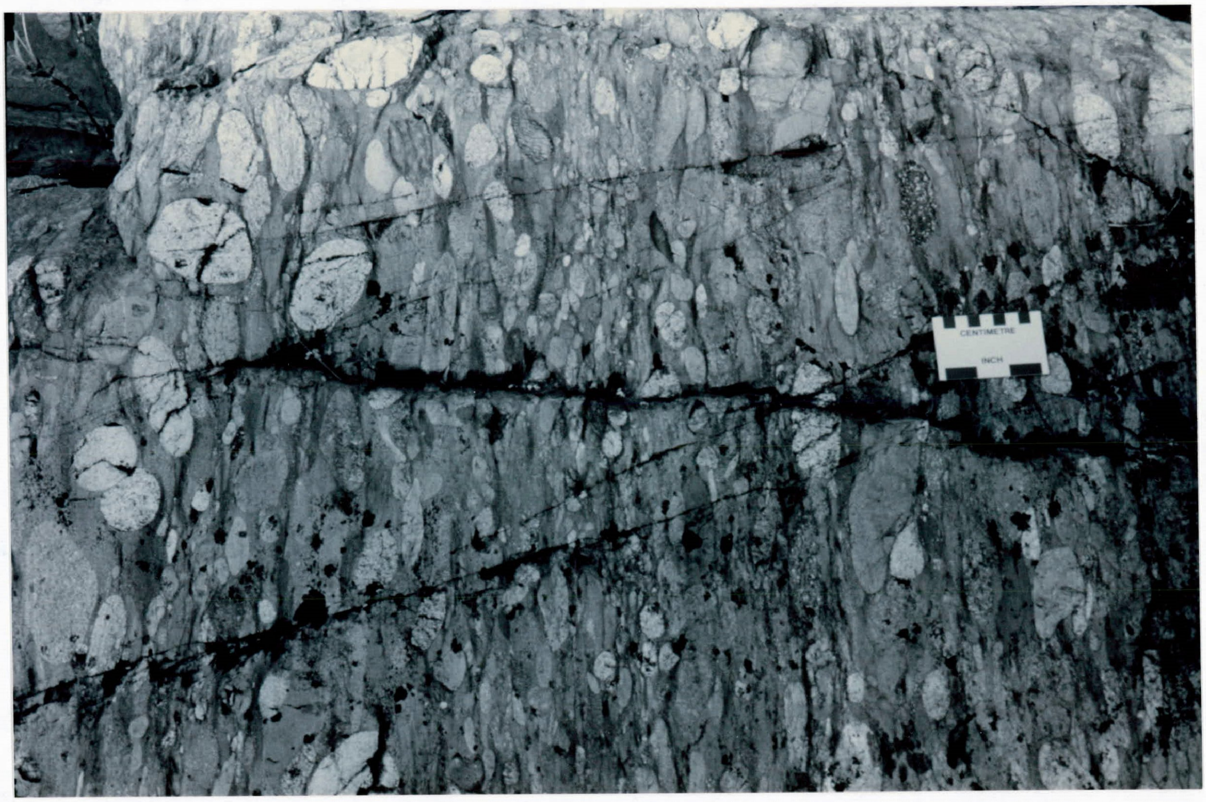
Fine-grained conglomerate. In Dorothea and Sandra Townships (Fig. 3.1) distinctly fine-grained conglomerate is common. Individual beds have fine-grained (4-32 mm), moderately to well sorted clast-supported frameworks (Figs. 6.8, 6.9).

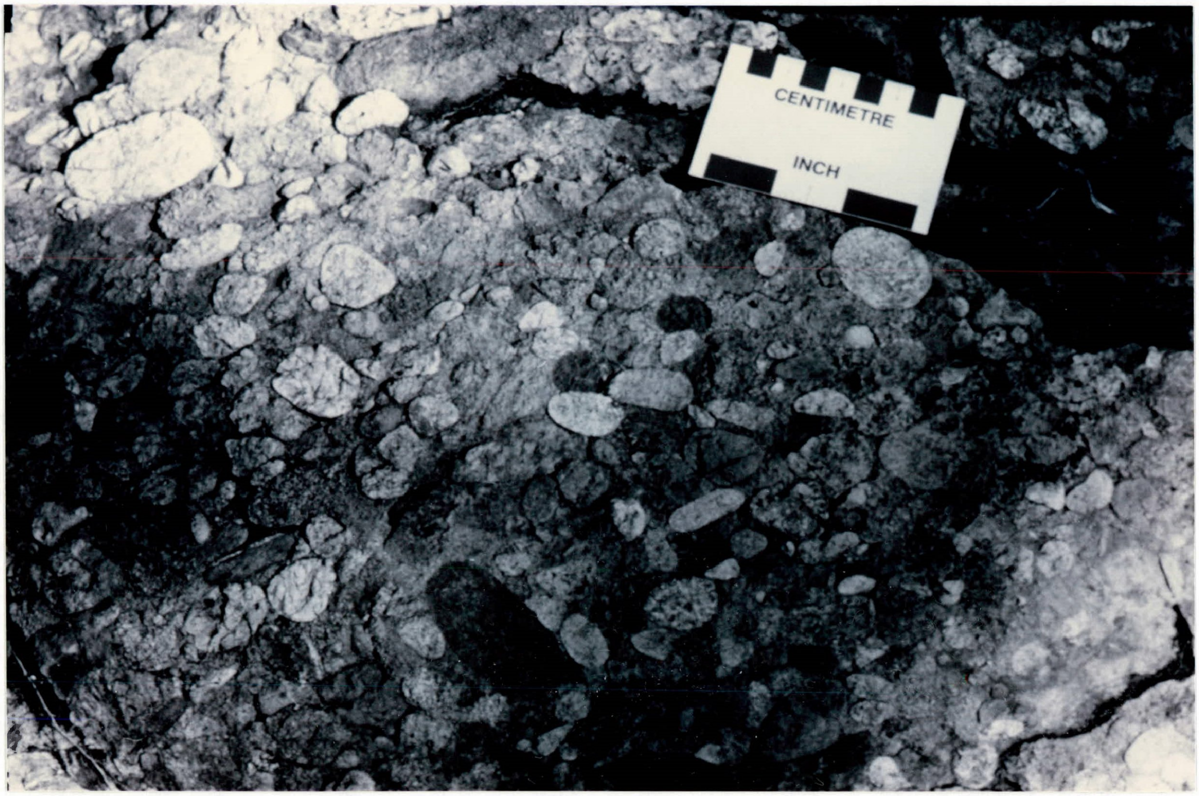
Sorting is frequently related to average clast size. Coarser beds or horizons of 32-64 mm pebbles, with or without cobbles, are poorly to moderately sorted, while the finer grained (4-32 mm) ones are moderately to well sorted and often cobble-free. Changes in average grain size and sorting reveal crude to well-developed stratification (beds,

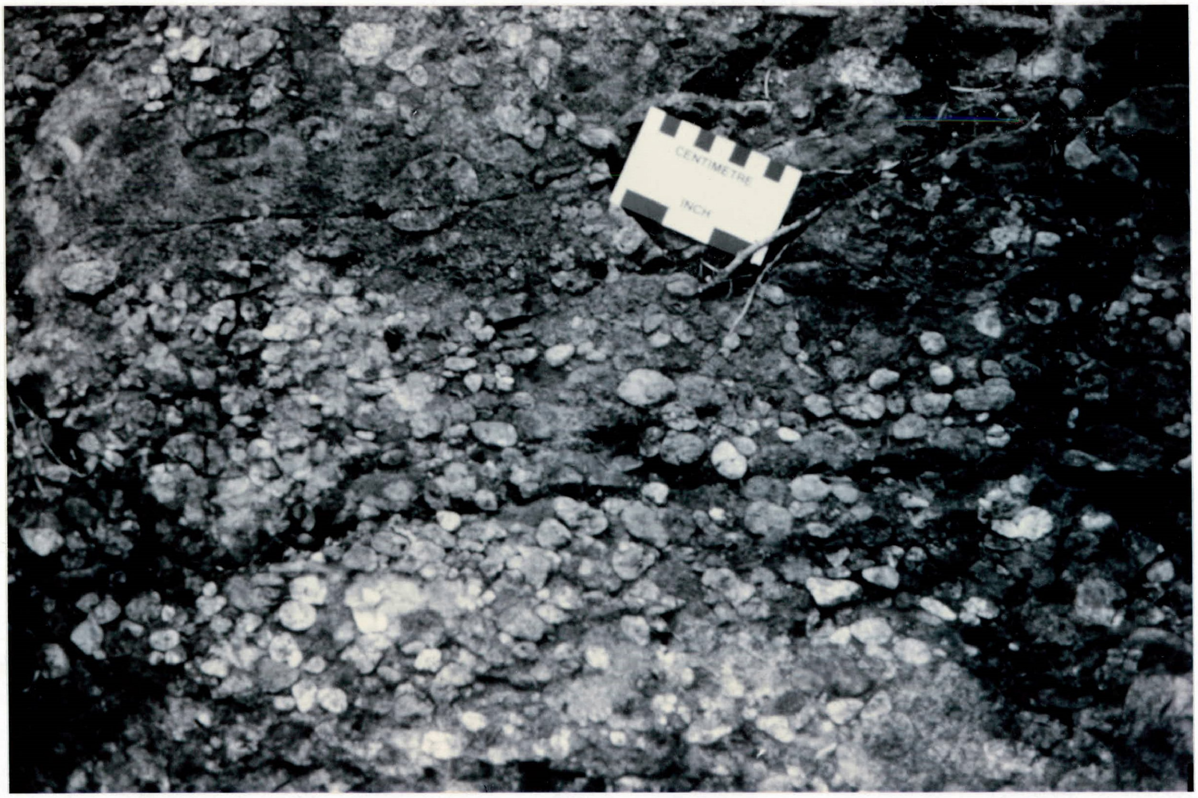
Fig. 6.6: Typical schistose. clast-supported polymict conglomerate with a sandstone matrix. Light coloured felsic volcanic clasts are abundant. Some clasts have been tectonically rotated. Outcrop #64. site of point count #21.


CENTIMETRE
INCH

Fig. 6.7: Crudely bedded, clast-supported polymict conglomerate. A vaguely defined horizon of finer pebbles (middle of photo) reveals the stratification, which has approximately the same strike as the schistosity here. Outcrop #64.


CENTIMETRE
INCH


CENTIMETRE
INCH


CENTIMETRE
INCH

Figs. 6.8, 6.9: Relatively fine-grained, clast-supported polymict conglomerate. Note the lack of schistosity. Compare the clast sizes and fabric with Figs. 2.3, 2.8, 6.6, 6.7, 7.6 and 9.1. Fig. 6.9 shows at its upper left a sandstone lens. Outcrop #13; Fig. 6.8 is the site of point count #4.

horizons) within the conglomerate.

In Walters Township (Fig. 3.1) the conglomerate locally has abundant finer (5-20 mm) pebbles and few cobbles. The D10 measurements (Table 6.2, Fig. 2.8) show the conglomerate to be finer here than in most of the assemblage to the east (in Leduc Township), and comparable to measurements from outcrops in Dorothea and Sandra Townships (Table 6.2). Outcrop #9 has one abnormal, 20 cm thick bed of very fine (2-6 mm) conglomerate with a moderately sorted, clast-supported framework.

Scours and associated features. In outcrop #13 fine conglomerate has scoured down into siltstone. The conglomerate is a 7 cm thick tapering bed, with a sharp south margin (erosive base) and a less sharp north margin (Fig. 6.10). The sandstone bed north of the conglomerate is identical to the conglomerate's matrix in texture, and contains siltstone clasts. These siltstone clasts, tiny granule chips to angular blocks 3 cm by 1 cm, are identical in composition to the siltstone south of the adjacent conglomerate bed.

A small scour (Fig. 6.11) about 10 cm deep is concave north and filled with fine (pebbles 2-20 mm; 2-10 mm ones most abundant) conglomerate.

Anomalous matrix. The sandy matrix of outcrop #53's conglomerate displays some horizontal bands rich in fine sand, in contrast to the usual coarse sand. These bands are probably cms thick, but it is difficult to identify "bed"

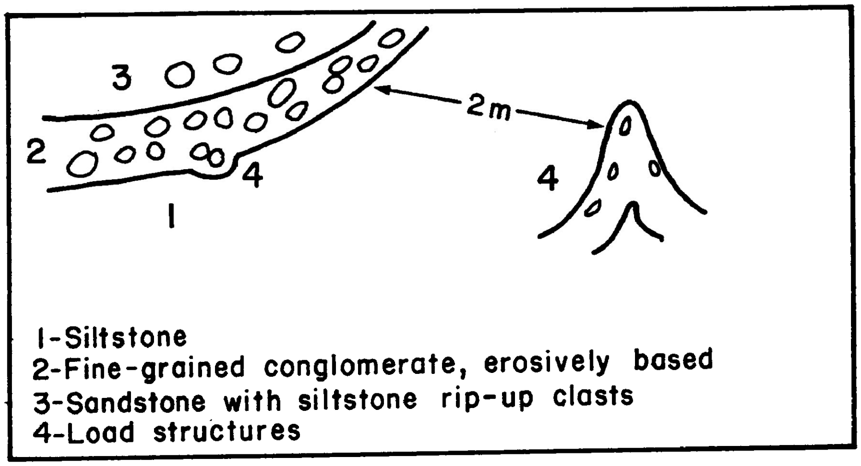
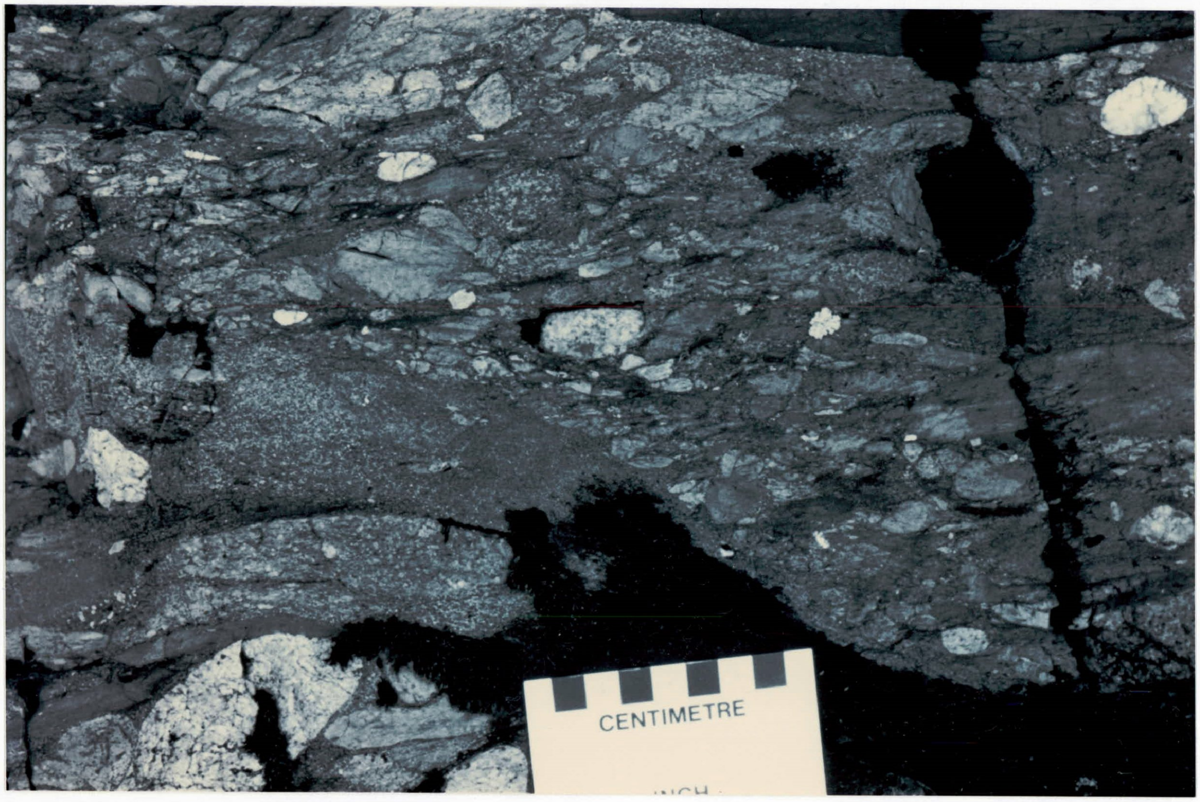

3
2
4
1
2m
4
1-Siltstone
2-Fine-grained conglomerate, erosively based
3-Sandstone with siltstone rip-up clasts
4-Load structures

Fig. 6.10: Scour bed (conglomerate), rip-up clasts and load structures. Sketch from field notes and photographs. Outcrop #13.


CENTIMETRE
INCH


CENTIMETRE
INCH

Fig. 6.11: Polymict conglomerate bed with a bed sole that scoured down into the underlying sandstone (at centre left). Scour is lined with fine conglomerate. Outcrop #65.

Fig. 6.12: Bottom: polymict conglomerate. Middle: massive sandstone. Top: planar, horizontal to low-angle, dominantly parallel laminated sandstone. Outcrop #69.

contacts within the matrix.

Some of the matrix's larger inter-pebble areas have an abnormally abundant concentration of granules. Triangular areas about 7 cm sq. have 10-15 granules, a poorly to very poorly sorted sandy matrix. One matrix-filling cluster of granules is 2 cm by 1 cm.

Within a bed in outcrop #64, a bed-parallel line separates sandy matrices of different character. South of the line the matrix is very coarse to fine-grained and poorly sorted, versus medium to fine-grained and moderately sorted matrix north of the line.

Fining-north beds. A 2.8 m thick bed in outcrop #70 displays crude coarse-tail grading, fining north. Measuring mostly the largest granitoid clasts, the southern 90 cm has the largest cobbles (60-140 mm, short axis), the middle 100 cm has fewer cobbles and smaller clasts (40-80 mm), and the northern 90 cm has small (15-35 mm) pebbles.

In the northern part of outcrop #31 two vaguely defined (crudely developed?) northward fining beds or units are present. The sequence from south to north is:

1) a conglomerate bed/unit 2.5 m thick, with few cobbles to the south, and fine-grained (mean deformed clast size 10-20 mm) to the north;
2) a tapering sandstone bed, up to 20 cm thick, with sharp bed contacts;
3) a 3.0 m thick conglomerate bed/unit, the coarser south part with abundant cobbles, including one

170/585 mm clast. This is an example of coarse tail grading; and

4) a 10 cm thick sandstone bed with sharp contacts, capped by conglomerate (edge of outcrop).

Stratification. Outcrop #66 contains a well-defined, 15 cm thick conglomerate bed that tapers laterally over a few metres to a pebble band, a vaguely defined unit separating sandstone beds. This tapering bed is part of a section (Fig. 6.3) rich in sandstone, pebble bands and thin conglomerate beds. Fig. 6.5 shows an example of the tapering of a bed associated with a paleoslope and lateral fining.

6.3 Sandstone Lithofacies

Sandstone beds interstratified with conglomerate are massive, plane laminated (Fig. 6.12), and planar cross-bedded (Figs. 6.13, 6.14). They are pebbly, poorly to moderately sorted, and coarse to fine-grained. The sandstone can be coarser (very coarse to coarse grains) where it is rich in pebbles.

Abnormally pebbly sandstone (e.g. outcrop #80) is transitional in texture to matrix-supported conglomerate, and is likely the result of superimposed, vaguely defined pebble bands.

Beds are up to 30 cm thick, and only rarely thicker, in the conglomerate-rich sections. Contacts are sharp. Outcrop #68 contains a sandstone bed with its north margin noticeably sharper than its south one, a stratification texture

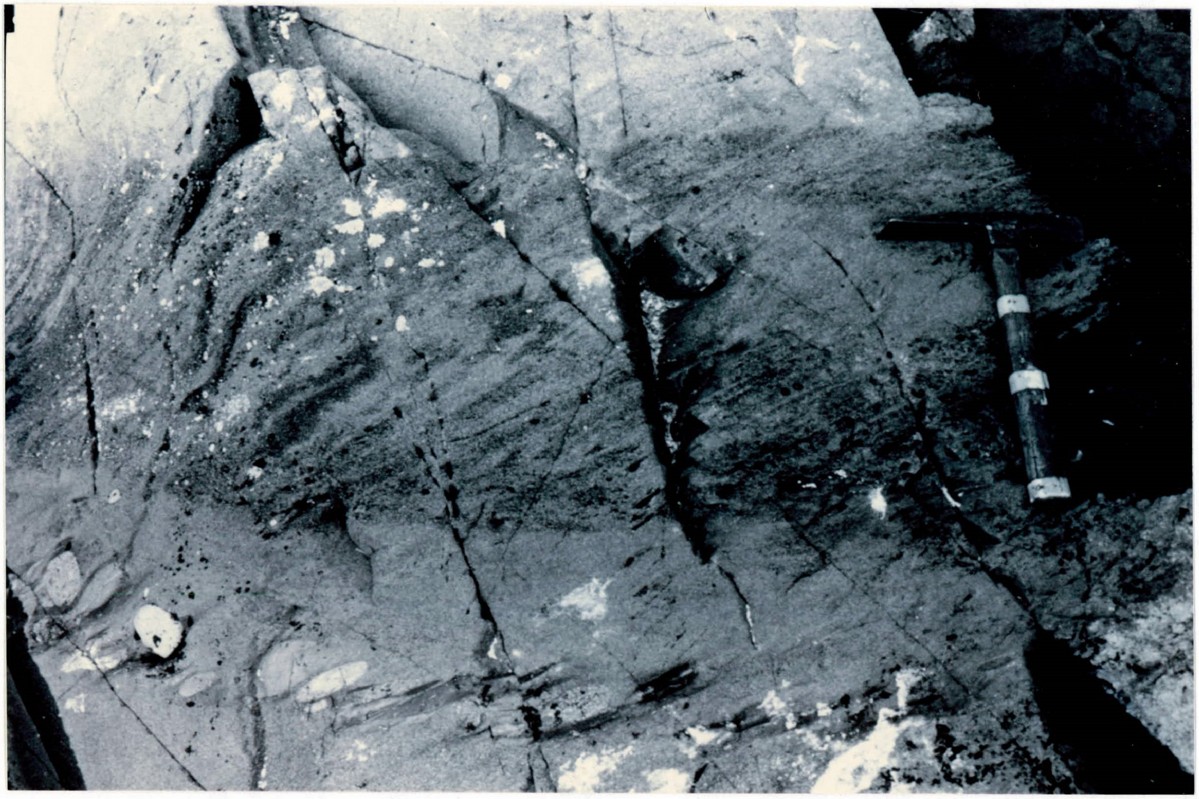


Fig. 6.13: Right: sandstone beds separated by a pebble band. Note the three-dimensional exposure of bedding (60° dip to left, to north) and the approximately vertical schistosity. Middle: planar cross-bed. Left: sandstone and polymict conglomerate (in bushes). Outcrop #67.

Fig. 6.14: Detail of Fig. 6.13. Bottom: pebble band. Note that the schistosity strikes north-east of the east-trending stratification, as is typical in the CMB. The pebble band is overlain by faintly plane laminated sandstone. Middle: planar tabular cross-bed. Outcrop #67.

previously discussed (Fig. 2.7).

Planar, horizontal and parallel laminae are defined by coarse-medium versus medium-fine-grained layers. The finer laminae may be darker. Some laminae are slightly (up to 10°) inclined.

Cross-lamination. Outcrop #42 contains a ripple in sandstone that has its foresets burying pebbles on the uneven top of a conglomerate bed (Fig. 6.15; the stratigraphic top is based on a residual lag bed 1.5 m away).

An unusual planar cross-bed. Part of this bed is shown in Fig. 6.16. Foreset lamination is best developed in the upper half of this set of cross-strata. The laminae are 5-10 mm thick, parallel and planar to slightly curved, being planar on a dm scale and concave north on a 0.5 m scale. The paleocurrent is to the southeast (after a simplistic single rotation to horizontal).

The cross-bed contains small pebbles, up to 1 cm in size. Fig. 6.16 shows a spot where a pebble has rolled down a foreset and has blocked coarse sand avalanching down the foreset slope. The coarse sand lamina thickens towards the pebble (to the southeast), as would be expected above an obstacle.

This bed has a north contact noticeably sharper than its south one.

A reactivation surface. A 35 cm thick planar cross-bed (Fig. 6.17) in outcrop #65 has well defined planar, inclined and parallel foreset laminae. Some toesets are tangential

Fig. 6.15: Middle: cross-laminae (at a high angle to conglomerate-sandstone bed contacts in rest of outcrop) produced by a ripple burying the large clast. Note the finer sandstone above the cross-laminae. Outcrop #42.

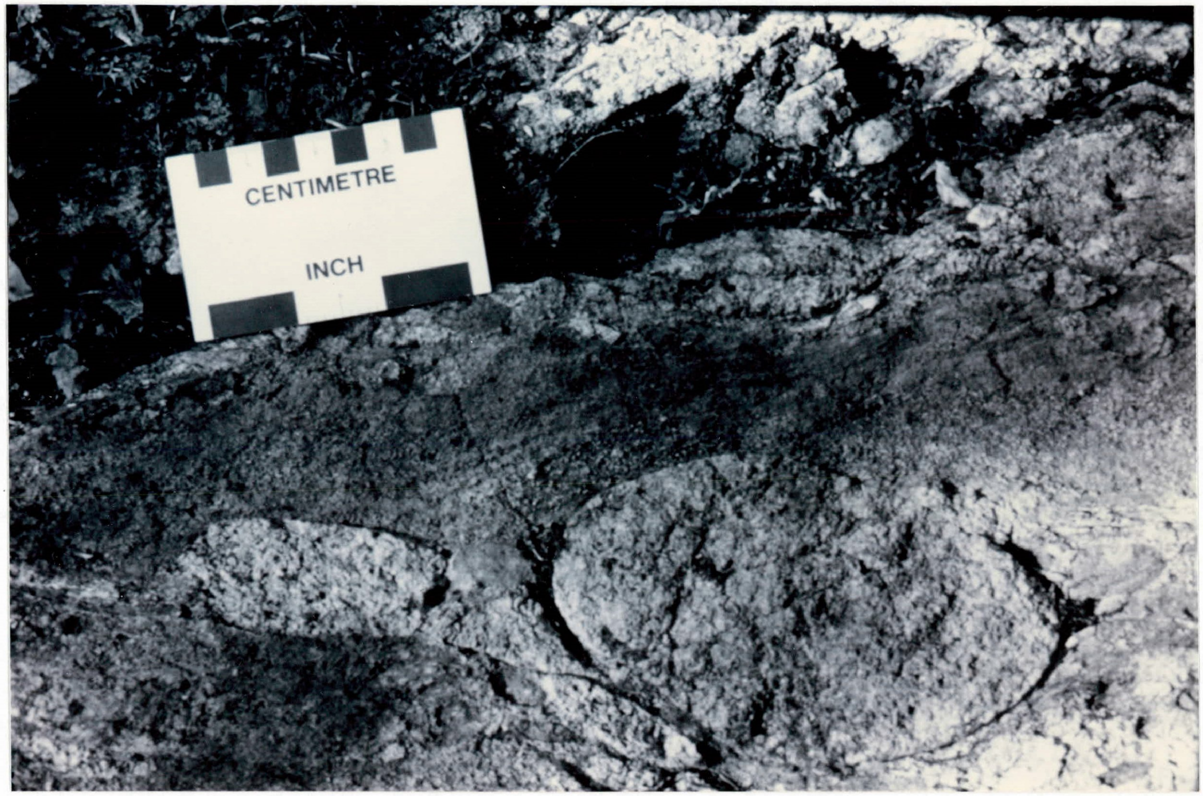

CENTIMETRE
INCH

Fig. 6.16: Planar cross-bedded sandstone. A pebble rolled down the foreset and later blocked coarser sand (which tapers to the upper left) avalanching down the foreset slope. Outcrop #53.

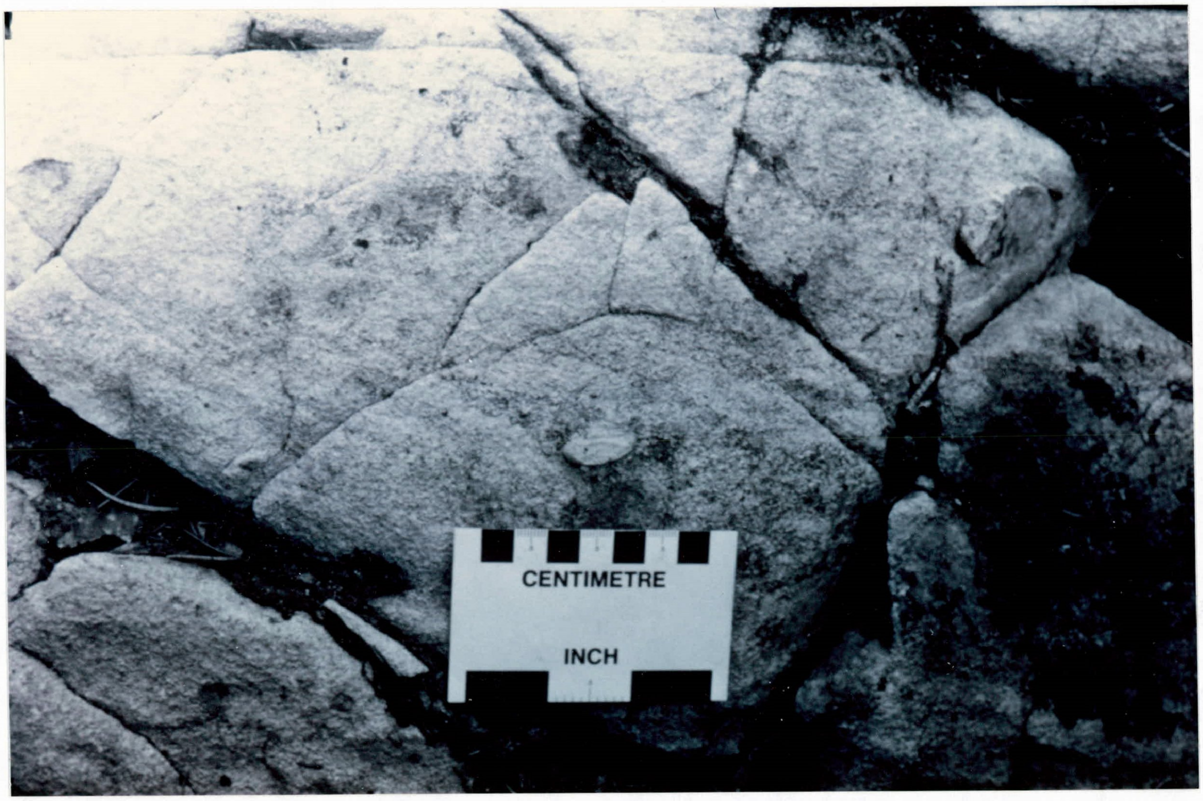
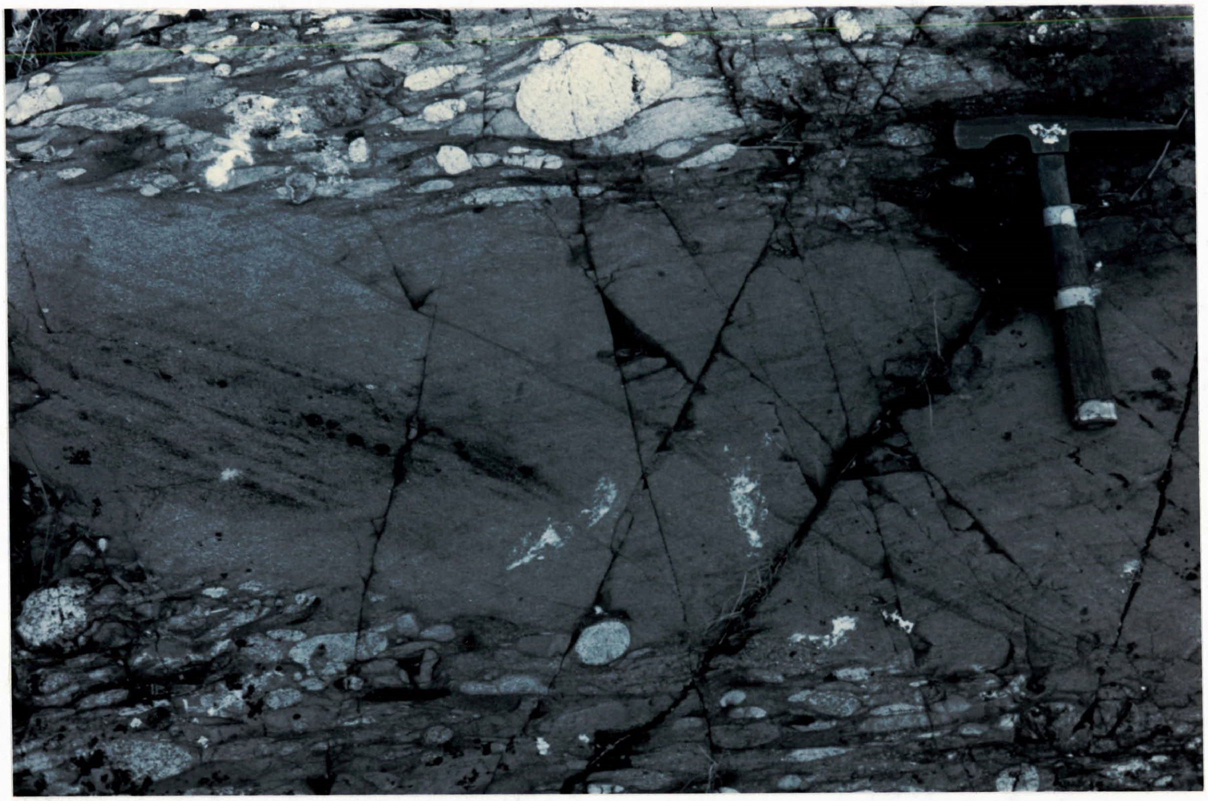
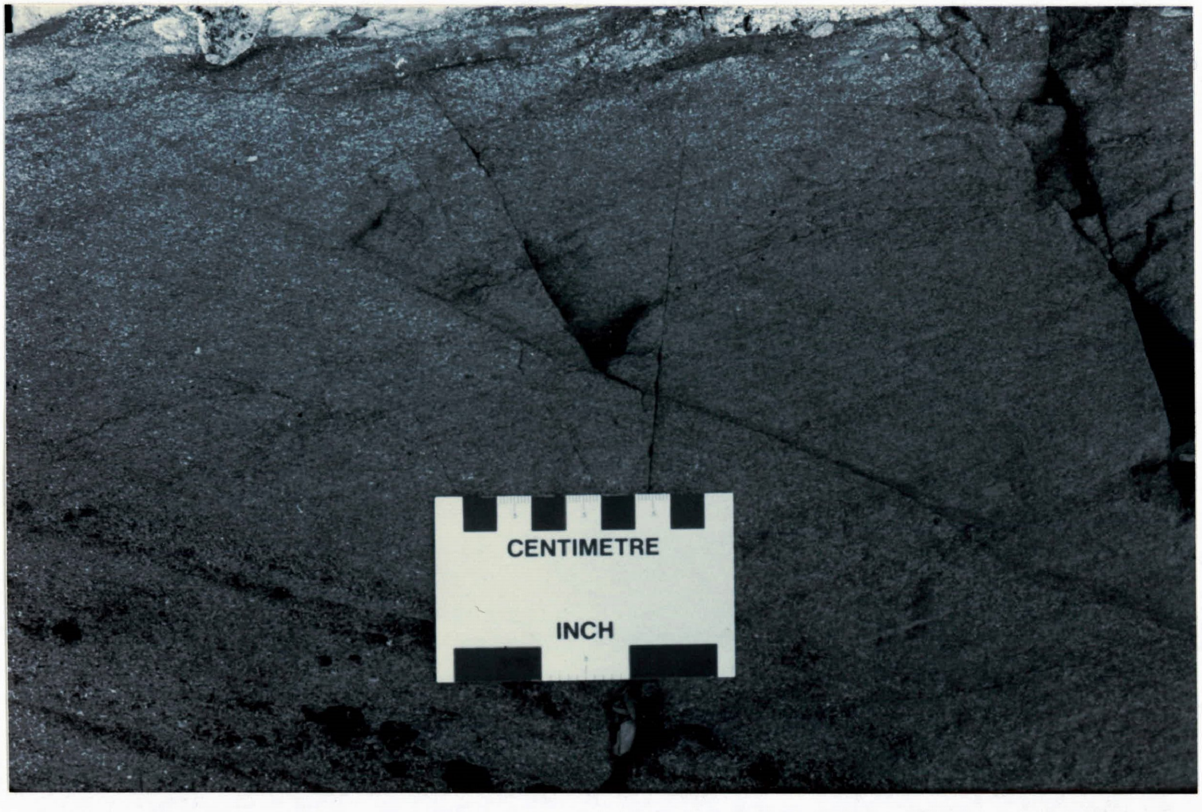

CENTIMETRE
INCH

Fig. 6.17: Planar cross-bedded sandstone between polymict conglomerate beds. Middle: darker, finer-grained lamina (upper left to lower right) is a reactivation surface. Note the toeset concavitv at the lower right. Outcrop #65.

Fig. 6.18: Detail of Fig. 6.17. The reactivation surface is the dark lamina (upper left to lower right). The reactivation surface and the younger (to the right) foresets are steeper than the older (to the left), lower angle, erosionally truncated foresets. Note the 1 cm deep concave depression in the reactivation surface at the lower right.





CENTIMETRE
INCH

and concave north.

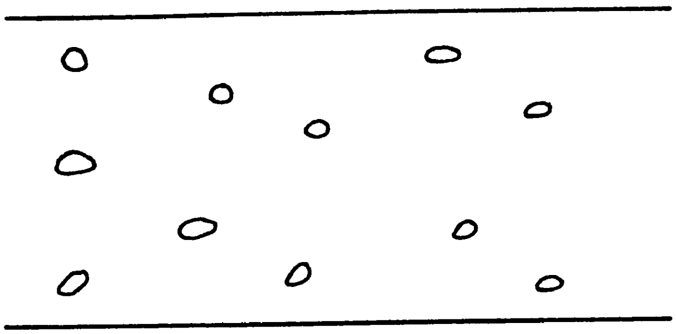
Figs. 6.17 and 6.18 show the reactivation surface (RS), a very small-scale angular unconformity within the cross-bed. The RS is defined by a darker, finer-grained lamina having a slight undulating topography of depressions up to 1 cm deep. To the left of the RS, foreset laminae are 1-5 mm thick, toesets are angular, and the sandstone is slightly coarser (richer in coarse sand) at the base (south margin) of the cross-bed, with a few granules along the coarser foresets. The RS truncates these earlier foreset laminae at a steeper angle (Fig. 6.18), and the later foresets (to the right in Figs. 6.17, 6.18) are parallel to the RS. These post-RS foreset laminae are 1-2 mm thick and are composed of slightly finer (pebble-free, less coarse sand) sandstone.

An unusual pebbly sandstone bed. Outcrop #42 contains a sharp-sided sandstone bed with a mid-bed, bed-parallel, laterally discontinuous pebble band, 1-2 pebbles (about 10 mm) thick. The northern part of the bed is pebble-free and has some vague, poorly continuous plane laminae, in contrast with the pebbly sandstone south of the pebble band (see Fig. 6.19).

6.4 Mudstone Lithofacies

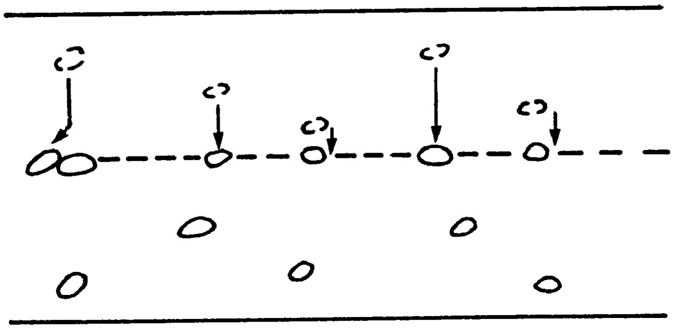
Mudstone. Three types of metamorphosed bluish grey to black mudstone are present: coarse siltstone, argillite, and slate. Mudstone beds interstratified with the conglomeratic sequences are usually only cms thick; such beds are present

(a)



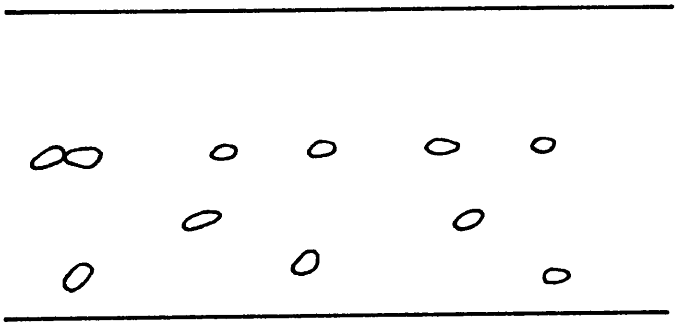
pebbly sand
bed deposited

(b)



bed partially eroded;
sand removed, pebbles
lagged behind

(c)



pebble-free sand
from upstream, deposited
on top of lag pebble
band

Fig. 6.19: Erosion and lag of the coarsest fraction (b), and subsequent burial by finer, better sorted sand (c) would give a way-up criterion. In outcrop #42, the inferred base of the two-bed unit is coarse to medium-grained, moderately to poorly sorted, pebbly sandstone, as opposed to medium to fine-grained, moderately to well sorted, non-pebbly sandstone north of (inferred to be above) the lag pebble band, as in (c).

in outcrops #13, 14, 15, 76 and 97.

Slaty mudstone at the base of Fig. 6.1a contains a few percent of very fine sand as coarser planar, horizontal and parallel laminae, from ultrafine to 1 mm thick.

Outcrop #56 contains a rare example of textural inversion. Pebbles are partially supported in a siltstone matrix in one bed only. Nearby pebbles probably "contaminated" a silt bed, an overlap of local facies.

<u>The northern mudstone unit</u>. This unit or horizon (Fig. 3.1) in the northern CMB has been outlined by Mackasey (1976).

Where not massive, the mudstone has some faint, very thin (less than 1 mm) planar, horizontal and parallel silty laminae. A linear valley, a large-scale example of recessive weathering, extends east-northeast from the northwest end of Leduc Lake (Fig. 3.1). The north and south contacts of this recessive unit were found within outcrop #31, giving an apparent (horizontal) thickness of about 120 m. Mackasey's (1976) map suggests that this mudstone horizon is thicker 2 km to the west (at outcrop #75).

<u>Mudstone clasts</u>. Good examples of mudstone clasts are exposed in outcrops #31 and 76. Because of the schistosity they are highly elongate and wispy. The clasts have a distinctive yellow weathered surface and are recessively weathered relative to the host sandstone. The mud chip clasts are 1-15 mm thick, and 2 to tens of cms long.

In outcrop #31 the mudstone clasts (mud chips) are at the contact between coarser sandstone and conglomerate to the south, and finer sandstone to the north, with the mud chips getting larger towards the conglomerate. Outcrop #76 contains a similar sequence of, from south to north, conglomerate, a 25 cm sandstone bed, and sandstone with mud chips.

In outcrop #31 the mud chips are immediately south of the area's northern mudstone unit.

Outcrop #122 contains mud chip clasts and fine to very fine-grained sandstone clasts, one laminated clast, and one vaguely defined mud clast band.

A mudstone clast conglomerate bed. Outcrop #20 contains a very unusual 90 cm thick bed of mudstone clasts 30-80 mm thick and up to 40 cm long (Fig. 6.20). This bed changes from partially matrix-supported in its southern part (base) to wholly clast-supported in its northern part.

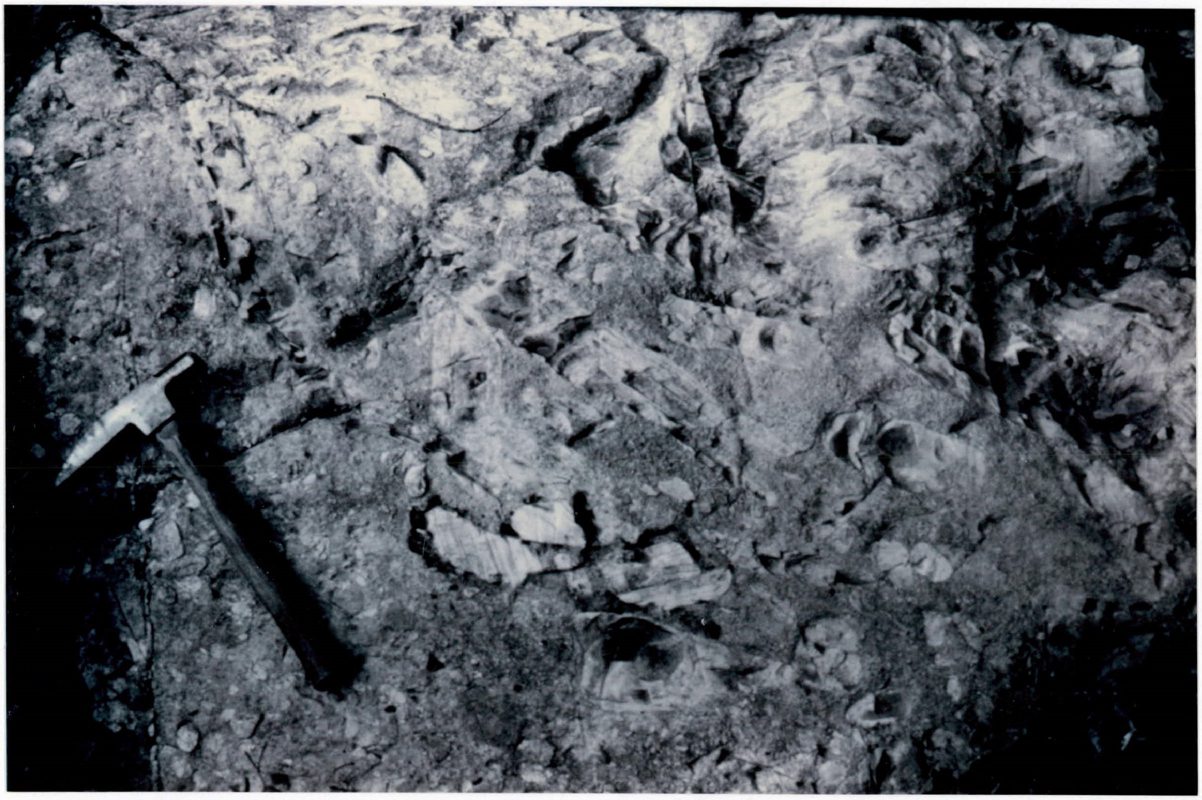
Most of the clasts are mudstone; small pebbles and abundant granules are present as part of the matrix to the large mud clasts. Some of these smaller pebbles appear to have indented the apparently once-soft mud clasts. The mud clasts sometimes contain fine or very fine sand. Many are plane laminated, with some graded laminae.

Nearby in the same outcrop, thin (up to 20 cm) conglomerate beds are rich in mud clasts.

The anomalous mud clast bed described above had:

1) a quasi-oligomict composition, versus adjacent beds of typical polymict pebble conglomerate;
2) an odd texture, both matrix and clast-supported;

Fig. 6.20: Intraformational (rip-up) mudstone clast conglomerate bed. Some of the rip-up clasts are laminated. Middle: partially matrix-supported framework. Upper right: clast-supported framework.



3) an extremely atypical composition for a coarse conglomerate bed in the study area; and

4) dented mudstone clasts.

Another example is a coarse to fine-grained sandstone bed in outcrop #56 that contains a 20 cm thick lens of mudstone clasts varying from clast to matrix-supported. The clasts are 4-80 mm and angular to subround. Some are laminated, with coarser laminae of fine to very fine sand. Note that the sand within the clasts is mostly finer than that of the host bed.

6.5 Units B and C

The conglomeratic assemblage has been subdivided into "Unit B" and "Unit C" in the Beatty Lake area (Fig. 3.1; Devaney and Fralick, 1985, Fig. 16.2). This terminology will be followed here for consistency and to later emphasize the vertical stratigraphic trends through the CMB in this area.

Unit B is distinguished by its:

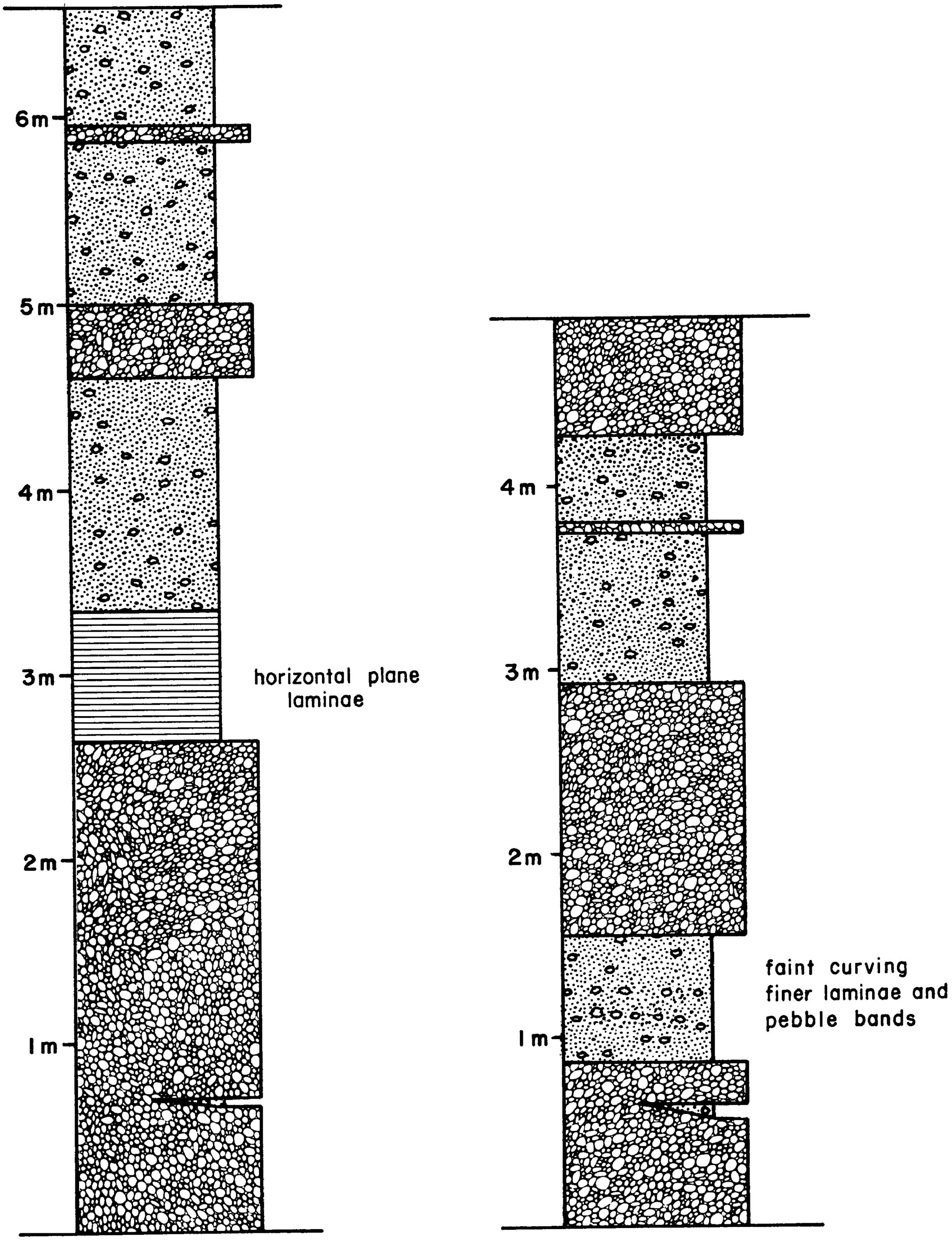
1) roughly equal amounts of conglomerate and sandstone in section (Table 6.3, Figs. 6.21, 6.22);

2) more abundant thin conglomerate beds in the sandstone-rich sections; and

3) rare mudstone: one bed, and numerous clasts.

The boundaries of Unit C were accurately outlined by Mackasey(1976). It is generally richer than Unit B in conglomerate, and lacks mudstone beds and clasts. The northernmost part of Unit C is consistently coarse and contains the

Fig. 6.21: Sections (a, b) of small islands, part of outcrop #83. Top to north. Legend in Fig. 5.2.


6m
5m
4m
3m
2m
1m
horizontal plane laminae
4m
3m
2m
1m
faint curving finer laminae and pebble bands

(a)

(b)

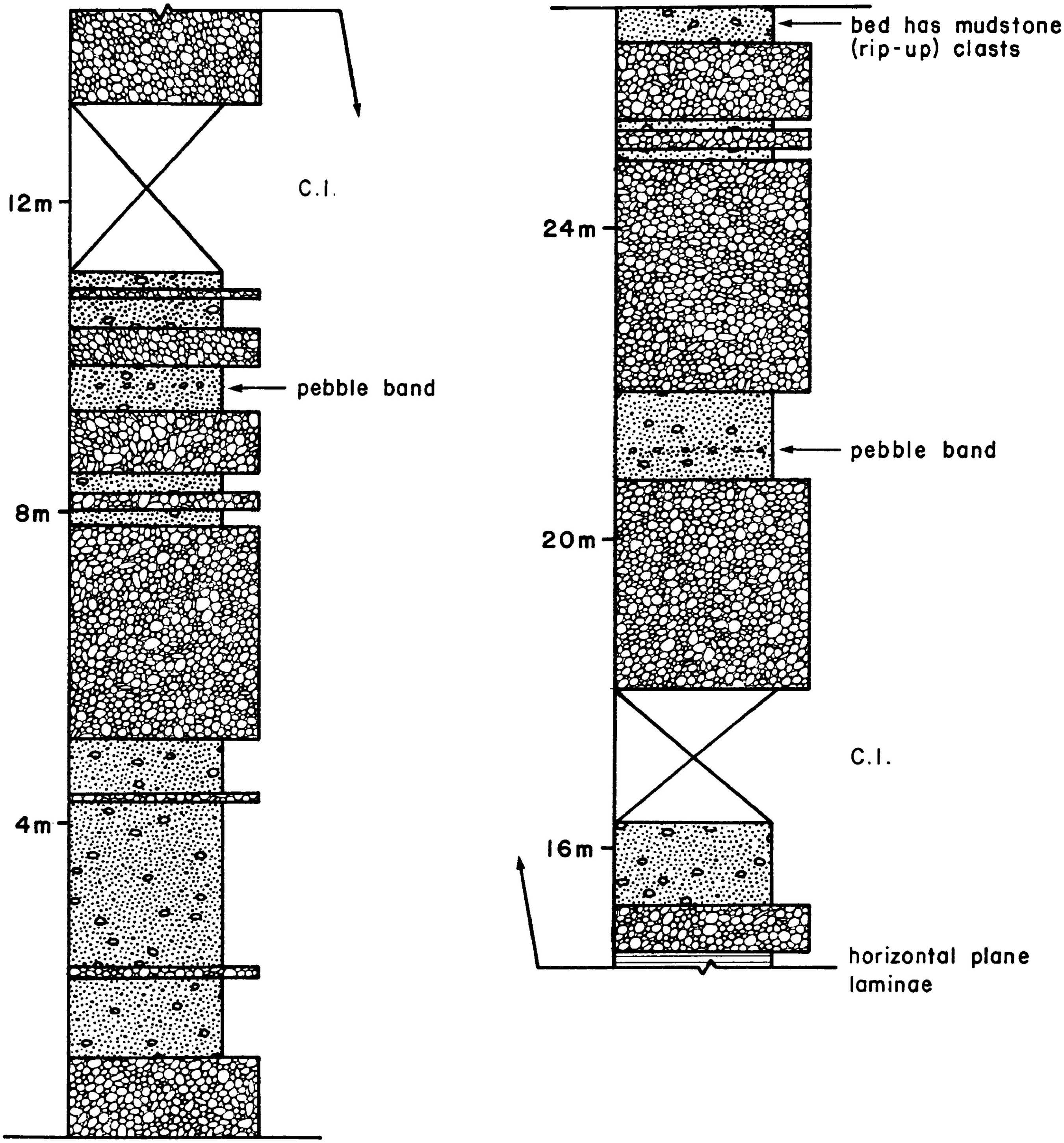

12m
C.I.
pebble band
8m
4m
bed has mudstone (rip-up) clasts
24m
pebble band
20m
C.I.
16m
horizontal plane laminae

Fig. 6.22: Section from east end of large island, outcrop #97. Top to north. Legend in Fig. 5.2.

coarsest exposure in the CMB (Tables 3.2, 6.3; Fig. 2.8).

Conglomerate lithofacies of Unit B. Thin conglomerate beds 8-25 cm thick (Fig. 6.23) are usually cobble-poor and gradational to well-developed pebble bands.

Beds up to 40 cm thick can be seen to taper out laterally. Usually the outcrops are too small to see such lensing, a small-scale lateral facies change to sandstone.

TABLE 6.3

Outcrop #	D10 (in mm)	Conglomerate : sandstone	Assemblage
	S L		
109	281/638		Unit C, bouldery conglomerate
70B	141/204		Unit C, (estim. 90-100% conglomerate)
70	130/195		
71	136/226		
67	94/161	48:52 of 8.5 m	Unit C, (parts are conglomerate-rich: > 90%)
65		87:13 of 6.6 m	
66		21:79 of 4.9 m	
64	85/116		
80	74/110	55:45 of 6.0 m	
81	144/217		
97	111/176	61:39 of 22.9 m	Unit B
83	141/222	58:42 of 4.8 m	
83		47:53 of 6.6 m	
86	120/172		
88	76/105		Heterogeneous Assemblage
105	68/ 97		
98	106/145		
103	90/142		
96	85/157		
104	65/ 87		
94	90/145		

Table 6.3: Quantitative criteria from the Beatty Lake area. Refer to Fig. 3.1, Table 6.2 and Section 11.12.

Fig. 6.23: Thin polymict conglomerate bed in a sandstone-rich outcrop section. Outcrop #84.

There is a possible example of cross-bedded conglomerate in outcrop #83. Within a 95 cm thick bed, crude stratification defined by finer (rich in 2-15 mm pebbles, and better sorted) and coarser (rich in 60 mm pebbles) layers is at a high angle, about 30°, to the local conglomerate-sandstone bed contacts. The two or three finer "foresets" are 0.5 m apart.

<u>Pebble bands</u>. These tend to be cobble-free and composed of smaller clasts, showing them not to be tectonically stripped-off portions of conglomerate beds (i.e. preferentially finer clast size is a primary feature). Pebble bands (syn. rows, stringers, lags) are most often bed-parallel (horizontal), and less often inclined up to 10° from the paleohorizontal. They reveal the multistoried nature of sandstone units metres thick (Figs. 6.3, 6.4, 6.21, 6.22). Vaguely defined, laterally discontinuous pebble bands (Fig. 6.24) can, when stacked closely in sequence, resemble a matrix-supported conglomerate.

The sand grains along pebble bands are usually quite coarse. Where clasts are several to tens of pebble lengths apart, this coarser sand often helps define such laterally discontinuous pebble bands situated between sandstone beds.

Thicker pebble bands, three or four pebbles thick, have better lateral continuity than vaguely defined bands usually one clast thick, the former being gradational to thin conglomerate beds.

<u>Sandstone lithofacies of Unit B</u>. Most of Unit B's

Fig. 6.24: Right: two polymict conglomerate beds separated by a sandstone lens (2 m wide, 7 cm thick). Left: three-dimensional exposure of pebble bands in sandstone, and vertical schistosity. Outcrop #83.

sandstone is moderately to poorly sorted, coarse or medium to fine-grained and pebbly, with rare fine grained, better sorted beds. Note that most of Unit B's sandstone is slightly coarser than the south part of Unit A and of the same grain size as the north part of Unit A.

The coarser (coarse to medium-grained) sandstone beds tend to be massive and pebbly, versus the finer (fine-grained), often plane laminated and pebble-free beds.

Examples of east-striking sharp bed contacts separating coarser, more poorly sorted beds to the north from finer beds adjacent to the south are common (Figs. 6.25, 6.26). This feature is present, but much less common, in Unit C, a consequence of Unit B's more abundant sandstone-rich sections. The coarser beds are richer in coarse sand and pebbles. Occasionally the northern "sandstone" is the sandy matrix of conglomerate bed margin, as in Fig. 5.23.

A very unusual variety of plane laminated sandstone is present in outcrop #82, with an odd regular spacing of alternating coarse, thicker laminae (5 mm thick, of medium-fine sand) and finer, thinner laminae (1 mm thick, of fine sand). Lateral continuity of the laminae is poor, only cms to tens of cms. It is not known if this is a foliation texture.

Mudstone lithofacies of Unit B. Fig. 6.27 shows a laminated siltstone bed located near the top of the section in Fig. 6.22. As siltstone was seen only in outcrop #97, such muddy beds are very rare components of Unit B's


CENTIMETRE
INCH

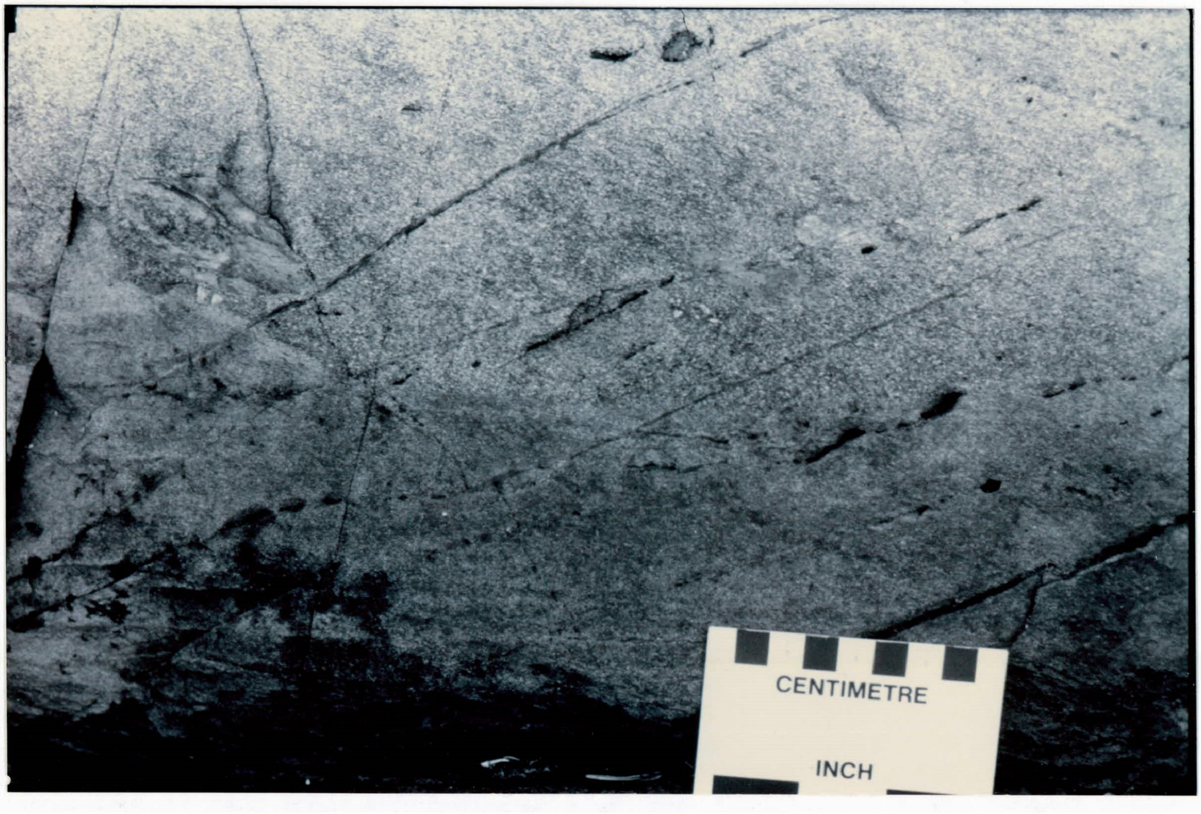

CENTIMETRE
INCH

Fig. 6.25: Top: sharp-based coarse to medium-grained (some very coarse sand grains and pebbles) sandstone with an intraformational mudstone clast overlies medium to fine-grained sandstone. Outcrop #84.

Fig. 6.26: Top: sharp-based coarse to fine-grained, pebbly sandstone in contact with fine-grained, faintly plane laminated sandstone. Outcrop #97.

Fig. 6.27: Plane laminated siltstone. Outcrop #97.


CENTIMETRE
INCH

conglomerate-sandstone assemblage.

Very small (cms thick, tens of cms long), laterally impersistent siltstone or very fine-grained sandstone units may be large rip-up clasts flattened along the schistosity, or tectonically interleaved bed "slices".

Mudstone clasts. These are usually brown to yellow and recessively weathered. They occur as isolated clasts within sandstone beds (Fig. 6.25), and rarely form bands. Short : long axis ratios are about 1:10, the most elongate of all the clast lithologies, as befits the softest clast lithology, that of unlithified mud during sedimentation, although some clasts may have been shale.

6.6 Anomalous Outcrop-scale Sequences

Small-scale fining-north sequences. These are very rare in the CMB and were seen only in outcrop #15. Three examples are given below.

Based on top directions from the loading of beds metres away, conglomerate is overlain by 20 cm of coarse pebbly sandstone, which is overlain by 50 cm of laminated sandy siltstone, capped by conglomerate. A few metres above, a similar sequence of sandstone and 35 cm of laminated sandstone/siltstone is sharply overlain by loaded conglomerate. Bed contacts are sharp.

At the inferred top of a sandstone bed is a single grain lamina of fine sand, abruptly overlain (to the north) by 1 cm of mudstone, in turn sharply overlain by conglomerate.

Sandstone-conglomerate transition. There is an apparent sedimentological transition from pebbly sandstone to conglomerate, defined by changes in maximum clast size, in outcrop #31. Over 10 m, sandstone with small (about 5 mm) pebbles changes northward (at 4 m) to sandstone with larger (about 50 mm) pebbles and, at 10 m, to clast-supported conglomerate.

6.7 Soft-sediment Deformation Features

Outcrop #15 contains an example of soft-sediment diapirism. Sand and finer, plane laminated silty very fine-grained sand intruded the overlying gravel (Fig. 6.28). It is odd that the laminae do not show any drag at the edge of this load structure, but the mushroom-shaped form of the conglomerate's south contact does not suggest any interpretation other than metre-scale loading. Soft-sediment injection of thin dikes and sills of coarser sand into finer sand (Fig. 6.29) was approximately synchronous with the loading.

The northward and upward injection of these dikes corroborates the tops to the north in this small outcrop shown by the loading and fining-north sequences.

In outcrop #13 scouring gives tops to the north. About 2 m from the scour, laminated siltstone and a band of pebbles have been drawn upward into a cusp about 25 cm high, forming a large "flame" structure (Fig. 6.10). The elongate pebbles are oriented perpendicular to the local bedding and are


CENTIMETRE
INCH

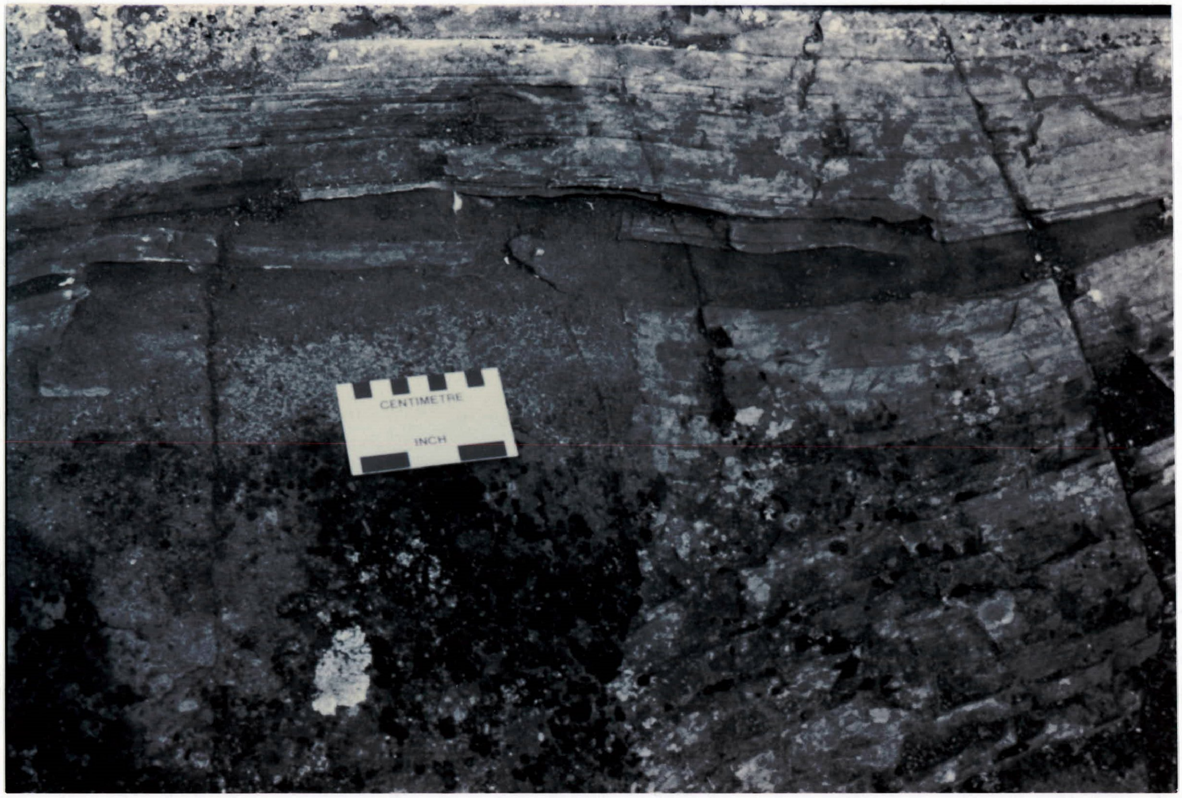

CENTIMETRE
INCH

Fig. 6.28: Middle and right: laminated sandstone. Left and top: clast-supported polymict conglomerate, partly obscured by lichen. An S-shaped contact separates the conglomerate (left) from the sandstone (middle). Outcrop #15.

Fig. 6.29: Soft sediment dike and sills. Darker, coarse to medium-grained sand intruded lighter and finer silty sand.

parallel to the cuspate, loaded laminae. No schistosity is visible in the outcrop.

6.8 Iron Formation Lithofacies

Iron formation laminae ("BIF") interlayered with siltstone laminae are present in outcrops #10 and 12. These are historic exposures (Coleman and Moore, 1907, 1908). The laminae are parallel, horizontal and planar, and of varying thickness, some thinner than 0.2 mm. Dark non-magnetic, lens-shaped laminae may be starved ripples, but this is difficult to determine for these tectonized (kinked) rocks. Some horizons cms thick are iron formation-dominant.

Both magnetite and hematite (red, non-magnetic) iron formation are present. Cherty-looking mudstone horizons and magnetite bands are both up to 2 cm thick.

In outcrop #12 these fine-grained clastic and chemical sediments are exposed in contact with the usual conglomerate and sandstone lithofacies.

Outcrop #79 is at the contact between the CMB and the mafic volcanic belt to the north. Within a 40 m interval conglomerate fines northward to typical sandstone (at 12 m), to mudstone (at 24 m), to iron formation interbedded with siltstone (at 39 m). Mackasey (1976) did not record iron formation or mudstone for this site. The sediment-mafic volcanic contact is tens of metres north in a depression covered by vegetation.

The iron formation is dark and only partly magnetic, possibly a mix of clay and magnetite. Laminated iron-rich bands are up to 4 cm thick, with some 1 mm thick planar, horizontal and parallel laminae. There is an odd, subequal spacing, 8-16 cm apart, of the thick (4 cm) bands.

6.9 Anomalous Lithofacies

A unique exposure. At the west end of outcrop #6 is a highly unusual exposure of conglomerate and pebbly sandstone. The conglomerate has a clast-supported, moderately sorted framework of very fine (2-16 mm) pebbles. Crude stratification is defined by coarser and finer layers. Thin conglomerate beds are similar to pebble bands, but bed contacts are difficult to identify because of vague gradations to pebbly sandstone, and heavy lichen growth. The tiny clasts are difficult to identify because of this heavy lichen cover. Although polymict, this lithofacies does not look at all similar to the usual conglomerate lithofacies seen elsewhere throughout the CMB and NMB.

An anomalous lithofacies assocation. Outcrop #43, a small shoreline exposure, is an unusual example of the usual conglomerate and sandstone in close spatial association with mudstone and graded sandstone.

Some sandstone beds are graded, with notably coarser south portions (bases) and faint laminae. All of the graded beds fine northward. Table 6.4 gives data for these graded beds. The basal 5 mm of the fourth bed listed in Table 6.4 have microlaminae (less than 1 mm thick).

One 70 mm thick mudstone bed contains microlaminae of very fine sand that may be the coarse-tail bases of graded laminae.

TABLE 6.4

Bed thickness (mm)	Coarser base	Finer top
8	M	VF
20	M	F
20	F	VF
20	VF	mud
65	M-F	F

Table 6.4: Graded bed characteristics, outcrop #43. M = medium-grained sand, F = fine-grained sand, VF = very fine-grained sand.

One bed showing pebbles mixed with mudstone, an unusual situation, is likely the result of tectonic interleaving, via slippage along the foliation. Gravel and mud fractions are mixed in one horizon only, at a site where bedding and foliation are collinear, in contrast to the rest of the outcrop in which conglomerate, sandstone and mudstone are preserved as distinct beds with the strike of bedding at a high angle to that of the foliation.

Outcrop #43 is located at the south contact of the northern mudstone unit (see Section 6.4), and may represent a facies change with small-scale tectonic modification.

CHAPTER 7: THE CMB's HETEROGENEOUS ASSEMBLAGE

7.1 Introduction

A heterogeneous assemblage of conglomerate, sandstone and mudstone at east-central Beatty Lake is known to extend for at least 1.2 km along strike (Fig. 3.1). Fig. 3.1 gives the locations of specific outcrops referred to in the text. For local details such as bedding and schistosity measurements, the reader should examine Mackasey's (1976) map.

Beds of all three of the above lithologies are:

1) both graded and ungraded;
2) cms to metres thick;
3) interbedded in very small (metres) outcrops; and
4) interbedded in sections as thin as tens of cms.

The variety of lithofacies present, the abundant graded beds, and the conglomerate's smaller D10 measurements (Tables 6.2, 6.3) make this assemblage unique.

The composition of the conglomerate is identical to the coarse conglomerate in the rest of the CMB (see Chapter 10).

7.2 Conglomerate Lithofacies

Beds of clast-supported polymict conglomerate have massive to crudely bedded, very poorly to moderately sorted frameworks. Most are ungraded, but some excellent examples of graded conglomerate are separately described below. Rare matrix-supported beds are also present.

Mean deformed clast size is in the 16-64 mm range. Table 6.3 and Fig. 2.8 give the D10 measurements for the assemblage's outcrops. Fine-grained (clasts up to 20 mm) beds have moderately sorted frameworks. Sandy matrices are coarse or medium to fine-grained.

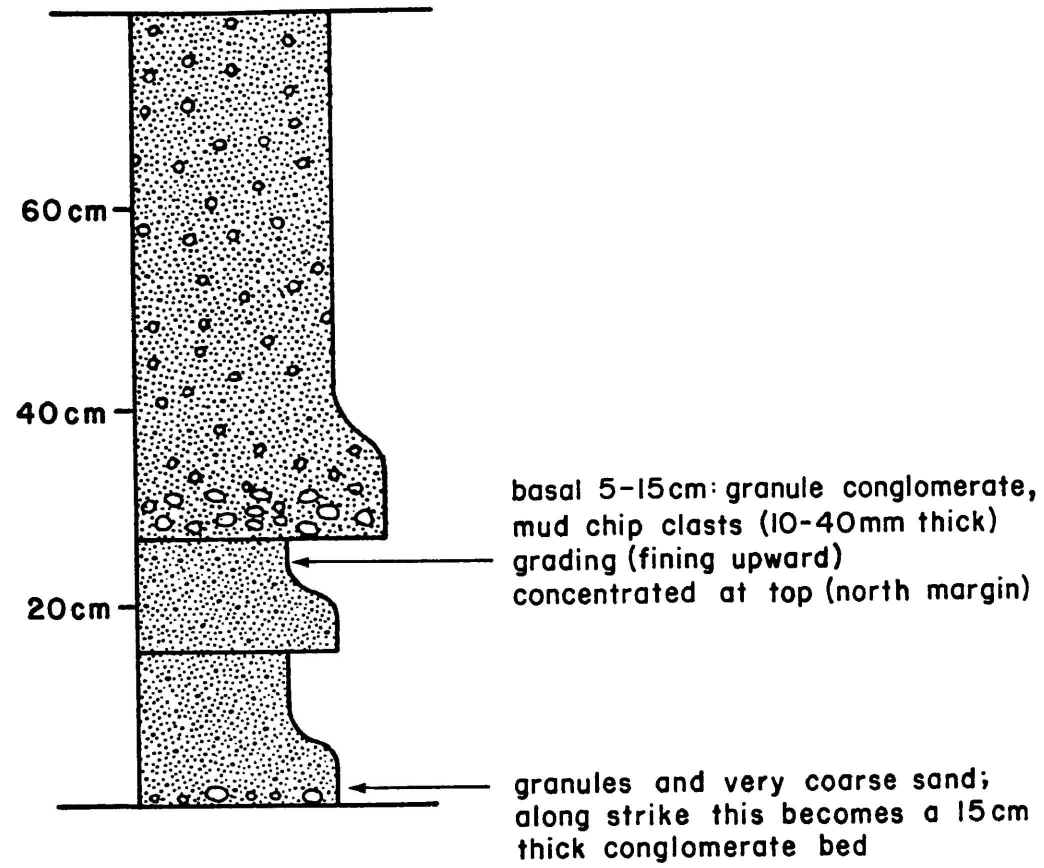
Beds are cms to metres thick, the thinnest ones being transitional to laterally discontinuous (vaguely defined) pebble bands and irregular lenses. Contacts with sandstone and mudstone beds are sharp. Horizons of finer (2-25 mm) pebbles suggest poorly defined beds within thicker units of amalgamated (vertically stacked) beds.

Graded conglomerate. The best examples of fining-north beds could also be described as pebbly sandstones with coarse-tail (see Reineck and Singh, 1980, Fig. 197b) grading developed near the beds' bases (Figs. 7.1 to 7.3). Fig. 7.2 shows one completely clast-supported, vaguely graded bed.

Two or three beds are probably inverse graded. Fig. 7.4 shows the best coarsening-north example, and Fig. 7.5 illustrates the results of a grading test done on a more subtle candidate.

Graded-stratified conglomerate. Figs. 7.6 and 7.7 show the one bed present of this lithofacies. The bed's coarsest clast is at its south margin. The rest of the bed is rich in:

1) 50-60 mm clasts 50 cm from the north margin;
2) 10-40 mm clasts 20 cm from the north margin; and
3) small (up to 10 mm) clasts in the pebble bands


60cm
40cm
20cm
basal 5-15cm: granule conglomerate,
mud chip clasts (10-40mm thick)
grading (fining upward)
concentrated at top (north margin)
granules and very coarse sand;
along strike this becomes a 15cm
thick conglomerate bed

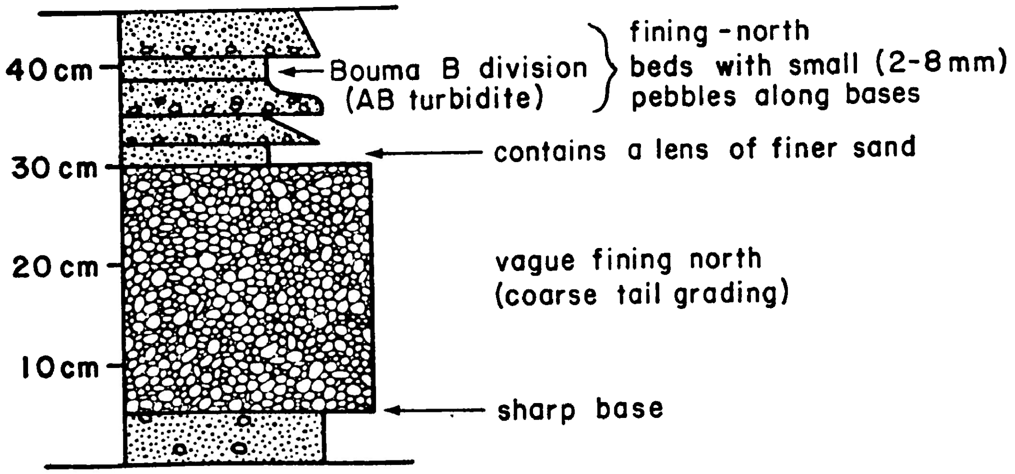
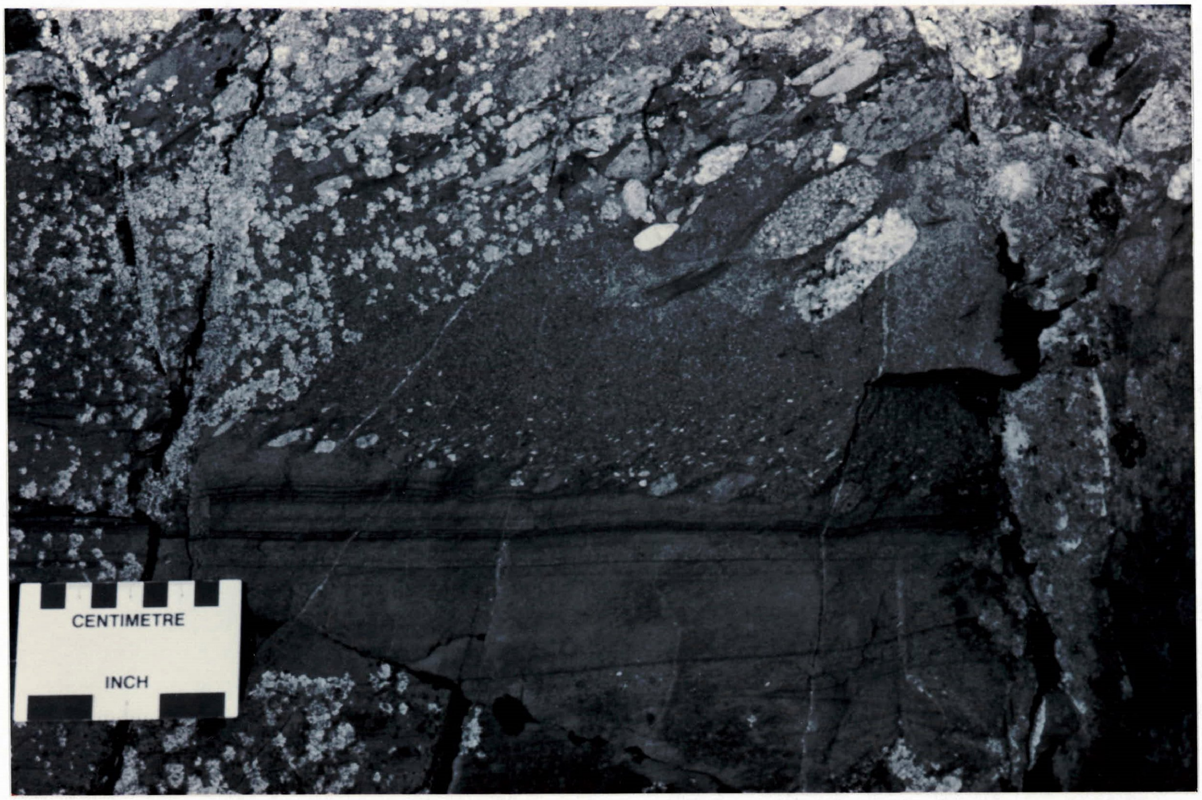

40cm
30cm
20cm
10cm
Bouma B division
(AB turbidite)
fining-north
beds with small (2-8mm)
pebbles along bases
contains a lens of finer sand
vague fining north
(coarse tail grading)
sharp base

Fig. 7.1: Section from outcrop #96. The graded beds fine northward (upward). Legend in Fig. 5.2.

Fig. 7.2: Section from outcrop #104. The graded beds fine northward (upward). Legend in Fig. 5.2.


CENTIMETRE
INCH

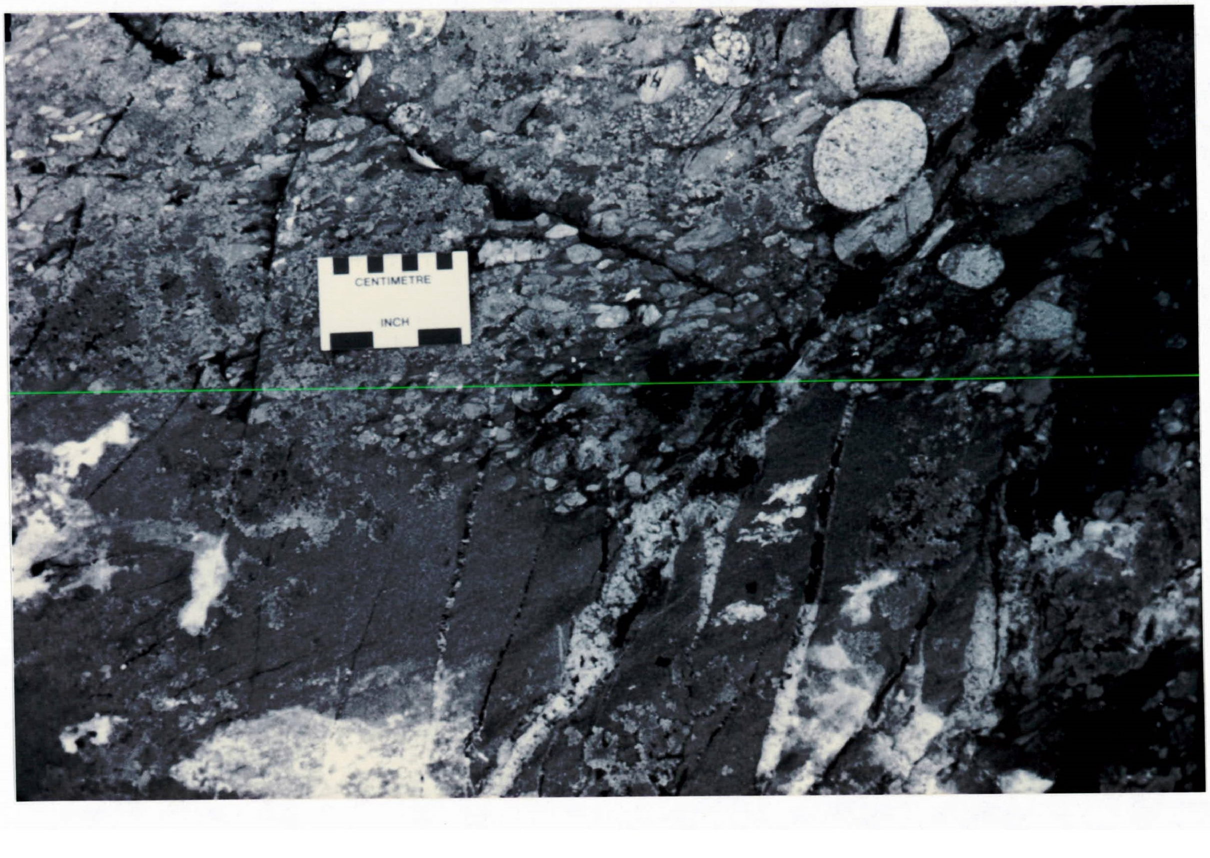
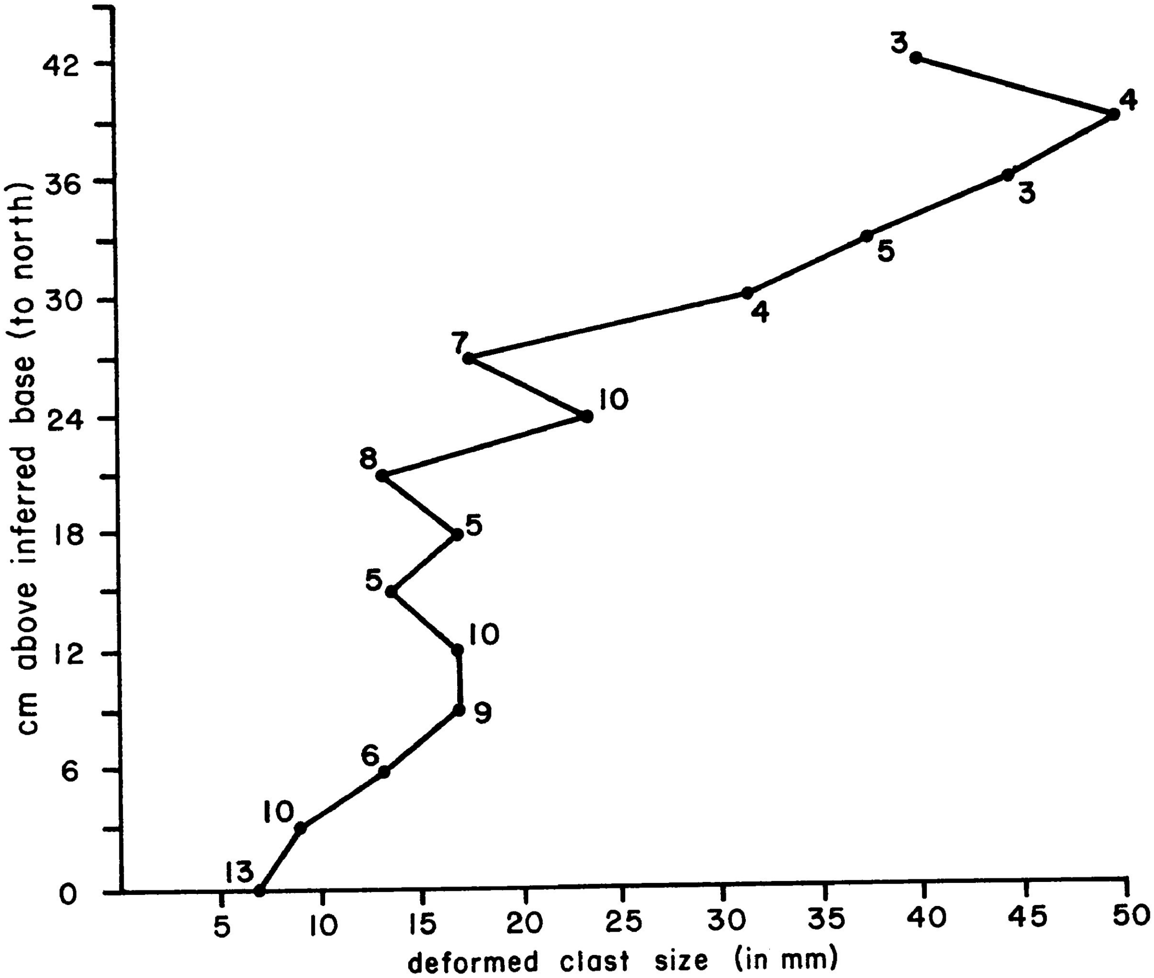

CENTIMETRE
INCH

Fig. 7.3: Middle: graded bed with: 1) erosive sole truncating underlying laminated sandstone; 2) conglomeratic base; and 3) sandstone top. Note the wispy mudstone clast at the base of the upper polymict conglomerate bed. The clasts show schistosity, not imbrication. Outcrop #98.

Fig. 7.4: Sharp-based, coarsening-north polymict conglomerate, probably an inverse graded bed. Bed sole is offset by a vein-filled fault. White marks are recent bird feces. Outcrop #98.



cm above inferred base (to north)
0
6
12
18
24
30
36
42
5
10
15
20
25
30
35
40
45
50
deformed clast size (in mm)
13
10
6
9
10
5
5
8
10
7
4
5
3
4
3

Fig. 7.5: Results of a grading test performed on a clast-supported conglomerate bed at the east tip of outcrop #98. A 130 cm x 42 cm grid, with 14 columns each 10 cm apart and 15 rows each 3 cm apart, was used. The data points given are average values, with the number beside each data point representing the number of non-matrix (clast) grid points in a row. The results show that the bed's south margin coarsens to the north.


CENTIMETRE
INCH


CENTIMETRE
INCH

Fig. 7.6: Graded-stratified polymict conglomerate bed. Top of bed shows the combined grading and stratification defined by horizontal pebble bands. Outcrop #98, site of point count #29.

Fig. 7.7: Detail of Fig. 7.6. Note the horizontal plane laminae in the sandstone bed.

forming the bed's gradational north margin (Figs. 7.6, 7.7).

Bands of small pebbles and thin sandstone beds alternate (Fig. 7.7), showing the combined grading and stratification. Distribution grading appears to be present, but the variable (lithology-dependent) deformation of the clasts makes this difficult to recognize, and the coarsest fraction (coarse tail) at a given level in the bed seems to illustrate best the grading to the north.

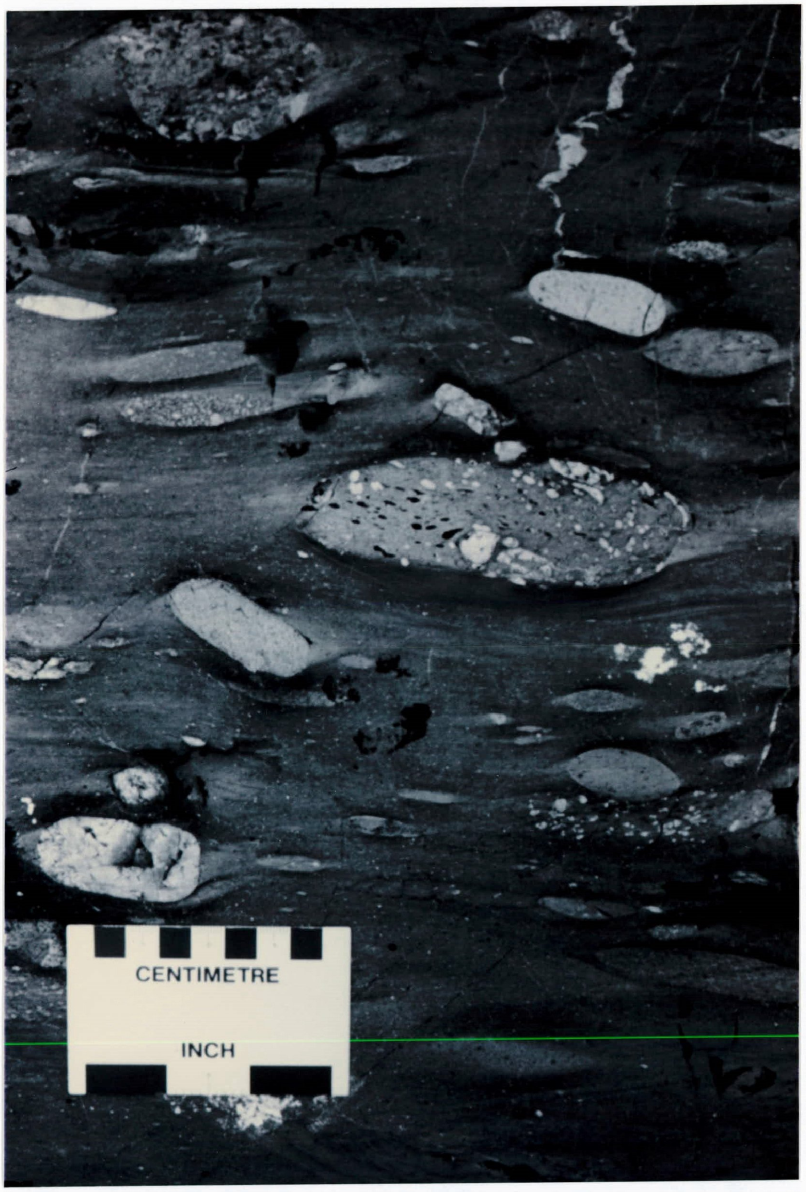
Muddy matrix-supported conglomerate. Three beds of this lithofacies, 3.7 m, 0.9 m and 0.2 m thick, are present in outcrop #98. Clasts are supported in an argillaceous matrix (Fig. 7.8) rich in very fine sand and dotted with a very low (1% ?) percentage of coarse sand. Parts of the matrix are laminated; laminated siltstone beds may be present within matrix-supported units (as opposed to one ingle matrix-supported bed), or may now be bed fragments streaked out along the foliation.

7.3 Sandstone Lithofacies

Sandstone beds and conglomerate matrices with a wide variety of grain sizes are present: pebbly coarse to fine-grained; medium to fine-grained; and fine to very fine-grained. The finer beds are better sorted and usually pebble-free. Graded beds are separately described below.

The coarser beds are most often massive. Planar, horizontal and parallel laminae are best developed in the

Fig. 7.8: Argillaceous matrix-supported polymict conglomerate. Matrix is sandy mudstone. Outcrop #98.


CENTIMETRE
INCH

finer beds, some as microlaminae (sub-mm thick). Very rare starved ripple laminae are defined by fine sand lenses 0.2-1.0 mm thick in a very fine-grained bed. Rare flasers of silt to fine sand, 1-2 mm thick and 3-4 cm wide, outline ripple troughs.

Beds are commonly up to 50 cm thick and have sharp planar contacts. Coarser sandstone may be in contact with the finer, siltier units.

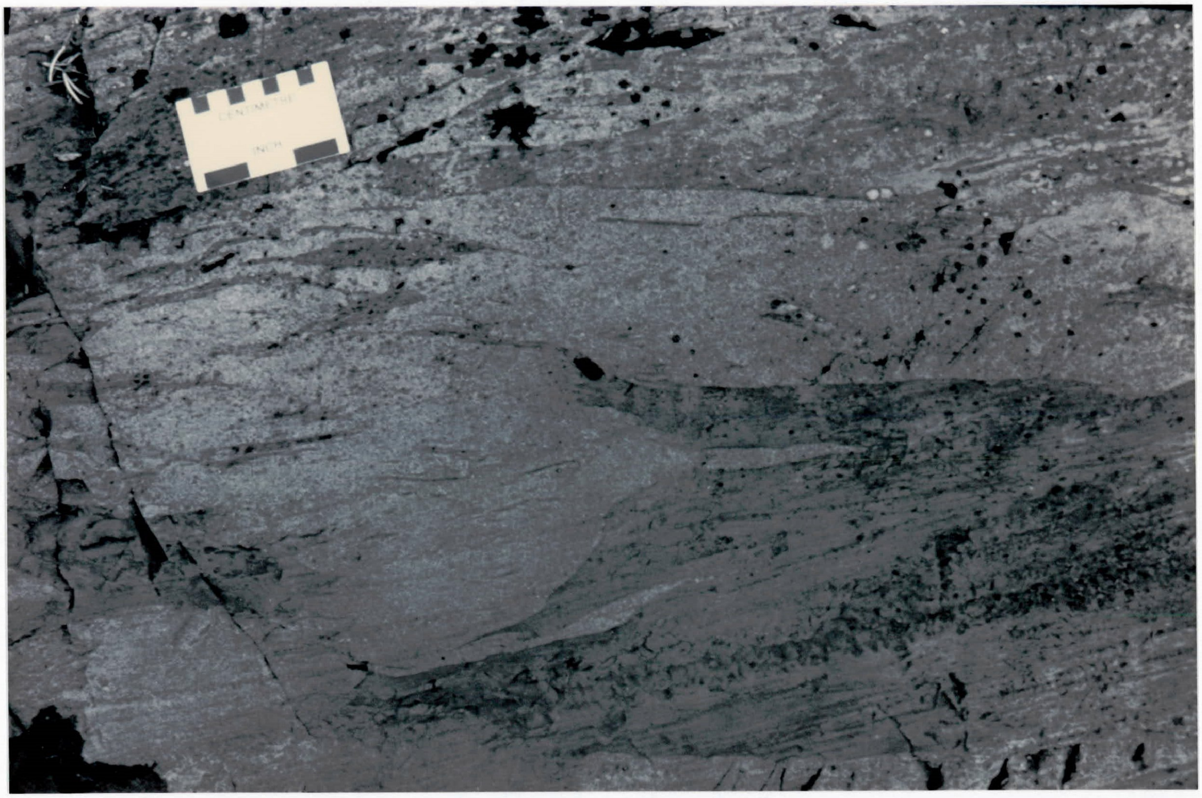
Some "lenticular" beds are present. Two bedding planes are inclined at 10° and 20°, and one bed thickens laterally over 7 m from 0.5 m to 1.3 m. These may be large scours or channels but, as usual, they may also be tectonized bed fragments.

Other features present are rare jasper sand grains, and penecontemporaneously deformed, interbedded coarser and silty finer sandstone (Fig. 7.9).

Graded sandstone. Graded beds are only rarely thicker than 20 cm; some 80-150 cm thick graded "units" may be single beds. Graded laminae are more common in coarse siltstone beds.

Sharp bases are characteristic. The beds' basal portions are coarse to fine-grained and massive, gradational upward to very fine-grained tops which may be plane laminated. Grading can be limited to the basal 0.5-1.0 cm of the thicker beds, particularly in pebbly sandstones showing coarse-tail grading of 2-25 mm clasts (Figs. 7.1 to 7.3). Stratigraphic tops to the north are by far the most common.

__Fig. 7.9:__ Soft-sediment deformation of sandstone. The coarser, lighter sandstone appears to have intruded and torn apart the finer, darker laminated sandstone and siltstone. Outcrop #98.


CENTIMETRE
INCH

The above assumes these beds fine upward, a fairly safe assumption.

No complete Bouma sequences were identified.

Sharp bed contacts with relatively coarser beds adjacent to the north are common, and are suggestive of sharp-based coarser beds in cases where the rest of the bed is obscured (e.g. by lichen, algae, soil).

In outcrop #104, to the north of a fine-grained sandstone bed and sharp bed contact, a very coarse to fine-grained sandstone bed or unit with numerous mudstone clasts displays a form of coarse-tail grading: the largest (up to 40 mm thick) clasts are near the inferred base, versus smaller (1-5 mm thick) ones 0.7-1.4 m higher up.

7.4 Mudstone Lithofacies

Coarse siltstone and mudstone, occurring as both beds and clasts, are common in the heterogeneous assemblage but are subordinate in amount to the conglomerate and sandstone lithofacies. The siltstone is most often planar, horizontal and parallel laminated (Fig. 7.10). Some microlaminae (sub-mm thick) are within thicker (mms) laminae that stand out owing to colour banding. Graded laminae are common: they are 1-5 mm thick, show tonal (light to dark) and grain size (very fine sand to mud) gradation, and usually fine northward. Some siltstone is transitional to very fine-grained sandstone, particularly as sandy laminae within siltstone beds.

Fig. 7.10: Plane laminated siltstone in contact with pebbly sandstone. Outcrop #98.


CENTIMETRE
INCH

Beds are cms to metres thick, some with lateral continuity of up to several metres, but in such tectonized exposures of heterogeneously layered sequences the bed thicknesses of the finest-grained layers are the most unreliable.

Mudstone clasts. These are typically 1-10 mm thick and 2-5 cm long, the largest examples being 4 by 30 cm and 10 by 50 cm. They weather black, brown, yellow and recessively. Some clasts contain very fine sand. Planar and parallel laminae and graded (fine or very fine sand to mud) laminae in the clasts are identical to those in nearby mudstone beds.

The clasts occur in sandstone, usually the coarser beds, and more rarely in clast-supported conglomerate frameworks. They occur more often in groups within a bed rather than singly. Small mudstone clasts are in some cases within beds interstratified with fine conglomerate beds or horizons.

CHAPTER 8: THE CMB's IDA LAKE ASSEMBLAGE

8.1 Introduction

A small number of outcrops in the Ida Lake area (Figs. 3.1, 3.2) form a lithofacies assemblage situated to the north of the CMB's conglomeratic assemblage. Fig. 3.1 gives the locations of specific outcrops referred to in the text. For local details such as bedding and schistosity measurements, the reader should examine Mackasey's (1975, 1976) maps.

The exposures of this sandstone-rich assemblage are characterized by a wide variety of sand grain sizes, distinctive very coarse-grained sandstone, and well-developed graded bedding.

8.2 Lithofacies Descriptions

Sandstone grain sizes. Sandstone beds with the following wide variety of grain sizes are present:

1) granule-rich to very coarse-grained;
2) poorly sorted, very coarse to medium-grained, with 0.5 cm mud chip clasts;
3) moderately to poorly sorted, coarse to fine-grained; this is the most common size range;
4) medium to fine-grained;
5) fine-grained;
6) silty fine to very fine-grained;

and also siltstone. The sandstones are usually massive, and less commonly laminated. The coarser varieties tend to be thinly (cms) bedded, versus the laminae of the finer sands. Sharp bed contacts are common. Sometimes a wide variety of grain sizes are interlaminated in 2-6 mm thick laminae or are interbedded, suggesting a rapidly fluctuating sand supply. Beds finer than medium-grained sand are pebble-free. Siltstone may be laminated.

Graded bedding. Outcrop #115 (Fig. 8.1) is composed of pebble-free, sharp-based (south bed contacts) beds with grading (fining north) in some of both the coarser and finer beds, mostly the coarser ones.

Outcrop #111 has several good examples of graded bedding (Fig. 8.2), all fining to the north (upward).

Outcrop #145, a roadside exposure by Pasha Lake, contains interbedded sandstone and mudstone. The beds are cms to tens of cms thick, with the sandstones generally thicker. Good examples of graded beds are present, all fining north (upward). One 3 cm thick bed fines from coarse to fine-grained sand to fine to very fine-grained, and a 6 cm thick bed fines from medium to fine-grained sand to fine to very fine-grained, with lamination in the upper part defined by single grain-thick layers of fine sand, probably an AB Bouma sequence.

Other features. Outcrop #111 contains one small concave north scour, and a mud chip band, a laterally discontinuous horizon of over ten 0.5 cm mud chip clasts distributed over 0.5 m along strike and near parallel to adjacent beds.

Jasper grains are occasionally seen in the sandstones.

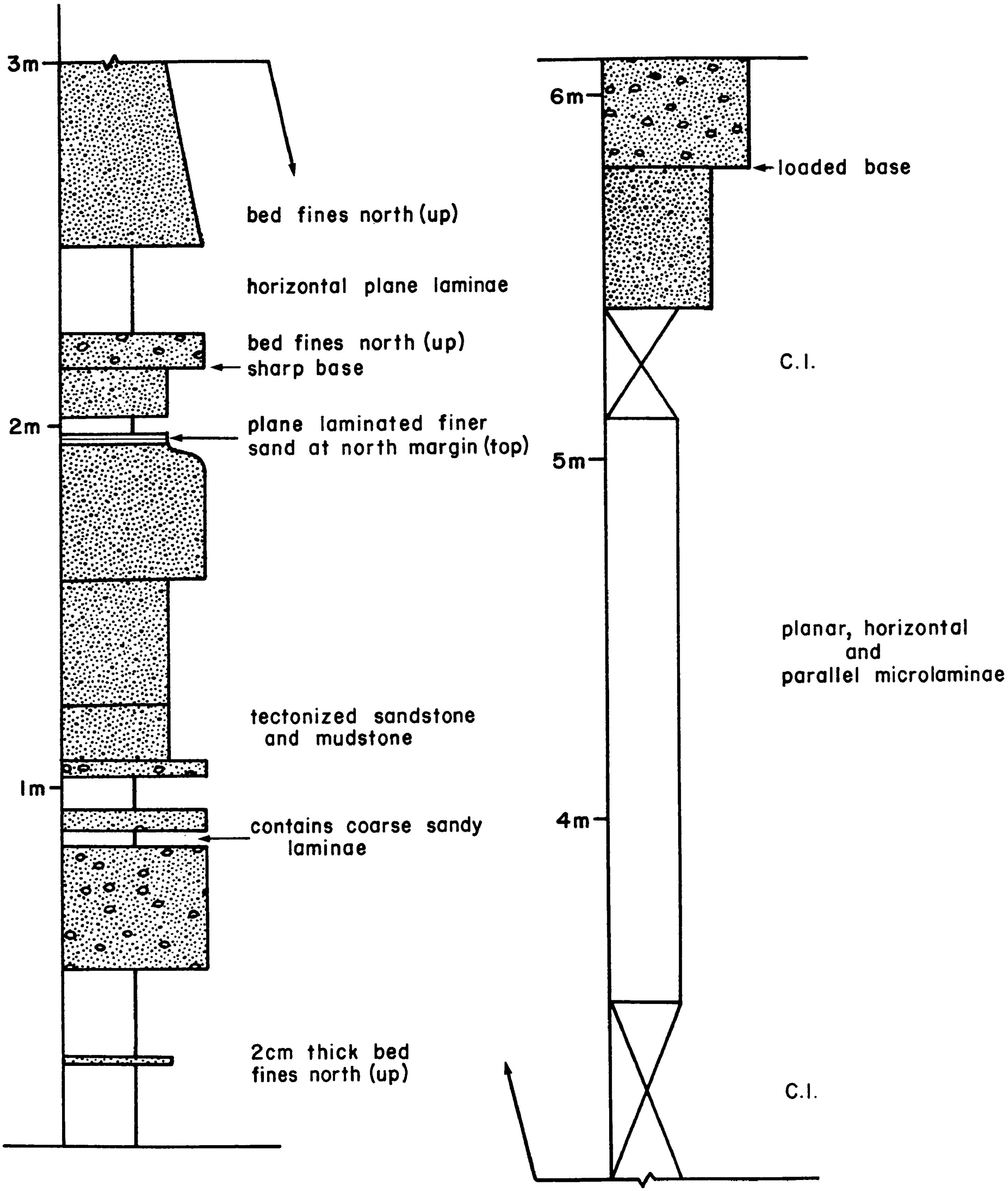
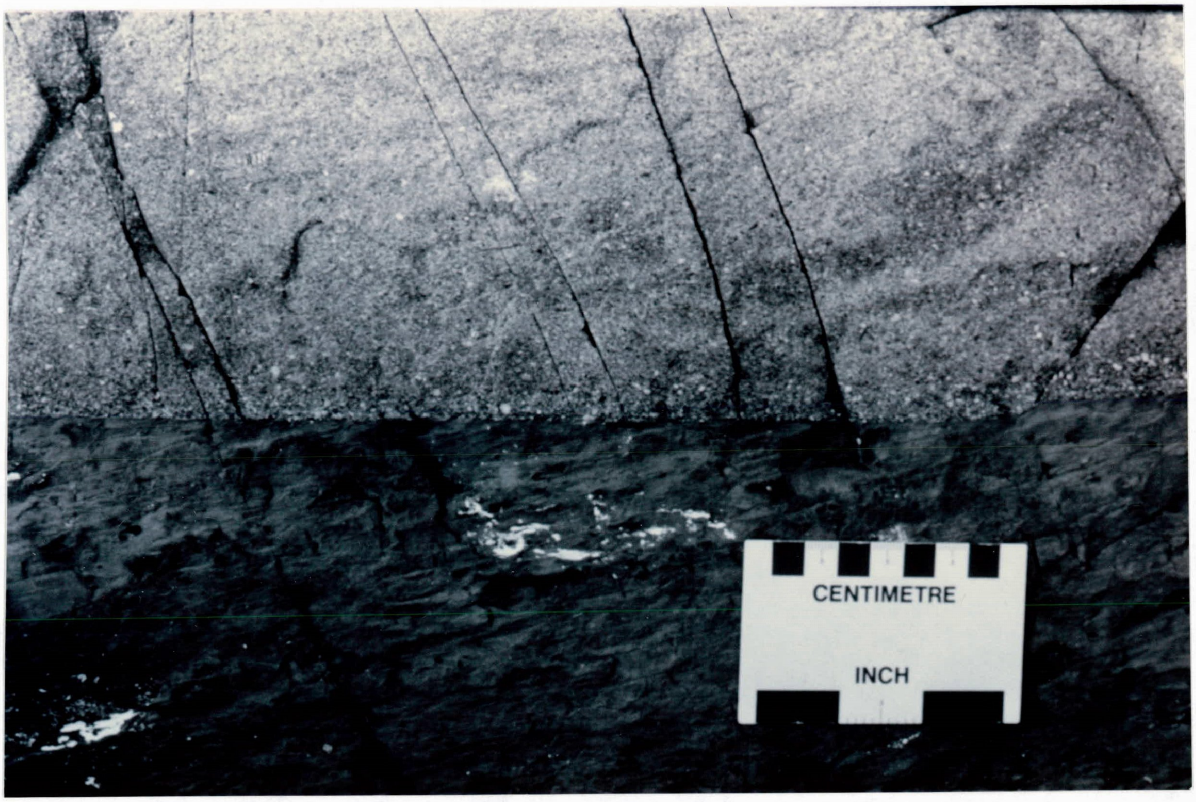

3m
2m
1m
bed fines north (up)
horizontal plane laminae
bed fines north (up)
sharp base
plane laminated finer
sand at north margin (top)
tectonized sandstone
and mudstone
contains coarse sandy
laminae
2cm thick bed
fines north (up)
6m
5m
4m
loaded base
C.I.
planar, horizontal
and
parallel microlaminae
C.I.

Fig. 8.1: Section from outcrop #115 (west bank of a creek). The graded beds fine northward (upward). Apparent horizontal thicknesses are given. Legend in Fig. 5.2.

Fig. 8.2: Dark, silty, very fine-grained sandstone overlain by sharp-based, coarse to medium-grained sandstone with coarse-tail grading (very coarse sand grains and granules along the base). Outcrop #111.


CENTIMETRE
INCH

CHAPTER 9: THE CMB's CORAL LAKE ASSEMBLAGE

9.1 Introduction

Fig. 3.1 shows the outcrops of the central metasedimentary belt (CMB) examined in east Leduc and Legault Townships (Fig. 1.1) and discussed in this chapter. This area's lithofacies assemblage will be referred to as the "Coral Lake (volcaniclastic) sediments" (Fig. 3.2). For local details such as bedding and schistosity measurements, the reader should examine the maps of Mackasey (1976) and Mackasey *et al* (1976a).

The Coral Lake sediments are not well-exposed. Outcrop #29 is the best exposure. Poor quality dirty and lichenous surfaces, deformation, and a low percentage of outcrop limited the data obtained and the interpretations derived.

Mackasey (1976) did not differentiate between the Coral Lake sediments and the obviously different polymict conglomerate and interbedded sandstone typical of most of the CMB (Chapters 3, 6 and 7), but Mackasey *et al* (1976a) did, terming the Coral Lake sediments:

> "Volcaniclastic conglomerate and associated sandstone", "composed mainly of volcanic material and displaying a relatively open framework".

These volcaniclastics have been included in the CMB as its easternmost exposures. They were mapped in the same way as the rest of the CMB, by examining sedimentological details; as Lajoie (1984, p.47) has stated, "The solution to the problem is to map volcaniclastic rocks as sediments, which they are." While superficially similar to the CMB's and NMB's conglomeratic assemblages, the Coral Lake volcaniclastic sediments display important compositional and textural differences.

The amount of volcaniclastic conglomerate, at 80-100% of the exposed surfaces, greatly exceeds the amount of interbedded volcaniclastic sandstone exposed.

Miscellaneous features. The D10 measurements (Table 6.2, Fig. 2.8) show the Coral Lake sediments' clasts to have greater L/S ratios than in most of the CMB.

There is a relatively extreme degree of clast elongation near the Jellicoe Fault (Mackasey, 1976), in outcrops #47 and 48 east of Leduc Lake (Fig. 3.1).

The Coral Lake sediments display an uncommon, red weathered surface at some sites. Some smooth weathered and fresh surfaces are bluish.

9.2 Conglomerate Lithofacies

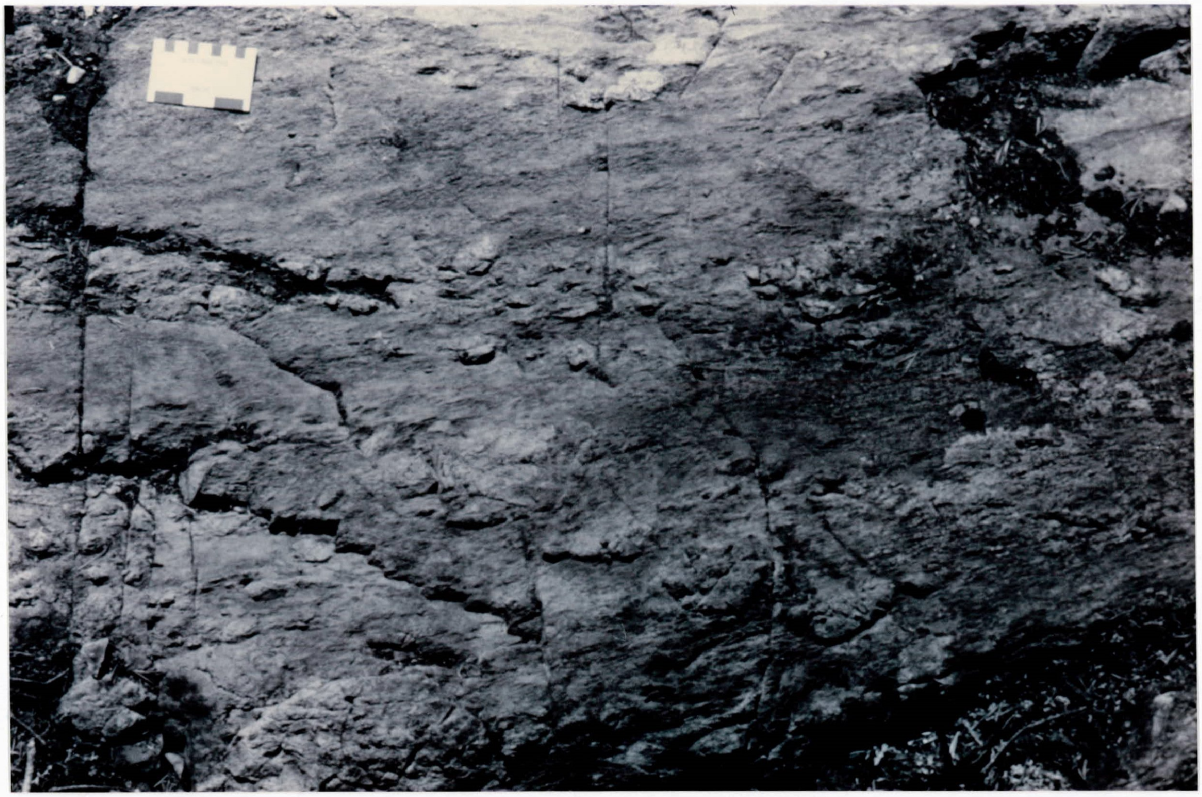
The schistose volcaniclastic conglomerate is massive to crudely bedded, clast to matrix-supported, and nearly oligomict (Figs. 9.1 to 9.3). Average deformed clast size is in the 16-64 mm (short axis) range. Cobbles are usually present, as observed in outcrop and shown by the D10 data (Table 6.2). Boulders are very rare. Some outcrops display a large variation in clast size, both along and across strike at a 10-100 m scale.

Fig. 9.1: Schistose, massive, clast-supported volcaniclastic conglomerate composed of felsic volcanic clasts. Outcrop #37. Compare this lithofacies with the polymict conglomerate in Fig. 6.6.


CENTIMETRE
INCH

Fig. 9.2: Interbedded clast-supported, oligomict volcaniclastic conglomerate and sandstone. Outcrop #29.

Fig. 9.3: Shadowed view of partially to wholly matrix-supported, very to extremely poorly sorted, oligomict volcaniclastic conglomerate. Outcrop #29.




CENTIMETRE
INCH

The ungraded clast frameworks are moderately to poorly sorted (Fig. 9.1, 9.2). Less common varieties of this facies, fine-grained and matrix-supported conglomerates, are described below.

Stratification. Well-bedded exposures are similar to the interbedded conglomerate and sandstone in the CMB and NMB. Crude bedding consists of coarser and finer layers; for example, one coarser layer is 30 cm thick and rich in 25-50 mm pebbles. Where the conglomerate is fine-grained, the crude bedding is akin to coarser pebble bands in a sandy matrix.

Where stratification is poorly defined, it may have been either originally so, or partially obliterated by shearing along the schistosity.

Fine-grained conglomerate. This finer variety of clast-supported conglomerate is locally common. Clasts range from 5-25 mm (short axis), with some spots rich in clasts less than 10 mm size. Sorting is poor to very poor, in contrast to the coarser, better sorted and more common conglomerate.

Matrix-supported conglomerate. The above fine-grained clast-supported conglomerate is occasionally gradational to a sandier, more poorly sorted, partially to wholly matrix-supported conglomerate (Fig. 9.3).

This matrix-supported variant is very poorly to extremely poorly sorted. Clasts tend to be smaller than in the adjacent clast-supported beds. There is a very subtle, gradual transition from clast-supported to matrix-supported areas; bed contacts are not easily recognized. Pebbly sandstone is frequently gradational to matrix-supported conglomerate.

9.3 Sandstone Lithofacies

Volcaniclastic sandstone beds (and conglomerate matrices) up to 30 cm thick are coarse to fine-grained and moderately to poorly sorted. Fine-grained beds are uncommon. Pebbly beds usually contain a few percent of small (5-20 mm) pebbles. The sand grains are angular to subround.

Sedimentary structures are extremely rare. Only two sets of cross-beds were seen, in outcrops #24 and 132.

Outcrop #62 contains some odd, faint laminae defined by medium versus fine-grained layers. Some graded laminae up to 1 cm thick grade from medium sand to silt and fine southward. Outcrop #132 has some plane laminated horizons cms thick.

9.4 Abnormal Exposures

Both of the large roadside outcrops of the Coral Lake sediments are atypical. Outcrop #132 is abnormally rich in felsic porphyry clasts, and contains a set of ripple cross-stratification. Outcrop #131 is better stratified and darker coloured than is normal.

Outcrop #131. This exposure of felsic volcaniclastic conglomerate is mostly matrix-supported, but is also clast-supported, especially in the better-stratified spots. The matrix sand is coarse to fine-grained, with some planar, horizontal and parallel laminae 1-3 mm thick. Some "laminae" in this outcrop may be structural features (shear bands?). More schistose horizons may have originally been muddier.

Stratification is shown by conglomerate-sandstone contacts, and sharp contacts of the fine-grained and coarser sandstone. The thicker conglomerate beds are 15-30 cm thick. Beds metres thick are probably present, but difficult to recognize in this tectonized outcrop. One possible inversely graded bed is present, a clast-supported conglomerate rich in 2-10 mm pebbles in its northern part, 30 mm pebbles in its middle, and a sandy southern part darkening (tonal gradation) south; the above features suggest tops south in this bed.

9.5 Conglomerate Clast Composition

Nearly all of the conglomerate´s clasts are of felsic volcanic composition (Figs. 9.1 to 9.3; see also the criteria in Table 10.1). The clasts weather to white, pink, light green, grey and blue. They are aphanitic to fine-grained; some coarser porphyritic and very rare ash-textured clasts are also present.

Very minor amounts (1-2%) of the following clast lithologies are present: felsic porphyry, mafic volcanic, jasper, and quartz. Mafic volcanic clasts are aphanitic, very dark green, and recessively weathered. Quartz and jasper comprise less than 1% of small (square metres) areas. The interbedded sandstone contains jasper grains.

The largest clasts are nearly always felsic volcanics, and rarely felsic porphyry.

Overall, the conglomerate is close to being oligomict (monomict) in composition. Although the minor constituents listed above make it not truly oligomict, felsic volcanic clasts strongly dominate (98-100%) the composition. Most individual outcrops contain felsic volcanic clasts only.

Outcrop #24 contains some spots with unusual compositions, somewhat similar to the normal polymict conglomerate of the CMB, with porphyry, chert, and mafic volcanic clasts, but no granitoid clasts. Point count #8 was done at this site (see Sections 10.3, 10.5).

CHAPTER 10: PROVENANCE STUDY OF THE NMB AND CMB

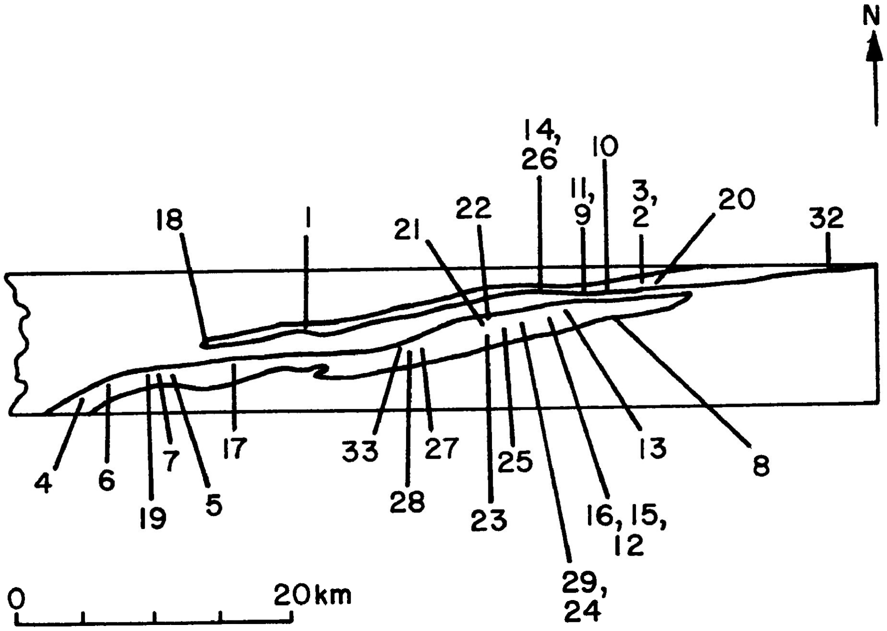
10.1 Introduction

Part of the detailed examination of the NMB and CMB involved an attempt to search for and document any regional trends in clast composition. Point counts of 500 non-matrix points per count were performed at 11 sites in the NMB and 20 sites in the CMB (Fig. 10.1) using the methods outlined in Section 1.8. Criteria for the field identification of the clast lithologies are given in Section 10.2. The tabulated results are presented in Sections 10.3 and 10.4 and discussed in Section 10.5.

Investigating sediment provenance via the study of gravel-sized clast lithologies is far more reliable than thin section examination of metamorphosed (greenschist facies) sand grains, which are in the study area typically strain-shadowed quartz, albite plagioclase, altered orthoclase, and chloritic, largely unrecognizable small rock fragments and matrix.

Point counting of clasts in the field and general field observations allow examination of many more clasts than laboratory methods but are obviously not as sophisticated as petrographic (thin section) study. Sampling of specific interesting clasts in the well-lithified metaconglomerate is very difficult; on most surfaces, half an hour with a hammer and chisel will usually result in a lot of rock flour but not

<u>Fig. 10.1:</u> Locations of the study area's numbered point counts (point count #1, 2, etc.) given in Tables 10.2 to 10.5, 10.8 and 10.9. Refer to Fig. 1.1 for the location map of study area.


N
14,
26
10
11,
9
3,
2
20
32
18
1
21
22
4
6
19
7
5
17
33
28
27
23
25
29,
24
16, 15,
12
13
8
0
20km

enough sample for a thin section.

Figs. 2.3, 6.8, 7.6, and 6.6 show parts of the outcrop surfaces of point counts #14, 4, 29 and 21, respectively.

10.2 <u>Field Identification of Clast Lithologies</u>

Table 10.1 gives the criteria used in identifying the various clast lithologies present. The colour index ranges given apply to the greenschist facies. No pillowed mafic volcanic clasts were seen. Mudstone clasts are generally very rare, but are locally common in the CMB, some being obvious intraformational (rip-up) clasts.

Generally, the granitoid and quartz clasts are the most mechanically competent (least elongate/oblate), and the mafic volcanic clasts the least competent and bent around more competent clast lithologies.

TABLE 10.1

Criteria for field (at the outcrop) identification of clasts in polymict conglomerate of the NMB and CMB

Felsic volcanic	- low colour index (<15%) - light green, whitish and pink weathered surfaces - some phenocrystic, flow banded clasts - rare ash-textured clasts - resistant-weathering compared to most other clasts - tonally gradational to clasts of intermediate colour index - texturally gradational to felsic porphyry and fine-grained granitoid clasts

Table 10.1 (Cont'd)

Mafic volcanic	- high colour index (>35%) - dark green fresh and weathered surfaces - aphanitic (almost always) - recessively weathered compared to all other clast lithologies and the conglomerate's quartzofeldspathic sandy matrix - tonally gradational to clasts of intermediate colour index
Felsic porphyry	- similar to criteria for felsic volcanic clasts - intrusive or volcanic origin uncertain
Granitoid	- usually leucocratic; whitish to pinkish weathered surfaces - usually equigranular and medium-grained (1-5 mm grains) - resistant-weathering compared to most other clasts - in relatively rare cases, possibly transitional to felsic porphyry - least elongate/oblate of all strained clasts, excepting quartz clasts
Chert	- red (jasper), black, and whitish varieties - commonly laminated - highly resistant to weathering - black chert clasts are very resistant to weathering and often display the best glacial striations (Pleistocene) of any clast lithology, showing that they are not mudstone clasts (which should be recessively weathered) - weakly magnetic black chert clasts are transitional to strongly magnetic magnetite iron formation clasts
Iron formation	- dark, magnetic varieties are magnetite-rich - red to purplish, rusty, non-magnetic varieties are likely hematitic - iron minerals may be interlaminated with either clastic or chemical sediment laminae
Quartz	- white to clear, vitreous lustre, Moh's hardness of 7, conchoidal fracture, most resistant to weathering - possible confusion with white chert clasts, but chert clasts are usually more elongate/oblate

Table 10.1 (Cont'd)

Other lithologies	- most often of medium-grained (1-5 mm), mafic to intermediate composition (dioritic, gabbroic, diabasic clasts)
Unknown	- "fubarite" clasts.

10.3 Point Count Results

Raw data for the NMB and CMB are presented in Tables 10.2 and 10.3, and in a partially grouped, more generalized form in Tables 10.4 and 10.5.

The following equation:

$$E = Z_{\alpha}/2 \sqrt{\frac{x/n(1-x/n)}{n}} \qquad (1)$$

where x = number of points counted for a given category (i.e. a clast lithology) in a given point count

n = 500 = total number of non-matrix points counted per count

$Z_{\alpha}/2$ = 1.96 for a 95% confidence interval

E = maximum error of the estimate (converted to a percentage in Tables 10.6, 10.7)

gives the maximum error of an estimate (Freund, 1984, p.326). Table 10.6 lists sample error values that can be applied to Tables 10.2 to 10.5, 10.8 and 10.9.

The point count locations (Fig. 10.1) are numbered in the order that the counts were done; because the sequence is geographically staggered/scattered, there is no systematic geographic drift in any possible bias.

Regarding the spacing of the point count localities (Fig. 10.1), conglomerate exposures in the NMB in most of

TABLE 10.2

Clast Lithologies by Percentage of Area in a Point Count

Township	Outcrop	Point Count	FV	FIV	FVG	MV	MIV	FPO	GR	RCH	BCH	LCH	QTZ	OTH
Sandra	#55	#18	58.6	8.6	4.6	2.6	3.8	4.4	10.4	–	0.4	0.2	0.4	6.0
Irwin	3	1	64.8	←	←	10.2	←	8.6	15.6	0.2	–	0.2	–	0.4
Leduc	35 N	14	70.6	9.0	5.0	3.0	4.2	3.2	4.4	–	0.2	–	0.2	0.2
	*35 S	26	47.2	13.6	5.6	11.2	5.8	4.8	11.0	0.4	–	–	0.4	–
	27 N	11	69.0	9.0	2.0	9.2	4.2	0.4	4.0	0.4	–	1.2	0.6	–
	27 S	9	86.8	2.6	4.0	–	0.4	0.6	4.8	0.2	0.2	0.2	–	0.2
	28	10	77.4	4.2	4.4	2.4	3.2	1.2	7.0	–	0.2	–	–	–
Legault	8	3	43.0	2.2	4.4	28.0	3.2	5.0	13.6	0.2	0.2	0.2	–	–
	7	2	28.8	6.4	1.4	44.4	9.4	0.4	8.0	0.2	0.2	0.2	0.6	–
	58	20	60.0	12.2	8.8	2.6	7.0	1.8	6.0	–	–	0.4	1.2	–
Colter	129	32	53.4	12.6	11.8	1.8	4.4	1.2	13.4	0.2	–	–	0.2	1.0

Table 10.2: Raw data from compositional point counts of conglomerate beds in the NMB.

FV = felsic volcanic
FIV = felsic to intermediate volcanic
FVG = felsic volcanic to granitoid
MV = mafic volcanic
MIV = mafic to intermediate volcanic
FPO = felsic porphyry
GR = granitoid
RCH = red chert
BCH = black chert
LCH = light coloured chert
QTZ = quartz
OTH = other

* Conglomerate "lens" within mafic volcanics on Mackasey's (1976) map; see Section 2.7.

TABLE 10.3

Clast Lithologies by Percentage of Area in a Point Count

Township	Outcrop	Point Count	FV	FIV	FVG	MV	MIV	FPO
Dorothea	#13	# 4	61.4	5.2	3.4	5.8	5.4	2.4
Sandra	17	6	56.2	6.0	11.2	0.4	2.8	2.6
	57	19	75.8	5.2	6.8	0.2	2.6	1.0
	20	7	46.8	12.4	6.8	12.6	5.8	4.2
	14	5	55.2	8.0	4.4	8.0	4.2	2.4
Irwin	53	17	70.6	10.6	3.2	3.2	3.4	2.4
Walters	9	33	64.8	7.8	8.4	–	2.8	2.6
	110 W	28	62.8	8.0	11.6	1.0	2.6	0.6
	110 E	27	57.8	8.0	11.4	1.0	2.2	3.4
Leduc	71	22	63.2	8.6	4.2	1.6	4.4	11.0
	64	21	66.0	10.0	2.6	4.2	1.0	7.6
	97	23	61.8	9.6	3.8	3.0	3.8	12.4
	83	25	63.6	9.2	1.6	2.8	2.8	10.6
	98	29	59.4	8.2	4.4	3.4	3.8	6.4
	103	24	60.2	10.2	4.4	3.0	4.6	11.8
	42	16	70.4	10.0	3.8	2.4	3.8	1.0
	31 S	15	64.2	5.4	5.0	6.2	4.8	1.4
	32	12	74.0	6.2	4.6	3.0	1.8	0.6
	31 N	13	74.6	5.0	6.4	1.8	2.4	1.6
	24	8	73.6	6.2	0.2	14.8	3.6	0.6

Township	Outcrop	Point Count	GR	RCH	BCH	LCH	QTZ	OTH
Dorothea	#13	# 4	8.2	–	0.2	0.4	0.4	7.2
Sandra	17	6	19.6	–	0.4	0.2	0.4	0.2
	57	19	6.2	0.6	0.6	–	1.0	–
	20	7	8.4	0.6	0.2	–	–	2.2
	14	5	15.4	–	0.2	–	1.4	0.8
Irwin	53	17	5.4	–	–	–	1.2	–
Walters	9	33	10.8	1.0	0.4	0.6	0.8	–
	110 W	28	10.2	0.6	0.2	–	0.6	1.8
	110 E	27	12.0	0.4	0.2	0.4	0.6	2.6
Leduc	71	22	6.0	–	–	–	–	1.0
	64	21	5.8	–	–	–	–	2.8
	97	23	4.6	0.4	0.2	–	0.4	–
	83	25	4.4	–	–	–	–	5.0
	98	29	8.4	–	–	0.2	0.4	5.4
	103	24	4.6	0.6	0.2	–	0.2	0.2
	42	16	7.4	0.4	–	0.2	0.2	0.4
	31 S	15	10.6	0.6	0.2	0.4	1.2	–
	32	12	8.4	0.6	0.2	0.2	0.4	–
	31 N	13	7.4	0.2	0.4	–	0.2	–
	24	8	–	–	–	0.6	0.2	0.2

Table 10.3: Raw data from compositional point counts of conglomerate beds in the CMB. Lithological abbreviations as in Table 10.2.

TABLE 10.4

Clast Lithologies by Percentage of Area in a Point Count

Township	Outcrop	Point Count	FV	MV	FPO	GR	CH	QTZ	OTH
Sandra	#55	#18	67.2	6.4	4.4	15.0	0.6	0.4	6.0
Irwin	3	1	64.8	10.2	8.6	15.6	0.4	-	0.4
Leduc	35 N	14	79.6	7.2	3.2	9.4	0.2	0.2	0.2
	35 S	26	60.8	17.0	4.8	16.6	0.4	0.4	-
	27 N	11	78.0	13.4	0.4	6.0	1.6	0.6	-
	27 S	9	89.4	0.4	0.6	8.8	0.6	-	0.2
	28	10	81.6	5.6	1.2	11.4	0.2	-	-
Legault	8	3	45.2*	31.2*	5.0	18.0	0.6	-	-
	7	2	35.2*	53.8*	0.4	9.4	0.6	0.6	-
	58	20	72.2	9.6	1.8	14.8	0.4	1.2	-
Colter	129	32	66.0	6.2	1.2	25.2	0.2	0.2	1.0
Average			67.2	14.6	2.9	13.7	0.5	0.3	-
Range (low to			35(?)	< 1	< 1	6	< 1	0	-
high)			89	54(?)	9	25	2	1	-

Table 10.4: Partially grouped data from Table 10.2; compositional point counts of conglomerate beds in the NMB.

FV = felsic (to intermediate) volcanic
MV = mafic (to intermediate) volcanic
FPO = felsic porphyry
GR = granitoid
CH = chert
QTZ = quartz
OTH = other

* Suspect results; possibly inconsistent observations (see discussion in text).

TABLE 10.5

Clast Lithologies by Percentage of Area in a Point Count

Township	Out-crop	Point Count	FV	MV	FPO	GR	CH	QTZ	OTH
Dorothea	#13	# 4	66.6	11.2	2.4	11.6	0.6	0.4	7.2
Sandra	17	6	62.2	3.2	2.6	30.8	0.6	0.4	0.2
	57	19	81.0	2.8	1.0	13.0	1.2	1.0	–
	20	7	59.2	18.4	4.2	15.2	0.8	–	2.2
	14	5	63.2	12.2	2.4	19.8	0.2	1.4	0.8
Irwin	53	17	81.2	6.6	2.4	8.6	–	1.2	–
Walters	9	33	72.6	2.8	2.6	19.2	2.0	0.8	–
	110 W	28	70.8	3.6	0.6	21.8	0.8	0.6	1.8
	110 E	27	65.8	3.2	3.4	23.4	1.0	0.6	2.6
Leduc	71	22	71.8	6.0	11.0	10.2	–	–	1.0
	64	21	76.0	5.2	7.6	8.4	–	–	2.8
	97	23	71.4	6.8	12.4	8.4	0.6	0.4	–
	83	25	72.8	5.6	10.6	6.0	–	–	5.0
	98	29	67.6	7.2	6.4	12.8	0.2	0.4	5.4
	103	24	70.4	7.6	11.8	9.0	0.8	0.2	0.2
	42	16	80.4	6.2	1.0	11.2	0.6	0.2	0.4
	31 S	15	69.6	11.0	1.4	15.6	1.2	1.2	–
	32	12	80.2	4.8	0.6	13.0	1.0	0.4	–
	31 N	13	79.6	4.2	1.6	13.8	0.6	0.2	–
	24	8	79.8	18.4	0.6	0.2	0.6	0.2	0.2
Average			72.1	7.4	4.3	13.6	0.6	0.5	–
Range (low to high)			59 81	3 18	< 1 12	< 1 31	0 2	0 1	– –

Table 10.5: Partially grouped data from Table 10.3; compositional point counts of conglomerate beds in the CMB. Lithological abbreviations as in Table 10.4.

TABLE 10.6

% of n (n = 500)	± x%	% of n (n = 500)	± x%
90	2.6	33	4.1
86	3.0	30	4.0
83	3.3	26	3.8
80	3.5	23	3.7
76	3.7	20	3.5
73	3.9	16	3.2
70	4.0	13	2.9
66	4.2	10	2.6
63	4.2	9	2.5
60	4.3	8	2.4
56	4.4	7	2.2
53	4.4	6	2.1
50	4.4	5	1.9
46	4.4	4	1.7
43	4.3	3	1.5
40	4.3	2	1.2
36	4.2	1	0.9

Table 10.6: Maximum errors of estimates (± x%) of a given percentage value (% of n points counted).

TABLE 10.7

Clast Lithology	% of NMB	% of CMB
Felsic volcanic	67.2 ± 4.1	72.1 ± 3.9
Mafic volcanic	14.6 ± 3.1 (8.4 ± 2.4)	7.4 ± 2.3
Felsic porphyry	2.9 ± 1.5	4.3 ± 1.8
Granitoid	13.7 ± 3.0	13.6 ± 3.0
Chert	0.5 ± 0.6	0.6 ± 0.7
Quartz	0.3 ± 0.5	0.5 ± 0.6

Table 10.7: Estimates of the average (mean) compositions of the NMB and CMB conglomerates, derived from Tables 10.4, 10.5 and 10.6.

TABLE 10.8

NMB DATA

Township	Out-crop	Point Count	FPS	FVP	FPG	FVG	FVA	MII	MIF	FEL	RIP	TOT
Sandra	#55	#18	-	-	-	-	-	6.0	-	-	-	6.0
Irwin	3	1	-	-	-	-	-	-	0.4	-	-	0.4
Leduc	35 N	14	-	-	-	-	-	-	-	0.2	-	0.2
	27 N	9	-	-	-	-	-	-	0.2	-	-	0.2
Colter	129	32	-	1.0	-	-	-	-	-	-	-	1.0

TABLE 10.9

CMB DATA

Township	Out-crop	Point Count	FPS	FVP	FPG	FVG	FVA	MII	MIF	FEL	RIP	TOT
Dorothea	#13	# 4	0.4	-	-	-	-	6.4	-	-	-	*6.8
Sandra	17	6	0.2	-	-	-	-	-	-	-	-	0.2
	20	7	0.2	-	-	-	-	0.6	-	0.2	1.2	2.2
	14	5	0.8	-	-	-	-	-	-	-	-	0.8
Walters	110 W	28	0.6	1.2	-	-	-	-	-	-	-	1.8
	110 E	27	-	2.6	-	-	-	-	-	-	-	2.6
Leduc	71	22	-	-	-	-	1.0	-	-	-	-	1.0
	64	21	-	-	-	-	2.8	-	-	-	-	2.8
	83	25	-	1.2	1.0	2.0	0.8	-	-	-	-	5.0
	98	29	-	5.0	-	-	-	0.4	-	-	-	5.4
	103	24	-	-	-	-	0.2	-	-	-	-	0.2
	42	16	-	-	-	-	-	-	-	-	0.4	0.4
	24	8	-	-	-	-	-	-	-	0.2	-	0.2

Tables 10.8 and 10.9: Breakdowns of the "other" subpopulations in Tables 10.2 to 10.5.

FPS = felsic pyroclastic or sandstone
FVP = felsic volcanic to porphyry
FPG = felsic porphyry to granitoid
FVG = felsic volcanic to granitoid
FVA = dark felsic volcanic or argillite
MII = mafic to intermediate "intrusive"
MIF = magnetite iron formation
FEL = felsite (chert/quartz/felsic volcanic)
RIP = intraformational rip-up clasts
TOT = total of other

* plus 0.4% unknown clasts.

Walters and Leduc Townships are too deformed for reliable clast identification, and in the CMB in western Walters Township the outcrops are too poor in quality to warrant point counting.

Table 10.7 compares the average composition of the NMB and CMB.

10.4 Minor Compositional Components

Tables 10.8 and 10.9 give breakdowns of the subpopulation "other" in Tables 10.2 to 10.5.

Minor components of the CMB. Clasts identified as felsic pyroclastics, based on their colour index, "sandy" texture and relative resistance to deformation, are sometimes laminated. Confusion of these with true (epiclastic) sandstone clasts is possible.

Chert clasts are red (jasper), white, and black. Three clasts with black and red laminae were found. One laminated, magnetic black chert is probably gradational to a lean iron formation. Magnetite iron formation clasts were identified by their strong magnetism.

Laminated siltstone clasts, mudstone clasts, and obvious rip-up (intraformational) clasts are present in the CMB's fluvial conglomerate. The former two may be either intrabasinal or extrabasinal.

Mafic to intermediate "intrusive" clasts are probably diorite or diabase, although coarse-grained basalt is also a possibility.

Multi-lithology clasts are granitoids with pre-erosion aplitic veins.

Three possible gneissic clasts were seen, but appear to be more like slightly foliated granitoids than high-grade gneisses (i.e. with paleosome and neosome).

Minor components of the NMB. Clasts identified as felsic pyroclastic, chert (red, black and white; including one jasper clast with black laminae, and two magnetic black cherts), magnetite iron formation, mafic to intermediate intrusive (Fig. 10.2), and granitoid and felsic volcanic clasts with pre-erosion quartz veins are similar to those of the CMB, described above.

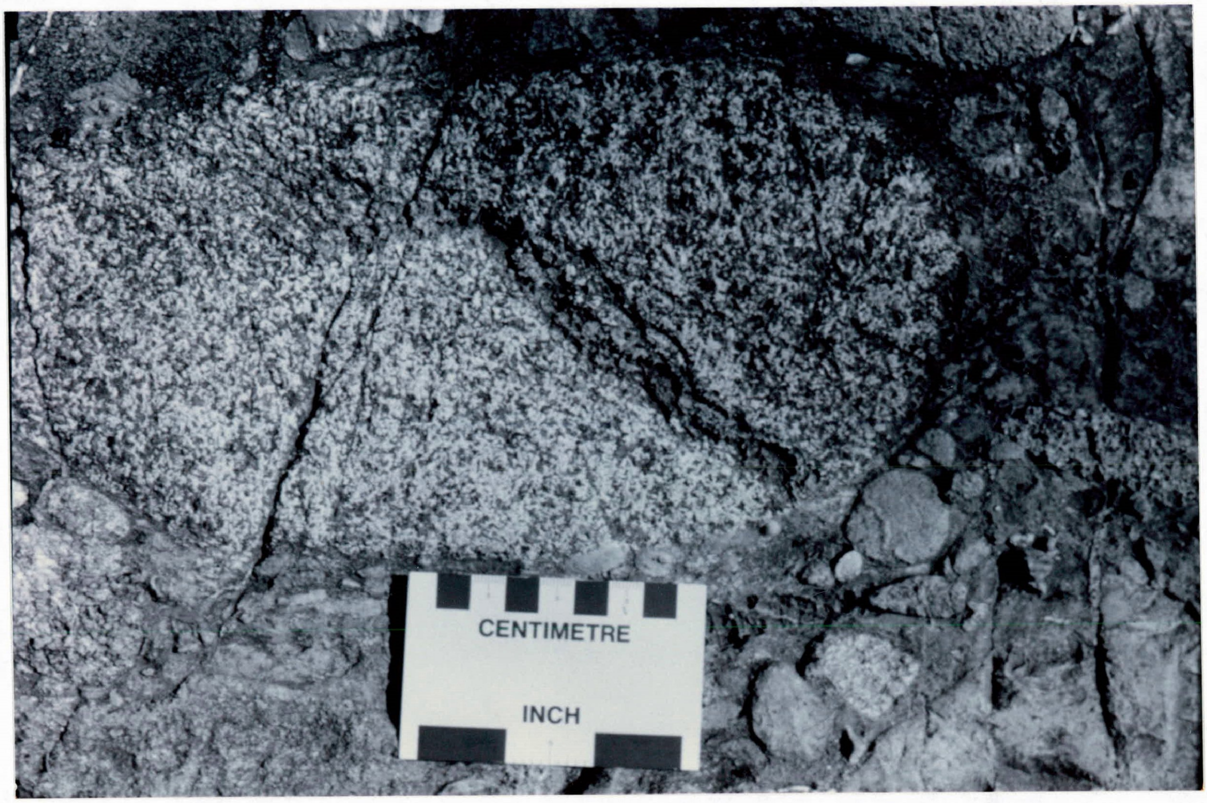
10.5 Interpretation of Compositional Trends and Provenance

Compositional correlation of the NMB and CMB. The results of the conglomerate point counts (Tables 10.4, 10.5, 10.8, 10.9) reveal little systematic variation along regional strike within each of the NMB and CMB. Whether or not this also reveals little variation along strike in the source terrain(s) is unknown because:

1) the precise geographic source of the clasts is unknown (see Section 13.3);
2) the amount of mixing of clasts along transport paths is unknown; and
3) the role of any possible easterly or westerly-directed transport systems is mostly conjectural.

Regional structural complexity, poor exposure (low density

Fig. 10.2: Mafic to intermediate "intrusive" clasts (the small clast at the middle right margin is similar to the large one). Outcrop #55.


CENTIMETRE
INCH

and quality), and a lack of distinctive marker lithologies (those with a clast and source lithology known, unique and unequivocal) preclude the identification of a specific source terrain. Indeed, the source terrain(s) may have been completely eroded away, leaving only its detritus in the NMB and CMB as a record of its presence. These points are discussed further in Section 13.3.

Table 10.7 indicates that there is a greater concentration (percent area) of mafic volcanic clasts in the NMB, but if the one abnormally high value (53.8%) in Table 10.4 is omitted there is overlap (8.4 ± 2.4% versus 7.4 ± 2.3%) between the NMB and CMB. This raises the question: are the percent concentrations of mafic volcanic clasts in point counts 2 and 3 really richer in such clasts and proportionally poorer in felsic volcanic clasts (Table 10.4), or are inconsistent clast identifications to blame for the anomalous results? The answer is probably the latter choice, although the mafic volcanic clast-rich basal conglomerate outcrop (Sections 2.3, 11.5) is only 1.0-1.2 km away.

The extremely good match between the NMB's and CMB's average composition (Table 10.7) strongly suggests that either they had source terrains of identical character, or the NMB and CMB are two now-separate halves of the same clastic depositional systems (paleo-) tract. Further discussion below and in Section 13.1 favours the latter choice.

The abnormally high concentration of the distinctive mafic to intermediate "intrusive" clasts (Fig. 10.2) in only the western portions of the NMB and CMB (Tables 10.8, 10.9), to the west of each belt's sections rich in aquabasinal facies in Irwin Township (see Sections 2.6, 4.2, 11.2, 11.6), defines an excellent compositional match between the two belts. These diabase/gabbro/diorite clasts are smaller in the CMB. Along with the nearly identical average compositions of the NMB and CMB (Table 10.7), and the CMB's generally finer-grained conglomerate (compare Tables 2.1, 6.2), these results also strongly suggest that either two different (NMB and CMB) systems had source terrains of identical composition, or the NMB and CMB are the proximal and distal parts of one system that was later tectonically separated; the presence of rip-up clasts (a more distal feature) in the CMB (Sections 6.4, 11.12; Table 10.9) and not the NMB suggests the latter option, a scenario elaborated on in Section 13.1. Evidence from the western parts of both the NMB and CMB suggests that an at least partially easterly-directed depositional system containing the distinctive mafic to intermediate "intrusive" clasts may have been separate from a system situated to the east, one that was at least partially westerly-directed and did not contain the distinctive clasts (see Section 12.4).

Weak trends along strike within the CMB. There is a greater variety of rare clast lithologies in the CMB than in the NMB (compare Tables 10.8 and 10.9), but nearly twice as many point counts were done in the CMB.

The CMB in Dorothea and Sandra Townships has trace amounts of felsic pyroclastic/sandstone clasts, and significant amounts of mafic to intermediate intrusive (diabase/diorite) (Table 10.9) and intraformational (rip-up) clasts. Outcrop observations (Section 6.4) show the presence of rip-up clasts better than Table 10.9 does.

There is a subtle enrichment in felsic volcanic clasts from west to east in the CMB (Table 10.5, Fig. 10.3), which may reflect increasing proximity to the Coral Lake "sediments" (Chapter 9), a possible source of such clasts. The two anomalously high values (81.0 ± 3.4% and 81.2 ± 3.4%) from point counts #17 and 19 (Table 10.5) are near the western CMB's north margin, possibly a result of greater proximity to a felsic volcanic source (see Sections 12.4, 13.3) in a prograded, CMB-scale section (see Sections 12.3).

The following are distinctive features of western Leduc Township's Beatty Lake area (point counts #22, 21, 23, 25, 29, 24 in Tables 10.5 and 10.9):

1) more abundant felsic porphyry clasts than in the rest of the CMB (Table 10.5);
2) unusual felsic volcanic clasts (Table 10.9; one type is also present several kms to the west, at point counts #27, 28); and
3) an unusual dark felsic volcanic, or possibly argillite, clast lithology (Table 10.9).

The anomalies may reflect proximity to the Coral Lake "sediments" (Chapter 9), a felsic volcanic clast-rich lithofacies assemblage and a possible source of clasts for

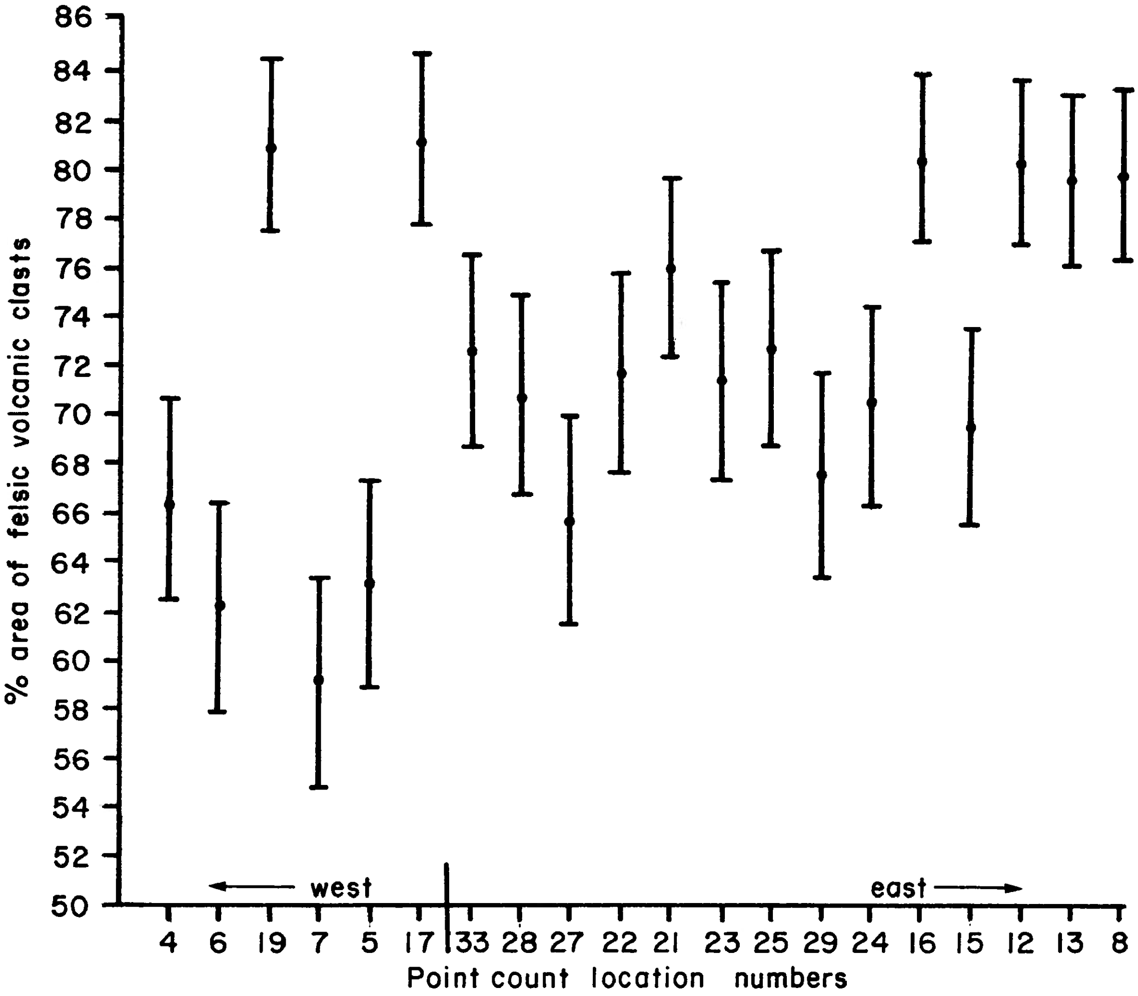

% area of felsic volcanic clasts
86
84
82
80
78
76
74
72
70
68
66
64
62
60
58
56
54
52
50
west
east
4
6
19
7
5
17
33
28
27
22
21
23
25
29
24
16
15
12
13
8
Point count location numbers

Fig. 10.3: Variation in percent area of felsic volcanic clasts in point counts performed on conglomerate beds in the CMB. Data points are from Table 10.5, with error bars (see Section 10.3).

the eastern CMB.

Point count #8 was performed on outcrop #24 (Tables 10.3, 10.5), part of the Coral Lake "sediments" (Chapter 9). This site has:

1) 98.2 ± 1.2% volcanic clasts;
2) 79.8 ± 3.4% felsic volcanic clasts, the remainder of the volcanic clasts being darker, more mafic to intermediate (?) clasts; and
3) virtually no granitoid component.

This volcanic clast-rich character is consistent with the Coral Lake "sediments" largely oligomict, felsic pyroclastic and/or volcaniclastic origin. Either not all of the Coral Lake conglomeratic "sediments" are oligomict, or some dark felsic clasts (e.g. black rhyolites) have been identified as mafic ones.

Implications of the clast lithologies present. Tables 10.2, 10.3, 10.8, 10.9 and field observations reveal that a wide variety of clast lithologies were supplied to the gravelly fluvial and submarine fan/fan delta front systems (see Chapter 11), a direct result of the variety of bedrock lithologies in the source region's drainage basins. An exception to the above would be a source region of polymict conglomerate, with some or all of the inferences about provenance applying to an earlier cycle of erosion, deposition and burial (no conglomerate clasts are present).

Volcanic and felsic clasts are by far the most abundant in both the NMB and CMB (Table 10.7), reflecting the importance of:

1) volcanism (both felsic and mafic);
2) felsic volcanism;
3) the unroofing of plutonic rocks; and
4) the possible relative enrichment of resistates (felsic clasts).

It is not surprising that abundant felsic volcanic clasts, including some possible pyroclastic ones, are present, since felsic volcanic edifices/centres are thought to have been the source of most Archean clastic material (Thurston *et al*, 1985; Ojakangas, 1985; Ayres and Thurston, 1985), an hypothesis which also accounts for the absence of extra-basinal sandstone clasts in the study area.

The presence of granitoid and felsic volcanic clasts with pre-erosion quartz or aplite veins requires two pre-erosion intrusive events or one multi-phase event.

In a previous study (Devaney, 1982), HF staining of sawn slabs containing granitoid clasts showed that some K-feldspar-bearing granite clasts are present in the easternmost NMB. This is important because early diapiric granitoids are thought to have been largely tonalitic, versus younger and often potassic granites, in Canadian Archean greenstone belt development (Ayres and Thurston, 1985).

The chert and iron formation clasts are probably from areas with a volcanic affinity, but whether specifically caldera lakes, interflow beds, or other aquabasinal deposits acted as sources is unknown. These clasts are found in the fluvial conglomerate of both the NMB and CMB, and require some syn- or pre-conglomerate aquabasinal source area(s). The CMB's lower (southern) iron formation-bearing aquabasinal units are thought to have been buried by younger fluvial deposits (Section 12.3), thus ruling them out as probable sources. Iron formation in the upper (northern) CMB in Dorothea (outcrops #10, 12) and Leduc (outcrop #79) Townships (Fig. 3.1) are possible intrabasinal source areas.

CHAPTER 11: INTERPRETATIONS OF DEPOSITIONAL PALEOENVIRONMENTS

11.1 Introduction

The NMB (Chapter 2) is interpreted in Sections 11.2 to 11.5. Interpretations of the CMB's lithofacies assemblages are given in Sections 11.6 to 11.15, following the same order that the CMB data was presented in Chapters 4 to 9. The CMB's sandstone assemblage (Chapter 5) is highly variable along strike, necessitating four separate interpretations (Sections 11.7 to 11.11).

Figs. 2.1 and 3.1 show the locations of specific outcrops referred to in the text.

Chapter 12 summarizes the sedimentological and stratigraphic trends in the CMB.

11.2 Interpretation of the NMB: Rejection of the Submarine Fan Hypothesis

It can be difficult to differentiate between fluvial and resedimented conglomerates in small (metres to tens of metres) exposures. Usually the lithofacies assemblages and stratigraphic context allow a fairly unambiguous interpretation, but in deformed terranes with small, scattered outcrops assemblage data and context are less reliable. In the absence of primary clast fabric data, such as imbrication (Walker, 1984a, Fig. 11), rejection of a submarine/subaqueous fan or fan-delta slope origin for the NMB is not as easy as for undeformed strata. Submarine fan deposits have been interpreted as products of braiding (Winn and Dott, 1977, 1979; Hein and Walker, 1982), with similar subaerial and subaqueous processes possibly causing confusion (Hein, 1984).

Throughout the NMB evidence of turbiditic processes is totally absent, except in parts of east Irwin Township (Fig. 1.1), where the previously described (Section 2.6) anomalous graded sandstone and mudstone are present but have not been found interbedded with the typical coarser conglomerate-sandstone assemblage. The anomalous strata may represent regionally minor facies changes (Section 2.6) to a small or large aquabasin.

The NMB's conglomerate-sandstone lithofacies assemblage lacks the following features:

1) abundant, well developed normal grading (fining upward) in conglomerates (Walker, 1975, 1978b, 1984a; Lowe, 1982; Hein, 1982, 1984; Ethridge and Wescott, 1984);

2) graded-stratified conglomerate (Davies and Walker, 1974; Rocheleau and Lajoie, 1974; Walker, 1975, 1978b, 1984a; Hein, 1982; Hein and Walker, 1982; R3-S1 of Lowe, 1982), a turbiditic product;

3) Bouma sequences, or any graded sandstones (Lowe, 1982; Hein, 1982; Walker, 1978b, 1984a);

4) fluid escape structures, such as dish structures (Lowe, 1975, 1982; Hein, 1982);

5) mudstone;

6) deep (up to 10 m) channels (Hein and Walker, 1982; Hein, 1984); and

7) large-scale slide marks, slumps and load structures (Winn and Dott, 1979; Johnson and Walker, 1979; Howell and Normark, 1982).

The absence of the facies and structures cited above argues very strongly against any submarine fan or fan-delta slope interpretation. None of the above features is unequivocal (except perhaps graded-stratified conglomerate), but the lack of all these together is nearly unequivocal, particularly as the NMB has been examined for 60 km along strike.

11.3 Interpretation of the NMB: Rejection of other Paleoenvironments

The NMB strata do not have any features specifically indicative of coastal or shallow marine gravel-bearing paleoenvironments (Nemec and Steel, 1984; Ethridge and Wescott, 1984; Bourgeois and Leithold, 1984; Phillips, 1984; Kleinspehn *et al*, 1984) or of glacial systems (Eyles and Miall, 1984). However, the NMB could possibly be a glacial outwash (fluvial) deposit. Correct but climatically/ environmentally ambiguous lithofacies are present, and there is no known evidence for Archean glaciation. Because of abrasion, glacially striated clasts often lose their striae rapidly in the bedload channels of braided outwash rivers. The author has spent months in modern glacial outwash environments and cannot recall seeing many striated clasts in gravel bars near the glaciers. Clast deformation will further remove such data from the geologic record.

11.4 Gravelly Braided River Depositional Processes and Products

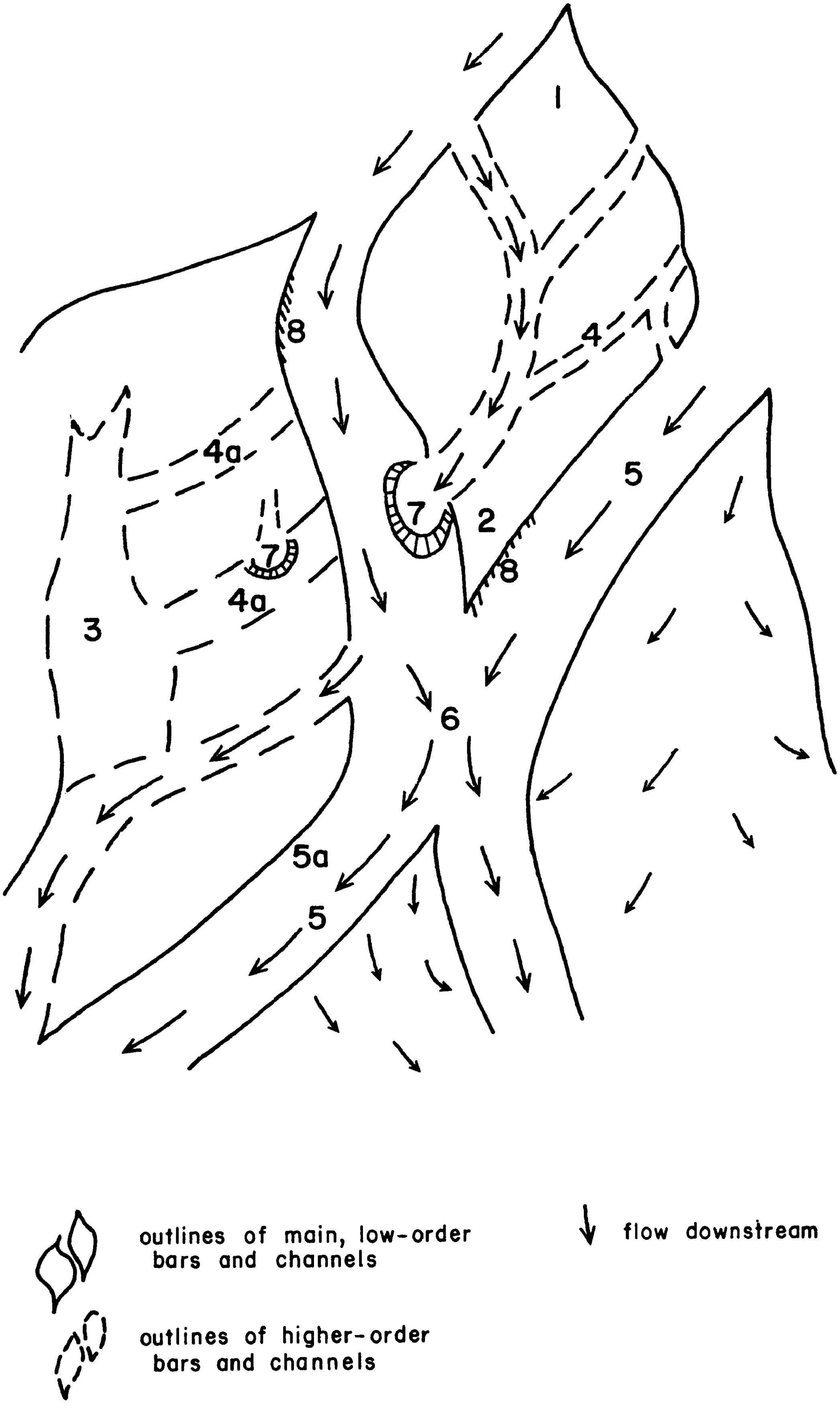
The following summary, based on both a literature survey and personal observations, is designed to account for the features present in the NMB (Chapter 2) and CMB's conglomeratic assemblage (Chapter 6).

Gravelly braided rivers are characterized by steep regional slopes, low sinuosity, strongly fluctuating discharge, intervals of rapid sedimentation, broad and shallow rapidly shifting channels, and submerged and exposed mid-channel bars composed of coarse, non-cohesive bedload (Miall, 1977, Church and Gilbert, 1975). In flow-perpendicular cross-sections the channels have high width : depth ratios (greater than 40; Miall, 1977) and alternate laterally with low-relief gravel bars, convex-up and often irregular bedforms. Bar terminology used below follows Miall (1977) and Collinson (1978).

Longitudinal (and diagonal) bars (Fig. 11.1) occur at all scales, from less than a metre to kilometres in length, with the largest ones probably compound forms (Miall, 1977). Thus a "bar" is usually both part of larger bars and contains smaller bars within its margins (Krigstrom, 1962; Bluck, 1974). Most deposition is actually the modification of older, larger bars by the creation of superimposed, newer and smaller bars and channels, and it is thus largely semantic to differentiate between the creation of new smaller longitudinal bars and the remolding or sculpturing of older, larger bars (see Bluck, 1974, p.219). Williams and Rust (1969) erected hierarchical orders of bars and channels in order to cope with this scale problem. Most of the longitudinal bars described in the literature appear to be about

Fig. 11.1: Schematic plan view of modern gravelly braided river longitudinal bars, and intervening channels. Based on figures in Krigstrom (1962), Williams and Rust (1969) and Bluck (1974, 1979), and personal observations. Scale is diagrammatic and variable; most low-order bars (term of Williams and Rust, 1969) are tens to hundreds of metres long.

1) bar apex (crest or head);
2) bar tail;
3) abandoned main channel (slough);
4) abandoned higher-order channels (chutes)
4a) abandoned transverse channels;
5) riffles;
5a) riffle margin;
6) pools;
7) microdeltas (chute lobes, chute bars); and
8) erosive bar margins.


1
4
8
4a
7
2
5
8
7
4a
3
6
5a
5
outlines of main, low-order bars and channels
outlines of higher-order bars and channels
flow downstream

10-100 m long, with occasional larger ones hundreds of metres long.

Longitudinal bar formation. Reasons for a river depositing gravelly sediment through a loss of competency may be related to extrinsic factors such as declining (waning) flood stage, or intrinsic factors such as channel widening or shallowing (Miall, 1977). Stream power is greater in the deeper main channels than on the adjacent shallower bar surfaces (Boothroyd and Ashley, 1975), the channels being most competent at their deepest points (Miall, 1977). Thus in the deepest active channels the coarsest clasts are present and more erosion takes place (Boothroyd and Ashley, 1975).

Except in debris flows, deposition of all but the finest gravel is usually by traction, forming clast-supported frameworks. The rapid and turbulent transport and deposition do not allow much local sorting, thus sorting of both the sand and gravel populations is generally poor, but this also depends on the inherited characteristics of the detritus supplied to the system.

Ideally, lag of coarser clasts in a channel acts as an obstacle, trapping other clasts and forming a new bar (Leopold and Wolman, 1957), but in most cases (especially because of scale) modification and extension of older, larger pre-existing bars is the rule. Also, the sudden nucleation of small groups of pebbles (quantum bar growth) may be as likely as clast-by-clast accretion.

River channels flow around both the small, newly created bars and the crests (apices) of large pre-existing bars; at such points of flow divergence velocity, slope and depth usually decrease, competency is lost and sediment is deposited, adding to the growing (aggrading and/or laterally accreting) bar (Krigstrom, 1962; Boothroyd and Ashley, 1975; Church and Gilbert, 1975), which may have the lobate outline of a diffuse gravel sheet (Hein and Walker, 1977). Water continues to carry the bed load to the downstream end of the bar during shallowing and partial emergence (Krigstrom, 1962).

Riffles, shallow stretches of supercritical (upper regime) flow, may be on submerged bar tops or flanks or shallow channel floors (McDonald and Banerjee, 1971; Gustavson, 1974; Church and Gilbert, 1975). Deeper water pools separate the riffle dominated reaches. Transverse ribs and stone cells testify to the upper flow regime conditions (McDonald and Banerjee, 1971; Gustavson, 1974; Boothroyd and Ashley, 1975; Koster, 1978). The cluster bedforms of Brayshaw (1984), nucleating about large obstacle clasts, are one of the smallest (less than 1 metre) gravel bar forms.

Bars may grow downstream, sideways (Smith, 1974; Bluck, 1974; 1979, Fig. 15) or upstream. They may fine downstream (Smith, 1970, 1974; Gustavson, 1974; Boothroyd and Ashley, 1975; Miall, 1977; Bluck, 1979) or show no pattern (Smith, 1974). If the downstream portion of a bar is finer grained than the adjacent channel, lateral migration of bar over channel facies may produce a fining-upward sequence (Smith, 1974), and progradation of the bar crest (head) over the downstream end (tail) may produce a coarsening-upward sequence (Bluck, 1979).

Bluck (1979, p.179) wrote:

> "bars are mere sheets of gravel at high
> stage, develop form on the falling stage
> and are dissected upon emergence."

As the flow stage decreases, deeper and more efficient channels are cut into the bar (Church and Gilbert, 1975), parts of the bar become subaerially exposed and flow is funnelled down the smaller (chute) channels that dissect and partly erode the growing bar (Krigstrom, 1962). Some flat bar surfaces may experience sheetflow, but as depth decreases the flow will become confined to channels.

On bar surfaces and at bar margins erosion and deposition are usually simultaneous, with some sites in dynamic equilibrium and others out of phase. Also, during declining flow stage, different bars may synchronously undergo high and low stage conditions, high stage in and near the main (most active) channel and low stage away from it (Bluck, 1979).

The rapidly deposited gravel, facies Gm of Miall (1977, 1978a), tends to be poorly sorted, massive to crudely horizontally bedded, and in the form of gravel sheets (Rust, 1972; Bluck, 1979) (lobes are rarely reported: Gustavson, 1974).

Cross-bedding is rare in proximal stream gravels (Church and Gilbert, 1975; Rust, 1978). Slipfaces may develop at the downstream ends of bars (Rust, 1972; Gustavson, 1974; Smith, 1974; Boothroyd and Ashley, 1975; Hein and Walker, 1977), the downstream avalanche-face progradation producing cross-bedding (facies Gp, Gt of Miall, 1978a) on the bar flank (out into the channel). It is easier to develop cross-bedding if the gravel is fine, sandy, or a bar tip sand wedge is present.

Diagonal bars. These are very similar and gradational to longitudinal bars, and Miall (1977) recommends that they be considered a sub-type of longitudinal bars.

A system of channels draining diagonally (in plan view) across a bar, resulting in lateral accretion preferentially to one side via the migration of riffle and/or foreset bar margins (Smith, 1974; Bluck, 1974, Fig. 7, 1979), depositing gravels and preserving sand and mud left behind in the inactive and less active channels (slough and transverse channels, respectively, of Bluck, 1974, Fig. 7).

Transverse bars. Active during and/or after peak discharges, these produce cross-bedding (e.g. facies Gp, Sp of Miall, 1978a) by avalanching along and at lower levels than the bar crest (Smith, 1974; Boothroyd and Ashley, 1975; Church and Gilbert, 1975; Hein and Walker, 1977; Miall, 1977).

Gravelly low-order channels. Because the large volume of clasts in the shallow river flow have a high resistance (i.e. bed roughness, boundary resistance) to the flow, deposition and selective scouring occur, with the resulting narrower and deeper channels being hydraulically more efficient (Church and Gilbert, 1975). The initiation of a bar causes flow diversion with

> "mutual interaction between the accumulation in the diverging streams and the erosion in the channels which become narrower because of the bars." (Krigstrom, 1962, p.338)

Scouring and waning discharge can dump channel-filling sediment (Church and Gilbert, 1975), which may fine upward (Smith, 1974). At the downstream ends of bars, flow convergence causes erosion in the deeper channels and pools, and erosion of the bar flanks. Deposition will occur in pools where velocity and competency decrease. Selective erosion may leave a gravel lag as an armoured channel floor or sides (Church and Gilbert, 1975). Winnowing and lag can produce a gravelly fining-upward section at a bar margin (personal observation).

The chutes and lobes of Southard *et al* (1984) are small, ephemeral within-channel features, the lobes being very small bar forms but also on a larger scale merely a sculpturing of larger bar tops.

At a small enough scale, that of smaller bars and not entire fans, the most active channels are at the lowest elevations relative to adjacent bars and higher, less active channels (Williams and Rust, 1969). But a channel floor may aggrade to up above the immediately surrounding area, the bars directly adjacent to such a channel acting as levees. Avulsion of these superelevated channels and "levee-bars" may take place, allowing the new channel to follow a shorter and steeper path, scouring new channels and depositing new bars (Krigstrom, 1962; Church and Gilbert, 1975; Miall, 1977).

The above processes can operate on a small scale, or may affect deposition on an entire fan or braidplain (Schumm, 1977).

The shifting channels result in mostly erosional bar margins (Boothroyd and Ashley, 1975; Church and Gilbert, 1975) and complex multistorey deposits (Southard *et al*, 1984) making it difficult to recognize bar-channel contacts in gravelly sections (Smith, 1974), to determine original bar sizes (Smith, 1974; Boothroyd and Ashley, 1975), and lowering the preservation potential of sands interbedded with the gravels (Rust, 1978). The lenticular nature of the beds is due to the great multiplicity of bars and channels present. Frequent erosional truncation by many rapidly deposited units metres to hundreds of metres in width and length, produces thin, flat to low-angled lenses of coarse-grained sediments, with some preserved steeper bar flanks and channel sides, and a lack of regional-scale, sheet-like and correlatable layers.

Sandy higher-order channels. Exposed bar surfaces are dissected by shallow channels, either active or abandoned. These "higher-order" channels are at higher topographic levels, the uppermost ones being active during only the larger flood discharges (Williams and Rust, 1969). The topographic levels between the dissecting channels "represent stages of progressive downcutting" (Miall, 1977, p.9), and are the product of the bars re-equilibrating to the post-flood, low-stage energy conditions. Terraces forming these topographic levels are usually within 0.3 m of each other vertically (Williams and Rust, 1969).

During lower flow stages and at a low enough flow competency thin (tens of cms) beds of sand (with little or no gravel) are deposited as patches or sheets up to tens of metres long on flat bar surfaces and infilling concave-up hollows, troughs and higher-order channels in the bars (Smith, 1974; Gustavson, 1974; Bluck, 1974; Boothroyd and Ashley, 1975). The beds are thus lenticular in cross-section; the stratigraphy is one of lenses and not large-scale (kms or more), laterally continuous sheets.

Sand and mud filter downward into the open matrix of the gravel frameworks, producing a bimodal gravel-sand size distribution. Precipitation may help matrix infiltration and removal of sand from the bar surface (Gustavson, 1974). The infilling matrix may obscure the gravels' crude bedding (Church and Gilbert, 1975).

Channels may fill either laterally or downflow (Bluck, 1974, 1979). As there is more vertical distance available for depositional processes to operate in, flow separation can occur and a greater variety of sedimentary structures can form in the higher-order channel filling sands than for the massive to plane bedded gravels (Church and Gilbert, 1975; Boothroyd and Nummedal, 1978).

Sand funnelled down channels forms upper flow regime plane beds and/or lower flow regime dunes and ripples (Bluck, 1974; Gustavson, 1974; Boothroyd and Ashley, 1975). Small bars may form at the downstream termini of the higher-order channels (Krigstrom, 1962; "chute bars" of Bluck, 1974, 1979). Planar, horizontally and parallel laminated sands, facies Sh of Miall (1978a), are usually interpreted as upper flow regime deposits, but some may be lower regime. Hydrodynamic fluctuations (e.g. velocity) account for the coarser and finer laminae. The dunes (megaripples) produce trough cross-beds, facies St of Miall (1978a); straight-crested dunes (sand waves) produce planar cross-beds, facies Sp of Miall (1978a). Some apparent horizontal plane lamination may in fact be flow-perpendicular sectional views of planar cross-strata (Reineck and Singh, 1980, Fig. 41). Lateral filling of channels forms bar edge sand wedges (Rust, 1972; Boothroyd and Nummedal, 1978), which are internally cross-bedded (facies St, Sp, Sr of Miall, 1978a), and small microdeltas (Bluck, 1974, 1979; Reineck and Singh, 1980, Fig. 286, 287; facies Sp, Sr, Gp of Miall, 1978a). Further deposition or reworking often produces a veneer of ripples and plane beds, facies Sr and Sh, respectively, of Miall (1978a). These may have mud drapes (Bluck, 1974, 1979; Boothroyd and Ashley, 1975; Miall, 1977), facies Fm of Miall (1978a). Most of the sand beds eventually preserved are likely partially truncated, their tops and/or margins removed by erosion, particularly by subsequent high-energy erosive events (i.e. sharp-soled, erosively based beds: Fig. 2.7).

Very low-angle (less than 10°) cross-beds may be:

1) antidune deposits;
2) scour fill with laminae conforming to the scour hollows' form;
3) crevasse splay deposits (e.g. a micro-fan, with a slope), all included in facies Sl of Miall (1978a);
4) tangential toesets of large cross-beds;
5) channel drapes (especially very large channels); or
6) views of planes oblique to bedding in dipping beds of plane laminated sand (a problem of exposure in some cases).

Massive sand (Sm) is thought to represent sudden deposition (e.g. from a waning sheetflood), but can also be merely the too-subtle preservation of structures or beds whose stratification was destroyed during fluid escape.

Much of the sand undergoes episodes of redeposition during the erosional dissection of bar top surfaces at low flow stages. Sand winnowed from higher topographic levels can leave behind a gravel lag or armour (Gustavson, 1974). Later, smaller bedforms may be superimposed on earlier, larger ones (Bluck, 1979). Falling stage erosion may form a reactivation surface, a small-scale angular unconformity within cross-beds, which can have a mud drape (Collinson, 1978, Fig. 3.19; Boothroyd and Ashley, 1975; Bluck 1979); post-erosion foreset development is the reactivation. Reworking of sands during still-stands (e.g. the lowest flow stage) forms sand spits along bar margins (Bluck, 1979).

As always, burial promotes preservation. Sand pods and lenses may be either units deposited as sand patches only cms to a few metres square in area, or eroded fragments of beds previously more continuous, or the discontinuous and tapering margins of predominantly thicker sand beds.

Stratification. Changes in flow stage, stream power and local competency produce changes in clast size. Deposition of coarser and finer clasts on flat bar surfaces forms horizontal to low-angled strata. Transverse ribs ("clast stripes"), lags below small (less than 0.5 m) waterfalls, diverging flow (competency loss dumps gravel at bar crest), and the winnowing of finer clasts along erosional bar margins all produce coarse bands of gravel on bar surfaces.

Thicker massive units, especially those several or more metres thick, should be compound (syn. amalgamated, or multi-storey) sets of beds since it is probably easier to aggrade (stack up) a sequence of massive beds totalling metres thick than to accumulate many such single, extremely thick flood units (deep channel fills or huge bars).

Bed transitions may be sharp or gradual in terms of abruptness of grain size change and/or percentage of gravel per unit (e.g. cm) of section. High-energy gravel deposition often produces sharp erosive bed soles, particularly in finer underlying beds. Erosive soles may be flat, gently inclined or curved, the products of scouring and channelling. Non-erosive deposition of sand on a fairly smooth pebbly surface will produce an abrupt gravel-sand boundary.

Gradational bed transitions defined by a change from clast-supported to matrix-supported gravel or pebbly sand may reflect:

1) non-erosive deposition of sand on an irregular pebbly surface;
2) synchronous deposition of sand and pebbles; or
3) multiple depositional events forming a compound unit (a multistory bed, but separate events not recognizable).

These are less common than sharp contacts.

Pebble bands. In a gravel-dominant system most of the sand will not be pebble-free, and pebbly sands are common. Pebbles deposited synchronously with sand should be small, less than 16 mm (Rust, 1978).

Many pebble bands (gravel lenses, rows or stringers) are lags. Clasts may lag behind in sandy channels following a decrease in flow competency and form a bed or band along a bedding plane or horizon. This implies a net preservation of sand; more sand remained than was eroded by the rapid gravel-depositing events. As an example, surging flows may roll or bounce clasts into sandy areas. Foley (1977) has described gravel lenses that form as "antidune dropout armor".

Residual lag horizons, bands or beds can form when scouring vortices or an increase in flow competency carries away sand but not heavier pebbles. For example, flow re-attachment in the lee of dunes forms scour troughs in which pebbles can be erosionally concentrated (Bluck, 1979, Fig. 4B). As the dunes and scour troughs migrate downstream, a layer of such pebbles may accumulate laterally (downflow), forming a band.

Inclined pebble bands outline channel floors or dune or bar surfaces (stoss slopes, lee foresets). Doeglas (1962, Fig. 23) has provided a photograph of cobbles moving down a bar flank and burying sand in the adjacent channel.

Eolian reworking. Deflation by winds can affect sediment texture (Church and Gilbert, 1975), with "fine sand and silt being winnowed out and wind-generated ripples formed at the (bar) surfaces" (Gustavson, 1974, p.386). Boothroyd and Nummedal (1978) have described distal inactive stream areas with eolian dune fields and wind-tidal flats which produce cross-stratified sands and post-flood laminated muds.

Overbank deposits. Overbank deposition may take place on the elevated and exposed parts of mid-channel bars, forming muddy chute channel-fills, or may be on a floodplain completely removed from the active river tract.

Gravel bars can act as levees (Fahnestock, 1963; Church, 1972; Church and Gilbert, 1975; Bluck, 1979), leading to possible channel fill, aggradation and superelevation of channels, a gravitationally unstable arrangement eventually corrected by avulsion into the overbank (Church and Gilbert, 1975; Miall, 1977). Crevasse splays, with potentially good stratification and sorting, may fine away from the node of avulsion (breach site) and can fine upward due to waning velocity (Church and Gilbert, 1975). Gravel lobes may be present on floodplains (Ritter, 1975). Fine sand, silt and clay settle from suspension in slack water pools (abandoned channels, or sloughs) forming drapes on underlying bedforms (Boothroyd and Ashley, 1975).

Modern braidplains have few muddy floodplain deposits, since fines tend to be flushed through the system in river channels (Smith, 1974; Church and Gilbert, 1975) and/or groundwater flow (Hjulstrom, 1955), and muddy channel-fills tend to be very thin (no deep channels available to fill, only shallow ones), of limited lateral extent, and ephemeral due to erosion and redeposition. Braided fluvial deposits in the ancient record also have little or no mudstone. This

> "paucity of fines ... can be related to ... the extreme vulnerability of overbank flood deposits to later fluvial and eolian erosion, and possibly to removal of fines ... as wash (suspended) load" (Long, 1978, p.313)

Obviously, in the Precambrian there was no vegetation to stabilize muddy overbank areas.

Section 11.12 discusses further muddy overbank processes and deposits.

11.5 Interpretation of the NMB

The lithofacies assemblage Gm, Sm, Sh and Sp (facies codes of Miall, 1977, 1978a; Sm is massive sand) present in nearly all of the NMB conforms to the following facies models for proximal braided rivers, including wet/humid alluvial fans:

1) the Scott type facies assemblage/vertical profile model of Miall (1977, 1978a);
2) The G2 facies assemblage of Rust (1978);
3) the proximal facies of Boothroyd and Ashley (1975) and Boothroyd and Nummedal (1978, Fig.5); and
4) that of Collinson (1978, pp.24,25).

The conglomeratic NMB strata have outcrop-scale Scott type vertical profiles produced by

> "longitudinal bar gravels with sand lenses formed by infill of channels and scour hollows during low water". (Miall, 1977, p.1)

Regarding preservation potential,

> "to be expected in ancient analogues, is an assorted array of laterally adjacent and juxtaposed remnants of bar and channel-fill sediments fortuitously preserved by the shifting channels". (Smith, 1974, pp.219,220)

Sections are composed of usually 90-100% conglomerate, facies Gm. Facies Sh is more common than Miall (1977, Table 5, Fig. 12; 1978a, Table 2, Fig.1) has implied, according to Boothroyd and Ashley (1975, Table 1) and Rust (1978, Table 1). Massive sandstone (Sm), despite being common (e.g. McGowen and Groat, 1971) has been ignored by most authors. Miall intended pebble bands to be covered by facies code Gm (personal communication, 1983), but some authors have called them "Glag" (e.g. Krapez, 1985).

Section 11.4 has described in detail the processes, some essential and others probable to unlikely, that contributed to the deposition and preservation of the NMB's conglomerate and sandstone.

Except in parts of Irwin Township discussed in Section 2.6 and 11.2, the NMB is the product of deposition by gravelly braided rivers which were part of alluvial fans and/or braidplains, larger-scale depositional environments or systems. The bouldery fluvial conglomerate exposures suggest proximality to their source, perhaps as parts of fans. The lack of muddy debris flow deposits may indicate a "fluvial fan" (wet alluvial fan; Schumm, 1977) paleoenvironment, in which fans are produced completely by braided river processes (McGowen and Groat, 1971; Miall, 1977, 1978a, b; Boothroyd and Nummedal, 1978). The term "humid alluvial fan" is to be avoided because of its strong climatic connotations. Sheet-flow fan conditions (Ballance, 1984) can produce sequences of coarse, massive conglomerate with little or no interbedded massive sandstone, such as are present in large parts of the NMB.

The distinction between alluvial fans and braidplains (Rust and Koster, 1984) is largely unimportant to this thesis. Fans are landforms with sedimentological processes and products often identical to those of braidplains (Miall, 1977, 1978a, b), lower-sloping and more two-dimensional environments. Also, fans can easily be distally and gradually gradational to a braidplain (Rust and Koster, 1984, Fig.6).

Very similar lithofacies descriptions and interpretations of other Archean alluvial fan-braided river conglomeratic sequences have been given by Eriksson (1978, 1981), Teal

(1979), Hyde (1980), Wood (1980) and Gordanier (1982). Ojakangas (1985) lists many other studies, including Lawson's (1913) early fluvial interpretation of the Seine Series.

A basal conglomerate. The mafic clasts in outcrop #63A's conglomerate were probably derived from the mafic volcanic belt now adjacent, allowing for possible later transcurrent relative movements (e.g. Williams, 1985). Thus the mafic clasts are thought to have been derived from the belt *below*, and tops at this site are to the north. The contact is interpreted to be a faulted or sheared unconformity and the outcrop's conglomerate a basal horizon of the NMB.

In the old-fashioned terminology of the early workers, this is a "Timiskaming-Keewatin" contact, and proves the Timiskaming (conglomerate) to be younger at this site.

One problem exists: the mafic clasts are of a coarser lithology than the pillowed mafic volcanics below (south of) the contact. Aside from poor exposure and possible structural complexities, this may be due to lithologic variation originally present in the mafic volcanics (models of Dimroth *et al*, 1985).

The direction of transport of these mafic clasts is unknown. The sedimentological characteristics present should negate any purely deformational origin for the mafic clasts (i.e. a fault breccia).

In terms of the regional geology, could such a basal contact between regional-scale lithologic belts be isoclinally folded, or does the basal conglomerate reveal the sedimentary polarity (tops to the north) of the NMB?

Where the NMB's southern contact has been seen elsewhere, in the Townships (Fig. 1.1) of Irwin (outcrop #4), Walters (outcrop #140), and Leduc (outcrop #134), it appears to be a fault contact, based on relatively extreme clast elongation, folding, linear topography, and Mackasey's (1975, 1976) maps. The sharp contacts at outcrop #35 (Leduc Township) are ambiguous.

11.6 Interpretation of the CMB's Fine-Grained Assemblage

Massive and laminated mudstone and siltstone were deposited by the settling from suspension of fines, with minor silty or sandy traction deposits (laminae, ripples). Most of the graded sandstone beds are thin-bedded turbidites. Massive, ungraded sandstone may be either sediment gravity flow or suspension (e.g. settled from eolian input) deposits.

The two anomalously thick sandstone beds in outcrop #53 are probably sediment gravity flow deposits (e.g. turbidites, fluidized sediment flows, grain flows: Rupke, 1978, Figs. 12.4, 12.5, 12.11).

Currents or waves could have formed the dunes or swales (trough cross-beds or swaley cross-strata), one graded bed, and fining and thinning-upward laminated sandstone-mudstone sequences (only tens of cms thick) preserved in outcrop #72. The swaley (outcrops #89, 72) and possible hummocky (outcrop #72) cross-strata should record storm wave activity (Walker, 1984b).

Magnetite and hematite iron formation laminae are usually interpreted to be chemical precipitates [that] settled from suspension.

The CMB's southern fine-grained assemblage is largely the product of sediment gravity flows (sandy turbidites) and fines settling from suspension in an aquabasin. Most of the assemblage was deposited below wave base in a prodeltaic, submarine fan or "abyssal" plain setting. Two outcrops display evidence of storm wave activity. The tectonized nature of the assemblages' exposures and sequences limits their interpretation.

Interpretation of the anomalous conglomerate. The unusual oligomict, fine-grained conglomerate is probably not related to the coarser, polymict conglomeratic assemblages elsewhere in the CMB, and may be a distal facies of the Coral Lake volcaniclastics (see Chapter 9). The two small exposures of this facies have no environmentally diagnostic criteria.

11.7 Interpretation of the CMB's Sandstone Assemblage: Sandra Township

There is no inequivocal paleoenvironmental interpretation for the rippled sandstone in outcrop #18 (see Section 5.2); its darker layers are slightly coarser, higher energy sands deposited within a silty sand background.

The interpretations given below are for the couplets and associated strata described in Section 5.2.

Hydrodynamic interpretation. Minor amounts of abnormally high-energy deposition processes are indicated by the medium-scale cross-stratification, rare pebble bands and granule laminae, very coarse sand, and scouring of laminae. The cross-bedded and plane laminated, medium to fine-grained beds tens of cms thick were deposited by sinuous-crested dunes (trough cross-beds), straight-crested sand waves (planar cross-beds), and flat areas (plane laminated sand). Pebble bands may be either directly deposited by traction (a lag due to a decrease in flow competency) or residual erosive lags (due to an increase in flow competency with removal of sand but not gravel). Imbricate granules should be traction deposits, particularly those bands displaying contact imbrication.

The sharp, sometimes erosive bases of Type B units reflect high-energy inflow relative to the underlying Type A units. The consistent style of Type B's stratification was produced by inflow of rippled sand, with regular turning off of the current shown by the mud laminae and flasers. Clusters of closely spaced mud laminae represent microfluctuations in sediment supply and deposition. The bulk of the Type B units are higher-energy deposits than Type A units, as shown by the former's coarser sand and larger ripples. The size of the ripples is probably a function of grain size and not directly related to current velocity.

Transition zones are relatively minor horizons; B and A-style sedimentation clearly dominated deposition of the couplets.

Type A units' stratification style is due to constant fluctuation about the lower flow regime boundary between plane laminae and ripples, producing the intimate mix of the two. The two anomalous, sharp-based, thicker massive beds within Type A units are likely grain flow or fluidized sediment flow deposits.

The very rare symmetrical ripples represent the minor influence of waves.

Some rapid sedimentation is indicated by sand dikes (dewatering, upward injection of sand) and load structures (ball-and-pillow, flame structures).

The Type B-A couplets are of variable thickness; identical processes produced thin (0.4 m) and thick (4 m) couplets. The relation of couplet thickness to proximality to a source is unknown. Grain sizes and sedimentary structures reveal waning energy conditions upwards through each couplet, the result of regular pulsating (section-scale) and progressively weakening (couplet scale) inflow events or phases. The time scale of the pulsating and waning is unknown. Inflow was consistently focused in the same direction, as shown by the broadly unimodal paleocurrent pattern.

In ascending outcrop #21A's coarsening-upward section, several features suggest a shallowing trend. The subtle coarsening of the sandstone is concomitant with the upward loss of its mud fraction. There is less mud in the coarser

(B-style?) sandstone, and a lower percentage of the section is composed of finer (A-style?) sandstone. This suggests the winnowing away of mud and finer sand. The general lack of well-defined, rhythmic Type B-A couplets probably reflects a greater variety of processes, which would be expected in shallower waters more influenced by waves and possibly tides. Some rare wave ripples are present.

Interpretation of delta front couplets. The following is partly taken from Devaney and Fralick (1985, pp.129,130).

The couplets are interpreted to be deltaic distributary mouth bar and/or distal bar deposits by the following process of elimination. The couplet-dominated sandstone lithofacies assemblage has:

1) no turbidites or other features typical of submarine fans;
2) no criteria strongly suggestive or diagnostic of shallow marine shelf or prodelta environments;
3) a rhythmic Type B-transition-Type A sequence, which does not suggest fluvial channels, in which the couplets would be susceptible to truncation by the initiation of new erosively-based channels;
4) a lack of other features suggestive of fluvial processes (e.g. the typical fining-upward profile and stratification of meandering stream sequences);
5) a lack of small-scale (metres, tens of metres) coarsening-upward sequences, as would be produced by interdistributary bay fill sequences (Elliott,

1974, 1978a);

6) couplets too thin to represent deltaic lobes; and

7) a broadly unimodal paleocurrent pattern that does not suggest either meandering stream or shallow coastal (non-deltaic) paleoenvironments, which have polymodal to widely varying patterns.

One would expect mudstone and turbidites towards prodeltaic bottomsets, and fluvial facies as delta plain topsets; the couplets fit conceptually in between, in a delta front or slope paleoenvironment.

The best exposed section's coarsening-upward character is typical of progradational deltaic sequences.

It is not possible to determine whether distributary mouth bar or distal bar processes dominated. Distal bars could easily have been supplied with much coarser input than is the norm for most examples in the literature.

The best description found by the author of such a likely fluvially dominated delta front setting is by Elliott (1978a, p.128). Briefly, sand is supplied from river mouths by floods and regular inflow. The flood beds usually have planar erosive bases, waning flow sequences, and rippled caps, the results of traction currents flowing downslope as underflows. Above wave base, post-flood waves can rework the sand and produce ripples and laminae in the background muddy silt. Near the top of the slope sequence, coarser, more thickly bedded sand should dominate.

The only problem the description cited above raises is: is it the couplets or the individual beds within the couplets that are the products of single-event sediment incursions such as floods? It seems unlikely that couplets up to several metres thick with mud drapes in their lower parts could be the result of single flood events, but the precise depositional processes and their temporal scale are unknown. For instance, is a long term, pulsating flood a viable scenario?

Type B units were probably produced by the repeated aggradation of flood-induced incursions/input of sandy rippled beds and massive gravity flows, both as beds cms to tens of cms thick. Deposition of each bed was followed by possible reworking and finally a capping by quiet water (settling from suspension) mud drapes and laminae. Eventually deposition of finer, more "distal" sand took over, forming the Type A units. It is not known whether the finer Type A units are more distal in an axial-to-lateral sense (Elliott, 1978a), as a result of distributary mouth and channel avulsion and/or lateral migration, or in a proximal-to-distal sense, with the Type A units representing finer grained background sedimentation periodically invaded from up the delta slope by coarser Type B sands. Also, whether the couplets were in or laterally adjacent to any subaqueous channels incised in the delta slope is unknown. Many submarine topographic features of delta fronts are of a much larger scale than the exposures studied, resulting in the lack of an obvious context of the couplets to any other hypothesized delta front subenvironments.

As discussed below, either a periodic or constant wave influence would be significant as well. The pebbly lags and cross-bedded and plane to low-angle laminated sandstone beds may be products of either abnormally high-energy fluvial inflow or storm wave action.

The couplets are part of a coarsening-upward sequence on two scales. Firstly, the couplets are an integral part of the best exposed section, an approximately 100 m thick subtly coarsening upward sequence. Secondly, this 100 m section is near the middle of the CMB, which is here more than 1 km thick, below (south of) the fluvial conglomeratic part, which also displays only north-topping beds.

The couplets are finer grained and better sorted than the coarse to fine-grained, moderately to poorly sorted sandstone interbedded with the fluvial conglomerate in the section above, as might be expected of a more distal deltaic succession. Also, shallow water wave reworking could have aided in improving the sorting of the sands.

It is difficult to find analogues to the couplets because the literature on deltas is somewhat biased towards the modern Mississippi delta and Phanerozoic bioturbated sequences. Vos' (1981a) subaqueous distributary channel Facies 2 is vaguely similar to Type B-A couplets. Facies 1 and 2 of Dimroth *et al* (1982a), from the Archean of the Abitibi Subprovince, are also somewhat similar, but still poor as direct analogues.

There is no evidence for a fan-delta interpretation, other than the likely presence of braided rivers on the delta plain.

The possible influence of shoreface processes. The couplets have grain sizes, sedimentary structures, and internal stratification (bedding on cm to tens of cms scales) similar to those of some lower shoreface deposits described by Eriksson (1979, p.165), Vos (1981b), McCubbin (1982, Fig. 23), and Dimroth *et al* (1982a). However, such a subenvironment does not account for the rhythmic nature of the couplets and their repetition in the sequence.

The lower shoreface is an area of subtidal wave and storm deposition, with background settling of fines from suspension (Elliott, 1978b; McCubbin, 1982; Reinson, 1984). Because they occupy much the same shallow subaqueous area in terms of water depth and proximity to shore, a delta front may sometimes function as a shoreface, and vice versa. Textbooks largely ignore this point, except to remark on the influence of storm events, and the reworking of distributary mouth bars into sheet sands in wave-dominated systems.

As fluvial input processes diminish in importance, wave and/or tidal influences increase. The couplets have no recognized unequivocal tidal features. In a delta system significantly affected by both river inflow and waves, there is a possible constant overlap of shoreface and delta front processes during the formation of sedimentation units. A constant and continual interplay of processes could create a hybrid influence more than an alternation of wave and then fluvial processes. Thus lower shoreface hydrodynamic processes, with the minor influence of adjacent subenvironments (upper shoreface, and the transition zone to offshore) may account for some small-scale (less than 1 m) features of the couplets, within the larger scale (in section) context of distributary mouth and/or distal bars.

Limitations to the deltaic interpretation. The outcrop observations are small-scale, lacking in information on lateral facies changes. The regional context is possibly suspect, because of both the CMB's scattered exposures, and possible structural disturbances. The inter-pretation is thus biased towards the well-exposed vertical profile.

The lack of good paleocurrent data is problematic, but is typical of Archean terranes (Ojakangas, 1985). It is difficult to recognize in outcrop ripples in the B-style coarser sandstones where not outlined by flaser drapes. Muddy bedding planes, which weather out preferentially, may possibly have differently oriented ripples than non-draped, mud-free ripples. Ripples on bedding planes are not always well-defined, because of possibly complex initial morphologies, or later partial erosion. The three-dimensional exposure is generally not good enough for detailed paleocurrent and bedform analysis.

11.8 Interpretation of the CMB's Sandstone Assemblage: Walters Township

The sandstone assemblage in Walters Township (Fig. 3.1, Section 5.3) has no paleoenvironmentally diagnostic characteristics at all. It may be largely fluvial, as it is along strike from exposures kms away in Leduc Township that are thought to be of fluvial origin (Sections 11.10, 11.11).

11.9 Sandy Braided River Processes and Products

The deposits of sandy braided rivers are characterized by, quite simply, a lot of cross-stratification. This is produced by the linguoid bars (transverse bars of Smith, 1970, 1971; sandwaves of Collinson, 1978), dunes and ripples that dominate the active river tracts.

Linguoid bars have high length : height ratios, continuous crest lines, and a lack of lee-side scour, versus dunes with low length : height ratios, sinuous and discontinuous crests, and scour troughs (Collinson, 1978). The former produce planar cross-strata (facies Sp), the latter trough cross-beds (facies St). Ripples are facies Sr, and plane laminated sand is facies Sh (facies codes of Miall, 1977, 1978a).

The following summarizes the processes and products of very shallow braided rivers; rivers with deeper channels and more topographic differentiation are described separately as the South Saskatchewan model. Like the coarser braided rivers described in Section 11.4, parameters such as high slope, rapid discharge fluctuations, abundant sand and non-cohesive banks (Walker and Cant, 1984) are important influences.

High stage conditions. High stage (flood) flow is broadly divergent over submerged, lobate linguoid bars (Collinson, 1970, 1978; Smith, 1971). These are usually up to 2 m thick, dip gently upstream, and advance downstream via avalanche face (slipface) progradation (Miall, 1977). Areas between bars may undergo scour and fill (Collinson, 1978), resulting in trough cross-bedded channels with basal lags (Cant, 1978).

Sediment is transported across the bars by smaller-scale ripples, sandwaves, dunes, plane beds and antidunes, resulting in lateral and downstream bar growth (Smith, 1971).

Erosion can undercut islands and floodplains, the caving banks producing muddy intraclasts (Cant, 1978).

Falling and low stage processes. Many of the basic processes of braiding in sandy rivers are broadly similar to those in gravelly braided rivers. Linguoid bars are modified by currents, waves and wind (Collinson, 1970), re-equilibrating them to lower energy flow stages.

Most authors note that as flow stage drops, bars become emergent and thus divert flow around them, causing braiding. Waning flood deposits may cause local channel aggradation and preservation (Collinson, 1978; Walker and Cant, 1984). The exposed or shallowly submerged (e.g. less than 1 m) bars are then dissected by higher-order (term of Williams and Rust, 1969), bar top channels (Collinson, 1978; Blodgett and Stanley, 1980), with lags possibly lining the floors of such chutes (Miall, 1977). Flow confined to channels erodes bar margins, producing low-angle erosion surfaces; further bar growth makes these reactivation surfaces (Collinson, 1970, Figs. 21, 27). Very shallow flow can plane off bar tops, forming plane laminated beds and local pebble lags (Blodgett and Stanley, 1980).

Lateral accretion and expansion of bars is often via microdeltas (facies Sp) building out from bars as lobes at the downflow termini of higher-order (chute), bar top channels (Collinson, 1970, Fig. 12; Smith, 1971; Boothroyd and Nummedal, 1978, Fig. 11c), which can be a form of bar front progradation (Blodgett and Stanley, 1980). Collinson (1978) suggests that a rapid falling stage will produce low-angle erosion/reactivation surfaces, versus a slow falling stage's bar dissection via bar top channels and microdeltas. Note that microdeltas are actually smaller-scale linguoid bars superimposed on their larger parents.

The low stage channel floors and bar surfaces are covered with small dunes, sandwaves, ripples and plane beds. Waves may terrace and remold bar margins (Collinson, 1970).

Convolute laminae form via liquefaction and flow of sediment (Reineck and Singh, 1980, p.89) and numerous other hypothesized processes. Coleman (1969, p.218) interpreted it to be produced by "increased shear stress on the bottom caused by a sudden rise in turbulence" in his study of a large, deep, sandy braided river.

Most of the mud is flushed through these high energy river systems as suspended load (Long, 1978; Walker and Cant, 1984). The mud that is present will settle from suspension in abandoned bar top channels and slack water areas of main channels (e.g. mud-draped slipfaces and rippled toesets; Collinson, 1978; facies Fm, Fl of Miall, 1978a). The resultant rare, thin and laterally impersistent mud lenses are usually reworked into intraclasts (Smith, 1970) via later desiccation and transport of mud chips, scouring, and bank caving. Any floodplains developed would have a very low preservation potential owing to large-scale (avulsion-controlled) and small-scale lateral channel migration and erosion.

On both exposed bar surfaces in the active river tract and inactive braidplains, winds may form dunes, ripples and plane beds, with rippled interdune runoff areas (Boothroyd and Nummedal, 1978). Massive eolian sand may accumulate in subaerial scour troughs, buried and preserved later by dunes (Collinson, 1970). More distal wind-tidal flats should also be cross-bedded, rippled and plane laminated, possibly with muds (Boothroyd and Nummedal, 1978).

The intermediate and low stage bar top channeling, microdelta growth (lobe accretion), and merging and superimposition of bars and bedforms lead to a complex and fragmentary record of deposition and erosion (Collinson, 1970; Smith, 1971; Boothroyd and Nummedal, 1978).

The South Saskatchewan model. The South Saskatchewan type of facies assemblage and profile model (Miall, 1978a; Cant, 1978, 1982; Walker and Cant, 1984) varies from the account of sandy braided rivers given above in that more trough cross-bedding is produced in deeper channels.

Channels floored by dunes (trough cross-beds) and lags separate linguoid bars (planar cross-beds) and sand flats (compound bars) at higher elevations in the active river tract (Cant, 1978). Cross-channel bars, a type of linguoid bar, form as microdeltas, from flow divergence over bars, and as larger bars oblique to the main braid channels (Walker and Cant, 1984). Sand flats 50 m - 2 km long are mantled with smaller bedforms, thus

> "The sand flat deposits comprise several planar crossbed sets along with minor amounts of parallel lamination and a thin veneer of small cross-beds and ripple cross-laminations." (Cant, 1978, p.637)

The idealized profiles produced (Walker and Cant, 1984, Fig. 17) are capped by vertical accretion floodplain deposits: mud, rippled sand, eolian sand. Such fining- and thinning-upward sequences may possibly be confused with those of meandering rivers. Braided sequences should have

> "more planar crossbed sets ..., more irregularities in grain size, trends, and in many cases less fine overbank material." (Cant, 1982, p.126)

The capping overbank muds are prone to erosion by the next, often erosively based, sequence. The rapid lateral migration of braided channels preserves fewer overbank deposits than in meandering sequences.

11.10 Interpretation of the CMB's Sandstone Assemblage: Unit A, Leduc Township

Conformity with sandy braided river facies models. Unit A's sandstone assemblage was deposited by sandy braided rivers. The processes that operated have been outlined above in Section 11.9.

Despite the lack of good three-dimensional exposure of Unit A's cross-beds, the assemblage best fits Miall's (1977, 1978a) Platte type facies assemblage and profile model, which represent non-cyclic sections rich in superimposed linguoid bars (planar cross-beds) deposited in "very shallow rivers or those without marked topographic differentiation" (Miall, 1977, p.48). The D1 profile of Boothroyd and Nummedal (1978, Fig. 5) is similar. Blodgett and Stanley (1980) suggest that thinning-upward sequences having a basal, high stage, thick planar cross-bed overlain by lower stage cross-beds, ripples and a mud drape may be more typical of Platte type conditions.

Where stacked planar cross-sets overlie trough cross-beds in Unit A, small South Saskatchewan profiles may be partly exposed. Miall's (1977, 1978a) South Saskatchewan type cyclic profile, Cant's (1978, Fig.5) idealized facies sequences, and Rust's (1978) S II facies assemblage are fining- and thinning-upward sequences rich in trough cross-beds and, according to Miall (1978a), containing less than 10% of gravelly section.

Rather than conforming to any specific ordered sequence, Collinson (1978, p.28) advocates

> "a random interbedding of trough and tabular sets with occasional interbedded units of ripple cross-lamination".

Unit A has no features strongly suggestive of a dominantly eolian origin (e.g. Hunter, 1977, 1985; Ahlbrandt and Fryberger, 1982). Intermittent eolian reworking during low flow stages could have been important in removing silt from exposed sand and producing a better-sorted sandstone.

Parts of Unit A could be shallow marine, an environment which can also be characterized by abundant cross-bedded sand (Elliott, 1978b; Johnson, 1978; Reinson, 1984; Walker, 1984b). Rapid facies changes over a low-sloping surface could easily produce thin, interfingering shoreface, distributary mouth bar, fluvial and eolian facies, all rich in cross-bedded sand. For example, Hine and Boothroyd (1978) have outlined models for fan-delta shorelines, with beach facies up to a few metres thick situated above fluvial or channeled barrier spit strata and below fluvial, eolian or wind-tidal flat facies. Apparent paleocurrents in Unit A define a broadly unimodal to the west trend, the unimodality being typical of braided fluvial systems, as opposed to the polymodal or widely dispersed paleocurrent patterns more suggestive of shallow marine (or aquabasinal) conditions (Long, 1978; Long and Young, 1978; Miall, 1984, pp.261-266). In Phanerozoic sequences the distinction between normally bioturbated and/or fossiliferous shallow or coastal marine strata and mostly non-bioturbated fluvial sandstone is usually much easier.

<u>Coastal facies</u>? The closely spatially associated lithofacies suite of:

1) apparent herringbone (bimodal-bipolar) cross-sets;
2) a 1.2 m thick horizon rich in plane laminae; and
3) rare mud laminae and flasers

in outcrop #100 suggest a marginal marine paleosubenvironment. The herringbone cross-sets should be an intertidal product (e.g. tidal inlets, flood and ebb deltas, tidal creeks) as they are the classic indicator of tidal conditions (Reineck and Singh, 1980, p.454; Miall, 1984, p.165). Plane laminated sands would, in such a setting, be widespread, particularly as foreshore (beach) or washover fan deposits. Mud laminae and flasers represent ettling from suspension during slack water periods.

Each of the above features in equivocal. Herringbone cross-bedding can be produced in rivers (e.g. Miall, 1984, p.165; Alam *et al*, 1985). Large lakes can have a small tidal range. Storms and seiches can direct cross-sets upstream within coastal distributary rivers. Plane laminated sand can form in virtually any environment, and is a fairly common fluvial facies. Mud laminae and flasers are similarly widespread.

<u>Unit A's coarsening-upward trend</u>. This is based on stratigraphic tops from cross-bed and ripple foresets/toesets, a slight upward increase in coarse sand, and a well-

defined coarsening upward of the pebbles and cobbles.

If the section through Unit A is even approximately continuous, the coarsening-upward trend implies a progradation of increasingly more competent (higher energy) river channels. This was probably concomitant with an increase in paleoslope (i.e. more proximal conditions); paleoclimatic change may also have been an influence. Note that the possible coastal facies are at the inferred base of Unit A's subtly prograded sequence.

11.11 Interpretation of the CMB's Sandstone Assemblage: Leduc Township

The fluvial interpretation of this assemblage's outcrops #41 and 76 is very similar to that outlined in Sections 11.9 and 11.10. Planar cross-beds were deposited by sandwaves (linguoid or transverse bars; Collinson, 1978; Miall, 1977). Gravel overpassing (process of Allen, 1983) the tops of these sand bars resulted in pebbles avalanching down the foreset slopes and finally resting on the foresets and toesets.

The massive sandstone beds are either sheetflood deposits, or units with obscured subtle structures and stratification.

The pebble bands are lag deposits; lag due to increasing or decreasing flow competency has been discussed in Sections 11.4 and 11.7. The thicker conglomerate beds are either thin channel floor or bar deposits. Beds wholly matrix-supported may be stacked (amalgamated) pebble bands, or, more likely, extremely poorly sorted, non-bimodal (sand deposited synchronously with pebbles) deposits.

The above processes and products and their relative abundance (Fig. 5.21) suggest a sandy braided river paleoenvironment, with a vertical profile resembling that of the Platte type model of Miall (1977, 1978a), with a possible significant sheetflood influence (Bijou Creek type profile of Miall, 1977, 1978a).

As in the interpretation for Unit A (Section 11.10), interpretation of a shallow marine setting cannot be ruled out for these exposures. The broadly unimodal paleocurrent data, although of poor quality, do not suggest shallow marine conditions (see Section 11.10).

In outcrop #41, toeset concavity and sharp (erosive?) south bed contacts (soles) give tops north.

Significance of the sandstone grain sizes. The finer, better sorted beds in outcrop #41 are products of local fractionation via winnowing from a coarser and more poorly sorted parent population. This parent population is the coarse to fine-grained, moderately to poorly sorted, pebbly sandstone. The abundant coarse to medium-grained beds are probably also part of the parent population. Selective transport of the finer sand grains removed that finer sand, leaving pebble bands (residual lags) associated with the coarser sand, and deposited downflow (downstream?) the derived subpopulation of fine, well sorted grains. These easily erodable finer units were preserved as beds only a few cms thick. The plane laminated sandstone, with its coarse-medium sand laminae versus its medium-fine ones, represents a partial hydrodynamic fractionation of the sand.

11.12 Interpretation of the CMB's Conglomeratic Assemblage

The conglomerate and interbedded sandstone in the CMB's conglomeratic assemblage (Chapter 6) are mostly very similar (to identical) to those in the NMB (Chapter 2). Since they often contain the same suite of lithofacies, facies Gm, Sm, Sh, Sp, and pebble bands (facies codes of Miall, 1978a), the interpretation of the CMB's conglomeratic assemblage is generally the same as that given previously for the NMB in Sections 11.2 to 11.5. Thus gravelly braided rivers deposited most of the northern CMB (Table 6.1, Figs. 3.1, 3.2) in the form of Scott to Donjek type profiles (models of Miall, 1978a).

The conglomeratic assemblage's interesting and unusual features are interpreted below.

Conglomerate facies. The finer grained and better sorted beds suggest a braided river paleoenvironment more distal than the NMB and some other coarser parts of the CMB. These finer conglomerates are products of the selective sorting and downslope movement of a fine fraction from the available gravels. This implies declining stream power and competency and, most likely, a correspondingly lower paleoslope. Abnormally fine conglomerate beds were likely winnowed from coarser gravel and locally re-deposited in channels or on bars.

Turbulent flow over a channel floor with high bed roughness produced vortices which scoured a stream bed, leaving a concave-up scour hollow (Fig. 6.11) that later filled with sediment.

The scour in outcrop #13 (Fig. 6.10) is concave up (north). The conglomerate bed has a sharp erosive sole. Rip-up clasts of the local siltstone are present in the sandstone overlying the conglomerate. The rip-ups and sharp sole suggest a scour channel and not a large load structure. Scouring formed a channel and removed muddy silt clasts from upstream. Pebbles were transported and deposited during the scouring event, soon followed by sand with rip-up clasts. Later deposition of the section above caused loading and soft sediment deformation.

Stratified matrices are produced by the infilling of open, clast-supported gravel frameworks by sand (Smith, 1974). Fine and coarse sand infiltrated between the framework clasts at different times, forming different layers. Granule-rich spots within the matrix are products of coarse, probably winnowed fractions of the pebbly sand supply being deposited by either sieve action (infilling a clast framework) or granules being trapped between larger clasts at the sediment-water interface (e.g. sand shadows).

The possibly cross-bedded conglomerate in outcrop #83 would be the result of slipface lateral accretion or downstream progradation of a transverse bar or the margin of a longitudinal bar (Miall, 1977; Hein and Walker, 1977).

The coarse-tail graded bed in outcrop #70 reflects a decrease in stream competency, via either channel fill or longitudinal bar aggradation, and may represent a waning flood sequence (Miall, 1977). Both fluvial gravel bars and their adjacent channels can produce vaguely graded units (Smith, 1974; Hein, 1984). Similar beds/units in the NMB and CMB are usually northward fining and are near to beds with reliable north-younging criteria; this would suggest the beds described for outcrop #31 are also fining-upward ones.

The wedging out of some conglomerate beds is a small-scale facies change, representing the lateral thinning of bars or channel fills. In Fig. 6.5, lateral tapering and fining down a paleoslope record a finer subpopulation of gravel winnowed from an adjacent, higher bar surface.

Sandstone facies. Modes of formation of pebble bands have been outlined in Section 11.4.

Beds that are coarser north of east-striking sharp bed contacts (e.g. Fig. 6.25) suggest erosive soles of higher-energy sedimentation events. Such a stratification texture gives tops to the north; no south-facing beds are found using this criterion. Coarser sand may travel and be deposited with pebbles, producing bed contacts marked by a change to coarser sandstone with a vaguely defined pebble band(s).

Cross-lamination in contact with underlying clasts is the product of ripples burying pebbles, a process commonly seen on many contemporary gravel bars during low river stages. On topographically higher bar levels sand and mud are winnowed from between pebbles and are deposited as close as only metres away on lower bar levels. This local reworking represents an equilibrium adjustment of sediment and bedforms to a lower energy river stage.

Fluvial conglomerate beds are often erosively based (Fig. 2.7), which accounts for the planar cross-bed in outcrop #53 having a sharp north bed contact versus its less sharp south contact, the latter being the result of sand passively burying a pebbly channel floor. Note that this bed contains three different criteria (foreset concavity, down-slope aspect, bed sole) for stratigraphic tops to the north.

Outcrop #65's reactivation surface (Figs. 6.17, 6.18) is a small-scale angular unconformity within a planar cross-bed. Subaerial exposure partially eroded microdelta or linguoid bar foresets, and later downstream or lateral accretion of slightly finer sand continued to build the bedform via slightly steeper foresets (Collinson, 1970).

Fig. 6.19 shows the processes hypothesized for the unusual pebbly sandstone bed described. The pebble band is interpreted to be a residual lag, an erosional enrichment in the coarsest fraction due to the winnowing away of sand but not pebbles. The (winnowed?) sand above the pebble band is finer grained and better sorted than below it, as might be expected of a subpopulation of grains derived from a more poorly sorted parent population.

A more random origin is also possible: the pebbles may have rolled into a sandy channel or hollow on a gravel bar surface, but this does not account for the pebbly-non-pebbly polarity of the bed.

Mudstone facies. Mudstone may have been deposited in fluvial overbank settings such as ponds in abandoned channels, often forming the last stage of a channel fill sequence (Cant, 1982), or in larger floodplain ponds or lakes and deltaic interdistributary bays. The plane laminae of mudstone beds and intraclasts and the graded laminae of the muddy intraclasts reflect processes in such overbank sites, away from active channels and influenced by the low-energy settling from suspension of muds and minor inflow (e.g. underflows, turbidity currents) of finer sand. Input to overbank areas would be via flooding, avulsion and/or crevasse splays, the breaching of gravel bar flanks acting as levees (Church and Gilbert, 1975; Bluck, 1979).

At a small scale, laterally shifting erosive channels will erode away most or all of the muddy overbank deposits in channel fills near an active river tract, but will not affect any laterally distant floodplain deposits present. It is when avulsion and the subsequent lateral shifting of active channels (possibly the entire river system or tract, if the node of avulsion is proximal enough: Leeder, 1978, Fig. 1) are on a fan or braidplain scale that floodplains are threatened by fluvial erosion. The re-occupation of an abandoned channel or the invasion of a new area on the floodplain/braidplain could remove part or all of the uppermost sediments there, producing rip-up clasts via scouring and bank caving. Because the lateral shifting of erosive channels is common at both the mall and large scales described above, very little mudstone will be preserved in a gravelly braided river system ("system" includes any floodplains). Eolian erosion is also a factor (Long, 1978); winds howling over unvegetated braidplains can transport clouds of silt and sand.

It is likely that not much mud was present at the depositional sites: mud carried in suspension may have bypassed the more proximal river stretches (Long, 1978), and may have been flushed out of the buried sand and gravel by groundwater flow (Hjulstrom, 1955).

The northern mudstone unit's lithology, extent and apparent thickness suggest an interval of muddy aquabasinal deposition; the thickness may be structurally influenced. Thick mudstone intervals with or without associated iron formation can also result from submergent (transgressive) aquabasinal conditions.

In the study area, mudstone bed fragments (rip-up clasts, intraclasts) are preserved more often than entire beds, showing the bed fragments to have a better preservation potential in this braided river paleoenvironment. Partial or no preservation following initiation of the next overlying and often erosionally based sequence accounts for the rarity of these typically thin beds.

Flood-deposited rip-up clast beds in outcrops #20 and 56 were deposited by the following processes. High-stage river flow (a flood) scoured away muddy overbank deposits. Caving away of a muddy river bank is also a possibility. The flood was competent enough to carry the cobble to boulder size mud blocks (rip-up clasts, or intraclasts), the former bank or overbank deposits, for a time. Rapid deposition is indicated by the matrix-supported lower part of the bed. Following this dumping of clasts and matrix, which froze in place immediately, preserving the matrix-supported texture, the flood waned enough to deposit mud clasts by traction, forming the clast-supported upper part of the bed, burying and preserving the lower part. Because the large mud clasts are intraformational they were not used in collecting the D10 measurements.

Outcrop #13 nearby contains excellent evidence of scouring and production of local rip-up clasts (see above).

Unit B. Donjek type facies assemblages and vertical profiles should have fining upward sequences of clast-supported conglomerate (facies Gm; 10-90% of section), cross-bedded sandstone (facies St, Sp) and, if not truncated by subsequent erosionally based beds/sequences, mudstone caps (facies Fl; model and facies codes of Miall, 1977, 1978a). These are the products of rivers that are "braided with marked differentiation of topographic levels" (Miall, 1977, p.45).

Unit B's conglomerate-sandstone assemblage fits the Donjek model somewhat. Outcrop sections (e.g. Fig. 6.22) show subequal amounts of conglomerate and sandstone; other exposures are conglomerate-rich (more than 80%). The apparent lack of cross-bedded sandstone is problematic. Small-scale folds are present in Unit B, and tectonism may have obscured or obliterated cross-strata. Alternatively, river channels may have been very shallow and wide, as opposed to the Donjek model's idealized 1-3 m deep channels (Cant, 1982, p.121). Low flow depth could account for a lack of flow separation and consequent lack of cross-beds (Church and Gilbert, 1975).

Fining-upward sequences are extremely rare in Unit B. Perhaps there was little mud available, or it was flushed through the river system (Long, 1978). Mudstone caps atop sequences are prone to erosion by new erosively-based channels, a process which likely did take place in Unit B as shown by its relatively abundant mudstone clasts, intraclasts of the reworked, abandoned, muddy channel fills. And as discussed above, the mixed bed load (gravel, sand) channels were perhaps not topographically differentiated enough to produce the elevated overbank areas (inactive or abandoned bar surfaces and channels) that would be mostly inactive and could accumulate muds. The presence of mudstone beds and intraclasts does, however, suggest more topographic differentiation of bars and channels in Unit B than in Unit C.

<u>Vertical sequences</u>. The small-scale fining-upward trends in outcrop #15 likely represent channel fill sequences, the standard interpretation for such fluvial facies. In the third example given, reworking by lower flow

regime conditions formed the single grain lamina, and later settling from suspension in a pond in an inactive channel deposited mud; only 1 cm of mudstone has been preserved following the subsequent gravel deposition and compaction.

The south-to-north "conglomerate-sandstone-finer sandstone with mud chips" sequences in outcrops #31 and 76 suggest local fining-upward sequences, with muddy channel fills or ponds preserved here as mud chip clasts. Desiccation cracks (mudcracks) commonly produce mud polygons or curls that can later be transported away to form mud chip/ flake conglomerates and sandstone with small mudstone clasts.

A simple lateral facies transition from sand-rich to pebble-rich sediments could have produced the vertical profile for the sandstone-conglomerate transition in outcrop #31. Fortuitous circumstances would be required for this sequence to form via the structural juxtaposition of non-conformable units.

<u>Iron formation facies</u>. The iron formation-rich sequences in outcrops #10, 12 and 79 were deposited by the settling from suspension of iron-rich chemical and muddy clastic sediments in a low-energy aquabasinal environment. Iron formation-dominant horizons cms thick indicate a short-term predominance of chemical deposition over clastic.

<u>Anomalous facies</u>. No interpretation is offered for the conglomerate in outcrop #6 because the small exposure contains only environmentally equivocal features.

The anomalous association of mixed coarse and fine clastics in outcrop #43 has no unequivocal environmental interpretation. The graded sandstone beds are most likely turbidites deposited in an aquabasinal setting. This outcrop is located at or near a possible facies change between fluvial conglomerate and the aquabasinal northern mudstone unit.

Distal facies and their D10 measurements. Many fine-grained conglomerate beds (e.g. Figs. 6.8, 6.9) are present in Dorothea and Sandra Townships (Fig. 3.1). This area also has mudstone beds, mudstone intraclast beds, sandstone-rich sequences, and fining-upward sequences. These five sedimentological features suggest more distal conditions than in the most conglomeratic portions of the assemblage, the parts with Scott type profiles (>90% conglomerate, no mudstone; model of Miall, 1977, 1978a). Interestingly, the D10 measurements for Dorothea and Sandra Townships (Table 6.2) are notably finer than many of the other CMB and NMB (compare Tables 2.1 and 6.2) conglomeratic outcrops. It appears that the D10 measurements actually reflect the more distal, finer-grained facies present, despite the potential difficulties involved in measuring the strained clasts (see Section 1.7).

Distal facies are similarly suggested by the presence of fine-grained conglomerate, mud chip intraclasts, and a mud chip band in Walters Township (Fig. 3.1). The D10 measurements for this area (Table 6.2, Fig. 2.8) are finer then those from many of the outcrops farther along strike to the east, and there is an apparent thinning of the conglomeratic assemblage towards Walters Township's finer D10 values. Once again, D10 measurements seem to confirm the independently interpreted distal nature of this portion of the conglomeratic assemblage.

In the Beatty Lake area (Fig. 3.1) some low D10 values correspond to relatively low conglomerate : sandstone ratios (Table 6.3, Fig. 2.8), but some other D10 measurements do not do so (Unit B in Table 6.3). A possible reason for this apparent discrepency is given in Section 12.5.

Because of the greater abundance of mudstone and finer sandstone in these distal facies,

1) more rip-up clasts (local bed fragments) are present than in sequences lacking mudstone beds, and the rip-up clasts are abundant enough in some cases to form intraformational conglomerates;
2) sequences fining upward to mud have been preserved; and
3) lithological contrasts are greater than in sequences with conglomerate and sandstone only, which allowed more loading to occur.

11.13 Interpretation of the CMB's Heterogeneous Assemblage

Hydrodynamic interpretation. Although parts of the heterogeneous assemblage's sequence are likely structurally modified by faulting, shearing and/or transposition of beds, it can be inferred that a wide variety of gravel, sand and mud facies were deposited and interbedded together. Some lithofacies may have been winnowed from within the subaqueous assemblage; in this case local slopes would have been significant (e.g. Postma, 1984, Fig. 28).

Changes in intergranular dispersive pressure and matrix buoyant lift determine the nature of high-density turbidity currents and density-modified grain flows. As these decelerate, deposition of clast-supported gravel is by traction carpet layers, frictional freezing, and settling from suspension (Lowe, 1982). With regard to the heterogeneous assemblage's conglomerate,

1) fallout from suspension and the freezing of traction carpet layers produced inverse-graded layers;
2) flow unsteadiness caused suspension sedimentation of graded gravel beds; and
3) reworking of sediment surfaces by a turbidity current's later flow formed stratified fine gravel and sand, the upper parts of the graded-stratified bed (in outcrop #98)(processes of Lowe, 1982).

The graded-stratified bed's alternating sand and traction gravel layers are the result of fluctuations in shear velocity and thus competence (Walker, 1975). Rapid deceleration of sandy turbidity currents deposited from suspension fine pebbles and granules at the bases of the normal graded pebbly sandstones (process of Hein, 1982).

Massive and ungraded conglomerate beds ("disorganized-bed" term of Walker, 1975, 1978b, 1984a) are more common than graded ones in the heterogeneous assemblage. Hein (1982, Table 4, p.280) lists four possible mechanisms leading to an absence of grading:

1) "complete recycling of material within the [turbidity] current";
2) complete mixing via "syn-depositional deformation (churning and (or) fluidization)";
3) a "very rapid depositional phase in relation to rate of loss of flow competence"; and
4) "grains were not free to move within the flow".

Crude bedding shows the vertical amalgamation of beds in multistoried units.

Surlyk (1984) has argued that the resedimented conglomerate models of Walker (1975, 1978b, 1984a), Hein (1982) and Lowe (1982) reflect the proximal and distal aspects of individual flows and flow maturity and not those of the entire depositional system. This is an important point to consider for cases in which sediment flows were derived from most or all of a system (e.g. a subaqueous fan, a delta front) that lacked a relatively fixed fan apex or had more than one feeder channel.

The freezing of flows in which clasts were fully or partly supported by matrix buoyancy and cohesiveness (Lowe, 1982) deposited outcrop #98's matrix-supported debris flows.

Massive sandstone may represent sediment gravity flows (e.g. fluidized flows, grain flows). Laminae in ungraded beds can be either suspension deposits or traction layers formed by currents, the latter including rippled laminae. Only rarely were mud-draped ripples preserved.

The graded sandstone beds record fining-upward deposition by low-density turbidity currents. An ideal Bouma sequence has a coarse base dropped out of suspension (Ta), followed by the flow's waning upward to traction (Tb,c), traction and/or suspension (Td) and suspension (Te) deposits; post-turbidite mud is included in the Te division (Lowe, 1982; Walker, 1984a). It is very difficult to recognize the lamination and small structures of the various Bouma divisions within the heterogeneous assemblage's foliated, thinly bedded, graded sandstone. The bed-parallel shearing typical of the Beardmore-Geraldton terrane according to Williams (1985) often obscures or destroys fine sedimentological details and bedding. This is why the graded sandstones have not been described in the more typical manner of other studies (e.g. Tabc, Tabd, Tae). Partial Bouma sequences such as Tabd and Tad are probably present in beds cms thick in outcrop #103. Archean thin-bedded turbidites observed in other parts of Superior Province almost never contain full and well-defined Bouma sequences, and the rippled Tc division is particularly rare.

The laminated siltstone with common graded laminae may be either the products of very low density turbidity currents (Stow and Bowen, 1980, Stow and Shanmugam, 1980; Chough and Hesse, 1980; Hesse and Chough, 1980) or pure suspension deposits (i.e. settled from overflows or interflows).

Mudstone intraclasts were supplied by either fluvial inflow to subaqueous areas or the disaggregation (hence rip-up clasts) of subaqueous mud beds via slumping and/or erosive currents. Since the mudstone clasts are identical to mudstone interbedded with the coarser lithofacies, the bed-disaggregating processes cited above likely operated.

In such tectonized exposures it is likely that the muddy and often laminated (closely spaced discontinuities) lithofacies are not exactly *in situ* but are bed fragments that have slipped along the foliation between the coarser conglomerate and sandstone beds. Also, some thick "beds" may in fact be large (metres, tens of metres) rip-up clasts or slump blocks flattened along the foliation. The occasional association of smaller mudstone clasts with smaller (about 1 cm) pebbles suggests paleohydraulic equilibrium (i.e. only small clasts transported) and argues against at least some of the clasts being beds fragmented by tectonic deformation.

The penecontemporaneously deformed sandstone suggests liquefaction and/or slumping of rapidly deposited, water-saturated sediment.

Gravity-winnowing from the walls of channels, gullies and slump scars (Prior *et al*, 1981; Wescott and Ethridge, 1982) or from debris flow lobes (Postma, 1984, Fig. 28) could produce small (metres) sections of rapidly varying character.

No unequivocal shallow marine conglomerate features (e.g. Nemec and Steel, 1984; Bourgeois and Leithold, 1984) are present.

<u>Fan-delta front or submarine fan?</u> The heterogeneous assemblage of resedimented conglomerates (deposited by high-density turbidity currents and other sediment gravity flow processes, including debris flows), sandstone turbidites, mudstone beds and intraclasts, and other facies was deposited as part of either a submarine fan or fan-delta front paleoenvironment. Distinction between these two settings depends on facies relationships, stratigraphic context, and the terminology used, as discussed below. Recent work on fan-deltas (see below) suggests that the heterogeneous assemblage should not be forced to fit the archetypical submarine fan facies model's framework (Walker, 1978b, 1984a) of feeder channel, distributary fan channels, suprafan lobes, levees and an outer fan or basin plain.

In a preface to a recent memoir on gravels and conglomerates, R. G. Walker (p. ix in Koster and Steel, 1984) notes that the volume's papers on fan-deltas create "the impression that fan-deltas commonly grade rather imperceptibly into submarine fans". This gradation has led to some confusion regarding the nomenclature of subaqueous (or submarine) fans and fan-deltas (Table 11.1). Some authors refer to the subaqueous parts of fan-deltas that contain resedimented conglomerates and turbidites as "fan-delta fronts/slopes" (Wescott and Ethridge, 1982; Nemec and Steel, 1984; Postma, 1984; Nemec *et al*, 1984; Porebski, 1984), while others term the same sites "submarine fans" (Ethridge and Wescott, 1984, Fig. 12; Surlyk, 1984). Discussion of semantics is relevant when one considers previous problems involving conflicting and obscure terminology for submarine fans (e.g. Nilsen, 1980, with replies by Walker (1980) and Normark (1980); Miall, 1984, pp.198,199). Also, new variants have recently been added, including the delta-fed submarine ramp facies model for sand-rich turbidite systems (Heller and Dickinson, 1985), and a bottomset-modified Gilbert-type delta (Postma and Roep, 1985).

If a fan-delta is considered to be a single system with subaerial and subaqueous parts, the term "subaqueous/ submarine fan" seems to neglect the linked subaerial component, whereas "fan-delta front/slope" emphasizes the connection between a fan-delta's plain and its subaqueous front/slope. If a fan-delta is considered to be a composite of functionally separate but spatially related systems, the subaerial and subaqueous parts each contain lobes fanning outward from their respective apices. After avulsion and the re-occupation of a formerly inactive area on a fluvial delta plain, a distributary mouth will supply or become the apex of a subaqueous (submarine) fan. Such a subaqueous fan will be superimposed onto and/or incised into the subaqueous part of the larger scale fan-delta. Multiple avulsion events and thus changes in the location of distributary mouths are to be expected, and individual subaqueous fan systems will laterally overlap and coalesce. In some cases subaqueous channels may allow sediment to bypass a delta front, constructing at a deeper level a separate subaqueous fan

TABLE 11.1

Facies Associations *	Single Depositional System	Depositional Systems Tracts			
		1	2**	3	4***
Alluvial fan-fluvial	fan-delta plain	felsic volcanic inland alluvial fans braidplain fan-deltas (coastal alluvial fans)	Scott type proximal gravelly braided river Donjek type medial braided river Platte or South Sask. distal sandy braided river	fan-delta (coastal alluvial fan)	proximal fan-delta lower fan-delta
Shoreline		shoreface	coastal		transition zone
Resedimented (turbidite)	fan-delta front/slope	fan-delta front submarine fan	submarine fan (upper, middle and lower fan)	either submarine fan or slope and submarine ramp	slope proximal submarine fan
Pelagic	prodelta	basin plain	basin plain	basin plain	distal submarine fan

Table 11.1: Terminology of some clastic depositional systems and tracts.

* Ojakangas, 1985
** Fluvial models of Miall, 1978a
*** Terms of Ethridge and Wescott, 1984, Fig.12

(Elliott, 1978a, Fig. 6.50C; Walker 1978b, Fig. 23).

In the conceptual frameworks outlined above, the single system approach emphasizes a subaqueous portion with multiple feeder (tributary) channels, whereas the multi-system view considers separate subaqueous fan systems, each having its own feeder channel.

The above discussion leads to complexities not considered by the current standard submarine fan model (Normark, 1978; Walker, 1978b, 1984a) such as:

1) multiple feeder channels (discussed by Heller and Dickinson (1985) for sandy systems);
2) separate and overlapping subaqueous fan systems (Walker, 1984a, p.184);
3) widespread slumping (Prior *et al*, 1981; Ethridge and Wescott, 1984; Postma, 1984, Fig. 10; Massari, 1984);
4) sediment gravity flows originating from all parts of a subaqueous fan, resulting in the interbedding of ideally "proximal" and "distal" resedimented conglomerates (Surlyk, 1984);
5) fan-delta fronts with very steep slopes (up to 25°) that generate sediment gravity flows (Postma, 1984; Postma and Roep, 1985); and
6) shallow marine, wave-worked conglomerates (Nemec and Steel, 1984; Ethridge and Wescott, 1984; Kleinspehn *et al*, 1984; Bourgeois and Leithold, 1984).

The overlap and interplay among fluvial, shallow aquabasinal (waves, tides) and deeper-water resedimenting processes may make fan-delta systems/models more complex than mere juxtapositions of alluvial fan, delta and submarine fan facies models. Also, Ito (1985) has described a regressive submarine fan to fan-delta sequence complicated by submarine volcanism.

Keeping in mind the provisions outlined above, fitting the heterogeneous assemblage into the standard submarine fan model (Walker, 1978b, 1984a) would allow interpretation of the conglomerate as feeder channel deposits incised into a slope or upper fan, the sandstone as distributary mid-fan channel and suprafan lobe deposits (distributary channels could also be conglomeratic), and the silty mudstone as channel fill, levee or overbank (inter-channel) deposits. It is unlikely that outer fan or basin plain muds were interbedded in this conglomerate-rich assemblage. In a terraced channel situation (Hein and Walker, 1982; Eriksson, 1982), conglomerates are regarded as subaqueous braided thalweg channel deposits, sandstone as terrace and levee deposits, and mudstones as levee and overbank deposits.

Since stratigraphic continuity and conformability cannot be assumed in a deformed Archean terrane (i.e. potential juxtaposition of fault slices) with a low density of exposure (i.e. few or no transitional contacts between units exposed), the stratigraphic context of the heterogeneous assemblage cannot be reliably inferred. The nature of the heterogeneous assemblage's exposure does not allow any differentiation between fan-delta front and submarine fan settings, which may be largely semantic anyway.

11.14 Interpretation of the CMB's Ida Lake Sandstone-Rich Assemblage

The very minor amount of outcrop exposed limits the paleoenvironmental interpretation.

The graded beds suggest turbidites that infilled part of an aquabasin below wave base. Beds and laminae with a wide variety of grain sizes (granules to mud) were probably deposited by a variety of processes. Grading, sharp bed soles, scouring and load structures all give stratigraphic tops to the north.

11.15 Interpretation of the CMB's Coral Lake Volcaniclastic Assemblage

Interpretation of the Coral Lake sediments is limited by the nature of the exposures: outcrops are generally poor in quality and quantity, and sedimentary structures and top indicators are almost non-existent.

The oligomict, volcaniclastic conglomerate and sandstone assemblage (Fig. 3.1) is similar in texture and stratification to the CMB's polymict conglomerate and sandstone assemblage (Chapter 6), implying similar depositional processes and paleoenvironments. Using Miall's (1977, 1978a) lithofacies codes and profile types, the lithofacies assemblage in the Coral Lake sediments of Gm, Gms, Sm (massive sandstone), Sh, Sp and Sr suggests an alluvial fan-braided river paleoenvironment with Scott to Trollheim-type vertical profiles. Braided stream deposits (bar, channel, and sheetflood) are a fundamental component of most alluvial fans (Bull, 1977; Collinson, 1978; Miall, 1978b, p.33; Nilsen, 1982; Rust and Koster, 1984).

Within the Coral Lake sediments, stratification is dominantly crude, suggesting the possible reworking of a felsic volcanic source by braided streams. Fine-grained conglomerate beds were separated from their coarser and more poorly sorted parent populations via local sorting processes such as winnowing from slopes and bars. The matrix-supported conglomerate, with its very poor sorting, suggests the deposition of a finer fraction as sandy debris flow deposits (Collinson, 1978; Nilsen, 1982; Rust and Koster, 1984). Sorting variations at 10-100 m scales may be due to a large-scale process, reflecting local slopes and/or channels. The sandstone beds generally lack sedimentary structures, probably the result of sheetflood (Collinson, 1978; Ballance, 1984; Hogg, 1982) or sediment gravity flow (e.g. grain flow) processes. It is difficult to determine whether or not the strata in outcrop #131 (Section 9.4) were deposited subaqueously, and how locally (in Fig. 3.1) important such a facies may have been.

The lack of muddy deposits may be either due to the presence of a coarse (pyroclastic?), felsic volcanic-rich source with little or no mud (or fine ash) supplied, or a result of highly proximal conditions in which weathering may not have had time to produce much mud (e.g. rapid sedimentation and transport on unvegetated slopes) or, more likely, mud was flushed through the transport system by streams and/or groundwater flow. Note that any debris flow deposits would thus have sandy matrices.

Lower paleoslopes would be suggested by more structured deposits (plane lamination, cross-bedding; facies models of Boothroyd and Nummedal, 1978, Fig. 5, and Rust and Koster, 1984, Fig. 3) and fewer or no debris flow deposits, because in more distal and usually lower sloping areas debris flows have a low preservation potential. The likely presence of debris flow and sheetflood deposits suggest relatively high paleoslopes, such as in an alluvial fan setting. The dominantly massive stratification units and probable fluvial deposits are consistent with a fan interpretation. Felsic volcanic edifices may be laterally and distally (downslope) transitional to alluvial fans (Fisher and Schmincke, 1984, Fig. 13.4; Ayres and Thurston, 1985, Fig.2; Kuenzi *et al*, 1979) or, where fans were/are not present, braided rivers (Davies *et al*, 1978).

Although the features listed and interpreted above are equivocal in terms of differentiating pyroclastic versus alluvial fan-fluvial processes, the lack of any unequivocal, purely volcanic textures and stratification slightly favours the alluvial fan-fluvial interpretation. Parts of the Coral Lake sediments, particularly the massive ones, may be the products of purely pyroclastic processes. The exposures are too poor to determine the amount of unreworked (pyroclastic) versus reworked (volcaniclastic) material. It is the nearly oligomict, felsic volcanic composition of the Coral Lake sediments' clasts that suggests a proximity to a felsic volcanic-rich source, and thus a possible volcaniclastic to pyroclastic origin.

Because of poor exposure, the uncertain and equivocal relationships of the region's lithologic belts (Fig. 1.1), and a lack of radiometric dating, it is unknown if felsic volcanism was ever even approximately synchronous with deposition of the Coral Lake ediments, although such a scenario is commonly inferred for Archean supracrustal sequences (Ojakangas, 1985; Ayres and Thurston, 1985) and would be an excellent way of providing the required supply of felsic volcanic clasts. The typically steep slopes of felsic volcanic cones/edifices and the abundance of coarse, poorly sorted sediment would favour the development of shallow braided streams.

Compositional reworking of a felsic volcanic source area can be:

1) minor, thus preserving the source's approximately oligomict, felsic volcanic clast composition; this is likely for the origin of the Coral Lake sediments;
2) partial, resulting in a compositional interfingering of oligomict and more polymict facies; this may be the reason for the odd conglomerate composition noted in outcrop #24 (Section 9.5); or

3) total, with thorough mixing from other sources producing a polymict conglomerate (and associated lithofacies), one possibly still dominated by felsic volcanic clasts; this may be the reason for the felsic volcanic-rich clast composition of the CMB's and NMB's polymict conglomerate.

Teal (1979; Teal and Walker, 1977; Blackburn *et al*, 1985) has described for the Archean Manitou Group in western Wabigoon Subprovince a felsic volcanic source reworked distally into finer alluvial fan and braided river deposits. Teal's volcanic-to-fluvial depositional systems tract may be partly similar to the CMB, but in the CMB there are no good clast size or paleocurrent trends to suggest that the Coral Lake volcaniclastic sediments are a more source proximal equivalent of the rest of the CMB.

CHAPTER 12: STRATIGRAPHY OF THE CMB

12.1 Vertical Trends Within the CMB

The differences between the following sections through the CMB (Fig. 12.1; see also Figs. 3.1, 3.2) are important enough to warrant their separate discussion. The paleoenvironmental interpretations given for the various lithofacies assemblages (see Chapter 11) do not depend in any way on their relative positions in the CMB.

Dorothea and Sandra Townships. A CMB-scale mudstone to sandstone to conglomerate coarsening-north sequence (Figs. 3.1, 3.2, 12.1) defines the progressive increase in average grain size. Within the sandstone-rich horizon, the deltaic sequence (Sections 5.2, 11.7) shows a subtle coarsening upward (to the north) trend defined by a slight increase in coarse sand and the concurrent loss of the muddy fraction. Sandstone interbeds to the north in the conglomeratic horizon are rich in coarser sand and are more poorly sorted. The well to moderately sorted, very fine-grained (silty) to medium-grained deltaic sandstone is likely a distal form of the moderately to poorly sorted, often pebbly, coarse to fine-grained braided river sandstone. The coarsest fraction present (D10) is only small pebbles in the deltaic sandstone, versus larger pebbles and cobbles in the fluvial conglomerate to the north. The percentage of conglomerate in the sequences is near zero in the deltaic sandstones (rare pebble

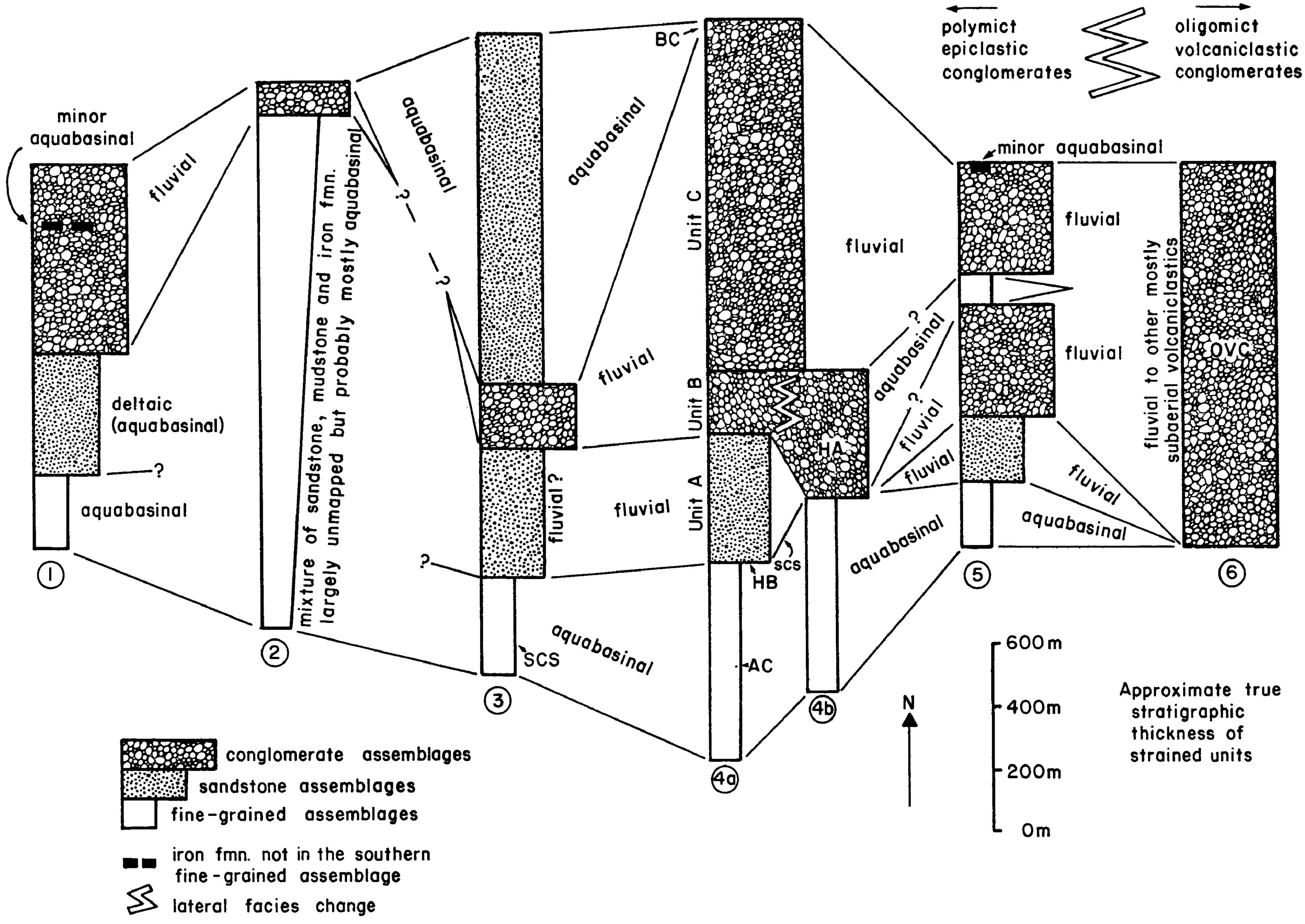

polymict epiclastic conglomerates
oligomict volcaniclastic conglomerates
minor aquabasinal
fluvial
deltaic (aquabasinal)
?
aquabasinal
①
mixture of sandstone, mudstone and iron fmn.
largely unmapped but probably mostly aquabasinal
②
aquabasinal
?
?
fluvial?
?
SCS
aquabasinal
③
aquabasinal
fluvial
fluvial
BC
Unit C
Unit B
Unit A
HB
scs
AC
HA
④a
④b
fluvial
? aquabasinal
? fluvial
fluvial
aquabasinal
minor aquabasinal
fluvial
fluvial
fluvial
aquabasinal
⑤
fluvial to other mostly subaerial volcaniclastics
OVC
⑥
N
600m
400m
200m
0m
Approximate true stratigraphic thickness of strained units
conglomerate assemblages
sandstone assemblages
fine-grained assemblages
iron fmn. not in the southern fine-grained assemblage
lateral facies change

Fig. 12.1: Lithostratigraphy and interpretation of the CMB.
Composite stratigraphic sections of:
1) Dorothea and Sandra Townships;
2) Irwin Township;
3) Walters Township;
4a) western Beatty Lake) western Leduc
4b) eastern Beatty Lake) Township;
5) Leduc Lake area, central Leduc Township; and
6) eastern Leduc and western Legault Townships.

AC = anomalous conglomerate
SCS = swaley cross-strata
HB = herringbone cross-strata
HA = heterogeneous assemblage (resedimented conglomerates, etc.)
OVC = oligomict volcaniclastic conglomeratic assemblage
BC = bouldery conglomerate

The lateral facies change from Unit B to the HA in Section 4 is a lithofacies change, with no change in clast provenance. The possible lateral facies change between Sections 5 and 6 is a mostly compositional (provenance) one.

bands) to about 75-100% in the fluvial conglomeratic portion.

Irwin Township. A fluvial conglomeratic assemblage at the CMB's north margin merely hints at a coarsening-north sequence like those better developed along strike to the east and west (Figs. 3.1, 3.2, 12.1). Aquabasinal ironstone is distributed throughout the southern half of the sequence (Mackasey, 1975).

Walters Township. The fairly well developed coarsening-north sequence in Walters Township (Fig. 3.1, 3.2, 12.1) is very similar in its lithostratigraphy and lithofacies assemblages to other sections of/through the CMB, except that is spans only the southern half of the CMB. The same south-to-north aquabasinal-to-fluvial transition appears to be present.

Stratigraphic top indicators are very rare in this section (township), except in the Ida Lake sandstone-rich assemblage.

Western Leduc Township (Beatty Lake area). The conglomerate : sandstone ratio of the outcrop sections, bed thicknesses, stratification features, sedimentary structures, mean and maximum (D10) grain/clast size, and textures (massiveness, nature of or lack of grading, degree of sorting) listed in the lithofacies descriptions and summarized in Table 12.1 suggest the interpretations given in Chapter 11 and summarized in Table 12.1.

A generally coarsening-north sequence spans the entire CMB thickness in the Beatty Lake area (Figs. 3.1, 3.2, 12.1;

Table 12.1) and is well-defined by both mean grain/clast size (lithology) and maximum deformed clast size (D10) measurements (Fig. 2.8; Tables 6.2, 6.3, 12.1). The paleoenvironmental facies are arranged in a south-to-north, subaqueous-to-subaerial and increasingly proximal to the north succession (Fig. 12.1, Table 12.1).

Parts of Unit C are interpreted to be medial facies via their conglomerate : sandstone ratios and D10 measurements (Table 6.3), but Unit C does not contain mudstone beds or clasts, likely an indication of slightly more proximal conditions in Unit C than in Unit B.

The presence of muddy channel fills and their reworked equivalents (muddy intraclasts) along with generally less conglomerate and thinner conglomerate beds suggest a more distal and possibly more topographically differentiated braided river paleoenvironment in Unit B than in Unit C (Miall, 1977, 1978a; Cant, 1982).

The distal fining reflected by some of the D10 measurements (Table 6.3) and Unit A's coarsest (pebbly) fraction and sandstone mean grain size resulted from the rivers' tendency to laterally fractionate their load into finer sizes downstream (Smith, 1970). There may be possible minor transitions to a South Saskatchewan type profile (Miall, 1977, 1978a; Cant, 1978) in Unit A (Table 12.1).

Units A, B and C lack unequivocal evidence of turbidite processes.

Possible intertidal (herringbone cross-beds) and shallow

TABLE 12.1

Lithofacies Assemblage	Objective Criteria	Interpreted Paleoenvironment
Unit C	conglomerate > sandstone thicker conglomerate beds Sm > Sh > Sp no mudstone beds or clasts	proximal (Scott type) to medial (Donjek type) gravelly braided rivers; longitudinal bars
Unit B	conglomerate ≈ sandstone some thinner conglomerate beds Sm > Sh mudstone beds and clasts	medial (Donjek type) gravelly braided rivers; longitudinal bars
Unit A	no conglomerate pebble bands, which coarsen N (up) Sp, St > Sh > Sm some finer-grained and better sorted sandstone than in Units B, C	distal (Platte type) sandy braided rivers; linguoid bars; coarse-tail upward coarsening of Unit A
outcrop #100, Unit A	herringbone cross-strata plane laminated horizons mud laminae and drapes	intertidal? coastal? (minor)
outcrop #89	swaley cross-strata	shallow part of an aquabasin; (storm) wave-influenced
Hetero-geneous assemblage	graded conglomerate and sandstone	resedimented conglomerates and sandstone turbidites of a subaqueous fan or fan-delta front
Southern fine-grained assemblage	mudstone, graded sandstone	suspension muds and sandstone turbidites of an outer subaqueous fan, prodelta or basin plain

Table 12.1: Interpretation of lithofacies assemblages in the Beatty Lake area. All tops are to the north, except in the southern fine-grained assemblage.

marine (swaley cross-strata) facies are present, and suggest a paleoshoreline between the aquabasinal assemblages to the south and the fluvial ones to the north.

D10 measurements (Fig. 2.8, Table 6.3) show that the resedimented (subaqueous) conglomerate beds are finer-grained than the fluvial (subaerial) ones in Units B and C and elsewhere in the CMB (Table 6.2), as might be expected of more distal clasts, a clast population that had travelled through a fluvial system and had been exposed to its sediment traps, selective sorting and abrasion processes. Point counts and field observations show that the fluvial and resedimented conglomerates have the same provenance (see Chapter 10).

East-central Leduc Township. The CMB-scale mudstone to sandstone to conglomerate-rich coarsening-north sequence (Figs. 3.1, 3.2, 12.1) defines the increase in average grain size. The thin conglomerate beds in the sandstone-rich lithofacies assemblage (Section 5.5) are finer grained than the more thickly bedded, coarser conglomerate beds in the conglomeratic assemblage (Section 6.2) to the north, as shown by D10 data (Section 5.5, Table 6.2) and field observations. The conglomerate in both the sandstone and conglomerate-rich assemblages has the same composition. Thus the contrasts in the conglomerate's average and maximum (D10) clast size, bed thickness, and its percentage in the local outcrop sections suggest that the sandstone-rich assemblage is more distal than the related conglomeratic assemblage. This proximal-distal relationship is based on standard sedimentological reasoning, summarized in the facies model of Boothroyd and Nummedal (1978, Fig. 5).

Stratigraphic top indicators are uncommon in the CMB in this area. Anomalous departures from the coarsening-north trend are the northern mudstone unit (Section 6.4) within the conglomeratic assemblage, and the iron formation (Section 6.8) at the CMB's north margin. It is not known whether or not the northern mudstone unit may have been tectonically interleaved/intercalated in the CMB sequence.

East Leduc and Legault Townships. The Coral Lake "sediments" are felsic pyroclastics or volcaniclastics deposited by braided rivers and sediment gravity flows reworking volcanic material: both primary and reworked facies may be present. The generally oligomict, felsic volcanic nature of the clasts contrasts markedly with the highly polymict conglomerate in the rest of the CMB (and in the NMB). There is no coarsening-north trend developed in this section/area (Figs. 3.1, 3.2, 12.1).

12.2 Summary of Top Indicators in the CMB

Stratigraphic facing (top) directions of individual beds are nearly all to the north in the CMB's generally east-striking beds. Only some of the thinly bedded graded sandstone beds (turbidites) in the Beatty Lake area (Figs. 3.1, 4.1) have (rare) tops to the south.

The following top indicators are present in the CMB and have been previously described and interpreted in this thesis:

1) conglomerate and sandstone beds with notably sharper south margins (erosive bed soles) than north margins;
2) beds that are coarser to the north of sharp bed contacts (soles) than the beds adjacent to the south (below) (a variation of 1) above);
3) concave-north scours, including a channel;
4) cancave-north cross-bed foresets, toesets and ripples, including swaley cross-stratification;
5) evidence of a downslope aspect (pebble rolled down) on a foreset layer;
6) a sandstone bed finer to the north of a probable residual lag pebble band than to the south;
7) fining-north sequences (interpreted to be fining-upward ones), one with loading and soft-sediment diking giving tops north;
8) local rip-up clasts derived from an adjacent bed to the south (below);
9) an intraformational conglomerate bed, with a south-to-north, matrix- to clast-supported texture attributed to flood waning;
10) fining north (interpreted to be normal grading, fining upward) in both pebbly and fine sandstone beds (most examples interpreted as turbidites);
11) a fining-north, graded-stratified conglomerate bed

(a fining-upward turbidite);

12) possible inverse graded conglomerate beds that coarsen north (upward) only near their presumed bases, characteristic of inverse grading in such turbidites according to Walker (1978b, p.942);

13) loading of coarser beds southward (downward) into finer beds; and

14) soft-sediment injection to the north (upward).

With the exception to 10) noted above, all of the top indicators in the CMB give tops to the north. It thus seems obvious, in light of the arguments presented in Section 12.1, that the CMB's coarsening-north sequence is best interpreted as a coarsening-upward one.

12.3 Interpretation of the CMB's Internal Stratigraphy

Regional-scale evidence of a preserved stratigraphy. The internal stratigraphy of the CMB (Chapters 3 to 9, Fig. 12.1; see also Mackasey, 1975, 1976) contains a southern aquabasinal lithofacies assemblage, a central aquabasinal to fluvial sandstone assemblage, and a northern fluvial conglomerate-dominant assemblage (Figs. 3.1, 12.1). These lithofacies assemblages persist from east to west within the 1-2 km thick CMB for about 34 km along strike (Figs. 1.1, 3.1, 3.2, 12.1), a (sub-)regional scale of lateral continuity and a well-defined stratigraphy. It is the lithofacies assemblages, not individual lithofacies (e.g. single beds), that define the linear (to curvilinear) regional stratigraphy.

The individual, objective sedimentological trends (e.g. mean grain/clast size, conglomerate : sandstone ratio) presented in Chapters 4 to 8 and summarized in Section 12.1 cross-cut the boundaries of the various lithofacies assemblages because the trends usually span the CMB's thickness of several assemblages.

Nowhere in this thesis has stratigraphic continuity (conformability) between the CMB's lithofacies assemblages been assumed. Similarly, continuity and conformability within the assemblages has not been assumed except for the well-exposed sandstone (deltaic) assemblage in outcrop #21A (Sections 5.2, 11.7).

The regional stratigraphy outlined above demands that the CMB sequence young either to the north or south, perpendicular to both the regional (tens of kms) and local (metres to hundreds of metres to kms) east-striking stratification. Since nearly all the top indicators face north (see Section 12.2), except in some isoclinally folded, fine-grained and thinly bedded lithologies preferentially located along the south margin of the CMB, the CMB's coarsening-northward sequence is interpreted to be a coarsening-upward (prograded) sequence.

It is interpreted that a coarsening-upward (prograded), subaqueous to subaerial, increasingly proximal to the north trend is preserved in the epiclastic portion of the CMB (Figs. 3.1, 12.1). This prograded stratigraphic sequence is capped by an aquabasinal unit in Walters Township (Figs.

3.1, 12.1).

The similarity of this generally coarsening-upward (mega-) sequence, spanning in most cases the entire CMB thickness (up to 2 km), to the sequence that would be produced by a mainly/dominantly prograding (regressive) coarse clastic depositional systems tract is quite striking (Table 12.1, Figs. 12.1, 12.2). Despite possible structural disturbances, trends identical to those seen in undeformed progaded sequences are present.

Tectonic stacking of fault slices has been proposed as an origin for the Beardmore-Geraldton terrane's linear arrangement of metavolcanic and metasedimentary belts (Fig. 1.1) by Williams (1985), but this does not necessarily apply to the stratigraphic sequences *within* each linear metavolcanic or metasedimentary belt.

There seems little doubt that the CMB's internal stratigraphy is not a sequence of tectonically and randomly juxtaposed segments. It is highly unlikely that structural telescoping of lateral facies equivalents could produce such a well-ordered vertical sequence in the CMB, particularly the one exposed in the Beatty Lake area (Figs. 3.1, 12.1; Table 12.1). Structural telescoping cannot be completely ruled out in the absence of a rigorous structural analysis of both the CMB and the entire Breardmore-Geraldton terrane, but the excellent fit of the coarsening-upward CMB section to fundamental sedimentologic and stratigraphic trends and patterns cannot rationally be ignored or de-emphasized. It

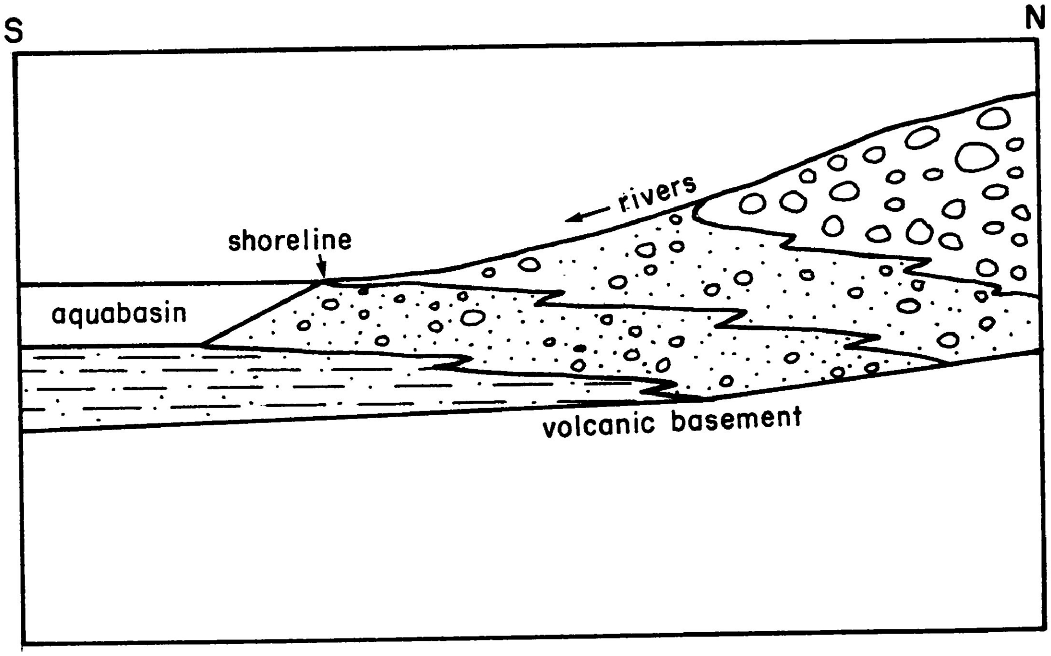

S
N
rivers
shoreline
aquabasin
volcanic basement

Fig. 12.2: Regressive, prograded clastic depositional systems tract and vertical sequence. Small fan-deltas may be located at the shoreline, or the entire clastic tract may comprise a large fan-delta.

seems nearly inescapable that the CMB sequence is a quasi-conformable one, a nearly continuous cross-section through a prograded clastic wedge or sheet ("wedge" assumes the usual thickening of fluvial coarse clastics towards their source) not seriously interally disturbed by tectonism.

The discussion above suggests that the use of stratigraphic context (Walther's Law) is, in a broad way, valid in analyzing the CMB. This leads to some interesting speculations, listed in Section 12.5.

Other Archean sequences. Teal (1979; Teal and Walker, 1977) has presented a stratigraphic synthesis of an Archean metasedimentary sequence over 2 km thick with a transition upward from subaerial to subaqueous paleoenvironments and descriptions and interpretations of lithofacies similar to the CMB's. The Manitou Group in western Wabigoon Subprovince is a megasequence in which:

1) "the sedimentary belt is essentially monoclinal" (p.2);
2) "neither the faulting nor the folding has interfered substantially ..." (p.22);
3) "folding affected ... fine-grained sedimentary rocks, but not the conglomerates below them" (p.24);
4) nearly all the stratigraphic top indicators face in one direction, north (Fig. 8);
5) "it is almost inescapable" (p.156) that the megasequence is not a fairly continuous (broadly

conformable) stratigraphic succession (Teal, 1979). (Generalizations 1, 2, 4 and 5 above may apply to this thesis' CMB as well.) Teal's synthesis has been widely accepted (e.g. Easton, 1984; Blackburn *et al*, 1985, p.105).

The Swaziland Supergroup of the Archean Barberton greenstone belt has a gradational (conformable) transition upward from turbidites (Fig Tree Group) to fluvial conglomerates (Moodies Group) (Eriksson, 1978, p.289; 1980, p.156), and coarsening-upward (prograded) fan-delta sequences hundred of metres thick (Eriksson, 1978).

The CMB (Fig. 12.1) and the studies of Teal, Eriksson and others strongly suggest that large-scale (up to kms thick) sedimentological/stratigraphic trends and sequences are sometimes relatively well preserved in Archean supracrustal units.

12.4 Discussion of Lateral Trends Within the CMB

As the preceding discussion of vertical stratigraphic trends has shown, a largely prograded sequence is apparently well-preserved in the CMB. It is now pertinent to examine the lateral variations within the overall prograded succession.

Fig. 3.2 outlines the CMB's regional stratigraphy. Fig. 3.1 shows the distribution of the lithofacies assemblages and the outcrops studied. Mackasey's (1975, 1976) maps show most aspects of the descriptive stratigraphy. Interpretive composite stratigraphic sections are given in

Fig. 12.1.

The aquabasinal assemblage has good continuity across the southern, lower portion of the CMB. Because of a lack of mapping, due to access problems, it is not known if the fluvio-deltaic sandstone assemblage present in Sandra, Walters (?) and Leduc Townships extends across Irwin Township. Mackasey (1975) recorded outcrops of sandstone, siltstone, mudstone and iron formation for most of this problematic area of the CMB. If deeper-water aquabasinal facies occupy most of the Irwin Township segment of the CMB they may be correlative to the turbiditic sandstone assemblage in the upper CMB of Walters Township (Chapter 8; Section 11.14; Fig. 12.1) which would imply that the fluvial and aquabasinal assemblages are broadly interfingering along strike in Walters and Irwin Townships.

The northern conglomeratic assemblage's apparent thinning along strike (Figs. 3.2, 12.1) is either a primary stratigraphic feature or an artifact of structural truncation, or both. Fluvial conglomeratic systems usually fine and thin distally (Smith, 1970; McGowen and Groat, 1971; Miall, 1977, 1978a, 1984, p.237; Boothroyd and Nummedal, 1978). The fining (D10 data) and thinning west from Leduc to Walters Township (Figs. 2.8, 3.2, 12.1) suggests a proximal to distal transition along regional strike, but low density of outcrop and the lack of exposed assemblage boundaries limit this interpretation. The D10 data are based on measurements of the largest clasts, which are mostly (90-100%)

granitoids, thus the data is not clast lithology-dependent.

Following the conglomeratic assemblage along strike in the directions of apparent thickening leads to the coarse pyroclastic or volcaniclastic "sediments" at the CMB's eastern terminus (Chapter 9), and felsic to intermediate pyroclastics in Dorothea Township, north of and immediately adjacent to the CMB (Mackasey, 1975). This is of interest as the CMB's polymict conglomerate is rich in felsic volcanic clasts (Chapter 10), and felsic volcanic edifices are thought to have been the source of most of the detritus that formed Archean clastic sedimentary units (Teal, 1979; Thurston *et al*, 1985; Ojakangas, 1985; Ayres and Thurston, 1985).

Sandra Township's sandstone assemblage (Sections 5.2) displays both apparent (two-dimensional cross-section) and true (three dimensional) paleocurrents nearly all to the east and southeast. Apparent paleocurrents in Leduc Township (Sections 5.4, 5.5) are mostly to the west. Combined with the apparent thinning of the conglomeratic assemblage toward Irwin Township, one cannot help but speculate that these vague trends suggest that the conglomeratic assemblage displays, in both the east and west CMB, transport and thinning away from felsic volcanic source areas. This could reflect either sedimentation by systems from the east and west, or the radial drainage component of systems (e.g. fan-deltas) that prograded from the north (progradation from the north is further discussed in Sections 13.2 and 13.3). The western CMB's distinctive mafic to intermediate "intrusive" clasts, present to the west of Irwin Township's aquabasinal facies-rich section (Figs. 3.2, 12.1) and paralleled by a similar trend in the NMB (Section 10.5), further support the concept of two different coarse clastic systems, one to the east and one to the west, within the overall CMB depositional systems tract.

Fine-grained lithostratigraphic units are present within the boundaries of the CMB's fluvial conglomeratic assemblage. Iron formation is present as a thin (thickness unknown, probably tens of metres or less) unit extending 2.7 km along strike in the northern, upper CMB of Dorothea Township (Section 6.8; Coleman and Moore, 1907; Mackasey, 1975), and at the north margin of the CMB in Leduc Township (Section 6.8). The approximately 120 m thick mudstone unit in Leduc Township (Section 6.4) is shown by Mackasey (1976) as a 2.9 km long lenticular horizon. This mudstone extends from the resedimented conglomeratic assemblage at Beatty Lake (Chapter 7; Section 11.13) to a fluvial conglomeratic area near Leduc Lake.

It is unclear whether the above aquabasinal facies (iron formation, mudstone) and the turbiditic sandstone assemblage in Walters Township (Chapter 8; Section 11.14), all sporadically present for kms along strike in the upper CMB, are the deposits of highly proximal inter-fan lakes (e.g. McGowen and Groat, 1971), lakes within a braidplain, or a marine transgression (regional submergence).

There is little significant variation in clast composition along strike within the CMB, just as in the NMB.

12.5 Speculative Stratigraphic Interpretations Based on Stratigraphic Context

Much of the discussion below pertains specifically to the Beatty Lake area; refer to Figs. 3.1, 12.1, and Table 12.1.

The three possibly coastal facies--herringbone cross-beds, mud laminae and flasers, and plane laminated sandstone--are closely spatially associated within outcrop #100 and are located at the local (Beatty Lake) and regional (CMB) boundary between subaerial (fluvial) and subaqueous (aquabasinal) paleoenvironments, near the middle of the CMB's section. One could not ask for a more appropriate site at which to find intertidal (herringbone) cross-beds and other associated coastal facies.

Outcrop #89's swaley cross-strata are very close to fluvial facies (outcrops #84, 85). As the swaley cross-strata probably reflect storm wave activity, a shallow aquabasinal setting above storm wave base can be inferred for outcrop #89, with a paleoshoreline necessary between the aquabasinal and fluvial facies. This paleoshoreline is probably exposed in outcrop #100 (as noted above), which is approximately 1.1 km to the west along strike (Devaney and Fralick, 1985, Fig. 16.2).

The subaqueously deposited resedimented (heterogeneous) assemblage was part of a fan-delta front/slope, or a submarine fan(s) below and linked to a fan-delta slope channel, or a submarine fan(s) with its source unrelated to the local paleoslopes (i.e. a more distant input system). Ojakangas (1985) mentions the possible contribution of slumping and sediment gravity flows on steep, subaqueous slopes of felsic volcanoes.

The resedimented (heterogeneous) assemblage is in the middle of the CMB, near where any inferred paleoshoreline should be. It is possible that resedimented conglomerate and sandstone turbidites may have been deposited at relatively shallow depths as discussed in detail in Section 11.13. This lends support to the hypothesis that the heterogeneous assemblage may have been at least partly deposited on a (subaqueous) fan-delta front/slope, rather than a very deep-water fan fed by a channel(s) incised into a delta front or basinal slope.

The presence of gravelly aquabasinal sediments in the heterogeneous assemblage and Unit A's non-conglomeratic character are not problematic. Gravelly sediment could have bypassed a coastal sandy fluvial system (e.g. via erosional incision of a river valley) and accumulated in an aquabasin, or a coastal gravelly braided river system debouched directly into an aquabasin at a different time than the intervals during which sandy river sediments were accumulating.

Units A, B, and C may be the distal, medial and proximal facies assemblages, respectively, of a prograding braidplain or alluvial fan. The gravelly fraction of Units A, B and C shows well the overall coarsening- and thickening-upward trend. Sections in Units A, B and C's outcrops are similar to those given by McGowen and Groat (1971) in their classic description and model of fluvial fan (syn. humid alluvial fan of Boothroyd and Nummedal, 1978; wet alluvial fan of Schumm, 1977) facies, and conform to the proximal-distal relations common in braided river systems (Table 12.2).

TABLE 12.2

Parameter	Proximal	Distal (Downstream)
Grain size (mean and maximum)	coarser (gravelly)	finer (sandy, and probably better sorted)
Stratification	horizontal	cross-strata
Bars	longitudinal	linguoid
Sand channels	shallow (often upper flow regime)	deeper (usually) (often lower flow regime)
Sandy facies	Sm, Sh, Sp	Sp, St

Table 12.2: Proximal-distal criteria for braided river systems. Based on Smith (1970), McGowen and Groat (1971), Boothroyd and Nummedal (1978), Collinson (1978).

The coarsest conglomerate exposure in the CMB, outcrop #109 (Tables 6.2, 6.3), is at the top of this prograded sequence. Its largest clasts are 400/850 mm and 320/1230 mm (short axis/long axis; long axis lies within the schistosity), with a D10 of 281/638 mm. The top of a prograded alluvial fan-fluvial sequence would indeed be an appropriate place to find such a bouldery conglomerate.

Fan-head entrenchment causes sediment to bypass the proximal parts of a fan and to be deposited on a new, active fan segment. These have been variously termed the new segment of segmented alluvial fans (Bull, 1977, Fig. 20), deposition at the fan toe (Schumm, 1977, Fig. 7-1), and secondary fans (Heward, 1978). Tectonism and climate can be extrinsic controls, and Schumm (1977, pp.246-252, pp.255-264) shows that geomorphic (slope) thresholds can control the growth of both dry and wet fans. In an experiment on fan growth, Schumm (1977, p.258) found that

> "The alluvial fan grew progressively with time, but growth at the fan head was episodic, being interrupted by periods of incision, sediment reworking, and down-fan distribution of sediment."

This episodicity suggests an intrinsically pulsating fan system. Such a fluctuating system would not show a simple down-fan fining trend, a trend which is considered characteristic of alluvial fans (e.g. Nilsen, 1982; Rust and Koster, 1984), but long term coalescence of fans and fan segments should minimize complications due to secondary fan growth (Heward, 1978).

This episodic secondary fan growth is a possible mechanism for delivering abnormally coarse sediment to normally finer, "distal" areas. This may account for the parts of Unit B that are coarser and richer in conglomerate than parts of Unit C, small-scale variations within the CMB's overall coarsening-upward sequence.

Determining which of volcanism, tectonism, climate, avulsion, or episodic fan growth was most significant is far beyond the resolution of the data.

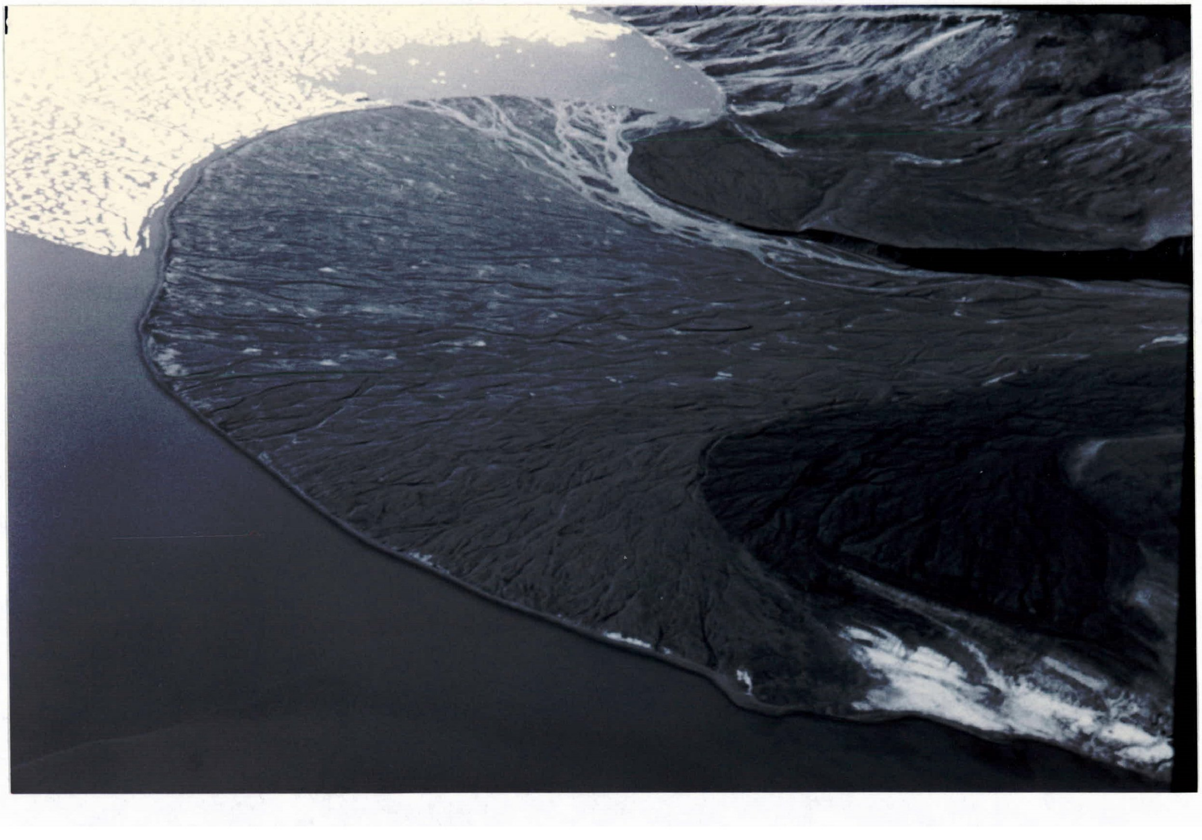
The entire CMB section of lithofacies assemblages (Table 12.1) could easily have been part of a prograding fan-delta depositional system (Figs. 12.2, 12.3; see also terms in Table 11.1). In this scenario, basin plain and prodelta muds and thinly bedded distal turbidites were buried by prograding (laterally accreting) fan-delta front/slope deposits, the resedimented (heterogeneous) assemblage. As the depositional regression continued, the shoreline was extended basinward and rivers flowed further basinward over a fan-delta plain, the subaerial part of a coastal alluvial fan or bajada (laterally coalesced fans or braidplain) gradually filling in the adjacent aquabasin (Figs. 12.2 to 12.4). Note that this use of context, the fan-delta model, was not mentioned by Ojakangas (1985, Fig. 19) in his otherwise excellent review of the Canadian Shield's Archean clastic sedimentation.

The general lack of coastal and shallow marine features is typical of Archean sedimentary successions (Ojakangas, 1985), and is likely the result of rapid facies changes from fan-delta plains to steep fan-delta fronts (deltaic topsets to foresets and/or bottomsets), with only a very restricted transition zone (beach, shoreface, etc.) in between (Wood, 1980; Ethridge and Wescott, 1984).

Synchronous uplift or volcanic edifice growth is helpful but not required in such progradational scenarios; for example, a eustatic sea level fall could expose previously

__Fig. 12.3:__ Fan-deltas prograding into sea water. Seven fan-deltas are present along about 20 km of coastline. The nearest fan-deltas show subaerial lateral coalescence. Gravelly braided rivers are actively flowing. Cañon Fiord, Ellesmere Island, Northwest Territories.

__Fig. 12.4:__ Detail of Fig. 12.3. Fan delta (coastal alluvial fan). Note the active river channels, inactive gravelly fan-delta plain, erosional incision into older deposits, radial paleocurrent pattern, and the very narrow beach. Oblique aerial photograph.



submarine topography to subaerial erosion. Indeed, if the rate of fluvio-deltaic sedimentation exceeded the relative rates of local subsidence and eustatic sea level rise, a local depositional regression of the shoreline would take place during a more widespread transgression.

CHAPTER 13: REGIONAL GEOLOGICAL INTERPRETATIONS, AND SUMMARY

13.1 Correlation of the CMB with the NMB

Both the data for individual point count localities and the closely similar average composition values of the NMB and CMB in Tables 10.4, 10.5 and 10.7 strongly suggest that the NMB and CMB had either the same source terrain, or source terrains of identical provenance. As the NMB and CMB also:

1) are closely adjacent, regionally parallel lithologic belts (Fig. 1.1);
2) have very similar to identical conglomeratic lithofacies assemblages and fluvial paleoenvironments (Section 11.12); and
3) have D10 measurements (Fig. 2.8; compare Tables 2.1 and 6.2) clearly showing the NMB to be coarser than the CMB, even with the NMB's clasts having typically greater L/S ratios;

it seems extremely probable that the CMB is a more distal, lithostratigraphic and chronostratigraphic equivalent of the NMB. This interpretation is further supported by distal features such as:

1) finer grained and better sorted conglomerate and sandstone (Smith, 1970; Miall, 1977; Boothroyd and Nummedal, 1978);
2) fluvial and aquabasinal sandstone-rich sections;
3) mudstone interbedded with fluvial conglomerate

(e.g. Donjek model of Miall, 1977, 1978a) and as intraformational mud chip clasts; and

4) abundant exposure of aquabasinal facies

which are present only in the CMB, with exception of the NMB's relatively minor anomalous aquabasinal facies (Sections 2.6, 11.2, 11.5). The CMB's aquabasinal resedimented conglomerate (Section 11.13) has the same provenance as all of: the local (Beatty Lake area) fluvial conglomerate, the rest of the epiclastic conglomeratic CMB, and the NMB.

It is thought that the formerly continuous NMB-CMB depositional systems tract (term of Miall, 1984, Fig. 1.1) was later separated by faulting and/or folding into the discrete metasedimentary belts shown in Fig. 1.1 and published geological maps (Mackasey, 1975, 1976; Stott, 1984a, b).

13.2 Correlation of the SMB with the CMB and NMB, and Regional Implications

The CMB is interpreted to be a coarsening-northward, generally coarsening-upward megasequence. It is postulated that the NMB may also be a north-younging sequence, based on the younging to the north of its basal conglomerate horizon (Section 11.5), and very few other local top indications based on sedimentary structures and stratification features.

The SMB is possibly also a formerly and dominantly north-younging megasequence. Recent structural studies of the SMB are those of Kehlenbeck (1983), Carter (1983), Anglin and Franklin (1985) and Williams (1985).

The relative positions of the various sedimentological facies present in the NMB, CMB and SMB are more important than local top directions in attempting to reconstruct a basinal paleogeography. The NMB is composed largely of the most proximal fluvial conglomerate. The CMB is a dominantly coarsening-upward sequence of:

1) lower and distal aquabasinal mudstone, turbiditic sandstone and minor iron formation;
2) medial fluvio-deltaic sandstone and minor resedimented conglomerate;
3) upper and proximal fluvial conglomerate-rich strata; and
4) in Walters and Irwin Townships (Fig. 1.1, Chapters 4, 8) only, an uppermost sandstone-mudstone assemblage (turbidites in Walters Township; Section 11.14).

The SMB is composed largely of aquabasinal turbiditic sandstone and mudstone with lesser amounts (less than a few percent) of iron formation and conglomerate (Mackasey, 1970a, b; Carter, 1983; Kehlenbeck, 1983; Anglin and Franklin, 1985; Barrett and Fralick, 1985; Williams, 1985; and personal observations). Note the similarity between the lower CMB and the SMB. Conglomerate beds in the SMB have the same provenance as those in the NMB and CMB, based on field observations of most of the conglomeratic outcrop areas in the SMB. It is thus highly probable that the SMB is the most distal component of a formerly linked tract (Langford, 1929;

Ayres, 1969; Mackasey, 1970a, 1972, 1975, 1976; Mackasey *et al*, 1974; Devaney and Fralick, 1985). According to Blackburn *et al* (1985, p.109), such regional subaerial to subaqueous, proximal to distal facies changes characterize the margins of western Wabigoon Subprovince. Later tectonism probably produced the now-separate belts (Fig. 1.1), as discussed in Section 13.1.

Any regional structural interpretation of the Beardmore-Geraldton terrane (Fig. 1.1) should account for:

1) the region's numerous northward dips of bedding (e.g. the CMB, many 45°-70° N dips);
2) the separation of the proximal fluvial NMB from the distal fluvial CMB;
3) the separation of the CMB from its more distal probable equivalent, the SMB; and
4) the SMB-CMB-NMB triad of metasedimentary belts mimicking both the CMB's coarsening-north trend and the CMB's aquabasinal to fluvial (proximal to the north) trend.

This thesis supports Mackasey's (1976, p.15) hypothesis that the NMB, CMB and SMB were

> "related to the same depositional basin and ...formed a laterally continuous succession before folding"

and Mackasey's (1975, p.30) postulation that the NMB "is an up-faulted segment" of the CMB. Williams (1985) has argued a case for underthrusting from the south and dextral shearing in order to account for the regional lithostratigraphic and structural patterns.

13.3 Potential Sources of the Clastic Sediments in the NMB and CMB

Much of the NMB is adjacent to a regional fault, the Paint Lake Fault (Fig. 1.1; Mackasey, 1975, 1976). The NMB's conglomerate clasts are coarsest immediately along the Paint Lake Fault in Walters Township (Fig. 1.1; Table 2.1; Fig. 2.8). Both the various lithologies present as clasts in the NMB and CMB (Tables 10.2, 10.3, 10.7, 10.8 and 10.9) and conglomerate and sandstone of similar lithofacies and provenance (Moorhouse, 1937; Amukun, 1980) are exposed today within the Onaman-Tashota area (Fig. 1.1). Braided rivers, including alluvial fan deposits, may have flowed generally southward from such a source area (e.g. Bruce, 1937). If the Paint Lake Fault was active during sedimentation, it could be an excellent way (e.g. a fault scarp) to produce proximal bouldery gravels adjacent to a fault, such as in Walters Township (Table 2.1, Fig. 2.8).

The Onaman-Tashota area (Fig. 1.1; Stott, 1984a, b) is the likely type of source region, one rich in felsic volcanic (particularly pyroclastic) and granitoid lithologies (Mackasey, 1975, 1976; Mackasey and Wallace, 1978). Because of its present geographic proximity to the Beardmore-Geraldton terrane (Fig. 1.1) it is tempting to speculate that the Onaman-Tashota area was the source terrain for the clastic sediments now present to the south, but the total movement along any regional east-west faults, such as the

Paint Lake Fault (Fig. 1.1; Mackasey, 1975, 1976), is unknown. In light of the recent research on accreted (or "exotic") terranes and transcurrent fault motions (Jones *et al*, 1977; Irving *et al*, 1980), great caution in interpretation is required.

Adding to the uncertainty regarding the clasts' felsic volcanic-rich provenance is the question of whether the erosion of the source terrain(s) was:

1) synvolcanic, the product of the erosion of active felsic volcanic complexes; or
2) post-volcanic, the product of the erosion of an area where volcanism has permanently terminated, leaving "volcanic highlands"; or
3) syntectonic to post-tectonic, the product of the erosion of a tectonically deformed (mountainous?) area.

The apparently very similar composition of clasts in the NMB and CMB and lithofacies in the Onaman-Tashota area does not preferentially favour any one of the above three choices. The lack of multiply foliated and high-grade metamorphic clasts (e.g. gneisses) slightly favours 1 and 2 above. Much petrographic study and radiometric age dating of both clasts and potential source areas would be required to pursue clues to such problems.

Can a completely eroded away, pre-existing Archean conglomeratic sequence be ruled out as a source of NMB and CMB clasts? Probably yes, because the felsic volcanic-rich provenance of the NMB and CMB conglomerates so closely approximates both the ikely composition of subaerially exposed Archean volcanic areas (Thurston *et al*, 1985; Ojakangas, 1985; Ayres and Thurston, 1985) and the felsic volcanic centres exposed today in Superior Province, and no conglomerate clasts are present.

Sediment transport from the north is implied by the regional positions and character of the NMB, CMB and SMB (Fig. 1.1), as noted previously by Langford (1929) and Ayres (1969). There is a general lack of evidence for any progradation from the east or west, and only weak evidence for such in the CMB (Sections 10.5, 12.4). An extremely limited number of paleocurrents in deformed strata of the NMB and CMB are to the east, south and west, but never to the north.

As the magnitudes of possible transcurrent motions (e.g. Williams, 1985) of the lithologic belts in Fig. 1.1 are unknown, the geographic source area of the NMB's and CMB's conglomerate clasts cannot be specified. The composition data, Tables 10.2 to 10.5, suggest a felsic volcanic centre. The Onaman-Tashota area could easily have been the source of clastic sediments. Remember also that the source area may have been completely eroded away (Section 10.5), leaving an epiclastic sequence rich in volcanic detritus as the indirect record of felsic volcanism (Ayres, 1983; Thurston *et al*, 1985; Ojakangas, 1985; Ayres and Thurston, 1985).

13.4 Relevance of This Study to Archean Stratigraphy and Sedimentology

This thesis adds to our knowledge of the Archean's Resedimented (Turbidite) and Alluvial Fan-Fluvial Facies Associations (status quo in Ojakangas, 1985). The results obtained are similar to those from other studies of Archean metasedimentary successions in Superior Province (e.g. Teal, 1979; Hyde, 1980; Wood, 1980; Gordanier, 1982; Dimroth *et al*, 1982b) and other shields (e.g. Eriksson, 1978, 1980, 1981), and are typical in that they reflect the importance of felsic volcanic sources of epiclastic material and a general absence of shallow marine shelf facies (Ojakangas, 1985; Ayres and Thurston, 1985).

The delta front, intertidal and shallow aquabasinal (marine?) sandstone facies of the CMB described and interpreted herein are important new finds. The author knows of no other reports from Superior Province of hummocky and swaley cross-strata or of delta front (couplets) facies, and has read of (and seen) only one other set of intertidal exposures in Superior Province (Parker, 1980). Also, this thesis' resedimented conglomerates are probably the first Archean ones to be interpreted as possible fan-delta front deposits (Devaney and Fralick, 1985), a result of the very recent emphasis on fan-delta processes and products (see Sections 11.13, 12.5; Wescott and Ethridge, 1980; Koster and Steel, 1984) and the apparent general applicability of stratigraphic context (Walther's Law) in analyzing the CMB sequence. Eriksson (1978, 1980), Wood (1980) and Ethridge and Wescott (1984) have briefly discussed the role of fan-deltas in the Archean.

Ayres (1969, p.312) recommended the Beardmore-Geraldton area as "probably the best place to attempt a sedimentological study" of the transitions at the edges of Superior Province's Subprovinces. The margins of western Wabigoon Subprovince are characterized by a transition from terrestrial to deep water deposits (Blackburn *et al*, 1985), a sedimentary polarity long recognized in eastern Wabigoon Subprovince's Beardmore-Geraldton terrane (Fig. 1.1) by Langford (1929), Ayres (1969) and Mackasey *et al* (1974). Goodwin (1977) inferred the presence of a basin margin in the Beardmore-Geraldton area and was criticized by Walker (1978a). One must be very careful in distinguishing fluvial and resedimented facies (pebbly sandstones and conglomerates) as they can have very similar appearances (Turner and Walker, 1973; Winn and Dott, 1977, 1979; Walker, 1978a, 1984a; Hein, 1984; Ojakangas, 1985), particularly in deformed Precambrian strata lacking imbrication and paleontological data. This thesis strongly suggests that Goodwin (1977) was partly correct in postulating a basinal subaerial-subaqueous facies transition in the Beardmore-Geraldton area. Many earlier workers (Tanton, 1935; Pye, 1968; Mackasey *et al*, 1974; Mackasey, 1975, 1976) have noted the probable lithostratigraphic correlation of the Beardmore-Geraldton terrane's NMB, CMB and SMB with similar units west of Lake Nipigon (Fig. 1.1); aeromagnetic trends seem to confirm this correlation (Mackasey *et al*, 1974; Mason *et al*, 1985, p.4), but this need not imply any exact chronostratigraphic equivalence.

The term "terrane" has been used because the possibly anomalous nature of the Beardmore-Geraldton area's Archean rocks may not be best summarized as an "Archean greenstone belt" (*status quo* in Condie, 1981). The Beardmore-Geraldton terrane (Fig. 1.1) may be a regional zone of shear (M.M. Kehlenbeck, personal communication, 1983; Williams, 1985) located at the southern margin of Wabigoon Subprovince.

The above use of the term "terrane" does not really conflict with its present usage either. Many of the principles of contemporary suspect terrane analysis apply equally well to Archean geology: formation- to supergroup-scale stratigraphic units may be allochthonous, and multi-disciplinary study (e.g. radiometric age dating, paleontology, sedimentology and volcanology, stratigraphic correlation, structural analysis) is required.

Teal (1979, p. 182, 183) stated:

> "It appears that South Africa, which shows good evidence of stabilized cratons in the Archean, had a tectonic behaviour similar to younger regions, resulting in a stratigraphic sequence similar to the Phanerozoic flysch-molasse sequence, whereas the Canadian Shield, which lacks evidence of stabilization, had a fundamentally different tectonic behaviour in the Archean, specifically, foundering of the basin after non-marine deposition and/or felsic pyroclastic volcanism."

Hyde (1980) referred to this record of foundering as an "up-down sequence". Both Teal and Hyde may have been premature in inferring the "fundamental style" of Archean sedimentary successions; this thesis is one of the many more studies needed in order to try to unravel and reveal the paleogeographic and tectonic evolution of Archean greenstone belts. "More work needs to be done ... the more we look, the more we see" (Ojakangas, 1985, p.41, 42).

13.5 Concluding Summary

Gravelly braided rivers, parts of fans and/or braidplains, deposited most of the northern metasedimentary belt (NMB) with a relatively minor (about 5-8% of areal extent) aquabasinal facies assemblage in Irwin Township (Fig. 1.1) only. A basal conglomerate is exposed in Legault Township (Fig. 1.1).

The easternmost 7 km of the exposed central metasedimentary belt (CMB: Fig. 1.1) is composed of generally oligomict, coarse felsic pyroclastics (primary) and/or volcaniclastics (reworked) deposited by volcanic, braided river and sedimentary gravity flow processes.

The CMB's epiclastic lithofacies assemblages, or depositional paleoenvironments, are laterally continuous on a sub-regional (tens of kms) scale. Sedimentological trends vertically cross-cut the lithofacies assemblages.

The relative positions of the assemblages/paleoenvironments suggest that the epiclastic portion of the 1-2 km thick CMB is a dominantly coarsening-northward, prograded aquabasinal to fluvial megasequence (Fig. 12.1), similar to stratigraphic successions produced by prograding fan-deltas. The role of fan-deltas in Archean basins has probably been underestimated in the past by many workers.

The lower CMB is an aquabasinal assemblage of mudstone, iron formation and fine, often turbiditic sandstone. The middle horizon of the CMB contains (a) a sandstone assemblage representing:

1) storm wave-influenced facies;
2) a delta front paleoenvironment;
3) sandy braided river paleoenvironments, with an extremely rare (one tiny rock knob) intertidal subfacies,

and (b) also has a resedimented conglomeratic assemblage that was part of a submarine fan or fan-delta front. The upper CMB is most often a conglomeratic braided fluvial assemblage, with relatively minor mudstone and iron formation of aquabasinal origin but unknown paleoenvironmental context. A limited exposure of a turbiditic sandstone-rich aquabasinal assemblage appears to overlie the fluvial conglomerate in Walters Township (Fig. 1.1) only, and aquabasinal facies may form most of the CMB in Irwin Township (Fig. 1.1).

The facies that are of possible intertidal, delta front and shallow marine storm wave-influenced origin are new finds important to Archean sedimentology and stratigraphy because of their general rarity in Superior Province. These facies are located in the CMB exactly where they should be present in a progaded stratigraphic section, above deep-water aquabasinal facies and below fluvial facies.

Within the CMB's generally prograded sequence, two separate systems (e.g. fan-deltas), one in the east and one in the west, may have been present.

Potential structural complexity partially limits confidence in the stratigraphic history outlined above, but does not affect the interpretations of the individual depositional paleoenvironments.

Sedimentological details (e.g. mean and maximum deformed particle size, bedding, sedimentary structures) and both compositional homogeneity of and minor variations in clast lithologies strongly suggest that the NMB is the source-proximal extension of the CMB. Later tectonism likely separated the formerly continuous NMB-CMB depositional systems tract into the discrete metasedimentary belts (regional to sub-regional scale lithostratigraphic units) exposed today. Measurements of the average dimensions of an outcrop's ten largest deformed clasts (D10), usually granitoids, were surprisingly effective in showing sub-regional proximal-to-distal trends.

The southern metasedimentary belt (SMB) is probably the ultradistal component of a formerly continuous NMB-CMB-SMB, fluvial to aquabasinal depositional systems tract. Sedimentary facies analysis can provide important clues to paleotectonic evolution of Archean volcano-sedimentary basins such as the one partially preserved in the western part of the

Beardmore-Geraldton terrane (Fig. 1.1)

Although the Onaman-Tashota area north of the study area (Fig. 1.1) is a potential source region, the exact source of the sediments in the NMB and CMB cannot be specified at this time. Landforms with felsic volcanic bedrock, pyroclasts or erosional detritus were the ultimate source of most of the clastic material supplied to the NMB-CMB systems. This thesis does not prove but is consistent with the contemporary view that subaerial felsic volcanic edifices active during the Archean supplied clastic sediment downslope to fluvial and more distal aquabasinal paleoenvironments.

13.6 Acknowledgements

Members of the Ontario Geological Survey have been influential in my Archean studies. Jim Pirie, Paul Bowen and Gary Beakhouse taught me by their example to carefully examine Archean rocks. Conversations over the years with Phil Thurston, John Wood, Henry Wallace, Sandy Colvine and James Macdonald have been most educational.

John Mason and Gerry White of the Ontario Ministry of National Resources Thunder Bay office have been very helpful.

Andrew Butler provided heroic field assistance. Borys Zayachkivsky helped out too.

Iain Hastie and Sam Spivak re-drafted the diagrams.

A grant (Research Agreement 176) from Energy, Mines and Resources, Canada, partially funded this study.

Last but not least, thanks to my advisor, Prof. Phil Fralick, for his friendship, ideas, money, help in the field and years of patience.

REFERENCES

Ahlbrandt, T.C. and Fryberger, S.G. 1982: Introduction to eolian deposits; p.11-47 in: Scholle, P.A. and Spearing, D., eds.: Sandstone Depositional Environments, Amer. Assoc. of Petroleum Geologists, Memoir 31.

Alam, M.M., Crook, K.A.W. and Taylor, G. 1985: Fluvial herring-bone cross-stratification in a modern tributary mouth bar, Coonamble, New South Wales, Australia. Sedimentology, 32, 235-244.

Allen, J.R.L. 1983: Gravel overpassing on humpback bars supplied with mixed sediment: examples from the Lower Old Red Sandstone, South Britian. Sedimentology, 30, 285-294.

Amukun, S.E. 1980: Geology of the Conglomerate Lake area, District of Thunder Bay. Ontario Geol. Survey, Report 197, 101 p.

Anglin, C.D. and Franklin, J.M. 1985: Gold mineralization in the Beardmore-Geraldton area of northwestern Ontario: structural considerations and the role of iron formation; p. 193-201 in: Current Research, Part A, Geol. Survey of Canada, Paper 85-1A.

Anhaeusser, C.A., Mason, R., Viljoen, M.J. and Viljoen, R.P. 1969: A reappraisal of some aspects of Precambrian Shield geology. Geol. Soc. Amer. Bull., 80, 2175-2200.

Ayres, L.D. 1969: Early Precambrian stratigraphy of part of Lake Superior Provincial Park, Ontario, Canada, and its implications for the origin of Superior Province. Unpublished PhD dissertation, Princeton Univ., 399 p.

Ayres, L.D. 1983: Bimodal volcanism in Archean greenstone belts exemplified by greywacke composition, Lake Superior Park, Ontario. Can. Jour. of Earth Sci., 20, 1168-1194.

Ayres, L.D. and Thurston, P.C. 1985: Archean supracrustal sequences in the Canadian Shield: an overview; p.343-380 in: Ayres, L.D., Thurston, P.C., Card, K.D. and Weber, W., eds.: Evolution of Archean Supracrustal Sequences, Geol. Assoc. of Canada, Special Paper 28.

Ayres, L.D., Thurston, P.C., Card, K.D. and Weber, W., eds. 1985: Evolution of Archean Supracrustal Sequences. Geol. Assoc. of Canada, Special Paper 28, 380 p.

Ballance, P.F. 1984: Sheet-flow dominated gravel fans of the non-marine middle Cenozoic Simmler Formation, central California. Sedim. Geol., 38, 337-359.

Barrett, T.J. and Fralick, P.W. 1985: Sediment redeposition in Archean iron formation: examples from the Beardmore-Geraldton greenstone belt, Ontario. Jour. Sedim. Petrol., 55, 205-212.

Beakhouse, G.P. 1984: Geology of the Grenville Lake area, Thunder Bay District. Ontario Geol. Survey, Open File Report 5513, 146 p. Acc. by Maps P2592, P2593.

Bell, R. 1869: Geological report. Geol. Survey of Canada, Summary Report, p. 313-364.

Blackburn, C.E., Bond, W.D., Breaks, F.W., Davis, D.W., Edwards, G.R., Poulsen, K.H., Trowell, N.F. and Wood, J. 1985: Evolution of Archean volcanic-sedimentary sequences of the western Wabigoon Subprovince and its margins: a review; p.89-116 in: Ayres, L.D., Thurston, P.C., Card, K.D. and Weber, W., eds.: Archean Supracrustal Sequences, Geol. Assoc. of Canada, Special Paper 28.

Blodgett, R.H. and Stanley, K.O. 1980: Stratification, bedforms and discharge relations of the Platte braided river system, Nebraska. Jour. Sedim. Petrol., 50, 139-148.

Bluck, B.J. 1974: Structure and directional properties of some valley sandur deposits in southern Iceland. Sedimentology, 21, 533-554.

Bluck, B.J. 1979: Structure of coarse grained braided stream alluvium. Trans. Royal Soc. of Edinburgh, 70, 181-221.

Boothroyd, J.C. and Ashley, G.M. 1975: Processes, bar morphology, and sedimentary structures on braided outwash fans, northeastern Gulf of Alaska; p.193-222 in: Jopling, A.V. and McDonald, B.C., eds.: Glaciofluvial and Glaciolacustrine Sedimentation, Soc. of Economic Paleontologists and Mineralogists, Sp. Pub. 23.

Boothroyd, J.C. and Nummedal, D. 1978: Proglacial braided outwash: a model for humid alluvial-fan deposits; p.641-668 in: Miall, A.D., ed.: Fluvial Sedimentology, Can. Soc. of Petroleum Geologists, Memoir 5.

Bourgeois, J. and Leithold, E.L. 1984: Wave-worked conglomerates - depositional processes and criteria for recognition; p.331-343 in: Koster, E.H. and Steel, R.J., eds.: Sedimentology of Gravels and Conglomerates, Can. Soc. of Petroleum Geologists, Memoir 10.

Brayshaw, A.C. 1984: Characteristics and origin of cluster bedforms in coarse-grained alluvial channels; p.77-85 in: Koster, E.H. and Steel, R.J., eds.: Sedimentology of Gravels and Conglomerates, Can. Soc. of Petroleum Geologists, Memoir 10.

Bruce, E.L. 1937: The eastern part of the Sturgeon River area. Ontario Dept. Mines, v.45, part 2, 1936, 1-59. Acc. by Map 45a, scale 1:63 360.

Buck, S. and Williams, H.R. 1984: Structural studies in the Geraldton area; p.208-212 in: Wood, J., White, O.L., Barlow, R.B. and Colvine, A.C., eds.: Summary of Field Work, 1984, by the Ontario Geological Survey. Ontario Geol. Survey, Misc. Paper 119.

Bull, W.B. 1977: The alluvial fan environment. Progress in Physical Geography, 1, 222-270.

Burrows, A.G. 1917: Longuelac to Jellicoe and Orient Bay, Ontario Bureau of Mines, v.26, 227-247.

Callander, R.M. 1970: Windigokan Conglomerate. B.Sc. Thesis, Univ. of Waterloo, 89 p.

Cant, D.J. 1978: Development of a facies model for sandy braided river sedimentation: comparison of the South Saskatchewan River and the Battery Point Formation; p.627-639 in: Miall, A.D., ed.: Fluvial Sedimentology, Can. Soc. of Petroleum Geologists, Memoir 5.

Cant, D.J. 1982: Fluvial facies models and their application; p.115-137 in: Scholle, P.A. and Spearing, D., eds.: Sandstone Depositional Environments, Amer. Assoc. of Petroleum Geologists, Memoir 31.

Card, K.D., Percival, J.A. and Coe, K. 1980: Progress report on regional geological synthesis, central Superior Province; p.61-68 in: Current Research, Part A, Geol. Survey of Canada, Paper 80-1A.

Carter, M.W. 1983: McComber and Vincent Townships, District of Thunder Bay; p.32-36 in: Wood, J., White, O.L., Barlow, R.B. and Colvine, A.C., eds.: Summary of Field Work, 1983, by the Ontario Geological Survey. Ontario Geol. Survey, Misc. Paper 116.

Chough, S.K. and Hesse, R. 1980: The northwest Atlantic mid-ocean channel of the Labrador Sea: III. Head spill vs. body spill deposits from turbidity currents on natural levees. Jour. Sedim. Petrol., 50, 227-234.

Church, M. 1972: Baffin Island sandurs. Geol. Survey of Canada, Bulletin 216, 208 p.

Church, M. and Gilbert, R. 1975: Proglacial fluvial and lacustrine environments; p.22-100 in: Jopling, A.V. and McDonald, B.C., eds.: Glaciofluvial and Glaciolacustrine Sedimentation, Soc. of Economic Paleontologists and Mineralogists, Sp. Pub. 23.

Coleman, A.P. and Moore, E.S. 1907: I. - Iron ranges east of Lake Nipigon. II. - Iron ranges east of Lake Nipigon; the ranges around Lake Windegokan. Ontario Bureau of Mines, v.16, p.105-148.

Coleman, A.P. and Moore, E.S. 1908: I. - Iron ranges east of Lake Nipigon. II. - Iron ranges east of Lake Nipigon: The Onaman iron ranges. Ontario Bureau of Mines, v.17, 136-189. Acc. by maps at scales 1:63 360, 1:31 680.

Coleman, J.M. 1969: Brahmaputra River: channel processes and sedimentation. Sedim. Geol., 3, 129-239.

Collinson, J.D. 1970: Bedforms of the Tana River, Norway. Geografiska Annaler, 52A, 31-56.

Collinson, J.D. 1978: Alluvial sediments; p.15-60 in: Reading, H.G., ed.: Sedimentary Environments and Facies, Elsevier.

Condie, K.C. 1981: Archean Greenstone Belts. Elsevier, 434 p.

Davies, D.K., Vessell, R.K., Miles, R.C., Foley, M.G. and Bonis, S.B. 1978: Fluvial transport and downstream sediment modifications in an active volcanic region; p.61-84 in: Miall, A.D., ed.: Fluvial Sedimentology, Can. Soc. of Petroleum Geologists, Memoir 5.

Davies, I.C. and Walker, R.G. 1974: Transport of deposition of resedimented conglomerates: the Cap Enrage Formation; Cambro-Ordovician, Gaspe, Quebec. Jour. Sedim. Petrol., 44, 1200-1216.

Devaney, J.R. 1982: Stratigraphy of the Volcan Lake Metaconglomerates: Archean braided river deposits. Undergraduate (GLG 471F) research project, Univ. of Toronto.

Devaney, J.R. and Fralick, P.W. 1985: Regional sedimentology of the Namewaminikan Group, northern Ontario: Archean fluvial fans, braided rivers, deltas, and an aquabasin; p.125-132 in: Current Research, Part B, Geol. Survey of Canada, Paper 85-1B.

Dimroth, E., Muller, W., Archer, P., Gobeil, A. and Allard, G.O. 1982a: Evidence for extensive shallow marine sedimentation in the Chibougamau area, Quebec; p.29-36 in: Current Research, Part A, Geol. Survey of Canada, Paper 82-1A.

Dimroth, E., Imreh, L., Rocheleau, M. and Goulet, N. 1982b: Evolution of the south-central part of the Archean Abitibi Belt, Quebec. Part I: Stratigraphy and paleogeographic model. Can. Jour. of Earth Sci., 19, 1729-1758.

Dimroth, E., Imreh, L., Cousineau, P., Leduc, M. and Sanschagrin, Y. 1985: Paleogeographic analysis of mafic submarine flows and its use in the exploration for massive sulphide deposits; p.203-222 in: Ayres, L.D., Thurston, P.C., Card, K.D. and Weber, W., eds.: Evolution of Archean Suprcrustal Sequences, Geol. Assoc. of Canada, Special Paper 28.

Doeglas, F.J. 1962: Structure of sedimentary deposits of braided rivers. Sedimentology, 1, 167-190.

Easton, R.M. 1984: Reconstruction of Precambrian volcanoes: a review of the Canadian literature; p.163-212 in: Easton, R.M. and Easton, M.G., eds.: Growth and evolution of volcanic edifices - with implications for Precambrian volcanoes. Geol. Assoc. of Canada, Short Course Notes, v.4.

Elliott, T. 1974: Interdistributary bay sequences and their genesis. Sedimentology, 21, 611-622.

Elliott, T. 1978a: Deltas; p.97-142 in: Reading, H.G., ed.: Sedimentary Environments and Facies, Elsevier.

Elliott, T. 1978b: Clastic shorelines; p.143-177 in: Reading, H.G., ed.: Sedimentary Environments and Facies, Elsevier.

Eriksson, K.A. 1978: Alluvial and destructive beach facies from the Archean Moodies Group, Barberton Mountain Land, South Africa and Swaziland; p.287-311 in: Miall, A.D., ed.: Fluvial Sedimentology, Can. Soc. of Petroleum Geologists, Memoir 5.

Eriksson, K.A. 1979: Marginal marine depositional processes from the Archean Moodies Group, Barberton Mountain Land, South Africa: evidence and significance. Precambrian Research, 8, 153-182.

Eriksson, K.A. 1980: Transitional sedimentation styles in the Moodies and Fig Tree Groups, Barberton Mountain Land, South Africa: evidence favouring an Archean continental margin. Precambrian Research, 12, 141-160.

Eriksson, K.A. 1981: Archean platform-to-trough sedimentation, east Pilbara Block, Australia; p.235-244 <u>in</u> Geol. Soc. of Australia, Sp. Pub. 7.

Eriksson, K.A. 1982: Geometry and internal characteristics of Archean submarine channel deposits, Pilbara Block, Western Austrailia. Jour. Sedim. Petrol., 52, 383-393.

Ethridge, F.G. and Wescott, W.A. 1984: Tectonic setting, recognition and hydrocarbon reservoir potential of fan-delta deposits; p.217-235 <u>in</u>: Koster, E.H. and Steel, R.J., eds.: Sedimentology of Gravels and Conglomerates, Can. Soc. of Petroleum Geologists, Memoir 5.

Eyles, N. and Miall, A.D. 1984: Glacial facies; p.15-38 <u>in</u>: Walker, R.G., ed.: Facies Models, Second Edition, Geoscience Canada, Reprint Series 1.

Fahnestock, R.K. 1963: Morphology and hydrology of a glacial stream - White River, Mount Ranier, Washington. U.S. Geol. Survey, Prof. Paper 422-A, 70 p.

Fisher, R.V. and Schmincke, H.-V. 1984: Pyroclastic Rocks. Springer-Verlag, 528 p.

Foley, M.G. 1977: Gravel-lens formation in antidune-regime flow - a quantitative hydrodynamic indicator. Jour. Sedim. Petrol., 47, 738-746.

Freund, J.E. 1984: Modern Elementary Statistics, Sixth Edition. Prentice-Hall Inc., 561 p.

Goodwin, A.M. 1977: Archean basin-craton complexes and the growth of Precambrian shields. Can. Jour. of Earth Sci., 14, 2737-2759.

Gordanier, W.D. 1982: Sedimentology and relationship to volcanology of Formation K, Favourable Lake metavolcanic-metasedimentary belt, northwestern Ontario. Unpublished M.Sc. Thesis, Univ. of Manitoba, 90 p.

Gustavson, T.C. 1974: Sedimentation on gravel outwash fans, Malaspina Glacier foreland, Alaska. Jour. Sedim. Petrol., 44, 374-389.

Hein, F.J. 1982: Depositional mechanisms of deep-sea coarse clastic sediments, Cap Enrage Formation, Quebec. Can. Jour. of Earth Sci., 19, 267-287.

Hein, F.J. 1984: Deep-sea and fluvial braided channel conglomerates: a comparison of two studies; p.33-49 in: Koster, E.H. and Steel, R.J., eds.: Sedimentology of Gravels and Conglomerates, Can. Soc. of Petroleum Geologists, Memoir 10.

Hein, F.J. and Walker, R.G. 1977: Bar evolution and development of stratification in the gravelly, braided, Kicking Horse River, British Columbia. Can. Jour. of Earth Sci., 14, 562-570.

Hein, F.J. and Walker, R.G. 1982: The Cambro-Ordovician Cap Enrage Formation, Quebec, Canada: conglomeratic deposits of a braided submarine channel with terraces. Sedimentology, 29, 309-329.

Heller, P.L. and Dickinson, W.R. 1985: Submarine ramp facies model for delta-fed, sand-rich turbidite systems. Amer. Assoc. of Petroleum Geologists Bull., 69, 960-976.

Hesse, R. and Chough, S.K. 1980: The northwest Atlantic mid-ocean channel of the Labrador Sea: II. Deposition of parallel laminated levee-muds from the viscous sublayer of low density turbidity currents. Sedimentology, 27, 697-711.

Heward, A.P. 1978: Alluvial fan sequence and megasequence models: with examples from Westphalian D - Stephanian B coalfields, Northern Spain; p.669-702 in: Miall, A.D., ed.: Fluvial Sedimentology, Can. Soc. of Petroleum Geologists, Memoir 5.

Hine, A.C. and Boothroyd, J.C. 1978: Morphology, processes and recent sedimentary history of a glacial-outwash plain shoreline, southern Iceland. Jour. Sedim. Petrol., 48, 901-920.

Hjulstrom, F. 1955: The ground water; p.234-245 in: Hjulstrom, F., Sundborg, A., Arnborg, A. and Johansson, J.: The Hoffellsandur - a glacial outwash plain. Geografiska Annaler, 37, 170-245.

Hogg, S.E. 1982: Sheetfloods, sheetwash, sheetflow or ...? Earth-Sci. Reviews, 18, 59-76.

Howell, D.G. and Normark, W.R. 1982: Sedimentology of submarine fans; p.365-404 in: Scholle, P.A. and Spearing, D., eds.: Sandstone Depositional Environments, Amer. Assoc. of Petroleum Geologists, Memoir 31.

Hunter, R.E. 1977: Basic types of stratification in small eolian dunes. Sedimentology, 24, 361-388.

Hunter, R.E. 1985: Subaqueous sand-flow cross strata. Jour. Sedim. Petrol., 55, 886-894.

Hyde, R.S. 1980: Sedimentary facies in the Archean Timiskaming Group and their tectonic implications, Abitibi greenstone belt, northeastern Ontario. Precambrian Research, 12, 161-195.

Irving, E., Monger, J.W.H. and Yole, R.W. 1980: New paleomagnetic evidence for displaced terranes in British Columbia; p.441-456 in: Strangway, D.W., ed.: The Continental Crust and Its Mineral Deposits, Geol. Assoc. of Canada, Special Paper 20.

Ito, M. 1985: The Ashigara Group: a regressive submarine fan - fan delta sequence in a Quaternary collision boundary, north of Izu Peninsula, central Honshu, Japan. Sedim. Geol., 45, 261-292.

Johnson, B.A. and Walker, R.G. 1979: Paleocurrents and depositional environments of deep water conglomerates in the Cambro-Ordovician Cap Enrage Formation, Quebec Appalachians. Can. Jour. of Earth Sci., 16, 1375-1387.

Johnson, H.D. 1978: Shallow siliciclastic seas; p. 207-258 in: Reading, H.G., ed.: Sedimentary Environments and Facies, Elsevier.

Jones, D.L., Silberling, N.J. and Hillhouse, J.W. 1977: Wrangellia - a displaced terrane in northwestern North America. Can. Jour. of Earth Sci., 14, 2565-2577.

Kehlenbeck, M.M. 1983: Structural studies in the Beardmore-Geraldton area; p.201-203 in: Wood, J., White, D.L., Barlow, R.B. and Colvine, A.C., eds.: Summary of Field Work, 1983, by the Ontario Geological Survey. Ontario Geol. Survey, Misc. Paper 116.

Kleinspehn, K.L., Steel, R.J., Johannessen, E. and Netland, A. 1984: Conglomeratic fan-delta sequences, Late Carboniferous-Early Permian, western Spitsbergen; p.279-294 in: Koster, E.H. and Steel, R.J., eds.: Sedimentology of Gravels and Conglomerates, Can. Soc. of Petroleum Geologists, Memoir 10.

Koster, E.H. 1978: Transverse ribs: their characteristics, origin and paleohydraulic significance; p.161-186 in: Miall, A.D., ed.: Fluvial Sedimentology, Can. Soc. of Petroleum Geologists, Memoir 5.

Koster, E.H. and Steel, R.J., eds. 1984: Sedimentology of Gravels and Conglomerates. Can. Soc. of Petroleum Geologists, Memoir 10, 441 p.

Krapez, B. 1985: The Ventersdorp Contact placer: a gold-pyrite placer of stream and debris flow origins from the Archean Witwatersrand Basin of South Africa. Sedimentology, 32, 223-234.

Krigstrom, A. 1962: Geomorphological studies of sandur plains and their braided rivers in Iceland. Geografiska Annaler, 44, 328-346.

Kuenzi, W.D., Horst, O.H. and McGehee, R.V. 1979: Effect of volcanic activity on fluvial-deltaic sedimentation in a modern arc-trench gap, southwestern Guatemala. Geol. Soc. Amer. Bull., 90, 827-838.

Laird, H.C. 1937: The western part of the Sturgeon River area. Ontario Dept. Mines, v.45, part 2, 1936, 60-117. Acc. by Map 45a, scale 1:63 360.

Lajoie, J. 1984: Volcaniclastic rocks; p.35-52 in Walker, R.G., ed.: Facies Models, Second Edition, Geoscience Canada.

Langford, G.B. 1929: Geology of the Beardmore-Nezah gold area, Thunder Bay District. Ontario Dept. Mines, v.37, part 4, 1928, 83-108. Acc. by Map No. 37k, scale 1:63 360.

Lavigne, M.J., Jr. 1983: Gold deposits of the Geraldton area; p.198-200 in: Wood, J., White, O.L., Barlow, R.B. and Colvine, A.C., eds.: Summary of Field Work, 1983, by the Ontario Geological Survey. Ontario Geol. Survey, Misc. Paper 116.

Lawson, A.C. 1913: The Archean geology of Rainy Lake re-studied. Geol. Survey of Canada, Memoir 40, 115 p.

Leeder, M.R. 1978: A quantitative stratigraphic model for alluvium, with special reference to channel deposit density and interconnectedness; p.587-596 in: Miall, A.D., ed.: Fluvial Sedimentology, Can. Soc. of Petroleum Geologists, Memoir 5.

Leopold, L.B. and Wolman, M.G. 1957: River channel patterns; braided, meandering and straight. U.S. Geol. Survey, Prof. Paper 282-B, 85 p.

Long, D.G.F. 1978: Proterozoic stream deposits: some problems of recognition and interpretation of ancient fluvial systems; p.313-341 in: Miall, A.D., ed.: Fluvial Sedimentology, Can. Soc. of Petroleum Geologists, Memoir 5.

Long, D.G.F. and Young, G.M. 1978: Dispersion of cross-stratification as a potential tool in the interpretation of Proterozoic arenites. Jour. Sedim. Petrol., 48, 857-862.

Lowe, D.R. 1975: Water escape structures in coarse-grained sediments. Sedimentology, 22, 157-204.

Lowe, D.R. 1982: Sediment gravity flows: II. Depositional models with special reference to the deposits of high-density turbidity currents. Jour. Sedim. Petrol., 52, 279-297.

Macdonald, A.J. 1984: Gold mineralization in Ontario, I: the role of banded iron formation; p.412-430 in: Guha, J. and Chown, E.H., eds.: Chibougamau - Stratigraphy and Mineralization, Can. Instit. of Mining, Special Vol. 34.

Mackasey, W.O. 1969a: Leduc Township, District of Thunder Bay. Ontario Dept. Mines, Prelim. Map P 540, scale 1:15 840.

Mackasey, W.O. 1969b: Walters Township, District of Thunder Bay. Ontario Dept. Mines, Prelim. Map P 539, scale 1:15 840.

Mackasey, W.O. 1970a: The Beardmore-Geraldton Belt; p.69-84 in: Proceedings of the 16th Annual Mtg., Institute on Lake Superior Geology.

Mackasey, W.O. 1970b: Eva Township, District of Thunder Bay. Ontario Dept. Mines, Prelim. Map P 601, scale 1:15 840.

Mackasey, W.O. 1970c: Summers Township, District of Thunder Bay. Ontario Dept. Mines, Prelim. Map P 602, scale 1:15 840.

Mackasey, W.O. 1972: Sturgeon River metavolcanic-metasedimentary formations in the Beardmore-Geraldton area; p.46-50 in: Franklin, J.M., ed.: The Precambrian rocks of the Atikokan-Thunder Bay-Marathon area. 24th Int. Geol. Congress, Field Excursion C34 Guidebook.

Mackasey, W.O. 1975: Geology of Dorothea, Sandra and Irwin Townships, District of Thunder Bay. Ontario Div. Mines, Geol. Report 122, 83 p. Acc. by Map 2294, scale 1:31 680.

Mackasey, W.O. 1976: Geology of Walters and Leduc Townships, District of Thunder Bay. Ontario Div. Mines, Geol. Report 149, 58 p. Acc. by Map 2356, scale 1:31 680.

Mackasey, W.O., Blackburn, C.E. and Trowell, N.F. 1974: A regional approach to the Wabigoon-Quetico Belts and its bearing on exploration in northwestern Ontario. Ontario Div. Mines, Misc. Paper 58, 30 p.

Mackasey, W.O., Edwards, G.R. and Cape, D.F. 1976a: Legault Township, District of Thunder Bay. Ontario Div. Mines, Prelim. Map P 1191, Geol. Series, scale 1:15 840.

Mackasey, W.O., Edwards, G.R. and Cape, D.F. 1976b: Colter Township, District of Thunder Bay. Ontario Div. Mines, Prelim. Map P 1192, Geol. Series, scale 1:15 840.

Mackasey, W.O. and Wallace, H. 1978: Geology of Elmhirst and Rickaby Townships, District of Thunder Bay. Ontario Geol. Survey, Geoscience Report 168, 101 p. Acc. by Map 2373, scale 1:31 680.

Mason, J.K. and McConnell, C.D. 1983: Gold mineralization in the Beardmore-Geraldton area; p.84-97 in: Colvine, A.D. ed.: The Geology of Gold in Ontario. Ontario Geol. Survey, Misc. Paper 110.

Mason, J., White, G. and McConnell, C. 1985: Field guide to the Beardmore-Geraldton metasedimentary-metavolcanic belt. Ontario Geol. Survey, Open File Report 5538, 73p.

Massari, F. 1984: Resedimented conglomerates of a Miocene fan-delta complex, southern Alps, Italy; p.259-277 in: Koster, E.H. and Steel, R.J., eds.: Sedimentology of Gravels and Conglomerates, Can. Soc. of Petroleum Geologists, Memoir 10.

McCubbin, D.G. 1982: Barrier-island and strand-plain facies; p.247-279 in: Scholle, P.A. and Spearing, D., eds.: Sandstone Depositional Environments, Amer. Assoc. of Petroleum Geologists, Memoir 31.

McDonald, B.C. and Banerjee, I. 1971: Sediments and bedforms on a braided outwash plain. Can. Jour. of Earth Sci., 8, 1282-1301.

McGowen, J.H. and Groat, C.G. 1971: Van Horn Sandstone, west Texas: an alluvial fan model for mineral exploration. Bureau of Economic Geol., Univ. of Texas at Austin, Report of Investigations 72, 57 p.

Miall, A.D. 1977: A review of the braided-river depositional environment. Earth-Sci. Reviews, 13, 1-62.

Miall, A.D. 1978a: Lithofacies types and vertical profile models in braided river deposits: a summary; p.597-604 in: Miall, A.D., ed.: Fluvial Sedimentology, Can. Soc. of Petroleum Geologists, Memoir 5.

Miall, A.D. 1978b: Fluvial sedimentology: an historical review; p.1-47 in: Miall, A.D., ed.: Fluvial Sedimentology, Can. Soc. of Petroleum Geologists, Memoir 5.

Miall, A.D. 1984: Principles of Sedimentary Basin Analysis. Springer-Verlag, 490 p.

Moorhouse, W.W. 1939: Geology of the south Onaman area. Ontario Dept. Mines, v.47, part 8, 1938, 30 p.

Nemec, W. and Steel, R.J. 1984: Alluvial and coastal conglomerates: their significant features and some comments on gravelly mass-flow deposits; p.1-31 in: Koster, E.H. and Steel, R.J., eds.: Sedimentology of Gravels and Conglomerates, Can. Soc. of Petroleum Geologists, Memoir 10.

Nemec, W., Steel, R.J., Porebski, S.J. and Spinnangr, A. 1984: Domba Conglomerate, Devonian, Norway: process and lateral variability in a mass flow-dominated, lacustrine fan-delta; p.295-320 in: Koster, E.H. and Steel, R.J., eds.: Sedimentology of Gravels and Conglomerates, Can. Soc. of Petroleum Geologists, Memoir 10.

Nilsen, T.H. 1980: Modern and ancient submarine fans: discussion of papers by R.G. Walker and W.R. Normark. Amer. Assoc. of Petroleum Geologists Bull., 64, 1094-1101.

Nilsen, T.H. 1982: Alluvial fan deposits; p.49-86 in: Scholle, P.A. and Spearing, D., eds.: Sandstone Depositional Environments, Amer. Assoc. of Petroleum Geologists, Memoir 31.

Normark, W.R. 1978: Fan-valleys, channels and depositional lobes on modern submarine fans: characters for the recognition of sandy turbidite environments. Amer. Assoc. of Petroleum Geologists Bull., 62, 912-931.

Normark, W.R. 1980: Modern and ancient submarine fans: reply. Amer. Assoc. of Petroleum Geologists Bull., 64, 1108-1112.

Ojakangas, R.W. 1985: Review of Archean clastic sedimentation, Canadian Shield: major felsic volcanic contributions to turbidite and alluvial fan-fluvial facies associations; p.23-47 in: Ayres, L.D., Thurston, P.C., Card, K.D. and Weber, W., eds.: Evolution of Archean Supracrustal Sequences, Geol. Assoc. of Canada, Special Paper 28.

Okada, H. 1971: Classification of sandstone: analysis and proposal. Jour. Geol., 79, 509-525.

Parker, J.R. 1980: The structure and environment of deposition of the Finmark metasediments, Thunder Bay, Ontario. Unpubl. B.Sc. thesis, Lakehead Univ., 90 p.

Parks, W.A. 1901: The country east of Nipigon Lake and River. Geol. Survey of Canada, Annual Report, v.14, part A, p.105-109.

Pettijohn, F.J. 1975: Sedimentary Rocks, 3rd edition. Harper & Row, 628 p.

Phillips, R.L. 1984: Depositional features of Late Miocene, marine cross-bedded conglomerates, California; p.345-358 in: Koster, E.H. and Steel, R.J., eds.: Sedimentology of Gravels and Conglomerates, Can. Soc. of Petroleum Geologists, Memoir 10.

Porebski, S.J. 1984: Clast size and bed thickness trends in resedimented conglomerates: example from a Devonian fan-delta succession, southwest Poland; p.399-411 in: Koster, E.H. and Steel, R.J., eds.: Sedimentology of Gravels and Conglomerates, Can. Soc. of Petroleum Geologists, Memoir 10.

Postma, G. 1984: Mass-flow conglomerates in a submarine canyon: Abrioja fan-delta, Pliocene, southeast Spain; p. 237-258 in: Koster, E.H. and Steel, R.J., eds.: Sedimentology of Gravels and Conglomerates, Can. Soc. of Petroleum Geologists, Memoir 10.

Postma, G. and Roep, T.B. 1985: Resedimented conglomerates in the bottomsets of Gilbert-type gravel deltas. Jour. Sedim. Petrol., 55, 874-885.

Prior, D.B., Wiseman, W.J., Jr. and Bryant, W.R. 1981: Submarine chutes on the slopes of fjord deltas. Nature, 290, 326-328.

Pye, E.G. 1968: Geology of Lac des Iles area, District of Thunder Bay. Ontario Dept. Mines, Geol. Report 64, 47 p. Acc. by Maps 2135, 2136, scale 1:31 680.

Reineck, H.-E. and Singh, I.B. 1980: Depositional Sedimentary Environments, Second Edition. Springer-Verlag, 549 p.

Reison, G.E. 1984: Barrier-island and associated strand-plain systems; p.119-140 in: Walker, R.G., ed.: Facies Models, Second Editions, Geoscience Canada, Reprint Series 1.

Ritter, D.F. 1975: Stratigraphic implications of coarse-grained gravel deposited as overbank sediment, southern Illinois. Jour. Geol., 83, 645-650.

Rocheleau, M. and Lajoie, J. 1974: Sedimentary structures in resedimented conglomerate of the Cambrian flysch, L'Islet, Quebec Appalachians. Jour. Sedim. Petrol., 44, 826-836.

Rupke, N.A. 1978: Deep clastic seas; p.372-415 in: Reading, H.G., ed.: Sedimentary Environments and Facies, Elsevier.

Rust, B.R. 1972: Structure and process in a braided river. Sedimentology, 18, 221-245.

Rust, B.R. 1978: Depositional models for braided alluvium; p.605-625 in: Miall, A.D., ed.: Fluvial Sedimentology, Can. Soc. of Petroleum Geologists, Memoir 5.

Rust, B.R. and Koster, E.H. 1984: Coarse alluvial deposits; p.53-69 in: Walker, R.G., ed.: Facies Models, Second Edition, Geoscience Canada, Reprint Series 1.

Schumm, S.A. 1977: The Fluvial System. John Wiley and Sons, 338 p.

Schwerdtner, W.M., Stone, D., Osadetz, K., Morgan, J. and Stott, G.M. 1979: Granitoid complexes and the Archean tectonic record in the southern part of northwestern Ontario. Can. Jour. of Earth Sci., 16, 1965-1977.

Smith, N.D. 1970: The braided stream depositional environment: comparison of the Platte River with some Silurian clastic rocks, north-central Appalachians. Geol. Soc. Amer. Bull., 81, 2293-3014.

Smith, N.D. 1971: Transverse bars and braiding in the Lower Platte River, Nebraska. Geol. Soc. of Amer. Bull., 82, 3407-3419.

Smith, N.D. 1974: Sedimentology and bar formation in the upper Kicking Horse River, a braided outwash stream. Jour. of Geology, 82, 205-224.

Southard, J.B., Smith, N.D. and Kuhnle, R.A. 1984: Chutes and lobes: newly identified elements of braiding in shallow gravelly streams; p.51-59 in: Koster, E.H. and Steel, R.J., eds.: Sedimentology of Gravels and Conglomerates, Can. Soc. of Petroleum Geologists, Memoir 10.

Stott, G.M. 1984a: Lake Nipigon Sheet, Thunder Bay District. Ontario Geol. Survey, Map P 257 (Rev.), Compilation Series - Preliminary Map, scale 1:126 720.

Stott, G.M. 1984b: Geraldton Sheet, Thunder Bay and Cochrane Districts. Ontario Geol. Survey, Map P 241 (Rev.), Compilation Series - Preliminary Map, scale 1:126 720.

Stow, D.A.V. and Bowen, A.J. 1980: A physical model for the transport and sorting of fine-grained sediment by turbidity currents. Sedimentology, 27, 31-46.

Stow, D.A.V. and Shanmugam, G. 1980: Sequence of structures in fine-grained tubidites: comparison of recent deep-sea and ancient flysch sediments. Sedim. Geol., 25, 23-42.

Surlyk, F. 1984: Fan-delta to submarine fan conglomerates of the Volgian-Valanginian Wollaston Forland Group, East Greenland; p.359-382 in: Koster, E.H. and Steel, R.J., eds.: Sedimentology of Gravels and Conglomerates, Can. Soc. of Petroleum Geologists, Memoir 10.

Tanton, T.L. 1919: Canadian Northern railway between Nipigon and Longuelac, northern Ontario. Geol. Survey of Canada, Summary Report, 1917, Part E, 1E-6E.

Tanton, T.L. 1935: Geology of the Sturgeon River area. Can. Mining and Metallurgical Bull., no.280, 341-348.

Teal, P.R. 1979: Stratigraphy, sedimentology, volcanology and development of the Archean Manitou Group, northwestern Ontario, Canada. PhD thesis, McMaster Univ. 291 p.

Teal, P.R. and Walker, R.G. 1977: Stratigraphy and sedimentology of the Archean Manitou Group, northwestern Ontario; p.181-184 in: Current Research, Part A, Geol. Survey of Canada, Paper 77-1A.

Turner, C.C. and Walker, R.G. 1973: Sedimentology, stratigraphy and crustal evolution of the Archean greenstone belt near Sioux Lookout, Ontario. Can. Jour. of Earth Sci., 10, 817-845.

Vos, R.G. 1981a: Sedimentology of an Ordovician fan delta complex, western Libya. Sedim. Geol., 29, 153-170.

Vos, R.G. 1981b: Deltaic sedimentation in the Devonian of western Libya. Sedim. Geol., 29, 67-88.

Walker, R.G. 1975: Generalized facies models for resedimented conglomerates of turbidite association. Geol. Soc. of America Bull., 86, 737-748.

Walker, R.G. 1978a: A critical appraisal of Archean basin-craton complexes. Can. Jour. of Earth Sci., 15, 1213-1218.

Walker, R.G. 1978b: Deep-water sandstone facies and ancient submarine fans: models for exploration for stratigraphic traps. Amer. Assoc. of Petroleum Geologists Bull., 62, 932-966.

Walker, R.G. 1980: Modern and ancient submarine fans: reply. Amer. Assoc. of Petroleum Geologists Bull., 64, 1101-1108.

Walker, R.G. 1984a: Turbidites and associated coarse clastic deposits; p.171-188 in: Walker, R.G., ed.: Facies Models, Second Edition, Geoscience Canada, Reprint Series 1.

Walker, R.G. 1984b: Shelf and shallow marine sands; p.141-170 in: Walker, R.G., ed.: Facies Models, Second Edition, Geoscience Canada, Reprint Series 1.

Walker, R.G. and Cant, D.J. 1984: Sandy fluvial systems; p.71-89 in: Walker, R.G., ed.: Facies Models, Second Edition, Geoscience Canada, Reprint Series 1.

Wescott, W.A. and Ethridge, F.G. 1980: Fan-delta sedimentology and tectonic setting - Yallahs fan delta, southeast Jamaica. Amer. Assoc. of Petroleum Geologists Bull., 64, 374-399.

Wescott, W.A. and Ethridge, F.G. 1982: Bathymetry and sediment dispersal dynamics along the Yallahs fan-delta front, Jamaica. Marine Geology, 46, 245-260.

Williams, H.R. 1985: An Archean forearc accretionary prism in the Superior Province of northern Ontario. Unpublished manuscript, 17 p.

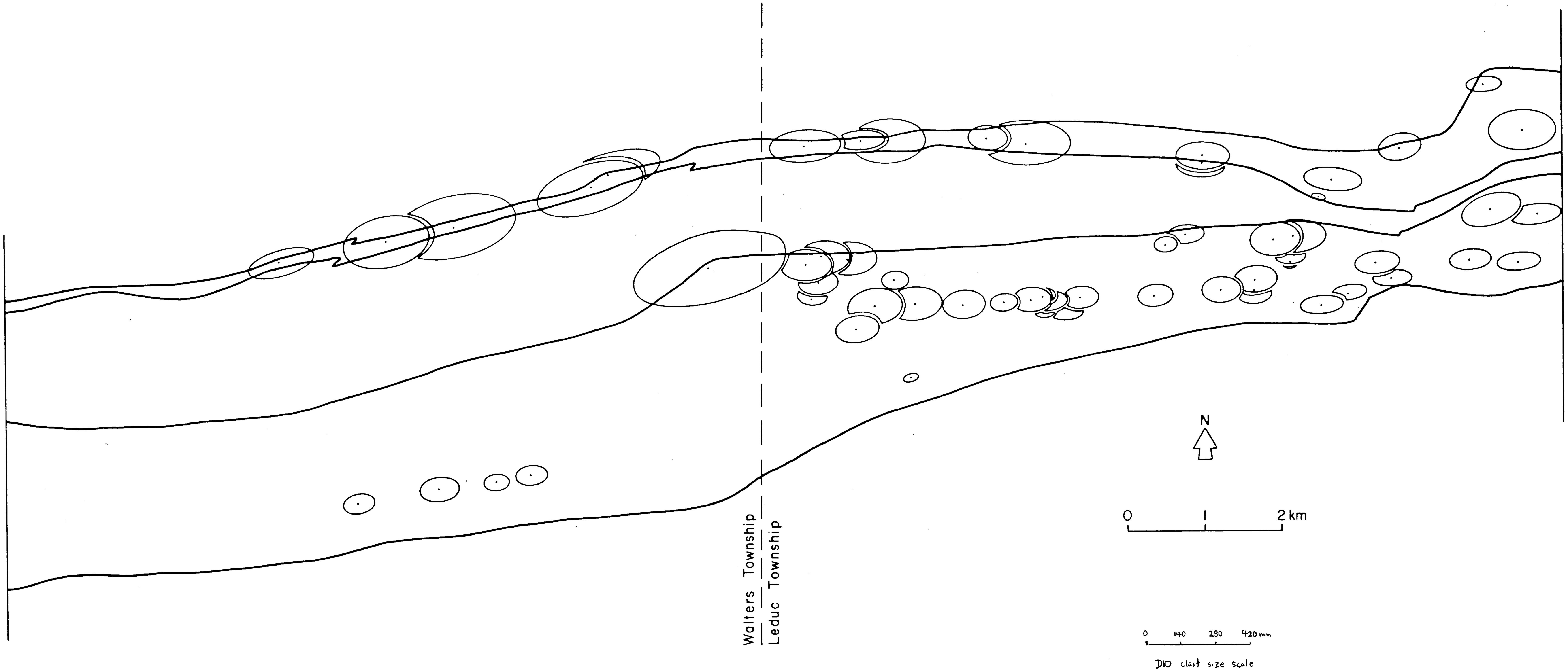
Williams, P.F. and Rust, B.R. 1969: The sedimentology of a braided river. Jour. Sedim. Petrol., 39, 649-679.

Wilson, A.W.G. 1910: Geology of the Nipigon Basin, Ontario. Can. Dept. of Mines, Geol. Survey Branch, Memoir 1, 152 p. Acc. by Map 1090.

Winn, R.D., Jr. and Dott, R.H., Jr. 1977: Large-scale traction-produced structures in deep-water fan-channel conglomerates in southern Chile. Geology, 5, 41-44.

Winn, R.D., Jr. and Dott, R.H., Jr. 1979: Deep-water fan-channel conglomerates of Late Cretaceous age, southern Chile. Sedimentology, 26, 203-228.

Wood, J. 1980: Epiclastic sedimentation and stratigraphy in the North Spirit Lake and Rainy Lake areas: a comparison. Precambrian Research, 12, 227-255.


Walters Township
Leduc Township
N
0
1
2 km
0
140
280
420 mm
D10 clast size scale