# SEDIMENTOLOGY AND LOCAL BASIN ANALYSIS OF THE LOWER CONCEPTION GROUP (HADRYNIAN), AVALON ZONE, NEWFOUNDLAND



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# SEDIMENTOLOGY AND LOCAL BASIN ANALYSIS OF THE LOWER CONCEPTION GROUP (HADRYNIAN), AVALON ZONE, NEWFOUNDLAND

BY

© Scott Gardiner B.Sc. (Hons.)

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Breaching Humpback whale photographed during the caplin run in mid-July, 1983,
at St. Vincent's, St. Mary's Bay, Newfoundland

Abstract

Argillites and medium to very fine grained sandstones in the Mall Bay (approx. 800 m thick) and Drook (approx. 1500 m thick) Formations of the lower Conception Group (Hadrynian) are well exposed along the coast on the east side of St. Mary's Bay, Avalon Peninsula, Newfoundland. These volcaniclastic sediments contain graded beds, convolute and parallel laminae, ripples, and rare floating shale clasts. A few sedimentary slide deposits are also present. Inferred sediment transport mechanisms are turbidity currents, low-density, low velocity lutite flows, cohesive gravity flows, and modified grain flows. Superimposed depositional processes include grain interaction and fluid escape.

Seven facies are recognized based on lithology, grain size, bed thickness, and sedimentary structures. Six of these have been grouped into facies associations I, II, and III, which are interpreted to represent sedimentation on basin plain/fan fringe, lower basin-slope, and lower mid-fan environments respectively, within a turbidite fan depositional system. The intensely faulted, chaotic nature of the seventh facies precluded interpretation.

Field measurements of solemarks (grooves and prodmarks), linguoid and straight-crested ripples, and laboratory measurements of sandstone fabric indicate a complex paleoflow pattern dominated by a

southwest-northeast trend. Paleocurrent directions obtained from sandstone fabric agree with those obtained from solemarks and ripples. Stratigraphic variation of paleoflow directions does occur. During deposition of the Mall Bay Formation, turbidity currents flowed toward the northwest and southwest, while marine currents flowing to the east reworked muds and silts on the adjacent lower basin-slope. During Drook deposition, turbidity current flow was initially toward the southwest, but later switched to the northeast from a new southerly source.

The average sandstone composition of the lower Conception Group is 39.6% plagioclase (albite), 36.6% volcanic rock fragments, 14.2% matrix, 8.2% quartz, 1.4% carbonate, and trace amounts of epidote and accessory heavy minerals (e.g., zircon). Based on electron microprobe data, albitization of feldspar is pervasive.

The dominant source terrane consisted of rhyolitic and basic volcanic rocks. There was also an older sedimentary source. A tectonic setting for these sandstones could not be determined using the QFL and $Q_mFL_t$ diagrams of Dickinson (1983) and coworkers. The sandstone composition is similar to that around island arcs beside both modern and ancient subduction zones, but the paucity of other features indicative of subduction, such as andesitic volcanism, melange, and ophiolites, precludes an unequivocal subduction-zone interpretation.

## ACKNOWLEDGEMENTS

My sincere thanks are extended to Dr. Rick Hiscott, supervisor, coach and editor, who gave generously of his time, near and afar, to give advice, helpful criticism, and to generally keep my ideas in touch with reality.

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# TABLE OF CONTENTS

Page

Title page . . . . . . . . . . . . . . . . .i

Abstract. . . . . . . . . . . . . . . . . . . . . . . . ii

Acknowledgements. . . . . . . . . . . . . . . . . . . . iv

Table of Contents . . . . . . . . . . . . . . . . . . . vi

List of figures . . . . . . . . . . . . . . . . . . . . x

List of tables. . . . . . . . . . . . . . . . . . . . .xii

1. INTRODUCTION . . . . . . . . . . . . . . . . . . . . .1

Unit, area of study, and previous work . . . . . . . .1

Statement of problem and objectives. . . . . . . . . .4

2. GEOLOGICAL SETTING . . . . . . . . . . . . . . . . . .6

General geology of Appalachians . . . . . . . . . . .6
in Newfoundland

Geology and stratigraphy of the Avalon Zone. . . . 10
in Newfoundland

Study area stratigraphy. . . . . . . . . . . . . . . 17

Depositional framework . . . . . . . . . . . . . . . 20

Deformational history and tectonic models. . . . . . 21

3. FACIES, FACIES ASSOCIATIONS AND INTERPRETATION . . . 26
Introduction . . . . . . . . . . . . . . . . . . . . 26
Facies A . . . . . . . . . . . . . . . . . . . . . 29
Description . . . . . . . . . . . . . . . . . . 29
Process interpretation. . . . . . . . . . . . . 51
Facies B . . . . . . . . . . . . . . . . . . . . . 53
Description . . . . . . . . . . . . . . . . . . 53
Process interpretation. . . . . . . . . . . . . 58
Facies C . . . . . . . . . . . . . . . . . . . . . 65
Description . . . . . . . . . . . . . . . . . . 65
Process interpretation. . . . . . . . . . . . . 65
Facies D . . . . . . . . . . . . . . . . . . . . . 68
Description . . . . . . . . . . . . . . . . . . 68
Process interpretation. . . . . . . . . . . . . 68
Facies E . . . . . . . . . . . . . . . . . . . . . 71
Description . . . . . . . . . . . . . . . . . . 71
Process interpretation. . . . . . . . . . . . . 81
Facies F . . . . . . . . . . . . . . . . . . . . . 81
Description . . . . . . . . . . . . . . . . . . 81
Process interpretation. . . . . . . . . . . . . 92
Facies G . . . . . . . . . . . . . . . . . . . . . 93
Description . . . . . . . . . . . . . . . . . . 93
Process interpretation. . . . . . . . . . . . . 93
Facies associations. . . . . . . . . . . . . . . . 93
Depositional environment . . . . . . . . . . . . . 97
Interpretation of facies association I. . . . . 97

Interpretation of facies association II . . . 99
Interpretation of facies association III. . .101
Interpretation of depositional environment. .104
Discussion . . . . . . . . . . . . . . . . . . . . .108

4. PETROGRAPHY. . . . . . . . . . . . . . . . . . . . . .111
Purpose. . . . . . . . . . . . . . . . . . . . . .111
Method . . . . . . . . . . . . . . . . . . . . . .111
Results. . . . . . . . . . . . . . . . . . . . . .112
Volcanic rock fragments . . . . . . . . . . .115
Feldspar. . . . . . . . . . . . . . . . . . . .124
Matrix. . . . . . . . . . . . . . . . . . . . .127
Palagonite. . . . . . . . . . . . . . . . . . .127
Quartz. . . . . . . . . . . . . . . . . . . . .128
Carbonate . . . . . . . . . . . . . . . . . . .131
Epidote . . . . . . . . . . . . . . . . . . . .136
Minor constituents. . . . . . . . . . . . . . .136
Interpretation of petrography. . . . . . . . . . .136
Implications for tectonic setting . . . . . .138

5. PALEOCURRENT ANALYSIS. . . . . . . . . . . . . . . . .149
Introduction . . . . . . . . . . . . . . . . . . .149
Method . . . . . . . . . . . . . . . . . . . . . .151
Errors . . . . . . . . . . . . . . . . . . . . . .158
Results. . . . . . . . . . . . . . . . . . . . . .159
Discussion and comparison to previous work . . . .168

6. SYNTHESIS AND LOCAL BASIN ANALYSIS . . . . . . . . . . .172

Introduction . . . . . . . . . . . . . . . . . . . . .172

Paleogeography and depositional history. . . . . . .172

Comparison to a modern example . . . . . . . . . . .182

Comparison to an ancient example . . . . . . . . . .184

Discussion . . . . . . . . . . . . . . . . . . . . . .186

7. CONCLUSIONS AND RECOMMENDATIONS. . . . . . . . . . . .187

Conclusions. . . . . . . . . . . . . . . . . . . . . .187

Recommendations. . . . . . . . . . . . . . . . . . . .190

REFERENCES. . . . . . . . . . . . . . . . . . . . . . . . .192

APPENDIX I: detailed stratigraphic sections . . . . . .216

APPENDIX II: nomenclature. . . . . . . . . . . . . . . .226

APPENDIX III: feldspar staining techniques. . . . . . . .229

N.B. The headings and subtitles within this thesis appear in text with the following hierarchy: (refer to the headings in chapter 3 as an example)

3. FACIES, FACIES ASSOCIATIONS, AND INTERPRETATION

<u>Introduction</u> (1st order heading-centered)

<u>Facies Association I</u> (1st order heading-centered)

<u>Facies A</u> (2nd order heading-the following sentence starts on the <u>next</u> line)

<u>Description</u> (3rd order heading-the following sentence starts on the <u>same</u> line)

# LIST OF FIGURES


| Figure | | Page |
|---|---|---|
| 1.1 | Study area, outcrop locations, general geology | 2 |
| 2.1 | Appalachian tectonostratigraphic zones | 7 |
| 2.2 | Late Proterozoic successions in North Atlantic | 9 |
| 2.3 | Geology of the Newfoundland Avalon Zone | 11 |
| 2.4 | Avalon Zone stratigraphy | 12 |
| 3.1 | Organizational diagram | 28 |
| 3.2 | Facies A: general characteristics | 31 |
| 3.3 | Facies A: general characteristics | 34 |
| 3.4 | Facies A: concretions | 36 |
| 3.5 | Facies A: bed thickness changes | 39 |
| 3.6 | Facies A: channel-like feature | 42 |
| 3.7 | Facies A: margin of the channel-like feature | 44 |
| 3.8 | Facies A: small scale cross lamination | 46 |
| 3.9 | Facies A: coherent sedimentary slide | 48 |
| 3.10 | Facies A: ball and pillow features | 50 |
| 3.11 | Facies B: general characteristics | 55 |
| 3.12 | Facies B: cross laminated siltstones | 59 |
| 3.13 | Facies B: diagenetic features | 62 |
| 3.14 | Facies C: general characteristics | 67 |
| 3.15 | Facies D: general characteristics | 70 |
| 3.16 | Facies E: general characteristics | 73 |
| 3.17 | Facies E: erosive channel features | 75 |
| 3.18 | Facies E: solemarks | 78 |
| 3.19 | Facies E: lateral bed thickness changes | 80 |

3.20 Comparisons between density flow deposits . . . . . 82

3.21 Facies F: bed thickness characteristics . . . . . . 84

3.22 Facies F: general characteristics . . . . . . . . . . 87

3.23 Facies F: convolute laminae . . . . . . . . . . . . . 89

3.24 Facies F: partially coherent sedimentary slide. . . 91

3.25 Paleoflow with respect to facies associations . . . 96

3.26 Schematic section of lower Conception Group . . . 105

3.27 Schematic depositional model. . . . . . . . . . . . 107

4.1 General petrographic characteristics of . . . . . . 114
volcaniclastic sandstones

4.2 Volcanic rock fragment types. . . . . . . . . . . . 120

4.3 Feldspar types. . . . . . . . . . . . . . . . . . . . 126

4.4 Quartz grain types. . . . . . . . . . . . . . . . . 130

4.5 Quartz grains and polycrystalline rock fragments. 133

4.6 Carbonate and accessory minerals. . . . . . . . . . 135

4.7 QFL and $Q_mFL_t$ diagrams for ancient sandstones . . . 143

4.8 QFL and $Q_mFL_t$ diagrams for ancient sandstones . . . 144

4.9 QFL diagram for modern sands. . . . . . . . . . . . 146

5.1 Least projection elongation . . . . . . . . . . . . 155

5.2 Paleoflow directions from field and lab data. . . 163

5.3 Paleoflow directions from field and lab data. . . 164

5.4 Paleoflow directions from field and lab data. . . 165

5.5 Overall paleoflow pattern in study area . . . . . . 166

6.1 Stage 1 of depositional history . . . . . . . . . . 174

6.2 Stage 2 of depositional history . . . . . . . . . . 175

6.3 Stage 3 of depositional history . . . . . . . . . . 176

# LIST OF TABLES

Table — Page

3.1 Summary of facies, facies associations and their interpretation . . . . . 27

4.1 Point count results, per cent minerals . . . . . . . 116

4.2 Types of volcanic rock fragments . . . . . . . . . . 118

4.3 QFL and $Q_mFL_t$ values to be plotted on the diagrams of Dickinson and Suzcek (1979) . . . . . . 142

5.1 Sandstone fabric results . . . . . . . . . . . . . . 160

- If I have seen further, it is by standing upon the shoulders of giants -

Sir Isaac Newton
February 5, 1675

# 1. INTRODUCTION

## Unit, area of study, and previous work

The Conception Group consists of an estimated 3 to 5 km thickness of Precambrian (Hadrynian) marine siliciclastic sediments (Williams and King, 1979) which are exposed on the Avalon Peninsula within the larger tectonostratigraphic Avalon Zone in Newfoundland (Williams, 1979). According to Hutchinson (1953), the type area is located east of Conception Bay. However, many of the best exposed and least deformed sections occur to the south; viz., the eastern side of St. Mary's Bay and eastward to Trepassey Bay (Fig. 1.1).

Five, well exposed, coastal stratigraphic sections in the western part of the Trepassey Map area of the southern Avalon Peninsula (Fig. 1.1) were measured. Despite inaccessibility, which precluded detailed measurement, thickness estimates and interpretation were still attempted for a sixth outcrop at Frapeau Pt. In addition, a small roadside outcrop near St. Stephen's was included in this study although its size hindered production of a stratigraphic section. Therefore, information from a total of seven outcrops was utilized in this thesis.

Previous study of the Conception Group has been limited to regional field mapping (Williams and King, 1979). The thinly bedded sequences exhibit graded bedding and sole marks, indicative of deposition by sediment

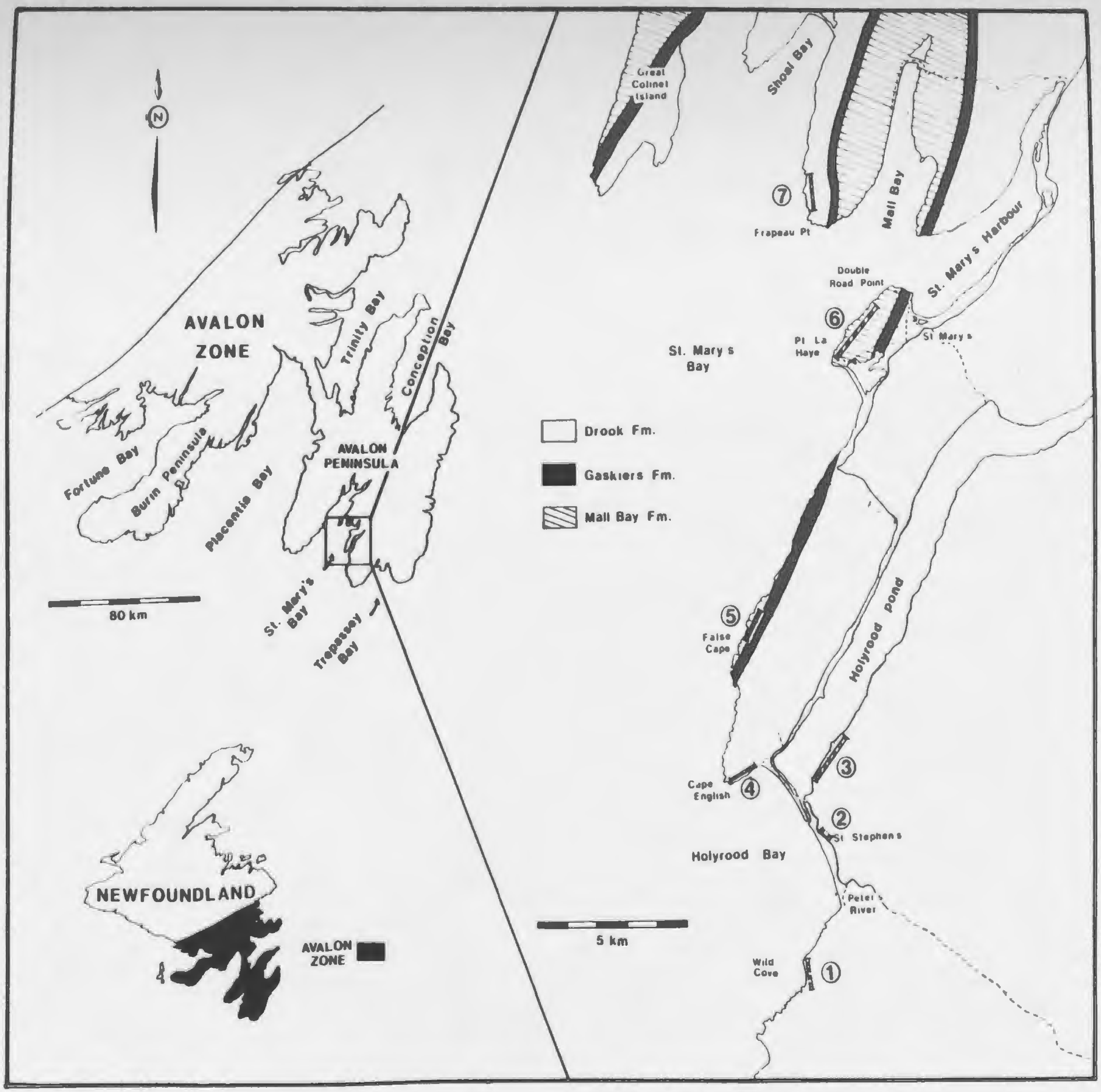



Figure 1.1: Location of study area, seven outcrops studied, and general geology. General geology after Williams and King (1979).

gravity flows, probably turbidity currents (King, 1982, p. 17). Broad generalizations about the style of sedimentation and stratigraphic relationships of the Conception Group with nearby volcanic and plutonic rocks have been made by Hughes (1970,1976), Hughes and Bruckner (1971) and King (1982, p. 19). These writers proposed that the deposition of Conception Group sediments was penecontemporaneous with construction and destruction of volcanic islands with associated plutonic rocks. However, no facies analysis or sedimentological interpretation of specific depositional environments was attempted.

Paleocurrent data for the Conception Group are sparse. Paleoflow directions have been obtained by previous workers from sole marks, cross-bedding and asymmetrical ripples (Williams and King, 1979; Misra, 1969a). However, most of these data were measured from stratigraphically different levels within the group and then combined, thus precluding an assessment of temporal trends. An alternate source of paleocurrent data is sandstone grain fabric (Spotts, 1964; Sestini and Pranzini, 1965; Onions and Middleton, 1968; Colburn, 1968; Hiscott and Middleton, 1980). An initial study by Stewart (1982) determined that the sandstone fabric of Conception Group sediments is a valuable indicator of paleoflow direction. For this reason, and because of the lack of detailed stratigraphic paleocurrent data, grain orientation and imbrication data have been obtained to corroborate paleocurrent measurements from solemarks and ripples in the field.

Two interesting features within the Conception Group have received further attention. Firstly, a till-like (mixtite) unit has been described with respect to structure, stratigraphy and, to a lesser degree, sedimentology (Harland, 1964; McCartney, 1967; Bruckner and Anderson, 1971; Henderson, 1972; and Gravenor, 1980). Inferred dropstones, chattermarked garnets, and striated and faceted stones in diamictites indicate a glacial origin for these sediments (Gravenor, 1980; Gravenor et al., 1984). The sedimentation model proposed by Gravenor (ibid.) involves failure of accumulated basal and supraglacial debris on a submarine slope and downslope transport as debris flows. Further study of this unit was not undertaken for this thesis.

Secondly, disc-shaped and frond-like impressions in the Mistaken Point Formation have been interpreted as primitive hydrozoan and metazoan fossils (Anderson and Misra, 1968; Misra, 1969b). The use of the till-like (mixtite) unit and the fossils for regional and intercontinental correlation has been attempted by King (1982, p. 17).

## Statement of problem and objectives

Because detailed sedimentology, including paleocurrent and basin analysis of the marine Conception Group clastics is poorly known, this study will attempt the following:

1) to perform an analysis of facies and facies associations in order to determine the specific environments of deposition;

2) to determine the details of the stratigraphic succession based on field criteria and petrography, thereby supplementing and/or refining present nomenclature;

3) to perform an analysis of paleocurrents, facies distribution and petrography in this area of the basin, leading to a better understanding of the local paleogeography and possible sediment source areas (provenance);

4) to obtain information which will aid in providing a framework for future synthesis of the relationship between Precambrian basin sedimentation and tectonic setting in the Avalon Zone of Newfoundland.

# 2. GEOLOGICAL SETTING

## General geology of the Appalachians in Newfoundland

Insular Newfoundland represents the northern extremity of the Appalachians in North America. When the effects of post-Triassic plate movements are removed, the orogen extends northeastward into the British and Scandinavian Caledonides (Church, 1969; Dewey, 1969; Kennedy et al., 1972; Williams, 1978). The orogen also extends southeastward along the Atlantic seaboard of the U.S.A. to Tennessee and Alabama until buried by Upper Paleozoic rocks and Mesozoic and Tertiary coastal plain sediments.

The Appalachian Orogen in Canada is divided into five tectonostratigraphic zones (Williams, 1979). From west to east, the names of the zones are as follows: Humber, Dunnage, Gander, Avalon and Meguma. With the exception of the Meguma Zone, all zones are present in Newfoundland (Fig. 2.1). Only the Humber and Avalon Zones can be traced along the full length of the orogen in North America (Williams, 1978).

The Newfoundland Avalon Zone is composed of Upper Proterozoic volcanic and sedimentary rocks which are relatively unmetamorphosed and undeformed. They are locally intruded by Precambrian granites. Submarine sampling on the Grand Banks (Lilly, 1965) and the Flemish Cap (Pelletier, 1971) indicates that Precambrian rocks, similar to those on land in the Avalon Zone, extend a

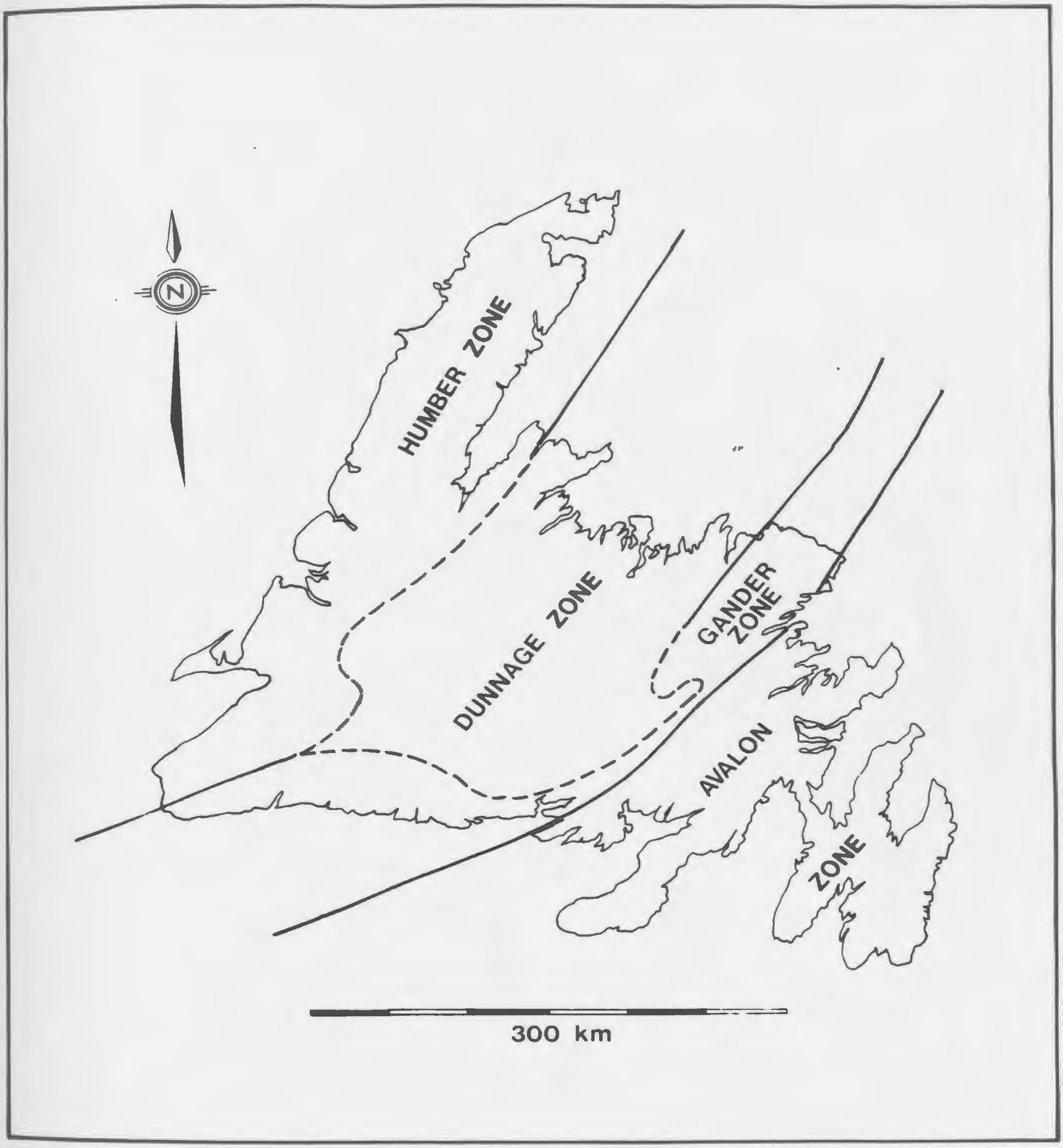



Figure 2.1: The four tectonostratigraphic zones of the Appalachian Orogen in Newfoundland. The Meguma Zone (not shown) is restricted to Nova Scotia (after Williams, 1979).

considerable distance eastward from insular Newfoundland. Therefore, the Avalon Zone may be far wider than the combined width of all other zones in the Appalachian Orogen (Williams, 1979). Haworth et al. (1978) used refraction seismic data to recognize a regionally continuous east-dipping layer of ultramafic rocks beneath the Avalon Zone. They interpreted this as indicative of southeastward subduction of early Paleozoic oceanic crust. The ophiolites of western Newfoundland may be correlative with this ultramafic layer (ibid.). Correlatives of the Avalon Zone occur in Nova Scotia, New Brunswick, Maine, Massachusetts, the Carolinas, Virginia and Georgia (Fig. 2.2). Possible correlatives have also been reported from England, Wales, Brittany, Spain, Portugal and northwest Africa (Fig. 2.2)(Hughes, 1972; Rast et al., 1976; Strong et al., 1978a,b; O'Brien et al., 1983). The Avalon Zone is discussed in greater detail below.

The Avalon and Meguma Zones are examples of suspect terranes (Williams and Hatcher, 1982,1983). They are difficult to explain by a simple Iapetus Ocean model because they were situated on the eastern side of the Iapetus Ocean (i.e., east of the Gander Zone) during Cambro-Ordovician time (Strong and Williams, 1972; Williams and Stevens, 1974; Stukas and Reynolds, 1974). Williams and Hatcher (1982) suggested that these suspect terranes were accreted to eastern North America by oblique


Figure 2.2: Distribution of the late Proterozoic successions around the Precambrian cratons of the North Atlantic region (after O'Brien et al., 1983).

convergence and/or transcurrent movements.

Geology and stratigraphy of the Avalon Zone in Newfoundland

The Upper Proterozoic (Hadrynian) rocks of the central and eastern Avalon Zone (Fig. 2.3) have been divided into three "assemblages" (King et al., 1974; Williams, 1979). These are as follows: (1) a basal assemblage of subaerial and submarine alkalic to tholeiitic volcanic rocks (e.g., Harbour Main Group) with associated volcaniclastics (Nixon and Papezik, 1979); (2) an intermediate assemblage of marine siliceous slates and greywackes (Conception Group, Connecting Point Group), locally interlayered with volcanics and containing a tillite (mixtite) unit in the southeast of the Avalon Zone; (3) an upper assemblage dominated by shallow-marine red sandstones, conglomerates and siltstones (St. John's, Signal Hill, Cabot, Hodgewater and Musgravetown Groups) with a local volcanic unit near the base (Bull Arm Formation).

Due to similar lithologies, geological setting, stratigraphic position and U/Pb age dates, several groups are thought to be equivalent (Fig. 2.4). For example, the Harbour Main Group in the east is considered to be equivalent to the Love Cove Group in the west (King et al., 1974). Due to lithologic similarities, the Connecting Point Group is considered to be equivalent to the Conception Group (ibid.). In addition, the St. John's and

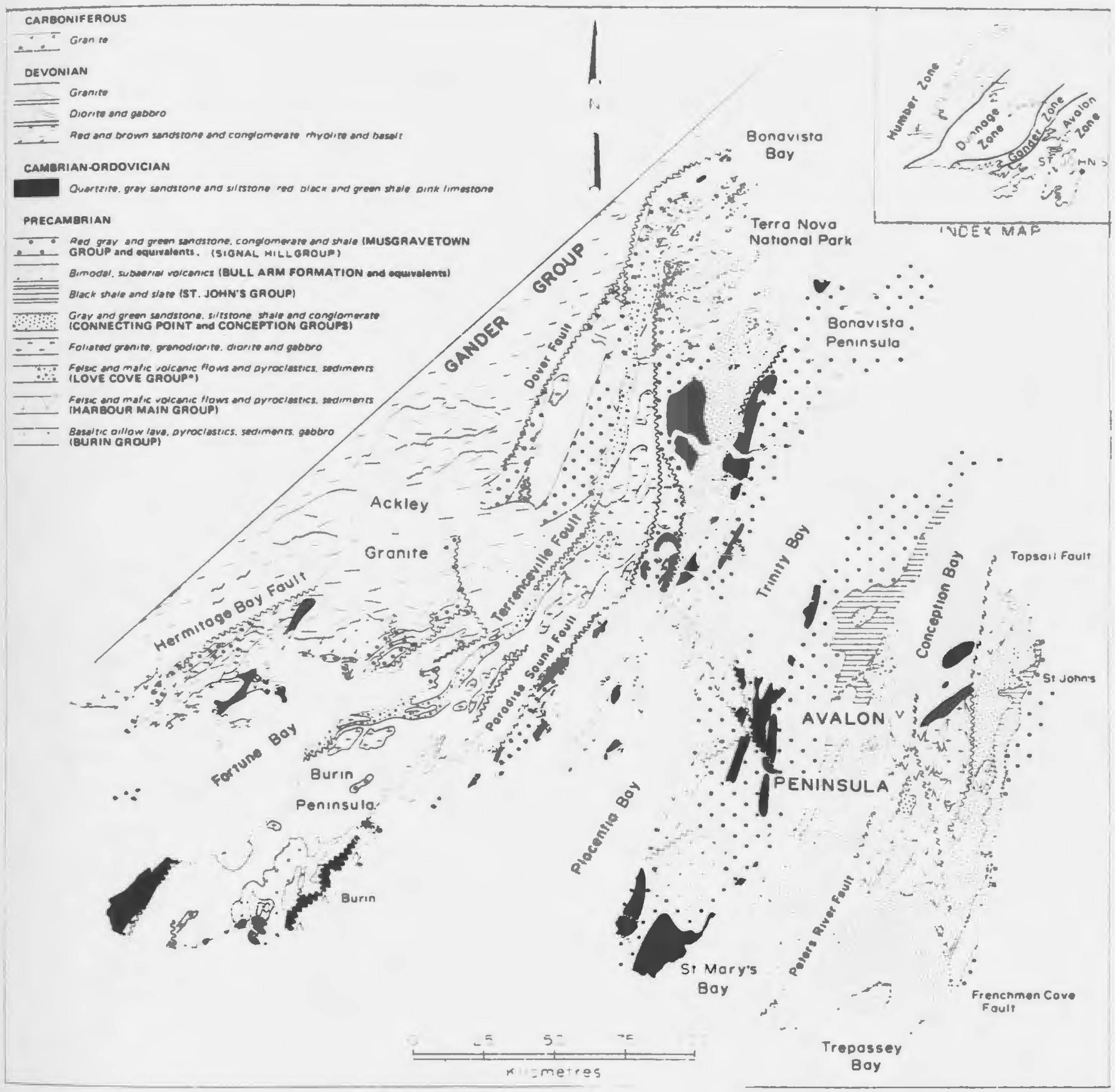



Figure 2.3: Geology of the Avalon Zone in Newfoundland (after King, 1982).

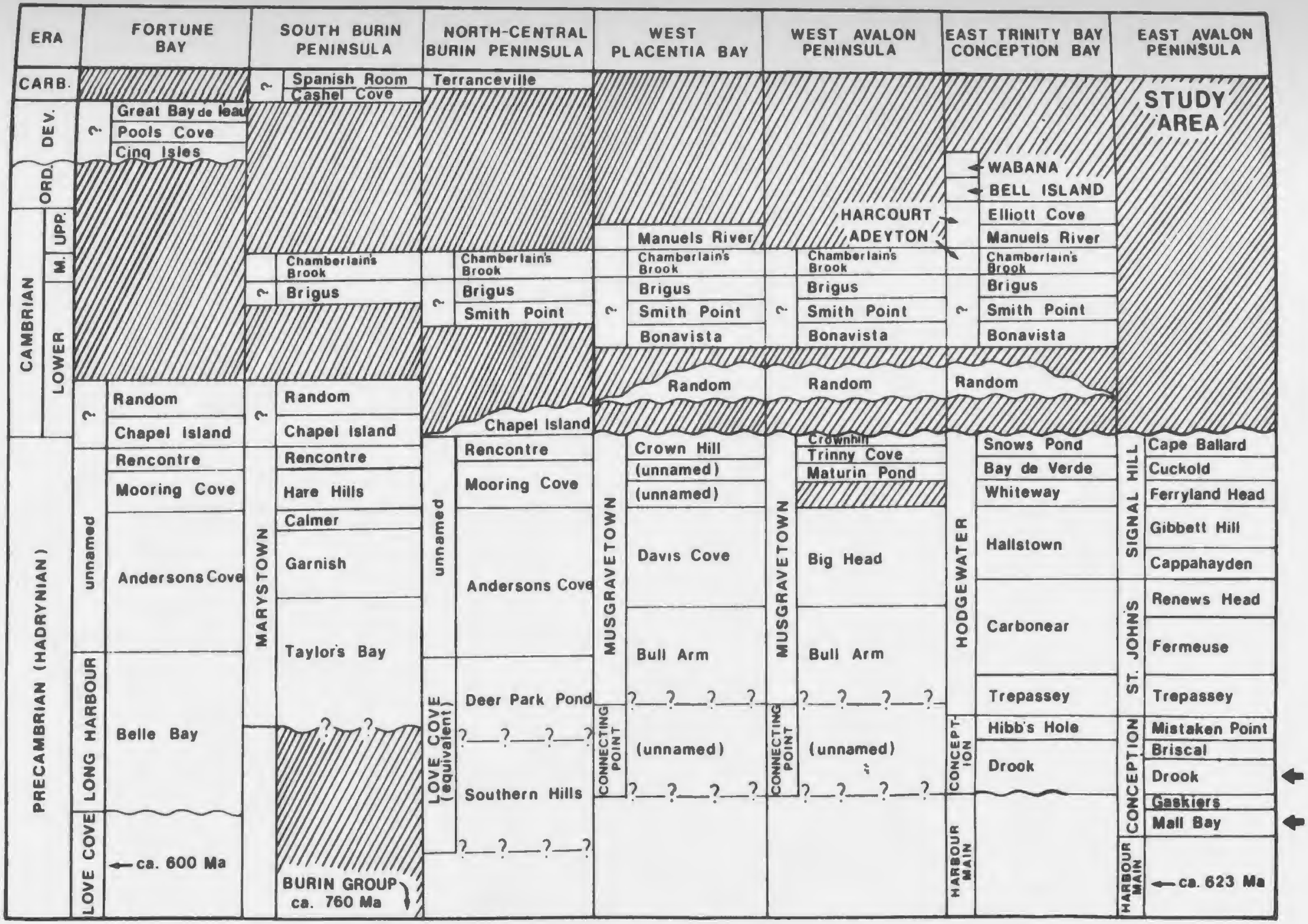



Figure 2.4: Stratigraphy and possible correlations from west to east across the Avalon tectonostratigraphic zone (modified after Strong et al., 1978; Laracy and Hiscott, 1982; Hiscott, 1982; and O'Brien and Taylor, 1983). Burin Peninsula geology after pers. comm. with D.F. Strong, S.J. O'Brien, and R.N. Hiscott). Refer to Fig. 1.1 for locations across the Avalon Zone.

Signal Hill Groups are thought to be equivalent to the Hodgewater and Musgravetown Groups (ibid.).

The Musgravetown Group in the central Avalon Zone contains volcanics of the Bull Arm Formation at its base. These have no recognized equivalent in the Hodgewater, St. John's or Signal Hill Groups to the east (King, 1982, p. 12). The Bull Arm may be relatively younger than other Upper Proterozoic volcanic suites (McCartney, 1967; King et al., 1972) or may be their stratigraphic equivalent (Bruckner, 1977). Strong (1979) states that the mafic and silicic volcanics of the Bull Arm Formation differ from the volcanic Harbour Main and Love Cove Groups in that they are completely subaerial, basalts are tholeiitic to calc-alkaline (Strong, 1977) and they are conformably overlain by the upper assemblage (Musgravetown Group) with no intervening middle assemblage (Conception Group). This strengthens the view that the Bull Arm Formation is significantly younger than the Harbour Main Group.

Submarine mafic volcanics (Burin Group) are present on the Burin Peninsula in the western Avalon Zone. This sequence is 4 to 5 km thick. A major gabbro sill, known as the Wandsworth Gabbro, intrudes the Burin Group. A U-Pb zircon age for the sill of 763 $\pm$ 2 Ma (Krogh et al., 1983) has been proposed as a minimum age for the Burin Group, assuming that the gabbro sill is comagmatic with the volcanics (Taylor, 1976; Strong et al., 1978a, b).

A unit of poorly sorted conglomerates, sandstones and thinly bedded argillites, previously assigned to the Rock Harbour Group by Strong et al., (1978) and O'Brien (1978), outcrops in the western Avalon Zone. O'Brien and Taylor (1983) reassigned these rocks to the Musgravetown Group. These deep-water sediments were originally interpreted to conformably underlie the volcanics of the Burin Group (Strong et al., 1978a; Strong, 1979). Hiscott (1981) suggested the reverse of this relationship; i.e., that the Rock Harbour Group rests unconformably on the Burin Group. Hiscott's conclusion regarding age is substantiated by U-Pb zircon dating of a clast from within the Rock Harbour Group which yielded an age of 623 ± 2 Ma (Krogh et al., 1983) which is younger than the minimum age of the volcanics of the Burin Group. Krogh's age data also support Hiscott's (1981) suggestion that the Rock Harbour Group may be a deeper marine facies of the Gaskiers Formation (mixtite) of the Conception Group, in light of similar clast compositions.

Harbour Main Group volcanic rocks yielded zircons dated at 622.6 ± 2 Ma (Krogh, pers. comm.). Zircons from the Harbour Main Group which cross-cuts a rhyolite plug have also been dated, yielding an age of 631 ± 2 Ma (ibid.). Zircons from the Love Cove Group volcanics, which are thought to be Harbour Main equivalents, have been dated by U-Pb at about 600 Ma (O'Driscoll and Gibbons, 1980;

Dallmeyer et al., 1981). The volcanics of the Harbour Main Group are intruded by the Holyrood Granite, which has a U-Pb age of 620.5 ± 2 Ma (Krogh et al., 1983).

The Conception Group is interpreted to conformably overlie the Harbour Main Group (Williams and King, 1979). McCartney (1967) interpreted this contact as a regional unconformity based largely on the presence of a boulder conglomerate containing clasts of the Harbour Main Group at the base of the Conception Group. This relationship is now viewed as a local unconformity because elsewhere Conception-like sediments are intercalated with Harbour Main volcanics, suggesting penecontemporaneous deposition (Williams and King, 1979).

The Upper Proterozoic volcanic, sedimentary and intrusive rocks of the Avalon Zone are unconformably overlain by shallow marine and terrestrial sediments of Cambro-Ordovician and Carboniferous age (King, 1982, p. 20). The flat-lying Paleozoic sediments are relatively undeformed and are indicative of deposition that took place on a stable platform or shelf (McCartney, 1967; O'Brien et al, 1977; Anderson, 1981; Hiscott, 1982). In stratigraphic order, the Eocambrian and lowest Lower Cambrian rocks are represented by the Rencontre, Chapel Island, and Random Formations (Fig. 2.4). The Rencontre and Chapel Island Formations are only present in the western Avalon Zone. The more widespread Random Formation is unconformably overlain by 2 to 3 km of Lower Cambrian to Lower Ordovician miogeoclinal shales and carbonates (McCartney, 1967; Ranger, 1978). The Lower Cambrian rocks include the Bonavista, Smith Point, Brigus, Chamberlains Brook, Manuels River, and Elliot Cove Formations. The latter five formations contain Acado-Baltic trilobite fauna (Hutchinson, 1962; King _et al._, 1974). The Cambrian rocks are conformably overlain by the Lower Ordovician Bell Island and Wabana Groups (Ranger, 1978). These rocks contain oolitic hematite and chamosite on Bell Island, Conception Bay, and were mined for iron ore for over 70 years ending in 1966.

Rare Devonian and Carboniferous sediments also occur in the Newfoundland Avalon Zone (King, 1982, p. 20). Unfossiliferous Devonian limestones, shales and conglomerates comprise the Cinq Isles Formation north of Fortune Bay (Fig. 2.3)(Calcutt, 1974). These rocks may be, in part, of Silurian age (ibid.). They are overlain disconformably by red conglomerates of the Pools Cove and Great Bay de l'Eau Formations (Williams, 1971; Calcutt, 1974).

Red conglomerate, sandstone, siltstone and caliche horizons comprise the Carboniferous rocks in the Avalon Zone. They outcrop at Spanish Room Point, Burin Peninsula (Strong _et al._, 1978; Laracy and Hiscott, 1982), and at Terrenceville (Fig. 2.3)(Howie and Barss, 1975, p. 40;

O'Brien and Nunn, 1980).

## Study area stratigraphy

The sedimentary origin of the rocks of the Avalon Zone was first documented by Jukes (1842). Rose (1952) defined the Conception Group as the thick sequence of sedimentary rocks overlying the Harbour Main Group and underlying the St. John's Formation of the Cabot Group. Locally, the Conception Group sediments are intercalated with, or lie unconformably on, the Harbour Main volcanics (Williams and King, 1979). Despite the minor intercalated sediments within the Harbour Main volcanics, the Conception Group is considered to be the oldest major sedimentary unit in the Avalon Zone (ibid.). Current estimates of the thickness of the Conception Group range from 3 to 5 km (King, 1982, p. 17). The dominant lithologies are grey to green volcaniclastics and tuffaceous sediments. Grain size ranges from argillite to very fine sandstone to granules. Boulder conglomerates, such as those at Turks Gut near Marysvale on Colliers Bay, are rare. In general, the sediments are relatively fine grained, siliceous, and break with a conchoidal fracture. They have been described informally as exhibiting a "cherty" appearance (Hughes, 1976).

Rose (1952) divided the Conception Group into two units, the "Conception slate" and the "Torbay slate".

Misra (1969a, 1971) proposed a three-fold division of the Conception Group with internal gradational boundaries. In stratigraphic order, from the oldest to youngest, Misra's units were known as the Drook Formation, the Freshwater Point Formation and the Cape Cove Formation. However, this nomenclature was based on a restricted area between Biscay Bay and Cape Race, and did not apply elsewhere in the Conception Group. Subsequent work by Williams and King (1979) modified the upper boundary of the Drook Formation so that it included most of the Freshwater Point Formation. The Drook Formation is the thickest (approx, 1500 m), and most areally extensive formation within the Conception Group. The cherty, siliceous siltstones and argillites of the Drook Formation characterize much of the Conception Group throughout the Avalon Peninsula (Williams and King, 1979, p. 4). In addition, the Cape Cove Formation was discarded and replaced by three new units. In stratigraphic order, from oldest to youngest, these are the Briscal, Mistaken Point and Trepassey Formations. The Trepassey Formation was reassigned to the St. John's Group, which overlies the Conception Group. Williams and King (ibid.) also introduced names for the two lowermost formations of the Conception Group; i.e., the Mall Bay and Gaskiers (mixtite) Formations. The Mall Bay formation name was proposed to designate all strata (approx. 800 m) underlying the Gaskiers Formation mixtites (ibid.). Figure

2.4 summarizes this stratigraphy. It is worth noting that the five formations of the Conception Group have not been recognized in inferred equivalent strata in the northern Avalon Peninsula or in the Connecting Point Group to the west (King, 1980).

The St. John's Group conformably overlies the Conception Group with a gradational boundary (King, 1982, p. 17). It is composed of three formations with an aggregate thickness of about 3 km (ibid.). In stratigraphic order, from oldest to youngest, these are the Trepassey, Fermeuse and Renews Head Formations (Williams and King, 1979). The Fermeuse and Renews Head Formations were previously mapped as the St. John's slates (Jukes, 1842) and the St. John's Formation (Rose, 1952). Williams and King (1979) proposed that the St. John's Formation be elevated to group status.

The base of the Signal Hill Group has an abrupt contact with the underlying St. John's Group due to a local erosional disconformity. The Signal Hill Group was elevated from earlier formation status by Williams and King (1979). It is composed of four formations with an aggregate thickness of approximately 5 km (King, 1982, p. 17; O'Brien and Taylor, 1983, Table 1). In stratigraphic order, from oldest to youngest, these are the Cappahayden, Gibbet Hill, Ferryland Head and Cape Ballard Formations (ibid.). Originally, the St. John's and Signal Hill

Groups were two of three formations within the Cabot Group (Rose, 1952). However, upon their elevation to group status, the name Cabot Group was abandoned.

## Depositional framework

The Conception Group is a flysch-like sequence of volcaniclastic rocks indicating widespread, marine, turbidite and pelagic (hemipelagic?) sedimentation (King, 1982, p. 17). The overlying black shale, slate and silty sandstones of the St. John's Group mark the beginning of a coarsening-upward cycle into the red sandstones and conglomerates of the Signal Hill Group (ibid.). The fine grained rocks of the St. John's Group have been interpreted as prodelta and delta front deposits (Williams and King, 1979; Anderson and King, 1980). Sporadic volcanic activity is indicated by the presence of minor tuff and ash beds (ibid.).

A transition from marine deltaic to subaerial alluvial plain sedimentation is recorded by the Signal Hill Group. These uppermost rocks are cross-bedded, exhibit soft-sediment deformational structures and are interpreted as fluvial channel sands (upper Ferryland Formation) and braided river deposits (Cape Ballard and Blackhead Formations)(Williams and King, 1979). Pebbles in the upper part of the Signal Hill Group are mainly volcanic and decrease in size southward. Paleocurrent studies indicate southward flowing streams (Anderson and King, 1980, p. 12). King (1980, 1982, p. 19) interprets this sequence as the product of upward shoaling in a marine basin (Conception Group) with subsequent progradation of deltaic and alluvial, terrigenous deposits (i.e., St. John's, Signal Hill Groups).

## Deformational history and tectonic models

The dominantly Precambrian rocks of the Avalon Zone are separated from the Gander Zone to the west by the Dover Fault (Blackwood and Kennedy, 1972) in the north and by the Hermitage Bay Fault (Blackwood and O'Driscoll, 1976) in the south. The age, direction, and nature of movement on these faults is unknown. The amphibolite facies metamorphic rocks of the Gander Zone contrast sharply with the greenschist facies rocks of the Avalon Zone. Dallmeyer *et al.* (1981) attributed this contrast to a significant uplift of the Gander Zone relative to the Avalon Zone. Concordant 40Ar/39Ar age spectra between 365 and 395 Ma for mylonites of the Dover Fault Zone suggest that rapid uplift occurred during Acadian (Devonian) orogenesis. The juxtaposition of the Gander Zone against the Avalon Zone was completed prior to the intrusion of the Ackley Batholith, dated by 40Ar/39Ar at 352 $\pm$ 10 Ma (Dallmeyer *et al.*, 1981, 1983). This intrusion straddles the boundary between the two zones (White, 1939). The "Straddling

Granite" was previously thought to also straddle the Hermitage Bay - Dover Fault system, and its Rb/Sr age of 504 ± 12 Ma (Blenkisop et al., 1976) was believed to correspond to the earliest juxtaposition of the Avalon and Gander Zones. However, Elias and Strong (1982) determined that the "Straddling Granite" is divisible into two distinct geochemical suites, the Indian Point and Hardy's Cove Granites. The Hardy's Cove rocks provided the 504 ± 12 Ma age for the "Straddling Granite". The Indian Point Granite intrudes the Gaultois Granite, which has been dated by Rb/Sr at 350 ± 18 Ma (Elias, 1981), indicating that the juxtaposition of the Avalon and Gander Zones may have been as late as the Early Carboniferous. This is in accord with the age determination of Dallmeyer et al. (1981) for the Ackley Granite. Combined, these age dates suggest that the Avalon Zone in Newfoundland was linked to the Gander Zone sometime before the Early Carboniferous. Although there is no field evidence, the two zones may have evolved contiguously, sharing a common basement since the Precambrian (Williams, 1979).

Two periods of tectonic deformation, in the late Precambrian and Devonian, have been identified in the Avalon Zone, although the nature and significance of each is poorly understood. The late Precambrian event is the "Avalonian" Orogeny of Lilly (1966) and Rodgers (1967). In Newfoundland, this orogeny is represented mainly by block faulting, granitoid emplacement and gentle folding. It is not associated with extensive metamorphism and penetrative fabric development (King, 1982, p. 4). The presence, in the late Precambrian, of an unconformity between folded Conception Group sediments overlain by younger sediments of the uppermost Signal Hill Group is additional evidence for Precambrian orogenesis; the tightest folding was apparently restricted to the vicinity of local thrust faults (Poole, 1967; Anderson et al., 1975). An angular unconformity at Bacon Cove between Cambrian strata and the underlying Conception Group, is further evidence for Precambrian deformation (Neale, 1972). Fletcher (1972) also provides evidence for broad, regional folds of Precambrian age in the southwestern Avalon Peninsula.

The most widespread deformation in the Avalon Zone in Newfoundland was produced by the Acadian Orogeny of Devonian age (Dallmeyer et al., 1981). It preceded deposition of Devonian sediments that lie unconformably on Upper Cambrian strata. This event is characterized by folding, faulting, greenschist facies regional metamorphism and granitoid plutonism (King, 1982, p. 6). The Acadian deformation is most intense in the western Avalon Zone where Lower Paleozoic and older rocks are asymmetrically folded and thrust toward the southeast (Strong et al., 1978a). Late Carboniferous (Hercynian) deformation is also well developed in the Maritimes and may occur in the Avalon

Zone of Newfoundland (Rast et al., 1975). Laracy and Hiscott (1982) noted tilting of Carboniferous strata on the Burin Peninsula, and speculated that strike-slip motions may have been responsible.

The Precambrian evolution of the Avalon Zone was genetically unrelated to subsequent Paleozoic evolution of the Appalachian Orogen (Williams and King, 1979; O'Brien et al., 1983).

Numerous tectonic models have been proposed to explain the Precambrian geology of the Avalon Zone along the Appalachian Orogen. These range from subduction models (Hughes and Bruckner, 1971, 1972; Hatcher, 1972; Rast et al., 1976) to Basin and Range type extensional tectonics (Papezik, 1970, 1972; Schenk, 1971; Baker, 1973; Strong et al., 1974; Giles and Ruitenberg, 1977; Nixon and Papezik, 1979) to continental extension and rifting (Rankin, 1975; Strong et al., 1978).

O'Brien et al. (1983) envision a tectonic model involving crustal rifting (circa 800 Ma) with associated localized partial melting and metamorphism. Limited crustal separation (circa 760 Ma) led to the restricted development of oceanic volcanics (Burin Group). Strong (1979) pointed out that these oceanic tholeites may be equivalent to the Proterozoic ophiolites of the margins of the West African Craton at Bou Azzer, Morocco (Leblanc, 1975). Subsequent closure led to the eruption of widespread subaerial volcanic suites, block faulting, granite plutonism, and local sedimentary basin formation (O'Brien *et al.*, 1983). The evidence to date is insufficient to confidently choose between the various tectonic settings proposed above.

# 3. FACIES, FACIES ASSOCIATIONS, AND INTERPRETATION

## Introduction

The purpose of this chapter is threefold. Firstly, each facies will be described. The criteria used to identify each facies are lithology, grain size, bed thickness, and sedimentary structures. Secondly, transport and/or depositional processes necessary to produce the observed facies characteristics are inferred. Finally, in order to facilitate an interpretation of depositional environment, facies which tend to occur together are grouped into facies associations (see Table 3.1 for a summary of facies, facies associations, and their interpretation). For discussion of facies and facies associations, the Mall Bay and Drook Formations are grouped together. Any differences between the two formations, with respect to facies or facies associations, will then be discussed. The synthesis of information from this chapter, from petrography (chapter 4), and from paleocurrent analysis (chapter 5), allows a local basin analysis (chapter 6) to be undertaken (Fig. 3.1).

Appendix I contains the legend used to denote the characteristics of each facies on the drafted stratigraphic sections. Appendix II contains a summary of the descriptive nomenclature used to define terms such as bed, layer, lamination, etc., and bed thickness.

A total of seven outcrops were examined (Fig. 1.1).

Table 3.1: Summary of facies, facies associations, and their interpretations

| FACIES | DESCRIPTION | FACIES PROCESS INTERPRETATION | FACIES ASSOCIATION | ASSOCIATION INTERPRETATION |
|---|---|---|---|---|
| A | -thinly bedded to medium bedded sandstone/argillite<br>-abundant concretions, ball and pillow structures | -turbidity currents<br>-coherent sedimentary slides (Nardin _et al._, 1979) | I | fan fringe/ basin plain |
| B | -parallel laminated (2-10 cm) argillite, some rare ripples<br>-small (1 m) channel-like feature | -low density, low velocity turbidity currents or lutite flows (Piper, 1978) | | |
| C | -wavy bedded siltstone/ argillite<br>-well developed linguoid ripples | -marine currents flowing transverse to the turbidite fan axis (not necessarily contourites) | II | lower basin-slope |
| D | -lenticular bedded siltstone/ argillite<br>-well developed straight-crested ripples | " " | | |
| E | -medium bedded to thickly bedded sandstone<br>-minor channels | -turbidity currents<br>-cohesive flows (Lowe, 1976a, 1982) | III | lobe/ lobe fringe |
| F | -thickly bedded sandstone<br>-abundant convolution | " "<br>-partially coherent sedimentary slides | | |
| A | -see above | | | |
| G | -intensely faulted, chaotic, coarse grained sandstone | -debris flow?? | | |

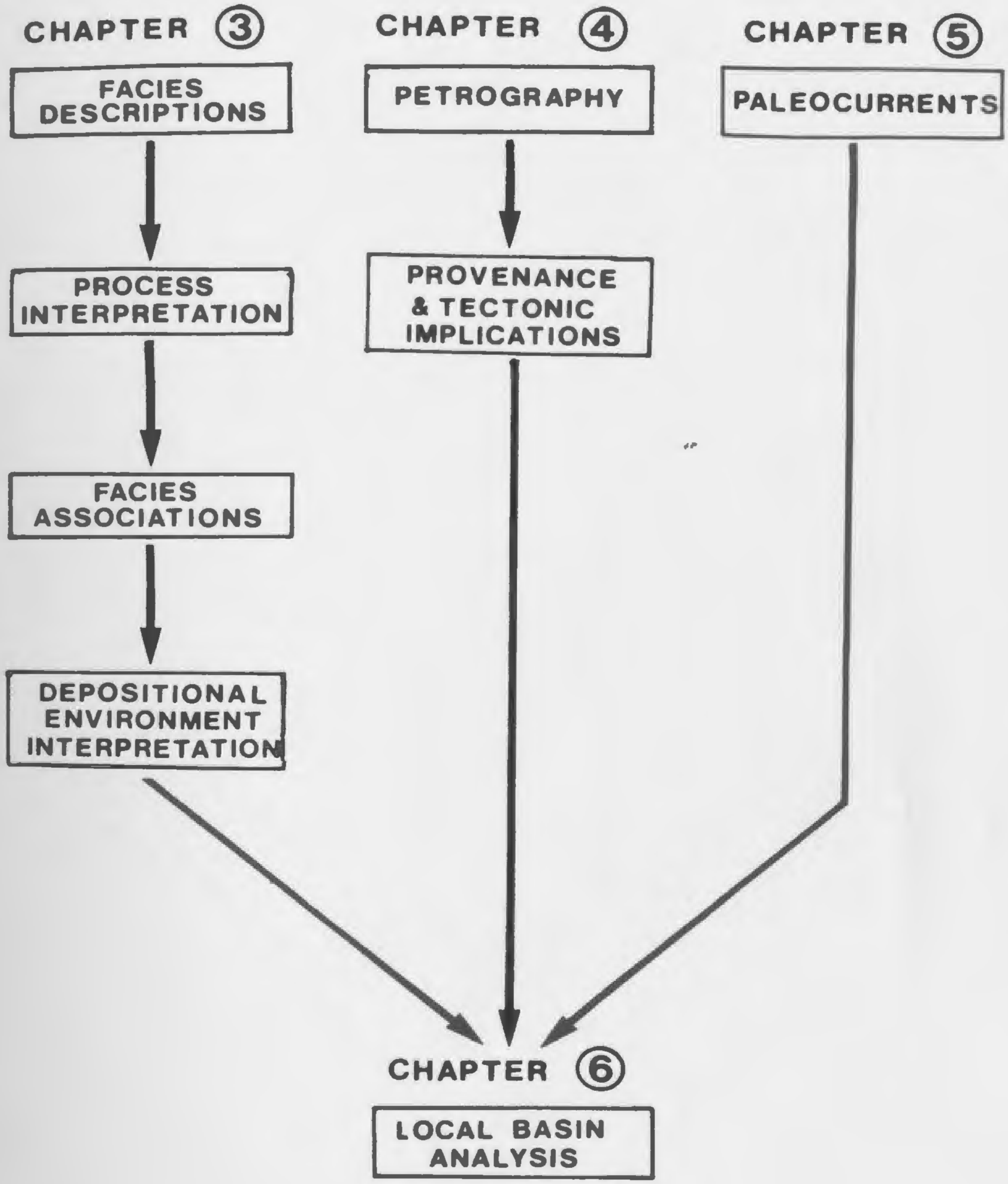



Figure 3.1: Organizational flow chart indicating the contents of chapters 3, 4, and 5. Chapter 6 is a synthesis of the information in previous chapters needed to perform a local basin analysis. The basin analysis discussed in this thesis represents only a small part of the entire Precambrian basin, and therefore is described as "local".

Five stratigraphic sections, with a cumulative thickness of 741 m, were logged in detail and sampled for sandstone fabric and petrographic study (Appendix I). Of the five detailed stratigraphic sections, two were measured in the Mall Bay Formation (Point La Haye, False Cape) and three were measured in the Drook Formation (Wild Cove, Holyrood Pond, and Cape English). The two remaining sections (Frapeau Point, St. Stephens) were examined in less detail because of inaccessibility and insufficient exposure, respectively.

### Facies A: thinly bedded to medium bedded sandstone/argillite

Description:

This facies consists of grey to buff coloured, graded and cross laminated, very fine to medium grained volcanic litharenites and quartzose sandstones interbedded with argillite (Fig. 3.2a). It is best exposed at Frapeau Point (Fig. 3.2b). The Bouma sequence (Bouma, 1962) is used to describe many of the facies characteristics. The sand to shale (argillite) ratio is approximately one, based on field estimates.

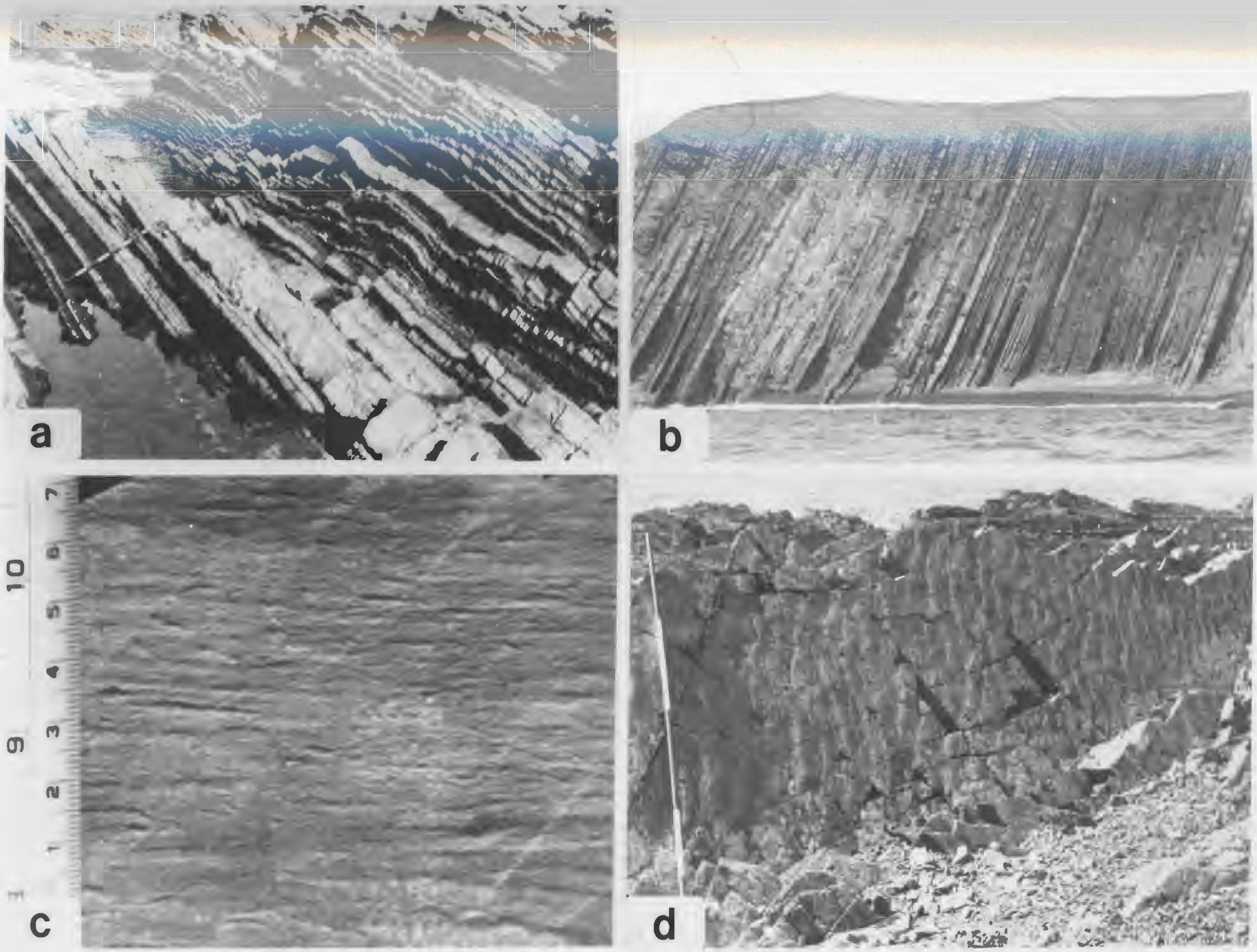
Sandstones are predominantly thinly bedded (3 to 10 cm), display subtle graded bedding (distribution grading) and parallel lamination with abundant solemarks (groove casts, striation casts, and prodmarks)(Fig. 3.2c) and

Figure 3.2: a. Thinly to medium bedded sandstones and argillites of Facies A at Point La Haye. Scale bar divisions equal 10 cm.

b. Facies A. Although inaccessible due to rough seas, this facies is best exposed at Frapeau Point. Cliffs are approximately 100 m high.

c. Solemarks include small groove and striation casts, prod marks, and scratch marks. Scale in centimetres (right) and inches (left).

d. Asymmetrical ripples on upper surface of sandstone bed indicate paleoflow to the left (south) at Point La Haye. Scale bar divisions equal 0.5 m.


a
b
c
d

asymmetrical ripple marks (Fig. 3.2d). Convolute lamination is also present, and is relatively more common than ripples in the Bouma C divisions of beds in this facies. Based on general observation, the size and character of solemarks does not appear to be related to bed thickness. Bouma DE sequences are present in thin beds (3 to 10 cm thick), with BCDE sequences in medium bedded (10 to 30 cm) sandstones (Fig. 3.3a). Beds are laterally continuous over distances up to 0.5 km, with no observed change in sedimentary structures or thickness. Typically, the beds occur in packets up to 3 m thick, which are characterised by 3 to 20 beds of equal thickness, capped by 1 or 2 thicker beds (Fig. 3.3 b,c).

Isolated carbonate concretions and concretion bands with pitted weathering surfaces occur in sand beds (Fig. 3.4 a,b) and are ubiquitous in all facies in the lower Conception Group. A few concretions contain pebble size shale clasts (Fig. 3.4c). The shale pebbles are enveloped by 1 to 2 cm, light coloured, diagenetic alteration rinds (Fig. 3.4d). A polished slab of a concretion reveals a carbonate-filled fenestral fabric arranged in a concentric pattern (Fig. 3.4e).

Lateral thinning of single beds was observed (Fig. 3.5 a,b,c,d). Typically, fine grained sandstone beds thicken from about 5 cm to 15 or 20 cm over a lateral distance of 0.5 to 1.0 m. They are often enveloped by

Figure 3.3: a. Medium bedded sandstones of Facies A at Point La Haye. Scale bar divisions equal 10 cm.

b. Sandstones of Facies A occur in packets up to 3 m thick. Packets are characterized by 3 to 20 beds of about equal thickness capped by 1 or 2 thicker beds. Scale bar divisions equal 10 cm.

c. Sketch of b) above. Location Point La Haye.

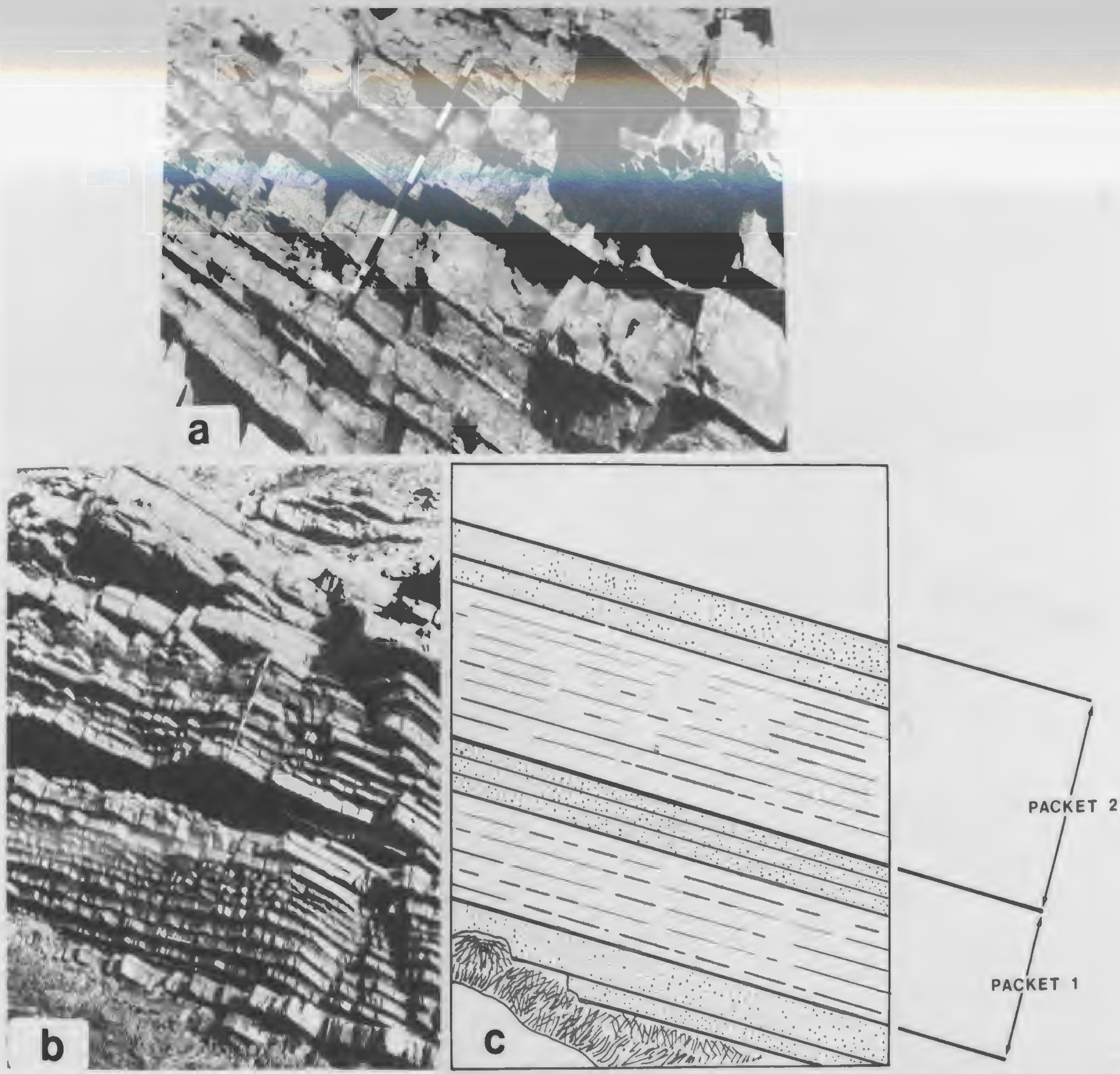

a
b
c
PACKET 2
PACKET 1

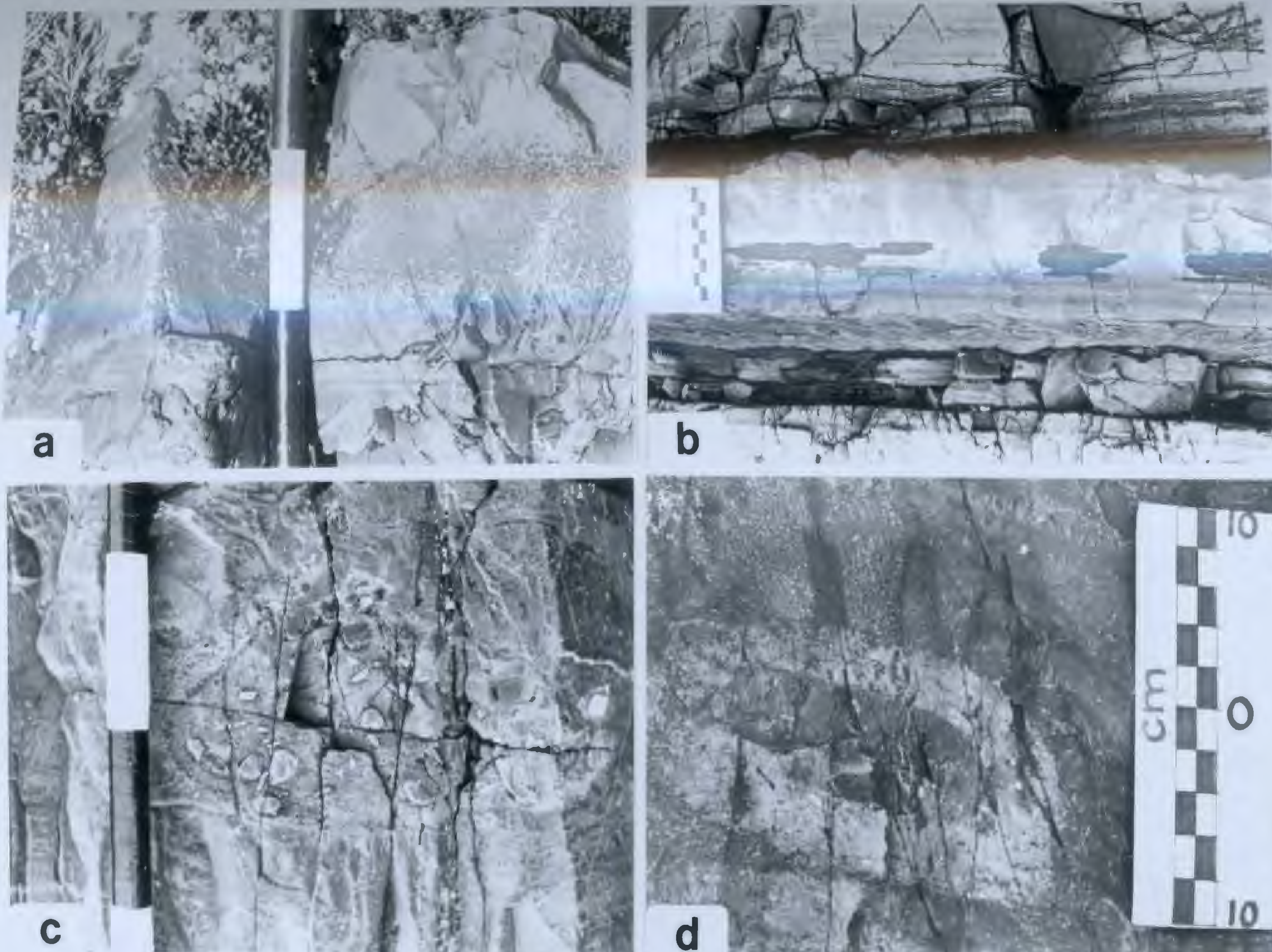
Figure 3.4: a. Pitted weathering surface exhibited by carbonate concretion in fine grained sandstone bed at Point La Haye. Scale bar divisions equal 10 cm.

b. A continuous horizon of carbonate concretions at Point La Haye. Scale divisions equal 2 cm.

c. Pebble size shale clasts within sandy carbonate concretion. Scale bar divisions equal 10 cm.

d. Discoloured "rind" around the perimeter of shale clasts, probably due to diagenetic alteration. Location is Cape English.

e. Carbonate-filled fenestral fabric exhibited by a slabbed portion of a concretion.


a
b
c
d
10
cm
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10

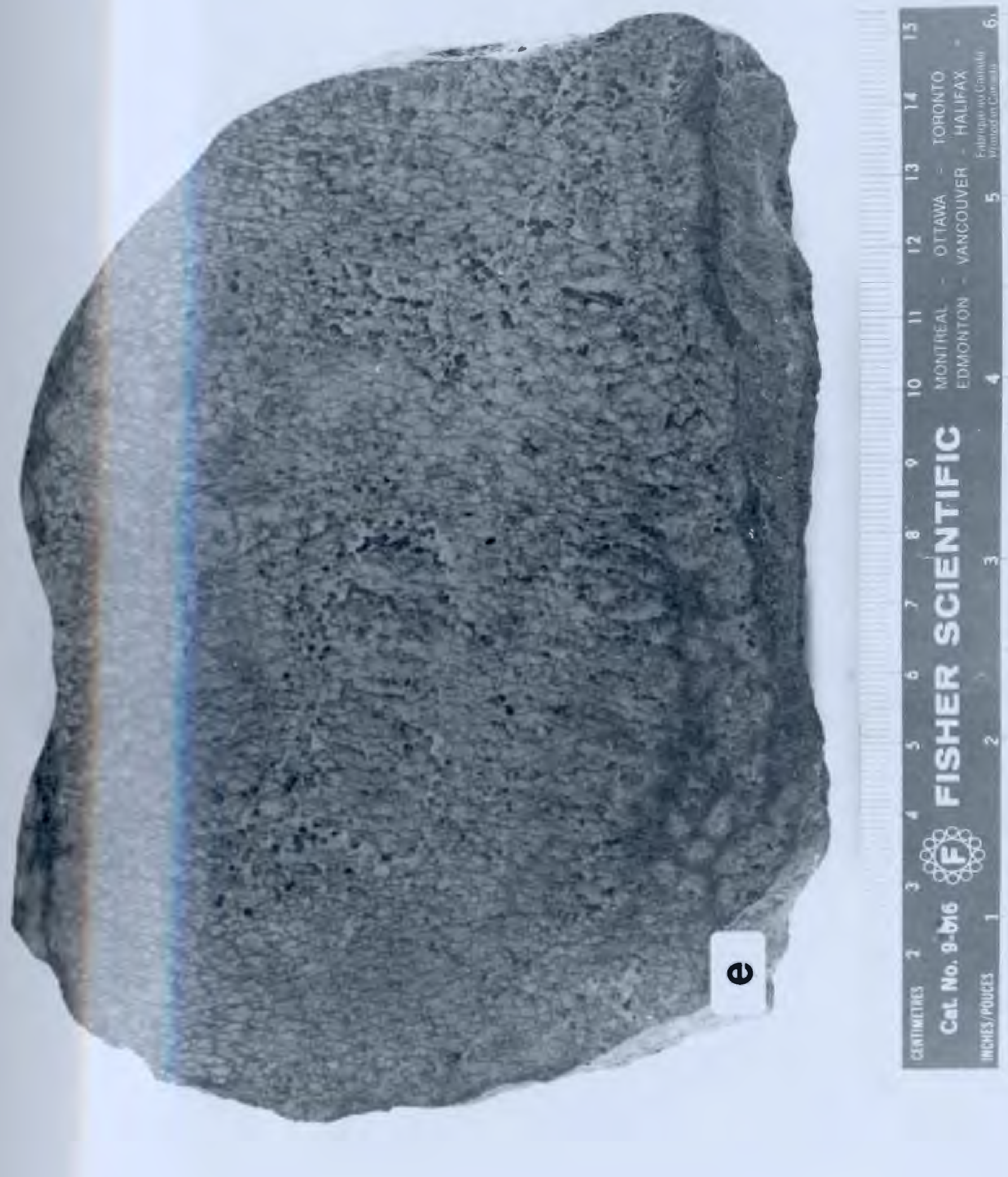

e
CENTIMETRES
FISHER SCIENTIFIC
MONTREAL - OTTAWA - TORONTO
EDMONTON - VANCOUVER - HALIFAX
INCHES/POUCES

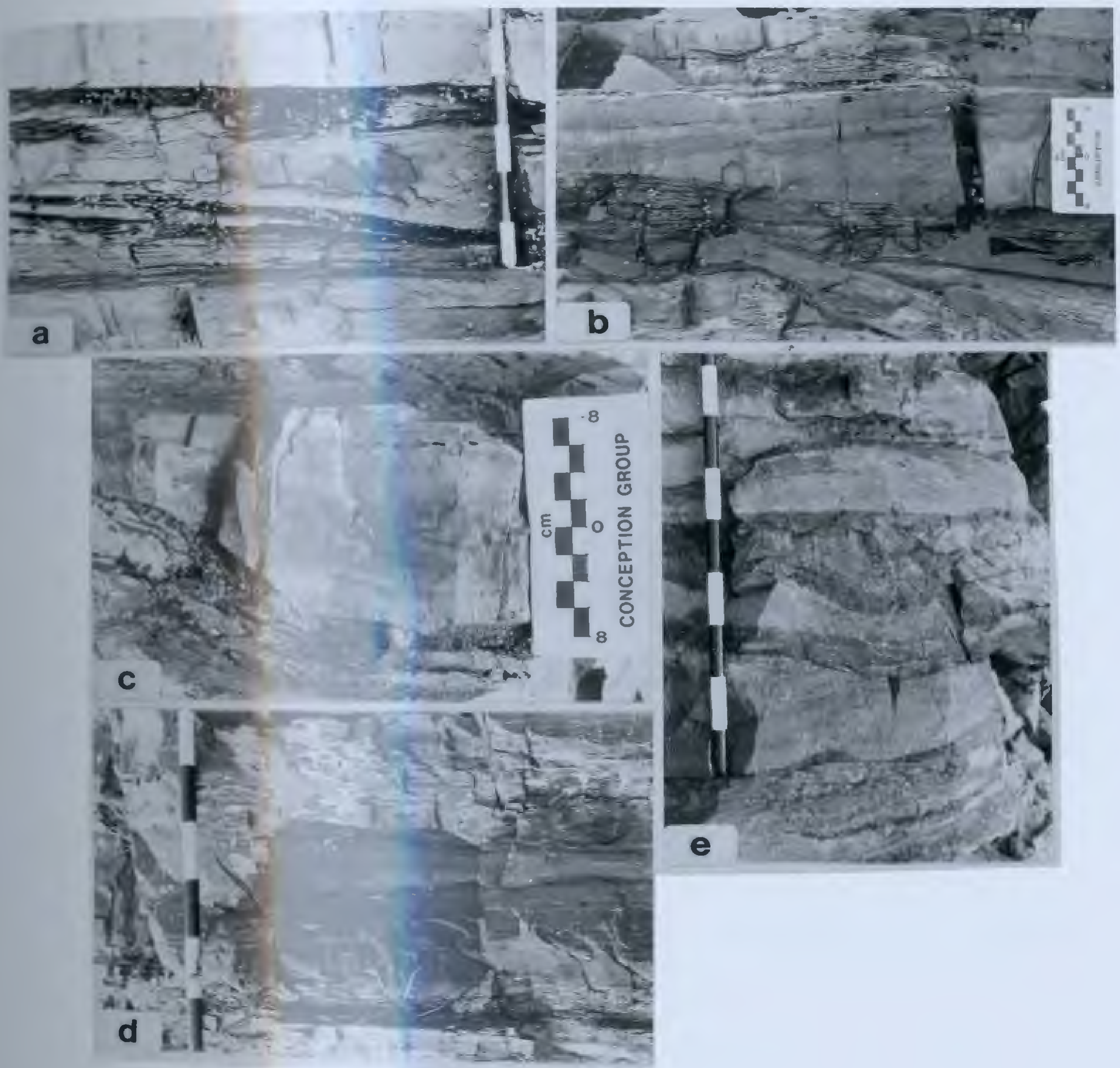
Figure 3.5: a. Lateral bed thickness changes in Facies A sandstones at Cape English. Scale bar divisions equal 10 cm.

b. Same as a) above. Scale divisions equal 2 cm.

c. Lateral thickness changes associated with a concretion (arrow) at Holyrood Pond.

d. Same as a) and b) above. Scale bar divisions equal 10 cm.

e. Lateral bed thickness changes are due to erosion and truncation of internal laminae caused by turbidity currents.


a
b
cm
CONCEPTION GROUP
c
d
e

argillite beds above and below (Fig. 3.5 a,b,d). In Fig. 3.5e, partially convoluted sand beds have been truncated. In one example, at Holyrood Pond, the thinning was related to the presence of a concretion (Fig. 3.5c).

Unique to this facies at Point La Haye are buff coloured, thin (3 to 10 cm), very fine to fine grained sandstones exhibiting lateral thickening, thinning and pinchouts into adjacent cherty argillites (Figs. 3.6, 3.7 a,b,c,d). Small scale cross lamination is present (Fig. 3.8). At the base of the thicker parts of these beds, conglomerate comprised of 1 to 4 cm shale clasts passes upward into millimetre- to centimetre-scale parallel laminated sandstone (Fig. 3.7b). Laterally, at the margin of the lensing bed, laminated argillite clasts are partially incorporated into the sand bed (Fig. 3.7 c,d). Fabric measurements on these lensing sandstones indicate no preferred orientation of grains.

A coherent slide, corresponding to a type A slide of Nardin _et al._ (1979), is present at Cape English (Fig. 3.9 a,b).

Unique to this facies at Cape English are two interesting features. Firstly, ball and pillow structures, as defined by Reineck and Singh (1980, p. 86) and Potter and Pettijohn (1977, p. 201), are well developed (Fig. 3.10a). This ball and pillow horizon is overlain by a pillow-like, discontinuous, sand layer, which is truncated

Figure 3.6: Channel-like, buff coloured, very fine to fine grained sandstones exhibiting lateral thickness changes in Facies A at Point La Haye. Scale bar divisions equal 10 cm.

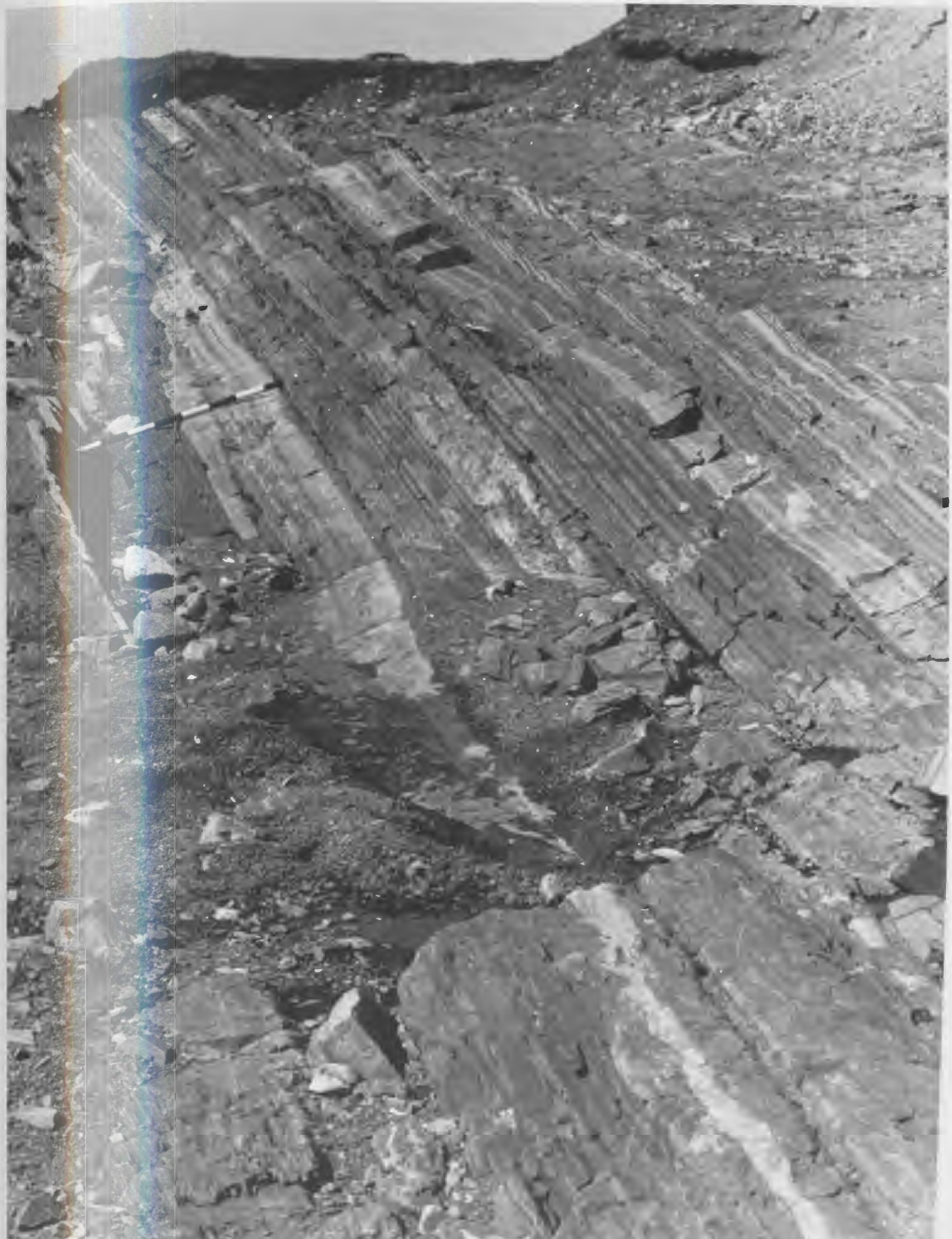


Figure 3.7: a. Parallel laminated argillite partially incorporated into the sandstone bed at the margin of the channel-like feature in Fig. 3.5. Scale bar divisions equal 10 cm.

b. Argillite rip-up clasts fully incorporated into the channel-like sandstone. Parallel laminated sandstone is present in thicker parts of of the bed. Scale bar divisions equal 10 cm.

c. Same as a) above (see arrow). Scale divisions in centimetres (left) and inches (right).

d. Close-up of a) above.

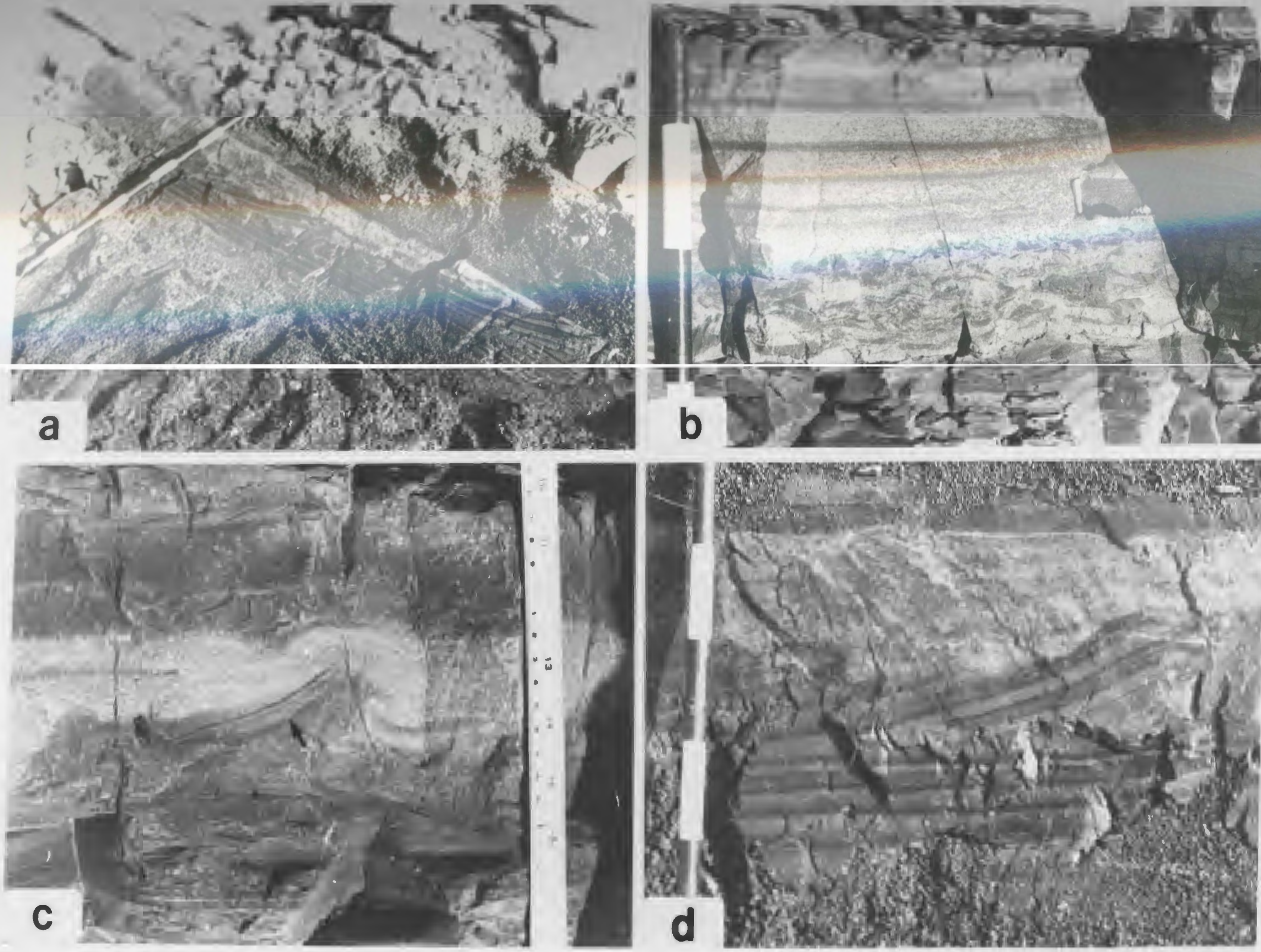

a
b
c
d

Figure 3.8: Small scale cross lamination in fine grained, buff coloured sandstones shown in Fig. 3.6. Scale in centimetres (top) and inches (bottom).


CENTIMETRES
Cat. No. 9-016
INCHES/POUCES
FISHER SCIENTIFIC
MONTREAL
EDMONTON
OTTAWA
VANCOUVER
TORONTO
HALIFAX

Figure 3.9: a. A sedimentary slide exposed in Facies A sandstone at Cape English. This slide is equivalent to a type "A" slide described by Nardin et al. (1979). Scale bar divisions equal 10 cm.

b. Same as a) above.

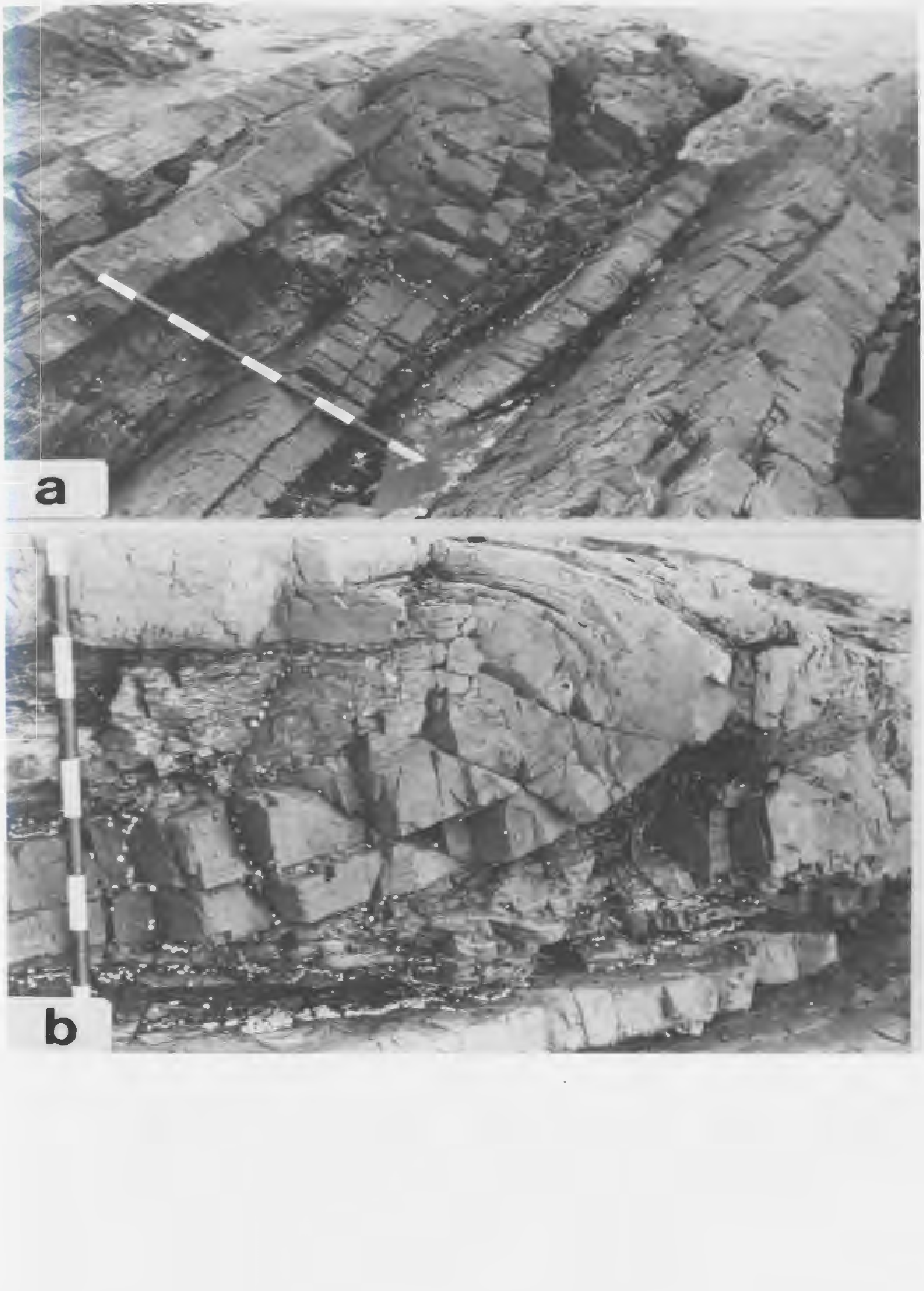
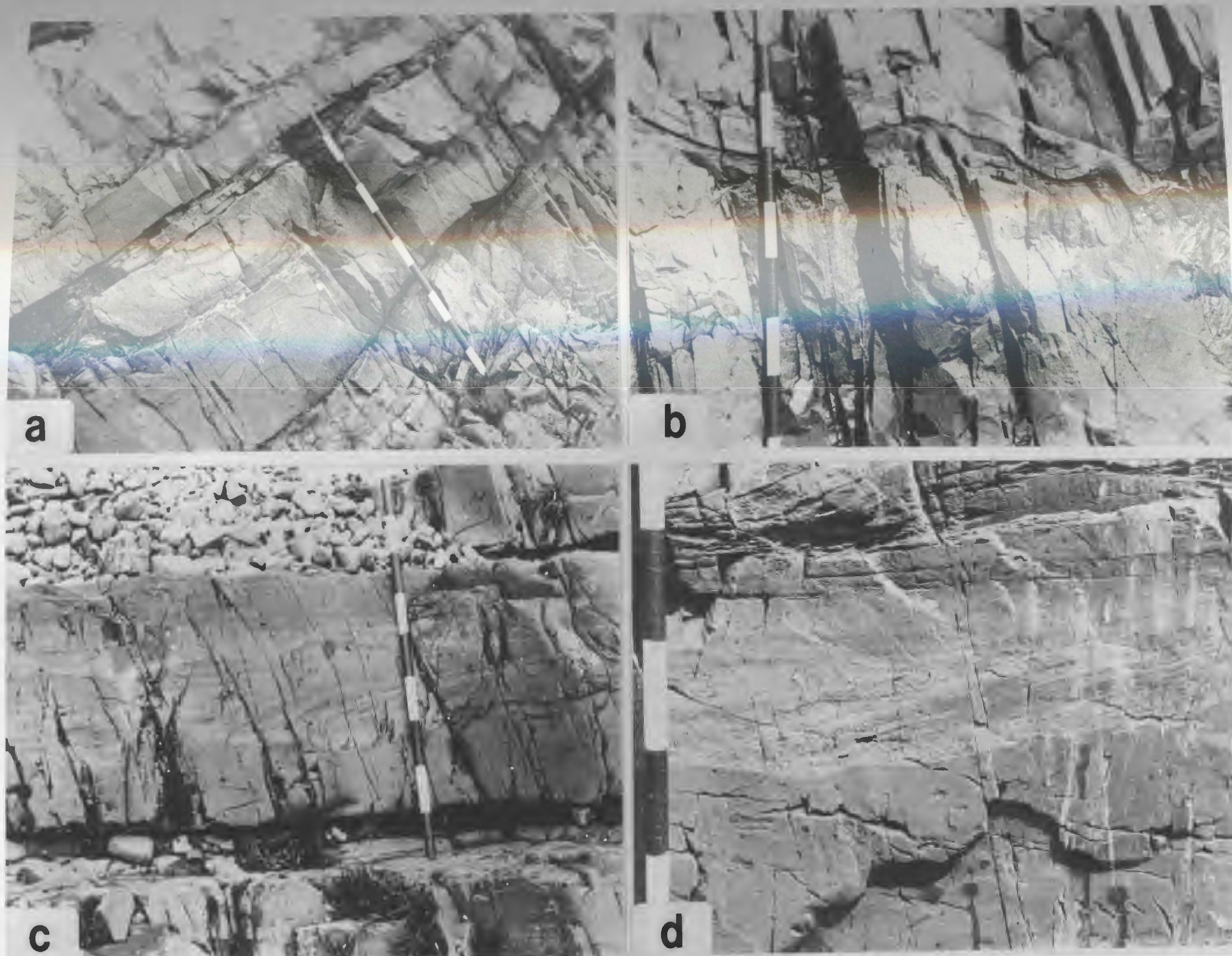

a
b

Figure 3.10: a. Ball and pillow structures in medium bedded sandstone of Facies A at Cape English. Scale bar divisions equal 10 cm.

b. Truncated, discontinuous, pillow-like sandstone layer just above the ball and pillow horizon. Scale bar divisions equal 10 cm.

c. Loading of laminae, producing irregular cross-bed sets. Scale bar divisions equal 10 cm.

d. Parallel cross-bed laminae with an uneven base, perhaps indicative of scour and fill. Scale bar divisions equal 10 cm.


a
b
c
d

by fine grained sandstone (Fig. 3.10b).

Secondly, cross-bed sets, 10-30 cm thick, occur in well sorted, medium grained sandstones (Fig. 3.10 c,d). The apparent paleoflow direction deduced from cross-bed dips is opposite to that of solemarks in adjacent beds (NW vs. SE). No other facies in the Mall Bay or Drook Formations display features of this type.

Process interpretation-Facies A:

Based on the presence of partial Bouma sequences, all beds are the product of deposition from turbidity currents. The lateral thinning of single beds in fig. 3.5 may be due to variable substrate erosion. These sandstone beds represent the filling of small depressions and topographic irregularities. These depressions may have been caused by scour from previous turbidity currents, or by scour beneath the erosive head of the same turbidity current from which the sand was deposited. The truncation of partially convoluted sand beds (Fig. 3.5e) may have been caused by erosion from subsequent turbidity currents resulting in lateral thickness variations. Bed thickness changes associated with concretions may indicate the effects of differential compaction.

Another example of small scale erosive downcutting and filling caused by turbidity currents are the lateral thickening and pinchouts exhibited by the very fine grained sandstones and siltstones in Figs. 3.6 and 3.7. Evidence for this includes partially and wholly incorporated argillite clasts and normal grading. Filling of the small channel does not appear to have been caused by one event. The stratification and alternation of shale and silt laminae throughout suggests that several, intermittent(?) flows were required to fill this eroded depression.

The fold axis of the coherent slide (Fig. 3.9 a,b) is oriented approximately in a southeast-northwest trend. The hooked shape suggests a paleoslope toward the southwest. The usefulness of single fold orientations for determining paleoslope is controversial (Lajoie, 1972), and the southeast sense of paleoslope deduced from this fold is given little weight.

The presence of the 10-30 cm thick cross-bed sets (Fig. 3.10 c,d) is enigmatic. In Fig. 3.10c, to the left of the scale bar, laminae seem to have been loaded, producing irregular, non-parallel laminae. In Fig. 3.10b, the laminae are regular, but the irregular base may indicate scour and subsequent fill. Cross bedded sandstones have been observed in deep-sea clastic rocks, particularly in channel sequences of the inner fan (Hiscott, 1980; Hein and Walker, 1982; Lowe, 1982).The laminae exhibited by this facies are probably not due to bedform migration. Using experimental data for bedform stability (velocity vs. grain size, Middleton and Southard, 1977, p. 7.37), a strong unidirectional current

(60 to 110 cm/s) is necessary to winnow and rework medium sand (0.3 to 0.5 mm) into bedforms of this scale. In modern studies of ocean currents, velocities up to only 40 cm/s have been measured (Reineck and Singh, 1980, p. 489).

The contrary paleoflow direction, the irregular laminae and bases, combined with the lack of modern evidence in ocean basins for the necessary flow velocities, suggests that these features are largely the product of loading of laminated sand (Fig. 3.10c) and/or scour and fill (Fig. 3.10d).

## Facies B: argillite

Description:

Parallel laminated, graded, green, brown, or purple, silty argillites (Fig. 3.11a) and chert, with abundant current ripples and rare, type C climbing ripples of Jopling and Walker (1968), comprise this facies. The upper and lower boundaries with associated facies are sharp. Argillite identical to Facies B occurs interbedded with thinly to medium bedded sandstones in Facies A. The distinction between argillites of Facies B and those within Facies A, is made on the basis of thickness. When argillite occurs in packets one meter thick or greater, it is considered to be Facies B.

Grading and lamination occur in layers with varying thicknesses from 1 to 10 cm (Fig. 3.11 b,c). Individual

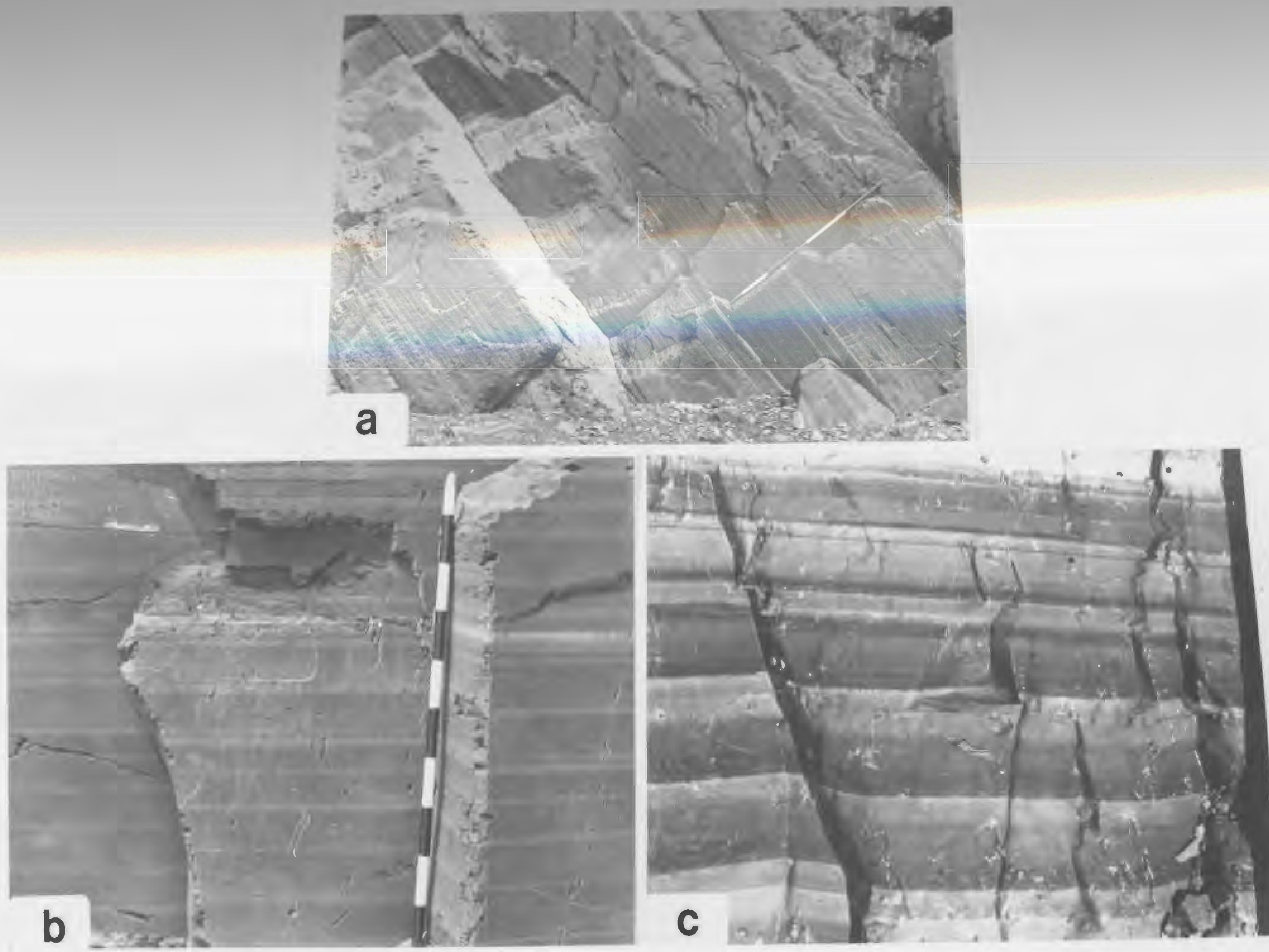
Figure 3.11: a. Parallel laminated, graded, green, brown, or purple silty argillites of Facies B at Wild Cove. Seventy per cent of the 263 m section is composed of this argillite. Scale bar divisions equal 0.5 m.

b. Close-up of laminated argillites shown in a) above.

c. Laminated argillites exhibiting colour change from dark green at the base to lighter green or buff colours at the top of each layer. Location is Point La Haye. Scale divisions in centimetres (left) and inches (right).

d. Close-up of slabbed section of Facies B argillite exhibiting typical Bouma sequence. The coarse grained sediment of the Bouma A division is usually very fine grained sandstone or silt.

e. Sketch of d) above to highlight the features of the Bouma sequence in this fine grained turbidite. Although not exhibited by this sample, the $E_1$ (laminated mud) and $E_2$ (graded mud) divisions described by Piper (1978) are present.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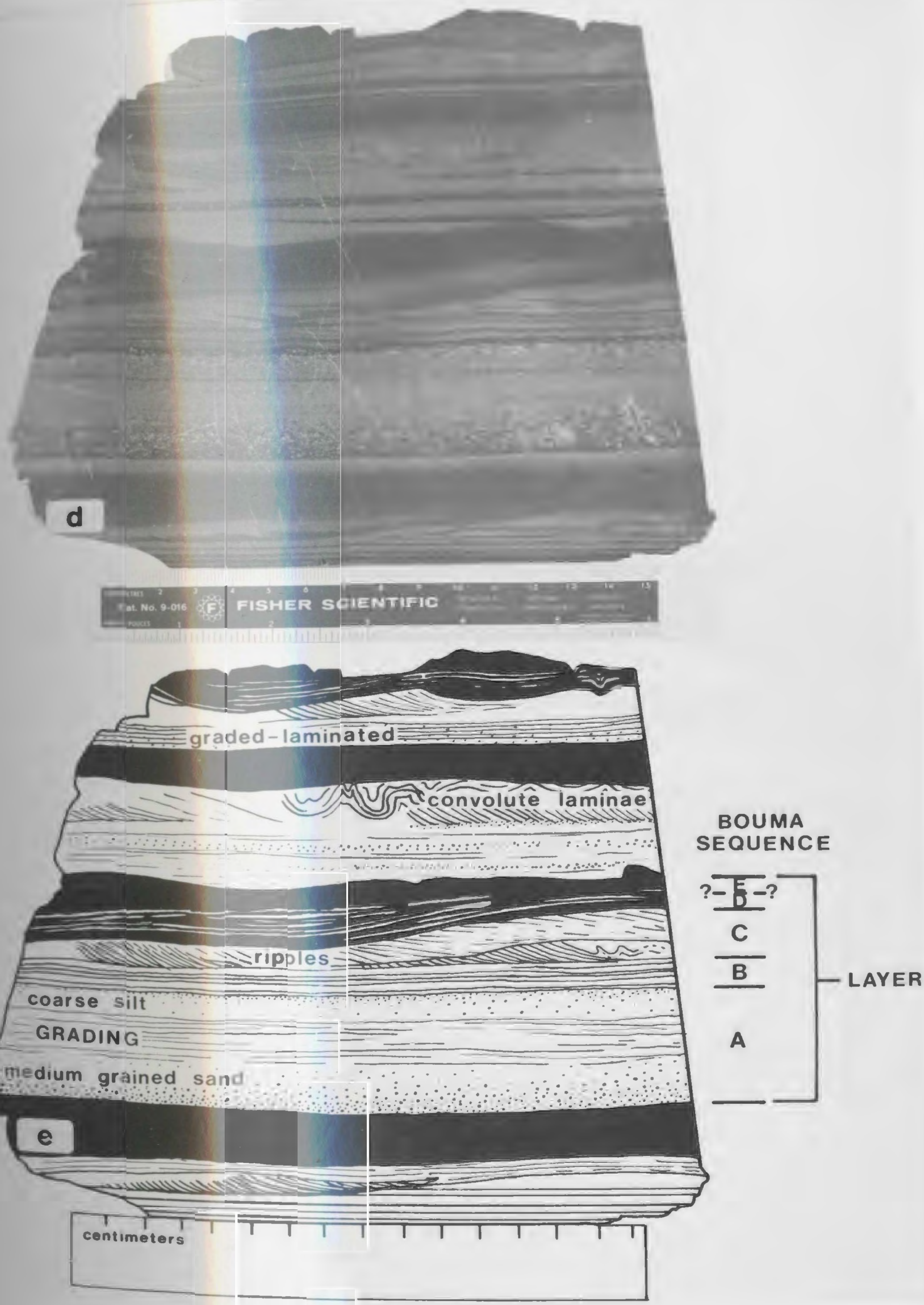
d
Cat. No. 9-016
FISHER SCIENTIFIC
graded-laminated
convolute laminae
ripples
coarse silt
GRADING
medium grained sand
e
centimeters
BOUMA
SEQUENCE
?– E –?
D
C
B
A
LAYER

layers exhibit parallel lamination on a millimetre to centimetre scale, although some are not internally laminated. Complete Bouma sequences are usually not developed. Typically, only Bouma C, D and E divisions exist.

Layers exhibit sharp bases with a very fine sand or silt fraction grading upward into siliceous shale (Fig. 3.11 d,e). The grading and lamination in the Bouma E division corresponds to the E1 (laminated mud) and E2 (graded mud) divisions of Piper (1978). Within these graded layers, colour also varies from dark shades at the base to lighter shades at the top (Fig. 3.11 c) or vice-versa (Fig. 3.11d). This colour variation was thought to be a reflection of decreasing grain size from the base of the layer to the top. However, no correlation between grain size and colour could be made in the field or by thin section study. Some individual silt laminae are laterally continuous for hundreds of meters. However, small scale (0.5 to 2.0 cm) erosional downcutting and truncation of laminae does occur. In some layers which contain a Bouma A division, very small scale (0.5 to 2.0 cm) convolute laminae are present in the upper part, probably equivalent to the Bouma C or D division.

Rare solemarks were observed, although the welded nature of the contact between successive layers, or beds (Appendix II), usually precluded direct observation. Cross lamination is present, but rare (Fig. 3.12 a,b,c,d). Beds exhibiting starved, asymmetrical ripples have the general appearance of lenticular bedding described by Reineck and Singh (1980, p. 114).

Several features are unique to this facies. Firstly, white, circular (1 cm diameter) markings, in discrete bands or horizons parallel to bedding, are present (Fig. 3.13a). There are also thin, white, discontinuous "streaks" parallel to bedding (Fig. 3.13b). Secondly, peculiar, irregular bedding occurs (Fig. 3.13c).In addition, concretions occur in this facies (Fig. 3.13d), but they are less common compared to all other facies.

<u>Process interpretation-Facies B:</u>

Most of the primary features of this facies are characteristic of the turbiditic silts described by Piper (1978) and Stow and Shanmugam (1980). The possibility of the presence of non-turbidite, hemipelagic mud layers (Hesse, 1975; O'Brien <u>et. al.</u>, 1980) cannot be discounted. Criteria based on biogenic structures or micro fauna character are inappropriate for this Precambrian unit. Because grading appears to be continuous from the base to the top of each argillite layer, it seems likely that the presence of non-turbidite, hemipelagic accumulation is volumetrically insignificant.

The paucity of directional sole markings, ripples, erosive bases, and the presence of fine grain sizes,

Figure 3.12: a. Cross laminated siltstone in siliceous shale of Facies B at Wild Cove. Paleoflow is from right to left. Scale divisions in centimetres (right) and inches (left).

b. Same as a) above.

c. Slightly larger scale bedforms in siltstone. Scale bar divisions equal 10 cm.

d. Close-up of c) above. Paleoflow based on cross stratification is from left to right. Scale divisions in centimetres (left) and inches (right).

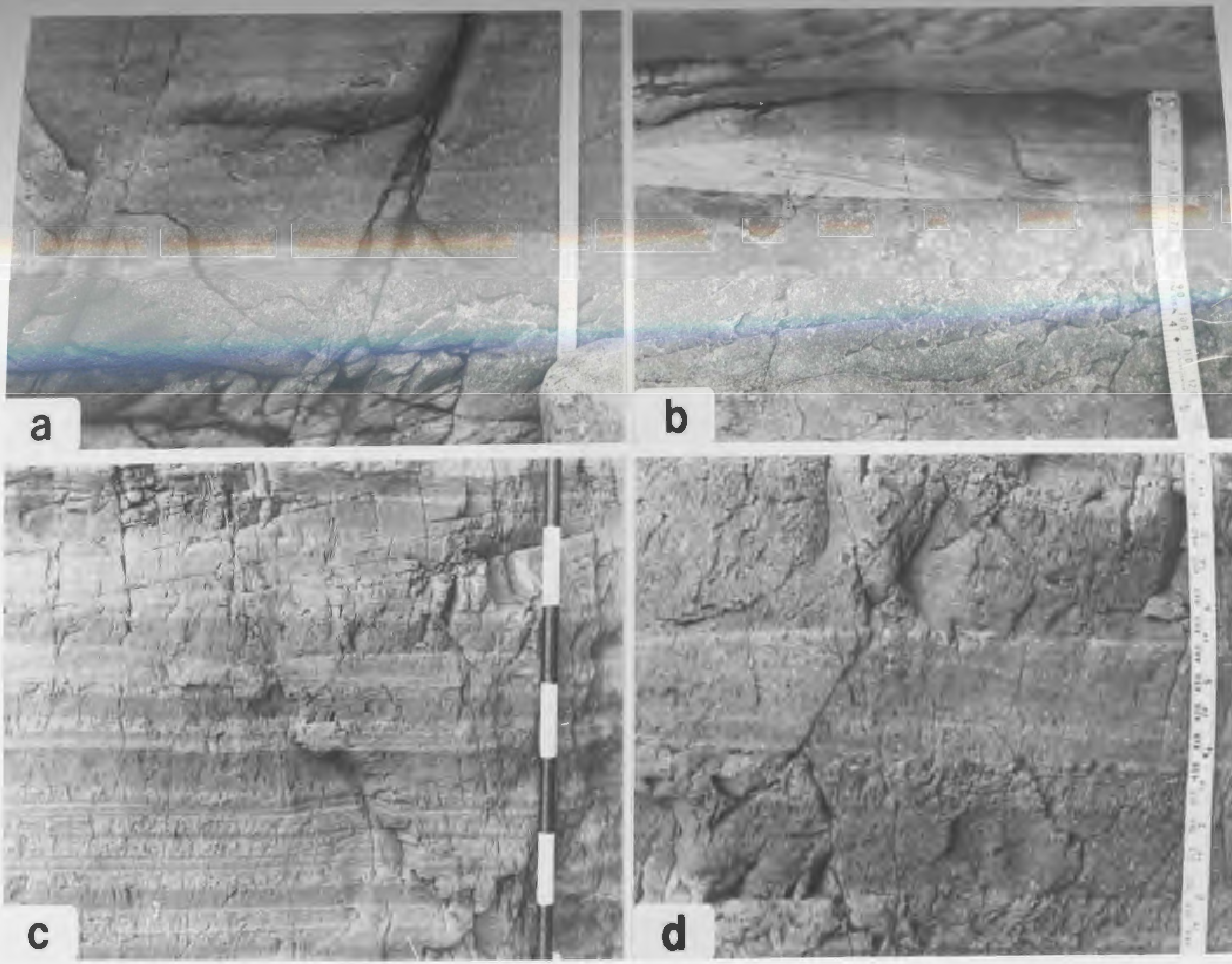
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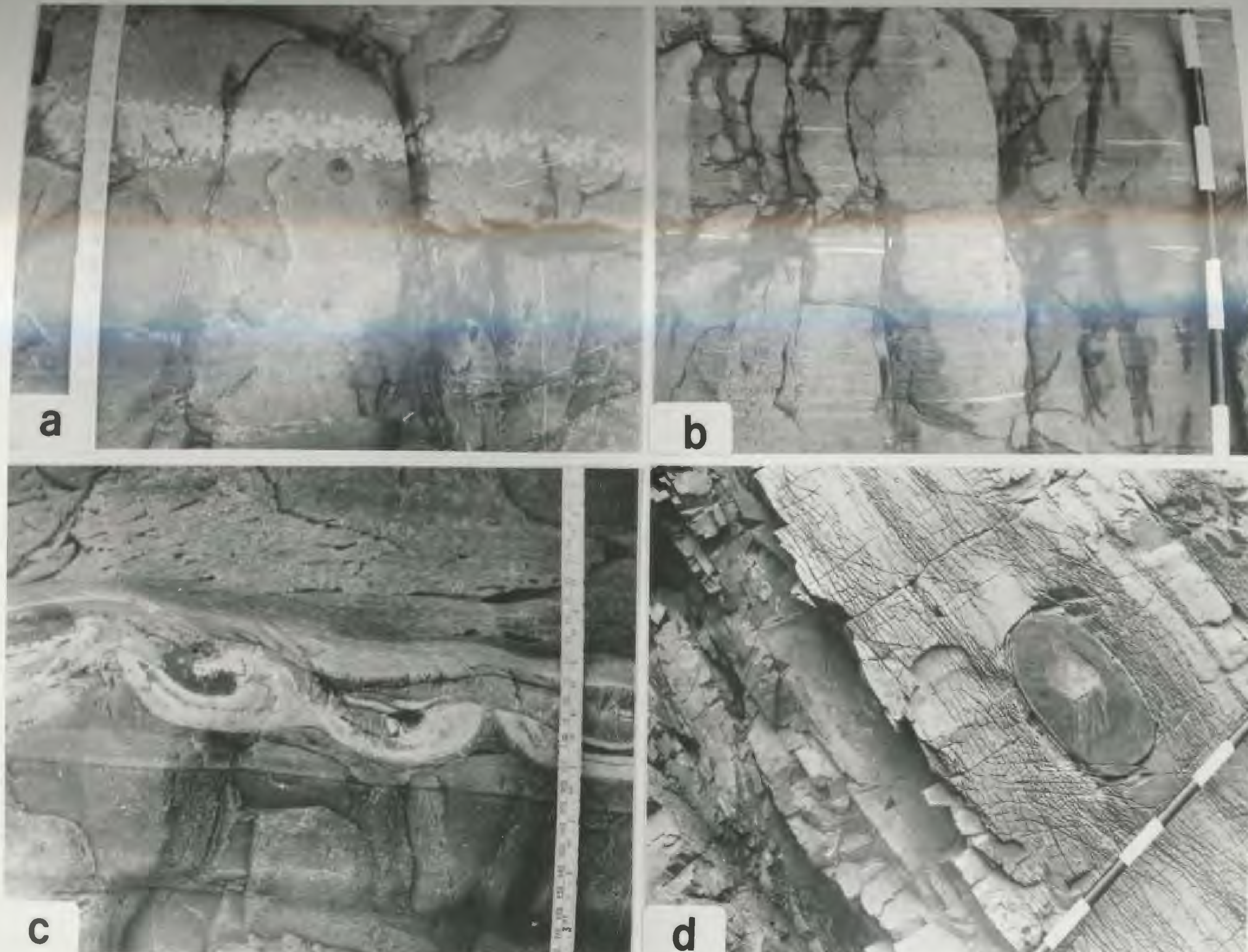
Figure 3.13: a. Discrete horizons of circular markings which occur parallel to bedding are unique to Facies B at Wild Cove. Scale divisions in centimetres (right) and inches (left).

b. Discontinuous white "streaks" occur parallel to bedding. These streaks and the circular markings shown in a) above may be of diagenetic origin. Scale bar divisions equal 10 cm.

c. Irregular bedding in Facies B at Wild Cove. This feature is similar to truncated ball and pillow structure formed by sediment dewatering and/or convolution. Scale in centimetres (left) and inches (right).

d. Large concretion in Facies B at Wild Cove. This concretion is larger than those with the pitted surfaces discussed in fig. 3.4. Scale bar divisions equal 10 cm.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lamination and grading, suggests deposition from low density, low velocity turbidity currents described by Gorsline et al. (1968); Moore (1969); McCave (1972); Shepard et al. (1972); and Stow and Bowen (1980). The origin of the lamination which defines the base of each 1 to 10 cm graded layer in Facies B, can be attributed to dilute, turbid flows, one for each layer. The laminations within each layer are generated by depositional sorting due to increased shear in the bottom boundary layer (Stow and Bowen, 1980). Alternatively, a multiple bursting cycle mechanism (Hesse and Chough, 1980), involving increased duration of stability of the laminar sub-layer flow as turbulence in the main current decayed, could be invoked to explain the millimetre- to centimetre-scale, internal, lamination of Facies B argillite.

Because the thin, white, discontinuous streaks (Fig. 3.13) are not related to grain size or composition, they may be of diagenetic origin. The irregular bedding (Fig. 3.13c) may be the product of sediment dewatering and/or convolution such as ball and pillow structure.

Because of the similarity of Facies B laminites to some modern contourites, some of the lamination may be produced by deposition from a nepheloid layer or bottom (contour) current (Ewing and Thorndike, 1965; Heezen, Hollister, and Ruddiman, 1966; Zimmerman, 1971; Bassin, Harris, and Bouma, 1972; Eittreim and Ewing, 1972; Ewing and Connary, 1972; Hollister and Heezen, 1972; McCave, 1972; Bouma and Hollister, 1973; Betzer, Richardson, and Zimmerman, 1974; Pierce, 1976; Stow and Bowen, 1978). Some deep water laminated muds are demonstrably of this origin (Stow, 1969; Wezel, 1969; Anketell and Lovell, 1976; Stow and Lovell, 1979).

Distinguishing between fine grained turbidites and contourites in the Precambrian Conception Group is difficult because some of the documented criteria of Bouma and Hollister (1973) and Stow and Lovell (1979) are related to organic material, and the remainder are not diagnostic. In the lower Conception Group, this problem is compounded by the difficulty of determining fabric in greenschist facies, cherty, volcanogenic, fine grained sediment. Piper (1978) states that the deposits of contour currents and lutite flows may often resemble mud turbidites, and there is no simple set of criteria to distinguish them. Nevertheless, it seems likely that the graded and laminated argillites of this facies are the products of dilute turbidity currents rather than contourites for the following reasons: a) the presence of climbing ripples as well as starved ripples indicates the presence of a sediment supply, as opposed to intermittent sediment reworking by contour currents; b) paleoflow directions obtained from ripples are identical to those obtained in thicker, sandier, turbidite beds, not transverse as might be anticipated for contourites; c) although beds appear to have a buff coloured, winnowed appearance in the field (Fig. 3.6), petrographic examination suggests no better sorting than in adjacent beds; and d) an improbable bottom current flow would have been required to produce successive graded silt laminae on a millimetre to centimetre scale (Piper, 1978).

Facies C: wavy bedded siltstone/argillite

Description:

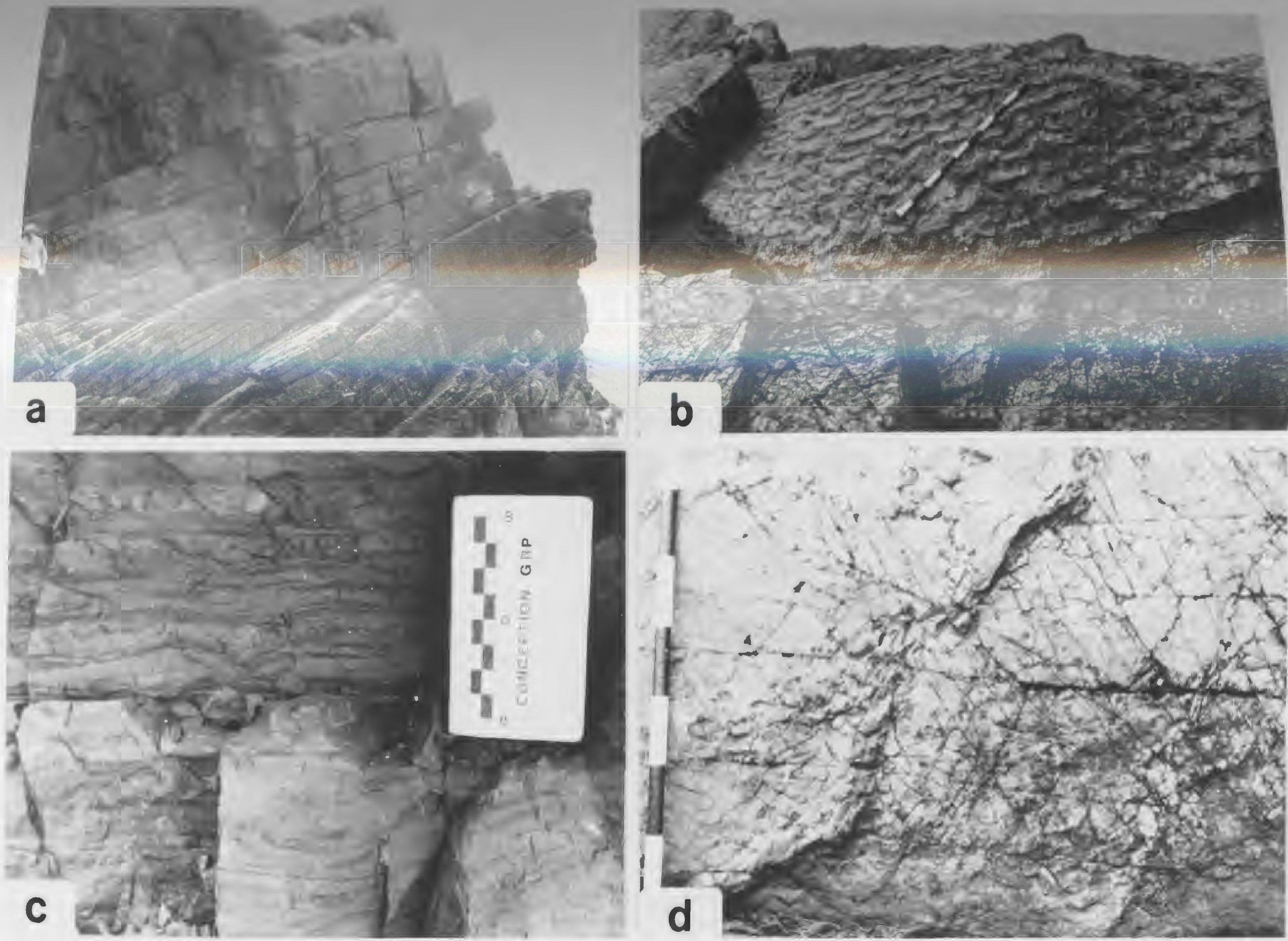
This facies is composed of wavy bedded (as defined by Reineck and Singh, 1980, p. 114; Reineck and Wunderlich, 1968), brown to purple argillites with intercalated siltstones up to one cm thick. Facies C contains linguoid current ripples which are well exposed on bedding surfaces (Fig. 3.14 a,b). Based on a few measurements and a visual comparison between ripples, these small-scale ripples have wavelengths of about 8 to 12 cm, amplitudes of about 2 cm, and maximum inclination of lee-side laminations of approximately 20°. The ripples measured are thought to be representative of all ripples in this facies. Convolute lamination is common (Fig. 3.14c). Rare, thin (1 to 3 cm), clastic dikes are present (Fig. 3.14d).

Figure 3.14: a. General aspect of Facies C at False Cape. Scale bar divisions equal 10 cm.

b. Linguoid ripples on bedding surface at False Cape. These ripples characterize the wavy bedded nature of Facies C. Flow direction is toward the bottom left. Tops to top left. Scale bar divisions equal 10 cm.

c. Wavy bedded siltstone/argillite of Facies C. Note convolute lamination. Scale divisions equal 2 cm.

d. Rare clastic dike on upper bedding surface. Scale bar divisions equal 10 cm.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Facies D: lenticular bedded siltstone/argillite

Description:

Brown to purple, lenticular bedded (as defined by Reineck and Singh, 1980, p. 114) argillites with intercalated siltstones up to one cm thick comprise this facies. Facies D is characterized by straight-crested ripples and an absence of convolute laminae (Fig. 3.15 a,b). Up to approximately 60% of the ripples or sand lenticles are connected in the horizontal and vertical directions (Fig. 3.15c). These unidirectional, asymmetric ripples show internal laminations which are emphasized by grain size changes from very fine and fine grained sandstone to siltstone/shale. Based on a few measurements of ripple dimensions, length to height ratios are less than 20. The ripples measured are thought to be a representative sample of all ripples in this facies. Therefore, these ripples can be distinguished as forming flat-lens lenticular bedding, as described by Reineck and Singh (1980, p. 115). A ripple morphology transitional between straight-crested and linguoid types is rare (Fig. 3.15d). It was not possible to establish a relationship between crest type and style of internal lamination of the ripples in Facies C and D.

Process interpretation–Facies C and D:

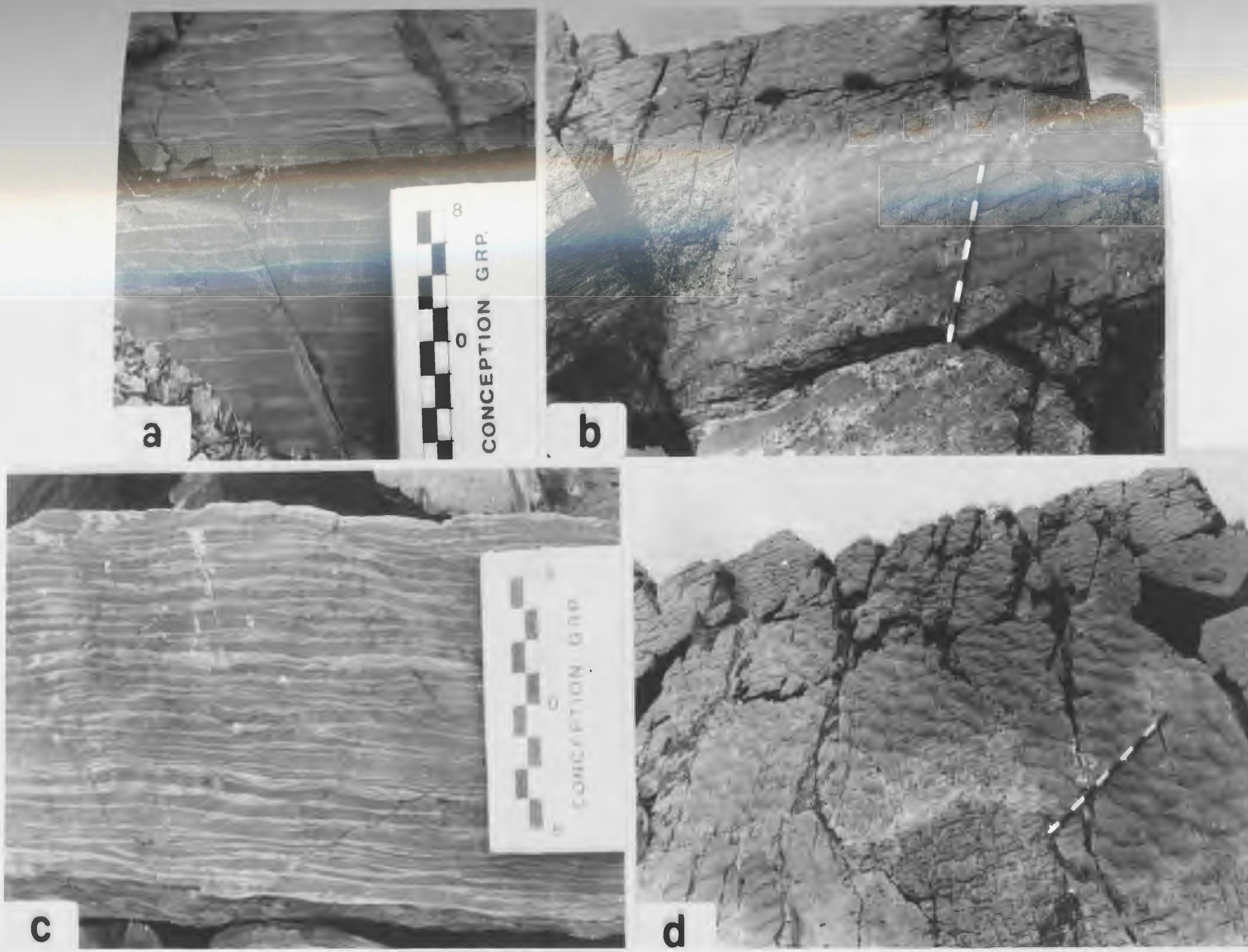
The silt and argillite content of Facies C and D is similar to that of Facies B. However, the presence of

Figure 3.15: a. Lenticular bedded siltstone/argillite of Facies D. Note the absence of convolute lamination and compare with Facies C in Fig. 3.13. Scale divisions equal 2 cm.

b. Straight crested asymmetrical ripples on bedding surface at False Cape. These ripples characterize the lenticular bedded nature of Facies D. Flow direction is toward the bottom left (east). Scale divisions equal 10 cm.

c. Up to 60% of the silt lenticles are connected producing rare flaser bedding transitional between the characteristics of Facies C and D. It has been grouped with Facies D based on the absence of convolute laminae. Scale divisions are 10 cm.

d. Ripple morphology transitional between straight crested and linguoid types. Flow toward the bottom left (east).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straight-crested and linguoid ripples in these facies, combined with a lack of grading and lamination, requires a different process interpretation. The tractional processes that formed these ripples can occur in virtually every marine environment. Speculation as to the specific origin of the ripples is only possible in the context of an interpretation of depositional environment (see the interpretation of facies association II that follows).

Facies E: medium bedded to thickly bedded sandstone

Description:

This facies consists largely of grey to olive green coloured, fine to coarse grained, medium bedded (10 to 30 cm) and thickly bedded (30 cm to 2.0 m), massive sandstones (Fig. 3.16 a,b,c,d) exhibiting sharp upper and lower boundaries with associated facies. At False Cape, this facies is tuffaceous based on thin section examination, and is buff, yellow to white in colour. These are volcanic clastics (Blokhina et al., 1959), or specifically, volcaniclastic sandstones (Fisher, 1961, 1966).

Based on visual estimates the sand to shale ratio is approximately 20 to 1. Classical Bouma sequences are poorly developed. Rare grading, lamination, shale clast horizons (Fig. 3.17a), and poorly developed solemarks occur within the thicker tuffaceous beds. The poor preservation of grain boundaries due to fracturing and

Figure 3.16: a. Medium bedded to thickly bedded sandstones of Facies E. Note the consistency of bed thickness both laterally and vertically. Location is Holyrood Pond. Divisions of scale bar in centre foreground equal 0.5 m.

b. Facies E at False Cape. No scale was available.

c. Thickly bedded part of Facies E at False Cape. No scale was available.

d. Medium bedded tuffaceous sandstones at False Cape. Scale bar divisions equal 10 cm.

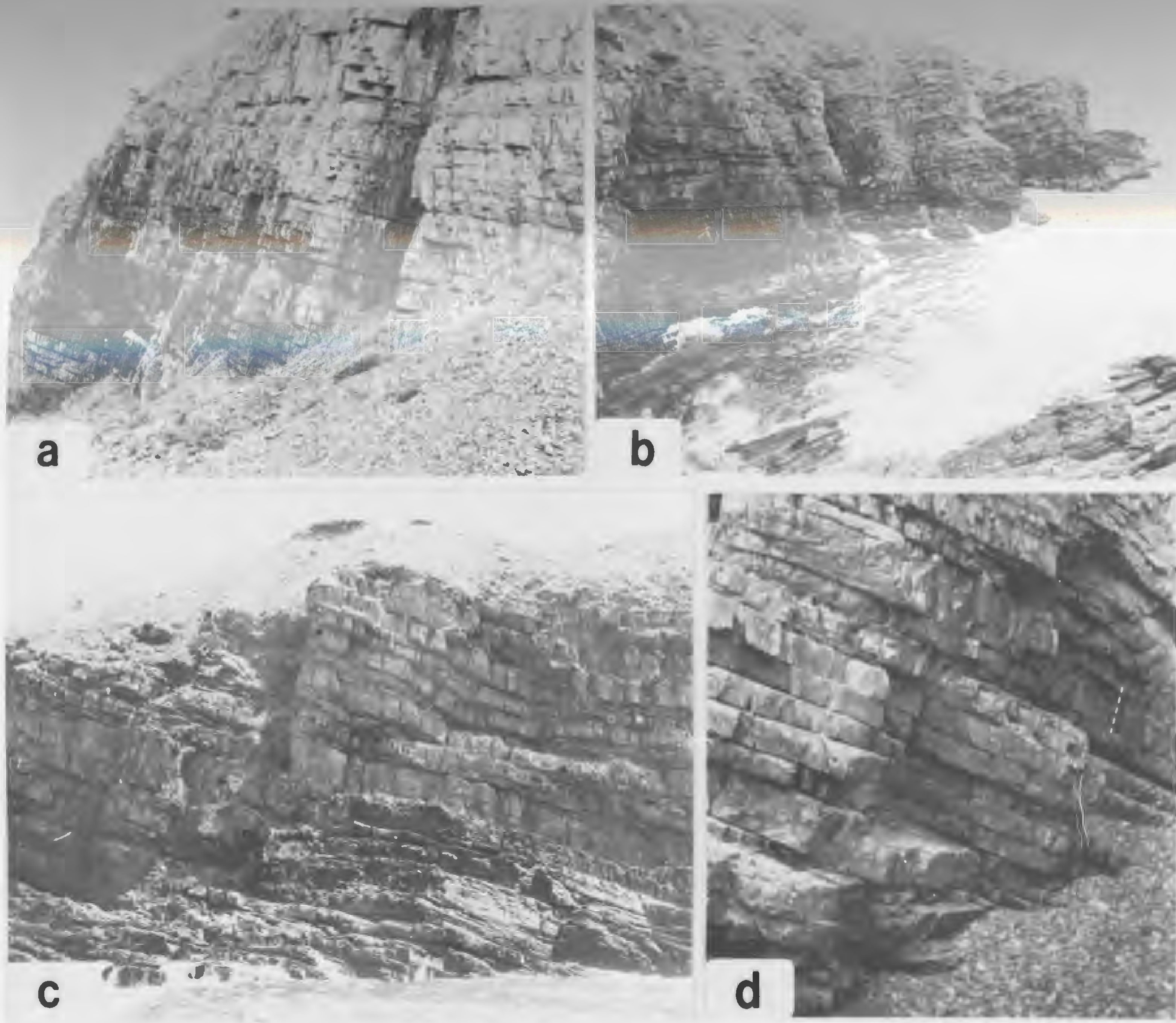
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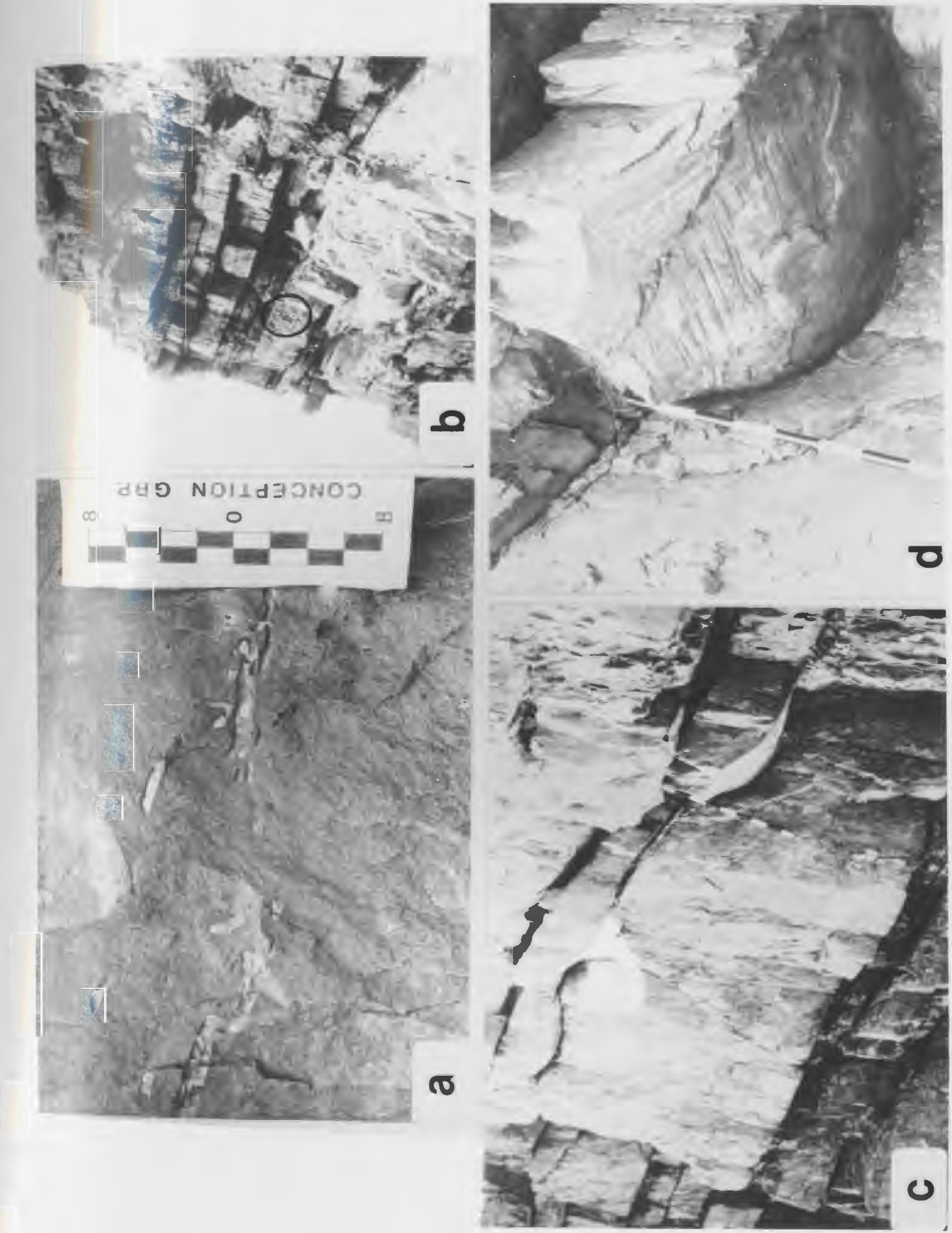
Figure 3.17: a. Shale clast horizon in tuffaceous sandstone of Facies E at False Cape. Scale divisions equal 2 cm.

b. Channel-like features (arrow) exhibiting lateral thickness changes in Facies E at False Cape. Scale (circled) bar divisions equal 10 cm.

c. Close-up of channel-like features in b) above. Note the thinning of the bed underlying the erosive channel-like features.

d. Solemarks, resembling longitudinal ridges described by Dzulynski and Walton (1965, p. 61-65), on the base of the channel-like feature.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alteration precluded fabric analysis in the tuffaceous sandstones.

Peculiar to the tuffaceous beds at False Cape, is the presence of channel-like features which exhibit lateral thinning and thickening (Fig. 3.17 b,c). A thin (1 to 4 cm) shale layer separates the lenticular-shaped sand from the underlying sand bed. The lower surface of the lenticular sand has solemarks which resemble longitudinal ridges (Fig. 3.17d) described by Dzulynski and Walton (1965, p. 61-65). These solemarks are thought to form parallel to current direction, which indicates an east-west sense of flow at False Cape. Poor preservation did not allow unequivocal paleoflow data to be obtained.

In contrast to the buff colour of the tuffaceous volcaniclastic sandstones at False Cape, the grey and green volcaniclastic sandstones that define Facies E elsewhere exhibit abundant solemarks, including prodmarks, groove and striation casts, and flutes (Fig. 3.18 a,b,c,d,e). Based only on visual comparison, the size and character of solemarks is not related to bed thickness.

Convolute lamination is present in the non-tuffaceous sandstones (Fig. 3.18f). In addition, minor amalgamation and associated thickening and thinning of beds occurs (Fig. 3.19 a,b,c). Sharp truncation surfaces, with associated pebble lags, indicate erosional downcutting (Fig. 3.19d).

Figure 3.18: Solemarks in sandstones of Facies E

a. Load marks and prod marks (lower left). Scale divisions equal 10 cm.

b. Scratch marks, groove and striation casts. Scale divisions equal 10 cm.

c. Flute marks. Scale in centimetres (below) and inches (above).

d. Groove casts. Scale divisions equal 10 cm.

e. Flute marks. Scale divisions equal 10 cm.

f. Convolute lamination is present in non-tuffaceous sandstones of Facies E at Holyrood Pond. Scale divisions equal 10 cm.

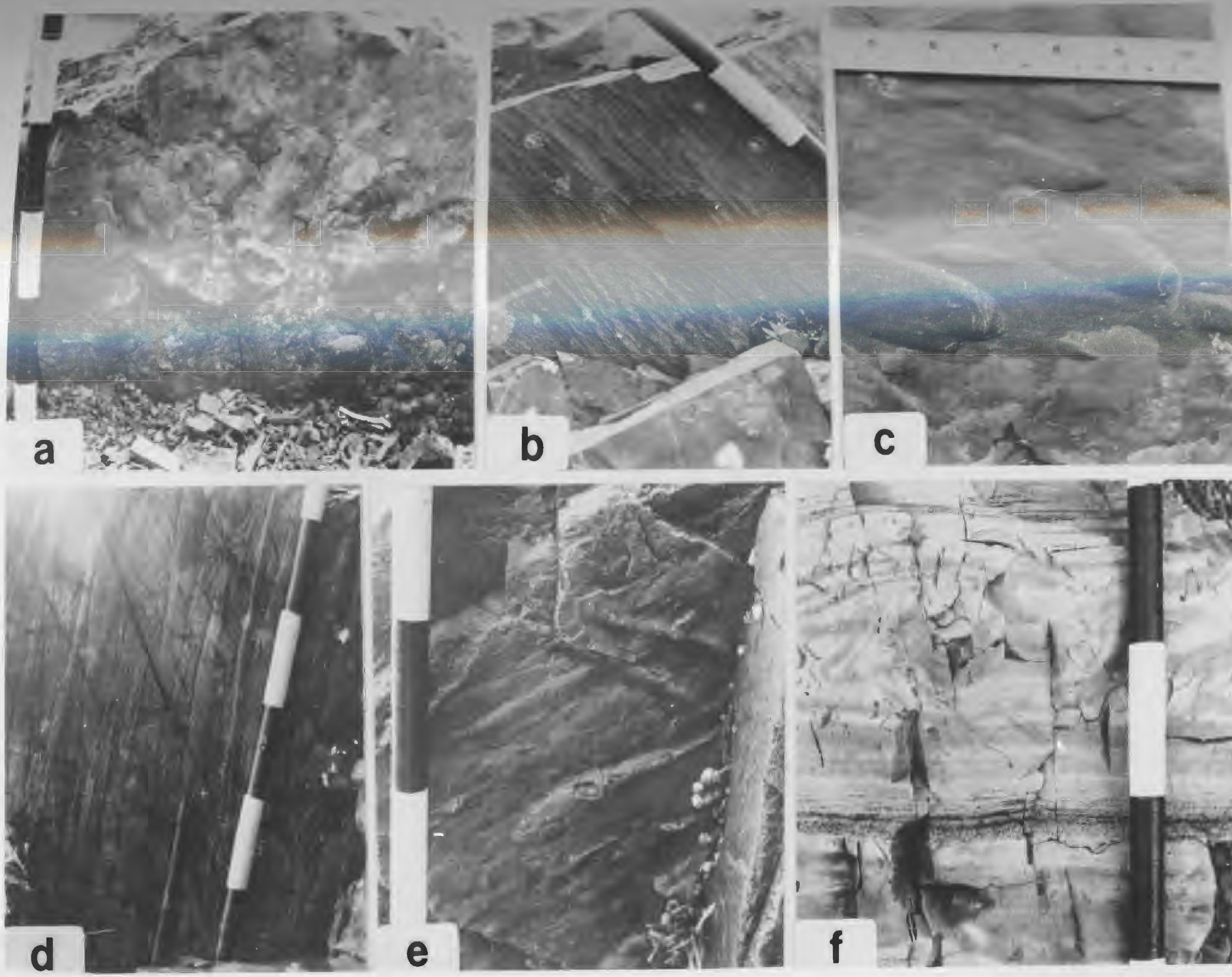
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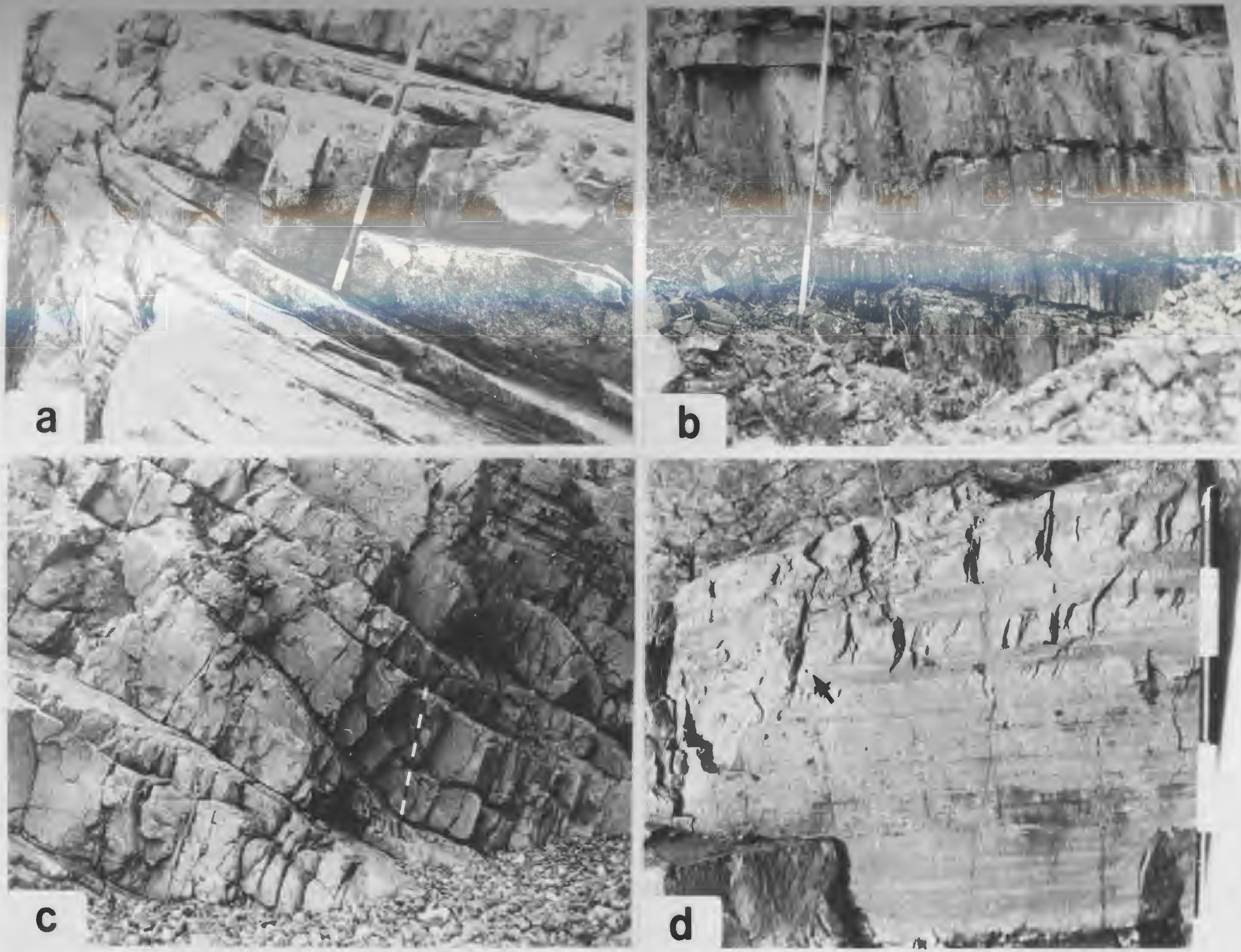
Figure 3.19: Examples of lateral thickness changes within Facies E.

a. At False Cape medium bedded tuffaceous sandstones pinch out. Scale bar equals 10 cm.

b. Small erosional downcutting in medium bedded, non-tuffaceous sandstones at Holyrood Pond. Note pinching out of lower beds to the right of the scale bar. Scale bar divisions equal 0.5 m.

c. Lateral bed thickness changes at False Cape.

d. At Holyrood Pond a sharp truncation surface (arrow) occurs with a pebble lag.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Process Interpretation-Facies E:

Abundant shale clasts "floating" within the bed, poorly developed grading, and diffuse lamination in moderately well sorted sandstone is typical of air-fall ash which has undergone later gravity flow (Sigurdsson et. al, 1980). The presence of "floating" clasts (Fig. 3.17a) and an absence of grading, was originally interpreted to be characteristic of grain flows (Stauffer, 1967). More recent work on sediment gravity flows (Middleton and Hampton, 1973; Lowe, 1976b, 1982) suggests that several processes of lift and/or grain support may have operated simultaneously, as described by Lowe (1979, p. 80). A comparison between the general characteristics of the density flow deposits of Facies E to the classical Bouma (1962) turbidite sequence and to the cohesive gravity flows of Lowe (1982, p. 292 and 294) shows similarities to both (Fig. 3.20). Based on these similarities, perhaps both turbidity currents and cohesive flows (or modified grain flows) were responsible for deposition of some beds within the lower Conception Group.

## Facies F: thickly bedded sandstone

Description:

Thick (0.5 to 4.0 m), medium to coarse grained, grey to green sandstones comprise this facies (Fig. 3.21a). Minor granule-size conglomerate is exposed at St. Stephen's (section 2)(Fig. 1.1). This facies exhibits

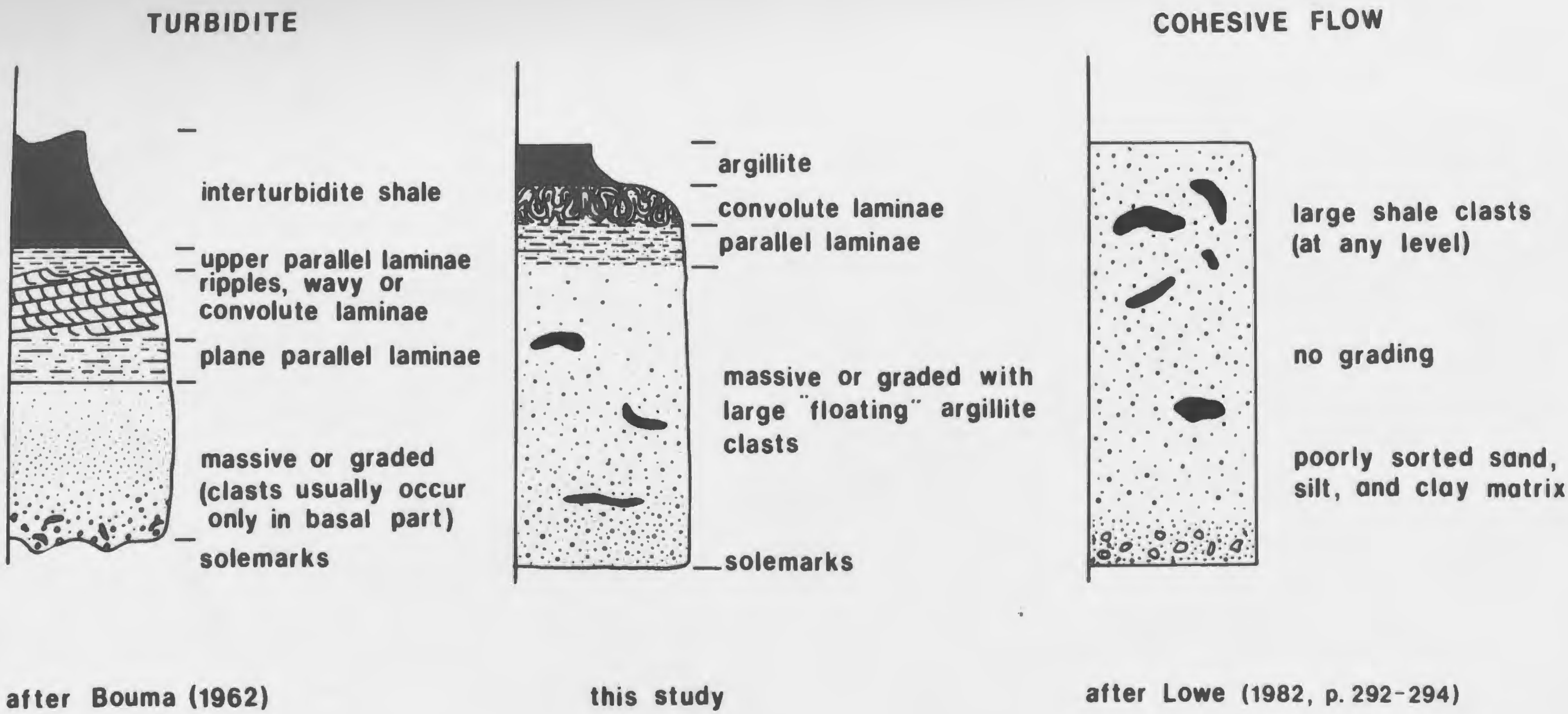
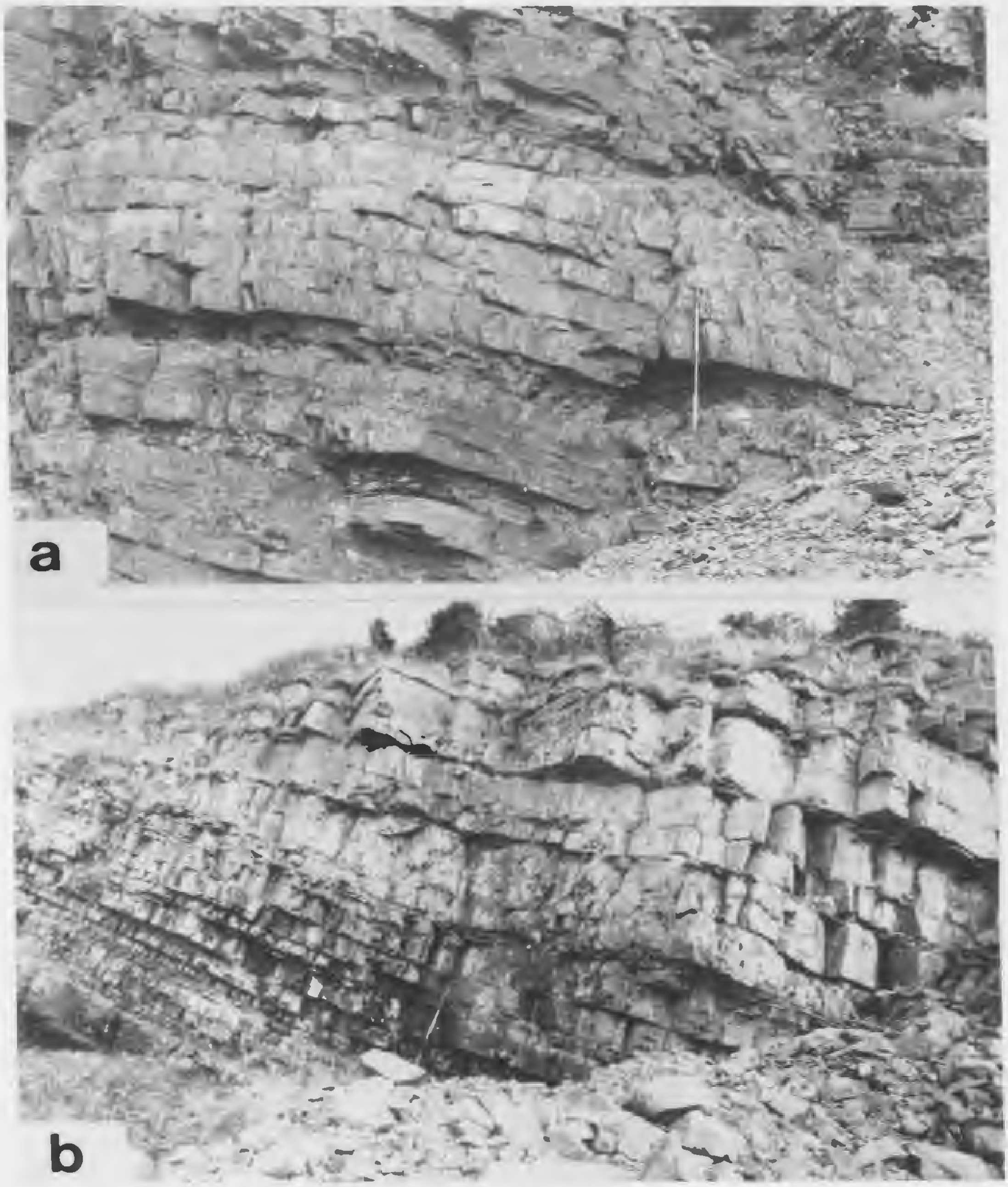



Figure 3.20: A comparison between the general characteristics of density flow deposits of this study, the turbidite sequence of Bouma (1962), and cohesive flows described by Lowe (1982).

Figure 3.21: a. Thickly bedded sandstones of Facies F at Holyrood Pond. Note the continuity of the bed thickness both laterally and vertically. Scale bar divisions equal 0.5 m.

b. Well developed thickening- and coarsening-upward cycle capped by thick beds of Facies F. Location is Holyrood Pond. Scale bar divisions equal 0.5 m.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Bouma AE sequences, multiple grading, large scale (up to 0.5 m) convolute laminae, scouring, and large (long axes up to 0.5 m) shale clasts. Sand to shale (argillite) ratios are about 20 to 1.

Bed thicknesses are variable. Note the lateral thinning of one bed in figure 3.21b. Typically, bed thicknesses range up to 4 m (Fig. 3.22a). The 3 to 4 m beds are composed of several amalgamated, graded layers, which produces a multiple graded appearance (Fig. 3.22b). Anomalously large, "floating", shale clasts (Fig. 3.22c) and shale clast horizons (Fig. 3.22d) occur in the middle of the 3 to 4 m thick beds.

The most characteristic feature of this facies is the presence of convolute lamination. There are two observed styles of convolute lamination: 1) convolute laminae may comprise all or part of a bed resulting in convolute bedding (Fig. 3.23a), or 2) convolute laminae may occur in layers, several of which may amalgamate to form a bed (Fig. 3.23 b,c). The scale of layers in the second type varies from about 10 to 15 cm to greater than 40 cm.

A small outcrop at St. Stephen's represents a partially coherent slide (Fig. 3.24 a,b), equivalent to a type B slide of Nardin <u>et al.</u> (1979). It consists of 1 to 3, 15 cm-thick, graded beds (distribution grading) of small-pebble and granule-size conglomerate surrounded by a matrix of medium to coarse grained sandstone. The normal

Figure 3.22: a. Four metre thick coarse sandstone bed of Facies F at Holyrood Pond. Scale bar divisions equal 0.5 m.

b. Amalgamation surface (arrow) between successive graded layers within the 4 m thick bed at Holyrood Pond. Scale bar divisions equal 10 cm.

c. Large shale clast parallel to bedding. This clast is "floating" within the 4 m thick bed. Scale in centimetres (below) and inches (above).

d. Horizon of smaller shale clasts (dark areas) adjacent to large clast shown in c) above. Scale divisions equal 2 cm.

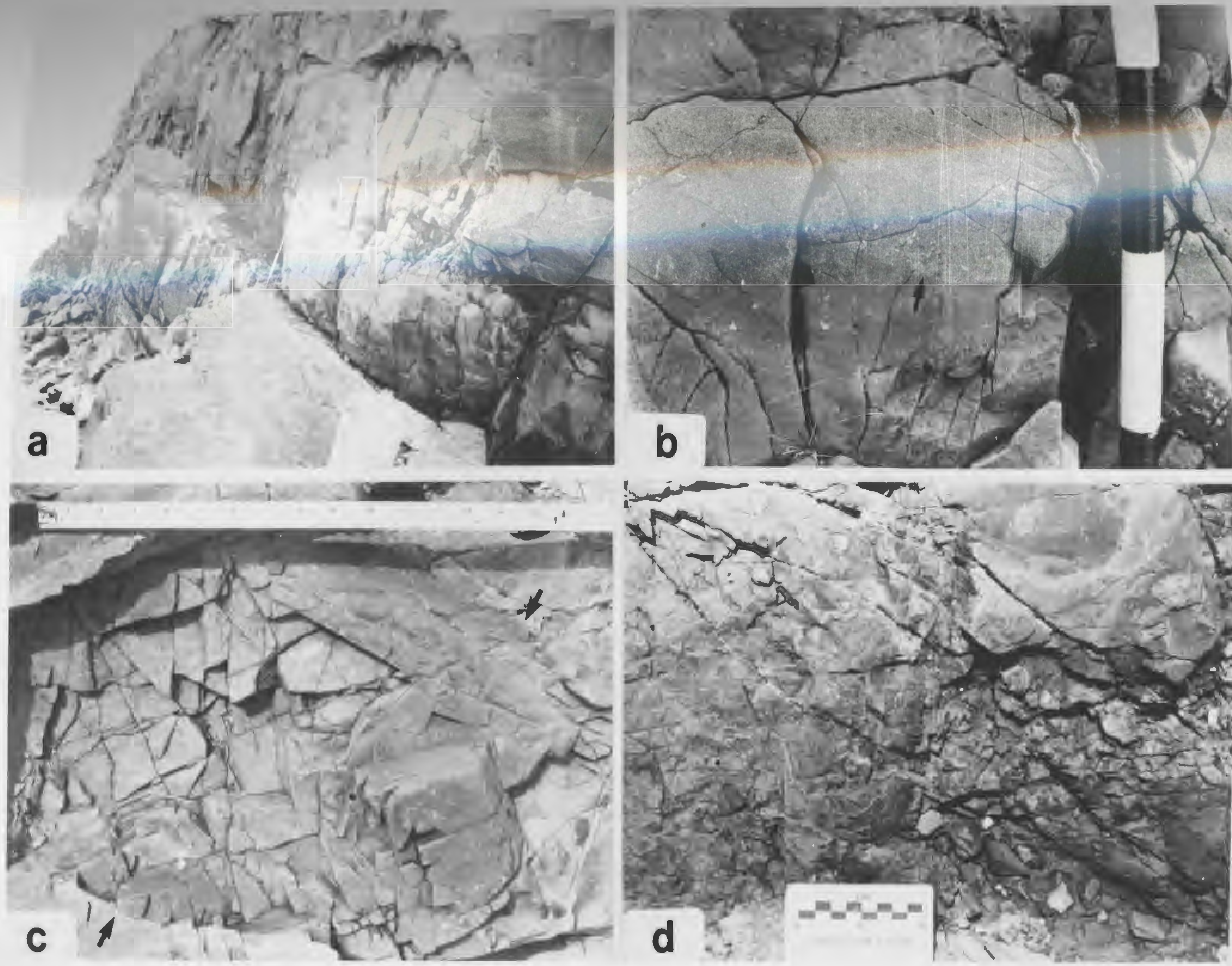
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Figure 3.23: a. Large scale convolute laminae resulting in convolute bedding in Facies F at Holyrood Pond. Scale bar divisions equal 10 cm.

b. Convolution of successive layers which amalgamate to form a bed.

c. Same as a) above. Subsequent erosion of the lower bed confirms that the convolution is syndepositional.

d. Same as b) above on a slightly larger scale.

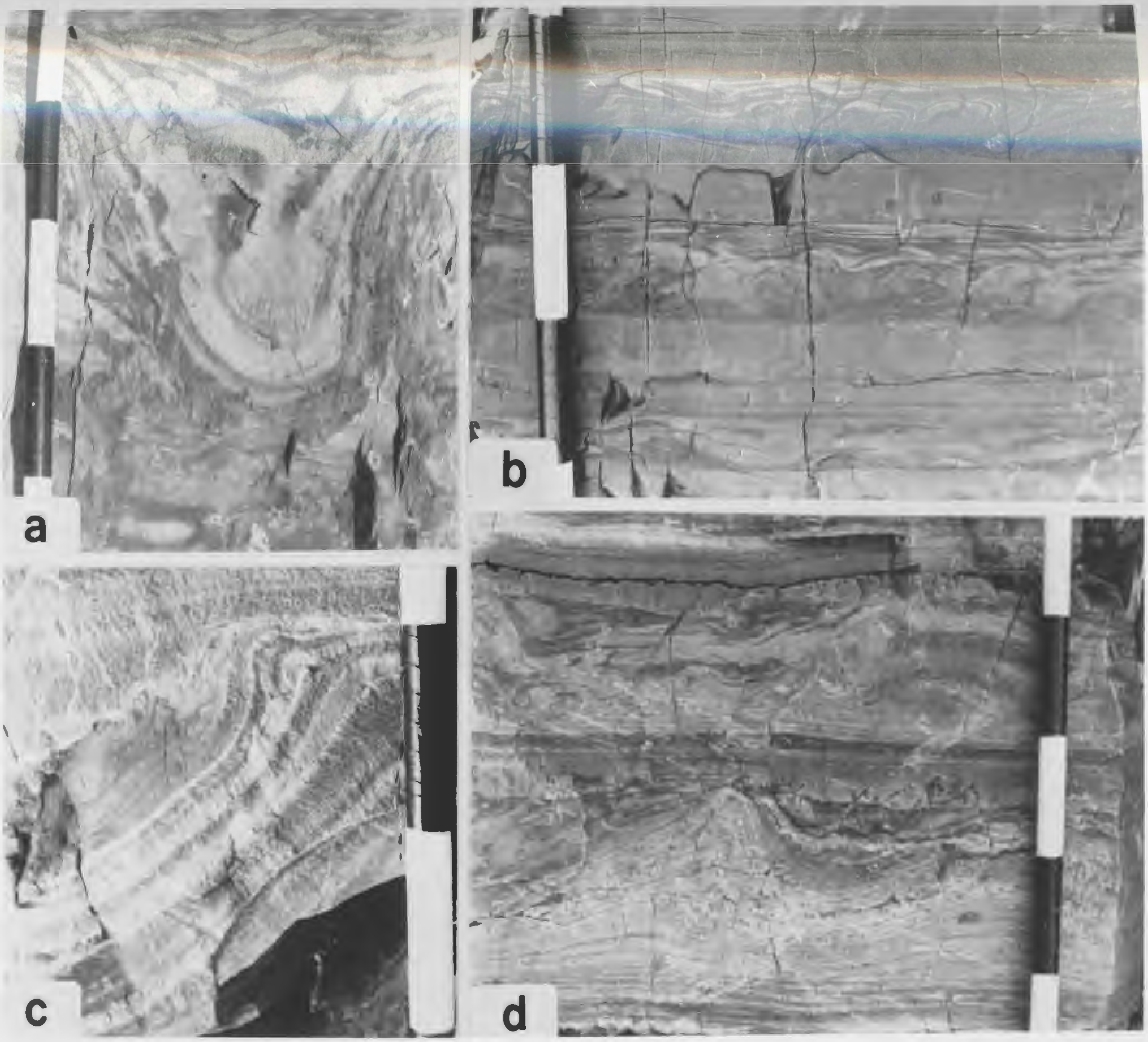
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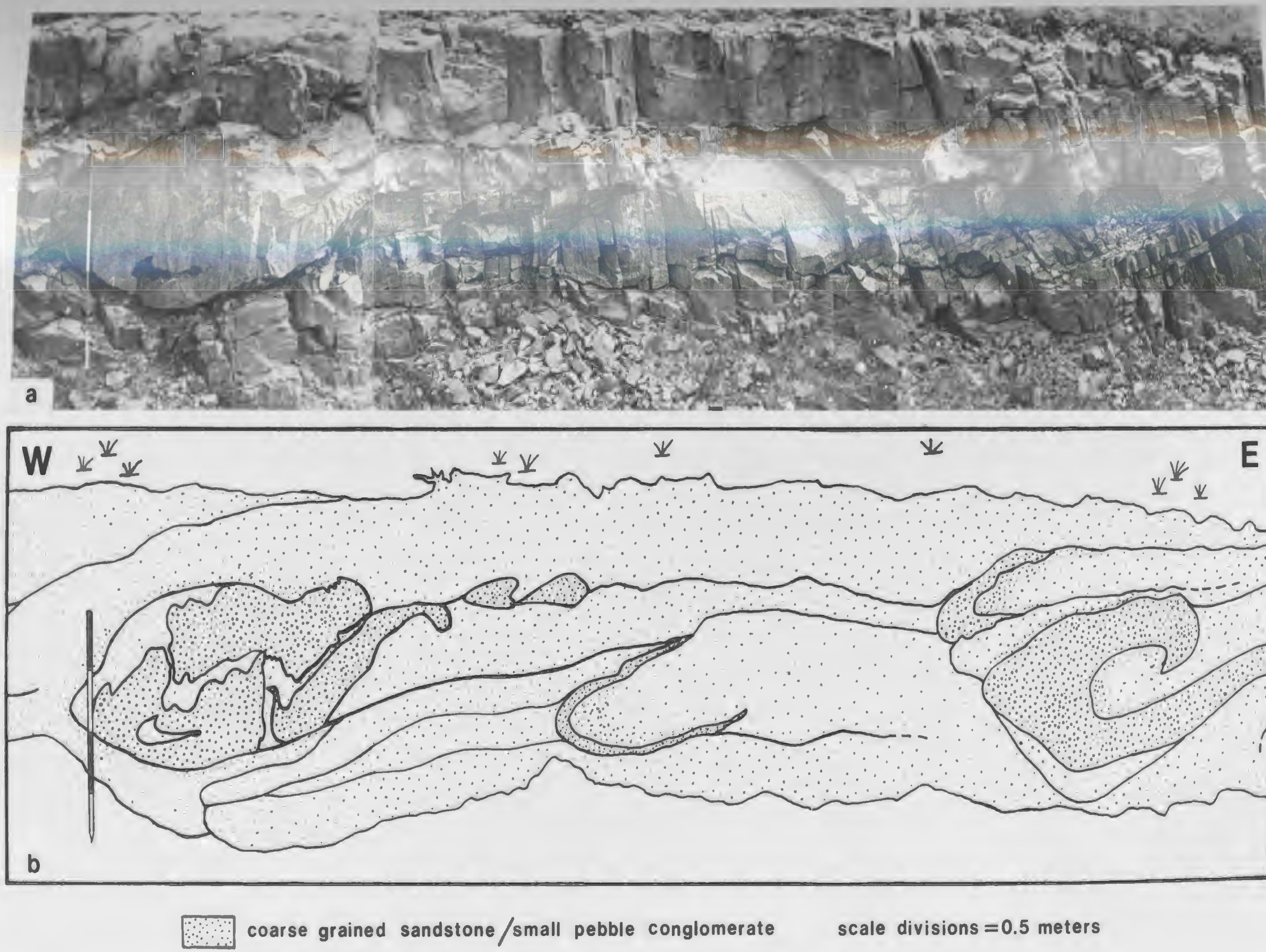
Figure 3.24: a. At a roadside outcrop in St. Stephens, a partially coherent slide, equivalent to a type B slide of Nardin et al. (1979), occurs. Scale bar divisions equal 0.5 m.

b. Sketch of a) above to highlight the three slide folds. The fourth slide at this location (mentioned in text) was not photographed but is situated to the left of the outcrop shown in this photo.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E
b
coarse grained sandstone / small pebble conglomerate
scale divisions = 0.5 meters

grading was used to determine that the slide involves four sedimentary recumbant folds. These hook-like structures are called "slump overfolds" by Crowell (1957, p. 998). The slide fold axes are oriented approximately in a north-south trend. The hooked shape may indicate a westerly direction of paleoslope. However, as discussed previously for Facies A, the usefulness of single fold orientations for determining paleoslope is controversial (Lajoie, 1972), and the westward sense of paleoslope deduced from these folds is given little weight.

Process interpretation-Facies F: Based on the presence of grading and Bouma AE sequences, deposition from turbidity currents is the dominant depositional mechanism for sandstones of this facies. Multiple graded beds up to 4 m thick, formed as a result of several successive turbidity flows. These "floating" clasts (Fig. 3.22c) may be evidence of high flow densities (Fig. 3.14), giving buoyant support, as described by Lowe (1976a,1976b,1982).

The origin of the convolute lamination is enigmatic (Potter and Pettijohn, 1977, p. 228). Possibly the convolute laminae in this facies are due to a combination of thixotropic transformation and loading due to the rising pressure of trapped fluids and reduced strength and density of liquified layers (ibid.). This explanation can be invoked when an entire bed, and all the Bouma divisions it contains, has undergone convolution. However, where convolution is exclusively contained in the Bouma C division, it may represent the result of forces which acted on newly deposited laminae which were still under the influence of a moving current (Dzulynski and Smith, 1963, p. 622).

Facies G: chaotic deposits

Description:

The abrupt upper boundary of the Mall Bay Formation with the overlying Gaskiers Formation is marked by low angle faults and a 1 to 2 m, intensely weathered, chaotic, conglomeratic zone. Clasts range in size from granules (1-2 mm) up to large pebbles (64 mm) within a matrix of medium to coarse grained sand. The clast composition varies from sandstone to volcanics with no apparent fabric developed. Many of the clasts are fractured and faulted.

Process interpretation:

This conglomeratic zone may represent a debris flow. A conclusive, sedimentologic, process interpretation was precluded because of the fracturing, faulting and associated intense weathering.

## Facies Associations

Facies associations are groups of facies that occur together and are considered to be genetically or environmentally related (Reading, 1980, p. 5). Facies associations provide additional evidence which makes environmental interpretation easier than treating each facies in isolation (ibid.).

Three facies associations are recognized in the Mall Bay and Drook Formations in the lower Conception Group (Table 3.1): I) a thinly bedded sandstone/argillite facies association, II) a rippled siltstone/argillite facies association, and III) a medium to thickly bedded sandstone facies association which is tuffaceous only in the Mall Bay Formation.

Facies association I consists of two facies: Facies A) thinly bedded to medium bedded sandstone/argillite; and Facies B) argillite (Table 3.1). The general aspect of facies association I is shown in Fig. 3.2 a,b and Appendix I, p. 187, 0-25 m. Facies association II also consists of two facies: Facies C) wavy bedded argillite/siltstone; and Facies D) lenticular bedded argillite/siltstone (Table 3.1). The general aspect of facies association II is shown in Fig. 3.14a and in Appendix I, p. 185, 0-35.5 m. Facies C and D were differentiated from eachother based on differences in amount of siltstone, presence of convolute laminae, and type of ripples. Three facies comprise facies association III (Table 3.1, Fig. 3.15, Appendix I, p. 183, 198-218 m): Facies E) medium bedded to thickly bedded sandstone; Facies F) thickly bedded sandstone; and Facies B) argillite (described previously). Facies G could not be assigned to a specific facies association because it is a single occurrence within the study area.

Facies association I is best exposed at Point La Haye, Cape English, and Frapeau Point. Excellent exposure of facies association II occurs at False Cape, with good exposure north of Point La Haye. Excellent exposure of facies association III exists at Holyrood Pond and Wild Cove. In the Mall Bay Formation at Point La Haye, northeast of the lighthouse along the coast toward Double Road Point, there is continuous stratigraphic exposure of all three facies associations. Facies association III is thinner at Double Road Point than at False Cape (6 m vs. 60 m).

Some of the interpretations of depositional environment that follow depend on paleocurrent information. For this reason, the paleoflow directions for each facies association are summarized here (Fig. 3.25). The paleocurrent data presented here represents only field data such as solemarks and ripples. A more comprehensive coverage of paleocurrents derived from both field and sandstone fabric measurements appears in chapter 5. The results of the detailed paleocurrent analysis in chapter 5 will be used to determine paleogeography as part of basin analysis (chapter 6).

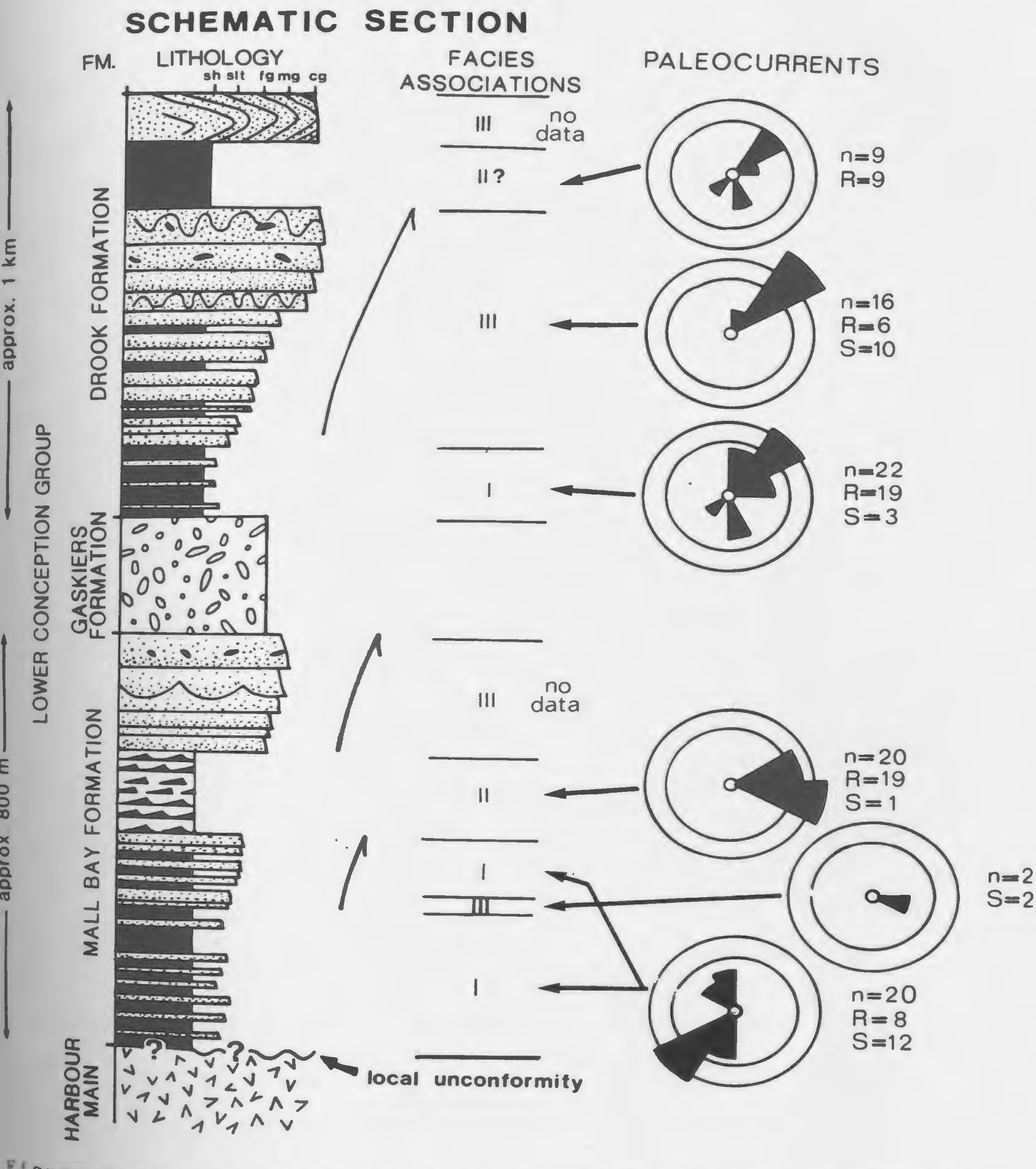



Figure 3.25: Relationship between paleocurrents (n= total number of bed averages represented, R=ripples, S=solemarks) and facies associations for the lower Conception Group. A more comprehensive paleocurrent analysis follows in chapter 5 (compare Figs. 5.2, 5.3, 5.4, and 5.5). On rose diagrams, outer ring=10 bed averages, inner ring=5 bed averages).

## Depositional Environment

### Interpretation of facies association I

Facies association I represents a transition between the fan fringe and basin plain settings of Mutti (1977) within, and adjacent to, the outer fan part of a turbidite fan depositional system (Table 3.1). Evidence for this interpretation includes the even and parallel bed surfaces, the fine to very fine sand and silt grain size, constant bed thickness, a low sand to shale (argillite) ratio estimated to be less than 0.5, and the presence of Bouma DE divisions (Appendix I).

The thin (3 m) interbedded packets represent the distal effects of fan aggradation or fan switching. Identical characteristics have been documented as evidence for fan fringe settings by Ricci Lucchi (1975) and Pickering (1983). Other workers might consider these 3 m-thick packets to represent asymmetric cycles (i.e., coarsening- and/or thickening-upward cycles). Hiscott (1981) states that many observed asymmetric cycles in turbidite sequences are fictitious. Also, Carr (1982) said "what we perceive as order in nature may be a reflection of our desire for order". For these reasons, and because unequivocal asymmetrical cycles are not evident, cyclicity and attendant environmental implications, do not play a role in interpretation of association I. Although this facies association is similar to the fan lateral margin deposits described by Pickering (1983), it does not contain: a) the characteristic small channels at various angles to the slope direction; b) lobe deposits associated with channels; or c) abundant clastic dikes and other soft-sediment deformation. Furthermore, the paleocurrent indicators of this association are consistently to the northwest and southwest (Fig. 3.25). This suggests that these thinly bedded deposits are distal, axis-parallel, basinward equivalents of the inner and middle fan rather than fan lateral margin deposits which occur away from the fan axis (Pickering, 1983).

The minor channels observed in Facies A are possible in fan fringe or basin plain settings when flows extend basinward of the mid-fan due to channel switching or flow-deflecting topography (Walker, 1978).

Some of the rippled and graded argillite of Facies B of this facies association may be characteristic of the lower basin-slope (see facies association II interpretation)(Pickering, 1982a; Pickering, 1983, Fig. 15). Similar characteristics, such as the presence of grading, solemarks, and rippled fine grained sand or silt, make it difficult to distinguish between deposits of the slope and those of the basin plain. Depending on the geometry and size of the submarine fans, the slope and basin plain need not be far removed from each other. This is especially true in the depositional and tectonic setting envisioned for these volcaniclastic rocks (see chapters 4 and 6), which is not that of a broad, stable, continental margin. Also, in the Precambrian, proper selection of mud and silt depositional environments is hindered by the absence of trace and body fossils and pelagic carbonates. If present, these features may indicate water depth in Phanerozoic rocks, particularly in the Mesozoic and Cenozoic.

Some argillite which has been grouped into this facies association may also represent mud blankets over submarine fan lobes (Hiscott, 1980). In particular, this would be anticipated for argillite packets which are associated with sandstone packets (lobes).

Interpretation of facies association II

Three depositional interpretations of facies association II will be considered: 1) interchannel or channel margin; 2) basin plain or rise; and 3) lower basin-slope.

Mutti (1977) presents diagnostic characteristics of interchannel and channel margin sediments. The lack of associated channel-fill sequences is strong evidence against a channel-related origin for facies association II. The absence of Bouma CE and BE sequences, small crevasse channels or splays and associated climbing ripples, also does not favour this interpretation.

Intermittent bottom currents could explain development of the lenticular and wavy bedded siltstones/argillites. Straight-crested and linguoid ripples have been documented in water depths from 490 m to 4000 m on the slope and basin floor off eastern North America (Heezen and Hollister, 1964; Reineck and Singh, 1980, p. 490). Ripples formed in these water depths are affected by bottom currents which produce flow patterns transverse to the turbidite fan axis. The easterly paleoflow direction of these ripples (Fig. 3.25) in strata of facies association II agrees with similar observations from the Recent.

If a basin floor depositional setting is accepted, then the stratigraphic position of this facies association relative to other associations becomes a problem. One would anticipate finding this argillite and intercalated rippled siltstone stratigraphically below the basin plain and fan fringe deposits of facies association I. This is not observed. Facies association II occurs stratigraphically between the inferred basin plain (facies association I) below, and inferred lobe/lobe fringe deposits (facies association III) above (see sections 5 and 6 in Appendix I). Although speculative, it seems more likely that this facies association represents current reworking of slope deposits (Table 3.1). Pickering (1982a) described facies very similar to these rippled siltstones/argillites (his Facies II) and interpreted them to be the result of low density, low velocity, turbidity currents on an upper basin-slope (Pickering, 1982a). This interpretation is based largely on stratigraphic relationships, where this facies is "sandwiched" between submarine fan deposits below and a fluvio-deltaic complex above (ibid.). Deposits similar to those of the upper basin-slope, could be anticipated on a lower basin-slope, but would be interbedded with channel and lobe deposits of the submarine fan (Pickering, 1983, Fig. 15) and would not show the effects of wave reworking. The stratigraphic position of facies association II suggests a location in the lower basin-slope area or in the transitional zone between upper and lower basin-slope, laterally adjacent to a submarine fan.

Interpretation of facies association III

The medium bedded to thickly bedded, massive, and channelized sandstones of this facies association represent deposition on the smooth and partially channelled portion of suprafan lobes, and on lobe fringes, in a lower mid-fan depositional setting (Table 3.1).

Evidence for this interpretation includes the presence of Bouma ABC(E) sequences, very coarse to medium grained sandstone, amalgamation, small ($\leq$1 m) erosive downcutting, and a few well developed coarsening- and thickening-upward cycles (see sections 3 and 5 in Appendix I). Typically, the thickly bedded sandstones (Facies F) constitute the upper portion of these cycles (Fig. 3.21b). Within these cycles, a transition from Facies B, through E to F, occurs over an approximate stratigraphic thickness of 30 m (see top of section 3 in Appendix I), although one cycle is only 10 m thick (150 to 160 m on section 3). Overall, cyclicity is rare, with about 5% to 10% of this association exhibiting cycles of this scale.

The observed lateral thickening and thinning of beds may be due to erosional downcutting or the topography of the mid-fan area (Fig. 3.21b). Lateral thickness changes of lobe deposits, on a similar scale, have been documented by Mutti et al. (1978) and Ricci Lucchi and Valmori (1980). Similar characteristics of lobes in the middle fan and parts of the outer fan have been observed in ancient studies by Pickering (1981), Mutti et al. (1977) and in the modern studies of Normark et al. (1979) and Normark (1978). The variation in bedding thickness of Facies E versus Facies F is a function of specific location within the mid-fan area and the extent of fan switching and development. Fan development in the basin is not only dependent on the amount of sediment supply. Changes in slope gradients due to penecontemporaneous volcanic and tectonic activity are also important factors. These factors will be discussed in chapter 6.

The distinction between lobe and lobe fringe interpretations is arbitrary. Packets of coarsening- and thickening-upward sandstones displaying some lateral thickness changes are interpreted as lobes, and medium bedded sheet sandstones that seldom or never display cyclicity, are interpreted as fringes of the lobes. In general, the lobe fringes are like those described by Ricci Lucchi (1975), Mutti (1977), and Pickering (1981,1983). The coarsening- and thickening-upward asymmetric cycles are of a medium scale (i.e., 10 m to 30 m), second-order, nature. Smaller scale (i.e., 2 m to 10 m), third-order cyclicity, like that of Ghibaudo (1980), disputed by Hiscott (1981), was not recognized. In addition to the medium scale cycles, there are first-order, overall coarsening- and thickening-upward cycles several hundreds of metres thick. A cycle of this order of magnitude is represented by the entire Mall Bay Formation and about 80% of the exposed Drook Formation within the study area.

Two types of lobe sequences have been described by Pickering (1981). The relative difference between Pickering's two lobe types is based on their vertical stratigraphic frequency within fan fringe deposits (ibid.). Attempts to apply Pickering's type I and II lobe sequences to the Conception Group proved very subjective. Two styles of lobe deposition do exist in the Conception Group, although they are not those described by Pickering (1981). Firstly, there are well developed thickening- and coarsening-upward cycles (Fig. 3.21b) about 10 m to 30 m thick (255 m to 265 m on section 3 in Appendix I) containing some beds that pinch out laterally. These pinchouts reflect the topography of the submarine fan lobes, being thickest toward the center of the lobe, and thinner toward the margin. Secondly, there are continuous vertical sequences up to 30 m thick (Fig. 3.16 a,b), containing beds of remarkably constant thickness, which display no asymmetric cyclicity (165 m to 180 m, 182 m to 196 m on section 3 in Appendix I). The bed thickness in this depositional style of lobe is approximately the average of the cumulative thickness of beds in the thickening upward cycles. The sandstone packets that typify the second style of lobe deposition are stratigraphically bounded by argillite packets of Facies B (155 m to 200 m on section 3 in Appendix I).These two styles of lobe deposition may represent the effects of lobe progradation (lateral building) and lobe aggradation (vertical building), respectively.

Interpretation of Depositional Environment

Facies analysis and interpretation of facies associations in the Mall Bay and Drook Formations suggests that deposition took place in mid-fan, outer fan, and basin plain settings within a turbidite fan depositional environment with some adjacent, non-fan, lower basin-slope deposits (Fig. 3.26). Based on the above interpretation of facies associations, the turbidite fan model of Walker (1976a) has been modified to include a volcanic island

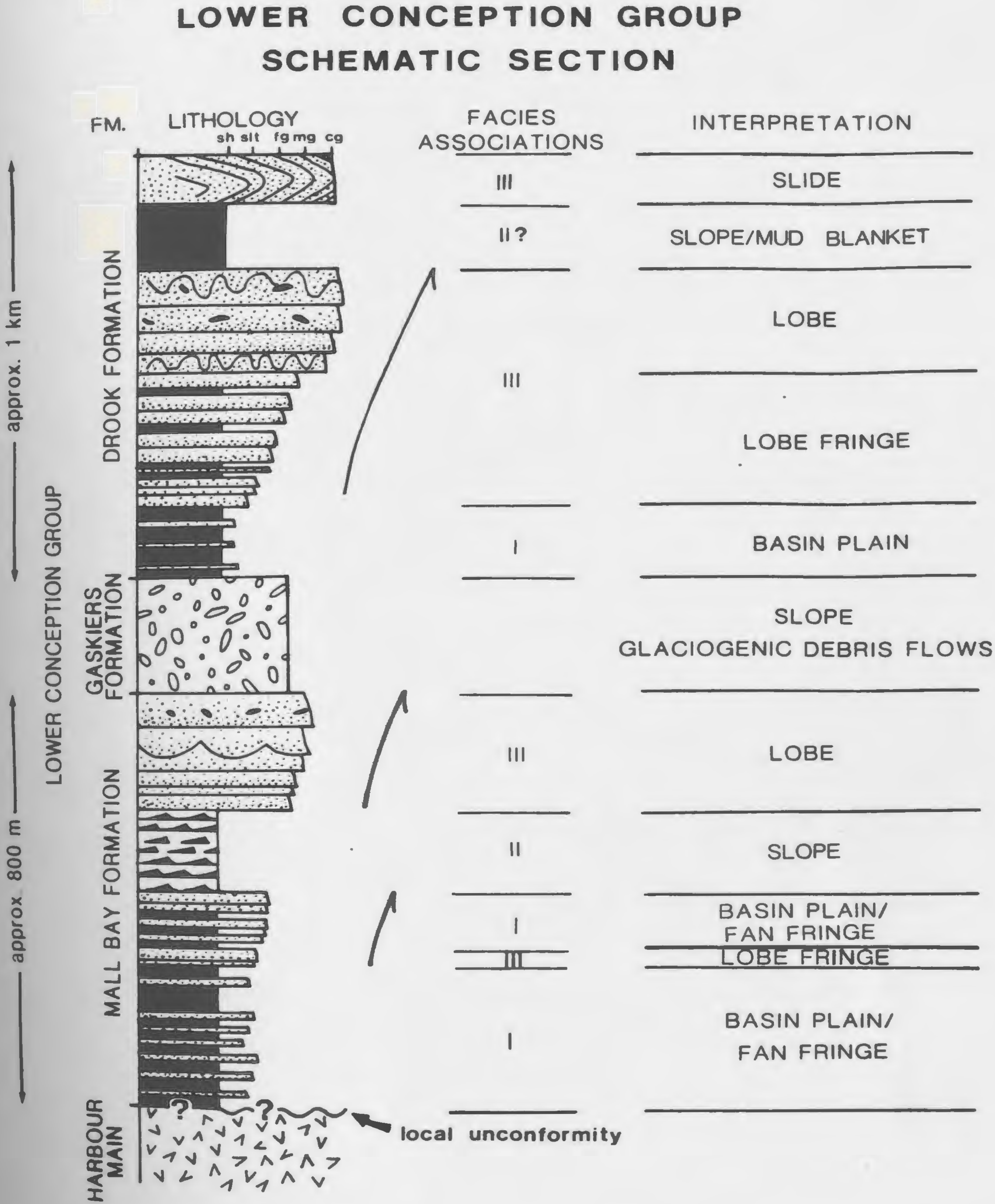



Figure 3.26: Schematic section of the approximate thicknesses, lithologies, facies associations and the inferred depositional environment within a turbidite fan depositional system. Note the overall first-order thickening- and coarsening-upward nature of both the Mall Bay and Drook Formations.

(Fig. 3.27). This model is not meant to indicate paleogeography. In chapter 6 this model will be incorporated into an interpretation of paleogeography and depositional history of the lower Conception Group based on paleocurrent and petrographic data. In addition, comparisons to modern and ancient examples will then be made.

The three recognized facies associations are common to both the Mall Bay and Drook Formations. In the Mall Bay Formation, there is clearly a coarsening- and thickening-upward sequence from facies association I at the base, upward through association II to association III at the top (Fig. 3.26). In the Drook Formation, this overall sequence of facies associations is not easily recognized. The reasons for this are twofold. Firstly, the Drook Formation is almost twice as thick as the Mall Bay Formation (1500 m vs. 800 m), and therefore seems to be composed of relatively thinner vertical sequences of associated facies. Secondly, unlike the continuous stratigraphic exposure of the Mall Bay Formation, the areal exposure of the Drook Formation is patchy, requiring that the overall sequence of facies associations be inferred from several separate but closely spaced outcrops.

Inner fan and upper mid-fan, large scale (i.e., 10's of metres) channel deposits described by Walker (1975b,1976), Hendry (1978), Hiscott (1980), Hein (1981),

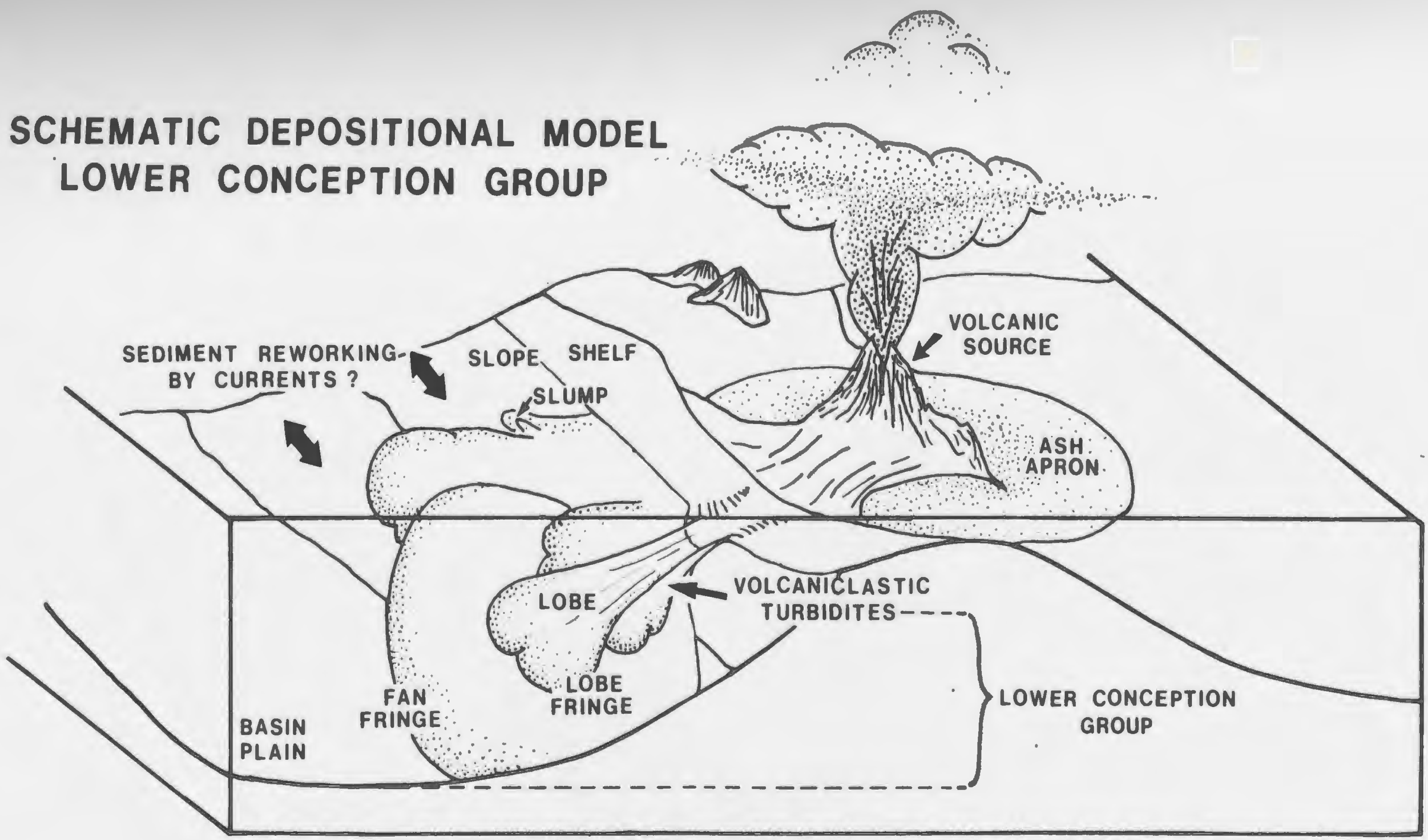



Figure 3.27: General schematic depositional model for the lower Conception Group involving sedimentation associated with a volcanic source area. Within this setting, that part which the lower Conception Group represents is as indicated. The relative position of the volcanic source or the paleoflow directions is not implied. The paleogeography will be discussed in chapter 6 (Figs. 6.1, 6.2, and 6.3).

Eriksson (1981) and Pickering (1982b), were not observed in the lower Conception Group. There are two possible reasons for the absence of large scale channels: a) channels do not exist in this part of the turbidite fan depositional system, or b) channels do exist, but are not exposed in outcrop. Using the facies model of Walker and Mutti (1973) and Walker (1976a) for the turbidite fan depositional system, abundant large channels are not anticipated basinward of the upper mid-fan. Therefore, proximal channelized deposits should exist either in the upper Conception Group (i.e., the Briscal or Mistaken Point Formations) or in the overlying St. John's Group (Fig. 2.4). Based on regional mapping, the St. John's Group was inferred to represent slope sediments (Williams and King, 1979, p. 14-15). If present, the channels may be difficult to observe in the field due to either: a) poor exposure, b) the scale of the channels being larger than the outcrop, or c) channel sides having gentle slopes making them inconspicuous.

## Discussion

It is difficult to "pigeon-hole" facies associations into the traditional turbidite fan environments; i.e., the inner, middle and outer fan settings of Walker (1978) and Normark (1978). Mutti (1977) and Mutti and Ricci Lucchi (1978) have attempted to bridge this gap by introducing terms such as fan fringe, lobe fringe, interchannel and channel mouth to describe the facies in transitional regimes. Pickering (1983) took their terminology a step further by recognizing three transitional regions between the major fan environments, and between fan and non-fan, which had not been documented previously. Pickering states that it has been tacitly assumed that because the transition regions possess mixed attributes of adjacent "defined" environments they require no explanation. The nomenclature for transitional environments of the above previous workers were used in this thesis.

The characteristics of deep-water sedimentation associated with volcanic islands are not well understood. Walker and Mutti (1973) admit that when interpreting ancient turbidite sequences in light of modern examples, biases tend toward submarine fans at stable continental margins. "We know little or nothing about turbidite types, or even bed thicknesses, in oceanic trenches and interarc and marginal basins" (Walker and Mutti, 1973, p. 134). This may be extended to include turbidites associated with volcanic islands in any tectonic setting. Also, Pickering (1982c, p. 44) states that "unfortunately, the submarine fan facies models and terminology are being overemphasized in active margin successions where fan morphology is generally unproven".

Depositional settings around volcanic island chains typically include narrow shelves and steep slopes (Dickinson, 1982). An absence of submarine canyons, channels and shelf deposits, may also be characteristic of volcanic island settings (Cas _et al._, 1981). Non-existent or very narrow shelves, with a maximum width of 9 km, have been observed for modern volcanic settings (Hayes, 1964, p. 69) and envisioned for Archean to Miocene ancient arc settings (Thornbury, 1969, p. 479; Ojakangas, 1972; Hayward, 1976; Hyde, 1980; Cas _et al._, 1981).

Without evidence for a style of sedimentation which differs from that of classical models for turbidite fan depositional systems, the interpretation of turbidite deposits associated with a volcanic source is limited to the model proposed (Fig. 3.27).

# 4. PETROGRAPHY

## Purpose

Petrographic analysis of sandstone composition may indicate provenance which may ultimately be useful in identifying tectonic setting in basin analysis. The main objective of this chapter is to determine tectonic setting of the lower Conception Group sandstones, if possible.

## Method

Eighty-six thin sections, cut perpendicular to bedding, were prepared and examined from samples taken from each of the seven outcrops studied (see Appendix I for stratigraphic sample locations). Point counting was performed on 20 of the 86 thin sections. Point counting is a non-biased method of determining the percentage composition of a rocks constituent minerals or grains. The 20 thin sections studied are believed to accurately represent the composition of the entire lower Conception Group because thin section examination revealed no systematic compositional variations in the Mall Bay or Drook Formations. To effectively distinguish between quartz and feldspar, and between feldspar types, thin sections were stained using the method of Quinn (1984)(see Appendix III) who substantially modified the versions advocated by Friedman in Carver (1971), Norman (1974) and Houghton (1980).

Point counting methods, criteria for distinguishing components, and problems in mentally removing diagenetic alterations to reconstruct original compositions, are discussed thoroughly by Dickinson (1970), Graham <u>et al.</u> (1976), Ingersoll (1978) and Ingersoll and Suczek (1979).

Only medium to coarse grained sandstones (modal grain size 0.3 to 1.0 mm) of Facies E and F were used for point counting. The composition of these sandstone grain sizes is thought to be representative of all lower Conception Group sediments. Grain identification is unreliable in fine grained sandstones or siltstones. Based on 250 counts, error is restricted to 5% or less (Van der Plas and Tobi, 1965).

## Results

The sandstones of the lower Conception Group are poorly sorted according to the classification of Folk (1974, p. 104) and may contain high proportion (but less than 25%) of matrix (30 micron maximum size limit)(Fig. 4.1 a,b). Many of the grains are extraordinarily well rounded using the classification of Powers (1953)(Fig. 4.1c). Tuffaceous samples are rare, are highly altered, and contain glass and other vitric material, producing a perlitic texture (Fig. 4.1d). In order of decreasing relative abundance, the minerals and constituents found are: volcanic rock fragments (L), plagioclase feldspar

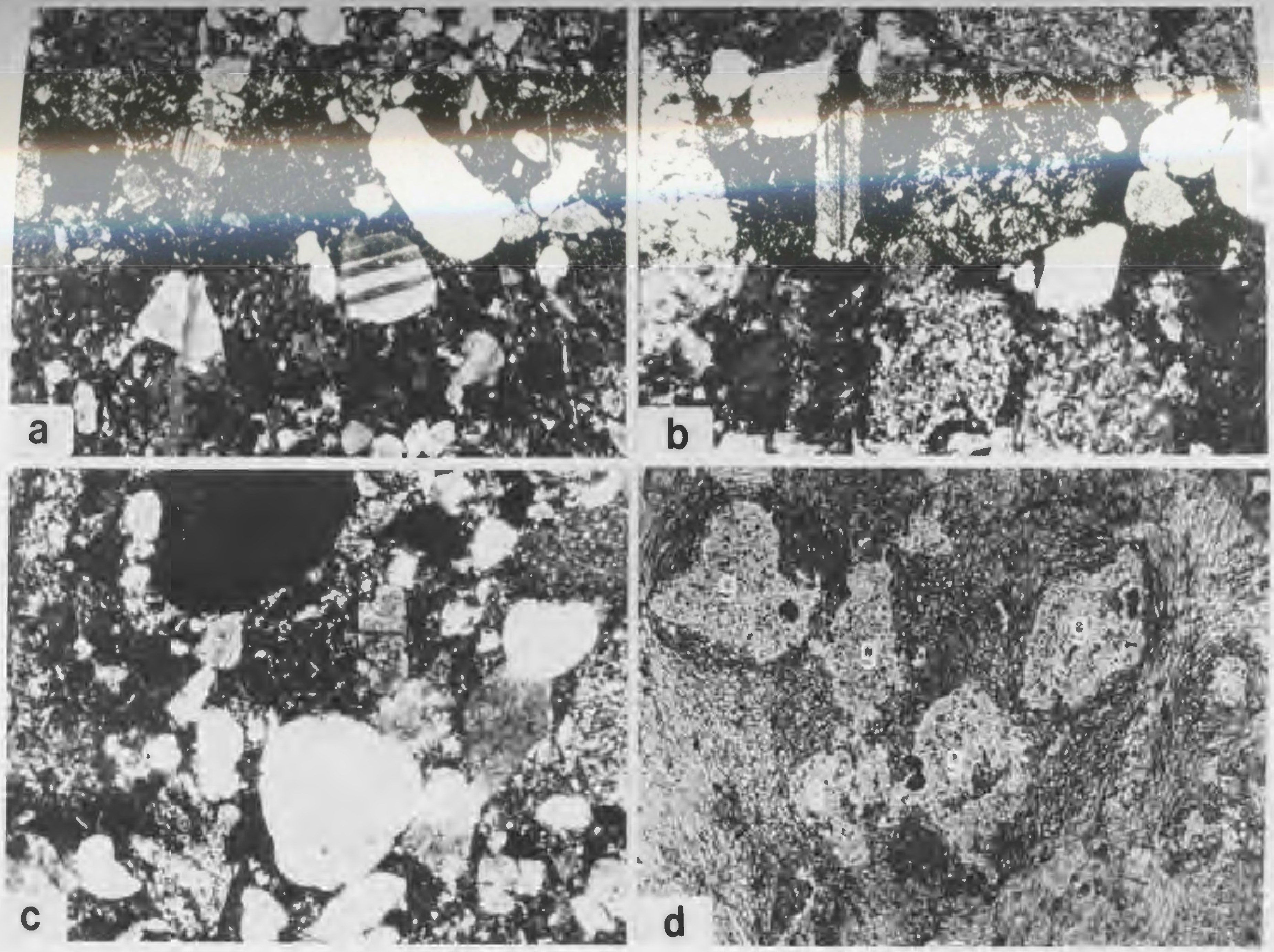
Figure 4.1: General petrographic characteristics of lower Conception Group sandstones

a. Poorly sorted volcaniclastic sandstone characteristic of the lower Conception Group. Note the difficulty in distinguishing between volcanic rock fragments from matrix. Sample HP-26, mag. 25X, crossed nicols.

b. Same as a) above. Note the abundance of volcanic rock fragments and twinned plagioclase. The light grains are untwinned plagioclase. Sample from the sedimentary slide at St. Stephens (section 2). Mag. 25X, crossed nicols.

c. Well rounded untwinned plagioclase and volcanic rock fragments. Sample from the sedimentary slide at St. Stephens (section 2). Mag. 25X, crossed nicols.

d. Tuffaceous sandstone containing intact glass fragments (g) surrounded by a fragmented matrix exhibiting perlitic texture. Sample FC-6. mag. 25X, plane polarised light.


a
b
c
d

(P), matrix, palagonite (or pumpellyite), quartz, carbonate and epidote. Many samples have approximately equal amounts of plagioclase and volcanic rock fragments. Vitric fragments, including devitrified shards, were counted as volcanic rock fragments.

Table 4.1 indicates point count results. Excluding samples that have undergone extensive diagenetic replacement of matrix and rock fragments by calcite, the sandstones of the lower Conception Group conform with the definitions of lithic arkose and feldspathic litharenite proposed by Folk (1974).

<u>Volcanic rock fragments</u>

Volcanic fragments are the most abundant detrital framework grain, comprising about 40 to 50% of all greywackes examined (Table 4.1). Seven types of volcanic fragments were recognized, as described below. Table 4.2 summarizes their characteristics and interpretation.

1) Coarse grained lathwork of plagioclase exhibiting intersertal or intergranular texture (Dickinson, 1970)(Fig. 4.2a)
2) Porphyritic plagioclase grains (Fig. 4.2b)
3) Vitric and vitrophyric grains, now composed of devitrified shards composed of feathery sheaves of chlorite, in a matrix of microcrystalline aggregate (Fig. 4.2c)
4) Banded rhyolite grains with some quartz (qtz) and

Table 4.1: Point count results, per cent minerals

| SAMPLE # | $L_v$ VOLCANIC ROCK FRAGMENTS | F PLAGIOCLASE | MATRIX | Q QUARTZ $Q_m$ | $Q_p$ | $Q_s$ | CARBONATE | EPIDOTE | HEAVIES |
|---|---|---|---|---|---|---|---|---|---|
| HP-3 | 30.4 | 39.6 | 20.0 | 6.0 | 0.0 | 0.0 | 0.0 | 5.6 | 0.0 |
| HP-4 | 26.0 | 36.8 | 15.6 | 1.6 | 0.4 | 1.2 | 0.0 | 2.0 | 0.0 |
| HP-12 | 19.6 | 49.2 | 16.8 | 4.4 | 0.8 | 0.8 | 4.4 | 3.2 | 0.0 |
| HP-13 | 34.4 | 36.8 | 20.4 | 4.0 | 0.4 | 0.0 | 0.0 | 3.2 | 0.0 |
| HP-15 | 17.6 | 53.2 | 16.4 | 4.8 | 0.0 | 0.4 | 0.8 | 6.8 | 0.0 |
| HP-18 | 40.0 | 27.6 | 21.6 | 4.4 | 0.0 | 1.6 | 0.0 | 3.6 | 0.0 |
| HP-19 | 32.8 | 38.0 | 16.4 | 4.8 | 1.2 | 0.4 | 0.0 | 6.0 | 0.0 |
| HP-20 | 47.6 | 31.2 | 14.8 | 3.2 | 0.0 | 0.0 | 0.4 | 2.8 | 0.0 |
| HP-22 | 46.8 | 29.6 | 14.4 | 5.2 | 0.0 | 0.8 | 0.4 | 2.8 | 0.0 |
| HP-25 | 28.8 | 35.6 | 18.8 | 14.0 | 0.4 | 0.4 | 0.0 | 2.4 | 0.0 |
| HP-26 | 36.8 | 28.0 | 18.8 | 7.2 | 0.4 | 0.8 | 1.2 | 5.6 | 0.0 |

n=250 point counts per thin section

(......continued)

Table 4.1: (continued)

| SAMPLE # | $L_v$ VOLCANIC ROCK FRAGMENTS | F PLAGIOCLASE | MATRIX | Q QUARTZ | | | CARBONATE | EPIDOTE | HEAVIES |
|---|---|---|---|---|---|---|---|---|---|
| | | | | $Q_m$ | $Q_p$ | $Q_s$ | | | |
| CE-14 | 55.2 | 22.8 | 9.6 | 6.0 | 0.4 | 0.0 | 2.8 | 0.0 | 3.2 |
| CE-17 | 60.4 | 25.6 | 1.6 | 9.2 | 0.8 | 0.4 | 2.0 | 0.0 | 0.0 |
| CE-18 | 40.8 | 40.8 | 10.4 | 0.4 | 0.0 | 0.4 | 6.0 | 0.4 | 0.8 |
| PH-2 | 19.2 | 53.2 | 14.8 | 5.2 | 0.8 | 0.0 | 3.2 | 0.0 | 3.6 |
| PH-4 | 28.0 | 44,0 | 7.6 | 16.0 | 0.8 | 0.0 | 0.8 | 0.0 | 2.8 |
| PH-7 | 21.2 | 46.8 | 18.0 | 9.6 | 0.4 | 0.0 | 0.0 | 0.0 | 4.0 |
| PH-18 | 41.2 | 33.6 | 15.2 | 1.6 | 0.4 | 0.0 | 4.0 | 0.0 | 4.0 |
| SS-1 | 46.0 | 34.8 | 0.8 | 0.8 | 0.8 | 1.6 | 1.2 | 1.2 | 1.6 |

n=250 point counts per thin section

Table 4.2: Types of volcanic rock fragments and their lithologic interpretation

| TYPE | DESCRIPTION | FIGURE REFERENCE | INTERPRETATION |
|---|---|---|---|
| 1) Lathwork grains | -plagioclase laths in intergranular and intersertal textures (Dickinson, 1970) | 4,2a | basalt |
| 2) Porphyritic plagioclase grains | - | 4.2b | basalt |
| 3) Vitric grains | -devitrified shards and glass fragments in a matrix of microcrystalline aggregate and palagonite | 4.2c | rhyolite |
| 4) Banded rhyolite porphyritic | -porphyritic grains containing quartz and plagioclase phenocrysts, some banding | 4.2d | rhyolite |
| 5) Microlitic grains | - | 4.2e | basalt |
| 6) Quartz-porphyritic grains | - | 4.2f | dacite-rhyolite |
| 7) Granophyric grains | -excellent development of graphic texture | 4.2g | recrystallised ignimbrite (V.S. Papezik, pers. comm.) or high level intrusion |
| 8) Spherulitic grains | -quartz and feldspar spherules formed by devitrification of felsic, glassy flows | 4.2h | rhyolite |

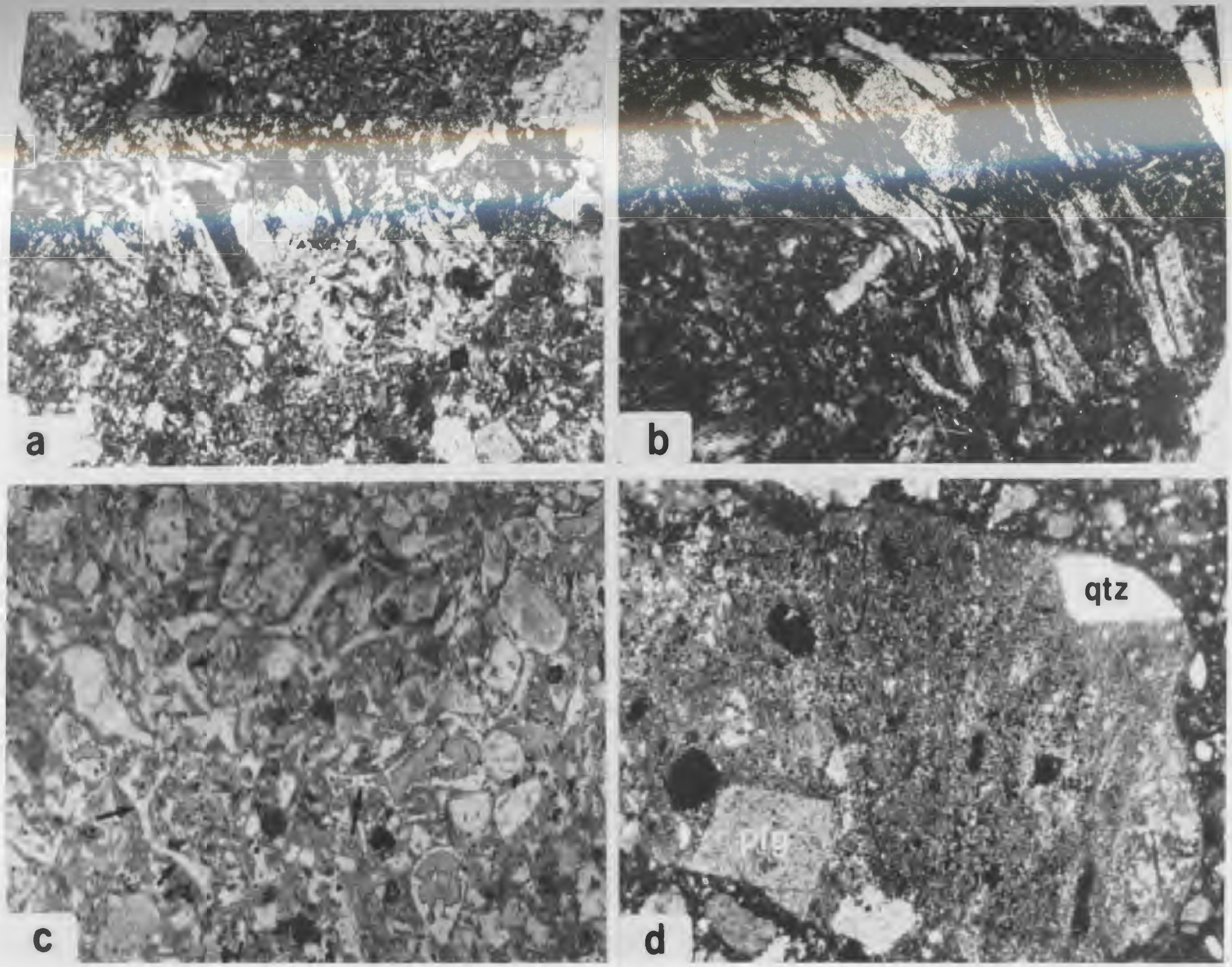
Figure 4.2: Types of volcanic rock fragments

a. Well rounded grain containing coarse grained plagioclase lathwork exhibiting intersertal texture. Grains of this type are classified as volcanic rock fragments when point counting. Sample HP-4, mag. 25X, crossed nicols.

b. Porphyritic plagioclase grains. Sample HP-4, mag. 100X, crossed nicols.

c. Vitric grains, now composed of devitrified glass shards (see arrows) in a matrix of palagonite (dark areas). Sample HP-21, mag. 25X, plane polarised light.

d. Porphyritic grains containing quartz (qtz) and plagioclase (plag) phenocrysts. Sample HP-3, mag. 100X, crossed nicols.


a
b
c
d
qtz
Plg

plagioclase (plag) phenocrysts (Fig. 4.2d)

5) microlitic grains (Fig. 4.2e)

6) Quartz-porphyritic grains (Fig. 4.2f)

7) Granophyric grains(Fig. 4.2g)

8) Spherulitic grains (Fig. 4.2h)

Interpretations are deduced from mineralogy, phenocryst types, grain size, and texture. Volcanic fragment types 1 to 3 are probably indicative of basaltic flows (Dickinson, 1970). An intermediate lava composition is reflected in type 5, while types 4,6,7 and 8 may represent flows of felsic composition. The granophyric texture (type 7) is common in recrystallized ignimbrites of the Harbour Main Group and in high level intrusives of the Holyrood Plutonic Complex (V.S. Papezik, pers. comm.). The granophyre is not thought to be the result of devitrification as described by Lofgren (1971). The proportions of fragments which are interpreted to be of basaltic and rhyolitic composition is in agreement with the observed composition of Harbour Main volcanic rocks described by Papezik (1970,1972), Nixon and Papezik (1979), and Cameron and Papezik (1984).

The degree of roundness is variable from subangular to well rounded, with most grains being rounded according to the classification of Powere (1953). Grains are either floating in matrix, or have tangential contacts. Volcanic rock fragments form the coarsest grain size, varying from

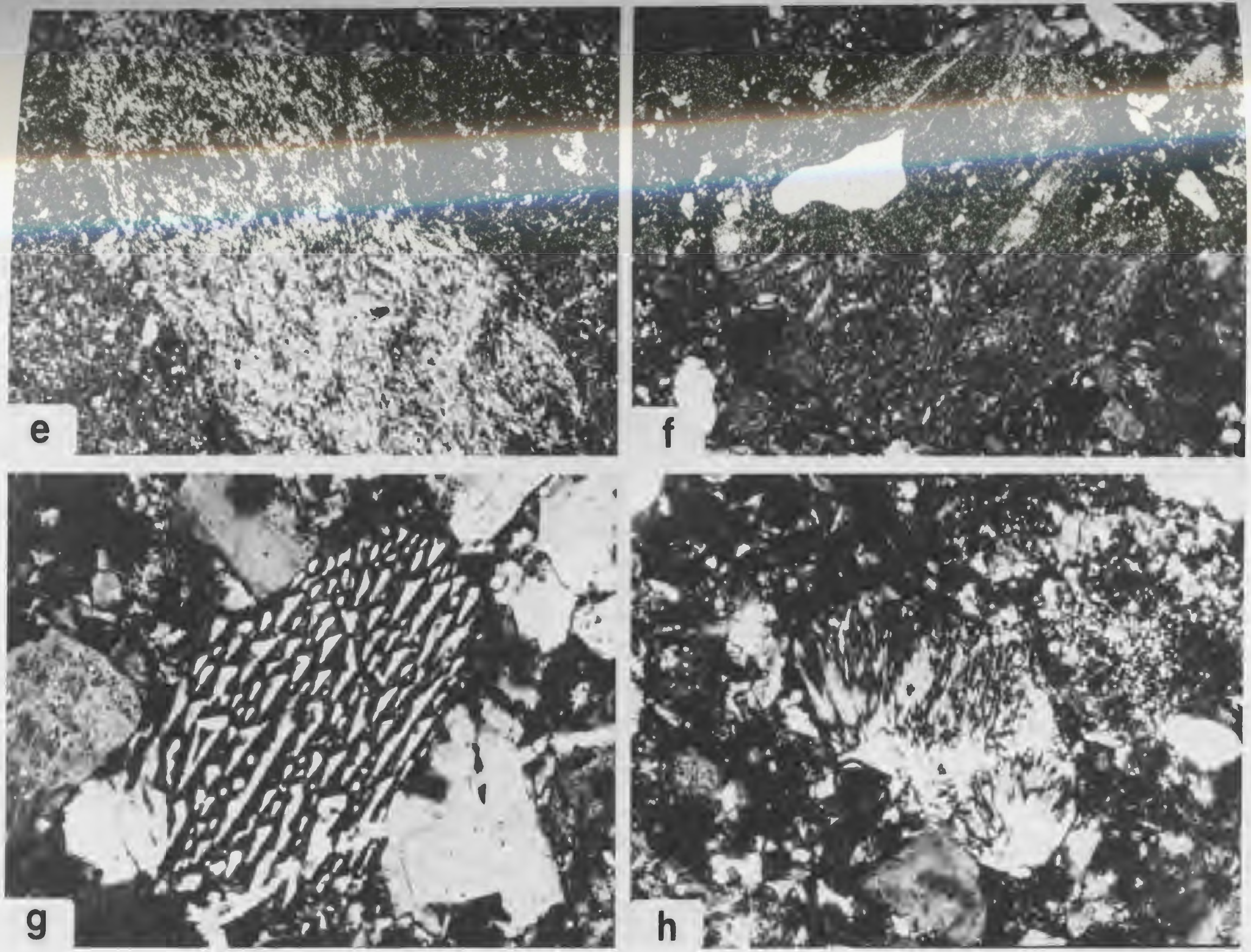
Figure 4.2: Types of volcanic rock fragments (continued)

e. Microlitic grains, sample HP-4, mag. 25X, crossed nicols.

f. Quartz-porphyritic grain, sample HP-4, mag. 25X, crossed nicols.

g. Well developed granophyric texture. Sample PH-4, mag. 100X, crossed nicols.

h. Spherulitic texture. Sampel HP-3, mag. 100X, crossed nicols.


e
f
g
h

2ø (0.25 mm) to larger than -2ø (4.0 mm).

Feldspar

Thin sections were stained with sodium cobaltinitrite and amaranth (red dye no. 2)(Appendix III) to identify potassium feldspar and plagioclase, respectively. No potassium feldspar was observed in the lower Conception Group. Plagioclase comprises an average of 37% of framework grains, with up to 53% present in two samples.

Three types of plagioclase grains are recognized. Two of these, representing end members, are: 1) untwinned, calcitized or sericitized, subangular to subrounded, lath-shaped grains up to 2 mm in length (Fig. 4.3a) which may contain acicular mineral inclusions, and 2) pristine, polysynthetically twinned (albite twinning), subangular, grains from 1ø (0.5 mm) to 3ø (0.125 mm) in size (Fig. 4.3b). Transitional grains, exhibiting characteristics of both types of plagioclase, are present. Most plagioclase observed is of type 1. A third grain type exhibits Carlsbad twinning and commonly contains calcite inclusions (Fig. 4.3c). A rare grain type, very similar to perthite, was observed (Fig. 4.3d). No zoned plagioclase crystals were observed. The Ca and Na content of the plagioclase was determined quantitatively for six samples using the electron microprobe (EMP). All plagioclase grains analysed were Na-rich, with little calcium and no potassium present.

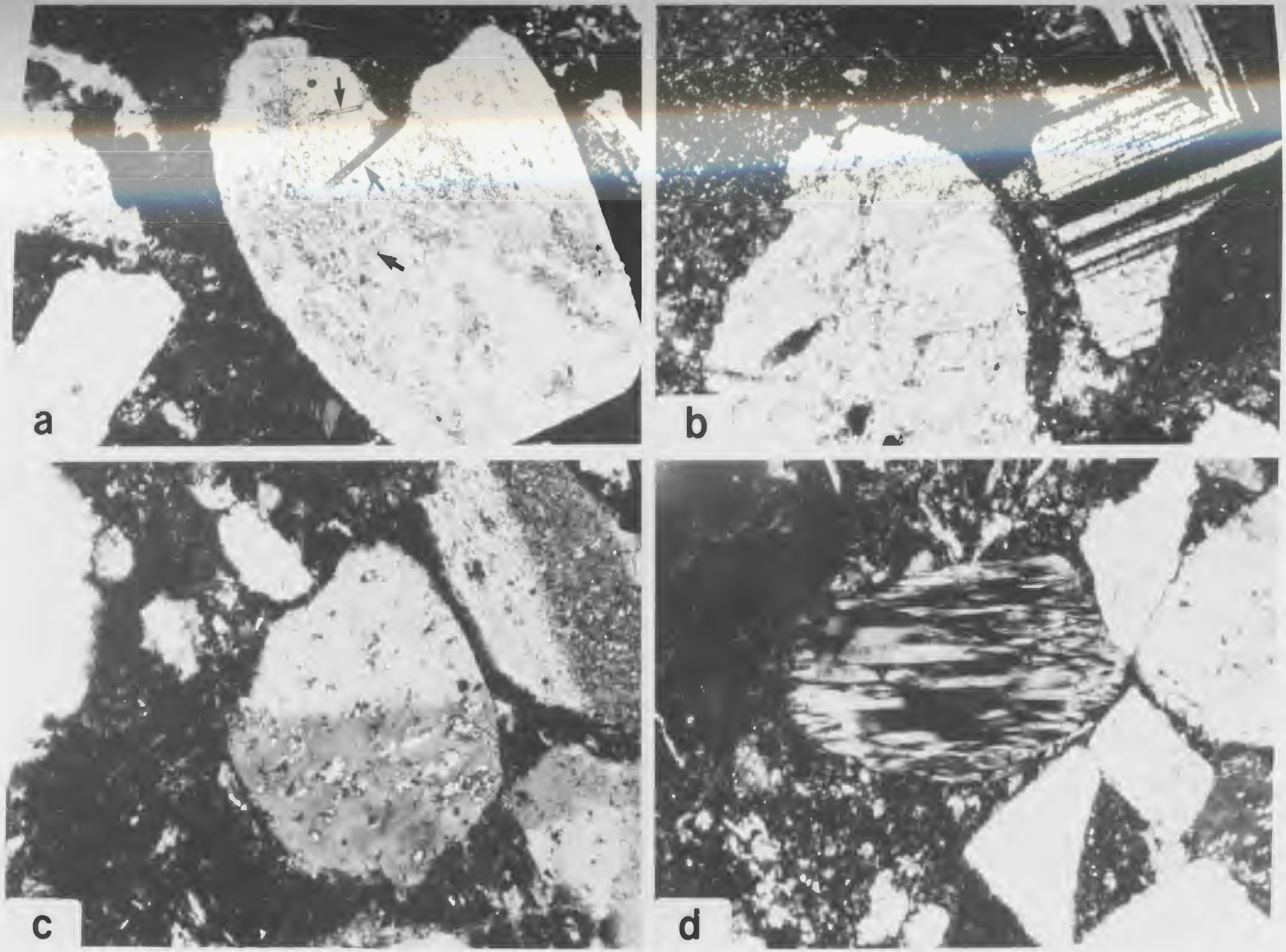
Figure 4.3: Feldspar grain types

a. Untwinned, seritized plagioclase grain which contains acicular mineral inclusions (see arrows). Sample HP-26, mag. 25X, crossed nicols.

b. Plagioclase grain exhibiting polysynthetic twinning (on right) with untwinned plagioclase grain (on left). Sample CE-17, mag. 25X, crossed nicols.

c. Plagioclase grain exhibiting Carlsbad twinning and high birefringent blebs due to calcitization or seritization. Sample HP-26, mag. 25X, crossed nicols.

d. Rare, unidentified grain, very similar to perthite. Sample HP-26, mag. 25X, crossed nicols.


a
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The feldspar composition of the six sandstone samples which were analysed is thought to be representative of the entire lower Conception Group.

Matrix

Matrix includes all material 30 microns or less in size (Dott, 1964). Not all of the matrix is primary. The formation of secondary matrix, by diagenetic alteration of framework grains, is possible (Cummins, 1962). The difficulty encountered when distinguishing between primary and secondary matrix is known as the "matrix question" or the "matrix problem" (ibid.). Grain boundaries were mentally reconstructed to outline primary grains, based on obvious textures. When material could not be identified as a primary grain or an alteration product, such as palagonite or calcite, it was counted as matrix. Constituent grains comprising the matrix produce a green or grey colour. Their mineralogy could not be positively identified, but probably includes chlorite, albite, sericite and clay minerals as in previous studies (Blatt, 1982, p. 229).

Palagonite (or pumpellyite)

It is difficult to distinguish palagonite from pumpellyite because the minerals are similar to one another and are very likely to occur in volcaniclastic rocks. Therefore, distinguishing between these two possibilities, based only on petrography, proved impossible. In general, palagonite is present in amounts up to 30% but comprises almost 50% of some samples, and was classified as matrix when point counting. This green, amorphous, mineraloid probably gives these sandstones their distinctive colour. It is typically found as a rim or zone within and around glass fragments in a massive form (Fig. 4.2c).

Quartz

Quartz grains are typically monocrystalline, angular to well rounded (Fig. 4.4a), with less than 5° of undulosity. Less than 1% of the quartz grains are polycrystalline. Most quartz varieties are clear and transparent, although some contain mineral, bubble, or dust inclusions. Some inclusions are highly birefringent minerals (mica?)(Fig. 4.4b), twinned plagioclase, or acicular shapes (rutile?). Boehm lamellae are present (Fig. 4.4c) which indicates intense strain deformation in the source area (Scholle, 1979, p. 9). Syntaxial overgrowths of silica cement are rare and are usually poorly developed, but may be well developed (Fig. 4.4d). The overgrowths are not well rounded and therefore probably formed *in situ*, not by sandstone recycling. Quartz grains rarely occur in mutual contact. Most contacts with surrounding lithic fragments are long or tangential, never sutured. In calcite-cemented sandstones from Point La Haye, quartz grains are corroded or partially replaced by calcite.

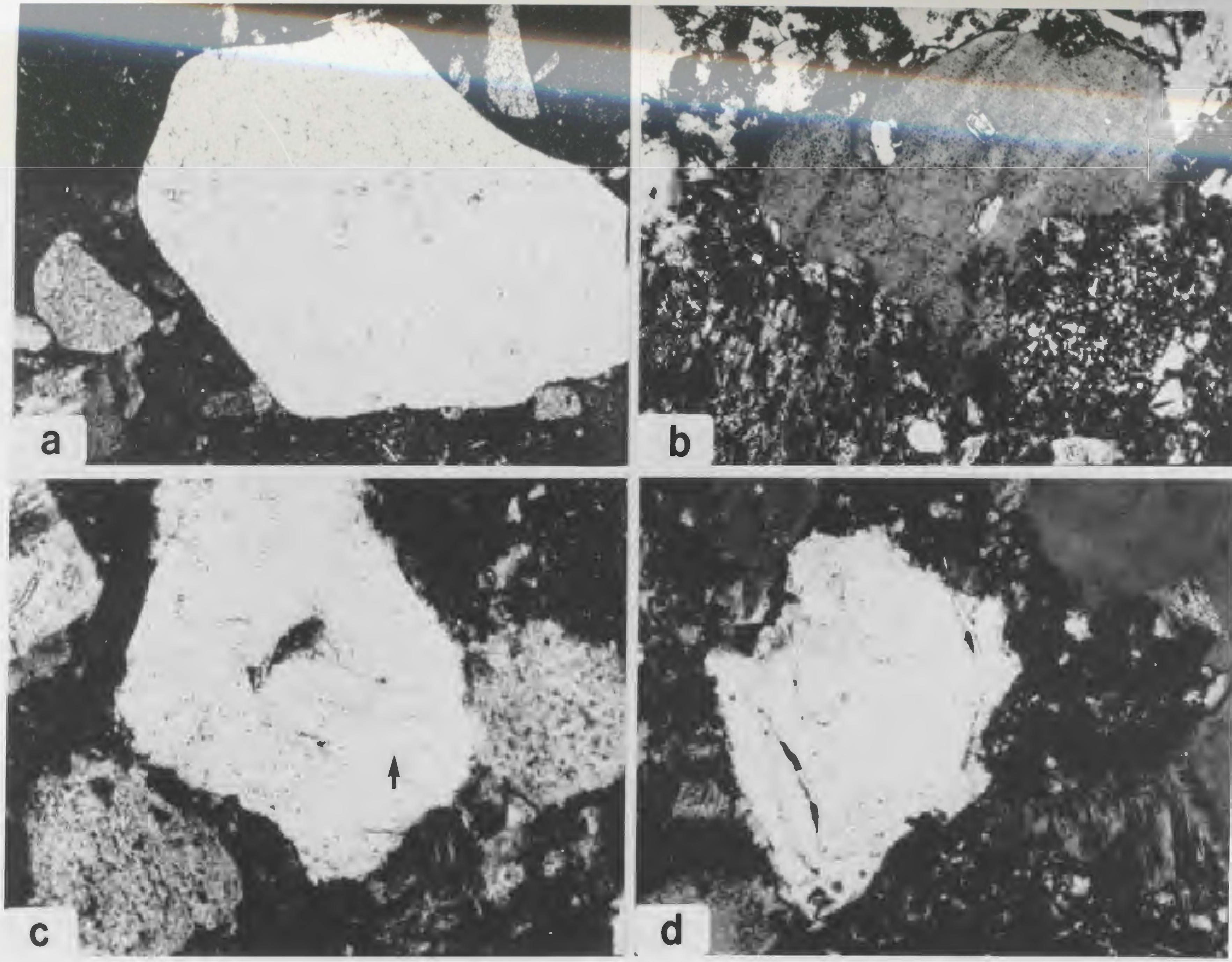
Figure 4.4: Quartz grains

a. Extremely well rounded quartz grain in poorly sorted sample taken from the sedimentary slide at St. Stephens (section 2). Mag. 25X, crossed nicols.

b. Mineral and dust inclusions in quartz grain. Sample HP-3, mag. 25X, crossed nicols.

c. Boehm lamellae (see arrow) in quartz grain indicating intense strain deformation in the source area (Scholle, 1979, p. 9). Sample HP-19, mag. 25X, crossed nicols.

d. Syntaxial overgrowths of silica cement on quartz grains is rare. Sample HP-3, mag. 25X, crossed nicols.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Three types of quartz grains that indicate their original provenance are i) elongate, polycrystalline grains of metamorphic origin which exhibit internal sutured boundaries (Fig. 4.5a), ii) quartz grains with embayed margins (Fig. 4.5b) indicative of a volcanic origin (Potter _et al._, 1972, p. 264-265), and iii) polycrystalline grains with some interstitial matrix and varying extinction angles indicating a sedimentary source (Fig. 4.5 c,d). Distinguishing between types i and iii (metamorphic vs. sedimentary clasts) is subjective. The presence of sutured intergranular contacts probably indicates a metamorphic origin (Blatt, 1982, p. 153-154). Provenance determination, using the method of Basu _et al._ (1975), was not attempted because of an insufficient number of quartz grains.

Monocrystalline quartz grain size is variable from 2Ø (0.25 mm) to 0Ø (1.0 mm). In general, quartz grains occur in the coarsest fraction of the grain population. Polycrystalline quartz varieties are relatively coarse, ranging from 0Ø (1.0 mm) to -2Ø (4.0 mm).

<u>Carbonate</u>

Carbonate rock fragments were not observed, and most carbonate appears to be of secondary origin. Although rare overall (4%), carbonate occurs largely as a replacement of plagioclase grains, mimicking twin lamellae (Fig. 4.6a), as isolated blebs in untwinned plagioclase (Fig. 4.3a),

Figure 4.5: Quartz grains and polycrystalline rock fragments

a. Elongate, polycrystalline grains of metamorphic origin which exhibit internal sutured boundaries. Sample HP-19, mag. 25X, crossed nicols.

b. Quartz grain with embayed margin characteristic of volcanogenic quartz crystals. Sample HP-19, mag. 25X, crossed nicols.

c. Polycrystalline, coarse grained sedimentary rock fragment. Some internal boundaries are slightly sutured but most are long contacts. Sample HP-19, mag. 25X, crossed nicols.

d. Polycrystalline, fine grained sedimentary rock fragment, similar to c) above. Sample from the sedimentary slide at St. Stephens (section 2). Mag. 25X, crossed nicols.

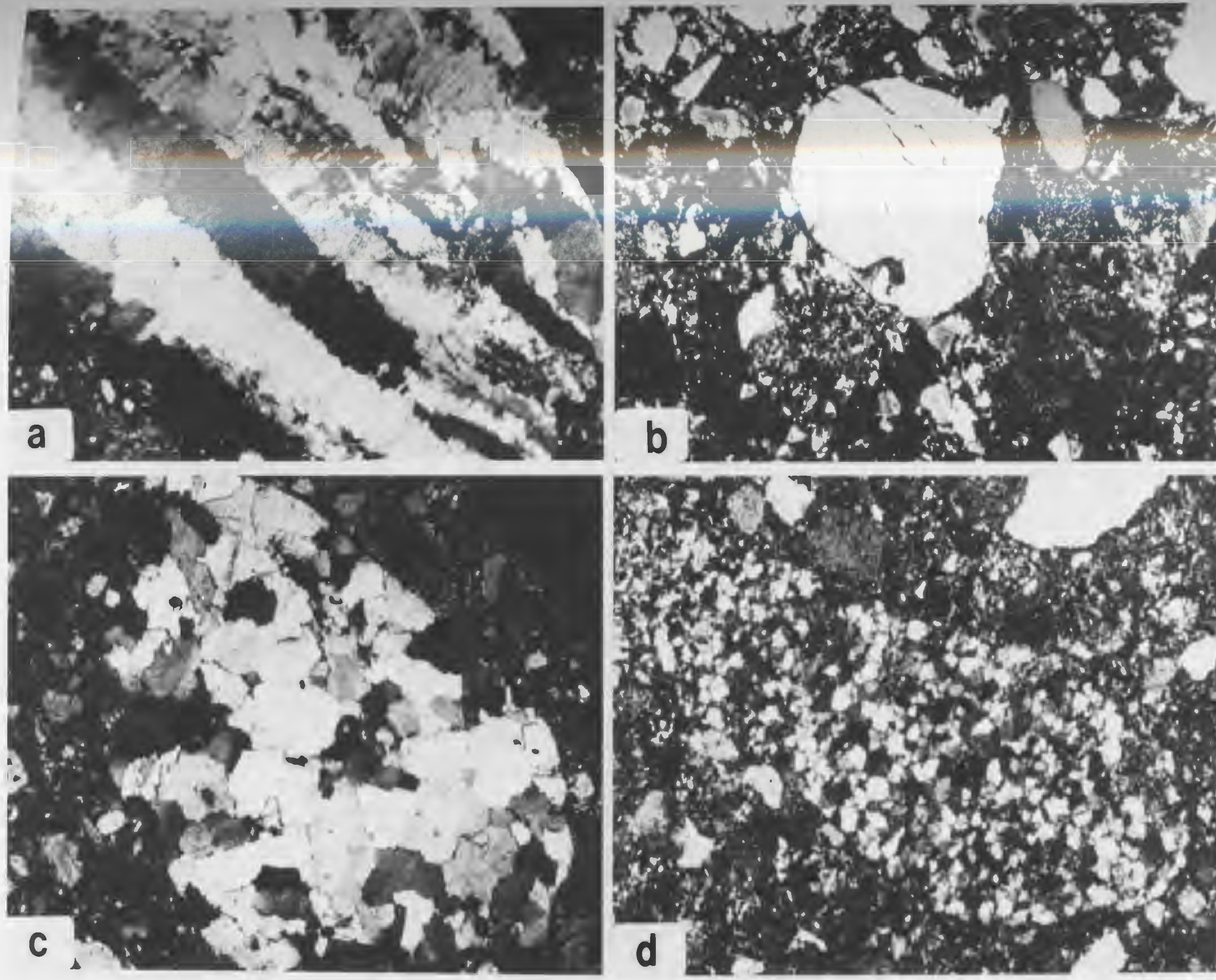
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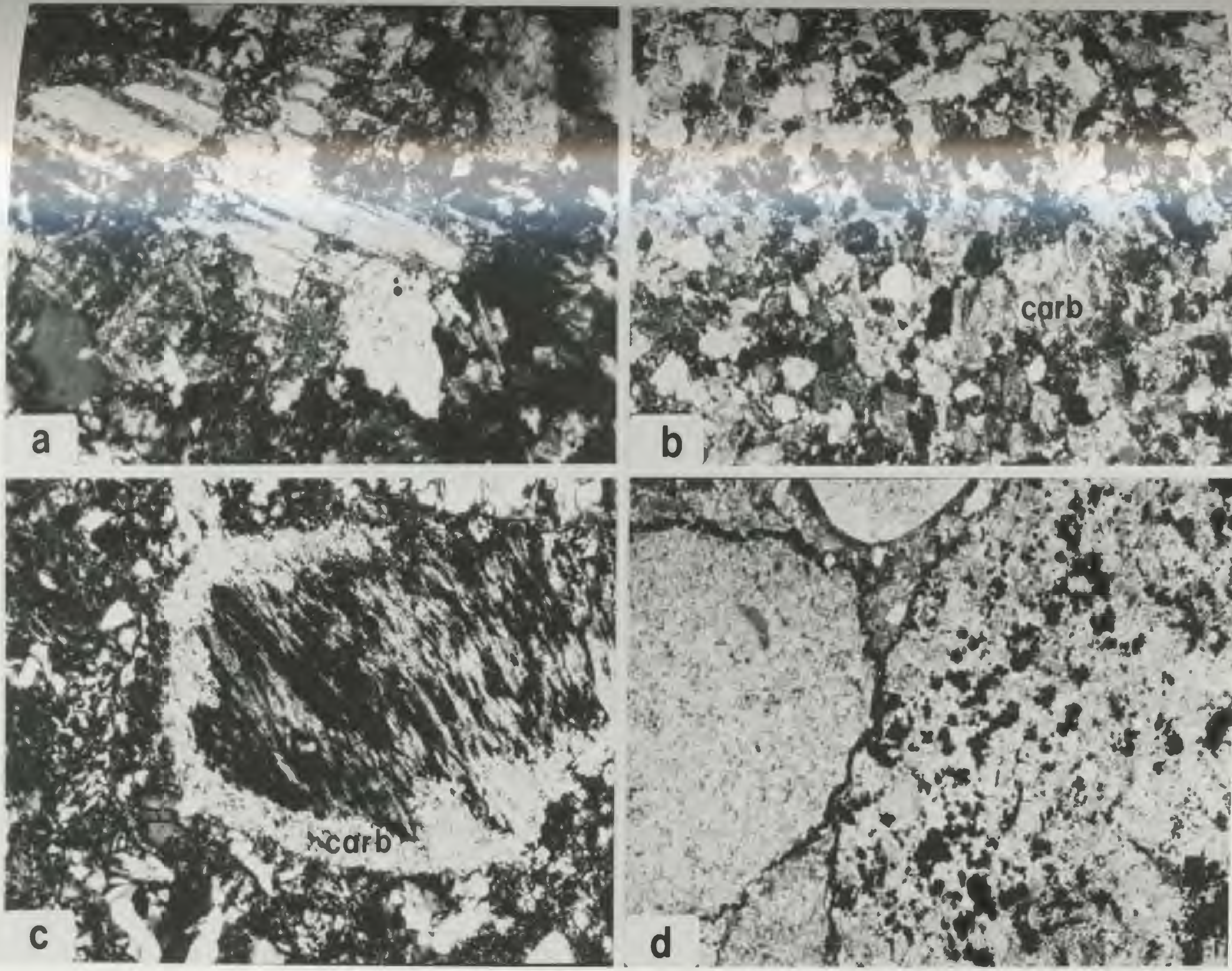
Figure 4.6: Carbonate and accessory minerals

a. Carbonate replacing plagioclase grain by mimicking twin lamellae. Sample HP-26, mag. 100X, crossed nicols.

b. Carbonate (carb) cemented sandstone. Sample PH-1, mag. 25X, crossed nicols.

c. Carbonate (carb) envelope due to replacement of the perimeter of a volcanic rock fragment. Sample CE-17, mag. 25X, crossed nicols.

d. Opaques occur around the perimeter of volcanic rock fragments. Sample HP-3, mag. 25X, plane polarised light.


a
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carb
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carb
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and as a cement in thinly bedded sandstones of Facies A (Fig. 4.6b) at Point La Haye. Calcite also occurs as envelopes around grains (Fig. 4.6c). Within concretions, carbonate also occurs in circular patterns resembling infilled fenestrae.

Epidote

The distinctive green colour, high relief, and low order birefringence, characterizes epidote. It usually occurs in and around altered glass shards (Fig. 4.2c) and imparts its green colour to these sandstones.

Minor constituents and other characteristics

Additional minerals, such as sphene, zircon, and opaques are present in trace amounts. Although extremely rare, zircon occurs in characteristic high relief grains, about 1.0 mm (0Ø) in size, with high birefringent colours. Opaques could not be identified as specific minerals.

## Interpretation of Petrography

Because volcaniclastic sandstones are usually deposited in zones of high heat flow and then subjected to circulation of groundwaters rich in alkalis and alkaline earth elements, they may suffer intense diagenesis at depths of only a few thousand metres (Blatt, 1982, p. 227). During diagenesis, several types of alteration occur, the most common being the formation of matrix composed of chlorite, sericite and minor quartz.

Boundaries between recognizable detrital grains and irresolvable clay paste are indistinct, probably reflecting the mineralogical transition between the two materials. The occurrence of calcite blebs in plagioclase may indicate the replacement of plagioclase which produces a clay matrix byproduct. This is a common reaction in volcaniclastic sandstones (Blatt, 1982, p. 229). Calcite has also replaced volcanic rock fragments, to produce the carbonate envelope around the grains. The unidentified irregular masses and aggregates may be also be alteration products of volcanic fragments (Fig. 4.6d).

The absence of potassium feldspar may reflect either: 1) an absence of potassium feldspar in the original source or 2) alteration of original potassium feldspar due to a reaction with seawater (Moore, 1979). Cowan (1974) suggested that the low potassium feldspar content of Franciscan greywackes in California could be due to alteration. Because the volcanic rock types 4,6,7 and 8 probably indicate a rhyolitic source, they should contain potassium feldspar. Its absence is likely due to alteration. Although the Na-rich plagioclase composition may be primary, diagenetic processes may be responsible.

The presence of sedimentary and metasedimentary clasts implies the presence of an older quartzose sedimentary source. Potential sedimentary source rocks include equivalents of the Helikian(?) carbonate and clastic platformal sequences of the Green Head Group (Wardle, 1978) or George River Group that are found in Nova Scotia and New Brunswick.

The extraordinary rounding exhibited by most volcanic and plagioclase fragments and some quartz grains requires explanation. Some of the well rounded quartz grains are undoubtedly recycled based on the presence of contemporaneous sedimentary fragments. The plagioclase grains originate as laths, the corners of which are probably abraded relatively quickly, producing a rounded appearance. Because many of the volcanic fragments contain mafic minerals, they might tend to weather and abrade quickly, producing the characteristic rounding.

## Implications for tectonic setting

Attempts to relate ancient sandstone composition to known tectonic settings have been made by Dickinson (1970, 1971a, 1971b), Graham *et al.* (1976), Dickinson and Suczek (1979), and Dickinson *et al.* (1983). Similar studies on modern sands are described by Sibley and Pentony (1978), Dickinson and Valloni (1980), Valloni and Maynard (1981), Maynard *et al.* (1982) and Yerino and Maynard (1984). If sandstone composition accurately reflects the geology of the source terrane, then petrographic analysis of the volcaniclastic lower Conception Group should allow one to infer the provenance and perhaps gain insight into its

Precambrian tectonic setting. Sandstone petrography cannot be used to interpret paleotectonic setting without incorporating all the available sedimentological, geochemical, geophysical, and structural data.

The Precambrian tectonic setting of the Avalon Zone is controversial. Numerous tectonic models, ranging from subduction to ensialic rifting, have been proposed (see chapter 2). The previous models are largely based on major and trace element geochemistry of volcanic rocks and field relationships. Plotting the Conception Group data (Table 4.3) on the QFL and $Q_mFL_t$ diagrams of Dickinson and his co-workers, an attempt is made to compare the inferred paleotectonic setting reflected in the sandstone composition with the subduction and rifting tectonic settings proposed by previous workers. Where data are available, the volcanology and sandstone composition from volcanic source areas in various tectonic settings will be summarized before attempting a comparison with Conception Group sandstones.

There are four settings where abundant volcanic rocks occur (Garcia, 1978): 1) marginal and backarc basins (e.g. Sea of Japan), 2) sites of intraplate volcanism (e.g. Hawaiian Islands, Tahiti), 3) sites of rifting (e.g. the mid-Atlantic ridge), and 4) in arcs above subduction zones (e.g. Aleutian Islands, Alaska). Although much is known about the volcanology and geochemistry of each, less is known about sedimentary models and sandstone composition in some of these modern settings, particularly mid-ocean ridges and intraplate volcanoes. The deep-sea drilling project (DSDP) and the international program of ocean drilling (IPOD) have examined turbidites associated with linear island chains (Kelts and Arthur, 1981; Clague, 1981) and active convergent margins (Von Huene, 1981). However, of the modern settings, only convergent margin sandstones have been examined mineralogically (Maynard et al., 1982; Yerino and Maynard, 1984). This has led to biases in tectonic interpretations of ancient sandstone compositions in favour of convergent margins.

No petrographic data is available for sandstones derived from volcanoes in intraplate and mid-ocean ridge settings. The petrography of sandstones derived from volcanoes above modern and ancient subduction zones and in associated backarc basins has received much attention. The overall similarity between Conception Group sandstone petrology with these modern and ancient settings, warrants comment.

The petrography of sandstones associated with four ancient subduction zones is available; viz., the Great Valley Sequence (Dickinson, 1982; Ingersoll, 1983) and the Franciscan Complex (Gilbert, 1973; Cowan, 1974) in California; New Zealand greywackes (Dickinson, 1971b); the Appalachian Ouachita and Black Warrior Basins (Graham et al., 1975); and around Nias Island, Indonesia (Moore, 1979). The Great Valley Sequence and the New Zealand greywackes were derived from volcanic and plutonic centers (magmatic arcs) and associated sedimentary and low-grade metamorphic rocks, on Andean-type continental margins. Dominant detrital components are quartz, volcanic rock fragments, and feldspar. The Ouachita and Black Warrior Basin sandstones were derived from a cratonic suture zone which resulted from the closing of the Iapetus Ocean (Graham et al., 1979). In contrast to this study, the Ouachita sandstones are quartzose, largely deficient in feldspar or volcanic rock fragments, which reflects their cratonic provenance. Nias Island sandstones have low proportions of volcanic lithic grains and greater amounts of sedimentary and metasedimentary grains relative to the Franciscan and New Zealand sandstone assemblages.

On the QFL (Fig. 4.7a) and $Q_mFL_t$ (Fig. 4.7b) diagrams of Dickinson and Suczek (1979), lower Conception Group sandstones (Table 4.3) cluster in the quartz-poor regions of the Great Valley Sequence and New Zealand forearc assemblages, far removed from the fields for sandstones of Nias Island and Ouachita and Black Warrior Basins. On the general QFL (Fig. 4.8a) and $Q_mFL_t$ (Fig. 4.8b) diagrams for ancient volcaniclastic sandstones of Dickinson and Suczek (1979) and Dickinson et al. (1983), lower Conception Group sandstones plot in the "transitional arc" region.

Table 4.3: QFL and $Q_mFL_t$ values (see also Figures 4.7 and 4.8)

| SAMPLE # | Q | F | L | $Q_m$ | F | $L_t$ |
|---|---|---|---|---|---|---|
| HP-3 | 13.0 | 47.7 | 39.4 | 11.1 | 48.7 | 40.2 |
| HP-4 | 15.9 | 50.8 | 33.3 | 8.4 | 55.3 | 36.3 |
| HP-12 | 8.2 | 67.6 | 26.4 | 6.1 | 67.6 | 26.3 |
| HP-13 | 5.9 | 49.2 | 44.9 | 5.4 | 49.5 | 45.2 |
| HP-15 | 6.8 | 70.0 | 23.2 | 6.3 | 70.4 | 23.3 |
| HP-16 | 16.3 | 33.2 | 50.5 | 15.6 | 33.5 | 50.9 |
| HP-18 | 8.3 | 37.5 | 54.2 | 6.1 | 38.3 | 55.6 |
| HP-19 | 8.3 | 50.3 | 43.4 | 6.3 | 50.3 | 43.4 |
| HP-20 | 4.0 | 38.0 | 58.0 | 4.0 | 38.0 | 58.0 |
| HP-22 | 7.3 | 35.9 | 56.7 | 6.4 | 36.2 | 57.4 |
| HP-25 | 18.7 | 44.9 | 36.4 | 17.9 | 45.4 | 36.7 |
| HP-26 | 11.0 | 38.5 | 50.5 | 10.0 | 38.9 | 51.1 |
| CE-14 | 7.6 | 27.0 | 65.4 | 7.2 | 27.1 | 65.7 |
| CE-17 | 10.7 | 26.6 | 62.7 | 9.7 | 26.9 | 63.4 |
| CE-18 | 1.0 | 49.5 | 49.5 | 1.0 | 49.5 | 49.5 |
| PH-2 | 7.6 | 67.9 | 24.5 | 6.8 | 68.5 | 24.7 |
| PH-4 | 37.0 | 31.5 | 31.5 | 37.0 | 31.5 | 31.5 |
| PH-7 | 12.8 | 60.0 | 27.2 | 12.4 | 60.3 | 27.3 |
| PH-18 | 2.6 | 43.8 | 53.6 | 2.1 | 44.0 | 53.9 |
| SS-1 | 3.8 | 41.4 | 54.8 | 1.0 | 42.6 | 56.4 |

n=250 point counts per thin section

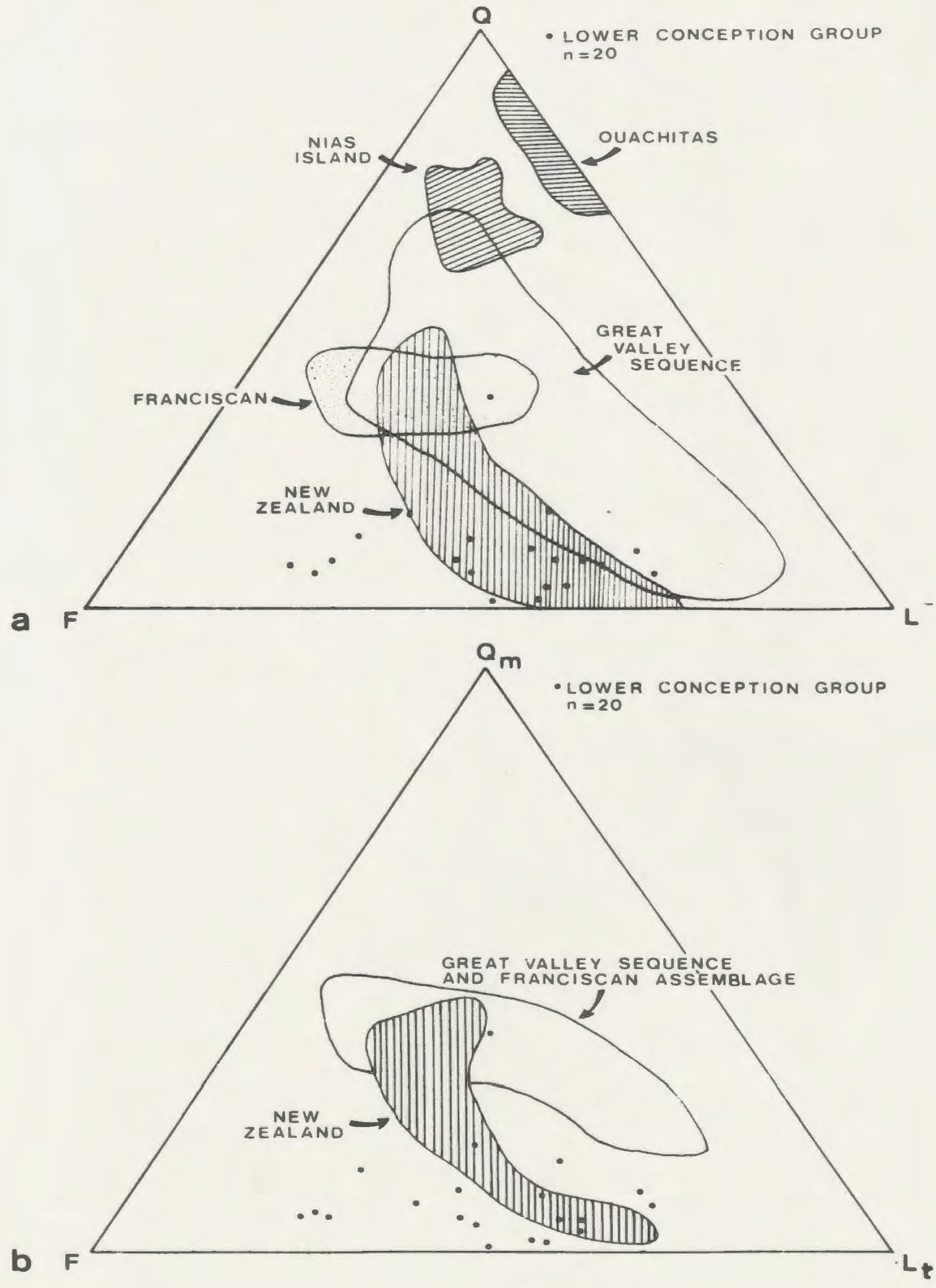


Figure 4.7: a. and b.
Using the QFL and $Q_mFL_t$ diagrams of Dickinson and Suzcek (1979), the lower Conception Group sandstones cluster in the quartz-poor region of the Great Valley Sequence and New Zealand Forearc assemblages.

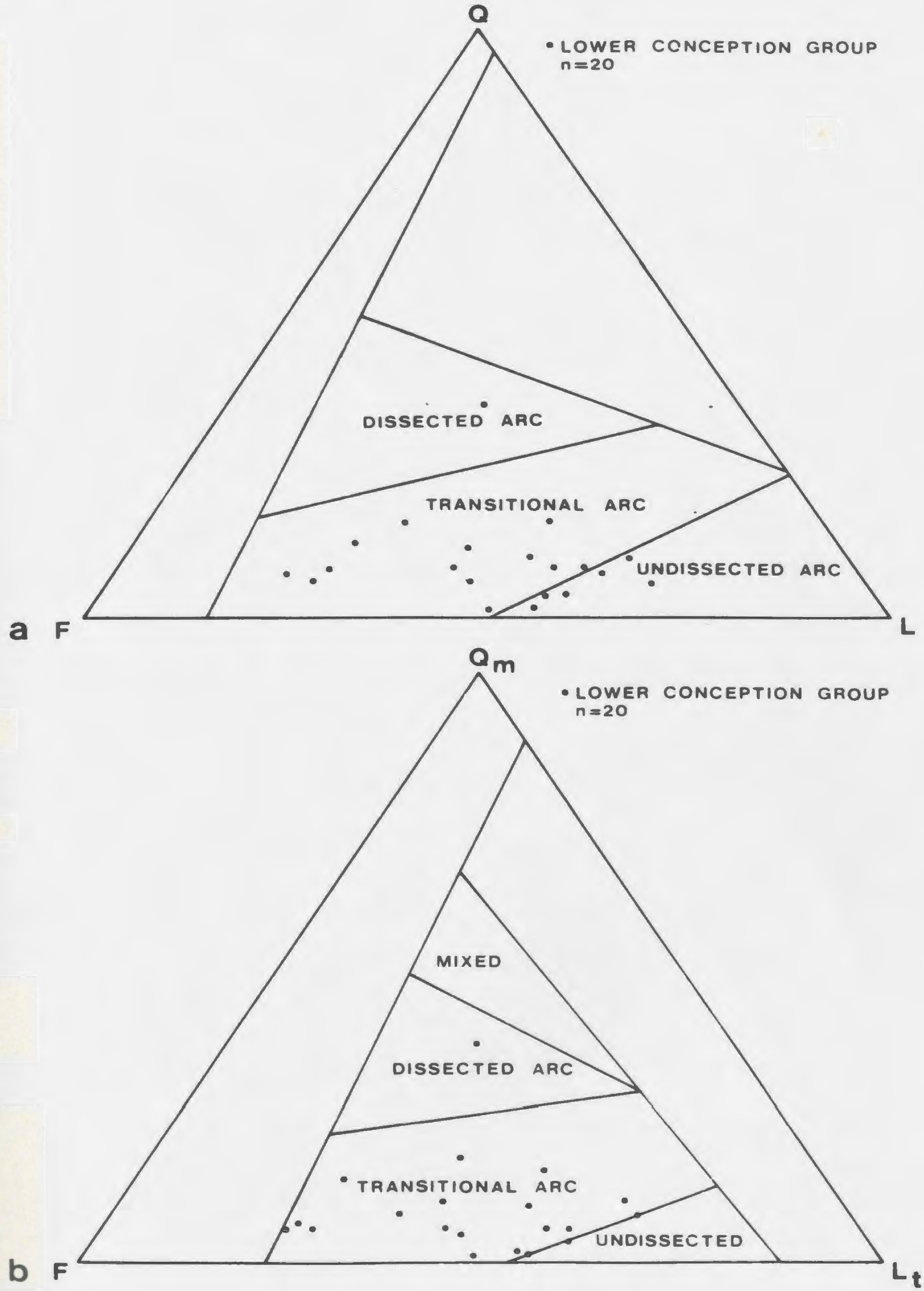


Figure 4.8: a. and b.
Using the general QFL and $Q_mFL_t$ diagrams of Dickinson and Suzcek (1979) and Dickinson et al. (1983), the lower Conception Group sandstones plot in the region of "transitional arcs".

Dickinson's term "transitional arc" is a term used to indicate a source area with partial erosion of volcanic rocks exposing underlying cogenetic batholiths. However, if the quartz grains plotted on this diagram are from continental basement or recycled sediments, his diagram may be misleading.

The plotted QFL data for modern sands in various tectonic settings (Maynard, Valloni and Yu, 1982; Yerino and Maynard, 1984) also produces distinct regions similar to the equivalent plots for ancient sandstones. The lower Conception Group sandstones plot in a region occupied by sands from backarc, forearc, and continental margin arc settings (Fig. 4.9).

The comparisons above indicate a striking similarity between detrital sand composition of the lower Conception Group and that found adjacent to both modern and ancient subduction zones. The main features that they have in common are the abundance of volcanic rock fragments (up to 50%), and plagioclase, with relatively little quartz.

Despite the petrographic similarities, there are several characteristic features of modern and ancient subduction zones which are not found in the lower Conception Group or associated volcanic and plutonic rocks. The absence of classical ophiolites, ophiolitic melanges, trench melanges, and extensive continental margin successions, combined with the lack of volcanics of

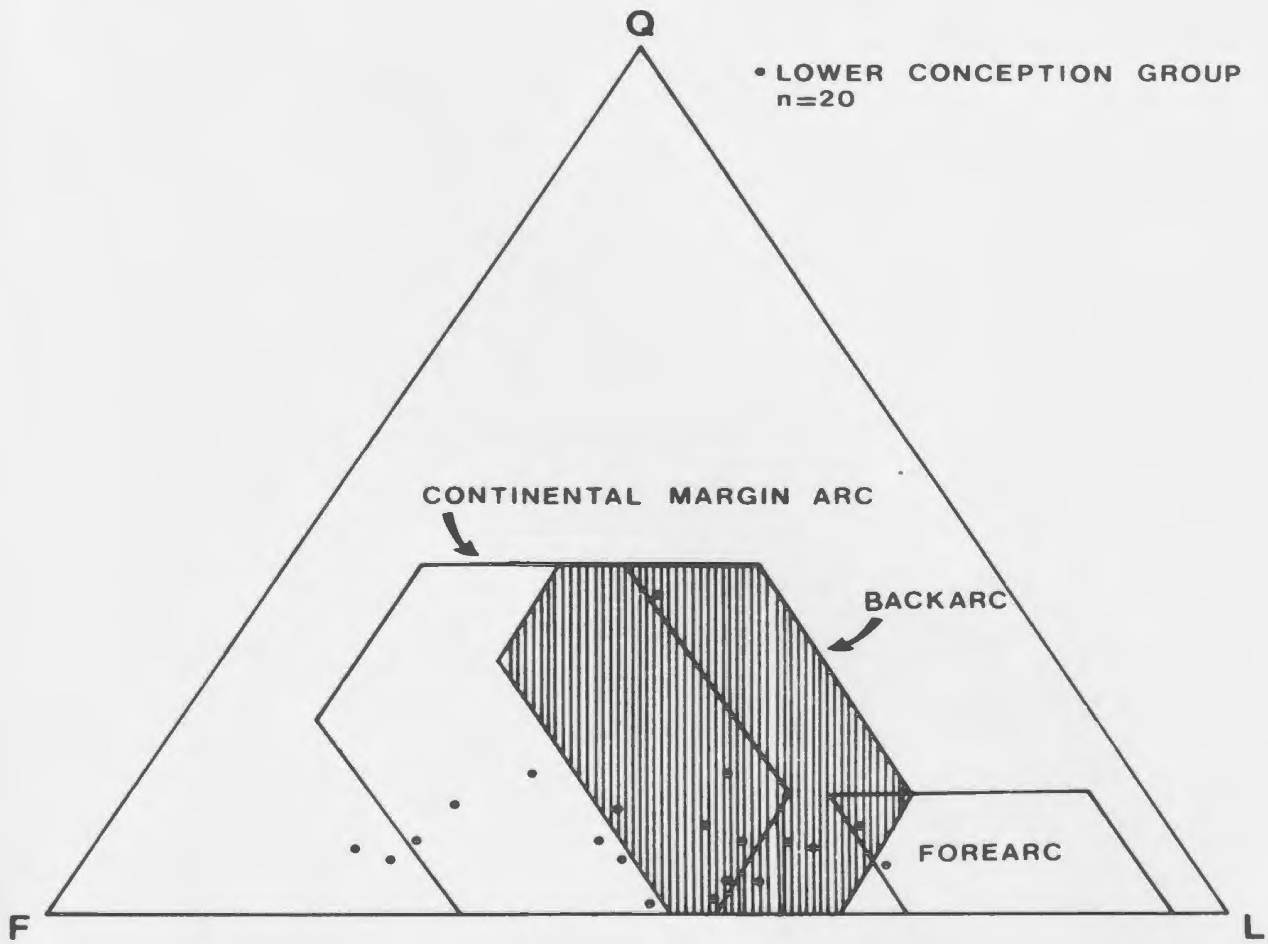


Figure 4.9: On the general QFL diagram for modern sand compositions (Maynard, Valloni and Yu, 1982; Yerino and Maynard, 1984) the lower Conception Group sandstones plot in the regions occupied by sands from backarc, forearc, and continental margin arc settings.

andesitic composition, suggests that the volcano-sedimentary sequences of the Newfoundland Avalon Zone did not form in a setting similar to that of classical Mesozoic-Cenozoic Andean arcs (King, 1982, p. 23; O'Brien et al., 1983).

Mack (1984) advises a cautionary approach when inferring tectonic setting based on sandstone composition. He determined four categories of anamolous sandstone compositions which have a tectonic interpretation inferred from Dickinson's diagrams that does not agree with tectonic setting determined exclusive of petrography. Velbel (1979) advises similar caution when interpreting sandstone petrographic data from forearc regions. Based on petrographic work on sandstones from the Barbados submarine ridge, Velbel suggests that their quartz-rich composition, which is considered to be anomolous for accretionary complexes associated with magmatic arcs, may be more prevalent in modern arc settings than previously suspected (ibid.). Also, because the regions depicting the various tectonic settings proposed on Dickinson's QFL diagrams are based only on arc-related sandstone suites, any attempts to use these diagrams will always result in an arc interpretation. However, because volcanoes also occur in non arc-related settings, such as mid-ocean ridge and intraplate settings, distinguishing between orogenic magmatic arcs and other volcanic settings using detrital sand composition is difficult, if not impossible.

Sandstone petrography of the lower Conception Group is consistent with a subduction arc-related model, but is not compelling evidence in favour of this setting. Equally appropriate tectonic settings for the Newfoundland Avalon Zone are the Basin and Range extensional regime proposed by Papezik (1970,1972) and Strong _et al._ (1974), the continental extension and rifting regime proposed by Rankin (1975) and Strong _et al._ (1978), and the tectonic model of O'Brien _et al._ (1983) which involves local and limited subduction during closure of small ocean basins in its final stage.

# 5. PALEOCURRENT ANALYSIS

## Introduction

Data used to analyze paleocurrent directions in all facies of the Mall Bay and Drook Formations were collected from field measurements of current ripples and sole markings. Also, oriented samples were collected for grain fabric study. Three types of solemarks were observed; flute casts, groove and striation casts, and less commonly, prod marks. While flutes and prod marks do provide unidirectional flow information, grooves and striations are used for sense of flow direction only. In some cases sandstone fabric samples, taken from just above the groove and striation casts, provided unidirectional flow information (see Appendix I). For fine grained lithologies (i.e., argillite) containing groove and striation casts, determination of unidirectional flow using fabric was not possible. Paleoslope directions determined from coherent and partially coherent sedimentary slides (see Nardin et al., 1979) were not given any weight when determining paleoflow and basin geometry.

The use of solemarks as paleocurrent indicators must be approached with caution. Although solemarks may indicate the flow direction at a given point on a bed, the flow direction may be variable along the base of the bed. These variations may indicate meandering flow patterns as described by Parkash and Middleton (1970).

Due to the nature of the welded contact between fine to coarse grained sandstone beds and the underlying argillite, soles are generally not exposed. For this reason, and to corroborate the paleocurrents obtained from the available solemarks, oriented samples of fine to coarse grained sandstones were analysed for grain fabric, an alternative paleocurrent indicator.

Assuming that the effects of slope are negligible, the long axes of elongate sand grains respond to unidirectional flow and usually parallel the current and are imbricated upcurrent (Potter and Pettijohn, 1977, p. 51). The stratigraphic location of numbered oriented samples is indicated in Appendix I. Wayland (1939) was the first to describe preferred orientation of quartz grains in undeformed sandstones although Becker (1893) previously noticed the effect in coarse stream sediment. However, Kopstein (1954) was the first to utilize sedimentary fabric in a systematic study of paleocurrents. Since then, numerous workers have used fabric to compare the relationship between paleoflow indicators obtained from solemarks and fabric (Smoor, 1960; Hand, 1961; Bouma, 1962; Spotts, 1964; Spotts and Weser, 1964; Sestini and Pranzini, 1965; Onions and Middleton, 1968; Colburn, 1968), and to examine changes of fabric vertically and horizontally within one bed (Rukavina, 1965; Onions and Middleton, 1968; Parkash and Middleton, 1970; Hiscott and

Middleton, 1980). Gibbons (1972) and Martini (1971) discuss the biasses and validity of grain counting methods for paleocurrent analysis.

Although not employed in this thesis, sedimentary fabric has also yielded information about depositional mechanisms of sedimentary units (Walker, 1975; Taira, 1976a,b; Hiscott and Middleton, 1980).

## Method

The method described by Pettijohn and Potter (1977, p. 252) for measuring solemark orientation was employed in this thesis. The azimuth of the solemarks on any given bed was constant. Therefore, averaging of the solemark azimuths was not necessary. Equal area rose diagrams for each type of paleoflow indicator (i.e., solemarks, ripples, grooves) are used to indicate paleoflow patterns. Each measurement which appears on a rose diagram represents the paleoflow on a single bed. Therefore, a rose diagram which summarizes 16 measurements represents the data from 16 different beds each of which may have had one or more solemarks.

Paleocurrent data from outcrops with inclined strata dipping greater than 25° were corrected for tectonic tilt using stereonet rotation procedures described by Philips (1967) and Ramsey (1967, p. 486-488). The axis of rotation employed was the strike of the beds. For strata tilted less than 25°, the maximum possible error between the measured and original azimuths is 3°, and no stereographic correction is required (Ramsey, 1961). The plunge of the folds in the study area is $\leq$ 10° and any associated error is negligible (ibid.).

Oriented sandstone samples were obtained from the base of normally graded or parallel laminated beds or layers, using a geology hammer, sledge hammer, chisel, Brunton compass and a marker pen. Where possible, basal bedding surfaces were sampled to enable accurate reorientation in the laboratory. On each sample, a strike and dip indicator, a "way-up" indicator, and a line representing the present horizontal were drawn in addition to the sample number. The sample number includes two letters, which represent the outcrop sample location (i.e., HP-12= Holyrood Pond). The maximum angular error associated with this sampling method is estimated to be $\pm$ 5 for both orientation and imbrication measurements.

The fabric of a sedimentary rock is defined as "the way in which the grains are put together to make an aggregate" (Pettijohn and Potter, 1972, p. 89). Alternatively, it is defined as "the orientation in space of the particles of which a sedimentary rock is composed" (AGI, 1972). Historically, several methods have been used to measure fabric. These methods include: a) direct measurement under the microscope using a 360° rotational stage with a Vernier scale, b) projecting the thin section onto a screen and tracing of elongate grains onto paper, c) photographing the thin section, d) producing an acetate peel, or e) measuring magnetic susceptibility anisotropy (Rees, 1965, 1966; Taira and Lienert, 1979). Photographing and tracing a projected thin section are advantageous methods because they provide a permanent record of the fabric.

The projection of thin sections, magnified up to 25X, 50X, or 100X (depending on grain size), was used in this study. Quartz, feldspar and rock fragments were outlined if they were encountered at intersection points on a 3 cm-square grid. Only grains with a b/a axis ratio of 0.7 or less were outlined. The use of this b/a axis ratio for a cutoff criterion was determined empirically, as outlined by Potter and Mast (1963, Table 3). Ideally, sand-size grains of equal composition should be used in fabric studies to avoid errors associated with different hydrodynamic properties due to varying composition and density contrasts. However, the highly variable and quartz-poor composition of these sandstones required measurement of all lithic fragments, plagioclase and quartz grains that met the size and elongation criteria.

Several methods have been utilized for measurement of the long dimension (a axis) of sand grains. These include: a) measurement of the azimuth of the longest apparent axis (the symmetry plane in grains with bilateral symmetry), b) least projection elongation (Dapples and Rominger, 1945)(Fig. 5.1) and c) center of area elongation (the longest line that can be drawn through the center of the area of the grain). Least projection elongation and the trace of the symmetry plane or longest apparent axis were used to measure 100 grains per sample. Within each bedding plane, azimuthal angles of oriented grains were recorded with respect to north in the range of 0° to 180°.

The 100 azimuth values were entered in data files on a Vax/Vms computer to test for a preferred orientation of grains. If no statistically preferred orientation exists, the fabric is said to be isotropic. Any statistically preferred orientation constitutes an anisotropic fabric. The Rayleigh test (Curray, 1956), F test (ibid.) and the chi-squared test are statistical methods which are used to test the significance of two dimensional orientation distributions based on the properties of a normal population. The Rayleigh test calculates vector magnitude in percent and the probability of obtaining a greater than normal amplitude by chance combination of random phases. The F test evaluates the significance of orientation distributions through the F ratio; a ratio of the variance of a uniform distribution about the vector mean, to the observed sample variance. The chi-squared test measures the significance of the differences between the observed

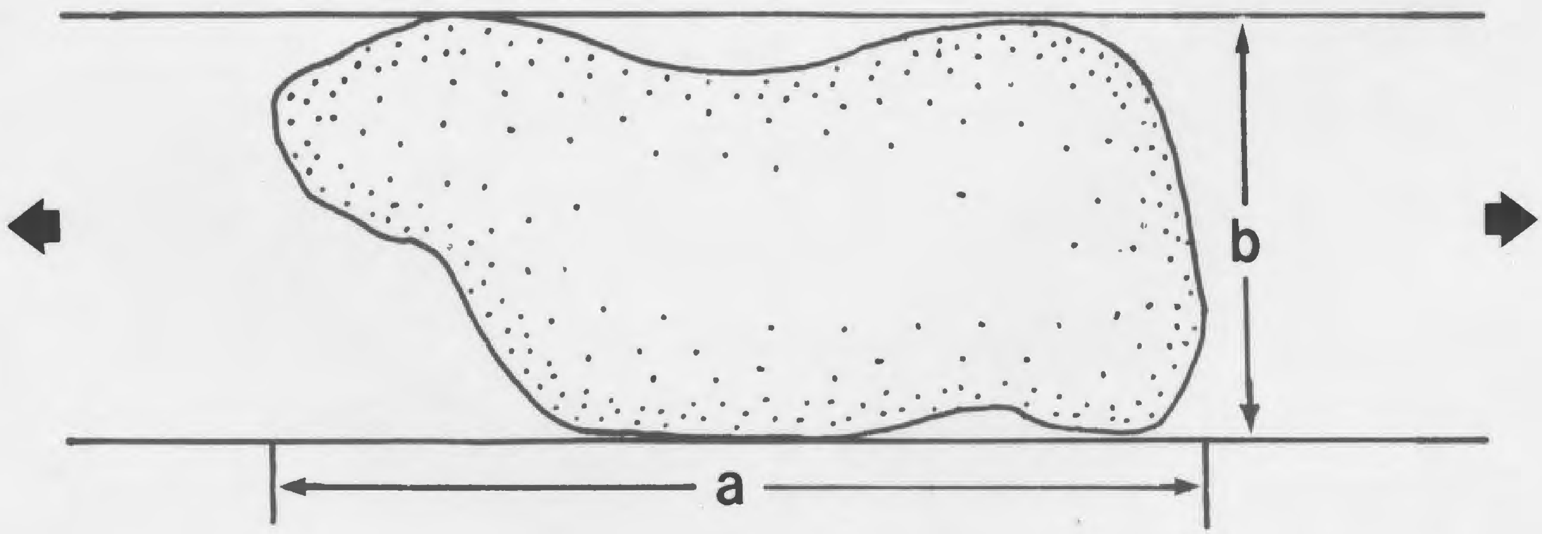


Figure 5.1: **Least projection elongation** (after Dapples and Rominger, 1945).

number of observations per group and that expected for a random sample.

Curray (1956) and Middleton (1965, 1967) state that these tests are too sensitive for a grain orientation study because they detect deviations from a uniform distribution that may not constitute a "preferred orientation". A modification of the chi-squared test suitable for grain orientation studies is the Tukey chi-squared test (Rusnak, 1957; Harrison, 1957; Middleton, 1965, 1967). Unlike normal chi-squared tests, the Tukey chi-squared test searches for a statistically significant departure from a uniform circular distribution which will result in a non-zero mean orientation vector (Davis, 1973). Directional vectors in a random circular distribution would cancel each other, leaving a vector mean of zero. Because data are recorded in groups from $0°$ to $180°$, a $2\theta$ transformation is necessary to give a unimodal test sample. In bimodal samples, a $4\theta$ transformation is conducted prior to subjecting data to the Tukey chi-squared test. Similar to the $2\theta$ transformation, the $4\theta$ transformation allows two vector modes at $180°$ to reinforce each other. Also, it allows two vector modes separated by $90°$ to reinforce each other for significance-level testing.

Vector magnitudes were calculated for each sample using the method of Curray (1956). This involves calculation of the magnitude of a north-south component

($\sum n\cos2\theta$) and an east-west component ($\sum n\sin2\theta$) for each observation or group of observations (ibid.). The following equations are used (Curray, 1956):

$$r = [(\sum n\sin2\theta)^2 + (\sum n\cos2\theta)^2]^{1/2}$$

$$L = 100r/\sum n$$

$$\bar{\theta} = (1/2)\arctan(\sum n\sin2\theta/\sum n\cos2\theta)$$

where: $r$ = resultant vector magnitude

$L$ = vector magnitude in per cent

$\theta$ = azimuth from 0° to 180° for each observation or group

$\bar{\theta}$ = azimuth of resultant vector

The data were processed using a Fortran IV program designed to: a) perform the Tukey chi-squared test, b) calculate a vector mean, c) calculate vector magnitude, d) find the significance level of the vector mean with two degrees of freedom operating and e) group grain orientations into 10° classes from 0° to 180°.

Thirty-nine thin sections were cut parallel to bedding to determine flow orientation. A significance level of 0.10 (90%) was arbitrarily chosen as the minimum cutoff criteria for a statistically significant preferred orientation. This means that there is a 1 in 10 chance that a sample for which the grain orientation has been calculated to be statistically significant, may not be significant. Alternatively, it also means that a sample for which the grain orientation has been calculated statistically and found not to be significant, actually may be significant. Only 23 of the 39 samples (59%) tested had a preferred orientation with an adequate significance level of 0.10 or 90%. These samples were then cut perpendicular to bedding, in the direction of the preferred grain orientation, in order to evaluate grain imbrication (see Spotts, 1964, p. 242).

## Errors

An error of $5^{\circ}$ is estimated for sample orientation in the field. In addition, when cutting and making thin sections, a potential maximum error of $\pm 5^{\circ}$ may be expected. This error occurs twice; once each for cuts both parallel and perpendicular to bedding. To find the error associated with operator inconsistency and bias in obtaining grain azimuths, an operator variation test was conducted. Two operators measured 100 fabric elements in two specimens. This process was repeated four weeks later. The resulting vector means of each slide obtained by both operators were equal to within $4^{\circ}$ of each other. One mean differed by $7^{\circ}$ from the previous trial, but by only $3^{\circ}$ from the other operator in the same trial. Therefore, the maximum error due to operator inconsistency is about $\pm 5^{\circ}$. The sum of the above errors is $\pm$ 15°. The error may be slightly less for imbrication results due to higher vector magnitudes.

In a study of this type, a potential $\pm$15° maximum error is not serious because; a) the data were grouped into 30 intervals (coarse divisions), and b) the overall paleocurrent trend will not be significantly altered because more significance can be attributed to consistent trends than to individual measurements. Furthermore, errors may occur antithetically, thereby cancelling each another out. This effect can reduce the expected cumulative error in any single sample. Based on the above, only those imbrication angles measured to be greater than 5° were considered to be significantly different from the horizontal.

## Results

Table 5.1 indicates the fabric determinations for all 39 thin sections studied. Significant grain orientation azimuths vary from 11° to 108°. Significant imbrication angles obtained from fabric measurements, range from 23° to 38°. Seventy-four per cent (74%) of imbrication results were less than 30°, 26% less than 20°, and 4% less than 10°. No apparent relationship exists between the angle of imbrication and vector magnitude for preferred orientations within sandstones of the Conception Group. Based on the paleoflow patterns determined for each outcrop location

Table 5.1: $\chi^2$ results for sandstone fabric

| | SAMPLE NO. | ORIENTATION VECTOR MEAN (degrees) | ORIENTATION VECTOR MAGNITUDE (percent) | ORIENTATION SIGNIFICANCE LEVEL (percent) | ORIENTATION RANK | IMBRICATION VECTOR MEAN (degrees) | IMBRICATION VECTOR MAGNITUDE (percent) | IMBRICATION SIGNIFICANCE LEVEL (percent) | IMBRICATION RANK |
|---|---|---|---|---|---|---|---|---|---|
| HOLYROOD POND | HP-2 | 005 | 1.52 | 50 | random | — | — | — | — |
| | HP-3 | 096 | 51.41 | 99 | excellent | 025 | 26.59 | 50 | random |
| | HP-4 | 042 | 32.56 | 99 | excellent | 038 | 55.61 | 99 | excellent |
| | HP-5 | 061 | 35.86 | 99 | excellent | 030 | 53.26 | 99 | excellent |
| | HP-7 | 084 | 27.41 | 99 | excellent | 024 | 35.73 | 70 | poor |
| | HP-9 | 077 | 28.60 | 99 | excellent | 027 | 20.19 | 50 | random |
| | HP-10 | 072 | 41.58 | 99 | excellent | 167 | 21.53 | 50 | random |
| | HP-12 | 026 | 17.88 | 90 | good | 036 | 37.11 | 80 | moderate |
| | HP-13 | 100 | 24.05 | 99 | excellent | 029 | 64.40 | 99 | excellent |
| | HP-15 | 048 | 26.75 | 99 | excellent | 023 | 36.33 | 80 | moderate |
| | HP-16 | 049 | 15.63 | 80 | moderate | 019 | 17.01 | 50 | random |
| | HP-18 | 011 | 19.37 | 95 | strong | 013 | 30.69 | 50 | random |
| | HP-19 | 108 | 25.77 | 99 | excellent | 031 | 71.00 | 99 | excellent |
| | HP-20 | 042 | 43.28 | 99 | excellent | 035 | 36.64 | 70 | poor |
| | HP-21 | 043 | 16.03 | 80 | moderate | 016 | 35.70 | 70 | poor |
| | HP-22 | 068 | 27.25 | 99 | excellent | 035 | 59.85 | 99 | excellent |
| | HP-25 | 035 | 31.66 | 99 | excellent | 027 | 40.48 | 95 | strong |
| | HP-26 | 066 | 36.36 | 99 | excellent | 023 | 41.60 | 95 | strong |
| | HP-27 | 087 | 12.41 | 50 | random | — | — | — | — |
| | HP-29 | 064 | 19.6 | 95 | strong | 001 | 26.54 | 50 | random |

Ranking hierarchy is based on significance level.

(continued. . . .)

| SIGNIFICANCE | RANK | | |
|---|---|---|---|
| 99 | excellent | 80 | moderate |
| 98 | very strong | 70 | poor |
| 95 | strong | 50 | random |
| 90 | good | | |

Table 5.1: (continued): $\chi^2$ results from sandstone fabric

| SAMPLE NO. | ORIENTATION VECTOR MEAN (degrees) | VECTOR MAGNITUDE (percent) | SIGNIFICANCE LEVEL (percent) | RANK | IMBRICATION VECTOR MEAN (degrees) | VECTOR MAGNITUDE (percent) | SIGNIFICANCE LEVEL (percent) | RANK |
|---|---|---|---|---|---|---|---|---|
| PH-2 | 126 | 9.17 | 50 | random | — | — | — | — |
| PH-3 | 143 | 15.28 | 50 | random | — | — | — | — |
| PH-4 | 007 | 24.06 | 50 | random | — | — | — | — |
| PH-5 | 162 | 20.77 | 50 | random | — | — | — | — |
| PH-6 | 128 | 7.00 | 70 | poor | — | — | — | — |
| PH-7 | 151 | 18.78 | 50 | random | — | — | — | — |
| PH-17 | 124 | 1.59 | 98 | very strong | 005 | 41.53 | 95 | strong |
| PH-18 | 110 | 18.59 | 50 | random | — | — | — | — |
| CE-3 | 044 | 42.16 | 95 | strong | 159 | 54.20 | 99 | excellent |
| CE-5 | 047 | 33.17 | 50 | random | — | — | — | — |
| CE-6 | 121 | 7.27 | 70 | poor | — | — | — | — |
| CE-7 | 033 | 44.74 | 99 | excellent | 046 | 58.73 | 99 | excellent |
| CE-9 | 068 | 27.91 | 50 | random | — | — | — | — |
| CE-10 | 030 | 40.89 | 95 | strong | 163 | 39.61 | 95 | strong |
| CE-11 | 008 | 27.98 | 50 | random | — | — | — | — |
| CE-12 | 054 | 32.55 | 70 | poor | — | — | — | — |
| CE-13 | 133 | 8.07 | 80 | moderate | 160 | 35.44 | 80 | moderate |
| CE-14 | 129 | 8.48 | 99 | excellent | 158 | 41.07 | 95 | strong |

(Figs. 5.2, 5.3, 5.4 and 5.5), using ripples, solemarks and fabric where applicable, the following observations were made:

Mall Bay Formation

i) Paleoflow obtained from solemarks, ripples and fabric are not significantly different from one another.

ii) Most paleoflow patterns are unimodal with rare bimodality. Dominant directions are to the southwest and northwest with strong easterly flow in the lower basin-slope deposits (recall facies association II).

iii) Mean current directions obtained from flutes, grooves or ripples contained in the same facies are not significantly different from one another.

iv) Only one of the nine fabric samples had a preferred orientation. This was probably a function of the fine grain size of the sediment. Grain sizes smaller than fine sand are difficult to accurately trace onto paper using the fabric analysis method employed in this thesis.

Drook Formation

i) Paleoflow is dominated by unimodal patterns toward the northeast with a few unimodal trends to the southeast and southwest.

ii) Excellent agreement exists between paleoflow

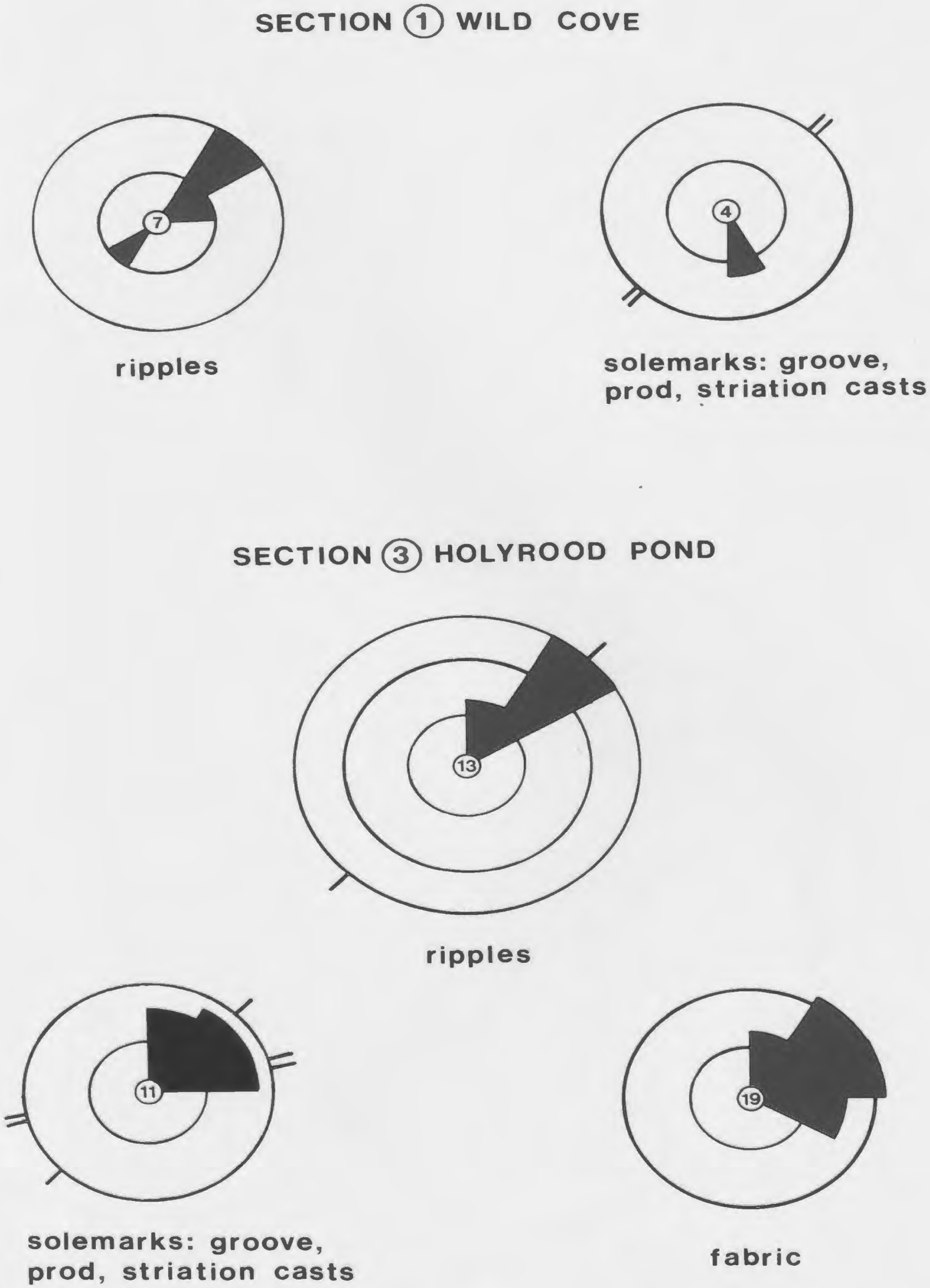


Figure 5.2: Paleocurrent rose diagrams. Circles of increasing diameter represent 1, 5, and 10 measurements respectively.

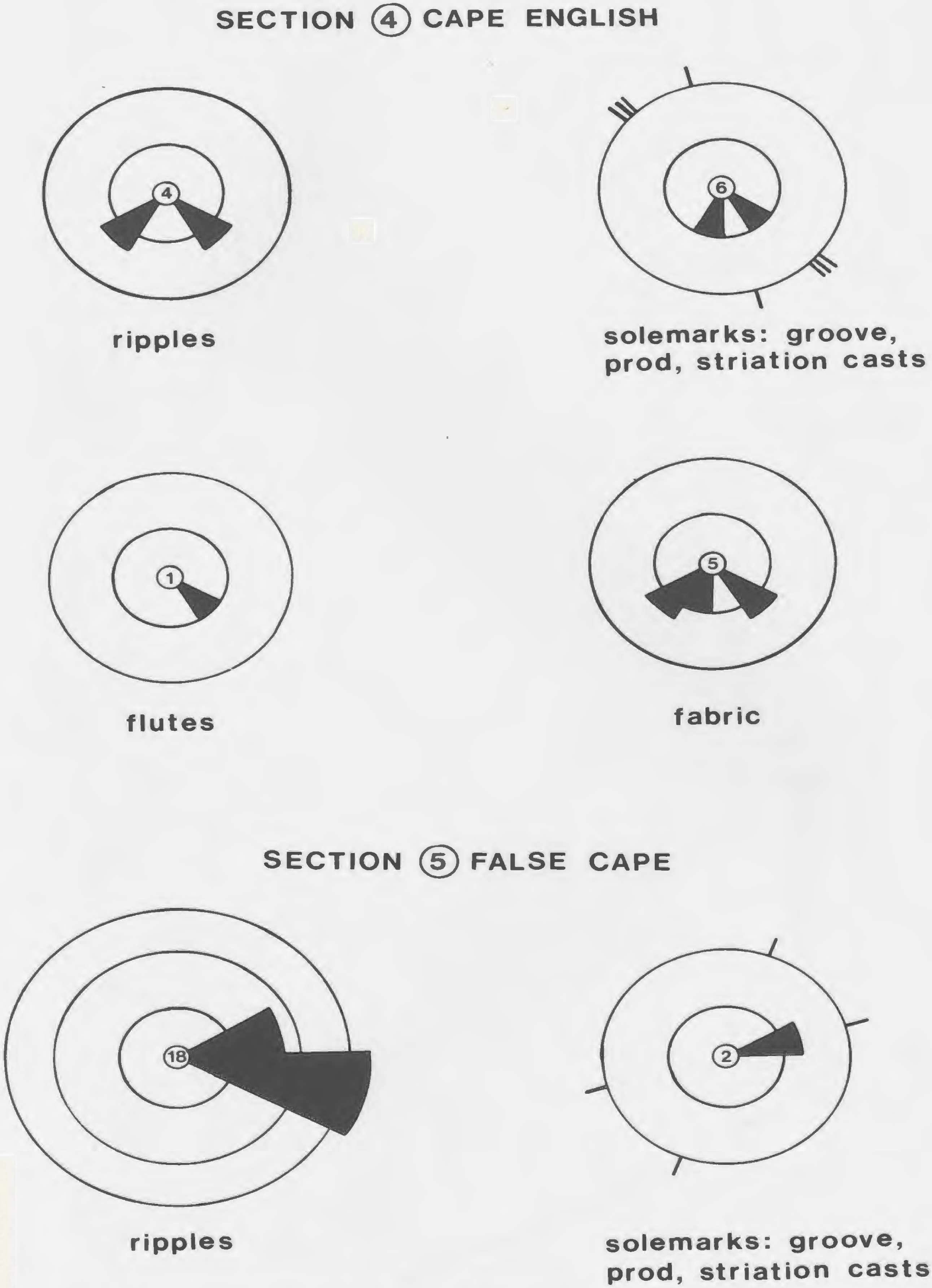


Figure 5.3: Paleocurrent rose diagrams. Circles of increasing diameter represent 1, 5, and 10 measurements respectively.

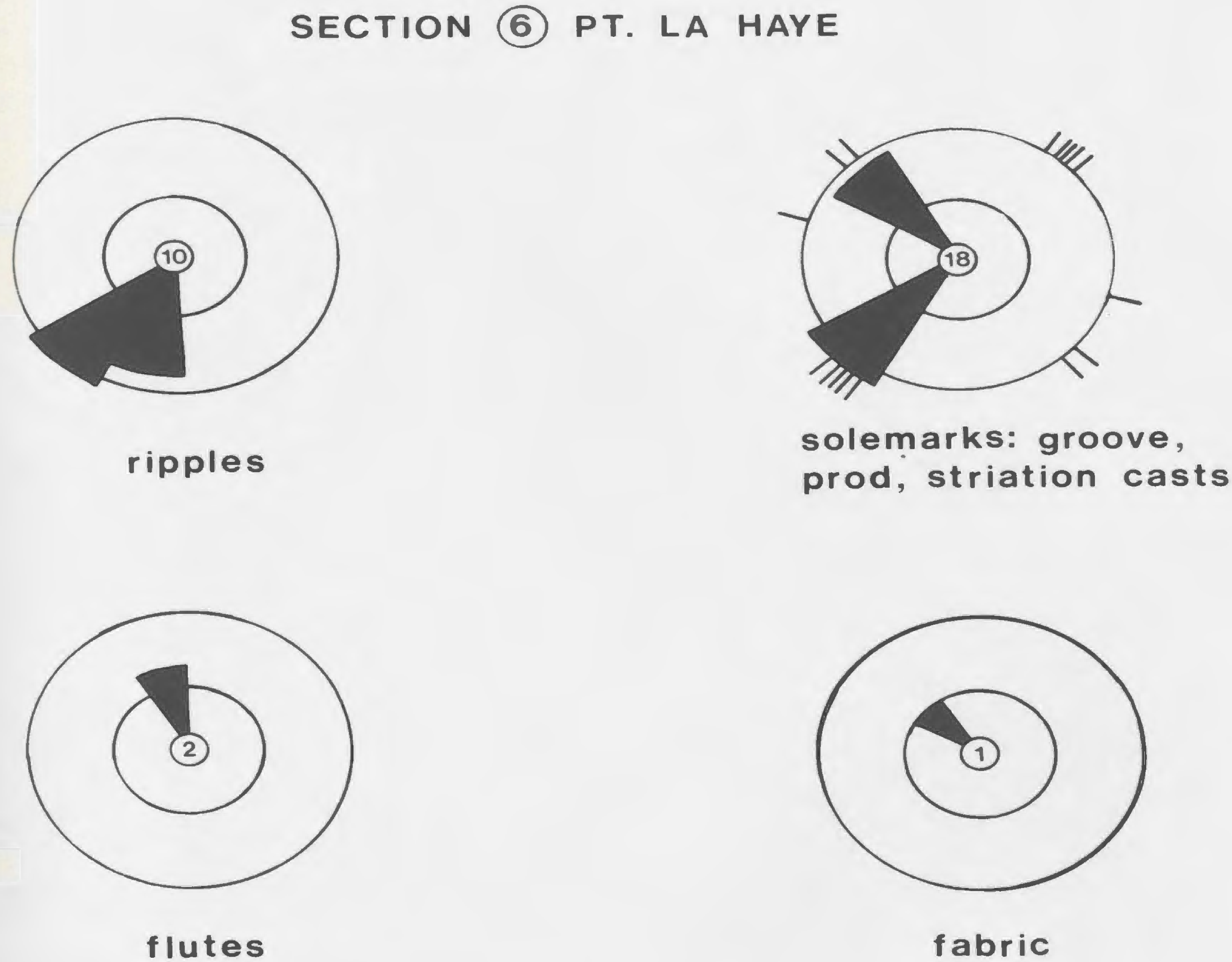


Figure 5.4: Paleocurrent rose diagrams. Circles of increasing diameter represent 1 and 5 measurements respectively.

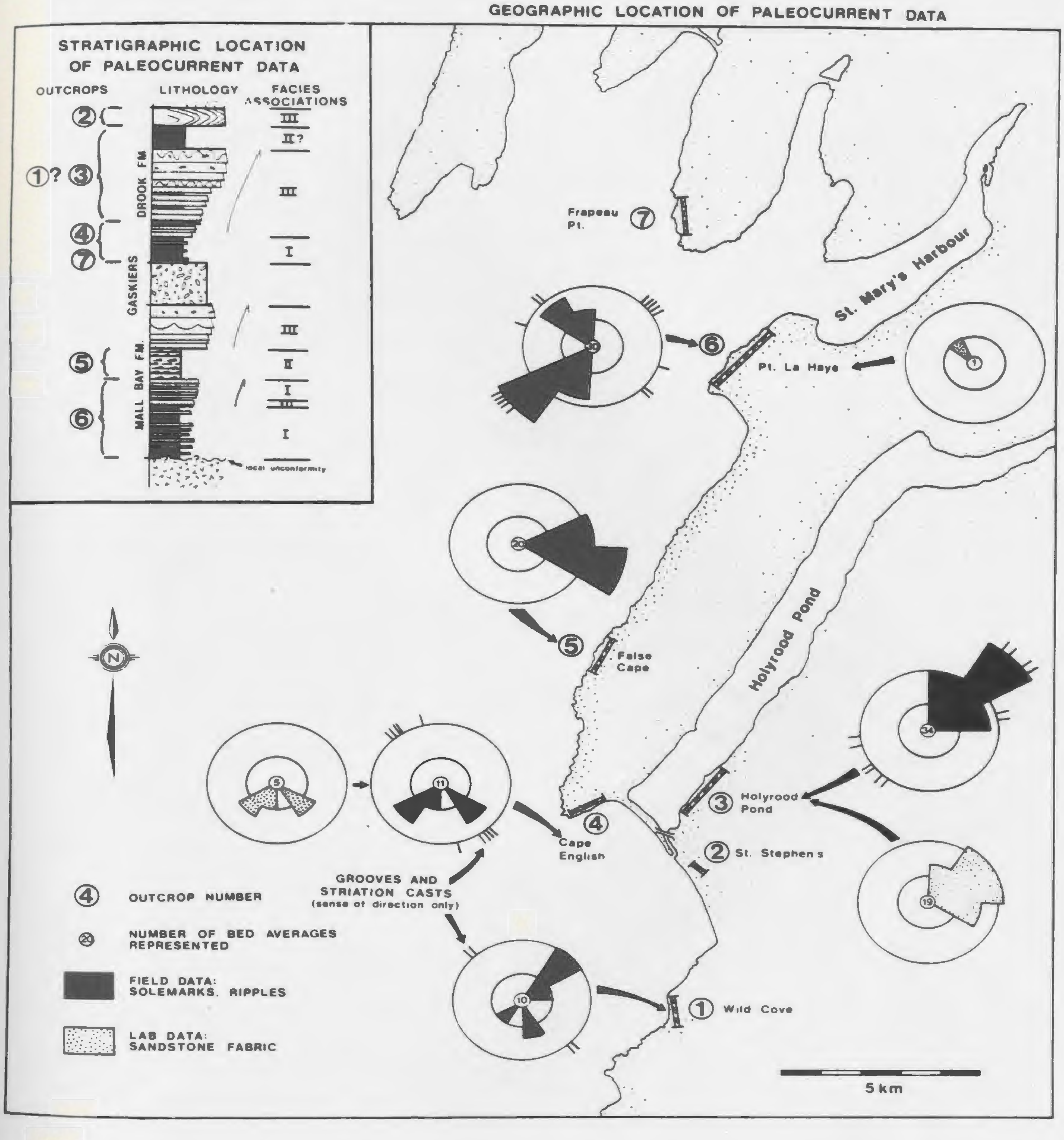


Figure 5.5: Geographic and stratigraphic location of paleocurrent data for the Mall Bay and Drook Formations. Note that the data from various outcrops is from different stratigraphic intervals (see key in inset top left). Refer to Fig. 3.25 for the relationship between facies associations and paleocurrents.

obtained from fabric and that deduced from solemarks and ripples.

iii) Grain fabric did vary as a function of grain size. Medium to coarse grained sandstones have the strongest anisotropic fabrics.

Equal area rose diagrams are presented separately for each type of sedimentary structure (Figs. 5.2, 5.3, 5.4). Alternatively, paleoflow data could have been grouped according to facies. Within each facies of both the Mall Bay and Drook Formations, the paleoflow indicators are in a consistent direction. Variations of paleoflow direction for the outcrops studied is largely stratigraphically controlled, not facies controlled (see Appendix I). For example, within the lower Mall Bay Formation, the overall flow varies from northwest to southwest at the base (at Point La Haye), while in the upper part of the formation (at False Cape), the flow is dominantly toward the east. The easterly paleoflow direction at False Cape is based exclusively on ripples and solemarks. No fabric analysis was attempted because it was impossible due to the highly altered, fragmented nature of the tuffaceous sandstones. This stratigraphic change in flow direction within the Mall Bay Formation may reflect newly initiated or recurring volcanism to the west. The evidence for this recurring volcanism is the presence of the tuffaceous detritus found only at the top of the Mall Bay Formation. Stratigraphic control of paleoflow direction is generally believed to indicate changing morphology and submarine topography of the basin with time due to tectonic controls such as penecontemporaneous volcanism, faulting, and compression or extension.

With the exception of facies association II at False Cape, the overall sediment dispersal pattern of the lower Conception Group in this part of the basin indicates a northeast-southwest trend (Fig. 5.5). The rippled argillite of facies association II, inferred to be lower basin-slope deposits, exhibits an easterly flow pattern which is approximately perpendicular to the overall trend. This transverse flow pattern is probably the result of sediment reworking by marine bottom currents or by low density, low velocity turbidity currents flowing across the slope, perhaps caused by spilling over the channel margin in the mid-fan area (see chapter 3).

## Discussion and comparison with previous paleocurrent work

It seems unlikely that the paleocurrent patterns observed reflect the effects of meandering turbidity currents described by Parkash and Middleton (1970). The evidence to support this is that virtually all beds in a randomly chosen stratigraphically continuous succession indicate the same paleoflow direction with only minor scatter (e.g., Appendix I, p. 183, 250-270 m and p. 187, 25-35 m and 35-45 m). If meandering currents were present, one would anticipate significant variation in paleoflow on a bed by bed basis. This was not observed.

Williams and King (1979) first obtained paleoflow information from field measurements during regional mapping of the entire Avalon Peninsula. Also, Stewart (1982) has studied the paleocurrent patterns in the Conception Group using sandstone fabric and field measurements. The location of Stewart's (1982) study area is the same as that of this thesis. However, only the outcrops at Point La Haye, Cape English, and Holyrood Pond are common to both theses. Although Williams and King (1979) and Stewart (1982) did not indicate the stratigraphic position of their paleocurrent results within those parts of the Mall Bay and Drook Formations which they studied, their observations still warrant comparison to those found in this thesis.

Firstly, the paleocurrent directions obtained from field measurements in this thesis are consistent with those of all previous workers. There are some discrepancies between Stewart's sandstone fabric results and those of the present study. These discrepancies will be discussed for each of the three outcrop locations mentioned above.

At Point La Haye, only one of the nine fabric samples in this study yielded a statistically significant preferred grain orientation. A paleoflow to the northwest was observed. In contrast, Stewart (1982) found two of seven fabric samples to yield a preferred orientation in a southwesterly direction. Stewarts's (1982) fabric data is in agreement with field observations in this thesis. Due to the small grain size (maximum size is fine grained sandstone) of lithologies at this location, sampling was limited. The discrepancy between the two fabric results may be simply due to error. In light of the potential for error when attempting to obtain useful data from these fine grained sediments, the results should not be given much weight.

At Cape English, sandstone fabric results differ from those obtained by Stewart (1982). While fabric results from this thesis indicate a southerly trend varying from $120^{\circ}$ to $240^{\circ}$, Stewart's study yielded northeasterly paleoflow directions ranging from $036^{\circ}$ to $075^{\circ}$. This disagreement is difficult to reconcile. In addition to the possibility of error, another plausible explanation is that Stewart sampled from different beds from within the 40 m section. Alternatively, it is conceivable that the fabric changes along the strike of these beds (cf. Parkash and Middleton, 1970), although this explanation is not favoured for reasons discussed earlier.

At Holyrood Pond the coarsest grained sandstones crop out and are the most suitable for fabric analysis within the study area. The field measurements and fabric results from this thesis are in agreement with each other and are unimodal to the northeast. Stewart's (1982) fabric results contained one paleocurrent reversal toward the southwest. He suggested that opposing grain imbrication may result from a turbidity current reflection from basin slopes or other submarine topography (cf. Kelling, 1964; Colburn, 1968; Pickering and Hiscott, in prep.). Again it is possible that Stewart (1982) sampled from different beds within the 279 m section. The moderate vector magnitude of 17.20 obtained for an imbrication of 25° in Stewart's paleocurrent reversal suggests that this data is probably not in error. It seems likely that Stewart's paleocurrent reversal is a very localized stratigraphic occurrence. The excellent internal agreement between the field measurements and the fabric results from this thesis, and the greater number of samples and field measurements taken, suggests that the dominant flow at Holyrood Pond is still to the northeast.

# 6. SYNTHESIS AND LOCAL BASIN ANALYSIS

## Introduction

Basin analysis is the structural, stratigraphic and paleogeographic study of large, regionally distributed packages of sedimentary rocks (Miall, 1983, p. 1). Therefore, the following information is needed (Potter and Pettijohn, 1977, p. 341):

i) the facies distribution and thickness within the basin at various stages of its filling

ii) the paleocurrent pattern

iii) the relation of facies and paleocurrents to both contemporaneous and later deformation

iv) the petrology of its fill

For the lower Conception Group, synthesis of information contained in chapters 3, 4 and 5 is required to accomplish basin analysis. Local basin characteristics will be extrapolated to approximate the regional paleogeography and tectono-sedimentary nature of this basin in late Precambrian time. Lower Conception Group sedimentation will then be compared to modern and ancient examples.

## Paleogeography and Depositional History

The local complexity of the paleocurrent and facies patterns suggests that several local volcanic and sedimentary source areas were present within, or adjacent to, the basin. Determining paleogeography of the lower

Conception Group is difficult because of: 1) a lack of dependable stratigraphic correlations, due to an absence of fossils and marker beds, 2) a lack of formation thickness patterns (isopach maps), and 3) rapid facies changes that occur over short distances.

The only previous model of paleogeography and depositional history is that described by King (1980, 1982, p. 19) for all late Precambrian sediments of the Avalon Peninsula. King's regional paleogeography of the lower Conception Group appears to be based solely on cursory facies examination and measurement of paleoflow indicators in widespread locations with little or no input from detailed paleocurrent analysis, facies analysis or petrography. While King's (1982) study provides a framework to work within, an attempt is made here to fit local details of paleocurrent trends and facies occurrences into the regional paleogeographic context of the basin. With this in mind, the following depositional history is suggested for the lower Conception Group (i.e., the Mall Bay, Gaskiers and Drook Formations).

Three hypothetical stages of basin filling are envisioned (Figs. 6.1, 6.2 and 6.3). In order of decreasing age they are: 1) deposition of the Mall Bay Formation in an overall first-order (>500 m), progradational, coarsening- and thickening-upward cycle, 2) deposition on the slope from debris flows of glaciogenic

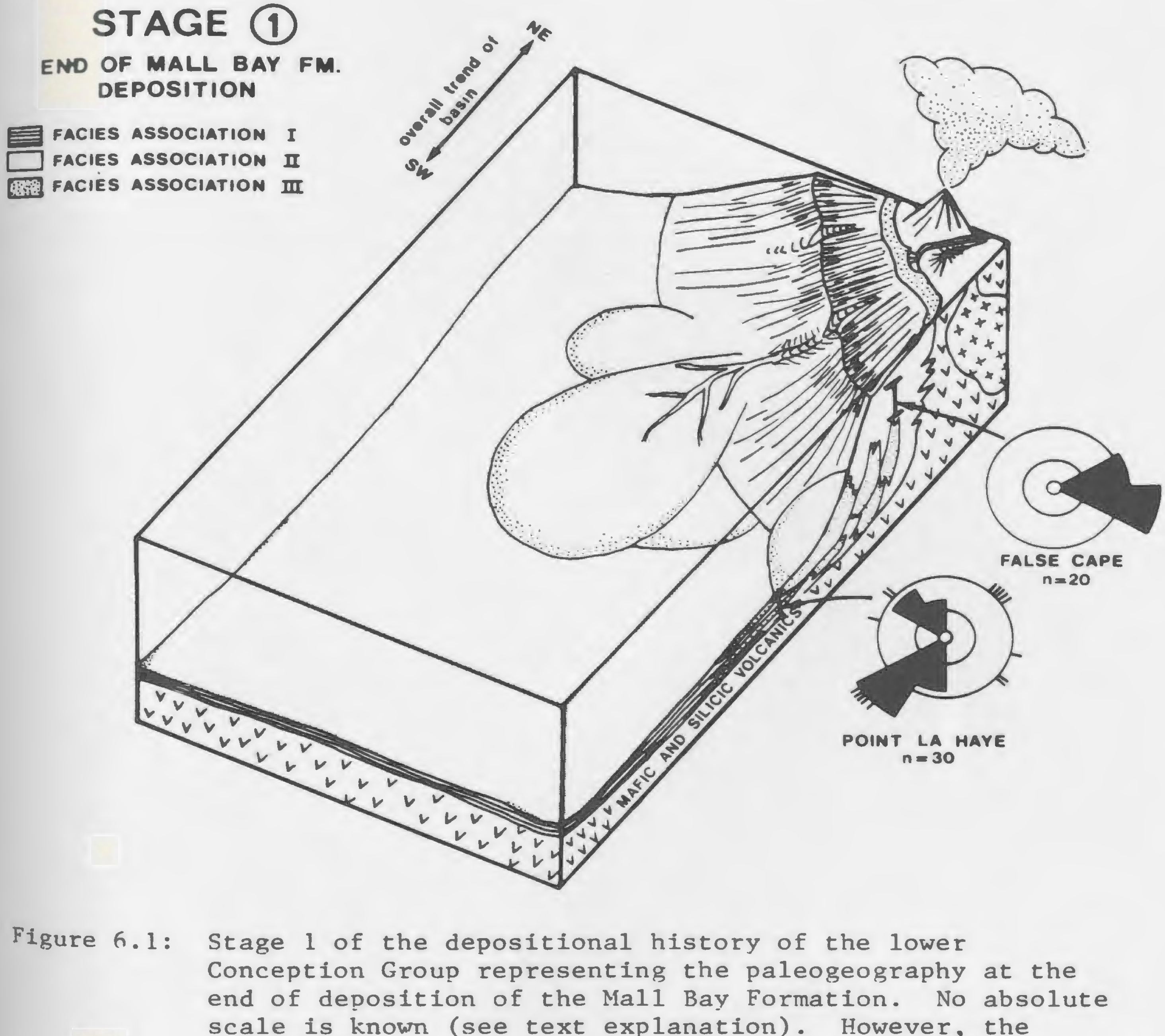


Figure 6.1: Stage 1 of the depositional history of the lower Conception Group representing the paleogeography at the end of deposition of the Mall Bay Formation. No absolute scale is known (see text explanation). However, the turbidite fan system may be relatively small, perhaps on the order of 60 to 100 km from the base of slope to fan fringe, comparable to the modern Astoria Fan (cf. Nelson et al., 1970). The legend for the facies associations refers to the vertical cross-sectional area not the plan view. Paleocurrent rose diagrams are for solemarks and ripples (n= number of bed averages presented).

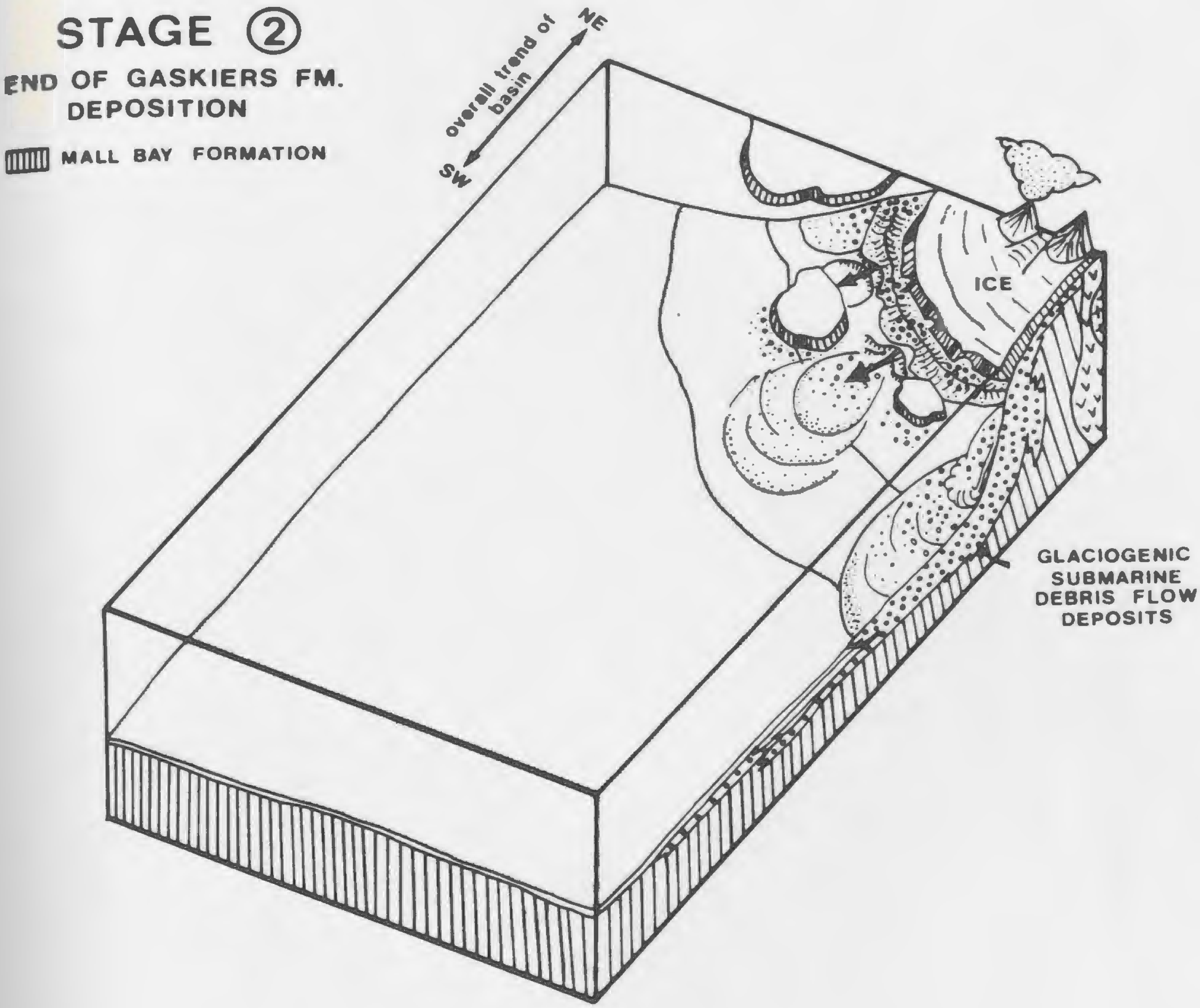


Figure 6.2: Stage 2 of the depositional history of the lower Conception Group representing the paleogeography at the end of deposition of the Gaskiers Formation (after Gravenor, 1980). No absolute scale is known (see text explanation).

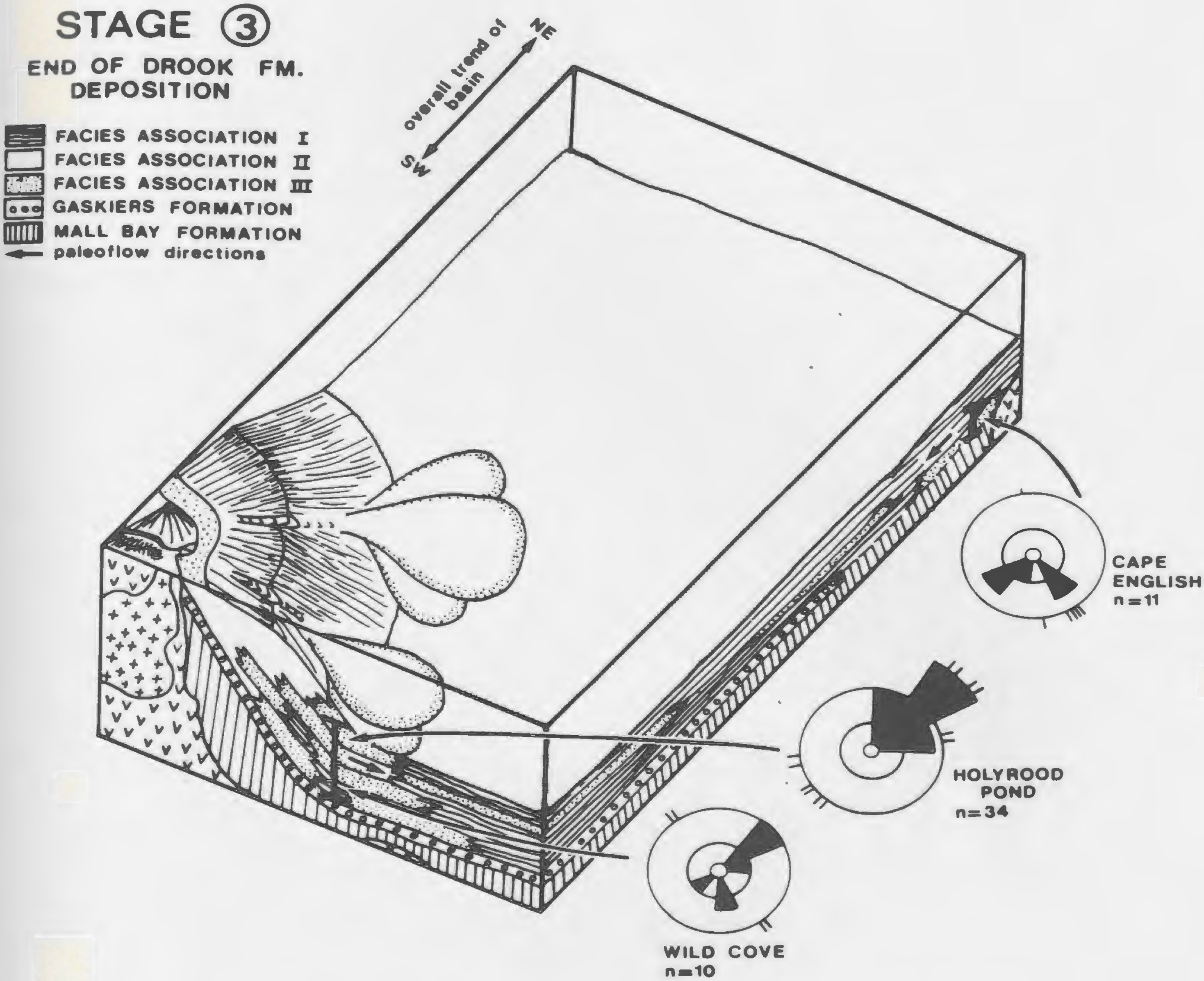


Figure 6.3: Stage 3 of the depositional history of the lower Conception Group representing the paleogeography at the end of deposition of the Drook Formation. No absolute scale is known (see text explanation). Refer to Fig. 6.1 for a discussion of the approximate scale of the turbidite fan. The legend for facies associations refers to the vertical cross-sectional area not the plan view. Paleocurrent rose diagrams are for solemarks and ripples (n= number of bed averages presented).

sediment (Gaskiers Formation) associated with a glacial event of unknown magnitude, and 3) deposition of the Drook Formation to produce an overall first-order (>500 m), coarsening- and thickening-upward cycle.

Stage 1: Deposition of the Mall Bay Formation (Fig. 6.1)

Volcanism was contemporaneous with slope and turbidite fan deposition. The volcanic source terrane was located to the east or northeast of the study area based on southwest and northwest paleoflow indicators observed in the Mall Bay Formation. The easterly paleoflow direction of the muddy slope sediments (facies assocation II at False Cape) is transverse to the inferred turbidite fan axis and probably reflects the effects of marine currents or low velocity turbidity currents operating on the lower basin-slope (see chapter 3). The presence of a gradual, progradational, coarsening- and thickening-upward, first-order cycle from the base to the top of the 800 m-thick Mall Bay Formation is attributed to fan progradation which may be related to basin subsidence, viz. the basin filling with a successive sequence of basin plain/fan fringe, lower basin-slope and lower mid-fan deposits (Fig. 3.24).

No fossil, carbonate, or other data is available to determine basin depth during deposition of the Mall Bay Formation. A speculative estimate of basin depth is on the order of 1 km, perhaps as little as 500 m depth. Calculating a meaningful sedimentation rate for the Mall

Bay Formation is impossible because the age span of the formation is not precisely known (M.M. Anderson, 1972, and pers. comm.). Also, because the tectonic setting of the lower Conception Group is not known (see chapter 4), it is not advisable to compare its sedimentation rate to known rates for turbidites in modern basins.

The size of the turbidite fan can not be accurately determined due to a lack of information regarding slope gradient, water depth, and sediment supply. A speculative estimate, based on the observed thickness of each facies association and the aggregate thickness of the Mall Bay and Drook Formations, suggests that this fan was relatively small. It may be comparable in size to the Astoria Fan (Nelson et al., 1970) and was probably on the order of 60 to 100 km from the base of slope to the limit of the outer fan.

The location of the volcanic island point sources on figures 6.1, 6.2, and 6.3 represents the simplest possible case. It is conceivable that the observed paleoflow directions resulted from feeder channels in these locations which comprised part of a much larger turbidite fan.

Stage 2: Deposition of the Gaskiers Formation (Fig. 6.2)

Any interpretation of the depositional history of the Mall Bay and Drook Formations must be compatible with the sedimentology of the intervening Gaskiers Formation diamictite. Detailed study of the Gaskiers Formation was not the objective of this thesis. Based on cursory outcrop examination, and the work of Gravenor (1980) and Gravenor et al. (1984), the diamictites were probably deposited by downslope transport and resedimentation of glacial sediment by debris flows. This slope interpretation fits well with the depositional model envisioned for the lower Conception Group; viz., deposition on the basin plain (facies association I) and lower mid-fan (facies association III) with adjacent deposition on the lower basin-slope (facies association II) to form the Mall Bay Formation. Also, the dominantly glaciogenic (versus volcanogenic) sedimentation of the overlying Gaskiers Formation mixtites occurs on a submarine slope associated with nearby volcanic islands. The evidence for volcanism contemporaneous with glacial sedimentation is the presence of volcanic detritus, air fall bombs, blocks, and lapilli, within the diamictite (Williams and King, 1979, p. 8; Gravenor, 1980, p. 1333).

The presence of diamictite at the top of the first-order, 800 m-thick, progradational, coarsening- and thickening-upward megacycle of the Mall Bay Formation may imply a regression in the late Proterozoic as suggested by Hughes and Bruckner (1971) and Anderson (1972). However, this progradational sequence is probably not attributable to a glacially-induced global eustatic sea-level lowering (i.e., regression) for two reasons. Firstly, significant changes in eustatic sea level would require glaciation of a very large scale (Shanmugam and Moiola, 1982), for which there is no evidence (Anderson and King, 1981, p. 766). If a more precise age date of Gaskiers Formation glacial event was known, then comparison with other large scale Precambrian glaciations might reveal whether this one is similar (R.J. Rogerson written comm.) However, the inferred nature of the glaciation responsible for Gaskiers Formation deposition is thought to be of local extent, probably a highland glaciation with mountainous volcanic islands as the ice centres (Anderson and King, 1981,p. 766). Secondly, although the time span of the observed progradational sequence is not accurately known (Anderson, 1972), the time represented is probably too long to be considered glacially induced (R.J. Rogerson pers. comm.). Regressions due to glacial events usually require a time scale on the order of thousands to tens of thousands of years (ibid.).

Relative sea level curves have not been derived for the late Proterozoic. The character of the sea level curves cannot easily be inferred using the concepts of Vail et al. (1977) for the lower Conception Group. This difficulty arises because turbidite fan development may be independent of sea-level fluctuations where rates of tectonic uplift/subsidence are large, such as in active continental margins (Klein, 1984, p. 49).

According to Moore et al. (1974) the origin of the progradational sequence may be due to: 1) non-glacially induced eustatic sea level changes, 2) local changes in sediment supply, or 3) local subsidence/uplift. In light of the nearby volcanism in this unknown tectonic setting, the most plausible explanation for the apparent regression is the interplay between sediment supply and uplift/subsidence due to tectonic activity. Although these factors may also cause a eustatic sea level change, in this case relative sea-level lowering is probably a local occurrence.

Stage 3: Deposition of the Drook Formation (Fig. 6.3)

Following deposition of the Gaskiers Formation diamictite, turbidite deposition on the basin plain/fan fringe resumed (Fig. 3.24). The abrupt transition from the underlying glaciogenic sediments to inferred basin plain deposits may represent a relative sea level rise. Alternatively, the Gaskiers Formation diamictites may have moved far enough downslope to reach the basin plain (Fig. 6.2). Also, some sediment may have fallen from ablating ice rafts or bergs which broke off from the main ice mass and drifted further out into the basin. The paleoflow indicators in the lower part of the Drook Formation (approx. 100 m to 300 m) reflect a source to the northeast (Fig. 4.5), probably relict from that which prevailed during Mall Bay deposition. Later during Drook deposition, a new volcanic source to the southwest provided detritus to the basin (Fig. 6.3). The overall first-order, progradational, coarsening- and thickening-upward nature of that part of the Drook Formation exposed within the study area is probably a combined result of fan progradation and basin subsidence.

## Comparison to a modern example

The volcanogenic sedimentary deposits associated with the Lesser Antilles Arc (Sigurdsson et al., 1980) are comparable to the Lower Conception Group with respect to the style of sedimentation. Although possible, the following comparison does not suggest that the lower Conception Group formed in a similar tectonic setting (i.e., a backarc basin).

Sigurdsson et al. (1980) concentrate on volcanogenic deposits from sediment gravity flows (turbidites, modified grain flows) in 100 deep-sea cores in the 2500 m-deep Grenada backarc basin. Examples of the similarities between the lower Conception Group sedimentary basin and the modern deposits of the Grenada Basin are:

1) The sediment composition in both basins is dominated by volcaniclastic sands and ash (now tuffaceous) layers supplied from volcanic islands. Grain sizes range from hemipelagics and clay-size to coarse grained material.

2) Grenada Basin sediments were deposited from modified grain flows and turbidity currents. Some deposits display "hybrid" characteristics which imply multiple flow mechanisms were operating during transport and/or deposition. These "hybrid" deposits contain crude inverse or normal grading and floating clasts. One core from the Grenada Basin contained a 2.5 m-thick, ungraded, massive, layer with some convolution and clay clasts. It is interpreted to be the product of a fluidized sediment flow, similar to some beds in the lower Conception Group (Fig. 3.19).

3) Some of the modern sediments exhibit characteristics resembling those of contourites, or, more broadly, resemble sediments affected by reworking by marine bottom currents.

4) Sediment lobes fan out between the active islands of Dominica, Martinique, St. Lucia, and St. Vincent, on the eastern flank of the Grenada Basin. The sediment for these lobes is supplied from slumps and other slope sediment failures which generate the various types of sediment gravity flows.

It is not yet known if fan channels and shelf-incised canyons play a significant role in transporting sediment to the base of slope in the Grenada Basin (S.N. Carey, pers. comm. to R.N. Hiscott, 1984). Until proven otherwise, it is assumed that channels are required to conduct sediment to the mid-fan and outer fan areas in both the Grenada Basin and in the Precambrian Conception Group Basin according to the present turbidite fan model of Walker (1976a).

## Comparison to an ancient example

The Upper Silurian to mid-Devonian deep-water volcaniclastics of the Kowmung Group in N.S.W. Australia described by Cas et al. (1981) show many similarities to the style of sedimentation exhibited by the lower Conception Group. Cas et al. (1981) do not disclose the tectonic setting of these sediments.

Sediments of the Kowmung Group have been grouped into two facies associations; viz., a mudstone association and a volcaniclastic association. The mudstone association represents the ambient, background sedimentation which is periodically interrupted by the deposition of the volcaniclastic facies association. The thick (metres to tens of metres) beds of the volcaniclastic association are sometimes amalgamated, and may exhibit diffuse layering and grading or may be ungraded. The fine grained sediment of the Bouma D and E divisions is commonly silicified, a characteristic also found in the argillites of the lower Conception Group.

The dominant transport mechanisms for the sandstones are inferred to be mass-flow processes involving grain support by fluid turbulence and dispersive pressure. Cas et al. (1981) interpret some of the cross-laminae measured in the mudstone facies association to be the products of contour currents. The effects of marine currents operating on the slope was also observed in the lower Conception Group (see Chapter 3-Facies C and D in facies association II).

Although subject to interpretation, a major contrast between this ancient example and the lower Conception Group is the suggestion of Cas et al. (1981) that the deposition of the coarsening- and thickening-upward first-order cycle in the Kowmung Group is not the product of a prograding submarine fan. Instead, they envision a blanket-like or sheet-like sedimentation style which produces a coarsening- and thickening-upward cycle by progressive replacement of the mudstone facies association by the volcaniclastic facies association. They suggest that this would occur within a setting containing a broad, submarine, volcaniclastic apron with several "sediment-influx points" (my quotes) adjacent to the subaerial volcanic rocks. However, within the Conception Group there are also thin (3m) sandstone packets (Fig. 3.2) which are deposited too rapidly to be considered the result of progradation of the entire slope (R.N. Hiscott written comm.). Unlike the interpretation of the Kowmung Group, these thinner packets are thought to be the result of lobe dynamics and fan progradation.

## Discussion

Admittedly, the seven outcrops studied in this thesis probably represent a small fraction of the entire Precambrian basin now exposed in the Avalon Zone. However, the facies patterns and paleocurrent trends suggest that this local basin analysis will be useful because it emphasizes the complexity within such a small area. Recognition of these complexities will help keep future larger-scale models of Precambrian basin evolution of the Newfoundland Avalon Zone in perspective.

# 7. CONCLUSIONS AND RECOMMENDATIONS

## Conclusions

1) Sediment transport mechanisms for deposition of the argillites and medium to very fine grained sandstones of the Mall Bay and Drook Formations of the lower Conception Group are inferred to be turbidity currents, low density, low velocity lutite flows, cohesive flows and modified grain flows. Deposition may also have involved fluid escape and grain interaction effects.

2) Seven distinct facies (A to G) were recognized. Six of these were grouped into three facies associations. The intensely faulted, chaotic nature of the seventh facies precluded a conclusive interpretation, but it may represent a debris flow deposit. Facies associations I, II, and III are interpreted to represent sedimentation in basin plain/fan fringe, lower basin-slope, and lobe/lobe fringe environments respectively, within a turbidite fan depositional system. No inner fan or upper mid-fan deposits were recognized within the study area. All facies associations represent deposition in relatively deep water below wave base, but no absolute water depth can be determined.

3) A complex paleoflow pattern was obtained from solemarks (flutes, prodmarks, groove and striation casts), straight-crested and linguoid ripples, and sandstone fabric. Paleoflow directions obtained from sandstone

fabric agree well with those obtained from field measurements. With the exception of an anomolous paleoflow direction at False Cape (section 5), an overall northeast-southwest trend is observed. In lower basin-slope deposits at False Cape, marine currents travelled parallel to the inferred shoreline (not necessarily contour currents) producing an easterly paleoflow pattern transverse to the inferred turbidite fan axis.

4) Lower Conception Group sandstones are volcaniclastic and their average composition is 39.6% plagioclase (albite), 36.6% basaltic and rhyolitic volcanic rock fragments in approximately equal proportions, 14.2% matrix (including palagonite), 8.2% quartz, 1.4% carbonate and trace amounts of epidote and heavy minerals (e.g., zircon). Based on electron microprobe data, the albitization of feldspars is pervasive. The dominant source terrane was volcanic, probably of rhyolitic and basic composition, in about equal proportions. Sedimentary rock fragments reflect the presence of a sedimentary source, perhaps equivalent to the (Helikian?) platformal sequence (Green Head and George River Groups) of Nova Scotia and New Brunswick.

5) There is no compelling evidence in favour of any particular tectonic setting. When the lower Conception Group sandstones are plotted on the QFL and $Q_mFL_t$ diagrams

of Dickinson and his co-workers, they occur in the region of transitional island arcs. They have a composition similar to inferred arc-derived ancient flysch sequences such as the Franciscan and Great Valley Sequences of California and the New Zealand Forearc Assemblage. The petrography of the lower Conception Group is also similar to compositional data from sands associated with volcanic arcs in modern subduction zones. Despite these similarities in sandstone composition, other features characteristic of classical Andean-type margins are absent from the regional geologic framework. No absolute sedimentation rate could be determined because accurate age span and thickness of the lower Conception Group are not known.

6) The depositional history of the lower Conception Group can be simplified into three stages. Initial basin filling (Mall Bay Formation) involved sediment supply from a northeasterly volcanic source. In stage 2, flysch sedimentation was interrupted by deposition of glaciogenic debris flows (Gaskiers Formation) on a submarine slope. Later, during deposition of the Drook Formation, another volcanic source of similar composition was introduced to the basin to the southwest. The facies produced are very similar to those formed during deposition of the Mall Bay Formation. The location of the sedimentary source is also to the southwest. The first-order coarsening- and

thickening-upward cycles observed in the Mall Bay and Drook Formations are probably the combined result of local sediment supply and basin uplift/subsidence rather than eustatic sea-level fluctuations.

## Recommendations

Several suggestions for future work on late Precambrian sediments in the Newfoundland Avalon Zone derive from this thesis. Firstly, a sedimentological and paleocurrent study of the remaining two formations (i.e., the Briscal and Mistaken Point Formations) in the Conception Group would help to further define the nature and size of the turbidite fan system. According to the fan model used in this thesis, the upper parts of the Conception Group should contain deposits of the upper mid-fan and inner fan. Further study of the Drook Formation in other localities would also be useful. Much of the Drook Formation is well exposed to the west of this study area, along the southern shore of the Avalon Peninsula. This continuous stratigraphic exposure might enable correlations to be made and also provide information on vertical and lateral facies changes.

Secondly, the overlying St. John's Formation (Fig. 2.4) requires a detailed sedimentological study. King (1982, p. 5, 17-19) suggests that the St. John's Group may represent pro-delta or delta front deposits. The sedimentary structures found in the fine grained sediments of the St. John's Group are also formed in slope depositional settings. A more detailed account of this part of the Precambrian basin fill is needed.

Finally, a comparison study of sediments equivalent to the Conception Group (i.e., Connecting Point Group) in the central and western parts of the Newfoundland Avalon Zone would give an idea of regional facies characteristics and paleogeography within the basin. Also, a comparison between the overlying units (i.e., the Musgravetown, Hodgewater, and Signal Hill/St. John's Groups) would eventually document the entire record of Precambrian basin filling in the Newfoundland Appalachians.

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# APPENDIX I

# LEGEND

## LITHOLOGY

- matrix supported shale clast conglomerate
- sandstone
- siltstone
- argillite
- wavy bedded sandstone and argillite
- lenticular bedded sandstone and argillite

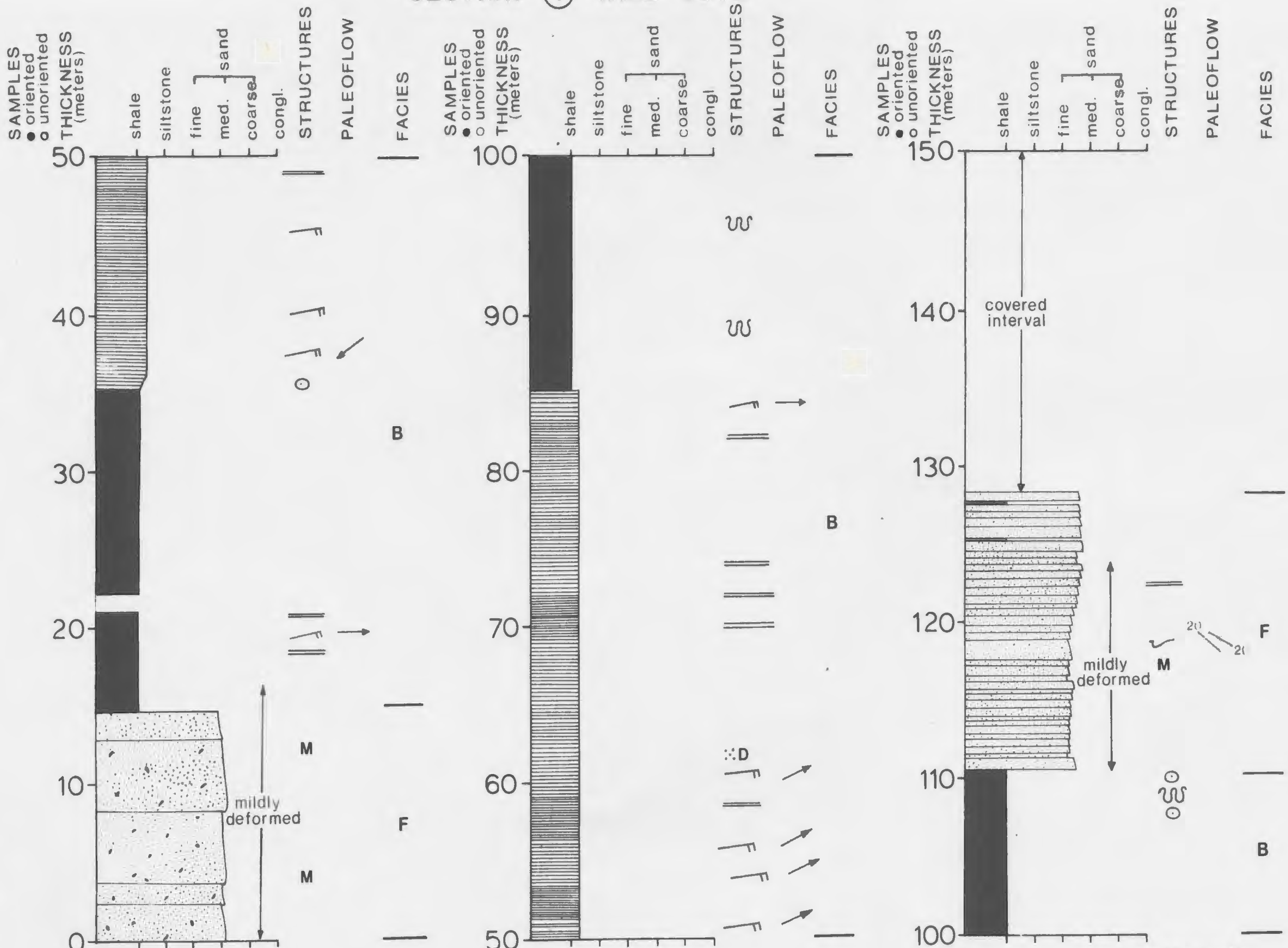
## STRUCTURES AND CONSTITUENTS

- M massive
- current ripples
- linguoid ripples
- climbing ripples
- parallel laminae
- vague parallel laminae
- undulating laminae
- grooves, flutes — solemarks
- normal grading
- n = number of measurements; solemark azimuth
- carbonate concretion
- convolute laminae
- ball and pillow
- shale clasts (relative size indicated)
- ∴ D diagenetic horizon
- scour and fill
- rippled top bedding surface
- pillar, dike — fluid escape structures
- amalgamated bedding
- n (if n > 100, no value of n is indicated) ripple azimuth

* covered interval = outcrop obscured by talus, grasses, etc. ∴ no observations

* inaccessible = outcrop clearly visible, general characteristics observed, no detailed examination

## SECTION ① WILD COVE

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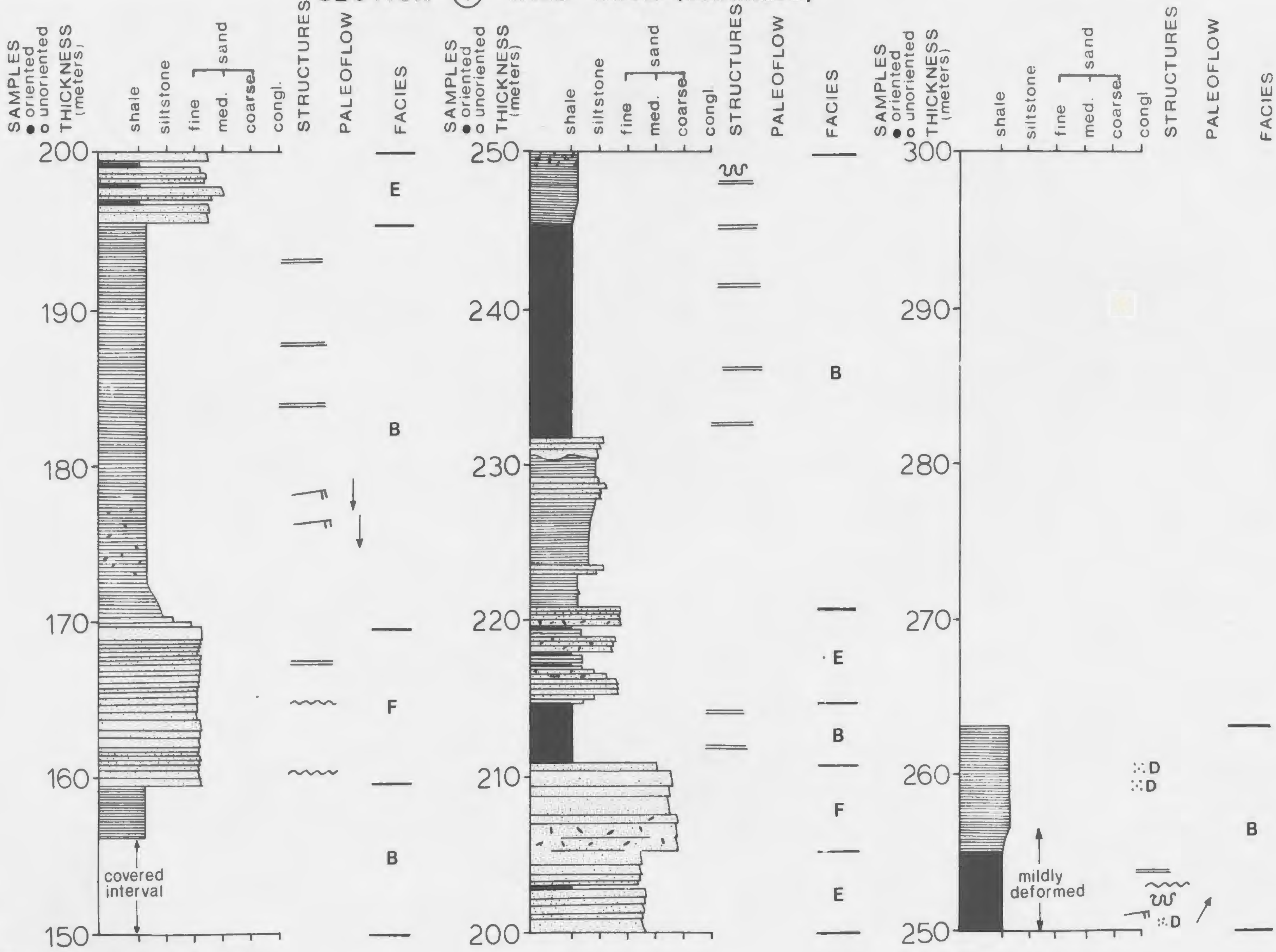

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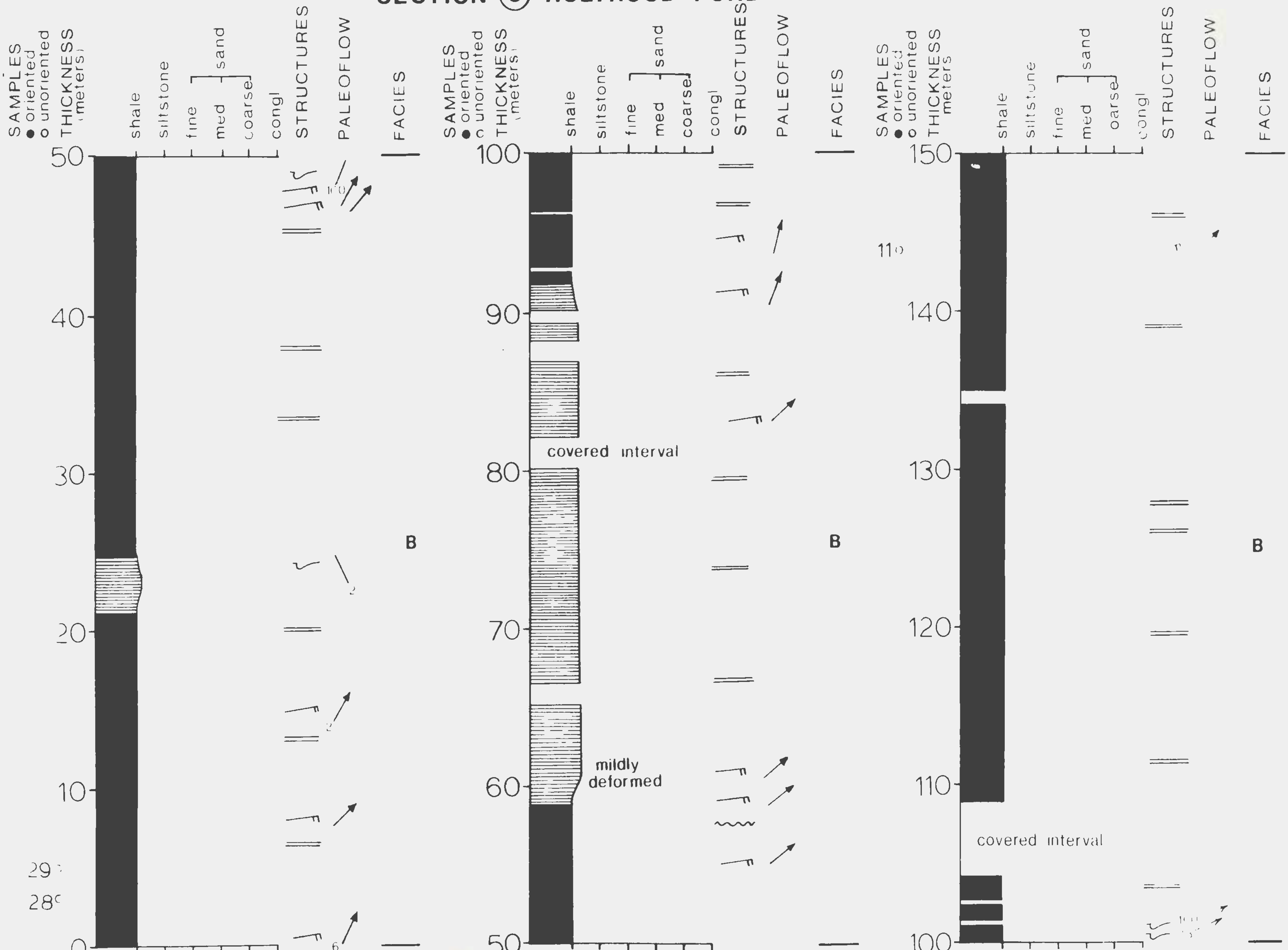


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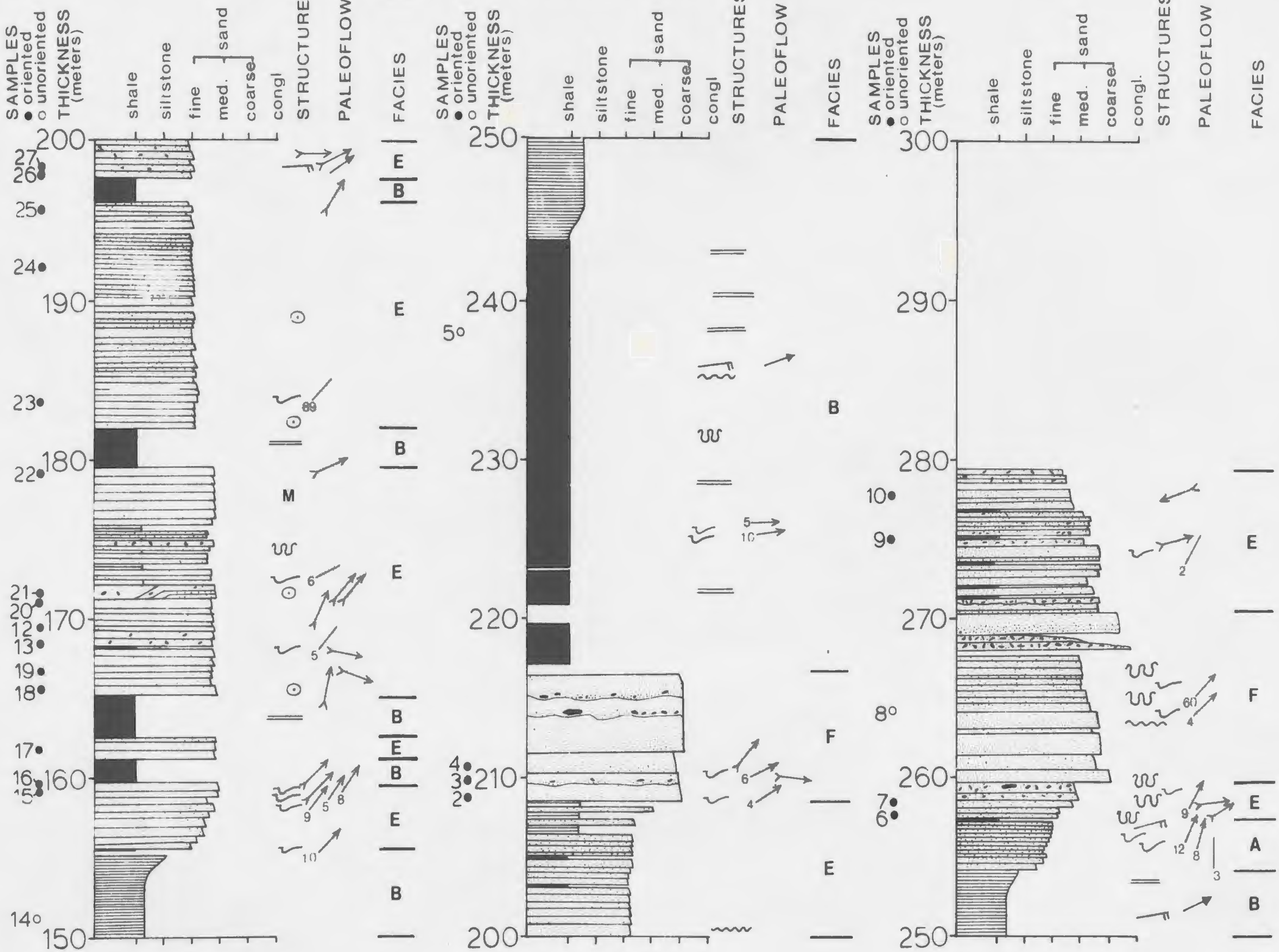

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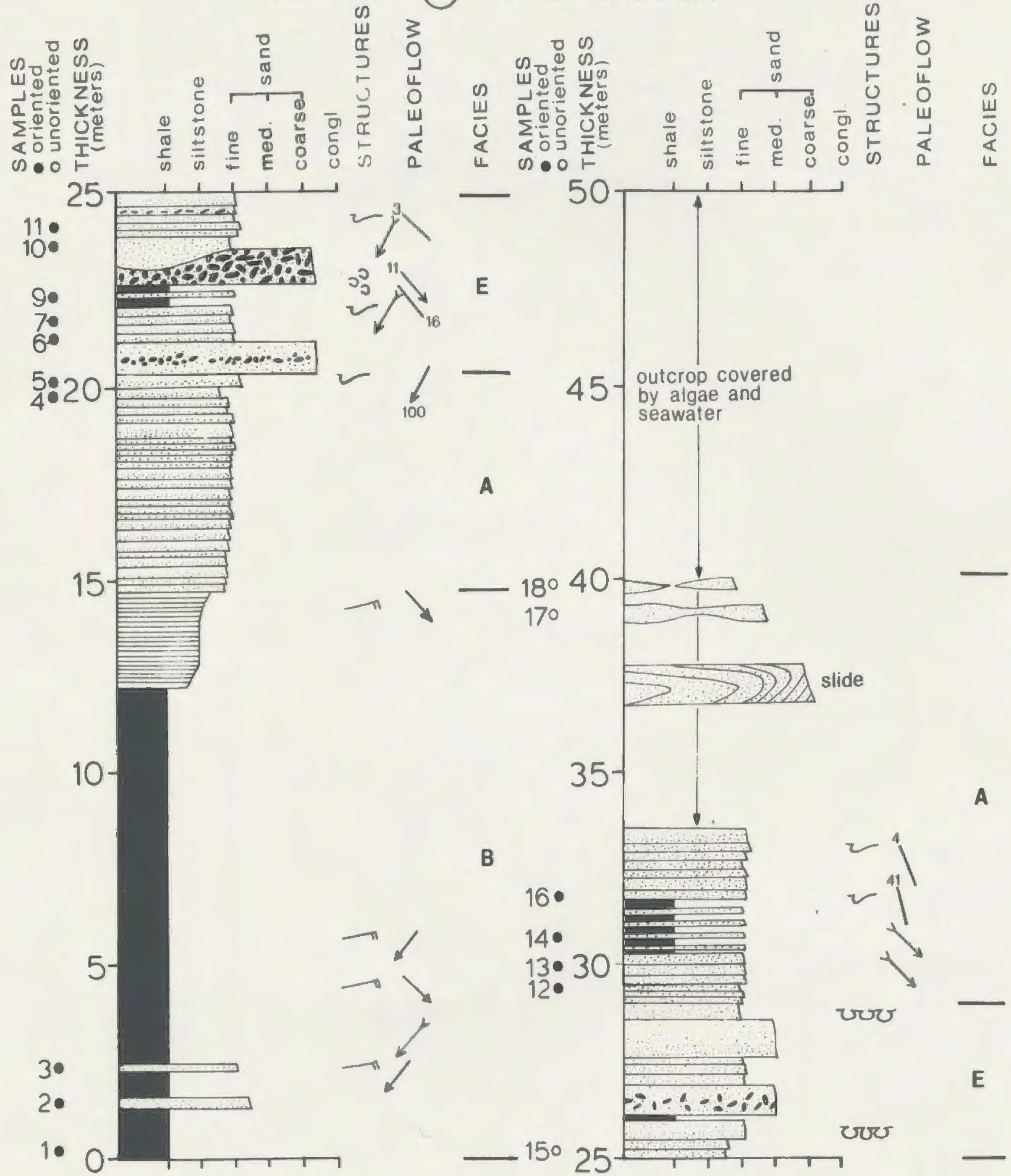

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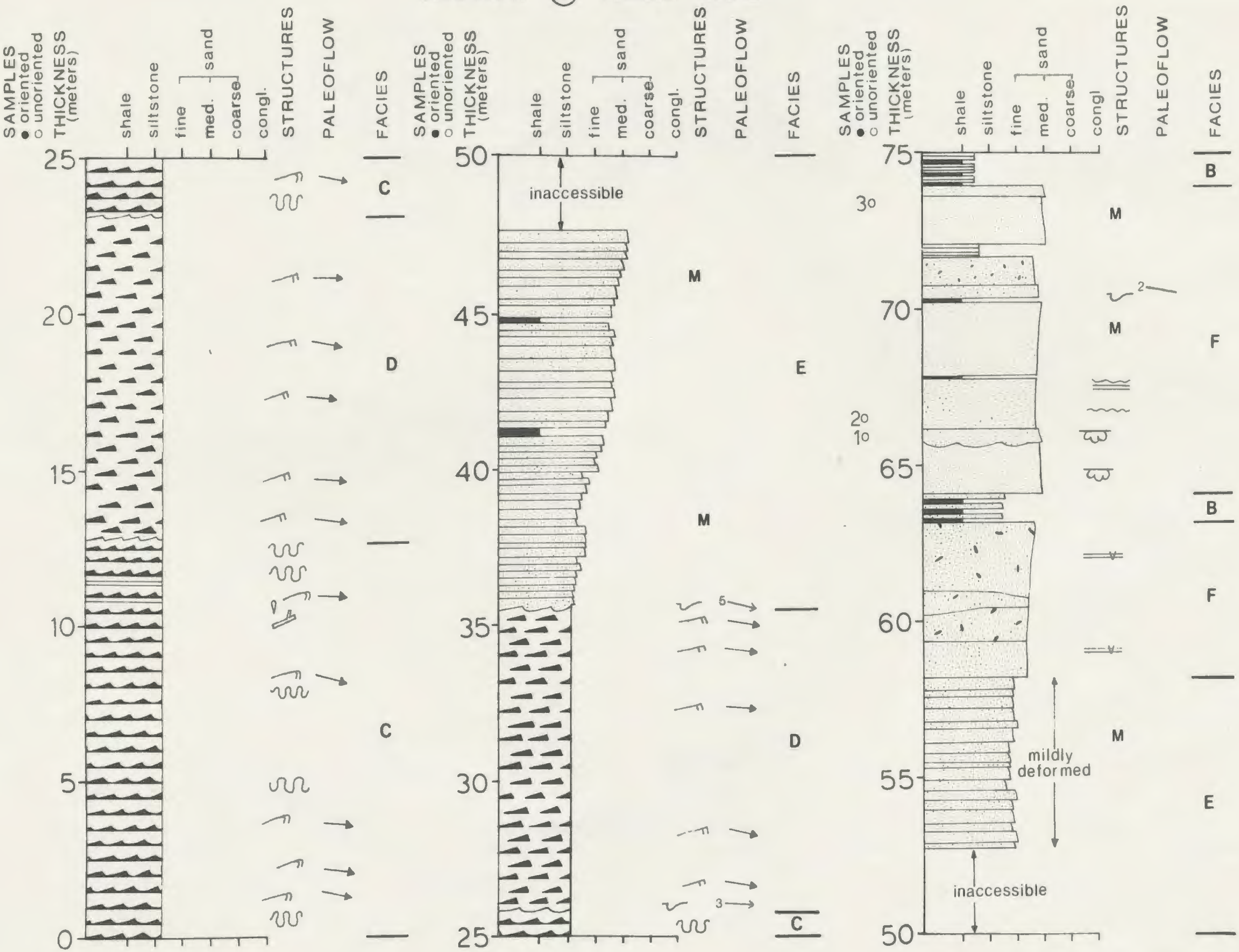


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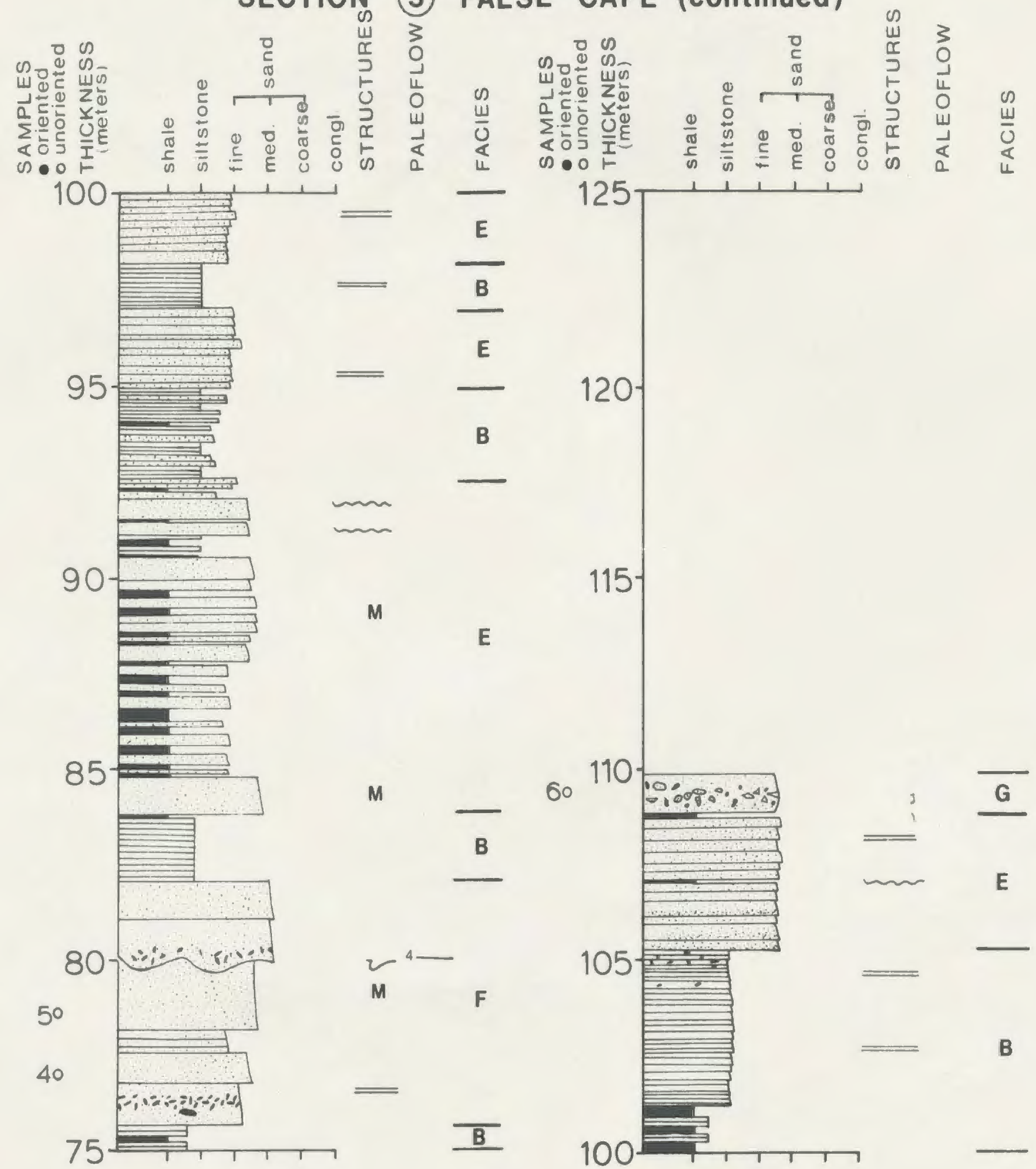

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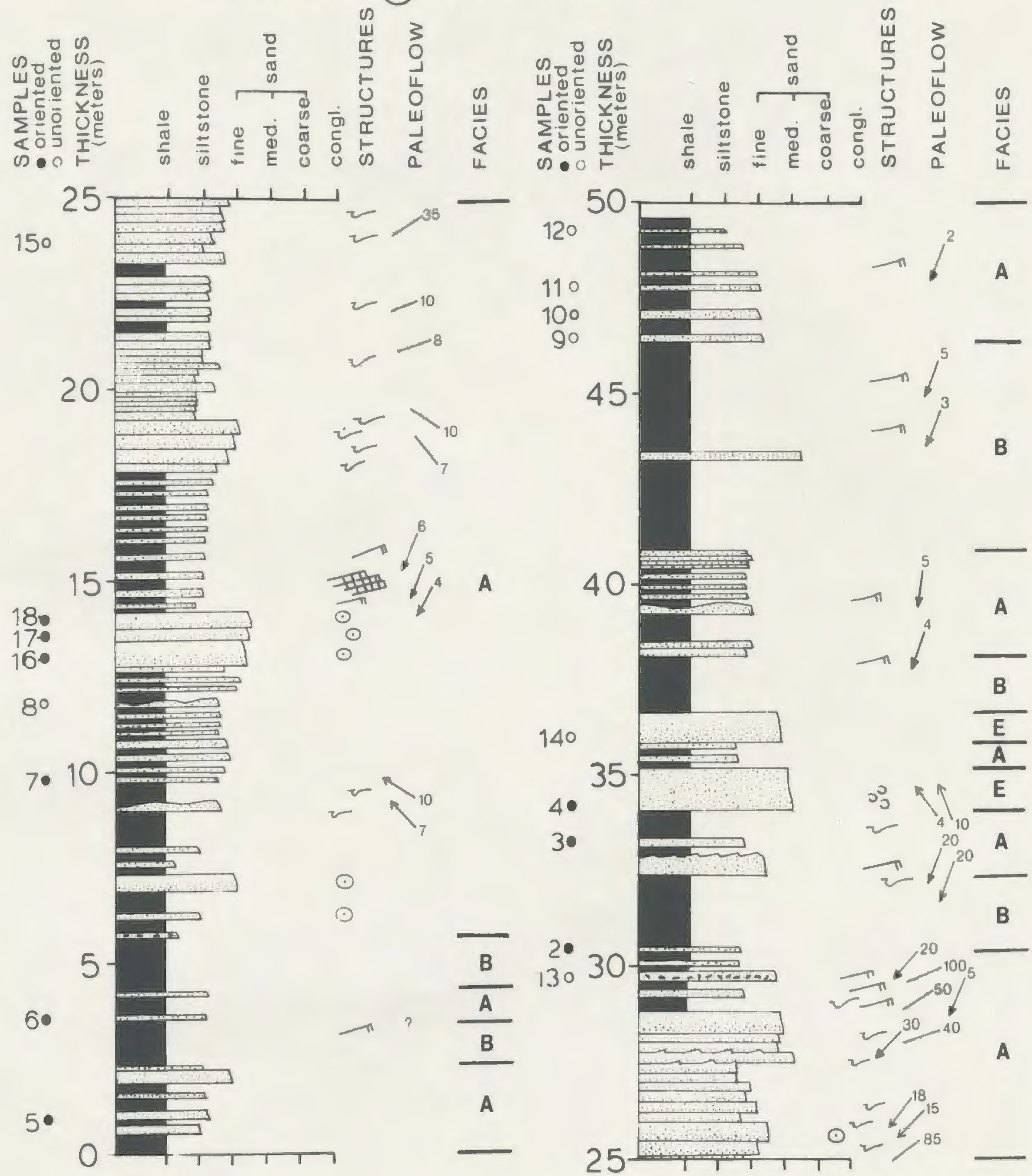


# APPENDIX II

## Descriptive nomenclature

A bed is a lithologically homogeneous sedimentary unit, or a unit exhibiting only minor internal lithological variation, which was originally deposited on a horizontal surface.

A sedimentary unit which is not broken by recognizable internal erosion surfaces or other discontinuities defines a layer (Wood and Smith, 1959; Stanley and Bouma, 1964; Ricci Lucchi, 1969). Layers represent deposition from one sedimentation event. A layer need not be structurally homogeneous, provided that the boundaries between different structural divisions are gradational. A bed can be comprised of one or more layers. When more than one layer is contained in a bed, it is termed an amalgamated bed. The boundary between layers in an amalgamated bed, constitutes an amalgamation surface.

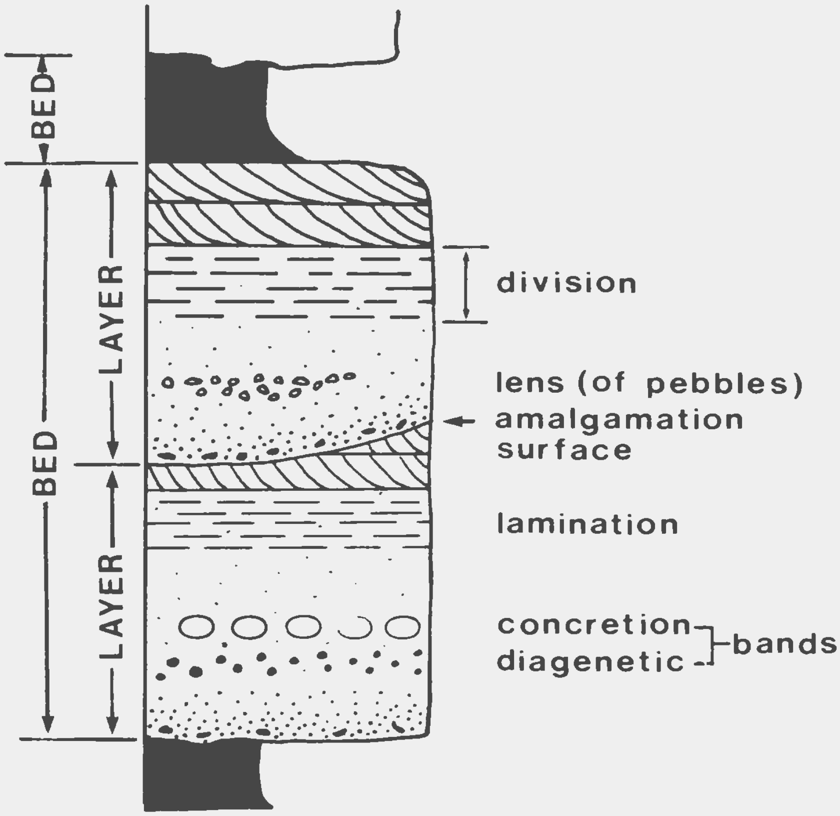
A division is a structurally homogeneous part of a layer. In turbidites, the 5 divisions of Bouma (1962) are usually employed or referred to. Recent studies of silt and mud turbidites has led to the classification of more divisions (Kuenen, 1964; Hesse, 1975; Piper, 1978).

A horizontally continuous part of a layer which differs from the remainder with respect to colour, presence of pebbles, concretions, or any anomolous characteristic, defines a band. For example, in this thesis, the terms used in facies description include concretion bands, diagenetic bands and colour bands.

A lamination is a relatively distinct alternation of materials which differ from each other with respect to grain size or composition on a scale of one cm or less.

APPENDIX II

## NOMENCLATURE




(in part, after Hiscott, 1977)

| | | |
|---|---|---|
| thinly bedded | =0.03–0.10 | meters |
| medium-bedded | =0.10–0.30 | " |
| thickly bedded | =0.30–4.00 | " |

# APPENDIX III

## Feldspar staining method

Uncovered thin sections must first be polished with 1200-grit abrasive. After the photomicrographs were taken, and before point counting, thin sections were stained with sodium cobaltinitrite and amaranth for determining the presence of potassium feldspar and plagioclase, respectively. Staining procedures have often been found to require additional comment by those who use them (Houghton, 1980). The following procedure is substantially modified from Norman (1974). In fact, Norman's (1974) procedure was found to be unsuccessful unless modified as described below.

1) Etch thin section for 30 seconds over 52 to 55% hydrofluoric acid (HF). A plastic ice cube tray with one segment filled almost to the top with HF was found to be suitable for etching. Do not rinse the thin section after etching. The HF should be changed every 45 minutes or so to ensure that it remains fresh.

2) Immerse the section in saturated sodium cobaltnitrite solution for 60 seconds. Rinse in a large beaker of tap water. Rinse again in a second beaker. Allow to dry.

3) Re-etch the section for 10 s over HF. Do not rinse.

4) Immerse the section for 15 seconds in saturated barium chloride solution. Gently rinse once in a beaker of tap water.

5) Using a dropper, cover the section with a saturated solution of amaranth (F.D. and C. red no. 2) for 15 seconds. Try to make certain that the solution is distributed evenly. Rinse in a beakerof tap water, then run section under a gentle stream of tap water. Allow to dry and apply a clear enamel spray (don't use glass cover slips; the mounting medium used with the glass cover slips tends to dissolve and disperse the stain).

Potassium feldspar will stain yellow, and plagioclase will stain pink or red. The red colour is more intense with increasing Ca/Na ratio of the feldspar. Pure albite is difficult to stain and may require substitution of calcium chloride for barium chloride in step 4) above.