MIDDLE AND UPPER KEWEENAWAN SILICICLASTICS OF THE LAKE SUPERIOR BASIN

By

IAN EDWARD TAYLOR, B.A.

A Thesis

Submitted to the School of Graduate Studies

in Partial Fulfilment of the Requirements

for the Degree

Doctor of Philosophy

McMaster University

MIDDLE & UPPER KEWEENAWAN SILICICLASTICS, LAKE SUPERIOR BASIN

DOCTOR OF PHILOSOPHY (1987)
(Geology)

McMASTER UNIVERSITY
Hamilton, Ontario

TITLE: Middle and Upper Keweenawan Siliciclastics of the Lake Superior Basin

AUTHOR: Ian Edward Taylor, B.A. (Cambridge University)

SUPERVISOR: Dr. G.V. Middleton

NUMBER OF PAGES: xx, 319

ABSTRACT

Middle and Upper Keweenawan sediments of the Lake Superior Basin were deposited by braided streams on large, gently sloping, fluvially-dominated alluvial fans.

The conglomeratic facies association (upper fan deposits) dominates the Keweenawan succession. The conglomerates are coarse and poorly sorted with fabrics between clast-supported and (sandy) matrix-supported. Sorting is so poor that the clast-matrix division is arbitrary, so the term c-m supported (from clast and matrix supported) is used. The conglomerates were deposited on longitudinal bars by simultaneous deposition of sand, granules, pebbles, cobbles and boulders. This style of deposition reflects high sediment yields and flashy stream behaviour resulting from the lack of terrestrial plants. Intercalated sandy lenses with low-angle cross-stratification were deposited in channels by antidunes or poorly developed dunes.

Conglomerates of the conglomerate-sandstone facies association (mid-fan) are better sorted and imbricated and show some planar cross-stratification, indicating that longitudinal and transverse bars were present. Intercalated sandstones are trough cross-stratified and flat-bedded.

The sandy facies association (lower fan) comprises trough cross-stratified sandstones (channel deposits) and flat bedded and rippled sandstones (bar-top deposits).

The abundance of trough cross-stratified sandstones implies that slopes on the Keweenawan fans were lower than those indicated by alluvial fan facies models. More attention should be paid to large, gently sloping, fluvially-dominated alluvial fans.

Abrupt transitions from proximal fan conglomerates to lacustrine sediments reflect volcanic damming. More gradual and poorly-defined vertical facies transitions resulted from changes in tectonism.

The Keweenawan streams were ephemeral, and desiccation features are abundant. These conditions enabled the development of fields of small transverse aeolian dunes, giving rise to large-scale cross-stratified well sorted sandstones. Aeolian palaeocurrents may indicate topographically constrained palaeo-trade winds.

ACKNOWLEDGEMENTS

Many thanks are due to Dr. Gerard Middleton, whose supervision has at all stages provided valuable guidance and insight.

Funding to meet field expenses and fees payable by foreign students in Canada has come from many sources. Trinity College, Cambridge, generously provided a Travelling Studentship. McMaster University provided support in the form of a Dalley Fellowship, and Tuition Supplements and Scholarships. The Provincial Government provided an Ontario Graduate Scholarship. Field expenses have been supported by an N.S.E.R.C. operating grant, and by a G.S.A. scholarship.

Of those that have provided assistance during the course of this study, only a few can be mentioned here. Simon Hopkins and Karen Dam were cheerful and efficient field assistants. Jack Whorwood's many photographic talents have contributed much to the Figures herein. Len Zwicker produced fine thin sections from sometimes poor samples.

Dr. Roger Walker and all the graduate sedimentologists at McMaster have participated in fruitful and encouraging discussions.

Most of all, thanks to my wife, Lynn, to whom I dedicate this work.

TABLE OF CONTENTS

CHAPTER		PAGE
1	INTRODUCTION	1
	1.1 Primary purpose of this study	1
	1.2 Further aims of this study	5
	1.3 Scope and method of this study	8
2	REGIONAL GEOLOGY AND PREVIOUS SEDIMENTOLOGICAL STUDIES	11
	2.1 Structure	11
	2.2 Stratigraphy	18
	2.3 The Keweenawan lavas	21
	2.4 Previous sedimentological studies of the Middle and Upper Keweenawan	24
3	DESCRIPTION AND INTERPRETATION OF FACIES	32
	3.1 Introduction	32
	3.2 Facies descriptions	32
	3.2.1 Conglomerate facies	33
	3.2.1.1 Massive, c-m supported, coarse conglomerate, C-1	34
	3.2.1.2 Massive, matrix-supported, coarse conglomerate, C-2	36
	3.2.1.3 Imbricated, c-m supported or clast-supported, coarse conglomerate, C-3	38
	3.2.1.4 Massive, c-m supported or clast-supported, fine conglomerate, C-4	39

3.2.1.5 Imbricated, clast-supported, fine conglomerate, C-5 39

3.2.1.6 Imbricated, matrix-supported, fine conglomerate, C-6 41

3.2.1.7 Small scale planar cross-stratified fine conglomerate, C-7 41

3.2.1.8 Medium scale planar cross-stratified fine conglomerate, C-8 42

3.2.1.9 Large scale planar cross-stratified fine conglomerate, C-9 43

3.2.1.10 Trough cross-stratified fine conglomerate, C-10 45

3.2.1.11 Horizontally stratified, fine conglomerate, C-11 47

3.2.2 Sandstone facies 47

3.2.2.1 Upper flat bed sandstone, S-1 47

3.2.2.2 Upper flat bed pebbly sandstone, S-2 48

3.2.2.3 Horizontally stratified sandstone, S-3 48

3.2.2.4 Trough cross-stratified sandstone, S-4 50

3.2.2.5 Trough cross-stratified pebbly sandstone, S-5 55

3.2.2.6 Trough cross-stratified very pebbly sandstone, S-6 55

3.2.2.7 Mixed trough cross-stratified and flat bedded sandstone, S-7 56

3.2.2.8 Mixed trough cross-stratified and flat bedded pebbly sandstone, S-8 56

3.2.2.9 Planar cross-stratified sandstone, S-9 56

3.2.2.10 Planar cross-stratified pebbly sandstone, S-10 58

3.2.2.11 Rippled sandstone, S-11 58

3.2.2.12 Mixed rippled and upper flat bedded sandstones, S-12 62

3.2.2.13 Large scale cross-stratified sandstone, S-13 62

3.2.2.14 Massive sandstone, S-14 63

3.2.2.15 Thin graded sandstone beds, S-15 63

3.2.3 Siltstone facies 65

3.2.3.1 Massive siltstone, St-1 66

3.2.3.2 Parallel laminated siltstone, St-2 66

3.2.3.3 Rippled siltstone, St-3 67

3.2.3.4 Mixed rippled and parallel laminated siltstone, St-4 67

3.2.3.5 Pebbly siltstone, St-5 68

3.2.4 Stromatolite-oncolite beds 68

3.3 Description and interpretation of facies associations 71

3.3.1 The conglomeratic facies association 71

3.3.2 The conglomerate-sandstone facies association 75

3.3.3 The sandy facies association 78

3.3.4 Rippled sandstone facies association 84

3.3.5 The silty facies association 86

3.3.6 Interpretation of the large scale cross-stratified sandstones (S-13) 87

3.3.7 Interpretation of the stromatolite-oncolite beds 88

4 DESCRIPTION AND DISCUSSION OF MEASURED AND SURVEYED SECTIONS 89

4.1 Methods of outcrop description and problems encountered 90

4.2 Measured and surveyed sections within the conglomeratic facies association 96

4.3 Measured and surveyed sections within the conglomerate-sandstone facies association 110

4.4 Sections within the sandy facies association 138

4.5 Sections within the rippled sandstone facies association 145

4.6 Section within the silty facies association 148

5 PRESENTATION AND DISCUSSION OF PALAEOCURRENT DATA 152

5.1 The data 152

5.2 The regional picture 152

5.2.1 The relationship of fluvial palaeocurrents and lava palaeoflow directions 154

5.2.2 The Copper Harbor Formation 159

5.2.3 The Nonesuch and Freda Formations 163

5.3 Large scale studies of palaeocurrent data 165

5.3.1 Sandstone Falls 165

5.3.2 Five Mile Point 167

6 PETROGRAPHY 171

6.1 Presentation and discussion of grain size data 171

6.1.1 Grain size distributions of conglomerate beds 172

6.1.2 Use of grain size data to distinguish aeolian and fluvial sandstones 179

6.1.3 Grain size trends and bed thickness - maximum clast size relationships 187

6.2 Diagenetic features of sandstones of the Copper Harbor Formation 192

6.2.1 Previous work; aims of this study 192

6.2.2 Methods of study 193

6.2.3 Sample compositions 195

6.2.4 Diagenetic histories of the non-calcareous sandstones (IT2, IT4 and IT15) 199

6.2.4.1 Compaction 199

6.2.4.2 Red staining 201

6.2.4.3 Epidote and chlorite cementation 204

6.2.4.4 Feldspar and quartz overgrowth 215

6.2.4.5 Calcite cementation 217

6.2.5 Diagenetic histories of the oncolithic sandstones (IT5 and IT6) 221

6.2.6 General conclusions regarding diagenesis 225

7 COMPARISON OF THE KEWEENAWAN SEDIMENTS WITH MODERN AND ANCIENT ALLUVIAL FAN AND BRAIDED RIVER SEDIMENTS 231

7.1 Comparisons with arid-region alluvial fans 231

7.2 Comparisons with humid-region alluvial fans 240

7.3 Comparisons with braided fluvial systems 260

7.4 Appropriateness of the alluvial fan - braided river distinction and the climatic division of alluvial fan types 268

8 TECTONIC, VOLCANIC AND CLIMATIC CONTROLS ON KEWEENAWAN ALLUVIAL FAN SEDIMENTATION 274

8.1 Some general remarks concerning alluvial fan behaviour 274

8.2 Evidence for tectonic and volcanic control of Keweenawan sedimentation 287

8.2.1 Basin-fill sequences 287

8.2.2 Volcanically controlled sequences 288

8.2.3 Small tectonically controlled sequences 289

9 SYNTHESIS AND CONCLUSIONS 294

REFERENCES 305

LIST OF FIGURES

FIGURE		PAGE
2.1	Middle and Upper Keweenawan geology of the Lake Superior basin	12
2.2	A generalised stratigraphy of the Lake Superior basin, modified from White (1966a)	13
2.3	Middle and Upper Keweenawan geology of the Keweenaw Peninsula, Upper Michigan	14
2.4	Outcrop area of Middle Keweenawan basalts in the Nipigon Bay area of Ontario	15
2.5	Geology of Isle Royale, Michigan	15
3.1	View of bedding plane of massive, c-m supported, coarse conglomerate (facies C-1)	35
3.2	Massive, c-m supported, coarse conglomerate (facies C-1)	35
3.3	Sandstone lens (bleached white) within massive, c-m supported, coarse conglomerate (facies C-1)	37
3.4	Imbricated, c-m supported or clast-supported, coarse conglomerate, and imbricated, clast-supported, fine conglomerate (facies C-2; C-3)	37
3.5	Shelter structures in massive, c-m supported or clast-supported, fine conglomerate (facies C-4)	40
3.6	Channel fill of massive, clast-supported, fine conglomerate (facies C-4)	40
3.7	Medium scale planar cross-stratified fine conglomerate (facies C-9), passing laterally into planar cross-stratified pebbly sandstone (facies S-10)	44
3.8	Large scale planar cross-stratified fine conglomerate (facies C-9)	44

3.9 Trough cross-stratified fine conglomerate (facies C-10) 46

3.10 Symmetrical gravel waves overlying horizontally stratified fine conglomerate (facies C-11) 46

3.11 Upper flat bed sandstone (facies S-1) 49

3.12 Mudchips on an erosive surface within the horizontally stratified sandstone facies (facies S-3) 51

3.13 A wave rippled surface within the horizontally stratified sandstone facies (facies S-3) 51

3.14 A bubble sandstone surface within the horizontally stratified sandstone facies (facies S-3) 52

3.15 Wrinkle marks within the horizontally stratified sandstone facies (facies S-3) 52

3.16 Desiccated mud drape within the horizontally stratified sandstone facies (facies S-3) 53

3.17 Rain prints on mud draped current ripples within the horizontally stratified sandstone facies (facies S-3) 53

3.18 Trough cross-stratified sandstone (facies S-4) 54

3.19 Trough cross-stratified pebbly sandstone (facies S-5) 54

3.20 Planar cross-stratified sandstone (facies S-9) 57

3.21 Planar cross-stratified sandstone (facies S-9) 57

3.22 Climbing ripples within the rippled sandstone facies (facies S-11) 59

3.23 A bedding plane within the rippled sandstone facies (facies S-11) with ripple fans 59

3.24 Ladderback linguoid ripples in the rippled sandstone facies (facies S-11) 61

3.25 Large scale cross-stratified sandstone (facies S-13) 61

3.26 Siltstone with thin graded sandstone beds (facies S-15) 64

3.27 Parallel laminated siltstone (facies St-2) 64

3.28 Plan view of stromatolite mounds 69

3.29 Oncolithic bed grains 69

3.30 Bedding surface covered with knobbly structures 80

3.31 A close-up view of the knobbly surface shown in Figure 3.30. 80

3.32 A knobbly surface showing a directional element 81

4.1 Section of the Copper Harbor Formation along-strike on the Keweenaw Peninsula 94

4.2 Index to symbols used in the stratigraphic sections 97

4.3 Stratigraphic section in the Copper Harbor Formation at Rainbow Cove, Isle Royale 98

4.4 Stratigraphic section in the Copper Harbor Formation at Horseshoe Harbor, Upper Michigan 100

4.5 Surveyed section in the Copper Harbor Formation, east of Dan's Point, Upper Michigan 102

4.6 Surveyed section of interflow sediments at Mamainse Point, Ontario. 105

4.7 Photograph of the uppermost very coarse conglomerate bed shown in Figure 4.6 106

4.8 Surveyed section and photograph of interflow sediments at Mamainse Point, Ontario 107

4.9 Stratigraphic section in the Copper Harbor Formation at Sandstone Falls on the Black River, Upper Michigan 111

4.10 Surveyed section in the Copper Harbor Formation at the base of the section shown in Figure 4.9 113

4.11 Stratigraphic section in the Copper Harbor Formation at Rainbow Falls on the Black River, Upper Michigan 116

4.12 Correlation of sections in the Copper Harbor Formation near Dan's Point, Upper Michigan 118

4.13 Aerial view of Dan's Point section 120

4.14 Large scale section of the thickest section near Dan's Point 122

4.15 The western stratigraphic section in the Copper Harbor Formation above the lava flow at Five Mile Point, Upper Michigan 123

4.16 Surveyed section from just above the 15 m level of the section in Figure 4.15 125

4.17 The eastern stratigraphic section in the Copper Harbor Formation above the lava flow at Five Mile Point, Upper Michigan 127

4.18 Surveyed section at the 9 m level of the section in Figure 4.17 129

4.19 Section at the base of the Copper Harbor Formation in Eagle River, Upper Michigan 131

4.20 Section between lava flows at Saxon Falls in the Montreal River, Northern Michigan/Wisconsin 133

4.21 Cliff at Moffat Strait, Nipigon Bay, Ontario 135

4.22 Section measured above the unconformity shown in Figure 4.21 136

4.23 Small surveyed section in sediments underlying lavas of the Osler Group to the west of the north end of Moffat Strait, Nipigon Bay, Ontario. 137

4.24 The lower stratigraphic section in the Copper Harbor Formation at Union Bay, Upper Michigan 139

4.25 The upper stratigraphic section in the Copper Harbor Formation at Union Bay, Upper Michigan 140

4.26 Possible aeolian ripple translatent strata 142

4.27 Stratigraphic section in the Copper Harbor Formation at Point Houghton, Isle Royale 143

4.28 Stratigraphic section in the Copper Harbor Formation below the lava flow at Five Mile Point, Upper Michigan 144

4.29 Stratigraphic section in the Copper Harbor Formation at Francis Point on Isle Royale 146

4.30 Stratigraphic section in the Freda Formation 147

4.31 Stratigraphic section in the Nonesuch Formation 149

4.32 Stratigraphic section in sediments between lavas of the North Shore Volcanic Group, Goodharbor Bay, Minnesota 150

5.1 Palaeocurrent roses for all Keweenawan outcrops 153

5.2 Bent pipe amygdales at the base of a basalt flow 155

5.3 Supplementary palaeocurrent roses for outcrops in the Copper Harbor Formation 160

5.4 Palaeocurrent roses for sections at Union Bay and in the Black River 162

5.5 Palaeocurrent map of Sandstone Falls 166

5.6 Palaeocurrent roses for aeolian and fluvial sandstones at Five Mile Point 168

5.7 Map showing palaeowinds 170

6.1 Conglomerate grain size histograms 173

6.2 Cumulative frequency - grain size curves for three conglomerate beds plotted on log-normal probability paper 175

6.3 Cumulative frequency - grain size curves for three conglomerate beds plotted on Rosin's law probability paper. 177

6.4 Grain size - frequency curves for eight Five Mile Point sandstones 181

6.5 Photomicrograph of specimen IT28, a fluvial sandstone from Five Mile Point 182

6.6 Photomicrograph of specimen IT15, an aeolian sandstone from Five Mile Point. 182

6.7 Photomicrograph of specimen IT18, a bimodal sandstone from Five Mile Point 183

6.8 An oncolith within specimen IT16 185

6.9 An oncolith within specimen IT15 185

6.10 Oncoliths in specimen IT5 186

6.11 A badly crushed oncolith in specimen IT5 186

6.12 Bed thickness / grain size profile for conglomeratic section in Eagle River gorge 188

6.13 Grain size and bed thickness data for conglomerates in Eagle River gorge 189

6.14 Log-log plot of Grain size and bed thickness data for conglomerates in Eagle River gorge 190

6.15 Compositions of samples studied in this project with "average" Copper Harbor Formation sandstones 198

6.16 Photomicrograph of specimen IT2 showing grains replaced by calcite and feldspar cement 200

6.17 Crushed volcanic fragments and long grain contacts in specimen IT2 202

6.18 Sutured grain contact in specimen IT2 202

6.19 Iron staining and chlorite rims in specimen IT2 203

6.20 S.E.M. view of a chlorite rim in specimen IT2 203

6.21 S.E.M. view chlorite rims in specimen IT2 206

6.22 Close-up S.E.M. view of the chlorite rim shown in Figure 6.21 206

6.23 Very close-up view of the chlorite rim shown in Figures 6.21 and 6.22 207

6.24 X-ray analysis of a chlorite rim in specimen IT2 208

6.25 Typical chlorite and smectite spectra 209

6.26 X-ray spectrum from a chlorite rim, specimen IT2 210

6.27 Tiny crystal aggregates in specimen IT4 211

6.28 X-ray spectrum from the spherical crystal aggregates shown in Figure 6.27 213

6.29 Vermicular chlorite in feldspar (specimen IT2) 214

6.30 Epidote and chlorite replacement of grains in IT4 216

6.31 Feldspar grain with overgrowths in IT2 216

6.32 Cathodoluminescence of specimen IT2 218

6.33 Cathodoluminescence of specimen IT2 219

6.34 An x-ray spectrum from calcite cement in IT2 220

6.35 Cathodoluminescence of specimen IT5 222

6.36 Cathodoluminescence of specimen IT6 224

6.37 Magnetite grain showing alteration 227

7.1 Sediments exposed in the walls of the fanhead channel on Badwater Fan, Death Valley. 235

7.2 Close up of the sediments shown in Figure 7.1 235

7.3 Close up of the sediments in Figures 7.1 and 7.2 236

7.4 Gravels within Cedar Creek alluvial fan, Montana 255

8.1 Evolution of a simple two dimensional alluvial fan - playa system during a hypothetical tectonic cycle 278

9.1 A summary diagram of depositional environments on the fluvially-dominated Keweenawan alluvial fans 298

LIST OF TABLES

TABLE		PAGE
4.1	Approximate stratigraphic positions of the sections described in this study	93
6.1	Results of chi-square tests of conglomerate grain size data for closeness of fit to log-normal and Rosin distributions	178
6.2	Locations of samples chosen for study of diagenetic features	194
6.3	Modal analyses of specimens IT2, IT4, and IT15	196
6.4	Modal analyses of the calcareous sandstone specimens, IT5 and IT6	197
7.1	Flow parameters and slopes at the dune - upper flat bed transition	248

CHAPTER 1

INTRODUCTION

1.1 Primary purpose of this study

Ancient alluvial fan sediments form thick successions in many fault-bounded sedimentary basins. Identification and interpretation of ancient alluvial fan sediments relies upon facies models, which aim to summarize the depositional environments, depositional processes, and sedimentary facies types and distributions that are typical of alluvial fans. This study considers the emphasis of existing alluvial fan facies models, and questions whether these models adequately represent the range of alluvial fan types, and suggests ways in which alluvial fan facies models may be improved. The Middle and Upper Keweenawan siliciclastic sediments of the Lake Superior Basin provide an example of alluvial fan sediments that do not comfortably conform to existing alluvial fan facies models, and so provide indications as to ways in which the models may be deficient.

Alluvial fan facies models, such as those presented by Rust and Koster (1984), Nilsen (1982) or Miall (1978), depict small, steep alluvial fans containing a large

component of debris flow deposits. This emphasis reflects the influence of the large amount of work on alluvial fans in the semi-arid regions of the southwestern United States. These now classic studies often used very small fans (e.g. Hooke, 1967, studied three fans, with an average radius of approximately 2 km). Small fans are the easiest fans to survey and comprehensively describe, and because there are many examples of small rivers debouching from mountain fronts, modern specimens of such fans are abundant. However, it is likely that large modern fans will produce large volumes of sediment, prominent in the rock record. The emphasis on the fans of the southwestern United States has concerned various workers, who have suspected that climate may be an important control on alluvial fan sedimentation. However, the resulting studies of fans in other climatic zones have not succeeded in altering the bias towards small fans. Wasson (1977a,b) described fans formed during the last glacial in periglacial conditions, but these are only about 1 km in radius. Ryder (1971) studied similar periglacial fans, but these too are very small (about 1.5 km in radius according to the radial profiles presented). Kochel and Johnson (1984) worked on Holocene humid-temperate region alluvial fans, but these are even smaller (fan profiles are not presented, but average fan areas are apparently less

than 1 km^2). Whilst these studies usefully address the question of climatic control on alluvial fan sedimentation, they do nothing to rectify the bias towards small, steep fans. It is therefore not surprising that the sedimentary facies described in these studies fit fairly well into the established alluvial fan model built using information about small steep fans.

The bias of alluvial fan models towards small steep alluvial fans is of concern because this type of model does not adequately describe sedimentary processes and facies on larger, flatter fans. A number of ancient alluvial fan deposits display features that are not accounted for by such models. Sandstone is a much more important component of ancient alluvial fan successions than the alluvial fan facies models would predict. Furthermore, in many deposits much of the sandstone is trough cross-bedded (e.g. in deposits described by Mack and Rasmussen, 1984; McGowen and Groat, 1971; Nadon and Middleton, 1985; Steel and Wilson, 1975; Williams, 1969). This is significant because dunes can only form on fans where flow is relatively deep and sustained, requirements which necessitate lower gradients than those of the fan model discussed above. This also implies that the fans were probably larger than the model predicts (Williams, 1969, actually documented radiating

palaeocurrent patterns indicating that fans were 30 to 60 km in length). In addition, it is evident that on such large, gently sloping fans, debris flow deposition is unlikely to be as important as on small steep fans (although variables other than slope and size may conspire so as to create exceptions).

This problem with alluvial fan models has been addressed by previous workers. In particular, studies of glacial outwash fans by Boothroyd and Ashley (1975) and Boothroyd and Nummedal (1978) aimed to provide a model describing fans with low gradients, dominated by streamflow processes. The Kosi fan of India (Gole and Chitale, 1966; Wells and Dorr, 1987) has also been used as an example of such fan types, and has been used as the basis for a model of sedimentation ~~by~~ Parkash et al. (1980). However, the Kosi fan is so vast (160 km in radius) that for many workers it is a moot point whether or not it is an alluvial fan. It would certainly be hard to identify a fan of this type with the information available in most of the rock record. The relevance to alluvial fans of the glacial outwash studies has also been questioned, precisely because these are large and of low gradient and are formed at glacier snouts rather than at mountain fronts, so the alluvial fan reviews of Rust

and Koster (1984) and Nilsen (1982) rejected the model of Boothroyd and Nummedal (1978).

The Upper and Middle Keweenawan siliciclastic deposits of the Lake Superior Basin were chosen as the subject of this study because previous work (particularly that of White and Wright, 1960, and Elmore, 1981) showed these to be alluvial fan deposits, consisting predominantly of fluvial sediments, with a significant proportion of trough cross-stratified sandstone as well as conglomerate. As such, these sediments, well exposed around the shores of Lake Superior, provided an opportunity for detailed study of alluvial fan deposits that do not conform to the fan model embedded in the literature.

1.2 Further aims of this study

In addition to the reasons outlined above, the Middle and Upper Keweenawan siliciclastics of the Lake Superior Basin merit study in order to clarify how volcanism, aeolian transport and the Proterozoic lack of terrestrial plants can influence fan sedimentation.

The association of volcanics and alluvial fans observed in the Lake Superior Basin also occurs elsewhere in the rock record. Bluck (1978) noted that volcanism coincided

with alluvial fan deposition in the Midland Valley of Scotland during Lower Old Red Sandstone times, and concluded that alluvial fans locally developed on the margins of a volcanic pile in the centre of the valley (shown in his Figure 5). Burggraf and Vondra (1982) noted that the sediments filling the East African Rift Valley, which include alluvial fan sediments, are interbedded with pyroclastic flows and lavas, and commented that these might have created topographic barriers, or formed localised sediment sources. Neither of these studies, however, investigated in detail the sedimentological consequences of volcanic activity synchronous with alluvial fan deposition. Daniels (1982) speculated that volcanic damming may sometimes have occurred in the Lake Superior Basin during Keweenawan times, but was unable to prove this hypothesis. The present study tries to isolate the effects that volcanism had on Keweenawan alluvial fan sedimentation from the effects of tectonism and climate. Since volcanics and alluvial fan sediments are associated in other parts of the rock record, conclusions from this work have implications reaching beyond the Keweenawan of the Lake Superior Basin.

The observations of Wolff and Huber (1973) and Elmore (1981) indicate that the streams draining Keweenawan alluvial fans were ephemeral. Desiccation features are

abundant and gypsum crystal casts are present in places. The ephemeral conditions implied by these features suggest that aeolian reworking of fan sediments was at times important on the Keweenawan alluvial fans. Elmore (1981) similarly concluded that such aeolian activity was likely, but did not conclusively identify aeolian sediments. This study attempts to separate the aeolian and fluvial components of fan sedimentation as completely as possible, and to describe the sedimentary relationships of the aeolian and fluvial deposits.

In Proterozoic times, terrestrial plant life was yet to evolve, so no roots were available to stabilize catchment slopes or strengthen river banks on fans, and no foliage was present to trap fines or to prevent aeolian winnowing of fan sediments. Quite apart from probable enhancement of aeolian transport under such conditions, fluvial sedimentation is likely to have been considerably affected by the hydrological regime that must have persisted in such an environment. Plants, by binding weathering products to catchment slopes, in at least three ways significantly influence sediment and water supply to fans. Firstly, plants reduce the total amount of sediment eroded from a catchment. Secondly, in permitting the formation of thick soils, plants enable weathering of slope material to progress further

before that material is removed from catchment slopes, thereby causing a decrease in the grain size of the sediment supplied to fans. Thirdly, such soils act like sponges, soaking up precipitation and reducing the amount of runoff and increasing the period over which runoff occurs. Thus, before terrestrial plant life, rivers would have been very flashy, and whilst in spate, would have been supplied with large amounts of relatively coarse sediment. Schumm (1968) considered the effect that the evolution of plants may have had on fluvial style, and concluded that braided streams would have been relatively abundant prior to terrestrial plant life. Cotter (1978) has provided some evidence from the rock record to support this hypothesis. The present study tries to identify some of the sedimentary features and comment on the types of depositional process that can result from an environment lacking plant life.

1.3 Scope and method of this study

This study considers all of the Middle and Upper Keweenawan sediments of the Lake Superior Basin. For this sort of study only the outcrops which reveal the sedimentary structures and facies relationships are valuable. Only outcrops meeting this requirement were intensively studied.

This restriction results in a strong information bias towards the southern shore of Lake Superior (sediments of the Copper Harbor Formation). However, outcrops all around the Lake Superior Basin were visited, and information from different areas has been amalgamated as much as possible. Whilst there is variation in the sediments around the basin, in general, it is the overall similarities which are striking. This is reinforced by the marked differences with sediments of the underlying Lower Keweenawan rocks.

Outcrops were studied by measuring stratigraphic sections, and correlating between outcrops where possible. By and large, the rapid lateral variations in sediment style, which reflect the fluvial environment of deposition, made such correlation impossible. For similar reasons, stratigraphic sections were found to be poor representations of large and complex outcrops, so where possible, such outcrops were surveyed using tape and compass to produce surveyed sections showing lateral facies relationships. Palaeocurrent data of all types was collected wherever available. Outcrops and thin sections have been analysed for grain size data where it appeared that such information might provide insights regarding depositional processes or controls on such processes. This study has not been concerned with provenance, since this has received much

attention from previous workers (summarized by Daniels, 1982), and discussion of diagenesis has been restricted to aspects not dealt with by earlier studies.

CHAPTER 2

REGIONAL GEOLOGY AND PREVIOUS SEDIMENTOLOGICAL STUDIES

2.1 Structure

The Lake Superior basin (Figure 2.1) is partly filled by a thick pile of basaltic lava flows immediately overlain by a thick clastic succession comprising, in order of deposition, the Copper Harbor Formation (the main focus of this study), the Nonesuch Formation, and the Freda Formation (Figure 2.2). These rocks are the fill of a failed Keweenawan rift system, and thus all have a closely related genesis.

The Lake Superior basin overlies the mid-continent gravity high, a linear pattern of gravity anomalies that reflects the presence of basalts and gabbros which filled an abortive Keweenawan rift system (Chase and Gilmer, 1973). The gravity high has an arcuate form, extending southwest from Lake Superior as far as Kansas, and southeast from Lake Superior into Michigan and maybe as far as Tennessee (Halls, 1978). The eastern arm consists of more subdued gravity anomalies than the western arm, probably because fewer volcanics are present in the east and these are more deeply

Figure 2.1 Middle and Upper Keweenawan geology of the Lake Superior basin. Structural features are shown as heavy lines. Ticks show the direction of dip of fault planes. The syncline between Isle Royale and the Keweenaw Peninsula is known as the Lake Superior Syncline. The key to the types of shading is as for Figure 2.3.

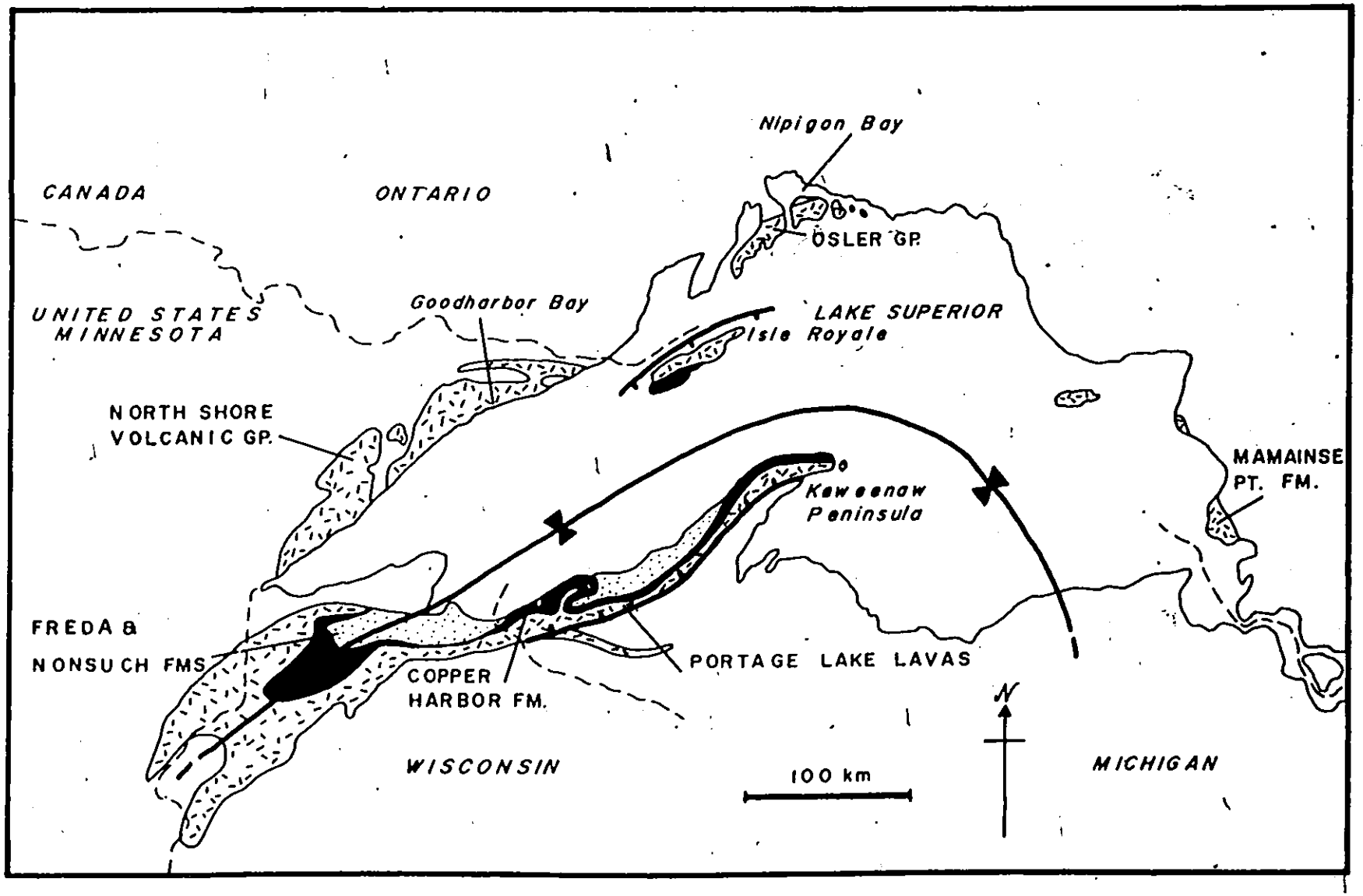

Nipigon Bay
CANADA
ONTARIO
OSLER GP.
UNITED STATES
MINNESOTA
Goodharbor Bay
LAKE SUPERIOR
Isle Royale
NORTH SHORE
VOLCANIC GP.
MAMAINSE
PT. FM.
Keweenaw
Peninsula
FREDA &
NONSUCH FMS
COPPER
HARBOR FM.
PORTAGE LAKE LAVAS
N
WISCONSIN
100 km
MICHIGAN

Figure 2.2 A generalised stratigraphy of the Lake Superior basin, modified from White (1966a). Thicknesses of the Copper Harbor Formation are discussed in the text.

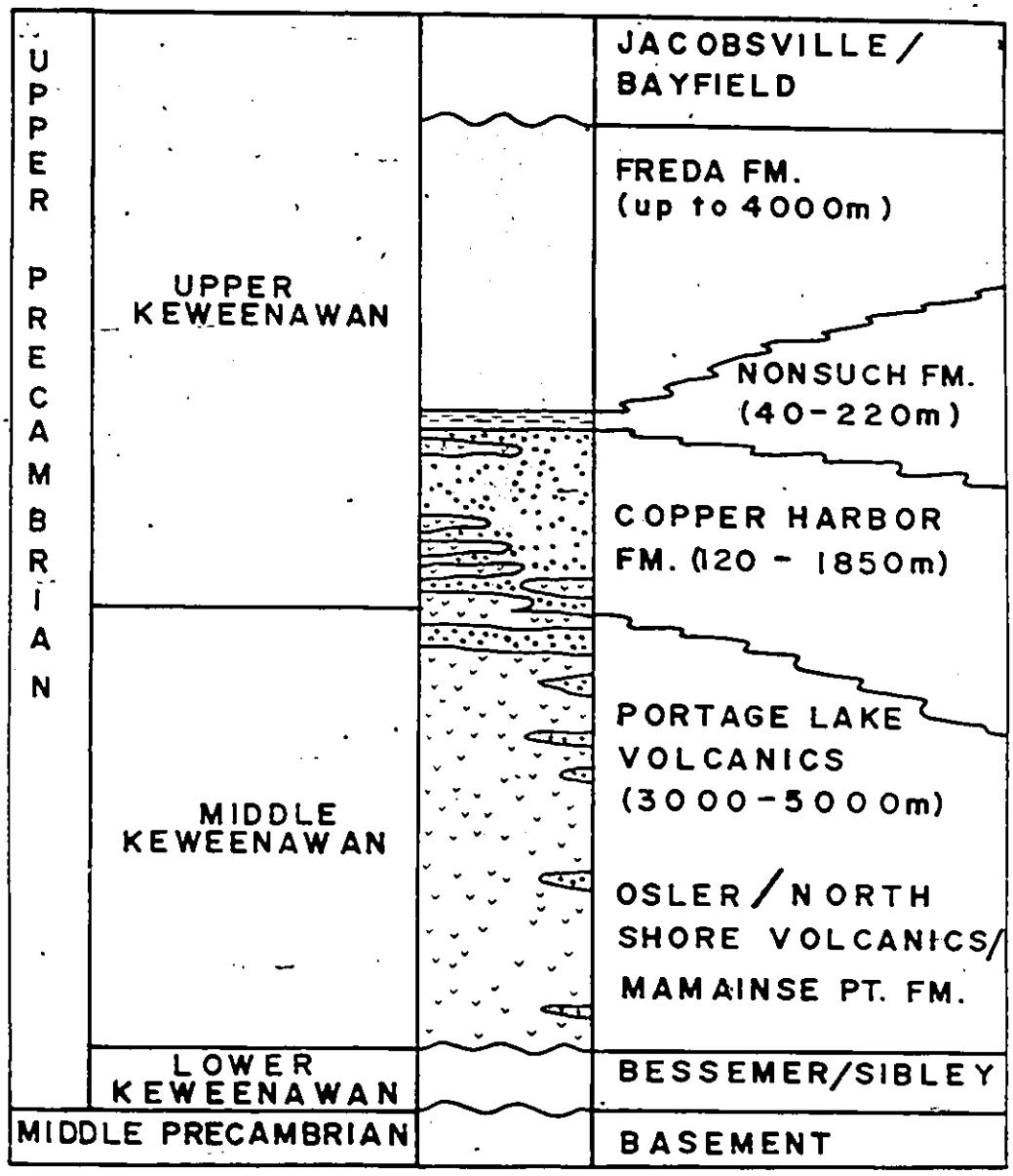
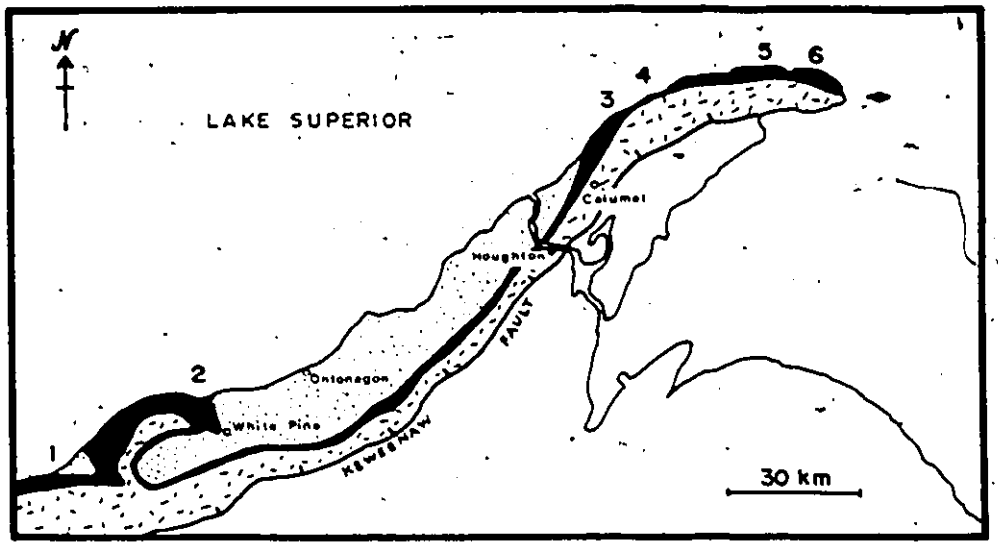

UPPER PRECAMBRIAN
UPPER KEWEENAWAN
MIDDLE KEWEENAWAN
LOWER KEWEENAWAN
MIDDLE PRECAMBRIAN
JACOBSVILLE/ BAYFIELD
FREDA FM. (up to 4000m)
NONSUCH FM. (40-220m)
COPPER HARBOR FM. (120 - 1850m)
PORTAGE LAKE VOLCANICS (3000-5000m)
OSLER/NORTH SHORE VOLCANICS/ MAMAINSE PT. FM.
BESSEMER/SIBLEY
BASEMENT

Figure 2.3 Middle and Upper Keweenawan/geology of the Keweenaw Peninsula, Upper Michigan. For a smaller scale location map, refer to Figure 2.1.



KEY

FREDA & NONSUCH FMS.

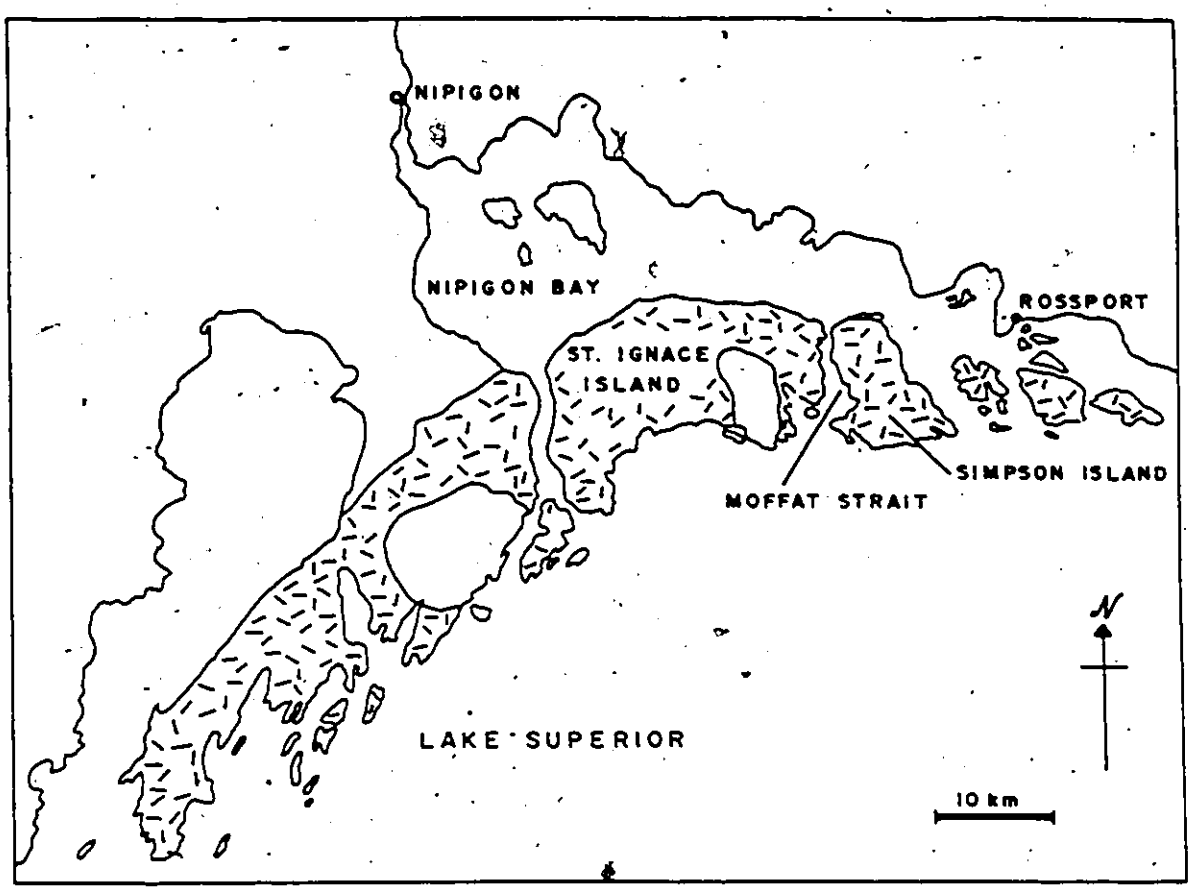
COPPER HARBOR FM.

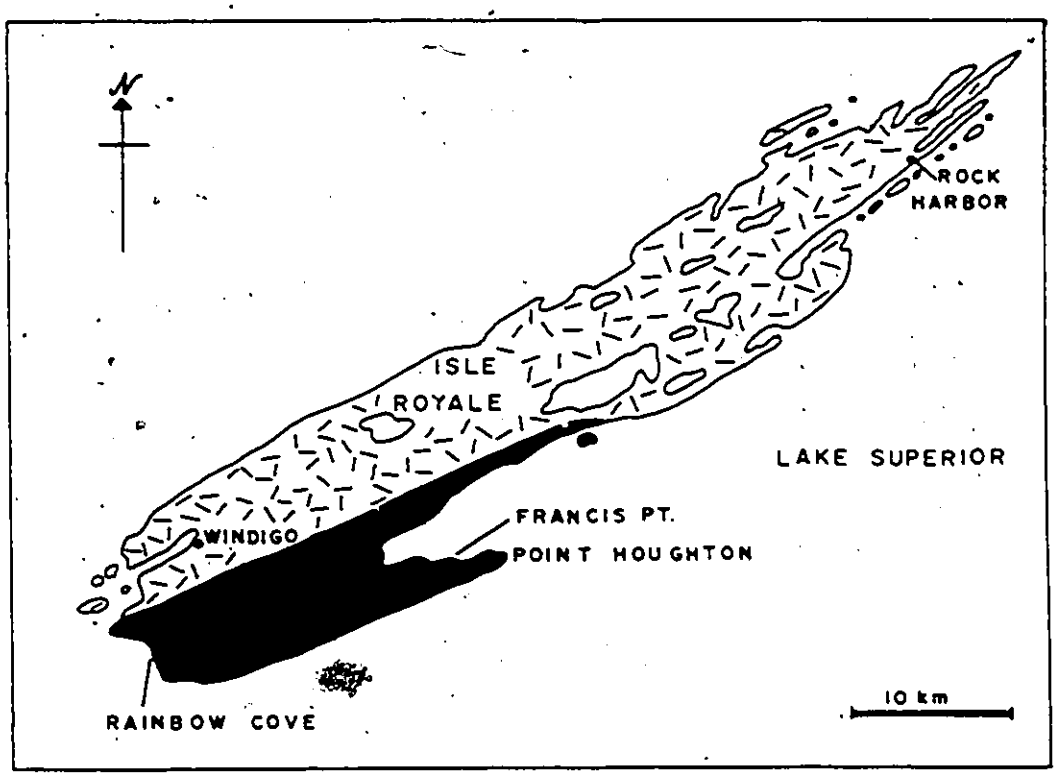
BASALTS (PORTAGE LAKE LAVAS)

1 *BLACK RIVER*
2 *UNION BAY*
3 *FIVE MILE PT.*
4 *EAGLE RIVER*
5 *DAN'S PT.*
6 *HORSESHOE HARBOR*

Figure 2.4 Outcrop area of Middle Keweenawan basalts in the Nipigon Bay area of Ontario. For a smaller scale location map, refer to Figure 2.1. Interflow sediments are thickest at the base of the lava pile, particularly at the north end of Moffat Strait.

Figure 2.5 Geology of Isle Royale, Michigan. For a smaller scale location map refer to Figure 2.1. Types of shading are as for Figure 2.3.


NIPIGON
NIPIGON BAY
ROSSPORT
ST. IGNACE ISLAND
SIMPSON ISLAND
MOFFAT STRAIT
N
LAKE SUPERIOR
10 km


N
ROCK HARBOR
ISLE ROYALE
LAKE SUPERIOR
FRANCIS PT.
WINDIGO
POINT HOUGHTON
RAINBOW COVE
10 km

buried (Hinze et al., 1982; Halls, 1978). The western arm consists of a series of sharply defined en-echelon segments, explained by Chase and Gilmer (1973) as rift spreading centres that were offset along sinistral transform faults. The best fitting pole of rotation has been calculated by these authors, and implies that the rift opened to a width of about 90 kilometres in the Lake Superior region.

The nature of the rift in the Lake Superior region was apparently rather complex, and some workers (e.g. Hinze et al., 1982) have considered the possibility that it may have been a triple junction, with a third arm extending in the direction of Lake Nipigon in Ontario. It can be shown with some certainty that more than one rift basin was present in the Lake Superior area in Keweenawan times (White, 1966a,b; Green, 1977; Halls, 1978). Hinze et al. (1982) have documented a major cross fault (the Thiel Fault) running north-northeast from just east of the Keweenaw Peninsula, cutting the Lake Superior basin into eastern and western portions, each with somewhat different geophysical characteristics. White (1966b) noted the thinness in outcrop of the lavas and overlying Copper Harbor Formation just west of the Michigan-Wisconsin border (near Mellen), and discussed geophysical evidence that these Formations are also thin along a line running north from this point. White

(1966b) therefore proposed a division along this line, between two basins, one to the southwest and one to the northeast, where the lavas and Copper Harbor Formation are much thicker.

The lavas and overlying sediments of the Copper Harbor, Nonesuch and Freda Formations, define a broad syncline (the Lake Superior syncline) with an axis roughly coincident with the axis of the lake. This syncline formed, at least partly, due to basin subsidence contemporaneous with lava extrusion and clastic sedimentation (Davidson, 1982). The overall synclinal structure is perturbed locally, in particular by an anticline near the Michigan-Wisconsin border (which forms the core of the Porcupine Mountains). The observations of Hubbard (1975) indicate that this folding must postdate Copper Harbor deposition.

The limbs of the Lake Superior syncline are cut by faults parallel to the basin margin (particularly the Isle Royale Fault north of Isle Royale and the Keweenaw Fault). These are high angle thrust faults and have caused the central portion of the basin to be raised some 2 to 4 km. It is not entirely clear whether or not these faults are genetically related to any normal faults that may have bounded the rift basin in Keweenawan times (Halls, 1978). However, White (1966a,b) pointed out that the present faults

cannot be wholly responsible for the original basin margins because the considerable numbers of Keweenawan lava clasts in the Copper Harbor Formation indicate that lavas were present beyond the present Keweenaw Fault. Hamblin (1958, pp. 46-49) noted exposures of Keweenawan basalts southeast of the Keweenaw Fault that substantiate the proposal that basaltic highlands were present in this area.

The Middle Keweenawan lavas and Upper Keweenawan sediments generally have shallow to moderate dips towards the axis of the Lake Superior syncline. Due to syndepositional subsidence, dips in general decrease upsection (White, 1968). On Isle Royale, dips are 5° to 30°. On the south shore dips are generally between 30° and 45°, but are locally steeper, becoming vertical in the area immediately west of the Michigan-Wisconsin border. The interflow sediments of the Minnesota shore, Nipigon Bay and Mamainse Point also have shallow to moderate dips into the basin.

2.2 Stratigraphy

Figure 2.2 shows a very generalised stratigraphy of the Lake Superior basin. There are many gaps in the rock record, so basinwide Keweenawan correlation is largely based

on U-Pb, Rb-Sr and K-Ar dates and on palaeomagnetic pole positions and reversals. Van Schmus et al. (1982) have summarized the data and concluded that Keweenawan igneous activity peaked 1,110 million years ago (±10 Myrs). Copper Harbor sedimentation occurred immediately after this, and may have continued until 1023 million years ago (±46 Myrs), at which time deposition of the Nonesuch Formation was taking place, according to a date quoted by Van Schmus et al. (1982). The Lower, Middle and Upper Keweenawan divisions (Halls, 1966) do not correspond to the palaeomagnetic polarity epoques, but are used here for the sake of simplicity and continuity.

Various authors (e.g. Hubbard, 1975) have chosen to omit the lithological terms sometimes included in the Freda and Nonesuch formation names. This practice is followed here and is extended to the Copper Harbor Formation. None of the formations are entirely of one lithology, and this style of terminology best accords with present nomenclature practice.

The various contacts between the lavas, the Copper Harbor Formation, the Nonesuch Formation and the Freda Formation are all conformable, and, grossly, gradational, with interfingering relationships. Interflow sediments similar to sediments of the Copper Harbor and Nonesuch Formations occur within the lava pile, and a number of lava

flows are present within the Copper Harbor Formation, indicating a gradual change from predominantly volcanic to predominantly clastic sedimentation. The transitions between facies of the Copper Harbor Formation, the Nonesuch Formation and the Freda Formation are also gradational rather than abrupt, indicating gradual changes from one environment to another (Daniels, 1982; Elmore, 1981).

The thicknesses of the formations vary from place to place within the basin. The Copper Harbor Formation is especially variable (from about 120 m to about 1850 m). The maximum thickness observed at the surface is on Isle Royale (about 1850 m according to Wolff and Huber, 1973). On the Michigan mainland (Figure 2.3) the thickness varies along strike, averaging 1000 m to 1500 m, but thinning over lava ridges in the Houghton region of the Keweenaw Peninsula and west of Ontonagon in the Porcupine Mountains (here thinning to 150 m according to White and Wright, 1960). The thinning near Houghton seems to be the result of differential subsidence, but the thinning in the Porcupine Mountains is probably due to rhyolite flows that piled up near their source (White and Wright, 1960). The Copper Harbor Formation and the underlying basalts also thin west of the Michigan-Wisconsin border, as mentioned in section 2.1. According to Hamblin and Horner (1961) the Copper Harbor Formation thins

to a minimum of 120 m in this area before thickening again to the west. White (1966b) interpreted this as an area of minimal subsidence.

There is also evidence that the Copper Harbor Formation thickens into the basin (Daniels, 1982), a result of syndepositional subsidence. However, White (1966a) noted that the lava flows also thicken into the basin and probably increase in abundance, so are likely to make up a larger thickness within the Copper Harbor Formation in the centre of the basin. Studies of gravity anomalies (Hinze et al., 1982) tend to support this conclusion, indicating that the total thickness of the Upper Keweenawan sediments under Lake Superior, interbedded with or overlying the lavas, does not greatly exceed that of the thickest sections seen on land.

2.3 The Keweenawan lavas

For more than a century the remarkable native copper deposits of the Keweenawan lavas and interflow sediments have excited attention. A comprehensive geological review of the Keweenawan copper belt was produced by Butler and Burbank (1929), shortly after the peak in copper production. The mines of the Keweenaw Peninsula are now worked out, but

interest in the lavas continues and various recent studies throw much light on Keweenawan palaeogeography.

Most of the lavas are basalts. Geochemically many are olivine tholeiites with compositions that are more often associated with modern mid-ocean ridge basalts (Green, 1972). Rhyolites are volumetrically significant, although often areally restricted (e.g. west of Ontonagon in the Porcupine Mountains), and intermediate volcanics are least abundant (Green, 1982). Many basalt flows are thick and areally extensive. White (1960) noted that one flow, the "Greenstone Flow", is in many places more than 300 m thick and crops out both on the Keweenaw Peninsula and on Isle Royale, in both localities continuing for tens of kilometres along strike. It therefore seems likely that the flood basalts were erupted from fissures (Green, 1982), and that slopes in the basin were generally low (White, 1960, estimated slopes less than 0.0038). White (1960) noted that thickening trends and bent pipe amygdale orientations indicate that basalt generally flowed away from the centre of the Lake Superior basin. The interflow sediments, however, were carried into the basin from the basin margins. White (1960) therefore proposed periodic slope reversals, clastic sedimentation occurring when volcanic outpourings did not keep pace with basin subsidence.

Green (1977, 1982) has identified several (up to eight) possible volcanic centres which were active in the Lake Superior basin at different stages during the Keweenawan. The basalt accumulations occurred in different basins within the Lake Superior basin. The sediments observed during this study are associated with a number of different volcanic plateaus. The lavas of the Osler Group in the Nipigon Bay area (Figures 2.1 and 2.4) were erupted relatively early in the rifting and were apparently roughly contained within that area. To the southwest, extensive volcanism at that time formed a lava plateau in the western Lake Superior area, nearly all of which is now buried, although some of these volcanics outcrop south of the Keweenaw Fault. Later, basalt plateaus formed both in the west (most of the North Shore Volcanic Group) and in the eastern Lake Superior area. The basalts at Mamainse Point belong to the latter plateau, and at that time, Michipicoten Island in northeastern Lake Superior was apparently an active volcano. The Isle Royale and Keweenaw Peninsula basalts (the Portage Lake Lava Series) and basalts to the southwest in Wisconsin were formed last and roughly contemporaneously. At that time, there was apparently a volcano west of Ontonagon in the Porcupine Mountains.

2.4 Previous sedimentological studies of the Middle and Upper Keweenawan

Sedimentological features of the Copper Harbor Formation throughout the Keweenaw Peninsula and on Isle Royale were first studied in detail by White and Wright (1960). These authors distinguished two facies. Their "conglomerate facies" was defined as conglomerate with subordinate sandstone. Palaeocurrent (crossbedding) meaurements within this facies on the Keweenaw Peninsula indicated deposition by streams flowing northwards into the basin (White (1952) had previously studied imbrication in an interflow conglomerate and had come to similar conclusions). Their other facies was found on the mainland only in the west of the area mapped, although at various stratigraphic levels. This "red facies" consisted of fine and medium sandstone. Palaeocurrent (crossbedding) measurements in this facies on the mainland apparently indicated flow to the south. These authors obtained relatively little data from Isle Royale, but their diagram (p. 86) would appear to indicate an easterly progression from their "conglomerate facies" to their "red facies". White and Wright (1960) interpreted their "conglomerate facies" on the mainland as a piedmont fan deposit formed at the foot of hills along the

southern basin margin. Their "red facies" was interpreted as a "flood plain deposit" where transport was, at least locally, southward.

Hamblin and Horner (1961) similarly came to the conclusion that the source of the Copper Harbor on the Keweenaw Peninsula was to the south and southeast, and considered the palaeocurrents to indicate that sediment came from the area of what is now the Huron Mountains (which form the broad peninsula to the south of the Keweenaw Peninsula). However, the data set is small, and standard deviations are large. These authors regarded the sediment coarseness and the variation in pebble lithologies as evidence that the source area was probably not more than 16 km distant. Hubbard (1972) showed that the pebbles of the Copper Harbor Formation must mainly be derived not from the immediately underlying lavas of the Portage Lake Lava Series but from older Keweenawan lavas now only exposed south of the Keweenaw Fault, suggesting significant tectonic movement to build highlands from these rocks before or during deposition of the Portage Lake Lavas. These highlands seem to have been gradually worn down, because, according to Hubbard (1972), the proportion of pre-Keweenawan detritus increases upsection in the Upper Keweenawan sedimentary rocks, becoming dominant in the Freda Formation. On Isle Royale,

Hamblin and Horner (1961) noted transport directions to the southeast and east, and thought that here too the source area was not far distant (to the west). These authors also concluded that the "major (north and south) current systems merged and flowed to the northeast".

An interflow conglomerate on Isle Royale was studied by Larson (1969), who concluded that flow directions were to the southeast. Detailed sedimentological work on the Copper Harbor Formation of Isle Royale has been done by Wolff and Huber (1973). Between Rainbow Cove in the west, and Point Houghton in the east (Figure 2.5) these authors found a marked decrease in grainsize, but a fourfold increase in formation thickness (from 460 m to 1850 m). Over the same distance, palaeocurrents (not necessarily all recorded at the same stratigraphic level) apparently indicated a swing from flow directed southeast to flow directed east and northeast. Wolff and Huber (1973) analysed conglomerate clast compositions and concluded that the source consisted of Keweenawan lavas. About half the volcanic clasts are felsic, the other half being basic. This proportion of felsic clasts, according to Wolff and Huber (1973), requires that the source was the North Shore Volcanic Group of Minnesota, where there is a significant proportion of felsic flows, indicating that there was uplift in that area during

deposition of the Copper Harbor Formation on Isle Royale and the Keweenaw Peninsula. Larson (1969) came to similar conclusions as to source area for the interflow conglomerate, also noting that certain granite fragments may have come from rocks in Ontario.

The uppermost part of the Copper Harbor Formation on the Keweenaw Peninsula was studied by Johnson (1973). The poor sorting and the matrix-rich fabric of these conglomerates led Johnson (1973) to interpret these conglomerates as the products of sandy debris flows. This is in contrast to interpretations of other workers, and usefully directs attention to the similarity between the deposits that can result from dilute debris flows and the sediments that can be deposited by concentrated and powerful stream flows.

The most recent work on the sediments of the Copper Harbor Formation is that of Elmore (1981, 1984) and Daniels (1982). These workers emphasized the gradational boundaries between the Copper Harbor, Nonesuch and Freda Formations, and interpreted these rocks as being parts of the same alluvial fan-floodplain-lacustrine system which evolved as time went on. In this interpretation the Copper Harbor Formation represents a fining upwards and distally fining alluvial fan complex, dominated by braided fluvial processes. Late during

deposition of the Copper Harbor, a distal lacustrine environment gradually developed and then completely transgressed the alluvial fans to produce the Nonesuch Formation. Daniels (1982) postulated that volcanic dams formed at this time, but was unable to prove this. A resurgence of fluvial activity later caused progradation of a braided fluvial system, represented by the Freda Formation. Lateral variations in clast composition were not found to be very significant, but like Hubbard (1972), Daniels (1982) noted that, proceeding upsection through the Upper Keweenawan sediments, fewer volcanic fragments and more plutonic fragments are found. The "average" Copper Harbor Formation sandstone is an immature litharenite (scheme of Folk, 1968), whereas the "average" Freda Formation sandstone is a more compositionally mature feldspathic litharenite (Daniels, 1982). This is consistent with Hubbard (1972) who hypothesized that the Keweenawan volcanic source was gradually worn away to expose pre-Keweenawan rocks. The heavy mineral suites, according to Daniels (1982), support this conclusion.

The Mamainse Point interflow sediments have been studied by Merk (1972) and Merk and Jirsa (1982), who considered these to be the products of fluvial deposition high on an alluvial fan. These authors concluded that the

source area was to the east of Mamainse Point (from the general sense of palaeocurrent indicators). Most of the conglomerate clasts were found to be derived from mafic Keweenawan volcanics. However, clasts of pre-Keweenawan basement were found to make up 27% of the rocks. Merk and Jirsa (1982) therefore concluded that a tectonically active highland of Keweenawan lava and pre-Keweenawan rocks existed somewhere to the east.

The interflow sediments of the North Shore Volcanic Group in Minnesota have been studied by Merk (1972) (whose study took in all of the interflow sediments of the Lake Superior area) and Jirsa (1980). Both of these studies focused on the sedimentary petrography and its implications for provenance. According to Jirsa (1980), the style of the interflow sediments, regional variation in sediment thickness, and the regional palaeocurrent pattern indicate that sedimentation occurred in two basins, one to the northeast, the other to the southwest, the division being at a point roughly bisecting the present outcrop belt of the North Shore Volcanic Group. These basins would correspond to those proposed by White (1966b) (mentioned in section 2.1). However, interflow sediments are not abundant in the North Shore Volcanic Group and outcrops are small, so all conclusions of this sort are based on very little data.

Similar problems obstruct the analysis of depositional processes and environments. The presence of fine grained sediment amongst the North Shore interflow sediments led Jirsa (1980) to conclude that deposition was by meandering as well as braided streams. To distinguish the two fluvial styles Jirsa (1980) relied on measurements of palaeocurrent variance. The data set is, however, too small to provide conclusive results. It is nevertheless evident that temporally and spacially there was considerably variation in the energy of the depositional environment, resulting in sediments ranging from laminated fines, interpreted by Jirsa (1980) as lacustrine, to sand-rich poorly sorted conglomerates, which Jirsa (1980, p. 49) interpreted as "the result of rapid dumping of sediment after relatively short distances of transport" (apparently in braided streams on alluvial fans, although whether deposition was by debris flows or by fluvial processes was not discussed by the author).

Near Minneapolis, Minnesota, sediments of the Solar Church Formation have been studied. These directly overlie lavas, and were probably deposited in the Keweenawan rift at nearly the same time as the Copper Harbor Formation, although in a separate basin. The general nature of these sediments is finer than the Copper Harbor Formation, but

Morey (1974) and Morey and Ojakangas (1982), with the aid of much subsurface data, have concluded that these are probably fluvial deposits, and that streams flowed east and west into the basin from its margins. These authors claimed to identify two types of fining-upward cycles (fining from sandstone to siltstone). The smaller cycles (1-15 m) were interpreted to be the result of channel migration (i.e. sedimentologically controlled). The larger cycles (up to 180 m thick) were interpreted to be the result of pulses of tectonic activity.

CHAPTER 3

DESCRIPTION AND INTERPRETATION OF FACIES

3.1 Introduction

Division of sedimentary rocks into various facies is the first step towards finding the processes by which sedimentation occurred, which may in turn point to the original sedimentary environment. Rocks within the same facies have the same aspect ("face") and can be assumed to have been deposited by similar processes. This does not mean that those rocks were deposited in the same environment, since similar processes operate in dissimilar environments. However, particular associations of facies may be peculiar to certain sedimentary environments.

3.2 Facies descriptions

The facies of the Copper Harbor Formation fall almost entirely into three categories: conglomerates, sandstones and siltstones. These are coded with prefixes C, S, and St respectively. On the Keweenaw Peninsula, stromatolite-oncolite beds are found within the Copper Harbor Formation,

and these are described separately. In a number of cases, the divisions between facies are gradational rather than sharp, because the rocks show a spectrum of sedimentary features.

3.2.1 Conglomerate facies

The Copper Harbor Formation conglomerates can broadly be divided between pebble-cobble-boulder conglomerates ("coarse conglomerates") and pebble-cobble conglomerates ("fine conglomerates"). In all the conglomerates, clasts are subrounded to well rounded fragments of felsic and basic volcanics in subequal proportions (Daniels, 1982; Elmore, 1981; White and Wright, 1960). Iron staining of both clasts and sandy matrix gives all the conglomerates a strong red colouration. Structure in the conglomerates is only visible where the orientation and condition of the outcrop permit. It must therefore remain somewhat uncertain exactly how many of the conglomerates labelled "massive" are indeed structureless. For details of the lateral relationships between conglomerate facies, the reader is referred to the surveyed sections in Chapter 4.

3.2.1.1 Massive, c-m supported, coarse conglomerate, C-1

The most abundant facies in the Copper Harbor Formation is massive, coarse conglomerate with fabrics on the borderline between clast-supported and (sand) matrix-supported. These are hereafter called c-m supported conglomerates (for clast and matrix supported). In fact, the clast-matrix division in these very poorly sorted rocks is almost arbitrary, but it can be noted that whilst many pebbles and cobbles may be in contact with one another, many others "float" in a very poorly sorted sand-granule matrix (Figures 3.1 and 3.2). Clasts in this facies have long axes up to 176 cm, although it is rare for clasts to be larger than 50 cm. Preferred clast orientation, if any, is a subhorizontal alignment of long axes of the largest clasts.

Individual beds are hard to identify, and where seen are not laterally extensive, being interrupted by sandstone lenses. A line perpendicular to dip and strike will, on average, intersect a sandstone lens once every 50 cm to 1 m, but the sandstone lenses may be more closely spaced. In some such beds, the coarsest clasts are at the base of the bed, but in the majority of beds even this crude coarse-tail grading is absent. In a few beds, the topmost layer or two of cobbles are imbricated. As already mentioned, this facies

Figure 3.1 View of bedding plane of massive, c-m supported, coarse conglomerate (facies C-1). Hammer for scale.

Figure 3.2 Massive, c-m supported, coarse conglomerate (facies C-1). Bedding is dipping at about 30° to the right of the picture. Note the very poor sorting, and the consequent difficulty distinguishing "clasts" and "matrix". Clast-support and matrix-support are not very useful terms here. Person for scale.

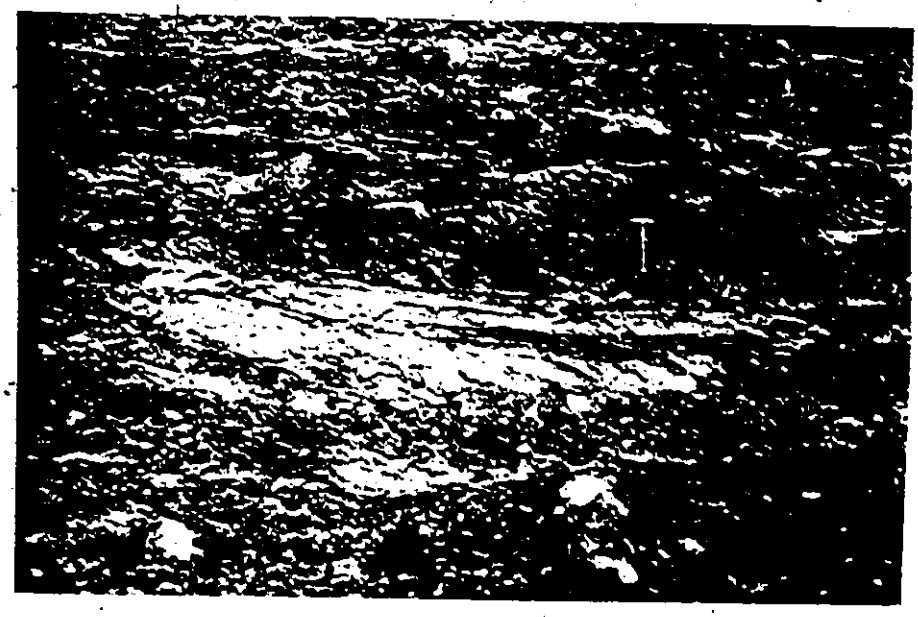
is invariably closely associated (vertically and laterally) with sandstone lenses (facies S-1, S-2, S-4, S-5), ranging from a few centimetres to a few tens of centimetres in thickness (Figure 3.3). The conglomerate itself commonly also contains numerous silty drapes, some with desiccation cracks. The sandstone lenses show two types of relationship with the conglomerate. Where sandstone infills a channel in the conglomerate basal and lateral contacts tend to be diffuse, but the upper contact with the conglomerate is invariably sharp due to erosion. Some of the erosive bases of conglomerate beds show gutter casts. Where sandstone was once a more laterally extensive sheet but has been reduced to isolated lenses by erosion, only the basal contact is diffuse. Some sandstone lenses fall between the two categories, showing a gradational contact with the conglomerate at one edge and an erosive contact with the conglomerate at the other (as if a channel filled with sandstone was partly eroded). These relationships are best shown by the surveyed sections in Chapter 4.

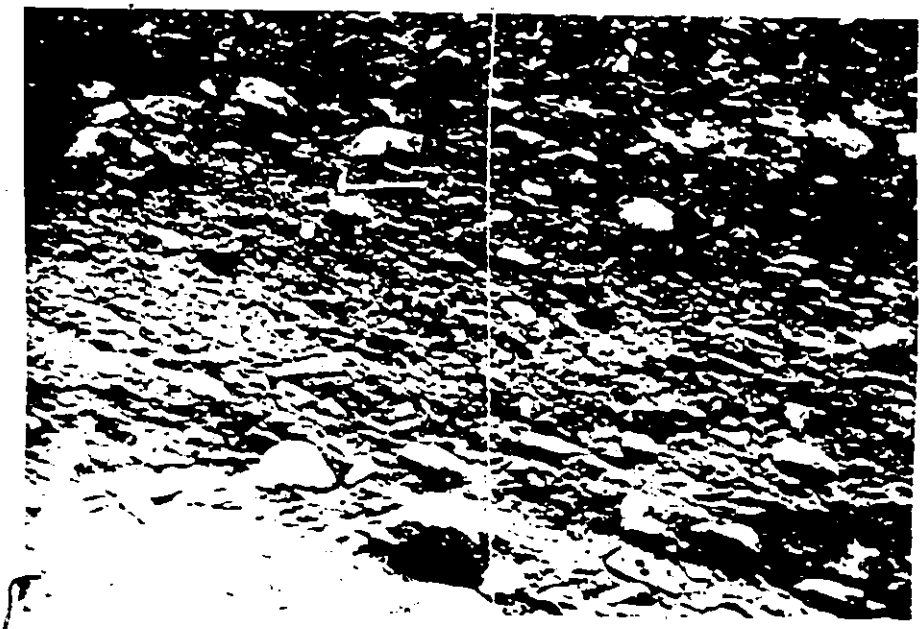
3.2.1.2 Massive, matrix-supported, coarse conglomerate, C-2

In rare cases (e.g. parts of the Five Mile Point section) coarse conglomerates are matrix-supported. The

Figure 3.3 Sandstone lens (bleached white) within massive, c-m supported, coarse conglomerate (C-1). Sandstones with slightly greater lateral extent are visible in the upper part of the photograph. All these sandstones show upper flow regime flat bed or poorly developed cross-stratification (facies S-1, S-2, S-4, S-5). Hammer for scale.

Figure 3.4 Imbricated, c-m supported or clast-supported, coarse conglomerate (top and bottom of picture), and imbricated, clast-supported, fine conglomerate (centre of picture). Bedding is horizontal. Hammer for scale.





matrix is invariably sandy and very poorly sorted, and clasts are subrounded or rounded and range up to about 25 cm in length. The conglomerate interfingers with sandstone lenses, as described for facies C-1. As with facies C-1, the only evident clast orientation is a tendancy for the clast long axes to be subhorizontal. The division between this facies and facies C-1 is not sharp, and all gradations in the relative amounts of clasts and matrix can be found. The bases of these beds are normally slightly erosive.

3.2.1.3 Imbricated, c-m supported or clast-supported, coarse conglomerate, C-3

This facies (Figure 3.4) is less common than facies C-1, but more common than facies C-2. Clasts range in size up to about 35 cm and are poorly to well imbricated with long axes perpendicular to flow. Sorting of clasts is generally better than in facies C-1. Many of the fabrics are again c-m supported, but generally with less sandy matrix than facies C-1, and in some cases showing complete clast support. Basal contacts, where beds can be defined, are generally slightly erosive, and the largest clasts may be concentrated at the base of the bed. As with facies C-1, beds closely interdigitate laterally and vertically with

sandy lenses, and silty drapes within beds are common. Where sandstone predominates, facies C-3 (and C-1) may only form stringers one or two cobbles thick. These have erosive bases, and in many cases, mud draped tops.

3.2.1.4 Massive, c-m supported or clast-supported, fine conglomerate, C-4

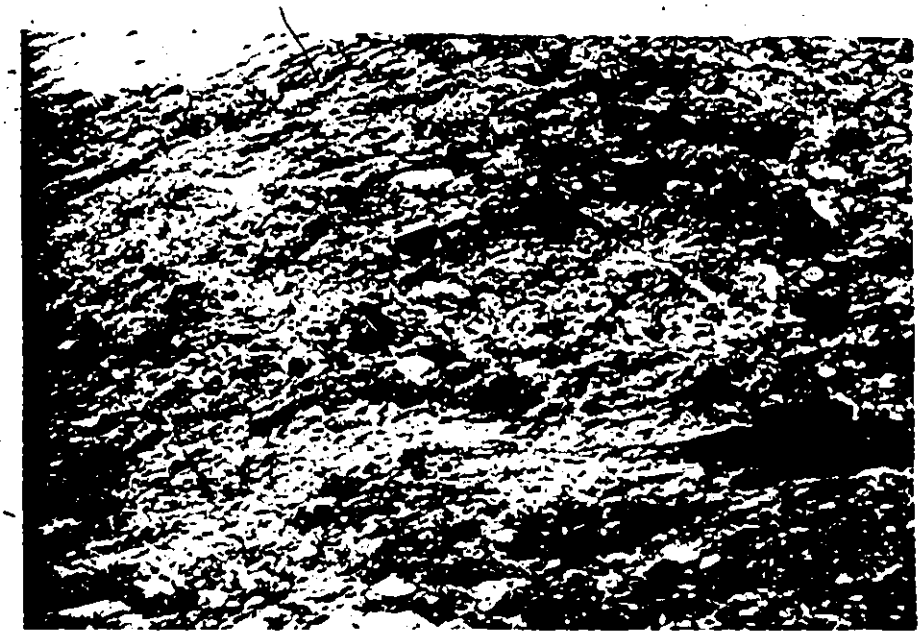
Clasts in this facies range up to about 15 cm in length. Sorting is moderate to poor. The conglomerate may have a c-m supported fabric, or may be completely clast-supported. Some beds contain sandy matrix only at bottoms of interstices (an example of such shelter structures is shown in Figure 3.5). This facies occurs in two situations: as fills of channels (up to 40 cm deep) in the tops of beds of the coarser more matrix-rich conglomerate facies (Figure 3.6), and as more laterally continuous sheets (up to about 30 cm thick). Some of the channel fills fine upwards into sandstone.

3.2.1.5 Imbricated, clast-supported, fine conglomerate, C-5

The largest clasts in this facies are about 15 cm long, and sorting of clasts is poor to moderately good.

Figure 3.5 Shelter structures in massive, c-m supported or clast-supported, fine conglomerate (facies C-4). Lens cap is 5 cm in diameter.

Figure 3.6 Channel fill of massive, clast-supported, fine conglomerate (facies C-4), picked out by calcite-filled interstices. Note also the poorly defined sandstone lens lower in the picture. Notebook is 20 cm long.



Clasts are imbricated with long axes perpendicular to flow. Sandy matrix may or may not be present, and in some cases shelter structures are formed, with matrix is present only as (geopetal) accumulations at the bases of interstices otherwise filled with sparry calcite. This conglomerate occurs in the same two situations as described for facies C-4. Figure 3.4 shows a laterally continuous sheet of this facies.

3.2.1.6 Imbricated, matrix-supported, fine conglomerate, C-6

This is an uncommon facies, and is gradational with facies C-5, but with more (sandy) matrix. The only occurrence is as relatively extensive sheets (up to 30 cm thick). At Five Mile Point, one such sheet passes laterally into facies C-5.

3.2.1.7 Small scale planar cross-stratified fine conglomerate, C-7

Planar cross-stratification in conglomerates of the Copper Harbor Formation shows a wide variation of scales. At the smallest end of the spectrum, pebble-cobble

conglomerates (clasts up to about 10 cm) may show planar cross-sets only 15 cm thick. Fabrics in these beds range from matrix-supported to clast-supported. At Sandstone Falls, one such conglomerate bed shows two clast orientations: at the base of the set, pebbles are lying on planar foresets, but at the top of the set, the pebbles are imbricated. The base of this set is slightly erosive, and the set shape is not perfectly tabular.

3.2.1.8 Medium scale planar cross-stratified fine conglomerate, C-8

Clasts in this facies are normally less than 15 cm long. Sets of cross-strata may have slightly erosive bases, and are generally 40-90 cm thick. Grain size and sorting varies between different foresets of the same bed. Sandy matrix may or may not be present, and in many examples, some of the planar foresets are clast-supported whilst others are largely sand and are matrix-supported (in which case the facies becomes gradational with facies S-10, planar cross-stratified pebbly sandstone). Grading may be normal, reverse, or absent. Pebbles generally are approximately aligned with b-axes parallel to foreset dip. Foreset dips may approach angle of repose, but most are somewhat less

steep. Three types of occurrence of this facies can be identified:

1. tabular beds, generally extending at least 10 m laterally (Figure 3.7);

2: channel fills (generally in coarser conglomerate), which generally fine upwards, and in which cases foresets are subparallel to the channel edge;

3. in one instance (at Sandstone Falls), a partial channel fill, in which foresets are almost perpendicular to the channel margin.

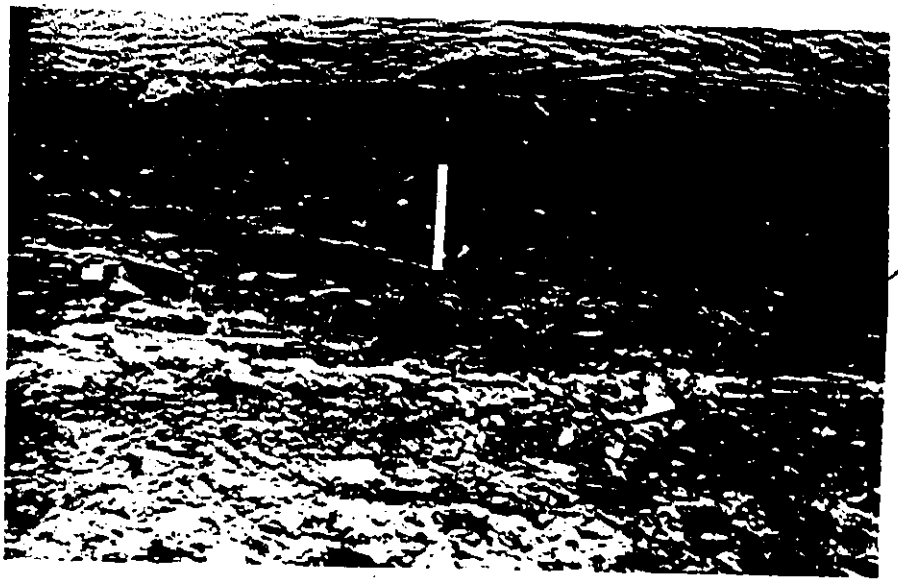
In one channel fill of the second type, at Horseshoe Harbor, conglomeratic cross-strata pass laterally into silt, which completes the fill.

3.2.1.9 Large scale planar cross-stratified fine conglomerate, C-9

At Rainbow Falls, Michigan, sets of planar cross-stratified conglomerate reach a thickness of 4 m (Figure 3.8). Foresets range from sandstone through (sandy) matrix-rich conglomerates to open-framework conglomerates without sandy matrix. Clasts are up to about 15 cm in length. Individual foresets vary greatly in thickness, with some conglomeratic "foresets" up to 50 cm thick. Some of the

Figure 3.7 Medium scale planar cross-stratified fine conglomerate (facies C-9), passing laterally into planar cross-stratified pebbly sandstone (facies S-10). Beneath the cross-stratified conglomerate is parallel-laminated siltstone (facies St-2), in turn underlain by a thin stromatolite bed (white) developed over massive, c-m supported, coarse conglomerate (facies C-1). 30 cm ruler for scale.

Figure 3.8 Large scale planar cross-stratified fine conglomerate (facies C-9). The horizontal conglomerate bed at top of outcrop is about 50 cm thick.



sandy foresets also reach this thickness, and internally show trough cross-bedding (truncated by the overlying conglomeratic foresets). Foreset dips are much less than angle of repose. A much smaller, but similar type of cross-stratification is seen at Five Mile Point. The thickness of this bed is only 1m, but again the "foresets" are thick, shallowly dipping, and consist of an alternation between trough cross-stratified sandstone and conglomerate.

3.2.1.10 Trough cross-stratified fine conglomerate, C-10

In conglomerates with pebble and granule grainsizes, trough cross-stratification is not uncommon (Figure 3.9). Sets are generally 40-50 cm thick, with largest pebbles at set bases. Such units laterally pass into more massive conglomerate or trough cross-stratified pebbly sandstone (facies S-5). On Isle Royale, at Rainbow Cove, one trough cross-stratified conglomerate unit extends laterally for more than 200 m. This particular bed is relatively well sorted, consisting very largely of granule size sediment.

Figure 3.9 Trough cross-stratified fine conglomerate (below notebook). These troughs are largely filled with granules and pebbles. Very little sand is present. Notebook is 20 cm long.

Figure 3.10 Symmetrical gravel waves overlying horizontally stratified fine conglomerate (facies C-II). The gravel waves are draped with silt. The ruler lying across the top of the gravel waves is 30 cm in length.

3.2.1.11 Horizontally stratified, fine conglomerate; C-11

Some fine (pebble and cobble) conglomerates show horizontal stratification. Stratification is outlined by changes in grain size and amount of sandy matrix. Completely sandy strata (up to 10 cm thick) may be present, and these may internally show parallel or cross-stratification. At three locations, near Dan's Point, this facies is associated with symmetrical gravel waves (wavelengh 25 cm, height 2.5 cm) draped with mud (Figure 3.10).

3.2.2 Sandstone facies

The Copper Harbor sandstones are red litharenites with grey heavy mineral laminae (Daniels, 1982). With the exception of the large scale cross-stratified sandstones at Five Mile Point, sorting is generally poor (according to the sorting divisions of Folk, 1974).

3.2.2.1 Upper flat bed sandstone, S-1

This facies contains sandstones of all grades, but most commonly consists of fine and very fine sandstones. The upper flow regime flat bed lineation is evident due to

slight grain size changes, best seen in the finer sandstones, where better sorting clarifies such variations. Horizontal surfaces show parting lineation, in some cases with heavy mineral shadows (Figure 3.11). In places, current crescents have formed around mud chips. A few centimetres of fine upper flat bed sandstone commonly are capped by a thin layer of very fine upper flat bed sandstone.

3.2.2.2 Upper flat bed pebbly sandstone, S-2

Less abundant than facies S-1, pebbly flat bed tends to be formed in the coarser sandstone grades. Pebbles and pebble clusters (and mud chips) may or may not show current crescents. As the number of pebbles decreases, this facies grades into facies S-1. With increasing amounts of granule-pebble material, the facies grades into horizontally stratified fine conglomerate (facies C-1).

3.2.2.3 Horizontally stratified sandstone, S-3

This facies normally occurs in very fine or fine sandstones, but exceptionally occurs in coarser sandstones. Upper flat bed, current ripples, and (minor) small scale trough cross-stratification are intimately associated.

Figure 3.11 Upper flat bed sandstone (facies S-1). In the centre of the photograph ripples are washing out into upper flat bed. Hammer for scale.

Slightly scouring horizons covered with mud chips and coarser sandstone are generally present (Figure 3.12). Horizontal surfaces may show casts of current ripples, or less commonly, of wave ripples (Figure 3.13), or current ripples with wave overprinting. Horizontal surfaces may otherwise display rib and furrow, or enigmatic "knobbly" structures (discussed in section 3.3.3). Bubble sandstone, with calcite infilling the bubbles, is in places also present (Figure 3.14). Some surfaces show wrinkle marks (Figure 3.15). These various types of surface may show mud drapes, commonly with desiccation cracks (Figure 3.16), and in exceptional cases with rain prints (Figure 3.17). Some pebbles may be present, particularly on the scouring mud chip surfaces. The different structures alternate on a vertical scale of several centimetres only.

3.2.2.4 Trough cross-stratified sandstone, S-4

Trough sets (Figure 3.18) vary in thickness, reaching 40 cm. Widths of troughs also vary widely, in some cases reaching 3 m. Mudchips may be present on foresets (in some places to the extent that one trough set is full of mud and fractures messily). Some mud draped trough scour hollows are preserved, a few with ripple casts. In places, a trough

Figure 3.12 Mudchips (bleached white) on a slightly erosive surface within the horizontally stratified sandstone facies (facies S-3). Bedding is approximately parallel to the plane of the picture. Pen for scale.

Figure 3.13 A wave rippled surface within the horizontally stratified sandstone facies (facies S-3). The light meter is 10 cm long.

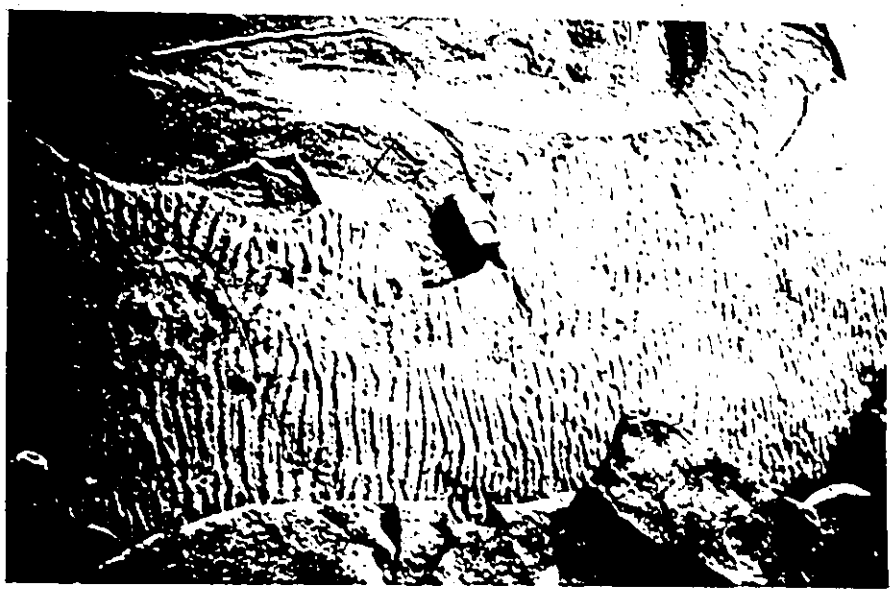


Figure 3.14 A bubble sandstone surface within the horizontally stratified sandstone facies (facies S-3). Note that some of the bubbles are filled with calcite. In vertical section, bubbles are seen to be flattened, indicating that they formed prior to compaction. Lens cap is 5 cm in diameter.

Figure 3.15 Wrinkle marks within the horizontally stratified sandstone facies (facies S-3). Lens cap is 5 cm in diameter.

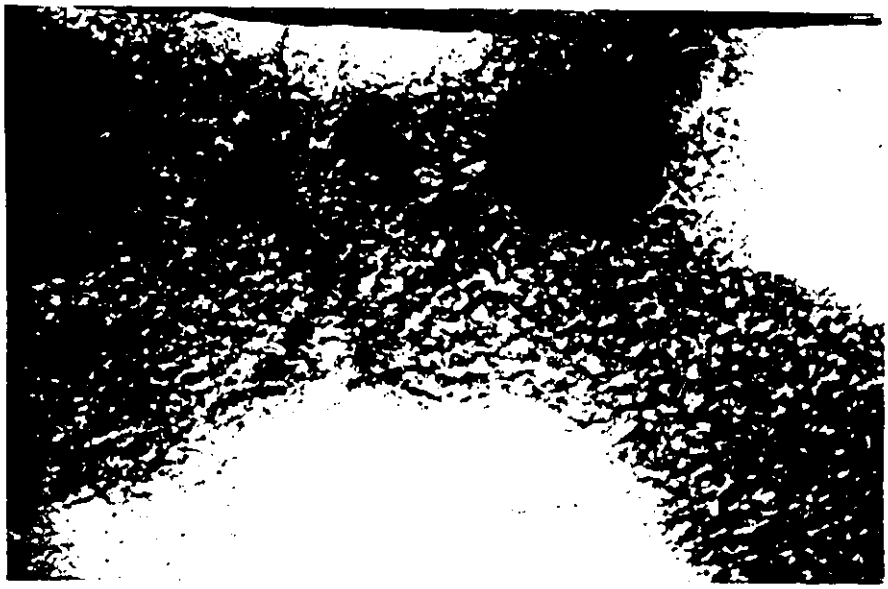
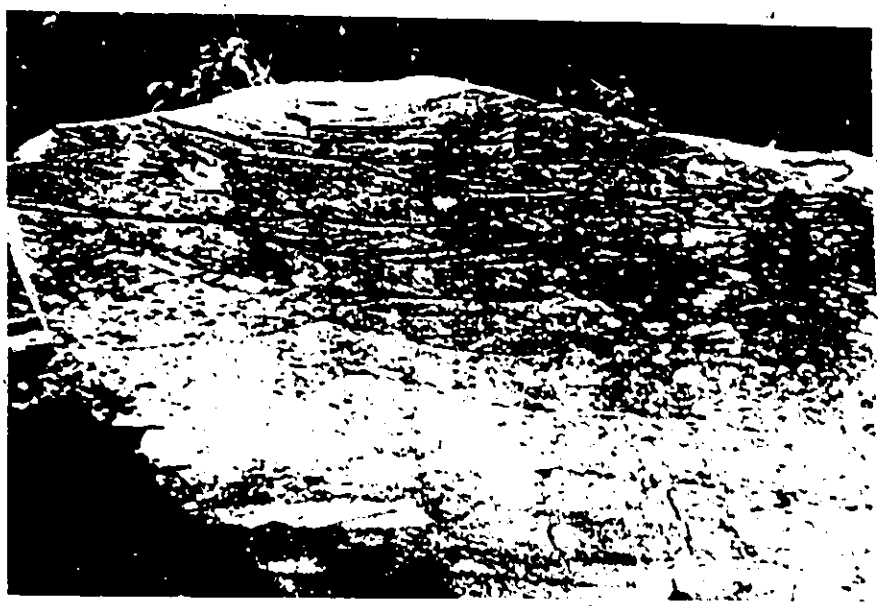


Figure 3.16 Desiccated mud draped surface within the horizontally stratified sandstone facies (facies S-3). The different colours are due to different degrees of bleaching. The conical depressions may be due to dewatering of the sediment. Lens cap is 5 cm in diameter.

Figure 3.17 Rain prints on mud draped current ripples within the horizontally stratified sandstone facies (facies S-3). Quarter for scale.

Figure 3.18 Trough cross-stratified sandstone (facies S-4). Hammer for scale.

Figure 3.19 Trough cross-stratified pebbly sandstone (facies S-5), overlain by upper flat bed sandstone (facies S-1). Notebook is 20 cm long.



partly filled with medium to coarse sandstone has a final fill of very fine sandstone.

3.2.2.5 Trough cross-stratified pebbly sandstone, S-5

With a decreasing proportion of pebbles, this facies grades into facies S-4. With increasing numbers of pebbles, this facies grades into trough cross-stratified very pebbly sandstone (facies S-6). Pebbles and cobbles (up to about 15 cm) concentrate at the bases of sets, but may also lie on foresets (Figure 3.19). Set thicknesses are generally less than 40 cm, but may reach more than 60 cm. As in facies S-4, mud clasts are common. In two localities, examples of climbing pebbly dunes are seen, where stoss sides are preserved.

3.2.2.6 Trough cross-stratified very pebbly sandstone, S-6

This facies is gradational with facies S-5, and, with increasing pebbles, with trough cross-stratified conglomerate (facies C-10). The matrix is invariably very poorly sorted medium or coarser sandstone, generally with a high proportion of granules. Sets are normally about 40 cm in thickness.

3.2.2.7 Mixed trough cross-stratified and flat bedded sandstone, S-7

In some locations, upper flat bedded sandstone and trough cross-stratified sandstone are too closely associated to be divided into different facies. Flat bed horizons are of very limited lateral extent (mostly only 1 m or less), but are abundant, and cut the tops of many of the trough sets. This gives rise to a mixed facies.

3.2.2.8 Mixed trough cross-stratified and flat bedded pebbly sandstone, S-8

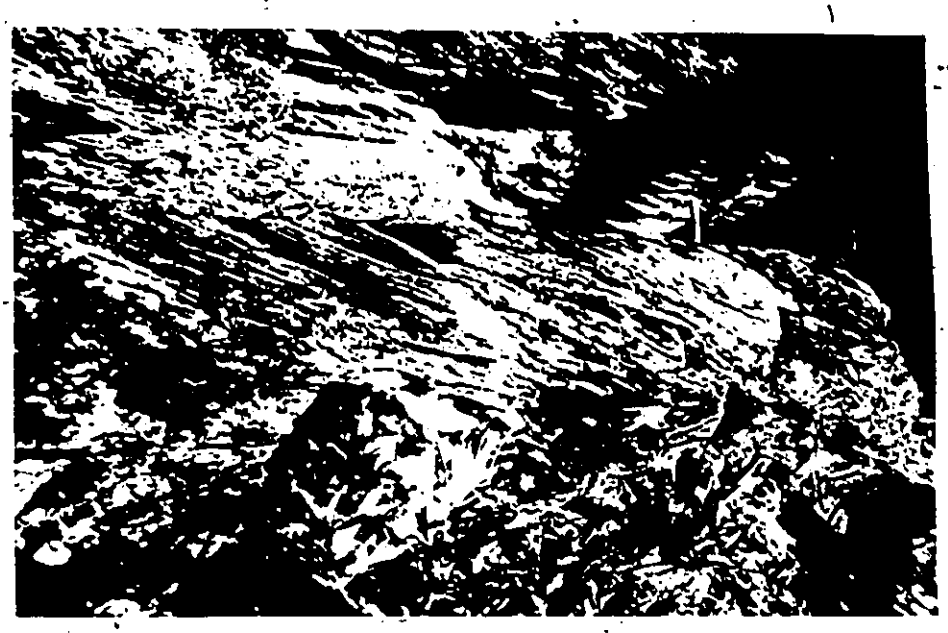
This facies arises in the same manner as S-7, but in pebbly sandstones.

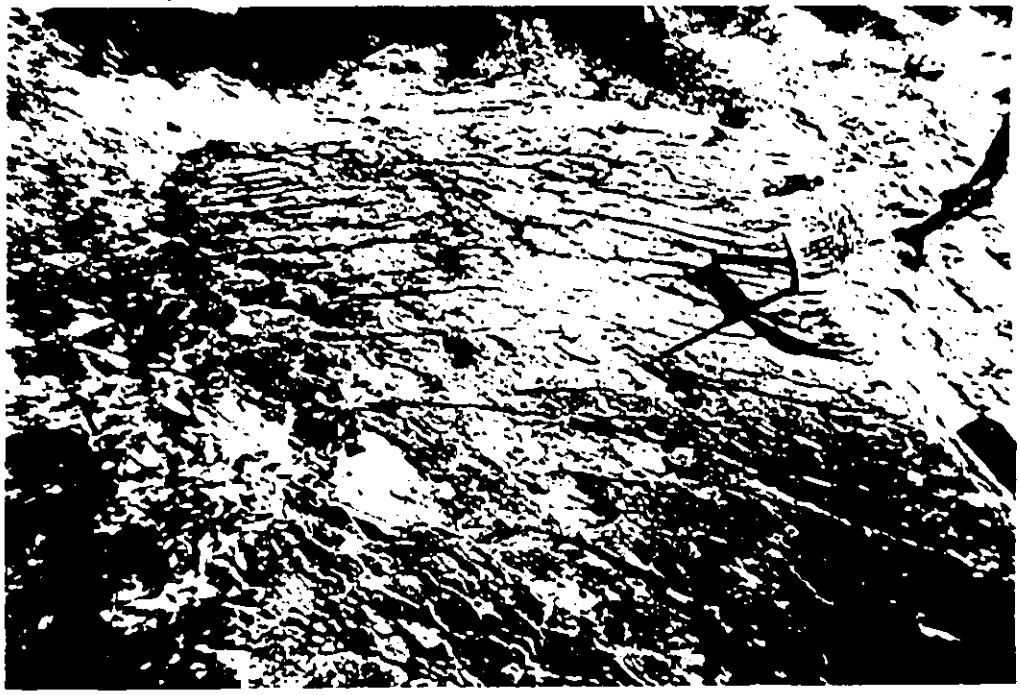
3.2.2.9 Planar cross-stratified sandstone, S-9

This is a relatively uncommon facies. Sets range in thickness from about 30 to 60 cm (Figures 3.20 and 3.21). Most foresets flatten out slightly at set bases. Original crest lines were apparently straight or slightly sinuous. The base of the set may be somewhat erosive, and set

Figure 3.20 Planar cross-stratified sandstone (facies S-9) viewed in section. Hammer (standing on set top) and backpack (red) for scale. See also Figure 3.21.

Figure 3.21 Planar cross-stratified sandstone (facies S-9). Plan view of the set shown in Figure 3.20. Hammer and backpack for scale.





boundaries are not normally tabular, but in most cases are slightly wedge shaped. Foreset dips generally approach angle of repose.

3.2.2.10 Planar cross-stratified pebbly sandstone, S-10

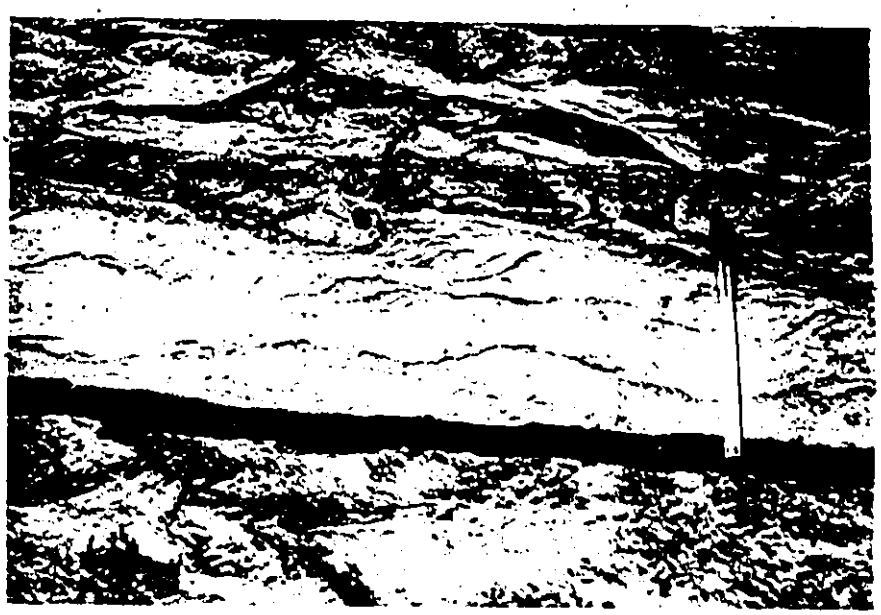
This facies is gradational with facies C-8 (medium scale planar cross-stratified fine conglomerate), and some examples pass laterally into that facies (Figure 3.7). Set thicknesses range from about 50 cm to 1 m. Set boundaries may be tabular or somewhat wedge shaped, but the extent of the set is ordinarily determined by the erosive base of overlying conglomerate. Pebbles may be concentrated at tops or bases of sets, or be spread throughout the set. This facies occurs in coarser sandstones than facies S-7, generally being medium or coarse sandstone.

3.2.2.11 Rippled sandstone, S-11

This facies is found in fine and very fine sandstones. Most ripples are current formed, but, in general, symmetrical forms are also present. Climbing ripples (Figure 3.22) may form part of the facies, but are not in general dominant, and it is uncommon for angles of

Figure 3.22 Climbing ripples within the rippled sandstone facies (facies S-11). Pen for scale.

Figure 3.23 A bedding plane within the rippled sandstone facies (facies S-11). The pen is resting on an angle of repose step downstream of which ripple fans have developed.



climb to be high enough to allow stoss side preservation. Most of the rippled layers making up this facies are only one ripple or so thick (1-2 cm), and the ripples tend to die upwards. Stacking of these layers forms the facies. Boundaries between ripple layers are flat or somewhat undulatory. Thin silt drapes between layers are common, and many show desiccation cracks. Mud chips may be present, either scattered throughout the ripples, in which case the fracture is very rough, or concentrated at slightly erosive coarser horizons overlain by ripples.

On horizontal surfaces, ripple casts or rib and furrow are present. At Presque Isle, large rippled surfaces (many square metres) are exposed, and can be seen to be at low angles to one another. The surfaces themselves undulate, and show occasional angle of repose steps (some spawning ripple fans as shown in Figure 3.23) and scour hollows, migration of which has given rise to 15 cm cross-sets in parts of the outcrop. One surface at Presque Isle shows a runnel about 10 cm deep and 1 m wide. The floor of the runnel shows linguoid ripple casts and wrinkle marks, and the edges show falling water level marks. Elsewhere at this outcrop, ladderback linguoid ripples are found (Figure 3.24).

Figure 3.24 Ladderback linguoid ripples in the rippled sandstone facies (facies S-11). During the very shallow last stages of flow, tiny ripples developed between the larger ripples. The compass is 7 cm wide.

Figure 3.25 Large scale cross-stratified sandstone (facies S-13). The hammer is resting on a reactivation surface (laminae of the overlying set truncate against the top lamina of the underlying set).

3.2.2.11 Mixed rippled and upper flat bedded sandstones, S-12

Facies S-11 may contain some flat bedded sandstone. As noted above, boundaries between ripple layers in facies S-11 tend to be fairly flat. Only in places, however, is upper flat bed parallel lamination visible at the bases of rippled layers. Where flat bed is present, it does not invariably pass up into ripples, but instead may be capped by a thin very fine sandstone or silty layer. The amount of flat bed present is very variable, and may dominate the facies, so that ultimately the facies is gradational with facies S-1.

3.2.2.13 Large scale cross-stratified sandstone, S-13

This facies is only seen at Five Mile Point. Cross-sets are up to 1.5 m in thickness (Figure 3.25). Although some large trough shaped set boundaries and trough shaped foresets are present, in general, crest lines of the original bedform seem to have been straight to slightly sinuous. Foreset bases are almost asymptotic to set boundaries. Foresets and set boundaries to different sets commonly meet one another at low angles, so it can be

difficult to distinguish different sets where the outcrop is small. Pebbles, if any are present, are only found at set bases.

3.2.2.14 Massive sandstone, S-14

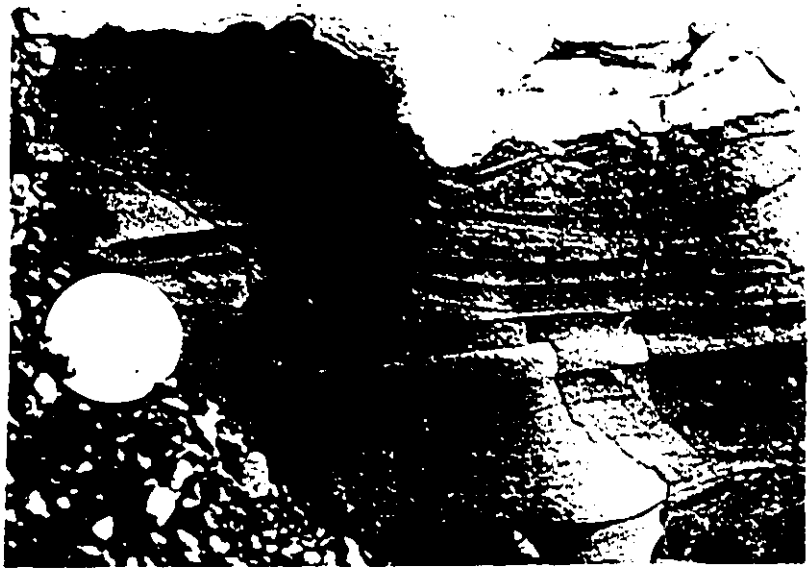
This is not an abundant facies, and it may be that certain of the beds labelled massive would show more structure if outcrop were better. All massive sandstones are fine, and rather better sorted than most of the Copper Harbor Formation, properties which tend to make structure, if present, hard to see. However, at Goodharbor Bay, where there are many fine sandstone tabular beds several centimetres thick, the structureless appearance is probably real (the depositional environment here would suggest that these massive sandstones are similar to turbidite A divisions). At Presque Isle, the section in the Freda Formation shows thicker massive sandstone beds (30 cm thick).

3.2.2.15 Thin graded sandstone beds, S-15

These are found in very fine or fine sandstones interbedded with silt (Figure 3.26). Graded beds are a few

Figure 3.26 Siltstone (red) with thin graded sandstone beds (facies S-I5), some of which are picked out by selective bleaching. Note the lateral continuity of the sandstones. This section coarsens upwards, and for the top 1 m or so is baked by heat from the overlying lava flow (grey with blocky fracture). The total height of the outcrop is about 7 m.

Figure 3.27 Parallel laminated siltstone (facies St-2). Note that some of the laminae are slightly erosive. This is one of the more sandy examples of this facies. The siltstone here is overlain by stromatolites. Quarter for scale.



centimetres thick or less, but laterally extensive, and are very slightly erosive. In their basal portions, which may be full of mud chips, the beds show upper flat bed lamination or are massive. Beds generally pass up into current ripples at the top of the bed. Less commonly, the top of the bed shows symmetrical wave ripples. These beds may be considered as individual layers of facies S-12, isolated in silt. Some outcrops show gradations between the two facies.

3.2.3 Siltstone facies

All siltstone facies are rare in the Copper Harbor Formation. Like the conglomerates and sandstones, the siltstones are a rich red colour, except where bleaching has formed white blotches. Nowhere in the Copper Harbor Formation are there rocks with clay but without significant amounts of silt (although this may reflect diagenesis rather than depositional process). As well as the facies outlined below, siltstone occurs throughout the Copper Harbor Formation as rip-up clasts and as drapes (commonly desiccated).

3.2.3.1 Massive siltstone, St-1

Absolutely massive siltstone is rare in the Copper Harbor Formation. There is generally sufficient very fine sand that some structure is visible. Of those siltstones that appear massive, many are at outcrops where conditions are not ideal, and where shaley parting is highly developed. It is suspected therefore that even these would show some fine lamination in good outcrops or in core. Nevertheless, it can be said that these beds in general contain less coarse silt and very fine sand than facies St-2 and St-3.

3.2.3.2 Parallel laminated siltstone, St-2

Most laminae are laterally extensive, but some are somewhat erosive and cut others (Figure 3.27). Laminae are generally less than 1 cm thick, but exceptionally reach 2 cm. Most are graded from finest very fine sandstone to silt or clay. The finer laminae may show desiccation cracks with coarser material from overlying laminae infilling cracks around curled up mudflakes. This facies occurs both as laterally extensive beds up to 30 cm thick (traceable 400 m at Horseshoe Harbor), and as channel fills. Where conglomerate overlies this facies, loading contorts the

laminae. In one small siltstone channel fill at Horseshoe Harbor a small slump has occurred at the side of the channel.

3.2.3.3 Rippled siltstone, St-3

Ripples in silt are picked out by their slightly sandier composition. Asymmetrical (current) ripples are most common, but wave ripples are also found at most outcrops where this facies occurs. Ripples may form laterally continuous layers (with approximately flat bases), or may be starved. The siltier laminae may show desiccation cracks infilled by sandier material from overlying rippled laminae.

3.2.3.4 Mixed rippled and parallel laminated siltstone, St-4

In places, there is a very close alternation between parallel laminated silt and rippled silt, giving rise to a mixed facies. This may be considered, depending on the proportion of sandy silt, as a very fine equivalent of facies S-12 or S-15.

3.2.3.5 Pebbly siltstone, St-5

At Goodharbor Bay, one siltstone bed 25 cm thick (and laterally extensive) was found, upon excavation, to contain dispersed rounded pebbles and cobbles up to 10 cm in length. The pebbles were supported by the matrix, and were volumetrically a minor part of the rock. In view of the sparsity of the pebbles, it may be that some of the massive siltstone beds at Goodharbor Bay, upon mining, would turn out to be pebbly siltstone.

3.2.4 Stromatolite-oncolite beds

On the northern Keweenaw Peninsula, near Dan's Point and at Horseshoe Harbor, algal beds are found. Stromatolite development is very variable, ranging from linked hemispherical calcite-rich mounds 30 cm high (Figure 3.28), through low mats a few centimetres or less thick with bumpy surfaces, to sandstones where calcite is very minor and the algal influence is only detectable as peculiar crinkly laminae. At Horseshoe Harbor, stromatolites form a horizon laterally continuous for more than 400 m. Stromatolites also occur as part of channel fills in conglomerate. Invariably

Figure 3.28 Plan view of stromatolite mounds. Lens cap for scale.

Figure 3.29 Oncolithic bed with pebble sized amalgamations of grains (centre of picture). Pen for scale.

associated with the stromatolites, underlying or overlying stromatolite beds, or as lateral equivalents, are oncolithic sandstones (Figure 3.29). Oncolites are also incorporated into the stromatolites along with sand and silt grains. In thin section, few grains are found to show laminae regular enough to merit the term oolith. Oncoliths are seeded on the (litharen[te) sandstone grains, and grapestone amalgamations are common. Some larger grains only show thick algal development above the grain, but some cobbles have completely envelopping algal coatings. Oncolithic sandstones may be trough cross-bedded, and isolated oncoliths, and algal coated pebbles may be found in sandstones and conglomerates with very little other organic input (these beds generally show large amounts of sparry calcite cement). However, a well developed stromatolite-oncolite layer is invariably nearby. The stromatolite mats show signs of a considerable ability to resist erosion, thin layers (1 cm thick) being laterally persistent, even where directly beneath coarse conglomerate. This suggests that the stromatolites resisted erosion rather like concrete.

3.3 Description and interpretation of facies associations

The Copper Harbor Formation consists almost entirely of five facies associations; the conglomeratic facies association, the conglomerate-sandstone facies association, the sandy facies association, the rippled sandstone facies association, and the silty facies association. Large scale cross-stratified sandstones (facies S-13) and stromatolite-oncolite beds have environmental implications of their own, and so are dealt with separately.

3.3.1 The conglomeratic facies association

As far as can be ascertained from outcrop, this facies association makes up about 70% of the Copper Harbor Formation. Some 80 to 90% of this facies association consists of massive, c-m supported, coarse conglomerate (facies C1). Imbricated, c-m supported, coarse conglomerate (facies C3), medium scale planar cross-stratified fine conglomerate (facies C8), and trough cross-stratified fine conglomerate (facies C10) are present from place to place, especially where the grainsize is somewhat finer. Sandstone lenses, tens of centimetres thick and several metres wide, are invariably present, and closely interfinger with the

conglomerate. The lateral and vertical relationships can best be seen by reference to Figure 4.5 in Chapter 4. Sandstone lenses may be cross-stratified, or flat bedded (facies S-1, S-2, S-4, S-5), although in many cases, the sandstone shows poorly developed low angle cross-stratification, perhaps indicating that dunes were beginning to form, but time or flow depths were insufficient. An alternative explanation is that these sandstones are the product of antidune deposition.

These sediments were deposited in a braided stream environment, the conglomerates as bars and the sandstones as channel fills. The conglomerate passes laterally into sandstone lenses with a contact that is almost invariably diffuse, conglomerate clasts dying out gradually, so that fingers of conglomerate reach into the sandstone lenses. The conglomerate beds themselves, upon close examination, are found to contain, at intervals of about 10 cm, very fine sandstone or siltstone drapes (maybe desiccated), and small sandstone lenses (several centimetres thick and tens of centimetres wide), maybe with cross-strata. For these reasons, it appears that the conglomerates were not deposited bed by bed by mass flow events. Each bed was built up by several depositional episodes, which were such that, at late stages of flow, sandstone deposition in channels and

conglomerate deposition on bars often went on simultaneously. If any debris flow deposition did occur, then debris flows must have been very thin (less than the distance between sandstone and siltstone partings in the conglomerate), and can only have deposited the rare beds which do not show a delicate lateral interfingering with sandstone lenses. Such a situation may exist in parts of the Mamainse Point outcrop, and possible debris flow deposition of these conglomerates is discussed in Chapter 4 with reference to the surveyed sections. The possibility that the biggest clasts in facies C-1 are a lag from debris flows now entirely reworked by high energy stream flow is discussed in Chapter 6 with reference to measured grain size distributions.

The general lack of structure in the conglomerate bars is probably a function of several factors. Firstly, the extremely poor sorting tends to obscure structure, and there is not a high proportion of tabular clasts to help emphasize imbrication or cross-bedding. Where structure is seen, it is almost invariably in finer, rather better sorted conglomerates. Secondly, the original barforms may not have had any angle of repose slip faces (perhaps similar to the diagonal and longitudinal bars of Hein and Walker, 1973). Thirdly, the conglomerate fabrics (see description of facies

C-I) show that sand, granules, pebbles, cobbles and boulders all were deposited simultaneously. This betokens very rapid deposition, without time for sorting to occur. Many clasts may not have rolled along the bed prior to deposition, so did not develop a transverse long axis orientation, and when clasts came to rest they did not have time to realign themselves before they were covered, and therefore were buried in relatively unstable positions. Thus, imbrication did not form. Imbricated conglomerates, as would be expected, are better sorted, are rather finer, and apparently were deposited more slowly. The massive, c-m supported, coarse conglomerate bars were probably deposited during stream flood conditions when sediment concentrations were high. It is not possible to say whether concentrations approached "hyperconcentration", as defined by Beverage and Culbertson (1964), since no criteria for establishing this are known (although Smith, 1986 has tried to make a start in this direction), but the fabrics indicate that the conventional hydrodynamic division between a rolling-saltating population of boulders, cobbles and pebbles, and a suspended population of granules and sand, is not meaningful in streams such as these were. Since sand was being deposited, sand was obviously present at the bed, and equally, it seems likely that pebbles and cobbles were

sometimes above the bed (indeed, could observation be made of such systems whilst in flood, it might be rather a subjective decision where to place the bed-sediment boundary). The possible reasons for this hydrological regime, with very flashy conditions and high sediment loads are discussed in Chapter 7.

From time to time, however, some bars did develop slip faces, giving rise to planar cross-stratification in the conglomerate. Gravel dunes sometimes formed, and produced trough cross-stratified conglomerate. The abundance of fine sandstone and siltstone drapes, some desiccated, is testimony to the ephemeral nature of the system.

This facies association was deposited high on the Keweenawan alluvial fans, upstream of the other facies associations.

3.3.2 The conglomerate-sandstone facies association

10 to 15% of the Copper Harbor Formation consists of this facies association. Conglomerate and sandstone facies are present in roughly equal proportions. The conglomerates are in general finer than in the conglomeratic facies association, and generally show more structure. All conglomerate facies may be present, with massive, c-m

supported, coarse conglomerate (facies C-1), imbricated, c-m supported or clast-supported, coarse conglomerate (facies C-3) and medium scale planar cross-stratified fine conglomerate (facies C-8) being the most common. Sandstones are generally, although not invariably, pebbly, with flat bedded and trough cross-stratified types being dominant (facies S-1, S-2, S-4, S-5, S-7, S-8). Trough cross-stratified very pebbly sandstone (facies S-6) and planar cross-stratified pebbly sandstone (facies S-10) are normally also present, and commonly grade laterally or vertically into conglomerate facies. In addition to close association of sandstone and conglomerate facies, there is generally a vertical alternation between mainly sandy and mainly conglomeratic facies on a scale of 5-15 m (further discussed in Chapter 4). Where thick section of the Copper Harbor Formation is seen in the Black River, Michigan, this facies association is stratigraphically between the conglomeratic facies association and the sandy facies association, passing into these somewhat gradationally.

These sediments were deposited in a braided river environment. Conglomerate barforms can be seen in Figure 4.10, Chapter 4, passing laterally into sand filled channels. Some conglomerate bars developed slip faces, giving rise to medium scale planar cross-stratified fine

conglomerate. Where sandstone deposited from suspension onto the foresets, angles of foreset dip were reduced and cobbles rolling from the bar to the top of the foreset tended to lodge near the brink, creating inverse grading. Where relatively sandy foresets prograded into a channel in which cobbles were moving, some cobbles rolled onto the base of the foresets, and normal grading resulted. Reworking of the tops of conglomerate bars created channel fills of finer, better sorted conglomerate (facies C-4, C-5). The small scale planar cross-stratified conglomerate perhaps formed in channels at low stages of flow when bars could not grow to great height. The large scale planar cross-stratified fine conglomerate (facies C-9) formed when large channels migrated laterally, and conglomerate and sandstone were alternately deposited at the channel margin. The origin of these large channels is discussed in Chapter 8. The trough cross-stratification in the sandstone indicates that this was not deposited by avalanching from the top of the bar, and the imbrication at Five Mile Point (Figure 4.16) indicates that the conglomerate, at least at the base of the sets, might also have been supplied from within the channel. The fine horizontally stratified conglomerate facies (C-11) seems to have formed within channels (perhaps from "diffuse gravel sheets" similar to those observed by Hein and Walker

(1977) in the Kicking Horse River). Symmetrical gravel waves only occur near stromatolites, and so are discussed separately.

This facies association was apparently a distal equivalent of the conglomeratic facies association.

3.3.3 The sandy facies association

This facies association makes up most of the remainder (10-15%) of the Copper Harbor Formation. Sandy facies form nearly all of the outcrop. Much of the section consists of an alternation between trough cross-stratified sandstones (facies S-4) and horizontally stratified sandstones (facies S-3) or upper flat bedded sandstones (facies S-1). Mixed upper flat bedded and trough cross-stratified sandstone (facies S-7) may make up part of the section. Rare medium scale planar cross-stratified fine conglomerate beds are present in some sections, associated with trough cross-stratified pebbly and very pebbly sandstones (facies S-5, S-6). At Union Bay, on and between the lines of sections, there are a few isolated sets of planar cross-stratified sandstone (facies S-9).

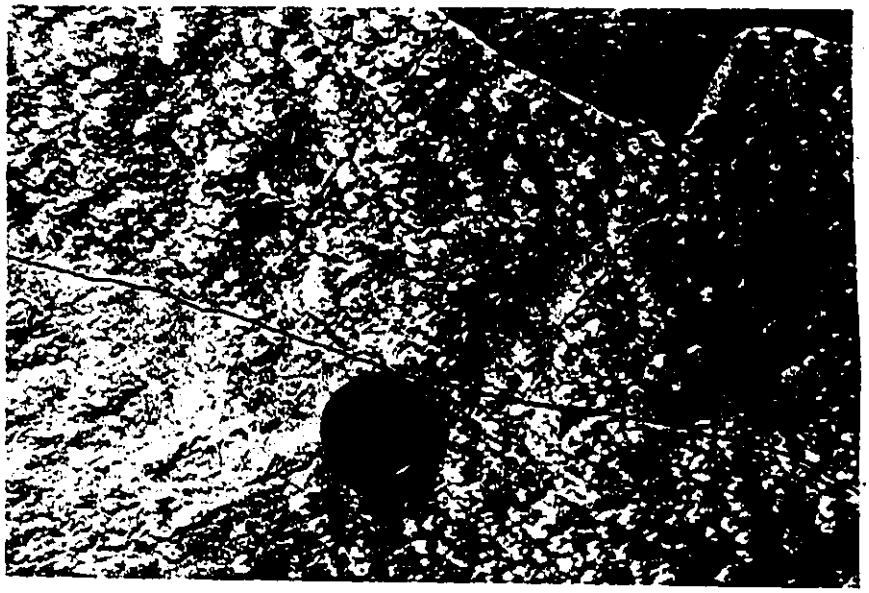
The depositional environment of these rocks is more distal than for the conglomeratic and sandstone-conglomerate

facies associations, and the braided rivers here were only rarely supplied with pebbly material. When pebbly material was present, conglomerate bars formed, and migration of slip faces gave rise to planar cross-stratification. For the remainder of the time, only sand was present. The close alternation between trough cross-stratified sandstone (S-4), horizontally stratified sandstone (S-3) and upper flat bedded sandstone (S-1), on a scale of 1 m or less, would imply that these facies were often lateral equivalents. The horizontally stratified facies contains abundant desiccation and very shallow water features, so each facies package was evidently deposited by several events on a relatively high part of the system (perhaps on bar tops, or in shallow channels across bars), whilst trough cross-stratified sandstone was deposited in deeper channels. "Knobbly surfaces" (Figures 3.30 and 3.31) have not been described elsewhere, and might either be dewatering structures akin to sand volcanoes, or might possibly be aeolian adhesion warts. Kocurek and Hunter (personal communications, 1985 and 1986) are not able to resolve this ambiguity. The appearance of the structures gives the impression of fluid having welled up from below, and would favour the former interpretation. The knobbly surfaces where the knobbly structures show a directional element (Figure 3.32) would then result from

Figure 3.30 Bedding surface covered with knobbly structures (discussed in the text) within the horizontally stratified sandstone facies (facies S-3). Notebook is 20 cm long.

Figure 3.31 A close-up view of the knobbly surface shown in Figure 3.30.

Figure 3.32 A knobbly surface within the horizontally stratified sandstone facies showing a directional element (as if water had flowed from left to right across the structures). Lens cap is 5 cm in diameter.



streamflow over submerged sand volcanoes. Bluck (1974) notes that sand volcanoes form in Icelandic braided outwash streams, and Williams and Rust (1969) note that such features may be submerged or subaerial, tending to form most often on bar flanks during falling stage, when bar porewaters are draining out. However, if this is indeed the origin of the knobbly surfaces, then it seems improbable that these would be most common within the horizontally stratified sandstone facies, which seems to have been deposited on the highest parts of the bars, where porewaters would not well up. The associated bubble sandstones and desiccated mud drapes (some with rain prints) indicate frequent subaerial exposure, and the wrinkle marks also indicate virtual dryness at intervals. All these features, which occur within a thickness of only 10 cm or so, would not be preserved if water depths at that point had often been sufficient to allow dunes to form, because migration of troughs would have wiped out such features. It therefore seems probable that the horizontally stratified facies (and knobbly surfaces) developed in small channels and on bar tops, whilst dunes, giving rise to trough cross-stratification, were restricted to deeper channels. These observations would tend to favour the adhesion wart interpretation of the knobbly surfaces.

As well as the close alternation between horizontally stratified sandstones (S-3) and trough cross-stratified sandstones (S-4), there is a weakly defined alternation between mainly trough cross-stratified parts of a section and mainly horizontally stratified parts of that section on a scale of about 5 m (ranging from 2 to 15 m). This is further discussed in Chapter 4. Trough sets are thickest in the predominantly trough cross-stratified part of the section, and planar cross-stratified sandstone (S-9), if present, is associated with the predominantly horizontally stratified parts of the system. Flat bedded sandstone ordinarily occurs as part of the horizontally stratified facies, but is also found in the parts of the section dominated by trough cross-strata, where it tends to be of somewhat coarser grainsize. This large scale of alternation is greater than the probable channel depths (as estimated from planar cross-strata, where seen, and by comparison with modern rivers such as the Rio Grande), and thus cannot be explained as an alternation between channel and bartop sedimentation. The system appears therefore to have periodically shallowed and deepened. Possible reasons for this are discussed later.

3.3.4 Rippled sandstone facies association

This facies association is a minor part of the Copper Harbor Formation (a few percent or less, as far as can be judged from outcrop, although more may possibly be present in the buried parts of the basin further from the rift margins). Most is seen on Isle Royale. On the mainland, the basal section of the Freda Formation at Presque Isle consists largely of this facies association. At Swedetown Creek, the Nonesuch Formation mainly consists of this association. The related genesis of the Copper Harbor Formation, the Nonesuch Formation, and the Freda Formation, has been dealt with at length by Elmore (1981), who, from the gradational contacts between these Formations and overall similarities in sedimentary style, inferred that these were all part of the same fan-floodbasin environment, which evolved as time went on.

Rippled sandstone and mixed rippled and upper flat bedded sandstone (facies S-11 and S-12) are dominant. Rippled silt, parallel laminated silt, massive siltstone and upper flat bedded sandstone (facies St-3, St-2, St-1, S-1) are also present. Massive sandstone, horizontally stratified sandstone, trough cross-stratified sandstone,and planar cross-stratified sandstone (facies S-14, S-13, S-4, S-9)

occur in minor amounts. In places, there is a fining upwards sequence from upper flat bedded sandstone to rippled sandstone to rippled siltstone (facies S-I to S-II to St-3) on a scale of about 2 to 3 m. More often, however, sequences are not evident or are only partially developed. The sequence does not represent deposition from a single flood event, being found to contain internally silt drapes, commonly desiccated. Rather, each ripple-silt couplet represents one flood event. It is more likely that the sequences represent approach or retreat of nearby channels (or some external influence, as discussed in Chapter 8). In the main, however, all the evidence is that relief in the system was very small, with flow depths rarely deep enough for trough cross-stratification to form. Channels were not abundant, and were shallow, with most flow occurring as sheetflood. Low bars occasionally developed to give planar cross-stratification. Migration of small dunes, scour hollows and steps in rippled surfaces (such as described for facies S-II) gave rise to small cross-strata from place to place. Flow was only intermittent, and desiccation was frequent. This floodplain environment was the distal equivalent of the braided river environments that gave rise to the coarser facies associations.

3.3.5 The silty facies association

This facies association is the least common. The section in interflow sediments at Goodharbor Bay on the Minnesota shore, and the section in the Nonesuch Formation at Potato River Falls, Wisconsin, contain the greatest thicknesses of this facies association.

Siltstone is volumetrically most important in this facies association, and may be massive, rippled or parallel laminated (facies St-1, St-3, St-2). About one quarter of the rock consists of turbidite-like sandstone beds a few centimetres or less in thickness (facies S-15). In sandier parts, rippled sandstone (facies S-11, S-12) may be developed. Sandstone bases are in some cases loaded into silts to produce flame structures. The upper part of the Goodharbor Bay section contains many desiccation cracks. In comparison with other facies association, the thin beds are here laterally very continuous (many tens of metres at least).

These rocks were deposited in a lacustrine environment (the possible origins of the lake are discussed in Chapter 8) by turbidity currents resulting from floods on the more proximal braided streams leading to the lake. Waves sometimes reworked the tops of the turbidite beds. The

pebbly siltstone bed at Goodharbor Bay (facies St-4) indicates that pebbly sediments at the lake margin were sometimes redeposited in deeper parts by debris flows.

3.3.6 Interpretation of the large scale cross-stratified sandstones (S-13)

This facies is interpreted as the product of aeolian transport and deposition. Sandstones of this type are only seen in the Five Mile Point sections. They are remarkable for the scale of the cross-stratification (up to 1.5 m), and for the absence of pebbles except at the set bases (in which respect they differ from the thickest trough sets seen elsewhere). These are features which would be expected in aeolian sandstones. Petrographic and palaeocurrent data provide strong support for this interpretation, and this evidence is discussed in detail in Chapters 5 and 6. The form of the sets and foresets, as discussed in section 3.2.2.13, indicates that the sand dunes responsible for deposition were straight crested to gently sinuous in form, with foresets flattening out to become asymptotic to underlying set boundaries.

Since this facies is found interstratified with the conglomerate-sandstone facies association, it is evident

that the dune field was not restricted to the most distal parts of the fans, but was present higher on the fans. This is not unexpected in an ephemeral fluvial environment, and abundant desiccation features indicate that flow on these fans was indeed very intermittent.

3.3.7 Interpretation of the stromatolite-oncolite beds

The formation of algal mats and oncolites necessitates standing water. As noted in section 3.2.4 ,this facies occurs both as channel fills and as laterally extensive sheets, indicating that from time to time, relatively extensive areas of the Keweenawan fans were flooded. The symmetrical gravel waves associated with the stromatolites result from oscillatory (wave) flow (Forbes and Boyd, 1987), and are additional evidence that the lakes must sometimes have had substantial fetch. The possible ways in which the water table on the fans could have been raised or lowered are discussed in Chapter 8.

CHAPTER 4

DESCRIPTION AND DISCUSSION OF MEASURED AND SURVEYED SECTIONS

4.1 Methods of outcrop description and problems encountered

The measured and surveyed sections of outcrops are here grouped and discussed according to the facies association they mainly comprise.

The emphasis of this study is elucidation of depositional processes and controls upon these. For this purpose it is necessary to have outcrops where sedimentary structures and facies relationships may be discerned. By no means all of the outcrops in the Copper Harbor, Nonesuch and Freda Formations fulfil this requirement, so the measured and surveyed sections described here were necessarily chosen selectively. In particular, the conglomeratic facies association is underrepresented, in that the thickness of section shown here is not commensurate with the proportion of the Copper Harbor Formation that consists of this facies association. This arises from the uninformative nature of most section in the conglomerate. Structure is difficult to see in the conglomerate, and requires especially good outcrop. Thick sections are available in some river gorges,

but these are generally lichen covered or occur where it is impossible to gain close access. Many coastal outcrops are good enough to make useful facies divisions, but the coastline is commonly parallel to strike, so the thickness of available section is small. In any event, the conglomeratic facies association and the conglomerate-sandstone facies association are but poorly described by vertical section because lateral variations are considerable (variations are normally very significant over 10m or less). In an endeavour to circumvent this problem, surveyed sections of some of the largest and best outcrops in these facies associations were constructed using compass and measuring tape. At most outcrops the bedding is dipping at 30° or so, and the outcrop surface is not horizontal. Photographs of the outcrop were taken wherever useful, but in general, appropriate vantage points were not available, and even those photographs obtained were not found sufficient to allow delineation of the facies boundaries, which are commonly ill-defined. It was therefore necessary to measure the positions of sedimentary features relative to some originally horizontal marker horizon, or, failing that, relative to a line positioned parallel to bedding. At some of the flatter outcrops, measurements could be made horizontally and later recalculated as stratigraphic

thicknesses by multiplying by the sine of the angle of dip. Other outcrops were better suited to direct measurement of stratigraphic thicknesses on surfaces perpendicular to dip and strike. In the latter cases, description of several such surfaces was required in order to cover the entire outcrop (i.e. the outcrop was stepped), and the different surveyed sections were joined at marker horizons. The position and size of a sedimentary feature (e.g. a sandstone lens) was obtained by measurement, and the feature was then sketched on the surveyed section. This method describes fairly well the relative positions, sizes and shapes of such features, and gives a relatively good impression of the overall aspect of the outcrop. However, it is necessarily rather inaccurate in terms of <u>absolute</u> thicknesses, particularly where the outcrop surface is uneven.

The considerable lateral variation of the sediments also makes the correlation between sections difficult. Correlation is, anyway, only meaningful over a relatively small distance, and most of the sections are too widely spaced. Dan's Point was the only location at which valuable lateral correlation could be made. Correlation problems are exacerbated by the occurrence of cross-faults. Butler and Burbank (1929), with the aid of much subsurface data, are able to document many small cross-faults on the Keweenaw

Peninsula cutting the basalt flows in the Houghton-Calumet region. This leads to the suspicion that such faults also exist in the overlying Copper Harbor Formation. At various outcrops on the Keweenaw Peninsula, small faults were indeed encountered. Only at Five Mile Point does throw appear to be more than a few metres, but it must be supposed that in the cover between outcrops other such faults exist.

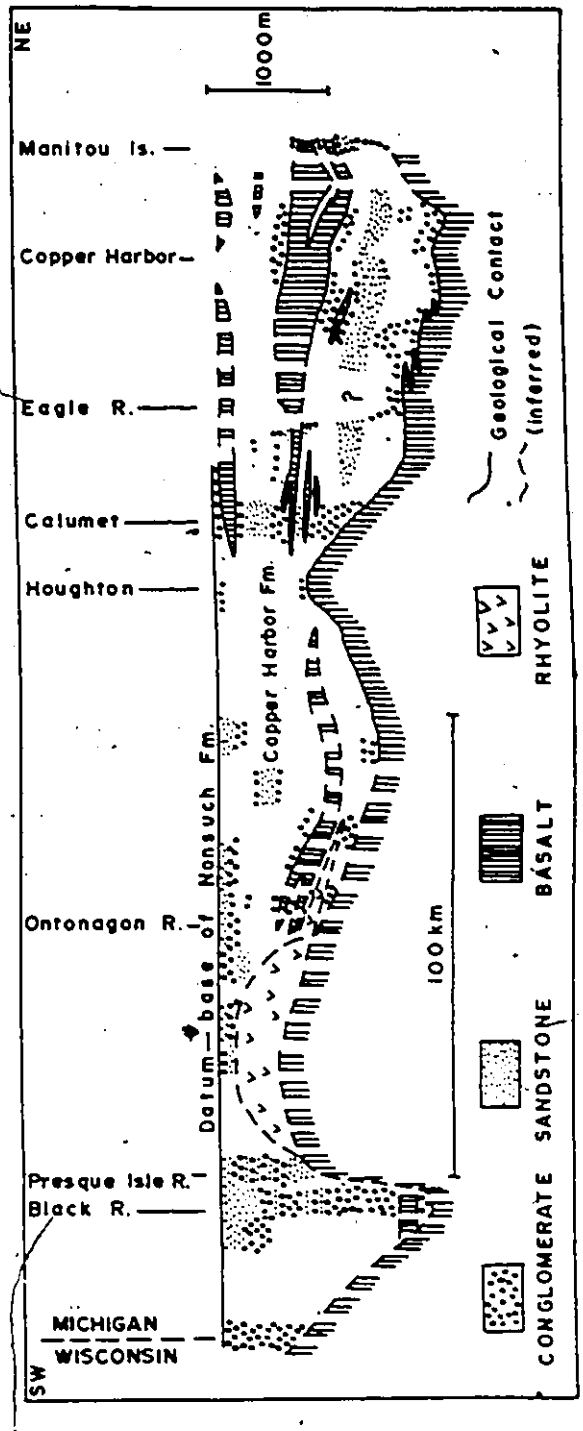
Estimated stratigraphic positions of the bases of the sections are given in Table 4.1, but the variations in the thickness of the Copper Harbor Formation along strike are considerable (see Chapter 2), so sections at the same stratigraphic height above the top of the lavas cannot be assumed to have been deposited at the same time.

The thick gorge sections through the Copper Harbor Formation have been studied by Elmore (1981), although unfortunately are not shown diagramatically. These give some impression of the overall variation in the sediments, but, for the reasons stated earlier, cannot provide much large scale information. White and Wright (1960), using data from various sources, produced a longitudinal section of the Copper Harbor Formation on the Keweenaw Peninsula. This is reproduced in modified form in Figure 4.1, to give an indication of the gross variation of the Copper Harbor Formation in this area. Near the Michigan-Wisconsin border,

Table 4.1 Approximate stratigraphic positions of the bases of sections described in this study. Calcualations were made with the aid of Quadrangle maps and other sources.

SECTION	STRATIGRAPHIC POSITION (GIVEN AS THE APPROXIMATE HEIGHT ABOVE THE BASE OF THE COPPER HARBOR FORMATION FOR SECTIONS WITHIN THAT FORMATION)
Presque Isle (10km east of the mouth of the Black River)	10m above the base of the Freda Formation
Swedetown Creek (near Houghton)	Within the Nonesuch Formation
Sandstone Falls (on the Black River)	850m
Rainbow Falls (on the Black River)	1300m
Union Bay (lower section)	800m
Union Bay (upper section)	925m (approximately the stratigraphic level of the top of the lower section)
Five Mile Point (lower section)	740m
Five Mile Point (upper sections)	860m
Eagle River	0m (base at the top of the Portage Lake Lava Series)
Dan's Point	1300m
Horseshoe Harbor	1300m
Rainbow Cove (Isle Royale)	460m
Francis Point (Isle Royale)	990m
Point Houghton (Isle Royale)	1310m
Saxon Falls (on the Montreal River)	Between lava flows near the top of the Portage Lake Lava Series
Goodharbor Bay	Between lavas of the North Shore Volcanic Group
Mamainse (maps only)	Between Middle Keweenawan lavas
Moffat Strait	Unconformably overlying the Sibley Formation, beneath lavas of the Osler Group.

Figure 4.1 Section of the Copper Harbor Formation along-strike on the Keweenaw Peninsula. Modified slightly from White and Wright (1960).


NE
SW
1000 m
Manitou Is.
Copper Harbor
Eagle R.
Calumet
Houghton
Ontonagon R.
Presque Isle R.
Black R.
MICHIGAN
WISCONSIN
Datum — base of Nonsuch Fm.
Copper Harbor Fm.
100 km
Geological Contact
(inferred)
RHYOLITE
BASALT
SANDSTONE
CONGLOMERATE

the diagram has been modified to include more information from section in the Black River (data of Elmore, 1981). The line of section passes south of White Pine, so the Union Bay area is not on the line of section. In the Union Bay area, the top 400m of the Copper Harbor Formation is sandy (Hubbard, 1975). Just west of Eagle River, the diagram has been slightly modified to include the sandstones found at Five Mile Point (data of this study), which underlie and overlie the extensive basalt flow shown in this portion of the diagram. Evidence for the laterally extensive sandstone body shown in this region is scanty. Cornwall (1954a), Cornwall (1955), and Cornwall and White (1955) find small scattered outcrops of this sand body, but these authors (upon whose work White and Wright, 1960, partly rely for their diagram) mainly infer its existence from a marked topographic low extending for more than 30km along strike. Much of this area could be underlain by shale, and some conglomerate is almost certainly present. The question mark on the diagram indicates a region where the supposed sand body lies offshore. The basalts have been detected offshore by aeromagnetic surveys, but this is not possible for the sandstone and conglomerate bodies. Where no pattern is present, the Copper Harbor Formation is presumed by White and Wright (1960) to be predominantly conglomerate.

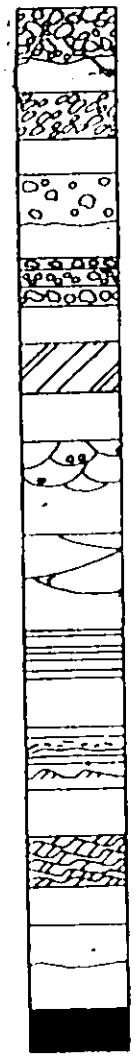
Various types of palaeocurrent measurement are presented with the measured and surveyed sections in this Chapter, but discussion of these is largely reserved for Chapter 5.

4.2 Measured and surveyed sections within the conglomeratic facies association

On Isle Royale, at the southern end of Rainbow Cove (location map of Figure 2.5), the conglomeratic facies association is exposed as low cliffs. This location is only accessible by boat or on foot from Windigo. The conglomerates here (Figure 4.3) show more structure and rather better sorting than generally seen on the Michigan mainland. Imbrication is present in several beds, and in these there is relatively good sorting of the cobble and small boulder sizes. Finer parts of the conglomerate show horizontal stratification (facies C-11), medium scale planar cross-stratification (facies C-8), or trough cross-stratification (facies C-10). The trough cross-stratified conglomerate at the 13.50 m level is a well-defined bed, predominantly formed from granule grain sizes, extending for more than 200 m laterally.

Figure 4.2 Index to the symbols used in the stratigraphic sections.

KEY TO SYMBOLS USED IN SECTIONS



massive, c-m supported or clast supported conglomerate

imbricated conglomerate

matrix-supported conglomerate

horizontally stratified conglomerate

planar cross-stratification

trough cross-stratification (set thicknesses approximately to scale), shown with pebbles at set bases

large scale cross-stratified sandstones (with set thicknesses drawn approximately to scale)

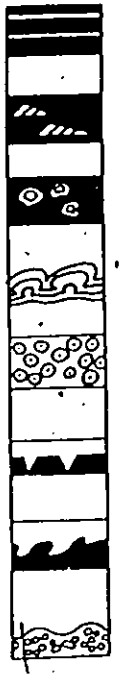
upper flow regime flat bed

the horizontally stratified sandstone facies, with upper flat bed, rippled surfaces, and scoured surfaces covered with mud chips

sandstones of the rippled sandstone facies

massive sandstone

massive siltstone



parallel laminated siltstone

rippled siltstone

pebbly siltstone

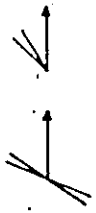
stromatolites

oncolites

desiccation cracks in siltstone filled with sandstone from the bed above

sandstone loaded into siltstone

symmetrical gravel waves

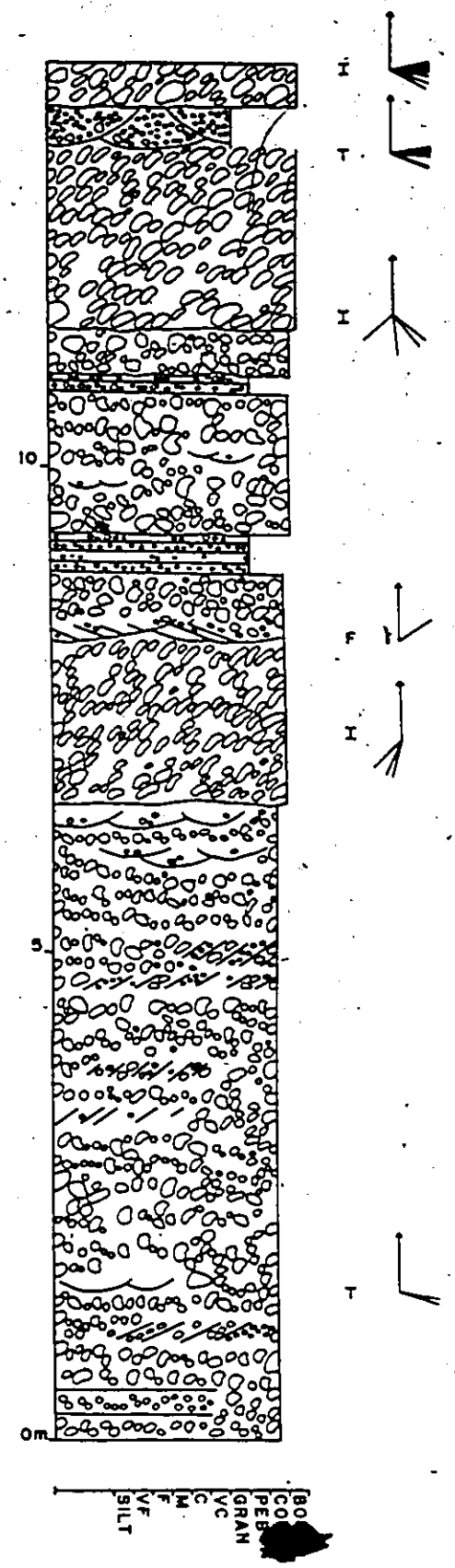


unidirectional and bidirectional palaeocurrent directions with north arrow

- T, trough axis measurements
- Tf, trough foreset dip direction measurements
- F, planar foreset dip direction measurements
- FB, upper flat bed lineation measurements
- R, assymetrical ripple measurements
- CC, current crescent measurements
- R&F, rib and furrow measurements
- I, imbrication measurements (please see text for explanation of method of measurement)

(due allowance has been made for tectonic tilt)

Figure 4.3 Stratigraphic section in the Copper Harbor Formation at the southern end of Rainbow Cove, Isle Royale.


0m
5
10
T
H
F
H
T
H
BO
CO
PEB
GRAN
VC
C
M
F
VF
SILT

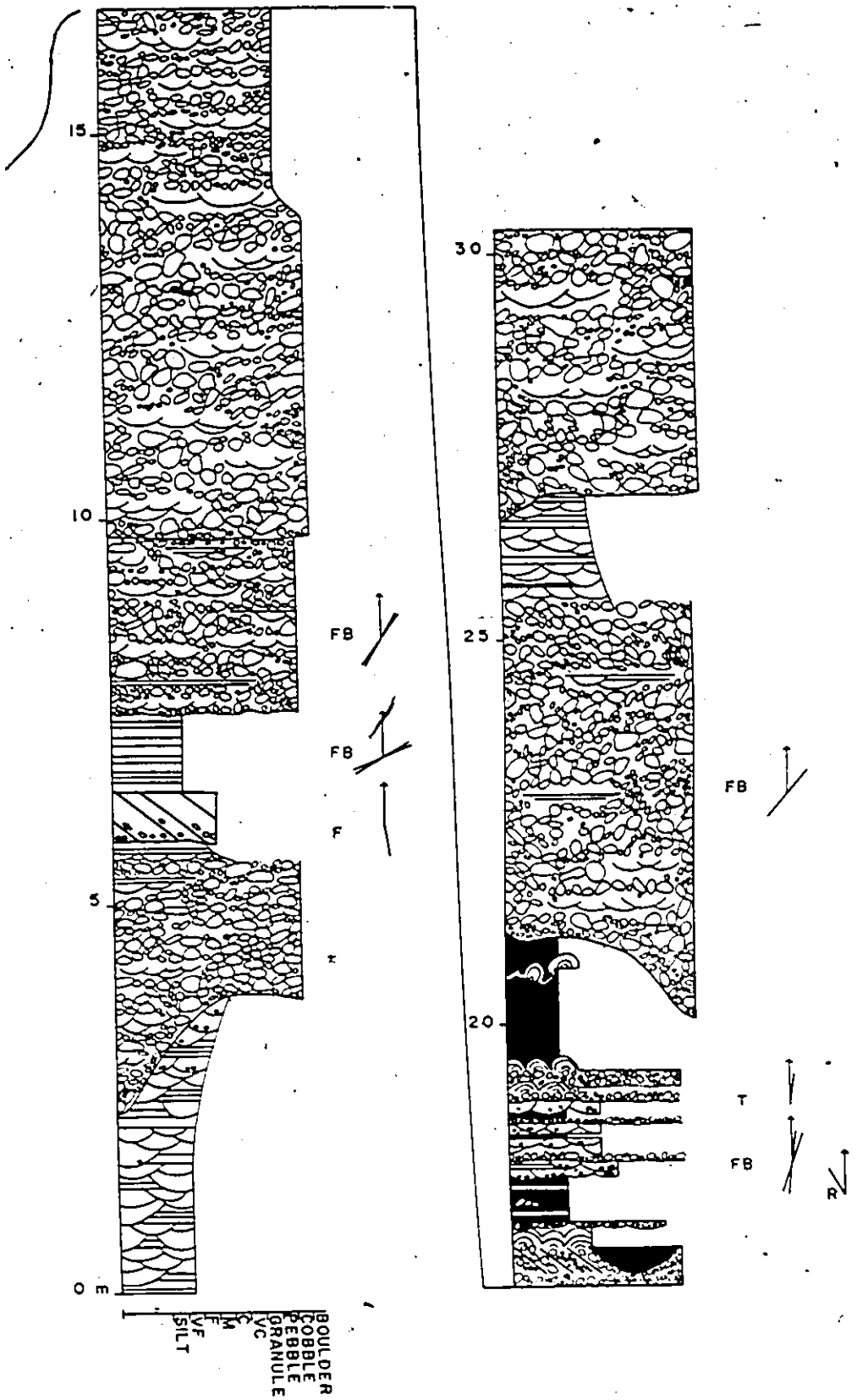
At the 8 m level, shallow scours appear to have been filled by a mixture of planar cross-stratified pebbly sandstone (facies S-10) and medium scale planar cross-stratified conglomerate (facies C-8).

In comparison, the section at Horseshoe Harbor on the Michigan mainland (Figure 4.4, location map of Figure 2.3) consists of proportionally more, massive, c-m supported, coarse conglomerate. The section is complicated by the occurrence of stromatolite-oncolite beds and siltstones near the 20 m level. At least one of the stromatolite horizons can be traced for 400 m laterally. The uppermost stromatolite bed and overlying siltstone have been distorted by loading due to the overlying conglomerate. This overlying conglomerate has an erosive base, with gutter casts indicating a north-northeasterly flow direction.

At the 17.50 m level, planar cross-stratification in the conglomerate passes laterally into a siltstone channel fill 7 m wide and 0.4 m deep. The channel fill consists of finely laminated mud, silt and sand, with several wave rippled surfaces, one slump fold, and small stromatolites at the channel base.

The lateral variations resulting from channelling are better described by surveyed sections, where these can be constructed (albeit laboriously). Outcrop at the headland to

Figure 4.4 Stratigraphic section in the Copper Harbor Formation at Horseshoe Harbor on the Keweenaw Peninsula, Upper Michigan.

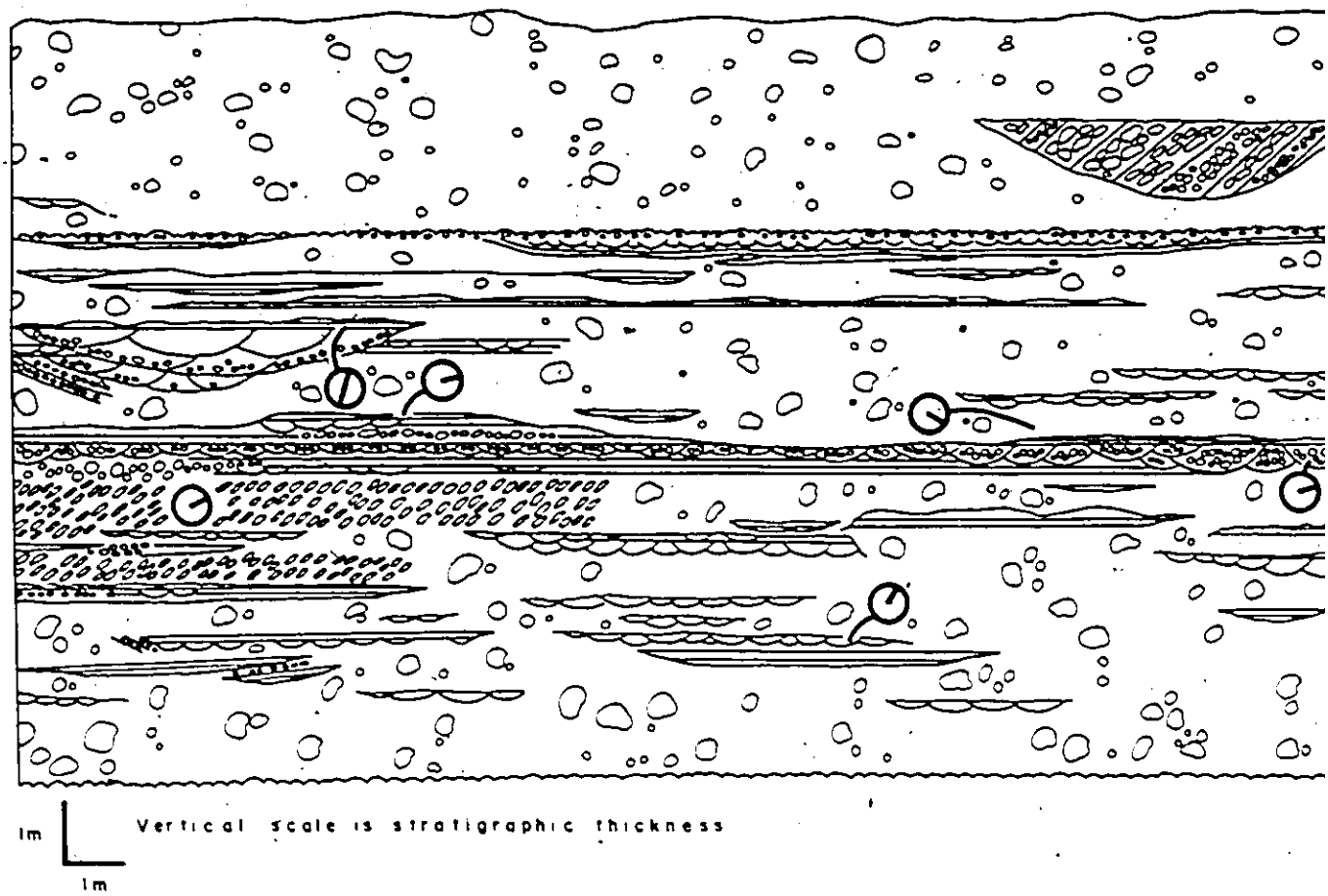

0
5
10
15
20
25
30
FB
FB
F
M
FB
T
FB
R
E
BOULDER
COBBLE
PEBBLE
GRANULE
VC
C
M
F
VF
SILT

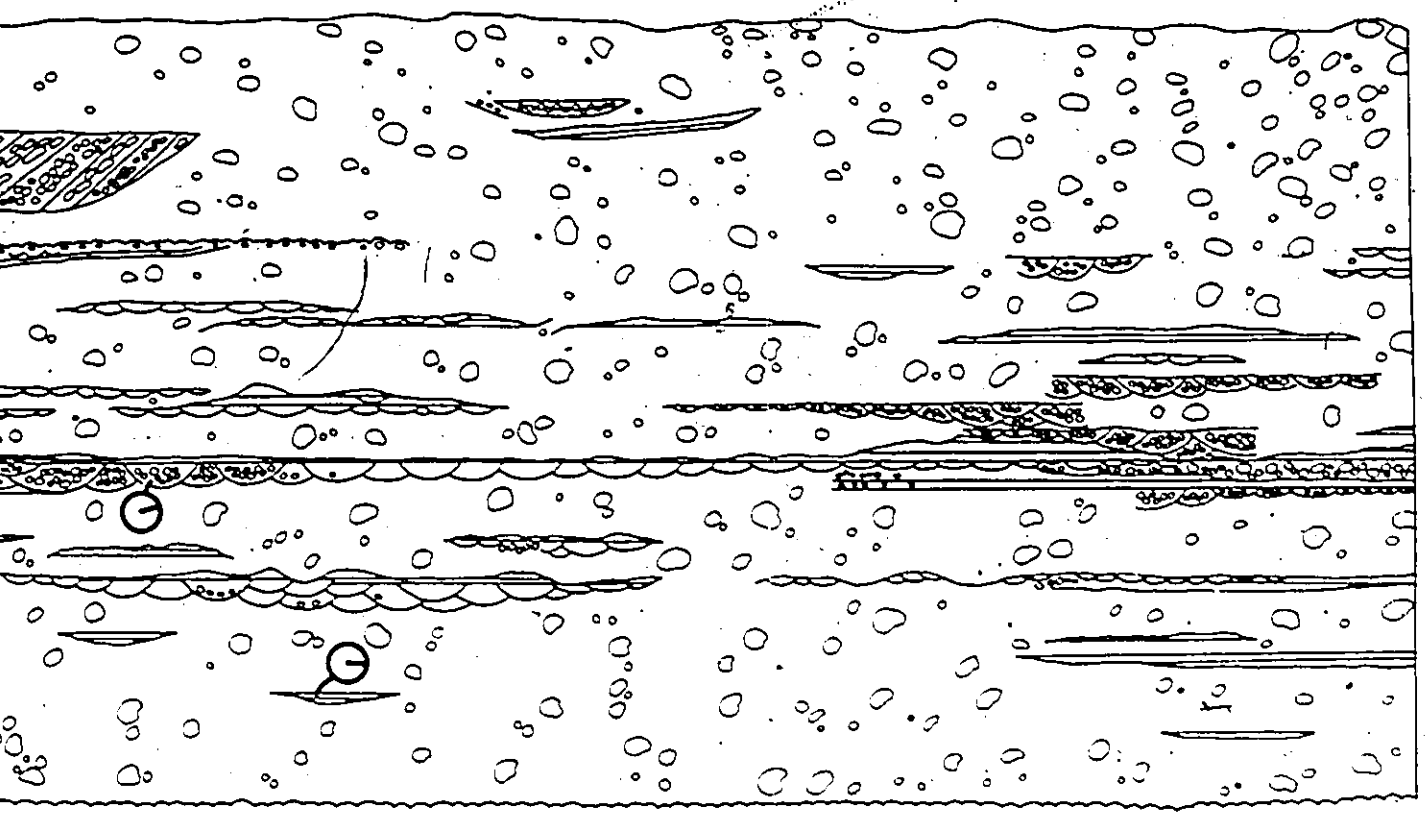
the east of Dan's Point, on the Keweenaw Peninsula, is suited to such treatment, and is shown in Figure 4.5. An attempt has been made here to preserve the facies symbols used in the sections (so, trough cross-stratification, for example, is everywhere drawn as troughs, regardless of flow direction relative to the plane of the diagram). However, for the sake of clarity, the massive, c-m supported, coarse conglomerate making up most of the outcrop is here drawn with relatively few clasts.

This particular outcrop shows a greater abundance of sandstone lenses than many outcrops in the conglomeratic facies association. The base of the diagram is at a laterally extensive stromatolite horizon. Another, less extensive stromatolite horizon occurs three quaters of the way up the diagram. This higher horizon is immediately overlain by symmetrical gravel waves very similar to those shown in Figure 3.10. Such features are also seen in the fine conglomerate in the west of the surveyed section immediately above the relatively extensive trough cross-stratified conglomerate bed. The crest lines of the gravel waves at these two horizons are 119° and 025° respectively.

The sandstone lenses give an impression of numerous shallow channels. Several lenses are present at most stratigraphic levels, and some of these probably existed at

Figure 4.5 Surveyed section from outcrop in the Copper Harbor Formation to the east of Dan's Point on the Keweenaw Peninsula, Upper Michigan.


1m
1m
Vertical scale is stratigraphic thickness



unidirectional palaeocurrents

bidirectional palaeocurrents

(up is north)
(plane of diagram is 276°)

the same time (i.e. a braided river environment). The lenses are not invariably horizontal. For example, the highest sandstone lens depicted, and the flat bed within, dip west, as if a channel was partly filled with sand at its eastern edge, but was finally filled with conglomerate showing no grain size contrast with the adjacent bar. In the upper part of the diagram, another conglomerate channel fill exhibits well defined planar cross-strata, indicating the migration of the adjacent bar slipface into the channel. This channel fill is easily seen because it has an open-framework texture filled with calcite.

Trough cross-stratified conglomerate is seen both as a laterally extensive bed in the lower centre of the diagram, and as discontinuous beds probably reflecting partial channel fills. It should be noted that the flow direction in the apparently extensive bed is close to the plane of the diagram, so this bed may show a channel form in another section.

Silty drapes occur at various levels in the outcrop, indicating periodic cessation of flow. Close above the extensive trough cross-stratified horizon there is bubble sandstone and a rippled surface with rain prints, so the rivers at times dried up completely.

The Mamainse Point outcrops, at the eastern end of Lake Superior (location map of Figure 2.1), contain fewer sandy lenses than the Dan's Point outcrops. Figure 4.6 shows part of the outcrop where sandstone is very minor. Both this surveyed section and the surveyed section in Figure 4.8 are from parts of the thickest interflow conglomerate package at Mamainse Point, which has a total thickness of some 300m. Figure 4.6 was constructed rather differently to the other surveyed sections in this Chapter, because the outcrop here was too uneven to survey. Instead, two sections were measured about 20 m apart, then the beds were walked out between the sections and sketched on the diagram.

No imbrication is seen at this outcrop, but flow directions at other outcrops at Mamainse would tend to indicate that flow was towards the southwest (i.e. not very different to the plane of the diagram). Beds may not therefore appear so continuous in a section at right angles to this diagram.

Sorting of clasts is worse than indicated by the diagram. The uppermost of the two very coarse conglomerate beds contains boulders up to 176 cm long (Figure 4.7). Neither of the two coarsest beds towards the top of the diagram are, however, the result of a single depositional event. Both contain sandy lenses, generally rather finer

Figure 4.6 Surveyed section from outcrop in interflow sediments at Mamainse Point, Ontario.

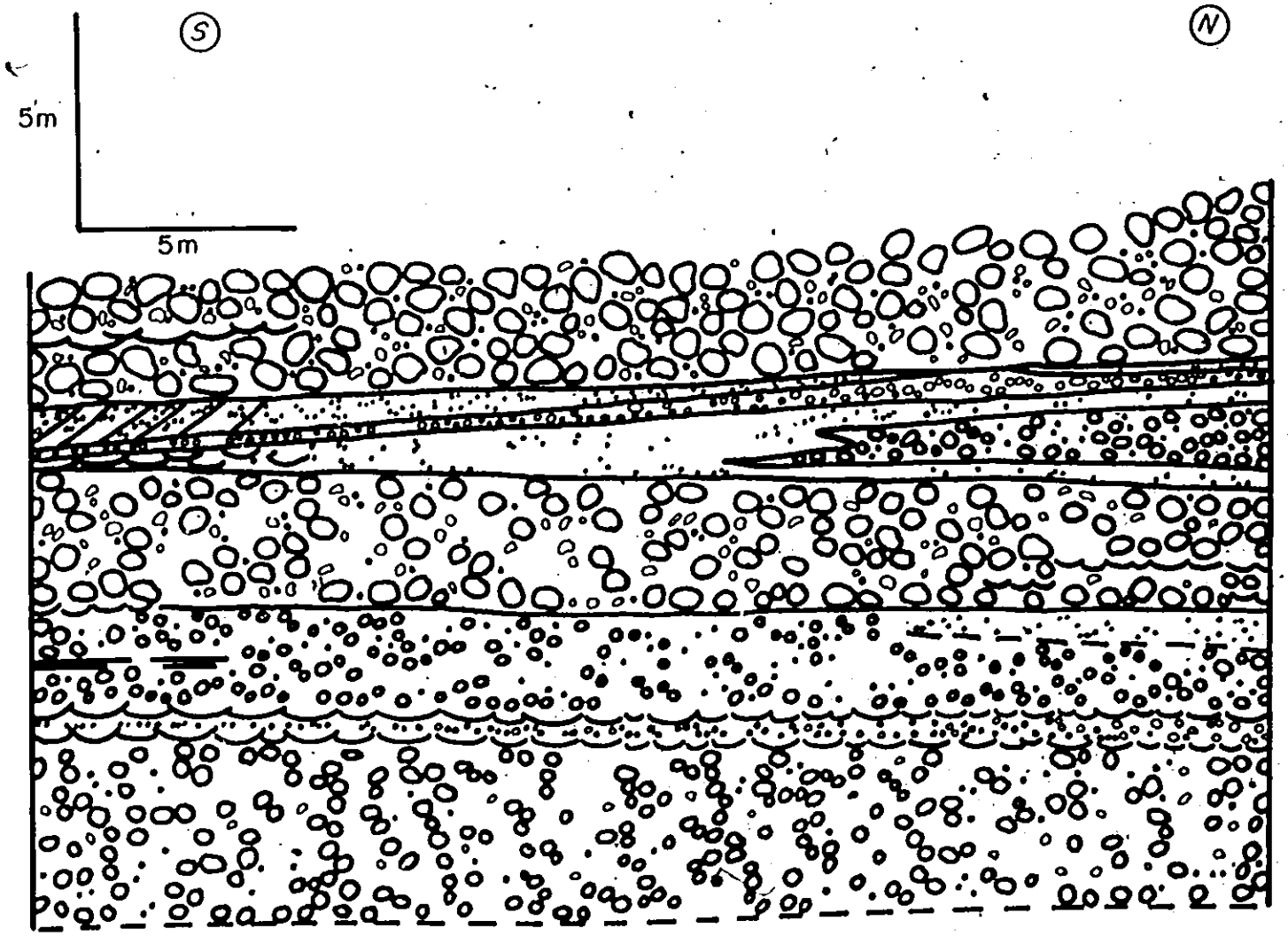
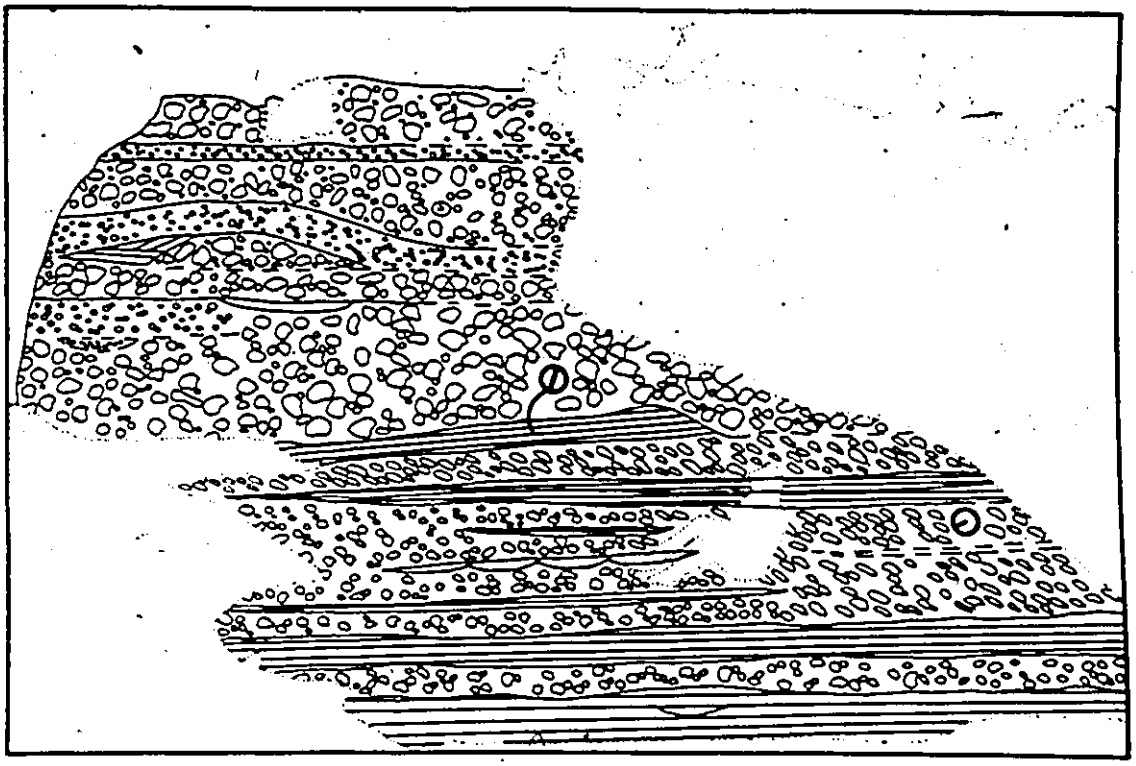

S
N
5m
5m

Figure 4.7 Photograph of the uppermost very coarse conglomerate bed shown in Figure 4.6. Note the exceptionally large boulders. Field assistant for scale.

Figure 4.8 Surveyed section and photograph of outcrop in interflow sediments at Mamainse Point, Ontario.



than the sandy matrix of the conglomerate, and other lenses are present but are too small to be shown on the diagram. These observations militate against a debris flow origin for these beds, but cannot rule out the possibility that deposition was by several thin debris flows, the tops of which were later fluvially reworked to produce sandstone lenses. The c-m supported fabrics of these conglomerates require that such debris flows were very clast-rich and poor in mud. Debris flow deposits of this type and deposits from concentrated streamflood are likely to be very hard to distinguish. The question also arises as to whether these were once more muddy debris flows but were entirely reworked by streamflow. The extremely poor sorting does not support this interpretation, unless it is supposed that the streamflow responsible for reworking was very high energy and waned very fast. This, however, leads back to a variant on the streamflood theme. A point count for grainsize was performed on the lower of the two very coarse conglomerate beds, and does not provide any evidence for reworking (in the form of a lag). This is further discussed in Chapter 6. It is therefore concluded that the dominant mode of deposition at this location (and for much of the conglomeratic facies association) was intense streamflood, perhaps with high sediment concentrations, which waned fast.

Figure 4.6 then indicates that several such events alternated with periods of lower energy deposition producing finer conglomerate with more structure (e.g. planar cross-stratification).

Stratigraphically below this, more of the same interflow conglomerate package is exposed just above the underlying basalt flow. Part of this outcrop was surveyed using the methods described in section 4.1. In this instance, it was also possible to obtain a photograph of the outcrop presenting an image approximately perpendicular to dip and strike. The final surveyed section was then prepared by tracing over the photograph then inserting facies symbols, etc., using the field notes. Both the photograph and the surveyed section are presented in Figure 4.8. This gives a fair indication of the degree of interpretation involved in, and the accuracy of the surveyed sections shown elsewhere in this Chapter.

The outcrop here has more sandstone lenses than in the stratigraphically higher outcrop described previously. Most lenses are approximately flat-bedded (facies S-1, S-2), but some cross-stratification, probably troughy, is present. Three quarters of the way up the outcrop, sandstone cross-stratification has apparently formed in the lee of a cobble-boulder mound (presuming that the palaeocurrents from

imbrication and flat bed lineation lower in the outcrop also apply here).

4.3 Measured and surveyed sections within the conglomerate-sandstone facies association

20 km east of the Michigan-Wisconsin border, the Black River gorge cuts through the entire thickness of the Copper Harbor Formation (here 1605 m according to Elmore, 1981). A good deal of the gorge section is poorly exposed or completely inaccessible to the prudent. At Sandstone Falls and Rainbow Falls the torrent broadens and the topography becomes less daunting. Figure 4.9 shows section measured at Sandstone Falls. Conglomerate and sandstone are present in roughly equal proportions. Most of the conglomerate is massive and c-m supported (facies C-1), but some imbrication is present near the base of the section. At the 52 m level an erosively based conglomerate bed wedges out into sandstone. This conglomerate appears to have had positive relief at the time of deposition (i.e., it was a barform). The same area is shown in plan view as a palaeocurrent map in Figure 5.5 of Chapter 5.

The sandstones show well developed trough cross-stratification and flat bed (facies S-5, S-1, S-2). A few

Figure 4.9 Stratigraphic section in the Copper Harbor Formation at Sandstone Falls on the Black River, Upper Michigan.

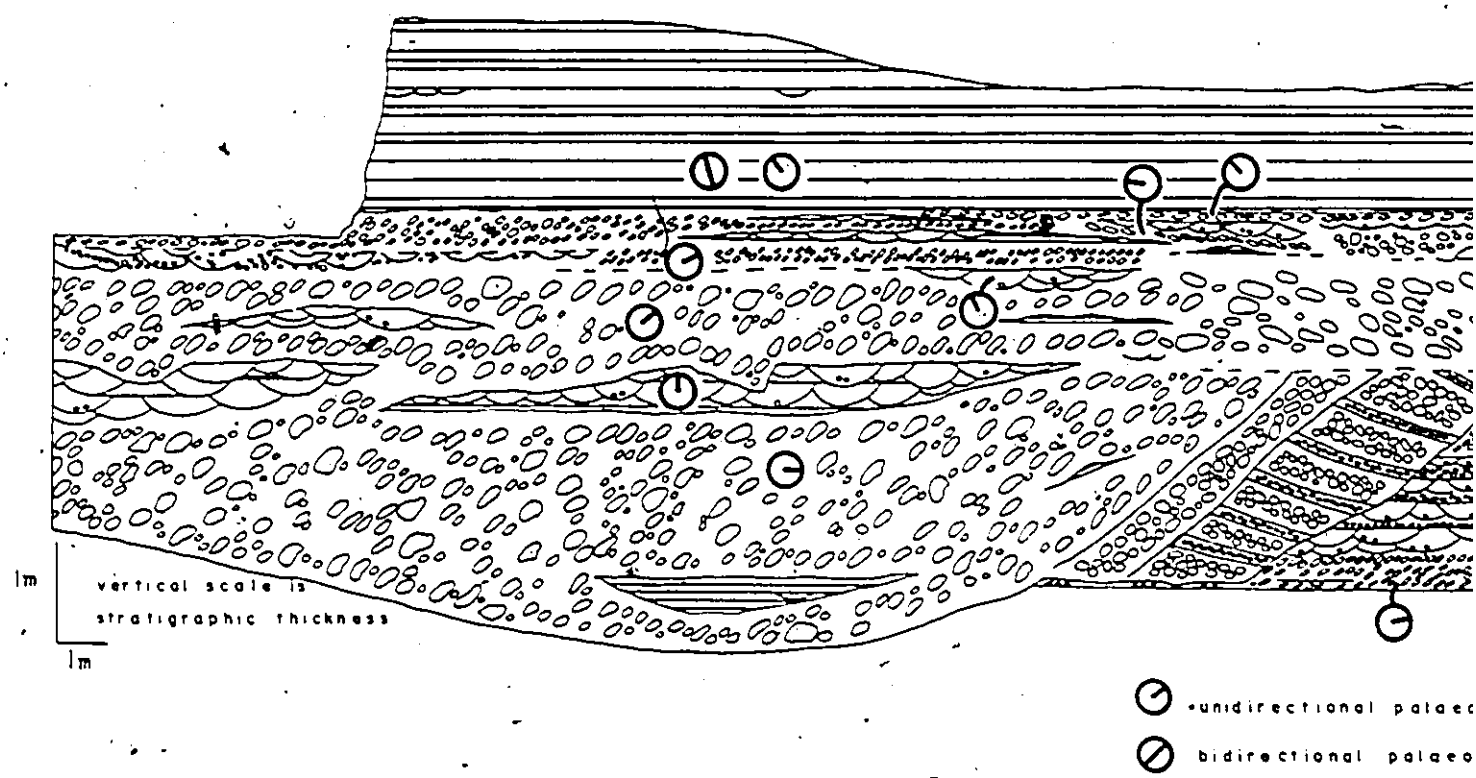
BOULDER
COBBLE
PEBBLE
GRANULE
VC
C
M
F
VF
SILT

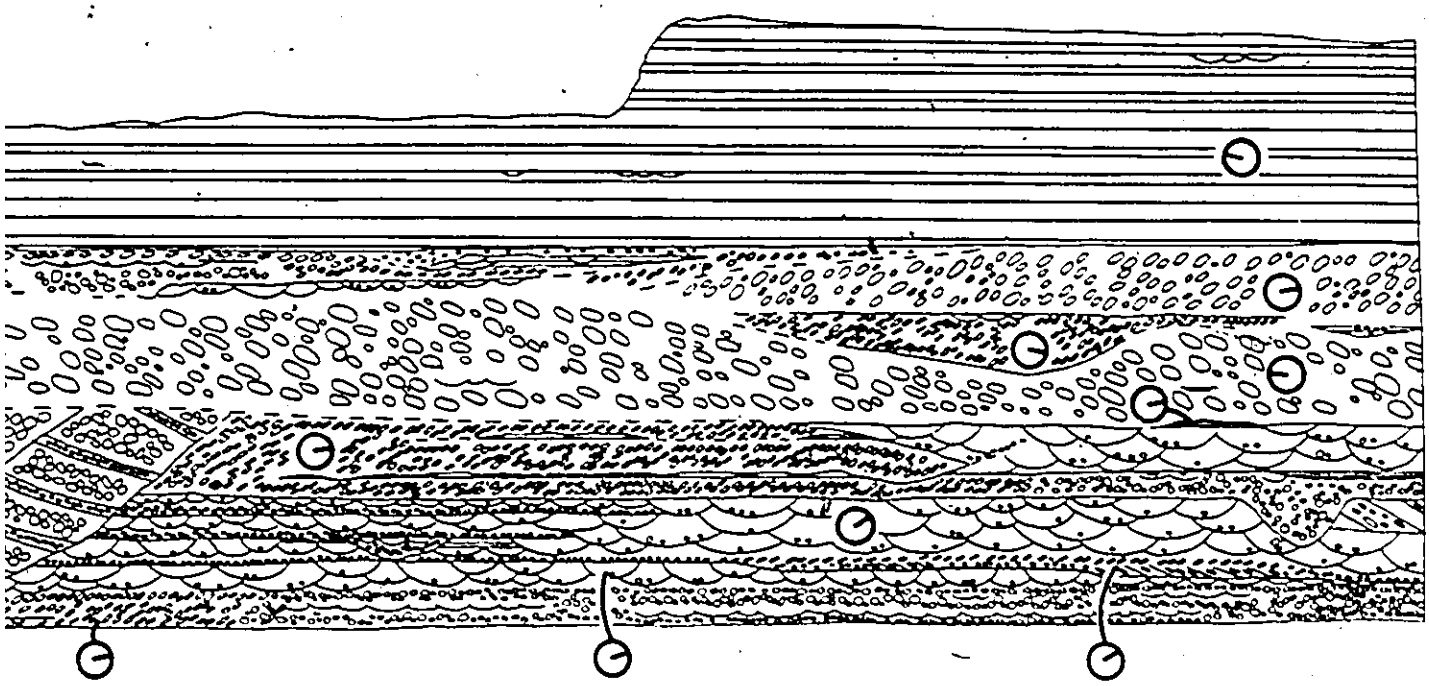
troughs show final fills of very fine sandstone, in some cases current rippled. Mud drapes are abundant.

The basal 7 m of the Sandstone Falls section is suitable for surveying. Figure 4.10 gives a much better impression of the aspect of the outcrop than the vertical section (which was measured at the eastern end of the outcrop). The most striking feature is the large conglomerate-filled channel in the western portion of the outcrop. This channel is about 2.5 m deep, and has a complex fill. At the eastern margin there is planar cross-stratified conglomerate. The planes of the foresets here seem to be approximately perpendicular to the channel wall (they are drawn as they appear in outcrop), as if this was a barform at the channel margin accreting downstream. This conglomerate is truncated to the west and overlain by massive fine conglomerate. This in turn is overlain by imbricated, c-m supported to clast-supported, coarse conglomerate (facies C-2) which constitutes most of the channel fill. The small flat bedded sandstone lens dipping parallel to the channel margin is evidence that filling continued to progress laterally by a succession of depositional events.

In this, and the other surveyed and measured sections, time (and in places outcrop quality) did not

Figure 4.10 Surveyed section from outcrop in the Copper Harbor Formation at the base of the measured section shown in Figure 4.9.


1m
vertical scale is
stratigraphic thickness
1m
unidirectional palaeo
bidirectional palaeo



tional palaeocurrents (up is north)

ional palaeocurrents (plane of diagram is 254°)

permit exhaustive point counting analysis of clast axis orientations in the imbricated conglomerates. The flow directions shown were obtained by choosing at random within a particular bed about 10 clasts which formed part of the dipping imbricate fabric and for which long axis orientations could be measured. The extensive coarse conglomerate bed in this Figure apparently shows a palaeocurrent reversal. This might be the result of accretion to both sides of a bar form now seen in cross section, or could be the result of accretion on two bars in areas of different flow directions. Note that this conglomerate bed is interrupted by a number of sandstone lenses, attesting to accretion during several depositional events. The top of this conglomerate has been reworked by lower energy flow to form an imbricated, clast-supported, fine conglomerate scour fill. Parts of this fill show shelter structures.

The pebbly sandstone in the lower eastern portion of the diagram shows well-developed trough cross-stratification. These sandstones pass laterally into fine, generally imbricated conglomerates that appear to have been low bar forms or gravel sheets.

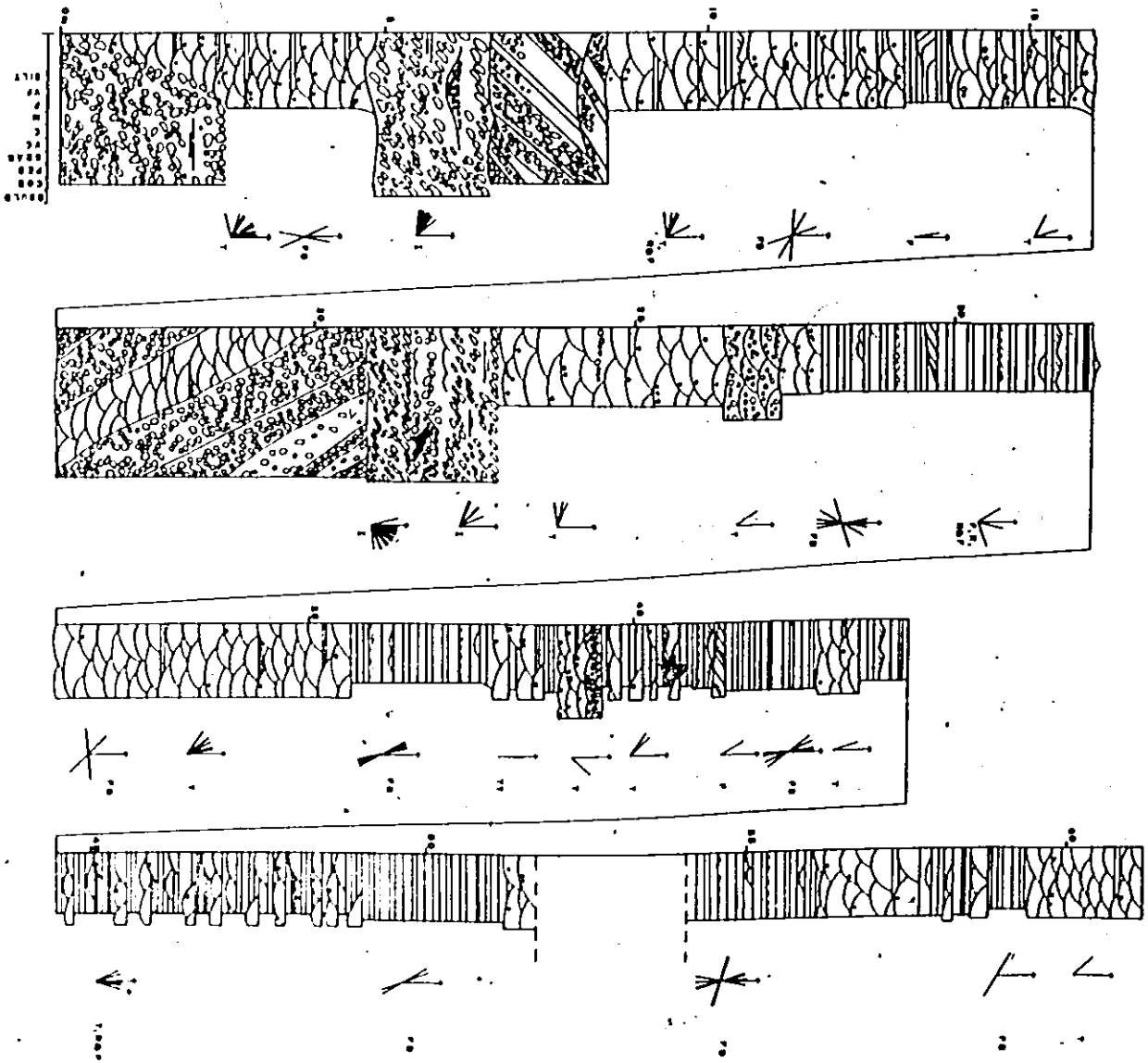
The thick sandstone body at the top of Figure 4.10 and the thick trough cross-stratified sandstone body at the

15-20 m level in Figure 4.9 can be correlated laterally over 100-200 m (at least), although outcrop is not continuous. These beds may therefore be similar in extent to that seen at Dan's Point (discussed below). The possible factors causing these changes in the style of sedimentation are discussed in Chapter 8.

The Rainbow Falls section is stratigraphically above the Sandstone Falls on the Black River. This section (Figure 4.11) covers the transition from the conglomerate-sandstone facies association to the sandy facies association. Above the top of this section, the Copper Harbor Formation is almost entirely sandy all the way to its contact with the Nonesuch Formation at Black River Harbor. Towards the top of the Rainbow Falls section there is a general decrease in grain size, and an increase in the amount of the horizontally stratified sandstone facies (S-3), reflecting a trend towards a shallower environment.

A remarkable feature of this section is the large scale planar cross-stratified fine conglomerate (facies C-9) at the 20 m level (see also Figure 3.8). This presumably reflects some sort of lateral accretion within an exceptionally deep channel. Cornwall (1954b) makes mention of "foreset beds of enormous size [that] occur near the base of the Copper Harbor Conglomerate at the eastern edge of the

Figure 4.11 Stratigraphic section in the Copper Harbor Formation at Rainbow Falls on the Black River, Upper Michigan.



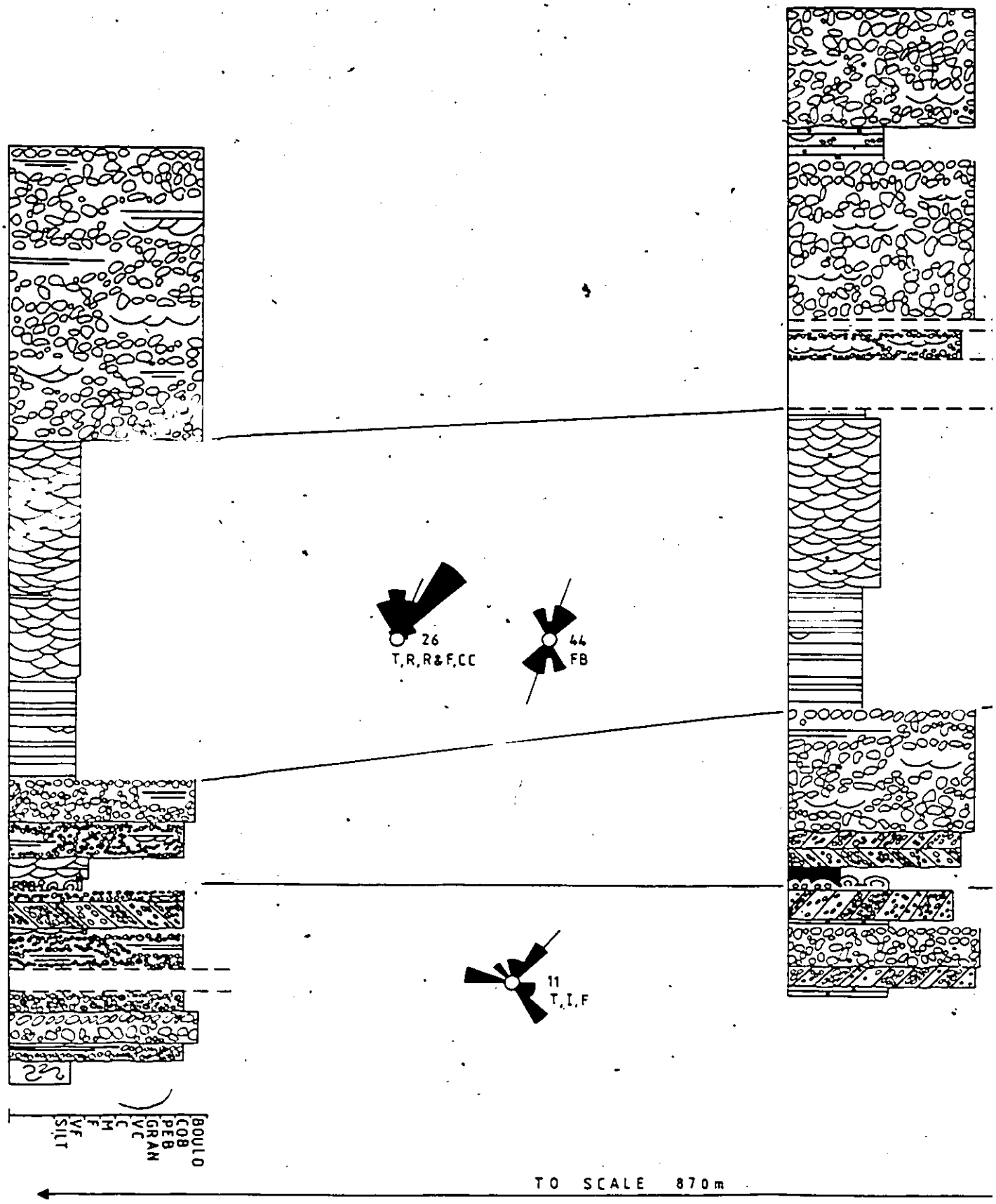
[Delaware] Quadrangle, where strikes of individual conglomerate ridges diverge appreciably from the strike of the basalt contact". These beds may be similar to those at Rainbow Falls.

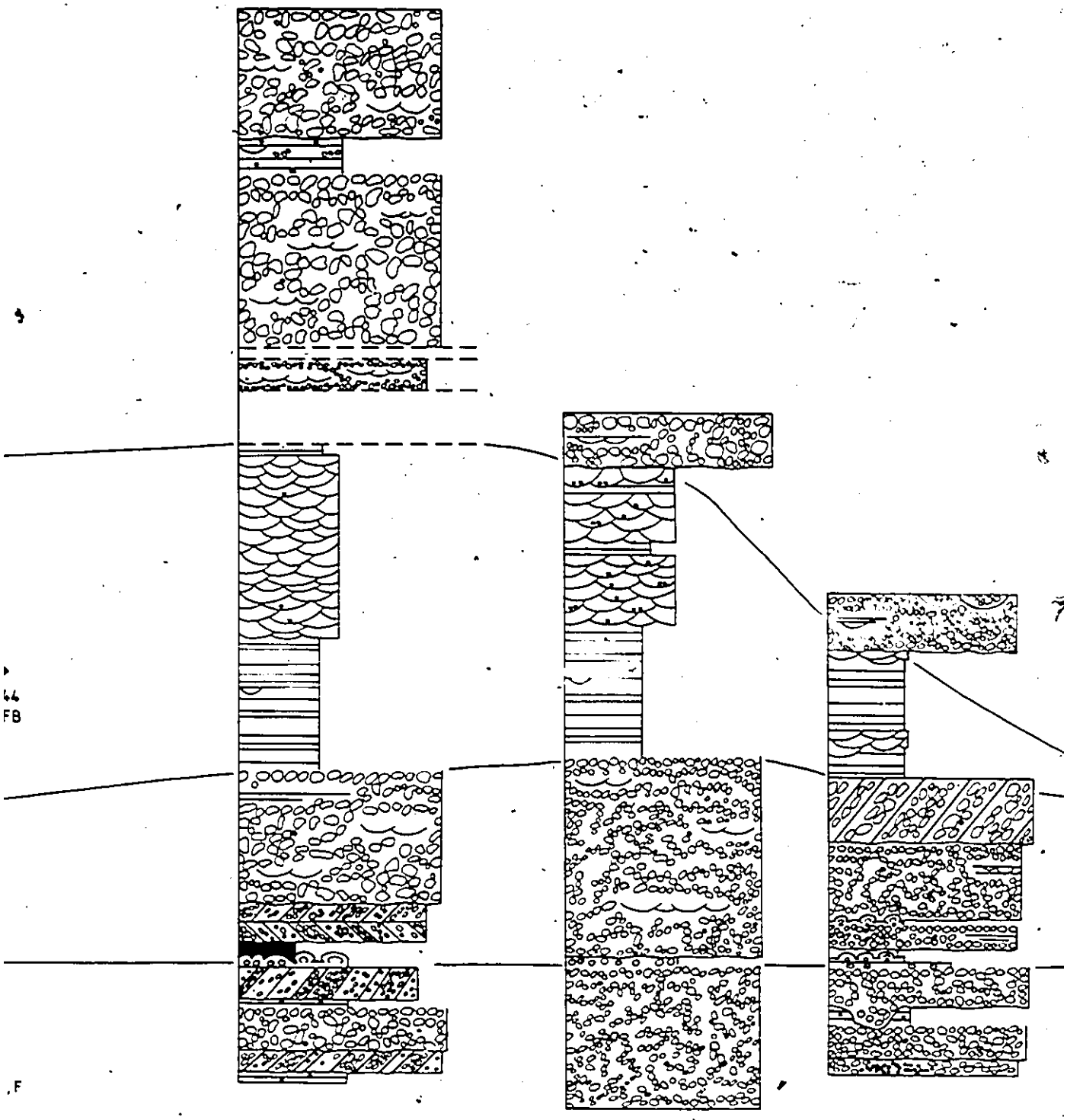
Medium scale cross-stratified fine conglomerate (facies C-8) occurs as channel fills at the 7 m level. Planar cross-stratified sandstones occur at several levels in the section (13 m, 29.50 m, 42 m), and are laterally equivalent to flat bedded sandstones. Planar foresets in some examples level out into such flat bed. In general, as at Union Bay, the planar cross-stratified sandstones are associated with the horizontally stratified sandstone facies (S-3).

Mention was made above of the laterally extensive sandstone body at Dan's Point. The indentation of the coastline in this area exposes the same section at five places along the strike, providing information about lateral facies variations over a total distance of 2.5 km. All the sections are readily accessible from the coast road.

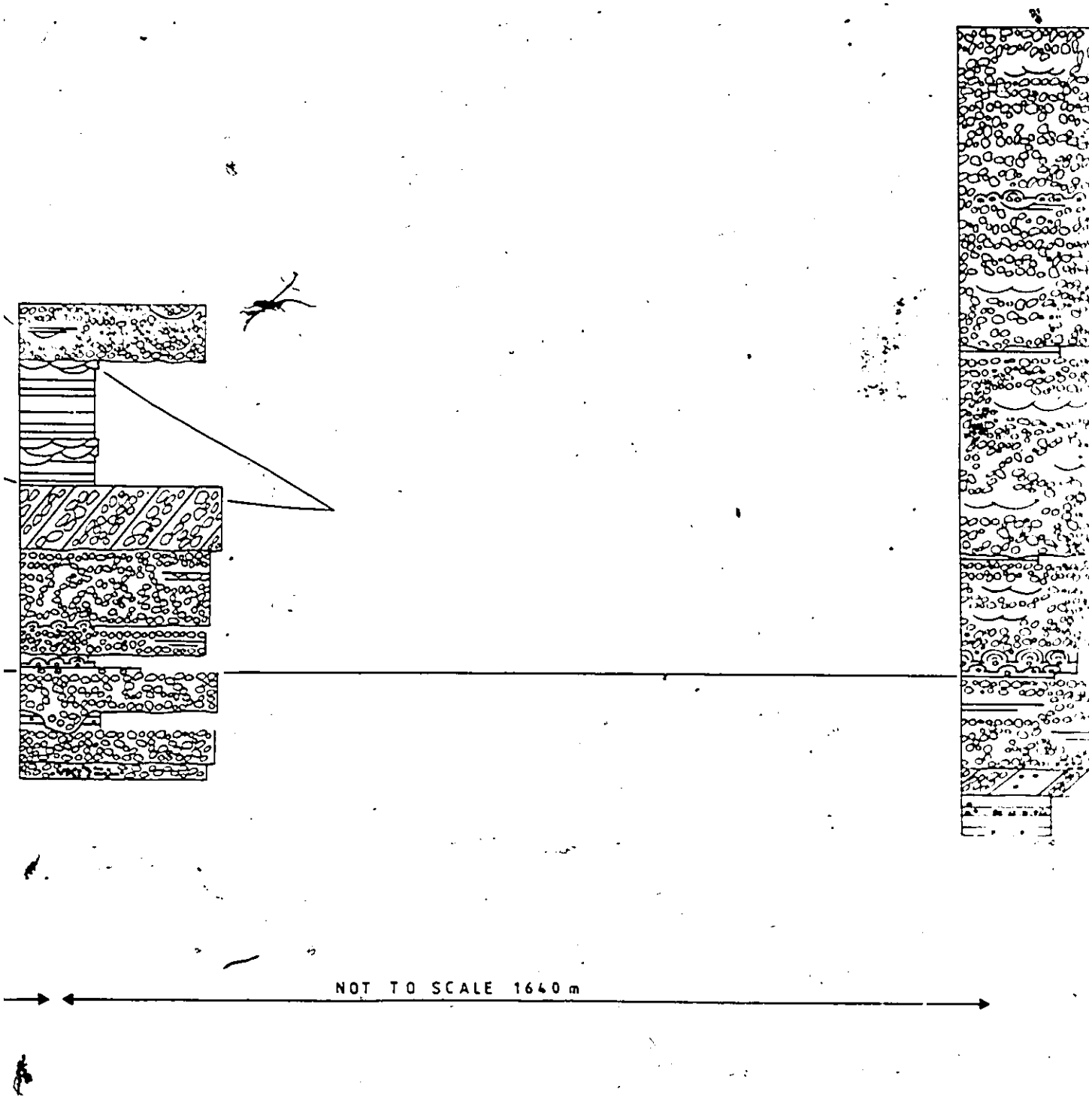
The datum used in construction of Figure 4.12 is a laterally extensive stromatolite-oncolite horizon. In the western part of the area (drawn to scale in the diagram), this can be easily located, although several small cross-faults (with throws of a few metres or less) have offset

Figure 4.12 Correlation of sections in the Copper Harbor Formation exposed near Dan's Point on the Keweenaw Peninsula, Upper Michigan.


26
T,R,R&F,CC
44
FB
11
T,I,F
SILT
VF
F
M
C
VC
GRAN
PEB
COB
BOULD
TO SCALE 870m



SCALE 870m


NOT TO SCALE 1640 m

certain of the sections in this area. Figure 4.13 is an aerial view of one of the outcrops in this area, showing the stromatolite horizon. The correlation between the western part of the diagram and the easternmost section (across an intervening bay) is somewhat less secure, but several observations support the correlation as made in Figure 4.12. Firstly, it is possible to walk someway down-section at the western side of the bay, and there are no other stromatolites present which could confuse the correlation. Secondly, it is possible to sight along strike across the bay, which affirms that the easternmost section is at the approximate stratigraphic position of the other sections (although this makes no allowance for any faults hidden by the bay). There is another stromatolite horizon towards the top of the easternmost section, but this is not laterally extensive. Even allowing for some slack in the correlation, there is no evidence at the eastern side of the bay for the thick sandstone seen to the west. This eastern outcrop is the same as that surveyed for Figure 4.5.

The thick extensive sandstone does not have a strongly channellised base, but the top has been eroded prior to deposition of the overlying conglomerate. The facies change from predominantly flat bedded sandstone to predominantly trough cross-stratified sandstone is

Figure 4.13 Aerial view of the easternmost Dan's Point section at which the thick sandstone body is present (see also Figure 4.12). The stromatolite horizon is visible as a white marker extending across the outcrop.

continuous across the sandstone, although the trough cross-stratified sandstone has been eroded from the eastern section. Within the sandstone, there is a slight coarsening trend to the east. In comparison, facies within the underlying conglomerate change considerably over the same distance.

The thickest section is shown in slightly more detail in Figure 4.14, with the addition of palaeocurrent measurements. In this diagram, gravel waves are depicted at the 24 m level. These have crest lines with an orientation of 277°, a wavelength of 25 cm, and a height of about 2.5 cm (see also Figure 3.10).

The siltstone underlying the stromatolites is parallel laminated and partly loaded, but the scale of the diagram only permits depiction of massive siltstone. Some of the sandier parts of the silt show poor parting lineation directed north-northeast. At the 19 m level, patchy calcite cementation has an appearance similar to caliche. The sandstones in this section weather lumpily as a result of rosette calcite cementation.

At Five Mile Point, there is a lava flow within the Copper Harbor Formation. Immediately above this, the sediments are of the conglomerate-sandstone facies association. Figure 4.15 shows the section here, measured

Figure 4.14 Large scale section of the thickest section near Dan's Point (see also Figure 4.12).

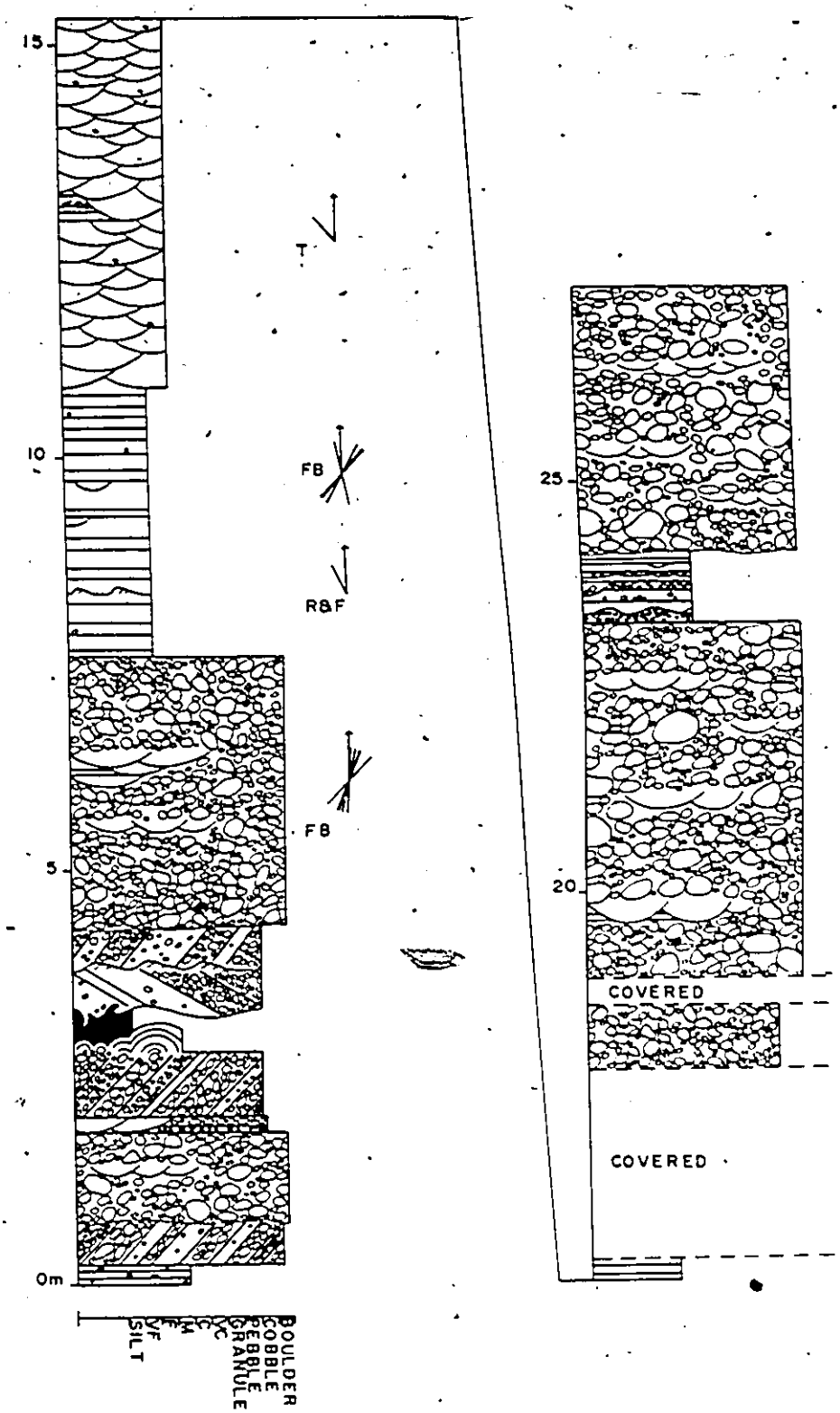

15
10
5
0m
T
FB
R&F
FB
25
20
COVERED
COVERED
SILT
VF
F
M
C
VC
GRANULE
PEBBLE
COBBLE
BOULDER

Figure 4.15 The western stratigraphic section in the Copper Harbor Formation above the lava flow at Five Mile Point, Upper Michigan.

15
10
5
0 M
T
T, R
FB
T
T
T
FB
30
25
20
FB
T
T
T
FB
FB
T
T
FB
FB
T
VCC
BOULDER
COBBLE
PEBBLE
GRANULE
VC
C
M
F
VF
SILT

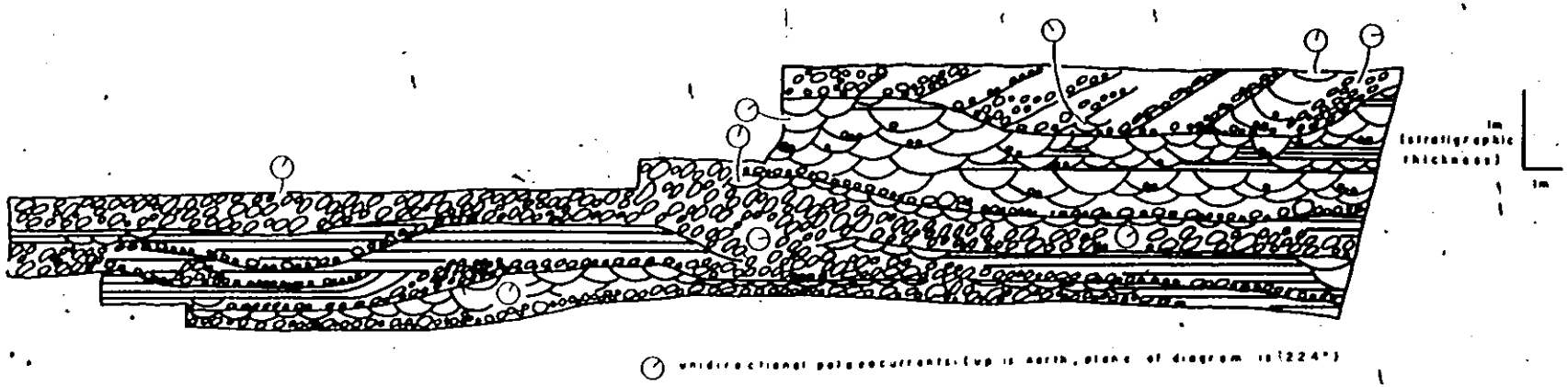
starting above the beach in the bay 0.5 km west of the lighthouse.

The large scale cross-stratified sandstones (facies S-13) near the 25 m and 35 m levels are interpreted as aeolian (see sections 3.3.6, 5.3.2 and 6.2). The palaeocurrents from these sandstones are markedly different from those of sandstones immediately above and below. Just below the lava this section is faulted. The throw is not known, cannot be deduced from the outcrop, and is not marked on the geological quadrangle map, but it is significant, because west of the fault, large scale cross-stratified sandstones are found in the same stratigraphic position as conglomerate east of the fault.

Two beds of massive, matrix-supported, coarse conglomerate are found near the 5 m level. For clarity the difference between the fabrics of these beds and other conglomerates has been exaggerated, but the difference in the amount of sandy matrix is not very large, and in every other respect these conglomerates are the same as other conglomerate beds (facies C-1).

Just above the 15 m level, the outcrop surface is approximately horizontal and can be surveyed (Figure 4.16). Much of the conglomerate here is imbricated, showing flow directions similar to those obtained from trough axes in the

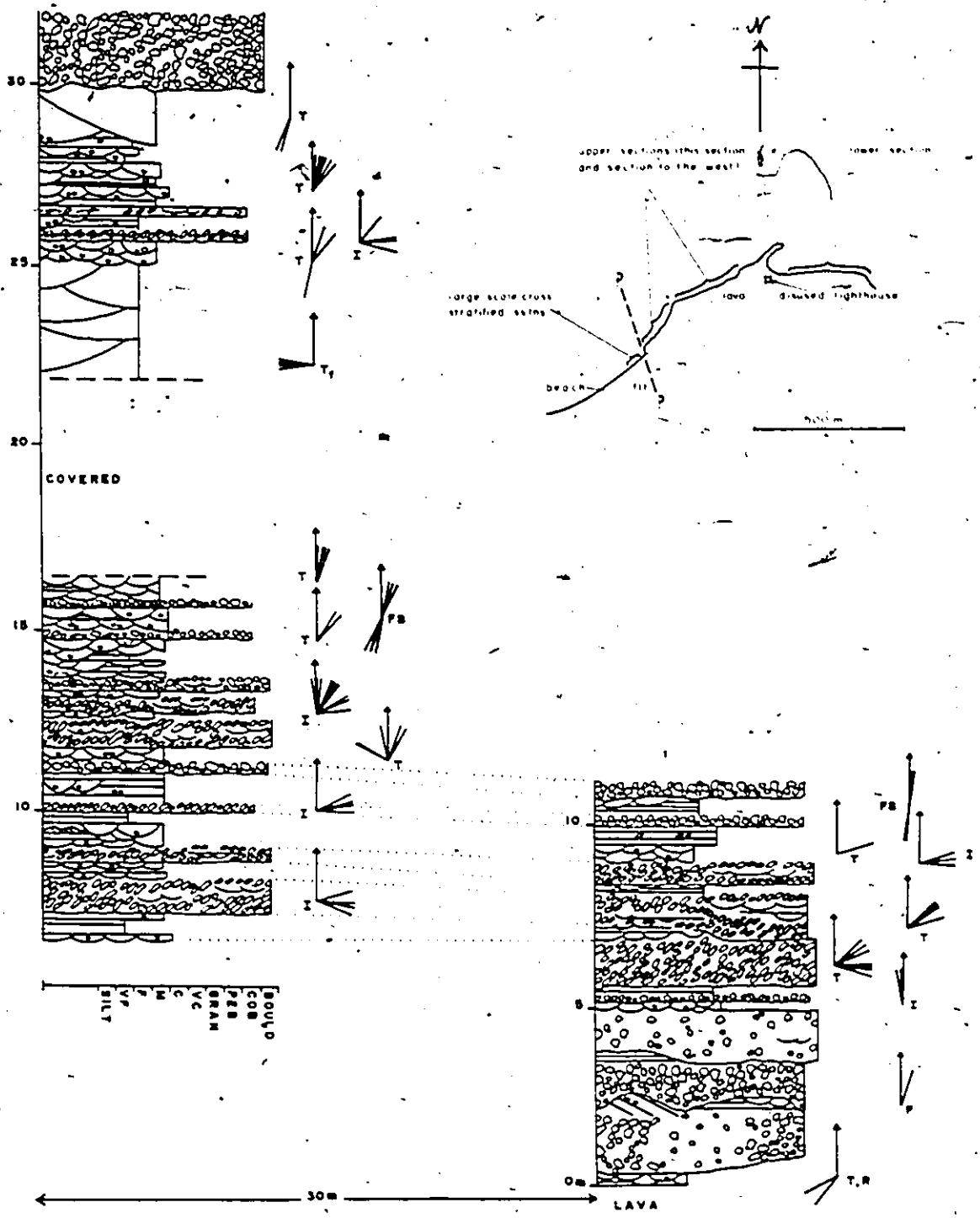
Figure 4.16 Surveyed section from outcrop in the Copper Harbor Formation just above the 15 m level of the section in Figure 4.15.


1m
(stratigraphic
thickness)
1m
unidirectional paleocurrents (up is north, plane of diagram is 224°)

sandstones. Facies relationships are complex, with abundant channelling. In the lower western portion of the diagram, flat-bedded sandstone drapes a channel wall. At the top of the diagram, an unusual channel fill is preserved. Planar cross-strata outline lateral accretion surfaces. Trough-cross-stratified sandstones indicate that the channel was partly filled by sand supplied from within the channel. Apparent imbrication in the conglomerate foresets might either reflect accretion of pebbles supplied from within the channel or pebbles that rolled down from the bar top. Either process could give the clast orientation recorded. However, the concentration of pebbles and cobbles at the set base would favour the former interpretation.

Due to a bend in the coastline, section at the same stratigraphic level is exposed to the east. This section is shown in Figure 4.17, with a sketch map showing the relative locations of the sections here. This section is drawn at a slightly different scale to the other sections, because a small cove interrupts the sequence. Beds can be seen underwater and thus traced across the cove. The total thickness of section here is slightly less than that to the west. This is probably due to a small fault in the covered part of the section. Several small faults cut the exposed

Figure 4.17 The eastern stratigraphic section in the Copper Harbor Formation above the lava flow at Five Mile Point, Upper Michigan.


30
25
20
COVERED
15
10
5
0m
LAVA
30m
SILT
VF
F
M
C
VC
GRAN
PEB
COB
BOULD
T
Σ
FS
T,R
N
upper section this section
and section to the west
lower section
large scale cross
stratified seams
lava
drowned lighthouse
beach
fill

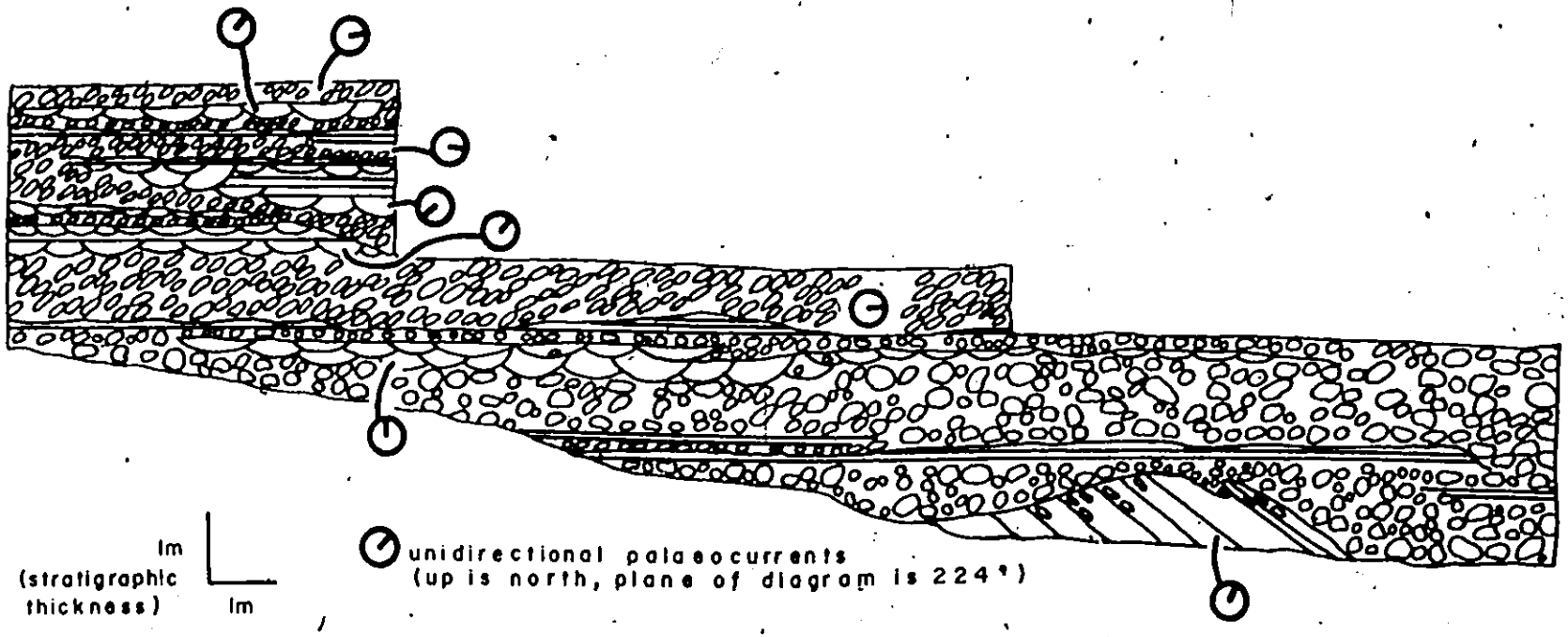
section, but corrections have been made for the throw on these.

More imbrication is seen in this section than in that to the west, probably, in part, a function of outcrop orientation and quality. At the base of the section, the sandstone contains basalt pebbles with textures similar to the underlying flow. Palaeocurrents here are reversed relative to those in most of the section.

The comments made regarding the matrix-supported conglomerates in the western section equally apply to the conglomerates near the base of this section. It might be noted that these conglomerates all occur at approximately the same stratigraphic level. Facies correlations between the eastern and western sections cannot be made with any reliability except where the sections come together at the headland, but the occurrence of matrix-supported conglomerates at this level in both the sections would seem to point to a related origin.

Below the 9 m level, the outcrop can be surveyed (Figure 4.18). The aspect of the outcrop in this surveyed section is generally similar to that surveyed on the other side of the point. The extensive conglomerate in the lower centre of the diagram is one of the beds mentioned above. This bed has a sparse pebble fabric, in places matrix-

Figure 4.18 Surveyed section from outcrop in the Copper Harbor Formation below the 9 m level of the section in Figure 4.17.


1m
(stratigraphic thickness)
1m
unidirectional palaeocurrents
(up is north, plane of diagram is 224°)

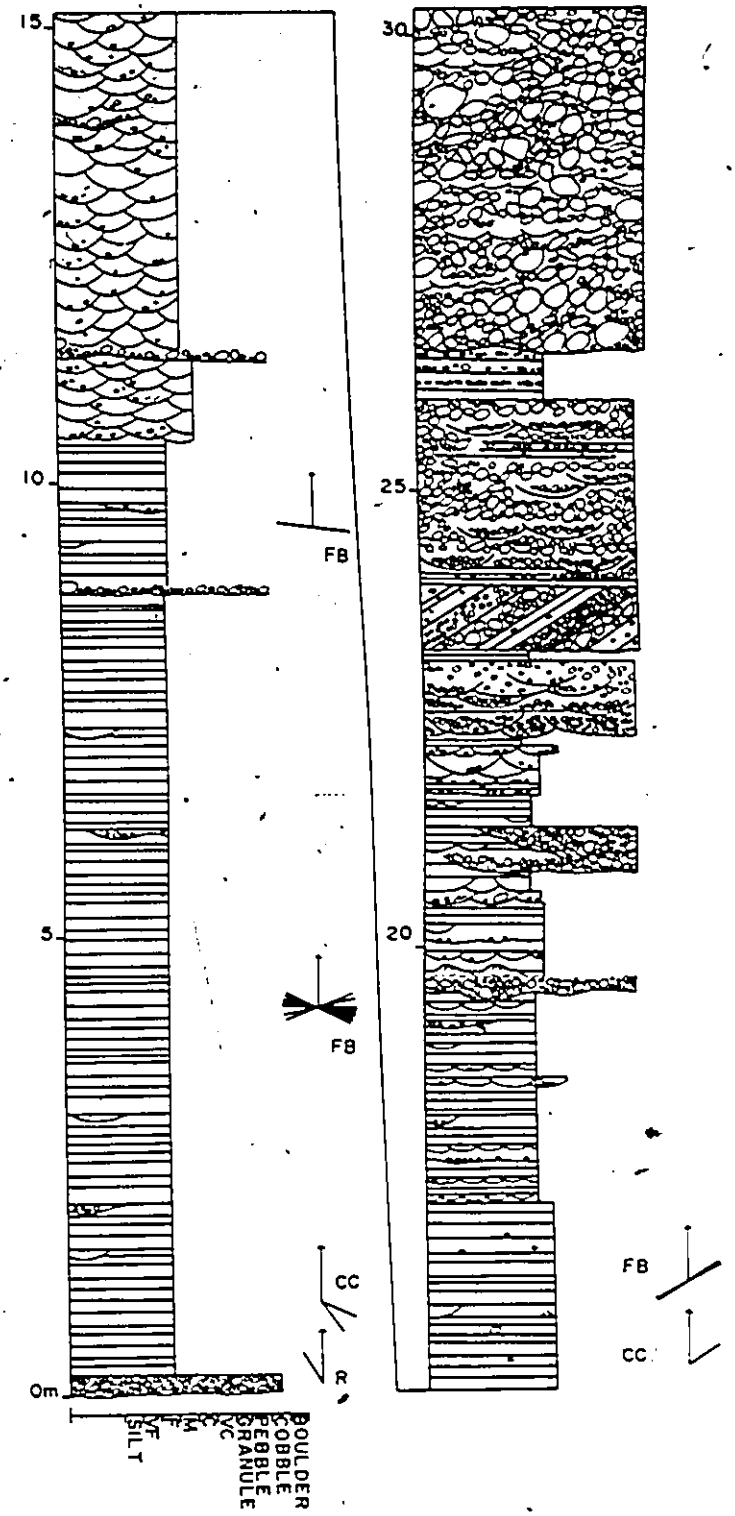
supported. The fabric is drawn on the surveyed section more nearly as it appears in outcrop. As at Mamainse Point, a debris flow origin cannot be ruled out, but cannot be proved.

Near the top of the diagram one trough appears to be oppositely directed to all the others. At the bottom of the diagram, there is some sort of sandy bar, with pebbles at the tops of foresets, where these have rolled over from the bar top and lodged.

At Eagle River on the Keweenaw Peninsula, Copper Harbor Formation sediments are exposed immediately above the top of the Portage Lake Lava Series. The section (Figure 4.19) starts just below the dam at the settlement of Eagle River.

As a whole, the section seems to record a deepening trend, with more trough cross-stratification higher in the section, perhaps representing gradual progradation of an alluvial fan over the lavas flooring the basin. In the lower, finer conglomerates near the top of the section, planar cross-stratification, horizontal stratification, and trough cross-stratification (facies C-8, C-11, C-10) are present. Above this, the conglomerate coarsens upwards into massive, c-m supported, coarse conglomerate. Similar coarse conglomerate is more poorly exposed for a further 90 m above

Figure 4.19 Section at the base of the Copper Harbor Formation in Eagle River on the Keweenaw Peninsula, Upper Michigan.


15
10
5
0m
30
25
20
FB
FB
CC
R
FB
CC
SILT
VF
F
M
C
VC
GRANULE
PEBBLE
COBBLE
BOULDER

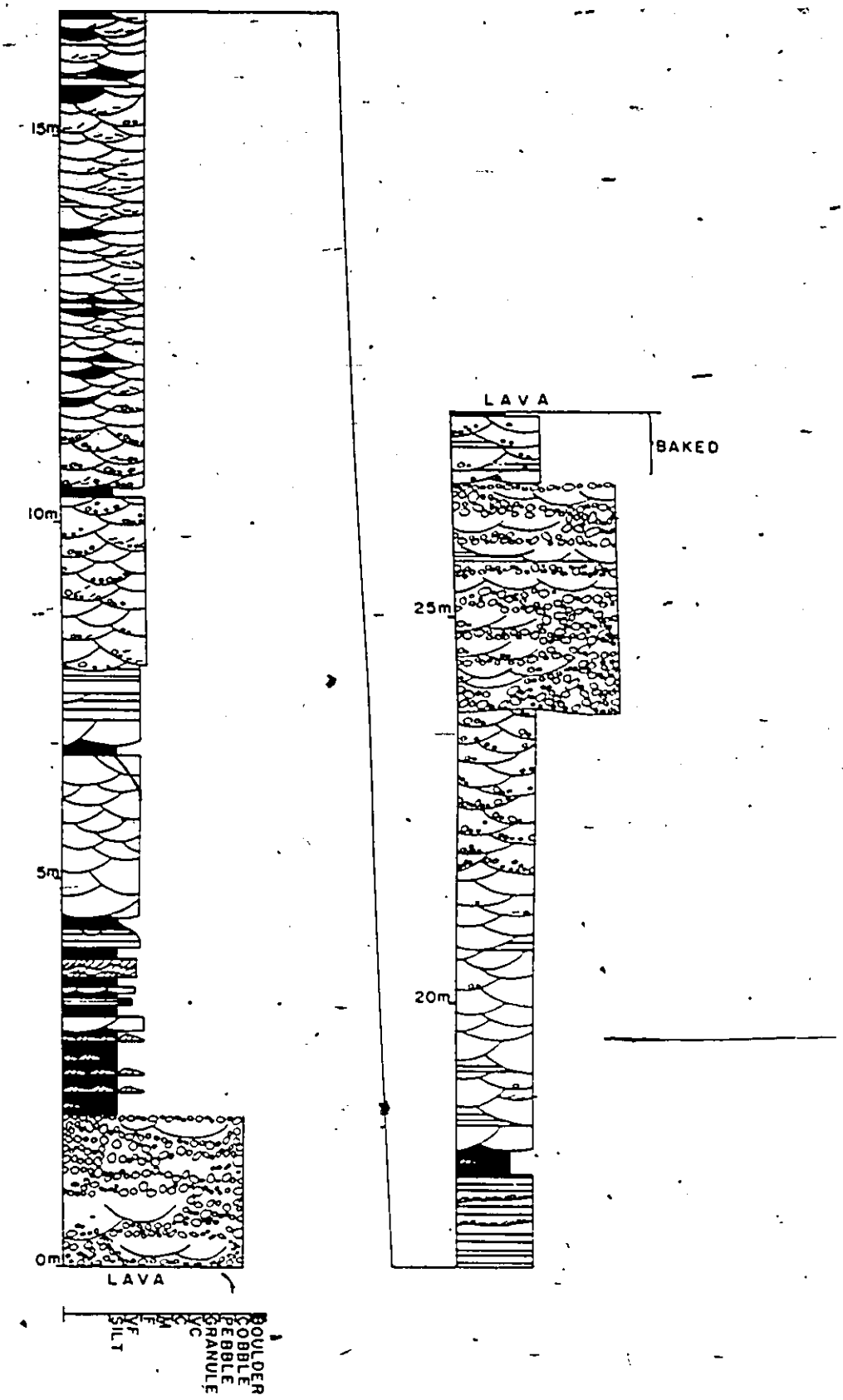
the top of the section (grain size trends for these conglomerates are discussed in Chapter 6).

Interflow sediments are exposed at Saxon Falls in the Montreal River, which in this area is the Michigan-Wisconsin border. A hydroelectric power scheme here has caused flow diversion resulting in exposure of bedrock in the channel floor (although it is well for the visitor to this outcrop to keep a wary eye out, since the sluice controllers make no allowance for wandering geologists). The bedding here is vertical, probably the result, at least in part, of movements on the nearby Keweenaw Fault. This dip and the planed-off character of the outcrop allow of no palaeocurrent measurements.

The section here (Figure 4.20) is interesting in that it shows abrupt changes from conglomerates, which presumably required fairly steep slopes for transport, to siltstones, which reflect ponding. In this respect, a comparison may be drawn with the Dan's Point and Horseshoe Harbor sections. The top metre or so of the section is baked by the overlying lava flow. The siltstones here have a purplish (rather than reddish) hue, and a sharp fracture.

Sediments below the Middle Keweenawan lava pile are exposed to the west of the northern end of Moffat Strait in the Nipigon Bay area (location map of Figure 2.4). The total

Figure 4.20 Section between lava flows towards the top of the Portage Lake Lava Series at Saxon Falls in the Montreal River, Northern Michigan/Wisconsin.


15m
10m
5m
0m
LAVA
BOULDER
COBBLE
PEBBLE
GRANULE
VC
C
M
F
VF
SILT
LAVA
BAKED
25m
20m

thickness of sediments here is about 120 m (Giguere, 1975). The sediments rest on quartzites of the Sibley Group with an angular unconformity (Figure 4.21).

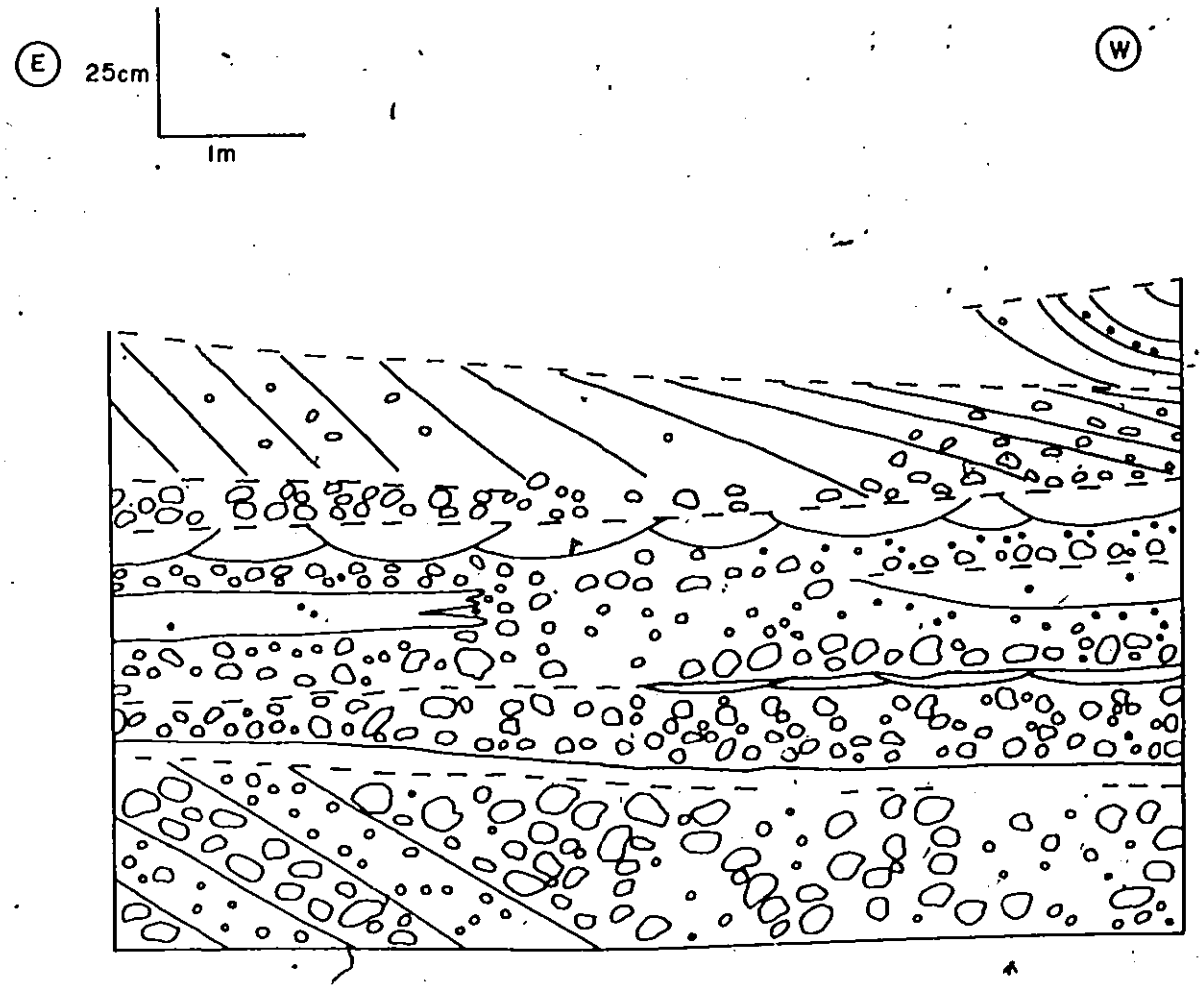
The short section in Figure 4.22 was measured at the base of the cliff in Figure 4.21. It is not possible to work above this level. The section base is at the top of the Sibley Group. Some of the cobbles and pebbles in the basal sandstone are Sibley quartzite. This is analogous to the conglomerate beds found immediately overlying lavas elsewhere in the Lake Superior basin, which normally contain more basic volcanic clasts than other conglomerates (a phenomenon also described in the subsurface of the Keweenaw Peninsula by Butler and Burbank, 1929). Fallen blocks from the cliff show wave ripples, desiccation cracks, and sandy surfaces covered with mud chips. This section is presented here, because the cliff section apparently coarsens upwards into conglomerate, which is exposed at ground level some distance along strike. Figure 4.23 is a detail of outcrop of this conglomerate (note the considerable vertical exaggeration).

Figure 4.21 Cliff at Moffat Strait, Nipigon Bay, Ontario, formed in sediments underlying volcanics of the Osler Group. At the base of the cliff, an erosion surface is visible, underlain by lighter coloured tilted sediments of the Sibley Group. The total height of the cliff is about 15 m.

Figure 4.22 Section measured above the unconformity shown in Figure 4.21.

2
1
0m
COB
PEB
GRAN
VC
C
M
F
VF
SILT

Figure 4.23 Small surveyed section from outcrop in sediments underlying lavas of the Osler Group to the west of the north end of Moffat Strait, Nipigon Bay, Ontario. Note the considerable vertical exaggeration resulting from the low angle between this surface and bedding.


E
25cm
1m
W

4.4 Sections within the sandy facies association

At Union Bay campground in the Porcupine Mountains State Park (location map of Figure 2.3), two sections are exposed in the sandy facies association. The lower and upper sections are presented in Figures 4.24 and 4.25 respectively. The base of the upper section is several hundred metres west of the top of the lower section (across an area covered by a beach), and is at approximately the same stratigraphic level.

Nearly all of the sandstones are fine grained. The bulk of the section consists of trough cross-stratified sandstones (S-4) and horizontally stratified sandstones (S-3). There is a poorly defined alternation between mainly trough cross-stratified parts of the sections and horizontally stratified parts, but this is not very regular. The horizontally stratified part of the upper section at the 37 m level contains planar cross-stratified sandstone (S-9). More of this facies is exposed along strike from the top of the lower section, where the outcrop broadens.

Around the 55 m mark, in the lower section, large scale cross-stratification is present. This is free of pebbles, and, although finer, in some respects is similar to the aeolian sandstones of the large scale cross-stratified

Figure 4.24 The lower stratigraphic section in the Copper Harbor Formation at Union Bay, Upper Michigan.

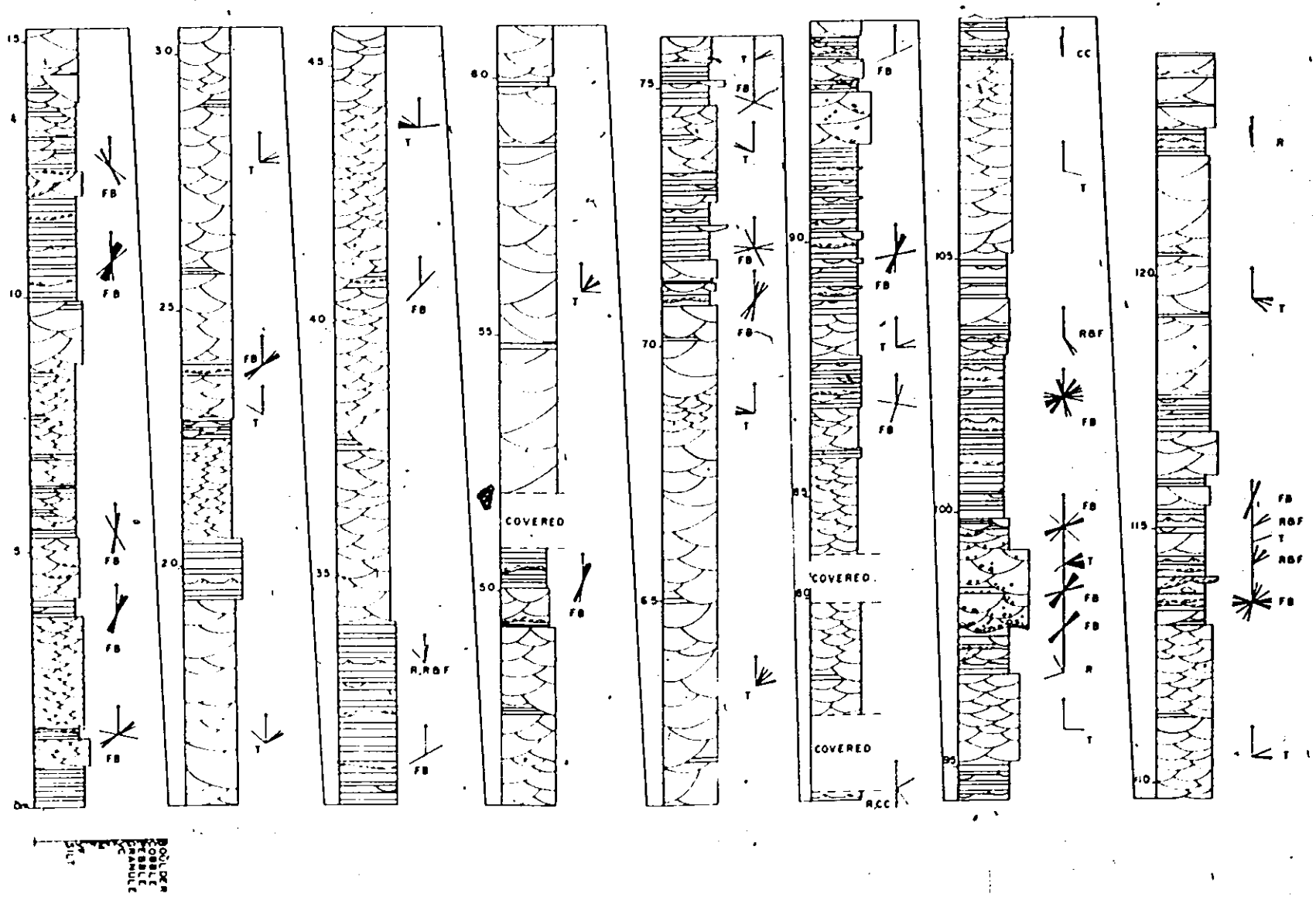
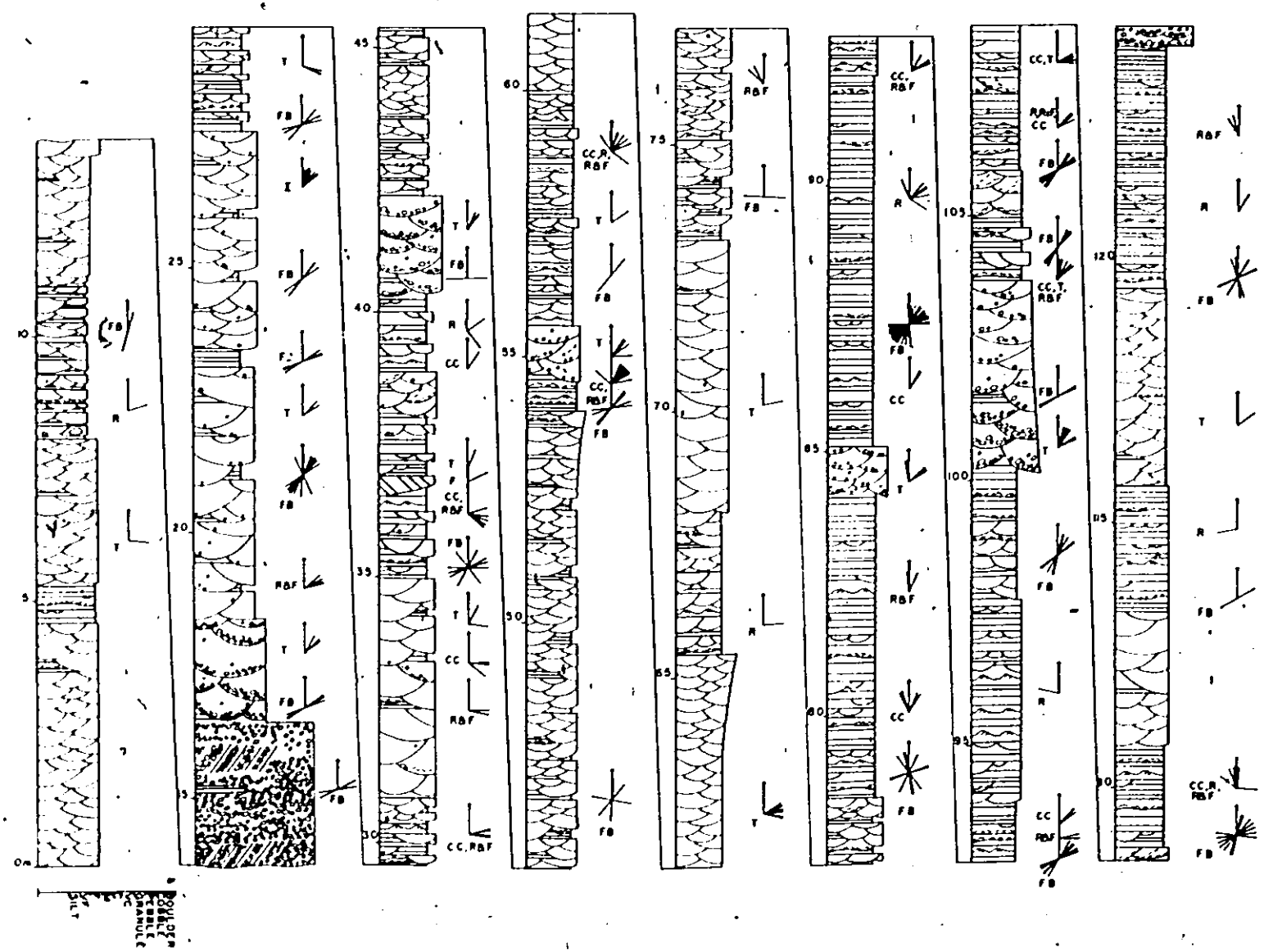

0
5
10
15
20
25
30
35
40
45
50
55
60
65
70
75
80
85
90
95
100
105
110
115
120
FB
T
R,R&F
COVERED
R,CC
CC
R&F
R
BOULDER
COBBLE
PEBBLE
GRANULE
SILT

Figure 4.25 The upper stratigraphic section in the Copper Harbor Formation at Union Bay, Upper Michigan. The base of this section is at approximately the same stratigraphic level as the top of the section in Figure 4.24.


0m
5
10
15
20
25
30
35
40
45
50
55
60
65
70
75
80
85
90
95
100
105
110
115
120
SILT
SAND
GRANULE
PEBBLE
COBBLE
BOULDER
FB
R
T
RBF
CC
CC,R,RBF
CC,RBF
CC,T,RBF
CC,T

facies (facies S-13) seen at Five Mile Point. Unfortunately, the nature of the outcrop here, and a lack of grainsize variation to outline structure, do not permit definite conclusions as to the origin of these sandstones. However, a surface at a low angle to bedding at the 52 m level shows structures which may be examples of critically climbing ripple translatent strata (terminology of Hunter, 1977), resulting from migration of aeolian ripples (Figure 4.26).

Similar section is exposed at Isle Royale at Point Houghton (Figure 4.27). This section may be reached on foot, with difficulty, or by boat, providing that great care is taken to avoid the reefs off the end of the Point. Many mud chips are present on foresets and at bases of trough cross-beds, indicating frequent cessation of flow.

Below the lava flow at Five Mile Point, sandy section is exposed (Figure 4.28). As well as horizontally stratified sandstones (facies S-3), trough cross-stratified pebbly sandstones (facies S-5) and trough cross-stratified sandstones (facies S-4), there is a large amount of large scale cross-stratified sandstone (facies S-13). Set thicknesses in the section are drawn approximately to scale. It may be noted that the large scale cross-strata contain no pebbles, and show flow directions different to the other sandstones in this section.

Figure 4.26 Possible ripple translatent strata of aeolian origin, as discussed in the text.

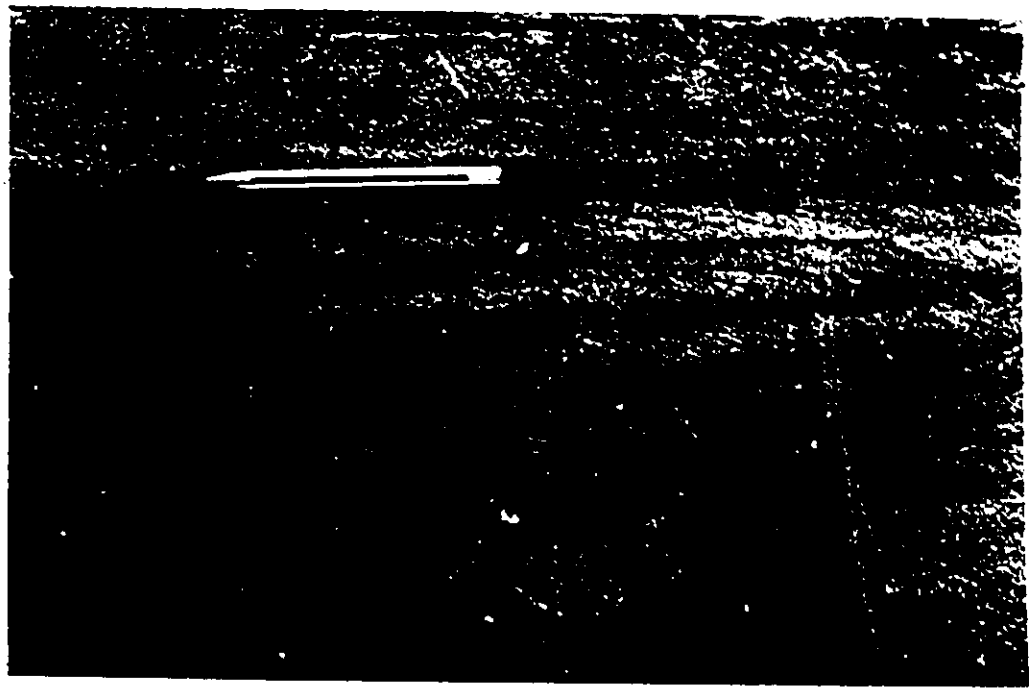


Figure 4.27 Stratigraphic section in the Copper Harbor Formation at Point Houghton, Isle Royale, Upper Michigan.

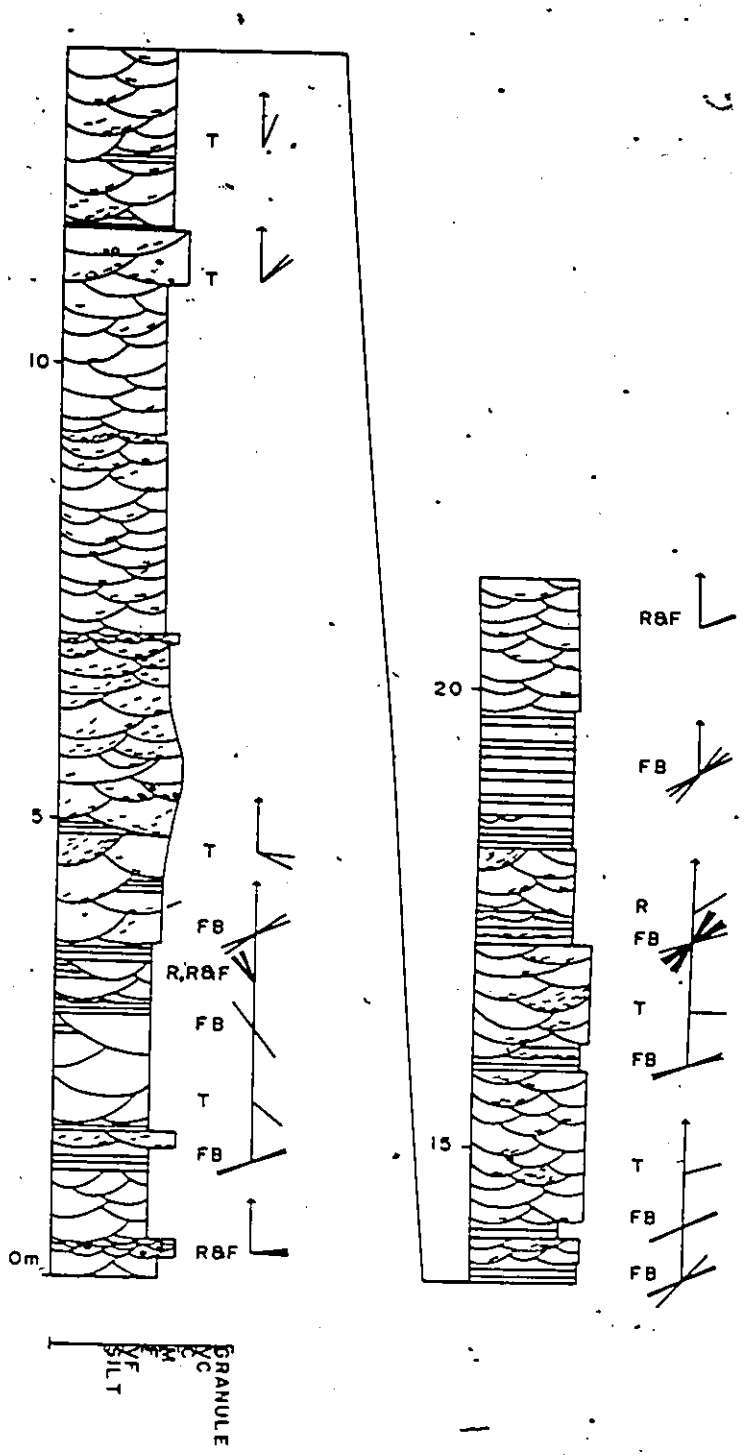
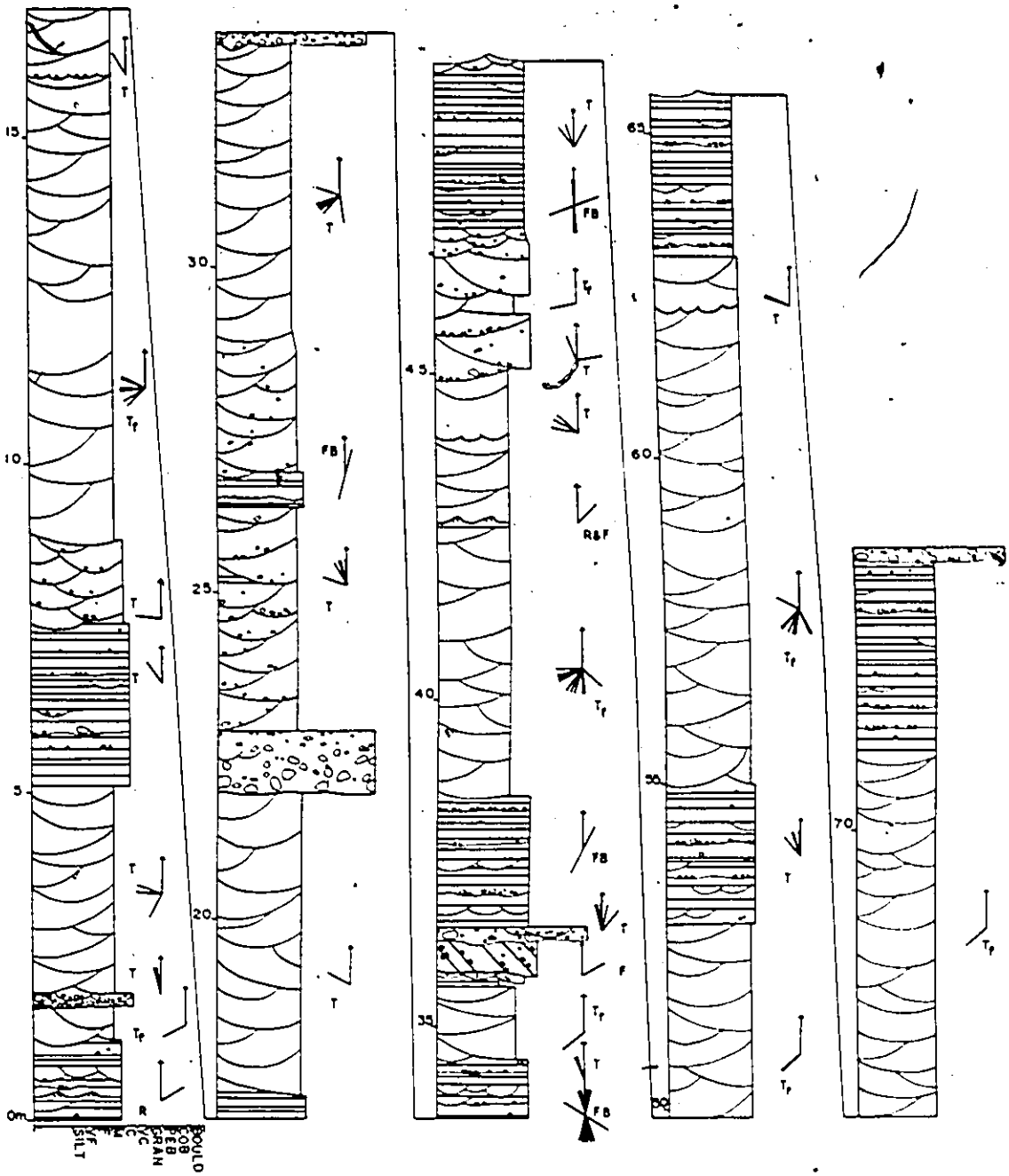

T
T
10
5
T
FB
R,R&F
FB
T
FB
R&F
0m
SILT
F
M
C
VC
GRANULE
R&F
20
FB
R
FB
T
FB
15
T
FB
FB

Figure 4.28 Stratigraphic section in the Copper Harbor Formation below the lava flow at Five Mile Point, Upper Michigan. See also Figures 4.17 and 4.15.


0m
5
10
15
20
25
30
35
40
45
50
55
60
65
70
BOULD
COB
PEB
GRAN
VC
C
M
F
VF
SILT
R
T
Tp
FB
R&F
F

4.5 Sections within the rippled sandstone facies association

Within the Copper Harbor Formation, the only section consisting mostly of the rippled sandstone facies association is that at Francis Point on Isle Royale. This is shown in Figure 4.29.

Much thicker section is found at the base of the Freda Formation along the coast from the mouth of the Presque Isle River (about 10 km east of the mouth of the Black River). This section (Figure 4.30) coarsens upwards, containing planar and trough cross-strata near the top. Most of the section is made up of sandstones of the rippled, and mixed rippled and upper flat bedded facies (facies S-11, S-12). Massive, rippled, parallel laminated, and mixed rippled and parallel laminated siltstones (facies St-1, St-3, St-2, St-4) form most of the remainder of the section. At the 6 m level, a scour surface is overlain by a sequence gradually fining from massive and flat bedded sandstones, through rippled sandstones, to rippled siltstones. This is not one event, since silt drapes are present within the sequence. This sequence, however, is not the rule, and in general there is only a poorly defined alternation between siltier and sandier parts of the section.

Figure 4.29 Stratigraphic section in the Copper Harbor Formation at Francis Point on Isle Royale, Northern Michigan.

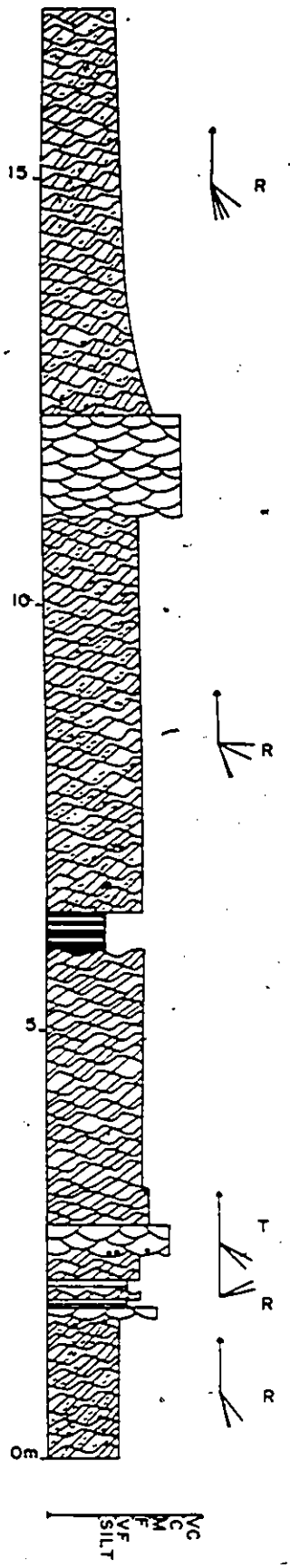
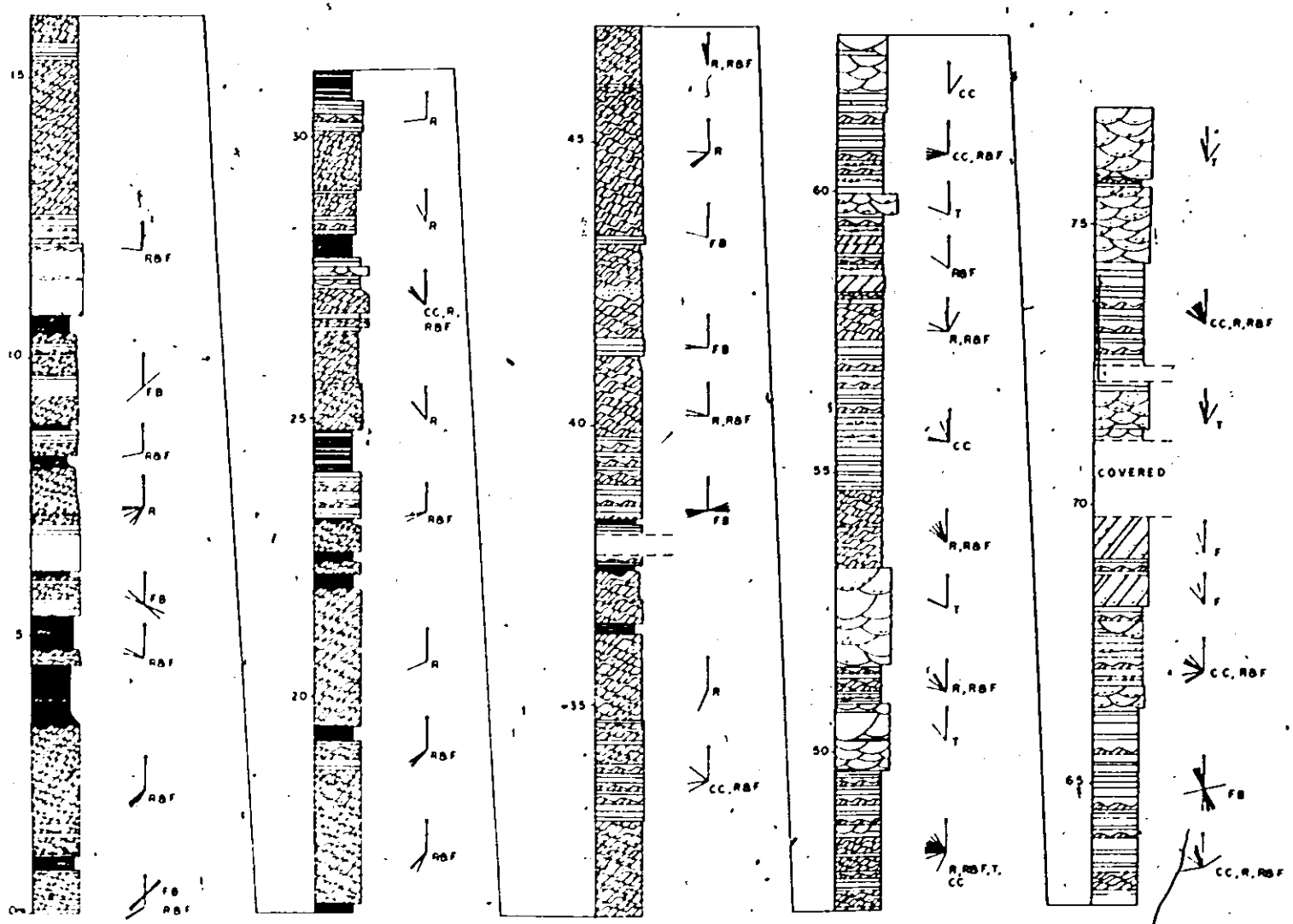

15
10
5
0m
R
R
T
R
R
VC
C
M
F
VF
SILT

Figure 4.30 Stratigraphic section in the Freda Formation. The section begins just west of the mouth of the Presque Isle River, some 10 km east of the mouth of the Black River, Upper Michigan.


15
10
5
0
RBF
FB
RBF
R
FB
RBF
RBF
FB
RBF
30
25
20
R
R
CC, R, RBF
R
RBF
R
RBF
RBF
45
40
35
R, RBF
R
FB
FB
R, RBF
FB
R
CC, RBF
60
55
50
CC
CC, RBF
T
RBF
R, RBF
CC
R, RBF
T
R, RBF
T
R, RBF, T, CC
75
70
65
T
CC, R, RBF
T
COVERED
F
F
CC, RBF
FB
CC, R, RBF

In Swedetown Creek, near Houghton on the Keweenaw Peninsula (location map of Figure 2.3), section is exposed in the Nonesuch Formation. This section (Figure 4.31) consists both of sandstones of the rippled sandstone facies association and of siltstones and sandstones of the silty facies association. The sandy parts of the section are similar to those sections described above. The siltier parts are similar to section at Goodharbor Bay, Minnesota (described below).

4.6 Section within the silty facies association

The only section consisting entirely of the silty facies association is that between lava flows at Goodharbor Bay, Minnesota, exposed in a cutting on the coast road (Figure 4.32). The base of this section is not exposed. The top is baked, and immediately overlain by a basalt flow, deposition of which caused some disruption of the sediment surface.

Many of the sandstones here (facies S-15) are similar in appearance to turbidites, but others seem to have been completely reworked, producing wave or current ripples. At the 9 m level, there is one bed of pebbly siltstone. More of

Figure 4.31 Stratigraphic section in the Nonesuch Formation measured in Swedetown Creek, near Houghton, Upper Michigan.

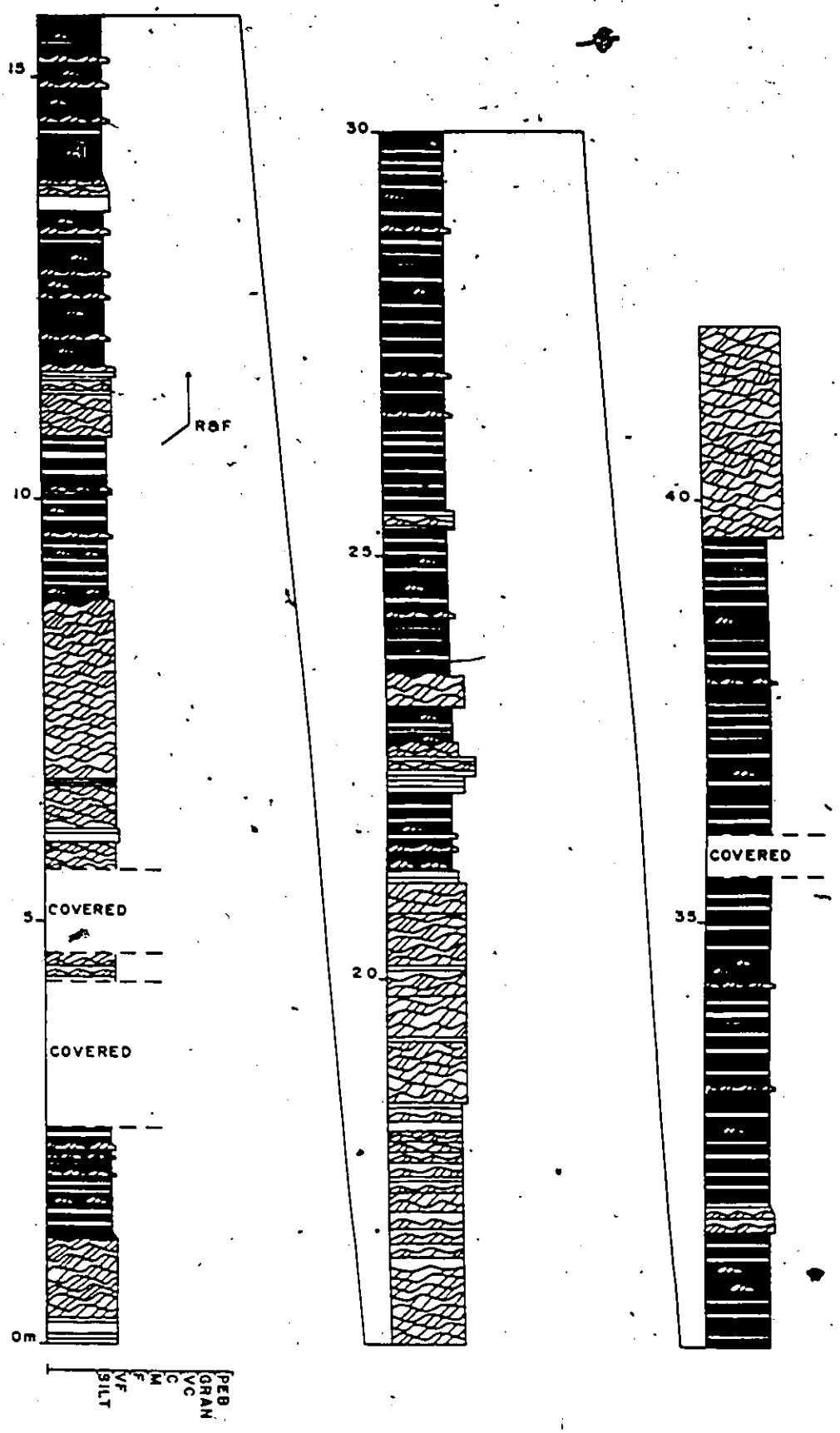

15
10
5
0m
COVERED
COVERED
R&F
30
25
20
40
35
COVERED
R&F
PEB
GRAN
VC
C
M
F
VF
SILT

Figure 4.32 Stratigraphic section in sediments between lavas of the North Shore Volcanic Group, exposed at Goodharbor Bay, Minnesota.

these may be present, but unless the outcrop is excavated, these appear massive.

Two coarsening upward sequences are seen towards the top of the section, terminated by trough cross-stratified (fluvial) sandstones. Towards the top of these sequences, desiccation features become more abundant, reflecting shallowing trends.

CHAPTER 5

PRESENTATION AND DISCUSSION OF PALAEOCURRENT DATA

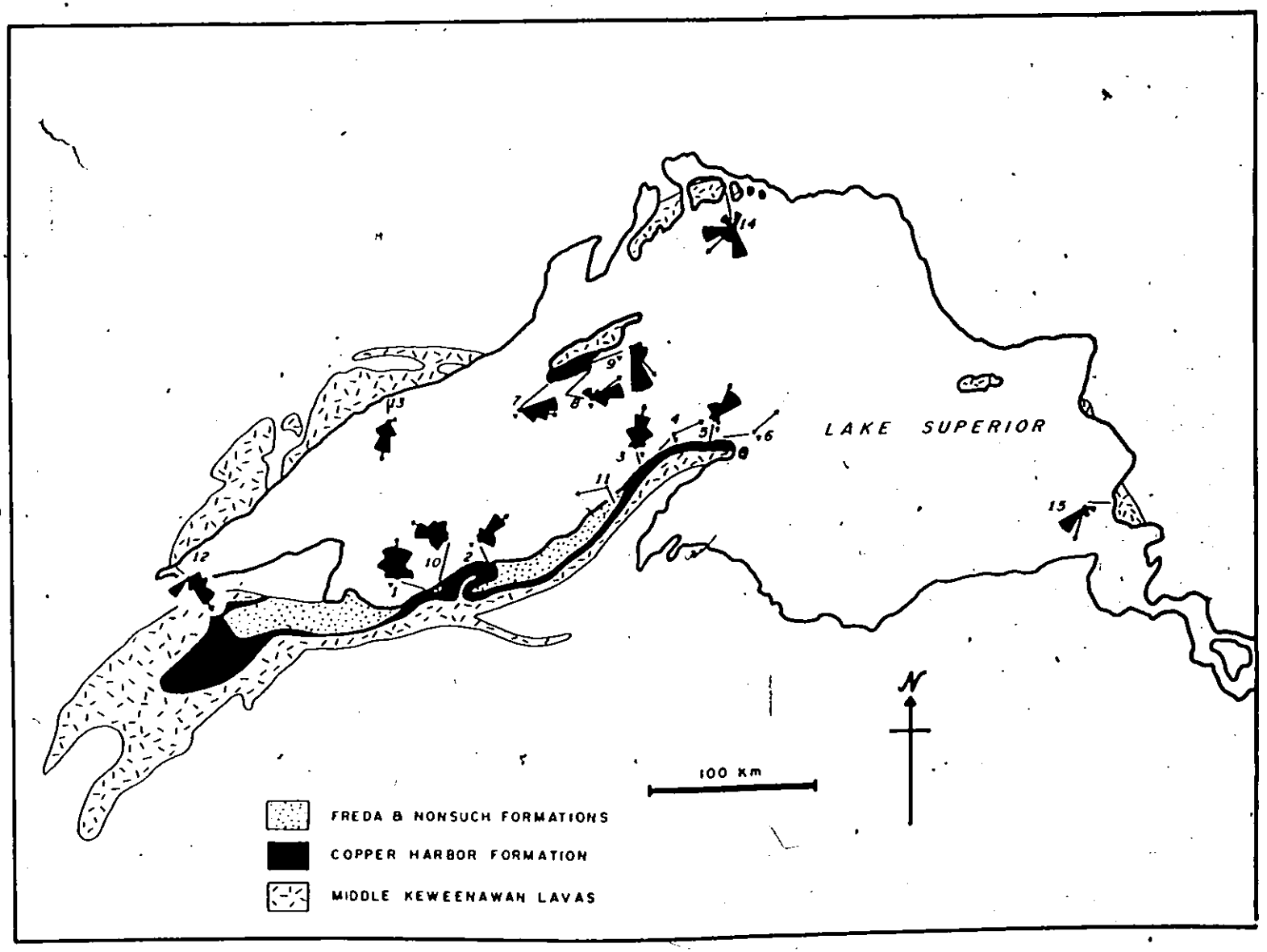
5.1 The data

In this Chapter, palaeocurrent data are presented for all the Middle and Upper Keweenawan sediments involved in this study. Palaeocurrents have been studied at outcrops all around Lake Superior, and provide palaeogeographical information for various parts of the Lake Superior basin at various times throughout the Middle and Upper Keweenawan. Taken as a whole, the palaeocurrent information is spread thinly, both areally over the Lake Superior basin and temporally over the stratigraphic column. This is the result of the paucity of outcrop. Data coverage is best for the Copper Harbor Formation on the Keweenaw Peninsula and on Isle Royale.

5.2 The regional picture

Palaeocurrent roses for all the sections described in Chapter 4 are shown in Figure 5.1 at their appropriate positions around the Lake Superior basin. In addition,

Figure 5.1 Palaeocurrent roses with vector means for Keweenawan outcrops described in this study. Roses for outcrops in the Copper Harbor Formation are marked with a small triangle. All types of unidirectional indicator are included, with the exception of imbrication. At Mamainse Point, imbrication is the only unidirectional indicator available, so here the rose is constructed using the current sense indicated by 5 imbricated beds. Locations are as follows: 1. Black River, data for both Sandstone and Rainbow Falls, 185 measurements; 2. Union Bay, data for both upper and lower sections, 266 measurements; 3. Five Mile Point, data for all sections, excluding aeolian sandstones, 119 measurements; 4. Eagle River, vector mean of 4 measurements; 5. Dan's Point, data for all 5 sections measured in this area, 38 measurements; 6. Horseshoe Harbor, vector mean of 3 measurements; 7. Rainbow Cove, 9 measurements; 8. Point Houghton, 16 measurements; 9. Francis Point, 16 measurements; 10. Mouth of Presque Isle River, Freda Formation, 148 measurements; 11. Swedetown Creek, Nonesuch Formation, vector mean of 2 measurements; 12. Leif Ericsson Memorial Park, Duluth, North Shore Volcanic Group, 16 measurements; 13. Goodharbor Bay, North Shore Volcanic Group, 9 measurements; 14. Moffat Straits, Osler Group, data for all outcrops, 8 measurements; 15. Mamainse Point, Mamainse Point Formation, 5 measurements.


LAKE SUPERIOR
N
100 Km
FREDA & NONSUCH FORMATIONS
COPPER HARBOR FORMATION
MIDDLE KEWEENAWAN LAVAS

palaeocurrent data are shown for an outcrop of interflow sandstones at the Leif Ericsson Memorial Park in Duluth. For sediments below the lavas at Moffat Strait, Ontario, and for the thick interflow sediment package at Mamainse Point, Ontario, where palaeocurrent data are sparse, the roses were constructed with data from the whole outcrop, not just data from the measured sections.

5.2.1 The relationship of fluvial palaeocurrents and lava palaeoflow directions

The unidirectional data for the interflow sediments at Mamainse Point are supported by flat bed parting lineation, four examples of which were observed, with a mean azimuth of 234°. One bent pipe amygdale occurrence was encountered during this study, and gives some indication of lava palaeoflow direction (Annels, 1973, made mention of such features, but unfortunately did not give flow directions). These amygdales indicate a lava flow direction of approximately 219°, and immediately below these, imbrication in conglomerate, although poor, apparently shows that rivers were flowing in a similar direction (Figure 5.2). Both lavas and rivers were presumably flowing down a regional palaeoslope. The lava flow direction does not

Figure 5.2 Bent pipe amygdales at the base of a basalt flow at Mamainse point, indicating flow from right to left. The underlying conglomerate has been baked to a purplish colour, and is poorly imbricated, indicating flow from right to left. Notebook is 20 cm long.

disagree with the Michipicoten Island source proposed for these basalts by Green (1977). It agrees better, however, with a source associated with Keweenawan dyke swarms to the northeast and north of Mamainse Point. These dykes are not mentioned by Green (1977), but were proposed as a lava source by Merk (1972).

The Moffat Strait palaeocurrent data are of variable orientation, but with flow directions to the south or southwest being dominant. Here too, bent pipe amygdales were observed at one location, indicating a lava flow direction of approximately 070°. Trough axes in the sediments immediately below indicate that rivers were flowing due south to south-southwest. This lava flow direction is different to most of those observed by Giguere (1975), who recorded eight examples of pahoehoe and bent pipe amygdales, with a vector mean of approximately 140°. Tanton (1931, pp.57-65), however, reported pahoehoe lava flow directions to the north and northeast.

The palaeocurrents in the interflow sediments on the Minnesota shore are to the south. In addition to the data shown, trough cross-bedding indicating a flow direction of about 176° was observed at a small outcrop about 9km south of Goodharbor Bay (about 2km south of the mouth of the Cascade River). The exposure at the Leif Ericsson Memorial

Park in Duluth consists of similar trough cross-stratified sandstones. No lava flow-directions for this area were recorded during this study, but Sandberg (1938) produced a palaeoflow diagram for the southern part of the North Shore Volcanic Group, constructed from 70 observations of flow indicators. This indicates that lavas flowed from east to west. Green (1977), however, made 241 observations and concluded that there is no dominant palaeoflow direction.

The only other location at which observations were made during this study of the relationship between lava palaeoflow directions and fluvial palaeocurrents was near the base of the Copper Harbor Formation in the Black River, Michigan. Bent pipe amygdales are visible at the base of the thick lava flow within the Copper Harbor Formation at Conglomerate Falls, indicating a flow direction of 244°. Immediately below this, imbrication in the conglomerate appears to indicate flow to the northeast (i.e. oppositely directed to the lava palaeoflow). The fracture of the baked sediments does not enable accurate palaeocurrent measurements, but this flow direction agrees with those from downstream (higher in the Copper Harbor Formation) at Sandstone and Rainbow Falls.

According to Green (1977,1982), the likelihood is that the Mamainse Point Formation, the Osler Group, the

North Shore Volcanic Group and the Portage Lake Lava Series accumulated in separate basins at different times. Apart from the Mamainse Point data, the palaeoflow directions of lavas and rivers differ (although it should be noted that the data are few). The situation is not, however, quite as straightforward as White (1960) envisages, with lava flowing out from the centre of the Lake Superior basin, and, after subsidence, rivers flowing from the basin margins in the opposite direction. This is most nearly true for the Keweenaw Peninsula. For the North Shore Volcanic Group and the Osler Group, the divergence of flows appears to be approximately 90°, and for the Mamainse Point Formation, lavas and sediments seem to have flowed in the same direction down a regional palaeoslope. All the petrological evidence (e.g. Annels, 1974; Merk and Jirsa, 1982; Giguere, 1975) indicates that sediments were partly derived from areas outside the lava fields, so rivers were generally flowing into the area of the Lake Superior basin. However, the palaeocurrent data would indicate that once inside the basin, rivers did not always flow perpendicular to the basin margins. This conclusion is borne out by data from the Copper Harbor Formation (see below).

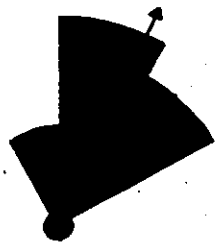
5.2.2 The Copper Harbor Formation

Palaeocurrent roses for the sections measured in the Copper Harbor Formation are shown in Figure 5.1. To supplement this undirectional data where such data are lacking (at Horseshoe Harbor and Eagle River) palaeocurrent roses for bidirectional data are presented in Figure 5.3. At two outcrops, good palaeocurrent data are available, despite a lack of section, and palaeocurrent roses for these locations are also presented in Figure 5.3. Checker Point is an outcrop of the rippled sandstone facies association just north (and downsection) of Francis Point on Isle Royale. The Veale Memorial Park outcrop is geographically and stratigraphically about midway between the Eagle River and Five Mile Point sections, and consists of trough cross-bedded and flat bedded sandstones (facies S-1, S-2, S-4, S-5). Some additional palaeocurrent data are also available just above the top of the Portage Lake Lava Series at Cumberland Point on Isle Royale, at the north end of Rainbow Cove. Imbrication here gives a mean palaeocurrent direction of 111° and five trough axes indicate a mean flow direction of 119°.

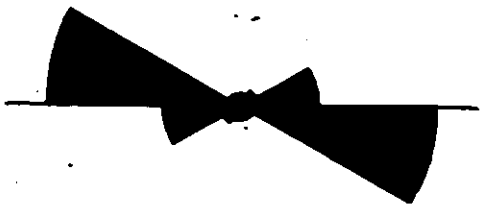
On the Keweenaw Peninsula, several locations yield one or two palaeocurrent indicators but no useful section.

Figure 5.3 Unidirectional and bidirectional palaeocurrent roses for outcrops in the Copper Harbor Formation to supplement the data shown in Figure 5.1. The numbers of measurements for the roses are: Checker Point, 17; Veale Memorial Park, 46; Eagle River, 15; Horseshoe Harbor, 9.

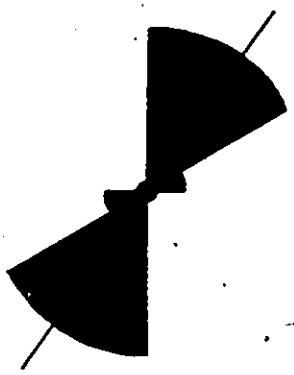
Checker Point



Veale Memorial Park



Eagle River



Horseshoe Harbor

At the base of the Copper Harbor Formation, where lavas and sediments interfinger in Owl Creek, east of Eagle River (southwest of Eagle Harbor), flat bed parting lineation has an azimuth of 048o. To the west of Copper Harbor, at Silver River Falls, stratigraphically just below the Lake Shore Traps (the thick lava flows within the Copper Harbor Formation in this area), asymmetrical ripples yield a palaeocurrent direction of 176o. At the Copper Harbor lighthouse, above the Lake Shore Traps, two imbricated beds show a mean palaeocurrent of 095o.

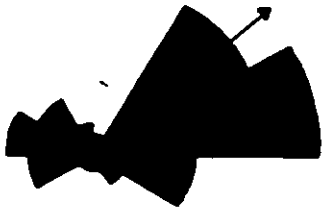
All the Copper Harbor Formation palaeocurrent measurements are in approximate agreement with previous workers' measurements, indicating flow to be generally towards the axis of the Lake Superior syncline, but the data presented here fairly clearly indicate an easterly palaeoslope during Copper Harbor times (to the northeast near the Keweenaw Peninsula and to the east near Isle Royale).

The palaeocurrent data for Union Bay and the Black River are broken down according to the measured sections in Figure 5.4. The vector mean is very similar for both the lower and upper sections at Union Bay, but the lower section shows more palaeocurrent variability. This variability may result in part from an aeolian influence for which no

Figure 5.4 Palaeocurrent roses for individual sections at Union Bay and in the Black River. The numbers of measurements for the roses are: Sandstone Falls, 94; Rainbow Falls, 91; Union Bay, lower section, 76; Union Bay, upper section, 190.

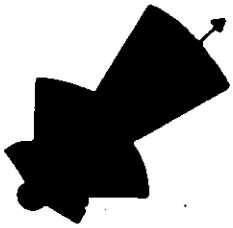
Sandstone Falls

Rainbow Falls



Lower Section

Union Bay



Upper Section

allowance has been made. The Black River data are possibly indicative of a westward swing in palaeocurrents as deposition continued. These data alone are not sufficient to be conclusive, but taken in conjunction with the data from the overlying Nonesuch and Freda Formations (see below), the evidence is in favour of a gradual and progressive palaeoslope change.

5.2.3 The Nonesuch and Freda Formations

Only one outcrop in each of these formations was closely studied during this investigation. The Swedetown Creek palaeocurrent vector for the Nonesuch Formation disagrees with that of Elmore (1981), but the number of observations concerned is very small. Elmore (1981), obtained data from Nonesuch Formation outcrops all over Upper Michigan, and generally recorded flow directions to the northeast, although with much variation in some areas. Daniels (1982) in a review of all the available data, also commented on the degree of variability in the Nonesuch data. The Freda Formation outcrop at Presque Isle measured during this study yielded a large number of palaeocurrent indicators. This flow direction agrees very well with palaeocurrent data for the whole peninsula collected by

Hamblin (1961) and Elmore (1981, appendices). Daniels' (1982) review also concluded that flow directions were to the west and northwest.

It therefore seems that during Upper Keweenawan times, there was a palaeoslope reversal, from east and northeast to west and northwest, with an intervening period of generally low or variable slopes during Nonesuch times.

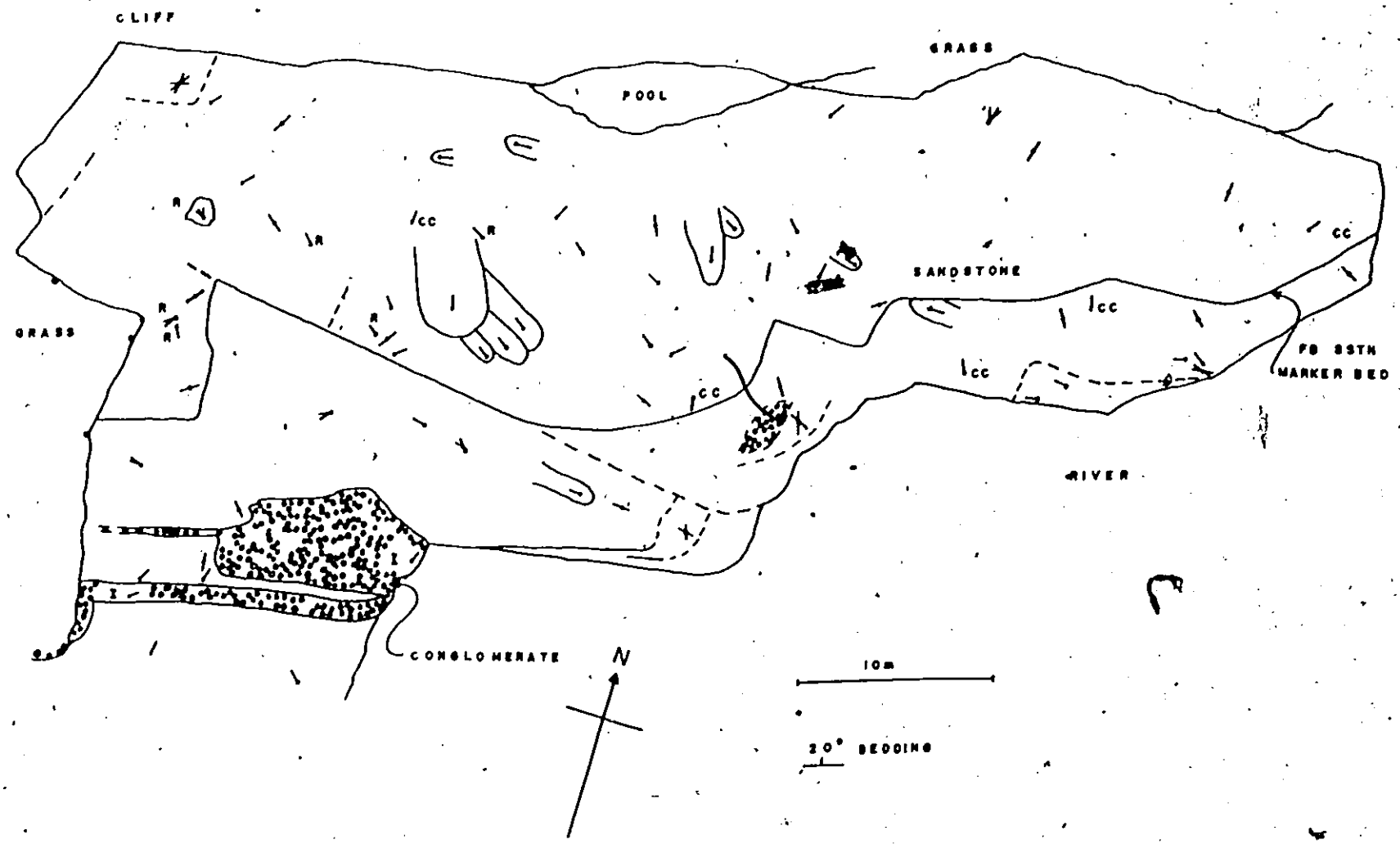
The basalts of the Mamainse Point Formation are unconformably overlain by sandstones and siltstones (seen in Mica and Alona Bays), thought to be correlative with the Freda Formation. A brief visit to these outcrops was made during this study and confirmed that sediments are of similar aspect to those of the Freda Formation at Presque Isle. One occurrence of climbing ripples indicated a palaeocurrent to the north, in agreement with observations of Ojakangas and Morey (1982), who recorded palaeocurrents to the north and northwest. It therefore seems that in this area too, there was a noticeable palaeocurrent swing prior to deposition of the Freda Formation, although it is not possible to say whether this is related to the change in slopes that occurred in the western Lake Superior area at this time.

5.3 Large scale studies of palaeocurrent data

5.3.1 Sandstone Falls

Around the 52 m level in the Sandstone Falls section (Figure 4.9), palaeocurrents are very variable. The outcrop is broad at this point, and was mapped with tape and compass. Figure 5.5 shows the resulting plan view of the outcrop. The outcrop here is not flat, and bedding dips at 22° in the direction indicated in the diagram. However, the flat bedded sandstone marker bed was originally close to horizontal and gives an indication of outcrop topography. Troughs large enough to be mapped are sketched in the appropriate position on the map. A considerable variation in palaeocurrents is evident, with troughs oppositely directed even though stratigraphically very close. Because this palaeocurrent variation is apparent in trough axis measurements as well as in smaller scale structures, it cannot be attributed to high stage - low stage variations. The only possible conclusion is that palaeoslopes were not high the time of deposition. Boothroyd and Nummedal (1978) noted that although Icelandic outwash fans are entirely covered with braided streams, the distal portions of Alaskan outwash fans may develop meandering streams. These authors

Figure 5.5 Large scale palaeocurrent map of part of the outcrop at Sandstone Falls. Unless otherwise marked (in which case the abbreviations are the same as those used for the measured sections), unidirectional palaeocurrents are troughs. Bidirectional palaeocurrents are upper flat bed lineations. The outlines of the largest troughs are sketched in. The bedding attitude indicated applies to the whole outcrop, and the topography of the outcrop, which is not flat, is apparent from the flat bedded sandstone horizon, which marks an original horizontal.


CLIFF
GRASS
POOL
GRASS
SANDSTONE
FB SSTN
MARKER BED
RIVER
CONGLOMERATE
N
10m
20° BEDDING

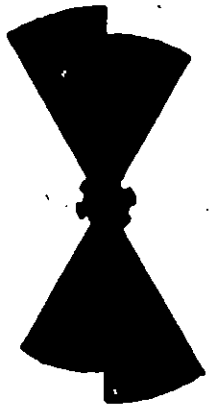
attributed the difference to vegetation. The slopes at the time of deposition of the Sandstone Falls sediments may have been sufficiently low that if plants had been present to stabilise river channels the river would have taken on a meandering morphology.

5.3.2 Five Mile Point

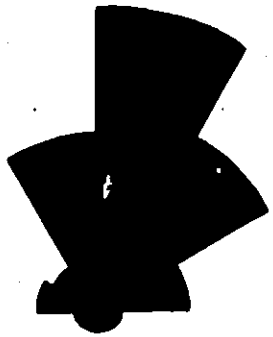
At the Five Mile Point outcrops, palaeocurrent data are a valuable tool for distinguishing aeolian and fluvial sediments (also discussed in sections 3.3.6, 4.3 with reference to Figure 4.15, and 6.1.2). Figure 5.6 shows that fluvial sediments were deposited by currents flowing northwards, whereas the sandstones of the large scale cross-stratified facies were deposited by dunes moving west. This direction of movement is opposed to the general palaeoslope for the Copper Harbor Formation (see section 5.2.3). The probable regional palaeowinds can be estimated from palaeomagnetic studies of the Copper Harbor Formation. Halls and Palmer (1981) and Dubois (1962) (the two studies cited as most reliable in the review of Halls and Pesonen, 1982) indicate that the Lake Superior region was about 20° north of the equator during Copper Harbor times, in the northeasterly tradewinds belt. Halls and Palmer (1981) were

Figure 5.6 Palaeocurrent roses for aeolian and fluvial sandstones at Five Mile Point. The numbers of measurements for the roses are: lower section, unidirectional fluvial, 33; lower section, bidirectional fluvial, 15; lower section, aeolian, 52; all sections, unidirectional fluvial, 119; all sections, aeolian, 95.

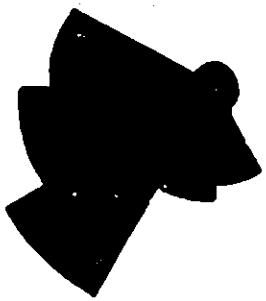
Unidirectional FLUVIAL Bidirectional



FLUVIAL

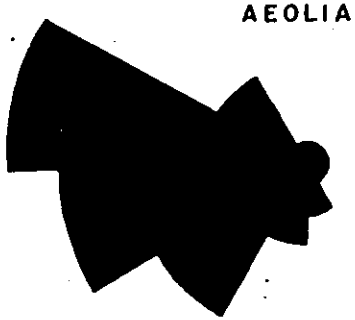


AEOLIAN



LOWER SECTION

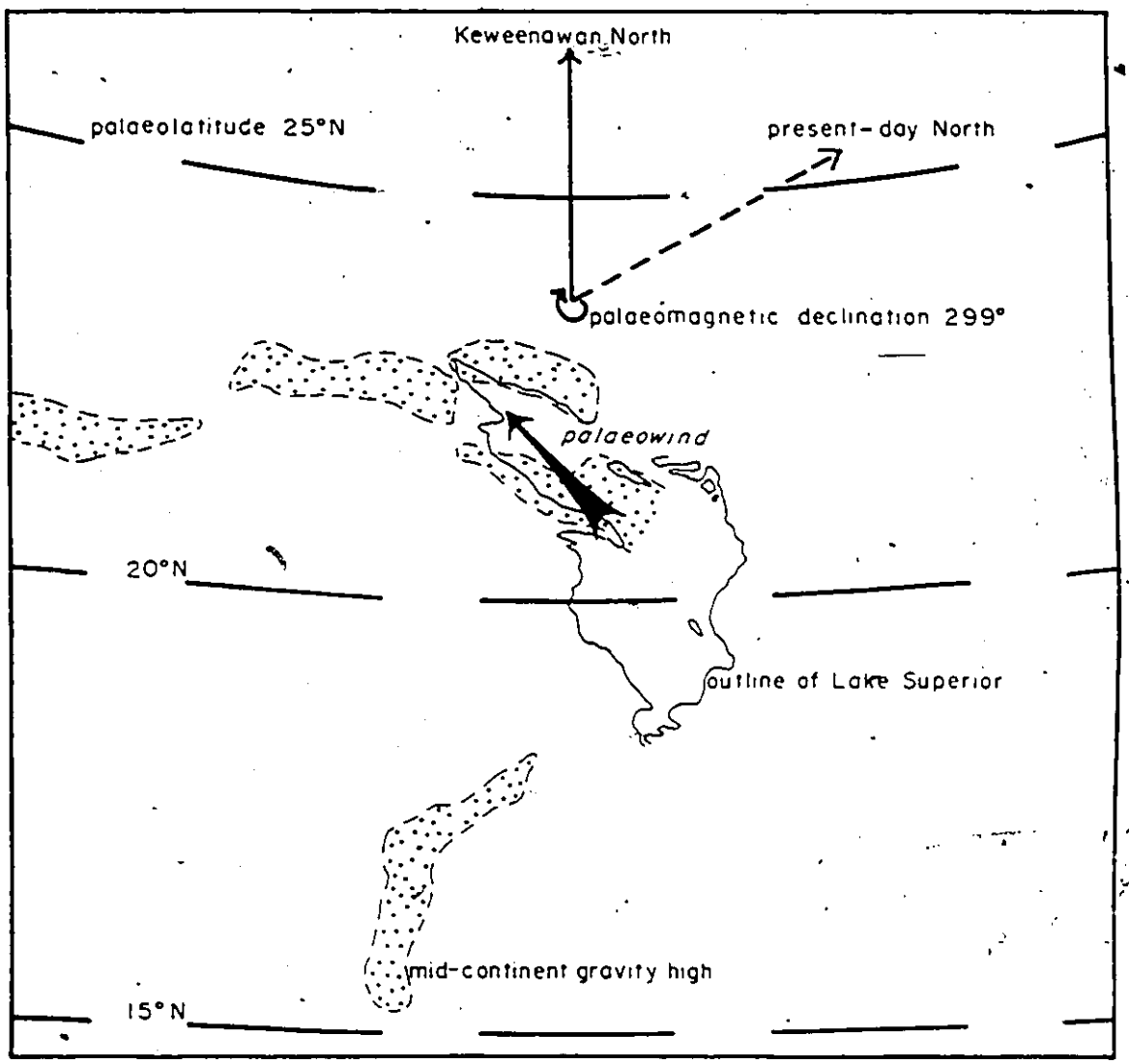
AEOLIAN



ALL SECTIONS

in fact studying the Lake Shore Traps, which pass between the Five Mile Point sedimentary sections, and so provide palaeomagnetic data for the exact time of deposition of these sediments. These authors' palaeomagnetic means are: declination 299o; inclination 37.9o, corresponding to a palaeolatitude of 21o. If this data is used to rotate the North American continent to its Keweenawan orientation then the result is as shown in Figure 5.7. The vector mean of all the Five Mile Point measurements of aeolian cross-strata (258o) is at 319o to Keweenawan north. This indicates a palaeowind blowing to the northwest rather than to the southwest as would be expected for palaeo-tradewinds. However, palaeowinds may have blown in this direction (along the rift valley) due to topographic constraint, so the data is not necessarily in conflict with palaeo-tradewinds.

Figure 5.7 Map showing the palaeowind direction as inferred from the aeolian cross-strata at Five Mile Point, drawn so that the orientations and palaeolatitudes are those of the Lake Superior region in Late Keweenawan time (constructed using the palaeomagnetic data of Halls and Palmer, 1981). The stippled pattern indicates the position of the mid-continent gravity high marking the line of the Keweenawan rift (from Chase and Gilmer, 1973). Gravitational anomalies are weakly defined in the eastern Lake Superior region.


Keweenawan North
palaeolatitude 25°N
present-day North
palaeomagnetic declination 299°
palaeowind
20°N
outline of Lake Superior
mid-continent gravity high
15°N

CHAPTER 6

PETROGRAPHY

The petrographic observations made during this study are of two types: a) measurements in outcrop and thin section of sediment grain sizes, which are discussed in the first part of the Chapter; and b) microscopic investigations of diagenetic features, which are discussed in the latter part of the Chapter.

6.1 Presentation and discussion of grain size data

Analysis of the grain size of clastic sediments, within particular beds or over larger stratigraphic intervals, potentially may provide information as to both sedimentary processes and factors controlling those processes (e.g. tectonic activity). The general lack of thick section in the Middle and Upper Keweenawan sediments of the Lake Superior basin precludes detailed analysis of grain size variations throughout Keweenawan time, although it is evident that there is no strongly defined and regular cyclicity of the type seen in the Hornelen Basin of Norway (Steel and Aasheim, 1978; Gloppen and Steel, 1981). This

Chapter is therefore mainly concerned with the information that grain size data can provide regarding sedimentary processes.

6.1.1 Grain size distributions of conglomerate beds

Grain size distributions of conglomerates were obtained by point counts at good outcrops (as mentioned in Chapter 4, outcrops are not often amenable to photographic analysis of such features). Three exceptionally well exposed conglomerate beds were chosen for grain size analysis, two of the massive, c-m supported, coarse conglomerate facies, one of imbricated, clast-supported, coarse conglomerate. The point counts were performed using a 30 cm or 40 cm grid spacing, the spacing being chosen so that few clasts were intersected twice. Lengths of the intermediate axes of clasts were recorded, these axes being those that would be measured during a sieve analysis.

The resulting grain size distributions are presented in Figures 6.1 and 6.2 as histograms and cumulative frequency curves plotted on log-normal probability paper. The Mamainse Point bed shows the best approximation to a log-normal distribution. A chi-square analysis of the grouped data confirms this fit at a 95% confidence level

Figure 6.1 Grain size histograms for three conglomerate beds. A: massive, c-m supported, coarse conglomerate at Mamainse Point. B: imbricated, clast-supported, coarse conglomerate at Five Mile Point. C: massive, c-m supported, coarse conglomerate at Horseshoe Harbor. The total numbers of grid points for each case were A: 194, B: 236, C: 278, of which, respectively, 76, 57, and 103, fell on "matrix" (considered for point counting purposes as material finer than 8mm). See text for discussion.

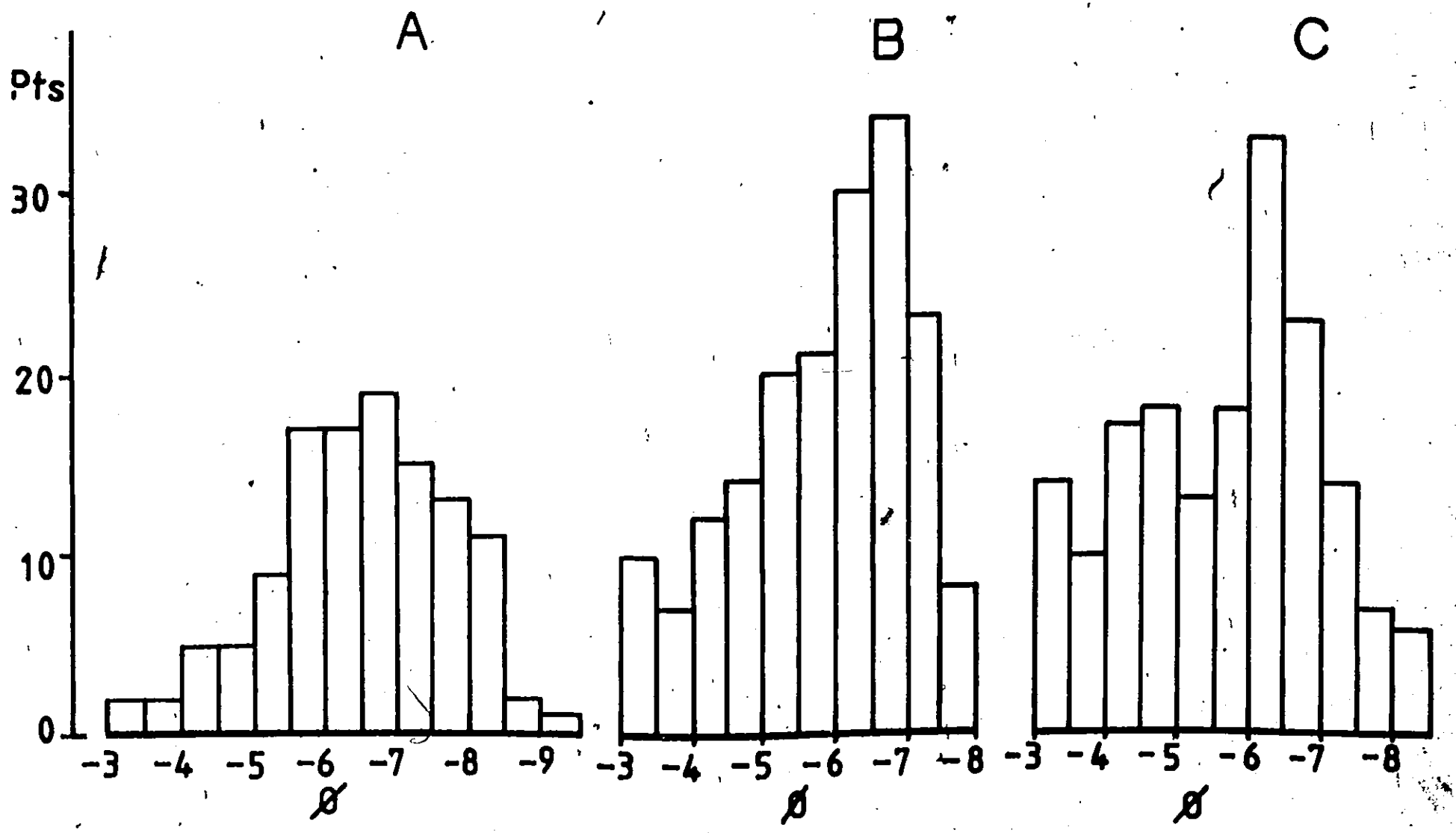
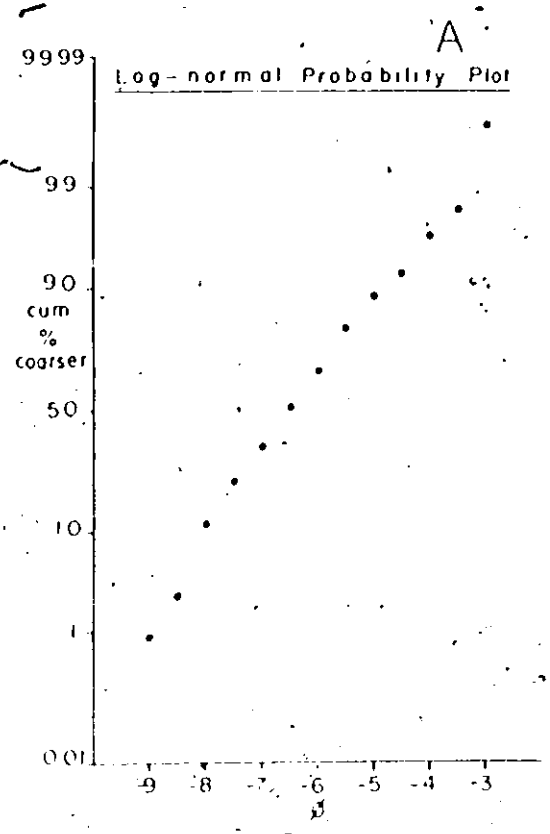
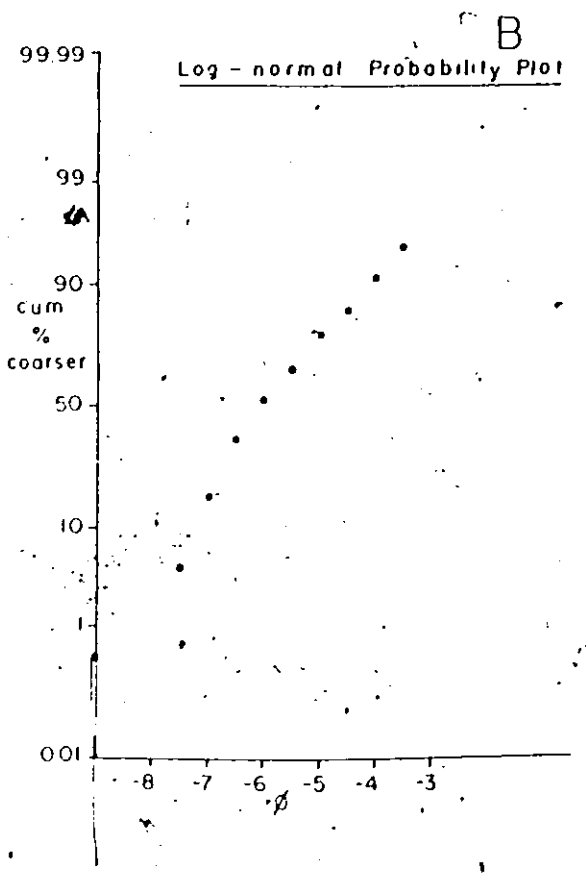
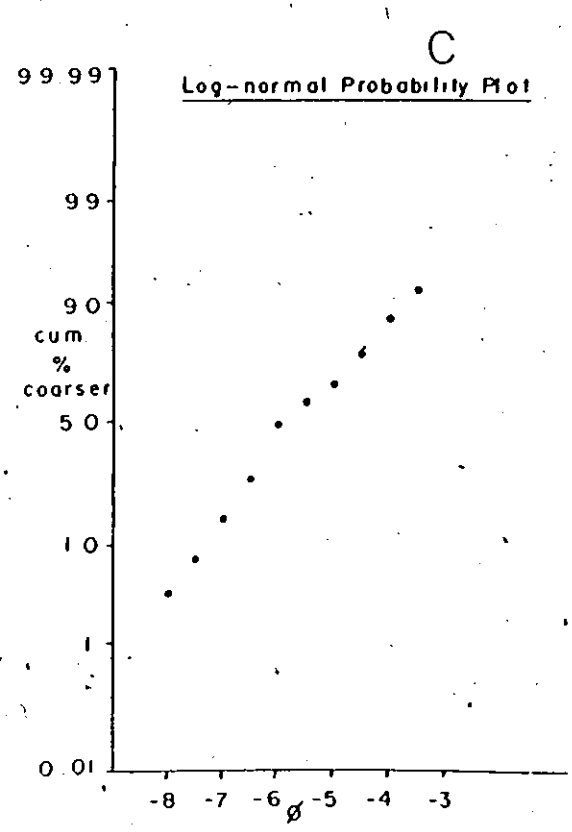

A
B
C
Pts
30
20
10
0
-3 -4 -5 -6 -7 -8 -9
-3 -4 -5 -6 -7 -8
-3 -4 -5 -6 -7 -8
Ø
Ø
Ø

Figure 6.2 Cumulative frequency - grain size curves for three conglomerate beds plotted on log-normal probability paper. A: massive, c-m supported, coarse conglomerate at Mamainse Point. B: imbricated, clast-supported, coarse conglomerate at Five Mile Point. C: massive, c-m supported, coarse conglomerate at Horseshoe Harbor. See also Figure 6.1 and the discussion in the text.


A
Log-normal Probability Plot
99.99
99
90
cum
%
coarser
50
10
1
0.01
-9
-8
-7
-6
-5
-4
-3
ø


B
Log-normal Probability Plot
99.99
99
90
cum
%
coarser
50
10
1
0.01
-8
-7
-6
-5
-4
-3
ø


C
Log-normal Probability Plot
99.99
99
90
cum
%
coarser
50
10
1
0.01
-8
-7
-6
-5
-4
-3
ø

(moment measures were used to generate the best fit log-normal distribution). The Horseshoe Harbor data is more noisy, perhaps because the outcrop here has been planed flat, so that it could not be guaranteed that intermediate axes were measured in every case. This distribution fails to fall within the 95% confidence interval when analysed for log-normality by a chi-square test. If, however, the misfit of the finest grain size interval is ignored, on the grounds that the high count here may reflect overlap with a finer population having a mode somewhere in the sand grain sizes, then the chi-square test no longer rejects the population at the 95% confidence level. The Five Mile Point distribution is clearly skewed, and for this reason fails to fall within the 95% confidence interval when analysed for log-normality by a chi-square test.

Some authors (e.g. Ibbeken, 1983) have suggested that grain size distributions of conglomerates deposited close to their source area should follow Rosin's law, which often describes size distributions resulting from fracture of rocks during weathering. Rosin's law has the form:

$$R = 100 \exp(-x/k)^n$$

where R is the percentage of the population larger than a grain size, x; n is a measure of sorting; and k is the 36.78th percentile, analogous to the mode of a log-normal

distribution (Kittleman, 1964). During transport, sorting processes might be expected to convert Rosin-distributed conglomerates to log-normally distributed conglomerates. Figure 6.3 shows cumulative frequency curves for the three point-counted conglomerate beds, plotted on Rosin's law probability paper. Using the maximum likelihood estimators (for such information about Rosin's distribution, Johnson and Kotz, 1970, Chapter 20 was consulted), best fit Rosin distributions were generated for the three cases, so that chi-square tests could be performed. The results are similar to the chi-square tests for log-normality. This is not unexpected, since the difference between Rosin's law distributions and log-normal distributions is relatively small. The Mamainse conglomerate again passes the chi-square test at the 95% confidence level. The Horseshoe harbor bed again fails unless the finest grain size interval is ignored. The Five Mile Point bed also fails, but passes if the finest grain size interval is ignored. The numerical results of the chi-square tests are presented in Table 6.1. It is evident that with this quality of data, it is not possible to distinguish Rosin's law distributions from log-normal distributions.

In Chapters 3 and 4, there is some discussion of the possibility that the massive, c-m supported, coarse

Figure 6.3 Cumulative frequency - grain size curves for three conglomerate beds plotted on Rosin's law probability paper. A: massive, c-m supported, coarse conglomerate at Mamainse Point. B: imbricated, clast-supported, coarse conglomerate at Five Mile Point. C: massive, c-m supported, massive conglomerate at Horseshoe Harbor.

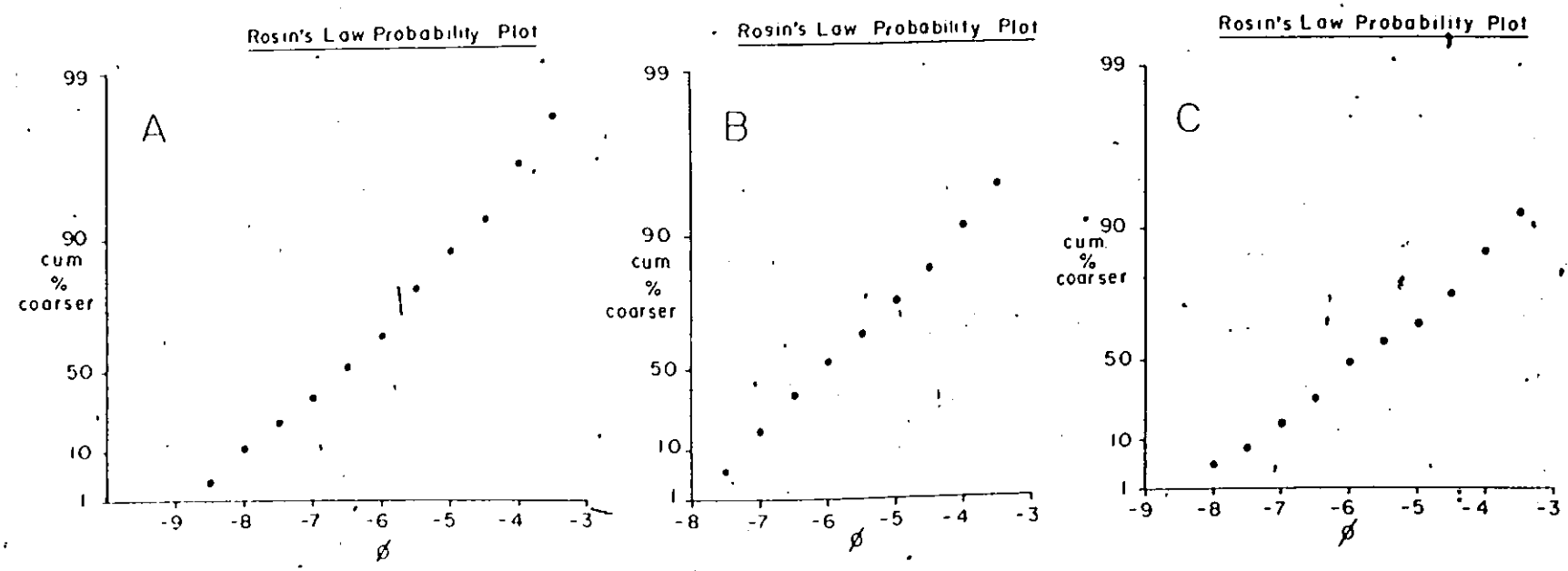

Rosin's Law Probability Plot
A
99
90
cum
%
coarser
50
10
1
-9
-8
-7
-6
-5
-4
-3
ø
Rosin's Law Probability Plot
B
99
90
cum
%
coarser
50
10
1
-8
-7
-6
-5
-4
-3
ø
Rosin's Law Probability Plot
C
99
90
cum
%
coarser
50
10
1
-9
-8
-7
-6
-5
-4
-3
ø

Table 6.1 Results of chi-square tests of conglomerate grain size data for closeness of fit to log-normal and Rosin distributions.

	Minimum Point	Fine Mesh Point	Not enough data
No. finite results	1[illegible]4	1[illegible]4	1[illegible]
No. 1/2 phi intervals	13	10	11
Degrees of freedom	10	7	8
Critical value of chi squared for 95% conf. score	18.31	14.07	[illegible]
Chi squared value for the best fit to a lognormal distribution	7.97	[illegible]	[illegible]
As above, but assuming that data for the finest 1/2 phi interval fits exactly	[illegible]	[illegible]	[illegible]
Chi squared value for the best fit to a Rosin distribution	5.94	[illegible]	[illegible]
As above, but assuming that data for the finest 1/2 phi interval fits exactly	5.94	[illegible]	[illegible]

conglomerate beds at Mamainse Point represent completely reworked debris flows. If this is the case, then it might be expected that there should be a course lag of boulders moveable by debris flow but not by stream flow. No such lag is evident in the grain size distributions.

The Five Mile Point imbricated bed shows a grain size distribution truncated at the coarse end. This is presumably the result of a competency limit. In contrast with the processes responsible for deposition of the massive, c-m supported, coarse conglomerate beds, which seem to have been capable of moving all the sediment sizes available, the processes that formed the imbricated, clast-supported, coarse conglomerate beds apparently could only move part of a grain size population supplied by more powerful processes (i.e. the streamfloods depositing the massive, c-m supported, coarse conglomerates).

6.1.2 Use of grain size data to distinguish aeolian and fluvial sandstones

In Chapter 5 it was shown that palaeocurrent analysis supports an aeolian interpretation of the large scale cross-stratified sandstones at Five Mile Point. To further test this interpretation, eight samples were point counted for

grain size in thin section. The apparent long and short clast axes were measured using an eyepiece graticule, and geometrical means were calculated from these to produce measurements comparable ~~to~~ those that would have been obtained from sieve analysis (method of Middleton 1962).

Frequency curves resulting from the point counts are presented in Figure 6.4. Photomicrographs of two of the sandstones are shown in Figures 6.5 and 6.6. The aeolian sandstones are, in general, better sorted than the fluvial sandstones. IT18 (Figure 6.7) is unique in that it consists of two populations, one with a mode in the coarse sand fraction, the other with a mode at the boundary of very fine sand and silt. Each population is relatively well sorted (the dominant population is as well sorted as the other aeolian sandstones). This kind of bimodal sandstone has been described by Folk (1968), who noted that such sandstones occur in deserts as interdune deposits. Folk (1968) considered that such sandstones result from aeolian winnowing of a poorly sorted population, whereby the easily moved sand sizes are removed, leaving behind coarse and very fine sandstone grains, which are resistant to wind erosion. It is perhaps relevant that the unimodal Five Mile Point sandstones consist predominantly of the grain sizes missing from the bimodal population of IT18. Thus, it would be

Figure 6.4 Grain size - frequency curves for eight Five Mile Point sandstones. Data points represent the number of grains in each 1/4 phi size interval. Data points for fluvial sandstones are shown as solid circles; data points for aeolian sandstones are shown as solid triangles. For each curve, the lowest data points shown correspond to a frequency of zero. See text for discussion.

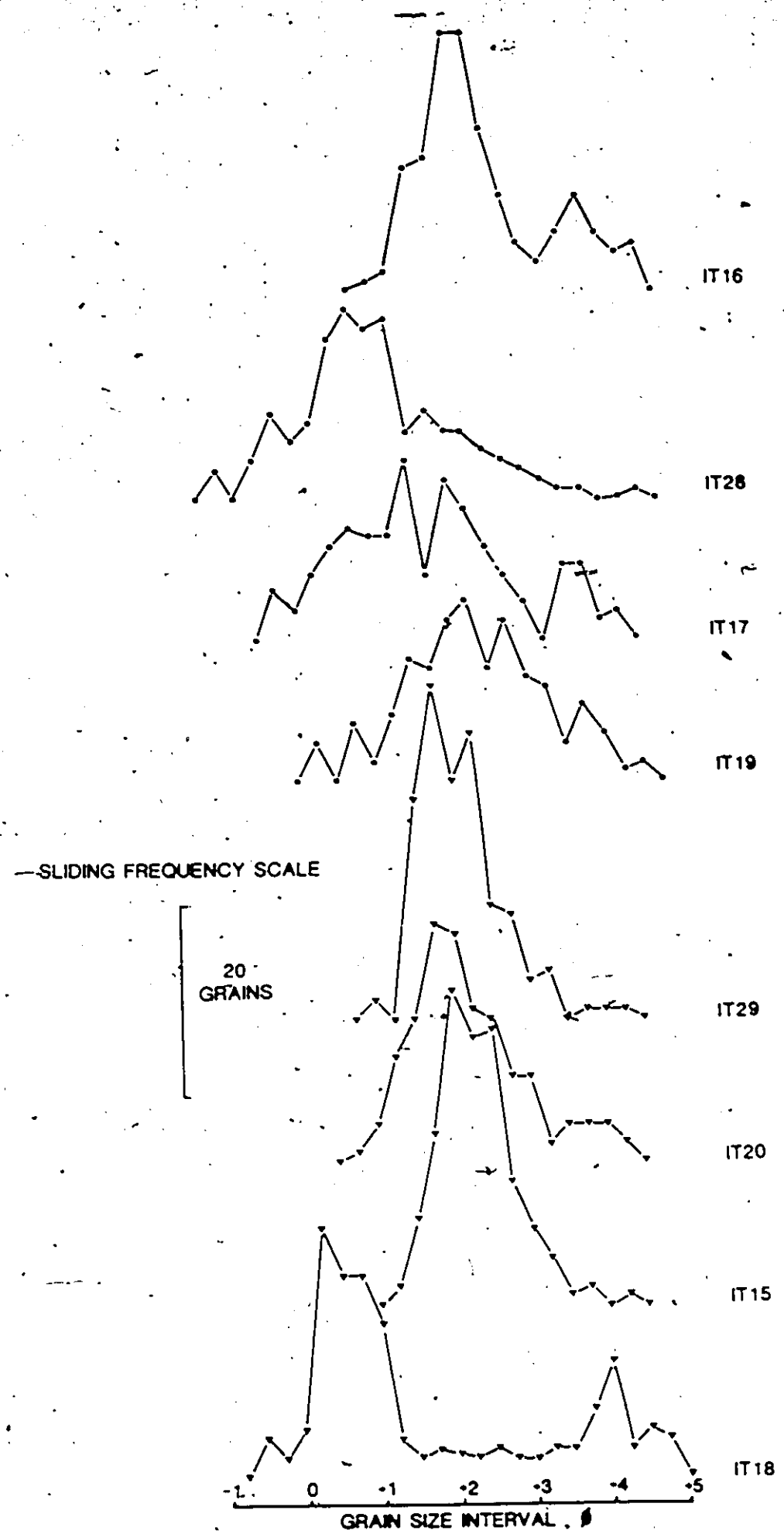
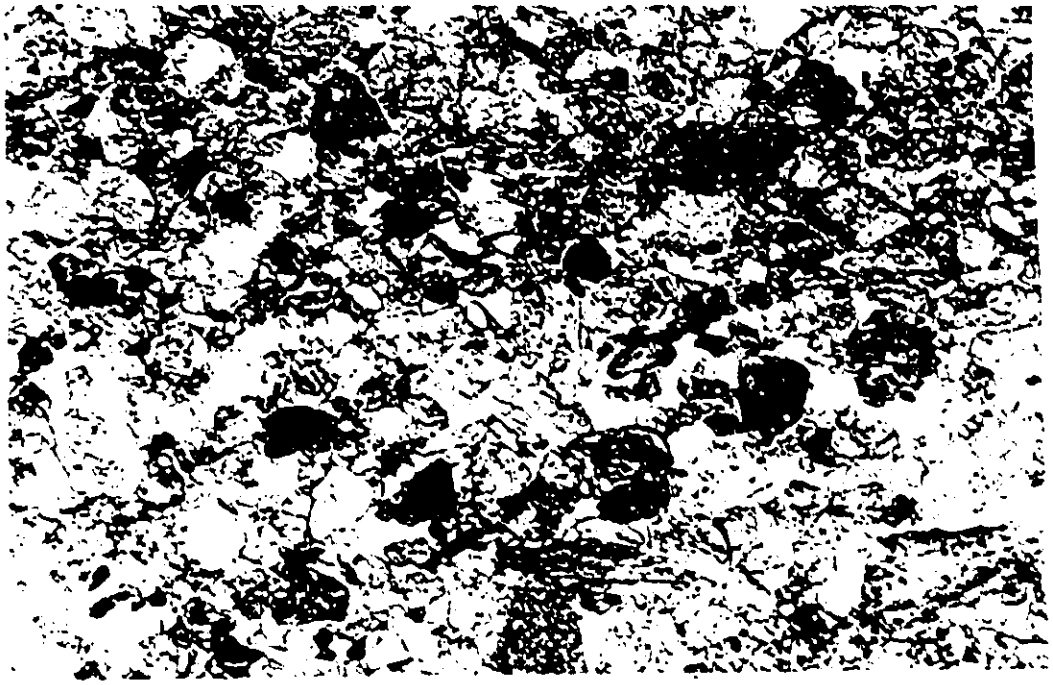

IT16
IT28
IT17
IT19
—SLIDING FREQUENCY SCALE
20
GRAINS
IT29
IT20
IT15
IT18
-1
0
+1
+2
+3
+4
+5
GRAIN SIZE INTERVAL, ϕ

Figure 6.5 Photomicrograph of specimen IT28, a fluvial sandstone from Five Mile Point. Note that the sorting is worse than IT15 (shown below in Figure 6.6.) The maximum width of the field of view is approximately 6.5 mm.

Figure 6.6 Photomicrograph of specimen IT15, an aeolian sandstone from Five Mile Point. Compare with Figure 6.5. The maximum width of the field of view is approximately 4 mm.



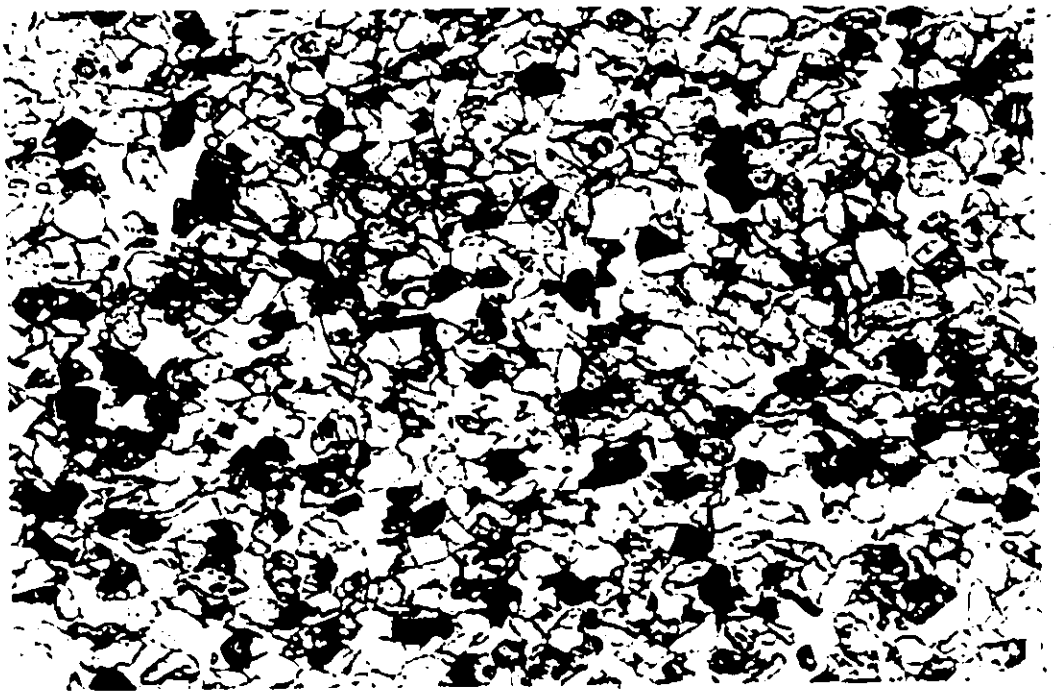
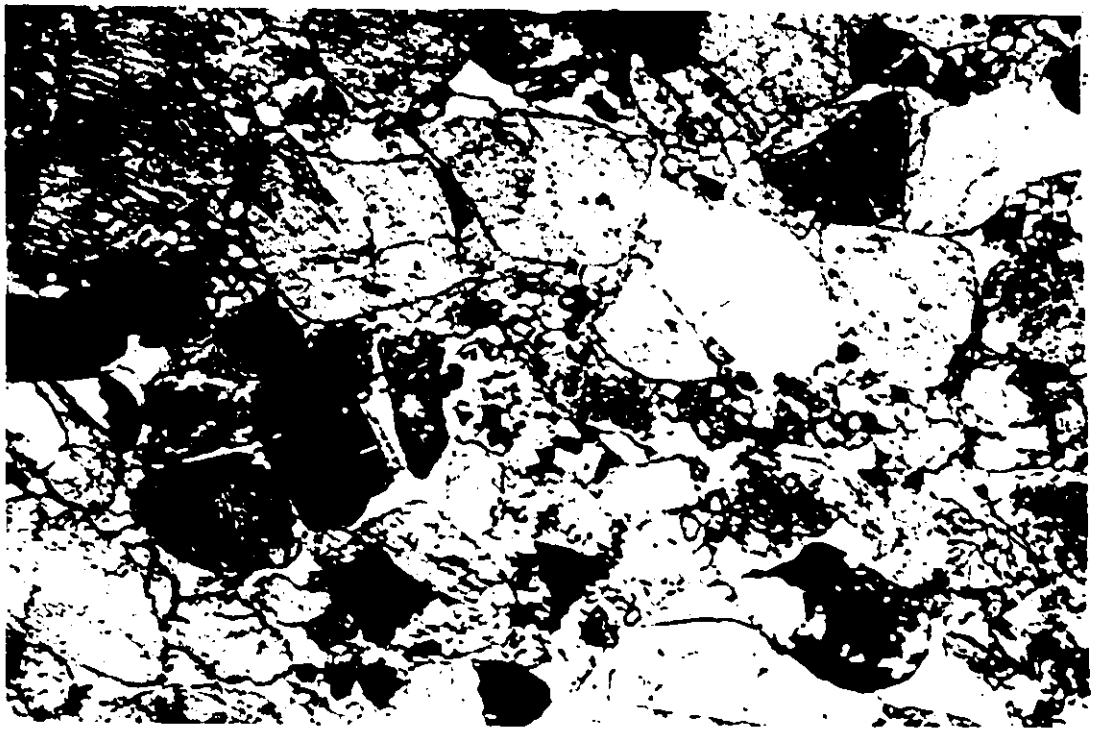


Figure 6.7 Photomicrograph of specimen IT18, showing the marked bimodality of this sandstone. The maximum width of the field of view is approximately 4 mm.



possible to produce all the observed grain size distributions of the aeolian sandstones by winnowing one of the poorly sorted fluvial sandstones (i.e. IT28, IT19 or IT17).

IT16 is better sorted than the other fluvial sandstones, and it may be that both fluvial and aeolian processes transported this sand prior to deposition.

IT29, IT20, IT15, IT16, and IT28 contain a few examples of what are here interpreted as rotted oncoliths (Figures 6.8 and 6.9). These appear similar to the oncoliths seen in sandstones associated with stromatolites (Figures 6.10 and 6.11). Since the stromatolites probably formed around lakes on the distal portions of the Keweenawan alluvial fans, the occurrence of such oncoliths at Five Mile Point would appear to indicate upslope transport, presumably by aeolian means. It should be noted, however, that Kalliokoski (1986), in a study of interflow sediments at Centennial mine, on the Keweenaw Peninsula, considered sand grains of similar appearance to the Five Mile Point oncoliths to have been formed during caliche development.

Figure 6.8 An oncolith within specimen IT16 from Five Mile Point. The maximum width of the field of view is 0.63 mm. Compare with Figures 6.9, 6.10 and 6.11.

Figure 6.9 An oncolith within specimen IT15 from Five Mile Point. The nucleus of this oncolith is a granular aggregate of carbonate grains, the only other occurrence of which is in close association with stromatolite-oncolite beds, where carbonate with the same texture occurs both as discrete grains and as oncolith nucleii. An example of the latter case is shown in Figure 6.11 for comparison.

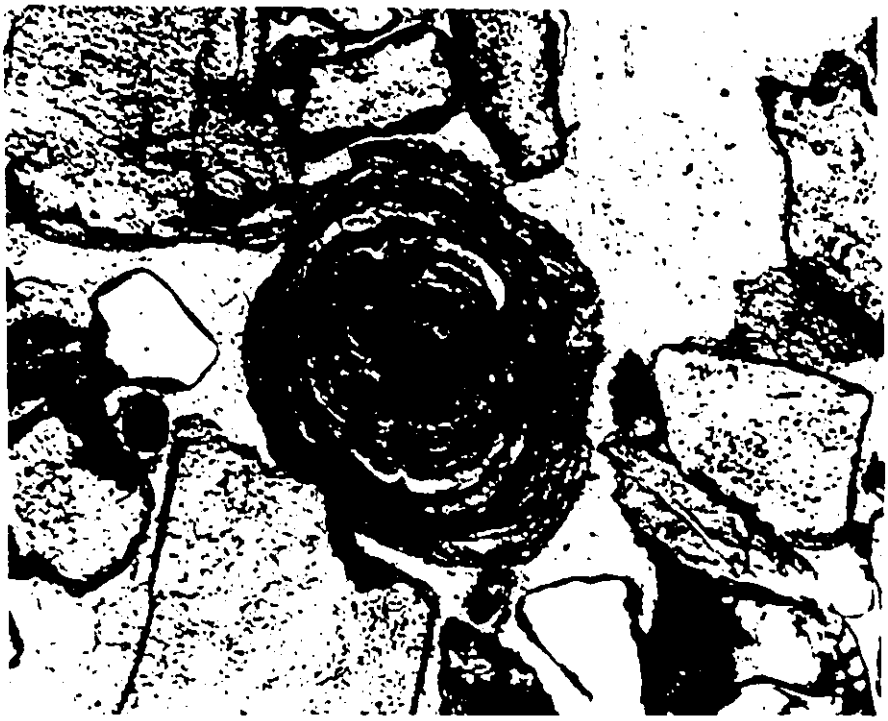
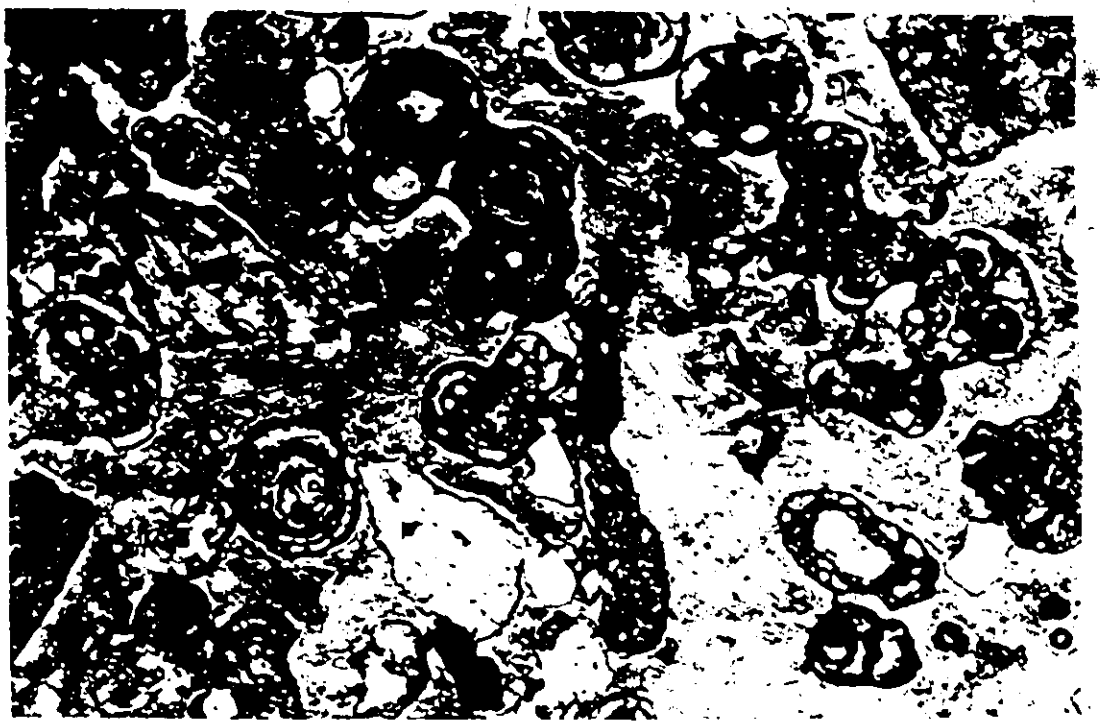
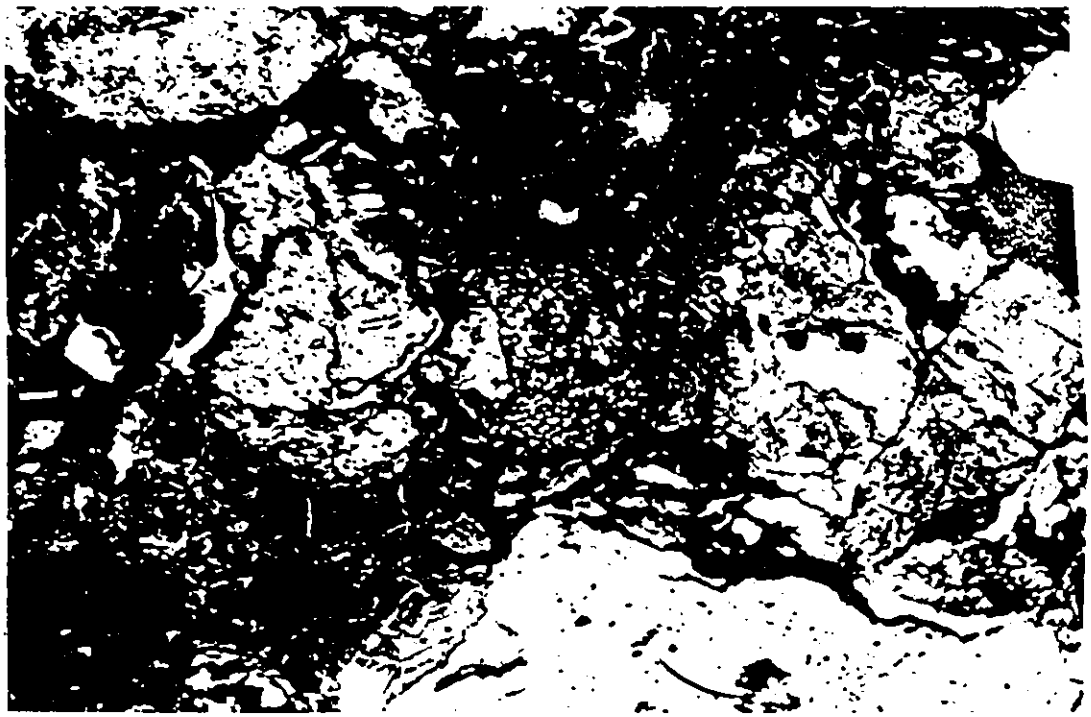


Figure 6.10 A low power photmicrograph of specimen IT5 from Horseshoe Harbor close to a stromatolite horizon. Oncoliths are seen in various states of preservation, some as amalgamated (grapestone) grains, an example of which is arrowed. The red stain is alizarin red S and potassium ferricyanide, which in this region of the slide is rather patchy. The maximum width of the field of view is 4 mm.

Figure 6.11 A badly crushed oncolith in specimen IT5, with a granular carbonate nucleus very similar to that seen in IT15 (Figure 6.9). The red stain is alizarin red S and potassium ferricyanide. The maximum width of the field of view is 1.6 mm.





6.1.3 Grain size trends and bed thickness - maximum clast size relationships

As mentioned earlier in this Chapter, grain size and bed thickness show no pronounced cyclical variations moving up-section through the Copper Harbor Formation. The gross fining-upwards nature of the Copper Harbor Formation was briefly mentioned in Chapter 4.

Analysis of bed thickness variation is hindered by the poorly defined nature of most conglomerate beds. At Eagle River, however, beds are relatively clearly defined, although their lateral extent is not apparent at this outcrop. Figure 6.12 shows grain size and bed thickness variations at Eagle River, presented as a weathering profile. This data tends to confirm the observation that no regular variations are present. Figures 6.13 and 6.14 graphically present the relationship between bed thickness and maximum particle size (measured as the average of the ten largest clasts in a bed). The data is very scattered, and least squares analysis of the log-log plot generates a regression line with gradient 0.9 or 2.7 depending whether grain size or bed thickness is treated as the independent variable. The correlation coefficent for the log-log plot is 0.61. Despite the scatter, there is a general increase in

Figure 6.12 Bed thicknesses and grain sizes for conglomeratic section in Eagle River gorge, presented as a weathering profile. The base of the profile is at approximately the same stratigraphic level as the top of the Eagle River measured section. The maximum particle size is the average of the long axes of the 10 largest clasts in each bed.

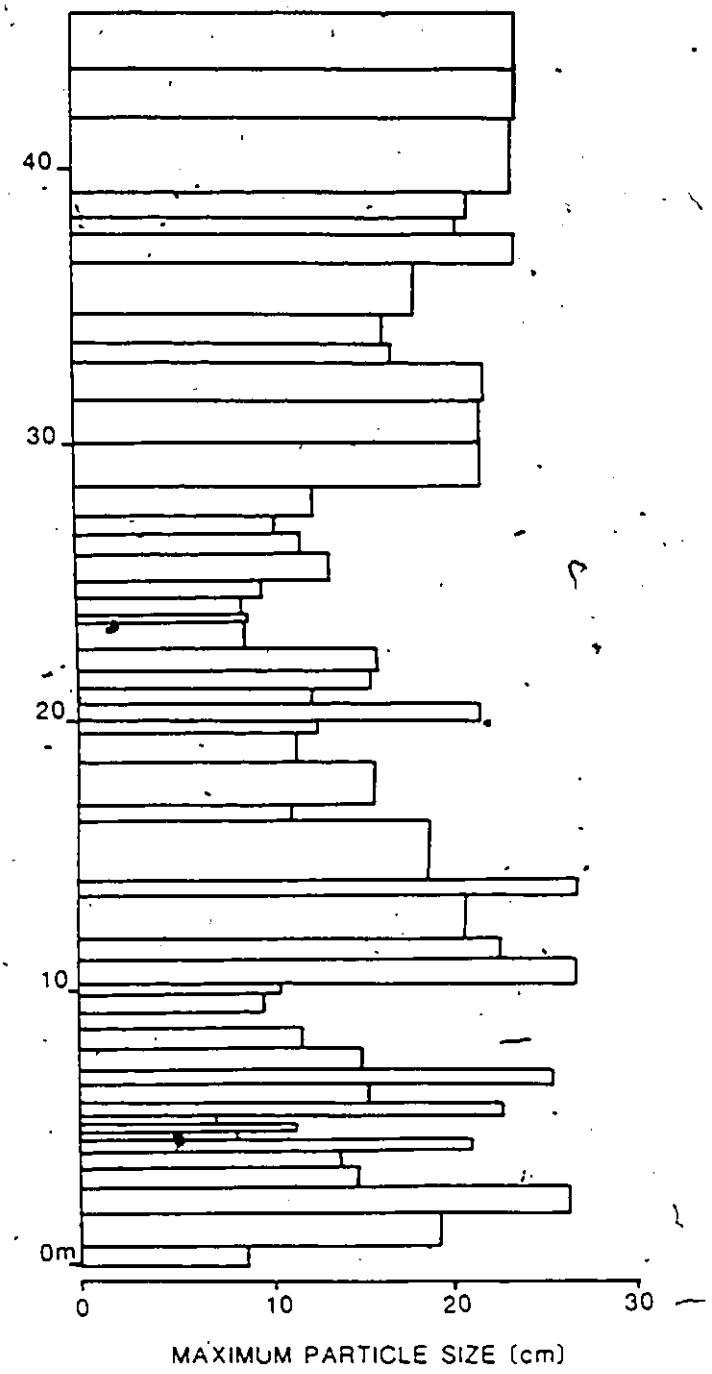

40
30
20
10
0m
0
10
20
30
MAXIMUM PARTICLE SIZE (cm)

Figure 6.13 Grain size and bed thickness data for conglomerates in Eagle River gorge. The maximum particle size is the average of the 10 largest clasts in each bed. See also Figures 6.12 and 6.14.

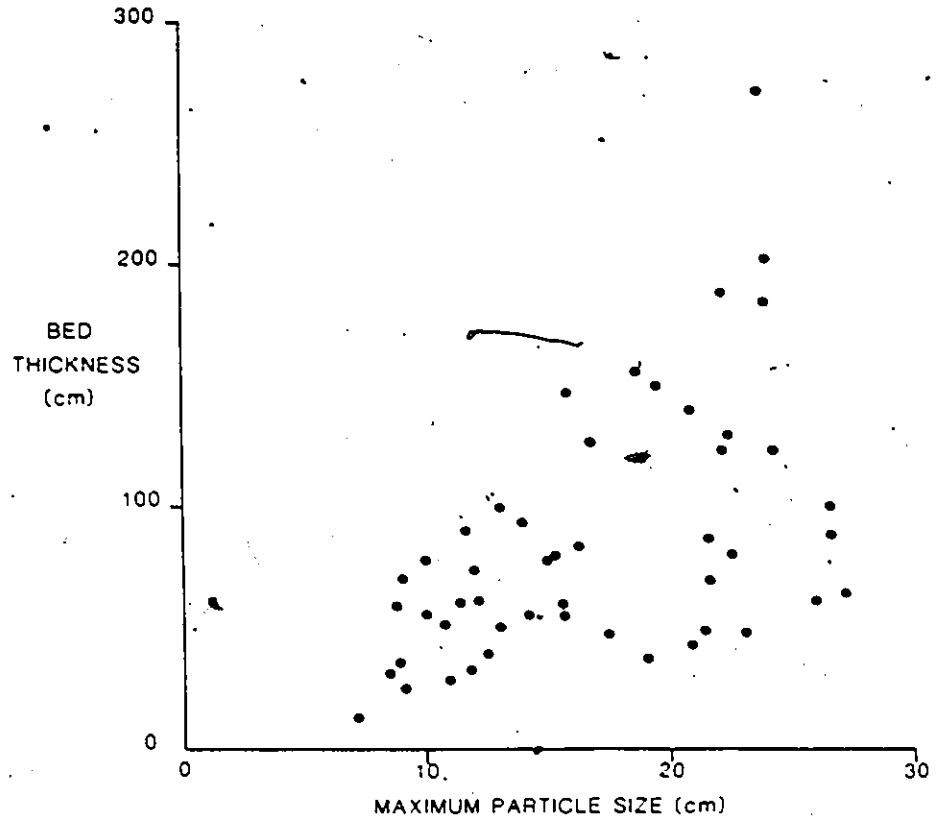
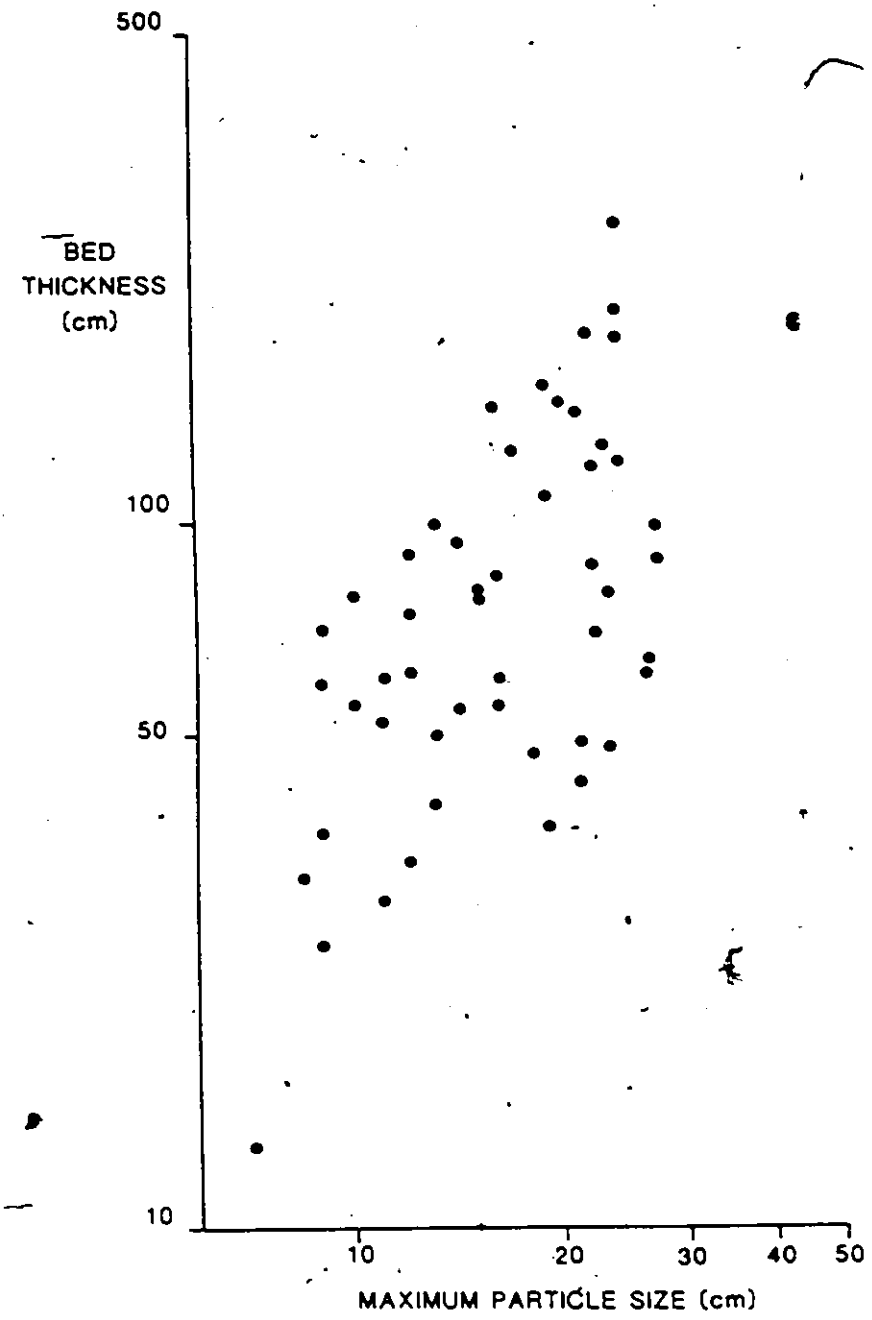

300
200
100
0
BED
THICKNESS
(cm)
0
10
20
30
MAXIMUM PARTICLE SIZE (cm)

Figure 6.14 Grain size and bed thickness data for conglomerates in Eagle River gorge, presented as a log-log plot. The maximum particle size is the average of the long axes of the ten largest clasts in each bed. See also Figures 6.13 and 6.12.


500
100
50
10
BED
THICKNESS
(cm)
10
20
30
40
50
MAXIMUM PARTICLE SIZE (cm)

maximum grain size with increasing bed thickness. Although such a trend has been noted for debris flows (e.g. Steel, 1974), and has been explained theoretically (Johnson, 1970), there is no theoretical explanation for the trend observed in the fluvially deposited conglomerates at Eagle River. One possible explanation is given below.

Shield's diagram, although not dealing with data for conglomerate grain sizes, would appear to indicate that for large grains, the maximum moveable grain size is approximately proportional to the basal shear stress provided by a stream.

$$\text{i.e. } T_o/(p_s-p)gD = 0.06 \quad \ldots\ldots 1$$

(where T_o is the basal shear stress, p_s and p are the sediment and water densities respectively, g is acceleration due to gravity, and D is the maximum diameter of (perfectly sorted) sediment grains that can be moved. For a river of depth, d, and slope, s, the basal shear stress is given by

$$T_o = pgsd \quad \ldots\ldots 2$$

If slope is constant (i.e. considering only one segment of an alluvial fan profile), then equations 1 and 2 would predict a proportionality between d and D. If the bed thicknesses measured in the field are assumed to be bar thicknesses, and it is further assumed that bar heights were approximately the same as river depths, then it follows that

there should be a proportionality between bed thickness and D. Quite clearly, this argument contains many dubious assumptions, but it is presented here as one possible explanation for the observed bed thickness - maximum particle size relationship.

6.2 Diagenetic features of sandstones of the Copper Harbor Formation

6.2.1 Previous work; aims of this study

The native copper deposits of the Portage Lake Lava Series have drawn much attention to the effects of diagenesis and low grade metamorphism displayed by the basalts and the interflow sediments. Butler and Burbank (1929), and more recently White (1968), have dealt with this subject at length. Broderick (1929) and Stoiber and Davidson (1959) have succeeded in finding some pattern to the mineralisation on a regional scale, but neither study extended to the overlying Copper Harbor Formation. Of the sedimentological studies that have focused on the Copper Harbor Formation, several have involved petrography, but generally only with the purpose of determining clast compositions and thus provenance (e.g. Daniels, 1982; Wolff

and Huber, 1973). Two recent studies have dealt with specific aspects of the diagenesis. Elmore (1981) investigated the evidence that the haematite stain in the Copper Harbor Conglomerate is diagenetic, and considered the origin of the palaeomagnetism. Kalliokoski and Welch (1985) and Kalliokoski (1986) have claimed to identify a caliche palaeosol in the interflow conglomerate exposed in Centennial Mine near Calumet.

This petrographic study therefore aims to describe the diagenetic history of some of the sediments of the Copper Harbor Formation, an area thinly covered by previous work. Five sandstone samples were chosen from outcrops on the Keweenaw Peninsula, this area being selected in order that comparisons might be drawn with the work that has been done on the copper-bearing rocks. Sample locations are given in Table 6.2. The study focuses in particular on the properties of the authigenic clays (abundant in places), the occurrence of feldspar cement, and some of the details of carbonate cementation.

6.2.2 Methods of study

Thin sections were initially studied using a petrographic microscope. The calcareous sandstones (IT5 and

Table 6.2 Locations of samples chosen for study of diagenetic features.

SAMPLE	LOCATION
IT2	Five Mile Point, western section above lava, 25 m level (section of Figure 4.15). The sample is from large scale cross-stratified sandstone (facies S-13).
IT4	Veale Memorial Park (stratigraphically and geographically midway between the Five Mile Point and Eagle River locations shown in Figure 1.3). The sample is from pebbly trough cross-stratified sandstone (facies S-5).
IT5	Horseshoe Harbor section, 19 m level (section of Figure 4.4). The sample is from pebbly trough cross-stratified sandstone (facies S-5) with a large amount of calcareous material.
IT6	Dan's Point, 3 m level (section of Figure 4.14). The sample is from an oncolithic part of a stromatolite-oncolite bed.
IT15	Five Mile Point, eastern section above lava, 29 m level (section of Figure 4.17). The sample is from large scale cross-stratified sandstone (facies S-13).

IT6) were stained with alizarin red S and potassium ferricyanide to distinguish different carbonates. Two of the relatively non-calcareous sandstones (IT2 and IT4) were stained with sodium cobaltinitrate, barium chloride and potassium rhodizonate to distinguish different feldspars. Sections were also observed under a luminoscope, operated at 12.5 kV, paying particular attention to cement stratigraphy and overgrowths. Specimens IT2 and IT4 were studied under S.E.M. at 15 kV to reveal details of clays and haematite staining. Samples were broken and mounted on S.E.M. stubs broken side up, and were coated with gold. Two samples of IT2 were used, one having been etched in 38% HCl to remove pore filling calcite cement and reveal clay pore linings.

6.2.3 Sample compositions

The modal composition of the samples is shown in Tables 6.3 amd 6.4, and is compared in Figure 6.15 with "average" compositions for the Copper Harbor sandstones calculated by previous workers. The samples studied evidently contain more feldspar than most Copper Harbor sandstones.

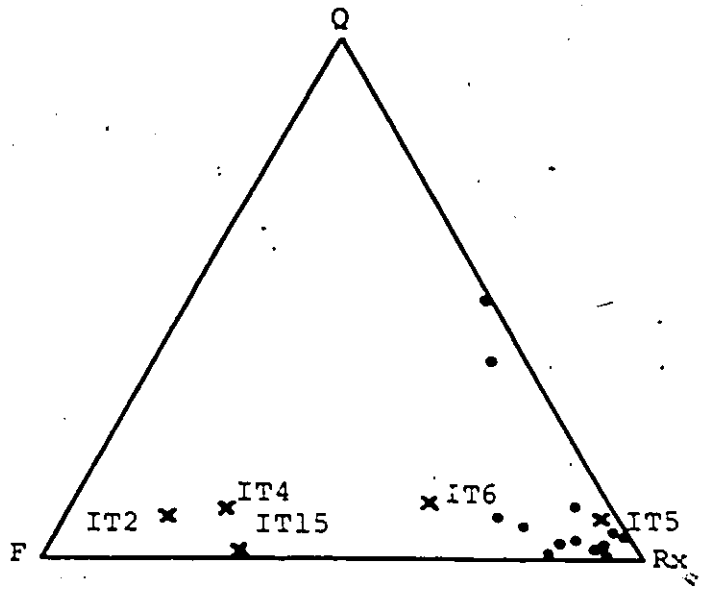
Table 6.3 Modal analysis of specimens IT2, IT4, and IT15. 300 point were counted for each analysis. For point counting purposes, untwinned feldspars were considered to be potassium feldspar (although IT2 and IT4 were stained for feldspar, the high degree of sericite alteration prevented the stain from clearly discriminating between plagioclase and potassium feldspar, the clay particles proving very receptive to the stains; as far as could be ascertained, the stains confirmed the dominance of potassium feldspar in IT2 and the dominance of plagioclase in IT4). Volcanic rock fragments range from felsic to basic. Some opaques show alteration laminae characteristic of magnetite, but some may be highly altered basic volcanic grains. Plutonic rock fragments have an approximately granitic composition. IT2 contains minute amounts of quartz cement around some quartz grains. IT4 contains tourmaline as an accessory.

	IT2	IT4	IT15
GRAINS			
Quartz	1	2	--
Plagioclase	9	10	6
K feldspar	31	7	31
Volc. frags.	12	10	18
Plut. frags.	5	5	<1
Opaques	4	27	7
Chert	4	1	<1
INTERGRANULAR CEMENT			
Calcite	14	2	10
Epidote	--	6	--
Chlorite	6	12	7
Feldspar	4	--	7
Red stain	5	3	7
MINERALS REPLACING GRAINS			
Calcite	<1	3	1
Epidote	--	8	2
Chlorite	5	4	4

Table 6.4 Modal analysis of the calcareous sandstone specimens, IT5 and IT6. 200 points were counted for each analysis. Amalgamated grains (grapestones) and possible ooliths (a few grains in IT6 only) were regarded as oncoliths for point counting purposes. Oncolith nucleii consist of feldspar and volcanic rock fragments, so the proportion of the rock consisting of these materials is greater than indicated. Comments are otherwise as for Table 6.3.

	IT5	IT6
GRAINS		
Quartz	2	2
Plagioclase	<1	--
K feldspar	<1	5
Volc. frags.	42	10
Oncoliths	17	44
Opaques	--	2
Sedimentary frags.	2	2
INTERGRANULAR CEMENT		
Calcite spar	36	34
Micrite	--	<1
Mud	--	<1

Figure 6.15 Compositions of samples studied in this project with "average" Copper Harbor Formation sandstones of previous workers (from Daniels, 1982).


Q
F
Rx
IT2
IT4
IT15
IT6
IT5

6.2.4 Diagenetic histories of the non-calcareous sandstones (IT2, IT4 and IT15)

Although the five samples come from fairly closely spaced locations, diagenetic histories vary considerably between samples. The oncolith bearing samples IT5 and IT6 display a very different style of diagenesis to the relatively non-calcareous samples IT2 and IT4, so the samples are treated in two groups. The following diagenetic events are presented in the order in which they appear to have affected the non-calcareous sandstones.

6.2.4.1 Compaction

Both these samples have experienced significant compaction, but IT4, which lacks structurally strong cements, has been compacted to a greater degree than IT2 and IT15, where feldspar and calcite cements have provided resistance to compaction. The amount of porosity lost as a result of compaction cannot be estimated accurately. From consideration of the evidence that grains have dissolved, leaving only the stained rims (Figure 6.16), it seems highly probable that many grains have dissolved completely without any stained rims surviving as evidence, thereby increasing

Figure 6.16 Photomicrograph of specimen IT2 showing grains replaced by calcite and feldspar cement (arrowed). Plane polarised light (top); cross polarised light (bottom). Field of view is about 1 mm.

the apparent percentage of intergranular cement. Consequently, minus-cement porosities, obtained by subtraction of intergranular cement (Table 6.3) are probably overestimates. The minus-cement porosities are 24% in IT4, 30% in IT2, and 31% in IT15. On the basis of the sorting in these rocks, porosities of the unconsolidated sands may be estimated to have been between 30% and 40% (Beard and Weyl, 1973). IT4, the most poorly sorted, probably had an initial porosity near 30%, whereas IT2 and IT15, which are better sorted, probably had initial porosities of 35% or more (using the comparators provided in Beard and Weyl, 1973). It therefore appears that compaction accounted for between 5% and 10% of the porosity loss in IT4 and somewhat less in IT2 and IT15. Evidence of compaction includes crushed volcanic rock fragments (Figure 6.17), many long contacts between grains (Figure 6.17), concavo-convex contacts and sutured contacts (Figure 6.18).

6.2.4.2 Red staining

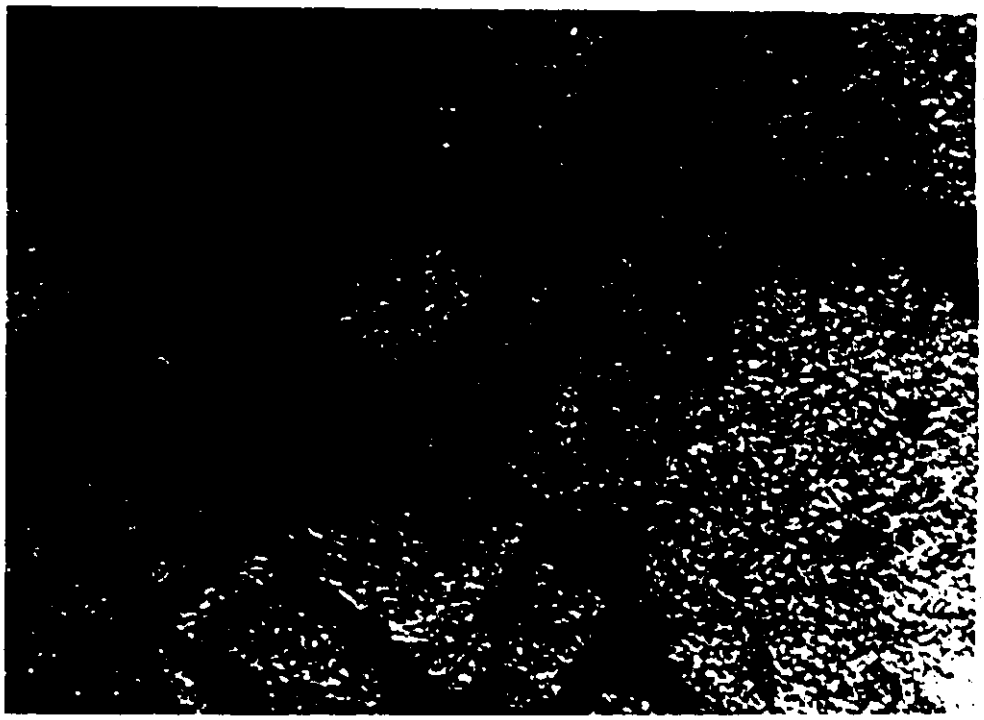
All the grains in these samples have red iron-stained rims. Figures 6.17 and 6.19 show that the stain is often present between grain contacts, indicating that at least some staining was early, prior to much of the compaction,

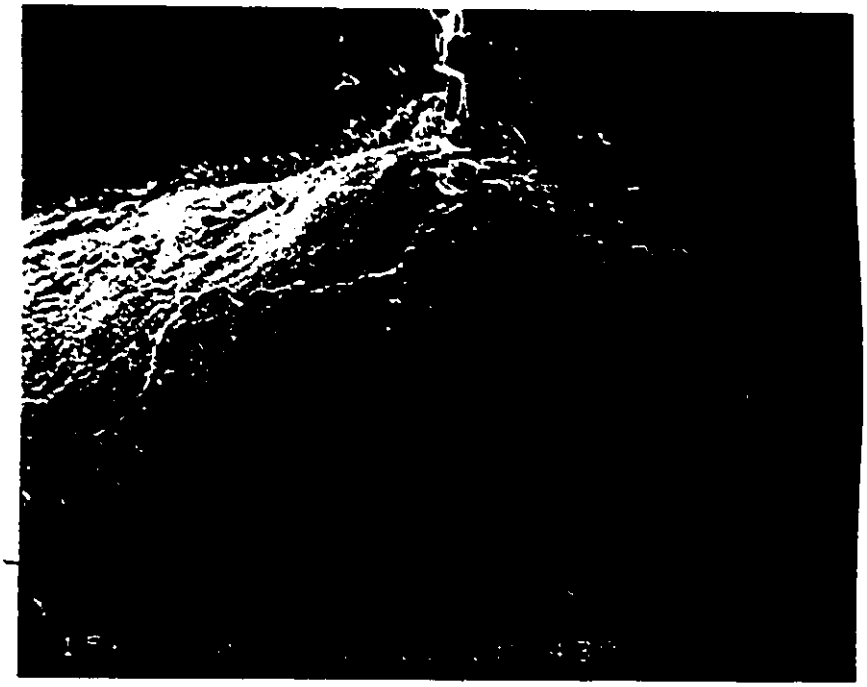
Figure 6.17 Crushed volcanic fragments and long grain contacts (arrowed) in specimen IT2. Plane polarised light. Maximum width of field of view is 1.6 mm.

Figure 6.18 Sutured grain contact (arrowed) in specimen IT2. Plane polarised light. Maximum width of field of view is approximately 0.6 mm.

Figure 6.19 Feldspar grains in specimen IT2 showing iron staining and chlorite rims (arrowed). Plane polarised light (the pastel shades in this photograph are artificial; the blue colouration of the top right of the field of view is due to epoxy impregnation of the sample). Maximum width of field of view is approximately 0.4 mm.

Figure 6.20 S.E.M. view of IT2 (unetched sample), showing chlorite rim, partly broken away from the underlying grain, and overlain by calcite cement. The white line (bottom centre) is a scale bar and is 18.2 microns in length.





and perhaps prior to deposition. S.E.M. investigations did not find any identifiable haematite crystals. This is probably the result of two factors:

1. much of the haematite may be present as tiny grains smaller than the limit of resolution (ultra-fine pigment of Walker et al., 1981), a conclusion in agreement with the investigations of Elmore (1981);

2. larger grains of haematite (and perhaps iron rich clay) have been engulfed by the chlorite that rims the grains, so when the chlorite rim breaks away from the grain to reveal the grain surface (Figure 6.20), any underlying haematite crystals are also torn away, never to be observed.

6.2.4.3 Epidote and chlorite cementation

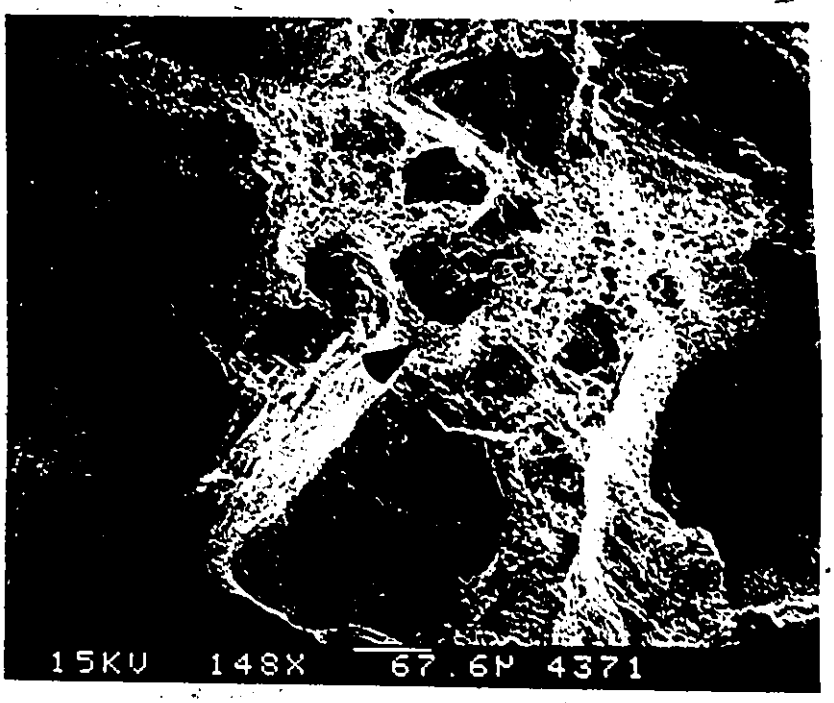
As mentioned above, the haematite stain is most often overlain by a chlorite rim (Figures 6.19, 6.20). This is not present at grain contacts and does not appear to have been crushed, and so post-dates the initial compaction. S.E.M. study shows that the chlorite is more widespread than thin section analysis alone would indicate, forming rims around the majority of grains. The etched specimen of IT2, in which it was possible to see down into pores where calcite had been present, revealed most about the chlorite rims. More

chlorite is present in IT4, but the structure is not so well seen because the pores are completely clogged and only fractured surfaces are available for study of the clays. Figures 6.21, 6.22, and 6.23 show the boxwork style of the chlorite rims, which are clearly authigenic in origin. Figure 6.20 shows a chlorite rim (in IT2, unetched) partly broken away, revealing the underlying (stained) grain surface, and overlain by calcite cement. X-ray analyses (e.g. Figure 6.24) indicate that the clay is probably a chlorite, although certain smectites have very similar compositions and may show a similar ragged appearance. Comparison with spectra from Welton (1984) (Figure 6.25) shows that the spectrum is very similar to chlorites observed in other rocks, but is also very similar to certain smectites, so the identification cannot be made solely on grounds of the spectral analysis. The anomalous birefringence colours and the green colouration of the rims in thin section are indicative of chlorite. Comparison of Figure 6.26 and Figure 6.24 shows that the chlorite nearest the grain surface may be more iron rich than that further into the pore.

Two other forms of clay are present. Figure 6.27 shows small spherical aggregates, which occur between chlorite flakes in specimen IT4. These cannnot be analysed

Figure 6.21 S.E.M. view of IT2 (etched specimen), showing chlorite rims to grains. The arrow indicates the location of Figure 6.22. The white scale bar (bottom centre) is 67.6 microns in length.

Figure 6.22 Close-up S.E.M. view of the chlorite rim shown in figure 6.21. The white scale bar (bottom centre) is 6.76 microns in length.


15KU 148X 67.6µ 4371

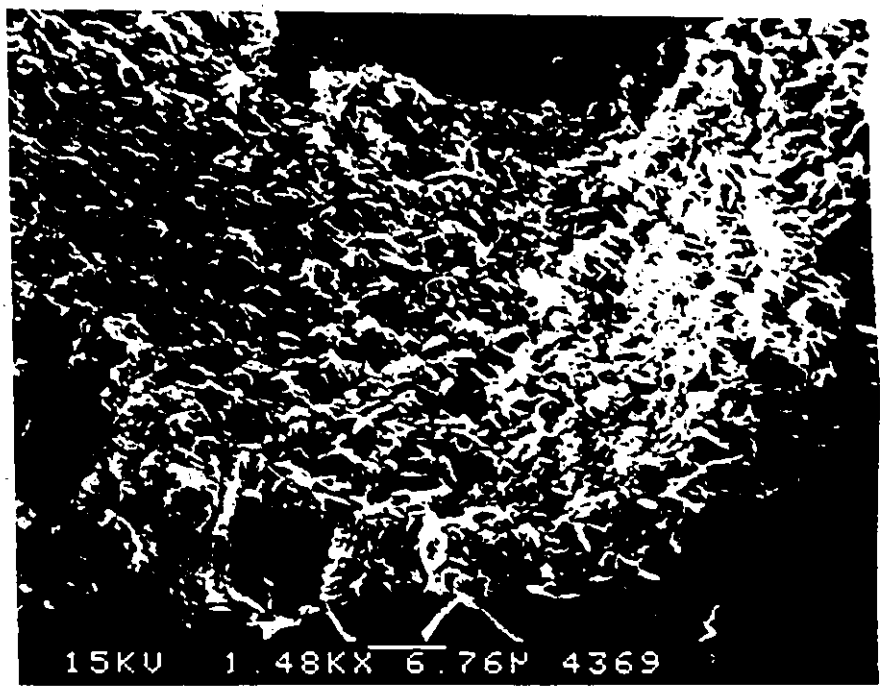

15KU 1.48KX 6.76µ 4369

Figure 6.23 Very close-up view of the chlorite rim shown in Figures 6.21 and 6.22. The white scale bar (bottom centre) is 1.71 microns in length.

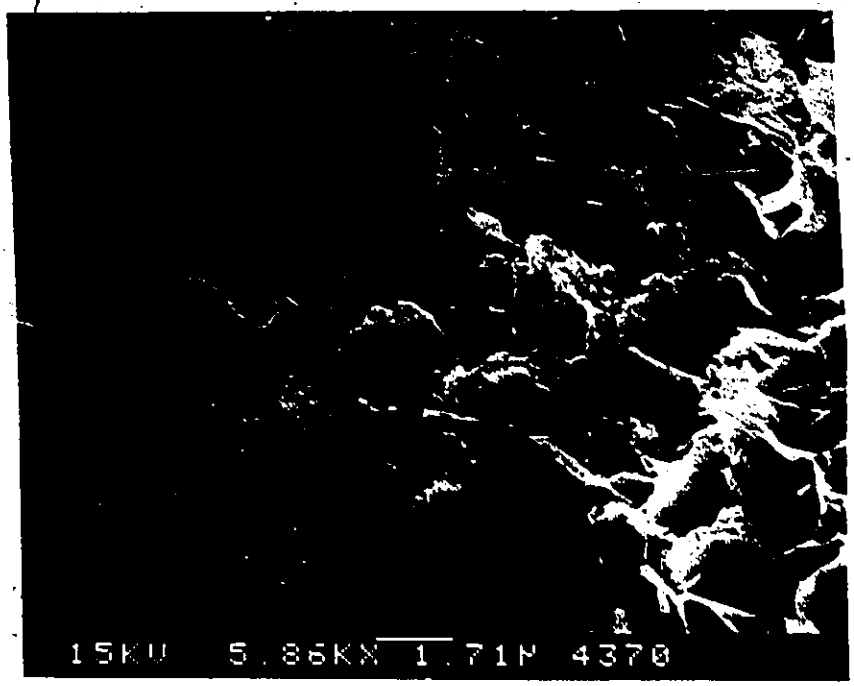

15KV 5.86KX 1.71M 4370

Figure 6.24 X-ray analysis of a chlorite rim in IT2, taken from the surface of the rim (the last stage of chlorite growth). The chlorine peak is a result of HCl etching.

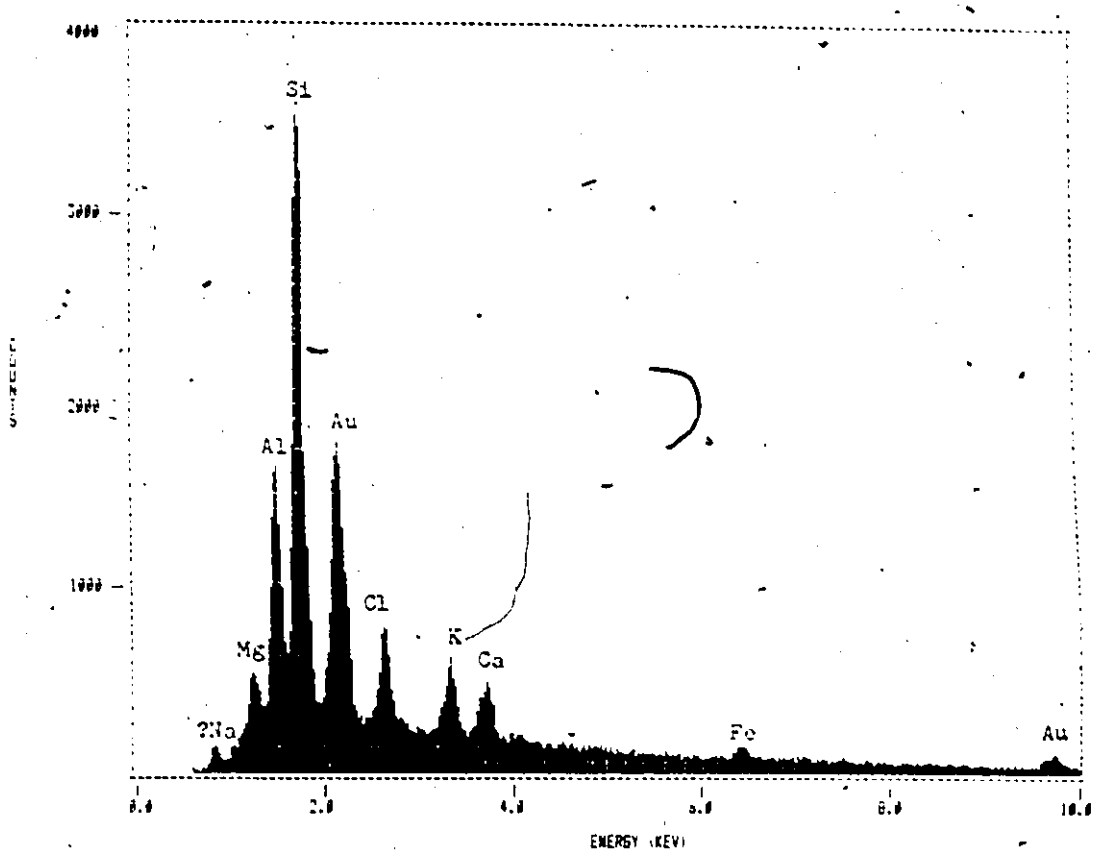
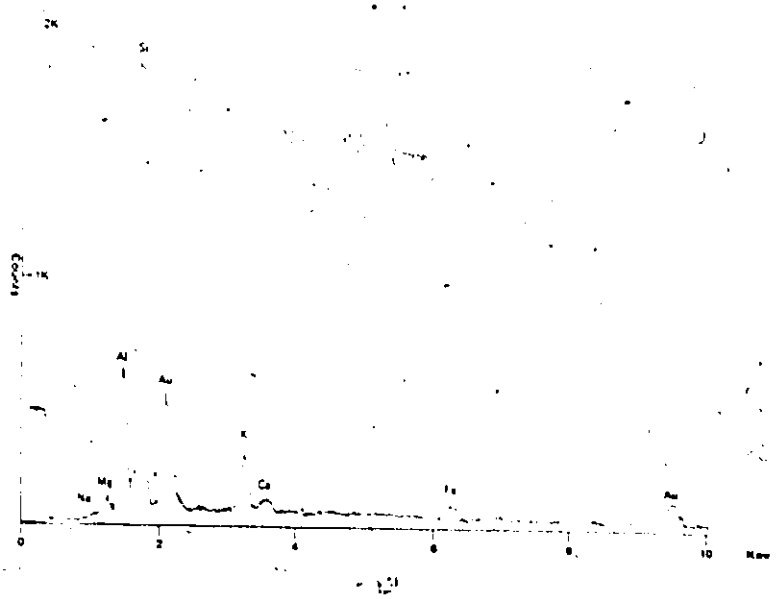

Si
Al
Au
Cl
K
Ca
Mg
Na
Fe
Au
ENERGY (KEV)

Figure 6.25 X-ray spectra from Welton (1984), indicating the difficulty of distinguishing chlorite and smectite. Cu and Cl are within other minerals close to the clay.

Energy Dispersive X—Ray Spectrum (EDX)

SMECTITE



Energy Dispersive X—Ray Spectrum (EDX)

CHLORITE

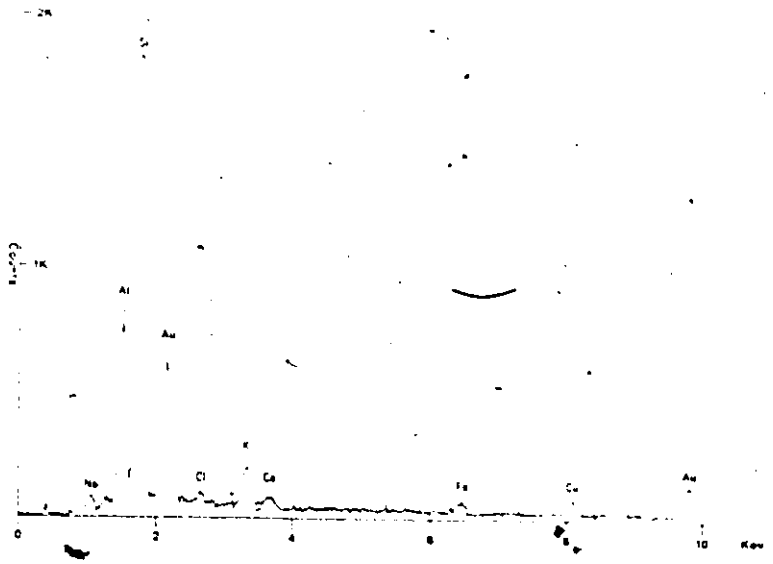


Figure 6.26 X-ray spectrum from chlorite just above the surface of the grain it is rimming (specimen I-T2).

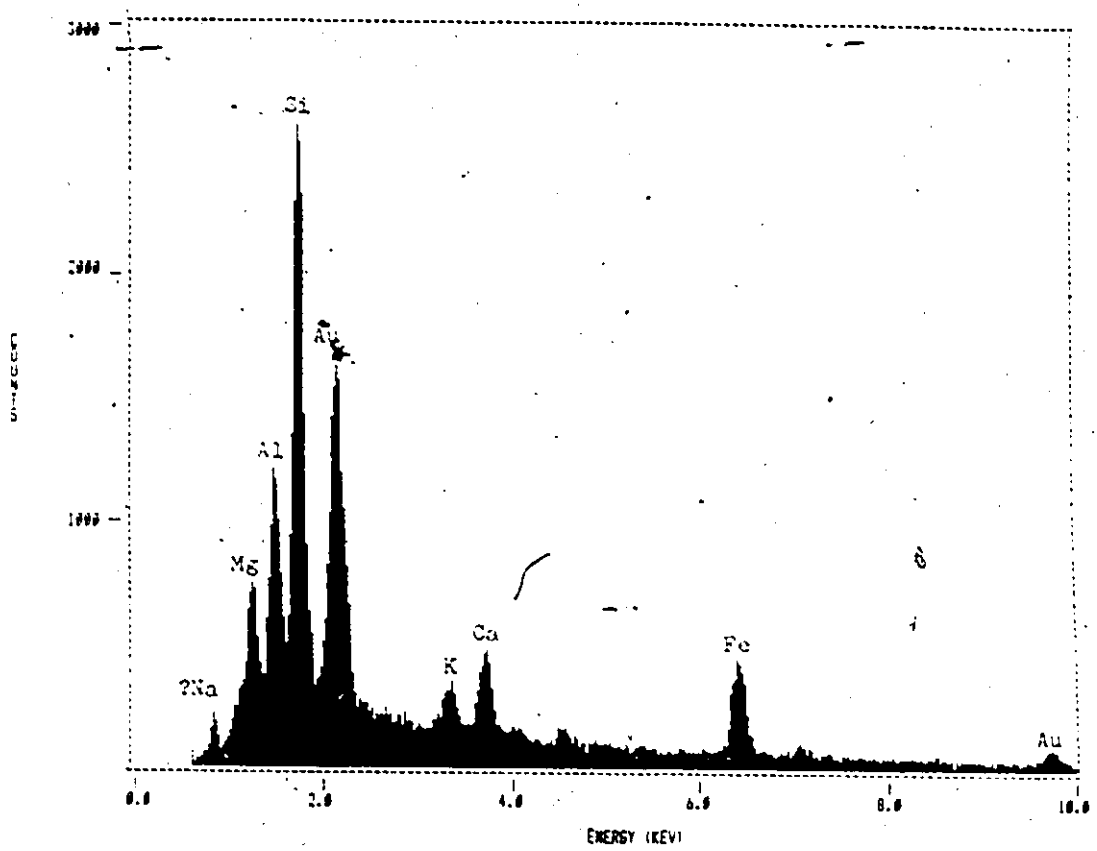
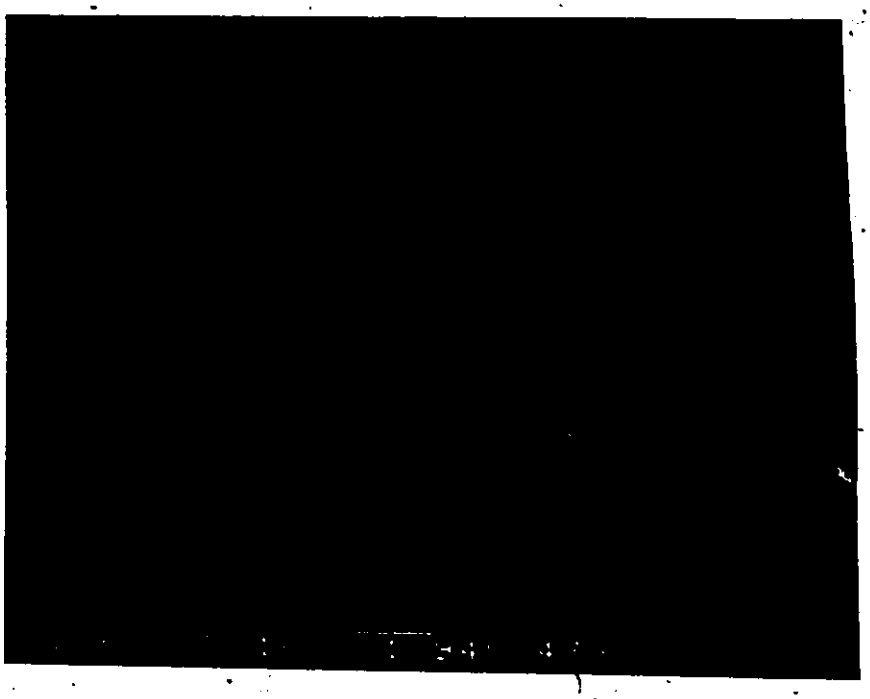

Si
Au
Al
Mg
2Na
K
Ca
Fe
Au
0.0
2.0
4.0
6.0
8.0
10.0
ENERGY (KEV)

Figure 6.27 Tiny crystal aggregates of ?clay between flakes of chlorite in specimen IT4. The white scale bar (bottom centre) is 1.94 microns in length.



by x-ray methods because the spheres are overshadowed by chlorite (and nearby grains). Consequently x-ray count rates are very low and probably only show the influence of surrounding clay. The spectrum obtained is therefore noisy and uninformative (Figure 6.28). The spherical aggregates are not framboids, because they do not consist of cubic pyrite crystals, but rather of small plates. Kaolinite sometime takes a nodular form (Byorlykke, 1983, his Figure 18d). There is no documentation of chlorite or haematite taking this form, but neither can be ruled out. Wilson and Pitman (1973) point out that various types of silica may show forms rather similar to this. Dissolution and alteration of the large number of volcanic fragments in this rock may have raised silica concentrations to high levels at certain stages during diagenesis.

In IT2 a vermicular form of chlorite is seen within a feldspar (Figure 6.29). Books and worms of crystals like this are often kaolinite, but the green colouration and anomalous birefringence colours indicate that this clay is chlorite. This form of chlorite is generally found within hydrothermal quartz (Blatt et al., 1980, Figure 8-4), but when the rock was later stained, this grain took a greenish (potassium feldspar) stain, and the optical figure appears to be biaxial. As Figure 6.29 indicates, the chlorite is

Figure 6.28 Noisy and probably unreliable x-ray spectrum from one of the spherical crystal aggregates shown in Figure 6.27.

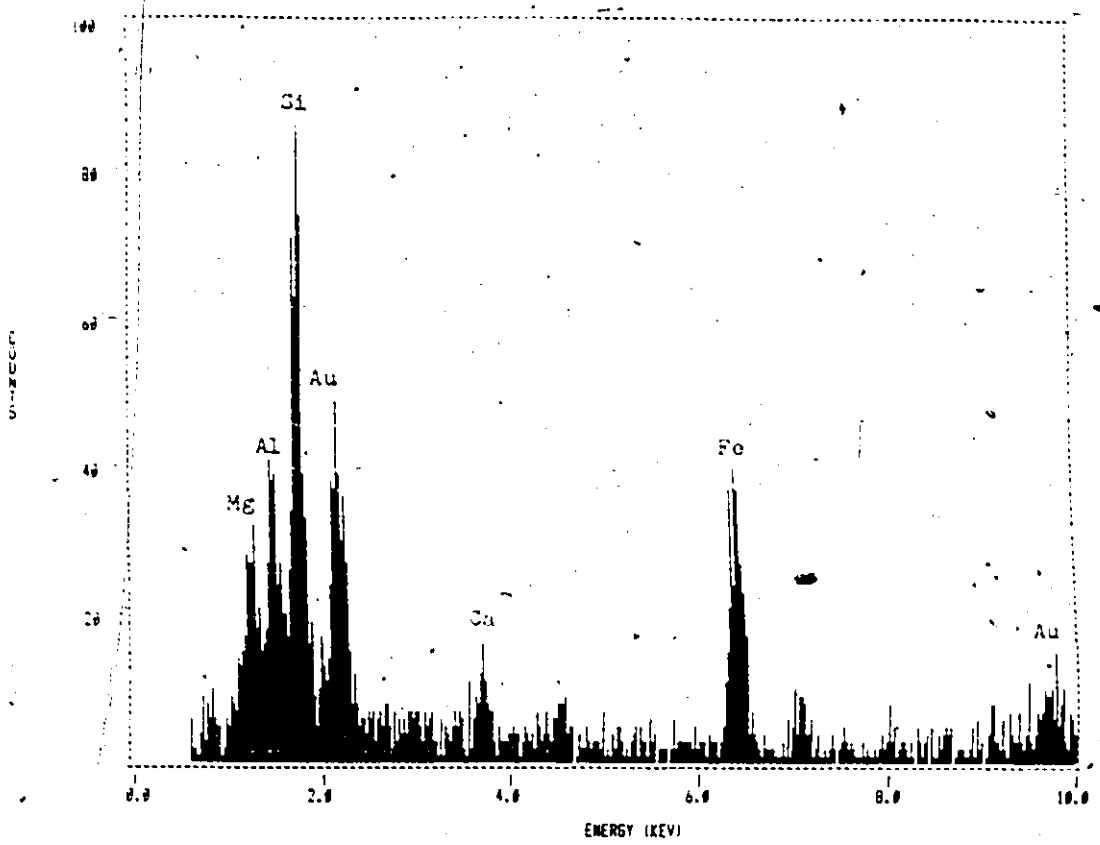

Si
Au
Al
Mg
Fe
Ca
Au
0.0
2.0
4.0
6.0
8.0
10.0
ENERGY (KEV)

Figure 6.29 Vermicular chlorite in feldspar (specimen IT2). Plane polarised light (top); cross polarised light (bottom). The maximum width of the field of view is approximately 0.25 mm.

concentrated near the edge of the grain, and may be a replacement of the feldspar.

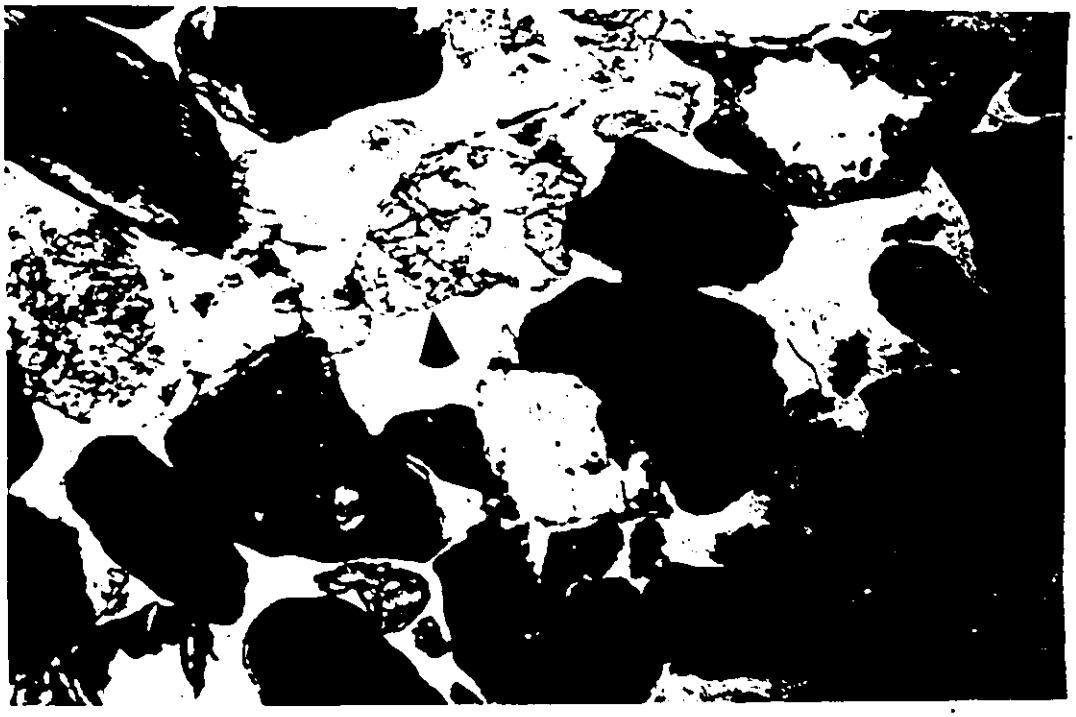
Epidote cement only occurs in IT4, where it is abundant. It both rims and replaces grains, in both cases apparently preceding chlorite growth. Figure 6.30 shows epidote rimming opaques, and replacing a completely dissolved grain, the rest of which is replaced by chlorite.

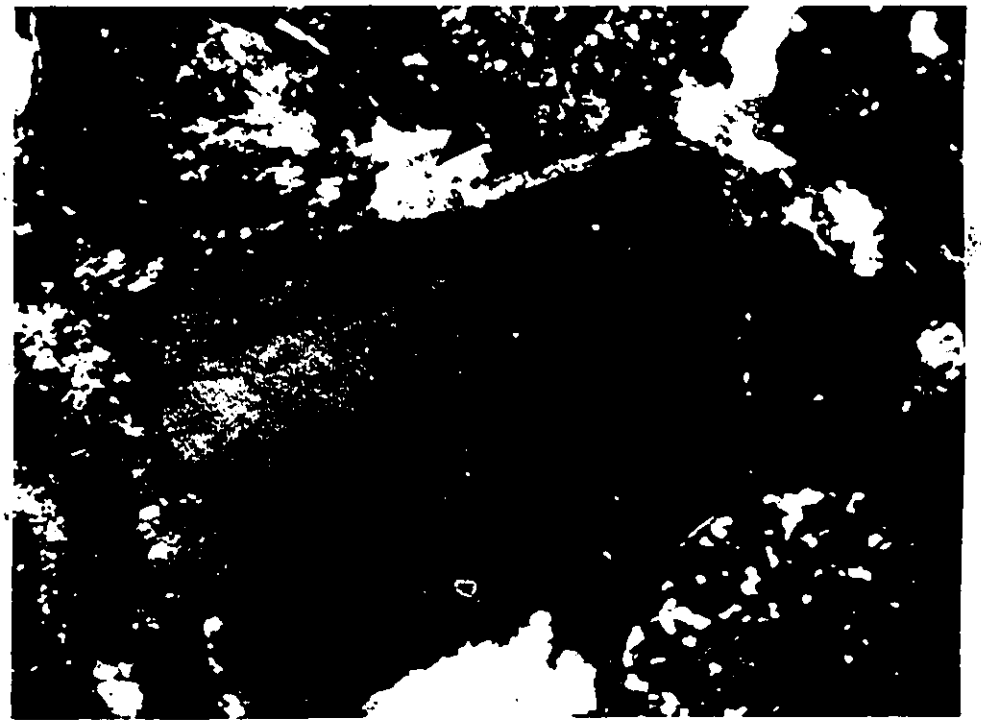
6.2.4.4 Feldspar and quartz overgrowth

Feldspar overgrowths form a significant proportion of the cement in IT2 and IT15 but are rare in IT4. Quartz overgrowths are rare in all specimens, reflecting the scarcity of quartz grains. Feldspar overgrowths postdate the chlorite rims, sometimes filling pores lined by chlorite. Figure 6.31 shows a potassium feldspar grain with overgrowths. As mentioned in the comments regarding Table 6.3, sericitisation of both feldspar grains and feldspar overgrowths is considerable, and renders staining unreliable. Some twinned plagioclase grains take up a predominantly greenish stain, supposedly the potassium feldspar stain. Under these circumstances, it is not possible to determine whether patchily stained feldspars, displaying both pink and greenish regions, are potassium

Figure 6.30 Epidote and chlorite replacement of grains in IT4. The arrow marks the boundary of a replaced grain. Plane polarised light (the strong green colour of the epidote in this photograph is not accurate). The maximum width of the field of view is approximately 1.6 mm.

Figure 6.31 A feldspar grain with overgrowths (arrowed) in IT2. Cross polarised light. The maximum width of the field of view is approximately 0.63 mm.





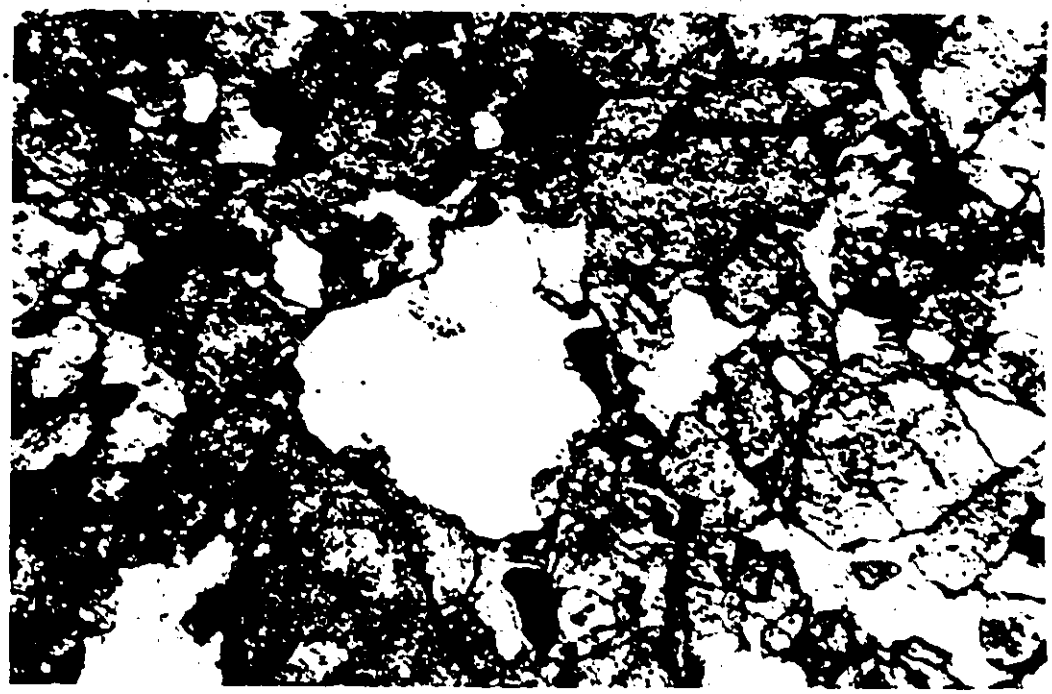
feldspars undergoing albitisation, poorly stained plagioclase or potassium feldspar grains, or perthites. Feldspar cements also show both stains, and similar problems apply. However, from observation of unsericitised feldspars (as in Figure 6.31) before and after staining, it is concluded that both potassium feldspar and plagioclase (probably albite) occur as overgrowths. The lack of quartz and feldspar overgrowth in IT4 probably results from early filling of pores by epidote and chlorite, leaving little remaining pore space to be cemented.

Under the luminoscope, many feldspars show only dull luminescence (Figures 6.32, 6.33), probably the result of alteration. Figure 6.32 shows that relatively clear feldspars do luminesce more brightly. Overgrowths tend to show only dull luminescence (Figure 6.33), probably because these also are altered.

6.2.4.5 Calcite cementation

The last cementing agent was calcite. This filled the last remaining pore space, and replaced grains and feldspar cement (Figure 6.16). In IT2, calcite also occurs in cross cutting veins. An x-ray spectrum taken from the cement, and the bright cathodoluminescence (Figures 6.33, 6.34),

Figure 6.32 Specimen IT2 under plane polarised light (top) and showing cathodoluminescence (bottom). Note the difference in luminescence between sericitised and clear feldspars. The maximum width of the field of view is 1.6 mm.



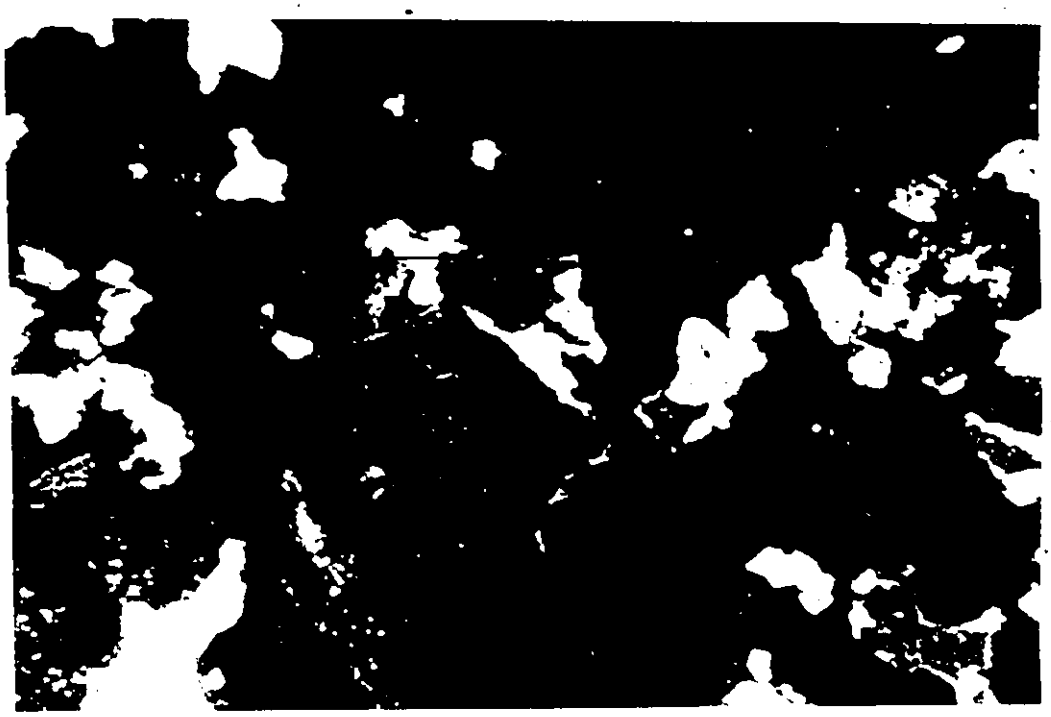
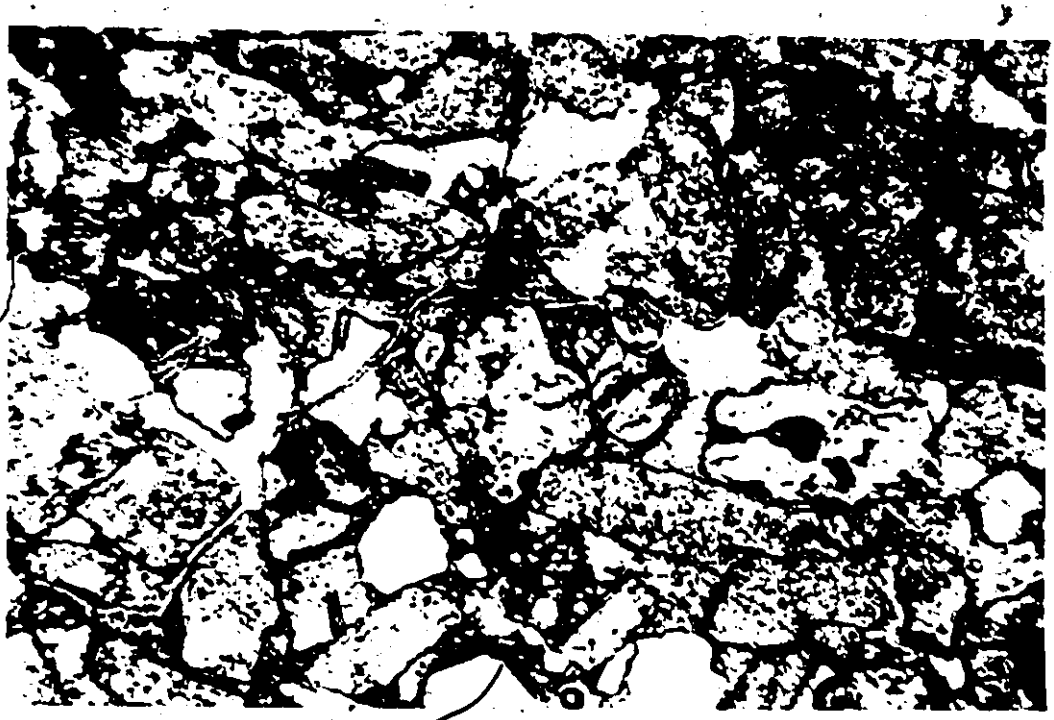


Figure 6.33 Specimen IT2 under plane polarised light (top) and showing cathodoluminescence (bottom). A feldspar overgrowth is arrowed. The yellow luminescence is due to calcite cement. The bluish colouration of some of the feldspars in the upper picture is strong cathodoluminescence not masked by the plane polarised light. The maximum width of the field of view is 1.6 mm.



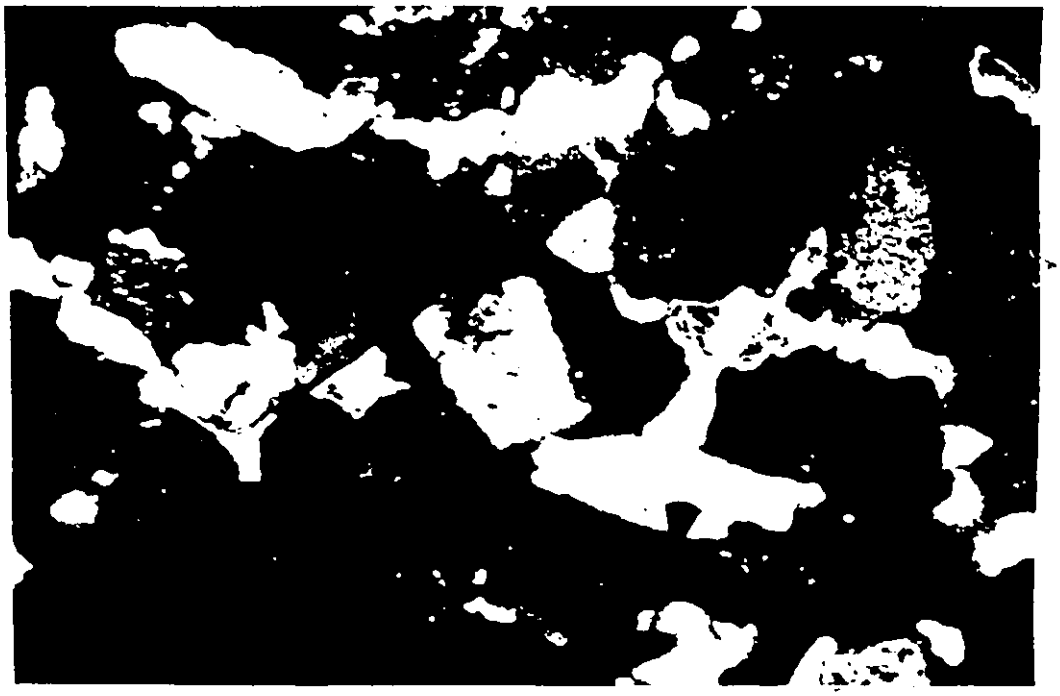
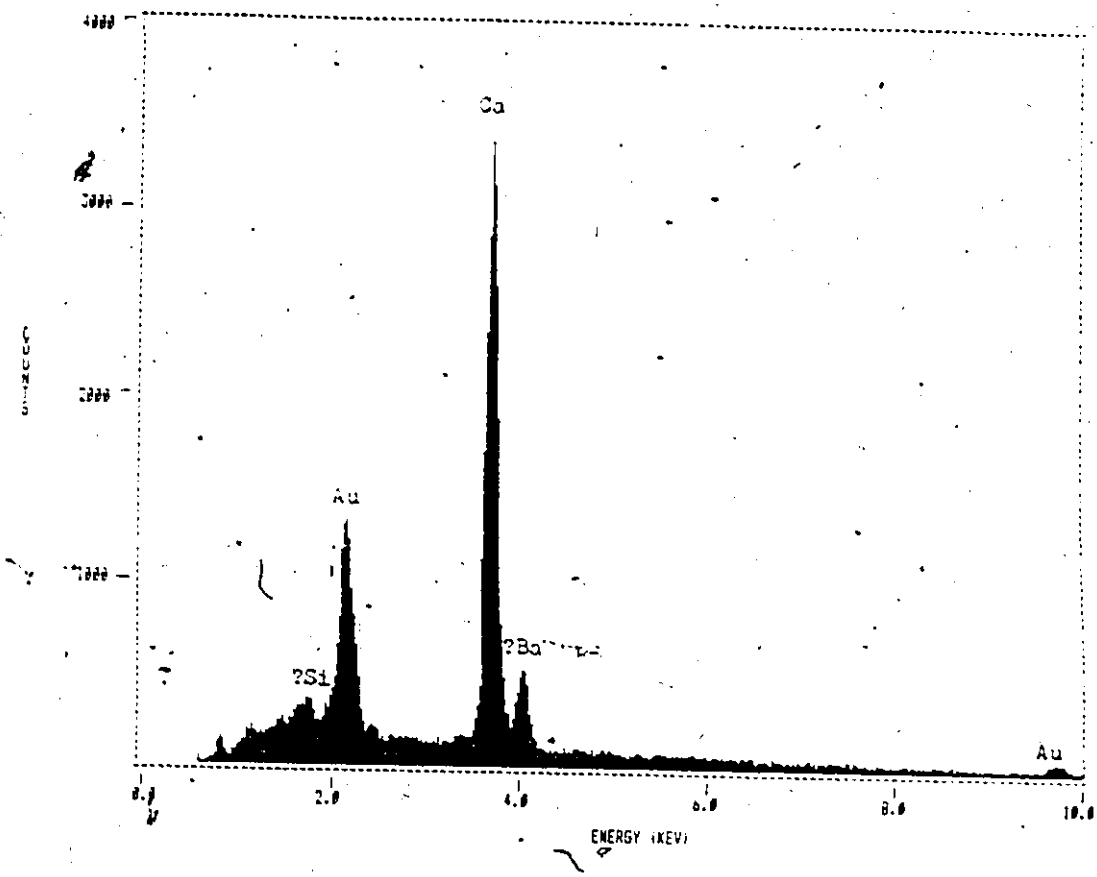


Figure 6.34 An x-ray spectrum taken from the calcite cement in IT2. No iron peak is observed.


Ca
Au
?Si
?Ba
Au
ENERGY (KEV)

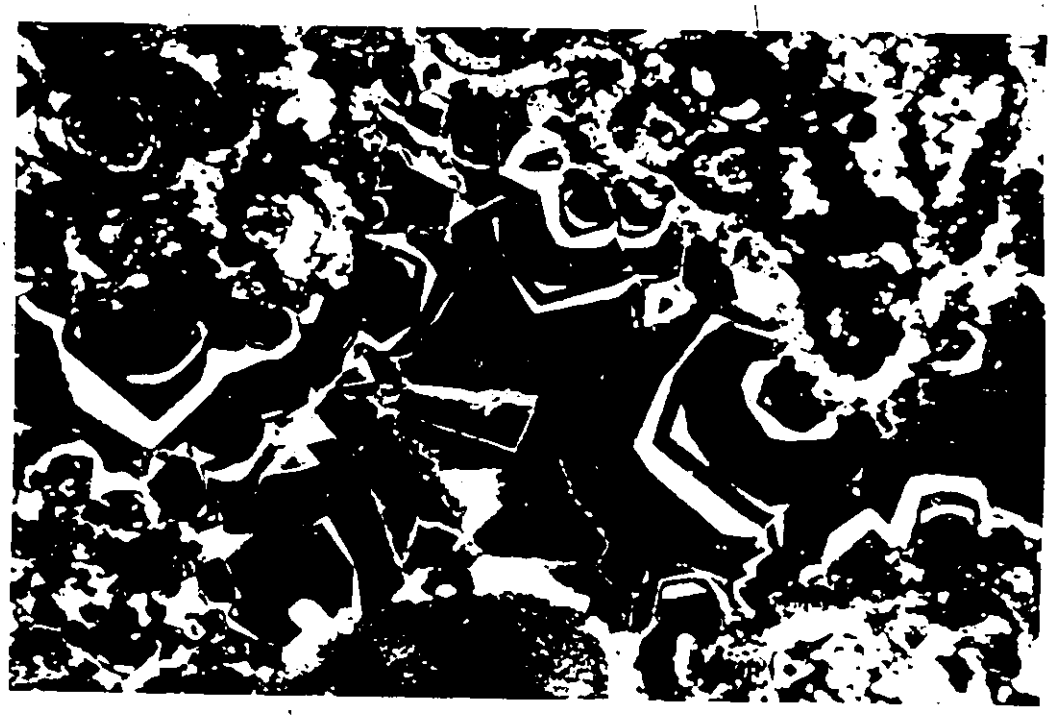
indicate that it contains very little iron, somewhat surprising in such an iron-rich rock.

6.2.5 Diagenetic histories of the oncolithic sandstones (IT5 and IT6)

The only cement present in these samples is calcite. Grains have iron stained rims as in IT2 and IT4.

IT5 shows many signs of compaction, with a large number of sutured and long grain contacts, and many crushed and broken volcanic rock fragments. It therefore appears that cementation was late, postdating much of the compaction. Oversize grains and "floating" grains are common, indicating that there was an episode of dissolution after some cementation. These pores are filled with sparry calcite displaying a striking stratigraphy when observed using a luminoscope. As Figure 6.35 shows, dull luminescence alternates with bright luminescence, defining 5 dull and 5 bright zones, the first consisting of brightly luminescent blocky grains with a fairly even (phreatic style) distribution. All of the calcite takes a bright pink stain, indicating that it is relatively low in iron. Nevertheless, the cathodoluminescence zonation may well reflect variations in iron levels, since very little iron causes considerable

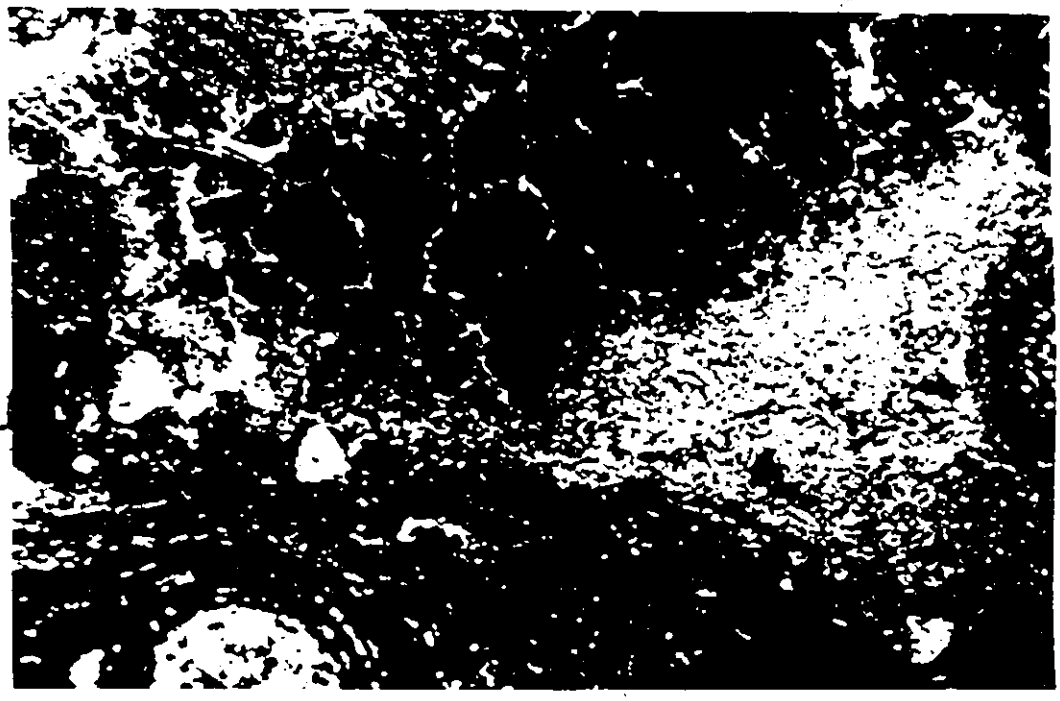
Figure 6.35 Specimen IT5 under plane polarised light (top) and showing cathodoluminescence of the calcite cement (bottom). The maximum width of the field of view is approximately 1.6 mm.



inhibition of luminescence (Amieux, 1982). If the zonation were entirely due to alternations in concentration of the activator (probably Mn^{2+} in view of the reviews of Amieux, 1982 and Nickle, 1978) about the threshold level, the dull zones might be expected to show a dull blue intrinsic luminescence, or no luminescence at all, rather than the brownish luminescence observed. The variation in iron levels probably reflects an alternation of relatively reducing and relatively oxidising conditions, the dark zones corresponding to cementation during reducing conditions, when (soluble) Fe^{2+} is available. There are, however, no "positive" or "negative" sequences (Amieux, 1982) reflecting any progressive change from oxidising to reducing conditions or vice versa.

In comparison with IT5, IT6 shows very few signs of compaction. An open structure has been retained, but oversized pores of the type seen in IT5 are not present. The larger pores are partly filled with geopetal accumulations of silicate silt and carbonate silt (micrite) grains which sifted down between sand grains after deposition. The geopetal micrite is well observed under cathodoluminescence (Figure 6.36). Following the same reasoning as for IT5, it appears that the carbonate silt was relatively non-ferroan, whereas the later sparry cement which plugs the pores was

Figure 6.36 Specimen IT6 under plane polarised light (top) and showing cathodoluminescence (bottom). Note geopetal silt (arrowed). The maximum width of the field of view is approximately 1.6 mm.



more ferroan. This is the situation that might be expected for a progressively more closed (increasingly reducing) system, where burial causes oxygen levels in the sandstone porewater to gradually diminish, so that Fe^{2+} becomes stable and is present in solution when the calcite cement is precipitating. As in IT5, all the calcite stains bright pink, indicating that iron levels are generally low.

6.2.6 General conclusions regarding diagenesis

It is evident that diagenesis of the Copper Harbor Formation must be considered, at least initially, on a local rather than on a regional scale. The diagenetic history is strongly influenced by the rock composition, which varies considerably within the Copper Harbor Formation. IT4, which contains a high proportion of opaques and volcanics, has large amounts of epidote and chlorite cement. Because relatively early chlorite fills most of the pores in IT4, calcite and feldspar cements are rare, and compaction has been a more important factor than in IT2 and IT15, where the feldspar grains are abundant and often show overgrowths which have resisted compaction. The oncolithic sandstone samples, although only 5km apart geographically and close stratigraphically, also show many differences in their

compaction and cementation histories. IT5 was apparently cemented later than IT6, and shows a more complex cement stratigraphy than IT6. The compositional differences between IT5 and IT6 do not appear sufficient to explain these diagenetic differences, and the presence of oncoliths in both samples would tend to indicate that the depositional environments of the two samples were not very dissimilar. The differences therefore result from different groundwater conditions after deposition, involving other factors in addition to the compositions of the specimens. The specimen compositions are important, however, as is seen from the considerable differences in diagenetic style between the non-calcareous and oncolithic sandstones.

The red stain probably results from alteration of iron bearing minerals and basic volcanic clasts (as argued by Walker et al., 1981; Walker, 1976). There is an abundance of such material in these rocks, and most appears to have altered (e.g. Figure 6.37). Elmore (1981) also concluded that much of the iron stain is diagenetic in origin.

The exclusive occurrence of non-ferroan calcite rather than ferroan calcite is not what might be predicted from the sample compositions, which are rich in iron. The x-ray spectra indicate that much of the chlorite is also rather iron-poor. The most probable explanation is that iron

Figure 6.37 Magnetite grain showing irregular boundaries and dissolution along certain crystallographic planes, now filled with calcite (specimen IT4, plane polarised light).

levels in the groundwater were low because the porewaters were never strongly reducing, so Fe^{2+} was not available to be taken into solution. The environment of deposition may have been the major factor in producing these conditions. In Proterozoic times plants were absent so fan surfaces and sediments were largely free of organic matter, which might otherwise have acted as a reducing agent. The alluvial fans contain virtually no shale, so were probably highly permeable. Topographic relief was high at the rift margins, where the fans developed. These factors would have favoured movement of relatively oxidising meteoric water to large depths in the sedimentary pile. These groundwaters could have carried calcium and carbonate for early calcite cementation, produced by weathering of the hinterland (which consisted of felsic and basic volcanics). In the case of IT6, the presence of stromatolites and oncoliths indicates that a distal lake was present at the time of deposition. Gypsum casts (Elmore, 1981) and abundant desiccation features indicate that some such lakes were hypersaline, so at the time of deposition of IT6 the products of weathering were probably very concentrated (perhaps partly by a fairly arid climate), which would explain the very early cementation of this sample. The calcite cement of IT5, although apparently later than in IT6, could have a similar

origin. The other samples, however, have been cemented by calcite late in diagenesis, and the carbonate might have been provided by other means. The Nonesuch Formation, which overlies the Copper Harbor Formation, is an organic rich, somewhat calcareous shale, but is much thinner than the Copper Harbor Formation (Figure 2) and cannot be appealed to as a source of calcite to cement a significant part of the Copper Harbor Formation. Calcareous sandstones, such as IT5 and IT6 are themselves cemented solid by calcite, and do not help explain calcite cement elsewhere. Albitization of plagioclase releases calcium ions, but as mentioned earlier, the feldspar stains are not sufficiently clear to estimate the extent of albitization. Boles and Franks (1979) have pointed out that the smectite-illite conversion releases various ions, including calcium. In the Copper Harbor Formation, smectite may have formed early in diagenesis as a result of alteration of volcanic fragments, and have later reacted to form chlorite. This reaction may or may not release calcium, depending upon the composition of the smectite.

Clay cementation is important in these rocks. Chlorite is the dominant clay and precipitated early in diagenesis, probably at the same time as the opaques and basic volcanic rock fragments were dissolving and altering.

Sericitic alteration of feldspars and feldspar cement is also important, but apparently occurred at a much later stage in the diagenetic history.

Although feldspars are altered to sericite and some show signs of dissolution and replacement by carbonate, there does not in general appear to be sufficient feldspar dissolution to account for all of the feldspar cement observed. This is, however, difficult to estimate, partly because sericitisation is so extensive and obscures feldspar features. If some other rock was acting as a source for the feldspar cement, then it ought to be possible to somewhere find examples of sandstones where feldspar dissolution is more extensive. It is not, however necessary to search for another rock to provide a source for the feldspar cement. A more likely source may be the volcanic material within IT2, much of which appears to have dissolved or altered. A wide range of volcanic types are present, and these fragments could provide all the silica, sodium and potassium required to precipitate the feldspar cement. Klein (1939) has documented the occurrence of authigenic microcline in amygdales in the Portage Lake Lava Series. The feldspar cement in the Copper Harbor Formation was noted by Elmore (1981), but has generally received little attention, and is not mentioned by Daniels (1982) or Wolff and Huber (1973).

CHAPTER 7

COMPARISON OF THE KEWEENAWAN SEDIMENTS WITH MODERN AND ANCIENT ALLUVIAL FAN AND BRAIDED RIVER SEDIMENTS

The Middle and Upper Keweenawan sediments of the Lake Superior Basin are compared in this Chapter with sediments characteristic of three types of environment: 1) arid-region alluvial fans; 2) humid-region alluvial fans; 3) braided fluvial systems. In each case both similarities and dissimilarities will be apparent. It is not implied that the three types of environment mentioned above do not to some extent overlap, or that intergradational environments are not of relevance, but the discussion is initially set up in this way because these environments have each been the focus of studies by previous workers. The merits and demerits of these environmental divisions are discussed at a later stage.

7.1 Comparisons with arid-region alluvial fans

The present-day alluvial fans in the semi-arid regions of the southwestern United States are the most intensely studied fans in the world (e.g. Blissenbach, 1954;

Bull, 1963 and 1964; Denny, 1967; Beaty, 1970; Hooke, 1967, 1968 and 1972; Hooke and Rohrer, 1979). Bull (1972; 1977) reviewed the observations and the ideas resulting from work on these fans. The reviews of Nilsen (1982) and Rust and Koster (1984) also drew heavily upon this work. The resulting model for (arid-region) alluvial fan sedimentation is a small steep fan, with a downfan facies progression (according to Rust and Koster, 1984) from: 1) imbricated, horizontally-stratified, clast-supported conglomerates (fluvial deposits) interbedded with massive, mud matrix-supported conglomerates (debris flow deposits), to: 2) planar cross-stratified fine conglomerates or pebbly sandstones interbedded with thin horizontally-stratified sandstones (all fluvial deposits), to: 3) laminated or massive muds (distal floodplain deposits). An arid-region alluvial fan model in the form of a vertical section was constructed by Miall (1978), who based his model on the Trollheim Fan of Deep Springs Valley, California, as described by Hooke (1967). 66% of this vertical profile model consists of debris flow deposits. Miall (1978) did suggest that there can be a downfan variation from this "Trollheim profile" to a "Scott profile". Such a downfan facies change would create a fan model similar to that envisaged by Rust and Koster (1984) and outlined above.

Miall (1978) did not, however, imply that this style of variation is a <u>necessary</u> feature of an arid-region alluvial fan model.

According to Nilsen (1982) an average (arid-region) alluvial fan slope is 5 degrees (0.087) and an average radius is less than 10 km. The Trollheim fan, the basis for the model produced by Miall (1978) is 2 km in length, and has four segments, with gradients between 14° (0.255) and 6° (0.105).

In comparing fan models, such as the arid-region alluvial fan model summarized above, with ancient sediments like the Keweenawan conglomerates and sandstones, several questions arise. Firstly, is the aspect of the sediments similar, or at least, similar enough for application of the model to be relevant? Secondly, if this is the case, can it be shown that the depositional processes, fan slopes and fan sizes were similar? Thirdly, if both these questions can be answered affirmatively, are the climatic and tectonic implications of the model of relevance to the ancient example?

The aspect of the Keweenawan sediments is markedly different to that of the sediments in the arid-region alluvial fan model summarized above. Firstly, debris flow deposits are absent. Secondly, much more sandstone is

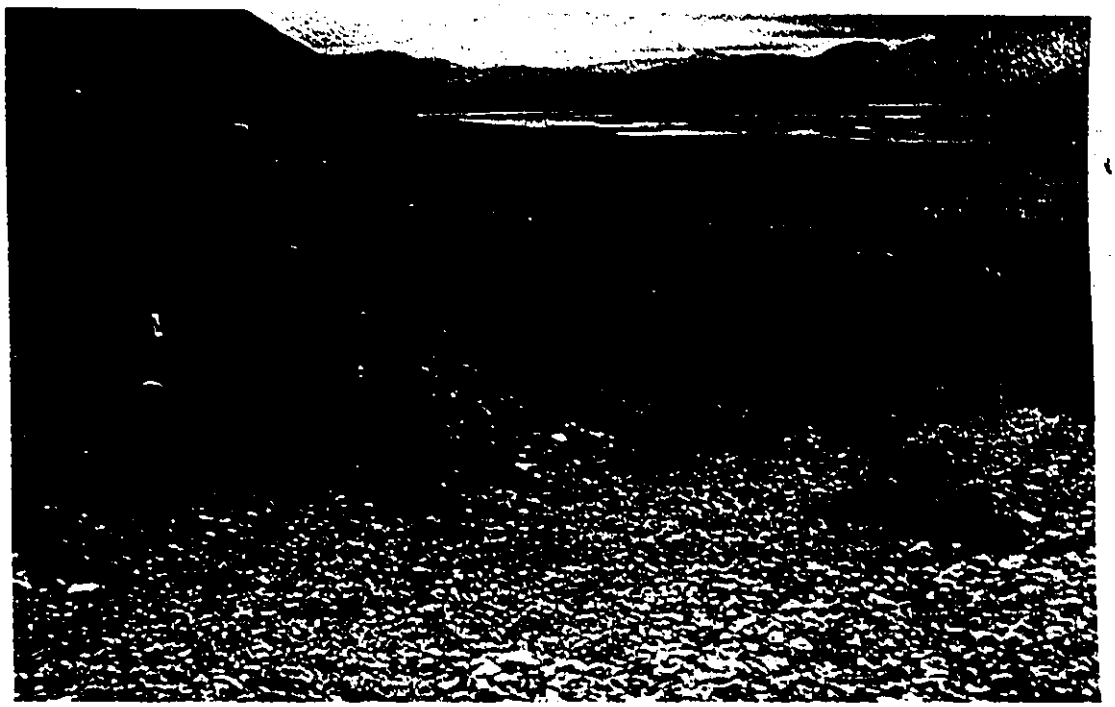
present in the Keweenawan sequence than is implied by the model, and furthermore, the sandstone is not of the facies described in the model. Particularly, various types of trough cross-stratified sandstone, which the model implies are absent or very minor, are abundant in the Keweenawan. The Keweenawan sediments in addition do not conform to the supposedly diagnostic criterion of Bull (1977) that sediment bodies within alluvial fan sequences should display predominantly sheet-like geometries. The surveyed sections of Chapter 4 show that channellisation is prevalent in the Keweenawan sediments.

These differences are sufficient to disqualify the model based on arid-region alluvial fans as an adequate analogue for the Keweenawan sediments. The implication is that Keweenawan fan processes, slopes and sizes were different from those of the model (these points are further discussed below and in section 7.2). This does not mean, however, that arid-region alluvial fans do not provide some useful points of comparison.

For example, Figures 7.1, 7.2 and 7.3 show some typical features of the fluvial deposits in Death Valley fans. Of especial interest is the relative lack of sorting in these deposits. Sand and gravel have not been well separated by the processes responsible for transport or

Figure 7.1 Sediments exposed in the walls of the entrenched fanhead channel on Badwater Fan, Death Valley. The photograph was taken looking south. Notebook is 18 cm in length. See also Figures 7.2 and 7.3. Photographed by G.V. Middleton.

Figure 7.2 Close up of the sediments shown in Figure 7.3. Debris flow deposits are not evident (although it is possible that debris flows transported the large boulders in Figure 7.1 to their present positions). Sand is present within gravel beds but sandy lenses of the type seen in the Keweenawan sediments are not present. This type of appearance is typical of fluvial sediments in Death Valley. Notebook is 18 cm in length. See also Figures 7.1 and 7.3.



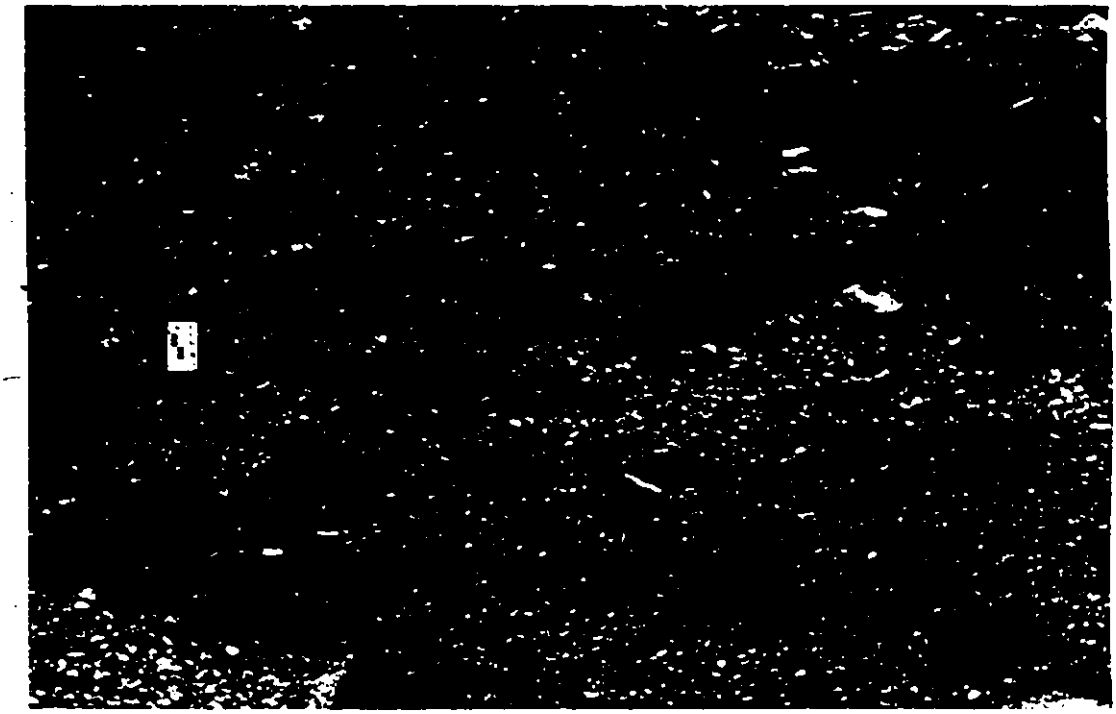
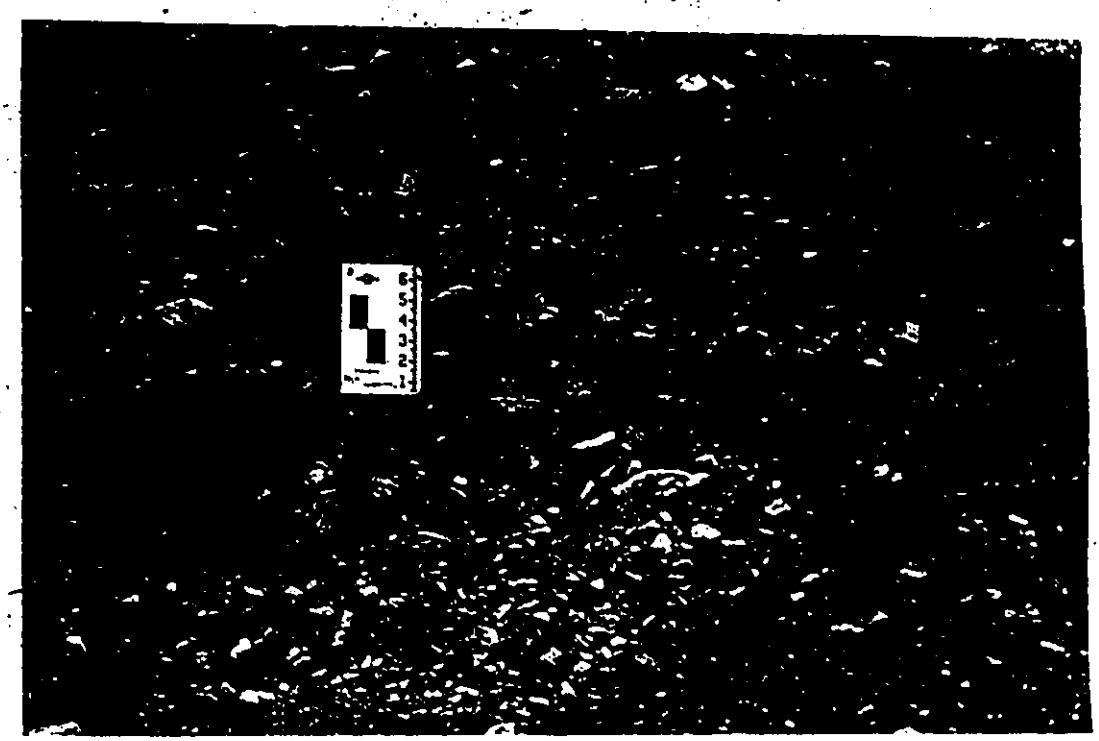


Figure 7.3 Close up of the sediments shown in Figures 7.1 and 7.2. Notebook is 18 cm long.



deposition. Sand lenses are not present, and the gravel beds contain much sand as well as pebbles, cobbles and boulders. On many of the Death Valley fans, the fluvial deposits present this kind of appearance for much of the length of the fan (G.V. Middleton, personal communications; my own observations from a one week field trip to Death Valley). This lack of sorting is in some respects similar to the Keweenawan conglomeratic facies association. However, the Keweenawan sediments do show a degree of sorting not shown by the Death Valley gravels, in that sand has been segregated to form the sandstone lenses that occur in all the conglomeratic outcrops.

A number of ancient alluvial fan successions are a fairly convincing fit to the arid-region alluvial fan model, as outlined above, but in addition show features that are not present in the model. Some of these features are also displayed by the Keweenawan sediments of this study.

Steel and Wilson (1975) (also Steel, 1974), and Mack and Rasmussen (1984) studied alluvial fan (red-bed) sediments of approximately Permian to Triassic age. Both studies presented evidence (such as caliche, mudcracks) that the palaeoclimate was mainly arid and that streams were ephemeral, and both alluvial fan successions contain a significant component of debris-flow deposits. In both

cases, in contrast to the model, trough cross-bedded sandstones were found to be an important part of the alluvial fan deposits. Steel and Wilson (1975, p.186) noted an "abundance of trough cross-stratification in sandstones and conglomerates". Mack and Rasmussen (1984, p.113) concluded that "the abundance of trough cross-bedded braided-stream facies indicates that transportation of sand and gravel as dune bed forms was a common process on the Cutler alluvial fan". These authors went on to suggest that the Cutler alluvial fan may therefore have been larger, and more gently sloping than the modern fans studied in the southwestern United States.

Other Triassic red-beds were studied by Nadon (1981), and Nadon and Middleton (1985). Here too calcite nodules and mudcracks indicate that depostion was by ephemeral streams in a semi-arid environment, although the climate was apparently sometimes wetter. As in the studies already mentioned, trough cross-stratified sandstones are prominent in the sedimentary succession, and Nadon and Middleton (1985, p.1191) suggested that "the fans were probably large, with relatively low gradients distally, permitting the formation of the abundant cross-bedding by lower regime flows". These particular sediments bear a further similarity

to the Keweenawan sediments of the Lake Superior Basin in that debris-flow deposits appear to be absent.

The trough cross-stratified gravel noted by Mack and Rasmussen (1984) and Steel and Wilson (1975) is another feature that these examples share with the Keweenawan sediments. Bluck (1965), again in a study of Triassic red-beds, has also recorded trough cross-bedded conglomerates. Here the trough cross-bedded conglomerates fill channels in the alluvial fan sequence and show trough axes directed approximately down the channel axes. In this example, and in contrast to the arid-region alluvial fan model, trough cross-bedding was found to be more abundant than planar cross-bedding.

To summarize, the Keweenawan sediments of the Lake Superior Basin fit very poorly into the arid-region alluvial fan model. However, it is also apparent that a number of the ancient examples of alluvial fans that clearly formed in arid or semi-arid climates, and would therefore be expected to most closely fit this model, do not entirely fit either. The misfit of these examples might be improved if the arid-region alluvial fan model is modified to account for the possible occurrence of larger, flatter fans.

7.2 Comparisons with humid-region alluvial fans

In view of the limitations of the arid-region alluvial fan model, Boothroyd and Nummedal (1978) and Boothroyd and Ashley (1974) documented the features of glacial outwash fans, with the aim of providing a humid-region alluvial fan model. This model has not gained universal acceptance. Nilsen (1982) and Rust and Koster (1984) preferred not to regard these outwash fans as fans, or even as relevant to fans. Galloway and Hobday (1983), however, in another review of alluvial fan deposits, treated this humid-region fan model on an equal basis with the arid-region fan model. In the remarks that follow it will be shown that this humid-region fan model is of some relevance to a number of ancient examples that have been interpreted as alluvial fan deposits, including the Keweenawan sediments of the Lake Superior Basin. Like the arid-region alluvial fan model, however, it is found lacking in certain respects.

The fans studied by Boothroyd and Nummedal (1978) are between 3 km and 30 km in length with upper fan gradients averaging 0.006 - 0.017 and lower fan gradients averaging 0.002 - 0.003. These fans are therefore

considerably larger and flatter than the alluvial fans that have been studied in semi-arid regions.

In the humid-region alluvial fan model, as presented by Boothroyd and Nummedal (1978), deposition is entirely by fluvial processes. Debris flows are not present. The fan surface aggrades as a result of braided stream sedimentation. The model is divided into upper, middle and lower fan segments primarily on the basis of grain size variations, although average slopes also differ between segments. On the upper fan, coarse gravels are predominant, and the largest clasts are more than 10 cm long. Gravels on the midfan contain clasts less than 10 cm long. On the lower fan the sediment consists entirely of sand. On the upper fan the streams contain low longitudinal bars lacking slip-faces, which give rise to deposits of well imbricated, crudely bedded, coarse gravels that otherwise lack structure. Sandy deposits are rare in the upper fan, but some thin interbeds of flat-bedded sand are present. On the mid-fan, longitudinal bar types still dominate, but tend to have higher relief and show more slip-faces, generally developed in sand. The midfan deposits therefore consist of gravels similar to those of the upper fan but finer, interbedded with a roughly equal amount of sand showing approximately planar cross-stratification, or, as on the

upper fan, showing flat-bed lamination. On the sandy lower fan both longitudinal and linguoid bars are present, and deposits consist of approximately planar cross-stratified sand interbedded with a roughly equal amount of current rippled sand and flat-bedded sand (deposited on bar surfaces). Other sediments on glacial outwash fans were described by Boothroyd and Nummedal (1978) and Boothroyd and Ashley (1975), but the downfan facies sequence outlined above is the essence of their humid-region alluvial fan model.

The absence of debris-flow deposits and the exclusive occurrence of braided stream deposits make this humid-region alluvial fan model a better analogue for the Keweenawan sediments than the arid-region alluvial fan model. The upper fan deposits of the model most closely compare with the conglomeratic facies association of the Keweenawan rocks. However, whereas imbrication and clast-support are ubiquitous in the conglomerates of the model, in the Keweenawan conglomeratic facies association imbricated beds are the exception and massive, c-m supported, coarse conglomerates are predominant. This probably indicates that deposition was faster on the Keweenawan fans, with less opportunity for sorting to occur, probably reflecting higher rates of sediment supply and more rapid waning of stream

flow. Boothroyd and Ashley (1975) did note, however, that the low relief bars on the upper fan (with thicknesses not much greater than the size of the largest clasts) formed by accretion of thin sheets of large clasts that could only be moved by "rare, very high flood flows". Consideration of Figure 4.5 indicates that this style of low relief bar without slipface was also the commonest form on proximal parts of the Keweenawan fans.

Sandstone lenses within the Keweenawan conglomeratic facies association seem to be more common than sand interbeds are in the upper fan of the model. In the model the sand interbeds are dominantly flat-bedded. Boothroyd and Ashley (1975) noted that sandy dunes do form in channels on the upper fan during falling stages of flow, but these appear to have a very poor preservation potential. In the Keweenawan, sandstone lenses may be flat-bedded or trough cross-bedded, but in many cases show a poorly developed low angle cross-stratification. This low angle cross-stratification may indicate that dunes were beginning to form but time or flow depths were insufficient, or might be due to antidune deposition under supercritical flow conditions. The latter explanation might fit well with the depositional conditions indicated by the interbedded conglomerates, since high sediment concentrations and

rapidly waning flows would favour preservation of antidune bedforms. It is otherwise difficult to explain why conglomerates deposited by flows more energetic than those of the humid-region fan model should be interbedded with more, rather than less sandstone than in the model (in general highly energetic conditions cause sand to bypass a particular area, but sand may be deposited in such conditions if sediment concentrations are excessively high).

The middle and lower fan deposits of the model may be compared respectively with the conglomerate-sandstone facies association and the sandy facies association of the Keweenawan sediments. However, there is an obvious contrast between the sedimentary structures in the model and those in the Keweenawan. Where the model would predict mainly planar cross-stratified sandstones, the Keweenawan sediments contain abundant trough cross-stratified sandstones. Boothroyd and Ashley (1975, p.214) recorded the presence of dune bedforms but found that these failed to leave trough cross-bedding as evidence of their existence: "trough cross-beds, the result of megaripple migration, are extremely uncommon in sandy longitudinal and linguoid bar deposits, although these bedforms are readily apparent on active bar surfaces observed during low and rising flow stages" (it is possible, however, that below the lowest channels on the

lower fan, where these authors were unable to trench due to thixotropic sediment, there is more trough cross-bedding). On one of the fans studied by these authors (the Scott Fan) the most distal portion of the fan drainage takes the form of a meandering stream, and there preservation of trough cross-beds is much more common.

These discrepancies between the humid-region fan model and the Keweenawan sediments necessitate further discussion of fan slopes and sizes. The palaeoslopes of the Keweenawan fans may be roughly estimated by consideration of the sedimentary structures present and outcrop information regarding channel geometries. Trough cross-stratified sandstones and flat-bedded sandstones with parting lineation are intimately associated at many outcrops, indicating that flow parameters were often close to the values at which a sandy bed undergoes a dune - upper flat bed transition. For example, at Dan's Point, on the Keweenaw Peninsula, trough sets are about 30 cm thick and lateral correlation (Figure 4.12) indicates that the sandstone was deposited by sheet-like flow. Palaeodepths were therefore probably no greater than 1 m, so data from flume studies may be applied to gain a first approximation to the palaeoslope (it must be noted, however, that because the palaeodepth constraints are poor the the analysis is necessarily incomplete). The flume

study of Simons et al. (1961) used water depths averaging 0.16 m and used a medium sand bed (similar grainsize to that of the Dan's Point sandstones). For dunes at the transition between dunes and upper flat bed, Simons et al. (1961) measured a dimensionless Chezy coefficient ($c/\sqrt{g}$) of approximately 12 and a Froude number (Fr) of about 0.7. If $\bar{u}$ is the mean stream velocity, d is the stream depth, s is the palaeoslope, and g is the acceleration due to gravity, then:

$$\bar{u} = (c/\sqrt{g})\sqrt{(gds)}$$

$$\text{and,} \quad Fr = \frac{\bar{u}}{\sqrt{(gd)}}$$

$$\text{so,} \quad Fr = (c/\sqrt{g})\sqrt{s}$$

giving an estimated palaeoslope,

$$s = [Fr/(c/\sqrt{g})]^2 = (0.7/12)^2 = 0.0034$$

This gradient, derived using flow parameters measured in a flume with water depths averaging only 0.16 m, probably represents a maximum palaeoslope. If water depths on the Keweenawan fans were greater, then the flat-bed to dune transition would have occurred at lower slopes. The Rio Grande is a river about 0.60 m deep with a medium sand bed and has been used as a natural flume. Culbertson and Dawdy

(1964) measured Froude numbers and Chezy coefficients in various reaches for conditions transitional between upper flow regime (flat-bed) and lower flow regime (dunes). If these values are used to estimate a possible palaeoslope by the method used above, then an average value of 0.0009 is obtained. The size of the trough cross sets at Dan's Point (30 cm) indicates that palaeodepths may have been closer to 0.60 m than 0.16 m, in which case, the palaeoslope estimated from the Rio Grande data may be a more suitable value. To give a measure of the uncertainty resulting from the indeterminate palaeodepth, Table 7.1 shows palaeoslopes estimated using the data of Simons et al. (1961) and Culbertson and Dawdy (1964), and also shows the range of slopes measured by these workers for which transitional conditions were recorded (Simons et al., 1961, ascribed much of their scatter to hysteresis, the reluctance of the bed to assume a new configuration, but there are also effects resulting from the fact that real flow is not two dimensional but three dimensional).

This range of possible palaeoslopes is towards the lower end of the range of gradients for modern glacial outwash fans (and is much less than the average gradient of 0.087 given by Nilsen (1982) for arid-region alluvial fans). The normal ranges of slopes for glacial outwash fans, as

Table 7.1 Flow parameters and slopes at the dune - upper flat bed transition, as observed by Simons, Richardsons and Albertson ,1961 (S.R.A.), and Culbertson and Dawdy, 1964 (C.D.).

STUDY	AVERAGE FLOW DEPTH (m)	AVERAGE FROUDE NUMBER, Fr, AT TRANSITION	AVERAGE DIMENSIONLESS CHEZY COEFFICIENT, $c/\sqrt{g}$, AT TRANSITION	MEASURED RANGE OF SLOPES AT TRANSITION	CALCULATED SLOPES AT TRANSITION $[Fr/(c/\sqrt{g})]^2$
S.R.A.	0.16	0.7	12	0.0025 - 0.0035	0.0034
C.D.	0.60	0.43	14.6	0.0006 - 0.0012	0.0009

estimated by Boothroyd and Nummedal (1978) were mentioned earlier, and taking the middle of these ranges gives an average value for proximal slope of 0.0115 and an average value for distal slope of 0.0025. A detailed analysis of slopes on the Scott outwash fan was given by Boothroyd and Ashley (1975). Average values for the different segments were not given in that study, but measurements from the longitudinal profile (their Figure 8) yield slopes of 0.0114 for the upper fan, 0.0038 for the middle fan and 0.0014 for the lower fan, which agree with the ranges of values given in their text. The values for the gradients of the middle and lower fan are close to the palaeoslopes calculated for the Keweenawan fans. Boothroyd and Ashley (1975, p.220) noted that "Froude number calculations for the midfan show most values below 0.60 ..... in the range of megaripples (dunes) to transition forms". Depths on the midfan recorded by these authors range up to about 60 cm. It is of note that these observations and conclusions correspond well to the estimates and conclusions deduced for the Keweenawan (without prior consultation of the data of Boothroyd and Ashley, 1975). Whilst this supports the validity of the approximate palaeoslope calculation, it only complicates the discrepancies between the sedimentary structures in the ancient and those in the modern. To explain the lack of

trough cross-stratification in the glacial outwash fans, Boothroyd and Ashley (1975) tentatively suggested that dunes might act only as a conduit for sediment, contributing sediment to bar slipfaces, but not being an accretionary bedform. Another explanation might be provided by the observation of Boothroyd and Ashley (1975) that antidune water-wave forms are very common on the upper and middle fan areas. This would indicate that even if dunes do form at some flow stages, they are likely to be wiped out at the highest stages of flow. It therefore would seem that to produce trough cross-stratified deposits it is necessary that flow conditions not only often allow dunes to form, but rarely are supercritical. This would necessitate slopes at the low end of the range of palaeoslopes calculated for the Keweenawan.

Other studies of glacial outwash deposits include Gustavson (1974) and Bluck (1974). Both of these studies describe sediments very similar to those of the humid-region alluvial fan model, as summarized above. Bluck (1974) recorded the presence of dunes, but did not comment on their likely preservation potential. However, Bluck (1974) noted that most sand deposition occurs during falling stage, after the main phase of gravel movement. During falling stage dunes may form in some of the higher channels, producing

sandy sediments which, when trenched, show trough cross-strata. The resulting deposits of sand are thin and at relatively high levels in the system, so their preservation potential is low; most will be removed during the next period of peak flow.

It therefore seems that the Keweenawan fans had gradients as low as, and perhaps lower than modern glacial outwash fans, such that trough cross-stratification could form and be preserved. This would imply that the Keweenawan fans were large (i.e. as large as or larger than modern periglacial fans, with radii of 30 km or more). Palaeogeographical considerations tend to support this conclusion. The original rift margin, from which the fans developed, cannot have been nearer to Lake Superior than the present outcrop limit of the basalts immediately underlying the fan sediments. These basalts were erupted within the rift valley and cannot have flowed (uphill) out of the rift valley. The basalts are not cut by major faults running parallel to the rift margin, which would be expected if these basalts had been uplifted by faulting to form a source area for the Keweenawan fans. It is not certain how much faulting there was at the rift margin, but even if uplift occurred by other means, the evidence of previous workers (presented in Chapter 2) indicates that the source for the

Keweenawan sediments was not the basalts immediately underlying, but older basalts and felsic rocks. On the Keweenawan Peninsula these source rocks are only exposed to the south of the Keweenaw Fault. From Union Bay in Upper Michigan where the sandy facies association crops out, the distance to the Keweenawan Fault along the mean palaeocurrent direction (NNE) is 36 km (Figure 2.3). This is a minimum value for the original fan radius. Consideration of Isle Royale (where the sediment source was the North Shore Volcanic Group of Minnesota) gives a minimum fan radius estimate of 27 km (Figure 5.1), and the true value could be much higher.

The problems in determination of preservation potential are a major drawback when studies of modern environments are extrapolated to provide models for analysis of ancient sediments. There is clearly a need to find greater thicknesses of section than can be provided by trenching. Fans in Tasmania, formed during the last glacial, have been described by Wasson (1977a,b). Due to incision there is some good section in these fans. These fans do not, however, fit the humid-region alluvial fan model. The downfan facies succession, which spans a distance of about 1 km only on these small fans, is from a mixture of debris flow deposits and fluvially deposited imbricated gravels to

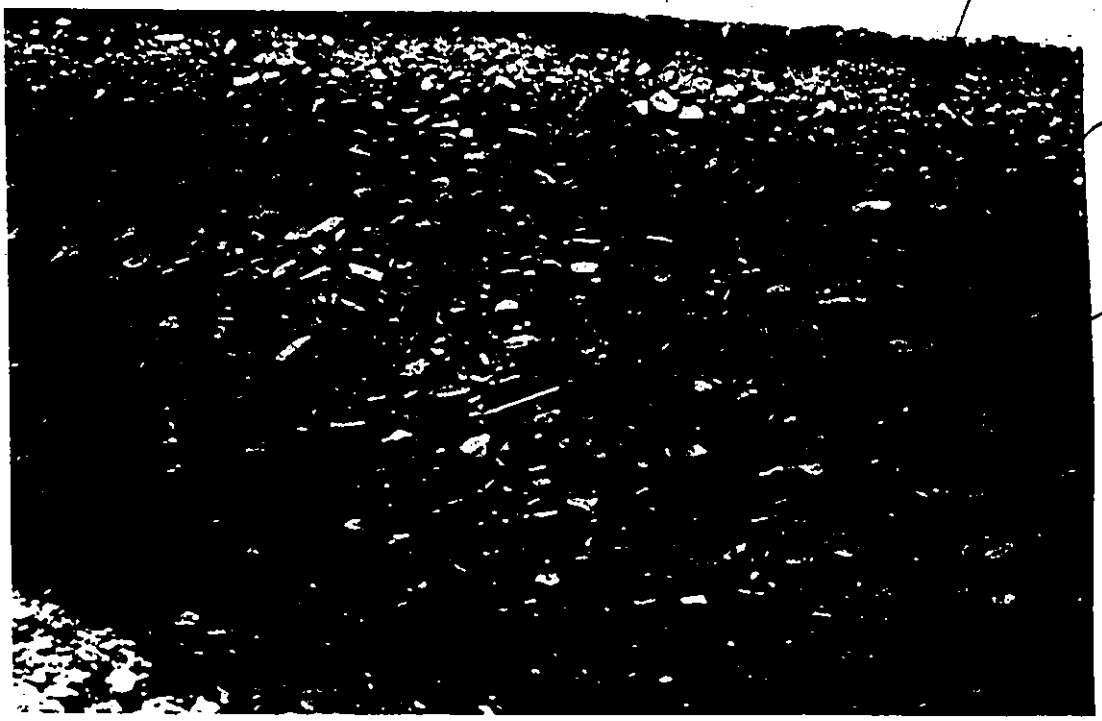
distal sheetflood silts and clays. This is more akin to the arid-region alluvial fan model than the humid-region alluvial fan model based on modern outwash fans. The differences probably result from the fact that the Tasmanian fans were not outwash fans, but formed in a periglacial environment where the glacial climate enhanced mass-wasting processes on the slopes of fan catchments. Fans of a similar age, with similar sediments, that formed in a similar environment in British Columbia have been described by Ryder (1971). Studies like this bring into question the validity of any simple climatic division of alluvial fan types. Although the Tasmanian and British Columbian fans probably experienced periglacial climatic conditions similar to those experienced today by the Alaskan and Icelandic outwash fans (described by Boothroyd and Nummedal, 1978), the supply of sediment and water to the Pleistocene fans was so different to the supply to the modern outwash fans that fan processes in these two different situations bear little resemblance. The question of how alluvial fans respond to climate is pursued further in section 7.4. The Tasmanian fans do, however, provide one valuable point of comparison with the Keweenawan sediments. Wasson (1977b) described a group of imbricated gravels which "lack [the] separation of fine and coarse sediments typical of much alluvium". The fabrics of

these gravels vary between clast-supported and matrix-supported, according to Wasson (1977b). It would seem that many of these sediments display fabrics which could best be termed c-m supported, as defined in Chapter 3 for the Keweenawan sediments. Wasson (1977b) noted that debris flows and fluvial processes should be viewed as end-members of a continuum, and went on to suggest (p.794) that the c-m supported gravels "were laid down in streams which experienced high concentrations of fines", a similar explanation to that proposed in this study for the massive, c-m supported, coarse conglomerate facies.

Figure 7.4 shows c-m supported gravels within Cedar Creek fan, in the Madison Valley, Montana. These fans are at present largely inactive, and, like the Tasmanian fans of Wasson (1977a,b), were built during more humid glacial periods. It might therefore be concluded that periglacial conditions tend to give rise to high sediment fluxes, producing highly concentrated streams which deposit c-m supported conglomerates, and furthermore, that this situation in some respects provides a parallel for the hydrological conditions that prevailed in the Keweenawan, prior to the advent of plant life.

The Proterozoic alluvial fan deposits of the Van Horn Sandstone were studied by McGowen and Groat (1971) with the

Figure 7.4 Gravels within Cedar Creek alluvial fan, Montana. The fan surface slopes to the right of the field of view. There is a poorly developed imbrication dipping to the left. The matrix is a poorly sorted sand. Note the variation between clast and matrix supported fabrics, and the generally poor sorting. The rule is extended 30 cm. Photographed by myself during an alluvial fan field seminar run by Indiana University.



aim of setting up a humid-region alluvial fan model based on the rock record. The information provided by the Van Horn Sandstone is, by itself, not fully sufficient to provide a very reliable model (indeed, no "model" intended for general application should be based on a single example), and the Van Horn model would benefit from input of information from other studies. The Van Horn model has, nevertheless, received some attention, and was mentioned by Galloway and Hobday (1983). It was also quoted by Nilsen (1982), despite being in most respects very similar to the humid-region alluvial fan model of Boothroyd and Nummedal (1978), which was disregarded by Nilsen as irrelevant to alluvial fans. According to McGowen and Groat (1971) the Van Horn alluvial fans were large and gently sloping, and were built entirely by braided stream deposition. The facies and their ascribed positions on the fan are similar to those of the humid-region alluvial fan model of Boothroyd and Nummedal (1978), with the notable difference that trough cross-stratified sandstones are abundant on the middle and lower fan, and even occur in the upper fan deposits.

Another Precambrian alluvial fan system was described by Williams (1969). This study documents relatively thoroughly the sediments of large alluvial fans deposited entirely by braided streams. In addition, this is one of the

rare examples where radiating palaeocurrents make it possible to reconstruct the original fan shape. In this case, the fans were between 30 and 60 km in length. Williams (1969) considered that the style of weathering indicates a humid climate with a dry season. The "facies" described by Williams (1969) are very similar to the facies associations defined for the Keweenawan sediments in this study. Sandstone is important in these fans, even in the upper fan deposits (as in the Keweenawan), and Williams (1969) noted a large component of trough cross-stratified sandstones.

The only studies of a modern fan as large as those described by Williams (1969) are those of the Kosi Fan (Gole and Chitale, 1966; Wells and Dorr, 1987). This fan is 160 km in length, and so is much larger than those of Williams (1969). The climate of the Kosi Fan is humid, and strongly seasonal. The fan profile (Wells and Dorr, 1987, p.53) shows two segments: the upper is 40 km in length with a gradient of 0.00185; the lower is 120 km in length with a gradient of 0.00035. The top 15-20 km of the fan are gravelly (Gole and Chitale, 1966). The rest of the fan consists predominantly of sand and fines. The modern Kosi River changes from a braided to a meandering form about 147 km from the fan head (Wells and Dorr, 1987, p.59). However, much of the fan consists of flood plain environments and smaller channels

which meander or have forms between braided and meandering (Wells and Dorr, 1987). Exposure of sediments on the Kosi Fan is very poor. Wells and Dorr (1987) trenched some areas, and noted the importance of floodplain deposits and point bars containing lateral accretion surfaces and rippled sands. Some trough cross-stratification is present, but reworking to produce rippled sand appears most common (although Wells and Dorr, 1987, noted the strong viewing bias towards tops of fluvial cycles). Pliocene sediments, deposited by large rivers cutting through the Himalayas (like an ancestral Kosi), have been described by Parkash et al. (1980). These were interpreted as the product of sedimentation on large alluvial fans consisting of an upper portion with braided streams depositing coarse sediments and a lower portion with meandering streams depositing sandy in-channel deposits and overbank fines in fining-upwards sequences.

A braided-meandering transition recorded in the Quaternary deposits of the Reno alluvial fan has been described by Ori (1982). Here the transition in fluvial style occurs over a distance of only 8 km. The upper fan was apparently covered by longitudinal gravel bars, and passed into a lower fan covered by gravelly and sandy point bars and overbank deposits. The gradient of this fan is 0.0025.

Neither the Reno alluvial fan or the Indian alluvial fans fit the humid-region alluvial fan model of Boothroyd and Nummedal (1978). These studies illustrate the possible occurrence of meandering streams on the lower parts of relatively flat alluvial fans. In the Keweenawan, this style of sedimentation does not appear to have been important. However, the (rare) large-scale planar cross-stratified fine conglomerates (facies C-9) were clearly the result of lateral accretion of deep channels, which may well have shown a meandering form.

In summary, therefore, the comparison of the Keweenawan sediments with humid-region alluvial fans yields the conclusion, that in a number of respects the Keweenawan sediments fit quite well with a humid-region alluvial fan model where deposition is entirely by braided streams. The amount of sandstone in the Keweenawan sediments, particularly that which shows trough cross-stratification, is, however in contradiction with the humid-region alluvial fan model of Boothroyd and Nummedal (1978). Since other ancient examples of alluvial fan sediments show the same type of misfit, there is probably a need to modify this humid-region alluvial fan model so that it incorporates the features of larger fans with lower slopes, on which trough cross-stratified sandstones can form and be preserved.

7.3 Comparisons with braided fluvial systems

The argument as to what is best designated a braided fluvial system and what is best designated an alluvial fan is deferred to section 7.4. For the present discussion, it is only necessary to recognise that alluvial fans are at least partly covered by braided streams. Since this is the case, it is of relevance to discuss the similarities between the Keweenawan sediments and braided stream sediments as described in various studies of the modern and ancient, and as summarized in various models of braided stream sedimentation. The aim here is not to try and discuss all of the wide spectrum of braided fluvial types, but to focus on the braided fluvial systems that bear some particular resemblance to the Keweenawan sediments.

The vertical profile models, constructed by Miall (1977; 1978) for various types of braided streams, provide a starting point for comparisons with the Keweenawan sediments. The "Trollheim" and "Scott" vertical profile models were briefly considered in the alluvial fan discussion, with respect to the conglomeratic facies association. The "Donjek" vertical profile model is the only mixed sand-gravel profile. In comparison with the Keweenawan conglomerate-sandstone facies association two major

differences appear. Firstly the conglomerate beds in the Keweenawan are frequently thick, too thick to be each the product of sedimentation by a single channel or channel system as implied by the vertical profile model. None of the (channel-fill) fining upwards sequences seen in the model are apparent in the Keweenawan. Secondly, planar cross-stratified sandstones are a prominent feature of the model, but in the Keweenawan there is more trough cross-stratified sandstone, and flat-bedded sandstone or sandstone of the horizontally stratified facies is abundant. The Keweenawan sandy facies association is most similar to the "South Saskatchewan" vertical profile model of Miall (1978), but here too, to produce a reasonable fit it is necessary to substitute flat-bedded sandstone and horizontally stratified sandstones for the planar cross-stratified sandstones in the model. The Keweenawan rippled sandstone facies association bears some similarity to the "Bijou Creek" vertical profile model of Miall (1978), but the flat-bedded sandstone in the model must be largely replaced by rippled sandstone to produce a reasonable resemblance. These differences are hardly surprising when it is considered that the models, if indeed they merit the term, are based on so few examples (in most cases one). Walker and Cant (1984) commented on the lack of studies of sandy braided fluvial systems that are

comprehensive and detailed enough to be used to build models, and noted that there is a need to integrate more studies of the ancient into the models. These points make it reasonable to consider the variant models that might be obtained by rolling together various of the models of Miall (1977; 1978). This approach will not, however, remove the misfits with the Keweenawan sediments, as outlined above. It is therefore necessary to search for other analogues to the Keweenawan.

Poorly sorted coarse conglomerates of the Pliocene White Channel Gravels (Morison and Hein, 1987) appear to be texturally similar to the massive, c-m supported, coarse conglomerates of the Keweenawan conglomeratic facies association. Morison and Hein (1987) described their "disorganised gravel" facies as containing a high matrix content of sand and mud, with rounded clasts, generally in clusters. These were interpreted as the deposits of "sediment-laden gravelly flows, which contained a large amount of suspended sediment (possibly hyperconcentrated flows)" (Morison and Hein, 1987, p. 207). The "massive to crudely stratified gravel" facies of these authors, was described as poorly sorted, clast-supported and matrix-filled. Their photograph of this facies (p.208) shows sorting sufficiently poor to make the clast-matrix division

unclear. As in the Keweenawan massive, c-m supported, coarse conglomerate facies, clasts are rounded, and the gravels contain discontinuous sandy interbeds. Most interestingly, the imbrication in this facies is mainly a-axis upstream, with only subordinate b-axis upstream imbrication (which I would interpret as an indication that flow was concentrated enough to shear clasts into flow parallel orientations). Morison and Hein (1987) tentatively interpreted this facies as the result of "deposition from either hyperconcentrated flood flows, or stream flows with very high sediment concentrations and high current velocities", although without defining the difference between these two mechanisms. As in the Tasmanian examples of Wasson (1977b), weathering at the time of White Channel deposition seems to have been enhanced by periglacial conditions, giving rise to large amounts of sediment. In the Keweenawan, the same effect was probably achieved by a lack of terrestrial plant life.

The "horizontally stratified conglomerate" facies of Steel and Thompson (1983) is another example of c-m supported conglomerate similar to the Keweenawan facies C-1. These authors concluded (p.351) that "the sand and cobbles were deposited more or less simultaneously, and from a flow in which there was a high sediment concentration and rapid

deposition". Steel and Thompson (1983) studied the Triassic Bunter Pebble Beds, which consists of a mixture of conglomerate and sandstone, with sandstone generally somewhat less abundant than conglomerate. This Formation provides a comparison for the conglomerate-sandstone facies association of the Keweenawan. The Bunter sandstones and pebbly sandstones are both planar cross-stratified and trough cross-stratified, but in contrast to the Keweenawan, where trough cross-strata dominate, the Bunter is dominated by planar cross-strata. As already mentioned, some of the Bunter conglomerate fabrics are comparable to those of the c-m supported Keweenawan facies. In general, however, much more planar cross-stratification is present in the Bunter conglomerates than in the Keweenawan. Some of these are very large sets (Figure 3 of Steel and Thompson, 1983, shows one set 7 m in thickness), and appear similar to the Keweenawan large-scale planar cross-stratified fine conglomerate facies. The planar cross-stratified Bunter conglomerates were interpreted as the product of bars in deep braided streams, although Steel and Thompson (1983) debated whether incision due to tectonic activity may have produced some of the larger sets filling deep channels. As in the Keweenawan, some of the sandstone - conglomerate alternation in the Bunter is on a gross scale (15-30 m) and is laterally

extensive (several kilometres), requiring an explanation in terms of factors other than sedimentological controls (such as tectonics).

The Rio Grande at the Texas - New Mexico border is at present a straight river, due to artificial bank stabilisation, but the stratification types formed in this river (as described by Harms and Fahnestock, 1965) are similar to those of the Keweenawan sandy facies association. The Rio Grande at this point generally averages less than 1 m in depth and has a slope of about 0.0006, less than the palaeoslope estimates calculated earlier for the Keweenawan. The nature of the Rio Grande river bed varies considerably from season to season. In the spring, upper flow regime flat bed is predominant, but by late summer, the bed is almost entirely covered with dunes. Since these differences persist even when discharge is unchanged, Harms and Fahnestock (1965) attribute the change to temperature (viscosity) variations. The river sediments, however, which are almost entirely sandy, show much more trough cross-stratification than flat bedding. Harms and Fahnestock (1965, p.108) observed, from trenches over 1 m deep, that "large scale trough cross-stratification is volumetrically the most significant stratification form, and is found at depth in every area, under bar or thalweg". Bars are present, with

slip-faces that give rise to planar cross-stratification, and bar tops are covered by thin deposits of rippled or flat bedded sandstone. However, these stratification forms are not found at depth, because as channels containing dunes change position bars are eroded and replaced by trough cross-stratified deposits. At high stage, upper flow regime flat bed or even antidune bedforms may form in the channels, but during falling stage, the bed reverts to a dune configuration, and dune troughs scour to such a depth as to remove stratification formed during higher stage.

Jurassic sheet sandstones in New Mexico, described by Campbell (1976), appear to be an ancient example of sediments similar in style to those of the Rio Grande. These rocks consist very largely of trough cross-stratified sandstones (although partly of a larger size than those in the Rio Grande or the Keweenawan), with minor amounts of flat-bedded and rippled sandstones.

The sedimentary sections proposed by Walker and Cant (1984) to summarise the deposits of the South Saskatchewan river can easily be modified so as to resemble the Rio Grande sediments or the New Mexico Jurassic sandstones. In the South Saskatchewan, dunes deposit trough cross-stratification in the channels. If the overall rate of vertical accretion were sufficiently low, structures formed

at higher levels in the system (on the bars) would be destroyed as channel positions changed, and the South Saskatchewan deposits would at depth (and in the rock record) consist very largely of trough cross-stratified sandstone. The important variables here are sediment supply and subsidence rate. For the Keweenawan sandy facies association, this style of explanation does not quite work. The sandy facies association contains not only trough cross-stratified and flat bedded sandstone, but also large amounts of the horizontally stratified sandstone facies. As argued in Chapter 3, this facies was apparently formed on bar tops or in channels at higher levels in the braided system. Ripples and flat bed are part of the horizontally stratified facies (although other structures are present as well) and are described as characteristic of bar tops by many workers (including Walker and Cant, 1984). It therefore appears that in the Keweenawan, high levels of the braided system were preserved. The relative paucity of planar cross-stratification therefore may indicate that many bars did not develop angle of repose slip-faces.

7.4 Appropriateness of the alluvial fan - braided river distinction and the climatic division of alluvial fan types

The question as to what should or should not be termed an alluvial fan has proved to have a subjective answer. There is no argument that deposits, ancient or modern, which fit the arid-region alluvial fan model with abundant debris flows, should be called alluvial fans. However, many workers, despite the name alluvial fan, have been reluctant to concede the existence of alluvial fans without debris flows. According to Rust and Koster (1984, p.65) "the basic model for alluvial fan deposition.....is characterised by rapid fining in the downslope direction and by the presence of debris flows". Rust (1978) disliked the application of the term "alluvial fan" to the outwash fans studied by Boothroyd and Ashley (1975), on the grounds that these originate at a glacier-front rather than a mountain front. This objection is an important one, since the vast majority of ancient fan deposits are not of glacial origin. However, it is difficult to avoid the effects of the Pleistocene climatic fluctuations when searching for modern examples to assist model building, and without the glacial outwash fans, there remain no reasonable modern analogues for ancient alluvial fan successions built entirely by

fluvial processes. Rust (1978) in fact quotes, as an ancient alluvial fan, the Van Horn Sandstone (McGowen and Groat, 1971), which is an example of a deposit for which the outwash fan is a much better analogue than the arid-region alluvial fan model. Nilsen (1982) also dismissed the glacial outwash fans but extensively discussed the study of McGowen and Groat (1971) and quoted Williams (1969). For these two studies (discussed in section 7.2) it is by no means clear that if these were present-day fluvial systems workers such as Rust (1978) and Nilsen (1982) would be inclined to term them alluvial fans, since neither contains debris flow deposits. However, both these studies, in which the topography of the surface underlying the alluvial fan was documented, fulfil the criterion of Rust (1978, p.606) that "fan deposits differ from those of rivers and plains in that they represent a more immediate response to adjacent high relief". This is an important criterion for distinguishing alluvial fans. Alluvial fans do have particular tectonic implications, which are often of value in reconstructing the geological history of a basin. If it is not recognised that there is a category of alluvial fans that form adjacent to areas of high relief but that do not contain debris flows, then tectonic information will be missed. It is of note that such fluvially-dominated fans comply with the definition by

Blissenbach (1954, p. 176) that "an alluvial fan is a body of detrital sediments built up by a mountain stream at the base of a mountain front".

The types of information obtainable from the ancient and the modern are to a large extent, complementary, so it is undesirable to base any fluvially-dominated alluvial fan model on studies of only one or the other. The fluvially-dominated alluvial fan model should be an amalgam of studies of the modern glacial outwash fans and studies of ancient alluvial fan successions, such as Williams (1969), Nadon and Middleton (1985), McGowen and Groat (1971), and this study. Any such model will clearly consist of braided stream environments and sediments. To identify an ancient deposit as a fluvially-dominated alluvial fan rather than simply a braided river requires information on a broader scale, such as evidence of downfan facies changes, the proximity of a basin margin with marked topographic contrast, or radiating palaeocurrent patterns. Such detailed palaeocurrent information is rarely available in the rock record, but the basinal setting and the downfan facies succession are often evident.

The division of alluvial fans into arid-region and humid-region types does not work. Wasson (1977a,b) and Ryder (1971) have documented Pleistocene alluvial fans which

formed in a relatively humid periglacial environment, but which contain many debris flow deposits. Pleistocene glacial outwash fans containing debris flow deposits were described by Cherven (1984). Kochel and Johnson (1984) described Holocene alluvial fans from temperate regions and these too contain debris flows. Cretaceous temperate-region alluvial fans with debris flows were described by Wells (1984). Conversely, Nadon (1981) and Nadon and Middleton (1985) described Triassic arid-region alluvial fans that do not contain debris flows. The present-day arid-region alluvial fans in Death Valley do not all contain a significant component of debris flows. For example, the cuts in Badwater fan, a small steep fan that would be expected to contain a large number of debris flow deposits, show only fluvial deposits (e.g. Figures 7.1, 7.2, 7.3). The margins of Death Valley are partly covered by large alluvial fans (including the confined "washes"), which would be prominent in the rock record, but in which cuts reveal only fluvial deposits.

The major problem with the climatic division of alluvial fan types is that fan processes, and thus deposits, do not depend <u>only</u> on climate. The different variables which cause particular fans to be built predominantly by debris flows or entirely by stream flows are not fully understood. The answer must lie in the ways in which sediment and water

are delivered to fans. Where large volumes of sediment are produced in the fan catchment and sediment transport events are few and far between, debris flows are likely to result. Where only small amounts of sediment are produced in the fan catchment and sediment transport events are frequent, fluvial deposits are likely to result. The rate of sediment production is a function not only of climate, but also of catchment slopes (a function of uplift rate), presence/absence of vegetation (a function of climate and geological period), and bedrock type (most important for small catchments which might be dominated by only a few bedrock types). The frequency of sediment transport events is primarily a function of climate, but the nature of individual events will vary with the size and slope of the catchment (large, gently sloping catchments producing more prolonged periods of flow) and with the presence/absence of vegetation (vegetation increasing the duration of flow events). The type of sediment produced in the fan catchment is another variable that can influence sediment transport processes. Pierson (1981) has shown that debris flows can form without large amounts of clay (Mount Thomas debris flow surges have less than 5 % clay by weight), so it is unlikely that a fan catchment will deliver so few fines that debris flows cannot form. However, the proportion of fines will

have a dramatic effect on the appearance of the resulting debris flow deposits. Vegetation also plays a role in this respect, because vegetation favours soil formation (i.e., favours the reduction of weathering products to a fine grade before their release from catchment slopes).

Even if a climatic division of fan types could be shown to work for modern fans (which would clearly require many more climatic divisions than just humid/arid), there is good reason to doubt that such a division would be at all applicable to Precambrian fans such as the Scottish Torridonian (Williams, 1969), the Van Horn Sandstone (McGowen and Groat, 1971), or the Keweenawan of the Lake Superior Basin, which formed at a time when hydrological regimes were everywhere subject to a lack of terrestrial plant life. Larsen and Steel (1978) described the Devonian Karlskaret Fan as debris-flow dominated. A division between debris-flow dominated and fluvially dominated alluvial fan types might avoid many of the problems caused by the arid-region/humid-region division, but it must also be recognised that fans form a spectrum running from small steep debris-flow dominated types to large, relatively flat fluvially-dominated types where sand is an important component.

CHAPTER 8

TECTONIC, VOLCANIC AND CLIMATIC CONTROLS ON KEWEENAWAN ALLUVIAL FAN SEDIMENTATION

8.1 Some general remarks concerning alluvial fan behaviour

This Chapter considers the information that the sediments of the Lake Superior Basin can provide regarding tectonism, volcanism and climate in Keweenawan times. Before these various "extrinsic" factors can be analysed, it is necessary to distinguish their effects from those of "intrinsic" (i.e., sedimentological) controls. This distinction is largely made by consideration of the scale of the variations in sedimentary style. For any single stratigraphic section, extrinsic controls can only be isolated with any certainty if the vertical variation in sediment type is relatively gross (on a scale of about 10 m or more). However, if information is obtainable as to the lateral extent of sediment bodies, then it may be possible to ascertain whether smaller variations are also extrinsically controlled.

Distinction of climatic, tectonic and volcanic controls on sedimentation is difficult because all three can

have similar effects. Even on present-day fans there can be scope for more than one interpretation. In general, however, climatic variations sufficient to cause marked alterations in sedimentary style are likely to occur over a longer time span than tectonic and volcanic variations (excluding the rapid and severe climatic variations that occur during glaciations). The coarseness of the Keweenawan sediments and the presence of interbedded basalts indicate that the Keweenawan environment was tectonically and volcanically active, so these two controls are likely to be the most important.

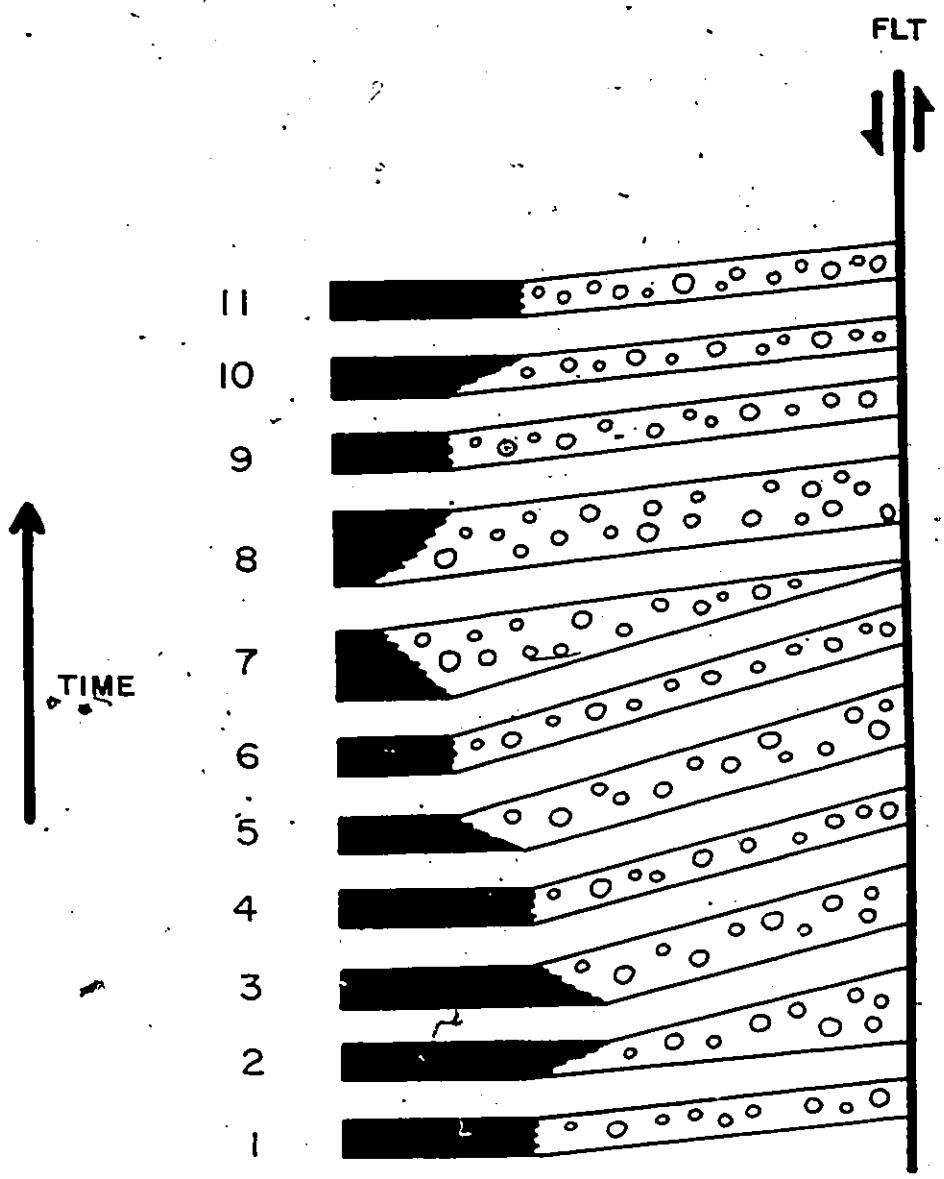
Volcanic damming is a base level alteration, and as such will produce effects which are immediate at the toes of fans and which gradually work back up the fan. It might therefore be anticipated that distal fan environments will show more and greater effects of volcanism than proximal fan sediments. Even if any volcanic dam were to last long enough for the base level change to effect the whole fan, the thickness of sediments involved might be expected to be greater at the fan toe than at the fan head. However, as argued in Chapter 7, the Keweenawan fans were relatively gently sloping, so, using the largest palaeoslope calculated in section 7.2 (0.0034), even a 30 km long Keweenawan fan was only about 100 m high (a maximum estimate, since the

calculation uses the maximum probable palaeoslope). The thickest Keweenawan flow is more than 300 m thick (White, 1960), indicating that a dam formed by just one lava flow could flood an entire Keweenawan alluvial fan and cause the whole fan surface to be draped with lacustrine sediments. In contrast, to tectonically flatten out a Keweenawan fan from base to apex requires that the fan head be dropped 100 m relative to the fan toe. In reality, the amount of faulting at the basin margin needed to produce this effect is much more, because the whole basin floor will take up the tilt. Tectonic flooding of the Keweenawan alluvial fans would therefore have been a more drawn out process than volcanic flooding and would be reflected by a more gradual change in the type of sediments deposited at any one point on the fan.

In general, the rate of tectonic movement between an alluvial fan and its catchment will vary through time. The variation will occur on several time scales, ranging from centuries with particularly high or low earthquake activity through millennia characterised by slow or fast rates of tectonic movement to periods of a million years or more during which mountain building is particularly intense or subdued. In response the size and slope of an alluvial fan will change. If there is a sufficiently protracted alteration of the tectonic intensity then the overall relief

and the average slope of the fan catchment will significantly alter, causing a change in the amount and coarseness of the sediment delivered to the fan. Figure 8.1 follows the evolution of an alluvial fan-playa system through a tectonic cycle consisting of an initial state of slow (constant) hinterland uplift to a state of fast (constant) uplift and returning to a final state of slow (constant) uplift the same as at the beginning of the cycle. The probable time scale of the cycle is discussed at a later stage. In the remarks that follow "steady-state" is used in the sense of Hooke (1968) to mean a state of dynamic equilibrium such that on average every point on the fan-playa surface is aggrading at the same rate and the fan radius is constant (any fan that is not in this state will continue to get larger and flatter or smaller and steeper). In (I) of Figure 8.1, the fan and playa are aggrading in steady state. The fan-playa boundary aggrades vertically at a position determined by the relative rates of supply of fan gravel and playa fines. In the two dimensional system shown the relative extents of the fan and playa imply a fan gravel / playa fines supply ratio of 2:1 (in steps (I) to (II) the playa fines are assumed to be supplied entirely from the fan catchment). The steady state shown in (I) implies that the uplift rate of the catchment has been constant for a time

Figure 8.1 Evolution of a simple two dimensional alluvial fan - playa system during a hypothetical tectonic cycle. The spotted pattern represents alluvial fan gravels; the solid black pattern represents playa fines. The stages I-II are <u>not</u> intended to represent equal time intervals, but show periods during which the alluvial fan is behaving in a particular fashion. For clarity the different stages are separated; if placed directly above one another, they would form a stratigraphic section. See text for a detailed discussion.


FLT
11
10
9
8
7
6
5
4
3
2
1
TIME

sufficient that at least the lower part of the catchment stream has graded to an equilibrium profile, so that the fan head is rising at a constant rate relative to some marker fixed in the fan-playa sediments. In (2) uplift increases to a faster (but constant) rate. As a result, the catchment stream is raised faster and grades to a new equilibrium profile. It is presumed that the catchment is sufficently large that there is some substantial time lag before the change in uplift rate influences slopes and relief in enough of the catchment to influence sediment supply to the fan (the validity of this assumption is discussed later). Since the mouth of the catchment stream is now rising at a faster rate a larger proportion of the sediment supply must be deposited at the fan head (otherwise a gap will result), thus causing the fan to steepen towards a new equilibrium profile. An extreme form of this readjustment occurs on the fans on the eastern side of Death Valley, where sediment is only being deposited on steep cones at the tops of fans (Hooke, 1972). Whilst the fan is changing towards its new equilibrium profile, a smaller proportion of the fan gravel is deposited on lower parts of the fan, with the consequence that the lowest parts of the fan begin to aggrade more slowly than the playa, so the fan-playa boundary moves towards the fan. However, once the fan has attained its new

equilibrium profile (stage (3) of Figure 8.1) the supply of fan gravel again becomes evenly distributed over the fan surface and the fan progrades until the fan-playa system is at steady state (4). Since the rate and type of sediment supply has not changed the fan-playa boundary in stage (4) is at the same distance from the fault as in stage (1) (it may be noted that although the slope has changed, the fan gravel deposited in the same time interval in the two different situations forms parallelpgrams which are of equal area when the fan radii are equal). At some stage, the increased rate of uplift will increase relief and slopes in the catchment to the extent that the amount and the coarseness of the sediment supplied to the fan-playa system is significantly altered. Stage (5) shows progradation of the fan over the playa in response to an increase in the fan gravel / playa fines supply ratio. It seems likely that the rate of this progradation will be markedly different to the rate of progradation shown in stage (3). Some of the problems of differentiating these two types of fan adjustment are discussed later. Stage (6) shows attainment of a new steady state, with a fan gravel / playa fines supply ratio of 4:1. Stages (7) to (11) are essentially a reversal of stages (1) to (6), ending at a final steady state identical to the initial steady state. In (7), the

uplift rate is decreased to its initial rate. As the fan grades towards a new equilibrium profile, proportionately more fan gravel is deposited at the fan toe, causing the fan to prograde over the playa. The fans on the western side of Death Valley provide an extreme example of this behaviour: these are incised at the head, with deposition only occurring at the fan toes. Once the new, shallower equilibrium fan profile is developed, the fan retreats (8) until steady state is reached (9). At some stage, the decreased uplift alters the overall state of the catchment to such a degree that the sediment supply decreases and becomes finer, causing the fan to retreat (10) until it reaches a steady state corresponding to a fan gravel / playa fines ratio of 2:1.

The hypothetical tectonic cycle described above would give rise to both coarsening and fining sequences (if downfan facies changes were significant). From the sedimentological point of view, it is of interest to know how the sequences produced during the different stages differ in style and scale. Stages (2) and (7), (3) and (8), (5) and (10) are essentially pairs of opposite steps, and the two members of each equivalent pair of sedimentary sequences will probably be of similar scale. To separate each of the types of sedimentary sequence requires a

detailed modelling of the catchment-fan-playa system involving simulation of the catchment stream profile, erosion mechanisms and other variables. This will not be attempted here. However, some approximate calculations are made below to try and determine whether it is possible to clearly separate sedimentary sequences resulting from the morphological response of the fan (stages (2), (3), (7) and (8)) from the sequences due to the overall alteration of the catchment and sediment supply (stages (5) and (10)). It may also be noted that, according to Figure 8.1, the initial response to an increase in uplift rate might be a fining-upwards sequence, and a coarsening upwards sequence may initially result from a decrease in uplift rate. This is not what is usually assumed in analysis of the rock record.

The scale of the sedimentary sequences resulting from changes in the overall catchment slope and relief will depend upon the relationship between sediment supply and relief and upon sedimentation rates on the fan. Schumm (1977, p. 22) provided a curve relating denudation rates to drainage basin relief. This is of exponential form, indicating an approximate doubling of denudation rate for every 5000 feet (1500 m) of added relief. It seems likely that an increase of 50% in the rate of sediment supply to the fan will sufficiently affect the fan sediments to be

detectable in the rock record. This will require about 700 m of additional hinterland relief, regardless of the initial relief. However, Ahnert (1970) found a linear relationship between denudation rate and relief. The conflict between these studies may result from the large difference in the size of the drainage basins considered (Schumm, 1977, only considers small drainage basins). Ahnert's (1970) linear relationship requires that, in order to estimate the change in sediment supply, there is some estimate of the initial relief. The Keweenawan hinterland was probably not highly mountainous, so an initial relief of about 1500 m is probably reasonable. Ahnert's (1970) relationship then indicates that to increase the sediment supply by 50%, an extra 750 m of relief are required. This value is very similar to that obtained using Schumm's (1977) relationship; however, if other values of the initial relief are used the difference in the two estimates may be much larger. If these calculations are accepted, at least as a first approximation, then the <u>minimum</u> time for sediment supply to be increased by 50% is the time needed for 700 m of uplift. Schumm (1977, pp. 30, 31) quoted 8 m per 1000 years as a typical rate of mountain uplift. In mountainous environments, erosion is also rapid. Schumm (1977, p.34) suggested that 1 m per 1000 years may be a reasonable figure

for denudation rates in the world's mountainous regions, so a typical rate of increase in mountain height during a period of tectonic activity is probably about 7 m per 1000 years. At this sort of rate, approximately 100,000 years are required to add 700 m of relief to the catchment and significantly influence sediment supply to the fan. In this period, using the sedimentation rate of 1 m per 1000 years estimated by Hooke (1972, p. 2093) for Death Valley fans, about 100 m of sediment are likely to have been deposited on the fans. The sedimentation rate determined by Beaty (1970) for Milner Creek alluvial fan in the White Mountains of California is about a magnitude less than that of Hooke (1972) (although, unlike Hooke, 1972, Beaty, 1970, has no core data). If this rate is used, only 10 m of sediment will have been deposited on the fans during the 100,000 years needed to markedly increase hinterland relief. Hence, the sedimentary sequences resulting from this (long term) kind of tectonic change are likely to be somewhere between 10 m and 100 m in thickness. Although these calculations are very approximate, it is of note that vertical movement on the fault to the east of Death Valley is about 7 m per 1000 years (Hooke, 1972, p. 2096), which agrees well with the kind of uplift rate assumed above. Furthermore, since the fans on the western side of Death Valley are of very similar

size to their catchment areas (Hooke, 1972, p. 2087), the denudation rate of 1 m per 1000 years indicated by Schumm (1977) and used above would produce the sedimentation rate of 1m per 1000 years observed by Hooke (1972), if all the sediment were to be deposited on the fan. Rates of sedimentation of the Copper Harbor Formation can be (very approximately) estimated from the ages given by Van Schmus et al. (1982) (as reviewed in Chapter 2). According to these authors, the top of the Portage Lake Lava Series may be dated at 1,110 ± 10 Myrs and the Nonesuch Formation above the Copper Harbor Formation may be dated at 1,023 ± 46 Myrs. If the Copper Harbor Formation is assumed to have been deposited in the 87 million years bracketed by these figures, then the average Formation thickness of about 1500 m (as discussed in Chapter 2) implies an average Copper Harbor sedimentation rate of about 0.02 m per 1000 years (although, it should be noted that the figure of 87 Myrs could be in error by about 50 Myrs). This sedimentation rate is a magnitude less than even the estimate of Beaty (1970), but since Nonesuch Formation sedimentation may have been very slow, and could account for much of the 87 million years, this is a <u>minimum</u> sedimentation rate for the Copper Harbor Formation. If this figure is real, however, then reasoning as before, sedimentary variations on a scale of just 1 m

could reflect relief variations resulting from long term changes in the rate of hinterland uplift. This scale of variation is less than channel depths on alluvial fans, so in this depositional environment sequences from one long-term tectonic episode will become confused with sequences from later long-term tectonic episodes, and will be thoroughly intermingled with any sequences resulting from morphological changes in the fan caused by shorter term tectonic changes. In Death Valley, the data of Hooke (1972) indicate that sedimentation rates are high, so it might be expected that long-term and short-term tectonic changes will be distinguishable in the sedimentary record. Hooke's (1972) profiles of Badwater fan indicate that the new steep segment is about 3 m thick. This gives some approximate idea of the scale of sedimentary sequence that might result from morphological changes on Death Valley fans, and is significantly less than the scale calculated earlier for sequences resulting from long-term tectonic variations in this area (however, no sequences will result unless there are significant downfan facies variations, and as briefly noted in Chapter 7, such variations are not prominent in Death Valley fans).

Previous studies of alluvial fans have not attempted to isolate the modes of alluvial fan behaviour which can

prevail long enough to determine the characteristics of a substantial thickness of the geological record. Figure 8 of Heward (1978) shows two scenarios: the first, intended to describe fan response to fast tectonic uplift, depicts a fan that is becoming progressively steeper and shorter (which ultimately would result in a vertical fan of zero radius); the second, intended to describe fan response to slow tectonic uplift, depicts a fan that is becoming increasingly flat and long (which ultimately would produce a horizontal fan of infinite radius). The hypothetical tectonic sequence presented in Figure 8.1 does depict fan behaviour that can continue indefinitely, and therefore is a more useful aid to interpretation of the geological record.

8.2 Evidence for tectonic and volcanic control of Keweenawan sedimentation

8.2.1 Basin-fill sequences

As mentioned in Chapters 2 and 4, the Copper Harbor Formation shows an overall fining upwards into the Nonesuch Formation, which in turn coarsens upwards into the Freda Formation. This pattern is most likely to reflect the long term tectonic trends. It is reasonable that the coarsest

sediments should coincide with and directly follow the main pulse of volcanic activity. After this, relief and slopes gradually decreased until a resurgence of tectonic activity gave rise to the Freda Formation. The Freda Formation is, overall, finer than the Copper Harbor Formation, so it appears that this second tectonic pulse was not as severe as the earlier phase. It cannot be said with certainty that climate was not also a long term sedimentary control. Palaeomagnetic measurements (as mentioned in Chapter 5), although necessarily imprecise, do indicate a general drift from palaeolatitudes of about 20° N during Copper Harbor sedimentation to very low latitudes during Nonesuch sedimentation. This movement may have caused an increase in precipitation, and could have contributed to the establishment of a relatively permanent distal lake.

8.2.2 Volcanically controlled sequences

The stromatolite-oncolite horizons and associated siltstones at Horseshoe Harbor and Dan's Point on the Keweenaw Peninsula are interbedded with relatively proximal fan deposits. As indicated in Chapters 3 and 4, some of these horizons are laterally extensive (2.5 km at Dan's Point), testifying to flooding of large areas of the fans.

The contacts between conglomerates and siltstones or stromatolite-oncolite horizons are sudden. The discussion in section 8.1 indicated the ease with which volcanism could cause fan flooding and the relative difficulty of tectonically flooding the fans. It was predicted that tectonic flooding, if it occurred, should produce gradual facies changes. The Horseshoe Harbor and Dan's Point sequences are therefore best explained as the product of volcanic dams formed by one or several flows.

The Goodharbor Bay interflow sediments were described in detail in Chapter 4. These also were interpreted as the result of volcanic damming. The two fining upwards sequences seen here are of a very similar thickness to many of the lava flows making up the North Shore Volcanic Group in this area, and are precisely what would result from two volcanic damming episodes occurring at time intervals such that the lake was just filled between eruptions.

8.2.3 Small tectonically controlled sequences

The conglomerate-sandstone facies association at Sandstone Falls, Rainbow Falls and Dan's Point consists of a conglomerate-sandstone alternation on a scale too coarse to be explained solely by sedimentological fluctuations.

Obvious sequences, coarsening or fining upwards, are not present. The Dan's Point sections show coarsening upwards sandstones capped by conglomerate. A similar style of sequence although thicker (15 m) is seen above the 26 m level in the Sandstone Falls section (Figure 4.9). The only other gradual coarsening upwards in this section is around the 65 m level. At Rainbow Falls, it is hard to distinguish sequences of any kind. In general then, sedimentary variations take the form of rather sudden conglomerate-sandstone alternations or weakly defined coarsening upwards sequences on a scale of about 10 - 15 m. The lateral extent of most of these alternations is not known, but the Dan's Point correlation and the lateral extent of the sandstone beds at the 5 and 17 m levels in the Sandstone Falls section (at least 200 m) indicate that the sedimentary variation is probably not simply the result of channel migration or other intrinsic factors. The vertical scale of the variation supports this conclusion. The operative extrinsic control is probably tectonic. The uncertainty about Keweenawan sedimentation rates allows that these variations be interpreted as the result of short-term changes in the fan morphology or the result of long-term changes in hinterland relief as tectonic activity varied in intensity. The latter explanation might fit better with irregular and poorly

developed sequences such as those observed. The former explanation is probably more applicable to alluvial fan successions dominated by well defined sequences (e.g. the Hornelen Basin of Norway described by Gloppen and Steel, 1981). These two different styles of fan succession probably result from tectonic regimes of differing intensity. Only in very tectonically active areas where sedimentation is fast and tectonic pulses large can the detailed imprint of tectonic pulses be preserved in the rock record. In less active areas, tectonic pulses are smaller and rates of sediment supply are less, so any sedimentary cycles resulting from short term tectonic variations will be small and ill-defined. If sedimentation rates are sufficiently low much fan sediment will be reworked during several successive tectonic pulses, blurring the effects of each one (corresponding to a situation where the layer of sediment deposited during each tectonic phase is so thin that channels cut down into sediments deposited during the previous phase). Hence, only the longer term variations in sedimentation (due to relief alteration) are preserved.

As noted in Chapter 3, the Union Bay sections in the sandy facies association display a weakly defined alternation between mainly trough cross-stratified sandstones and mainly horizontally stratified sandstones, on

a scale varying between about 2 and 15 m. The top part of the Rainbow Falls section (Figure 4.11), which is also of the sandy facies association shows a similar alternation. The scale and generally vague style of this alternation is similar to that of the conglomerate-sandstone facies association, and so is interpreted as having a common cause. In general, since the Copper Harbor formation thickens basinwards, sedimentary cycles would also be expected to thicken in this direction. However, as pointed out in Chapter 2, much of the subsidence in the basin centre was probably taken up by lava flows.

The rippled sandstone facies association was discussed in Chapter 3, where it was pointed out that some fining upwards sequences are present, with a scale of about 2 - 3 m. These are very poorly developed, and only one thick section was measured in this facies association. It is not possible on the basis of this evidence to determine whether these sedimentary variations resulted from tectonic movements or from purely sedimentary variations such as migration of active parts of a fan or migration of channels. It may be noted, however, that the scale of the sedimentary variation in the rippled facies association is smaller than that of the tectonically controlled variation in the coarser facies associations.

The origin of the large scale planar cross stratified fine conglomerate seen at Rainbow Falls is debatable. The sets are 4 m thick, and the presence of trough cross stratified sandstones within the stratification points to a lateral accretion origin. Although channel depths decrease downfan, if the water does not end up by simply evaporating on a playa or entering the groundwater system, then there must come a point when channel size increases again to form a trunk stream draining the valley. However, the Rainbow Falls section does not consist of the most distal facies associations in the Copper Harbor Formation. Furthermore, overbank fines of the type expected with a large meandering river are entirely absent. It therefore seems probable that this exceptionally large channel represents incision due to a long period of tectonic quiescence.

CHAPTER 9

SYNTHESIS AND CONCLUSIONS

In Middle and Late Keweenawan times, large fluvially-dominated alluvial fans developed at the margins of the mid-continent rift valley. Several lines of evidence indicate that the depositional environment is best considered as a piedmont of alluvial fans rather than simply as a braidplain (the debate over this type of distinction was reviewed in section 7.4). The coarseness and the thickness (up to 1850 m) of the sediments together indicate a tectonically active environment, and the work of Wolff and Huber (1973) on Isle Royale has shown that facies change markedly over a relatively small distance in the downcurrent direction (in the 24 km between Cumberland Point and Point Houghton there is a change from a considerable predominance of conglomerate to a marked predominance of sandstone). As far as can be ascertained from geophysical data (e.g. that of Hinze et al., 1982), the sediments are close to the original rift margins, and the palaeocurrent measurements of this and other studies indicate that in general dispersal was away from those rift margins (although it is evident that there was an overall basin palaeoslope to the east or northeast).

Outcrop is not sufficiently abundant to define radiating palaeocurrent patterns, but the other features outlined above indicate that an alluvial fan interpretation is most appropriate.

The Keweenawan sediments comprise five facies associations. The conglomeratic facies association is the most abundant. The conglomerate-sandstone facies association and the sandy facies association make up the bulk of the rest of the sediments, the remainder being of the rippled sandstone facies association and the silty facies association. The conglomeratic, conglomerate-sandstone and sandy facies associations are interpreted as braided stream sediments that at the time of deposition were laterally equivalent, forming a fining downfan succession. More distally these environments passed into a floodplain where depositional topography was subdued, giving rise to sediments of the rippled sandstone facies association. Sedimentation in ephemeral distal lakes, when and where present, produced sediments of the silty facies association.

Deposition on the highest parts of the fans, giving rise to the conglomeratic facies association, occurred largely on longitudinal bars without slipfaces. Boulders, cobbles, pebbles and sandstone were deposited rapidly and simultaneously producing poorly sorted c-m supported

conglomerates. This mode of deposition predominated because streams rose and waned rapidly and whilst in flood were heavily loaded with sediment. These conditions probably resulted partly from the lack of plants in Proterozoic times. During waning flow sandstones were often deposited in channels producing flat-bedded and low angle cross-stratified sandstones. The latter sediments may be the deposits of partially formed dunes or antidunes.

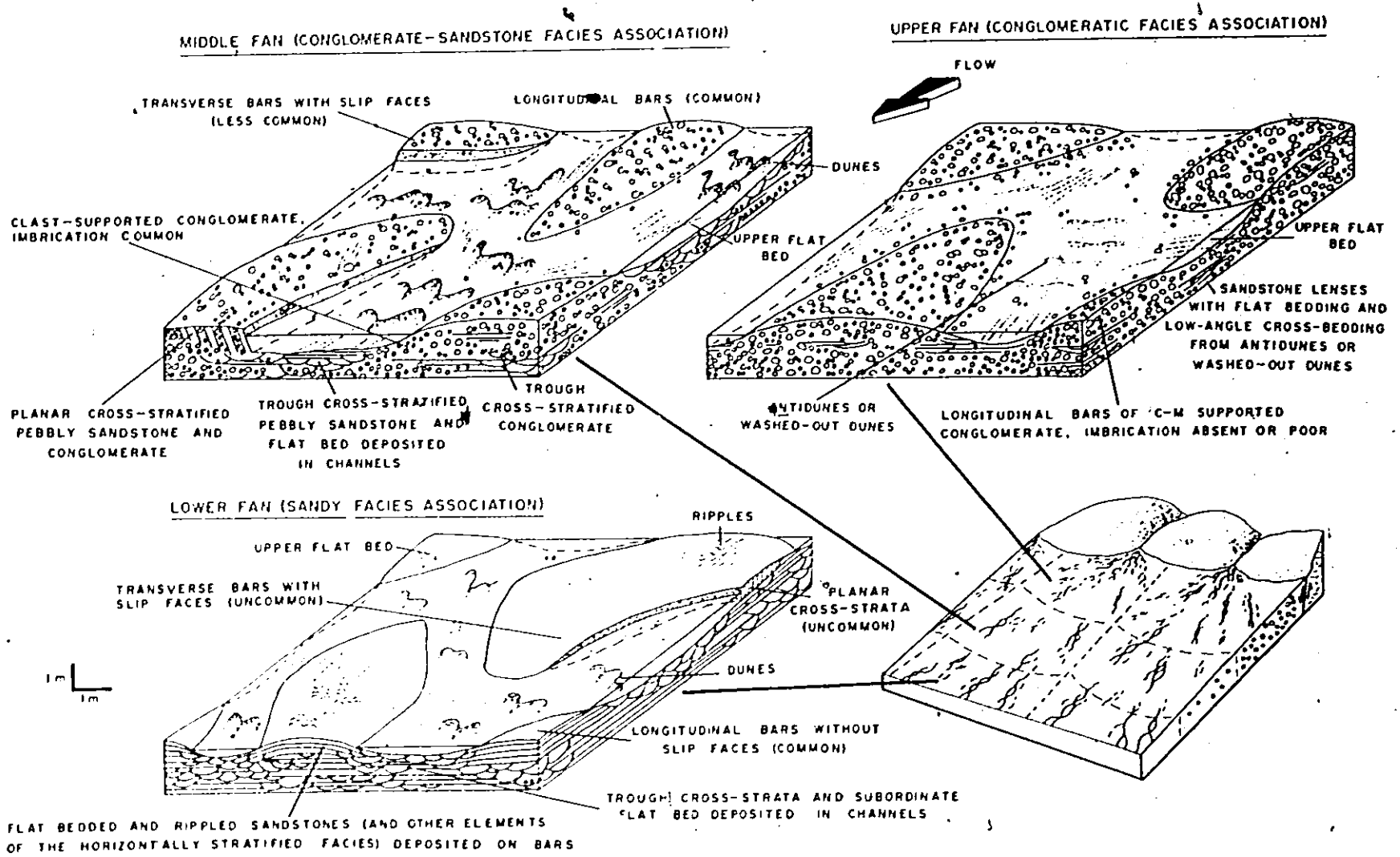
The conglomerate-sandstone facies association was deposited on the mid-fan. As on the upper fan, longitudinal bars without slipfaces produced c-m supported conglomerates, but some bars possessed slipfaces, giving rise to planar cross-stratified conglomerates and pebbly sandstones. Imbrication is more common in mid-fan deposits than on the upper fan. Trough cross-stratified sandstones, pebbly sandstones and conglomerates were deposited by dunes in the channels. Flat bedded sandstones formed both in channels and at higher levels. The alternation between sandstone and conglomerate was partly controlled by tectonism (discussed below).

The lower fan environment was almost entirely sandy. Trough cross-stratified sandstones and subordinate flat bedded sandstones were deposited in channels. The horizontally stratified sandstone facies (flat bedded and

rippled sandstones, mud drapes with desiccation cracks, scoured surfaces with mudchips, knobbly surfaces) was deposited on bar tops. The rarity of planar cross-stratification indicates that bars generally lacked angle of repose slip faces. As in the conglomerate-sandstone facies association, the facies alternation bears a tectonic imprint (discussed below).

The depositional environments of the three major Keweenawan facies associations are reconstructed in the summary diagram of Figure 9.1. In comparison with modern glacial outwash fans (Boothroyd and Nummedal, 1978), the Keweenawan environment gave rise to much more trough cross-stratification and correspondingly less planar cross-stratification. Imbrication is also less evident in the Keweenawan, partly because deposition was often fast, producing c-m supported conglomerates. The "facies" defined by Williams (1969) for Torridonian fan sediments appear to be very similar to the Keweenawan facies associations described here, although the Torridonian conglomerates apparently are better sorted than the Keweenawan conglomerates. The Van Horn sandstone described by McGowen and Groat (1971) also appears similar to the Keweenawan. McGowen and Groat (1971), however, apparently find better defined trough cross-stratification in their "proximal

Figure 9.1 A summary diagram of depositional environments on the fluvially dominated Keweenawan alluvial fans.


MIDDLE FAN (CONGLOMERATE-SANDSTONE FACIES ASSOCIATION)
TRANSVERSE BARS WITH SLIP FACES (LESS COMMON)
LONGITUDINAL BARS (COMMON)
DUNES
CLAST-SUPPORTED CONGLOMERATE, IMBRICATION COMMON
UPPER FLAT BED
PLANAR CROSS-STRATIFIED PEBBLY SANDSTONE AND CONGLOMERATE
TROUGH CROSS-STRATIFIED PEBBLY SANDSTONE AND FLAT BED DEPOSITED IN CHANNELS
TROUGH CROSS-STRATIFIED CONGLOMERATE
UPPER FAN (CONGLOMERATIC FACIES ASSOCIATION)
FLOW
UPPER FLAT BED
SANDSTONE LENSES WITH FLAT BEDDING AND LOW-ANGLE CROSS-BEDDING FROM ANTIDUNES OR WASHED-OUT DUNES
ANTIDUNES OR WASHED-OUT DUNES
LONGITUDINAL BARS OF C-M SUPPORTED CONGLOMERATE, IMBRICATION ABSENT OR POOR
LOWER FAN (SANDY FACIES ASSOCIATION)
RIPPLES
UPPER FLAT BED
TRANSVERSE BARS WITH SLIP FACES (UNCOMMON)
PLANAR CROSS-STRATA (UNCOMMON)
1m
1m
DUNES
LONGITUDINAL BARS WITHOUT SLIP FACES (COMMON)
TROUGH CROSS-STRATA AND SUBORDINATE FLAT BED DEPOSITED IN CHANNELS
FLAT BEDDED AND RIPPLED SANDSTONES (AND OTHER ELEMENTS OF THE HORIZONTALLY STRATIFIED FACIES) DEPOSITED ON BARS

facies" than present in the sandstone lenses of the Keweenawan conglomeratic facies association, and their "distal facies" contains more planar cross-stratification than the sandy facies association of the Keweenawan. It is not clear whether the Van Horn conglomerates are as poorly sorted as the Keweenawan conglomerates, but McGowen and Groat (1971) did note that their "massive boulder beds" display very little separation of sand and gravel fractions. These three examples are the alluvial fan systems most similar to the Keweenawan alluvial fans, and together with the Keweenawan may be considered to provide a model for sedimentation on large, gently sloping, fluvially-dominated alluvial fans. It is of note that like the Keweenawan, the two ancient examples are of Proterozoic age, and therefore formed in an environment without terrestrial plant life. Schumm (1968) has noted how the absence of plants in pre-Devonian times would have tended to create flashy rivers supplied with large amounts of sediment, particularly coarse sediment. These are conditions which encourage streams to adopt a braided morphology, and indicate that large alluvial fans built by braided stream processes might have been relatively common in pre-Devonian times.

Both the conglomerate-sandstone facies association and the sandy facies association display a weakly defined

facies alternation on a scale of about 2 - 15 m. Clear coarsening or fining upwards cycles, such as might be expected to form as the result of relatively short-term tectonic pulses, are not evident. The observed facies variation is therefore interpreted as the result of relatively long-term changes in tectonic activity, sufficiently protracted to alter the overall relief and slopes of the fan catchment.

The close intercalation in outcrops on the Keweenaw Peninsula of lacustrine sediments (with stromatolite horizons) and proximal fan sediments resulted from damming of the fan drainage by lava flows. The sharp conglomerate - siltstone contacts indicate sudden environmental changes, which are exactly as would be expected to result from volcanic damming, but which cannot be explained in terms of tectonism without proposing unreasonably large and rapid tectonic movements (as discussed in Chapter 8). Interflow lacustrine sediments at Goodharbor Bay on the Minnesota shore are interpreted as having a similar origin.

Streamflow on the fans was intermittent, giving rise to a large number of desiccated mud drapes. These conditions led to aeolian reworking of fan sediments. Aeolian sandstones are a prominent part of the section at Five Mile Point on the Keweenaw Peninsula. These display large scale

cross-stratification (with sets up to 1.5 m), lack pebbles and are better sorted than the intercalated fluvial sandstones. Palaeocurrent means for the aeolian strata diverge by 90o from fluvial palaeocurrent means. Several types of dunes could have produced the aeolian strata. On a bedding plane, traces of foresets are straight to slightly curved. In vertical section, the bases of foresets are asymptotic to set boundaries, and some large trough forms are present. The dominant type of dune was therefore probably a dune with a slightly sinuous crest line. The consistency of the aeolian palaeocurrents tends to indicate that dune slipfaces were predominantly transverse to the wind direction. The close intercalation with fluvial sediments implies that dune fields probably did not exist for a time sufficient to allow very large dunes to develop. The aeolian strata were therefore probably deposited by some variety of relatively small transverse dune with somewhat sinuous crest lines (similar to those on Icelandic glacial outwash fans described by Hine and Boothroyd, 1978). Assuming that this interpretation is correct, then the aeolian palaeocurrents indicate deposition by winds blowing along the rift valley, perhaps topographically constrained palaeo-tradewinds. Aeolian strata cannot be conclusively identified elsewhere in the Copper Harbor Formation, but

several sections display knobbly surfaces which might have originated by adhesion of wind-blown sediment to wet sand (adhesion warts of Kocurek and Fielder, 1982). At Union Bay, large scale cross-strata and critically climbing ripple translatent strata also may be of aeolian origin.

The present study points to two areas where further work would be of value. Firstly, there is a need to investigate how sedimentation occurs in powerful, high concentration streams. Smith (1986, 1987) has considered some of the features of modern and ancient sediments deposited in such conditions. Pierson and Scott (1985) documented a transition from debris flow to hyperconcentrated streamflow in a modern lahar. Beverage and Culbertson (1964) and Nordin (1963) have described some aspects of hyperconcentrated streams with sand-silt sediment loads. However, all of these studies are fairly extreme examples, and there could be useful study of streams which, even if not hyperconcentrated, are powerful and carry large amounts of poorly sorted sediment. Sorting must be taken into account. A sediment load of boulders, cobbles, granules and sand will not move in the same way as a load of sand only. Sediment transport theory must also take account of sorting. Admittedly, the task is daunting, but present theory sheds little light on how poorly sorted sediment

loads behave. For example, a potentially fruitful approach might seem to be the treatment of bedload as a grainflow sheared by the overlying fluid, as advocated by Bagnold (1956). Hanes and Bowen (1985), using this approach, have constructed a model which enables prediction of the thickness of grainflow resulting from a given shear. However, even taking maximum probable values of upper fan slope and flow depth for the Keweenawan (0.01 and 3 m respectively), use of Hanes and Bowen's equation 11a and Figure 6 predicts a grain flow layer only some 3 cm thick. This is clearly not very meaningful in a system where a large part of the sediment being transported is greater than 3 cm across. The problem probably lies in the theoretical assumption of perfect sorting. Hanes and Bowen (1985) make much use of the dimensionless Shield's stress, which is rendered dimensionless using the sediment grain size. For sediments such as those of the Keweenawan, where sorting is far from perfect, it is not apparent what is the appropriate grain size to use to make calculations involving the Shield's stress, and it seems most unlikely that it will be the mean (in fact, a sediment grain size had to be estimated to obtain the above prediction of the thickness of the grain flow layer, but whatever value is used, the calculated thickness is a few centimetres or less). Flume experiments

using coarse poorly sorted sediments and computer modelling are two possible routes which may lead to a better understanding of how poorly sorted sediments are transported.

The second area where further study would be of value is that dealt with in Chapter 8. To understand exactly how alluvial fans respond to varying rates of tectonic movement and to identify the sedimentary sequences that can result it is necessary to attempt a detailed simulation of the catchment-fan-playa system. This too is an area which might best be approached by computer modelling. Price (1972) has constructed a model of alluvial fan sedimentation, but this does not allow for the variation of uplift rates that will occur during the accumulation of a thick succession of alluvial fan sediments. There is also a need for more investigation of downfan facies changes. As briefly noted in Chapters 7 and 8, such changes are not prominent in Death Valley, yet it is generally assumed that coarsening or fining upwards sequences can be produced by the lateral migration of facies belts on alluvial fans.

REFERENCES

AHNERT, F., 1970. Functional relationships between denudation, relief and uplift in large mid-latitude drainage basins. American Journal of Science, v. 268, pp. 243-263.

AMIEUX, P.,1982. La cathodoluminescence: methode d'etude sedimentologique des carbonates. Bulletin des Centres Recherches, Exploration-Production Elf-Aquitaine, v. 6, pp. 437-483.

ANNELS, R.N., 1973. Proterozoic flood basalts of Eastern Lake Superior: the Keweenawan volcanic rocks of the Mamainse Point area, Ontario. Geological Survey of Canada, Paper 72-10, 51 pp.

_____, 1974. Keweenawan volcanic rocks of Michipicoten Island, Lake Superior, Ontario. Geological Survey of Canada, Bulletin 218, 141 pp.

BAGNOLD, R.A., 1956. The flow of cohesionless grains in fluids. Philosophical Transactions of the Royal Society, London, v. A249, pp. 235-297.

BEARD, D.C. and WEYL, P.K., 1973. Influence of texture on porosity and permeability of unconsolidated sand. American Association of Petroleum Geologists, Bulletin, v. 57, pp. 349-369.

BEATY, C.B., 1970. Age and estimated rate of accumulation of an alluvial fan, White Mountains, California, U.S.A.. American Journal of Science, v. 268, pp. 50-77.

BEVERAGE, J.P., and CULBERTSON, J.K., 1964. Hyperconcentrations of suspended sediment. Journal of the Hydraulics Division, Proceedings of the American Society of Civil Engineers, v. 90, pp. 117-128; discussion v. 91, pp. 385-388.

BJORLYKKE, K., 1983. Diagenetic reactions in sandstones. In Parker, A. and Sellwood, B.W., eds., Sediment Diagenesis, D. Reidel, pp. 169-213.

BLATT, H., MIDDLETON, G.V. and MURRAY, R., 1980. The Origin of Sedimentary Rocks (second edition). Prentice-Hall, New Jersey, 782 pp., pp. 290-291.

BLISSENBACH, E., 1954. Geology of alluvial fans in semiarid regions. Geological Society of America, Bulletin, v. 65, pp. 175-189.

BLUCK, B.J., 1965. The sedimentary history of some Triassic conglomerates in the Vale of Glamorgan, South Wales. Sedimentology, v. 4, pp. 225-245.

_____, 1974. Structural and directional properties of some sandur deposits in southern Iceland. Sedimentology, v. 21, pp. 533-554.

_____, 1978. Sedimentation in a late orogenic basin: the Old Red Sandstone of the Midland Valley of Scotland. In Bowes, D.R., and Leake, B.E., eds., Crustal Evolution in Northwest Britain and Adjacent Regions. Geological Journal Special Issue 10, pp. 249-278.

BOLES, J.R. and FRANKS, S.G., 1979. Clay diagenesis in the Wilcox Formation sandstones of southwest Texas: implications of smectite diagenesis on sandstone cementation. Journal of Sedimentary Petrology, v. 49, pp. 55-70.

BOOTHROYD, J.C., and ASHLEY, G.M., 1975. Processes, bar morphology, and sedimentary structures on braided outwash fans, northeastern Gulf of Alaska. In Jopling, A.V., and McDonald, B.C., eds., Glaciofluvial and Glaciolacustrine Sedimentation, Society of Economic Palaeontologists and Mineralogists, Special Publication 23, pp. 193-222.

_____, and NUMMEDAL, D., 1978. Proglacial braided outwash: a model for humid alluvial fan deposits. In Miall, A., ed., Fluvial Sedimentology, Canadian Society of Petroleum Geologists, Memoir 5., pp. 641-669.

BRODERICK, T.M., 1929. Zoning in Michigan copper deposits and its significance. Economic Geology, v. 24, pp. 149-162, 311-326.

BULL, W.B., 1963. Alluvial fan deposits in western Fresno County, California. Journal of Geology, v. 71, pp. 243-251.

_____, 1964. Geomorphology of segmented alluvial fans in western Fresno County, California. United States Geological Survey, Professional Paper 352-E, pp. 89-129.

_____, 1972. Recognition of alluvial-fan deposits in the stratigraphic record. In Hamblin, W.K., and RIGBY, J.K., eds., Recognition of ancient sedimentary environments, Society of Economic Palaeontologists and Mineralogists, Special Publication 16, pp. 63-83.

_____, 1977. The alluvial-fan environment. Progress in Physical Geography, v.1, pp. 221-269.

BURGGRAF, D.R., Jr., and VONDRA, C.F., 1982. Rift valley facies and palaeoenvironments: and example from the East African Rift System of Kenya and southern Ethiopia. Zeitschrift fur Geomorphologie, v. 42, pp. 43-73.

BUTLER, B.S., and BURBANK, W.S., 1929. The copper deposits of Michigan. United States Geological Survey, Professional Paper 144. 238 pp.

CAMPBELL, C.V., 1976. Reservoir geometry of a fluvial sheet sandstone. American Association of Petroleum Geologists, Bulletin, v. 60, pp. 1009-1020.

CHASE, C.G. and GILMER, T.H., 1973. Precambrian plate tectonics: the midcontinent gravity high. Earth and Planetary Science Letters, v. 21, pp. 70-78.

CHERVEN, V.B., 1984. Early Pleistocene glacial outwash deposits in the eastern San Joaquin Valley, California: a model for humid-region alluvial fans. Sedimentology, v. 31, pp. 823-836.

CORNWALL, H.R., 1954a. Bedrock geology of the Lake Medora Quadrangle, Michigan. United States Geological Survey, Geological Quadrangle Map GQ52.

_____, 1954b. Bedrock geology of the Delaware Quadrangle, Michigan. United States Geological Survey, Geological Quadrangle Map GQ51.

_____, 1955. Bedrock geology of the Fort Wilkins Quadrangle, Michigan. United States Geological Survey, Geological Quadrangle Map GQ74.

_____, and WHITE, W.S., 1955. Bedrock geology of the Manitou Island Quadrangle. United States Geological Survey, Geological Quadrangle Map GQ73.

COTTER, E., 1978. The evolution of fluvial style, with special reference to the Central Appalachian Paleozoic. In Miall, A.D., ed., Fluvial Sedimentology, Canadian Society of Petroleum Geologists, Memoir 5, pp. 361-383.

CULBERTSON, J.K., and DAWDY, D.R., 1964. A study of fluvial characteristics and hydraulic variables, Middle Rio Grande, New Mexico. United States Geological Survey Water-Supply Paper 1498-F, 74 pp.

DANIELS, P.A., Jr., 1982. Upper Precambrian sedimentary rocks: Oronto Group, Michigan-Wisconsin. In Wold, R.J. and Hinze, W.J., eds., Geology and Tectonics of the Lake Superior Basin, Geological Society of America, Memoir 156, pp. 107-133.

DAVIDSON, D.M., Jr., 1982. Geological evidence relating to interpretation of the Lake Superior Basin structure. In Wold, R.J. and Hinze, W.J., eds., Geology and Tectonics of the Lake Superior Basin, Geological Society of America, Memoir 156, pp. 5-14.

DENNY, C.S., 1967. Fans and pediments. American Journal of Science, v. 265, pp. 81-105.

DUBOIS, P.M., 1962. Palaeomagnetism and correlation of Keweenawan rocks. Geological Survey of Canada, Bulletin 71, 75 pp. (particularly pp. 57, 27-30)

ELMORE, R.D., 1981. The Copper Harbor Conglomerate and Nonesuch Shale, sedimentation in a Precambrian intracontinental rift, Upper Michigan. Ph.D. thesis, University of Michigan, 192 pp.

_____, 1984. The Copper Harbor Conglomerate: a late Precambrian fining-upward alluvial fan sequence in northern Michigan. Geological Society of America, Bulletin, v. 95, pp. 610-617.

FOLK, R.L., 1968. Bimodal supermature sandstones: product of the desert floor. Genesis and Classification of Sedimentary Rocks, Proceedings of Section 8, 23rd International Geological Congress, pp. 9-32.

_____, 1974. Petrology of Sedimentary Rocks. Hemphill's Book Store, Austin, Texas. pp. 42, 127.

FORBES, D.L., and BOYD, R., 1987. Gravel ripples on the inner Scotian shelf. Journal of Sedimentary Petrology, v. 57, pp. 46-54.

GALLOWAY, W.E., and HOBDAY, D.K., 1983. Terrigenous Clastic Depositional Systems. Springer-Verlag, New York, pp. 423. Particularly Chapter 3: alluvial fan systems, pp. 25-50.

GIGUERE, J.F., 1975. Geology of Saint Ignace Island and adjacent islands, district of Thunder Bay. Ontario Division of Mines, Geological Report 118, 35 pp.

GLOPPEN, T.G., and STEEL, R.J., 1981. The deposits, internal structure and geometry in six alluvial fan - fan delta bodies (Devonian - Norway) - a study in the significance of bedding sequence in conglomerates. In Ethridge, F.G., and Flores, R.M., eds., Recent and Ancient Nonmarine Depositional Environments: Models For Exploration, Society of Economic Palaeontologists and Mineralogists, Special Publication 31, pp. 49-69.

GOLE, C.V., and CHITALE, S.V., 1966. Inland delta building activity of Kosi River. Journal Hydraulics Division, Proceedings of the American Society of Civil Engineers, v. 92, pp. 111-126.

GREEN, J.C., 1972. North Shore Volcanic Group. In Sims, P.K. and Morey, G.B., eds., Geology of Minnesota: A Centennial Volume. Minnesota Geological Survey, St. Paul, pp. 294-332.

_____, 1977. Keweenawan plateau volcanism in the Lake Superior region. In Baragar, W.R.A., Coleman, L.C. and Hall, J.M., eds., Volcanic Regimes in Canada, Geological Association of Canada, Special Paper 16, pp. 407-422.

_____, 1982. Geology of Keweenawan extrusive rocks. In Wold, R.J. and Hinze, W.J., eds., Geology and Tectonics of the Lake Superior Basin, Geological Society of America, Memoir 156, pp. 47-55.

GUSTAVSON, T.C., 1974. Sedimentation on gravel outwash fans, Malaspina Glacier foreland, Alaska. Journal of Sedimentary Petrology, v. 44, pp. 374-389.

HALLS, H.C., 1966. A review of the Keweenawan geology of the Lake Superior region. In Steinhart, J.S. and Smith, T.J., eds., The Earth Beneath the Continents. American Geophysical Union, pp. 3-27.

_____, 1978. The late Precambrian central North American rift system - a survey of recent geological and geophysical investigations. In Ramberg, I.B. and Neumann, E.R., eds., Tectonics and Geophysics of Continental Rifts: N.A.T.O. Advanced Study Series, C, 37. D. Reidel, Boston, pp. 111-123.

_____, and PALMER, H.C., 1981. Remagnetization in Keweenawan rocks. Part II: lava flows within the Copper Harbor Conglomerate, Michigan. Canadian Journal of Earth Sciences, v.18, pp. 1395-1408.

_____, and PESONEN, L.J., 1982. Palaeomagnetism of Keweenawan rocks. In Wold, R.J. and Hinze, W.J., eds., Geology and Tectonics of the Lake Superior Basin, Geological Society of America, Memoir 156, pp. 173-203. (particularly appendix 1).

HAMBLIN, W.K., 1958. Cambrian sandstones of northern Michigan. Michigan Department of Conservation, Geological Survey Division, Publication 51. 149 pp., pp. 46-49 particularly.

_____, 1961. Paleogeographic evolution of the Lake Superior region from Late Keweenawan to Late Cambrian time. Geological Society of America, Bulletin, v. 72 pp. 1-18.

_____, and HORNER, W.J., 1961. Sources of the Keweenawan conglomerates of northern Michigan. Journal of Geology, v. 69, pp. 204-211.

HANES, D.M., and BOWEN, A.J., 1985. A granular-fluid model for steady intense bed-load transport. Journal of Geophysical Research, v. 90, pp. 9149-9158.

HARMS, J.C., and FAHNESTOCK, R.K., 1965. Stratification, bed forms, and flow phenomena (with an example from the Rio Grande). Society of Economic Palaeontologists and Mineralogists, Special Publication 12, pp. 84-115.

HEWARD, A.P., 1978. Alluvial fan sequence and megasequence models: with examples from Westphalian D - Stephanian B coalfields, Northern Spain. In Miall, A.D., ed., Fluvial Sedimentology, Canadian Society of Petroleum Geologists, Memoir 5, pp. 669-702.

HEIN, F.J., and WALKER, R.G., 1977. Bar evolution and development of stratification in the gravelly, braided, Kicking Horse River, British Columbia. Canadian Journal of Earth Sciences, v. 14, pp. 562-570.

HINE, A.C., and BOOTHROYD, J.C., 1978. Morphology, processes, and recent sedimentary history of a glacial-outwash plain shoreline, Southern Iceland. Journal of Sedimentary Petrology, v. 48, pp. 901-920.

HINZE, W.J., WOLD, R.J., and O'HARA, N.W., 1982. Gravity and magnetic anomaly studies of Lake Superior. In Wold, R.J. and Hinze, W.J., eds., Geology and Tectonics of the Lake Superior Basin, Geological Society of America, Memoir 156, pp. 203-221.

HOOKE, R.LeB., 1967. Processes on arid-region alluvial fans. Journal of Geology, v.75, pp. 438-460.

_____, 1968. Steady-state relationships on arid-region alluvial fans in closed basins. American Journal of Science, v. 266, pp. 609-629.

_____, 1972. Geomorphic evidence for Late-Wisconsin and Holocene tectonic deformation, Death Valley, California. Geological Society of America, Bulletin, v. 83, pp. 2073-2098.

_____, and ROHRER, W.L., 1979. Geometry of alluvial fans: effect of discharge and sediment size. Earth Surface Processes, v. 4, pp. 147-166.

HUBBARD, H.A., 1972. Source of pebbles of volcanic rocks in the Middle and Upper Keweenawan conglomerates of northern Michigan. Journal of Geology, v. 80, pp. 627-629.

_____, 1975. Geology of Porcupine Mountains in Carp River and White Pine quadrangles, Michigan. Journal of Research, United States Geological Survey, v. 3, pp. 519-528.

HUNTER, R.E., 1977. Basic types of stratification in small eolian dunes. Sedimentology, v. 24, pp. 361-387.

IBBEKEN, H., 1983. Jointed source rock and fluvial gravels controlled by Rosin's law: a grain-size study in Calabria, South Italy. Journal of Sedimentary Petrology, v. 53, pp. 1213-1231.

JIRSA, M.A., 1980. The petrology and tectonic significance of interflow sediments in the Keweenawan North Shore Volcanic Group of Northeastern Minnesota. M.S. Thesis, University of Minnesota, 125 pp.

JOHNSON, A.M., 1970. Physical Processes in Geology. Freeman, Cooper and Co., 577 pp., particularly Chapters 12, 13 and 14.

JOHNSON, N.L., and KOTZ, S.K., 1970. Continuous Univariate Distributions, Volume 1. Houghton Mifflin, Boston, 300 pp., particularly pp. 250-271.

JOHNSON, R.L., 1973. The petrology, pebble morphology and origin of the Outer Conglomerate (Upper Member), Copper Harbor Conglomerate, Keweenaw Peninsula, Michigan. M.S. Thesis, Michigan Technological University, Houghton, Michigan.

KALLIOKOSKI, J., 1986. Calcium carbonate cement (caliche) in Keweenawan sedimentary rocks (1.1 Ga), Upper Peninsula of Michigan. Precambrian Research, v. 32, pp. 243-259.

_____ and WELCH, E.J., 1985. Keweenawan-age caliche palaeosol in the lower part of the Calumet and Hecla Conglomerate, Centennial Mine, Calumet, Michigan. Geological Society of America, Bulletin, v. 96, pp. 1188-1193.

KITTLEMAN, L.R., 1964. Application of Rosin's distribution in size-frequency analysis of clastic rocks. Journal of Sedimentary Petrology, v. 34, pp. 483-502.

KLEIN, I.,1939. Microcline in the native copper deposits of Michigan. American Mineralogist, v. 24, pp. 643-650.

KOCHEL, R.C., and JOHNSON, R.A., 1984. Geomorphology and sedimentology of humid-temperate alluvial fans, central Viginia. In Koster, E.H., and Steel, R.J., eds., Sedimentology of Gravels and Conglomerates. Canadian Society of Petroleum Geologists, Memoir 10, pp. 109-122.

KOCUREK, G. and FIELDER, G., 1982. Adhesion structures. Journal of Sedimentary Petrology, v. 52, pp. 1229-1241.

LARSON, V., and STEEL, R.J., 1978. The sedimentary history of a debris-flow dominated, Devonian alluvial fan - a study of textural inversion. Sedimentology, v. 25, pp. 37-59.

LARSON, R.J., 1969. The Mott Island Conglomerate, Isle Royale National Park, Michigan. M.Sc. Thesis, Michigan Technological University, Houghton, Michigan.

MACK, G.H., and RASMUSSEN, K.A., 1984. Alluvial fan sedimentation of the Cutler Formation (Permo-Pennsylvanian) near Gateway, Colorado. Geological Society of America, Bulletin, v. 95, pp. 109-116.

McGOWEN, J.H., and GROAT, C.G., 1971. Van Horn Sandstone, West Texas: an alluvial fan model for mineral exploration. Report of Investigations No. 72, Bureau of Economic Geology, The University of Texas at Austin, 57 pp.

MERK, G.P., 1972. Provenance and tectonic inferences concerning the Keweenawan interflow sediments of the Lake Superior region. Ph.D. Thesis, Michigan State University, 126 pp.

_____, and JIRSA, M.A., 1982. Provenance and tectonic significance of the Keweenawan interflow sedimentary rocks. In Wold, R.J. and Hinze, W.J., eds., Geology and Tectonics of the Lake Superior Basin, Geological Society of America, Memoir 156, pp. 97-105.

MIALL, A.D., 1977. A review of the braided-river depositional environment. Earth Science Reviews, v. 13, pp. 1-62.

_____, 1978. Lithofacies and vertical profile models in braided river deposits: a summary. In Miall, A.D., ed., Fluvial Sedimentology, Canadian Society of Petroleum Geologists, Memoir 5, pp. 597-604.

MIDDLETON, G.V., 1962. Size and sphericity of quartz grains in two turbidite formations. Journal of Sedimentary Petrology, v. 32, pp. 725-742.

MOREY, G.B., 1974. Cyclic sedimentation of the Solar Church Formation (Upper Precambrian; Keweenawan), southeastern Minnesota. Journal of Sedimentary Petrology, v. 44, pp. 872-884.

_____, and OJAKANGAS, R.W., 1982. Keweenawan sedimentary rocks of eastern Minnesota and northwestern Wisconsin. In Wold, R.J. and Hinze, W.J., eds., Geology and Tectonics of the Lake Superior Basin, Geological Society of America, Memoir 156, pp. 135-146.

MORISON, S.R., and HEIN, F.J., 1987. Sedimentology of the White Channel Gravels, Klondike area, Yukon Territory: fluvial deposits of a confined valley. In, Ethridge, F.G., Flores, R.M., and Harvey, M.D., eds., Recent Developments in Fluvial Sedimentology, Society of Economic Palaeontologists and Mineralogists, Special Publication 39, pp. 205-216.

NADON, G.C., 1981. The stratigraphy and sedimentology of the Triassic at St. Martins and Lepreau, New Brunswick. M.Sc. Thesis, McMaster University, Hamilton, Ontario, 279 pp.

_____, and MIDDLETON, G.V., 1985. The stratigraphy and sedimentology of the Fundy Group (Triassic) of the St. Martins area, New Brunswick. Canadian Journal of Earth Science, v. 22, pp. 1183-1203.

NICKLE, E., 1978. The present status of cathode luminescence as a tool in sedimentology. Minerals Science Engineering, v. 10, pp. 73-100.

NILSEN, T.H., 1982. Alluvial fan deposits. In Scholle, P.A., and Spearing, D., eds., Sandstone Depositional Environments. American Association of Petroleum Geologists, Memoir 31, pp. 49-86.

NORDIN, C.F., 1963. A preliminary study of sediment transport parameters, Rio Puerco near Bernardo, New Mexico. United States Geological Survey, Professional Paper 462-C, 21 pp.

OJAKANGAS, R.W., and MOREY, G.B., 1982. Keweenawan sedimentary rocks of the Lake Superior region: a summary. In Wold, R.J. and Hinze, W.J., eds., Geology and Tectonics of the Lake Superior Basin, Geological Society of America, Memoir 156, pp. 157-163.

ORI, G.G., 1982. Braided to meandering channel patterns in humid-region alluvial fan deposits, River Reno, Po Plain, (northern Italy). Sedimentary Geology, v. 31, pp. 231-248.

PARKASH, B., SHARMA, R.P., and ROY, A.K., 1980. The Siwalik Group (molasse) - sediments shed by collision of continental plates. Sedimentary Geology, v. 25, pp. 127-159.

PIERSON, T.C., 1981. Dominant particle support mechanisms in debris flows at Mount Thomas, New Zealand, and implications for flow mobility. Sedimentology, v. 28, pp. 49-60.

_____, and SCOTT, K.M., 1985. Downstream dilution of a lahar: transition from debris flow to hyperconcentrated streamflow. Water Resources Research, v. 21, pp. 1511-1524.

PRICE, W.E., Jr., 1972. A random walk simulation model of alluvial fan deposition. Technical Report 7, Technical Reports on Hydrology and Water Resources, University of Arizona, Tucson, 225 pp.

RUST, B.R., 1978. Depositional models for braided alluvium. In Miall, A.D., ed., Fluvial Sedimentology, Canadian Society of Petroleum Geologists, Memoir 5, pp. 605-625.

_____, and KOSTER, E.H., 1984. Coarse alluvial deposits. In Walker, R.G., ed., Facies Models (second edition), Geoscience Canada Reprint Series 1., pp. 53-69.

RYDER, J.M., 1971. The stratigraphy and morphology of para-glacial alluvial fans in south-central British Columbia. Canadian Journal of Earth Science, v. 8, pp. 279-298.

SANDBERG, A.E., 1938. Section across Keweenawan lavas at Duluth, Minnesota. Bulletin of the Geological Society of America, v. 49, pp. 795-830.

SCHUMM, S.A., 1968. Speculations concerning paleohydrologic controls of terrestrial sedimentation. Geological Society of America, Bulletin, v.79, pp. 1573-1588.

_____, 1977. The Fluvial System. John Wiley and Sons, New York, 338 pp., particularly pp. 21-35.

SIMONS, D.B., RICHARDSON, E.V. and ALBERTSON, M.L., 1961. Flume studies using medium sand (0.45mm). United States Geological Survey, Water-Supply Paper 1498-A, 76 pp.

SMITH, G.A., 1986. Coarse-grained volcaniclastic sediment: terminology and depositional process. Geological Society of America, Bulletin, v. 97, pp. 1-10.

_____, 1987. The influence of explosive volcanism on fluvial sedimentation: the Deschutes Formation (Neogene) in Central Oregon. Journal of Sedimentary Petrology, v. 57, pp. 613-629.

STEEL, R.J., 1974. New Red Sandstone floodplain and piedmont sedimentation in the Hebridean Province, Scotland. Journal of Sedimentary Petrology, v. 44, pp. 336-357.

_____, and AASHEIM, S.M., 1978. Alluvial sand deposition in a rapidly subsiding basin (Devonian, Norway). In Miall, A.D., ed., Fluvial Sedimentology, Canadian Society of Petroleum Geologists, Memoir 5, pp. 385-412.

_____, and WILSON, A.C., 1975. Sedimentation and tectonism (Permo-Triassic) on the margin of the North Minch Basin, Lewis. Journal of the Geological Society, London., v. 131, pp. 183-202.

_____, and THOMPSON, D.B., 1983. Structures and textures in Triassic braided stream conglomerates ("Bunter" Pebble Beds) in the Sherwood Sandstone Group, North Staffordshire, England. Sedimentology, v. 30, pp. 341-367.

STOIBER, R.E., and DAVIDSON, E.S., 1959. Amygdule mineral zoning in the Portage Lake Lava Series, Michigan copper district. Economic Geology, v. 54, pp. 1250-1277, 1444-1460.

TANTON, T.L., 1931. Fort William and Port Arthur, and Thunder Cape map-areas, Thunder Bay district, Ontario. Canada Department of Mines, Geological Survey, Memoir 167, 222 pp. (pp. 57-65 particularly).

VAN SCHMUS, W.R., GREEN, J.C. and HALLS, H.C., 1982. Geochronology of Keweenawan rocks of the Lake Superior region: a summary. In Wold, R.J. and Hinze, W.J., eds., Geology and Tectonics of the Lake Superior Basin, Geological Society of America, Memoir 156, pp. 165-171.

WALKER, R.G. and CANT, D.J., 1984. Sandy fluvial systems. In Walker, R.G., ed., Facies Models (second edition), Geoscience Canada Reprint Series 1., pp. 71-89.

WALKER, T.R., LARSON, E.E., and HOBLITT, R.P., 1981. Nature and origin of haematite in the Moenkopi Formation (Triassic), Colorado Plateau: a contribution to the origin of magnetism in red beds. Journal of Geophysical Research, v. 86, pp. 317-333.

WALKER, T.R., 1976. Diagenetic origin of continental red beds. In Falke, H., ed., The Continental Permian in Central, Western and Southern Europe, Proceedings of the N.A.T.O. Advanced Study Institute, Mainz, Germany, pp. 240-282.

WASSON, R.J., 1977a. Catchment processses and the evolution of alluvial fans in the lower Derwent Valley, Tasmania. Zeitschrift fur Geomorphologie, v. 21, pp. 147-168.

_____, 1977b. Last-glacial alluvial fan sedimentation in the Lower Derwent Valley, Tasmania. Sedimentology, v. 24, pp. 781-799.

WELLS, N.A., 1984. Sheet debris flow and sheetflood conglomerates in Cretaceous cool-maritime alluvial fans, South Orkney Islands, Antarctica. In Koster, E.H., and Steel, R.J., eds., Sedimentology of Gravels and Conglomerates. Canadian Society of Petroleum Geologists, Memoir 10, pp. 133-145.

_____, and DORR, J.A., Jr., 1987. A reconnaisance of sedimentation on the Kosi alluvial fan of India. In, Ethridge, F.G., Flores, R.M., and Harvey, M.D., eds., Recent Developments in Fluvial Sedimentology, Society of Economic Palaeontologists and Mineralogists, Special Publication 39, pp. 51-61.

WELTON, J.E., 1984. S.E.M. Petrology Atlas. Methods in Exploration Series, American Association of Petroleum Geologists, 237 pp.

WHITE, W.S., 1952. Imbrication and initial dip in a Keweenawan conglomerate bed. Journal of Sedimentary Petrology, v. 22, pp. 189-199.

_____, 1960. The Keweenawan lavas of Lake Superior, an example of flood basalts. American Journal of Science, Bradley Volume, v. 258-A, pp. 367-374.

_____, 1966a. Geologic evidence for crustal structure in the western Lake Superior basin. In Steinhart, J.S., and Smith, T.J., eds., The Earth Beneath the Continents: American Geophysical Union, Geophysical Monograph 10, pp. 28-41.

_____, 1966b. Tectonics of the Keweenawan basin, western Lake Superior region. United States Geological Survey, Professional Paper 524-E, 23 pp.

_____, 1968. The native copper deposits of northern Michigan. In Ridge, J.D., ed., Ore Deposits of the United States, 1933-1967. American Institute of Mining, Metallurgy and Petroleum Engineers, v. 1, pp. 305-325.

_____, and WRIGHT, J.C., 1960. Lithofacies of the Copper Harbor Conglomerate, northern Michigan. United States Geological Survey, Professional Paper 400-B, pp. 5-8.

WILLIAMS, G.E., 1969. Characteristics and origin of a Precambrian pediment. Journal of Geology, v. 77, pp. 183-207.

WILLIAMS, P.F., and RUST, B.R., 1969. The sedimentology of a braided river. Journal of Sedimentary Petrology, v. 39, pp. 649-679.

WILSON, M.D. and PITTMAN, E.D., 1977. Authigenic clays in sandstones: recognition and influence on reservoir properties and palaeoenvironmental analysis. Journal of Sedimentary Petrology, v. 47, pp. 3-31.

WOLFF, R.G., and HUBER, N.K., 1973. The Copper Harbor Conglomerate (Middle Keweenawan) on Isle Royale, Michigan, and its regional implications. United States Geological Survey Professional Paper 754-B, pp. 1-15.